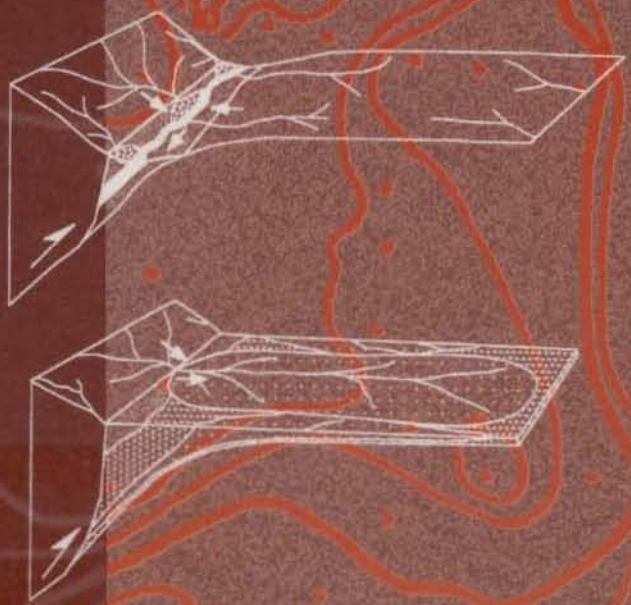
A. D. Miall

Principles of Sedimentary Basin Analysis



3rd, updated and enlarged Edition

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For
Charlene

Preface

Basin analysis is a process that requires the collection and integration of a wide variety of different types of data. Many review articles and books are available that deal with some of the individual specialized fields that together constitute basin analysis, such as stratigraphic methods, facies analysis, geophysical methods, sedimentation and tectonics, and sequence stratigraphy; but this is one of the few books that has attempted, now through three editions, to show how these different types of data and the varying types of analytical methodology used in basin analysis need to be brought together to tackle specific basin-analysis problems. The emphasis in this book continues to be on the traditional geological skills of observation, description and synthesis.

Some of the major components of basin analysis have stabilized over the last decade. Facies analysis and the definition of process-response models (Chap. 4) constitute an area that has reached a level of maturity and stability to the extent that only minor refinements are now to be expected, although many more case studies exploring local variability will continue to be generated. The methods of lithostratigraphy and biostratigraphy (Chap. 3) have also changed little in recent years, whereas the widespread acceptance of sequence-stratigraphic principles and the introduction of allostratigraphic mapping methods (Chap. 6) have led to fundamental changes in the nature of regional basin analysis. These developments, in turn, have lent support to the continuing complex, interdisciplinary effort to improve the accuracy and precision of the geological time scale and its practical applicability to local and regional problems of dating and correlation (Chap. 3). Many questions about the extent and origins of stratigraphic sequences require precise chronostratigraphic correlation, and much remains to be done in this area. The refinements in data collection brought about by the increasing resolution of the reflection-seismic method (Chaps. 5, 6) continue to demonstrate a complexity in the preserved stratigraphic record that challenges our powers of description and interpretation.

Methods for numerical and graphical modeling of sedimentary basins and basin fills continue to provide exciting new insights, such as the new links emerging between climate, sediment supply and basin subsidence rates, but the causes of crustal subsidence and uplift and the relationship of these motions to mantle processes and plate tectonics seem to have been largely worked out.

The most significant changes in basin analysis methods since publication of the second edition of this book are in the area of sequence stratigraphy (Chap. 6). The definitions of systems tracts had only just appeared when the last edition of this book was being completed, and the application of sequence models to an analysis of the evolving paleogeography of a basin was a subject in its infancy. In addition, we now have a much more complete base of case studies from which to assess the many and varied mechanisms that generate stratigraphic sequences, so that the rather simplistic ideas about global eustasy that were popular a decade ago can now be re-evaluated. All this has required complete rewriting of Chapters 6 and 8 of this book, whereas the remaining chapters have been substantially rewritten while retaining their original form, and have been updated with new and useful case studies. Developments in such fields as dynamic topography and basin inversion have been added to Chapter 7, and improvements in our understanding of foreland basins and in the tectonic processes of the Precambrian constitute some of the more substantial changes to Chapter 9.

Much work remains to be done in the area of basin analysis. The search for non-renewable resources will continue into the foreseeable future, with many frontier basins remaining to be explored, while the task of extracting additional petroleum reserves from the heterogeneities in mature fields provides basin analysts with some of their most complex challenges. The development of water resources, the search for sediment-hosted minerals, and the need to integrate basin histories into broad-based studies of regional, national and continental geological histories will continue to create a demand for basin analysis and the skills with which to perform it. The need for this new edition of the book is therefore clear, and it is to be hoped that it will continue to provide a useful reference manual and resource of ideas for advanced students and professionals.

Acknowledgments. The writing of the first edition of this book took about 3 years, but the ideas and experience on which it was based had been accumulating since I first discovered geology at the age of 15 years, during an optional one-period-a-week course at Brighton Grammar School. I am deeply indebted to the late Arnold Berry, geography master at the school, for his lectures and field trips, which rapidly convinced me I had found my vocation. I received every encouragement during my undergraduate years at the University of London (Queen Mary and Westfield College), where instruction in stratigraphy and historical geology was given by Professor Kirkaldy.

My introduction to modern sedimentological principles came through the stimulating graduate seminars held by Brian Rust at the University of Ottawa. Later, the opportunity to gain extensive practical experience was made possible by Don Stott, who took me on staff at the Geological Survey of Canada. Writing of this book began shortly after I moved to the fertile academic climate of the University of Toronto, an appointment brought about by the efforts of Frank Beales, Geoff Norris, and David Strangway.

Many colleagues helped me formulate my ideas on basin analysis during the 1970s and 1980s, and I am grateful to them for their stimulating company over the years. Foremost among these are Don Campbell and John Stuart-Smith at J.C. Sproule and Associates and my former Survey colleagues Jim Dixon, Ashton Embry, Ulrich Mayr, Ross McLean, Ray Thorsteinsson, and Hans Trettin. More recently, my understanding of sequence stratigraphy has been aided by contact with individuals such as Phil Allen, Makoto Ito, Dale Leckie, Dag Nummedal, Guy Plint, and Roger Walker.

For specific advice relating to specialized parts of this book, I am indebted to Norm Evensen (radiometric dating), Gary Jarvis (geophysical basin models and plate processes), George Klein (backarc basins and shelf depositional systems), and Geoff Norris (stratigraphic methods).

Photographs were contributed by Jim Dixon, Ashton Embry, Rolf Ludvigsen, Brian Pratt, A. Easton Wren, and the Royal Ontario Museum. My particular thanks to David Rudkin at the museum for his assistance with the photographs of fossils in Chapter 2. Neil Ollerenshaw and the staff at GSC, Calgary, are thanked for assistance in obtaining two illustrations for the sections on the dipmeter. About one-third of the figures in the book were drafted by Subash Shanbhag and the remainder in the first edition were reproduced from other sources by photographer Brian O'Donovan. Their efforts and their patience during the long gestation of the manuscript are much appreciated.

During preparation of the first edition, Geoff Orton and Winston Mottley worked as research assistants for short but crucial periods. Typing and word processing were carried out by Diane Gardner and Naomi Frankel, with the final text prepared at the University of Toronto Secretarial Services unit under the supervision of Vera Baker. The text of the second and third editions were prepared by the author, thanks to the magic of Windows '95 and Microsoft Word.

Andy Baillie and Tony Tankard undertook the task of critically reading the entire manuscript of the first edition. Their comments were most useful and are much appreciated. In addition, the author is grateful to Roger Walker (who read parts of

Chap. 4), Geoff Norris (Chap. 3), and Ray Ingersoll (Chap. 9), who provided some much-needed specialized commentary.

Comments from students and colleagues were of considerable use in the preparation of the second edition of the book, as were published reviews of the first edition, which made several valuable points. A one-semester administrative leave from teaching and other duties in the fall of 1988 enabled me to complete the initial draft, and I am grateful to the University of Toronto for making this time available to me. The entire manuscript was then critically read by Ray Ingersoll. I am honored by his willingness to undertake this considerable task, and I wish to express my considerable gratitude for the conscientious and thoughtful way in which he completed it. While we do not agree on all of the finer points of basin analysis, his comments have been invaluable in helping me to refine the final draft. However, the content of the book, including any errors or omissions, remains my responsibility.

Chapters 6 and 8 of the third edition were critically read by Guy Plint and, once again, Ray Ingersoll provided critical reading of the material dealing with sedimentation and tectonics – Chapters 7 and 9. The assistance of these colleagues is much appreciated.

Finally, I must thank my wife Charlene, whose assistance, advice, and support throughout my career have been irreplaceable. I am particularly grateful for her love and patience during my long obsession with "the book."

Toronto, July 1999 Andrew D. Miall

Preface to the First Edition

This book is intended as a practical handbook for those engaged in the task of analyzing the paleogeographic of ancient sedimentary basins. The science of stratigraphy and sedimentology is central to such endeavours, but although several excellent textbooks on sedimentology have appeared in recent years, little has been written about modern stratigraphic methods. Sedimentology textbooks tend to take a theoretical approach, building from physical and chemical theory and studies of modern environments. It is commonly difficult to apply this information to practical problems in ancient rocks, and very little guidance is given on methods of observation, mapping, and interpretation.

In this book, theory is downplayed and the emphasis is on what a geologist can actually see in outcrops, well records, and cores and what can be obtained using geophysical techniques. A new approach is taken to stratigraphy, which attempts to explain the genesis of lithostratigraphic units and to de-emphasize the importance of formal description and naming. There are also sections explaining principles of facies analysis, basin mapping methods, depositional systems, and the study of basin subsidence history and thermal history, so important to the genesis of fuels and minerals. Last, an attempt is made to tie everything together by considering basins in the context of plate tectonics and eustatic sea-level changes.The vast and important subjects of paleoclimatology, diagenesis, and sedimentary geochemistry are not dealt with here, except for some aspects of diagenetic change that can be related to burial history. None of these areas is crucial to the main theme of the book, although geochemical data are becoming increasingly important in the study of chemical sediments and their depositional environments. The economoc applications of basin analysis (e.g., exploring sediment-mineral associations and types of stratigraphic hydrocarbon traps) are not discussed in detail, although some implications for exploration are pointed out at several places in the book.

It is hoped that this book will be useful to those engaged in exploration for the various nonrenewable fuel and mineral resources and to students and other geologists carrying out local or regional basin analysis studies, beginning with undergraduates about to leave for their first field school. The book is divided into two parts. Part 1 comprises Chapters 1 to 6; it deals with the stratigraphy and sedimentology of basins and is written mainly at an undergraduate level. Part 2 consists of Chapters 7 to 10 and deals with the global controls of basin development, including plate tectonics. This part of the book is addressed to senor undergraduates, graduate students, and professional geologists; a working knowledge of the principles of plate tectonics is assumed for this final part of the book. It is intended that the book be used throughout an individual's training and professional employment.

Andrew D. Miall

Preface to the Second Edition

Five years between the editions of this book might not seem like a very long time, but advances in some areas of basin analysis have been very rapid. In order for this book to remain a useful textbook, it became clear very soon after the initial publication in 1984 that revisions would have to be undertaken. Accordingly, I began to keep files of critical references – these now number more than one thousand – and rewritting was started in the fall of 1987.

The new developments that have occurred, or at least have become more important since 1984, include the publication of new stratigraphic codes and the widespread application of the stratotype concept to chronostratigraphic subdivisions, new research into the significance of stratigraphyic completeness, unconformity-bounded sequences and sequence stratigraphy, fission-track dating as applied to stratigraphic problems, developments in our understanding of facies architecture, including the erection of hierarchis of bounding surfaces for many types of deposits, 3-D seismic methods, Milankovitch cycles, evolution in our thinking regarding the importance of tectonic versus eustatic controls of regional and global cycles, evolution of ideas regarding the interpretations about sea-level change that can be made from seismic stratigraphic records, and many refinements of geophysically based basin-subsidence models.

These developments have required a virtually complete rewriting of Chapters 7 and 8 and substantial additions and minor modifications to most of the remaining chapters. Although it is hoped that this has recreated a fairly up-to-date book, continuing developments are inevitable, and the files of material for the third edition have already been started.

Andrew D. Miall

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Part 1
The Stratigraphy and Sedimentology of Sedimentary Basins: Introduction to Part 1

Part 1 of this book deals with the techniques required to describe the sedimentary fill of a basin. We first examine briefly the evolution of modern stratigraphic methods and the kinds of basin-analysis projects a geology is likely to encounter (Chap. 1). This chapter concludes with some general suggestions for a research methodology and guidelines for the reporting of results. We then describe and illustrate the nature of the stratigraphic-sedimentologic data base that can be built up from outcrop and well (sample, core and petrophysical log) evidence (Chap. 2).

A major part of a basin-analysis exercise must be to compare and correlate evidence from different parts of a basin, in order to build up a picture of the overall basin architecture. Traditional methods for carrying out this process include lithostratigraphy and biostratigraphy. To these must be added magnetostratigraphy and chemostratigraphy, which aid in correlation and which have contributed, with radiometric dating and biostratigraphy, to the construction of a global geological time scale (Chap. 3).

In Chapter 4, we turn to the core subject in the analysis of paleogeographic evolution of a basin: the analysis of the sediments for the evidence they contain about sedimentary environments and depositional processes. This is called facies analysis. Chapter 5 describes how we can extend the analysis from a few outcrops or drill holes to the analysis of an entire basin, using various mapping techniques. The analysis of reflection-seismic data is introduced in this chapter.

Analysis of reflection-seismic data led to a major revolution in the documentation and interpretation of the large-scale architecture of sedimentary basins in the 1970s, with the development of a new methodology for description and interpretation termed sequence stratigraphy. Part 1 concludes with a description of these new methods (Chap. 6). The analysis of stratigraphic sequences has, in turn, opened up new lines of research into regional and global sedimentary controls. These are the subject of Part 2 of the book.

Introduction

1.1 Scope and Purpose of Book

Most of the world's nonrenewable fuel resources and many of its metals and minerals are derived from sedimentary rocks ranging in age from Archean to Cenozoic. Exploration for such resources requires an understanding of their relationship to host strata, whether they be primary deposits formed at the same time as the sediments, such as coal and placer minerals, or postdepositional deposits, whose distribution depends on porosity and permeability trends in the sediments, such as epigenetic ore deposits and petroleum.

The study of such host strata is known by the convenient term **basin analysis**, and its most important product is the documentation of the **subsidence history, stratigraphic architecture**, and **paleogeographic evolution** of a sedimentary basin. The work may include many components, among which the most important are stratigraphy, structure, and sedimentology. The rocks may be examined in outcrops or in wells, or the data may be acquired by geophysical methods. Most basins nowadays are described and interpreted in the context of a framework of sequence stratigraphy.

Basin analysis requires an understanding of many diverse geological specialties and an ability to assess the relationships between varied types of evidence, including lithostratigraphy, biostratigraphy, facies analysis, sequence stratigraphy, stable-isotope geochemistry, organic geochemistry, geophysics, petrophysics, and numerical and graphical modeling. A complex array of detailed technical information may be available for any given research or exploration project, and it is up to the project leaders or district geologists to make sense of all these services. These individuals must develop the broad overview and understand enough about the data input to re-examine both their models and the data itself when things cannot be made to fit.

It is hoped that this book will be useful for advanced undergraduate and postgraduate courses, and that it may provide some alternative ideas for thesis research projects. In sedimentary geology, too many research projects deal in great detail with very small areas, ignoring any regional context, or concentrate on developing a particular technique or specialty, without considering the importance of other types of data.

There is a continuing need for a modern text dealing with basin analysis. The enormous advances made in sedimentology, seismic methods, sequence stratigraphy geochemistry, and geophysical basin modeling; the evolution of the depositional-systems and systems-tract method of stratigraphic analysis; the documentation of sea-level changes; the documentation of many types of global stratigraphic cyclicity; the impact of plate tectonics; and the power of computer modeling have brought about a revolution in the methods and results of basin analysis during the last 30 years. Most recent texts in the area have been specialized, dealing with sedimentology (Walker and James 1992; Reading 1996), stratigraphic methods (Schoch 1989; Blatt et al. 1991; Boggs 1995; Prothero and Schwab 1996), sequence stratigraphy (Emery and Myers 1996; Miall 1997), and the tectonics of sedimentary basins (Busby and Ingersoll 1995). The purpose of this book is to show how these subdisciplines fit together into the modern science of basin analysis. Many of the exciting new discoveries about the sedimentary history of the earth have been described in a short and highly readable book by Hallam (1981), whereas, in this book, the focus is on the methods of analysis and synthesis that are being used to make these discoveries. The important book by Allen and Allen (1990) covers much the same ground as this book, but with a much greater emphasis on quantitative methods.

The various elements of the synthesis, such as facies models, trend analysis, seismic stratigraphy, sequence stratigraphy, etc., are touched on in this book, but the coverage of these could not attempt to be exhaustive, because the intent is to show how the elements contribute to the whole. Practical methods are emphasized, and theory is downplayed. Thus, some readers may find coverage of their favorite topic unsatisfactory, and I hope the provision of lists of selected references will serve as an adequate substitute. An example of the practical approach used here

is in the treatment of sedimentary structures. Most sedimentology textbooks start by dealing in greater or lesser detail with fluid mechanics, continue on to discuss sediment dynamics and bedform generation, arriving at sedimentary structures at the end of the discourse. This is a correct theoretical approach, but most such treatments make only passing reference to the fact that, because of sedimentary lag effects (Allen 1974) and the ubiquitous presence of minor internal erosion surfaces, it is usually very difficult to make use of this theoretical material when dealing with a practical problem in ancient rocks. The approach used here is to describe what can actually be seen in the sediments first, then to discuss what kinds of interpretations can be made from them.

The book is divided into two parts. Part 1 deals mainly with descriptive aspects of sedimentary basins, including their stratigraphy, sedimentology, and related areas of geophysics, such as seismic facies. The topics of the chapters range from the simple observational details described in Chapter 2 to the problems of stratigraphic and facies analysis (Chaps. 3, 4) and the mapping and paleogeographic analysis of large-scale depositional systems and sequence stratigraphy (Chaps. 5, 6). Part 2 focuses on the major sedimentary controls, including mantle processes, plate tectonics, and orbital forcing of climates. We deal with crustal behavior and the origin of sedimentary basins (Chap. 7), regional, global and astronomical controls on stratigraphic cyclicity (Chap. 8), and, through a discussion of the many possible styles of plate interaction, we explore the various basin configurations and evolutionary patterns that plate-tectonic processes can create (Chap. 9).

Those engaged in a practical basin problem will, it is hoped, find themselves turning back and forth among the chapters, and the organization of the book is not intended to imply any particular order in which a basin analysis should be performed.

1.2 The Modern Revolution in Stratigraphy

1.2.1 Traditional Stratigraphy

The subject of stratigraphy once provided uninspired teachers with unrivaled opportunities to bore their students to distraction. The traditional method of teaching the subject was to choose selected parts of the world and laboriously ascend the local stratigraphic column, cataloguing the names of the principal stratigraphic units and documenting their lithology and fossil content. This may have been coupled with an attempt to relate stratigraphic style to theories of cratonic and geosynclinal sedimentation, but the main emphasis was on formation and group and member names, and more names. The result was the layer-cake view of stratigraphy sediments, as geometrically uniform blankets bounded by the sharp vertical lines of the correlation table.

There is no excuse for the continued use of such an approach. Formation names, except for those of particular historical interest or those of units that demonstrate a stratigraphic principle of importance, should be left to the local specialist. The many scientific developments touched on previously now permit and, indeed, require a completely different genetic approach to the subject of stratigraphy. Far from being a dull, descriptive art, stratigraphy should now be the discipline that represents the pinnacle of our achievements in studying sedimentary rocks, based on the most fundamental understanding of global sedimentary and tectonic processes. It is one of the principal aims of this book to outline the elements of this science.

Unfortunately, in the present-day real world of stratigraphy, we have to deal with a wide range of different approaches to the subject. In the heavily populated parts of the developed countries, sedimentary geology is cluttered with a myriad of local stratigraphic names built up since the 19th century by dedicated workers describing local sequences within the compass of isolated map areas. Old concepts of basin development and a poor understanding of depositional environments and facies relationships meant that earlier stratigraphic units were established in the absence of any real understanding of their origins. Arguments about stratigraphic correlation and terminology have been interminable since the classic confrontation between Murchison and Sedgwick over the Welsh lower Paleozoic. P.D. Krynine once said "stratigraphy is the triumph of terminology over common sense." The basic problem – the need (or desire) for the geologist to establish a formal stratigraphy, with named units, before he or she has had the opportunity to examine all the evidence – has remained unchanged for over a century. Ideally, stratigraphic units should be established on the basis of a basin-wide perspective, but they rarely are. Local terminologies continue to be proposed by geologists studying limited areas, while paying little or no attention to the regional framework. As noted below, the advent of sequence stratigraphy will probably change all this.

1.2.2 Modern Developments

Until about 1960, geologists had only two, quite different practical tools for studying stratigraphy. These were lithostratigraphy and biostratigraphy

(see Chap. 3). The increased pace of deep exploration drilling, particularly since World-War II, did not at first change stratigraphic practices fundamentally, except that microfossils of all kinds became essential biostratigraphic tools to complement the older studies of macrofossils, and subsurface logging methods rapidly evolved (Chap. 2). Geophysical tools were limited to an early version of the seismic reflection method and a few primitive petrophysical methods (spontaneous potential, resistivity).

Beginning gradually, around 1960, six fundamental changes have occurred in sedimentary geology, so that the discipline of stratigraphy, as practiced in new frontier basins, particularly those offshore, is a completely new science, bearing little relationship to the subject described in such classic textbooks as Dunbar and Rodgers (1957) or Krumbein and Sloss (1963). An attempt to summarize the application of these new techniques is shown in Fig. 1.1.

Among the most fundamental changes have been the refinements in chronostratigraphy – the study of the absolute age of rock units that have come about during the last two decades. Of particular importance has been the integration of radiometric, magnetostratigraphic, and biostratigraphic data. Dating of sedimentary rocks using radiometric techniques is a complex subject. In part, the results depend on the prior establishment of a refined relative-age scale using biostratigraphic methods. Harland et al. (1964) edited a milestone compilation of the research and its results for the Phanerozoic, and it became a standard reference in the field, although revisions have long since rendered that particular publication obsolete. The chronostratigraphic framework is now good enough that reasonably accurate ages can be given for at least the major subdivisions of the periods, e.g., for the upper, middle and lower Devonian. In some case, particularly for Mesozoic and Cenozoic rocks, ages of individual stages are now fairly reliable within ranges of potential error of a few million years or less (Chap. 3).

For the post-Triassic period, a quite different dating technique has become available. This is magnetic reversal stratigraphy or **magnetostratigraphy**. Vine and Matthews (1963) were the first to demonstrate that the periodic reversal in the earth's magnetic field is recorded in a systematic manner in the new basaltic crust developed at oceanic spreading centers. As the new crust is carried away from either side of these linear spreading axes, it carries the record of the reversals with it as a kind of permanent tape recording. Detailed marine magnetic surveys run across ocean

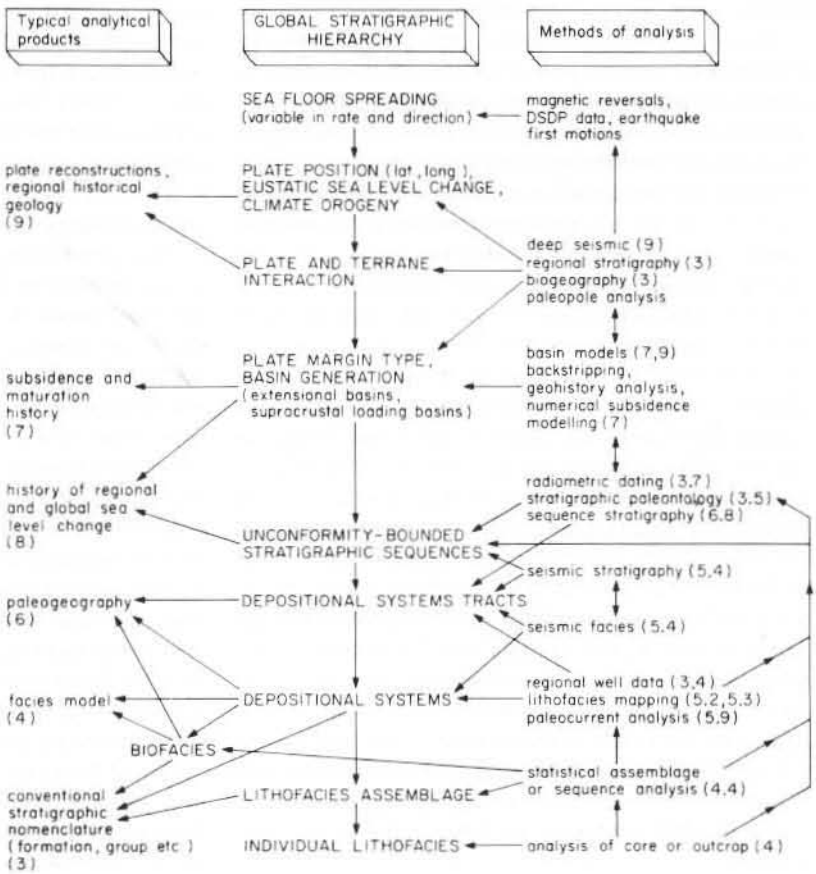


Fig. 1.1. Basin analysis flow chart. Numbers in parenthesis refer to chapters or sections where the subject is discussed

floors in the direction of spreading have produced a documentation of the reversal history that can be dated by biostratigraphic and radiometric means and correlated across all the world's oceans. Distinctive episodes in this reversal sequence are called **magnetic anomalies**, and these have been numbered in sequence. Attempts to develop similar records in stratigraphic rocks have been hampered by the problems of nonsequences and erosion surfaces. Few of the intervals within the reversal sequence are distinctive, although some normal or reversed polarity intervals are markedly longer or shorter than others and may therefore be recognizable in an incomplete profile. The method may be used for local correlation and tied to the global time scale by use of radiometric control points; but, the older the rocks are, the more difficult it is to compare these results to the standard oceanic sequence. Early developments are discussed by Watkins (1976) and Kennett (1980). The current state of the art is surveyed in Chapter 3.

The oldest undisturbed oceanic crust on earth is Jurassic (preserved remnants off the eastern United States, east Africa, and adjacent to the Mariana Trench), and, therefore, pre-Jurassic reversal stratigraphy must be derived entirely from sedimentary or volcanic successions, with all their contained sedimentary breaks. It remains to be seen whether a reliable magnetic chronology can be developed for these.

The second of the six major changes has been the evolution of sedimentology into a mature science, capable of explaining the origin of sedimentary rocks through facies studies and facies models. Stratigraphy has long been thought of as a descriptive science, providing documentation of past events on earth but without adding much to our store of fundamental, theoretical knowledge of global processes – a good example of the "stamp collecting" variety of science that was derided by Lord Rutherford and his ilk, early in this century. However, this is no longer the case. The first modern revolution, in my view, was the explosion of knowledge regarding the physical and chemical processes of sedimentation that began in the late 1950s and early 1960s, stimulated and, in large measure, carried out by petroleum geologists, such as the famous Shell Development Group in Houston and the group led by Harold Fisk at Esso. The emergence of the flow-regime concept as a unifying theory to explain hydrodynamic sedimentary structures (Harms and Fahnestock 1965; Simons et al. 1965) and the detailed documentation of sedimentary products and processes in carbonate environments, such as the Bahamas Banks and the Persian Gulf (Ham 1962), were among the many developments that formed the basis for the revolution in the generation of **process-response models** or **facies models** in the 1960s and 1970s.

The science of sedimentology matured from a study of thin-section petrography, one of its primary focuses in the 1930s, into an analysis of facies, facies successions, sedimentary processes, and regional stratigraphic patterns (Dunbar and Rodgers 1957; Pettijohn 1962; Krumbein and Sloss 1963; Potter and Pettijohn 1963). The facies-model concept was first discussed at a conference in 1958 (Potter 1959), and the first facies models were those of Oomkens and Terwindt (1960) for tidal channels, Bernard et al. (1962) and Allen (1963) for fluvial point bars, Bernard et al. (1962) for regressive shoreface sands, and Bouma (1962) for sandy turbidites. Textbooks edited by Reading (1978) and Walker (1979) were among the first to synthesize the new developments in the field of facies analysis and process-response models, including the many facies models that began to appear at this time. It is now possible to interpret and predict the composition, geometry, and orientation of virtually all stratigraphic units using this approach, based on a wide range of studies in modern environments and ancient rock units (Chap. 4).

This revolution was built on the Lyellian basis of "uniformitarianism", which at first placed undue emphasis on slow, evolutionary, steady-state processes of the type that can be observed all around us every day. It took some time for the importance of rare convulsive or catastrophic events to become apparent, because our first-hand knowledge of such events is so incomplete (Ager 1981, 1993; Dott 1983; Hsü 1983; Clifton 1988). The gradual piecing together of the evidence for the terminal Cretaceous impact event, since the classic paper by Alvarez et al. (1980), has had a significant effect on our views of the rare event in sedimentology (and many other sciences).

This genetic approach to stratigraphy was carried a step further in the **depositional-systems** method, the third major development. This was developed by the Texas Bureau of Economic Geology (Fisher and McGowen 1967; Frazier 1974; Brown and Fisher 1977) and builds on an extension of Walther's Law, which states that, in a conformable succession, the only facies that can occur together in vertical succession are those that can occur side by side in nature (Middleton 1973). A prograding delta is a good illustration of this. It contains prodeltaic, delta front (bay and distributary mouth), and delta plain (distributary channel and overbank) facies, which occur side by side in that order and the products of which occur together in the same order in vertical succession. A depositional system is this complete package of environments and its sedimentary products. Depositional systems are bounded by nonconformities or by facies transitions into adjacent, genetically unrelated systems, such as the passage of a deltaic coast into a clastic-sediment-starved carbonate shelf. Use of the depositional-sys-

tems concept enables predictions to be made about the stratigraphy of large masses of sediment, because it permits interpretations of the rocks in terms of broad paleoenvironmental and paleogeographic reconstructions (Chap. 5). This technique has now become part of **sequence stratigraphy**, as described below and in Chapter 6. Sequences can be correlated regionally as packages of unconformity-bounded strata that record local or regional changes in relative sea level. They may correlate with global standards, indicating a eustatic control (Chap. 8). Sequence-stratigraphic methods have now become a standard part of basin analysis methodology, as noted below.

The fourth revolution, that of **plate tectonics**, began at about the same time as that in sedimentology, but carried different kinds of implications for sedimentary geologists. That the structure, stratigraphy, and facies of sedimentary basins could be reliably related to plate-tectonic evolution was suggested by Dewey and Bird (1970), and was thoroughly explored by Dickinson (1974) and by Bally and Snelson (1980) in papers that still make for useful reading. From these studies, geologists learned the importance of horizontal plate motions for the generation of vertical movements of the earth's crust. The power to make logical predictions of stratigraphy from plate kinematics – the documented passage of a tectonic plate through various tectonic events and climatic zones – became apparent in papers such as that by Berger and Winterer (1974). The subject of tectono-stratigraphic models is examined in Chapter 9.

Another new field, that of **geodynamics**, has developed from investigations relating plate tectonics to deeper crustal and mantle processes. As part of the development of plate-tectonic theory, geophysicists have been evaluating the rheological behavior of the earth's crust and its response to extensional and contractional processes, thermal changes, and the imposition of sedimentary loads. Work by Sleep (1971) and Watts and Ryan (1976) led to the classic study of McKenzie (1978), which was among the most important in this field. It was the first fully quantitative model that related basin subsidence to crustal processes, in this case the crustal attenuation and thermal changes accompanying continental rifting. More recently, the work of Gurnis (1992) has explained how mantle convection affects the surface elevations of continents, a process he termed **dynamic topography**, and has provided a powerful mechanism to explain the vertical motions of continents (**epeirogeny**). These papers provided the basis for the fourth major development, the emergence of a series of powerful techniques for numerical simulation and computer modeling of basin evolution. This has brought about a profound change in our understanding of how basins "work," as discussed in Chapter 7 (see also Sect. 8.6).

The fifth major development has been the evolution of modern techniques of **seismic stratigraphy.** Several major petroleum companies were working on techniques for seismic processing and interpretation in the 1960s and 1970s, including Shell, Exxon, and British Petroleum, but it was the work of an Exxon team, led by Peter Vail, that quickly received prominence (Payton 1977). Seismic reflections derived from sedimentary rocks are essentially chronostratigraphic correlation lines (Vail et al. 1977c). Modern acquisition and processing techniques can now enhance seismic records to such an extent that the fine architectural details of entire basins can be seen and mapped. In this way, major depositional systems can be delineated and related to each other. Subsurface geologists can therefore recognize the major stratigraphic subdivisions in a basin before the first well has been drilled and can describe their composition and make intelligent predictions about internal variation at an early stage of exploration drilling. By being able to see the whole picture before becoming bogged down in detail, the work of the stratigrapher is made immeasurably easier (Chap. 5). Contrast this with the random collection of partially exposed, incomplete sections that the surface geologist has to work with. Application of the seismic stratigraphic approach, particularly in frontier basins, has resulted in many new ideas about stratigraphic relationships (Mitchum et al. 1977) and the effects of sea-level changes (Vail et al. 1977a, b). Building on the work of Sloss (1963), Vail and his co-workers claimed that the sediments in many basins could be divided into discrete sequences that may be capable of correlation from basin to basin and even around the world. Global (**eustatic**) sea-level changes would be the only possible mechanism for such cyclicity. Recent work has developed several important alternatives to the global-eustasy model, as examined in Chapter 8.

One of the impacts of the new seismic methods was to stimulate a revitalization of interest in all forms of stratigraphic cyclicity, leading to the emergence of **sequence stratigraphy,** the sixth major development in sedimentary geology. Many of our ideas regarding the relationships between sedimentation, sea level, and tectonics had emerged in the 19th century and the early part of the 20th century through the work of such individuals as Suess, Gressly, Walther, Barrell, and Grabau. However, modern concepts of sequence stratigraphy essentially began with the work of Wheeler (1958, 1959), Sloss et al. (1949), and Sloss (1963). Wheeler attempted to systematize our concepts of time, as applied to the stratigraphic record, while Sloss and his colleagues searched for regional order in the wealth of descriptive stratigraphic detail from the North American continent and found a pattern of continent-wide

transgression and regression pointing to repeated sea-level changes. As pointed out by Ross (1991), all the basic ideas that form the modern framework of sequence stratigraphy were essentially in place by the early 1960s, but the revolution did not take place then, partly because sedimentologists became engrossed with the process-response model and, in part, because of a lack of the necessary regional stratigraphic data.

Vail's work contained two important new concepts: first, stratigraphic architecture, on a large scale, could be related to cycles of rising and falling sea level and is, therefore, predictable (Posamentier et al. 1988; Posamentier and Vail 1988); and second, these cycles could be correlated globally, suggesting that global sea-level change, or eustasy, is the major controlling process. It was pointed out most clearly by Carter et al. (1991) that these concepts are independent and need to be considered so by working geologists. This is a point of some importance, because, while the sequence-architecture approach to stratigraphy has become a powerful and increasingly discriminating method of analysis (Chap. 6), the global-eustasy model is becoming increasingly controversial (Sect. 8.5). As discussed in Chapter 8, there are at least five types of regional or global stratigraphic cycles present in the rock record. The development of sequence models has therefore provided stratigraphers with powerful tools for description, interpretation, and prediction.

1.2.3 The New Stratigraphic Method

Developments in the six areas discussed in Section 1.2.2 are far from complete, but already they require an entirely new approach to the procedures for defining, correlating, and interpreting stratigraphic units.

In the past, the **formation** has been the foundation of stratigraphic procedure, mainly because, for geologists working on the ground, it was the most convenient scale of unit for mapping purposes (Chap. 3). The proliferation of formations has hindered rather than advanced the development of stratigraphy as a science. In new or frontier areas (where the ground is still uncluttered), the geologist should now leave the definition of formations to the end of the work, or ignore this step altogether. The first step should be the establishment of the framework of major sequences through seismic methods (Chap. 5), by the use of detailed lithostratigraphic correlation and biostratigraphic zonation (Chap. 3), and by subsurface mapping based on petrophysical logs and cores (Chap. 5). The next step is to interpret each sequence in terms of its component systems tracts, and depositional systems (Chap. 6) using sedimentological data, the principles of facies analysis (Chap. 4), and basin-mapping methods (Chap. 5). Work in tectonically deformed or in Precambrian rocks may have to begin at the depositional-systems stage. Stratigraphic data may be used in a **backstripping** procedure to evaluate basin subsidence history and thermal maturation (Chap. 7). Within each depositional system, it should then be possible to recognize units of formation rank on the basis of genetic criteria, using the sequence and/or paleogeographic synthesis developed in the preceding stage. If required, these formations can then be defined and named for the convenience of local mappers or prospect developers (Chap. 3), but actual naming becomes a refinement of lower priority when using this approach, rather than the culmination of the stratigrapher's efforts. It has been argued that, given the growing importance of sequence stratigraphy, it may be more meaningful to use sequences as the basis for the documentation, subdivision, and naming of a stratigraphic succession, rather than conventional lithostratigraphic formations. The principles of **allostratigraphy** are available for use in this context, as set out in Section 3.6.

In the 1960s, the science of stratigraphy began the move away from the descriptive and qualitative to the predictive and quantitative. Studies such as those noted above provide descriptive stratigraphy with an underpinning of theoretical knowledge about basinal controls – the surface environmental aspects that comprise the study of sedimentology, and the deeper crust and mantle processes that are the realm of geophysics. The foundations for a powerful new technique – that of numerical modeling of earth processes aided by graphic simulation – began to appear in the late 1970s, and is discussed in Chapter 8. The combination of plate tectonics, geodynamics, and sedimentology became a powerful one, forming the basis of what came to be known in the 1970s as basin analysis, the study of the origin and development of sedimentary basins, including their depositional and structural history. Vast numbers of regional basin studies have been undertaken using the techniques that evolved at that time.

Although it overlaps in time and, to some extent, scientifically, with the revolution in sequence stratigraphy, the *Decade of North American Geology Project* (DNAG) of the Geological Society of America and the Geological Survey of Canada, which was planned as a major commemoration of the centenary of the Geological Society of America in 1988, is in large measure a product of the explosion of research techniques and energy that took place earlier, in the 1960s and 1970s, of which basin analysis was an important part. DNAG succeeded in bringing together hundreds of scientists to generate a modern synthesis of the geological history of the continent and its adjacent

oceans, and has resulted in a collection of some 28 regional volumes, 6 volumes of field guides, and various other products. The *European Geotraverse* has been a similarly ambitious project (Blundell et al. 1992). In both cases, stratigraphy and basin analysis form a large part of the research product.

1.3 Time in Stratigraphy

Sedimentologists and stratigraphers study sedimentary events over time periods extending through 17 orders of magnitude, from the long-term global changes in plate distribution, atmospheric composition, and oceanic geochemistry that take place over billions of years (10^9 years), to the entrainment and displacement of individual sand grains in traction carpets, events that take but a few seconds (3 s = 10^{-7} years) (Miall 1991). Different parts of this time spectrum have become the special interest of various scientific groups (and are considered in different parts of this book). Consider this wide variety of observations:

1. Those interested in fluid hydraulics and the development of bedforms are concerned with periods of a few seconds to a few hours (10^{-7} years to 10^{-3} years; Chap. 4).
2. Those who study sedimentary facies and depositional models have to take into account diurnal changes in air and water circulation and tidal effects at one extreme (10^{-3} years), the effects of **dynamic events**, such as floods and storms (durations of 10^{-3} years to 10^{-2} years, spaced over intervals of 10^{-1} years to 10^{1} years), and, at the other extreme, the growth, migration, and progradation of various subenvironments, such as river channels, tidal flats, delta lobes, and crevasse splays (10^{1}–10^{4} years; Chap. 4).
3. Conventional stratigraphy – the mapping and formal definition of formations, members and groups, and sequence stratigraphy – deals with sedimentary packages and depositional systems that may take 10^{2} years to 10^{7} years to accumulate (Chaps. 3, 5, 6).
4. For those concerned with the establishment of a geologic time scale, chronostratigraphic precision in the dating of stratigraphic events is rarely better than about ±1 m.y. (10^{6} years), this being the highest level of resolution that can normally be achieved with biostratigraphic data. In exceptional cases, for example, where radiometric determinations on numerous superimposed volcanic ash beds permit a calibration of volcanic events against sedimentary thicknesses, a precision of 10^{5} years may be achieved. For parts of geologic time, the same precision is now starting to be achieved with the use of magnetostratigraphy (Chap. 3) and, for the Neogene, the availability of a chemostratigraphic scale based on oxygen isotope variations has provided an even finer degree of resolution.
5. There is increasing recognition of the importance of **astronomic (orbital) forcing** as a control on climate. Commonly, in the geologic past, this has been responsible for the growth and decay of continental ice caps, with consequent major effects on climate and global sea level and, hence, on stratigraphic cyclicity, notably the development of the cyclic deposits termed cyclothems. These are the so-called **Milankovitch effects**, which deal mainly with time periods of 10^{4}–10^{5} years (Chap. 8). The study of these cycles and the time scale that may be derived from them constitutes **cyclostratigraphy**.
6. There is increasing evidence for global correlation of major sequences with durations in the range of 10^{7} years. The driving force for these sequences is thought to be eustasy, driven by changes in the global average rate of sea-floor spreading. This affects spreading-ridge volumes and hence the volume of the ocean basins (Chap. 8).
7. Geophysical modeling of sedimentary basins has, in recent years, focused on the flexural and rheological behavior of the earth's crust in response to stretching, heating, and loading events, and on the regional (epeirogenic) changes in continental elevation caused by mantle thermal processes (dynamic topography). Subsidence, driven by these mechanisms, is slow and the evolution of the large-scale architecture of a basin is a process that takes 10^{6}–10^{7} years to complete (Chap. 7).
8. The assembly and dispersion of supercontinents on the earth's surface is a product of the thermal behavior of the core and mantle. Evidence is accumulating to support a long-term cyclicity during Proterozoic and Phanerozoic time, with a periodicity in the order of 10^{8} years (Chap. 8).

The disparities in the time scales that form the background to the work of these various groups have given rise to some tensions in the development of basin analysis. For example, the study of fluid hydraulics in the laboratory has yielded elegant phase diagrams that permit interpretations to be made of instantaneous flow conditions in the development of sequences of hydrodynamic structures (crossbedding, parting lineation, etc.). However, such analyses relate only to the few minutes or hours when the individual bedform was being deposited and can tell us nothing about the usually much longer periods of time (up to several or many years) represented by nondeposition or erosion; they also tell us nothing

about the preservation potential of the various bedforms.

There has been considerable debate in recent years regarding the completeness of the stratigraphic record and the importance of the "rare event" in geology. Ager (1981) was among the first to bring these ideas to the attention of geologists, and other important essays on the subject have been presented by Gretener (1967), Sadler (1981), Dott (1983), and Reading (1986). Ager (1981) pointed out that the sedimentary record is more gap than record, that in most environments the total time represented by the actual beds is only a fraction (sometimes a small fraction) of the total time represented by the overall succession. Considerable lengths of time are represented by bedding planes, diastems, and disconformities. This has given rise to problems in reconciling sedimentation rates measured in modern settings with those deduced from the ancient record. McKee et al. (1983) used radionuclides to study sedimentation rates on the modern continental shelf and found that the calculated rates varied depending on the scale of measurement used. Nuclides with slow decay rates yielded lower sedimentation rates than those with rapid decay rates. Slower decay rates permit more episodes of nondeposition to be recorded in the sediments. Miall (1991) collated data on sedimentation rates in a wide range of modern and ancient sedimentary basins and found that measured rates in modern settings are typically an order of magnitude higher than those calculated from the geologic record (Sect. 3.7.7).

The concept of **event stratigraphy** has emerged, which emphasizes the importance of short- and long-term dynamic phenomena in earth's history. Dynamic events, such as hurricanes and flash floods, may seem to be rare events on a human time scale, but have a high probability of recurrence when considered on a geologic time scale. Such events are typically responsible for deposition of disproportionately large volumes of sediment in a sedimentary basin, when measured against the slow accumulation that takes place during "normal" conditions. We consider these ideas further in Chapter 4.

Longer term "events" include changes in climate, tectonics, and global sea level, which are responsible for widespread cyclicity in the rock record. These have several ultimate causes and range in periodicity over several orders of magnitude (10^4–10^7 years). These cycles, or **sequences**, are described in Chapter 6, and their causes are discussed at length in Chapter 8.

The methodologies of event stratigraphy and sequence stratigraphy are becoming increasingly central to stratigraphic correlation and the documentation of basin evolution. Short-term events may leave marker beds that find great usage in lithostratigraphic correlation (Sect. 3.4.3). The record of longer term events, stratigraphic sequences, provides information on changes in sea level that may be interpreted in terms of regional processes driven by tectonism, or global events driven by eustatic processes, such as orbital forcing (Miall 1997). These subjects are discussed in Section 3.6 and Chapter 8.

1.4 Types of Project and Data Problems

Data collection and analysis procedures are, of course, determined by the nature of the project. The following are some typical basin-analysis problems, with a brief discussion of the data collection potential.

1.4.1 Regional Surface Stratigraphic Mapping Project

Work relating to regional surface stratigraphic mapping is one of the primary functions of government surveys, intent on providing complete map-sheet coverage of their area of responsibility, both as a service to industry and as a basis for expert advice to government economic planners. Similar regional surveys are commonly undertaken by industry as a preliminary to detailed surface or subsurface exploration, although their studies are rarely as thorough. Many academic theses are also of this type.

Many government surveys are carried out by individuals who are not specifically trained in the analysis of sedimentary basins, the idea being that members of the survey should be generalists, capable of mapping anything. This is the old British tradition, and it is an unfortunate one because it means that the individual survey officer cannot possibly be aware of all the skills that are now available for mapping work in sedimentary rocks, nor are they encouraged to take the time for the specialized observations, which would make their work so much more effective. The argument that this is "left for the academics to do later" is not always satisfactory, for it is commonly the case that the stratigraphy of a succession can only be clarified by those who thoroughly understand its sedimentology. Many fruitless arguments about stratigraphic terminology can be avoided if this is realized at the beginning of a mapping endeavor. A team approach to a project is often an ideal solution. For instance, the research synthesis edited by Besly and Kelling (1988) is an excellent example of an integrated basin-analysis approach to a broad and complex research problem, in this case, the origin of a major series of sedimentary basins in northwest Europe.

Another argument that is sometimes heard about geological mapping is that once it is done it does not

need to be done again. Political pressures to disband state geological surveys have been based on this argument, which, of course, ignores the fact that maps constantly need to be updated to incorporate new data, or redrawn to encompass new concepts. It is likely that, eventually, many standard surface geological maps will be redrawn using sequence concepts instead of lithostratigraphic terminology.

The basis of all surface basin-analysis projects is the careful compilation of vertical stratigraphic sections. These are described by the geologist in the field, who also collects samples for subsequent laboratory analysis, taking care to label each sample according to its position in the section. Where should such sections be located? The choice depends on a variety of factors. First, they should be typical of the area in which they are found and should be as free as possible of structural deformation. Obviously, they should also be well exposed and, ideally, free of chemical or organic weathering, which disturbs or obscures textures and structures.

The best sections for regional correlation are those that include several stratigraphic intervals, but it is rare to find more than a few of these except in exceptionally well exposed areas. Short, partial sections of a stratigraphic unit can provide much valuable sedimentological data, such as facies and paleocurrent information, although it may not be possible to locate them precisely within a stratigraphic framework unless they can be correlated by a marker bed or structural interpretation (Chap. 3). Exposures of this type tend to be ignored by the regional mapper, but they should not be ignored by those intent on producing an integrated basin analysis, because they add to the database and may provide many useful sedimentological clues.

The great advantage of surface over subsurface studies is the potential, given adequate clean outcrop, to see medium- to large-scale sedimentary features, such as crossbedding, and larger scale architectural elements, including channels, large bar deposits, bioherms, etc., that may be difficult or impossible to identify in a drill hole. Large outcrops can be used to construct lateral profiles (Sects. 2.2.2, 4.5.9). These, of course, add immeasurably to any basin interpretation, particularly paleogeographic aspects, such as the size, geometry, and orientation of depositional elements, e.g., channels. The geologist should also always be searching for lateral variations in lithologies, fossil content, or sedimentary structures, as these changes may provide crucial control for paleogeographic interpretations.

The disadvantage of studies carried out exclusively at the surface is that most of the rocks in any given basin are buried and may be inaccessible to observation over very large areas. Many basins are depositional basins, in the sense that they preserve, in the present day, essentially the same outline as during sedimentation. The rocks exposed at the surface, especially around the margins of the basin, may have a quite different thickness and facies to those preserved at the center and may show erosion surfaces and unconformities not present in the center because of the tendency of basin margins to be affected by a greater degree of tectonic instability. This is illustrated in Fig. 1.2. A basin analysis carried out under such circumstances might therefore produce very incomplete or misleading results.

1.4.2 Local Stratigraphic-Sedimentologic Mapping Project

A common incentive for a detailed local study is the occurrence of some highly localized economic deposit, such as an ore body or coal seam. A geologist will examine every available outcrop within a few square

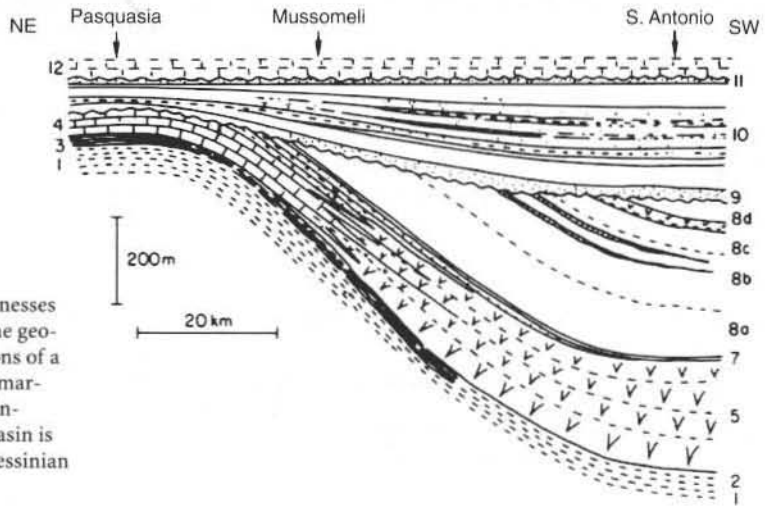


Fig. 1.2. Contrasts between stratigraphic thicknesses and facies at basin center and basin margin. The geology of the deep, hydrocarbon-producing regions of a basin might be quite different from that at the margins, so that surface geology gives little useful information on what lies below. The North Sea Basin is an excellent example. This example is of the Messinian evaporite basin, Sicily. (Schreiber et al. 1976)

kilometers of the deposit and may also supplement the analysis with logs of diamond drill holes. Another type of local study is an academic thesis project, particularly at the master's level. A small-scale project is chosen because of the time and cost limits imposed by academic requirements.

Many of the comments given in the preceding section apply to local studies, but there are some additional complications that often arise because of the nature of such projects. A common fault is the emphasis on local features to the exclusion of any real consideration of regional implications. The geologist will erect a detailed local stratigraphy and fail to show clearly how it relates to any regional framework that may have been established, or there may be an overemphasis on certain selected parameters chosen, perhaps, as a training exercise, to the exclusion of others.

To set against these problems is the advantage a detailed local project may offer for carrying out very complete, sophisticated paleogeographic reconstruction. Rather than relying on selected stratigraphic sections, the geologist may be able to trace out units on foot or study their variation in a close network of diamond drill holes (Fig. 1.3). In this way, the detailed variations within, for example, an individual reef or channel network or a coal swamp, can be reconstructed. Typically, there will be more time to spend developing architectural data from large outcrops. Such reconstructions are particularly valuable because they provide a mass of three-dimensional data against which to test theoretical facies models and sequence concepts.

Diamond-drill holes are produced by a process of continuous coring. They therefore provide a complete lithologic and stratigraphic sample through the units of interest. The small diameter of the core (2–5 cm) does not permit recognition of any but the smallest sedimentary structures (Fig. 1.8), but the close drill-hole spacing of a few hundred meters or less means that very detailed stratigraphic correlation is usually possible.

Unfortunately, a tradition of retention and curation of diamond-drill-hole core has rarely emerged in government or industrial organizations except, perhaps, for a few key holes in areas of particular interest. The core is normally discarded once the immediate interest in the area has subsided, and much valuable material is lost to future workers. The mining industry seems to prefer it this way, but what is good for corporate competitiveness is not necessarily the best method for developing a national stratigraphic database.

1.4.3 Regional Subsurface Mapping Project

Exploration activity in the petroleum industry is now mainly of the regional subsurface mapping type. Companies may initially send surface geological mapping parties into the field, but ultimately the most thorough field studies in a given field area are likely to be done by government survey organizations. Much industry activity is now located in offshore regions for which surface geological information is, in any case, sparse, unobtainable, or irrelevant. As noted previously, the beds exposed at the edge of a basin may bear little relationship to those buried near its center (Fig. 1.2).

Regional subsurface work is based initially on geophysical data and subsequently on test drilling. Gravity and aeromagnetic information may provide much useful information on broad structural features, particularly deep crustal structures (Fig. 1.4; Chap. 5). Refraction seismic lines may be shot for the same purpose. More detailed structural and stratigraphic data are obtained from reflection-seismic surveys (Chaps. 5, 6, 8), and these provide the basis for all exploration drilling in the early phases of basin development. Deep reflection surveys during the last decade have profoundly changed our ideas about deformed belts (Chap. 9). Seismic shooting and processing is now a highly sophisticated process, and its

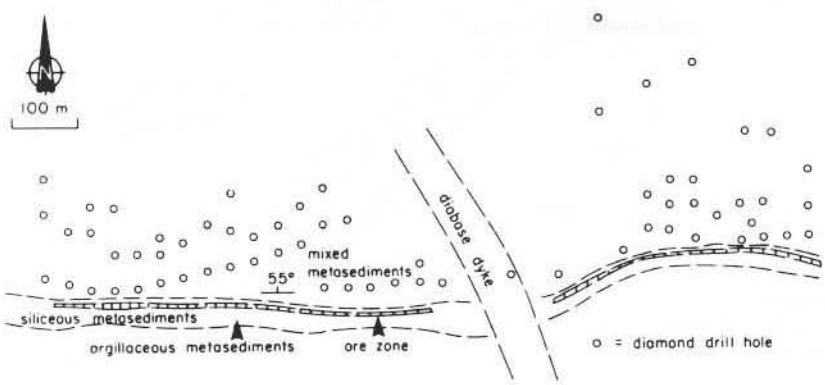


Fig. 1.3. A typical diamond drill hole (DDH) network across a mining property, a gold prospect in Precambrian metasediments, northern Ontario

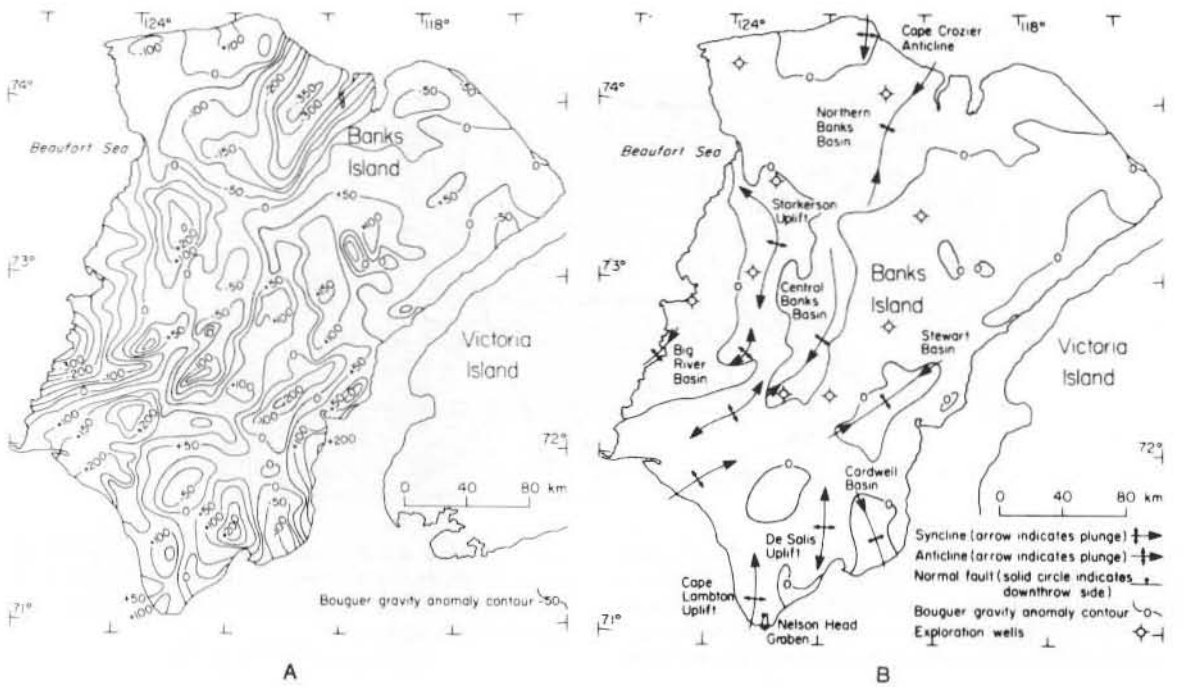


Fig. 1.4. Use of regional gravity data (A Bouguer anomalies) in defining broad structural features (B), Banks Island, Arctic Canada. In this case, exploratory drilling of the nine wells shown indicated that negative anomalies corresponded closely to the outline of Mesozoic-Cenozoic sedimentary basins up to 2-km thick, resting unconformably on Devonian strata. Gravity data from Stephens et al. (1972), interpretation from Miall (1979)

practitioners like to talk about seismic stratigraphy as if it can provide virtually all the answers, not only about structure but about the stratigraphic subdivision of a basin, regional correlation, and even lithofacies (Fig. 1.5). However, the seismic method is only one exploration tool, and its results must be tested against those derived in other ways. For example, test drilling often shows that stratigraphic correlations predicted from seismic interpretation are incorrect. Problems of stratigraphic velocity resolution, the presence of low-angle unconformities, and the obscuring effects of local structure can all introduce errors into seismic interpretation. Resolution of bedding units is still relatively crude, as illustrated in the comparison between typical seismic wave forms and the scale of actual bedding units shown in Fig. 1.6 (see also Sect. 5.4.1). Dating and correlation of seismic sequences (Chap. 8) depend heavily on biostratigraphic and geophysical log information from exploration wells.

Exploration wells, especially the first ones to be drilled in a frontier basin, are as valuable in their own way as space probes sent out to study the planets; it is unfortunate that, in many countries, the resulting data are not treated with the same respect afforded space information, but are regarded as the private property of the organization that paid for the drilling. In a competitive world, obviously, a company has the right to benefit from its own expenditures but, in the long-term, knowledge of deep basin structure and stratigraphy belongs to the people and should eventually be made available to them. In Canadian frontier areas, well samples and logs must be deposited with the federal government and released for public inspection 2 years after well completion. Two-year competitive advantage is quite long enough for any company in the fast-moving world of the oil industry, and, after this time period, the well records become part of a national data repository which anybody can use – with obvious national benefits. Seismic records are released after 5 years.

The nature of the stratigraphic information derived from a well is both better and worse than that derived from surface outcrops. It is better in the sense that there are no covered intervals in a well section, and such sections are generally much longer than anything that can be measured at the surface (perhaps exceeding 6000 m), so that the stratigraphic record is much more complete. The disadvantage of the well section is the very scrappy nature of the actual rock record available for inspection. Three types of sample are normally available:

1. Cuttings. These are produced by the grinding action of the rotary drill bit. These are generally less

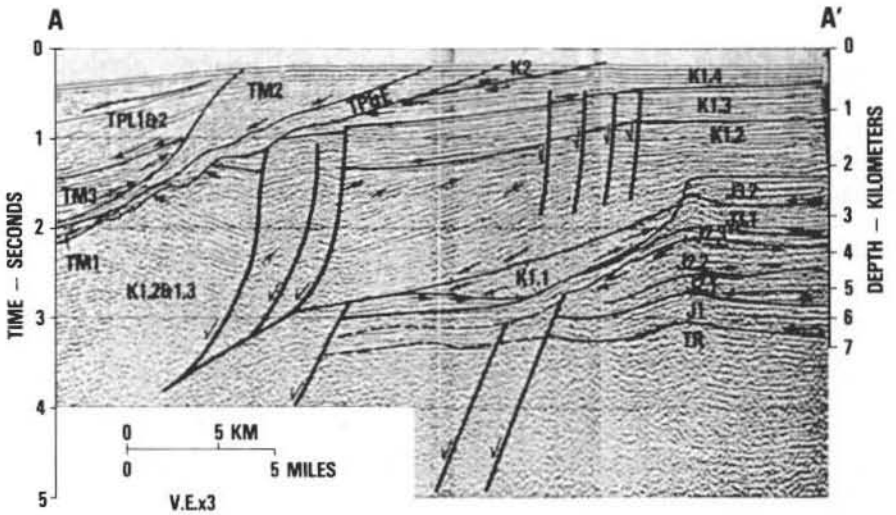


Fig. 1.5. Example of an interpreted seismic line across the continental margin of North Africa (Mitchum et al. 1977). This is a famous section and has been reproduced in many textbooks as one of the first examples published to show the wealth of stratigraphic detail that could be extracted from the data. Note the presence of several angular unconformities, buried paleotopography, the internal architectural details of individual stratigraphic units, and the system of standard coded stratigraphic units employed by the authors



Fig. 1.6. Scale of a typical seismic wave form compared with an outcrop. (Based on an idea from A.E. Pallister and A.E. Wren)

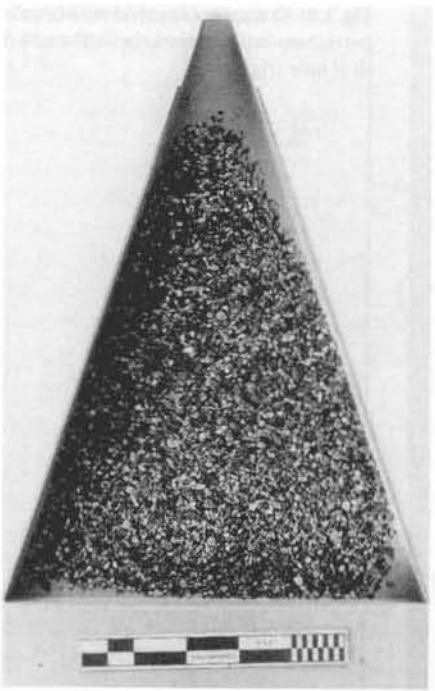


Fig. 1.7. Typical well cuttings. (Photograph courtesy of J. Dixon)

than 1 cm in length (Fig. 1.7) and can therefore only provide information on lithology, texture, and microfossil content. The North American practice is to collect from the mud stream and bag for examination samples at every 10 ft (3 m) of drilling depth. When drilling in soft lithologies, rock may cave into the mud stream from the side of the hole many meters above the drill bit, so that the samples become contaminated. Also, cuttings of different density may rise in the mud stream at different rates, which is another cause of mixing. It is thus necessary to observe the first appearance rule, which states that only the first (highest) appearance of a lithology or fossil type can be plotted with some confidence. Even then, depth distortions can be severe.

2. Full-hole core (Fig. 1.8). The rotary drill bit may be replaced by a coring tool when the well is drilling through an interval of interest, such as a potential or actual reservoir bed. However, this type of core is expensive to obtain, and it is rare to find that more than a few tens of meters, perhaps only a few meters, of core are available for any given hole.
The advantage of a core is that, because of the large diameter (usually in the order of 10 cm), it permits a detailed examination of small- to medium-sized sedimentary structures. Macrofossils may be present, and trace fossils are usually particularly well seen. The amount of sedimentological detail that can be obtained from a core is thus several orders of magnitude greater than is provided by chip samples. However, it is frustrating not to be able to assess the scale significance of a feature seen in a core. An erosion surface, for example, may be the product of a local scour, or it may be a major regional unconformity, but both could look the same in a core. Hydrodynamic sedimentary structures might be present, but it may not be possible to interpret their geometry and, except in rare instances where a core has been oriented in the hole, they provide no paleocurrent information. Orientation can sometimes be deduced if the core shows a structural dip that can be determined from regional structural data or dipmeter logs.
3. Side-wall cores. These are small plugs extracted by a special tool from the wall of a hole after drilling has been completed. These cores are rarely available to the geologist, because they are used in porosity/permeability tests and are disaggregated for caving-free analysis by biostratigraphers. In any case, their small size limits the amount of sedimentological information that they can yield.

In addition to the samples and core, each exploration hole nowadays is subjected to an extensive series of petrophysical logging methods, which provide records by direct analog tracing or digitization. The description and interpretation of such logs has been the subject of several textbooks and cannot be dealt with exhaustively here. Log information is discussed further in Section 2.4, and the dipmeter in Sections 5.4.3 and 5.9.6. The following are a few preliminary remarks discussing the utility of logs in a subsurface data-collection scheme.

Most geophysical tools measure a single physical property of the rock, such as its electrical resistivity, sonic velocity, and gamma radioactivity, etc. (Fig. 1.9). These properties reflect lithology, and can therefore be used, singly or in combination, to interpret lithology. Because measurements with modern tools have depth resolution of a few centimeters, they are of great potential value in deriving accurate, depth-controlled lithologic logs free of the problems of sample caving. The response of a single tool is not unique to each rock type; for example, many different formations will contain rocks with the same electrical resistivity, and so it is not possible to interpret lithology directly from a single log type. However, such interpretations may be possible from a combination of two or more logs, and attempts have been made to automate such interpretations based on computerized calculation routines from digitized log records. Unfortunately, the physical properties of rocks and their formation fluids vary so widely that such automated interpretation procedures can only

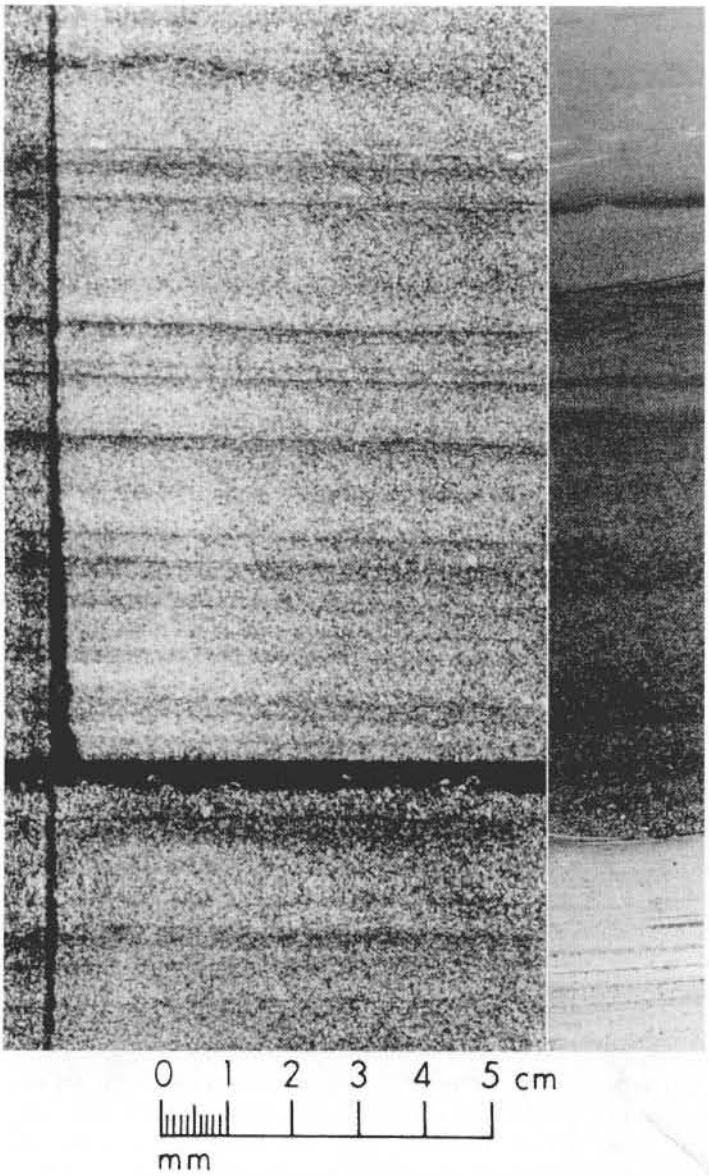


Fig. 1.8. Examples of typical core from a petroleum exploration hole (*left*) and a diamond drill hole (*right*)

be successful if they are adapted to the specific conditions of each basin. They thus lose much of their exploration value, but become of considerably greater importance once an initial scatter of exploration wells has become available to calibrate the logs.

A particularly common use to which geophysical logs are put is in stratigraphic correlation. Log records through a given unit may have a distinctive shape, which a skilled geologist can recognize in adjacent holes. Correlation may, therefore, be possible even if details of lithology are unknown and, indeed, the establishment of correlations in this way is standard practice in subsurface work (Sect. 3.4.3; Chap. 5). Similarly, log shape is a useful tool in environmental interpretation, though it must be used with considerable caution (Sect. 4.5.8).

Logs are also of considerable importance in calibrating seismic records – seismic velocities are routinely derived from sonic logs and used to improve seismic correlations (Sect. 5.4.1).

Regional subsurface work may lead to the development of a list of petroleum **plays** for an area. These are conceptual models to explain the generation-migration-trapping system for petroleum pools. For example, a Devonian reef play, which might be based on the occurrence in Devonian strata of porous, dolomitized reef masses enclosed in mudrock, would provide an ideal series of stratigraphic traps. The development of a petroleum play leads to the evolution of an exploration philosophy, which summarizes exploration experience in choosing the best combination of exploratory techniques and the type of data requir-

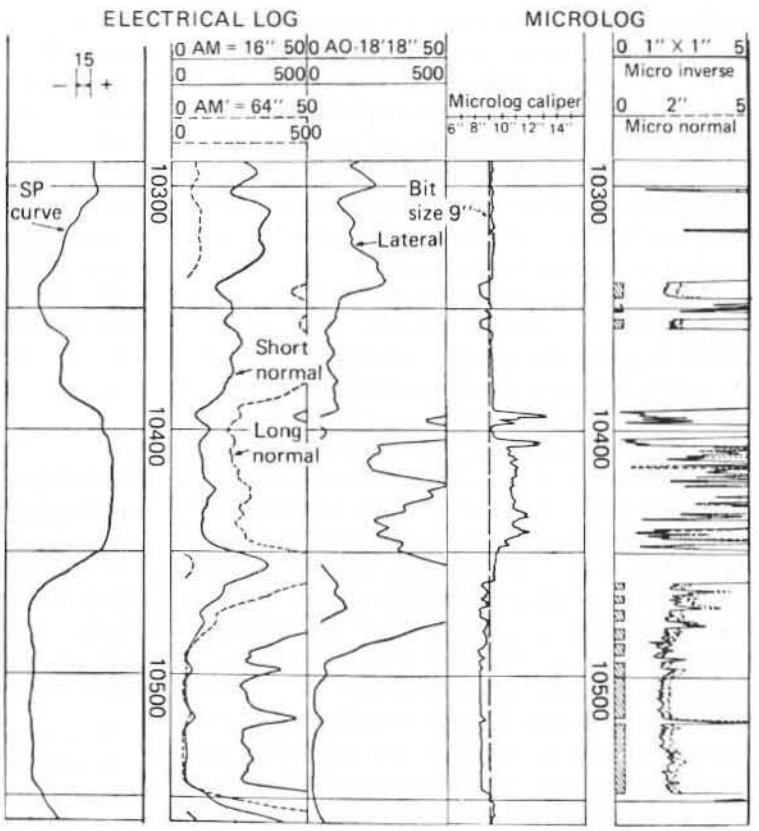


Fig. 1.9. Typical geophysical log suite from a petroleum exploration hole. (Krumbein and Sloss 1963)

ed to locate individual pools within the play area. At this stage, exploration moves from the play stage to the **prospect** stage. The first wells drilled in an area might have been drilled on specific prospects, with a specific play in mind, but it is rare for early wells to be successful. Prospect development is the next phase in an exploration program.

It is now common practice in petroleum exploration programs to carry out studies of burial history and thermal maturation of the stratigraphic section in a basin. This is based on what are now called backstripping procedures. Subsidence history is reconstructed using computer programs that reverse the process of sedimentation, stripping off each layer in turn, decompacting the remaining section, and calculating the isostatic balance. In this way, the component of subsidence due to tectonic effects can be isolated from that caused by the weight of sediment, and this, in turn, can be related to the rheological behavior of the crust underlying the basin (Chap. 7). Burial history can be related to thermal history in order to document the organic maturation of the sediments and, in this way, the timing of petroleum generation and release can be evaluated.

Another change during the last decade is that stratigraphic work is now commonly geared to the recognition and correlation of stratigraphic sequences (Chap. 6). These are then used to assess regional paleogeographic histories and cycles of local to regional sea-level changes. The latter may be compared with regional and global standards (Chap. 8), although this is of less importance to the exploration geologist than the production of a useable regional stratigraphic framework.

1.4.4 Local Subsurface Mapping Project

A network of closely spaced holes a few kilometers (or less) apart may be drilled to develop a particular petroleum or mineral prospect. These may be in virtually virgin territory or may be step-out wells from a known pool or deposit in a mature petroleum basin or mine. It is at this stage in the exploration work that geological skills come most strongly into play. In the early exploration phases, virtually the only information available is geophysical; almost all early petroleum exploration wells are drilled on potential structural traps. Once a hydrocarbon-bearing reservoir is located, the next problem is to understand why it is there, how large and how porous it is, and in what direction it is likely to extend. Interpretations of depositional environments, applications of facies models, and reconstructions of facies architecture and paleo-

geographic evolution may become crucial in choosing new drilling sites. Similar procedures must be followed in order to develop stratabound ore bodies. In the case of offshore petroleum exploration, the exploitation of a new petroleum pool may be achieved by directional drilling of many wells from a central platform, and great pressure may be placed upon the development geologists to design a production strategy based on extremely limited data.

The types of data available for local subsurface mapping are discussed in Section 1.4.3. To these should be added three-dimensional seismic surveying, which is becoming an increasingly valuable tool for local structural and stratigraphic mapping (Sect. 5.4.1). At this point, it is pertinent to add a few cautionary notes on the uses of subsurface data. Because of its inherent limitations, it is natural that exploration geologists would wish to exploit the data to the full extent but, paradoxically, this can lead to a very limited approach to an exploration problem. Geologists may be so impressed with a particular interpretive technique that they may tend to use it to the exclusion of all others. This has led to many false interpretations. It cannot be overemphasized that, in basin analysis, as, no doubt, in every area of geology, every available tool must be brought to bear on a given problem. Examples of techniques in the area of sedimentology that have been over utilized include the use of vertical-profile studies, grain-size analysis, and grain-surface-texture analysis to interpret depositional environments.

The use of vertical profiles was first described in detail in print by Visher (1965), although the technique was being explored by several petroleum companies in the 1950s; it followed from **Walther's Law**, which stated, in effect, that only those environments and facies found side by side in nature could be represented in the same order in a vertical stratigraphic profile (Visher 1965; Middleton 1973). Visher, and many others since (including this writer), have devoted many pages of print to documenting the various types of vertical profile and demonstrating their environmental significance. Many of these profiles have characteristic log signatures but, unfortunately, few of these signatures are unique, and mistakes will arise (and have arisen) where undue emphasis is placed on log shape as an environmental indicator. This is discussed further in Chapter 4 (see also Miall 1988).

In the 1970s, grain-size analysis of clastic rocks consumed an enormous amount of research energy. The basic idea was that hydrodynamic sorting of a clastic sediment leaves an imprint on the grain-size distribution, which varies depending on the nature of the hydrodynamic process. The rock may contain a traction load and a suspension load population or the entire grain mass may have been carried in turbulent suspension. It is thought that these sorting processes have subtly different effects in the different clastic environments, such as rivers, deltas, or seas, depending on the nature of the dominant transport process (unimodal currents, waves, wind, etc.). Numerous statistical techniques, including Gaussian-curve statistics and curve shape, discriminant functions, and factor analysis, have been proposed that, it is hoped, would provide infallible guides to the depositional environment. However, it has been demonstrated in many replication studies that none of the methods is infallible, because of multiple grain recycling and diagenesis.

1.5 Summary of Research and Reporting Procedures

Vail et al. (1991) recommended a systematic methodology for tackling a local or regional tectonic-stratigraphic analysis. The following is a modified version of this listing, adapted to suit the practicalities that occur to this writer (see Miall 1997, Chap. 17).

1. Develop an allostratigraphic framework based on detailed lithostratigraphic well correlation and/or seismic facies analysis.
2. Develop a suite of possible stratigraphic models that conform with available stratigraphic and facies data.
3. Establish a regional sequence framework with the use of all available chronostratigraphic information.
4. Determine the relationships between sequence boundaries and tectonic events by tracing sequence boundaries into areas of structural deformation, and documenting the architecture of onlap/offlap relationships, fault offsets, unconformable discordancies, etc.
5. Establish the relationship between sequence boundaries and regional tectonic history, based on plate-kinematic reconstructions.
6. Refine the sequence-stratigraphic model. Subdivide sequences into depositional systems tracts and interpret facies.
7. Construct regional structural, isopach, and facies maps; interpret paleogeographic evolution; and develop plays and prospects based on this analysis.
8. Develop detailed subsidence and thermal-maturation history by backstripping/geohistory analysis.

It is suggested that subsidence and maturation analysis (Step 8; Chap. 7) be carried out at the conclusion of the detailed analysis, rather than near the beginning, as proposed in the Exxon approach. The reason is that a complete, thorough analysis requires the input of a considerable amount of stratigraphic data.

Corrections for changing water depths and for porosity/lithification characteristics, which are an integral part of such analysis, all require a detailed knowledge of the stratigraphic and paleogeographic evolution of the basin.

The structure of a final report should also follow a logical order. A summary of the typical contents of a regional report is given below. Elements of this structure commonly are missing in the first drafts of graduate theses and reports submitted for publications in journals, or are not presented in a logical order. Not all components may be required for every individual study.

1. Introduction to the project. Why has this research been undertaken? What regional problem or scientific question has been addressed?
2. A description of the area covered by the report. This is normally illustrated by the use of a map. A project area may be defined geographically (mapping quadrant, state boundary) or geologically (tectonic province, sedimentary basin).
3. The geological setting of the area. The reader must be able to understand the purpose of the report within the context of the regional geological history. If appropriate, a description of the geological setting may include a brief outline of the plate-tectonic history of the region. Significant prior work on the problem at hand or on the regional geology of the area may be referenced.
4. The stratigraphy of the area, typically including a stratigraphic table. The basis for regional correlations and age designations may be provided. If the purpose of the report is to describe the sequence stratigraphy of the region, prior lithostratigraphic analyses should be briefly summarized, with references to this previous work
5. Location of study sections, wells used, or samples collected. These should be located by annotated points on a map, and other details may be provided if pertinent, such as ground-penetrating radar (GPR) data or national map grid references for key sections. If highly detailed location information is required, this may be placed in an appendix. Annotated aerial photographs may be used to locate key outcrop sections.
6. Research methods should be described. This may include, where relevant, sample collection and preparation procedures, field mapping and measurement methods, laboratory procedures, and data-reduction techniques. Be sure to reference the source of the methods used. Lengthy descriptions of methodology, such as laboratory standardization procedures, may be placed in an appendix. Methods that have been used before do not have to be described in detail, but the reader should be able to understand what you have done without needing to read a body of previously published literature.
7. Presentation of observations and/or laboratory results. As far as possible, this presentation should be separate from any interpretation. Data should be illustrated without interpretation, e.g., uninterpreted seismic sections, raw stratigraphic sections. This may be difficult, for example, where observations have been guided by a new interpretive hypothesis. The use of systems-tract designations in the description of stratigraphic sections is a good example of a supposedly descriptive approach that has generated many problems (see Chap. 6).
8. Discussion of results, interpretations, and new hypotheses or theories arising from the project. Suggestions for further work.
9. Summary and conclusions. The major results. What new insights does this research provide? What new research or new problems does the project highlight? Do not introduce new references at this point.

There are also some simple rules to follow regarding illustrations:

1. All locations and geological regions referenced in the text must be named on a map, preferably near the beginning of the report.
2. All symbols on all figures must be explained, preferably by legends on the figures themselves. The old European and Russian practice of providing symbols in numbered boxes that are explained by numbered notes in the caption is cumbersome and not to be recommended.
3. Lettering on figures should not be smaller than 1-mm high.
4. Photographs, whether of outcrops, samples, thin-sections, scanning-electron-microscope (SEM) images, etc., must include a scale.
5. Figures, tables, and plates must be numbered in the order by which they are referenced in the text.

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The Stratigraphic-Sedimentologic Database

2.1 Introduction

A successful basin analysis requires the collection and integration of several, perhaps many, different kinds of data. Direct observation of the rocks may or may not be fundamental to the study. In the case of a surface geological project, they will be preeminent, though perhaps supplemented by geochemical and geophysical information plus laboratory analysis of collected samples. For subsurface petroleum studies, actual rock material available for examination may be very limited, consisting of well cuttings from rotary drilling plus a few short cores. Petrophysical well logs and regional seismic lines may provide at least as important a part of the total database. Investigations for strata-bound ores and minerals typically employ networks of diamond drill holes from which a continuous core normally is available. This provides a wealth of material for analysis, although certain types of observation, such as analysis of sedimentary structures, may be difficult or impossible in such small-diameter cores. In this chapter, we discuss the collection and description of stratigraphic and sedimentologic data from outcrops and drill holes.

2.2 Describing Surface Stratigraphic Sections

2.2.1 Methods of Measuring and Recording the Data: Vertical Stratigraphic Sections

The simplest way to record the details of a surface outcrop is by measuring and describing a vertical stratigraphic section. Ideally, the location of the section should be chosen to include important stratigraphic features, such as formation contacts, but, in practice, the location is commonly determined by accessibility, e.g., the presence of bars or beaches allowing us to walk along a river cut, or a negotiable gully cutting through a cliff section. Only those geologists who use their profession as an excuse to practice their favorite sport of mountaineering will be able to apply sound geological principles to the choice of section. The rest of us take what we can reach.

In reconnaissance work, rapid measurement and description techniques are acceptable. For example, a hand-held altimeter (aneroid barometer) may be used in conjunction with dip measurements to reconstruct stratigraphic thicknesses using simple trigonometry. Another method that is commonly described in field handbooks is the pace-and-compass technique, suitable for estimating thicknesses across relatively level ground, given accurate stratigraphic dip. The same distances may be measured from maps or air photographs. Long experience with these methods has shown that they are not very reliable; errors of up to 50% can be expected.

By far the simplest and most accurate method for measuring a section is the use of a Jacob staff or "pogo stick." The stick is constructed of a 1.5 m wooden rod, with a clinometer and sighting bar (Fig. 2.1). The



Fig. 2.1. A "pogo stick" for measuring surface sections, showing the sighting bar and clinometer on the upper 50 cm of the rod

clinometer is preset at the measured structural dip and can then be used to measure stratigraphic thickness as fast as the geologist can write down descriptive notes (Fig. 2.2). The best technique is to use two persons. The senior geologist observes the rocks and makes notes, while the junior (who can be an inexperienced student) "pogos" his or her way up the section, recording increments of 1.5 m on a tally counter and collecting samples. The length 1.5 m is convenient for all but the tallest or shortest persons, although it can be awkward to manipulate on steep slopes. The only skill required by the pogo operator is the ability to visualize the dip of the strata in three dimensions across whatever terrain the geologist may wish to traverse. This is important so that the pogo can at all times be positioned perpendicular to the bedding, with the line of sight extending from the sighting bar parallel to the bedding.

It is far preferable to measure up a stratigraphic section rather than down, even though this often means an arduous climb up steep slopes. Many geologists working by helicopter in rugged terrain have made their traverses physically easier by working downhill wherever possible. But not only is it difficult to manipulate a pogo stick downward in a section, it makes it more difficult for the geologist to comprehend the order of events he or she is observing in the outcrops.

The geologist should search for the cleanest face on which to make observations. Normally, this should be weathered and free of vegetation, talus, or rain wash. Most sedimentary features show up best where they have been etched out by wind or water erosion, or where a face is kept continuously clean and polished by running water, as in a river bed or an intertidal outcrop. Such features rarely show up better on fresh fracture surfaces, so a hammer should only be used for taking samples. Carbonates may benefit from etching with dilute acid. The geologist should methodically examine both vertical cuts and the topside and underside of bedding planes; all may have something to reveal, as described later in this chapter.

It is also useful, on larger outcrops, to walk back and examine them from a distance, even from a low-flying helicopter or from a boat offshore, as this may reveal large-scale channels, facies changes, and many other features of interest. To document such large outcrops, a different technique may be used, which may conveniently be termed lateral profiling to distinguish it from vertical profiling. We discuss this in the next section.

In the interests of maximum efficiency, the geologist obviously should ensure that all the necessary measurements, observations, and sample collections are made during the first visit to a section. It may be useful to carry along a checklist to refer to each time a lithologic change in the section requires a new bed description. Many geologists have attempted to carry this process one step further by designing the checklist in the form of a computer-processible data card or by recording the data in the form of a computer code. There are three problems with this:

1. If attempts are made to record every piece of field information on the computer file, the resulting file is likely to be very large and cumbersome. Storage and retrieval programming may consume far more time than the original field work unless the geologist can draw on some pre-existing program package. This leads to the second problem.
2. If data are to be coded in the field or if a pre-existing program system is to be used, it means that decisions will already have been made about what data are to be recorded and how they are to be recorded before the geologist goes into the field. This may be satisfactory if, as a result of some previous descriptions or reconnaissance work, the geologist knows in advance what is likely to be found, but in the case of isolated field areas such foreknowledge may not be available. In this case, there is a certain risk involved in having the observation system designed in advance.
3. Individual geologists vary in their interests and in the observations they make and, of course, the

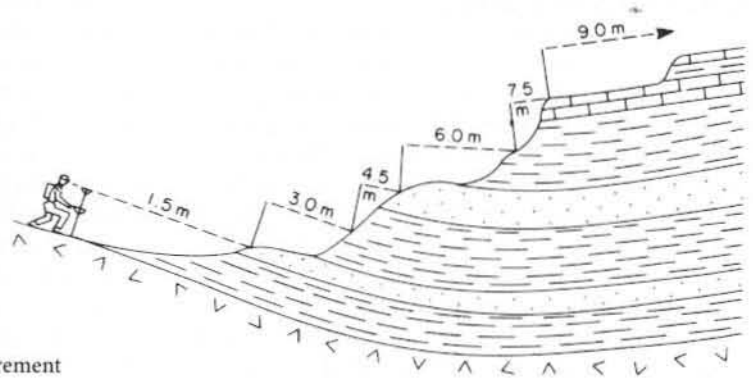


Fig. 2.2. Use of a pogo stick in section measurement

rocks are highly variable, so it is not possible to design a single, all-purpose, section-measuring software package. Specialized systems have to be designed for specific projects, and although this leads to expense in programming and debugging, it means that the program can be designed for the specific type of output required. It should not be forgotten, though, that programming, as such, is a technical skill, not a scientific one. Students and other workers who write programs for their research get few rewards for producing a workable program, only for the interesting scientific ideas their programming allows them to test.

An example of a successful computerized section-description system is that devised by P. F. Friend and his students for studying Devonian, nonmarine, clastic sequences in east Greenland (Alexander-Marrack et al. 1970; Friend et al. 1976). They designed a form with spaces to be filled in by the geologist in the field (Fig. 2.3). Marks were made in the appropriate tracks with a soft pencil, corresponding to the observed properties. These cards could be fed directly into a computer using an optical-mark page reader. The data form provides a choice of items to be filled in within each small block outlined by a solid line. Areas of the card delimited by heavy horizontal lines across the full width of the page correspond to individual bedding units recognized in the field. Items encoded on the card include thickness of unit, bedding status (exposed/covered), color, grain size, pebbles (present/absent), nature of basal contact, presence of carbonate, internal structures, and paleocurrent measurements. The range of possible observations is limited by the card. For example, the method would not be suitable if the section was found to contain thick limestone intervals because, although carbonate content can be indicated, there are no spaces to record typical carbonate sedimentary features, such as ooliths, birdseye structures, stromatolites, etc. Also, the method is suitable only for recording vertical stratigraphic changes. The presence of lateral facies changes or the exposure of a large channel could not be recorded in this system.

The particular system designed by Friend worked very well for a team of about a dozen geologists working over three field seasons, but it is unlikely that it could be applied anywhere else without modification. Unless a geologist is committed to doing much the same type of field geology for a long period, the investment of time and resources into designing and setting up such a system may not be worth the effort. In some cases, the field records must, in any case, be modified following laboratory work. This is particularly the case with carbonate sediments, which commonly are best studied in polished hand specimens or thin sections. In this case, much of the effort of the field geologist must be devoted to ensuring that a sufficiently rigorous sample-collection routine is observed. In the long run, it may be simpler to stay with an old-fashioned field notebook for field observations. Computer files can be built at a later date from corrected data using a file structure designed for a specific purpose. For example, structural data, paleocurrent data, and vertical facies-succession data lend themselves readily to processing in special-purpose computer files that can fully exploit one of the main advantages of the computer, namely its ability to carry out complex or repetitive numerical manipulations with great rapidity.

2.2.2 Methods of Measuring and Recording the Data: the Construction of Lateral Profiles

Some stratigraphic units are essentially tabular at the scale of the outcrop and can be quickly and accurately documented using vertical profiles in the way outlined in the previous section. However, some types of sedimentary assemblage contain complex facies changes that may be at a scale small enough to be observed in individual outcrops, especially in large outcrops. For example, a reef core, with its reef-front talus slope and back-reef lagoonal deposits, or a large fluvial or submarine-fan channel, with its fill of complex bar deposits, may be spectacularly displayed in a road cut or mountainside. The measurement of a few vertical sections across such an outcrop is a quite inadequate way to document the wealth of facies detail that may be available. Petroleum companies have developed a considerable interest in such large outcrops because of their use as potential analogs of subsurface reservoir units. Internal heterogeneities of these reservoir analogs and their porosity and permeability characteristics can be studied, for example by the use of minipermeameters (Miall and Tyler 1991; Doyle and Sweet 1995).

To document the details in a large outcrop, it may be necessary to construct a lateral profile, a long section that encompasses the full vertical stratigraphic height of the outcrop and also extends along the strike as far as possible to illustrate the facies changes. This may be constructed by careful surveying, but a much quicker method is to make use of photographic mosaics of the outcrop. The geologist moves well back from the exposed face, perhaps as much as several hundred meters in the case of a very large outcrop, and carries out a traverse parallel to the face, taking a series of overlapping frames until he or she has covered the entire outcrop. By taking care to remain at the same distance from the face, each frame will be at approximately the same scale. The same end can be

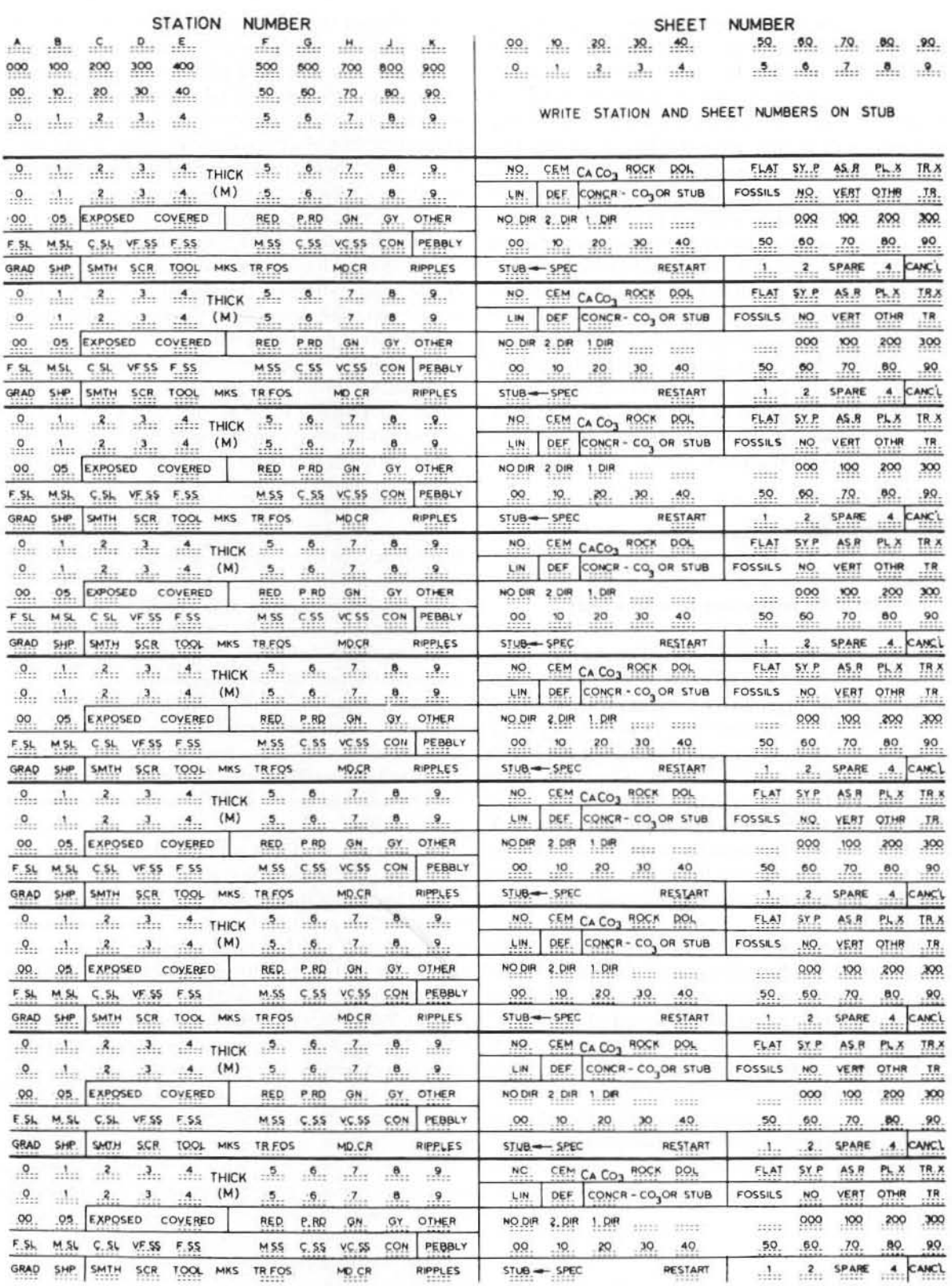


Fig. 2.3. A form for field recording of stratigraphic sections that can be processed by a computer. *Heavy horizontal lines* define the input for each descriptive unit. Items are entered by making a pencil mark in the *dashed-line tracks*. (Friend et al. 1976)

achieved by taking photographs from a boat, or even from an aircraft flying low over the outcrop.

The prints are carefully overlapped and mounted in the office in order to construct a mosaic, which can then be used in the field as a kind of topographic map base on which to enter stratigraphic and sedimentologic detail (Fig. 2.4). Some scale distortions inevitably arise. Outcrops are rarely flat, and projections and gullies will not fit together precisely in the mosaic because of the differing perspectives of adjacent frames. Such distortions are trivial relative to the immense amount of detail that can be shown on such profiles and, in any case, if accurate measurements of individual features are required, they should be made in the field and not from the photograph. The photographs may also be digitized, and can then be manipulated with a computer to adjust for scales, contrast, etc. Techniques for documenting and interpreting lateral profiles are discussed further in Section 4.5.9.

2.2.3
Types of Field Observation

2.2.3.1
Subdivision of the Section into Descriptive Units

This is a subjective operation based on the rock types present, the quality and accessibility of the exposure, and the amount of detail required in the description. Very detailed descriptions may require subdivision into units containing (for example) a single mudstone lens or crossbed set, and will therefore be on the order of a few centimeters or tens of centimeters thick. Thicker units can be defined by grouping similar rock types, but sedimentologically useful detail may be lost thereby. For each unit, the kinds of observations listed in the succeeding paragraphs are made where appropriate.

In the case of lateral profiles (Fig. 2.4), among the most valuable kinds of observation that can be made is the documentation and classification of the various kinds of bounding surfaces that separate stratigraphic units. These range from the simple bedding-plane surfaces that separate individual crossbed sets, through the surfaces that bound channels and bars, to the major (usually horizontal) surfaces that delimit mappable stratigraphic units (formations, members, stratigraphic sequences, etc.). A discussion and classification of bounding surfaces is presented in Section 4.5.9.

2.2.3.2
Lithology and Grain Size

Lithologic classification of clastic rocks can usually be done satisfactorily by visual observation in the field, without the necessity of follow-up laboratory work. Classification is based on grain size (Table 2.1), which is easily measured on the outcrop. For sand-grade rocks, it is useful to take into the field a grain size chart (Fig. 2.5) or a set of sand samples, each representing one phi-class interval through the sand-size ranges. These are used for comparison purposes and permit recognition of the main sand-grade subdivisions: very fine, fine, medium, coarse, and very coarse. Many tests by the author and others have shown that such observations provide adequate, accurate information on the modal size range of the sandstones. The description should be modified by appropriate adjectives if sorting characteristics require it; for example, pebbly coarse-grained sandstone, silty mudstone, etc. For the purpose of regional facies analysis, this is usually the only kind of grain-size information required. Skewness, kurtosis, and other statistical measures are not needed. The use of laboratory grain-size analyses is briefly discussed again in Section 4.5.1.

Siltstone and mudstone can be distinguished in a hand specimen by the presence or absence of a gritty texture, as felt by the fingers or the tongue. This is, of course, a crude method and should be checked by making thin sections of selected samples. However, field identifications of this type commonly are adequate for the purpose of facies analysis.

Fine-grained rocks, including those consisting of a mixture of sandstone, siltstone, and mudstone, are difficult to classify and describe. Dean et al. (1985) discussed the methods used in the Deep Sea Drilling Project based on smear slides of soft sediments made onboard ship.

For conglomerates, maximum clast size is often a useful parameter to measure. This is typically estimated by taking the average of the ten largest clasts visible within a specified region of an outcrop, such as a given area of a certain bedding plane. For thick conglomerate units, it may be useful to repeat such measurements over regular vertical intervals of the section. It is also important to note the degree of sorting, clast shape and roundness, matrix content, and fabric of conglomerate beds. For example, does the conglomerate consist predominantly of very well-rounded clasts of approximately the same size, or is it composed of angular fragments of varying size and shape (breccia)? Do the clasts "float" in abundant matrix, a rock type termed **matrix-supported conglomerate**, or do the clasts rest on each other with minor amounts of matrix filling the interstices – **clast-supported conglomerate**? These features are discussed at length in Chapter 4.

Carbonate rocks commonly cannot be described adequately or accurately in outcrops and require description based on thin sections or polished sections

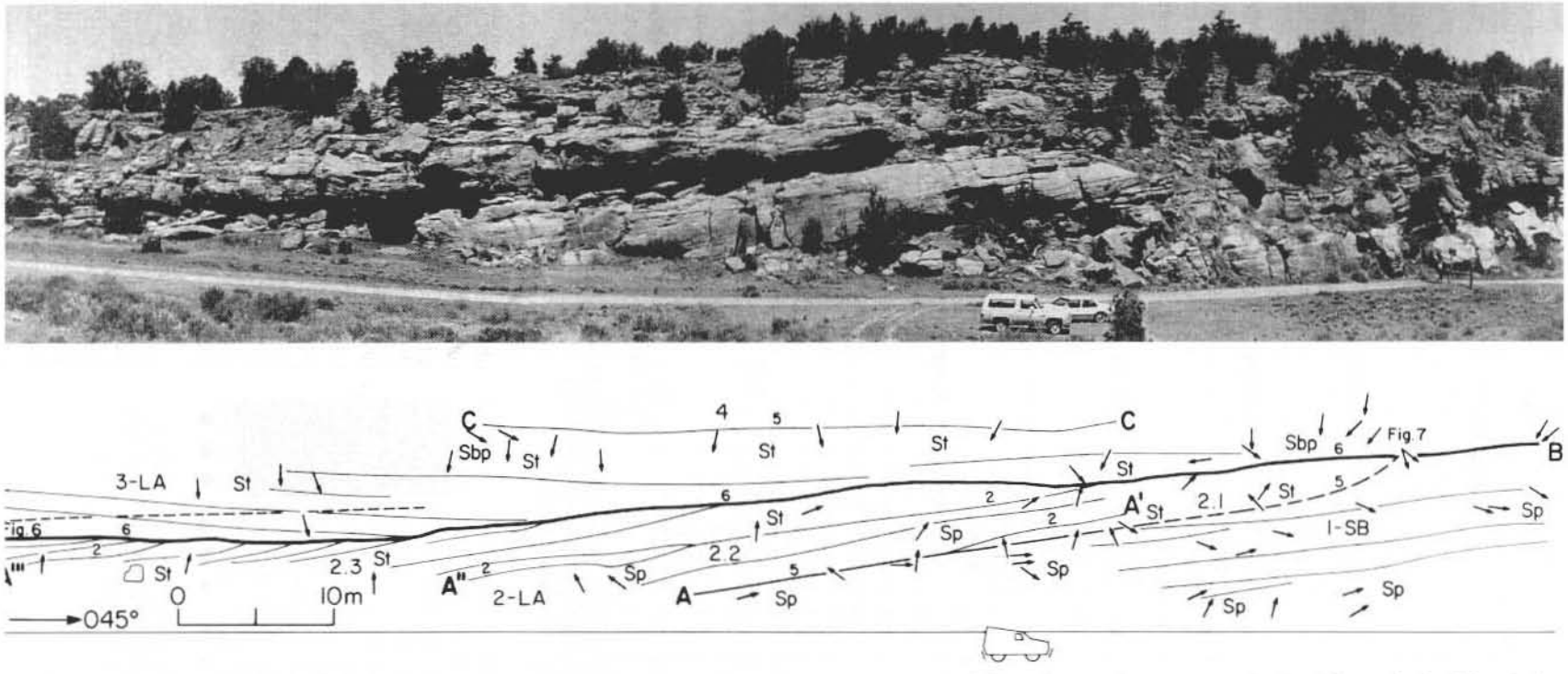


Fig. 2.4. An example of a lateral profile. *Above* is the outcrop photomosaic, and *below* is the interpretation made from photo and outcrop examination. The methods of description and interpretation of such profiles are discussed further in Section 4.5.9. This example illustrates a succession of compound bar deposits in a fluvial system. A major internal bounding surface (labeled *B*) indicates a pause in deposition and the initiation of a new channel. Beds above this surface are downlapped onto it and are interpreted as lateral accretion surfaces. Surface B may be a sequence boundary. (Miall 1991)

Table 2.1. Standard grain-size scales for carbonate and clastic sediments[a]

Limiting particle diameter				Clastic rocks			Carbonate rocks	
m	mm	ϕ units[b]					Transported constituents	Authigenic constituents
	2048	−11		V. large	Boulders	Gravel	V. coarse calcirudite	Extremely coarsely crystalline
1	1024	−10		Large				
	512	− 9		Medium				
	256	− 8		Small				
	128	− 7		Large	Cobbles			
10^{-1}	64	− 6		Small				
	32	− 5		V. coarse	Pebbles		Coarse calcirudite	
	16	− 4		Coarse				
10^{-2}	8	− 3	Micrometers (μm)	Medium			Medium calcirudite	
	4	− 2		Fine				
	2	− 1		V. fine			Fine calcirudite	Very coarsely crystalline
10^{-3}	1	0		V. coarse	Sand		Coarse calcarenite	Coarsely crystalline
	1/2	+ 1	500	Coarse				
	1/4	+ 2	250	Medium			Medium calcarenite	
	1/8	+ 3	125	Fine			Fine calcarenite	Medium crystalline
10^{-4}	1/16	+ 4	62	V. fine			V. fine calcarenite	
	1/32	+ 5	31	V. coarse	Silt		Coarse calcilutite	Finely crystalline
	1/64	+ 6	16	Coarse			Medium calcilutite	
10^{-5}	1/128	+ 7	8	Medium			Fine calcilutite	Very finely crystalline
	1/256	+ 8	4	Fine			V. fine calcilutite	
	1/512	+ 9	2	V. fine				Aphanocrystalline
10^{-6}					Clay/mud			

[a] After Friedman and Sanders, 1978; Fok, 1968. Reproduced, with permissin, from G.M. Friedman and J.E. Sanders, Principles of Sedimentology, © 1978, John Wiley and Sons, Inc., New York.

[b] Phi (ϕ) scale is given by $-\log_2$ of particle diameter.

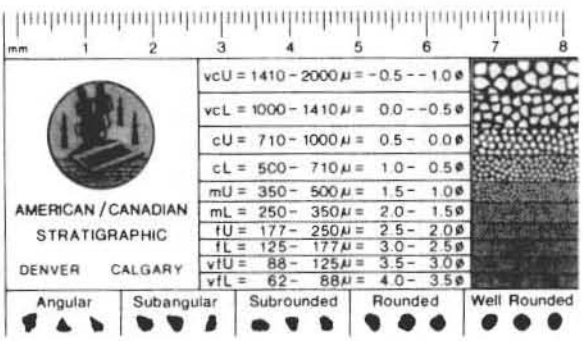


Fig. 2.5. A grain-size-comparison chart for use in the field or for logging well cuttings or cores. (Reproduced with permission of the American Stratigraphic Company)

observed under a low-power microscope. Among the reasons for this are the ready susceptibility of carbonate rocks to fine-scale diagenetic change and the fact that, in many cases, weathering behavior obscures rather than amplifies such changes, as seen in outcrops. Another important reason for not relying on outcrop observation is that some of the types of information required for carbonate facies analysis are simply too small to be seen properly with the naked eye. These include mud content, certain sedimentary textures, and biogenic features.

Field geologists traditionally take a dropper bottle of 10% hydrochloric acid with them to test for carbonate content and to aid in distinguishing limestone from dolomite (on the basis of "fizziness"). However, for research purposes, the test is quite unsatisfactory, and the geologist is advised to abandon the acid bottle (and stop worrying about leakage corroding packsacks).

Dolomite commonly can be distinguished from limestone by its yellowish weathering color in the field, but a better field test is to use alizarin red-S in weak acid solution. This reagent stains calcite bright pink but leaves dolomite unstained. In both hand specimens and thin sections, use of this reagent can reveal patterns of dolomitization on a microscopic scale.

Because of the problems associated with carbonate rocks, discussed previously, the geologist is advised not to rely on field notes for facies analysis of these rocks, but to carry out a rigorous sampling program and, using observations made on polished slabs or thin sections, supplement (and correct) the field notes. Sampling plans are discussed later in this section. Laboratory techniques for studying carbonates are described by Wilson (1975; Chap. 3).

Evaporites are difficult to study in surface outcrops. They are soft, recessive, and, commonly, poorly exposed except in arid environments. Like carbonates, they are highly susceptible to diagenetic change so that field observations must be supplemented by careful laboratory analysis.

Mixed carbonate-clastic sediments are common and are typically dealt with as if they were carbonate or clastic, which may not be the most effective way to emphasize subtle lithologic characteristics. Mount (1985) discussed the problems of classifying these rocks and suggested some methodological approaches.

2.2.3.3 *Porosity*

Porosity and permeability are of particular interest if the rocks are being studied for their petroleum potential. Observations in surface outcrops may be of questionable value because of the effects of surface weathering on texture and composition, but the geologist should always break off a fresh piece of the rock and examine the fracture surface, because such observations commonly constitute the only ones made. The geologist should distinguish the various types of porosity, such as intergranular (in detrital rocks), intercrystalline (in chemical rocks), and larger pores, such as vugs, birdseye texture, moulds of allochems, such as oolites or pellets, fossil moulds, fracture porosity, etc. Porosity types should be reported in terms of the estimated percentage they occupy in the bulk volume of the sample.

More accurate observations may be made from thin sections, and samples may be submitted to a commercial laboratory for flow tests if required. Measurements of relative permeability using minipermeametry equipment are now routinely made on outcrop profiles of reservoir analogs.

2.2.3.4 *Color*

Color may or may not be an important parameter in basin analysis. Individual lithologic units may display a very distinctive color that aids in recognition and mapping. Sometimes it even permits a formation to be mapped almost entirely from the air using helicopter observations with a minimum of ground checking. However, the sedimentological meaning and interpretation of color may be difficult to resolve.

Some colors are easily interpreted – sandstones and conglomerates commonly take on the combined color of their detrital components: pale grays and white for quartzose sediments, pinks for feldspathic sandstones, and darker colors for lithic rocks. As noted, limestones and dolomites may also be distinguished using color variations. However, color is strongly affected by depositional conditions and diagenesis, particularly the oxidation-reduction balance. Reduced sediments may contain organically derived carbon and Fe^{2+} compounds, such as sulfides, imparting green, or drab gray colors. Oxidized sediments may be stained various shades of red, yellow, or brown by the presence of Fe^{3+} compounds such as hematite and limonite. However, local reducing environments, such as those created around decaying organisms, may create localized areas or spots of reduction color. Color can change shortly after deposition as shown, for example, by Walker (1967) and Folk (1976). Moberly and Klein (1976) found that oxidation and bacterial action cause permanent color changes when fresh sediments, such as deep-sea cores, are exposed to the air. Leaching by groundwaters can also drastically change formation colors.

Thus, the problem is to decide how much time to devote to recording color in the field. Ideally, each descriptive unit in the stratigraphic section should be studied for color using a fresh rock-fracture surface and comparisons to some standard color scheme, such as the U.S. National Research Council Rock-Color Chart (Goddard et al. 1948). In practice, for the purposes of facies and basin analyses, such precision is not required. Simple verbal descriptions, such as pale gray, dark red-brown, etc., are adequate. More precise descriptions may be useful if detailed studies of diagenetic changes are to be undertaken, but recent work has shown that such studies may give misleading results if carried out exclusively on surface exposures because of the effects of recent weathering (Taylor 1978).

2.2.3.5 *Bedding*

An important type of observation, particularly in clastic rocks, is the thickness of bedding units. The

thickness relates to the rate of environmental change and to depositional energy. In some cases, bed thickness and maximum grain size are correlated, indicating that both are controlled by the capacity and competency of single depositional events. Bed-thickness changes may be an important indicator of cyclic changes in the environment, and sedimentologists frequently refer to **thinning-upward** and **fining-upward** or **coarsening-** and **thickening-upward cycles**. It is important to distinguish bedding from weathering characteristics. For example, a unit may split into large blocks or slabs upon weathering, but close examination may reveal faint internal bedding or lamination not emphasized by weathering. Bedding can be measured and recorded numerically, or it can be described in field notes semiquantitatively using the descriptive classification given in Table 2.2.

2.2.3.6 *Sedimentary Structures Produced by Hydrodynamic Molding of the Bed*

Sedimentary structures include a wide variety of primary and post-depositional features. All individually yield useful information regarding depositional or diagenetic events in the rocks, and all should be meticulously recorded and described in the context of the lithology and grain size of the bed in which they occur. The assemblage of structures and, in some cases their orientation, can yield vital paleogeographic information.

Inorganic sedimentary structures can be divided into three main genetic classes, as shown in Table 2.3. These are described briefly in order to aid their recognition in the field, but a discussion of their origin and interpretation is deferred to later chapters. Useful texts on this subject include Potter and Pettijohn (1977), Allen (1982), Harms et al. (1982), Collinson and Thompson (1982), and Ashley (1990).

Sediment carried in turbulent suspension by mass gravity-transport processes, such as debris flows and turbidity currents, is subjected to internal sorting processes. When the flow slows and ceases, the sorting may be preserved as a distinct texture termed **graded bedding**. Grading commonly consists of an upward decrease in grain size, as illustrated in Fig. 2.6; this is termed **normal grading**. However, certain sedimentary processes result in an upward increase in grain size termed **inverse grading**.

Clastic grains can be divided into two classes on the basis of their interactive behavior. **Cohesive** grains are those that are small enough that they tend to be bound by electrostatic forces and, thus, resist erosion once deposited on a bed. This includes the clay minerals and fine silt particles. A range of ero-

Table 2.2. Scale of stratification thickness[a]

Very thickly bedded	Thicker than 1 m
Thickly bedded	30–100 cm
Medium bedded	10–30 cm
Thinly bedded	3–30 cm
Very thinly bedded	1–3 cm
Thickly laminated	0.3–1 cm
Thinly laminated	Thinner than 0.3 cut

[a] From H. Blatt, G.V. Middleton, and B.C. Murray, *Origin of Sedimentary Rocks*, 2nd edn. 1980, p. 128. Reprinted by permission of Prentice-Hall, Inc., Englewood Cliffs, New Jersey, USA.

Table 2.3. Classification of inorganic sedimentary structures

I. Hydrodynamic molding of the bed
- *A. By mass gravity transport*
  - 1. Graded bedding
    - a. Normal
    - b. Reversed
- *B. By noncohesive flow*
  - 1. Lamination
  - 2. Cross-lamination (amplitude < 5 cm)
  - 3. Crossbedding (amplitude > 5 cm)
  - 4. Clast imbrication
  - 5. Primary current lineation
  - 6. Fossil orientation

II. Hydrodynamic erosion of the bed
- *A. Macroscopic*
  - 1. Scours
  - 2. Channels
  - 3. Low-relief erosion surfaces
- *B. Mesoscopic*
  - 1. Intraformational breccias
  - 2. Hardgrounds
  - 3. Lag concentrates
  - 4. Flutes
  - 5. Tool markings
  - 6. Rain prints

III. Liquefaction, load, and fluid-loss structures
- 1. Load cast
- 2. Flame structures
- 3. Ball, pillow, or pseudonodule structures
- 4. Convolute bedding
- 5. Syndepositional faults and slumps
- 6. Growth faults
- 7. Deformed crossbedding
- 8. Dish and pillar structures
- 9. Sand volcanoes
- 10. Injection features
  - a. Dikes
  - b. Mud lumps
  - c. Diapirs
- 11. Syneresis features
- 12. Desiccation cracks
- 13. Ice- and evaporite-crystal casts
- 14. Gas-bubble-escape marks
- 15. Teepee structures
- 16. Ptygmatic, enterolithic, and chicken-wire structures; gypsum/anhydrite

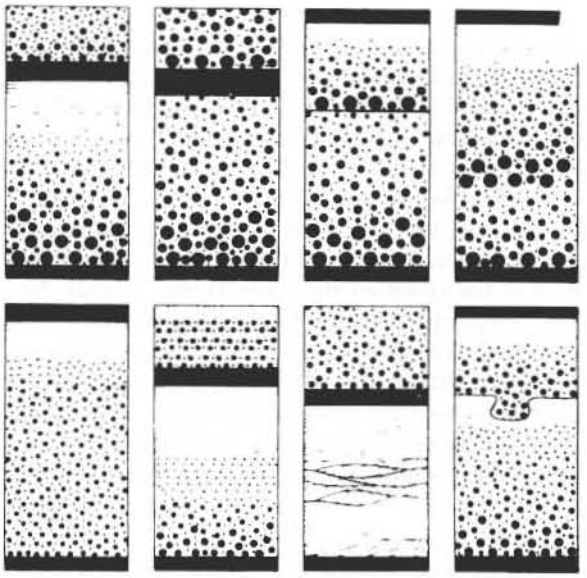


Fig. 2.6. Various types of graded bedding. (Pettijohn et al. 1972)

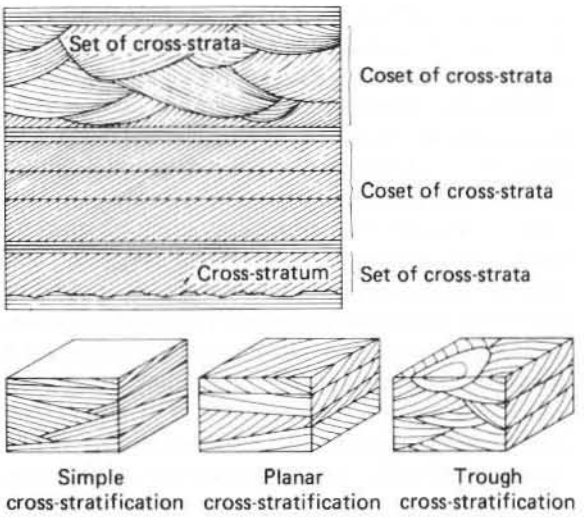


Fig. 2.7. Terminology of crossbedding. (Allen 1963)

sional sedimentary structures is present in such rocks (Table 2.3), as discussed later. Larger clastic grains of silt to cobble size, including siliciclastic, evaporite, and carbonate fragments, are **noncohesive.** They are moved by flowing water or wind as a traction carpet along the bed, or by intermittent suspension. The dynamics of movement cause the grains to be moulded into a variety of **bedforms** that are preserved as **crossbedding** within the rock.

There are three main classes of bedforms and crossbedding found in ancient rocks:

1. Those formed from unidirectional water currents, such as are found in rivers and deltas, and oceanic circulation currents in marine shelves and the deep sea.
2. Those formed by oscillatory water currents, including both wave- and tide-generated features. Although the time scale of current reversal is, of course, quite different, there are comparable features among the structures generated in these different ways.
3. Those formed by air currents. Such currents may be highly variable; the structure of the resulting deposits will be correspondingly complex. However, examination of ancient wind-formed (eolian) rocks indicates some consistent and surprisingly simple patterns (Chaps. 4, 5).

Recognition in an outcrop or in a core of the diagnostic features of these crossbedding classes is an invaluable aid to environmental interpretation (as discussed at length in Chap. 4) and, therefore, crossbedding structures must be examined and described with great care wherever they are found.

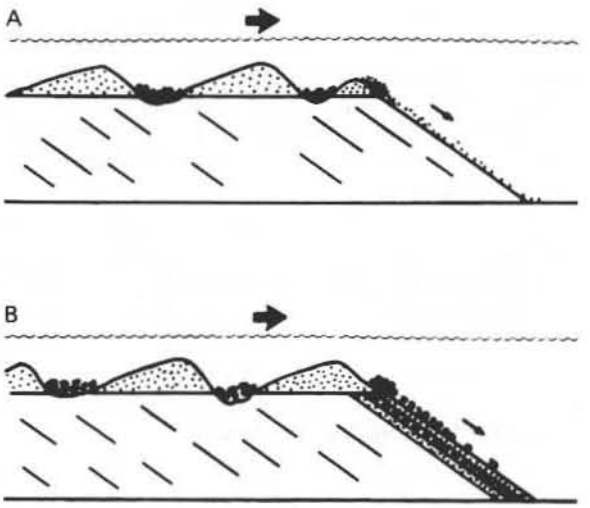


Fig. 2.8 A, B. Growth of foresets, advance of a flat-topped dune, and development of size sorting by migration of ripples along the bedform top. Note increase in foreset grain-size from (A) to (B) as the ripple crest reaches the avalanche face. (Smith 1972)

The components of crossbedding are illustrated in Figs. 2.7–2.10. A **foreset** represents an avalanche face down which grains roll or slump or are swept down by air or water currents. Figure 2.8 shows one of many types of bedform morphology, illustrating the way grains advance up the up-current (stoss) side of the bedform and are fed to a continuously advancing or prograding down-current foreset (lee) surface. In this case, the grains are transported by ripples that contain internal foreset structure. The hydrodynamics of this process are examined further in Chapter 4. Continuous deposition produces repeated foreset bedding or lamination as the bedform accretes laterally,

resulting in a crossbed **set** (Fig. 2.7; McKee and Weir 1953). A **coset** is defined as a sedimentary unit made up of two or more sets of strata or crossbedding separated from other cosets by surfaces of erosion, nondeposition, or abrupt change in character (McKee and Weir 1953). Note that a coset can contain more than one type of bedding.

When describing crossbedding, attention must be paid to seven attributes, as illustrated in Figs. 2.9 and 2.10. All but the last of these were first described in detail in an important paper on crossbedding classification by Allen (1963). In the field, crossbeds are classified first according to whether they are solitary or grouped. Solitary sets are bounded by other types of bedding or crossbedding; grouped sets are cosets consisting entirely of one crossbed type. Scale is the next important attribute. In water-laid strata, it is found that a bedform amplitude of about 5 cm is of hydrodynamic significance (Allen 1982; Ashley 1990) and, accordingly, this amplitude is used to subdivide crossbeds into small- and large-scale forms. An assumption is made that little or none of the top of a bedform is lost to erosion prior to burial; generally, the amount lost seems to increase in approximate proportion to the scale of the bedform or the thickness of the crossbed structure. Forms thinner than 5 cm are termed **ripples**, whereas forms larger than 5 cm have been given a variety of names, reflecting in part a diversity of hydrodynamic causes and in part a considerable terminological confusion. Ashley (1990) proposed several recommendations for a simplification of the terminology that have become widely ad-

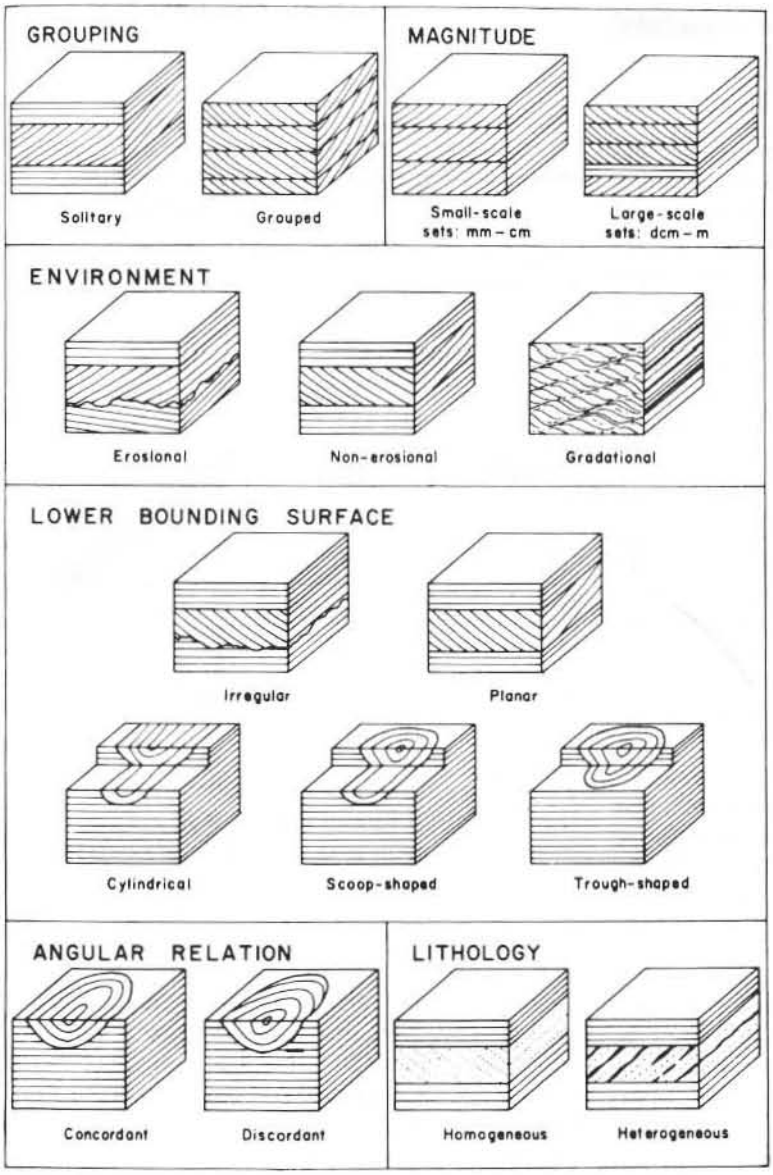


Fig. 2.9. Criteria used in the description and definition of crossbedding types. (Allen 1963)

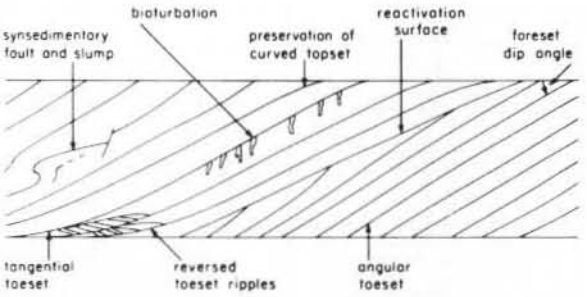


Fig. 2.10. Internal structures of crossbedding

opted. Most forms larger than ripples are now termed **dunes**. This will be discussed further in Chapter 4.

Most crossbed sets contain foresets that terminate at the base of the set, in which case the foresets are said to be **discordant**. In rare cases where the crossbeds are parallel to the lower bounding surface, as occurs in some sets with curved lower surfaces, the crossbeds are described as **concordant**.

The crossbeds may show either homogeneous or heterogeneous lithology. Homogeneous crossbeds are those composed of foresets whose mean grain size varies by less than two phi classes. Heterogeneous (or heterolithic) crossbeds may contain laminae of widely varying grain size, including interbedded sand and mud or sand and gravel, possibly even including carbonaceous lenses.

The minor internal structures within crossbeds are highly diagnostic of their origin (Fig. 2.10). The dip angle of the foreset relative to the bounding surface is of considerable dynamic significance. Are the foresets curved, linear, or irregular in sections parallel to the dip? Is the direction of dip constant, or are there wide variations or reversals of dip within a set or between sets? Do the sets contain smaller-scale hydrodynamic sedimentary structures on the foresets and, if so, what is the dip orientation of their foresets relative to that of the larger structure? What is the small-scale internal geometry of the foresets – are they tabular, lens, or wedge shaped? Do they display other kinds of sedimentary structures, such as trace fossils, synsedimentary faults, or slumps? Are **reactivation surfaces** present? These represent minor erosion surfaces on bedforms that were abandoned by a decrease in flow strength and then reactivated at some later time.

These seven attributes can be used to classify crossbed sets in the field. It is time consuming to observe every attribute of every set, but it is usually possible to define a limited range of crossbed types that occur repeatedly within a given stratigraphic unit. These can then be assigned some kind of local unique descriptor, enabling repeated observations to be recorded rapidly in the field notebook. One way to classify crossbed sets is to use a published classification system, of which that erected by Allen (1963) is the most complete. This is illustrated in Fig. 2.11. However, this classification does not encompass any of the internal structures illustrated in Fig. 2.10. Allen's classification is no longer widely used, although a few of his Greek-letter names are well known among sedimentologists and are used as a convenient shorthand. Examples of crossbedding in outcrops are illustrated in Fig. 2.12. Modern methods of facies classification that encompass variations in crossbed type are discussed in Section 4.3, and some other types of crossbedding not illustrated here are described in Section 4.5.4.

A vital component of basin analysis is an investigation of sedimentological trends, such as determining the shapes and orientations of porous-rock units. **Paleocurrent analysis** is one of several techniques for investigating sedimentary trends based, among other attributes, on the size, orientation, and relative arrangement of crossbedding structures. Therefore, when describing outcrop sections, it is essential to record the orientation of crossbed sets. The procedures for doing this and the methods of interpretation are described in Section 5.9.

Crossbedding represents a macroscopic orientation feature, but each clastic grain is individually affected by a flow system and may take up a specific orientation within a deposit in response to flow dynamics. The longest dimension of elongated particles tends to assume a preferred position parallel or perpendicular to the direction of movement and is commonly inclined upflow, producing an imbricated or shingled fabric. This fabric may be present in sand-sized grains and can be measured optically, in thin section (Martini 1971) or using bulk properties such as dielectric or acoustic anisotropy (Sippel 1971). In recent years, paleomagnetic data have also been recognized to contain much useful information relating to primary sedimentary fabrics. Eyles et al. (1987) discussed magnetic orientation and anisotropy data with reference to the depositional processes of till and till-like diamict deposits. Oriented specimens must be collected in the field for such an analysis (Sect. 2.2.4). In conglomerates, an **imbrication** fabric commonly is visible in an outcrop and can be readily measured by a visual approximation of average orientation or by laborious individual measurements of clasts. Fig. 2.13 illustrates imbrication in a modern river bed. As discussed in Section 5.9, it has been found that in nonmarine deposits, in which imbrication is most common, the structure is one of the most accurate paleocurrent indicators (Rust 1975).

In sand-grade material, grain sorting is responsible for generating another type of fabric that is also an excellent paleocurrent indicator. This is **primary**

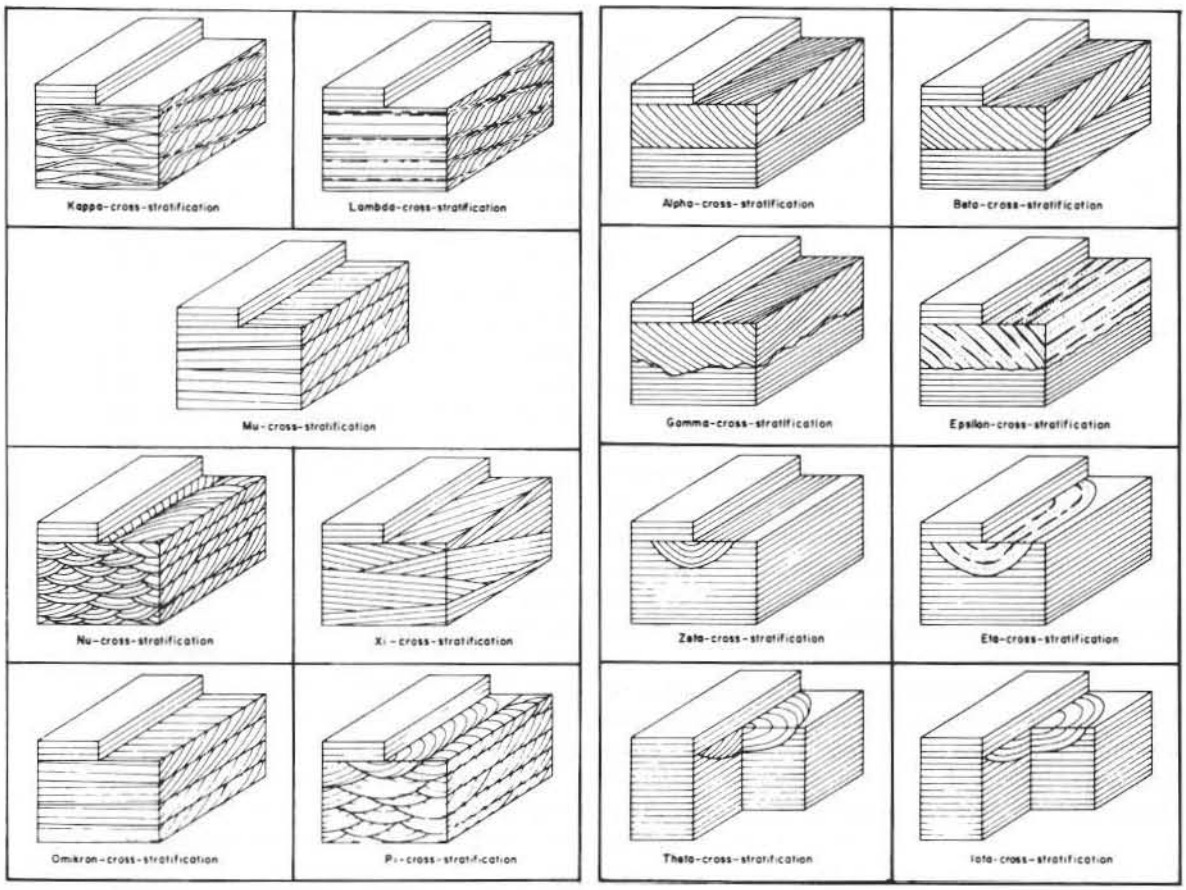


Fig. 2.11. The 15 crossbedding types defined by Allen (1963), distinguished using the criteria set out in Fig. 2.9

current lineation, also termed **parting lineation** because it occurs on bedding-plane surfaces of sandstones that are flat bedded and usually readily split along bedding planes. An example is illustrated in Fig. 2.14. Primary current lineation is the product of a specific style of water turbulence above a bed of cohesionless grains, as are the various bedforms that give rise to crossbedding (Sect. 4.5.4). Therefore, it has a specific hydrodynamic meaning and is useful in facies analysis as well as paleocurrent analysis.

Rather than the bed itself, objects such as plant fragments, bones, or shells may be oriented on a bedding plane. This should be observed, if possible, but interpretation commonly is not easy, as discussed in Section 5.9.

2.2.3.7
Sedimentary Structures Produced by Hydrodynamic Erosion of the Bed

A wide variety of erosional features is produced by water erosion of newly deposited sediment. These result from changes in water level or water energy in response to floods, storms, tides, or wind-driven waves and currents. They can also result from evolutionary change in a system under steady equilibrium conditions. These processes result in the development of various types of **bounding surfaces** in the rocks. Recognition and plotting of these features in outcrop sections are important components of facies analysis and, with adequate exposure, orientation studies may contribute significantly to the analysis of depositional trends.

These features range in size up to major **river** and **tidal channels**, **submarine canyons**, and **distributary channels** several kilometers across and tens or hundreds of meters deep, but large features such as these can rarely be detected in the average small outcrop. They may be visible in large outcrop sections, where they can be documented using lateral profiles (Chap. 4), and on seismic sections (Chap. 6), and it may be possible to reconstruct them by careful lithostratigraphic correlation and facies analysis of scattered outcrops, but this is beyond the scope of our immediate discussion. At the outcrop scale, there are two types of small-scale erosional features to discuss, those that truncate one or more bedding units and

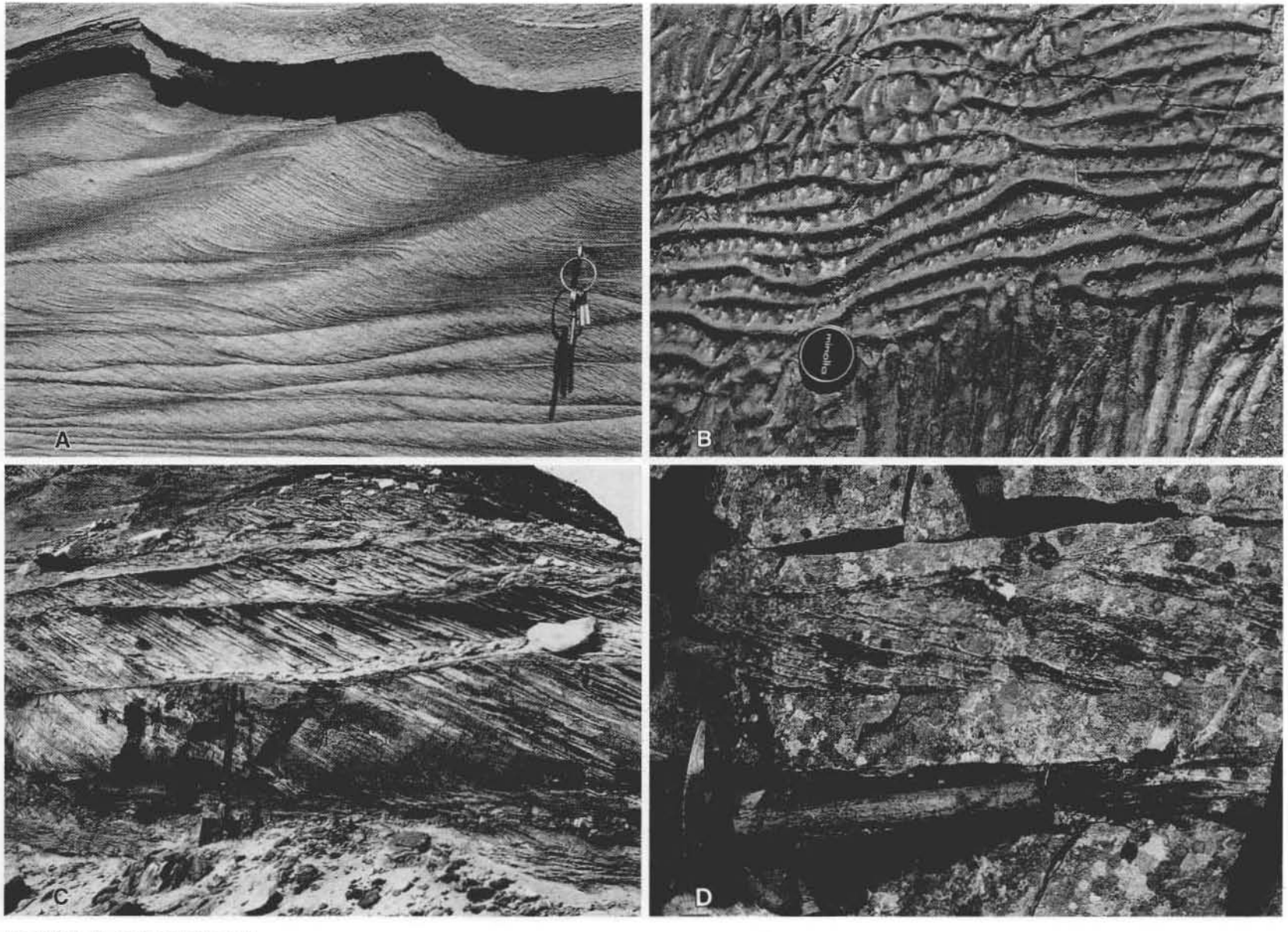


Fig. 2.12 A – D. see p. 38 for legend

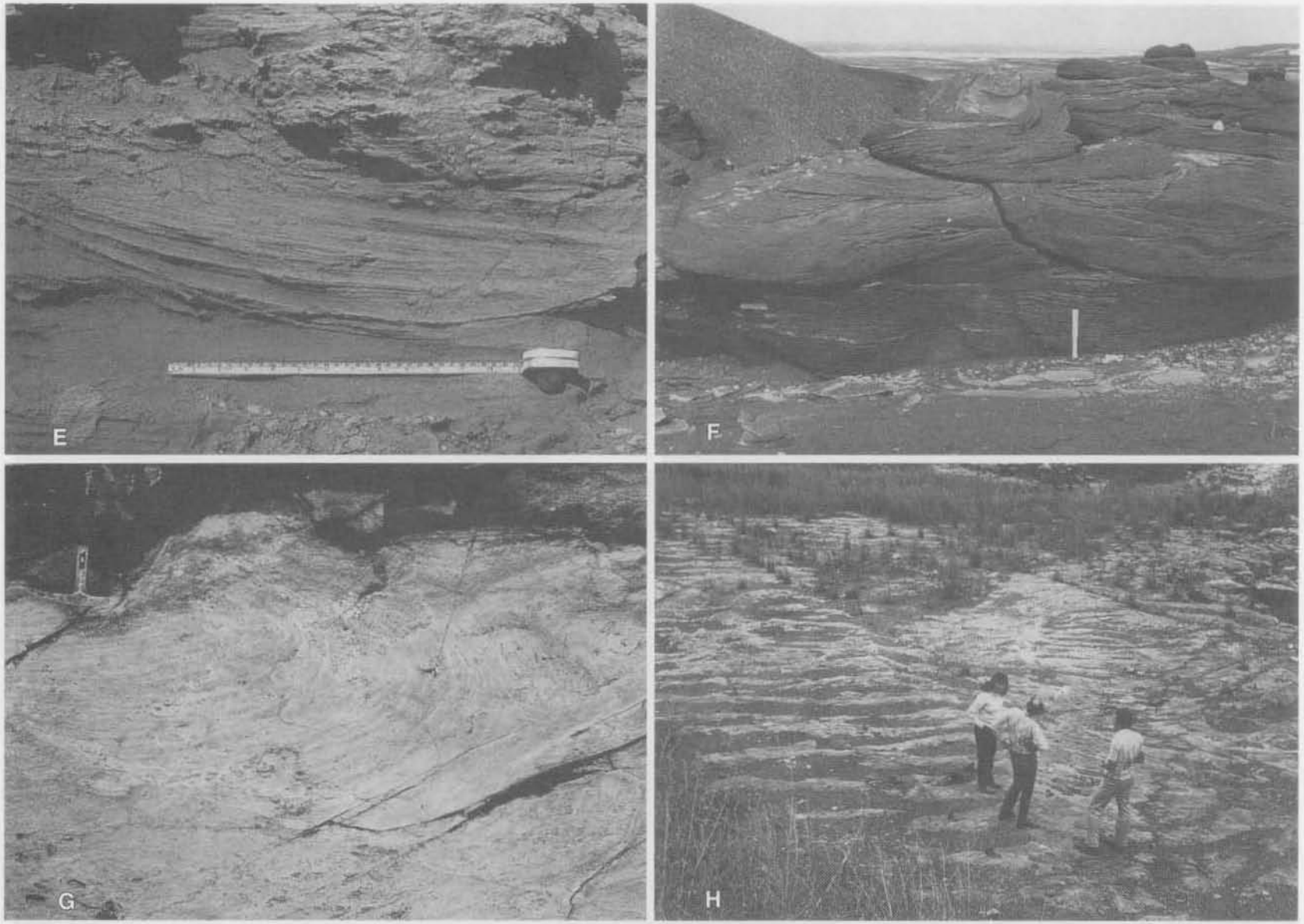


Fig. 2.12 E–H. see p. 38 for legend

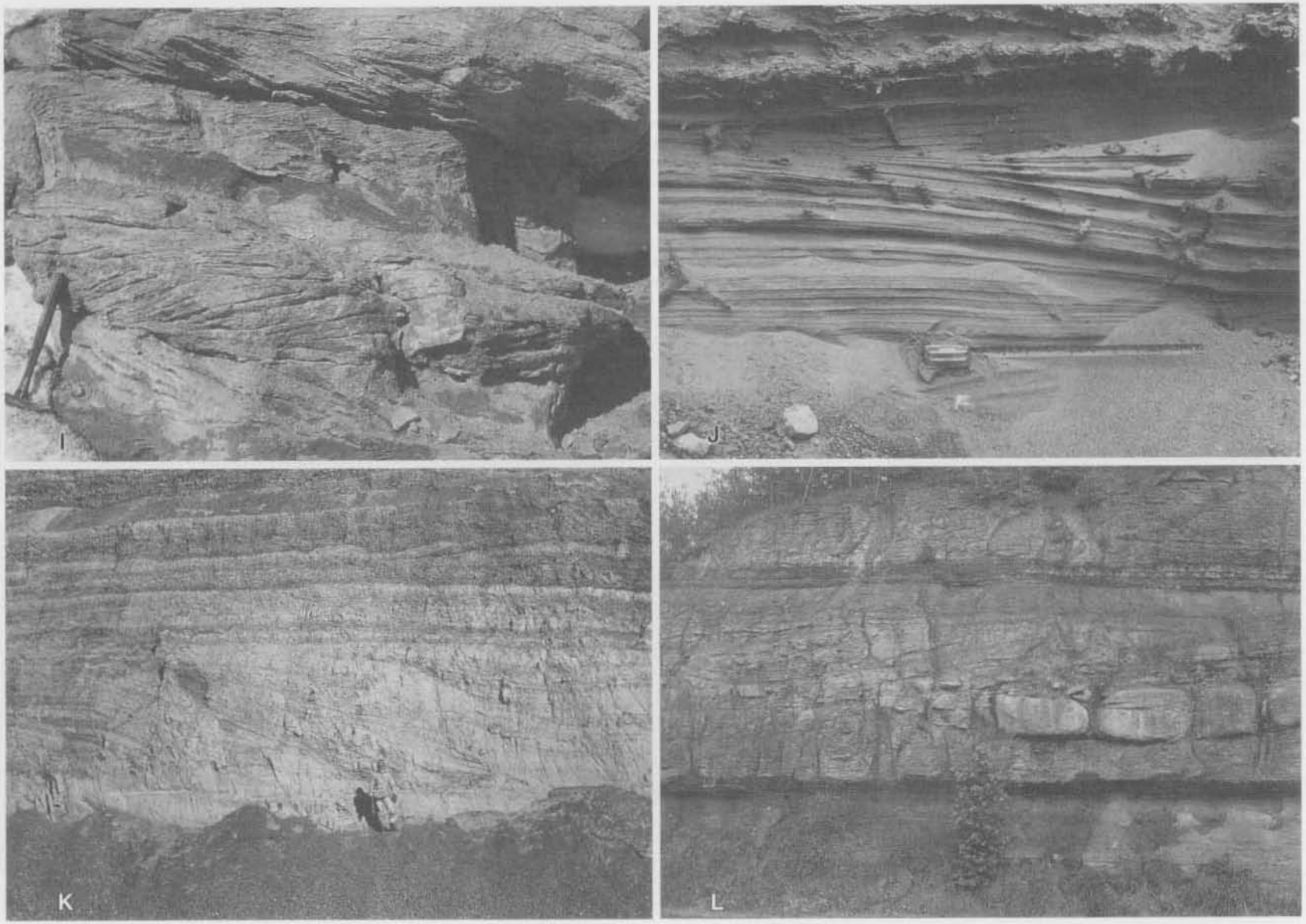


Fig. 2.121 – L. see p. 38 for legend



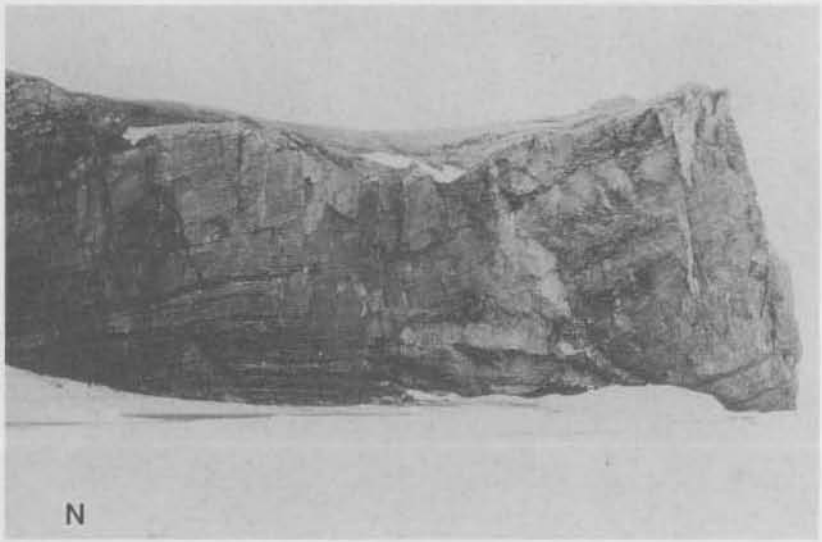


Fig. 2.12 A – N. Types of crossbedding, as observed in outcrops. **A** Climbing ripple crosslamination. Note the increase in the angle of climb, a common feature that indicates decrease in flow strength and increase in detritus settling from suspension. Fluvioglacial environment, Pleistocene, Ottawa. **B** Ladderback ripples. An older ripple set is visible as short segments of ripple crest preserved in the troughs of a younger set formed in a perpendicular direction. Fluvial environment, Karoo Supergroup, Beaufort West, South Africa. **C** A coset of planar cross-strata (alpha-type crossbedding of Fig. 2.11). Note the consistency of orientation. Fluvial environment, Cretaceous, Banks Island, Arctic Canada. **D** Planar crossbed set in a clastic limestone. Foresets are emphasized by chert lenses. Shallow marine environment, Mississippian, northeast Alaska. **E** Oblique section through a solitary trough crossbed set with trace fossils. Shallow marine environment, Cretaceous, Banks Island, Arctic Canada. **F** A coset of trough cross-strata (pi-type crossbedding of Fig. 2.11). Fluvial environment, Devonian, Somerset Island, Arctic Canada. **G** Bedding-plane view of a trough crossbed. The curved foresets dip toward the left. Fluvial environment, Huronian Supergroup (Proterozoic), near Elliot Lake, Ontario. **H** Bedding-plane exposure of straight-crested megaripples. Shallow marine environment, Karoo Supergroup, near Durban, South Africa. **I** Herringbone crossbedding: five planar crossbed sets showing reversals of flow direction. Shallow marine environment, Permian, Ellesmere Island, Arctic Canada. **J** Low-angle, curved crossbedding, possibly hummocky cross-stratification, with the trace fossil *Ophiomorpha*. Shallow marine environment, Eocene, Banks Island, Arctic Canada. **K** Large-scale crossbedding (epsilon-type crossbedding of Fig. 2.11). Fluvial point bar, Eocene, Ellesmere Island, Arctic Canada. **L** Large-scale crossbedding (epsilon-type crossbedding of Fig. 2.11). Fluvial point bar, Carboniferous, Alabama. **M** Large-scale eolian crossbedding. Jurassic Navajo Formation, Zion National Park, Utah. **N** Talus slope of a reef composed of coarse, crinoidal calcarenite and calcirudite. Devonian, Princess Royal Islands, Arctic Canada



Fig. 2.13. Clast imbrication in a modern river, indicating flow from left to right. Ellesmere Island, Arctic Canada

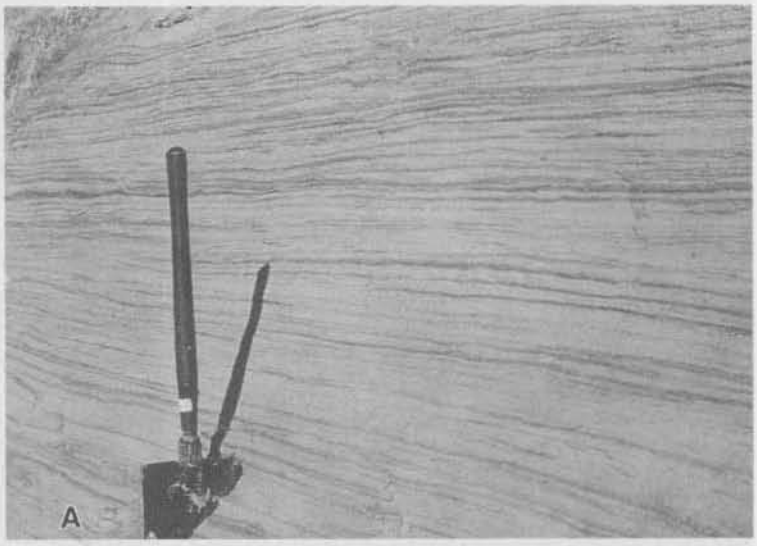




Fig. 2.14. **A** Plane bedding in sandstone containing dark, sand-sized, comminuted carbonaceous debris. Eocene, Ellesmere Island, Arctic Canada. **B** Bedding-plane exposure of a sandstone such as that in **A**, showing parting lineation. upper Proterozoic, Banks Island, Arctic Canada

those that scour or pit the bedding plane without significantly disrupting it.

The first type includes, in decreasing order of scale, **channels**, **scours**, low-relief **erosion surfaces**, and **rill markings**. Typical examples are illustrated in Fig. 2.15. These may be classified as macroscopic erosion features (Table 2.3). Channels and scours are usually filled by sediment that is distinctly different in grain size and bedding characteristics from that into which the channel is cut. The channel fill is almost invariably coarser than the eroded strata, indicating, as might be expected, that the generation of the channel was caused by a local increase in energy level. Fig. 2.15A illustrates an exception to this, where a channel was abandoned and subsequently filled by fine sediment and coal under low-energy conditions.

It is a common error to confuse trough crossbedding with channels. Troughs are formed by the migration of trains of dunes or sinuously crested megaripples (Sect. 4.5.4). They rest on curved scour surfaces, but these are not channels. The scours are formed by vortex erosion in front of the advancing dunes and are filled with sediment almost immediately. Channels, on the other hand, may not be filled with sediment for periods ranging from hours to thousands of years after the erosion surface is cut, and so the cutting and filling of the channel are quite separate events.

Erosion surfaces may exhibit little erosional relief, which may belie their importance. In the nonmarine environment, sheet erosion, wind deflation, and pedimentation can generate virtually planar erosion surfaces. In subaqueous environments, oceanic currents in sediment-starved areas, particularly in abyssal depths, can have the same result. Exposed carbonate terrains may develop **karst** surfaces with extensive cave systems. At the outcrop scale, careful examination of erosion surfaces may reveal a small-scale relief and the presence of features such as infilled **desiccation cracks**, basal intraformational or extrabasinal **lag gravels** (Fig. 2.15E), fissures filled with sediment from the overlying bed, zones of bioclastic debris, etc. In some subaerial environments, soil or weathering profiles may have developed, including the development of **caliche** or **calcrete** (Fig. 2.15I, J) and the presence of surfaces of nondeposition. In carbonate environments, surfaces of nondeposition commonly develop subaqueously. **Hardgrounds** are organically bored surfaces that may be encrusted with fossils in growth positions. Alternatively, they may be discolored by oxidation, giving a red stain, or blackened by decayed algal matter. Long-continued winnowing of a surface of nondeposition may leave **lag concentrates** or **condensed sections** consisting of larger particles blackened by algal decay and possibly including abundant phosphatized fossil material (Fig. 2.15F; Wilson 1975). Condensed sections may be generated by rapid transgression, which forces coastal sediment sources to undergo retreat (Chap. 6). In continental-slope deposits, giant slumps and slides are common and are particularly well exposed as **intraformational truncation surfaces** in deep-water carbonate sediments. An example is illustrated in Fig. 2.15K.

It may be difficult to assess the length of time missing at erosion surfaces – some may even represent major time breaks detectable by biostratigraphic zonation. In any case, a careful search for and description of such features in the field is an important part of section description. The recognition and mapping of erosion surfaces is a critical element in the study of sequence stratigraphy, as discussed at greater length in Chapter 6.

Mesoscopic erosional features fall mainly into a class of structure termed **sole markings**. These are features seen on the underside of bedding planes, usually in sandstones, and they represent the natural casts of erosional features cut into the bed below, which is typically siltstone or mudstone. They attest to the erosive power of the depositional event that formed the sandstone bed but, beyond this, most have little facies or environmental significance. However, they can be invaluable paleocurrent indicators.

Flute markings are formed by vortex erosion, typically at the bases of turbidity currents, although they have also been observed at the bases of fluvial channels. Erosion is deepest at the up-current end of the scour and decreases down current so that, in a flute cast, the high-relief nose of the cast points up current. In rare examples, vortex flow lines may be perceived in the walls of the flute (Fig. 2.16A). Flutes generally are on the order of a few centimeters deep.

Tool markings are a class of sole structure formed by the erosional impact of large objects entrained in the flow, including pebbles, plant fragments, bone, and shell material. The many varieties that have been observed have been assigned names that indicate the interpreted mode of origin. They include **groove, drag, bounce, prod, skip, brush**, and **roll markings**. A few examples are illustrated in Fig. 2.16B, which shows the strongly linear pattern on the bed, thus providing excellent paleocurrent indicators.

2.2.3.8 *Liquefaction, Load, and Fluid-Loss Structures*

Clay deposits saturated with water are characterized by a property termed **thixotropy**: when subjected to a sudden vibration, such as that generated by an earthquake, they tend to liquify and lose all internal strength. This behavior is responsible for generating a variety of structures in clastic rocks. Clay beds commonly are interbedded with sand or silt and, when liquefied, the coarser beds have a higher density than

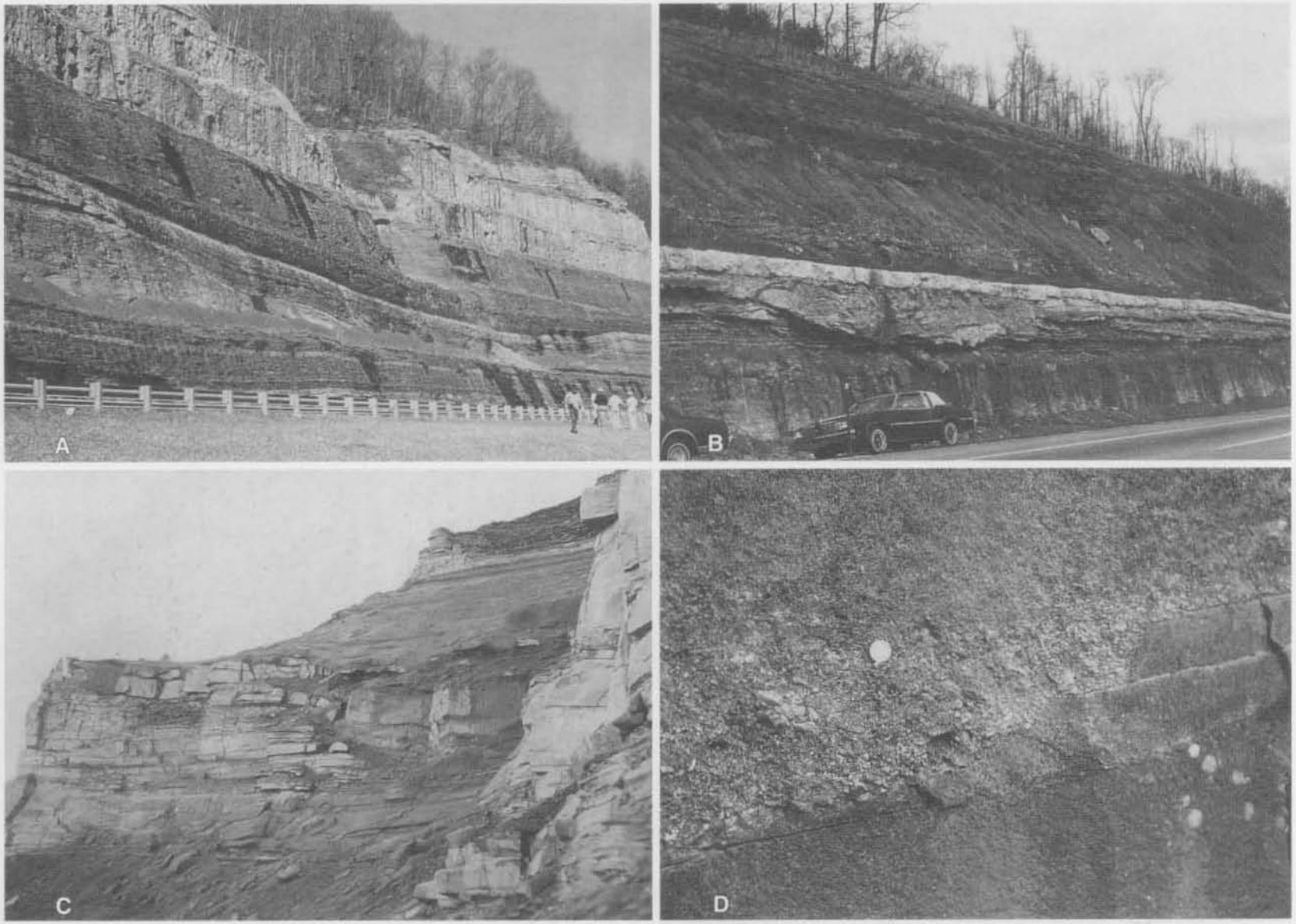


Fig. 2.15 A – D. see p. 42 for legend





Fig. 2.15 A–K. Macroscopic erosional features. **A** Abandoned fluvial channel showing levees and fine-grained fill. Carboniferous, Kentucky. **B** Tidal channel filled with calcarenite, cut into finer-grained tidal-flat deposits. Carboniferous, Kentucky. **C** A fluvial cutbank. Tertiary, Bylot Island, Arctic Canada. **D** Close up of a stepped scour surface at the base of a fluvial channel. Permian, southern Poland. **E** Intertidal carbonate mudflat deposits containing several scour surfaces covered by bioclastic debris (*above*) and a desiccated layer (*below*) showing incipient brecciation. Compare the latter with photo **G**. Silurian, Somerset Island, Arctic Canada. **F** Discontinuity surface with manganiferous nodules and crusts. Lower Cretaceous, Provence, France. **G** Intraformational breccia produced by breakup and reworking of a desiccated dolomite bed on a tidal flat. Ordovician, Somerset Island, Arctic Canada. **H** Rill marks produced by water seepage from an exposed fluvial bar. Karoo Supergroup, Beaufort West, South Africa. **I** Lowermost bed is caliche formed by subaerial weathering of a shallow marine carbonate bed. It is followed by a coarse-grained lag deposit formed by a storm (*center* of view, below overhanging bed). Carboniferous, Kentucky. **J** A caliche breccia. Cretaceous, Provence, France. **K** A slide surface overlain by a large slumped mass of carbonate sediments. Cretaceous, Provence, France

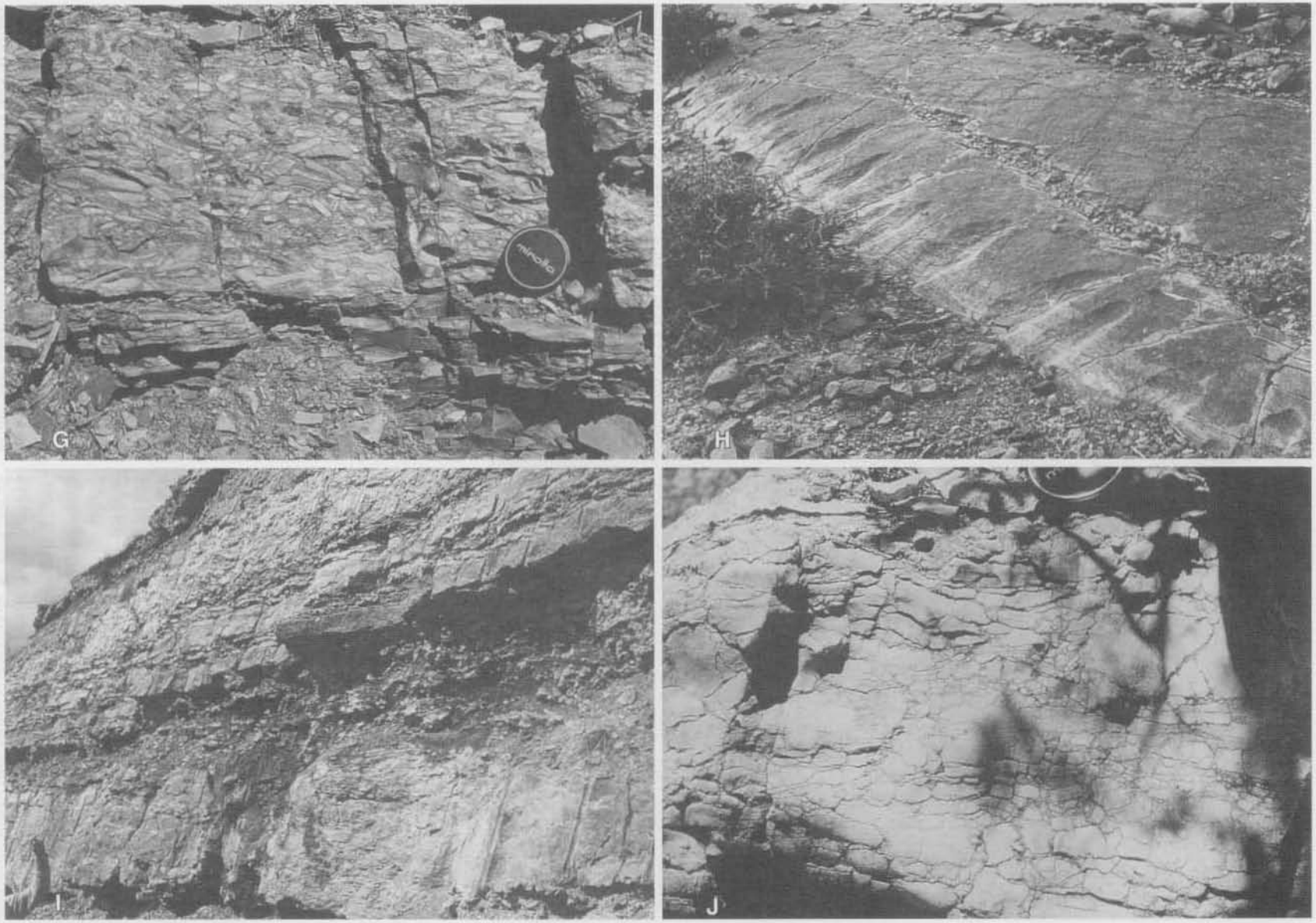


Fig. 2.15G–J. see p. 42 for legend

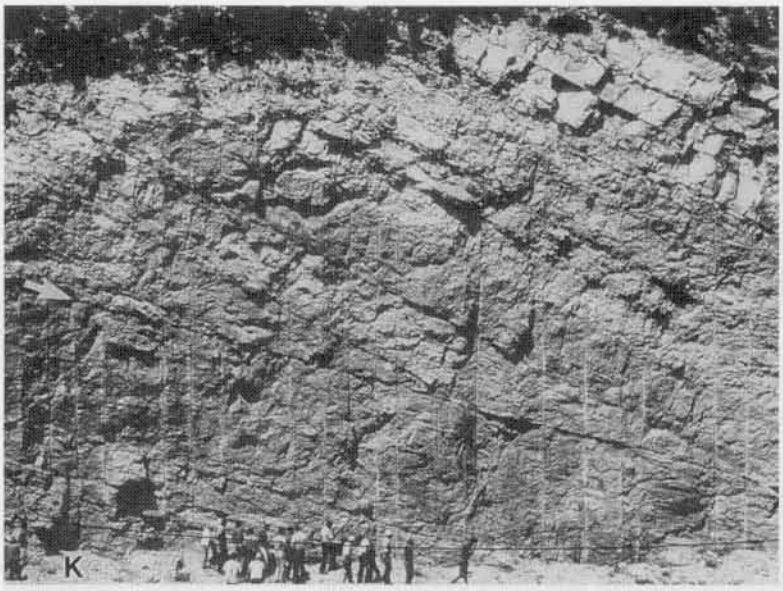


Fig. 2.15K. see p. 42 for legend

the clay and tend to founder under gravity. This may or may not result in the disruption of the sand units. Where complete disruption does not take place, the sand forms bulbous shapes, termed **load structures**, that project into the underlying clay. These are usually best seen in ancient rocks by examining the underside of a sandstone bed (Fig. 2.16B). They are, therefore, a class of sole structure, though one produced without current movement. Clay wisps squeezed up between the load masses form pointed shapes termed **flame structures** because of a resemblance to the shape of flames. These are best seen in cross section. Occasionally, loading may take place under a moving current, such as a turbidity flow, and load structures may then be stretched, possibly by shear effects, into linear shapes paralleling the direction of movement.

Commonly, load masses become completely disrupted. The sand bed may break up into a series of ovate or spherical masses that sink into the underlying bed and become surrounded by mud. Lamination in the sand is usually preserved in the form of concave-up folds truncated at the sides or tops of each sand mass, attesting to the fragmentation and sinking of the original layer. Various names have been given to these features, including **ball**, **pillow**, and **pseudonodule structures** (Fig. 2.17H). These structures rarely have a preferred shape or orientation and are not to be confused with slump structures, which are produced primarily by lateral rather than vertical movement.

In many environments, sediment is deposited on a sloping rather than a flat surface (large-scale examples of this are visible on seismic reflection surveys and are called **clinoforms** – see Chaps. 5, 6), such as the subaqueous front of a delta, which is something like a very large-scale foreset built into a standing body of water. The difference is that, once deposited, such material usually does not move again as individual grains because the angle of the slope is too low. Typically, large-scale submarine fans and deltas exhibit slopes of less than 2°. However, the sediment in such environments is water saturated and has little cohesive strength. Therefore, slopes may become oversteepened, and masses of material may be induced, by shock-induced failure, to slump and slide downslope. Undoubtedly, the thixotropic effects described previously facilitate this process. The result is the production of internal shear or glide surfaces and deformed masses of sediment, termed **slump structures**. Failure surfaces may be preserved as **syndepositional faults**. Small-scale examples of these features are commonly seen in outcrops, as illustrated in Figs. 2.17I and J. Some examples of **convolute bedding** may be produced this way, although others are the result of water escape, as discussed below.

Structures produced by failure and lateral movement commonly retain an internal orientation with a simple geometric relationship to the orientation of the depositional slope. This could include the elongation of slump masses and the orientation (strike) of slide surfaces (both parallel to depositional strike) or the asymmetry, even overturning, of folds in convolute beds. Recognition of these geometric properties in an outcrop is important because it helps distinguish structures from those of different origin, and orientation characteristics obviously have potential as paleoslope indicators.

Very-large-scale slumps and syndepositional faults are developed on major deltas. The latter are

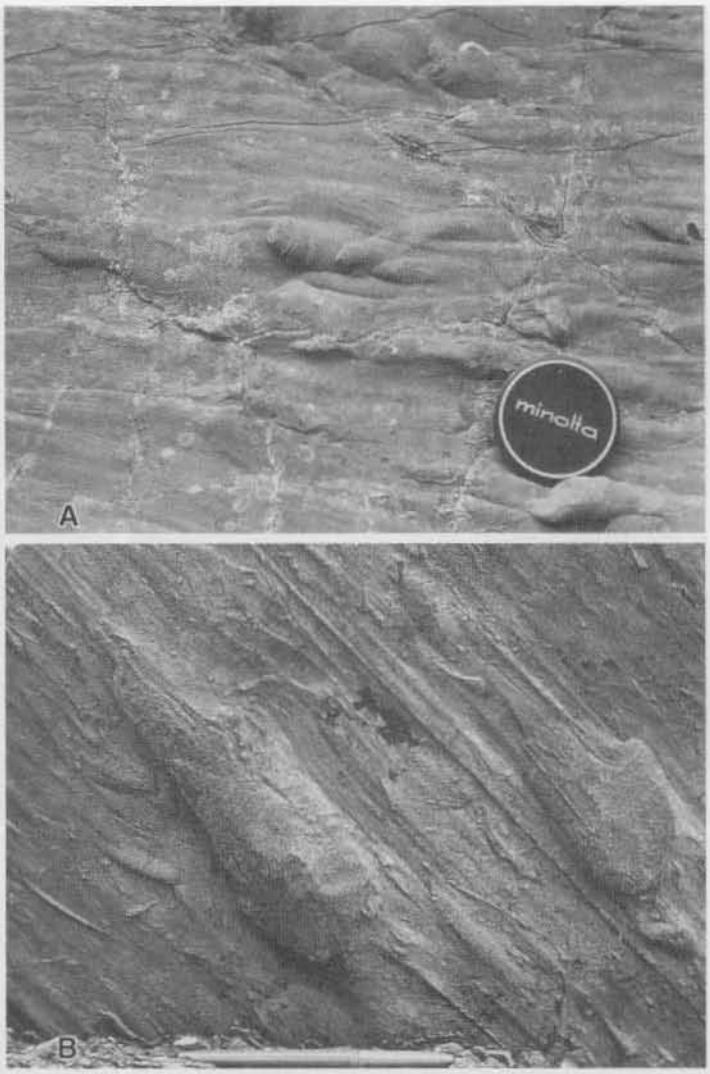


Fig. 2.16 A, B. Sole markings. **A** Flute markings. Note vortex flow lines on the large flute near the *center* of view. Cretaceous, Provence, France. **B** Tool markings, mainly groove casts and load casts. Cretaceous, Provence, France

termed **growth faults** (Chaps. 6, 9). **Olisthostromes** are giant slumps developed on tectonically active continental slopes (Chap. 9).

Deformed or **overturned crossbedding** (Fig. 2.17K) is developed in saturated sand beds by the shearing action of water or turbid flow across the top of the bedform. The upper few centimeters of the crossbedded unit move down current by a process of intragranular shear, and foreset lamination is overturned as a result, producing an up-current dip. Obviously, to produce this structure the shearing current must have a similar orientation to that of the current that generated the crossbedding. Deformed crossbedding is common in fluvial and deltaic environments.

As additional sediment is laid on top of saturated deposits, grains within the substrate begin to settle and pack more tightly. Pore waters are expelled in this process and move upward or laterally to regions of lower hydrostatic pressure. Eventually, they may escape to the surface. This process may take place slowly if sediment is being deposited grain by grain and the fluid movement leaves little or no impression on the sediments. However, if loading is rapid, a much more energetic process of fluid loss takes place, and the sediment itself will be moved around in the process. The result, in sand-grade deposits, is a group of features called **dish** and **pillar structures** (Fig. 2.17A). Dishes are produced by escaping water breaking upward through a lamination and turning up the edges; pillars record the vertical path of water flow moving to the surface. Dishes may be up to 50 cm in diameter. These structures are particularly common in the deposits of sediment gravity flows, such as fluidized flows, in which sediment emplacement is rapid. They are produced by water escape as a flow ceases movement, the loss of the lubricating effect of water itself being one of the main reasons why the flow stops.

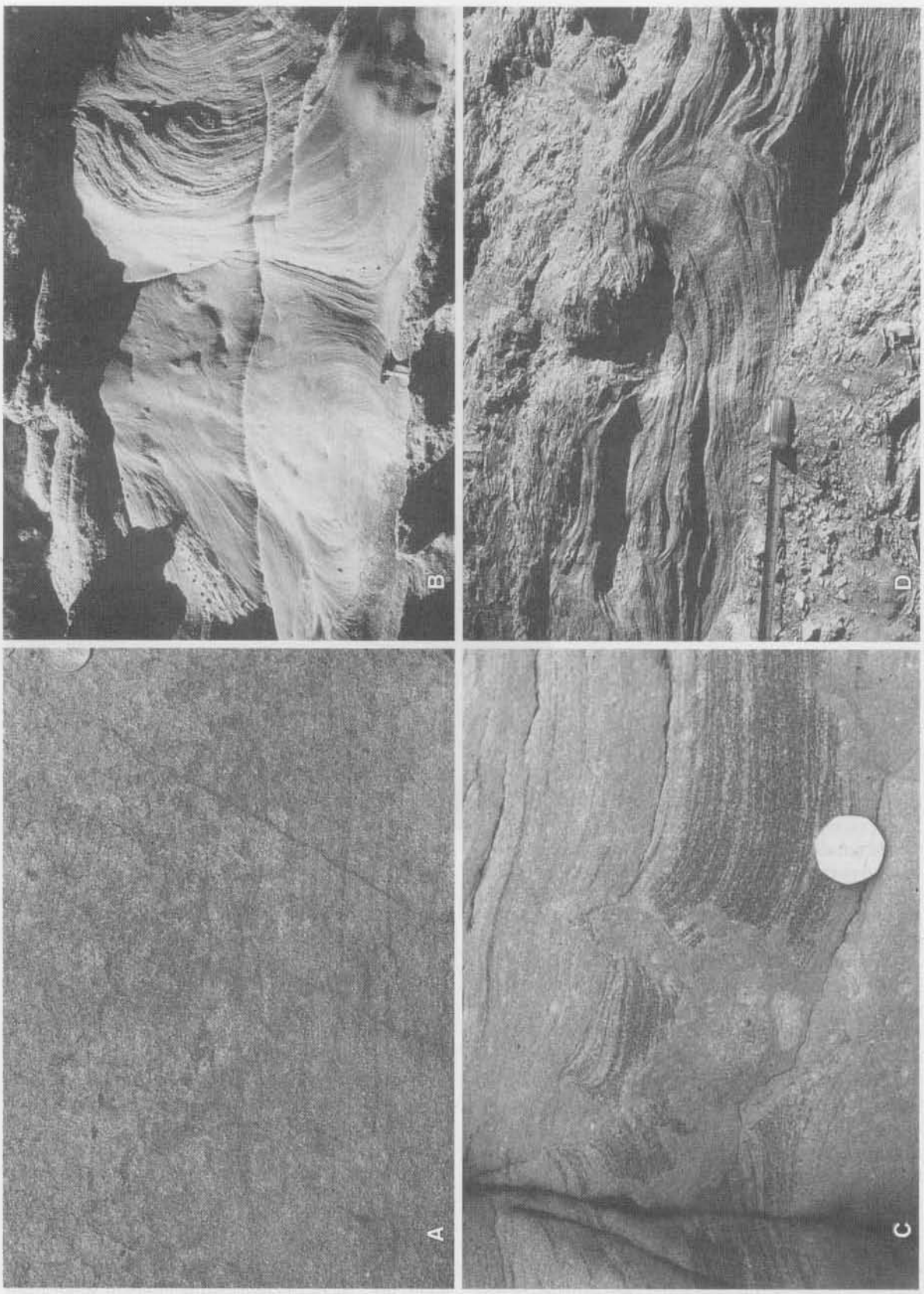


Fig. 2.17 A–D. see p. 49 for legend

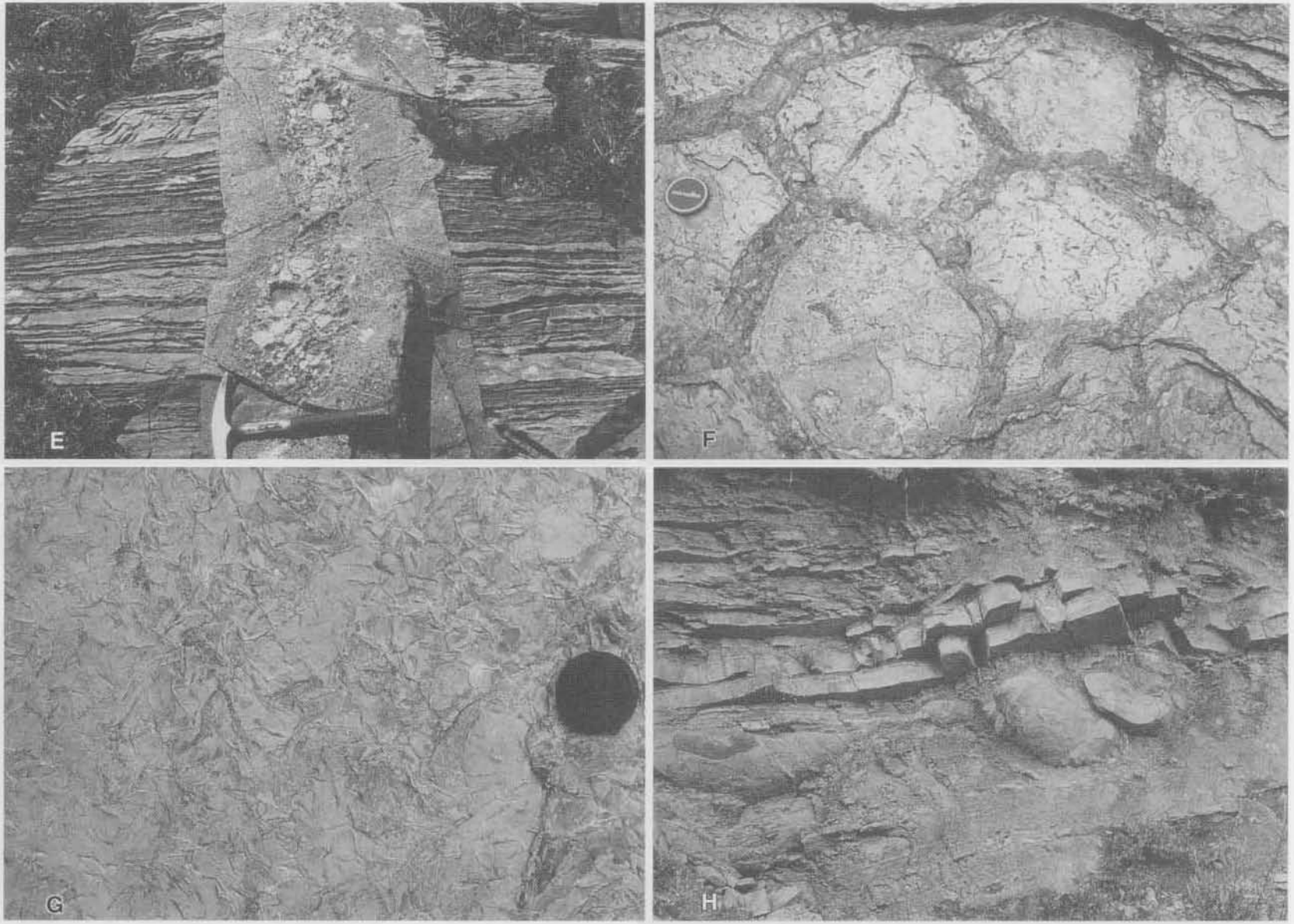


Fig. 2.17 E–H. see p. 49 for legend

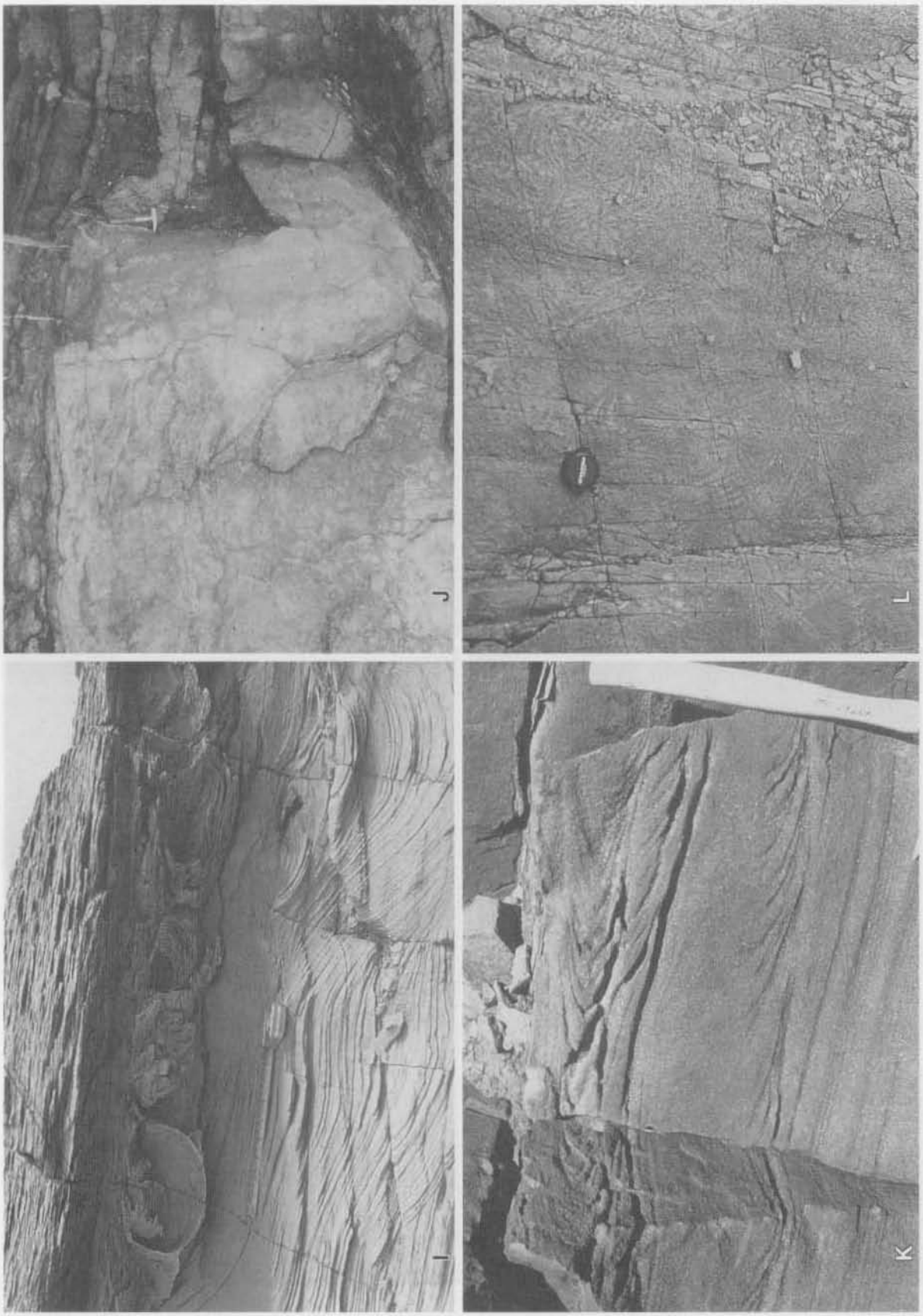


Fig. 2.171 – L. see p. 49 for legend

However, dish and pillar structures have also been observed in fluvial and other deposits and are, therefore, not environmentally diagnostic. Obviously, they can only be seen in deposits containing lamination and will not be present if the sand is uniform in texture.

Fluid movement within a bed is an additional cause of convolute lamination. In this case, the laminations are folded by internal shear, and may occasionally be broken through by pipes (Fig. 2.17 B, C). At the sediment-water interface, escaping water may bubble out as a small spring, building up a miniature **sand volcano** (Fig. 2.17 D).

The emplacement of relatively more dense material over a lower-density layer was discussed previously as the cause of load casts. In addition to the downward movement of the denser material, this situation can be the cause of upward movement of lighter sediment, which is injected, together with contained pore fluids, into the overlying rocks. This can occur on a large scale, producing **diapirs** of evaporite or mud, both of which flow readily under the overburden weight of a few hundred meters of sediment. These diapirs may be several kilometers across and may extend up for several kilometers through overlying deposits. Evaporite diapirs commonly develop on continental margins (Chaps. 6, 9). Mud diapirs are a characteristic feature of deltas, where coarser deltaic sediment is dumped rapidly on marine mud deposits. Pore fluids may be sealed in the sand beds by this process, with the resulting high overburden pressures leading to overpressuring. Exploration drilling into such deposits must include the use of blow-out preventers to guard against the explosive release of pore fluids and gases when an overpressured bed is penetrated by the drill.

On a smaller scale, the same injection process can generate **clastic dikes,** consisting of sheets of sandstone or conglomerate (siliciclastic or carbonate) cutting through overlying or underlying beds (Fig. 2.17 E). The host rocks usually are sharply truncated and not internally deformed, indicating that they were at least partially lithified prior to intrusion. Some dikes intrude along fault planes. Some are intensely folded, suggesting deformation by compaction and further dewatering after injection.

Desiccation cracks are readily recognized by even the untrained eye (Fig. 2.17 F). They are one of the best and most common indicators of subaerial exposure in the rock record (note the spelling of "desiccation", a word that is almost invariably misspelled by students). They may penetrate as deep as a few meters into the underlying rocks (although a few centimeters is more typical) and are normally filled by sediment from the overlying bed. **Teepee structures** are a variety of large desiccation cracks caused by limestone or evaporite expansion on tidal flats. Desiccated beds on tidal flats may peel or curl as they dry, and disrupted fragments commonly are redeposited nearby as an **intraclast breccia** (Fig. 2.15 G). A subtly different kind of shrinkage feature, termed a **synaeresis structure** (Fig. 2.17 G), may be distinguished from desiccation cracks by two principal differences: unlike desiccation cracks, they do not normally form continuous polygonal networks across bedding planes, but may appear as a loose assemblage of small worm-like relief markings on a bedding surface; second, they do not show deep penetration into the substrate but appear to rest on the bedding plane in which they are found. Synaeresis structures are common in such environments as lakes, lagoons, and tidal pools, where salinity changes may be frequent. As noted in Section 4.5.6, these features were thought to be subaqueous shrinkage cracks but have now been reinterpreted as gypsum pseudormorphs

Evaporation and freezing may cause the development of crystals of evaporite salts and water, respectively, on the depositional surface, particularly on alluvial flood plains, supratidal flats, and lake margins. Evaporite crystals and nodules may be preserved in the rock record and, of course, major evaporite deposits are common; but individual crystals commonly are replaced by pseudomorphs, or are dissolved and the cavity filled with silt or sand, forming a **crystal cast**. Such structures are a useful indicator of subaerial exposure and desiccation but do not necessarily imply long-term aridity. Gypsum and halite are

Fig. 2.17 A–L. Sedimentary structures produced by liquefaction, load, or fluid loss. **A** Dish structures in a grain-flow deposit. Cretaceous, Provence, France. **B** Pipe structure produced by penecontemporaneous vertical water escape in a fluvial sandstone. Devonian, northeast Scotland. **C** A set of small water-escape pipes in a lacustrine sandstone. Devonian, northeast Scotland. **D** Small sand volcano created by water escape at the depositional surface. Eocene, Banks Island, Arctic Canada. **E** A clastic dike filled with sandstone and conglomerate. Huronian Supergroup (Proterozoic), Espanola, Ontario. **F** Desiccation cracks in a lacustrine deposit. Devonian, northeast Scotland. **G** Synaeresis markings. Silurian, Somerset Island, Arctic Canada. **H** Ball and pillow structures in a turbidite unit. Cretaceous, Provence, France. **I** Penecontemporaneous slump produced by failure on a delta front slope. Eocene epoch, Banks Island, Arctic Canada. **J** Load and drape structures caused by the presence of massive crinoidal limestone in thin-bedded limestones and mudstones. Devonian, Princess Royal Islands, Arctic Canada. **K** Overturned crossbedding caused by fluid shear of a saturated sandstone. Proterozoic, Banks Island, Arctic Canada. **L** Ice-crystal casts in a fluvial floodplain pond. Karoo Supergroup, Beaufort West, South Africa

the two commonest minerals to leave such traces. Gypsum forms blade-shaped casts and halite characteristic cubic or "hopper-shaped" structures. Ice casts may be formed in soft sediments during periods of freezing but have a low preservation potential (Fig. 2.17L).

Particularly distinctive evaporite structures on supratidal (sabkha) flats are termed **ptygmatic, enterolithic,** and **chicken-wire structures.** These are caused by in-place crystal growth, expansion, and consequent lateral compression of evaporite nodules, possibly aided by slight overburden pressures. Enterolithic structure is so named for a resemblance to intestines; chicken-wire structure is caused by squeezing of carbonate films between the nodules; ptygmatic folds may be caused largely by overburden pressures (Maiklem et al. 1969).

Last, **gas-bubble-escape structures** should be mentioned. These are produced by carbon dioxide, hydrogen sulfide, or methane escaping from buried, decaying organic matter. Gas passes up through wet, unconsolidated sediment and forms bubbles at the sediment–water interface, leaving small pits on the bedding surface. These structures have been confused with **rain-drop imprints** but form subaqueously, as may be apparent from associated features preserved in the rocks. True rain-drop imprints probably are very rare. Further information and a discussion of this class of structures have been presented in a review by Mills (1983).

2.2.3.9 *Fossils*

Body fossils are obviously among the most powerful environmental indicators to be found in sedimentary rocks, and should be observed and identified with care. Paleontology and paleoecology are specialized subjects, a detailed discussion of which is beyond the scope of this book; however, those engaged in describing outcrop sections for basin-analysis purposes should be able to make use of such information as they can gather. A complete and thorough paleontological–paleoecological examination of an outcrop section may take several hours, days, or even weeks of work, involving the systematic examination of loose talus and the breaking open of fresh material or sieving of unconsolidated sediment in the search for a complete suite of fossil types. Extensive suites of palynomorphs or microfossils may be extracted by laboratory processing of field samples. Many apparently unfossiliferous or sparsely fossiliferous stratigraphic intervals have been found to contain a rich and varied fauna or flora by work of this kind, but it is the kind of research for which few basin analysts have the time or inclination.

We are concerned, in this book, with reconstructing depositional environments and paleogeography. Fossils can be preserved in three different ways that yield useful environmental information.

1. In-place life assemblages include invertebrate forms attached to the sea bottom, such as corals, archaeocyathids, rudistids, some brachiopods and pelecypods in growth positions, some bryozoa, stromatoporoids, stromatolites, and trees. In-place preservation usually is easy to recognize by the upright position of the fossil and presence of roots, if originally part of the organism. This type of preservation is the easiest to interpret because it permits the drawing of close analogies with similar modern forms, in the knowledge that the fauna or flora almost certainly is an accurate indicator of the environment in which the rocks now enclosing it were formed. Examples are illustrated in Fig. 2.18.
2. Environmental indicators almost as good as in-place life assemblages are examples of soft-bodied or delicately articulated body fossils preserved intact (Fig. 2.19). These indicate very little transport or agitation after death, and preservation in quiet waters, such as shallow lakes, lagoons, abandoned river channels, or deep oceans. The Cambrian Burgess Shale in the Rocky Mountains near Field, British Columbia, contains one of the most famous examples of such a fossil assemblage, including impressions of soft, nonskeletal parts of many organisms, such as sponges, that are not found anywhere else. The Jurassic Solnhofen Limestone of Germany is another good example.
   The bones of vertebrate animals tend to disarticulate after death because of the decay of muscle and cartilage and the destructive effects of predators. Nevertheless, entire skeletons of bony fish, reptiles, and mammals are commonly found in certain rock units, indicating rapid burial under quiet conditions. The Solnhofen Limestone is a well-known example containing, among other fossils, the entire skeleton and feather impressions of a primitive bird. Nevertheless, such fossil assemblages must be interpreted with care, because a limited amount of transportation is possible from the life environment to the site of eventual burial. Presumably, the bird of the Solnhofen Limestone lived in the air, not at the sediment–water interface where it was deposited!
3. Much more common than either of the foregoing are death assemblages of fossils that may have been transported a significant distance, perhaps many kilometers from their life environment. These commonly occur as lag concentrates of shelly debris such as gastropod, pelecypod,

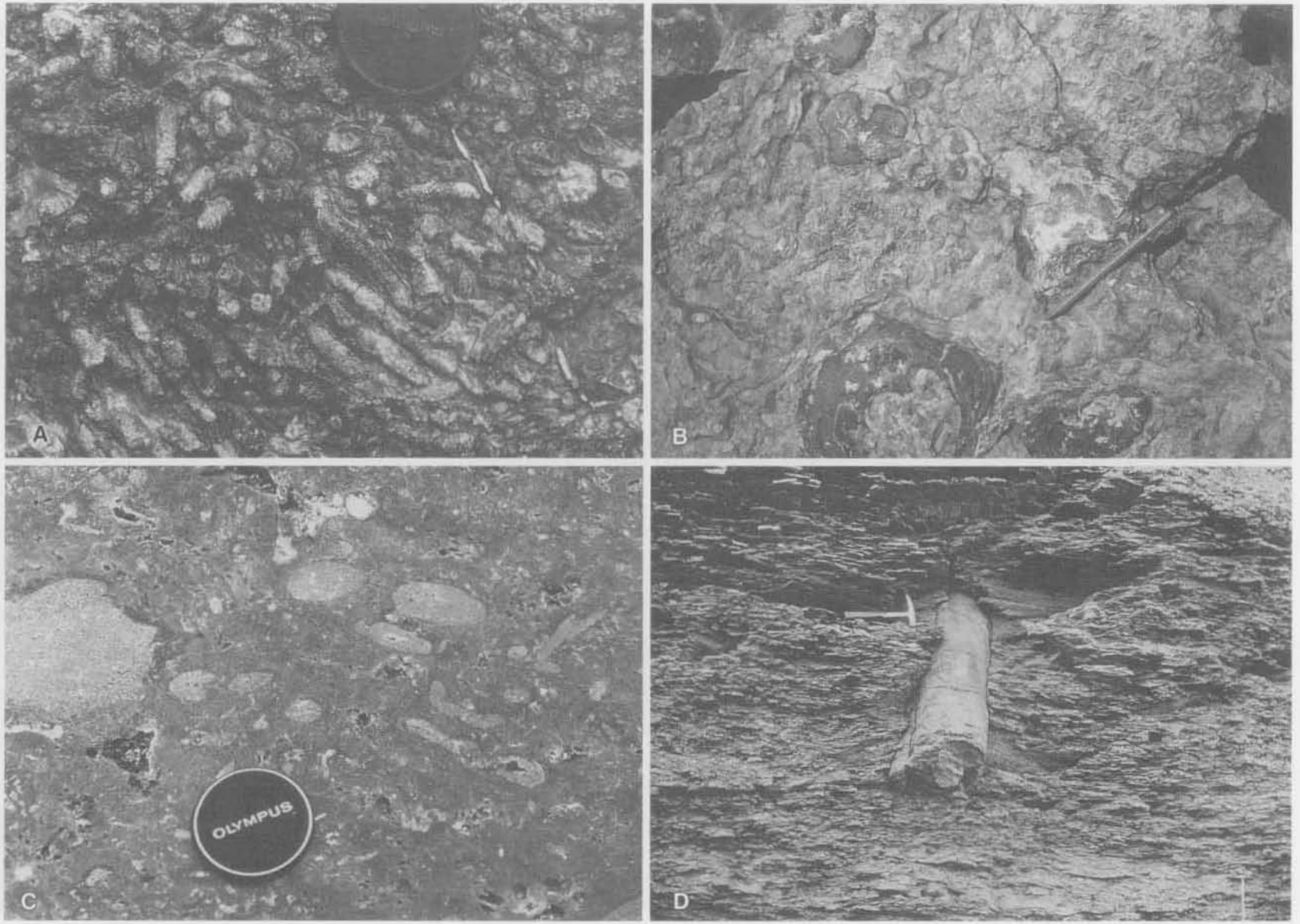


Fig. 2.18 A – D. see p. 52 for legend

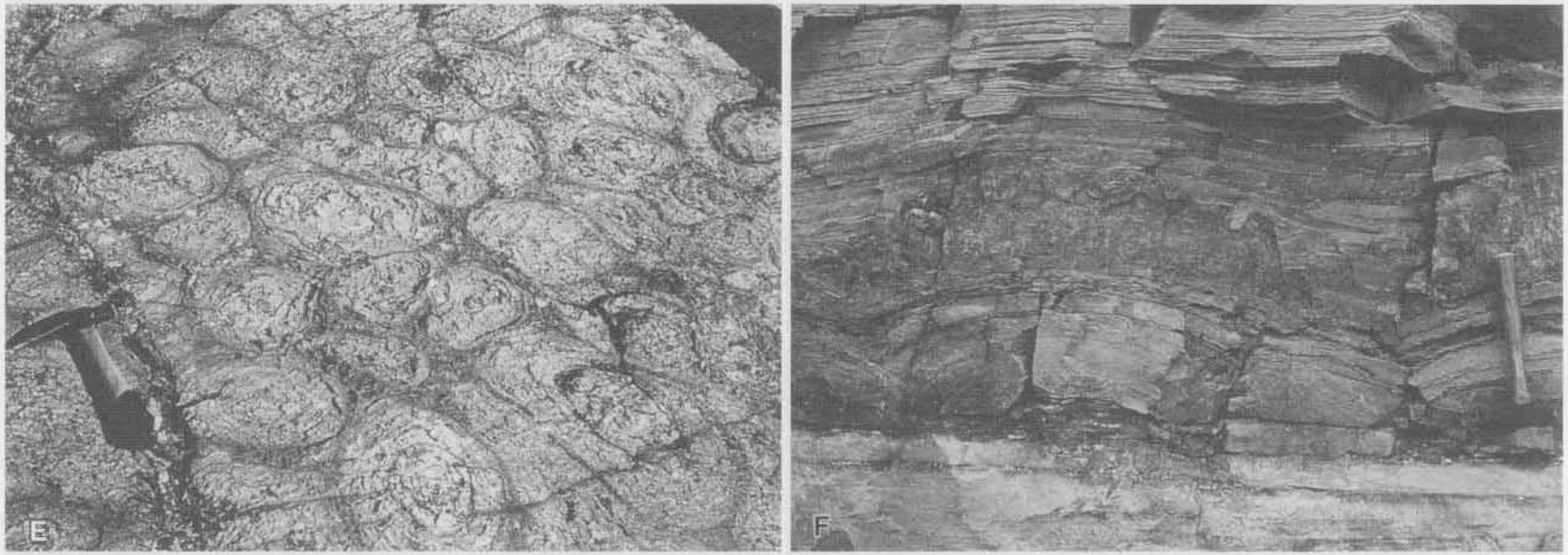


Fig. 2.18 A–F. In-place fossils (life assemblages). **A** Biostrome containing *Disphyllid* and *Thamnopora* corals. Devonian, northern Spain. **B** Coral biostrome. Devonian, near Norman Wells, Northwest Territories, Canada. **C** Typical reef facies. Boundstone composed of calcified sponges and encrusting stromatolites of Capitan reef. Permian, Guadalupe Mountains, Texas. **D** Tree in growth position. Pennsylvanian, Kentucky. **E** Domal stromatolites, bedding-plane exposure. Proterozoic, Dismal Lakes, Northwest Territories, Canada. **F** Domal stromatolites capped by finger stromatolites. Silurian, Somerset Island, Northwest Territories, Canada. (Photos **A** and **C** courtesy of B. Pratt)

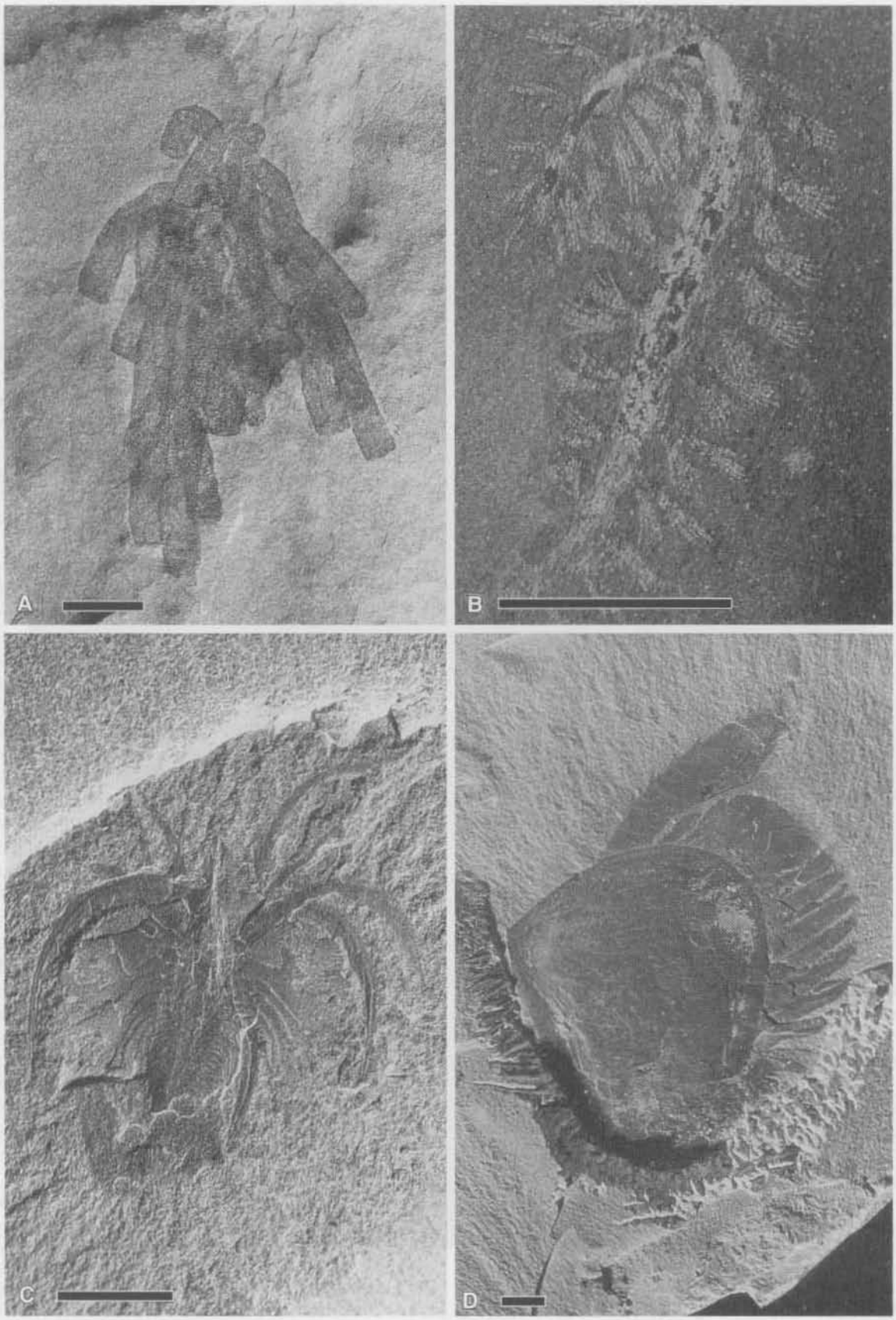
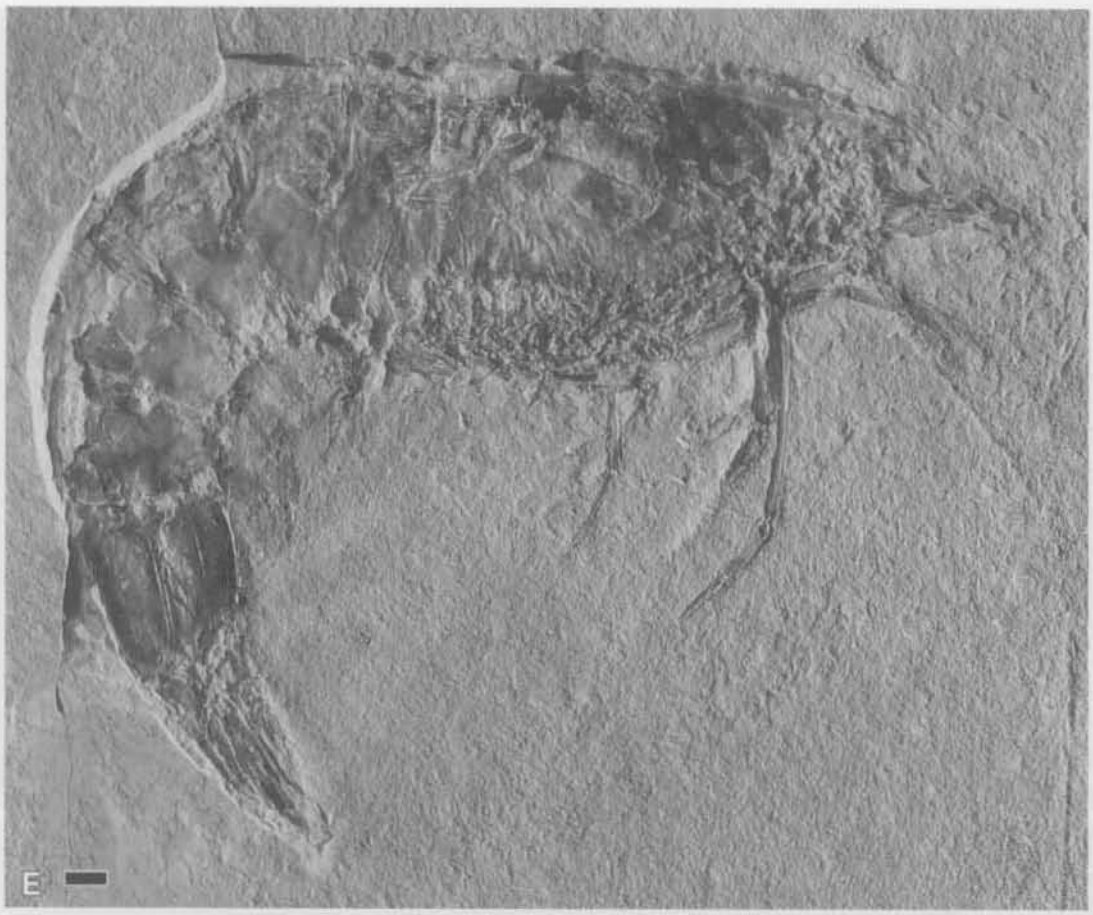


Fig. 2.19 A – D. see p. 55 for legend



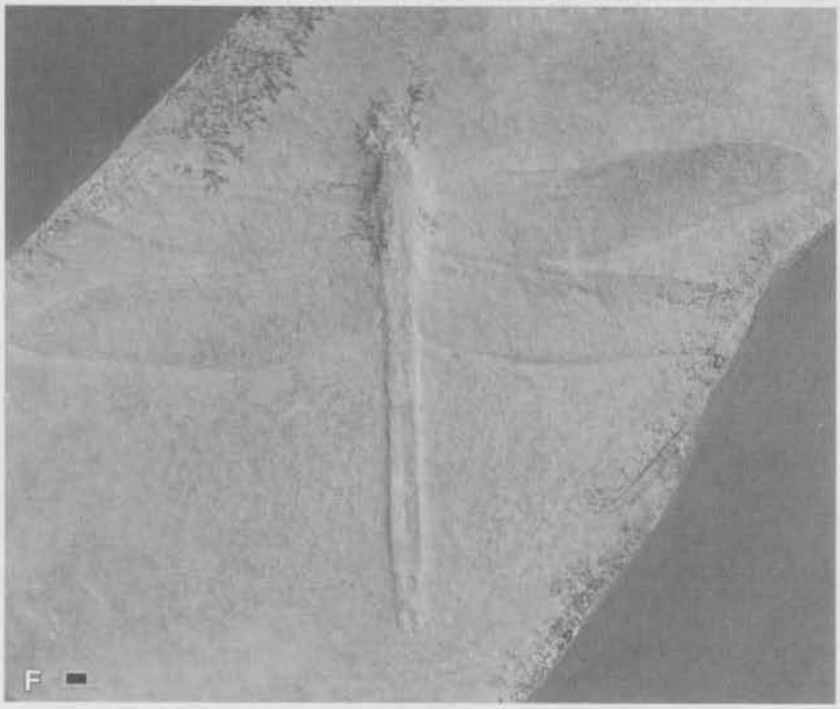


Fig. 2.19 E, F. see p. 55 for legend

brachiopod, or trilobite fragments, and fish bones or scales (Fig. 2.20). Such concentrations may be abundant enough to be locally rock forming as, for example, the famous Silurian Ludlow Bone Bed of the Welsh borderlands. Such assemblages normally indicate a channel-floor lag concentration or the product of wave winnowing and can usually be readily recognized by the fact that fragment grain size tends to be relatively uniform.

Transported body fossils may not occur as concentrations but as scattered, individual occurrences, in which case each find must be interpreted with care. Did it live where it is now found or was it transported a significant distance following death? The environmental deduction resulting from such an analysis may be quite different depending on which interpretation is chosen. Evidence of transportation may be obvious and should be sought; for example, broken or abraded fossils may have traveled significant distances. Overturned corals, rolled stromatolites (including oncolites), and uprooted tree trunks are all obviously transported.

These problems are particularly acute in the case of microfossils and palynomorphs that, on account of their size, are particularly susceptible to being transported long distances from their life environment. For example, modern foraminiferal tests are blown tens of kilometers across the supratidal desert flats of India and Arabia, and shallow-water marine forms are commonly carried into the deep sea by sediment gravity flows, such as turbidity currents. Detritus eroded from earlier stratigraphic units may also include derived fossil material. Environmental interpretations based on such fossil types may, therefore, be very difficult, though it may still be possible if the analysis is carried out in conjunction with the examination of other sedimentary features. In fact, it was the occurrence of sandstones containing shallow-water foraminifera interbedded with mudstones containing deep-water forms in the Cenozoic strata of the Ventura Basin, California, that was one of the principal clues leading to the development of the turbidity-current theory for the origin of deep-water sandstones.

Because of the great variety of life forms preserved as fossils, the subject of paleoecology is a large and complex one. Detailed studies are for the specialist, and a complete treatment is beyond the scope of this book. The reader is referred to such texts as Hallam (1973), McKerrow (1978), and Dodd and Stanton (1981). A few examples are discussed in Chapter 4.

2.2.3.10
Biogenic Sedimentary Structures

Footprints, burrows, escape burrows, and resting, crawling, or grazing trails are abundant in some rock units, particularly shallow-marine deposits; all may yield useful environmental information, including water depth, rate of sedimentation, and degree of agitation. Even the nondescript structure **bioturbation**, which is ubiquitous in many shallow marine rocks, can be interpreted usefully by the sedimentologist. Footprints are, of course, best seen on bedding-plane surfaces, as are many types of feeding trails and crawling traces (Fig. 2.21). Burrows are better examined in vertical cross section and are most visible either in very clean, fresh, wetted rock surfaces or in wind- or water-etched weathered outcrops (Fig. 2.22). Stromatolites are a distinctive component of many carbonate successions, particularly those of Precambrian age, and have been the subject of much detailed study.

Information to record on trace fossils includes morphology, size, attitude, abundance, internal structures, and lithofacies associations. Farrow (1975) provided a useful review of field and laboratory techniques for studying trace fossils. Table 2.4 is a classification of some common invertebrate trace fossils (from Simpson 1975). Interpretations of this information are discussed in Section 4.5.7. Useful texts on trace fossils include Hofmann (1973), Frey (1973, 1975), Crimes and Harper (1977), Basan (1978), Ekdale (1978), and Pemberton et al. (1992).

2.2.4
Sampling Plan

The amount of sampling to be carried out in outcrop section depends on the nature of the problem at hand, as discussed in Section 1.4. We are not concerned here with sampling of ore, hydrocarbon source beds, or coal to be analyzed for economic purposes, but the sampling required to perform a satisfactory basin analysis. Sampling is carried out for three basic purposes: (1) to provide a suite of typical lithologic

Fig. 2.19 A–F. Preservation of delicate structures, including soft parts. A–D are from the middle-Cambrian Burgess Shale, Yoho National Park, British Columbia; E and F are from the upper Jurassic Solnhofen Limestone, Bayern, Germany. *Bar scales* are all 5 mm in length. A *Vauxia gracilenta* Walcott, a sponge. B *Burgessochaeta setigera* Walcott, a polychaete worm. C *Marrella splendens* Walcott, an arthropod. D *Canadaspis perfecta* Walcott, a crustacean; E *Antrimpos speciosus* Munster, a shrimp. F *Aeschnogomphus intermedius*, a dragonfly. (All photos courtesy of D. Rudkin and the Royal Ontario Museum)

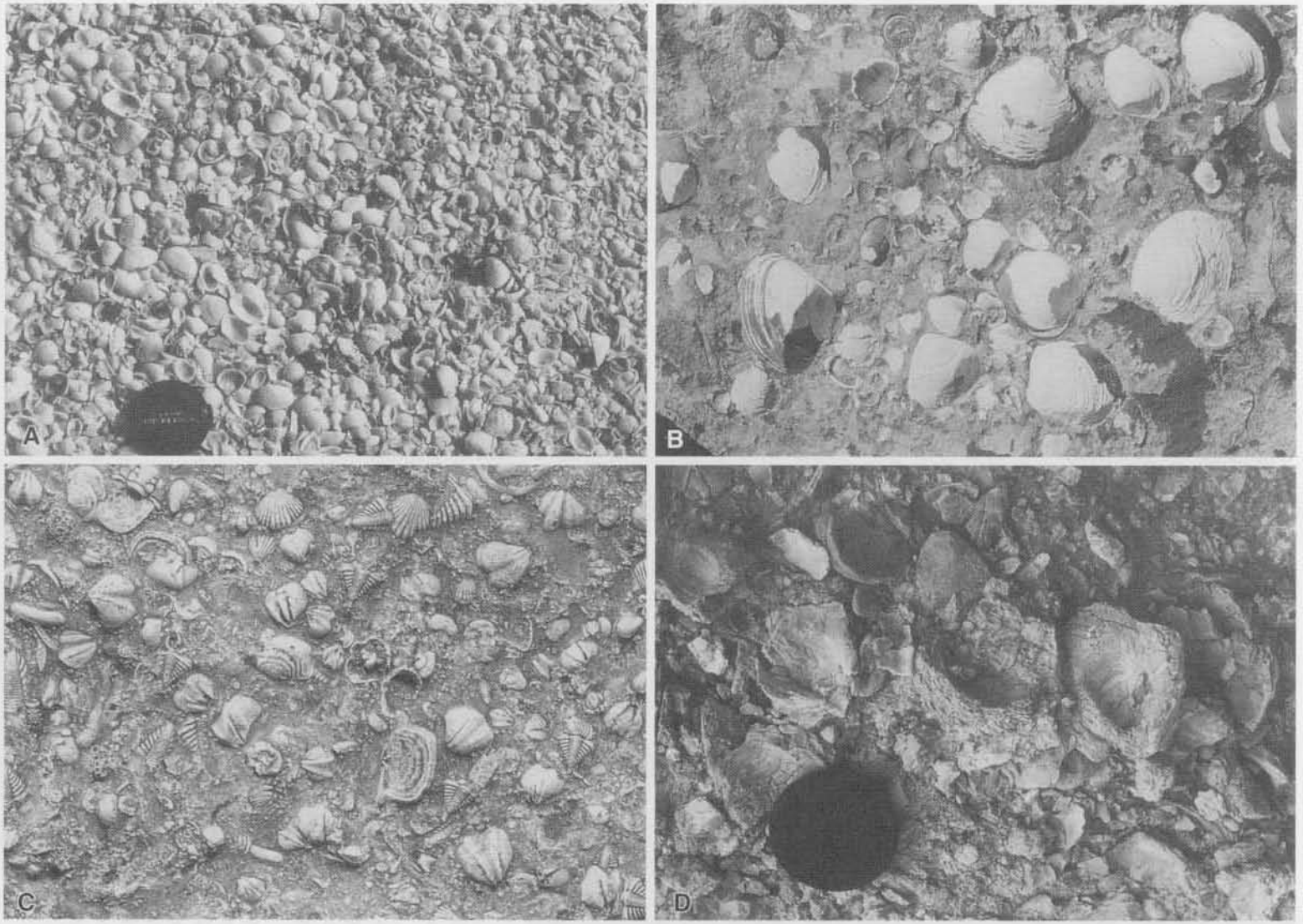


Fig. 2.20 A – D. Fossil assemblages. All but **H** are death assemblages. **A** Modern shell beach. Sanibel Island, Florida. **B** Pelecypods in shallow-marine sandstone. Cretaceous, Banks Island, Northwest Territories, Canada. **C** Trilobites and brachiopods on a bedding plane. Silurian, southern Ontario. **D** Oyster beach deposit. Pleistocene, Mallorca. (Photo **C** courtesy of R. Ludvigsen; photos **D, F** courtesy of B. Pratt)

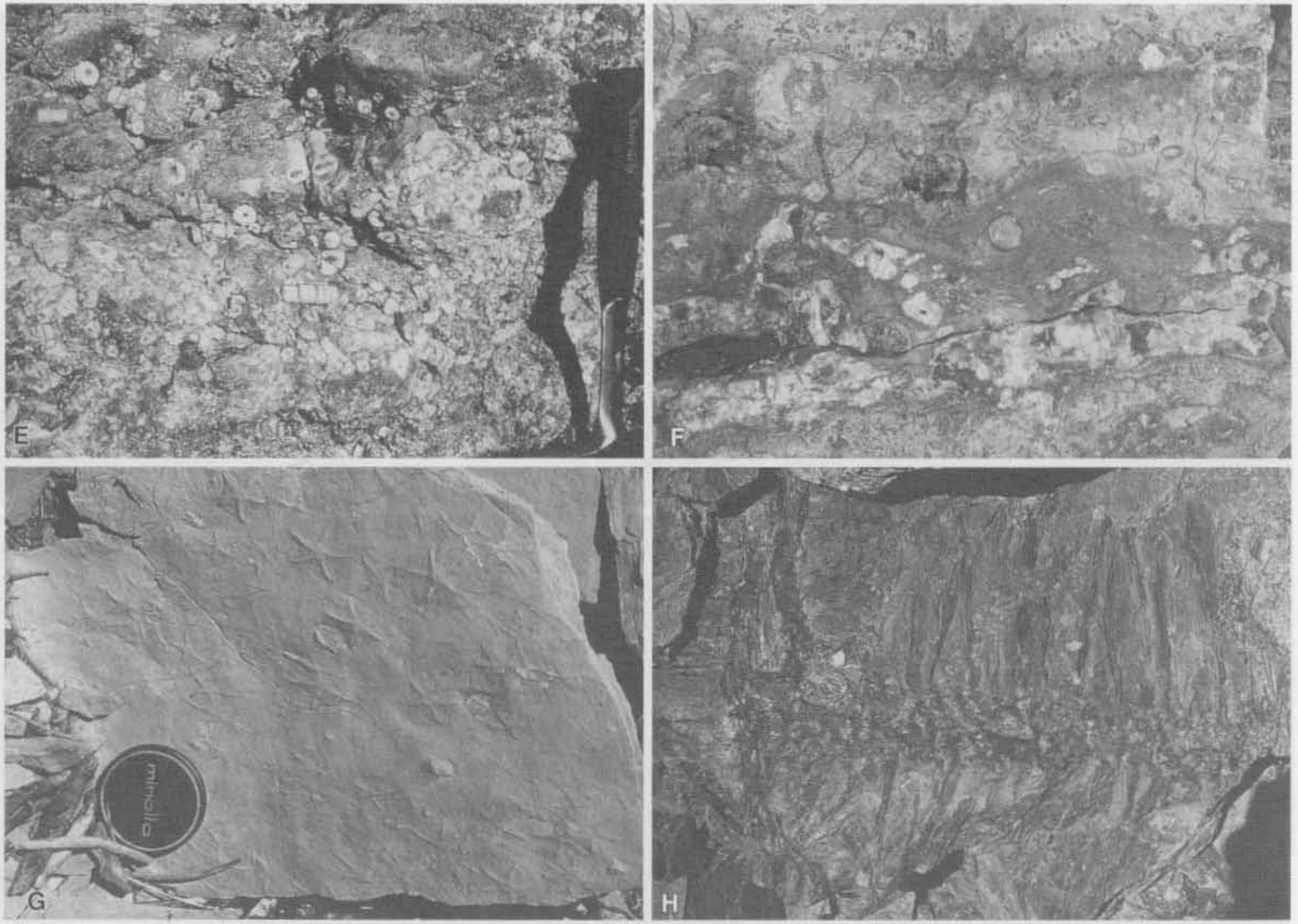


Fig. 2.20 E–H. (continued) **E** Crinoidal reef debris. Devonian, Princess Royal island, Northwest Territories, Canada. **F** Brachiopods, corals, ostracodes and other bioclastic debris in vertical section. Devonian lagoonal limestone, southern Ontario. **G** Starfish in shallow marine sandstones of Molasse marine sediments. Miocene, near Digne, France. **H** Plant fragments are typically preserved as fragments in a death assemblage, but this is an exception. The photograph shows in-place rootlets of *Stigmaria* species. Pennsylvanian, Kentucky. (Photo **C** courtesy of R. Ludvigsen; photos **D, F** courtesy of B. Pratt)

Fig. 2.21 A – D



Fig. 2.21 A – E. Trace fossils as exposed on bedding-plane surfaces. **A** Interconnected burrows in Ordovician limestone, Ottawa. **B** Short burrows in Ordovician sandstone. Boothia Peninsula, Arctic Canada. **C** Cross sections of vertical, tubular *Skolithus* burrows. Karoo Supergroup, near Durban, South Africa (see Fig. 2.22 A). **D** Casts of bird footprints. Cretaceous flysch, Provence, France. **E** Markings of a resting or swimming crustacean. Karoo Supergroup, Beaufort West, South Africa

samples illustrating textures, structures, or distinctive fossils on a hand-specimen scale; (2) to provide a set for laboratory analysis of petrography and maturation using polished slabs, the optical microscope, and possibly other tools, such as X-ray diffraction analysis and the scanning electron microscope; and (3) to gather macrofossil and lithologic samples for microfossil or palynological examination to be used for studying biostratigraphy.

2.2.4.1 *Illustrative Samples*

The choice of samples is usually simple and can be based on a trade-off between how much it would be useful to take and how much the geologist can physically carry. Large samples showing sedimentary structures or suites of fossils may be an invaluable aid in illustrating the geology of the project area to the geologist's supervisor or for practical demonstration at a seminar or a poster display at a conference. Unless field work is done close to a road or is being continuously supported by helicopter, it is rarely possible to collect as much as one would like.

2.2.4.2 *Petrographic Samples*

Before carrying out a detailed sampling program for petrographic work, the geologist should think very carefully about what the samples are intended to demonstrate. Here are some typical research objectives:

1. Grain-size analysis of siliciclastic and carbonate clastic rocks as a descriptive parameter and as an aid to interpreting the depositional environment
2. Petrography of detrital grains, including heavy minerals, as an aid to determining sediment sources
3. Petrography of carbonate grains as an aid to determining the depositional environment
4. Studies of grain interactions, matrix, and cement of both carbonate and clastic grains in order to investigate diagenetic history
5. Studies of detrital grain fabric in order to determine paleocurrent patterns or, in certain cases, as an aid in interpreting the depositional environment
6. Studies of thermal basin maturity using clay-mineral characteristics, vitrinite reflectance, microfossil color, or fluid inclusions
7. Samples for paleomagnetic study, for use in developing a reversal stratigraphy, a paleopole, or for studying diagenesis

The geologist must define the scope of the problem before collecting any samples, otherwise it may later be discovered that the collection is unsatisfactory. Work in remote regions is excellent training in such planning exercises, because only rarely is there a chance to return to an outcrop a second time. For the purpose of most regional studies, it is useful to ex-

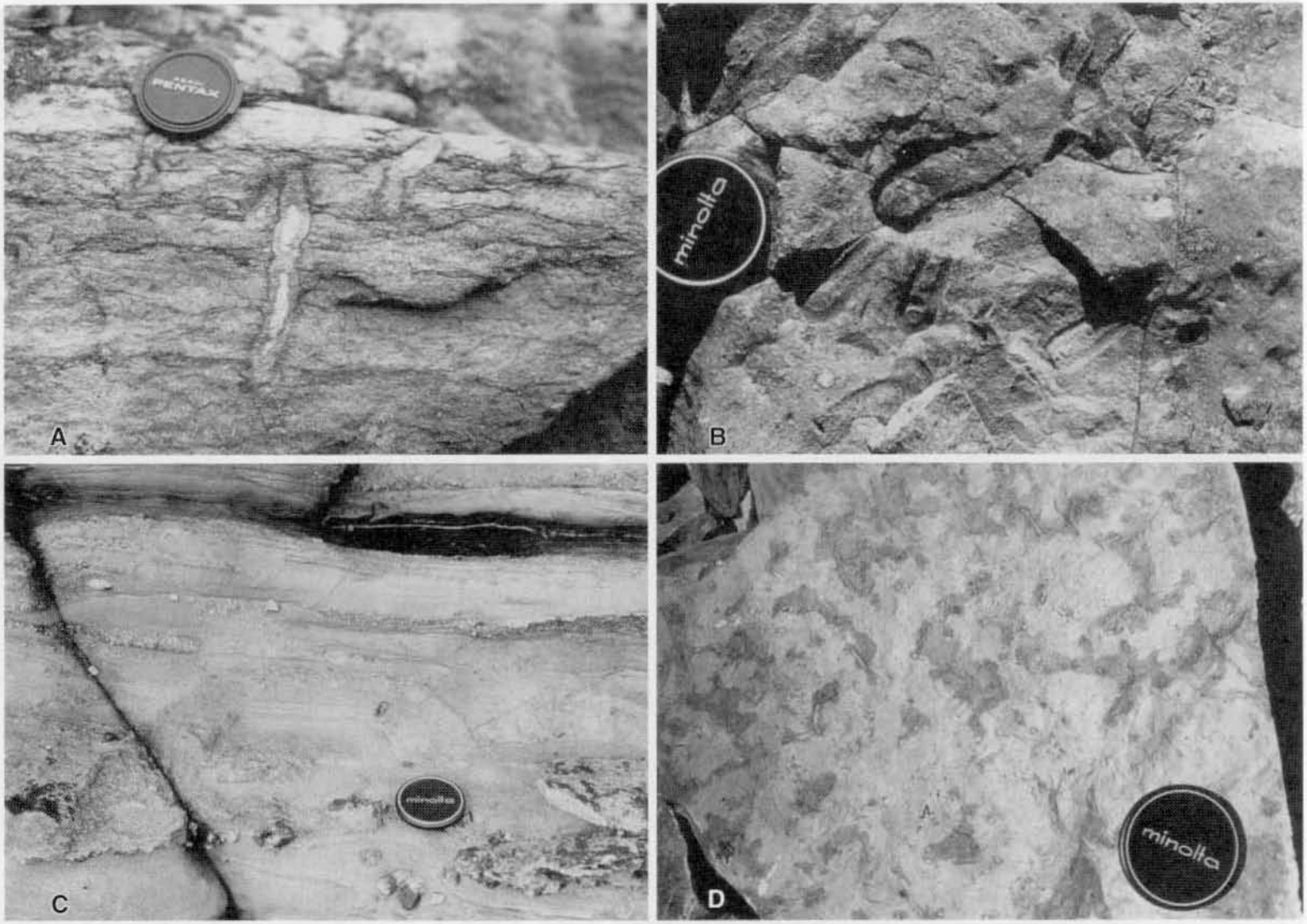


Fig. 2.22 A – D. Trace fossils as exposed in cross section. **A** *Skolithus,* Karoo Supergroup, near Durban, South Africa (see Fig. 2.21 C). **B** *Diplocraterion.* Cretaceous, Provence, France. **C** Sloping lungfish burrows. Devonian, northeast Scotland. **D** Mottling produced by bioturbation in a dolomite, plus individual small burrows. Ordovician, Boothia Peninsula, Arctic Canada

Table 2.4. Classification of trace fossils[a]

Rank IV	Rank III	Rank II	Rank I	Examples
Track-like trace on the bedding plane		Prods or scratches, all alike	Clustered scratches	*Paleohelcura*
			Rows of prods	*Tasmanadia*
		Prods or scratches of different kinds	Rows of prods	*Kouphichnium*
Trail-like trace on the bedding plane	Freely winding	Simple trail	No ornament	*Gordia*
			Transverse ornament	*Climactichnus*
		Bilobed trail	Transverse ornament	*Cruziana*
		Trilobed trail	Transverse ornament	*Scolicia*
	Windings in contact with one another; pattern on the bedding plane	Simple trail	No ornament	*Helminthoidea*
		Bilobed trail	Transverse ornament	*Nereites*
Radially symmetrical in a horizontal plane	Without vertical axial structure	Five-rayed	Rays are grooved	*Asteriacites*
		Multirayed	Club-shaped rays	*Asterosoma*
	With vertical axial structure	Circular outline	Conical depression	*Histioderma*
		Multirayed	Radial branches	*Lennea*
Tunnels and shafts	Of uniform diameter	Vertical	Isolated	*Tigillites*
			En masse	*Skolithos*
		Horizontal	Winding	*Planolites*
		U-shaped		*Arenicolites*
		Regularly branching		*Chondriles*
	Variable diameter	Irregular network		*Thalassinoides*
Forms having a spreite		U-shaped	Vertical plane	*Diplocraterion*
			Horizontal plane	*Rhizocorallium*
		Spiral	Inclined plane	*Zoophycos*
		Branched	Vertical plane	*Phycodes*
Pouch shaped		Smooth surface		*Pelecypodichnus*
		Transverse ornament		*Rusophycus*
Miscellaneous		Net pattern		*Palaeodictyon*

[a] From Simpson (1975).

amine petrographic variations vertically through a sequence and areally within a single stratigraphic unit. For example, the composition of detrital grains in a sequence may show progressive vertical changes, recording erosional unroofing of a source area or the switching of source areas. Sampling should be adequate to permit statistical documentation of this. Samples taken every 10–50 m through a sequence will normally suffice. For diagenetic and fabric studies and for paleomagnetic work, more detailed sampling may be required. Paleomagnetic research normally requires several samples from a single locality to permit checks for accuracy.

The most detailed sampling program is required for carbonate sequences, particularly those of shallow marine origin, which show the most facies variation. Laboratory work on polished slabs or thin sections is required for a reliable facies description of most carbonate rocks, and this may call for sampling every meter, or less, through a section.

For certain purposes, it is necessary that the sample be oriented; that is, it should be marked in the field so that its position in space can be reconstructed in the laboratory. For paleomagnetic and fabric studies, this need is obvious, but it may also be useful for petrographic purposes where, for example, it is necessary to examine cavity-filling detrital matrix to determine the time of filling relative to tectonic deformation, as an aid to determining structural top, or for studying microscopic grain-size changes related to bedding, e.g., graded bedding. To collect an oriented sample, the geologist selects a projecting piece of the outcrop that is still in place, not having been moved by frost heave, exfoliation, or other processes, and yet is still removable by hammer and chisel. A flat face on this piece is measured for orientation and marked



Fig. 2.23. Marks to show orientation on a sample before removal from the outcrop

by felt pen before removal (Fig. 2.23). A more detailed discussion of the methods and purpose of sample orientation is given by Prior et al. (1987).

How much should be collected at each sample station? A few hundred grams is adequate for most purposes. Thin sections can be made from blocks with sides less than 2 cm. Grain mounts of unconsolidated sediment can be made from less than 20 g. Where a particular component is sought, such as disseminated carbonaceous fragments for vitrinite-reflectance measurements, it may be necessary to take a larger sample to ensure that enough fragments are included for a statistical study at each sample station. Samples for paleomagnetic study are collected in a variety of ways, including the use of portable drills, which collect cores about 2 cm in diameter. Alternatively, the geologist may wish to take oriented blocks about $4 \times 84 \times 15$ cm, from which several cores can be drilled in the laboratory. Oriented specimens of unconsolidated sediment are collected by means of small plastic core boxes or tubes pushed into an unweathered face by hand. Textures may be preserved by on-site infiltration with resin (see also Prior et al. 1987).

How do we ensure that samples are truly representative? There is a conflict here between statistically valid experimental design and what is practically possible. Statistical theory requires that we take samples according to some specific plan, such as once every 10 m or 30 m (for example) through a vertical section, or by dividing a map area into a square grid and taking one sample from somewhere within each cell of the grid. Using these methods, we can satisfy the assumptions of statistical theory that our samples are truly representative of the total population of all possible samples. In practice, we can never fully satisfy such assumptions. Parts of any given rock body are eroded or too deeply buried for sampling. Exposures may not be available where sampling design might require them, or a particular interval might be covered by talus. An additional consideration is that the very existence of exposures of a geological unit might be governed by weathering factors related to the parameter the geologist hopes to measure. For example, imagine a sandstone bed formed at the confluence of two river systems, one draining a quartz-rich terrain and one a quartz-poor terrain. The quartzose sandstone intervals may be quartz cemented, much more resistant to erosion, and therefore much more likely to crop out at the surface. Sampling of such a unit might give very biased petrographic results.

In carrying out a basin analysis, we often deal with large areas and considerable thicknesses of strata. Petrographic, textural, and maturity trends are usually strong enough to show through any imperfections in our sampling program. We collect what we can, taking care that our measurements are controlled by the appropriate geological variables, for example, that counts of detrital components are all made on the same grain-size range (see Chap. 5).

2.2.4.3 *Biostratigraphic Samples*

The study of any fossil group for biostratigraphic purposes is a subject for appropriate specialists who, ideally, should collect their own material. However, this may not be possible, and the geologist often is required to do the collecting.

Unless a unit is particularly fossiliferous, the collecting of macrofossils can rarely be performed in a fully satisfactory way by the geologist, who is also

measuring and describing the section. The search for fossils may take a considerable amount of time, far more than is necessary for other aspects of the work. In practice, what the geologist usually ends up with are scattered bits and pieces and spot samples of more obviously fossiliferous units, in which it may be fortuitous whether or not there are any species of biostratigraphic value present. Given adequate time – for example, the two or three field seasons required for dissertation research – the geologist may be able to spend more time collecting samples and familiarizing him or herself with the fauna and/or flora; in reconnaissance mapping exercises, however, this is usually impracticable.

The increased sophistication of subsurface stratigraphic analysis by the petroleum industry has led to a greatly expanded interest in fossil microorganisms. Groups such as conodonts, acritarchs, foraminifera, palynomorphs, diatoms, and radiolaria have been found to be sensitive stratigraphic indicators and commonly have the inestimable advantage of occurring in large numbers so that biostratigraphic zonation can be based on statistical studies of taxon distribution. Microfossils are extracted by deflocculation or acid dissolution of suitable host rocks. Most useful fossil forms are pelagic or are distributed by wind (palynomorphs) so that, potentially, they may be found in a wide variety of rock types. However, their occurrence is affected by hydrodynamics in the depositional setting and postdepositional preservation. Palynomorphs may be rare in sandstones because they cannot settle out in the turbulent environments in which sand is deposited, but they are abundant in associated silts, mudrocks, and coal. Radiolarians and other siliceous organisms may be entirely absent in mudstones but abundantly preserved in silts, cherts, and volcanic tuffs because they are dissolved in the waters of relatively high pH commonly associated with the formation of mud rocks. Conodonts are most commonly preserved in limestones and calcareous mudstones and may be sparse in dolomites because dolomitization commonly occurs penecontemporaneously in environments inimical to conodonts, such as sabkha flats.

Armed with advice of this kind from the appropriate specialist, the geologist can rapidly collect excellent suites of samples for later biostratigraphic analysis and can cut down on the amount of barren material carried home at great effort and expense only to be discarded. Advice should be sought on how much material to collect. Normally, a few hundred grams of the appropriate rock type will yield a satisfactory fossil suite, but more may be required for more sparse fossiliferous intervals; for example, several kilograms may be needed if conodonts are to be extracted from unpromising carbonate units.

Samples should be collected at regular vertical intervals through a section, preferably every 10–50 m. This will vary with rock type to permit more detailed sampling from condensed units or particularly favorable lithologies. Such sample suites permit the biostratigrapher to plot range charts of each taxon and may allow detailed zonation. Scattered or spot samples have to be examined out of context and may not permit very satisfactory age assignments.

2.2.5 Plotting the Section

A stratigraphic section can be published in the form of a written description, but this is not a very effective use of the information. It is required for the formal description of type and reference sections of named stratigraphic units, but it is doubtful that much is to be gained by reading such a description. The same information can be conveyed in a much more compact and digestible form as a graphic log with a central column for lithology and adjacent columns for other features of importance. Such logs have the added advantage that they can be laid side by side, permitting comparisons and correlations between sections from several locations.

It is possible to devise logging techniques with columns and symbols to convey every scrap of lithologic and petrographic information plus details of fossils, sedimentary structures, paleocurrent measurements, and even chemical composition. An example is illustrated in Fig. 2.24. Many companies and government organizations print blank logging forms of this type for use by their geologists to facilitate the logging process and to standardize the results of different workers. Johnson (1992) developed a comprehensive, all-purpose logging form that he offered as a standard for field workers.

However, such logs have the disadvantage that significant information may be lost in a welter of detail. If a main purpose of a log is to permit visual comparisons between different sections, it is advisable to simplify the logs so that the critical features stand out from the page. For the purpose of basin analysis, the most important data are those that carry paleogeographic information such as lithology, grain size, sedimentary structures, and fossil content. Paleocurrent data can be added to these but commonly are treated separately (Chap. 5). As we shall see in Chapter 4, the vertical succession of facies is often of crucial importance in interpreting the depositional environment, and so it is helpful to emphasize this in the logs.

For clastic rocks, one of the most useful techniques is to vary the width of the central lithology column according to grain size; the wider the column, the

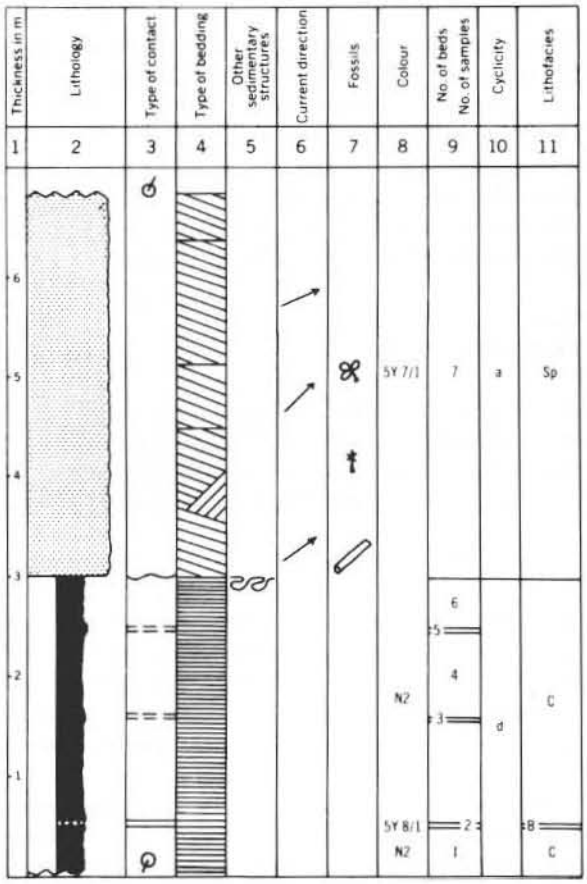


Fig. 2.24. Example of a complex log showing all lithologic information in symbolic form

coarser the rocks. Examples are illustrated in Fig. 2.25. Many other examples are given by Reading (1986). This method imitates the weathering profile of most clastic rocks, as muddy units tend to be less resistant and to form recessive intervals, in contrast to the projecting buttresses and cliffs of coarser sandstones and conglomerates. Drawing the column in this way enables rapid visual comparisons between sections and also permits instant recognition of gradual trends, such as upward fining and sharp breaks in lithology, as at an erosion surface. Subtleties, such as changes in sorting or a bimodal grain-size distribution, cannot be displayed in this way but are rarely as important from a facies perspective and may, in any case, be accompanied by other kinds of facies change, such as in a sedimentary-structure assemblage, which can be readily displayed in visual logs.

The variable-column-width technique has not been widely used for carbonate rocks, but there is no reason why it should not be. However, grain size is subject to changes by diagenesis, and this may make interpretation more difficult.

Within the column, various patterns can be used to indicate lithology, including symbols for sedimentary structures and fossils. For siliciclastic sequences consisting of interbedded mudstone, siltstone, sandstone, and conglomerate, the column width conveys most of the necessary lithologic information, and the body of the column can be used primarily for structures and fossils. Some loggers split the column in two, one side for lithology and one for structures, but this is rarely as visually successful. Very little need be placed outside the column, which preserves an uncluttered appearance and increases the graphic impact of the log. Symbols, abbreviations, and other plotting conventions are discussed in Section 2.3.4.

At what scale should the logs be drawn? This depends on what it is they are intended to demonstrate. Detailed, local sedimentological logs may require a scale of between 1 cm:1 m and 1 cm:10 m. Regional stratigraphic studies can be illustrated in large fold-out diagrams or wall charts at scales in the order of 1 cm:10 m to 1 cm:50 m, whereas page-sized logs of major stratigraphic sequences can be drawn (grossly simplified) at scales as small as 1 cm:1000 m. In the petroleum industry, the scale 1 inch:100 ft has long been a convenient standard for subsurface stratigraphic work. This translates to a convenient approximate metric equivalent of 1 cm:10 m, although it is actually closer to 1 cm:12 m. Metric units should always be used, preferably in multiples of 10.

2.3 Describing Subsurface Stratigraphic Sections

2.3.1 Methods of Measuring and Recording the Data

Subsurface sections are logged and described using three types of data: well cuttings, core, and petrophysical logs (Sects. 1.4.3 and 1.4.4). All three are typically available for the large-diameter holes drilled by petroleum exploration companies. Diamond-drill holes (DDHs) provide a continuous core but nothing else. The logging techniques and the results obtainable are therefore different.

2.3.1.1 *Examination of Well Cuttings*

Samples stored in company and government laboratories are of two types: washed and unwashed. Unwashed cuttings consist of samples of all the material that settles out of the mud stream into a settling pit at the drill site. Unconsolidated mudrocks may disperse completely into the mud stream during drilling, in which case little of them will be preserved except as coatings on larger fragments or occasional soft chips.

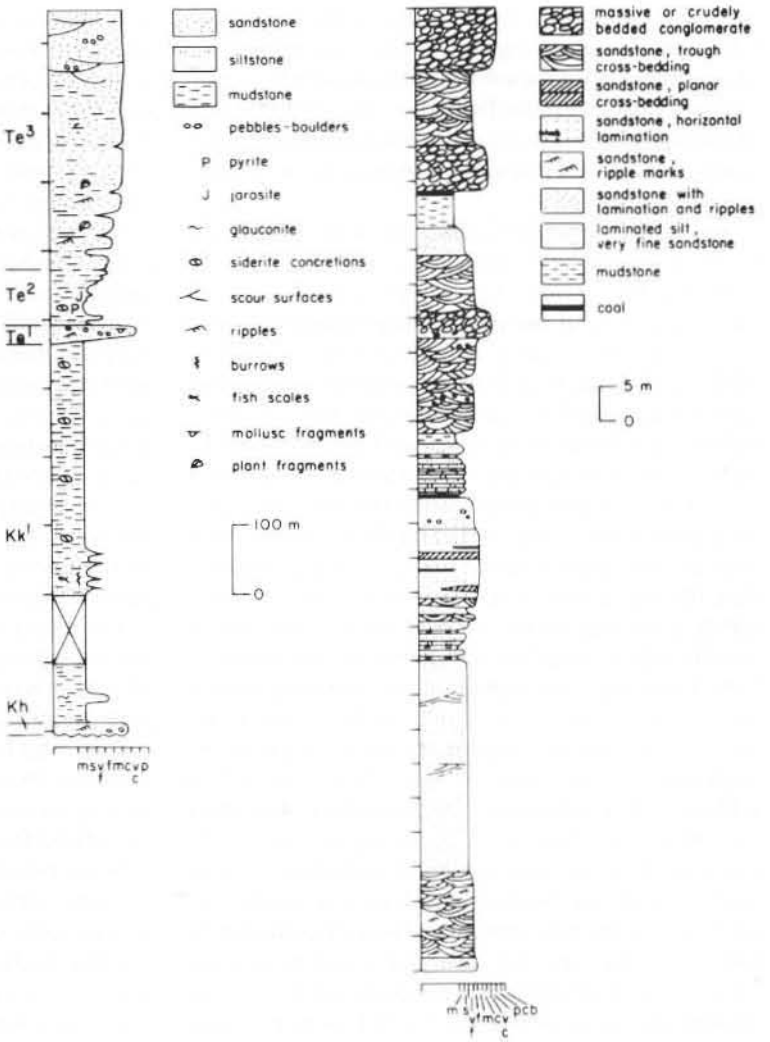


Fig. 2.25. Examples of graphic stratigraphic logs drawn to emphasize clastic sedimentological features. *Kh*, *Kk*^{1}, etc., are formation codes. Grain size is given by the width of the column, as indicated by the device at base of the column. *m* Mud; *s* silt; *vf*, *f*, *m*, *c*, and *vc* very fine, fine, medium, coarse, and very coarse sand; *p* pebbles; *c* cobbles; *b* boulders

Washed cuttings are those from which all mud has been removed. The washing process makes the cutting examination process easier, but it further biases the distribution of rock types present if the drill penetrated any unconsolidated muddy units. Stratigraphic well logging is normally carried out on the washed cuttings, whereas palynological and micropaleontological sampling is done on the unwashed material. Stratigraphic logging techniques are described below. A detailed guide and manual was published by Low (1951) and is well worth reading. Many companies also provide their own manuals. McNeal (1959) has some useful comments, and the American Association of Petroleum Geologists has also issued a logging guide (Swanson 1981).

As described in Section 1.4.3, samples are collected at the well site and bagged every 10 ft (3 m). The bag is labeled according to the depth of recovery by the well-site geologist, who makes allowances for the time taken for the mud and cuttings to rise to the surface. Measurements are normally given as "depth below K. B."; K. B. stands for kelly bushing, a convenient measurement location on the drilling platform a few meters above ground level. The altitude above sea level of this point is determined by surveying, so that these drilling depths can be converted to "depths subsea." On offshore rigs, K. B. is 25 – 30 m above sea level.

For various reasons, not all the cuttings in any given bag may be derived from the depth shown. The problems of caving and variable chip density have been referred to in Section 1.4.3. These problems are particularly acute in soft or unconsolidated rocks, and samples from such a sequence may consist of a heterogeneous mixture bearing little relation to such stratigraphic detail as thinly interbedded units of contrasting lithology. The loss of the muds from the cutting suite compounds the problem. There are several ways in which these problems can be at least partially resolved. Caved material may be obvious by the large size of the fragments or by its exotic litho-

logy or fossil content. For example, in one well in the Canadian Arctic, I logged Jurassic pelecypods and foraminifera 270 m below the unconformable contact of the Jurassic with the Devonian. The geologist will gradually become familiar with the formations under study, and will then readily recognize such obvious contamination.

A more powerful tool is available to the logger, and that is to study the cuttings in conjunction with the suite of petrophysical (wire-line) logs from the hole. These logs record various physical properties as a measuring tool is slowly run the length of the hole. Modern petrophysical logging methods are capable of resolving lithologic variations over vertical intervals of a few decimeters or less and can therefore be used, with practice, to interpret lithologies in conjunction with the well cuttings. A description of the common petrophysical methods and their uses is given later in this chapter. Once the geologist is familiar with the petrophysical response of the various lithologies in the hole under examination, it is possible to use the logs to adjust or correct the sample description. A lithologic log may therefore be drawn up that bears only a loose relationship to the material actually present in the sample bags. Such a log is an interpretive log and should be clearly labeled as such. It is likely to be more useful in basin interpretation than a log which simply records the cuttings dogmatically, particularly in the case of poorly consolidated beds or those with rapid vertical lithologic variations. Soft muds will be entirely unrecorded in a straight sample log, which may give the geologist a very inaccurate picture of the subsurface stratigraphy. All this can be allowed for in an interpretive log, but interpretations can be wrong, and the geologist must be aware of it when using this type of record.

Figure 2.26 illustrates a comfortable, convenient laboratory arrangement for examining well cuttings.



Fig. 2.26. Laboratory arrangements for examining well cuttings. (Geological Survey of Canada, Calgary; photo courtesy of J. Dixon)

The samples are tipped from the sample bag or vial into a metal tray (Fig. 1.7) and observed under a low-power, reflected-light binocular microscope. Immersing the samples in water may aid observation, particularly as dust adhering to the chips can be washed off. The most useful magnification range for such a microscope is from about ×5 to ×50. The critical petrophysical logs should be unfolded to the appropriate depth interval and placed at one side of the microscope. It is advisable to scan rapidly the samples from several tens of meters of section before beginning the detailed description. Like standing back from an outcrop section, this gives the geologist the opportunity to perceive major lithologic variations. These may be correlated with changes in the petrophysical-log response, permitting precise depth control.

As the samples are described, the information may be recorded directly on a preprinted logging form or written out in note form. If the log is to be an interpretive one, several tens of meters of section should be examined before plotting the graphic log so as to give the geologist time to digest what is being observed. It may be necessary, for publication or other purposes, to produce a written sample description, but, as discussed in Section 2.2.3, these are difficult for a reader to absorb and are likely to be seldom used. The Geological Survey of Canada now publishes them in microfiche form to save paper and space.

Many petroleum- and mining-exploration companies and service companies (such as the American Stratigraphic Company, Canadian Stratigraphic Service Ltd., and International Geosystems Corporation) now use computer-processible logging forms. An example of a form designed for studying outcrops of nonmarine clastic rocks was discussed earlier (Sect. 2.2.1). Similar forms have been designed for well logging, and the data are then stored in digital form in data banks. Retrieval programs may be available that can use these data for automated log plotting, and the same data bank can be exploited for automated plotting of maps and sections, as discussed in Chapter 5.

2.3.1.2 *Examination of Core*

The large-diameter core produced by petroleum drilling is stored either in a company laboratory or an official government repository. North American practice is to divide the core into 2.5-ft (0.8-m) lengths, which are stored side by side, two to a box. Top and bottom should be marked on the box. The core usually consists of a series of short pieces broken from each other by torque during the drilling process. The well-site geologist should number or label each piece with respect to its position in the box and its

orientation because, unless this is done, once a piece is moved it may be very difficult to restore it to its correct position, and serious errors may be introduced in reconstructing the vertical lithologic succession. Diamond-drill-hole (DDH) cores are normally stored in 5-ft (1.6-m) lengths, five to a box. The same remarks apply with regard to the position and orientation of core pieces. DDH cores are rarely brought from the field back into the office, except for crucial holes. They are examined and logged in the field and, then, are commonly abandoned. They may even be tipped out of the box to prevent competitors from taking a look. This is a great waste of research material, but the practice seems unlikely to change.

Cores are most conveniently examined in a laboratory specifically designed for this purpose (Fig. 2.27). Core boxes are laid on roller tables so that they can be readily loaded and unloaded using a fork-lift vehicle. A movable platform is positioned above the core boxes, on which are placed the microscope, petrophysical logs, note book, etc.

The grinding action of the core barrel during drilling may smear the core surface and obscure lithologic features; sometimes, this can be rectified by washing the core with water or even dilute hydrochloric acid. An even more useful technique is to have the core cut longitudinally with a rock saw, creating a flat section. This should always be wetted with water or etched lightly with dilute acid before examination. The etching technique is particularly useful when examining carbonate rocks, as it tends to generate a fine relief between grains and cement or different carbonate minerals.

Where petrophysical logs are available, it is important to correlate core lithology with log response. This exercise may reveal that parts of the core are missing, perhaps as a result of fragmentation of soft lithologies. Such information is of importance in attempts to reconstruct a detailed vertical lithologic profile.



Fig. 2.27. A core laboratory. (Geological Survey of Canada, Calgary; photo courtesy of J. Dixon)

For a discussion of description and plotting routines, refer to the previous section on well cuttings. Essentially the same methods are used, except that many more features are visible in core, such as bedding features, sedimentary structures, and macroscopic trace and body fossils. Orientation of core for geological purposes was discussed by Davison and Haszeldine (1984) and Nelson et al. (1987).

2.3.2 Types of Cutting and Core Observation

Large-scale features, including most sedimentary structures and the subtleties of bedding, cannot be identified in well cuttings. They are partly visible in cores, but cores usually provide only frustratingly small snapshots of major features, and core research is rather like trying to describe elephant anatomy by examining a piece of skin with a microscope. The following notes are given in the same format as for surface sections so that contrasts with the latter can be emphasized. The description of field observation techniques should be referred to where appropriate.

2.3.2.1 *Subdivision of the Section into Descriptive Units*

In the case of petroleum exploration wells, subdivision of the section is best carried out by core and sample examination in conjunction with petrophysical logs. The combination is a powerful one and yields good generalized stratigraphic subdivisions with precise depth control. For DDH cores the absence of petrophysical logs is compensated by the availability of a continuous core, and stratigraphic subdivision is simple. Because so many features, such as sedimentary structures, cannot be observed in cuttings, descriptive units in the subsurface tend to be thicker and more generalized than those observed in the outcrop. However, for core, the focus on what are really very small samples and the attempt to maximize the use of limited amounts of information tends to lead to very detailed descriptions. Examination of surface sections based on cuttings and those on cores, particularly short petroleum cores, require very different concepts of scale. These should be borne in mind when a basin-analysis exercise calls for the correlation of all three types of data.

2.3.2.2 *Lithology and Grain Size*

These can be observed satisfactorily, using the same techniques as described under surface sections, in cuttings for all rock types except conglomerates.

Conglomerates cannot be adequately studied in cuttings where the clast size is larger than the cutting size. It may not be possible to ascertain which cuttings represent clast fragments and which matrix, and no observations of clast grain size can be made. Remember, also, that unconsolidated silts, muds, and evaporites may not be represented in well cuttings. These may require identification using petrophysical logs.

Well cuttings commonly contain contaminants that the logger should discard or ignore. Many are easy to recognize, such as metal pipe shavings or bit fragments. Oily substances, such as pipe dope or grease, may coat some fragments but can usually be distinguished from natural oil stains by the fact that they coat the cuttings and do not penetrate them. Drilling mud may also coat the cuttings, particularly poorly washed samples. Casing cement may appear as a flood of cuttings at certain levels, where the hole was reentered following the setting of the casing. Cement can be easily mistaken for sandy, silty, or chalky carbonate. Finally, foreign materials, such as feathers, sacking, seeds, cellophane, perlite, or coarse mica flakes, may be present. These are used in the drilling mud to clog large pores and prevent loss of mud circulation. Usually, there are few restrictions on lithology and grain-size determinations in core except where particularly coarse conglomerates are present.

2.3.2.3 *Porosity*

See the discussion in Section 2.2.3. Observations and measurements from subsurface rocks are more reliable than are those from outcrops because of the complications of surface weathering.

2.3.2.4 *Color*

See the discussion under the heading of surface sections (Sect. 2.2.3).

2.3.2.5 *Bedding*

For core, see the discussion under the heading of surface sections (Sect. 2.2.3). Bedding cannot be seen in well cuttings except for fine lamination, whereas core provides good information on bedding variation. Caution should be exercised in attempts to extract any quantitative information about bed thickness from cores because of the possibility of core loss, as discussed earlier.

2.3.2.6 *Sedimentary Structures*

Very few, if any, structures can be observed in well cuttings. However, a wide range of structures is visible in core. Large structures, such as major erosion features and giant crossbedding, may be difficult to discern because the small sample of the structure visible in the core may easily be confused with something else. For example, thick crossbed sets could be mistaken for structural dip if the upper or lower termination of the set against a horizontal bedding plane cannot be determined. Dipmeter interpretations and formation microscanner observations may help here, as discussed in Sections 5.4.3 and 5.9.6. Erosion surfaces are practically impossible to interpret in small outcrops or cores. The break may indicate anything from a storm-induced scour surface to an unconformity representing several hundred millions of years of nondeposition. The presence of soils or regoliths below the erosion surface is about the only reliable indicator of a major sedimentary break (except, of course, where there is structural discordance).

Ripple marks and crossbed sets up to a few decimeters thick are more readily recognizable in core, and the reader should turn to Section 2.2.3 for methods of study. The large-scale geometry of crossbed sets is difficult to interpret in core. For example, the difference between the flat shape of a foreset in planar crossbedding and the curved surface of a trough crossbed is practically impossible to detect in core, even large-diameter petroleum core. Curvature of a typical trough crossbed across such a core amounts to about 2°. Sensitive dipmeter logs have considerable potential for interpreting crossbedding in the subsurface, but the methodology of interpretation has yet to be fully worked out. Dipmeter logs permit some paleocurrent measurements to be made in the subsurface (Sect. 5.9.6), something which cannot now be done without an oriented core, and the availability of the latter is practically zero.

Small-scale erosion features such as flutes, tool markings, and rain prints, and other bedding features such as desiccation cracks and syneresis markings, are difficult to find in core because bedding plane sections are rare, and the geologist is discouraged from creating additional sections by breaking up the core. Liquefaction, load, and fluid-loss structures are commonly visible in core and, except for the larger features, should be readily interpretable.

2.3.2.7 *Fossils*

Fossil fragments are commonly visible in well cuttings, but are difficult to recognize and interpret. The

best solution is to examine them in thin section, where distinctive features of internal structure may be apparent. A program of routine thin-section examination of fossiliferous sequences may be desirable and, particularly if the rocks are carbonates, this can be combined with lithologic analysis. An excellent textbook by Horowitz and Potter (1971) discusses the petrography of fossils in detail.

Macrofossils can rarely be satisfactorily studied in core except in the case of highly fossiliferous sections, such as reefs and bioherms. The reason is that the chance sections afforded by core surfaces and longitudinal cuts do not necessarily provide exposures of a representative suite of the forms present and, unlike the case for sparsely fossiliferous surface outcrops, there is no opportunity to break up more rock in a search for additional specimens. Fragments may be studied in thin section, as described previously, but the limitations on the quantity of material still apply.

Both cores and cuttings may be used by biostratigraphic specialists, who extract palynomorphs and microfossils from them by processes of deflocculation or acid dissolution. Much useful ecological information may be obtainable from these suites of fossils. For example, Mesozoic and Cenozoic foraminifera were sensitive to water depth (as are modern forms), and documentation of foram assemblages through a succession can permit a detailed reconstruction of the varying depths of marine depositional environments (Sect. 4.5.7).

2.3.2.8
Biogenic Sedimentary Structures

As in the case of most other mesoscopic features, very little can be seen in cuttings, whereas cores commonly contain particularly well-displayed biogenic sedimentary structures. Those confined to bedding planes may not be particularly easy to find, whereas burrows usually are easy to study and can provide invaluable environmental interpretation. Refer to the notes and references in Section 2.2 and to Section 4.5.7.

2.3.3
Sampling Plan

As in the case of surface sections, we are concerned here with sampling in three main categories for basin-analysis purposes: illustrative lithologic samples; samples for laboratory petrographic analysis using thin and polished sections, X-ray diffraction, etc.; and samples for biostratigraphic purposes.

Very limited quantities of material are available for any kind of sampling in subsurface sections. The cuttings and core stored in the laboratory are all that will ever be available and, once used, cannot be replaced. Thus, they should be used with care, and permission should always be sought from the appropriate company or government agency before removing any material for research purposes.

A core may provide excellent illustrative material for demonstrating lithology, sedimentary structures, and facies sequences. The Canadian Society of Petroleum Geologists has established a tradition of holding a "core conference" every year in one of the government core laboratories in Calgary [other societies, such as the Society for Sedimentary Geology (SEPM), are now copying this idea]. Each contributor to the conference provides a display of a selected core from a producing unit and uses this as a basis for presenting an interpretation of its geology, with emphasis on depositional environments, diagenesis, and petroleum migration history. The educational value of these conferences is inestimable, and they are always well attended. Some have resulted in the publication of well-illustrated proceedings volumes, such as Shawa (1974), Lerand (1976), and McIlreath and Harrison (1977). The SEPM has now published 20 "core workshop" notes.

Clearly, the use of well cuttings as illustrative material of this type is limited. However, cuttings may be sampled routinely for petrographic studies, using etching and staining techniques on the raw, unwashed cuttings or preparing polished or thin sections. Fortunately, very small samples are adequate for this kind of work.

Sampling for biostratigraphic purposes should always be carried out on the unwashed rather than the washed cuttings. Depth control is, of course, better for core, but in petroleum wells there is rarely adequate core for routine biostratigraphic sampling. Very rarely, a petroleum exploration well may be drilled by continuous coring for stratigraphic research purposes, and these provide ample sample material free of the problems of sample caving and depth lag. The same is true, of course, for DDH cores. The quantity of material required for biostratigraphic purposes depends on the fossil type under investigation. A few hundred grams is usually adequate for palynological purposes, whereas to extract a representative conodont suite from sparsely fossiliferous carbonate sediments may require several kilograms of material. In the latter case, samples from several depth intervals may have to be combined.

2.3.4
Plotting the Section

Well logging is a routine procedure, and most organizations provide standard forms for plotting graphic logs with a set of standard symbols and abbrevia-

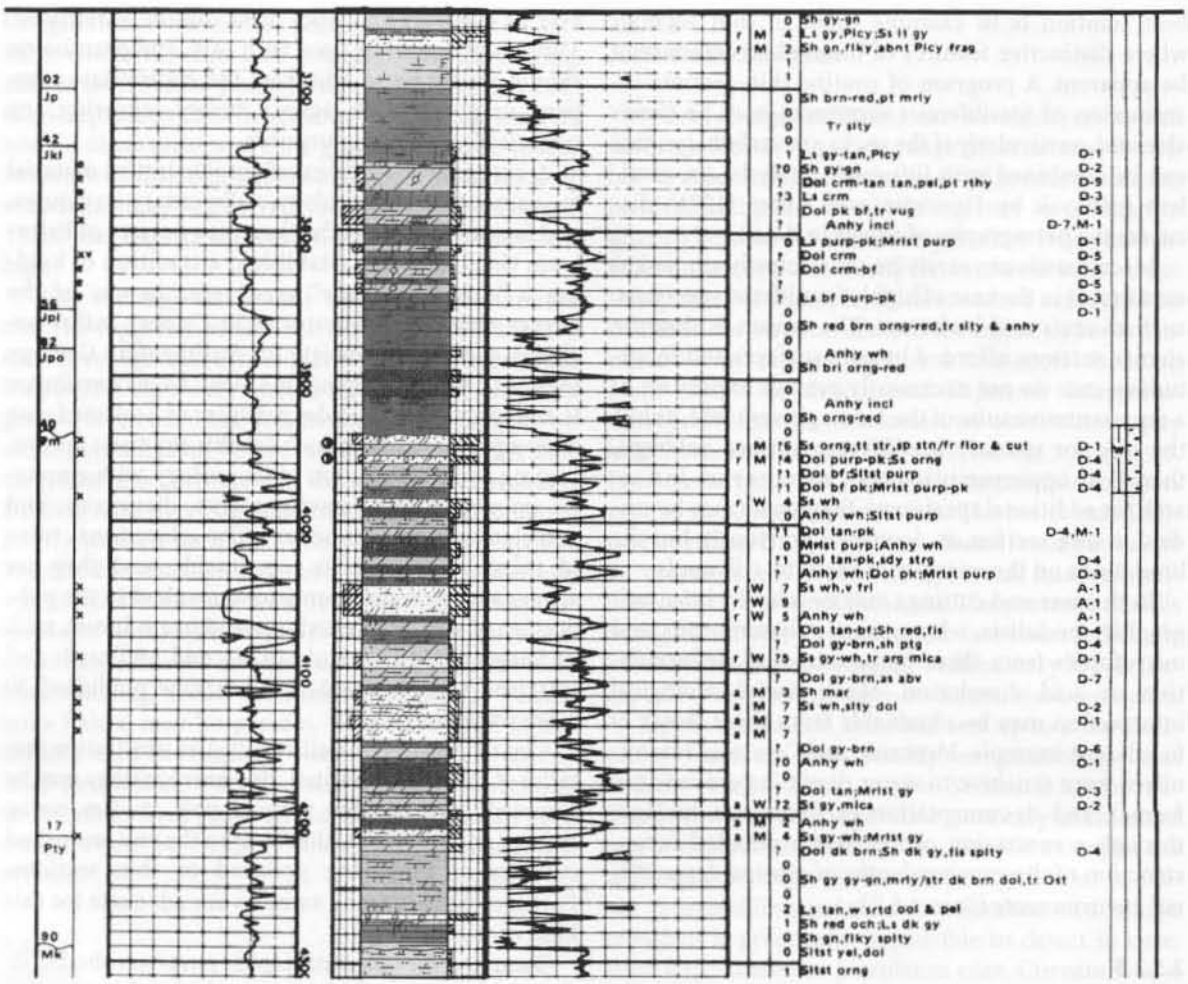


Fig. 2.28. Example of part of a log produced by the American/Canadian Stratigraphic Company

tions. That used by the American Stratigraphic Company and Canadian Stratigraphic Services is typical. An example is illustrated in Fig. 2.28. Lithology is shown by color in a column near the center of the log, and accessories, cements, fossil types, and certain structures are shown by symbols. To the left of this column are columns for formation tops and porosity type and grade (amount). Formation tops are shown by a formation code. They are interpreted and may be subject to revision. Porosity type is given in a standard code or symbol and porosity grade by a crude graph based on visual estimates. The depth column may be used with symbols to indicate hydrocarbon shows or stains. To the right of the lithology column is a column for crystal or detrital grain size based on visual estimates or measurements against a grain-size comparison set (Fig. 2.5). Both to the left and right of the lithology column are spaces for selected petrophysical logs. On the right-hand side of the log there is a space for typing in an abbreviated description of each lithologic interval. The American/Canadian Stratigraphic Company system uses a list of more than 450 abbreviations, covering almost every conceivable descriptive parameter. Remaining columns are used for grain rounding and sorting and for engineering data.

Much of this detail is not necessary for basin-analysis purposes. Fig. 2.29 illustrates the more limited range of symbols and codes used by Tassonyi (1969) in a study of the subsurface stratigraphy of the Mackenzie Basin in northern Canada. Fig. 2.30 illustrates one of his graphic logs. This style emphasizes lithology and other important stratigraphic variables.

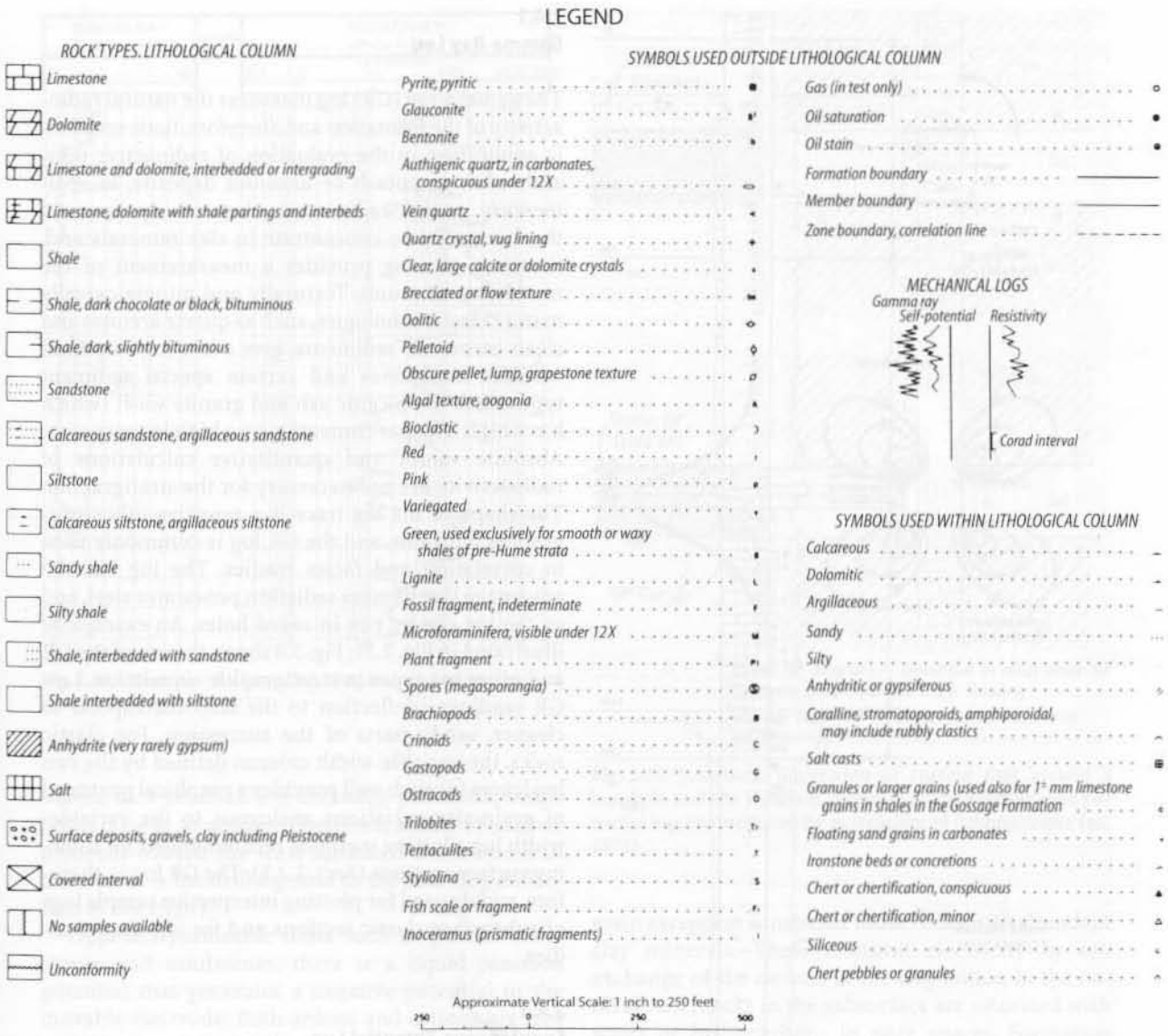
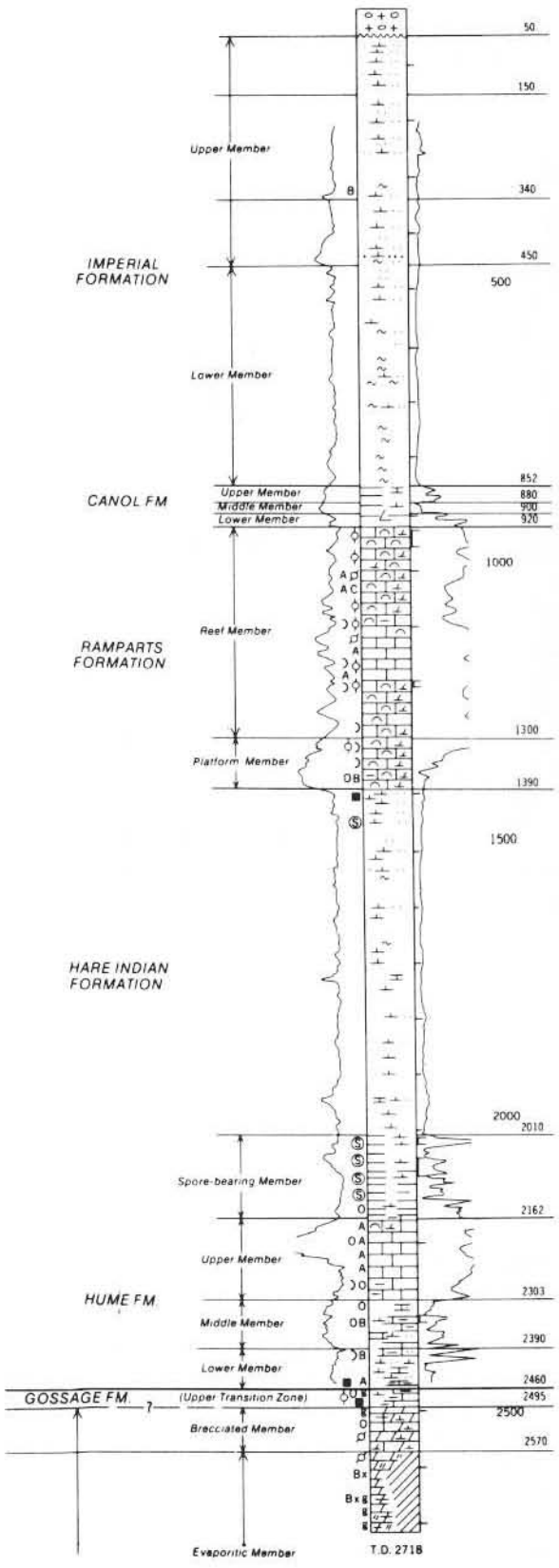


Fig. 2.29. Codes and symbols used in a regional subsurface study of Paleozoic and Mesozoic stratigraphy in the Mackenzie Valley region, Northwest Territories (Tassonyi 1969)

2.4 Petrophysical Logs

A wide range of physical parameters can be measured using tools lowered down a petroleum exploration hole. These give information on lithology, porosity, and oil and water saturation. In many cases, the measurements are not direct but require interpretation by analogy or by correlating values between two or more logs run in the same hole. The subject of petrophysics is a highly advanced one and is beyond the scope of this book. Some approaches are described by Pirson (1977); other are given in the various interpretation manuals issued by the logging companies. Other excellent texts that discuss geological interpretations are those by Cant (1992) and Doveton (1994), and Whittaker (1998) provided an up-to-date review of petrophysical logging methods.

Petrophysical logs are used routinely by stratigraphers and basin analysts to provide information on lithology and to aid stratigraphic correlation, as discussed in Sections 1.4.3 and 3.4.3. In this section, some of the principal log types are described briefly, and some demonstrations of their utility are given.



2.4.1 Gamma-Ray Log

The gamma-ray (GR) log measures the natural radioactivity of the formation and, therefore, finds economic application in the evaluation of radioactive minerals such as potash or uranium deposits. In sedimentary rocks, radioactive elements (potassium, thorium) tend to concentrate in clay minerals and, therefore, the log provides a measurement of the muddiness of a unit. Texturally and mineralogically mature clastic lithologies, such as quartz arenites and clean carbonate sediments, give a low log response, whereas mudstones and certain special sediment types, such as volcanic ash and granite wash (which has a high feldspar content), give a high log response. Absolute values and quantitative calculations of radioactivity are not necessary for the stratigrapher. The shape of the log trace is a sensitive lithostratigraphic indicator, and the GR log is commonly used in correlation and facies studies. The log has the advantage that gamma radiation penetrates steel, and so the log can be run in cased holes. An example is illustrated in Fig. 2.31; Fig. 3.8 shows the use of the GR and other log types in stratigraphic correlation. Low GR readings (deflection to the left) correspond to cleaner, sandy parts of the succession. For clastic rocks, the variable-width column defined by the two log traces for each well provides a graphical portrayal of grain-size variations analogous to the variable-width log-plotting methods recommended for drawing surface sections (Sect. 2.2.5). The GR log is, therefore, widely used for plotting interpretive sample logs of subsurface clastic sections and for log-shape studies.

2.4.2 Spontaneous-Potential Log

The spontaneous-potential (SP) log curve is a recording of the potential difference between a movable electrode in the borehole and the fixed potential of a surface electrode. SP readings mainly record currents of electrochemical origin, in millivolts, and are either positive or negative. There are two separate electrochemical effects, as illustrated in Fig. 2.32. When the movable electrode is opposite a muddy unit, a positive current is recorded as a result of the membrane potential of the mudstone. The latter is permeable to most cations, particularly the Na^+ of saline formation

Fig. 2.30. A subsurface log drawn to emphasize stratigraphically important details using the codes and symbols given in Fig. 2.29. Devonian period, Mackenzie Valley, Northwest Territories. Spontaneous potential and resistivity logs have been redrawn beside the lithologic column. (Tassonyi 1969)

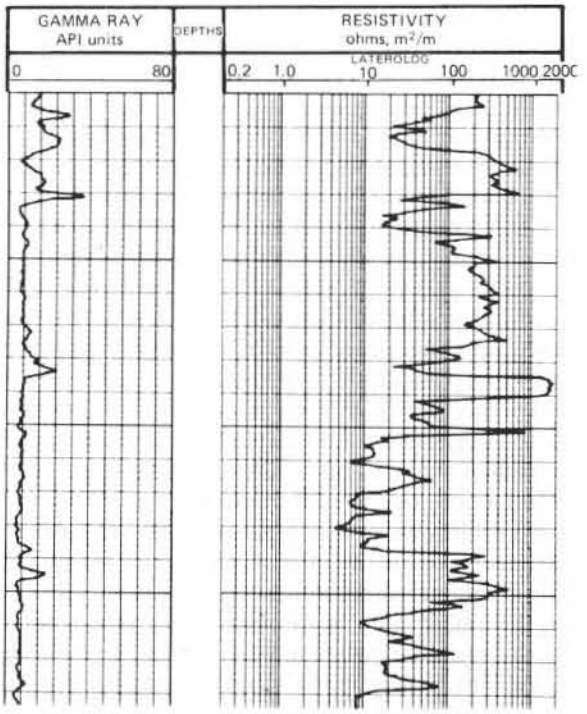


Fig. 2.31. Example of a laterolog. (Reproduced by permission of Schlumberger Inc. 1972)

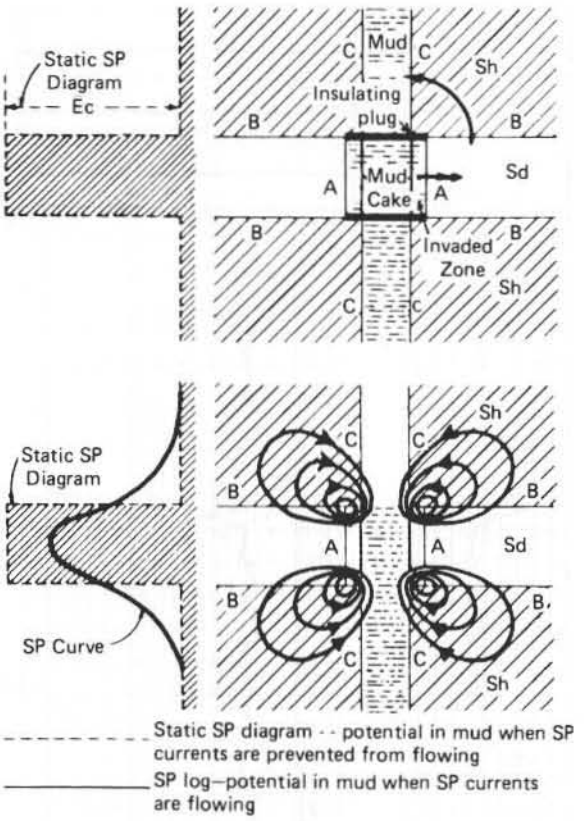


Fig. 2.32. Schematic illustration of current flow around a borehole and the resultant response of the spontaneous potential log. (Reproduced by permission of Schlumberger Inc. 1972)

waters, as a result of ion-exchange processes, but is impermeable to anions. Therefore, a flow of cations proceeds toward the least saturated fluid, which in most cases is the drilling mud in the hole (log deflection to the right).

Opposite permeable units such as porous limestones and sandstones, there is a liquid junction potential that generates a negative potential in the movable electrode. Both anions and cations are free to diffuse into the drilling mud from the more concentrated saline formation waters, but Cl^- ions have the greater mobility, and the net effect is a negative charge (deflection to the left).

The SP-log trace, particularly in clastic sequences, is very similar in shape to that of the GR log (examples are given in Figs. 2.33, 3.8, 3.9), and the two may be used alternatively for correlation purposes, as illustrated in Fig. 3.8. The SP curve is normally smoother, and this log type does not offer the same facility for identifying thin beds. SP deflections are small where the saltiness and resistivities of the formation fluids and the drilling mud are similar.

2.4.3 Resistivity Logs

Most rock types, in a dry state, do not transmit electric currents and are, therefore, highly resistive. The main exception consists of those rocks with abundant clay minerals. These transmit electricity by ion exchange of the cations in the clay lattice. In the natural state, rocks in the subsurface are saturated with water or hydrocarbons in pore spaces. Formation waters are normally saline and, thus, act as electrolytes. Therefore, the resistivity depends on the salinity and continuity of the formation waters. The latter depends, in turn, on porosity and permeability, so that resistivity is lowest in units such as clean sandstone and vuggy dolomite and highest in impermeable rocks such as poorly sorted, dirty, silty sandstones and tight carbonates. Evaporites and coal are also highly resistive. Metallic ores have very low resistivity. Oil is highly resistive and so, under certain conditions, resistivity tools may be used to detect oil-saturated intervals.

A wide variety of resistivity-measurement tools has been devised, as listed below. Published logs are commonly identified by the appropriate abbreviation, as shown in Fig. 3.8.

A. Electrical survey (ES)
B. Laterolog (LL)

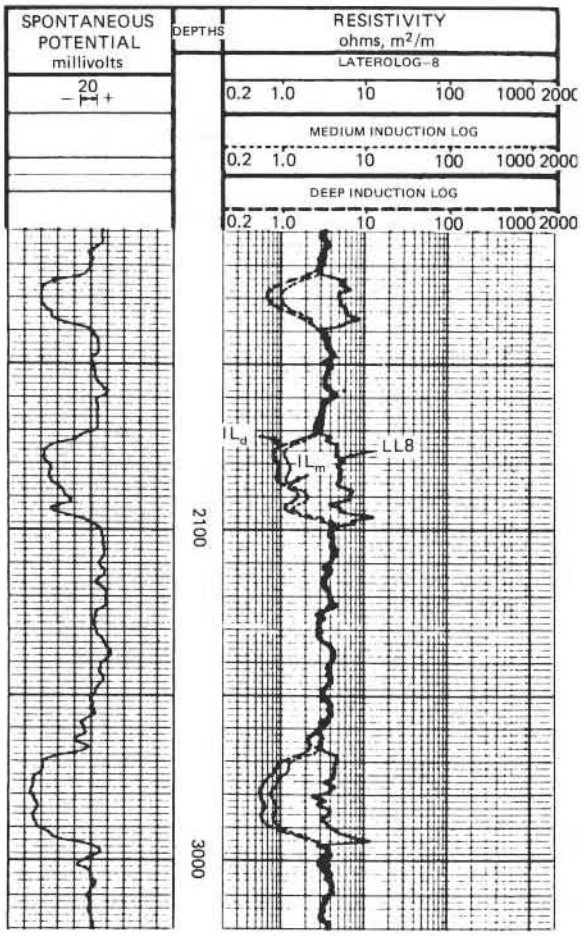


Fig. 2.33. Example of a dual-induction laterolog. (Reproduced courtesy of Schlumberger Inc.)

C. Induction electrical survey (IES)
D. Dual-induction laterolog (DIL)
E. Microlog
F. Microlaterolog

Laterolog, microlog, and microlaterolog are registered trade names for tools developed by Schlumberger Inc.

The conventional ES was, together with the SP log, the only logging tool available for many years during the early days of petrophysical logging before the war. In an ES, an electrical current is passed into the formation via an electrode, and this sets up spherical equipotential surfaces centered on the electrode–rock contact. Three additional electrodes are positioned on the tool to intersect these surfaces at set distances from the first electrode. The standard spacing is:

A. Short-normal electrode: 40.6 cm (16 inches)
B. Medium-normal electrode: 1.63 m (64 inches)
C. Lateral electrode: 5.69 m (18 ft 8 inches)

The wider the spacing the deeper the penetration of the current into the formation. This permits comparisons between zones close to the hole permeated by drilling mud, and uninvaded zones farther out. The wide spacing of the electrodes also means that the ES tool is not sensitive to thin beds, and so it is not a very satisfactory device for stratigraphic studies.

The LL uses an arrangement of several electrodes designed to force an electrical current to flow horizontally as a thin sheet out from the borehole. A monitoring electrode measures a variable current that is automatically adjusted to maintain this pattern as the tool passes through lithologies of variable resistivity. This device is much more responsive to thin beds. An example, run together with a GR survey and plotted on a logarithmic scale, is shown in Fig. 2.31.

Induction logs were developed for use with oil-based drilling muds which, because they are nonconductive, make the use of electrodes unsatisfactory. A high-frequency alternating current is passed through a transmitting coil, creating a magnetic field that induces a secondary current to flow in the surrounding rocks. This, in turn, creates a magnetic field that induces a current in a receiver coil. The strength of the induced current is proportional to the formation conductivity.

The DIL survey is a combination of two induction devices and an LL device, with different formation penetration characteristics. It is normally run with a SP tool. An example is illustrated in Fig. 2.33. Separation between the three resistivity curves occurs opposite permeable units, where the low resistivity of the saline formation waters contrasts with the higher resistivity of the zone close to the borehole, which has been invaded by low-salinity drilling mud. The deep-penetration induction log (IL_d), therefore, gives the lowest resistivity reading and the shallow penetration laterolog (*LL8*) the highest. The presence of these permeable zones is confirmed by the SP curve, which has a pattern very similar to that of the induction logs in the example shown.

The microlog is the most sensitive device for studying lithologic variations in thin-bedded sequences. Its primary petrophysical purpose is to measure the resistivity of the invaded zone. The principle is as follows: during the drilling process, mud enters permeable beds and hardens on the surface as a mud cake up to about 1 cm thick. No mud cake is formed opposite impermeable units. The microlog tool consists of two closely spaced electrodes. Opposite the mud cake, they record the low resistivity of the mud itself, whereas, opposite impermeable units, they record the generally higher resistivities of uninvaded rocks. The presence of mud cake is confirmed by a caliper log, which is sensitive to the slight reduction in hole diameter when a mud cake is present. Muddy

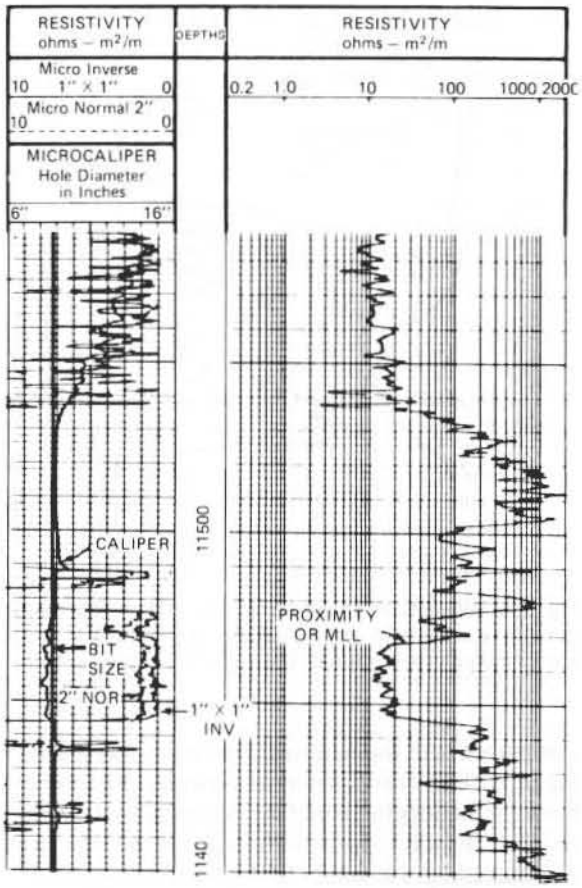


Fig. 2.34. Example of a microlog. (Reproduced courtesy of Schlumberger Inc.)

units commonly cave and give a very erratic caliper log. The microlog readings are also likely to be erratic because of the poor electrode contact. All these responses are illustrated in Fig. 2.34. The microlaterolog is a more sensitive version of the laterolog. Use of both the microresistivity devices permits accurate determinations of permeable sandstone and carbonate thickness, which are of considerable use in regional subsurface-facies studies (Chap. 5).

2.4.4 Sonic Log

The sonic tool consists of a set of transmitters for emitting sound pulses and a set of receivers. The fastest path for sound waves to travel between transmitters and receivers is along the surface of the hole – in the rock itself rather than through the mud or the actual tool. The time of first arrival of the sound pulses is, therefore, a measure of formation density, which depends on lithology and porosity. The sonic log is normally run with a GR tool.

The sonic tool has two main uses: the estimation of porosity where lithology is known, and the calibration of regional seismic data (Sect. 5.4.1). For the latter purpose, a computer in the recording truck at the well head integrates the travel time over each increment of depth as the survey is run. Every time this amounts to 1 ms, a small pip appears on a track down the center of the log.

2.4.5 Formation-Density Log

The formation-density tool contains a radioactive source emitting gamma rays. These penetrate the formation and collide with it in a process known as Compton scattering. The deflected gamma rays are recorded at a detector on the tool. The rate of scattering is dependent on the density of the electrons in the formation with which the gamma rays collide. This, in turn, depends on rock density, porosity, and composition of the formation fluids.

2.4.6 Neutron Log

For this log, a radioactive source emitting neutrons is used. These collide with the nuclei of the formation material, with a consequent loss of energy. The greatest loss of energy occurs when the neutron collides with a hydrogen nucleus, and so the total loss of energy depends mainly on the amount of hydrogen present, either in formation waters, hydrocarbons, or bound water in clay minerals, gypsum, etc. The detector measures either the amount of scattered low-energy neutrons or the gamma rays emitted when these neutrons are captured by other nuclei.

2.4.7 Crossplots

Where the formation is known to consist of only two or three rock types, such as sandstone-siltstone-mudstone or limestone-dolomite, combinations of two or more logs can be used to determine lithology, porosity, and hydrocarbon content. These crossplots are, therefore, of considerable stratigraphic use where only generalized lithologic information is available from well cuttings.

For clean, nonmuddy formations, combinations of the sonic, formation-density, and neutron logs are the most useful. For example, Figs. 2.35 and 2.36 show the neutron-density combination. These logs are commonly calibrated in terms of apparent limestone porosity; that is, if the rock is indeed limestone, its porosity has the value indicated. For other lithologies, the porosity estimate will be in error but, by reading

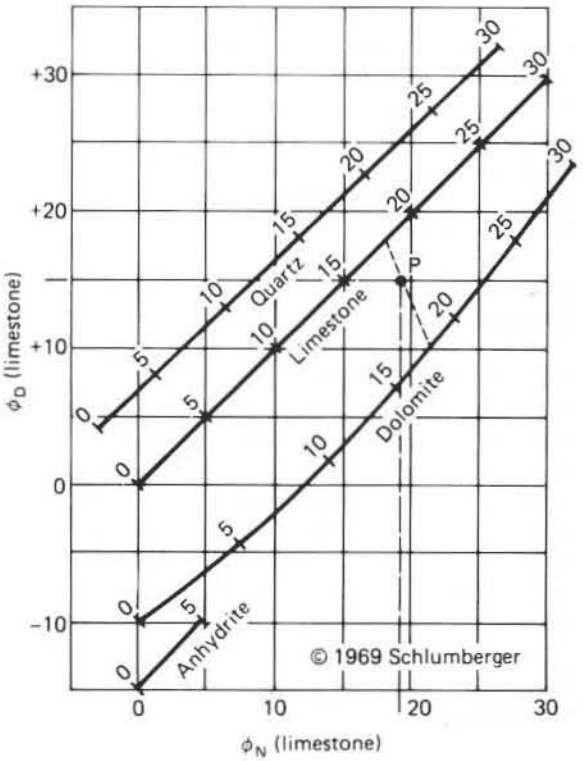


Fig. 2.35. A neutron (ϕN)-density (ϕD) crossplot, showing how readings of the two logs can be used to determine lithology and porosity. (Reproduced courtesy of Schlumberger Inc.)

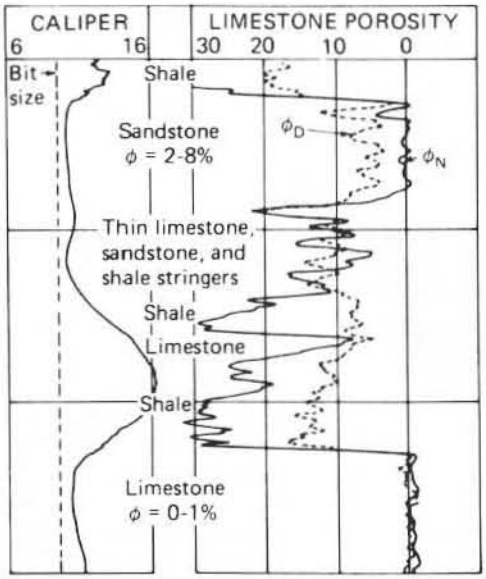


Fig. 2.36. Example of a neutron-density overlay, illustrating how curve separation and deflections can be used to determine lithology. (Reproduced courtesy of Schumberger Inc.)

values for both logs, it is possible to determine both lithology and correct porosity. The curves in Fig. 2.35 give ranges of actual porosity readings for four principal rock types. To show how this graph can be used, compare it with the neutron-density overlay given in Fig. 2.36. Such overlays may be provided on a routine basis by the logging company or can be redrawn on request. The thick sandstone interval at the top can be recognized by the distinctly higher readings on the density curve. Values range from about 4 to 12 porosity units, whereas those on the neutron curve are mostly close to zero. Examination of Fig. 2.35 shows that only sandstone can give this combination. The thick limestone at the bottom of Fig. 2.36 is indicated

Table 2.5. Interpretation of density-neutron overlays, assuming single mineral lithologies and liquid-filled formations[a]

Curve relationship	Approximate difference	Additional confirming criteria	Sonic-log confirmation (if no secondary porosity)	Probable matrix material	Approximate true porosity
At high resistivities:					
$\phi_D \gg \phi_N$	40 p.u.	ϕ_N=4, ϕ_D=43.5	$\Delta t = 67$	Salt	$\phi \approx \phi_N - 4$
	$\phi_{SNP} > 40$,	Low gamma ray,	$110 < \Delta t < 140$	Coal	
$\phi_D > \phi_N$	$50 < \phi_D < 80$, $1.3 < \varrho_b < 1.5$	high resistivity			
$\phi_D > \phi_N$	5–6 p.u.		$\Delta t - 1.4\phi_N \approx 58$	Sandstone	$\phi \cong 45\ (\phi_D + \phi_N)/2$
$\phi_D = \phi_N$	0 p.u.		$\Delta t - 1.4\phi_N \approx 47$	Limestone	$\phi = \phi_D = \phi_N$
$\phi_D < \phi_N$	8–13 p.u.		$\Delta t - 1.4\phi_N \approx 41$–$43$	Dolomite	$\phi \cong 0.5\ (\phi_D + \phi_N) + 2.5$ (for $\phi > 8$ p.u.) $\phi \cong 0.7\ \phi_N$ (for $\phi < 8$ p.u.)
$\phi_D < \phi_N$	16 p.u.		$\Delta t - 1.4\phi_N \approx 50$	Anhydrite	$\phi \cong \phi_N$
$\phi_D \ll \phi_N$	10–30 p.u.	See gamma-ray log		Shale	
At high resistivities:					
$\phi_D \ll \phi_N$	28 p.u.	$\phi_N = 49$, $\phi_D = 21$	$\Delta t = 52.0$	Gypsum	$\phi \cong \phi_D - 21$

p.u., limestone porosity units; Δt, interval transit time; ϱ_b, bulk density

[a] Reproduced courtesy of Schumberger Inc.

by the near coincidence of the two curves. Dolomite or anhydrite would be suggested by relatively higher neutron readings. Mudstones can commonly be recognized by very high neutron readings relative to density because of the abundant water in the clay mineral lattice. This would be confirmed by the gamma ray response or the SP or caliper log, if available. Table 2.5 summarizes the use of the sonic, density, and neutron logs in lithology identification.

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Stratigraphic Correlation

3.1 Introduction

Whether the geologist is dealing with roadside outcrops, subsurface data from a petroleum play, or regional seismic lines, one of the first problems to be encountered is that of stratigraphic correlation. In order to reconstruct depositional environments and paleogeography or to trace a unit of economic interest, the geologist must be able to define a stratigraphy and trace it from one location to another. The procedures for carrying this out are the subject of this chapter.

Formal stratigraphic practices, including the definition of formations and of stages, had their origins in nineteenth-century field geology and have evolved into a set of carefully defined procedures for naming and correlating the various kinds of stratigraphic units. They have also been applied successfully to subsurface well data. These methods are mainly based on detailed lithostratigraphic and biostratigraphic information, the collection of which is discussed in Chapter 2. Other important aids to correlation, including radiometric dating and magnetic-reversal stratigraphy, are discussed in Section 3.7.

A quite different approach to correlation is taken by those dealing with regional reflection seismic data. Seismic work in frontier regions and the deep-reflection profiles now being produced by groups such as the Consortium for Continental Reflection Profiling (COCORP) provide sweeping regional cross sections, within which correlations at the detailed level may be far from clear.

With outcrop and well data, the problem may be to establish the regional framework from a mass of local detail, whereas, with seismic data, it is the detail that is hard to see. The ideal combination is, of course, a basin with a network of seismic lines tied to key exploration holes. If there are outcrops too, so much the better, although, as pointed out in Section 2.2.1, outcrops around a basin margin may not be very representative.

These differences in data type and scale have led to two different approaches to stratigraphic correlation. In industry exploration work in frontier regions, particularly in the great offshore basins, a rather informal, pragmatic approach is taken to such topics as biostratigraphic zonation and the naming of formations. The broad picture can be derived from seismic cross sections, and the details gradually resolve themselves as more well data become available. Application of various basin-mapping methods (Chap. 5) and use of the genetic, depositional systems approaches and sequence stratigraphy concepts (Chap. 6) are of particular value here.

In the absence of seismic data, it is necessary to construct the forest from the individual trees. The stratigraphic framework in most well-explored (mature) basins was built up this way, and the work has usually been accompanied by considerable controversy as local specialists have argued about the relationships between the successions in different parts of a basin or between outcrop and subsurface units. Sequence concepts can now make this task easier, although the work is usually not without its problems.

Whether a basin-analysis exercise starts from seismic sections or from outcrop work, it is necessary, eventually, to document the fine detail of the stratigraphy by establishing a formal lithostratigraphic and (if in Phanerozoic strata) a biostratigraphic framework. Such detail is beyond the needs of most exploration companies and is an area of research commonly taken over by state geological surveys and individuals in academic institutions, although the database and expertise built up in industry may form an essential component. This is one area in which the government core and data repositories referred to in Chapter 2 can prove their usefulness.

Every local biostratigraphic, radiometric, and magnetostratigraphic study can potentially contribute to the long-term effort to perfect a global chronostratigraphic (time) scale. Several commissions and subcommissions and numerous working groups of the International Union of Geological Sciences (IUGS) have been tackling this problem for many years, as have many national groups. Some of the results are reported in this chapter, and the interested reader may wish to examine the IUGS journal *Episodes*, which reports on the activities of these various groups and announces important publications. Some

references to specific articles in this journal are given elsewhere in this chapter.

This chapter is intended to provide an introduction to practical research methods. Chronostratigraphic (including biostratigraphic) research must form an integral part of any ongoing basin analysis project unless it is strictly local in scope. The work is usually performed by specialists. Lithostratigraphic correlation methods are also described here, although the procedures for erecting formal, named units (included here for consistency and completeness) can be left to advanced stages of the analysis. Such naming is best carried out by an individual with sedimentological training so that the depositional systems and sequence concepts described in Chapter 6 can be incorporated into the work.

3.2 Types of Stratigraphic Unit

Rocks may be described in terms of any of their physical, chemical, organic, or other properties, including lithology, fossil content, geochemistry, mineralogy, electrical resistivity, seismic velocity, density (gravity), magnetic polarity, or age. Theoretically, any of these properties may be used for description and correlation, and most are so used for various purposes. In practice, lithology is one of the most important criteria; fossil content is also crucial for rocks of Phanerozoic age. Magnetic polarity is finding increasing application as a correlation tool, particularly for rocks of the younger Mesozoic and Cenozoic eras, and radiometric ages are used to assign absolute ages to biostratigraphic, magnetic, and other chronostratigraphic units. Other geophysical properties are used in the early reconnaissance stages of exploration of a sedimentary basin. Not all these properties will necessarily give rise to the same correlations of a given rock body; for example, it is commonly difficult to relate geophysical properties precisely to lithology. Therefore, no single type of stratigraphic unit can be used to define all the variability present in nature.

Reflection-seismic data have been widely used in the exploration of sedimentary basins since the 1970s, making use of the concepts of seismic stratigraphy (Sect. 5.4.1) and with particular application to the interpretive field of sequence stratigraphy (Chap. 6). Sheriff (1976) pointed out that for reflecting events to be distinguished on seismic data they must each represent a clear velocity contrast and must be at least the equivalent of a quarter wavelength apart. At shallow depths, velocities are in the range of 1.5–2.5 km/s and reflections are relatively high frequency (about 5–100 Hz), so that a quarter wavelength is on the order of 5–12 m. At greater depths, typical reflection wavelengths increase considerably. Therefore, stratigraphic resolution is fairly coarse (Fig. 1.6). Early work in seismic sequence stratigraphy had portrayed sequence boundaries as distinctive reflection surfaces correlatable over wide distances (Mitchum et al. 1977), but improved acquisition and processing techniques have demonstrated that the broad, through-going correlations on which this early work were based may be suspect. Cartwright et al. (1993) demonstrated that many of the critical features of sequence architecture, including the sequence boundaries, break down upon detailed examination into complex reflection patterns representing local facies variability, and major through-going surfaces may not be present (see also Miall 1997; Sect. 2.2). Seismic data are essential for studying large-scale stratigraphic features such as depositional systems (Sect. 5.4.1; Chap. 6) and regional (or global) sequences (Sect. 8.3), but are of less use in the development of the refined stratigraphic subdivisions that are the subject of this chapter. The reader interested in the application of seismic methods to basin analysis should, therefore, turn to the chapters noted.

Two basic categories of stratigraphic information are essential for the complete documentation of the stratigraphic record: (1) descriptive lithic units; and (2) geochronologic information, dealing with correlation and the age of the strata (Harland 1993).

The most important types of stratigraphic units are:

A. *Lithostratigraphic units:* these are based on observable lithologic features, including composition and grain size, and possibly also including certain basic sedimentological information, such as types of sedimentary structures and cyclic sequences.
B. *Biostratigraphic units:* these are based on fossil content. Life forms evolve with time, permitting subdivision of strata on the basis of changes in the fauna or flora they contain.
C. *Unconformity-bounded units:* these are units bounded above and below by unconformities. They may consist of any kinds of rocks, be they igneous, metamorphic, or sedimentary. Unconformity-bounded sedimentary successions form the basis for sequence stratigraphy.
D. *Chronostratigraphic units:* these comprise an interpretive stratigraphy, in contrast to the first three types of unit, which are strictly descriptive. Chronostratigraphy concerns itself with correlation and the age of the strata, which may be determined by a variety of means, of which the most important are fossil content, radiometric dating, and magnetic polarity.

Both lithostratigraphic and biostratigraphic units tend to be local in extent. Lithologic character de-

pends on the depositional environment, sediment supply, climate, rate of subsidence, etc., all of which can vary over short distances. Lithostratigraphic units are diachronous to a greater or lesser degree; that is, they represent a different time range in different places, reflecting gradual shifts in environment, for example, during transgression or regression. The limits of a stratigraphic unit are either its erosional truncation at the surface or beneath an unconformity or a facies change into a contemporaneous unit of a different type. A special type of lithostratigraphic unit is that formed by stratigraphic events, which have widespread depositional effects within very short time spans (Ager 1981). Examples of such events are volcanic-ash falls, the deposits of violent storms and tsunamis, and regional or global sea-level changes. As discussed in Section 3.4.3, recognition of such processes and their results can form the basis for a special kind of stratigraphy, termed event stratigraphy, which may have considerable chronostratigraphic significance.

Biostratigraphic units are based on fauna or flora, the distribution of which is ecologically controlled. Also, contemporaneous faunas located in ecological niches that are similar but geographically isolated may show subtly different evolutionary patterns, making comparisons or correlations between the areas difficult (this is the subject of **biogeography**). All life forms evolve with time, so that faunas and floras show both spatial and temporal limits on their distribution.

The unconformities that demarcate unconformity-bounded units are caused by subaerial and marine erosion during times of low sea level. These events typically are widespread. They may be caused by tectonic elevation of the basement or by eustatic changes in sea level. In either case, the unconformity surfaces define units of considerable lateral, potentially even global, extent (although there has been controversy on this latter point, as discussed in Sect. 8.5). These units, therefore, provide an excellent basis for the regional subdivision of a basin fill, and their interpretation throws considerable light on regional tectonic evolution. Sequence stratigraphy has become the method of choice for subsurface mapping by the petroleum industry precisely because of its practical utility in focusing on and documenting these unconformity-bounded successions (Chap. 6).

Chronostratigraphy attempts to resolve the difficulties in regional and global correlation by establishing a global, time-based reference frame. However, the accuracy of chronostratigraphic correlation is only as good as that of the time-diagnostic criteria on which it is based. Error and imprecision are still rife, as indicated in Section 3.7.

The evolution of these four types of units has had a long and complex history (Hancock 1977), and there is still not unanimous agreement on definitions. Hedberg (1976), Hancock (1977), and Harland (1978, 1993) discussed some of the early practical and philosophical problems. Nineteenth-century geological practice did not distinguish lithology from age, causing severe correlation problems wherever a facies change or a diachronous boundary occurred. More recently, there has been controversy over whether the rocks or the more abstract evidence of age that they contain should form the primary basis of chronostratigraphy. The discussions are likely to seem somewhat academic and theoretical to the average basin analyst and will not be discussed at length here. A useful summary of the problems, with particular reference to practical nomenclature difficulties in the Gulf Coast Paleogene succession, was provided by Siesser (1984). Holland (1986, 1998) surveyed the recent global effort to establish standard stratigraphic scales and touched, once again, on the debate about the meaning, value, and relationships between chronostratigraphic and biostratigraphic units. With the introduction of several new tools for determining relative and absolute ages (Sect. 3.7), the debate has become less important.

3.3 Stratigraphic Procedures

3.3.1 North American and International Stratigraphic Codes

Stratigraphic geology was founded by the English geologist William Smith toward the end of the eighteenth century, although the oldest principle in stratigraphy, the Law of Superposition of Strata, was developed by Steno in the seventeenth century. Since the early 1800s, a wide variety of local stratigraphies, stratigraphic methods, and philosophies have evolved in different parts of the world. Inevitably, this has generated much controversy, ranging from disagreements over local correlations to arguments about what criteria should be used to define the composition and limits of global chronostratigraphic units. The first serious attempt to resolve the broader problems was started in 1952 at the 19th International Geological Congress in Algiers. At this time, the International Commission on Stratigraphic Terminology was created. Since 1965, this commission, now called the International Subcommission on Stratigraphic Classification (ISSC), has worked under the aegis of the International Union of Geological Sciences (IUGS). Several drafts of a guide to procedures and terminology were produced and published by this commission, based on numerous discussion

meetings at International Geological Congresses plus correspondence with various official national geological bodies and individual specialists. More than 125 individuals from over 40 countries and 39 official national or regional committees were involved in this work. The culmination of this effort was the publication in 1976 of the *International Stratigraphic Guide*, a guide to stratigraphic classification, terminology and procedure, edited by H.D. Hedberg (1976). The purpose of this guide was to set out those practices and procedures that have been broadly accepted by the international stratigraphic community. It was hoped that, eventually, this guide would be universally adopted, although at present there are many national and regional codes that differ to some degree from that of the ISSC.

Much work remained to be done at the time the ISSC guide was published, particularly in the areas of Precambrian and Quaternary stratigraphy and the relatively new field of magnetic reversals. Various reports and recommendations concerning particular stratigraphic procedures have been published since the ISSC guide by specialist working groups of the IUGS. Some of these are referred to here.

In 1983, a new *North American Stratigraphic Code* (NASC) was published, providing definitions of several new kinds of unit but otherwise following the recommendations of the ISSC Guide as much as possible (North American Commission on Stratigraphic Nomenclature, 1983). The new kinds of units defined in the NASC include **lithodemic units**, which are intrusive, highly deformed or highly metamorphosed rocks; **magnetostratigraphic units**, which are based on remanent magnetic properties and are discussed in Section 3.7.5 of this book; **pedostratigraphic units**, which are fossil soil zones; and **allostratigraphic units**, which are unconformity-bounded units discussed further in Section 3.6 of this book. Allostratigraphic units may be considered a variety of an important and rather more general type of stratigraphic unit that has not been formalized in stratigraphic codes at this time; this is **event stratigraphy**, discussed in Section 3.4.3. Finally, the NASC erected an entirely new type of unit based on age determinations, called the **diachronic unit**, which "comprises the unequal spans of time represented either by a specific lithostratigraphic, allostratigraphic, biostratigraphic, or pedostratigraphic unit, or by an assemblage of such units." Diachrons may be subdivided, in order of decreasing rank, into episodes, phases, spans, and clines. The introduction of these terms seems rather superfluous to this writer, and I predict that they will not be widely used by basin analysts. An older term that was introduced with the same purpose is **magnafacies**, defined in the American Geological Institute (AGI) *Glossary of Geology* as "a major, continuous, and homogeneous belt of deposits that is distinguished by similar lithologic and paleontologic characters and that extends obliquely across time planes or through several defined chronostratigraphic units." Bridge and Gordon (1985) used the term "Catskill magnafacies" for one of the Paleozoic clastic wedges derived from the Appalachian orogen. For large, relatively homogeneous but diachronous bodies of rock such as syntectonic clastic wedges, this term would seem to be appropriate and useful.

Not included in the NASC are **petrofacies** and **petrosomes**, which are lithostratigraphic units characterized by detrital petrographic composition. Recognition of such units may provide a useful mapping tool (Sects. 5.5 and 9.4; Dickinson and Rich 1972).

A second edition of the ISSC guide was published in 1994 (Salvador 1994). As noted in the preface to this new edition, the guide has been very successful in stabilizing the evolution of concepts and terminology and providing a framework upon which many national stratigraphic codes have been built. The first edition was translated into Chinese, French, Russian, Spanish, and Turkish. In the second edition, new material of relevance to this book include sections dealing with unconformity-bounded units and magnetic-polarity units, plus an extensive bibliography.

The procedures outlined in this chapter are based mainly on the ISSC guide and the North American code, to which the reader is referred for further discussion regarding particular points of interest or concern. Owen (1987) provided a useful discussion and review of the NASC and offered many practical tips that the reader may also find useful.

3.3.2 Establishing Named Units

The basis for all stratigraphic documentation consists of named lithostratigraphic and biostratigraphic units (and now, allostratigraphic units). The establishment of a system of names should not be the first objective of the basin analyst, as explained in Section 1.2.3, but without the units it is difficult to draw geological maps or write intelligible reports.

Named units are based on the establishment and description of a **type section** or **stratotype**. Detailed procedures for the selection of a stratotype have been described by Cowie (1986). Briefly, these are as follows. The stratotype may be a well-exposed surface section, a mine section, or a well section, chosen to typify the unit. The stratotype should be as complete as possible. It should, preferably, include the upper and lower contacts of the unit and should be as nearly as possible continuous and unfaulted. Surface sections should not include long covered intervals or be heavily weathered or only exposed under exceptional

circumstances (such as a river bed exposed during drought). Subsurface sections should be accessible via a record of well cuttings or (better still) core stored in a permanent repository. For both surface and subsurface sections, as much descriptive documentation as possible should be available in the public records of a state geological survey, museum, or university. This should include precise location data, descriptive logs, geophysical logs (for subsurface sections), and as much sample material as can conveniently be curated. Important stratotypes, such as those defining the major periods and series, are decided by working groups of the International Commission on Stratigraphy, as discussed further in Section 3.7.3.

Additional reference sections may be designated to illustrate any features not covered in the type section. Preferably, where a unit occurs in both the surface and subsurface, the type section should be chosen at the surface, but it may be useful to designate a subsurface reference section, especially as this will serve to establish the geophysical-log character of the unit. Cope (1996) argued for the acceptance of what he termed **secondary standard stages** "for use in those cases faunal differences [with the standard stages] make correlation by all normal means impossible."

A named unit is not considered established until it is published in a recognized, refereed, scientific medium. This could include national and international scientific journals and state geological-survey publications. Local journals, bulletins, and special reports issued by museums or universities are rarely a desirable way to establish a stratigraphic unit, because they commonly do not receive sufficiently wide circulation and may not be rigorously edited. Names used only in company reports or in trade magazines are not considered established but may, of course, be widely used in internal, informal communication. Abstracts of talks presented at meetings are also not acceptable as a method of formal publication. Many informal suites of named or numbered sequences are now being published in the stratigraphic literature. Though this is useful for purposes of communication, such names must remain "unofficial" and subject to revision until they have been formally documented and named following the procedure set out here. Named allostratigraphic units are now being established to accomplish this purpose, as noted below.

The purpose of these publishing rules is to ensure that only the most useful, well-defined names receive public circulation. The literature is full of poorly-defined stratigraphic names, and much ink has been wasted on controversies over stratigraphic terminology. For this reason, it is advisable that, before attempting to establish or revise any stratigraphic nomenclature, the worker submit the proposals to a formal local or national stratigraphic nomenclature committee (this is a condition of publication for most state geological surveys). Many journal editorial boards include such a committee, and the state geological surveys invariably do. The advice might seem somewhat bureaucratic, but geological surveys are in business to establish the local geological framework, and their employees usually include many of the experts on the local stratigraphy. Their view of the subject tends to be more regional in scope than that of the average graduate student or mine geologist, and they are usually better acquainted with stratigraphic procedures. (However, the brash young student may delight in an opportunity to overturn the shibboleths of the established order, as personified by a body of aging civil servants. Thus is progress achieved – sometimes.)

The published description of a new or revised stratigraphic unit should include:

1. The name, type, and rank of the unit
2. The location of type and reference sections, including a map or aerial photo
3. A detailed description of the unit at the type section, including its nature and height in the section or the well depth of the contacts
4. Comments on the local or regional extent of the unit and its variability
5. A graphic log of the unit (optional, but desirable, for lithostratigraphic units only, and should include geophysical logs for subsurface sections)
6. A statement of the location of curated reference material
7. A discussion of the relationship of the unit to other contemporaneous stratigraphic units in adjacent areas

New or revised stratigraphic names should only be proposed after a careful evaluation of existing terminology and an examination of regional variability beyond the confines of the immediate project area. Both biostratigraphic and lithostratigraphic units have commonly been established on the basis of very detailed work in a very small area, such as a mine or a few closely spaced outcrops. The geologist should always stand back from such detail and attempt to visualize it in a regional context. A local stratigraphic terminology may be useful for in-house company communication, but it may not be worth burdening the geological literature with such detail.

Informal units are commonly established in the reconnaissance stage of basin exploration; for example, the first successful well in a frontier basin may give its name to an important hydrocarbon reservoir, such as the Leduc reef or Parsons sand. Whether such names should be formalized and published depends on further stratigraphic work,

and the extent and correlation of the unit may not be apparent until a much later stage of exploration. Industry geologists rarely have the time for the detailed lithostratigraphic and biostratigraphic follow-up work required, and it may be best left to university or geological survey workers to follow in the wake of such drilling and perform the necessary scientific research.

Seismic records are commonly used to delineate provisional stratigraphic units. An exploration group may establish an informal, in-house convention of coloring prominent reflectors in a consistent way. They then refer (internally) to the "red marker" or the "blue marker". Modern seismic stratigraphy permits a more detailed and rigorous subdivision than this, as explained in Chapters 5, 6, and 8. Seismic sequences and subsequences may be recognized and correlated using very little well control. It is dangerous, however, to assign too much weight to such interpretations until the detailed, scientific follow-up work has been carried out. Too often, seismic records are miscorrelated or misinterpreted because of the presence of subtle unconformities, problems of matching reflectors across faults, or poor reflector quality (Cartwright et al. 1993). Formal stratigraphic units should never be established without extensive, hard lithostratigraphic or biostratigraphic information from a network of wells or outcrops. The suite of coded stratigraphic sequences that formed an essential part of the Exxon global-cycle charts (Vail et al. 1977; Haq et al. 1987, 1988) is flawed because it has always lacked this essential documentation.

A few words are in order at this point regarding the correct use of the subdivisons early, middle, and late versus lower, middle, and upper. The first set of terms refers to time, the second set to rocks. Thus, it is not correct to refer to lower Jurassic time, just as it is not correct to refer to lower Wednesday. Likewise, it is not appropriate to call the lower part of the Confusion Formation the Early Confusion. A related problem is whether these subdivision terms should be capitalized. Wherever formal subdivison has been carried out, the answer is yes, they should. For example, the Early and Late Cretaceous have long been defined and recognized in the geological literature and should be so capitalized. However, for historical reasons, there has never been a middle Cretaceous. Events during this strictly nonexistent period could be referred to as having occurred during the middle part of the Cretaceous, without a capitalized "Middle." Owen (1987) provided a table listing all formal (and therefore capitalized) subdivisions of the major geochronologic and chronostratigraphic units, but because of continuing refinements in the time scale, this list will undoubtedly need to be changed in the future.

3.3.3 The Six Steps Involved in Dating and Correlation

Six main "steps" are involved in the dating and correlation of stratigraphic events (Miall 1994, 1997, Chap. 13). Figure 3.1 summarizes these steps and provides generalized estimates of the magnitude of the uncertainty associated with each aspect of the correlation and dating of the stratigraphic record. Some of these errors may be cumulative, as discussed in the subsequent sections. The assignment of ages and of correlations with global frameworks is an iterative process that, in some areas, has been underway for many years. There is much feedback and cross-checking from one step to another. What follows should be viewed, therefore, as an attempt to break down the practical business of dating and correlation into more readily understandable pieces, all of which may be employed at one time or another in the unravelling of regional and global stratigraphies. The main steps are as follows:

1. Identification of the units or stratigraphic events to be correlated and development of regional correlation frameworks. These are normally based on lithostratigraphy or sequence stratigraphy, with correlations constrained by other data such as biostratigraphic zonation. Determining the position of events such as sequence boundaries may or may not be a straightforward procedure and may require the application of facies mapping or sequence-mapping techniques (Chaps. 4, 6).
2. Determining the chronostratigraphic significance of unconformities. Unconformities, such as sequence boundaries, represent finite time spans that vary in duration from place to place. In any given location, this time span could encompass the time span represented by several different sedimentary breaks at other locations. Resolving such problems may require that some of the other steps be completed, particularly step 3.
3. Determination of the biostratigraphic framework. One or more fossil groups is used to assign the selected event to a biozone framework, and zones are defined and correlated from section to section. Error and uncertainty may be introduced because of the incompleteness of the fossil record. Graphic correlation or other quantitative techniques may be employed (Sect. 3.5).
4. Assessment of relative biostratigraphic precision. The length of time represented by biozones depends on such factors as faunal diversity and rates of evolution. It varies considerably through geological time and between different fossil groups.
5. Correlation of biozones with the global stage framework. The existing stage framework was,

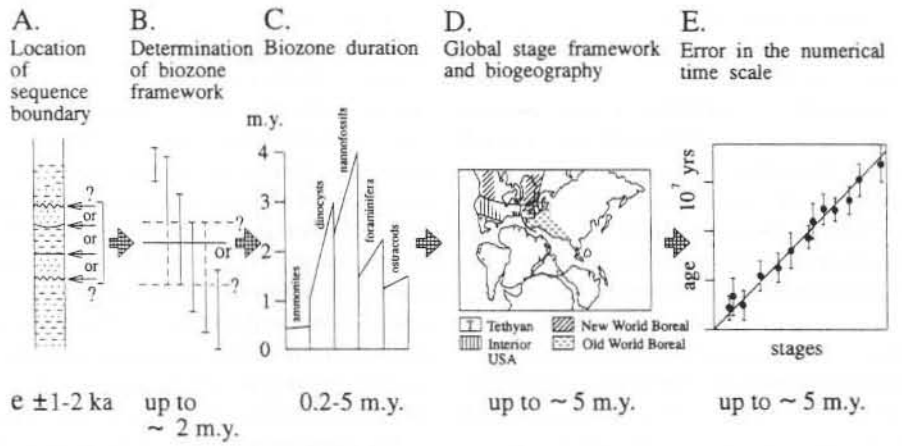


Fig. 3.1 A–E. Steps in the correlation and dating of stratigraphic events. *e*, typical range of error associated with each step. **A** Selection and local correlation of the stratigraphic event in question (e.g., a sequence boundary). **B** Assignment of the boundary event to the biozone framework. An incomplete record of preserved taxa (almost always the case) may lead to ambiguity in the placement of biozone boundaries. **C** The precision of biozone correlation depends on biozone duration. Shown here is a simplification of Cox's (1990) summary of the duration of zones in Jurassic sediments of the North Sea Basin. **D** The building of a global stage framework is fundamental to the development of a global time scale. However, global correlation is hampered by faunal provincialism. Shown here is a simplification of the faunal provinces of Cretaceous ammonites shown on a mid-Cretaceous plate-tectonic reconstruction. This diagram is based on Kennedy and Cobban (1977) and Kauffman (1984). **E** The assignment of numerical ages to stage boundaries and other stratigraphic events contains inherent experimental error and the error involved in the original correlation of the datable horizon(s) to the stratigraphic event in question. Diagrams of this type are a standard feature of any discussion of the global time scale (Haq et al. 1987, 1988; Harland et al. 1990). The establishment of a global, biostratigraphically-based sequence framework involves the accumulation of uncertainty over steps **A** through **D**. Potential error may be reduced by the application of radiometric, magnetostratigraphic, or chemostratigraphic techniques that, nonetheless, contain their own inherent uncertainties (step **E**)

with notable exceptions, built from the study of macrofossils in European type sections. Correlation with this framework raises questions about environmental limitations on biozone extent, our ability to inter-relate zonal schemes built from different fossil groups, and problems of global faunal and floral provinciality.

6. Assignment of absolute ages. The use of radiometric and magnetostratigraphic dating methods plus the increasing use of chemostratigraphy (oxygen and strontium-isotope concentrations) permits the assignment of absolute ages (in years) to the biostratigraphic framework. Such techniques also constitute methods of correlation in their own right, especially where fossils are sparse (Sect. 3.7).

3.4 Lithostratigraphy

3.4.1 Recognition of a Lithostratigraphic Subdivision

In Chapter 2, what to observe in the rocks was shown and some ideas were given on how to record and present the data. This information can now be used to establish a lithostratigraphic framework. The geologist should have plotted graphic logs of the surface and subsurface sections and, for subsurface sections, should have a suite of geophysical logs, if available. In many cases, an obvious stratigraphic breakdown will have become apparent as the data-collection work has proceeded, and the individual geologist or exploration group may have started to formulate informal units to aid mental comprehension or for internal communication. Where regional seismic records are available, the broad regional framework may already be apparent. This process of subdivision should now be carried out in a slightly more rigorous way by comparing the logs and making a preliminary attempt at correlation.

In each section, it will be possible to recognize one or more (perhaps many) distinctive lithostratigraphic units based on physical properties such as mineralogical composition and grain size. Bedding characteristics, sedimentary structures, cyclic sequences, and fossil content may be used as secondary criteria. Fossils, although primarily of importance as biostratigraphic indicators, may also have importance as facies indicators in lithology studies (Chap. 4). They may be of rock-forming abundance, as in bioherms and coquinas, or may be present as minor but distinctive components of a lithostratigraphic unit. The latter applies to both body fossils and trace fossils. Lithological variations based on diagenetic changes,

such as color changes, dolomitization, or emplacement of epigenetic ores, should not be used for lithostratigraphic subdivision.

The recognition of a lithostratigraphic unit is a two-part process requiring subdivision of vertical sections and correlation between sections. The exercise of correlation is one of the most important aspects of basin analysis. As discussed in Section 3.4.3, lithostratigraphic correlation, ideally, should not be divorced from biostratigraphic correlation and considerations of basin structure and the local or regional depositional system. The geologist is, therefore, required to perform several mental tasks at once. Omission of these overlapping procedures can lead to serious lithostratigraphic miscorrelations, as an examination of older literature for any well-studied sedimentary basin will demonstrate.

In recent years, a powerful concept termed event stratigraphy has evolved from the recognition that, in many sedimentary basins, careful study of depositional processes and meticulous correlation have revealed that part or all of a sedimentary succession may be dominated by distinctive stratigraphic events such as widespread marker beds or discontinuities caused by regional/global sea-level change (Ager 1981). Such events may have considerable chronostratigraphic significance and may, therefore, provide the basis for the stratigraphic subdivision of the basin. However, there is a danger of circular reasoning in stratigraphic correlation if the development of an event-stratigraphic framework is used as a justification for correcting age assignments arrived at by biostratigraphic techniques, as was done in some of the early sequence-stratigraphic work (Miall 1986; Sect. 8.3.2).

3.4.2
Types of Lithostratigraphic Units and Their Definition

Based on the formation, which is the primary lithostratigraphic unit, a hierarchy of units has been developed:

A. Group
B. Formation
C. Member
D. Tongue or lentil
E. Bed

3.4.2.1
The Formation

An important convention has been established that all sedimentary rocks should be subdivided (when sufficient data have been collected) into formations. No other types of lithostratigraphic subdivision need be used, although convenience of description may require them.

What is a formation? This is a question with no final answer. The degree of lithologic variability required to distinguish a separate formation tends to reflect the level of information available to the stratigrapher. Formations may be only a few meters or several thousands of meters in thickness; they may be traceable for only a few kilometers or for thousands of kilometers. Formations in frontier basins usually are completely different in physical magnitude from those in populated, well-explored basins, such as much of western Europe and the United States. As work in frontier basins proceeds, some of the larger formations may be subdivided into smaller units and the ranking of the names changed (see below).

The most important criteria for establishing a formation are its usefulness in subdividing stratigraphic cross sections and its "mappability". For reconnaissance mapping, a thin unit that cannot accurately be depicted at a scale of, for example, 1:250,000, is of little use. For more detailed work, mappability at a scale of 1:50,000 or even 1:10,000 may be a more useful criterion. Problems of consistency arise when detailed work is conducted around a mine site within what is otherwise a poorly explored frontier basin (Sect. 1.4.2).

Formations should not contain major unconformities, although minor disconformities may be acceptable. The contacts of the formation should be established at obvious lithologic changes. These may be sharp or gradational. An unconformity is a logical choice for a formation contact. Where lithologies change gradually, either vertically or laterally, it may be difficult to choose a logical place to draw the contact. For example, a mudstone may pass up into a sandstone through a transitional succession, with sandstone beds becoming thicker and more abundant upward. The mudstone–sandstone formation contact could be drawn at the oldest thick, coarse sandstone (with thickness and coarseness carefully spelled out), at the level where sandstone and mudstone each constitute 50% of the section, or at the youngest extensive mudstone bed. The choice is arbitrary, and it is immaterial which method is selected as long as the same method is used as consistently as possible throughout the extent of the formation.

Other problems of definition arise where there are lateral lithologic changes requiring definition of a new formation. A simple diachronous contact is not a problem, but, where the two units intertongue with one another, it may be virtually impossible to draw a simple formation contact. One solution is to give each tongue the same name as the parent formation. A section passing through the transition region may then show the two formations succeeding each other several times. The only problem this causes is if formation contact and thickness data are stored in a data

bank and used in automated contouring programs. Without additional input from the operator, a computer program might not be able to handle this type of data. Other alternatives are to define the whole transitional rock volume in terms of one of the parent units, to separate the transitional lithologies as a separate lithostratigraphic entity, or to give separate tongues their own bed, tongue, or member names. Published stratigraphic codes do not specify any rigid rules for the solution of such problems (although the NASC provides some useful suggestions on pages 856 and 857; North American Commission on Stratigraphic Nomenclature 1983). The main criteria should be practicality, convenience, and consistency. An illustration of interpreted stratigraphic relationships in the form of a cross section or cartoon will usually serve to clarify any ambiguity. Some examples are discussed in Section 3.4.3.

3.4.2.2 *The Group*

All other stratigraphic units are based on the formation. A group consists of two or more formations related lithologically. In the past, named groups have been established for thick and varied successions without first defining the constituent formations. This is not recommended practice. Groups should not contain major unconformities.

The component formations of a group may not everywhere be the same. Lateral facies changes requiring definition of different formations can occur within a single group. In contrast, a component formation may extend laterally from one group to another. Groups are normally defined for regions of complex stratigraphy. Toward the basin margin or basin center, the component formations may lose their individuality, in which case the group may there be "demoted" to a formation, while still retaining the same name.

The terms supergroup and subgroup are occasionally used to provide an additional hierarchy of subdivision. Usually, there are historical reasons for this; some of the higher-ranking names may have started out as member, formation, or group names, with reclassification and promotion being required as additional work demonstrated the need for further subdivision.

3.4.2.3 *The Member*

This is the next ranking unit below the formation. Not all formations need to be divided into members, and formal names need to be used for only a few, one, or none of the constituent members, depending on convenience or the level of information available.

There are no standards for the thickness or extent of members, and it is commonly difficult to decide whether to define a given lithostratigraphic unit as a member or a formation. However, recommended practice is that all parts of a succession be subdivided into formations, so this is the best level at which to start. A member cannot be defined without its parent formation. For mapping and other purposes, it commonly is convenient to establish informal units, such as the lower sandstone member, that do not require formal names.

3.4.2.4 *Tongue or Lentil*

These are similar to members. Because of their geometric connotations, the terms are useful for parts of formations where they interfinger with each other. Formal names may be established for one, several, or all such units, depending on convenience and practicality.

3.4.2.5 *Bed*

This is the smallest formal, named unit in the hierarchy of lithostratigraphic units. Normally, only a few parts of a stratigraphic succession will be subdivided into named beds. Coal seams in mine areas, prominent volcanic-tuff horizons, and other marker beds are typical examples. Certain strata-bound ore-bearing beds, such as placer units may also be named.

3.4.2.6 *Choice of Names*

When establishing a named unit, it is standard practice to give it a geographical name chosen to suggest the location or areal extent of the unit. This may be a river, lake, bay, headland, hill, mountain, town, village, etc. Permanent names are preferable. Subsurface work in frontier basins, particularly in offshore areas, may rapidly use up all the available names, in which case the name of the well chosen as the type section may be used. Failing this, names may have to be invented.

In most cases, the geographical name will be followed by the rank designation, for example, Wilcox Group, Pocono Formation. For beds, this is commonly not done, particularly in the case of coal seams, for which a complex mine terminology may have evolved. Many older stratigraphic units use a lithologic term instead of a rank term, such as Gault Clay, Dakota Sandstone, or Austin Chalk, but this is not recommended, because the rank of the unit is not clear from the name alone.

Workers should beware of renaming units without justification or using a geographic name that has already been employed in a different context. Geological-survey organizations commonly retain a file of current and obsolete stratigraphic names that the worker may wish to consult.

3.4.2.7
Changes in Rank

Lithostratigraphic units may be changed in rank as the level of knowledge improves. For example, the Cornwallis Group in the Canadian Arctic Islands started as the Cornwallis Formation and was raised to group rank when it was realized that it contained three mappable units of formation rank. Conversely, the Eureka Sound Group was named for a thick and varied clastic succession, but it was never subdivided into named constituent formations by the original author and was reduced to formation status. The unit has now been subdivided and has been formally redefined as a group once again. When a unit is raised in rank, the original name should not be used for any of the subdivisions but is best retained for the higher ranking unit or abandoned altogether.

3.4.3
Mapping and Correlation Procedures

The term "correlation" means establishment of the mutual relationships existing between stratigraphic units. When used in the context of lithostratigraphic units, the term implies that there are physically continuous rock bodies of more or less constant character that can be traced from one part of a basin to another. Such units may be diachronous. When dissimilar lithostratigraphic units are said to be correlated, a time connotation is implicit. That is, two units may be formed in different environments and contain different lithologic characteristics but may be said to be correlated because they are known to have been formed at the same time on the basis of biostratigraphic or other chronostratigraphic evidence.

Correlations based strictly on lithostratigraphic evidence should be viewed with caution, because similar units may be deposited at different times in different areas and have no physical connection with each other. Many mistakes have arisen this way when geologists have attempted to correlate widely spaced sections.

There are only three conditions under which lithostratigraphic correlation should be attempted without supporting biostratigraphic or chronostratigraphic information:

1. Where the beds can be physically traced out on the ground
2. Where there are distinctive, readily recognizable marker beds or where facies studies indicate considerable lateral bed continuity (included here are cases where the succession can be subdivided on the basis of an event stratigraphy)
3. In Precambrian rocks, where only very generalized chronostratigraphic data (radiometric dates) may be available

3.4.3.1
Surface Mapping

In areas lacking vegetation cover, such as in arid or arctic climates, lithostratigraphic units can commonly be traced in outcrops on the ground, from the air, or from aerial photographs. In areas of simple to moderate structural complexity, photogeological reconnaissance using stereoscopic air-photo coverage can produce preliminary maps of large areas very quickly. Standard aerial photos taken from an elevation of a few thousand meters and printed at a scale of about 1 cm:1 km are the best for this purpose. Satellite images are at too small a scale (although remote-sensing technology is developing at a rapid rate, and this statement may soon be out of date). Most of the Canadian Arctic has been mapped by the Geological Survey of Canada using conventional aerial photographs. The photoreconnaissance maps are used to determine areas of structural complexity or uncertainty and locations where the stratigraphy appears to be particularly well exposed. These places are then visited on the ground in order to resolve structural problems and measure detailed stratigraphic sections. Marked facies changes, such as a carbonate–mudstone transition at a shelf edge, may be readily apparent and easily mapped. This is lithostratigraphic mapping in its purest form. Coherent units may be distinguished and mapped, but, at this stage, there is little or no information on age or possible diachronism of the units. An example of Arctic lithostratigraphic mapping based on aerial photos is shown in Fig. 3.2.

Where there is a significant cover of soil, surficial deposits, or vegetation, it may still be possible to carry out a similar type of mapping by noting changes in ground texture, soil color, or vegetation type or subtle changes in relief reflecting varying resistance to erosion of the different units. For example, the stratigraphic succession may be studied in detail in good outcrops along river cuts and mapped out between the rivers by tracing the outcrops into covered regions where the unit shows up as a resistant ridge or a belt of distinctively colored subsoil fragments. Remote sensing techniques, including conventional aerial photographs, plus the use of satellite data, may be of considerable importance in

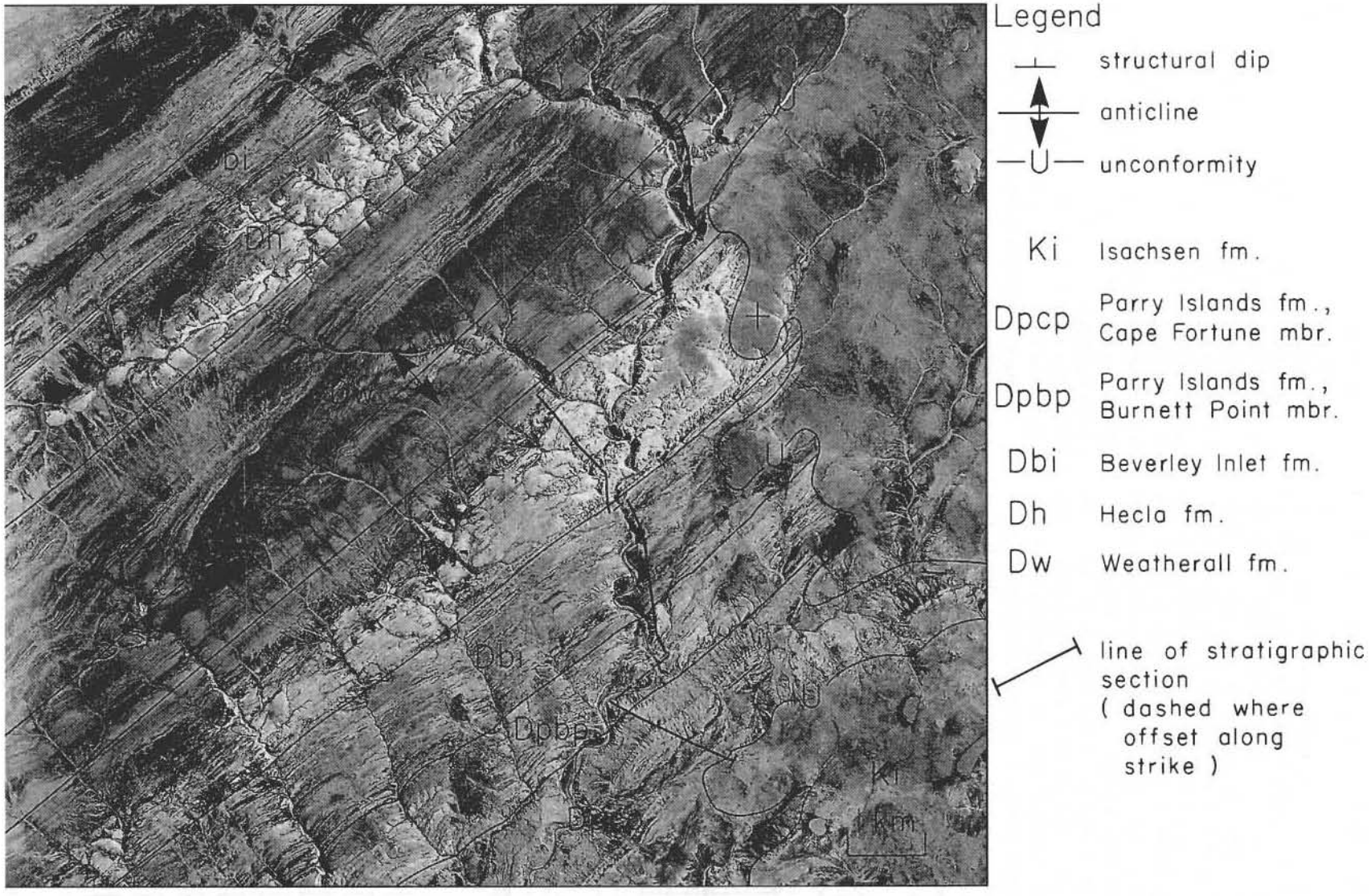


Fig. 3.2. Use of vertical aerial photographs for lithostratigraphic mapping. An example from southeastern Melville Island, Canadian Arctic. (Courtesy of A.F. Embry)

lithostratigraphic mapping, as discussed briefly in Section 5.7.

In areas of dense vegetation cover, thick surficial deposits, or complex structure, detailed surface mapping may not be possible. Lithostratigraphic correlation is then carried out in very much the same way as subsurface work, attempting to match isolated sections by whatever means possible (Sect. 3.4.3.3). Formal lithostratigraphic units, such as formations and groups, may be defined, but it may be difficult to plot their distribution accurately on a surface-geological map. This is the case with many temperate or tropical regions. In areas such as Western Europe, much use is made of soil augers and temporary exposures, such as trenches dug for pipelines, to supplement outcrop information for mapping purposes. For further details on mapping procedures, the reader is referred to standard field manuals such as that by Compton (1962).

In areas of complex facies changes, detailed ground mapping may be a very productive exercise for the basin analyst, because work of this type can provide significant clues for environmental and paleogeographic reconstructions that may be applicable to large parts of the basin. At this stage, lithostratigraphic documentation is only preliminary to detailed facies and trend studies such as are described in Chapters 4, 5, and 6. An example is illustrated in Fig. 3.3. This section is one of many compiled for the Carboniferous, coal-bearing, deltaic sediments of Kentucky and West Virginia by Horne et al. (1978) and Horne and Ferm (1978). They are based on extensive artificial cuts along railroad lines and highways. The series of vertical rectangles outlines detailed sections plotted on the ground across an exposure more than 5 km wide. These are connected by lithostratigraphic correlation lines, most of them traced out on foot in the field. Unornamented areas represent areas not exposed or not

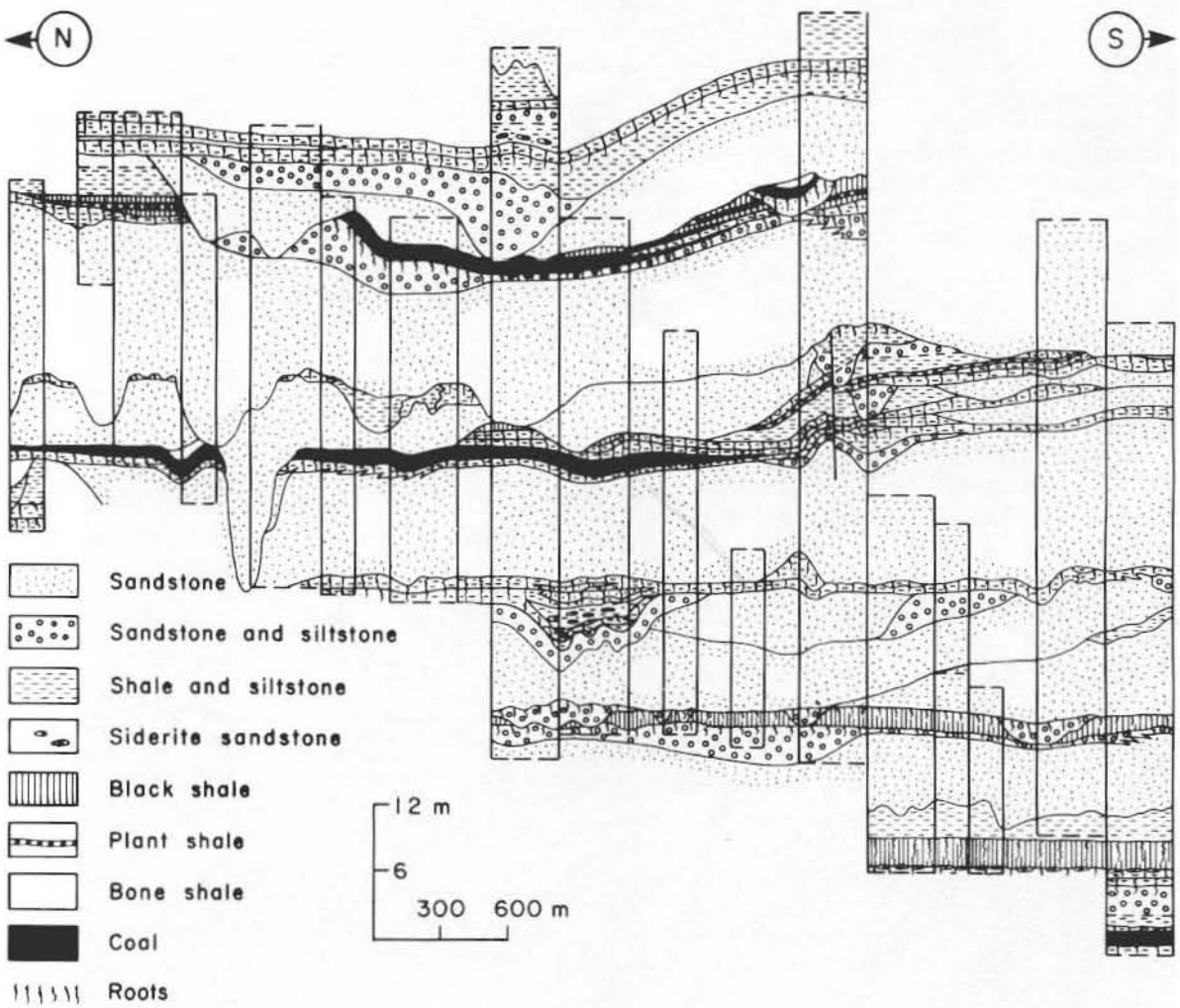


Fig. 3.3. Example of a detailed lithostratigraphic study carried out on well-exposed roadside outcrops. Carboniferous period, Kentucky. (Horne et al. 1978)

studied. The section reveals a series of sandstone beds 5–15 m thick, many with irregular, sometimes deeply channeled bases. They are separated from each other by siltstone, mudstone, and coal beds. Horne et al. (1978) and Horne and Ferm (1978) discussed the lithostratigraphic succession, lateral variability, and facies data of this and other large exposures and interpreted the rocks as the deposits of a coastal fluviodeltaic plain.

Figure 3.4 A illustrates a carbonate example. Jones and Dixon (1975) measured detailed, closely spaced cliff sections at Port Leopold in Northern Somerset Island in the Canadian Arctic. The rocks are Silurian shelf carbonates of the Leopold Formation and show considerable lateral facies changes over distances of only a few tens of meters. Key beds can be traced on foot from section to section, permitting a detailed correlation to be documented. Lithostratigraphic variations in this and adjacent sections were used by Dixon and Jones (1978) to erect local members within the Leopold Formation.

3.4.3.2 *Event Stratigraphy*

Many stratigraphic successions contain marker beds or discontinuity surfaces that are known to have chronostratigraphic significance. They may be of purely local distribution or may have regional or even global significance. The documentation of such features has come to be known as event stratigraphy. As discussed at length in Section 6.8 and Chapter 8, event stratigraphy on a continental and global scale is providing a framework for a new and dynamic way of interpreting global basin evolution.

An example of the value of recognizing stratigraphic events at a relatively local scale is illustrated in Fig. 3.4 B. Extension of the lithostratigraphic work of Dixon and Jones (1978) along well-exposed cliff sections showed that, over a distance of about 130 km, the member boundaries are markedly diachronous. Marker beds consisting of distinctively weathering units of varying lithology could be physically traced along the cliffs, permitting tight correlation of the sections (the sections in Fig. 3.4 A correspond to Sect. 11 in Fig. 3.4 B). In addition, it was found that pulses of quartz-sand detritus could be correlated parallel to the marker beds. On regional evidence, the line of section is interpreted to be parallel to the line of sedimentary strike relative to the detrital source, whereas the various shallow shelf environments shifted gradually to the south at the same time (left in Fig. 3.4 B), generating diachronous lithostratigraphic units.

Ash beds have long been recognized and used as stratigraphic-event marker beds. Examples are discussed by Johnson et al. (1979; Sect. 3.7.5.2), Hsü et al. (1980), and Allen and Williams (1978). Problems with this method arise if more than one ash is present, and special laboratory techniques may be required to determine distinctive geochemical and mineralogical features or other characteristics of each ash bed to permit correct correlation (Westgate 1980).

Several other lithofacies may be suitable for use as chronostratigraphic markers, although all show a certain degree of diachroneity, and their precision as markers of geologic time varies. In carbonate basins, abrupt shifts in lithofacies may occur in response to sea-level changes or other environmental changes, and these may make good markers (e.g., the **drowning unconformities** of Schlager 1989). Such changes are the basis for the definition of a new kind of cycle, the **punctuated aggradational** cycle (PAC) of Goodwin and Anderson (1985). Sediment gravity flows are almost instantaneous deposits and, therefore, have chronostratigraphic potential. The problem is to correlate them correctly between sections (Sect. 3.4.3.3). A good example of catastrophic sediment-gravity-flow deposition providing the basis for a local event stratigraphy was published by Bürgisser (1984). Kleverlaan (1987) described a widespread "seismite" formed as a result of seismic shock. Shiki (1996) described the importance of tsunamis as geological events. Major et al. (1996) described the products of a major hurricane and their preservation potential. Hardgrounds are surfaces of nondeposition or very slow sedimentation and may be traceable over many kilometers (Bathurst 1976; Stoakes 1980). Surfaces of submarine nondeposition or condensed sedimentation have been interpreted as stratigraphic events indicating rapid regional or global sea-level change (Vail et al. 1984) but, as noted by Miall (1986), there are doubts about the appropriateness of using such beds for intercontinental correlation. Transgressive sandstones, such as those occurring above an unconformity or in the abandonment phase of a delta (Elliott 1978), may be widespread, relatively thin, glauconitic, mature quartzose units readily recognizable in the sedimentary record. However, they may be markedly diachronous. Widespread shale beds commonly develop during high stands of sea level. In sequence-stratigraphic terminology these are called **maximum flooding surfaces**, and they commonly provide excellent marker beds for surface and subsurface correlation and mapping (e.g., Galloway 1989; Underhill and Partington 1993). Sageman (1996) described Cretaceous skeletal limestone beds within the Western Interior Seaway that he interpreted as the product of storm activity during low stands of sea level. He was able to trace them for hundreds of kilometers.

Miall (1979) used a distinctive bituminous mudstone unit (basal Kanguk Formation) to map the up-

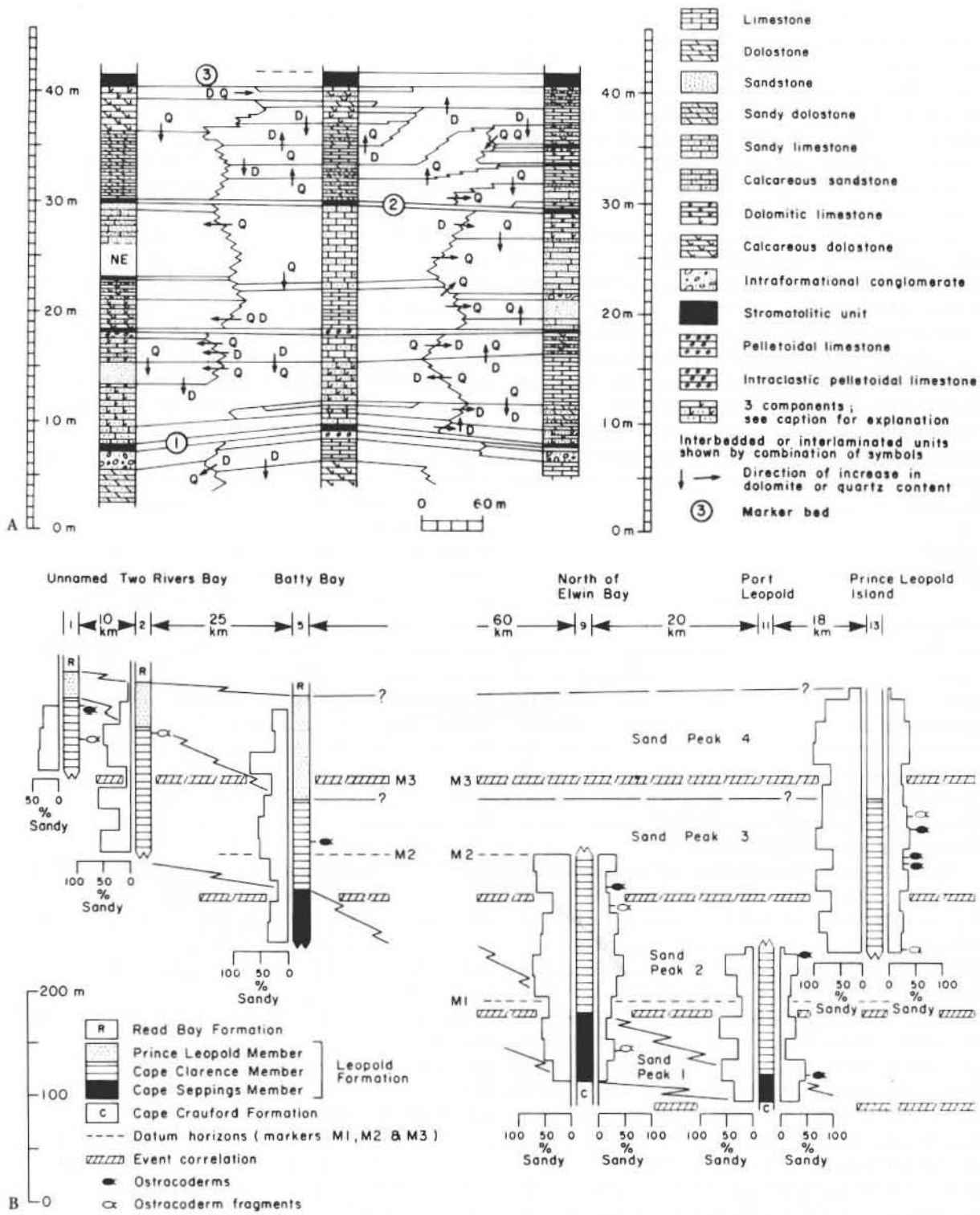


Fig. 3.4. **A** Use of stromatolitic marker beds (numbered *1*, *2*, and *3*) to correlate closely spaced sections showing considerable lateral facies changes. The three-component lithology consists of interbedded quartz, micritic limestone and dolomite. Silurian period, Somerset Island, Canadian Arctic (Jones and Dixon 1975). **B** Chronostratigraphic correlation of regional stratigraphic sections using weathering markers and correlation of sand peaks. The sections shown in **A** are combined here as section 11. Chronostratigraphic correlation demonstrates the diachroneity of the lithostratigraphic units. Silurian period, Canadian Arctic. (Dixon et al. 1981)

per–lower Cretaceous contact in Banks Island in the western Canadian Arctic. The bituminous content, which reflects a euxinic event in the basin, has oxidized to a bright red, and this enables the contact to be mapped from a low-flying helicopter. The unit also has a distinctive subsurface log signature. This same marker bed is present throughout much of Arctic North America (Miall 1979). Hallam and Bradshaw (1979) commented on the sedimentary-tectonic interpretation of black mudstones and oolitic ironstones and their stratigraphic significance. Many other types of stratigraphic events were discussed by Einsele and Seilacher (1982), such as storm deposits, cyclic limestone–shale rhythms, and turbidites.

Schultz (1982) pointed out that, where a rock body is delimited by marker beds or by well-defined biostratigraphic horizons, it has a local chronostratigraphic significance. He proposed the term **chronosome** for such units and suggested that definition and naming of individual chronosomes would have considerable value for basin-analysis work. A chronosome is distinct from a formation in that it can contain lateral lithologic (facies) changes and is delimited by time planes. Therefore, it defines a single episode or period of deposition. The term **chronozone**, as defined by Salvador (1994), is virtually a synonym for chronosome.

A particularly instructive example of event stratigraphy is the stratigraphic synthesis of the Cardium Formation, an important hydrocarbon-producing unit in the Alberta Basin, Canada, developed by Plint et al. (1986) as part of a detailed regional surface-subsurface study of the many producing fields in the area. Numerous local studies had, over the years, led to a confusing welter of local informal terminologies for sandstone horizons and marker units within the Cardium Formation and much controversy regarding the depositional environments of the various facies. Routine but meticulous lithostratigraphic correlation of subsurface records led A. G. Plint to the recognition that this unit, which is only about 100 m thick, contains at least seven basin-wide erosion surfaces, indicating the occurrence of this number of events of erosion and transgression (Fig. 3.5). Not only does this new framework provide a rational basis for basin-wide correlation, but it also throws a wholly new light on the depositional history of the formation. The unit consists largely of a series of basin-wide coarsening-upward cycles capped by sandstone units containing hummocky cross-stratification, overlain in some areas by lenticular conglomerate beds. The cycles, up to the level of the sandstones, are readily interpreted as the product of a shoaling shelf environment, with the sediment surface gradually building up to storm-wave base. The problem is to fit the conglomerate into this interpretation. These coarse deposits are tens to several hundreds of kilometers from the basin margin, and the problem of transporting coarse detritus out this far in a shelf setting has been discussed in the Cardium literature for many years. The new interpretation by Plint provides a simple resolution to the problem. The conglomerates are not a conformable cap to the coarsening-upward cycles, as had always been thought; instead,

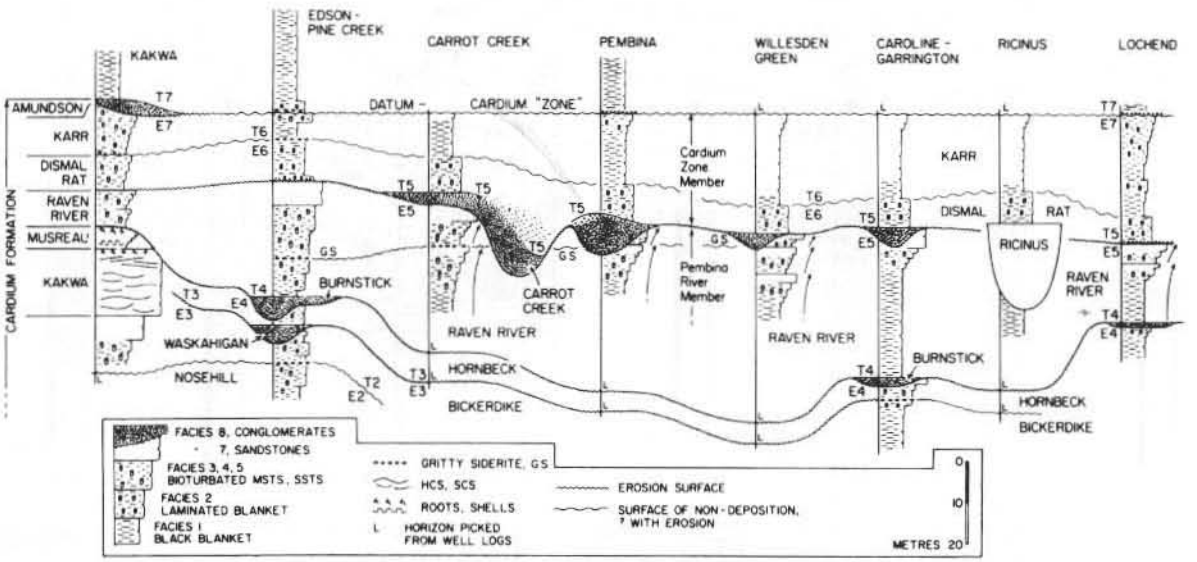


Fig. 3.5. Correlation of typical sections of the Cardium Formation, Alberta, based on a framework of event stratigraphy. The sections are tied to a series of seven basin-wide erosion/transgression surfaces, of which surfaces 2–5 are present in this section. The conglomerate beds were formerly interpreted as the conformable tops to coarsening-upward successions, but are now interpreted as being in unconformable contact with underlying sandstones. (Plint et al. 1986)

each conglomerate lens rests on one of the erosion surfaces. Therefore, in all probability, they are beach deposits, transported basinward during regression of the shoreline and concentrated during the initial transgression at the beginning of a new cycle of relative sea-level rise.

Considerable discussion has, in recent years, focused on several apparently spectacular events in Earth history, especially the various sudden faunal extinctions that are supposed to have occurred during the Phanerozoic. These are not only spectacular stratigraphic events, notably those that mark the end of the Permian and the end of the Cretaceous, but their interpretation has important consequences for the way we view Earth history in general. A discussion of such events is well outside the scope of this book except for one important subject: that of stratigraphic completeness. Stratigraphic correlation and reconstruction and the development of facies and paleogeographic syntheses depend on the appropriate comparisons between contemporaneous sequences for their veracity. As Ager (1981) has pointed out forcefully, the stratigraphic record is commonly "more gaps than record"; in many instances, it may be biased by the preservation of the products of rare events, with long, intervening periods of time unrepresented. We return to this point in Section 3.7.8.

3.4.3.3 *Section Matching*

Correlations cannot be walked out or mapped out in the subsurface, and lithostratigraphic work depends on the presence of distinctive units in well records or cores that can be recognized and traced from one section to the next. The best correlations are those based on widespread marker beds or other events, as discussed previously, because of their chronostratigraphic value. With or without these, the geologist must develop visual skills in pattern recognition, placing graphic lithostratigraphic logs or petrophysical logs beside each other, comparing all the various attributes present, and judging which correlations permit the closest fit. This process is performed routinely by subsurface-exploration geologists. Petrophysical (wire-line) logs are particularly useful for such purposes. However, without chronostratigraphic markers, the geologist must always watch for the possibility of diachronism and lateral facies change. Similar correlation procedures may be necessary for surface work, although, here, the task tends to be somewhat easier because of the greater range of lithostratigraphic parameters that can be examined in outcrops.

Even the most detailed biostratigraphic zonation is usually of little value for the level of correlation required at the scale of diamond drill hole (DDH) exploration of a mining property or exploration of a petroleum prospect. The art of section matching is, therefore, an important one for the explorationist.

Stratigraphic sequences vary considerably in the ease with which lithostratigraphic correlation may be performed at the member or smaller-scale level. In many depositional environments, facies changes are numerous and, without marker beds, section matching may be impossible. This is the case with fluvial and deltaic deposits and the sediments of most other

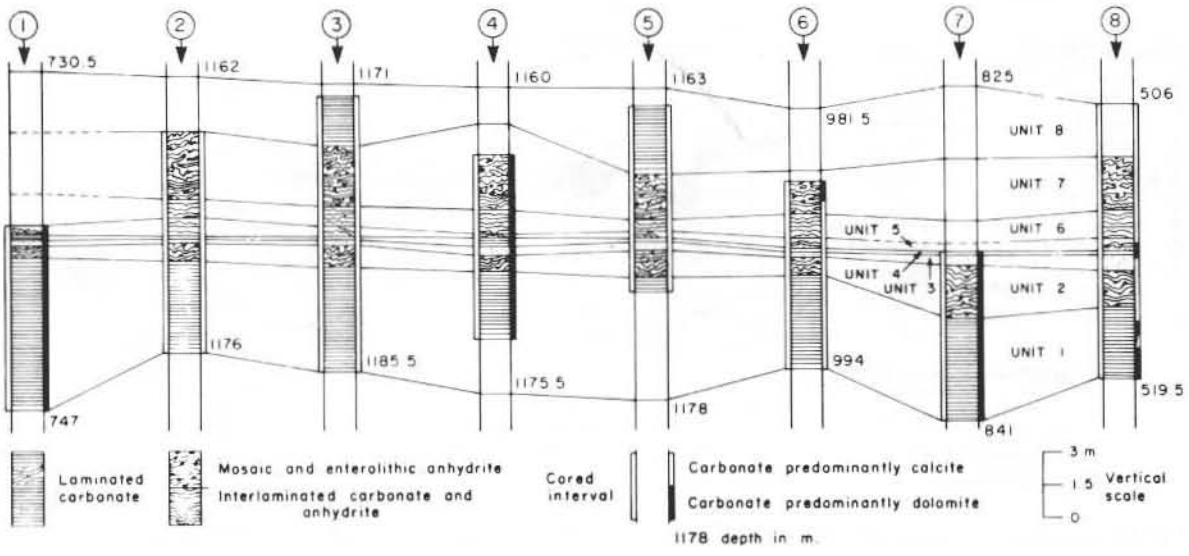


Fig. 3.6. Detailed lithostratigraphic correlation of cored subsurface units consisting of thinly interbedded carbonates and evaporites. (Wardlaw and Reinson 1971)

coastal clastic and carbonate environments. Submarine fan deposits also show considerable facies variation. Environments that may give rise to more lateral consistency include lakes (away from any marginal clastic facies), low-energy carbonate and evaporite shelves, some clastic and carbonate (sabkha) tidal flats, and deep oceanic abyssal plain or trench deposits. In some of these cases, practically every sedimentation unit is a marker bed.

Figure 3.6 illustrates carbonate–evaporite correlations made in the middle Devonian sediments of Saskatchewan (Wardlaw and Reinson 1971). The sections were plotted from cores derived from wells extending more than 160 km across the basin. The authors were lucky to locate so many cores through the same thin interval. The consistency of facies and thickness is remarkable. Correlation of some Miocene basin plain turbidites is shown in Fig. 3.7. Ricci-Lucchi and Valmori (1980) measured a series of outcrop sections along the axis of a turbidite basin in the northern Apennines. Transport directions were toward the southeast (left to right). A considerable facies change takes place down the basin, so a sandy section such as *#1* could not be correlated with a mud-dominated section such as *#12*. However, by using the intermediate sections, it is possible to trace the lateral facies change of virtually every bed. The amount of fieldwork required to establish this correlation network was considerable. Deep-sea cores in some modern turbidite basins have demonstrated correlations over distances up to 500 km (Bennetts and Pilkey 1976).

Many attempts have been made to use statistical techniques for lithostratigraphic correlation. Me-

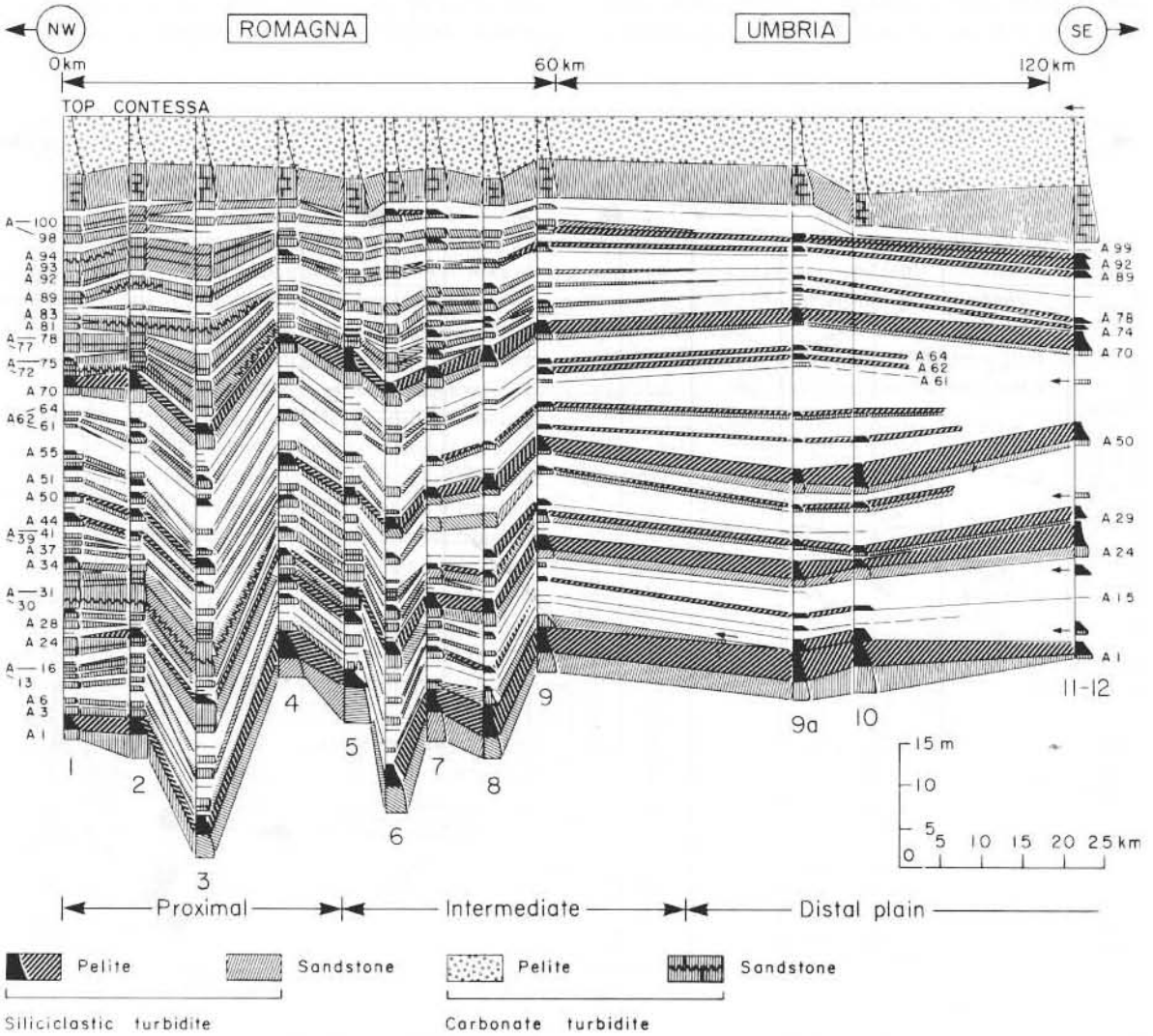


Fig. 3.7. Detailed lithostratigraphic correlation of outcrop sections through basin-plain turbidites. The line of section is approximately parallel to the paleoslope. Blank intervals are pelites. Cenozoic, Italy. (Ricci-Lucchi and Valmori 1980)

thods such as Fourier analysis have been used to characterize the cyclicity of certain attributes, such as petrographic variation or various petrophysical properties, in a succession (Preston and Henderson 1964). Various pattern-matching techniques have been used to determine the best fit between two or more sections (Davies and Ludlum 1973; Smith and Waterman 1980). However, this approach is not recommended by the author for the following reasons. In cases such as the turbidite and evaporite examples described previously, correlation is obvious enough that it can be made manually. Where the correlation is not obvious, statistical techniques are not likely to help, because numerical manipulation cannot easily accommodate thickness and facies changes. These are better dealt with by applying sedimentological principles.

Lithostratigraphic correlation at the formation and member level is routine in surface and subsurface work. The use of geophysical log signatures to correlate well sections is illustrated by Fig. 3.8, an example from the Milk River Formation (upper Cretaceous) of Alberta (Myhr and Meijer-Drees 1976). The geologist does not need to know much about the actual lithologies penetrated to carry out such correlations, nor is it necessary to interpret the actual values of the log deflection. Curve shape is sufficiently distinctive to permit precise, reliable correlation until a facies change alters the pattern. Figure 3.9 demonstrates a case where simple log matching cannot be performed reliably (Athabasca Oil Sands, Alberta; Carrigy 1971). This is a common problem in channelized fluvial, deltaic, and submarine fan deposits. Petroleum geologists frequently use formation-pressure data to test the interconnectedness of reservoir units. Pressure is plotted against well depth. Continuous (and therefore correctly correlated) units yield data points falling on a straight line. Nonconnected units show up as isolated points or as separate but parallel straight lines. An example is shown in

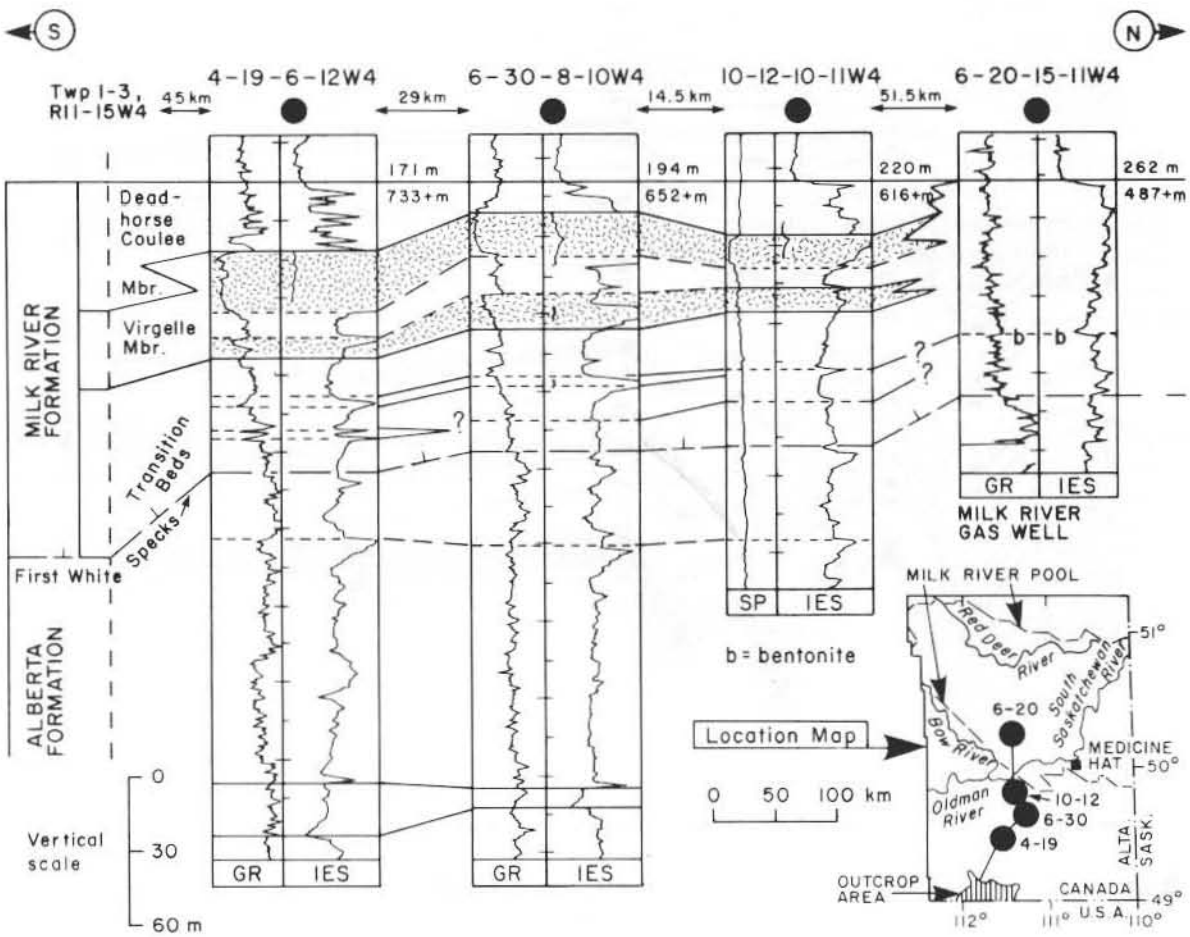


Fig. 3.8. Use of gamma-ray and induction-electric-survey logs to correlate subsurface sections. This diagram shows that, where lithostratigraphic units retain similar thickness and lithology, they can readily be correlated using geophysical logs. Cretaceous, Alberta. (Myhr and Meijer-Drees 1976)

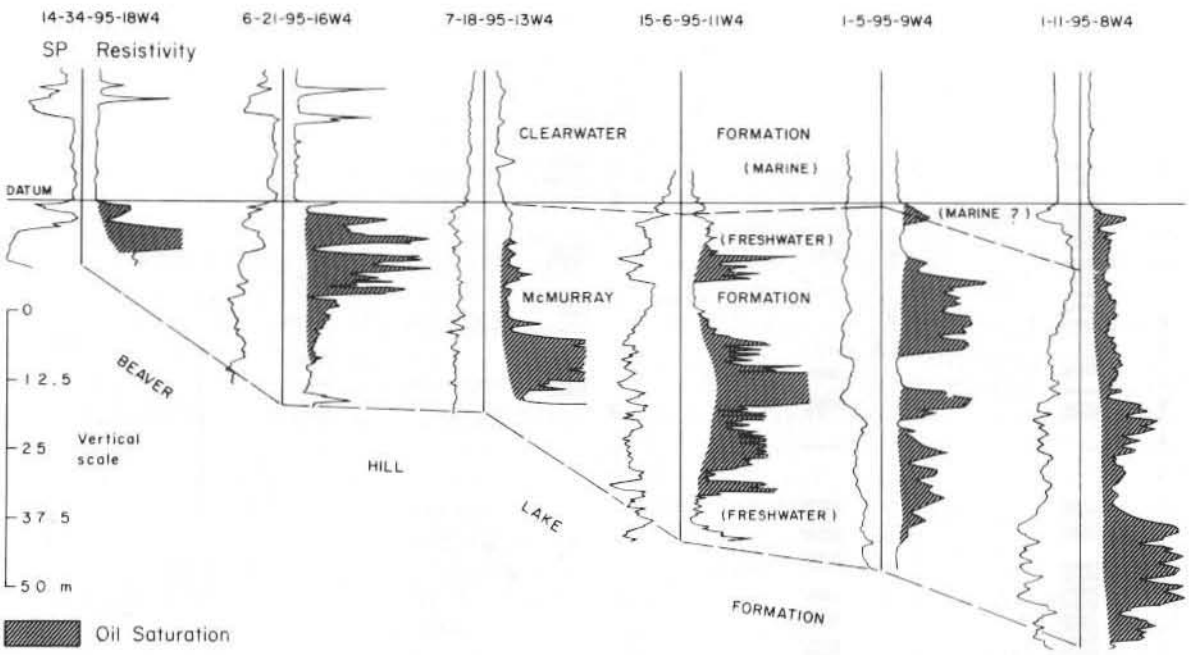


Fig. 3.9. In contrast to Fig. 3.8, this diagram illustrates the difficulty of correlating sections that show marked lateral facies changes. Fluvial deposits of Athabasca Oil Sands, Alberta. Length of section is 200 km. (Carrigy 1971)

Fig. 3.10. Without such data, it is easy to make spurious correlations that may only be revealed as incorrect after many more wells have been drilled.

A particularly ambitious attempt at lithostratigraphic correlation is illustrated in Fig. 3.11 (Young 1979). The rocks are cratonic and pericratonic successions of Proterozoic age. A few radiometric dates are indicated and the correlations are, therefore, loosely controlled by chronostratigraphy. However, virtually all the dashed correlation lines are based on more or less detailed lithostratigraphic section matching. The correspondence between sections *3* and *4* is particularly close (these sections are about 1200 km apart). Imagine a geologist attempting to match Phanerozoic sections, some of them up to 20 km thick, across half a continent using less than a dozen biostratigraphic control points! The attempt would be greeted with scorn. However, for Precambrian strata, this represents one of the best examples of regional lithostratigraphic correlation available. Note the recognition of three unconformity-bounded sequences, *A*, *B* and *C*, each on the order of 300 m.y. in duration, and compare these with the Phanerozoic sequences discussed in Chapter 8.

The application of the principles of sequence stratigraphy in the matching and correlation of sections is now routine for surface and subsurface work. Sequence models provide a predictive framework for the recognition of facies assemblages and stratigraphic architectures that can be interpreted in terms of systems tracts, and these constitute components of sequences that may be traceable throughout a sedimentary basin. The methods are explained in detail in Chapter 6.

In conclusion, it is worth repeating that the best lithostratigraphic subdivision and correlation is bas-

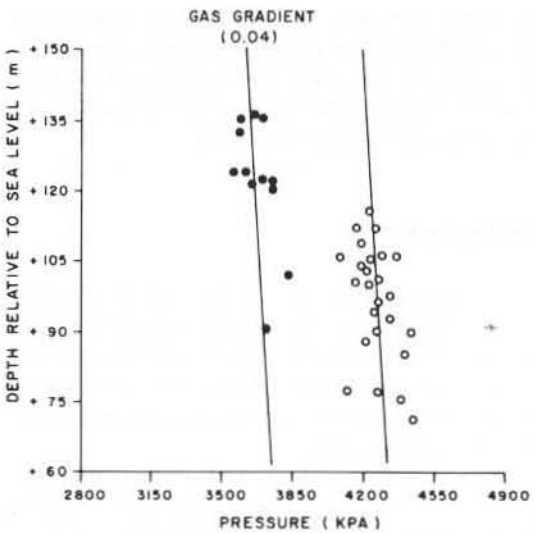


Fig. 3.10. Use of a pressure–depth plot to test lithostratigraphic correlation. The points fall into two groups, indicating that they represent two channel-fill sandstone bodies isolated from each other by fine-grained units. The data were derived from lower Cretaceous fluvial sandstones over an area of about 1900 km^2 in Alberta. (Putnam and Oliver 1980)

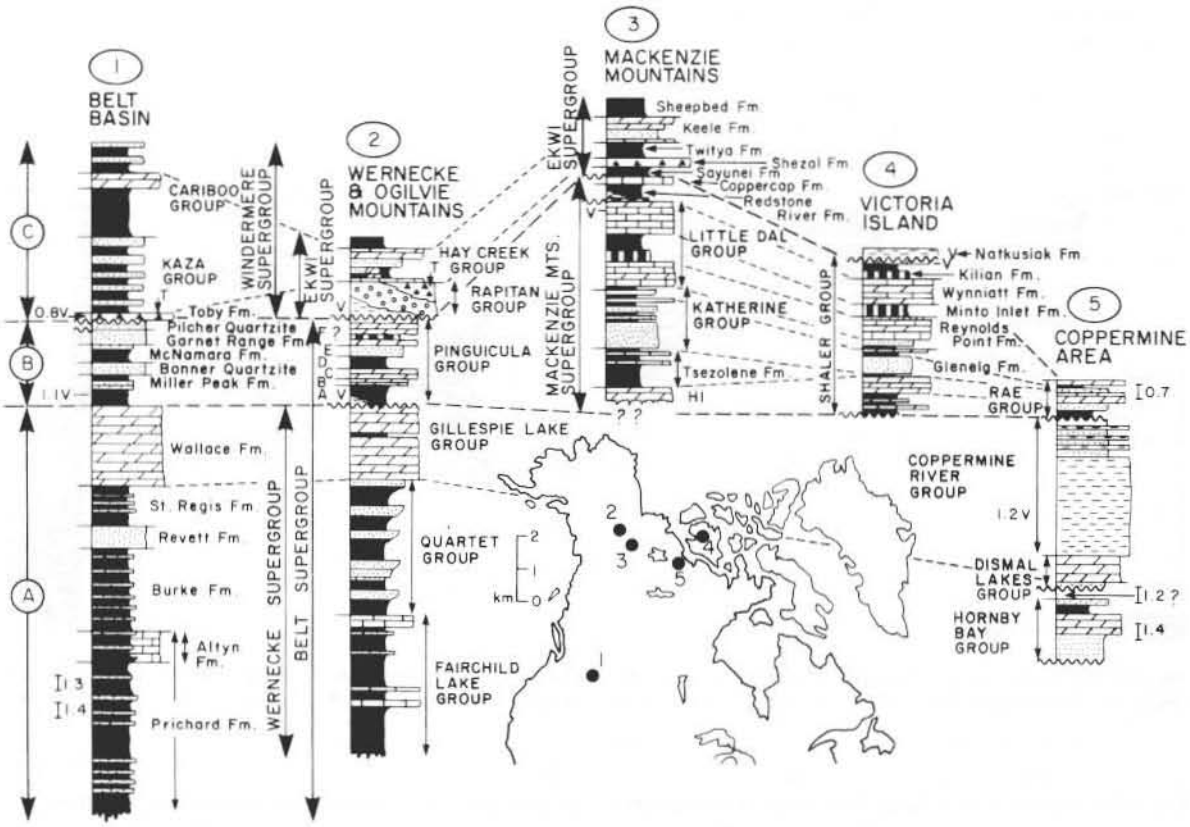


Fig. 3.11. Lithostratigraphic correlation of Proterozoic sedimentary sections. North American craton. (Young 1979)

ed on an understanding of the depositional systems within which the rocks were laid down. Lithostratigraphy is a descriptive, empirical science, but correlations can be made more comprehensive and are likely to be more reliable if they are erected using genetic, sedimentological principles.

3.5 Biostratigraphy

Biostratigraphy is the study of the relative arrangement of strata based on their fossil content. Descriptive or empirical biostratigraphy is used in erecting zones for local or regional stratigraphic correlation and forms the basis for a global system of chronostratigraphic subdivision.

Fossil content varies through a stratigraphic succession for two main reasons: evolutionary changes and ecological differences, such as changes in climate or depositional environment. Biostratigraphy should be based only on evolutionary changes, but it is always difficult to distinguish these from changes that take place in a biostratigraphic assemblage as a result of ecological modifications, and this problem is a cause of continuing controversy for many fossil groups.

Obviously, biostratigraphy can only be studied and a classification erected where fossils are present. This rules out much of the Precambrian strata, although some use has been made of stromatolites for correlation, particularly by Russian geologists, and fossil microbe communities hold promise (Sect. 3.7.9.2). Even in the Phanerozoic, there are many rock units for which the fossil record is very sparse, and biostratigraphic subdivision is correspondingly crude. This is particularly the case in nonmarine strata or those (particularly carbonates) in which fossil remains have been destroyed by diagenesis.

Biostratigraphy is a study for specialists. Refined work requires intimate knowledge of the phylogeny of a large number of fossil groups and their regional or global distribution. To accumulate this knowledge may take half a lifetime, and the subject is an excellent example of a science in which the practitioner seems to spend inordinate amounts of time learning more and more about less and less. Some of the leading authorities in a particular fossil group may be able to discuss the cutting edge of their research with

only half a dozen other colleagues around the world. This gives them considerable value if one happens to find their kind of fossil, but it may somewhat restrict their scientific scope. Geologists engaged in basin analysis of Phanerozoic strata are very rarely such specialists. Biostratigraphers are, therefore, either employed by many organizations to provide these specialized service skills or they function independently as consultants. They may be engaged, much of the time, in pursuing paleontological research but are able to provide biostratigraphic diagnoses for selected fossil types over a specified age range.

Professional biostratigraphic work may take a great deal of field and laboratory time. Sections that a sedimentologist may dismiss as sparsely fossiliferous may yield hundreds or even thousands of specimens to the careful collector. Laboratory extraction of microfossils or palynomorphs may yield similar numbers. It is this kind of work that is necessary for modern, refined biostratigraphic studies. Much of the submitted material, particularly that from frontier exploration wells, may itself provide the basis for new biostratigraphic zoning schemes.

Basin analysts should understand what they are getting when they submit their own material to a specialist for identification. Commonly, all they are interested in is the age of the enclosing rock, information that can be used for correlation purposes. Age is a chronostratigraphic interpretation based on taxonomic descriptions, but, commonly, there are problems with fossil identification or interpretation, particularly where the fossil record is sparse or the material is from a new, poorly studied area. The purpose of this section, therefore, is to discuss some of the problems of the biostratigraphic record and to describe the methods biostratigraphers use in plying their trade.

3.5.1 The Nature of the Biostratigraphic Record

3.5.1.1 *Biostratigraphy and Evolution*

Basically, biostratigraphy is an empirical, descriptive science. Biostratigraphic subdivision and correlation is based on the gradual changes through time of faunas and floras. Such changes reflect evolution, but it is not necessary to understand the evolutionary development (phylogeny) of a group of animals in order to use the faunal information, just as English history is linked to the age range of kings and queens, although it is not necessary to know the family relationships of these individuals in order to make use of the chronology. However, it improves one's understanding of biostratigraphic information if it can be related to evolutionary changes. Useful discussions of evolution have been given by Eldridge and Gould (1977) and Fahraeus (1986). Pearson (1998) explored the relationship between evolutionary processes and the nature of biostratigraphic correlation.

Biologists and paleontologists have described three types of evolutionary pattern. There remains a controversy over the relative importance of these three styles, but there seems little doubt that all three exist in nature, because well-documented examples can be quoted to demonstrate each.

The first was termed **phyletic gradualism** by Eldridge and Gould in 1972 (their 1977 paper is a useful, more recent discussion). It refers to long-term evolutionary change, typically in response to geographical, climatic, or other environmental pressures. Certain varieties of a species may be favored by these changes, so that there is a gradual adjustment in the stock until a distinctive new species appears. Eldridge and Gould (1977) doubted the reality of this process. They maintained that population groups are genetically conservative and that change will only take place on the fringe of the species range, where ecological pressures are most acute. According to this view, most species are stable most of the time, and the hope of paleontologists to be able to continually refine biostratigraphic subdivision by perfecting our knowledge of evolutionary lineages is unlikely to be realized. They suggested that unidirectional environmental change rarely continues for long enough to force evolution, and they quote several examples of supposed lineages that have now been discredited, for example, the famous coiling of the mollusk *Gryphaea* throughout the Jurassic. Trueman (1922) claimed that the shell size, thickness, and degree of coiling could be related to position on an evolutionary lineage, and hence age, with a considerable mathematical precision. Later work (Hallam 1959; Burnaby 1965; Gould 1972) showed that the coiling and other changes were related entirely to ecological (environmental) factors and that similar coiling behavior recurred wherever conditions favored it. Trueman's lineage and the chronology arising from it turned out to be based on circular reasoning.

However, the disproof of one example does not discredit the entire idea. McKerrow (1971) quoted the even earlier work of Rowe (1899) on the Cretaceous echinoid *Micraster*. Rowe was one of the first to attempt to define a detailed evolutionary lineage; his conclusions are shown in Fig. 3.12. The recognition of the succession of species was based on careful stratigraphic collecting in the English Chalk. More recently, Kermack (1954) confirmed the general pattern outlined by Rowe but placed a somewhat different interpretation upon it. He determined that the changes discovered by Rowe represent a gradual broadening of the ecological niche occupied by *Micraster* over a

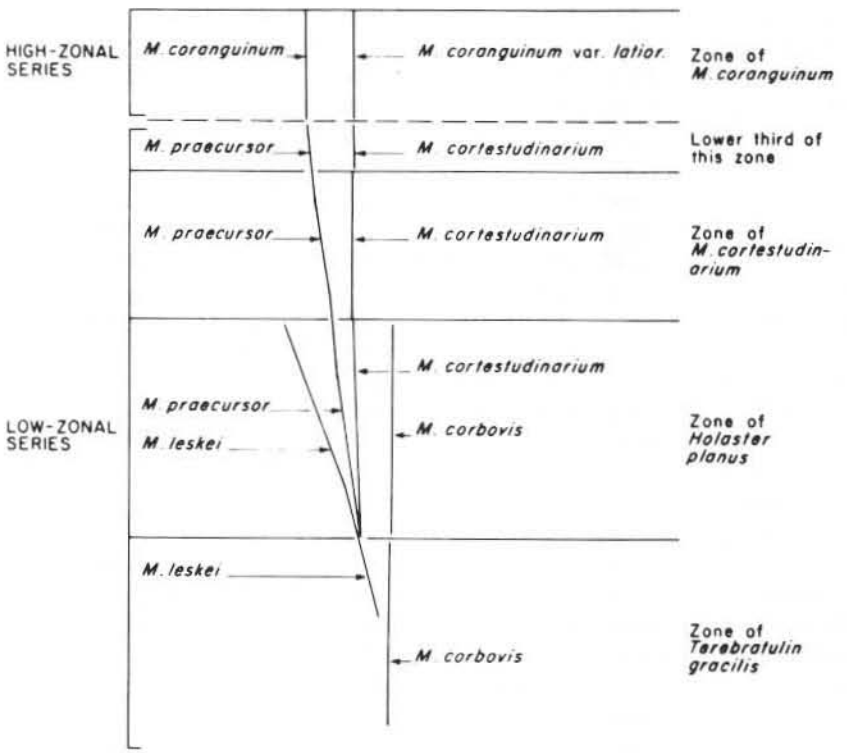


Fig. 3.12. Interpreted lineages of the echinoid *Micraster* in the British Chalk. (upper Cretaceous; Rowe 1899)

period of about 15 m.y., without any perceptible environmental change. Some varieties of the genus developed a gradually deeper burrowing habit. However, these are not separate lineages, because intermediate (interbreeding) varieties are present at every stage (Fig. 3.13). Nevertheless, this seems to represent a good example of phyletic gradualism, and one that has been of biostratigraphic use for over 80 years.

Kauffman (1977) described two examples of phyletic gradualism in Cretaceous pelecypods. Figure 3.14 illustrates a series of histograms of height: width ratios of the *Inoceramus pictus* lineage, derived from populations collected at about 40-cm intervals

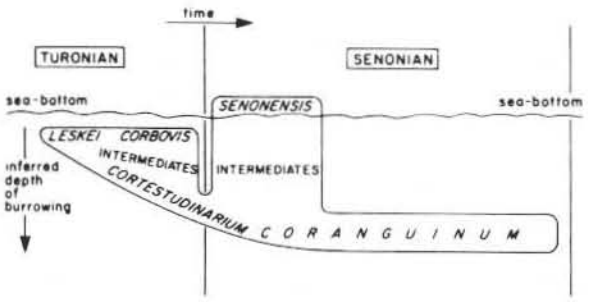


Fig. 3.13. Reinterpretation of *Micraster* phylogeny in terms of a change in its ecological adaptation. It is suggested that a deeper burrowing habit developed with time in some forms, but that intermediate forms were at all times in existence. (Kermack 1954; McKerrow 1971)

(lower graph), and the number of growth ridges in the first 25 mm of shells of *Mytiloides labiatus* (upper graph). The data permit subdivision of the population into species (S), subspecies (SS), and morphological zones (MZ), as indicated in the adjacent columns.

Two types of environmental adaptation can occur: that which is accompanied by permanent genetic change and that which can, to some extent (never precisely), reverse itself to recreate the same variety or race of a species more than once, whenever the same environmental conditions are repeated (homeomorphy). Clearly, the first type is the only one of use to biostratigraphers, but the literature is replete with ambiguous biostratigraphic determinations that may be falsely based on diachronous environmental change. For example, this has been a serious problem with the ammonites (Kennedy and Cobban 1977), one of the best of biostratigraphic indicators.

Taxa that evolve by phyletic gradualism have the most potential for refined biostratigraphic zonation, but they require specialist study to recognize the very subtle changes between the varieties. This type of work is usually beyond the abilities of the generalist basin analyst.

The second style of evolution was named **punctuated equilibrium** by Eldridge and Gould (1977). This concept is based on the premise that, in a successful, widely distributed taxon, the population is genetically conservative. Evolution is thought to occur only where extreme variants are selected by environ-

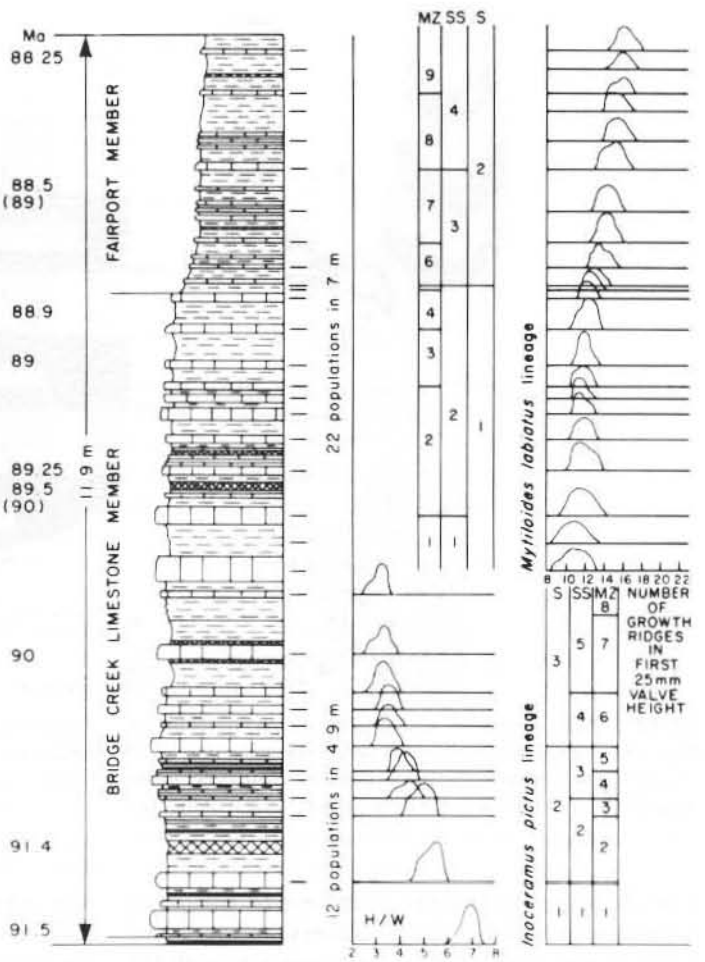


Fig. 3.14. Evolutionary trends in two pelecypod species, Cretaceous rock of Western Interior. The symbol *X* in the lithologic column indicates bentonite beds used for radiometric age dating. Suggested systematics and zonal subdivisions are shown in the numbered columns: *MZ* Morphological biozone; *SS* subspecies; *S* species. (Kauffman 1977)

mental pressures on the fringes of the species range. Rather than a gradual adaptation to an ecological niche or a broadening of a species range by extending slowly into subtly different niches, as in phyletic gradualism, the hypothesis of punctuated equilibrium proposes the spasmodic occurrence of bursts of relatively rapid evolutionary change. Extreme variants of a species can only evolve into a new species if they become isolated by changes in environment, climate, or geography, as through the rifting and drifting apart of continental plates. Also, the catastrophic extinction of organisms by bolide impacts or other catastrophes empties out many ecological niches and permits rapid adaptive radiation and the explosive development of many new taxa in the period immediately following these extinctions. The differences between the two concepts of evolution are illustrated in Fig. 3.15.

Sylvester-Bradley (1977) gave an example of punctuated equilibrium based on the work of Howarth (1973) on the lower Jurassic ammonite genus *Dactylioceras.* Over a few meters of strata, the whorl shape and rib pattern show considerable variation, yet careful collection showed that each range of variation is confined to a distinct stratigraphic interval, indicating sudden small evolutionary changes. Some of the histograms in Fig. 3.14 also show sudden changes and, in fact, these pelecypods are thought to show both styles of evolution at different times. Punctuated equilibrium is a style of evolution that tends to produce major but infrequent taxonomic changes (index fossils) of a type that can commonly be readily recognized for biostratigraphic purposes without specialist training.

Sylvester-Bradley (1977) proposed a third style of evolution, which he termed **reticulate speciation** (Fig. 3.15). This combines, on a small-scale, the mechanisms of both the other two evolutionary styles. Sylvester-Bradley offered the modern common vole as an example of a taxon that has evolved in this way. The vole is distributed virtually globally and has numerous varieties, reflecting adaptation to local variations in climate, vegetation, altitude, isolation on islands, etc. These varieties evolved in response to rapid global changes following the Pleistocene ice age, and they demonstrate the rapidity with which geographical and ecological changes may bring about evolu-

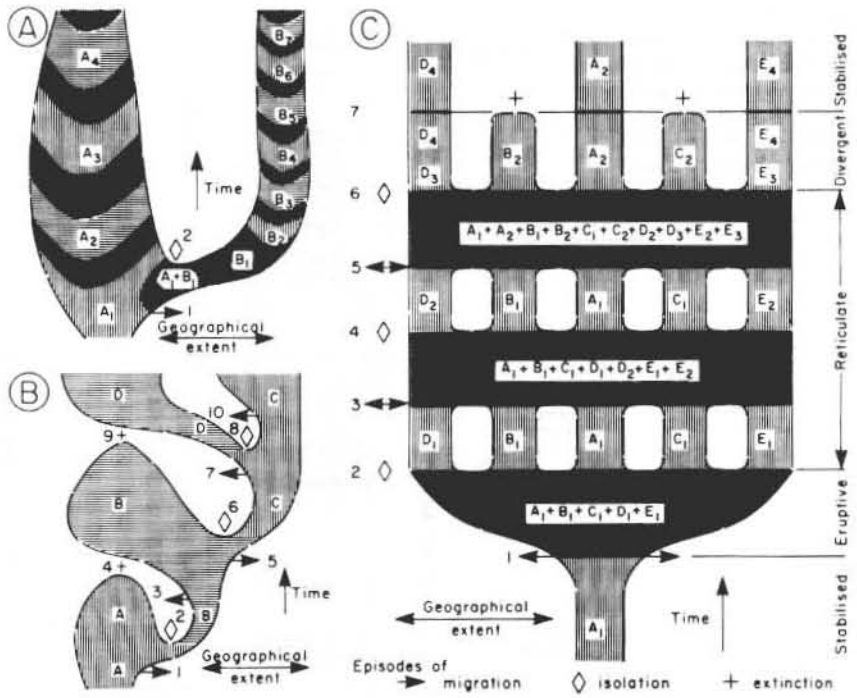


Fig. 3.15. Three models of evolution: **A** phyletic gradualism; **B** punctuated equilibrium; and **C** reticulate speciation. A, B, etc., refer to successive varieties or species; 1, 2, etc., refer to the chronology of events. In **B**, the same area at the left of the diagram is successively occupied by three species (A, B, and C), which evolve elsewhere and migrate in. (Sylvester-Bradley 1977)

tion. The apparent stability of the species of several taxonomic groups for several million years or more at times during the geological past contrasts with the rapid adaptability of the modern vole. To what extent reticulate speciation will be recognized for fossil groups remains to be determined. However, to recognize this style of evolution would seem to require an immense bank of detailed descriptive data and, therefore, it is very much a subject of study for specialists.

3.5.1.2 *Biostratigraphy and Biogeography*

Biogeography is the study of the geographical distribution of taxa, reflecting the restriction of ranges due to ecological variations and the geographical isolation of populations. Two topics are included under this heading: the facies control of faunas and floras and the problem of faunal and floral provinces.

Some taxa are adapted to a benthonic (bottom-dwelling) mode of life and others to a nektonic (swimming) or planktonic (floating) habit. Nektonic or planktonic forms are greatly to be preferred for biostratigraphic purposes because of the likelihood of their being more widely distributed and, therefore, more practically useful. Benthonic forms tend to be more facies dependent because of their need for certain water conditions or sediment types for feeding and dwelling behavior. However, in practice, benthonic forms are widely used by professional biostratigraphers. Even such static forms as corals, burrowing molluscs, or anchored brachiopods have been found to be invaluable for zoning the deposits of the continental shelves. Many benthonic taxa have a planktonic larval stage that ensures wide distribution via marine circulation. Conversely, many planktonic forms, such as the graptolites, are too fragile to survive in agitated, shallow-water environments and are, therefore, just as facies-bound as their benthonic contemporaries. In practice, virtually every taxonomic group has some biostratigraphic utility, although considerable problems may arise in attempts to determine the relationships between the various facies-bound faunas unless environmental fluctuations cause lithofacies of different types, with their accompanying faunas or floras, to become interbedded. Where well exposed, such mixed successions are of great value in establishing a global chronostratigraphic framework, as discussed in Section 3.7.

Classic examples of facies-bound faunas widely used by biostratigraphers are the **shelly** and **graptolitic** faunas of the lower Paleozoic strata. The shelly fauna actually includes two more or less distinct subfaunas, one in the inner, shallower shelf dominated by

brachiopods and the other on the outer shelf, characterized by trilobites. The graptolitic fauna is confined mainly to low-energy deposits of the continental slope, rise, and abyssal plain (Berry 1977).

A good example of facies control of what might appear at first sight to be a recurrent, biostratigraphically controlled fauna is provided by the brachiopod communities of the upper Ordovician to middle Silurian strata of the Welsh Borderlands. Ziegler et al. (1968) showed that, at many localities, there is a sequence of assemblages containing, in upward stratigraphic order, *Lingula*, *Eocoelia*, *Pentamerus*, *Stricklandia*, and *Clorinda*, followed by a graptolitic fauna. Careful correlation of these sections using graptolites showed that the brachiopod sequence is markedly diachronous. This is shown in Fig. 3.16, in which the graptolite zones are shown by horizontal correlation lines labeled A1 to C7. One might ask, why are the graptolites trusted more than the brachiopods for the purposes of chronostratigraphic correlation? The answer is that brachiopods are benthonic organisms known to be prone to facies control, whereas graptolites are tried and tested biostratigraphic indicators. The correlations shown in Fig. 3.16 are supported by additional work on two of the brachiopod genera, *Eocoelia* and *Stricklandia*. When examined in detail, evolutionary trends can be detected within the populations of these genera as they are traced from west to east across the line of the section shown in Fig. 3.16 (and other sections not shown). The overall interpretation of these faunal data is that the sections reveal a gradual eastward transgression and deepening of the water and successive brachiopod communities representing some (unknown) ecological adjustment to increased depths. There is no obvious relationship between brachiopod assemblage and sediment type in this case (we return to these brachiopod assemblages in Section 8.3.2 to discuss their use in unravelling cyclic changes in sea level).

Faunal provinces are much discussed by biostratigraphers. To the uninitiated, such concepts as the Malvinokaffric Province or the Tethyan or Boreal Realms seem nebulous, even mystical. Their description is usually accompanied by global distribution maps showing apparently random dots for each field occurrence of a selected taxon. Rarely are the patterns of dots for different taxa similar, and the innocent reader may find it difficult to determine exactly what it is that defines each province. Clearly, the understanding of what constitutes a given faunal province requires a great deal of specialist knowledge, but, before the advent of plate tectonics, even the specialists had difficulty in comprehending why many of these provinces existed. There was much discussion of the appearance and disappearance of strange, narrow "land bridges" to explain the merging and divergence of provincial variations.

The ammonites provide an excellent example of the various biogeographic styles that can occur in organisms, some offering considerable advantages to the biostratigrapher, others a severe hindrance. Many ammonites underwent a planktonic larval stage that may have lasted from hours to weeks. Where this occurred, it would have been of some importance to the distribution of the species. Not all ammonites showed this. Distribution patterns and varying degrees of facies independence show that some ammonites were benthonic in adult life habitat, some were nektonic, and some may have been planktonic. Their facies distribution and provincial tendencies thus varied considerably. Some ammonites may have drifted long distances after death. The modern Nautilus, the only living relative of the ammonites, has a buoyant shell after death, and observations in modern oceans suggests that the shell may drift for hundreds, if not thousands, of kilometers. Geologically, this could be of great importance, but Kennedy and Cobban (1977) suggested that many ammonites, in fact, became rapidly waterlogged after death and did not float appreciable distances.

Kennedy and Cobban (1977) summarized much of the data for Cretaceous ammonite distribution and concluded that there were five types of faunal provinces.

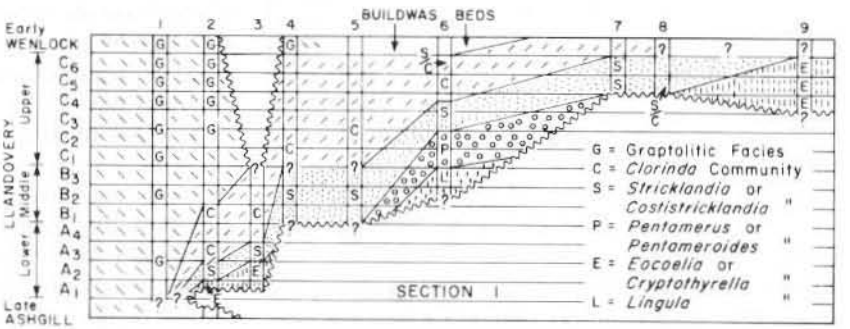


Fig. 3.16. The succession of brachiopod communities in each of these sections is the same (L, E, P, S, C), but use of graptolite biostratigraphy (biozones A1 to C7) shows that they are markedly diachronous and, therefore, facies controlled. Silurian, Welsh Borderland. (Ziegler et al. 1968; McKerrow 1971)

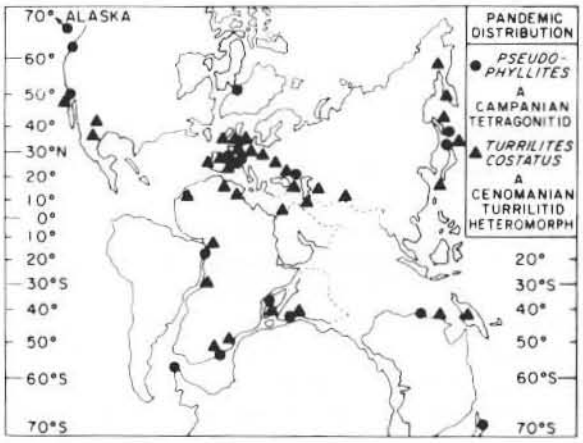


Fig. 3.17. Examples of ammonite genera showing pandemic distributions, plotted on mid-Cretaceous plate-tectonic reconstruction of Smith et al. (1973). (Kennedy and Cobban 1977; reprinted with permission of the publisher)

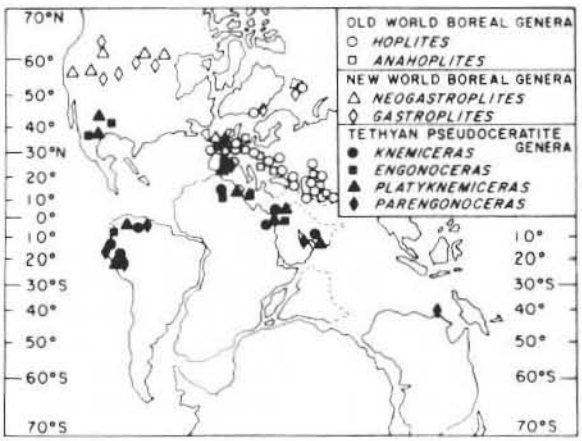


Fig. 3.18. Examples of latitudinally restricted and endemic distributions. (Kennedy and Cobban 1977; reprinted with permission of the publisher)

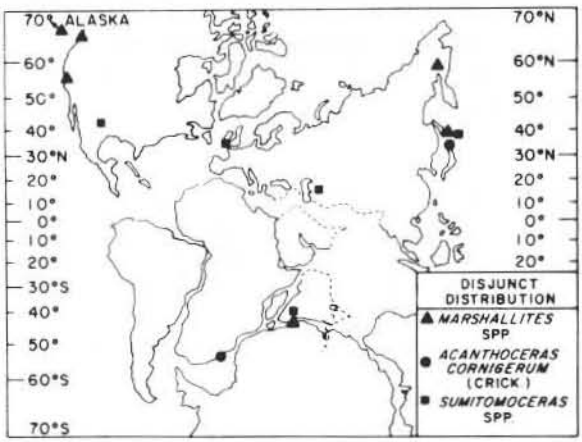


Fig. 3.19. Examples of disjunct distributions. (Kennedy and Cobban 1977; reprinted by permission of the publisher)

Some genera have a virtually worldwide, or **pandemic**, distribution; two examples are shown in Fig. 3.17, plotted on a plate-tectonic reconstruction of the mid-Cretaceous period. These and succeeding illustrations are examples of those dot-plot diagrams, but careful comparison between them reveals some real differences in distribution. Note that, in Fig. 3.17, the two genera have been recorded in every continent except Antarctica (sparse data) and South America. No obvious geographical isolation or latitudinal control seems likely. Pandemic taxa would seem to offer the best possibilities for global correlation. They are relatively facies independent, but it turns out that many are long-ranging forms and, thus, of limited biostratigraphic usefulness.

Some ammonites have **latitudinally restricted** distributions, reflecting their preference for waters of a certain temperature or salinity and their tolerance of seasonal fluctuations. Some examples are shown in Fig. 3.18. They define two provinces, the northern, colder-water Boreal province (*open dots*) and the more tropical Tethyan province (*closed dots*). In many parts of Europe and North America, faunal fluctuations through stratigraphic successions between Tethyan and Boreal (and other) faunas have been cited as evidence of the existence of connecting seaways and transgressions across otherwise barren areas.

Longitudinal restrictions on distribution, such as the presence of land masses or large ocean basins, are a cause of further provincialism. These are added to latitudinal restrictions in the generation of the third type of faunal province: **endemic** distributions. Note that, in Fig. 3.18, the Tethyan genera show no longitudinal restriction, whereas the four Boreal genera are typical endemic taxa, restricted to either Eurasia or North America (these are examples chosen to illustrate a point and should not be taken to define a universal difference between the Boreal and Tethyan provinces). Endemic ammonites have been shown to have evolved rapidly and, thus, are of prime biostratigraphic importance, although their provincialism has hindered intercontinental correlation.

Disjunct distributions are those of scattered but nevertheless widely distributed taxa, three examples of which are shown in Fig. 3.19. The distributions are not thought to represent inadequate data or severe facies control but probably reflect very low population densities.

As noted, some ammonite taxa may drift in oceanic currents after death. In extreme cases, where an endemic form is involved, such drifted or **necrotic** distributions may prove invaluable for long-distance correlation. An example is shown in Fig. 3.20.

Many examples of faunal provincialism and facies control could be quoted. The type of province de-

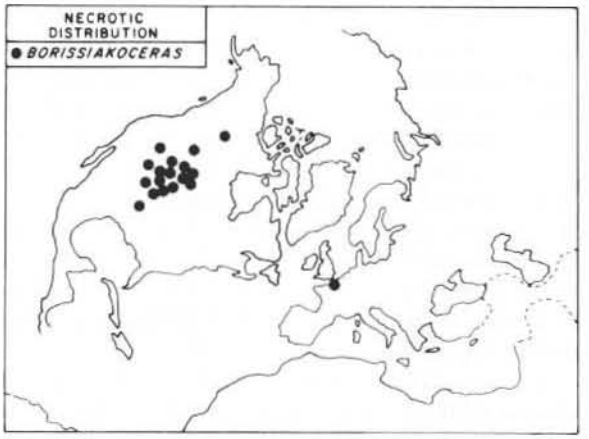


Fig. 3.20. Example of a necrotic distribution. Note lone occurrence in France. (Kennedy and Cobban 1977; reprinted with permission of the publisher)

pends on the mobility of the organism and the degree of fragmentation of the continents. For example, faunas and floras tended to be less provincial during the period when continents were combined into Pangea (Valentine 1977). The brief discussion of provinces here is intended to provide only background information on some of the typical problems encountered by biostratigraphers when dealing with faunal and floral distributions. For further examples and more specialized discussion, the reader is referred to such texts as Middlemiss et al. (1971), Hallam (1973), Hughes (1973), Kauffman and Hazel (1977), and Gray and Boucot (1979).

3.5.1.3
Life, Death, and Reworked Assemblages

When fossils are used as biostratigraphic indicators, it is assumed, if not explicitly examined and stated, that they represent the same time period as the rock in which they occur. Some conglomerates contain fossils as clasts, indicating a dramatically different age; for example, I have found Permian corals as pebbles in an Early Cretaceous sandstone and Silurian fossils within limestone boulders in a Paleocene fanglomerate. However, these are obvious examples and easy to recognize. Where the indicated age is not so different, the biostratigrapher must be on guard against using such reworked material. For example, Barnes et al. (1976) described attempts to relate graptolite faunas in shales of the Ordovician Cow Head Group of Newfoundland to trilobites and conodonts contained in limestone boulders with the shales. The limestone and its shelly faunas were derived from the nearby shelf and slumped into the deep basin as olistho-stromes (this term is discussed in Sect. 9.3.2). The interbedding of the two faunas is biostratigraphically useful, but it must be assumed that the boulders are slightly younger than the shales and may themselves not always be the same age. Several tests are used by biostratigraphers to distinguish life and death assemblages.

If enough specimens can be collected or measured in place, their size distribution may indicate whether the population is a life assemblage (**biocoenosis**) or has been reworked into a death assemblage (**thanatocoenosis**). Boucot (1953) argued that many marine invertebrates have high reproductive and initial death rates. This results in a skewed size distribution histogram, with very large numbers of larval and juvenile forms, a moderate number of adults, and a much smaller number of large, gerontic individuals. In the initial stages, the organism may not develop hard parts capable of preservation but, nevertheless, a fossil biocoenosis is expected to show a skewed distribution curve. The action of wave and current sorting on dead shell material will tend to winnow out smaller individuals, and a transported assemblage is likely to be well sorted, with a more nearly normal, bell-shaped size distribution histogram.

For pelecypods and brachiopods, which consist of two shells held together by soft muscle that decays after death, two additional tests can be made. The ratio of articulated (still joined) to disarticulated shells and the ratio of opposite valves may be indicative. The chief factors affecting disarticulation are the strength and duration of bottom currents versus the rapidity of postmortem burial. Where the shells have separated, they tend to respond differently to currents because, in most cases, opposing valves are of different size or shape. Sorting may, therefore, tend to build a concentrated deposit consisting preferentially of one or the other of the valves of a given species (Boucot 1953).

The presence of a death assemblage accumulated by selective sorting, as described here, is not necessarily a biostratigraphic problem. Such assemblages may be produced over a period of only a few weeks or months or even in a single day during a violent storm – time periods of insignificant biostratigraphic importance. However, at times of slow sedimentation or during the re-erosion of relatively recent deposits, a death assemblage of the wrong age may be produced. For example, in areas of negligible sediment supply, Kennedy and Cobban (1977) reported that as many as five ammonite zones may be mixed in a single condensed sequence. After burial, the process of fossilization involves diagenetic changes such as replacement of the original aragonite by calcite, phosphate, pyrite, or silica, filling of living chambers by cement, and complete dissolution of the shell. Such changes may take place relatively rapidly. Storm re-erosion,

deepening of a submarine canyon, or erosion during a regression may re-expose beds bearing such fossils that are now sufficiently resistant to act as clasts in sedimentation processes.

Additional observations can be made to distinguish such death assemblages. The degree of chemical alteration and the amount of breakage, abrasion, rounding, etc., are all likely to be much greater than that of other fossils enclosed with them that were living immediately prior to burial.

Certain microorganisms seem to be particularly susceptible to such reworking. Palynomorphs are highly resistant to destruction (among other ordeals, they survive laboratory extraction processes that make use of hydrofluoric acid). For example, modern sediments of the Mississippi Delta contain abundant derived Cretaceous planktonic foraminifera from the midcontinent (Otvos and Bock 1976); Cenozoic deposits in the Canadian Arctic commonly contain more derived Cretaceous and Devonian taxa than indigenous species. Biostratigraphers may have great difficulty seeing through smoke screens such as these. Of course, the taxa commonly show obvious taxonomic differences, but the degree of alteration and abrasion may also be a useful clue. Palynomorphs undergo a color change during burial as a result of progressive oxidation (Staplin 1969, 1977; Sect. 7.6). The color changes are retained during reworking and can be distinctive where intermixed with younger, fresher material.

The drilling process may produce reworked microorganism assemblages by the process of caving or recycling, as discussed in Chapter 2. Microorganisms may even be present in the drilling mud. Techniques similar to those discussed above can be used to recognize and discount such material during subsurface biostratigraphic evaluation.

3.5.1.4 *Conclusions*

The intent of this section has been to show that a single fossil occurrence or a suite or succession of fossils does not necessarily reveal simple or accurate biostratigraphic information. The biostratigrapher may suffer from the geological equivalent of a whiteout – unable to move for lack of a fixed datum. First, it must be demonstrated that the assemblage represents the same time period as the host stratum. Next, the assemblage or, better still, sequence of assemblages must be sought in more than one location, otherwise the objective of correlation is not being achieved. However, even if the same assemblage or sequence is found elsewhere, does this prove contemporaneity, or are the particular fossils diachronous facies-bound forms? A consistent relationship between fossil type and sediment type might suggest the latter. If the assemblage of interest cannot be located in another section, does this mean the rocks are of a different age, the same age but a different facies, or the same (or different) age and a different faunal province? How do we correlate a nonmarine vertebrate assemblage with a marine ammonite or foraminiferal assemblage? What if there is more than one apparently useful biostratigraphic indicator, yielding contradictory correlations? Which one do we trust? Figure 3.16 is a simple example of this in which the answer is clear (the brachiopods are environment-dependent), but far more complex cases could be presented.

The answers to all this come from years of painstaking work by many specialists. We will return to some of these questions in Section 3.7, when it should become clearer (if it is not already) why few basin analysts are their own biostratigraphers.

3.5.2 Biozones

The basic biostratigraphic unit is the biozone, which can be defined as the body of strata containing a prescribed fauna (flora) or faunal (floral) element. Ideally, taxa are chosen that have a short time span, are geographically widely distributed, as independent as possible of facies, and abundant. As discussed in this section, a variety of biozone types have been erected to make the best use of the different ways in which faunas and floras appear and disappear. The concept of the overlapping distributions of organisms in time and (eco-)space is shown in Fig. 3.21.

The historical evolution of much of this nomenclature was entertainingly discussed by Hancock (1977). Further discussions of the various biozone types and biostratigraphic practices are contained in Hedberg (1976), Harland et al. (1972), and other stratigraphic guides. Ludvigsen et al. (1986) discussed the relationship between biozones and biofacies (see also

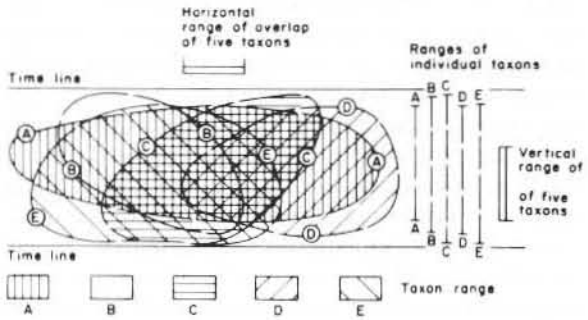


Fig. 3.21. Variations in the extent of a concurrent-range biozone, depending on the number of taxa and degree of concurrence required. (Hedberg 1976; reprinted with permission of the publisher)

Sect. 4.4.1). In practice, many biostratigraphic publications do not make completely clear which biozone type is in use. Biozones are named after prominent or otherwise important taxa or are given formal code designations. The basis for the construction of biozones (biozonation) is the range chart (e.g., Fig. 3.21), in which the stratigraphic extent of each species is plotted.

3.5.2.1 *Assemblage Biozone*

The assemblage biozone is a body of strata characterized by a certain assemblage of fossils without regard to their ranges. The limits of the zone are drawn within the rock body to encompass all the strata that contain most or all the components of the assemblage. Judgment must be used where some members of the assemblage are absent. The assemblage may consist of whatever fauna or flora are present but, in practice, the assemblage normally is restricted to one type of fossil group, otherwise the distribution of the assemblage might be adversely affected by diverse ecological requirements of the individual components.

In fact, the assemblage-biozone concept is hampered by the fact that no attention is paid to ranges. Assemblages may be entirely ecologically controlled and represent little more than fossil communities that recur whenever conditions are appropriate. Obviously, with time the fauna (flora) will evolve and, thus, taxonomic differences will become apparent.

3.5.2.2 *Range Biozone*

The range biozone is the body of strata representing the total range of occurrence (horizontal and vertical) of a particular species, genus, etc. A modification of this is the **acme biozone**, which is defined on the basis of the exceptional abundance of the chosen taxon. Difficulties may arise from dependence on a single species, because its occurrence may become sparse at the limits of its range, making it difficult to decide where to draw the boundary. The concept of acme should, preferably, be defined in some quantitative manner. Both total range and acme may be affected by local ecological (facies) considerations. For these reasons, it is better to rely on more than one taxon, as in the concurrent-range biozone.

3.5.2.3 *Concurrent-Range Biozone*

The concurrent-range biozone is defined as the body of strata characterized by the overlapping ranges of two or more selected taxa. This is one of the most commonly used types of biozone. Modern work normally makes use of many species and may also incorporate abundance data in order to avoid problems of local facies variation. This is easy to do in the case of microorganisms that tend to occur in great abundance in many rocks.

Decisions must be made as to how many of the selected taxa must be present in a sample in order to define the zone. As Hedberg (1976) pointed out, each taxon will probably have a different distribution in space and time (Fig. 3.21), so some judgment is required. This is where the extensive knowledge and experience of the biostratigrapher becomes important.

Figure 3.22 is an example of palynological zonation of two wells through a Cretaceous succession in Delaware (from Doyle 1977). The wells are represented by their gamma-ray logs, with sample collection depths given in feet. The ranges of the principal angiosperm pollen types are shown by vertical bars, shown dashed where identification is uncertain. Concurrent-range zones I to IV are shown by horizontal dotted lines and are numbered at the top. It was found that the zones could be most easily defined on the basis of first (oldest) appearance of a taxon, partly because extinct species tended to be reworked and partly because taxa were found to die out slowly at the upper limit of their range. Work of this type required the counting and documentation of several hundreds of individual pollen grains in each sample. Several or many complete sections through the succession of interest may be required before the data are adequate for the definition of the biozones. Range charts, such as that in Fig. 3.22, must be prepared for each and carefully compared.

Figure 3.23 shows a series of concurrent-range zones that have been erected for the Maastrichtian (upper Cretaceous) Chalk of northwest Europe, based on belemnites, coccoliths, brachiopods, foraminifera, and ammonites (from Surlyk and Birkelund 1977). Zones are given arbitrary (but permanent) numbers or are named after distinctive species within the zone. This diagram summarizes an immense amount of work. More than 1200 species are present in the Danish Maastrichtian. Surlyk and Birkelund (1977) collected 10–30 kg of rock every 2–3 m in all major Danish sections. Washing and boiling-freezing treatment allowed the fossils to be picked out clean. About 70,000 brachiopods were obtained in this way. Each 10-kg sample contained 100–4000 specimens of all taxa.

A variation on the concurrent-range biozone is the **Oppel zone**, named after the nineteenth-century German geologist who introduced many of the modern ideas about zoning methods. The Oppel zone is, essentially, a concurrent-range biozone in which supplementary biostratigraphic criteria are used in its

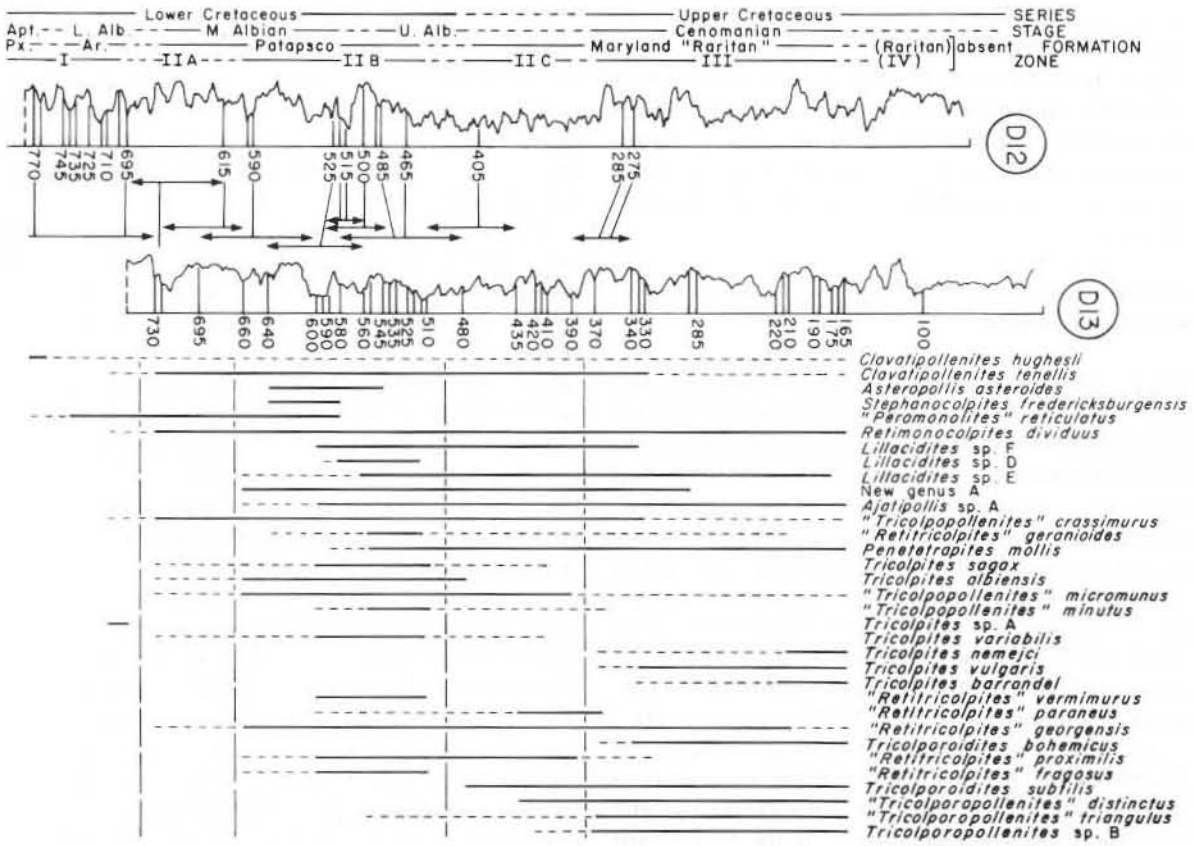


Fig. 3.22. Use of palynological concurrent-range biozones to correlate two subsurface wells (as shown by gamma-ray logs). Correlation brackets (*double-headed arrows*) terminate just above and below samples in the other well that bracket the age of the indicated sample. (Doyle 1977; reprinted with permission of the publisher)

definition, such as abundance data or the presence or range of some unrelated taxon. Many older zoning schemes were of the Oppel type, though they may not have been so identified by the use of his name.

3.5.2.4 *Lineage Biozone*

This may be defined on the basis of the evolutionary sequence of a particular taxon. It may be a simple linear change, such as the pelecypod shape and growth-ridge trends shown in Fig. 3.14, or the branching or radiation of a taxon into new varieties or lineages, such as the adaptation of the echinoid *Micraster* to new ecological niches, shown in Fig. 3.13. Several examples of Cretaceous foraminifera lineage biozones were given by Van Hinte (1976b). As discussed in Section 3.5.1, the style of evolution (in particular, phyletic gradualism versus punctuated equilibrium) will markedly affect the ease with which lineage zones may be established. The most rapid evolution is most likely to occur on the fringes of the taxon's geographic range. Zonation can best be established there, although it may be of little use in studying conservative stock in a stable environment at the center of the range. Ideally, lineage biozones should permit the establishment of very refined biostratigraphic subdivisions, but the work of recognizing and discriminating patterns in a series is complex and can be subjective (Hay and Southam 1978).

3.5.2.5 *Other Biozone Terms*

Many other terms have been used by biostratigraphers for variations on the main zone types described here (George et al. 1969; Van Hinte 1969; Hay and Southam 1978; Hedberg 1976). For the total range of a taxon, the terms range zone, **acrozone**, and **total-range zone** have been used. Acme zones have also been called **epiboles**, **peak zones**, and **flood zones**. Local range zones may be called **teilzones** or **topozones**. Lineage zones are sometimes called **phylozones**, **morphogenetic zones**, or **evolutionary zones**.

Some biostratigraphic methods rely on the **first-appearance datum** (FAD) and **last-appearance da-**

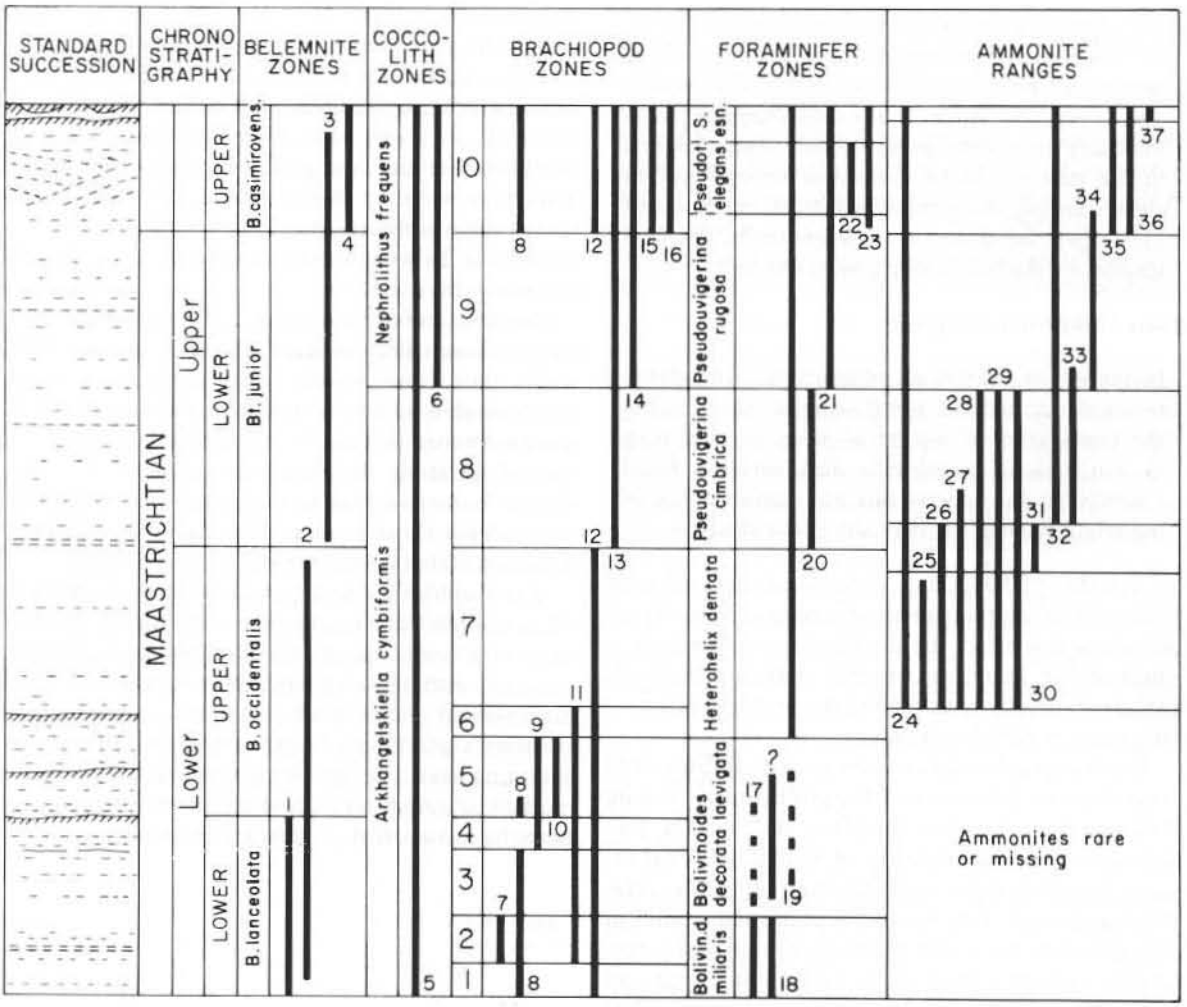


Fig. 3.23. Concurrent-range biozones of Maastrichtian Chalk in northwest Europe. Numbers beside each range bar refer to individual species. (Surlyk and Birkelund 1977; reprinted with permission of the publisher)

tum (LAD) of key, widely distributed taxa. Berggren and Van Couvering (1978) discussed some examples. Statistical treatment of this kind of information is made use of in the graphic correlation technique (Sect. 3.5.3.1).

3.5.3 Quantitative Methods in Biostratigraphy

During the last 30 years, a rather different approach to biostratigraphic subdivision has evolved, contrasting with the sometimes subjective zoning methods described in the preceding section. Excellent syntheses have been provided by Agterberg (1990) and Geux (1991). The new approach relies on quantitative techniques, including the use of multivariate statistics. It is claimed that these methods are more objective, in that the biostratigraphic subdivisions emerge from normal analysis rather than interpretation, extrapolation, generalization, and judgment. Such methods have not received universal acceptance; there is an argument widely heard in geology that "if you cannot see a trend in the data, no amount of statistics is going to prove that one exists". More importantly, critics of the methods point out that many techniques deal with matching or **seriation**, not correlation. Matching is simply the process of determining the degree of comparison between data sets and could result in an apparently close match between sets having similar biofacies but very different ages (Sect. 4.4.1). Seriation is the determination of the order of events and, again, this may have little or nothing to do with correlation.

Geux (1991), in discussing the use of multivariate methods, quoted Millendorf and Heffner (1978), who stated:

This approach ignores the effects of faunal gradation within an isochronous unit with respect to geographic position. Thus, if the faunal composition of such an isochronous unit changes across the study area, samples taken from distant points in the unit might be dissimilar enough not to cluster. Simply, the greater the lateral variation and the larger the distance between them, the less similar are the two isochronous samples.

Geux (1991) himself stated:

In one way or another, all methods based on global resemblance between fossil samples end in fixing the boundaries of statistical biofacies, and they do not make it possible to find, within a fossil assemblage, the species that are characteristic of the relative age of the deposits under study.

As Agterberg (1990; Chap. 2) demonstrated, not only biofacies but sampling methods and questions of preservation also affect the distribution of fossils. It is questionable, therefore, whether statistical methods can assist directly with solving the problem of assessing error in global correlation.

Biostratigraphic data sets are extremely difficult to treat objectively because of the problems of missing data due to incomplete sampling and missing sections, the varying reliability of individual observations, lateral changes resulting from biofacies variations, and so on. Very few of the published studies in this area have been able to produce improved correlations without a great deal of data processing and computer analysis involving considerably more effort than traditional methods. Is it worth it? The debate continues (for an incisive criticism of one of the books quoted in this section, see Shaw and Miller 1985). These criticisms cannot be applied to the first of the techniques to be described, which is routinely used in some major petroleum companies, and it is clear that statistical techniques are essential to exploit the vast amounts of micropaleontological and palynological data that are being accumulated during routine petroleum-exploration drilling.

3.5.3.1
The Graphic Correlation Technique

This method was first described in a landmark book by Shaw (1964). Useful explanations of the technique were given by Miller (1977) and Edwards (1984, 1985). Mann and Lane (1995) edited a research collection devoted to the application of this topic to practical problems in basin analysis. The example used herein has been borrowed from Miller (1977).

As with conventional biostratigraphy, the graphic method relies on the careful field or laboratory recording of occurrence data. However, only two items of data are noted for each taxon: the first (oldest) and last (youngest) occurrence (the FAD and LAD). These define a local range for each taxon. The objective is to define the local ranges for many taxa in at least three complete sections through the succession of interest. The more sections that are used, the more nearly these ranges will correspond to the total (true) ranges of the taxa. To compare the sections, a simple graphical method is used.

One particularly complete and well-sampled section is chosen as a **standard reference section**. Eventually, data from several other good sections are amalgamated with it to produce a **composite standard reference section**. A particularly thorough paleontologic study should be carried out on the standard reference section, as this enables later sections, for example, those produced by exploration drilling, to be correlated with it rapidly and accurately.

The graphic technique, which will now be described, is used both to amalgamate data for the production of the composite standard and for correlating the standard with new sections. Figure 3.24 shows a two-dimensional graph in which the thicknesses of two sections X and Y have been marked off on the corresponding axes. The first occurrence of each taxon is marked by a *circle* on each section and the last occurrence by a *cross*. If the taxon occurs in both sections,

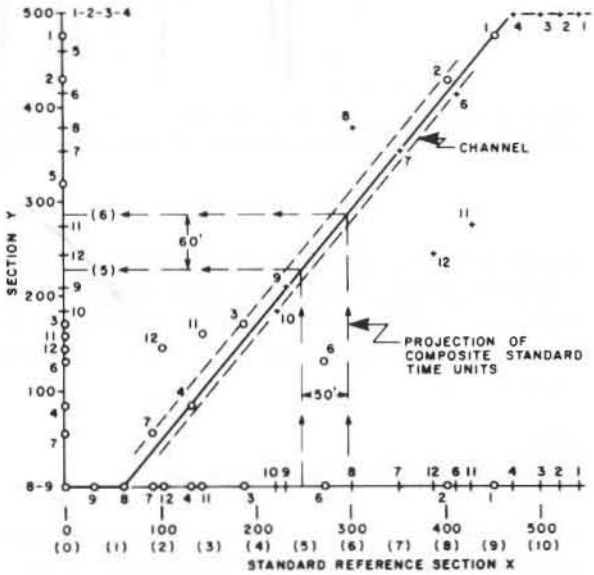


Fig. 3.24. This and the next three figures illustrate Shaw's (1964) graphic correlation method, as discussed by Miller (1977). This plot shows the distribution of first occurrences (*open circles*) and last occurrences (*crosses*) in two sections and positioning of the line of correlation. The channel is the zone on either side of the line of correlation encompassing the observation error. (Kauffman and Hazel 1977; reprinted with permission of the publisher)

points corresponding to first and last occurrences can be drawn within the graph by tracing lines perpendicular to the X and Y axes until they intersect. For example, the plot for the top of fossil 7 is the coincidence of points X = 350 and Y = 355.

If all the taxa occur over their total range in both sections and if sedimentation rates are constant (but not necessarily the same) in both sections, the points on the graph fall on a straight line called the line of correlation. In most cases, however, there will be a scatter of points. The X section is chosen as the standard reference section, and ranges will presumably be more complete there. The line of correlation is then drawn so that it falls below most of the first-occurrence points and above most of the last-occurrence points. First-occurrence points to the left of the line indicate late first appearance of the taxon in section Y. Those to the right of the line indicate late first appearance in section X. If X is the composite standard, it can be corrected by using the occurrence in section Y to determine where the taxon should have first appeared in the standard. The procedure is shown in Fig. 3.25. Arrows from the first occurrence of fossil 6 show that, in section X, the corrected first appearance should be at 165 ft (49.5 m). The same arguments apply to the points for last appearances. Corrections of the kind carried out for fossil 6 (and also the last occurrence of fossil 8) in Fig. 3.25 enable refinements to

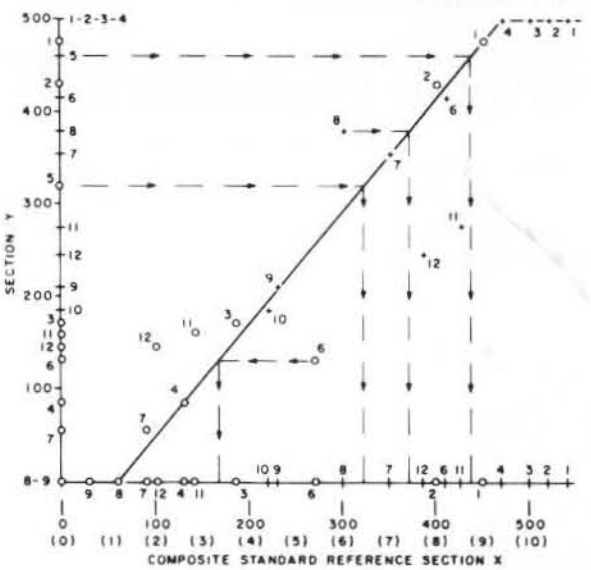


Fig. 3.25. The method used to compile a composite standard reference section. Data from new sections may be used to extend the range of occurrence of taxa that do not show their full range in the standard section (lowest occurrence of species 6, highest occurrence of species 8) and may also be used to transfer data onto the standard section, such as the range of species 5, which does not occur in the latter. (Miller 1977; reprinted with permission of the publisher)

be made to the reference section. Combining several sections in this manner is the method by which the composite standard is produced. Data points can also be introduced for fossils that do not occur in the reference section. In Fig. 3.25, *arrows* from the first and last occurrences of fossil 5 in section Y show that it should have occurred between 320 ft (96 m) and 437 ft (128 m) in Sect. X.

If the average, long-term rate of sedimentation changes in one or other of the sections, the line of correlation will bend. If there is a hiatus (or a fault) in the new, untested sections (Sect. Y), the line will show a horizontal terrace. Obviously, the standard reference section should be chosen so as to avoid these problems as far as possible. Harper and Crowley (1985) pointed out that sedimentation rates are, in fact, never constant and that stratigraphic sections are full of gaps of varying lengths (Sect. 3.7.7). For this reason, they questioned the value of the graphic correlation method. However, Edwards (1985) responded that, when due regard is paid to the scale of intraformational stratigraphic gaps versus the (usually) much coarser scale of biostratigraphic correlation, the presence of gaps is not of critical importance. Longer gaps, of the scale that can be detected in biostratigraphic data (e.g., missing biozones) will give rise to obvious bends in the line of correlation, as noted previously.

The advantage of the graphic method is that, once a reliable composite standard reference section has been drawn up, it enables chronostratigraphic correlation to be determined between any point within it and the correct point on any comparison section. Correlation points may simply be read off the line of correlation. The range of error arising from such correlation depends on the accuracy with which the line of correlation can be drawn. Hay and Southam (1978) recommended using linear regression techniques to determine the correlation line, but this approach assigns equal weight to all data points instead of using one standard section as a basis for a continuing process of improvement. But, as Edwards (1984) noted, all data points do not necessarily have equal value; the judgment and experience of the biostratigrapher are essential in evaluating the input data. For this reason, statistical treatment of the data is inappropriate.

Figures 3.26 and 3.27 illustrate an example of the use of the graphic method in correlating an upper Cretaceous succession in the Green River Basin, Wyoming, using palynological data (from Miller 1977). The composite standard reference section has been converted from thickness into **composite standard time units**, by dividing it up arbitrarily into units of equal thickness. As long as the rate of sedimentation in the reference section is constant, these time units will be of constant duration, although we

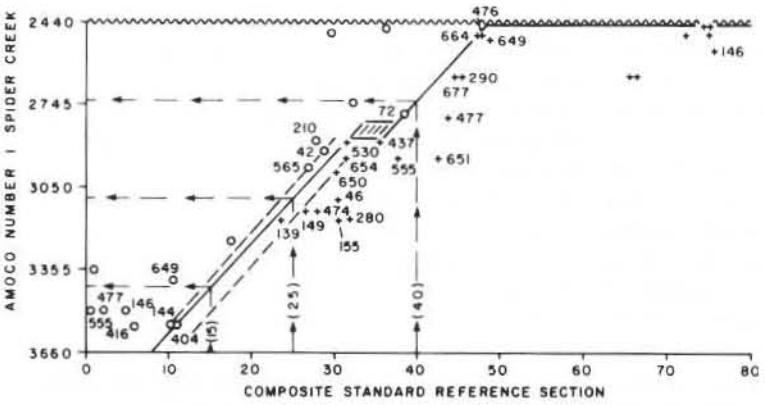


Fig. 3.26. An example of the use of the graphic method, showing a plot of data from one well against the composite standard. The break and shaded area on the line of correlation are interpreted as an unconformity. The reference section has been divided into arbitrary thickness units (composite standard time units). (Miller 1977; reprinted with permission of the publisher)

cannot determine, by this method alone, what their duration is in years. Figure 3.26 shows the method for determining the position of selected time lines on each test section and, in Fig. 3.27, the time units are used as the basis for drawing correlation lines between four such sections. Note the unconformity in each illustration and the variation in sedimentation rates in Fig. 3.27.

The value of the graphic method for correlating sections with highly variable lithofacies and no marker beds is obvious, and it is perhaps surprising that the method is not more widely used. An important difference between this method and conventional zoning schemes is that zoning methods provide little more than an ordinal level of correlation (biozones, as expressed in the rock record, have a finite thickness that, commonly, cannot be further subdivided), whereas the graphic method provides interval data (the ability to make graduated subdivisions of relative time). Given appropriate ties to the global time frame, the composite standard time units can be correlated to absolute ages in years and used to make precise interpolations of the age of any given horizon (such as a sequence boundary) between fossil occurrences and tie points. The precision of these estimates is limited solely by the accuracy and precision obtainable during correlation to the global standard. MacLeod and Keller (1991) provided excellent examples of this procedure, and their results suggest an obtainable precision of less than ±100 ka.

3.5.3.2
Use of Multivariate Statistics

Many different multivariate statistical techniques have been used, including cluster analysis, principal-

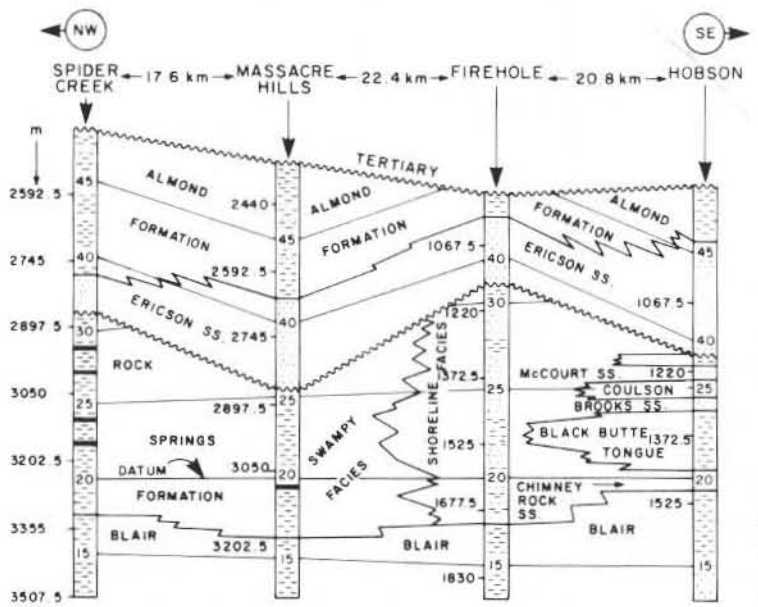


Fig. 3.27. Correlation of four wells in an area of marked lateral-facies change using the composite standard time units from Fig. 3.26. (Miller 1977; reprinted with permission of the publisher)

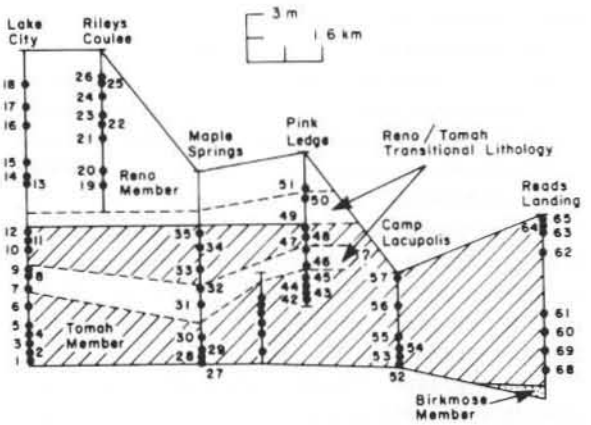


Fig. 3.28. This and the next two diagrams illustrate the use of cluster analysis as an aid to biostratigraphic zonation. This figure shows the lithostratigraphic correlation of some Cambrian rocks in Minnesota and the positions of numbered trilobite collections. (Hazel 1977; reprinted with permission of the publisher)

components analysis, principal-coordinates analysis, and multidimensional scaling (Hazel 1977). Hay and Southam (1978) discussed a technique for quantifying the relative biostratigraphic value of a taxon based on facies independence and vertical and horizontal range. Other refinements are also covered in their paper. The relationship between biostratigraphy and biofacies was explored by Ludvigsen et al. (1986). Several other approaches and useful discussions are given by Harper (1981), Edwards (1982), Agterberg and Gradstein (1981), and in articles in the books by Cubbitt and Reyment (1982), Gradstein et al. (1985), Agterberg (1990), and Geux (1991).

A single example of these applications is given here, a cluster analysis evaluation of trilobite data from an upper Cambrian unit in Minnesota (Hazel 1977). Sixty-five samples were collected from seven

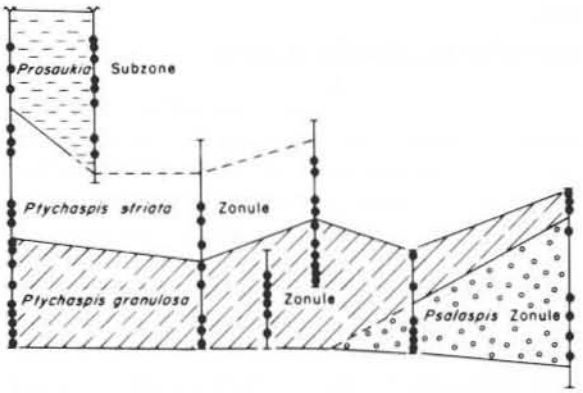


Fig. 3.30. Occurrence in the sections of the clusters outlined in Fig. 3.29, defining four assemblage zones. (Hazel 1977; reprinted with permission of the publisher)

sections and were found to contain 24 species (Fig. 3.28). A Q-mode analysis, which compares all the samples with each other on the basis of presence–absence data for each species, was performed. A table of similarity-coefficient values was produced; this table enables a dendrogram to be drawn, as in Fig. 3.29. Each major cluster corresponds approximately to an assemblage biozone, the distribution of which is shown by transferring the ornamentation in Fig. 3.29 back to the stratigraphic cross section, as shown in Fig. 3.30. Different methods of calculating the similarity coefficients will produce slight differences in clustering but, in general, the method can reproduce objectively and very rapidly what would otherwise be a laborious manual procedure.

Quantitative methods are useful for the treatment of large quantities of data, such as the Maastrichtian Chalk study referred to previously (Sect. 3.5.2; Fig. 3.23). They are particularly valuable for studies of microfossils and palynomorphs, which tend to occur in great abundance.

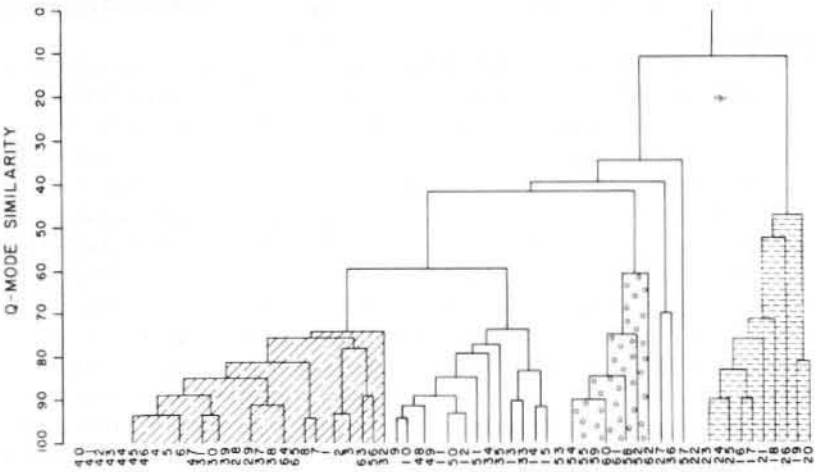


Fig. 3.29. Q-mode dendrogram using 65 samples located as shown in the previous diagram and containing 24 trilobite species. (Hazel 1977; reprinted with permission of the publisher)

3.6 Unconformity-Bounded Units

Unconformities have commonly served as convenient boundaries for various types of stratigraphic units. In many areas, regional unconformities have long been used to define natural subdivisions of the stratigraphic record. For example, the sequences of Sloss et al. (1949) were defined as operational units separated by "marked discontinuities in the stratal record of the craton, which may be traced and correlated for great distances on the objective bases of lithologic and faunal 'breaks'..." These sequences formed the basis for the regional and global cycles discussed in Chapters 6 and 8. The use of unconformities as boundaries is now avoided in the definition of chronostratigraphic units for reasons explained in Section 3.7.3. However, as a way of clarifying geological relationships and important stratigraphic events throughout a particular region, the definition of a separate class of unconformity-bounded units has distinct benefits. Increasingly, it is being recognized that the stratigraphic record contains units of regional and possibly even global extent caused by widespread changes in sea level or by regional tectonic events (Emery and Myers 1996; Miall 1997). Chapter 6 contains a detailed description of the sedimentary geology of stratigraphic sequences and Chapter 8 is devoted to an examination of the allocyclic (tectonic, eustatic, and other) mechanisms that lead to the development of regional and global cycles. Here, we discuss briefly the formalization of the descriptive stratigraphy of such units, based on proposals by the International Subcommission on Stratigraphic Classification (ISSC, 1987) and the definitions contained in the *North American Stratigraphic Code* (NASC).

Unconformity-bounded units include other types of unit within them, such as biostratigraphic and lithostratigraphic units. They are not the same as these and are not necessarily equivalent to chronostratigraphic units, because the ages of the bounding unconformities may change from place to place. However, unconformity-bounded units have a certain chronostratigraphic significance because, with certain unusual exceptions, all the rocks below an unconformity are older than all of those above, and time lines do not cross unconformity surfaces. (A possible exception to these rules is where a disconformity surface is caused by submarine erosion via a deep oceanic current that changes position with time as a result of the changing configuration of the ocean basin, but this is not a problem that is likely to be encountered very frequently. Ravinement surfaces are also time-transgressive.)

Formal proposals for the definition of unconformity-bounded units were given by the ISSC (1987; see also Salvador 1994) and are also contained in the NASC. The term **sequence**, as originally used by Sloss et al. (1949), is not recommended by these authorities, because this term has come to be used in a number of slightly but significantly different ways. The ISSC (1987) recommended the formal term **synthem**, a proposal followed by Salvador (1994), whereas the NASC proposed the definition of **allostratigraphic units**, including alloformation, allogroup, and allomember. Conventions for formal naming of such units are the same as those for other stratigraphic units, as set out in Section 3.3.2.

Several groups of workers are now explicitly employing an allostratigraphic methodology. For example, Autin (1992) subdivided the terraces and associated sediments in a Holocene fluvial floodplain succession into alloformations. Walker and his coworkers have employed allostratigraphic terminology in their study of the sequence stratigraphy of part of the Alberta Basin, Canada. Their first definition of unconformity-bounded units is described in Plint et al. (1986), where the defining concepts are referred to as event stratigraphy, following the developments of ideas in this area by Einsele and Seilacher (1982). Figure 3.5 is a summary of their stratigraphic synthesis. Walker (1990) discussed some of the practical problems in making use of sequence and allostratigraphic concepts. Explicit use of allostratigraphic terms appears in later papers by this group (Plint 1990). A recent text on facies analysis that builds extensively on the work of this group recommends the use of allostratigraphic methods and terminology as a general approach to the study of stratigraphic sequences (Walker 1992). Martinsen et al. (1993) compared lithostratigraphic, allostratigraphic and sequence concepts as applied to a stratigraphic succession in Wyoming. As they were able to demonstrate, each method has its local advantages and disadvantages.

3.7 Chronostratigraphy and Geochronometry

Geochronometry is the study of the continuum of geologic time. Geologic events and rock units may be fixed within this time frame by a variety of methods, of which radiometric dating is the most direct.

Chronostratigraphy is the study of the standard stratigraphic scale comprising the familiar eras, periods, and ages (e.g., Paleozoic, Triassic, Campanian). For the Phanerozoic, biostratigraphy is the main basis of the method, but radiometric methods and magnetic-reversal stratigraphy are also widely used, and chemostratigraphy is becoming increasingly important. Radiometric dating is the only chronostratigraphic tool available for the Precambrian, but it is also of considerable importance in providing a cali-

bration scale of biostratigraphic subdivisions in the Phanerozoic (Harland et al. 1964, 1990; Berggren et al. 1995b). Magnetic reversals are now of great importance for studying upper Cretaceous and Cenozoic strata but are difficult to use in older rocks owing to the difficulty of obtaining complete reversal sequences and the problems of postdepositional modification (Kennett 1980).

There has been controversy over the relative meaning of chronostratigraphic and biostratigraphic zone concepts. A biozone can only be erected and used for correlation if the fossils on which it is based are present in the rocks. However, Hedberg (1976) defined a chronozone as a zonal unit embracing all rocks formed anywhere during the range of a specified geological feature, such as a local biozone. In theory, a chronozone is present in the rocks beyond the point at which the fossil components of the biozone cease to be present as a result of lateral facies changes. Hedberg (1976) used a biostratigraphic example to illustrate the chronozone concept but, clearly, if the fossil components are not present, a chronozone cannot be recognized on biostratigraphic grounds, and its usefulness as a stratigraphic concept may be rather hypothetical (Johnson 1992). However, other means may be available to extend the chronozone, including local marker beds or magnetostratigraphic data.

Hedberg (1976) suggested that the stage be regarded as the basic working unit of chronostratigraphy because of its practical use in interregional correlation. As Hancock (1977) and Watson (1983) pointed out, Hedberg neglected to mention that all Phanerozoic stages were first defined on the basis of groups of biozones and historically, therefore, they are biostratigraphic entities. In practice, therefore, so long as biostratigraphy forms the main basis of chronostratigraphy, no useful purpose is served by treating them as theoretically different subjects. Stages may be defined by more than one system of biozones, which extends their range and reduces their facies dependence but, although this improves their chronostratigraphic usefulness, it does not change them into a different sort of unit. Where other methods, such as radiometric dating or magnetostratigraphy, are precise enough to challenge the supremacy of biostratigraphy, a case could be made for a separate set of chronostratigraphic units. For times from the upper Cretaceous to the Recent epoch, this point is now being reached, and the global time scale, based on correlations among the three main dating methods, is attaining a high degree of accuracy (Sect. 3.7.9). However, instead of a new set of chronostratigraphic units, this correlation research is being used to refine the definitions of the existing, biostratigraphically based stages. The stage is, therefore, evolving into a chronostratigraphic entity of the type visualized by Hedberg. These ideas are embodied in the second edition of the *International Stratigraphic Guide* (Salvador 1994).

However, the debate continues (Holland 1986). For example, Ludvigsen and Westrop (1985) stated that "recognition of a stage is limited by the occurrence of the component zones. Consequently, zonal and stadial successions will differ in separate regions and on different continents." They proceeded to propose three new stages for the Cambrian period based on their work in the Rocky Mountains of Canada. Cope's (1996) discussion of the need for "secondary stratigraphic standards" is an acknowledgment of the value of such additional formal units.

The following sections deal with four main subjects: the establishment of stages and higher-order units from biostratigraphic data, the problem of how to define their boundaries, the use of radiometric data, and the use of magnetic reversals as chronostratigraphic tools. Modern work in seismic stratigraphy indicates that seismic reflections have some value as chronostratigraphic markers. This is touched on later (and in Chap. 8). Finally, a brief discussion of the development of a global time scale is presented as an introduction to the research literature.

3.7.1 Chronostratigraphic and Geochronometric Scales

Chronostratigraphy refers to rocks, geochronometry to geological time. Distinction between the two tends to become blurred, because the same names are used for both. Table 3.1 lists the terms used for the hier-

Table 3.1. Hierarchy of chronostratigraphic and geochronometric terms

Chronostratigraphy	Geochronometric	Examples
Eonothem	Eon	Phanerozoic
Erathem	Era	Mesozoic
System	Period	Cretaceous
Series	Epoch	Upper (late) Cretaceous
Stage	Age	Campanian
Chronozone	Chron	*Orbitoides tissoti*

archy of subdivisions of the two measurement systems, with examples of some names in common use (based on Hedberg 1976). Schulz (1982) proposed to add the term chronosome to this list, as noted in Section 3.4.3. The term is proposed for the body of rock enclosed between two markers that may approximate isochronous surfaces. The markers may be biostratigraphic, magnetostratigraphic, or physical features such as ash beds. A chronosome is, therefore, analogous to Hedberg's (1976) chronozone except that, in the original definition of the latter, biostratigraphic criteria were emphasized. Geologists should attempt to make clear whether they are referring to the time scale or the rock record when using any of these terms. One commonly used method is to employ the qualifiers early, middle, and late when referring to geologic time, and lower, middle, and upper, when referring to the rock. These terms should be capitalized when formal subdivisions have been defined. Geologists should never use lithostratigraphic names in a time sense, for example, Catskill time, meaning the time during which the Catskill delta was deposited, because lithostratigraphic units are diachronous and have no precise time meaning.

The higher-ranking terms (periods, eras, and eons) have mostly been with us since the nineteenth century, and it is assumed that most readers are familiar with at least some of the colorful and controversial history behind their definition (e.g., Berry 1968). Many of the period names are British in origin and, it turns out, could hardly have been defined in a worse place because of gaps in the section, atypical facies, and so on. The Silurian-Devonian boundary is the classic example. The original definition of this demarcation point within a marine to nonmarine transition in the sediments of the Welsh borderlands has led to the creation, during about the last 30 years, of almost an entire geological industry or subdiscipline devoted to deciding where the Silurian-Devonian boundary should be drawn stratigraphically and which section to use as the stratotype (McLaren 1970, 1973). This particular problem will be returned to later when discussing boundaries.

Of more immediate practical importance to basin analysts is how the lower ranking units – the stages – are erected, and what their use by biostratigraphers means. This will now be discussed.

3.7.2 The Stage

The evolution of the stage concept was a confused one. In the mid-nineteenth century, D'Orbigny and Oppel were the first to use the term with essentially its present meaning (Hancock 1977).

Stages are, essentially, convenient groupings of biozones. Stage boundaries may be drawn at the top or base of a particularly well or widely developed biozone or at a prominent faunal change. Many of our modern stage names were rather loosely defined, perhaps on only a handful of taxa when originally established, but have subsequently been refined by more and more detailed studies using many different life forms. Biozones are now usually established using only the members of a single phylum (except, perhaps, Oppel zones), whereas many stages have now been defined in many different ways. For example, Devonian stages are based mainly on brachiopods, corals, trilobites, fish, conodonts, and palynomorphs; Cretaceous stages are based mainly on ammonites, pelecypods, brachiopods, foraminifera, nannofossils, and palynomorphs.

Most fossil groups tends to be facies bound, and the key to successful stage correlation is to locate sections where more than one useful fossil group is present or where repeated facies variations cause ecologically incompatible taxa to occur close together by interdigitation. This is largely a matter of chance. Much depends on lucky exposure of the right section, and the literature is replete with obscure geographic localities that have attained a specialized kind of fame because of the excellence of the biostratigraphic work carried out there. Examples are such small Welsh towns as Llanvirn and Llandeilo (Ordovician), the Eifel district, Belgium (Devonian), and the Barrandian area of the Czech Republic (Silurian-Devonian boundary). Geologists worldwide knew about Maastricht long before the average European politician had heard about this minor city. Most stages are named after such places. The end result of this extended effort is that many stages can now be recognized globally, with varying degrees of confidence. Plate-tectonic movements cause faunal provincialism to vary in many taxonomic groups simultaneously, so the perfection of this correlation varies from place to place and time to time. For example, Berry (1977) reported that correlation of Early and middle Ordovician graptolitic faunas between Europe and North America has been fraught with controversy because of faunal provincialism. At that time, the proto-Atlantic Ocean (Iapetus) was at its widest.

As an example of modern work in establishing stages by means of detailed faunal work on several animal groups, a brief discussion is presented here of the Pridolian and Lochkovian stages, as defined in the Barrandian area of the Czech Republic. Chlupác (1972) published a detailed description of the faunas and revised the earlier biostratigraphic subdivisions of these rocks. His faunal list for the two stages includes over 300 species, of which graptolites, conodonts, and trilobites constitute the most important

biostratigraphic indicators. Other groups providing subsidiary control include eurypterids, phyllocarids, ostracodes, echinoids, cephalopods, gastropods, pelecypods, and brachiopods. Figure 3.31 illustrates one of several short but critical sections measured through the Pridolian-Lochkovian boundary near Klonk. Limestone beds are numbered 1 to 53. This section has now been established as the stratotype of the Silurian-Devonian boundary and, therefore, also serves as a good example of a chronostratigraphic boundary, as discussed in the next section.

The Pridolian-Lochkovian succession in this area consists mainly of thinly interbedded, grayish-black, calcareous mudstone and grayish black to dark gray, fine-grained (micritic), skeletal, platy weathering limestone, with subordinate beds of pale limestone and coarser, detrital limestone (calcarenite to calcirudite). Graptolites are abundant in the mudstones; trilobites and conodonts occur sparsely in the limestones, particularly in the paler, less muddy units, and become more abundant north of Klonk, where the rocks undergo a facies change into a predominantly pure-carbonate succession. This interbedding of different facies with their contrasting faunas is one of the most important features of these central European sections from the point of view of biostratigraphic stratotype definition.

Two graptolite biozones have been recognized in the Pridolian stage and form the main basis for the definition of the unit. They are the *Monograptus ultimus* zone below and the *M. transgrediens* zone above. The latter does not reach to the top of the Pridolian strata, but is followed by a graptolite interregnum containing only sparse, nondiagnostic forms (Fig. 3.31). The base of the Lochkovian is defined by the sudden widespread appearance and abundance of *M. uniformis* in bed 20. Other species of *Monograptus* and *Linograptus* appear in the upper part of the lower Lochkovian strata, while in the upper Lochkovian, *M. hercynicus* is typical. Some of these species occur in the pure carbonate facies, permitting close correlation with the shelly fauna. It is interesting to note that, in spite of the effort biostratigraphers have made to formalize their biozone types, Chlupác's work is typical of many in that no attempt is made to state what kind of biozone is in use. The older of these graptolite biozones appear to be single-taxon range biozones, with concurrent-range biozones for the upper part of the lower Lochkovian and the upper Lochkovian.

The trilobite *Warburgella (Podolites) rugulosa rugosa* is of primary importance in delineating the lower boundary of the Lochkovian. In the Klonk section, it appears in limestone bed 21, immediately above the first appearance of *M. uniformis* in the upper part of bed 20 (Fig. 3.31).

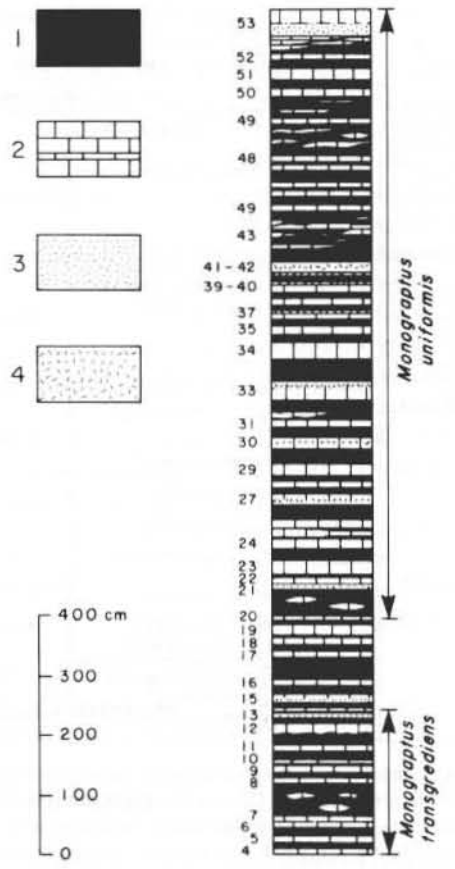


Fig. 3.31. The Silurian-Devonian section at Klonk, Czech Republic, used as the boundary stratotype for the Pridolian-Lochkovian stages and of the Silurian-Devonian. Limestone beds have been assigned numbers for reference. Ranges of two key graptolite species are shown at *right*: *1* grayish-black calcareous shales; *2* grayish-black and dark gray, fine-grained, platy limestones (micritic, skeletal); *3* paler gray, granular, fine-grained limestones (skeletal, micritic); *4* coarse, detrital, crinoidal limestones (skeletal, micritic, with biorudite components). (Chlupác 1972)

The conodont *Icriodus woschmidti* defines a range biozone corresponding approximately with the lower part of the Lochkovian, although in the Barrandian area it ranges down through the graptolite interregnum into the top of the *M. transgrediens* biozone. Conodonts are not common in the somewhat argillaceous facies at Klonk.

This discussion could be extended considerably into a consideration of other faunal groups and some of the subsidiary species that define concurrent-range zones. It is to be hoped, however, that by this time the reader can discern the main threads of a procedure that has now been followed innumerable times by many different workers.

Barnes et al. (1976) discussed the correlation between graptolites, conodonts, trilobites, and brachiopods for the Ordovician rocks of Canada. This is an



Fig. 3.32. An attempt to provide a chronostratigraphic framework for the Ordovician of Canada by relating various biozone schemes to each other. Note the use of two stage nomenclatures, one erected for North American cratonic sediments, the other the classic European, mainly graptolite-based nomenclature. (Reproduced, with permission, from Barnes et al. 1976)

interesting case in that the benthonic fauna was used to establish a North American stage nomenclature in the craton (the Richmond, Maysville, etc., in Fig. 3.32), whereas graptolites occur mainly in deeper water deposits and were correlated with the classic British stages (Ashgill, Caradoc, etc.) in spite of the difficulties of faunal provincialism across the Iapetus Ocean referred to earlier. To be of any regional use, these biostratigraphic schemes had to be integrated with each other, and this depended on finding locations, as at Klonk, where the different biofacies are interbedded. In order to cover the entire Ordovician System, it was necessary to study partial sections in the Canadian Rocky Mountains, Nevada, Texas, Newfoundland, the St. Lawrence Lowlands, and parts of the Canadian Arctic Islands. One of the key sections was the lower to middle Ordovician Cow Head Group of Newfoundland. Here, trilobites and conodonts occur in limestone boulders slumped from the shelf into the deep water basin, where graptolite-bearing shales buried them. Integration of the biozone schemes had to allow for the fact that the boulders were probably slightly younger than the enclosing shales. Elsewhere, transgressions and regressions caused intermingling of faunas from different facies and different faunal provinces.

3.7.3 Chronostratigraphic Boundaries

It would seem to be a logical assumption that the best place to draw a boundary between stages and higher-ranging chronostratigraphic units would be at a hiatus where the fauna changes abruptly. Many units have, in the past, been so defined, using major unconformities or even lithostratigraphic changes for boundary demarcation. Others have been defined on the basis of the extinction of particular faunal groups, causing great consternation when these fossils were later found in much younger rocks. McLaren (1970) discussed the Frasnian-Famennian boundary within the Late Devonian. Although this boundary has not been well defined worldwide, it seems that there was wholesale replacement of faunas across it, including the disappearance of tentaculitids, the near extinction of stromatoporoids, and the complete extinction of many coral, brachiopod, and trilobite groups. Other well-known mass extinctions occurred at the ends of the Permian and Cretaceous periods. These are obviously significant events and good places to draw boundaries, so long as they are real and not the product of biostratigraphic generalization from inadequate data. There has been much literature

discussion of these extinctions and of possible catastrophic causes, such as meteorite impact or supernovae (Alvarez et al. 1980; Silver and Schultz 1982; Clifton 1988).

There are few dramatic boundaries of this type, however, and most stratigraphers now agree that chronostratigraphic boundaries should be "defined whenever possible in an area where 'nothing happened'" (McLaren 1970). Hedberg (1976) stated that the boundary stratotypes of a stage should be within sequences of continuous deposition – preferably marine – and both should be associated with distinct marker horizons, such as biozone boundaries, that can be readily recognized and widely traced as isochronous horizons. The problem with using a hiatus as a chronostratigraphic boundary is that a hiatus represents a break in time that is bound to be represented by sediments and their contained fossils somewhere else. When these sediments are eventually discovered, how are they to be classified chronostratigraphically?

Even an apparently continuous sequence might later be discovered to contain a time break. For this reason, there is now general (but not universal) agreement that boundary stratotypes should serve as the base of the unit above, rather than the top of the unit below (Carter 1974). If it is subsequently realized that the boundary stratotype is a hiatus, sediments spanning the time break are automatically assigned to the older unit. Therefore, no revision of the stratotype itself is necessary. This is now known colloquially as the concept of the **topless stage**, an easy term to commit to memory.

Once international agreement has been reached, a boundary stratotype should not need to be revised, because the precise position of the boundary is less important than that there is agreement on where it happens to be. All possible biostratigraphic means may be used to trace the boundary into different regions. To physically fix the boundary for future study, the practice of emplacing a permanent marker in the stratotype outcrop has recently been started (Ager 1964). These have come to be termed **golden spikes**, although some other metal, causing less of a drain on research funds (and less prone to theft), is used in practice.

The first golden spike was inserted in bed 20 of the section at Klonk (Fig. 3.31). By international agreement, the first appearance of *M. uniformis uniformis* was chosen as the base of the Lochkovian stage and as the base of the Devonian System (McLaren 1973).

It is particularly apt that the first golden spike should have been inserted at such a place as Klonk. This happy accident of onomatopoeia celebrates two events – the establishment of a better method for resolving boundary problems, and the voluntary euthanasia of the Silurian-Devonian boundary industry.

As noted in several recent reviews (Carter 1974; Bassett 1985 and other papers in the June 1985 issue of *Episodes*; Holland 1986, 1998), much of Phanerozoic stratigraphy is now being revised and formalized using the boundary stratotype and topless-stage concepts. Virtually all this work is heavily based on biostratigraphy, although radiometric age data are now providing invaluable supplementary information in the form of an approximate numerical time scale in years. For the Cenozoic and perhaps the younger Mesozoic, magnetostratigraphy has also assumed great importance. Only in these cases are chronostratigraphic units in any sense conceptually different from biostratigraphic units. Some of the recent work in this area is discussed in Section 3.7.9.

Problems remain in applying the topless-stage approach to well cuttings. As discussed in Sections 1.4.1 and 2.3.1 the nature of the drilling process is such that "tops" are the only reliable kind of stratigraphic data.

3.7.4 Radiometric Dating

3.7.4.1 *Radiometric Methods*

Several naturally occurring isotopes of common elements are radioactive and, by a process of radioactive decay, produce one or a series of daughter products. The decay process ceases when a stable isotope is produced. The rate of decay for each radioactive series is constant – it is conventionally expressed by the **half-life** of the radioactive isotope. This is the length of time required for half the atoms in a given quantity of the element to decay. Half-lives range from fractions of a second to hundreds of millions of years.

Minerals containing the radioactive isotopes are formed by crystallization in an igneous melt or by authigenesis in a sediment. At this point, their isotopic composition is frozen in, so that any subsequent changes are, in principle, the result only of radioactive decay through time. By measuring the relative proportions of the radioactive isotope remaining and of new daughter products produced, it is possible to calculate the age of the crystal. In practice, there are many problems that have to be resolved, including the possibility of contamination by weathering, outward diffusion and escape of daughter products (yielding ages that are too young), and resetting of the clock by later thermal or tectonic events. Folding or fracturing of the rock, or metasomatism, can cause such changes, so the measurements are best carried out in cratonic areas or on fold belts that have not been subsequently disturbed. Igneous minerals are set when

they cool past their **Curie point**, or closure temperature, which is typically a temperature of a few hundred degrees. It ranges from about 130 °C for microcline to more than 800 °C for zircon. For large plutonic bodies, it may be a considerable period of time after the intrusion and initial crystallization before the minerals all pass their Curie point. Very deep burial and regional or contact thermal metamorphism can change the apparent age. Radiometric ages commonly record the time of primary cooling, a subsequent homogenizing event, or an intermediate date between the two.

In most cases, the amount of isotope material available for analysis is extremely small, and laboratory methods permit a reproducibility of only about 3% at best (Lambert 1964; Harland et al. 1990). McDougall (1978) reported that a 3% error between the results of different laboratories working on the same material is not uncommon. A more fundamental source of error is disagreement about the decay constants of the various radioactive isotopes. Armstrong (1978) showed that calculated ages may change by at least 3%, depending on the decay constant used. Steiger and Jager (1978) reported on an attempt to standardize these values. Because of these sources of error, the decay constant used and the calculated experimental error should always be quoted and should be borne in mind when using a radiometric age in stratigraphic work. A 3% error is equal to ±1.5 m.y. at 50 Ma or 15 m.y. at 500 Ma. These are significant quantities, as we shall see in a later section when we discuss the global time scale.

Two main radiometric methods are used in sediments, the potassium–argon (K–Ar) method, using the isotopes ^{40}K and ^{40}Ar, and the rubidium–strontium (Rb–Sr) method, using the isotopes ^{87}Rb and ^{87}Sr. Early stratigraphic work in this field is described in Harland et al. (1964) and Cohee et al. (1978), and more recent developments are discussed by Harland et al. (1990). Harland et al. (1990), Parrish and Roddick (1985), and Hole (1998) provide up-to-date reviews. ^{14}C methods are used for sediments back to about 70,000 years ago. For dating the primary crystallization of orogenic materials, uranium–lead methods are preferable and are widely used in studies of the Precambrian (Stockwell 1973, 1982). Modern methods of dating Precambrian rocks may yield zircon ages that are reproducible to within a few tens of millions of years or less, which has provided tight constraints for the reconstruction of Proterozoic plate-tectonic evolution of shield areas (Hoffman 1989). The K–Ar method is somewhat better for younger rocks up to less than 1 Ma, and the Rb–Sr method for rocks older than about 10 Ma, but both are widely used on rocks from throughout the Precambrian and Phanerozoic strata. Both may be used either for whole-rock analyses or for selected minerals, particularly biotite, muscovite, amphibole, pyroxene, chlorite, feldspars, and zircon, plus a few sedimentary minerals, as discussed later. Carrying out individual mineral determinations and whole-rock analyses provides a useful crosscheck of the results. For example, because of the varying resistance to ^{87}Sr diffusion, Moorbath (1964) reported the following empirical age relationships for typical specimens of granitic igneous rocks: whole-rock age is greater than microcline age, which is greater than muscovite age, which is greater than biotite age. Radiometric dates used in modern geological time scales are primarily K–Ar dates derived from the study of volcanic rocks (Harland et al. 1990).

A somewhat newer technique is **fission-track dating**. The isotope ^{238}U, which occurs in zircon, sphene, apatite, and volcanic glass, emits α particles that produce crystal damage. When polished surfaces of the crystals are etched, the damage shows up as short tracks, the density of which can be counted to determine the age of the rock. The principles are discussed in Fleischer et al. (1974), and Parrish and Roddick (1985) have provided a useful short review.

An exciting new tool, a variation on K–Ar dating, is the use of a laser microprobe to perform $^{40}Ar/^{39}Ar$ dating on diagenetic clays (Dong et al. 1995; Clauer and Chaudhuri 1996; Hassanipak and Wampler 1996). This method is also beginning to be used to improve the dating of glauconites, which are among the few widely occurring datable authigenic materials in the sedimentary record (Smith et al. 1993, 1998). Earlier work on this suite of minerals had long been suspect because of argon loss during irradiation, which yielded ages too young. New laboratory procedures are eliminating this problem by dating single crystals. As pointed out by Smith et al. (1998), this technique offers, for the first time, the potential for reliable direct radiometric dating of parts of the sedimentary record that also contain fossils, providing the potential for cross-correlation and mutual error reduction. The introduction of the new Sensitive High-Resolution Ion MicroProbe (SHRIMP) technique is also likely to revolutionize the accuracy and precision of radiometric dating. Application to Phanerozoic problems has only just begun (Claqué-Long et al. 1995).

3.7.4.2 *Uses in Stratigraphy*

Sedimentary rocks can be dated radiometrically in three different ways:

1. Igneous intrusions within the sediments can be dated. Although widely used, this method can never provide very accurate ages because of uncer-

tainties about the stratigraphic relationship between the igneous rocks and the sediments. The intrusions are, obviously, younger than the rocks they intrude, but this provides only a minimum age for the sediments. If truncated at an unconformity, the igneous rocks provide a maximum age for any sediments that follow. Many Precambrian successions have been dated by bracketing relationships of this type.

2. Volcanic flows and pyroclastic beds, including ignimbrites, tuffs, agglomerates, bentonites, and volcaniclastic rocks interbedded with the sediments can be dated. Most volcanic materials provide several of the most readily dated mineral types listed previously. Interbedded volcanics provide much more precise stratigraphic information than do intrusive rocks because it can be safely assumed that their indicated age is very nearly the same as that of the enclosing sediments. The method is widely used, as indicated by a few examples discussed later. However, many successions lack any evidence of volcanism.
3. Authigenic sedimentary minerals can be used. Glauconite is by far the most widely used mineral, although illite, sylvite, bone, limestone, and fluorite (in fact, any rubidium- or potassium-bearing mineral) may be used within the limits of analytical accuracy when dealing with very minute quantities. Numerous glauconite dates have now been published, and work is continuing to improve the measurement technique (Berggren et al. 1978; Odin 1978; Smith et al. 1993, 1998). The mineral is prone to argon loss with mild heating or overburden pressure, so glauconite dates are most reliable for the Cenozoic era and, with care, the younger Mesozoic, particularly for cratonic sequences or others that have not been deeply buried (newer techniques are eliminating this problem, as noted above). Odin (1978) claimed that even Precambrian sediments preserved under these conditions may yield reliable K–Ar dates on glauconite. Glauconite is rare as a detrital mineral, but this should always be checked lest spurious results are obtained. Both Rb–Sr and K–Ar methods may be used, but the latter is preferred. Glauconite is widely distributed in shallow-marine deposits, which are also facies in which many of the best biostratigraphic indicators occur, so glauconite dates have been widely used to provide absolute ages of biozones and stages.

It should be obvious from the above that precise radiometric dating of sediments can only be carried out indirectly by the study of diagenetic phases (a technique still in its infancy), by the use of glauconites in certain marine facies, or where volcanic rocks are present. This includes only a small proportion of the total volume of sediments. It is, therefore, unlikely that radiometric methods will ever replace biostratigraphy for studies of Phanerozoic sediments. However, they are proving of inestimable value in placing ages (in years) on the biostratigraphic time scale.

An interesting application of isotopic decay studies is to plot cooling histories of orogenic belts in order to study uplift rates. These data can then be used to investigate unroofing histories and the yield of detrital particles that contribute to the development of clastic wedges in the adjacent basins. Fission tracks anneal and disappear at closure temperatures that are ideally suited to studies of shallow burial. For volcanic glass, the temperature is less than 90°C, ranging up to about 290°C for sphene (Parrish and Roddick 1985). Parrish (1983) showed that, in the Coast Mountains of British Columbia, fission track ages increase with topographic elevation. The interpretation is that where uplift rates are highest, as in the areas of the highest mountains, the rocks now exposed at the surface were uplifted through their closure temperatures the soonest, yielding the oldest fission track ages. Cerveny et al. (1988) reported on a detailed study of the rapid uplift rates of the Himalayan Mountains in Pakistan.

Efforts to quantify geological time are as old as geology itself. Early attempts are described in several papers in Harland et al. (1964), which is the proceedings volume of a major symposium held to bring together modern work in the field. The symposium was held in honor of Arthur Holmes, the main pioneer in the use of radiometric methods in studying the geological time scale. At the time of this symposium, geologists were fairly confidently assigning ages to series or even stages, although these values have since been revised, and the literature is full of minor disagreements and inconsistencies that have yet to be eliminated. This has not prevented geologists from making elaborate correlations, calculations of sedimentation rates, and so on based on radiometric dates without due regard for their experimental error. Jeletzky (1978) and others have argued forcefully against this practice.

However, radiometric dates are not used in isolation from each other or from other types of data. A sequence of dates from several ash or bentonite beds interbedded with sediments can provide better information through averaging and by checking the dates against the thickness of bracketed sedimentary rocks, assuming constant sedimentation rates. An example of this approach is illustrated in Fig. 3.14. Kauffman (1977) obtained dates from most of ten bentonite beds occurring over 12 m of section, and tied the dates to 34 pelecypod collections. Checks of internal consistency of this type give the procedure a much

greater degree of reliability. The use of magnetostratigraphy as an additional aid is discussed in Section 3.7.5. None of these additional aids is available for the Precambrian, and chronostratigraphy for this major part of Earth's history remains relatively crude (Sect. 3.7.9.2).

3.7.5 Magnetostratigraphy

3.7.5.1 *Development of the Technique*

The study of the magnetic properties of rocks is one of the most recent additions to the range of tools available to stratigraphers for correlating sedimentary successions. The first major development, the erection of a satisfactory geomagnetic polarity scale for the last 4.5 Ma, was published by Cox (1969), and in the 30 years since then magnetostratigraphy has become almost equal in importance to biostratigraphy and radiometric dating in establishing a time scale for the last 160 Ma of Earth's history (Fig. 3.41). However, it seems at present unlikely that the method will be of more than local use for pre-upper Jurassic rocks.

The principle of magnetic stratigraphy is that

> during deposition of sediments or cooling of molten rock, magnetic iron oxide minerals align themselves with any magnetic field that then exists. This preferred alignment of grains results is a bulk property of the rock referred to as **remanent magnetism**, which in essence acts as a geomagnetic field recorder. Initially it was observed in sequences of volcanic rocks on land that the north and south poles of the earth's magnetic field have reversed fairly frequently. For instance our present polarity began 7×10^5 years ago, and before that for about one million years, the poles were reversed except for several short episodes of normal polarity. Magnetic stratigraphy, or magnetostratigraphy, is based on the reversals in the earth's magnetic field recorded in sediments, and the utilization of these changes for stratigraphic and chronological purposes (Kennett 1980).

Magnetic properties have also been used in two other ways that are not our immediate concern in this book: (1) for reconstructing ancient pole positions in order to track plate trajectories across the globe relative to an assumed stationary geomagnetic axis and (2) for analyzing sea-floor spreading patterns by using the polarity reversal pattern as it is imprinted on newly formed oceanic lavas. The advantage of the polarity-reversal sequence to stratigraphers is that the reversals are synchronous, worldwide events. The disadvantage is that there is little to distinguish one polarity episode from another except its duration. This makes it difficult to correlate the reversals to each other, especially as time can be distorted in sediments by variations in sedimentation rate and the presence of missing sections (Johnson and McGee 1983; Hall and Butler 1983; Sect. 3.7.7). To resolve this problem, magnetostratigraphy has become closely tied to biostratigraphy and radiometric dating for geochronological purposes.

The work on the reversal chronology has followed two major trends. Vine and Matthews (1963) and Morley and Larochelle (1964) discovered that linear magnetic anomalies on the sea floor could be correlated across midocean ridges, showing that they recorded the pattern of growth of oceanic crust. Heirtzler et al. (1968) developed a polarity time scale from these data for the last 79 Ma by plotting the width of the anomalies and assuming constant sea-floor spreading rates. Later, more accurate work on deep-sea drilling cores showed that this assumption was incorrect (Odin and Curry 1985), and the polarity scale has required revision.

The other approach to reversal chronology has been to carry out detailed sampling and reversal studies in volcanic and sedimentary rocks. Initially, the reversals were dated using radiometric methods. Cox (1969) published a revised scale for the last 4.5 Ma based mainly on measurements of volcanic lavas (Fig. 3.33). This work is still, for the most part, generally accepted (Opdyke 1972). Named polarity events in this table are based on various outcrop localities. The first use of deep-sea cores predated the Deep-Sea Drilling Project (DSDP), and the cores were very short (see review by Kennett 1980). Harrison and Funnel (1964) were the first to combine biostratigraphy and chronostratigraphy in an attempt to correlate a reversal event. Later, Opdyke et al. (1966) studied some longer cores, up to 12 m in length, and were able to correlate reversal events with the land-based lava sequence. Radiolarian zones closely parallel the reversal correlations (Fig. 3.34). The DSDP started in 1968 and began to have an important effect on chronostratigraphy. However, difficulties were encountered in establishing magnetic stratigraphy directly from the cores because of drilling disturbance and bioturbation. The practice developed of calibrating polarity and biostratigraphic data by correlating deep marine with exposed on-land sections and dating the latter radiometrically. This work has not been without its problems because of the variations in biofacies between various oceanic areas caused by water-mass differences and variations between oceanic and on-land sections. The latter commonly contain a shallower water fauna. Extension of a reliable magnetic chronology to older Cenozoic and Cretaceous se-

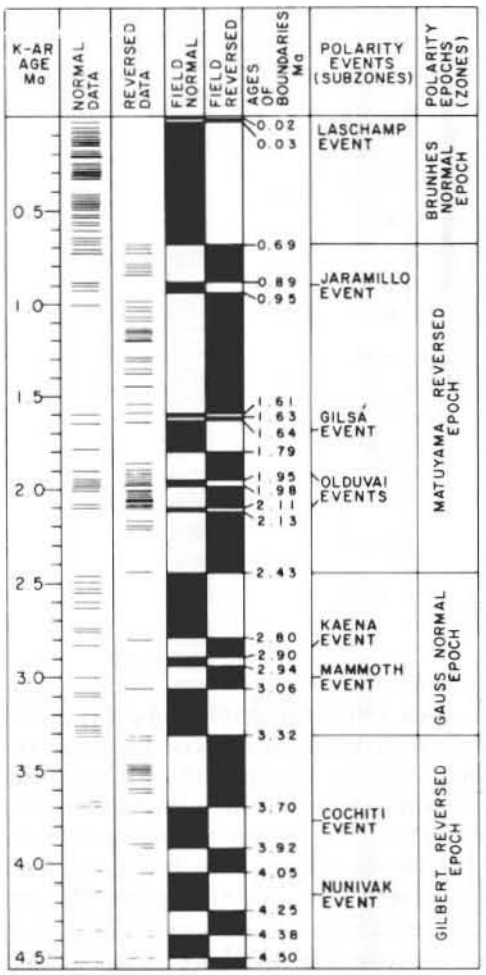


Fig. 3.33. The sequence of magnetic reversals for the last 4.5 Ma, based on studies of radiometrically dated lavas (Cox 1969). The ages of some boundaries and the occurrence of some of the reversal subzones have been revised in recent years. (Watkins 1976 and text)

quences has depended on the study of a few well-preserved sections in Italy, New Zealand, and a few other areas (Ryan et al. 1974; Sclater et al. 1974; Lowrie and Alvarez 1981; Harland et al. 1990). Modern time scales are referenced in Section 3.7.9.

In 1972, the Subcommission on a Magnetic Polarity Time Scale (SMPTS) was established as part of the International Commission on Stratigraphy. The deliberations of this group (Subcommission on a Magnetic Polarity Time Scale 1973), plus those of the Subcommission on Stratigraphic Classification, have resulted in a formal nomenclature for magnetostratigraphy. The new hierarchy of reversal units and their approximate duration are given in Table 3.2.

3.7.5.2 *Uses in Stratigraphy*

Magnetostratigraphy can be used by basin analysts in two main ways. First, it may provide part of the data used to correlate local sections with the global chronostratigraphic standard. Second, it may be used to correlate local sections with each other without regard to the global scale.

The first type of use requires detailed knowledge of the chronostratigraphic database. This is now excellent for the Cenozoic but decreases in quality rapidly back into the Mesozoic. The work of SMPTS extended the reversal scale, with numbered anomalies and **chrons**, as far back as the late Cretaceous period (Watkins 1976). Compilations for earlier in the Cretaceous and Jurassic periods are available but are based on combining scattered marine and on-land sections. The oldest undisturbed oceanic crust yet described is middle Jurassic (Larson and Pitman 1972; Gradstein 1981). Continuous anomaly sequences and virtually hiatus-free sedimentary cover are available from the oceans back to this time period, but for earlier rocks it is necessary to turn to incomplete on-land sequences. No reliable reversal sequence is, therefore, available for pre-middle Jurassic times, and it is likely to require many years of work before one becomes available.

For the purpose of local stratigraphic correlation, no knowledge of the global scale is required. However, because reversal events are not unique, it is not possible to correlate them by matching sequences from different stratigraphic sections unless they contain particularly distinctive long or short polarity intervals. Some supplementary criteria may be required to assist in matching, such as marker beds or biostratigraphic zonation. Picard (1964) and Irving (1966) were among the first Western workers to use paleomagnetic correlation for sediments on the continents. The technique has become widely used for nonmarine sediments because of the scarcity

Table 3.2. Nomenclature of magnetostratigraphy

Magnetostratigraphic units	Old chronologic terms (abandoned)	New geochronologic units	Approximate durations (years)
Polarity subzone	Polarity event	Subchron	10^4–10^5
Polarity zone	Polarity epoch	Chron	10^5–10^6
Polarity superzone	Polarity period	Superchron	10^6–10^7

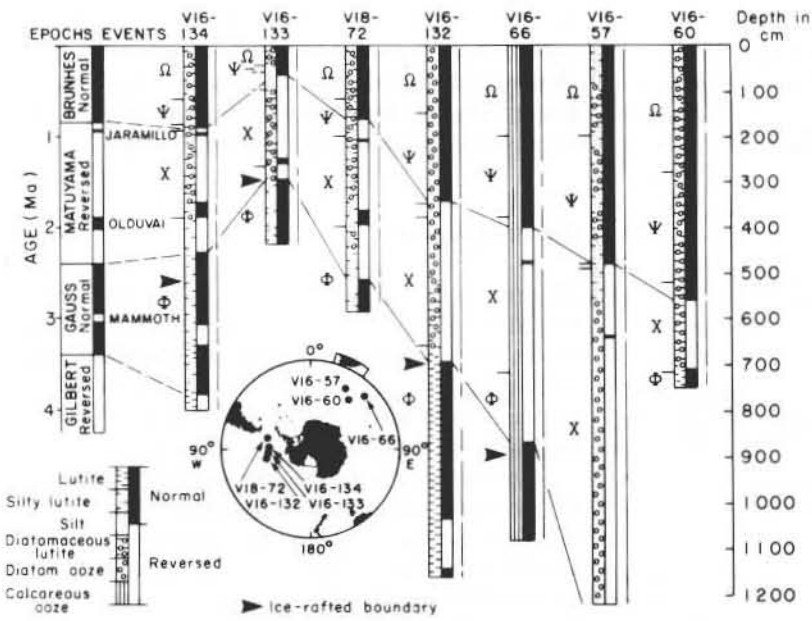


Fig. 3.34. Correlation of magnetic stratigraphy and radiolarian zones in seven cores from the Antarctic. (Opdyke et al. 1966)

of other means of precise correlation. However, problems of postdepositional modification of the magnetic remanence have introduced complications (Turner 1980).

A single example of the use of magnetostratigraphy in a practical field problem is described briefly here. For many years, a team has been exploring the nonmarine Siwalik Group (Oligocene-Quaternary) of Pakistan, in part because of its rich vertebrate fauna and in part because of the information the sediments yield about the tectonics of the Himalayas, from which the sediments were derived. Magnetostratigraphy, coupled with radiometric dating of several ash beds, has provided a useful means of local correlation between sections that show marked lateral facies changes (Keller et al. 1977; Barndt et al. 1978; Johnson et al. 1979). It was also possible to propose a correlation with the global scale. Fig. 3.35 illustrates the correlation of three closely spaced sections in the Pabbi Hills area, near Jhelum. From three to five oriented rock specimens were collected from each of 113 sites within the sections. These were subjected to laboratory tests to determine the stability of the field and the absence of magnetic overprinting, and the pole positions obtained were corrected for structural dip. The reversal zones were correlated with the standard scale of Opdyke (1972) using the following argument. The oldest remains of *Equus* (horse) were found at the 400-m level in the composite section (Fig. 3.35). The oldest occurrence of *Equus* in North America is dated 3.5 Ma and in Asia 2.5 Ma. It is considered unlikely that the Pabbi Hills fossils are older than 3.5 Ma. Note that the fossil locality is in a short normal-polarity sequence within a long reversed-polarity interval. Only two such dominantly reversed intervals are present in the magnetostratigraphic scale, the Matuyama and the Gilbert zones. The Gilbert zone extended from 5.1 to 3.3 Ma, and the Matuyama from 2.41 to 0.70 Ma. (Revised ages of Opdyke 1972; modified from Cox 1969; see his chart in Fig. 3.33 for comparison). The evidence of *Equus* suggests that this is the Matuyama zone. The two short normal events then correlate with the Olduvai and Jaramillo subzones. The Gilsa event (subzone) of Fig. 3.33 is not universally recognized and does not appear on Opdyke's (1972) chart.

Four composite sections have been correlated using the presence of two tuff horizons (Visser and Johnson 1978) and the polarity zones, as shown in Fig. 3.36. The importance of this work is that it permits precise local and global correlation of vertebrate localities and accurate calculations of sedimentation rates and provides accurate control for studying sedimentological characteristics, basin architecture, and tectonic events. Some of these aspects have been explored in later papers by this research group and other workers in the area. For example, Johnson et al. (1985) were able to determine the various time scales represented by fluvial cycles and to explore the implications for rates of channel wandering and the nature of tectonic and climatic controls on sedimentation. Behrensmeyer (1987) developed detailed two-dimensional reconstructions of the stratigraphic architecture as a basis for an examination of the taphonomy of the vertebrate remains.

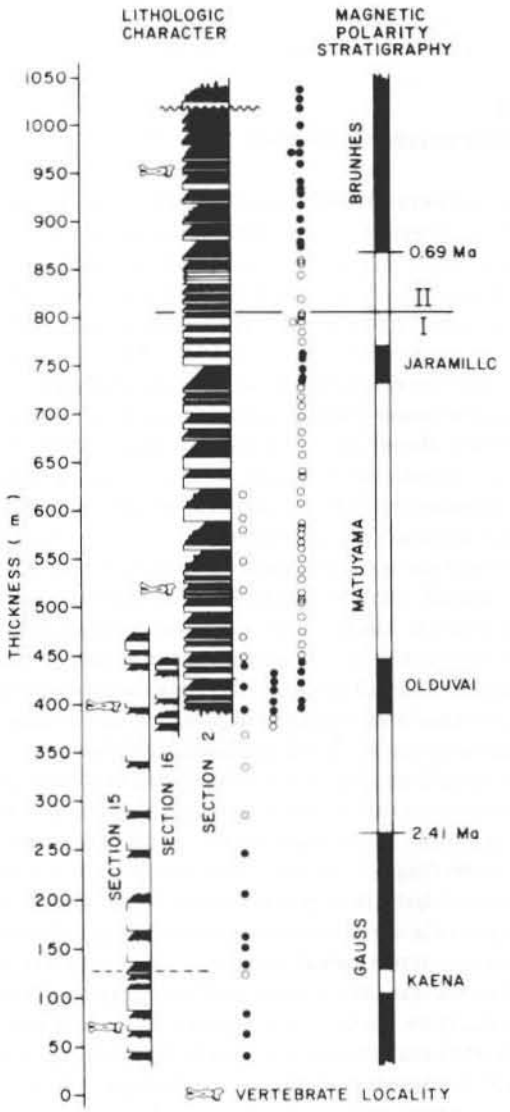


Fig. 3.35. Schematic lithology and magnetic-reversal stratigraphy, composite Siwalik section, Pabbi Hills area, Pakistan. Most of the section consists of fluvial fining-upward cycles. Correlation of the reversals with the standard sequence of Cox and others is discussed in the text. (Johnson et al. 1979)

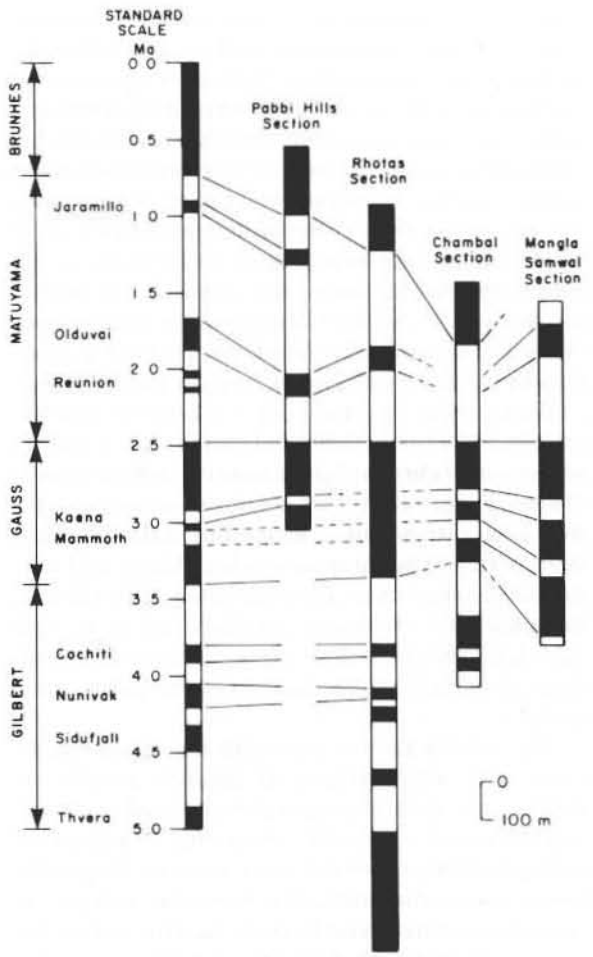


Fig. 3.36. Correlation of local reversal sequences in four composite Siwalik sections, with the standard sequence at *left*. Two tuff horizons were also used for correlation. (Johnson et al. 1979)

3.7.6 Chemostratigraphic-Dating Techniques

Variations in the oxygen- and strontium-isotope composition of sea-water provide the basis for two important additional dating methods. Oxygen-isotope stratigraphy has become an essential technique for dating the late Cenozoic stratigraphic record. The method depends on measurements of the $^{16}O/^{18}O$ ratio. Because ^{16}O is the lighter of the two isotopes, water molecules containing this light oxygen are preferentially evaporated from seawater. During times of ice-free global climate they are recycled to the oceans, and the isotopic ratio remains in a stable balance corresponding to the natural proportions of the two isotopes in the hydrosphere. However, glacial ice is composed of condensed ^{16}O-enriched water, so that continental ice buildups are preferentially enriched in ^{16}O while the oceans are depleted in this isotope, with the result that the ^{18}O content of the oceans is increased. Sediments preserve the isotopic ratios of the oxygen that existed in the hydrosphere at the time the sediments were formed. Emiliani (1955) was the first to demonstrate this oxygen cycle, and numerous studies since his have confirmed the reality of the mechanism. We now know that the $^{16}O/^{18}O$ ratio is a sensitive indicator of global ocean temperatures and can, therefore, be used as an analog recorder of ice volumes (Hays et al. 1976; Shackleton and Opdyke 1976; Matthews 1984, 1988). The measurements are made on the carbonate in foraminiferal tests. Matthews (1984) suggested a calibration value of ^{18}O

variation of about 0.011‰ per meter of sea-level change. Precise calibrations are impossible because of uncertainties regarding the volume and content of floating pack ice and diagenesis of foraminiferal tests. It is also important to take into account the lag between the maximum eustatic high and the highest surface ocean temperatures, which occur some 10^3 years later. The ^{18}O record provides an excellent time scale, because fluctuations in ice volumes are caused by climate changes that are driven by **orbital forcing** – the so-called **Milankovitch mechanisms** (Imbrie 1985), which are now known to be characterized by a precise cyclicity, at least for the latter part of the Cenozoic era (Chap. 8). A formal ^{18}O chronology has now been established, consisting of a series of numbered **chrons** (Martinson et al. 1987, Williams 1988; Williams et al. 1988), and many studies nave now been carried out in which this chronology has been used in stratigraphic studies (Kamp and Turner 1990; Feeley et al. 1990; see Miall 1997, Chap. 8, for additional examples). As discussed in Section 3.7.9.1, the ^{18}O chronology is now forming an important component of the Late Cenozoic global time scale.

The $^{87}Sr/^{86}Sr$ ratio in sea-water has varied significantly with time, providing the basis for another excellent time scale. Measurements are made on fossil shell material, especially belemnites, brachiopods, and conodonts, which are more resistant than other fossil types to diagenetic alteration. The isotopic ratio has varied in a crudely cyclic manner within the range 0.7070 to 0.7090 during the last 450 m.y., so that some intermediate values correspond to several different ages on the curve. It is necessary, therefore, to determine an approximate age for the sample first, using other methods. Modern compilations of the data, and discussions of error and precision, were provided by Smalley et al. (1994) and McArthur (1998). An example of the application of strontium-isotope dating to a problem in sequence stratigraphy is given by Sugarman et al. (1993). Smalley et al. (1994) claimed that the dating precision is under 2 m.y. for parts of the Mesozoic and Cenozoic eras.

Carbon-isotope stratigraphy is another chemostratigraphic technique that is gaining increasing attention (Jenkyns 1995; Vahrenkamp 1996). Grötsch et al. (1998) demonstrated excellent correlations between the ^{13}C record in successions of Cretaceous platform limestones and basinal shales in Italy, and also between these sections and a basinal succession in Abu Dhabi. They showed that diagenesis may have increased the spread of readings from the carbonate section but did not substantially modify the overall record. They argued that shifts in ^{13}C content are related to the rate of absorption of the lighter ^{12}C isotope in marine deposits, particularly during the formation of black shales (Sect. 4.6.9), and suggested that this occurs at times of rising sea level.

3.7.7 Other Correlation Techniques

Improvements in seismic techniques in the 1970s led to the suggestion that seismic reflectors are chronostratigraphic markers because they are generated at lithologic discontinuities that reflect widespread environmental change or stratigraphic events, such as turbidite deposits (Vail et al. 1977). This formed the basis for the revolution in seismic stratigraphy, from which the study of sequence stratigraphy has evolved. However, there are exceptions to this general rule, where rapid local facies changes can generate diachronous reflections and diachronous disconformities, such as ravinement surfaces (Sect. 6.3). It has also been shown from modern, high-resolution seismic records that, in detail, many apparently simple stratigraphic relationships are in fact complex, reflecting numerous local facies changes, and that in many cases it is difficult to carry correlations through such successions (Cartwright et al. 1993). These points are discussed in more detail by Miall (1997; Sect. 2.2).

Detrital composition of sandstones may have local chronostratigraphic value. Variations in composition through time, for example, as the result of the systematic unroofing of a source area, can lead to the generation of distinctive **petrofacies** in the basin fill. This may provide a useful subsurface mapping technique, because compositional variations can be recognized in thin sections made from well cuttings. A detailed case study by Dickinson and Rich (1972) was among the first to make use of these ideas. Morton and Hurst (1995) demonstrated that the technique can be applied to heavy-mineral data, in a study of a petroleum reservoir unit in the North Sea basin. Other mineralogical and chemostratigraphic techniques are described in a collection of research articles edited by Dunay and Hailwood (1995).

Astronomical forcing of the earth's climate leaves sedimentary signatures of many forms. The ^{18}O record is, as noted in the previous section, dependent on seawater temperatures that are largely controlled by this process over 10^4–10^5-year time periodicities. Other signatures include various chemical cycles and variations in clastic sediment input into sedimentary basins. House (1985) proposed that the record of orbital frequencies preserved in the rock record could assist in the refinement of a geologic time scale by providing calibration of biostratigraphic data with a 10^4- to 10^5-year precision. This has become a fruitful area of research (House and Gale 1995), but much remains to be done to clarify possible changes with time in orbital frequencies (Miall 1997; Sect. 10.2.3),

and the broad framework of absolute ages within which refined cyclostratigraphic determinations can be carried out. So far, a reliable **cyclostratigraphic (astrochronologic)** time scale is only available for the youngest Cenozoic strata, back to 5 Ma (Hilgen 1991; Berggren et al. 1995a). Milankovitch cycles are described in more detail by Miall (1997), and a brief treatment of this topic is given in Chapter 8.

3.7.8 Stratigraphic Completeness

Our entire stratigraphic edifice rests on the assumption that rocks are correlatable, given units formed at the same time in different places under different depositional environments. The reconstruction of stratigraphic and paleogeographic histories, the erection of regional and global sequence stratigraphies, and the development of a global time scale are based on this assumption. However, increasingly detailed stratigraphic studies indicate that we must proceed here with considerable caution.

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). Therefore, although time is continuous, the stratigraphic record of time is not. Time, as recorded in the stratigraphic record, is discontinuous on several time scales. Breaks in the record range from such trivial events as the nondeposition or erosion that takes place in front of an advancing bedform (a few seconds to minutes), to the nondeposition due to drying out at ebb tide (a few hours), to the summer dry periods following spring run-off events (several months), to the surfaces of erosion corresponding to sequence boundaries (tens to at least hundreds of thousands of years), to the longer breaks caused by tectonism, up to the major regional unconformities generated by orogeny (millions of years). There is a similarly wide variation in actual rates of continuous accumulation, from the rapid sandflow or grainfall accumulation of a cross-bed foreset lamina (time measured in seconds) 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 for hundreds or thousands of years or more).

Sadler (1981) documented this variation in sedimentation rates in detail, using 25,000 published records. His synthesis showed that measured sedimentation rates vary by eleven orders of magnitude, from 10^{-4} to 10^{7} m/ka. This huge range of values 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. Because of this, it is invariably found that the longer the interval of time represented by a geological section the smaller the calculated sedimentation rate. Algeo and Wilkinson (1988) concluded, following a similar discussion of sedimentation rates, that in most stratigraphic sections only about one thirtieth of elapsed time is represented by sediment. Even in the deep oceans it has been demonstrated that, in some cases, "there is almost as much geological time represented by unconformities as there is time represented by sediments" (Aubry 1995).

Miall (1991) suggested that the sedimentary time scale constitutes a natural hierarchy corresponding to the natural hierarchy of temporal processes (diurnal, lunar, seasonal, geomorphic threshold, tectonic, etc.). The sedimentary breaks within and between supposedly continuous successions may represent significant lengths of time, much longer than is suggested by calculations of long-term sedimentation rates. This opens the possibility that the successions that were deposited during the time span between the breaks may not actually correlate in time at all, and this has profound implications for the precision of isotopic age determinations and the validity of matching reversal patterns in magnetostratigraphic records. Physical tracing of stratigraphic units by use of marker horizons, mapping of erosion surfaces and sequence boundaries, etc., may confirm the existence of a regional correlatable framework, but the deposits could, in principle, be markedly diachronous; correlation between basins, where no such physical tracing is possible, should be viewed with extreme caution.

3.7.9 The Global Time Scale

Basin analysts engaged in any kind of project, be it local or regional in scope, using surface or subsurface data, require the most accurate dating available for their stratigraphic sections. This is vital for local, internal correlation within a project area, when relative ages may suffice. It is even more essential if attempts are to be made to relate local events to a regional, continental, or global chronology. The aim of this chapter has been to discuss the principles behind the various techniques used in stratigraphic correlation. In this section, a brief survey will be presented of the time scales that have been established for the geological column. Use of any of these scales requires "professional help," to borrow the words of the advice columnists. They summarize an enormous amount of interdisciplinary effort extending back, in the case of biostratigraphy, for nearly 200 years. This discussion can, therefore, focus only on some of the most recent review articles in which the data have been collated and synthesized. Those interested in the details of a

particular time period must search through the appropriate research literature. Harland et al. (1990) provided a comprehensive and readable synthesis, with many references.

Figure 3.37 illustrates the process of collating chronostratigraphic data from a variety of sources around the world in order to construct a global stage framework calibrated with absolute ages. A Cretaceous example is used. During much of the Mesozoic, the presence of the wide east–west-oriented Tethyan Ocean, separation of the North and South American continents, and the broad latitudinal extent of the American continents led to faunal provincialism in benthic and some pelagic fossil groups. The Tethyan, Boreal, and other provinces indicated in Fig. 3.37 are based on the distribution of Cretaceous ammonites. Only a limited number of the actual data sites is shown.

3.7.9.1
The Phanerozoic

The first basis on which the Phanerozoic time scale was erected was the relative scale of biostratigraphy. More recently, radiometric dating has provided ages in years for biostratigraphically based subdivisions. These ages are now assigned, with varying degrees of confidence, to most Phanerozoic stages. One of the first complete compilations of this work was that published by Harland et al. (1964). Chapters in this book deal individually with every period from the Cambrian to the Tertiary. The compilation by Harland et al. (1990) is, currently, the most authoritative source of data, reference sources, and assigned ages, although refinements and revisions continue. A recent collection of research results was assembled by Berggren et al. (1995b).

For post-Triassic times, a great deal of chronostratigraphic information is now available, based on information from oceanic sediments, including magnetostratigraphic and chemostratigraphic data. In addition to the long-established ammonite zones, biostratigraphic schemes have been established for nannofossils, radiolaria, benthic and planktonic foraminifera, and palynomorphs. Magnetic-reversal sequences in ocean-floor sediments were used in early attempts to establish time scales by assuming constant sea-floor-spreading rates (Heirtzler et al. 1968),

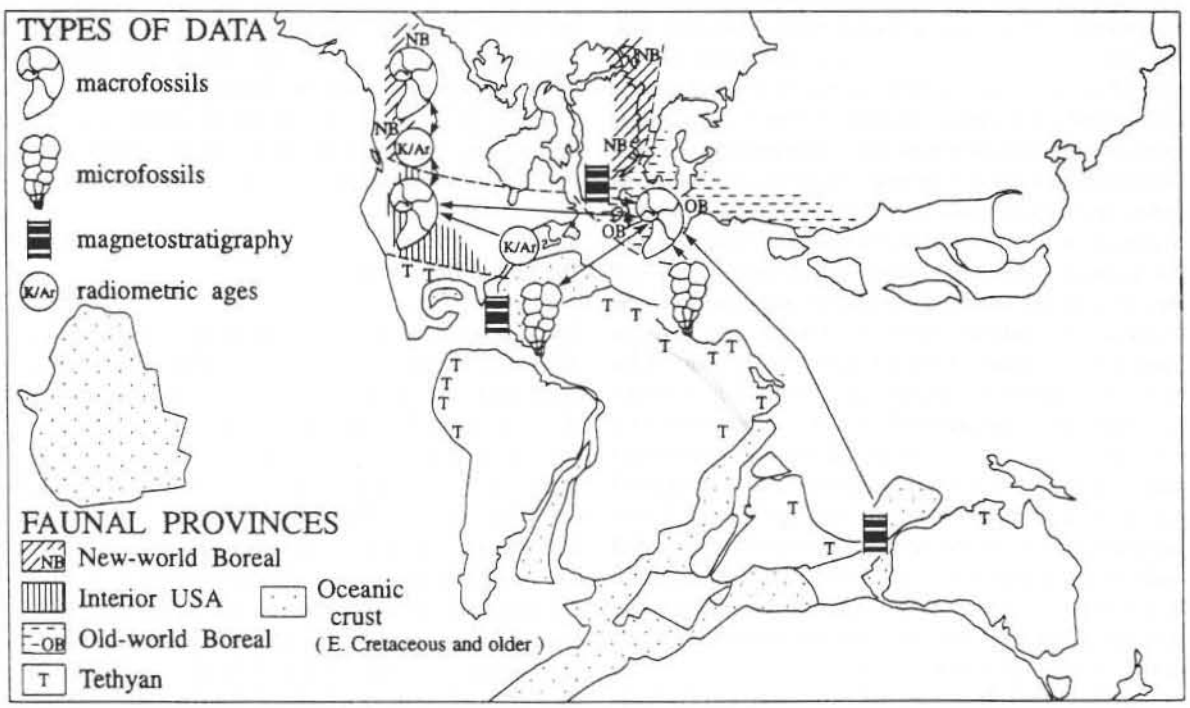


Fig. 3.37. The process of constructing a global time scale based on the collation of biostratigraphic, magnetostratigraphic and radiometric data. This diagram is loosely based on the construction of the Cretaceous time scale by Van Hinte (1976b) but, of necessity, shows only a few of the key data points used in the compilation. Macrofossil data from the original nineteenth-century type sections in Europe comprise a critical base for the time scale. The western interior of North America was also a key area because of the wealth of well-studied sections there (e.g., Obradovich and Cobban 1975). A few other data points, many representing Deep-Sea Drilling Project data, are indicated. Biostratigraphic correlation is complicated by the problem of faunal provincialism. A very simplified distribution of Cretaceous ammonite faunal provinces (from Kennedy and Cobban 1977; Kauffman 1984) is shown

but Odin and Curry (1985) demonstrated that this assumption is inappropriate and has led to serious error. Recent work has tied reversal events to biostratigraphic and radiometric data.

Berggren (1972) and Van Hinte (1976a, b) were among the first to attempt to bring together all available biostratigraphic, radiometric, and magnetostratigraphic information in order to construct integrated time scales. These served as the standard scales for the Cenozoic, Jurassic and Cretaceous periods, respectively, for several years. For example, Van Hinte's scales were used in the attempt to calibrate global seismic stratigraphic sequences, as discussed in Chapter 8. They were regarded as authoritative syntheses, but an examination of the database shows how tentative they really were. This point cannot be over emphasized, because there is a tendency to use such scales as precise time indicators, assuming that every biozone has been accurately radiometrically dated. This was and still is far from being the case. Only the major system boundaries and some sets of stages have been established by the definition of global stratotypes (Holland 1998). The practical requirement of locating complete, biostratigraphically rich sections that are also amenable to "absolute" dating by use of radiometric or magnetostratigraphic methods is a surprisingly difficult one to fulfill. In most cases, the ages of biozone or stage boundaries are determined by interpolation between reliably dated "tie points," usually by assuming constant sedimentation rates. Another approach has been to assume that, within suites of biozones that are all based on similar types of fauna (such as the graptolite zones of the Paleozoic or the ammonite zones of the Mesozoic), the individual biozones are all of equal duration. Both such assumptions are commonly wrong, but such approaches constitute a useful first step until more precise data can be brought to bear on the problem.

The problem of accuracy and precision in the global time scale and the potential for correlation error were discussed at length by Miall (1994, 1997; Chap. 13). Cope (1993) provided a useful discussion of biostratigraphic precision. Locally, a precision of ±100 ka may be achieved in Cretaceous and Cenozoic sections, if, for example, there are numerous datable volcanic horizons or an abundant ammonite record (Kauffman et al. 1991). However, this is rarely the case. Harland et al. (1990) indicated that the range of possible ages for stage boundaries in the Cretaceous vary by as little as ±4 m.y. for the Albian-Cenomanian boundary (the difference between the likely minimum and maximum possible ages) and as much as ±25 m.y. for the Aptian-Barremian boundary. They assigned an overall average 2% uncertainty to the calibration of the Phanerozoic scale (±2 m.y. at 100 Ma). These error values relate to the best available global data calibrated by several independent means, yet they reveal a residual imprecision that would not permit the dating of any given stratigraphic event, even in one of the global stratotypes, to better than ±2 m.y. In their new Mesozoic time scale, Gradstein et al. (1995) provided revised estimates of Mesozoic stage boundaries, with uncertainties expressed in millions of years to two standard deviations. These increase from ±0.5 m.y. for some of the late-Cretaceous stage boundaries, to ±2.6 m.y. for the Jurassic–Cretaceous boundary, to errors increasing from ±3 m.y. to ±4.8 m.y. for stages extending back from the late Jurassic to the early Triassic.

Two standard versions of the geologic time scale have received a particular prominence. The Decade of North American Geology Project of the Geological Society of America (DNAG) required that a standard geologic time scale be available for all contributors to the project, and an *ad hoc* advisory committee was formed in 1982 to advise on this matter. The result was a standard scale published in *Geology*, which has been widely reproduced (Fig. 3.38). Second, the various versions of the Elsevier time table have, over the years, proved extremely useful to the working geologist. Because of its compactness and comprehensiveness, it should be (and very nearly is) on every geologist's office wall. A fourth edition was prepared by Haq and van Eysinga (1987). In addition to radiometric ages, this chart also shows the relationship to each other of various systems of biostratigraphic stage nomenclature.

There are numerous minor differences between these and other "standard" time scales. Figure 3.39 illustrates the differences in assigned ages of Cretaceous stage boundaries in eight recent authoritative time scales. The differences between these scales reflect the different weights assigned to different data sets by the various authors and the different methods of interpolation and extrapolation between fixed points. Residual imprecisions in these scales and the experimental error involved in dating individual sections require that the working geologist be cautious in attempts at long-distance correlation. That the early modern work in sequence stratigraphy did not pay adequate attention to these points has been the cause of much debate and confusion regarding the global causes of sequence generation (Chap. 8); in particular, it underlies the heated debate that has taken place about global eustasy and the global cycle chart (Sect. 8.5). It is for this reason that it is quite inappropriate for summary charts, such as that published by Harland et al. (1990), to include the current Exxon global-cycle chart as if it represents a standard of geologic time equal in value and significance to the biostratigraphically-based stage framework.

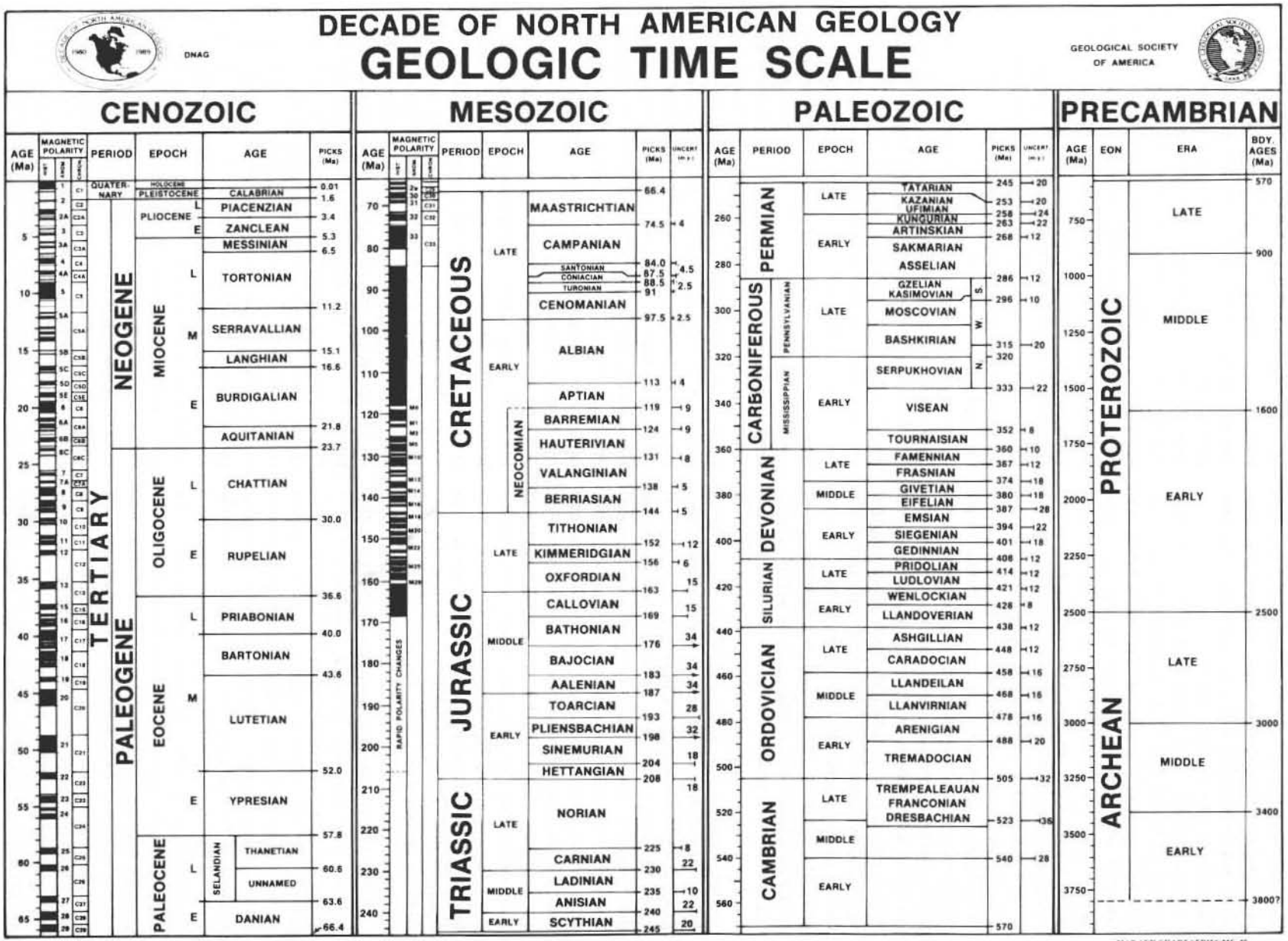


Fig. 3.38. The Decade of North American Geology time scale. (Palmer 1983)



Fig. 3.39. A comparison of Cretaceous stage-boundary ages in eight recent time scales. "This paper" refers to the source of this diagram, a revised Mesozoic time scale by Gradstein et al. (1995)

The ^{18}O record, especially where it has been "tuned" by reference to the orbital frequencies, is providing a highly precise time scale for the late Cenozoic era. This scale has now been extended back to 5 Ma and has been used to calibrate magnetostratigraphic reversal events to within 10^4 years (Hilgen 1991; Fischer 1995; Gale 1998). Many researchers are using this scale in studies of high-frequency sequence stratigraphy (**cyclostratigraphy**) of the Late Cenozoic glacioeustatic sedimentary record (Sects. 3.7.6, 8.3.4.1).

3.7.9.2 *The Precambrian*

It has taken considerable time to develop a standard global time scale for the Precambrian. There are several obvious reasons for this:

1. The biostratigraphic basis for chronostratigraphic subdivision is inadequate except, perhaps, for about the last 100 m.y. of the Precambrian
2. There are very few continuous stratigraphic sections through sedimentary or low-grade metamorphic rocks that are capable of accurate dating and that would, therefore, serve as global chronostratigraphic stratotypes
3. Radiometric determinations must be made mainly on igneous rocks or metamorphic events, which commonly provide only minimum or maximum ages for sedimentary successions. The determinations themselves are subject to possible errors as high as ± 100 Ma.
4. Most chronologies that have evolved are based on igneous, metamorphic, or tectonic events for specific, local shield areas. One of the lessons of plate tectonics is that these are extremely unlikely to have any global significance (Chap. 9).

These difficulties have led to controversy regarding the appropriate method of dealing with Precambrian time (Harland 1983; Nisbet 1985). One of the largest and best-exposed areas of Precambrian rocks is the Canadian Shield. Here, a Precambrian time scale has been erected using an approach rather different from that usually used for the Phanerozoic. Stockwell (1973, 1982) redefined the old terms Archean and Proterozoic eons and subdivided the latter into three

eras. The boundaries for these eons and eras were drawn at the end of major orogenic events, as defined by the cooling phase of late orogenic granites or pegmatites. Each of these orogenies was defined in a type area of the Canadian Shield. The ages of these events were based on numerous U/Pb and Rb/Sr whole-rock determinations. These do not give precisely the same results and, because of uncertainty in the ^{87}Rb decay constant at the time the scale was developed, two columns for Rb/Sr ages were given. This scale was found to be satisfactory for use throughout the Canadian Shield and is the official scale used in all maps and reports of the Geological Survey of Canada. However, as Harland (1983) noted, the use of a type area rather than a single stratotype for each subdivision was inappropriate, and the failure to come to grips with the revisions in decay constants in the later paper lessened the precision of the scale. As a result of these problems, plus the obvious provinciality of the orogenic record on which it is based, the Stockwell time scale has not received much use outside Canada. Recent work by the Geological Survey has, however, confirmed the usefulness of the scale for Canadian purposes (Fig. 3.40). Okulitch (1988) noted a change in emphasis away from boundaries dated on the basis of orogenic events, in part because of the diachroneity of orogeny. The preference now is for boundaries dated with reference to specific type sections, an approach more comparable to that used for the Phanerozoic.

James (1972) published the deliberations of a committee of the U.S. Geological Survey that reached quite different conclusions. This committee recommended that the subdivision of the Precambrian should be purely temporal (geochronometric), setting aside the principle of subdivision based on stratotypes. It also rejected the idea of using major orogenies as time markers, pointing out that these may last for tens of millions of years. The committee rejected terms such as Archean and Proterozoic on the basis that there was no agreement on their definition. Similar conclusions had earlier been reached by Goldich (1968), who proposed a scale consisting of equal time slices independent of any specific rock units or areas. However, James's (1972) committee chose not to go quite this far, dividing the Precambrian into four units based on groupings of sedimentary episodes, orogenies, and times of plutonism (Fig. 3.40).

This same approach was adopted by the IUGS Working Group on the Precambrian for the United States and Mexico (Harrison and Peterman 1980, 1982). They returned to the terms Archean and Proterozoic and subdivided each into three eras, as shown in Fig. 3.40. Note that most of the proposed boundaries differ from others shown in this figure (Australian and Russian scales are included for comparison). This points up the difficulty – perhaps futility – of attempting to reach a tectonically based consensus on the chronostratigraphy of what are, in effect, randomly accreted terranes having widely varying tectonic histories.

These problems can be exemplified by the definition of the Archean–Proterozoic boundary. The

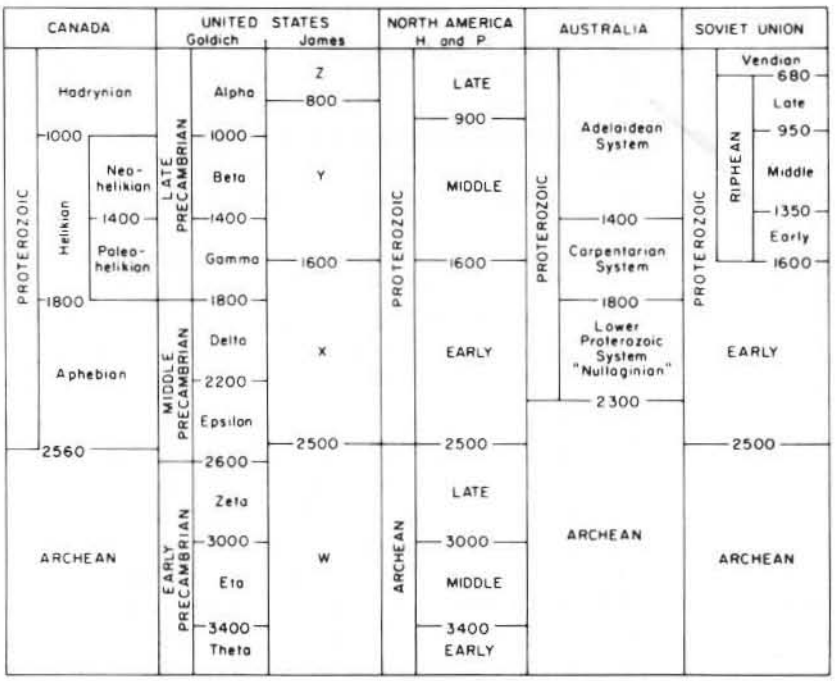


Fig. 3.40. Comparison of various Precambrian time scales. (Compiled from Dunn et al. 1966; Goldich 1968; James 1972; Stockwell 1973; Preiss 1977; Harrison and Peterman 1980, 1982)

IUGS Commission on Precambrian Stratigraphy recommended that this be fixed at the conveniently rounded-off value of 2.5 Ga (James 1972). At best, this is a convenient convention, but it has raised difficulties. It is widely recognized that crustal processes were quite different in the Archean and Proterozoic eons (Sect. 9.5.1), but the transition from one crustal style to the other occurred at different times in different parts of the globe. The 2.5-Ga boundary seems appropriate for North America, but one of the key indicators of Proterozoic-style crust, the development of recognizable, stable cratons with shallow-water cover sediments, can be documented in rocks as old as 3.0 Ga, possibly 3.5 Ga, in southern Africa (Swaziland Supergroup; Sect. 9.5.2; Windley 1984). As with the Phanerozoic, however, the achievement of international agreement about the positioning of a boundary is much more important than the actual location of that boundary (in time or space), and the setting of the Archean-Proterozoic boundary at 2.5 Ga seems now to have achieved almost universal acceptance.

Turning now to the Proterozoic, there seems to be considerable promise that a practical basis for a broad biostratigraphic subdivision will be developed within a few years. Russian geologists have long used acritarchs (Timofeev 1973) and stromatolites (Semikhatov 1974) as biostratigraphic indicators. Their subdivision of the Precambrian, based in part on stromatolite morphology for the post-1600 Ma period, is given in Fig. 3.40. However, stromatolite assemblages in Proterozoic rocks elsewhere do not permit very precise correlation with the Russian standard (Preiss 1977). Hofmann (1977) pointed out that the nomenclature of stromatolite study is very confused and needs to be properly quantified before this subject can be properly tested. Preiss (1977) stated "since the classification of stromatolites is based on morphological features not of micro-organisms, but of the organo-sedimentary structures built by them, it would be inappropriate to use a biological binomial nomenclature," a practice that is still common. The sedimentary environment exerted a major control on stromatolite morphology, and similar morphologies were built by different microbiotas at different times during the Proterozoic (Schopf 1977).

A more promising approach may be the study of the microbial communities that built the stromatolites. Schopf (1977) reported the recognition of potential index fossils and crude range biozones for the Proterozoic, and further refinements are anticipated. These organisms occur fairly frequently in Proterozoic sediments and, therefore, the method has the potential to be of considerable practical use. Hofmann (1987) provided a useful recent review of Precambrian biostratigraphy.

International agreement regarding the position of the Precambrian–Cambrian boundary has yet to be reached (Cowie 1985). The base of the Cambrian period has long been treated as coincidental with the base of the Phanerozoic, this being the time when abundant shelly faunas first appear in the rock record. However, intensive paleontological work on upper Proterozoic strata in many parts of the world has revealed a considerable wealth of fossil material, and there have been pressures to revise the base of the Phanerozoic downward. For example, between about 670 Ma and the beginning of the Cambrian period, the globe was characterized by a distinctive assemblage of soft-bodied metazoan organisms, termed the Ediacarian fauna, after the Ediacara Hills in South Australia. Elements of the fauna have been found in many parts of the world. Cloud and Glaessner (1982) proposed the erection of a new geochronologic/chronostratigraphic unit, the Ediacarian Period/System, and they further proposed that, because the fauna shows clear evolutionary links to those in the Phanerozoic eon and represents the first known multicellular (metazoan) animal life, the base of the Phanerozoic eon should be redefined downward to include the Ediacarian. They argued that, because the Ediacarian cannot be both Phanerozoic and Precambrian, the name of the latter should be changed to the old term Cryptozoic. The definition of an Ediacarian Period and System is an excellent way to recognize advances in chronostratigraphy, but the accompanying suggestions to revise the Phanerozoic eon and rename the Precambrian have not found wide acceptance. Further elaboration and modification of these ideas were given in a later paper Glaessner (1984).

The Precambrian–Cambrian Boundary Working Group has now decided to consider the Ediacaran fauna as Precambrian (Cowie 1985). It had decided, at a meeting in 1978, to place the boundary "near the evolutionary changes signaled in the rocks by the appearance of diverse fossils with hard parts," but at a later meeting (1983) changed this to "the lowest known appearance of diverse shelly fossils with a good potential for correlation" (Cowie 1985). A candidate for the boundary stratotype had been selected at Meishucun, China, but doubts about the suitability of this section remain, and a final decision is pending.

In 1990, the International Union of Geological Sciences approved a new standard chronometric scale for the Precambrian eon, which is reproduced here as Fig. 3.41. This was reported by Plumb (1991), who stated "the Subcommission on Precambrian Stratigraphy recommends strictly a chronometric subdivision of Precambrian time, in which boundaries are defined simply by precise isotopic ages that are expressed in years and have neither direct reference to any body of rock nor plus or minus error

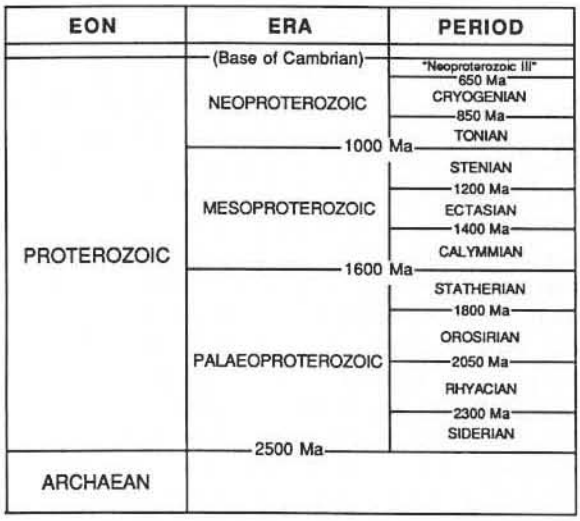


Fig. 3.41. The International Union of Geological Sciences-approved chronometric subdivision of Precambrian time. (Plumb 1991)

bars." The unit labeled "Neoproterozoic III" is intended to be provisional until such time as the Working Group on the Terminal Precambrian System defines and names a new chronostratigraphic unit and stratotype to define the top of this unit.

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Facies Analysis

4.1 Introduction

The study and interpretation of the textures, sedimentary structures, fossils, and lithologic associations of sedimentary rocks on the scale of an outcrop, well section, or small segment of a basin comprise the subject of facies analysis. Several excellent texts deal extensively with facies analysis methods and process-response facies models, for example, the introductory texts by Walker and James (1992) and McLane (1995), and the more advanced treatments by Reading (1996) and Prothero and Schwab (1996). Therefore, little would be gained by presenting a complete discussion of facies models. Accordingly, this chapter focuses on a discussion of analytical methods and reviews the kinds of information that can be obtained from sediments based on the observations described in Chapter 2. It is hoped that this material will provide students with an introduction to modern facies-modeling methods. Practice on specific ancient examples can then be carried out with one of the advanced texts (mentioned above) in hand, and the final section of this chapter should serve to complement these by providing a discussion and an update of some of the most recent advances in our ideas.

To extend a facies analysis to an entire basin, it is necessary to come to grips with problems of stratigraphic correlation and to apply various basin-mapping techniques. Stratigraphic methods are described in Chapter 3, and basin mapping, including the use of isopach and facies-ratio maps, paleocurrent analysis, and so on, is dealt with in Chapter 5. The end product of this work is a paleogeographic synthesis depicting an interpretation of the stratigraphic and geographic evolution of the basin through time. Certain large-scale basin-fill patterns have emerged over the years from work of this kind. These are referred to as depositional systems and are described in Chapter 6. The difference between facies analysis and depositional-systems analysis is essentially one of scale. However, the demarcation is not clear cut, and some overlap between Chapters 4 and 6 is unavoidable. This point is returned to in Section 4.7. Facies analysis is an essential complement to the practice of sequence stratigraphy, which is described in Chapter 6.

4.2 The Meaning of Facies

The meaning of the word facies has been much debated in geology (Moore 1949; Teichert 1958; Krumbein and Sloss 1963). It is widely used in sedimentary geology but also has a somewhat different meaning in the area of metamorphic petrology (Fawcett 1982). Anderton (1985), Reading (1996; Chap. 2), and Pirrie (1998) provided excellent discussions of the modern sedimentological uses of the term and the methods of interpreting individual facies and facies relationships.

The word facies is now used in both a descriptive and an interpretive sense, and the word itself may have either a singular or plural meaning. Descriptive facies include **lithofacies** and **biofacies**, both of which are terms used to refer to certain observable attributes of sedimentary-rock bodies that can be interpreted in terms of depositional or biological processes. (When used without a prefix in this book, the word facies is intended to mean either lithofacies or biofacies.) An individual lithofacies is a rock unit defined on the basis of its distinctive lithologic features, including composition, grain size, bedding characteristics, and sedimentary structures. Each lithofacies represents an individual depositional event. Lithofacies may be grouped into **lithofacies associations** or **assemblages**, which are characteristic of particular depositional environments. These assemblages form the basis for defining **lithofacies models**; they commonly are cyclic. A biofacies is defined on the basis of fossil components, including either body fossils or trace fossils. The term biofacies is normally used in the sense of an assemblage of such components. For the purpose of sedimentological study, a deposit may be divided into a series of facies units, each of which displays a distinctive assemblage of lithologic or biologic features. These units may be single beds a few millimeters thick or a succession of beds tens to hundreds of meters thick. For example, a river deposit may consist of decimeter-thick beds of a conglome-

rate lithofacies interbedded with a crossbedded, sandstone facies. Contrast this with the biofacies terms used to describe the fill of many major, early-Paleozoic basins. Commonly, this may be divided into units hundreds of meters thick comprising both a shelly biofacies, containing such fossils as brachiopods and trilobites, and a graptolitic biofacies. At the other extreme, J. L. Wilson (1975) recommended the use of microfacies in studying thin sections of carbonate rocks and defined 24 standard types.

The scale of an individual lithofacies or biofacies unit depends on the level of detail incorporated in its definition. It is determined by the variability of the succession, the nature of the research undertaken (basin-wide reconnaissance versus a detailed local study), or the availability of rock material for examination. Facies units defined on the basis of outcrop, core, well-cutting, or geophysical criteria tend to refer to quite different scales and levels of detail. Geophysicists in the petroleum industry refer to seismic facies, but this is not comparable to the small-scale type of facies discussed in this chapter (seismic facies are referred to in Chapters 5 and 6). Modern, high-resolution, shallow, seismic surveying coupled with side-scan sonar imaging is providing a powerful tool for the analysis of facies compositions and geometries in modern environments and is beginning to have a major impact on the understanding of shelf and slope sedimentary environments (Sects. 4.6.7, 4.6.9). Increasing attention is being paid to the three-dimensional geometry of facies units, particularly in outcrop studies and subsurface studies involving reservoir development (Sect. 4.3.4). To a large extent, the scales at which facies units are defined reflect criteria of convenience. Thus, the term is a very flexible and convenient one for descriptive purposes.

The term facies can also be used (usually for lithofacies assemblages) in an interpretive sense for groups of rocks that are thought to have been formed under similar conditions. This usage may emphasize specific depositional processes, such as till facies or turbidite facies. Alternatively, it may refer to a particular depositional environment, such as shelf carbonate facies or fluvial facies, encompassing a wide range of depositional processes.

Widespread use is still made of two 19th-century, Swiss, stratigraphic terms that have acquired a generalized facies meaning encompassing lithologic characteristics, depositional environment and tectonic setting. The first of these is the flysch facies, comprising marine sediments, typically turbidites and other sediment-gravity flow deposits, formed on tectonically active continental margins. The molasse facies consists of nonmarine and shallow-marine sediments, mainly sandstones and conglomerates, formed within and flanking fold belts during and following their elevation into mountain ranges. Both of these facies types may make up major stratigraphic units hundreds or thousands of meters thick and extending for hundreds or thousands of kilometers. Continued use of the terms flysch and molasse is not recommended because of ambiguities about their implications for the tectonic setting. The vague association with orogenic belts is inadequate now that we have the theory of plate tectonics to assist us in interpreting the relationship between sedimentation and tectonics (Chap. 9).

Lithostratigraphy and lithofacies analysis are two contrasting approaches to the study of sedimentary rocks. The first is the traditional descriptive approach. The second is based on detailed facies descriptions, which provide the basis for the genetic study of sediments using facies models. Lithofacies and biofacies analysis must be used to assist in stratigraphic studies because, by understanding the depositional environments and paleogeography existing at the time a rock unit was formed, we are much better equipped to make predictions and extrapolations about lateral changes in thickness and composition. Obviously, this will be invaluable for correlation purposes and can make for a much more logical definition of formal lithostratigraphic units. Biofacies analysis is crucial in the definition and comprehension of biostratigraphic units.

Lithostratigraphy and facies analysis are essential components of the study of **sequence stratigraphy**, which is a method of regional stratigraphic analysis based on the subdivision of stratigraphic successions into unconformity-bounded units that are generated by regional or global changes in the relative sea level. Some of the sedimentological techniques of sequence stratigraphy are discussed in Section 4.4.5 and Chapter 6, and the subject is returned to in Chapter 8, in which we discuss some of the models of global cyclicity that have emerged during the last decade and the mechanisms of sea-level change that give rise to sequences.

4.3 Recognition and Definition of Facies Types

4.3.1 Philosophy and Methods

In order to make sense of the lithologic variability present in most sedimentary basins, it is necessary to generalize, categorize, and simplify what we see in well sections and outcrops. It is necessary because the human brain can only absorb and process a limited amount of detail at any one time. In sedimentology, we find that much of the variability disguises a limited range of basic lithofacies and biofacies types and

that variations between these types represent minor, random, environmental fluctuations or are even the result of accidental exposure or the position of a thin-section cut. The existence of this natural pattern is what makes facies studies, facies modeling, and paleogeographic reconstruction possible. For example, for many years, sedimentologists categorized most deep-sea deposits in terms of only five basic lithofacies. These were the T_a, T_b, T_c, T_d, and T_e divisions of the classic Bouma turbidite sequence (Bouma 1962). This Bouma model served sedimentologists well, and was used as the type example in Walker's (1976; subsequently repeated in Walker and James 1992) definition of facies models. However, it has always been known that some beds do not fit the pattern, as discussed further below. The standard submarine-fan model, which evolved in the 1970s (Mutti and Ricci-Lucchi 1972; Walker 1978), interpreted Bouma sequences as occurring mainly on the outer, nonchannelized part of a fan. The lithofacies spectrum for submarine fans and for other types of deep-sea sand deposit is now rather more complex, as discussed in Section 4.6.9.

The deposits of many other depositional environments have proved to be amenable to some simple, empirical classifications. For example, Miall (1977, 1978, 1996) showed that most river deposits could be described using about 20 lithofacies types. J. L. Wilson (1975), in a major review of Phanerozoic carbonate rocks, concluded that most could be described satisfactorily by drawing from a list of only 24 standard microfacies. Eyles et al. (1983) proposed a simple lithofacies classification scheme for the description of glacial marine and glacial lacustrine deposits.

Facies modelers have yet to categorize all sediment types in every variety of depositional environment; for example, no one has yet attempted to erect a universal lithofacies scheme for clastic rocks comparable to the scheme of J. L. Wilson (1975) for carbonates. Similarly, many fossil groups await detailed biofacies analysis. For this reason, every basin analyst should study each basin with a fresh eye and, at least in the preliminary stages of the research, erect a local facies scheme without too much dependence on previously published work. Slavish adherence to such readily available research keys may result in minor but critical lithofacies types and lithofacies relationships being missed or forced to fit an inappropriate mold. New facies models or a better definition of old ones will never come about if geologists are content merely with such replication studies. Bridge (1993) has expressed this concern forcefully with respect to the facies scheme now in widespread use for fluvial deposits (Sect. 4.3.3), and Shanmugam (1997) has argued strongly that the Bouma-sequence concept has been seriously misused in the study of deep-sea sandstones. Others, such as Walker (1990, 1992) and Reading and Levell (1996), are strongly in favor of the simplifying and categorizing methods of facies analysis. The problem that concerns Bridge (1993) and Shanmugam (1997) is that hasty or lazy field observation can too easily lead to incorrect facies classification. For example, Shanmugam (1997) argued that for too long sedimentologists have been applying the Bouma-sequence concept and its fivefold subdivisions to deposits that exhibit few of the key characteristics of the original sequence. Many successions contain no graded basal unit and contain internal breaks in sedimentation (indicating that they are not the product of single depositional events), evidence of deposition from other types of sediment-gravity flow, such as rafted clasts, that would indicate debris-flow deposition, and so on. In fact, such practices have become so common, in Shanmugam's (1997) view, that he refers to a phenomenon that he terms the "turbidite mind set."

Hummocky cross-stratification (HCS) is an excellent example of a structure that sedimentologists had been looking at without seeing for many years, until it was focused upon and given a name by Harms et al. (1975). Suddenly, it was realized that this distinctive style of crossbedding is characteristic of many storm-generated ancient deposits. Its recognition has led to the reappraisal of numerous shallow-marine sequences, to the extent that the term became something of a cliché in the early 1980s (Byers and Dott 1981). This is a good example of critical advances in sedimentological interpretation being missed because of the widespread lack of critical, independent observation.

How, then, does the basin analyst perform a facies analysis of an undescribed rock sequence or succession of fossils? The method for measuring and describing stratigraphic sections is discussed in detail in Chapter 2. For the purpose of lithofacies analysis, the focus must be on recognizing associations of attributes that are repeated through the section (or parts of the entire basin). Lithofacies may be distinguished by the presence of bedding units with a characteristic sedimentary structure or structures, a limited grain-size range, a certain bed thickness and, perhaps, a distinctive texture or color (color is subject to diagenetic change and should not be used as a primary criterion in definition; Sect. 2.2.3). Biofacies represent associations within the same stratigraphic interval of a limited suite of genera or species. Biofacies studies may be carried out on single taxonomic groups, because many of these, if examined by specialists, can yield highly detailed paleoecological (and, hence, depositional) information. The definition of biofacies is an exercise in defining assemblages and will be discussed in the next section.

In order to recognize these associations of lithologic attributes or fossil types, it may be useful to set up a checklist and tabulate their occurrence. Field-data-recording forms, such as those discussed in Chapter 2, may be useful for this purpose. Statistical analysis is commonly used by paleoecologists to establish biofacies from such tabulated data. It is less used for erecting individual lithofacies, because many of the common associations of lithologic attributes are well-enough known to be recognized without the aid of statistics. However, the key to success is very careful, empirical observation.

4.3.2 Field Examples of Facies Schemes

Two examples of lithofacies schemes that have been erected from field data are given in this section to illustrate the methods. A simple biofacies association is also described.

Cant and Walker (1976) described a Devonian fluvial section in eastern Quebec and subdivided it into eight lithofacies. The section is shown in Fig. 4.1, in which lithologic symbols are keyed to the lithofacies scheme. Note the use, in this illustration, of the variable-width column-plotting technique for drawing stratigraphic sections, which is discussed in Section

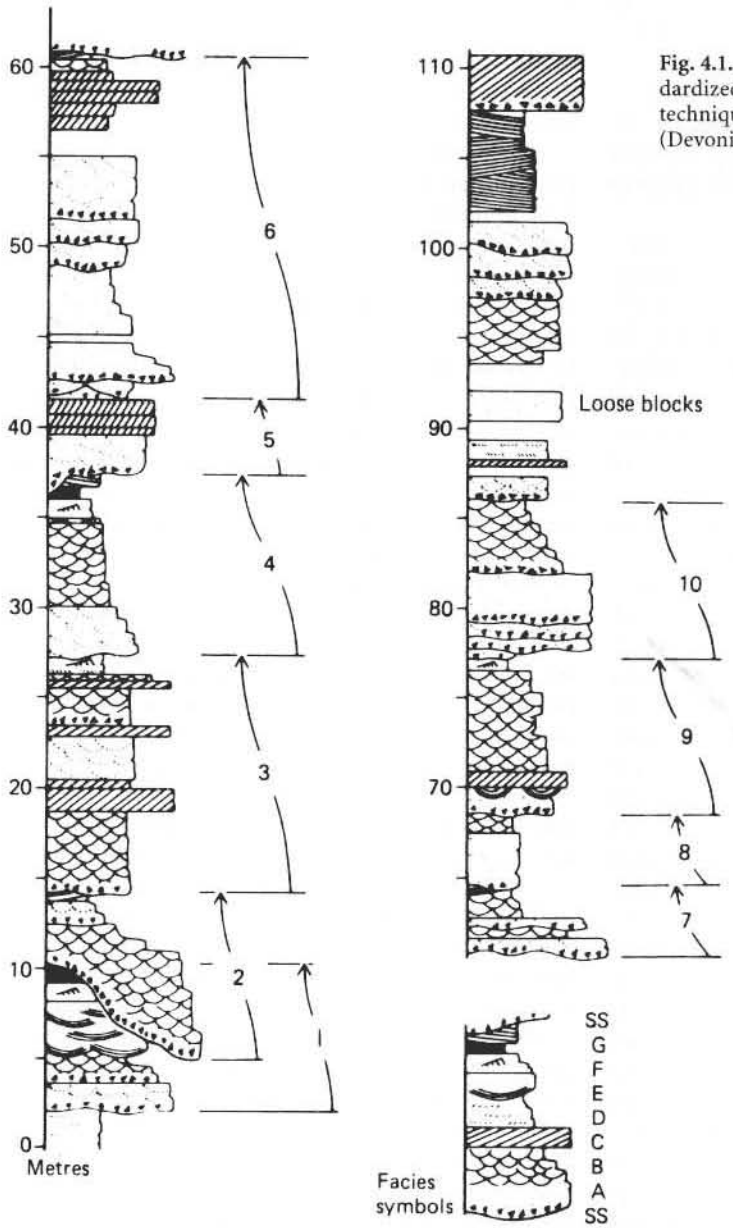


Fig. 4.1. A stratigraphic section plotted using a standardized facies scheme and the variable-width-column technique. Fluvial cycles of the Battery Point Formation (Devonian), Quebec. (Cant and Walker 1976)

2.2.5. Here are some examples of their lithofacies descriptions, edited in order to focus on salient features.

> Well defined trough crossbedded facies (*B*): This facies is composed of well-defined sets of trough crossbedding ..., with trough depths averaging 15 to 20 cm (range 10 to 45 cm). The troughs are regularly stacked on top of each other, but in some individual occurrences of the facies, trough depths decrease upward ... The sets are composed of well-sorted medium sand ... A few of the coarser sets have granules and pebbles concentrated at their bases.
>
> Asymmetrical scour facies (*E*): This facies consists of large, asymmetrical scours and scour fillings, up to 45 cm deep and 3 m wide ... The scours cut into each other and into underlying troughed facies (*A* and *B*), and occurrences of the asymmetrical scour facies have a flat, erosionally truncated top.... The main difference between the scour fillings and the two troughed facies (*A* and *B*) lies in the geometry of the infilling strata. In the asymmetrical scours, the layers are not at the angle of repose, but are parallel to the lower bounding surface.
>
> Rippled sandstone and mudstone facies (*F*): This facies includes cross-laminated sandstones ..., and alternating cross-laminated sandstones and mudstones. [An example of the latter] is 1.5 m thick and consists of three coarsening-upward sequences, which grade from basal mudstones into trough cross-laminated fine sandstone and finally into granule sandstone. The sandstones capping each coarsening-upward sequence have sharp, bioturbated tops.

A variety of criteria has been used in defining these lithofacies, but the authors refrained, at this stage, from making any environmental interpretation.

A carbonate example was described by R. C. L. Wilson (1975), who studied the upper Jurassic Oolite Series (actually a formation, but an incorrect, older nomenclature is still in use) of Dorset, England. He recognized four lithofacies, as shown in Fig. 4.2. Note in this illustration the relationship between lithostratigraphic and lithofacies units. The four lithofacies are as follows (described using the carbonate classification of Folk 1962):

1. Coarsening-upward units shown by two beds: (1) the *Chlamys qualicosta* bed of intramicrite–oomicrite–oosparite–poorly-washed biosparite and (2) pisolite, consisting of quartz sands and phyllosilicate lay–intramicrite–oomicrite–oosparite-oncolites.
2. Crossbedded sets of oosparite showing 20 to 25° dips and sharp contacts either with phyllosilicate clays with nodular micrites or bioturbated oolite. Some minor flaser bedding and clay drapes over current ripples also occur.
3. Association of *Rhaxella* biomicrites.
4. Sheet deposits (5–10 cm) and large accretion sets (30 cm) of oomicrite and biomicrite with subsidiary oosparite and biosparite. Shell debris often shows imbricate structure, and the oomicrites are texturally inverted sediments, being a mixture of extremely well-sorted oolites in a micrite matrix. Some sets showing alternating current directions occur.

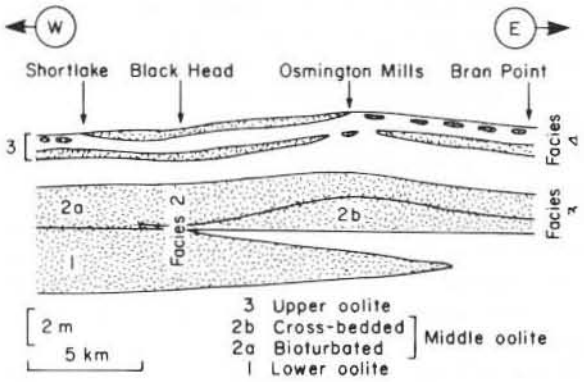


Fig. 4.2. Carbonate facies and their distribution in the Osmington Oolite (Jurassic), southern England. (R. C. L. Wilson 1975)

The careful reader will note two points about these descriptions. First, a mixture of lithofacies and biofacies criteria is used. This commonly is desirable in fossiliferous successions where the fossils are distinctive, rock-forming components. Second, several of the lithofacies consist of different carbonate rock types interbedded on a small scale. Each of these could be described as a lithofacies in its own right, in which case Wilson's four facies become facies assemblages. It is a question of which scale of description is the most suitable for the purpose under consideration. Those engaged in analyzing entire basins or major stratigraphic intervals in a basin cannot afford to spend too much time on fine detail. Decisions must be made with the core or outcrop in front of the analyst about how thick the descriptive units are going to be and how to combine thin beds for the purpose of lithofacies description. Generally, lithofacies thicknesses on the order of a few decimeters to a few meters have been found to be the most useful. Smaller-scale subdivisions may be erected for selected examples of well-exposed sections or continuous cores, if desired, to illustrate particular points in a facies description.

Wilson suspected that his Oolite Series beds were tidal in origin, and he tabulated various features of each of the four facies using a range of criteria suggested by Klein (1971) for recognizing tidal deposits (Fig. 4.3). Many of these make specialized use of paleocurrent evidence, which is discussed in Sec-

		Facies of Osmington Oolites			
		1	2	3	4
A.	1. Cross stratification with sharp set boundaries		■		■
	2. Herringbone cross-stratification			■	■
	4. Parallel laminae	■			
	5. Complex internal organization of dunes, etc.		■	■	■
B.	6. Reactivation surfaces		■		■
	7. Bimodal distribution of set thicknesses		■		
	9. Unimodal distribution of dip direction of cross strata		■	■	
	10. Orientation of cross-strata parallels sand-body trend			?	
C.	13. Small current ripples to larger current ripples			■	
	19. Symmetrical ripples			■	
D.	Flaser bedding			■	
F.	35 & 36 channel log deposits			■	
H.	Burrowing, etc.	■	■	■	

Fig. 4.3. Distribution of selected lithofacies attributes in the Osmington Oolite. (R.C.L. Wilson 1975)

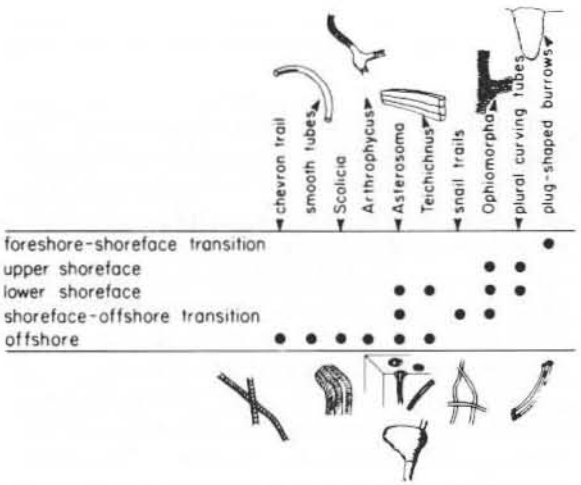


Fig. 4.4. Distribution of trace fossils in a Cretaceous, marginal-marine sandstone. (Howard 1972)

tion 5.6. Figure 4.3 illustrates the use of a checklist for lithofacies description. This can be a useful technique for field observation if the geologist suspects in advance what it is he or she is about to study. It is then possible to incorporate all the special observational detail and environmental criteria available from studies of the appropriate modern environment or other ancient rock units. In doing so, there is, of course, the danger that other features might be missed, and interpretation becomes a self-fulfilling prophecy. Tables of this kind are not recommended for illustration of written or oral presentation of a lithofacies study except, perhaps, as an appendix. They are visually difficult to absorb, and the geologist does better to provide summary descriptions, such as those given previously, plus photographs and graphic logs.

Howard (1972) gave an excellent example of the use of trace fossils in erecting simple biofacies for the purpose of environmental interpretation. He studied the upper Cretaceous, shallow-marine rocks exposed at Book Cliffs, Utah, and erected five loosely defined trace-fossil biofacies. In conjunction with lithofacies criteria, this enabled Howard to suggest environmental interpretations for each assemblage, as shown in Fig. 4.4. For example, the stratigraphically lowest facies consists of thoroughly bioturbated silt with the variety of trace fossils listed. Cleaner sands are associated with abundant *Ophiomorpha*, which is a shrimp burrow. In the shallowest waters, the turbulent conditions were unsuitable for most invertebrates. Note the occurrence of only one trace-fossil type in the foreshore–shoreface transition. In the deposits of the foreshore, no trace fossils were recorded. The use of statistical techniques in erecting more complex biofacies assemblages is discussed in Section 4.4.

4.3.3 Establishing a Facies Scheme

The examples in the previous section serve to illustrate the methods used in three different kinds of basin problems to define facies types. The objective of the basin analyst should be to erect a facies scheme that can encompass all the rock types present in his or her project area. If the project encompasses several or many stratigraphic units deposited under widely varying environments, it may make more sense to erect a separate scheme for each lithostratigraphic unit, because the lithofacies and, most likely, the biofacies assemblages will be different for each such unit.

Facies schemes should be kept as simple as possible, otherwise, they defeat the whole purpose of carrying out a facies analysis. Some authors have subdivided their rocks into 20 or 30 lithofacies and erected subclasses of some of these. This gives the appearance of meticulous research and great analytical precision, but such schemes are difficult to understand or absorb. Remember that the purpose of a facies analysis is to aid in interpretation and, in basin analysis, this is often best accomplished by judicious simplification. It is rare that many more than half a dozen distinct lithofacies occur together in intimate stratigraphic association. Commonly, two to four lithofacies occur in repeated, monotonous succession through tens or hundreds of meters of section. In defining only a few facies to cover thick and varied successions, the problem of what to group together and what to separate is always present. What should be done about thin beds? Should units that show gradational contacts be split and, if so, how? The only

answer is to do what works. One approach is to compile a very detailed description of the first well or section of a new project and to use the resulting notes to erect a preliminary facies scheme; this is then employed for describing each new section. It has the advantage that, instead of writing down basic descriptions for every bedding unit, it may be adequate simply to assign it to the appropriate facies and record in one's notes any additional observations that seem necessary, such as grain-size differences or additional sedimentary structures relative to the original descriptive scheme. Thus, the scheme itself may be modified as one goes along. This approach has worked well for the author for many years of studying clastic sedimentary sequences in outcrops and cores in the Canadian Arctic and elsewhere. It is more difficult to apply this method to carbonate rocks because of their finer-scale variability and the difficulty of seeing all the necessary features in weathered outcrops. One approach is to take into the field a rock saw and other necessary equipment for constructing etched slabs or peels. These can then be examined and described as field work proceeds. (J. L. Wilson 1975).

Friend et al. (1976) avoided these problems entirely by recording only attributes of a sequence, not lithofacies as such. They then used statistical techniques to group these attributes into lithofacies assemblages, as discussed in Section 4.4. Their field-recording methods are described in Section 2.2.1.

Anderton (1985) suggested that facies criteria be ranked in order of priority, permitting an increasingly refined subdivision of the facies units as the work proceeds. His suggested ranking for channelized clastic deposits (as an example) is:

Table 4.1. Facies classification. (Miall 1996)[a]

Facies code	Facies	Sedimentary structures	Interpretation
Gmm	Matrix-supported, massive gravel	Weak grading	Plastic debris flow (high-strength, viscous)
Gmg	Matrix-supported gravel	Inverse to normal grading	Pseudoplastic debris flow (low strength, viscous)
Gci	Clast-supported gravel	Inverse grading	Clast-rich debris flow (high strength), or pseudoplastic debris flow (low strength)
Gcm	Clast-supported, massive gravel	–	Pseudoplastic debris flow (inertial bedload, turbulent flow)
Gh	Clast-supported, crudely bedded gravel	Horizontal bedding, imbrication	Longitudinal bedforms, lag deposits, sieve deposits
Gt	Gravel, stratified	Trough crossbeds	Minor channel fills
Gp	Gravel, stratified	Planar crossbeds	Transverse bedforms, deltaic growths from older bar remnants
St	Sand, fine to v. coarse, may be pebbly	Solitary or grouped trough crossbeds	Sinuously crested and linguoid (3-D) dunes
Sp	Sand, fine to v. coarse, may be pebbly	Solitary or grouped planar crossbeds	Transverse and linguoid bedforms (2-D dunes)
Sr	Sand, very fine to coarse	Ripple crosslamination	Ripples (lower flow regime)
Sh	Sand, v. fine to coarse, may be pebbly	Horizontal lamination, parting or streaming lineation	Plane-bed flow (critical flow)
Sl	Sand, v. fine to coarse, may be pebbly	Low-angle (< 15°) crossbeds	Scour fills, humpback or washed-out dunes, antidunes
Ss	Sand, fine to v. coarse, may be pebbly	Broad, shallow scours	Scour fill
Sm	Sand, fine to coarse	Massive, or faint lamination	Sediment-gravity flow deposits
Fl	Sand, silt, mud	Fine lamination, v. small ripples	Overbank, abandoned channel, or waning flood deposits
Fsm	Silt, mud	Massive	Back swamp or abandoned channel deposits
Fm	Mud, silt	Massive, desiccation cracks	Overbank, abandoned channel, or drape deposits
Fr	Mud, silt	Massive, roots, bioturbation	Root bed, incipient soil
C	Coal, carbonaceous mud	Plants, mud films	Vegetated swamp deposits
P	Paleosol carbonate (calcite, siderite)	Pedogenic features	Soil with chemical precipitation

[a] *v.*, very; *D*, dimensional.

1. Bases of major channels or scoured surfaces
2. Tops and bases of nonclastic lithologies (e.g., coal, limestone)
3. Abrupt changes from a section composed predominantly of one grain size to a section composed of another (e.g., sandstone to shale, fine sandstone to coarse)
4. Changes of internal structure within units of similar grain size

With continual study of the same kind of rocks, a standard set of facies classes will gradually emerge and can be used for rapid field or laboratory description. For the writer, this led to the erection of a generalized lithofacies scheme for describing the deposits of braided rivers based on field studies of various ancient deposits and a review of literature on both ancient deposits and modern rivers. The initial scheme contained ten lithofacies types (Miall 1977). Further research by the writer and use by other workers led to the addition of about ten more lithofacies types (Miall 1978), but the original ten include most of the common ones. Numerous workers around the world have now applied this scheme to a wide variety of fluvial deposits and to the fluvial components of deltaic successions. Table 4.1 lists these lithofacies, showing the codes used for note taking and a sedimentological interpretation of each. The lithofacies codes consist of two parts, a capital letter for modal grain size (*G*, gravel; *S*, sand; *F*, fines) and a lowercase letter or letters chosen as a mnemonic of a distinctive texture or structure of each lithofacies. The three lithofacies B, E, and F of Cant and Walker (1976), discussed in the previous section, are St, Ss, and Fl in this scheme. Le Blanc Smith (1980) has developed this fluvial facies scheme still further by incorporating additional structures and information on grain size.

J.L. Wilson's (1975) microfacies scheme for carbonates contains 24 types. Figure 4.5 illustrates his standard legend, facies numbers, and abbreviated description. Figures 4.6 and 4.7 illustrate the use of these two schemes in drawing stratigraphic sections.

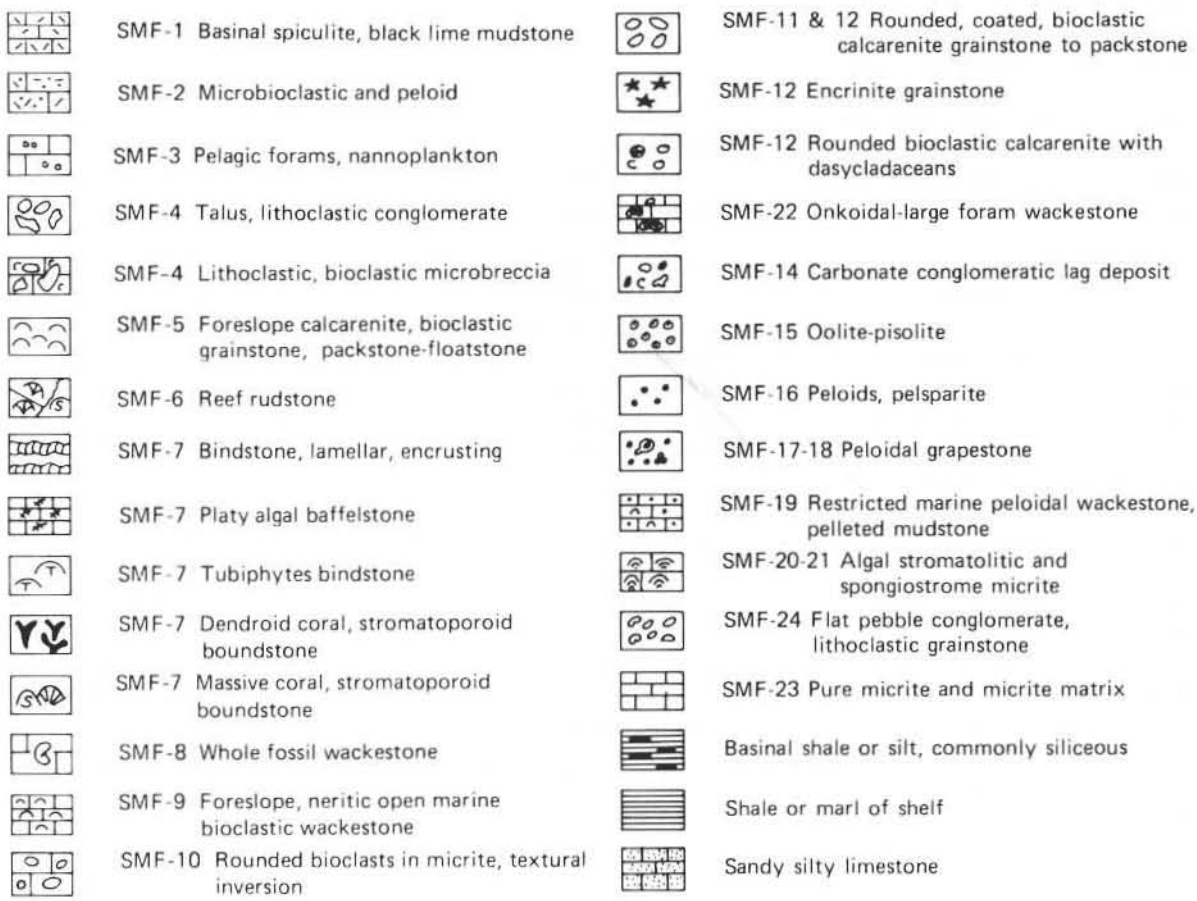


Fig. 4.5. The standard carbonate microfacies scheme of J.L. Wilson (1975)

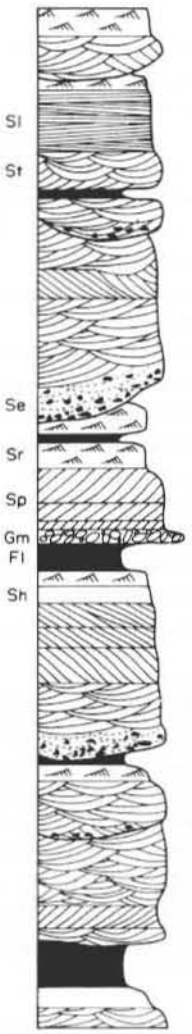


Fig. 4.6. Use of the fluvial lithofacies scheme of Table 4.1 for plotting a section in braided fluvial deposits. (Miall 1978)

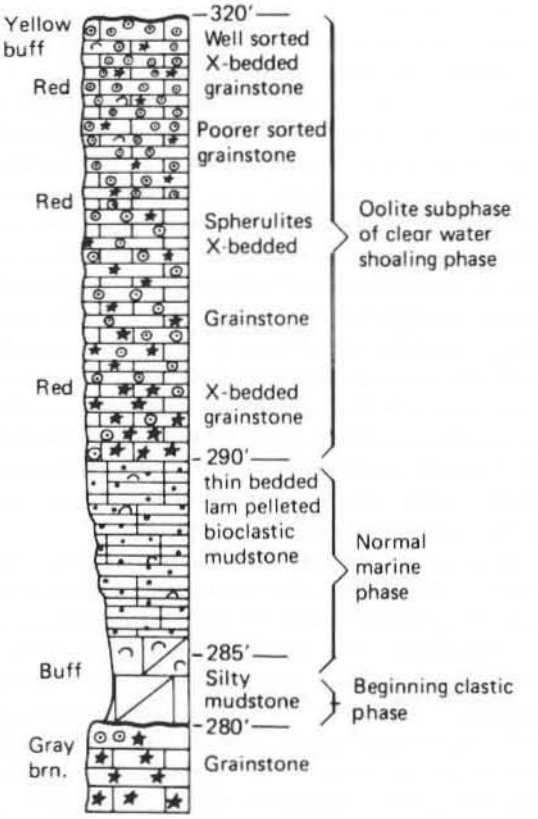


Fig. 4.7. Use of the carbonate microfacies scheme for plotting a limestone section. (J. L. Wilson 1975)

4.3.4 Facies Architecture

Many of the facies-analysis techniques discussed in this chapter focus on the use of vertical facies relationships, as derived from the measurement of stratigraphic sections (Sects. 4.4.2, 4.5.8). Although section measuring has evolved into a sophisticated art, it is essentially a one-dimensional approach to the study of what commonly are very complex three-dimensional (3-D) rock bodies. Increasingly, it has been realized in recent years that vertical sections are inadequate for the description of facies variations in many environments, and techniques are evolving for the description and classification of facies units in three dimensions. At the surface, this can be accomplished by the search for and description of large, laterally extensive outcrops. In the subsurface, lithostratigraphic correlation of closely spaced cores, coupled with high-resolution seismic data (especially 3-D seismic data, if available), may serve this purpose. The results are important not only to facilitate the development of more quantitatively accurate reconstructions of depositional environments but to provide a usable database with which development geologists and engineers can model fluid flow in reservoir units.

Outcrop techniques were described by Anderton (1985) and Miall (1988a,b, 1996), and field examples were given by Allen (1983), Miall and Tyler (1991), and Miall (1988b, 1996). Ideally, the geologist should search for long natural or artificial cliff faces in areas of minimal structural disturbance. The entire outcrop should be photographed to permit the construction of a photomosaic using the techniques described in Section 2.2.2.

Once developed and mounted on card or board, the mosaic can be taken back to the field and used as a base map on which to plot detailed sedimentological observations. These should be entered on a transparent overlay (e.g., mylar). The geologist should exploit the detail available in locations of excellent outcrop to measure and describe as many facies cri-

teria as possible, taking care to enter these on the mosaic with as precise as possible a record of their locations. Locations and orientations of hydrodynamic sedimentary structures and the position and attitude of various bounding surfaces between sedimentary units are among the most important kinds of observations to make. For average road cuts or river cliffs a hundred meters or so long and a few tens of meters high, this fieldwork may take as long as a day or two to complete. For large outcrops, it may be useful to employ the service of an assistant to carry out the actual plotting because, once one is standing on a large outcrop, it may be difficult to see precisely where one is on the mosaic. The assistant can return as closely as possible to the location from which the photograph was taken and enter measurements on the profile that are communicated by voice (or radio) from the geologist on the outcrop.

If necessary, vertical stratigraphic sections can be constructed in the office from such lateral-profile data. However, the wealth of detail yielded by a good-quality lateral profile may make such section drawing superfluous. As noted by Anderton (1985), "Photographic techniques are sufficiently important in facies description that if one ever has to choose between taking a camera and a hammer into the field, take the camera." Figure 2.4 is an example of a lateral profile constructed using these techniques; a brief discussion of the kinds of facies observations that can be made from profiles is presented in Section 4.5.9.

4.4 Facies Associations and Models

4.4.1 The Association and Ordering of Facies

The term **facies association** was defined by Potter (1959) as "a collection of commonly associated sedimentary attributes", including "gross geometry (thickness and areal extent); continuity and shape of lithologic units; rock types ..., sedimentary structures, and fauna (types and abundances)." A facies association (or assemblage) is, therefore, based on observation, perhaps with some simplification. It is expressed in the form of a table, a statistical summary, or a diagram of typical stratigraphic occurrences (e.g., a vertical profile). A **facies model** is an interpretive device that is erected by a geologist to explain the observed facies association. A facies model may be developed at first to explain only a single stratigraphic unit, and similar units may then be studied in order to derive generalized models.

Lithofacies group together into assemblages because they represent various types of depositional events that frequently occur together in the same overall depositional environment. For example, a submarine-fan environment typically contains canyon, channel, levee, overbank, and proximal-slope subenvironments, each of which produces a distinctive lithofacies. These lithofacies become stacked into stratigraphic units because the environments shift through time, permitting different lithofacies to accumulate along any given vertical axis. The nature of these environmental shifts is often predictable, which means that the resulting lithofacies successions are equally predictable. For example, on the submarine fan, channels may shift in position, and subenvironments within a turbidity current may move (very rapidly) downslope. Deltas and tidal flats prograde seaward, river channels and tidal inlets migrate, and so on. These processes may repeat themselves many times. This is the basis for the principle of cyclic sedimentation. It was clearly understood by one of the founders of modern sedimentary geology, Johannes Walther. He enunciated what has come to be known as Walther's Law, the "rule of succession of facies," which states:

> The various deposits of the same facies area and, similarly, the sum of the rocks of different facies areas were formed beside each other in space, but in a crustal profile we see them lying on top of each other ... it is a basic statement of far-reaching significance that only those facies and facies-areas can be superimposed, primarily, that can be observed beside each other at the present time (Walther, 1893–1894, as translated from German by Middleton 1973).

The value of this rule to basin analysts is that it means lateral facies relationships can be predicted by studying them in vertical succession. Stratigraphic sections, such as outcrops measured up cliffs or drill-hole sections, comprise most of our database. Lateral facies relationships are much more difficult to study, because they tend to be gradual and stretched out over considerable distances. Few outcrops are long enough to reveal such features in their entirety. Paleogeographic reconstruction, therefore, relies heavily on the examination of vertical profiles.

It must be stressed that Walther's Law only applies to continuous sections without significant stratigraphic breaks. An erosion surface may mark the removal of the evidence of several or many subenvironments. It is, therefore, important to observe the nature of contacts between lithofacies units when describing sections. Contacts may be sharp or gradational. Sharp contacts may or may not imply significant sedimentary breaks. Periods of erosion or nondeposition may be revealed by truncated sedimentary structures, extensive boring, penecontemporaneous deformation, hardgrounds, or submarine diagenesis.

Given this constraint, the study of cyclic sedimentation has been one of the most popular and fruitful in sedimentology because, as Reading (1986) stated, "it enabled geologists to bring order out of apparent chaos, and to describe concisely a thick pile of complexly interbedded sedimentary rocks." Interpretations of cyclic sedimentation form the basis of most facies models.

Statistical methods have been used for defining and describing lithofacies and biofacies associations. Cluster-analysis methods were used by Imbrie and Purdy (1962) and Purdy (1963) to document and describe the associations of facies components on the modern Bahama Platform (Figs. 4.8, 4.9). Markov-chain analysis has been used to explore the vertical ordering of clastic lithofacies, particularly in studies of fluvial deposits (Miall 1973; Harper 1984; Fig. 4.10). Multivariate statistical methods have also been used in the erection of biofacies associations (Mello and Buzas 1968; Ludvigsen 1978). These methods were described at some length in the second edition of this book, but this material has been omitted here because such techniques, having served their purpose, have now fallen into disuse. After more than 30 years of intensive research, the major facies associations in the most important sedimentary environments have now been recognized (Walker and James 1992; Galloway and Hobday 1996; Reading 1996), and such exploratory techniques are no longer needed. Attention has shifted to the investigation of the 3-D attributes of sedimentary successions (including architectural-element analysis) at the outcrop scale (Sect. 4.5.9), and sequence stratigraphy at the basin scale (Chap. 6). The following is a brief summary of two statistical methods that were popular in the 1960s and 1970s.

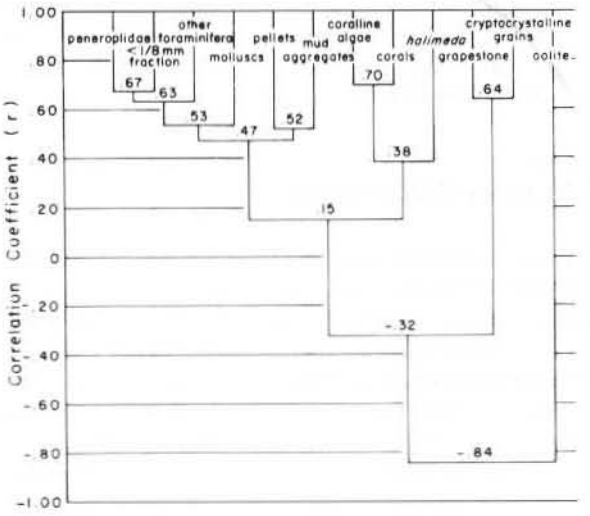


Fig. 4.8. Cluster analysis (R-mode) of facies components on the modern Bahama Platform. (Purdy 1963)

Imbrie and Purdy (1962) and Purdy (1963) carried out a landmark study of carbonate facies that involved a statistical analysis of data from modern sediments of the Great Bahama Bank. Under the microscope, Purdy performed point-count analyses of 218 samples from the northern part of this bank. He recognized 12 major organic and inorganic components: coralline algae, *Halimeda* (calcareous alga), Peneroplidae (a family of Foraminifera), other foraminifera, corals, molluscs, fecal pellets, mud aggregates, grapestone, oolite, cryptocrystalline grains, and silt grains under 1/8 mm in size. Cluster analysis was then used to determine which components tended to occur together; the results are shown in Fig. 4.8. From this information, five major lithofacies were defined and, by referring back to the sample data, these could be mapped as shown in Fig. 4.9. Purdy (1963) interpreted the lithofacies data in terms of physical and chemical processes operating over the bank. The work has been widely used and quoted as an example of a lithofacies study in carbonate rocks.

There has been considerable controversy over the methods used to define sedimentary cycles. Early attempts were subjective and qualitative. The culmination of the analysis was commonly the erection of a single model or ideal cycle (Visher 1965). This was criticized by Duff and Walton (1962) on the grounds that, in fact, this ideal cycle may rarely be represented in the section. Random environmental changes, local reversals of the sequence, or the effects of some external event, such as a flash flood or tectonic pulse, commonly intercede to modify the succession. The many analyses of the Carboniferous coal-measure cyclothems prior to about 1960 illustrate this difficulty well but cannot be entered into here for reasons of space (modern ideas about these cycles are discussed in Chap. 8). The technique of Markov-chain analysis was popular in the 1970s for the analysis for cyclicity. This has the virtue of a certain statistical rigor. Like the multivariate techniques described in the preceding paragraph, it results in a grouping of lithofacies into an assemblage, but it also has the additional property of revealing the order in which the lithofacies tend to occur. It can reveal not only a single-model cycle, but also the several or many statistically most probable variations on this theme. A Markov processes is one "in which the probability of the process being in a given state at a particular time may be deduced from knowledge of the immediately preceding state" (Harbaugh and Bonham-Carter 1970). The methods have been described by Miall (1973) and Harper (1984).

The analysis starts with a transition-count matrix. This is a two-dimensional array that tabulates the number of times all possible vertical lithofacies transitions occur in a given stratigraphic succession. A

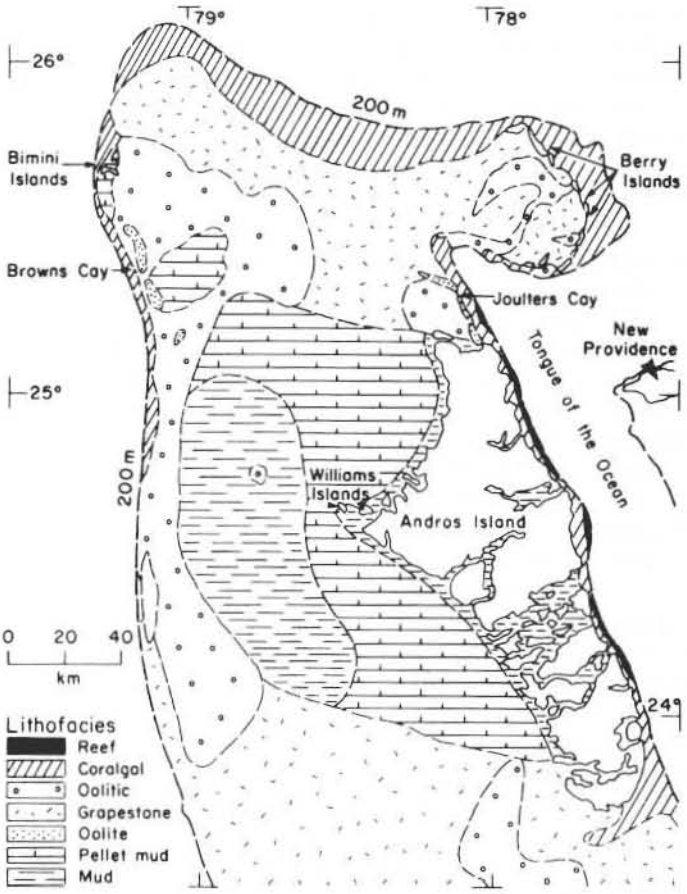


Fig. 4.9. Distribution of lithofacies of the Bahama Platform. (Purdy 1963)

series of simple mathematical operations, easily performed on a hand-held electronic calculator, generates a table showing the probabilities of each lithofacies transition and, from this, it is generally a simple exercise to build up a path diagram showing lithofacies relationships, as shown in Fig. 4.10.

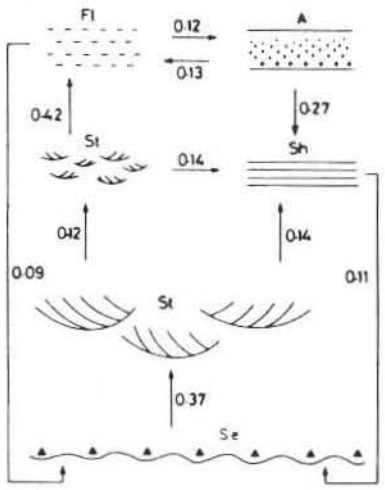


Fig. 4.10. Path diagram for a Devonian fluvial unit in the Canadian Arctic. (M. Gibling, in Rust 1978)

There is a temptation to use the Markov-chain method to reduce complex stratigraphic data to single, simple, cyclic models, such as Cant and Walker's (1976) "summary sequence." This can be very misleading, because it repeats the old mistake noted by Duff and Walton (1962) of concentrating on "ideal" or "model" cycles. This is not the object of the exercise. The strength of the method is its ability to show quantitatively how lithofacies relationships vary from such models.

4.4.2 The Theory of Facies Models

The concept of a facies model as a summary of a depositional environment and its products is a recent one in sedimentology. The term was first discussed at a conference reported by Potter (1959), but was used in the sense we now imply by the term facies assemblage. The difference is critical. A facies assemblage is essentially descriptive, whereas a facies model attempts to provide an interpretation of a particular type of facies assemblage in terms of depositional environments. Much use is usually made of comparisons with actual modern environments, so it is now

common practice to refer to these mental constructs as "actualistic facies models" [although Shea (1982) has criticized the use of the word "actualistic", partly on etymological grounds].

Walker (1992) described the process of constructing a facies model, using submarine fans as an example. Data from various sources (modern sediments, ancient deposits) are collected and the wealth of information sifted to determine the important features that they have in common, The local variability is "distilled" away (Fig. 4.11). Various statistical procedures may be used to achieve this, such as those discussed in the previous section. The objective is to produce a general summary of submarine fans. But, as Walker (1992) stated, "what constitutes local detail, and what is general? Which aspects do we dismiss, and which do we extract and consider important?"

Walker stated that a facies model should fulfill four functions:

1. It must act as a **norm** for purposes of comparison
2. It must act as a **framework** and **guide** for future observations
3. It must act as a **predictor** in new geological situations
4. It must act as a basis for **environmental interpretation**

This is illustrated in Fig. 4.11. Facies models are a powerful tool for the interpretation of poorly exposed sediments, because they suggest certain critical observations or clues for interpreting the sedimentary record. As with all such summaries and models, however, uncritical use may lead to loss of information or misinterpretation, because it is tempting to observe strata in terms of a preconceived model. If a facies model is being used properly in new field situations, each use may generate a refinement of the model, or it may lead to the recognition of situations where the particular model is inappropriate. The development of a new model could then follow.

What does a facies model look like? The answer depends on how the model was constructed. Paleogeographic sketches, block diagrams, and vertical profile logs are typical components of a published facies model. Some models are based primarily on

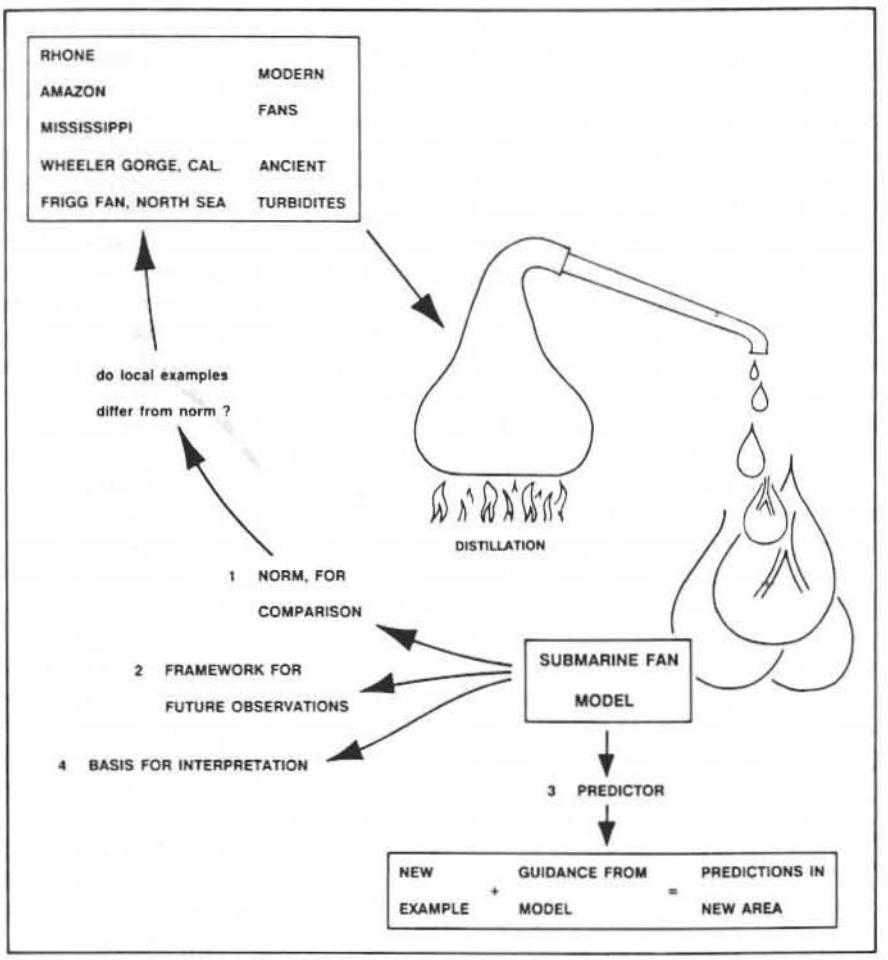


Fig. 4.11. The development of a facies model, using submarine-fan deposits as an example. (Walker 1992)

geomorphology (e.g., deltas, alluvial fans), others on transport processes (e.g., turbidite, glacial till, and eolian dune models), and others on organic processes (e.g., reefs). Many are governed by climatic and tectonic variables, which have not, in all cases, been exhaustively explored.

Many models are based primarily on the study of modern environments, for example, the three-fold subdivision of deltas erected by Galloway (1975) used maps of typical deltas to convey information on lithofacies geometry. For continental environments, there is no problem seeing and describing a wealth of detail in modern settings, and this has led to the recognition that, for fluvial deposits, at least sixteen distinct fluvial styles can be described (Miall 1996). Only in the last four or five years has the same stage been reached for eolian environments, as noted in Section 4.6.4. Rapid evolution in eolian facies models has been brought about by improved understanding of eolian bedforms (Sect. 4.5.4).

Facies models for subaqueous environments are, at present, only loosely based on studies of modern settings, because the techniques of marine geology (geophysical profiling, shallow coring) have not, until recently, permitted very precise descriptions of lithofacies geometry and relationships. The first comprehensive submarine-fan model of Mutti and Ricci-Lucchi (1972) was based mainly on well-exposed Cenozoic fan deposits in Italy, and interpretations of other ancient deposits figure prominently in the first detailed discussions of this environment (Walker 1978). Models for storm-dominated shelf deposits incorporate HCS (Hamblin and Walker 1979), a structure not reliably documented from modern environments until the 1980s. Howard (Byers and Dott 1981) explained how it may be obscured in cores of recent sediments by our methods of sampling the sea bottom and offered what may be the first observation from a modern offshore region (Howard and Reineck 1981). Application of side-scan sonar techniques to the study of submarine depositional systems is now bringing about something of a revolution in our knowledge of shelf and deep-marine environments, yielding knowledge that is beginning to have a major impact on facies models of channels, bedforms, and other depositional features (Bouma et al. 1985).

Many of these points are reviewed at greater length in Section 4.6, but this summary should serve to illustrate the varied state of the art of facies modeling at the present day. It is very much a case of work in progress and, by careful studies of newly explored stratigraphic units, basin analysts can potentially add much to our fund of knowledge.

4.4.3 The Present as the Key to the Past and Vice Versa

We have all heard the cliche "the present is the key to the past." It was one of the great generalizations to emerge from James Hutton's enunciation of the principle of uniformitarianism toward the end of the 18th century. Charles Lyell's work a half century later seemed to nail it down forever as a cornerstone of the still-unborn science of sedimentology. Yet it is only true in a limited sense. Undoubtedly, the study of modern depositional environments provides the essential basis for modern facies studies, but there are at least a dozen major problems that emerge to confuse this work. It turns out that the past is also a very important key to the present, that many geomorphological processes can best be understood by adopting a geological perspective and looking at the ancient record. The art of facies modeling is, therefore, a two-way process.

The greatest advantage of studying modern environments is that we can observe and measure sedimentary processes in action. We can measure current strength in the rivers and oceans; we can observe at least the smaller bedforms moving and evolving; we can measure temperatures and salinities and study the physics and chemistry of carbonate sedimentation and the critical effects of organic activity in the photic zone (shallow-depth zone affected by light penetration in the sea); we can sample evaporite brines in inland lakes and deep oceans, such as the Red Sea, and deduce the processes of concentration and precipitation; we can drill shallow core holes and study the evolution of the sedimentary environment through the Recent to the present day. Many physical, chemical, thermodynamic, and hydrodynamic sedimentary models have resulted from this type of work since the 1960s. As noted in the previous section, for some sedimentary environments, work of this type has constituted the major advance in our understanding. Examples of early facies models that depended largely on studies of modern environments include those for deltas (Coleman and Wright 1975; Galloway 1975), barrier islands (Bernard et al. 1959; Dickinson et al. 1972), tidal inlets (Kumar and Sanders 1974), clastic tidal flats (Van Straaten 1951, 1954; Evans 1965; Reineck and Singh 1980), tidal deltas (Hayes 1976), and sabkhas (Kinsman 1966, 1969; Shearman 1966).

Some of the most difficult problems to resolve when using actualistic models stem from the inadequate time scale available to us for observation purposes. A very persistent geologist may be stubborn enough to pursue the same field project for 10 years, perhaps even 20 years. Aerial photographs might push observations of surface form to as far back as

about 1920, at the earliest. Old maps may go back for another 100 years or more but become increasingly unreliable. Weather records may be available back into the 19th century; stream-gauge data have been collected for only a few decades. It is difficult to assess the relevance of a 100-year record to a geological unit that may have taken a million or more years to accumulate. Have the last 100 or so years been typical? Were the same sorts of processes occurring at the same rates in the distant geological past? As discussed in the following paragraphs, the answer is a qualified maybe.

The most important aspects of this question are the difficulties of assessing the importance of ephemeral events and judging the preservation potential of deposits we can see forming at the present day. Dott (1983) and Reading and Levell (1996) have called attention to the need to distinguish between normal and catastrophic sedimentation. Normal processes persist for the greater proportion of time. Net sedimentation is usually slow. It may be nil or even negative if erosion predominates. Normal processes include pelagic settling, organic growth, diagenesis, and tidal and fluvial currents (Reading and Levell 1996). Reading and Levell (1996) distinguished catastrophic processes as those that "occur almost instantaneously. They frequently involve 'energy' levels several orders of magnitude greater than those operating during normal sedimentation." They may deposit a small proportion of the total rock and give rise to only an occasional bed, or they may deposit a large proportion of the total rock and become the dominant process of deposition. Examples of catastrophic processes include hurricanes, flash floods in rivers, and sediment-gravity flows. Although geologists have studied many modern flash-flood deposits and the effects of several recent hurricanes (Hayes 1967), we cannot be sure that their magnitude and frequency at the present day are the same as those in some past period of interest without attempting to obtain some geological perspective by studying the ancient record. The most violent and geologically important event may be one that only occurs every 200 years or 500 years and has not yet been seen. Sediment-gravity flows are thought to be the chief agent of erosion of submarine canyons but, in spite of years of oceanographic observation, no major flows have actually been observed there (Shepard et al. 1979).

Ager (1981) and Dott (1983) argued that many stratigraphic sequences contain more gaps than record (Sect. 3.7.7) and that significant proportions of the sedimentary record are deposited in a very short time by particularly violent dynamic events. Such events (hurricanes, sediment-gravity flows, flash floods, etc.) are rare and difficult to study in action. Most of our energy as sedimentologists is expended in studying the less violent processes that occupy most of geological time but may contribute far less sediment to the total record.

Studies of modern environments also suffer from the fact that many of the deposits we see forming at the present day are lost to erosion and never preserved. Thus, the geological record may be biased. The bias may be in favor of more deeply buried sediments, which the geologist, scratching the surface with trenches and box cores, never sees. For example, Picard and High (1973) published a detailed study of the sedimentary structures of modern ephemeral streams, but many of the structures are sufficiently ephemeral to be rarely, if ever, found in the ancient record. Fluvial flash-flood deposits commonly are preferentially preserved, because they infill deep scours below the normal level of fluvial erosion (Miall 1996; Chap. 8). Many of our fluvial facies models are based on studies of modern rivers in upland regions undergoing net degradation. How relevant are they to research in some of the great, ancient, alluvial basins that, by the thickness of preserved deposits, demonstrate a long history of aggradation? Facies models for barrier and shoreface environments are subject to similar constraints. For years, the classic Galveston Island (Texas) model of coarsening-upward beach-accretion cycles dominated geological thinking (Bernard et al. 1959) but, more recently, it was realized that the barrier sediments may be removed by lateral migration of deep tidal inlets and that many barriers consist of superimposed inlet deposits with a superficial skin of wave-formed shoreline sediments (Kumar and Sanders 1974). Hunter et al. (1979) argued that many shallow, subtidal deposits are systematically removed by rip currents and are never preserved in the geological record.

Some of the best studies of modern environments are those that use shallow drill cores to penetrate into pre-Recent deposits. Such sediments can be said to have "made it" into the geological record, and yet they can be placed in the context of a still-extant and, presumably, little-modified modern environment. Some work on modern turbidites (Bennetts and Pilkey 1976), anastomosed rivers (Smith and Smith 1980), and reefs (Adey 1975; Adey et al. 1977; Shinn et al. 1979) is of this type.

In two important ways, the present is quite unlike the past and is, therefore, an unreliable laboratory for reconstructing sedimentary environments. The Pleistocene ice age generated several geologically rapid changes of sea level, culminating in a major rise and transgression during the last 10,000 – 20,000 years. Second, the present configuration of continents and oceans is a unique pattern, different from any in the past because of the long history of sea-floor spreading, rifting, subduction, and suturing. This plate

movement has had an important effect on some of the broader aspects of facies models (Sect. 9.1).

Because of the recent sea-level rise, modern continental-shelf, shallow-marine, and coastal-plain deposits around the world have been formed under transgressive conditions. Sea level was approximately 150 m lower during the Plio-Pleistocene glacial phases, rivers graded their profile to mouths located near the edge of present continental shelves, and carbonate platforms, such as the Bahamas, were exposed to subaerial erosion and may have developed extensive karst systems. Submarine canyons were deeply entrenched by active subaerial and submarine erosion (McGregor 1981; Shepard 1981). During the rapid transgressions that followed in interglacial phases, shoreline sands were continually reworked. On the Atlantic shelf off North America, extensive barrier islands were formed and receded into their present position (Swift 1975a). River valleys were drowned, and filled with estuarine and deltaic deposits. Submarine canyons were commonly deprived of their abundant supply of river-borne detritus, as this was now deposited on the landward side of a widening and deepening continental shelf. This had a drastic effect on the rate of growth of some submarine fans. Carbonate sedimentation began afresh in warm, detritus-free waters over resubmerged platforms, but the local water depths, circulation patterns, and facies distribution may have been partly controlled by erosional topography (Purdy 1974; although this is disputed by Adey 1978). Therefore, we have excellent modern analogues for studies of rapid transgressions in the geological record, but few for rapid regressions or periods of stillstand. The Mississippi and other large deltas are good examples of regressive deposits built since the last ice age, but they represent an environment that may be characterized by unusually rapid progradation. Some have maintained that most modern continental shelves are covered by relict sediments, implying that their study may not be of much geological relevance (Emery 1968), but more-recent work has shown that the dynamic effects of tidal currents and storms do, in fact, result in continual change (Swift et al. 1971).

Rapid transgressions may have occurred commonly during other ice ages in the geological past, as discussed in Chapter 8, but most other changes in the geological past were somewhat slower. For two lengthy periods in the Phanerozoic, our suite of modern analogues is particularly inadequate. These were the times of high global sea-level stand during the Ordovician–Silurian and again in the Cretaceous, when vast areas of the world's continents were covered by shelf seas. We simply do not have modern equivalents of these huge inland seas, and many of the large shelf seas that do exist, such as the Bering Sea and Yellow Sea, have yet to be studied in detail. The tidal range may have been amplified over these shelves, and wave energy may have been reduced by friction over the shallow bottom. R.N. Ginsburg (Byers and Dott 1981) discussed the problem of interpreting Cambro-Ordovician carbonate banks of the North American craton. These are up to hundreds of kilometers long and thousands of kilometers wide. "The dilemma is how the vast extent of the banks, most of which suggest carbonate production and deposition in but a few meters or less of water, could all be bathed in normal marine or slightly restricted water. Would not such banks form major circulation barriers?" (Ginsburg, as reported by L.C. Pray; Byers and Dott 1981). Perhaps the banks are actually diachronous and developed by seaward progradation, or they may have been crossed by "irregular to channelized deeper water areas facilitating water circulation." Tidal currents certainly would have been able to assist with the latter. We return to this point in Section 6.6.

Climatic patterns and the network of oceanic currents are controlled by global plate configurations. In many respects, our present geography is unique. Therefore, we have modern analogues for situations that may not have existed in the past and, conversely, we cannot replicate certain conditions that did exist. For example, the Mesozoic Tethyan Ocean and the Pangea supercontinent had profound effects on climate and water circulation and, hence, on sedimentation patterns, and we have only generalized models for interpreting them. Some of these broader effects on sediment facies are considered in Chapters 8 and 9. The study of paleoclimates is beyond the scope of this book but is considered in an excellent review by Frakes et al. (1992).

The last group of problems with the actualistic modeling method arises from the important effects organisms have on sedimentary processes. Plants and animals have, of course, both evolved profoundly since Archean times. Therefore, in many environments, sedimentation is controlled or modified by sediment–organism relations that did not exist, or were different, in the geological past. Many authors, beginning with Schumm (1968), have speculated on the implications of the evolution of land plants on fluvial patterns. Vegetation has a crucial effect on stabilizing channel banks, colonizing bars and islands, and controlling chemical weathering, sediment yield, and discharge fluctuations. Our typical braided and meandering fluvial facies models reflect these effects as they have been studied in temperate, humid, and hot and cold arid climates. Fluvial facies models seem likely to be quite different for thickly vegetated, humid, tropical climates (Miall 1996) but have yet to be studied in detail. Going back in time, the evolution of

grasses in Mesozoic time must have changed geomorphic patterns profoundly. Abundant land vegetation is thought to have first appeared in the Devonian and, prior to that time, the majority of river channels may have been unconfined, ephemeral, and braided, as in modern arid regions. Modern deserts are commonly used as analogues for pre-Devonian rivers, but there is no reason why they had to be arid. However, we have no modern analogue for a humid, vegetation-free environment with which to study the pre-Devonian except, possibly, the south coast of Iceland.

Turning to shallow-marine environments, the same kinds of difficulties apply. H. E. Clifton (Byers and Dott 1981) pointed out that salt-marsh vegetation, so important on modern tidal flats, did not exist prior to the Cenozoic. Similarly, foraminifera, which provide sensitive bathymetric indicators for younger sediments, particularly those of Cenozoic age, did not exist in the Paleozoic. James and Kendall (1992) discussed the implications of these various evolutionary developments on changes in carbonate-depositional environments.

The ecology of forms that are now extinct may be difficult to interpret, which makes them less useful in facies studies (Sect. 4.5.7). Instead of providing independent, unambiguous environmental information, as do still-living forms, it may be necessary to interpret them with reference to their sedimentary context, which itself may be of uncertain origin. Functional morphology is studied to determine probable habits, but many uncertainties may remain. In the Precambrian, most sedimentary units are entirely devoid of fossils and, here, problems of lithofacies interpretation without supporting fossil evidence may become acute. Many Precambrian units have been reinterpreted several times for this reason as different environmental criteria are brought to bear on particular problems. Long (1978) discussed many of these difficulties with reference to the recognition of fluvial deposits in the Proterozoic.

A thorough review of Precambrian clastic sedimentation systems by Eriksson et al. (1998) documented several major differences in the preserved Precambrian record relative to that of the Phanerozoic. Marine-shelf deposits are characterized by very uniform suites of sediments lacking distinctive vertical trends. Fluvial deposits are dominated by those of braided style. Foreshore deposits are rare, and eolian dune deposits appear to be absent in rocks older than about 1.8×10^9 years.

4.4.4 To Classify and Codify or Not?

We now know, in the 1990s, a great deal about the origins of sedimentary rocks and have amassed a considerable published record of case studies documenting sedimentary processes, facies and facies assemblages in a wide variety of modern settings, plus a wealth of detailed descriptions and interpretations of ancient units. These data are in danger of overwhelming the practitioner, and summaries and classifications are more needed than ever.

Stephen Jay Gould (1989) stated: "Classifications are theories about the basis of natural order, not dull catalogues compiled only to avoid chaos." In a descriptive science like geology, there will always be a need for classifications. These serve two purposes: they attempt to create order out of apparent chaos as aids to memory, and they are an attempt to understand genesis. The best classifications are those that highlight differences in origins of geological features and are, therefore, particularly useful for purposes of prediction and extrapolation, for example, in geological exploration.

There will always be a creative tension between those who focus on the minute differences among geological attributes and those who like to simplify and generalize. This is the tension between those who seek constantly to perfect our understanding of descriptive detail and of processes (the purists), and those who think they know enough details for the time being and want to move onto larger problems (the synthesizers). It is the same as the debate among taxonomists between the "lumpers" and the "splitters". Facies classification and facies models offer plenty of fuel for debate between the advocates of both sides of this argument. Anderton (1985) provided an excellent discussion of the various levels of complexity that are embraced by the modeling process (also Pirrie 1998).

As proof of the failure of the idea of classification, the purists like to point to classifications that have failed. For example, Bridge (1993) used Mike Leeder's (personal communication) example of the geochemical classification of igneous rocks proposed by Shand (1947), a complex system that did not catch on and was not used, no doubt because it was not based on any understanding of genetic reality that withstood the test of time. Likewise, Kay's (1951) classification of geosynclines was widely used until the 1960s, at which time it virtually vanished, to be replaced by classifications based on plate tectonics. Ingersoll and Busby (1995) now propose 27 varieties of sedimentary basin, based mainly on plate-tectonic principles– more types than there were geosynclines. Is this being objected to? No; in fact the contrary (Dickinson 1997), because it is now recognized that, at least according to current plate-tectonic concepts, this is a genetically based classification and, therefore, has deep meaning, much like a phylogenetic classification of organisms. Basin classifications are discussed in Chapter 9.

Bouma turbidite should look like that we can readily appreciate how far off the track many sedimentological descriptions and interpretations have strayed, when someone like Shanmugam comes along and brings us up short with careful, detailed observation. It turns out that most deep-marine sands may not be turbidites at all, but this does not prove that Bouma was wrong or that the concept of the facies model is wrong, only that these elegant concepts have been applied too carelessly.

Sequence stratigraphy, both as a body of concepts for the teaching of stratigraphy and as a research tool, has been heavily dependent on models, especially the by-now virtually classic papers of Posamentier et al. (1988), Posamentier and Vail (1988), and Van Wagoner et al. (1990). Yet this area, too, has been bedeviled by the misunderstanding and misuse of models, to the extent that the original proponents of the models have had to become apologists for them, pointing out, in such overview papers as Posamentier and James (1993) and Weimer and Posamentier (1993), that the models were never intended to be of universal application.

There is no doubt that classifications and models in sedimentology have been used as research tools in too facile a manner by many workers, but this does not mean that the approach is wrong, only that all due caution is required in the application of the models. Classifications and models that genuinely reflect variations in the genesis of sedimentary units are likely to survive, thrive, and be used, possibly with numerous revisions and accretions. Those that do not will be ignored.

4.4.5 Facies Analysis and Sequence Stratigraphy

As noted by Nummedal (1987),

> The interpretation of the sedimentary geological record has been greatly stimulated over the past few years by rapid conceptual advances in "sequence stratigraphy", i.e., the attempt to analyze stratigraphic successions in terms of genetically related packages of strata. The value of the concept of a "depositional sequence" lies both in the recognition of a consistent three-dimensional arrangement of facies within the sequence, the facies architecture, and the regional (and inter-regional) correlation of the sequence boundaries. It has also been argued that many sequence boundaries are correlatable globally, and that they reflect periods of sea-level lowstand, i.e., sequence boundaries are subaerial erosion surfaces.

Sequence stratigraphy is not a radically new concept. For example, many of the sedimentological ideas were discussed by Frazier (1974), and the first assemblage of sequences, in the modern sense, was established by Sloss et al. (1949). However, the study of sequences has received considerable impetus since developments in seismic stratigraphy in the late 1970s reminded geologists of the ubiquity of sea-level changes throughout the geologic past.

Sequence stratigraphy is a combination of lithostratigraphy, allostratigraphy, and facies analysis. The vertical arrangement of facies is interpreted in terms of Walther's Law, leading to reconstructions of the lateral movement through time of depositional environments (Chap. 6). Sequences typically consist of transgressive and regressive half sequences that together comprise repeated cyclic successions. Up to this point, the study of sequences does not differ greatly from the analysis of vertical profiles for the purpose of documenting autogenic depositional events, for example, the progradation and lobe switching of a deltaic coastal-plain complex. However, as noted by Nummedal, careful regional lithostratigraphic, allostratigraphic, and chronostratigraphic correlation demonstrates that, typically, the sequences extend laterally for much greater distances than could be attributed to autogenic controls. The influence of a regional or global allogenic control, such as changes in relative sea level, may then be suspected (Chap. 8).

4.5 Review of Environmental Criteria

In Chapter 2, a summary of major environmental indicators is given under the heading of what to look for in outcrop and subsurface sections. This list is by no means exhaustive; for example, it concentrates almost exclusively on what can be seen with the naked eye. However, it includes many of the sedimentary features vital to a generalized environmental interpretation and, hence, an effective basin analysis. In this section, some indications are given of how to use this information. Beginners at the art of basin analysis or those who wish to understand what their specialist colleagues are up to may find this section both enlightening and confusing. The numerical precision or well-defined statistical error of laboratory-oriented geological studies is not to be found in the area of facies interpretation. Statistical methods may be used to aid in analyzing the composition of facies assemblages, but the business of interpreting their meaning depends heavily on qualitative study. The sedimentologist must be aware of the meaning and limitations of all the facies criteria visible and must be able to weigh the evidence of all these against each other. A good knowledge of published facies models and modern analogues is, of course, essential.

A common basin-analysis problem occurs when the geologist is faced with a new outcrop or core showing certain assemblages of lithologies, textures, structures, and, perhaps, fossils and may have no idea which facies model to turn to for assistance in making an environmental interpretation. The intent of this section is to review very briefly the types of interpretation that can be made from the principal kinds of sedimentological observation in order to provide some environmental clues and an entry into some of the crucial literature. Discussions of the physical and chemical conditions of formation of most sediment types are beyond the scope of this book.

In some of the notes that follow, siliciclastics and carbonates are treated separately, because they require a different approach. This is particularly the case with interpretations of grain size and texture. For others, such as hydrodynamic sedimentary structures, there is no reason not to consider all rock types under the same heading. Structures such as crossbedding and ripple marks are commonly regarded as the domain of the sandstone specialists, but they also occur in carbonates and evaporites. Chemical sediments are often studied by geologists whose first interest is in the chemistry of their formation and diagenesis but who could usually benefit from the approach taken to sedimentology by clastic sedimentologists.

The emphasis throughout is on features that are most useful for constructing depositional environments and paleogeography. Diagenesis and geochemistry are not dealt with in this book.

It cannot be overemphasized that very few sedimentological criteria have an unambiguous environmental interpretation. Many years ago, it was thought that ripple marks only occurred in shallow water, until oceanographers started taking photographs of the bottoms of the oceans. More recently, dish structures were thought to be indicators of submarine grain flows but have now been found in fluvial and other deposits (Nilsen et al. 1977). The common trace fossil *Ophiomorpha* is a good indicator of shallow-marine environments, but J. Coleman (pers. comm., 1976) reported finding it many miles inland, in the deposits of the modern Mekong River. Marine it certainly is, but the Mekong River has an extensive marine salt wedge that flows far upstream during high tide. *Ophiomorpha* has also been found in outer shelf environments (Weimer and Hoyt 1964). Reliance should never be placed on a single structure or feature of the rocks or on a single method of analysis for making environmental interpretations. The geologist must review and assess all the evidence available. Detailed descriptions of actual facies models are not given here but, in Section 4.6, a review of some of the most recent developments in the area is offered for more advanced students.

4.5.1 Grain Size and Texture

In a general sense, the grain-size of a clastic sediment indicates the relative amount of energy required to emplace the grains in their final resting place. This energy might have been derived from the force transmitted by air or water movement, or it may represent downward movement under gravity. Most clastic sediments represent a combination of both of these processes.

Grain-size interpretation in siliciclastic rocks tends to be much simpler than it is in carbonates and evaporites, because diagenesis in these chemical sediments frequently obscures original grain relationships. Grain size may be related entirely to in-place primary or diagenetic crystal growth and not at all to transport processes. The size of allochems in carbonate sediments is commonly determined by their organic origins. However, some chemical sediments show evidence of having behaved as clastic detritus at some stage in their formation, so that considerations of grain dynamics might provide useful environmental information.

There are two ways in which general grain-size data yield environmental information in siliciclastic sediments. Local vertical variations in mean or maximum grain size are frequently cyclic or rhythmic, and these, coupled with variations in sedimentary structures, are powerfully diagnostic. Analysis of vertical profiles for cyclic patterns was discussed earlier in this chapter (Sect. 4.4.1) and is dealt with in greater detail in Section 4.5.8. The second aspect of grain-size information is the environmental information contained in the size distribution of individual samples. This subject has received an enormous amount of attention from sedimentologists, who have proposed a wide variety of statistical techniques for distinguishing the deposits of various depositional environments on the basis of some supposed environmental signature retained in the sample. Most of this attention has been directed toward the study of sandstones. It has long been recognized that the hydraulic sorting effects of waves, wind, and unidirectionally flowing water result in the movement of different populations of grains. Most sandstones are mixtures of several populations, but it has long been the hope that the right analytical methods would infallibly recognize these subpopulations based on some size or sorting criteria, enabling the depositional environment to be recognized. All such techniques depend on careful laboratory size analyses, preferably carried out using sieves, a setting tube or, failing this in the event of lithified samples, counts of grains in thin sections. Various methods were discussed by Visher (1969), Solohub and Klovan (1970), Glaister and

Nelson (1974), Friedman and Sanders (1978), and Friedman (1979). None of these methods is fully reliable because of the problem of inherited size distributions in the case of second-cycle or multicycle sands and the effects of diagenesis (cementation, secondary porosity development) on lithified sandstones. Extensive laboratory work and data analyses are required to complete a rigorous grain-size analysis. Even then, the results are usually ambiguous and, in this writer's opinion, do not justify the great effort expended. Accordingly, the use of this tool is not normally recommended. The method does, perhaps, have some uses in the analysis of well cuttings or sidewall cores where these are the only samples available, because the technique can be applied to small samples, whereas most of the observational techniques discussed here cannot. Even here, visual inspection of petrophysical logs may provide equally reliable environmental information in a fraction of the time.

A useful technique in the study of conglomerates is the measurement of maximum clast size. Normally, this is determined by averaging the intermediate diameter of the ten largest clasts present at the sample level. Such measurements have been used in the study of grading and cyclic changes in subaerial alluvial fans (Steel 1974; Gloppen and Steel 1981) and in subaqueous, resedimented conglomerates (Nemec et al. 1980). It is found that there is a direct relationship between maximum particle size and bed thickness for deposits formed by subaerial or subaqueous debris flows, indicating that the beds were formed by single depositional events without subsequent reworking (Bluck 1967; Nemec et al. 1980).

Grain size, fabric, and texture are, together with sedimentary structures and vertical profiles, important criteria for distinguishing flow type in the study of sediment-gravity flows and their deposits. The recognition of particular grain-size populations in a deposit, the internal sorting and grading of each population, and the vertical and lateral relationships of the populations to each other have been interpreted in terms of grain-support mechanisms and their evolution during the passage of flow events (Middleton and Hampton 1976; Lowe 1979, 1982; Postma 1986). Magnetic fabrics are now used to help distinguish sediment-gravity-flow deposits from other types of deposits. For example, Shor et al. (1984) used this technique in an attempt to distinguish turbidites from contourites. Gravenor and Wong (1987) and Eyles et al. (1987) employed magnetic-fabric data in studies of glacigenic deposits (see below).

Grain-size and textural studies in conglomerates were used by Walker (1975) to suggest models of deposition on submarine fans (Fig. 4.12). The disorganized-bed model is thought to represent rapid, clastic deposition on steep slopes, perhaps in the feeder submarine canyon at the head of a fan. Debris flows passing to the inner fan are thought to exhibit first an inversely to normally graded texture, passing downcurrent into a graded–stratified type. Inverse grading develops as a result of dispersive pressure, possibly including a "kinetic sieve" mechanism, whereby smaller clasts fall down between the larger ones. Hein (1982) has extended this work to pebbly and massive sandstones.

Similar studies of texture and fabric in glacigenic deposits have helped to distinguish the various mechanisms of formation of poorly sorted conglomerates. Early workers referred to all glacigenic conglomerates as **till** (tillite is the lithified variety), a deposit formed by the accumulation of unsorted debris below or in front of grounded glacial ice. However, careful facies analysis of many Quaternary and more ancient glacial deposits has shown that most are water laid (Sect. 4.6.10). Analyses of clast fabric and the magnetic fabric of the matrix are now being used as a supplement to outcrop-facies studies in order to determine depositional processes (Eyles et al. 1987; Fig. 4.75). Glacial conglomerates may have originated as tills but moved down local depositional slopes as slumps or sediment gravity flows or by the accumulation of material moved into the basin by ice rafting.

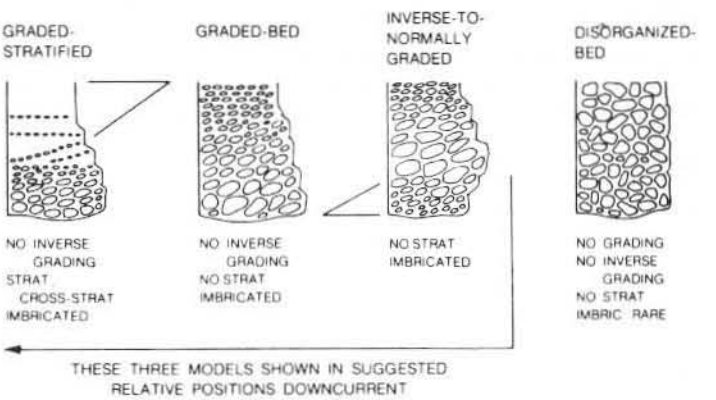


Fig. 4.12. Stratigraphic–textural-fabric models for resedimented conglomerates. (Walker 1975)

Fabric studies can help distinguish these processes (Fig. 4.75). Eyles et al. (1983) suggested referring to all poorly sorted conglomerates by the nongenetic term **diamictite** (unlithified variety: diamict or diamicton), and they erected a lithofacies classification for diamictites that places primary emphasis on the presence or absence of grading and stratification and the density of clast packing (Sect. 4.6.10). Schultz (1984) applied this same classification to the analysis of diamictites formed in a subaerial alluvial-fan setting.

In the conglomerate deposits of alluvial fans and other gravelly rivers, grain-size variations commonly reveal crude stratification and large-scale crossbedding (Sect. 4.5.4). The stratification is the product of longitudinal-bar growth or superimposition of debris flows. Individual bar deposits may show an upward decrease in grain size (Miall 1996). True grading is rare in fluvial gravels formed by traction transport.

Many resedimented conglomerates and pebbly sandstones contain a clast fabric with the long (a) axis dipping upstream, indicating that they were deposited from a dispersed sediment mass without bedload rolling (Walker 1975; Hein 1982). This contrasts with the common imbrication of flat clasts in conglomerates deposited by traction transport, such as in gravelly rivers (Rust 1972). The fabric is, therefore, a useful indicator of transport mode.

For most carbonate sediments, a different approach to grain size and texture must be taken. Most grains, both fine and coarse, are locally produced by organic activity. The mean grain size or the size of the largest grains may mean little in terms of local hydraulics. Dunham (1962) has pointed out that it is more useful to focus on the fine material in the rock, because this is an accurate measure of the strength of winnowing currents. Dunham (1962) devised a system for classifying the texture of carbonate rocks based on whether the sediment is a self-supporting grain framework or whether the grains are enclosed in a micrite matrix. The spectrum of packing categories erected by Dunham is shown in Table 4.2. The terms **mudstone**, **wackstone**, **packstone**, **grainstone**, and **boundstone** are widely used for describing the textures of carbonates. In general, the amount of current winnowing energy implied by the rock type name increases from left to right. A similar carbonate-textural spectrum was devised by Folk (1962), who used a two-part terminological system (Fig. 4.13). The suffixes **micrite** and **sparite** are used for the predominant matrix type. Sparry rocks are packstones and grainstones from which the micrite matrix has been removed by winnowing and replaced by coarse calcite cement. Documenting textural relations in this way is an important first step in subdividing the various subenvironments of carbonate platforms. J. L. Wilson (1975; Chap. 1) discussed the subject at greater length.

Schreiber et al. (1976) and Schreiber (1981) have shown that many evaporite deposits, particularly those formed in marginal marine environments, can be treated as clastic deposits, in that they show similar bedding and sedimentary structures. The same is true for some carbonate rocks, particularly oolite sand shoals (Ball 1967) and carbonate turbidites (Mountjoy et al. 1972), which are formed by clastic sorting and redistribution processes although consisting of carbonate particles. Considerations of grain size (cyclic or lateral changes, grading) and texture (packing) might provide useful environmental information in these cases.

Table 4.2. Classification of carbonate rocks according to depositional texture. (Dunham 1962)

Depositional texture recognizable					Depositional texture not recognizable
Original components not bound together during deposition				Original components were bound together during deposition, as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or or questionably organic matter and are too large to be interstices	Crystalline carbonate (Subdivide according to classifications designed to bear on physical texture or diagenesis)
Contains mud (particles of clay and fine silt size)			Lacks mud and is grain supported		
Mud supported		Grain supported			
Less than 10% grains	More than 10% grains				
Mudstone	Wackstone	Packstone	Grainstone	Boundstone	

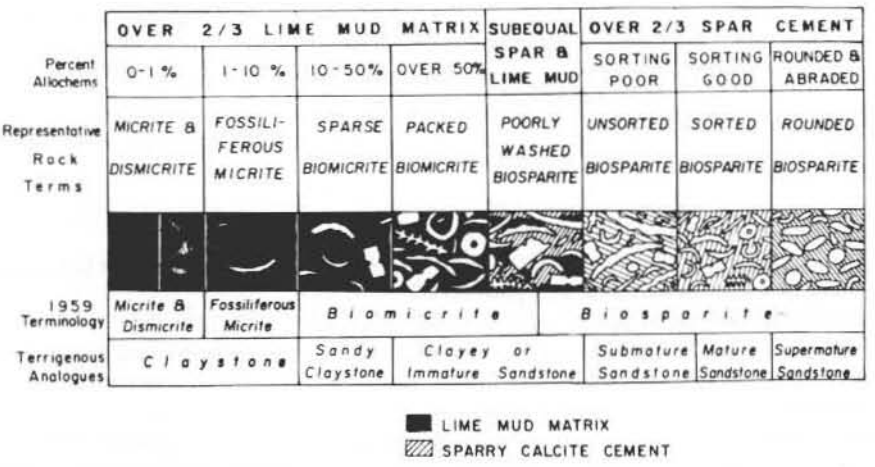


Fig. 4.13. The textural spectrum in limestones. (Folk 1962)

4.5.2 Petrology

The composition of the major rock-forming constituents of siliciclastic sediments (including mudrocks) is not directly diagnostic of environment. Krynine (1942) suggested that many sandstone types are characteristic of particular tectonic–sedimentary environments, for example **arkoses** (feldspathic sandstones) supposedly represent nonmarine sediments derived from granitic orogenic complexes, and **greywackes** (lithic sandstones) represent early "geosynclinal" sedimentation. These interpretations have been discarded, although Dickinson and Suczek (1979) and Schwab (1981) have shown that sandstone composition may closely reflect the plate-tectonic setting of the basin (Sect. 9.4). However, this does not necessarily translate into depositional environment. For example, volcaniclastic forearc sediments may be deposited in fluvial, lacustrine, marginal-marine, shelf, or submarine-fan settings, depending on the continental-margin configuration (Chap. 9).

Certain minor components of sandstones may be strongly suggestive of the depositional environment. For example, glauconite pellets form only in shallow-marine environments (Odin and Matter 1981) and are rare as detrital or resedimented grains. Carbonaceous debris from plants is typical of nonmarine environments. Paleosol development is indicated by lenses of calcium carbonate (caliche, calcrete) and other minerals. Abundant, red iron staining indicates oxygenated environments, typically either the preservation of oxidized states in detrital particles (Van Houten 1973) or the production of oxidized colors during early diagenesis (Walker 1967). Red beds are, therefore, mostly indicative of nonmarine or high, intertidal environments (Turner 1980), although there are exceptions (Franke and Paul 1980).

Within a given basin, detrital composition may reflect variations in source area or depositional environment. These data may, therefore, be used as a paleocurrent indicator (Sect. 5.5) and may assist in the stratigraphic correlation of units formed under the same hydraulic conditions. Davies and Ethridge (1975) showed that detrital composition varied between fluvial, deltaic, beach, and shallow-marine environments on the Gulf Coast and in various, ancient rock units as a result of hydraulic sorting and winnowing processes and chemical destruction. Thus, although composition is not environmentally diagnostic, it may be useful in extending interpretations from areas of good outcrop or core control into areas where only well cuttings are available.

For carbonate sediments, in contrast to siliciclastics, petrographic composition is one of the most powerful environmental indicators. Carbonate-facies studies, therefore, require routine thin-section petrographic analysis, nowadays supplemented by the use of the scanning electron microscope and cathodoluminescence (Wilson 1975; Scholle 1978; Flügel 1982; Machel 1985). Most carbonate grains are organic in origin, including micrite mud derived from organic decay or mechanical attrition, sand-sized and larger particles consisting of organic fragments, fecal pellets, grapestone, and ooliths, all of which are produced in part by organic cementation processes, and boundstones or biolithites, formed by framework-building organisms. Most carbonate particles are autochthonous; therefore, an examination of the composition of a carbonate sediment is of crucial importance.

These and other differences between carbonate and siliciclastic sediments are summarized in Table 4.3 (from James 1984a). Textural classifications of carbonate sediments were discussed in the previous section (Fig. 4.13; Table 4.2) and have been expanded

Table 4.3. Differences between siliciclastic and carbonate sediments. (Reproduced, with permission, from James 1984a)

Carbonate sediments	Siliciclastic sediments
The majority of sediments occur in shallow tropical environments	Climate is not a constraint. Sediments occur worldwide and at all depths
The majority of sediments are marine	Sediments are both terrestrial and marine
The grain size of sediments generally reflects the size of organism skeletons and calcified hard parts	The grain size of sediments reflects the hydraulic energy in the environment
The presence of lime mud often indicates the prolific growth of organisms whose calcified portions are mud-size crystallites	The presence of mud indicates settling out from suspension
Shallow-water lime-sand bodies result primarily from localized physicochemical or biological fixation of carbonate	Shallow-water sand bodies result from the interaction of currents and waves
Localized buildups of sediments without accompanying change in hydraulic regime alter the character of surrounding sedimentary environments	Changes in the sedimentary environments are generally brought about by widespread changes in the hydraulic regime
Sediments are commonly cemented on the sea floor	Sediments remain unconsolidated in the environment of deposition and on the sea floor
Periodic exposure of sediments during deposition results in intensive diagenesis, especially cementation and recrystallization	Periodic exposure of sediments during deposition leaves deposits relatively unaffected
The signature of different sedimentary facies is obliterated during low-grade metamorphism	The signature of sedimentary facies survives low-grade metamorphism

to include compositional details of framework-building organisms by Embry and Klovan (1971) and Cuffey (1985). Interpretation of ancient carbonate sediments is complicated by the fact that the organisms that generate carbonate particles have changed with time. However, there are many similarities in form and behavior between modern and extinct groups, so that actualistic modeling can usually be carried out with caution (Sect. 4.5.7). Another problem is that diagenetic change is almost ubiquitous in carbonate rocks and may obscure primary petrographic features (Sect. 4.6.8). Ginsburg and Schroeder (1973) showed that some carbonates are converted contemporaneously from reef boundstones or grainstones to wackestones by continual boring followed by infill of fine-grained sediment and cement. Mountjoy (1980) suggested that many carbonate mud mounds may owe their texture to this process.

Most environmental interpretation of carbonates is based on thin-section examination using a microfacies-description system such as that erected by Wilson (1975; Fig. 4.5 of this book). Wilson was able to define a set of standard facies belts for the subenvironments of a carbonate platform and slope, each characterized by a limited suite of microfacies reflecting the variations in water depth, water movement, oxygenation, and light penetration (Sect. 4.6.8). A classic example of a carbonate petrology study, the analysis of modern sediments of the Bahama Platform, is described briefly in Section 4.4.1 (Figs. 4.8, 4.9). Each facies assemblage can readily be related to environmental variables, such as the quiet-water, pelletoidal facies and the high-energy, skeletal sand and oolite lithofacies.

Many carbonates consist of dolomite rather than limestone. In most cases, the dolomite is clearly a replacement, as indicated by the presence of dolomite rhombs penetrating allochemical particles such as shell fragments or ooliths. In other cases, there may be evidence for a primary or penecontemporaneous origin, which may be environmentally useful information. Primary dolomite crusts form in association with evaporites in areas where seawater is evaporated at high rates, such as on supratidal flats and in shallow, tidal lagoons. Modern examples include parts of Bonaire, sabkhas on the Persian Gulf, Deep Springs Lake, California, and Coorong Lagoon, Australia (Blatt et al. 1980; Friedman 1980). The dolomite–evaporite association, together with the evidence of certain evaporite or desiccation textures (Sect. 4.5.6), is strongly environmentally diagnostic. The dolomitization processes depends on "seepage refluxion" or "evaporative pumping" of sea water to the surface under hot, arid conditions (Adams and Rhodes 1960). However, this is only one mechanism for the production of dolomite. Most dolomite probably forms beneath the surface by a diagenetic mixing of sea water and fresh water, a process known as the Dorag model (Badiozamani 1973; Land 1973). In rare cases, dolomite can also occur as detrital grains.

The composition of evaporite minerals is not a good guide to the depositional environment of eva-

porites. A sample of normal seawater, if evaporated to dryness, yields a sequence of precipitates in the following order: calcite, gypsum, halite, epsomite, sylvite, and bischofite. However, the composition of the final deposit in nature may vary considerably because of the effects of temperature, the availability of earlier-formed components for later reaction, and the rate and nature of replenishment of the water supply. Evaporites may form in a variety of marine and nonmarine environments (Schreiber et al. 1976; Schreiber 1981; Kendall 1992), and it is not their composition as much as their internal structure and lithologic associations that is the best clue to the depositional environment (e.g. the association with penecontemporaneous dolomite mentioned previously).

The formation of many chemical sediments, such as chalk, chert, phosphates, and glauconites, depends on factors such as organic activity and ocean-water oxygenation and temperature (Gorsline 1984). These factors are controlled, in part, by large-scale, oceanic circulation patterns, which are being studied in order to develop predictive models of facies development (Parrish 1983).

Certain other chemical deposits contain useful environmental information. Chert is common as a replacement mineral in carbonate sediments, where it forms nodules and bedded layers commonly containing replacement casts of fossils, ooliths, etc. Knauth (1979) suggested that such chert was formed by the mixing of fresh and marine waters in a shallow, subsurface, marginal-marine setting. Chert also occurs in abyssal oceanic sediments in association with mafic and ultramafic igneous rocks. Radiolarians, sponge spicules, and diatoms are common. These are some of the typical components of **ophiolites**, which are remnants of oceanic crust and indicate a former deep-water environment (Grunau 1965; Barrett 1982).

Iron-rich rocks occur in a variety of settings. Their chemistry is controlled partly by Eh and pH conditions. Pyrite and siderite are common as early-diagenetic crystals and nodules in the reduced environment of organic-rich muds, particularly in fluvial or coastal (deltaic, lagoonal) swamps. Occasionally, such deposits may be present in rock-forming abundance, as in the pisolitic-bog iron ores. Plant remains, impressions, and replaced (petrified) wood are commonly associated with these forms of iron. Pyrite is also associated with unoxidized, disseminated, organic particles in the black muds of anoxic lake and ocean basins.

Organic-carbon-rich black shales are of considerable importance as petroleum source beds. Arthur et al. (1984) reviewed their origins and significance in terms of oceanic organic productivity, sedimentation rate, and oxygenation of ocean-bottom waters (Fig. 4.14). Ultimate controls on sedimentation are plate-tectonic configurations, which determine global sea level and climate and which control the nature of oceanic circulation patterns. Such shales cannot, therefore, be interpreted strictly on a local basis but must be considered in a regional or even a global context.

Hematite and chamosite iron ores are locally important in the Phanerozoic. Most are oolitic and display typical shallow-water sedimentary structures. As Van Houten (1985) noted, these ironstones are particularly abundant in the Ordovician and Jurassic record but display widely varying paleogeographic and paleoclimatic settings, so their use as facies indicators is uncertain at present. The iron is probably an early diagenetic replacement. Precambrian iron formations are widespread in rocks between about 1.8×10^9 years and 2.6×10^9 years old. Their mineralogy is varied and unusual (Eichler 1976); it may reflect formation in an anoxic environment (Cloud 1973), although this has been disputed (Sect. 9.5.2).

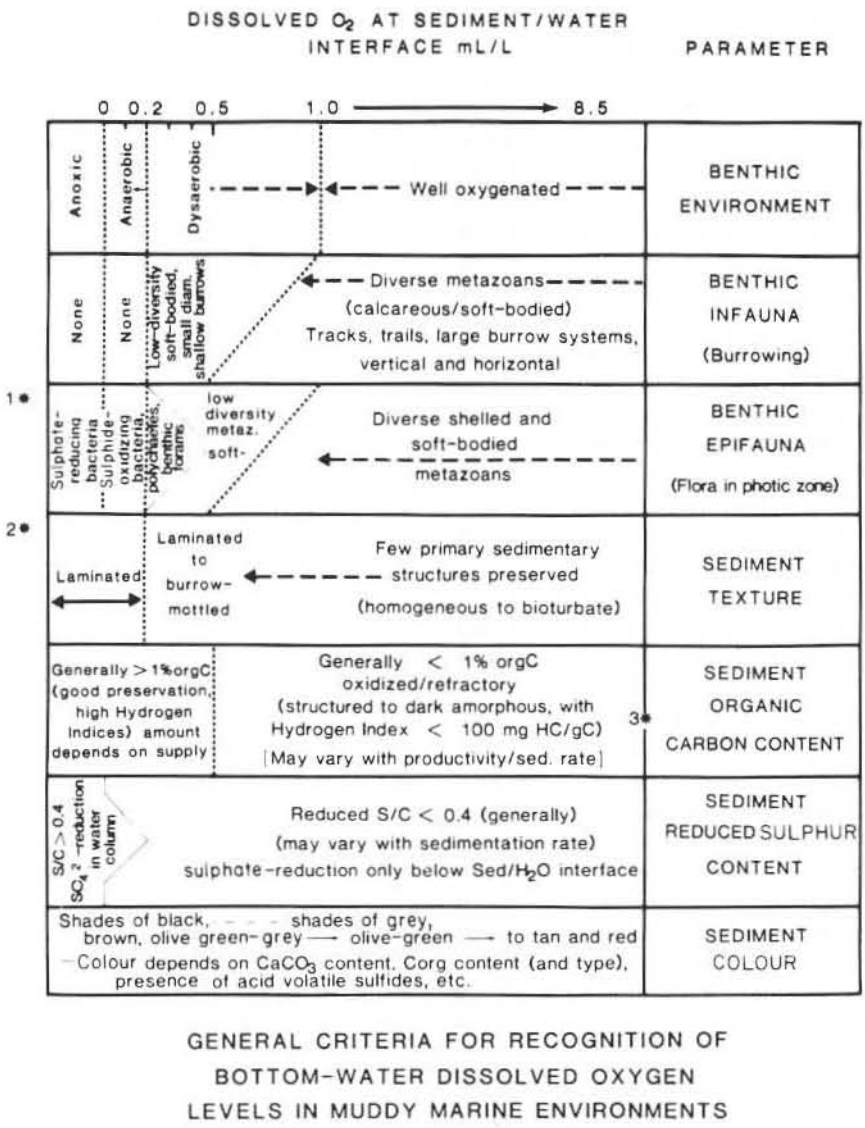
Manganese- and phosphate-rich rocks are locally important. Both commonly occur as crusts and replacements on disconformity and hardground surfaces, where they are taken as indicators of nondeposition or very slow sedimentation. Phosphates are particularly common on continental margins in regions of upwelling oceanic currents. More detailed discussions of these sediments are given by Blatt et al. (1980).

Coal always indicates subaerial swamp conditions, usually on a delta plain or river floodplain. However, lacustrine coals and coal formed in barrier-lagoon settings have also been described. Calcrete (or caliche) occurs in alluvial and coastal environments and is an excellent indicator of subaerial exposure (James 1972; Bown and Kraus 1981; Wright 1986).

4.5.3 Bedding

Are the various rock types present in a stratigraphic unit interbedded in major packages several or many meters thick, or are they interlaminated on a scale of a few centimeters or millimeters? Is the bedding fine or coarse? Is it flat, disturbed, undulatory, distinct, or indistinct (gradational)? These questions, while rarely providing answers that are uniquely diagnostic of a depositional environment, may provide important supplementary information. In a general way, the bed thickness is proportional to the depositional energy level.

Finely laminated sediments are mostly formed in quiet-water environments. X-radiography and scanning electron microscopy are commonly used to define lamination and very-small-scale sedimentary structures in fine-grained rocks (several papers in



GENERAL CRITERIA FOR RECOGNITION OF BOTTOM-WATER DISSOLVED OXYGEN LEVELS IN MUDDY MARINE ENVIRONMENTS

Fig. 4.14. Oxygen content at the sediment-water interface and generalized relationships between types of benthic organisms, sedimentary structures, water chemistry, mineralogy, organic content, and sediment color. Notes: 1*, certain polychaete organisms and foraminifera may inhabit surficial environments where benthic metazoans are excluded; 2*, the degree of lamination depends, in part, on seasonal variations in clastic input and productivity. (Arthur et al. 1984; reproduced by permission of the Geological Society, London)

Stow and Piper 1984; O'Brien 1990). Such sediments may include laminated pelagic mudrocks, prodelta deposits, deep-water evaporites, thin-bedded basin-plain turbidites, and delta-plain lagoonal and fluvial overbank muds. Laminated, fluvial sheet flood sandstones are an exception to this pattern (Sect. 4.5.4). Thicker beds form in a variety of high-energy, wave- or current-dominated environments. Reef rocks, formed in extremely high-energy conditions, may lack bedding entirely.

An interbedding of contrasting lithofacies may be environmentally indicative. For example wavy, flaser, and lenticular bedding (**tidal bedding**; Reineck and Wunderlich 1968) record the alternation of quiet-water mud sedimentation and higher-energy flow conditions under which rippled sand is deposited. This can occur during tidal reversals on exposed mudflats, on fluvial floodplains, or below the normal wave base on the shelf at depths affected by infrequent storm waves. Another common bedding association is that produced by the alternation of storm- and fair-weather processes on the shoreface. A sequence, basal gravel → laminated sand → bioturbated or rippled sand, indicates storm suspension

followed by decreasing energy levels and then a return to low-energy wave activity and bioturbation during periods of fair weather (Kumar and Sanders 1976).

As noted in Section 4.5.1 maximum grain size and bed thickness are commonly correlated in deposits formed by individual sediment-gravity-flow events. This can provide invaluable interpretive data. Bedding surfaces that bound depositional units have a particular importance in facies analysis, as discussed in Section 4.5.9.

4.5.4 Sedimentary Structures Produced by Hydrodynamic Molding of the Bed

The interpretation of hydrodynamic sedimentary structures is one of the most important components of facies analysis, particularly in siliciclastic sediments. Certain carbonate and evaporite deposits contain similar structures and can be studied in the same way.

The basis for interpreting structures formed in aqueous environments is the **flow-regime concept.** This fundamental theory states that flow of a given depth and velocity over a given bed of noncohesive grains will always produce the same type of bed configuration and, therefore, the same internal stratification. If such structures are predictable, their presence can be used to interpret flow conditions. These fundamental ideas were first enunciated following an extensive series of flume experiments by Simons and Richardson (1961) and were developed further by Simons et al. (1965), Harms and Fahnestock (1965), Southard (1971), and Harms et al. (1975, 1982) for use by geologists. Valuable syntheses of hydrodynamic sedimentary structures were provided by Collinson and Thompson (1982) and Allen (1982), and reviews of later developments were provided by Leeder (1983), Ashley (1990), and Allen (1993).

It is now realized that bedforms are controlled mainly by three parameters, sediment grain size, flow depth, and flow velocity, and a series of experiments has demonstrated the sequences of bedforms produced as these parameters are varied. For example, Fig. 4.15 shows the stability fields of ripples, dunes, and other bedforms for flow depths of about 20 cm. A welter of terms for various types of bedforms in different environments was clarified by Ashley (1990) following lengthy debate by a research group under the auspices of the Society for Sedimentary Geology (SEPM), previously known as Society of Economic Paleontologists and Mineralogists. Two broad classes of flow-transverse bedforms are now defined, 2-D forms, which occur at lower flow velocities, and 3-D forms, which occur at higher velocities for a given grain size. There is a continuum of sizes of these bedforms from small-scale ripples to very large dunes more than 10 m in height. A natural discontinuity in size occurs at a spacing (wavelength) of 0.5–1.0 m, which reflects the structure of the turbulence that generates these bedforms (Leeder 1983). Forms smaller than this are termed ripples; Ashley (1990) recommended that those larger should all be termed **dunes.** Small-scale ripples, dunes, and sand waves are forms out of phase with surface water movement; indeed, their form may bear no relation to surface water patterns at all. These bedforms have traditionally been

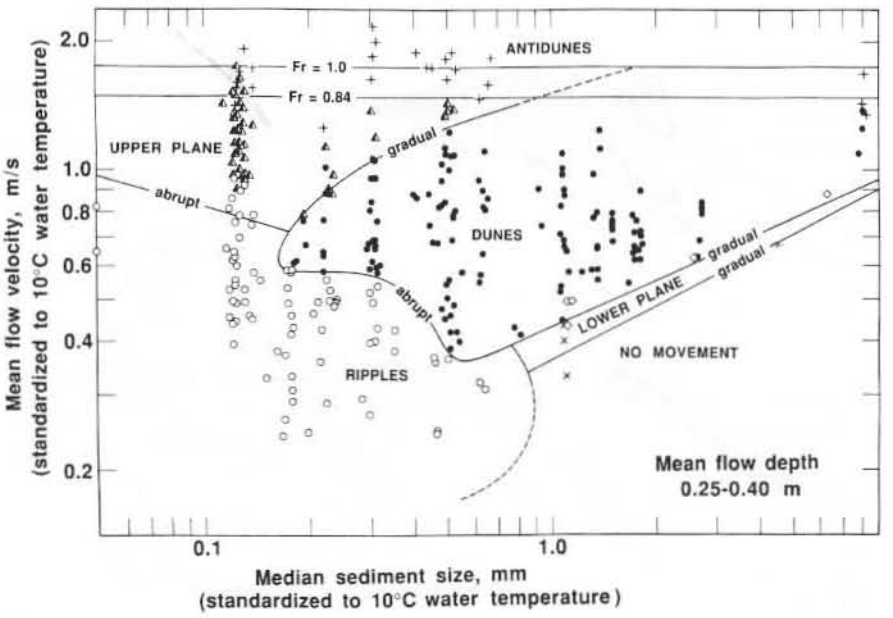


Fig. 4.15. Stability fields of bedforms in sand and silt in a flow depth of 25–40 cm. Bed-phase symbols: × no movement on plane bed; ○ ripples; • dunes; ◇ lower-flow-regime plane bed; △ upper-flow-regime plane bed; + antidunes. *Fr* Mean-velocity Froude Number. (Ashley 1990)

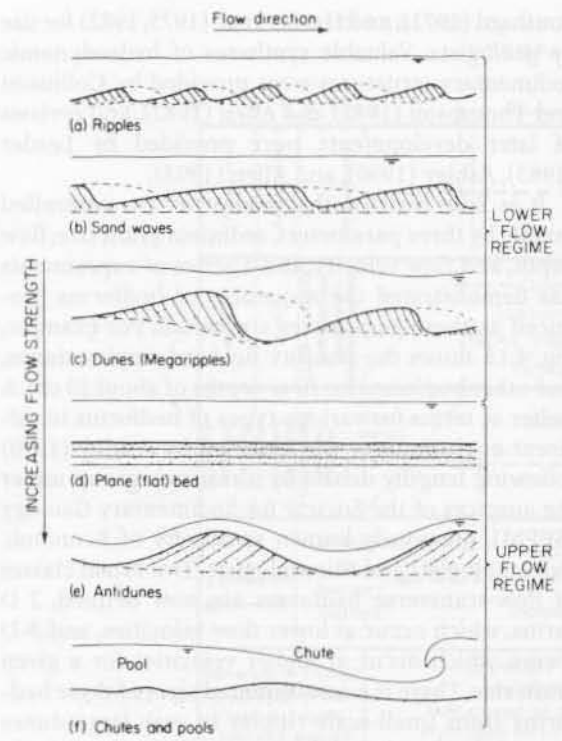


Fig. 4.16. The flow-regime concept, illustrating the general succession of bedforms that develops with increasing flow velocity. *Dashed lines* indicate areas of flow separation. Note internal stratification. (After Simons et al. 1965; Blatt et al. 1980)

classified as **lower flow-regime** forms (Fig. 4.16). The **upper flow regime** is characterized by **antidunes** and standing waves, which are in phase with surface water waves. The transition from the lower to the upper flow regime is marked by a streaming out of transverse turbulent eddies into longitudinal eddies. An intermediate **upper flat-bed condition** is marked by streaming flow, which aligns the sand grains and produces primary current lineation (parting lineation; Fig. 2.14B).

How can these flume data be used to interpret ancient sediments? First, Allen (1968) and Harms et al. (1975) demonstrated the relationships between bedforms and sedimentary structures. For example, planar tabular crossbedding is produced by the migration of straight-crested megaripples, such as sand waves (what are now termed 2-D dunes), whereas trough crossbedding develops from the migration of 3-D dunes (Fig. 4.17). Allen (1968) demonstrated the dependence of dune and ripple shape on water depth (Fig. 4.18). Second, the flow-regime concept may be used to interpret ordered sequences of sedimentary structures in terms of gradations in flow conditions. Examples of applications to fluvial point-bar deposits, Bouma turbidite sequences, and wave-formed sedimentary structures are discussed below. This is by no means an exhaustive listing. For example, the concepts have been adapted by Dott and Bourgeois (1982) to the interpretation of hummocky crossbedding, a product of storm-wave activity (below and Sect. 4.6.7).

Allen (1970) discussed flow patterns around a meander bend, in particular the effect on bedforms of the helical flow that develops in each meander, resulting in the movement of grains obliquely up the point-bar slope (Fig. 4.19). Decreasing depth and velocity up this slope result in decreasing grain size and scale of sedimentary structures. Using flow-regime data, such as that shown in Figs. 4.15 and 4.16, Allen (1970) was able to predict the types of sedimentary structures from depth–velocity–grain-size conditions. His series of hypothetical profiles is shown in Fig. 4.20.

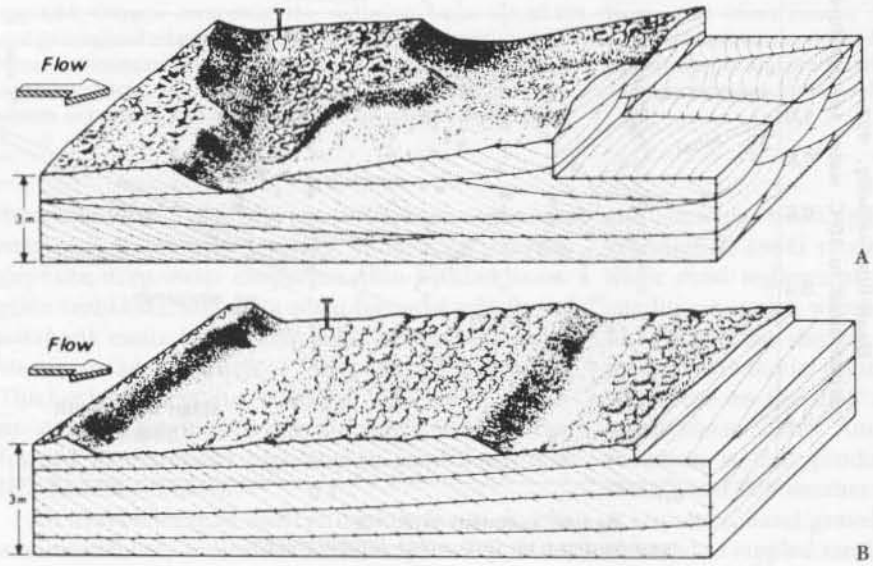


Fig. 4.17 A, B. Relationships between bedforms and sedimentary structures. **A** Linguoid (three-dimensional) dunes and trough crossbedding. **B** Sand waves (two-dimensional dunes) and planar crossbedding. (Harms et al. 1975)

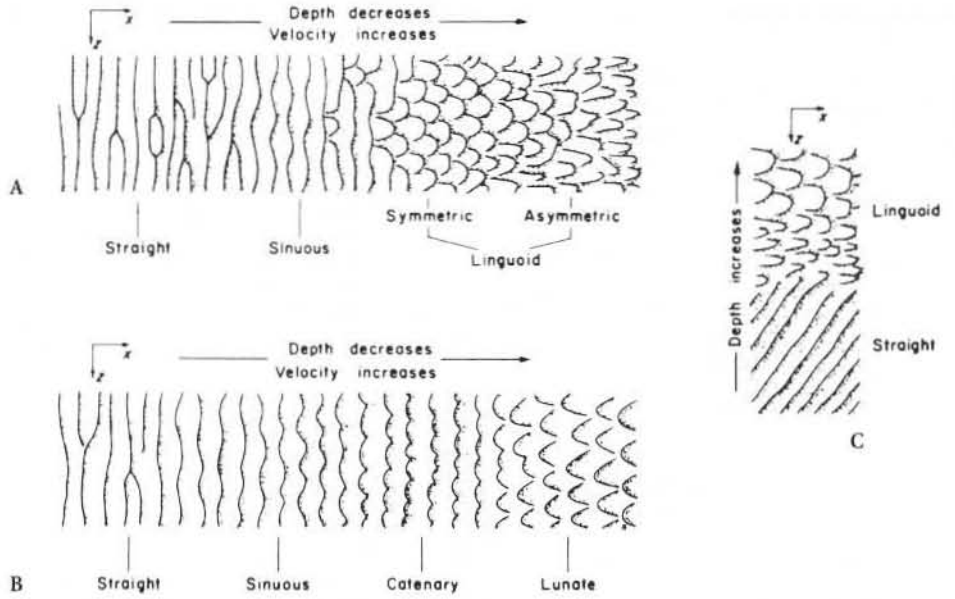


Fig. 4.18 A–C. Variations in bedform morphology with depth and velocity. **A** Large-scale ripples (dunes or megaripples). **B** Small-scale ripples. **C** Large-scale ripples where depth varies transverse to flow. (Allen 1968)

These can be matched to real examples of fining-upward cycles in the Devonian of Wales and the Appalachian region, demonstrating that Allen's model was of considerable value in reconstructing paleohydraulic conditions. Many elaborations of this model have now been developed (Miall 1996 for summary).

The Bouma sequence of thin-bedded, outer submarine-fan turbidite deposits also contains a succession of structures that can be interpreted in terms of flow-regime. The basal (*A*) member (Fig. 4.21) is formed by grains settling from suspension. Flow velocities decrease upward, so that the plane-bedded unit (*B*), which commonly contains parting lineation, is formed under upper-flow-regime, flat-bed conditions, and the rippled unit (*C*) represents the lower flow regime (Harms and Fahnestock 1965). The silty unit (*D*) is deposited from the dilute tail of the turbidity current as flow ceases altogether. This interpretation has been of considerable use in understanding the mechanics of turbidity currents.

Clifton et al. (1971) carried out one of the first detailed studies of the sedimentary structures that form on coastlines under breaking waves. They recognized a direct relationship between wave type, resulting water motion, and structure type (Fig. 4.22). The gradation from asymmetric ripple to outer, planar facies represents a shoreward increase in orbital velocity and a transition from lower flow-regime ripples through a dune facies to an upper-flow-regime plane-bed condition. These structures all dip landward. The inner, rough facies is characterized by seaward-dipping ripples and dunes of the lower flow regime, and the inner, planar facies by plane beds, antidunes, and standing waves formed under high-energy, upper-flow-regime, shallow-swash conditions. These facies all move up and down the shore

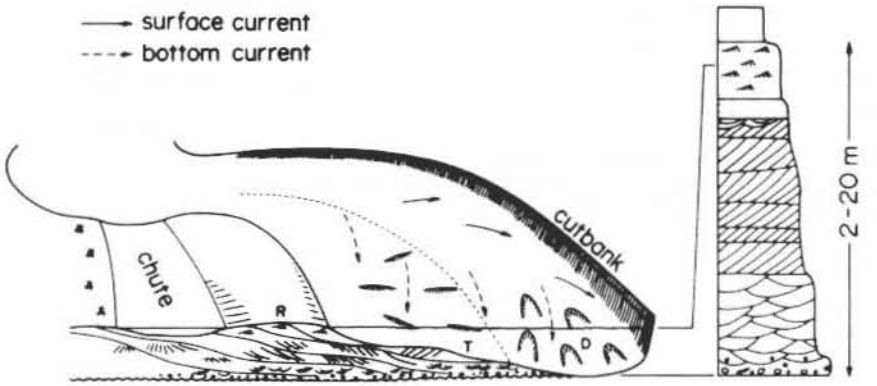


Fig. 4.19. Facies model for sedimentation on a point bar by lateral accretion inside a migrating meander. *D* Dunes; *T* transverse bars or sand waves; *R* ripples

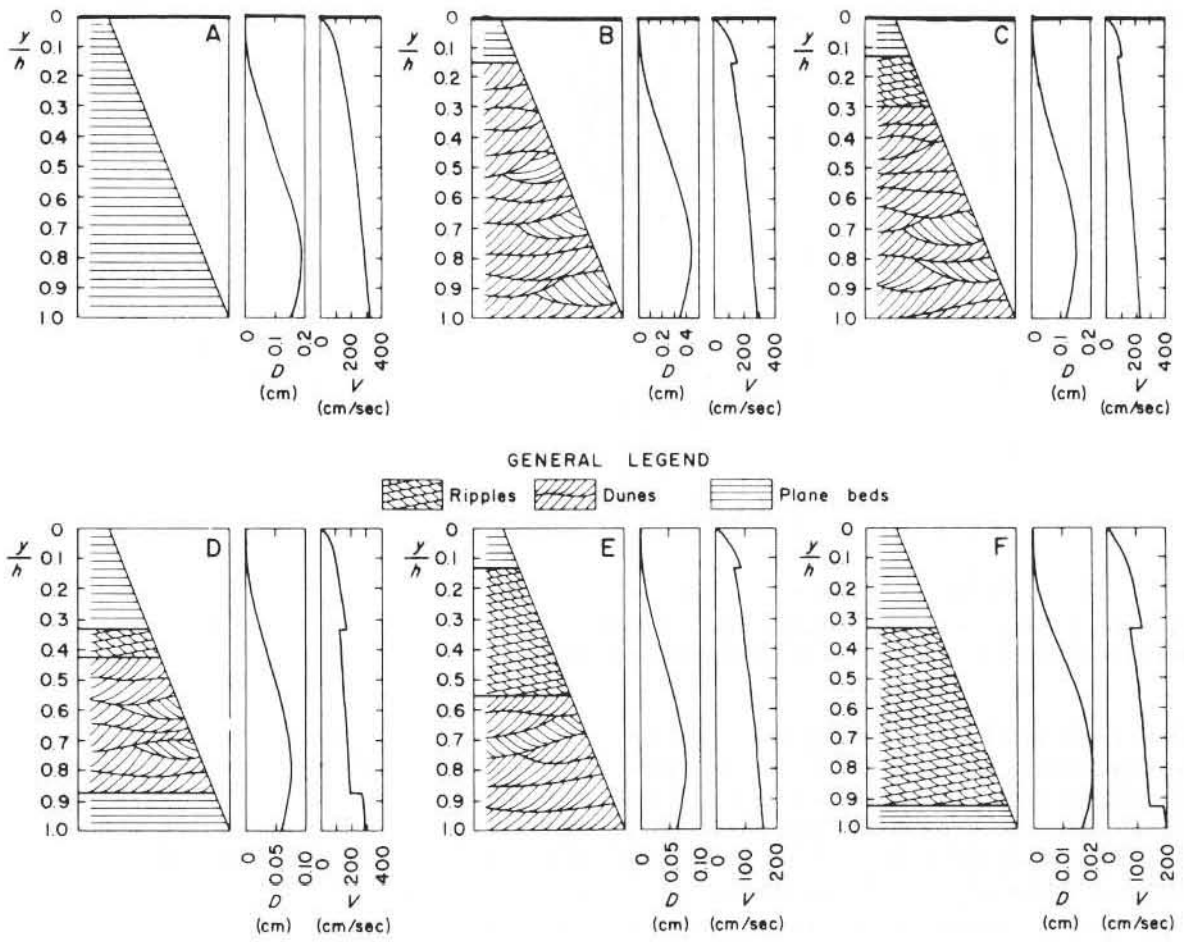


Fig. 4.20 A–F. Hydraulic model for point-bar sedimentation, showing variations in the vertical profile reflecting variations in grain size, *D*, and flow velocity, *V*. *y/h* indicates position on point bar with respect to total depth. (Allen 1970)

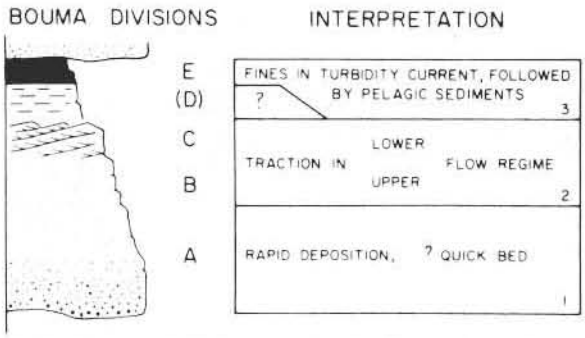


Fig. 4.21. Hydraulic interpretation of a Bouma turbidite sequence. Unit *D* represents deposition from the dilute tail of the turbidity current. (Walker 1978)

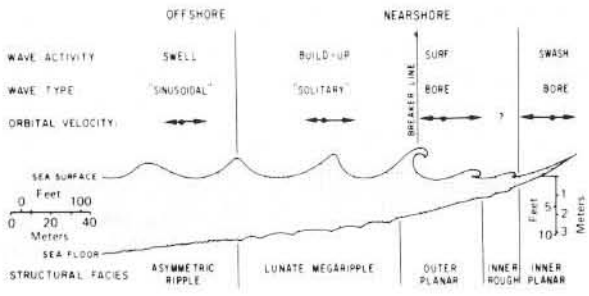


Fig. 4.22. Distribution of sedimentary structures (structural zones or facies) beneath shoaling waves. (Clifton et al. 1971)

with the rise and fall of the tide, producing a complex but distinctive series of structural assemblages that Clifton et al. (1971) showed could be recognized in the ancient record.

Many environmental deductions can be made from the details of the internal structure of hydrodynamic sedimentary structures and from orientation (paleocurrent) information. Three general groups of structures can be distinguished:

1. Structures formed by unimodal water currents in rivers, deltas, parts of ebb and flood tidal deltas in inlets, submarine fans, and continental slopes (contour currents)
2. Structures formed by reversing (bimodal) water currents, such as tides and wave oscillations in shelf and marginal-marine environments and in lakes
3. Structures formed by eolian currents in coastal dune complexes, inland sand seas, and some alluvial–lacustrine environments

Unimodal currents are readily recognized from unimodal foreset orientations (Figs. 2.12A, C, D, F, G), but such patterns are not necessarily environmentally diagnostic. For example, as discussed in Section 4.6, it has been found that in areas of strongly reversing currents, such as tidal inlets and their associated deltas, ebb and flood currents are segregated into different parts of the system. Structures in a single outcrop of a tidal delta may, therefore, be misinterpreted as fluvial in origin, based on structure type and paleocurrent patterns. Simple paleocurrent models, such as those of Selley (1968), should, therefore, be used with caution. Other evidence, such as faunae, might yield clues as to the correct interpretation. Crossbedding structures may contain evidence of stage fluctuation in the form of reactivation surfaces (Collinson 1970), as shown in Fig. 4.23. These are erosion surfaces formed during a fall in the water level but, again, they are not environmentally diagnostic, as water levels rise and fall in rivers, deltas, and tidal environments.

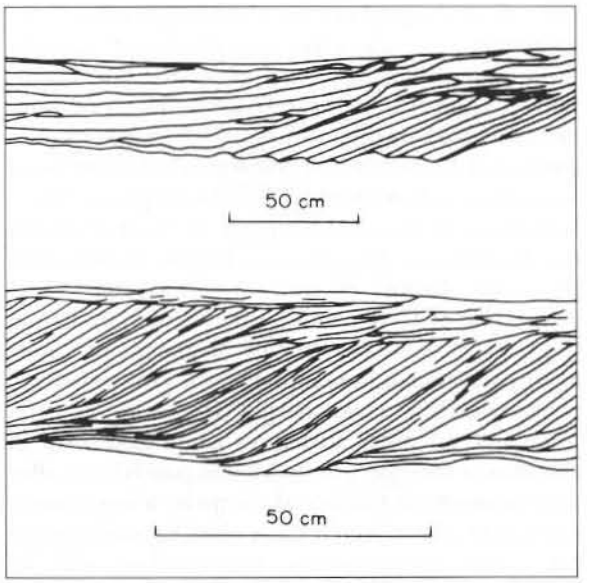


Fig. 4.23. Planar crossbed sets, showing reactivation surfaces. (Miall 1977; after Collinson 1970)

Reversing currents can be recognized from such structures as herringbone crossbedding (Fig. 2.12I) or wave-ripple cross-lamination (Fig. 4.24), in which foreset dip directions are at angles of up to 180° to

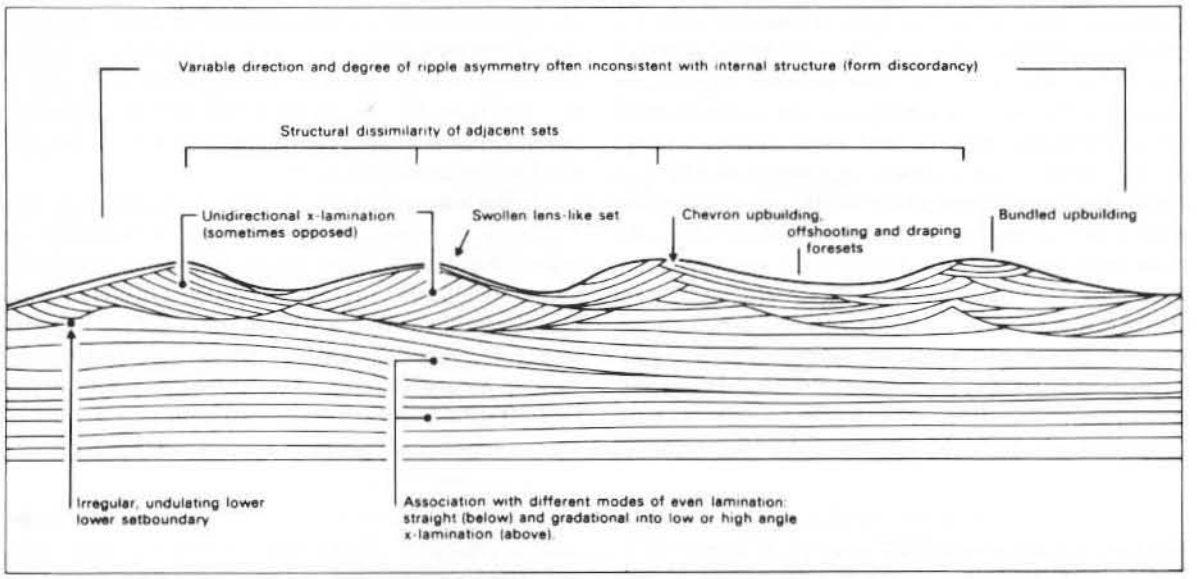


Fig. 4.24. Distinctive characteristics of wave-formed ripples. (De Raaf et al. 1977; Reading 1986)

each other. Herringbone crossbedding is a classic indicator of reversing tidal currents, but it can also form under oscillatory, wave-generated flow conditions (Clifton et al. 1971) and even in fluvial environments, where bars migrate toward each other across a channel. Because of the segregation of ebb and flood currents in estuaries and inlets, herringbone crossbedding is, in fact, not common in many marginal marine deposits. The reversing ripples, chevron ripples, lenticular foresets, and variable symmetry and orientation of wave-formed ripple cross-lamination (Figs. 2.12B, 4.24) are strongly diagnostic of a low-energy wave environment, such as a gently shelving marine beach or a lake margin (De Raaf et al. 1977; Allen 1981). Similar structures could also form in abandoned meanders or floodplain ponds in an alluvial environment but would comprise a less conspicuous part of the overall succession. In many marginal-marine environments, crossbedding will be formed by both waves and tides, resulting in very complex paleocurrent patterns (Klein 1970). Careful documentation of structure types and their orientations may be necessary to distinguish the precise environment and mode of origin, but such work may also yield invaluable information on sand-body geometry, shoreline orientation, beach and barrier configuration, etc., as discussed in Sect. 5.6.

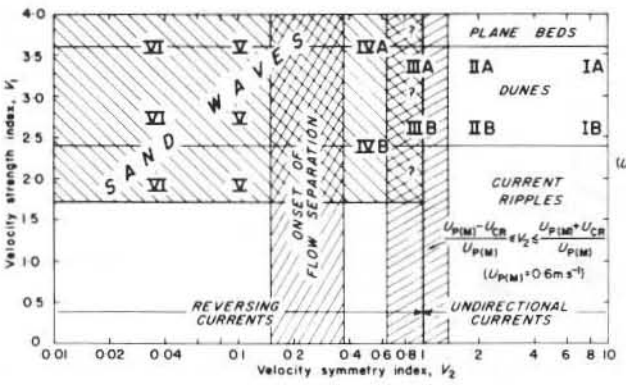
Sand waves that develop in tidal environments range from 1 m to 15 m in thickness and have characteristic internal structures, as modeled by Allen (1980, 1982). As noted earlier, ebb and flood currents in tidal settings are commonly segregated from each other so that, at any given location, relative strengths of the tidal stages may be quite different. The variations in tidal asymmetry give rise to important differences in the internal morphology of the resulting bedforms (Fig. 4.25). Strongly unimodal currents produce bedforms similar to those in fluvial settings (class I in Fig. 4.25). As tide–velocity asymmetry decreases, the weak reversing current is represented by reactivation surfaces and mud drapes (classes II–IV). With a more nearly symmetrical ebb and flood, the internal structure of the sand wave becomes very complex, with the migration of small-scale bedforms in both directions across the main accretion face (classes V and VI). Careful analysis of the thicknesses and numbers of sand layers in the bundles between reactivation surfaces can yield information regarding the nature of the local tidal patterns, including diurnal and lunar cycles (Visser 1980; Williams 1989). Many other subtle indicators of tidal influence were described by Nio and Yang (1991) and Shanley et al. (1992).

Recognition of eolian crossbedding (Fig. 2.12M) was, until recently, one the major problems in clastic sedimentology. Much reliance was placed on the idea that eolian dunes are large (tens of meters high), resulting in very-large-scale crossbedding. However, giant crossbed sets have now been recognized in fluvial (Coleman 1969; Jones and McCabe 1980) and shelf (Jerzykiewicz 1968; Nio 1976; Flemming 1978) environments, so this argument is no longer valid. Two important developments have led to a revolution in the study of eolian strata since the mid-1970s. First, Brookfield (1977) and Gradzinski et al. (1979) examined the mechanics of dune construction and migration and presented some useful ideas on the nature of large-scale crossbed bounding surfaces. Second, Hunter (1977a, b, 1980, 1981) and Kocurek and Dott (1981) studied the details of sand movement by wind on modern and ancient dunes and showed that several distinctive crossbedding and lamination patterns are invariably produced. In addition, Walker and Harms (1972) and Steidtmann (1974) carried out useful, detailed facies studies of ancient eolian units. All this work has brought us to the point where eolian crossbedding should now be relatively simple to recognize, even in small outcrops.

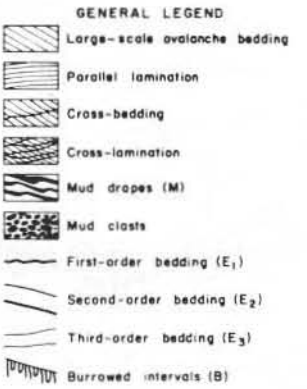
Gradzinski et al. (1979) illustrated the gross structure of dune deposits in the Tumlin Sandstone (Triassic of Poland), showing the presence of three types of bounding surfaces (Fig. 4.26). The main bounding surface is formed by the migration of transverse dunes and the truncation of underlying sets; second-order surfaces are commonly shallow scoop shapes bounding cosets of cross-strata; and third-order surfaces, analogous to reactivation surfaces, develop by changes in turbulent eddy currents flowing along the dune slip face. These currents commonly generate wind ripples with crests oriented parallel to the dip of the slip face. The recognition and use of bounding surfaces in the analysis of clastic sediments is discussed further in Section 4.5.9.

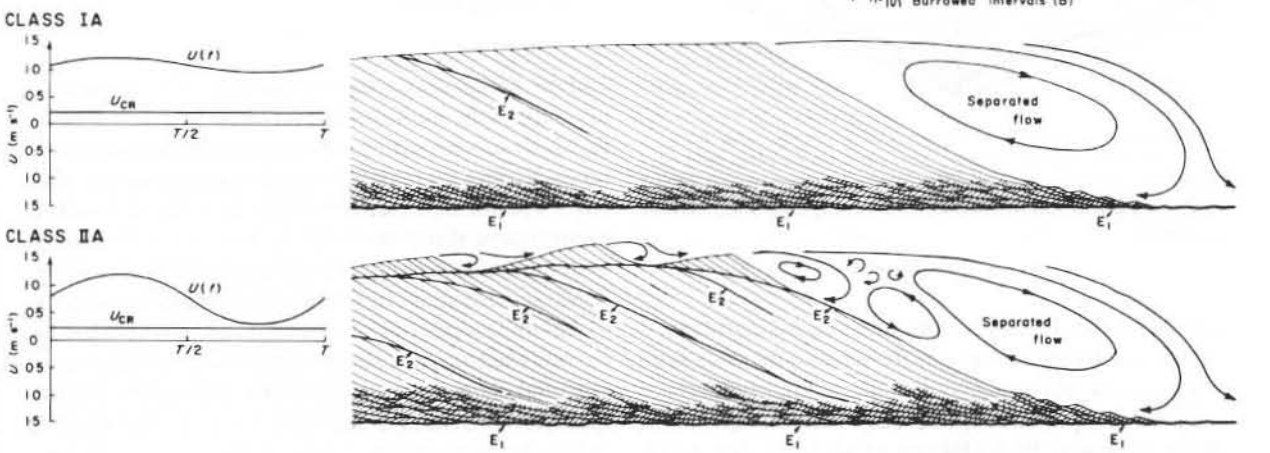
Hunter's contribution has been to demonstrate the variety of mechanisms by which sand is deposited on eolian dunes. Climbing ripple migration, grainfall, and sandflow are the principal processes, all of which can be readily recognized from the details of internal structure. For example, sand-flow crossbedding is formed by avalanching of noncohesive sand on slip faces. It forms units that are distinctly lens shaped in cross section and wedge shaped downdip. More re-

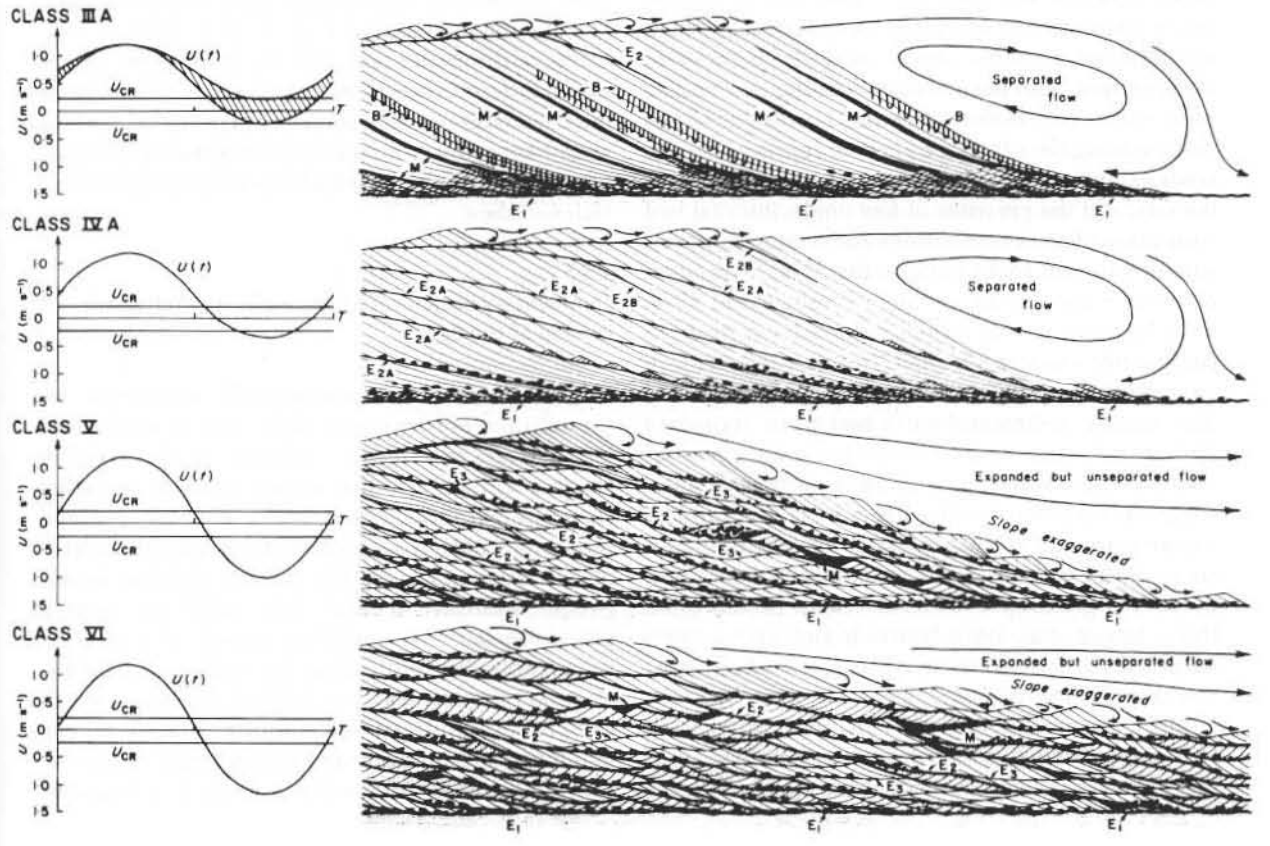
Fig. 4.25. A model to explain the variations in internal structure in tidal sand waves produced by variations in tidal time–velocity asymmetry. U is the current velocity, U_{cr} indicates the critical velocity required to initiate grain movement, and $U(t)$ indicates the tidal current velocity varying with time T. (Allen 1980)



REGIME DIAGRAM







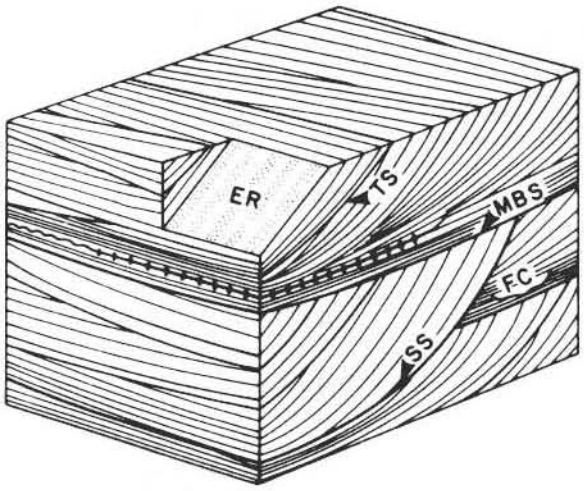


Fig. 4.26. Structure of a typical ancient eolian dune deposit. *MBS* Main bounding surface; *SS* second order bounding surface; *TS* third order bounding surface; *ER* eolian ripples; *FC* fluvial channel or sheet flood deposits. (After Gradzinski et al. 1979)

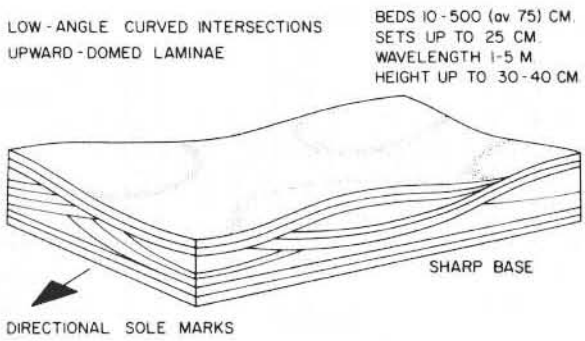


Fig. 4.27. Hummocky cross-stratification. (Hamblin and Walker 1979; Walker 1984)

cently, Hunter (1985) has pointed out the importance of the sand-flow process in the development of subaqueous crossbedding.

The last structure that should be mentioned in this brief review is HCS (Harms et al. 1975; Fig. 4.27), which has been recognized in many ancient shelf deposits (Byers and Dott 1981; Dott and Bourgeois 1982). The structure is readily recognizable in small outcrops but requires careful analysis in order to distinguish it from normal crossbedding in core. The most distinctive features are the low angle of the crossbedding (less than the angle of repose), the presence of both convex-up and concave-up curvature in the sets, and the presence of low-angle, internal bed truncations. Facies associations and stratigraphic position in ancient rocks indicate that it is the product of storm-wave activity in the inner shelf, and a few identifications of this structure have been made in comparable modern settings (Howard and Reineck 1981; Swift et al. 1983). Swift et al. (1983) suggested that marine sedimentologists had been recording HCS in side-scan sonar records for some time but interpreting the structures as dunes or megaripples. It now seems likely that the structure is produced by a combination of oscillatory water movement induced by storm activity and a unimodal current drift, possibly a geostrophic current set up by the same storm (Swift et al. 1983; Nottvedt and Kreisa 1987; Allen 1993). HCS has become a widely used environmental indicator (Sect. 4.6.7).

Most discussions of bedforms and sedimentary structures (including the preceding paragraphs of this review) deal with individual trains of structures and the conditions under which they form. This simple approach ignores the fact that, in nature, bedforms of several scales are commonly superimposed on each other and migrate simultaneously or alternately (Jackson 1975; Ashley 1990). In some situations, bedforms of different scales represent simultaneous responses to different scales of fluid turbulence within the same system. In other settings, the interactions of different types of bedforms represent responses to currents acting at different times (e.g., diurnal and seasonal changes in wind pattern). Examples of both types can be readily observed in modern and ancient eolian deposits, where the resulting cross-stratification structures may be extremely complex. Rubin (1987) developed a computer model to simulate graphically the interaction of up to three trains of bedforms of different scales moving at different speeds and in different directions. Many of the models can be matched with real-life examples in outcrops, and the work has produced an invaluable manual of structure types, a manual that will find considerable use in field facies studies.

4.5.5 Sedimentary Structures Produced by Hydrodynamic Erosion of the Bed

Few of these are environmentally diagnostic, although their presence may add weight to an interpretation made from other evidence. As discussed in Chapter 2, there are two main classes of erosional structure at the outcrop scale, macroscopic and mesoscopic. Macroscopic structures comprise channels, scours, and low-relief erosion surfaces. Examples are illustrated in Fig. 2.15 A – D. They can occur in practically any environment as a result of current activity. More diagnostic than the channel itself is the channel fill, which can be analyzed in terms of vertical profile, lithology, sedimentary structures, etc. Subordinate features on an erosion surface may also provide some environmental information. Lag con-

centrates of pebbles, bioclastic debris, or phosphatized fossils may be interpreted in terms of such processes as condensed sedimentation or wind deflation, storm erosion and deposition, or submarine nondeposition (hardgrounds). Desiccated surfaces may break up, yielding breccias of partly lithified mud, silt, or carbonate clasts, which can then be interpreted as the result of subaerial exposure (Sect. 2.2.3).

Another term for macroscopic erosion surfaces is bounding surfaces. Their physical extent, geometry, and facies associations have become important components of the architectural subdivision and classification of clastic deposits (Sects. 2.2.2, 4.3.4, 4.5.9), and also feature in the definition and subdivision of stratigraphic sequences (Chap. 6).

Mesoscopic erosional features include the variety of sole markings described and illustrated in Chapter 2. These are abundant in submarine-fan deposits and submarine-basin-plain sediments, particularly in sandy turbidites. Flute marks, in particular, indicate a style of vortex turbulence that is common at the base of turbidity currents (Fig. 2.16 A). However, they have been observed in a wide range of other settings, including fluvial traction-current deposits. Turbidity currents are not restricted to the deep oceans but are common in lakes and subaqueous ice-margin environments.

4.5.6 Liquefaction, Load, and Fluid-Loss Structures

A few of these are environmentally diagnostic, others are practically ubiquitous in occurrence. Load and slump structures, convolute bedding, and other products of thixotropy, can occur wherever sediments are saturated, which is to say practically anywhere. Water-escape features, such as dish and pillar structures (Fig. 2.17 A – D), are also found in the deposits of many environments (Nilsen et al. 1977), although they are particularly abundant in fluidized sediment-gravity flows.

As noted in Chapter 2, subtly different sedimentary structures have been attributed to desiccation and synaeresis. Desiccation cracks are an indicator of subaerial exposure; synaeresis has been attributed to volume changes resulting from salinity fluctuations in tidal, lagoonal, or lacustrine settings (Donovan 1973), but a re-examination of one of the classic locations of synaeresis cracks, the Orcadian Basin in Scotland, suggested that the "cracks" are gypsum pseudomorphs (Astin and Rogers 1991). Desiccation may be accompanied by the breakup of the uppermost beds and their resedimentation as an intraformational breccia. Shinn (1983) provided a useful review of both the terminology of some of the small-scale structures that develop on tidal flats and the kinds of interpretations that can be made from them. Displacive evaporite growth in sabkhas produces a variety of distinctive structures that are useful diagnostic indicators (Sect. 2.2.3).

4.5.7 Paleoecology

In Chapter 2, we discuss briefly the three main types of body-fossil preservation, those in growth positions, soft-bodied organisms that would not survive significant transportation, and the far more common death assemblage. In Section 3.5.1, we touch on the problems of the biostratigraphic interpretation of death assemblages and reworked fossils and the difficulties of correlating faunas and floras limited in their distribution by factors of ecology or biogeography (not necessarily the same thing). In Section 5.6, we discuss the use of oriented fossils as paleocurrent indicators. All these points are relevant when we focus on the use of body fossils as indicators of depositional environment.

Paleoecology is the study of the relationships between fossil organisms and their environment. In most paleoecological research and in the resulting books and papers, the primary focus is on the animals, with the use of such criteria as species inter-relationships and fossil-sediment characteristics in the analysis of the ecology of some fossil species or community. Such research is typically carried out by specialists whose background is usually paleontology or biology, and these individuals may have little or no interest in regional stratigraphy or basin history. Here, we wish to reverse the process, to use fossils as paleogeographic indicators. In the Phanerozoic record, fossils are among the most powerful environmental indicators available but, of course, their distribution and usefulness in the Precambrian is very limited. Useful texts on this subject have been provided by Ager (1963), Boucot (1981), and Dodd and Stanton (1981).

Even the simplest paleoecological observations can be invaluable. The presence of abundant bioturbation or (for example) a brachiopod fauna may be the key evidence in distinguishing the deposits of tidal channels and their point bars, tidal inlets, and deltas from certain fluvial and delta-plain facies, which can be sedimentologically very similar (Barwis 1978). At the other extreme are sophisticated statistical studies of particular communities or fossil groups, such as the trilobite-population analysis of Ludvigsen (1978). Most paleoecological studies by stratigraphers and basin analysts fall somewhere between these extremes.

Fossil data are subject to two biases that may distort our observations. First, there is the bias of un-

skilled or incomplete collection, which is particularly likely to arise when paleoecological analysis is attempted by the nonspecialist or when it is rushed through, as in a reconnaissance survey, by a prospecting team. Ager (1963) demonstrated this with data assembled by B. W. Sparks, who used two methods to collect molluscs from Quaternary deposits in southern Britain (Fig. 4.28). A bulk sample was analyzed by sieving, and the distribution was compared with that of a collection made by hand picking from the outcrop surface. The bias in the second collection in favor of larger or more brightly colored species is startling.

The second bias is that introduced by the geological obstacle courses organisms are put through before they end up under the geologist's hammer. The various hazards and possible paths through them are summarized in Fig. 4.29. The study of this flow diagram is itself the subject of a specialized science called **taphonomy**. Many of the processes of transportation, removal, or breakup of a fossil depend on sedimentological processes and are, therefore, of interest to basin analysts. Figure 4.30 illustrates the fate of dinosaur remains in Cretaceous fluvial deposits of the Red Deer Valley, Alberta (from Dodson 1971). The state of preservation may yield much information on local transport energy and, hence, on the depositional environment.

As shown by a thematic set of papers in the journal *Palaios* (Brett and Baird 1986; Speyer and Brett 1986), the study of taphonomy is growing into a useful subdiscipline. Although valuable in a negative sense, revealing the role of postmortem processes in biasing the fossil record, taphonomic studies can also provide useful information on the physical and chemical processes occurring after death and during burial (**biostratinomy**) and those that occur during early diagenesis (fossil diagenesis; Brett and Baird 1986). Speyer and Brett (1986) showed that the varying states of preservation of fossil trilobites (disarticulation, fragmentation) records useful information regarding energy levels in the depositional environment and proposed that these variations be formalized into taphonomic facies or **taphofacies**.

One of the first and most obvious observations that must be made in a paleoecological study is the relation of biofacies to lithofacies. The relationships may be obvious, including marked contrasts in facies, such as that shown in Fig. 4.31 (from Hecker 1965). Here, there is a red-bed lithofacies characterized by fish, a brackish lithofacies with Lingula, and marine deposits with corals, stromatoporoids, and abundant bioturbation. This pattern of interbedded sediments and fossil types is characteristic of Early Paleozoic continental–marine transitions around the world.

Many excellent paleoecological–sedimentological studies have been made of carbonate reefs, such as the Permian reef complex of the Guadalupe Mountains in the southwestern United States (Newell et al. 1953), Silurian reefs of the Great Lakes (Lowenstam 1950, 1957), and Devonian reefs of Alberta (Andrichuck 1958; Klovan 1964). Figure 4.32 illustrates the lithofacies and biofacies subdivisions of some Silurian carbonate banks (J. L. Wilson 1975; Sellwood

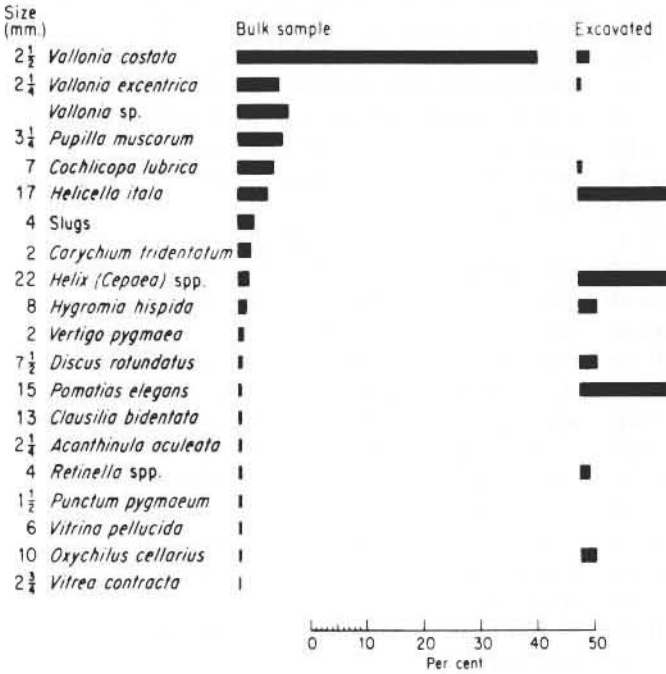


Fig. 4.28. Distribution of species in two collections made from the same outcrop by two different methods. (Ager 1963; reproduced with permission of the publisher)

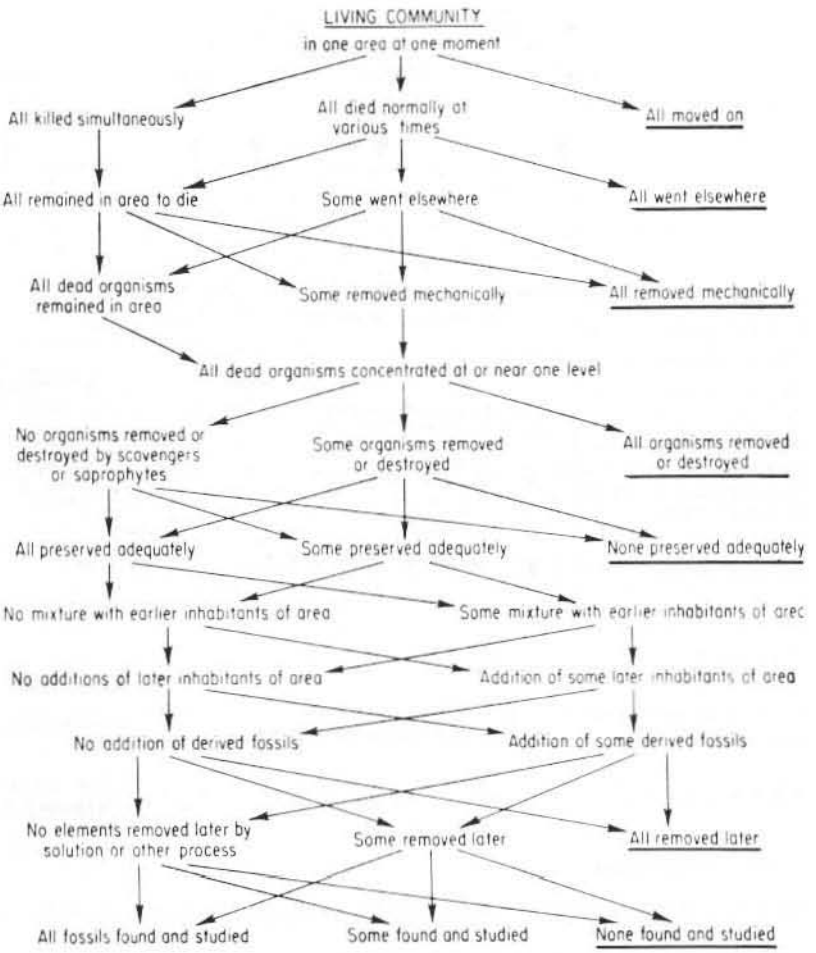


Fig. 4.29. The various possibilities for the preservation or destruction of a living animal community, and its eventual collection as a fossil assemblage. (Ager 1963; reproduced with permission of the publisher)

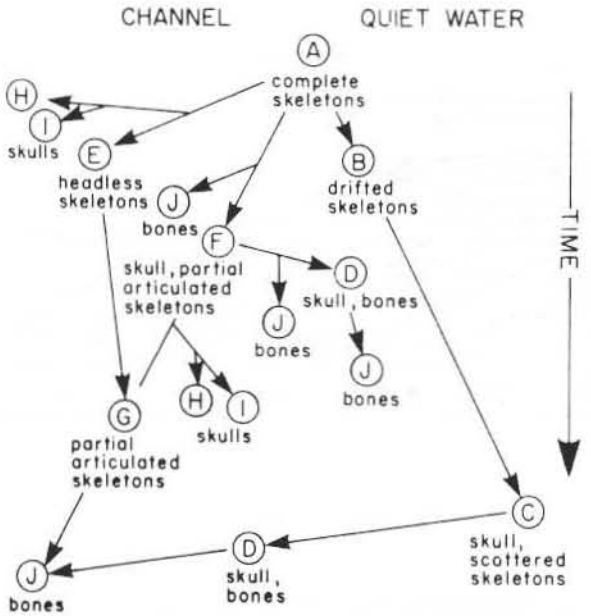


Fig. 4.30. The taphonomy of dinosaur skeletons based on studies of the Cretaceous rocks of Red Deer Valley, Alberta. (Dodson 1971)

1986). More subtle fossil–sediment relationships may only be apparent after careful statistical studies, such as the brachiopod analysis carried out by Jones (1977). In some cases, biofacies are not obviously related to lithofacies, as in the Welsh Paleozoic brachiopod assemblages discussed in Section 3.5.1.

How are these observations on biofacies actually used to indicate the depositional environment? In four main ways: (1) by comparison with the ecology of living relatives, (2) by deductions about functional morphology, (3) by deductions from position and preservation, and (4) by deductions from assemblages and associations. These lines of reasoning are not intended to be mutually exclusive and, indeed, in many ancient deposits, there is an overlap in their application.

To start with the first point, comparisons with modern relatives become more difficult the further back in time we recede, as we trace evolutionary lines back to more and more distant ancestors for whom there is no guarantee of unchanged habits. This is where examination of functional morphology, position, and associations might provide invaluable supportive

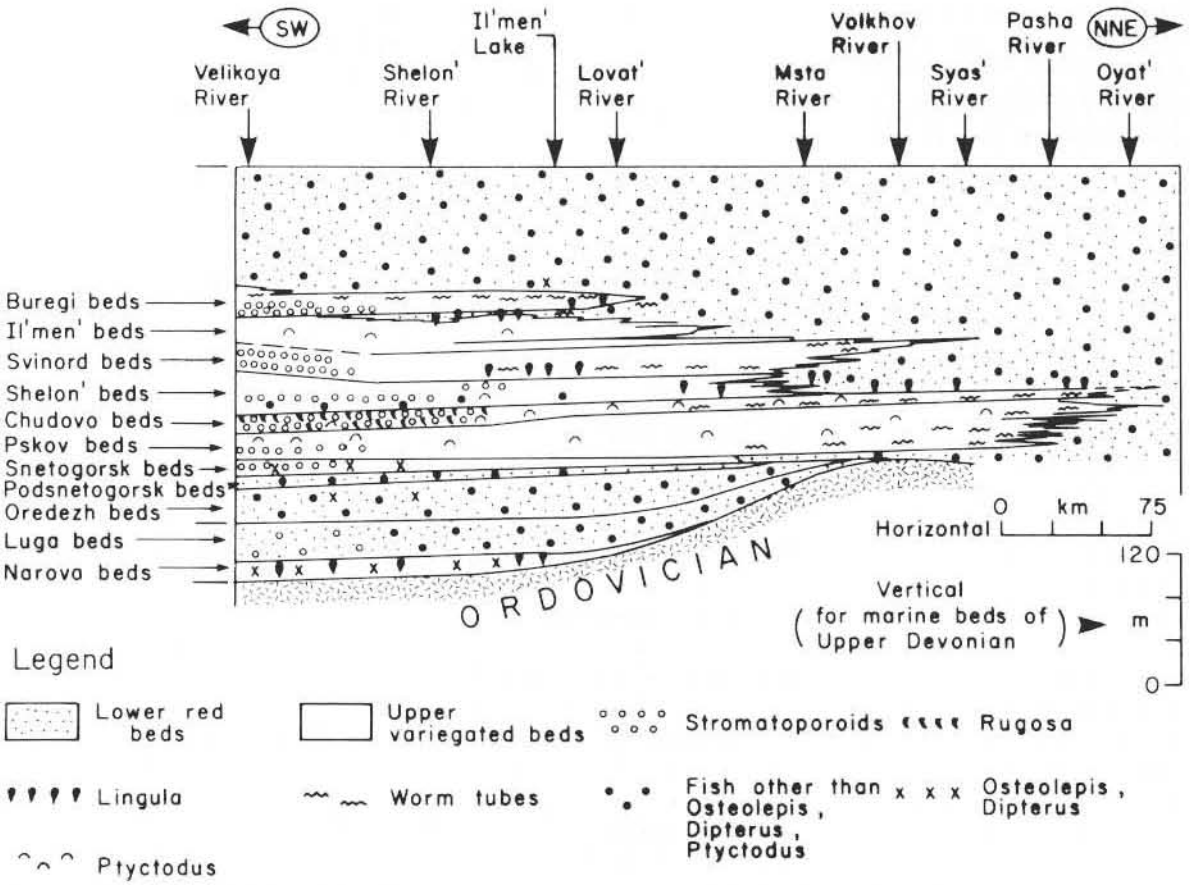


Fig. 4.31. Lithofacies–biofacies relationships for some upper Devonian rocks in Russia. (Hecker 1965)

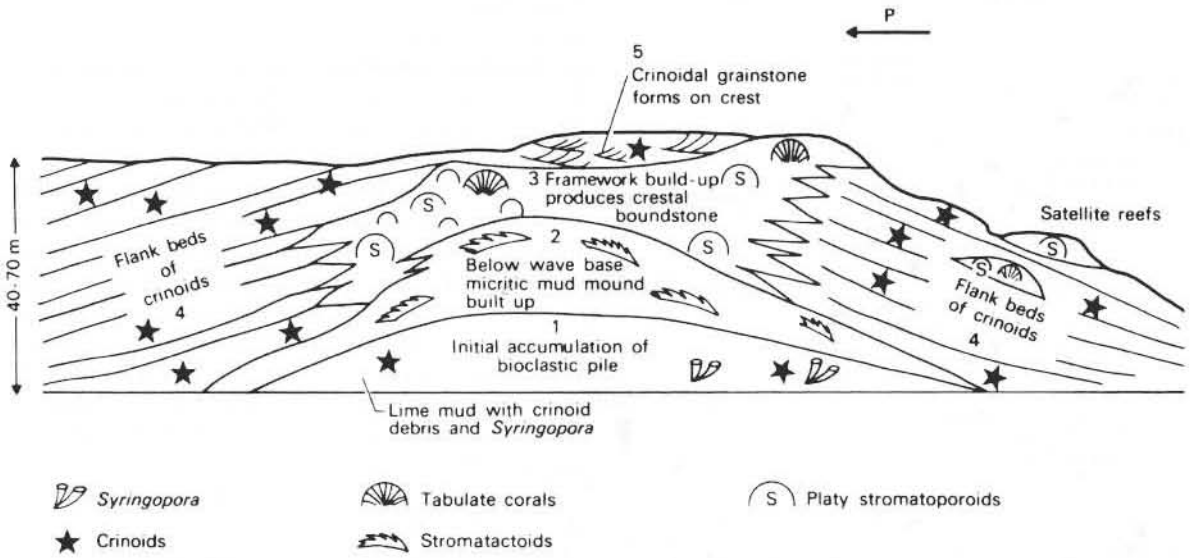


Fig. 4.32. Lithofacies-biofacies composition of carbonate buildups in the mid-Silurian shelf of the American Mid-west (Sellwood 1986; after J. L. Wilson 1975, reproduced with permission from J. L. Wilson)

evidence. The corals are a classic example of a group of organisms that has undergone marked evolutionary change during the Phanerozoic yet seems to have continued to inhabit rather similar ecological niches throughout this time. Modern hermatypic (colonial, reef forming) corals inhabit depths of water down to about 100 m but are only abundant above about 15 m. They are best adapted to grow in waters between 18 and 36 °C, with an optimum of 25–29 °C. Strong sunlight, a continuous nutrient supply, minimal salinity variations, and clean, sediment-free water are also necessary. These requirements restrict colonial corals to a present-day latitudinal range of 30° south to 30° north, with variations depending on the extent of cold or warm oceanic currents, sediment input from large rivers, etc. (James and Bourque 1992). For Mesozoic and Cenozoic corals, which are closely related to modern forms (orders Scleractinia and Alcyonaria), environmental deductions based on modern ecology seem safe. Paleozoic corals belong to two different orders, Rugosa and Tabulata, yet their gross appearance, lithofacies, and biofacies associations (particularly the symbiotic association with algae) remain similar, as is evident from the various Paleozoic reef studies quoted previously. Heckel (1972) suggested, however, that they may have occupied less agitated environments, with the now-extinct stromatoporoids filling part of this niche. Table 4.4 summarizes some other modern–ancient comparisons for carbonate-producing organisms.

A group of organisms that has become extremely important in recent years for unraveling the history of continental margins is the Foraminifera. Comparisons of ancient and modern assemblages yield invaluable paleobathymetric data for Cenozoic and later Mesozoic times, as discussed below.

Many ancient groups, such as the trilobites, graptolites, and many vertebrates, including early fish, amphibians, and reptiles, are now extinct, as are numerous genera and species of still-thriving phyla. Therefore, comparisons with the ecology of modern relatives become tenuous or impossible. Ecological and environmental interpretations may then depend partly or wholly on deductions made from functional morphology. For example, in Section 3.5.1, we discuss the evolution of the Cretaceous echinoid genus *Micraster*, which developed divergent stocks demonstrating habits of deeper and deeper burrowing into the Chalk substrate. This genus is now extinct, but Nichols (1959) was able to demonstrate the change in burrowing habit by careful examination of their functional morphology and comparison with similar features in modern echinoids. *Micraster* tests show an increase with time in the number of pores required for respiratory tube feet, an obvious adaptation to deeper burial.

There has been much debate about the habitats of trilobites, which show considerable variation in morphology. Some have large eyes, suggesting dim light (deep water?), others are blind and may have lived below the photic zone. Long horizontal spines may have been a device to inhibit sinking in a soft substrate, and the same effect may have been achieved by the broad, flat exoskeletons of some species. Others were burrowers, for example Bathyrus, which has a concave-up wedge shape adapted to wriggling into the

Table 4.4. Modern and ancient carbonate-producing organisms and their sedimentary products. (Reproduced, with permission, from James 1984a)

Modern organism	Ancient counterpart	Sedimentary aspect
Corals	Archaeocyathids, corals, stromatoporoids, byrozoa, rudistid bivalves, hydrozoans	The large components (often in place) of reefs and mounds
Bivalves	Bivalves, brachiopods, cephalopods, trilobites, and other arthropods	Remain whole or break apart into several pieces to form sand and gravel-sized particles
Gastropods, benthic foraminifers	Gastropods, tintinids, tentaculitids, salterellids, benthic foraminifers, brachiopods	Whole skeletons that form sand- and gravel-sized particles
Codiacean algae – *Halimeda*, sponges	Crinoids and other pelmatozoans, sponges	Spontaneously disintegrate upon death into many sand-sized particles
Planktonic foraminifers	Planktonic foraminifers, coccoliths (post-Jurassic)	Medium sand-sized and smaller particles in basinal deposits
Encrusting foraminifers and coralline algae	Coralline algae, phylloid algae, renalcids, encrusting foraminifers	Encrust on or inside hard substrates, build up thick deposits, or fall off on death to form lime-sand particles
Codiacean algae – *Penicillus*	Codiacean algae, *Penicillus*-like forms	Spontaneously disintegrate upon death to form lime mud
Blue-green algae	Blue-green algae (especially in pre-Ordovician)	Trap and bind fine-grained sediments to form mats and stromatolites

substrate. Ludvigsen (1978) discussed these and other morphological features and related them to trilobite biofacies zones and their position on the Ordovician platform of Western Canada.

A widely used climatic indicator is the shape of tree leaves. It is known from present-day distributions that, in tropical regions, the leaves of dicotyledonous trees are entire (smooth) whereas, in temperate regions, they are more commonly dentate. Edwards (1936) showed that this characteristic could be used to interpret fossil floras, even for extinct or unidentifiable forms. Examples are illustrated in Fig. 4.33.

Factors of abundance, state of preservation, and position may yield important environmental information. Fossil forms that normally grew attached to a substrate may occur as disoriented, broken, and abraded individuals or as hydraulically accumulated shell beds, suggesting storm-wave activity. Isolated individuals may be found out of context; for example, I have found rare, abraded, orthocone cephalopods and compound corals in laminated stromatolitic dolomites formed on a Silurian tidal flat in the Canadian Arctic. Ballance et al. (1981) reported finding coconuts in Miocene turbidites and suggested that they may have been transported offshore by tsunamis, the same events that probably triggered the turbidity currents. Fossils found in place, such as coral colonies, pelecypods in their burrows, and upright trees attached to their roots, are particularly diagnostic of depositional environment (so long as their habits are known), because such occurrences are usually devoid of any ambiguity about possible transportation into an unrelated environment. However, Jeletzky (1975) and Cameron (1975) have engaged in heated controversy over the significance of shallow-water invertebrate fossils found in nodules in a turbidite succession on Vancouver Island. Are the nodules still in place where they were formed around the fossils, or is there evidence of abrasion indicating transportation to deeper water? The geologist should be prepared for the most unlikely event. For example, upright trees, weighed down by boulders in their roots, have been found in the flood deposits of Mt. St. Helens and offshore near islands in the South Pacific, where they were emplaced by storms.

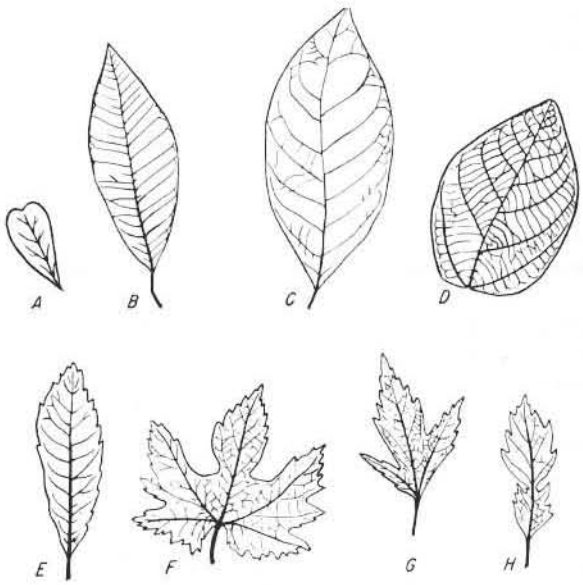


Fig. 4.33 A – H. Leaf shape as a reflection of climate. **A – D** Tropical plants typically have entire margins; assemblage from lower Eocene, southern England. **E – H** Temperate plants typically have dentate margins; assemblage from Miocene, southern France. (Edwards 1936; from Ager 1963, reproduced with permission from the publisher)

Preferred orientations of attached or transported fossils may have paleocurrent significance. Corals, crinoids, stromatolites, and other immobile organisms may grow in particular directions in response to current conditions, and accumulations of elongated fossils, such as graptolites, gastropods, logs, or belemnites, may be aligned hydraulically. This is discussed further in Chapter 5. Deductions based on preservation or position require particularly careful field observations and normally cannot be made on loose talus.

Perhaps the most reliable paleoecological interpretations are those based on assemblages and associations in an entire fossil community. Deductions made from extinct groups can be checked against information yielded by those with living relatives, and a picture of ecological relationships within the community can be attempted. Some forms may yield very generalized environmental information, including many pelagic groups, such as the graptolites, whose occurrence is governed more by energy conditions at the site of postmortem deposition than by habitat during life. Interpretations based on tolerant groups may be finely tuned if other, more selective species can be found. Particularly accurate interpretations may be possible if species with different but slightly overlapping habitat ranges can be found together. For example, in the Lower Cretaceous of northern Texas, arenaceous foraminifera are associated with the oyster *Ostrea carinata* in deposits assumed to be close to the shoreline, but the association only occurs in limestones and marls, not in mudstones. Modern arenaceous foraminifera are known to be tolerant of brackish water. *O. carinata*, therefore, appears to be tolerant of reduced salinities but not of turbid water. By contrast, the oyster *Gryphaea washitaensis* occurs in all rock types but never in association with arenaceous foraminifera. Therefore, it seems to have preferred open-marine conditions but had no preference for either turbid or nonturbid water (Laugh-

baum 1960; Ager 1963). Another example is the association of ostracodes, gastropods, and the brachiopod *Lingula* in some Silurian dolomite beds in the Canadian Arctic. This is interpreted as a brackish-water tidal-flat assemblage and contrasts strongly with the rich coral, brachiopod, crinoid, and trilobite assemblage of interbedded open-marine limestones (Miall and Gibling 1978). Heckel (1972) provided tables relating the modern distribution of fossilizable invertebrate groups to water salinity, turbidity, and depth (Fig. 4.34) and, within each group, individual genera or species may have much more restricted tolerance ranges, as in the various examples given above.

Some of the most detailed assemblage studies carried out for basin-interpretation purposes are those on ancient reefs of all ages and on Cenozoic foraminifera. For example, Fig. 4.35 shows the vertical changes in assemblage through Niagaran reefs of the Great Lakes (Lowenstam et al. 1956). The reefs were built on a soft substratum in quiet water and gradually extended up into the zone of breaking waves.

Studies of foraminifera paleoecology began in the Cenozoic of the Gulf Coast and California in the 1930s (Natland 1933). The initial stimulus for the work was the need to correlate petroleum-bearing sediments, and this remains one of its most important applications. However, the same data are now finding increasing application in the interpretation of the subsidence history of continental margins, a subject of considerable relevance to our attempts to understand the mechanisms of plate tectonics. An excellent example of this is the work on subsurface Cenozoic strata of the continental margin off Labrador and Baffin Island by Gradstein and Srivastava (1980). They distinguished four biofacies and interpreted their environments based on associated faunas and comparisons with other modern and ancient foraminifera assemblages. Considerable use was made of Deep-Sea-Drilling Project (DSDP) data. The biofacies are as follows:

1. Nonmarine: spores and pollen, no foraminifera.
2. Shallow neritic: marginal marine to inner shelf (<100 m deep). Diagnostic foraminiferal assemblages are of very low generic and specific diversity, with rare or no planktonics. In the late Neogene sections, *Cibicidiodes*, *Elphidium*, *Cassidulina*, *Bulimina*, *Melonis*; gastropods and bryozoans occur in low numbers.
3. Deep neritic: 100–200 m water depths. Foraminiferal generic and specific diversity varies, planktonics occur locally. Cenozoic benthonic genera include *Uvigerina*, *Pullenia*, *Gyroidina*, *Alabamina*. Paleogene assemblages are impoverished.

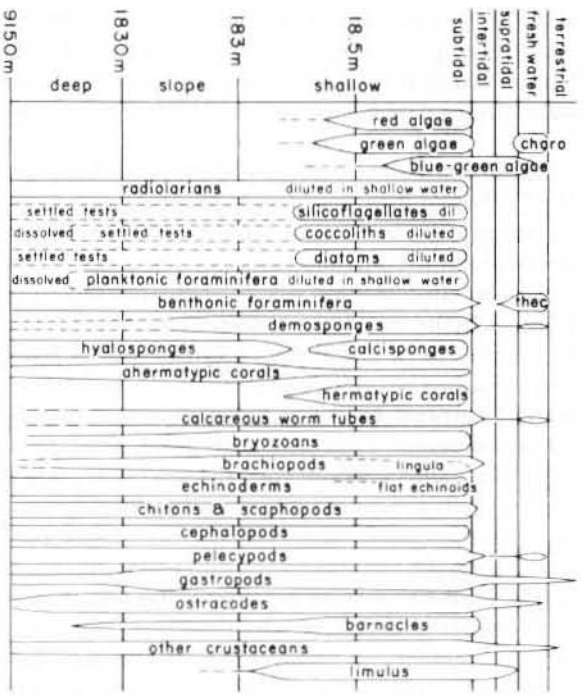


Fig. 4.34. Depth distribution of common types of fossil invertebrates. (Heckel 1972)

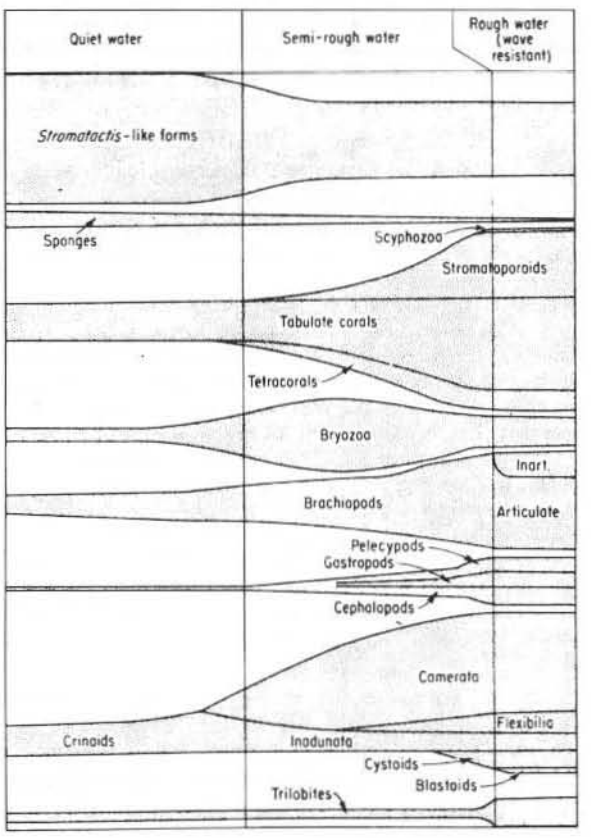


Fig. 4.35. Vertical variations in fossil assemblages of Niagaran reefs, Great Lakes. (Lowenstam et al. 1956)

4. Bathyal: upper slope, water depths 200–1000 m. Foraminiferal generic and specific diversity is high, and the assemblages are dominated by coarse, often large-sized, agglutinated taxa. Over 50 species, including rich *Cyclammina* species. Locally, planktonics are abundant (mostly turborotaliid and globigerinid forms). In a few wells, Paleogene calcareous benthonics, including *Pleurostomella*, *Osangularia*, *Stilostomella*, and *Nuttalides*, occur in low numbers.

Gradstein and Srivastava (1980) constructed subsidence-history plots for five wells off the Newfoundland–Labrador coast, one of which is shown in Fig. 4.36. The line defining the upper boundary of the shaded area represents changes in water depth with time, as determined from biofacies data. The lower line, defining the locus of the pre-Cretaceous basement unconformity, was constructed by subtracting the cumulative sediment thickness from the water-depth curve. It can be seen that subsidence rates have varied markedly since the Cretaceous, the fastest subsidence occurring in the Eocene, which geophysical evidence indicates was a time of active sea-floor spreading in the immediately adjacent oceanic crust. Data such as this may make important contributions to the techniques of **backstripping** or **geohistory** analysis. These are methods used to reconstruct basin subsidence as an aid to the analysis of crustal behavior during basin formation (Sect. 7.3.2). Benedict and Walker (1978) attempted a similar type of paleobathymetric analysis of a Paleozoic sequence using chemical, sedimentary, and paleoecological criteria.

Trace fossils are widely used environmental indicators in marine sediments, where they are most abundant. They have one advantage over body fossils in that they are almost invariably in place; derived and transported trace fossils are very rare. They are well preserved and readily studied in drill cores and have, therefore, made a substantial contribution to subsurface environmental analysis Trace fossils may be used in two ways: (1) to indicate local sedimentation and erosion patterns and (2) to indicate the depositional environment. Trace-fossil assemblages have become valuable tools in the analysis of discontinuities and their allostratigraphic significance. The study of **ichnofacies** has become a specialized form of paleoecology. It is widely used in subsurface studies because of the ease with which trace fossils may be studied in cores. Many recent review articles and textbooks have been written on this subject, among the most useful of which are those by Frey and Pemberton (1985) and Pemberton et al. (1992).

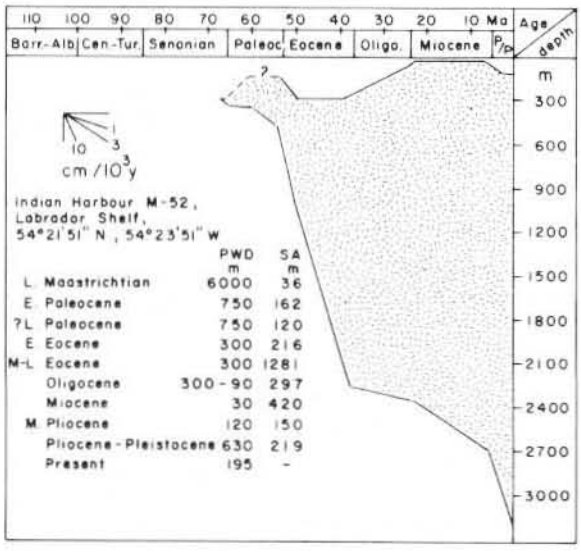


Fig. 4.36. Subsidence curve for an offshore well, Labrador shelf. See text for explanation. (Gradstein and Srivastava 1980)

Howard (Frey 1975) discussed the ways in which trace fossils may be used to evaluate local erosion and sedimentation. The density of bioturbation in a bed varies inversely with the rate of sedimentation so that, where sediment supply is low and invertebrate life abundant, primary structures such as bedding may be completely destroyed. Beds formed by rapid sedimentation, such as the passage of a turbidity current or a storm, may contain little or no penetration by organisms except, perhaps, for escape burrows made vertically through the bed. Types of bioturbation may vary from one lithology to the next because of different behavior patterns of the organisms in response to different sediment types. Some examples of specific responses of selected burrow types to varying sedimentation patterns are shown in Fig. 4.37, based on observations by Goldring (1964).

Important and widely quoted work by Seilacher (1967) showed how trace fossils could be used to interpret paleobathymetry. A set of five biofacies (now termed ichnofacies) were erected. These have now been expanded to nine. Six of these are shown in Fig. 4.38. In the shallowest waters, waves and currents keep nutrients in suspension. Animals such as *Skolithus* are subject to violent conditions and build deep, vertical burrows. In less turbulent waters, faunal diversity increases, sediment feeding becomes more important, and a variety of crawling and grazing feeding trails appear (*Cruziana* biofacies). Below the wave base, sedimentation is slower, the oxygen content of the sediments lower, and nutrient supply more sparse. Sediment-mining organisms of the *Zoophycos* and *Nereites* biofacies are characteristic. The latter, with its complex but highly systematic grazing patterns, is particularly typical of abyssal submarine-fan deposits. The other three assemblages include a nonmarine *Scoyenia* biofacies, a "softground" *Psilonich-*

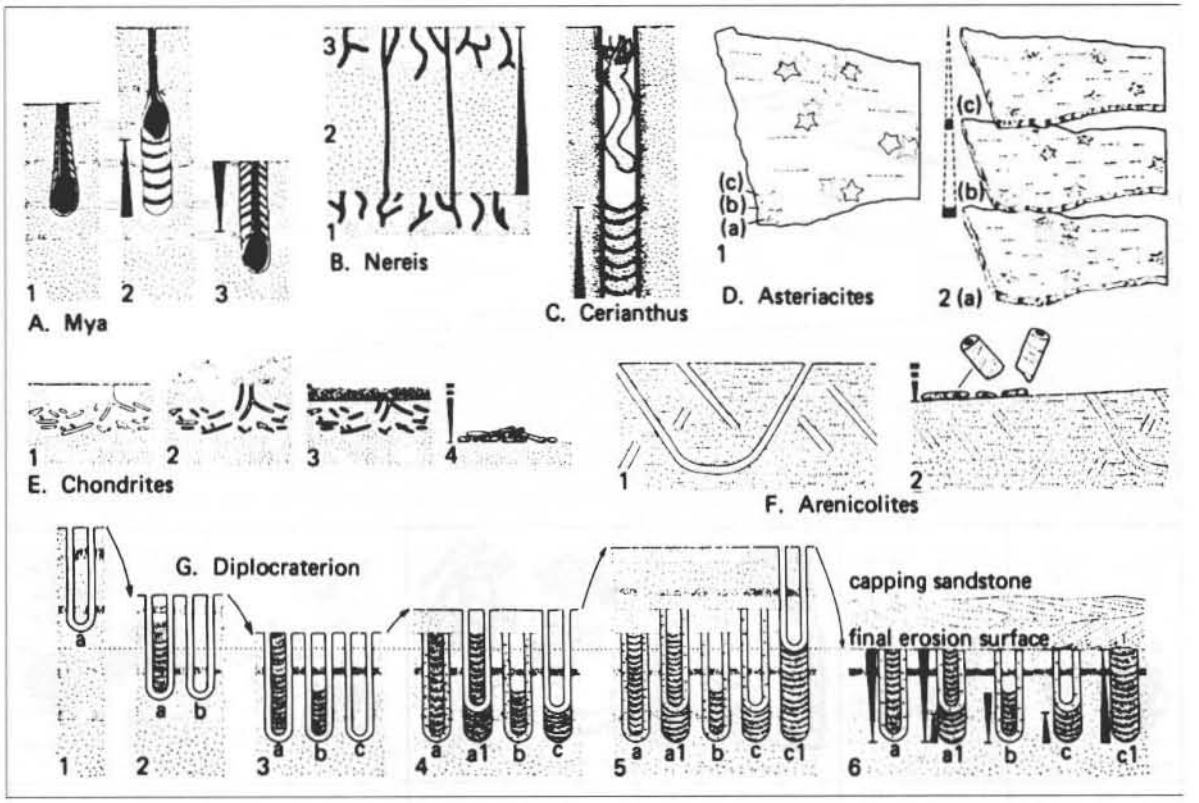


Fig. 4.37 A–G. Amount of sedimentation or erosion, as indicated by adjustments to depth and modes of preservation of various lebensspuren. Heights of *solid arrows* show amount of deposition or erosion. A Movement pattern of the pelecypod *Mya*, which has a single siphon. With a stationary sedimentary surface (*1*), the growing organism gradually burrows deeper; the bottom of the resulting structure is wider than the top. With rapid sedimentation (*2*), the organism migrates toward the surface, leaving an infilled burrow the width of the shell. With degradation of the surface (*3*), the organism migrates downward, leaving a burrow of the same width as in *2* but having a different internal structure. B Movement pattern of the polychaete worm *Nereis*. An older colonized surface (*1*) is rapidly covered by sediment (*2*) and, during deposition, paths of escape are directed upward. With stabilization, the new colonization surface (*3*) has irregular normal burrows. Structures in (*1*) and (*3*) are generally mucus lined; in (*2*), they are unlined. C Movement pattern of the anemone *Cerianthus*, an organism dwelling in a single tube. With sedimentation, the animal moves upward, leaving an unfilled or passively filled burrow. A similar pattern might be expected in the traces of *Skolithus* and *Monocraterion*. D Movement represented by the trace fossil *Asteriacites lumbricalis*, resting place of a stelleroid. With sedimentation, the animal migrates upward, in stages *a–c*; shown are a combined (*1*) and separate (*2*) plan of all impressions. E Preservation patterns of the trace fossil *Chondrites*. The animal's tunnel system (*1*) is infilled (*2*) following a change in the type of sediment being deposited (bed-junction preservation). Slight degradation of the surface (*3*) removes the proximal shafts before further sediment of a different type accumulates (concealed bed-junction preservation). Renewed degradation of the surface winnows away sediment, leaving mucus-lined infilled tunnels as burial preservations (*4*). F Preservation pattern of the trace fossil *Arenicolites curvatus*. A sedimentary surface containing open U tubes (*1*) has been degraded (*2*), the mucus-cemented tube fragments accumulating in an intraformational conglomerate; sediment has filled the tubes. G Movement pattern represented by the trace fossil *Diplocraterion yoyo*. In the upper Devonian Baggy Beds of North Devon, this trace occurs in the various configurations shown in stage *6*; all have been truncated to a common erosion surface. Repeated phases of erosion and sedimentation (*1–5*) evidently led to the development of the various types. Stage *1* shows the development of a burrow (*a*). With degradation of the surface, this tube migrates downward and, at intervals, new tubes (*b*, *c*) are constructed (*2*, *3*). Sedimentation follows (*4*, *5*), but some tubes are abandoned. In stage *6*, all tubes are abandoned, and erosion reduces them to a common base. (Goldring 1964)

nus assemblage, the *Skolithus* and *Cruziana* assemblages, and the *Glossifungites* biofacies, which develops on firm but unlithified substrates. The details of the biofacies distributions vary from basin to basin and depend on sediment type, water temperature, salinity, and circulation patterns. Frey and Pemberton (1984) pointed out that

> local sets of environmental factors are most important in controlling the distribution of tracemakers, whether or not these parameters occur at specific water depths. For instance, many of the estuarine point bars ... exhibit a high-energy, channelward side typified by a *Skolithus* associa-

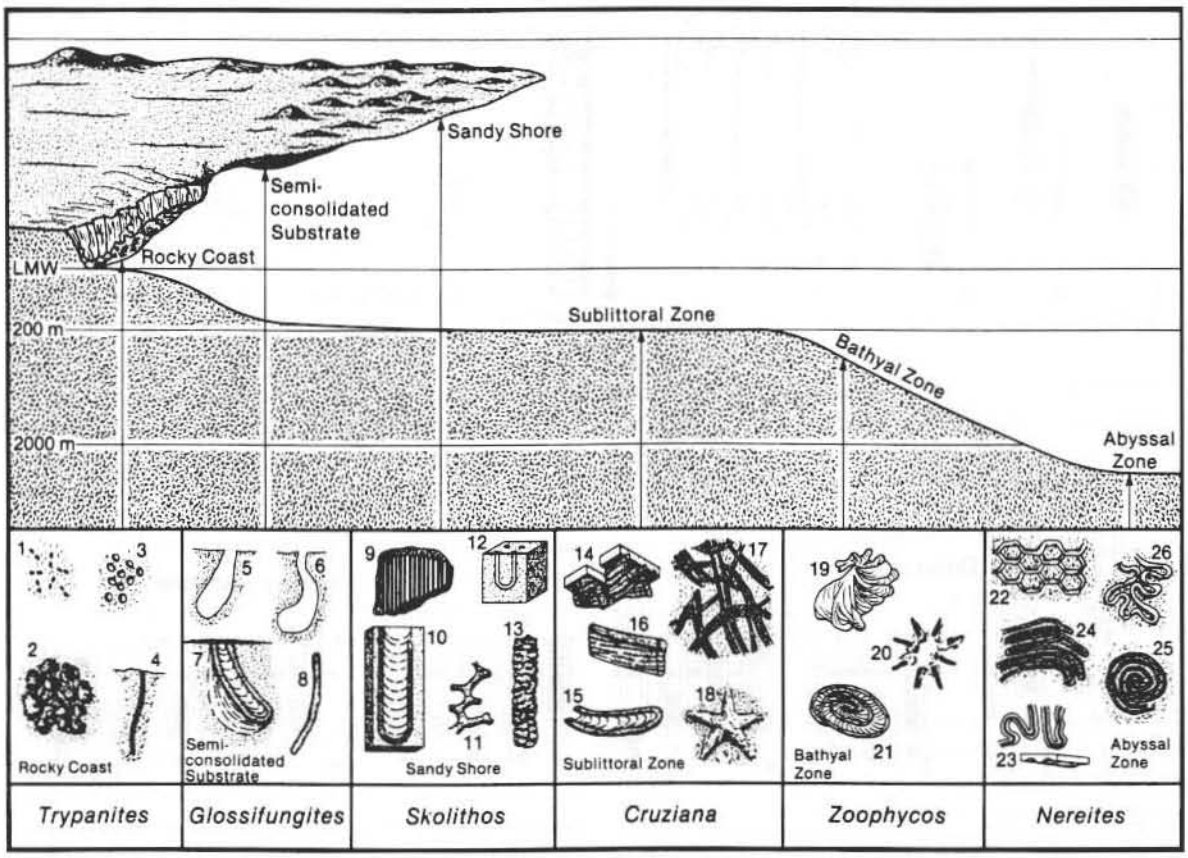


Fig. 4.38. Six marine trace fossil ichnofacies shown in representative environmental settings. Typical trace fossils include *1, Caulostrepsis*; *2, Entobia*; *3*, Un-named echinoid borings; *4, Trypanites*; *5, 6, Gastrochaenolites*; *7, Diplocraterion*; *8, Psilonichnus*; *9*; *Skolithus*; *10, Diplocraterion*; *11, Thalassinoides*; *12, Arenicolites*; *13, Ophiomorpha*; *14, Phycodes*; *15, Rhizocorallium*; *16, Teichichnus*; *17, Crossopodia*; *18, Asteriacites*; *19, 21, Zoophycos*; *20, Lorenzinia*; *22, Paleodictyon*; *23*; *Taphrhelminthopsis*; *24, Helminthoida*; *25, Spirorhaphe*; *26, Cosmorhape*. (Pemberton et al. 1992)

tion and a low-energy bankward side typified by a *Cruziana* association ... The respective associations occur in close proximity, at the same stratigraphic or bathymetric datum.

Also, as Howard (Basan 1978) pointed out, different organisms may make very similar structures and, conversely, the same species may make different structures when engaged in different activities or when interacting with substrates of different composition. Therefore, interpretations must always be made in conjunction with other facies studies, and the geologist would be wise to consult published work on trace-fossil assemblages in other basins of a similar age.

In carbonate environments, algal mats and stromatolites generally are excellent indicators of intertidal deposition, although recent discoveries of stromatolites in subtidal settings has indicated that an environment hostile to browsing organisms may be the critical control. According to James (1984b), their "upper limit is controlled by climate; in arid areas they cannot grow above the high intertidal into the supratidal zone, whereas in areas of high rainfall where the supratidal zone is moist or flooded for days at a time, mats are prolific. The lower limit is more variable and appears to be controlled by the presence of gastropods that eat algae." In hypersaline zones, gastropods cannot survive, and mats grow into the subtidal zone. After deposition and burial, mats commonly rot away, but they leave voids as a result of the disappearance of organic materials, entrapped gas, or shrinkage. The resulting, distinctive type of porosity has been referred to as laminoid fenestrate, loferite, or bird's eye. Stromatolites are present well back into the Precambrian and may have lived in similar environments throughout this time.

4.5.8 Vertical Profiles

The importance of Walther's Law and of the vertical profile in facies analysis is discussed in Section 4.4.1.

Methods of statistical analysis of cyclic sedimentation are also described briefly in this section. The recognition of cyclic sequences has become one of the most widely used tools for reconstructing depositional environments in the subsurface. One of the reasons for this is that cyclic changes are commonly readily recognizable in petrophysical logs and are often interpreted without access to core or well cuttings. Whether such interpretations are always correct is another question. The sedimentological literature is full of shorthand references to **fining-upward** or **coarsening-upward** cycles or to **fining-and-thinning** or **coarsening-and-thickening upward**. The grain size, bed thickness, and scale of sedimentary structures are commonly correlated in clastic rocks, so that the cyclicity may be apparent from several types of observation. French sedimentologists tend to use the terms positive and negative cycles; dipmeter analysts recommend coding cyclic changes diagrammatically in red and blue, but I am embarrassed to have to admit that I am not sure if fining-upward cycles are positive and red or negative and blue or the opposite and (or) vice versa. Such terms are obviously not helpful if one cannot remember which way they are used, and a simple descriptive terminology seems preferable.

An important warning: cyclicity may be in the eye of the beholder. Many facies interpretations in sedimentology place considerable emphasis on the recognition of repeated, ordered vertical change in facies characteristics, and many well-known facies models have been built upon such interpretations. However, as Hiscott (1981) demonstrated, with respect to submarine fan deposits, the presence of cyclicity may be based on very subjective interpretations. Anderton (1995) argued that most submarine fan deposits are non-cyclic. He stated "the human brain is very good at pattern recognition. It seldom misses a real pattern and can often see patterns in random nose. Thus, objective correlation tends to be over-correlation." Wilkinson et al. (1998) took issue with the interpretations of cyclicity in certain shelf-carbonate successions, descriptions of which have formed the basis for some important studies in sequence stratigraphy. They argued that other natural, random processes can be demonstrated to generate very similar patterns. The solution may be to apply rigorous, statistical tests to the data. Harper (1998) provided such tests for the examination of thickening- and thinning-upward patterns.

There are two common, basic types of cycle: those indicating an increase in transport energy upward and those demonstrating a decrease. Both types can be caused by a variety of sedimentary, climatic, and tectonic mechanisms. Beerbower (1964) divided these into **autocyclic** and **allocyclic** controls. To avoid the connotation of cyclicity some writers prefer the terms **autogenic** and **allogenic**. Autocyclic mechanisms are those that result in the natural redistribution of energy within a depositional system. Examples include: the meandering of a channel in a river, tidal creek, or submarine fan; subaerial flood events; subaqueous sediment-gravity flows; channel switching on a subaerial or submarine fan or a delta (avulsion); storms; and tidal ebb and flood. All of these can potentially produce cyclic sequences. Allocyclic mechanisms are those in which change in the sedimentary system is generated by some external cause. Eustatic sea-level and climatic changes, and tectonic control of basin subsidence, sediment supply, and paleoslope tilt are the principal types of allocyclic mechanisms. These are large-scale basinal sedimentary controls and are dealt with mainly in Chapters 6 and 8. A sedimentary basin may be affected by several of these processes at the same time, so it is not uncommon to find that there are two or three scales of cyclicity nested in a vertical profile. Allocyclic cycles tend to be thicker and more widespread in their distribution than autocyclic cycles. The latter are generally formed only within the confines of the subenvironment affected by the particular autocyclic process. This assists the geologist in distinguishing and interpreting sedimentary cycles, a matter of some importance in the definition of various scales of sequences and **parasequences**, but such interpretations may be far from easy.

Vertical profiles formed a crucial component of the first facies models, including the point-bar model of Allen (1963) and Bernard and Major (1963), the barrier-island model of Bernard et al. (1959), and the Bouma (1962) turbidite sequence. They were the focus of a classic paper on facies models by Visher (1965), and their recognition in the subsurface has been covered in many articles and textbooks (Fisher et al. 1969; Pirson 1977; Sneider et al. 1978; Selley 1979). They still provide an essential basis of many modern facies models (Walker and James 1992; Reading 1996), because they provide a simple way of synthesizing many different types of observation, including all the environmental criteria described in this chapter.

Vertical profiles have found an important use in recent years in the definition and documentation of stratigraphic sequences. As sea level rises and falls, coastal and shallow-marine depositional environments may migrate landward and seaward, respectively. The result is a corresponding vertical succession of facies assemblages recording the drowning and re-establishment of the various coastal environments. Coarsening-upward and fining-upward cycles may demonstrate the presence of deltaic, estuarine, or beach-barrier systems at the shoreline, depending

on the local paleogeography. Shoaling-upward successions bounded by marine flooding surfaces have been termed parasequences by sequence stratigraphers, and the nature of their vertical stacking is an important component of sequence analysis (Van Wagoner et al. 1987). Where younger successions extend successively further into the basin, they are said to exhibit a **progradational stacking pattern**, and this is interpreted to indicate a rate of sediment input more rapid than the rate at which depositional space (**accommodation**) is being generated by basin subsidence or sea-level rise. **Aggradational** (vertically stacked) and **retrogradational (backstepping)** patterns indicate a balance between sediment delivery and accommodation and a deficit in sediment delivery, respectively. The latter condition is particularly common during transgression (Chap. 6). Lithostratigraphic and chronostratigraphic techniques may be used to correlate sequences and their component facies successions across a basin in order to determine the regional history of sea-level change (Chaps. 6, 8).

Coarsening- and thickening-upward cycles are the most varied in their origins. Several distinct types are produced by coastal regression and progradation (lateral accretion), where there is a gradation from low-energy environments offshore to higher energy in the shoaling-wave and intertidal zones. Examples are illustrated in Fig. 4.39. Other types are formed where there is a steep slope and abundant sediment supply, and the flow system that develops over it attempts to grade itself to a balance by filling in the basin margin with a wedge of sediment. Coarsening-upward cycles are formed under these circumstances by prograding submarine fans and alluvial fans, particularly where the relief is maintained or even increased by active tectonic uplift. Other examples of coarsening-upward cycles (not illustrated) are those produced by crevasse splays in fluvial and deltaic settings, washover fans building from a barrier landward into a lagoon, and fluvioglacial sequences formed in front of an advancing continental ice sheet.

Fining- and thinning-upward cycles commonly occur in fluvial environments as a result of lateral channel migration (point-bar succession) or vertical channel aggradation. Alluvial fans may also show fining-upward cycles where they form under conditions of tectonic stability. These three types are shown in Fig. 4.40. Other illustrated examples are the

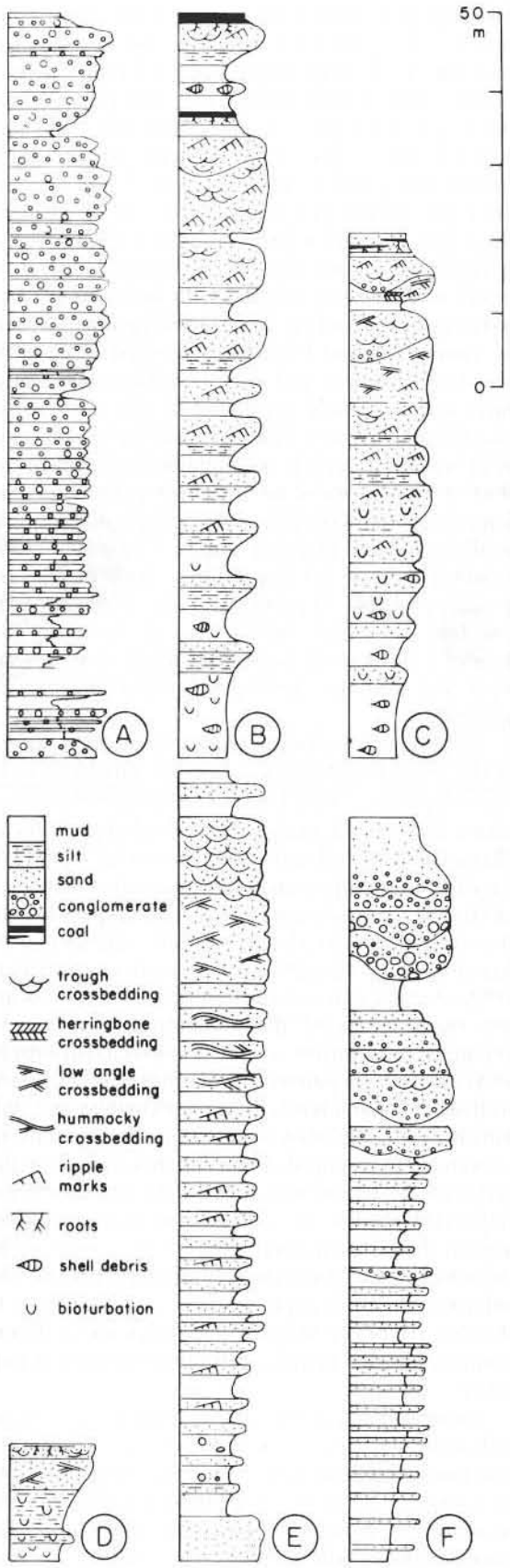


Fig. 4.39 A–F. Typical examples of thickening- and coarsening-upward profiles. **A** Prograding alluvial fan (Steel et al. 1977). **B** River-dominated delta (Miall 1979). **C** Wave-dominated delta (Miall 1979); **D** Barrier island – the Galveston Island model (Davies et al. 1971). **E** Prograding, storm-dominated shoreline (Hamblin and Walker 1979). **F** Submarine fan. (Walker 1984b)

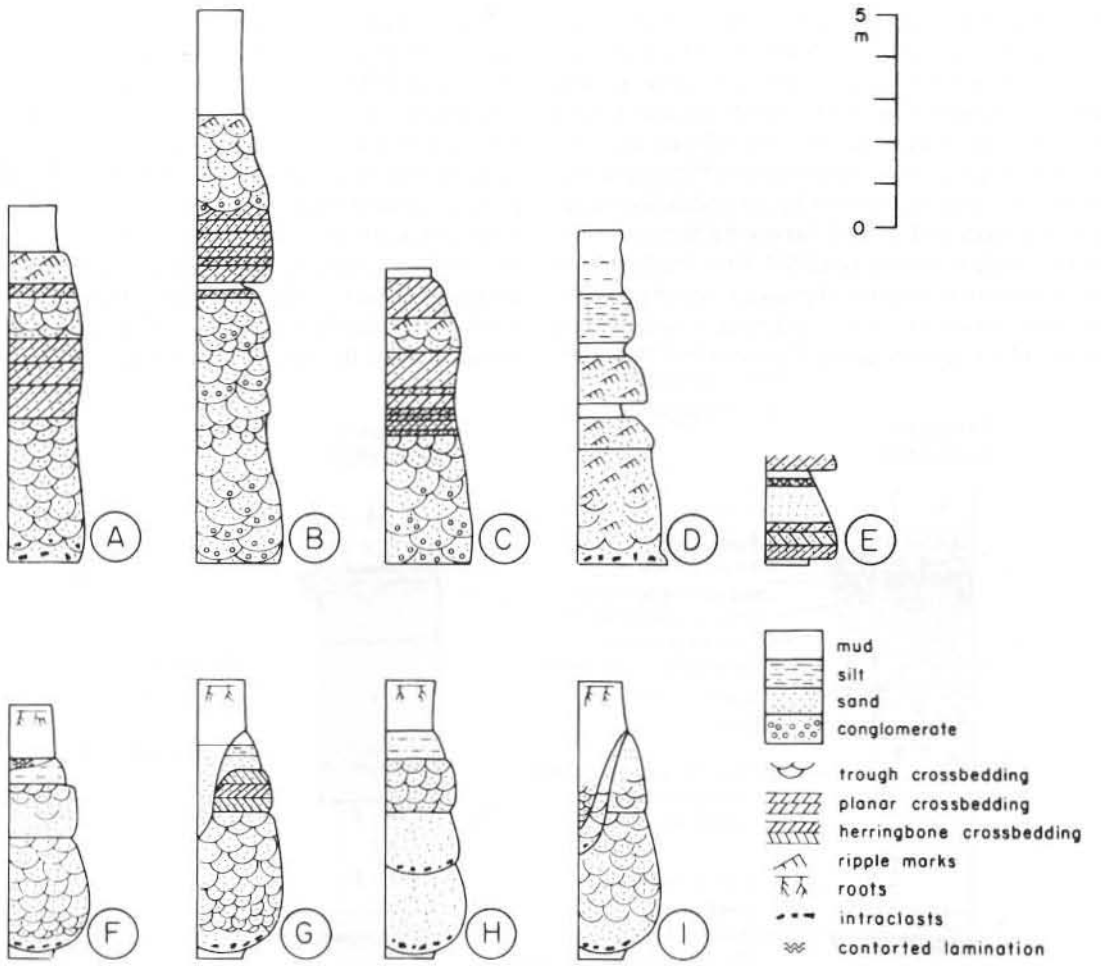


Fig. 4.40 A–I. Typical examples of thinning- and fining-upward profiles. **A** Sandy, braided river. **B, C** Point bars in high-sinuosity rivers. **D** Degrading alluvial fan. **E** Sandy tidal flat. **F–I** Tidal-creek point bars. (**A–D** from Miall 1980; **E** from Klein 1971; **F–I** from Barwis 1978)

tidal-creek point-bar and intertidal beach progradation sequences. Sediments deposited by catastrophic runoff events, including fluvial flash floods and debris flows and many types of subaqueous sediment gravity flows, also show a fining-upward character.

Many of these cycles are superficially similar, and it may require careful facies and paleocurrent studies to distinguish them. Information on lateral variability may be crucial, but this may be difficult to obtain in the case of subsurface studies.

For carbonate environments, less emphasis has traditionally been placed on the vertical facies succession or profile and more on the grain type, faunae, and structures of individual beds. Assemblages of such attributes are commonly environmentally diagnostic whereas, in the case of siliciclastic sediments, much ambiguity may be attached to their interpretation, and such additional features as vertical profile and lateral facies relationships assume a greater importance. The range of environments in which carbonates are formed is much narrower than that of siliciclastics; they are mainly confined to shallow continental shelves, platforms, or banks and adjacent shorelines and continental-margin environments. Nevertheless, enormous variability is apparent in these various settings, particularly in shallow-water and coastal regions, and this is another reason why standard vertical-profile models have not become as popular generally as they have with clastic sedimentologists.

James (1984b) and Ginsburg (1975) discussed shoaling-upward successions formed in shallow-subtidal to supratidal settings. These are common in the ancient record, reflecting the fact that the rate of carbonate sedimentation is generally much greater than the rate of subsidence. Shallowing-up sequences, therefore, repeatedly build up to sea level and prograde seaward. Lateral shifts in the various suben-

vironments are common. James (1984b) offered four generalized sequences as models of vertical profiles that could develop under different climatic and energy conditions (Fig. 4.41). Ginsburg and Hardie (1975) and Ginsburg et al. (1977) developed an exposure index representing the percent of the year an environmental zone is exposed by low tides. By studying tide gauges and careful surveying of part of the modern Andros Island tidal flat, they were able to demonstrate that a variety of physical and organic sedimentary structures is present over a surprisingly narrow tidal exposure zone. This idea has considerable potential for interpreting shoaling-upward sequences (Fig. 4.42), as demonstrated by Smosna and Warshauer (1981). Lofer cycles of the Alpine Triassic developed under conditions of fluctuating water level. Most sedimentation occurs during progradation, and these are, therefore, unusual in being deepening-upward cycles (Fischer 1964; Wilson 1975). The vertical profile is illustrated in Fig. 4.43. The style of carbonate cyclicity is a sensitive indicator of the balance between carbonate sedimentation and sea-level change and has received considerable attention in recent years because of the potential for the cyclic-

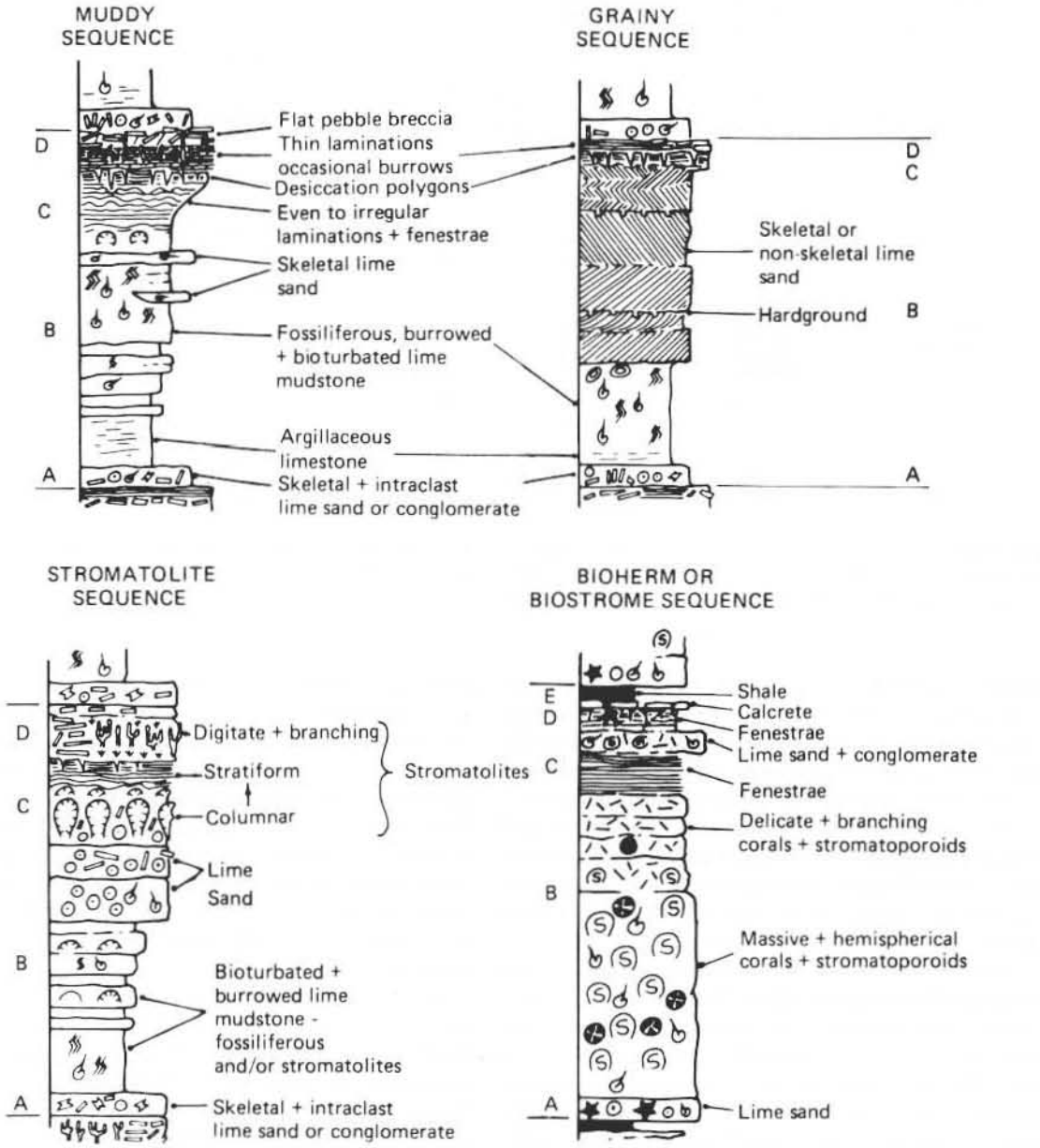


Fig. 4.41. Examples of shoaling-upward carbonate succession formed in shallow subtidal to intertidal environments. (James 1984b)

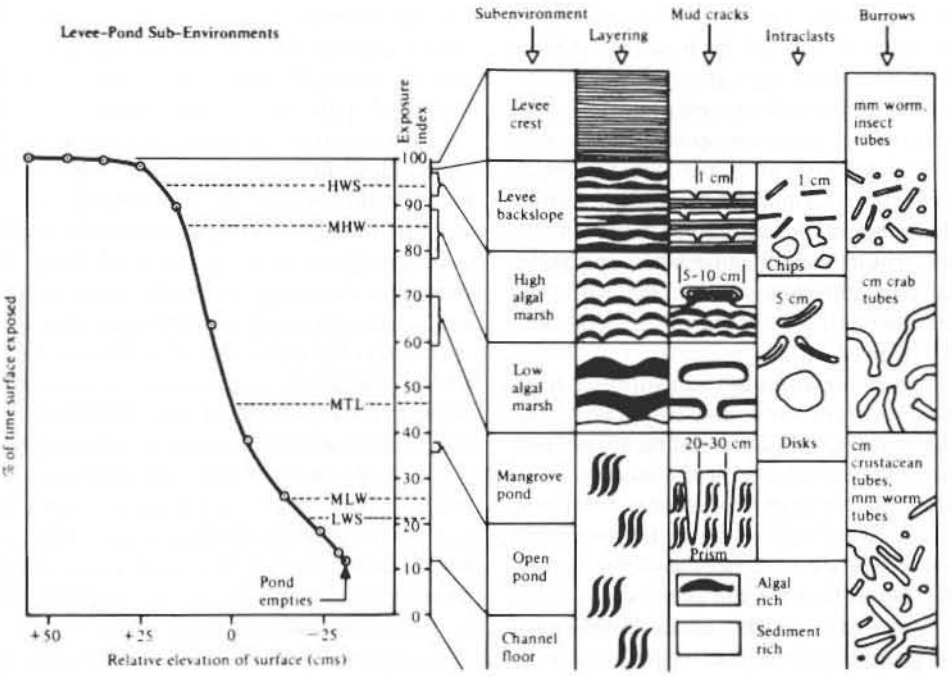


Fig. 4.42. Zonal distribution of sedimentary structures on a tidal flat and their relationship to the exposure index. Symbols are self explanatory except that the sigmoidal ornament at the base of the layering column indicates bioturbation. (Ginsburg and Hardie 1975)

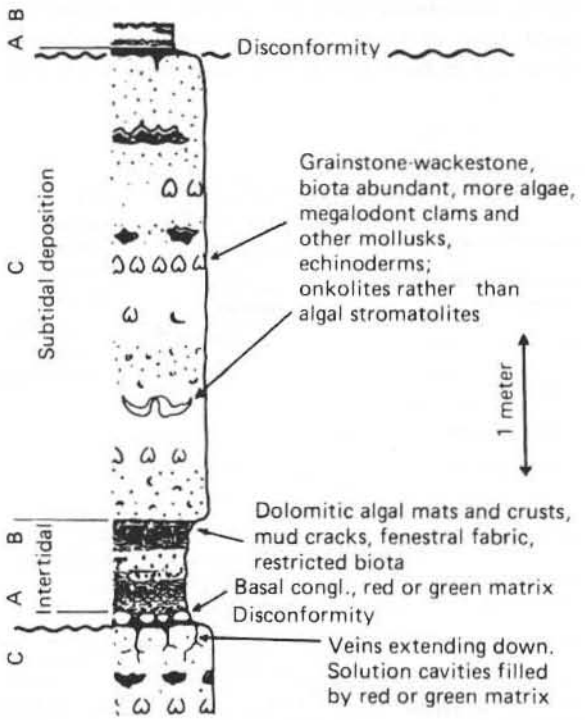


Fig. 4.43. A typical Lofer cycle. The C member typically contains cavities produced by desiccation and solution during the succeeding drop in sea level. (Fischer 1964; J. L. Wilson 1975)

ity to throw light on allocyclic sequence-generating mechanisms, including climate changes driven by orbital-forcing mechanisms (Chap. 6).

Carbonate buildups or reefs may contain an internal cyclicity that is the result of upward reef growth. James and Bourque (1992) suggested that the vertical profile may show an upward transition from an initial pioneer or stabilization phase to colonization, diversification, and domination phases, characterized by distinctive textures and faunae. In practice, most ancient reefs are the products of numerous sedimentation episodes separated by diastems or disconformities, attesting to fluctuating water levels (e.g., upper Devonian reefs of Alberta; Mountjoy 1980). Analysis of vertical profiles of repeated cyclic patterns may not, therefore, be very helpful for basin-analysis purposes, although such work may be useful for documenting small-scale patterns of reef growth (Wong and Oldershaw 1980).

Deep-water carbonates comprise a variety of allochthonous, shelf-derived breccias and graded calcarenites, contourite calcarenites, and hemipelagic mudstones cut by numerous intraformational truncation (slide) surfaces (Cook and Enos 1977; McIlreath and James 1984; Coniglio and Dix 1992). Slope sedimentation is commonly most rapid during times of high sea level, when abundant carbonate material

is being generated on the platform and shed from the margins, the process termed **highstand shedding** (Schlager 1991). The lithofacies assemblages are distinctive, but variations in slope topography and the random occurrence of sediment-gravity flows seem to preclude the development of any typical vertical profile. Organic stabilization and submarine cementation of carbonate particles probably prevent the development of carbonate submarine fans comparable to those formed by siliciclastic sediments, with their distinctive channel and lobe morphology and characteristic vertical profile.

Cyclic sequences are common in evaporite-bearing sediments, reflecting a sensitive response of evaporite environments to climatic change, brine level, or water chemistry. Vertical-profile models are, therefore, of considerable use in environmental interpretation. One of the most well known of these is the coastal sabkha, based on studies of modern arid intertidal to supratidal flats on the south coast of the Persian Gulf (Shearman 1966; Kinsman 1969; Kendall 1992). Coastal progradation and growth of displacive nodular anhydrite results in a distinctive vertical profile that has been widely applied (indeed, overapplied) to ancient evaporite-bearing rocks (Fig. 4.44). Kendall (1992) discussed variations in this profile model, reflecting differences in climate and water chemistry that arise in other coastal and playa-lake margin settings.

As noted elsewhere, evaporites can occur in a variety of other lacustrine and hypersaline-marine settings. They mimic many kinds of shallow- to deep-marine carbonate and siliciclastic facies, and a range of sedimentary criteria is required to demonstrate origin. The vertical profile is only one of these but may be useful, particularly when examining subsurface deposits in cores. For example, sulphates that accumulate below the wave base commonly display a millimeter-scale lamination interbedded with carbonate and organic matter and possibly including evaporitic sediment-gravity-flow deposits (Kendall 1992). The latter may even display Bouma sequences (Schreiber et al. 1976). Shoaling-upward intertidal to supratidal cycles have been described by Schreiber et al. (1976) and Vai and Ricci-Lucchi (1977) in Messinian (upper Miocene) evaporites of the Mediterranean basin (Fig. 4.44). Caution is necessary in interpreting these cycles, because they may not indicate a buildup or progradation under stable water levels but instead may be the product of brine evaporation and falling water levels. Many cycles of recharge and evaporation have been proposed for major evaporite basins such as the Mediterranean (Hsü et al. 1973).

Lacustrine environments are, in general, characterized by a wide variety of vertical profiles, reflecting many cyclic processes involving changes in water level and water chemistry. Many of these contain a chemical sediment component. Lakes are highly sensitive to climate change, and their sediments have, therefore, become important in the investigation of orbital-forcing mechanisms (Chap. 8). Van Houten (1964) described a shoaling-upward, coarsening-upward type of cycle in the Lockatong Formation (Triassic) of New Jersey. The cycles are about 5-m

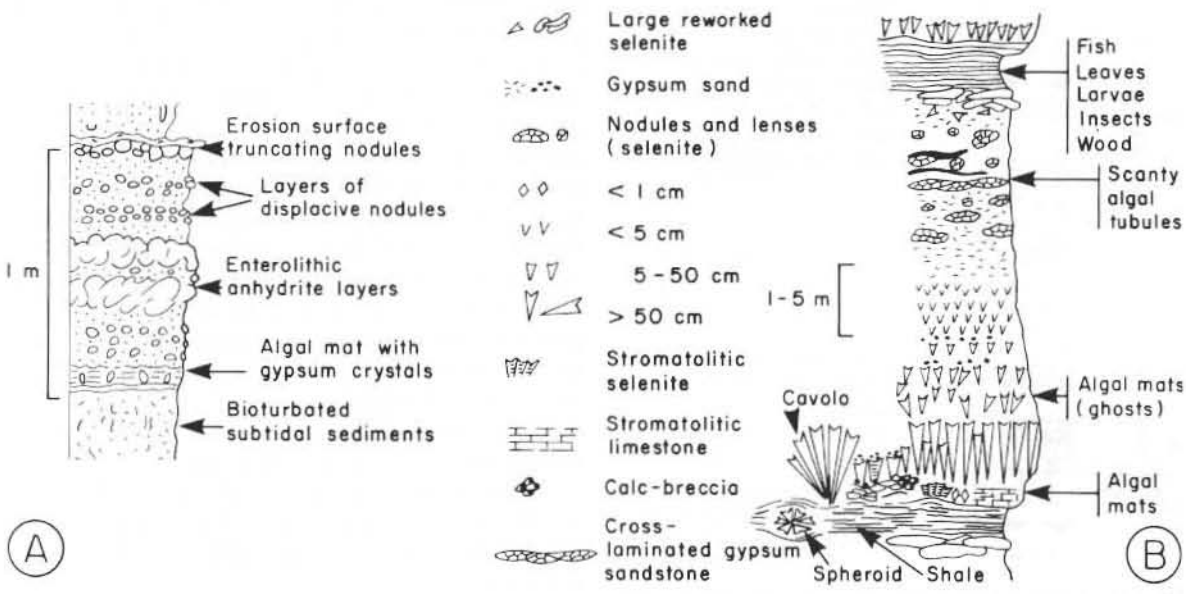


Fig. 4.44 A, B. Two typical evaporite cycles. **A** Coastal sabkha (Shearman 1966; Kendall 1984). **B** Subtidal to intertidal cycle formed by desiccation. (Vai and Ricci-Lucchi 1977)

thick and consist, in upward order, of black, pyritic mudstone, laminated dolomitic mudstone, and massive dolomitic mudstone with desiccation cracks and bioturbation. Chemical cycles that have an upper member of analcime-rich mudstone are also present. The cycles are interpreted as the product of short-term climatic change, with differences between the two types of cycle reflecting a greater tendency toward humidity or aridity, respectively. Eugster and Hardie (1975) described transgressive–regressive playa-margin cycles in the oil-shale-bearing Green River Formation of the Rocky Mountains. Donovan (1975) erected five profile models for cycles that occur as a result of changes in lake level and fluvial–deltaic lake-margin progradation in the Devonian Orcadian Basin of northern Scotland. Numerous other examples could be quoted.

Much has been written about the recognition of cyclic sedimentation from petrophysical logs (Fisher et al. 1969; Skipper 1976; Pirson 1977; Cant 1984, 1992). At present, the technique is best suited to the study of clastic cycles. As explained in Section 2.4.5, the gamma-ray, spontaneous-potential, and resistivity logs are sensitive indicators of sand–mud variations and are ideally suited to the identification of fining- and coarsening-upward cycles. These appear as bell-shaped and funnel-shaped log curves, respectively, and various subtleties of environmental change may be detected by observing the convexity or concavity and smoothness versus serration of the curves, the presence of nested cycles of different thicknesses, and so on. Fisher et al. (1969) published a series of typical profiles for coastal and marginal-marine clastic environments based on examples from the Gulf Coast (Fig. 4.45). Other examples from the Beaufort–Mackenzie Basin (Young et al. 1976) are illustrated in Fig. 4.46. These curves are commonly interpreted in the absence of cores or cuttings. As should be apparent from the preceding pages, similar cycles may be produced in different environments, so this is a risky procedure. However, by paying close attention to appropriate facies models and scale considerations (cycle thickness, well spacing), good paleogeographic reconstructions can be attempted. The availability of cores in a few crucial holes may make all the difference. Figure 4.47 illustrates a fluvial fining-upward cycle with the typical bell-shaped log profile. The log has been tied to a core that confirms the expected vertical changes in grain size and sedimentary structures.

4.5.9 Architectural Elements and Bounding Surfaces

One of the fundamental objectives of stratigraphy is to describe and interpret the 3-D architecture of the sediments filling sedimentary basins. On the largest scale, this involves the disciplines of lithostratigraphy, biostratigraphy, allostratigraphy and chronostratigraphy (Chap. 3), the reconstruction of major depositional sequences, perhaps with the aid of regional seismic sections (Chaps. 5, 6), an assessment of the roles of sea-level change (Chap. 8), and the various mechanisms of basin subsidence (Chap. 7). At an intermediate to small scale, basin architecture depends on the interplay among subsidence rates, sea level, and the autocyclic processes governing the distribution and accumulation of sediments within a particular range of depositional systems. Exploration for stratigraphic petroleum traps and for many types of strata-bound ore bodies requires close attention to basin architecture at the intermediate scale. For example, the search for and interpretation of subsurface trends in potential reservoir rocks is a major preoccupation of petroleum geologists. Since the late 1950s, many improvements that facilitate this task have been made in techniques of basin analysis.

For some time, we have been quite good at predicting the locations and trends of such major depositional entities as alluvial fans, barrier reefs, various types of delta, and barrier-island sand bodies (to take some examples at random). However, one of the outcomes of modern sedimentological research on depositional processes has been a much-improved understanding of the composition and architecture of the various facies assemblages formed within each type of depositional system. We now have the ability, therefore, to construct much better models of the heterogeneities that exist within clastic reservoirs. Conventional techniques of sedimentological analysis and facies modeling place primary emphasis on the use of Walther's Law and the interpretation of vertical profiles, particularly for the study of clastic sediments (Walker and James 1992; Reading 1996). While these have proved to be powerful techniques and have led to much-improved stratigraphic and paleogeographic syntheses, they have not provided an adequate basis for the quantitative modeling of depositional systems. In addition, the data have not found much application in the one area where detailed sedimentological information might be expected to have made its most useful contribution: the documentation of heterogeneities in petroleum reservoirs for the purpose of improved petroleum production, particularly during the design of enhanced-recovery programs.

The vast recent literature based on current facies-model techniques and the appearance of several definitive textbooks on the subject might have created the impression that facies modeling is virtually a completed task, that we have at last reached the "mopping up" phase, which was so prematurely described in the

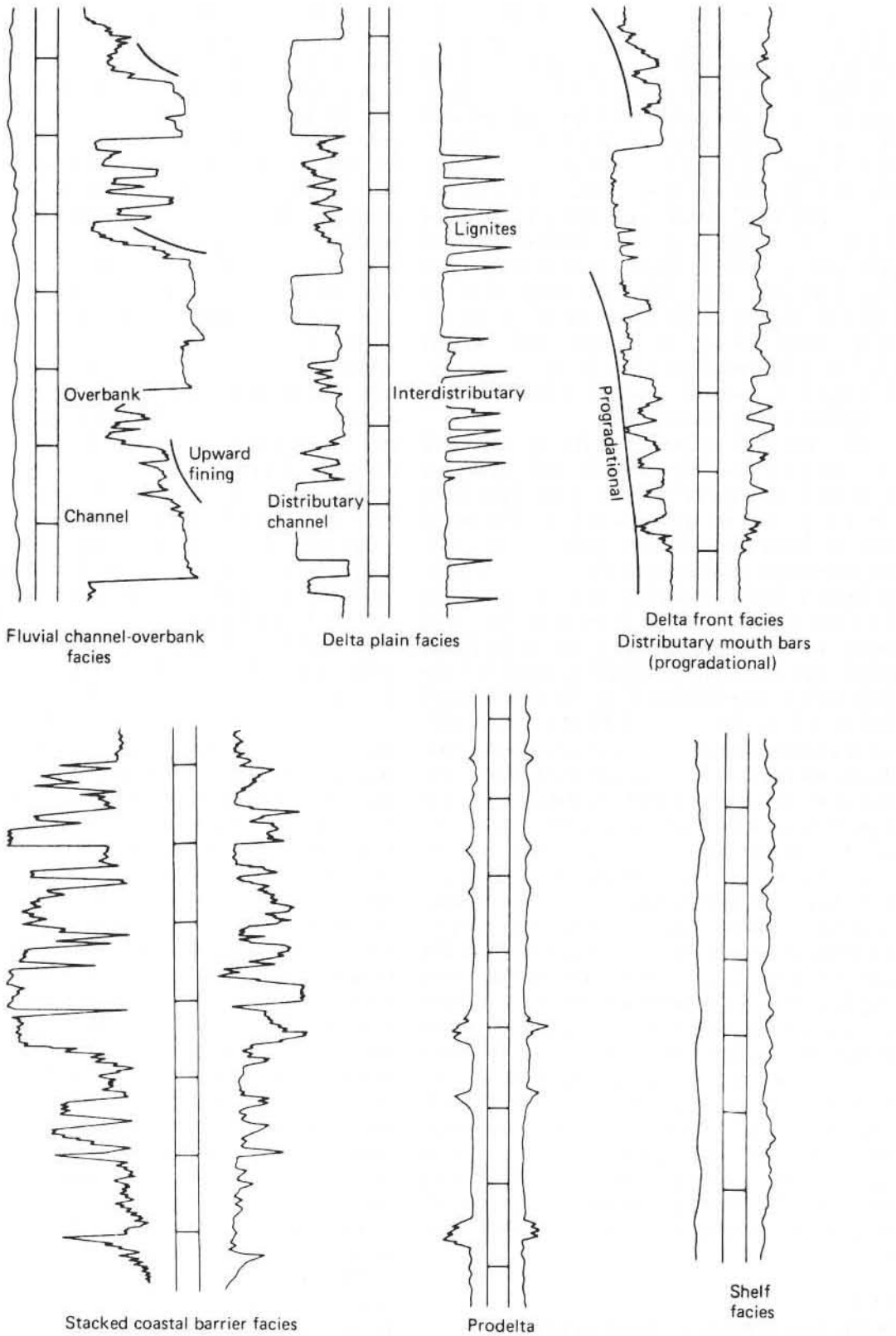


Fig. 4.45. Examples of typical petrophysical log profiles through coastal-plain and shelf clastic sequences based on examples from the Gulf Coast. *Left* Log: gamma ray or spontaneous potential (S.P.); *right* log: resistivity. *Center bar* shows scale subdivisions of 30 m (100 ft). (Fisher et al. 1969)

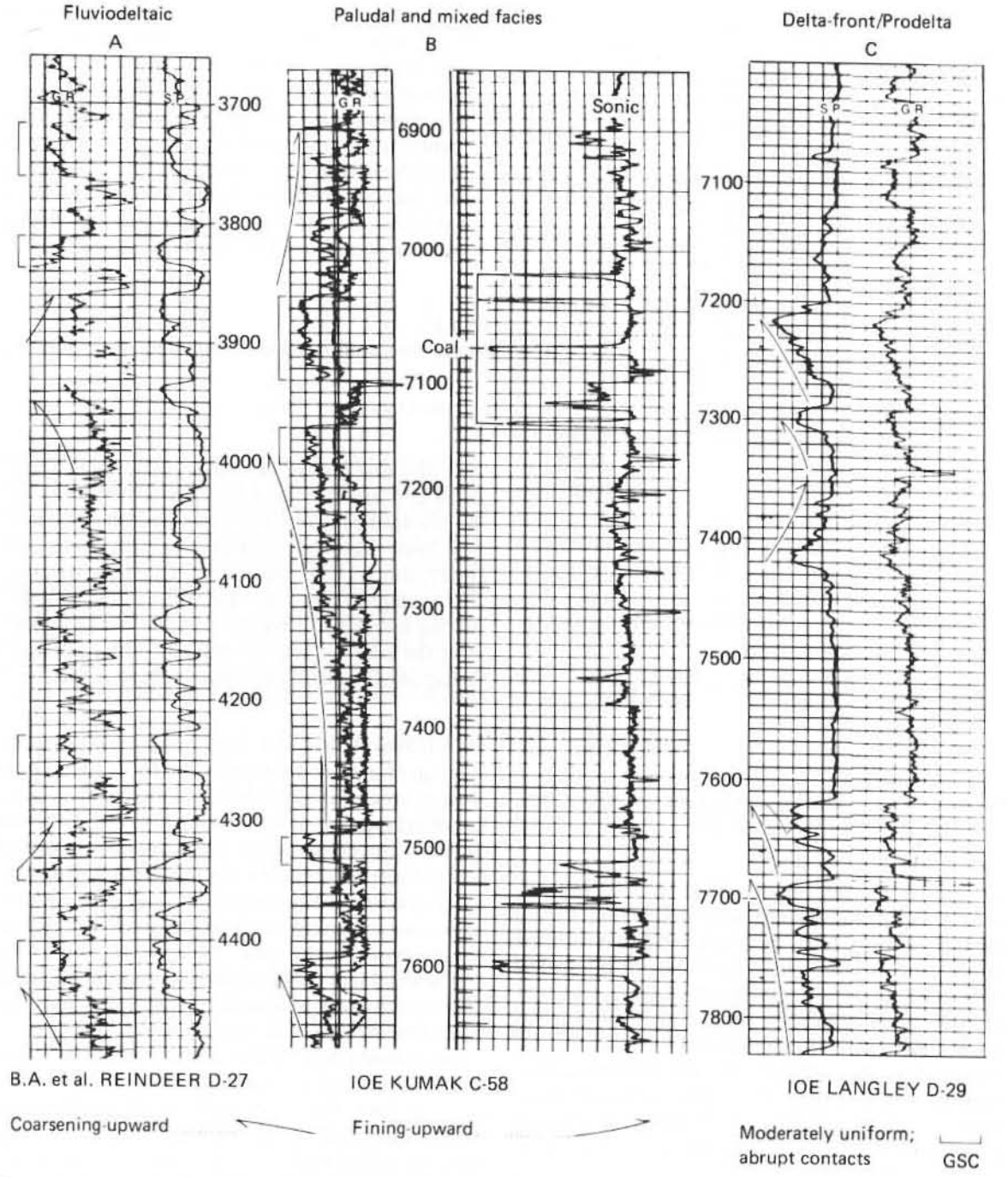


Fig. 4.46. Some examples of actual log profiles from the Beaufort–Mackenzie Basin. (Young et al. 1976)

case of turbidites by Walker (1973). However, the lack of success in the two areas noted above suggests that this is not the case. The key to further progress will be an evolution in our methods from the essentially one-dimensional approach of the vertical profile to a fully quantitative 3-D system of facies description. In this section, we examine some possible approaches to this objective.

Quantitative modeling of clastic depositional systems is facilitated by a better understanding of the types of depositional units or elements that comprise stratigraphic sequences, including data describing, in three dimensions, their scales, lithofacies compositions, length of time required for accumulation, preservation potential, and spacing in time and space. A more immediate practical application of these data is

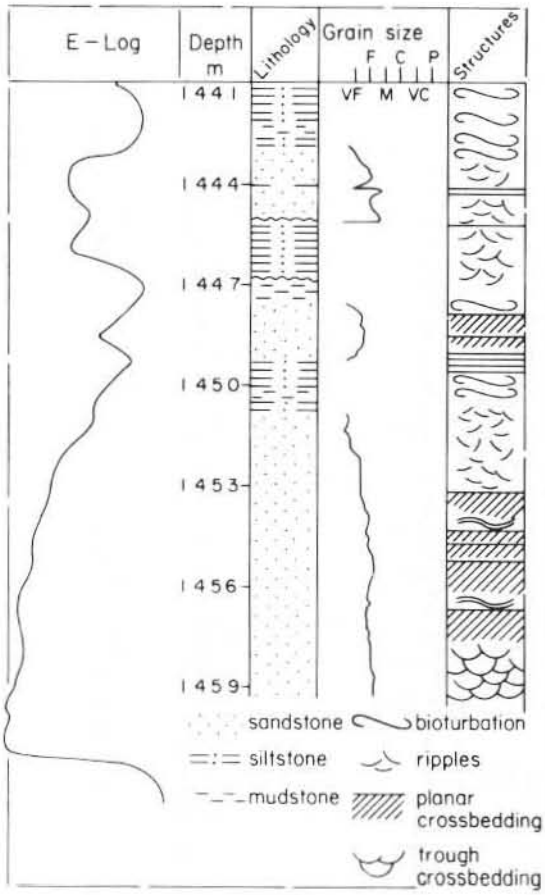


Fig. 4.47. Core and log character of a fluvial fining-upward cycle. (Lerand and Oliver, in Shawa 1974)

to improve the information used by reservoir engineers and geologists in their models of fluid flow in clastic reservoirs. It has been estimated that complex internal architectures are responsible for intrareservoir, stratigraphic entrapment of an average of 30% of the original oil in place, amounting to as much as 100 billion barrels of movable, nonresidual oil in the United States alone (Tyler et al. 1984). A better understanding of the architectural complexities of petroleum reservoirs would facilitate improved primary production and would increase the success rate of enhanced-recovery projects.

There are two important, inter-related ideas (Miall 1988a, b, c).

1. The first is the concept of **architectural scale**. Deposits consist of assemblages of lithofacies and structures over a wide range of physical scales, from the individual small-scale ripple mark to the assemblage produced by an entire depositional system. Recent work, particularly in eolian, fluvial, tidal, and turbidite environments, suggests that it is possible to formalize a hierarchy of scales. Depositional units at each size scale originate in response to processes occurring over a particular time scale and are physically separable from each other by a hierarchy of internal **bounding surfaces**.
2. The second is the concept of the **architectural element**. An architectural element is a lithosome characterized by its geometry, facies composition, and scale, and it is the depositional product of a particular process or suite of processes occurring within a depositional system.

4.5.9.1 *Architectural Scale and Bounding-Surface Hierarchies*

As noted by Allen (1983):

> The idea that sandstone bodies are divisible internally into "packets" of genetically related strata by an hierarchically ordered set of bedding contacts has been exploited sedimentologically for many years, although not always in an explicit manner. For example, McKee and Weir (1953) distinguished the hierarchy of the stratum, the set of strata, and the coset of sets of strata, bedding contacts being used implicitly to separate these entities.

Allen (1966) showed that flow fields in such environments as rivers and deltas could be classified in a hierarchical order. His hierarchy was designed as an aid to the interpretation of variance in paleocurrent data collected over various areal scales, from the individual bed to large outcrops or outcrop groups. The hierarchy consists of five categories, small-scale ripples, large-scale ripples, dunes, channels, and the integrated system, meaning the sum of the variances over the four scales. Miall (1974) added the scale of the entire river system to this idea and compiled some data illustrating the validity of the concept (Sect. 5.6.4).

Brookfield (1977) discussed the concept of an eolian bedform hierarchy and tabulated the characteristics of four orders of eolian bedform elements: draas, dunes, aerodynamic ripples, and impact ripples. These four orders occur simultaneously, superimposed on each other. Brookfield showed that this superimposition resulted in the formation of three types of internal bounding surface. His first-order surfaces are major, laterally extensive, flat-lying or convex-up bedding planes between draas (**macroforms**, in the terminology of Jackson 1975). Second-order surfaces are low- to moderately-dipping surfaces bounding sets of cross-strata formed by the passage of dunes across draas (**mesoforms**). Third-order surfaces are reactivation surfaces bounding bundles of laminae within crossbed sets and are caus-

ed by localized changes in wind direction or velocity (mesoforms to **microforms**).

Similar hierarchies of internal bounding surfaces have been recognized in some subaqueous bedforms. Reactivation surfaces (Figs. 2.10, 4.23; Collinson 1970) may develop as a result of changes in stage or flow direction or under conditions of constant stage, where bedforms randomly interact (McCabe and Jones 1977). They have been documented in large-scale, fluvial crossbed sets (Jones and McCabe 1980) and in tidal sand waves (Allen 1980). In the latter case, the sand waves rest on horizontal surfaces analogous to Brookfield's first-order surfaces and contain reactivation surfaces of second order that relate to individual tidal cycles. Third-order surfaces bound individual crossbed sets (Fig. 4.25). Shurr (1984) developed a fivefold hierarchy of morphological elements of shelf sandstone bodies, ranging from the lithosome (widespread stratigraphic units) to individual facies packages. Yang and Nio (1989) carried out a similar type of study on some ancient tidal-sandstone bodies.

Brookfield's (1977) development of the relationships among the time duration of a depositional event, the physical scale of the depositional product, and the geometry of the resulting lithosome was a major step forward that has been of considerable use in the analysis of eolian deposits. Brookfield (1977), Gradzinski et al. (1979), and Kocurek (1981) showed how these ideas could be applied to the interpretation of ancient eolian deposits. Kocurek (1988a) has found that first-order surfaces include two types of surfaces, the most laterally extensive of which he terms **super-surfaces**. Characterization of eolian deposits, therefore, now requires a fourfold hierarchy of bounding surfaces.

Several workers have attempted to develop a breakdown of the range of physical scales present in fluvial deposits. Allen's (1983) study of the Devonian Brownstones of the Welsh Borders represents the first explicit attempt to formalize the concept of a hierarchy of bounding surfaces in fluvial deposits and makes reference to Brookfield's work in eolian strata as a point of comparison. Allen described three types of bounding surfaces. He reversed the order of numbering used by Brookfield (1977), such that the surfaces with the highest number are the most laterally extensive. No reason was offered for this reversal, but the result is an open-ended numbering scheme that can readily accommodate developments in our understanding of larger-scale depositional units, as discussed below. First-order contacts, in Allen's scheme, are set boundaries, in the sense of McKee and Weir (1953). Second-order contacts "bound clusters of sedimentation units of the kinds delineated by first-order contacts." They are comparable to the coset boundaries of McKee and Weir (1953), except that more than one type of lithofacies may comprise a cluster. Allen (1983) stated that "these groupings, here termed complexes, comprise sedimentation units that are genetically related by facies and/or paleocurrent direction." Many of the complexes in the Brownstones are macroforms, in the sense defined by Jackson (1975). Third-order surfaces are comparable to the major surfaces of Bridge and Diemer (1983). No direct relationship is implied between Allen's three orders of surfaces and those of Brookfield because of the different hydraulic behavior and depositional patterns of eolian and aqueous currents.

Miall (1988a) found it useful to expand Allen's classification to a sixfold hierarchy to facilitate the definition of fluvial macroform architecture and to include the largest (basin-scale) heterogeneities in the classification (Fig. 4.48). This hierarchy is summarized in Table 4.5 and is compared with other examples of architectural hierarchies erected for eolian deposits (Brookfield 1977; Kocurek 1988a), coastal–estuarine sand waves (Allen 1980), HCS cycles (Dott and Bourgeois 1982), shelf deposits (Shurr 1984), and turbidite depositional systems (Mutti and Normark 1987). The various rankings are correlated mainly on the basis of the time duration represented by the deposits or the length of time represented by the bounding surfaces between them. An example of an outcrop lateral profile showing the application of the sixfold bounding surface classification is shown in Fig. 2.4. Note that the depositional units listed in Table 4.5 represent a range of 12 orders of magnitude in time scale. They also represent at least 14 orders of magnitude in size (area). Note also that, although this hierarchy has, at the time of writing, only been applied to fluvial deposits (Miall and Tyler 1991), there is considerable potential for similar subdivision of all clastic deposits, including an adaptation of the other hierarchical systems illustrated in the table. Miall (1988b) noted the similarities of fluvial deposits to some coarse-grained submarine-fan deposits, particularly the channel fills, and suggested ways of applying the architectural concepts described here to these rocks. These ideas were picked up by Clark and Pickering (1996), who provided numerous illustrations of modern and ancient deep-marine channel systems. Deposits of ranks 6 to 8 are discussed further in Chapter 6, where the concepts of sequence stratigraphy are considered. Causes of cyclicity in the stratigraphic record (first- to fifth-order cycles) are discussed in Chapter 8. A critique of this hierarchy and some alternative proposals were presented by Bridge (1993), and these ideas are, in turn, discussed by Miall (1995, 1996; Chap. 4). Yang and Nio (1989) and Nio and Yang (1991) presented a modified version of this system for application to tidal-sandstone bodies.

First- and second-order surfaces in the sixfold fluvial hierarchy record boundaries within microform

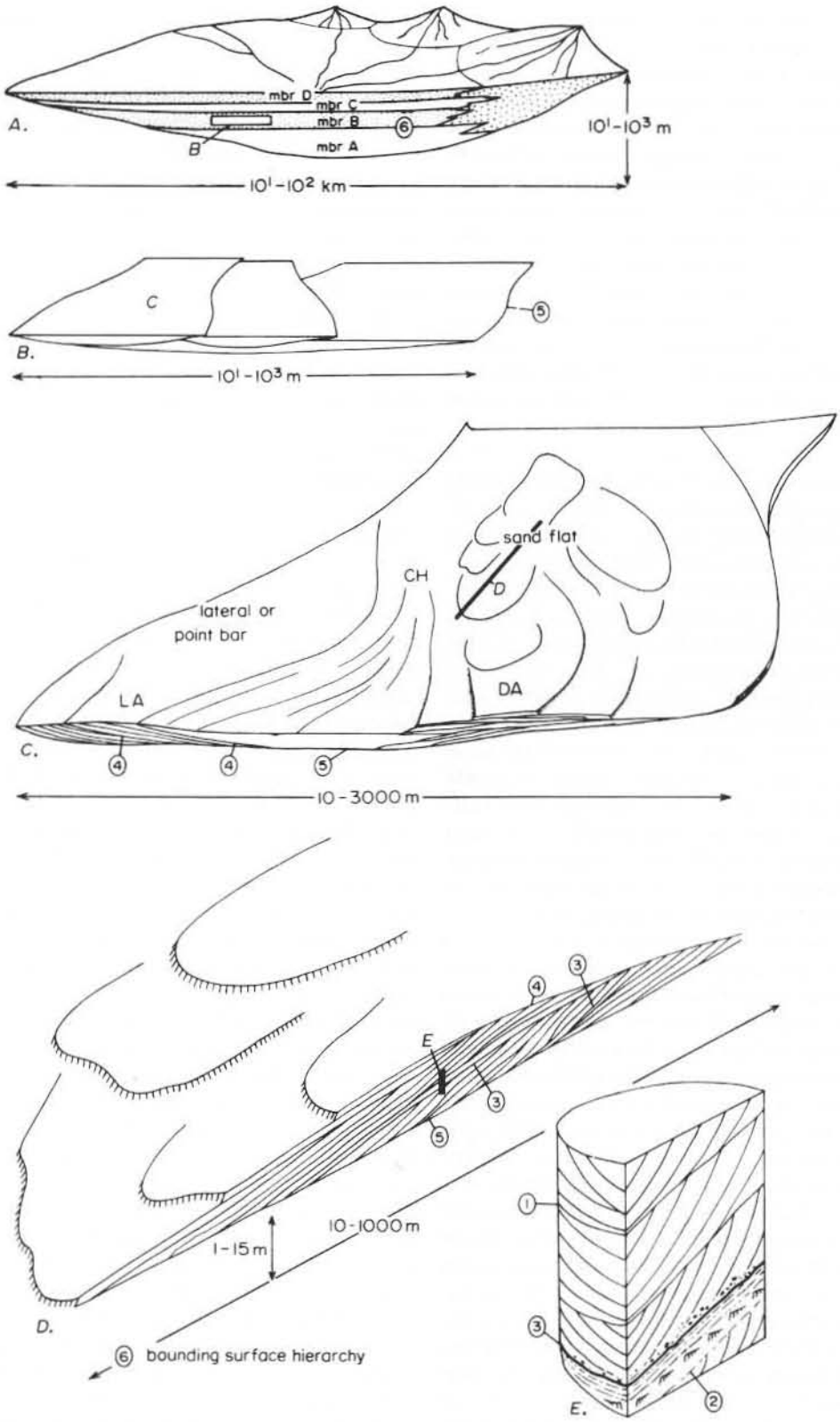


Fig. 4.48 A–E. A sixfold hierarchy of depositional elements and bounding surfaces for fluvial deposits. Diagrams **A** to **E** represent successive enlargements of part of a fluvial unit in which ranks of bounding surfaces are indicated by *circled numbers*. In **C**, *two-letter codes* are for architectural element types (Fig. 4.49). A similar hierarchy could probably be developed for channelized deposits in other environments. (Miall 1988a)

Table 4.5. Hierarchies of architectural units in clastic deposits[a]

Group	Time scale (years)	Examples of processes	Fluvial, deltaic (Miall, Kocurek)	Eolian (Brookfield)	Coastal, estuarine (Allen)	Shelf (Dott and Bourgeois, Shurr)	Submarine fan (Mutti and Norinark)
1A	10^{-5}–10^{-4}		Ripple (microform)		Ripple [E3 surface]	[3 Surfaces in HCS]	
1B	10^{-3}	Diurnal tidal cycle	Diurnal dune incr., react. surf.		Tidal bundle [E2 surface]	[2 Surfaces in HCS]	
2	10^{-2}–10^{-1}	Neap–spring tidal cycle	Dune (mesoform)		Neap–spring bundle [E2], storm layer	HCS sequence [1 surface]	
3	10^{0}–10^{1}	Seasonal to 10-year flood	Macroform growth increment	Reactivation [3rd order] surface	Sand wave, [E1], major storm layer	HCS sequence [1 surface]	
4	10^{2}–10^{3}	100-Year flood	Macroform, e.g. point bar, levee, splay	Dune [2nd-order surface]	Sand-wave field, washover fan	[Facies package (V)]	Macroform [5]
5	10^{3}–10^{4}	Long-term geomorphic processes	Channel, delta lobe	Draa [1st-order surface]	Sand ridge, barrier island, tidal channel	[Elongate lens (IV)]	Minor lobe, Channel levee [4]
6	10^{4}–10^{5}	5th-order (Milanko-vitch) cycles	Channel belt	Erg [supersurface]	Sand-ridge field c-u cycle	[Regional lentil (III)]	Major lobe [turbidite stage: 3]
7	10^{5}–10^{6}	4th-order (Milanko-vitch) cycles	Depo. system, alluvial fan, major delta	Erg [supersurface]	C-u cycle	[Sandstone sheet (II)]	Depo. System [2]
8	10^{6}–10^{7}	3rd-order cycles	Basin-fill complex		Coastal plain complex	[Lithosome (I)]	Fan complex [1]

[a] Hierarchical subdivisions of other authors are given in square brackets. Names of authors are at the head of each column. c-u, coarsening upward; depo., depositional; HCS, hummocky cross-stratification; incr., increment; react. surf., reactivation surface

and mesoform deposits (Fig. 4.48). The definition of *first-order surfaces* is unchanged from Allen (1983). *Second-order surfaces* are simple coset bounding surfaces, in the sense of McKee and Weir (1953). Third- and fourth-order surfaces are defined when architectural reconstruction indicates the presence of macroforms, such as point bars or sand flats (Fig. 4.48). *Third-order surfaces* represent growth increments of macroforms, such as the reactivation surfaces of Jones and McCabe (1980). Individual depositional units (storeys or architectural elements) are bounded by surfaces of *fourth-order* or higher rank. *Fifth-order surfaces* are surfaces bounding major sand sheets, such as channel-fill complexes (Fig. 4.48). They are generally flat to slightly concave upward but may be marked by local cut-and-fill relief and by basal lag gravels (these were termed third-order surfaces by Allen 1983 and are the major surfaces of Bridge and Diemer 1983). *Sixth-order surfaces* are surfaces defining groups of channels or paleovalleys. Mappable stratigraphic units, such as members or submembers, are bounded by sixth-order surfaces (Fig. 4.48; not defined by Allen 1983). Miall (1997, Table 15.3) suggested that sixth-order surfaces, as defined here, are equivalent to the "F" surfaces of Nio and Yang (1991) and to the bounding unconformities of allomembers or submembers, in the sense used by some workers in the area of sequence stratigraphy (Chap. 6). *Seventh-order surfaces* define individual depositional systems. *Eighth-order surfaces* bound major basin-fill complexes. Stratigraphic sequences are bounded by surfaces of sixth- and higher-order rank.

Surfaces of fifth and higher order are potentially the easiest to map in the subsurface because of their wide lateral extent and essentially simple, flat, or gently curved, channelized geometry. Many examples of the mapping of channel and paleovalley surfaces have, in fact, been reported in the literature (Busch 1974). Doyle and Sweet (1995) reported on an attempt to apply Miall's hierarchical system of bounding surfaces to a subsurface study of a fluvial sandstone succession. Considerable potential now exists for mapping these surfaces with 3-D seismic data, as described by Brown (1986). All higher-order surfaces may appear very similar in cores. They are best differentiated by careful stratigraphic correlation between closely spaced cores, an objective that is best achieved in the most intensively developed fields, where well spacing may be a few hundred meters or less. Third- and fourth-order surfaces may be recognized by gentle depositional dips (usually $< 10°$).

Identification and correlation of these various bounding surfaces can clearly make a major contribution to the unraveling of the complexities of a channelized depositional system. They are likely to be particularly useful in the recognition and documentation of macroforms, about which much remains to be learned. Further discussion and illustration of the fluvial bounding-surface hierarchy, as applied to outcrop examples of fluvial deposits, is given by Miall (1996).

The basis for a universal scheme with which to subdivide complex, 3-D rock bodies is beginning to emerge, as suggested by Table 4.5, and the development of a common terminology may become necessary. Thus, the eolian super-surfaces of Kocurek (1988b) are equivalent to Miall's (1988a) sixth-order surfaces, whereas Brookfield's (1977) first-order eolian surfaces could, perhaps, be equated with fifth-order surfaces. Similarly, in the shelf environment, tidal sand waves and sand banks also contain a hierarchy of facies units and internal bounding surfaces (Allen 1980; Shurr 1984; Berné et al. 1988; Harris 1988). The basal E1 surfaces of tidal sand waves (Allen 1980) are equivalent to Miall's third-order surfaces in terms of their time significance, and E2 surfaces are of first or second order, depending on their origin as diurnal or spring–neap erosion surfaces. The basal erosion surfaces of sand banks (Harris 1988) are of fifth order, whereas the internal master bedding planes are equivalent to fourth-order surfaces. Note that these rankings are based on time significance, not on geometry. E2 surfaces of sand waves and master bedding surfaces of sand banks both show depositional dips and appear superficially similar to Miall's third-order surfaces (compare Fig. 4.48 with Figs. 4.25, 4.61, and 4.62) but represent different processes acting over different time periods. These differences make the erection of a common terminology difficult. The rankings of Mutti and Normark (1987) for turbidite systems are also shown in Table 4.5.

4.5.9.2 *Architectural Elements*

Application of the bounding-surface concept permits the subdivision of a clastic succession into a hierarchy of 3-D rock units. This facilitates description and also makes it easier to visualize the appropriate physical extent and time duration of the processes that controlled sedimentation at each level of the hierarchy. Rock description involves the definition of lithofacies and the recognition of facies assemblages.

Most deposits may be subdivided into several or many types of 3-D bodies characterized by distinctive lithofacies assemblages, external geometries, and orientations (many of which are macroforms). Allen (1983) coined the term **architectural element** for these depositional units, and Miall (1985) attempted a summary and classification of the current state of knowledge of these elements as they occur in fluvial

deposits, suggesting that there are about eight basic architectural elements in fluvial depositional systems (Fig. 4.49). Figure 4.50 illustrates outcrop examples of some characteristic elements in a Jurassic fluvial deposit in New Mexico. Two interpretive processes are involved simultaneously in the analysis of outcrops that contain a range of scales of depositional units and bounding surfaces: (1) the definition of the various types and scales of bounding surfaces and (2) the subdivision of the succession into its constituent lithofacies assemblages, with the recognition and definition of macroforms and any other large features that may be present.

In general, the most distinctive characteristic of a macroform is that it consists of genetically related lithofacies, with sedimentary structures showing similar orientations and internal, minor bounding surfaces (first- to third-order of the classification given previously) that extend from the top to the bottom of the element, indicating that it developed by long-term lateral, oblique, or downstream accretion. A macroform is comparable in height to the depth of the channel in which it formed and, in width and length, is of a similar order of magnitude to the width of the channel. However, independent confirmation of these dimensions is difficult in multistory sandstone bodies, where channel margins are rarely preserved and the storeys commonly have erosional relationships with each other.

The definition of a macroform in any given outcrop is, in part, an interpretive process. Some types of macroforms, such as the lateral accretion deposits that comprise the typical point bar, are by now so well known that their recognition in outcrops would be classified by some workers as a descriptive rather than an interpretive exercise. Other types of macroform are less well known. The recognition of a macroform may depend, in part, on the type of bounding

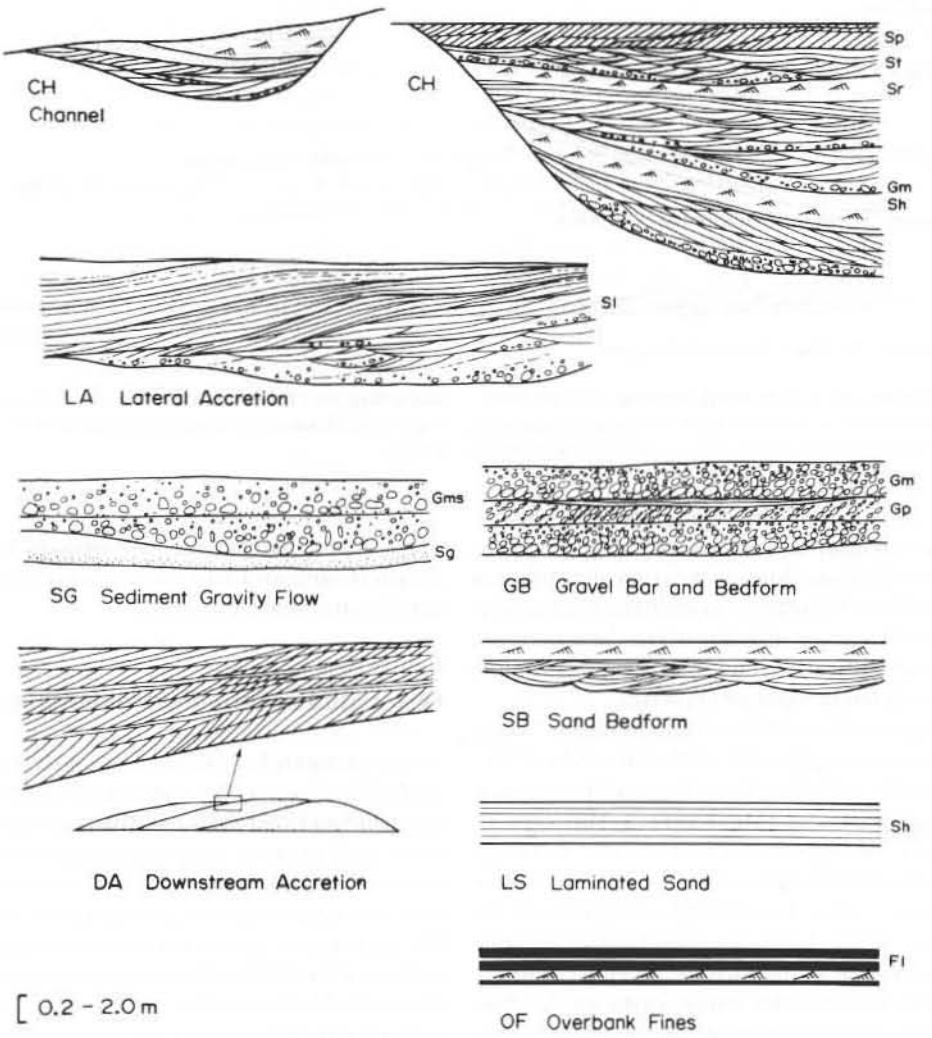


Fig. 4.49. Principal architectural elements in fluvial deposits. (Modified from Miall 1985)

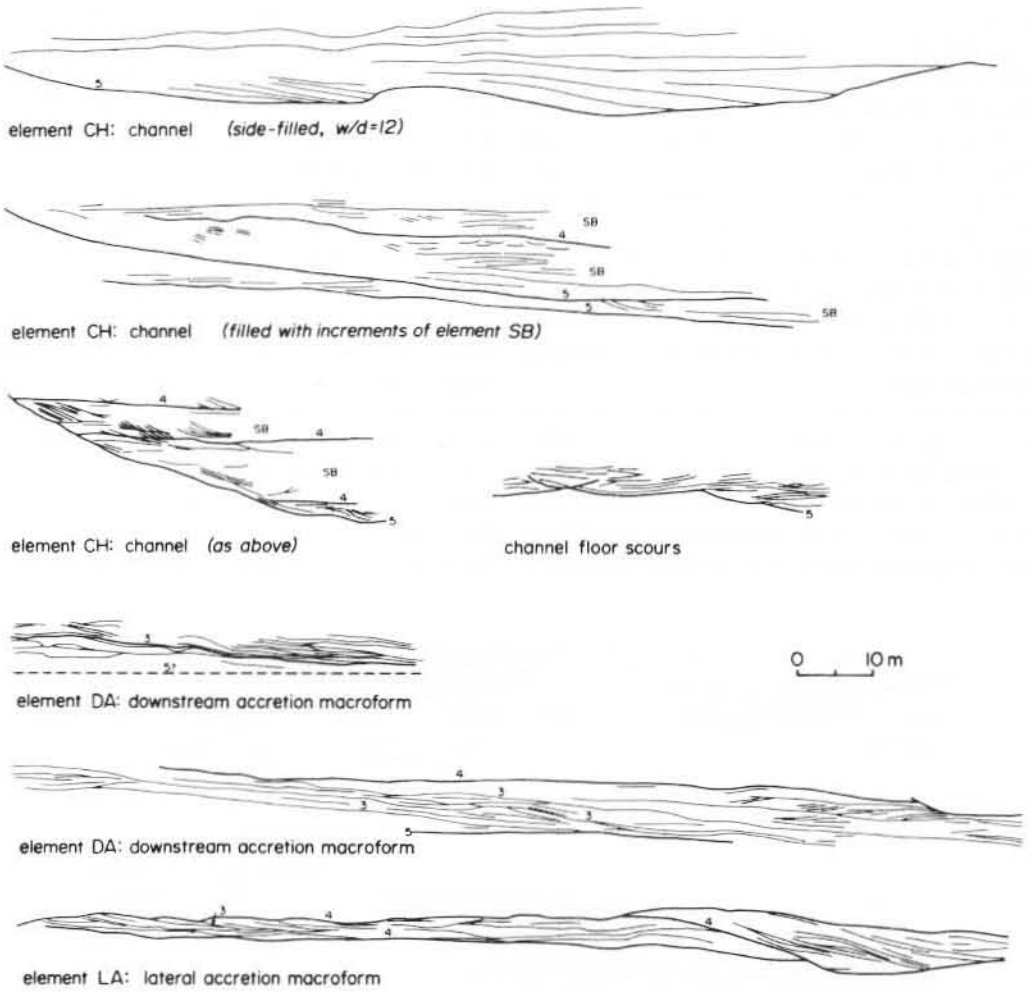


Fig. 4.50. Examples of architectural elements in the Westwater Canyon Member of the Morrison Formation (Jurassic), San Juan Basin, New Mexico. Bounding surfaces are numbered according to rank. Element types are those indicated in Fig. 4.49, lithofacies codes are those listed in Table 4.1. (Miall 1988c)

surface that encloses it. Conversely, the appropriate classification of a bounding surface may depend on a description of the lithofacies assemblage and geometry of the beds above and below it. For these reasons, description, classification, and interpretation cannot always be completely separate exercises.

It may be possible to document the range in scales and the potential mappability of the hierarchy of depositional units and their bounding surfaces in any given stratigraphic unit (Miall 1988c). This type of subdivision is of importance in reservoir development. However, hierarchies of this type do not have a fixed range of scales. For example, in parts of the Donjek River, Yukon, large compound bars (the first-order bars of Williams and Rust 1969, using an earlier classification; fourth-order macroforms in the present scheme) are in the order of 200–400 m in downstream length. Comparable features in the Brahmaputra River, Bangladesh, are several kilometers in length (Coleman 1969) – an order of magnitude larger (Bristow and Best 1993).

4.6 Facies Models for a New Millenium

Much stratigraphic work is now carried out under the umbrella of sequence stratigraphy, within which the mapping and correlation of the major bounding surfaces and systems tracts constitutes the major research activity (Chap. 6). The definition of sequences and the assignment of stratigraphic successions to the appropriate systems tracts is very much dependent upon the ability of the geologist to perform a sophisticated facies analysis and to recognize the depositional environments of the sediments under study. The collection of observations of facies criteria, such

as those described in the previous sections, and the assembly of this evidence into a complete, local environmental interpretation – the local "summary of the environment," to use Walker's (1992) terminology – relies, in part, on the ability of the geologist to make comparisons with various modern and ancient analogues. The availability of suites of standard facies models facilitates this task. Such facies models are also invaluable teaching aids.

The description of facies models has occupied many textbook writers in recent years. Some of this work is referenced in the following sections. In general, however, two major texts stand out above all others. That edited by Reading (1996) is by far the most complete and authoritative. It is now in its third edition. This book is aimed at a graduate-student and professional audience. At an advanced undergraduate level is the work by Walker and James (1992). The current edition of this book– its third since the first edition appeared in 1979 – is a complete recasting of earlier versions, written to take into account and to emphasize the recent focus on sea-level change as one of the primary controls on environmental change and facies architecture. Other recent books, for example those by Friedman et al. (1992), Galloway and Hobday (1996), and Prothero and Schwab (1996), also provide excellent coverage of this subject area. It would be superfluous to attempt to improve on this wealth of source material, nor is there space for it here. However, it would seem useful to review some of the current developments as a guide to research in the area for advanced students.

There has been an evolution in thinking regarding the relationships between processes in depositional environments, resulting in some changes in the way facies models are described and classified in these books. In the case of coastal environments, where fluvial, wave, and tidal processes interact, earlier editions of the Walker and James book described deltas of all types in a single chapter, whereas, in the most recent edition (Walker and James 1992), coastal deposits formed under tidal influence are not considered to be "deltaic" under normal circumstances, falling more typically into an estuarine classification. A different approach is taken in the Reading book. Deltas and other types of coastline (e.g., barrier systems) were described in separate chapters in early editions of the book but, in the most recent edition (Reading and Collinson 1996), this separation is considered to be arbitrary, and all types of clastic coastline are described under a single heading. Apart from this, few significant changes have occurred in the field of facies studies since the late 1980s. Most of the major developments in our understanding of process-response models had been achieved by that time, and progress during the 1990s has been primarily incremental, resulting in refinements and minor improvements to the existing body of knowledge, brought about primarily by the accumulation of case studies.

4.6.1 Fluvial Environments

Models of fluvial sedimentation represent an attempt to combine geomorphological observations of modern rivers with interpretations of ancient successions. Since the first International Symposium on Fluvial Sedimentology in 1977, there have been about a dozen major symposia devoted partly or entirely to fluvial deposits. Developments that were presented at these meetings, and others described in various review papers and short courses, have been summarized in the major review chapter by Collinson (1996) and the book by Miall (1996). There have been four major strands to fluvial research during the last decade: a move away from the proliferation of facies models erected mainly on the supposed distinctiveness of vertical profiles toward models based on 3-D facies characteristics (architecture); an increased interest in the control tectonics has over fluvial architecture; a new emphasis on the information to be obtained from the fine-grained facies and paleosols deposited in overbank areas; and a considerable increase in interest in the internal architecture of fluvial sandbodies and its relationship to porosity and permeability patterns – a topic of considerable concern in the fields of petroleum-production geology and hydrogeology (water supply, toxic-waste transmission).

Given the accessibility of the modern fluvial environment for geological and geomorphological study, it is probably not surprising that a wealth of descriptive material is now available for application to sedimentological studies. However, there remain significant gaps in our knowledge of modern rivers. Tropical rivers, including very large examples, such as the Amazon, the Orinoco, the Mekong, and the Congo, remain virtually unstudied from the point of view of facies models.

The facies classification and coding system introduced by Miall (1977, 1978) and updated in 1996 (Table 4.1) is now virtually in standard use for fluvial studies, although, as noted in Section 4.3.1, not all workers agree with the standardized approach to classification implicit in this system (Bridge 1993). An increasing number of workers are employing architectural methods for the 3-D documentation and classification of fluvial deposits in the surface and subsurface, particularly for channel-fill sandstone and conglomerate bodies (Miall and Tyler 1991; Miall 1996; Chap. 6). The classification of bounding surfaces proposed by Miall (1988; Sect. 4.5.9) has also

found considerable use, although it has, as expected, proved difficult to make much use of the classification in subsurface work (Doyle and Sweet 1995). A simplified classification of channel style, showing their dependency on sediment grain size and other factors and the range of typical architectural elements present in each type is shown in Fig. 4.51. However, researchers are encouraged to devise their own classifications to suit the particular needs of their research.

Detailed work on modern floodplains, such as the study of the Mississippi by Farrell (1987), and classification studies, such as that by Nanson and Croke (1992), have contributed to a much better understanding of floodplain processes and sedimentation during the last decade. Description and classification of ancient floodplain deposits in terms of facies assemblages and architectural elements were attempted by Eberth and Miall (1991) and Platt and Keller (1992), and the importance of this environment for coal development has been examined by Flores (1981). Important work on the genesis of paleosols and their stratigraphic context has been carried out by Kraus and Bown (1988) and Wright (1986 1992). From the point of view of basin analysis, probably the most interesting study of overbank deposits is that of

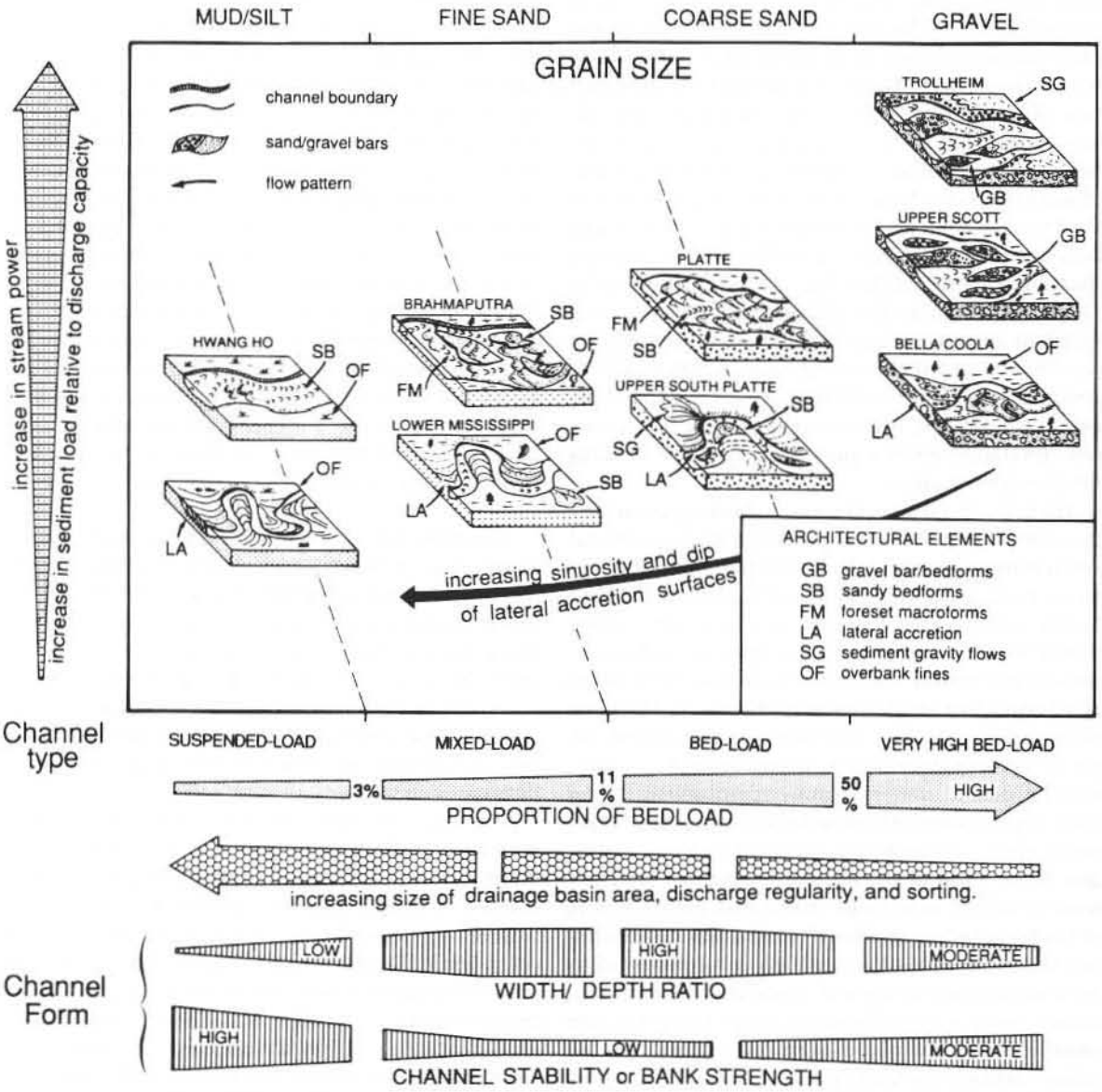


Fig. 4.51. Relationship between grain size and fluvial-channel pattern. Architectural elements are those of Miall (1985; Fig. 4.49). Note that FM (foreset macroforms) has now been replaced by DA (downstream-accretion element). (Orton and Reading 1993)

Willis and Behrensmeyer (1994), who used an extremely detailed stratigraphic study of part of the Miocene Siwalik succession in Pakistan as a basis for the building and testing of various models of the development of floodplain stratigraphies.

Rivers draining transversely into a sedimentary basin commonly deposit a cone-shaped wedge of detritus at the basin margin, with the river diverging into a series of radiating distributaries. These two features would be the key elements of most geologists' definition of the term **alluvial fan**. However, there is considerable controversy among sedimentologists as to the scope of the definition beyond this initial point. For some geologists, the term has no unique facies sense. As noted by Miall (1992), there are small, steep, gravel-dominated fans, within many of which debris-slow processes are important: there are sandy systems deposited by braided streams; there are giant modern fans, such as the Kosi, of India, which grades from boulder conglomerate near the mountains to fine sand–silt–mud 140 km downslope at its distal end; and there are fine-grained, arid systems termed **terminal fans** deposited by ephemeral flows that percolate and dissipate into the substrate at their distal fringes. Geomorphologically, these are all alluvial fans and fall under the definitions of this term provided by Bates and Jackson (1987) and Miall (1990, 1992). A recent review by Stanistreet and McCarthy (1993) followed this inclusive approach. They described a new class of large, low-energy fan based on the example of the modern Okavango fan of Botswana and erected a simple classification of alluvial fans, including all the types mentioned above. This is described below. They also discussed comparisons with submarine fans, a class of depositional system about which there has also been problems of definition.

Contrasting with this approach is that of McPherson and Blair (1993) and Blair and McPherson (1994). These authors accepted many of the standard components of the geomorphic definition of fans, noted above, including their conical, radiating form, their steep depositional slopes, and their association with high-relief source areas. They pointed out a significant contrast between fans and most other fluvial systems. The latter typically exhibit concave-up transverse profiles, the channel being the lowest point, whereas fan systems are characterized by convex-up transverse profiles, which lead to overbank flow and sheet flooding of bedload. However, these authors argue that there is a distinct break in the range of longitudinal slopes exhibited by the various types of fluvial distributary system. They recommended limiting the term alluvial fan to steep systems that have a range of slopes between 1.5° and 25° (slope: 0.026–0.466). These fans are characterized by hyperconcentrated flows, sheet floods, and sediment-gravity-flow deposits, especially debris flows. Traction-current deposits are rare to absent. According to the same authors, fluvial depositional systems having slopes in the range 0.4–1.5° (0.007–0.026) are rare, with most meandering and braided fluvial systems, including the radial, distributary systems that many authors would classify as fans, exhibiting slopes less than 0.4°. This provides a natural break, which they proposed to emphasize in their classification. They recommended excluding depositional systems on slopes of $< 1.5°$ (< 0.026) from the category of alluvial fans, mainly on the basis of a historical analysis of the evolution of the term fan. They pointed out that the original work of such researchers as Drew (1873) and Surell (1841, 1870), which helped to establish the term "fan", dealt with small, steep, coarse-grained fans (Sect. 2.2.2). Heward's (1978) important review of fan depositional cycles also focuses on coarse-grained fans, although he did not discriminate or subdivide fans on the basis of sediment-transport mechanisms. Blair and McPherson (1994) suggested that the inclusion of fan-shaped glacial-outwash dispersal systems under the heading of the term alluvial fan serves no useful purpose.

Following this approach, fine-grained, fan-shaped deposits, such as those of the Kosi River, terminal fans, and those not associated with mountainous uplifts, would not be included under the heading of alluvial fans. They would be termed distributary fluvial systems or braid deltas. While this proposal is clear and may be applied with some consistency (at least to modern environments, where slopes can be measured), it leaves out of the classification many depositional systems and deposits that have traditionally been termed fans.

Use of the restricted definition of Blair and McPherson (1994) will lead to the recognition of fans in the ancient record on the basis of their facies characteristics alone, the deposits being coarse-grained debris flow and related facies banked against an uplifted source area. Such classification already is implied by the use of such terms as **fanglomerate** for ancient conglomerates even where evidence of the geomorphology of the original depositional system is not available. Such a designation will likely be suggested without evidence of the distributary nature of the fluvial system, a feature that is difficult to prove in the rock record without detailed paleocurrent evidence. Whether the use of the Blair and McPherson (1994) definition in this way will prove to be useful and will serve to reduce controversy and confusion remains to be learned from experience (Miall 1996; Sect. 8.3).

Fluvial deposition is influenced by three major external (allogenic) forces, climate changes, tectonism, and sea-level change. The last of these is discussed in

Chapter 6. In a lengthy review, Miall (1996) demonstrated that there are very few characteristics of the clastic components of fluvial deposits that are climatically sensitive. Probably the most climatically distinctive lithofacies assemblage is that of the sandy terminal fan, a type of deposit formed in arid environments where runoff infiltrates into the depositional substrate, resulting in a downstream loss of competency and capacity. A general model for these deposits, which correspond to the **ephemeral braid delta** of Blair and McPherson (1992), was developed by Kelly and Olsen (1993). Chemical and biochemical sediments of the overbank environment are more climatically sensitive, as they include such features as paleosols, desiccation features, evaporites, or coal.

Tectonism is an all-pervading control on fluvial deposition, controlling regional depositional slope and sediment supply. As noted in the second edition of this book, the mapping of paleomagnetic-reversal isochrons can be a particularly powerful method for constructing a detailed local stratigraphy, permitting the details of facies and architecture to be related to variations in fluvial style, subsidence rate, the activity of bounding faults, and so on. Leeder et al. (1996a, b), working in the Cenozoic deposits of the Rio Grande Rift, and D.W. Burbank, in several studies of the Siwaliks of Pakistan (Burbank et al. 1992), have shown how fluvial style and paleogeography evolved in response to local tectonism.

4.6.2 Deltaic Environments

A classification of deltas that became generally accepted is the "triangular" classification of Galloway (1975), who divided deltas into wave-, tide-, and river-dominated types. Extensive, comparative studies of deltas on a worldwide basis led Coleman and Wright (1975) to recognize a wide range of variability in deltaic form and process. The classifications and processes described by Coleman and Wright (1975) and Galloway (1975) have been extensively used in studies of ancient deltas, particularly studies of coal sedimentology in river-dominated deltas (Horne et al. 1978; Flores 1981), although McCabe (1984) has questioned the validity of the deltaic interpretations of coal generation because of the abundance of clastic detritus that enters the peat swamps of many deltaic systems.

Recent work on deltaic sedimentation was summarized by Bhattacharya and Walker (1992), Orton and Reading (1993), and Reading and Collinson (1996). A collection of papers on coarse-grained deltas was edited by Colella and Prior (1990). New classifications of deltas were offered by Postma (1990) and Orton and Reading (1993). Nemec and Steel (1988) edited a collection of papers on fan deltas.

As noted by Orton and Reading (1993), in addition to recognizing the importance of the three major processes governing sediment distribution on deltas, the grain size of the sediment supply is also a critical factor in determining the configuration of the delta and the architecture of the resulting deposits. McPherson et al (1987) were among the first to attempt to incorporate grain size into delta classification, with the recognition of coarse-grained types termed **fan deltas** and finer-grained varieties termed **braid deltas** and **common deltas**. Theses ideas were developed further and generalized in the classifications by Postma (1990) and Orton and Reading (1993). Postma (1990) focused on the nature of the fluvial feeder system and the water depth in the receiving basin, which reflects the setting of the delta at the margin of a broad continental shelf or on the shelf-margin or slope. By not incorporating data dealing with wave- and tide-generated sediment redistribution, this classification departs from Galloway's emphasis on the processes of the redistribution of sediment at the delta front. The classification of Orton and Reading (1993) builds on that of Galloway to incorporate data on the grain size of the sediment load (Fig. 4.52). Orton and Reading also provide a thorough discussion of the importance of other variables, such as delta-plain slope, the control of grain size and wave height on the tendency of wave energy to be dissipated or reflected at the shoreline, and the effects of these variables on the resulting sedimentary deposit.

The variations in delta type are reflected in the vertical profile and the 3-D shape and distribution of the resulting deposits. As noted by Reading and Collinson (1996), deltaic successions, which commonly are cyclic in character (Figs. 4.39, 4.45), occur on three scales:

> Large-scale allocyclic sequences involving the entire delta system are caused by tectonic and climatic changes in the hinterland, by major diversion of rivers upstream, and by eustatic sea-level changes. Medium-scale sequences are cause by switching of delta lobes or distributaries within a stable depocentre. Small-scale sequences result from differential subsidence on the delta plain, lacustrine delta formation, crevassing of distributary channels and migration of tidal channels. (Reading and Collinson 1996).

"Small-scale sequences", in Reading and Collinson's (1996) terminology, correspond to parasequences, in the new sequence terminology (Van Wagoner et al. 1987; Posamentier et al. 1988; Van Wagoner 1995). These are defined as "relatively conformable successions of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative sur-

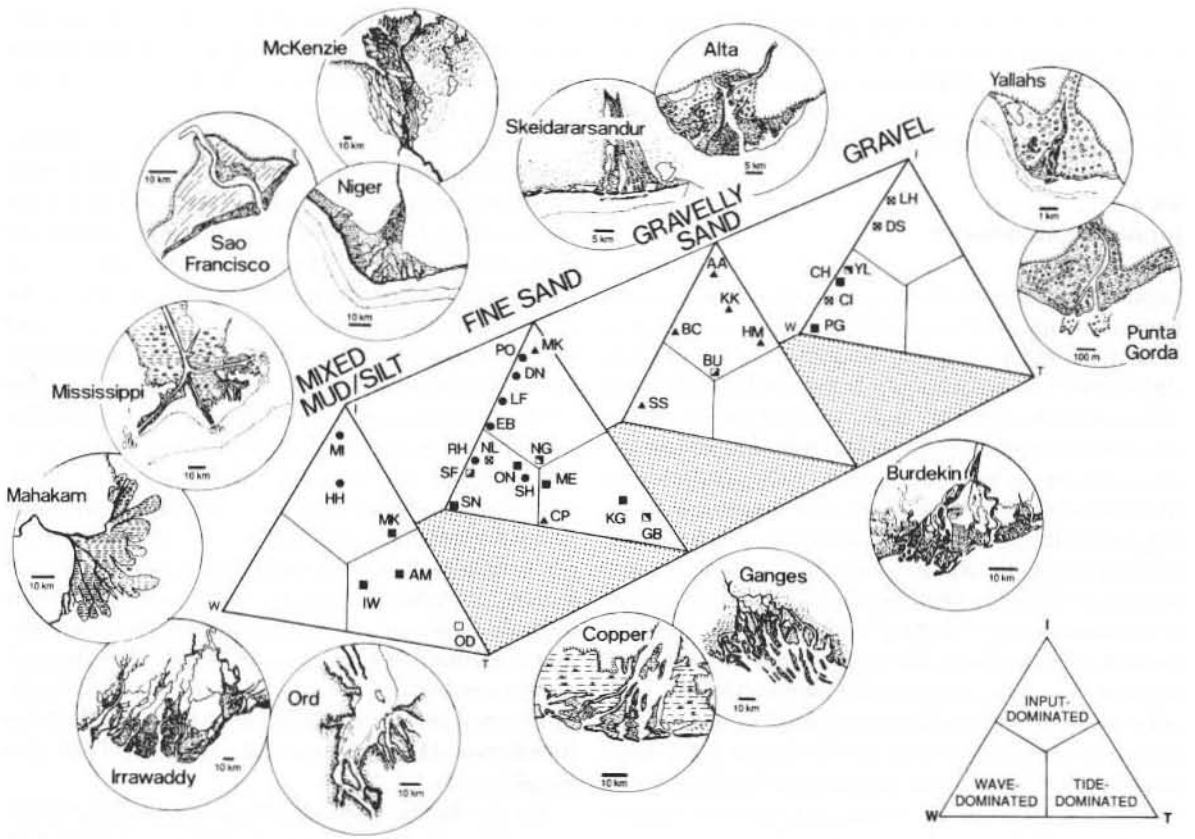


Fig. 4.52. A classification of deltas in which Galloway's (1975) "triangular" classification is amplified by the incorporation of data on the dominant grain size of the sediment delivered to the delta front. Examples of deltas are given by abbreviations as follows: *AA* Alta; *AM* Amazon; *AR* Amur; *BC* Bella Coola; *BU* Burdekin; *CH* Chao Pharya; *CL* Colorado; *CP* Copper; *DN* Danube; *DS* Dead Sea; *EB* Ebro; *GB* Ganges/Brahmaputra; *HH* Huanghe (Yellow); *HM* Homathko; *IW* Irrawaddy; *KG* Klang; *KK* Klinaklini; *LF* Lafourche; *LH* Liaohe; *ME* Mekong; *MI* Mississippi; *MK* Mackenzie; *NG* Niger; *NL* Nile; *OD* Ord; *ON* Orinoco; *PG* Punta Gorda; *PO* Po; *RH* Rhône; *SF* Sao Francisco; *SH* Shoalhaven; *YL* Yallahs. (Orton and Reading 1993)

faces" (Sect. 6.3). Examples would be the deltas lobes formed following crevassing of distributaries on river-dominated deltas such as the Mississippi (Coleman and Gagliano 1964). It is currently a problem and a source of confusion that, in the subsurface, "parasequences" formed by autocyclic processes of delta abandonment and switching cannot readily be distinguished from the more widely distributed deltaic sequences formed by relative sea-level change. Both types of succession yield similar vertical profiles. Delta-lobe progradation develops coarsening-upward successions comparable to highstand systems tracts, to use the sequence-stratigraphic terminology. When these are abandoned as a result of delta switching, continued subsidence and sediment reworking commonly leads to local transgression and the deposition of a thin "abandonment" facies consisting of shelf muds and thin, reworked sands. These compare to the transgressive systems tract of standard sequence models (Chap. 6).

Sequence concepts assist us with the interpretation of deltaic sedimentation in another way. As summarized by Bhattacharya and Walker (1992), the relative importance of delta-front depositional and reworking processes is affected by slope configuration and sea-level change. During times of high sea level, deltaic deposition typically is confined to shallow-water environments on the shelf. Such settings tend to be characterized by rapid lobe progradation and switching in a fluvially dominated environment. Deltas deposited during sea-level lowstands tend to be located at the edge of the shelf, where wave and storm influences are stronger. Growth faults are common at the shelf margin and can lead to the development of thickened sand bodies oriented parallel to the shelf margin. Deltaic sedimentation tends to be suppressed during times of rising sea level, and clastic influence is confined to aggrading fluvial floodplains. Where coastal deposition does occur, it typically is wave- and tide-influenced. Estuarine sedi-

mentation is common, filling the incised valleys that were formed by subaerial erosion during the preceding sea-level fall. Tidal sand bodies perpendicular to the shore can result, and some ancient examples have been termed tide-dominated deltas

4.6.3 Lacustrine Environments

Lakes are among the most varied of all depositional environments, even though they occupy a relatively small percentage of the earth's surface at the present day (about 1%, according to Talbot and Allen 1996). There are no universal facies models for the lacustrine environment, and recent books and review articles on this subject are not so much syntheses of ideas as catalogs of case studies, each example varying markedly from the next in terms of lithofacies, geochemistry, thickness, and extent (Matter and Tucker 1978; Ryder 1980; Picard and High 1981; Katz 1990; Anadón et al. 1991b). The documentation of this variability is, perhaps, the most important achievement of recent facies studies of lake sediments. This is by no means an academic subject. Vast petroleum reserves occur as oil shales in the Green River Formation of the Uinta Basin and as liquid petroleum in China (e.g., Daqing Oil Field), the western United States, West Africa, and elsewhere (Ryder 1980).

Most major lake deposits originated when a sedimentary basin was tectonically isolated from the sea. The advent of plate tectonics has helped explain the origins of these basins, many of which are surprisingly broad and deep. Typical tectonic settings for lakes include rifts (Lake Baikal, East African Rift System), transform plate margins (Dead Sea, Cenozoic basins of California), remnant ocean basins (Black Sea?), some foreland basins (Cenozoic Uinta Basin), and intracratonic depressions (Lake Eyre). Tectonic control of sedimentation is discussed at length in Chapter 9. Other lakes owe their origins to ice damming, volcanic damming, glacial scour. Temporary lakes form in various fluvial and coastal settings and may be the site for minor but distinctive deposits.

Another major sediment control is water chemistry. This can vary widely, reflecting variations in the inflow/evaporation balance, nature, and quantity of dissolved riverine sediment load, temperature, water level, and water-body structure (presence of stratification, seasonal overturning, etc.). As Talbot and Allen (1996) noted, lakes are very sensitive to climatic change. They provided a useful review of this topic and suggested a broad classification of lacustrine deposits into those formed in dilute lakes of hydrologically open basins and those formed in saline lakes of closed basins. However, with variations in climate or tectonic setting, lakes may change from one of these states to the other. Hardie et al. (1978) discussed subenvironments and subfacies that exist in saline lakes and showed how lacustrine brines evolve during evaporative concentration (Fig. 4.53).

Analysis of lake sediments may involve an examination of sediments and processes more commonly associated with marine or alluvial environments, including turbidites, deltas, evaporitic–carbonate mud flats, and fluvial sheet floods. These facies may be complexly interbedded within small stratigraphic thicknesses because of marked variations in water level that result from tectonic or climatic fluctuations (Donovan 1973, 1975; Link and Osborne 1978; Smoot 1978). Analysis of such deposits requires considerable sedimentological versatility. Recent research on continental-shelf environments has focused on the effects of storms and tides, and some of this work has found a parallel in the study of lacustrine sediments. For example, HCS, once thought to be an indicator of violent, long-fetch, marine storm waves, has now been documented in modern and ancient lake sediments (Duke 1985; Eyles and Clark 1986). Shallow-water lacustrine carbonates showing evidence of exposure and pedogenesis have been termed **palustrine limestones** (Freytet and Plaziat 1982; Platt and Wright 1992).

Recent interest in lacustrine sedimentation has focused on the African rifts (Frostick et al. 1986; Katz 1990; Tiercelin 1991) and Mesozoic–Cenozoic rift basins in China (Lao Qiuyuan and Gao Wenxue 1984). The Chinese basins are of considerable importance because their sediments are highly petroliferous, and the African rifts have received considerable attention, in part because of their value as well-exposed modern analogues for ancient lacustrine environments and potentially petroliferous sequences.

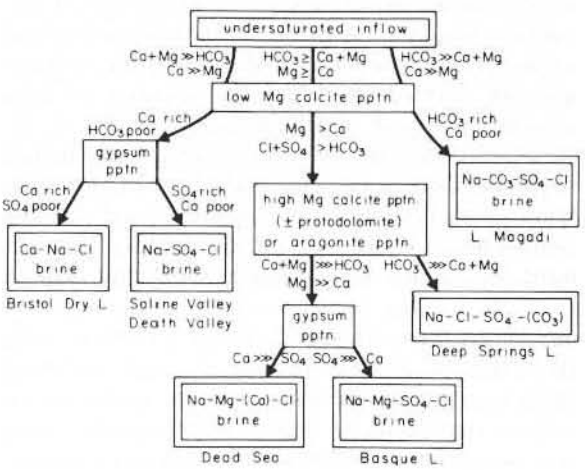


Fig. 4.53. Evolution of brines by concentration in saline lakes. (Hardie et al. 1978)

Because of their sensitivity to climate change and tectonism and the commonly rapid lateral and vertical facies changes, lacustrine sediments have been much studied by those interested in the processes of allogenic cyclic sedimentation. The processes of **orbital forcing** of climate change (the so-called **Milankovitch mechanisms**) have been studied in lacustrine sediments since their influence was first suspected in the Eocene Green River Shales of Wyoming (Bradley 1929). Recent studies of such cyclicity include work on Miocene deposits of northeast Spain (Anadón et al. 1991a) and the Triassic Newark Group of the eastern United States (Van Houten 1964; Olsen 1986, 1990).

4.6.4 Eolian Environments

The publication of satellite imagery of some of the large and inaccessible deserts of Africa and Asia (McKee 1979) demonstrated a complexity of dune types that has not been recognized in the ancient record. However, it has become apparent that many of these dune fields are not currently active, except that minor modifications of the giant bedforms (draas) are taking place. The major episodes of sand movement were during the Pleistocene ice age. This curtails the usefulness of modern-process studies of giant-dune formation. Fossil dune deposits demonstrate a surprising similarity and relative simplicity of form, as demonstrated by the useful summary papers in McKee (1979). We are only now beginning to understand why.

Since the early 1980s, research by Hunter, Brookfield, Kocurek, Rubin, Gradzinkski, and others on the internal structure of modern and ancient dunes and the processes of sand accumulation has provided us with a wide range of diagnostic environmental criteria with which to recognize and interpret ancient dune deposits (Sects. 4.5.4, 4.5.9). This work seems to have been largely responsible for stimulating an enormous increase in the amount of research being carried out on eolian facies, leading to the publication of several special journal issues and symposium volumes devoted to this topic (Brookfield and Ahlbrandt 1983; Hesp and Fryberger 1988; Kocurek 1988b; Fryberger et al. 1990; Pye and Lancaster 1993) plus many other individual papers too numerous to mention. Up-to-date summaries of the sedimentology of eolian deposits have been provided by Brookfield (1992) and Kocurek (1996). Allen (1993) briefly reviewed recent work on eolian bedforms. Eolian deposits are of some importance to the petroleum industry, because they constitute excellent reservoir rocks. Much of the natural gas produced from the North Sea Basin is trapped in the upper Paleozoic Rotliegendes Sandstones. Studies of these and other reservoirs has helped to stimulate the publication of much practical research on the characteristics of eolian reservoirs (North and Prosser 1993). The subsurface characteristics of eolian deposits, a discussion of their diagenesis, and some details on North American eolian production are provided by Fryberger et al. (1990).

Studies of eolian facies and architecture have been considerably assisted by developments in the understanding of their contained bounding surfaces, including the first- to third-order surfaces of Brookfield (1977) and the supersurfaces of Talbot (1985). It is now realized that bounding surfaces are developed by a wide variety of processes, including bedform climb, wind deflation, and regional changes in erg controls, such as climate and sand supply. Fryberger (1993) reviewed this topic and provided many illustrations of the various examples. He listed twelve types of surface ranging from primary ripple lamination to sequence boundary surfaces.

A second important research strand has developed from an interest in the dynamics of eolian sand transport and has led to a much-improved appreciation of the formation and preservation of cross-stratification in dune deposits (Rubin and Hunter 1982, 1983, 1985; Rubin 1987). Discussion regarding the interpretation of a single ancient deposit, the Rotliegendes Sandstone of the North Sea basin, illustrates the importance of this type of work (Glennie 1982, 1985; Steele 1985; Sneh 1988). Paleocurrent patterns, a major source of data in the paper by Glennie (1982), may be misleading if modern ideas regarding eolian bedform generation are not taken into account. Glennie used paleocurrent dispersal as a basis for interpreting dune type based on the assumption that dip directions indicate wind orientation. However, it has been shown that the direction of dip of major cross-stratification surfaces may be oblique to prevailing wind directions and that much sand may be carried obliquely along or across a dune surface, forming wind ripples (climbing, translatent strata) or small dunes (Fig. 4.54). The abundance of dune faces dipping at less than the angle of repose is a good indication that such processes are dominant (Rubin and Hunter 1985; Sneh 1988). Longitudinal dunes probably do not yield bimodal paleocurrent patterns, as has long been assumed. They form by slow lateral migration oblique to prevailing wind directions and probably internally develop simple, unimodal cross-stratification geometries (Rubin and Hunter 1985). The computer model for simulating complex dune forms that was developed by Rubin (1987) has been widely used (as was intended) for the analysis of real examples in the ancient record.

Many natural eolian systems, including many producing petroleum reservoir units, are interbedded

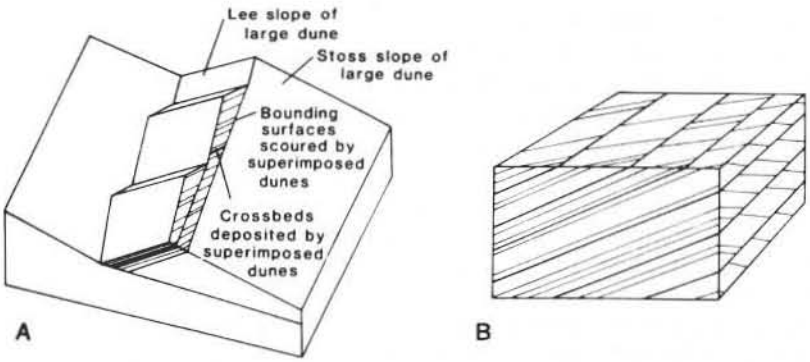


Fig. 4.54 A, B. Compound crossbedding generated by migration of dunes along the slope of a larger bedform. **A** Dune morphology and bedform migration. **B** Block diagram showing resulting crossbedding. (Rubin and Hunter 1983)

with the deposits of ephemeral streams, which may or may not contain fine-grained units. The fluvial and eolian facies tend to have very different facies and architecture and display different porosity and permeability characteristics. Most of the "eolian" papers in North and Prosser (1993) describe this type of mixed assemblage. A useful study of modern fluvial–eolian interaction was provided by Langford (1989), and a companion paper discusses an ancient example (Langford and Chan 1989).

4.6.5 Clastic Shorelines

The environments that are classified under this heading are: (1) dynamic coastlines, influenced by waves, tides, or storms, where there is an abundant, clastic sediment supply to build beaches, spits, and barriers, and (2) their associated subenvironments. Estuaries are also discussed in this section. Deltas are discussed separately in Section 4.6.2.

Recent textbooks treat clastic shorelines in different ways. All types of shoreline environment are included in a single chapter on this topic in the third edition of the classic Reading and Collinson textbook (1996), whereas, in the new "facies models" book (Walker and James 1992), the subject matter is subdivided into three chapters: "Deltas", "Transgressive Barrier Island and Estuarine Systems" and "Tidal Depositional Systems." This leads to a certain amount of overlap in the case of estuarine environments. An excellent compendium of recent research on tidal sedimentation was compiled by Smith et al. (1991).

The subdivision of coasts into domains on the basis of tide range (microtidal, mesotidal, and macrotidal; Hayes 1976) has been a very useful concept, but Hayes (1979) and Davis and Hayes (1984) showed that it is the relative effectiveness of waves and tides, not their absolute energy levels (wave height, tide range), that is important Thus, it is possible for a microtidal coast to be tide dominated if wave energy is low enough. Boothroyd et al. (1985) showed how this idea could be used to develop a modified classification of various coastal regions around North America and elsewhere (Fig. 4.55).

Much of the recent work in the area of sequence stratigraphy has focused on shoreline deposits because of their obvious sensitivity to sea-level change, and this has added a new dimension to our knowledge of their long-term responses to environmental change. For example, a long-standing controversy regarding the origins of barrier systems and their preservation had been largely resolved by Swift at al. (1985) and Niederoda et al. (1985) in a pair of detailed papers dealing with U.S. Atlantic-coast barriers that demonstrated rather conclusively the ability of barriers to migrate landward during transgression, feeding on their own sand to maintain volume. Sequence-stratigraphic reconstructions clearly confirm this process. As illustrated in Chapter 6, the transition from lowstand to transgression is accompanied, in many cases, by backstepping or retrogradation of coastal depositional systems and the partial stripping away of strandplain and barrier systems by the process of ravinement, with the sediment so released forming new beach systems further landward. Rather fundamental differences in coastal morphology are largely dependent on whether the coast is regressive or transgressive, as shown in Fig. 4.56.

Reinson (1984) and Hine and Snyder (1985) showed that subtidal inlet deposits and distributary-channel fills are the most likely facies to be preserved during transgression, as these are below the level of wave scour – the so-called **ravinement surface** (Figs. 6.7, 6.8). Reinson (1992) gave a useful discussion of the types of facies architecture and vertical succession that are developed during transgression and regression, and pointed out that such deposits are highly variable, depending on the local rate of sea-level change, sediment supply, local coastal configuration, riverine input, etc. Penland et al. (1985) developed a general model for barrier formation and drowning during the abandonment of delta lobes based on their studies of the Mississippi delta (Fig. 4.57). These two

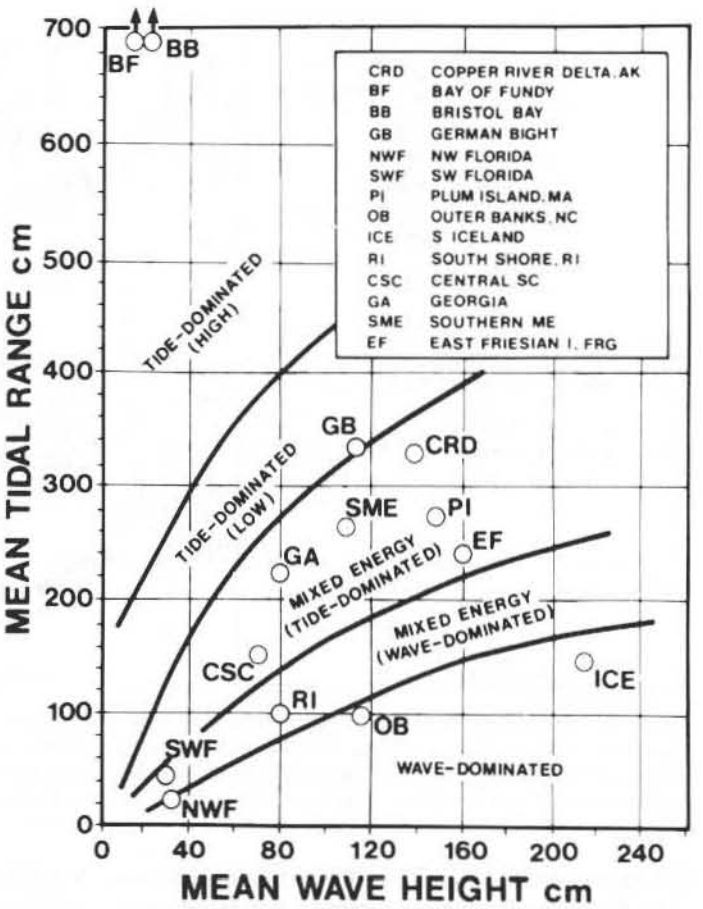


Fig. 4.55. Classification of coastal areas based on mean tide range and mean wave height. (Boothroyd et al. 1985; modified from Hayes 1979)

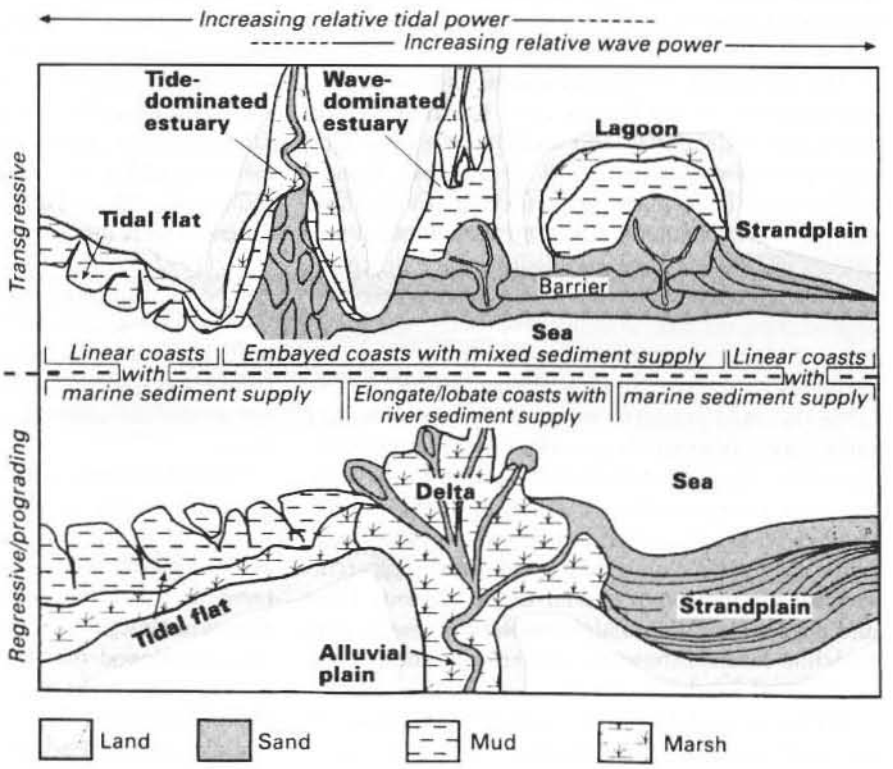


Fig. 4.56. Morphology of transgressive and regressive coasts, showing the variations in morphology dependent on variations in wave and tide power and sediment supply. (Reading and Collinson 1996)

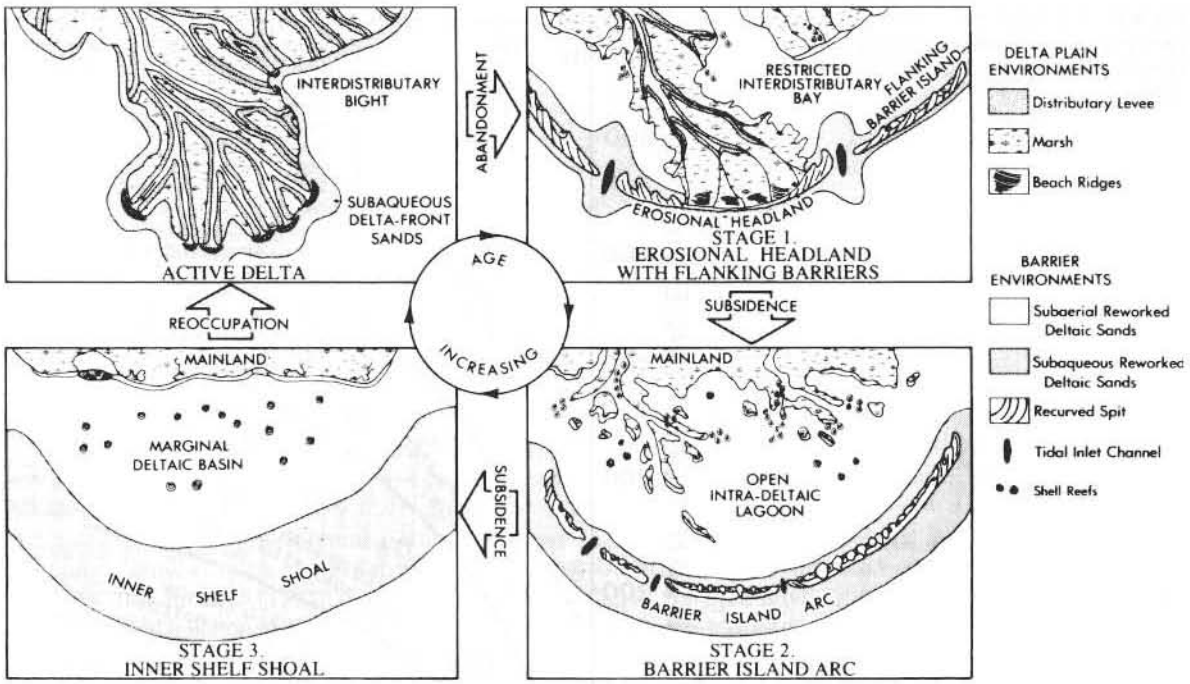


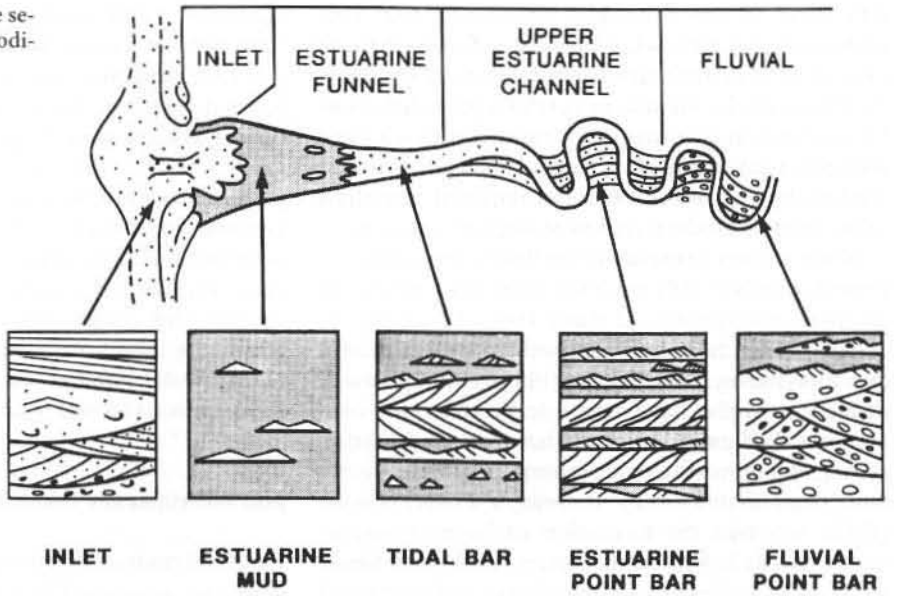
Fig. 4.57. The evolution of barrier systems above abandoned delta lobes. Barriers develop when a delta lobe is abandoned (*upper pair of diagrams*) but are progressively drowned or retreat shoreward as a result of subsidence (*lower left*). The model is based on research on the Mississippi Delta, but the ideas could be more generally applied to studies of transgressive–regressive changes, such as those that occur during cycles of sea-level change. (Penland et al. 1985)

discussions provide the basis for useful general models of transgressive and regressive coastal sequences, and the ideas are now part of the standard systems-tract models used in sequence stratigraphy (Chap. 6).

The subdivision of barrier systems into such subenvironments as the barrier core, beach shoreface, tidal inlets, tidal deltas, and washover fans has long been part of sedimentological terminology, but attempts have been made to further subdivide these morphological elements based on very detailed studies of geomorphology and sedimentary processes (Boothroyd et al. 1985). These subdivisions are now approaching the level of detail proposed in the architectural-element classification of fluvial systems (Sect. 4.6.1), except that the study by Boothroyd et al. (1985) is based primarily on surface geomorphology rather than sedimentary processes. A few attempts to apply this type of detailed classification to the sediments have been published. Galloway (1986) used petrophysical-log character and porosity–permeability patterns in petroleum reservoirs to develop an architectural subdivision of barrier-island deposits. Hine and Snyder (1985) used high-resolution seismic data to define coastal lithosomes in Cenozoic coastal successions off South Carolina.

A type of coastal depositional system that has only received detailed sedimentological study in recent years is the estuary. The Dutch tidal flats are estuarine and have been classic areas for the study of tidal sedimentation since the 1950s, but research there has tended to focus on individual depositional processes and products and has not led to the development of any overall models of estuarine sedimentation. Harris (1988) described the geomorphology, patterns of sand transport, and bedform development in several estuaries and discussed general models of estuarine sediment infilling. Fully developed facies models for estuaries were offered by Allen (1991) and Dalrymple et al. (1992, 1994b). They make use of our improved ability to recognize tidal signatures in siliciclastic sedimentary succession (Nio and Yang 1991); their models describe the gradual lateral transition from completely marine assemblages to fully fluvial deposits at the innermost end of the estuary (Figs. 4.58, 4.59).

Studies of Cretaceous units in the Western Interior Plains of North America, especially in the Alberta Basin, led to the recognition of a distinctive facies assemblage, for which Thomas et al. (1987) coined the term **inclined heterolithic stratification** and the acronym IHS. This facies consists of sand–mud couplets developed under tidal influence on the point bars of tidal creeks and tidally influenced rivers. Depositional dips reflect the formation of this facies by lateral accretion. Particularly spectacular examples of



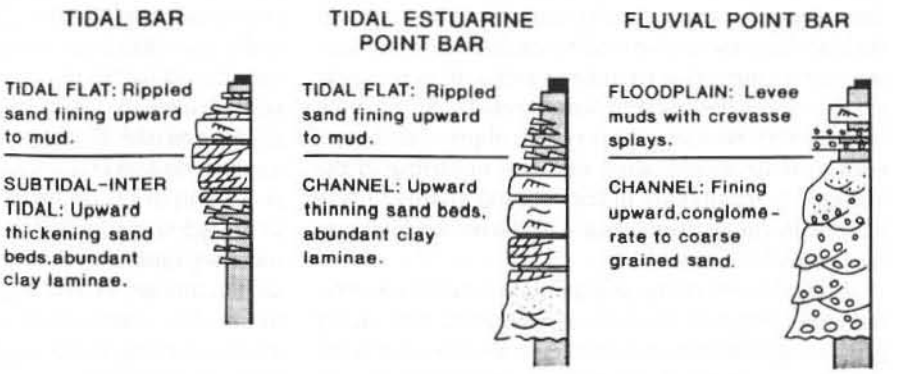


Fig. 4.58. A model of estuarine sedimentation. (Reinson 1992, modified from Allen 1991)

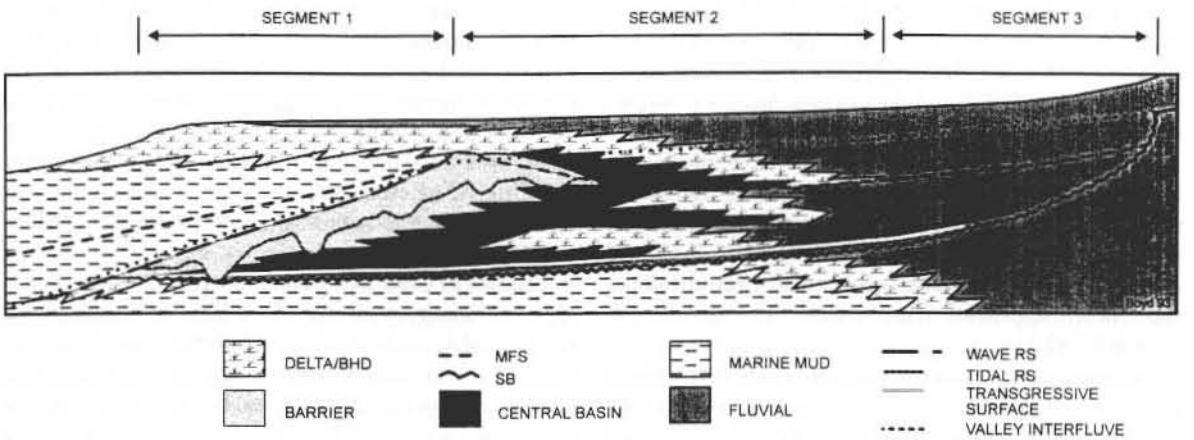


Fig. 4.59. Diagrammatic cross section along the axis of an estuarine, incised-valley fill. (Dalrymple et al. 1994b)

IHS occur in the Athabasca Oil Sands near Fort McMurray, Alberta, where tidally influenced point bars up to 25-m thick are exposed (Smith 1987, 1988a, b). The review by Thomas et al. (1987) provided a useful summation of ideas regarding estuarine-channel sedimentation, many of which had been known since the Dutch work of the 1950s but required reanalysis in the light of modern sedimentological concepts.

Many studies of estuarine bedforms have now appeared. Allen's (1980) model of tidal sand waves was an invaluable theoretical study (Sect. 4.5.4), the results of which have been applied to several ancient deposits (Blakey 1984; Teyssen 1984), although many of these examples have been interpreted in the context of a shelf environment rather than an estuarine setting. Modern estuarine examples of sand waves have been described by Dalrymple (1984). Harris (1988) reviewed the formation of larger estuarine bedforms, including sand ribbons and sand banks (tidal current ridges of other authors), and confirmed earlier ideas regarding the evolution of the latter by tidally driven race-track circulation of sand around them. Superb images of tidal sand waves (though in a shelf setting) were obtained by high-resolution seismic surveying, echo sounding, and side-scan sonar surveying off the coast of Normandy by Berné et al. (1988, 1991). Studies of this type will provide the basis for the next generation of facies modeling of the ancient record because of the detailed picture of surface bedform patterns and 3-D facies architecture they provide.

Interest in estuarine sedimentation received a tremendous impetus when it was realized that many producing sandstone reservoirs in the Western Interior basin, particularly in the lower Cretaceous Mannville Group of Alberta, occur in broad linear "channels" and contain evidence of tidal influence during sedimentation. Careful mapping revealed that these channels are deeply incised valleys that form part of regional erosion surfaces and that they originated as estuaries at the margins of the Western Interior Seaway (Rosenthal 1988). The new models of estuarine sedimentation that have now appeared (Dalrymple et al. 1992) are the result of this renewed interest, and a research volume devoted to the sedimentology and petroleum geology of incised valleys has now appeared (Dalrymple et al. 1994a; Sect. 6.5.1.4).

Another useful focus in coastal research has been an increased interest in high-energy, gravel-dominated coasts. Nemec and Steel (1984) provided a useful review of the textures, fabrics, sedimentary structures, and stratification features of conglomerates in fluvial and wave-dominated settings, including those formed by both traction and debris-flow mechanisms. The ability to distinguish these various environments and processes is particularly important in the case of coarse-grained fan deltas (Sect. 4.5.2), but conglomerates also occur on nondeltaic, wave-formed beaches. There they tend to show a well-developed seaward-dipping stratification, marked grain-size and clast-shape sorting, and seaward-dipping imbrication. Bourgeois and Leithold (1984) and Leithold and Bourgeois (1984) described these characteristics in some detail and illustrated several modern and ancient examples. An unusual deposit, a crossbedded conglomerate formed in a high-energy seaway by tidal processes, was described by Phillips (1984). Tidal bundles, evidence of current reversal, mud drapes, and reactivation surfaces are all present.

4.6.6 Arid Shorelines and Evaporites

Excellent reviews of the sedimentology of evaporites and their associated facies have been given by Kendall (1992) and Kendall and Harwood (1996), and this material will not be repeated here. Two important depositional models were developed in the late 1960s and early 1970s: the sabkha model, based on work carried out on the modern south coast of the Persian Gulf (Kinsman 1966, 1969; Shearman 1966; Purser 1973), and the desiccated-deep-basin model, based on outcrop and DSDP studies of the Messinian (Miocene) evaporites of the Mediterranean sea (Hsü et al. 1973; Schreiber et al. 1976; Ryan 1976). These two models profoundly affected our understanding of the sedimentology of arid regions, and the refinement of the models continues to this day. No other developments of comparable significance have occurred in this area of sedimentology in recent years. Warren and Kendall (1985) compared the sedimentology of sabkhas and subaqueous (salina) evaporites. The former are matrix dominated, with evaporite occurring mainly as diagenetic nodules, whereas salinas are evaporite dominated, composed of primary evaporite deposits that commonly show primary sedimentary structures. Sabkhas tend to form over paleotopographic highs, whereas salinas form in depressions, a distinction that may be of considerable use in paleogeographic reconstruction.

Studies of evaporites are hampered by several factors, including, typically, (1) a lack of detailed biostratigraphic control, which inhibits detailed correlation; (2) the susceptibility of evaporites to severe diagenetic change; and (3) removal of evaporites in the subsurface by dissolution, to the extent that entire stratigraphic units may disappear (Kendall 1992).

Many attempts have been made to model evaporite precipitation using calculations of evaporation rates and the rates of brine replacement by surface flow and subsurface seepage. These attempts are hamper-

ed by a lack of appropriate modern analogues where present-day processes can be studied. Several ancient evaporite deposits are far more extensive than anything forming now, which restricts the usefulness of modern-process studies (Kendall and Harwood 1996).

Kendall (1988) emphasized the importance of interpreting evaporites in their stratigraphic context and making full use of all available sedimentologic and stratigraphic data. For example, the long-known apparent association between basin-margin reefs and basin-center evaporites (as, for example, in the Silurian of the Michigan Basin) has generated a lengthy debate regarding the apparent incompatibility of these facies. As shown in Section 6.7, a simple model of cyclic sea-level change and evaporative drawdown comparable to that which occurred in the Mediterranean, readily explains this association. (However, the model seems fated to be continually rediscovered by those who do not review all available information; Cercone 1988.) Kendall (1988, 1992) and Kendall and Harwood (1996) are at pains to point out both that individual evaporite basins may pass through several different environmental conditions as brine concentration and evaporite sedimentation take place, and that apparent contradictions in the sedimentologic evidence can be resolved if this simple observation is taken into account. For example, evidence of deep-water sedimentation, such as lack of wave or current reworking, is not incompatible with a desiccated-deep-basin model of sedimentation, because the basin presumably starts out full and will already have developed considerable brine concentration and be precipitating evaporites while still containing a considerable residual water body. Kendall (1988) argued that Walther's Law cannot be applied to evaporites. Vertical successions of evaporite facies normally do not indicate lateral arrangements of environments but reflect basin-wide changes.

4.6.7 Clastic Continental Shelves

Our knowledge of the sedimentology of continental shelves underwent a major evolution in the 1980s. There were two main reasons for this: (1) improvements in the techniques of marine geology (high-resolution shallow-seismic and side-scan sonar surveying) that began to provide vastly improved data relating to shelf bedforms and facies distributions, and (2) shelf (and shoreline) deposits are one of the major components of stratigraphic sequences generated by regional (and global?) sea-level changes, recognition of which led to a revitalized interest in shelf processes and sedimentary controls on a basin scale (Chaps. 6, 8). Intensive study of shelf circulation processes, mainly by D.J.P. Swift and his colleagues on the Atlantic shelf of the United States and the British group (A.H. Stride, R.H. Belderson, N.H. Kenyon, and coworkers) on the shelves surrounding the British Isles, has led to a considerably improved understanding of tidal and storm-driven circulation and the resulting sediment transport and bedform-generation patterns.

These new ideas are summarized in a variety of recent review articles (McCave 1985; Walker and Plint 1992; Johnson and Baldwin 1996). Shelf environments and deposits have also been the focus of several symposia, and much important research data are contained in the proceedings volumes (Tillman and Siemers 1984; Knight and McLean 1986).

In general, the consensus that has emerged is that most shelves are either tide dominated (e.g., North Sea, Georges Bank) or storm dominated (e.g., US Atlantic shelf, Gulf Coast). A few areas of the continental shelf, such as the Agulhas Bank off southeast Africa, are dominated by strong, unidirectional oceanic currents that spill up onto the shelf (Flemming 1980; Martin and Flemming 1986). In tide-dominated shelves, the principal sedimentary processes are the development of sand waves and sand banks (ridges) whereas, in storm-dominated shelves, the main process is geostrophic flow (Fig. 4.60), leading to the development of storm cycles containing HCS (Walker and James 1992; Johnson and Baldwin 1996) and storm-generated, tidal sand ridges.

Many studies of shelf and estuarine bedforms have now been made using side-scan sonar and high-resolution seismic techniques (Luternauer 1986; Berné et al. 1988, 1991; Harris 1988; Davis and Balson 1992; Fig. 4.61), but the results have not yet led to a coherent synthesis of these features. Shelf sand waves and tidal sand ridges are the dominant forms but, as McCave (1985) noted, the terminology is confused, and he doubted whether these are genetically different types of bedforms, or whether they simply reflect changes in a single species of bedform in response to differences in depth and velocity. Belderson (1986) suggested that tidal sand sheets consisting mainly of sand-wave deposits are a lower-energy facies than the sand-bank facies. Sand waves are typically up to 7 m high, but may be as high as 15 m (Fig. 4.61). The internal details of their stratification reveal much regarding tidal-velocity asymmetry and the nature of local tidal cycles (diurnal, semidiurnal, neap–spring, etc.; Allen 1980; Visser 1980). Sand ridges or sand banks (Harris 1988) are up to 50 m high and 2 to 3 km wide. Little is known regarding their internal architecture except for the presence of gently dipping master bedding surfaces (Houbolt 1968; Fig. 4.62). Tidal sand banks are commonly oriented at an angle between 7° and 15° relative to local tidal-current direc-

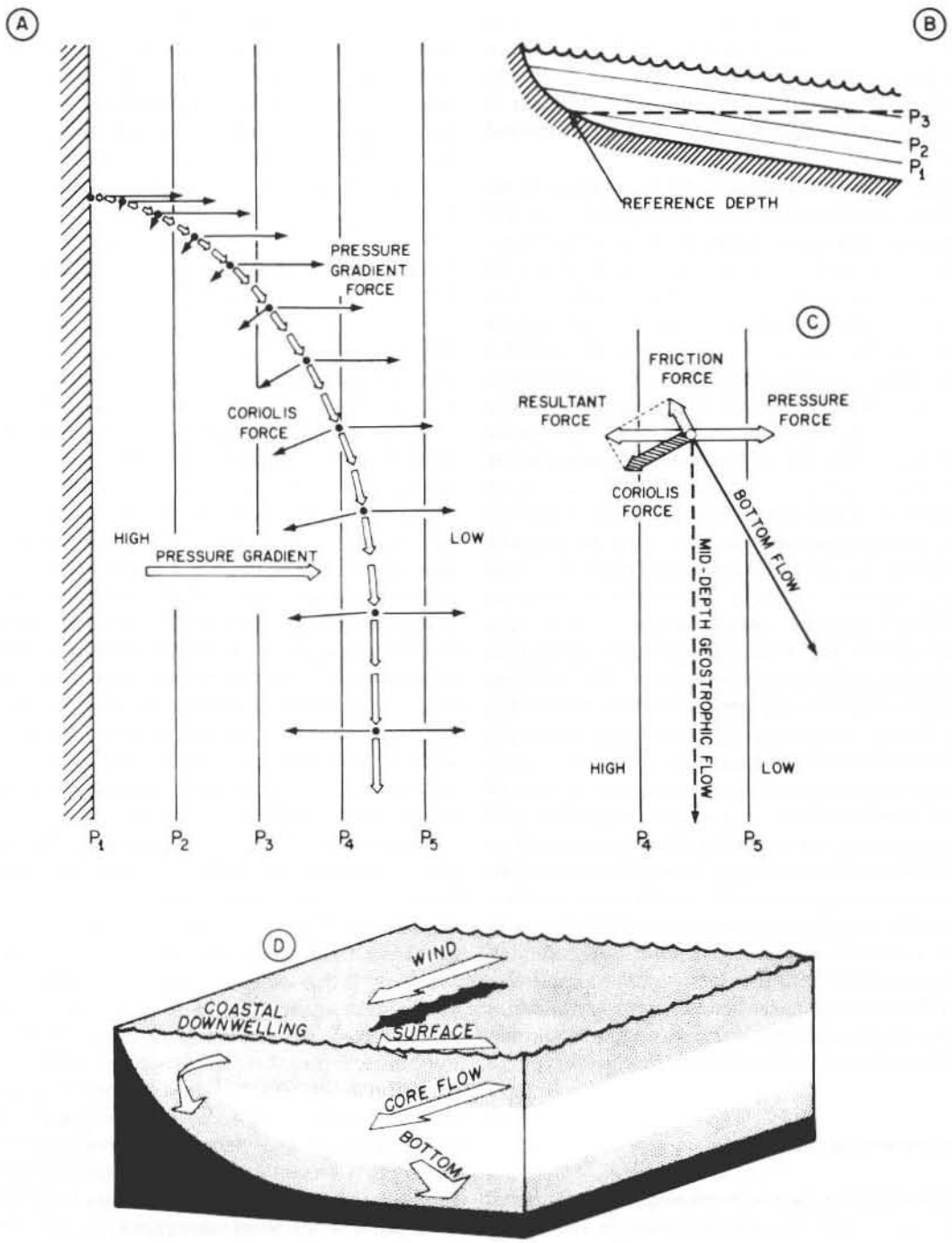


Fig. 4.60 A–D. Geostrophic flow on the continental shelf. Storm winds directed toward the coast pile up the sea against the shore (coastal setup). This generates a pressure gradient, the result of which is a compensating downwelling current that moves obliquely offshore and is deflected by the Coriolis effect. Sustained current velocities of 30 cm/s occur, with maximum velocities up to 2 m/s. (Swift 1976; Swift and Rice 1984)

tions (Harris 1988). They are formed by circular movement of sand detritus under the influence of tides.

Belderson (1986) discussed the differences between tide- and storm-generated sand banks and disagreed with the conclusion of Swift (1975b) that these are essentially the same type of structure. Storm-generated sand ridges are oriented at higher angles to the direction of main flow (typically 35–40°). They are generally smaller than tidal sand banks (average height 7 m), have smaller lengths and

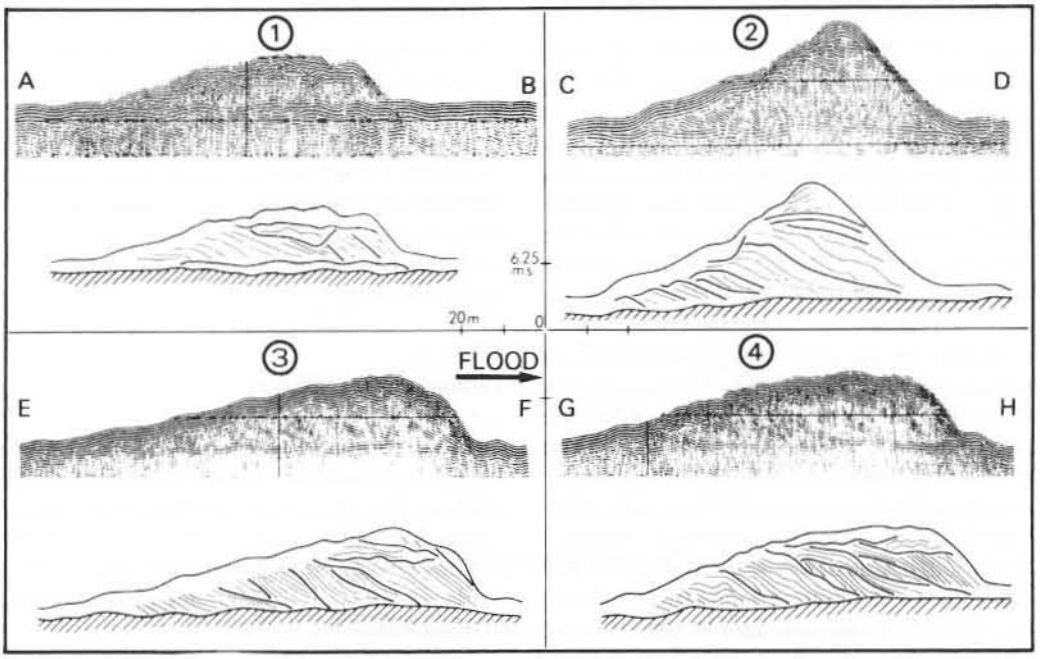


Fig. 4.61. Seismic profiles across typical tidal sand waves off the coast of Normandy, France. The vertical scale of 6.25 ms is approximately equivalent to 5 m. Vertical exaggeration is, therefore, about ×4. These bedforms are classified as types IIIB and IVB of Allen (1980). (Berné et al. 1988)

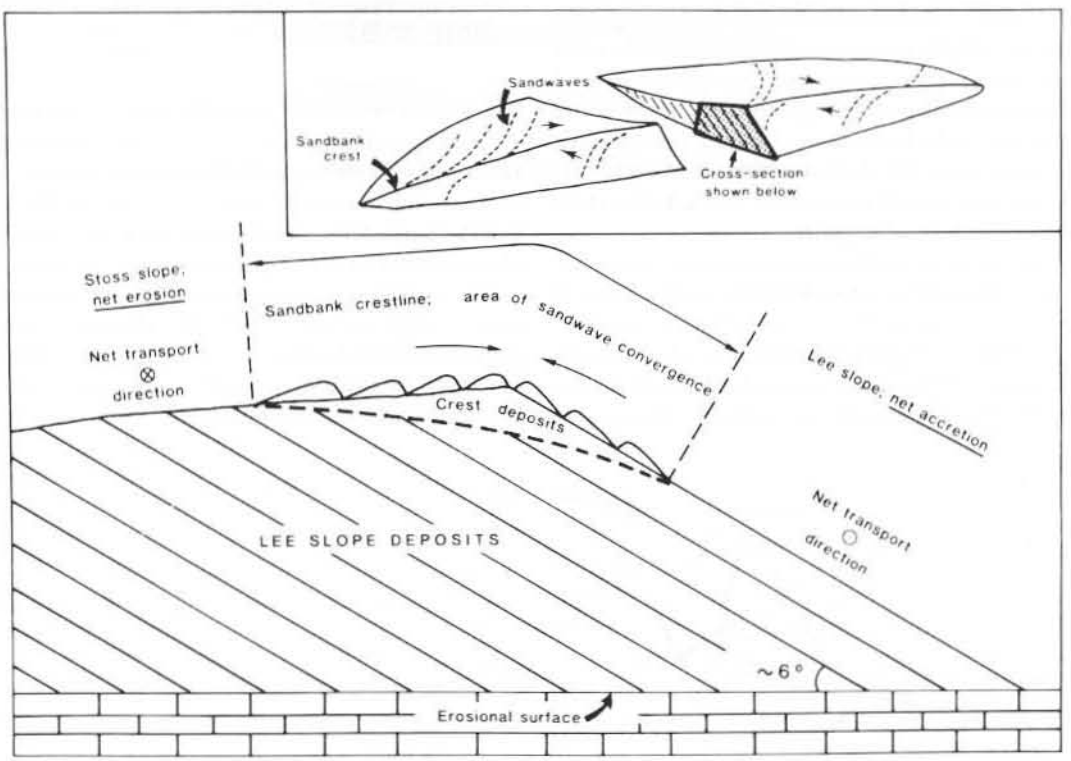


Fig. 4.62. Model for the generation of tidal sand banks. The main body of the bank results from deposition on the lee slope, forming master bedding planes (fourth-order bounding surfaces; Table 4.5) and consisting internally of sand-wave cross-stratification. (Harris 1988)

spacing, and have lower depositional slopes on the sides (less than 2°, versus 6° for tidal banks). A model for the internal structure of storm-generated sand banks was developed by Swift et al. (1986). They rest on lag deposits. On the downcurrent side of the ridge, high-energy, graded sets are deposited. These are succeeded upward by HCS deposits and then by cross-stratified sets formed from the migration of megaripples. This last facies caps the ridges and blankets the upcurrent side. The vertical and lateral sequence reflects decreasing flow strength (Fig. 4.63).

Many studies of ancient shelf deposits have attempted to reconstruct offshore sand bars using modern sand ridges as analogues (Evans 1970; Brenner 1980; Boyles and Scott 1982; Tillman and Martinson 1984; Brenner et al. 1985), but in none of these studies is the comparison entirely satisfactory, because of the paucity of information from modern banks on which to base the comparisons. Most ancient deposits previously interpreted as offshore sand bars have been reinterpreted during the last decade based on sequence-stratigraphic concepts. As first demonstrated by Plint (1988) and as reviewed by Walker and Plint (1992), most can now be demonstrated to have formed as shoreface deposits during times of falling sea level, when shorelines moved many kilometers out across the shelf. Many such deposits rest on surfaces of regressive marine erosion and are, therefore, "sharp-based". Most have been partially eroded by wave scour, which developed a regional ravinement surface during the subsequent transgression, and have become stratigraphically isolated, in the sense that they are underlain and overlain by fine-grained shelf clastics, typically shale. Walker and Plint (1992) proposed a new model, termed the **incised-shoreface model**, for this type of deposit.

The importance of storms as geologic agents has become widely appreciated with the publication of the useful reviews by Aigner (1985) and Brenchley (1985). Following, in part, the earlier work of Einsele and Seilacher (1982), they focused attention on the large body of sedimentary information to be obtained from detailed study of sedimentary events, such as storm deposits or **tempestites**. For example, they demonstrated independently the use of various small-scale facies criteria to establish proximality indicators for storm sequences (Fig. 4.64). Myrow and Southard (1996) re-examined the interaction of the three major processes that combine to form tempestites (waves, geostrophic flow, and density currents) and generated a predictive depositional model for these deposits.

4.6.8 Carbonate Environments

Carbonate sediments are primarily biochemical in origin. Light intensity, temperature, and nutrient supply are critical controls and limit the zone of rapid carbonate sedimentation to depths of less than about 20 m (Fig. 4.65 A). Sedimentation follows what Schlager (1992) called the **law of sigmoidal growth** (Fig. 4.65B):

> Growth is slow at first, then accelerates, often exceeding the rate of change in the forcing function; finally, growth decreases as the system reaches the limits of the newly formed niche. In carbonate sedimentology, this pattern is known as the start-up, catch-up, keep-up stages of growth ... (Schlager 1992)

The interaction of the controls illustrated in Fig. 4.65 is the reason why most carbonate sedimentation takes place in the so-called **carbonate factory** of the shallow-marine continental shelf. The study of carbonate facies cannot be divorced from considerations of diagenesis. In many environments, carbonate sediments start to undergo cementation and alteration as soon as they are deposited. The ubiquity of organic processes in carbonate seas introduces a significant complication, and many carbonate deposits are modified penecontemporaneously by submarine cemen-

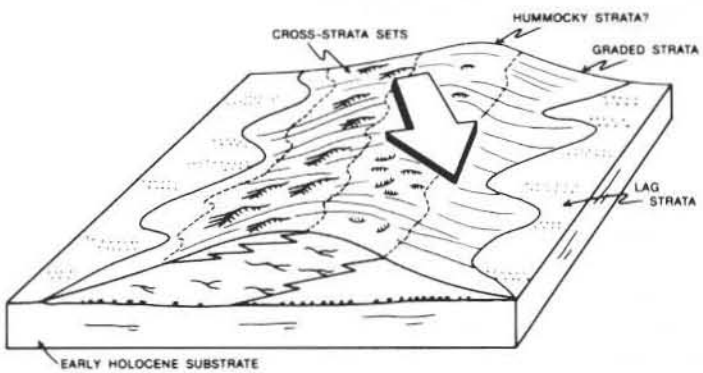


Fig. 4.63. Model for the internal architecture of storm-generated sand banks, based on studies of the New Jersey shelf. *Large arrow* shows flow direction. (Swift et al. 1986)

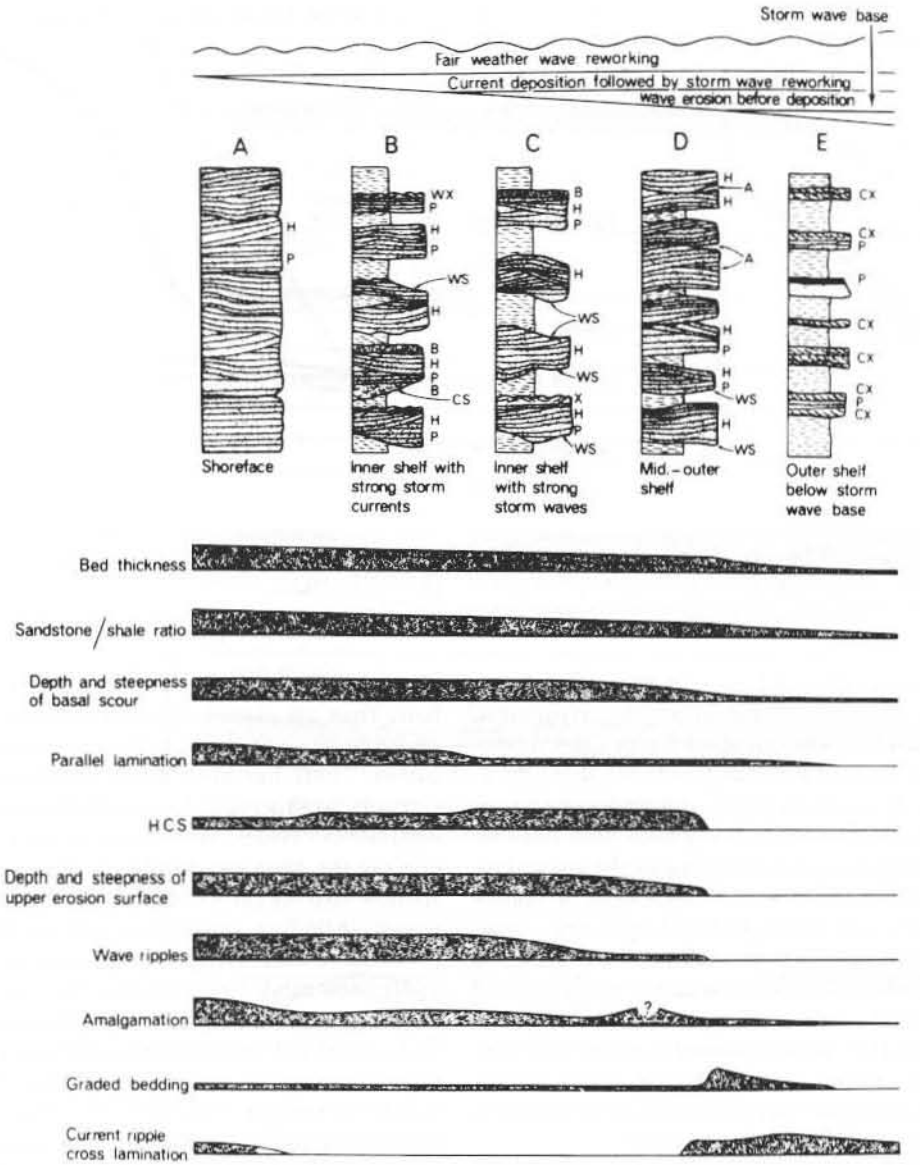


Fig. 4.64 A–E. Facies changes in storm sequences with distance from shore and progressively greater depth of water. *WS* Wave scour; *CS* current scour; *B* basal lag; *G* grading; *P* planar lamination; *H, HCS* hummocky cross-stratification; *WX* wave cross-lamination; *CX* current cross-lamination; *M* bioturbation; *A* amalgamation. (Brenchley 1985)

tation and other types of early diagenesis. For example, reef and carbonate-slope buildups commonly have steeply dipping sediment–water interfaces, much steeper than the angle of repose of loose talus, sometimes even vertical or overhanging. This is due to organic binding and early cementation. Most research now makes extensive use of special microscopic techniques, such as cathodoluminescence and scanning electron microscopy, plus a range of geochemical techniques, including the analysis of stable isotopes and trace elements, in order to track these diagenetic changes (Dickson 1985). Cathodoluminescence is widely used to see through diagenetic modifications to primary depositional fabrics (Machel 1985). Analyses of oxygen and carbon isotopes are used to determine depositional temperature and to distinguish early from late diagenetic cementation (Arthur et al. 1983). In many other respects, carbonate-facies studies can be pursued in a similar manner to siliciclastic studies, in that many carbonate deposits are accumulated by clastic processes controlled by waves, tides, and gravity.

One of the most important contributions to the study of carbonate facies models was the book by J. L. Wilson (1975) on Phanerozoic carbonates. Wilson demonstrated that carbonate sediments could be

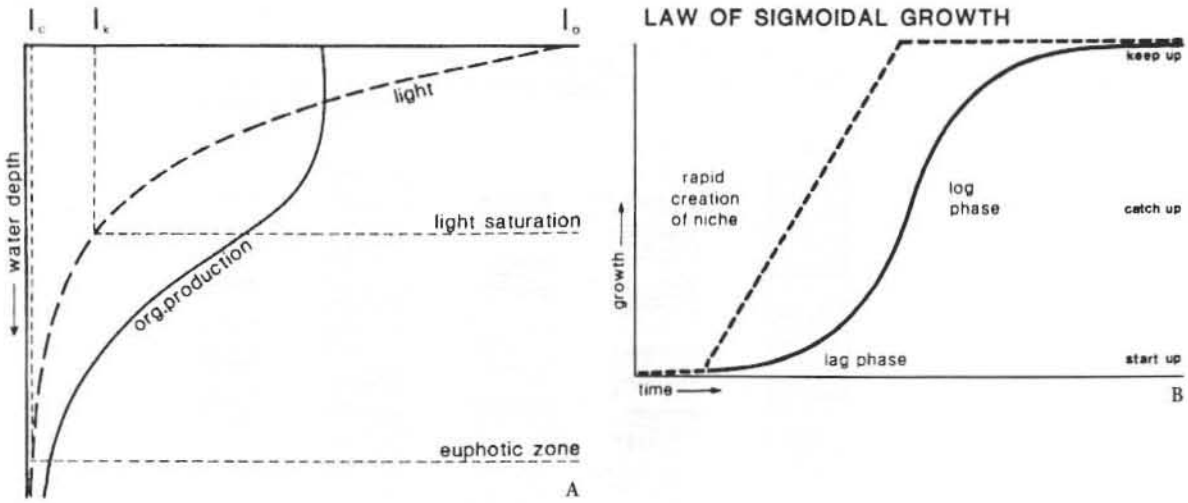


Fig. 4.65. A Growth rates for carbonate sediments, as dependent on light intensity. B The law of sigmoidal growth in carbonate sediments. Populations of organisms respond to the opening up of new living space slowly, at first. In most situations, sediment production can exceed the rate of increase of new space but is limited by the rate of growth of space. (Schlager 1992)

described using a standard set of 24 microfacies types (Sect. 4.3.1). He showed that lithofacies, structures, and biota defined nine standard facies assemblages that could be used to describe practically all carbonate rocks in the ancient record, although not all nine assemblages are present in every rock unit. Wilson's models are widely used in interpreting the ancient record. More recent syntheses and reviews of carbonate sedimentation have been provided by Scholle et al. (1983a), Tucker (1985), James and Walker (1992), and Wright and Burchette (1996). As noted by Wright and Burchette (1996) "the development of small-scale facies models, the essential tools for facies sedimentology, [have] all but ceased and few, if any, refinements have taken place in many of the simple models developed through the 1960s and 1970s."

Sellwood (1986) suggested that five major environmental zones could be recognized in shallow-marine settings: supratidal flats, intertidal flats, the marine shelf, the carbonate sand belt, and the reef belt. To these should be added the continental slope and pelagic environments, most of which are deep marine (Jenkyns 1986). In many respects, shelf and slope carbonate environments are similar to their siliciclastic counterparts. Thus, the sedimentology of carbonate shelves involves the study of storm deposits, some containing HCS (Aigner 1985; Handford 1986) and carbonate sand shoals similar to siliciclastic sand waves and banks (Hine 1977; Cook et al. 1983). Carbonates also occur as eolianites, most of which originated as coastal-dune deposits (McKee and Ward 1983).

Shoaling-upward cycles are common in shelf and tidal environments (James 1984b; Wright and Burchette 1996) and have been the subject of much debate. They are commonly referred to as **meter-scale cycles** because of their characteristic range of thicknesses. There has been active discussion about the controls on this type of cyclicity for some time. The arguments revolve around autogenic processes, including the rates of carbonate sedimentation in the shallow-marine carbonate factory, and allogenic processes, including subsidence and sea-level changes (Wilkinson 1982; Wright 1984; Jones and Desrochers 1992; Pratt et al. 1992; Wright and Burchette 1996). Shoaling-upward cycles are probably caused primarily by tidal-flat progradation and lateral shifting of subenvironments under conditions of slow but steady subsidence (Fig. 4.66). Ginsburg (1971) suggested that infilling of new accommodation space on a slowly subsiding platform would generate lateral tidal-flat progradation but would eventually fill all available space, and the carbonate factory would shut down until subsidence had generated new space. A lag in the renewal of carbonate-sediment generation would permit flooding, and renewed progradation and upward shoaling (James 1984b; Wright 1984). However, this model cannot explain subaerial exposure of cycle tops and rhythmic or periodic changes in cycle thickness (Goldhammer et al. 1990). The importance of sea-level change as an additional factor in this process is increasingly being emphasized (Wright and Burchette 1996). Interaction between the rate of development of new accommodation space and the rate of carbonate-sediment generation leads to variations in the style of carbonate cycles and in their regional geometry, as summarized in Fig. 4.67. Computer-simulation models are now routinely used

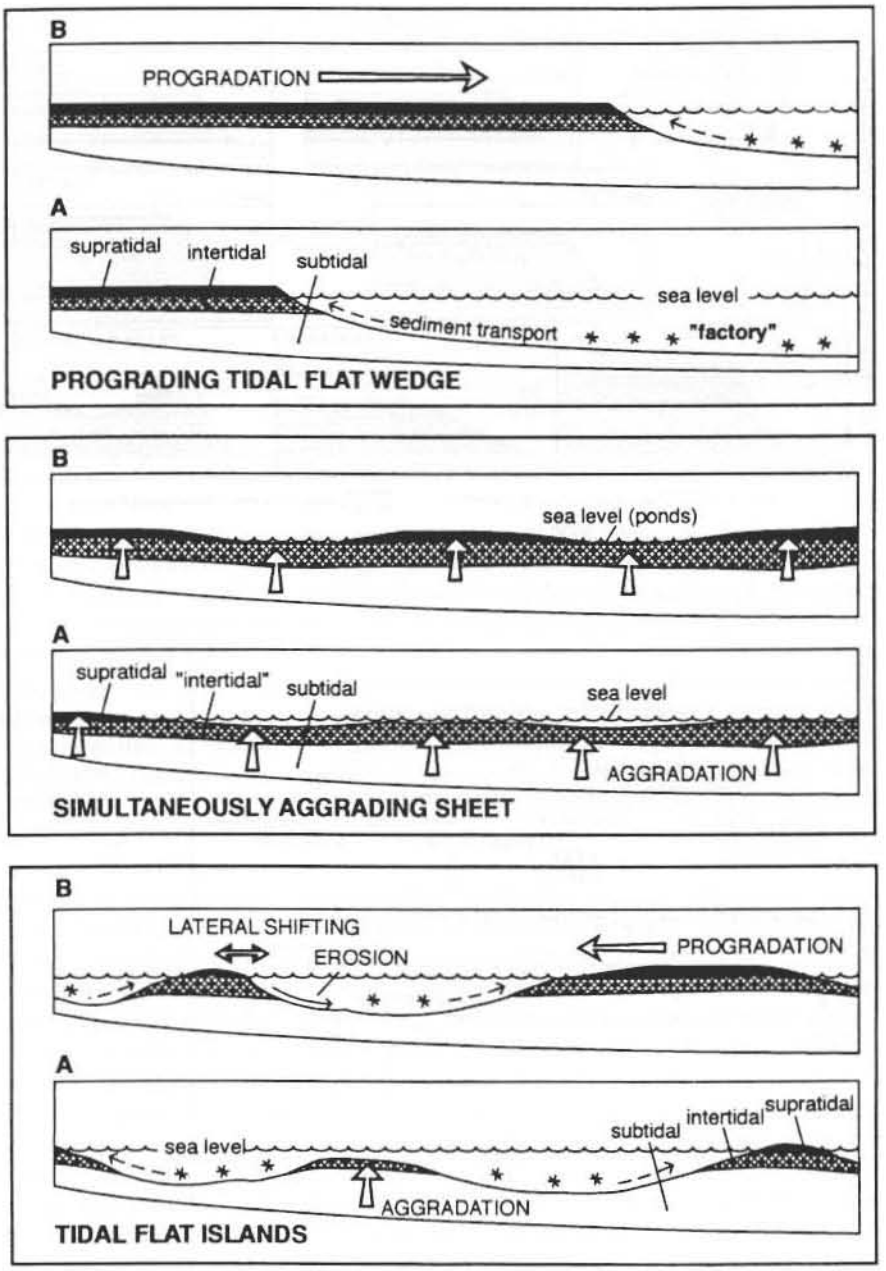


Fig. 4.66. Mechanisms that have been proposed for the generation of shallow-water, meter-scale carbonate cycles. (James et al. 1992)

to explore these variables (Grotzinger 1986; Read et al. 1986).

One of the more significant developments in recent years has been a recognition of the importance of carbonate platform sedimentation in cool-water environments, such as the west coast of Ireland and the Great Australian Bight (Lees 1975; Nelson 1988; James and Bone 1991; Boreen and James 1995). Where the ambient temperature falls below 15 °C, hermatypic corals are absent, and the carbonate-building association is dominated by bryozoa, foraminifera, red algae, and molluscs. The rate of sediment accumulation is about an order of magnitude slower than in warm-water settings (James and Bone 1991).

Reefs continue to receive considerable attention from carbonate geologists, both because of their intrinsic ecological interest and because of their importance as petroleum reservoirs. Excellent, recent reviews of reef sedimentology have been given by James (1983, 1984c) and James and Bourque (1992). James divided reefs into five zones (Fig. 4.68) and subdivided the reef core into four growth stages (Fig. 4.69).

The morphology of modern reefs is controlled, in part, by the Pleistocene karst surfaces on which they

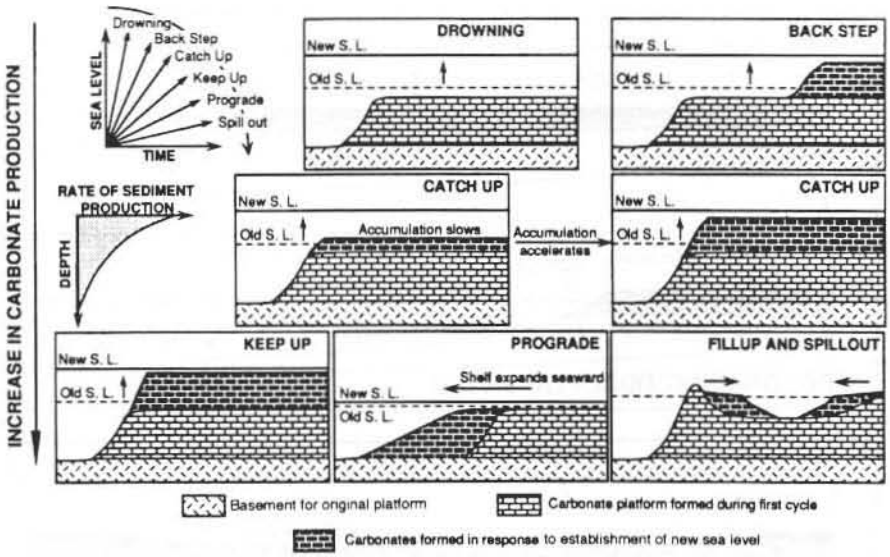


Fig. 4.67. Variations in cycle type and shelf architecture that reflect variations in the rate of sediment production. (Jones and Desrochers 1992)

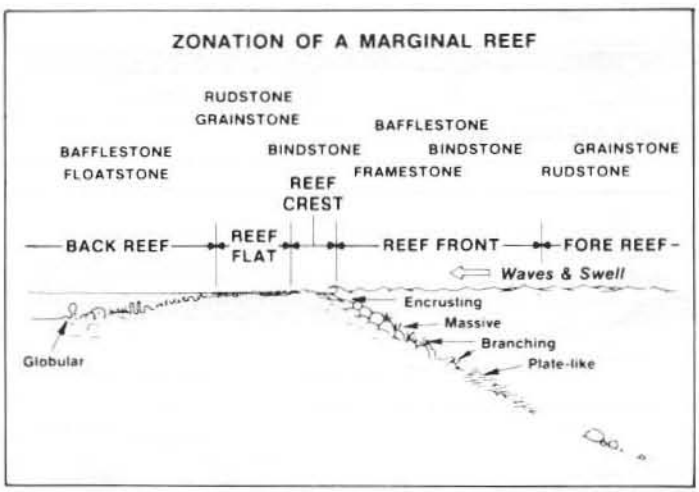


Fig. 4.68. Environmental zones of a marginal reef (James 1984c). Rock types that form in each zone are shown, using the terminology of Dunham (1962) and Embry and Klovan (1971)

STAGES OF REEF GROWTH



STAGE	TYPE OF LIMESTONE	SPECIES DIVERSITY	SHAPE OF REEF BUILDERS
DOMINATION	bindstone to framestone	low to moderate	Laminate encrusting
DIVERSIFICATION	framestone (bindstone) mudstone to wackestone matrix	high	domal massive lamellar branching encrusting
COLONIZATION	baffiestone to floatstone (bindstone) with a mud stone to wackestone matrix	low	branching lamellar encrusting
STABILIZATION	grainstone to rudstone (packstone to wackestone)	low	skeletal debris

Fig. 4.69. Four stages of growth of a reef core. (James 1984 c)

have grown (Purdy 1974; Sellwood 1986). For example, Purdy suggested that atolls formed because, during low-sea-level stands, more dissolution occurred in the center of an exposed platform than at the margins. The latter, therefore, became an elevated rim, and subsequent reef growth maintained this relief. Adey (1978) disputed this interpretation. He showed that vertical reef accretion can be very rapid, up to 15 m/1000 years, but that actual rates are dependent on nutrient availability, wave energy, and water clarity. These factors control the composition of the biota, some components of which grow faster than others. Reef shape throughout the later Cenozoic, therefore, reflects environmental variables as well as erosional topography. This has important implications for interpreting the ancient record.

Studies of many reefs (Mountjoy 1980; Stoakes 1980) show that their development was controlled by a series of eustatic sea-level changes, such as are described in Chapter 8. The reefs consist of several or many partially developed or eroded shoaling-upward cycles, commonly separated by diastems. Basin-fill sediments between the buildups consist either of fine detrital carbonates plus mudstone, as in many of the Devonian reefs of Alberta, or evaporites, as in, for example, the Silurian pinnacle reefs of the Michigan Basin (Mesolella et al. 1974). These reefs had a considerable relief above the basin floor, and the basin-fill sediments banking against them are, therefore, considerably younger. The contact is that of an onlap rather than interfingering. This may be difficult to demonstrate from scattered wells but is clearly shown on high-quality seismic records (Brown 1981). Many reef bodies are commonly termed **stratigraphic reefs** because the reef mass results from stratigraphic stacking of individual bioherms, which may themselves have had little relief. Each bioherm may display several of the growth stages illustrated in Fig. 4.69, separated from each other by hardground or karst surfaces. The recognition of the cyclic nature of deposition has considerable consequences for the interpretation of major reef bodies. They may have been deposited during short intervals of rapid buildup, alternating with longer intervals of stillstand, submarine cementation, and hardground development, or even exposure, karstification, and vadose diagenesis. Walther's Law can only be used with extreme caution in interpreting facies relationships in such complex deposits.

The organic components of reefs have changed dramatically with time, from the stromatolite reefs of the Precambrian and the simple archaeocyathid mounds of the Early Cambrian to the fully diversified reefs of Devonian and younger rocks. Periods of abundant reef growth (James 1983, 1984c; James and Bourque 1992) appear to coincide with times of long-term, high sea level (Chap. 8), whereas times of sparse reef growth occurred when sea levels were low, shelf areas smaller, and reef communities underwent extinctions and reorganizations (Sheehan 1985).

Many important advances in carbonate facies studies have been made in deep-water deposits as a result of improvements in research methods in the deep sea, including the development of submersibles and improvements in seismic methods and drilling technology. The DSDP began studying carbonate slopes and platforms in 1985. These studies demonstrate "that all deep carbonate bank margins are not simply areas of accumulation of 'fore-reef talus' and pelagic sediments" (Mullins and Neumann 1979). The work of Mullins and Neumann (1979), Schlager and Chermak (1979), the many excellent papers in Cook and Enos (1977), the reviews by Cook et al. (1983), Cook and Mullins (1983), Enos and Moore (1983), and and the summary articles by McIlreath and James (1984), Coniglio and Dix (1992), Schlager (1992), and Wright and Burchette (1996) describe a variety of distinctive facies types, topographic patterns, and facies assemblages, reflecting tectonic control of the bank edge and the effects of waves and oceanic currents in dispersing the sediment.

Bank margins vary from steep, possibly fault-controlled cliffs to gentle slopes extending for up to hundreds of kilometers. Coarse-grained talus-slope breccias occur on the steeper slopes, and are flanked by a variety of sediment-gravity-flow deposits, including debris flows, liquified or fluidized flows, and turbidites. These are interbedded with calcareous pelagic and hemipelagic sediments composed of planktonic organic detritus or reworked, fine platform sediment. Deep-water bioherms may occur on some slopes. These deposits are commonly cut by spectacular intraformational truncation surfaces.

The facies assemblage on any particular margin may vary markedly along strike, depending on the bank topography, strength of waves, and oceanic currents. Detailed study by Mullins and Neumann (1979), based on the Bahama Bank margin, revealed seven slope types. McIlreath and James (1984) proposed four basic slope models. Perislope talus breccias are coarser and thicker adjacent to cliff margins. Detritus may be carried into the basin by submarine canyons, but commonly there is a series of small gullies cutting the slope, fed from numerous sources on the bank rather than a single point source. These gullies channel the sand into a series of ephemeral, overlapping, and coalescing calcarenite lobes rather than single, large submarine fans. Carbonate-sediment gravity flows typically contain less mud than siliciclastic flows and, therefore, do not travel so far out into the basin plain but tend to terminate rather abruptly at the foot of the slope. Based on these and other differences between carbonate and siliciclastic

slopes, Mullins and Cook (1986) and Colacicchi and Baldanza (1986) developed a carbonate-apron model for carbonate-slope deposits. On windward margins, sediment accumulation is more rapid than on leeward margins, where it may be swept away. Ocean currents may rework the debris into contourites or remove it altogether, producing an erosion surface. The importance of sea-level change as a control on carbonate-slope development is discussed in Chapter 6.

Pelagic facies were first recognized in carbonate rocks of upper Silurian age but did not become volumetrically important until the evolution of planktonic foraminifera and coccoliths in the Jurassic. It has been estimated that, in the last 100 m.y., pelagic sediments have constituted 67% of world carbonate deposition (Hay et al. 1976). They were particularly important during the late Cretaceous high-sea-level stand (Chap. 8), when vast areas of Europe and North America were covered by a detritus-free epeiric sea in which the distinctive chalk facies accumulated (Scholle et al. 1983b). Carbonate deposits are sparse below depths of between 4 and 5 km in modern oceans because of the dissolution of calcite in deep water. This depth, known as the **carbonate compensation depth** (CCD) varies depending on temperature and CO_2 content, the latter being controlled by organic productivity and the age of the water mass (Scholle et al. 1983b).

In certain areas of the continental shelf, there is an intimate association between carbonate and siliciclastic sediments. For example, on steep tropical coasts, such as those bordering parts of the Red Sea and Puerto Rico, coarse-grained fan deltas are the typical clastic facies. Commonly, reefs develop on temporarily inactive fan lobes. On broad shelves, such as that on which the Great Barrier Reef of Australia is situated, clastic facies dominate coastal areas, whereas carbonate sedimentation is predominant in outer shelf areas. Studies of the relationships between these contrasting facies have, perhaps, not received as much attention as they should, a point that encouraged Doyle and Roberts (1988) to compile a useful synthesis of research in the area. To those studying sequence stratigraphy and the influence of sea-level change, carbonate–clastic transitions are of particular interest because of the sensitivity of carbonate sedimentation to changes in detrital influx, which is very dependent on coastal slopes and the position of the shoreline with respect to the shelf margin.

4.6.9 Clastic Continental Slope and Deep-Oceanic Environments

Rapid developments in our knowledge of submarine fans and other deep-marine turbidite systems occurred in the mid-1980s, with several milestone papers having appeared during that period. A vast body of DSDP information and other data has been generated from modern deep-sea fans. The use of side-scan sonar and high-resolution seismic techniques have revolutionized our knowledge of fan physiography, providing the ability to see individual depositional systems and architectural elements on a scale comparable to that of low-level aerial photographs of features on land (Gardner et al. 1996). Leg 96 of the DSDP, run in the fall of 1983, was specifically planned to examine the deep-sea fan of the Mississippi and was the first sustained attempt to drill deep holes into coarse-grained fan-channel deposits. Leg 155 was drilled over the Amazon fan, also with the objective of investigating sedimentary processes on a large submarine fan (Flood et al. 1995). Other large fans, such as the Indus, have also received much attention in recent years. Much outcrop information from Italy, Alaska, California, and elsewhere has also been collected during the last two decades. Much of this new information was synthesized at a special meeting in 1982, convened by A. H. Bouma. With the assistance of W.R. Normark and N.E. Barnes, he formed COMFAN, the Committee on Fans, and the proceedings of that meeting were presented in a special issue of Geomarine Letters (Normark et al. 1983). A revised and expanded version of this compilation, including the first results of DSDP Leg 96, later appeared in book form (Bouma et al. 1985). The stimulus provided by the availability of this huge, consolidated body of data led to the publication of several other important reviews (Mutti 1985; Stow 1985, 1986; Mutti and Normark 1987; Shanmugam and Moiola 1988; Stow et al. 1996). A careful reading of these recent papers and books is essential for an up-to-date comprehension of submarine-fan sedimentology.

The simplistic, all-purpose submarine-fan models of Mutti and Ricci-Lucchi (1972) and Walker (1978) are all but obsolete. The enormous ranges in size, shape, physiography, and facies composition of modern fans (Fig. 4.70) render such models virtually meaningless except as introductory teaching aids. Also called into question are some of the terms used to describe submarine fans. In particular, Shanmugam and Moiola (1988), in a very useful review, demonstrated the inconsistencies and contradictions that have crept into the literature in the use of such terms as upper, middle, and lower for fan subdivisions, the many varied ways in which the term lobe has been employed, and the confusion that has arisen over the concepts of transport efficiency. All such terms should probably be abandoned, although Shanmugam and Moiola (1988) attempted to arrive at compromise definitions of the three fan subdivisions (Fig. 4.71):

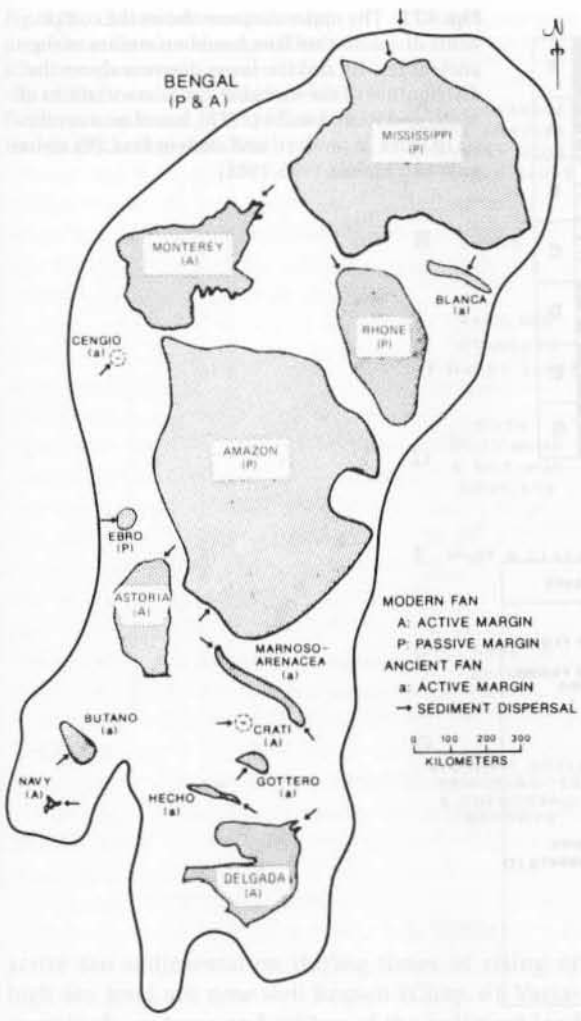


Fig. 4.70. Outline of selected submarine fans and turbidite systems, showing variability in size and shape. Note the absence of a true fan shape in many cases. (Bouma et al. 1985)

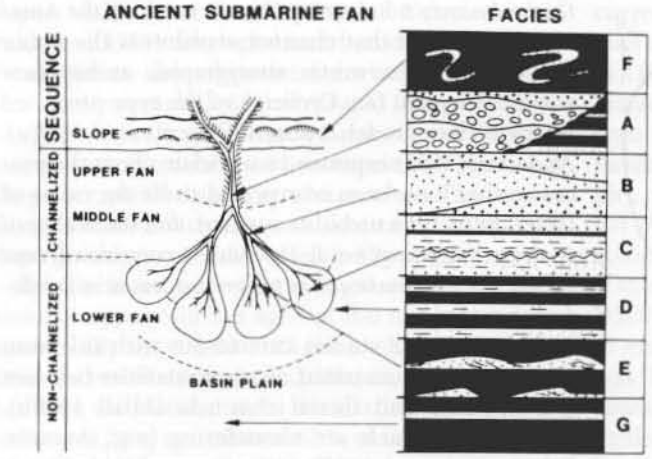
(1) the upper fan [can be recognized] by the presence of a major feeder channel (canyon); (2) the middle fan by a network of distributary channels and associated overbank deposits; and (3) the lower fan by lobes or sheet sands.

Even this subdivision will not work everywhere. For example, on the Bengal Fan, distributary channels extend from the top to the bottom of the fan.

Also under attack are some of the basic concepts on which early fan models were based, such as the idea that fan lobes develop by progradation and that they always develop coarsening- and/or thickening-upward sequences. Vertical aggradation may, in fact, be the dominant process, especially where the fan is banked against a fault-bounded basin margin. In such cases, vertical trends in grain size and bed thickness may not develop. Recent work on the Amazon fan showed that channel avulsion is the major process that determines stratigraphic architecture within this giant fan. Cyclicity, of the type predicted by the Walker model, is absent (Hiscott et al. 1997a). Anderton (1995) expressed scepticism about the processes that have been interpreted to be the cause of facies cyclicity in turbidite systems, and the reality of such cyclicity in general. He made a convincing case for the predominance of random processes in fan deposition.

Physiographic studies carried out with side-scan sonar have demonstrated many similarities between fan channels and fluvial channels (Miall 1988b). Many fan channels are meandering (e.g., Amazon fan; Damuth et al. 1983), and a few are braided (Orinoco fan; Belderson et al. 1984). Fluvial terminology, such as ridge and swale topography, levee, and crevasse, is routinely used to describe physiographic features on fans. Flood and Damuth (1987) carried this analogy to its logical conclusion in their study of the Amazon fan by calculating many of the parameters that have been used to describe fluvial channels, such as meander wavelength and radius of meander curvature and their relationships to channel width and depth. They offered some intriguing comparisons between the fan channels and fluvial channels. Similarities between the deposits of deep-sea fans and fluvial systems are striking. For example, Hein and Walker (1982) developed a model of a deep-sea braided channel from their study of the Cap Enragé Formation of Quebec, and Hein (1984) used the data from this research as a basis for a detailed comparison between coarse-grained fluvial and deep-sea facies.

Ancient fan deposits have, for many years, been assigned to fan subenvironments using the facies associations A to G defined by Mutti and Ricci-Lucchi (1972) and the interpreted relationships between these facies associations and depositional environments. However, Shanmugam and Moiola (1988; also Bouma et al. 1985) showed that the distribution of these associations is not as simple as had long been assumed. All occur in more than one subenvironment, and facies F and G can occur anywhere on a fan (Fig. 4.71). The facies descriptions and their grouping into associations have been re-examined by several authors and expanded to include considerably more detail (Stow 1985; Pickering et al. 1986). The table developed by these authors is shown in Fig. 4.72 and contains 41 discrete subfacies. A detailed facies classification of conglomerates was developed by Surlyk (1984). Available data from modern fans and ancient deposits documenting the five main subenvironments of fans – channels, scours, overbank deposits, lobes, and channel-lobe transitions – were summarized by Mutti and Normark (1987). It is clear that



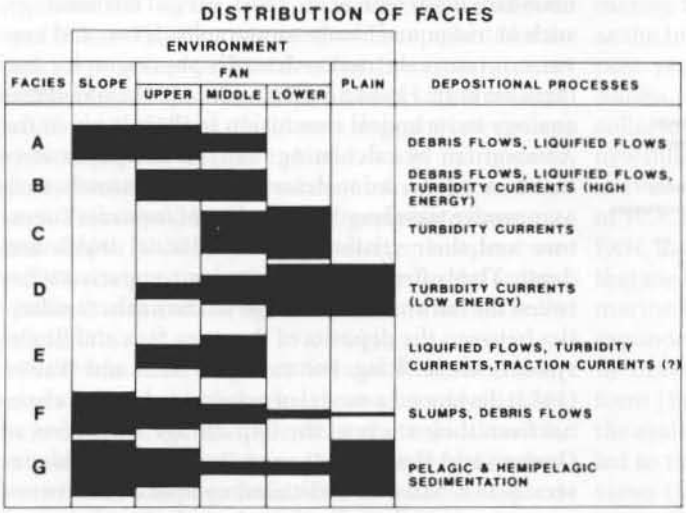


Fig. 4.71. The *upper diagram* shows the components of submarine fans, based on studies of the ancient record, and the *lower diagram* shows the distribution of the turbidite facies associations of Mutti and Ricci-Lucchi (1972), based on a synthesis of work in modern and ancient fans. (Shanmugam and Moiola 1985, 1988)

we still need to collect a great deal more stratigraphic and sedimentologic information from modern fans if we are ever to attempt a reliable updating of the Mutti–Ricci-Lucchi approach.

Most fan descriptions and models are based on the assumption of a single, canyon-fed point source for the fan deposit. For example, such models are used to explain the down-fan changes in grain size, texture, and fabric of sediment-gravity flows (Nemec et al. 1980; Porebski 1984). However, it is now realized that many turbidite systems have line sources and are, in this sense, comparable to the carbonate-apron deposits described in Section 4.6.8. Ancient examples have been described by Chan and Dott (1983) and Heller and Dickinson (1985). Multiple sources and overlapping and mixing of detritus on the continental slope may explain the variability in conglomerate facies in the ancient example described by Surlyk (1984), a deposit that does not seem to fit any preconceived ideas of a point-source fan model.

Recent syntheses of submarine-fan sedimentology emphasize the multiple controls on fan character. The primary control is that of tectonic setting (Shanmugam and Moiola 1988), which is discussed in Chapter 9. The main distinction is between the large, mud-rich fans of mature, divergent plate margins and the small, sand-rich fans of convergent margins. Classifications of fans based on shape (radial versus elongate; Stow 1985) are not very discriminating, because shape is a dependent variable reflecting the tectonic setting of the basin. An attempt to incorporate new data regarding the complexity of sediment sources into fan classifications and to discriminate fans according to grain-size variability, led Reading and Richards (1994) to propose a new classification (Fig. 4.73).

Fans tend to be most active during times of low sea level, when terrigenous detritus may be fed directly to the shelf edge (Mutti 1985; Shanmugam and Moiola 1985), but this is not always the case, and examples of

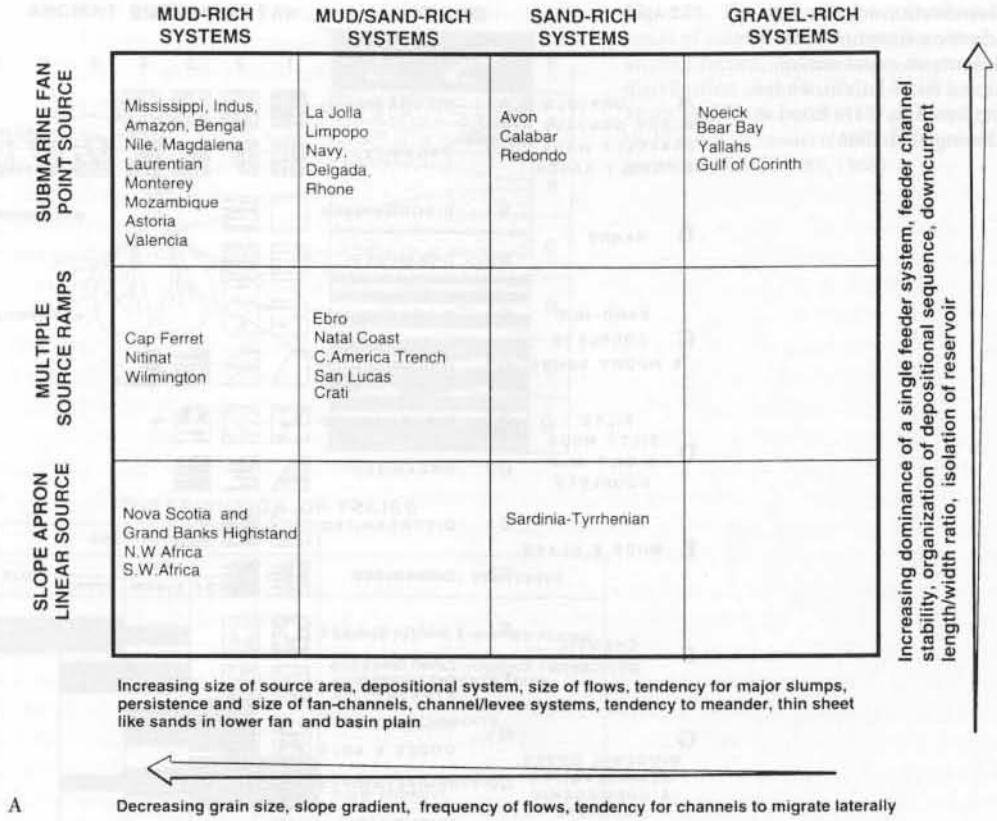


Fig. 4.72. Classification scheme for deep-water clastic sediments. This scheme represents an expansion of that of Mutti and Ricci-Lucchi, whose facies associations A to G are listed at the *left*. (Pickering et al. 1986)

active fan sedimentation during times of rising or high sea level are now well known (Chap. 6). Variations in the volume and caliber of the sediment load (partly dependent on sea level) led Mutti (1985) to propose three types of turbidite systems: type I deposits consist mainly of thick-bedded, unchannelized sandstone lobes and sheets; type II deposits consist of channelized sandstones with associated thin- to thick-bedded lobes; type-III deposits are mud rich, consisting of small-scale sandstone channel fills and thick mudstone and siltstone levee–overbank deposits. These types may succeed each other as turbidite stages in the same basin fill as a result of tectonism or sea-level change, resulting in a type of sequence stratigraphy. Mutti and Normark (1987) attempted to come to grips with the variable scales of depositional elements in turbidite systems and their many controls by subdividing the deposits into a fivefold hierarchy of architectural units. This is shown in Table 4.5 and discussed in Section 4.5.9.

Submarine fans are not the only important type of deposit in deep-marine environments, although the volume of literature devoted to them might tend to suggest that this is the case. Continental-slope aprons, contourites, basin-plain channel deposits, pelagites, and hemipelagites are, locally, of considerable importance, and a significant body of research has been carried out on them during the last decade. Hesse (1984), Stow and Piper (1984), and Stow and Faugères (1993, 1998) edited invaluable compilations of research in this area and other useful information, including recent syntheses of facies characteristics and models, is provided in the reviews by Pickering et al. (1986), Stow (1986), and Jenkyns (1986), and the study of some California borderland basins by Hein (1985). A study guide and reference source for the examination of shales and other mudrocks was compiled by Potter et al. (1980).

The principal agent of sediment transport into the deep sea is, of course, sediment-gravity flows. Classifications of flow types and the body of knowledge describing the facies characteristics of their deposits have essentially stabilized following the work of Middleton and Hampton (1976) and Lowe (1979, 1982). Recent work is focusing on a refinement of the details of internal structure of sediment-gravity-flow deposits, analysis of changes in flow support mechanisms down flow, and the development of facies models for environments and deposit types that had been neglected previously. Detailed studies of indivi-



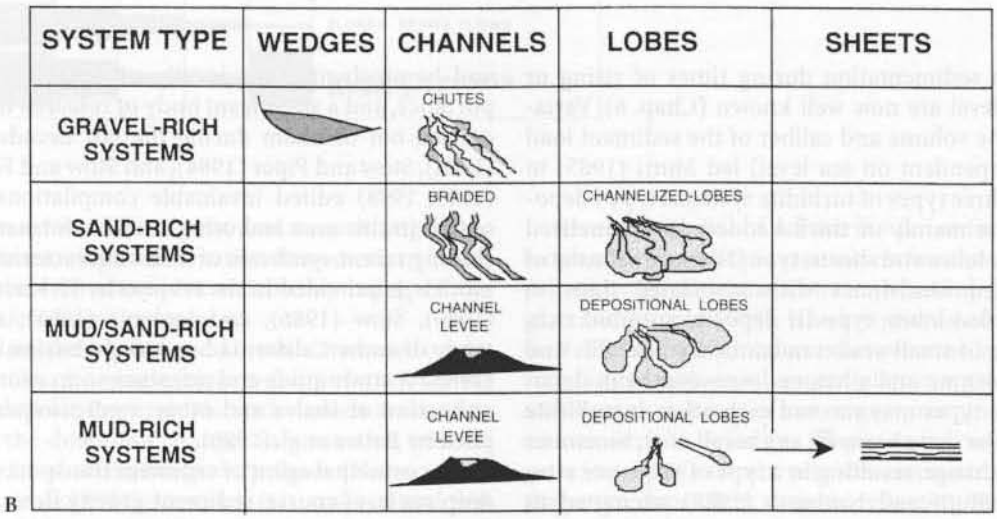


Fig. 4.73. A Classification of modern turbidite systems in deep-water basin margins, based on the volume and calibre of the sediment, and the nature of the sediment supply system. B Architectural elements of deep-water depositional systems. (Reading and Richards 1994)

dual flow deposits are now appearing. In the case of studies of Recent deposits, this type of research is facilitated by side-scan sonar studies of surface physiography, which provide a dramatic picture of the complex sequence of events that typically accompanies each flow event (Bowen et al. 1984; Prior et al. 1984; Piper et al. 1985, 1988). Hein (1985) developed a facies model for fine-grained sediment-gravity-flow deposits; this model attempted to take into account variations in flow fluidity and sedimentation rate be-

tween proximal and distal depositional sites. Kneller (1995) explored some of the expected complexities in the flow of turbidity currents and the likely sedimentary products: facies and facies successions that do not fit the classic Bouma model. The effects of variations in the flow-triggering process, topographic variability in the path of the flow, flow diversion around obstacles, and the development of internal waves within turbidity currents are among the complexities examined in this interesting paper.

A major controversy was started by Shanmugam (1996, 1997), Shanmugam and Moiola (1995), and Shanmugam et al. (1994, 1995), who questioned the interpretation of many ancient deposits as turbidites. They re-examined the facies criteria by which some classic turbidites have been defined, such as the Pennsylvanian Jackfork Group of the Ouachita Mountains and the Cenozoic fans of the North Sea Basin. They argued that internal evidence indicated laminar rather than turbulent flow in many cases and proposed that many of these deposits would be better defined as sandy debris flows. These ideas prompted a vigorous discussion that involved a careful restating of the facies evidence and its significance (Hiscott et al. 1997b; Slatt et al. 1997). Most workers in the field do not agree with the proposals by Shanmugam and his colleagues, but the detailing by Shanmugam of careful facies observations that indicate discrepancies with respect to the standard Bouma facies model have yet to be effectively answered.

Fine-grained turbidites were studied by Piper (1978), who subdivided the mudstone E division of the Bouma (1962) turbidite sequence into four facies to facilitate the recognition and description of these fine-grained rocks: facies TE1 (laminated), TE2 (graded), TE3 (ungraded), and F (pelagic mud). X-radiography is necessary to detect the necessary textural details. An ideal sequence for fine-grained turbidites, comprising the nine subdivisions T0 to T8, was proposed by Stow and Shanmugam (1980).

Fine-grained turbidites are the subject of several papers in Stow and Piper (1984). For example, Hill (1984) noted that unconfined, dilute turbidity currents can be deflected by the Coriolis effect and could flow parallel to continental-slope contours. The resulting deposits might be very difficult to distinguish from contourites deposited by traction currents. Stow et al. (1984) discussed the application of the concept of the ideal sequence of Stow and Shanmugam (1980) to the Lower Paleozoic Halifax Formation of Nova Scotia and documented various partial cycles recording proximal and distal variations in turbidite depositional processes.

Significant volumes of sediment are transported to oceanic basin plains by deep-sea channel systems, which head on the continental slope and extend for hundreds to thousands of kilometers across the virtually flat basin floors (Chough and Hesse 1976; Carter 1988; Hesse and Rakofsky 1992). The North Atlantic Mid-Ocean Channel is the longest, at 3800 km (Chough and Hesse 1976; Hesse and Rakofsky 1992; Hesse et al. 1997). The channels are eroded and filled by turbidity currents and develop meandering or braided courses. They have levees that are rendered asymmetric by the current deflections caused by the Coriolis effect. Some of these deep channels survive for tens of millions of years, their slope being maintained by the slow, thermally driven crustal subsidence that takes place as the ocean floor is carried away from spreading centers.

We are now beginning to appreciate the importance of continuous, slow circulation in the deep oceans. Fine-grained, deep-oceanic sediments are deposited by a variety of mechanisms other than simple pelagic settling. Many pelagic muds may, in fact, be mud turbidites or contourites (Gorsline 1984). X-radiography is an invaluable technique for the detection of lamination and textural detail in such deposits, and granulometric analysis is routinely employed to study vertical grain-size trends (e.g., grading). Nepheloid layers have been recorded in many locations, thermal currents, with their gyres, are present almost everywhere on the sea floor, and even some surface currents have effects on the ocean bottom, such as the Gulf Stream, eddies of which are known to reach down to the floor of the Atlantic Ocean (Gorsline 1984). Dilute turbidity currents and tidal currents are common in canyons. Water movements of all types increase in velocity where funneled through constrictions, such as canyons and narrow ocean straits. For example, thermal undercurrents, such as the Western Boundary Undercurrent of the North Atlantic, commonly flow at velocities of 30 cm/s, peaking at 1 m/s for short periods, but the warm, saline waters that flow out of the Mediterranean Sea at great depth reach velocities of 3 m/s in the Straits of Gibralter (Stow and Holbrook 1984). Such currents are responsible for transporting and depositing large quantities of sediment. For example, there are numerous sediment drifts in the North Atlantic basin, including some on the continental slopes (the classic contourites) and some on the deep-ocean floors. Some of these drifts are hundreds of kilometers long, tens of kilometers wide, and several hundred meters thick. They may be either sandy or muddy. Stow and Lovell (1979) provided a useful review of these deposits, and several recent papers are given in Stow and Piper (1984), including those by Stow and Holbrook (1984) and Gonthier et al. (1984) on the North Atlantic region. Two useful, edited collections of papers on contourites were provided by Stow and Faugères (1993, 1998).

Pelagic deposits of the deep oceans include various oozes, which lithify to limestones and claystones, and fine-grained terrigenous clastics. Nannofossils are common; many sediments are organic rich. As noted previously, silt or mud turbidites from very distant sources are commonly interbedded with true pelagic deposits. Organic-rich black shales are an important component of many deep-basin deposits and are of considerable interest because of their potential as petroleum source beds. They are described in several papers in Stow and Piper (1984). Considerable discussion continues regarding the causes of the anoxic conditions that are necessary for the preservation of organic materials (Arthur et al. 1984; Robertson 1984). Robertson (1984) summarized the main processes:

> (1) additional surface water productivity and thus extra oxygen demand to combust organic material; (2) increased rate of terrigenous organic matter input; (3) reduced bottom circulation; (4) increased overall sedimentation rate; (5) reduced solubility of oxygen in warmer waters.

Three main models have been used to explain the well-known black shales of Cretaceous age in the Atlantic Ocean basin (Robertson 1984):

1. The formation of a stratified ocean with anaerobic bottom water similar to the modern Black Sea
2. The intersection of the oxygen minimum with bottom topography (continental slope or sea mounts), with the production of organic-rich sediment in this zone
3. A high input of organic matter leading to a high oxygen demand within the sediment (ambient water may remain oxidizing)

A decreased thermal gradient between the equator and the pole at times of high sea level reduces thermohaline ocean circulation at times of high sea level, and this could lead to a reduction in general oxygen levels. Changes in climate will also affect water temperatures, organic productivity, the frequency and volume of input of terrigenous detritus and organic matter from the continents and, thus, the sedimentation rate and the rate of combustion of organic matter (Arthur et al. 1984). Grötsch et al. (1998) postulated that positive shifts in $\delta^{13}C$ values in the marine sedimentary record occur when the lighter isotope, ^{12}C, is preferentially deposited as black shale and, by using the $\delta^{13}C$ record as a chemostratigraphic correlation tool (Sect. 3.7.6), they were able to demonstrate that these shifts occurred during times of rising sea level.

Rhythmicity in fine-grained deposits is commonly on a very fine scale and may yield details regarding small-scale tectonic or climatic changes (Bourrouilh and Gorsline 1984; Robertson 1984). Thornton (1984) attributed rhythmic lamination in the sediments of the modern Santa Barbara Basin to the seasonal input of sediment from the continental margin and referred to the resulting deposits as nonglacial varves. Similar varve-like laminations were documented in the Cretaceous deposits of the western North Atlantic Ocean by Robertson (1984) and attributed to "some combination of variations of input of plant material and surface plankton productivity." However, these cycles are interpreted to be of 10^2–10^4 years duration and may result from astronomical climatic forcing affecting oceanic circulation and sediment input from the continents (Milankovitch cycles; Chap. 8).

It is useful to conclude with a remark by Gorsline (1985):

> The cutting edge of sedimentology today is much concerned with mass balances, scales, and the influence of cyclic forces on sedimentary accumulations.

4.6.10 Glacial Environments

Glacigenic deposits were among the last to be studied using facies-analysis methods. The analogue method of facies interpretation, in this case the use of Quaternary environmental and facies data, did not, at first, work well for glacigenic deposits. There are several reasons for this. A considerable body of specialized knowledge has built up from the study of Quaternary sediments, but much of this is stratigraphic in nature, rather than sedimentologic and, therefore, offers little assistance in the interpretation of the ancient record. A thorough understanding of ancient glacigenic deposits was also hindered by the tendency to describe all coarse-grained, poorly sorted, glacigenic deposits as "till". While this term means, primarily, deposition directly from grounded ice, some authorities have used a much broader definition. For example, Dreimanis (1985) stated that till "is formed and deposited by primary glacial and also by closely associated penecontemporaneous secondary non-glacial processes." It required much sedimentological investigation to document those secondary processes and to demonstrate that, in many basins, they were dominant. Most ancient glacigenic deposits formed by the dispersal and redeposition of glacigenic materials by a wide variety of processes in almost every conceivable environment, and it has taken some time for sedimentologists to sort out these various processes, to recognize their importance, and to devise criteria for their recognition in the rock record. Eyles et al. (1983) were among the first to develop systematic facies cri-

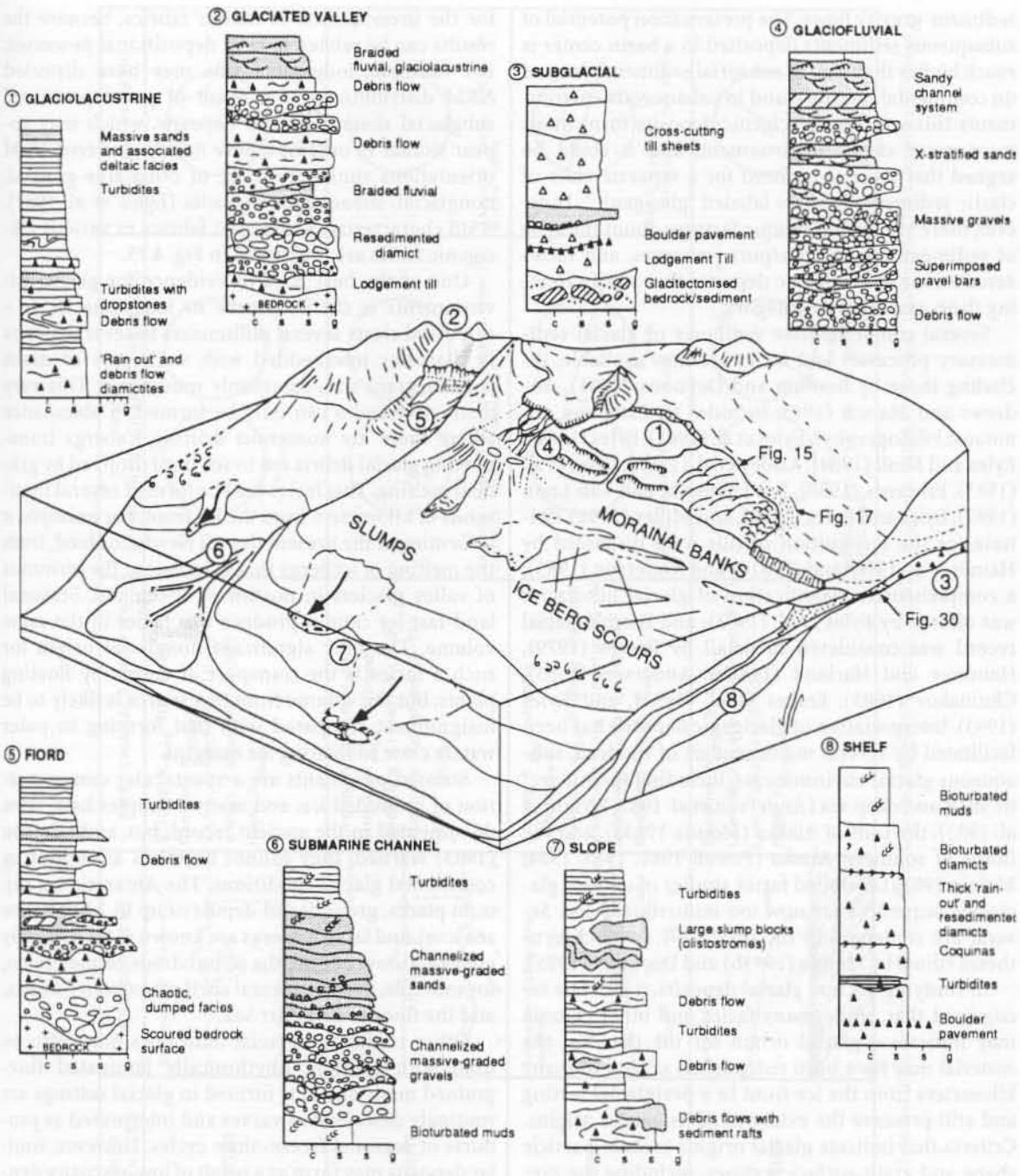


Fig. 4.74. Depositional environments and typical vertical profiles of facies deposited during a single phase of glacier advance and retreat in various glacioterrestrial and glaciomarine environments. (Eyles and Eyles 1992)

teria for this purpose (Fig. 4.74). Most authors now recommend use of the nongenetic term **diamict** (or **diamicton** or **mixton**) for the description of poorly sorted gravel–sand–silt–mud mixtures and **diamictite** (or **mixtite**) for the lithified variety. The term **till** is then restricted for use, in an interpretive sense, to mean only those deposits that can be demonstrated to have been laid down directly by the action of grounded ice and not subsequently reworked (lodgement till).

The bulk of Quaternary deposits are continental, and tills (in the restricted sense of a primary glacial deposit) are common. However, it is now becoming clear that, in the ancient record, subaqueous resedimented glacigenic deposits are much more important, including ice-rafted material and the deposits of

sediment-gravity flows. The preservation potential of subaqueous sediments deposited in a basin center is much higher than that of subaerial sediments formed on continental interiors, and in subaqueous environments tills are rare. Glacigenic deposits form in almost every clastic environment, and it could be argued that there is no need for a separate class of clastic sedimentary facies labeled "glacigenic". However, there are several unique features about the type of sediment supply, transport processes, and facies associations of glacigenic deposits that justify treating them as a separate category.

Several comprehensive syntheses of glacial sedimentary processes and facies are now available, including those by Boulton and Deynoux (1981), Andrews and Matsch (1983; includes an extensive annotated bibliography), Eyles et al. (1983), Eyles (1984), Eyles and Miall (1984), Ashley et al. (1985), Eyles et al. (1985), Edwards (1986), Brodzikowski and Van Loon (1987), Eyles and Eyles (1992), and Miller (1996). Criteria for the recognition of tills were discussed by Hambrey and Harland (1981a) and Anderson (1983); a comprehensive classification of glacial lithofacies was offered by Eyles et al. (1983); and Earth's glacial record was considered in detail by Frakes (1979), Hambrey and Harland (1981b), Anderson (1983), Chumakov (1985), Frakes et al. (1992), and Eyles (1993). Interpretation of glacigenic deposits has been facilitated by several major studies of modern, subaqueous glacial environments, including the Antarctic shelf and deep sea (Anderson et al. 1983; Wright et al. 1983), the Gulf of Alaska (Molnia 1983a), and the fiords of southern Alaska (Powell 1981, 1983, 1984; Molnia 1983a). Detailed facies studies of ancient glacigenic sequences are now too numerous to list. Several are contained in the invaluable research syntheses edited by Molnia (1983b) and Deynoux (1985).

In studying ancient glacial deposits, it must be recognized that, while many facies and other criteria may indicate a glacial origin for the detritus, the material may have been redeposited several to many kilometers from the ice front in a **periglacial** setting and still preserve the evidence of its glacial origins. Criteria that indicate glacial origins include particle shape and grain-surface textures, including the presence of striae on clasts. True tills (in the restricted sense of the word) may be distinguished from other diamicts, including ice-rafted deposits and sediment-gravity flows, by such sedimentologic and stratigraphic criteria as internal grain-size sorting (e.g., grading) and sedimentary structures, clast and matrix fabrics, bed thickness, geometry and lateral extent, and facies associations (Fig. 4.74). Magnetic characteristics, including the anisotropy of magnetic susceptibility (AMS) and the orientation of natural remanent magnetism (NRM), have become useful tools for the investigation of matrix fabrics, because the results can be related to final depositional processes. For example, lodgement tills may have distorted NRM distributions as a result of postdepositional subglacial shear. Rain-out deposits, which may appear similar in outcrops, have tightly clustered NRM orientations similar to those of other fine-grained, nonglacial, subaqueous deposits (Eyles et al. 1987). AMS characteristics and clast fabrics in various glacigenic facies are illustrated in Fig. 4.75.

One of the best lines of evidence for glacial environments is the dropstone or lonestone facies – dispersed clasts several millimeters to several meters in diameter interbedded with sediments of much smaller grain size, commonly mudstones. This very distinctive facies can only be formed in abundance where there are numerous drifting icebergs transporting glacial debris out to sea until dropped by gradual melting. This facies may be formed several thousands of kilometers from the ice front. For example, it is forming at the present day off Newfoundland, from the melting of icebergs that originate at the terminus of valley glaciers in northwest Greenland. Seasonal land-fast ice cannot produce this facies in the same volume. The only significant nonglacial origin for such a facies is the transport of debris by floating plants, but the volume from this source is likely to be insignificant compared with that forming in polar waters close to floating ice margins.

Striated pavements are a spectacular demonstration of grounded ice, and many examples have been documented in the ancient record; but, as Anderson (1983) warned, they cannot be taken as indicating continental glacial conditions. The Antarctic ice cap is, in places, grounded at depths of up to 2 km below sea level, and large icebergs are known to form deeply gouged grooves at depths of hundreds of meters on, for example, the continental shelf off eastern Canada, and the floor of Beaufort Sea.

Other traditional glacial indicators must also be used with caution. Rhythmically laminated fine-grained muds and silts formed in glacial settings are routinely described as varves and interpreted as products of seasonal freeze–thaw cycles. However, similar deposits may form as a result of low-viscosity density interflows and underflows and from slump-generated surge currents, and the number of rhythms does not necessarily bear any relationship to the number of years during which the deposit accumulated. Ashley et al. (1985) described the variations between these different types of rhythm in some detail, with many excellent photographs. Mackiewicz et al. (1984) coined the term **cyclopel** for varve-like laminated interflow and underflow deposits.

Sand-wedge structures may indicate subaerial permafrost conditions, but Eyles and Clark (1985) de-

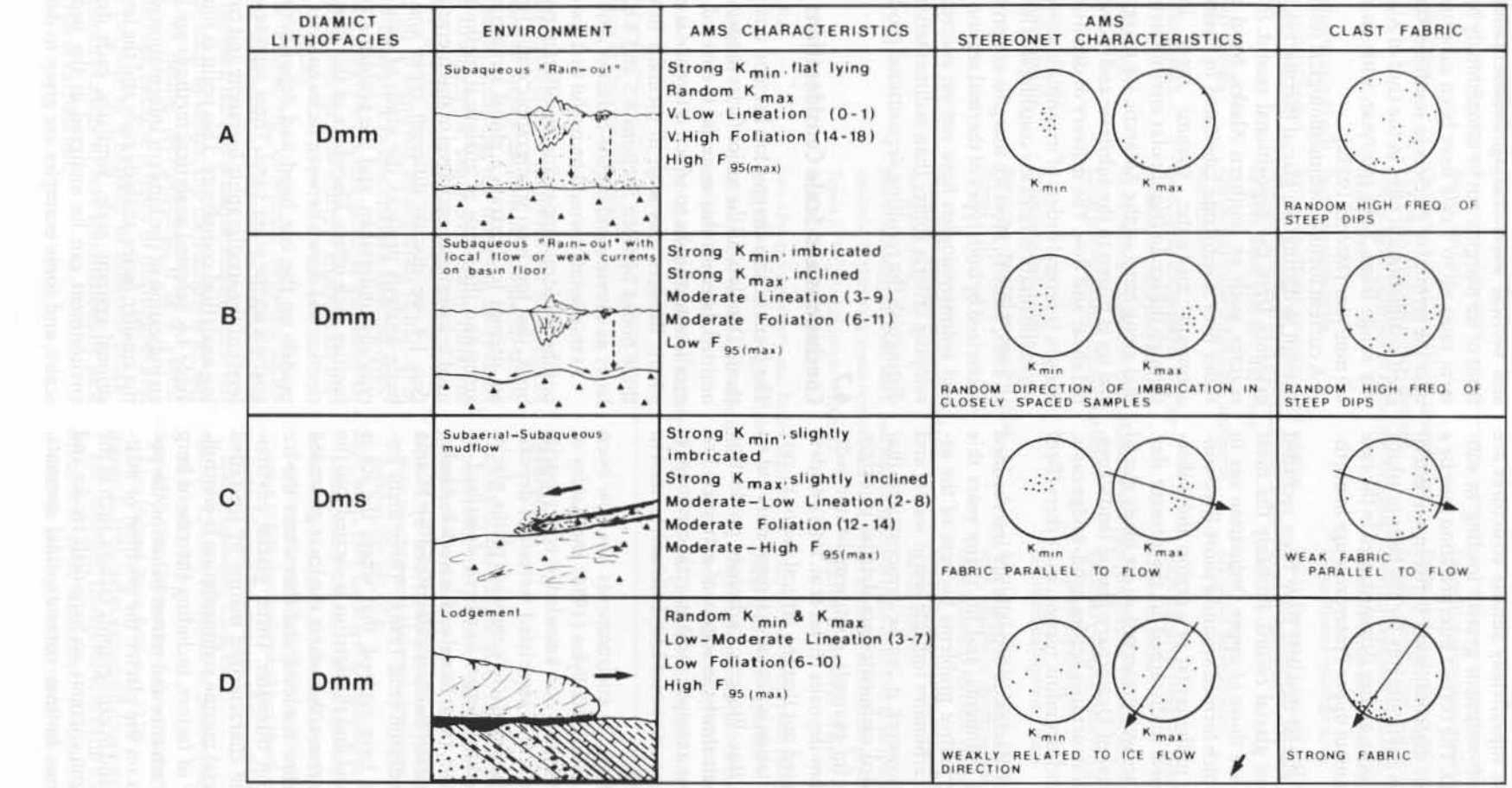


Fig. 4.75 A–D. Selected diamict lithofacies and their depositional environments, showing magnetic susceptibility and clast fabric characteristics. (Eyles et al. 1987)

monstrated that superficially similar structures are produced by soft-sediment gravity loading in subaqueous settings. The correct interpretation may be a critical factor in discriminating between terrestrial and subaqueous environments for the beds in which the structures occur, as they demonstrated in the case of a glacigenic unit of upper Proterozoic age in northwest Scotland.

Continental glacial deposits have been recorded from the ancient glacial record. Probably the most well preserved are those of upper Ordovician age in North Africa, which formed around a short-lived continental ice cap in the center of a craton and, subsequently, remained undisturbed to the present day. Terrestrial tills are well preserved, as are many glacial landforms typical of Quaternary glacial landscapes, such as ice-push structures, kettles, sand-wedge casts, polygonal ground, drumlins, pingos, and eskers (Beuf et al. 1971).

Some ancient glacigenic deposits are interbedded with carbonate sediments, and for many years this has raised interpretive problems because of the assumption that carbonates indicate warm waters and low latitudes. However, it can be demonstrated that some interbedded carbonates are detrital in origin and formed as, for example, continental-slope sediment-gravity-flow deposits (Eyles et al. 1985). Anderson (1983) pointed out that the critical factor in carbonate sedimentation is not water temperature but an absence of detritus. Biogenic carbonate production takes place, albeit slowly, in waters of extremely low temperature. For example, the Antarctic continent is ringed with carbonate oozes and shell hash formed in waters of 0–8 °C.

A variety of glacial environments has now been identified and described. Eyles (1984) and Ashley et al. (1985) synthesized our knowledge of terrestrial glacial landforms and associated terrestrial depositional settings (valley glaciers, outwash plains, glacial lakes). Fiord sedimentation has been described by Powell (1981, 1983, 1984) based on his work in Alaska. Shelf sedimentation has been described by Molnia (1983a) and Anderson et al. (1983). Three main ice-margin models have evolved: that where the ice is grounded on land and the detritus is carried into the sea by outwash rivers, that where the ice is grounded on the shelf below sea level, and that where the ice margin floats on tidewater. Distal glacial environments are those characterized mainly by ice-rafted sediments. Glacial marine sedimentation is controlled by a variety of factors, including the rate of berg calving, glacier advance and retreat behavior, the position of debris on the glacier, the position of meltwater streams, and oceanographic factors, such as the control of bottom currents on berg-drift tracks and sediment plumes, bottom currents, tidal currents, and vertical water mixing. Sedimentation rates in front of ice margins can be astonishingly high. Short-term rates up to 9 m/a have been measured in bays near the termini of glaciers in southern Alaska, while shelf sedimentation rates in the Gulf of Alaska typically range from 2–10 mm/year, with an average of 4.5 mm/year (Molnia 1983a).

A current focus of sedimentological research is the attempt to distinguish glacial thermal regimes using evidence from the depositional record. In subarctic regions, such as southern Alaska, basal glacier ice slides over a substrate lubricated by meltwater, and abundant meltwater streams move in tunnels through the ice. In frigid polar environments, such as that along most of the Antarctic ice margin, the glacial ice is frozen to the substrate and moves by internal shear and flow. The delivery of detritus to the ice margin is several orders of magnitude slower, and basin sedimentation rates are considerably less (Eyles et al. 1985, Table I). Most ice margins are probably characterized by both types of thermal regime over time, and sedimentologists have not yet succeeded in developing criteria (other than sedimentation rates) to distinguish the resulting depositional products.

4.7 Conclusions and Scale Considerations

The focus of this chapter has been on analytical methods. Details of all the major facies models have been omitted, because this material is covered in several excellent textbooks to which reference is made elsewhere. Readers may wish to mentally insert one of these books between Sections 4.5 and 4.6, using the latter as a review and update of recent developments.

At this point, it would be useful to draw together a number of ideas, appearing in different parts of this book, that touch on stratigraphic scale. Figure 1.1 is an attempt to illustrate a global stratigraphic hierarchy and the kinds of analytical methods that are used to investigate each level of this hierarchy. In Section 1.4, we discuss different types and scales of basin-analysis projects, the kinds of data collection typically undertaken, and the problems and opportunities each offers. Elsewhere in this chapter, a distinction was drawn between facies analysis and facies models on the one hand and depositional-systems analysis on the other hand. These represent different levels of the stratigraphic hierarchy and corresponding analytical complexity. This point is illustrated in Table 4.6. Selected analytical methods are listed with an indication of the kinds of information obtained at the smaller, facies-analysis scale and the larger, depositional-systems scale. Similarly, each depositional environment can be analyzed at the two different scales, and some examples are given to demonstrate

Table 4.6. Scales of sedimentological analysis

1. Scales	Facies analysis	Depositional-systems analysis
2. General definitions	Outcrop or local-scale facies assemblages and environmental interpretations, within-basin (autogenic) cyclic mechanisms, facies models	Basin-wide paleogeography and stratigraphic architecture, allogenic processes; contemporaneous tectonics, sea-level change, climate change, sequence stratigraphy
3. Selected analytical methods	Facies-analysis scale	Depositional-systems scale
Sedimentary structures and paleocurrents	Bar and bedform geometry and relative arrangement, local hydraulics, identification of river, wave, and tide influence	Basin dispersal patterns and relationship to tectonic elements
Vertical-profile studies	Environmentally diagnostic, autogenic mechanisms	Allogenic mechanisms
Petrology (siliciclastic)	Authigenic components may be environmentally diagnostic	Basin dispersal patterns, plate-tectonic setting of source rocks
Petrology (carbonate)	Basis of most lithofacies identifications and paleogeographic reconstructions at all scales	
Paleoecology	Local environmental interpretations	Regional paleobathymetry, gross salinity and temperature changes
4. Selected environments	Facies-analysis scale	Depositional-systems scale
Fluvial	Bar type, channel geometry, local fluvial style	Areal and stratigraphic variations of fluvial style, relation of dispersal patterns to contemporaneous tectonic elements
Eolian	Dune type and geometry, interdune environments	Erg paleogeography, relation to playa and fluvial systems
Deltaic	Discrimination of river, wave, and tide influence is main objective at all levels of analysis, as these are main controls on all scales of cyclicity and fancies geometry	
Shelf	Types of HCS [a] and storm sequence, tidal sand-wave evolution	Relative dominance of tides versus storms
Continental rise	Fan-channel geometry and facies, channel migration and lobe progradation	Canyon-fan–basin-plain relations, paleogeography of fan complexes
Carbonate platform	Local environments, tidal exposure, reef paleoecology	Discrimination of windward and leeward shelf margins, reef distribution
Evaporite basins	Recognition of deep- or shallow-water origin	Large-scale cyclicity and relation to water-level fluctuations

[a] HCS, hummocky cross-stratification.

this idea. The table is not exhaustive but is offered as an illustration of these scale considerations. To tackle the larger scale, including the mapping of systems tracts and stratigraphic sequences, many different basin-mapping techniques are required, which is why the discussion of sequence stratigraphy (Chap. 6) is preceded by Chapter 5. In this chapter, also, the question of sedimentary-structure scale is returned to, and its implications for paleocurrent analyses are discussed (Sect. 5.6.4).

To some extent, the difference between facies models and depositional systems is artificial; the boundary between the two is vague, and for some environments, such as deltas, the distinction is all but impossible to define. Nevertheless, it is a useful approach to take, because it helps to distinguish and clarify the purpose of a number of different procedures we perform more or less simultaneously as a basin analysis is carried out.

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Basin-Mapping Methods

5.1 Introduction

Facies analysis, as described in Chapter 4, begins with the art of interpreting single outcrops or, at most, small areas, such as a well-exposed cliff section. Paleogeographic analysis of an entire sedimentary basin requires the documentation of many such outcrops (or well sections), their correlation by stratigraphic methods (Chap. 3), and the use of various sedimentological mapping techniques. The latter are the subject of this chapter.

Stratigraphic units undergo thickness and facies changes across a basin, reflecting contemporaneous paleogeography, subsidence patterns, and location of sediment sources. Most of these aspects of basin development are controlled by tectonics, the development of bounding faulted or folded uplifts, and subsidence caused by thermal contraction of continental margins, basement failure, or thrust-sheet loading (Chaps. 7, 9). Eustatic sea-level changes are also of profound paleogeographic importance (Chap. 8). Such effects leave their imprint on sediments via their control on paleoslope by the structuring of land masses that guide oceanic currents, wave advance, or air flow and by the development of depocenters and their control on water depths or rates of internal scour.

The art of basin mapping is one of reconstructing paleogeography and fill geometry from very limited evidence. Being skilled at data synthesis, interpretation, and extrapolation is of prime importance. Maps drawn only from outcrop data are rarely sufficiently accurate or useful for subsurface prediction, because they do not contain an adequate, three-dimensional distribution of data points.

Two broad classes of data are used most frequently in basin mapping. The first class deals with facies thickness and ratio data and petrographic information. Such data reflect environment and paleogeography but suggest no pattern until they have been measured at many control points and the results contoured. This approach is discussed in Sections 5.2, 5.3, and 5.5. Geophysical methods are discussed in Section 5.4, and computer methods for manipulating these data are discussed in Section 5.8. The second class of data refers to oriented objects or structures within the rocks. Each of these has directional (paleogeographic) significance, although statistical methods are required to extract local or regional trends from the data sets. The study of this class of data is what is normally implied by the term **paleocurrent analysis.** It is discussed in Section 5.6. Remote-sensing techniques are now being used in experimental and reconnaissance studies of sedimentary basins and their economic deposits. They are discussed briefly in Section 5.7. Methods of regional synthesis and data presentation are discussed towards the end of the chapter. Computer mapping methods and map reliability are discussed in Section 5.8. The use of stratigraphic cross sections is illustrated in Section 5.10. Examples of the synthesis of these various types of information are described in Section 5.11. Many of these mapping methods may be used simultaneously, and the order of presentation in this chapter is not intended to suggest a rigid sequence of mapping procedures. For example, it is commonly advisable to start a detailed stratigraphic analysis by constructing cross sections in order to clarify correlations. Many other specialized techniques are discussed by Potter and Pettijohn (1977) and North (1985) and in numerous research publications, but those dealt with here are the most widely used in regional basin studies for the purpose of petroleum, coal, and ore prospecting and evaluation.

5.2 Structure and Isopach Contouring

Structure-contour and isopach maps are as essential as a stratigraphic correlation table for interpreting basin history. They may or may not contain much direct paleogeographic information, but they are required in order to illustrate basin shape and orientation and basin-fill geometry.

Structure contours are drawn to illustrate the attitude of selected stratigraphic horizons within the basin. They are always drawn relative to sea level. Control points are derived from whatever surface and subsurface information is available. Depth data in petroleum-exploration wells are converted to depth

subsea by subtracting the elevation of the kelly bushing (K.B.) of the drill rig. It may be useful to illustrate fill geometry by drawing several structure contour maps at carefully selected horizons. Commonly, these are chosen at major unconformities that demarcate discrete stratigraphic packages within the basin fill. Most important of all is that drawn on the basement unconformity. The term basement is a useful one in basin studies, but it should be used with caution, as it may mean different things to different people. Normally, the term refers to the base of the structurally conformable sequence that the geologist is engaged in mapping. This is usually an unconformity, and it may be a profound one with, for example, Mesozoic or Cenozoic rocks resting on Precambrian. Elsewhere, the basement surface may be within the Phanerozoic, for example, the basement of the Sverdrup Basin in the Canadian Arctic is of Cambrian to Devonian age. The basin fill itself is Mississippian to Tertiary and rests on a profound unconformity developed during a major middle to late Devonian orogeny. Petroleum geologists commonly refer to **economic basement**, which means the depth of the section beyond which all hope for oil and gas has been abandoned. It is drawn at an unconformity across which major diagenetic, structural, or metamorphic changes are apparent.

Structure contours are, of course, of prime importance in the exploration for structural petroleum traps. In the initial stages of exploration, they may be drawn entirely from seismic data, using depths of prominent reflectors plotted as two-way travel time (the time taken for the sound wave to travel down to the reflector and back; Sect. 5.4.1). As velocity information from wells becomes available, these times are converted to depths and the maps redrawn.

On a broad scale, structure contours within the basin (particularly at deeper levels) reveal the locations of sub-basins, depocenters, and axes of uplift. They may outline subtle syndepositional topography that had important effects on local paleogeography and facies patterns. Regional trends may reveal much about crustal characteristics and subsidence mechanisms (Chap. 7). Detailed local maps may reveal drape structures over channel fills, reefs, or drainage nets eroded into unconformity surfaces. Detailed well control may yield this information directly, or it may be reconstructed and interpreted from magnetic, gravity, or seismic data. Several examples of structure maps showing some of these features are illustrated in Section 5.4. A particularly well-documented example of a subsurface map showing a widespread erosion surface was provided by Leggitt et al. (1990) based on the extensive subsurface data available in the Alberta Basin.

In some areas, it may be useful to emphasize local structure (fault blocks, drapes, etc.) by subtracting regional trends from the local data. Computer methods are particularly useful for this purpose, as discussed in Section 5.8.

Paleogeological maps or **subcrop maps** are maps showing the units outcropping at an unconformity surface. They can only be reconstructed from extensive outcrop or subsurface data. Such maps reveal the extent of tectonic deformation and erosion that occurred before deposition of the post-unconformity strata. They may show buried folds or faults and are, therefore, useful for reconstructing ancient stress patterns. Busch (1974) illustrated examples where detailed well control permitted the construction of paleogeological maps that showed drainage patterns, just as do geological maps of present-day land surfaces underlain by gently dipping strata and cut by a network of stream valleys. **Worm's-eye-view maps** show the distribution of units that onlap an unconformity. Their main use is in illustrating the pattern of basin fill, shifting shorelines, or gradual burial of a pre-existing erosional topography.

An isopach is a line of equal thickness. Isopach maps should be drawn for an entire basin fill and for selected stratigraphic intervals of interest in order to illustrate basin-fill geometry. In areas of structural complexity, it may be impossible to draw meaningful

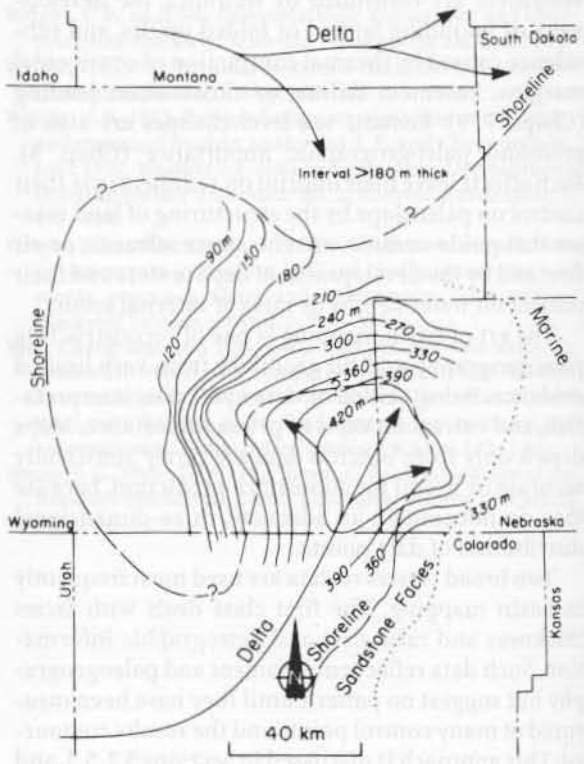


Fig. 5.1. An isopach map outlining an upper Cretaceous deltaic depocenter, Rocky Mountains area, Wyoming–Colorado. (Weimer 1970)

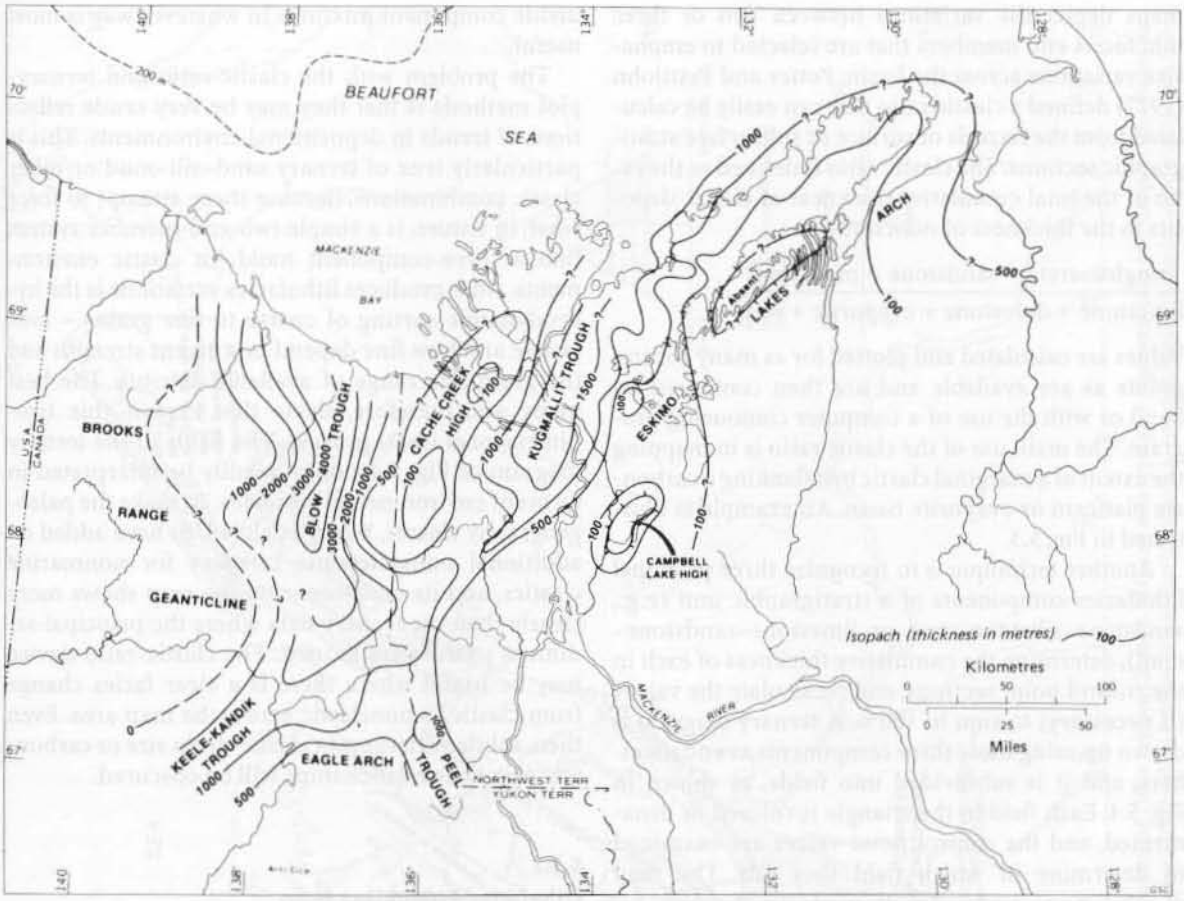


Fig. 5.2. Isopach map of Barremian–Albian strata, Mackenzie Delta region, Canada, which delineates depocenters and local structural highs. (Young et al. 1976)

or sedimentologically useful structure-contour maps, whereas isopach maps reveal the basin fill in its original, undeformed form and reveal something about contemporaneous structure. Isopachs are particularly useful for illustrating the location and outline of major depocenters, such as large deltas (Fig. 5.1), and for delineating syndepositional intrabasinal upwarps, or schwelle, over which the section is thinned (Fig. 5.2). Isopachs are usually constructed only for units bounded by conformable contacts. If they are drawn for unconformity-bounded units, interpretations must take into account the effects of postdepositional structural deformation and erosion. Isopachs will also be distorted if the unit to be plotted cannot be precisely correlated.

5.3 Lithofacies Maps

A variety of methods has been devised to map lithofacies variations. Most have been in use for many years, but the techniques for constructing and interpreting them have become more refined as more has been learned about depositional environments.

When constructing facies maps, the geologist should always work with as thin a stratigraphic unit as is feasibly possible. The objective of the exercise is to reconstruct paleogeography, but in every basin this was constantly changing. Delta and submarine-fan distributaries switched, shorelines transgressed or regressed, carbonate platforms prograded or were deeply scoured by hurricanes, and so on. By working with a thin unit, we can hope to reconstruct the paleogeography at some more or less frozen instant in time. If the interval is too thick, all the subtleties will be lost in a sea of generalizations. This is why the refinements of stratigraphic correlation discussed in Chapter 3 are so important.

5.3.1 Multicomponent Maps

Lithofacies maps may be plotted either as ratios or as isopachs of selected lithofacies components. Ratio

maps depict the variations between two or three lithofacies end members that are selected to emphasize variations across the basin. Potter and Pettijohn (1977) defined a clastic ratio that can easily be calculated from the records of surface or subsurface stratigraphic sections. The clastic ratio is defined as the ratio of the total cumulative thickness of clastic deposits to the thickness of nonclastics:

$$\frac{\text{conglomerate} + \text{sandstone} + \text{mudstone}}{\text{limestone} + \text{dolostone} + \text{evaporite} + \text{coal}}$$

Values are calculated and plotted for as many control points as are available and are then contoured by hand or with the use of a computer contouring program. The main use of the clastic ratio is in mapping the extent of a marginal clastic belt flanking a carbonate platform or evaporite basin. An example is illustrated in Fig. 5.3.

Another technique is to recognize three principal lithofacies components of a stratigraphic unit (e.g., sandstone–siltstone–mud or limestone–sandstone–mud), determine the cumulative thickness of each in the control point sections, and recalculate the values (if necessary) to sum to 100%. A ternary diagram is drawn up using these three components as end members, and it is subdivided into fields, as shown in Fig. 5.4. Each field in the triangle is colored or ornamented, and the control point values are examined to determine in which field they fall. The map can then be plotted and contoured with the same colors or patterns. The subdivision of the triangle into fields is arbitrary and may be chosen to group or subdivide component mixtures in whatever way is most useful.

The problem with the clastic-ratio and ternary-plot methods is that they may be very crude reflections of trends in depositional environments. This is particularly true of ternary sand–silt–mud or other clastic combinations, because these attempt to force what, in nature, is a simple two-end-member system into a three-component mold. In clastic environments, what produces lithofacies variations is the hydrodynamic sorting of coarse to fine grains – how coarse and how fine depend on current strength and the grain-size range of available detritus. The best maps are, therefore, those that exploit this two-dimensional (2-D) process. The fields in the ternary diagram of Fig. 5.4 cannot readily be interpreted in terms of environmental variables. To make the paleogeography clearer, Young et al. (1976) have added an additional nonsystematic category for nonmarine clastics, and its occurrence on the map shows more clearly than the ternary data where the principal sediment sources are located. The clastic-ratio device may be useful where there is a clear facies change from clastic to nonclastic across the map area. Even then, subtle variations in clastic-grain size or carbonate–evaporite relationships will be obscured.

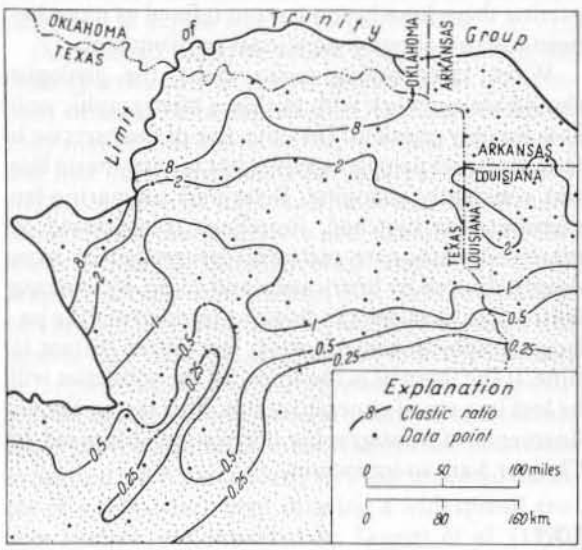


Fig. 5.3. A clastic-ratio map for part of the Cretaceous section of the Gulf Coast region. Note higher concentration of clastics near northern edge of basin, which is the direction of the sediment source. (Potter and Pettijohn 1977)

5.3.2 Lithofacies-Assemblage Maps

There are two solutions to the problems outlined above that provide better lithofacies mapping. They represent opposite extremes of complexity, but both are extremely useful in illustrating the full range of paleogeographic variability in a basin. The first method is to plot the distribution of lithofacies assemblages. The assemblages may be defined in any of the ways discussed in Chapter 4, using descriptive criteria or statistical methods. Once the assemblages are defined, however arbitrary or mathematical the process, each control point is assigned to one of the assemblages, and the results are plotted and contoured. The classic Bahama Platform lithofacies map referred to in Chapter 4 (Fig. 4.9) is a good example of this type of map. Another example, from an ancient shelf, is illustrated in Fig. 5.5. (from Murris 1980) and shows a northeastward gradation across the Arabian craton from a marginal, coastal-plain clastic belt to shelf- and deeper-water-carbonate areas. Murris has defined transitional zones between his lithofacies assemblages, such as the mixed clastic–carbonate shelf. If this creates a more meaningful map, there is nothing wrong with the approach, but it may require arbitrary decisions about cutoffs between the lithofacies assemblage classes.

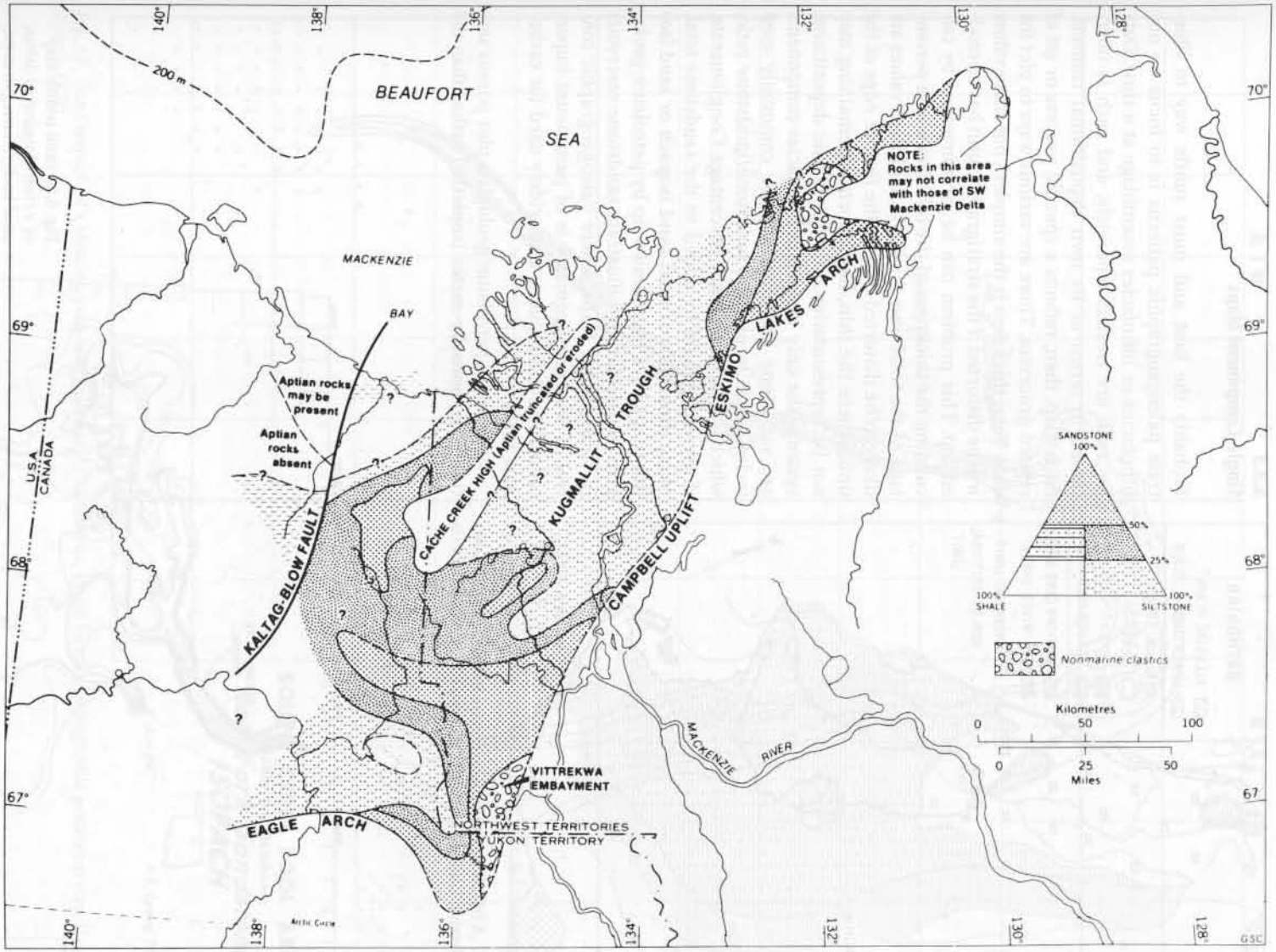


Fig. 5.4. A ternary lithofacies plot of early Aptian sediments, Mackenzie Delta region, Canada. Maps of this type give a general idea of transport directions and depositional energy but, in this case, the clearest information is given by the distribution of nonmarine clastics, an additional lithofacies subdivision that is not part of the ternary plot. (Young et al. 1976)

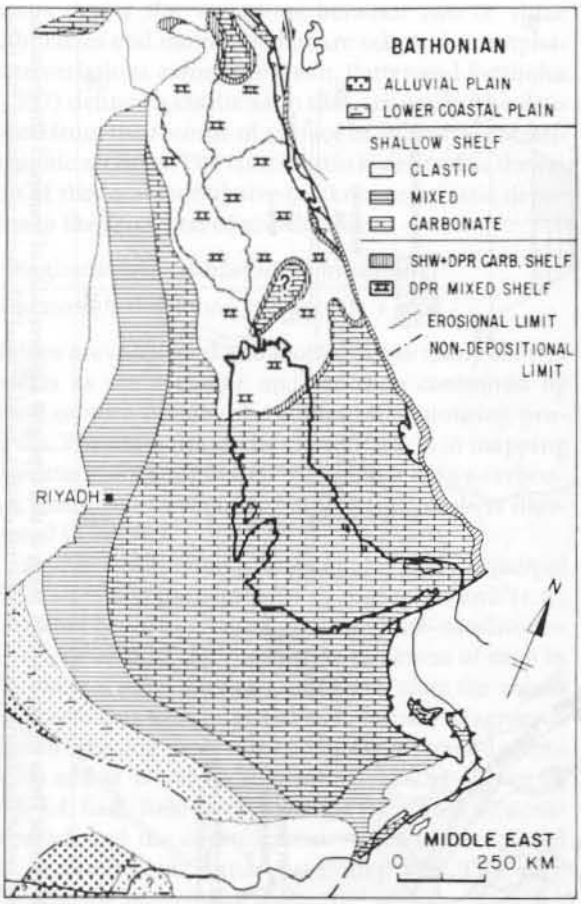


Fig. 5.5. A lithofacies assemblage map; Bathonian sediments of the Persian Gulf region. *SHR* Shallow; *DPR* deeper. (Murris 1980)

5.3.3 Single-Component Maps

Probably the best and most subtle way to illustrate paleogeographic patterns is to focus on one lithofacies or lithofacies assemblage at a time. Data for each are treated separately, and each is interpreted in terms of its own depositional controls. Each map, then, reflects a specific process or set of related processes. There are various ways to plot the data. Total thickness is the simplest, but these values will be distorted if the stratigraphic unit has an eroded top. This problem can be circumvented by calculating the thickness of the lithofacies as a percentage of the total unit thickness, and then values are likely to be distorted only at the feather edge of the unit, where the thin, basal portion remaining may not be representative. In many clastic depositional systems, the only important lithofacies components are sandstone and mudstone. A commonly used technique is to plot the sandstone/mudstone ratio, which is a variation of the percentage. Conglomerate, if present, should be added to the sandstone total. Many thousands of **total-sand isopach** or **sand isolith** maps have been drawn up by petroleum geologists, particularly to illustrate sandstone reservoirs in stratigraphic traps, where paleogeographic control of reservoir geometry is of paramount importance. The technique is also widely used for carbonate rocks.

Commonly, petroleum geologists plot porous section (e.g., **porosity-meter isopleths**) rather than the

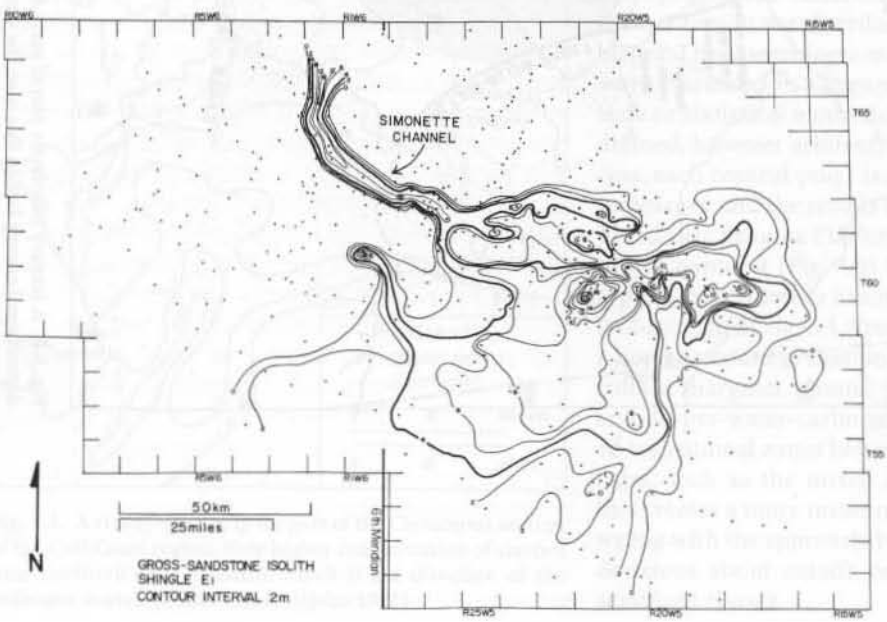


Fig. 5.6. Sand isolith map of a river-dominated deltaic sand body, Dunvegan delta, Alberta. Two shoestring sand bodies are identified; the larger one is named the Simonette Channel (Bhattacharya 1991). See Fig. 6.30 for a paleogeographic interpretation of this unit

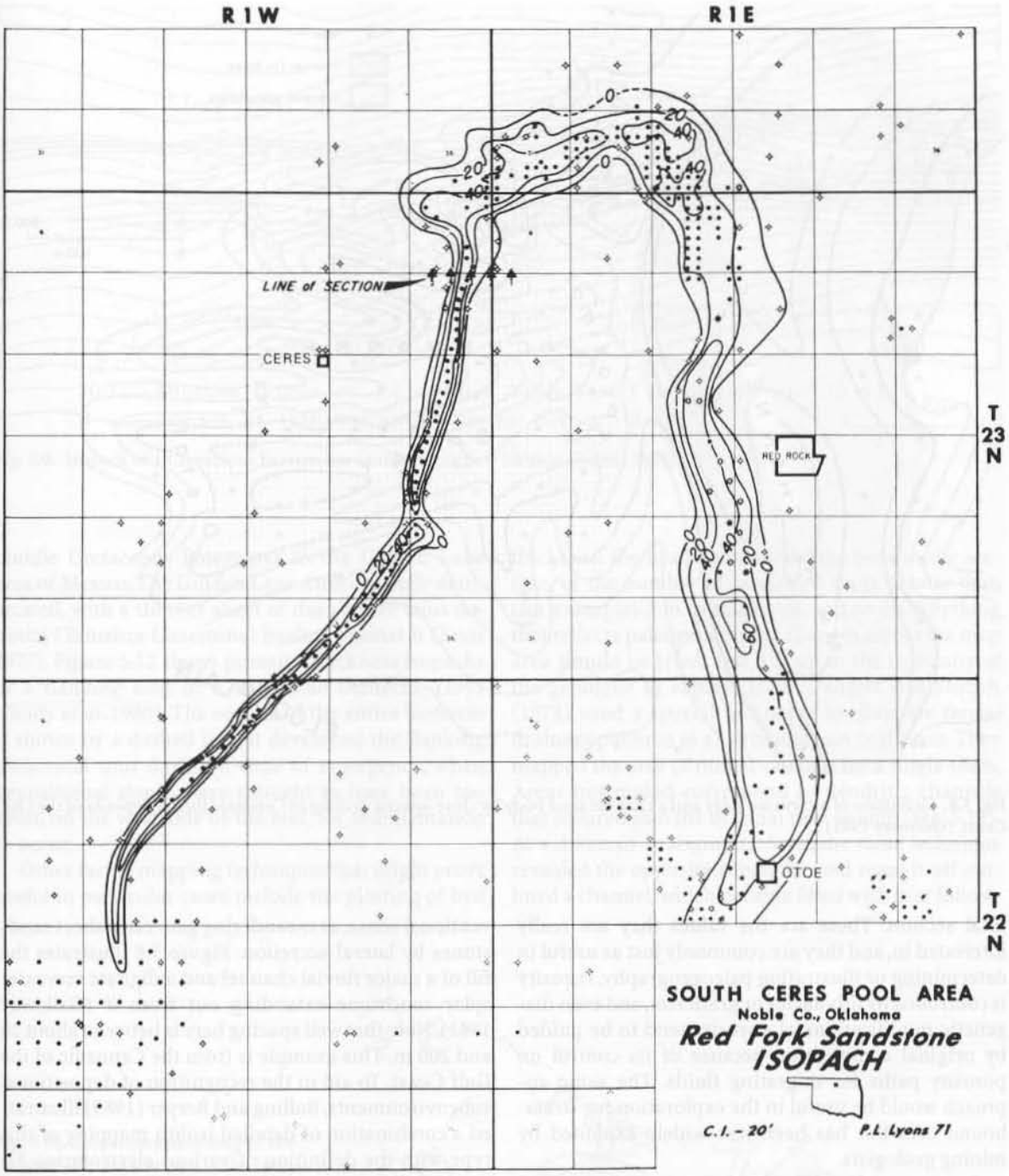


Fig. 5.7. Sand isopach of a Mississippian valley-fill sand in Oklahoma, which forms a stratigraphic petroleum trap. (Lyons and Dobrin 1972)

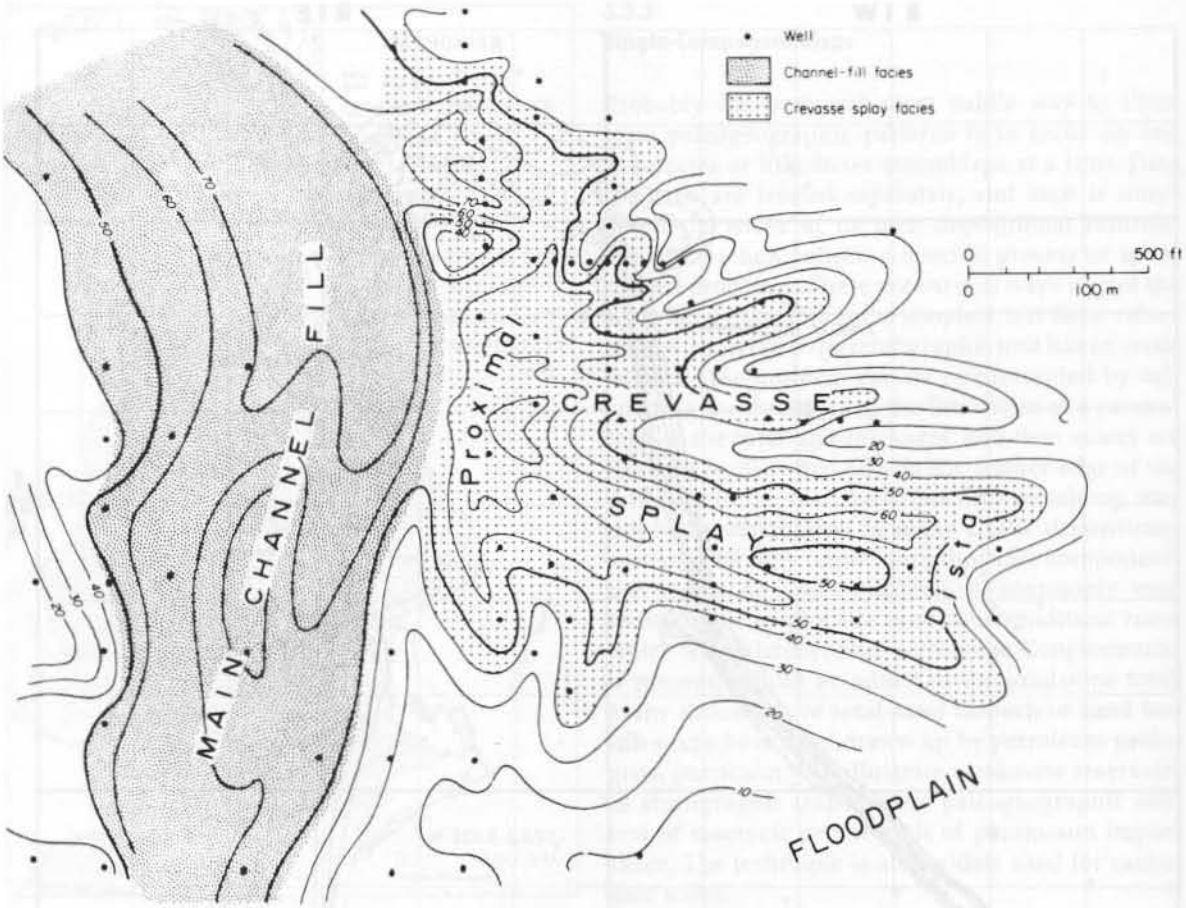


Fig. 5.8. Definition of a crevasse splay and a channel sand body by close-spaced drilling in Cenozoic fluvial deposits of the Gulf Coast. (Galloway 1981)

total section. These are the values they are really interested in, and they are commonly just as useful in determining or illustrating paleogeography. Porosity is controlled by lithofacies or grain size, and even diagenetic modifications of porosity tend to be guided by original composition because of its control on porosity paths for migrating fluids. The same approach would be useful in the exploration for stratabound ores but has been less widely exploited by mining geologists.

Some examples of these one-component maps are given in Figs. 5.6 – 5.12. Figure 5.6 illustrates the use of sand-thickness data in mapping delta systems on a regional scale (Bhattacharya and Walker 1992). Two ancient, river-dominated deltas are shown, with the isoliths defining the location and configuration of "shoestring sands". Figure 5.7 shows total-sand isopachs that define a valley-fill sandstone of Mississippian age (Lyons and Dobrin 1972). The large scale and the U-shaped plan geometry of the sandstone body show that it is not a meandering channel in the conventional sense, as meandering generates sheet sandstones by lateral accretion. Figure 5.8 illustrates the fill of a major fluvial channel and a digitate crevasse-splay sandstone extending out from it (Galloway 1981). Note that well spacing here is between about 20 and 200 m. This example is from the Cenozoic of the Gulf Coast. To aid in the recognition of depositional subenvironments, Bulling and Breyer (1989) illustrated a combination of detailed isolith mapping of this type with the definition of various electrofacies. Figure 5.9 illustrates a different type of environment, a linear, barrier-island sand trend in a Cretaceous unit (San Juan Basin, New Mexico; Sabins 1972). The barrier sand has been separated from related facies for plotting purposes. Superimposition of sand isopachs and structural contours shows that the barrier is aligned parallel to structural strike. Figure 5.10 illustrates, using permeability data, the outline of a coarse-grained, braided delta system.

Some carbonate examples are illustrated in the next two maps. Figure 5.11 shows the distribution of

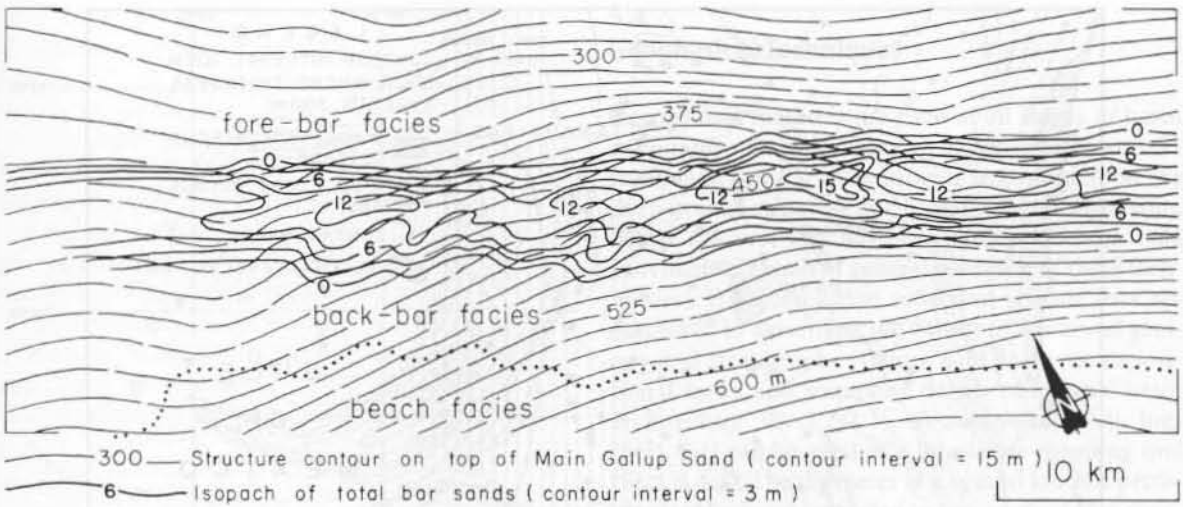


Fig. 5.9. Isopach of a Cretaceous barrier-bar sandstone in New Mexico. (Sabins 1972)

middle Cretaceous limestones in the Golden Lane area of Mexico. The Golden Lane Atoll is clearly demarcated, with a thinner sheet of deep-water talus deposits (Tamabra Limestone) banked against it (Enos 1977). Figure 5.12 shows porosity-thickness isopachs of a flanking unit in a Paleocene bioherm, Libya (Brady et al. 1980). The outline of the entire bioherm is shown by a dashed line. It developed the flanking biohermal unit during a time of emergence, when depositional slopes were thought to have been too great, on the west side of the reef, for sedimentation to occur.

Other facies-mapping techniques that might prove useful in particular cases include the plotting of bed thickness, the number of sandstone beds in the section, or the number of sandstone units greater than (for example) 5 m in thickness, and so on. Anything that reflects paleogeographic changes across the map area should be tried, and it is up to the ingenuity of the geologist to explore these. Padgett and Ehrlich (1978) used a special technique to illustrate fluvial drainage patterns in a Carboniferous coal basin. They mapped the area of mined-out coal for a single seam. Areas not mined correspond to dendritic channels that scoured into the original peat swamp (Fig. 5.13). At a different stratigraphic level, the same technique revealed the opposite effect: the coal seam itself outlined a channel, which became filled with peat follow-

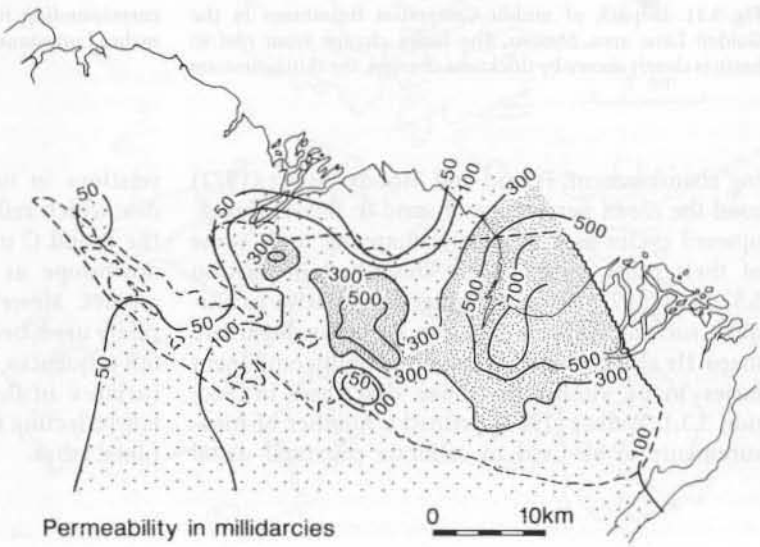


Fig. 5.10. Isopermeability map of the braided-channel complex in the Sadlerochit Formation, Alaska North Slope, in millidarcies. The Prudhoe Bay field is outlined by the *dashed line*. The *wavy line* indicates the faulted margin of the field. High permeability values are yielded by the coarse, texturally more mature deposits formed in proximal reaches of the complex. The crudely lobate shape of the area of high permeability corresponds to the distribution of the sand sheet, in the form of a large, fan-shaped, braided delta complex. (Wadman et al. 1979)

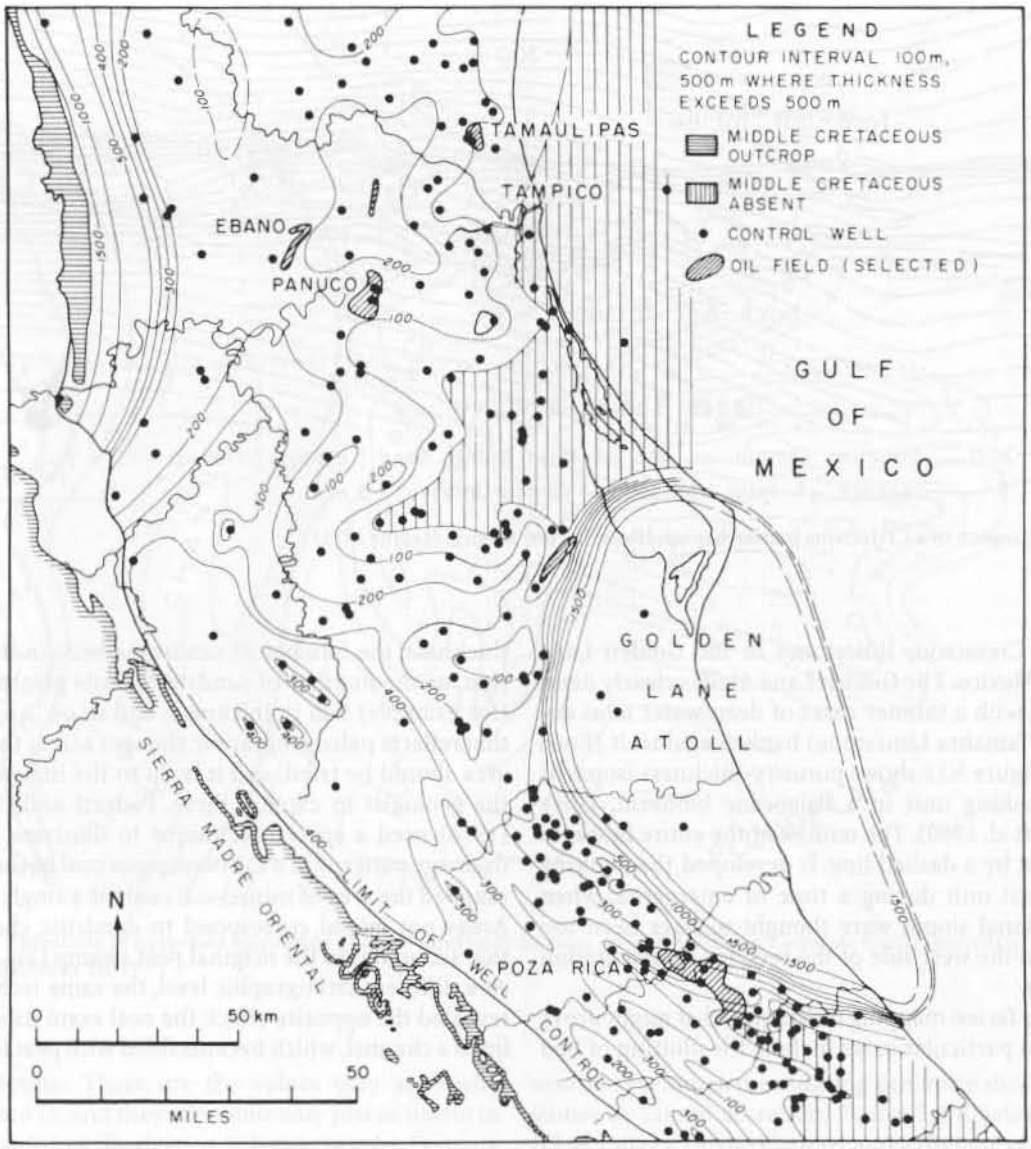


Fig. 5.11. Isopach of middle-Cretaceous limestones in the Golden Lane area, Mexico. The facies change from reef to basin is clearly shown by thickness changes, the thin limestone corresponding to the deep-water talus deposits of the Tamabra Limestone. (Enos 1977)

ing abandonment. Friend and Moody-Stuart (1972) used the mean percentage of sand in fluvial fining-upward cycles as a lithofacies-mapping tool (some of their other methods are described in Section 5.5). Rees (1972) discussed other quantitative techniques, such as distance-function or facies-departure maps. He also provided examples of multicomponent facies maps similar to those described in Section 5.3.1. Walker (1967) defined a number of measurements to be used to calibrate proximal–distal relations in turbidite beds, including the ABC index, which reflects the fact that first the A and then the B and C units in a Bouma sequence wedge out downslope as turbidity-current flow strength decreases. However, this mapping approach is now rarely used, because it is recognized that submarine-fan sequences may have several sources, and a wide variance in flow directions is possible on any one fan, reflecting the complex topography of the depositional lobes.

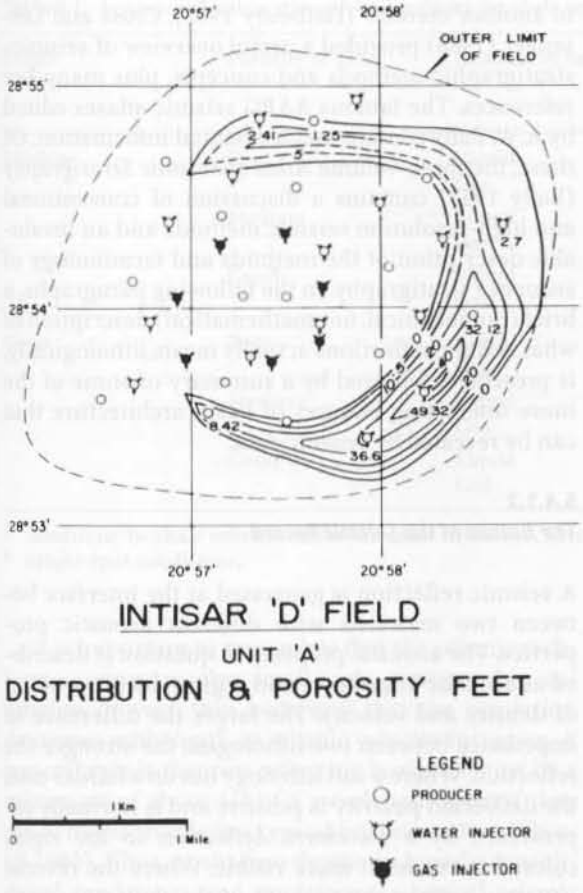


Fig. 5.12. Isopach of a porous section in a flanking unit, Intisar bioherm (Paleocene), Libya. (Brady et al. 1980)

5.4 Geophysical Techniques

Geophysical methods are used at all stages in basin exploration. Regional gravity, aeromagnetic, and refraction-seismic studies provide information on the deep-crustal structure and broad basin configuration before the first well has been drilled (Sharma 1986 provided an excellent general overview of these techniques). High-resolution reflection-seismic data are employed to determine the details of structural geology and stratigraphy (Chaps. 6, 8) and even depositional facies and some fine details of internal basin architecture (Sect. 5.4.1). Petrophysical well logs (Sect. 2.4) can be used as a lithofacies-mapping tool (Sect. 5.4.2). The dipmeter is a special kind of petrophysical tool and is treated in two parts of this chapter (Sects. 5.4.3, 5.6.6).

As exploration of a basin proceeds, geophysical and lithologic data derived from wells or diamond drill holes can be used to improve interpretations of the gravity or seismic data. In fact, such information is essential if seismic records are to be calibrated and their processing finely tuned to reveal the maximum amount of information. Even in well-explored basins, seismic interpretation of the areas between wells or in deeper parts of the section may be of considerable value, so a basin analyst may be using many different mapping techniques simultaneously and interactively.

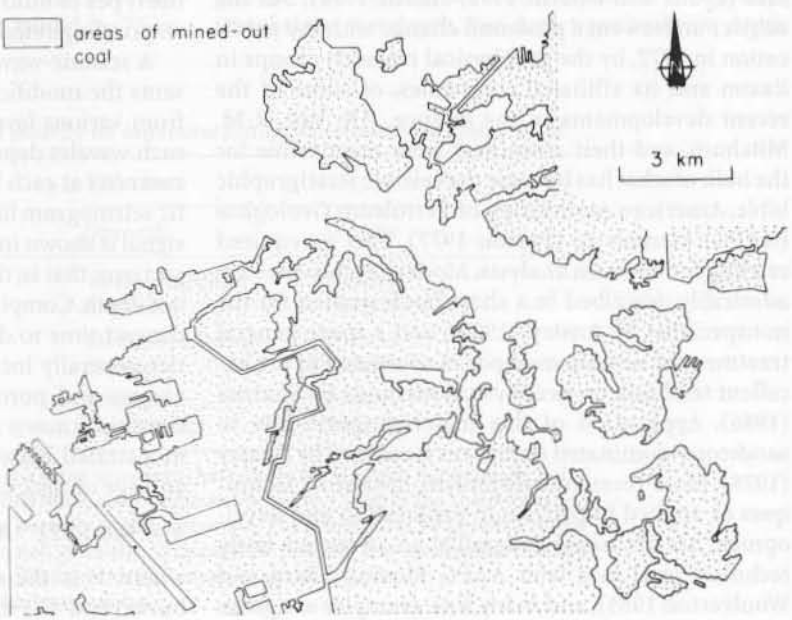


Fig. 5.13. Mined-out areas of a Carboniferous coal seam, West Virginia. *Unshaded areas* delineate a sandstone-filled channel system. (Padgett and Ehrlich 1978)

5.4.1 Reflection-Seismic Maps and Sections

5.4.1.1 *Introduction*

For many years, seismic-data collection and interpretation have absorbed a significant percentage of the exploration budget of petroleum companies because of the central role seismic maps and sections have played in delineating potential structural traps. In the 1970s, data acquisition and processing methods improved to the extent that we can now see many details of internal stratigraphic geometry, and these can be interpreted in terms of depositional environment, paleogeography, and sequence stratigraphy. In fact, sequence stratigraphy started life as **seismic stratigraphy**, first practiced by major oil companies, such as Exxon (Vail et al. 1977) and Shell (Shell Oil Company 1987). Seismic modeling is also an important field of development. Seismic techniques have, therefore, become an even more vital tool for basin analysts. The terms **seismic stratigraphy** and **seismic facies** are much in use among geologists; enthusiasts like to suggest that they are in possession of the ultimate weapon for exploring our piece of inner space. Enough is known about facies geometries and reflection patterns that, indeed, based only on skilled interpretation of good seismic records, a lot can be said about basin-fill composition and style and, hence, basin history. Seismic facies studies are largely an exercise in making interpretations from stratigraphic geometry.

In earlier years, many useful papers were published on the use of seismic methods in stratigraphic studies (Lyons and Dobrin 1972; Sheriff 1976), but the subject underwent a profound change with the publication in 1977, by the geophysical research groups in Exxon and its affiliated companies, of some of the recent developments in the science. P.R. Vail, R.M. Mitchum, and their coworkers were responsible for the bulk of what has become the seismic stratigraphic bible, American Association of Petroleum Geologists (AAPG) Memoir 26 (Payton 1977). This is required reading for all basin analysts. Modern techniques are admirably described in a short book written for the nonspecialist by Anstey (1982), and a more general treatment of seismic methods is contained in the excellent textbook on geophysical methods by Sharma (1986). Application of this material specifically to sandstone-dominated sequences is covered by Anstey (1978). More recent developments in seismic techniques as applied to petroleum exploration and development are discussed in detail at an advanced, fairly technical level in a later AAPG Memoir (Berg and Woolverton 1985), and many fine examples are given in another memoir (Halbouty 1982). Cross and Lessenger (1988) provided a useful overview of seismic-stratigraphic methods and concepts, plus many key references. The famous AAPG seismic atlases edited by A.W. Bally contain much essential information. Of these, the three-volume *Atlas of Seismic Stratigraphy* (Bally 1987) contains a discussion of conventional and high-resolution seismic methods and an invaluable description of the methods and terminology of sequence stratigraphy. In the following paragraphs, a brief (nontechnical, nonmathematical) description of what seismic reflections actually mean, lithologically, is presented, followed by a summary of some of the more interesting features of basin architecture that can be revealed by seismic data.

5.4.1.2 *The Nature of the Seismic Record*

A seismic reflection is generated at the interface between two materials with different acoustic properties. The acoustic property in question is described as acoustic impedance and is given by the product of density and velocity. The larger the difference in impedance between two lithologies, the stronger the reflection. Where a soft lithology lies on a harder one, the deflection polarity is positive and is normally represented by a waveform deflection to the right, colored to render it more visible. Where the reverse situation occurs, the polarity of the deflection is to the left (left blank). This is the origin of the familiar, variable-area **wiggle traces**. Colors are now commonly used to exaggerate amplitude differences, and these may emphasize major facies variations. Tables 5.1 and 5.2 list various combinations of lithology, the types of fluid saturant in pore spaces, and the corresponding reflection characteristics.

A seismic-wave trace recorded at a geophone represents the modification of a source pulse by reflection from various layers below the surface. The nature of each wavelet depends on the nature of the impedance contrasts at each bed contact. An example of a synthetic seismogram illustrating the production of a seismic signal is shown in Fig. 5.14. Such seismograms are time sections; that is, their long axis is proportional to time, not depth. Complex calculations must be performed to convert time to depth. This is because seismic velocities generally increase downward as rock density increases and porosity diminishes. An empirical relationship, known as the **Faust equation**, has been demonstrated between seismic velocity and the depth and age of rock units (Sharma 1986):

$$V = 46.5\,(ZT)^{1/6}\ \text{m/s}\,,$$

where V is the seismic velocity, Z is the depth of burial, and T is the geological age in years.

Table 5.1. Seismic reflection strength and polarity for shale resting on sandstone.[a] (Anstey 1978)

Depth	Porosity of sand	Saturant in sand	Reflection	
			Strength	Polarity
Shallow	Poor	Liquid	Strong	Positive
		Gas	Medium	Positive
	Medium	Liquid	Medium	Positive
		Gas	Weak	Positive
	Good	Liquid	Weak	Positive or negative
		Gas	Very strong[b]	Negative
Deep	Poor	Liquid	Strong	Positive
		Gas	Medium	Positive
	Medium	Liquid	Medium	Positive
		Gas	Weak	Positive or negative
	Good	Liquid	Weak	Positive or negative
		Gas	Weak to medium	Negative

[a] Sandstone-to-shale reflections, for the same situations, have the same strength but opposite polarity.
[b] Bright-spot conditions.

It is important to remember that the seismic technique cannot resolve small-scale stratigraphic anomalies or very thin beds and that the resolution decreases with depth as seismic velocity increases. A general rule is that two reflecting horizons must be a minimum of about 1/4 of a wavelength apart if they are to be resolvable in a typical seismic section (Sheriff 1985). Thus, at shallow depths in loosely consolidated sandstones and mudstones, a typical seismic velocity would be on the order of 1800 m/s, and the typical frequency would be around 60 Hz. One quarter wavelength is, therefore, 7.5 m. Resolution is much poorer in older and deeper formations. Consider, for example, a Paleozoic carbonate section with a seismic velocity of 4500 m/s and frequency of 15 Hz. The thinnest unit resolvable in this situation would be 75 m. This lower limit is much larger (thicker) than many of the important facies details that can be seen in outcrops and cores, as shown dramatically in Fig. 1.6. High-resolution techniques are available for studying sediments at shallow depths and are becoming widely used in detailed studies of stratigraphic architecture in modern environments (Bouma et al. 1987), but they are not yet available for the study of deep-basin structure and stratigraphy.

Published seismic sections show a series of vertical wave traces, each representing a recording made at a shot point. These are usually located about 25–50 m apart. Today, each trace is obtained by combining (**stacking**) the records of many adjacent geophones. The vertical scale of the section is **two-way-travel time**, not depth. For each time horizon, depth

Table 5.2. Seismic-reflection strength and polarity for sandstone resting on carbonate.[a] (Anstey 1978)

Porosity of sand	Saturant in sand	Reflection	
		Strength	Polarity
Poor	Liquid	Weak to medium	Positive
	Gas	Medium	Positive
Medium	Liquid	Medium to strong	Positive
	Gas	Strong	Positive
Good	Liquid	Strong	Positive
	Gas	Very strong[b]	Positive

[a] Carbonate-to-sandstone reflections, for the same situations, have the same strength but opposite polarity. No distinction of depth is made in this table, since depth has less effect than in the shale-to-sandstone case. The table assumes negligible porosity in the carbonate. The likely effect of porosity in the carbonate is to weaken all reflections but not, generally, to change the polarity. It is conceivable that the interface between a tight sand and a gas-filled porous carbonate could be negative.
[b] Also bright-spot conditions, but geologically uncomon.

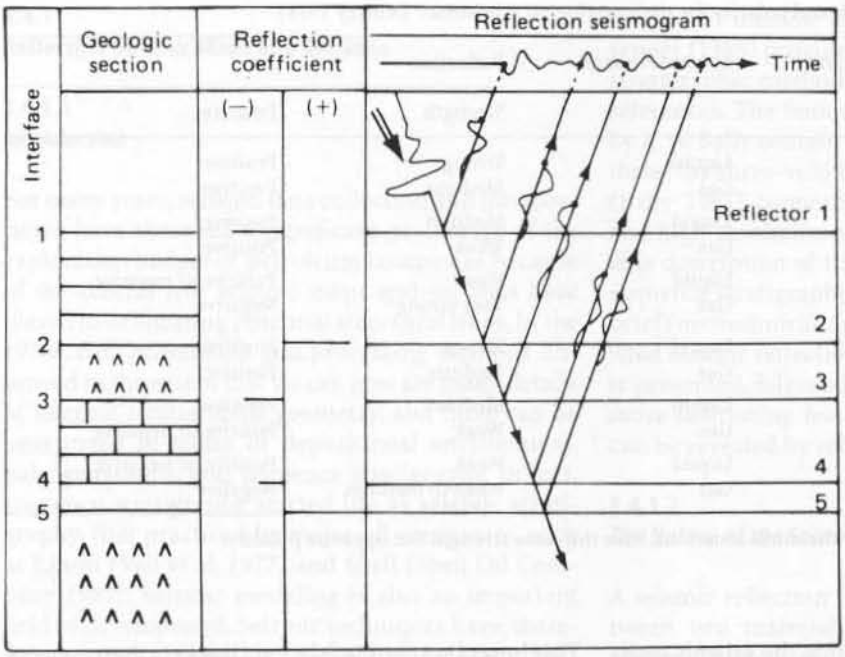


Fig. 5.14. Synthetic reflection seismogram generated from a simple geological section. At each interface, the input wavelet is modified by the reflection effects. (Sharma 1986)

varies according to velocity, as noted previously, so that the structure revealed on a section may be spurious. For example, a mass of high-velocity rock, such as a dense carbonate reef or channel fill, may overlie what looks like a gentle anticline. This is a case of **velocity pull-up**, where the more rapid acoustic transmission through the dense material returns reflections from the underlying layers sooner than occurs on either side. Computer processing can produce a depth-corrected section if accurate velocity information is available. This is discussed briefly below.

It is important to remember that it is contacts that produce reflections, not rock bodies as such. For example, tight, cemented sandstone and dense, nonporous carbonate have higher impedances than most mudrocks, so a contact of one of these with an overlying mudstone will yield a strong positive reflection. Conversely, a gas-filled sandstone has a very low impedance. Where it is overlain by a mudstone, there will be a strong negative deflection. This appears as a blank area in the seismic record and is the basis of the **bright-spot technique** for locating gas reservoirs. However, there are many conditions where interbedded units may have contrasting lithology but very similar impedances, so they yield little or no seismic reflections. Examples would include a contact between a tight sandstone and a porous carbonate or between a very porous oil- or water-saturated sandstone and a mudstone. Tables 5.1 and 5.2 demonstrate a range of such conditions.

A very important assumption, first emphasized by Vail et al. (1977) and reemphasized by Cross and Lessenger (1988), is that, in most cases seismic reflections from conformable bedding surfaces are chronostratigraphic in nature. Facies contacts are diachronous, but they are too gradual to yield reflections. Facies changes may be deduced from subtle lateral changes in reflection characteristics within a single bed. This assumption is of profound importance in the study of basin structure and stratigraphy and the analysis of stratigraphic sequences (Chaps. 6, 8).

Imagine an angular unconformity with lateral facies changes in the overlying unit and a gently dipping, truncated sequence of different units below. The seismic reflection from the unconformity could vary from strongly positive to strongly negative, with some areas completely invisible. This would be a difficult reflection to follow accurately on seismic records (Fig. 5.15).

Seismic records are also complicated by **multiples** and **diffractions**. Multiples are reflections from a shallow layer that are reflected downward again from a near-surface level such as the base of the weathering zone. This internal reflection is then redirected toward the surface from the original point and appears in the seismic record as a parallel layer several tenths

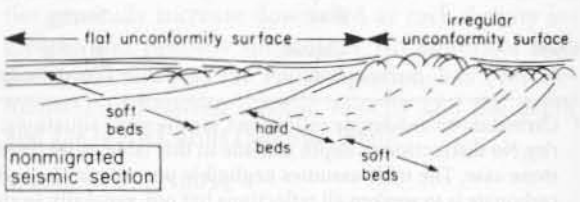


Fig. 5.15. Seismic multiples, diffractions, and reflections developed by an unconformity surface with variable impedances

of a second lower down in the section. Multiples may be strong enough to obscure deep reflections, but they are now relatively easy to remove by processing. Diffractions are reflections from steeply dipping surfaces, such as faults, channel margins, and erosional relief on an unconformity (Fig. 5.15). Their appearance in the record occurs because seismic waves are not linear, laser-like energy beams but spherical wave fronts. The pulse from each shot point, therefore, sees reflection events over a circular region (the **Fresnel zone**), the diameter of which can be shown by calculation to range from about 100 to 1200 m, depending on depth and wavelength. A reflection on the periphery of the wave front will be seen later than one at the same depth directly below the shot point and, if at a high angle from the horizontal, may produce complex internal reflections. The resulting diffraction patterns may be very pronounced. They are a useful indicator of a steep reflecting event but may confuse other reflections and can be removed by a processing routine called **migration**, which repositions the reflections in their correct spatial position. The circular reflection area also accounts for the fact that diffractions can be produced by reflectors off the line of section, for example, if the seismic line skims past a reef body or a buried fault-line scarp.

The seismic record can be considerably improved and interpretation facilitated by the use of two special techniques: the construction of **synthetic seismograms** and the use of **vertical seismic profiling** (VSP) (Hardage 1985). Synthetic seismograms are generated by the conversion of sonic and density data, from petrophysical logs, into a series of reflection coefficients (Sect. 2.4 for a discussion of petrophysical logs). Vertical seismic profiling is the recording and analysis of seismic signals received from a geophone lowered downhole. Signals generated at the surface are recorded as they descend and as they return to the surface. The geophone is moved in increments, the length of each being set at less than half the shortest wavelength of the signals being analyzed. A new recording is made at each position. A comparison between synthetic, VSP, and field-recorded data is shown in Fig. 5.16. Both synthetic and VSP data are extremely valuable for the purposes of calibrating the seismic record. For example, they enable generation of a detailed record of seismic velocities that can be used to produce depth-corrected cross sections. However, for various reasons, the VSP technique is far superior. It can, for example, be used to extend the calculations below the total depth (TD) of the well, which synthetic seismograms cannot (Fig. 5.16). The

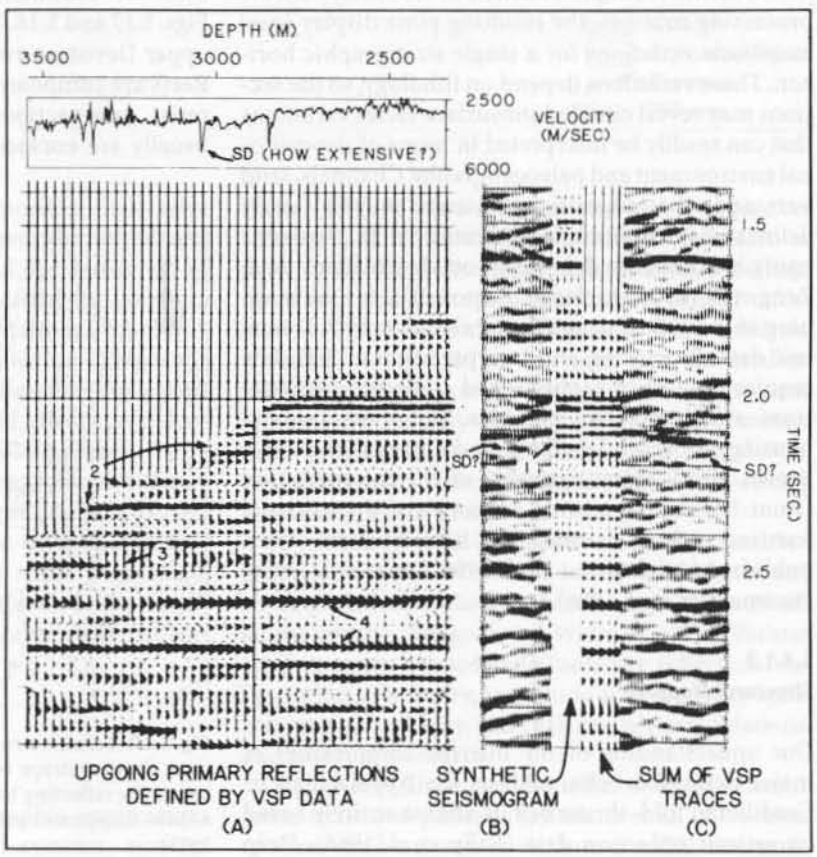


Fig. 5.16. Comparison of vertical seismic profiling (VSP) data with a synthetic seismogram and that produced by normal, surface field recording. The surface section is split in two at the location of a well down which the synthetic record and the VSP traces were generated. At *left* is shown the VSP traces, and *above* is the sonic log of the well. Total well depth is at 2.41 s. Notes: *1* calcareous sandstone (see text); *2* curvature of VSP reflections implies structural dip; *3* termination of reflection implies the presence of a fault near the borehole; *4* VSP data from below total depth. (Hardage 1985)

synthetic method relies on well-log data, which only reflect conditions immediately adjacent to the hole. In the case of poor hole conditions or seismic reflectors having a very small horizontal extent (less than one Fresnel zone), the synthetic method may not provide a record typical of the area. For example, in Fig. 5.16, there is a discontinuous, calcareous sandstone (denoted by the letters *SD*) at a depth of about 3070 ft. It is visible as a strong reflector at about 2.15 s in the synthetic seismogram, but not on the VSP trace, indicating that the bed does not extend very far from the borehole. Other possible occurrences of the bed are shown by the letters *SD* on the field record. VSP data are now routinely used to improve subsurface imaging around critical boreholes, to carry out precise time–depth conversions, and to remove multiples.

An important advance in seismic methods during the last two decades is the use of three-dimensional (3-D) seismic data (Brown et al. 1980, 1982; Brown 1985, 1986). A tight grid of seismic lines is run across an area at line spacings of a few tens of meters. This permits the plotting of horizontal seismic sections at selected depths by correlation of key reflectors from line to line. Correlation of key reflectors can be carried out by computers using automatic tracking techniques, and this provides a facility for plotting seismic maps of selected stratigraphic horizons. Where the structure is simple or can be removed by suitable processing routines, the resulting plots display areal amplitude variations for a single stratigraphic horizon. These variations depend on lithology, so the sections may reveal detailed subsurface facies variations that can readily be interpreted in terms of depositional environment and paleogeography. Channels, sand bars, and other depositional features may be clearly delineated (an example is illustrated below). By developing synthetic models based on electrofacies data, Zeng et al. (1996) explored the potential for the mapping of individual depositional elements by 3-D seismic data. Processing and interpretation of 3-D data require the use of sophisticated computer workstations and interactive graphics. Color display and plotting are essential (Savit and Changsheng 1982; Brown 1986). Many examples of the interpretation (from horizontal seismic sections) of depositional features, such as channels and bars, have now been published (Armentrout et al. 1996; Lopez et al. 1996; Posamentier et al. 1996)

5.4.1.3 *Structural Mapping*

Our understanding of the internal complexities of major deformed belts, such as the Appalachian or Cordilleran fold–thrust belt, is almost entirely based on seismic-reflection data (Bally et al. 1966). Deep seismic penetration using the most sophisticated of industry techniques has formed the basis for such large-scale, multidisciplinary projects as COCORP (the Consortium for Continental Reflection Profiling) in the United States and LITHOPROBE in Canada. It has also produced evidence of dramatic crustal shortening in areas such as the Cordilleran foreland region of Wyoming and the southern Appalachians, and huge low-angle detachment faults in areas of extension, such as the Atlantic borderlands and the Basin and Range Province. This has profound implications for the plate-tectonic interpretation of major suture zones and other plate margins and is discussed in Chapter 9.

In the area of stratigraphic-sedimentologic basin analysis, seismic data may be of considerable assistance in delineating the configuration of important internal surfaces, bearing in mind the limitations on the seismic record noted above. Some examples of the kinds of data useful to the basin analyst are discussed in Section 5.2.

The best reflections on a seismic record may be traceable for hundreds or thousands of kilometers through the available grid of seismic lines. It is a common procedure to plot and contour the two-way-travel times of such reflections, producing an analogue structural map. Two examples illustrating features of sedimentological interest are shown in Figs. 5.17 and 5.18. The first of these shows drape over upper Devonian reefs in Alberta (from Evans 1972). Reefs are composed of carbonate accumulations that resist compaction under burial pressure. They usually are encased in less resistant offreef litholo-

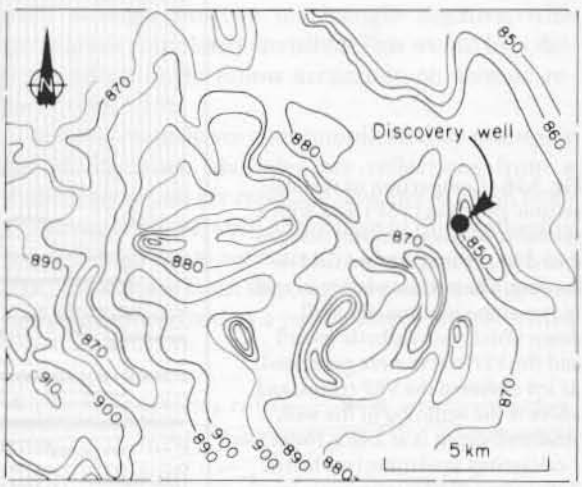


Fig. 5.17. Seismic structure map, contoured in two-way travel time, showing drape over middle Devonian (Zama) reefs, Alberta. The reflecting horizon is the top Slave Point Formation, a shale–limestone contact about 200 m above the reefs. (Evans 1972)

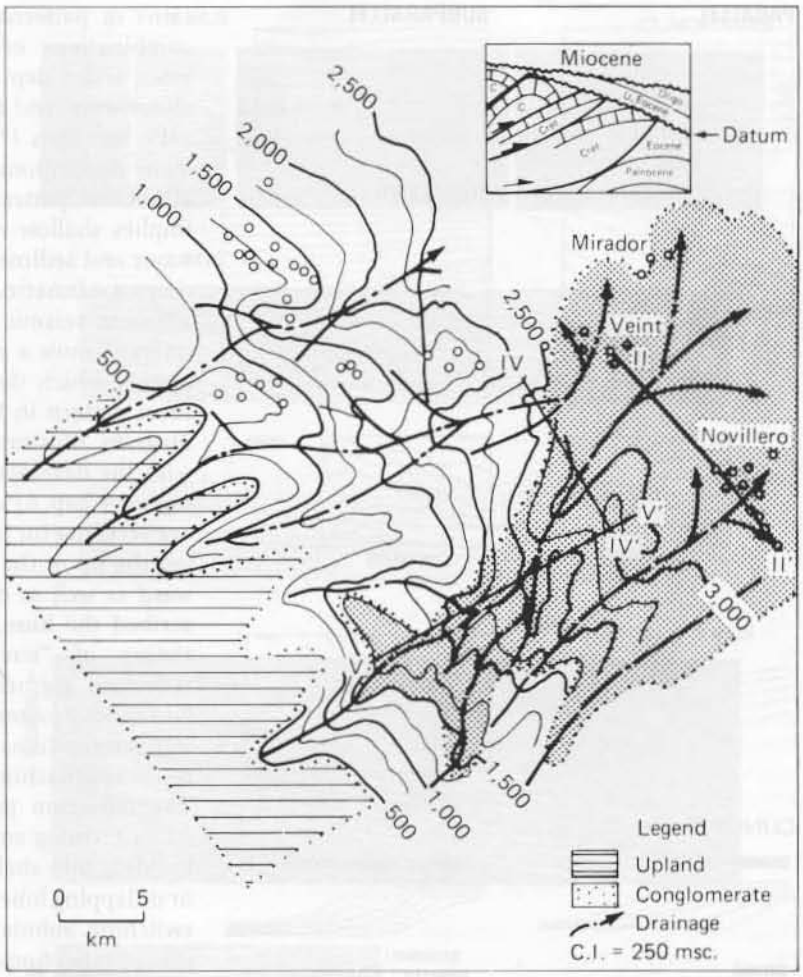


Fig. 5.18. Seismic structure map defining dendritic drainage channels on an unconformity surface, Cretaceous–Tertiary contact, Veracruz Basin, Mexico; *msc* milliseconds. (Helu et al. 1977)

gies, such as mudrocks or evaporites. Successive bedding units, therefore, drape over the initial reef units, and the drape may extend for hundreds of meters upsection. In Western Canada, the compaction-drape effect of Devonian reefs may commonly be seen in structure contours drawn on Cretaceous horizons, and it seems likely that the subtle variations in subsidence rates even had slight effects on paleogeography for several hundred million years.

Structure contours drawn on unconformities may reveal erosional relief, buried fault-line scarps, etc. Figure 5.17 illustrates a dendritic drainage pattern eroded into uplifted Cretaceous rocks in Mexico. Conglomerates derived from the upland to the southwest were deposited as a fanglomerate wedge downslope (Helu et al. 1977).

5.4.1.4 *Seismic Facies*

Seismic data are used in stratigraphic applications primarily to define depositional systems, systems tracts (Chap. 6) and regional and global sequences (Chaps. 6, 8). However, the internal configuration or pattern of reflections within such seismically defined units may contain much essential deposition information. This is the subject of seismic facies analysis. It should be repeated at this point that the smallest features resolvable on conventional industry seismic sections are typically several meters or tens of meters in vertical amplitude (Fig. 1.6 and discussion above), so that the term 'facies' has a quite different meaning from that used in Chapter 4.

A seismic facies is defined as an areally restricted group of seismic reflections whose appearance and characteristics are distinguishable from those of adjacent groups (Sangree and Widmier 1977). Various attributes may be used to define facies: reflection configuration, continuity, amplitude and frequency spectra, internal velocity, internal-geometrical relations, and external 3-D form.

Figure 5.19 illustrates the main styles of seismic-facies reflection patterns (Mitchum et al. 1977). Most of these are best seen in sections parallel to deposi-

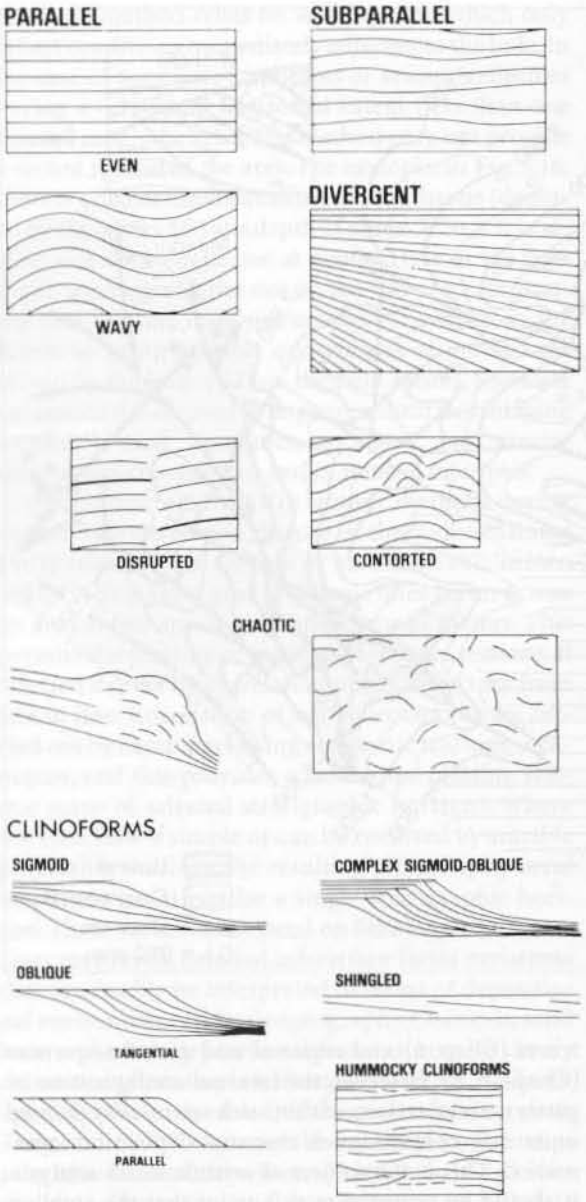


Fig. 5.19. Typical seismic reflection patterns illustrating the concept of seismic facies. (Mitchum et al. 1977)

tional dip. Parallel or subparallel reflections indicate uniform rates of deposition; divergent reflections result from differential subsidence rates, such as in a half-graben or across a shelf-margin hinge zone. Prograding reflections comprise an important class of seismic-facies patterns. They are particularly common on continental margins, where they represent deltaic or continental-slope outgrowth. The dipping surfaces, like large-scale crossbed foresets, are termed **clinoforms**. As noted in Chapter 6 they demonstrate that, in many basins, lateral accretion is as important a part of basin filling as vertical aggradation. Variations in patterns of progradation reflect different combinations of depositional energy, subsidence rates, water depth and sea-level position. **Sigmoid clinoforms** tend to have low depositional dips, typically less than 1°, whereas **oblique clinoforms** may show depositional dips up to 10°. **Parallel-oblique clinoform** patterns show no topsets. This usually implies shallow-water depths with wave or current scour and sediment bypass to deeper water, perhaps down a submarine canyon that may be revealed on an adjacent seismic cross section. Many seismic sequences show a very complex prograding stratigraphy, of which the **complex, sigmoid-oblique clinoform** pattern in Fig. 5.19 is a simple example. This diagram illustrates periods of sea-level stillstand, with the development of truncated topsets (termed toplap; Chap. 6) alternating with periods of relative sea-level rise (or more rapid basin subsidence), allowing the lip of the prograding sequence to build upward as well as outward. Mitchum et al. (1977) described the **hummocky clinoform** pattern as consisting of "irregular discontinuous subparallel reflection segments forming a practically random hummocky pattern marked by nonsystematic reflection terminations and splits. Relief on the hummocks is low, approaching the limits of seismic resolution ... The reflection pattern is generally interpreted as strata forming small, interfingering clinoform lobes building into shallow water," such as the upbuilding or offlapping lobes of a delta undergoing distributary switching. Submarine fans may show the same hummocky reflections. **Shingled clinoform** patterns typically reflect offlapping sediment bodies on a continental shelf. Fulthorpe and Austin (1998) showed how detailed mapping of clinforms using a grid of seismic lines throws considerable light on continental-margin transport and depositional processes.

Chaotic reflections may reflect slumped or contorted sediment masses or those with abundant channels or cut-and-fill structures. Disrupted reflections are usually caused by faults. Lenticular patterns are likely to appear most frequently in sections oriented perpendicular to depositional dip. They represent the depositional lobes of deltas or submarine fans. Using the large-scale bedding patterns observed in the very large carbonate outcrops of the Italian Dolomites, Doglioni et al. (1989) elaborated on the definition of these seismic, stratal facies architectures.

Some examples of these patterns are illustrated in Figs. 5.20 and 5.21. All the features shown in Fig. 5.19 are several hundreds of meters thick and several kilometers or even tens of kilometers across, which explains why they were not properly documented until they were seen on high-quality seismic sections. Few individual outcrops are big enough to reveal them and, on stratigraphic cross sections constructed

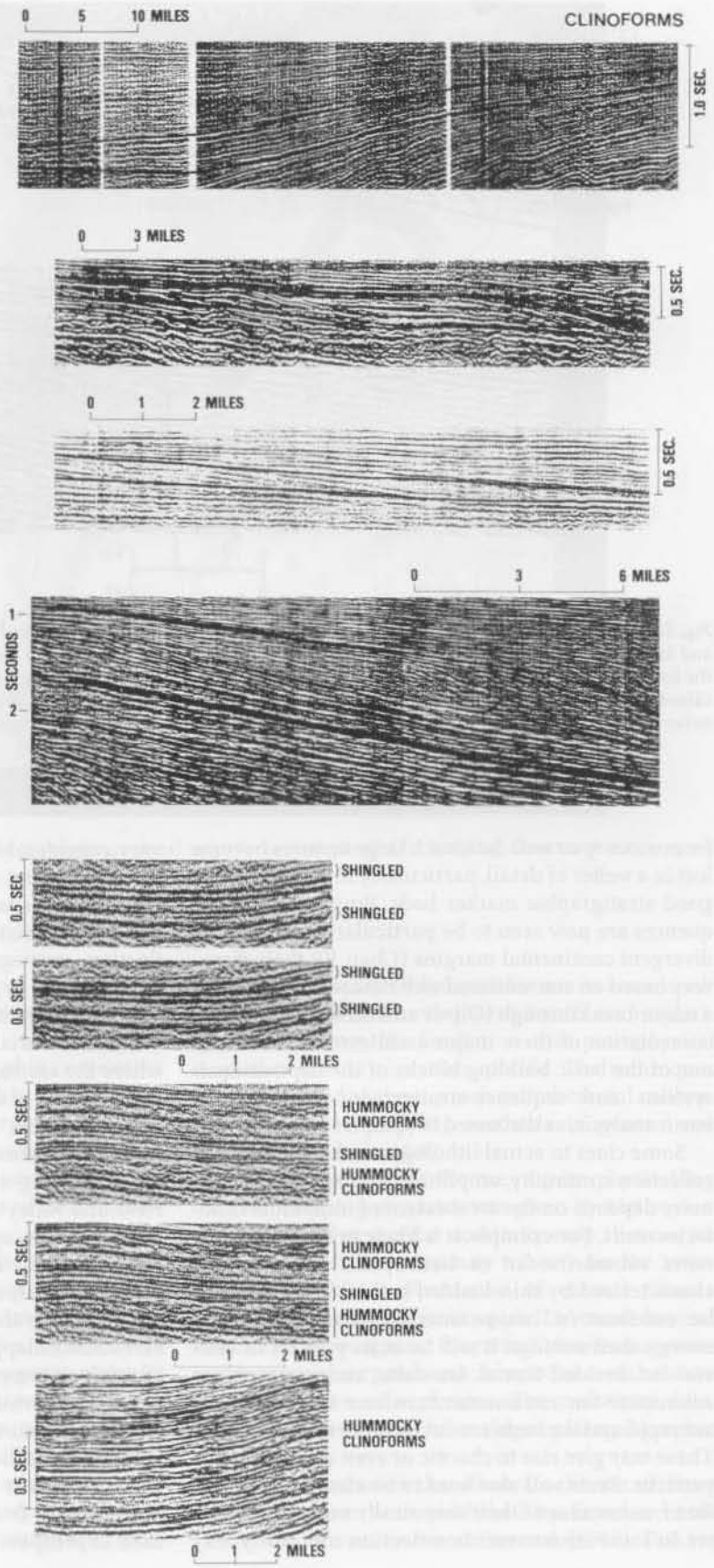


Fig. 5.20. Examples of seismic lines, showing typical seismic facies patterns. (Mitchum et al. 1977)

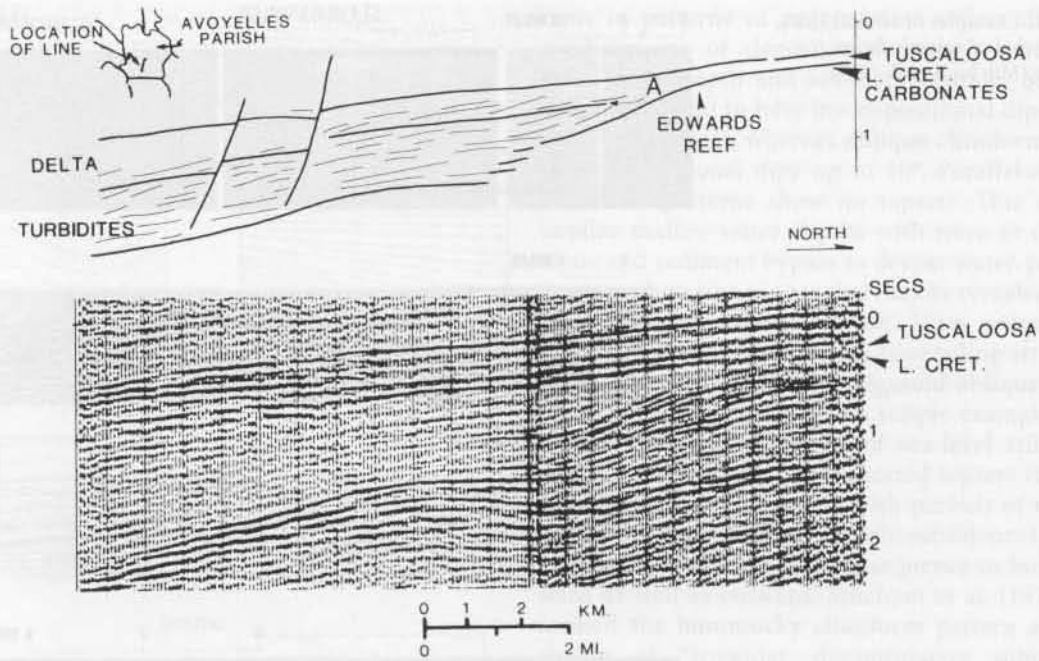


Fig. 5.21. Example of a seismic line to illustrate stratigraphic and facies mapping. A lower Cretaceous reef grades (toward the *left*) into a steep forereef slope. The upper Cretaceous Tuscaloosa Formation is a clastic unit. Turbidites at the base may onlap the reef, although this is not clear from the seismic data. Surface *A* marks a facies contact with younger deltaic deposits, which drowned the reef. The turbidites may be contemporaneous, downdip facies equivalents of the deltaic deposits. Alternatively, the prominence of the A reflector suggests it may be a surface of toplap and a sequence boundary. (Berg 1982)

from outcrop or well data, such large features become lost in a welter of detail, particularly in the absence of good stratigraphic marker beds. Progradational sequences are now seen to be particularly common on divergent continental margins (Chap. 9); their discovery based on conventional well data was regarded as a major breakthrough (Oliver and Cowper 1963). Documentation of these major architectural features is one of the basic building blocks of the depositional-systems and sequence-stratigraphic approach to basin analysis, as discussed in Chapter 6.

Some clues to actual lithology can be gained from reflection continuity, amplitude, and spacing. Continuity depends on the areal extent of individual lithofacies units. For example, it is likely to be excellent in outer submarine-fan or basin-plain environments characterized by thin-bedded turbidites. It will also be excellent in basin-center lacustrine and low-energy shelf settings. It will be at its poorest in alluvial-fan, braided-fluvial, fan-delta, and coarse, inner submarine-fan environments, where facies changes are rapid and the beds are cut by numerous channels. These may give rise to chaotic or even reflection-free patterns. Reefs will also tend to be chaotic or reflection free because of their seismically massive character. In some environments, reflection continuity may vary considerably throughout the section. For example, prograding deltas may be characterized by discontinuous or hummocky clinoform reflections capped by a persistent horizontal or gently dipping reflection corresponding to the thin, fine-grained, transgressive deposits of the abandonment phase.

Amplitude and spacing are controlled by vertical lithofacies variability. They will be at their highest where the section is characterized by cyclic sequences consisting of contrasting lithologies and where the cycle thickness is greater than the limit of seismic resolution (namely, a few meters or tens of meters). Fluvial fining-upward cycles (thick, meandering-river and valley-fill examples), prograding deltaic or barrier cycles, and the coarsening-upward cycles of submarine-fan lobes are good examples.

Two interpreted examples of seismic stratigraphy and facies are also shown here. Figure 5.21 illustrates turbidites onlapping a reef body, and the deposition of a deltaic complex across the top of it. Figure 5.22 is a black-and-white rendering of a colored diagram (in which the features are much clearer) showing a horizontal 3-D section following a depositional surface that contains a deltaic distributary channel. Such horizontal 3-D seismic sections are now routinely used in prospecting for stratigraphic traps.

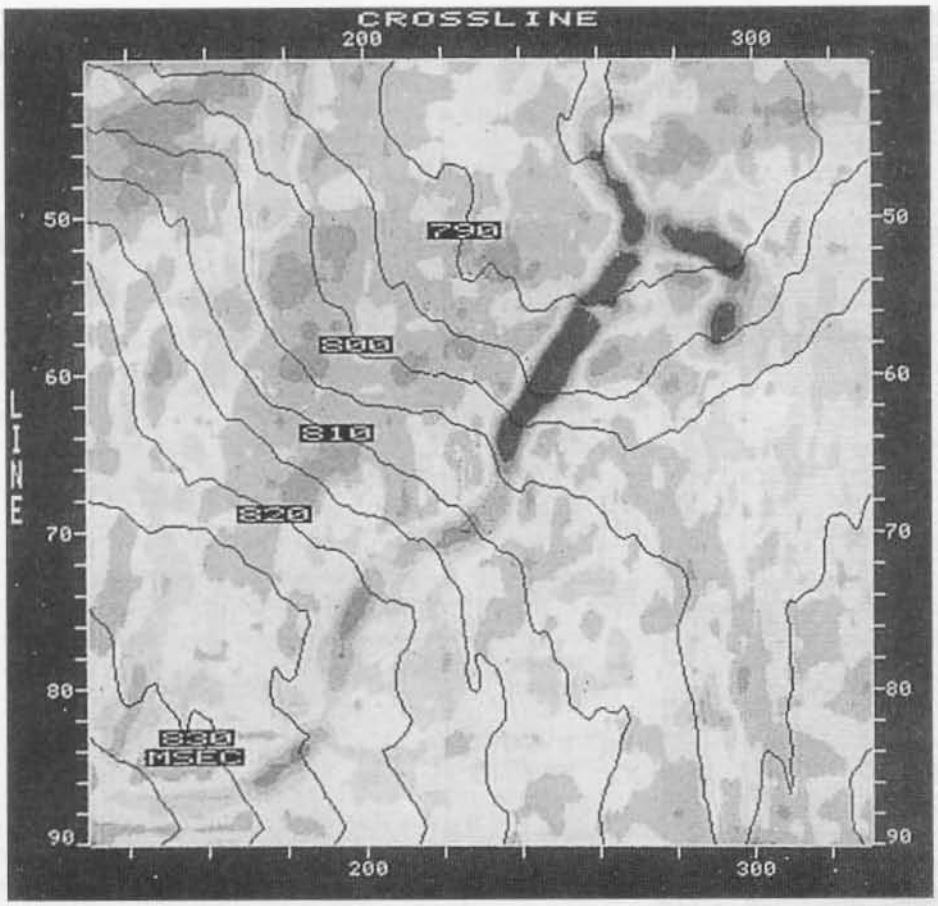


Fig. 5.22. Example of a channel seen on a three-dimensional seismic section parallel to bedding. Structural contours are superimposed. (Brown 1985)

5.4.2 Lithofacies Mapping with Petrophysical Logs

Well-log data are used for stratigraphic correlation and for lithologic interpretation. They may also be used directly for facies-mapping purposes, although this is a technique that has not been widely employed. As noted briefly in Section 2.4, various combinations of logs can be employed in crossplots to yield lithologic information. These relationships can be converted into computer algorithms and, where log data are digitized and stored in data banks, a powerful, automated mapping technique becomes available (Fig. 5.23). Digital log data can also be displayed and manipulated using interactive computer-graphics routines, a facility that permits easy comparisons for purposes of stratigraphic-correlation, normalizing of logs made at different times using different hole conditions, and many others (Srivastava et al. 1982). Well-log service companies have devoted considerable energy to devising and marketing automated processing and display techniques for use in basin analysis and petroleum development, but all such techniques suffer from the limitation that petrophysical responses are very location specific. The techniques cannot be used without much initial careful calibration (computer mapping methods in general are discussed in Section 5.8).

An early example of the use of petrophysical-log data for mapping purposes was given by McCrossan (1961). He constructed a resistivity map that can be used to predict the proximity to reefs for the lower and middle Ireton Formation of central Alberta (Fig. 5.24). The Ireton is an upper Devonian basinal shale that drapes against the reefs. Calcareous content and thus resistivity increase close to the reefs because of the presence of reef talus. The resistivity data, therefore, provide a useful analogue paleogeographic map.

Two recent studies that illustrate the use of these techniques have been published by Doveton et al. (1984) and Bornemann and Doveton (1983). Two re-

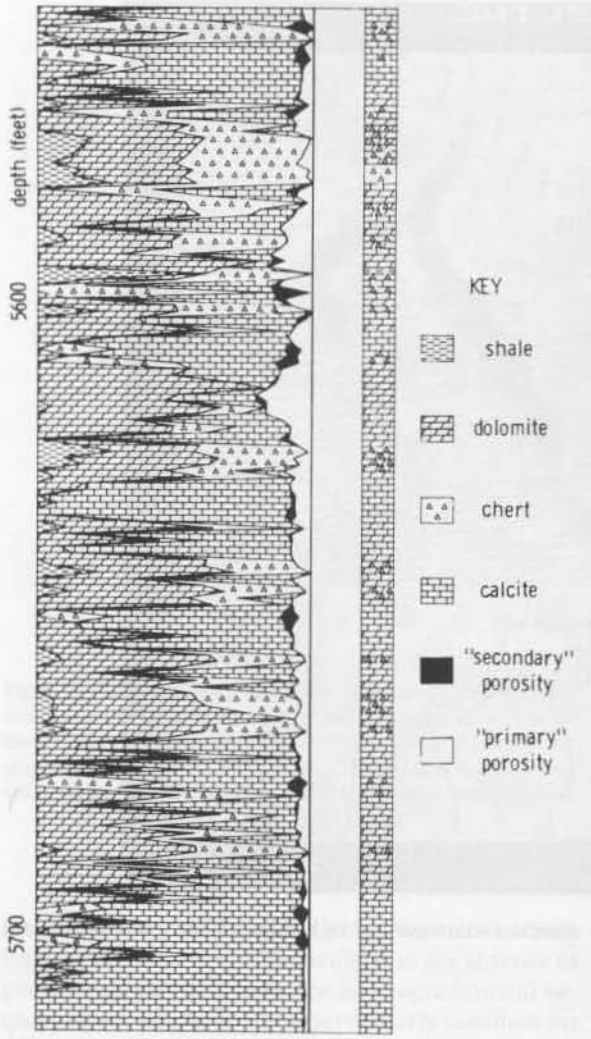


Fig. 5.23. Mineralogical profile and porosity in a well section, computed, using matrix algebra, from petrophysical logs. (Bornemann and Doveton 1983)

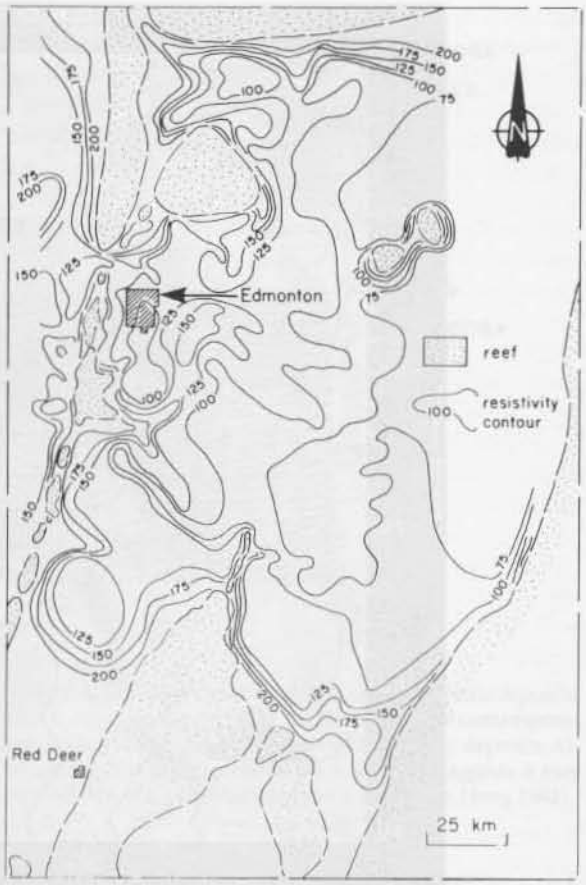


Fig. 5.24. Resistivity map of the Ireton Formation (Devonian), central Alberta. High values indicate an increase in carbonate content adjacent to reef bodies. (McCrossan 1961)

cent, comprehensive manuals published by Schlumberger (Serra 1985; Frank 1986; and more recent updates, available directly from the company) provide many examples but must be read with caution, as they tend not to emphasize the limitations inherent in the use of petrophysical data and interpretations made largely from vertical profiles.

The recognition of characteristic log "shapes" and distinctive vertical-profile character in siliciclastic deposits (**electrofacies**) may be a useful tool for correlation purposes and an aid for interpreting depositional environment. The technique is almost as old as modern sedimentology itself (Nanz 1954). The technique depends on the distinctiveness of the vertical profile through a sandstone body and our ability to interpret this profile in terms of depositional processes and environment. The well-known, bell-shaped, gamma-ray or spontaneous-potential log response yielded by a typical fluvial fining-upward cycle is a classic example. Visher et al. (1971) used log character and shape to discriminate various fluvial and deltaic environments in the Bartlesville Sandstone of Oklahoma. Wightman et al. (1987) reported on a similar type of exercise applied to the Mannville Group of the Lloydminster area, Alberta. Using log shape, Hamilton and Galloway (1989) discriminated various fluvial styles in a Permian fluvial unit in Australia (Fig. 5.25). Bulling and Breyer (1989) used a combination of electrofacies and detailed sand-isolith maps to define individual depositional elements, such as crevasse-splay sands, washover fans, bar-finger sands, etc. (cf. Fig. 5.8). Zeng et al. (1996) used a similar approach as a basis for the construction of synthetic, 3-D seismic models in order to explore the potential of the seismic technique to identify individual lithosomes in the subsurface.

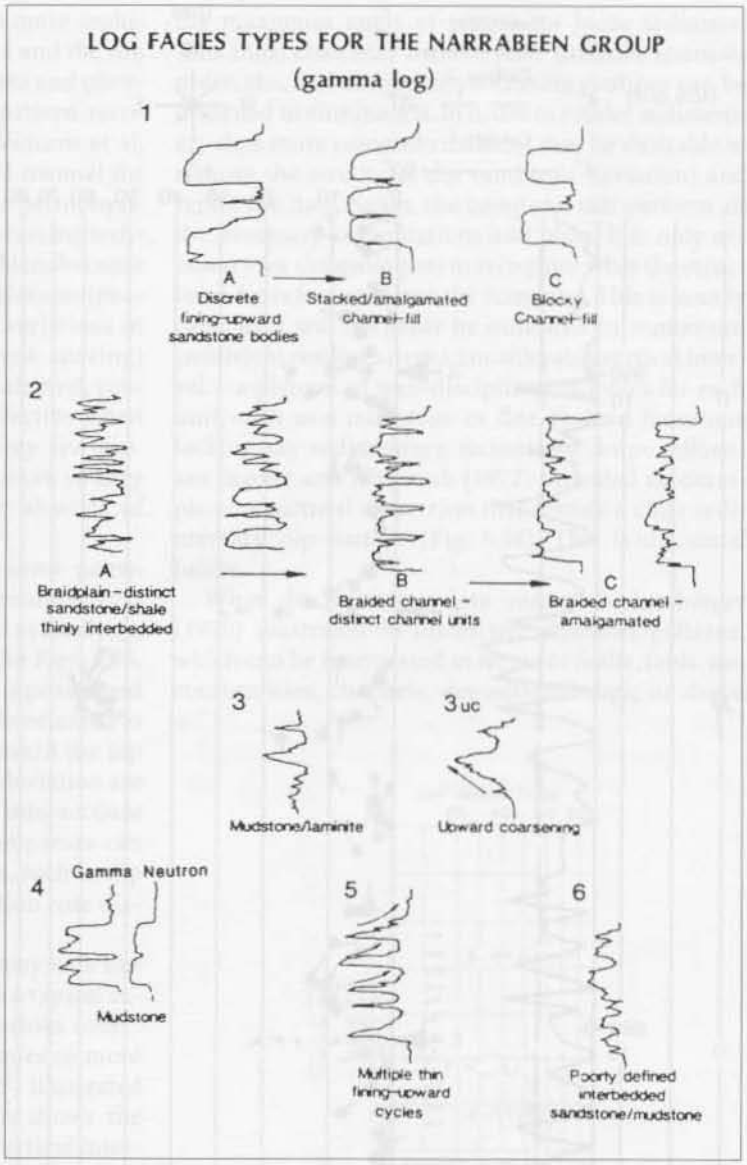


Fig. 5.25. Log facies types for the Permian Narrabeen Group, Sydney Basin, Australia. (Hamilton and Galloway 1989)

5.4.3 The Dipmeter

One of the most sensitive petrophysical logs is the microresistivity device, which can record lithologic units as thin as a few centimeters. The dipmeter tool consists of four microresistivity recorders arranged at 90° to each other around the hole. Comparison of the resulting curves will yield many points at which correlations of the same bed (or other types of hole intersection) can be made between two, three, or all four curves (Fig. 5.26). Any slight differences in depth of the correlated bed reflect dip, and it is simply a matter of applying the principle of the three-point problem, familiar to all first-year undergraduates, in order to solve for dip azimuth and angle. The dipmeter will respond to any kind of dipping surface in the rocks and, therefore, has considerable potential in structural, facies, and paleocurrent analysis, as shown below. A more sophisticated variation of this tool is the Formation MicroScanner (FMS), which generates a continuous resistivity image of the entire hole that is plotted "unwrapped" as a strip log. Dipping surfaces that intersect the hole appear as curves on the FMS log, and online computer manipulation of these logs can be performed to calculate their dip and orientation (Dueck and Paauwe 1994; Carr-Crabaugh, et al. 1996).

The dipmeter tool has been used as an aid in structural mapping for many years (Stratton and Hamilton

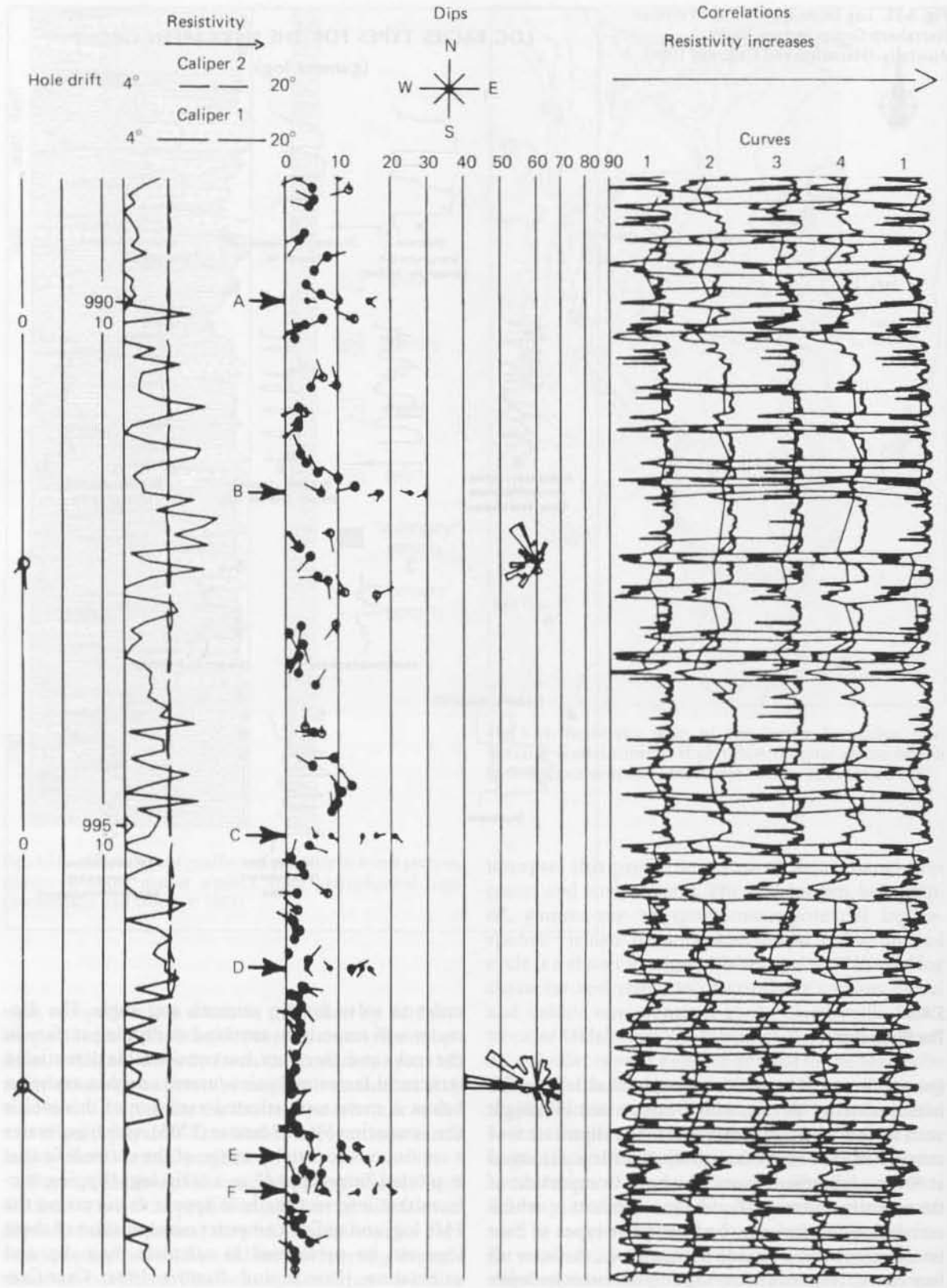


Fig. 5.26. Typical computer plot of dipmeter microresistivity recording (*right*) and calculated dips (*tadpole plots*). A hole-deviation plot and summary rose diagrams are also shown. (Vincent et al. 1977; copyright, Society of Petroleum Engineers-American Institute of Mining, Metallurgical, and Petroleum Engineers)

1947), but the method has become much more sophisticated with the advent of digitized logs and the use of the computer for correlating the curves and plotting the results. For example, modern pattern-recognition techniques were described by Vincent et al. (1977). Serra (1985) provided a detailed manual for the interpretation of dipmeter and other petrophysical data based on modern computer-processing techniques. Data processing is a difficult problem because of the possibility of spurious curve correlations (producing noise), which are aggravated by variations in lithology (and, hence, correlatable event spacing) with depth. As is the case with the seismic tool, processing and interpretation are most effective when adapted to specific local conditions. Very few dipmeter patterns have a unique interpretation, so they cannot be correctly interpreted in the absence of lithologic data.

Dipmeter logs are routinely run by some petroleum companies in order to assist with prospect evaluation. The data are normally presented as a **tadpole plot**, examples of which are shown in the Figs. 5.26, 5.28 and 5.30. The head of each tadpole is positioned on a grid to show dip angle, and the tail is oriented to indicate azimuth, north always being toward the top of the plot. The azimuth and dip of hole deviation are also usually shown and must be taken into account for detailed interpretations. Logging companies can also provide other kinds of data displays, such as dip histograms, stereonet plots, and orientation rose diagrams for selected intervals.

At first sight, a typical dipmeter plot may look like a confusing welter of detail. Figure 5.26 is a typical example. The first task is to separate the various sources of dip, using simple graphical techniques or more complex statistical routines. Rider (1978) illustrated the use of a histogram method, which shows the range of dips recorded over a selected vertical interval (Fig. 5.27). In this case, the readings fell into three clearly defined classes: low structural dips, sedimentary dips, and those higher than 35°, which is about the maximum angle of repose for loose sediment. This third class may include joint surfaces, spurious noise, etc., and computer-processing routines can be designed to eliminate it. In order to render sedimentary dips more comprehensible, it may be desirable to remove the structural dip (and hole deviation) and replot the data. Again, the computer can perform all the necessary computations and plots. It is only necessary for the geologists to recognize what the structural dip is for input into the computer. This is usually clear, as it will normally be indicated by numerous, consistent readings over a considerable vertical interval – a platoon of well-disciplined tadpoles for each unit, such as a mudstone or fine-grained limestone lacking any sedimentary, tectonic, or drape influences. Jageler and Matuszak (1972) provided an example of structural correction that reveals a clear sedimentary dip pattern (Fig. 5.28). This is discussed below.

What do dipmeter data mean? Schlumberger (1970) illustrated 34 distinctive dipmeter patterns, which can be interpreted in terms of faults, folds, unconformities, channels, depositional dips, or drape

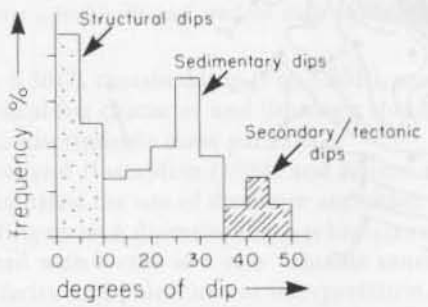


Fig. 5.27. Histogram of calculated dips for a selected stratigraphic interval, showing three classes of dip. (Rider 1978)

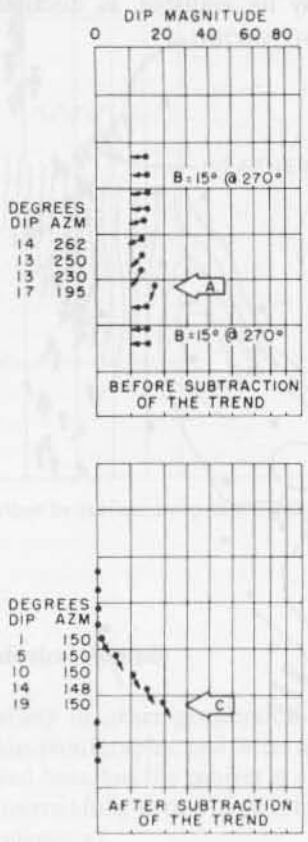


Fig. 5.28. Correction of dipmeter readings for regional structure, which revealed a typical sedimentary pattern of an upward decrease in dip. (Jageler and Matuszak 1972)

over massive sedimentary bodies. Gilreath (1977) discussed some of the more common examples. An exhaustive compilation based on modern facies-modeling studies of vertical-profile data was given by Serra (1985). Rider (1978) and Shields (1974) discussed the use of the dipmeter for interpreting depositional environments and crossbedding. The difficulty is that few of the possible patterns have a unique interpretation and, without detailed local lithofacies and structural information from cores, outcrops, or seismic data, it may not be possible to make effective use of the information. Many of the references listed above suggest using the dipmeter to recognize crossbedding patterns but, until the development of microresistivity devices, this was strictly hypothetical, because log sensitivity was adequate only to see the largest crossbed structures, such as eolian draas. Rider (1978) pointed out that crossbedding may contain several kinds of dipping surfaces (foresets, set bounding planes, reactivation surfaces, scours) and that dipmeter sampling through a typical crossbedded interval with set thicknesses in the order of a few decimeters may produce very scattered and unintelligible patterns. However, where the dipmeter can be correlated with a core, a powerful paleocurrent-analysis tool may be available, as discussed in Section 5.6.6. Very little published work of this type is available, possibly because of its confidential nature. Fault and fold interpretations are not discussed here, for reasons of space.

Jageler and Matuszak (1972) gave an excellent example of the use of the dipmeter in delineating a reef, based on the depositional dips of the reef itself and drape over the resistant reef mass. Figure 5.29 illustrates two structural-contour maps, one drawn from well information alone and one using dipmeter information to control local contour orientation and spacing. The reef was first discovered when well no. 1 was drilled, revealing a 35° dip, probably representing the windward talus slope.

Dipmeter readings commonly show a regular upward or downward increase in angle while maintaining constant azimuth. Examples are illustrated in Fig. 5.30. An upward increase in dip angle over a thickness of a few tens of meters may indicate a fault or surface creep below an unconformity surface (Fig. 5.30 D). An upward decrease in dip has a greater variety of causes. Sediments onlapping an irregular erosion surface, draped over a reef (Fig. 5.30 C) or sandstone pod (Fig. 5.30 B), or infilling a channel or valley (Fig. 5.30 A) may show similar patterns. Associated data, such as the random readings within the

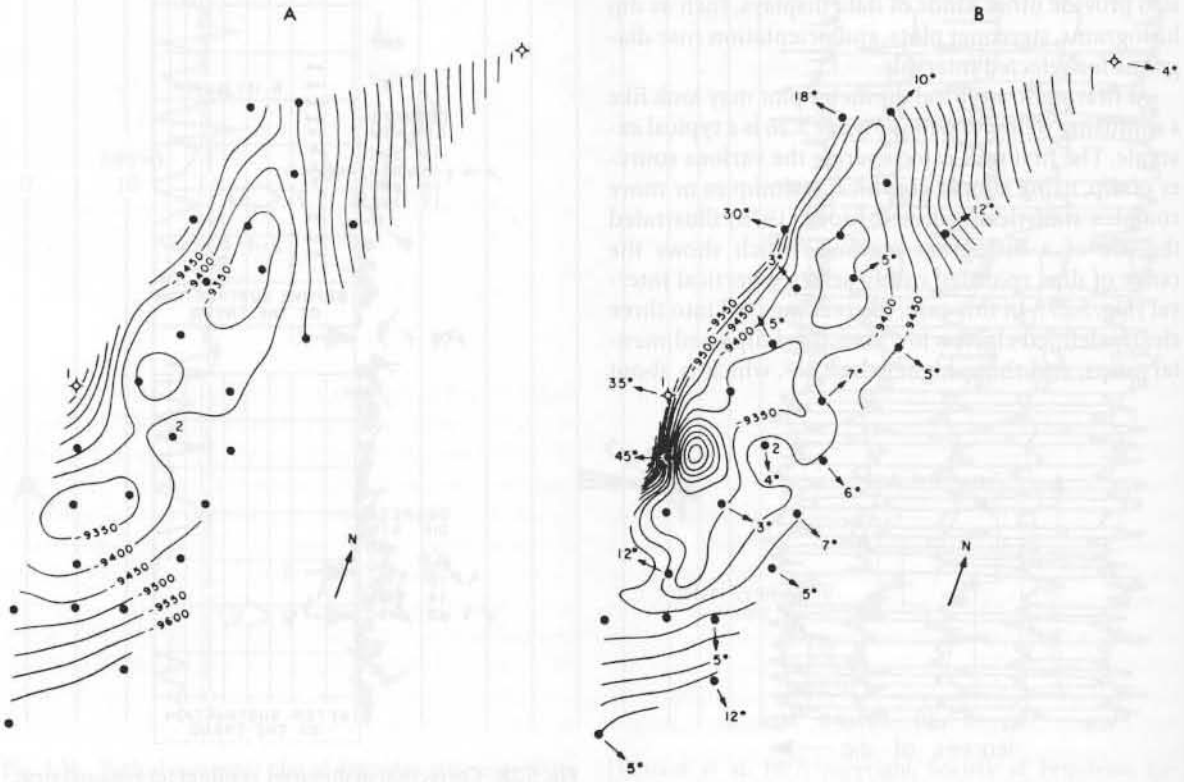


Fig. 5.29. Use of a dipmeter to correct and refine a structure-contour map of a reef body, west Texas. A Map drawn using well data only. B Dipmeter readings have been added and the map replotted. (Jageler and Matuszak 1972)

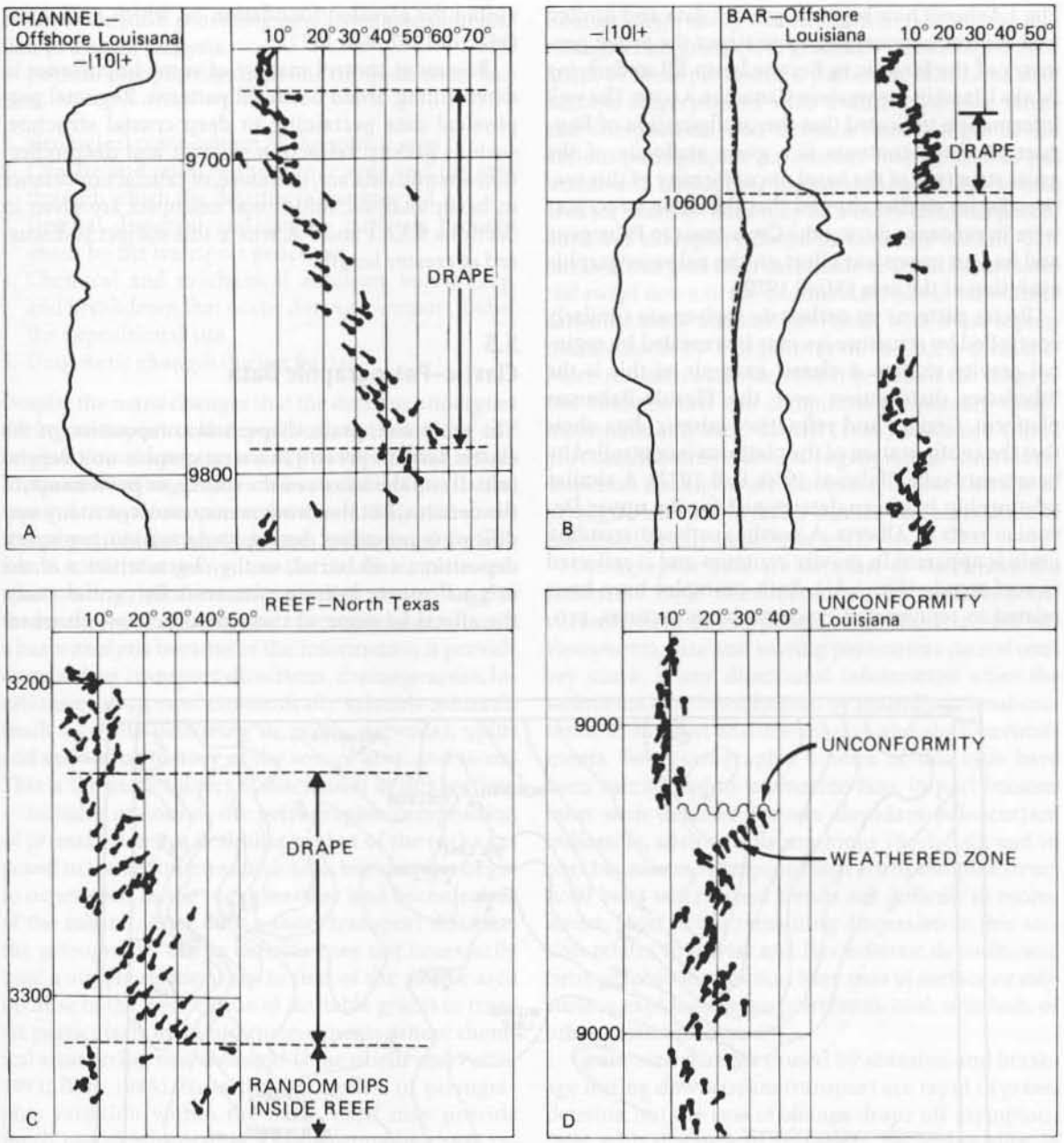


Fig. 5.30 A – D. Examples of typical sedimentary-dipmeter patterns. **A** Drape in a channel fill. **B** Drape over a bar deposit. **C** Drape over a reef body and random dips within the reef. **D** Dip distortion by surface creep in a weathering zone. (Gilreath 1977)

reef (Fig. 5.30C), crossbedding (Fig. 5.30B), and the petrophysical-log character and lithology, should all be used to discriminate these patterns.

Davison and Haszeldine (1984) and Nelson et al. (1987) discussed the use of dipmeter and other data for orienting cores. A dipmeter that has been correlated in detail with a core is a very valuable source of data for facies and paleocurrent interpretation. The use of the dipmeter in paleocurrent analysis is described in Section 5.6.6.

5.4.4 Other Geophysical Methods

A wide variety of other geophysical techniques is used in basin-stratigraphic and facies mapping. A few are discussed here, but the subject is a large one and cannot be treated fully in this book (Sharma 1986, for a general reference).

Gravity and magnetic data are commonly used to delineate broad basin configuration. For example,

Fig. 1.4 shows how regional gravity data and limited well control were used to reconstruct the broad geometry of the Jurassic to Eocene basin fill underlying Banks Island in the western Canadian Arctic. The well information indicated that the configuration of Bouguer anomaly contours is a good analogue of the gross structure of the basal unconformity of this section. Facies studies showed that the same structures were in existence during the Cretaceous to Paleogene and had an important effect on the paleogeographic evolution of the area (Miall 1979).

Facies patterns on carbonate shelves are similarly controlled by structure, as may be revealed by regional gravity studies. A classic example of this is the lithofacies distribution over the Florida-Bahamas platform. Gravity and refraction-seismic data show that the configuration of the platform is controlled by basement faults (Talwani 1960; Ball 1972). A similar relationship has been determined for the upper Devonian reefs of Alberta. A north–northeast-trending grain is apparent in gravity contours and is reflected in reef trends (Fig. 5.31). Both examples have been related to rejuvenated Precambrian structures, providing the elevated foundation on which reefs grow (Martin 1967; Ball 1972).

Basement control may be of some importance in determining broad basin-fill patterns. Regional geophysical data pertaining to deep crustal structure, such as gravity, refraction seismic, and deep-reflection seismic data are, therefore, of crucial importance in basin analysis. Additional examples are given in Sections 9.3.6.1 and 7.8, where this subject is discussed at greater length.

5.5 Clastic–Petrographic Data

The grain size, grain shape, and composition of the clastic detritus present in a stratigraphic unit depend initially on the nature of the source, or **provenance**, of the detritus, but the detritus may undergo many modification processes during its derivation, transport, deposition, and burial, so the characteristics of the original source become obscured. By careful study, the effects of some of these modification processes

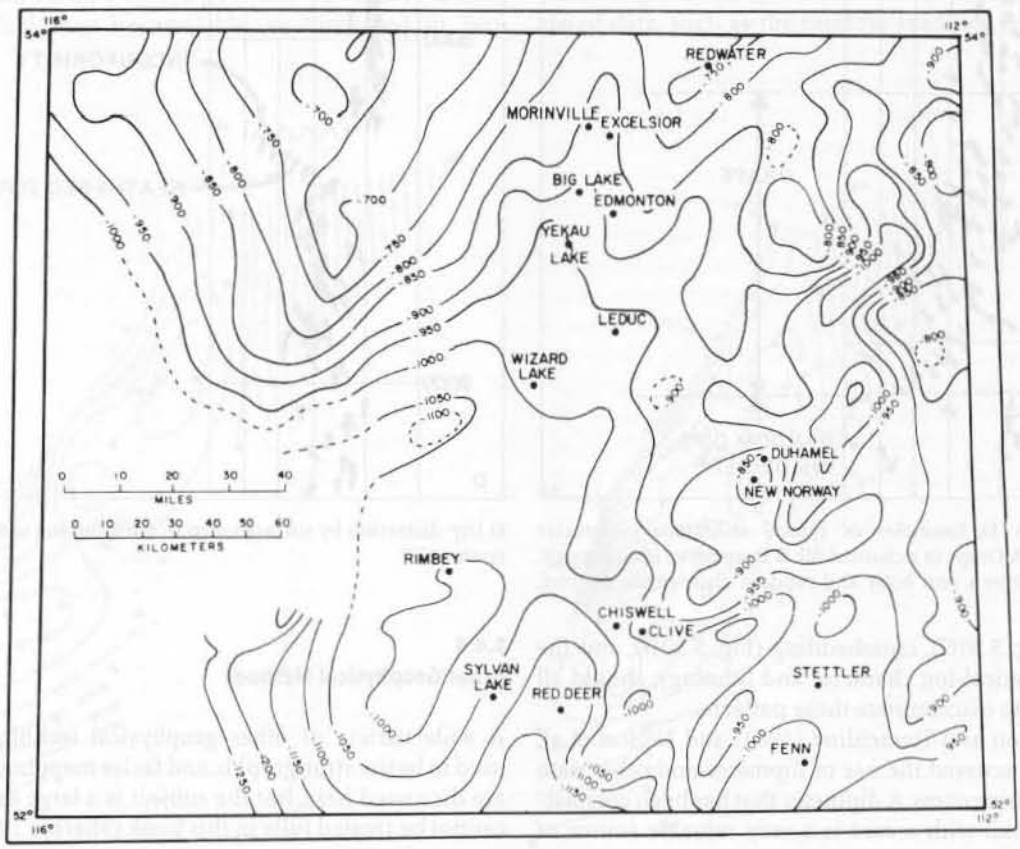


Fig. 5.31. Bouguer gravity-anomaly map of central Alberta showing a north–northeasterly grain reflecting structure in the Precambrian basement. Most Devonian reefs are located on or near gravity highs (e.g., Redwater, Leduc, Duhamel, New Norway, Clive, Sylvan Lake, and Wizard Lake). (Ball 1972)

can be identified, providing much valuable information to a basin analysis.

Detrital composition depends on these main factors:

1. Geological composition of the source area
2. Climate and relief of the source area and the area through which the detritus is transported
3. Detrital dispersal and mixing patterns brought about by the transport processes
4. Chemical and mechanical abrasion, winnowing, and breakdown that occur during transport and at the depositional site
5. Diagenetic changes during burial

Despite the many changes that the detritus undergoes during transport and deposition, the nature of the original source area can usually be identified. Certain compositional ranges are characteristic of particular types of tectonic source environment. This is the subject of **petrofacies analysis** and is discussed in Section 9.4. Without going into the details of plate-tectonic setting, information regarding the geology of the local source area may be of considerable value in a basin analysis because of the information it provides regarding transport directions, drainage areas, locations of sources of economically valuable minerals (such as those occurring in placer deposits), uplift and **unroofing history** of the source area, and so on. This is the main subject of discussion in this section.

Initially, of course, the petrographic composition of primary detritus is similar to that of the rocks exposed in the sediment source area, but changes begin to occur even as the rocks weather and become part of the subsoil. After only a short transport distance, the petrography of the detritus may not necessarily bear a simple relationship to that of the source area because of the destruction of unstable grains in transit, particularly in humid environments, where chemical weathering may be rapid (Franzinelli and Potter 1983; Basu 1985). Nevertheless, studies of petrographic variation within the basin itself may provide much useful information. Downcutting by rivers reveals progressively older rocks, so the derived sediments may display an **inverted stratigraphy**, older sediments containing detritus from the young cover rocks and the upper strata containing material derived from the local basement (Graham et al. 1986). Documentation of this unroofing sequence may help clarify the timing of orogeny, volcanism, plutonism, rates of uplift, etc. (Cerveny et al. 1988). Areal variations in petrographic composition depend on transport directions which, in turn, reflect tectonic control of the paleoslope. As clastic sediments are dispersed by currents or sediment-gravity flows, they are subjected to size sorting, abrasion, breakage, and rounding. They may retain a distinctive detrital composition, although easily weathered minerals, such as ferromagnesian grains, will gradually be destroyed. As a result of these processes, a particular lithosome may become fingerprinted with a distinctive size range and composition, and downcurrent changes in the various petrographic parameters may be used as paleocurrent indicators. Interpretation of the source area for marine rocks may be a more complex procedure. For example, submarine fans may contain deltaic detritus plus shelf carbonate or siliciclastic material swept down to the continental rise via submarine canyons. Many authors have dealt with these topics; the textbook by Pettijohn et al. (1973) is a classic. More recently, McBride (1985) reviewed the diagenetic changes that can complicate provenance determinations, and Ricci-Lucchi (1985) discussed transport considerations and paleogeographic controls on dispersal. Examples are discussed below. The use of petrographic data in facies studies and correlation is discussed briefly in Section 4.5.2.

Petrographic data are simplest to use where the sediments have been deposited by unidirectional currents, as in fluvial, deltaic, and submarine-fan environments. Size and sorting parameters do not convey much, if any, directional information when the sediments have been formed by multidirectional currents, as in most shallow-marine and shelf environments. Few petrographic studies of this type have been carried out in submarine fans, in part because most such deposits contain abundant paleocurrent indicators, such as sole markings (Sect. 5.6), and in part because most exposed fans are in complex structural belts where areal trends are difficult to reconstruct. Most of the remaining discussion in this section relates to fluvial and fluviodeltaic deposits, and petrographic studies may have uses in surface or subsurface exploration for petroleum, coal, uranium, or other economic deposits.

Grain-size changes caused by abrasion and breakage during downstream transport are rapid in gravel detritus, but the rate of change drops off asymptotically with distance of transport, becoming slow in coarse sand and negligible in fine sand to silt. Much experimental work has been carried out in an attempt to quantify the rate of change so that grain-size in the field could be interpreted directly in terms of distance of transport. However, each detrital lithology responds differently to current transport, and rivers vary markedly in competency at any one point, so research of this type is unlikely to meet with any success. Nevertheless, mapping of relative grain-size changes in conglomeratic and coarse, sandy sediments has considerable use as a paleocurrent indicator.

What measurements should be made? For grain-size studies in conglomeratic rocks, the simplest and

most useful mapping parameter is maximum clast size, which is usually given as the average of the intermediate clast diameter of the 10 largest clasts present at a sample point. For sand-sized grains, mean or maximum grain-size may be used. Both can be measured in an outcrop or in a thin section by visual estimation against a grain-size chart (Fig. 2.5). This permits measurements to the nearest 0.5 ϕ, which is quite accurate enough for our purposes (sieve analyses and other laboratory studies are time consuming and usually add little).

Detrital composition of conglomerates can be studied by counting and identifying 100 or more clasts at each sample point. Sandstone composition should be studied in thin section using a moveable microscope stage and a point counter. It is recommended that at least 300 grains be identified and counted per sample point. The stage should be set to advance at each count by an amount equal to or slightly greater than the mean grain diameter. The number of points counted can be adjusted according to the compositional variability of the sample. Pure quartz arenites do not need such intensive study, but immature, lithic arenites with many varieties of rock fragments may need more detailed examination. Care should be taken to recognize and distinguish **extrabasinal** and **intrabasinal** detritus, particularly in the case of carbonate and volcanic fragments. Zuffa (1985b, 1987) discussed criteria for distinguishing these classes of particles. Mineral types have different densities, so they vary in their hydrodynamic behavior. They also have different fragmentation patterns. The detrital composition of a clastic rock may, therefore, vary markedly with grain size (Fig. 5.32), so, for mapping purposes, it is advisable to control measurements according to a standard grain-size range, for example 1ϕ class. These methods will usually result in a measurement repeatability of two or three percentage points or better, and this, again, is accurate enough for our purposes. Analytical methods are discussed in detail by Dickinson (1970), Pettijohn et al. (1973), and Zuffa (1985a,b, 1987). It is important to standardize laboratory procedures in order to facilitate comparisons between data sets. Zuffa (1985b, 1987) and Ingersoll et al. (1984) recommended use of the so-called **Gazzi-Dickinson point-counting method**, which counts coarse grains (>0.0625 mm) in terms of their component mineral grains rather than as rock fragments (Fig. 5.33). The method also requires that the observer attempt to reconstruct the original composition of altered grains. This leads to a better clustering of data points and increased accuracy in provenance determinations. However, there is not universal agreement that the Gazzi-Dickinson method is suitable for all purposes. For example, Suttner and others (1985), in two separate discussions of the paper by Ingersoll et al. (1984), pointed out that the method of classifying coarse grains results in a loss of certain types of data, such as information regarding lithic grains. Careful study of these discussions and the replies by Ingersoll and his colleagues is recommended for anyone contemplating detailed petrographic work. Howard (1993) discussed the statistics of clast counting, sources of error, and field procedures.

Much use has been made of heavy minerals (density greater than about 2.8) in provenance studies, even though most sands typically contain less than 1% by weight. They are extracted from loose or disaggregated sand by separation in dense liquids, such as bromoform, and examined in grain mounts. Distribution is affected by hydraulic sorting during transport and by intrastratal solution following deposition (Morton 1985a). Considerable skill is needed to identify grain types correctly, and the technique is, in general, more difficult to use than thin-section analysis of the light fraction.

Grain-size and compositional measurements may be averaged over a vertical section at each mapping

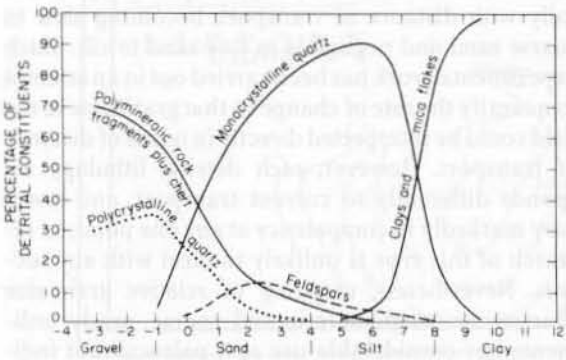


Fig. 5.32. Relationship between grain size and composition of the detrital fraction in siliciclastic sediments. (Blatt et al. 1980, reprinted with permission of Prentice-Hall Inc.)

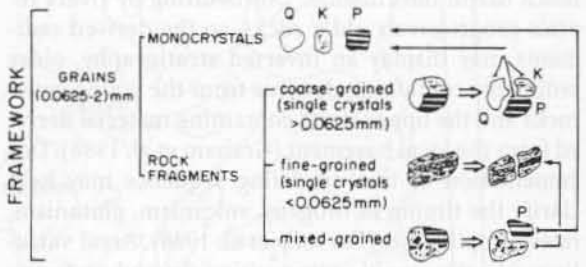


Fig. 5.33. Flow chart for the point-counting technique (Gazzi–Dickinson method) used to minimize the dependence of rock composition on grain size. Coarse rock fragments are counted in terms of their constituent mineralogical components wherever possible. *Q* Quartz; *K* potassium feldspar; *P* plagioclase feldspar. (Zuffa 1985b)

location or restricted to individual, correlatable stratigraphic units. Plotting of both areal and vertical changes may provide useful paleogeographic information, and it is up to the basin analyst to exploit the data in the most effective way. Petrographic data may be analyzed manually, graphically, or statistically to define petrographic assemblages or provinces typifying a particular source-area geology (Suttner 1974). Alternatively, particular mineral species may be treated separately.

Some workers have concentrated on identifying and mapping varieties of quartz, feldspar, zircon, or other grain types, requiring a great deal of highly specialized research but possibly providing provenance information unobtainable in any other way (Allen 1972; Heim 1974; Schnitzer 1977; Helmold 1985). Sophisticated analytical techniques are increasingly being used to define mineral varieties based on trace-element composition (Matter and Ramseyer 1985; Morton 1985a,b). For example, Morton (1985b) employed an electron microprobe to discriminate three populations of detrital garnets in a stratigraphic unit and related them to probable sources. Heller and Frost (1988) explored a huge new area: the study of stable- and radiogenic-isotope distributions in detrital grains, and their use in fingerprinting source terranes.

Resistant grains, such as quartz and chert, plus the heavy minerals zircon, tourmaline, rutile, and garnet, are commonly recycled through several episodes of erosional derivation, sedimentation, lithification, uplift, and re-erosion. Usually, they show abundant evidence of this in the form of a high degree of rounding and the presence of authigenic overgrowths. It is usually difficult or impossible to determine the sources of such grains because, in most cases, they could have come from several different stratigraphic units in the source area. Variations in color, crystal habit, or inclusions may help but require more detailed research (Zuffa 1985b, 1987).

Extensive treatments of sandstone petrology and its application in sedimentological studies are given by Pettijohn et al. (1973) and Blatt et al. (1980). Zuffa (1985a) edited a useful compilation of research and review papers, several of which are quoted above. In the remainder of this section, some examples of field studies are discussed to illustrate the kind of contribution petrography can make to regional basin analysis.

Friend and Moody-Stuart (1972) studied the Devonian fluvial sediments of Spitsbergen and drew a variety of lithofacies and petrographic maps. Figures 5.34 and 5.35 illustrate two of these. Mean grain size was calculated for each locality by averaging the grain-size values obtained at 25-cm intervals up each section. The rocks were divided into three stratigraphic units, and a separate map was drawn for each (Fig. 5.34). The data points were contoured using a trend-surface analysis method (Sect. 5.8). Figure 5.35 shows detrital orthoclase content in the sandstones, based on measurements in over 200 thin sections. Statistical studies by the authors showed that part,

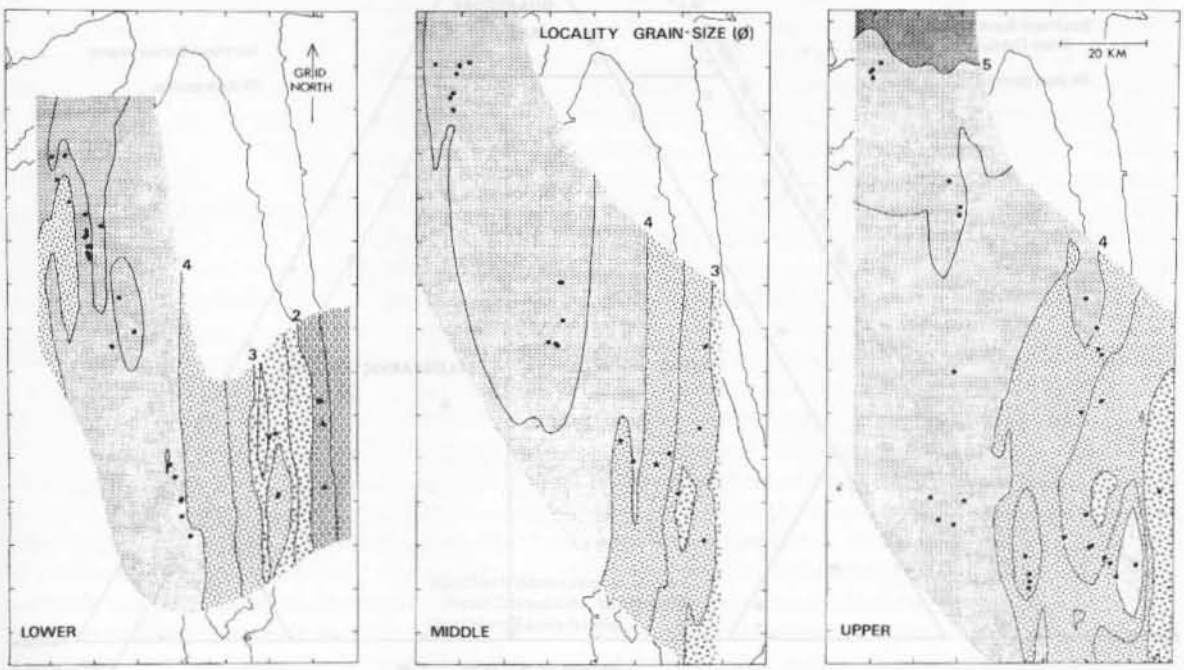


Fig. 5.34. Locality mean grain size for each of three stratigraphic subdivisions of the fluvial Wood Bay Formation (Devonian), Spitsbergen, contoured by trend-surface analysis. (Friend and Moody-Stuart 1972)

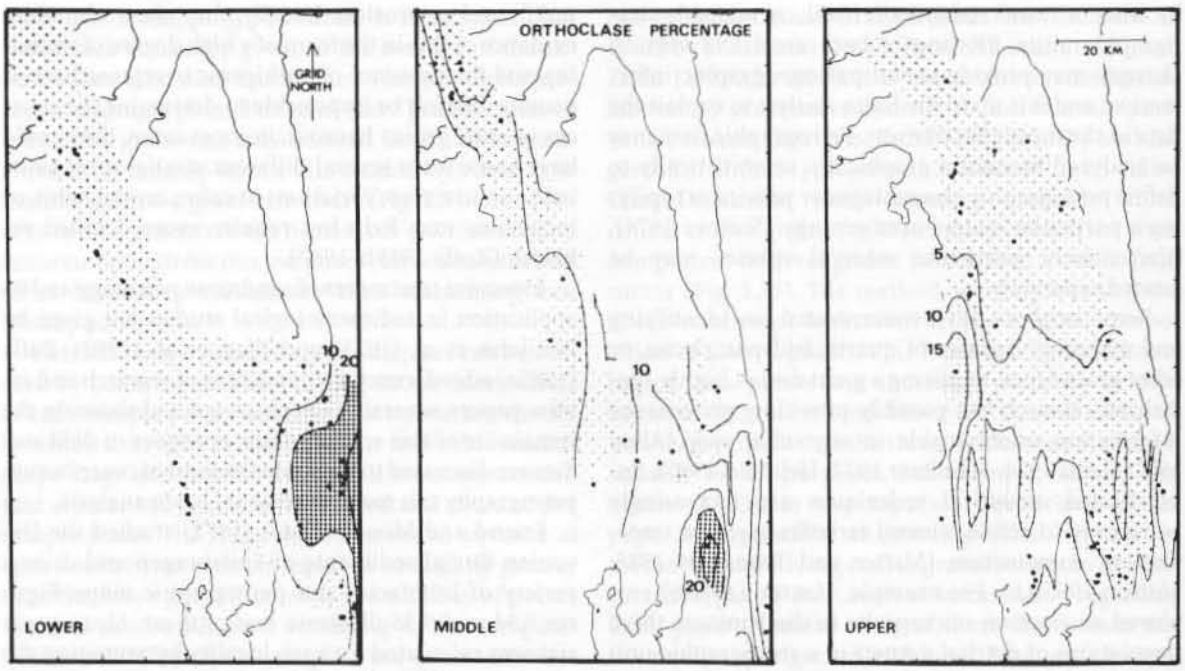


Fig. 5.35. Orthoclase content, plotted and contoured as in Fig. 5.33. (Friend and Moody-Stuart 1972)

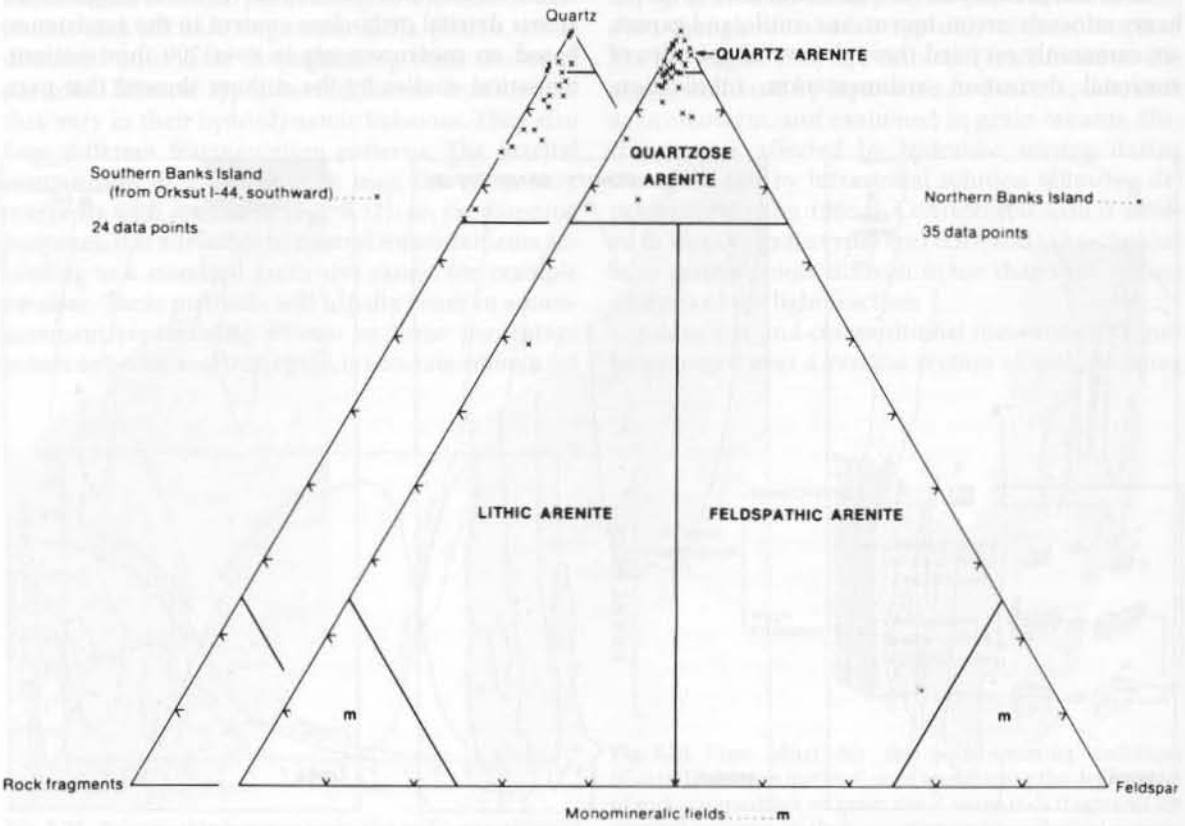


Fig. 5.36. Detrital petrography of the fluvial Isachsen Formation (lower Cretaceous) of Banks Island, Arctic Canada. The two plots are for the north and south halves of the island. The sandstone classification is that of Okada (1971). (Miall 1979)

but not all, of the variation in orthoclase content could be explained as being due to differences in grain size – the feldspar tends to be concentrated in the coarser sandstones. What all these maps show is that sediments were fed into the basin from the west, south, and southeast, with river systems flowing generally northward.

Miall (1979) analyzed the petrography of the Isachen Formation, a lower Cretaceous braided-river deposit in Banks Basin, Arctic Canada. The bulk of the formation consists of texturally and mineralogically mature, medium to coarse sandstone showing abundant evidence of a recycled origin. Paleocurrent evidence indicates northward-flowing trunk rivers with tributaries entering from the east (the west side of the basin is covered by younger strata but may also have been a source area). Initial attempts to recognize petrographic provinces in the sandstone were unsuccessful. Ternary plots show that the samples from the north and south ends of the basin contain essentially the same mineral composition (Fig. 5.36). Cluster analysis revealed a weak trend for feldspar to be more common in the north and sedimentary rock fragments are slightly more abundant in the south. This was then emphasized by replotting the thin-section data in terms of three minor components (Fig. 5.37); these plots show that chert is also slightly more abundant in the north. It was found that a few rare components, such as detrital quartz–feldspar intergrowths, quartz with mica, tourmaline, or zircon inclusions, and zoned quartz, are all present almost exclusively in the north end of the basin. Analyses of these data in terms of potential sediment sources resulted in the map shown in Fig. 5.38. The bulk of the quartz sand and the sedimentary-rock fragments were derived from mature Proterozoic sediments, chert was fed into the north end of the basin from various lower Paleozoic sources, and the feldspar plus rare quartz and feldspar components appear to have been derived from a small Archean granodiorite pluton. The Melville Island Group is a middle to upper Devonian sandstone unit that outcropped over wide areas adjacent to the north end of the basin but appears to have contributed little detritus. It is too fine grained and has the wrong texture and mineralogy.

Rahmani and Lerbekmo (1975) carried out a detailed heavy-mineral analysis of the Cretaceous–Paleocene molasse sandstones derived from the Cana-

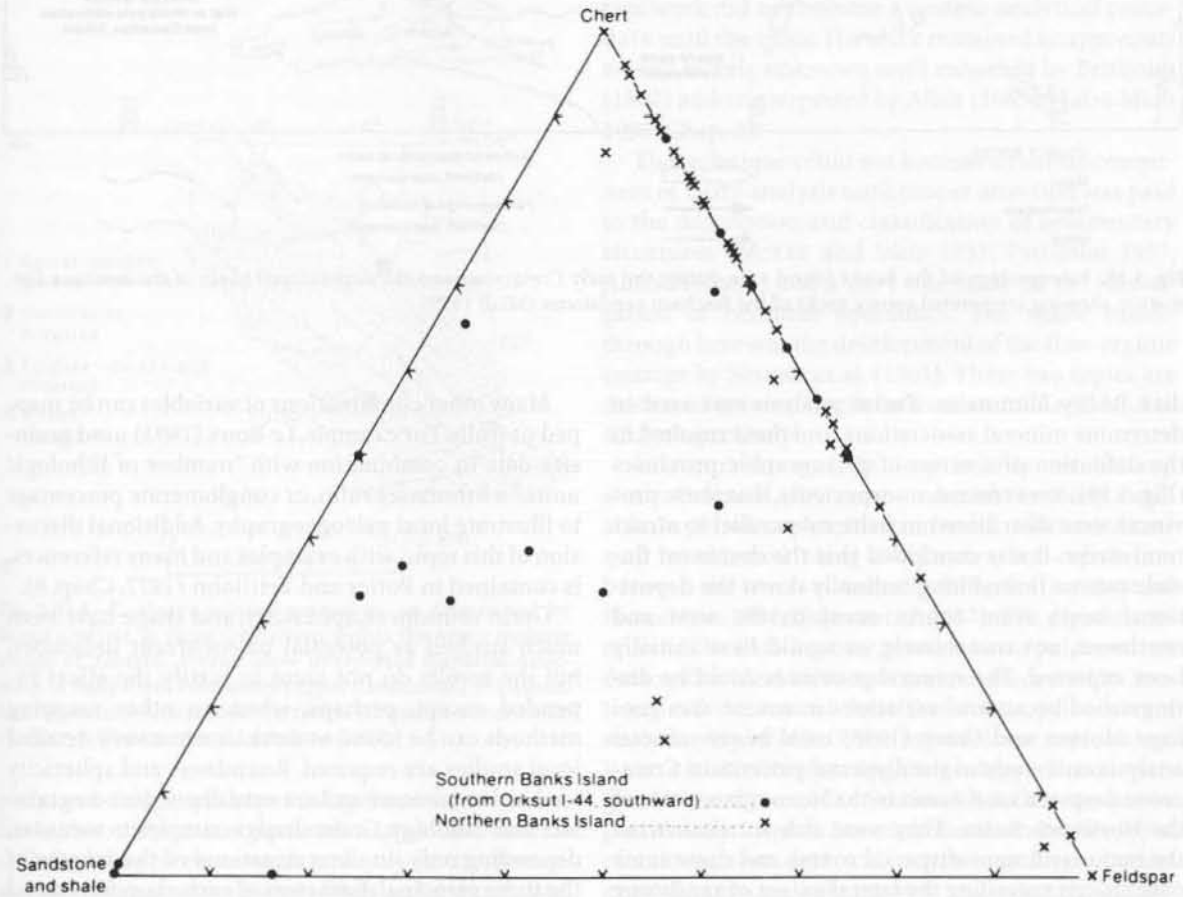


Fig. 5.37. Distribution of minor components in the Isachsen Formation, Banks Island (Miall 1979)

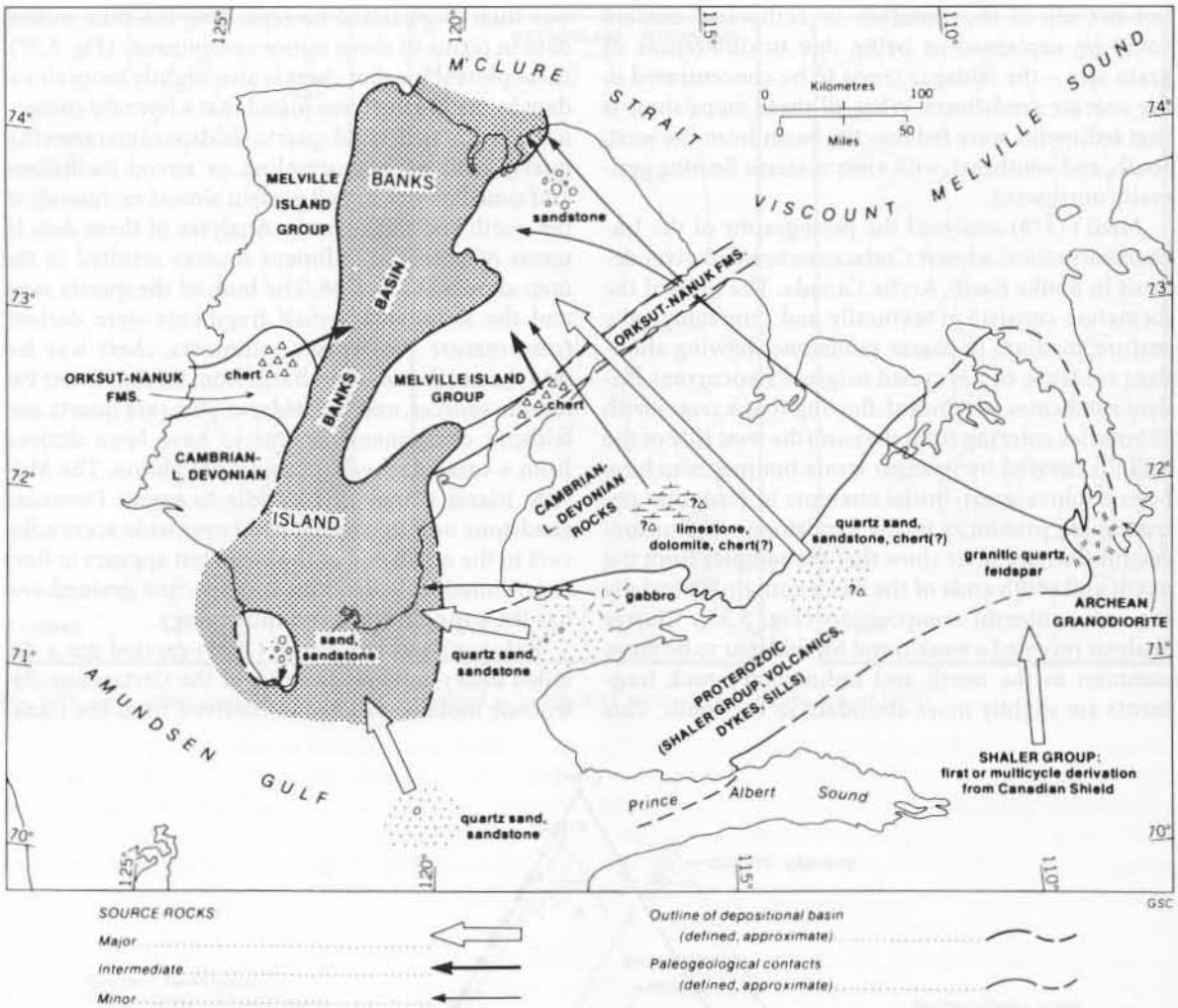


Fig. 5.38. Paleogeology of the Banks Island area during the early Cretaceous and the depositional basin of the Isachsen Formation, showing interpreted source rocks of the Isachsen sandstones (Miall 1979)

dian Rocky Mountains. Factor analysis was used to determine mineral associations, and these resulted in the definition of a series of petrographic provinces (Fig. 5.39). It was found, unexpectedly, that these provinces were distributed in belts subparallel to structural strike. It was concluded that the dominant fluvial systems flowed longitudinally down the depositional basin from source areas to the west and northwest, not transversely, as would have initially been expected. The mineral provinces could be distinguished because of variations in source-area geology. Morton and Grant (1998) used heavy-mineral analysis on a study of the dispersal patterns of Cretaceous deep-sea sandstones in the Norwegian sector of the North Sea Basin. They were able to reconstruct the major sediment-dispersal routes and draw some conclusions regarding the lateral extent of sandstone bodies.

Many other combinations of variables can be mapped usefully. For example, Le Roux (1993) used grain-size data in combination with "number of lithologic units," a lithofacies ratio, or conglomerate percentage to illustrate local paleogeography. Additional discussion of this topic, with examples and many references, is contained in Potter and Pettijohn (1977, Chap. 8).

Grain roundness, sphericity, and shape have been much studied as potential paleocurrent indicators, but the results do not seem to justify the effort expended except, perhaps, when no other mapping methods can be found to work or when very detailed local studies are required. Roundness and sphericity are hard to measure and are very dependent on grain-size and lithology. Grain shape is simpler to measure, depending only on determinations of the lengths of the three principal diameters of each clast. Sneed and Folk (1958) recognized various classes of compact,

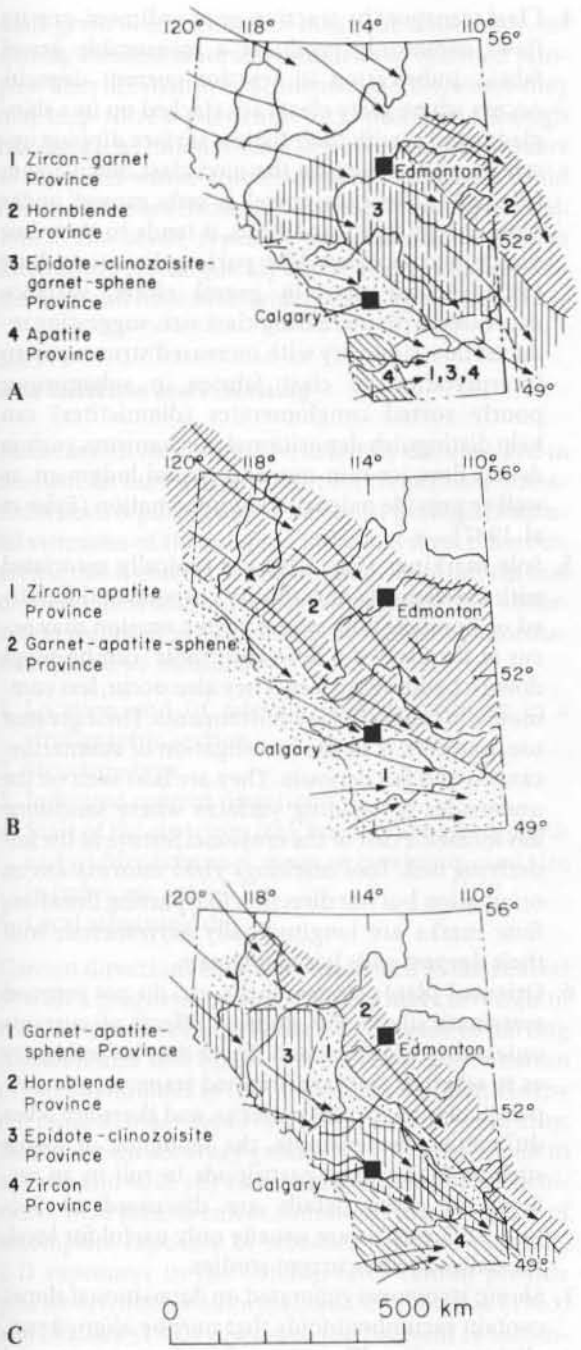


Fig. 5.39 A–C. Heavy-mineral provinces, as determined by factor analysis, in three formations within the upper molasse wedge of Alberta. *Arrows* show interpreted dispersal directions. **A** Belly River Formation (upper Cretaceous). **B** Edmonton Formation (Cretaceous–Tertiary). **C** Paskapoo Formation (Tertiary). (Rahmani and Lerbekmo 1975)

platy, and elongated shapes using these data. It seems likely that only in conglomerates could the data be collected accurately enough and only in conglomerates are downstream changes likely to be significant. For example, Bluck (1965) found that, in the deposits of a small Triassic alluvial fan, there was a downstream increase in the proportion of platy clasts, because these offered flatter surfaces for more effective sediment transport. Ehrlich and Weinberg (1970) proposed the use of automated Fourier-analysis methods for grain-shape characterization, with promising results.

5.6 Paleocurrent Analysis

5.6.1 Introduction

The art of paleocurrent analysis was invented by the quintessential 19th-century English amateur Henry Clifton Sorby, who published his first paper on the subject in 1852. He understood the formation of "ripple drifted" and "drift bedded" structures (climbing ripples and large-scale crossbedding, respectively) and, in seven years, claimed to have made about 20,000 observations, a total that has probably never been equaled. Most of Sorby's work remained unpublished. He was far ahead of his time, as paleocurrent work did not become a routine analytical procedure until the 1950s. His work remained unappreciated and largely unknown until exhumed by Pettijohn (1962) and reinterpreted by Allen (1963b) (also Miall 1996; Chap. 2).

The technique could not become a routine component of basin analysis until proper attention was paid to the description and classification of sedimentary structures (McKee and Weir 1953; Pettijohn 1957; Allen 1963a, b) and a start was made on the investigation of bedform hydraulics. The major breakthrough here was the development of the flow-regime concept by Simons et al. (1965). These two topics are discussed at some length in Sections 2.2.3 and 4.5.4 to 4.5.5, respectively.

Paleocurrent analysis is primarily an outcrop study. Oriented core can be used but is rarely available. In petroleum-exploration wells, unoriented core may be oriented using a dipmeter, but correlation between the two is commonly difficult (Davison and Haszeldine 1984; Nelson et al. 1987). The dipmeter itself and the FMS log may provide some paleocurrent data under ideal circumstances, as discussed in Section 5.6.6. Because of these limitations, paleocurrent analysis may be of little use in petroleum exploration. It can assist in interpreting the geology of basin margins but, as discussed in Section 1.4.1, this may not be particularly helpful. For coal, metal, and other types of mining that deal with surface or large, underground outcrops, paleocurrent analysis may, in contrast, be extremely useful.

The technique can provide information on four main aspects of basin development:

1. The direction of local or regional paleoslope, which reflects tectonic-subsidence patterns
2. The direction of sediment supply
3. The geometry and trend of lithologic units
4. The depositional environment

Examples of these applications are discussed in the ensuing paragraphs.

5.6.2 Types of Paleocurrent Indicators

Sedimentary structures and fabrics used in facies and paleocurrent studies are described and illustrated in Section 2.2.3. The following notes explain briefly how they yield current directions.

1. For ripple marks and crossbedding (Fig. 2.12), the inclination of foreset directions is generally downcurrent because of the grain-avalanching mechanism. Needless to say, it is not always this simple. Smith (1972) demonstrated that planar crossbed sets in rivers commonly advance obliquely to the flow direction. Trough crossbeds do not yield accurate flow directions unless the analyst can observe the orientation of the trough axis (see below). Hunter (1981) discussed the flow of air around large eolian dunes and suggested that these, too, commonly advance obliquely to the wind direction. The use of statistical procedures and appropriate facies data usually circumvents these problems (this is also discussed below).
2. Channels and scours (Fig. 2.15) occur in many environments and may indicate the orientation of major erosive currents, such as those generating river or tidal channels and delta or submarine-fan distributaries. However, the larger channels, which are those most likely to be of regional significance, are usually too large to be preserved in outcrops. They may be readily apparent on seismic records, as discussed in Chapter 6.
3. Parting lineation or primary-current lineation, the product of plane-bed flow conditions (Fig. 2.14), is only visible on bedding-plane exposures. It usually yields directional readings of low variance, because it forms during high-energy flow in river, delta, or tidal channels, when bars or other obstructions to flow are under water and flow sinuosity is low. The structure indicates orientation but not direction of flow because of the ambiguity between two equally possible readings at 180° to each other. Usually, this can be resolved with reference to other structures nearby.
4. Clast transport by traction or in sediment-gravity flows commonly produces a measurable gravel fabric. Imbrication in traction-current deposits occurs where platy clasts are stacked up in a shingled pattern, with their flattest surface dipping upstream and resting on the next clast downstream (Fig. 2.13). Because gravel is only moved under high-energy flow conditions, it tends to show low directional variance like parting lineation. Rust (1975) found that, in gravel rivers, variance decreased with increasing clast size, suggesting reduced flow sinuosity with increased stream power. Interpretation of clast fabrics in subaqueous, poorly sorted conglomerates (diamictites) can help distinguish depositional mechanisms, such as debris flow, ice rain-out, and glacial lodgment, as well as provide paleocurrent information (Eyles et al. 1987).
5. Sole markings (Fig. 2.16) are typically associated with the deposits of turbidity currents and fluidized or liquified flow, where vortex erosion may occur at the base of a flow, and "tools" can be swept down a bedding surface. They also occur, less commonly, in other clastic environments. Their greatest use, however, is in the investigation of submarine-canyon and fan deposits. They are best seen on the undersides of bedding surfaces where sandstone has formed a cast of the erosional feature in the underlying bed. Tool markings yield information on orientation but not direction like parting lineation; flute marks are longitudinally asymmetric, with their deepest ends lying upstream.
6. Oriented plants, bones, shells, etc. do not respond systematically to the aligning effects of currents unless they are elongated. There may be ambiguity as to whether they are oriented transverse or perpendicular to current patterns, and there are other difficulties, for example, the tendency of fossils such as high-spired gastropods to roll in an arc. Some of these details are discussed in Section 5.6.5. Fossils are usually only useful for local, specialized paleocurrent studies.
7. Slump structures generated on depositional slopes contain recumbent folds that may be aligned parallel to strike. They are, therefore, a potential paleoslope indicator, particularly on the relatively steep slopes of prodeltas and actively prograding continental margins. Friction at the margin of the slump may cause it to rotate out of strike alignment, so a statistical approach to measurement is desirable. Potter and Pettijohn (1977, Chap. 6), Rupke (1978), and Woodcock (1979) reviewed their use as paleoslope indicators.

Several other paleocurrent indicators are reviewed by Potter and Pettijohn (1977), including some, such as

sand-grain orientation and magnetic anisotropy, requiring detailed laboratory analysis of oriented samples. They are mainly of academic interest, where they may help solve a particular local problem, although laboratory techniques have improved in recent years to the point where studies of magnetic fabric could become routine (Hrouda and Stránik 1985; Eyles et al. 1987). The seven types of paleocurrent indicators listed above can be quickly and routinely measured in the field and have been widely used in basin studies.

5.6.3 Data Collection and Processing

Paleocurrent data should be carefully documented in the field. In the past, paleocurrent trends have sometimes been reported on the basis of a geologist's mental estimates of the range of indicated directions but, even if this is correct, it may result in a significant loss of useful information. For every paleocurrent observation recorded in field notes, the following information should be included:

1. Location and (if relevant) precise position in a stratigraphic section
2. Structure type
3. Indicated current direction
4. Scale of the structure (thickness of crossbed, depth and width of channel, mean or maximum clast size in imbricate gravels)
5. Local structural dip

Current directions should be measured to the nearest 5° with a magnetic or sun compass and corrected to true north wherever necessary. In the case of parting lineation and tool markings, the correct orientation of two possibilities at 180° can usually be identified by referring to other types of current structures nearby. Measurement accuracy greater than ± 5° is difficult to achieve and is, in any case, rarely important. Difficulties in field measurement commonly arise because of incomplete exposure of crossbed sets. For example, 2-D exposures in flat outcrop faces cannot provide precise orientation information. DeCelles et al. (1983) and Bradley (1983) discussed the problem of incomplete or partial exposure of trough crossbed sets and ripple crosslamination, and offered some solutions involving field measurement techniques and statistical treatment of the data. Indicated current directions may need to be corrected for structural dip and fold plunge, otherwise significant directional distortions may result. This should be carried out as soon as possible, preferably after every day's work, in order that possible errors can be detected while there is still time to rectify them in the field. For linear structures, such as parting lineation or sole markings, structural dips as high as 30° can be safely ignored, as they result in errors of less than 4°. However, the foreset-dip orientation of crossbedding is significantly affected by structural dip, and should be corrected wherever the structural dip exceeds 10°. This subject is discussed further by Potter and Pettijohn (1977, Chap. 10), who illustrated a correction technique using a stereonet. Ramsay (1961) provided graphical solutions. Parks (1970) presented computer routines for the necessary trigonometric calculations. Wells (1988) provided a practical guide to field procedures and recommended the use of a pocket calculator to perform structural corrections and data synthesis in the field.

Sooner or later the question will arise: how many readings should be made? There is no single or simple answer to this problem, because it depends on how many measurable current indicators are available for observation and what the objectives of the study are. Olson and Potter (1954) discussed the use of grid-sampling procedures and the random selection of structures to measure, followed by calculation of reliability estimators to determine how many readings are necessary in order to be sure of determining correct directional trends. In this study, they were concerned only with determining regional paleoslopes. We now understand a great deal about air and water flow patterns in different depositional environments, and a case could be made for measuring and recording every visible sedimentary structure. Such detailed data can be immensely useful in amplifying environmental interpretations and clarifying local problems, as discussed in Section 5.6.5. Some selection may have to be made in areas of particularly good exposure. A practical compromise is to record every available structure along measured stratigraphic sections and to fill in the gaps between sections with spot (gridded or random) samples. This procedure permits the elaboration of both local and regional paleocurrent trends. When constructing lateral profiles of large outcrops, it is essential to document these with abundant paleocurrent determinations. Such data provide the third dimension to a 2-D outcrop and yield essential information regarding the shape and orientation of architectural elements (Sect. 4.5.9).

If the trend itself is important, 25 readings per sample station is commonly regarded as the minimum necessary for statistically significant small samples. However, the same or fewer readings plotted in map or section form can yield a great deal of environmental detail, whether or not their mean direction turns out to be statistically significant. Several hundred or a few thousand readings may be necessary for a thorough analysis of a complete basin.

A variety of statistical data-reduction and data-display techniques is available for paleocurrent work. The most common approach is to group data into

subsets according to stratigraphic or areal distribution criteria, display them visually in current rose diagrams, and calculate their mean and standard deviation (or variance). The method of grouping the data into subsets has an important bearing on the interpretations to be made from them, as discussed in the next section. A current rose diagram is simply a histogram converted to a circular distribution. The compass is divided into 20, 30, 40, or 45° segments, and the rose is drawn with the segment radius proportional to number of readings or percent of total readings. A visually more correct procedure is to draw the radius proportional to the square root of the percent number of readings, so that the segment area is proportional to percent. Examples are given in Fig. 5.40. Wells (1998) warned that the choice of origin for rose diagrams and the method of subdivision of the segments can affect the visual appearance of the resulting diagram and the orientation of the modes. The calculation of means, which is done from raw data, is not affected, but Wells' warning indicates the need for caution in the visual interpretation of rose diagrams.

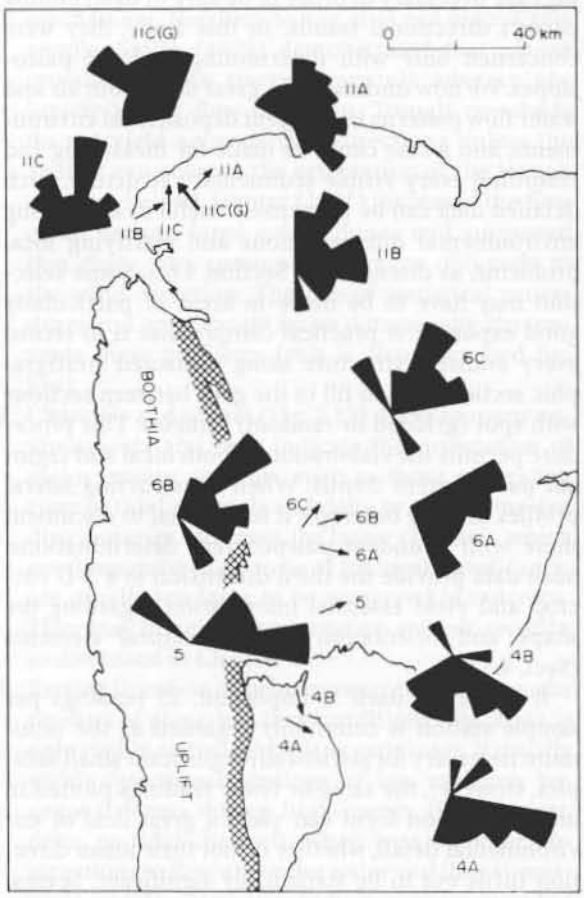


Fig. 5.40. Typical regional paleocurrent map, showing current rose diagrams and vector-mean arrows for each of ten outcrop areas, with station numbers. The map represents 165 field readings. Devonian fluvial sandstones, Somerset Island, Arctic Canada. (Miall and Gibling 1978)

Potter and Pettijohn (1977) and Curray (1956) discussed arithmetic and vector methods for calculating mean, variance, and vector strength, and statistical tests for randomness. Miall (1976) and Cant and Walker (1976) demonstrated the display and interpretation of data collected from vertical stratigraphic sections (Sect. 5.6.5).

Statistical procedures, such as moving averages and trend analysis, are available for smoothing local detail and determining regional trends. A moving-average map is constructed from gridded data, as shown in Fig. 5.41. An arbitrary grid is drawn on the map; the mean current direction for the data in each group of four squares is then calculated and shown by an arrow at the center of this area. Each data point is thus used four times (except at the edges of the map). Examples of this method applied to paleocurrent data have been published by Pelletier (1958), Sturm (1971), Andersen and Picard (1974), and Seeland (1978). Trend analysis requires rather more complex calculation. Agterberg et al. (1967) used this technique in the study of a Triassic delta.

The basin analyst should be wary of becoming too deeply enmeshed in the refinements of statistical methods. The use of probability tests is a useful curb on one's wilder flights of interpretive fancy, but there is a vast literature on the statistics of the circular distribution that seems to detract attention from very simple questions: do the data make geological sense? Can they be correlated with trends derived from other methods, such as lithofacies mapping or petrographic data? Moving-average and trend analyses are useful techniques for reducing masses of data to visually appealing maps, but they inevitably result in a loss of much interesting detail.

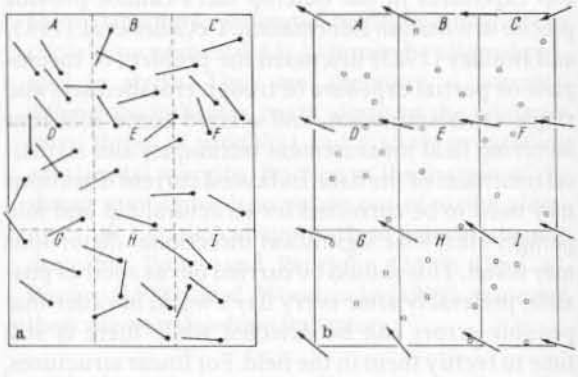


Fig. 5.41. Construction of a moving-average paleocurrent map (*right*) from raw data (*left*). (Potter and Pettijohn 1977)

5.6.4 The Bedform Hierarchy

A geologist collects paleocurrent measurements over individual bedding planes, through a local sequence in a quarry, along a lengthy dipping section in a river cut, or from several such sections throughout a map area. These readings differ from each other by varying amounts at different levels of this sampling hierarchy, but what do these differences mean? Statistically this is a classic example of a problem in analysis of variance (Olsen and Potter 1954; Kelling 1969), but what do the differences mean geologically? The larger the sampling area or the thicker the sampled section, the greater will be the number of depositional events that contributed to the data set and, normally, it will mean an increase in directional variance. This generalization can be systematized by the concept of the **bedform hierarchy**, as first noted by Allen (1966, 1967). Flow fields and the sedimentary structures arising from them span up to six orders of magnitude, ranging from individual ripples to entire depositional systems. Allen (1966) and Miall (1974) applied this idea to river systems (Fig. 5.42). In Table 5.3, an attempt is made to expand it to other environments. Jackson (1975) showed that these ranks depend on three orders of time scale and physical scale and proposed that they be grouped into three dynamic types:

1. **Microforms** are structures of rank 6, which are controlled by turbulent eddies in the fluid-boundary layer. The time scale of variation is on the order of hours to seconds.
2. **Mesoforms** are structures of rank 5 and, to some extent, rank 4. They form in response to what Jackson (1975) termed **dynamic events**, such as hurricanes, seasonal floods, or eolian sand storms, when disproportionately large volumes of sediment are moved in geologically instantaneous

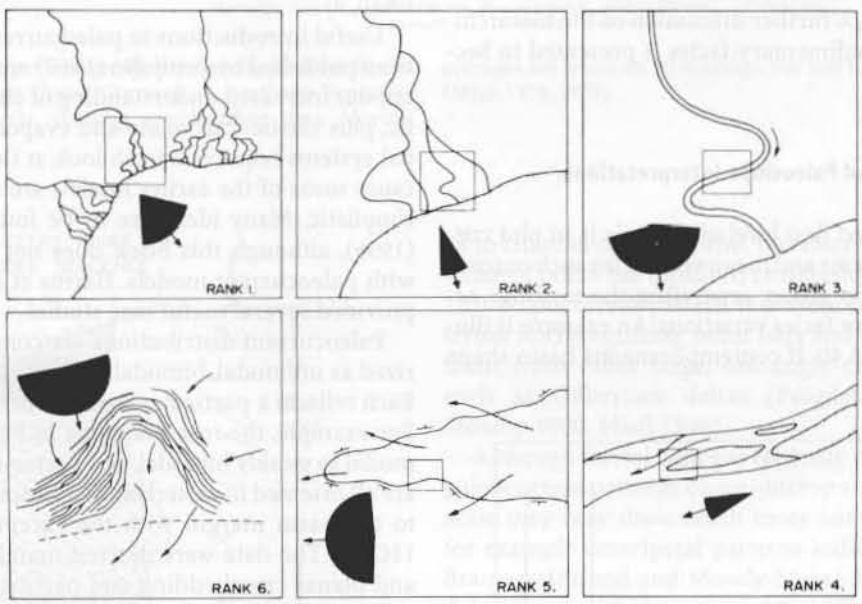


Fig. 5.42. The bedform hierarchy, illustrated for a fluvial–deltaic system. Current rose diagrams illustrate directional variability for the structures of each rank, which must be added to that of the next larger structures to determine total variability of the structures of each rank. *Small squares* show the area enlarged in the next lower rank. The structures are: *rank 1* fluvial–deltaic system; *rank 2* channels within a delta; *rank 3* channel reaches; *rank 4* lateral and mid-channel bars; *rank 5* crossbedding flanking a bar; *rank 6* ripple crests on a bar surface, oriented around a sheet runoff gully. (Miall 1974)

Table 5.3. The bedform hierarchy and its application to three clastic environments

Rank	Dynamic type	Fluvial systems	Tidal inlets and deltas	Submarine fans
1	Macroform	River system	Inlet system	Fan system
2	Macroform	Meander belt	Inlet channel	Individual fan
3	Macroform	Channel reach	Ebb or flood delta	Channel, suprafan lobe
4	Macroform	Lateral, point bar	Spill-over lobe, ebb spit	Sediment-gravity flow
5	Mesoform	Large-scale crossbedding	Large-scale crossbedding	Flute
6	Microform	Ripple	Ripple	Vortex marks

time periods. The system may remain virtually unchanged between dynamic events.

3. **Macroforms** represent the long-term accumulation of sediment in response to major tectonic, geomorphic, and climatic controls.

It is important, for mapping purposes, to attempt to relate the sampling scale to this hierarchy so the sources of directional variance can be interpreted intelligently. For example, a small quarry might be located entirely within the deposit of a single point bar or tidal delta, whereas a township (6 miles or 9.6 km square) could encompass an entire delta lobe or barrier-inlet system. When interpreting small-scale structures (ranks 5 and 6, the usual stuff of paleocurrent analysis) over large areas, it is important to remember that directional variance within each rank is summed to that of the higher ranks as far as is locally appropriate. Miall (1974) demonstrated the implications of this for fluvial–deltaic systems, such as those shown in Fig. 5.42, by compiling data for the various ranks summed to that of the single river or meander belt (Fig. 5.43). A further discussion of the hierarchical nature of sedimentary facies is presented in Section 4.5.9

5.6.5 Environment and Paleoslope Interpretations

A recommended first level of analysis is to plot current rose diagrams and mean vectors for each outcrop or local outcrop group, separating the readings according to major facies variations. An example is illustrated in Fig. 5.40. If contemporaneous basin shape

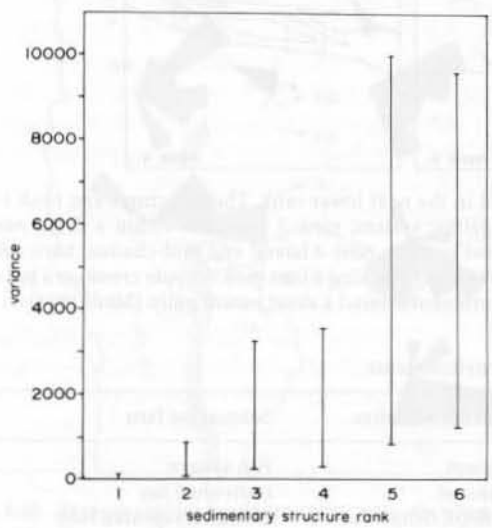


Fig. 5.43. The range of directional variance shown by sedimentary structures of different ranks. Based on a compilation of data for fluvial systems by Miall (1974)

and orientation can be deduced from these data or from other mapping techniques, the paleocurrent data can be used interactively with lithofacies and biofacies criteria to interpret the depositional environment and to outline major depositional systems, such as deltas or submarine-fan complexes. The geologist should examine the relationships between mean current directions in different lithofacies and in different outcrops. The number of modes in the rose diagrams (modality) and their orientations with respect to assumed shoreline or lithofacies contours are also important information. Another useful approach is to plot individual readings or small groups of readings collected in measured sections at the correct positions in graphic section logs (Figs. 5.44, 5.45, 5.46). Some examples of how to use these data are discussed below, and many regional case studies are considered in Chapters 6 and 9. Potter and Pettijohn (1977) provided extensive examples and an annotated bibliography. Advanced methods for use in the analysis of fluvial systems are described by Miall (1996, Chap. 9)

Useful introductions to paleocurrent models have been published by Pettijohn (1962) and Selley (1968), but our increased understanding of clastic (siliciclastic, plus clastic–carbonate and evaporite) depositional systems requires a fresh look at the problem, because some of the earlier models are now seen to be simplistic. Many ideas are to be found in Reading (1996), although this book does not deal explicitly with paleocurrent models. Harms et al. (1975, 1982) provided several useful case studies.

Paleocurrent distributions are commonly categorized as unimodal, bimodal, trimodal, or polymodal. Each reflects a particular style of current dispersion. For example, the rose diagrams in Fig. 5.40 are unimodal to weakly bimodal, and vector-mean directors are all oriented in easterly directions, roughly normal to the basin margin with the exception of station 11C(G). The data were derived mainly from trough and planar crossbedding and parting lineation, and the environment of deposition is interpreted as braided fluvial (Miall and Gibling 1978). In some areas, for example, stations 6A, B, and C, there is a suggestion of a fanning out of the current systems, indicating possible deposition on large, sandy alluvial fans. The anomalous data set of station 11C(G) was derived from giant crossbed sets ranging from 1 to 6 m in thickness. They may be eolian dunes or large sand waves in a trunk river draining the basin east of Boothia Uplift. Field evidence is inconclusive, but the paleocurrent data can be reconciled with either interpretation.

In high-sinuosity rivers, point bars dip at high angles to channel trends, whereas minor structures migrating down the point bars are oriented subparal-

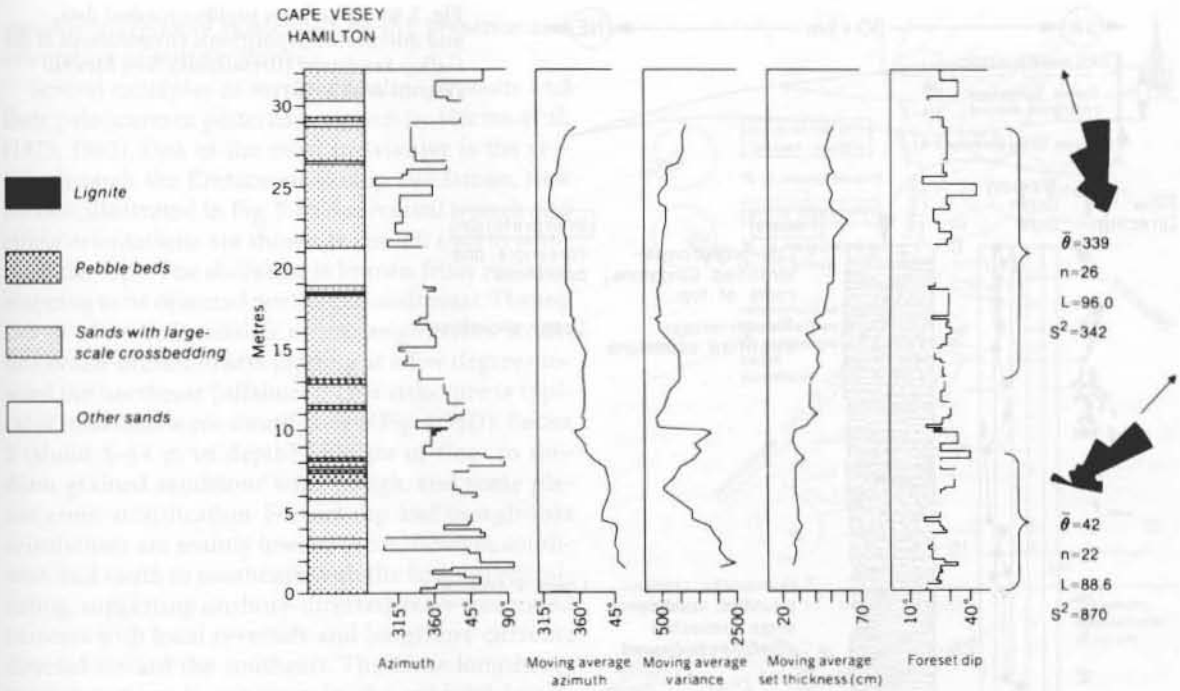


Fig. 5.44. Detailed vertical profile through the braided, fluvial Isachsen Formation (lower Cretaceous), northern Banks Island, Arctic Canada, showing crossbedding data. Moving averages are based on 10 readings. See text for interpretation. (Miall 1976, 1979)

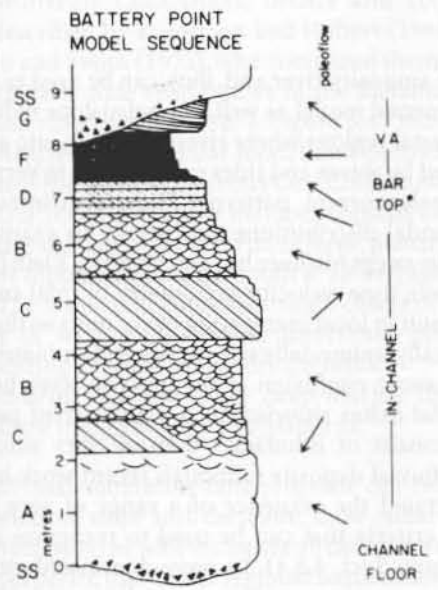


Fig. 5.45. Generalized vertical profile model for the fluvial Battery Point Formation (Devonian), Quebec, showing variation of crossbed orientations in different lithofacies types (north toward top of page). *VA* Vertical accretion deposits. (Cant and Walker 1976)

lel to channel direction and, therefore, parallel to the strike of point-bar (epsilon) crossbedding (Fig. 4.19). This paleocurrent pattern is a useful diagnostic criterion for recognizing point bars and distinguishing them from other large, low-angle crossbed types, such as Gilbertian deltas (Puigdefábregas 1973; Mossop 1980; Miall 1996).

Although fluvial deposits typically yield unimodal paleocurrent patterns on an outcrop scale, on a larger scale they may show much more complex patterns, for example centripetal patterns indicating internal drainage (Friend and Moody-Stuart 1972) or bimodal patterns with the two modes at 90° and occurring in different lithofacies assemblages. Rust (1981) reported two examples of the latter case. In each case, one mode, occurring in coarse conglomerates, was interpreted as the product of alluvial fans prograding transversely out from the basin margin. The other mode, in interbedded sandstones, represented a trunk-river system draining longitudinally.

Miall (1976, 1979) reported a similar case of interacting river systems, in which the respective lithofacies could only be separated by documenting subtle differences in crossbedding characteristics. Figure 5.44 displays data collected from a 32-m-thick section through a lower Cretaceous braided-stream deposit at the margin of Banks Basin, Arctic Canada.

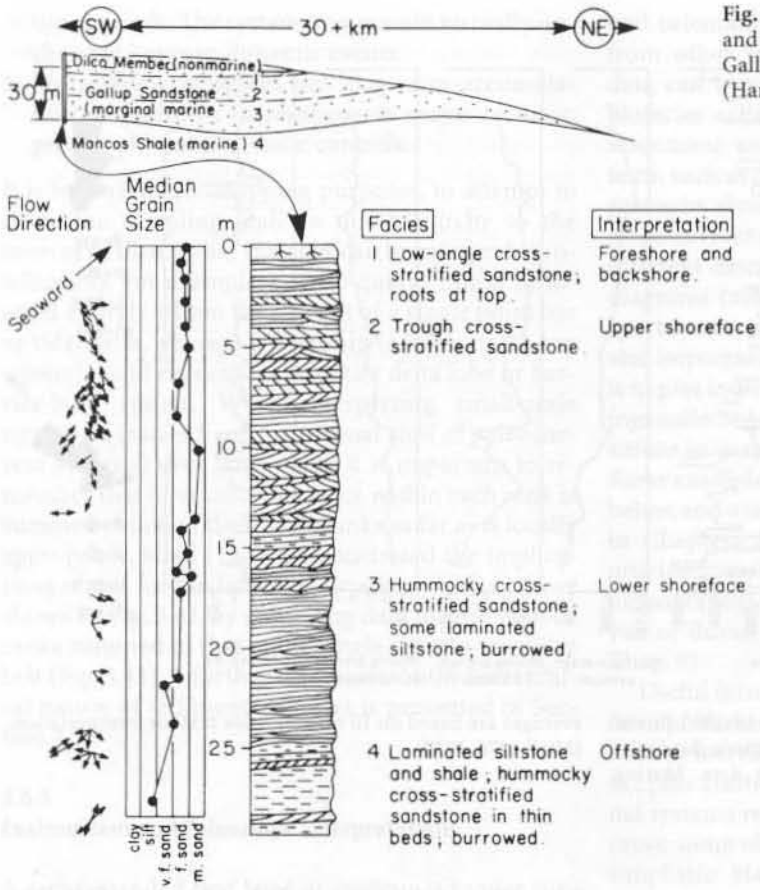


Fig. 5.46. Lithofacies profile, crossbed data, and interpreted depositional environments of the Gallup Sandstone (Cretaceous), New Mexico. (Harms et al. 1975)

The sediments consist mainly of superimposed planar crossbed sets deposited from large, 2-D dunes. At around the 10-m level in the section, they become thicker, shift in mean orientation by 63°, and show less directional variance. These changes were interpreted as the result of rejuvenation of a tributary river draining into the basin from the southeast, possibly as a result of movement on one of the faults bounding the basin.

It is a common observation in fluvial deposits that planar crossbeds show a higher directional variance than trough crossbeds (Miall 1977). Cant and Walker (1976) and Walker and Cant (1979) compared a modern river with an ancient deposit and showed why this is so. The deposits, in both cases, consisted of a series of fining-upward cycles, a model example of which is shown in Fig. 5.45. Trough crossbeds in the Devonian unit show rather consistent current directions flowing toward the west–northwest and were interpreted as the product of 3-D dunes migrating down the deeper channels. Planar crossbeds are oriented at high angles to this trend as a result of deposition from large, lobate bars (2-D dunes) migrating across the channel. This lithofacies and paleocurrent pattern is distinctive of certain types of low-sinuosity river and, thus, can be used as an environmental model as well as a paleoslope indicator.

Coastal regions where rivers debouch into an area affected by waves and tides can give rise to very complex paleocurrent patterns. Bimodal, trimodal, or polymodal distributions may result, for example, in the tide-swept offshore bar described by Klein (1970). However, time–velocity asymmetry of tidal currents can result in local segregation of currents so that they are locally unimodally ebb- or flood-dominated. This could cause confusion at the outcrop level, because the tidal deltas showing such paleocurrent patterns may consist of lithofacies that are very similar to some fluvial deposits (although recent work has demonstrated the existence of a range of very subtle facies criteria that can be used to recognize a tidal signature; Sect. 4.5.4). In wave-dominated environments, current reversals generate distinctive, internal ripple-lamination patterns and herringbone crossbedding, as illustrated in Fig. 4.24. Paleocurrent analyses of these bimodal crossbeds may yield much useful information on the direction of wave attack and, hence, on shoreline orientation. The term paleoslope means little in this environment, because there are a diversity of local slopes and because waves and tides

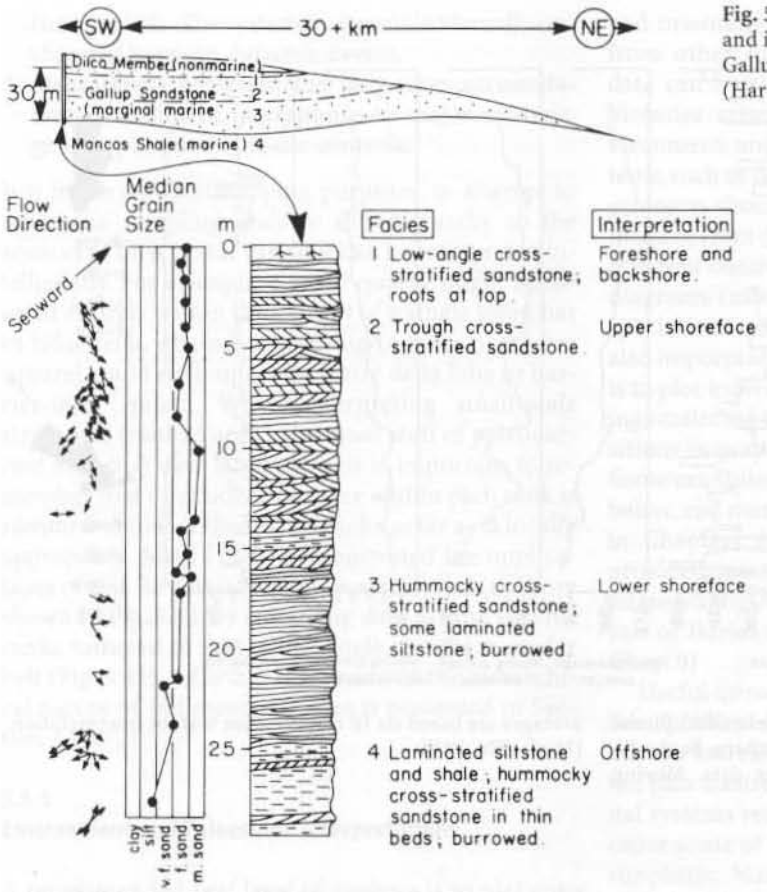


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are only marginally influenced by the presence and orientation of bottom slopes.

Several examples of sandy shoreline deposits and their paleocurrent patterns are given by Harms et al. (1975, 1982). One of the most instructive is the section through the Cretaceous Gallup Sandstone, New Mexico, illustrated in Fig. 5.46. Individual trough and ripple orientations are shown at the left (north is toward the top). The shoreline is known from regional mapping to be oriented northwest–southeast. The top 2–3 m of section consists of low-angle, cross-stratified beach-accretion sets dipping at a few degrees toward the northeast (offshore). This structure is typical of intertidal wave-swash zones (Fig. 4.39D). Facies 2 (about 3–14 m in depth) consists of fine- to medium-grained sandstone with trough, and some planar, cross-stratification. Foreset-dip and trough-axis orientations are mainly toward the northwest, southwest, and south to southeast, with the latter predominating, suggesting onshore-directed, wave-generated currents with local reversals and longshore currents directed toward the southeast. The same longshore-current pattern is apparent in the subtidal lower-shoreface deposits, which are dominated by storm-generated hummocky cross-stratification. Wave ripples are present in the lowest facies, again showing predominantly onshore-directed currents.

Similar paleocurrent patterns in some ancient, shallow-marine calcarenites, oolites and coquinas were described by Knewtson and Hubert (1969) and Striklin and Smith (1973), who compared them to the modern carbonate sand bodies in the Bahamas that were described by Ball (1967). Figure 5.47 shows the reconstructed depositional model and paleocurrent pattern of the lower Cretaceous Cow Creek Limestone in central Texas. The upper member consists of 4.5 m of calcarenite and coquina forming large, planar crossbed sets dipping toward the south at an average angle of 7°. The lower unit contains trough cosets oriented mainly in west–southwesterly directions, probably representing subtidal longshore currents.

Submarine-fan and other deep-marine deposits show three main paleocurrent patterns:

1. Individual submarine fans prograde out from the continental slope and, therefore, show radial paleocurrent patterns with vector-mean directions oriented perpendicular to the regional basin strike (Stow et al. 1996). This pattern is typical of many continental margins, particularly divergent margins.
2. In narrow oceans and many arc-related basins, deep-water sedimentation takes place in a trough oriented parallel to tectonic strike. Sediment-gravity flows, particularly low-viscosity turbidity currents, emerge from submarine fans and turn 90° to flow longitudinally downslope, possibly for hundreds of kilometers. Current directions may be reversed by tilting of the basin. Many examples of this pattern have been published (Trettin 1970; Hesse 1974). Figure 5.48 illustrates a lower Paleozoic example in the Canadian Arctic, in which the 90° turn in flow direction was meticulously documented. Pickering and Hiscott (1985) used detailed grain-orientation measurements as a basis for reinterpreting anomalous paleocurrent measurements in a turbidite sequence. It was shown that supposed backset crosslamination attributed to upcurrent antidune migration was, in fact, the product of turbidity currents that reversed direction as a result of deflection and reflection from the basin margins and internal topographic highs. Here is a case where, to improve the sophistication of a basin-analysis case study, subtle use was made of paleocurrent data.
3. Contour currents or boundary undercurrents flow parallel to continental margins and generate paleocurrent patterns oriented parallel to basin margins. Stow and Lovell (1979) and Stow and Holbrook (1984) discussed the use of facies criteria and paleocurrent data in distinguishing these deposits.

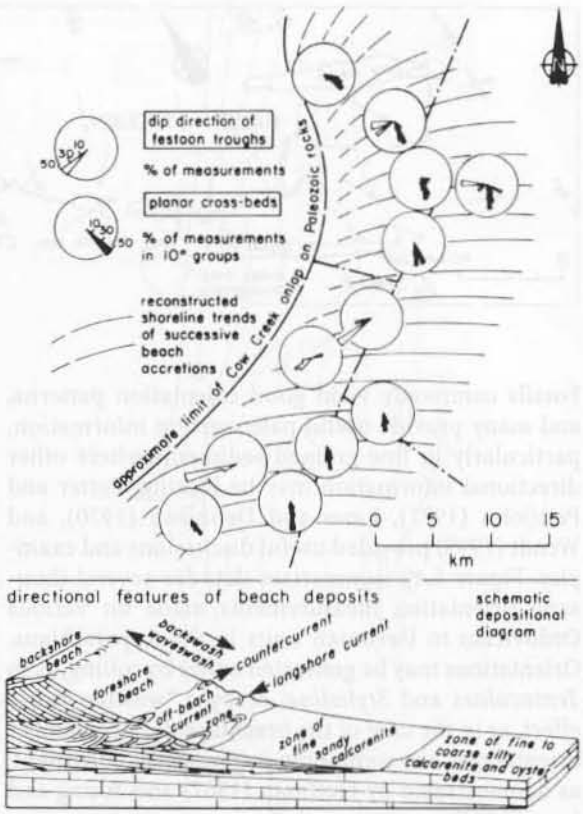


Fig. 5.47. Two types of crossbedding in the Cow Creek Limestone (lower Cretaceous), central Texas, and their interpretation. (Stricklin and Smith 1973)

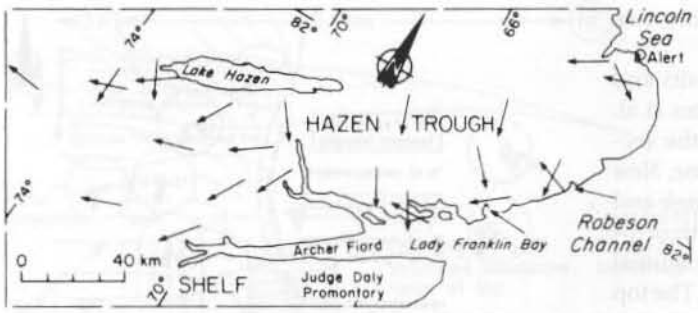


Fig. 5.48. Paleocurrent trends revealed by sole markings in the lower Paleozoic turbidities of Hazen Trough, northern Canadian Arctic. Note the 90° swing in direction from transverse (southeasterly) to longitudinal (southwesterly). (Trettin 1970)

Fossils commonly yield good orientation patterns, and many provide useful paleocurrent information, particularly in fine-grained sediments where other directional information may be lacking. Potter and Pettijohn (1977), Jones and Dennison (1970), and Wendt (1995) provided useful discussions and examples. Figure 5.49 summarizes data for several thousand orientation measurements made on various Ordovician to Devonian units in the Appalachians. Orientations may be generated either by rolling, as in *Tentaculites* and *Styliolina*, or by a "weather-vane" effect, as in the case of the brachiopods. Stromatolite orientations are commonly impressively consistent, as demonstrated by Hoffman (1967) and Young and Long (1976).

The examples of paleocurrent studies discussed in this section show that the depositional environment, bedform hierarchy, and sampling scale all affect the consistency, variance, and modality of the data. This fact should lay to rest the widely quoted remark by Potter and Pettijohn in the first edition (1963) of their book (repeated in the 1977 edition) that, with regard to crossbedding, "the most common variance of fluvial–deltaic deposits is in the range 4000 to 6000 ... whereas the smaller sample of marine deposits is the range 6000 to 8000." This supposed difference has been used as a criterion for distinguishing depositional environments, although use of all the facies-analysis and mapping methods described in Chapters 4 and 5 of this book should have rendered the approach obsolete. Long and Young (1978) re-examined the idea and compiled more data, modifying the variances to 1000 to 4000 for fluvial rocks and 4000 to 8000 for marine rocks. They were mainly concerned with the use of the method for studying Precambrian rocks, where some analytical methods, such as biofacies analysis, are unavailable (except for stromatolites). However, the problem of the source of the variance remains, and anyone who has time to measure enough crossbeds for a variance study has time to carry out a proper basin analysis.

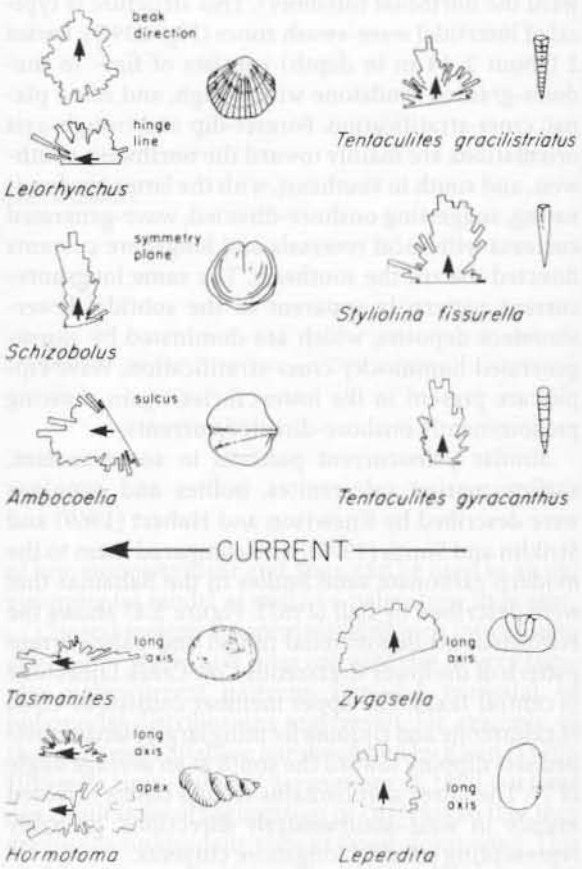


Fig. 5.49. Examples of orientation measurements made on fossil invertebrates in Ordovician to Devonian sediments of the Appalachians. (Jones and Dennison 1970)

5.6.6 The Dipmeter Again

The principles of dipmeter data collection, processing, and presentation, and the uses of these data in facies studies are presented in Section 5.4.3. In this section, some brief comments and examples are provided, showing the use of the tool in paleocurrent analysis.

Nurmi (1978) reported that the use of computer-processing routines, such as those described by

Vincent et al. (1977), may result in as many as 14 dip readings over a 30-cm interval. Such detailed results require large-scale, graphical tadpole displays, and scales of 1:40 or even 1:24 are typical, which contrasts with the 1:1200 scale (1 in = 100 ft) commonly used for subsurface work in the North American petroleum industry. If readings can be obtained at nearly 2-cm intervals, the dipmeter could theoretically see structures as small as ripples. However, these would not normally yield planar surfaces when cut by a normal-diameter petroleum exploration hole and, therefore, the processing routines could not recognize them. The dipmeter can see medium- to large-scale crossbed foresets, which offers considerable potential for paleocurrent analysis, but even here there are many pitfalls. Planar crossbed foresets should be particularly easy to recognize in dipmeter logs, particularly if they display the thickness range and orientation consistency of those illustrated in Fig. 5.44. Trough crossbeds would be less easy to identify correctly, because a hole penetrating the edge of a set will yield an apparently spurious orientation, and the trough axis will be nearly horizontal. Set bounding surfaces, reactivation surfaces, channels, slumps, overturned crossbeds, and the climb surface of climbing ripples may all yield apparently spurious readings. In practice, therefore, although practically every paper on dipmeter applications trumpets the possibilities for paleocurrent analysis, there are very few good published examples available (this may be an industry confidentiality problem). Only where the dipmeter log can be correlated with core – itself a tricky procedure – are reliable results likely to be obtained. Two examples are illustrated here. Many other examples, with numerous color illustrations, are given by Serra (1985). The Formation MicroScanner (FMS) log may be used instead of the dipmeter to evaluate the presence and orientation of dipping surfaces, but the same cautions apply with regard to the planarity of the surfaces and the many different origins of such surfaces.

Rider (1978) illustrated a dipmeter analysis of a channel sandstone. Figure 5.50 shows sedimentary dips (the central mode in Fig. 5.27) within a channel sandstone. According to Rider, crossbed-set thickness is too small to have been picked up in his analysis. These dips, therefore, could represent set bounding surfaces, minor channels, drape over bars, etc. The southerly dips along the channel axis may be reactivation surfaces in large, multiset bar forms (macroforms). Westerly dips oriented toward the channel margin could be of the same origin, as midchannel bars in sandy rivers commonly migrate close to the bank, leaving a small slough (Cant and Walker 1976).

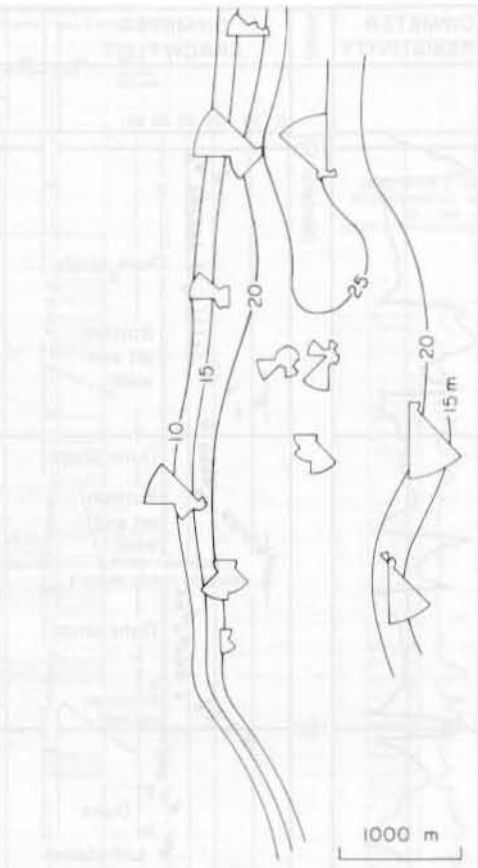


Fig. 5.50. Dipmeter and isopach data for a channel sandstone. (Rider 1978)

Some of the best dipmeter patterns are obtained from the large, consistently oriented crossbed sets of ancient, eolian dune fields. Nurmi (1978) gave an example from the Rotliegendes Sandstone (Permian) of the southern North Sea Basin (Fig. 5.51). It is instructive to compare these dipmeter readings with the dune block-diagram model illustrated in Fig. 4.26. The long, curved bottom sets are particularly obvious in the dipmeter data, as is the overall westerly directed paleocurrent pattern. However, care must be taken with interpretations from such data. The Rotliegendes eolian crossbedding has been the subject of a debate based on different interpretations of the dynamics of draa migration and crossbed preservation. Glennie (1982) interpreted tightly clustered dipmeter crossbed data as the product of migration of transverse dunes with crests oriented perpendicular to wind direction. However, Sneh (1988) applied the ideas of Rubin and Hunter (1985) to a reinterpretation of these rocks and suggested that the data indicate a dominance of obliquely oriented dunes, requiring a subtly different paleogeographic interpretation (Sect 4.6.4).

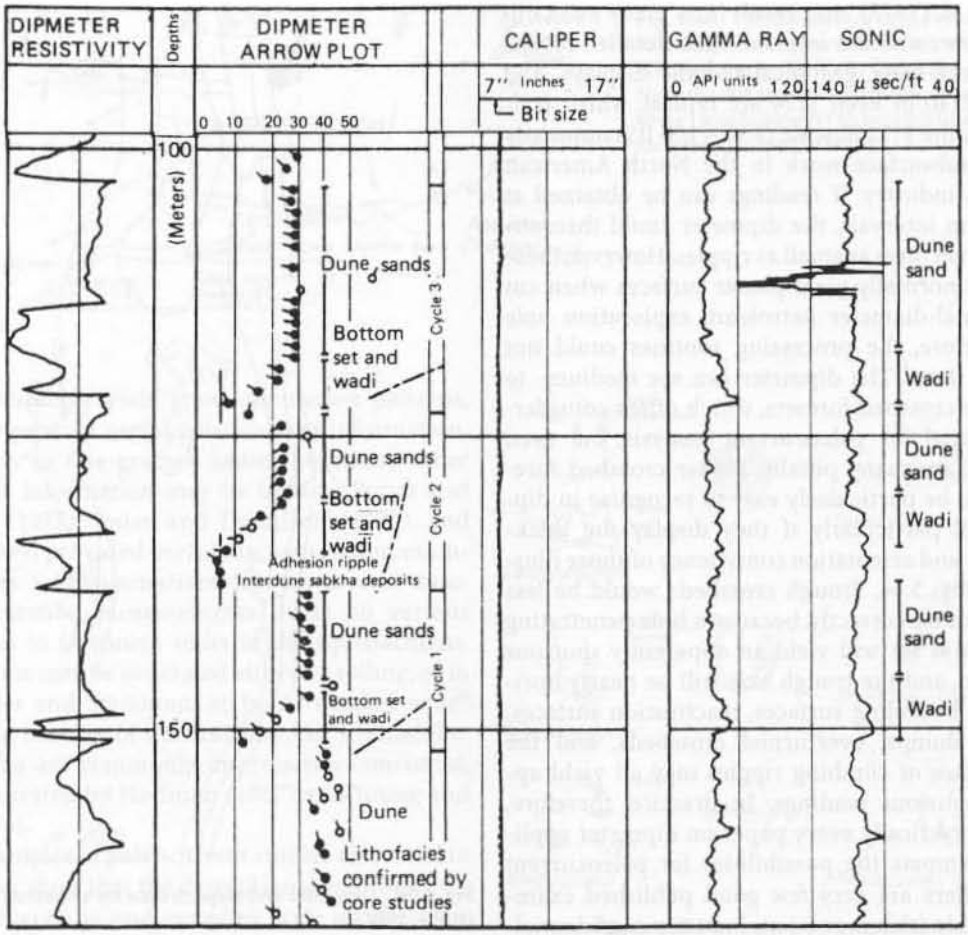


Fig. 5.51. Dipmeter data for a section through eolian crossbedding and associated facies in the Rotliegendes Sandstone (Permian), southern North Sea. (Nurmi 1978)

5.7 Remote Sensing

Remote sensing "refers to all of the arts and techniques of measurement and interpretation of phenomena from afar ... specifically ... the use of electromagnetic radiation to obtain data about the nature of the surface of the earth" (Goetz and Rowan 1981). Conventional aerial photography is included under this heading and has been used as a mapping tool for many years (Fig. 3.2). Bussell (1998) provided an excellent review of modern techniques, and Sgavetti (1992) illustrated the use of remote sensing in the stratigraphic mapping of the Pyrenees. Likewise, the use of airborne geophysical techniques for mapping and prospecting has a long-established history (e.g., aeromagnetic mapping of igneous bodies and associated mineral deposits). Today, the term remote sensing is generally understood to mean mapping from satellites although, typically, satellite imagery is combined, wherever possible, with that obtained from the air and the ground (the latter, once called field work, is now termed "ground truthing"). Modern sensing tools include a wide variety of airborne and satellite-borne, sophisticated detectors that record and transmit data in digital form, which permits the use of various techniques of image enhancement and selective display. Multispectral scanners, spectrometers, and radar have all been used. Data points (pixels) in satellite imagery may now be as small as 9 m across. Most geologists are familiar with Landsat imagery, but there are now many varieties of Landsat data available, and many other tools are in use or in development (Settle 1984; Lang 1985; Paylor et al. 1985; Lang et al. 1987).

However, the question must be asked, what services can remote sensing actually perform as a contribution to basin analysis? The most obvious application is in the exploration of areas where, for economic or political reasons, airborne or ground-mapping

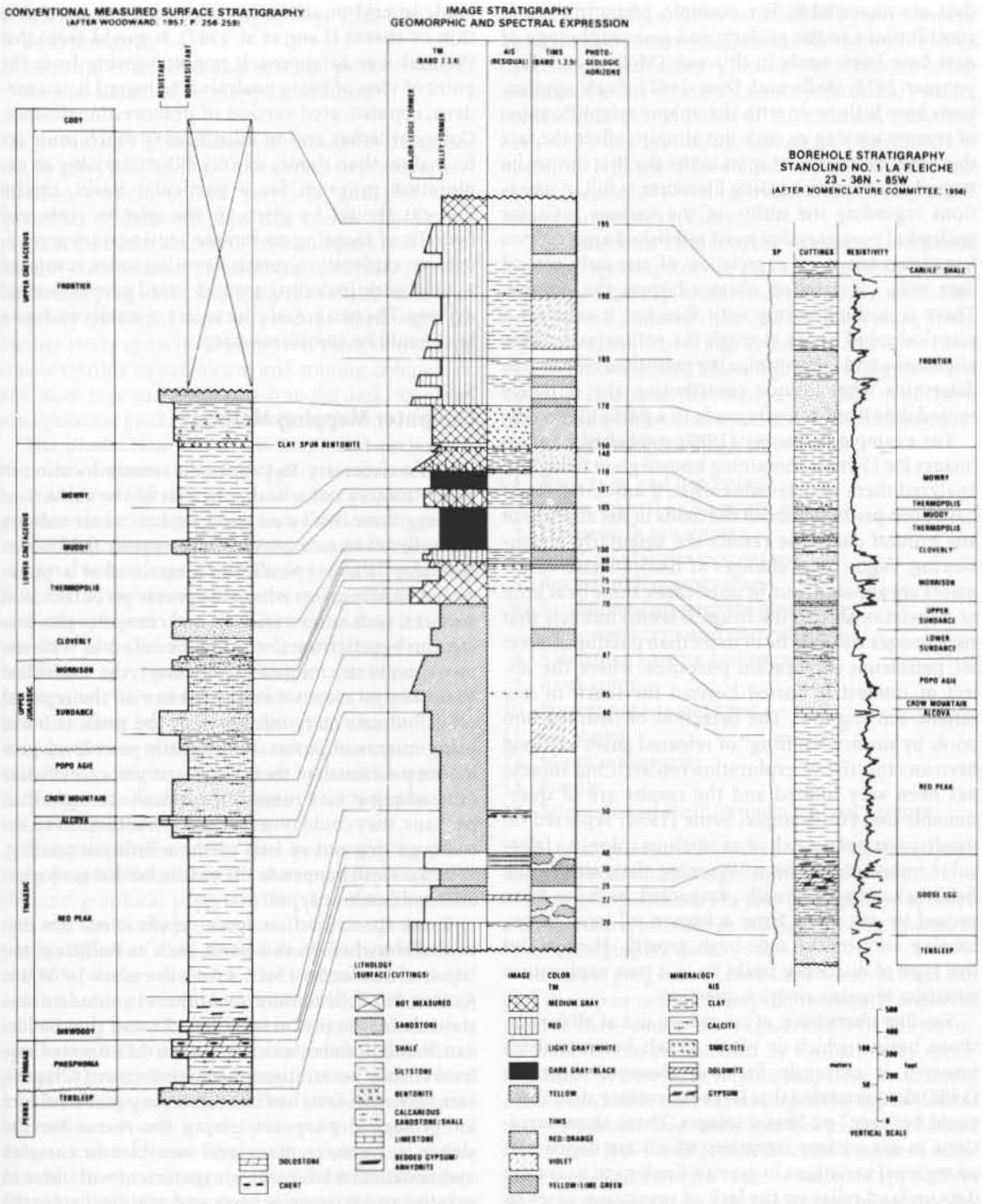


Fig. 5.52. Correlation of a conventional surface lithologic section (*left*) and subsurface borehole log (*right*) with image stratigraphy obtained from three airborne instruments. *TM* Thematic mapper (Landsat); *AIS* airborne imaging spectrometer; *TIMS* thermal-infrared, multispectral scanner (also airborne). (Lang et al. 1987)

data are unavailable. For example, many important contributions to the geology and geomorphology of Asia have been made in this way (Molnar and Tapponnier 1975; Wells and Dorr 1987). Such applications have little to do with the unique scientific value of remote sensing as such but simply reflect the fact that satellites represent spies in the sky that cannot be turned off. Remote-sensing literature is full of assertions regarding the utility of the various tools for geological purposes, but most published applications have been based on correlation of remotely sensed data with information obtained from the ground. There is nothing wrong with this, but it requires a sceptical mind to see through the enthusiasm of the aficionado and to scrutinize the published research to determine what unique contribution the remotely sensed data itself actually made to a particular study.

For example, Halbouty (1980) published Landsat images for 15 areas containing known giant fields and analyzed them to determine what, if anything, could have been predicted about the fields in the absence of any ground data. The results are singularly unconvincing. Vague tonal changes or disconnected lineaments are apparent but, in most cases, these bear little or no relationship to the fields. It seems unlikely that such images will ever be of more than passing interest for petroleum-exploration purposes, where the object of interest is buried beyond the reach of any remote-sensing tool. The detection of hydrocarbon pools by surface "sniffing" of released gases has long been an objective of exploration research, but success has been very limited and the results are of questionable use. For example, Settle (1984) reported on the discovery of a patch of anomalous color in a false-color image of southern Wyoming that, after some fieldwork, was eventually suspected to have been caused by gas seeps from a known oil pool below, causing stunting in sage-bush growth. How useful this type of discovery could be in a true exploration situation remains an open question.

Satellite photos are, of course, no use at all for offshore basins, which is where much basin-analysis research is currently focused. However, Bostrom (1989) demonstrated that large subsurface structures could be "seen" on Seasat images. These show variations in sea-surface elevation, which are dependent on regional variations in gravity. Geological use of the data on land relies on the lack of vegetation cover or on the ability to correlate vegetation with soil characteristics, which requires ground checking. Spectral analysis can now discriminate mineralogical characteristics, so long as the rocks are actually exposed, and can, therefore, be used to generate spectral logs or image stratigraphy of exposed sequences (Fig. 5.52). It is, perhaps, not surprising that the greatest successes in the use of these techniques have been made in arid or semiarid areas, where vegetation is thin or absent (Lang et al. 1987). It would seem that the best way to approach remote sensing from the point of view of basin analysis is to regard it as a modern, sophisticated version of photoreconnaissance. Colors, whether real or false, convey much more information than shades of gray. When devising an exploration program for a particular basin, careful thought should be given to the relative costs and benefits of focusing on remote sensing during preliminary exploration versus devoting more resources to field work, including ground-based geophysics and drilling. The two are, of course, not mutually exclusive but should be complementary.

5.8 Computer Mapping Methods

It is not necessary to travel to a remote location in order to carry out a basin analysis of the subsurface geology there. Well data and geophysical records are typically taken to a convenient company field office. Nowadays, these typically are centralized at large, regional headquarters where the provision of technical services, such as core analysis and computer processing, can benefit from the economies of scale. With improvements in computer technology, the individual basin analyst may not even need to visit the regional office but can carry out much of the work at home using microcomputers. While many petroleum geologists pass much of their basin-analysis career without looking at very many real rocks, the thought that, perhaps, they could now do their job without even actually getting out of bed seems a little disquieting. Whatever will happen to the old cliché that geologists are hard outdoor types?

Some specialized computer applications are discussed elsewhere in this book, such as Rubin's simulation of bedforms (Sect. 4.5.4), the work by Watts, Karner, Bond, Beaumont, and others to simulate crustal-subsidence processes (Chap. 7), and that by Jordan, Kendall, and others to simulate the effects of sea-level change on stratigraphic architecture (Chap. 8). Simulation, in fact, has become a very powerful tool for geological purposes, giving the researcher the ability to convolve many real variables in complex mathematical relationships, experiment with these in iterative and interactive ways, and rapidly display the results. In addition, there are data-management and special processing routines used in various geophysical applications, especially seismic surveying and petrophysical logging (Sects. 2.4, 5.4).

In this section, we briefly discuss the use of the computer as a data-handling tool in the mapping and interpretation of sedimentary basins. Computers are used for the purposes of:

1. Storage and housekeeping of large data files, such as well records
2. Retrieval of selected data sets and performance of various statistical or mapping routines

The advantages of working with computer databases are, by now, well known – the ability to retrieve, manipulate, correct, update, correlate, and plot large quantities of data with ease and speed. This facility permits the basin analyst to crosscompare and correlate any combination of variables, to rapidly test different mapping ideas, to display the results visually in a variety of forms, and to produce hard copy for further study or archival purposes. Work of this type is now routine in petroleum and mining companies, and most now make use of advanced desk-top data-manipulation packages and **workstations**.

The production of various maps and sections is the most common computer application in basin analysis. Modern software can deal with complex stratigraphic situations, such as faults and stratigraphic truncations. Some software systems are designed to accept a wide range of different types of mapping data, including semiquantitative (e.g., ordinal) data and values based on probability estimates. It is also possible to build trends into contouring routines, which is a very valuable facility for geological purposes, where structural or facies trends or grain may be a dominant feature of a sedimentary basin. This is done by the use of algorithms in contouring routines that fit the data to elliptical or spheroidal surfaces (with operator-selected shapes and orientations) rather than spherical surfaces.

Many improvements in data processing have become possible because of the development of interactive and graphical programs for data handling. The operator sits at the terminal and plays with the data, testing different ideas to find out which works best and tells him/her the most. The workstation concept, which has long been used in seismic data processing, is becoming more widely used in other basin-analysis applications. The workstation is a terminal linked to a powerful microcomputer or mainframe with several monitors to display different data sets and which uses interactive software to permit rapid manipulation of data, calculation, and display. The computer system installed in the drilling ship JOIDES Resolution (see October 1985 *Geotimes*) is a good example of a powerful but flexible system that uses workstations to perform a wide variety of data-management and research functions. Many trade journals and magazines, such as *Geobyte*, published by the AAPG, and AAPG *Explorer*, report on computer applications in stratigraphic and geophysical basin analysis.

Among the most important types of data for basin analysts is the well file. For approximately the last 30 years, files have been established by government agencies and commercial houses for the well records of the major sedimentary basins. These files, covering areas such as the Western Canada Sedimentary Basin, contain millions of items of information, including each well's name, location, completion date, name of operator, geological formation tops, drill-stem test results, depths of cored intervals, and possibly much more (Farmer 1981; Clark 1981).

The American Stratigraphic Company and Canadian Stratigraphic Service routinely log most wildcat and many other exploration wells in North America, totaling several hundred thousand wells. For many years, they have offered these logs to industry and other users on a commercial basis. These logs are now digitized and stored in computer-processable data files. In addition to the basic information listed above, the files also contain the following lithologic information:

1. Rock type and percentage
2. Grain, crystal or fragment size
3. Framework percentage
4. Rock-accessories percentage
5. Rock-builders percentage
6. Sandstone-components percentage
7. Diagenesis type and percentage
8. Sandstone sorting and rounding
9. Rock color
10. Porosity grade and type
11. Oil shows
12. Minor accessories
13. Fossils
14. Minerals and their frequency of occurrence

Each of these items is recorded for all descriptive units or beds, which range from 30 cm to several tens of meters in thickness. The files are regularly updated and, therefore, provide an efficient storage system of basic stratigraphic data.

Retrieval programs are available to extract and list, plot in map or section form, and contour, if required, most of these data items (Figs. 5.53, 5.54). The retrieval interval can be specified by formation, by specified slices of a formation (**slice maps**), or by defining a fixed thickness interval ending at the top or base of the formation. Two excellent books by Robinson (1982) and Jones et al. (1986) described modern mapping procedures and the types of software application that are now available. Two examples of papers describing specialized data banks and the kinds of mapping projects carried out using them are given by Hughes (1984) and Putnam and Klovan (1987). The number and variety of computer programs available for mapping purposes have, in fact, become bewildering in recent years, as is known by anyone who has attended a large trade show recently, such as the exhibits area at an AAPG convention.

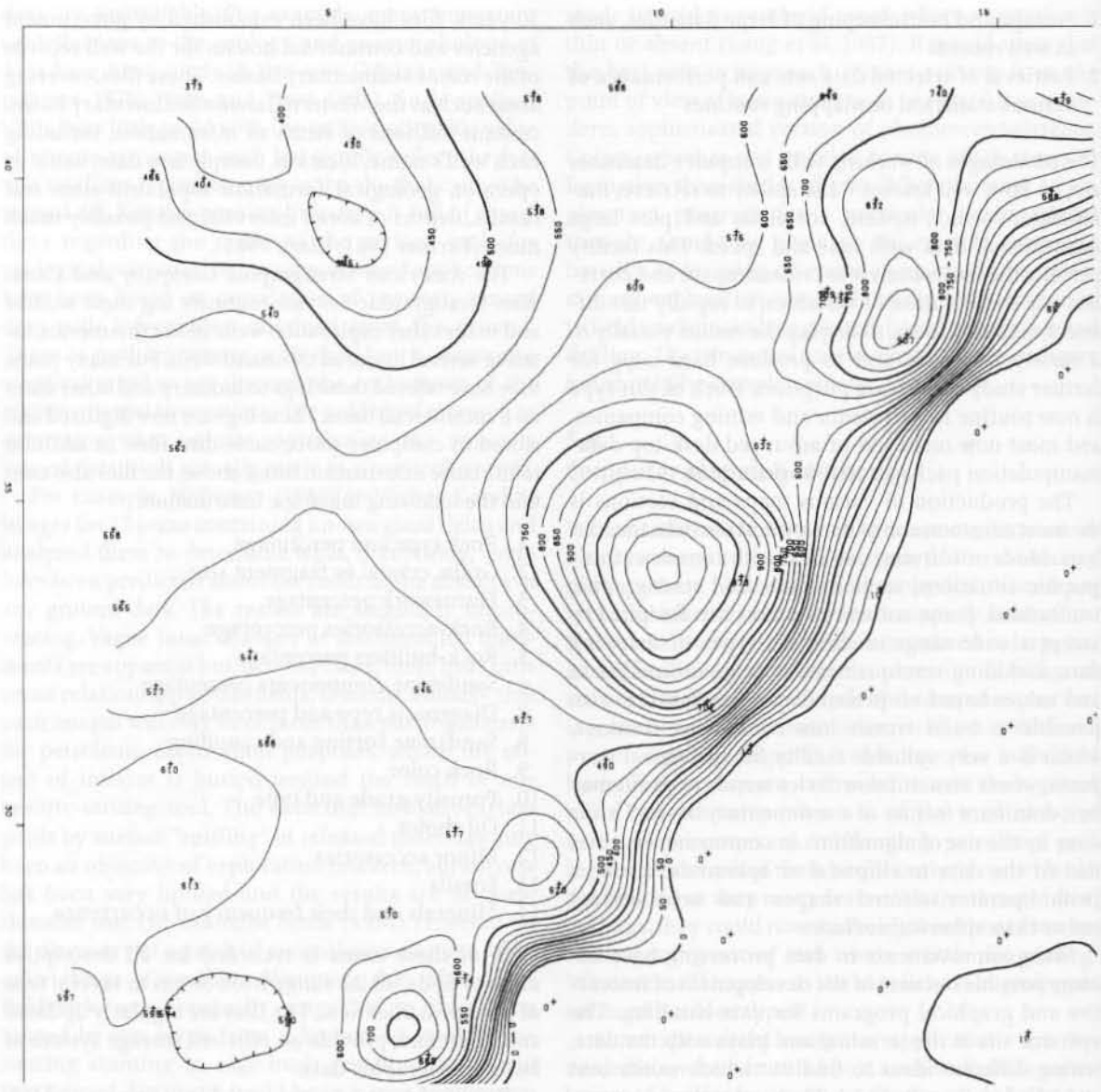


Fig. 5.53. A typical example of an isopach map plotted and contoured by a computer. (Courtesy of Canadian General Electric Corporation and Scientific Computer Applications Inc.)

One of the principal advantages of all these computer-based data-management systems is that they serve to standardize field and core observations with the provision of checklists and spaces on input forms to fill out each item. Standard codes and estimation procedures are used, so data from different areas, or collected by different operators, can readily be combined, if required, for basin-mapping purposes.

The main virtue of computer-produced maps and sections is the ease and rapidity with which they can be drawn. A typical exploration problem might be concerned with one or two formations penetrated by anywhere from several dozen to several thousand wells. To plot and contour all these maps by hand is a laborious procedure. The data can be retrieved and plotted by computer in minutes.

Whether computer-contoured maps and sections are better than hand-drawn ones is another question. For maps with numerous control points, they are probably equally meaningful but, where the control is sparse, there probably is little to be gained from using the computer other than its speed and neatness. Some computer experts would not agree. They point out that there may be numerous ways of contouring a given data set and suggest that only a clinically unbiased computer algorithm can produce an objective

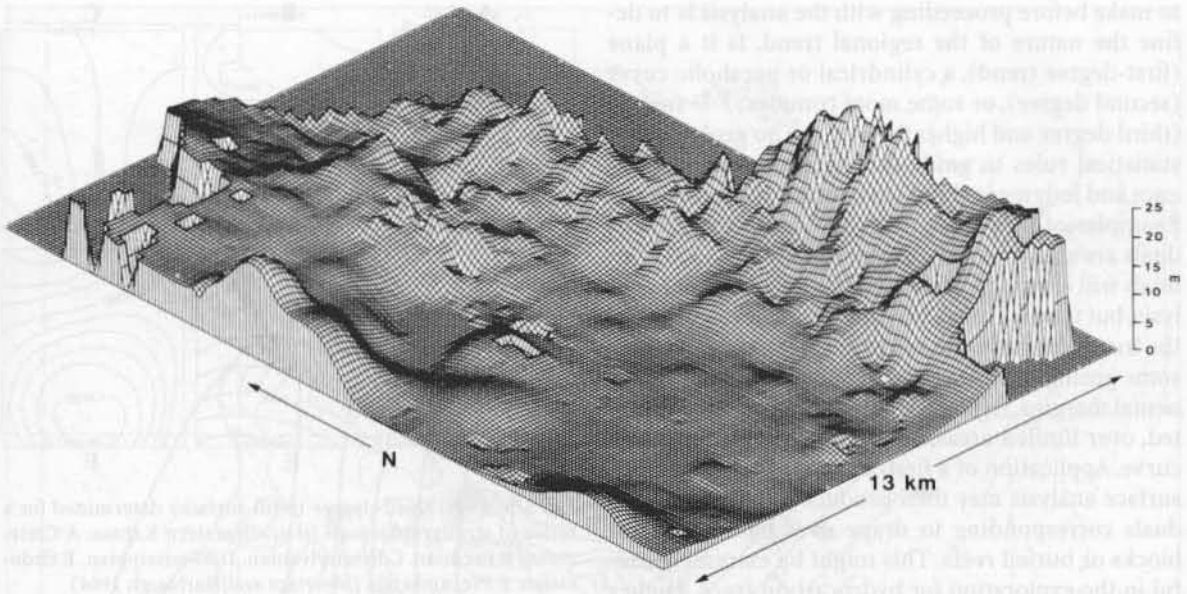


Fig. 5.54. Example of a perspective diagram drawn by computer from X-Y-Z data. Modern software can manipulate many such surfaces simultaneously, and perform rotation to any angle (Hughes, 1984)

map. However, data points on an isopach or lithofacies map are not arbitrary digits but reflections of subtle trends that only a skilled basin analyst can interpret. From regional information, paleocurrent data or some piece of core data, it might be possible to predict shoreline orientation or channel or reef trends, and the contours can be drawn accordingly. Such maps demonstrate the analyst's ideas about an area and, in the exploration business, the generation of ideas that go beyond established, routine facts is the most important contribution a geologist can make. The ideas may reflect several or many different interpretations but, although many companies may explore the same piece of territory using essentially the same data, it is usually the one with the best geological ideas that finds the mine or brings in the oil producer. The ability, using modern interactive-mapping routines, to change the contouring algorithm and to revise the data online also adds powerful new tools for testing ideas and refining the results and makes the computer even more useful.

A routine problem encountered in commercial data files is the accuracy of their stratigraphic information. Most retrieval programs for petroleum-exploration data are keyed to formation tops but, as we discuss in several places throughout this book, formations may be unreliable guides to paleogeography. They may be difficult to correlate and, particularly in some of the better explored sedimentary basins, they may be based on a local stratigraphic section and of little use regionally (Chap. 3). Application of the depositional-systems approach to stratigraphy may demonstrate that many formations have a quite arbitrary meaning (Chap. 6). The basin analyst may, therefore, be required to recheck all the relevant formation tops used in the data bank or input his or her own based on an up-to-date interpretation of the stratigraphy. The use of allostratigraphic mapping methods and sequence terminology should resolve this problem.

Once this task is completed, the data are available for processing by any kind of statistical data-reduction technique, many of which have broad geological applications. The use of multivariate statistics in facies analysis is discussed in Section 4.4; an example of the use of factor analysis in studying heavy-mineral dispersal trends is discussed in Section 5.5. This type of work and all the various multicomponent and ratio maps discussed in Section 5.3 are readily produced by statistical manipulation of a data bank. A particular favorite with users of well files is **trend-surface analysis**. The technique is described by Davis (1973) and Jones et al. (1986).

The purpose of trend-surface analysis is to separate map data into two components: regional trends and local fluctuations. The regional component can then be subtracted mathematically, leaving residuals that correspond to the local variation. The definition of regional and local is subjective and can be adapted to the particular mapping exercise at hand. For example, local fluctuations might be defined as channel- and crevasse-sand accumulations within a delta lobe or the individual lobes within a large delta complex. The most important decision the geologist has

to make before proceeding with the analysis is to define the nature of the regional trend. Is it a plane (first-degree trend), a cylindrical or parabolic curve (second degree), or some more complex, 3-D surface (third degree and higher)? There are no geological or statistical rules to guide this decision, only experience and judgment about what makes the most sense. Examples of how this affects the definition of residuals are shown in two dimensions in Fig. 5.55. Residuals will always be generated in a trend-surface analysis, but they will have no geological meaning unless the trend that has been extracted also corresponds to some geological reality. For example, on many continental margins, regional structure can be approximated, over limited areas, by a dipping plane or gentle curve. Application of a first- or second-degree trend-surface analysis may then produce a pattern of residuals corresponding to drape over basement fault blocks or buried reefs. This might be extremely useful in the exploration for hydrocarbon traps. Higher order trend-surface analysis will also result in complex patterns of residuals, but these may have no geological meaning whatever.

A geologist may be more interested in the regional trend than in the residuals, in which case derivation of ever more complex polynomial surfaces might be

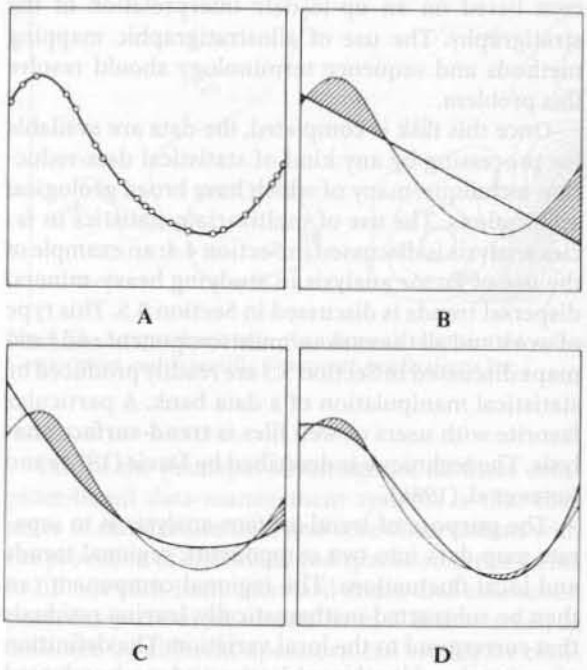


Fig. 5.55 A–D. Fitting a curve to data points – a two-dimensional representation of the trend-surface analysis process. Residuals are *shaded*. A Curve drawn through the original data. B Straight-line (plane) fit. C Parabolic fit. D Cubic fit. Note the different positions of the residuals in each case. (Davis, 1973; Reproduced with permission)

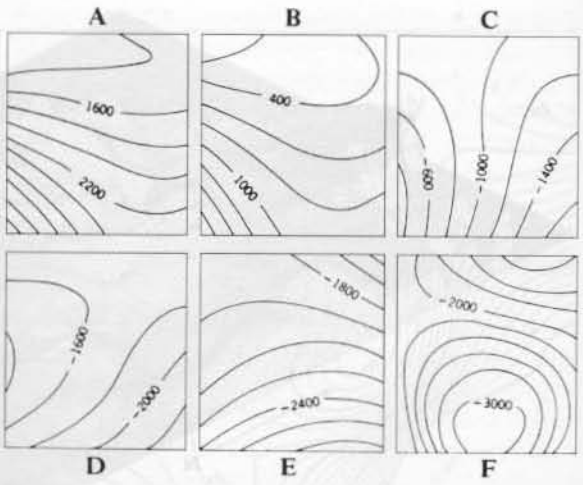


Fig. 5.56 A–F. Third-degree trend surfaces determined for a series of stratigraphic tops in northwestern Kansas. A Cretaceous. B Permian. C Pennsylvanian. D Mississippian. E Ordovician. F Precambrian. (Merriam and Harbaugh 1964)

required. Statistical tests are available to determine the goodness of fit of these surfaces to the original data. Again, the decision about which surface is most appropriate must be based on geological grounds. The highest-degree surfaces may fit the data very accurately, but only by accommodating all the local variability. Such a result is not likely to be very useful. Davis (1973) pointed out that realistic-looking trends can be produced by random data, and he discussed statistical tests for determining where to stop the analysis.

Two examples of trend-surface analysis are given in Figs. 5.56 and 5.57. The first of these shows a series of third-degree trend surfaces of stratigraphic tops in northwestern Kansas (from Merriam and Harbaugh 1964). The second example shows a fourth-degree trend and residuals for sand distribution in a Pennsylvanian delta (from Wermund and Jenkins 1970). The regional trend corresponds to the major lobe or depocenter, and individual channel fills and bar fingers are clearly shown. Moving averages and other, more advanced, computer map-analysis techniques are discussed by Miller and Kahn (1962), Davis (1973), and Jones et al. (1986).

5.9 Sampling Density and Map Reliability

Most lithostratigraphic maps are plotted from inadequate data. In any kind of exploration program, it is necessary to draw conclusions and make predictions based on a limited number of outcrops or exploratory holes. The subsurface geologist will commonly be required to make decisions about where to place the

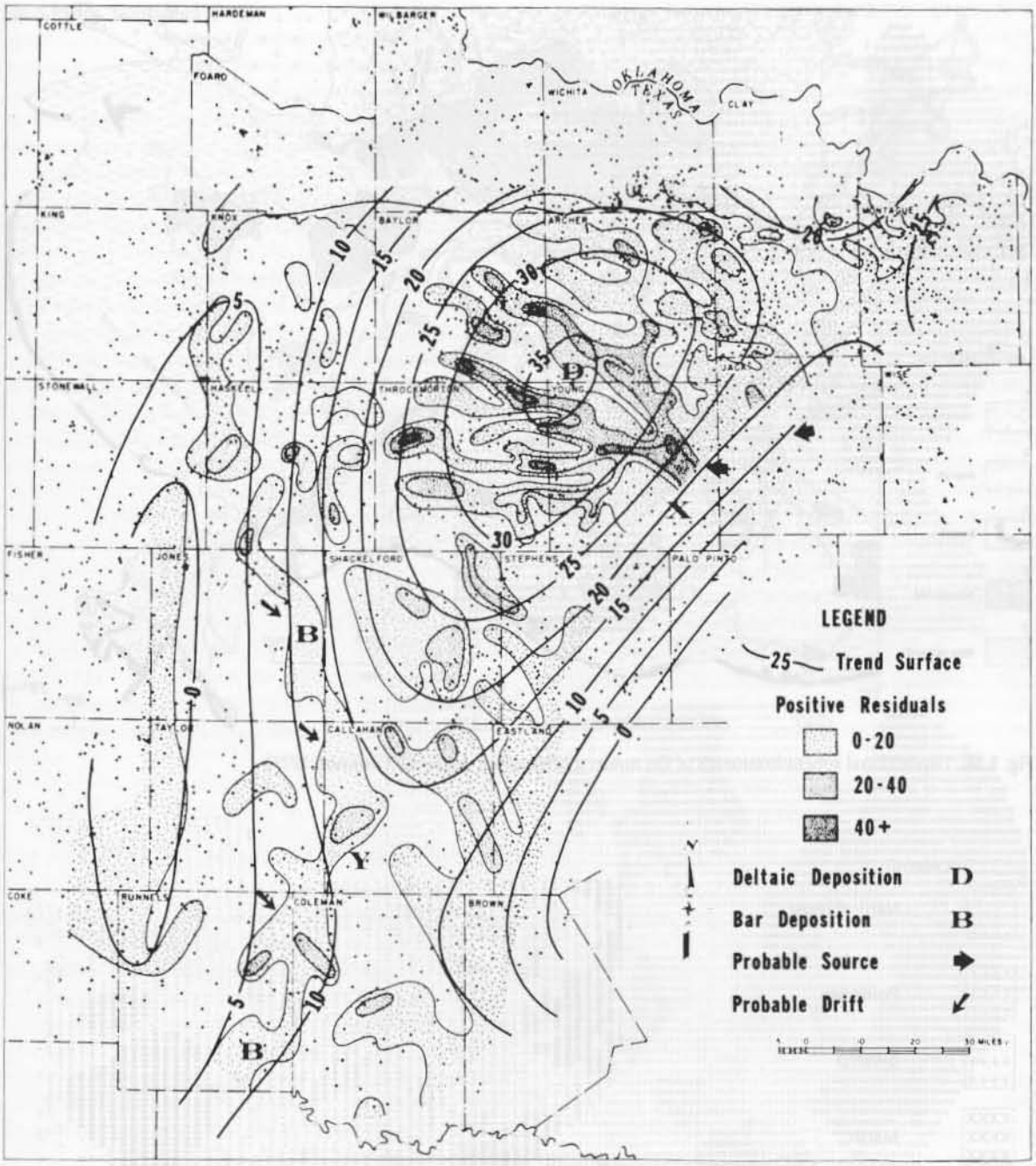


Fig. 5.57. Trend-surface analysis of a Pennsylvanian delta in north-central Texas. The deltaic depocenter is outlined by a fourth-degree surface drawn on sand thicknesses. Channel fills and other individual sandstone bodies are outlined by the residuals. (Wermund and Jenkins 1970)

next hole in order to pursue underground a particular facies type or a stratigraphic hydrocarbon or mineral prospect.

There are two sorts of questions to be considered here:

1. What sort of sample density, spacing, or distribution pattern is adequate to ensure that our lithofacies, petrographic, or seismic maps are an accurate reflection of reality?
2. Given all the mapped distribution and trend information available, what is the likely value of one or a few more sample points (i.e., exploration holes)?

These questions are the delight of statisticians, but the answers require consideration of some geological

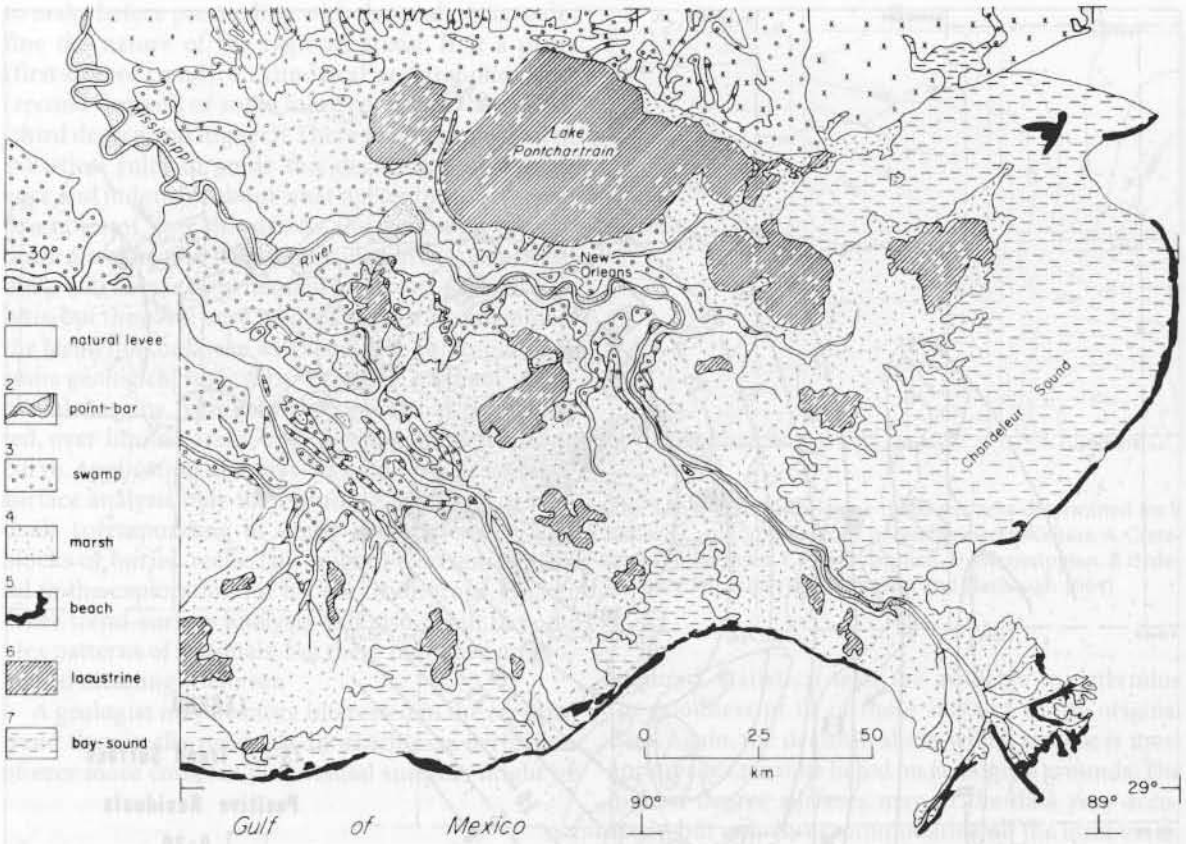
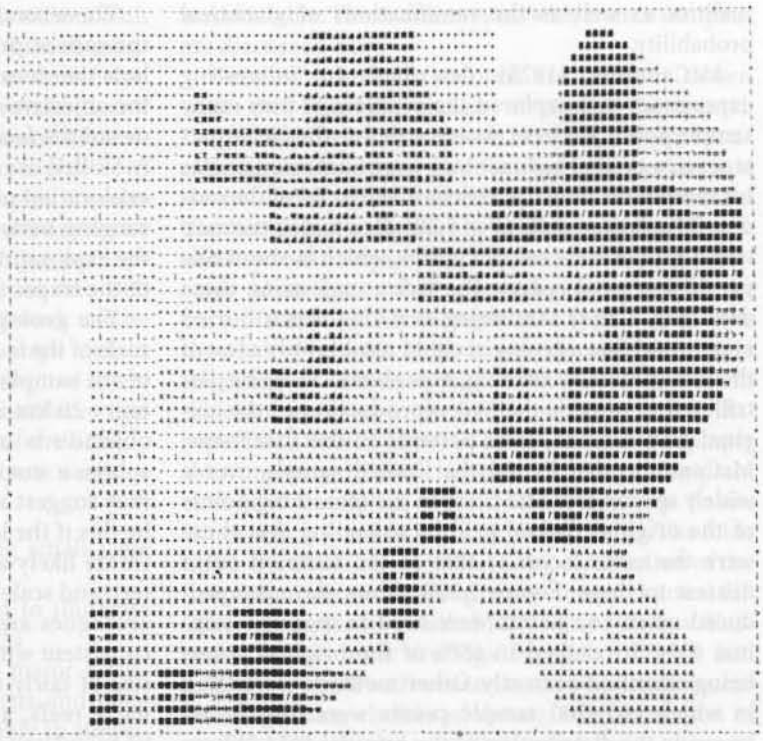


Fig. 5.58. Depositional subenvironments of the modern Mississippi delta. (McCammon 1975)

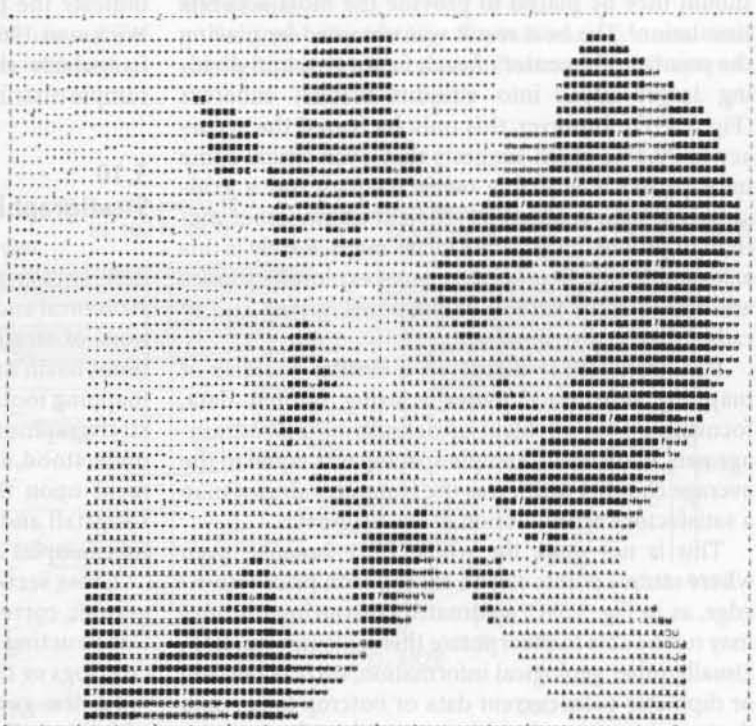


Fig. 5.59. Reproduction of the map pattern of Fig. 5.58, using a limited number of sample points. See text for explanation of the three methods used, (McCammon 1975)

Fig. 5.59b, c. Continued



b



c

realities as well as the ramifications of statistical probability.

McCammon (1975) described an interesting experiment that explored the problem of how many sample points were necessary to define the facies patterns in a typical sedimentary environment, using as his sample the fluvial and deltaic subfacies of the modern Mississippi delta (Fig. 5.58). He divided the map into a grid of 4034 sample points, which is about one per 10 km^2, and redrew the facies map using these data (Fig. 5.59A). Assuming that this was a buried system and that a geologist could sample only a few of the total points by drilling, how should these be distributed to provide the best reproduction of the original grid-derived facies pattern? In the first maps, McCammon distributed the limited sample over a widely spaced grid and filled in the remaining points of the original 10-km^2 grid by assuming that these were the same facies as that of the sampled point nearest to them. Figure 5.59B shows the map produced when 167 points were used to map the original 4034. It resulted in 65% of the original points being classified correctly. Other methods were tried in which 100–200 sample points were distributed knowing the distribution of the original 4034. Where should they be placed to provide the most accurate simulation? The best result was obtained by placing the points in the center of each facies area and dividing larger areas into equidimensional subareas (Fig. 5.59C). However, this only increased the accuracy to 70.8%, which suggests that, even where some information is available on facies patterns, a grid-sampling plan is about as efficient as any other method in producing accurate facies maps. Rarely is this achieved in practice, because most exploratory holes are drilled for a variety of purposes, giving rise to rather patchy distributions.

Robinson (1981) discussed a similar exercise of mapping subsurface patterns using limited data, focusing on the problem of delineating paleodrainage nets. Only where sample spacing was equal to the average channel width was the computer-drawn map a satisfactory reproduction of the original.

This is not quite the whole story because, even where sample points can be chosen with prior knowledge, as in Fig. 5.59C, automated contouring routines may not be able to incorporate this same information. Usually, other geological information, such as outcrop or dipmeter paleocurrent data or outcrop facies data that can be extrapolated into the subsurface, can be used to define facies trends. This is where facies-analysis techniques can be most useful. Known basement-fault control or ancient-shoreline orientation may help improve the map, and geologists also have the advantage that they can use Walther's Law and other facies-model principles to improve the interpretation.

The second of our questions is an important one in the case of prospect exploration. The one additional hole that locates an economic prospect has achieved the objective of the facies-mapping exercise, whether or not the facies patterns have been mapped correctly. Skilled use of geology and geophysics in petroleum exploration results in a ratio of successful to dry holes ranging between about 1:4 and 1:10, depending on the exploration maturity of the basin and the subtlety of the traps.

The geologist must constantly bear in mind the scale of the facies units being mapped versus the scale of the sampling pattern. For example, grid sampling every 20 km is not likely to be very successful if the objective is to map fluvial point bars and crevasse splays; a map showing lobate sand-isolith contours may suggest a simple delta pattern, but is unlikely to be this if the lobes are hundreds of kilometers across (more likely a coastal-plain complex). Constant pattern and scale comparisons with appropriate modern analogues are necessary to keep the interpretations consistent with reality. The relationship between the size of fairly regularly shaped features, such as pinnacle reefs, and the spacing of control points can be calculated statistically for sample grids in order to indicate the probability of interception (Singer and Wickman 1969; Davis 1973). This is not possible for irregularly shaped geological features or random sample distributions.

5.10 Stratigraphic Cross Sections

Reference is made at numerous points in this book to structural and stratigraphic cross sections as illustrations of stratigraphic correlations, facies interpretations, basin models, etc. Such sections are a standard mapping tool. However, there are several points about stratigraphic cross sections that are not well-enough understood, and so it would seem to be useful to comment upon them at this point. The short book by Langstaff and Morrill (1981) also provides many useful examples and ideas.

Cross sections are, of course, vital tools in stratigraphic correlation and in structural interpretation. Constructing stratigraphic cross sections from outcrop logs or the first exploratory wells in a basin provides the geologist with a kind of workbench on which to refine basin interpretations. Typically, the logs are pinned to the office wall or loosely taped to a large card, and correlation lines are drawn in using soft pencil or pins and colored thread. Days may be spent contemplating this pastiche, rearranging the correlation lines, and raising or lowering the logs to try out different stratigraphic relationships. Many

thoughtful employers now provide their offices with large corkboards or magnetic walls and small magnets to facilitate this process (and to save on redecoration bills). The results of this work commonly end up as large, folded charts tucked in at the end of company reports or published monographs.

Initially, the focus will be on stratigraphic correlation but, as analysis proceeds, cross sections can be used to study detailed facies relationships; perhaps individual lithosomes can be traced from one well section to the next. Ideas gained in this work can be used as a basis for drawing lithofacies maps and, in fact, maps and sections should be constructed interactively and simultaneously, as far as possible, so that the geologist can build up a 3-D image of the basin.

There are essentially three types of cross sections that are useful in basin-analysis work:

1. Detailed lithosome sections showing small-scale basin architecture
2. Generalized facies correlations used to illustrate depositional systems or systems tracts
3. Basin cross sections drawn to show major stratigraphic sequences, gross facies patterns, and some elements of regional structure, possibly in simplified, cartoon form

In all cases, the act of drawing up the cross section from raw data may provide the geologist with unforseen insights. These three types, with some examples, are discussed next.

Detailed lithosome sections are typically drawn with the following range of scales: horizontal, 1 cm = 40 m to 5 km; vertical, 1 cm = 5 – 300 m. Vertical exaggeration may vary for convenience, but it is always useful in order to clarify facies relationships. The horizontal scale may be arbitrary, with well sections spaced at equal intervals, but there is no particular advantage to this approach, and it may introduce distortions. An example of a type-1 section is shown in Fig. 5.60; it illustrates the complex interfingering of Mississippian shelf limestones with sabkha evaporites in the Williston Basin, North Dakota (from Wilson 1980). A different kind of example showing two deltaic channels cut into delta-plain coals and floodplain fines in the Maracaibo Basin, Venezuela, is given in Fig. 5.61 (Busch 1974). The folds below the channels were caused by compaction of the fine sediments and loading of the denser channel fill. Other examples of detailed lithosome sections illustrated in this book are Figs. 3.3 – 3.8, 3.27, 4.2, and 4.32. It is an instructive exercise to compare these, examining scales, choices of datum, documentation of section controls, styles of illustrating facies changes, and so on.

Sections illustrating depositional systems and stratigraphic sequences are illustrated in Chapter 6.

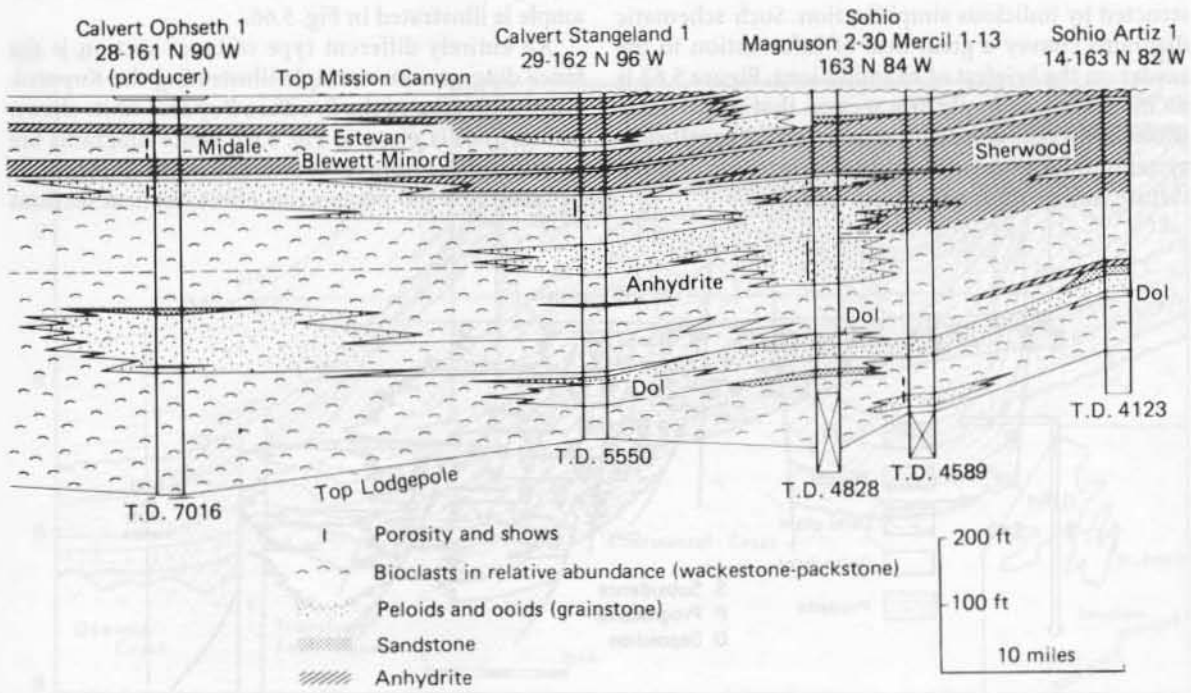


Fig. 5.60. Example of a detailed stratigraphic cross section, showing the relationship of major lithosomes, Mississippian of North Dakota. (Wilson 1980)

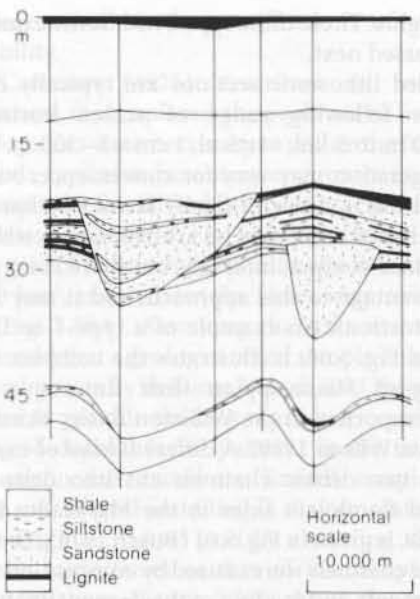


Fig. 5.61. A stratigraphic cross section revealing deltaic channels in the Maracaibo Basin, Venezuela. (Busch 1974)

The following are typical scales: horizontal, 1 cm = 3 to 30 km; vertical, 1 cm = 50–500 m. Seismic sections are often illustrated at these scales. Examples are given in Figs. 5.62 and 5.63, showing clastic wedges on the Gulf Coast and southern Africa. Both of these examples show how successful illustrations can be constructed by judicious simplification. Such schematic diagrams convey a great deal of information to the reader on the briefest of examinations. Figure 5.63 is an interesting example of a section that contrives to show three types of stratigraphy at once: depositional systems or facies relationships, stratigraphic nomenclature, and geochronologic correlations.

Small-scale basin cross sections are constructed using the following scales: horizontal, 1 cm = 5–100 km; vertical, 1 cm = 1–5 km. These are almost invariably structural cross sections but, in addition to illustrating structural style, they can be used to delineate major stratigraphic sequences and gross facies relationships. They are usually extended to the basement and may contain geophysical data pertaining to deep-crustal structure. Figure 5.64 shows the structure and stratigraphy of the continental margin off Newfoundland, based on seismic and limited well data (from McWhae 1981). Figure 5.65 illustrates the reconstructed stratigraphy of the mainly Precambrian Hecla Hoek geosynclinal trough, Svalbard (from Birkenmajer 1981). This is an example where structure has been omitted because of its complexity. Stratigraphic relations are clearly seen, yet the reader is left with an uncomfortable feeling of not knowing how close to the presently preserved reality this cross section really is.

The Svalbard illustration is one step closer to a variation of the regional reconstruction known as a schematic, stratigraphic cross section. This is a combination of a correlation table and a facies diagram. Vertical and horizontal scales are usually variable and arbitrary, although the vertical scale may be drawn using radiometrically based time or chronostratigraphic increments (e.g., stages) rather than thickness. Thus, these diagrams may contain gross spatial distortions, but they can be extremely useful in conveying correlation and facies information. An example is illustrated in Fig. 5.66.

An entirely different type of cross section is the fence diagram. An example illustrating the Kuparuk Formation (one of the Prudhoe Bay area units shown in Fig. 5.66) is given in Fig. 5.67. Fence diagrams are an attempt to improve on conventional 2-D diagrams by creating a 3-D perspective effect. Vertical sections

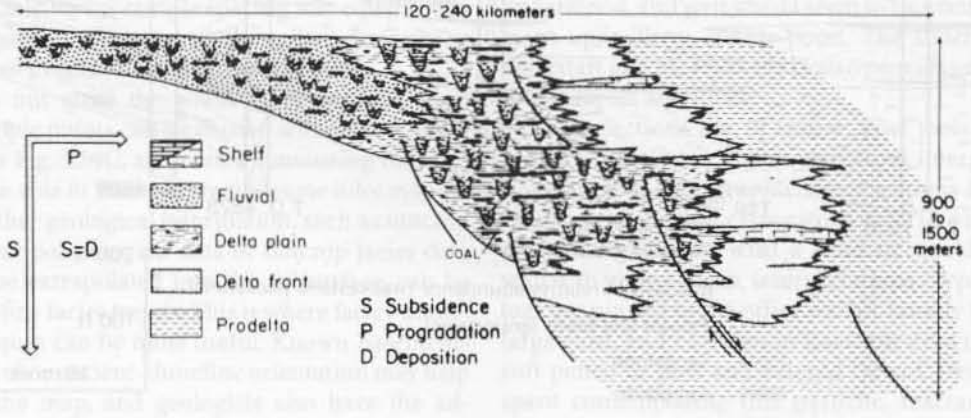


Fig. 5.62. Generalized stratigraphic cross section through the Eocene sediments of the Gulf Coast, showing inter-relationships of the various depositional systems. (Brown and Fisher 1977)

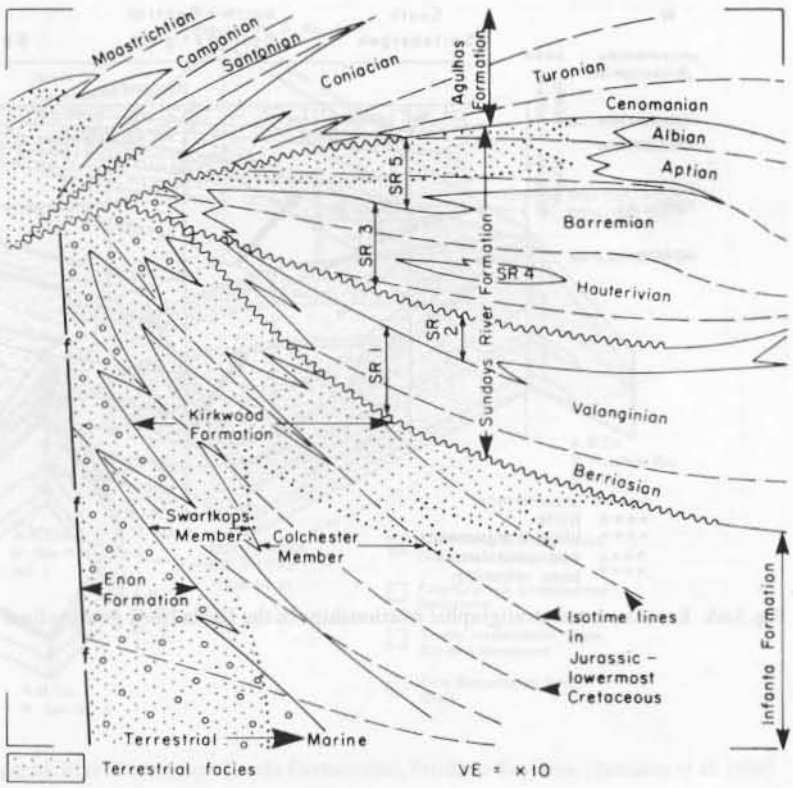


Fig. 5.63. Generalized stratigraphic relationships of Jurassic–Cretaceous strata, southern Africa, showing depositional systems, stratigraphic nomenclature, and geochronology. (Winter 1979)

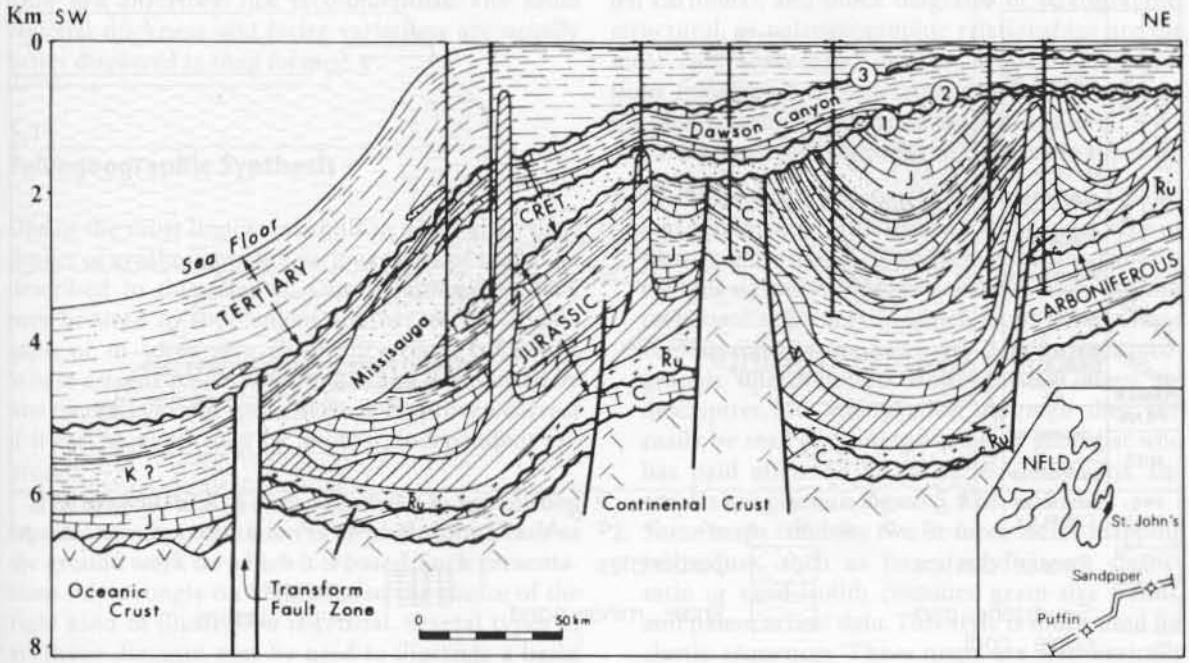


Fig. 5.64. Generalized basin cross section showing schematic structure and facies, continental margin off Newfoundland (McWhae 1981)

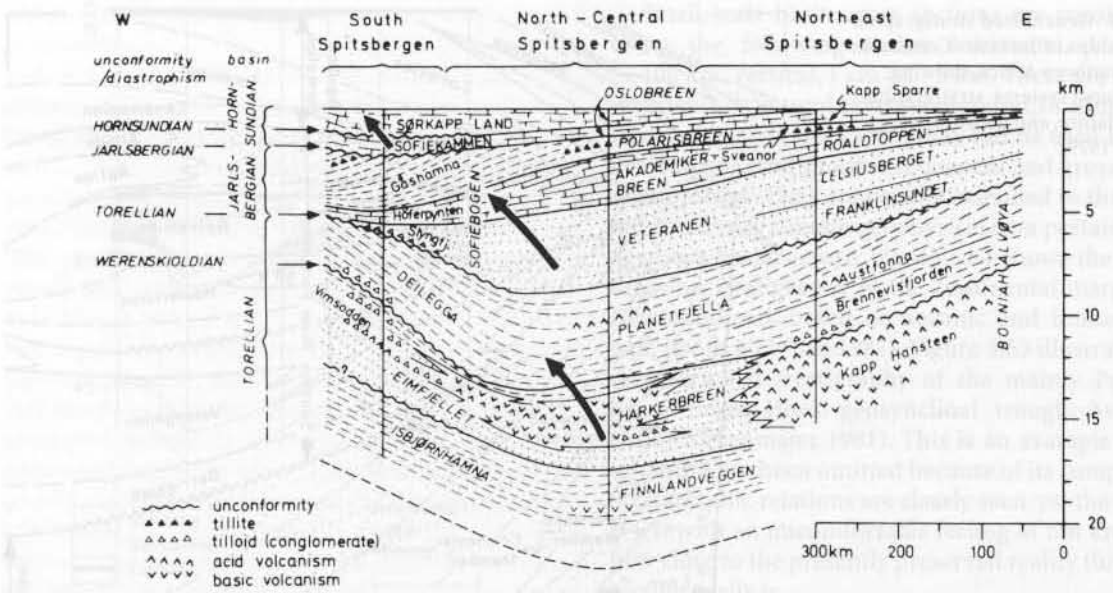


Fig. 5.65. Reconstructed stratigraphic relationships in the Hecla Hoek geosynclinal trough, Svalbard. (Birkenmajer 1981)

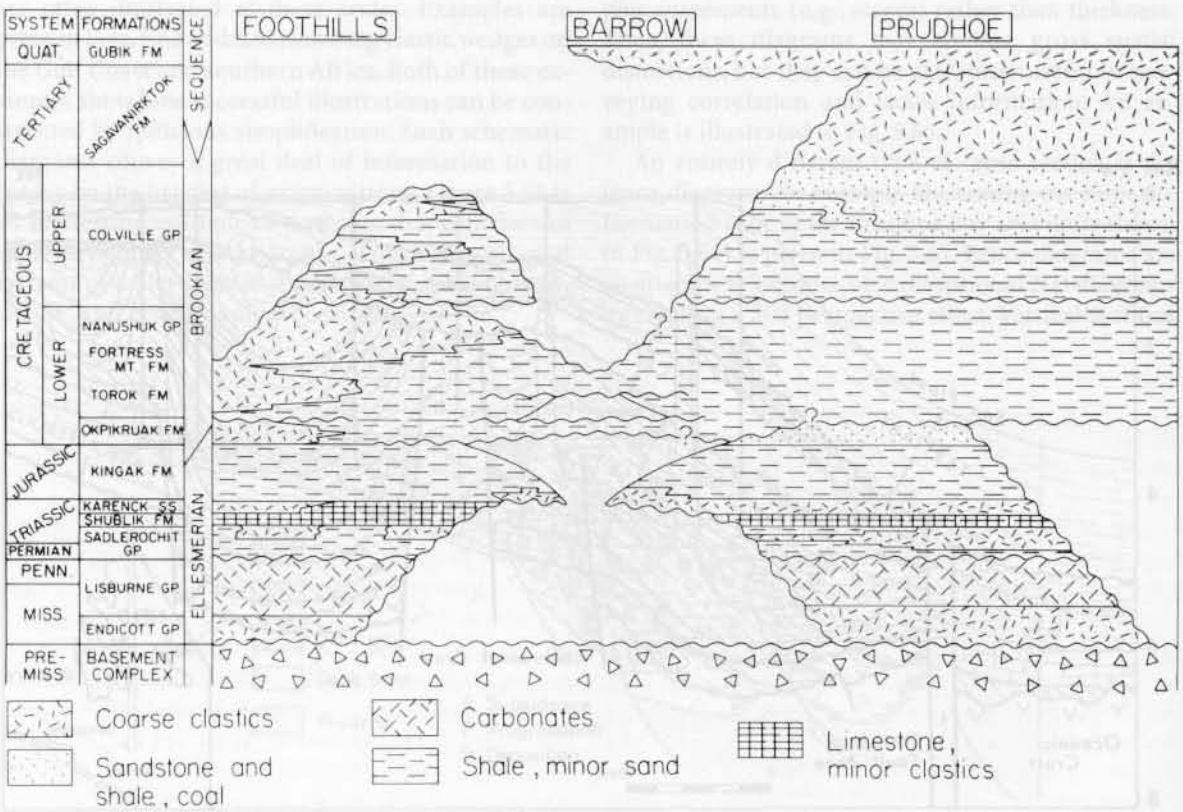


Fig. 5.66. Example of a schematic stratigraphic cross section or generalized correlation diagram showing lithofacies assemblages, North Slope of Alaska. (Jamison et al. 1980)

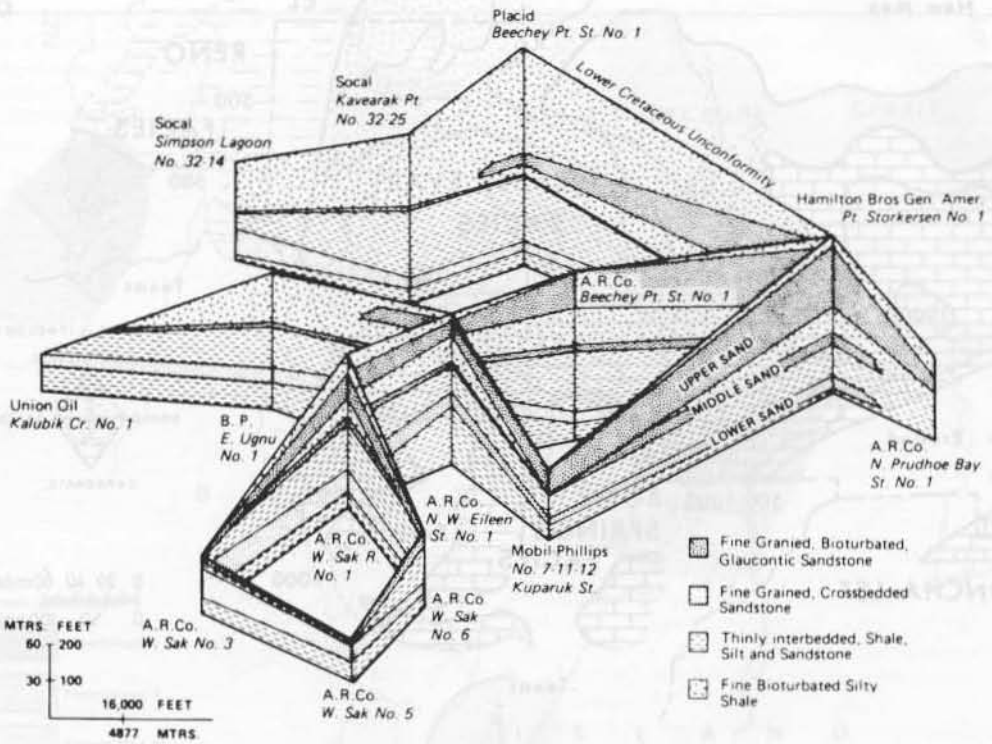


Fig. 5.67. Example of a fence diagram, Kuparuk River Formation (lower Cretaceous), Prudhoe Bay area. (Jamison et al. 1980)

are placed in their correct relative positions as far as is possible. These diagrams are hard to read, because parts of the section are hidden by "fences" in front. They are, therefore, not recommended. The same regional thickness and facies variations are usually better displayed in map form.

5.11 Paleogeographic Synthesis

One of the most important skills in basin analysis is the art of synthesizing the various types of map data described in this chapter. One mapping approach may be used to supplement another, to fill in data gaps, or to serve as a check on overall reliability. Where a vague paleogeographic trend is indicated by one method, it is far more likely to be proved correct if it can be confirmed by another, independent approach.

The presentation of such a synthesis in a published report or oral presentation is as important a task as the ground work on which it is based. Such presentations rely strongly on diagrams, so the choice of the right kind of illustration is crucial. Several types of synthesis diagram may be used to illustrate a basin model, and experienced researchers know that the work of designing and compiling such diagrams commonly provides them with valuable new insights. Judicious simplification may be very useful.

Interpretive maps, cross sections (commonly called cartoons), and block diagrams of stratigraphic, structural, or paleogeographic relationships are the most commonly used devices. Sequential illustrations may be used to demonstrate basin evolution through tectonic events or changes in sea level.

Paleogeographic maps are of three main types:

1. There are maps showing basic facies data, labeled and ornamented with regard to lithofacies type or stratigraphic name. Isopachs may be superimposed. This style of map is particularly widely used for carbonate sequences in which facies subdivisions contain most of the necessary data for paleogeographic interpretation. However, such maps are descriptive, not interpretive, although they can easily be read and interpreted by a geologist who has paid attention to the facies definitions. Examples are given in Figs. 5.5, 5.68, and 5.69.
2. Some maps combine two or more facies-mapping techniques, such as facies subdivisons, clastic-ratio or sand-isolith contours, grain-size trends, and paleocurrent data. This style is much used for clastic sequences. These maps are also basically descriptive and may need explanatory captions. Examples are illustrated in Figs. 5.70 and 5.71.

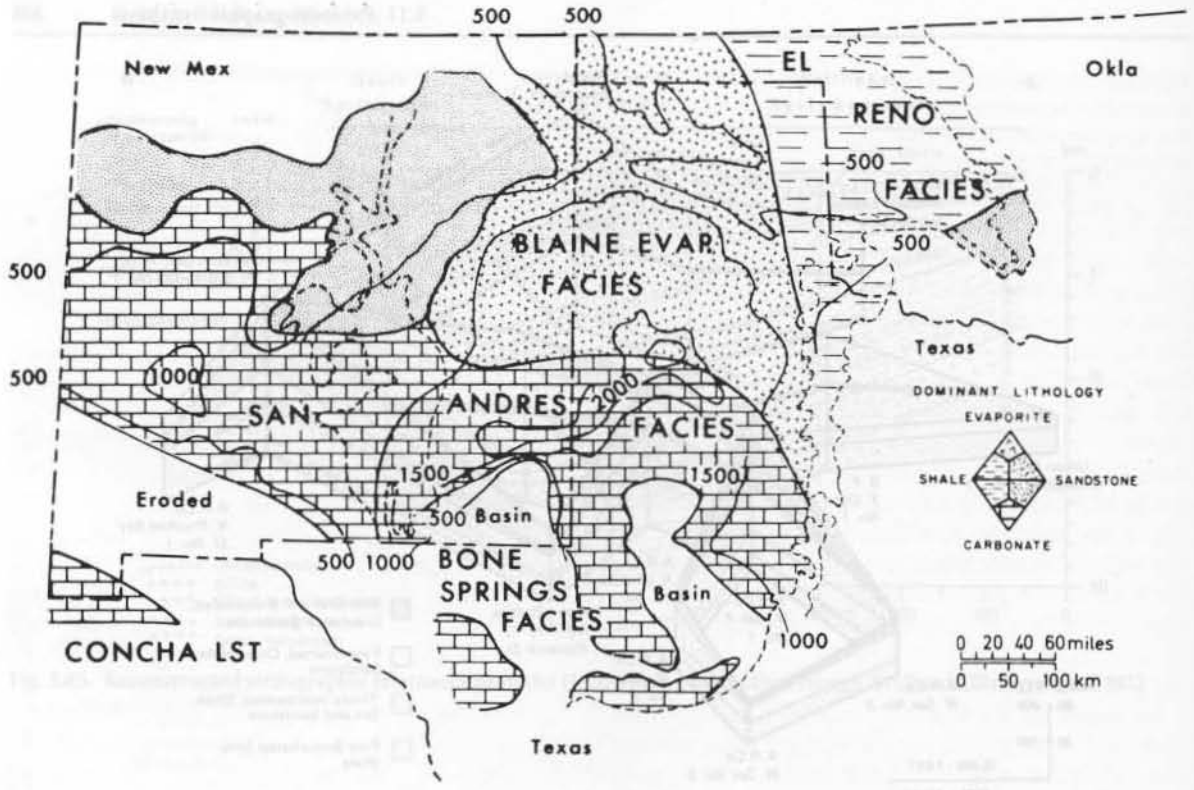
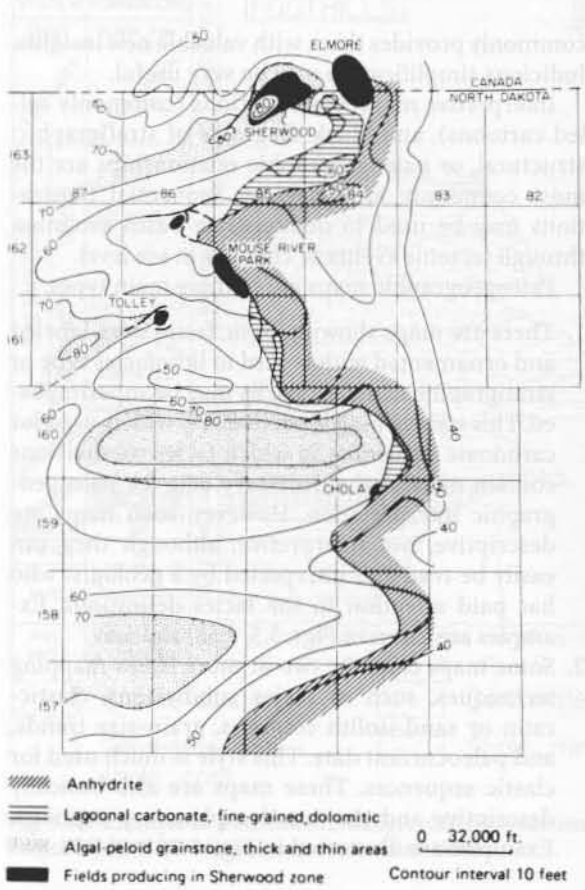


Fig. 5.68. Lithofacies and isopach map of middle Permian sediments, Texas and New Mexico. The San Andres and Concha Formations are shelf limestones, and the Bone Springs is a black, laminated, basinal limestone. (Meissner 1972)



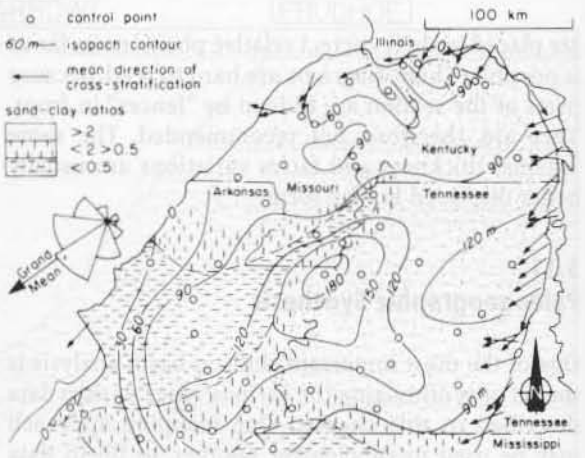


Fig. 5.70. Isopach, lithofacies, and paleocurrent data of upper Cretaceous McNairy Formation, upper Mississippi embayment. Paleocurrent data represent vector means for 35 sampling areas based on 920 readings made on basin-margin outcrops. A fluvial system is interpreted to have fed the basin from the northeast, and the area where the sand:clay ratio is greater than 2 represents a delta plain, passing into a prodelta region in the south. This map is from one of the earliest integrated basin-analysis studies in North America. (Pryor 1960)

Fig. 5.69. Lithofacies and isopach map of Mississippian sediments, North Dakota. Linear shoals of grainstone pass eastward (updip) into evaporite, creating stratigraphic petroleum traps. Figure 5.60 is an east–west cross section through the north part of this map area. (Wilson 1980)

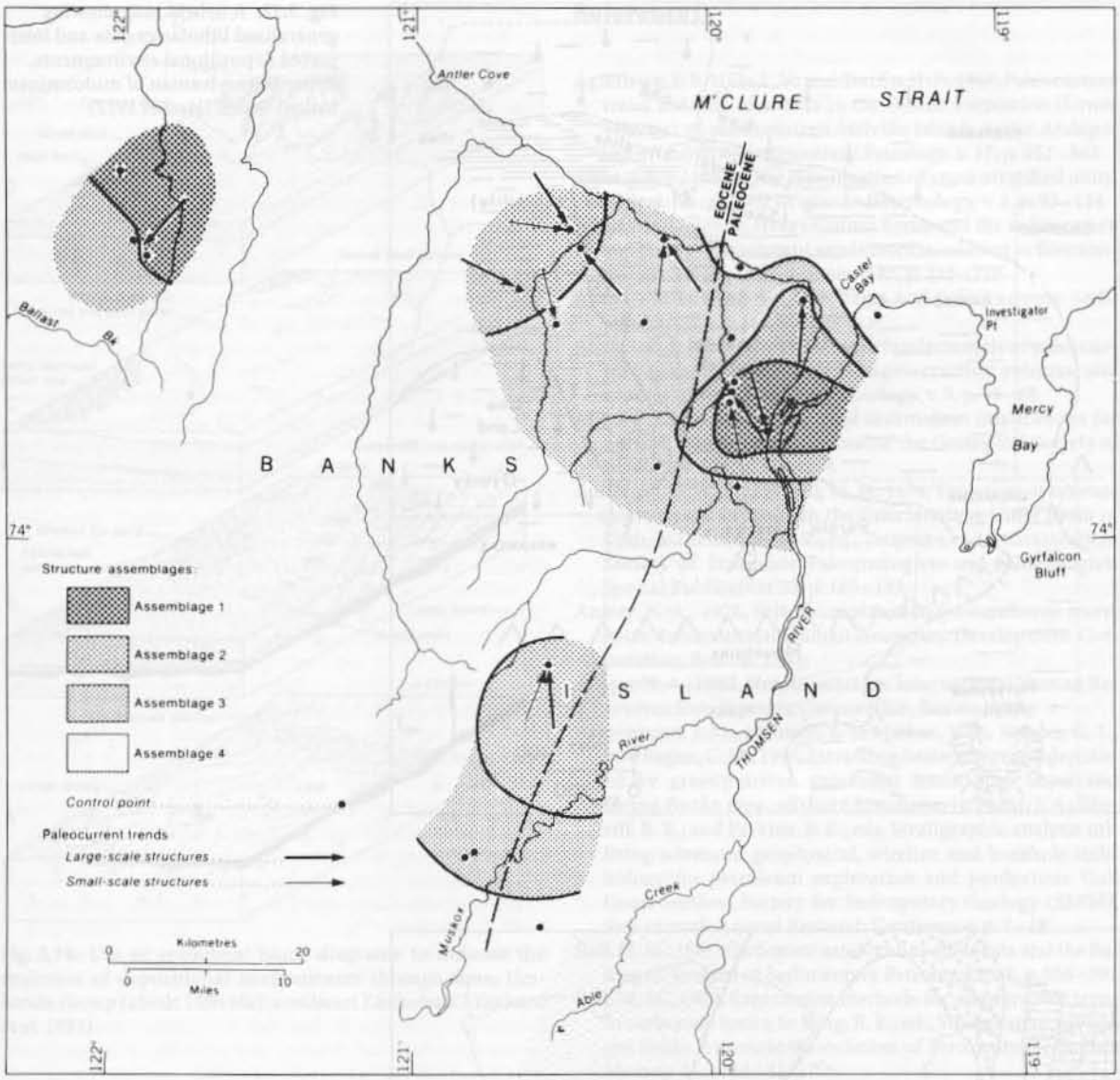


Fig. 5.71. Mapping bird's-foot-shaped deltas in the Eureka Sound Formation (Paleocene–Eocene), Banks Island, Arctic Canada. Structure assemblages are: *1* large-scale crossbedding, including epsilon sets (delta plain); *2* planar and trough crossbedding (proximal delta front); *3* ripple marks (distal delta front); *4* few current structures (prodelta). Paleocurrent patterns appear to be random but coincide well with the mapped distribution of sedimentary structures. (Miall 1979)

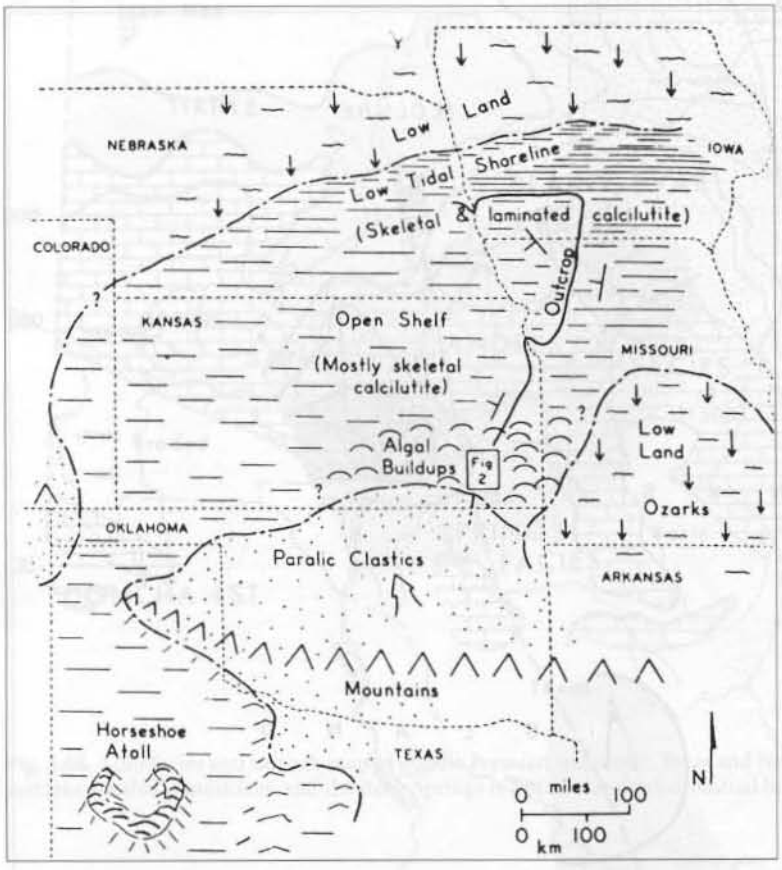


Fig. 5.72. A hybrid map showing generalized lithofacies data and interpreted depositional environments, upper Pennsylvanian of midcontinent United States. (Heckel 1972)

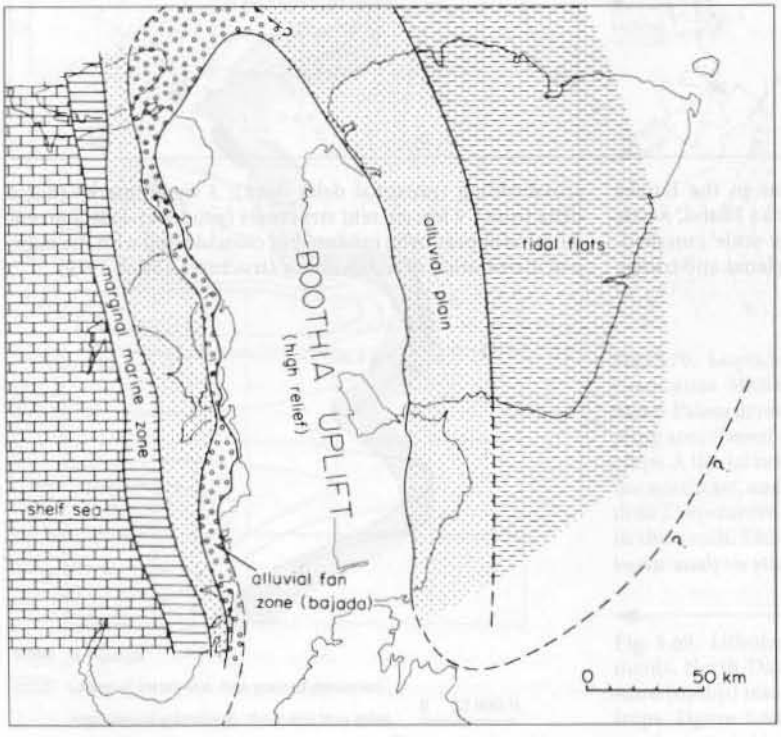


Fig. 5.73. Paleogeographic map for Somerset Island and eastern Prince of Wales Island during the earliest Devonian, showing the detrital apron flanking the Boothia Uplift. (Miall and Gibling 1978)

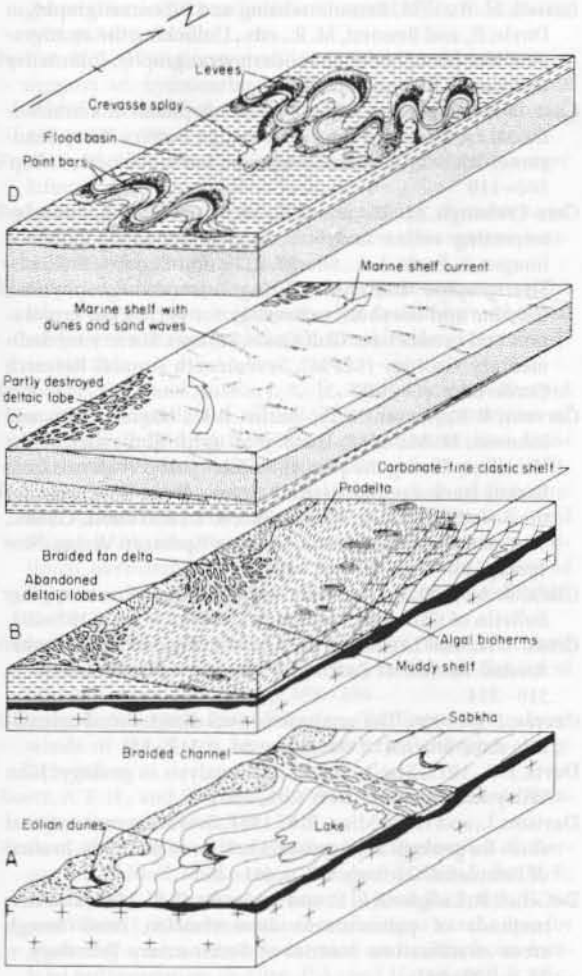


Fig. 5.74. Use of sequential block diagrams to indicate the evolution of depositional environments through time, Umkondo Group (about 1800 Ma), southeast Zimbabwe. (Tankard et al. 1982)

3. There are also interpretive maps, in which descriptive data are very generalized or may be omitted altogether. Two examples are provided in Figs. 5.72 and 5.73.

Block diagrams are very useful devices for summarizing paleogeographic interpretations. They are drawn to schematically show depositional environments, and an artist's touch is needed to make them look convincing. Some examples are shown in Fig. 5.74, which illustrates another valuable technique, the use of sequential diagrams to demonstrate changing conditions during the formation of a major stratigraphic unit. Care should be taken, using an index map, to indicate what particular piece of earth space is covered by the diagrams.

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Sequence Stratigraphy

6.1 Introduction

Sequence stratigraphy developed initially as a tool for stratigraphic correlation (Sloss et al. 1949; Sloss 1963). The sequence-stratigraphic approach to basin analysis became widely used after Peter Vail and his Exxon colleagues demonstrated that sequences could be defined and correlated by the use of reflection-seismic data. Since the ground-breaking publication by Vail et al. (1977), sequence stratigraphy has evolved into one of the central, theoretical underpinnings of the science of stratigraphy. It comprises a set of predictions that can be made about the facies and architecture of sediments formed in a particular range of environments under specified allogenic controls. It is the predictability of sequences that give them their value for regional correlation and for one of the main practical applications of such correlation – petroleum exploration. Sequence stratigraphy represents a relatively new paradigm in geology. The development of this paradigm has completely revolutionized stratigraphy since the 1980s. The principal hypothesis is that stratigraphic successions may be subdivided into discrete sequences bounded by widespread unconformities. There are two parts to the hypothesis.

First, it is suggested that the driving force which generates sequences and their bounding unconformities, be it eustatic sea-level change, tectonic movements, or some other process, also generates predictable three-dimensional stratigraphic architectures. In recent years, stratigraphic research, guided by sequence models, has brought about fundamental improvements in our understanding of stratigraphic processes and the controls of basin architecture. Sequence models have provided a powerful framework for mapping and numerical modeling, enabling the science of stratigraphy to advance with rapid strides. Nummedal (1987) wrote:

> The interpretation of the stratigraphic record has been greatly stimulated over the past few years by rapid conceptual advances in "sequence stratigraphy," i.e., the attempt to analyze stratigraphic successions in terms of genetically related packages of strata. The value of the concept of a "depositional sequence" lies both in the recognition of a consistent three-dimensional arrangement of facies within the sequence, the facies architecture, and the regional (and inter-regional) correlation of the sequence boundaries. It has also been argued that many sequence boundaries are correlatable globally, and that they reflect periods of sea-level lowstand, i.e. sequence-boundaries are subaerial erosion surfaces. (Nummedal 1987)

The major steps forward in the development of sequence-architecture models are also attributed to the work of the Exxon research group. The pair of papers by Posamentier et al. (1988) and Posamentier and Vail (1988) and the book by Van Wagoner et al. (1990) set out most of the basic ideas upon which present-day models have been built. Despite the many controversies that have developed as other workers have set out to apply sequence concepts, it is important not to lose sight of the fundamental changes in stratigraphic geology that were made by the Exxon group, particularly Peter Vail and his colleagues, including Bob Mitchum, Henry Posamentier, and John Van Wagoner. Much credit should accrue to these individuals.

This chapter is concerned with the details of sequence architecture and the sequence models that have emerged as powerful tools for subsurface description and prediction. The study of sequence stratigraphy has demonstrated the importance of a wide range of processes for the generation of cyclic sequences, including eustasy, tectonics, and orbital forcing of climate change. These are discussed in Chapter 8.

To a considerable extent, the great strides that have been made in the development of sequence stratigraphy have been obscured by a controversy emerging from the second part of the sequence stratigraphic paradigm, referred to at the end of the above quote. Vail et al. (1977) proposed that sequences are superior to other forms of chronostratigraphic data as chronostratigraphic indicators, and that they therefore comprise the ideal basis for a superior standard

of geologic time. The main theoretical basis for the paradigm is the supposition that global stratigraphic successions are controlled primarily by eustatic sea-level changes. However, as discussed in Chapter 8, the basic premise of the paradigm remains unproved.

The currently accepted definition of **sequence** is that proposed by Vail et al. (1977). It is a "stratigraphic unit composed of a relatively conformable succession of genetically related strata ... bounded at its top and base by unconformities or their correlative conformities". However, the term "sequence" has had a checkered history. Sloss (1963) defined stratigraphic sequences as "rock-stratigraphic units of higher rank than group, megagroup, or supergroup, traceable over major areas of a continent and bounded by unconformities of interregional scope." This usage has not received universal acceptance. In some literature, "sequence" has been used as a synonym for "succession", and it has been employed for cyclic or unconformity-bounded units of varying dimensions and for time-spans ranging from thousands of years to hundreds of millions of years. In view of the importance of sequence concepts to stratigraphy, it is now recommended that the term "sequence" only be used to refer to unconformity-bounded successions, and that such synonyms as "succession" be used, where necessary, to refer informally to thicknesses of strata in other contexts where sequence concepts are not being invoked. For example, the deposits of a delta lobe on a continental margin should be referred to as a deltaic succession, not a deltaic sequence.

A useful history of the development of modern sequence stratigraphy was provided by the main protagonist, L. L. Sloss (1988), who described the evolution of the field based on the studies by himself, his colleagues, and his students, of the cratonic sedimentary cover of North America. His most important early paper (Sloss 1963) established the existence of six unconformity-bounded sequences (Chap. 8), which he named using Indian tribal names (they are now sometimes referred to as **Sloss sequences**) in order to distinguish them from the conventional litho- and chronostratigraphic subdivisions, the latter having been mainly imported from Europe. This paper built on earlier work by Sloss et al. (1949) and was paralleled independently by Wheeler (1958, 1959a,b, 1963), who is credited with the popularization of chronostratigraphic charts, in which stratigraphic cross sections are plotted with a vertical time axis rather than a thickness axis. These diagrams are now referred to as **Wheeler diagrams**. They are a useful technique for demonstrating the range in age of stratigraphic units and their bounding unconformities from place to place.

Many of the ideas incorporated into current models of continental-margin sequence architecture were developed by workers analyzing the Cenozoic record of the Gulf and Atlantic coasts of the United States. Curray (1964) was among the first to recognize the relationships between sea level and sediment supply. He noted that fluvial and strandplain aggradation and shoreface retreat predominate under conditions of rising sea level and low sediment supply, whereas river entrenchment and deltaic progradation predominate under conditions of falling sea level and high sediment supply (Morton and Price 1987). Frazier (1974) subdivided the Mississippi deltaic successions into transgressive, progradational, and aggradational phases and discussed autogenic (delta switching) and glacio-eustatic sedimentary controls (Figs. 6.1, 6.2). Brown and Fisher (1977), in a paper that actually deals with seismic data and appears in American Association of Petroleum Geologists (AAPG) Memoir 26, summarized the ideas of an important group of stratigraphers at the Bureau of Economic Geology, University of Texas (including W. L. Fisher, L. F. Brown Jr., J. H. McGowen, W. E. Galloway, and D. E. Frazier), which later became an integral part of the Exxon sequence-stratigraphy interpretive framework: the use of regional facies concepts to define depositional systems and systems tracts.

Sequence stratigraphy remained a subject of relatively minor, academic interest throughout the 1960s and 1970s (Ross 1991) until improvements in reflection-seismic acquisition and processing procedures, and conceptual breakthroughs in interpretation led Peter Vail (a former graduate student of Sloss) and his colleagues at Exxon Corporation to discover that a wealth of stratigraphic detail could be interpreted from seismic sections, in addition to the structural data for which seismic data had long been used. Over a period of more than a decade in the 1960s and 1970s, Vail and his coworkers studied seismic-reflection data, and the methods and results eventually appeared in AAPG Memoir 26, a major publication that is now required reading for all advanced students of stratigraphy (the history of development of these ideas is described in a personal memoir by Vail 1992). Seismic stratigraphy makes use of the concept that seismic reflections parallel bedding surfaces and are, therefore, of chronostratigraphic significance (with a few significant exceptions), thus enabling widespread correlation to be readily accomplished (although, because of the relatively coarse resolution of the seismic method, such correlations are not necessarily very precise). This contrasts with the correlation techniques used in older methods of conventional outcrop basin analysis, in which lithofacies contacts are known to be typically diachronous. The resulting lithostratigraphic classification may have limited local applicability and can only be integrated into a reliable chronostratigraphic framework with consid-

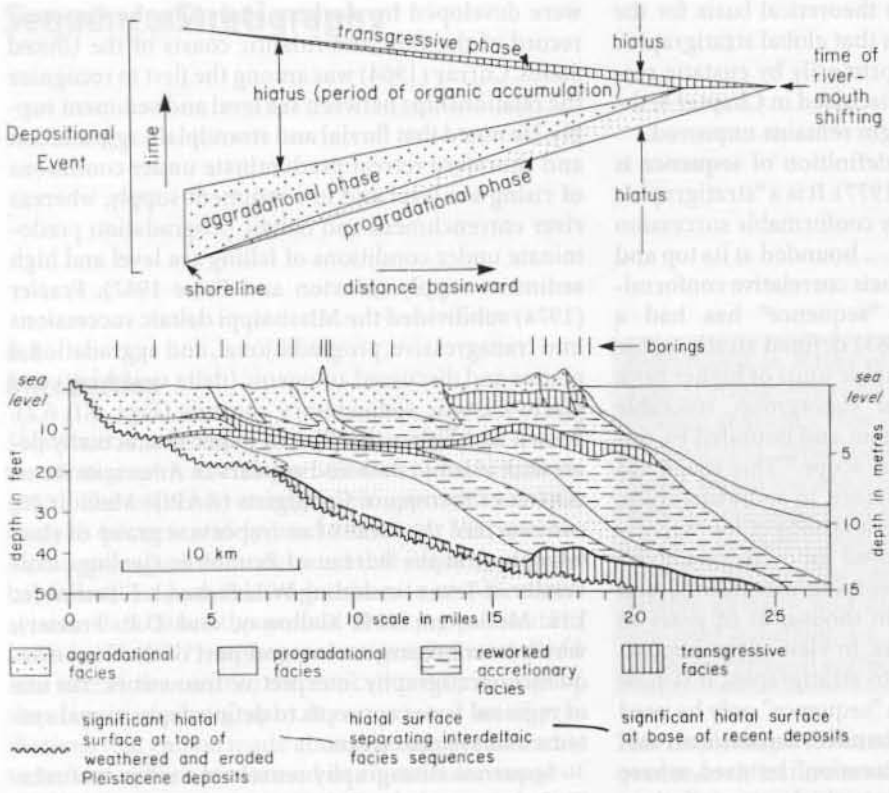


Fig. 6.1. The sequence-stratigraphic concepts of Frazier (1974), based on analysis of the Mississippi Delta (below)

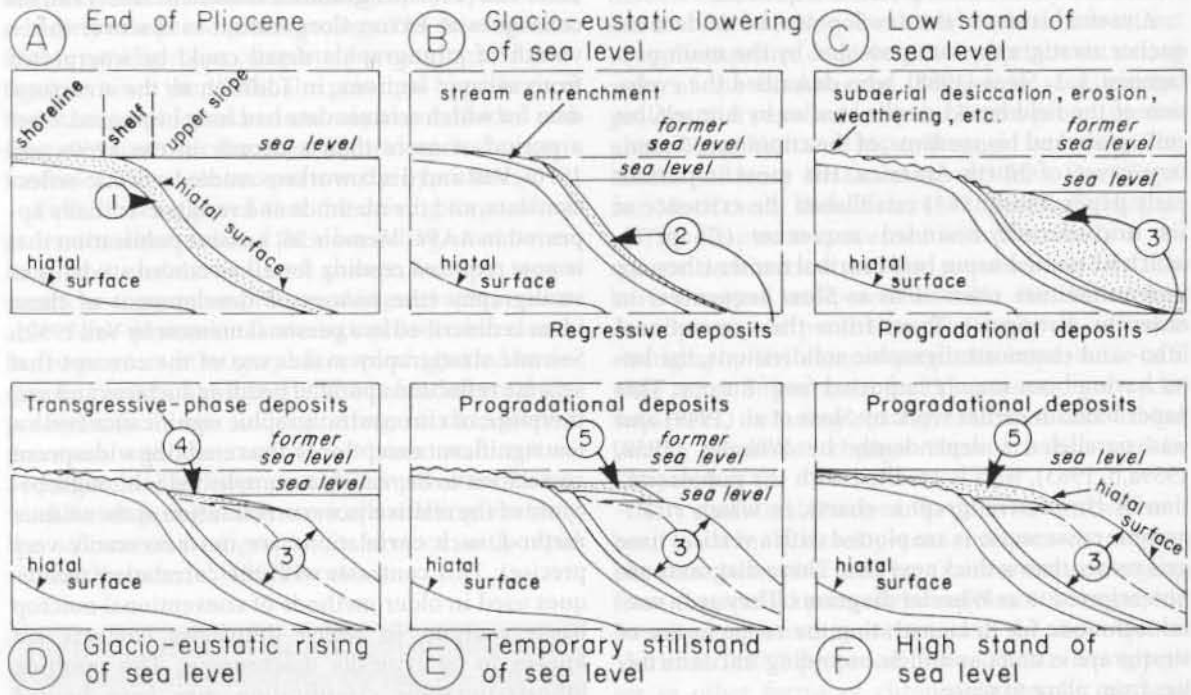


Fig. 6.2. A depositional event in the Mississippi Delta, as shown by Frazier (1974). Holocene sediments contain several or many such events as a result of the rapid changes in sea level during the ice age

erable effort. The practical utility of the concepts of sequence stratigraphy for basin studies and regional, even global, correlation (Payton 1977) was quickly realized, and sequence stratigraphy commenced a rapid rise in popularity that has yet to abate.

Seismic stratigraphy permitted two major practical developments in basin analysis: the ability to systematize and define complex basin architectures in considerable detail and the ability to recognize, map, and correlate unconformities over great distances. Architectural work led to the development of a special terminology for defining the shape and character of stratigraphic surfaces. A particular emphasis came to be placed on the nature of bedding terminations because of the significance these carry with regard to the processes of progradation, aggradation, and erosion. Seismic work by Vail and his team eventually led to the recognition of the sequence-stratigraphic model for the interpretation of seismic records and the building of regional and global stratigraphic syntheses. The methods have become immensely popular and are now, in fact, the standard method for regional stratigraphic analysis.

The methods of seismic sequence stratigraphy were documented in detail by Vail (1987) and, subsequently, two important papers and a book by members of the Exxon team demonstrated how sequence concepts could be applied to the interpretation of outcrops, petrophysical logs, and cores (Posamentier et al. 1988; Posamentier and Vail 1988; Van Wagoner et al. 1990). This body of work has also been enormously influential.

The history of sequence stratigraphy has been dealt with in a number of recent publications from several different points of view. Sloss (1988) provided a personal view. Ross (1991) pointed out that all the essential ideas that form the basis for modern sequence stratigraphy were in place by the 1960s. Dott (1992) edited a collection of papers discussing the development of the concept of eustasy. Miall (1997) provided a detailed review of the entire field covered here in Chapters 6 and 8, with particular emphasis on the controversies and remaining questions surrounding sequence development and correlation. Emery and Myers (1996) provided a detailed description of modern sequence-architecture models and depositional processes, but they adhered closely to the original Exxon models and do not discuss any of the controversial areas of the subject.

In the discussion that follows, it is assumed that the reader is familiar with a basic knowledge of facies analysis and facies models (Chap. 4). As noted in Section 4.6, modern treatments of facies models have been provided in several recent textbooks, so that material is not covered here.

6.2 Stratigraphic Architecture

The concept of **accommodation space** is fundamental in sequence stratigraphy. Jervey (1988) defined it as "the space made available for potential sediment accumulation [where] in order for sediments to be preserved, there must be space available below base level (the level above which erosion will occur)." (The word "accommodation" implies space; therefore, the use of the word "space" in the term "accommodation space" is, strictly speaking, superfluous.) In the marine environment, sea level corresponds closely to base level, although some processes (beach regression, delta progradation, reef growth) can generate limited thicknesses of deposit (which have low long-term preservation potential) above sea level. In nonmarine settings, the profile of a graded river is the surface that controls sedimentary accommodation. Here, the fluvial response to erosive processes and the input of new sediment is to maintain a profile in a state of dynamic equilibrium. Sequence architecture and the development of the various systems tracts depend primarily on the changes in accommodation generated by external forces, including eustasy, tectonic uplift and subsidence, and climate change. In broad terms, such processes as progradation, aggradation, and retrogradation depend on the rate of change of accommodation relative to the rate of sediment supply.

The law of superposition of strata is one of the oldest principles of sedimentary geology (dating back to Steno in the 17th century). It states that, in any succession of strata, the beds were deposited in a simple vertical order from base to top. This very obvious idea is the basis of the layer-cake model of stratigraphy, which sees sediments as more or less uniform blankets or wedges, although perhaps showing thickness or facies changes from one part of a basin to another. Clearly, on the large scale, at the level of the system or series (Table 3.1), this is generally true. However, it overlooks an important point that has become increasingly clear as our knowledge of depositional processes and our ability to construct detailed stratigraphic cross sections have improved. This point is that, in many environments, lateral accretion (progradation) of sediments is as important or more important than vertical aggradation. In a few environments, such as some cratonic carbonate shelves, evaporite basins and pelagic seas, sedimentation is uniform and develops stratigraphic units with an essentially tabular geometry (Sect. 3.4.3.3). However, in most other environments, distribution of the sediment is constrained by some factor, such as river grade, base level, wave base, or depth of marine light penetration, and sediment piles build out as well as (or instead of) up. This is true of alluvial fans, sub-

marine fans, prograding shorelines, and many carbonate banks, but is nowhere so clearly illustrated as in prograding deltas and on continental shelves, which show rapid lateral accretion on dipping **clinoform** surfaces (Fig. 5.19). Stratigraphic superposition is a complex process, and it results in a diversity of styles of stratigraphic architecture. For example, time increments represented by thick clinoform units on the prograding delta or shelf may be represented by a thin reworked zone or even a hiatus on the delta plain or the shelf platform, respectively. Also, the locus of progradation may switch laterally as a result of slope changes or channel diversion, so that a clinoform set may be followed laterally by a hiatus, with the succeeding interval of time being represented by strata elsewhere. Sediment depocenters and stratigraphic relationships are markedly affected by subsidence and uplift and by changes in sea level, all of which generate offlap and onlap patterns. Unconformity surfaces may or may not be of regional significance. Many are temporary zones of sediment bypass or subaerial exposure during episodes of low relative sea level. They may or may not be related to tectonic movements or to eustatic sea-level changes. However, relative sea-level changes are extremely common in the geologic record (Chap. 8).

Seismic stratigraphers have developed a range of terms to describe bedding relationships within the stratigraphic packages that are formed during individual depositional episodes. These are illustrated in Fig. 6.3 (from Mitchum et al. 1977); the seismic facies patterns shown in Figs. 5.19 and 5.20 should be added to these.

Onlap and offlap are familiar terms and have been precisely defined by Mitchum (1977): **onlap** is "a base-discordant relation in which initially horizontal strata terminate progressively against an initially inclined surface, or in which initially inclined strata terminate progressively updip against a surface of greater initial inclination." **Offlap** is the progressive offshore shingling of sedimentary units within a conformable sequence, in which each successively younger unit leaves exposed a portion of the older unit on which it lies (Gary et al. 1972). These are both varieties of **lapout**, a general term for "lateral termination of strata at their depositional pinchout" (Mitchum 1977). At the upper boundary of a depositional package, this is termed **toplap**: "termination of strata against an overlying surface mainly as a result of nondeposition (sedimentary bypassing) with perhaps only minor erosion" (Mitchum 1977). It should be remembered that seismic resolution is such that thin beds cannot be discriminated, and apparent toplap may actually be a perfectly conformable, but attenuated, sequence (Sect. 5.4.1).

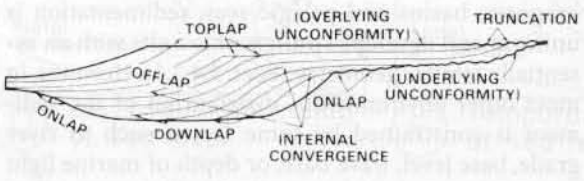


Fig. 6.3. Terminology for stratigraphic terminations, as developed by Mitchum et al. (1977) from seismic data

At the base of a depositional package, lapouts are termed **baselap**. Varieties of this pattern are termed onlap (defined above), downlap, and uplap. **Downlap** is "a base-discordant relation in which initially inclined strata terminate downdip against an initially horizontal or inclined surface" (Mitchum 1977). **Uplap** is a term used by Brown and Fisher (1977) to refer to depositional terminations against growth structures, such as active diapirs, graben faults, or growth faults.

The geometry or architecture of a depositional package or stratigraphic sequence is controlled by the interrelationship of three processes:

1. The rate of sediment supply (clastic) or sediment generation (chemical)
2. The rate of basin subsidence
3. The rate of eustatic sea-level change

Facies changes, lapout relationships, and hiatuses would provide a complete picture of the interrelationship of the first two processes were it not for the complications introduced by the third. Much recent work (Chap. 8) shows that long-term and short-term changes in global sea level have probably been occurring almost continuously, at least during the Phanerozoic. However, documentation of the timing and amount of such changes is far from complete, and we are left with a great potential for confusion in unraveling the local stratigraphic record. For example, given adequate sediment supply and relatively slow subsidence, it is possible for sedimentary progradation and regression to occur, even during a rise in sea level (Pitman 1978). Conversely, rapid subsidence of a sediment-starved shelf could cause a transgression even during a time of falling sea level. It is, therefore, important to attempt to distinguish local events that can be interpreted directly from the sedimentary record, such as transgression and regression, from their ultimate cause, which is generally much more difficult to establish. Figure 6.4 illustrates the relationship between the rates of deposition and subsidence in a deltaic complex, assuming stationary sea level (from Curtis 1970). This should be compared with Fig. 6.2, which shows a model based on sea-level change. There are a few techniques that can be used to assess whether sea-level changes are local, regional, or global in scope. Where apparent changes in sea level can be precisely correlated regionally or globally, it is safe

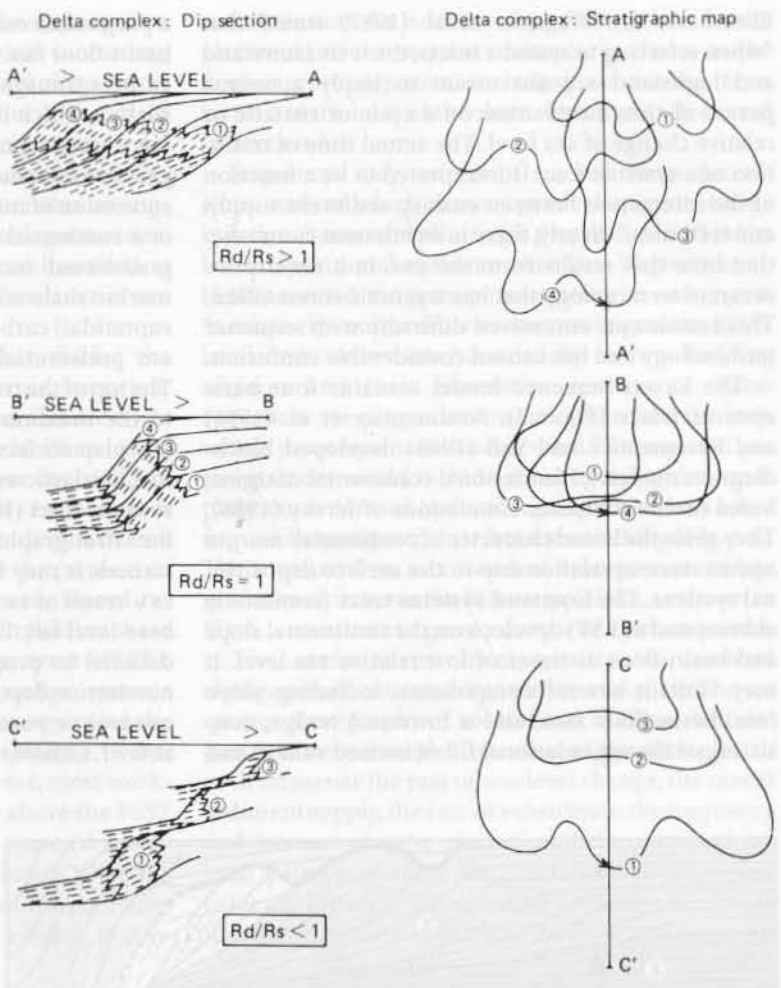


Fig. 6.4 A–C. Relationship between the rate of deposition (*Rd*) and the rate of subsidence (*Rs*) in a delta complex. **A** Progradational. **B** Aggradational. **C** Retrogradational. Maps at *right* show successive positions of delta front. (Curtis 1970)

to assume that the change is eustatic (Chap. 8). Regionally, isopach patterns may be indicative. Uniform unit thicknesses may indicate eustatic controls, whereas regional thickness variations may indicate tectonic controls and can probably be correlated with movements on known structures.

Figure 6.4 illustrates the importance of the concept of accommodation in sequence generation. Figure 6.4A illustrates a basin in which the rate of accommodation generation is low or is outpaced by sediment supply. Outbuilding, or **progradation**, is the normal depositional response. In Fig. 6.4B, the rate of accommodation generation is in balance with sediment supply, resulting in **aggradation**. In Fig. 6.4C, the rate of accommodation generation is rapid, outpacing sediment supply so that **retrogradation**, or **backstepping**, takes place. The condition in Fig. 6.4A is common when relative sea level stabilizes at a low or high point in the sea-level cycle. Conditions in Figs. 6.4B and C are common during rising relative sea level and are typically (but not necessarily) accompanied by transgression.

6.3 Depositional Systems and Systems Tracts

As noted above, the concept of depositional episodes and the depositional-systems basin-analysis methodology were developed largely in the Gulf Coast region. A **depositional system** is defined as "a three-dimensional assemblage of lithofacies, genetically linked by active (modern) or inferred (ancient) processes and environments" (Posamentier et al. 1988). A **systems tract** is defined as

> a linkage of contemporaneous depositional systems. Each is defined objectively by stratal geometries at bounding surfaces, position within the equence, and internal parasequence stacking patterns. Each is interpreted to be associated with a specific segment of the eustatic curve (i.e., eustatic lowstand-lowstand wedge; eustatic rise-transgressive; rapid eustatic fall-lowstand fan, and so on), although not defined on the basis of this association (Posamentier et al. 1988).

Elsewhere, Van Wagoner et al. (1987) stated that "when referring to systems tracts, the terms lowstand and highstand are not meant to imply a unique period of time or position on a cycle of eustatic or relative change of sea level. The actual time of initiation of a systems tract is interpreted to be a function of the interaction between eustasy, sediment supply, and tectonics." Clearly, there is an inherent contradiction here that results from the use, in a descriptive sense, of terminology that has a genetic connotation. This remains an unresolved difficulty with sequence terminology and has caused considerable confusion.

The Exxon sequence model contains four basic systems tracts (Fig. 6.5). Posamentier et al. (1988) and Posamentier and Vail (1988) developed block-diagram models of extensional continental margins, based on the computer simulations of Jervey (1988). They show the broad character of continental-margin architecture in relationship to the surface depositional systems. The **lowstand systems tract** (commonly abbreviated as **LST**) develops on the continental slope and basin floor at times of low relative sea level. It may contain several components, including slope fans, basin-floor fans, and a lowstand wedge, consisting of the aggradational fill of incised valleys and a progradational wedge which may downlap onto the basin-floor fan. The top of the lowstand tract is marked by a **transgressive surface**, possibly a ravinement surface, which is a surface of wave scour formed during transgression (see below). Above this is the **transgressive systems tract** (**TST**). This may be a very thin succession of marine shales, a transgressive gravel lag, or a condensed succession, or it may consist of retrogradational successions of shelf deposits, including marine shale and sandstone, or platform (subtidal to supratidal) carbonates. Coals in coastal-plain settings are preferentially developed during transgression. The top of the transgressive systems tract corresponds to the **maximum flooding surface** (**MFS**) and is a downlap surface, commonly a very widespread shale in siliciclastic systems, above which is the **highstand systems tract** (**HST**). This third tract forms the top of the stratigraphic sequence although, in some instances, it may be considerably reduced in thickness as a result of erosion accompanying the next cycle of base-level fall. The highstand tract is, typically, aggradational to progradational and consists of shelf to nonmarine deposits arranged in successive facies successions or parasequences (the latter term is defined below). Clinoform architectures are characteristic.

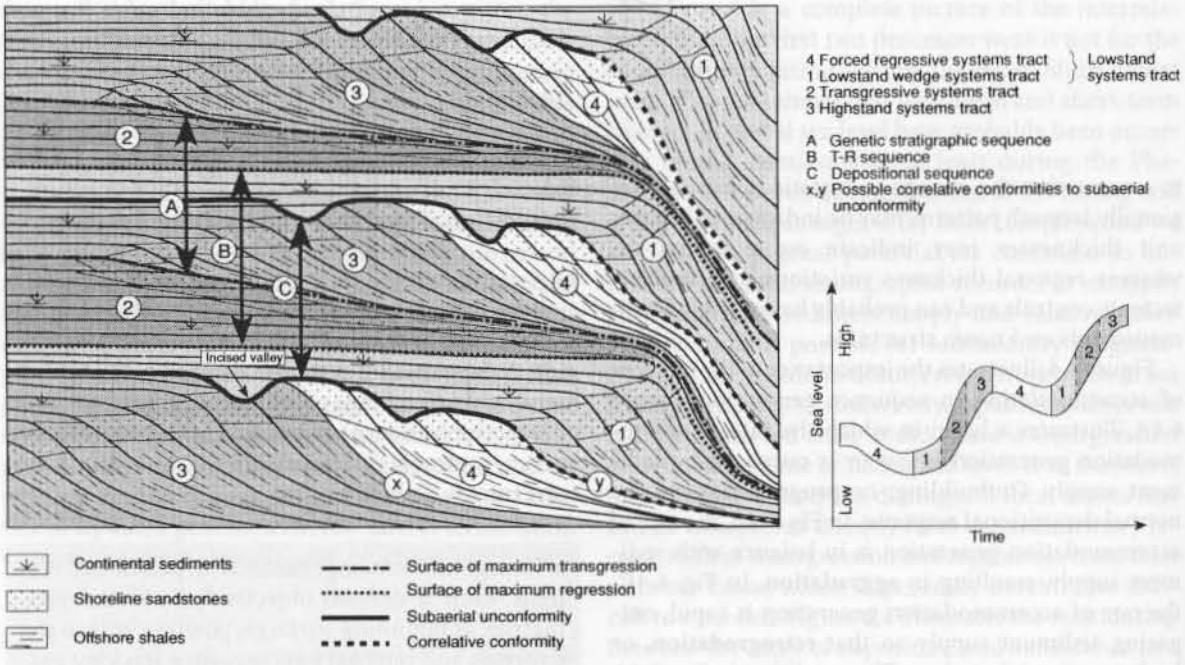


Fig. 6.5. Cartoon of sequence architecture on an extensional continental margin. The surface of maximum transgression is also termed the maximum flooding surface (*MFS*). The surface of maximum regression corresponds to the first marine flooding or transgressive surface (*TS*). Systems tracts are indicated by numerals. Systems tracts *1–3* correspond to the original systems tracts of Posamentier and Vail (1988). The forced-regressive wedge systems tract (*4*) has also been termed the falling-stage systems tract and was defined by later workers (see text). The *arrows* with letters *A–C* indicate three methods of defining sequences based on the choice of bounding surfaces (see text). At *right* is shown the approximate relationship of these systems tracts to segments of the graph of change in relative sea level. (Helland-Hansen and Gjelberg 1994)

Shelf-margin systems tracts may be deposited at times of slow fall in relative sea level when (according to the Exxon models) sea level does not drop below the edge of the continental shelf. This is the main characteristic of a so-called type-2 sequence, as defined in the next section. The shelf-margin tract consists of shelf and slope clastics or carbonates arranged in aggradational or progradational geometries, resting on the sequence boundary and bounded at the top by a transgressive surface. This tract is not shown in Fig. 6.5.

Recent work has demonstrated that, during the falling stage of the relative sea-level curve, important erosional and depositional events occur that were not included in the original Exxon models. Plint (1988, 1991) mapped **sharp-based sandstones** deposited in shoreface environments above erosion surfaces that were formed during regression. He demonstrated that these regressive sandstones were the product of falling sea level, and termed the process **forced regression**. Hunt and Tucker (1992) incorporated these beds into a new **falling-stage systems tract** (**FSST**), (also termed the **forced-regressive wedge systems tract** or **FRWST**). A problem then arose regarding the placement of the sequence boundary relative to this systems tract. As discussed in Section 6.4, most workers now draw the sequence boundary above the FSST (the subaerial unconformity and its seaward extension in Fig. 6.5), because this surface forms when the base-level reaches its lowest point, and it coincides with (is coeval with) the nonmarine surface of subaerial erosion. An additional problem relates to the placement of the sequence boundary relative to the deep-marine deposits, including submarine fans, located at the base of the slope. This is also discussed in Section 6.4.

Sequences may consist of stacked facies successions, each of which shows a gradual upward change in facies character, indicating a progressive shift in the local depositional environments. Van Wagoner et al. (1987) erected the term **parasequence** to encompass "a relatively conformable succession of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative surfaces ... Parasequences are progradational and therefore the beds within parasequences shoal upward." Examples of parasequences would include shoaling-upward strandplain sands formed by local regression or delta lobes formed by autogenic distributary switching (Fig. 6.6). The topmost 5-m succession of transgressive to aggradational facies in the Mississippi delta, as shown in Fig. 6.1, constitutes a good example of a parasequence.

The environments most sensitive to sea-level change are, of course, the coastline and shelf. The sequence stratigraphy of coastal and shallow-shelf deposits is very complex because of the interplay of such factors as the rate of sea-level change, the rate of sediment supply, the rate of subsidence, the frequency and amount of wave erosion, and the nature of the local paleogeography. Six generalized stratigraphic columns through transgressive systems tracts were illustrated by Nummedal and Swift (1987) as exam-

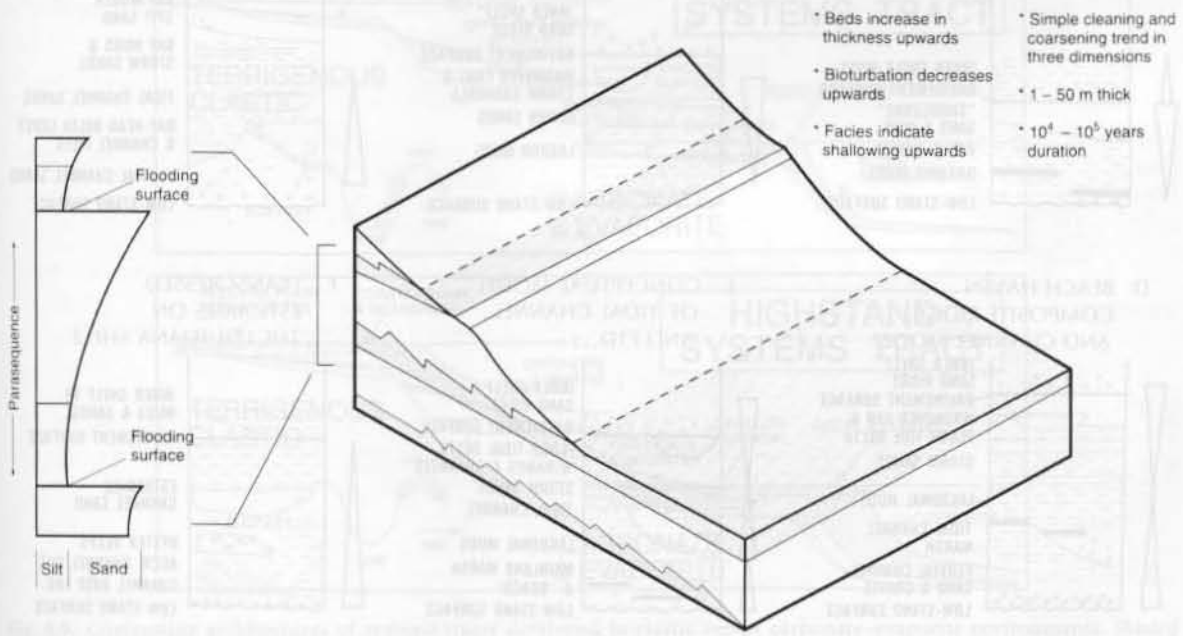


Fig. 6.6. Idealized parasequence representing a shallow-marine environment in a wave- or storm-dominated setting, where upward coarsening indicates shallowing with time. (Emery and Myers 1996)

ples of this complexity (Fig. 6.7). Coarsening-upward beach profiles and fining-upward estuary-fill or tidal-channel profiles are among the types of local succession that may be present.

Note the presence of a **ravinement surface** overlain by shelf sands or muds in most of these diagrams. This surface is the surface of equilibrium on the shelf formed in response to the local wave and current regimes. "As sea-level rises the shoreface profile translates upward and landward through a process of erosional shoreface retreat. The ravinement surface cut by this process will physically rise toward the basin margin. It will also become younger in the same direction" (Nummedal and Swift 1987; Fig. 6.8). The ravinement surface is only one of several types of diastem that may be encountered in transgressive systems tracts. Others include channel scours and marine erosion surfaces. These can make the definition of sequence boundaries difficult. The ravinement surface, in particular, is commonly a major erosion surface that separates markedly different facies, but it forms within the transgressive systems tract and is, therefore, not a sequence boundary, unless the depth of erosion at the ravinement surface is such that no transgressive deposits are preserved. In the latter case, the ravinement surface is coincident with the erosion surface at the top of the underlying regressive parasequence.

Posamentier and Allen (1993) demonstrated that, in some instances, near the base of the continental slope and at the toe of the "last clinoform" of the older regressive deposits, this transgressive surface may be covered by a wedge of fine-grained sediment shed seaward by ravinement erosion. They referred to this as the **healing phase** of the transgression, "because the area seaward of the relatively steeply dipping shoreface or delta front is partly infilled or 'healed over' as a lower-gradient shelf equilibrium profile is re-established."

Carbonate and clastic depositional systems respond very differently to sea-level change (Sarg 1988; James and Kendall 1992; Schlager 1992). Figure 6.9 illustrates these differences, which are described briefly here. At times of sea-level lowstand, terrigenous clastics are carried across the continental shelf – a process known as **sediment bypass** – leading to exposure, erosion, and the development of incised valleys. Submarine canyons are deepened and sand-rich turbidite systems build submarine fan complexes on the slope and the basin floor. Carbonate systems essentially shut down at times of lowstand, because the main "carbonate factory" (the continental shelf) is exposed and commonly undergoes karstification. A narrow shelf-edge belt of reefs or sand shoals may occur, while the deep-water basin is starved of sediment or possibly subjected to hyperconcen-

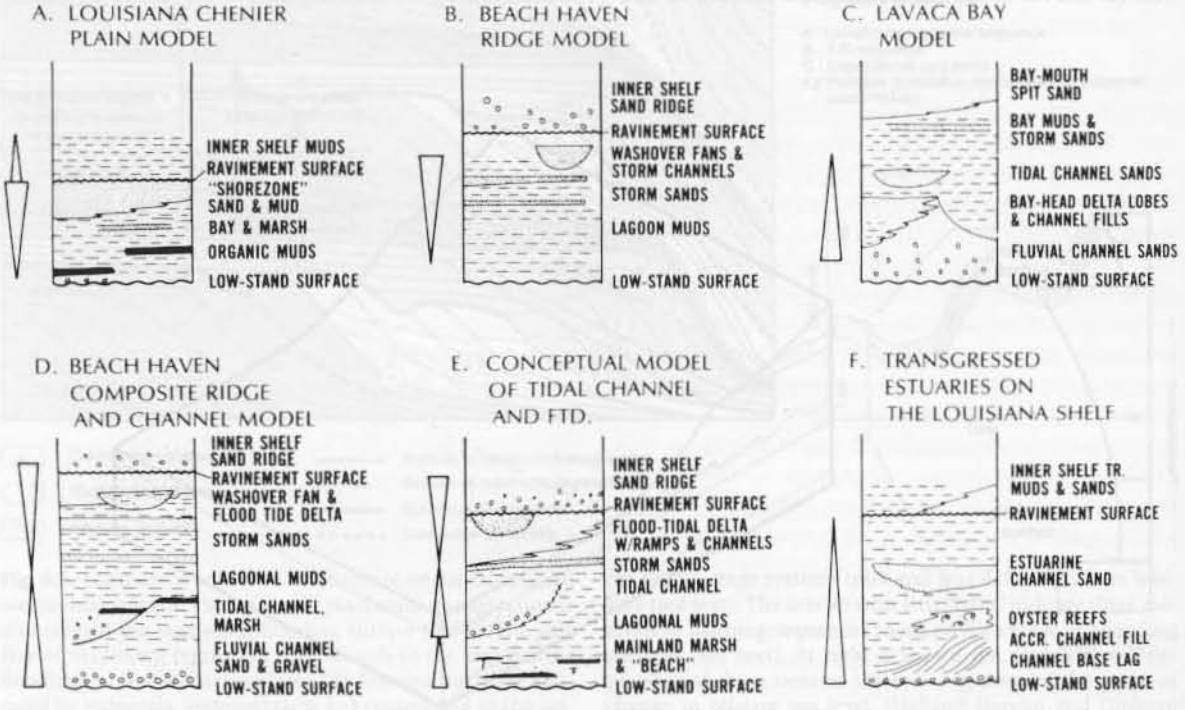


Fig. 6.7. Stratigraphic columns from selected Holocene marginal-marine settings to illustrate transgressive systems tracts. (Nummedal and Swift 1987)

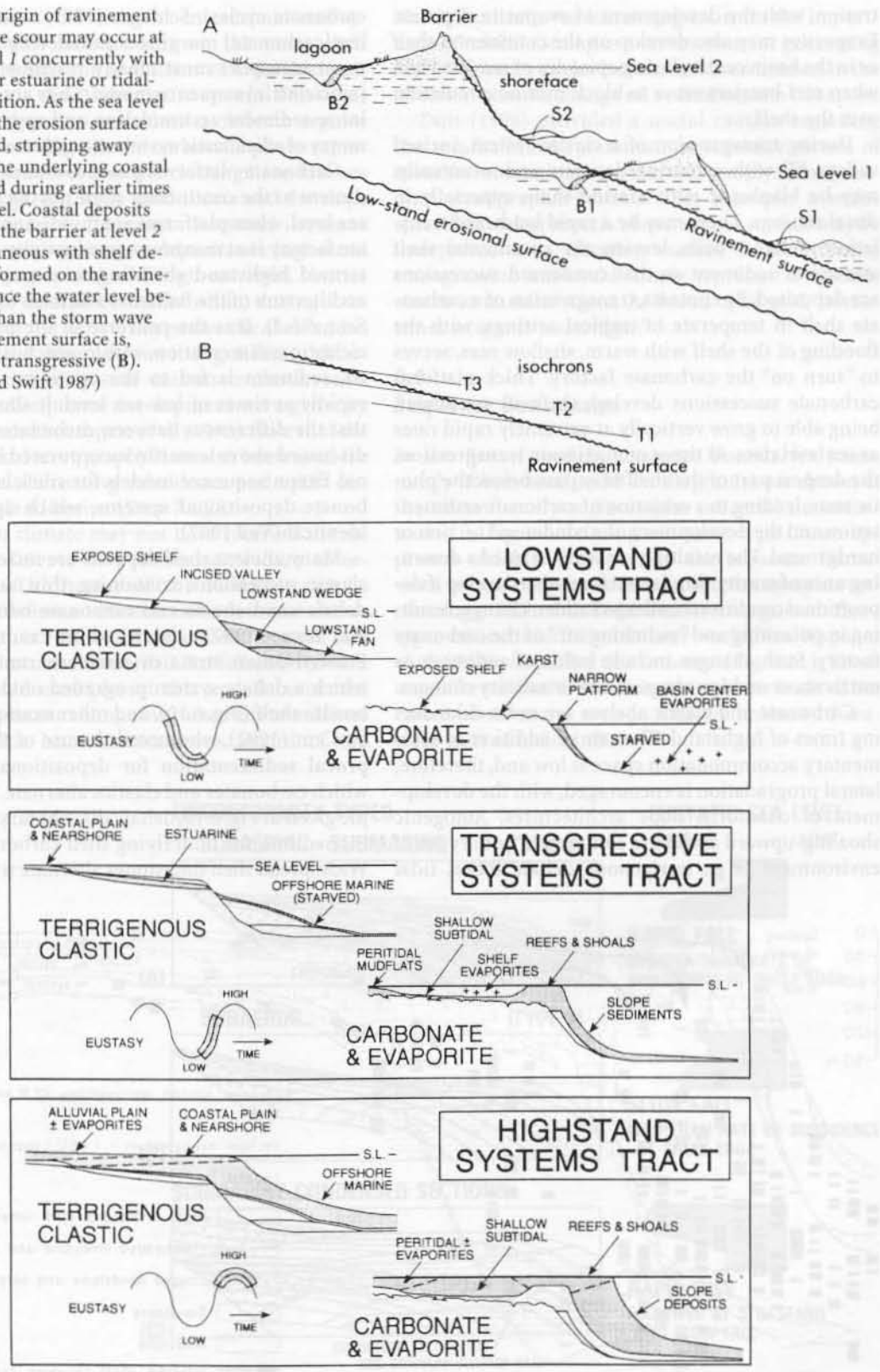


Fig. 6.8 A, B. Origin of ravinement surfaces. A Wave scour may occur at time of sea level *1* concurrently with back-barrier (or estuarine or tidal-channel) deposition. As the sea level rises to level *2*, the erosion surface moves landward, stripping away some or all of the underlying coastal deposits formed during earlier times of lower sea level. Coastal deposits formed behind the barrier at level *2* are contemporaneous with shelf deposits that are formed on the ravinement surface once the water level becomes deeper than the storm wave base. The ravinement surface is, therefore, time-transgressive (B). (Nummedal and Swift 1987)

Fig. 6.9. Contrasting architectures of systems tracts developed in clastic versus carbonate–evaporite environments. *Shaded areas* indicate the sediments deposited during the time interval indicated on the graph of sea-level change. (James and Kendall 1992)

tration, with the development of evaporite deposits. Evaporites may also develop on the continental shelf or in the basin centre during episodes of sea-level fall, when reef barriers serve to block marine circulation over the shelf.

During transgression of a clastic system, incised valleys fill with estuarine deposits and, eventually, may be blanketed with marine shale, especially in distal settings. There may be a rapid landward translation of facies belts, leaving the continental shelf starved of sediment so that condensed successions are deposited. By contrast, transgression of a carbonate shelf in temperate to tropical settings, with the flooding of the shelf with warm, shallow seas, serves to "turn on" the carbonate factory. Thick platform carbonate successions develop, reefs, in particular, being able to grow vertically at extremely rapid rates as sea level rises. At times of maximum transgression, the deepest part of the shelf may pass below the photic zone, leading to a cessation of carbonate sedimentation and the development of a condensed section or hardground. The resulting surface is termed a **drowning unconformity**. Similar surfaces also develop if depositional conditions undergo sudden change, resulting in poisoning and "switching off" of the carbonate factory. Such changes include influx of sediment or nutrients or sudden temperature or salinity changes.

Carbonate and clastic shelves are most alike during times of highstand. The rate of addition of sedimentary accommodation space is low and, therefore, lateral progradation is encouraged, with the development of clinoform slope architectures. Autogenic shoaling-upward cycles are common in both types of environment (e.g., terrigenous deltaic lobes, tidal carbonate cycles). Schlager (1992) stated: "Prograding [carbonate] margins dominated by offshore sediment transport most closely resemble the classical [siliciclastic] sequence model. They are controlled by loose sediment accumulation and approach the geometry of siliciclastic systems (e.g., leeward margins)."

Carbonate platforms produce and export most sediment to the continental slope during highstands of sea level, when platforms are flooded and the carbonate factory is at maximum productivity. This process, termed **highstand shedding**, is exemplified by the architecture of the Bahamas Platform (Schlager 1992; Sect. 6.6.3). It is the converse of the pattern of siliciclastic sedimentation, within which, as already noted, sediment is fed to the continental slope most rapidly at times of low sea level. It should be noted that the differences between carbonates and clastics discussed above were not incorporated into the original Exxon sequence models for siliciclastic and carbonate-depositional systems, which appear almost identical (Vail 1987).

Many ancient shelf deposits are mixed carbonate-clastic successions, containing thin sand banks or deltaic sand sheets and carbonate banks. Galloway and Brown (1973) described an example from the Pennsylvanian strata of northern central Texas, in which a deltaic system prograded onto a stable carbonate shelf (Fig. 6.10), and other examples are given by Cant (1992), who noted the use of the term **reciprocal sedimentation** for depositional systems in which carbonates and clastics alternate. In the example given in Fig. 6.10, deltaic distributary channels are incised into the underlying shelf carbonate deposits. Widespread shelf limestones alternate with the clastic

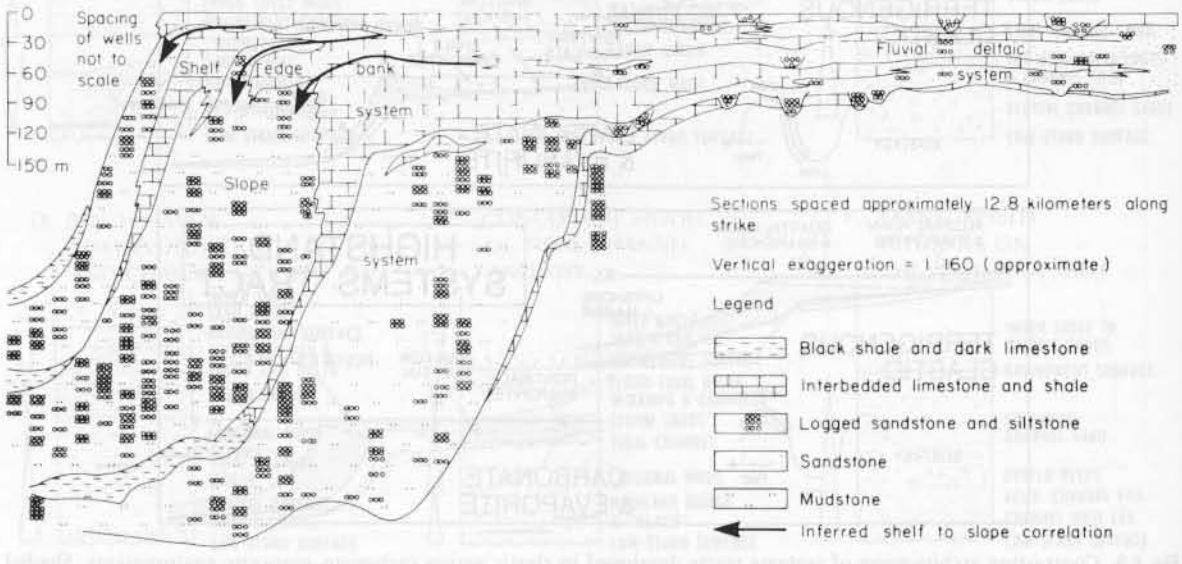


Fig. 6.10. A mixed carbonate–clastic shelf-slope depositional system, formed as a result of alternating periods of high and low sea level. Pennsylvanian, north-central Texas. (Galloway and Brown 1973)

sheets and also occur in some interdeltaic embayments. Carbonate banks occur on the outer shelf edge, beyond which the sediments thicken dramatically into a slope clinoform clastic system. This association of carbonates and clastics reflects regular changes in sea level, with the carbonate phase representing high sea level and the clastic phase low sea level. The deltaic and shelf-sand sheets and the slope clinoform deposits represent lowstand systems tracts, whereas the carbonate deposits are highstand deposits. During episodes of high sea level, clastics are trapped in nearshore deltas while, during lowstands, much of the detritus bypasses the shelf and is deposited on the slope (arrows in Fig. 6.10). Additional examples of reciprocal carbonate-clastic sedimentation and sequence development were given by Dolan (1989) and Southgate et al. (1993). Note that the term "reciprocal stratigraphy" has also been used for the pattern of alternating subsidence and uplift observed in foreland basins (Sect. 7.4.2).

Changes in climate may not have received the attention they deserve as possible controllers of sequence style. The nature of the sediment supply, especially the balance between carbonate and clastic sedimentation, is very much dependent on such factors as temperature and rainfall. Ruffell and Rawson (1994) provided examples of this and documented the effects of climate change over intervals of millions of years in the Jurassic-Cretaceous record of the North Sea basin. An increasing focus on the importance of orbital forcing as a mechanism for generating high-frequency cycles has also aroused interest in the effects of climate change on sedimentation (Chap. 8).

Dott (1996) provided a useful caution regarding the too-facile interpretation, in sequence terms, of some facies relationships that may simply record random stratigraphic "events". For example, a widespread intraclast breccia above a scour surface may be interpreted as a flooding surface formed during transgression but could simply represent a random storm or tsunami event. As Freud once said, "sometimes a cigar is just a cigar."

6.4 Sequence Boundaries

In the Exxon models, sequence boundaries (commonly abbreviated as SB on diagrams) are drawn at the unconformity surfaces (the subaerial unconformities in Fig. 6.5 defining sequence type C), an approach which readily permits the sequence framework to be incorporated into an allostratigraphic terminology. Vail and Todd (1981) recognized three types of unconformity, but later work (Van Wagoner et al. 1987) simplified this into two (Fig. 6.11). These are differentiated from each other on the basis of the extent of subaerial erosion and the amount of seaward shift of facies belts.

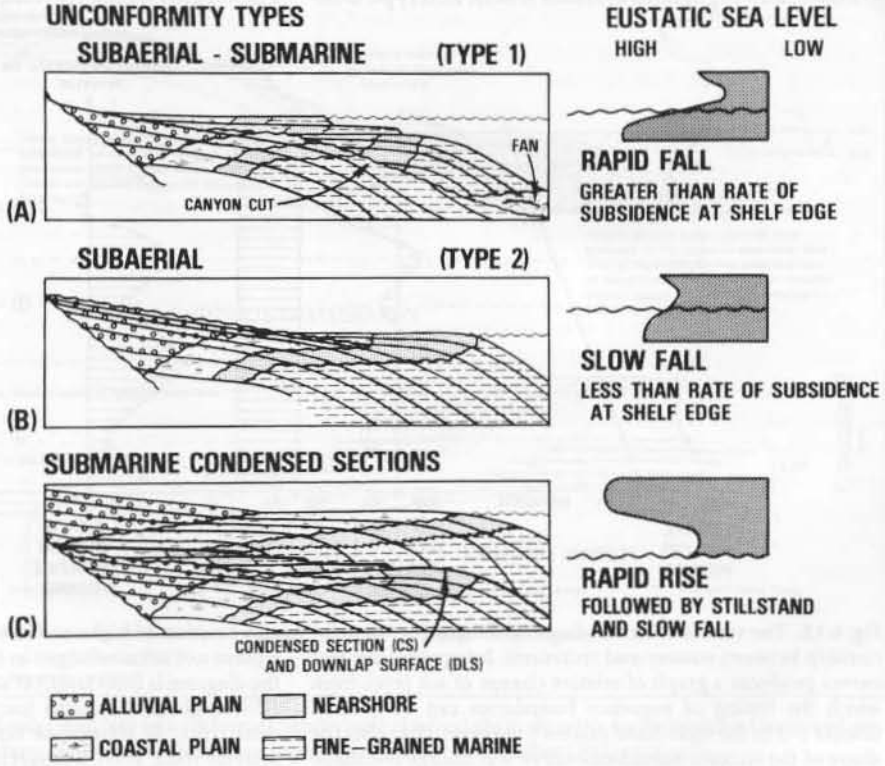


Fig. 6.11. Definition of type-1 and type-2 unconformities. (Vail et al. 1984)

A **type-1 unconformity** develops where eustatic sea-level fall is more rapid than tectonic subsidence. The coastline may move out to near the shelf edge, and extensive subaerial erosion takes place, with the development of incised fluvial valleys on the shelf and the deepening of submarine canyons on the continental slope. Clastic detritus is transported down these fluvial and canyon systems to the base of the continental slope, forming extensive lowstand systems tracts. At type-1 unconformities, facies belts undergo a substantial basinward shift. Highstand deposits below the unconformity may be deeply eroded. In carbonate systems, exposure of the platform may lead to the development of widespread karst systems and internal dissolution, with considerable erosion of the platform margin and downslope movement of carbonate breccias and turbidites (Sarg 1988). These are essentially the conditions shown in the "lowstand" panel of Fig. 6.9.

A **type-2 unconformity** develops when relative sea level falls slowly, resulting in a gradual seaward shift of facies tracts but only minor subaerial exposure and erosion. According to Vail (1987) and Vail et al. (1991), a shelf-margin systems tract is formed under these conditions. Type-2 unconformities are much more difficult to identify in seismic and outcrop records, because they are not characterized by deep erosion or major facies shifts.

Vail et al. (1991) subdivided sequences into type-1 sequences, which are bounded at the base by a type-1 unconformity and are composed of lowstand, transgressive, and highstand systems tracts, and type-2 sequences, which are bounded at the base by a type-2 unconformity and consist of shelf-margin, transgressive, and highstand systems tracts. Whereas the type-1/2 terminology for unconformities is now widely used, extending the terminology to sequences has not received widespread acceptance.

Seaward, where they pass into the deep-marine realm, it is hypothesized that sequence boundaries become conformable. The equivalent surface is termed the **correlative conformity** and, theoretically, there is no reason why sequences should not be mapped and correlated on the basis of such surfaces. However, recognizing these surfaces within conformable successions, in the absence of distinctive facies characteristics, is not likely to be a very practical exercise and may depend entirely on extrapolation from the shallow-marine realm based on the available chronostratigraphic information or the tracing of reflectors on seismic cross sections. Neither method can provide a sequence framework as tightly constrained as one based on the physical tracing or correlation of unconformity surfaces or distinctive facies changes in core or on petrophysical logs. Allostratigraphic terminology is, strictly speaking, not applicable to such conformable sequences.

It is important to note that sequence boundaries develop throughout the falling stage and lowstand of the relative sea-level curve, and their ultimate expression may not reflect the times when eustatic sea level is actually at its lowest. The simplest case is shown in Fig. 6.12, in which tectonic subsidence and eustasy are

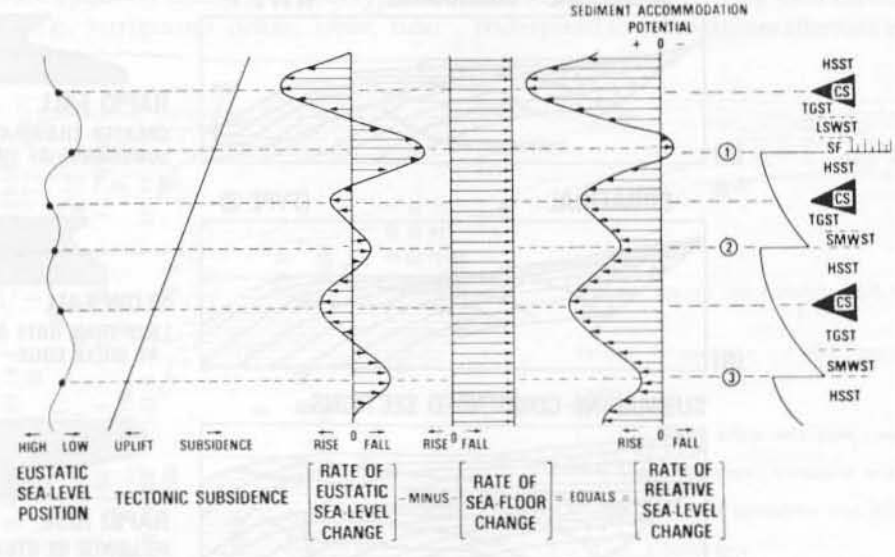


Fig. 6.12. The standard Exxon diagram illustrating the relationship between eustasy and tectonism. Integrating the two curves produces a graph of relative change of sea level, from which the timing of sequence boundaries can be derived (events 1–3 in the *right-hand column*). However, changing the shape of the tectonic subsidence curve will change the shape and position of highs and lows in the relative sea-level curve, a point not acknowledged in the Exxon work. This version of the diagram is from Loutit et al. (1988). *CS* Condensed section; *HSST* highstand systems tract; *LSWST* lowstand wedge systems tract; *SF* submarine fan; *SMWST* shelf-margin wedge systems tract; *TGST* transgressive systems tract

integrated to show an overall graph of the rate of change of relative sea level. In this ideal or model case, sequence boundaries (numerals 1–3 at the right of Fig 6.12) represent the time of maximum rate of fall of sea level. As is indicated in most of the Exxon diagrams, where the eustatic curve approximates a sine curve, the sequence boundary corresponds to the inflection point in the falling curve. After this time, according to the Exxon model, tectonic subsidence outpaces eustasy and relative sea level rises. This point assumes considerable importance when attempts are made to use sequence boundaries to date eustatic sea-level change and derive global-cycle charts. The offset in time between the eustasy inflection point and the time of lowest relative sea level will vary depending on the rate of sea-level change and the rates of tectonic subsidence and sediment supply. These rates each vary widely depending on the nature of tectonic and eustatic controls, as discussed in Chapter 8.

Care must be taken to map and interpret unconformities correctly. Not all breaks in sedimentation are caused by changes in sea level, and some erosional breaks are time transgressive. For example, ravinement surfaces are commonly more prominent in a stratigraphic section than subaerial unconformities, yet they form during transgression and are time transgressive (Fig. 6.8).

Hunt and Tucker (1992) pointed out several problems with the placement of the sequence boundary relative to the falling stage and lowstand deposits in the original sequence models of Posamentier et al. (1988) and Posamentier and Vail (1988). They illustrated these problems with two diagrams, which are reproduced here as Figs. 6.13 and 6.14. Two particular problems need to be addressed.

1. Falling-stage deposits were not recognized in these early models, for reasons that are not clear but may relate to the fact that they tend to be thin and, therefore, may not be recognizable on seismic-reflection lines. Recognition of the importance of this systems tract raised the question of where to draw the sequence boundary. Posamentier et al. (1988) and Posamentier and Vail (1988) had related changing accommodation rates and facies shifts to the inflection points in the sea-level curve. Following this approach, Posamentier et al. (1992) suggested that the sequence boundary represents the beginning of the sea-level fall and proposed that the boundary be drawn below the deposits formed during the falling stage. The problems with this approach are that this contact is truncated by the regressive surface of marine erosion formed during the falling stage and may, in

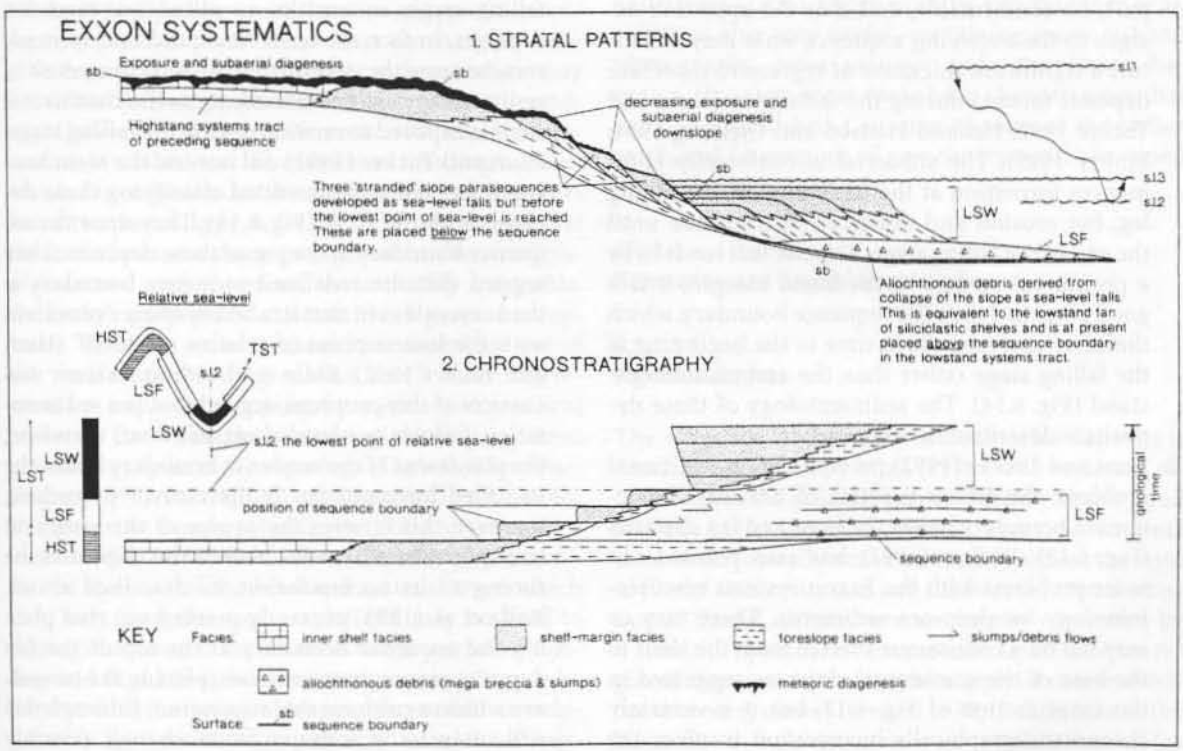


Fig. 6.13. Cross section and Wheeler diagram of a carbonate sand-shoal-rimmed shelf, showing facies, systems tracts, and positioning of the sequence boundary, according to the Exxon terminology. (Hunt and Tucker 1992)

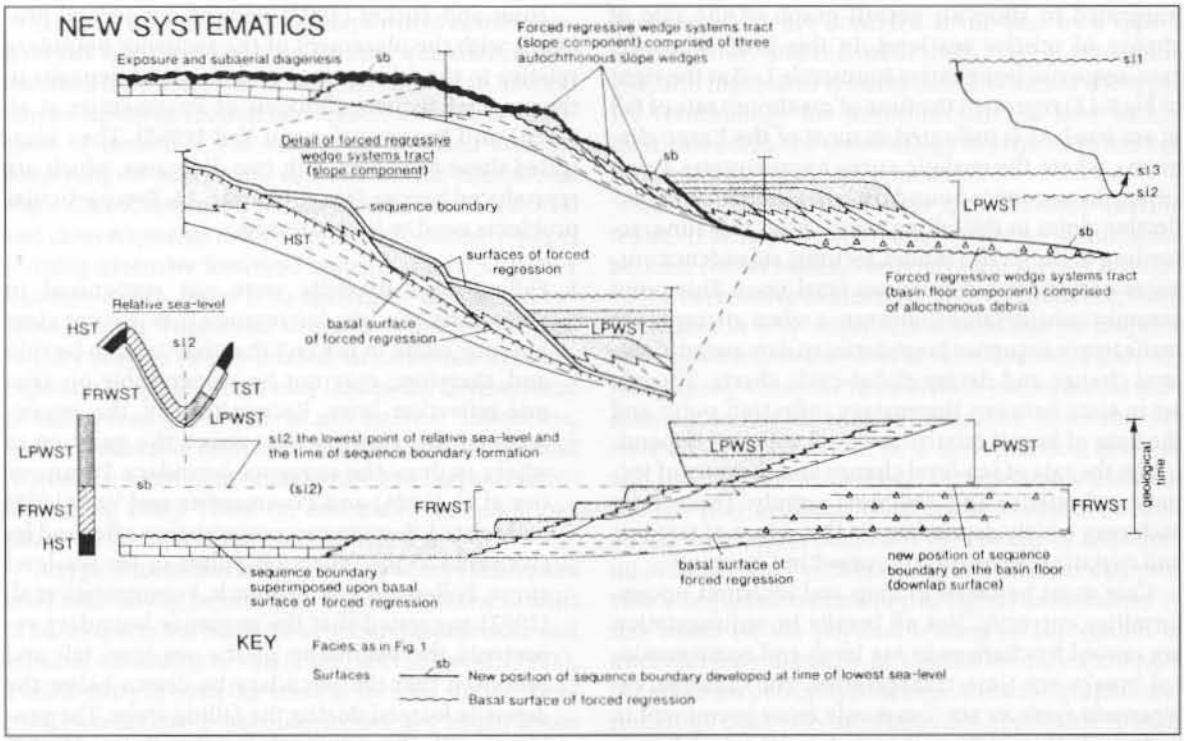


Fig. 6.14. A revised version of Fig. 6.13, showing a reinterpretation of the position of the sequence boundary relative to the lowstand fan deposits – what Hunt and Tucker (1992) termed the lowstand prograding-wedge systems tract

part, be conformable, and that the approach assigns to the overlying sequence what may constitute a significant thickness of regressive shoreface deposits formed during the falling leg (Hunt and Tucker 1992; Helland-Hansen and Gjelberg 1994; Embry 1995). The subaerial unconformity commences formation at the beginning of the falling leg, but erosion and downcutting continue until the end of the falling stage. Because this tends to be a prominent surface at the basin margin, it is a good place to define the sequence boundary, which therefore corresponds in time to the beginning of the falling stage rather than the end of the highstand (Fig. 6.14). The sedimentology of these deposits is described in Sect. 6.5.2.2.

2. Hunt and Tucker (1992) pointed out an additional problem: the Exxon practice of drawing the sequence boundary below the lowstand fan deposits (Fig. 6.13). Walker (1992) has also pointed out some problems with the Exxon systems tract terminology for deep-sea sediments. There may or may not be a continuous surface from the shelf to the base of the continental slope, as suggested in the cross section of Fig. 6.13, but it is certainly chronostratigraphically inconsistent to place the sequence boundary at the base of these deposits (Fig. 6.13). Just as the FSST is formed during the falling stage, so are the so-called lowstand fan deposits. In fact, the term "lowstand fan" is incorrect, because the deposits are formed as a result of sediment bypass and reworking as the continental shelf is exposed to erosion during the falling stage. Hunt and Tucker (1992) did not use the term lowstand fan and recommended classifying these deposits with the FSST (Fig. 6.14). They drew the sequence boundary at the top of these deposits. They argued that the redefined sequence boundary is then consistent in that it is "everywhere coincident with the lowest point of relative sea level" (Hunt and Tucker 1992). Kolla et al. (1995), in their discussion of this proposal, argued that fan sedimentation follows sea-level drop and that, therefore, the placement of the sequence boundary below the so-called lowstand fan is the correct procedure; however, this ignores the source of the sediment transported down to the base of the slope and the timing of its accumulation, as described above. Kolla et al. (1995) correctly pointed out that placing the sequence boundary at the top of the fan deposits means, in many cases, placing the boundary within a conformable succession, although this will often be at a major facies change, possibly within a condensed succession. This they found problematic, preferring to follow the historical

precedent of drawing sequence boundaries at unconformities, even though this may disregard the elementary sedimentology of the falling-stage systems tract and even though the term "correlative conformity" was introduced by Mitchum et al. (1977) to include the type of situation where an unconformable sequence boundary passes laterally into a conformable succession. Other former members of the Exxon team (Mitchum et al. 1993) have correctly noted that the basin-floor fan is formed during the "early phase of a relative fall of sea level", yet persisted in placing the sequence boundary below it.

An alternative sequence model, termed the **genetic stratigraphic sequence**, was defined by Galloway (1989), building on the work of Frazier (1974; Figs. 6.1, 6.2). Although Galloway stressed supposed philosophical differences between his model and the Exxon model, in practice, the difference between them is simply one of where to define the sequence boundaries. The Exxon model places emphasis on unconformities, but Galloway (1989) pointed out that, under some circumstances, unconformities may be poorly defined or absent and, in any case, are not always easy to recognize and map because they may correspond to channel-scour surfaces within a complex of mutually erosive sand bodies constituting the falling-stage and lowstand systems tracts. This is particularly likely to be so in the case of type-2 unconformities and, in particular, within the marginal-marine depositional systems, where sedimentation may be virtually continuous across the sequence boundary. Galloway's (1989) preference is to draw the sequence boundaries at the maximum flooding surface, which corresponds to the highstand downlap surfaces (sequence type A in Fig. 6.5). He claims that these surfaces are more prominent in the stratigraphic record and, therefore, more readily mappable. Galloway's proposal has not met with general acceptance. However, some workers have found his use of the maximum flooding surface much more convenient for sequence mapping, for a variety of practical reasons. For example, it may yield a prominent gamma-ray spike in wireline logs (Underhill and Partington 1993), correspond to widespread and distinctive goniatite bands (Martinsen 1993), or provide a more readily traceable marker, in contrast to the surface at the base of the lowstand systems tract, which may have irregular topography and may be hard to distinguish from other channel-scour surfaces (Gibling and Bird 1994). In nonmarine sections it may be hard to find the paleosol on interfluves that correlates with the sequence-bounding channel-scour surface (Martinsen 1993). In some studies (e.g., Plint et al. 1986; Bhattacharya 1993), it has been found that ravinement erosion during transgression has removed the transgressive systems tract, so that the marine flooding surface coincides with the sequence boundary.

Embry (1993) defined sequences in shelf and slope successions on the basis of their contained transgressive surfaces and referred to them as **T-R sequences** or **transgressive-regressive sequences** (sequence type B in Fig. 6.5). In most cases, these transgressive surfaces are readily mappable, because they represent a distinct change in facies. Commonly, they merge landward with and are contemporaneous with the subaerial erosion surface (although this is not shown in Fig. 6.5). Embry's sequences contain only transgressive and regressive systems tracts, which avoids the problems of boundary definition referred to later in this chapter. However, facies and architectural criteria commonly are available to permit a breakdown into the full suite of four systems tracts, and it is a matter for debate and practical convenience which method the user prefers to follow.

6.5 Systems Tracts and Sequences in Siliciclastic Deposits

A detailed treatment of systems tracts and sequences in siliciclastic deposits has been provided by Emery and Myers (1996). Older work by Van Wagoner et al. (1990) is also very useful, containing many outcrop photographs, cross sections and interpretive diagrams. However, some ideas have changed since that work was published. Critiques of some of the earlier work, and discussions of new concepts, are contained in Miall (1997; Chap. 15).

6.5.1 Nonmarine and Coastal Sequences

6.5.1.1 *Fluvial Sequences*

The sequence stratigraphy of nonmarine deposits is described in detail by Miall (1996), who reviewed recent work in the area, particularly that by Shanley and McCabe (1994). In inland basins, especially in areas close to tectonically active sediment sources, the style of fluvial sedimentation and the broader characteristics of fluvial architecture are largely controlled by tectonism. Episodes of source-area uplift steepen regional slopes and increase the supply of coarse detritus to river systems (Heward 1978). Subsidence in the basin due to fault downthrow or tectonic loading (as in a foreland basin; Chap. 7) may divert rivers or even cause them to pond in internal drainage systems. Sea-

level change is an important sediment control in coastal-plain basins. It may be driven by eustasy or by uplift or subsidence driven by tectonism. A relative lowering of sea level may lead to enhanced erosion and to valley incision. A rise in sea level generates additional accommodation for sediment, with the results described below (Miall 1991; Schumm 1993). Climate change may also be an important sedimentary control. The sediment yield from the source area, the type and density of vegetation, and the amount of surface runoff are all dependent on climate. An estimate of the relative importance of these three controls is shown in Fig. 6.15.

Fluvial depositional systems are sensitive to changes in accommodation. The preservation potential of floodplain deposits relative to that of channel-fill sediments depends in large part on variations in this parameter. Where accommodation generation is static or slow, channel systems meander or avulse back and forth across the valley floor through their own deposits, incising laterally into the floodplain and replacing it with the channel fills, which typically consist of sand or gravel. The results may be stacked and laterally-amalgamated channel-fill deposits with few fine-grained intervals. Where accommodation generation is rapid, lateral channel migration is accompanied by stratigraphic climb and, if there is a lengthy return period between successive reoccupations of a given position on the valley floor, significant thicknesses of floodplain deposits may have had time to accumulate. Rapid accommodation generation driven by base-level rise favors avulsion and the development of anastomosed fluvial styles (Törnqvist 1993).

Near the coast, increases in accommodation may result in vertical aggradation of the river systems, marine transgression, or both. This can be observed in the valley fills beneath modern rivers that have undergone the Holocene post-glacial rise in sea level. During the glacial lowstand, rivers became incised as they graded themselves to the sea-level lowstand and, at high latitudes, the coarse sediment supply that was delivered by fluvioglacial rivers typically formed braided river systems. During the post-glacial sea-level rise, many river valleys became drowned to form estuaries. Where sediment supply was adequate, particularly near the end of the initial period of rapid rise, fluvial coastal-plain aggradation resulted in dramatic changes in fluvial style from braided to meandering or anastomosed styles, as in the modern Mississippi and Rhine valleys, respectively (Törnqvist 1993). The formation and burial (preservation) of coal seams takes place primarily during periods of rapid accommodation generation (Bohacs and Suter 1997).

Inland, beyond the influence of sea-level change, accommodation increases may be brought about by basin subsidence or source-area uplift. A coarsening of the sediment load and a steepening of the fluvial profile are common and can result in a change in fluvial style to braided systems or to alluvial fans that prograde out over the basin floor (Heward 1978). On the floodplain, away from the influence of channels, sedimentation may be slow enough to permit the development of soils but, if aggradation is rapid, these are unlikely to be very well developed.

A reduction in accommodation, as at the end of a cycle of sea-level rise, leads to a reversal of these trends. Channel incision and reworking are favored. Channel sand bodies may become laterally amalgamated. Upland interfluve areas may be characterized by long periods of nondeposition, with the development of mature soils (McCarthy and Plint 1998).

These observations have been formed into generalized sequence models by several workers, including Wright and Marriott (1993), Shanley and McCabe (1994), and Currie (1997). An example of these models is illustrated in Fig. 6.16. Incised valleys (the se-

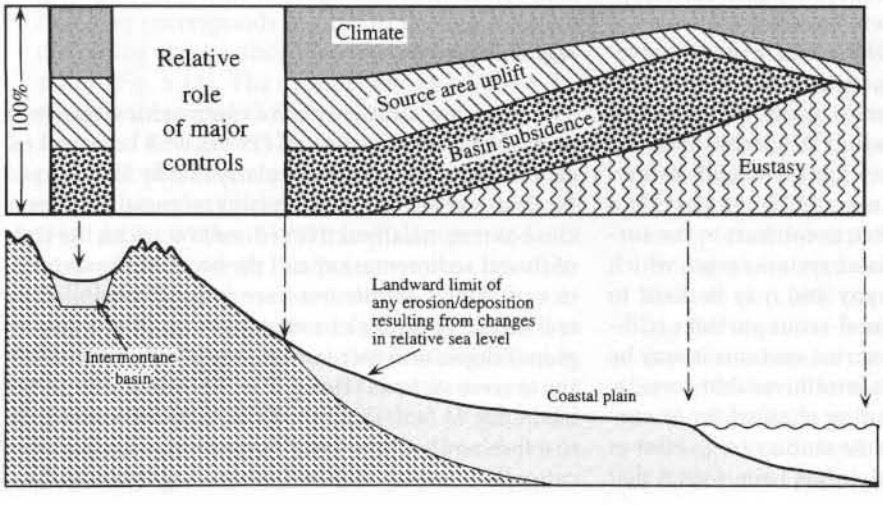


Fig. 6.15. The relative importance of the major controls on fluvial sequence architecture, showing how the balance of control varies between source area and shoreline. Intermontane basins (*left*) are, of course, unaffected by sea-level change. (Shanley and McCabe 1994)

quence stratigraphy of which is also discussed in Sect. 6.5.1.4) form during the falling stage of a sea-level cycle (Fig. 6.16A). These may be flanked by terrace remnants. The valleys begin to fill with sediment once the sea-level curve bottoms out and begins a slow rise, corresponding to the lowstand systems tract. In estuarine settings, a subsequent increase in the rate of accommodation generation is indicated by the appearance of a tidal influence on clastic sedimentation, particularly tidal bedding, rhythmic tidal couplets, inclined heterolithic stratification, and various other characteristic tidal signatures in large-scale crossbedding (Shanley et al. 1992). These tidally influenced beds are assigned to the transgressive systems tract by Shanley and McCabe (1994; Fig. 6.16C). In their model, the tidal signature (and hence the LST–TST contact) appears at about the time the incised valley is filled, but this transition could take place at any time and will tend to be later in inland parts of the valley system, reflecting the time taken for transgression to gradually climb inland. In other examples of nonmarine sequences, no tidal influence is present, because the beds were formed too far inland, beyond the reach of the sea even during the maximum flooding event. The diachronous nature of the fluvial–tidal transition emphasizes the important fact, commonly misunderstood, that systems-tract boundaries are not chronostratigraphic surfaces.

Systems-tract terminology is based on references to sea level, but it needs to be questioned whether the standard systems-tract terminology is appropriate for use in nonmarine deposits, where the effects of sea-level change may be remote or absent altogether and where the major architectural changes are controlled by tectonism or climate change. For nonmarine systems tracts, Currie (1997) proposed a different terminology that avoids reference to sea level. In his model, fluvial sequences comprise three systems tracts. The **degradational systems tract** develops during times of falling base level and negative accommodation generation, corresponding to the falling-stage and lowstand systems tracts of marine sequences. The **transitional systems tract** occurs during times of initial rise in base level and the beginning of accommodation generation, corresponding to the early transgressive systems tract. Fluvial styles may undergo major architectural changes during this interval. The **aggradational systems tract** is equivalent to the late transgressive to highstand systems tract. Currie (1997) showed how these ideas could be applied to an analysis of the Morrison (Jurassic) and Cedar Mountain (Cretaceous) formations of Wyoming and Utah (Fig. 6.17).

The sequence stratigraphy of fluvial deposits is rendered more complex by the fact that cycles of aggradation and degradation, with the formation of regional sequence boundaries and regional variations in facies architecture, are caused not just by base-level changes but also by changes in discharge and sediment supply driven by climate change. In coastal fluvial deposits, the effects of base-level change and climate change are superimposed and are complex in nature. The major effects of climate change on river systems are alteration of the discharge regime and,

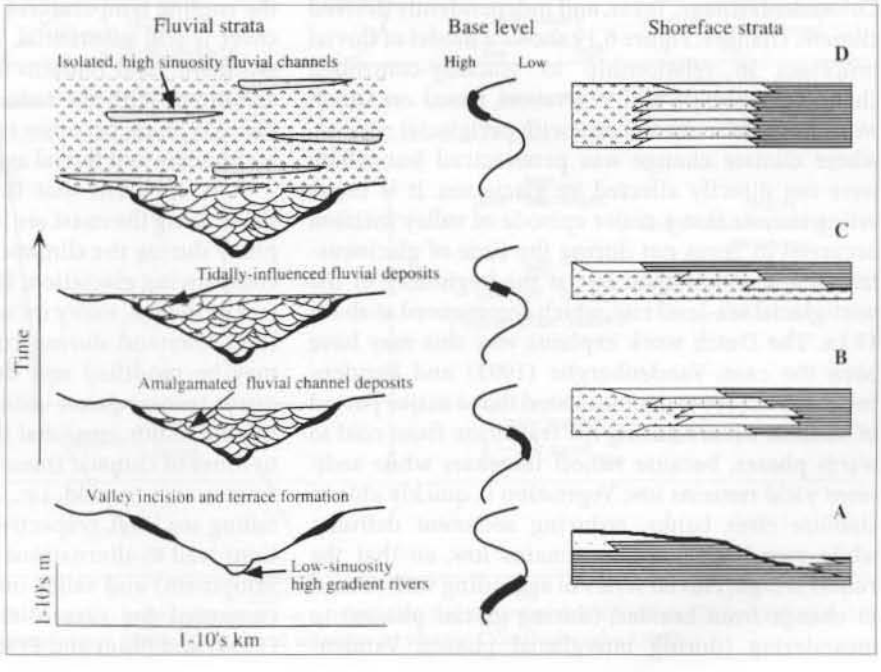


Fig. 6.16 A–D. The fluvial sequence model of Shanley and McCabe (1994), showing the relationship between shoreface and fluvial architecture and base-level change. A Falling stage systems tract, with development of incised valley and fluvial terraces. B Lowstand systems tract. C Tidal influence indicates the beginning of the transgressive systems tract. D Highstand systems tract

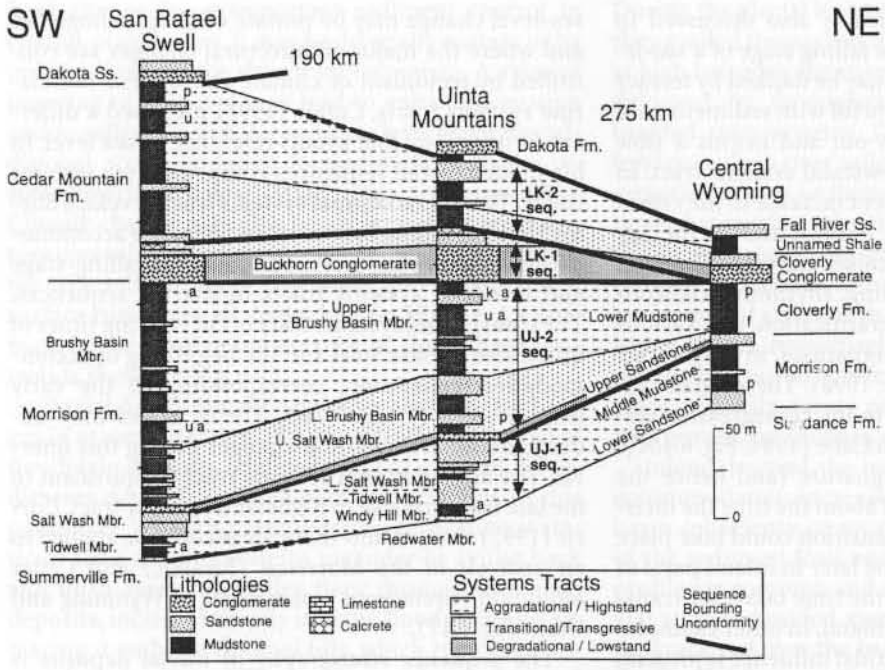


Fig. 6.17. Regional correlation and sequence diagram of the Upper Jurassic-Lower Cretaceous rocks of east-central Utah and central Wyoming. *Lowercase letters* refer to beds dated chronostratigraphically. (Currie 1997)

coupled with changes in vegetation cover, alteration of the sediment yield. Because of lags in the response to climatic forcing, and various geomorphic threshold effects, these changes do not occur in phase with the base-level changes. In areas close to continental ice caps, glacioeustatic base-level changes are also influenced by the isostatic responses to ice-sheet loading and unloading.

Figure 6.18 shows the correlation between the glacial to post-glacial alluvial stratigraphy of the upper Colorado drainage, Texas, and independently derived climatic changes. Figure 6.19 shows a model of fluvial processes in relationship to glacially-controlled changes in climate and vegetation, based on Dutch work. Both these studies deal with periglacial regions, where climate change was pronounced but which were not directly affected by glaciation. It is interesting to note that a major episode of valley incision occurred in Texas not during the time of glacioeustatic sea-level lowstand but at the beginning of the post-glacial sea-level rise, which commenced at about 15 ka. The Dutch work explains why this may have been the case. Vandenberghe (1993) and Vandenberghe et al. (1994) demonstrated that a major period of incision occurs during the transition from cold to warm phases, because runoff increases while sediment yield remains low. Vegetation is quickly able to stabilize river banks, reducing sediment delivery, while evapotranspiration remains low, so that the runoff is high. Fluvial styles in aggrading valleys tend to change from braided (during glacial phases) to meandering (during interglacial phases; Vandenberghe et al. 1994). With increasing warmth, and consequently increasing vegetation density, rivers of anastomosed or meandering style tend to develop, the former particularly in coastal areas where the rate of generation of new accommodation space is high during the period of rapidly rising base level (Törnqvist 1993; Törnqvist et al. 1993; Fig. 6.20). 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, sediment yield remains low, and channel incision is enhanced. With the reduction in vegetation cover as the cold phase becomes established, sediment deliveries increase and fluvial aggradation is re-established.

It is apparent that fluvial processes inland and those along the coast are likely to be completely out of phase during the climatic and base-level changes accompanying glaciation. Within a few tens of kilometers 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, finally, it becomes buried. Inland, major erosional bounding surfaces correlate to times of climatic transition from cold to warm and from warm to cold, i.e., during times of rising and falling sea level, respectively. Minor climatic fluctuations lead to alternations of aggradation (terrace development) and valley incision, as has now been documented for several Gulf Coast rivers by Blum (1994) and Blum and Price (1994).

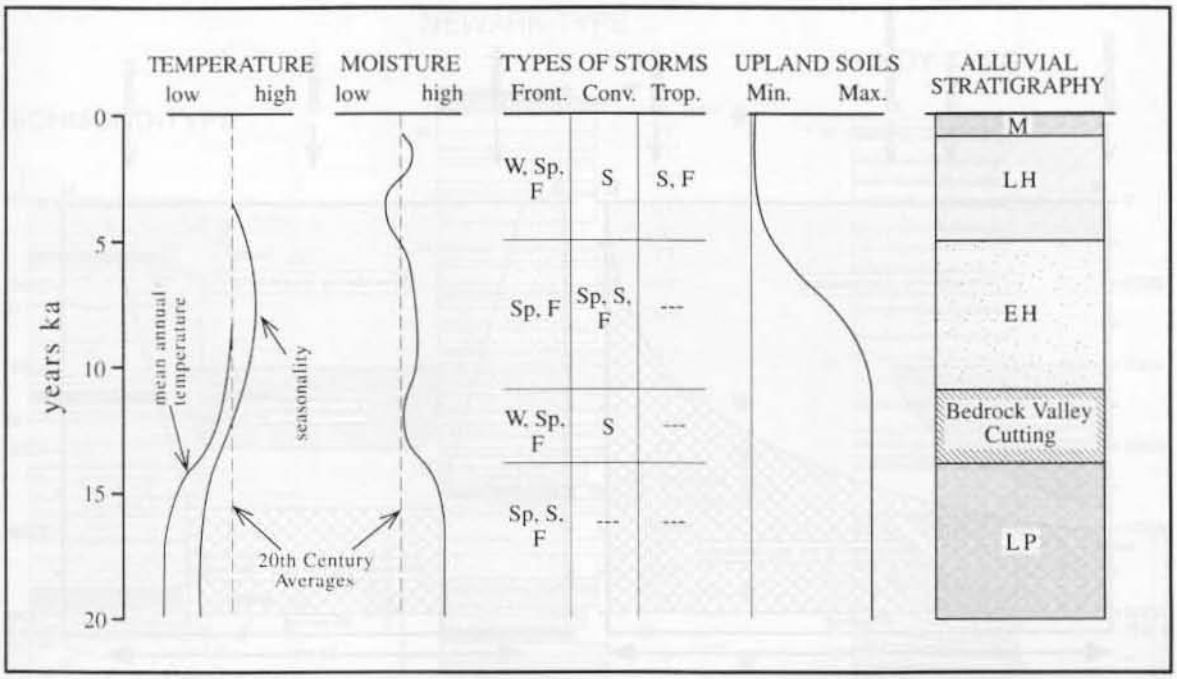


Fig. 6.18. Late-Pleistocene-Holocene climatic and environmental changes on the Edwards Plateau, Texas, correlated with the alluvial stratigraphy of the upper-Colorado drainage of the same area. *Fronts* Mid-latitude cyclones; *conv* convectional storms; *trop* tropical storms. Seasons are given by: *W* winter; *Sp* spring; *S* summer; *F* fall. The soil curve represents initial thick red clay formation during the late Pleistocene, and its gradual removal during the Holocene. *LP* Late Pleistocene; *EH* early Holocene; *LH* late Holocene; *M* modern alluvial deposits. (Blum 1994)

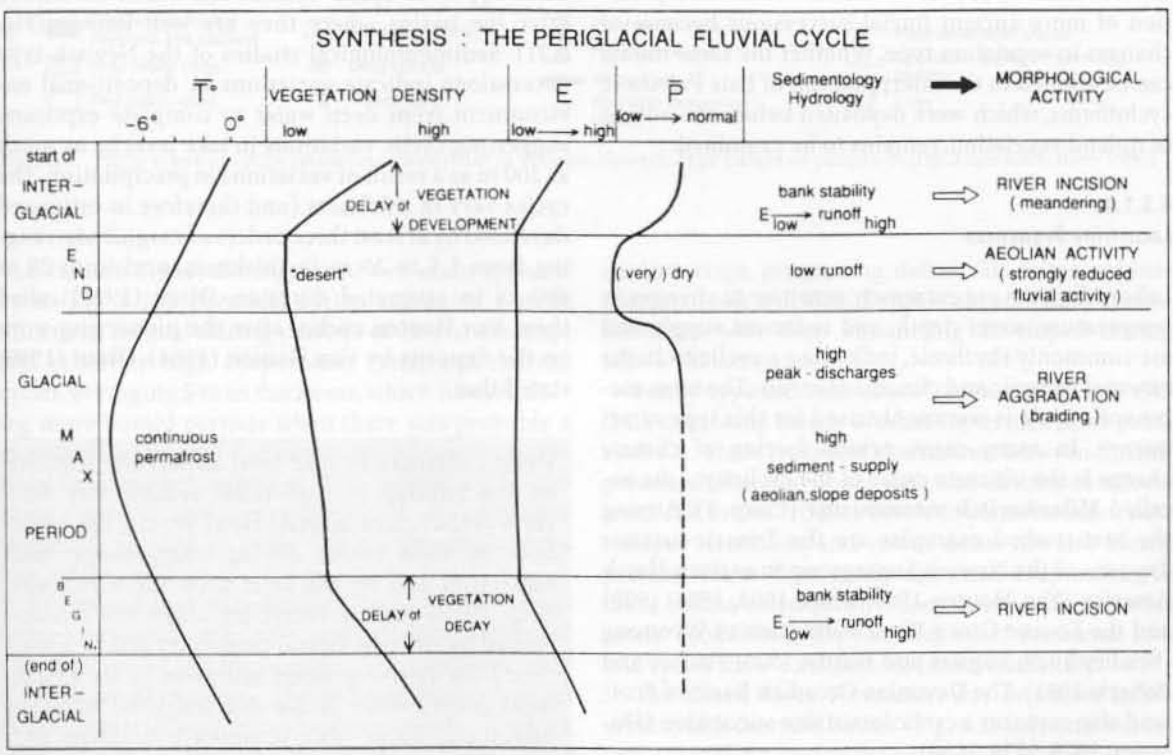


Fig. 6.19. A model showing the relationship between temperature, vegetation density, evapotranspiration (*E*), precipitation (*P*), and sedimentary processes in river systems during glacial and interglacial phases. (Vandenberghe 1993)

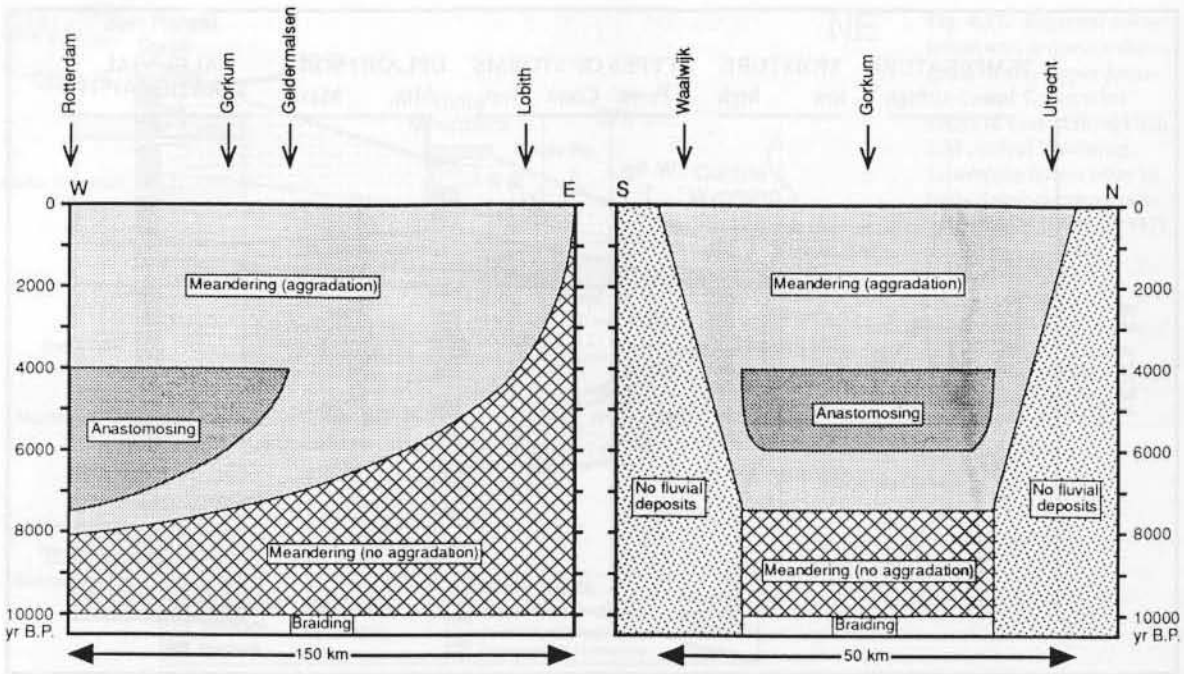


Fig. 6.20. Time-space model showing the Holocene development of channel patterns in the Rhine-Meuse delta. The "W-E" diagram is longitudinal; the "S-N" diagram is a transect across the alluvial valley. (Törnqvist 1993)

Although it provides a fascinating explanation for fluvial processes during the Late Cenozoic, this body of ideas must be used with caution for the interpretation of more ancient fluvial successions because of changes in vegetation type. Whether the same model can be applied to the interpretation of Late Paleozoic cyclothems, which were deposited before the advent of upland vegetation, remains to be examined.

6.5.1.2 *Lacustrine Sequences*

Lake sediments are extremely sensitive to changes in temperature, water depth, and sediment supply and are commonly rhythmic, indicating a cyclicity in the external tectonic and climatic controls. The term **meter-scale cycle** is commonly used for this type of sequence. In many cases, orbital forcing of climate change is the ultimate cause of the cyclicity – the so-called **Milankovitch mechanisms** (Chap. 8). Among the best-studied examples are the Triassic-Jurassic deposits of the Newark Supergroup in eastern North America (Van Houten 1964; Olsen 1984, 1986, 1990) and the Eocene Green River Formation of Wyoming (Bradley 1929; Eugster and Hardie 1975; Fischer and Roberts 1991). The Devonian Orcadian Basin of Scotland also contains a cyclic lacustrine succession (Donovan 1975, 1978, 1980).

The Newark basins are a series of extensional basins developed during the initial, rifting phase of the breakup of Pangea and the opening of the modern Atlantic Ocean. Olsen (1990) recognized three types of cyclic succession, which he named after the basins where they are well known (Fig. 6.21). Sedimentological studies of the Newark-type successions indicate variations in depositional environment from deep water to complete exposure, suggesting cyclic variations in lake level by as much as 200 m as a result of variations in precipitation. The cycles vary in thickness (and therefore in estimated duration) by at least three orders of magnitude, ranging from 1.5 to 35 m in thickness, and from 20 to 400 ka in estimated duration. Olsen (1990) called these **Van Houten cycles**, after the pioneering work on the deposits by Van Houten (1964). Olsen (1990) stated that

> Drastic changes in lake level apparently inhibited the buildup of high-relief sedimentary features (other than alluvial fans) within the basin both by wave action during transgression and regression and by the brief time the water was deep. Consequently, lacustrine strata are characterized by extreme lateral continuity and by a tendency for coarse-grained sediment to be absent from deeper water facies and restricted to basin margins ... Large-scale sequence boundaries and large deltas apparently are absent from the main basin fill.

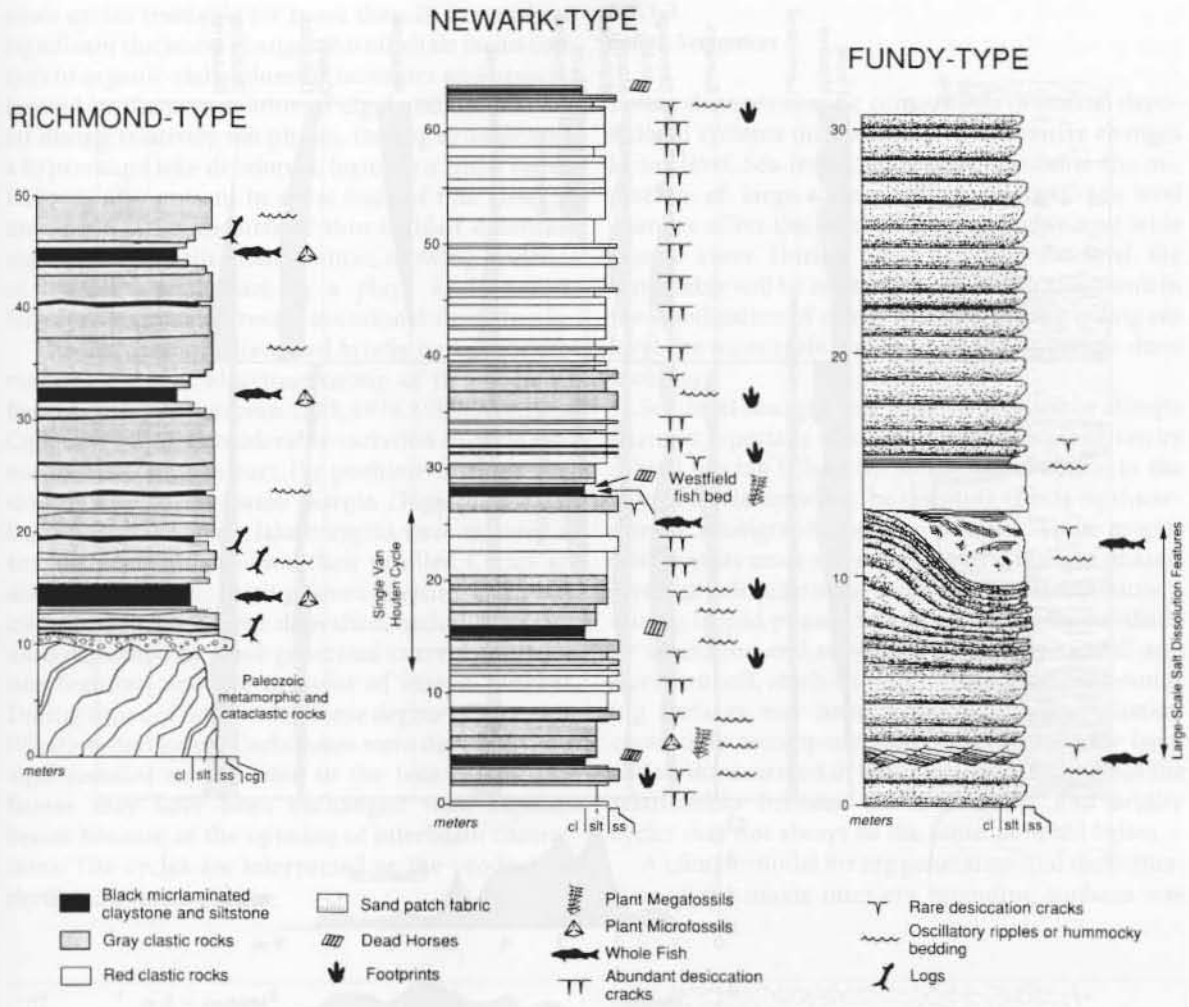


Fig. 6.21. Three types of cyclic lacustrine succession in Triassic Newark-type basins of eastern North America. (Olsen 1990)

Van Houten (1964) distinguished two main types of cycle (Fig. 6.22): chemical cycles, averaging 3 m in thickness, formed during periods of closed drainage and relatively high evaporation rates, and detrital cycles, averaging 5 m in thickness, which formed during more humid periods when there was probably a run-off flow through that maintained low concentrations of dissolved sediment. The detrital cycles coarsen upward and are interpreted as regressive sequences formed by basin filling following a rapid rise in lake level.

Richmond-type successions (Fig. 6.21) are characterized by significant coals, and by bioturbated and microlaminated siltstones and sandstones containing no evidence of exposure. The successions are thought to indicate relatively humid environments, in which complete basin desiccation was rare. During some cycles the lake may have remained deep long enough that high-relief sedimentary features were built, including large, prograding deltas, fan deltas, submarine channels and fans. Large-scale erosional unconformities with deep channeling developed during lake-level lowstands.

Fundy-type successions are "characterized by a cyclicity consisting mostly of what are termed sand-patch cycles ... These represent alternations between shallow perennial lakes and playas with well-developed efflorescent salt crusts" (Olsen 1990). Gypsum nodules, salt-collapse structures, and eolian dunes are also locally present. Water flow rarely exceeded evaporation in these basins; consequently, deep-lake conditions were rarely established, depositional relief remained low, and cycles show considerable lateral continuity.

The Green River Formation was deposited in several arid lacustrine basins (Fig. 6.23). Annual varves and solar-band cycles are prominent in these deposits (Fischer and Roberts 1991) but are not discussed here. Milankovitch cycles are well developed, with

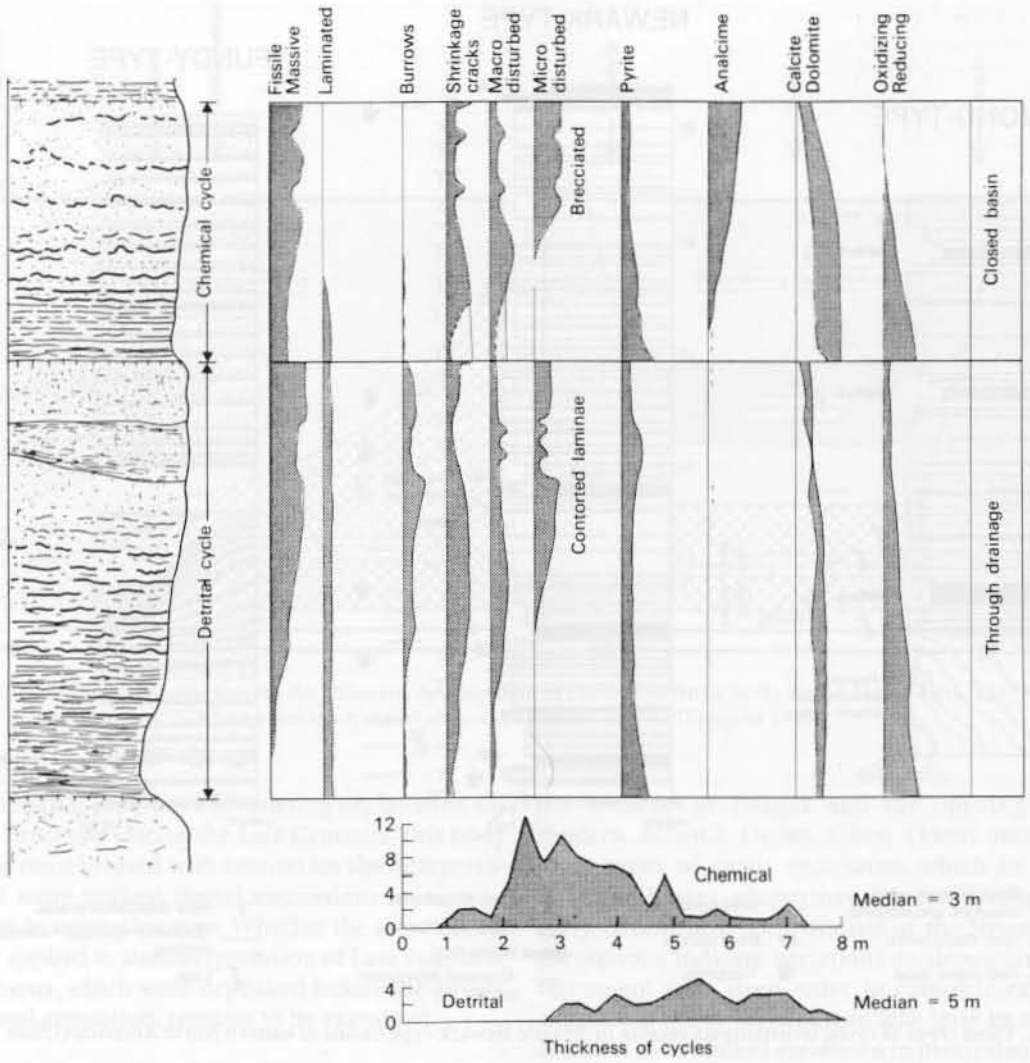


Fig. 6.22. Model of detrital and chemical cycles in the Lockatong Formation (Triassic), Newark Basin, New Jersey. (Van Houten 1964)

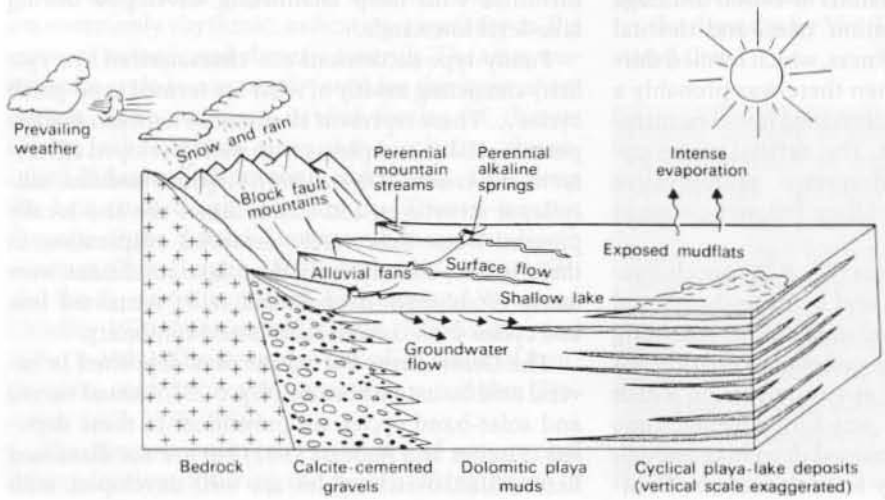


Fig. 6.23. Schematic block diagram showing depositional framework of the Wilkins Peak Member, Green River Formation. (Allen and Collinson 1986, after Eugster and Hardie 1975)

some cycles traceable for more than 20 km without significant thickness changes. An oil-shale facies consists of organic-rich dolomitic laminites and breccias formed by the preservation of algal and fungal matter during relatively wet phases. During drier phases, a hypersaline lake developed, forming a trona facies. Halite is also present in some beds of this facies. A marlstone facies consists of thin-bedded dolomitic mudstones with silt–mud laminae, showing evidence of frequent desiccation in a playa environment. Silt–mud laminae represent occasional sheet floods.

The last example discussed briefly here is the Devonian Caithness Flagstone Group of the Orcadian Basin, Scotland (Donovan 1975, 1978, 1980; Allen and Collinson 1986). Considerable variation in cycle type occurs, reflecting, in part, the position of the succession relative to the basin margin (Figs. 6.24, 6.25). During low lake levels, lake margins were exposed to erosion, and rivers incised their profiles. Cycles are dominantly clastic. During times of rising lake level, transgressive beds were deposited, including clastic units deposited by wind-generated currents, desiccation features, and the deposits of inland sabkhas. During times of deep water, some degree of lake stratification developed. Carbonates were deposited, and algal remains accumulated in the lake centre. Fish faunas may have been exchanged with adjacent basins because of the opening of interbasin connections. The cycles are interpreted as the products of rhythmic climate change.

6.5.1.3 *Eolian Sequences*

Eolian deposits may be components of coastal depositional systems that are affected by relative changes in sea level. Sea-level change may also cause the migration of large-scale sand seas (**ergs**). Sea-level changes affect the level of the water table over wide coastal areas. During times of rising sea level, the water table will be relatively high, which may result in the stabilization of dunes whereas, during falling sea level, the water table may fall, permitting greater dune mobility.

Sea-level changes may be accompanied by climate changes, especially when both processes are driven by orbital forcing (Chap. 8). As discussed above, in the case of fluvial systems, the resulting effects on the sequence stratigraphy may be complex. Dune migration is at its most active during arid climate phases, given an adequate sand supply, and is at a minimum during humid phases, when dunes may be stabilized by vegetation and subject to erosion by rainfall and fluvial runoff, at which times widespread erg bounding surfaces may form. Variations in precipitation commonly accompany glacial cycles and have been widely documented in the Quaternary record, but the relationship between sea-level cycles and aridity cycles may not always be the same, as noted below.

A climate model for erg generation and the formation of the major inter-erg bounding surfaces was

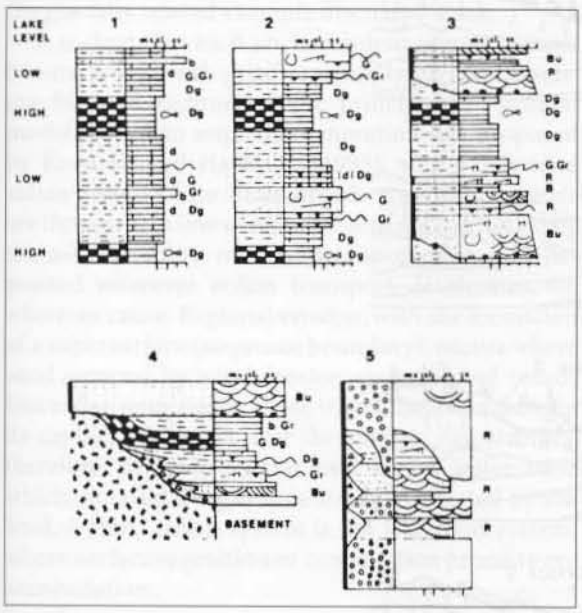


Fig. 6.24. Styles of cyclic sedimentation in the Caithness Flagstones, Scotland. Five types of cycle may be present, depending on the location of the section relative to the lake margin. (Donovan 1978)

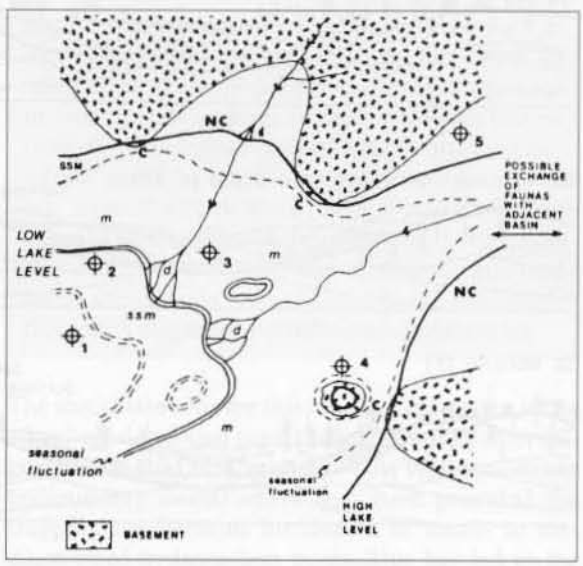


Fig. 6.25. Model of Orcadian basin margin, showing fluctuation in lake levels. Numbered circles indicate suggested positions of sedimentological profiles illustrated in Fig. 6.24. (Donovan 1978)

first proposed by Talbot (1985). These surfaces were termed **superbounding surfaces** by Kocurek (1988). Application of the model, with discussion of a range of examples, was provided by Clemmensen and Hegner (1991), Kocurek et al. (1991), Brookfield (1992), Kocurek and Havholm (1993), and Kocurek (1996).

A careful study of the coastal ergs of Mauritania led Kocurek et al. (1991) to develop the sequence model shown in Fig. 6.26. Erg migration and dune generation was most active during the so-called Ogolian climate phase, between 20 and 13 ka. This was an arid phase corresponding to a northern-hemisphere glacial maximum. Ergs were degraded and stabilized during the subsequent interglacial period, which was more humid. The main inter-erg bounding surface was formed at this time. Its shape partly reflects the geometry of the underlying draas. Dunes were reactivated a few thousand years ago. Note that, in this example, the main bounding surface, which would be designated stratigraphically as the sequence boundary, developed during the interglacial phase (a time of high sea level) and is therefore 180° out of phase, in terms of the sea-level cycle, with bounding surfaces formed by subaerial erosion of the continental shelf during glacioeustatic lowstands. Kocurek et al. (1991) also cautioned that draa sands and the interdraa lacustrine and coastal sabkha sediments were not formed simultaneously but represent entirely different events developed at different parts of the climatic cycle.

Blakey and Middleton (1983) reconstructed a cycle of coastal erg migration and deflation in the Coconino Sandstone (Permian) of Arizona (Fig. 6.27), which developed in response to a combination of climate and sea-level changes. The superbounding surfaces form during periods of low sea level and regional deflation. They are overlain by coastal sabkha deposits formed during transgression. Ergs are established at times of high accommodation and gradually expand seaward during the next cycle of relative sea-level fall.

Multiple climatic cycles driven by Milankovitch processes are recorded in the Rotliegende Sandstone (Permian) of the North Sea Basin, which was a giant

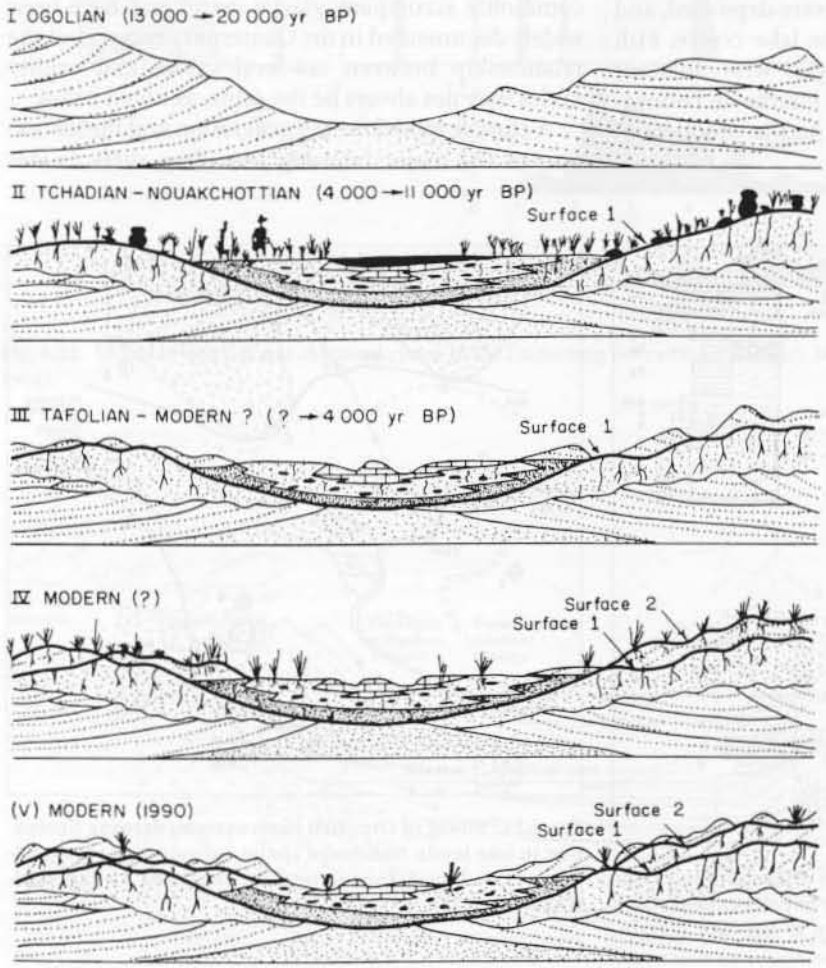


Fig. 6.26. A model for the development of the Akchar Erg, Mauritania. See text for explanation. (Kocurek et al. 1991)

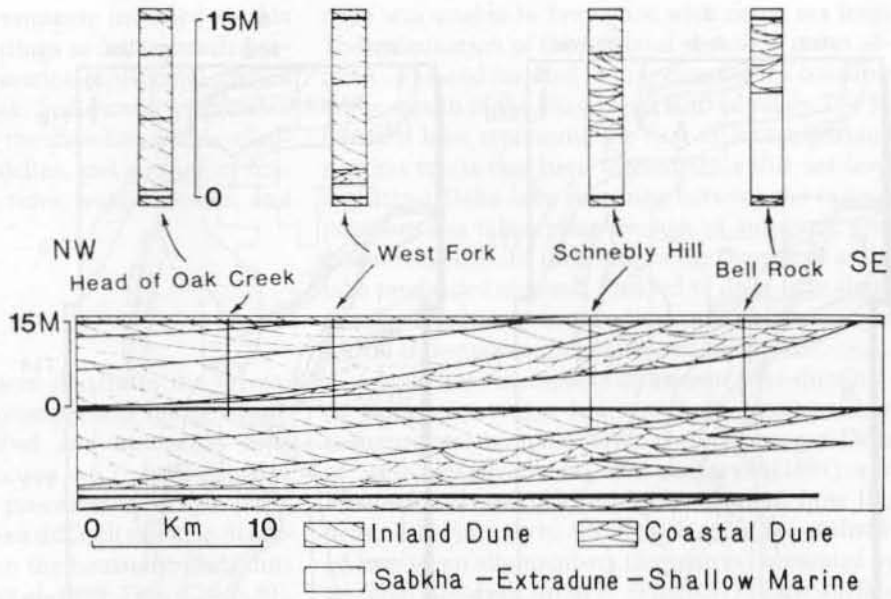


Fig. 6.27. Correlation of super surfaces and major facies in the Coconino Sandstone, Arizona. (Blakey and Middleton 1983)

inland basin isolated from the sea at that time. Here, widespread eolian dune deposits correspond to times of aridity, when water levels in the basin were low. The sequence boundaries were generated during the time of maximum rate of fall, as aridity was undergoing a rapid increase. Eolian deposits are, therefore, most important in the highstand and overlying lowstand deposits and are least important during the transgressive phase, when sabkha and fluvial deposits are dominant (Yang and Nio 1993). This contrasts with the glacially related example discussed above.

It is clear that each ancient eolian sequence must be interpreted with great care, applying basic stratigraphic and sedimentologic principles. A general model for eolian sequence generation was proposed by Kocurek and Havholm (1993), who subdivided eolian systems into three broad types. Dry systems are those with a low water table, and are controlled by the aerodynamics of sand transport. Ergs are deposited wherever eolian transport decelerates, for whatever cause. Regional erosion, with the formation of a supersurface (sequence boundary), occurs where sand removal by wind erosion exceeds sand influx. Wet eolian systems are those where the water table or its capillary fringe is near the surface. Erg stability, therefore, depends on the level of the water table which, in coastal regions, is strongly affected by sea level. A third eolian system is the stabilized system, where surface vegetation or cementation promote erg accumulation.

6.5.1.4 *Incised Valleys and Estuaries*

As noted by Dalrymple et al. (1994) in the preface to their volume on incised valley systems,

> Incised valleys were not widely recognized prior to the 1980s, although there are notable exceptions. Most early workers forced the isolated, incised-valley deposits along an unconformity into a single continuous unit, ignored them by including them within larger stratigraphic units, or interpreted them as deltaic distributaries or non-incised fluvial channels. In the last decade, intense interest in the influence that changes in accommodation space have on stratigraphic organization has focused attention on incised-valley systems, because they are one of the most visible records of major decreases in accommodation. In practical terms they are also a significant key to the identification of sequence-bounding unconformities.

The initial stimulus for this renewed interest in incised valleys developed from the realization of their importance in the Cretaceous fill of the Western Canada Sedimentary Basin, where they have provided the trapping mechanisms for dozens of small- to medium-sized hydrocarbon pools. This has led to the mapping of a large number of very complex and widespread valley systems at many stratigraphic levels by petroleum geologists, aided by the extensive public subsurface database available for this basin (Fig. 6.28). Because of the interaction of terrestrial

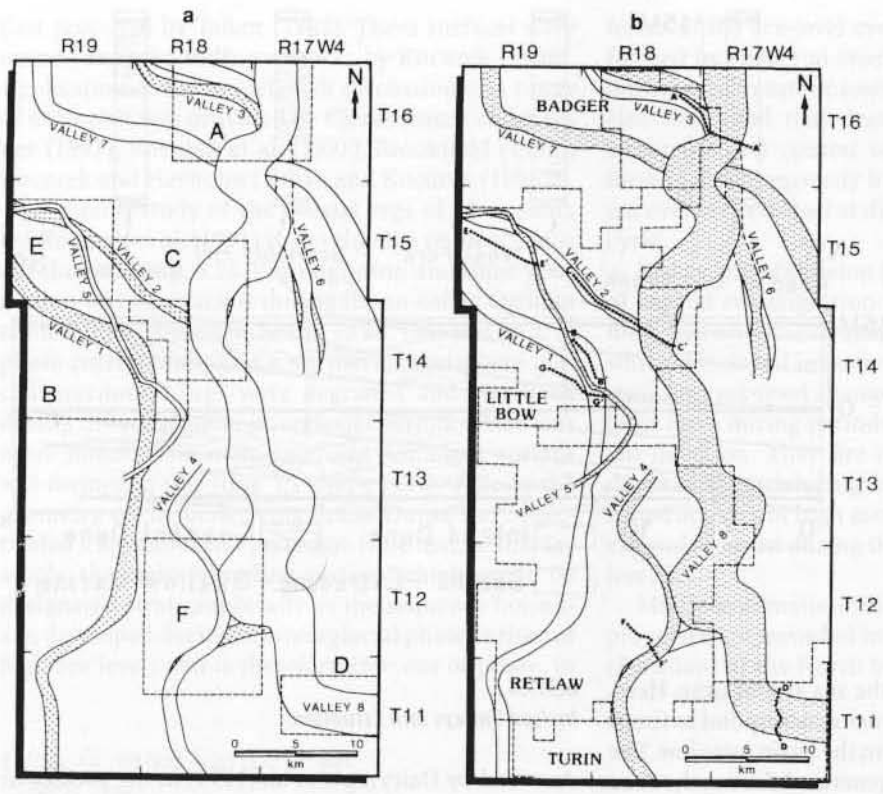


Fig. 6.28. Valleys in the Glauconite Sandstone, Upper Mannville Group (Lower Cretaceous), southern Alberta. (Wood and Hopkins 1992)

and marine processes in their filling, and because their relatively large local accommodation space permits preservation of a larger part of each accommodation cycle than do adjacent areas, incised valleys are particularly rich in information about the various factors influencing sedimentation. Commonly, the flanks of incised valleys pass laterally into erosion surfaces (sequence boundaries) along which much or all evidence of a cycle of stratigraphic cut and fill may otherwise be lost.

Incised valleys are, typically, estuarine during their fill phase. Excellent models of estuarine sedimentation have been developed by Allen (1991) and Dalrymple et al. (1992) based, in part, on work in the Gironde (France) and Bay of Fundy (Canada) estuaries, respectively. This work has shown that estuaries can be subdivided into four main parts (Fig. 4.58), with the stratigraphic fill subdivisible into three broad segments (Fig. 4.59). The recognition of tidal signatures in an otherwise fluvially dominated succession may be the key to the recognition of marine transgressive deposits and the distinction between simple fluvial valleys and marine-influenced estuaries. Such tidal signatures may take the form of inclined heterolithic stratification, or crossbedding showing tidal couplets and rhythmic, sigmoidal foresets (Sects. 4.5.4, 4.6.5).

6.5.2 Marginal-Marine and Shallow-Marine Sequences

More attention has probably been paid to siliciclastic sequences in marginal-marine and shallow-marine environments than those of any other setting. In part, this is probably because of the sensitivity of the environment (and, hence, of the deposits) to sea-level change. The deposits, therefore, provide ideal field laboratories for the testing of sequence concepts. In addition, the focus on these deposits reflects their importance to petroleum geologists. Much research has been carried out on units such as the Cretaceous Viking and Cardium Formations in Alberta, which are important petroleum reservoirs and which have been the sources of several important developments in the area of sequence stratigraphy. The Brent Group in the North Sea Basin has performed a similar function for European petroleum geologists. In addition, the famous Book Cliffs section of east-central Utah and the Gallup Sandstone of the San Juan Basin in New Mexico, which consist of suites of Cretaceous sequences, have been the sites of innumerable field trips for petroleum geologists and the locations of many critical studies of shallow-marine to nonmarine sequence stratigraphy by American, Canadian, British, Scandinavian, and other geologists.

The depositional environments included in this section are such coastal settings as deltas, beach-barrier systems, and shallow-marine continental shelves out to the shelf-slope break. Sedimentary processes include fluvial input along the shoreline, the development of river-dominated deltas, and a range of marine processes, including tides, waves, storms, and geostrophic currents.

6.5.2.1 *Deltas*

The Mississippi-delta system illustrates the formation of sequences and parasequences under conditions of changing sea-level and autogenic delta switching. As noted in Section 4.6.2, distinguishing autogenic from allogenic processes in deltaic parasequence generation can be a difficult exercise. Six sequences have developed on the Louisiana shelf during the last 700 ka (Boyd et al. 1989; Figs. 6.29, 6.30). Sequence 1 is interpreted to represent the lowstand associated with the Illinoian glacial stage. Sequences 5 and 6 correspond to the post-Wisconsinan sea-level rise since 18 ka. The Mississippi-delta complex developed during this final stage of sea-level rise. The first three lobes (numbered 1–3 in Figs. 6.29, 6.30) are interpreted as transgressive systems tracts. They backstep onto the shelf because sediment supply to the delta was unable to keep pace with rising sea level. The culmination of the sea-level rise occurred at about 3–4 ka and resulted in the retreat of the coastline to the mouth of the Mississippi alluvial valley. The St. Bernard lobe represents the first of four highstand systems tracts that have formed since this sea-level highstand. Delta-lobe switching between the various positions has taken place because of autogenic processes related to the flattening of the river slope as the lobe prograded seaward. This led to delta-lobe abandonment, sediment compaction and subsidence, and a local transgression, as illustrated in Fig. 4.57.

An excellent example of an ancient river-dominated delta system that has now been interpreted in sequence-stratigraphic terms is the Dunvegan Delta of west-central Alberta. Bhattacharya (1991) and Bhattacharya and Walker (1992) showed how this heterolithic unit up to 300 m thick could be subdivided into seven allomembers (sequences) separated by regional flooding surfaces (Fig. 6.31). Each surface extends over areas of up to tens of thousands of square kilometers and is typically represented in core by a sharp demarcation between nonmarine heterolithic sandstones and mudstones with roots below, and marine mudstones above. Evidence of erosion or transgression in the form of lag deposits is typically absent above these sequence boundaries. Following rapid transgression, the upper part of each allomem-

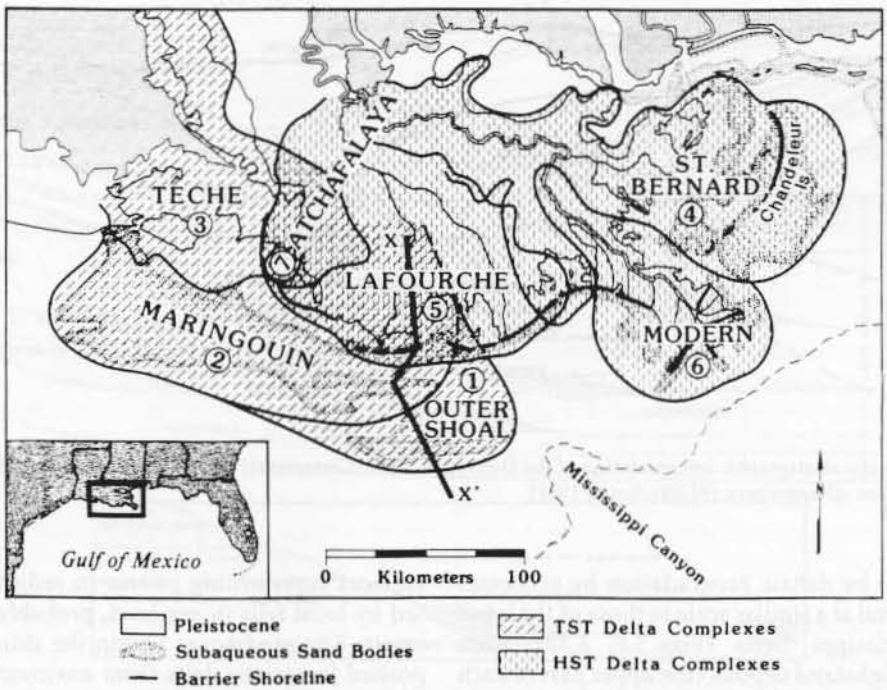


Fig. 6.29. The delta lobes of the Holocene Mississippi River. Lobe *1* is dated at about 9 ka and represents the first of three transgressive systems tracts formed following the end of the Wisconsinan glacial stage. Lobes *4–7* are highstand systems tracts (Boyd et al. 1989). The cycles of sea-level change documented for this area are shown in Fig. 8.24

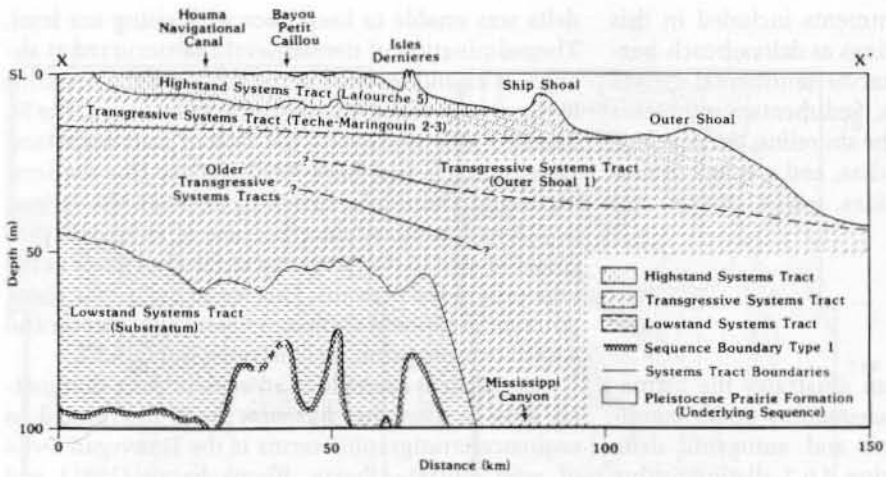


Fig. 6.30. Cross section along line X-X' in Fig. 6.29, showing the systems tracts of the Mississippi-delta complex. The "type-1" sequence boundary at the base of the complex is dated at about 18 ka and was formed by subaerial erosion during the last Wisconsinan lowstand. It is followed by a lowstand systems tract formed between 18 and 9 ka. (Boyd et al. 1989)

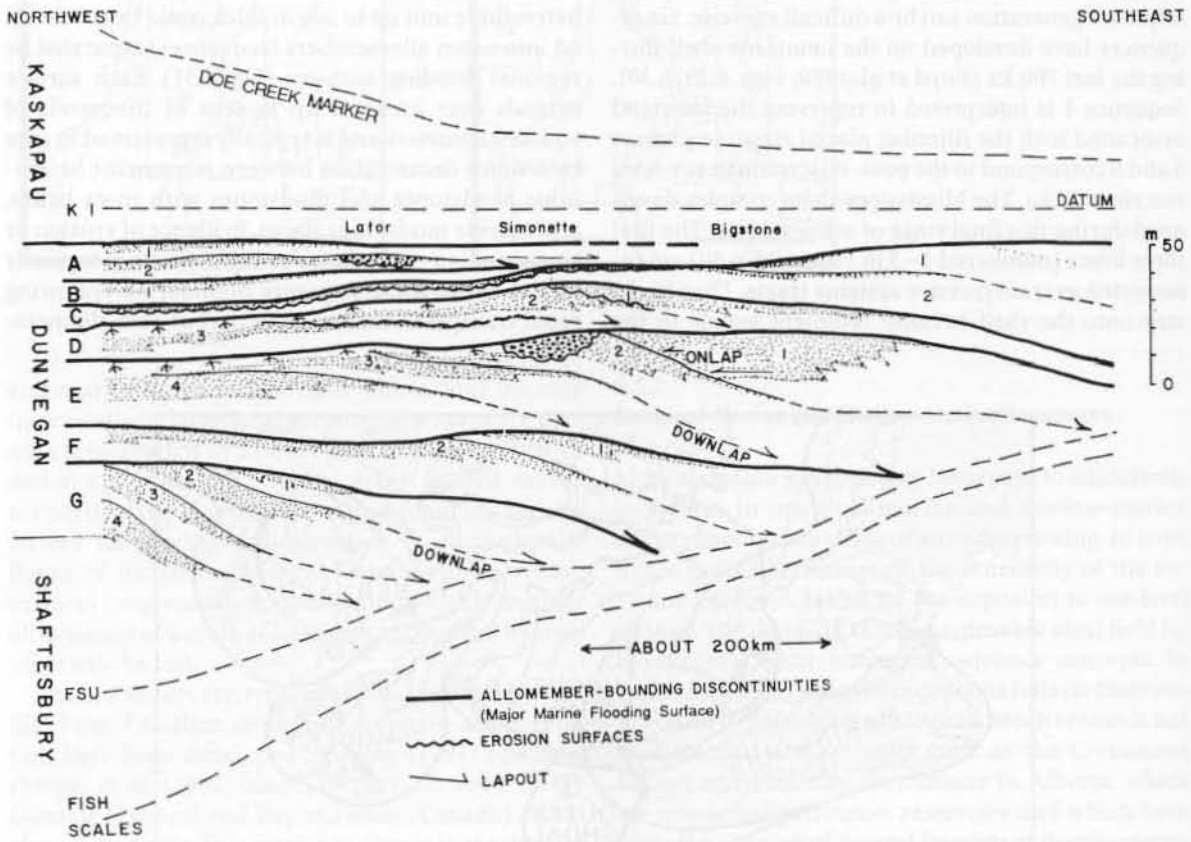


Fig. 6.31. Sequence-stratigraphic interpretation of the Dunvegan Delta (Cretaceous) of west-central Alberta, showing the subdivision into seven allomembers. (Bhattacharya 1991)

ber is formed by deltaic progradation by processes comparable and at a similar scale to those of the lobes of the Mississippi Delta (Figs. 5.6, 6.32). Each prograding highstand deposit (the upper part of each allomember) may be subdivided into a series of "shingles" constituting a progradational parasequence set. Shingles are bounded by minor erosion surfaces representing pauses in sedimentation caused by local falls in sea level, probably of autogenic origin. The sandstones within the shingles were deposited in various delta-front environments, including distributary-channel and mouth bar settings. In a few localities, the fluvial feeder channel for the distributaries is preserved in the form of a coarser-

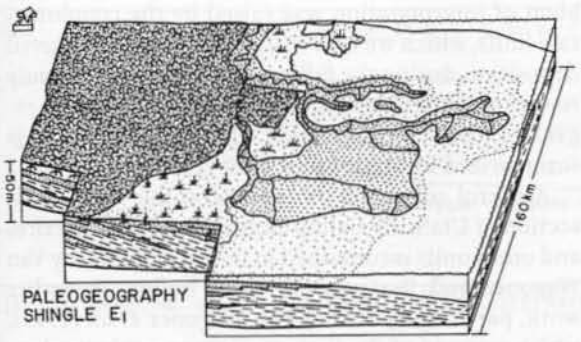


Fig. 6.32. Reconstruction of the formation of shingle E1 in the Dunvegan delta, as the product of progradation of a river-dominated delta comparable in scale to those of the modern Mississippi delta. (Bhattacharya 1991)

grained, erosionally based unit incised into the top of the shingle. For example, the U-shaped channel fill shown at the top of shingle E2 in Fig. 6.31 is interpreted as the deposit of the feeder channel for shingle E1.

As noted in Section 4.6.2, wave- and storm-influenced deltas are commonly developed at shelf-margin settings during lowstand, whereas tide-influenced deltas are most likely to develop during transgression. The facies assemblages of wave-influenced deltas are similar to those of beaches, the sequence stratigraphy of which is discussed below. Tidal deposits are also typical of estuaries, which are described in Section 6.5.1.4 above. Using the techniques described in Section 5.3.3, the deltaic form of these deposits may be revealed by isolith mapping of unit thickness or sand content.

6.5.2.2 *The Coast-Shelf Interface*

The architecture of sequences formed in marginal-marine to shallow-marine environments is highly sensitive to surface marine processes, including waves, tides, and storms, and to sea-level change. One of the first such units to be analyzed from a sequence-stratigraphic perspective was the Cardium Formation of Alberta. As explained in Section 3.4.3.2, the definition of a suite of through-going surfaces of erosion and transgression in the Cardium sandstone by Plint et al. (1986) was a major breakthrough that led to a complete reappraisal of the paleogeographic history of this unit. Another important contribution by Plint (1988, 1991) was the recognition of the potential importance of the falling stage of the sea-level cycle in the generation of shallow-marine sequences. As noted in Section 6.3, falling-stage deposits did not form part of the original suite of systems tracts defined by Posamentier and Vail (1988) and Posamentier et al. (1988).

The first sequence-stratigraphic interpretation of the Cardium Sandstone is shown in Fig. 3.5. A more recent interpretation of a different part of Alberta is given in Fig. 6.33, and Plint's concepts regarding falling-stage deposition are illustrated in Figs. 6.34 and

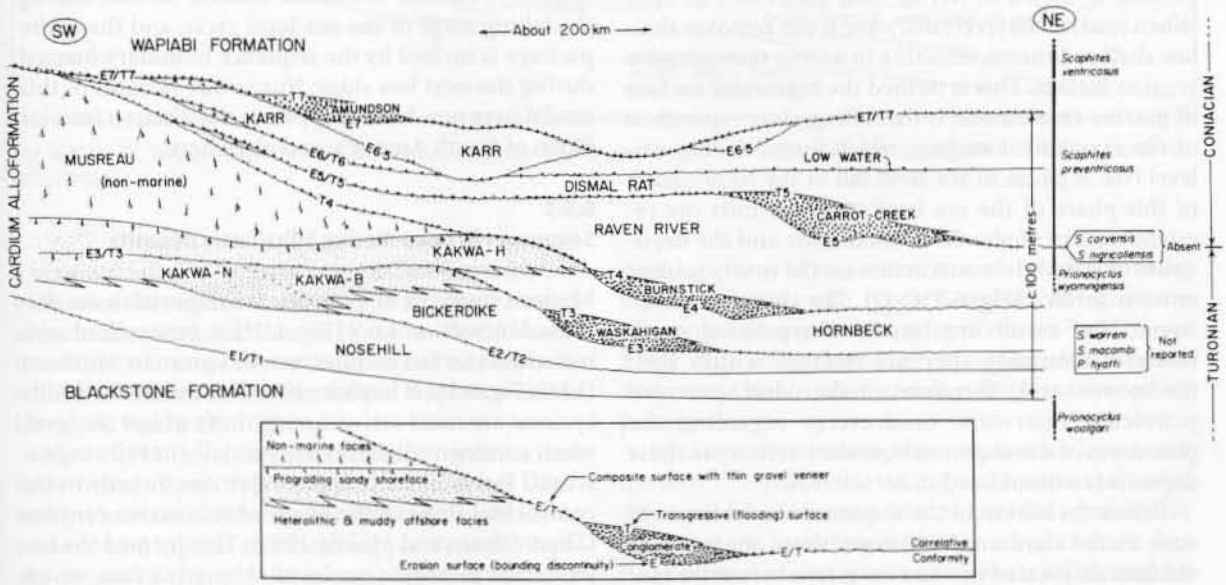


Fig. 6.33. Allostratigraphy of the Cardium Formation, Alberta Basin. *E* Surface of erosion; *T* surface of transgression. In many places, these surfaces are coplanar, but, where they are separate, they record the transgressive modification of falling-stage to lowstand shoreface conglomerate bodies. The seven E/T surfaces define a suite of virtually basin-wide allomembers. (Walker and Plint 1992)

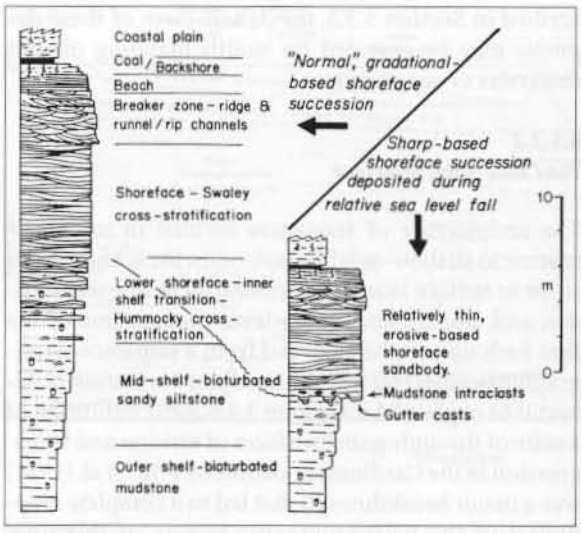


Fig. 6.34. Wave-dominated shoreface deposits, characterized by upward-coarsening successions of mudstone, siltstone, and sandstone with abundant hummocky cross-stratification (*HCS*). The column at right shows a "sharp-based sandstone" in which the shoreface-shelf transition is missing at an erosion surface. The formation of this succession is illustrated in Fig. 6.35. (Walker and Plint 1992)

6.35 (based on his most recent work with R.G. Walker). The upward-coarsening profile shown at left in Fig. 6.34 is the type of succession formed by beach deposition and regression during an episode of continuous sedimentation. Shoreface deposits fill the available accommodation by lateral aggradation. This process is shown at left in each panel of Fig. 6.35. When relative sea level falls, wave scour removes shallow shelf sediments, resulting in a time-transgressive erosion surface. This is termed the **regressive surface of marine erosion** and is the falling-stage equivalent of the ravinement surface, which forms during sea-level rise. A pause in sea-level fall or the termination of this phase of the sea level cycle, permits the re-establishment of shoreface conditions and the deposition of a shoreface succession on the newly formed erosion surface (Fig. 6.35C, D). The shoreface sandstones that result are termed **sharp-based sandbodies**. Commonly, they are isolated within shelf mudstones and, therefore, make ideal potential petroleum reservoirs. Controversy regarding the placement of the sequence boundary relative to these deposits is summarized in Section 6.4.

Before the advent of the sequence model for units such as the Cardium Sandstone, these sharp-based sandstones, located in some cases tens to hundreds of kilometers seaward from the presumed basin-margin shoreline, were interpreted as "offshore bars," but no convincing depositional mechanism had been proposed for the origins of these units. The same problem of interpretation was raised by the conglomerate units, which we now realize are formed by fluvial deposition during the falling stage and subsequently reworked by transgression. The ongoing debate regarding the reinterpretation of these shelf units is summarized by Walker and Plint (1992).

A useful collection of articles on the Book Cliffs section of Utah, the Gallup Sandstone of New Mexico, and other units is contained in the book edited by Van Wagoner and Bertram (1995). It builds on earlier work, particularly that of Van Wagoner et al. (1990), which was one of the first attempts to apply modern sequence concepts to core and outcrop data. Some of the papers in Van Wagoner and Bertram (1995) reveal a heavy influence of the Exxon "school" in the use of terminology, models, and references to the eustasy model of sequence generation but, whether these models are appropriate or not (see discussion of them in Chapter 8), the book is packed with field documentation and field-based sequence interpretations and is, therefore, an invaluable reference source.

Figure 6.36 illustrates the large-scale architecture of the Book Cliffs section, a classic example of a suite of progradational shallow-marine to nonmarine sequences, and Fig. 6.37 illustrates a general model for the Gallup Sandstone of New Mexico. The latter shows how the sequence boundary is followed by estuarine deposits formed during transgression. Contemporaneous shelf deposits are stripped off by ravinement, forming a widespread ravinement erosion surface. Highstand deposits are represented by the overlying shoreface sandstones, which overlie a regressive surface of marine erosion formed during the falling stage of the sea-level cycle, and the entire package is incised by the sequence boundary formed during the next low stage. Numerous variants of this model have now been mapped in the Western Interior Basin of North America, and elsewhere.

6.5.3 Sequences in Deep-Marine Siliciclastic Deposits

Modern concepts of deep-sea fan deposition are discussed in Section 4.6.9 (Fig. 4.73). A generalized seismic model for fan architecture was given by Mitchum (1985; Fig. 6.38) It has long been known that turbidite systems are most active during times of low sea level, when continental shelves are partially or fully exposed and fluvial systems deliver detritus directly to the continental slope or the heads of submarine canyons (Shanmugam and Moiola 1982). This formed the basis for the sequence model of submarine fans, which placed them in the lowstand and transgressive systems tracts (Posamentier and Vail 1988; Posamentier et al. 1988). However, Hunt and Tucker (1992) pointed out that submarine-fan systems are deposited during

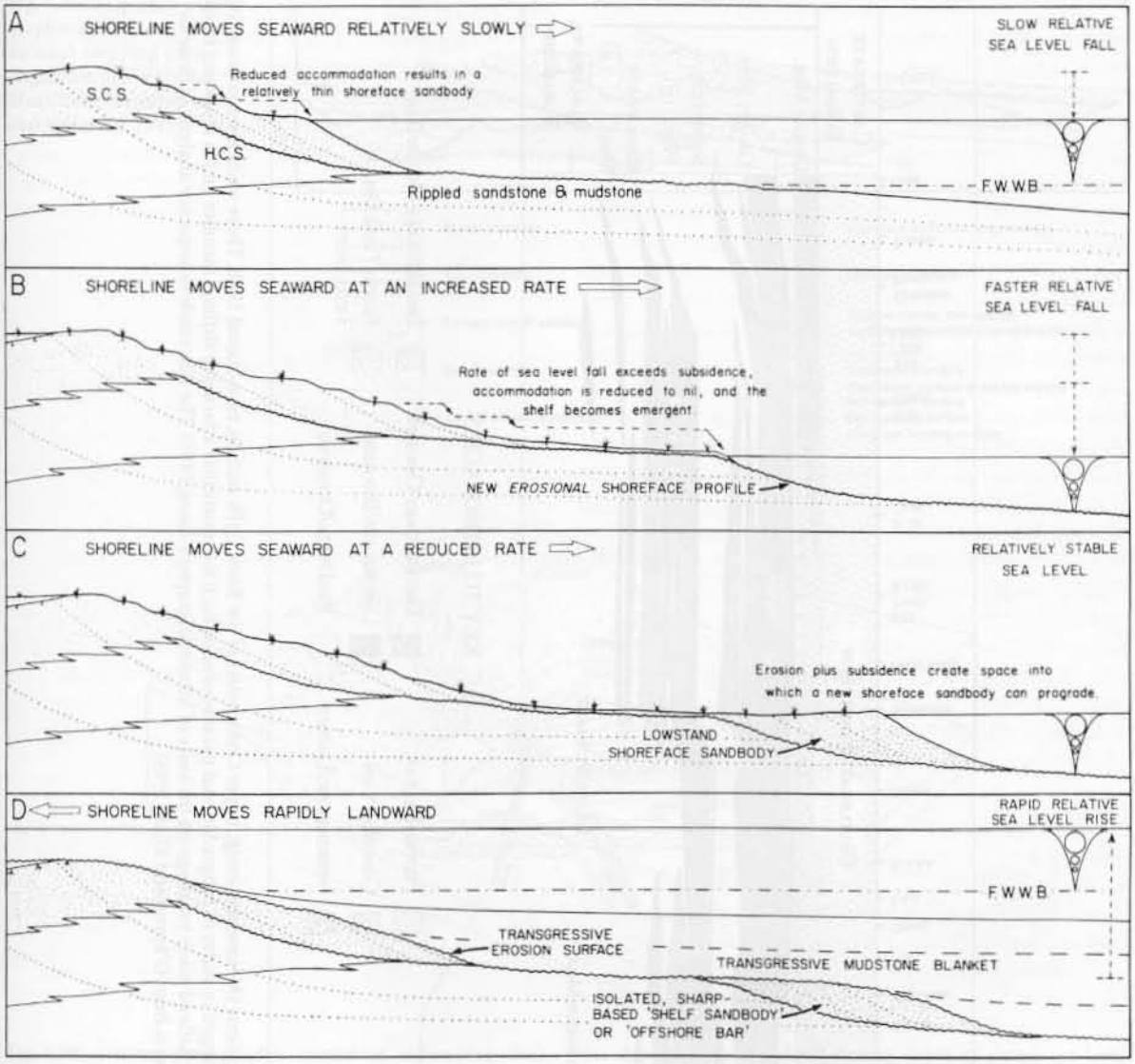


Fig. 6.35 A–D. Interpretation of the formation of sharp-based sandstones by marine erosion during time of falling sea level. (Plint 1988)

a sea-level fall as a result of progressive exposure and erosion of the continental shelf. The sequence boundary, which records the time of lowest sea level, should therefore be located above the lowstand fan deposits, not below them, as in the Exxon sequence model (Sect. 6.4; Figs. 6.13, 6.14).

Mutti (1985) subdivided turbidite systems into three broad types. Type-I deposits consist of thick-bedded lobes of sandstone, formed at times of lowest sea level and highest sediment supply. Type-III deposits consist of channel–levee complexes formed during higher stands of sea level, when detrital supply is smaller. Type II is an intermediate style, when the submarine fans may contain well-developed, thin-bedded lobe fringes. These three broad types of assemblage form in succession to each other, according to Mutti (1985), with types II and III downlapping onto type I during the rising leg of the sea-level curve. These late-stage turbidite systems also backfill canyons at this time, forming a deep-marine onlap pattern.

It is important to note, however, that sea level is not the only control on the timing and location of turbidite systems. Kolla and Macurda (1988) pointed to other factors, including the tectonic and climatic settings of the receiving basin. Canyons, which may be the main source of sediment feeding a fan, may indeed be initiated during lowstands by fluvial erosion of the shelf, but they may continue to evolve and deepen by submarine mass wasting during the sub-

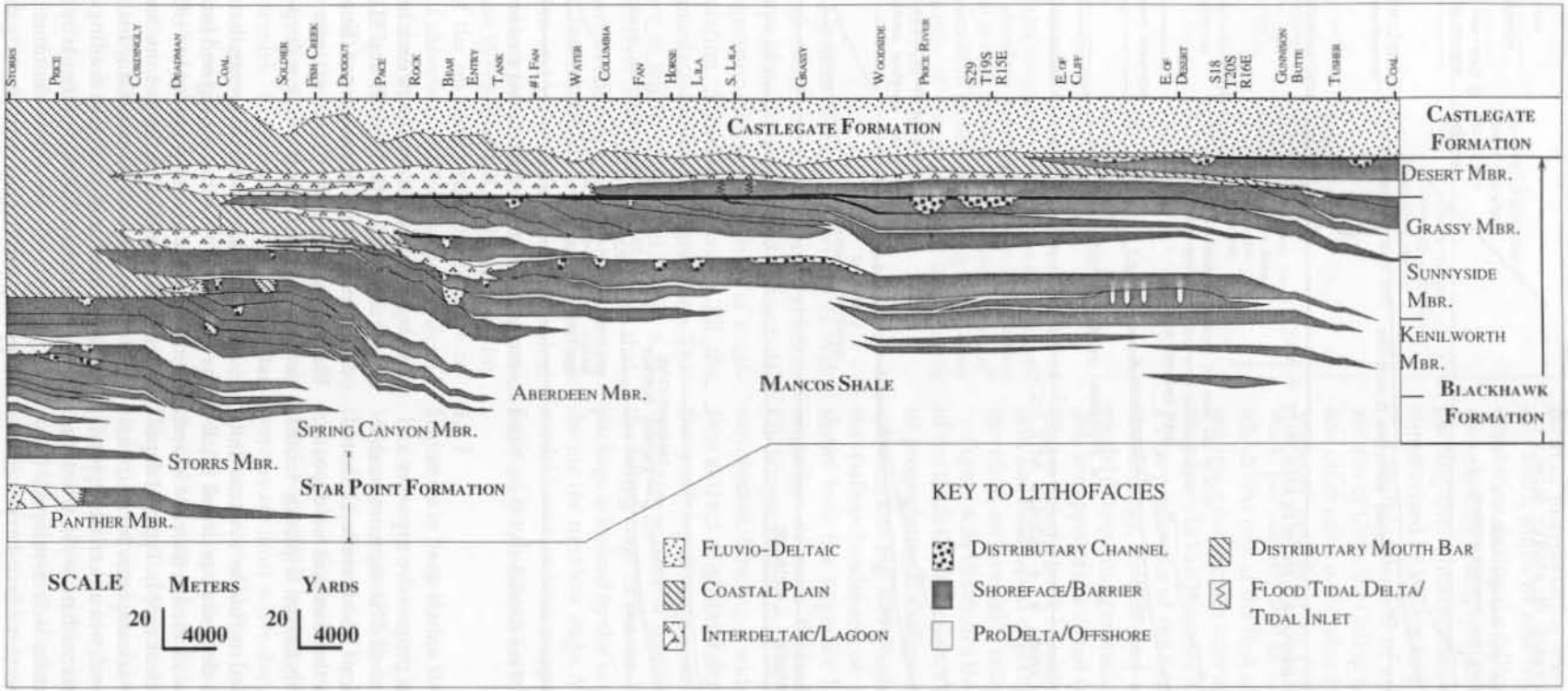


Fig. 6.36. Stratigraphy of the Star Point and Blackhawk Formations (Mesaverde Group, upper Cretaceous) of the Book Cliffs section, east-central Utah. These units are composed of a succession of coastal plain, shoreface, and shelf deposits arranged in suites of progradational parasequences. Each set constitutes a formally defined member and is separated from the next set by a widespread marine shale. The members were defined many years ago on the basis of lithostratigraphy (Young 1955). The first modern sequence analysis of these deposits was provided by Van Wagoner et al. (1990). This diagram is from O'Byrne and Flint (1995)

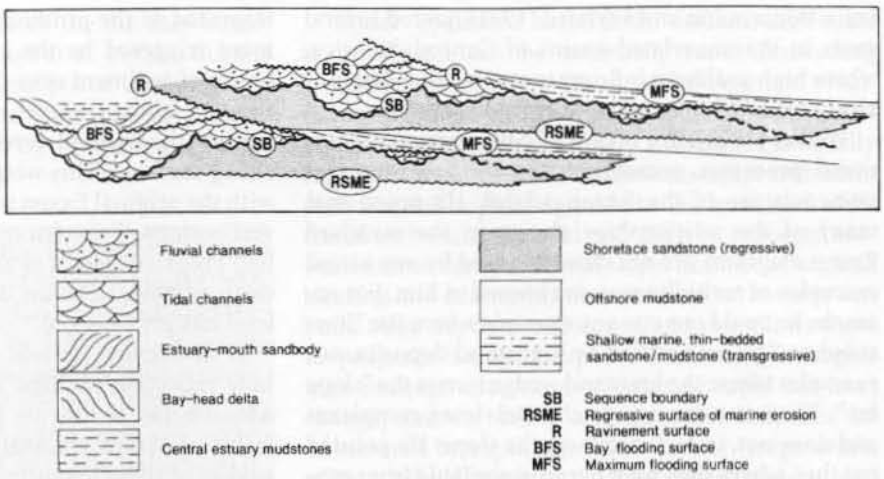


Fig. 6.37. Sequence-stratigraphic model for two forward-stepping parasequences of the Gallup Sandstone. (Nummedal and Molenaar 1995)

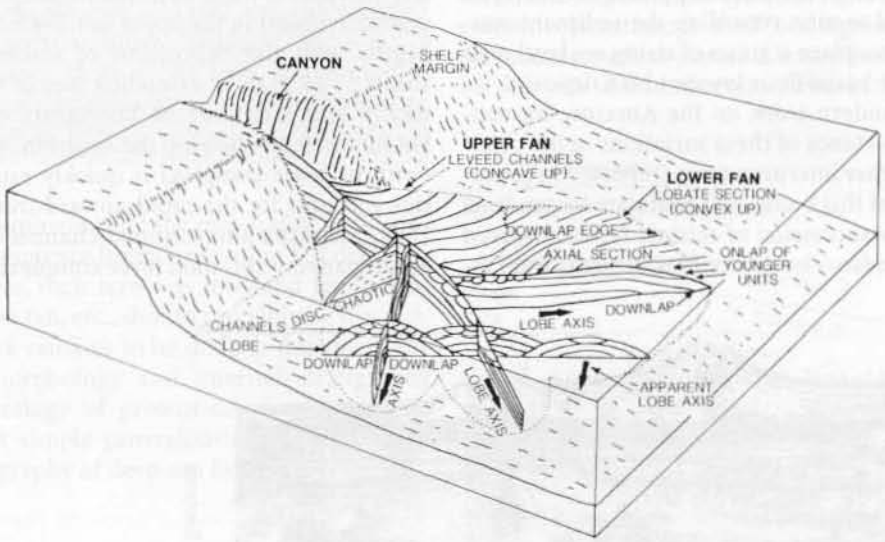


Fig. 6.38. Seismic-stratigraphic model of submarine fans. Downlapping lower fans are interpreted, in this model, as deposits of the lowest stand of sea level (type I of Mutti 1985). Channel-levee complexes of the upper fan were thought to be the deposits that occurred during times of rising sea level (type III of Mutti 1985). (Kolla and Macurda 1988, after Mitchum 1985)

sequent sea-level rise. Fan sedimentation may continue during sea-level highstand if the continental shelf is narrow and if a sediment supply to the continental slope is maintained during sea-level rise (e.g., the Congo and Magdalena fans: Emery and Myers 1996). Gorsline (1996) provided detailed documentation of major turbidite events triggered by seismic events or major floods in the Santa Monica Basin of California that have occurred during the post-glacial sea-level highstand of the last 500 years.

The three-fold subdivision of fan depositional styles defined by Mutti (1985) may not apply to the evolution of very large fans, such as the Indus, Bengal, and Amazon fans, where the depositional slope and sediment supply are less sensitive to sea-level change than in the small fans that formed the basis of Mutti's classification. Underhill (1991) pointed out the importance of local tectonism in controlling depositional architecture. He related growth patterns in Jurassic fan deposits in the Moray Firth area of the North Sea basin to tilting of the basin floor by rotational block faulting. He argued that Vail and Todd (1981) and Vail et al. (1984) had missed this relationship and incorrectly attributed downlapping architecture to lowstand progradation.

The control of sediment supply by volcanism is also an important determining factor in controlling the timing and facies architecture of turbidite sys-

tems. Winsemann and Seyfried (1991) quoted several cases in the arc-related basins of Central America, where high sediment influxes overcame the effects of rising sea level to form major turbidite systems.

Walker (1992), in a review of deep-marine depositional processes, commented on the systems-tract nomenclature of the Exxon school. He noted that many of the relationships shown in the standard Exxon diagrams are not demonstrated by any actual examples of turbidite systems known to him. For example, he could not cite any example where the "lowstand wedge" was buried by highstand deposits, nor examples where the lowstand wedge buries the "slope fan". The latter consists of channel–levee complexes and does not, in fact, occur on the slope. He pointed out that, where data have become available from modern fans such as the Mississippi fan, the architecture is quite different. In the Mississippi fan, for example, the condensed section recording the sediment starvation that takes place at times of rising sea level rests directly on the basin-floor lowstand fan deposits.

Detailed modern work on the Amazon fan confirms the importance of these variations in the original model. Earlier interpretations (reported in the second edition of this book) had postulated a sea-level control for the succession of seismic facies observed in the Amazon fan. The debris-flow deposits were interpreted as the products of large-scale mass movement triggered by the movement of increased volumes of sediment over the shelf–slope break during times of falling sea level, with subsequent upward fining into the channel–levee complex corresponding to rising sea level. This would have been in accordance with the original Exxon models. However, detailed investigations of the fan on leg 155 of the Ocean Drilling Program (Flood et al. 1995b) concluded that "no clear relation between debris-flow activity and sea level has yet emerged."

A distinctive seismic facies termed "high-amplitude reflection packets" (HARP) was defined in the Amazon fan sediments (Fig. 6.39). Commonly, this facies underlies channel–levee complexes. Detailed studies of these deposits by Hiscott et al. (1997), based on coring and petrophysical logging of leg-155 holes, interpreted these deposits as the products of repeated avulsion in the upper fan. Each avulsive event begins with the deposition of stacked sandstone sheets 5–25 m thick extending tens of kilometers laterally and hundreds of kilometers downfan (the HARP facies). Following the avulsion, a new channel (with its levee complex) is quickly established, and this accounts for the rapid upward transition of the HARP deposits into confined-channel deposits with accompanying silt–mud levee complexes.

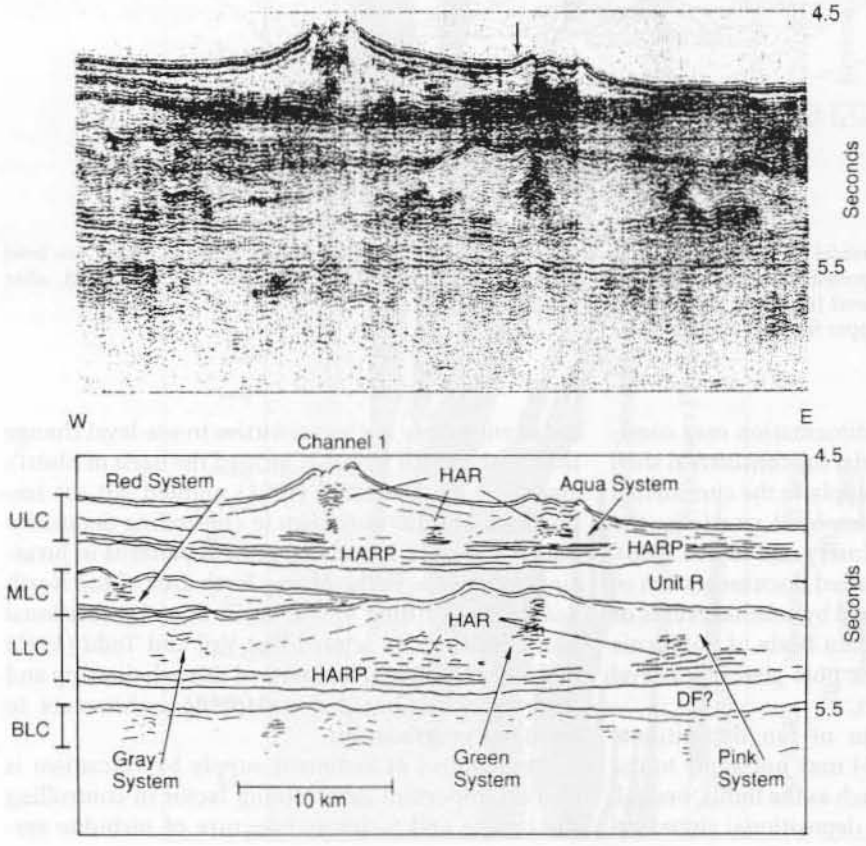


Fig. 6.39. Seismic section and interpretation of a section across the modern Amazon fan, modified from the original interpretation of Manley and Flood (1988). *HAR*[*P*] High-amplitude reflection [packet]; *LC* levee complex; prefix *U*, *M*, *L*, *B* upper, middle, lower, bottom; *DF* debris flow. *Unit R* is also a debris flow deposit. (Flood et al. 1995a)

Each major levee complex of the fan may correspond to a glacial stage (Flood et al. 1995b), but the complexes themselves are composed of fining-upward HARP–levee successions that represent avulsive events occurring every 5–10 k.y. Flood et al. (1995a) concluded:

> Thus the relationship of fan architecture to sea level-change (Posamentier et al. 1988; Mutti 1992) remains rather speculative. It is uncertain whether sequence boundaries can be traced off the shelf onto submarine fans at the base of channel-levee complexes, as proposed by Weimer (1989) for the Mississippi fan. On the Amazon fan and some other modern fans (e.g., Mississippi, Indus), latest Pleistocene sediment in piston cores indicates that the sandy lower-fan subenvironment ("basin-floor fan" of Van Wagoner et al. 1988) and the muddy middle- to upper-fan subenvironment ("slope fan" of Van Wagoner et al. 1988) form contemporaneously, at least throughout a period of rising sea level. (Flood et al. 1995a)

Given these comments and the reappraisal of the position of the sequence boundary within fan deposits, as noted above, such terms as lowstand fan, basin-floor fan, slope fan, etc., should probably be discarded. Much work remains to be done to document the surface geomorphology and internal stratigraphy and sedimentology of present-day fans before we should accept simple generalizations about the sequence stratigraphy of deep-sea fans.

6.6 Systems Tracts and Sequences in Carbonate Deposits

The sequence stratigraphy of carbonate systems is considered in detail by Tucker and Wright (1990), Schlager (1992), Handford and Loucks (1993), and Emery and Myers (1996). Useful summaries are provided in Walker and James (1992).

Carbonate sedimentation depends on a wide variety of depositional controls (Sect. 4.6.8). Factors such as temperature, carbonate saturation, salinity, water depth, nature of water currents, light penetration, water turbidity, nature of sediment substrate, and rates of relative sea-level change all determine sedimentation patterns, and these vary widely and, in part, independently. Subsidence rates and water-circulation patterns depend strongly on the structural setting and, in turn, control the thickness, configuration, and content of carbonate-depositional systems.

Under optimum conditions, calcium-carbonate accumulation can be extremely rapid (exceeding 1 m/ka, according to Wilson and D'Argenio 1982). Such conditions include warm, clear, shallow water of normal marine salinity. Unless subsidence is equally rapid and keeps pace with sedimentation, a carbonate platform will rapidly aggrade to sea level, and additional carbonate material may then be exported seaward across the continental slope. Because of organic binding and submarine cementation, carbonates can build steep depositional slopes, and this allows them to prograde seaward rapidly. Continental margins may undergo drowning at times of rapid sea-level change, and carbonate sedimentation may also be terminated by rapid changes in water temperature, suspended sediment, or nutrient concentrations, and this will also affect the overall configuration of the carbonate-depositional systems. Figure 6.40 shows a common situation: a shelf undergoing differential subsidence, such as occurs at a continental margin or on one side of an intracratonic basin. Figure 6.41 il-

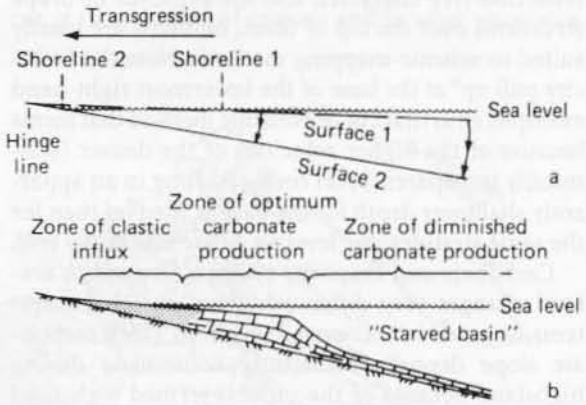


Fig. 6.40. Generation of shelf-to-basin topography by hinging shelf margin. The lower diagram shows the generation of a thick shelf sequence in the zone of optimum carbonate production. (Meissner 1972)

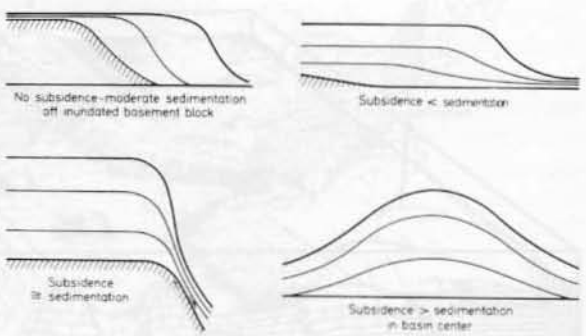


Fig. 6.41. Stratigraphic profiles in carbonate-depositional systems reflecting variations in rates of subsidence and sedimentation. (Wilson 1975)

lustrates variations in these growth patterns, depending on relative rates of subsidence and cementation. They are very reminiscent of the seismic clinoform patterns discussed in Chapter 5 (Fig. 5.19).

Whereas the analysis of clastic systems revolves around the analysis of detrital depocenters, the crucial component of many carbonate systems is the **carbonate buildup**. Wilson (1975) defined the latter as "a body of locally formed (laterally restricted) carbonate sediment which possesses topographic relief. This is a general and useful term because it carries no inference about internal composition." Buildups include offshore banks or barriers and shelf-margin buildups, which may or may not be predominantly organic in origin. Also included are the various kinds of organically constructed wave-resistant features termed reefs. Figure 6.42 illustrates some of the carbonate buildups most easily recognized from seismic interpretation, and Fig. 6.43 shows some typical seismic facies patterns that develop in and around buildups. Because of their usually strong velocity contrast with enclosing sediment, their internally massive and reflection-free character, and the presence of drape structures over the top of them, buildups are ideally suited to seismic-mapping methods. Note the "velocity pull up" at the base of the lowermost right-hand example, an artifact of the seismic method that forms because of the higher velocities of the denser (seismically transparent) reef rock, resulting in an apparently shallower depth for the base of the reef than for the same stratigraphic level on either side of the reef.

Carbonate and evaporite systems respond to sea-level changes very differently than siliciclastic systems, as noted in Section 6.3 (Fig. 6.9). Thick carbonate slope deposits commonly accumulate during highstand because of the process termed highstand shedding. The accommodation on a carbonate platform can be rapidly filled by a "carbonate factory"

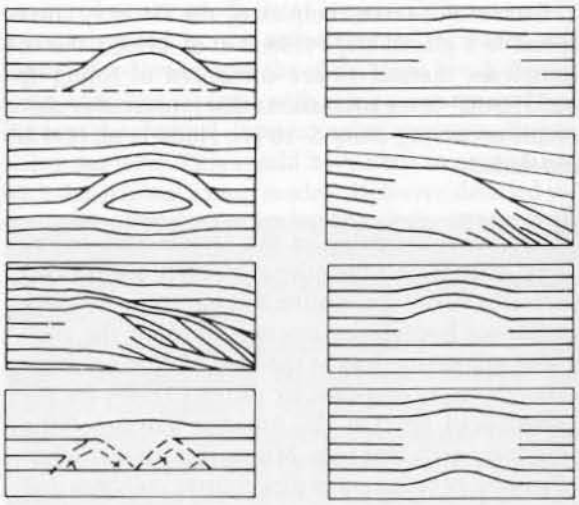


Fig. 6.43. Typical seismic reflection patterns of carbonate buildups. (Bubb and Hatlelid 1977)

operating at full speed at highstand, and surplus sediment feeds the clinoforms of a prograding margin. This is illustrated below with reference to the Alberta Basin and the Bahama Platform. During lowstand, platform exposure may lead to erosion and karstification, with carbonate production confined to a narrow margin. This contrasts with the thick accumulations of clastic slope and basin deposits that are funneled onto the slope by submarine canyons.

There are, in the broadest sense, four types of carbonate platform: epeiric platforms, shelves, isolated platforms, and ramps (Wright and Burchette 1996, based on Read 1985). There are some variations on this classification. Thus, Emery and Myers (1996), following Tucker and Wright (1990), classified carbonate systems as follows: ramps, rimmed shelves, escarpment margins, and detached platforms. How-

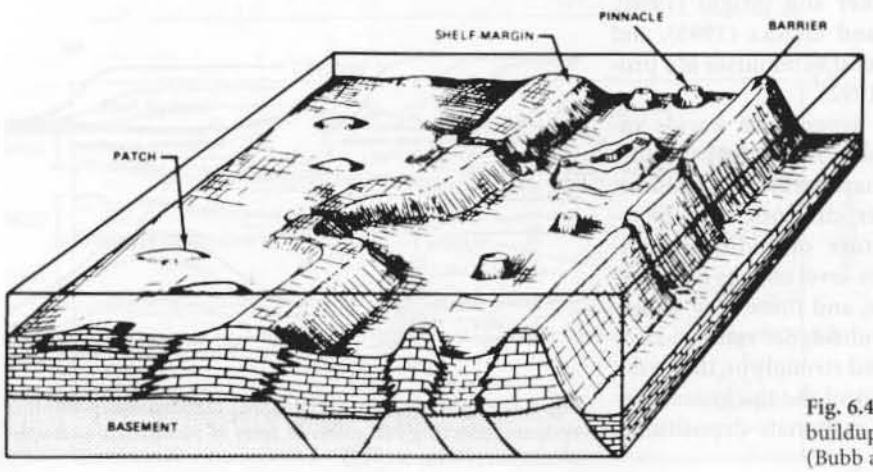


Fig. 6.42. Principal types of carbonate buildup recognizable from seismic data. (Bubb and Hatlelid 1977)

ever, "escarpment margins can be thought of as rimmed shelves without a significant accretionary slope facies" (Emery and Myers 1996) and are, therefore, variants of the rimmed-shelf class. Read (1985) subdivided these classes further. The various ramp and rimmed-shelf carbonate-depositional systems described here can evolve from one type to another, as a result of tectonism (faulting, crustal flexure), sea-level rise, or clastic influx. Kendall and Schlager (1981) and Read (1985) provided detailed discussions.

6.6.1 Epeiric Platforms

There are no modern examples of the vast platform carbonate blankets that developed in the geologic past during times of high global sea-level, when large areas of the world's continents were covered by epeiric seas. Good examples are the Ordovician strata of North America (Fig. 6.44, after Clark and Stearn 1968; Sanford and Norris 1973), the Cretaceous Chalk of Western Europe, and Permian to Tertiary carbonate successions of the Arabian craton south of the Arabian Gulf (Murris 1980; Fig. 6.45). Lithostratigraphic correlations in some units can be traced for hundreds of kilometers and, in the case of the North American interior, formed the basis for the early recognition of regional stratigraphic sequences, as discussed in detail in Chapter 8.

Detailed study of part of the Lower Ordovician succession in Newfoundland by Pratt and James (1986) led them to disagree with prevailing views (Shaw 1964) that these platform-interior deposits are a product of deposition on broad tidal flats within an essentially tideless sea. Pratt and James (1986) developed a depositional model consisting of a mosaic of subtidal to supratidal environments, in which thin and laterally-impersistent shoaling-up cyclic sequences were deposited. They suggested that the epeiric sea was filled with small islands and characterized by tides with a tidal range of at least a few meters. They proposed that the cycles were localized and autogenic in origin (Fig. 4.66).

6.6.2 Rimmed Shelves

Particularly thick platform and buildup deposits occur at the margins of cratons and in what were once

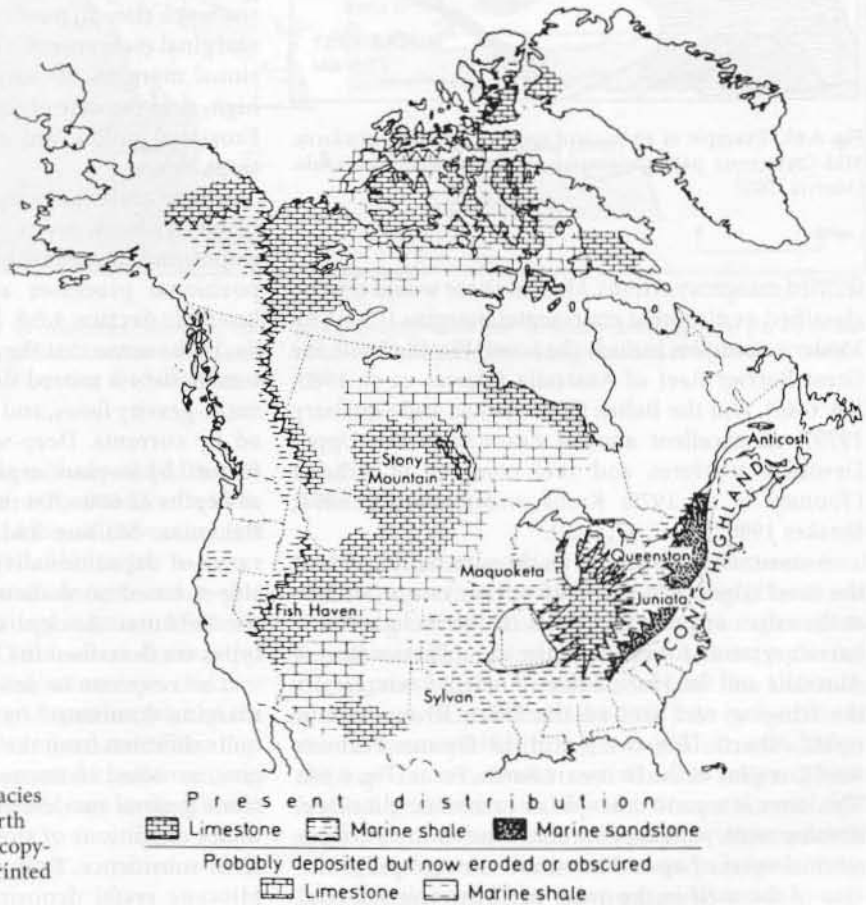


Fig. 6.44. Distribution and lithofacies of Upper Ordovician strata in North America (Clark and Stearn 1968; copyright John Wiley & Sons Ltd., reprinted with permission)

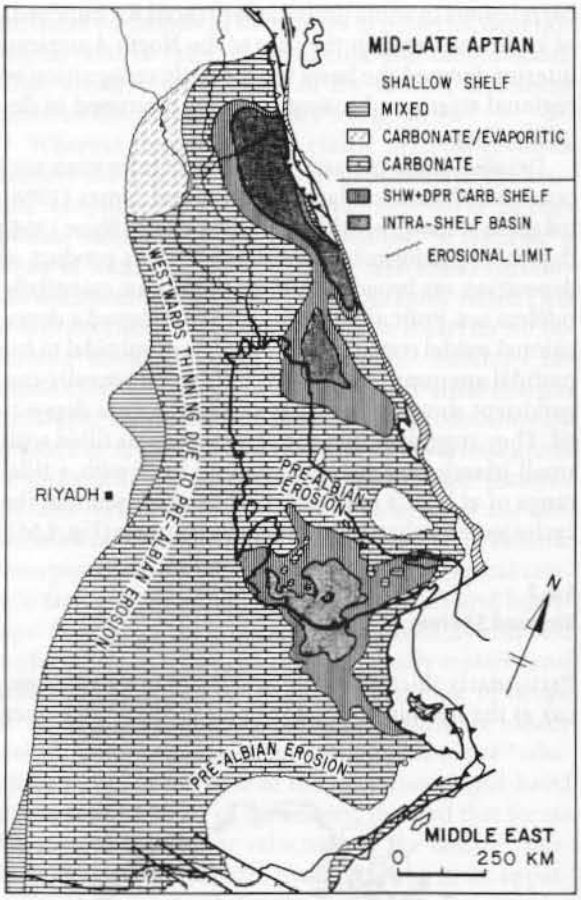


Fig. 6.45. Example of an ancient epeiric carbonate platform. Mid-Cretaceous paleogeography of the Arabian Peninsula. (Murris 1980)

termed miogeosynclines. Most of these would now be classified as divergent continental margins (Chap. 9). Modern examples include the South Florida Shelf, the Great Barrier Reef of Australia (Davies et al. 1989; Fig. 6.46), and the Belize Shelf (James and Ginsburg 1979); an excellent ancient example is the Upper Devonian platform and reef complex of Alberta (Toomey et al. 1970; Krebs and Mountjoy 1972; Stoakes 1980; Figs. 6.47, 6.48).

A common location for carbonate buildups is at the outer edges of major platforms or ramps, such as at the edges of cratonic blocks. These include linear barrier systems, for example, the Great Barrier Reef of Australia and the Florida Keys. Ancient examples are the fringing reef around the Peace River cratonic uplift, Alberta (Fig. 6.47), and the famous Permian Reef Complex of the Delaware Basin, Texas (Fig. 6.49). The latter is a good example of carbonate clinoform development, with depositional dips up to 30°, depositional relief of up to 700 m, and seaward progradation of the shelf in the order of 20 km (Newell et al. 1953; Wilson 1975). Small carbonate buildups occur within platforms or cratonic basins in which moderate subsidence takes place. There are very few modern examples of this setting, except perhaps parts of the Arabian Gulf (Purser 1973). However, ancient examples are numerous and include Silurian and Devonian patch reefs and Mississippian Waulsortian mud mounds (Wilson 1975).

Platform margins in carbonate systems are of three broad types (Schlager and Ginsburg 1981; Read 1985; Tucker and Wright 1990): erosional, by-pass, and accretionary (Fig. 6.50). Wilson (1975) subdivided accretionary margins into three types. Variation between them probably depends on wave and current energy, nutrient supply, and tectonics. Type III is favored by initially steep (faulted) margins, rapid sedimentation, and upwelling currents bringing in abundant nutrients for organic framework growth. Type II reflects lesser wave or current energy, an environment less favorable for framework building. Water movement is inadequate to remove much of the organic debris, which accumulates as lime sand. Type I represents the lowest-energy environment and includes significant accumulations of lime mud. "Bypass margins occur in areas of rapid upbuilding where shallow-water sedimentation keeps pace with sea-level rise. Bypassing may be associated with a marginal escarpment" (Read 1985). In the case of erosional margins, the escarpment may be up to 4 km high, as in the case of the Blake-Bahama Escarpment. Erosional gullies feed abundant detritus to the talus slope below.

Major carbonate-depositional systems hundreds of meters thick occur in deep water, on continental slopes, and in oceanic basins. Some details of the depositional processes and products are described briefly in Section 4.6.8. Most of the deposits are clastic, in the sense that they consist of shelf-derived carbonate debris moved downslope by slumps or sediment-gravity flows, and they are commonly winnowed by currents. Deep-water bioherms (lithoherms) formed by in-place organic growth and cementation at depths of 600–700 m also occur on slopes off the Bahamas. Mullins and Neuman (1979) erected a range of depositional-systems models for carbonate slopes, based on shallow seismic and coring work off the Bahamas. Ancient examples of several of these types are described in Cook and Enos (1977).

The response to sea-level change of continental margins dominated by carbonate sedimentation is quite different from the response of siliciclastic margins, as noted in Section 6.3. Figure 6.51 illustrates some general models of sequence stacking patterns under conditions of slow, moderate, and rapid long-term subsidence. Examples of these are seen in the Miocene reefal deposits of Mallorca (Fig. 6.52). A

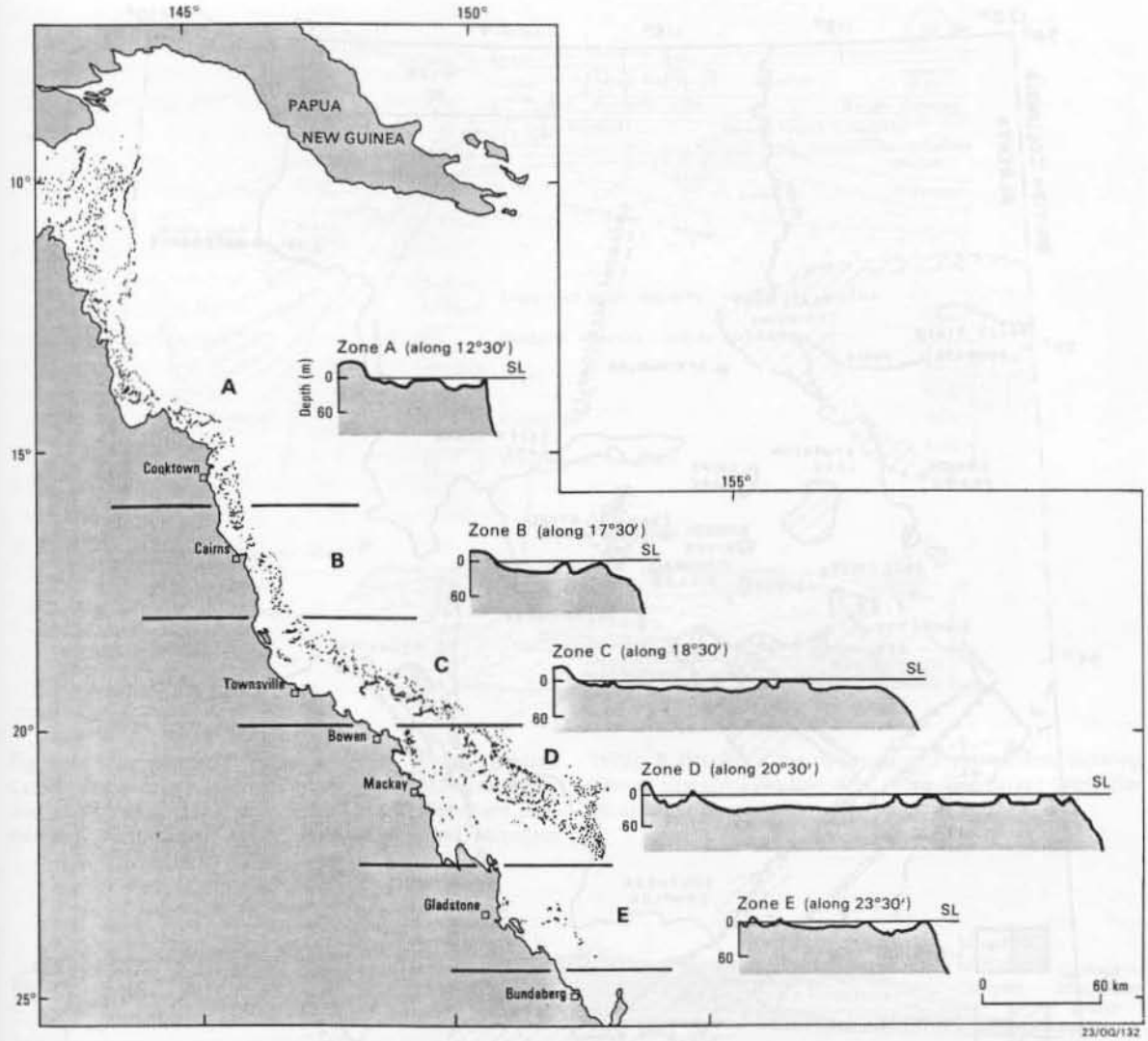


Fig. 6.46. The northeast margin of Australia, showing the location of the patch reefs that together constitute the Great Barrier Reef, and cross sections illustrating the variations in physiography. Note the variation from rimmed (*A, D*) to unrimmed (*C, E*) platform configurations. (Davies et al. 1989)

complete 3-m.y. sea-level cycle is illustrated in this interpretation. Another style of sedimentation is illustrated by the details of the Alberta Devonian reef complexes shown in Fig. 6.47B. The Leduc reefs are interpreted as a transgressive systems tract that developed at a time of vertical aggradation. The shelf margin backstepped tens of kilometers during this phase of sea-level rise. Slope sediments from this new margin constitute the highstand systems tract of the Ireton Formation, Camrose Member, and Nisku Formation. The Ireton is a fine-grained clastic unit and acts as the cap rock for many reef traps in the Alberta Basin.

6.6.3 Isolated Platforms

Isolated platforms occur above faulted continental remnants, such as the Bahamas Platform (Eberli and Ginsburg 1989), and above volcanic seamounts (oceanic atolls). Spectacularly exposed examples occur in the Middle Triassic rocks of the Dolomite Mountains, Italy (Blendinger 1986). As in the case of the Bahamas platform, these show different characteristics on windward and leeward sides. Cross-bedded grainstone shoals and discontinuous algal reefs occur on the windward margins, while subtidal sand-flat deposits occur on leeward margins. Talus breccias surround the platforms and exhibit well-

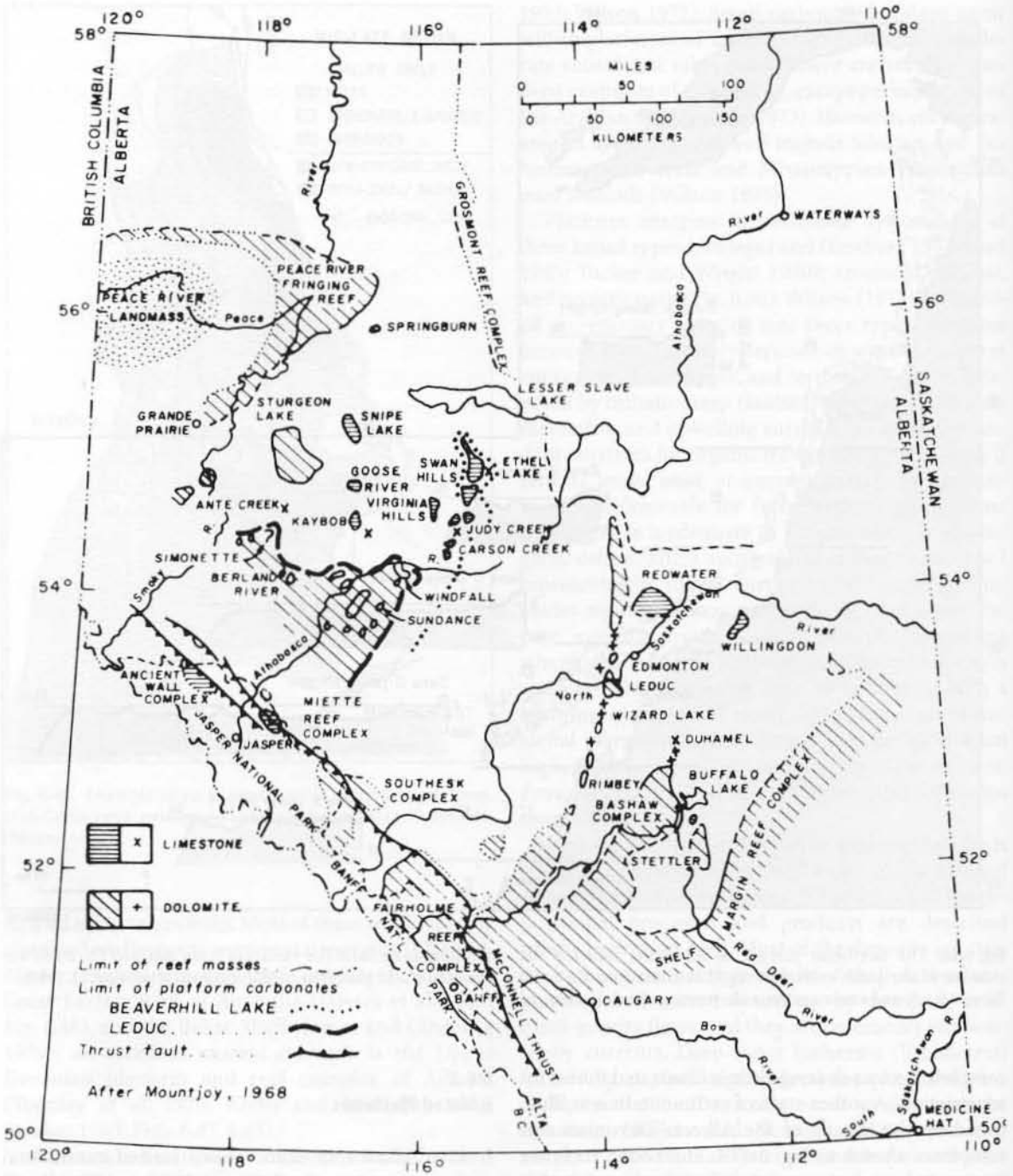


Fig. 6.47. Distribution of Upper Devonian carbonate buildups in Alberta. (Toomey et al. 1970)

exposed clinoform bedding with basinward dips of 30–40° (Fig. 6.53). At the base of the slope, these interfinger with basinal radiolarian micrites. On windward margins, these are interbedded with sediment gravity-flow deposits. The platforms remained stationary and accreted vertically because of syndepositional faulting.

This contrasts markedly with the Bahamas platform, which, though originally controlled by faults, expanded laterally toward the Florida margin by spectacular lateral clinoform progradation (Figs. 6.54, 6.55). Windward margins are steep and show chaotic seismic facies. Leeward margins evolved from steep, fault-bounded margins into bypass margins

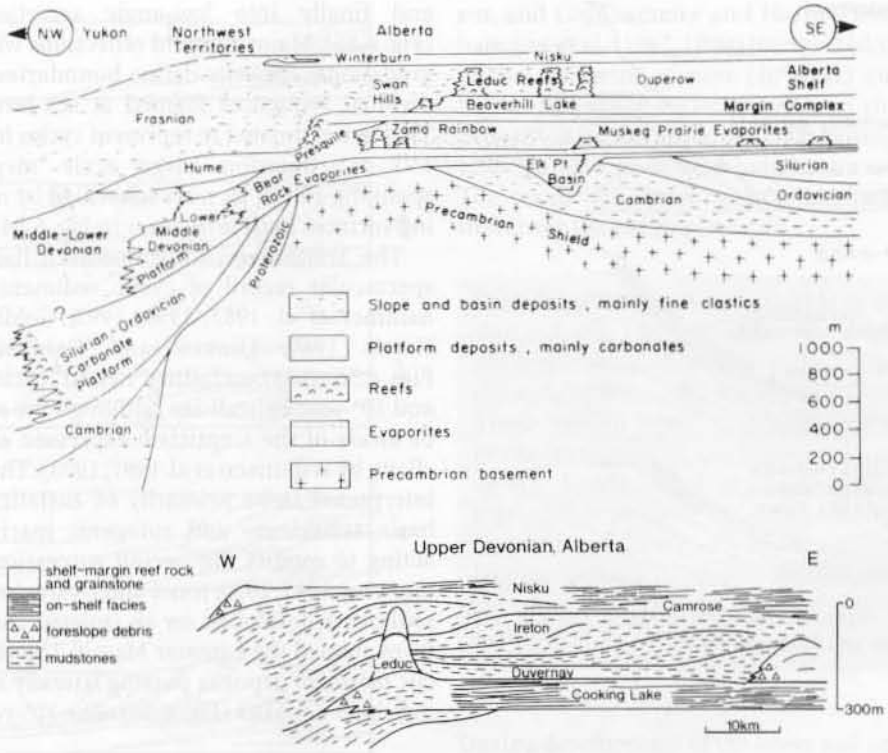


Fig. 6.48. Diagrammatic cross sections across Western Canada Sedimentary Basin, illustrating the general distribution of Devonian carbonate complexes and carbonate platforms. **A** Generalized cross section (Krebs and Mountjoy 1972). **B** Details of the upper part of the section, showing lithostratigraphic relationships. (Hunt and Tucker 1993, after Stoakes 1980)

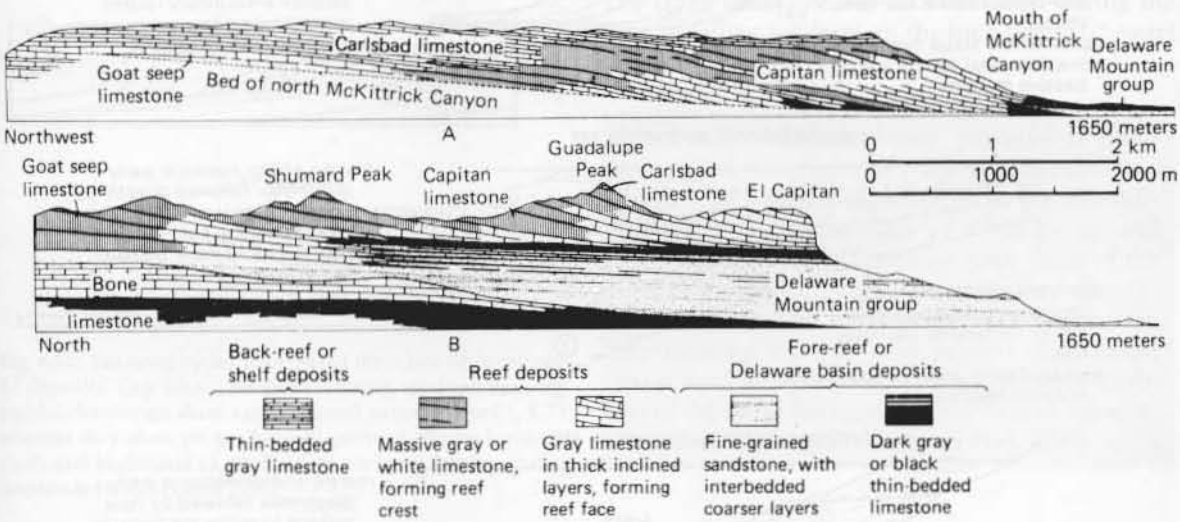


Fig. 6.49. Two sections through the Permian reef complex of the Guadalupe Mountains, west Texas. (King 1948)

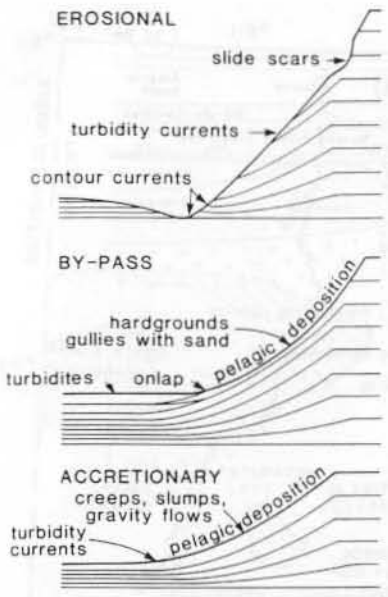


Fig. 6.50. Three types of platform margin. (Tucker and Wright 1990; Hunt and Tucker 1993)

and finally into low-angle accretionary slopes (Fig. 6.55). Major sigmoid reflections within the progradational deposits define boundaries between individual sequences formed at sea-level lowstands. They are estimated to represent cycles in the order of 1–2 m.y duration. Larger scale "megasequences" spanning 10–20 m.y. are identified by major bounding surfaces (*letters in boxes* in Fig. 6.54).

The Triassic record of northern Italy contains a spectacular record of cyclic sedimentation (Goldhammer et al. 1987, 1990, 1993; Goldhammer and Harris 1989; Hinnov and Goldhammer 1991; Figs. 6.56–6.58), including "nested" cycles of 10^4-, 10^5- and 10^6-year episodicity (although the reader should be aware of the scepticism expressed about this cyclicity by Wilkinson et al. 1997, 1998). These cycles are interpreted to be primarily of eustatic origin, with basin subsidence and autogenic marine processes acting to modify the overall succession (Goldhammer et al. 1987, 1990; Jones and Desrochers 1992). The sediments developed on an isolated carbonate platform, named the Latemar Massif. They consist of cyclic platform deposits passing laterally into prograding slope deposits (Fig. 6.56). The 10^4-year cycles re-

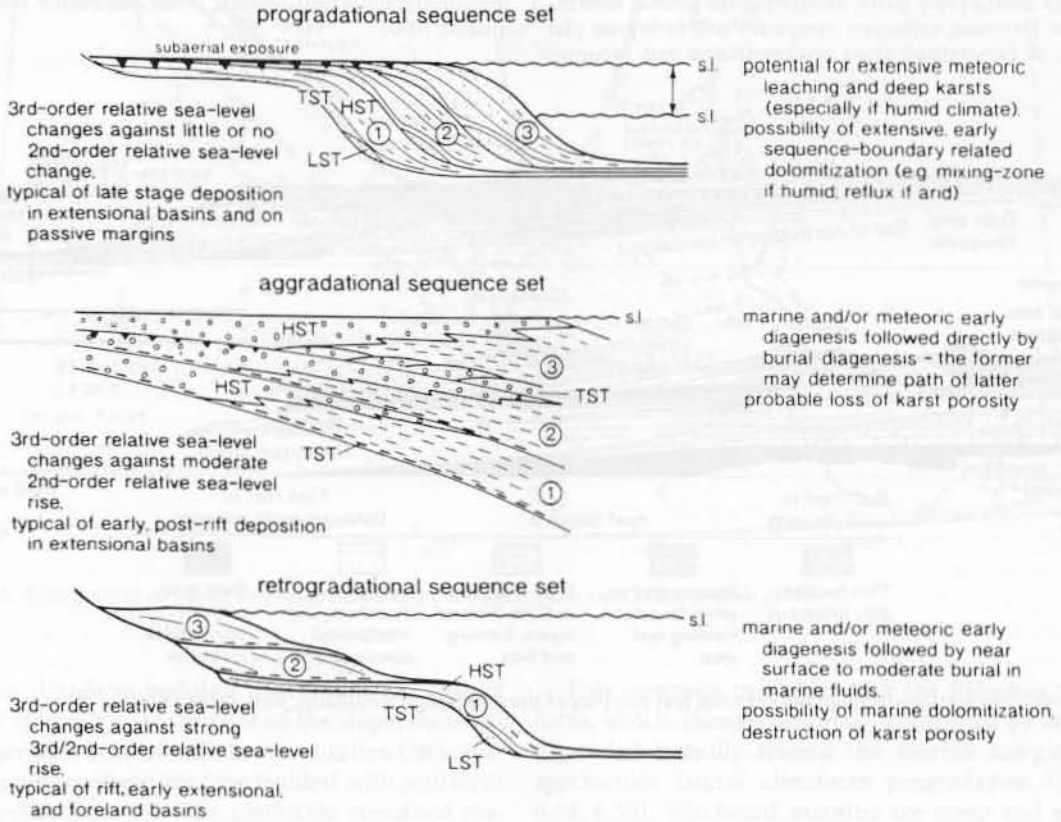


Fig. 6.51. Sequence-stacking patterns on carbonate platforms under varying conditions of long-term increases in accommodation space. (Tucker 1993)

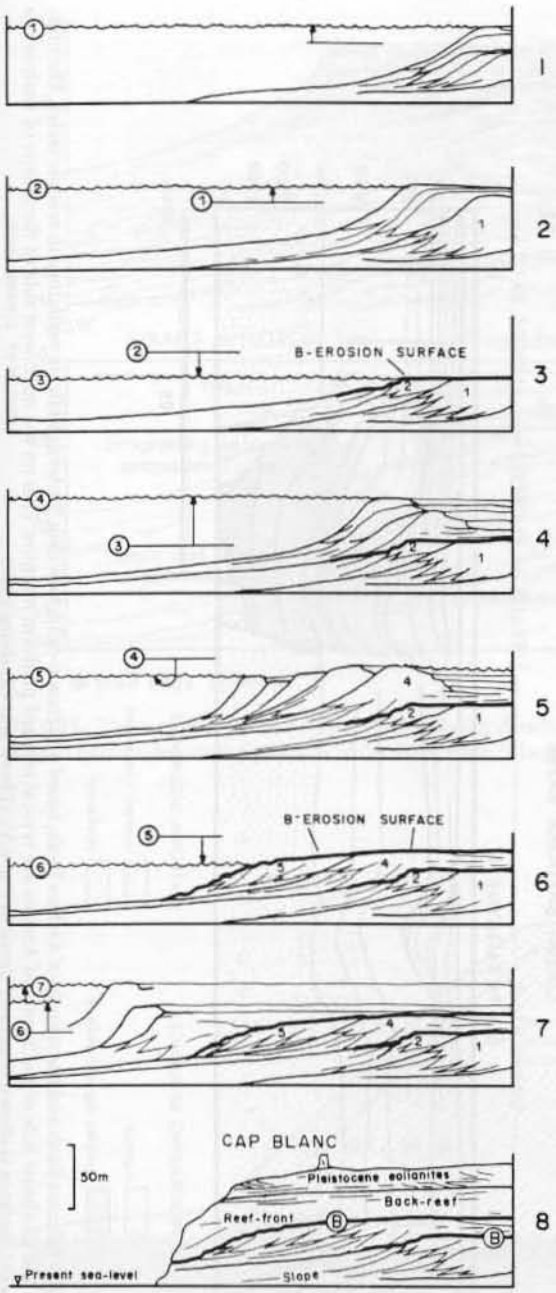


Fig. 6.52. Sea-level cycles interpreted from late Miocene reefal deposits, Cap Blanc, Mallorca. During sea-level rise, sigmoidal clinoforms show aggradational patters (times 1, 4, 7), whereas they show progradational patterns during lowstand (1, 4) and highstand (2, 5). Erosion surfaces develop during lowstands (3, 6). (Pomar 1991)

cord abrupt shallowing and exposure, with the development of subtidal platform deposits that underwent exposure and vadose diagenesis. The cycles in the lower and upper cyclic facies average 0.7 m in thickness. In the lower platform facies and the tepee facies, cycle compositions and thicknesses are different, and Goldhammer and Harris (1989) and Goldhammer et al. (1987, 1990) interpreted this as a result of the overprint of slow (10^6-year) eustasy on the higher-frequency cycles of sea-level change. During deposition of the lower platform facies, the area was undergoing a prolonged long-term rise in sea level. The result was the maintenance of subtidal conditions over the entire platform:

> Subtidal sedimentation is unable to keep up, and thus not every beat of fifth-order high-frequency sea-level can 'touch down' on the platform top and subaerially expose the top of the sediment column. These 'missed beats' of subaerial exposure result in the formation of thick (average approximately 10 m), fourth-order amalgamated megacycles. During this phase of subtidal carbonate aggradation, subaerial exposure occurs infrequently, and lengthy periods of marine submergence promote abundant sydepositional marine diagenesis (Goldhammer et al. 1990).

During development of the lower and upper cyclic facies, the rate of long term (10^6-year or "third-order") sea-level change was slow. Subtidal sedimentation was able to keep pace with net sea-level change, which was almost entirely due to subsidence. Every high-frequency (10^4-year, "fifth-order") oscillation in sea level is, therefore, recorded by subaerial exposure. The tepee facies records sedimentation during the time of falling sea level on the long-term (10^6-year) cycle.

> Third-order accommodation potential is minimized, but in a realm of continuous subsidence, the net effect is close to but not quite that of a still-stand. Now, sedimentation easily keeps up and very thin cycles form (thinner than those of the lower cyclic facies), which are grouped into fourth-order condensed megacycles with tepee tops representing more lengthy periods of subaerial exposure. The tepees form owing to prolonged subaerial exposure during the falling limb of a fourth-order oscillation. (Goldhammer et al. 1990)

6.6.4 Ramps

This distinctive type of depositional system was first recognized by Ahr (1973), and subsequent detailed treatments of this environmental model have been provided by Burchette and Wright (1992), Tucker et al. (1993), and Wright and Burchette (1996). Ramps

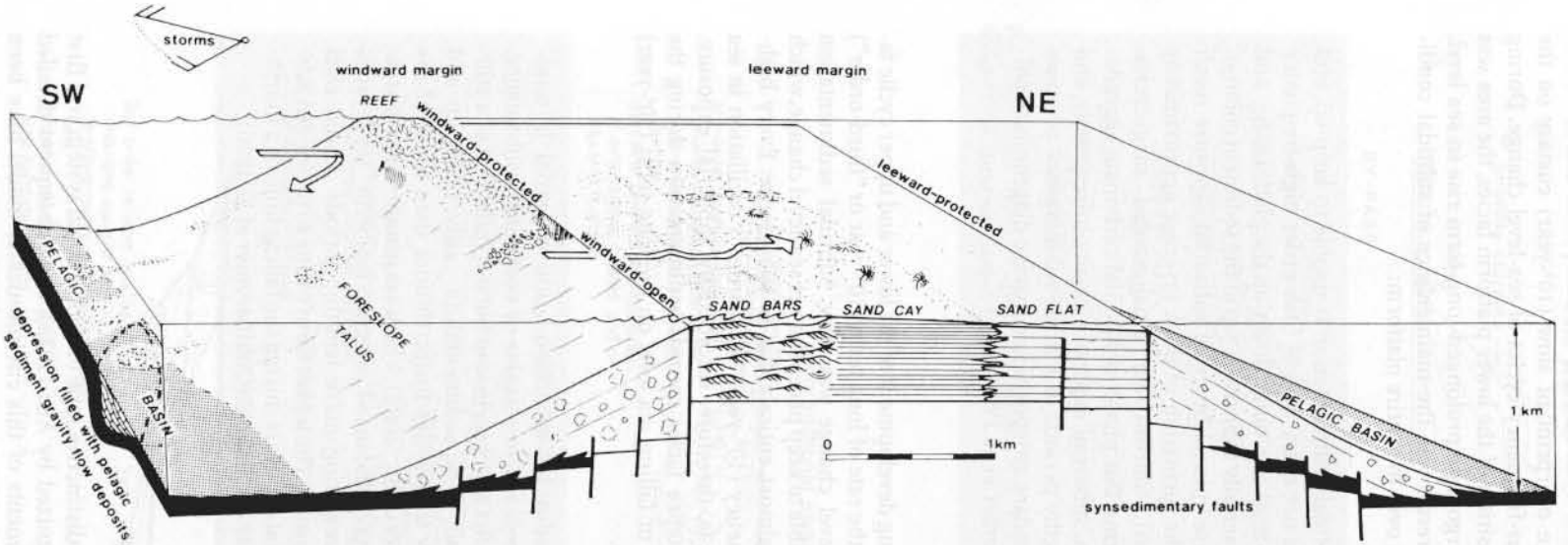


Fig. 6.53. Model of a stationary, isolated, and fault-bounded platform, based on exposures in the Triassic rocks of the Dolomite Mountains, Italy. (Blendinger 1986)

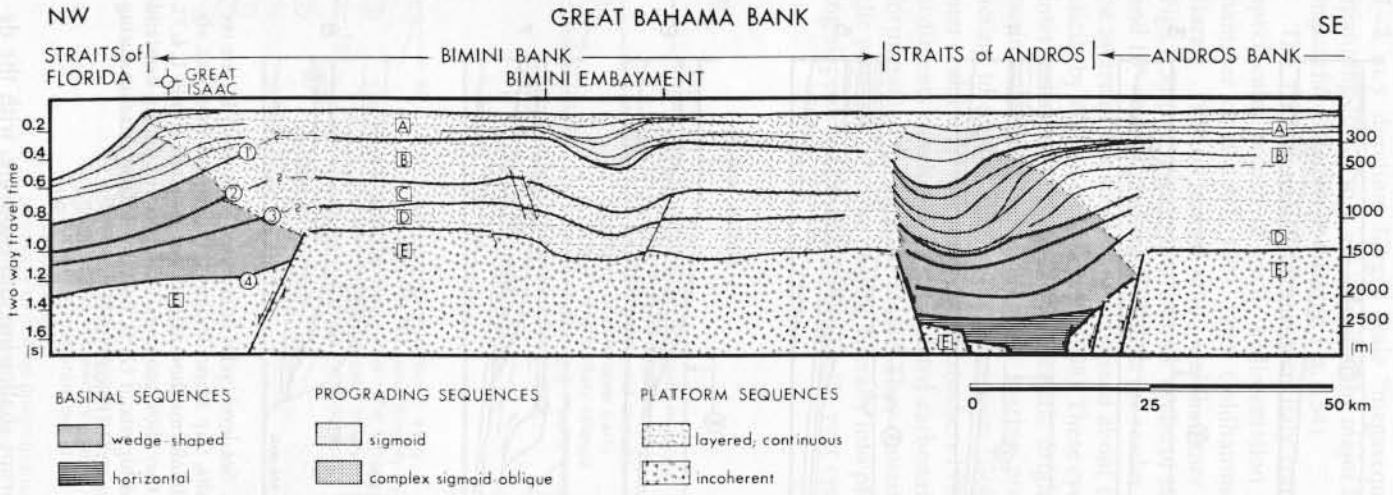


Fig. 6.54. Schematic cross section across Great Bahama Bank and the margin of Andros Bank, based on seismic-reflection data. Patterns indicate variations in seismic facies. The margin facing the Straits of Florida and that on the southeast side of the Straits of Andros are typical leeward platform margins. That on the northwest side of the Straits of Andros is a windward margin. The evolution of the leeward margins is shown in greater detail in Fig. 6.55. (Eberli and Ginsburg 1989)

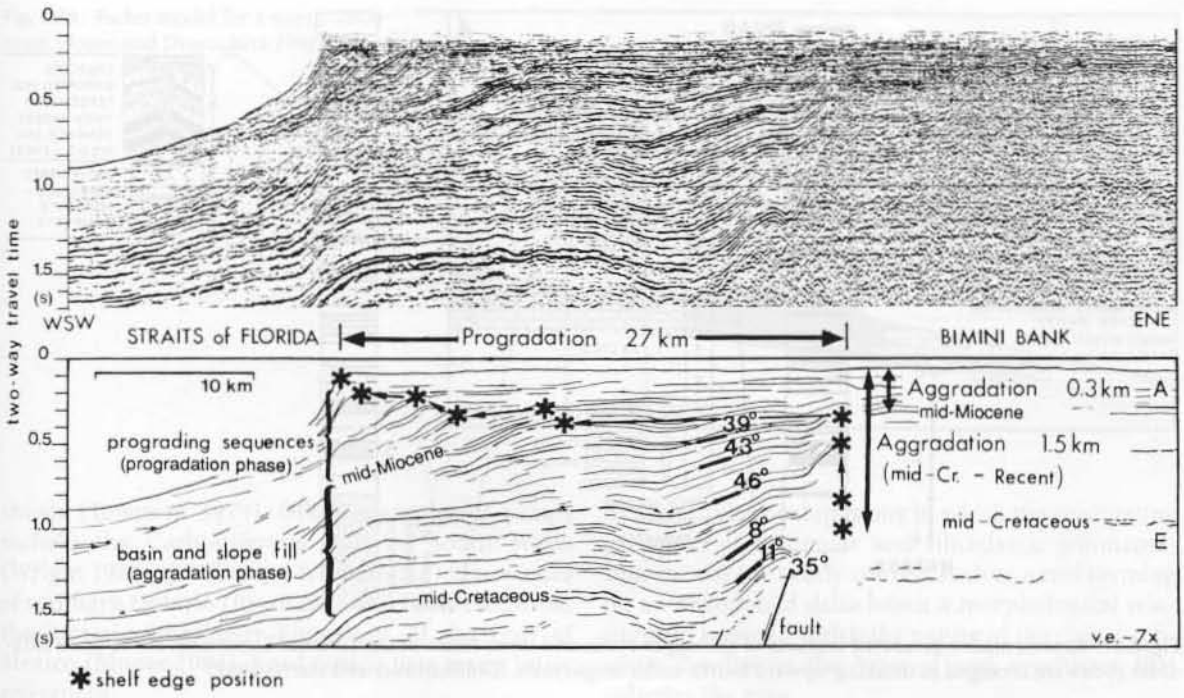


Fig. 6.55. The west margin of Bimini Bank, showing details of margin evolution. *Angles* document the decrease in continental slope. *Asterisks* denote the position of the shelf edge. (Eberli and Ginsburg 1989)

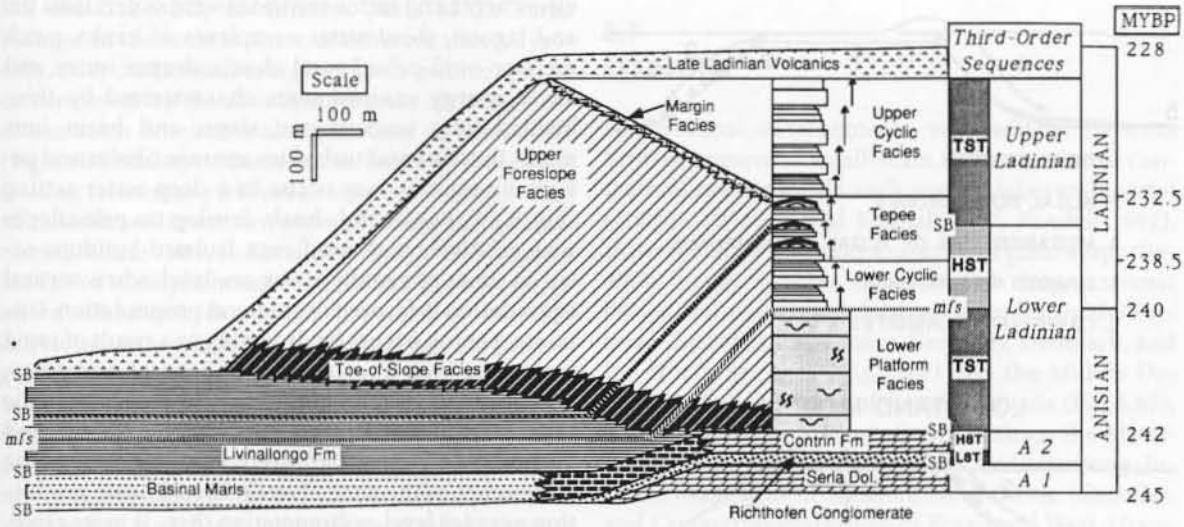


Fig. 6.56. Stratigraphic relations of the Triassic Latemar platform and slope deposits. Dolomite Mountains, northern Italy. (Hinnov and Goldhammer 1991)

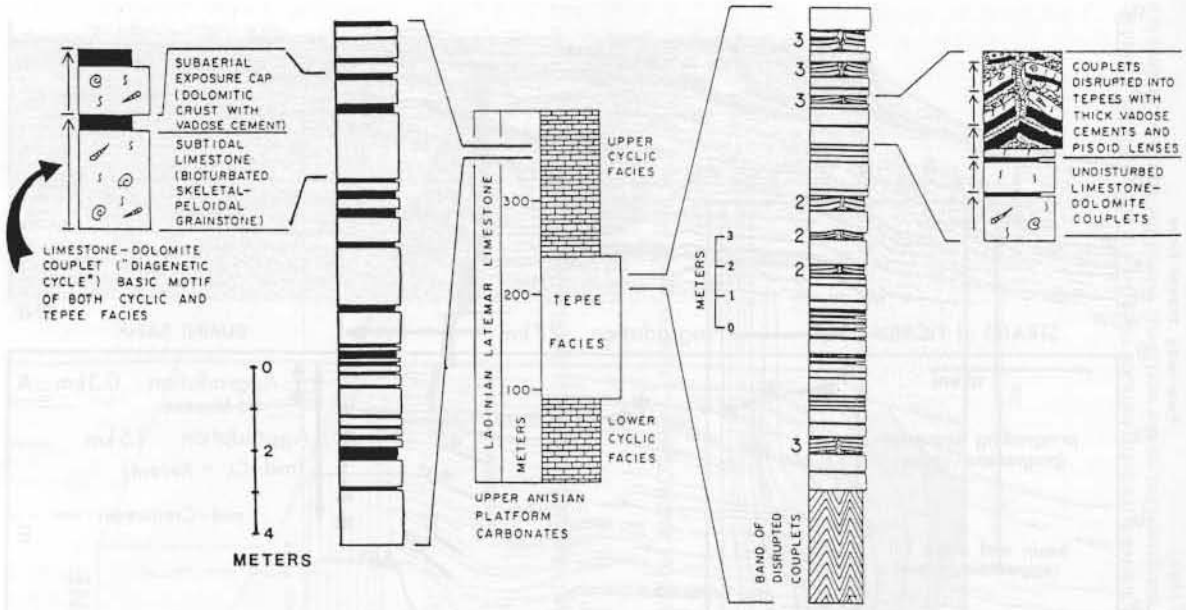


Fig. 6.57. Representative measured sections of the upper cyclic and tepee facies of the Latemar Limestone. Meter-scale fifth-order cycles are arranged in thinning-upward fourth-order megacycles. (Goldhammer and Harris 1989)

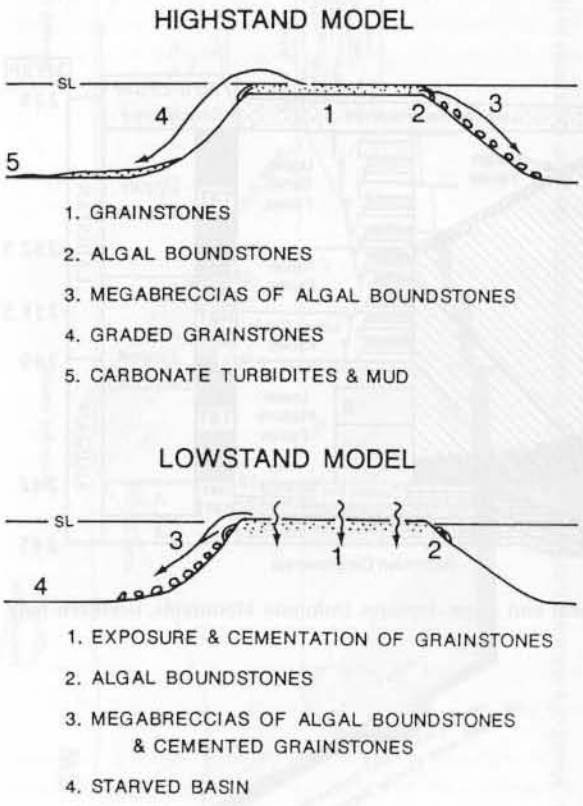


Fig. 6.58. A model for Latemar cycles, showing depositional conditions under highstand and lowstand conditions. (Goldhammer and Harris 1989)

are tabular or wedge-shaped masses built on the flanks of positive areas down regional paleoslopes. They typically have slopes less than 1°. Typical environments and facies are, in seaward order: tidal flat and lagoon, shoal-water complexes of banks, patch reefs or ooid-peloid sand shoals, deeper water, and lower energy environments characterized by thin-bedded lime wackestones, slope, and basin lime muds. Breccias and turbidites are rare. Shales and pelagic limestones may occur in a deep-water setting (Fig. 6.59). Banks and shoals develop on paleoslopes with relatively high gradients. Isolated buildups occur at times of rapidly rising sea level, when vertical aggradation is favored over lateral progradation. Distal steepening occurs by drowning as a result of rapid rise in sea-level or as a result of flexural downwarp. Low-angle, shingled clinoforms are a characteristic seismic expression of ramp margins (Hunt and Tucker 1993; Emery and Myers 1996). Because of the occurrence of the zone of optimal carbonate production near sea level, sedimentation there is more rapid, and there may be a tendency for a ramp to develop a buildup pattern such as that illustrated in Fig. 6.40. Such a buildup may have a virtually flat top and is then termed a carbonate platform or shelf.

The Portland Group (upper Jurassic) of southern Britain is a good example of a carbonate ramp developed at the margin of a cratonic uplift or swell. The succession is unusual in that it contains no significant buildups, only algal (stromatolitic) mounds or ridges in a lagoonal setting, and oolite

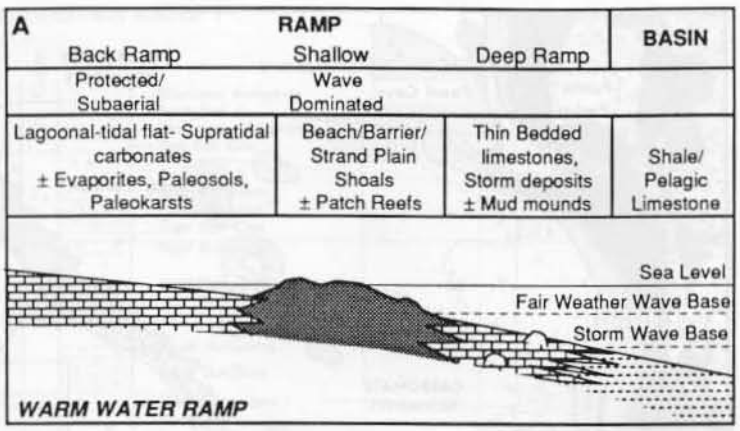


Fig. 6.59. Facies model for a warm-water ramp. (Jones and Desrochers 1992)

shoals (Townson 1975). Other examples of ramps include the Carboniferous rock of South Wales (Wright 1986), the Trenton (Ordovician) Limestones of southern Ontario (Brookfield and Brett 1988), and the Jurassic Smackover Limestone of the Gulf of Mexico (Moore 1984). Read (1985) lists many other references.

6.6.5 Mixed Carbonate-Clastic Systems

Carbonate and clastic sediments commonly accumulate simultaneously in different parts of the same basin or in the same location at different times. Typically, on a continental margin, clastic sedimentation dominates along the coastline, whereas carbonate sedimentation dominates along the mid to outer shelf (Fig. 6.60). Studies of the relations between these contrasting facies may yield much information regarding siliciclastic sediment supply and dispersal mechanisms (Mount 1984; Roberts 1987; Doyle and Roberts 1988). For example, broad zones of mixing may occur where siliciclastic detritus is dispersed by currents driven by trade winds (e.g., the shelf east of Nicaragua) or by monsoons (e.g., the Sunda Shelf of the Java Sea). Sediments of contrasting type may also be interposed by storms and tides. Contrasting with this are areas such as the Gulf of Suez and the Gulf of Elat, where active rifting has led to the development of steep-sided, fault-bounded basins and where marine dispersal processes are weak. Reefs have developed there as fringing facies on the outer margins of small fan deltas, and carbonate–clastic transitions occur over lateral distances of as little as a few tens of meters.

Ancient examples of reefs associated with delta lobes, beaches, deltaic channels, and stream-mouth bars were described by Santisteban and Taberner (1988). They recognized two types of carbonate–clastic relationships: dynamic and morphological. A dynamic relationship is one in which the contrasting processes of carbonate and siliciclastic sedimentation coexist (or nearly coexist, such as a reef forming on an abandoned delta lobe); a morphological relationship is one in which the nature of the clastic substrate conditions the type of reef organisms that colonize the area.

Carbonate and clastic sedimentation may alternate on a continental margin as a result of sea-level change and tectonism (Fig. 6.10). This has been termed reciprocal sedimentation (Sect. 6.3).

6.7 Evaporites

Depositional environments vary widely. Deposits form in numerous small-scale, shallow-water or continental environments, such as playa lakes and coastal sabkhas (Warren and Kendall 1985; Kendall 1992). However, there are many examples of giant evaporite-depositional systems that occur in large cratonic basins and small ocean basins. Examples are the Permian Zechstein of Western Germany, Denmark, and the North Sea area (Fig. 6.61) and the Middle Devonian Elk Point Basin of western Canada (Fig. 6.62), both of which occur in cratonic settings, the Messinian (Miocene) evaporites of the Mediterranean, Jurassic evaporites of offshore Nova Scotia (Canada), and Cretaceous evaporites of Brazil and West Africa, which developed during the early stages of Atlantic sea-floor spreading.

There has been a great deal of debate about the origin of these saline giants regarding whether they represent sabkha evaporites on a large scale or deep-water deposits. Excellent discussions of this problem are given by Schreiber (1986) and Kendall (1992). At present, the evidence seems to point to a similar origin: the shallow-water deep-basin model for most, if not all of these deposits. The model became popular

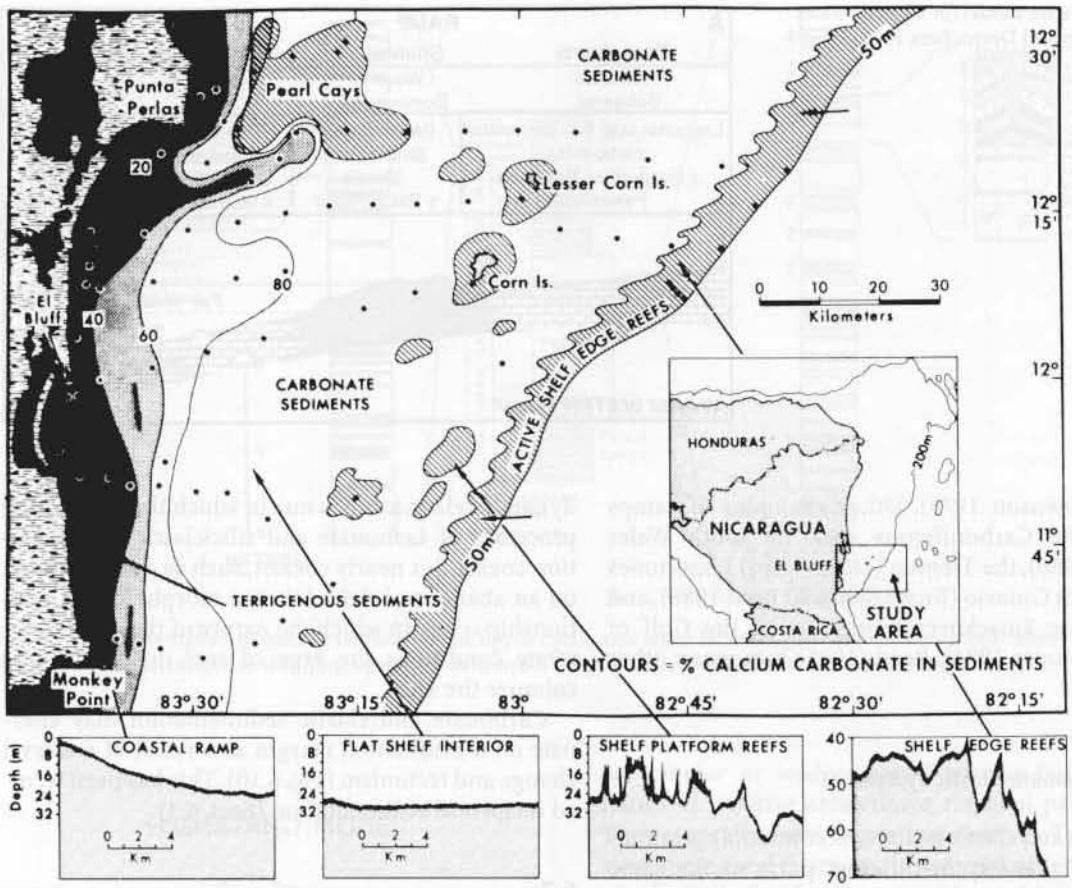


Fig. 6.60. Mixed carbonate–clastic sedimentation on the shelf east of Nicaragua. Contours indicate the percentage of calcium carbonate in bottom sediments. *Hatched areas* indicate mid-shelf and shelf margin reefs. Typical echo-sounder profiles are shown at *bottom*. (Roberts 1987)

following the work of Hsü et al. (1973), who invoked it in order to explain the widespread Messinian (Miocene) evaporites of the Mediterranean Sea basin. The general model is that, initially, deep topographic basins contain deep-water masses of normal salinity. These are cut off from oceanic circulation by elevation of a tectonic land barrier, growth of a reef barrier, eustatic lowering of sea level, or a combination of these processes. Under appropriate climatic conditions, evaporation exceeds riverine and reduced marine input, and the basin becomes completely desiccated. This process explains the presence of both quiet (deep?)-water laminated evaporites, some of which can be correlated over considerable distances (Fig. 3.6), and a range of shallow-water and subaerial, sabkha-like deposits, all in the same basin.

The success of the desiccated-Mediterranean model of evaporite sedimentation has led to a reaction in some quarters, and the suggestion that we should not abandon the examination of alternate models (Dietz and Woodhouse 1988). Indeed, not all scientists are convinced of the necessity of evaporating the Mediterranean to dryness in order to produce the Messinian evaporites. Sonnenfeld (1985) undertook a critical review of all the evidence and concluded that the model is inadequate. He proposed, instead, that the Messinian evaporites were deposited from a hypersaline sea that had restricted circulation as a result of lowered sea-level, but that retained a two-way exchange with the Atlantic Ocean (Sonnenfeld and Finetti 1985).

The presence of reefs and evaporites in the same basin is readily explained by fluctuations in sea level (Mesolella et al. 1974). Reefs grow during periods of deeper water and normal salinity and are exposed to erosion and karstification during phases of evaporative drawdown of the water level. This process may be repeated several or many times as sea levels fluctuate.

The application of sequence concepts to the interpretation of evaporite deposits has helped to clarify the dynamics of evaporite sedimentation and the relationship between carbonate and evaporite-depositional systems. As shown in Fig. 6.9, evaporites typically develop at two stages in a sea-level cycle. At low-

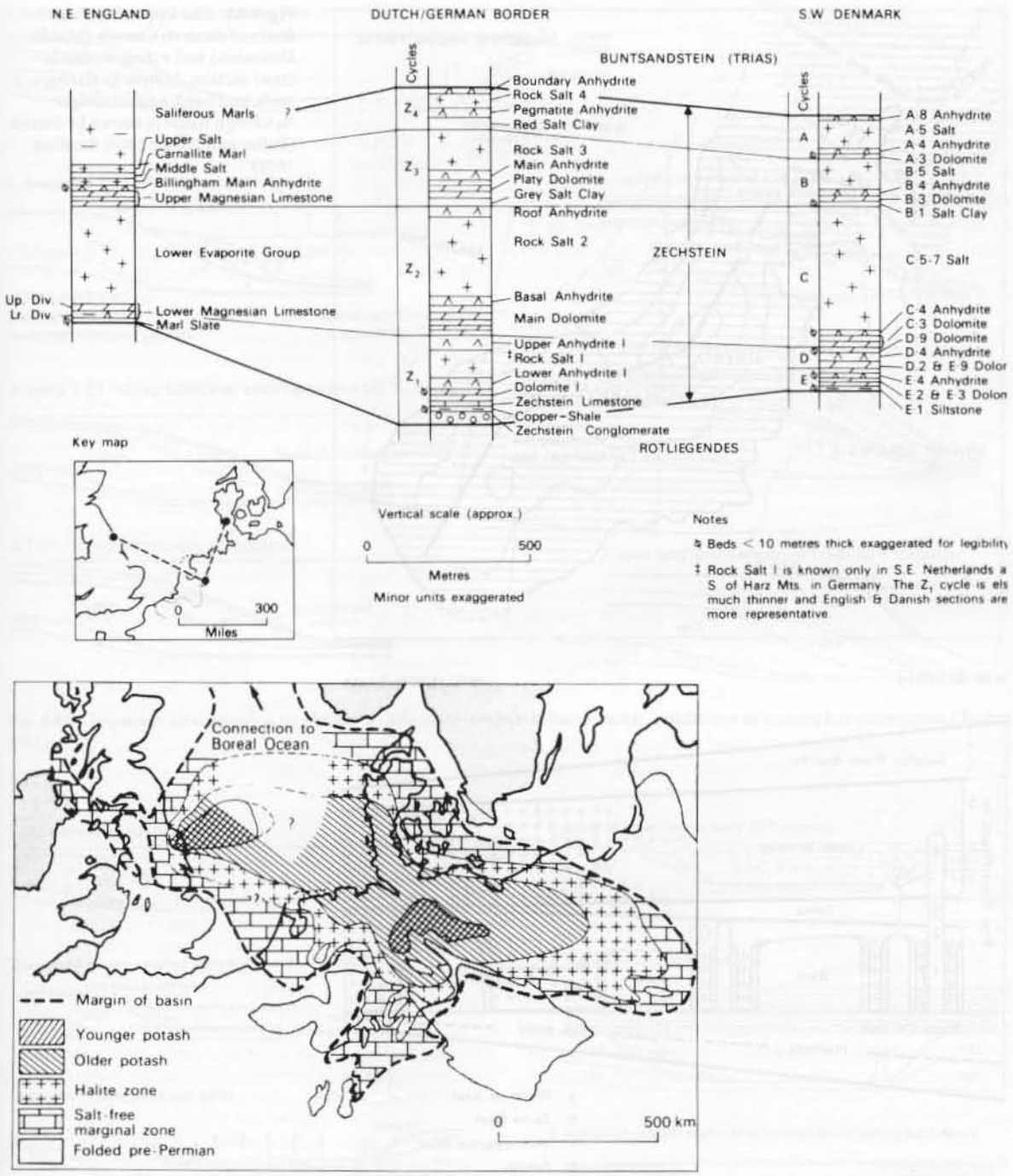


Fig. 6.61. Lithofacies distribution in the Zechstein Basin of western Europe (Upper Permian) and a cross section through the deposits. (Borchert and Muir 1964; Brunstrom and Walmsley 1969; Reading 1978)

stand, when oceanic circulation through an interior basin may be inhibited by basin-margin barriers, such as reef complexes, evaporation may exceed water influx under hot, arid climatic conditions, and widespread evaporites may form at such times. Many of the saline giants seem to have developed under such conditions. Secondly, during transgression, the initial flooding of the continental shelf may give rise to hypersaline conditions in which shallow-water or sabkha evaporites may develop. Tucker (1991) developed these models further and applied them to the Zechstein (Permian) evaporite–carbonate succession

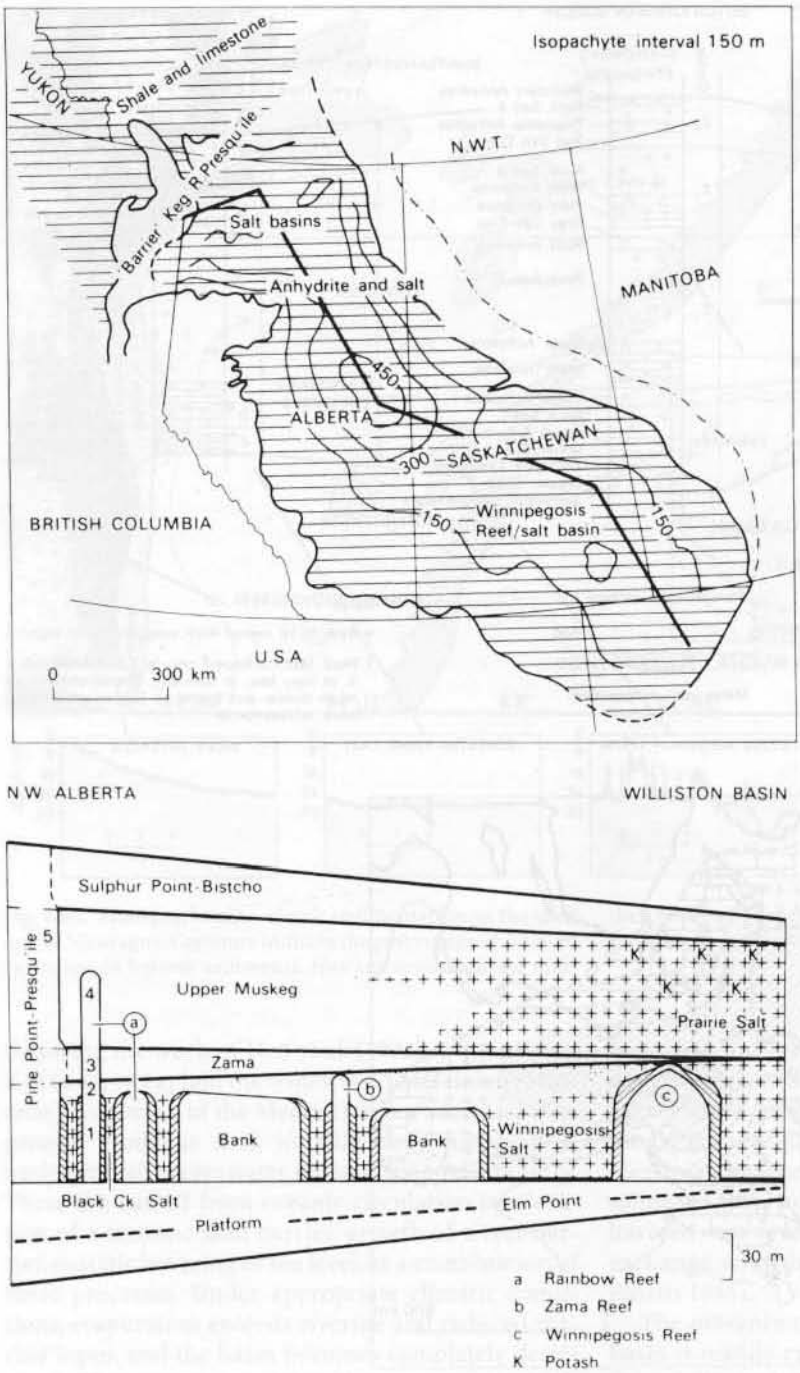


Fig. 6.62. The Upper Elk Point Basin of western Canada (Middle Devonian) and a diagrammatic cross section. Anhydrite flanks, reefs, and bank edges (*oblique hatching*); halite is shown by *crosses*. (Fuller and Porter 1969; Reading 1978)

of northern England and the North Sea basin. He developed two alternative models (Figs. 6.63, 6.64). In the first of these, incomplete evaporative drawdown leads to the development of thick gypsum wedges blanketing the basin margin. Resedimented gypsum in the form of sediment gravity flows may be deposited on the basin floor, where laminated basinal carbonates, gypsum, or organic-rich sediments may be accumulating. Sabkha-type deposits develop on the basin margin during transgression or late highstand, times at which accommodation is low and circulation of marine waters at the surface may be inhibited. In the second model, complete drawdown may lead to a much more complete desiccation at the basin centre, giving rise to deposition of halite and potash salts. Erosional or karst features may be much more prominent at the sequence boundary under such conditions.

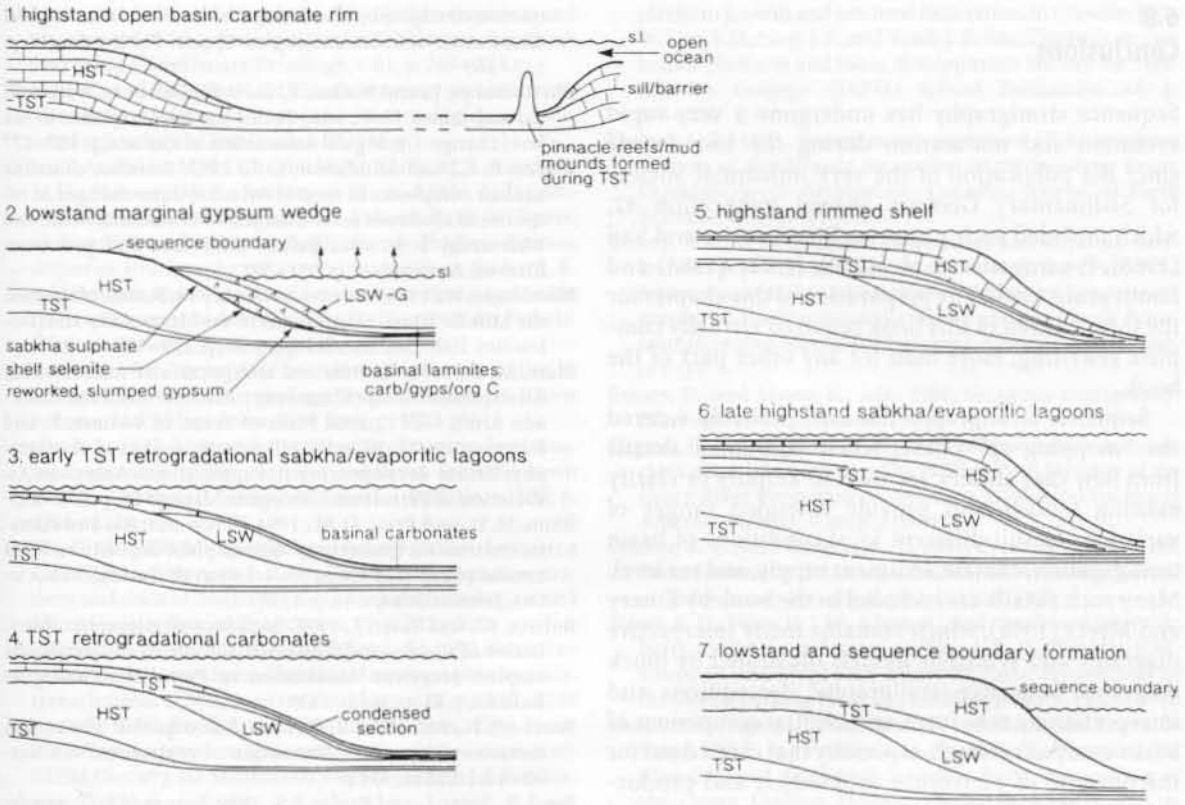


Fig. 6.63. Sequence-stratigraphic model for a carbonate–evaporite basin under conditions of incomplete drawdown. (Tucker 1991)

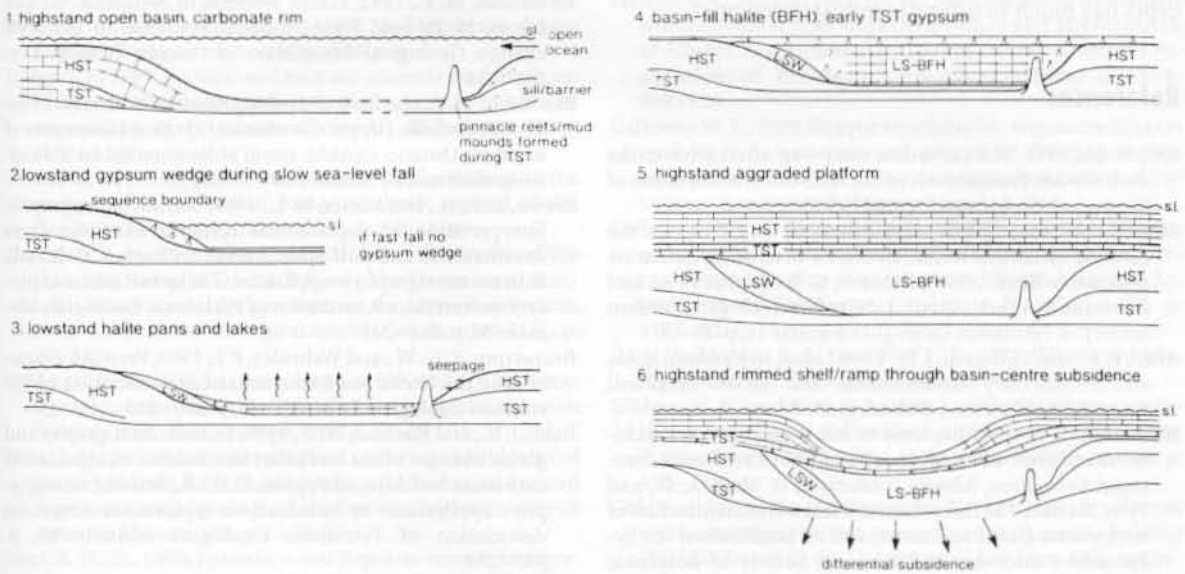


Fig. 6.64. Sequence-stratigraphic model for a carbonate–evaporite basin under conditions of complete drawdown. (Tucker 1991)

6.8 Conclusions

Sequence stratigraphy has undergone a very rapid evolution and maturation during the last decade since the publication of the very influential Society for Sedimentary Geology Special Publication 42, which included such papers as Posamentier and Vail (1988), Posamentier et al. (1988), Jervey (1988) and Loutit et al. (1988). The preparation of this chapter for the third edition of this book required virtually complete rewriting, more than for any other part of the book.

Sequence stratigraphy has now probably entered the "mopping-up" phase, when additional details from new case studies are used to amplify or clarify existing models and provide extended ranges of variations to suit different local conditions of basin configuration, climate, sediment supply, and sea level. Many such details are included in the book by Emery and Myers (1996), which contains many interpretive diagrams and synthesis models illustrated by block diagrams. Sequence-stratigraphic descriptions and interpretations now form an essential component of basin-analysis research, especially that carried out for the purpose of petroleum exploration and production. Stratigraphic synthesis is now normally carried out using sequence principles, whether or not the formal methods of allostratigraphy are employed. As shown by Cross et al. (1993), even at the level of the individual reservoir, a knowledge of sequence concepts has much to offer the industry scientist.

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Part 2
Global Controls of Sedimentary Basin Development

The position, subsidence behavior, and architecture of a sedimentary basin can now be seen to be dependent on the interaction of a range of global and astronomical controls.

The major cause of change in the earth's crust is the radiogenic heat engine, which drives mantle convection and generates the geomagnetic field. Convection distributes heat and drives plate tectonics. Density inhomogeneities occur in the mantle depending on the location of continents and subduction zones, and these lead to the formation of large-scale geoid anomalies (Anderson 1984) and to mantle convection, responsible for "dynamic topography" and the process of epeirogeny (Gurnis 1988). Since at least Proterozoic time, continental fragments, large and small, have drifted across the earth's surface, assembling and disassembling over long periods of time (hundreds of millions of years) (Windley 1984). Major changes in the configuration of the continents have small effects on the moment of inertia of the rotating globe, leading to slow changes in the geoid and changes in the mantle relative to the spin axis. True polar wander may occur. The assembly of the Pangea supercontinent in the late Paleozoic is thought to have caused a thermal blanketing effect over a large part of the earth and a consequent major geoid anomaly. Eventually, breakup was probably induced by thermal instability and membrane stresses caused by the rotational bulge (Anderson 1984). Pangea may not have been the first supercontinent to have formed on the earth's surface (Hoffman 1991; see Sects. 8.2.1, 8.3.1).

Rapid continental separation is accompanied by the generation of large thermal anomalies at spreading centers, with consequences for reduction in ocean basin volumes and the elevation of global sea levels. Evidence is now accumulating for a direct correlation between times of fast sea-floor spreading and times with a quiet magnetic field, during which there are few reversals. Sheridan (1987) attributed these observations to the effects of the eruption of mantle plumes from the core. During times of plume eruption, fast mantle convection leads to rapid sea-floor spreading, a relatively cooler core with weaker convection there, and consequently, a weaker magnetic field with few reversals. The mantle plumes gradually decay as heat builds up in the core. Turbulence there leads to the generation of a stronger magnetic field with many reversals, while, in the mantle, slower convection is accompanied by slower sea-floor spreading and reduced rates of continental breakup and collision. These cycles take 60–100 m.y. to complete, a length that compares with that of the 10^6-year cycles of sea-level change discussed in Chapter 8.

Dispersing and colliding continents interact according to predictable tectonic patterns, most of which include the development of major hollows in the earth's crust, which become receptacles for sediment. There are two major styles of sedimentary basin: those caused by crustal extension during divergent plate movements and those caused by contraction and crustal thickening during convergent plate movements (extension may also occur during convergent movements). The contrasting styles of subsidence and their dependence on the thermal and rheological properties of the crust and mantle are discussed in Chapter 7. The large-scale stratigraphic architecture of sedimentary basins is controlled largely by their plate-tectonic setting (Chap. 9).

The widespread development of shallow seas during times of high eustatic sea level has an ameliorating effect on global climate but leads to a slowing of oceanic circulation and a consequent lowering of oxygenation levels. Conversely, the passage of large continental masses across polar regions dramatically affects global air and water circulation patterns, increasing thermal oceanic circulation while tending to trigger hemispheric glaciation. Large-scale regional uplift, such as occurs during the breakup of supercontinents, may also help to trigger widespread continental glaciation. The sensitivity of the earth to climatic change, particularly glaciation, also depends on the global distribution of solar radiation, which is controlled by a series of cyclic changes in the orientation of the earth's axis relative to the sun and in the shape of the earth's orbit around the sun. Rhythmic changes over periods of tens to hundreds of thousands of years are the result and can be detected in the stratigraphic record in the form of various types of cyclicity. This is particularly apparent where climatic change leads to continental glaciation, because of the major withdrawal of

water from the oceans and the lowering of global sea-level that results. This we also examine in Chapter 8.

Many of the paleogeographic factors that control the evolution of sedimentary basins, including basin shape, subsidence patterns, water depths, and location and elevation of adjacent landmasses (sediment sources) are controlled by the plate-tectonic setting of the basin. Two additional factors affect the basin architecture: (1) the unique geological history of the plate on which the basin is located, which controls the configuration of the local basement and the geography of surrounding plates, and (2) the latitude of the basin, which is the major control on climate and, therefore, one of the major controls of climate-sensitive sediments, such as carbonates, evaporites, and eolian sands. The plate-tectonic control of sedimentary basins is discussed in Chapter 9. The basins described in this chapter illustrate some of the unique controls of plate history and latitudinal setting over which plate-tectonic processes are overprinted.

One of the most important controls on the nature of the sedimentary record is the continual incremental change over geologic time in the accommodation space for sediment. This is the space made available by an increase in water depth in the sea or by the upward translation of river profiles by source-area uplift on land. Changes in accommodation are generated by a multiplicity of causes, including eustatic sea-level changes driven by changes in ocean-basin volume and glacioeustasy, and changes in basin elevation driven by tectonism, including continental-scale epeirogeny (dynamic topography), regional extensional subsidence, and crustal loading. The rises and falls of relative sea level in basins lead to transgressions and regressions and to predictable changes in large-scale stratigraphic architecture. The documentation and interpretation of these changes is the study of sequence stratigraphy (Chap. 6). The predictability of sequences in the subsurface has become a powerful tool for petroleum exploration, and unravelling the causes of accommodation changes has triggered much invaluable research into long- and short-term allogenic processes (Chap. 8).

From core, through mantle to lithosphere, and including the oceans and the atmosphere, the earth is an integrated dynamic system. Sedimentary basins on the lithosphere are but one aspect of this system. To understand them, we must know more about the Earth than sedimentology and stratigraphy. Conversely, the stratigraphic record in basins is an invaluable record of past tectonic, climatic, and other events that have been important in shaping geologic history. Their study helps us to learn more about Planet Earth.

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Tectonism and Sedimentation: Principles and Models

7.1 Introduction

There are three crustal stress environments in which sedimentary basins are created (Fig. 7.1): extensional basins, in which the axis of maximum stress is vertical, and contractional and shear basins, in which this axis is horizontal. The various plate-tectonic settings in which these configurations occur are discussed in Chapter 9.

The processes or **driving mechanisms** that actually create basins in these stress environments are of three types. In the case of extensional basins, heating of the lithosphere from below and stretching and thinning of the crust are the critical controlling factors. One or more of these processes, in combination, leads to the generation of rift basins and downwarps filled with thick sedimentary prisms on divergent plate margins (Sect. 9.3.1). The same processes also control the architecture of extensional basins formed in shear settings, such as pull-apart basins (Sect. 9.3.3). These mechanisms are discussed in Section 7.3.

Contractional processes cause crustal thickening, for example, by the stacking of thrust slices or nappes. This leads to increased local loads and consequent isostatic subsidence, thus generating a moat bordering the margins of the load. Foreland basins are the classic examples of basins caused by crustal loading (Sects. 9.3.2 and 9.3.4), and the same type of mechanism also occurs in transpressional settings, where part of the shear stress may be released by reverse

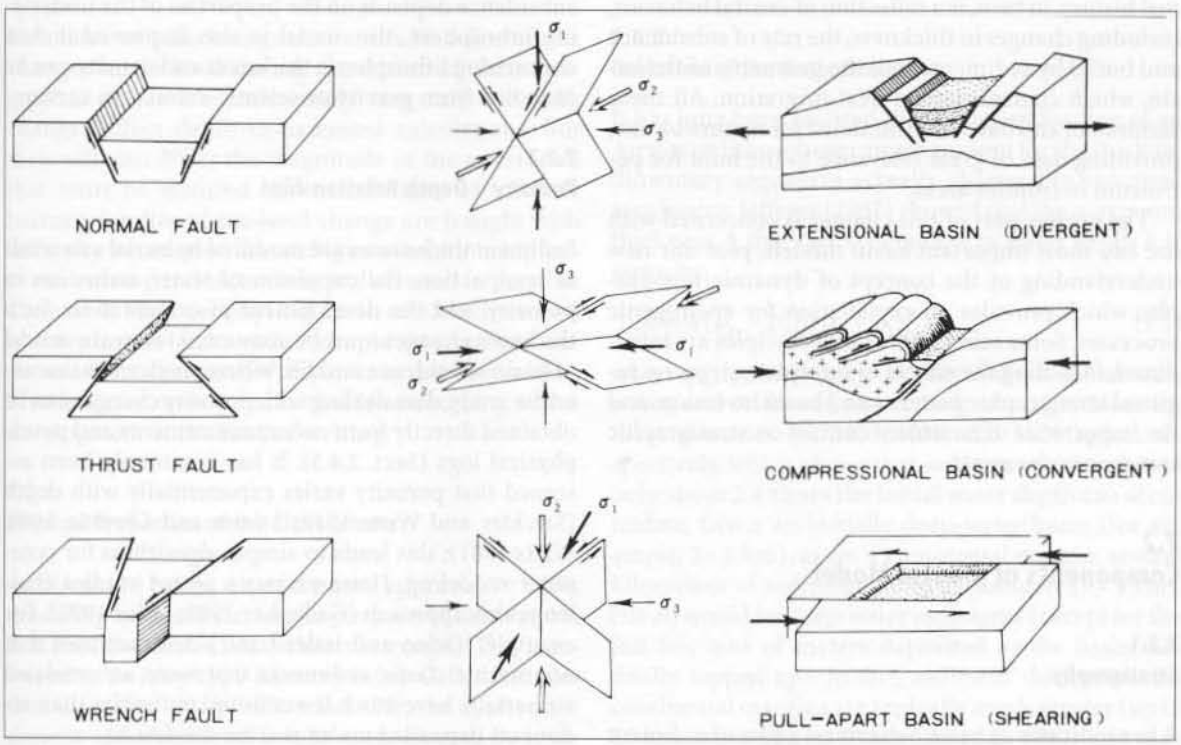


Fig. 7.1. The three principal stress environments under which sedimentary basins are produced. (Liu 1986; reprinted with permission of the American Association of Petroleum Geologists)

faulting. These processes of basin formation are discussed in Section 7.4.

Geophysicists have been studying the processes of basin formation with increasing success since the early 1970s, and our understanding of the two classes of basin models outlined above has now reached a considerable degree of sophistication. Procedures have been developed for modeling sedimentary basins using computer programs. These feed measured or estimated structural and stratigraphic data into sets of physical equations based on selected values of crustal thickness, strength, heat flow, isostatic response, and other characteristics. By varying the crustal properties in the model, the synthetic basins can be made to resemble real basins. This resemblance tells us that we have identified the appropriate input parameters for crustal behavior. Such models have considerable predictive value. They provide us with a physical understanding of mantle and lithosphere processes, plate-tectonic processes, isostasy, sea-level change, and, in the case of the newest models, the effects of climate change on sediment supply. This knowledge can then be applied to the interpretation of other basins, and the modeling process may yield valuable information on parameters of key importance in the understanding of various economically important deposits, such as hydrocarbons. For example, petroleum generation and migration are dependent on the thermal history of the basin. The thermal history, in turn, is a reflection of crustal behavior, including changes in thickness, the rate of subsidence and burial by sediments, and the geometry of the basin, which controls lateral heat migration. All these factors can and have been modeled for mature basins, providing data of great relevance to the hunt for petroleum in frontier areas.

The greater part of this chapter is concerned with the two most important basin models, plus our new understanding of the concept of dynamic topography, which provides an explanation for epeirogenic processes. Some other important principles are introduced, including the effects of intraplate stress on regional stratigraphic patterns and basin inversion, and the importance of basement control on stratigraphic trends in cover strata.

7.2 Components of a Basin Model

7.2.1 Stratigraphy

A key indicator of basin behavior is a plot of sediment thickness against age. Changes in the slope of this curve with time yield essential information about subsidence rates and sediment supply that can then be matched to rates of subsidence predicted from physical models. Therefore, a thorough documentation of the stratigraphic fill of a basin is of fundamental importance. At least one complete section down to the basement, with the best available chronostratigraphic subdivisions, must be available. For example, considerable advances in the understanding of extensional basins came about following the drilling of several Continental Offshore Stratigraphic Test (COST) wells off New York in the mid 1970s (Scholle 1977, 1980). Application of new **backstripping** techniques (Sect. 7.3.2) to these wells by Steckler and Watts (1978) provided the key understanding of **Atlantic-type margins** that has become the standard for the interpretation of extensional basins. Techniques for describing stratigraphic sections are discussed in Chapter 2 of this book, and their correlation is dealt with in detail in Chapter 3.

7.2.2 Basin and Crustal Geometry

In order to model a basin, we must know something about its shape and internal architecture, including the position and structure of key stratigraphic units. These parameters can be reproduced in some of the models discussed below, permitting a powerful check on the validity of input parameters. Because basin subsidence depends on the properties of the underlying lithosphere, the model is also improved if data concerning lithospheric thickness and density can be obtained from gravity or seismic-refraction surveys.

7.2.3 Porosity – Depth Relationships

Sediment thicknesses are modified by burial as a result of compaction, the expulsion of water, reduction in porosity, and the development of cementation. Such thickness changes must be accounted for in any model of basin subsidence and fill. Where modern basins are under study, data dealing with porosity changes can be obtained directly from core measurements and petrophysical logs (Sect. 2.4.5). It has commonly been assumed that porosity varies exponentially with depth (Steckler and Watts 1978; Sclater and Christie 1980; Watts 1981); this leads to simple algorithms for computer modeling. However, more recent studies challenge this approach (Gallagher 1989; Issler 1992). For example, Nadon and Issler (1997) demonstrated that nonmarine clastic sediments that were accumulated subaerially have much lower initial porosities than sediments deposited under marine conditions.

In the case of ancient basins, porosity changes relating directly to initial burial may have been com-

pletely distorted by subsequent cementation and diagenesis. Thus, Bond and Kominz (1984) developed techniques for delithification to deal with the Paleozoic continental-margin sediments of the southern Canadian Cordillera preserved on what was, during sedimentation, the cratonic edge of North America. By removing the effects of lithification, they were able to compare the basin's history to that of other extensional margins and make many useful deductions concerning the evolution of the continental margin.

7.2.4 Paleobathymetry

If deductions are to be made about tectonic controls, the position of the basement during burial must be documented. Vertical changes of thousands of meters can occur and can lead to quite different shapes for the subsidence–time curves for a basin fill. Estimates of paleobathymetry can be made from biofacies data, but commonly this is difficult, and the results are fraught with error. They must be attempted if a basin model is to be better than an order-of-magnitude approximation (Célérier 1988). A brief discussion of this topic is given in Section 4.5.7.

7.2.5 Sea-Level Change

Estimates of paleobathymetry must be adjusted for changes in sea level during deposition of the stratigraphic succession under study. Not only will such changes affect depth-to-basement calculations, but they will also affect the magnitude of the water load that must be factored into calculations of isostatic balance. Studies of sea-level change are fraught with difficulty because of the ambiguous relationship between eustasy and tectonism, as discussed in detail in Chapter 8. Most geophysical basin models discussed in this chapter make rather simple assumptions about sea-level change that may or may not be justified. This is a controversial area that is undergoing active development at the time of writing.

7.2.6 Physical Properties of the Lithosphere

As discussed below, a useful initial approach to basin modeling is to assume, for the purpose of calculation, that the lithosphere has no lateral strength. This is referred to as **Airy-type isostasy** (after G. B. Airy, who first discussed this problem in 1855); the applied sediment and water load is fully compensated by localized displacement of the denser mantle, which is assumed to be liquid and, therefore, to have no strength. An example would be where the lithosphere is cut by numerous steep faults, allowing localized response to loading. However, only in special circumstances is this actually the case. Under most conditions, the lithosphere has measurable physical strength, such that the effects of the load are spread beyond the point or area over which it is applied. This is the basis for the **flexural model** of lithospheric behavior. Geophysicists have found it useful to model the lithosphere much as mechanical engineers model structural materials. Under some conditions, the lithosphere behaves as an elastic plate (the **elastic-plate model**); it responds to loads by bending, but tends to resume its original configuration if the load is removed. However, other models demonstrate that the lithosphere may have a measurable viscosity; if a load is applied for long enough, part of the elastic response is released as irreversible strain. Under these conditions, the lithosphere is said to be viscoelastic (the **viscoelastic-plate model**). The viscosity and, hence, the nature of the lithospheric response are temperature dependent. The lithosphere becomes less viscous and, therefore, more rigid when moved away from a heat source such as an oceanic spreading center. This has critical implications for the architecture of the resulting sedimentary basin.

7.3 Extensional Basins

7.3.1 Evolution of the Basic Model

It has long been realized that sediment loading of an Airy-type lithosphere cannot account for the thick sedimentary sequences actually observed in sedimentary basins. Jeffreys (1962) showed that the maximum thickness A that can accumulate in water depth h is given by:

$$A = h(\varrho_m - \varrho_w)/(\varrho_m - \varrho_s), \tag{7.1}$$

where ϱ_m, ϱ_w, and ϱ_s are the densities of the mantle, water, and sediment, respectively. Average values for these densities are 3400, 1030, and 2400 kg/m^3, respectively, which shows that sediment thicknesses of only about 2.4 times the initial water depth can accumulate. Given an initially deep-water basin (for example, 2–3 km), as on a continental margin, several kilometers of sediment could accumulate (5–7 km), but all would be deep-water sediments (except for the last few tens of meters deposited as the basin was finally topped up). In fact, sediment thicknesses on continental margins are typically much greater (up to 16 km, according to Beaumont et al. 1982) and may contain substantial volumes of shallow-water sediments, particularly on the landward side of the basin,

indicating that there, at least, water depths were never more than a few tens of meters.

It is now known that the Airy model is only applicable where the crust is cut by numerous steep faults and can thus be treated as a stack of unconnected fault slices with no transmission of imposed load between the slices. Such a condition may occur in the early stages of rift development. In most places, the crust has lateral strength and responds to imposed loads by elastic bending. Walcott (1970 a, b, 1972) carried out several important early studies of the physics of the lithosphere and arrived at many useful conclusions regarding the flexural rigidity, viscosity, and thickness of the lithosphere. For example, it has long been known that, because of the elastic properties of the lithosphere, the load stress of a major ice cap produces a depression that extends far beyond the limits of the ice itself. This depression is typically occupied by glacial lakes. The flexural rigidity D of an elastic sheet is a measure of its stiffness, or resistance to bending, under the action of a bending or flexural couple. For a lithosphere plate of thickness T, Young's modulus E, and Poisson's ratio σ (following Walcott 1970b),

$$D = ET^3/12(1 - \sigma^2). \qquad (7.2)$$

Elastic up-bending of the lithosphere at the edge of the load produces an uplifted forebulge at the margins of the depression (Fig. 7.2). Walcott (1970b) summarized several studies of flexural rigidity that were used in later basin-modeling experiments. For example, Watts and Ryan (1976) investigated the subsidence history of the Gulf of Lion, France, and the Cape Hatteras area of North Carolina. They progressively removed the sediment load from the basin using backstripping procedures (described below) in order to isolate any tectonic, driving sediment load. In general, real basins contain much greater volumes of shallow-water sediment than are predicted by simple flexural models.

Two approaches to this problem emerged: thermal hypotheses and crustal-thinning hypotheses. Bott (1978) gave a useful review of the early developments of these ideas. The thermal hypothesis was developed mainly by Sleep (1971). He proposed that the heating that accompanies continental splitting causes thermal expansion, consequent reduction in density, and isostatic uplift of the continental margins. Surface erosion of the uplifted margins then thins the crust so that, when it cools and subsides, it sinks below its original level, creating space for sediments. Continental crust and oceanic crust have similar cooling patterns. The depth to the top of the crust is dependent on age and follows a time-dependent, exponential cooling curve, such that subsidence is a linear function of $t^{1/2}$, t being the age of the crust (Sclater et al. 1971; Sleep 1971). Studies of several continental margins confirmed the validity of this relationship, indicating that the thermal model explains much of the subsidence (Keen 1979). However, there remains a serious space problem. Unless significant erosion of the up-domed crust occurs, the crust will simply subside to its original level following the thermal event, and no space for sediments will have been created. Isostatic considerations show that only about half the thickness of the crust removed by thinning or erosion can be replaced by sediment deposited up to sea level, a condition that cannot possibly explain the case of major Atlantic margin-type basins, where sediment thicknesses of 5–10 km are typical (Bott 1978).

Some attempt has been made to circumvent the space problem by proposing additional subsidence due to seaward creep of lower crustal material or phase changes, metamorphism, or intrusion of igneous material into the lower crust, all of which would increase the density and, thus, cause additional

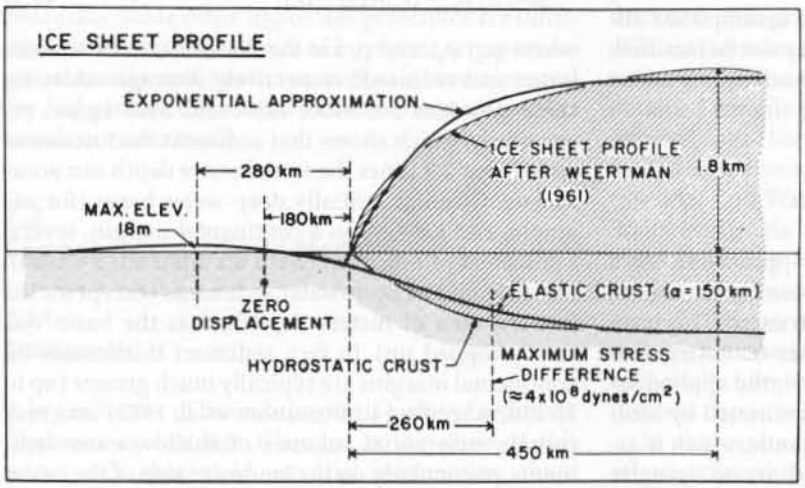


Fig. 7.2. Profile across an ice front, showing the depression, or moat, in front of the ice and the forebulge beyond. The moat is the site of glaciolacustrine sedimentation, and the forebulge undergoes erosion. Typical values for widths and depths were calculated by Walcott (1970a)

subsidence. The igneous-intrusion idea now forms the basis for an important variety of continental-margin model, as discussed below. However, the other proposals are difficult to model and have little support in the form of independent field evidence for the events that are supposed to have occurred. The thermal model cannot explain intracratonic basins, such as the Michigan or Williston Basins, for which there is no evidence of an initiating thermal event (their origin is further discussed in Sections 7.5, 9.3.6).

The other group of factors to be considered in basin formation includes those that lead to subsidence as a result of crustal thinning. That such thinning occurs has long been known (Artemjev and Artyushkov 1971). Bott (1978) discussed the various mechanisms whereby the crust might be thinned, including brittle failure (faulting) at the surface and ductile creep in the lower lithosphere. McKenzie (1978) discussed the growing evidence for the importance of listric normal faults in extensional tectonic environments, an importance that has been confirmed by recent deep-crustal seismic experiments in areas such as the Bay of Biscay (Montadert et al. 1979) and the Basin and Range Province of the United States (Chap. 9).

The main breakthrough in the development of a fully operational model for extensional basins was made by McKenzie (1978). He focused attention on the thermal consequences of the stretching and thinning of continental crust during continental separation. His model is illustrated in Fig. 7.3. At time $t = 0$, a unit length of continental lithosphere (length = α) is extended by a factor β to length $\beta\alpha$. The lithosphere is thinned by the factor $1/\beta$. Isostatic compensation causes upwelling of the hot asthenosphere, with a resultant increase in the geothermal gradient. Cooling of this hot material back to the equilibrium gradient that existed before stretching causes an increase in density and consequent subsidence. Hsü (1982) suggested that this processes be termed **Pratt subsidence**, in honor of J.H. Pratt, who in 1864 was the first to discuss thermal isostasy. McKenzie (1978) showed that, where the initial thickness of the continental crust is greater than 18 km, subsidence will take place, although areas of thin crust may actually undergo net elevation. Much research has been and is being conducted to determine the values of β in different basins. It has been shown that this quantity varies across a given basin, from unity at the basin edge to infinity at the edge of a rifted continental plate adjacent to oceanic crust. Average values of β equal to 2 or less seem to typify most rift basins, except where the crust is significantly widened by the intrusion of mafic dikes (Dewey 1982), as occurs on some continental margins adjacent to active spreading centers (Royden et al. 1980).

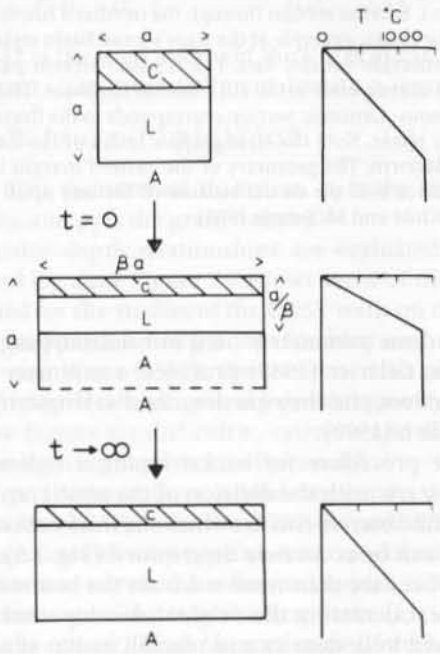


Fig. 7.3. The stretching and thermal subsidence model of McKenzie (1978). See text for explanation

The sequence of initial rapid continental stretching followed by slower, thermally driven subsidence combined with flexural loading gives rise to the classic **Steer's-Head** or **Texas-Longhorn model** of basin geometry (Fig. 7.4). This geometry is characteristic of failed rifts, aulacogens (Sect. 9.3.1), and some intracratonic basins (Sect. 9.3.6). The development of oceanic crust on divergent continental margins splits the basin into two halves, which evolve separately on either side of the ocean (Sect. 9.3.1).

With the basic physical model that evolved from the work of Walcott, Sleep, Watts, and McKenzie, geophysicists now had a powerful tool for computer modeling of real-world basins. Stratigraphic and other data are backstripped to reveal tectonic driving mechanisms, and these are fitted to equations of crustal behavior involving subtle modifications of the flexural model to incorporate possible geographical and temporal variations in flexural rigidity, viscous versus elastic behavior, internal heat sources, such as radioactivity, varying rates of sediment supply, and the thermal-blanketing effects of the sediments, which have low heat conductivity. Some of these refinements are discussed in Section 7.3.3.

7.3.2 Backstripping Techniques

Backstripping is a technique for progressively removing the sedimentary load from a basin, correcting for compaction (plus lithification, if necessary), paleo-

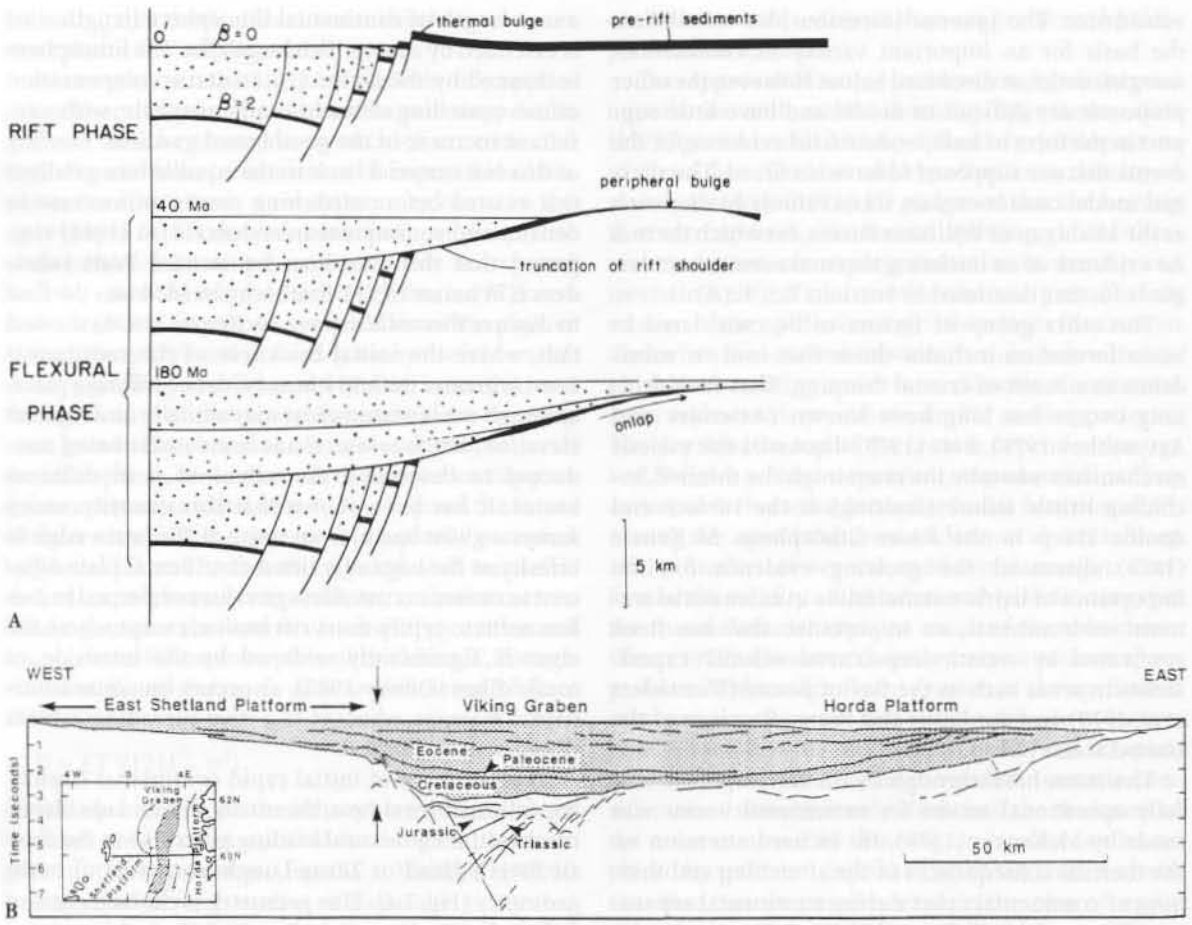


Fig. 7.4 A, B. The **Steer's-Head** or **Texas-Longhorn model** of basin geometry. **A** Idealized model showing the two phases of basin development. An initial "rift phase" is generated by rapid crustal stretching, resulting in listric normal faulting and ductile flow of the lower lithosphere, with Airy-type isostatic subsidence. The thermal anomaly generated during this phase causes an uplifted thermal bulge, which typically is eroded subaerially, forming an unconformity. The thermal anomaly then decays, resulting in relatively slower subsidence over a much broader area than the initial rift, creating the "horns" of the model – the so-called "flexural phase" of basin development. The origin of the stratigraphic onlap during the flexural subsidence phase is discussed in the text (from Dewey 1982; reprinted with permission of the Geological Society, London). **B** Cross section through the northern North Sea Basin – the classic example of the steer's-head basin style, drawn from reflection-seismic data. The Triassic–Jurassic part of the section corresponds to the rift phase of the basin. The shaded Cretaceous–Cenozoic section corresponds to the flexural subsidence phase. Note the stratigraphic onlap of the East Shetland Platform. The geometry of the eastern margin is not in accordance with the model because of Tertiary uplift of Norway. (White and McKenzie 1988)

bathymetry, and changes in sea level, in order to reveal the tectonic driving mechanisms of basin subsidence. The load is fitted to an Airy-type or flexural-subsidence model, and the residual subsidence that is revealed can then be related to thermal behavior and changes with time in crustal properties. The technique was developed by Sleep (1971) and was first explored in detail by Watts and Ryan (1976). Steckler and Watts (1978) applied the McKenzie (1978) stretching model to offshore stratigraphic data from the continental margin off New York in a paper that has become a standard work on the subject. Gallagher (1989) considered the inherent potential for error in the various parameters used in backstripping calculations. Célérier (1988) provided a summary of the procedures, and they are described at length by Allen and Allen (1990).

The procedure for backstripping a sedimentary basin starts with the division of the stratigraphic column into increments for which the thickness and age range can be accurately determined (Fig. 7.5). These time slices are then removed from the basement one by one, calculating the original decompacted thickness and bulk density and placing its top at a depth below sea level corresponding to the average depth of water in which the unit was deposited (W_{d1}; Fig. 7.6 B).

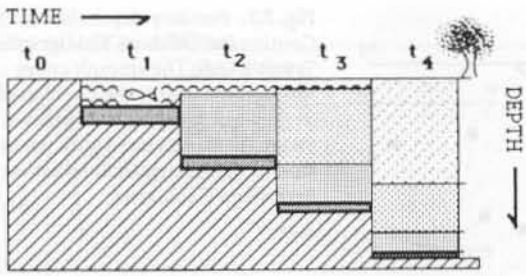


Fig. 7.5. Time slices in the subsidence history of a basin. Note that water depth, sedimentation rate, and compaction all vary with time. Each increment of sediment is compacted beneath the weight of successive increments, and this effect must be removed for each time slice in the backstripping procedure. (Mayer 1987)

The isostatic subsidence caused by the weight of this sediment can then be calculated, and the depth to the surface on which the sediment was deposited (Y_1) is calculated, with only the weight of the water as the basement load (Fig. 7.6). Each successively older unit is removed in turn, and the decompacted thicknesses and bulk density of the remaining units are calculated and positioned with respect to contemporaneous estimated water depths, and so on, up the column. These steps correspond to the incremental unloading from t_4 to t_0 shown in Fig. 7.5. Water depths must also be corrected for any eustatic changes in sea level, Δ_{SL}. The equation for determining Y is, as given by Steckler and Watts (1978):

$$Y = S^* \left[\frac{(\varrho_m - \varrho_s)}{\varrho_m - \varrho_w} \right] + W_d - \Delta_{SL} \left[\frac{\varrho_m}{(\varrho_m - \varrho_w)} \right] \qquad (7.3)$$

where S^* is the total sediment thickness at each time increment i. The mean sediment density ϱ_s is given by

$$\varrho_s = [\Sigma_i [\phi_i \varrho_w + (1 - \phi_i) \varrho_g] T_i] / S^* , \qquad (7.4)$$

where T_i is the interval thickness, ϕ_i is the interval porosity, and ϱ_g is the grain density.

Porosity–depth relationships are evaluated from core and log data. Figure 7.7 shows some of the data compiled for the studies of the COST wells on the Atlantic continental margin by Steckler and Watts (1978) and Watts (1981). Corrections must also be made for sea-level change and paleobathymetry. Both of these factors are difficult to estimate and may introduce significant error into the model.

Figure 7.8 shows three subsidence curves derived by Steckler and Watts for the COST B-2 well. The first curve assumes an Airy-type isostatic model, whereas the other two curves use a flexural model. Note that the tectonic subsidence in the first curve is less than that in the others. This is because some of the load, in the case of the flexural models, is compensated over a wide area of the continental margin rather than entirely at the base of the column. The difference between the two flexural models stems from differences in the assumed position of the shelf-edge during the Eocene, a detail that need not concern us here. Tectonic subsidence, driven by long-term crustal thinning and thermal effects, should be smooth. In fact, the tectonic subsidence curves in Fig. 7.8 are much

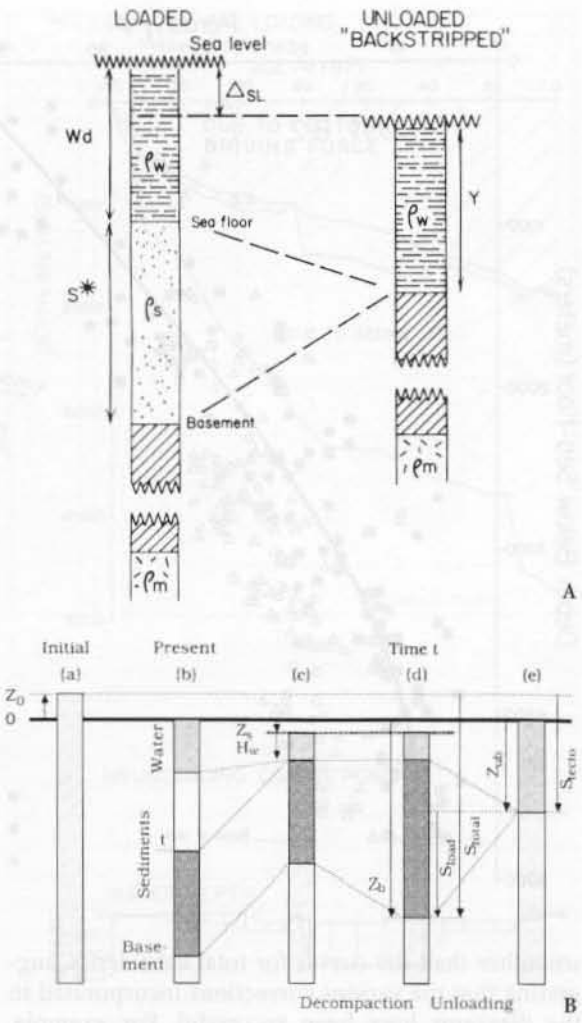


Fig. 7.6. **A** The basic relationships between a loaded sedimentary section and an unloaded or backstripped section. *Symbols* are those used in Eqs. (7.3) and (7.4) and are explained in the text (from Steckler and Watts 1978). **B** The steps involved in backstripping. *a* The initial position of unloaded basement (*dotted line*) showing initial basement depth, Z_0. *b* Present-day filled basin. Stratigraphic analysis permits subdivision of the section into slices that are successively removed in the backstripping process. One such slice, between the horizon dated t and basement, is shown by *dark shading*. *c* Reconstruction of the elevation of this unit at time t, corrected for water depth H_w (paleobathymetry) and sea level Z_s. *d* Correction of unit thickness for compaction indicates true basement depth Z_b. *e* Subtraction of the compacted thickness (S_{load}) from the total subsidence (S_{total}) reveals the subsidence due to tectonism (S_{tecto}). (Célérier 1988)

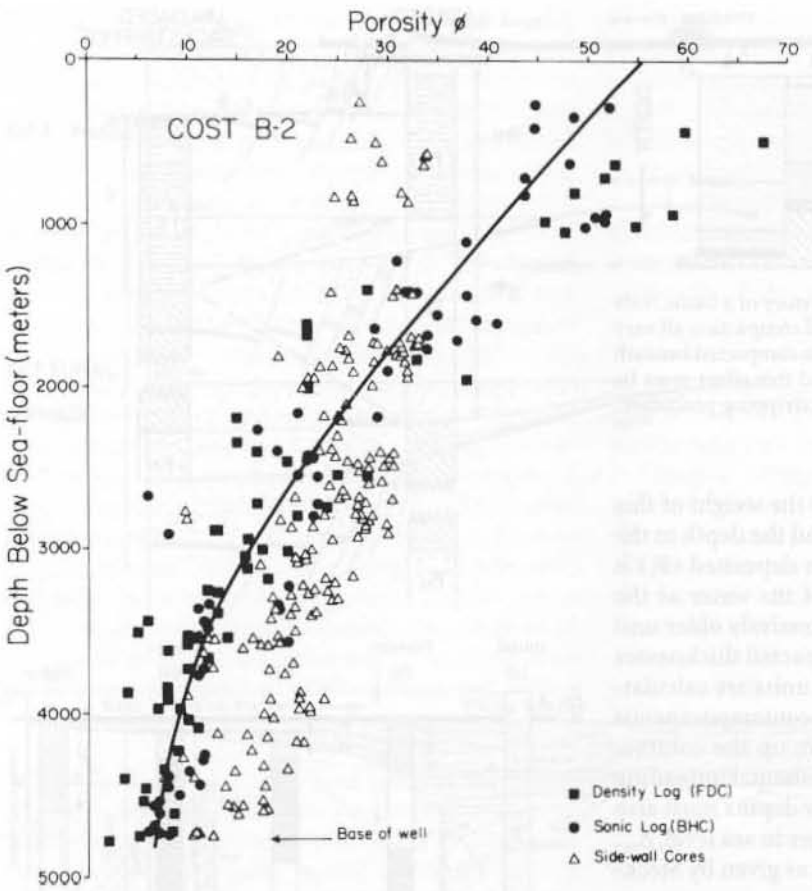


Fig. 7.7. Porosity-depth data for the Continental Offshore Stratigraphic Test B-2 well. The smooth curve through the data is the curve used in the backstripping calculations. (Watts 1981; reprinted with permission of the American Association of Petroleum Geologists)

smoother than the curves for total subsidence, suggesting that the various corrections incorporated in the diagrams have been successful. For example, Steckler and Watts (1978) suggested that the sharp change in the rate of total subsidence at about 80 Ma is the result of changes in the sedimentation rate accompanying a peaking in sea level, followed by its fall in the latest Cretaceous. This kink in the curve is eliminated in the tectonic subsidence curves.

Bond and Kominz (1984) applied the same techniques to the lithified Paleozoic sediments of the North American cratonic margin of the southern Canadian Cordillera and developed techniques for delithifying ancient sediments. They then used their curves to throw light on the question of the timing of continental separation and divergent-margin development. Sclater and Christie (1980) addressed the problem of correcting for porosity changes in basins such as the North Sea, where depth–porosity relationships are distorted by overpressuring. An interesting suite of backstripping diagrams for coastal Cenozoic basins of California was published by Dickinson et al. (1987). They showed how to maximize the usage of incomplete data and to plot the diagrams in such a way as to clarify possible ranges of error. Other useful studies have been performed by Royden and Keen (1980), Watts (1981), Watts et al., (1982), Beaumont et al. (1984), Meyer (1987), and Gallagher (1989).

Bond et al. (1989) introduced the terms **first reduction** (R1) and **second reduction** (R2) for the products of backstripping analysis. The R1 curve represents the removal of the effects of sediment loading, compaction, and water-depth changes from the cumulative thickness curve. As noted by Bond et al. (1989), "this step is the basis for identifying the tectonic component of subsidence and comparing it with subsidence models ..." The R2 curve is derived by extracting the best-fit curve for tectonic subsidence (such as the exponential thermal-subsidence curve for extensional-margin basins) and plotting the residual. Ideally, this reveals the nature and magnitude of external events superimposed on the tectonic subsidence. When R2 curves from widely separated areas are closely correlated and compared, they may reveal the timing and importance of eustatic sea-level changes, although errors in the dating of the stratigraphic successions and in the assumptions made during delithification reduce the accuracy and precision that might be expected from such an analysis.

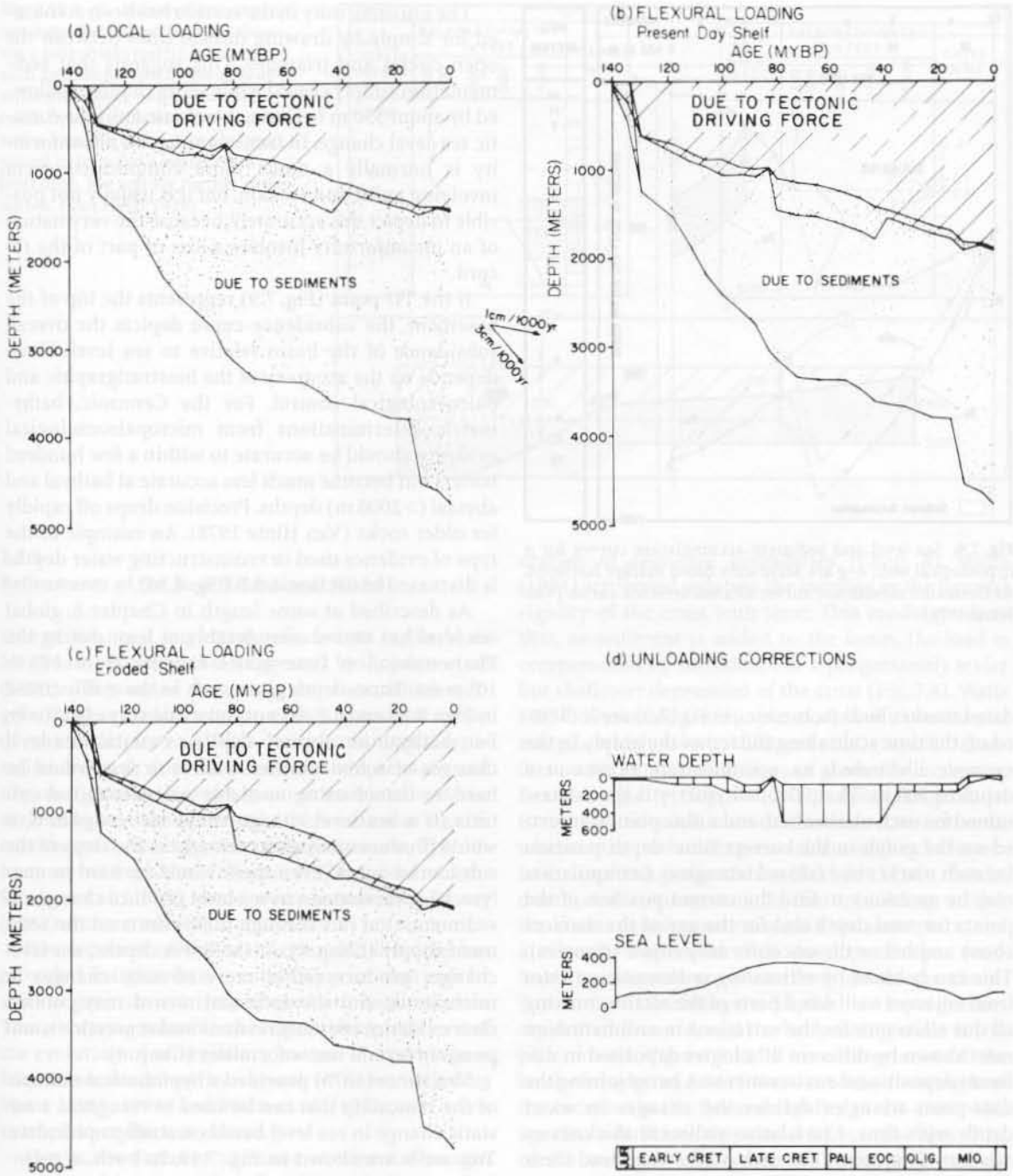


Fig. 7.8. Subsidence curves for the Continental Offshore Stratigraphic Test B-2 well. Subsidence caused by sediment and water loads is shown in the lower part of each diagram, with that resulting from tectonic driving forces shown above. Corrections for water depth and sea level are shown separately and are subtracted from the sediment load to give the tectonic subsidence. The differences between the three curves are discussed in the text. (Steckler and Watts 1978)

A procedure somewhat analogous to backstripping was termed **geohistory analysis** by Van Hinte (1978). This singularly ambiguous phrase was intended by Van Hinte to apply to the method of constructing time–depth plots for continuous stratigraphic sections. Much valuable stratigraphic detail can be extracted from well data using his methods. An example is illustrated in Fig. 7.9, and its method of construction is as follows. The chronostratigraphy of the section is determined as accurately as possible. Well-

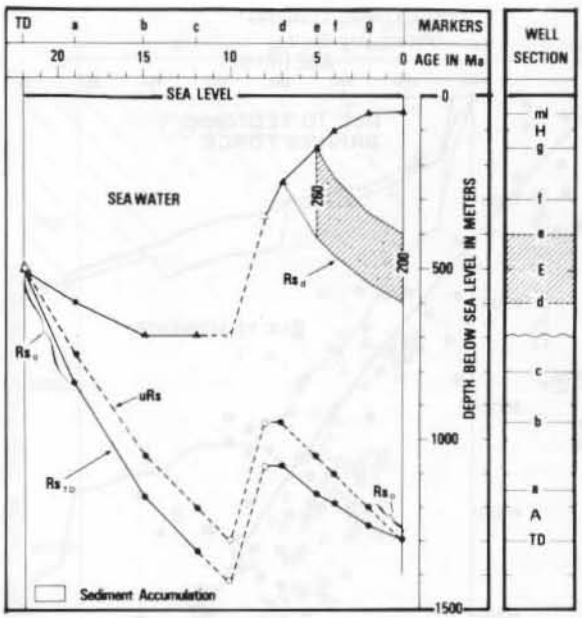


Fig. 7.9. Sea-level and sediment-accumulation curves for a hypothetical well; *a–g* are accurately dated marker horizons. *Rs* Corrected subsidence curve; *uRs* uncorrected curve. (Van Hinte 1978)

dated marker beds (a, b, c, etc., in Fig. 7.9) are indicated on the time scale along the top of the graph. In the example illustrated, an unconformity occurs at a depth of 700 m. The paleobathymetry is then determined for each marker bed, and a data point is inserted on the graph in the correct time–depth position for each marker bed (closed triangles). Extrapolation may be necessary to find the correct position of the points for total depth and for the age of the surfaces above and below the unconformity (open triangles). This can be done by estimating sedimentation rates from adjacent well-dated parts of the section, making all due allowance for the variations in sedimentation rates shown by different lithologies deposited in different depositional environments. A curve joining the data-point triangles defines the changes in water depth with time. Cumulative sediment thicknesses are then determined for each marker bed, and these values are plotted below the corresponding point on the water-depth curve (closed circles). A curve joining these points defines the subsidence history of the total-depth point in the section. Two curves are shown in Fig. 7.9. The **uRs** curve is for uncorrected subsidence, whereas the **Rs** curve is for subsidence corrected according to compaction. Van Hinte (1978) provided a method for determining compaction corrections, but Keen (1979) argued that it is not important to make this correction unless the section contains thick intervals of mudstone.

The unconformity in the section has been accounted for simply by drawing dashed lines between the open circles and triangles. This suggests that sedimentation simply ceased while water depths shallowed by about 350 m because of tectonic uplift or eustatic sea-level change. In fact, of course, an unconformity is normally a much more complicated event involving uplift and erosion, but it is usually not possible to depict this accurately, because the very nature of an unconformity involves a loss of part of the record.

If the TD point (Fig. 7.9) represents the top of the basement, the subsidence curve depicts the overall subsidence of the basin relative to sea level. Much depends on the accuracy of the biostratigraphic and paleoecological control. For the Cenozoic, bathymetric determinations from micropaleontological evidence should be accurate to within a few hundred meters but become much less accurate at bathyal and abyssal (> 2000 m) depths. Precision drops off rapidly for older rocks (Van Hinte 1978). An example of the type of evidence used in reconstructing water depths is discussed in Section 4.5.7 (Fig. 4.36).

As described at some length in Chapter 8, global sea level has varied considerably, at least during the Phanerozoic, on time scales ranging from 10^4 to 10^8 years. Time–depth plots, such as those illustrated in Figs. 4.36 and 7.9, do not show this very effectively. For bathyal or abyssal depths, eustatic sea-level changes of a few hundred meters or less would be hard to detect using available paleoecological criteria. If a sea-level change could be recognized, it would produce anomalous changes in the slope of the subsidence curve. Even these would be hard to analyze, because changes in sea level produce changes in sedimentation rate through their effects on the sediment supply (Chap. 8). At shallower depths, sea-level changes produce rather more obvious changes in microfauna, and the sediment record may contain clear evidence of transgressions and regressions, and possibly several unconformities (Chap. 6).

Van Hinte (1978) provided a hypothetical example of the reasoning that can be used to recognize a eustatic change in sea level based on stratigraphic data. Two wells are shown in Fig. 7.10. In both, a paleobathymetric break has been recognized at about 2.7 Ma. Calculation of sedimentation rates immediately above and below the break in each well suggests a time gap of 30 ka in well 1 and 600 ka in well 2. The break in well 2 clearly represents an unconformity, whereas that in well 1 may only reflect inaccuracies introduced by changes in the sedimentation rate. A simple explanation is that the paleobathymetric break represents a drop in sea level. Estimates from well 1 suggest that it amounted to about 700 ft (213 m).

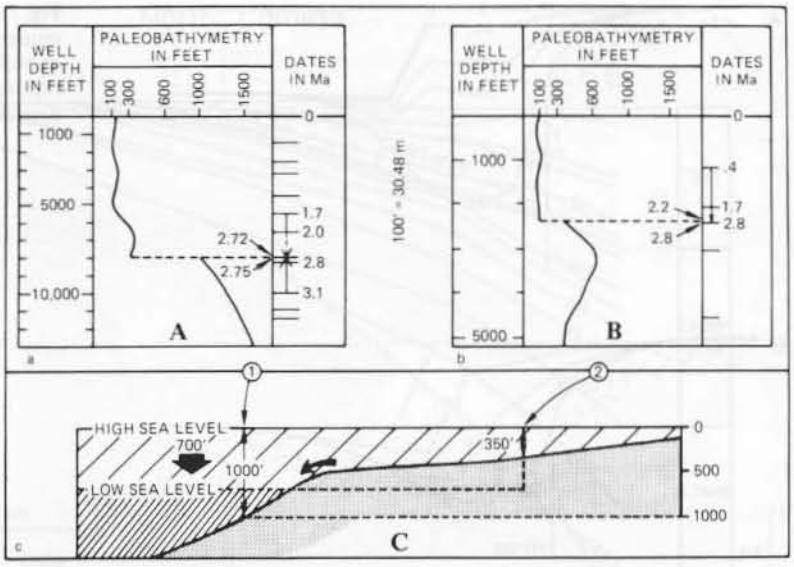


Fig. 7.10 A–C. Paleobathymetry of two hypothetical wells (A, B) showing a paleobathymetric break in each and estimation of the ages of sediments above and below the break by extrapolation of sedimentation rates. In C, the data are interpreted in terms of a eustatic sea-level drop. See text for further explanation. (Van Hinte 1978)

7.3.3
Refinements of the Basic Extensional Model

One of the most important refinements in the basic McKenzie model is that suggested by Royden et al. (1980). They noted that, on some continental margins, there is evidence for an early phase of intrusion of mantle diapirs and dikes during continental separation. Not only does the intrusion of dense mafic and ultramafic material result in considerable subsidence early in basin development, but there are also significant thermal consequences of such a process. The differences between the McKenzie (1978) model and the Royden et al. (1980) model were discussed by Dewey (1982) and are not elaborated here.

Development of the basic extensional model has provided insights into crustal behavior that can be used to generate a variety of synthetic basin models. These serve as valuable checks on interpretations of the evolution of sedimentary basins. The techniques have been described by several workers, including Watts (1981, 1989) and Watts et al. (1982), and a few examples are discussed here.

Distinctive features of many continental-margin extensional basins include basement onlap toward the continental margin (Fig. 7.4) and oceanward progradation and downlap. The origin of the onlap was controversial during the late 1970s and the 1980s, as it was variously attributed to a simple increase in accommodation as a result of eustatic sea-level rise by many seismic stratigraphers (especially Vail et al. 1977) and to thermally driven flexural subsidence by other geophysicists (Watts 1981). This debate was, for a while, at the center of the controversy regarding the importance of eustatic sea-level change in the generation of sequence boundaries (Chap. 8). Watts (1981) attributed onlap to the increase in the flexural rigidity of the crust with time. This model predicts that, as sediment is added to the basin, the load is compensated by subsidence of a progressively wider but shallower depression of the crust (Fig. 7.4). Watts (1981) illustrated two simple models of a continental margin, which assume different stratigraphic behaviors under similar conditions of gradually increasing flexural rigidity (Fig. 7.11). The upbuilding model is typical of carbonate-dominated environments, whereas the outbuilding model shows the characteristic clinoform progradation (Sect. 5.4.1) of a clastic-dominated environment. In the upbuilding model, the load is continuously added in the same place, and a virtually horizontal shelf develops. Deeper (outboard) sediments are back-tilted toward the shoreline. In the outbuilding model, the load is progressively shifted seaward, and the back-tilting effect is less pronounced. Figure 7.12 is an enlargement of the landward edge of one of these models, showing how the increase in the flexural rigidity of the crust with age that is built into this model leads to the spreading of the sediment and water load over progressively wider areas of the crust, causing coastal onlap. A comparison of this model with a cross-section through an actual coastal plain (Fig. 7.13) shows that the model successfully mimics real stratigraphic architecture. This has some important implications, because coastal onlap is one of the key indicators used to postulate rise in sea level in seismic stratigraphic models of sea-level change.

It is important to note that the process of thermal relaxation modeled by Watts (1981) takes tens of millions of years, whereas the coastal onlap described by

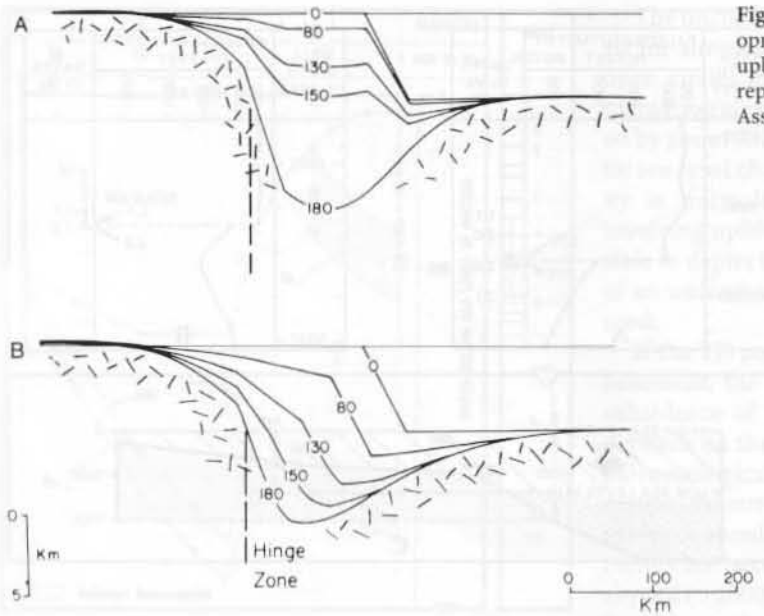


Fig. 7.11 A, B. Two simple models for the development of a continental margin, based on **A** upbuilding and **B** outbuilding. (Watts 1981; reprinted with permission of the American Association of Petroleum Geologists)

Vail et al. (1977) is that of "third-order" cycles that have durations in the million-year range. They are, therefore, distinctly different processes, and the basin analyst needs to be aware of both when interpreting stratigraphic architectures. A tectonic mechanism that generates onlap and offlap at "third-order" rates (million-year episodicity) is discussed in Section 8.2.3.1.

White and McKenzie (1988) argued that the flexural rigidity of continental crust does not, in fact, increase significantly with time, and suggested that a model involving differential stretching of the lithospheric mantle and the crust would generate the observed onlap. They also pointed out that the vertical component of onlap in some basins is much too great to be accounted for by sea-level change alone; for example, in the North Sea Basin (Fig. 7.4B) it totals 2 km. Watts (1989) showed that, by varying the rate of change of flexural rigidity and by incorporating a pulsed sediment supply over periods of millions to tens of millions of years, a variety of basinal architectures could be modeled. Onlap attributable to changes in flexural rigidity is more likely to be significant on young continental margins, because the rate of increase in elastic thickness of the crust is greater in such settings, reflecting the presence of a larger ther-

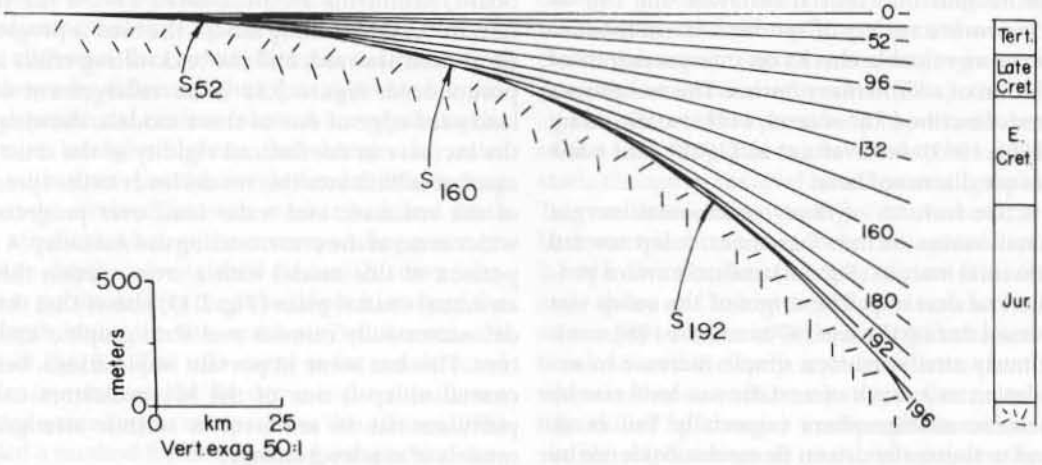


Fig. 7.12. Enlargement of the landward edge of the stratigraphy generated in flexural models, such as those illustrated in Fig. 7.11. Note the progressive onlap of younger sediments onto the coastal plain. (Watts 1981; reprinted with permission of the American Association of Petroleum Geologists)

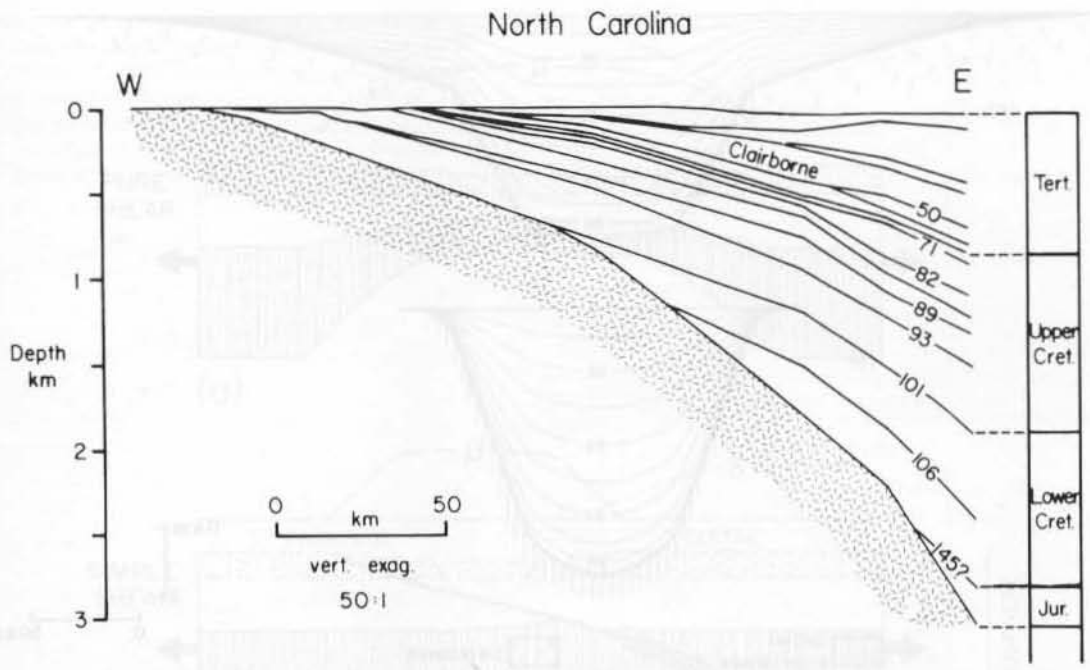


Fig. 7.13. Observed stratigraphy of the Atlantic coastal plain of North Carolina. Note the coastal onlap of Cretaceous strata and compare this with the synthetic model of Fig. 7.12. The significance of this is discussed in the text. (Watts 1981; reprinted with permission of the American Association of Petroleum Geologists)

mal anomaly. Some onlap is always likely to occur as a result of flexural loading (assuming a steady sediment supply) given the effectively limitless accommodation for lateral outbuilding of a continental margin facing a steadily widening ocean. Given the complex interplay of tectonic, eustatic, and other forces acting to generate the broad architecture of extensional basins, it is now recognized that it is simplistic to interpret eustatic sea-level changes directly from patterns of onlap and offlap, as initially suggested by Vail et al. (1977; Chap. 8).

Bond and Kominz (1984) modeled the shape and subsidence history of the North American divergent margin of southern Canada using the techniques of Watts (1981) and compared it with a palinspastically restored cross-section through this margin. They found that the flexural model could not account for the great thickness of the lower Paleozoic section actually preserved, nor for the presence of a thin wedge of sediments extending hundreds of kilometers beyond the margin onto the cratonic interior. The thickness problem probably relates to the use of incorrect parameters for crustal thickness and thermal properties and to inadequate information concerning the geometry and subsidence history of the westernmost, now-deformed part of the margin. The existence of the extensive cratonic cover beyond the limits of the basin is probably the result of regional or global sea-level rise and consequent transgression. That sea-level change can be deduced from flexural basin models is important and is discussed further in Chapter 8.

Walcott (1970b) discussed the question of whether the lithosphere is an elastic or a viscoelastic solid. His results suggest that, for short-term loads, the lithosphere behaves in an elastic manner, whereas, over geologically long-term periods (10^6 to 10^9 years), the response to loading is viscoelastic. This question was re-evaluated by Watts et al. (1982) using the basin-modeling techniques discussed above. They examined various combinations of thermal and stretching character and different configurations of flexural rigidity and viscoelastic strain release. Two of their synthetic models are illustrated in Fig. 7.14. Model (a) is a simple elastic model, in which the initial subsidence is based on stretching by a factor of 2 ($\beta=2$), with the applied load compensated by Airy-type isostasy. Subsequent basin fill is based on elastic flexure and shows the characteristic onlap of the basin margin as flexural rigidity increases with time or as the load increases with time and the flexural wavelength increases. The viscoelastic model (b) commences with the same stretching event, but the increasing elastic thickness of the plate with time caused by cooling is virtually counteracted by viscous stress release, and a narrow, deep basin results, with minimal marginal onlap. Most actual basins are closer in architecture to model (a), which is similar in

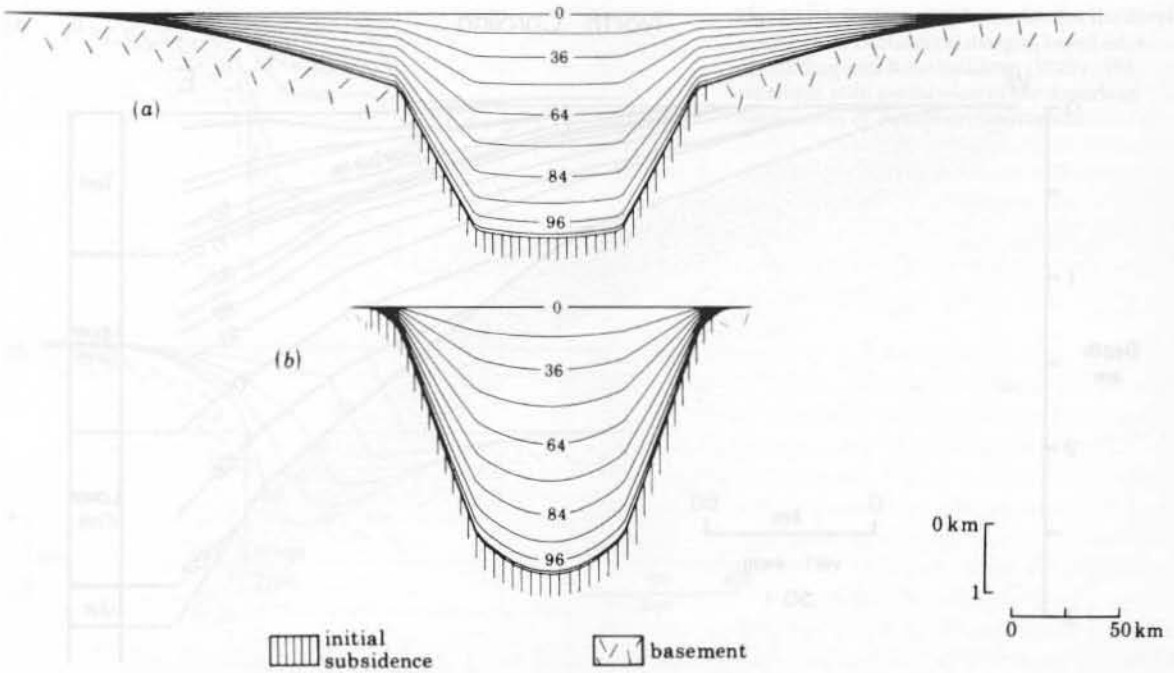


Fig. 7.14 A, B. Synthetic configurations of extensional basins. Initial subsidence is based on a stretching and Airy-type isostatic loading response, with later subsidence and sedimentation during the thermal relaxation phase calculated using an **A** elastic- and a **B** viscoelastic-plate model. (Watts et al. 1982)

shape to the Steer's-Head model of Fig. 7.4. Watts et al. (1982) concluded that elastic models can satisfactorily explain most extensional basins, but that local crustal conditions at the time of loading need to be examined. As we shall see, some foreland basin models indicate a viscoelastic response, so this question is far from settled.

The McKenzie (1978) model of continental stretching assumes uniform extension throughout the lithosphere, independent of depth. However, Royden and Keen (1980) have demonstrated that this may not be correct. The crust (upper lithosphere) deforms by brittle failure, whereas the lower lithosphere deforms by ductile flow. They modeled this behavior assuming a decoupling of the upper and lower lithosphere at a level assumed to be located at the crust–mantle boundary. Crustal extension, δ, was commonly found to be less than subcrustal extension, β. More recent models of extensional basin subsidence now take this two-layer model of lithospheric rheology into account (Watts and Thorne 1984).

The different amounts of extension in the upper and lower lithosphere create a space problem, a difficulty which was not fully explored by Royden and Keen (1980). However, later work on extensional tectonics has suggested a possible solution. The Basin and Range province of the southwestern United States is an area of extensional tectonics that has yet to lead to the development of oceanic crust. Excellent exposure and deep-penetration seismic lines across this area have provided a picture of extensional structural geology very different from the so-called **pure-shear** model of McKenzie (1978). It is suggested that the many gently dipping, commonly listric faults that characterize this area merge downward into a master detachment surface that dips gently through the entire crust and into the lower lithosphere. This is the **simple-shear** model of lithospheric extension (Wernicke 1985). In the pure-shear model, upper and lower lithospheric extension are assumed to take place over the same part of the rifted margin, whereas, in the simple-shear model, the distribution of extension is quite different (Fig. 7.15). Application of this model to specific continental margins is an area of active current research, as discussed in Chapter 9.

Loup and Wildi (1994) analyzed subsidence of the Paris Basin and compared it to other northwest European basins. None shows the simple concave-upward subsidence-versus-time curve that would be expected from a simple extensional basin. Most are characterized by short concave segments with periods of slow or no subsidence, intervals of accelerated subsidence, and intervals of uplift. Few of the events are synchronous over wide areas of northwest Europe. These results are interpreted as the product of rifting, transtension, and basin inversion resulting from regional plate-tectonic movements and subcrustal processes, including mantle plumes, to-

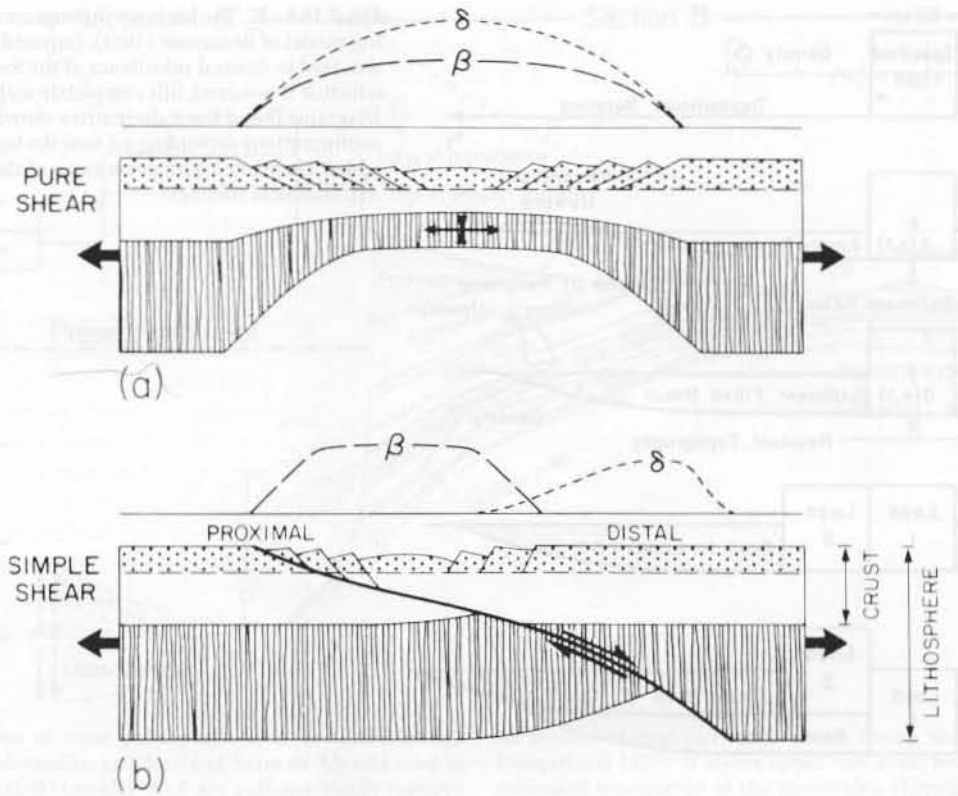


Fig. 7.15. A Pure shear and B simple shear models of lithospheric extension. The distribution of β and δ is shown schematically. (Keen et al. 1987, based on Wernicke 1985)

gether with intraplate stresses resulting from all these processes.

A recent review of the mechanics of passive-margin subsidence was provided by Bott (1992). Bond and Kominz (1988) reviewed the evolution of our knowledge of extensional continental margins from the beginnings of geosyncline theory in the middle of the 19th century.

7.4 Basins Produced by Supracrustal Loading

7.4.1 Evolution of the Basic Model

Barrell (1917) was the first to realize that "the thick nonmarine strata of the Gangetic plains accumulated in space made available by subsidence of the Indian crust beneath the mass of thrust plates of the Himalayan Range" (Jordan 1995). Price (1973) revived the concept of regional isostatic subsidence beneath the supracrustal load of a fold–thrust belt that generates the marginal moat we now term a **foreland basin**. It is clear that the crust must have mechanical strength for a wide foredeep, such as the Alberta Basin or the Himalayan foreland basin, to be created. Airy-type isostatic response to the supracrustal load cannot explain the formation of these basins. The classic architecture of a foreland basin is defined by the isopachs of the sediment fill, which is that of an asymmetric lozenge, with a depocenter adjacent to the location of the crustal load, tapering along the strike and also thinning gradually away from the orogen towards the craton. Good examples of this pattern are illustrated in Figs. 9.42A and 9.71.

Beaumont (1981) and Jordan (1981) were the first to propose quantitative flexural models for foreland basins, constraining the models with detailed knowledge of the structure and stratigraphy of the studied basins. Their papers have become the standard references on the subject. Beaumont's model is based on the incremental addition of loads of estimated height and density to a grid of 50 × 50 km squares in the area of the fold-thrust belt, followed by calculation of the predicted flexural response using different values of elastic thickness and flexural rigidity (Figs. 7.16 and 7.17). Actual load heights were estimated from the history of thrust faulting in the southern Rocky Mountains. Both elastic and viscoelastic models were attempted. In the first, J, the relaxation time, is infin-

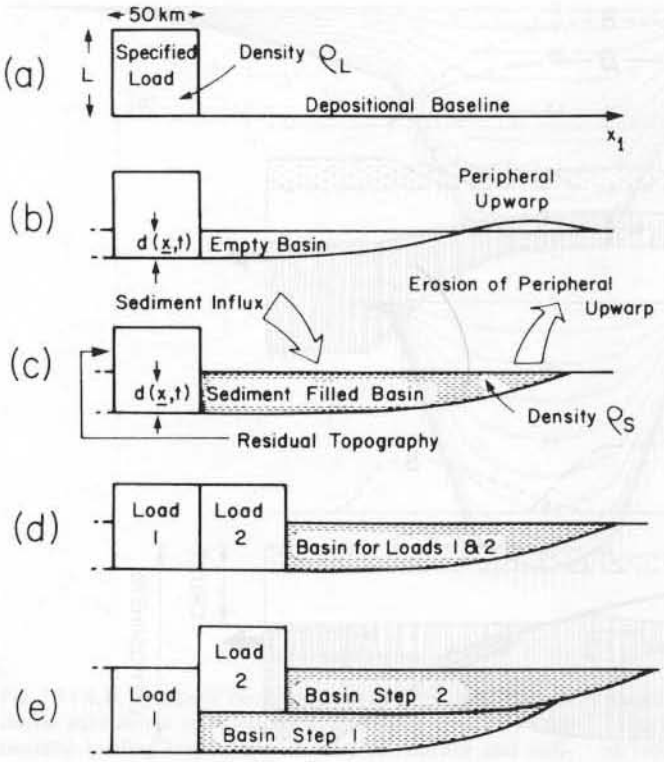


Fig. 7.16 A–E. The basis for the supracrustal-loading model of Beaumont (1981). Imposed loads 1, 2, etc., lead to flexural subsidence of the foredeep, which, it is assumed, fills completely with sediment. Diagrams **D** and **E** are alternatives showing different configurations depending on how the loads are added. (Reprinted with permission of the Royal Astronomical Society)

ity whereas, in the viscoelastic models, J is some value less than infinity. Some of the results of these experiments are summarized below.

Beaumont (1981) constructed three cross sections through the foreland basin and adjacent fold–thrust belt. One of these, the most stratigraphically complete, is shown in Fig. 7.18. Attempts to synthesize this cross section formed the main part of Beaumont's experiments, and the results were then cross checked against the other two sections. Two broad, coarse, nonmarine to shallow-marine syntectonic clastic wedges (molasse; Sects. 9.1, 9.3.4.2) were formed in the area represented by the illustrated section. These are the Kootenay-Blairmore succession and the Belly River-Paskapoo succession. Beaumont (1981) pointed out that each records episodes of uplift and nondeposition. Each of the four named units are, therefore, thought to have been caused by separate episodes of thrust faulting to the west. These dated periods of loading alternating with episodes of tectonic quiescence (uplift, erosion) formed the basis for the increments used to create the foreland basin in the model experiments.

A preliminary approximation regarding flexural rigidity can be made by comparing the three elastic models of Fig. 7.17 to the actual width and depth of

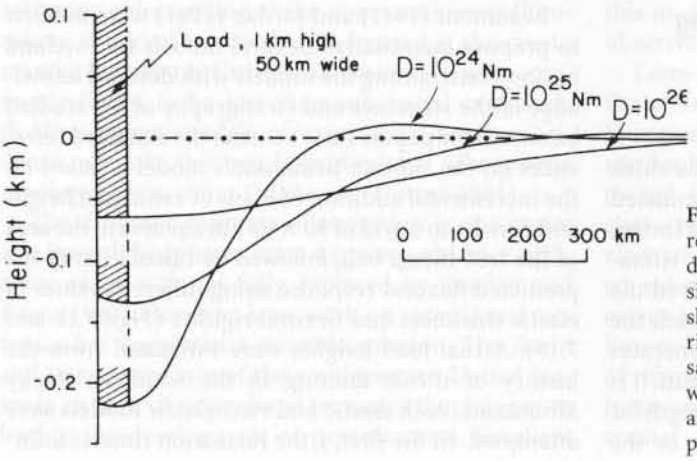


Fig. 7.17. Basins formed on an elastic lithosphere in response to a load 1 km high and 50 km wide, with a density of 2400 kg/m^3. Sediments of the same density fill the foredeep. Three basin configurations are shown, corresponding to three conditions of flexural rigidity. Each of these basins has approximately the same cross-sectional area but differs in the degree to which the load is spread out over greater surface areas of the crust. (Beaumont 1981; reprinted with permission of the Royal Astronomical Society)

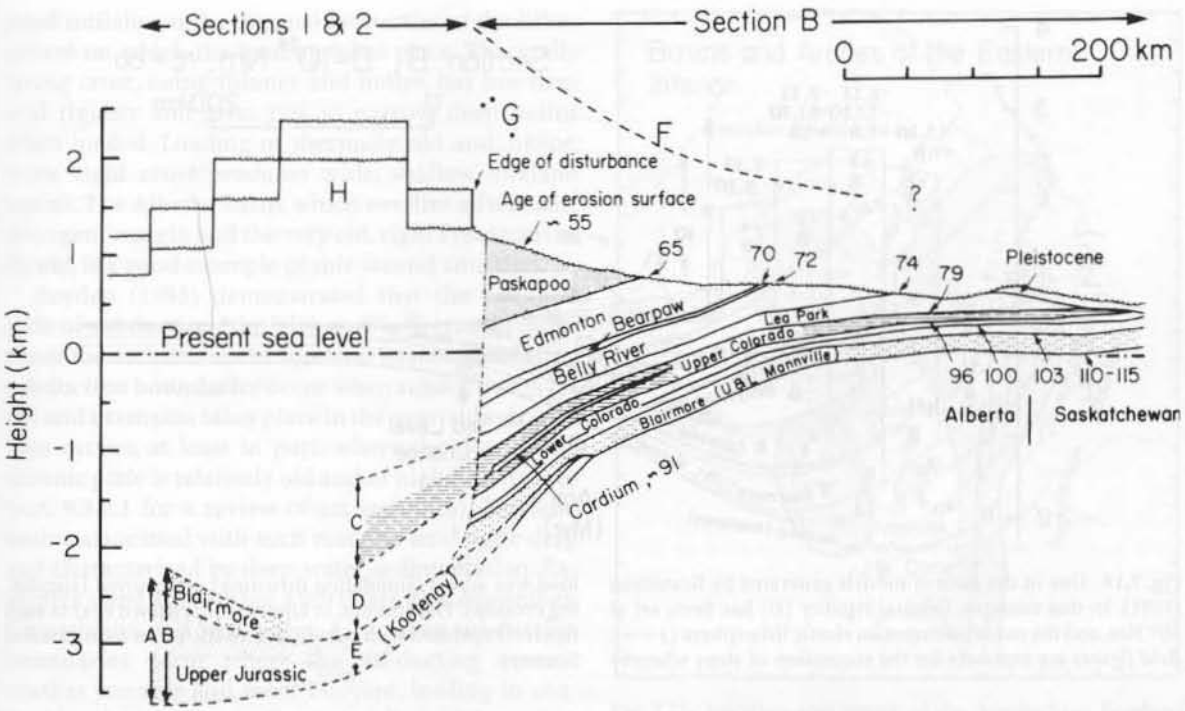


Fig. 7.18. One of three palinspastic cross-sections through the Rocky Mountains and foreland basin of Alberta used in Beaumont's (1981) model. *A*–*E* are palinspastically restored stratigraphic thicknesses. *F* and *G* are estimates of the original depositional surface prior to late Cenozoic erosion, based on studies of coal moisture content (using the method of Hacquebard 1977). *H* shows upper and lower bounds on the estimated topography of the mountains. (Reprinted by permission of the Royal Astronomical Society)

the Alberta Basin. Such a comparison suggests that a rigidity on the order of 10^{24}–10^{25} Nm is the most appropriate. A more rigid crust would lead to a much shallower and wider basin than is actually observed. A detailed example of one of these elastic models is given in Fig. 7.19. Here, the existence of all the stratigraphic units is correctly predicted, but thicknesses are inaccurate, the units mostly dip the wrong way in the deepest part of the basin, and several of the younger units, e.g., the Belly River-Edmonton, incorrectly pinch out toward the eastern margin of the basin. These defects arise because the center of the basin does not subside deeply enough. An example of a viscoelastic model is given in Fig. 7.20. The viscous relaxation generates a somewhat deeper basin, more closely approximating the actual palinspastically restored configuration shown in Fig. 7.18. The match of thicknesses and dips with the actual configurations is better, in this case.

Eustatic sea level was not a primary input into these models, but Beaumont's modeling methods permitted various sea levels to be tested to see which best fit the data. It was found that sea levels tens of meters higher than those of the present day are required, as shown by the values for sea-level change in Figs. 7.19 and 7.20.

Figure 7.17 shows three configurations for the forebulge, depending on variations in flexural rigidity. As viscous relaxation takes place, effective rigidity is reduced, and the forebulge migrates basinward and becomes steeper-sided and narrower. Beaumont touched on this point in his 1981 paper, pointing out that the Sweetgrass Arch of southern Alberta and northern Montana has the right configuration and position to be the forebulge of the Alberta Basin. In a later paper, discussed below, Quinlan and Beaumont (1984) addressed the behavior of the forebulge in detail and showed that a study of the structure and stratigraphy of the sediments overlying the forebulge can serve as a check on the validity of quantitative foreland-basin models.

The preferred viscoelastic model generated by Beaumont (1981) for the Alberta foreland basin contrasts with the elastic model, which seems to be more suitable for extensional basins, and it also contrasts with the elastic model used in foreland basin studies by Jordan (1981) and Hagen et al. (1985). This conflict regarding the physics of continental lithosphere may, however, be more apparent than real. As Beaumont (1981) pointed out, crustal properties depend on the local geological history and will vary markedly with age, thickness, and thermal history. Basin reconstruc-

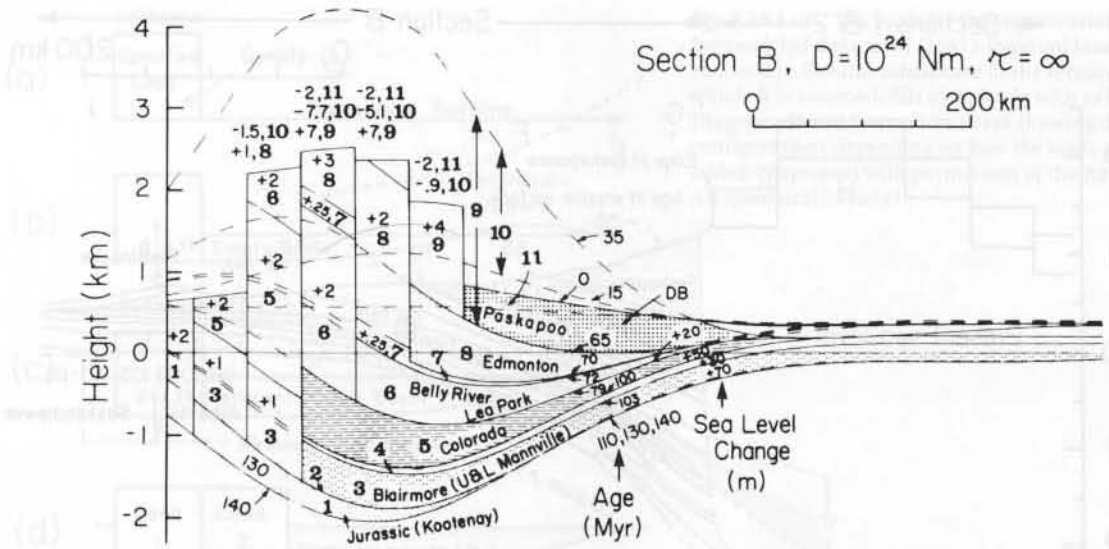


Fig. 7.19. One of the suite of models generated by Beaumont (1981). In this example, flexural rigidity (D) has been set at 10^{24} Nm, and the model assumes an elastic lithosphere ($\tau = \infty$). *Bold figures* are numbers for the succession of steps whereby load was added (simulating thrusting) or removed (simulating erosion). The amount, in kilometers, is shown next to each figure. (Reprinted with permission of the Royal Astronomical Society)

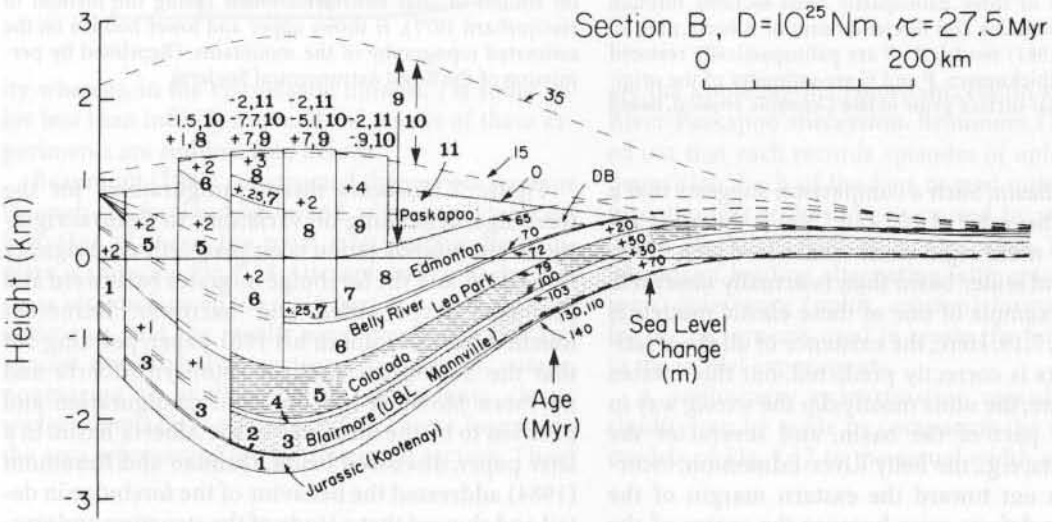


Fig. 7.20. A viscoelastic model for the Alberta Basin. In this example, τ, the relaxation time, is set at 27.5 Ma. (Beaumont 1981; reprinted with permission of the Royal Astronomical Society)

tions dealing with geologically short time periods, say less than 10 m.y., may satisfactorily be based on an elastic model because of the lengthy time necessary for viscous relaxation to become measurable. Much more work on individual basins is necessary before a full understanding of lithospheric behavior will be possible.

7.4.2 Refinements of the Basic Supracrustal Loading Model

Recent work on the geodynamics of retroarc foreland basins was summarized by Jordan (1995) and DeCelles and Giles (1996), and the tectonics and sedimentation of collision-related foreland basins was described by Miall (1995). Beaumont et al. (1982) and Stockmal et al. (1986) showed that the width and depth of basins produced by supracrustal loading de-

pend initially on the thermal properties of the lithosphere on which the loading takes place. Thermally young crust, being thinner and hotter, has low flexural rigidity and gives rise to narrow, deep basins when loaded. Loading of thermally old and, hence, more rigid crust produces wide, shallow foreland basins. The Alberta Basin, which overlies a Paleozoic divergent margin and the very old, rigid Precambrian Shield, is a good example of this second situation.

Royden (1993) demonstrated that the rate and style of subduction (the "slab-pull" effect) also influences the architecture of foreland basins. **Retreating subduction boundaries** occur when subduction is rapid and extension takes place in the over-riding plate. This occurs, at least in part, where the downgoing oceanic plate is relatively old and of high density (see Sect. 9.3.2.1 for a review of arc tectonics). Foreland basins associated with such margins tend to be deep and characterized by deep-water sedimentation. Examples cited by Royden (1993) are the Apennines, Carpathians, and Hellenides. **Advancing subduction boundaries** occur where the subducting oceanic crust is younger and more buoyant, leading to contractional tectonics in the overriding plate and to shallower foreland basins filled with shallow-marine to nonmarine sediments, as in the Andean, Alpine, and Himalayan basins. Variations in subsidence rate and in the architecture of the evolving basin may reflect heterogeneities in the basement (Waschbusch and Royden 1992).

An ambitious attempt to extend Beaumont's (1981) two-dimensional foreland-basin model into three dimensions was presented by Quinlan and Beaumont (1984) and improved and expanded by Beaumont et al. (1988). The elegant studies of the Appalachian foreland basin by Beaumont and his colleagues (Fig. 7.21) confirmed the basic characteristics of the model discussed above and paid considerable attention to the dynamics of forebulges and the relation of foreland basins to contemporaneous intracratonic basins. The flexural relationships among these various tectonic elements is modeled in Figs. 7.22 and 7.23, based on the presumed elastic behavior of the crust. It can be seen that, where foreland and intracratonic basins are sufficiently separated, the forebulges adjacent to these depressions interact to produce higher or wider uplifts. Where the basins are close together, however, the forebulges are out of phase and depressed. These details were incorporated in the computer models of Quinlan and Beaumont (1984) and can be tested by careful study of the stratigraphy and structure of the Paleozoic rocks of the area, as discussed below.

Quinlan and Beaumont (1984) employed a temperature-dependent, viscoelastic lithospheric model. Viscous flow, leading to relaxation, occurs first in

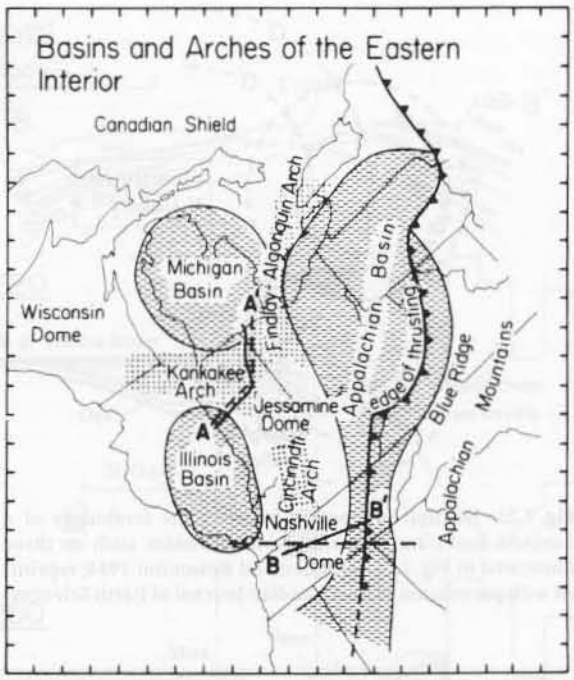


Fig. 7.21. Location and extent of the Appalachian Foreland Basin and associated cratonic basins and arches. The edge of thrusting is the outcrop trace of the edge of the deformed belt. (Quinlan and Beaumont 1984; reprinted with permission of the Canadian Journal of Earth Sciences)

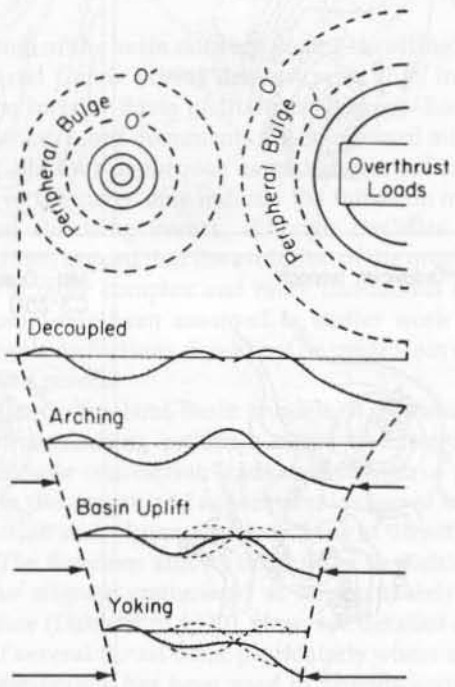


Fig. 7.22. Flexural interaction between a foreland basin (*right*), a peripheral forebulge, and an intracratonic basin. (Quinlan and Beaumont 1984; reprinted with permission of the Canadian Journal of Earth Sciences)

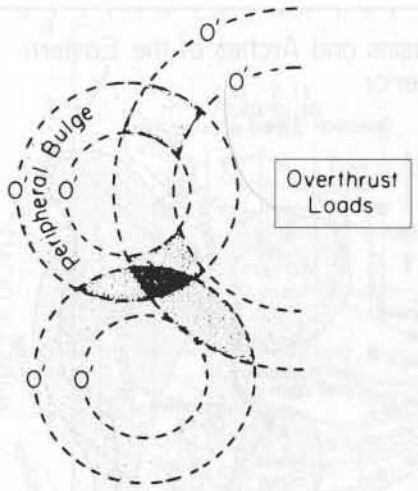


Fig. 7.23. Flexural interaction between the forebulges of a foreland basin and two intracratonic basins such as those illustrated in Fig. 7.21. (Quinlan and Beaumont 1984; reprinted with permission of the Canadian Journal of Earth Sciences)

the lower, hotter part of the lithosphere, with relaxation gradually propagating upward into cooler and more viscous regions. They divided mid-Ordovician to Permian time into 14 time steps and the foreland basin, intracratonic basins and thrust belt into a total of 20 × 23 grid squares, each 88.9 × 88.9 km. Loading, subsidence, erosion, and sedimentation were calculated using Beaumont's (1981) method. The Michigan and Illinois basins were assumed to subside at a specified exponential rate.

Figure 7.24 shows part of a series of maps produced as output by the model. On the left are actual maps of the foreland basin, and in the center and on the right are four syntheses developed using two different models of crustal behavior. The squares on the right of these four maps are the grid squares used to model supracrustal loading, with load figures given in kilometers. The amounts of these loads were determined by trial and error to best explain the stratigraphic record. Figure 7.25 shows a set of cross sections through

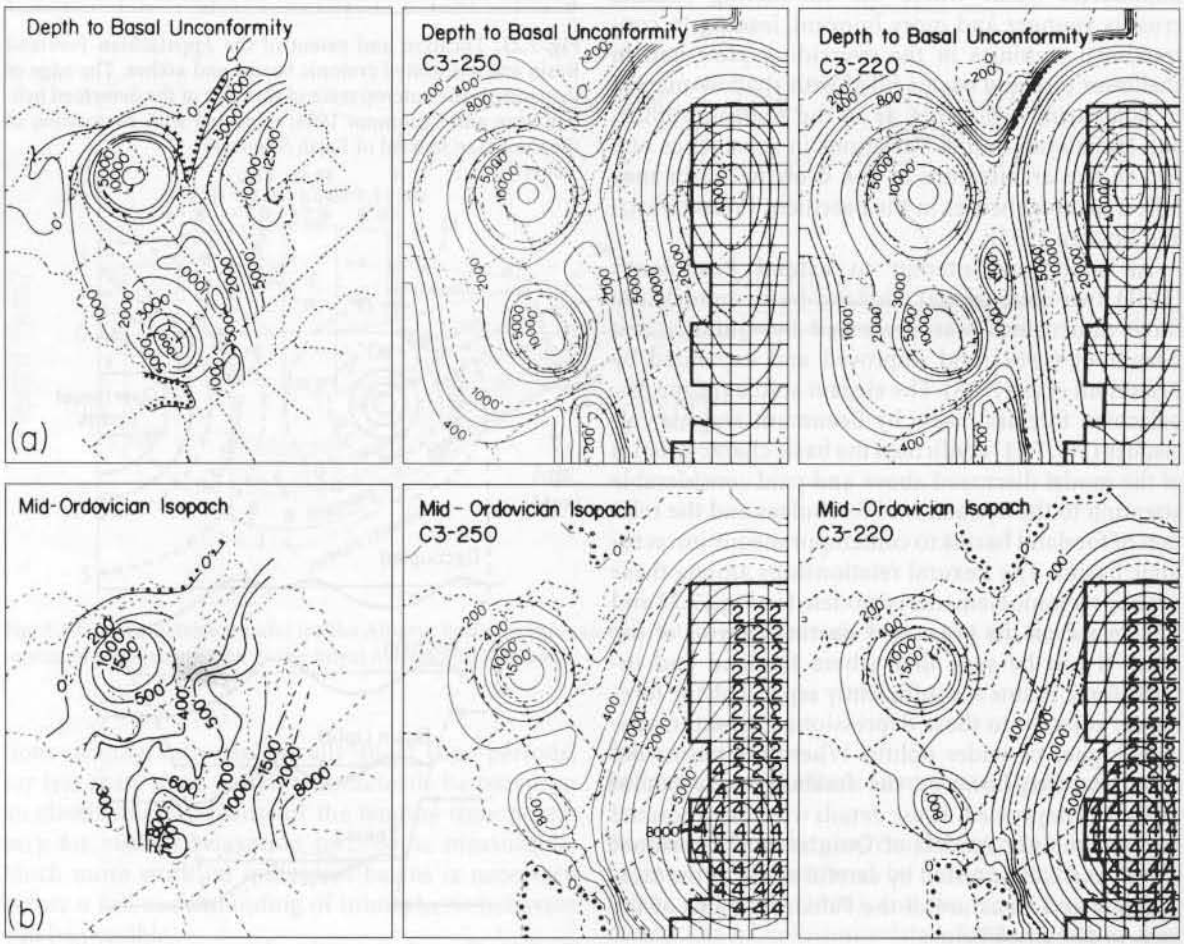


Fig. 7.24. Two sets of maps produced by Quinlan and Beaumont (1984) comparing the results of two model experiments with actual structural and isopach maps documenting the early stages of the Taconic overthrusting event. (Reprinted with permission of the Canadian Journal of Earth Sciences)

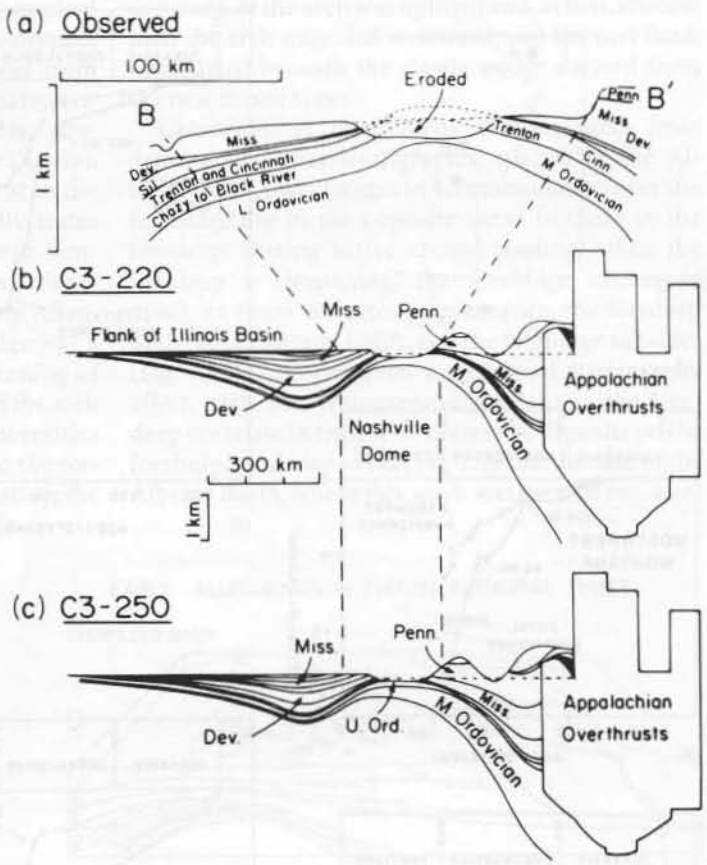


Fig. 7.25. Cross section through the Nashville Dome (*a*), compared with two synthetic cross sections (*b*, *c*) produced by Quinlan and Beaumont. (1984; reprinted with permission of the Canadian Journal of Earth Sciences)

one of the intracratonic arches, which is treated as a forebulge in the model.

DeCelles and Giles (1996) subdivided foreland basins into four "depozones", the **wedge-top** (or **thrust-top**), **foredeep**, **forebulge**, and **back-bulge** areas. The wedge-top zone is located above thrust ramps at the leading edge of the fold–thrust belt, which may include "blind" structures, that is, those that do not outcrop at the surface (Figs. 9.69, 9.70). These four zones are not fixed in position but migrate in response to the changing flexural effects of fold–thrust-belt tectonism, as noted below.

As shown by Van Hinte (1978), the shape of subsidence curves produced by backstripping or geohistory analysis can be indicative of subsidence behavior and basin type. Subsidence rates in foreland basins may be an order of magnitude faster than in the thermal subsidence stage of extensional basins, though they are comparable to the subsidence rates of initial rifting (Allen et al. 1986). Examples of subsidence curves for foreland basins were given by Cross (1986) and Homewood et al. (1986), and are illustrated in Fig. 7.26. Allen et al. (1986) noted that many (but not all) foreland basins yield convex-upward curves, indicating increasing rates of subsidence at a given point in the basin. This reflects the migration of the basin outward during thrusting. DeCelles and Currie (1996) demonstrated that, in the Western Interior Basin of Utah, the Jurassic–Eocene subsidence record documents the cratonward migration of the four depozones named above. Inflection points in the curve may indicate the initiation of individual thrusting events, although DeCelles and Giles (1996) argued that the evolution of the orogenic wedge is more complex and more continuous than had commonly been assumed in earlier work and that simple deductions should not be made from such inflection points.

In classic foreland-basin models, it is assumed that thrust-faulting gradually steps basinward, in part, because contraction leads to expulsion of pore fluids in the orogen and consequent increased internal friction and, ultimately, the locking of thrust surfaces. The foredeep and its orogen-fed depositional systems migrate cratonward at approximately the same rate (Dahlstrom 1970). However, detailed analysis of several thrust belts, particularly where magnetostratigraphy has been used to provide accurate dates of proximal foreland-basin sediments, has shown that this is not always the case; **out-of-sequence thrusting** may occur, leading to more complex patterns of basin fill, as in the Himalayan

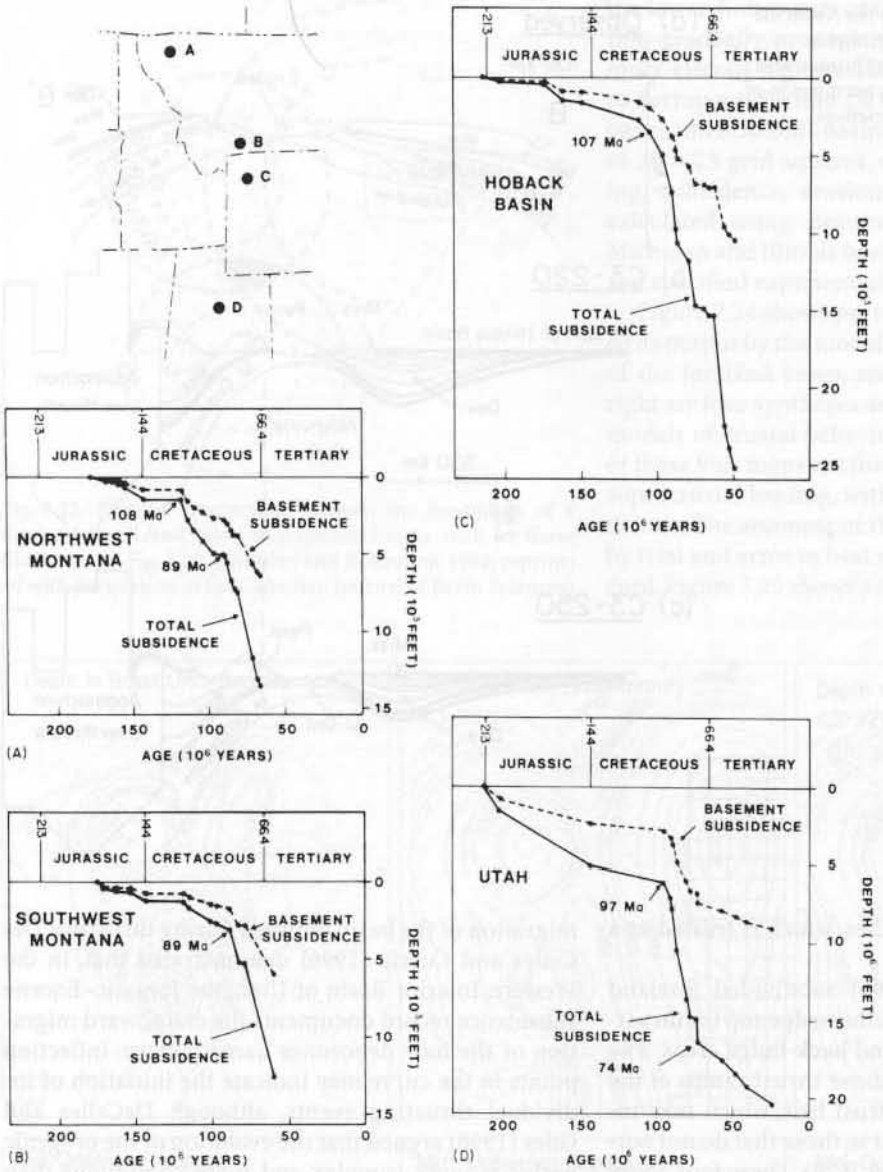


Fig. 7.26 A – D. Geohistory plots for four locations in the Mesozoic Sevier foreland basin of the western United States. (Cross 1986)

foreland basin of northern Pakistan (Burbank and Raynolds 1984, 1988).

It is a common assumption that the coarse, commonly conglomeratic top to these stratigraphic pulses records the time of maximum tectonism (the **syntectonic model** of sedimentation), but this may be incorrect. It depends (among other factors) on which response to thrusting and supracrustal loading is most rapid: crustal flexure and subsidence or the development of an organized drainage net to carry detritus from uplifted source areas. In another scenario, the so-called **antitectonic model**, the time of thrusting and crustal loading may be one of basin deepening, with the deposition of fine-grained deposits, the coarsening-upward succession recording the gradual establishment of drainage nets, and the building out of coastal-plain depositional systems, possibly long after the triggering pulse of tectonism has ceased (Heller et al. 1988; Heller and Paola 1992). Other scenarios are possible, depending on the flexural response to loading (which depends on crust thickness and age), rates of erosion and transportation (climate and source-area variables), and local paleogeography, which governs the ultimate distribution of the detritus (e.g., removal from the basin by through-going rivers or entrapment by basement topography). The relationships among these variables are complex, and researchers should beware of drawing simplistic conclusions (Jordan et al. 1988; Heller and Paola 1992).

During periods of tectonic quiescence, gradual viscous relaxation may cause the foredeep to deepen, while the forebulge migrates basinward, away from the craton, causing the foredeep to become narrower. The stratigraphy in the area of the forebulge is a sensitive indicator of these events, as shown by Quinlan and Beaumont (1984) and Tankard (1986a, b) in the case of the Appalachian Basin. Figure 7.27 illustrates the evolution of the Waverley Arch of eastern Kentucky from a late Mississippian (late Acadian) relaxation phase to a middle Pennsylvanian (early Allegheny) phase of thrust flexure. The forebulge was a site of carbonate sedimentation at the beginning of this period but, during relaxation, the crest of the arch migrated eastward, causing multiple unconformities and the onlapping of carbonate wedges onto the western flank of the arch. During renewed thrusting, the east flank of the arch was uplifted and, at first, eroded; then the arch migrated westward, and the east flank was buried beneath the clastic wedge derived from the new thrust sheet.

Catuneanu et al. (1997b) demonstrated, from detailed sequence-stratigraphic studies in the Alberta Basin, that changes in accommodation over the forebulge are in the opposite sense to those in the foredeep. During active crustal loading, when the foredeep is deepening, the forebulge undergoes uplift. At times of tectonic quiescence, the foredeep undergoes isostatic uplift, and the forebulge subsides (Fig. 7.28). This generates a **reciprocal stratigraphy** effect, such that transgressive deposits in the foredeep correlate in time with regressive deposits on the forebulge, and vice-versa (Fig. 7.29). In the case of the Alberta Basin, where this work was carried out, a se-

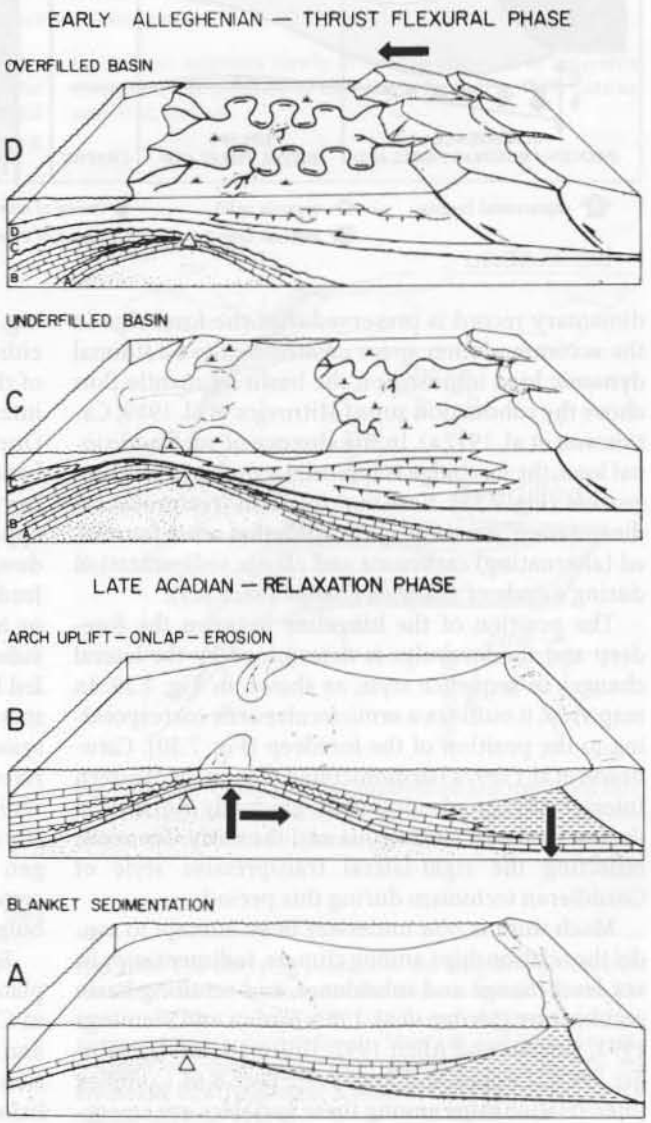


Fig. 7.27 A – D. Mississippian–Pennsylvanian evolution of the Waverley Arch of eastern Kentucky, part of the forebulge adjacent to the Appalachian foreland basin. (Tankard 1986b)

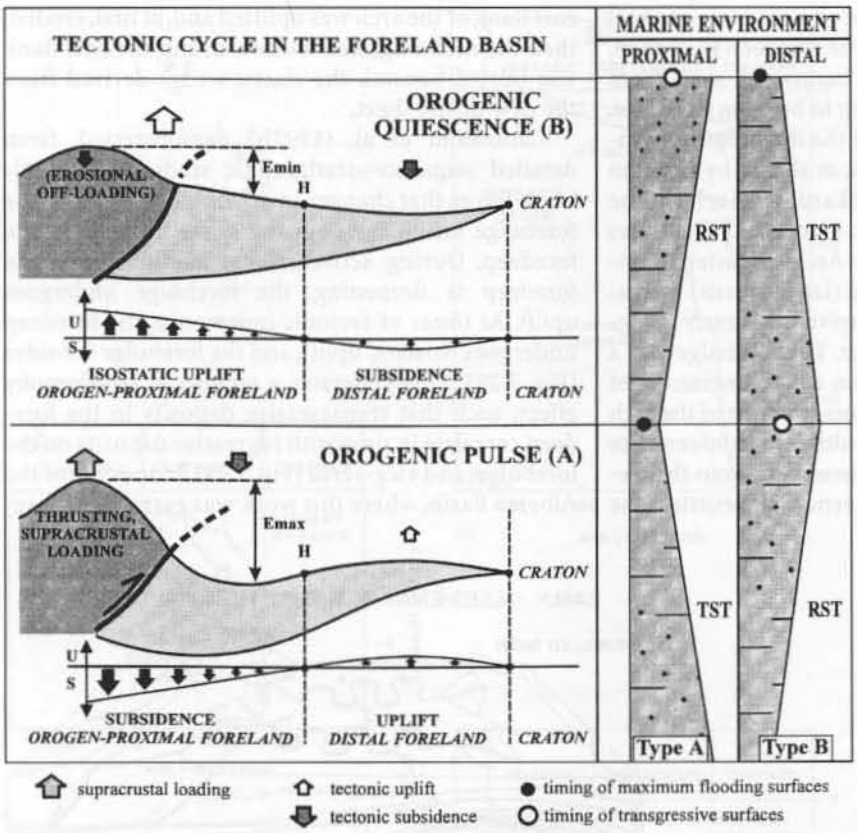


Fig. 7.28. Tectonic model of reciprocal sedimentation in foreland basins. *TST* Transgressive systems tract; *RST* regressive systems tract. (Chap. 6; Catuneanu et al. 1997b)

dimentary record is preserved over the forebulge in the accommodation space created by the additional **dynamic load** imposed on the basin by mantle flow above the subduction zone (Mitrovica et al. 1989; Catuneanu et al. 1997a). In the absence of such additional load, the forebulge is typically a zone of uplift and erosion (Fig. 7.25). Note that the term "reciprocal sedimentation" is used in quite a different sense for mixed (alternating) carbonate and clastic sedimentation during a cycle of sea-level change (Sect. 6.3).

The position of the hingeline between the foredeep and the forebulge is determined by the lateral changes in sequence style, as shown in Fig. 7.29. In map view, it outlines a semicircular area corresponding to the position of the foredeep (Fig. 7.30). Catuneanu et al. (1997a) demonstrated that, in the Western Interior, the foredeep shifted gradually northward during the latest Cretaceous and the early Cenozoic, reflecting the right-lateral transpressive style of Cordilleran tectonism during this period.

Much work is now underway in an attempt to model the relationships among climate, sediment supply, sea-level change and subsidence, and resulting basin architecture (Jordan et al. 1988; Jordan and Flemings 1991; Sinclair and Allen 1992; Hoffman and Grotzinger 1993; Dorobek and Ross 1995; Sect. 8.6). Complex inter-relationships among these variables are emerging. For example, the sediment supply to the basins on either side of an orogen may be very different because of the position and orientation of the basins and the intervening orogen with respect to prevailing winds. One basin may be in an upwind position, with adiabatic rainfall on the adjacent orogen flanks generating an enhanced sediment supply. The basin on the opposite side of the orogen may then be in rain shadow, with correspondingly reduced sediment supply. Jordan (1995) classified foreland basins as overfilled or underfilled, with respect to the balance between subsidence and sedimentation (Fig. 7.31). **Underfilled** basins are those receiving relatively limited sediment supply from the orogen. They may have lakes or seaways at their centers or be occupied by trunk rivers draining axially through the basins. Additional sediment may be shed into the basins from the distal (forebulge) side. A high sediment supply from an orogen may lead to the **overfilled** condition, with river systems extending across the entire basin and forebulge and onto cratonic (backbulge) areas.

Foreland basins are developed in two quite distinct plate-tectonic settings. These settings are described in Chapter 9, and additional details of sedimentation and tectonics in foreland basins are given in Section 9.3.2.7 (retroarc basins) and Section 9.3.4.7 (collision-related foreland basins).

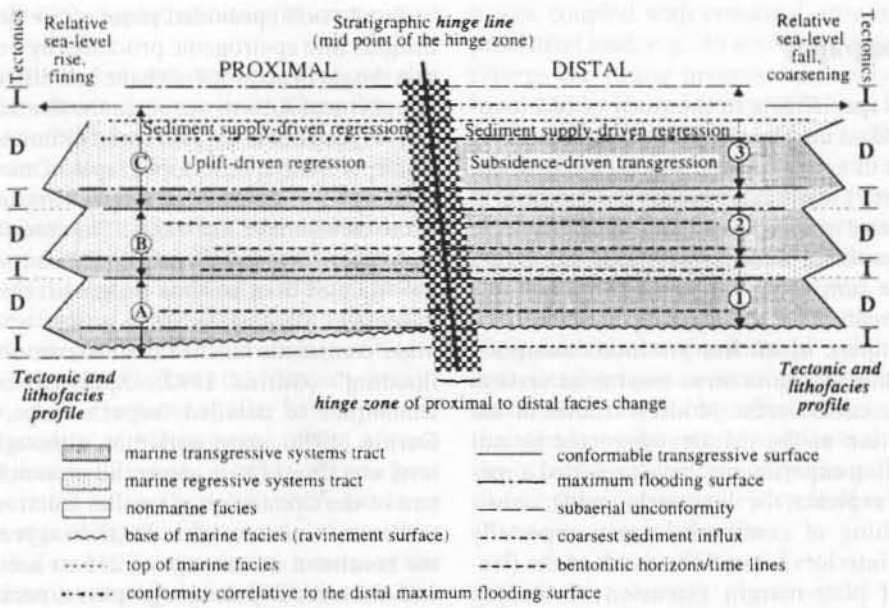
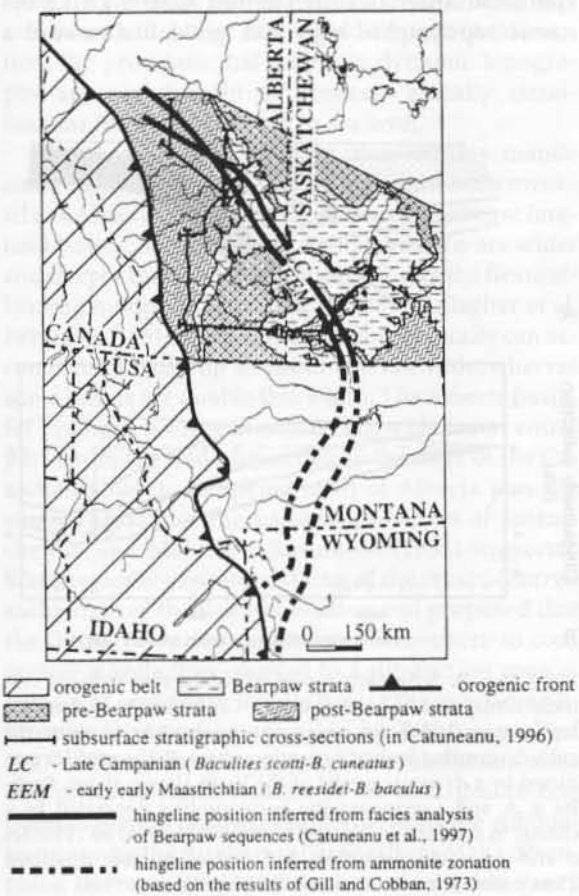


Fig. 7.29. Diagrammatic representation of reciprocal sedimentation. Transgressive deposits in the proximal part of the basin (the foredeep) correlate with regressive deposits in the distal part of the basin (the forebulge). The hingeline between these areas migrates slowly in time in response to long-term changes in the position of the crustal load. (Fig. 7.30; Catuneanu et al. 1997a)



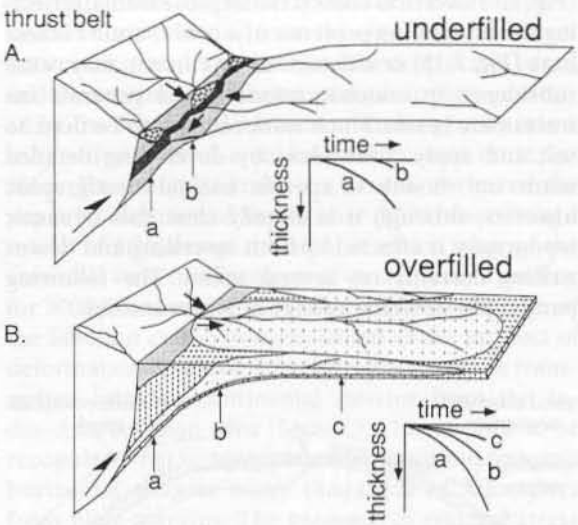


Fig. 7.31 A, B. Comparison of underfilled and overfilled foreland basins. Insets show simple subsidence plots for selected points within the basin. (Jordan 1995)

Fig. 7.30. The changing position of the hingeline between the foredeep and the forebulge in the Western Interior Basin. The position of this hingeline in Canada was determined by sequence-stratigraphic studies. In Montana and Wyoming, the position was determined from shorelines indicated by the ammonite biostratigraphic studies of Gill and Cobban. (1973; Catuneanu et al. 1997a)

7.5 Dynamic Topography

Stratigraphers specializing in the study of continental interiors (Sloss and Speed 1974) have for a long time appealed to a process that was termed **epeirogeny** by Gilbert (1890). The modern definition of epeirogeny (Bates and Jackson 1987) defines it as "a form of diastrophism that has produced the larger features of the continents and oceans, for example plateaux and basins, in contrast to the more localized process of orogeny, which has produced mountain chains." The definition goes on to emphasize vertical motions of the earth's crust. Modern studies of the thermal evolution of the mantle, supported by numerical modeling experiments, have provided a mechanism that explains the long-term uplift, subsidence, and tilting of continental areas, especially large cratonic interiors beyond the reach of the flexural effects of plate-margin extension or loading (Gurnis 1988, 1990, 1992; Mitrovica et al. 1989; Burgess and Gurnis 1995; Burgess et al. 1997). These studies have shown that the earth's surface is maintained in the condition known as **dynamic topography**, reflecting the expansion and contraction of the crust resulting from thermal changes in the underlying mantle (Fig. 7.32). Vertical contractional movements, reflecting the underlying presence of a cold mantle current limb (Fig. 7.33) or a downwelling current, may cause subsidence in cratonic interiors and generate intracratonic basins. Much work remains to be done to test and apply these ideas by developing detailed numerical models of specific basinal stratigraphic histories, although it is already clear that dynamic topography is affected by both upwelling and downwelling currents on several scales. The following paragraphs describe a range of recent studies.

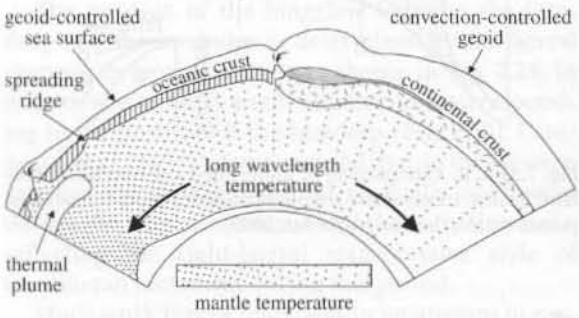


Fig. 7.32. The generation of dynamic topography and controls on sea-level change and regional, vertical crustal motion by shallow and deep mantle convection. Widespread heating beneath supercontinents (at *right*) generates continental uplift and an elevated geoid. More localized depression of the geoid is caused by the subduction of cold slabs of oceanic crust. (Gurnis 1992)

Bond (1978) provided some of the first important insights into epeirogenic processes by demonstrating that the earth's continents have had different histories of uplift and subsidence since the Cretaceous time. It is now possible to explain these differences using the concepts of dynamic topography. One of the most striking anomalies revealed by his hypsometric work is the elevation of Australia. "The interior of Australia became flooded by nearly 50% between 125 and 115 Ma and then became progressively exposed between 100 Ma and 70 Ma at a time when nearly all other continents reached their maximum Cretaceous flooding" (Gurnis 1992). Applying backstripping techniques to detailed isopach maps, Russell and Gurnis (1994) estimated that, although global sea level was about 180 m above the present level near the end of the Cretaceous, a smaller fraction of the Australian continent was flooded than at present. Raising the continent an average of 235 m accounts for the end-Cretaceous paleogeography, superimposed on a 180-m-high sea level. This result is related to cessation of subduction on the northeast margin of Australia at about 95 Ma. Subduction had generated a dynamic load by the presence of a cold crustal slab at great depth, and the cessation of subduction allowed uplift. The northward migration of the continent as it split from Antarctica then moved Australia off a dynamic topographic high and geoid low, toward a

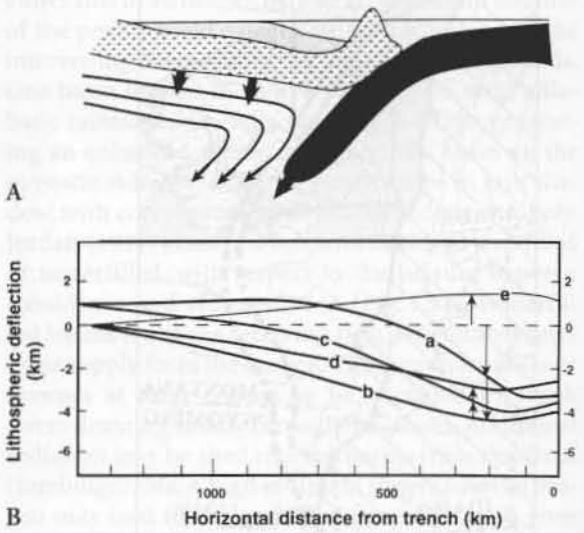


Fig. 7.33 A, B. Schematic model of subduction-induced cratonic tilting and basin deepening – an example of dynamic topography. **A** Overdeepening of a foreland basin above the cold, descending limb of a mantle current. **B** The profiles predicted by a dynamic model of the basin shown above. Profiles *a*, *b*, and *c* represent the basin profiles generated by a change in subduction dip angle from 60° to 25° to 60°. Profiles *d* and *e* are the deflection and surface profiles predicted 25 m.y. after subduction ceases. (Gallagher et al. 1994)

lower dynamic topography, resulting in continental subsidence.

The concepts of dynamic topography have also been invoked to explain cratonic basin formation and subsidence. As noted by Hartley and Allen (1994), there have been at least two major periods in Earth history when suites of interior basins formed within large cratons. Both periods were associated with the breakup of supercontinents. The first of these was during the late Proterozoic and Early Paleozoic, when the Williston, Hudson Bay, Illinois, Michigan, and other basins formed in the cratonic interior of North America. The second period was the Mesozoic breakup of Pangea, when a series of similar basins formed within the continent of Africa. Some of the African basins are undoubtedly related to plate-margin processes and, as noted below (Sect. 7.8), there has been much debate regarding the importance of reactivation of inherited crustal weaknesses as a cause of the North American basins (Quinlan 1987; Sect. 9.3.6). However, Hartley and Allen (1994) suggested that small-scale convective downwelling, decoupled from the large-scale motion, may be a significant factor in basin formation. They found strong evidence for this process in the formation of the Congo Basin. The stratigraphic histories of these suites of cratonic basins are similar but not identical (Quinlan 1987) and, as with the other examples discussed in this section, the processes that maintain dynamic topography are not thought to generate globally simultaneous (eustatic) changes in sea level.

The thermal consequences of secondary mantle convection above a subducting slab have been invoked as a cause of enhanced subsidence in retroarc foreland basins. Some retroarc foreland basins are wider and deeper than can be accounted for by the flexural-loading model (Mitrovica et al. 1989; Gallagher et al. 1994; Fig. 7.33). Supracrustal loading typically can account for a basin up to about 400 km wide, whereas some basins are double this width. The Alberta basin, for example, is more than 1000 km wide in its central part (from the fold–thrust belt to the edge of the Canadian Shield), occupying most of Alberta plus the southern parts of the adjacent provinces of Saskatchewan and Manitoba. Beaumont (1982) suggested that a regional basinward tilting of the crust occurred at the time of the flexural loading and proposed that the tilting was a response of the lithosphere to convective mantle flow coupled to a subduction zone. A pattern of secondary mantle flow in the overriding lithospheric plate was suggested by Toksöz and Bird (1977). Flow takes place toward the subduction zone and is drawn down parallel to the cold, descending oceanic plate. The crust is tilted toward the subduction zone by the mechanical drag effects of the downgoing current. When subduction ceases, buoyancy forces, coupled with erosional unroofing of the supracrustal load, e.g., an accreted orogen, combine to reverse the tilting process, leading to uplift of the basin. Mitrovica and Jarvis (1985) and Mitrovica et al. (1989) modeled this process and showed that the width of the crust affected by the tilting increases as the subduction angle decreases. At subduction angles of 20°, the tilt effect extends more than 1500 km from the thrust front. The degree of tilting will also be affected by the flexural rigidity of the overriding lithosphere. The cessation of crustal shortening accompanies the termination of subduction and its associated mantle convection currents, so that the mechanical down-drag effect ceases, and the basin then tends to rebound (Mitrovica et al. 1989). The entire cycle takes a few tens of millions of years to complete and would generate a cycle of relative sea-level change on a regional scale.

7.6 Intraplate Stress

"Ridge push" and "slab pull" are informal terms that have been used for some time in the plate-tectonics literature to refer to the horizontal (in-plane) forces associated with, respectively, the horizontal compressional effects resulting from the elevation differences between a spreading center and the deep ocean basin, and the tensional effects on an oceanic plate generated by the downward movement under gravity into a subduction zone of a cold slab of oceanic crust. In-plane (horizontal) stress is a central theme of the paper by Molnar and Tapponnier (1975) describing a model of Himalayan collision, in which most of the Cenozoic structural geology of west China, extending for 3000 km north of the Indus suture to the edge of the Siberian craton, was explained as the product of deformation resulting from intraplate stresses transmitted into the continental interior from the India–Asia collision zone (Sect. 9.3.4.10). It came to be recognized that tectonic plates can store and transmit horizontal stresses many thousands of kilometers from plate margins. The presence of residual stress fields in continental interiors has long been known from such evidence as the development of active joints ("break outs") in exploration holes (summary in Cloetingh 1988). Stress data for northwest Europe are shown in Fig. 7.34. Cloetingh (1988) provided similar data for the India–Australia plate, which revealed high compressive stresses, particularly in the northeast Indian Ocean, associated with the northward collision and attempted subduction of this plate beneath Eurasia. There, the seafloor has been deformed into broad flexural folds as a result of intraplate compressive stress. Basin inversion is another

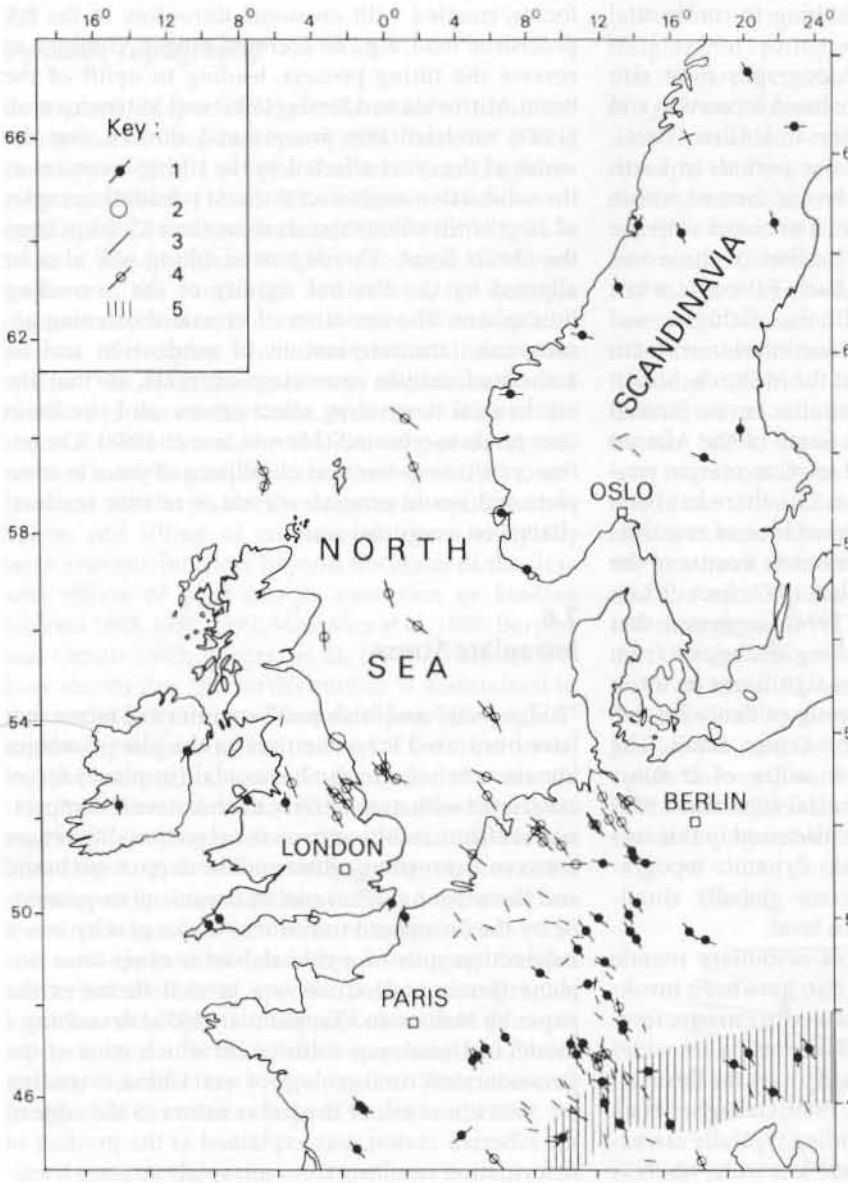


Fig. 7.34. Compilation of observed maximum horizontal stress directions in the northwest European platform. *1* In situ measurements; *2* horizontal stresses equal in all directions, as determined from in situ measurements; *3* determinations from earthquake focal-mechanism studies; *4* well break outs; *5* location of Alpine fold belt. (Cloetingh 1988)

consequence of intraplate stresses (next section). Cloetingh (1988) suggested that:

> The observed modern stress orientations show a remarkably consistent pattern [in northwest Europe], especially considering the heterogeneity in lithospheric structure in this area. These stress-orientation data indicate a propagation of stresses away from the Alpine collision front over large distances in the platform region.

In the late 1980s, the study of in-plane stress evolved into the World Stress Map Project, under the auspices of the International Lithosphere Program. Zoback (1992) provided a report on this project as one of a series of papers discussing intraplate stress. She compiled over 7300 in situ stress-orientation measurements and provided a series of maps documenting the results. Zoback (1992) and Richardson (1992) confirmed the observation that stress orientations are remarkably consistent over large continental areas, including areas that are characterized by considerable crustal heterogeneity. Richardson (1992) pointed out that, on a global scale, orientation measurements are most readily interpreted with respect to the location and orientation of active spreading centers and suggested that ridge-push forces are the most important in determining in-plane stress. Most stresses are compressional, with tensional stresses

having been recorded mainly in areas of high topography. Stresses associated with crustal collision are locally important, but the patterns of stress indicate a complex relationship between intraplate stress and tectonic deformation.

Intraplate stresses impose long-wavelength, low-amplitude flexures on the continental crust that have important effects in the generation and modification of stratigraphic sequences. These are described in Section 8.2.3.1.

7.7 Basin Inversion

Basin inversion is defined as the uplift by compression or transpression of a basin along its controlling fault system, resulting in uplift and partial exhumation of the basin fill (Fig. 7.35). It represents a specific type of response to the patterns of regional intraplate stresses described in the previous section. Typically, only a few of the faults in a basin may be reactivated, and the deeper parts of the basin may be left largely undisturbed. Basin inversion may be a precursor to orogenic deformation. Inversion is of considerable importance to the petroleum industry, because it commonly generates new hydrocarbon traps, and because of the effects the contractional movements have on fluid pressures and fluid transmission along fractures and faults. Useful collections of research articles on this topic were provided by Cooper and Williams (1989) and Buchanan and Buchanan (1995).

An example of regional inversion was described by Brodie and White (1995), who explained the present configuration of the north-western European continental shelf, including the large area of uplift corresponding to the British Isles. Given the history of Mesozoic extension of the north-western European margins and the distance of this region from the Cenozoic Alpine orogen, it might have been expected that the area presently occupied by Britain would consist of flat-lying Mesozoic–Cenozoic strata covered by the seas of a broad continental shelf (i.e., no Britain!). The islands owe their existence to widespread uplift by igneous underplating during the development of the Tertiary Igneous Province. Geological evidence for this consists of the lack of thick rift-fill sediments of Permo-Triassic age (the numerous Permo-Triassic basins in and around the British landmass are all too shallow, based on the criteria of the standard extensional-basin models), high maturation levels in the pre-Tertiary section, and the offlapping pattern of the Jurassic–Cretaceous section in southern and eastern Britain (and offshore in Moray Firth), indicating broad uplift to the north and west and removal of some 2–3 km of the cover succession from northern England and Scotland. This uplift took place in the Paleocene, and the resulting detritus (including much reworked Chalk) was transported into deep basins, such as the Viking Graben, where it now forms some of the best hydrocarbon reservoir units in the northern North Sea Basin.

The term "inversion" is more typically applied to individual basins on scales of a few tens to hundreds of kilometers. Most inverted basins achieve this configuration by contractional movements oblique to the orientation of the original extensional tectonism. Lowell (1995) stated that "it seems inescapable that external horizontal or far-field forces along and transmitted within tectonic plates are required for inversion." Transpressional tectonism is commonly the cause of inversion, although thermal and isostatic uplift may also be important. Some faults are reactivated by compression and others having very similar geometries are not. Sibson (1995) described the

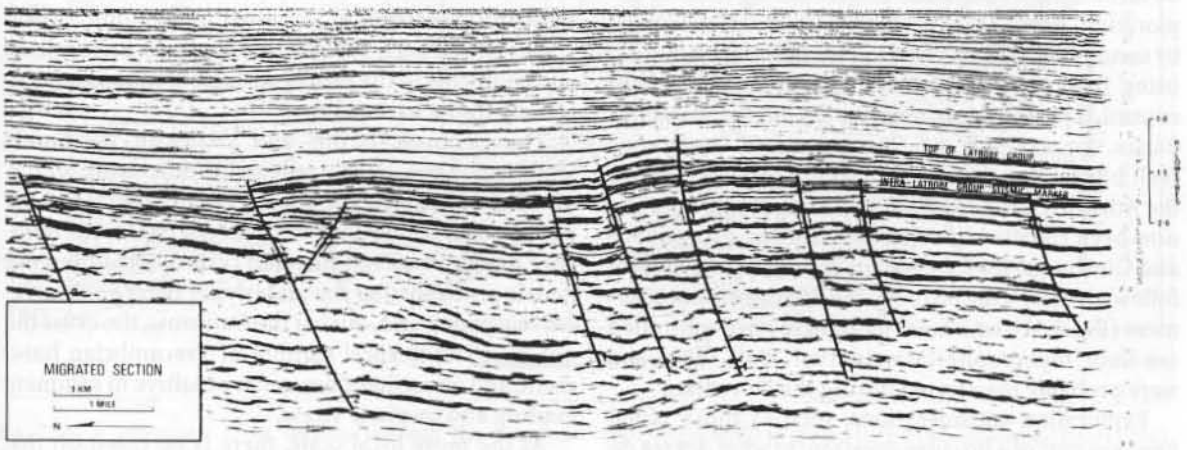


Fig. 7.35. An example of basin inversion from the Gippsland Basin, Australia. Faults labeled '2' were reactivated during inversion of the Latrobe Group (upper Cretaceous–upper Eocene), while faults labeled '1' remained as normal faults. (Lowell 1995)

process known as fault-valve release, whereby faults oriented at high angles to the contractional stresses may move suddenly, resulting in temporary dilation and the release of large volumes of overpressured fluids. Such events can commonly be reconstructed by the careful study of extensional vein systems and their calcite or quartz infills. Lowell (1995) provided numerous cross sections of inverted basins, most based on reflection-seismic data, that display examples of inversion from around the world. Detachment faults have been reactivated as thrust faults, and normal faults have undergone reversal of slip or have become strike-slip faults, commonly displaying positive flower-structure configurations in cross section. MacGregor (1995) demonstrated that the timing of inversion and the accompanying development of structural traps relative to the timing of petroleum generation and migration is critical in the formation of many large oil and gas fields.

7.8 Basement Tectonic Control

Anticlinal petroleum traps caused by draping over an underlying feature, such as a reef or buried hill, constitute a class of trap that has long been known (Levorsen 1967). The search for such traps may focus attention on the paleogeographic conditions that give rise to such resistant objects, such as a barrier-reef trend, or it may stimulate mapping of the paleogeomorphology of unconformity surfaces (Martin 1967; Halbouty 1982). The involvement of basement in the localization of anticlinal structures is commonly investigated, and such research may prove fruitful, as in the discovery of the giant Hassi Messaoud field in Algeria (North 1985). Static basement highs, including buried topography, and structural features that are capable of reactivation, such as fault blocks, may be included as exploration targets during frontier exploration. Ball (1972) discussed the control exerted by tectonics on the localization of carbonate deposits, using Bouguer gravity maps to illustrate basement control. The thick carbonate piles of the Bahamas Banks, for example, are located above Cretaceous fault blocks that formed during the initial rifting of the Atlantic Ocean (Ball 1972), a relationship that has now been clarified by seismic-reflection data (Eberli and Ginsburg 1989). Devonian reef trends in Alberta follow density contrasts in the Precambrian basement (Fig. 5.31), which are thought to have controlled sea-floor topography (Martin 1967; Ball 1972) and were probably reactivated during the Devonian.

Exploration considerations, such as those noted here, are part of a broader question: to what degree do basement structures influence basin histories in the cover rocks? There are two parts to this question:

1. Can the reactivation of basement weaknesses, such as ancient rift zones or sutures, be invoked to explain cratonic uplifts and basins?
2. Can basement reactivation be invoked to explain local paleogeographic trends in cover rocks?

The answer to the second question is definitely yes, whereas the degree to which basement tectonism controls the broader architecture of the cratonic cover remains controversial. Many researchers have pointed to the coincidence between the localization of intracratonic basins and arches in the cover strata and rifts or sutures in the underlying Precambrian basement, but the topic remains controversial, and many, such as Aitken (1993), are not convinced of the significance of such coincidences. Aitken (1993) discussed the Phanerozoic basins and arches of the cratonic cover of Western Canada, but the points he raised have a general applicability. The subject is discussed in greater depth in Section 9.3.6.1. Maps such as those of Sanford et al. (1985) are typical of a theme of comparative analysis that has been termed "trendology" – the linking of disparate objects (basement structures, mineral showings, oil pools) into trends that purport to reveal underlying causality, although commonly the actual mechanism of causation is not clear. Air-photo analysis of structural lineaments is commonly a part of such trend analysis. It can be useful in detecting the regional structural grain and may well be of relevance in analyzing the effects of intraplate stress, but the ease with which such analyses may be performed sometimes seems to impart a greater significance to the results than they deserve. The sweeping cratonic trends of Sanford et al. (1985) would not now be predicted from our modern knowledge of tectonic or epeirogenic processes, although we now know that, regionally, the forces associated with intraplate stress may reactivate or otherwise modify basement or cover structures and can lead to structural reactivation, as noted in Sections 7.6 and 7.7. In detail, such reactivation tends to be localized and conforms to the locally variable trends of heterogeneity in the basement.

The Cenozoic basins and arches of the Sunda Shelf, Indonesia, clearly reflect the underlying "grain" of relict arc complexes (Fig. 9.39), in contrast to the equivocal evidence for Precambrian basement influence on the Western Canadian craton. The difference here is probably that the underlying relict arcs are relatively young and, during the Cenozoic, the crust did not have the flexural rigidity of Precambrian basement, and were, therefore, more sensitive to sediment loading and in-plane stress.

At the more local scale, there is no question that paleogeographic patterns within individual basins, isopach and isolith trends, paleocurrent-dispersal

trends, and so on, reflect basement control, and a recognition of the importance of such controls may become an important guide to mapping and interpretation. For example, Hart and Plint (1993) and Plint et al. (1993) discussed tectonic mechanisms that governed construction of the Smoky Group in the foreland basin of Alberta and British Columbia. The overall architecture of the Smoky Group was interpreted as a product of a varying rate of flexural subsidence and associated forebulge movement on a 10^6-year time scale (Plint et al. 1993). Bevels on certain erosion surfaces within the Cardium Formation were interpreted as the result of flexure over reactivated basement faults and resulting shoreline incision (Hart and Plint 1993). Donaldson et al. (1998) interpreted subtle drape effects within Cretaceous strata of northwestern Alberta as the result of reactivation by foreland-basin tectonism of faults cutting the Precambrian basement. Brandley et al. (1996) confirmed the interpretation of Ball (1972), noted above, that heterogeneities in the Precambrian basement of Alberta were reactivated and influenced sedimentation during the Phanerozoic. They demonstrated that thickness and facies trends parallel terrane boundaries in the basement. Pang and Nummedal (1995) demonstrated the importance of basement heterogeneity as a control of subsidence patterns in the Western Interior Basin of the United States. Lawton (1986a, b) used evidence of sandstone detrital sources and paleocurrent patterns to determine the location of uplifted terrains and regional paleoslopes that confirmed the importance of basement movements during the deposition of Cretaceous and lower-Cenozoic alluvial and coastal-plain deposits derived from the Sevier and Laramide orogens of Utah. The incipient movements of basement-involved Laramide structures were clearly documented. This was well illustrated by a sedimentological study by Guiseppe and Heller (1998), who demonstrated changes in fluvial architecture across the downdip end of the San Rafael Swell in Utah. This structure began to rise during the Late Cretaceous. The beginning of the movement can be documented by changes in fluvial style. The architecture of channel sandstone bodies that were deposited before movement does not vary across the arch whereas, once Laramide uplift began, river systems were diverted to the flanks of the arch, and the deposits formed above the arch are dominated by floodplain facies.

A particularly instructive example of basement control is the evolution of the Paleozoic Paradox Basin of Utah and Colorado and its Mesozoic cover. The Paradox Basin was generated during a phase of Pennsylvanian tectonism in the southwestern United States triggered by the collision of Laurasia and Gondwana (Kluth and Coney 1981). This intraplate

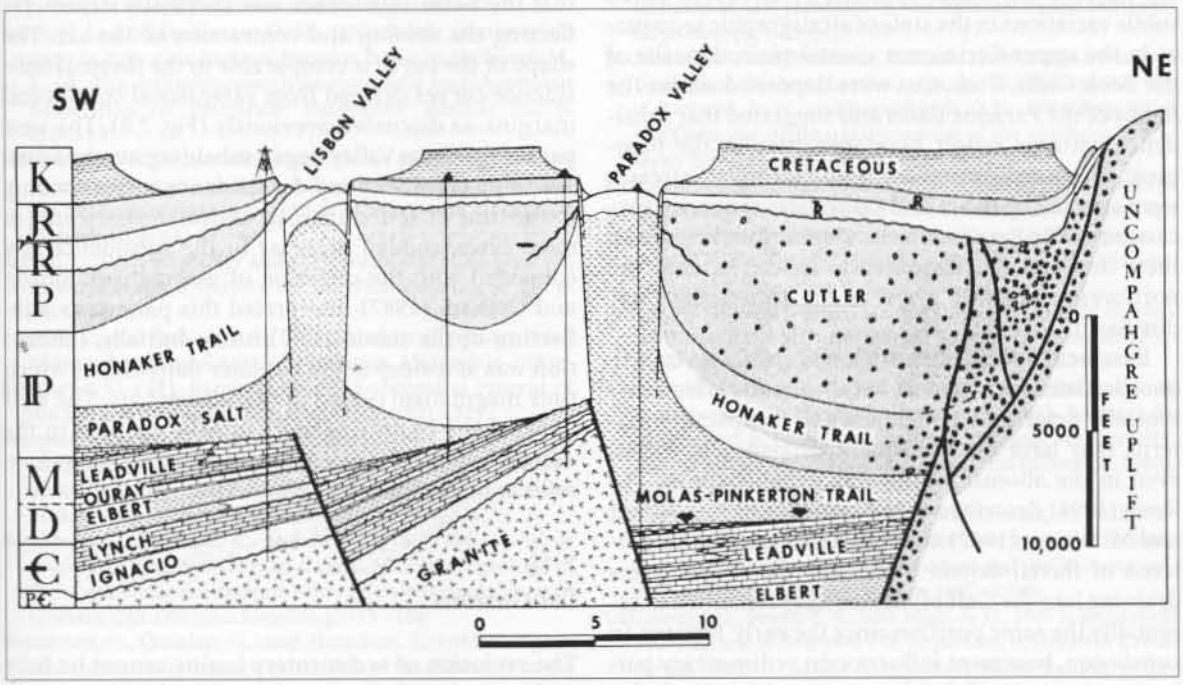


Fig. 7.36. A schematic cross section through the Paradox Basin, Utah, showing rejuvenated basement faults, overlying salt anticlines, the bounding Uncompahgre fault, and the great thickness of Cutler group clastics that were shed from the Uncompahgre uplift to the northeast (Stevenson and Baars 1986). Movements on the anticlines continued into the Jurassic (see text)

movement reactivated northwest–southeast-trending Precambrian structures and created the Ancestral Rockies, including the Paradox Basin and the adjacent Uncompahgre Uplift. The basin filled with evaporites and carbonates, the latter primarily along the flanks of the basin. In the middle Pennsylvanian, the Uncompahgre Fault, along the southwestern flank of the uplift, accumulated up to 6 km of displacement as an equivalent thickness of coarse clastic debris was shed from the uplift south and west into the basin to form the Cutler Group (Baars and Stevenson 1982; Stevenson and Baars 1986). The sediment load on the salt initiated flowage and caused the formation of giant salt anticlines along northwest–southeast trends (Fig. 7.36), guided by movement on reactivated Precambrian faults. Movement on these anticlines was largely completed by Triassic times, but recent research on the sedimentology of the Triassic and younger cover rocks has revealed much evidence for subtle basement influence, in this case, the continued upward movement of the salt. Hazel (1994) demonstrated the presence of intraformational unconformities in the Triassic Chinle Formation that developed as uplift continued on buried salt anticlines near Moab, Utah. Hazel (1994) also demonstrated the effects of this movement on fluvial architecture and dispersal patterns in the Chinle Formation. Bromley (1991) carried out a similar study, which showed such movement continuing into the Jurassic in the overlying Kayenta Formation. Yoshida et al. (1996) noted subtle variations in the style of stratigraphic sequences in the upper-Cretaceous coastal-plain deposits of the Book Cliffs, Utah, that were deposited across the flanks of the Paradox Basin and suggested that subsidence patterns reflect heterogeneities in the basement, where sediment loading and intraplate stresses were affected by the mixed carbonate–evaporite succession of the Paradox Basin. Depositional trends in these deposits also appeared to reflect, in part, the northwest–southeast "grain" of the underlying Paradox Basin.

Basement influence on thickness and facies trends, once initiated, may persist because of the effects tectonically determined thickness and compaction patterns may have on subsequent depositional cycles, even in the absence of obvious tectonic control. Le Roux (1994) described three examples of this. Fisher and McGowen (1967) showed, based on isopach patterns of fluvial-deltaic units, that the major rivers draining into the Gulf of Mexico have remained in essentially the same position since the early Tertiary. In conclusion, basement influence on sedimentary patterns can commonly be demonstrated, but the influence may be subtle, and the geologist should be wary of extrapolating trends too far.

7.9 Application of the Modeling Techniques to Other Types of Basin

As discussed in Chapter 9, sedimentary basins form in diverse plate-tectonic settings. However, at present, it appears that the two classes of basin models discussed in Sections 7.3 and 7.4 represent the most common processes whereby basins subside. The same processes of crustal extension, thermal contraction, and supracrustal loading probably lead to the same types of crustal response in whatever combination they occur. However, very few geophysical basin studies have been performed in other types of basin; therefore, much research remains to be carried out in this area.

A single case study may usefully be summarized here. Moxon and Graham (1987) used backstripping procedures to evaluate the subsidence history of the Great Valley, a well-preserved Late Jurassic-Paleogene forearc basin in California (see Sect. 9.3.2 for a discussion of the structure, stratigraphy, and tectonic settings of forearcs). They found that the basin could be subdivided into two parts on the basis of the age and shape of the subsidence curves. The east part of the basin showed relatively slow subsidence in the form of an asymptotic curve. The commencement of subsidence coincided with the cessation of arc magmatism immediately to the east. The conclusion is that the basin subsidence was thermally driven, reflecting the cooling and contraction of the arc. The shape of the curve is comparable to the thermal-subsidence curves derived from extensional continental margins, as discussed previously (Fig. 7.8). The west part of the Great Valley began subsiding much earlier, with a phase of very rapid subsidence corresponding to the time of active arc magmatism. A sharp and, in most cases, sudden decrease in the subsidence rate coincided with the cessation of magmatism. Moxon and Graham (1987) interpreted this pattern as a reflection of the subduction history. Initially, subduction was at a steep angle but later flattened, at which time magmatism ceased in the adjacent arc. The west flank of the basin was mechanically coupled to the descending slab, and its subsidence rate, therefore, reflects the tectonic history.

7.10 Conclusions

The evolution of sedimentary basins cannot be fully understood without consideration of their tectonic setting and structural evolution. In a general sense, all sedimentation is syntectonic, and all tectonism is syndepositional. The broader features of basin archi-

tecture depend on the plate-tectonic setting, as discussed in Chapter 9, but many aspects of basin stratigraphy and paleogeography are controlled by contemporaneous structural features activated by local extensional, transcurrent, or contractional tectonism, or by in-plane stresses transmitted from an orogen or other type of plate margin, and commonly modified by the response of the local basement to these stresses. Some of these tectonic processes are well-enough understood to have been replicated by numerical models, as described in Sections 7.3–7.5, and such models, by allowing us to explore various combinations of input data, are providing important insights into stratigraphic architectures formed under a wide range of real-world conditions (Sect. 8.6).

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Regional and Global Stratigraphic Cycles

8.1 Overview of Recent Developments

The development of the science of planetary geology, our increasing familiarity with views of the earth taken from satellites and from outer space, and the increasing sophistication of geophysical techniques for exploring the earth's interior have all encouraged scientists to adopt a planetary perspective on questions of global history and current problems of biological, climatic, and geologic change (Anderson 1984; Maxwell 1984). From this has evolved the "Gaia" concept and the **earth-systems-science** approach to the study of our planet (Skinner and Porter 1995). Early sequence studies evolved from the model of **global eustasy**, in which it was hypothesized that sequence stratigraphies around the world were controlled primarily by global changes in sea level (Vail et al 1977). If this is indeed the case, it permits us to build a global stratigraphic template for correlation based on regional stratigraphic successions.

Developments in the field of sequence stratigraphy have revolutionized the regional study of sedimentary rocks during the last decade. Sequence architecture models predict the distribution of sedimentary facies based on the response of depositional environments to changes in base level (Chap. 6). These models have provided a powerful basis for regional subsurface stratigraphic prediction and correlation, and have become widely used in the field of petroleum geology. The main credit for this revolution goes to Peter Vail, who brought his early ideas to fruition while working for Exxon Corporation (Vail et al. 1977), and to his graduate supervisor, the late Larry Sloss at Northwestern University, who established modern sequence concepts based on his extensive stratigraphic studies (Sloss et al. 1949; Sloss 1963).

However, these developments in sequence stratigraphy have proved controversial (Kerr 1980; Miall 1986; Gradstein et al. 1988), and this has stimulated a considerable body of research into basin subsidence processes (Chap. 7), the causes of sea-level change, and the response of depositional systems to the various allogenic forcing processes, tectonism, eustasy, and climate change. Early sequence-architecture models (Posamentier et al. 1988; Posamentier and Vail 1988) proved to be simplistic. For example, in the case of fluvial deposits, the responses of rivers to base-level change and to other influences, such as climate change and tectonism, is much more complex than had been suggested, and sequence models for this environment are, therefore, still evolving (Miall 1991a, 1996; Schumm 1993; Wescott 1993; Blum 1994; Shanley and McCabe 1994; Sect. 6.5.1.1). Studies of carbonate sedimentation on continental platforms and slopes show that carbonate environments respond very differently to sea-level change than do clastic environments and are very sensitive to other controls, such as changes in water temperature and suspended-sediment concentration (Sect. 6.3). Architectural models must be adapted accordingly. In particular, carbonate sedimentation is most active during sea-level highstands, developing thick platform deposits and slope debris aprons (the products of "highstand shedding") in contrast to clastic deposits, the thickest accumulations of which are commonly those deposited on the continental slope during lowstand (James and Kendall 1992; Schlager 1992; Sect. 6.3). The architectural implications of erosion and oceanward sediment transport during falling base level is also a recent addition to the body of sequence concepts, leading to modifications in our understanding of the timing of the evolution of beach–barrier systems and submarine fans (the "forced regressions" and "falling-stage systems tracts" of Posamentier et al. 1992; Hunt and Tucker 1992; Sects. 6.3, 6.4). The original sequence models were developed for extensional continental margins, and it is now being shown that, in other types of tectonic setting (especially foreland basins), sequence architecture and composition are quite different (Swift et al. 1987; Jordan and Flemings 1991; Posamentier and Allen 1993).

Questioning of the global eustasy model for sequence architecture has led to two distinct developments. One critical test of global eustasy has been exploration of the global correlation of sequence successions, and this has stimulated a renewal of interest in the global time scale, the data upon which it is based, and the accuracy and precision with which it can

be applied (Sect. 3.7). Miall (1991a, 1992, 1994) argued that the chronostratigraphic record is still too imprecise to permit reliable tests of global correlation of the higher frequency cycles – those of 1–10 m.y. duration, the so-called third-order sequences, which constitute the basis for recent versions of the Exxon curves (Haq et al. 1987). The question of global-cycle correlation and the story of the Exxon **global-cycle chart** are discussed in Section 8.4.

The one major exception to the problem of chronostratigraphic imprecision is the late Cenozoic, during which glacioeustatic fluctuations generated systematic variations in oxygen-isotope concentrations on a 10^4–10^5-year time scale (the so-called **Milankovitch band;** Shackleton and Opdyke 1973), and this is providing the basis for a reliable time scale for the last few million years of geologic history (Imbrie et al. 1984; Williams 1988, 1990; Sect. 3.7.9.1) and for the growth of a new discipline, that of **cyclostratigraphy** (discussed below).

The second development has been a renewed interest in the deep structure of the earth and the fundamental mechanisms that control sea-level change and the elevation of the continents. On most continents, deep-reflection seismic experiments that were begun in the 1970s by government–industry consortia (Oliver 1982) have produced profiles of the continental crust that reveal the structural complexity of orogens and continental margins down to depths of tens of kilometers, typically to the Mohorovičić discontinuity. The major discovery was demonstration of the existence of extensional and contractional detachment surfaces cutting to the base of the crust at steep angles from the horizontal, and the overthrusting of continental margins by tens to hundreds of kilometers in collisional orogens (Allmendinger et al. 1987; Mooney and Braile 1989; Blundell et al. 1992; Chap. 9; Figs. 9.43, 9.61, 9.87, 9.88). These data have profoundly influenced our understanding of the mechanisms that operate to influence sea-level change.

A variety of tectonic mechanisms has been explored by theoretical studies, numerical simulation, and the accumulation of careful, regional case studies. A new field, called **computational geodynamics**, has emerged, "in which computer models of mantle convection are used in the interpretation of contemporaneous geophysical observations like seismic tomography and the geoid as well as of time-integrated observations from isotope geochemistry" (Gurnis 1992). These developments are of great significance, and their implications for stratigraphy have yet to be fully realized. Gurnis (1988, 1990, 1992) developed the concept of **dynamic topography**, in which the earth's elevation and that of the oceans is related to the thermal properties of the mantle. The heat that accumulates beneath supercontinents can generate continental-scale upwarps of up to 1 km over periods of 100 m.y., whereas subduction of cold oceanic slabs and the downwelling that occurs as continental fragments converge generates regional continental depression of several hundred meters. These processes are the explanation for **epeirogeny** – the vertical motions of the continents, a feature commonly interpreted from the stratigraphic record (Sloss 1963; Bond 1976, 1978), but hitherto unexplained (Sect. 7.5). To determine their effects on sea level at a given location, they must be integrated with changes in the sea level itself (eustasy), the causes of which are discussed later.

In another series of developments, it has been shown that horizontal, **in-plane stresses** generated by the kinematics of plate tectonics impose long-wavelength, low-amplitude flexural effects on continental and oceanic plates, leading to regional tilts and broad upwarps and downwarps (Cloetingh et al. 1985; Cloetingh 1988; Cloetingh and Kooi 1990; Sects. 7.6, 8.2.3.1). Basin inversion is also caused by this mechanism (Sect. 7.7). Periodic brittle failure of the crust under a continuously-applied compressive stress in converging-plate environments may lead to periodic adjustments to the flexural load in adjacent basins (Peper et al. 1992; Waschbusch and Royden 1992), and this may prove to be one of the most important mechanisms for the development of high-frequency stratigraphic sequences (those of 10^4–10^6-year duration). In fact, the "need" for eustasy as a mechanism to generate sequences is receding (Peper 1994; Yoshida et al. 1996), and this makes global correlation, as a test of eustasy, even more important.

At the same time, a rigorous evaluation of early ideas about orbital forcing – the so-called Milankovitch effects – has confirmed that the processes are real (Berger et al. 1984), and much work on the geological record is demonstrating that sequences in the Milankovitch band have developed at many times in the geological past (Fischer 1986; de Boer and Smith 1994a; Sect. 8.2.4.1). Orbital forcing profoundly affects climate and, at certain times, climate change has been severe enough to trigger major episodes of continental glaciation, leading to periodic drawdown of sea levels. Even without sea-level change, orbital forcing can generate significant stratigraphic cyclicity through its effects on oceanic and atmospheric circulation, evapotranspiration, and organic productivity (Perlmutter and Matthews 1990; Gale 1998).

The purpose of this chapter is to review the stratigraphic record, given these recent developments in geology and geophysics. A more detailed review of the stratigraphic record and of sequence-generating mechanisms was given by Miall (1997).

8.2 Causes of Stratigraphic Cyclicity

The duration and episodicity of geological events and stratigraphic cyclicity span at least sixteen orders of magnitude, ranging from the repeat time of burst-sweep cycles in turbulent boundary layers (10^{-6} years), to plate-tectonic cycles involving the formation and breakup of supercontinents (10^{9} years; Miall 1991b; Einsele et al. 1991; Figs. 8.1, 8.2). A subdivision of stratigraphic cycles into four broad, genetically based classes has evolved from the duration-based hierarchy of Vail et al. (1977). A summary of this classification is shown in Table 8.1, with notes on causal mechanisms. These notes are expanded upon in this section.

It must be emphasized that the fourfold subdivision is quite arbitrary. The rank-order classification given in the right-hand column is used, nowadays, purely for convenience in referring to the time scales of the various cycle types, but this usage should be discouraged because of increasing evidence for the overlap in the durations of the major types of cycle. Drummond and Wilkinson (1996) carried out a quantitative study of the duration and thickness of stratigraphic sequences and confirmed the opinions of Carter et al. (1991). Their major conclusion was that "discrimination of stratigraphic hierarchies and their designation as nth-order cycles may constitute little more than the arbitrary subdivision of an uninterrupted stratigraphic continuum." Ideas about the causal mechanisms are still actively evolving, and it is becoming clear that many are not independent. For example, mantle convection drives both vertical motions of the crust and causes variations in the volume of the ocean basins, which affects eustasy. The resulting effects on sea level at a given point are complex.

8.2.1 Supercontinent Cycles

A cycle of several hundred million years duration is now thought to be caused by the thermal effects relat-

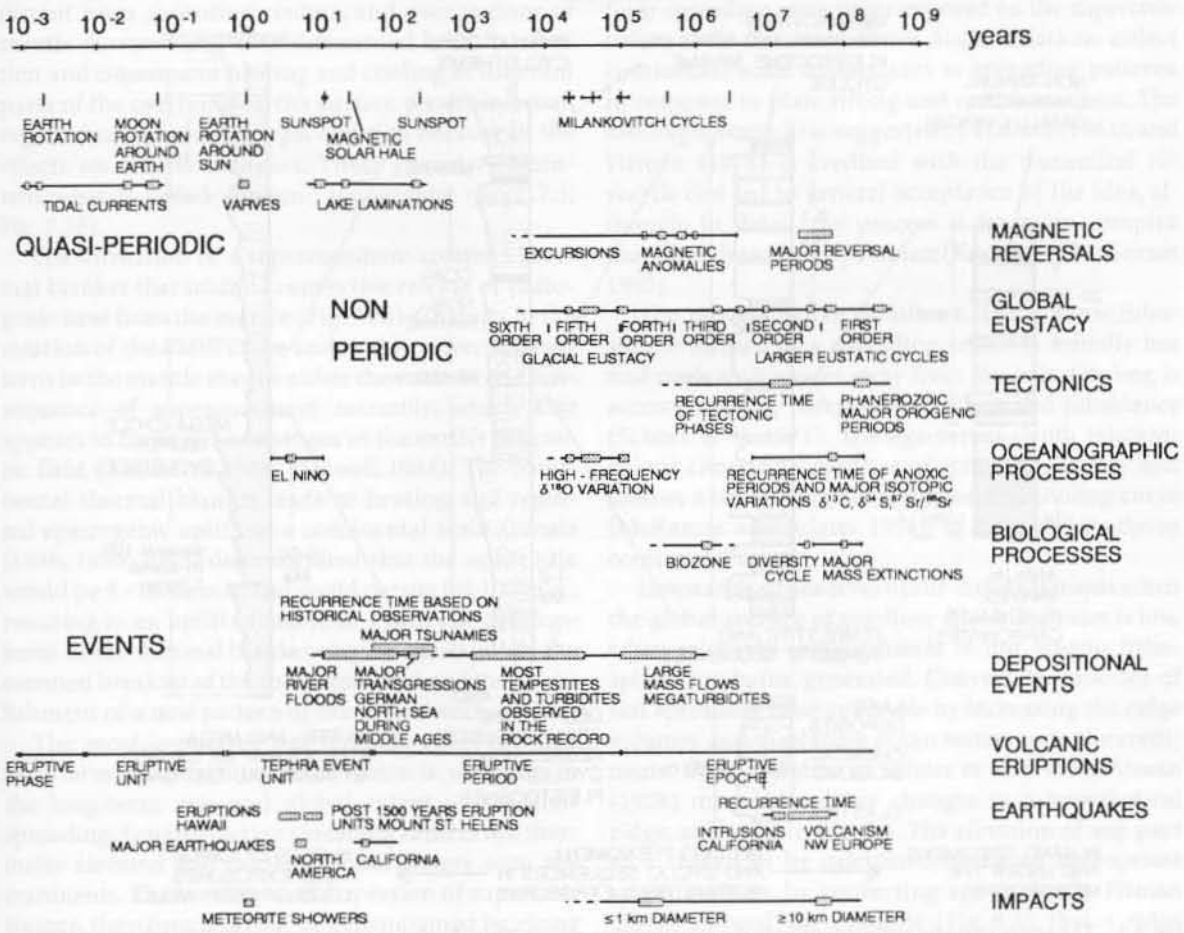


Fig. 8.1. Recurrence time of periodic and episodic processes and events in geology. (Einsele et al. 1991)

Table 8.1. Stratigraphic cycles and their causes

Sequence type	Duration (m.y.)	Other terminology
A. Global supercontinent cycle	200–400	First-order cycle (Vail et al., 1977)
B. Cycles generated by continental-scale mantle thermal processes (dynamic topography), and by plate kinematics, including: 1. Eustatic cycles induced by volume changes in global mid-oceanic spreading centres 2. Regional cycles of basement movement induced by extensional downwarp and crustal loading	10–100	Second-order cycle (Vail et al., 1977), supercycle (Vail et al., 1977), sequence (Sloss, 1963)
C. Regional to local cycles of basement movement caused by regional plate kinematics, including changes in intraplate-stress regime	0.01–10	3rd- to 5th-order cycles (Vail et al. 1977) 3rd-order cycles also termed: megacyclothem (Heckel 1986), mesothem (Ramsbottom 1979)
D. Global cycles generated by orbital forcing, including glacioeustasy, productivity cycles, etc.	0.01–2	4th- and 5th-order cycles (Vail et al. 1977) Milankovitch cycles, cyclothem (Wanless and Weller 1932), major and minor cycles (Heckel 1986)

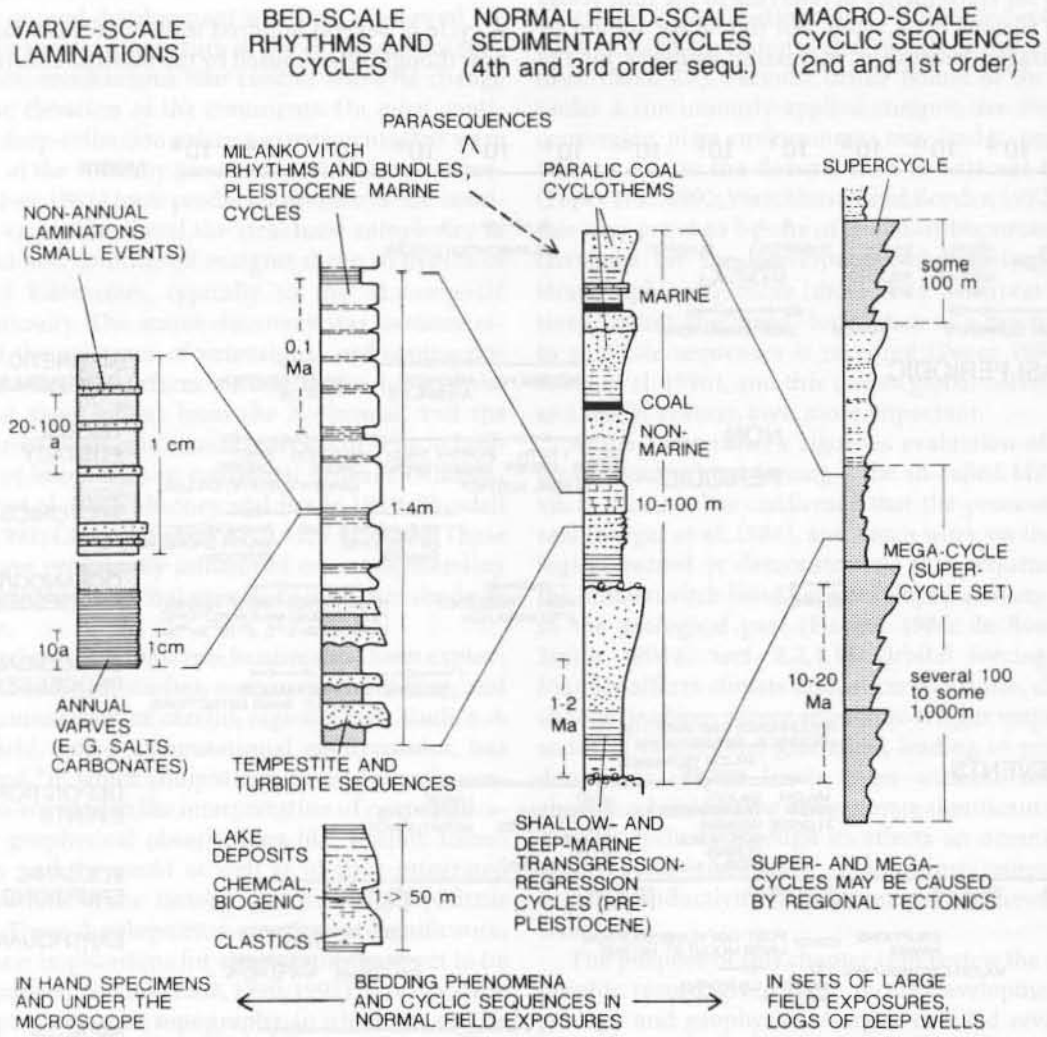


Fig. 8.2. The scales of cyclic sedimentation in the stratigraphic record. This chapter is concerned with all but the first of these types (varve-scale laminations). (Einsele et al. 1991)

ed to the formation and breakup of supercontinents. Worsley et al. (1984, 1986) and Veevers (1990) discussed the Phanerozoic record, and Hoffman (1991) and Dalziel (1992) presented plate-tectonic scenarios for the assembly and dispersal of the major continents that would allow us to extend the interpretation back to about 1 Ga. Some of the details of these cycles are described in Section 8.3.

The major cause of change in the earth's crust, including the driving force of this long-term cyclicity, is the radiogenic heat engine, which generates mantle convection and creates the geomagnetic field. Convection distributes heat and drives plate tectonics. Continental crust is heated adjacent to sea-floor spreading centers where new, hot, oceanic crust is generated. The effect of crustal heating is to cause uplift. This occurs along the flanks of new continental-rift systems (e.g., parts of present-day East Africa) and above mantle plumes. Heat also is thought to build up beneath large continental masses. Thermal doming beneath supercontinents may elevate the crust by as much as 1 km over periods of 100 m.y. Subsidence takes place over cooling areas of the earth's crust, such as areas of aging oceanic crust distant from spreading centers, and over regions of mantle downwelling. This differential heat distribution and consequent heating and cooling of different parts of the overlying earth's surface results in broad regional uplifts, downwarps, and tilts because of the effects on crustal densities. These processes maintain what is called dynamic topography (Sect. 7.5; Fig. 7.28).

The formation of a supercontinent creates a thermal blanket that inhibits convective release of radiogenic heat from the mantle (Fig. 7.28). Changes in the rotation of the earth's core and in the convective patterns in the mantle may be either the cause or the consequence of supercontinent assembly, which also appears to be linked to changes in the earth's magnetic field (Anderson 1984; Maxwell 1984). The continental thermal blanket leads to heating and regional epeirogenic uplift on a continental scale. Gurnis (1988, 1990, 1992) demonstrated that the uplift rate would be 5–10 m/m.y. and could persist for 100 m.y., resulting in an uplift of 500 m to 1 km. The development of the thermal blanket may be the cause of the eventual breakup of the supercontinent and the establishment of a new pattern of mantle convection.

The most important stratigraphic effect of these long-term plate-tectonic movements is variation in the long-term rate and global extent of sea-floor spreading. Lengthy, active spreading centers are thermally elevated and displace ocean waters onto the continents. The breakup and dispersion of supercontinents, therefore, tends to be accompanied by rising sea levels, such as those that characterized the earth during the rapid opening of the Atlantic, Indian and Southern Oceans during the Cretaceous. Plate collisions lead to the development of a new supercontinent, a lull in sea-floor spreading (with a consequent global sea-level lowstand), and the gradual buildup of heat beneath the new continental mass, leading eventually to the initiation of a new round of rifting and continental dispersal.

8.2.2 Mechanisms with Episodicities of Tens of Millions of Years

8.2.2.1 *Eustatic Cycles*

The interregional cycles of Sloss (1963) are the classic examples of what Vail et al. (1977) termed second-order cycles. There is increasingly convincing evidence that many of these are global in scope and are generated by cycles of eustatic sea-level change or by interrelated tectonic and eustatic cycles lasting several tens of millions of years (Miall 1997, Chap 6). The primary mechanism is probably variations in global sea-floor spreading rates superimposed on the supercontinent cycle discussed above. Such variations reflect continental-scale adjustments to spreading patterns in response to plate rifting and collision events. The mechanism was first suggested by Hallam (1963), and Pitman (1978) is credited with the theoretical research that led to general acceptance of the idea, although, in detail, the process is far more complex than first described by Pitman (Kominz 1984; Gurnis 1992).

The process works as follows. The oceanic lithosphere formed at a spreading center is initially hot and cools as it moves away from the axis. Cooling is accompanied by thermal contraction and subsidence (Sclater et al. 1971). The age-versus-depth relationship is constant regardless of spreading history and follows a time-dependent, exponential cooling curve (McKenzie and Sclater 1971), as does the overlying continental crust.

Lowstands of sea level occur during periods when the global average of sea-floor spreading rates is low, when relatively small volumes of hot oceanic lithosphere are being generated. Conversely, episodes of fast spreading raise sea levels by increasing the ridge volumes and displacing ocean waters onto the continents. Using the data of Sclater et al. (1971), Pitman (1978) modeled volume changes in a hypothetical ridge, as shown in Fig. 8.3. The elevation of any part of a ridge can be calculated, using an appropriate spreading rate, by converting age to depth. Pitman (1978) showed, for example (Fig. 8.3), that a ridge spreading at 60 m/ka will have three times the volume

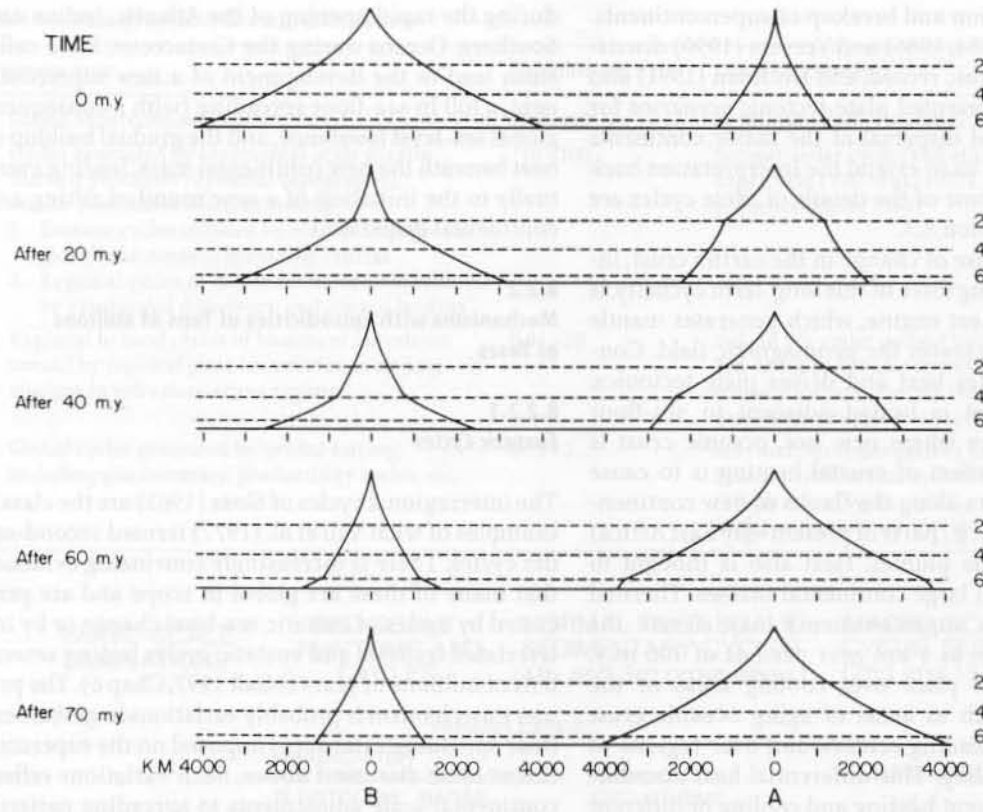


Fig. 8.3 A, B. Profiles of spreading ridges showing the effects of different spreading rates on volume at 20, 40, 60, and 70 m.y. after the initial condition. **A** Profile of a ridge that has been spreading at 20 m/ka for 70 m.y. and changes to 60 m/ka at time zero. After 70 m.y., the ridge has three times its starting volume. **B** Ridge that has been spreading at 60 m/ka for 70 m.y. and changes to 20 m/ka at time zero. After 70 m.y., its volume has been reduced to one-third of its starting volume. (Pitman 1978, 1979)

of one spreading at 20 m/ka, provided that these rates last for 70 m.y. This is the time taken for the oldest (outermost) part of the ridge to subside to average oceanic abyssal depths of 5.5 km, by which time the ridge has achieved an equilibrium profile. The total length of the world midoceanic ridge system is about 45,000 km (Hays and Pitman 1973; Pitman 1978), and Pitman (1978) argued that, allowing for the shape of the continental margins, measured spreading rates can account for eustatic sea-level changes up to a maximum rate of 0.01 m/ka. Later compilations (Pitman and Golovchenko 1991) demonstrated average spreading rates of 0.002–0.003 m/ka for the Jurassic to mid-Tertiary. This is fast enough to generate Sloss-type cycles, those with episodicities of tens of millions of years.

8.2.2.2 *Tectonostratigraphic Cycles*

Cycles of base-level change can be generated by crustal movements at continental margins. On extensional margins, crustal thinning and post-rift cooling lead to subsidence on a time scale of tens of millions of years (McKenzie 1978; Watts 1981, 1989; Sect. 7.3). Clastic wedges are also generated over periods of millions to tens of millions of years by flexural loading during convergent and strike-slip tectonics, especially in foreland basins, where such wedges of nonmarine "molasse" are common (Beaumont 1981; Jordan 1981; Allen and Homewood 1986; Sect. 7.4). Tectonic cycles generated by regional plate-tectonic events are regional to continental in scope (not global) as demonstrated by several regional studies (Tankard and Welsink 1987; Hubbard 1988).

On extensional continental margins, sequence boundaries may represent tectonic events, such as uplift related to rifting, the rift-drift breakup unconformity, and so on (Sect. 9.3.1). The evolution of extensional continental margins is shown diagrammatically in Fig. 7.4. There is an initial rapid phase of continental stretching, which may be completed within a few million years, in the simplest case (although some basins, such as the North Sea, undergo

protracted or repeated rifting events over tens of million of years). This is the rift phase shown in Fig. 7.4. Deformations typically are brittle at the surface, taking the form of extensional faults. These may consist of repeated graben or half-graben faults, and may be listric, allowing for considerable extension of the crust. Beneath the brittle crust, the lithosphere is plastic and extends by stretching. This generates decompression melting and causes hot asthenospheric material to rise closer to the surface, resulting in heating and uplifting of the rift margins, as shown in Fig. 7.4. The uplift is equivalent to a relative fall in sea level, the rate and magnitude of which has been modeled by Watts et al. (1982). The rift phase is then followed by a much longer cooling and flexural-subsidence phase, which lasts for tens of millions of years. Thermally driven subsidence, accompanied by water and sediment loading, leads to downflexure of the continental margin. The flexural phase involves a much broader area of the continental margin than the rift phase, which accounts for the classic steer's-head cross section of extensional-margin basins (Fig. 7.4). Cooling and subsidence is rapid at first, but the rate decreases asymptotically. The result is a subsidence curve that is concave up (Fig. 7.8). This sequence of events may be repeated as successive rifting events occur during the breakup of a major continent.

The question of causality of 10^7-year cycles is particularly difficult to resolve in foreland basins, which, by their very nature, owe their origins to regional tectonic activity. The processes of continental collision, terrane accretion, nappe migration, thrust movement, and imbricate-fault propagation thicken and flexurally load the crust while generating tectonic highlands. Erosional unroofing of the highlands subsequently causes isostatic uplift. These processes lead to episodic cycles of relative changes of sea level on time scales of 10^4–10^7 years. It has long been accepted that these cycles are responsible for large-scale molasse pulses (Miall 1978; Van Houten 1981). The numerical models of Jordan (1981) and Beaumont (1981) provide the necessary theoretical background for explaining subsidence in terms of tectonism. Allen et al. (1986) and Cross (1986) plotted subsidence curves that illustrated the typical subsidence patterns of foreland basins. Subsidence rates tend to increase with time, and the curves commonly are characterized by one or more sharp inflection points where the rate of subsidence rapidly increases as a result of a specific thrust-loading event (Fig. 7.26). Subsidence rates may also vary over a 10^6- to 10^7-year time scale because of heterogeneities in the underlying basement that is undergoing subduction. The rate of flexural response to loading is governed largely by the strength and elasticity of the underlying crust and, if this varies laterally, subsidence rates will vary as different parts of the crust are loaded during contractional movements, including fold-thrust development and subduction (Waschbusch and Royden 1992).

It has commonly been assumed that wedges of coarse sediment prograding from a basin margin are "syntectonic" and, in many earlier studies, the dating of such wedges has been used to infer the timing of major orogenic episodes (see summary and references in Miall 1981; Rust and Koster 1984). It can now be seen that this interpretation is simplistic. In fact, the relationship between tectonism and sedimentation is complex and depends on the balance between a range of controls, including sediment supply and sediment type, and the configuration and rigidity of the flexed basement (Sect. 7.4.2). Some recent studies have demonstrated that tectonism does not necessarily coincide with the progradation of wedges of coarse sediment but may precede such progradation by a significant period of time (Blair and Bilodeau 1988; Heller et al. 1988; Jordan et al. 1988; Heller and Paola 1992; Sect. 7.4.2). In other cases, sediment input and progradation rates are adequate to keep pace with thrust-sheet loading (Sinclair et al. 1991).

Posamentier and Allen (1993) developed a useful model for the sequence stratigraphy of foreland-ramp-type basins, in which the relationship between eustatic sea-level change and flexural subsidence was explored (Fig. 8.4). They suggested that, in many foreland basins, close to the fold–thrust belt there may be a proximal region where subsidence is always faster than the rate of eustatic sea-level change. In this area, eustasy does not lead to the generation of subaerial (type-1) unconformities. Subsidence and sedimentation are more or less continuous but may vary in rate. This area they term zone A. In more distal regions, the rate of flexural subsidence is lower and, in this area, termed zone B, eustatic fall may outpace flexural subsidence, and a type-1 unconformity will be the result. The distal margin of the basin is marked by a tectonic hinge, beyond which vertical motion of the forebulge is in the opposite direction to that of the basin. The location where the rate of tectonic subsidence and the rate of eustatic fall within the basin are equal is termed the equilibrium point. It moves basinward of the tectonic hinge during times of falling sea level and toward the forebulge during times of rising sea level. The boundary between zone A and zone B is defined as the most proximal position reached by the equilibrium point, which it will attain at the inflection point during the falling leg of the sea-level curve (Fig. 8.4). No time scales were suggested for this model. Long-term foreland basin subsidence extends over tens of millions of years, but localized loading events may generate cycles of subsidence and uplift over much shorter time periods (Table 8.2; Sect. 8.2.4.2), creating

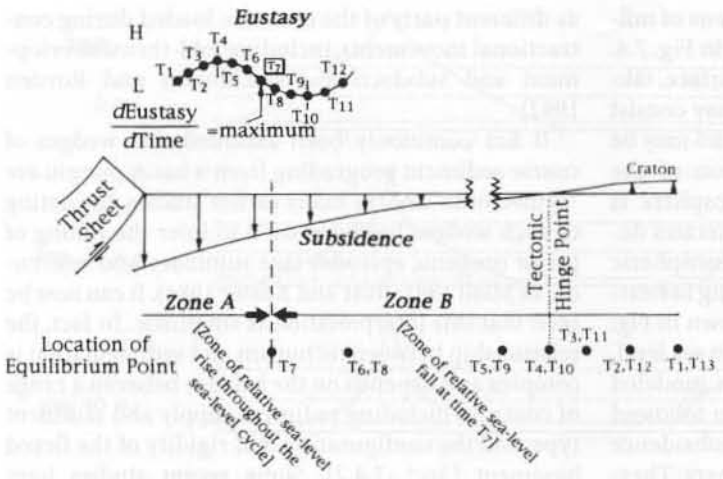


Fig. 8.4. The foreland-ramp model of Posamentier and Allen (1993). The rate of flexural subsidence decreases away from the thrust sheet. The equilibrium point is the point where the rate of flexural subsidence and the rate of sea-level change are equal. It moves across the basin as sea level rises and falls, as shown by the position of points *T1* to *T13*. The location of the most proximal position of the equilibrium point defines the boundary between zones *A* and *B*

the potential for tectonically driven cyclicity with a wide range of frequencies.

Subduction generates secondary mantle convection above the downgoing slab. The cold downwelling current induces a "dynamic load" that widens and deepens the foreland basin. Variations in the subsidence history cause variations in the magnitude of this process, which seems capable of generating broad uplift and downwarp movements over time scales in the tens of millions of years over basin widths of hundreds to a few thousands of kilometers (Sect. 7.5).

Other local tectonic processes that may affect relative sea level and develop tectonically related sequences, and systems tracts include strike-slip faulting, fault-block rotation, diapirism, the rise of mantle plumes, etc. Examples of these processes are discussed by Miall (1997, Sect. 11.2.2), particularly with respect to the sequence stratigraphy of the North Sea Basin.

8.2.2.3 *Conclusions*

The processes discussed here that lead to eustatic sea-level changes, such as changing rates of sea-floor spreading, also lead to tectonic adjustments of the continents (e.g., initiation of the rifting–thermal subsidence cycle, orogenic suture and uplift, subsidence

Table 8.2. 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).

over cold subducting slabs) so that actual changes in sea level in any given basin may have multiple causes that are not necessarily readily interpretable in terms of eustasy. Much work remains to be done to document and globally correlate these cycles in order to evaluate the various mechanisms. The fact that these various processes are not independent and may be in or out of phase depending on local plate-tectonic and basinal conditions cautions against simplistic models of stratigraphic responses to sea-level change.

8.2.3 Mechanisms with Million-Year Episodicities

Cycles of this type form the main basis for the Exxon global-cycle chart (Haq et al. 1987), but the existence of a global framework of such cycles has been questioned (Miall 1997, Part IV; Sect. 8.4), and no satisfactory mechanism of synchronous global change on a million-year time scale has been demonstrated. An increasing body of evidence is indicating that cycles of this type can be generated by tectonic mechanisms and are, therefore, likely to be regional to continental in extent, but not global. If this is the case, they cannot be used as the basis for a global stratigraphic time scale.

8.2.3.1 *Intraplate Stress*

The **intraplate-stress** model of Cloetingh (1988) (also termed **in-plane stress;** Sect. 7.6) is finding widespread application in the study of 10^6-year stratigraphic cyclicity. It is based on the concept that the lateral movement of tectonic plates exerts lateral stresses on the crust that are expressed as low-amplitude, long-wavelength flexures. Cloetingh et al. (1985) were the first to recognize the significance of in-plane stress as a control on basin architecture. They argued "that variations in regional stress fields acting within inhomogeneous lithospheric plates are capable of producing vertical movements of the Earth's surface or the apparent sea-level changes ... of a magnitude equal to those deduced from the stratigraphic record."

The important contribution which Cloetingh et al. (1985) made was to demonstrate, using numerical modeling, that horizontal stresses modify the effects of existing, known, vertical stresses (thermal and flexural subsidence, sediment loading) on sedimentary basins, enlarging or reducing the amplitude of the resulting flexural deformation. The principles are illustrated in Fig. 8.5. Cloetingh and coworkers demonstrated that a horizontal stress of 1–2 kbar, well within the range of calculated and observed stresses resulting from plate motions, may result in a local uplift or subsidence of up to 100 m, at a rate of up to 0.1 m/ka. Compressional stresses generate uplift of the flanks of a sedimentary basin and increased subsidence at the center. Extensional stresses have the reverse effect (Fig. 8.5). The magnitude of the effect varies with the flexural age (rigidity) of the crust as well as the magnitude of the stress itself.

The stratigraphic results of this process are extremely important. Figure 8.6 models the effects of imposing a horizontal stress of 500 bars on a continental margin undergoing long-term, thermally induced subsidence. Cloetingh (1988) stated, with reference to this model:

> When horizontal compression occurs, the peripheral bulge is magnified while simultaneously migrating in a seaward direction, uplift of the basement takes place, an offlap develops, and an apparent fall in sea level results, possibly exposing the sediments to produce an erosional or weathering horizon. Simultaneously, the basin center undergoes deepening (Fig. 8.6b), resulting in a steeper basin slope. For a horizontal tensional intraplate stress field, the flanks of the basin subside with its landward migration producing an apparent rise in sea level so that renewed deposition, with a corresponding facies change, is possible. In this case the center of the basin shallows (Fig. 8.6c), and the basin slope is reduced.

These stratigraphic features are locally identical in architecture and duration to the "third-order" onlap and offlap patterns from which Vail and his coworkers derived their global cycle charts (Sect. 8.4) although, on a basinal scale, sequence architectures show important differences. For example, build-up of contractional stresses over an extensional margin results in uplift of the continental shelf, and offlap, and may lead to the generation of an erosional unconformity at the same time the basin center undergoes deepening. Increased tilting of the continental margin may increase the tendency for slope failure and mass wasting, with increased potential for large-scale submarine-fan deposition at the base of the slope. Conversely, an increase in extensional stress may result in flooding (enhanced onlap) of the continental shelf and uplift of the basin floor. Some submarine unconformities may be the result of changes in the erosional patterns of oceanic currents brought about by such changes in ocean-floor configuration. The typical indicators of relative sea-level change that are used in sequence analysis (unconformities, onlap, offlap, transgressions, progradation, retrogradation, etc.) may, therefore, be generated by changes in regional stress regime and could be out of phase by a

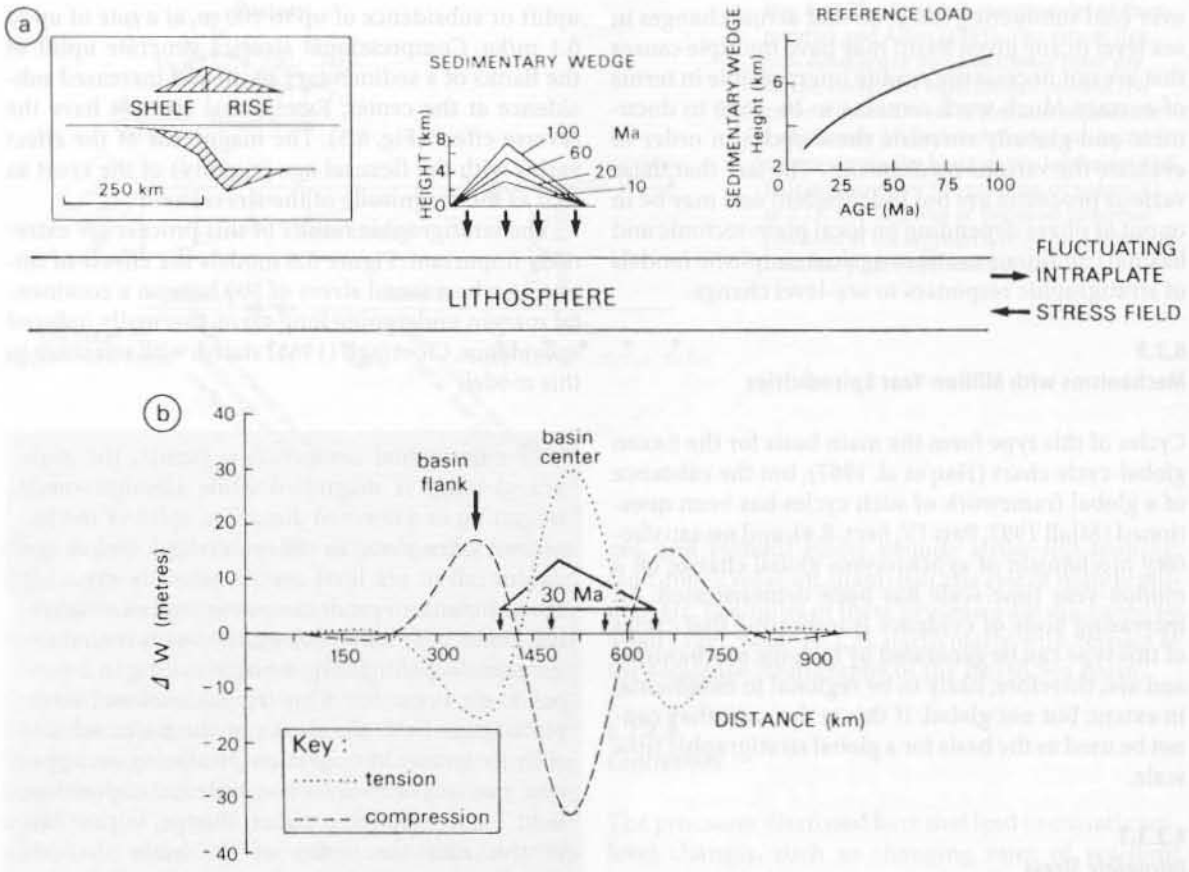


Fig. 8.5 A, B. Intraplate stress: principles of the geophysical model. **A** At *left* a simple shelf-slope-rise sedimentary wedge at a continental margin is shown. The thickness and age of this wedge are shown at *center* and *right*. The resulting flexural deformation attributable to a horizontal stress of 1 kbar acting on this continental margin is shown in **B**. Compressional and tensional stresses yield equal but opposite effects. (Cloetingh 1988)

half cycle between the basin margin and the basin center, although this may be impossible to demonstrate from the limited chronostratigraphic evidence that is commonly available. The kind of local, outcrop-scale analysis that is used as the basis for many sequence analyses (Van Wagoner et al. 1990) could potentially, therefore, give very misleading results. As Karner (1986) pointed out,

> ... basin margin and interior regions should experience opposite baselevel movements. The nature and distribution of sediment cyclicity across either a passive margin or intracratonic basin therefore offers an excellent opportunity to test the concept and importance of lateral stress-induced baselevel variations.

Figure 8.7 illustrates simplified, hypothetical subsidence curves for three positions within a basin undergoing thermally-induced subsidence in which the subsidence has been affected by changes in intraplate compressive stress. As pointed out by Cathless and Hallam (1991) in opposition to Cloetingh's hypotheses, the elevation changes produced within any given sedimentary basin are small in areal extent, limited by the flexural wavelength of the basement underlying the basin. However, the point they missed themselves is that the actual intraplates stresses are much more widespread – they may be transmitted for thousands of kilometers and will simultaneously affect all basins within that plate.

Cloetingh and Kooi (1990) developed a paleostress curve for the U.S. Atlantic margins based on studies of regional plate kinematics and developed these ideas further in a subsequent paper (Cloetingh and Kooi 1992). These studies indicate changes in plate configuration and rotation poles at about 2- to 16-m.y. intervals. Such changes would have resulted in changes in the in-plane stress and flexural responses

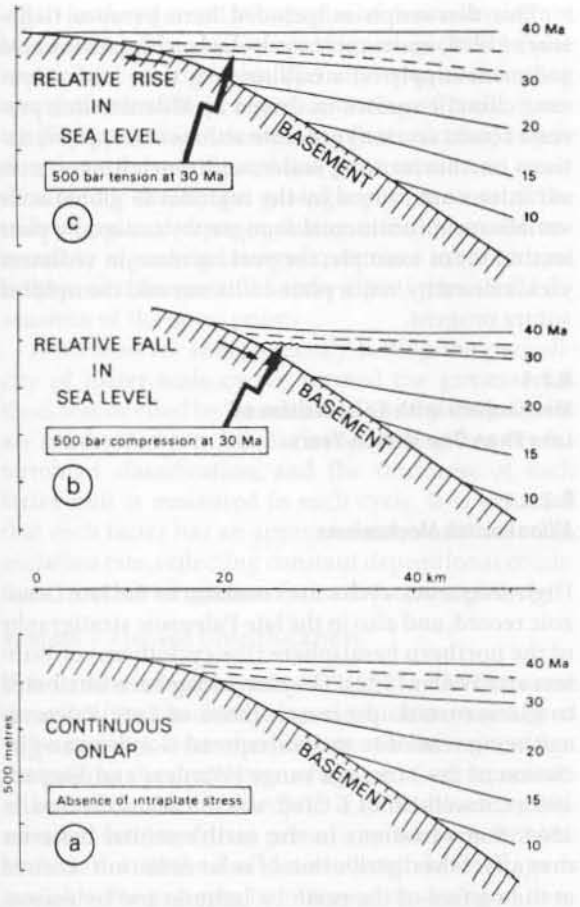


Fig. 8.6 A–C. Idealized stratigraphy at a basin margin underlain by 40 Ma of lithosphere. **A** The result of continuous onlap as a result of long-term cooling and flexural loading in the absence of intraplate stress (cf. Fig. 7.12). In **B**, imposition of a 500-bar compressive intraplate stress at 30 Ma induces short-lived uplift, offlap, and the development of an unconformity on the basin flank. In **C**, extensional stress generates enhanced subsidence and an increase in the rate of onlap. The stratigraphy in **B** and **C** indicate a relative fall and rise in sea level, respectively. (Cloetingh 1988)

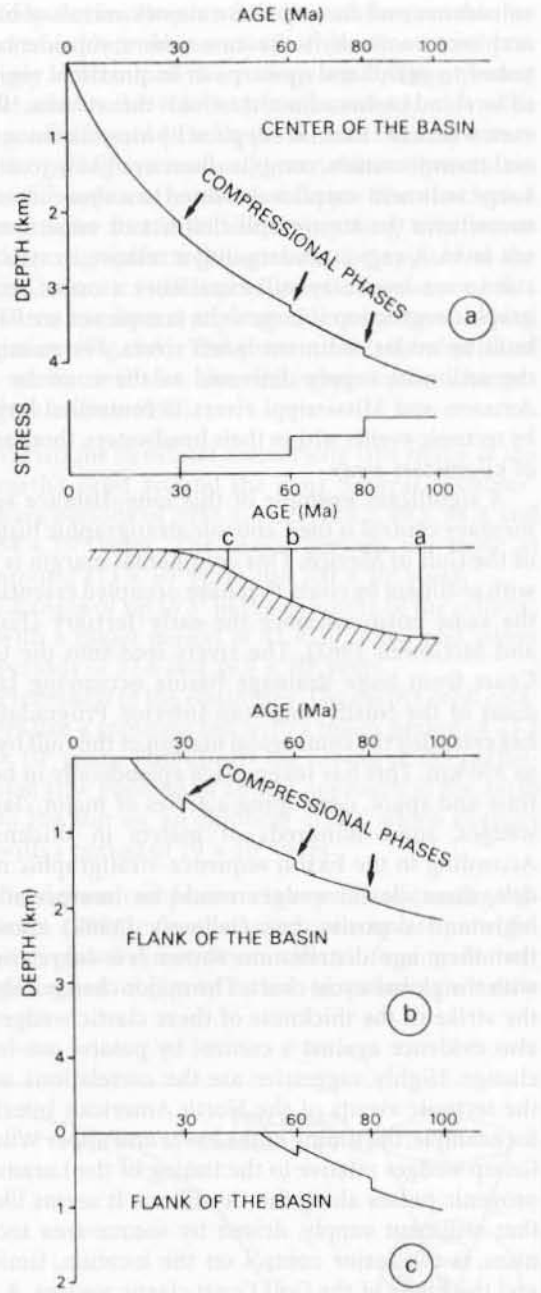


Fig. 8.7 A–C. Effect of intraplate compressive stress on subsidence curves at three different positions within a sedimentary basin. In the basin center (**A**), the effect of compression will be to enhance subsidence, whereas on the flanks of the basin (**B**, **C**), enhanced subsidence changes to uplift as the flexural node migrates during long-term flexural widening of the basin. (Cloetingh 1988)

of the continents, resulting in regional transgressions and regressions. In the case of foreland basins, it has long been known that flexural loading by nappes and thrust-sheet stacks is the main cause of long-term (10^7-year) basin subsidence (Price 1973; Beaumont 1981; Jordan 1981), but it is now being suggested that intraplate stress may act on much shorter time scales in response to load changes brought about by the movement of individual structures (see next section).

8.2.3.2 *Sediment Supply*

Major variations in sediment supply may have a significant impact on sequence architecture. Sediment supply is controlled primarily by tectonics and climate. In geologically simple areas, where the basin is fed directly from the adjacent margins and source-area uplift is related to basin subsidence, supply considerations are likely to be directly correlated to basin

subsidence and eustasy as the major controls of basin architecture. Such is the case where subsidence is yoked to peripheral upwarps or in proximal regions of foreland basins adjacent to fold–thrust belts. However, where the basin is supplied by long-distance fluvial transportation, complications are likely to arise. Large sediment supplies delivered to a shoreline may overwhelm the stratigraphic effects of variations in sea level. A region undergoing a relative or eustatic rise in sea level may still experience a major stratigraphic regression if large delta complexes are being built by major sediment-laden rivers. For example, the sediment supply delivered to the coast by the Amazon and Mississippi rivers is controlled largely by tectonic events within their headwaters, thousands of kilometers away.

A significant example of this long-distance sedimentary control is the Cenozoic stratigraphic history of the Gulf of Mexico. This continental margin is fed with sediment by rivers that have occupied essentially the same positions since the early Tertiary (Fisher and McGowen 1967). The rivers feed into the Gulf Coast from huge drainage basins occupying large areas of the North American Interior. Progradation has extended the continental margin of the Gulf by up to 350 km. This has taken place episodically in both time and space, developing a series of major clastic wedges, some hundreds of meters in thickness. According to the Exxon sequence-stratigraphic models, these clastic wedges would be interpreted as highstand deposits, but Galloway (1989) showed that their age distribution shows few correlations with the global-cycle chart. The major changes along the strike of the thickness of these clastic wedges is also evidence against a control by passive sea-level change. Highly suggestive are the correlations with the tectonic events of the North American Interior, for example, the timing of the lower and upper Wilcox Group wedges relative to the timing of the Laramide orogenic pulses along the Cordillera. It seems likely that sediment supply, driven by source-area tectonism, is the major control on the location, timing, and thickness of the Gulf Coast clastic wedges. A secondary control is the nature of local tectonism on the continental margin itself, including growth faulting, evaporite diapirism, and gravity sliding. As shown by Shaub et al. (1984) and as re-emphasized by Schlager (1993), variations in deep-marine sediment dispersal in the Gulf of Mexico show patterns very similar to the coastal and fluvial variations illustrated by Galloway (1989). Large-scale submarine-fan systems are, therefore, also dependent on considerations of long-term sediment-supply variation, which may be controlled by plate-margin tectonism, in-plane-stress regime, and dynamic topography.

This discussion is included here because Galloway's (1989) instructive study deals with variations in sediment supply on a million-year time scale. However, climatic variations driven by Milankovitch processes (next section) generate sediment-supply variations on shorter time scales, and much longer-term variations are caused by the regional to global scale variations in continental topography caused by plate tectonics, for example, the vast increase in sediment yield caused by major plate collisions and the uplift of suture orogens.

8.2.4 Mechanisms with Episodicities of Less Than One Million Years

8.2.4.1 *Milankovitch Mechanisms*

High-frequency cycles are common in the late Cenozoic record, and also in the late Paleozoic stratigraphy of the northern hemisphere (the cyclothems of Wanless and Weller 1932). They have long been attributed to glacioeustatic processes, those of Late Paleozoic age being related to the widespread Gondwanan glaciation of the same age range (Wanless and Shepard 1936; Crowell 1978). J. Croll was the first to realize, in 1864, that variations in the earth's orbital behavior may affect the distribution of solar radiation received at the surface of the earth by latitude and by season, and the Serbian mathematician Milutin Milankovitch carried out the necessary calculations in the 1930s. Emiliani (1955) was the first to obtain direct observational evidence for the processes, in the form of variations in the marine, Pleistocene, oxygen-isotope record. As noted above, orbital forcing, the so-called Milankovitch processes, are now universally accepted as the cause of high-frequency (10^4- to 10^5-year) periodicity (Berger et al. 1984; de Boer and Smith 1994a). In fact, as noted above, cyclostratigraphy is becoming the basis for a new standard of geologic time for the late Cenozoic.

However, as discussed earlier (see also below), it is now known that tectonic processes can drive cycles of base-level change with a periodicity within the range of what is commonly (but, in this case, misleadingly) referred to as the Milankovitch time band. In fact, as demonstrated by Algeo and Wilkinson (1988), many autogenic and allogenic processes generate cycles that can be demonstrated to have time spans within the Milankovitch time range and, in a later paper, Wilkinson et al. (1998) demonstrated that frequency spectra very similar to those associated with Milankovitch periodicity may be generated by entirely different processes. In this later study, they demonstrated this point by analyzing the frequency of goal scor-

ing in hockey games! Where glacial processes cannot be directly proven by isotopic signatures, it is unwise to attribute stratigraphic cyclicity to orbital-forcing mechanisms without rigorous testing of cyclic spectra (Fischer 1986). Attempts to relate calculated cycle durations to specific orbital parameters are questionable for the distant geologic past because of the wide margin of error in dating techniques and uncertainty regarding changes over time in the orbital frequencies. Berger and Loutre (1994) provided a discussion of the latter point.

A method for independently testing the periodicity of meter-scale cycles, termed the gamma method, was devised by Kominz and Bond (1990). Cycles are subdivided into lithofacies according to a predetermined classification, and the thickness of each facies unit is measured in each cycle. It is assumed that each facies has an approximately constant accumulation rate, reflecting constant depositional conditions. A value for gamma for each facies is given by

gamma = elapsed time/thickness.

If it can be assumed that each cycle has the same duration (the absolute value of which does not need to be known), a set of equations can be written that determines the proportion of elapsed time represented by each facies in each cycle. If these values are reasonably consistent, then periodicity has been demonstrated. If a method of determining absolute age is available, for example, by radiometric dating of bentonites that bracket the section under study, then the absolute duration of the cycle period can be determined and, if appropriate, this can be related to orbital frequencies. Kominz (1996) used the gamma method to explore the accuracy of the time series previously determined for a deep-sea drilling project drill core in the North Atlantic and found a wide range of errors, indicating the inherently noisy nature of the geologic record, e.g., variable sedimentation rates for apparently identical facies. The conclusion seems to be that it will be a very long time before cyclostratigraphic techniques can be applied to the older Cenozoic (still less the pre-Cenozoic) record.

There are several separate components of orbital variation (Fig. 8.8). The present orbital behavior of the earth includes the following cyclic changes (Schwarzacher 1993):

1. Variations in orbital eccentricity (the shape of the earth's orbit around the sun). Several "wobbles" have periods of 2035.4, 412.8, 128.2, 99.5, 94.9, and 54×10^3 years. The major periods are those at around 413×10^3 years and 100×10^3 years.
2. Changes of up to 3° in the obliquity of the ecliptic, with a major period of 41×10^3 years and minor periods of 53.6 and 39.7×10^3 years.
3. Precession of the equinoxes. The earth's orbit rotates like a spinning top, with a major period of 23.7×10^3 years. This affects the timing of the perihelion (the position of closest approach of the earth to the sun on an elliptical orbit), which changes with a period of 19×10^3 years.

Each of these components is capable of causing significant climatic fluctuations, given an adequate degree of global sensitivity to climate forcing. For example, when obliquity is low (rotation axis nearly

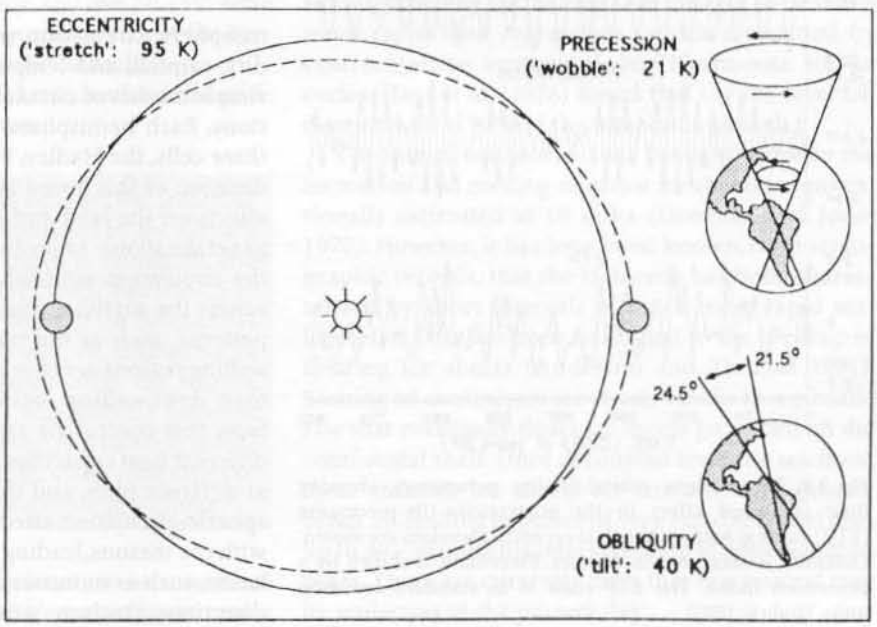


Fig. 8.8. Perturbations in the orbital behavior of the earth, showing the causes of Milankovitch cyclicity. (Plint et al. 1992, after Imbrie and Imbrie 1979)

normal to the ecliptic), more energy is delivered to the equator and less to the poles, giving rise to a steeper latitudinal temperature gradient and lower seasonality. Variations in precession alter the structure of the seasonal cycle by moving the perihelion point along the orbit. This changes the earth–sun distance at every season, thus changing the intensity of insolation at each season. "For a given latitude and season typical departures from modern values are on the order of ±5%" (Imbrie 1985). Because the forcing effects have different periods, they go in and out of phase. One of the major contributions of Milankovitch was to demonstrate these phase relationships on the basis of laborious time-series calculations. These can now, of course, be readily carried out by computer (Fig. 8.9). The success of modern stratigraphic work has been to demonstrate the existence of curves of temperature change and other variables in the Cenozoic record that can be correlated directly with the curves of Fig. 8.9. For this purpose, sophisticated time-series spectral analysis is performed on various measured parameters, such as oxygen-isotope content (Fig. 8.10) or cycle thickness. These methods are the basis for cyclostratigraphy (House 1985; Gale 1998).

How do orbital variations affect climate? The major control of global climate is the coupled ocean–atmospheric circulation, which, in turn, controls humidity, rainfall, and temperature. Figure 8.11 presents simple models of circulation for three climatic conditions. Each hemisphere of the earth is encircled by three cells, the Hadley, Ferrel, and Polar cells. Consideration of this broad atmospheric structure and its effects on the land and sea lead to a series of useful generalizations regarding regional climate. Where the circulation established by the three cells passes across the earth's surface, it sets up persistent wind patterns, such as the trade winds. Atmospheric upwelling regions are characterized by greater humidity than downwelling regions because of adiabatic effects that control air saturation. Land and sea have different heat capacities and, therefore, heat and cool at different rates, and this can set up regional atmospheric-circulation effects. The cells change position with the seasons, leading to seasonal changes in circulation, such as monsoonal changes in prevailing wind directions. Onshore winds carry moisture, which is

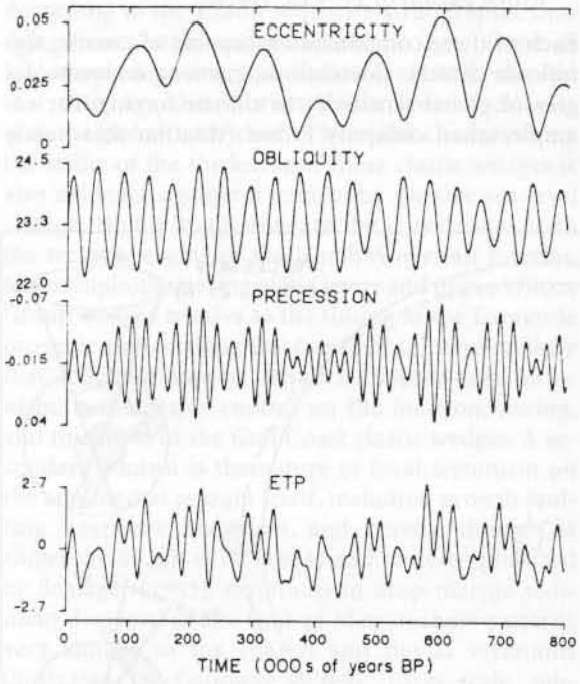


Fig. 8.9. Three major orbital-forcing parameters, showing their combined effect in the eccentricity–tilt–precession (ETP) curve at *bottom*. Absolute eccentricity values are shown. Obliquity is measured in degrees. Precession is shown by a precession index. The ETP scale is in standard-deviation units. (Imbrie 1985)

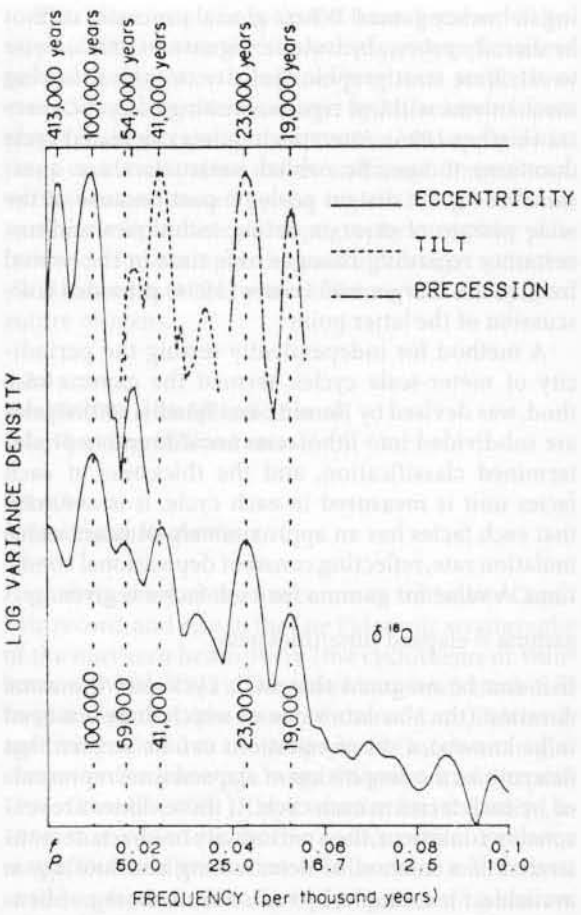


Fig. 8.10. Variance spectra over the past 800×10^3 years of orbital variations (*top*; from data shown in Fig. 8.9), and ^{18}O content, measured in foraminiferal tests (*bottom*). Note the close correlation between the two curves. (Imbrie 1985)

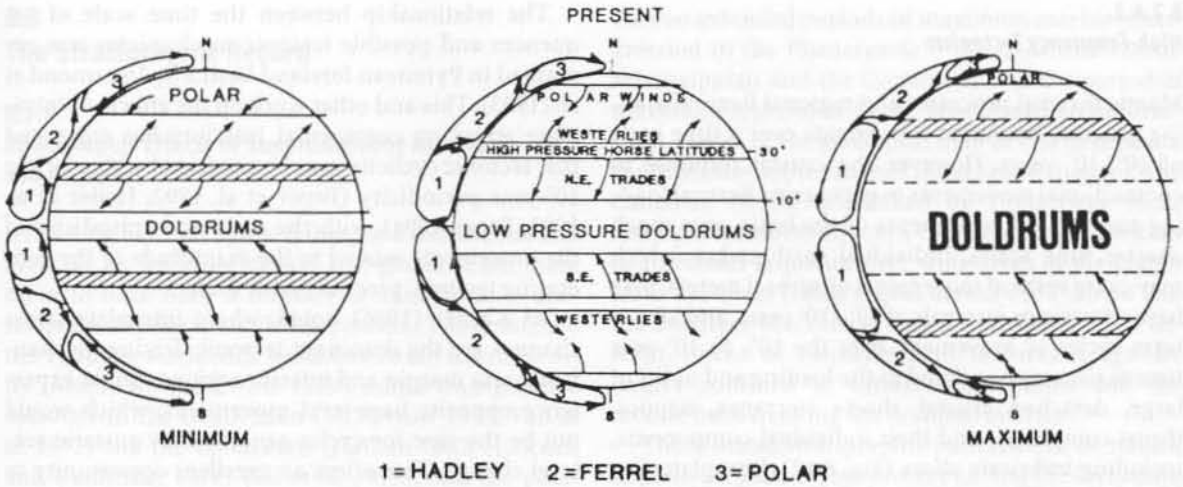


Fig. 8.11. Atmospheric circulation patterns for three climatic conditions. The present climate (*center*) is intermediate between true glacial and interglacial conditions. (Perlmutter and Matthews 1990)

released as rainfall when forced to rise over a high relief. Windward slopes, therefore, tend to be wetter than leeward slopes, and the climate on the east and west sides of continents may be quite different.

During climatic minimums, the downwelling region between the Hadley and Ferrel cells is located at about 15–35° latitude and is an area of warming and high evaporation rates. Deserts are, therefore, characteristic. Monsoon zones show little latitudinal shifting with the seasons, which also limits the amount of moisture transported to this zone. As conditions change to the climatic maximum, the Hadley–Ferrel downwelling zone shifts poleward, and the belt of deserts moves to 35–40°. Hadley and monsoonal circulation increase in efficiency, and the 10–35° zone becomes more humid. The upwelling arm of the Ferrel–Polar cell moves to about 70° latitude.

de Boer and Smith (1994b) summarized the orbital effects on climate as follows:

> At low latitudes, close to the equator, the influence of the cycle of precession, modulated by the varying eccentricity of the Earth's orbit, is dominant and causes latitudinal shifts of the caloric equator ... In turn, this causes significant shifts of the boundaries between adjacent climate zones. At mid-latitudes (20–40°) the orbital variations affect the relative length of the seasons and the contrast between summer and winter, and hence of monsoon intensity ... Toward higher latitudes (>40°) the effect of the varying obliquity becomes more prominent.

In detail, the response of the climate to astronomical forcing at any point on the earth's surface is extremely complex, reflecting various sensitivity factors, such as the relative positions and sizes of sea and land masses. Their latitude affects air and water circulation patterns, and their relative position affects humidity (and, hence, rainfall), monsoon effects, and so on. These complexities are exemplified by Barron's 1983 discussion of the difficulties inherent in attempting to reconstruct the warm, equable climates of the Cretaceous. In general, there is a lag effect between the astronomical force and the system's response, although, on the kiloyear scale under consideration in this chapter, the lag effect is probably relatively unimportant. An exception is the buildup and melting of major ice sheets. In general, glacioeustatic transgressions (caused by ice melting) are thought to be much more rapid than regressions (which are caused by continental ice formation). For Pleistocene 100-ka cycles, Hays et al. (1976) found that the sea-level fall occupied 85 to 90% of the total cycle period.

The rate of sea-level change brought about by the formation and melting of major ice sheets is conventionally estimated at 10 m/ka (Donovan and Jones 1979). However, it has long been known, from stratigraphic records, that the Holocene has been characterized by short intervals of much more rapid sea-level rise. This has been attributed to the breakup of floating ice sheets (Anderson and Thomas 1991). Melting of continental ice sheets results in a eustatic rise that eventually floats ice sheets grounded on the continental shelf. Once decoupled from the sea floor, these massive ice sheets are unstable and quickly break up, leading to pulses of very rapid sea-level rise, up to 30–50 m/ka for intervals of up to a few hundred years. There are currently fears that this process may be underway at the present day.

8.2.4.2 *High-Frequency Tectonism*

Mantle thermal processes and regional flexural loading generate basement movements over a time scale of 10^6–10^7 years. However, the crustal response to contractional movements may generate flexural loading and vertical movements of the basin over much shorter time scales. Individual earthquakes, which may cause vertical movements of several meters, may have recurrence intervals of 10^1–10^2 years, and longer term cycles of movement, over the 10^3- to 10^5-year time scale, may be related to the loading and uplift of large, detached crustal sheets (terranes, nappes, thrust complexes) and their individual components, including imbricate slices (Fig. 8.12). Intraplate (in-plane) stress may transmit the effects of such tectonism throughout the basin and beyond. Ancient foreland basins contain many examples of stratigraphic sequences formed over 10^4- to 10^6-year time scales, and an examination of these processes is, therefore, essential.

The relationship between the time scale of sequences and possible tectonic mechanisms was examined in Pyrenean foreland basins by Deramond et al. (1993). This and other work on the effects of intraplate stress on continental interiors has suggested that tectonic cyclicity may be generated with a 10^4- to 10^7-year periodicity (Peper et al. 1992; Heller et al. 1993; Peper 1994), with the scale and episodicity of the movements related to the magnitude of the generating tectonic process (Table 8.2).

As Karner (1986) noted, where intraplate stress changes are the dominant tectonic driving mechanism, basin margin and interior regions should experience opposite base-level movements, which would not be the case for cycles generated by eustatic sea-level change. This offers an excellent opportunity to test generative mechanisms through careful dating and correlation of the stratigraphic record but, as noted above, except in certain special cases, this is still beyond our ability. Possible examples of the evidence of the intraplate-stress mechanism in the stratigraphic record are discussed in Section 8.3.

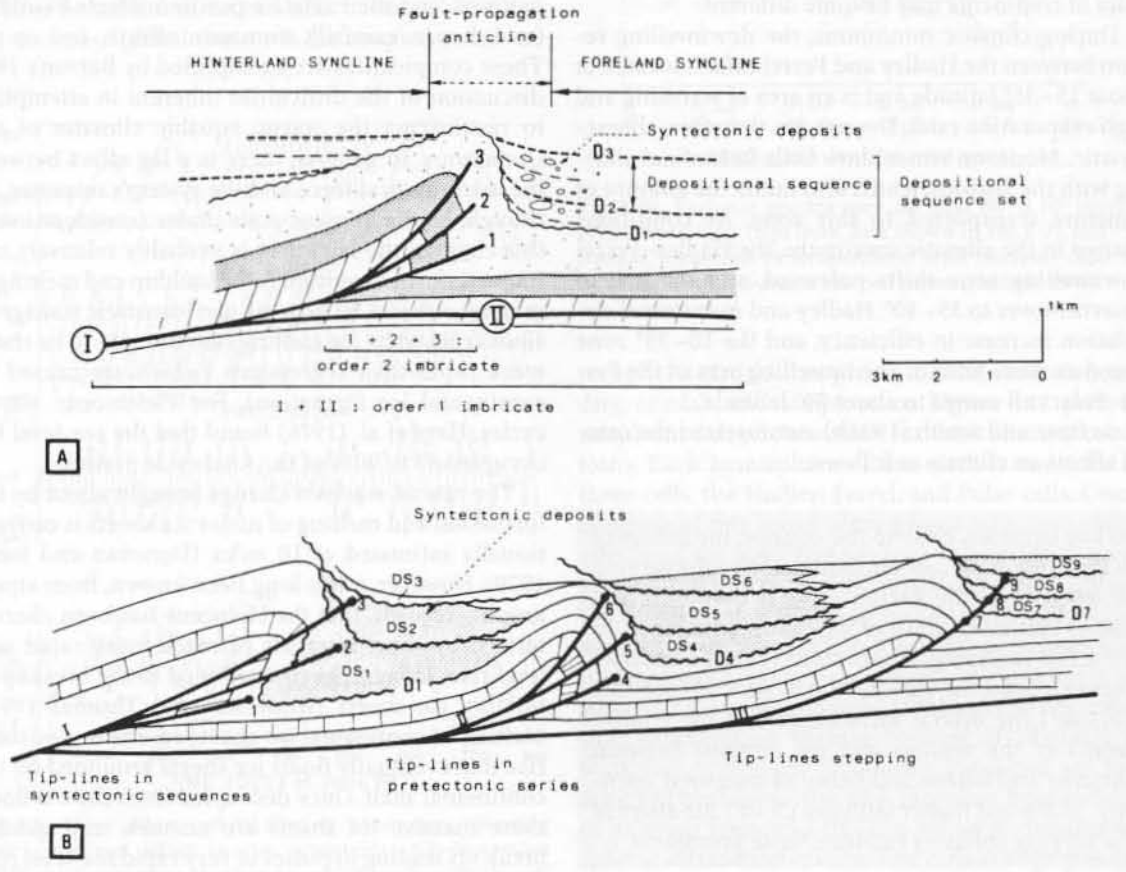


Fig. 8.12 A, B. The relationship between the development of a fold–thrust belt and the stratigraphy of the adjacent foreland basin. Unconformities, numbered *D1–D9*, develop as imbricate thrust slices develop. Uplift of each slice is recorded by a corresponding numbered unconformity in the basin. (Deramond et al. 1993)

8.3 The Stratigraphic Record

8.3.1 Stratigraphic Effects of Supercontinent Assembly and Dispersal

It has long been recognized that, on a continental and even on an intercontinental and global scale, there seem to have been a number of synchronous, correlatable stratigraphic developments. These include the virtually worldwide basal Cambrian unconformity (Matthews and Cowie 1979), major transgressive episodes in the Ordovician (McKerrow 1979; Vail et al. 1977) and the Cretaceous (Hallam 1963; Hancock and Kauffman 1979; Vail et al. 1977), and the post-Cretaceous regression (Hallam 1963). The late Cretaceous transgression was first recognized as being exceptionally prominent by Suess (1906). He suspected that it was only one of several such events in the Phanerozoic related to worldwide changes in sea level, and he named such movements eustatic.

Taking a longer view, such workers as Holmes, Gastil, Dearnley, Stockwell, Sutton, Fitch, and Miller demonstrated a global episodicity in periods of magmatism and plutonism throughout geologic time (Williams 1981). Sutton (1963) coined the term **chelogenic cycle** for this episodicity.

A cycle of sea-level rise and fall with a duration of about 300 m.y. is illustrated in Fig. 8.13. The method by which Vail and his coworkers constructed this curve is discussed in Section 8.4. The cycles include the two extended periods of maximum marine transgression in the Phanerozoic – the late Cambrian to Mississippian and the Cretaceous – and a period of maximum regression in the Pennsylvanian to Jurassic. A glance at the geological map of North America confirms the importance of these broad changes. The Canadian Shield is flanked by Cretaceous rocks resting unconformably on a Cambrian or Ordovician to Devonian sequence over wide areas of the craton from the Great Lakes region across the Prairies into the Beaufort-Mackenzie region and the Arctic Platform. Rocks of Pennsylvanian to Jurassic age are largely confined to intracratonic basins and the mobile belts flanking the cratonic interior.

These broad stratigraphic patterns can be related to plate-tectonic events accompanying the formation and breakup of Pangea. Worsley et al. (1984), following the earlier work of Sutton, Condie, and Windley, argued that the formation of a supercontinent creates a thermal blanket that inhibits convective release of radiogenic heat from the mantle. The development of a thermal blanket may be the cause of the eventual breakup of the supercontinent following the establishment of a new pattern of mantle convection (Sect. 8.2.1).

The total length of rifting continental margins and sea-floor spreading centers increases during the breakup of a supercontinent and is accompanied by active subduction, plutonism, and arc volcanism on the outer, convergent plate margins of the dispersing fragments. Major eustatic transgressions occur because of the displacement of ocean waters by ther-

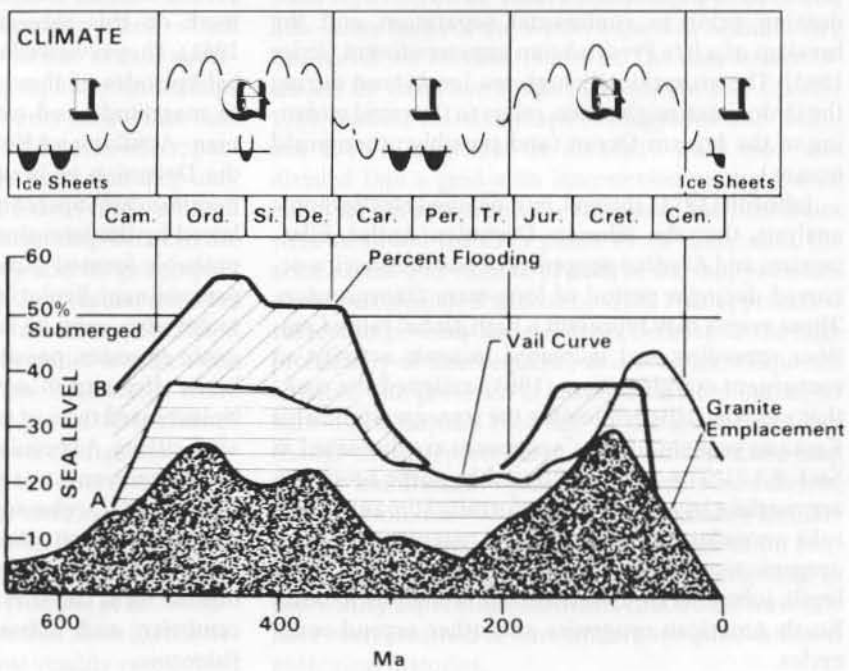


Fig. 8.13. First-order cycles during the Phanerozoic, including sea-level change (from Vail et al. 1977), percent flooding of the continents (from Fischer 1981), and volume of granite emplacement (from Engel and Engel 1964). The generation of the sea-level curve is discussed in Section 8.3. This diagram, which was compiled by Worsley et al. (1984), also shows fluctuations in global climate that accompanied these changes. *I* Icehouse climate; *G* greenhouse climate (discussed in text)

mally elevated young oceanic crust and active spreading centers in the new Atlantic-type oceans. Heller and Angevine (1985) argued that, during the first 50–100 m.y. after the initiation of the breakup of a supercontinent, the global average age of oceanic crust decreases because of the active development of Atlantic-type oceans. This will lead to a rise in sea level, without any change in global average spreading rate.

At times of continental assembly, low average rates of spreading may occur, indicating ridge reordering following major continental collision and suturing events. Collision results in crustal shortening, which has the effect of increasing the ocean-basin volume. At the end of a supercontinent assembly cycle, large areas of old, and therefore cool, and subsided oceanic crust will underlie the world's oceans (Worsley et al. 1984). Suturing might, therefore, be expected to correlate with, or precede, times of low sea level (Valentine and Moores 1970, 1972; Larson and Pitman 1972; Vail et al. 1977; Schwan 1980; Heller and Angevine 1985).

With these generalizations in mind, it is instructive to examine briefly the Phanerozoic record. The Cambrian transgression may reflect the breakup of the Eocambrian supercontinent (Matthews and Cowie 1979; Donovan and Jones 1979). It has been suggested that the late Precambrian Sparagmite sequence of Norway was formed in rifts representing the incipient Iapetus (proto-Atlantic) Ocean (Bjorlykke et al. 1976). Late Proterozoic glaciation, the evidence for which is widespread in Greenland and Scandinavia, may be related to the formation of regional ice caps on rift margins elevated by thermal doming prior to continental separation and the breakup of a late Precambrian supercontinent (Eyles 1993). The exceptionally high sea-level stand during the Ordovician might, then, relate to the rapid widening of the Iapetus Ocean (and possibly other world oceans).

Johnson (1971) showed, by a detailed stratigraphic analysis, that the Silurian–Devonian Antler, Ellesmerian, and Acadian orogenies of North America occurred during a period of long-term transgression. These events may represent a high global rate of sea-floor spreading and increased tectonic activity on convergent margins. Sloss (1963) assigned the rocks that were deposited following the transgression to his Kaskasia sequence (Sloss' sequences are discussed in Sect. 8.3.2). The rocks deformed by these orogenies are overlain by a major unconformity (the sub-Absaroka unconformity of Sloss 1963), suggesting a post-orogenic reordering of spreading axes and a fall of sea levels. Johnson (1971) extended this analysis to other North American orogenies and other second-order cycles.

A long-term period of low sea level occurred following the Caledonian–Acadian and Hercynian–Appalachian suturing of Pangea between the Devonian and Permian (Schopf 1974), probably corresponding to a time of slow spreading. The drifting of this supercontinent over the south pole is thought to have been the cause of increasingly continental-type climates, leading to the Late Devonian–Permian glaciations of Gondwana (Crowell 1978; Caputo and Crowell 1985; Eyles 1993) and the widespread Pennsylvanian to Jurassic eolian facies of the United States and Europe (Kocurek 1988).

Pangea began to fragment in the Triassic, and this is the cause of the very widespread suite of rift basins filled with nonmarine (fluvial, lacustrine) deposits in eastern North America, North Africa, and western Europe. Worldwide transgressions during the Jurassic and Cretaceous probably reflect the progressive splitting of Pangea. North America and Africa rifted apart in the mid-Jurassic, South America and Africa in the Early Cretaceous; North America and northwestern Europe split in the mid-Cretaceous, as did Africa and Antarctica; India and Madagascar rifted apart in the Late Cretaceous, and the North Atlantic split extended northward between Greenland and Scandinavia in the early Paleocene (summary and data sources in Bally and Snelson 1980; Uchupi and Emery 1991).

There is currently considerable discussion regarding the nature of possible earlier (Precambrian) supercontinent cycles. This is beyond the scope of the present book. Worsley et al. (1984) argued that there is fragmentary evidence for several first-order cycles during the Precambrian back to at least 2 Ga (earlier work on this subject is summarized by Williams 1981). They reviewed the evidence for four major global episodes of thermotectonic activity comparable in magnitude and extent to the combined Caledonian–Acadian and Hercynian–Appalachian events of the Devonian to Permian. These orogenic episodes may indicate supercontinent assembly. They were followed by the intrusion of major dike swarms, which probably formed during the initial rifting of the supercontinent. Evolutionary milestones in the biologic realm also seem to have postdated the major orogenic episodes, possibly indicating the explosion of biotic diversity in newly flooded shelf seas, caused by increased rates of sea-floor spreading 50–100 m.y. after rifting. A speculative reconstruction of late Precambrian events was offered by Hoffman (1991) and Dalziel (1992), who suggested that, at around 700 Ma, Laurentia was situated at the center of a supercontinent that subsequently dispersed and "turned inside-out" to form Gondwana around the end of the Precambrian and, subsequently, Pangea in the mid-Paleozoic.

8.3.2 Cycles with Episodicities of Tens of Millions of Years

8.3.2.1 *Cratonic Cycles*

The wealth of subsurface data in North America led Sloss et al. (1949) to recognize four continent-wide stratigraphic sequences in the Cambrian to Jurassic. Later, this list was expanded to six to include the Jurassic to Recent (Sloss 1963). Subsequently, detailed comparisons have been made between the Phanerozoic record of Western Canada and the Russian Platform (Sloss 1972) and between selected intracratonic and pericratonic basins on these continents (Sloss 1979). Soares et al. (1978) recognized a similar chronology of stratigraphic sequences in the intracratonic basins of Brazil. The fact that these sequences were later documented in the cratonic and pericratonic regions of three continents suggested that they are, in fact, global in extent, and this conclusion later received strong support from the Exxon work on subsurface stratigraphy using seismic methods (Vail et al. 1977).

How are these sequences defined in the rock record? Sloss (1963) defined them as "rock stratigraphic units of higher rank than group, megagroup or supergroup, traceable over major areas of a continent and bounded by unconformities of interregional scope." Their recognition can only be achieved by studying subsurface stratigraphic successions in the centers of sedimentary basins, where the record can be assumed to be relatively complete. At the basin margins and on the flanks of positive elements, there are commonly many local unconformities, which may be of only local significance but which, nevertheless, obscure broader stratigraphic relationships (Fig. 1.2). Research carried out only on surface outcrops will, of necessity, be biased by this effect, particularly in the case of the older units that may only crop out at the basin margins. Sloss (1963) pointed out that there is no apparent relationship between the prominence of an unconformity and its geographic extent or regional importance. Spectacular angular unconformities may be the product of very localized syndepositional tectonism, the evidence of which disappears completely within a few kilometers (Riba 1976; Miall 1978). In contrast, regional unconformities may be the product of sea-level change unaccompanied by tectonic tilting. They are characterized by very low structural discordance and, in many cases, little erosional relief. Very careful facies studies or biostratigraphic zonation may be required to recognize this kind of unconformity (Runkel et al. 1998).

The six major unconformities that Sloss (1963) defined in North America are most readily recognized within the craton and become obscured within the mobile belts, on the margins of the continent. The development of each unconformity was followed by a major transgression and onlap, beginning on the craton margins and extending into the major intracratonic basins and, eventually, onto the margins of the Canadian Shield. Sloss noted that the transgressive phase of each sequence tends to be better preserved than the regressive phase at the end of the cycle, because the transgressive units were protected by successively overlapping sediments. By contrast, the top of each cycle was immediately exposed to erosion as the sequence was terminated by regression. The six sequences are shown diagrammatically in Fig. 8.14.

The names of the sequences were derived by Sloss et al. (1949) and Sloss (1963) from North American Indian tribal names. Figure 8.14 shows the approximate time ranges of the sequences as they vary from west to east across the continent. The diagram does not extend into either the Cordilleran or Appalachian regions, where stratigraphic relationships have been complicated by tectonics. Note that the Sauk and Zuni sequences are characterized by slow transgression and rapid regression, whereas the Absaroka and Tejas sequences, which were formed during times of marked cratonic mobility, show a rapid basal transgression and a long regressive phase.

The initial recognition of the six cratonic sequences was based on careful lithostratigraphic and biostratigraphic correlation. In later papers, Sloss (1972, 1979) took a different approach and showed that a similar sequence chronology could be recognized in Europe and Russia. For example, in his 1972 paper, Sloss reported on an analysis of detailed isopach and lithofacies maps of the Western Canada Sedimentary Basin and the Russian Platform. The data source included 29 Canadian maps (McCrossan and Glaister 1964) and 62 Russian maps (Vinogradov and Nalivkin 1960; Vinogradov et al. 1961). Each map was divided into a grid with intersection points spaced about 60 km apart, and the thickness and lithofacies were recorded for each point. From these data, the areal extent and volume of each of the mapped units could be calculated and compared. This approach is subject to possibly serious error because of the high probability of intersequence and even intrasequence erosion. The presence of a single isolated outlier or fault block beyond the edge of the main basin can change the interpreted former area of extent of a map unit by hundreds of square kilometers. Nevertheless, the data from the two areas show remarkable similarities, and the detailed statistical documentation confirms that Sloss' six sequences can be recognized in two widely separated continents that would formerly have been assumed to have undergone quite different geological histories.

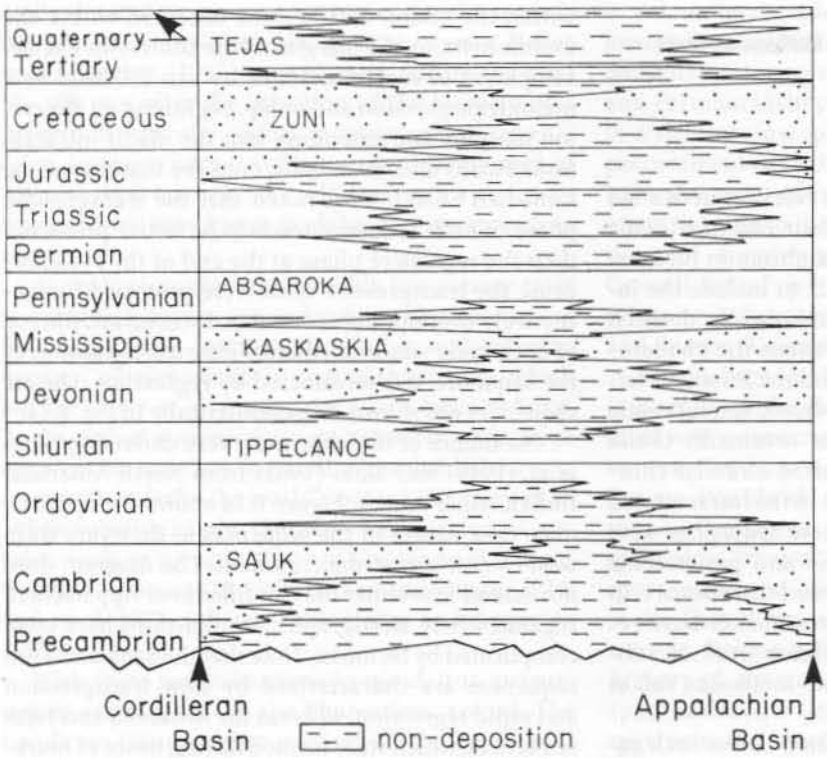


Fig. 8.14. The six sequences of Sloss (1963)

Intracratonic correlation of these long-term sequences is not always possible. Bond (1978) provided some of the first important insights into epeirogenic processes by demonstrating that the earth's continents have had different histories of uplift and subsidence since Cretaceous time. Stratigraphers specializing in the study of continental interiors (Sloss and Speed 1974) have, for a long time, appealed to a process that was termed epeirogeny by Gilbert (1890), which focuses on the vertical motions of the earth's crust. Modern studies of the thermal evolution of the mantle, supported by numerical-modeling experiments, have provided a mechanism that explains the long-term uplift, subsidence, and tilting of continental areas, especially large cratonic interiors beyond the reach of the flexural effects of plate-margin extension or loading (dynamic topography; Sect. 7.5). Much work remains to be done to test and apply these ideas by developing detailed numerical models of specific basinal stratigraphic histories, although it is already clear that dynamic topography is affected by both upwelling and downwelling currents on several scales.

It is now possible to explain observed epeirogenic effects using the concepts of dynamic topography, as discussed in Section 7.5. Russell and Gurnis (1994) discussed the broad tilting and basin development of the Australian interior. Burgess and Gurnis (1995) and Burgess et al. (1997) developed preliminary models for Sloss-type cratonic sequences invoking combinations of eustatic sea-level change and dynamic topography.

8.3.2.2 *Regional Tectonostratigraphic Cycles*

In many basins, tectonic influences are indicated by the presence of angular unconformities, faults that terminate at sequence boundaries, changes in isopach patterns that indicate changes in sediment-transport direction, changes in clastic sources, etc. (Embry 1990). Such structural features are commonly used as a means to subdivide the stratigraphy into **tectonostratigraphic sequences** spanning millions to tens of millions of years. They typically reflect major steps in the plate-tectonic evolution of the area, such as the transition from rifting to thermal subsidence on extensional continental margins. A comparison is offered here between the sequence stratigraphy of some extensional-continental-margin basins, using examples that have been studied by researchers independent of the Exxon group, who did not set out specifically to "test" the Vail global-cycle curves: Beaufort–Mackenzie Basin, Canada (Dixon and Dietrich 1990; McNeil et al. 1990), Beaufort Sea, Alaska (Hubbard 1988), and the Grand Banks, Newfoundland

(Tankard and Welsink 1987; Welsink and Tankard 1988; Hubbard 1988). In several of these syntheses, plate-kinematic events and tectonic subsidence phases are indicated. The results from these basins may be compared with the data from the Atlantic margin of the United States and the British Isles, which several workers have used as "tests" of the Vail curves (Poag and Schlee 1984; Poag and Ward 1987; Olsson 1991). A summary of the ages of the major sequence boundaries in these various sources is shown in Fig. 8.15. The disparity between these various columns is remarkable, especially considering that North Atlantic data were over-emphasized in the construction of the Exxon curves. Very few of the "supercycle" or "supercycle-set" boundaries in the Haq et al. (1987, 1988) charts are actually represented in the six basin columns shown.

It has long been known that foreland-basin strata are characterized by the intertonguing of marine and nonmarine strata, indicating episodic regression and transgression. For example, Weimer (1960) recognized four regional regressive-transgressive cycles in his classic work on the upper Cretaceous of the Western Interior of the United States, plus many minor events that he did not, at that time, attempt to correlate. Many nonmarine (molasse) successions consist of several separate wedges of coarse detritus that formed by episodic progradation into the basin (Miall 1978; Van Houten 1981; Blair and Bilodeau 1988). The timing of these wedges relative to tectonic episodes is a matter for detailed local study. They may or may not be syntectonic, as explained in Section 7.4.2.

Stockmal et al. (1992) suggested a relationship between the docking history of terranes on the west coast of Canada and the generation of clastic pulses in the foreland basin in Alberta (Fig. 8.16). The accretion process caused lithospheric delamination, which resulted in the terranes being emplaced onto the continental margin above a major detachment surface (Figs. 9.87, 9.88). This increased their flexural reach and led to the generation of multiple pulses of contractional thrusting and flexural subsidence in the foreland basin, some hundreds of kilometers inboard from the continental margin.

8.3.3
Cycles with Episodicities in the Million-Year Range

A wealth of stratigraphic data has accumulated for cycles having durations and episodicities of a few million years. They have been recorded in a wide variety of Phanerozoic basins in many different tectonic settings. They constitute the main basis of the Exxon global-cycle charts, where they are termed "third-order cycles" (Haq et al. 1987, 1988). A small selection of these is described in this section in order to illus-

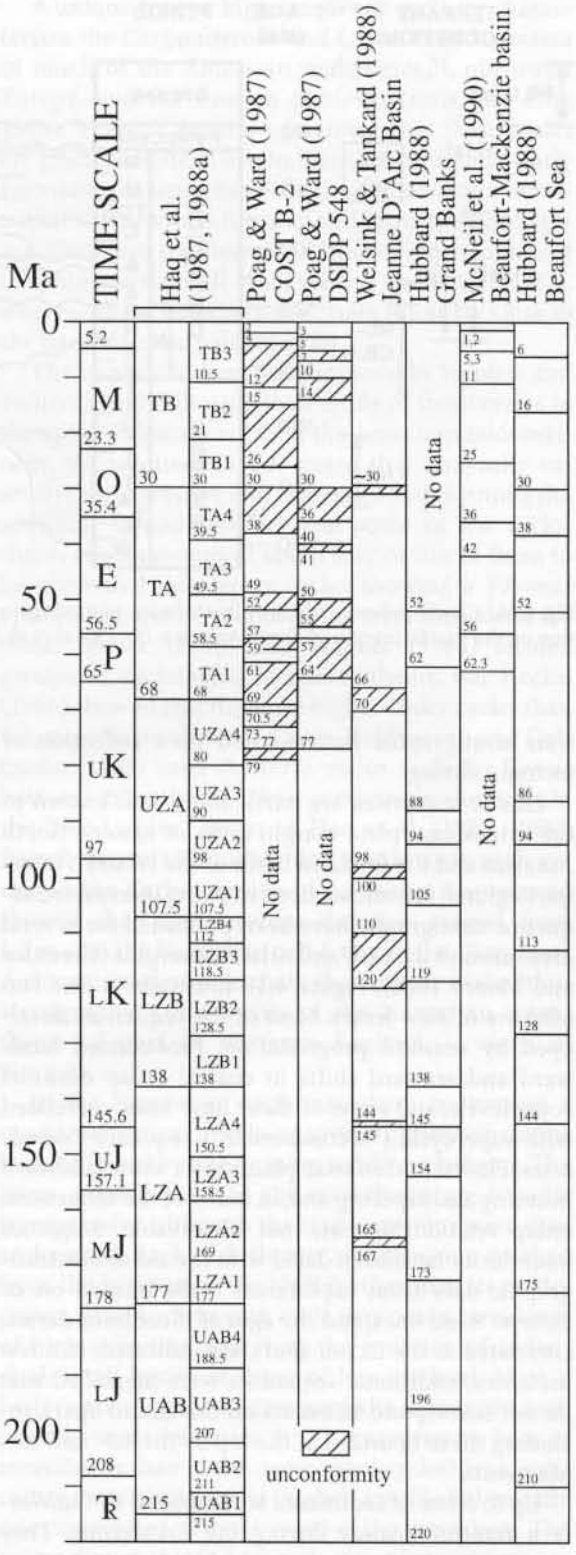


Fig. 8.15. Summary of the ages of sequence boundaries in various extensional-margin basins. Data derived from sources quoted in the text. (Miall 1997)

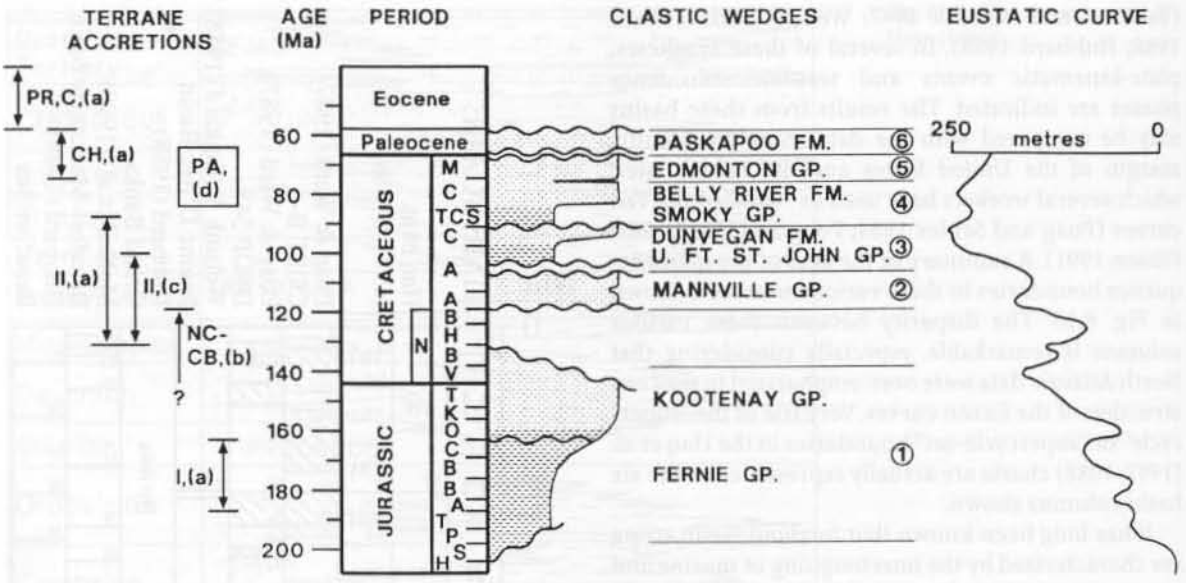


Fig. 8.16. Clastic pulses 1–6 in the Alberta basin, plotted against the global-cycle curve and the estimated times of terrane accretion on the Pacific margin of North America. (Stockmal et al. 1992)

trate stratigraphic patterns and their reflection of tectonic setting.

Clastic sequences are particularly well known in the extensional plate-margin areas of eastern North America and the foreland basin of the Western interior. Regional seismic sections showing interpreted sequence stratigraphy have been published for several areas around the Gulf and Atlantic margins (Greenlee and Moore 1988). Figure 8.17 illustrates a line run offshore of New Jersey. Most of the sequences developed by seaward progradation. Pronounced landward and seaward shifts in coastal onlap occur at some levels, and some of these have been correlated with supercycle ("second-order") sequence boundaries. Fluvial and coastal-plain strata show landward thinning and tapering and, in many of the sequences, onlap relationships are not observable. Sequence boundaries have been dated with the aid of biostratigraphic data from exploratory wells located on or close to these lines, and the ages of these boundaries, correlated to the Exxon chart, are indicated. In a few instances, additional sequences were identified that do not correspond to events on the Exxon chart, including those bounded at the top by the 4.7- and 5.1-Ma events.

Up to 5 km of sediments accumulated in the Western Interior Seaway during the Cretaceous. They constitute a classic "clastic wedge" (Figs. 8.18, 8.19), as this term was defined by Sloss (1962). Weimer (1960) was the first to recognize that the Upper Cretaceous section constitutes a succession of large-scale transgressive–regressive cycles with 10^6-year episodicities. Figure 8.20 illustrates Weimer's (1986) most recent synthesis of the stratigraphy and age of Cretaceous cycles in the Western Interior Seaway, including some of the key stratigraphic names from the Rocky Mountain and other basins. Major inter-regional unconformities and their ages in Ma are indicated on this diagram. They do not correlate in any particularly obvious way with the sequence boundaries in the Exxon charts, unless allowance is made for errors of up to 1–2 m.y., in which case they all correlate. The relationship between transgressive–regressive cycles and tectonism in the foreland basin was discussed by Fouch et al. (1983), Kauffman (1984), and Lawton (1994). Yoshida et al. (1996) described the Castlegate Sandstone of Utah, in which sequences that were generated in response to variations in the long-term rate of basin subsidence have higher-frequency sequences nested within them (Sect. 8.3.4.4).

Upper-Cambrian strata in various parts of North America are characterized by prominent large-scale cycles termed **grand cycles** by Aitken (1966, 1978), based on his work in the Rocky Mountains of Alberta. Similar cycles have subsequently been documented in the Great Basin, Nevada, the Northwest Territories, and southwestern Newfoundland (Chow and James 1987). The cycles consist of a lower "shaly" half cycle and an upper carbonate half cycle. In Newfoundland, the cycles are 100–200 m thick whereas, in the Rocky Mountains, they may exceed 700 m in thickness. Three superimposed cycles in Newfoundland are illustrated in Fig. 8.21. The shaly half cycles were thought, by Chow and James (1987), to represent de-

position on outer-platform, muddy tidal flats, whereas the upper carbonate half cycles originated as ooid sand-shoal complexes.

Each of the half cycles consists of a succession of meter-scale cycles. For example, the lower half cycle consists of interbedded units of ripple-laminated limestone, flaser to lenticularly bedded limestone–shale units, and grey shale. In the upper half cycle, there is a small-scale gradation between ooid calcarenite and carbonate laminite. These small-scale cycles are mostly of an autogenic shoaling-up type.

Chow and James (1987) interpreted the grand cycles as the products of variations in the rate of sea-level rise superimposed on an overall long-term rise. However, their sequence model has recently undergone a significant revision (Cowan and James 1993) based in part on a more widespread study of the regional stratigraphy, including correlation of the original grand-cycle profile with thinner successions deposited on the inboard area of the platform (Fig. 8.22). In the original Chow and James (1987) model, the lower, "shaly" half cycles were interpreted as representing relatively deep-water deposits formed during episodes of rapid rise, whereas the deposits of the upper half cycles were thought to have formed during periods of slow rise or stillstand, when sedimentation was able to fill the accommodation space. However, Cowan and James (1993) demonstrated that some of the components of the grand cycles are not depth-related facies and cannot, therefore, be used in the recognition and interpretation of grand cycles. The distribution of shale is thought to be related to variations in climate, not to transgression, and sand represents reworking of coastal eolian facies. The new sequence model for grand cycles assigns quite different interpretations to the various facies components. A facies termed "ribbon rock" has been recognized, consisting of thinly interbedded calcisiltite and either shale or dololutite. This facies is interpreted as forming during transgressive to highstand phases of the sea-level cycle. Much of the shelf was exposed during lowstand or underwent peritidal sedimentation, commonly under desiccated conditions. Oolitic facies commonly formed on the outboard platform during lowstand.

Variations in thickness and facies between the grand cycles in the different regions reflect variations in sedimentation and subsidence rates and variations in the width of the shelf. Estimates from biostratigraphic correlation of the grand cycles suggest that they each spanned 9–15 m.y. Therefore, they may be classified as short second-order cycles or long third-order cycles, using the original Vail et al. (1977) terminology (which demonstrates one of the inadequacies of a terminology based simply on duration). The cause of the sea-level changes is, at present, unclear.

A unique type of high-frequency cyclicity characterizes the Carboniferous and Lower Permian strata of much of the American midcontinent, northwest Europe, and the Russian platform (Ross and Ross 1988). There is general agreement that these cycles are glacioeustatic in origin. The Carboniferous–Early Permian corresponds to the time when major continental ice caps were forming and retreating throughout the great southern Gondwanan supercontinent (Caputo and Crowell 1985; Veevers and Powell 1987), whereas the areas where the cycles occur lay close to the late Paleozoic paleoequator.

The term cyclothem was proposed by Wanless and Weller (1932) following their study of these cycles in the upper Paleozoic rocks of the American midcontinent. Ramsbottom (1979) noted that unusually extensive transgressive and regressive beds forming the sequence boundaries between some of the cyclothems enable groups of about four or five of them to be combined into larger cycles showing a 10^6-year episodicity, and he proposed the term mesothem for these. Moore (1936) and Wagner (1964) termed groups of cyclothems megacyclothems, but Heckel (1986) showed that these are higher-order cycles than the mesothems discussed here. Holdsworth and Collinson (1988) used the term major cycle for Ramsbottom's mesothems. They compare in duration to the "third-order" cycles of Haq et al. (1987, 1988). Ramsbottom (1979) identified nearly forty such cycles in the Carboniferous of northwest Europe and showed that their average duration ranged from 1.1 m.y. in the Namurian to 3.6 m.y. in the Dinantian. A chronostratigraphic chart showing the main lithostratigraphic components of the Namurian mesothems of part of northern England, is given in Fig. 8.23.

In the Namurian, each mesothem consists of a muddy sequence containing several cyclothems at the base, followed by one or more sandy cyclothems. The lower, muddy parts of the mesothems are broadly transgressive, although the transgressions were slow and pulsed. Each cyclothemic transgression reached from the basin on to the shelf further than its predecessor. Basal beds in each cycle may contain evidence of high salinities, reflecting the isolation of individual, small basins at times of low sea level. Marine beds containing distinctive goniatite faunas are, supposedly, more extensive at the transgressive base of mesothems than in the component cyclothems, indicating more pronounced (higher-amplitude) eustatic rises associated with the onset of the mesothem. The regression at the end of each mesothem appears to have occurred rapidly. The sandy phase commonly commences with turbidites and is followed by thick deltaic sandstones (commonly called grits in the British Namurian; Fig. 8.23). The cycles may be capped

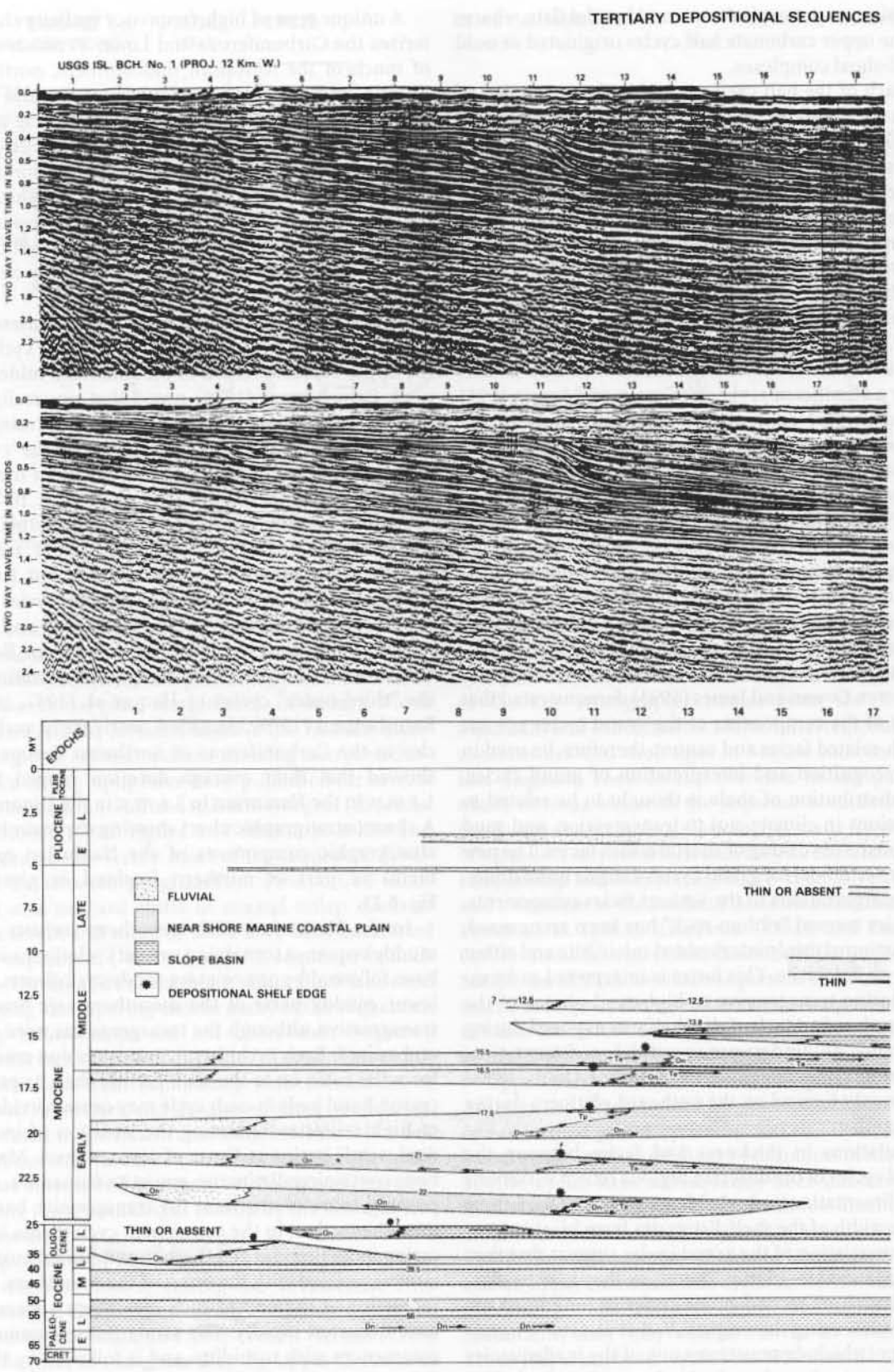


Fig. 8.17. Seismic line oriented northwest–southeast, perpendicular to the continental margin off New Jersey, showing uninterpreted line, interpreted line, and chronostratigraphic charts. (Greenlee and Moore 1988)

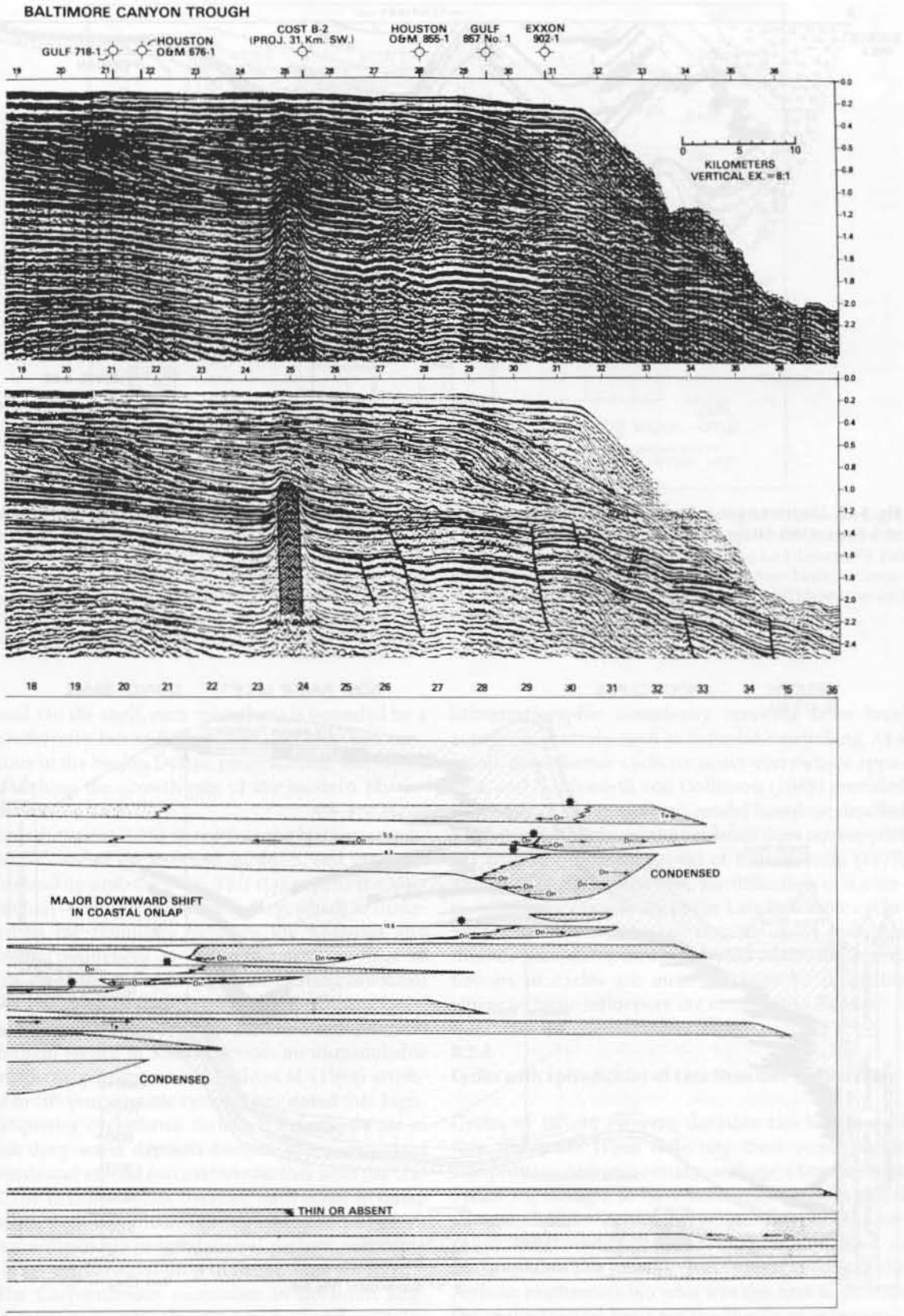


Fig. 8.17 (continued)

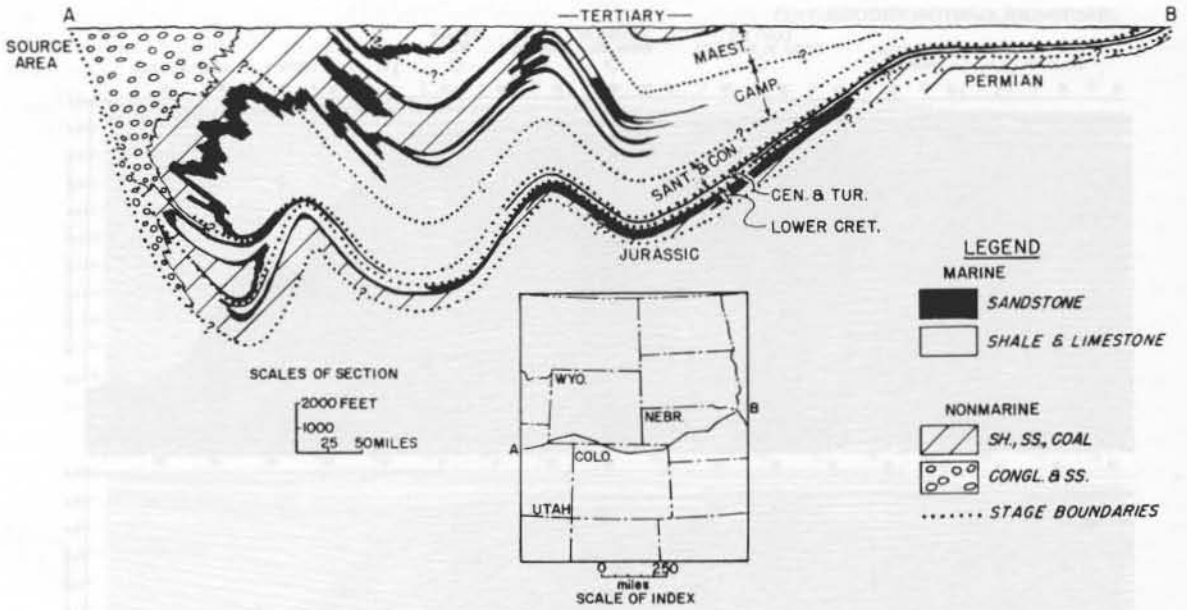


Fig. 8.18. Diagrammatic, restored cross section through the Upper Cretaceous rocks of the Western Interior Seaway, flattened with respect to a datum at the base of the Tertiary. (Weimer 1970)

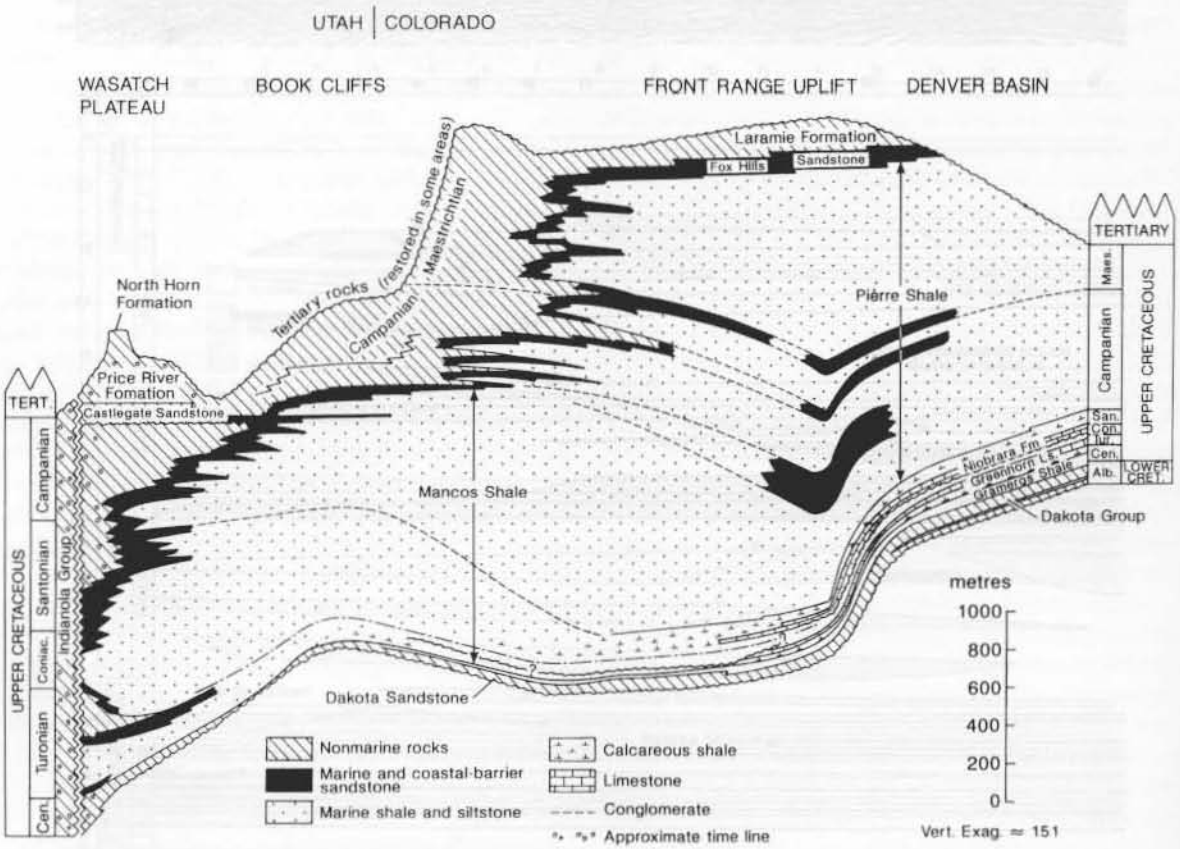


Fig. 8.19. Enlarged portion of Fig. 8.18, showing details of the stratigraphy of the upper Cretaceous clastic wedge of Utah and Colorado. (Molenaar and Rice 1988)

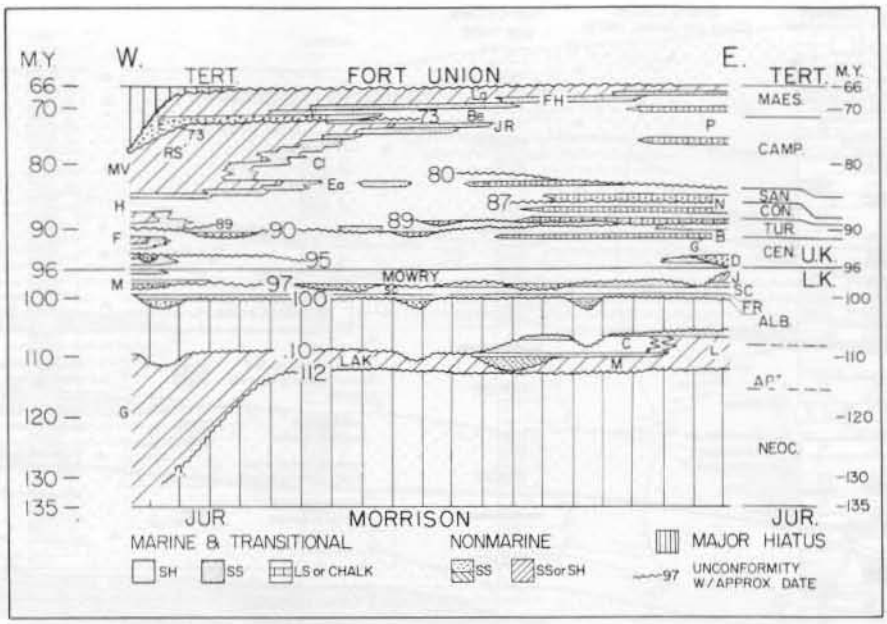


Fig. 8.20. Diagrammatic west–east cross section through the Western Interior Seaway of the Rocky Mountains, from Wyoming–Montana in the west to eastern Colorado–Black Hills–eastern Alberta in the east, showing stratigraphic positions and approximate dates of major transgressive units and inter-regional unconformities. Formations or groups to the west are: *G* Gannett; *SC* Skull Creek; *M* Mowry; *F* Frontier, *H* Hilliard, *MV* Mesaverde; *RS* Rock Springs; *E* Ericson; *Ea* Eagle; *Cl* Claggett; *JR* Judith River; *Be* Bearpaw; *FH* Fox Hills; *La* Lance. To the east, formations are: *L* Lytle; *LAK* Lakota; *FR* Fall River; *SC* Skull Creek, *J* and *D* sands of Denver basin; *G* Greenhorn; *B* Benton; *N* Niobrara; *P* Pierre; *M* and *C* McMurray and Clearwater of Canada. (Weimer 1986)

by coal. On the shelf, each mesothem is bounded by a disconformity, but sedimentation probably was continuous in the basins. Deltaic progradation was rapid, approaching the growth rate of the modern Mississippi delta.

Ramsbottom (1979) noted that the Namurian mesothems were of the shortest duration, and many do not extend up onto the shelf. This stage spans the Mississippian–Pennsylvanian boundary, which is designated as the boundary between the Kaskasia and Absaroka sequences in North America, a time at which sea level was at a long-term, eustatic lowstand (Sloss 1963).

A subsurface analysis of the middle Pennsylvanian cyclothem record in Kansas reveals an unmistakable "mesothem" pattern, which Youle et al. (1994) attributed to 10^6-year eustatic cycles. They noted that higher-frequency cyclothems form a transgressive set in which deep-water deposits become more important upwards and extend successively further onto the craton. The first few cycles overstep each other to onlap Mississippian basement. The highstand sequence set shows a distinctly progradational pattern, recording the long-term gradual drop in the average sea level.

The Carboniferous succession in northern England, the type area for the mesothem model, contains evidence for considerable local tectonism and for lithostratigraphic complexity resulting from local autogenic controls, such as delta-lobe switching. As a result, mesothemic cyclicity is not everywhere apparent, and Holdsworth and Collinson (1988) provided a critique of the mesothem model based on detailed local studies. In places, the evidence does not support the simple mesothem model of Ramsbottom (1979; also Leeder 1988). However, identification of a comparable form of cyclicity in the Late Paleozoic cyclothem succession of Kansas, as noted above, indicates that the mesothem concept should not be discarded. Groups of cycles are more likely to be mappable where tectonic influences are minor, as in Kansas.

8.3.4 Cycles with Episodicities of Less Than One Million Years

Cycles of 10^4- to 10^5-year duration can be grouped into five main types reflecting their stratigraphic composition, tectonic setting, and age. Many of these cycles are thought to have been generated by global climate changes driven by orbital forcing (the so-called Milankovitch mechanisms), as described in Section 8.2.4. The term derives from the name of the Serbian mathematician who was the first to provide the mathematical basis for the theory of astronomical forcing (Milankovitch 1930). However, there is in-

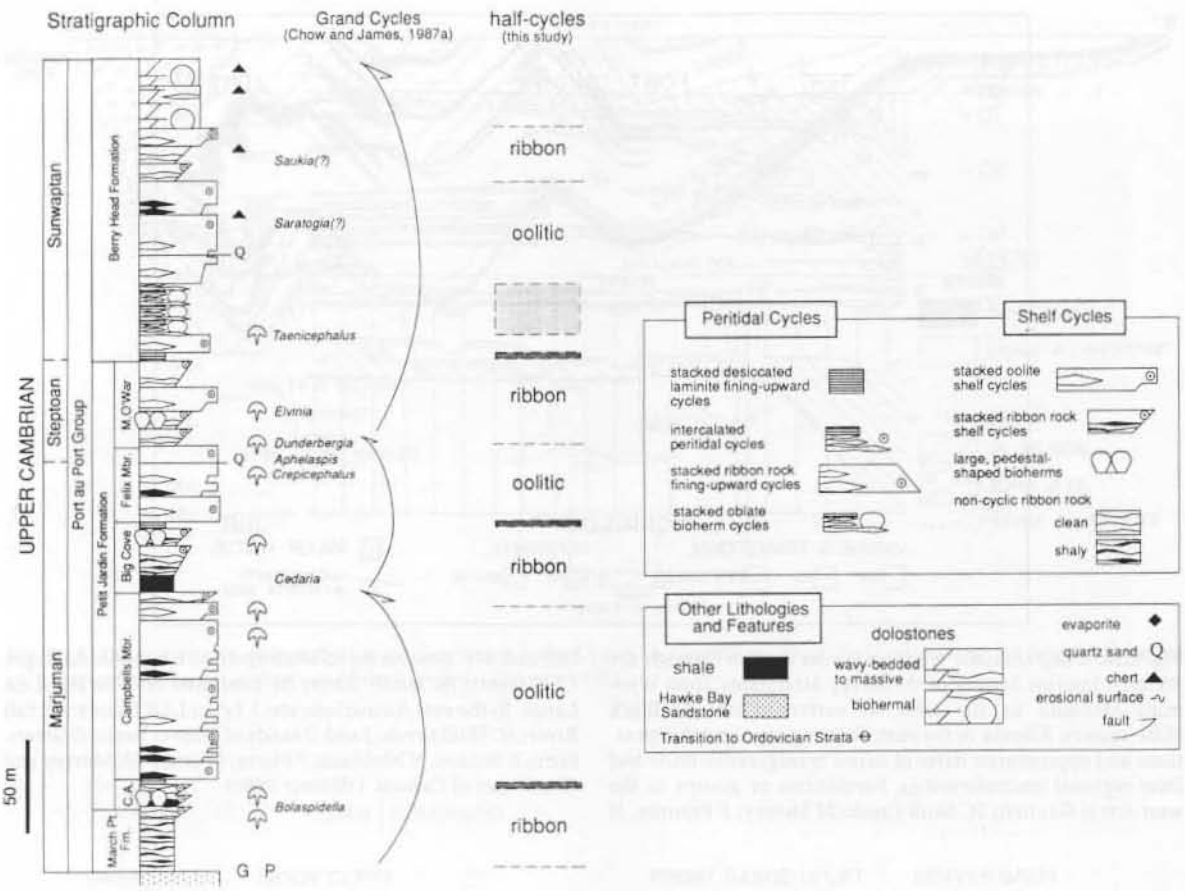


Fig. 8.21. Grand cycles of middle–late Cambrian age, southwest Newfoundland, as originally defined by Chow and James (1987), and showing the new facies subdivision of Cowan and James (1993)

creasing evidence of the tectonic origin of some of the clastic cycles in foreland basins.

It has been common practice to subdivide high-frequency sequences into those of fourth order, with durations in the 10^5-year range, and those of fifth-order, with durations of 10^4 years. This two-fold subdivision has been based largely on interpretations that invoke low- and high-frequency Milankovitch mechanisms. However, this classification should now be abandoned. There is increasing evidence for other sequence-generating mechanisms that do not readily fall into simple temporal classifications, and it has now been demonstrated that sequence thicknesses and durations have log-normal distributions that lack significant modes (Drummond and Wilkinson 1996). As demonstrated by Schwarzacher (1993), orbital forcing includes a periodicity of 2.035 m.y. (related to orbital eccentricity), but few examples of this have been described in the literature. Most examples of orbital periodicities demonstrated from the rock record fall between the 413×10^3-year eccentricity period and the 19×10^3-year precession period.

This book does not deal with cycles in the so-called solar band or calendar band of geological time. The solar band refers to cyclicity in the 10^1- to 10^2-year range, including the sun-spot cycle and its possible geological effects (e.g., el Niño current changes). The calendar band refers to cyclicity relating to Earth's seasonal rhythms (freeze-thaw, spring runoff, varves, etc.) and the tidal and other effects driven by the moon (Fischer and Bottjer 1991).

The five broad classes of cycle or sequence having episodicities of less than 1 million years are:

1. Neogene clastic cycles of continental margins generated by glacioeustasy
2. Pre-Neogene carbonate and clastic cycles that probably were generated by climatic forcing, of which eustasy may be a component
3. Late Paleozoic cyclothems of glacioeustatic origin
4. Lacustrine clastic and chemical rhythms driven by climatic forcing
5. Clastic cycles of foreland basins driven by high-frequency tectonic processes

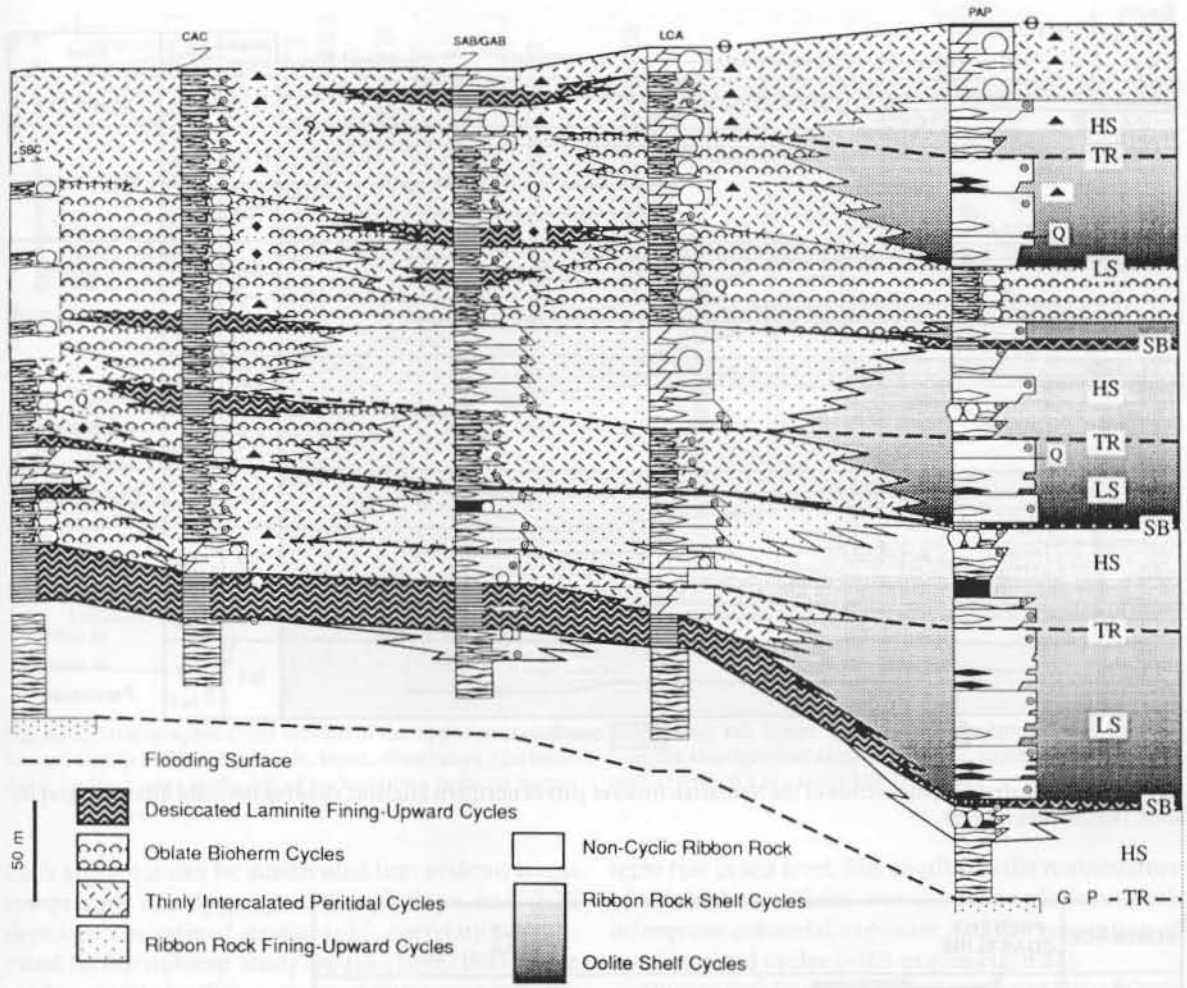


Fig. 8.22. Correlation of upper Cambrian platform rocks from inboard (*left*) to outboard (*right*) settings. The original grand-cycle study of Chow and James (1987) was based on the Port au Port Peninsula section (*PAP*) at *right*. *SB* Sequence boundary; *LS* lowstand systems tract; *TR* transgression; *HS* highstand systems tract. Compare the sequence classification of the *PAP* section with the grand cycles of Fig. 8.20. (Cowan and James 1993)

Types 1, 2, 3, and 5 are briefly documented and described here. Lacustrine sequence stratigraphy is summarized in Section 6.5.1.2. Additional details are given in Miall (1997, Chap. 8).

8.3.4.1
Neogene Clastic Cycles of Continental Margins

The Mississippi delta complex is a good example of the first type of high-frequency sequence (Figs. 6.29, 8.24). Six sequences have developed on the Louisiana shelf during the last 700×10^3 years (Fig. 8.24). Dating of the sequences is difficult in the absence of datable samples. Sequence 1 is interpreted to represent the lowstand associated with the Illinoian glacial stage. Sequences 5 and 6 correspond to the post-Wisconsinan sea-level rise since 18 ka. The Mississippi delta complex developed during this final stage of sea-level rise (Fig. 8.25). The first three sequences (Fig. 8.24; lobes 1–3 in Fig. 6.29) are interpreted as transgressive systems tracts. They backstep onto the shelf, because sediment supply to the delta was unable to keep pace with rising sea level. The culmination of the sea-level rise occurred at about 3–4 ka and resulted in the retreat of the coastline to the mouth of the Mississippi alluvial valley. The St. Bernard lobe represents the first of four highstand systems tracts that have formed since this sea-level highstand. Delta-lobe switching between the various positions has taken place because of autogenic processes related to the flattening of the river slope as the lobe prograded seaward.

The Wanganui Basin of North Island, New Zealand contains a particularly complete Pliocene–Pleistocene sedimentary record. Coastal cliff sections have

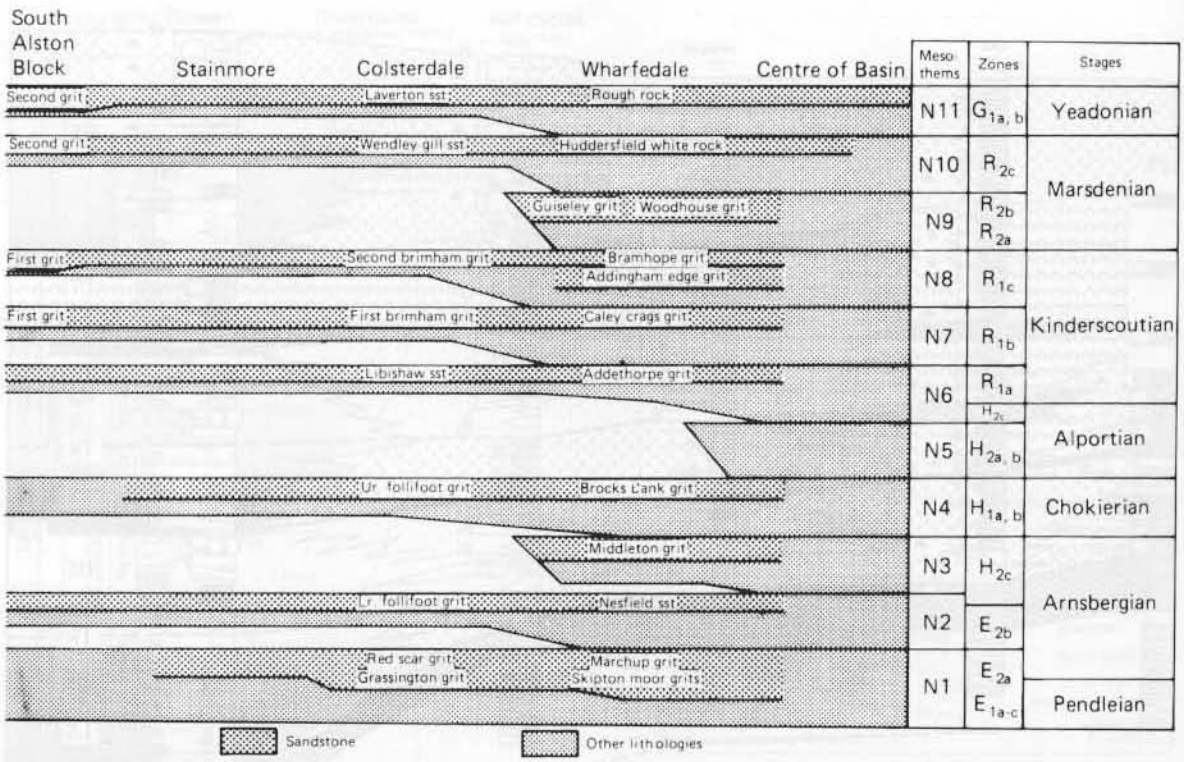


Fig. 8.23. Chronostratigraphic section of the Namurian rocks of part of northern England, showing the main lithostratigraphic units. (Ramsbottom 1979)

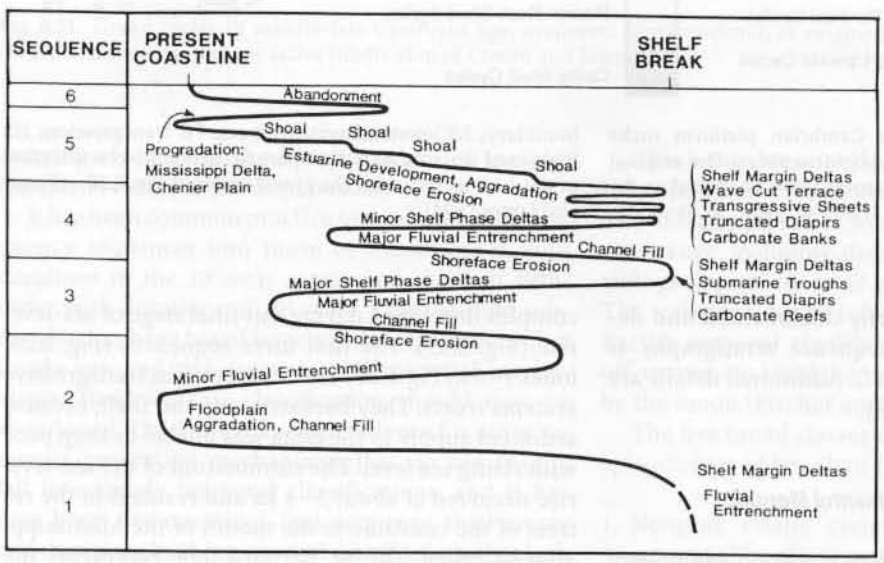


Fig. 8.24. Late Quaternary depositional sequences, sea-level changes since about 0.7 Ma, and associated sedimentary events on the Louisiana shelf. (Suter et al. 1987)

been interpreted in terms of 10^5-year cyclicity by Kamp and Turner (1990), Carter et al. (1991), and Naish and Kamp (1997), while correlative cycles also are present in offshore sediments, as revealed by seismic surveys (Carter et al. 1991). Neogene high-frequency cycles showing 10^4-year cyclicity have also been well documented in forearc and backarc settings on the continental margins of Japan (Ito 1992, 1995; Ito and O'Hara 1994; Masuda 1994). This succession consists of six sequences spanning the 0.4-Ma to 0.8-Ma period and estimated to have durations ranging from 45,000 years to 50,000 years (Fig. 8.25).

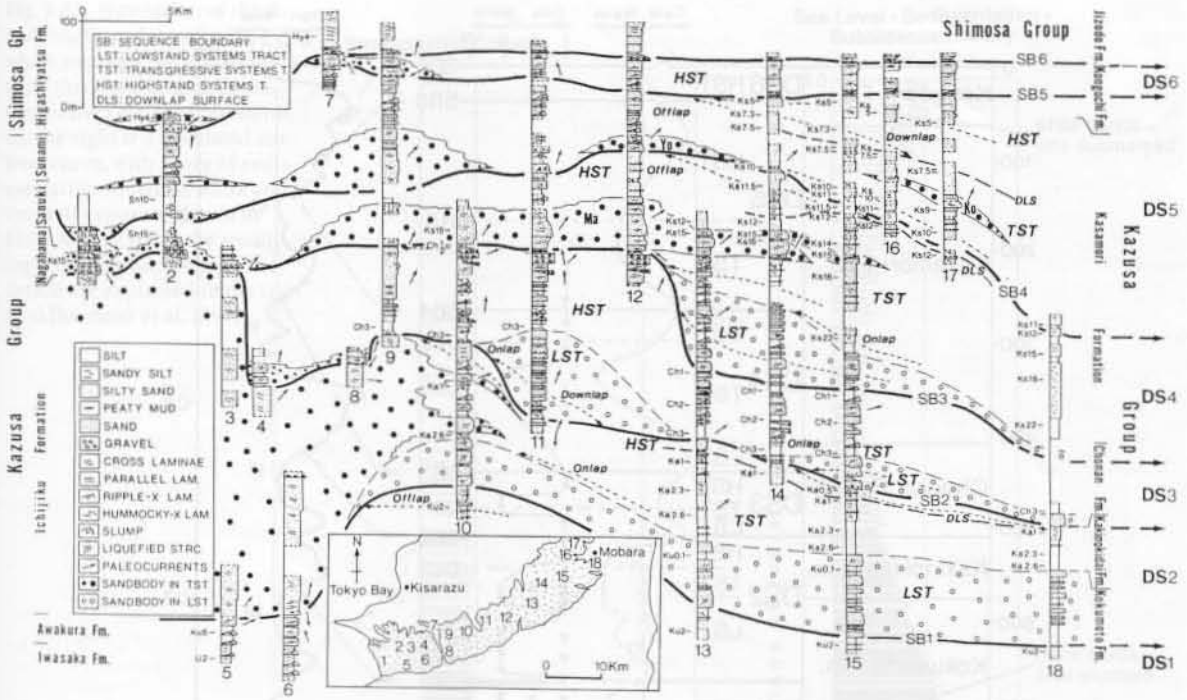


Fig. 8.25. Stratigraphic cross section of the upper part of the Kazusa Group in Boso Peninsula, Japan, illustrating glacioeustatic cycles. Codes to the *left* of each column indicate names of volcanic ash layers used for chronostratigraphic correlation. Six stratigraphic sequences are recognized, spanning the period from 0.4 Ma to 0.8 Ma. (Ito 1992)

Each sequence can be subdivided into systems tracts comprising mainly progradational slope and shelf deposits. The refined stratigraphic correlations indicated in the outcrop study by Ito (1992, 1995) were made possible by the presence of numerous volcanic ash beds, which can be correlated on the basis of their geochemical signatures and ages determined using fission tracks. A similar high precision has been obtained for the New Zealand sections. Ito (1992) showed how seawater temperatures fluctuated during sedimentation; he correlated these data with the oxygen-isotope record (Fig. 8.26).

8.3.4.2 *Pre-Neogene Carbonate-Dominated Cycles*

High-frequency sequences have been recorded in a wide variety of depositional and tectonic settings in the pre-Neogene stratigraphic record. Those of the Triassic in the Italian Dolomites are among the best known and are described in Section 6.6.3.

These cycles are interpreted as a result of the overprint of slow (10^6-year) eustasy on higher frequency cycles of sea-level change (Goldhammer et al. 1987, 1990, 1993; Goldhammer and Harris 1989; Hinnov and Goldhammer 1991; Jones and Desrochers 1992). During deposition of the Lower Platform Facies (Fig. 6.56), the area was undergoing a prolonged long-term rise in sea level. The result was the maintenance of subtidal conditions over the entire platform, with infrequent subaerial exposure and the generation of amalgamated cycles (~0.5 m.y. in Fig. 8.27).

During development of the Lower and Upper Cycle Facies (Fig. 6.56), the rate of long-term (10^6-year or "third-order") sea-level change was slow (~0 and 2.3 m.y., respectively, in Fig. 8.27; these were rhythmic and condensed megacycles). The subtidal sedimentation was able to keep pace with the net sea-level change, which was almost entirely due to subsidence. Every high-frequency (10^4-year, "fifth-order") oscillation in sea level is, therefore, recorded by subaerial exposure. The Tepee Facies (Fig. 6.57) records sedimentation during the time of falling sea level on the long-term (10^6-year) cycle (~1.5 m.y. in Fig. 8.27).

8.3.4.3 *Late-Paleozoic Cyclothems*

The first cycles to be described in the geologic literature were the Mississippian Yoredale cycles of the English Pennines area (Wilson 1975). They represent only part of a lengthy and widespread coal-bearing cyclic succession that spans much of the Carboniferous and Permian and extends throughout northwest Europe (Ramsbottom 1979). The classic Pennsylvanian cyclothems of the American Midcontinent

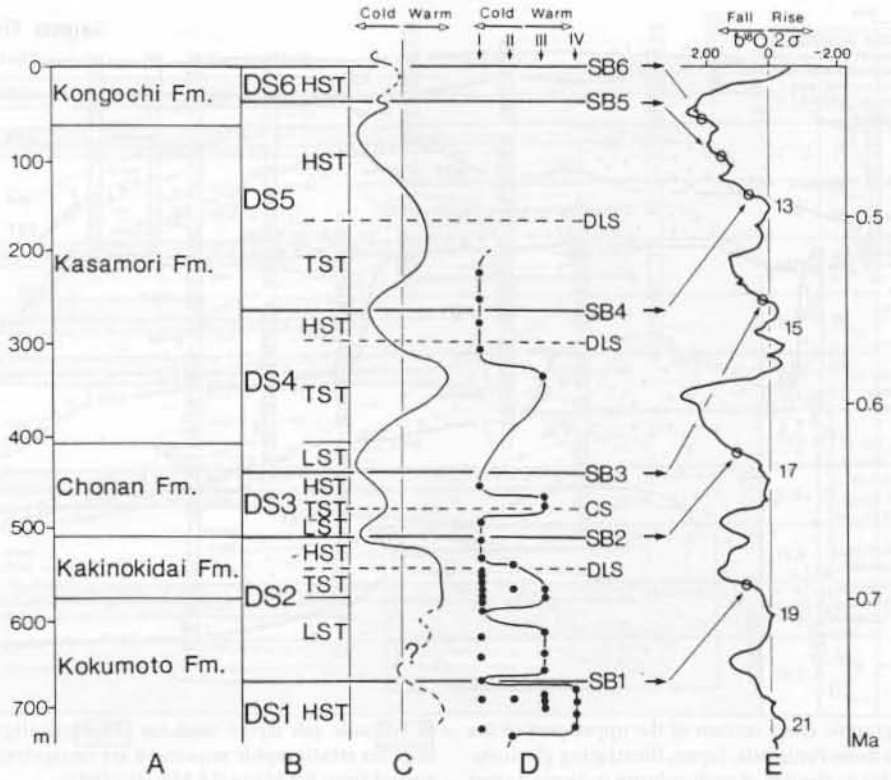


Fig. 8.26. Correlation of the Kazusa Group, Boso Peninsula, Japan (*columns A, B*), with the oxygen isotope record (*column E*). Also shown are temperature records derived from benthic molluscs (*column C*) and planktonic molluscs (*column D*). Compiled by Ito (1992)

were among the first cycles to be described in North America (Fig. 8.28). Wanless and Weller (1932) coined the term cyclothem for them. They have also been called **Klüpfel cycles**, after Klüpfel (1917). These cycles in the northern hemisphere are now interpreted primarily as the product of glacioeustasy (Crowell 1978); they developed in response to the major Carboniferous-Permian glaciation of the Gondwanan supercontinent. Klein and Willard (1989), Klein and Kupperman (1992), and Klein (1994) have also pointed out the importance of regional tectonism in the development of cyclothems, particularly with regard to the relative importance of a clastic component in the successions. Those developing closer to the Appalachian orogenic landmass, the so-called Appalachian-type cyclothem, are clastic dominated and commonly contain significant coals. The classic Illinois-type cyclothem is illustrated in Fig. 8.28.

8.3.4.4 *Clastic Cycles of Foreland Basins*

Shallow-marine to nonmarine clastic sequences with 10^4- to 10^6-year episodicities are common in the Cretaceous strata of the Western Interior of North America. They constitute a distinctive type of cycle, in part because they occur within a foreland basin, with its characteristic pattern of subsidence and sediment supply. It is not known whether the tectonic setting of these cycles is significant in terms of the controlling cyclic mechanisms. Foreland basins are, of course, tectonically highly active, and tectonic influences on sequence development seem to be indicated, as discussed at some length in Sections 8.2.3 and 8.2.4. However, controversy remains, and some researchers have referred to possible glacioeustatic causes (Plint 1991; Elder et al. 1994) despite the very limited evidence for glaciation during the Cretaceous.

Two examples of sequences in the Alberta Basin are briefly described here as examples of this type of sequence. The Dunvegan Formation represents a major delta complex up to 300 m thick, which prograded into the Alberta Basin from the northwest over a period of about 1.5 m.y. (Bhattacharya 1988, 1991; Bhattacharya and Walker 1991). It can be subdivided into seven allomembers, which are separated from each other by widespread flooding surfaces (Fig. 6.31). The allomembers have each been mapped over an area in the order of 30,000 km^2. Each allo-

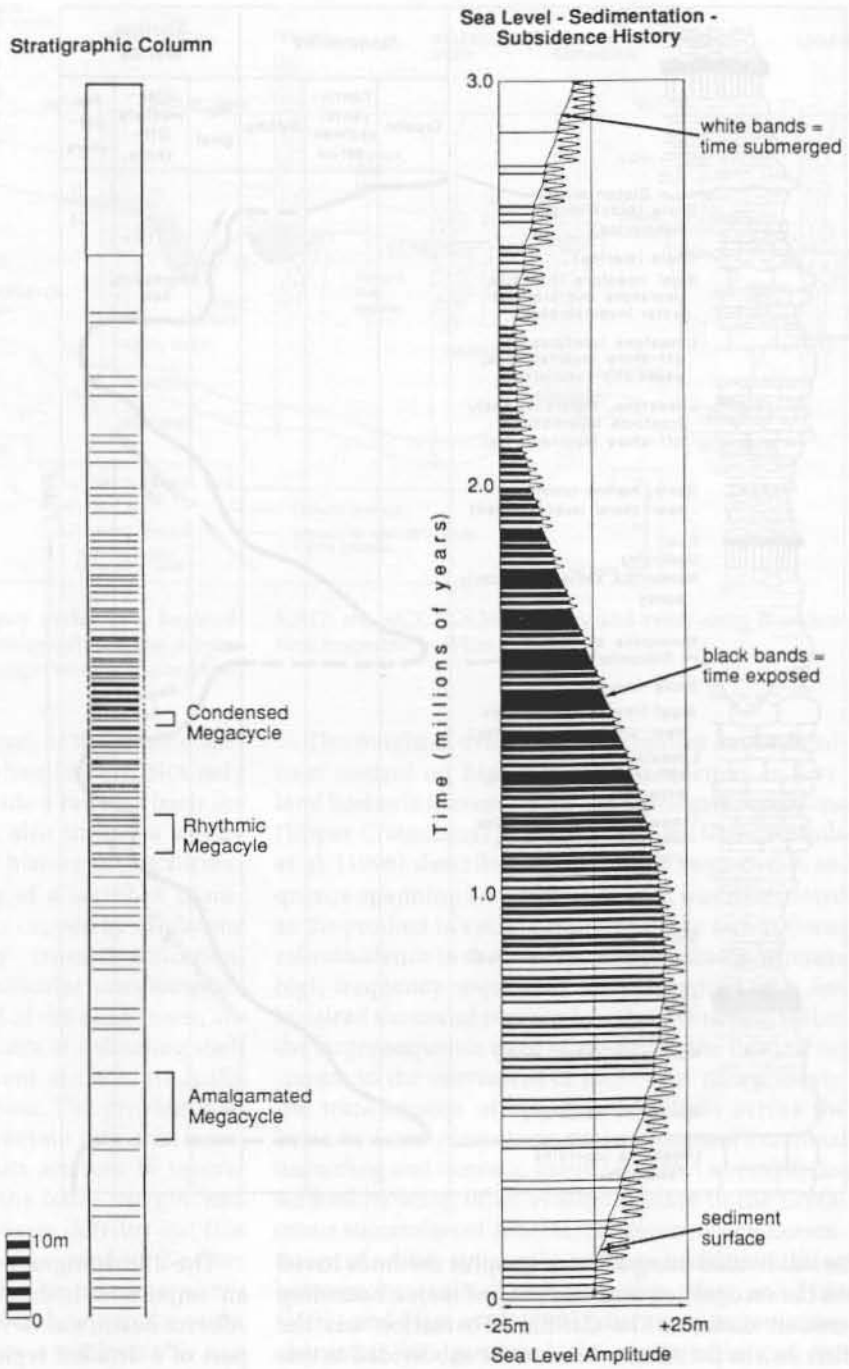


Fig. 8.27. Simulation of third- and fourth-order cyclicity, which explains the variations in cycle thickness and composition in the Latemar Limestone. On the right is a simulated sea-level curve, with times of sedimentation shown in *white* and times of exposure shown in *black*. On the *left* is the resulting stratigraphic section. The details are explained in the text. (Goldhammer et al. 1990)

member ranges up to 80 m in thickness and represents a depositional episode about 200×10^3 years in duration. The bounding marine-flooding surfaces are attributed to "allocyclically controlled relative rises in sea level, probably caused by a rate increase in tectonically induced basin subsidence" (Bhattacharya 1991). The allomembers consist of shingled clinoform deposits up to 30 m thick, each shingle comprising a heterolithic deltaic complex representing 10^4 years of sedimentation. The deltas are of a river-dominated type and compare in scale and composition (except for being somewhat more sandy) with the delta lobes of the modern Mississippi delta. Offsetting of the shingles within each allomember probably results from autogenic distributary switching, as in the modern Mississippi.

The Cardium and Viking formations of Alberta each consist primarily of shelf deposits, and both may

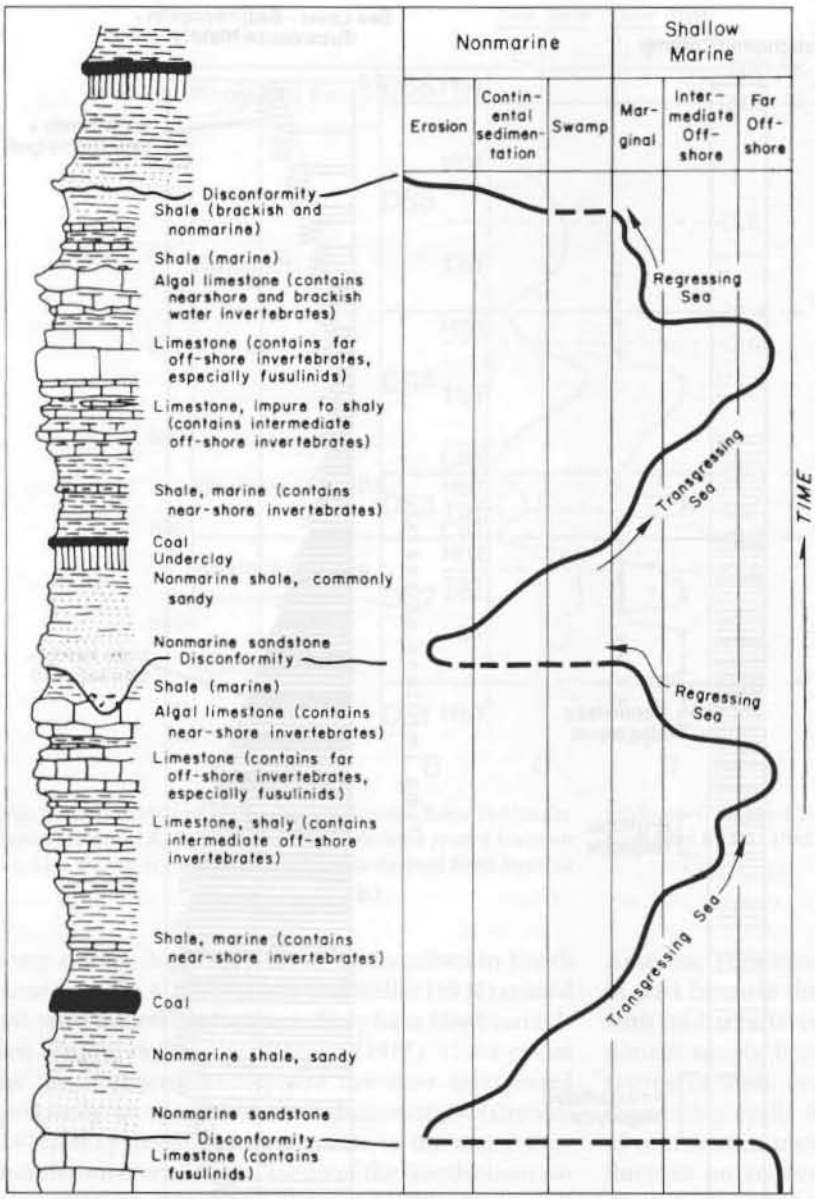


Fig. 8.28. Two typical Illinois-type Carboniferous cyclothems, showing the interpretation in terms of transgression and regression. (Crowell 1978; based on Moore 1964)

be subdivided using allostratigraphic methods based on the recognition and mapping of major, bounding erosion surfaces. The Cardium Formation was the first unit in the Alberta basin to be subdivided in this way (Plint et al. 1986), and recognition of the architectural style of the formation constituted a major breakthrough in foreland-basin geology when the 1986 paper was published. Indeed, the stratigraphic concepts were considered controversial, and some diverging opinions were published (Rine et al. 1987). Subsequent detailed papers on the Cardium Formation include those by Bergman and Walker (1987) and Walker and Eyles (1988). The allostratigraphy of the Viking Formation has been described by Boreen and Walker (1991).

The allostratigraphy of the Cardium Formation, an important hydrocarbon-producing unit in the Alberta Basin, was developed by Plint et al. (1986) as part of a detailed regional surface–subsurface study of the many producing fields in the area. Numerous local studies had, over the years, led to a confusing welter of local informal terminologies for sandstone horizons and marker units within the Cardium Formation and much controversy regarding the depositional environments of the various facies. Routine but meticulous lithostratigraphic correlation of subsurface records led Plint to the recognition that this unit, which is only about 100 m thick, contains at least seven basin-wide erosion surfaces, indicating the occurrence of this many events of erosion and trans-

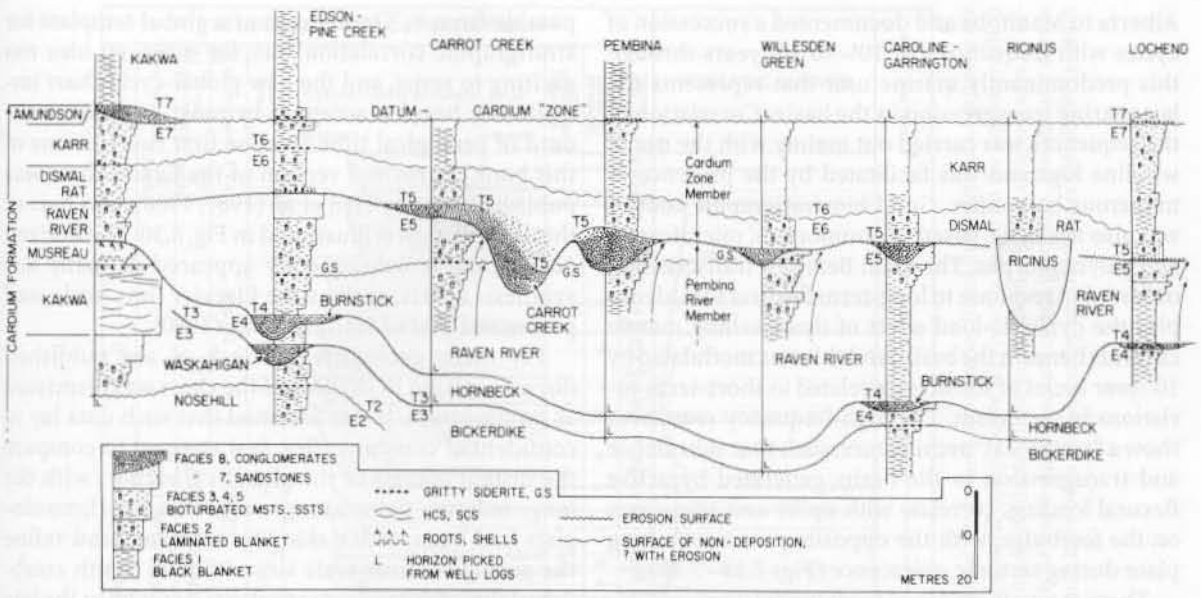


Fig. 8.29. Example of high-frequency cycles in a foreland-basin succession: the Cardium stratigraphy of the Alberta Basin. Surfaces of erosion and transgression are numbered *E2/T2*, etc. *HCS*, *SCS* Hummocky and swaly cross-stratification, respectively. (Plint et al. 1986)

gression (Fig. 8.29). Some, at least, of the erosion surfaces can be traced for more than 500 km. Not only does this new framework provide a rational basis for basin-wide correlation, but it also throws a wholly new light on the depositional history of the formation. The unit consists largely of a series of basin-wide coarsening-upward cycles capped by sandstone units containing hummocky cross-stratification, overlain in some areas by lenticular conglomerate beds. The cycles, up to the level of the sandstones, are readily interpreted as the products of a shoaling shelf environment, with the sediment surface gradually building up to a storm-wave base. The problem had always been to fit the conglomerate into this interpretation. These coarse deposits are tens to several hundreds of kilometers from the basin margin, and the problem of transporting coarse detritus out this far in a shelf setting had been discussed in the Cardium literature for many years. The new interpretation by Plint provides a simple resolution to the problem. The conglomerates are not a conformable cap to the coarsening-upward cycles, as had previously been thought; each conglomerate lens rests on one of the erosion surfaces. They are therefore, in all probability, beach deposits that originated as fluvial detritus transported basinward during regression of the shoreline and concentrated along temporary shorelines during the initial transgression at the beginning of a new cycle of relative sea-level rise. The seven Cardium sequences span about 1 m.y.; therefore, the average duration of each sea-level cycle is about 140×10^3 years.

The weight of evidence for tectonism as the dominant control on high-frequency sequences in foreland basins is increasing. In the Castlegate Sandstone (Upper Cretaceous) of the Book Cliffs, Utah, Yoshida et al. (1996) described two types of sequence. A sequence spanning an estimated 5 m.y. was interpreted as the product of cyclic variations in the rate of flexural subsidence in the Western Interior Basin, whereas high-frequency sequences, with durations of a few hundred thousand years or less that are nested within the larger sequence were attributed to the flexural response to the movement of individual thrust sheets, the transmission of the flexural effects across the basin by intra-plate stress, and subsequent erosional unroofing and isostatic uplift. Leggitt et al. (1990) described beveling of an erosion surface in the Cretaceous succession of Alberta, the geometry and orientation of which appears to indicate intrabasin tilting between intervals of sedimentation. Hart and Plint (1993) and Plint et al. (1993) discussed tectonic mechanisms that governed the construction of the Smoky Group on the Alberta–British Columbia border. The overall architecture of the Smoky Group was interpreted as a product of a varying rate of flexural subsidence and associated forebulge movement on a 10^6-year time scale (Plint et al. 1993). Bevels on certain erosion surfaces within the Cardium Formation were interpreted as the result of flexure over reactivated basement faults and resulting shoreline incision (Hart and Plint 1993). Catuneanu et al. (1997a, b) examined the sequence stratigraphy of the Bearpaw Formation across the Western Interior Basin from

Alberta to Manitoba and documented a succession of cycles with frequencies of 10^4- to 10^6 years through this predominantly marine unit that represents the last marine transgression in the basin. Correlation of the sequences was carried out mainly with the use of wireline logs and was facilitated by the presence of numerous bentonites. Good biostratigraphic control was also available, based on ammonites, microfossils, and palynomorphs. The main Bearpaw transgression occurred in response to long-term flexural subsidence plus the dynamic-load effect of downwelling mantle currents beneath the basin, and this was modulated by 10^5-year cycles of subsidence related to short-term variations in tectonism. The high-frequency sequences show a "reciprocal" architecture, such that subsidence and transgression in the basin, generated by active flexural loading, correlate with uplift and regression on the forebulge, with the opposite processes taking place during tectonic quiescence (Figs 7.28–7.30).

There is ample evidence from modern areas of active tectonism (e.g., New Zealand Alps, Alpine orogen of Europe, Himalayan peripheral basin, Banda Arc; Miall 1997, Sect. 11.3) that rates of tectonism and the episodicity of uplift and subsidence in modern foreland basins are of the right order of magnitude to accommodate the high-frequency cycles documented in Utah, Alberta, and elsewhere within the Western Interior of North America. It remains for future research to provide the tight chronostratigraphic correlation between tectonic episodes and the sequence stratigraphy that would confirm tectonism as the major depositional control.

8.4 Cycle Correlation and the Global Eustasy Model

Although the basic components of sequence stratigraphy were in place by the 1960s (Ross 1991), this radically new approach to the study of stratigraphy did not become popular for more than a decade. There are two main reasons for this. First, during the 1960s and 1970s, sedimentologists were preoccupied mainly by autogenic processes, the process-response model, and the implications of plate tectonics for large-scale basin architecture. Second, geologists lacked the right kind of data. It was not until the advent of high-quality seismic-reflection data in the 1970s and the development of the interpretive skills required to exploit these data that the value and importance of sequence concepts became widely appreciated. When Vail et al. (1977) published their ideas in American Association of Petroleum Geologists Memoir 26, the global eustasy model was an integral part of the new paradigm and, in fact, seemed to be inseparable from it. The concept of a global template for stratigraphic correlation was, for most, an idea too exciting to resist, and the new global-cycle chart immediately became accepted by most as a new standard of geological time (see the first two editions of this book). A revised version of the Exxon chart was published later by Haq et al. (1987, 1988), and part of this compilation is illustrated in Fig. 8.30. The derived eustatic curve subsequently appeared in many key synthesis charts, such as the Elsevier time-scale wall poster and that of Harland et al. (1990).

For many geologists, the lack of any published documentation in support of the chart was dismissed as unimportant. It was assumed that such data lay in confidential company files. Few stopped to compare the instant success of the global-cycle chart with the long, tedious, painstaking work that geochronologists had been undertaking to construct and refine the geological time scale since William Smith established the basic concepts of biostratigraphy in the late 18th century. If they did pause, it was, no doubt, to conclude that the new paradigm of global eustatic control had rendered all this earlier work largely irrelevant. Vail et al. (1977) had set out this paradigm clearly in their first publication, and this has remained their key concept:

> One of the greatest potential applications of the global cycle chart is its use as an instrument of geochronology. Global cycles are geochronologic units defined by a single criterion-the global change in the relative position of sea level through time. Determination of these cycles is dependent on a synthesis of data from many branches of geology. As seen on the Phanerozoic chart ..., the boundaries of the global cycles in several cases do not match the standard epoch and period boundaries, but several of the standard boundaries have been placed arbitrarily and remain controversial. Using global cycles with their natural and significant boundaries, an international system of geochronology can be developed on a rational basis. If geologists combine their efforts to prepare more accurate charts of regional cycles, and use them to improve the global chart, it can become a more accurate and meaningful standard for Phanerozoic time (Vail et al. 1977).

The fundamental problem with this paradigm is that the global-cycle chart is, itself, initially based on the existing global time scale, with all its imperfections (sequences are dated using fossils, radiometric dates, etc.); yet, when the data are inadequate or conflicts arise, biostratigraphic and other independent data bearing on sequence age are subordinated to the pre-

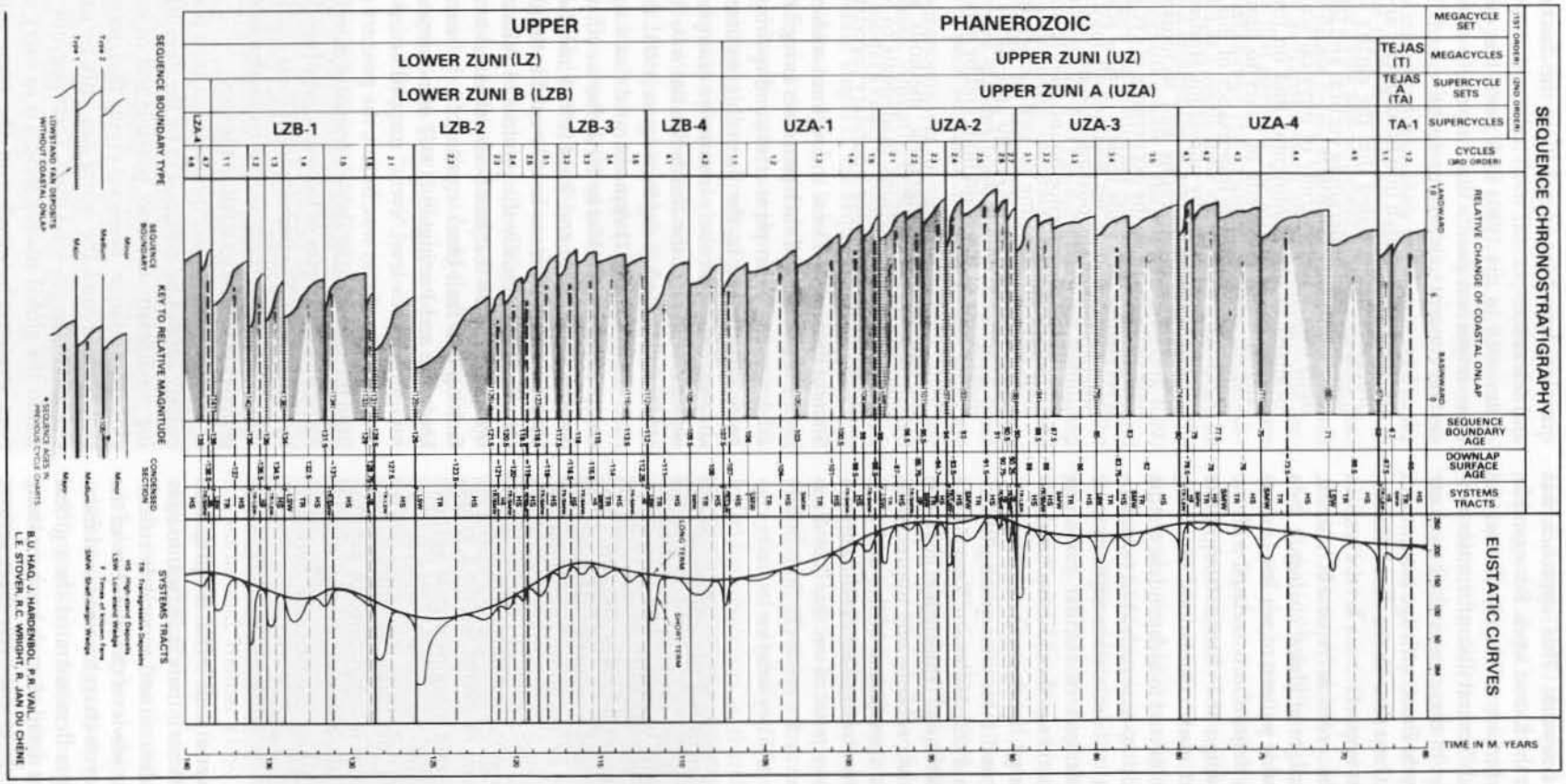


Fig. 8.30. Example of part of the revised global-cycle chart of Haq et al. (1987, 1988), showing the Cretaceous time interval. Cycles and supercycles listed in the *left hand columns* are part of the "global" sequence template that was first proposed by Vail and his coworkers in 1977. The "relative change of coastal onlap" curve is primarily based on the documentation of onlap and offlap architectures observed in seismic-reflection and well records, virtually none of which have been made available in the public domain. The assigned ages of sequence boundaries and downlap surfaces are shown without estimates of possible error, the omission of which gives a false impression of accuracy and precision. The eustatic curves at *right* are derived by making various assumptions about the response of coastal facies to sea-level change, some of which have been challenged by various writers (Miall 1991a; Hunt and Tucker 1992)

existing sequence framework. This approach has been used throughout the Exxon work. For example, with reference to the Jurassic of the North Sea, Vail and Todd (1981) stated: "several unconformities cannot be dated precisely; in these cases their ages are based on our global cycle chart, with age assignment made on the basis of a best fit with the data." The assumption is made that important sea-level events in any given stratigraphic section represent eustatic events. From the paradigm of global eustasy, it then follows that a comparable pattern of sea-level events in other sections is, by definition, correlated with the original event, even when the chronostratigraphic data may not support such correlations. How can such correlations be subjected to independent testing if the data upon on which to base such a test (e.g., biostratigraphic zonation) are declared in advance to be untrustworthy? The dangers of circular reasoning and self-fulfilling hypotheses should have been obvious to scientists at large, but this was not the case.

Few workers addressed these conceptual difficulties. Miall (1986, 1991a, 1992), Miller and Kent (1987), Christie-Blick et al. (1988), and Gradstein et al. (1988) pointed out some of the problems and imprecisions in chronostratigraphic correlation. Miller and Kent (1987), while arguing for the need for careful chronostratigraphic correlation, pointed out that "the durations of the third-order cycles are at the limit of biostratigraphic resolution." They went on to state:

> We agree that in order to test the validity of the third-order cycles it is not necessary to establish that *every* [their emphasis] third-order cycle is precisely the same age on different margins. Haq and others (1987) utilized a sequence approach to recognize third-order events above known datum levels. Assuming that they observed the same patterns on different margins, their observation of the same ordinal hierarchy of events within a given time window on different margins argues against a local cause and points to eustatic control. ... However, the simple matching of third-order cycles between locations is complicated by gaps in the records, uncertainties in establishing datum planes, and the ability to discriminate between these cycles at the outcrop level.

Miall (1992) asked the questions: "What if Vail and his colleagues are wrong? What if there is, in fact, no suite of global cycles dependent on and accurately reflecting a history of eustatic sea-level change?" Few of the "events" in the global-cycle chart have received independent testing that has demonstrated their global significance and, where detailed work has been done on specific intervals of the global-cycle chart, serious questions have been raised about the chart's validity and relevance.

Research in the 1980s and 1990s has established three important points that throw the entire global eustasy concept into serious doubt:

1. Geophysical research into lithospheric and mantle processes has revealed that the earth's surface is very dynamic, undergoing vertical and lateral motions over a wide variety of physical scales, from localized to intercontinental, and on time scales ranging from hundreds of years to hundreds of millions of years. The importance of this research has been confirmed by numerous studies of local and regional stratigraphy and tectonism, which have demonstrated the occurrence of cycles of sea-level change on a wide variety of time scales quite independent of any possible eustatic process.
2. The timing of sequence boundaries is not dependent only on the time of eustatic sea-level lowstands but may be markedly affected by tectonic movements of the basin floor and by variations in sediment supply.
3. Current chronostratigraphic methods are, with some specific exceptions, inadequate to provide a global test of sequence-boundary correlation.

Some of the wealth of local and regional studies bearing on the importance of such processes as regional flexural downwarping, in-plane stress, and epeirogeny have been summarized in the preceding sections of this chapter. Two particularly instructive examples are the regional studies of the North Sea Basin by Underhill (1991) and Underhill and Partington (1993). In the first case, Underhill (1991) demonstrated that a significant onlap event in the Moray Firth Basin, off northeast Scotland, which Vail and Todd (1981) had used as the basis for one of their sea-level "events" in the global-cycle chart, was actually the product of submarine-fan onlap in an area that had been subjected to repeated faulting and fault-block rotation. In the second case, Underhill and Partington (1993) demonstrated that a mid-Jurassic sea-level "event" mapped across most of the North Sea Basin was, in fact, the product of short-lived thermal doming above a transient mantle plume. Many other examples of tectonic control are described by Miall (1997). The importance of these studies is that they demonstrate many alternative mechanisms for the generation of stratigraphic sequences.

If tectonic and other mechanisms are a reality, and eustasy cannot be assumed to be the dominant driving mechanism, the only possible test of eustasy is global correlation. Sea-level events that occur simultaneously around the world can only be caused by changes in the sea-level itself. Therefore, one possible test of the global eustasy model is to carry out rigorous stratigraphic studies of the same interval in

many parts of the world that had different, independent tectonic histories. Even this type of test is not without its problems. Miall (1997, Fig. 11.1 B) demonstrated that, where tectonic subsidence and eustatic sea-level change have comparable rates, a synchronous eustatic signal is unlikely to be preserved in the geological record. Local and regional variations in subsidence rates and sediment supply can also generate significant local variations in the timing of such processes as transgression and regression – the mechanisms that generate the stratigraphic signatures of sea-level change. For example, consider the Mississippi Delta as a small-scale example of what might be happening around the margins of a large basin. The active, prograding lobe of the modern delta (Fig. 6.29) constitutes a parasequence analogous to a highstand systems tract. By contrast, the St. Bernard lobe of the delta (labeled lobe 4 in Fig. 6.29) has been abandoned and is undergoing compactional subsidence and transgression. In sequence stratigraphic terms, a transgressive systems tract is now developing there across a type-2 sequence boundary. The lobes are contemporaneous, yet, in sequence terms, they are out of phase. Similarly, Leckie (1994) demonstrated how the modern coast of South Island, New Zealand, is in different stages of the sequence cycle, reflecting local variations in rates of subsidence and sediment supply. A further discussion of the nonuniqueness of stratigraphic architectures is provided in Section 8.5.

Even setting aside or allowing for such difficulties, we are not yet in a position to carry out the necessary, precise global tests. The construction of the global time scale is discussed at length in Section 3.7, where the difficulties are made apparent (see Miall 1997, Chap. 13, for a more detailed treatment of this topic). Figure 3.22 illustrates the imprecision in correlation between two nearby wells even where a rich palynoflora is available for biozonation. Figure 3.39 indicates the uncertainties in the dating of Cretaceous stage boundaries by comparing a range of recent time-scale compilations. The difficulties were expressed succinctly by Ricken:

> Without a precise time control the depositional mechanisms forming beds and sequences cannot be sufficiently understood ... timing has remained an elusive problem. Too many inaccuracies are involved in resolving stratigraphic durations, including a large range of error in radiometric age determinations, poor biostratigraphic as well as magnetostratigraphic resolution, and an incompleteness of sedimentary sections. As a result, time estimates are commonly imprecise, and the range of error is often larger than the actual time span considered ... (Ricken 1991)

Harland et al. (1990) indicated that the range of possible ages for stage boundaries in the Cretaceous varies by as little as ± 4 m.y. for the Albian–Cenomanian boundary (the difference between the likely minimum and maximum possible ages) and as much as ± 25 m.y. for the Aptian–Barremian boundary. They assigned an overall average 2% uncertainty to the calibration of the Phanerozoic scale (2 m.y. at 100 Ma). These error values relate to the best available global data calibrated by several independent means, yet they reveal a residual imprecision that would not permit the dating of any given stratigraphic event, even in one of the global stratotypes, with a precision better than 2 m.y. In their new Mesozoic time scale, Gradstein et al. (1995) provided revised estimates of Mesozoic stage boundaries, with uncertainties expressed in million years to two standard deviations. These increase from 0.5 m.y. for some of the Late Cretaceous stage boundaries, to 2.6 m.y. for the Jurassic–Cretaceous boundary, to errors increasing from 3 m.y. to 4.8 m.y. for stages extending back from the Late Jurassic to the Early Triassic. The so-called "third-order" sequences that form the main basis for the global-cycle chart average 1.5 m.y. in duration in the late Cenozoic, increasing to about 3 m.y. in the Jurassic (Dickinson 1993). The sequences are therefore, in most cases, of lesser duration than the range of error that must be assigned to the ages of their boundaries. Inter-regional correlation of individual sequences is, therefore, not possible, and the entire basis for the global-cycle chart must be called into question. Some workers have succeeded in developing highly refined time scales for individual basins or regions. For example, Kauffman et al. (1991) claimed a precision of 100×10^3 years for Cretaceous ages in parts of the Western Interior Basin of North America. However, until such precision is available for correlation into tectonically unrelated regions, such data will not provide the basis for a test of global eustasy.

Given these problems, Miall (1992, 1997) suggested that the Exxon global-cycle chart is too flawed to be fixed and should be abandoned, including the system of supposedly global sequences identified by codes, which appears on each version of the chart (e.g., UZA-4, etc., in Fig. 8.30). However, there are many who do not agree with this suggestion. Posamentier and Weimer (1993) agreed that chronostratigraphic error is a serious problem, yet they argued that the global-cycle chart should be retained for correlation in frontier basins, because it is most useful to stratigraphers when there is no age dating available. Their suggestion is that, where we know nothing about a basin, the global-cycle chart is a good place to start. In other words, when building a stratigraphic database, start by building backwards

from a predetermined answer using circular reasoning. The problem is left for the readers of this book and other future stratigraphic practitioners to resolve.

8.5 Architectural Indicators of Tectonism, Sediment Supply, and Sea-Level Change: The Role of Stratigraphic Simulation

As noted in the previous section, along any given basin margin, different systems tracts may be forming at the same time because of local variations in subsidence and sediment supply. Together, such variations may be enough to interfere with the generation of a clear signal of eustatic sea-level change. As discussed earlier in this chapter, tectonism may generate relative changes in sea level of a rate and magnitude similar to those of eustasy, making it difficult to identify the ultimate causes of the sea-level change. Tectonism and climate change may also cause major changes in sediment supply that affect sequence architecture independently of the other controls. In fact, separating the relative importance of the three major controls on sequence architecture – tectonism, eustasy, and sediment supply – is typically very difficult, because each of the controls, acting in different combinations, can generate stratigraphic results that may appear very similar. For geologists attempting to understand the origins of a basin fill as part of a program of petroleum exploration or as part of a study of sea-level change, this ambiguity is a serious problem.

The importance of the three variables has long been known (Curray 1964). Burton et al. (1987) demonstrated that we cannot obtain absolute values of any of these variables unless we already know the magnitude of two of them (it is basically a simple problem of algebra: known and unknown variables in equations). Information on sediment accumulation rates can normally be derived from stratigraphic data, but the other two variables are elusive, and Burton et al. (1987) concluded, after an extensive review of various research methods, that "although an accurate eustatic sea-level variation chart would be a boon to stratigraphers ... such a sea-level chart cannot be made."

An alternative that is being pursued with increasing vigor and sophistication is that of computer-based numerical modeling. The values for missing variables can be estimated and used as input into computer models in order to simulate stratigraphic architecture. The results are illustrated graphically. Ranges of values for the three variables may be used as input in order to develop families of possible solutions. These are then compared with the actual basin architectures. Numerical modeling of this type permits the application of sensitivity tests to gauge the relative importance of the variables. There are two main types of model: **forward models**, which simulate sets of processes and responses given predetermined input variables, and **inverse models**, which use the structure of a forward model to simulate a specific result, such as an observed basin architecture (Cross 1990). Such comparisons do not provide definitive results, but they do help to eliminate unlikely solutions.

The study of stratigraphic simulation is a rapidly growing field, and there is now an abundant literature on this subject. The first significant modeling study was that of Jervey (1988) and is of some importance, because it provided the theoretical foundation for the sequence-stratigraphic models of Posamentier et al. (1988) and Posamentier and Vail (1988). A good introduction to modern approaches and the problems involved is provided in various chapters of the book edited by Cross (1990). Individual studies that describe the development and output of particular computer programs have been provided by Read et al. (1986), Jordan and Flemings (1991), Kendall et al. (1991), Reynolds et al. (1991), Sinclair et al. (1991), Lawrence (1993), Steckler et al. (1993), and Flemings and Grotzinger (1996), among others.

Most computer models are two-dimensional, simulating cross sections through continental margins. They do this by the development of geometric shapes, in which the cross-sectional area is dependent on input sediment volumes and the rates of differential subsidence and sea-level change (forward modeling). The models are therefore crude geometric approximations of stratigraphic units.

The actual building of stratigraphic units in the various computer models is carried out in various ways. The physical processes of sedimentation are, of necessity, simplified because of their complexity, and some critical information may be provided as input rather than calculated as part of the simulation, for example, depositional slopes. Burton et al. (1987) added sediment from the sides of the model using an arbitrary triangular distribution to simulate the downslope decrease in sediment volumes. Different slopes were used for sand and mud to simulate the different transport efficiencies of these two main lithologies. Their first model only simulated marine sedimentation. The model of Reynolds et al. (1991) built stratigraphic units by adding a constant volume of sediment with each time step and distributing it according to predetermined slopes. The model is restricted to marine deposition, differentiating between the continental shelf and slope but not including nonmarine deposition. Compaction and

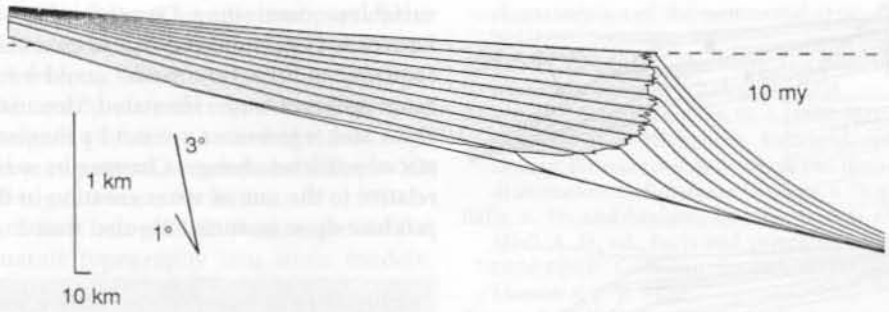


Fig. 8.31. The building of depositional surfaces in the Cornell model. Sediment enters from the side (*left*) and is moved across the surface by application of a diffusion equation. Two major depositional environments are modeled. The *dotted area* shows nonmarine deposition; the *unornamented area* is the site of marine deposition. (Jordan and Flemings 1991)

subsidence are added steps. The Cornell University model (Jordan and Flemings 1991) is somewhat more sophisticated than both the approaches described above. Figure 8.31 shows the building of a basic succession of depositional units on an extensional continental margin. Marine and nonmarine deposition are shown. Sediment is moved across the continental margin using a diffusion equation, and the depositional slope is not fixed but is calculated based on "the diffusional interactions of subsidence and sediment flux" for the two environments, using transport coefficients for marine and nonmarine deposition. Slope is, therefore, a product of the model, not an input parameter. This makes the model more realistic, although it still represents a crude generalization of reality. For example, it does not take into account along-shore sediment transport, and the transport coefficient is based on deltaic sedimentation, which, Jordan and Flemings (1991) argued, provided a useful simplification for clastic-rock-dominated coasts. Subsidence and compaction are also built into the model. Note, in this figure, the steepening through time of the marine–nonmarine facies contact, a process explained by Muto and Steel (1992). Jordan and Flemings (1991) modeled both extensional and ramp-type foreland-basin continental margins, which differ in terms of the relationship between sediment input points and subsidence styles. Figure 8.32 illustrates the evolution of their extensional-margin model from the starting point shown in Fig. 8.31 through one complete 10^6-year cycle of sea-level change. A chronostratigraphic (Wheeler) diagram of the results is given in Fig. 8.33.

One of the most useful outcomes of these modeling exercises is the insight it has provided about the timing of sequence boundaries. Jordan and Flemings (1991) stated that:

> Depending on the particular parameters used in the model, the time of reversal of onlap (i.e., the age of the sequence boundary) occurs between the times of maximum rate of fall and the eustatic lowstand ... For example, in the case of a narrower, more rapidly subsiding basin with less sediment supply ... the ages of the sequence boundaries much more closely approximate the times of maximum rate of fall of sea level ... factors that shift the age of the sequence boundary from the age of lowstand toward the time of maximum rate of fall of sea level are ... more rapid subsidence, lower sediment flux, more efficient nonmarine transport ..., longer periodicity of eustatic change, and smaller amplitude of eustatic change.

The phase lag between sea-level change and stratigraphic response has been noted before (Pitman and Golovchenko 1988), but the experiments of Jordan and Flemings provide what is probably the most thorough computer-model investigation of this phenomenon. Depending on the balance of the various input parameters, Jordan and Flemings (1991) demonstrated that "the sequence boundary for an identical sea level history could be of different ages and the ages could differ by as much as 1/4 cycle." Helland-Hansen et al. (1988) and Steckler et al. (1993) also commented on this point. This directly contradicts the Exxon models based on the work of Jervey (1988), who asserted that sediment supply affects shoreline position but not the timing of sequence boundaries. It is an important result, because the presumed synchronicity of sequence boundaries is the basis for the Exxon global-cycle chart. However, it has now been demonstrated that the stratigraphic record of a given 10^6-year ("third-order") sea-level cycle could vary by as much as 1/4 of a wavelength (that is, by up to several million years). Sequence boundaries are, thus, inherently imprecise recorders of sea-level change.

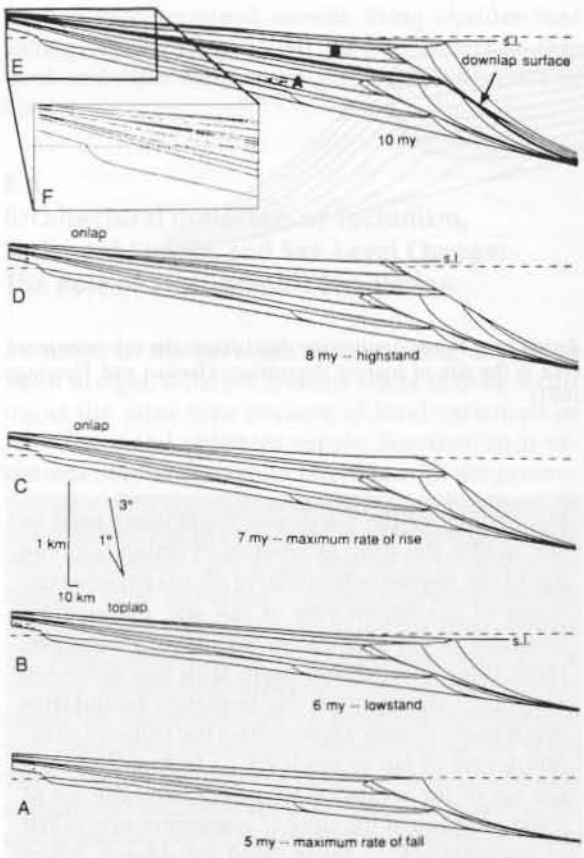


Fig. 8.32. The model of Fig. 8.31 extended through a complete 10^6-year ("third-order") cycle of fall and rise. *Dashed line at right* indicates sea level at the time of maximum rate of sea-level fall. Subsequent positions are indicated by the line marked *s.l.* (Jordan and Flemings 1991)

Flemings and Grotzinger (1996) demonstrated that clastic cycles generated by variations in sea-level change and those generated by variations in sediment flux are very similar, which confirms the difficulty set out by Burton et al. (1987) regarding the problem of resolving the relative importance of the three major variables controlling stratigraphic architecture. Lawrence (1993) showed that several elements of the sequence-architecture model could be generated in quite different ways. He stated "downlap, transgressions and regressions are not by themselves diagnostic of sea level change. Changes in sediment supply relative to the rate of space creation in the basin may produce these features." He also stated:

> ambiguities in interpreting relative sea level from two-dimensional lines may be reduced in many cases by incorporating three-dimensional data. For example, in three dimensions, erosion and marked downcutting in the nonmarine section may indicate lowering of relative sea level. Large changes in thickness along strike will point toward shifting depocenters (Lawrence 1993).

Jordan and Flemings (1991) also applied their modeling programs to foreland basins, where the basin configurations, the subsidence styles, and the sediment-input parameters are all quite different. Particularly interesting was their comparison of a foreland basin affected only by a cycle of eustatic sea-level change with constant subsidence versus a basin affected only by episodic thrusting with no sea-level change. Episodic thrusting generates cycles of progradation formed during thrust-induced subsidence phases, separated by proximal unconformities formed during episodes of tectonic quiescence. Dynamic models of foreland basins predict that thrust-induced subsidence in the proximal part of the basin occurs contemporaneously with uplift over the forebulge, and vice-versa (Sect. 7.4), and the graphical model of Jordan and Flemings (1991) confirms this, generating what Catuneanu et al. (1997a, b) termed "reciprocal stratigraphy" (Sect. 8.3.4.4). Contrasting with this architectural pattern is that of a basin affected by cycles of eustatic sea-level change and conti-

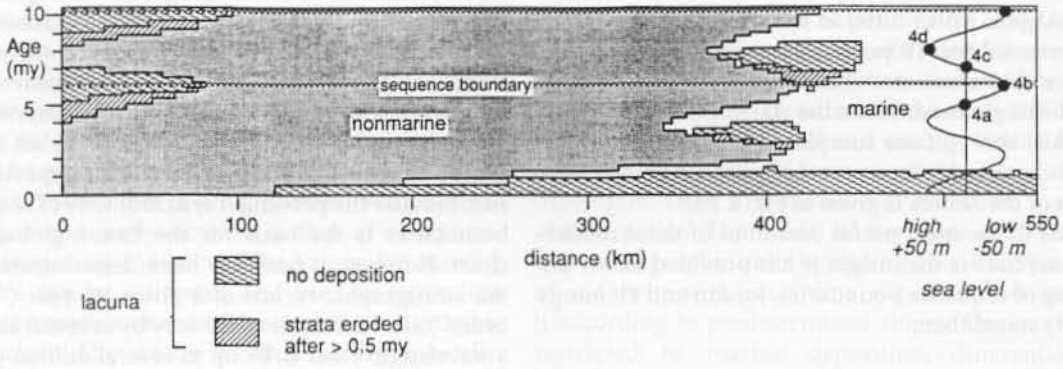


Fig. 8.33. Chronostratigraphic (Wheeler) diagram of the continental-margin architecture simulated in Fig. 8.32. The input sea-level curve is shown at *right*, *4a* to *4d* refer to the steps *A* to *D* in Fig. 8.32. (Jordan and Flemings 1991)

nuous tectonic subsidence. The proximal part of the basin is little affected by sea-level changes because of the rapidity of tectonic subsidence, and sedimentation tends to be continuous, whereas the stratigraphy over the forebulge is a sensitive record of eustatic cycles and preserves a pattern of onlap and offlap, with closely spaced angular unconformities.

Attempts are now being made to incorporate the effects of dynamic topography into some models. Burgess and Gurnis (1995) and Burgess et al. (1997) modeled Sloss-type sequences in which they combined cycles of eustatic sea-level change with broad cratonic tilts induced by mantle thermal processes. Johnson and Beaumont (1995) described a new three-dimensional model for foreland basins that incorporates variations in sediment supply and sediment transport.

8.6 Conclusions

Sequence stratigraphy has changed the face of sedimentary geology. Investigations of sequence architecture and its relationship to depositional processes have revitalized sedimentology, given added impetus to the development of geophysical, tectonic, and stratigraphic models of basin development, and provided a focus for the interdisciplinary work being carried out to improve the global time scale. This is very much a work in progress, as a comparison of this chapter with the versions in the first two editions of this book will reveal. Further research will undoubtedly contribute further to our knowledge of how earth systems work and is likely to constitute one of the most active areas of basin-analysis research over the next few years.

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Sedimention and Plate Tectonics

9.1 The Basin-Model Concept

Until the 1960s, sedimentary basins were explained and categorized in terms of **geosyncline** theory (Dott 1974, 1978; Mitchell and Reading 1978). Such classic books as those by Kay (1951), Krumbein and Sloss (1963), and Aubouin (1965) had a profound impact on geologists and formed the basis for all large-scale interpretations. However, we can now see that these and other studies, although meticulously descriptive, could not ultimately explain why or how most basins formed or why there were recurrent structural styles or lithofacies assemblages. Beginning with the development of plate tectonics in the 1960s, much has now become clear. The kinematics of modern plate movements have been documented in some detail and have provided geologists and geophysicists with a reliable data bank from which to build and constrain models of deep-lithospheric behavior. Beginning with the basin models of McKenzie, Beaumont, and others in the late 1970s (Chap. 7), our understanding of crustal and mantle processes has led to the development of the science of geodynamics, by which surface tectonic processes and events may be related to processes deep in the earth's interior.

Most sedimentary basins can now be explained in terms of plate-margin processes, and their structure and stratigraphy have become more comprehensible. Dewey and Bird (1970) were among the first to demonstrate the applicability of plate-tectonic concepts to basin development, and this began what became the complete abandonment of the geosyncline concept. Important papers by Dickinson (1974) and Bally and Snelson (1980) explored many of the details of the relationships among plate kinematics, basin subsidence, and sedimentation, and these papers are still worth reading. More recently, modeling of mantle thermal behavior by Mitrovica et al. (1989) and Gurnis (1988, 1990) has demonstrated the effects of deep-seated thermal convection on vertical motions of the crust and has provided a basis for the understanding of epeirogenic processes and plate-interior (intracratonic) basins (Sect. 7.5).

Much old terminology should now be entirely abandoned, a point recently emphasized by Ingersoll and Busby (1995). Such terms as *eugeosyncline* and *geanticline* – in fact all terms originating with the geosyncline classification of Kay (1951) – should be discarded, because the definitions of the terms cannot be correlated with modern geodynamic concepts. Other general terms, including *intermontane basin* and *successor basin*, are also so imprecise as to be of little value. Ingersoll and Busby (1995) also pointed out the confusions over the use of the terms *foredeep*, *foreland* and *hinterland*. Many workers have attempted to adapt old data and ideas that predate plate tectonics to the new concepts. The definition and interpretation of *flysch* (Lajoie 1970; Reading 1972) and *molasse* (Van Houten 1973; Miall 1981a) are good examples of such adaptation. They are well-known terms that describe readily recognizable sedimentary–tectonic associations. However, this is as far as the adaptation has gone, because modern plate-tectonic theory has shown that they represent a much wider variety of tectonic settings than formerly thought, and there is no agreement as to how the two terms should now be used (Miall 1984). They should also be abandoned, except, perhaps, in Swiss-type areas. However, I disagree with Ingersoll and Busby (1995) regarding the use of the term **terrane**. Although terrane concepts may have become confused in their application in certain parts of the world (Busby and Ingersoll both trained in and live in California, where the terrane approach has generated much controversy), and many of the principles of terrane analysis are not as original as claimed (Şengör and Dewey 1991), the study of terrane geology (**accretionary tectonics**) has made an enormous difference in our understanding of complex orogens, such as the Cordilleras and the Alpine–Himalayan belt and is, therefore, an integral part of books such as this.

Useful discussions of basin classification based on plate-tectonic principles were given by Bally (1984), Klein (1987), and Ingersoll (1988), but none of these incorporated the new ideas regarding mantle thermal processes and dynamic topography (Sect. 7.5) and are, therefore, incomplete. The most comprehensive recent treatment of the tectonics of sedimentary basins is the book by Busby and Ingersoll (1995). The introductory chapter by Ingersoll and Busby

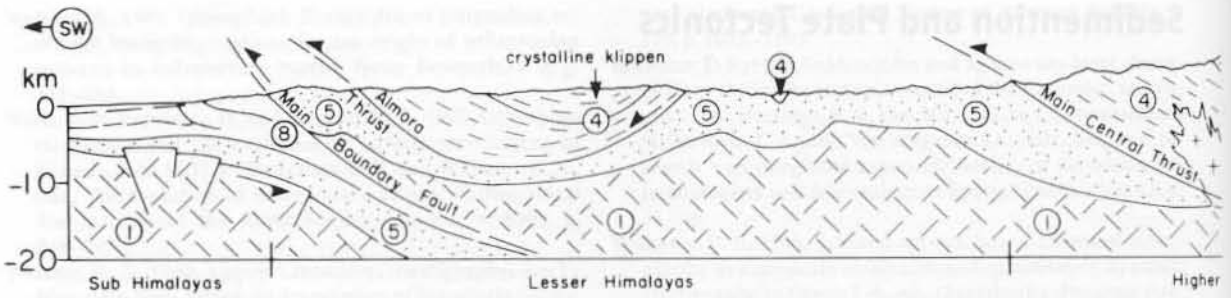


Fig. 9.1. Structural cross section across the Himalayas to illustrate juxtaposition in a suture zone of sediments formed in more than one plate setting: *1* Precambrian Indian Shield; *2* post-suture granites; *3* ophiolites; *4* Central Crystallines (Tethyan basement); *5* Tethyan Precambrian-Paleozoic sediments; *6* Tethyan upper-Precambrian–Paleozoic sediments; *7* Tethyan upper-Paleozoic–Mesozoic; *8* post-suture Tertiary. (Gansser 1964, reprinted with permission of John Wiley & Sons Ltd.)

(1995) is a concise summary of, and introduction to, the book.

Since its first edition, one of the themes of the present book has been the idea that, through the application of judicious simplification and skillful synthesis, we can systematize the descriptions of depositional systems (their facies assemblages and architecture), structural geology, petrology, and plate-tectonic setting into a series of **basin models** for the purpose of interpreting modern and ancient sedimentary basins. Dickinson (1980, 1981) used the term **petrotectonic assemblages** with the same meaning. These basin models are, then, a powerful tool for interpreting regional plate-tectonic history.

This process has been underway in a piecemeal manner for many years. For example, Mitchell and McKerrow (1975) and Graham et al. (1975) attempted to explain the complex Paleozoic stratigraphy of southern Scotland and the Appalachians–Ouachitas, respectively, with reference to modern plate interactions in Burma and India. Their references to modern analogues and the erection of local basin models are similar to the methods used in facies analysis. There are enough such local studies available that we can now develop a set of useful generalizations.

Many disparate skills and a wealth of data may now be brought to bear on problems of basin analysis and interpretation. Conventional stratigraphic and sedimentologic data are only the beginning. Reflection-seismic data may be extremely important. In fact, the advent of deep-crustal-reflection profiling helped to revolutionize basin analysis and geodynamic studies. The discoveries of the Consortium for Continental Reflection Profiling (COCORP) in the United States, of the LITHOPROBE Project in Canada, and of the European Geotraverse, British Institutions Reflection Profiling Syndicate (BIRPS) and Etude de la Croûte Continentale et Océanique par Réflexion et Réfraction Sismique (ECORS) in Europe are the most spectacular illustrations of this. Deep-seismic-reflection profiling revealed major thrust faults and extensional (detachment) faults extending to the base of the crust in many orogenic and taphrogenc belts and extensional plate margins. Hundreds of kilometers of overthrusting have been proved in orogenic belts, such as the Appalachians and the Himalayas, and it has been demonstrated that many terranes are entirely detached (delaminated) from their lower crustal roots (Oliver 1982; Allmendinger et al. 1987a, b; BIRPS and ECORS 1986; McCarthy and Thompson 1988; Blundell, et al. 1992; Cook et al. 1995).

Paleontological data also have much to contribute to basin studies. Changes in faunal provincialism may indicate relative plate movements, while vertical biofacies successions may yield much important data regarding changing water depths – vital information in the unraveling of subsidence histories (Chap. 7) and sea-level change (Chap. 8). Plutonism, volcanism, metamorphism, and metallogeny all reflect plate-tectonic settings (Hsü 1982), and integration of all these types of data may be essential for arriving at viable local interpretations and, ultimately, for developing satisfactory basin models (Hsü 1994; Harper 1998). The need to integrate a wide range of data is particularly important in areas of accretionary terranes, as discussed in Section 9.3.5.

9.2 Basin Classification

Most sedimentary basins can now be readily classified in terms of three criteria:

1. The type of crust on which the basin rests
2. The position of the basin relative to plate margins
3. Where the basin lies close to a plate margin, the type of plate interaction occurring during sedimentation

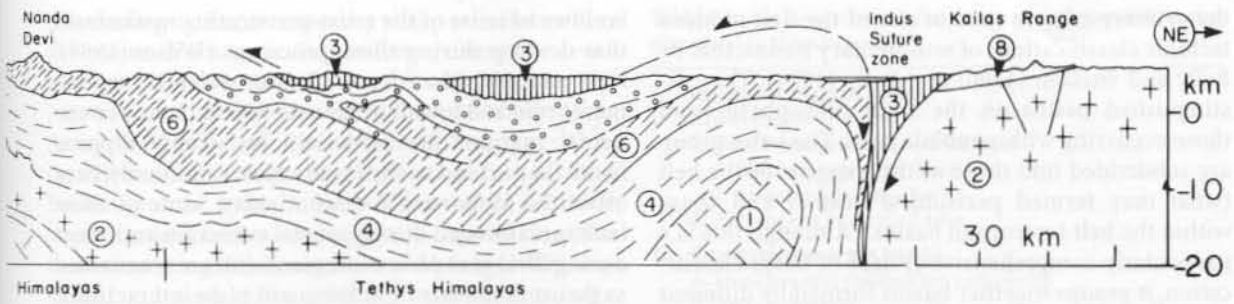


Fig. 9.1 (continued)

Plate-tectonic theory has shown us that all three of these parameters can change with time. This is a fundamental discovery, because it means that basins having several or many different origins are normally juxtaposed in the same orogenic belt. Many of the confusions of geosyncline theory can be resolved using this mobilist approach. For example, the collision of India and Asia has juxtaposed the shield and cratonic covers of two continents and has involved, in the intense deformation of the suture zone, thick Phanerozoic sediments originally developed as divergent (or passive) margin wedges on the flanks of the Tethyan Ocean, remnants of volcanic arcs, oceanic crust and overlying pelagic sediments (ophiolites), and immense thicknesses of detrital sediments derived from uplifts generated during and following continental collisions (Gansser 1964). Thus, the orogen contains at least four entirely different types of sedimentary basin (Fig. 9.1), and neither basin classification nor interpretation will be meaningful unless the entire tectonic history can be unravelled and each basin placed in its correct plate-tectonic context.

Mitchell and Reading (1978) pointed out that many basins as old as Jurassic can be regarded as modern, in the sense that they still occupy essentially the same plate-tectonic setting they did when formed. For example, the basins flanking the Atlantic Ocean, developed at various times from the Jurassic to the Tertiary, can be so classified because the Atlantic is still, as it was then, an actively spreading ocean. Conversely, all but the most recent basins in the North American Cordillera are ancient, in the sense that plate-tectonic patterns there have undergone several profound changes since the Triassic, and more are to be expected as the East Pacific Rise continues to interact with the westward-drifting North American plate. Numerous ancient basin types are, therefore, present in the Cordilleran region.

As discussed elsewhere in this book, geological interpretations of Jurassic and younger sediments are considerably facilitated by the evidence of sea-floor spreading preserved in magnetic anomalies and fracture zones on the ocean floors. Analysis and interpretation of this evidence have provided historical data on the position of rotation poles and the direction and rate of plate movement (Cox and Hart 1986), all of which considerably simplify basin analysis. The superb study of Indonesian tectonics by Hamilton (1979) is an excellent example of this approach. For the most recent events, this method can be supplemented by the development of fault-plane solutions from earthquake records. For ancient basins, the same procedure is sometimes attempted in reverse. For example, Phillips et al. (1976) deduced plate vectors for the British area during the Caledonian closure of the Iapetus Ocean, based on the orientation of structural surfaces, volcanic centers, the age of magmatic activity, and so on. It is the interpretation of complex ancient sutures such as this that offers the greatest challenge to basin analysts, and it is here that the availability of basin models should be the most useful.

The first of the three criteria for basin classification deals with the type of basement below the sedimentary basin. Oceanic crust normally sits between 2.5 km and 6 km below sea level, depending on its age (the thermal contraction of oceanic crust as it retreats from a spreading center is discussed in Section 8.2.2.1; Fig. 8.3), subsiding to as much as 11 km below sea level in trenches at subduction zones. Sediments deposited on oceanic crust are, therefore, all of deep-marine facies. Continental crust ranges from a few hundred meters below sea level up to the highest mountains, although the depth of the sedimentary basement varies considerably, depending on sediment thickness and structural shortening. Basins floored by continental crust are therefore typically shallow-marine to nonmarine. However, this simple subdivision breaks down at plate margins, particularly at divergent margins, where continental crust may subside because of attenuation, listric normal faulting, or ductile creep; conversely, sedimentation can add as much as 18 km to crustal thickness (usually compensated in part by subcrustal erosion or creep; Sects. 7.3, 9.3.1).

The position of the basin relative to the plate margin (the second of our classification criteria) is one of

the primary criteria used in one of the first modern tectonic classifications of sedimentary basins, that by Bally and Snelson (1980) and Bally (1984). They distinguished basins on the rigid lithosphere from those occurring within mobile belts. The latter group are subdivided into those at the margins of the belt (what they termed perisutural basins) and those within the belt (episutural basins). Although this is a particularly comprehensive system of basin classification, it groups together basins formed by different plate-margin processes. Since all sedimentary basins except those located on the cratonic basement are shaped by such processes, it is, perhaps, not the most satisfactory approach for use by basin analysts. Furthermore, as noted by Klein (1987), work subsequent to that of Bally and Snelson (1980) has indicated that some of their basin types (e.g., the Pannonian and Chinese types) are not as unique as formerly supposed and can be reclassified into more generalized categories.

In this chapter, basin classification is loosely based on the Wilson cycle of opening and closing oceans, and this emphasizes plate-margin behavior as the primary classification criterion. The primary importance of plate-margin behavior has been confirmed by the work of Cloetingh (1988), who showed that plate-margin stresses can be transmitted for thousands of kilometers, influencing stress fields and uplift-subsidence patterns (and hence local sea level) throughout plate interiors and generating a form of stratigraphic cyclicity (Sect. 8.2.3.1). Indeed, much mid-plate activity (faulting, basin formation, uplift, etc.) can be explained with reference to the effects of plate-margin processes propagated into or transmitted across plate interiors. However, not all mid-plate activity can be explained in this way. As argued in Section 7.5 and as discussed in more detail later in this chapter (Sect. 9.3.6), vertical motions of the crust unrelated to plate-margin processes (**epeirogenesis**, also termed **dynamic topography**) are now known to be of considerable regional significance.

We now know that the Wilson cycle is more complex than was first assumed in such classic papers as that by Dewey and Bird (1970). It does not effectively take into account the extensive oblique slip that occurs between many plates. Also, many orogenic belts are now seen to have been developed by processes of terrane accretion over hundreds of millions of years, rather than by the collision of major continents (the Indus suture may be something of an exception). This does not invalidate the Wilson cycle but requires that we use it to unravel the simultaneous movements of many small plates in addition to a few large ones. In other ways, recent work has re-emphasized the value of the Wilson-cycle concept. There is, in general, a tendency for continents to rift and rejoin along similar lines because of the crust-penetrating weaknesses that develop during these processes (Wilson 1988). For example, the deep-reflection-seismic experiments referred to above have shown that many continental margins are underlain by steeply dipping faults that extend to the lower crust. Commonly, they offset the Mohorovičić discontinuity. Some of these faults are initiated during crustal extension and shear during divergent plate movements but are reactivated as thrust faults during subsequent plate interactions.

One of the consequences of structural reactivation commonly is basin **inversion,** which is defined as the uplift, by compression or transpression, of a basin along its controlling fault system, resulting in uplift and partial exhumation of the basin fill (Buchanan and Buchanan 1995; Sect. 7.6). Typically, only a few of the faults present in the basin may be reactivated, and the deeper parts of the basin may be left largely undisturbed. Basin inversion may be a precursor to orogenic deformation. Inversion is of considerable importance to the petroleum industry, because it commonly generates new hydrocarbon traps and because of the effects the contractional movements have on fluid pressures and fluid transmission along fractures and faults.

As a result of the complexities noted above, some orogens are thought to be largely allochthonous (e.g., the southern Appalachians; Cook et al. 1979). More complex reactivations may occur. For example, the Wessex Basin of Britain developed during the late Carboniferous–Cretaceous by extensional reactivation of Variscan (Devonian–Carboniferous) thrust faults. During the Alpine orogeny (late Cretaceous–early Tertiary), renewed contraction along the same suite of faults brought about uplift and inversion of parts of the basin, while other areas underwent renewed subsidence as foreland basins (Karner et al. 1987; Kusznir et al. 1987). Other examples of reactivation are discussed later in this chapter.

As discussed below, plate-tectonic setting affects the stratigraphic composition and architecture of basin fills in many ways, both obvious and subtle. Carbonate sediments tend to be most common on mature, divergent margins (assuming climatic suitability) because of the lesser relief along these types of continental margins (e.g., the southern rim of the Tethyan Ocean in Africa and Arabia during the Mesozoic) and, hence, the smaller volume of detrital sediment than on young rifted margins, transform margins, or convergent margins, which tend to be characterized by sharp relief (Hay et al. 1988). The exceptions are at the mouths of major rivers, many of which tend to empty along divergent margins (e.g., the clastic-dominated Gulf Coast of Texas and Louisiana, compared to the carbonate-dominated coast of Yucatan). Another exception is the Red Sea, a young

rifted margin characterized by sharp relief but where the major drainage is directed away from the coast, and reefs develop on abandoned fan-delta lobes. Major rivers commonly flow along the axes of rifts (e.g., Amazon, Mississippi, Rio Grande, Rhine, upper Nile) or within forelands, paralleling sutures (e.g., Ganga, Brahmaputra, Indus). Submarine fans, similarly, show marked contrasts that partly depend on plate-tectonic setting. Shanmugam and Moiola (1988) suggested a broad differentiation between the large, mud-dominated systems of divergent margins and the smaller, sand-dominated fans of convergent and transform margins. There are exceptions to this; for example, the giant Bengal fan, although flanking the divergent margin of India, owes its origin to the convergent tectonics that led to the India–Asia suture. The petrography of clastic detritus is highly dependent on the plate-tectonic setting of the source area, a fact that has allowed the study of this topic to become a useful adjunct to basin-analysis studies (Sect. 9.4). Geochemical composition of detrital sands has been less well studied but, as would be expected, shows parallel trends (Bhatia 1983).

Ingersoll and Busby (1995) listed five subsidence mechanisms that generate accommodation space for sediment:

1. Crustal thinning as a result of extensional plate movements, including listric (detachment) faulting and creep in the lower crust (Sect. 7.3).
2. Supracrustal loading, particularly by magmatism or the stacking of nappes and thrust sheets at convergent plate margins and sutures (Sect. 7.4).
3. Mantle–lithosphere thickening during cooling or underthrusting.
4. Asthenospheric flow; subsidence above cold, downwelling convection currents (Sects. 7.5, 9.3.6).
5. Sedimentary and volcanic loading. The effects of such loading add to the subsidence initiated by the other mechanisms but may be the dominant process in settings such as remnant ocean basins, where tectonic subsidence processes are of minor importance.

Based on plate-tectonic setting, basins can be conveniently divided into five categories (Dickinson 1974; Table 9.1). As emphasized by Klein (1987), the nature of the basement on which the basin rests is an important defining parameter, as this governs subsidence behavior and the overall characteristics of the sedimentary environments:

1. Divergent-margin basins (Sect. 9.3.1) include intracratonic rifts that precede sea-floor spreading, divergent-margin–ocean-margin wedges, aulacogens, and failed rifts. These are all extensional basins that rest on continental or transitional crust.

Table 9.1. Classification of the types of sedimentary basins discussed in this chapter

I.	**Divergent-margin basins**
	A. Rift basins
	B. Ocean-margin rises and terraces
	1. Red Sea-type (youthful)
	2. Atlantic-type (mature)
	C. Aulacogens and failed rifts
II.	**Convergent-margin basins**
	A. Trenches and subduction complexes
	B. Forearc basins
	C. Intra-arc basins
	D Backarc basins
	E. Retroarc (foreland) basins
	F. Satellite (piggyback) basins
III.	**Transform- and transcurrent-fault basins**
	A. Basin setting:
	1. Plate boundary transform fault
	2. Divergent-margin transform (transfer) fault
	3. Convergent-margin transcurrent fault
	4. Suture-zone transcurrent fault (escape tectonics)
	B. Basin type:
	1. Basins in braided fault systems
	2. Fault termination basins
	3. Pull-apart basins in en echelon fault systems
	4. Transrotational basins
IV.	**Basins developed during continental collision and suturing**
	A. Peripheral (foreland or foredeep) basins (on downgoing plate)
	B. Remnant ocean basins
	C. Hinterland foreland, strike-slip, and graben basins (on overriding plate). Broken foreland intermontane basins
V.	**Intraplate basins**
	A. Cratonic basins (floored by continental crust)
	B. Oceanic islands, seamounts, plateaus (floored by oceanic crust)

Ingersoll and Busby (1995) include ocean-margin basins in the intraplate category. Once true oceanic crust has been generated and separates the two sides of a former rift system, these basins are "intraplate"; however, their structure and evolution are entirely due to their origin on divergent plate margins, and the Ingersoll and Busby classification is not used here.

2. Convergent-margin basins (Sect. 9.3.2) are basins related to magmatic arcs, including trenches and forearc, intra-arc, interarc, backarc, and retroforeland (retroarc) basins. Retroforeland basins are developed by supracrustal loading of the continental crust. Forearc, interarc, and backarc basins develop by extension and may be floored by continental or oceanic crust. Trenches develop by the downbending of the under-riding oceanic plate.
3. Basins associated with transform plate margins, transfer faults, and megashears (Sect. 9.3.3) may rest on either continental or oceanic crust and may be either extensional or contractional.

4. Basins generated during the process of continental collision and suturing (Sect. 9.3.4) include profo-reland (foredeep, peripheral) basins (contractional basins developed on the continental crust of the subducting plate), graben and wrench-fault basins (extensional basins developed on the continental crust of the over-riding plate), and remnant ocean basins (developed on oceanic crust). Many major orogenic belts consist of accreted terranes or microplates, and these raise particular problems of basin analysis (Sect 9.3.5).
5. Intraplate basins. The origin of cratonic basins located on continental crust (Sect. 9.3.6) may reflect long-term, deep-seated, intraplate processes, subsidence above downwelling mantle convection currents, or reactivation of ancient plate-margin lineaments. They are mostly of extensional origin. Cratonic basins may be markedly affected by processes at plate margins.

9.3 Basin Models

This section, comprising the main body of the chapter, presents a systematic description of the major basin models. For each basin type, the following essential information is discussed:

1. The plate-tectonic processes generating the basin
2. The mechanism of crustal subsidence
3. The structural geology of the basin
4. The typical evolutionary development of depositional systems

The discussion is based primarily on modern examples (in the sense defined in Sect. 9.2), and selected examples of interpreted ancient equivalents are briefly described or referred to. Precambrian basins are dealt with separately (Sect. 9.5). Theoretical thermal, mechanical, and other models of basin subsidence are described briefly, but space does not permit an extended treatment of this topic, which is dealt with at greater length in Chapter 7.

9.3.1 Divergent-Margin Basins

A distinctive suite of basins develops where plates rift and separate as a result of sea-floor spreading. Continental margins under these conditions are referred to as divergent or passive margins. The latter term is intended to suggest a tectonic contrast with the highly active character of convergent margins, but it is a misleading comparison. Several long-term, deep-seated, tectonic processes occur on so-called passive margins. These processes have profound effects on basin formation and evolution, so the term "passive" is a poor one. Crustal extension during backarc spreading develops very similar structural styles and basin architecture (Sect. 9.3.2.5), and the same processes may also occur in the hinterlands of suture zones (Sect. 9.3.4.10). Recent books edited by Frostick et al. (1986), Beaumont and Tankard (1987), Coward et al. (1987), and Landon (1994) contain many useful research papers dealing with extensional tectonics and sedimentation. Reviews of the rifting process and of rift basins have been given by Ruppel (1995), Şengör (1995), and Leeder (1995). The development of continental-margin terraces and rises on mature extensional margins is reviewed by Bond et al. (1995). Bond and Kominz (1988) provided a historical review of the development of concepts about "passive" margins.

In this section, we deal with basins formed during orthogonal spreading. Several important classes of basins also form by oblique spreading, controlled by transform and transfer faults. These are discussed in Section 9.3.3. A model for divergent margins is summarised in Table 9.2.

Table 9.2. Basin model for divergent continental margins[a]

Evolutionary stage	Structure-stratigraphy	Sedimentology
3	Few active faults (except growth faults in deltas), seaward-dipping, prograding (clinoform) wedges; regional unconformities and onlap–offlap relationships reflecting major stratigraphic cycles. Progradation may extend far onto oceanic crust	A. Continental coast-shelf-slope-rise clastic depositional systems B. Carbonate platform or ramp depositional systems C. Mixed carbonate-clastic systems showing crude cyclicity
2	Some active faults, blanket deposits in basins, wedge-out and drape over horsts. Sedimentation on oceanic or continental basement	A. Marginal fluvial, lacustrine, fan delta deposits with basin center containing either A1, thick evaporites, or A2, black, organic shales B. Carbonate platform and basinal pelagic deposits, some starved sequences
1	Rifted-arch and rim basins, many active faults, half-grabens, sag basins	Fluvial and lacustrine sediments, interbedded with mafic volcanics

[a] For orthogonal spreading only; see Section 9.3.3 for discussion of oblique spreading.

9.3.1.1 *Tectonic Review*

The availability of deep-reflection-seismic lines across several continental margins, beginning in the early 1980s, brought about a substantial change in our ideas about the mechanisms of crustal extension. For many years, the prevailing philosophy was based on the Sleep–Mackenzie model of pure shear, which is discussed in Section 7.3 (Fig. 7.15a). The rheological consequences of this model, including thermal changes, isostatic behavior, subsidence, and sedimentation rates, have been explored for such continental margins as the Atlantic margins of the United States, Canada, and France, and the Mesozoic–Cenozoic development of the North Sea Basin, and have been found to provide reasonably satisfactory explanations of basin architecture and evolution (Sect. 7.3), with some exceptions, as discussed below. However, seismic data indicate that a different model is required for many regions of continental extension. The model of simple shear illustrated in Fig. 7.15b evolved mainly from work in the Basin and Range Province of the United States, which is not an extensional continental margin in the normal sense of this term (some models treat it as a case of backarc spreading) but has been found to be a superb natural laboratory for the examination of processes of crustal extension (Sect. 9.3.2.6). The consequences of the two models are examined below.

Pertinent points to be discussed here regarding continental extension include:

1. The location of the initial rifts
2. The mechanism of rifting
3. The mechanism of subsidence
4. Rift kinematics (the extent of rifting, rift propagation, and relief generation)

Şengör and Burke (1978) and Baker and Morgan (1981) classified rifts into two categories: those resulting from differential stresses during plate evolution (the **passive-mantle hypothesis**) and those produced by convective upwelling in the mantle (the **active-mantle hypothesis**). Active rifting may be indicated by broad regional uplift and volcanism prior to rifting, as in Ethiopia and Kenya (Jarvis 1984; Fairhead 1986), whereas other parts of the East African Rift System may have developed by passive rift propagation from these areas (Jarvis 1984). Passive rifting may lead to upwelling of asthenospheric material and decompressional melting, resulting in the widespread outpouring of plateau basalts following crustal extension. In practice, distinguishing the two types of rifting may be difficult, and ancient rifts may display evidence of having passed through both conditions (Ruppel 1995).

Why do rifts occur where they do? Intraplate stresses may generate rifts along old lines of weakness (passive rifting), possibly because the crust is thinner there because of some previously existing rift or suture. The reopening of the modern Atlantic Ocean close to the Iapetus suture (Wilson 1966) is a good example of this. Similarly, southeast Africa separated from Antarctica in the Cretaceous (Hallam 1980) along the site of a Paleozoic failed rift (Natal Embayment; Hobday and Von Brunn 1979; Tankard et al. 1982). McConnell (1977, 1980) demonstrated that the Cenozoic East African Rift System follows a Precambrian taphrogenic lineament that may be Archean in origin. Baffin Bay, an incipient ocean, parallels Precambrian (Hadrynian) dyke swarms in Baffin Island (Fahrig et al. 1973), suggesting a great age for this lineation also. Veevers (1981) cited other examples.

Rifting may also be initiated by the upwelling of hot mantle (active rifting). Wilson (1963) suggested that this was a localized process, occurring in what he called **hot spots.** Morgan (1971, 1972) developed this idea into the **mantle plume** hypothesis. Upwelling causes uplift along the incipient rift, particularly over the mantle plumes, where the uplift has a domal shape with a relief of up to 4 km. Basaltic volcanism precedes and follows the uplift. Uplift generates horizontal extensional stresses, and it is these that may lead to the failure of the crust through extensional faulting. Burke and Dewey (1973) suggested that most triple-point plate junctions are initiated over these uplifts and that sea-floor spreading proceeds by the formation of spreading centers and transform faults linking these triple points, although Leeder (1995) points to several examples of rifts underlain by mantle plumes (as evidenced by outpourings of plateau basalts), that were not followed by continental separation. Many rifts may be generated by a combination of active and passive rifting processes. Underhill and Partington (1993) documented the presence of a transient plume and the resulting uplift during the Jurassic at what became the junction of the Viking Graben, Central Graben, and Moray Firth Rift in the North Sea Basin. This is an excellent study of the effects of plume activity on the details of regional stratigraphy and provides a tectonic explanation for a major sequence boundary in the North Sea.

On a detailed scale, the location of spreading centers sometimes seems puzzling. Why did the East Pacific Rise not extend through oceanic crust, but propagate into the continental United States, generating the ribbon of continental crust now called Baja California? Similarly, why did the Atlantic spreading center split the Lomonosov Ridge away from the Barents Shelf? Why was a fragment of continental crust, the Blake Plateau, left behind by Atlantic spreading in the Mesozoic? In the case of the Blake Plateau, the isola-

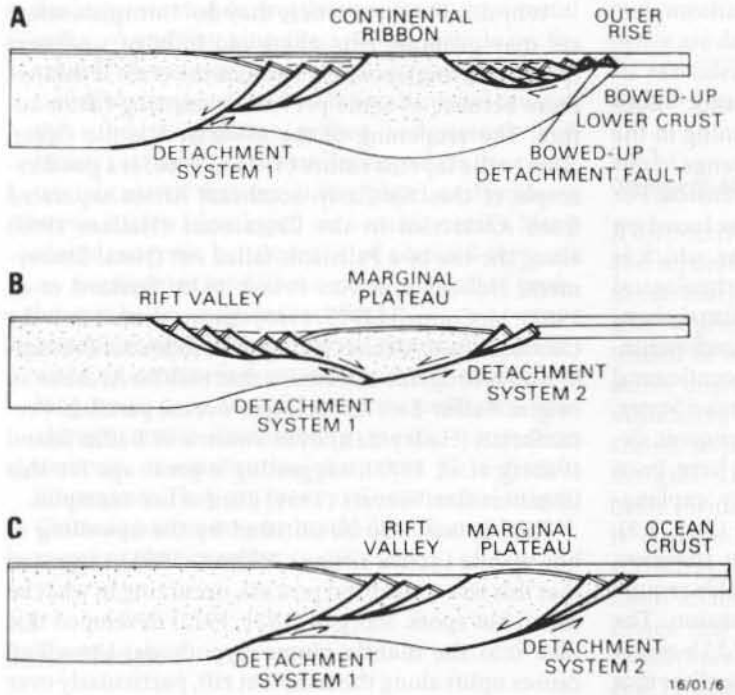


Fig. 9.2 A–C. The separation of continents by crustal detachment faults. More than one fault may develop, showing how marginal plateaus, internal rift valleys, and isolated continental ribbons may be created. (Lister et al. 1986, reproduced with permission of Geological Society of America)

tion of this fragment has been explained as a result of a jump in the spreading center, leaving an aborted rift separating the plateau from the main continental mass (Sheridan et al. 1981). Vink et al. (1984) demonstrated that continental lithosphere is weaker than oceanic lithosphere so that, even when oceanic lithosphere is generated along a rift, further brittle failure may take place preferentially within the continental crust. However, a model of simple shear may provide a more general explanation of the development of these continental plateaus and ribbons (Fig. 9.2). In the case of the East Pacific Rise, the spreading center was overridden by the relative westward drift of North America following the opening of the Atlantic Ocean in the Jurassic and may be responsible for some tectonic events in the Rocky Mountain states (Wilson 1988).

In the simple-shear model, the two separating margins have quite different structural styles and evolutionary development (Fig. 9.3). On the lower-plate margin, tilt blocks and half graben develop above the master detachment zone. These correspond to the rift phase in the classic Texas-Longhorn model of divergent-margin development (Fig. 7.4). Bounding faults are listric, a structural style that has, in fact, been recorded on many continental margins (Fig. 9.4). As the load of the upper plate is removed (leaving only the fragments comprising the tilt blocks), the lower-plate margin is, at first, isostatically uplifted, and the graben and sags remain nonmarine.

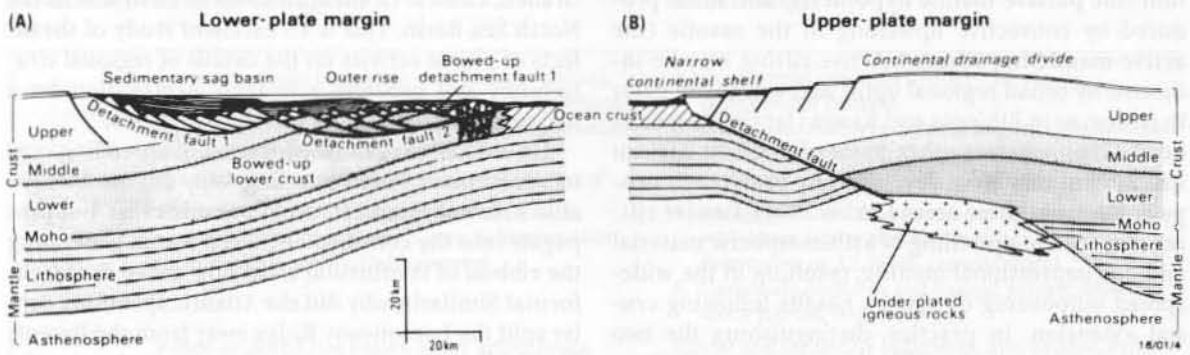


Fig. 9.3 A, B. Detachment-fault (simple-shear) model of continental extension and breakup. Two types of margin develop. The footwall becomes the lower-plate margin (**A**), while the hanging wall becomes the upper-plate margin (**B**). See text for explanation. (Lister et al., 1986, reproduced with permission of Geological Society of America)

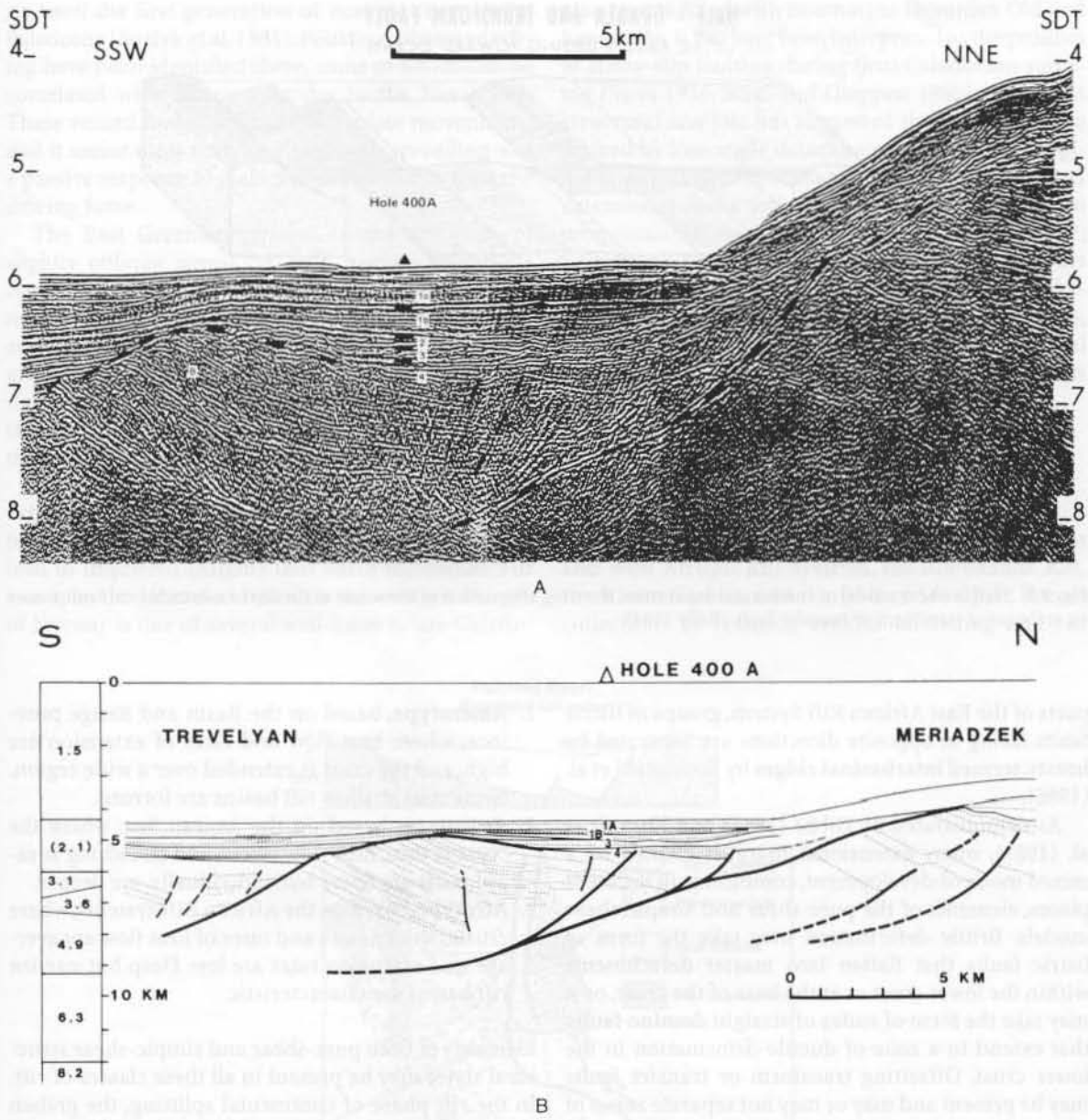


Fig. 9.4A, B. Seismic-line (A) and interpreted-line drawing (B) across northern Gulf of Biscay, showing listric faults developed during the early rift phase of continental separation (early Cretaceous). Numbered units are: *1a* Quaternary–upper Pliocene; *1b* Pliocene–lower Miocene; *2* Miocene–upper Paleocene; *3* Maastrichtian–Campanian; *4* upper Albian–upper Aptian. (Montadert et al. 1979; Bally et al. 1981)

Eventually, with continued continental separation, long-term thermal subsidence ensues, and the rifts subside and are covered by sediment (flexural phase in Fig. 7.4; units *1–3* in Fig. 9.4). The evolution of the upper-plate margin is different. The removal of the lower plate exposes the base of the upper plate to a hot, less dense asthenosphere, resulting in uplift and underplating by igneous rocks (Fig. 9.3B). The seaward edge of the margin undergoes extensional faulting, and it will also be covered by sediment and subside, so that the shallow structure, after several tens of millions of years, may appear quite similar to that of the lower-plate margin.

In plan view, other complexities may be apparent. The master detachment fault or faults may not dip in the same direction everywhere along the separating margins but may alternate in sense across orthogonal, steeply dipping transcurrent faults (Fig. 9.5). The latter have been referred to as continental transform faults (Bally 1984) or **transfer faults** (Gibbs 1984). In

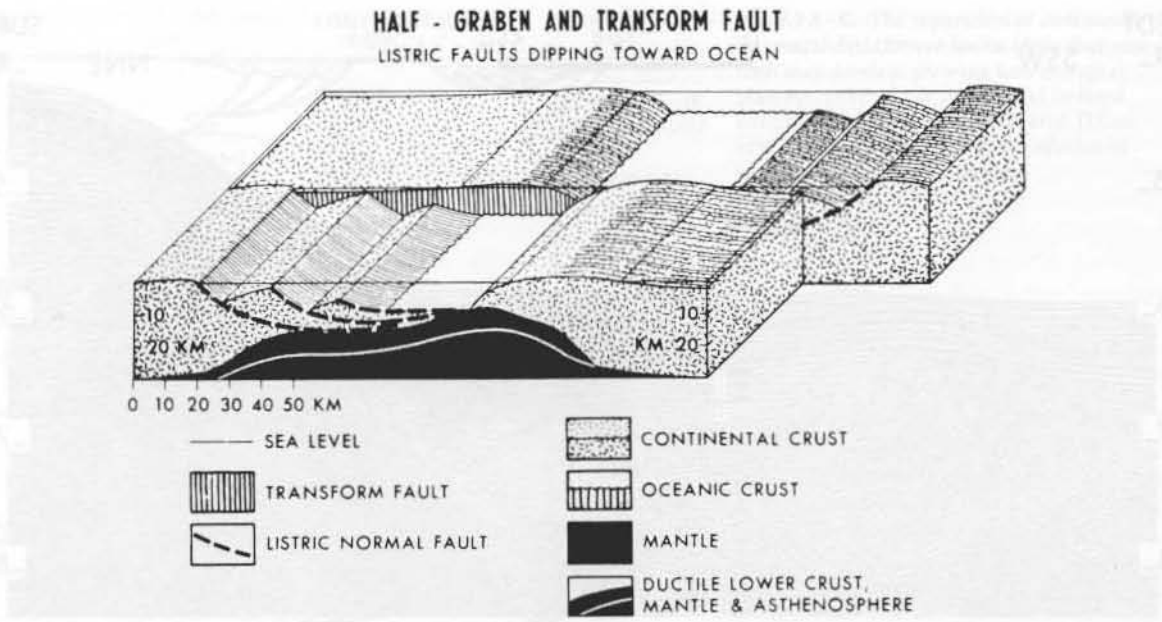


Fig. 9.5. Simple-shear model of continental separation, showing alternation of the sense of the dip of a detachment fault across an orthogonal transform or transfer fault. (Bally 1984)

parts of the East African Rift System, groups of listric faults facing in opposite directions are separated by horsts, termed **interbasinal ridges** by Rosendahl et al. (1986).

As demonstrated by Gibbs (1987) and Kusznir et al. (1987), many extensional margins formed by a mixed mode of development, combining, in different places, elements of the pure-shear and simple-shear models. Brittle deformation may take the form of listric faults that flatten into master detachments within the lower crust or at the base of the crust, or it may take the form of suites of straight **domino faults** that extend to a zone of ductile deformation in the lower crust. Offsetting transform or transfer faults may be present and may or may not separate zones of contrasting structural style or facing direction. Basins that form over the rift may develop complex collapsed and internally faulted ramps as extension continues. **Sag basins** form over individual detachments (Fig. 9.3) and over isostatically subsided zones of the upper-plate margin. Subsidence and sedimentation rates are rapid in rift basins. Tiercelin and Faure (1978) reported fault displacements and accumulation of fluvial–lacustrine sediments in the Ethiopian Rift and Danakil Depression at rates of 0.2–1.0 m/10^3 years. Clastic sedimentation rates locally reach 3.5 m/10^3 years, and accumulation rates are very high for lacustrine halite precipitation, possibly as much as 10–35 m/10^3 years.

Şengör (1995) classified rifts into three broad classes:

1. **Amerotype**, based on the Basin and Range province, where heat flow and rates of extension are high, and the crust is extended over a wide region. Numerous shallow rift basins are formed.
2. **Aegeotype**, based on the Aegean Sea, where the crust is thin, heat flow is low, and stretching is rapid. Rifts are fewer but, individually, are deeper.
3. **Afrotype**, based on the African Rift systems, where crustal thicknesses and rates of heat flow are average and extension rates are low. Deep but narrow rift basins are characteristic.

Elements of both pure-shear and simple-shear structural styles may be present in all these classes of rift. In the rift phase of continental splitting, the graben floors are initially above sea level and are floored by nonmarine sediments and volcanics. The duration of this phase seems to vary considerably, depending on regional spreading patterns. In many basin models, it is assumed that this phase is completed in a few million years (Sect. 7.3). Veevers (1981) suggested that true oceanic crust may be recognizable from ophiolite generation after 20 m.y. or less. However, where the plates are constrained by convergent plate movements driven by surrounding active spreading centers, as in the case of Africa at the present day, the rift phase may persist for a considerable period (Fairhead 1986). Rift basins bordering Baffin Bay span the Barremian to Eocene (60 m.y.; Miall et al. 1980; McWhae 1981), while those on East Greenland were intermittently active for about 170 m.y., from the early Trias-

sic until the first generation of oceanic crust in the Paleocene (Surlyk et al. 1981). Fourteen phases of rifting have been identified there, some of which can be correlated with rifting episodes in the North Sea. These record long-continued intraplate movements, and it seems clear that, here, sea-floor spreading was a passive response to plate movement rather than the driving force.

The East Greenland basins record a history of slightly oblique spreading, with marine conditions appearing first in the south and gradually extending northward by a process of rift propagation (Surlyk and Clemmensen 1983). Careful dating of the stratigraphic, magmatic, and structural history of the opening of the central Atlantic Ocean, which separated the United States from Africa, has shown that these margins also are diachronous (Withjack et al. 1998).

As noted at the beginning of this section, crustal extension and rifting occur in several distinct plate-tectonic settings, an understanding of which may lead to improved insights into basin formation. For example, the Hornelen Basin on the western margin of Norway is one of several well-known, late-Caledonian basins filled with nonmarine Devonian Old Red Sandstone. It has long been interpreted as the product of strike-slip faulting during final Caledonian suturing (Steel 1976; Steel and Gloppen 1980), but recent structural analysis has suggested that the basins are floored by low-angle detachment surfaces that originated as Caledonian thrusts and were reactivated as extensional faults following the end of Caledonian orogenesis (Hossack 1984; Seranne and Seguret 1987; Sect. 9.3.4). Extensional stresses may have been part of Variscan backarc spreading (Ziegler 1982). Strike-slip movement is apparent at high structural levels, because the detachment faults are scoop-shaped, and extensional faulting appears as strike-slip movement where the edges of the fault rise to the surface.

9.3.1.2 *Rift Basins*

Well-studied rift systems include the modern East and West African Rift Systems, the Rio Grande Rift, the Rhine Graben, and the Gulf of Suez Rift. Recent collections of research articles on rifting and rift

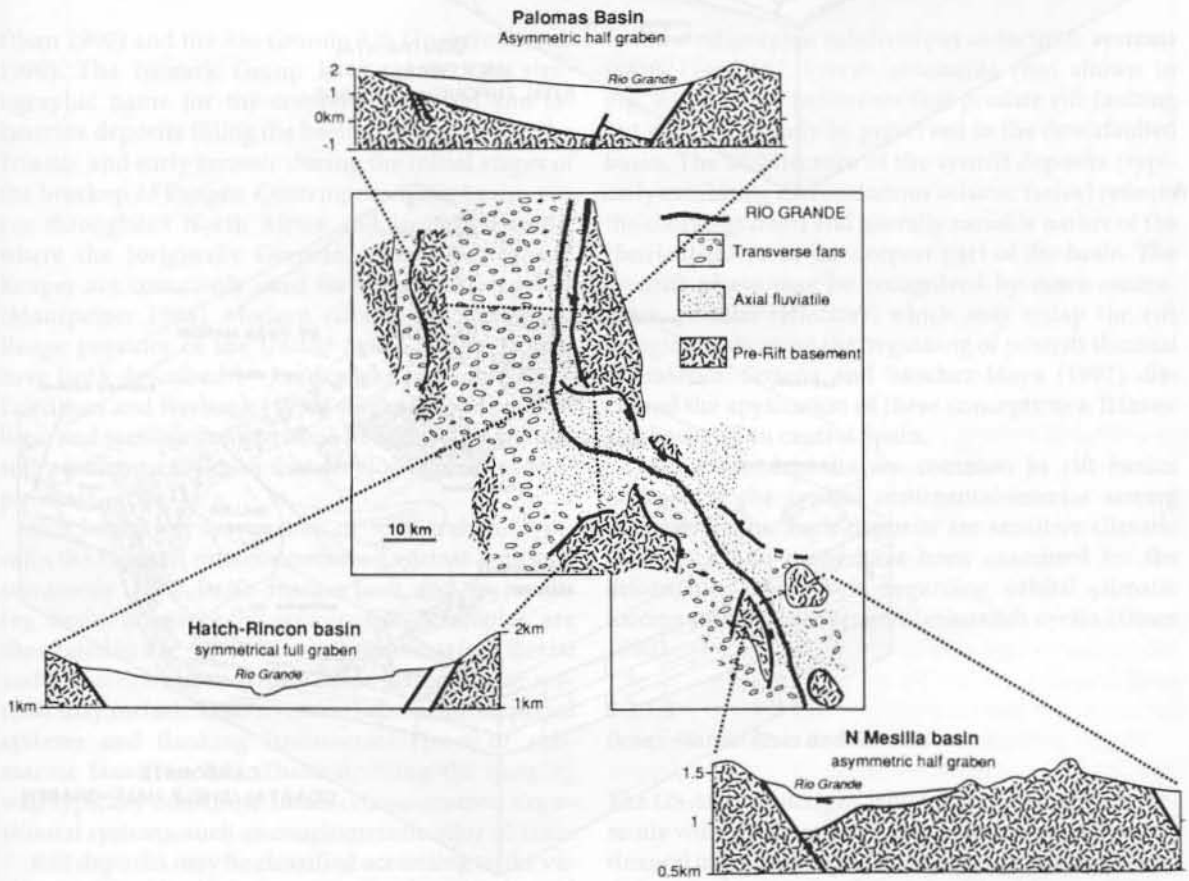
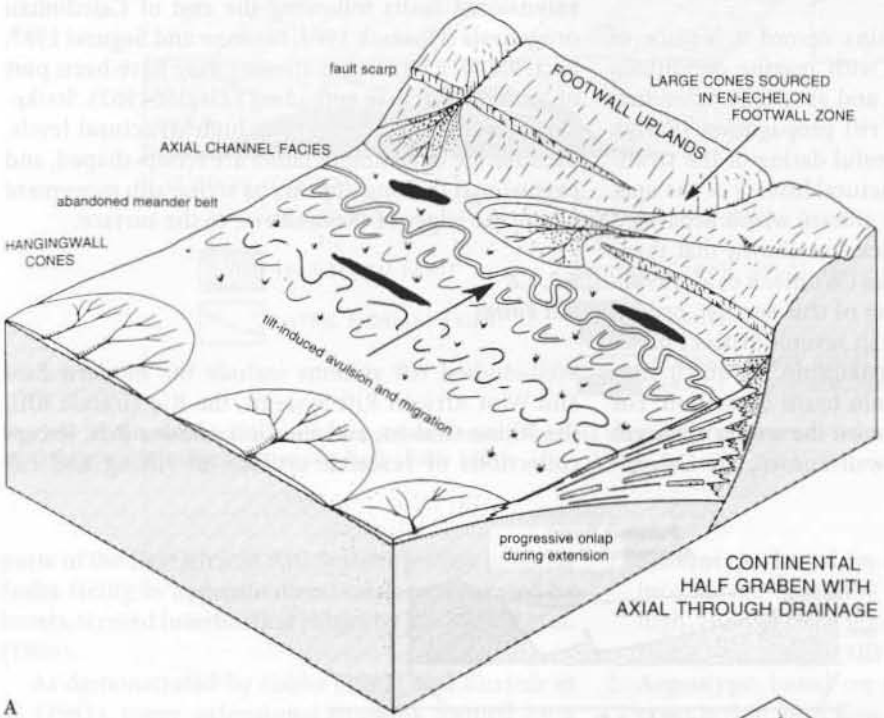


Fig. 9.6. The southern Rio Grande Rift, showing variations in the asymmetry of the basin and the distribution of the resulting depositional systems. Where the basin is asymmetric, the axial Rio Grande River is offset toward the deepest part of the basin, and flanking transverse alluvial fans extend across the relatively gently dipping footwall ramp. (Leeder 1995)

basins include Gangi (1991), Ziegler (1992), and Cloetingh et al. (1994). Depositional processes in the African rifts have been well described by Frostick et al. (1986), and several papers in Katz (1990). Models for rift sedimentation were developed by Leeder and Gawthorpe (1987), Leeder (1995), and Lambiase (1990), and some ideas regarding the relationship between sedimentation and tectonics, with illustrations of characteristic seismic facies, were provided by Prosser (1993). Among the most well-known ancient rift systems are the Newark Group of eastern North America (Manspeizer and Cousminer 1988;



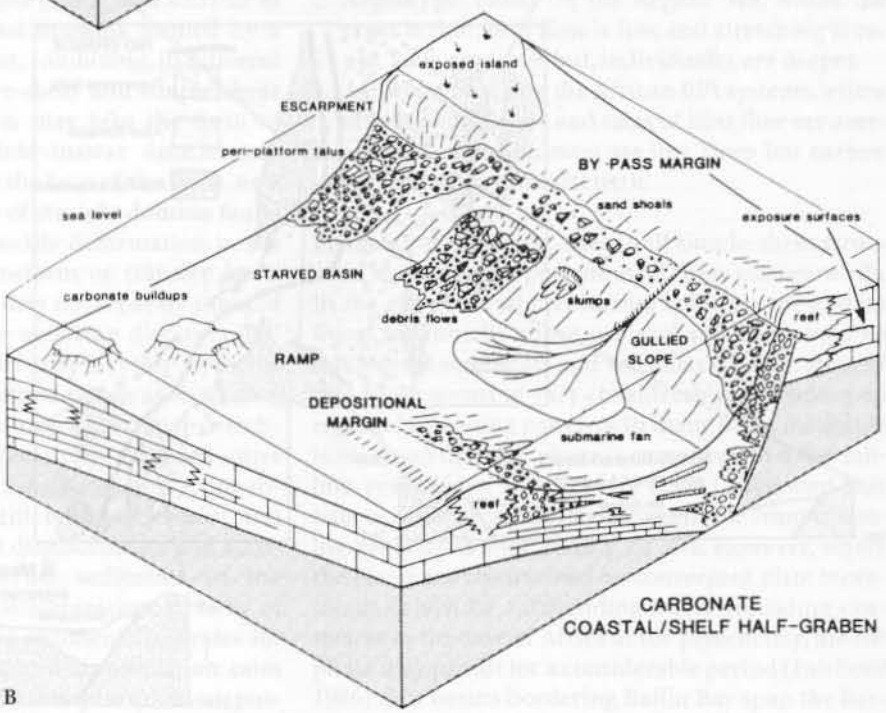


Fig. 9.7 A, B. Examples of depositional models for rift basins. **A** A nonmarine system, with an axial (through-going) river system and flanking alluvial fans. **B** A marine system, showing the development of a carbonate ramp on the hanging wall, and submarine talus slopes flanking the footwall fault scarp. (Leeder and Gawthorpe 1987)

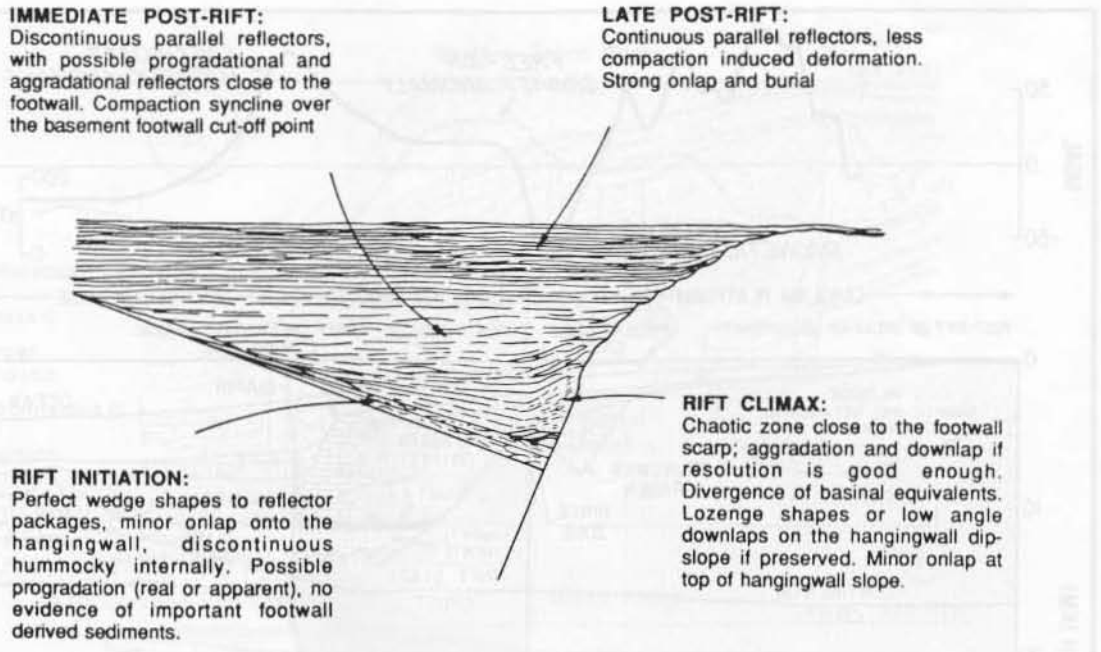


Fig. 9.8. An idealized seismic section interpreted in terms of tectonic systems tracts. (Prosser 1993)

Olsen 1990) and the Rio Grande Rift (Ingersoll, et al. 1990). The Newark Group is the American stratigraphic name for the complex of fluvial and lacustrine deposits filling the basins that formed in the Triassic and early Jurassic during the initial stages of the breakup of Pangea. Contemporaneous basins occur throughout North Africa and western Europe, where the (originally German) terms Bunter and Keuper are commonly used for the rift fill deposits (Manspeizer 1988). Modern rifts in the Basin and Range province of the United States and in Greece have been described by Leeder and Jackson (1993). Friedman and Burbank (1995) discussed sedimentologic and tectonic comparisons between rifts and the supradetachment basins that develop in areas of simple shear.

Rift basins occupy graben or half graben. Typically, the footwall rotates downward against a curved, commonly listric, basin-margin fault, and the resulting basin is markedly asymmetric. Examples are illustrated in Fig. 9.6. Rifts may be nonmarine (fluvial and lacustrine) or marine. Clastic depositional systems may include axial rivers or submarine dispersal systems and flanking (transverse) rivers or submarine fans (Fig. 9.7). Those draining the hanging wall typically constitute small, coarse-grained depositional systems, such as conglomeratic alluvial fans.

Rift deposits may be classified according to the various stages of basinal tectonism. Landon (1994) used the terms prerift, synrift, and postrift. Prosser (1993) developed a more detailed subdivision and referred to the stratigraphic subdivisions as **tectonic systems tracts** (Fig. 9.8). Prerift sediments (not shown in Fig. 9.8) include sediments that predate rift faulting but which may only be preserved in the downfaulted basin. The architecture of the synrift deposits (typically exhibiting discontinuous seismic facies) reflects the coarse-grained and laterally variable nature of the clastic deposits in the deepest part of the basin. The postrift phase may be recognized by more continuous, parallel reflectors, which may onlap the rift margins, indicating the beginning of postrift thermal subsidence. Sopeña and Sánchez-Moya (1997) discussed the application of these concepts to a Triassic rift-basin fill in central Spain.

Lacustrine deposits are common in rift basins because of the typical continental-interior setting of these basins. Such deposits are sensitive climatic indicators, and many have been examined for the information they yield regarding orbital climatic forcing and the resulting Milankovitch cycles (Olsen 1990).

9.3.1.3 *Ocean-Margin Rises and Terraces*

The US Atlantic margin, which has developed concurrently with sea-floor spreading orthogonal to the continental margin, is regarded as the type example of an extensional continental margin (Sheridan and Grow 1988; Sheridan 1989; Fig. 9.9). This margin formed the basis for the pure-shear extensional-margin model

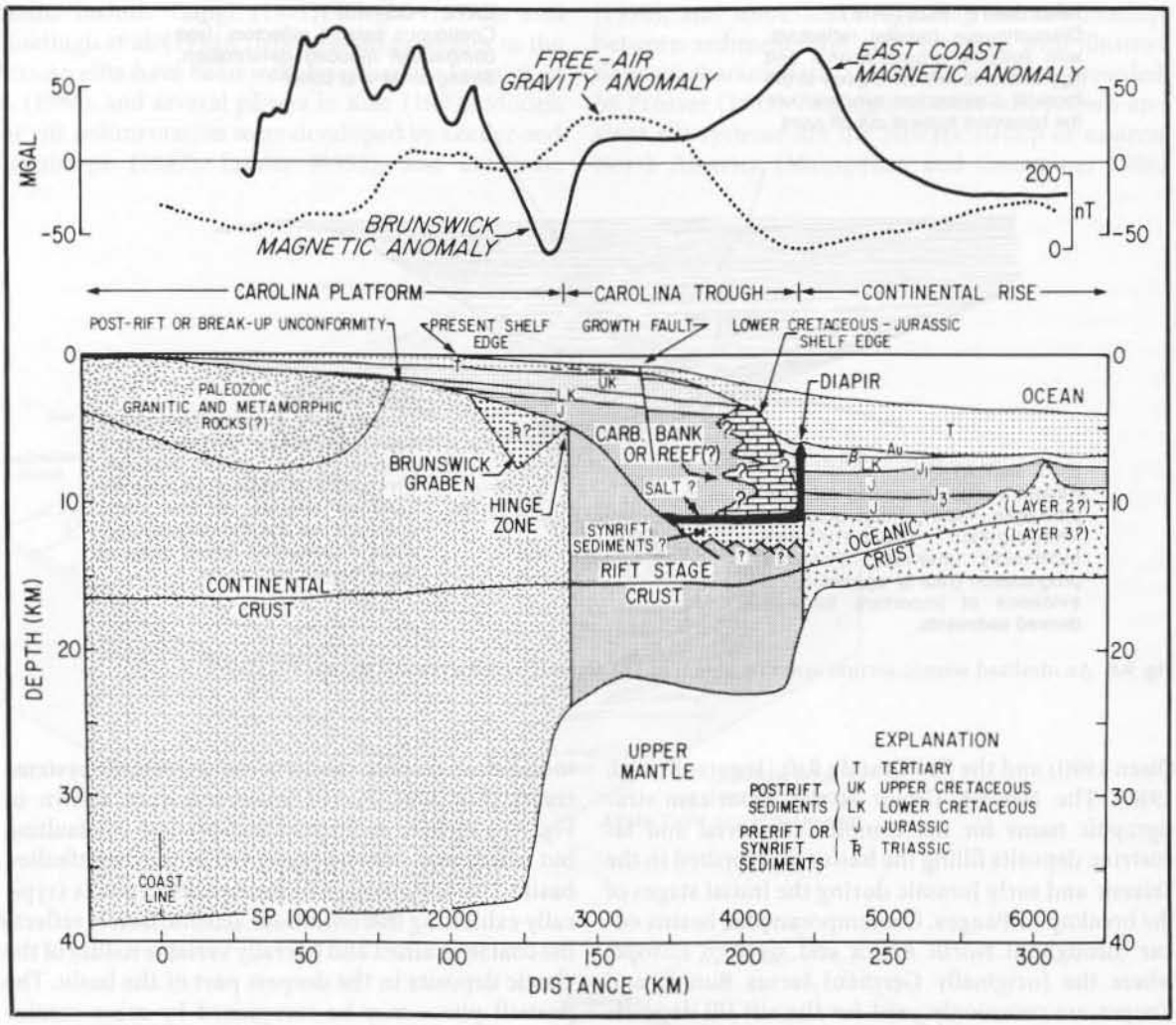


Fig. 9.9. Cross section across the most typical segment of the US Atlantic margin across the coast of South Carolina. (Hutchinson et al. 1982)

described in Section 7.3. The margin can be divided into three zones based on basement composition: (1) a landward portion underlain by normal continental crust, (2) an intermediate portion beyond the hinge zone of the basin, which is underlain by continental crust that has been thinned by ductile creep or shear or modified by magmatic intrusion, and by transitional oceanic crust consisting of sedimentary units and basaltic rocks, and (3) the most seaward portion, underlain by oceanic crust (Bond et al. 1995).

The overlying sediment prism is asymmetric, consisting of a seaward-thickening wedge of sediment that may be up to 18 km thick, and is characterized by two distinct internal structural styles, reflecting the two stages of basin development (Fig. 9.10). The rift basins that form in the initial stages of continental breakup are, typically, deeply buried at the base of the section, except at the landward limit of the margin (for example, the exposed Newark basins of the US margin; Fig. 9.11). The rift basins typically are truncated by a regional unconformity that records a short-lived flexural rebound event that occurred at the time of continental breakup, when the generation of oceanic crust commenced at the center of the rift (Falvey 1974). This is termed the **breakup unconformity** (Bond et al. 1995) or the **postrift unconformity** (Withjack et al. 1998). Above this unconformity is the stratigraphic section generated by thermal subsidence and flexural downwarp of the underlying, thinned edge of the continental crust (Sect. 7.3). This section normally onlaps the continental margin, which attests to the gradual deepening and widening of the basin. In a seaward direction, these deposits thicken and may be characterized by a clinoform architecture, reflecting the development of the wedge by seaward progradation.

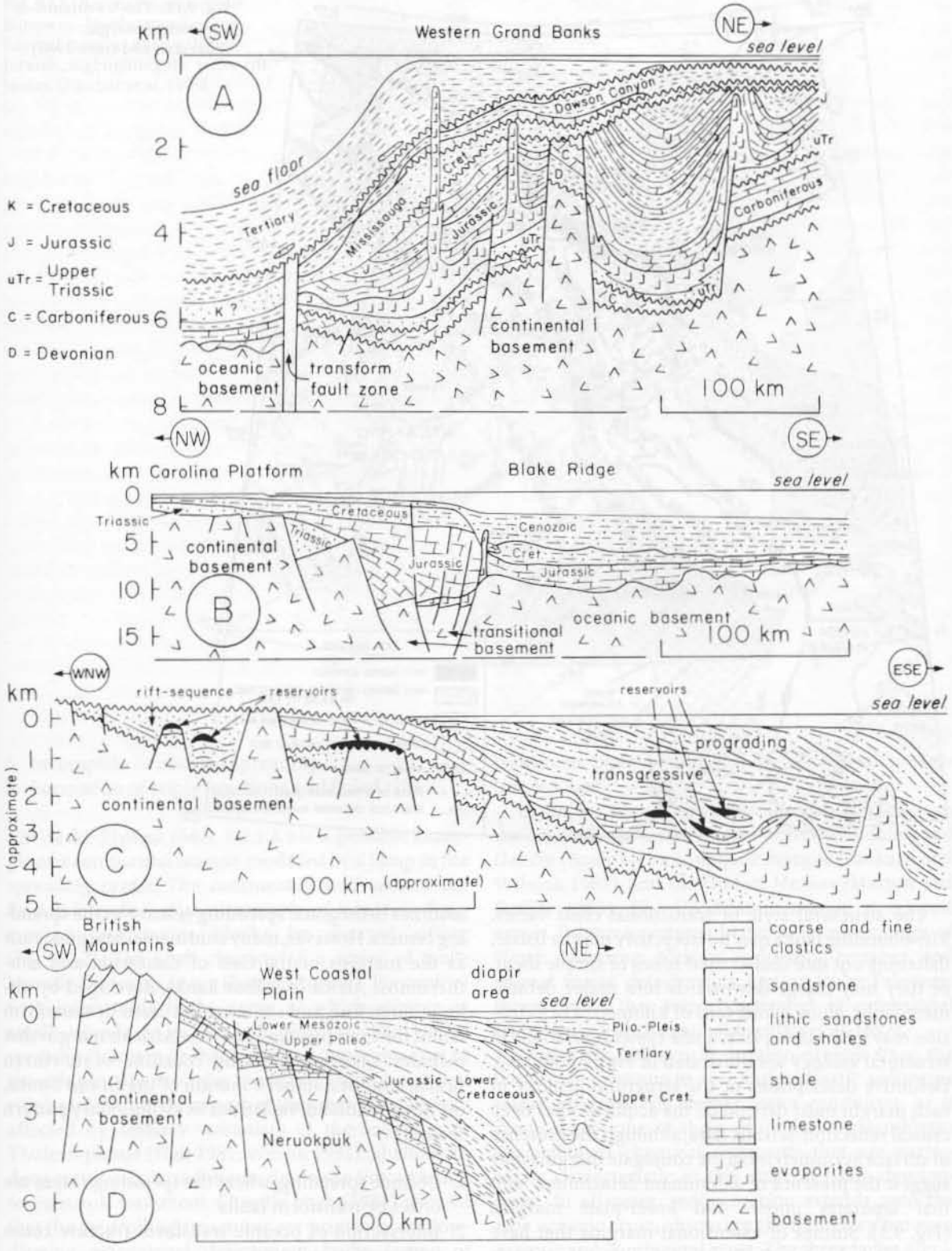


Fig. 9.10 A–D. Examples of mature continental-margin rises. **A** Grand Banks, Newfoundland (McWhae 1981); **B** Caroline Trough, off Cape Fear, North Carolina (Sheridan et al. 1981); **C** an idealized section across the Brazilian continental margin (Ponte et al. 1980); **D** Beaufort Mackenzie Basin, northern Canada. (Hawkings and Hatlelid 1975)

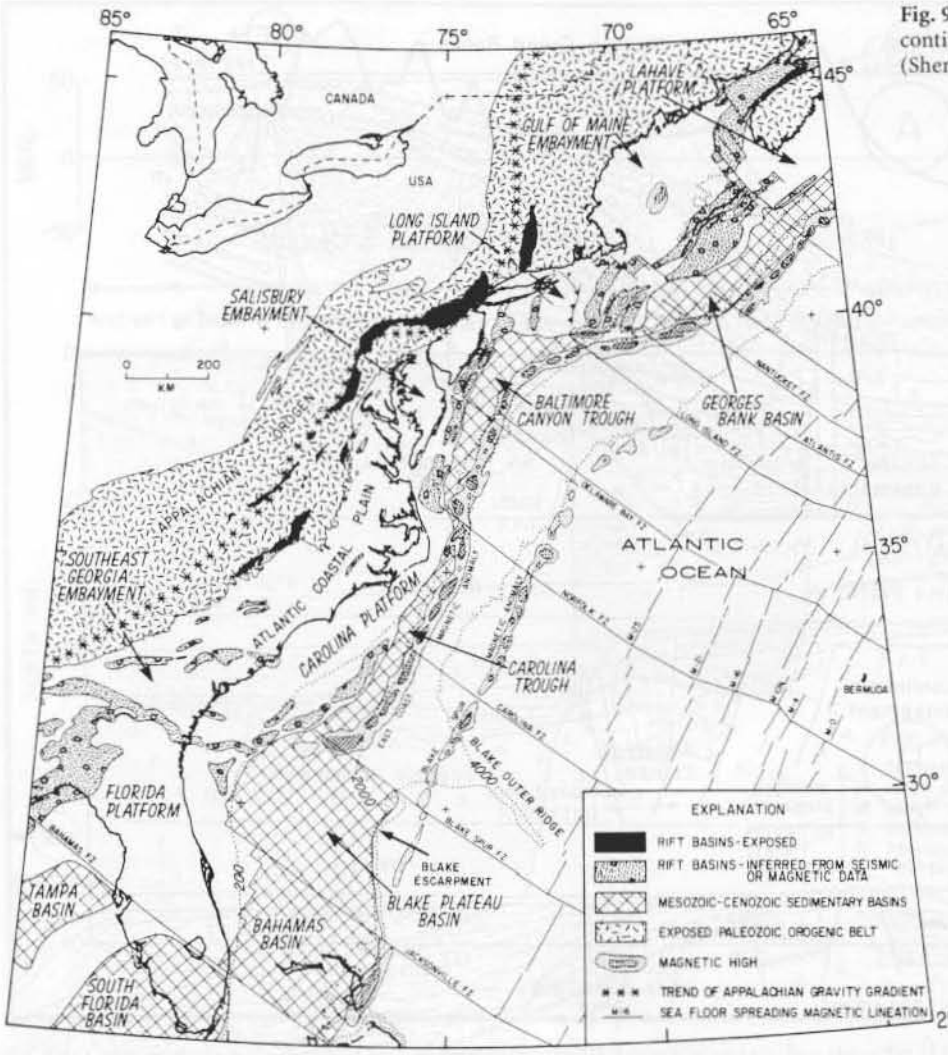


Fig. 9.11. The US Atlantic continental margin. (Sheridan and Grow 1988)

The structural style of transitional crust varies. Rift-bounding faults may be steep, they may be listric, flattening out into mid-crustal zones of simple shear, or they may merge downwards into major detachment faults, along which tens of kilometers of extension may have taken place. Some typical variations in structural geology are illustrated in Figs. 9.2 and 9.3. Definitive descriptions of the structural geology of each margin must depend on the acquisition of deep crustal reflection-seismic data, although the presence of surface asymmetries in the conjugate margins may suggest the presence of a dominant detachment fault that separates upper- and lower-plate margins (Fig. 9.3). Studies of extensional margins that have been carried out so far indicate that the structure is highly variable.

The standard model for extensional continental margins, which formed the basis for the pure-shear extensional-margin model described in Section 7.3, assumes orthogonal spreading relative to the spreading centers. However, many continental margins, such as the margins of the Gulf of California and southernmost Africa (Agulhas Bank), developed by oblique spreading, and some are bordered by transform faults, for example, the east–west African margin that includes Ghana, the facing coastline of northern Brazil, and the southern margin of the Grand Banks, off Newfoundland. Variations in evolutionary pattern include:

1. Oblique spreading, where the spreading ridges are offset by transform faults
2. Intersection of oceanic transform fracture zones with continental margin
3. Changes in position of spreading centers or in spreading direction
4. Fragmentation of the continental margin into major fault blocks

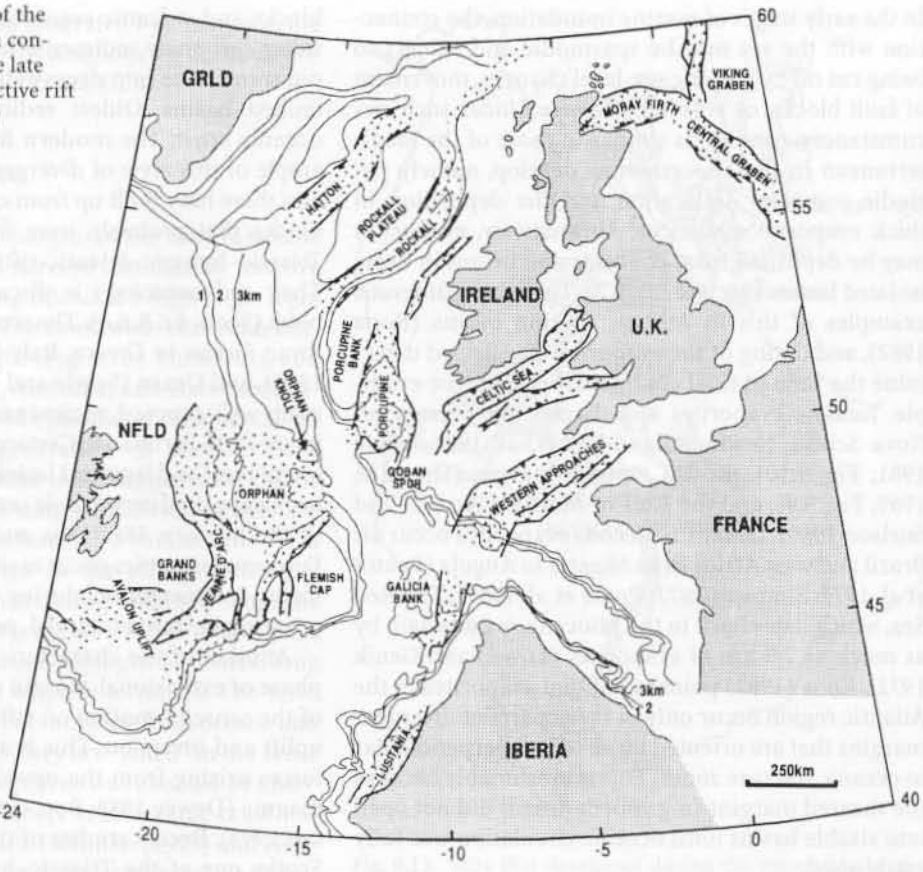


Fig. 9.12. Reconstruction of the European–North American continental margins during the late Jurassic, highlighting the active rift basins. (Sinclair et al. 1994)

5. Incomplete or aborted spreading
6. Formation of triple junctions and failed rifts

The Blake Plateau (Sect. 9.3.1.1.) is a possible example of a continental margin modified by a jump in the spreading center. The continental shelf around the British Islands is a good example of a margin fragmented into major blocks by aborted rifting (Fig. 9.12). The shelf consists of a series of fault-bounded troughs (e.g., Hatton, Rockall, Porcupine) with intervening highs, some of which emerge as small islands (e.g., Rockall, Faroes). Short-lived spreading centers probably existed between some of these highs (the troughs, therefore, qualify as failed rifts), and the northern part of the area was markedly affected by Tertiary volcanism of the Icelandic or Thulean plume (Hall 1981; Wilson 1981). Shallow- to deep-water clastic sediments drape the Precambrian to Paleozoic basement. Cheadle et al. (1987) showed that the controlling structures are primarily shallow-dipping extensional detachment faults (some of which cut deep into the basement) and wrench faults. Many of the faults are reactivated Paleozoic structures, including thrust faults formed during Caledonian suturing. Examples of oblique spreading and transform plate boundaries are discussed in Section 9.3.3.3.

Examples where the simple-shear model has been used include the Grand Banks of Newfoundland and Galicia (Spain) – its conjugate margin (Tankard and Welsink 1987), and the Gulf of Mexico (Marton and Buffler 1993). The US Atlantic margin developed across Paleozoic sutures between an assemblage of North American, African, and exotic terranes, and several major detachments appear to be Paleozoic thrust faults that were reactivated as extensional faults during the Jurassic (Sheridan et al. 1993).

During the late-synrift or proto-oceanic stage, the basin may continue to accumulate clastic sediments under progressively deeper water conditions, or it may contain one of three other facies assemblages consisting of chemical sediments that are particularly characteristic of this stage of basin development. In all cases, sedimentation extends onto the new oceanic crust, obscuring the boundary between oceanic and continental crust. The three facies are:

1. Evaporites
2. Black, organic shales
3. Starved-basin carbonates and pelagics

In the early stages of marine inundation, the connection with the sea may be spasmodic and subject to being cut off by eustatic sea-level changes, movement of fault blocks, or volcanic barriers. Under such circumstances, conditions similar to those of the Mediterranean in the Miocene may develop, namely periodic complete desiccation and the deposition of thick evaporite sequences. Alternatively, evaporites may be deposited from concentrated brines in deep, isolated basins (Sects. 4.6.6, 6.7). There are numerous examples of this in Atlantic margin basins (Rona 1982), and dating of the evaporites has helped determine the time of final continental rifting. For example, Jurassic evaporites and diapirs are present off Nova Scotia, Newfoundland (McWhae 1981; Wade 1981; Fig. 9.10), the US Atlantic margin (Sheridan 1989; Fig. 9.9), and the Gulf of Mexico (Worrall and Snelson 1989). Lower Cretaceous evaporites occur off Brazil and west Africa from Nigeria to Angola (Pautot et al. 1973; Simpson 1977; Ponte et al. 1980). The Red Sea, which dates back to the Miocene, is underlain by as much as 7.5 km of evaporites (Lowell and Genik 1972). Rona (1982) pointed out that evaporites in the Atlantic region occur only in those parts of divergent margins that are oriented more or less perpendicular to oceanic fracture zones. This is presumably because the sheared margins (e.g., north Brazil) did not open into sizable basins until oceanic circulation was fully established.

At several times during the Phanerozoic, including the lower Paleozoic, Jurassic, and Cretaceous, extensive deposits of black organic shale (2–15% organic carbon by weight) formed in the world's oceans. The causes of this distinctive deposit include high organic productivity and restricted marine circulation. Youthful oceans, where volcanic ridges or upfaulted basement sills are present to inhibit current movement, are ideal environments for this type of sedimentary accumulation, and Arthur and Schlanger (1979) showed that many of the Cretaceous deposits were located in such a tectonic setting. However, this facies is by no means restricted to young oceans; for example, the best modern analogue for the euxinic environment is the Black Sea, an oceanic embayment within a suture zone. Arthur and Schlanger (1979) argued that, at least during the Cretaceous, "equable climates and low latitudinal thermal gradients would have led to sluggish circulation and therefore slow renewal of oxygen to deep-water masses." This could help explain black-shale accumulation in other tectonic settings (Sect. 4.5.2).

Where continental-margin relief is low, sediment supply is minimal, and sedimentation may be dominated by platform and pelagic carbonates and other fine-grained, pelagic facies. Facies distribution is controlled by basement topography. Uplifted fault blocks and volcanic seamounts are sites of shallow-water carbonate sedimentation, which may show rapid transitions into deep-water facies in intervening, faulted basins. Oldest sediments rest directly on oceanic crust. The modern Bahamas are a good example of this style of divergent margin. The carbonates there have built up from continental-margin fault blocks that probably were first formed during the Triassic–Jurassic Atlantic-rifting episode (Sect. 7.8). Their sedimentology is discussed elsewhere in this book (Sects. 4.6.8, 6.6). The southern margin of the Tethyan Ocean in Greece, Italy (Bernoulli and Jenkyns 1974), and Oman (Searle and Graham 1982) provides many well-exposed, ancient examples. The rocks range in age from Permian to Cretaceous. Those in Oman are mainly reef and lagoonal facies and are thought to have been deposited on volcanic seamounts similar to those of the modern Mauritius and Reunion Islands. The European examples occur mainly on continental basement. Sedimentation during the Jurassic was slow, resulting in various starved, pelagic facies.

Another of the characteristics of the proto-ocean phase of extensional-margin development is reversal of the sense of motion on rift faults, leading to basin uplift and inversion. This is attributed to ridge push forces arising from the upwelling of asthenospheric magma (Dewey 1988; Bott 1992; Boldreel and Andersen 1993). Recent studies of the Fundy Basin of Nova Scotia, one of the Triassic–Jurassic rifts that developed during the initial rifting of the modern Atlantic Ocean, have demonstrated this effect there (Withjack et al. 1998).

The fully mature **extensional** or **divergent margin** that evolves about 10^7 m.y. after the start of sea-floor spreading is characterized by a decrease in active faulting (except for growth faults in deltas) and an increase in the scale of continental-margin depositional systems. The early rift basins are buried by immense wedges of carbonate and/or clastic sediment, which may show evidence of both lateral progradation and vertical upbuilding. Much of the information we now have on large-scale depositional systems is derived from extensional-margin basins of this type (e.g., Gulf of Mexico, Brazil), and the reader is referred to Chapter 6 for a discussion. Sea-level change profoundly influences the stratigraphic architecture, and extensional-margin basins were where many of the ideas about sequence stratigraphy were first developed. Similarly, the best evidence for the existence of regional and global stratigraphic cycles is derived from divergent margins, as described in Chapter 8. Widespread unconformities are common in the continental-terrace sediment wedge. At one time, it was thought that these reflected a history of repeated eustatic sea-level changes (Vail et al. 1977), but it is now thought that many of these unconformities are the product of

regional tectonic warping generated by changes in the in-plane stress regime as the neighboring ocean develops by sea-floor spreading (Sect. 8.2.3).

9.3.1.4 Aulacogens and Failed Rifts

Bordering the divergent margins of the world's oceans are numerous rifts and narrow, tectonically inactive marine troughs extending inland at a high angle from the continental margin. The northwest-European borderland is a good example (Fig. 9.12). These rifts, and others bordering Africa, Australia, and elsewhere, developed during short-lived episodes of continental extension prior to the generation of oceanic crust (Fig. 9.13). The continental margin may later collide with another continental margin following subduction of the intervening ocean, forming a suture. This event may reactivate structures in the rift, producing a mildly deformed trough trending obliquely from the suture. These types of rifts are now called **aulacogens**, a term coined by the Russian geologist N. Shatski long before the development of plate tectonics. The term **failed rift** is retained for rifts striking from oceans into continents (Burke 1977). They are "failed" in the sense that their tectonic development was aborted by changes in plate kinematics before true oceanic crust was developed along their axes. Hoffman (1973) and Hoffman et al. (1974) described the structural and stratigraphic evolution of aulacogens (although some of their examples have now been reinterpreted), and Burke (1977) summarized the structural complexities of various basin types. Şengör (1995) referred to this class of basin as **fossil rifts**.

The first plate-tectonic interpretation of these rifts was proposed by Burke and Dewey (1973), who showed how divergent margins could develop from triple junctions. One arm of a ridge–ridge–ridge (RRR) junction commonly "fails" after a few million years of continental rifting, in the sense that spreading is aborted shortly before or after the development of true oceanic crust, while complete continental separation occurs on the other two arms. The failed arms form deep troughs extending at a high angle away from the continental margin. The Benue Trough in Nigeria has become widely regarded as a classic example of a failed rift (Hoffman et al. 1974). However, Wilson and Williams (1979) and Benkhelil (1982) noted the presence of three Atlantic fracture zones striking into the trough, and Benkhelil (1982) suggested that the trough is a complex pull-apart basin. In support of this analysis, he presented evidence of structural patterns inconsistent with simple, orthogonal extension. The triple-junction model is no longer regarded as an essential prerequisite for the generation of rifts. Their development is recognized



Fig. 9.13. Rifts that developed during the generation of the Atlantic Ocean by sea-floor spreading. This is a conceptual diagram. The rifts were not all active at the same time. (Burke 1976a)

as an inevitable consequence of the stretching and fragmentation of a continental plate under extensional stress. Brewer et al. (1983) pointed out that failed rifts and aulacogens may, like other types of rift, be located along earlier lines of weakness (Sect. 9.3.1.1). Their locations and orientations may not, therefore, reflect sea-floor spreading trajectories in any obvious way.

The structural and stratigraphic evolution of failed rifts and aulacogens parallels that of divergent margins, except that the basin continues to be floored by continental crust and, therefore, does not subside to abyssal depths. A genetic model of crustal extension and upwelling of hot asthenosphere was suggested for the Oklahoma Aulacogen (Proterozoic–Permian) by Hoffman et al. (1974) and for the Viking-Central Graben of the North Sea (Triassic–Cretaceous; Figs. 9.14, 9.15) by Sclater and Christie (1980). This model provides for a simple two-phase development: an initial graben subsidence during crustal extension and a downwarping phase during thermal contraction and flexural loading (Steer's-Head or Texas-Longhorn model of basin geometry; Fig. 7.4). The intrusion and loading effect of dike swarms may

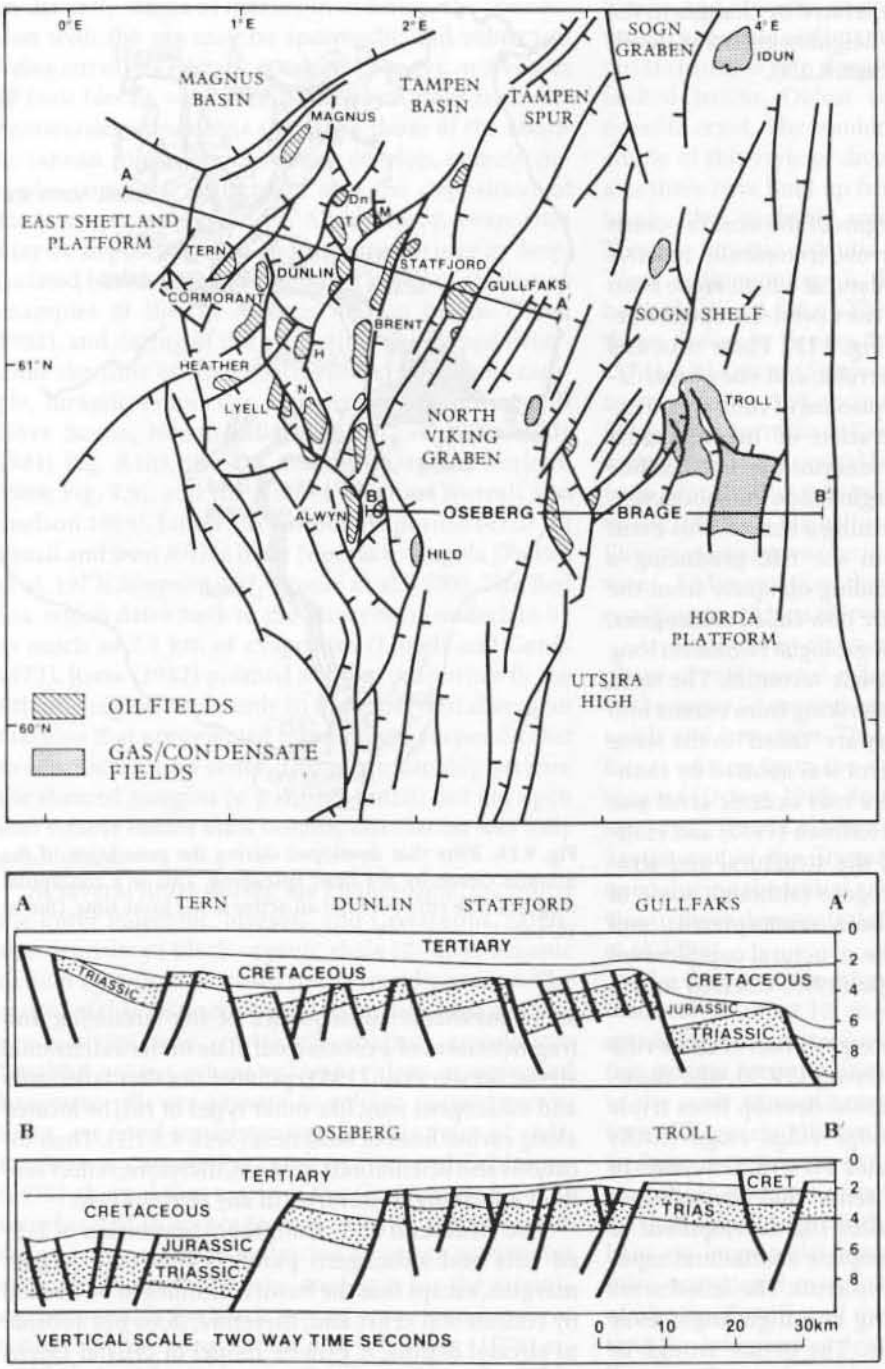


Fig. 9.14. Structural geology of the north Viking Graben, showing the numerous tilted fault blocks that form the structural traps for the Brent and other Triassic–Cretaceous oil fields. (Parsley 1984)

be important in some aulacogens (Burke 1976a, 1977). The duration of these phases is partly controlled by the constraints of the mantle-upwelling-and-cooling model and partly by regional stress patterns imposed by plate-spreading vectors. For example, the Viking–Central Graben (Figs. 9.14, 9.15) shows several periods of rifting (Ziegler 1980, 1982) that can be correlated with rifting episodes in the east-Greenland divergent margin (Surlyk et al. 1981). The graben was also affected by strike-slip faulting (Fig. 9.15). Beach et al. (1987) suggested that the North Viking Graben was initiated by low-angle detachment faulting in the Triassic–early Jurassic, but was modified by domino faulting (which offset the Mohorovičić discontinuity and contains a strike-slip component) during the late Jurassic–early Cretaceous. A 50%

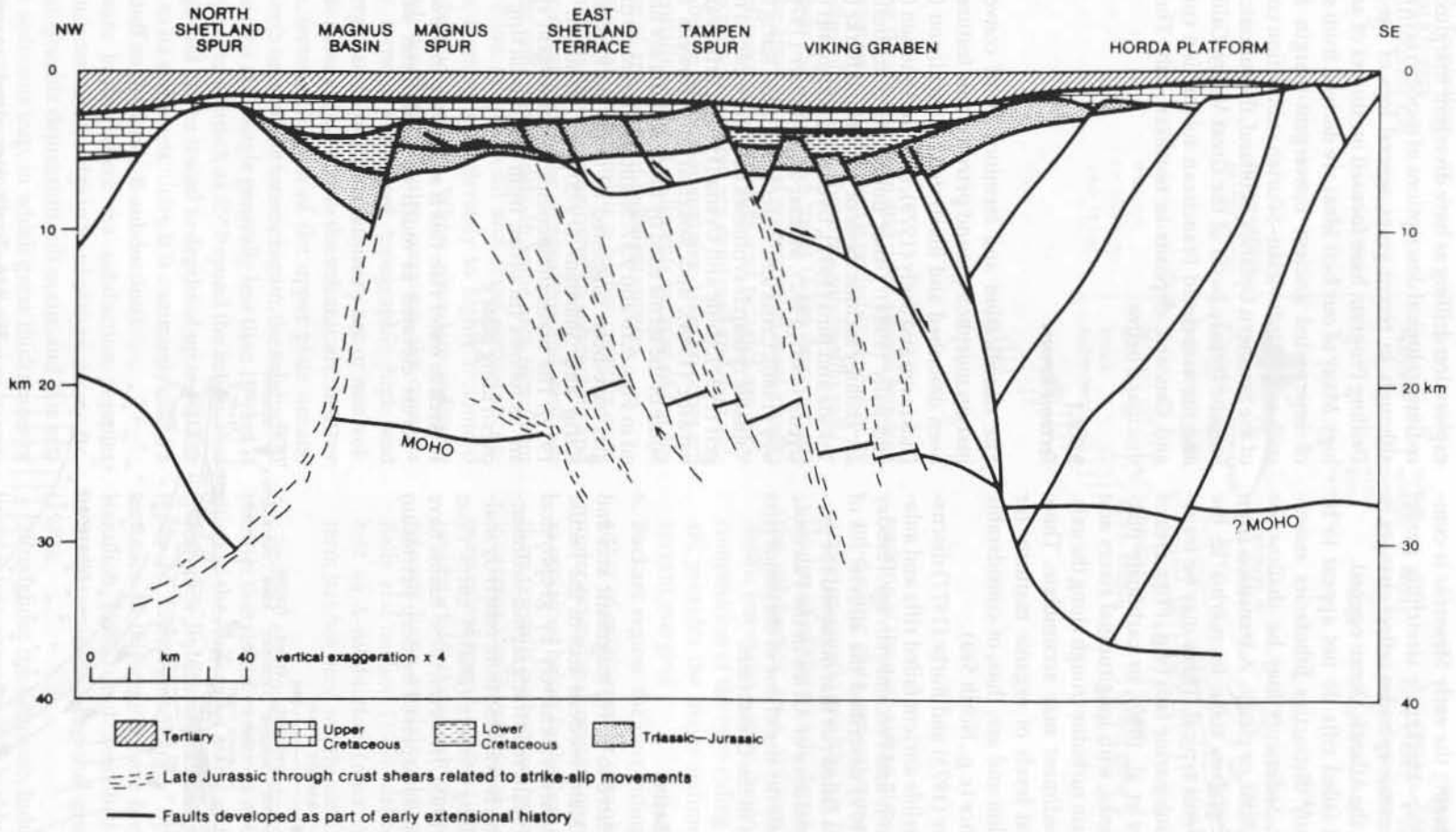


Fig. 9.15. Geological model for a seismic section across Viking Graben and adjacent regions of the North Sea Basin. Early extensional faults are shown by *solid lines*. These were modified and offset by steeper faults (*dashed lines*), which had a strike-slip component of movement during the late Jurassic–early Cretaceous. Upper Cretaceous and Tertiary sediments were formed during the flexural phase of basin formation. (Beach et al. 1987)

crustal extension during the early Mesozoic is compatible with the Sleep–Mackenzie stretching model (Sect. 7.3). These tectonic episodes reflect changes in plate movement as the Atlantic Ocean opened.

Aulacogens and failed rifts do not appear to be characterized by any distinctive lithofacies assemblage or sequence. Sediments may be shallow- or deep-marine, carbonate or clastic. A transition from nonmarine in the graben stage to marine in the downwarp stage seems typical. There may be transverse alluvial fans, submarine fans (e.g., Frigg area of North Sea; Heritier et al. 1980), or carbonate platforms along the flanks, with longitudinal rivers and deltas or a basin-plain turbidite trough along the axis. Up to 10 km of sediment may accumulate. These typically show ideal levels of organic maturity for petroleum generation and are, thus, of considerable economic importance (e.g., North Sea).

Burke and Dewey (1973) and Burke (1977) discussed a variety of possible ancient failed rifts and aulacogens of Phanerozoic and Precambrian age. Hobday and Von Brunn (1979) described the alluvial fill of Natal Embayment, a failed rift that occupied the present southeast coastal area of Africa in the Paleozoic and was subsequently the site of one of the major rifts through Gondwana in the Cretaceous.

9.3.2 Convergent-Margin Basins

These are basins related to active magmatic arcs and subduction zones. Many modern arcs in the Pacific have now been studied extensively by geophysical methods, and some scattered stratigraphic–sedimentological information is available from relatively shallow Deep-Sea Drilling Project (DSDP) cores (Von Huene 1981; Klein 1985). However, these basins have not been as extensively explored by deep petroleum exploration drilling as have divergent margins. Good sedimentological descriptions of modern arcs are few although, in recent years, several legs of the Ocean Drilling Program have focused on studies of arc geology. Many of our best ideas are derived from studies of interpreted ancient convergent-margin basins, such as the Ordovician–Silurian subduction complex of the Southern Uplands, Scotland, the Jurassic to Paleogene forearc basin of the Great Valley, California, and the associated Franciscan subduction complex, and Cenozoic deposits in northern Italy. These are discussed below.

9.3.2.1 *Tectonic Review*

The classification and terminology of convergent-margin morphologic and petrotectonic features have been described and illustrated by Dickinson (1974), Dickinson and Seely (1979), Bally and Snelson (1980), and Bally (1984). Other important contributions to arc geology include those of Wilson and Burke (1972), Toksöz and Bird (1977), Uyeda and Kanamori (1979), Uyeda (1981, 1982), Molnar and Atwater (1978), Dewey (1980), Cross and Pilger (1982), and Karig (1982). Valuable research syntheses have been edited by Leggett (1982), Hilde and Uyeda (1983), Carlson and Kobayashi (1984), and Kokelaar and Howells (1984). Some of the terms used by those authors are illustrated in Figs. 9.16 and 9.17. Slightly different terminologies have been suggested by other workers, such as Karig and Sharman (1975) and Hamilton (1977, 1979). The main components of convergent systems are as follows (in order, from the subducting to the over-riding plate):

1. **Arch** or **outer rise**: this is an upwarp in the oceanic crust elevated as much as a kilometer or more

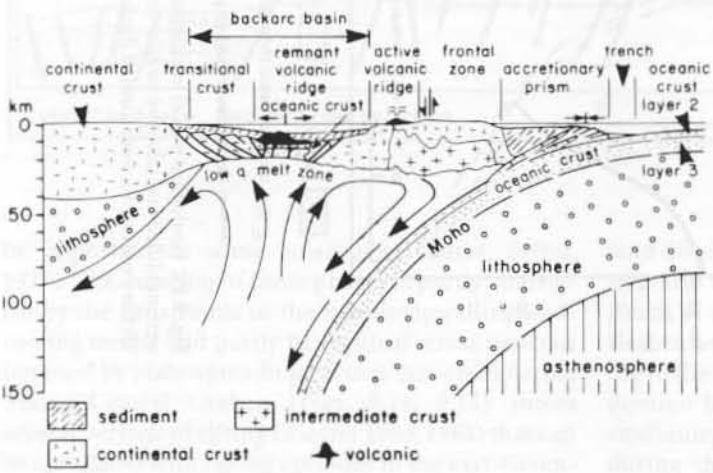


Fig. 9.16. Terminology related to a convergent margin involving a continental over-riding plate. A model for the development of a back-arc basin is shown based on Toksöz and Bird (1977; also Bally and Snelson 1980)

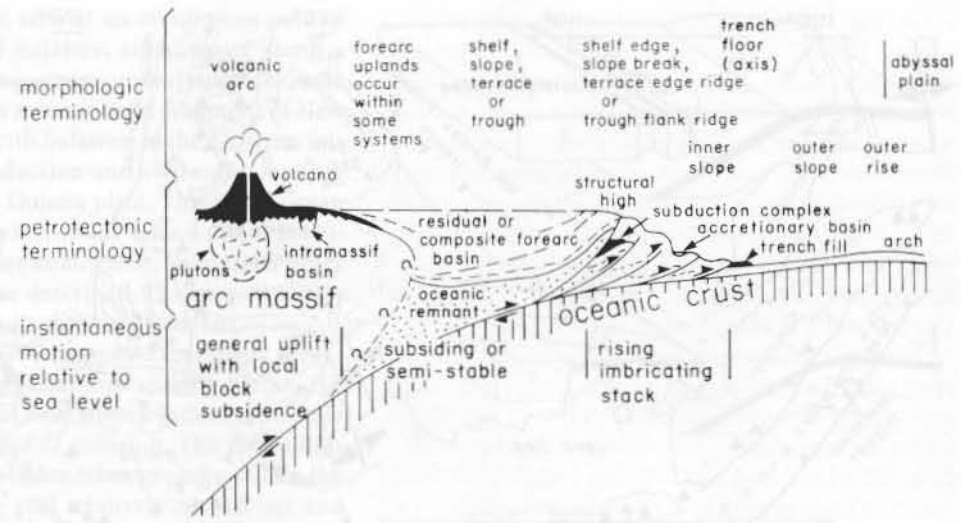


Fig. 9.17. Terminology related to a forearc region. (Dickinson and Seely 1979)

above the abyssal plain and up to several hundred kilometers wide. The arch is probably generated by bending of the crust above the subduction zone.

2. **Trench**: this is a trough up to 11 km deep in front of the arc and the final repository of most arc-derived sediment. Turbidites and pelagics are typically present here but are deformed and over-ridden by the subduction complex. Many trenches are essentially empty of sediments because of the absence of any adjacent sediment supply.
3. **Subduction zone**: this is a highly complex zone formed by the shearing between subducting and over-riding plates. The composition and structure of the zone vary widely. In some arcs, significant accretion of trench-fill and slope sediments occurs, forming a moderately to highly tectonized subduction complex. Where abundant sediment supply leads to extensive accretion, the complex may form a prominent topographic high and emerges as an outer **accretionary arc** or **non-volcanic outer arc** parallel to the volcanic arc. In other arcs, tectonic erosion of the upper plate occurs. Beneath most of the forearc region, the subduction zone dips at an angle generally less than 10°, but it steepens to as much as 85° beyond the subduction hinge and can be traced to depths of up to 600 km due to its active seismicity. It is commonly called a **Benioff zone** (hence, **B-subduction**).
4. **Forearc basin**: where the subduction complex forms a terrace or ridge in front of the volcanic arc, a topographic depression will occur, and this acts as a sediment trap, in which great thicknesses of sediment may occur. The forearc region ranges from 50 to 300 km in width.
5. **Magmatic arc**: where the subduction zone dips to a depth of between about 100 and 150 km, partial melting of the over-riding plate and possibly of the subducting plate generates magma that feeds an active magmatic arc. The arc lies at least 100 km landward of the trench, depending on the dip of the subduction zone. Volcanism is typically andesitic but varies widely, depending mainly on the composition of the over-riding plate. Uplift of this arc provides the major sediment source on convergent margins.
6. **Backarc region**: the over-riding plate may consist either of continental or oceanic crust and may be under extensional, neutral, or contractional stress, as explained below. In the case of extensional arcs, there typically is a **backarc basin** floored by oceanic or transitional crust. An old term for such basins is **marginal seas**. A contractional arc is characterized by a **fold–thrust belt** adjacent to a **foreland** or **retroarc basin** overlying the cratonic hinterland. Bally and Snelson (1980) defined the fold–thrust belt as **A-subduction** (after Ampferer), but the term has not been widely used.

Arc geology encompasses a daunting array of complexities that reflect variations in the age and composition of the converging plates and variations in plate trajectories. Indonesia provides examples of most types of convergent-margin behavior. The plate tectonics of this region have been summarized by Hamilton (1977) and described in greater detail in a masterpiece of regional geology by Hamilton (1979).

Throughout the Cenozoic, Indonesia and the Philippines have experienced a highly complex history as a result of the convergence of three major plates. The India–Australia plate is moving northward and approaching Asia at a velocity of about 10 cm/year,

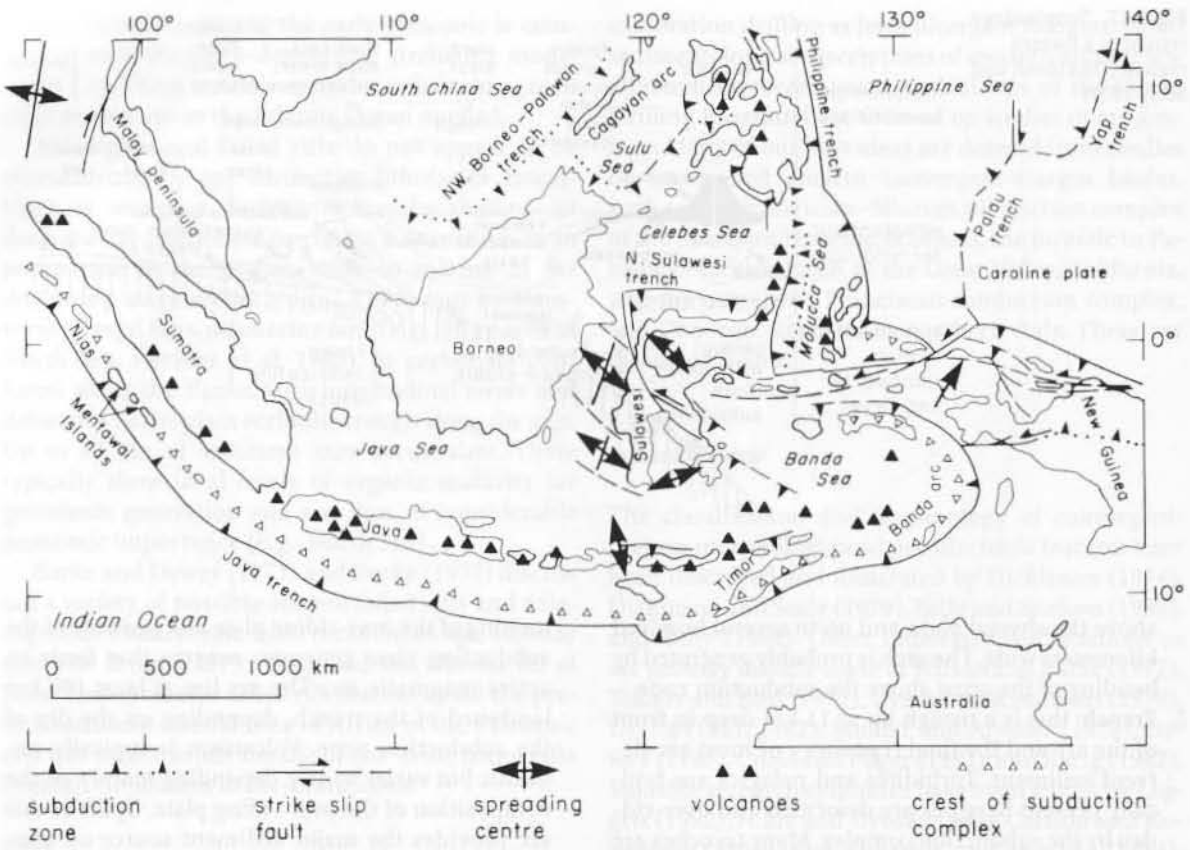


Fig. 9.18. Major Cenozoic tectonic elements in the Indonesian region. (Hamilton 1977, 1979)

while the Pacific plate is moving relatively westward. The Indonesia–Philippine region

> has responded with rapidly changing patterns of subduction, spreading, migration, strike-slip faulting, oroclinal folding, and rotation. After arcs have collided with each other or with continents, convergence continues at a new subduction zone on the outside of the collision aggregate. Some arcs die without collision, some arcs reverse polarity, and others migrate and change shape as spherical plates inflect through them (Hamilton 1977).

Some of the major Cenozoic tectonic elements in the area are shown in Fig. 9.18. The Java–Sumatra arc is a classic example of a convergent margin, showing an active volcanic arc, a foreland floored by continental crust to the north, a deep-marine forearc basin, a deep trench partly filled with turbidites of the Bengal Fan, and a well-developed outer accretionary arc consisting of uplifted trench-fill sediments between the trench and the forearc basin. Where the sediment supply from the Bengal Fan has been high, such as the trench along the southern margin of Sumatra, the outer arc is massive and has been uplifted to form the Mentawai Islands. Toward the east, the forearc region crosses the island of Timor, which consists mainly of a subduction complex containing Permian to Paleogene shallow-water strata representing the northernmost edge of the Australian continental plate. The arc has collided with a major continental mass, and the subduction zone may be undergoing the process of "flipping", a reversal in polarity, with a new southward-dipping subduction zone appearing to the north of the present volcanic arc. This arc continues eastward as the Banda arc, which curves northward against New Guinea. Hamilton (1977; 1979, Fig. 77) suggested that the curvature resulted from a collision with New Guinea coupled with a continued tendency for the subduction hinge (the bend in the oceanic crust at the outer rise) to migrate over the oceanic plate.

The Sangihe and Halmahera island arcs face each other in the Molucca Sea (Fig. 9.18) and are in the process of colliding. Collision is complete in the north (in Mindanao) but incomplete in the south. North Sulawesi is undergoing clockwise rotation bounded by an active, sinistral, strike-slip fault to the west and a subduction zone to the north. Convergence on this zone

decreases eastward to zero at an orocline or pole of rotation. Borneo and Sulawesi rifted apart along a mid-Tertiary spreading center underlying Makassar Strait. East Sulawesi is a continental fragment of New Guinea that sutured with Sulawesi in the Miocene following westward subduction and strike-slip faulting of the Australia–New Guinea plate. This small ocean is flanked by the giant Mahakam delta, a major petroleum province (Verdier et al. 1980). Numerous other complications could be described. They are all superimposed on earlier, quite different patterns.

Hamilton (1977, 1979) suggested that many megasutures, such as the Appalachians and the Alpine–Himalayan system, would have shown Indonesian-style tectonics in early stages of collision. The difficulties inherent in resolving Indonesian geology, where the tectonic elements are still relatively spread out and undeformed, pale by comparison with the task of sorting out a major suture zone, where thousands of square kilometers of terranes, microplates, arc complexes, and oceanic crust have been compressed into narrow, linear orogenic belts.

We discuss suture zones and microplates in greater detail in Sections 9.3.4 and 9.3.5. The purpose of introducing the topic at this stage is to make the point that convergent margins cannot be properly understood on the basis of a simple two-dimensional analysis such as the cross sections of Figs. 9.16 and 9.17. These imply orthogonal convergence, which is practically never the case, and tell us little about variations in subduction behavior in response to sediment supply and sediment loading, highly oblique convergence, variations in the age of oceanic crust, presence of small (but still subductable) masses, such as volcanic seamounts on the subducting plate, the effects of subducting spreading ridges or oceanic fracture zones (old transform faults), or the nature of backarc spreading. Some of these points are discussed below.

It has long been recognized that convergent margins in the western Pacific are subducting mainly Mesozoic crust older than about 100 Ma and that most show a history of present or relatively recent backarc spreading. Conversely, in the eastern Pacific, subducted crust is mainly younger than about 50 Ma and the arcs there are located on continental crust with active backarc contraction (retroarc forelands and fold–thrust belts). A significant point about Pacific arcs is that, in general, the older the subducted crust, the steeper the dip of the Benioff zone. The most widely accepted model for these variations in convergence patterns is that of Molnar and Atwater (1978) and Dewey (1980). They attributed the contrasts in arc type to the thermal properties of the downgoing slab: the older it is, the colder and more dense it is and, hence, the faster its rate of sinking. It is suggested that, where the subducting crust is old, it sinks faster

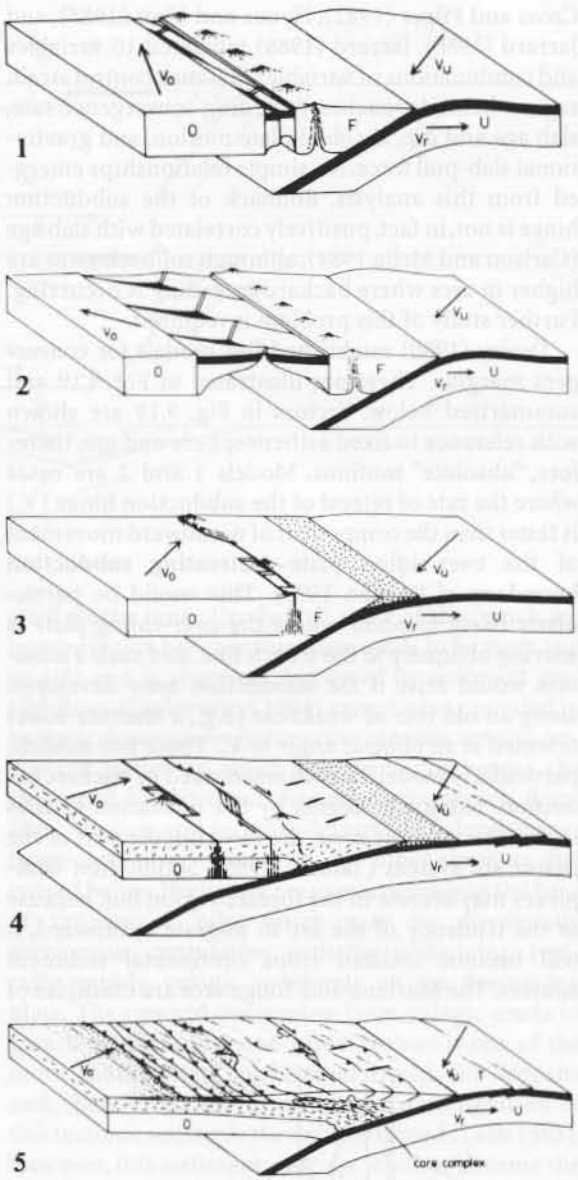


Fig. 9.19. Five models for the kinematics of a convergent margin, based on a vector analysis. V_o velocity of over-riding plate (O); V_u velocity of under-riding plate (U); V_r velocity of migration of the subduction hinge; F frontal (forearc) part of the arc. (Dewey 1980)

than the adjacent plates can converge. The subducting hinge rolls back (oceanward), carrying with it the forearc region of the over-riding plate. This leads to intra-arc and backarc spreading. An alternative model is that of Toksöz and Bird (1977) and McKenzie (1978), in which the spreading, when it occurs, is caused by secondary mantle convection induced by the downward motion of the slab. Various other parameters affect the subduction process, as described by

Cross and Pilger (1982), Hynes and Mott (1985), and Jarrard (1986). Jarrard (1986) tabulated 10 variables and combinations of variables that may control strain regime behind trenches, including convergence rate, slab age and dip, absolute plate motion, and gravitational slab-pull force. No simple relationships emerged from this analysis. Rollback of the subduction hinge is not, in fact, positively correlated with slab age (Carlson and Melia 1984), although rollback rates are higher in arcs where backarc spreading is occurring. Further study of this problem is required.

Dewey (1980) established five models for convergent margins. These are illustrated in Fig. 9.19 and summarized below. Vectors in Fig. 9.19 are shown with reference to fixed asthenosphere and are, therefore, "absolute" motions. Models 1 and 2 are cases where the rate of retreat of the subduction hinge (V_r) is faster than the component of oceanward movement of the over-riding plate (**retreating subduction boundary** of Royden 1993). This would be particularly likely to occur where the over-riding plate is moving obliquely to the trench line, and such a situation would arise if the subduction zone developed along an old line of weakness (e.g., a fracture zone) oriented at an oblique angle to V_o. These two models, particularly model 2, are characterized by backarc extension. Extension begins by the formation of rifts within the volcanic zone, the most ductile part of the island arc system (Tamaki 1985). Subduction complexes may accrete in the forearc region but, because of the tendency of the arc to migrate oceanward, it will become isolated from continental sediment sources. The Mariana and Tonga arcs are examples of these **extensional arcs**. The Bonin arc illustrates the beginning of the extension process. There, a system of rifts is developing along the volcanic arc and is expected to provide the locus for backarc spreading in the future (Tamaki 1985).

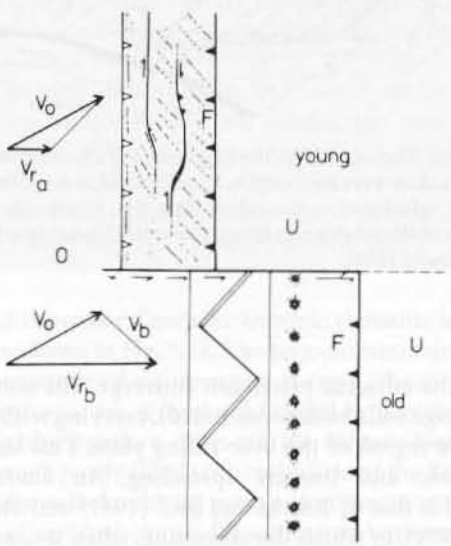


Fig. 9.20. Segmentation of a subduction zone where young and old oceanic crust are juxtaposed along a fracture zone. (Dewey 1980)

Model 3 is that of a **neutral arc**, where oceanward movement of the over-riding plate is balanced by the oceanward retreat of the subduction hinge. A strike-slip fault may be present along the arc (where the crust is weakest) to accommodate oblique convergence (Fitch 1972), but backarc spreading does not occur except where this fault is offset. The Alaska–Aleutian and Sumatra arcs are of this type and are characterized by well-developed subduction complexes. The Andaman Sea is an example of interarc spreading developed as a result of transform offset in a neutral arc.

Models 4 and 5 are **contractional arcs** and are thought to occur where the subducting crust is young. The hinge-line rollback (V_r) is, therefore, slow and the forearc region is under active contraction (**advancing subduction boundary** of Royden 1993). Tectonic erosion of the over-riding plate may occur in the subduction zone, and some of the shear stress is accommodated by thrusting in the oceanic plate. In model 4, a discrete forearc region is maintained, and oblique convergence gives rise to strike-slip movement within the arc. In model 5, the whole frontal region of the overriding plate is involved in contractional tectonics. Cordilleran-type nappes and foreland A-subduction occur. Peru and the Canadian–western-US Cordillera fit this model at certain stages of their development.

A convergent margin may show variations in style along strike if the age of the crust varies. For example, the subduction zone will be segmented where it intersects a fracture zone (Fig. 9.20). Subduction of older crust on one side of the fracture may develop an extensional margin with backarc spreading, whereas the arc facing younger crust may be neutral or contractional. This model illustrates that backarc spreading can be terminated along strike, particularly at a cross-cutting transform fault. This, in fact, occurs at several places in the western Pacific (Hamilton 1979; Dewey 1980). Other variations in style are caused by subduction of aseismic ridges or seamounts and by differences in the volume of sediment fed to the trench (Cross and Pilger 1982).

Dewey (1980) showed that arc behavior may also vary with time at a given convergent margin as the age of the subducted crust becomes progressively older or younger. The latter case is illustrated in Fig. 9.21. This adds much useful new detail to the concept of the Wilson cycle of opening and closing oceans. It demonstrates how backarc basins may open and close and accounts for the episodic nature of sub-

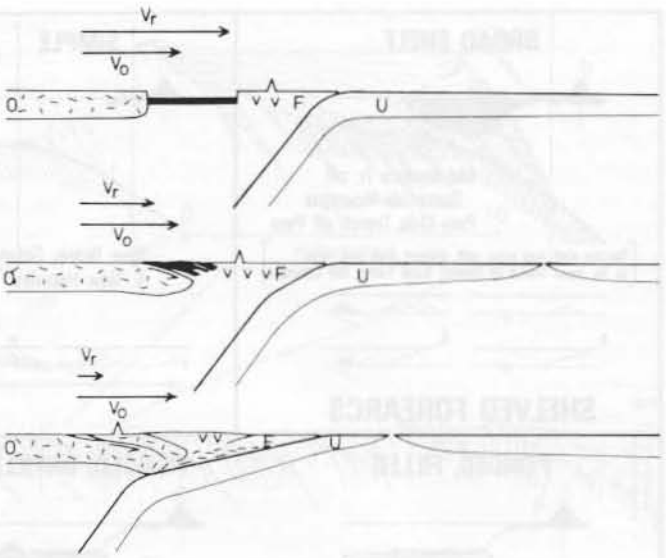


Fig. 9.21. Possible sequence of events during a Wilson cycle as progressively younger oceanic crust is subducted. A backarc basin is first opened and then closed. (Dewey 1980)

duction–accretion or orogenic activity in the foreland. There is not space to dwell here on the parallel implications for volcanism and plutonism. Busby et al. (1998) briefly described the evolution of the convergent margin constituting Baja California, which they interpret as an arc that has evolved from extensional to contractional largely as a result of the decreasing age of the subducting oceanic lithosphere. With this introduction to the complexity of arc tectonics, we shall now examine the main types of convergent-margin basin listed in Table 9.1.

9.3.2.2 *Trenches and Subduction Complexes*

In complex accreted terranes or megasutures, the identification of subduction complexes is one of the surest indicators of the previous existence of a convergent margin, as their structural and stratigraphic character are unique. If the arc magmatic belt can also be recognized and the two proved to be contemporaneous, the polarity of the subduction zone is then known. The existence of paired metamorphic belts (high *P*–low *T* and low *P*–high *T*) in arcs is also a well-known criterion (Miyashiro 1980).

In general, the sedimentary characteristics of trenches and trench inner slopes are as follows (Underwood and Bachman 1982; Underwood and Moore 1995). The oldest sediments in a trench are pelagic muds formed on the oceanic plate seaward of the trench and rafted into the trench by convergent plate movements. These may be overlain by a greater or lesser thickness of axially transported trench turbidites and other mass-flow deposits, their thickness and facies depending primarily on proximity to a terrigenous sediment source. Most sediments that are deposited on the inner (landward) slope of the trench are hemipelagics, because this slope tends to be steep (up to 11°) and is, therefore, bypassed by sediment-gravity flows (Underwood 1984) except where ponded in tectonic depressions. Submarine canyons are also important routes by which sediment can bypass the slope and be delivered directly to the trench floor. Mass wasting and tectonic deformation develop melanges and olistostromes on the inner slope, as discussed below. High pore pressures develop at the base of the trench pile, which may be dominantly arenaceous, contrasting with the underlying, typically muddy, pelagic sediments on the downgoing plate. The upward coarsening from pelagic muds to trench-fill turbidites and olistostromes is one of the more characteristic features of trench-fill deposits and, therefore, a useful criterion for recognition of this tectonic setting in the ancient record (Lash 1985). However, this sediment contrast tends to become the décollement at the leading edge of the subduction complex, so that the pelagics tend to be subducted, whereas the trench turbidites are, to a greater or lesser degree, accreted to the subduction complex (Karig 1982). Commonly, the downgoing plate is cut by extensional faults developed by flexure in the subduction hinge. These may have structural relief of more than 500 m. Graben act as sediment traps, whereas horsts may serve as rigid protuberances, tectonically eroding the underside of the upper plate or, alternatively, becoming sheared off, so that slivers of oceanic crust and overlying pelagic sediments become incorporated into the subduction complex (Karig 1982; Hilde 1983).

Dickinson and Seely (1979) identified "three distinct structural styles [that] may be formed as discrete belts or slabs within the subduction complex":

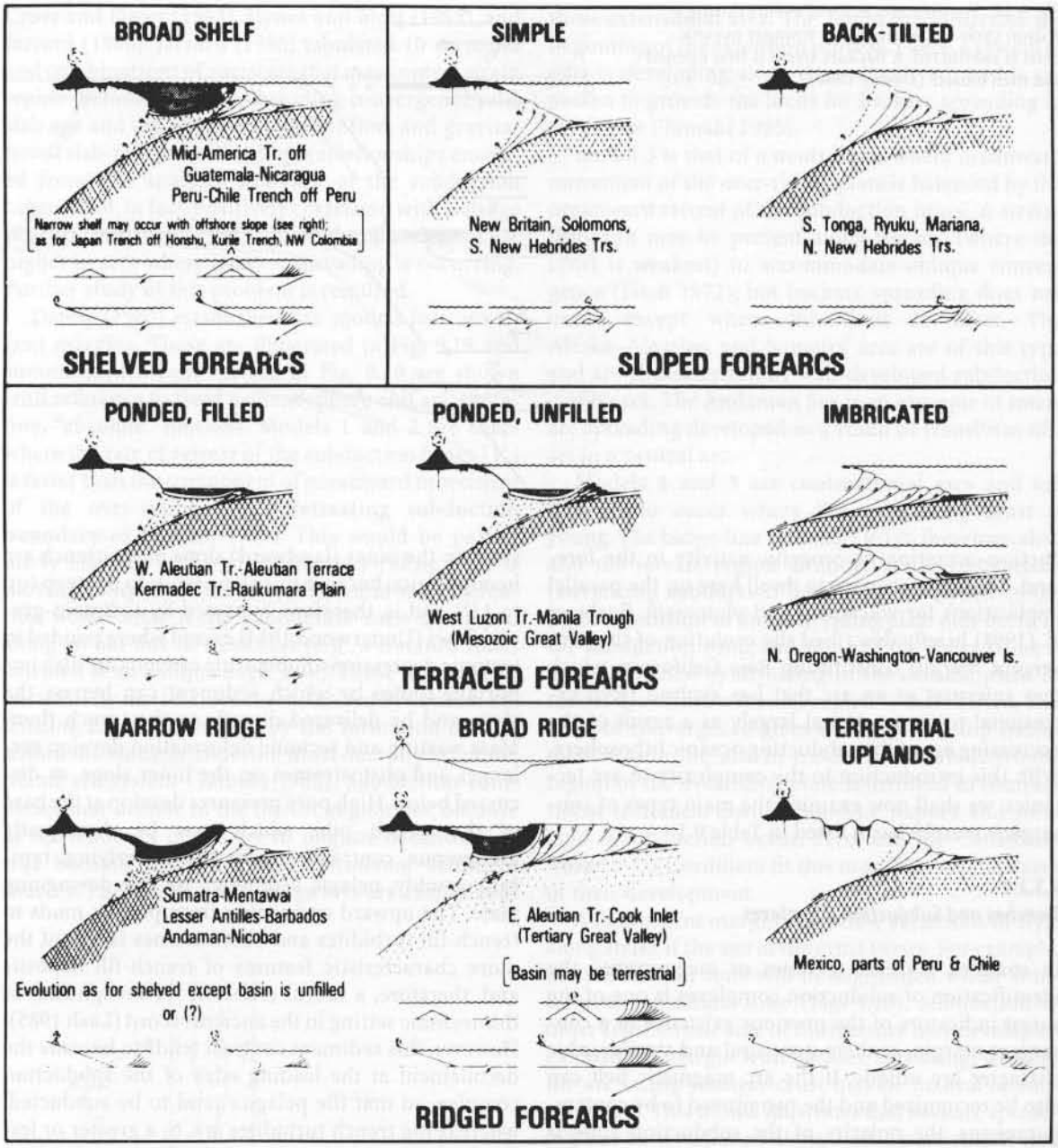


Fig. 9.22. Configuration of modern forearc regions. (Dickinson and Seely 1979)

1. Bedded sequences, which may be isoclinally folded, in which stratification is largely preserved and metamorphic fabrics are not widely developed, although metamorphic minerals may largely replace original constituents
2. Metamorphic tectonites, schistose or semischistose, in which a pervasive metamorphic fabric that largely supplants original textures exerts the dominant control on structural relations
3. Chaotic **melange**, in which pervasive mesoscopic shear fractures have disrupted the original rock into isolated tectonic inclusions immersed in a sheared matrix

Important constituents in any of these types may include slivers of oceanic crust (**ophiolites**) or giant slide masses (**olistostromes**). Rapid rates of sedimentation in the trench are recorded in areas that have large accretionary prisms, but the relationship is not a linear one (Karig 1982). Some variations in the relative scale and style of subduction complexes and other forearc elements are shown in Fig. 9.22.

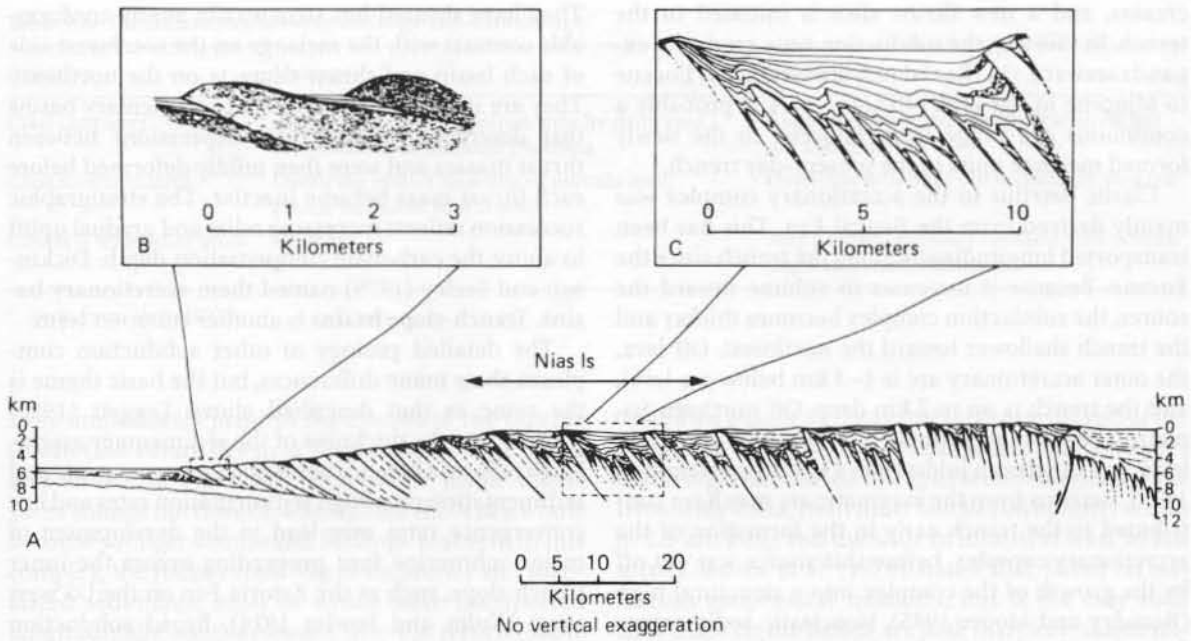


Fig. 9.23. Seismic profiles across the subduction complex of Sunda arc. This section is entirely below sea level, but the approximate along-strike position of Nias Island is shown in *profile A*. (Moore and Karig 1976; Karig et al. 1978)

Some subduction complexes are dominated by folded, bedded sequences. Modern examples include the Shikoku subduction zone off southwestern Japan (J.C. Moore and Karig 1976) and the Aleutian Kodiak–Kenai subduction complex of Alaska (Moore and Wheeler 1978; Von Huene 1978; Fisher and Von Huene 1980). An ancient example is the Ordovician–Silurian subduction complex of the Southern Uplands, Scotland (Leggett 1980). Other complexes consist largely of chaotic melanges, such as the Mentawai Islands complex of the Sunda arc off Sumatra (Karig et al. 1978; G.F. Moore and Karig 1980) and much of the Jurassic–Paleogene Franciscan Complex of California.

It is not yet clear whether Dewey's (1980) arc models can be extended to embrace differences in the style of the subduction complex. G.F. Moore and Karig (1980) suggested that pervasive shearing results from rapid dewatering under high stress. In the Shikoku subduction zone, folded sediments in the lower part of the subduction complex are overconsolidated as a result of tectonic dewatering (J.C. Moore and Karig 1976). Hamilton (1979) reported the existence of active mud volcanoes on top of several Indonesian subduction complexes, attesting to rapid dewatering. Differences in the intensity of internal deformation, from moderate (recumbent or isoclinal folding, and thrust faulting) to intense (shearing and production of chaotic melange), can possibly be related to the rate of dewatering, and this to arc style, be it extensional, neutral, or compressional.

The Sunda arc off Sumatra has long served as a classic illustration of an island arc because of the work there in the 1930s and 1940s, mainly by Dutch geologists (van Bemmelen, Umbgrove, and Vening Meinesz). Detailed geological and geophysical studies by Moore and Karig (1976, 1980) and Karig et al. (1978) provided a detailed picture of the unique geological style of an outer accretionary arc. Figure 9.23 is an interpreted seismic reflection profile showing the position of Nias Island, one of the Mentawai Islands, which owe their existence to the uplift of the subduction complex. The structure is dominated by imbricate thrust faults developed in a melange of pervasively sheared sedimentary-rock debris. The matrix consists of comminuted sandstone and siltstone plus sheared mudstone, and contains variously shaped blocks from millimeters to tens of meters in length. Some inclusions are fractured or have slickensides on the surface. The included blocks consist mainly of turbidite sandstone and siltstone, together with mudstone, rare conglomerate, clasts of chert, red shale, and limestone. Up to 20% of the inclusions are basalt fragments, including pillow lava and, very rarely, other ophiolite fragments, such as peridotite, dunite, and serpentinite.

The cross section shows that the thrust slices steepen progressively away from the trench. It is now generally accepted that this is the result of continuous underthrusting by the oceanic plate, which tilts the thrust slices landward. As tectonic compaction increases, the resistance of the rocks to shearing in-

creases, and a new thrust slice is initiated in the trench. In this way, the subduction zone gradually expands seaward. On Nias Island, the melange is Eocene to Miocene in age and, offshore, there is probably a continuous age range from Miocene to the newly formed melange units in the present-day trench.

Clastic detritus in the accretionary complex was mainly derived from the Bengal Fan. This has been transported longitudinally along the trench since the Eocene. Because it increases in volume toward the source, the subduction complex becomes thicker and the trench shallower toward the northwest. Off Java, the outer accretionary arc is 1–3 km below sea level, and the trench is up to 7 km deep. Off northern Sumatra, the outer accretionary arc is partly above sea level, and the trench is less than 4 km deep (Hamilton 1979). Detritus from the magmatic arc may have contributed to the trench early in the formation of the accretionary complex, before this source was cut off by the growth of the complex into a structural high (Beaudry and Moore 1985). Nonclastic sedimentary fragments in the complex (chert, red shale, limestone) are pelagic sediments that covered the oceanic crust before it was carried like a conveyor belt into the subduction zone. The oceanic crust itself is represented by the mafic and ultramafic rock fragments.

As shown in Fig. 9.23, resting on and folded into the accretionary complex are smaller sedimentary basins a few kilometers across. On Nias Island, these are Miocene to Pliocene in age and consist of coarsening-upward sequences 3 to 5 km thick. At the base are thin-bedded, carbonate-poor siltstones and shales. These are overlain by thin-bedded calcareous and terrigenous turbidites and then by calcareous and terrigenous sandstones and conglomerates. The beds are folded into tight, asymmetric, chevron folds. They have sheared but structurally nearly conformable contacts with the melange on the southwest side of each basin and thrust contacts on the northeast. They are interpreted as the fill of sedimentary basins that developed as structural depressions between thrust masses and were then mildly deformed before each thrust mass became inactive. The stratigraphic succession reflects increasing relief and gradual uplift to above the carbonate compensation depth. Dickinson and Seeley (1979) named them **accretionary basins**. **Trench-slope basins** is another common term.

The detailed geology of other subduction complexes show many differences, but the basic theme is the same as that described above. Leggett (1980) showed that the thickness of the sedimentary assemblage reflects a balance between convergence rate and sedimentation rate. High sedimentation rates and low convergence rates may lead to the development of major submarine fans prograding across the inner trench slope, such as the Astoria Fan on the US west coast (Kulm and Fowler 1974). Broad subduction complexes, such as that south of Kenai Peninsula, Alaska, may contain large, shallow accretionary basins. Those off Alaska are at least 100 km wide and are accumulating sediment in shelf and coastal environments (Von Huene 1978; Dickinson and Seely 1979). In Hawke Bay, New Zealand, an accretionary basin is accumulating coastal and fluvial sediments (Ballance 1980).

The analysis of the Southern Uplands of Scotland by Leggett (1980) and Leggett et al. (1982) is an excellent example of the application of a well-defined basin model to a hitherto incomprehensible piece of geology (Fig. 9.24). Following Mitchell and McKerrow (1975), this is now recognized as a subduction complex developed above a northwest-dipping Benioff

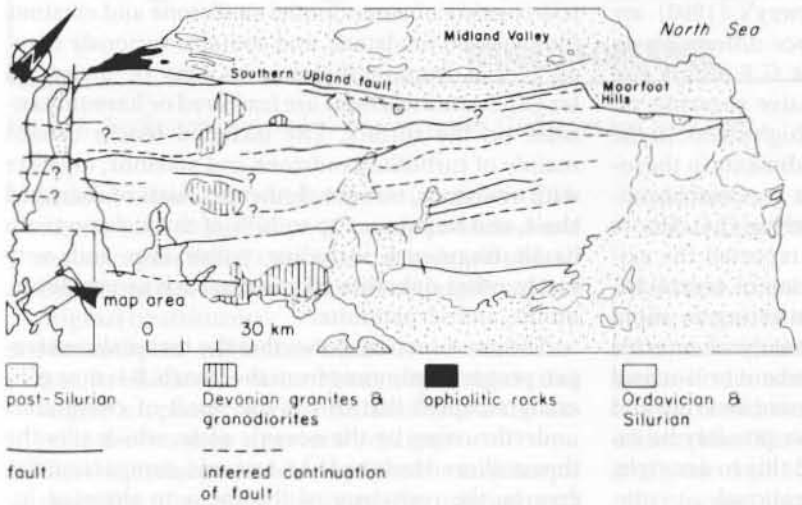


Fig. 9.24. Ordovician–Silurian outcrops of the Southern Uplands, showing the locations of the major thrust faults within the subduction complex. Thrust slices mainly dip northwest. The Iapetus suture is located approximately along the southeast edge of map. (Leggett 1980)

Table 9.3. The differences between melange and olistostrome[a]

Criterion	Melange	Olistostrome
Clast character	Angular, fractured, sheared; may be deformed into boudins or smeared phacoids	Angular to rounded, may be fractured, but not sheared
Clast source matrix	Overlying and/or underlying unit sheared; plastic intrusion into cracks	Overlying unit only, not necessarily sheared
Contacts with other units	Sheared	May have sedimentary contacts with normal slope or trench sediments

[a] Data from Hsü (1974).

zone immediately prior to the closure of the Iapetus Ocean (the suture lies near the southeast edge of the area shown in Fig. 9.24). Each of the major thrust slices youngs northward, whereas the slices as a whole become younger southward. Melange is absent in this complex, the rocks consisting of turbidites and associated sediments, most of which were transported longitudinally southwestward into the trench. Some sediment derivation from the rising accretionary arc itself can be demonstrated, just as younger (Pliocene) beds in Nias Island contain material derived from the uplifted melange. This uplift also contributed sediment northward to the forearc basin in both cases. The Southern Uplands complex contains a major slice of oceanic rocks, the Ballantrae Ophiolite, which was incorporated into the complex during an early stage of subduction (early Ordovician). Later work has introduced several complications into this analysis, such as the possibility of terrane accretion (Sect. 9.3.5) and evidence for major strike-slip displacement (Murphy and Hutton 1986; see also discussions and replies by Anderson et al. 1986), but the basic model remains useful. In other subduction complexes, such as many of those in Indonesia, ophiolitic material is more abundant, and the rocks show evidence of high-pressure, relatively low-temperature metamorphism (blueschist facies). Glaucophane, jadeite, and prehnite–pumpellyite are typical metamorphic minerals (Blake and Jones 1974, 1981; Hamilton 1979).

The term **melange** has two meanings, and it is important to distinguish them. As used here, it implies a pervasively sheared and fractured rock mass, but some authors extend the term to include ill-sorted deposits that have been derived by submarine slumping. The correct term for these is **olistostromes**. Hsü (1974) provided a useful discussion of the difference between tectonic melange and sedimentary olistostromes. These are summarized in Table 9.3. Olistostromes result from the failure of oversteepened slopes. Rapid sedimentation or advancing thrust sheets could trigger major slope failure, and both are typical of active subduction zones. Moore et al. (1976) described a giant olistostrome on the north end of the Sunda arc in the Bay of Bengal, which they interpret as having been caused by rapid sedimentation on the Irrawaddy delta. Individual blocks (olistoliths) are up to 2.8 km long, and the slide is, locally, at least 30 km across. Moore et al. (1976) stated that, based on numerous geophysical transects, this is the only such slide mass on the Sunda arc and, therefore, major olistostromes may not be common in subduction complexes. However, Field and Clarke (1979) showed that active continental slopes may contain numerous small slides a few hundreds of meters or less in width, and that these are beyond the resolving power of conventional seismic equipment.

Distinguishing a melange from an olistostrome in any given outcrop may not always be easy, because some rock masses could be the product of both processes. An olistostrome could be produced by the slumping of a melange from an active thrust sheet, then reincorporated in the melange by continued thrusting (Moore et al. 1976). Resolving the age of a melange may also be difficult, because it may contain only derived (recycled) fauna. An olistostrome could contain blocks that are older than the matrix, as determined by contained fossils, whereas the reverse is only possible in a melange. For example, in some Swiss melange blocks containing an early-Tertiary nummilitic, fauna are set in a matrix of shale yielding a late Cretaceous fauna (Hsü 1974). Normally, any contained fauna in a melange will only indicate a maximum age for the tectonism.

9.3.2.3 *Forearc Basins*

In transverse profile, forearc regions may be described as sloped, terraced, shelved, benched, or ridged (Dickinson 1995). Forearc sedimentation is minor on sloped forearc margins, which are characterized by a continuous slope between the volcanic arc and the trench.

Forearc basins of geologically significant size develop where the subduction complex forms a topo-

graphic ridge and sediment trap corresponding to the outer-arc high. They may be up to 100 km wide and several hundred kilometers long. Dickinson (1995) termed these terraced, shelved, or benched, depending on whether they are deep-marine, shallow-marine, or terrestrial, respectively. Dickinson and Seely (1979) subdivided forearc basins into two subtypes, **residual basins** floored by attenuated continental crust or a fragment of obducted oceanic crust, and **constructed basins** floored by the subduction complex. Distinction of the two subtypes is difficult from geophysical data and may not be possible from outcrop information in ancient, uplifted basins. In fact, Dickinson (1995) notes that, commonly, the nature of the crustal substratum cannot be determined by geophysical or other means, and he does not use the terms for the two subtypes noted above. Most basins are probably composites of the two subtypes. They show essentially the same style of structural and stratigraphic evolution and can be dealt with together.

Dickinson and Seely (1979) noted two other minor types of forearc basin. **Accretionary basins** are the relatively small basins of ponded sediment on the inner trench slope, such as are exposed on Nias Island (Fig. 9.23). These are discussed in the preceding section and are not dealt with further. The second type are extensional **intramassif** or **intra-arc** basins, which occur within the magmatic arc. These basins may be of considerable size, particularly where a broad forearc region has developed over continental crust, as in Chile. They are typically bounded by normal faults, the orientation of which is not necessarily parallel to arc trends but may be determined by pre-existing basement grain. Such is the case in Chile, where bounding faults and the alignment of andesitic arc eruptions are oriented either northwest or northeast, reflecting pre-existing Gondwana architecture (Katz 1971). The Osorno Basin is one of the largest of these basins. It is over 200 km long and, locally, contains 1100 m of Tertiary strata or 1300 m of Quaternary sediments, although not in the same location because of differential movement. The strata are mainly nonmarine, deposited by rivers whose alignment is determined by the same structural grain. However, repeated block movements related to subduction have caused local facies changes, and some of the strata are shallow marine (Katz 1971).

Forearc basins tend to become wider and shallower with time. Initially, the forearc region may consist of a simple seaward slope, depending on the location of the arc (old oceanic fracture zone, continental margin, etc.). Sediments are carried directly (transversely) into the trench and feed a growing subduction complex. As ponding commences, water depths decrease from abyssal to shallow, and environments may become nonmarine. At the same time, the basin becomes wider as the subduction complex advances oceanward. The forearc fill onlaps the magmatic arc, although the contact may be faulted and, on the other side of the basin, has a sedimentary or partly sheared contact with the subduction complex. The magmatic arc may tend to retreat landward, particularly if the angle of subduction decreases over progressively younger crust. These processes have been well documented in the Great Valley of California, a late Jurassic to early-Tertiary forearc basin (Ingersoll 1979; Moxon and Graham 1987; Sect. 7.9).

Basal sediments in a residual forearc basin are invariably deep marine, consisting of pelagic trench deposits resting on ophiolite. Constructed basins may be somewhat shallower. In both cases, the first major depositional systems are submarine fans that build transversely across the basin from the magmatic arc. In the case of intraoceanic arcs, which are largely submerged (e.g., Mariana arc), the derived deposits will be entirely deep marine, whereas arcs built on continental crust may contain a significant proximal shallow-marine to nonmarine component (e.g., the west coast of Oregon and Washington). In general, the submarine fans show a gross upward coarsening, indicating overall progradation but, in detail, both coarsening- and fining-upward trends are present as the fans switch position because of the interplay of active tectonics, sea-level change, and sedimentation (Ingersoll 1978a, 1979). Sediment thicknesses are, typically, considerable. The modern Sumatra trough (Figs. 9.25, 9.26) contains at least 6 km of basin fill (Karig et al. 1978), whereas the Great Valley Group is about 15 km thick (Ingersoll and Dickinson 1981). Sedimentation rates in the latter ranged from about 0.2 to 0.4 m/10^3 years.

The growth of the outer accretionary arc gradually delineates a seaward margin to the basin, and turbidity currents are deflected by the rising topographic barrier, turning 90° at the fringe of the submarine fan wedge to flow longitudinally along the basin axis. This is well documented in the Great Valley Group (Ingersoll 1979). The outer arc may itself become a sediment source, as occurred in the Great Valley during the Oligocene and in the Silurian forearc basin north of the Southern Uplands Fault, Scotland (Leggett 1980).

The youngest sediments in a forearc basin may be shallow marine or nonmarine. For example, in the Great Valley, the Cretaceous–Tertiary transition was marked by the progradation of deltas in the Sacramento Valley whereas, in the adjacent San Joaquin Valley, turbidite sedimentation continued until the end of the Cretaceous (Ingersoll and Dickinson 1981). This was followed by the deposition of an extensive nonmarine succession in the Cenozoic (Bartow 1988).

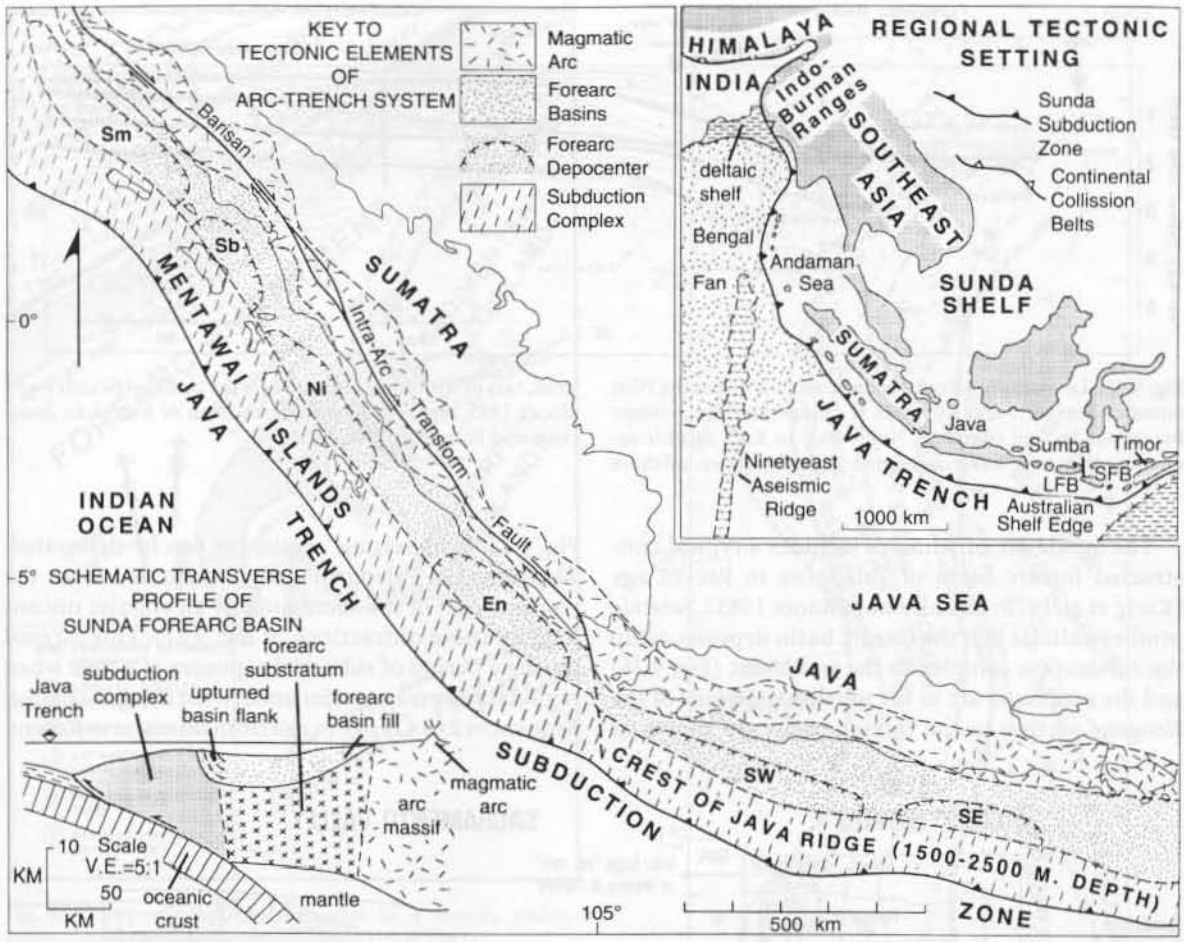


Fig. 9.25. Sunda Arc of Sumatra and Java, showing the position of the Sumatra–Java forearc basin. Forearc depocenters include: *Sm* Simeulue; *Ni* Nias; *Sb* Siberut; *En* Enggano; *SW* southwest Java; *SE* southeast Java. *Inset at upper right* shows the regional setting. *LFB* Lombok forearc basin; *SFB* Savu forearc basin. (Dickinson 1995)

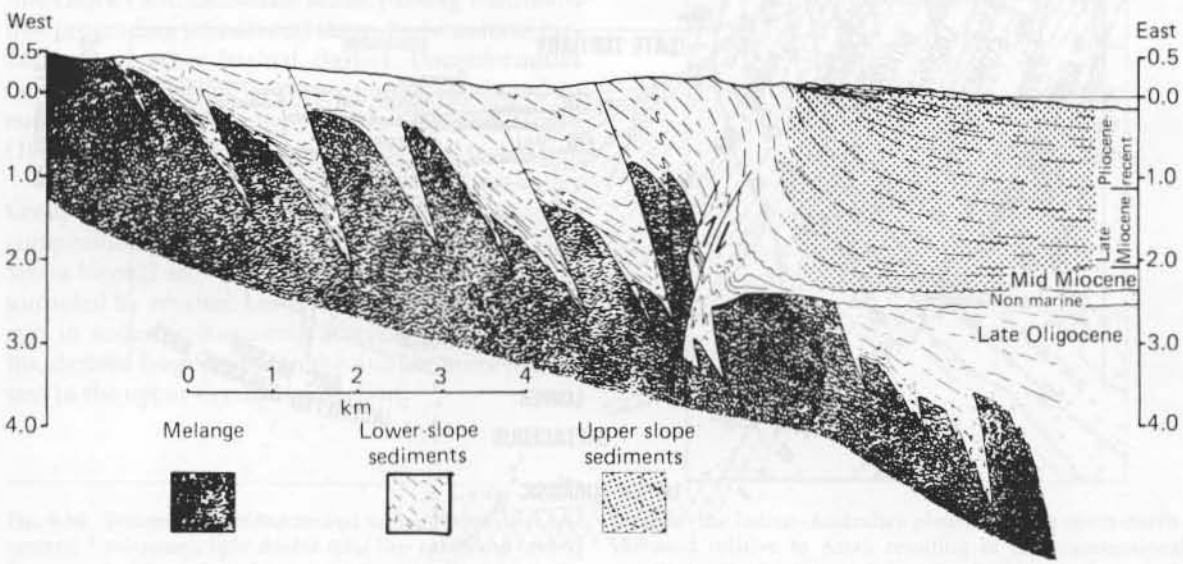


Fig. 9.26. Drawing from a seismic-reflection profile across the Sunda forearc basin near Nias Island, showing the relationship between the forearc basin and the subduction complex. (Karig et al. 1978)

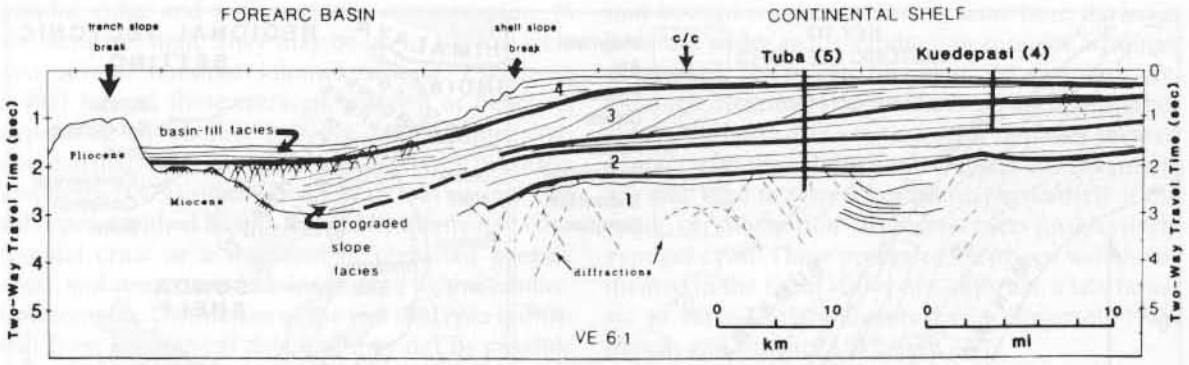


Fig. 9.27. Line drawing from a seismic section across the West Sumatra Forearc basin to where it onlaps the trench–slope break (subduction complex). Numbers *1* to *4* are seismic sequences that have been correlated with data from offshore wells, two of which are indicated on the profile. (Beaudry and Moore 1985, reproduced with permission of American Association of Petroleum Geologists)

The Sunda arc of Sumatra includes a typical constructed forearc basin of Paleogene to Recent age (Karig et al. 1978; Beaudry and Moore 1985). Seismic profiles indicate that the forearc basin deposits onlap the subduction complex to the southwest (Fig. 9.26) and the magmatic arc to the northeast. Details of the Neogene section in the forearc basin are shown in Fig. 9.27. Four seismic sequences can be delineated. The lowest is Paleogene and is separated from the younger part of the succession by an angular unconformity (note diffractions in Fig. 9.27). This formed during a period of subaerial exposure at a time when regional plate trajectories underwent a major change. Sequences 2 to 4 range in age from Miocene to Recent.

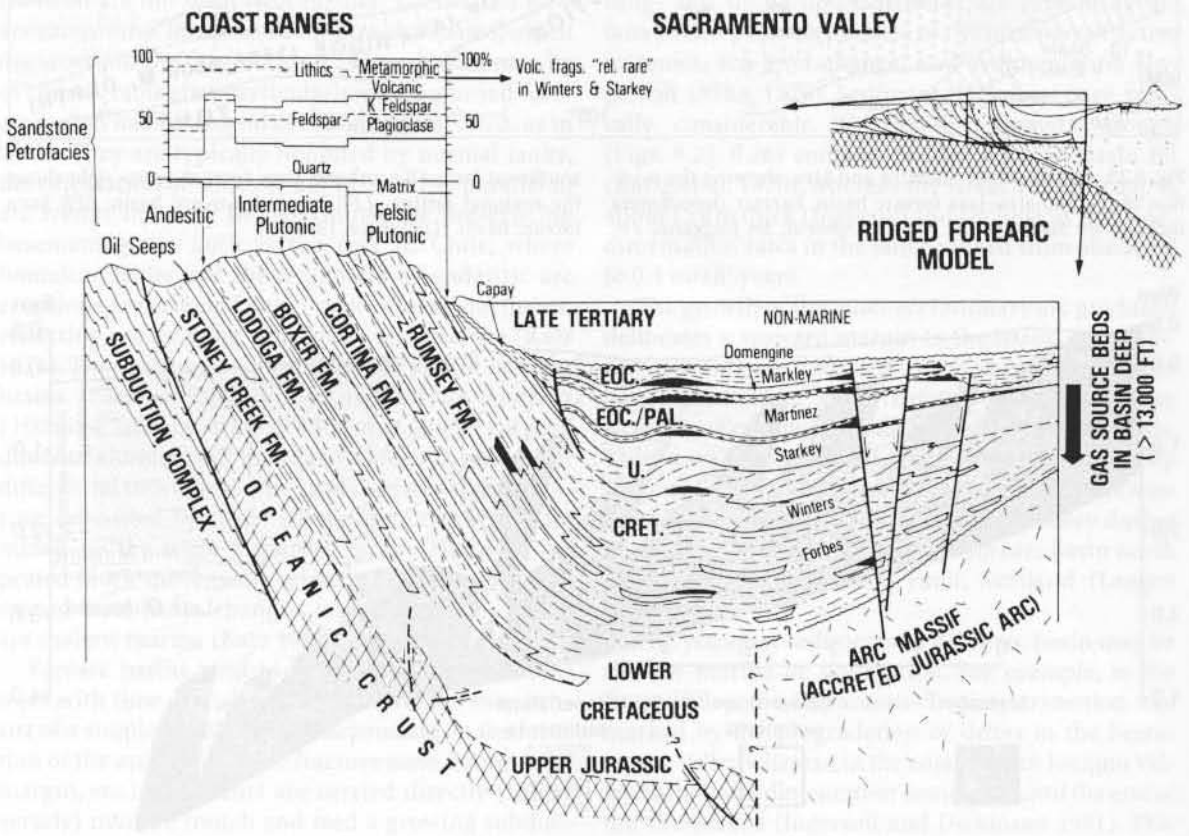


Fig. 9.28. Schematic cross section through the Great Valley Sequence in the Sacramento Valley, California, showing the variation in petrofacies through the Cretaceous section. (Graham 1981)

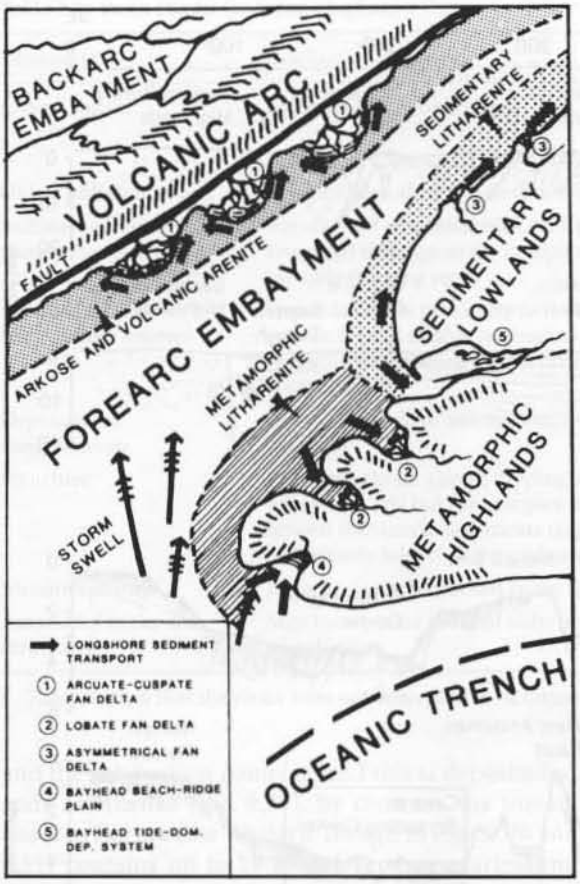


Fig. 9.29. Depositional environments in a forearc embayment, Cook Inlet, Alaska. (Hayes and Michel 1982)

They comprise gently dipping shelf successions of fine clastics and carbonate muds, passing basinward into prograding (clinoform) slope–basin units of turbidites and other basinal clastics. Unconformities between the seismic sequences were attributed to eustatic sea-level changes by Beaudry and Moore (1985) but could also be the result of local tectonism.

A schematic cross section through the Great Valley Group of California is shown in Fig. 9.28. Sandstone composition changes through the section, as the Sierra Nevada arc – the major sediment source – was unroofed by erosion. Lower Cretaceous deposits are rich in andesitic fragments, whereas plutonic detritus, derived from deeper in the arc, are more important in the upper Cretaceous section.

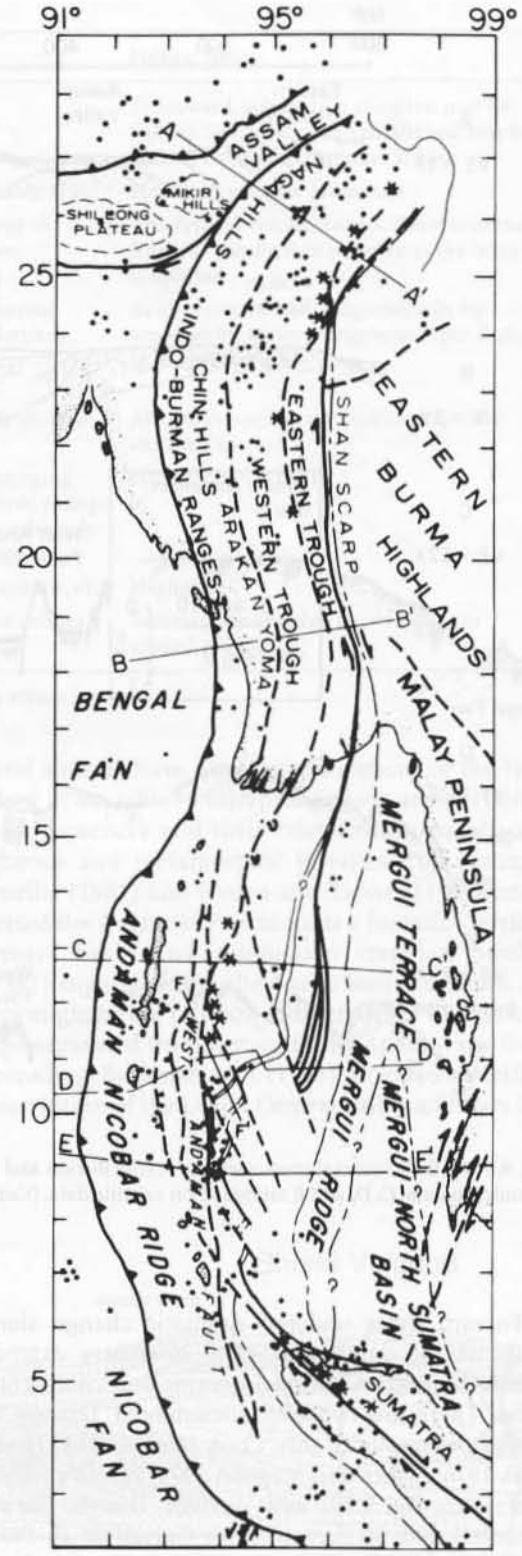


Fig. 9.30. Tectonic map of Burma and Andaman Sea. *dots* epicenters; * volcanoes; *light double solid line* spreading center; *heavy dashed line* edge of continental crust; *lines A–A'* etc. lines of the cross sections in Fig. 9.31. Subduction at the trench extending from the Indo-Burman Ranges southward is highly oblique (the Indian–Australian plate is moving north-northeastward relative to Asia), resulting in the transtensional opening of the interarc Andaman Sea basin; note the spreading center with transform offsets between latitudes 7° and 11°. (Curray et al. 1978)

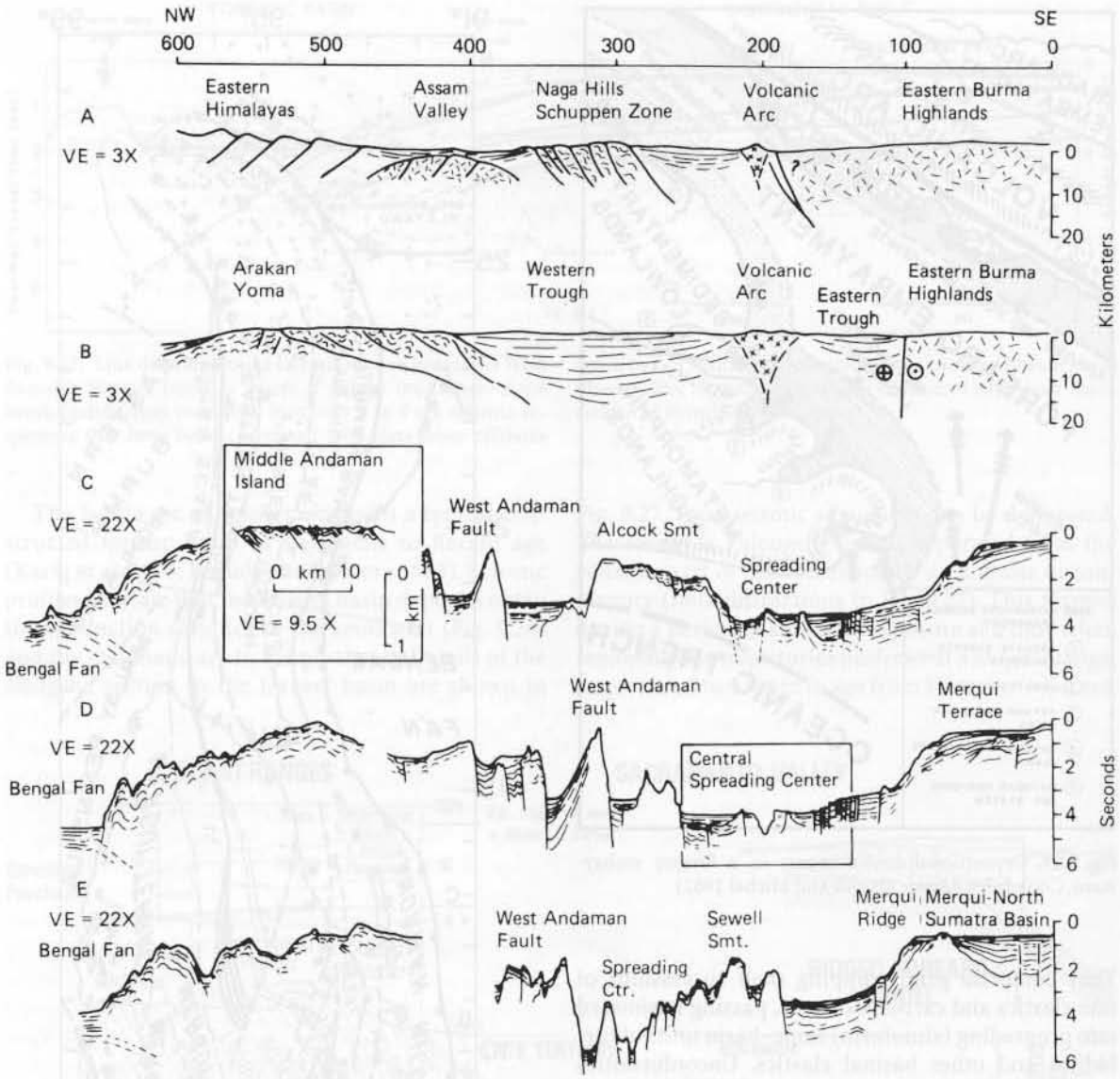


Fig. 9.31 A–E. Structural cross sections across Burma and Andaman Sea. Locations are shown in Fig. 9.30. **A** and **B** are based on outcrop data; **C, D,** and **E** are based on seismic data. (Compiled by Curray et al. 1978)

Forearc styles undergo dramatic change along strike if the convergent plate boundary extends across the contact between oceanic and continental crust. There are two well-documented Tertiary to modern examples of this: Cook Inlet, Alaska (Hayes et al. 1976; Fisher and Magoon 1978; Hayes and Michel 1982; Fig. 9.29) and Western Trough, Burma (Mitchell and McKerrow 1975; Curray et al. 1978; Bannert and Helmcke 1981; Figs. 9.30, 9.31). In both cases, the subduction complex increases considerably in thickness along strike toward land and has become uplifted to form a major landmass. The availability of a greater volume of sediment to fill the trench is the main reason for this. Second, the continental part of the forearc basin has become a major alluvial valley receiving sediment both from the magmatic arc and the subduction complex. Rivers drain transversely into the basin from both sides, feeding a longitudinal river that empties into the marine part of the forearc basin along strike.

In the case of Cook Inlet, although fluvial sedimentation has probably been continuously active since the Tertiary (Hayes et al. 1976), sedimentation has not outpaced subsidence, and the inlet remains a marine embayment (Fig. 9.29). At present, it is an estuary receiving sandy sediment from both the arc

Table 9.4. Basin model for forearc regions

Parameter	Subduction complex	Forearc basin
Lower contact	Ophiolites; may be sheared	To seaward, subduction complex; may be faulted. To landward, magmatic arc; may be faulted. Ophiolites common at base
Oldest sediments	Pelagic chert, abyssal limestone; on oceanic crust	Similar to younger sediments
Sediments from transverse sources	Arc-derived volcaniclastics, may give way to reworked melange as the complex grows; possible fringing reefs	Arc-derived volcaniclastics. Some material from the subduction complex as the latter is uplifted
Sediments from longitudinal sources	Depends on the proximity to the continental source. Fluvial and/or submarine-fan detritus may give rise to thick trench fill and large complexes	As above, redirected longitudinally by topography. Major terrigenous input if close to the continent
Depositional environments	Abyssal oceanic to nonmarine; typically shallows upward	Abyssal oceanic to nonmarine; typically shallows upward
Structure[a]	Imbricate thrust sheets, dipping and younging toward the arc, but the complex as a whole youngs toward the trench. Sediments may be moderately to intensely folded or sheared to melange	Simple, open syncline
Metamorphism[a]	High pressure; blueschist facies (glaucophane, etc.)	Negligible
Associated rocks and structures	May incorporate sheared slabs or thrust masses of ophiolite	Sediments may onlap arc volcanics or eroded plutons

[a] This assumes that the rocks have not been further deformed in a suture zone.

and the subduction complex, and this is deposited as spits and deltas (Fig. 9.29). By contrast, the forearc basin in Burma (the Western Trough in Figs. 9.30 and 9.31) contains up to 17 km of Tertiary marine and fluvial sediments derived from Burman and Tibetan highlands to the north and east and the Indo-Burman ranges (subduction complex) to the west. The Irrawaddy River flows southward along the axis of the basin and has built a major, tide-dominated delta over the continental margin.

An attempt at a summary basin model for forearc regions is presented in Table 9.4. In recent years, several authors have used generalizations of the type given in the table to interpret various ancient deformed sequences and their relationships to adjacent igneous and metamorphic terranes. For example, Smellie (1981) and Hyden and Tanner (1981) interpreted the Antarctic Peninsula as a Jurassic–Tertiary forearc basin and subduction complex. Smellie (1981) suggested that the Karoo Basin of South Africa might be the retroarc counterpart of this system, now separated from Antarctica by Atlantic sea-floor spreading. Bachman et al. (1983) provided a detailed description of the Luzon Central Valley, a forearc ba-

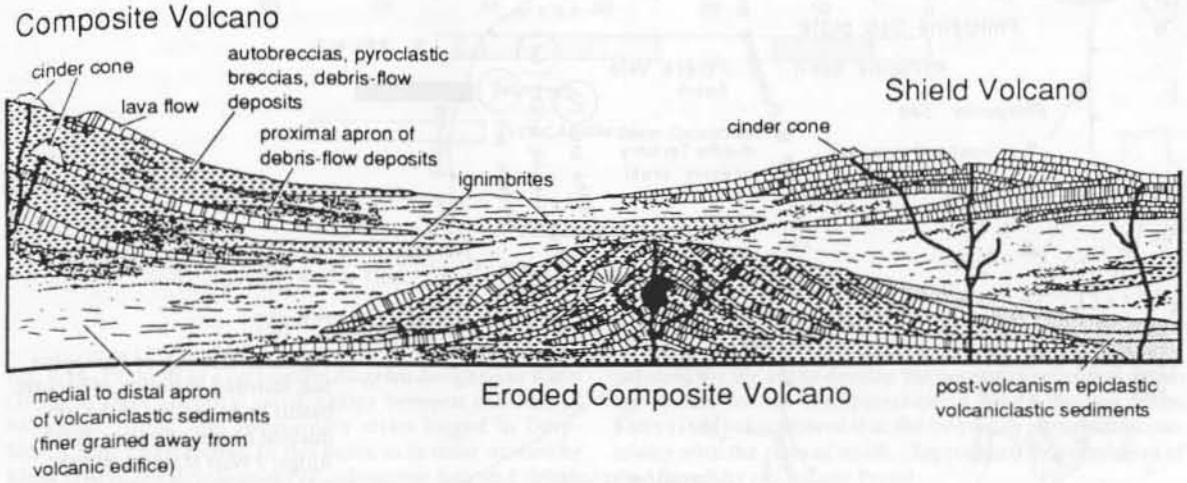


Fig. 9.32. Schematic illustration of facies relationships in a subaerial intra-arc setting. (Smith and Landis 1995)

sin showing similarities to the Great Valley of California.

9.3.2.4 *Intra-Arc Basins*

Volcanic arcs may be up to 200 km wide and may include depressions bounded partly by volcanoes and partly by faults. Depressions may be generated by the isostatic response to the mass of the arc and also by magma withdrawal and caldera generation and by extensional faulting. The resulting basins fill with volcanic flows and volcaniclastic debris that typically are deposited in nonmarine settings. The stratigraphic evolution of these basins typically is complex, reflecting the complex volcanic and structural history of the arc, which can include reversal of arc polarity and the passage of a given area of the arc through forearc, backarc, and interarc phases. Several areas in the Philippines have undergone such a complex evolution. Other arcs that include significant intra-arc volcaniclastic accumulations include the late-Cenozoic succession of the Cascade volcanic range in Washington and Oregon, and the Taupo volcanic zone in North Island, New Zealand.

An illustration of the typical facies relationships in intra-arc basins is given in Fig. 9.32. A thorough review of the sedimentary geology of this class of basin was provided by Smith and Landis (1995)

9.3.2.5 Interarc and Backarc Basins on Oceanic or Transitional Crust

Backarc spreading accompanies the rifting away of a magmatic arc from a continental margin in an extensional-arc setting (Dewey 1980). The duration of rifting typically ranges from 5 to 16 m.y. in most backarc

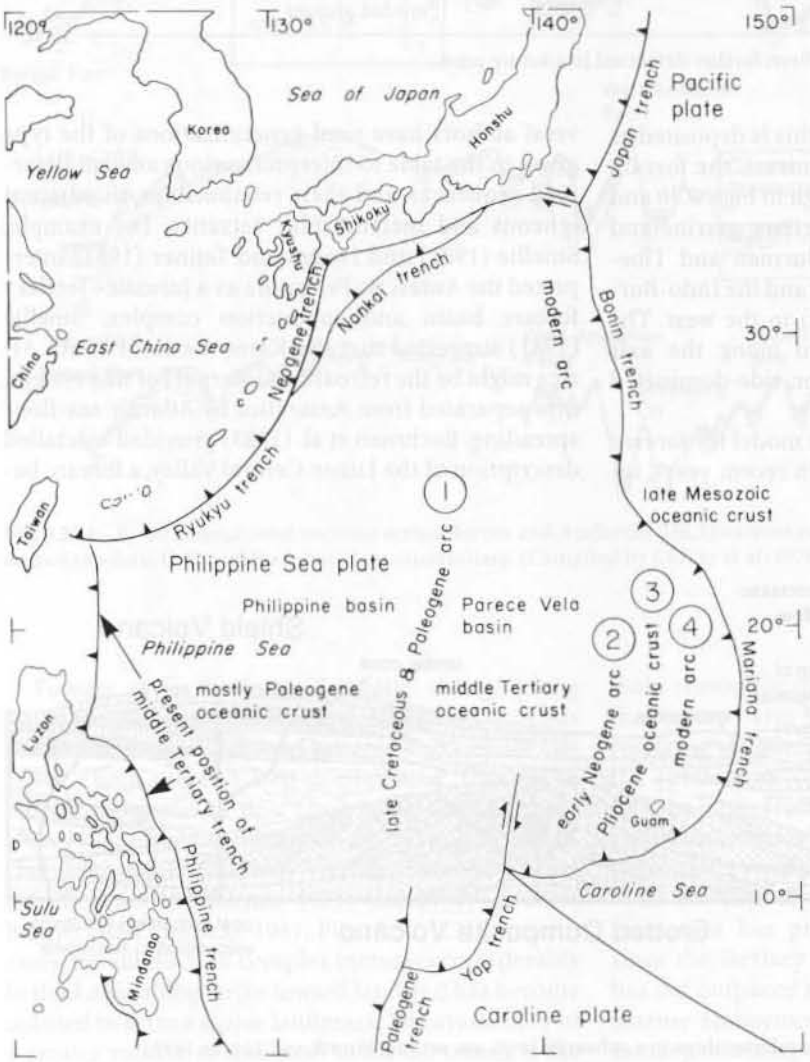


Fig. 9.33. Geology of Philippine Sea, showing evolution of backarc basins by migration of the subduction hinge. *1* Kyushu–Palau Ridge; *2* West Mariana Ridge; *3* Mariana Trough; *4* Mariana Ridge. (Hamilton 1979)

basins (Klein 1985). The resulting small ocean basins are commonly called **marginal seas**. Some backarc basins are floored by thinned continental crust or transitional crust, such as the South China Sea and the Japan Sea. Others are largely floored by oceanic crust, such as the West Philippine Basin and the Parece Vela Basin. The Okinawa Trough is a good example of an incipient backarc basin developing on continental crust (Letouzey and Kimura 1986). Useful compilations of research on volcanic, sedimentary, and tectonic processes in backarc basins have been provided by Carlson and Kobayashi (1984), Kokelaar and Howells (1984), and Marsaglia (1995).

Backarc basins may also be generated by the trapping of oceanic crust during episodes of plate reorganization. The Bering Sea and the West Philippine Sea Basin are cited as examples of this process (Marsaglia 1995). Hsü (1988, 1989) interpreted the modern Black Sea and Caspian Sea basins and the Cenozoic Tarim, Qaidam, and Junggar Basins in western China as examples of this type of basin. As noted in Section 9.3.4.8, Ingersoll and Busby proposed including these basins in their new class of **dormant ocean basins**.

In complex areas, such as the western Pacific, many basins are located between two island arcs and have been described as interarc basins. However, it is rare for there to be two active arcs with the same polarity in close proximity except in areas of complex microplate interaction. In most cases, there is only one active arc, and the basin behind this is better described as a backarc basin. Arcs undergo extension during rapid, oceanward rollback of the subduction hinge. This is accommodated by the development of rifts along the weakest part of the arc, the magma conduits, and zones of intrusion beneath the volcanic ridge. A new backarc basin is created, bounded by the active arc and an older, remnant arc (Karig 1971a, b).

The clearest example of arc migration is the Philippine Sea (Fig. 9.33), where the subduction zone has moved eastward twice since the Cretaceous according to Karig (1971a, b), Karig and Moore (1975), and Hamilton (1979). At the surface, the arc rifted along the line of volcanic activity, and the backarc region was abandoned and became inactive, while the forearc region migrated rapidly oceanward into a new, temporarily more stable position. The Kyushu–Palau Ridge was, thus, split in the mid-Tertiary, and the West Mariana Ridge was split in the Pliocene. The currently active Mariana Ridge now contains volcaniclastics of Eocene to Miocene age, indicating that the older part of the arc has indeed migrated oceanward from the Kyushu–Palau Ridge. The modern Bonin Arc contains rifts, indicating that it may be splitting in the same way (Tamaki 1985; Marsaglia 1995).

Sedimentary models for backarc basins have been discussed by Karig and Moore (1975), Carey and Sigurdsson (1984), Klein and Lee (1984), Klein (1985), and Marsaglia (1995). The depositional environments in backarc basins are deep marine, except along their margins. Stratigraphic–sedimentologic information from modern basins is, therefore, limited to seismic data, scattered DSDP holes, and piston cores. A volcaniclastic submarine-fan apron progrades into the basin from the arc and interfingers with pelagic clays. Calcareous organic debris may be important where the volcanic detritus decreases in abundance toward the mainland side of the basin. As

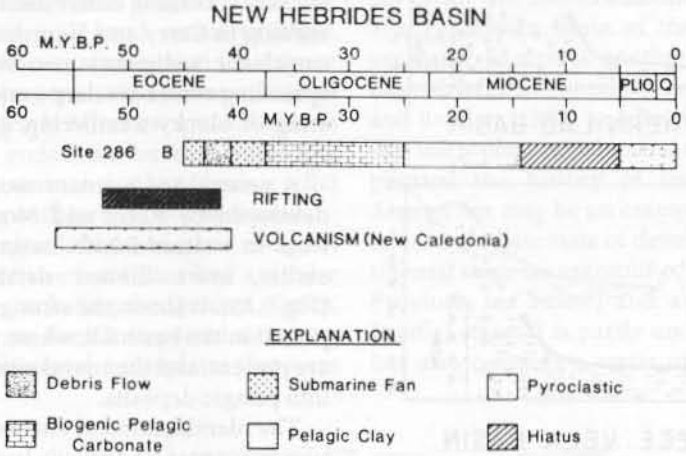


Fig. 9.34. Example of a correlation diagram designed by Klein (1985) to show temporal relationships between episodes of volcanism, rifting, and sedimentary styles logged in Deep-Sea-Drilling Project cores. In this basin, as in most studied by Klein (1985), the development of submarine fans and debris flows followed a lengthy episode of arc volcanism. Time is required for the arc to develop the necessary relief and drainage system for the transportation of debris into the basin. Klein (1984) also showed that the frequency of turbidites correlates with the rates of uplift. (Reproduced by permission of the University of Chicago Press)

the basin subsides, the floor eventually drops below the carbonate compensation depth, and younger basin fill is carbonate-free. Where the basin is developed against a continental margin, a wedge of continent-derived sediment forms, showing deep- to shallow-marine transitions in depositional environments. This should display an extensional structure and a stratigraphy similar to that of divergent continental margins.

Klein and Lee (1984) and Klein (1985) synthesized available DSDP data from backarc basins in the western Pacific and showed that, when each basin is examined in detail, the sedimentation patterns and their relationships to episodes of arc magmatism and rifting vary considerably (Fig. 9.34). They identified nine types of depositional systems: submarine fan, debris flow, silty basinal turbidite, biogenic pelagic carbonate, resedimented carbonate, biogenic pelagic silica, pyroclastics, hemipelagic and turbid-layer clay, and pelagic clay. Klein and Lee (1984) stated:

> Debris flows and submarine fan systems tend to accumulate after uplift of andesitic volcanic arc sources following development of a regional drainage network. Biogenic pelagic systems tend to be controlled by biological productivity and are indicators of back-arc basin movement through such productivity zones by regional crustal shift. Hemipelagic and turbid-layer clay systems are derived from terrigenous sources and their volume is dependent on climatically controlled sediment yield from such sources. Resedimented carbonate systems tend to occur during times of active basinal faulting.

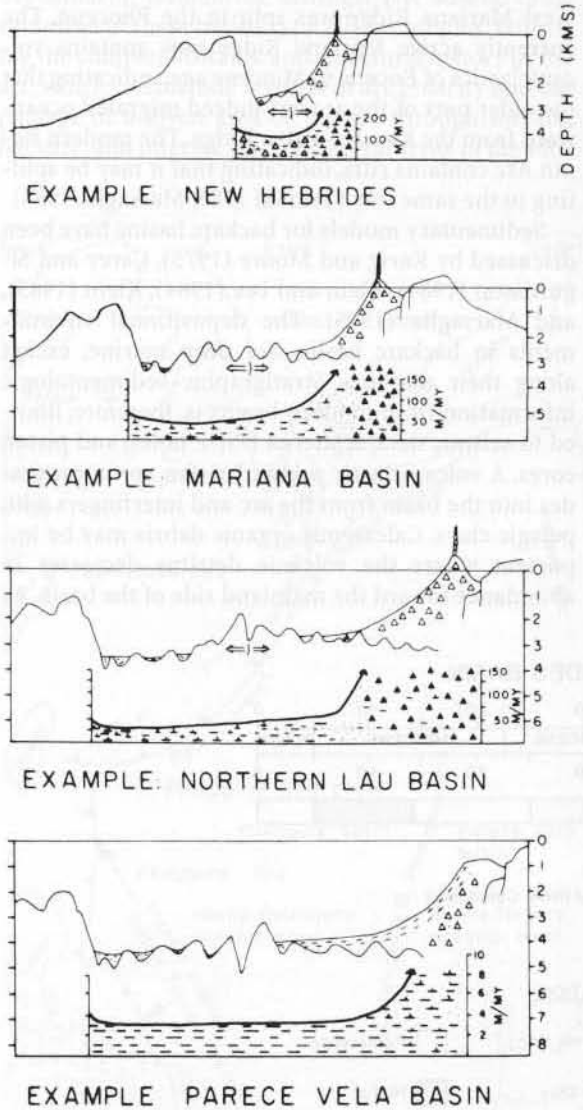


Fig. 9.35. Depositional model for the evolution of backarc basins, based on four basins in various stages of development, from oldest (*bottom*) to youngest (*top*). At the base of each section is a diagram showing sedimentation rate and lithofacies, assuming no continental sediment sources. (Karig and Moore 1975)

In modern mature backarc basins, sediment thicknesses are typically on the order of 2–3 km in the basin center, except where there is a particularly active local sediment source. Thicknesses in the range of 5–6 km may be expected in the volcaniclastic wedge, depending on the duration of active backarc spreading. The presence of multiple volcanic sediment sources in the arc and the absence of well-defined canyons commonly lead to the development of a slope-apron deposit behind the arc, rather than discrete submarine fans (Carey and Sigurdsson 1984). The contribution of volcanic and sedimentary rocks from the backarc spreading center itself is less well known. According to Carey and Sigurdsson (1984), typical volcaniclastic sediments associated with the backarc spreading center are deep-water hyaloclastites consisting of blocky-weathering, glassy, basalt sand and silt.

A general sedimentary model for backarc basins developed by Karig and Moore (1975) from their work in western Pacific basins, though based on an earlier, more limited database, remains useful (Fig. 9.35). It shows the strong asymmetry that is expected in the basin fill, where no continental sources are present, and the lateral passage away from the arc into pelagic deposits.

The identification of ancient backarc basins in suture zones may be difficult, because the elements that define them tend to be deformed (Leitch 1984). For example, an arc–continent collision would probably result in the reversal of arc polarity and the subduction of backarc oceanic crust. This topic is discussed further in Section 9.3.4. However, Hsü (1988, 1989)

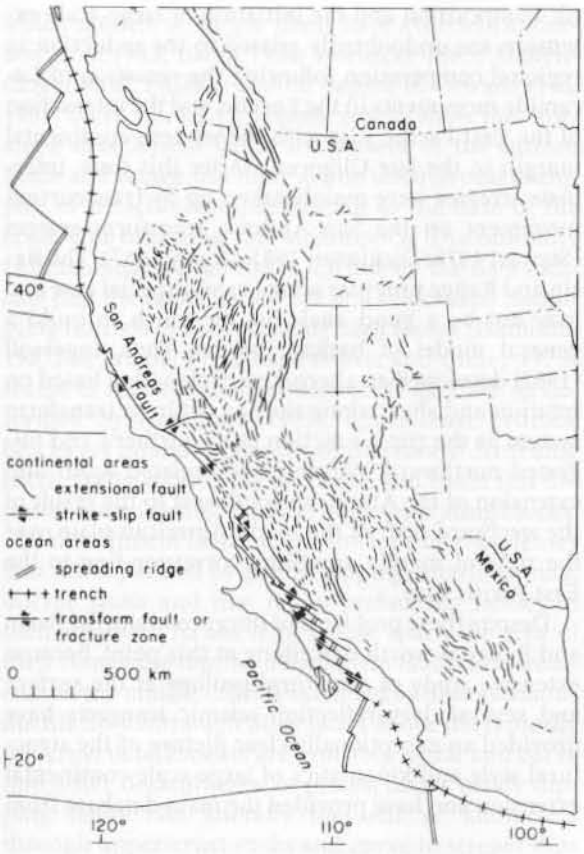


Fig. 9.36. Distribution of extensional faults in the western Cordillera, defining the area of the Great Basin. The San Andreas transform fault and other elements of the North American plate boundary are also shown. (Stewart 1978)

argued that, in certain circumstances, relict backarc basins could be preserved within orogenic belts. The South China Sea Basin is a good recent example, formed initially by continental extension (cf. Aegean Sea Basin) and with the generation of oceanic crust during the Oligocene–early Miocene. Arc collision in the middle Miocene ended the backarc spreading phase, and the basin has since that time been a relict basin, gradually filling with terrigenous sediment.

Hsü (1988, 1989) argued that several of the great intermontane basins of western China had a similar history. Most of these, including the Qaidam, Tarim, and Junggar basins, have been interpreted as having been formed over continental microplates that were trapped between orogenic belts during closure of the Tethyan Ocean (Zhang et al. 1984). Exposures of Precambrian basement and Paleozoic cratonic cover around the margins of these basins seem to confirm this hypothesis. However, Hsü (1988, 1989) suggested that the thin crust and large magnetic anomalies under these basins are more suggestive of basins floored by oceanic crust and showed that stratigraphic and facies data are consistent with this interpretation. Hsü (1988, 1989) discussed criteria for distinguishing such relict backarc basins from other types of basins present within megasutures, such as foreland and pull-apart basins.

9.3.2.6
Backarc Basins on Continental Crust

Where the arc is located on the leading edge of a continental plate, there may or may not be backarc spreading, depending on whether the arc is extensional or contractional. According to Dewey's (1980) model, this depends mainly on the age of the subducted crust (but see Sect. 9.3.2.1). Thus, the Andes do not show backarc spreading, although small, superficial, extensional intermontane basins are present, probably as a result of warping during subduction.

However, some other areas of the world show what might be termed incipient or aborted backarc spreading, which may or may not lead to full-scale extension and the development of oceanic crust. An example is the Cagayan Basin on Luzon, in the Philippines (Mathisen and Vondra 1983). In this basin, 900 m of nonmarine volcaniclastic sediment accumulated during the Plio-Pleistocene. The basin deepened and widened during sedimentation as a result of increasing relief of the arc and extensional faulting in the basin. The Pannonian Basin of the Carpathian region in southeast Europe is another example of a backarc basin floored by continental crust, although Burchfiel and Royden (1982) and Royden et al. (1983) showed that microplate collision and strike-slip faulting complicated the history of basin development. The Aegean Sea may be an example of a backarc basin in an intermediate state of development between a continental stage (as exemplified by the Basin and Range Province, see below) and a fully oceanic (western Pacific) stage. It is partly underlain by oceanic crust but also contains a series of elongate graben filled

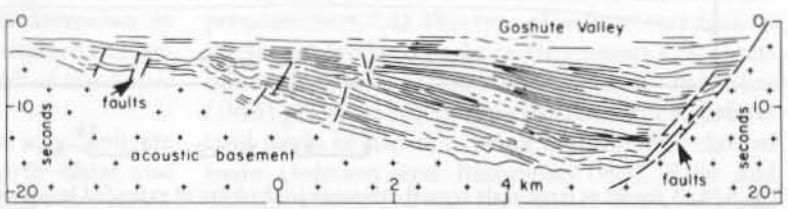


Fig. 9.37. Seismic cross section through a typical fault-bounded basin in the Great Basin, Goshute Valley, Nevada. (Bally and Snelson 1980)

with 600–1000 m of nonmarine to shallow-marine Pliocene to Quaternary clastic sediment (Biju-Duval et al. 1978; McKenzie 1978).

A particularly well-studied example of a backarc basin on continental crust is the Basin and Range province of the western United States and northern Mexico (Fig. 9.36). This area consists of a series of tilted fault blocks and half-graben basins with up to 3 km of nonmarine, upper-Cenozoic sediment (Fig. 9.37). Fault-generated relief between uplifted blocks and the basement floor of the basins is 2–5 km, and the entire area has been elevated by 2–3 km (Stewart 1978) and extended to at least double its original width (Hamilton 1987). The province is characterized by anomalously thin crust and high heat flow. It contains abundant early-Cenozoic calc-alkaline and silicic volcanics, whereas younger volcanics are bimodal, including the voluminous Columbia Plateau basalts, of Miocene age. Rifting began locally in the Eocene, but the main phase of active extension and basin subsidence did not commence until the mid to late Miocene. Hypotheses of the origin of the Great Basin involve mantle upwelling over a subducted East Pacific Rise, a wide zone of transcurrent faulting in a dextral megashear, or backarc spreading. Scholz et al. (1971), Stewart (1978), Eaton (1984), and Hamilton (1987) favored the last of these. The change in volcanic composition and the initiation of large-scale extension are undoubtedly related to the reduction in regional compression following the cessation of Laramide movements in the Eocene, and the interaction of the East Pacific Rise with the western continental margin in the late Oligocene. After this time, interplate stresses were mainly taken up by transcurrent movement on the San Andreas transform system (Stewart 1978; Dickinson 1981; Coney 1987). The Basin and Range province seems to be a special case and may not be a good analogue on which to build a general model of backarc basins. Thus, Ingersoll (1982) developed an alternative explanation based on rotation and shear along the San Andreas transform system as the triple junction at its northern end migrated northward. Wilson (1988) related uplift and extension of the American southwest to the result of the westward drift of the North American plate over the zone of mantle upwelling corresponding to the East Pacific Rise.

Despite these problems of interpretation, the Basin and Range is worth describing at this point, because extensive study of structural geology at the surface and several deep-reflection-seismic transects have provided an exceptionally clear picture of the structural style and kinematics of large-scale continental extension and have provided the main database from

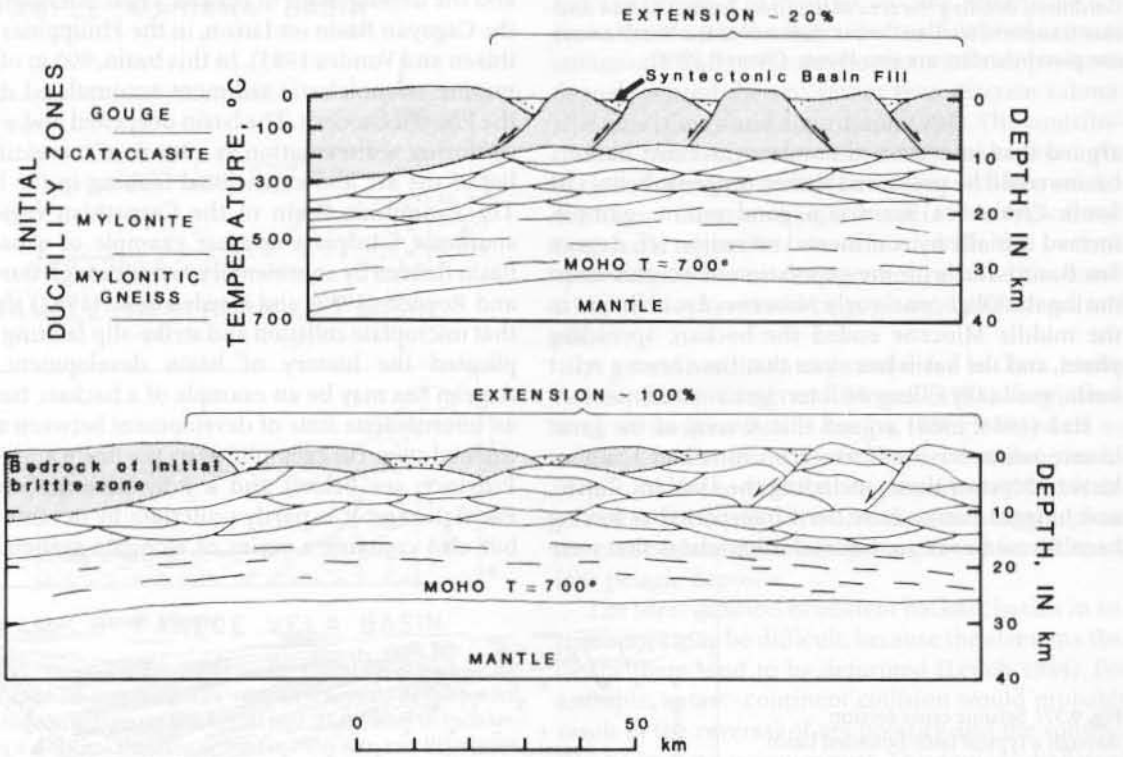


Fig. 9.38. A model of large-scale crustal extension in the form of extended lenses or crustal boudinage. (Hamilton 1987)

which models of simple shear have evolved (Anderson et al. 1983; Eaton 1984; Wernicke 1985; Allmendinger et al. 1987a,c; several papers in Coward et al. 1987; Sects. 7.3, 9.3.1). Seismic results show that low-angle detachment faults are present in the eastern Basin and Range, but there are no unequivocal examples of detachments penetrating to the base of the crust and offsetting the Mohorovičić Discontinuity (Allmendinger, et al. 1987a, c), nor do the data indicate that the faults flatten out at a midcrustal decoupling horizon. A model of extended lenses (Hamilton 1987) or crustal boudinage (Davis and Coney 1979) seems to receive the most support (Fig. 9.38). As described by Hamilton (1987), upper-crust bedrock blocks are bounded laterally by downward-flattening listric faults, rotated planar faults, and basin fills deposited in widening gaps between separating blocks. The detachment faults define the tops of great lenses that are separated by gently dipping, anastomosing, ductile faults and that retain prefaulting fabrics in their interiors. As the lenses slide apart, the area of their composite top, the detachment faults, is increased, and the middle crust is extended by discontinuous ductile flow. Although most faults above the typically midcrust detachments are probably listric and curve into other detachments, in places, major gently dipping faults rise laterally for tens of kilometers through upper-crust rocks and curve to steeper dips only at shallow depth.

Detachments may be reactivated Paleozoic or Mesozoic thrust faults (Allmendinger et al. 1987a). Core complexes are lenses that developed by ductile processes at mid- to lower-crustal levels and, subsequently, have been exposed by removal of upper plates and by uplift (Fig. 9.38). Seismic transects across much of the central Basin and Range show a characteristic pattern of "numerous horizontal and gently dipping reflections in the middle and lower crust," which Allmendinger et al. (1987b) assigned to their zone 4, one of four characteristic crustal seismic fabrics revealed by deep-seismic reflection (see Sect. 9.3.4 for a further discussion of these fabrics).

An alternative approach for developing a general model of sedimentation for this category of basin is to examine various ancient mobile belts. Ensialic backarc basins may be expected to have survived somewhat more intact than those developed over more readily subductable oceanic crust. Three possible ancient examples are the Cambrian to Silurian (Caledonian) Welsh Basin, the Devonian to Carboniferous (Hercynian) basins in western Europe, and the lower-Paleozoic Hazen Trough of the Arctic Islands.

The Welsh basin developed behind a volcanic arc trending northeastward through North Wales and contains at least 12 km of volcaniclastics and continent-derived turbidites and mudstones. It shows a transition to shallow-marine continental shelf facies toward the southeast (Phillips et al. 1976; Siever and Hager 1981; Kokelaar et al. 1984). Middle Devonian to Carboniferous rocks along a trend from South Devon, England, to the Ardennes are characterized by pelagic mudstones and carbonates that Leeder (1976, 1982) interpreted as the fill of a backarc basin above a north-dipping subduction zone. Northward in England, these pass into block-faulted basins, the outlines of which are strongly controlled by earlier lines of weakness. Hazen Trough in the Arctic Islands contains radiolarian chert and thin-bedded turbidites derived (transversely) from a magmatic arc to the northwest. The southeast margin of the basin is well exposed and shows a complete transition through slope facies to shelf carbonates (Trettin et al. 1972). Similarities among these three examples include the following, which may be regarded as the essential elements of a basin model:

1. A deep-marine basin fill derived from either or both sides of the basin
2. A transition into slope and shelf facies on the continental margin of the basin, analogous to that of divergent plate margins
3. A magmatic arc on the other side of the basin
4. Little evidence of contemporaneous tectonics, except normal faulting
5. A basement of normal or attenuated continental crust

9.3.2.7 *Retroarc Foreland Basins*

In the model of contractional arcs, the landward margin of the arc is deformed into a fold–thrust belt (Fig. 9.19, model 5; after Dewey 1980). The crust may be shortened and thickened by a factor of two, giving rise to a major mountain belt exposing the metamorphic and plutonic core rocks of the arc as well as the upthrusted masses of older rocks comprising the local continental basement. The loading of the craton with superimposed thrust sheets creates a downward flexure of the lithosphere (Price and Mountjoy 1970; Beaumont 1981; Jordan 1981). This basin is filled with sediment derived from the adjacent mountains, and the sediment load itself causes further crustal depression (Sect. 7.4). This type of sedimentary basin is a **retroarc foreland basin** in Dickinson's (1974) terminology. Jordan (1995) and DeCelles and Giles (1996) provided recent reviews. The term **retrofore-land basin** is also now being used for this class of basin (Johnson and Beaumont 1995). Bally and Snelson (1980) refer to the crustal shortening and

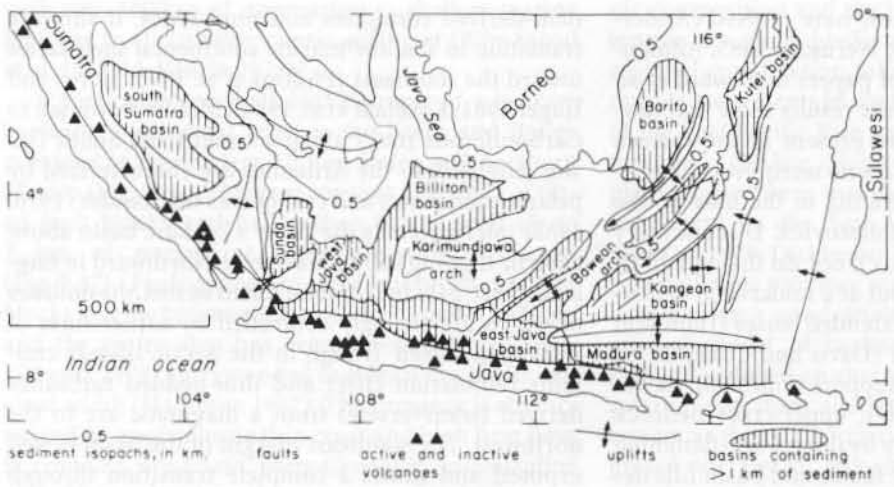


Fig. 9.39. Tectonic elements of Sunda Shelf. (Ben-Avraham and Emery 1973)

underthrusting of continental basement as A-subduction after the Austrian geologist Ampferer, who described these processes in the early part of the century. The term is perhaps misleading, because it is doubtful if much actual subduction of continental crust occurs. Retroforeland basins are briefly described in this section, but considerable more detail regarding the structural and stratigraphic evolution of foreland basins is presented in Section 9.3.4.7.

Other retroarc basins occur behind arcs where there was apparently no collisional event to cause crustal shortening. Examples include the Alberta Basin (MacQueen and Leckie 1992; Mossop and Shetsen 1994), Sevier Basin of Utah and Idaho (Jordan 1981; Cross 1986; DeCelles and Currie 1996), Andean basins (Stabler 1990), including the Magellanes Basin of southern South America (Biddle et al. 1986) and the Bermejo Valley of northwest Argentina (Johnson et al. 1986; Jordan et al. 1988), and the series of basins behind the Sunda arc in Sumatra, Java, and the Sunda Shelf (Ben-Avraham and Emery 1973). Hamilton (1985) attributed crustal thickening in contractional arcs "to the gravitational rise and spread of batholithic plutons."

Retroarc foreland basins should not be confused with proforeland or peripheral foreland basins, which are located on the downgoing plate adjacent to continental sutures but display similar structural and stratigraphic histories to retroarc basins. Examples of this second class of basin include the Appalachian Basin, Alpine foreland basins of Germany, Switzerland, France, Spain, etc., and the Indo-Gangetic basin of northern India and Pakistan. This category of basin is discussed in Section 9.3.4.

Mitrovica et al. (1989) and Gurnis (1992) demonstrated that subsidence in retroarc foreland basins may be amplified by a "dynamic, slab-driven" effect resulting from the "viscous coupling between the base of the continental plate and downward circulating mantle-wedge material that is entrained by the subducting slab" (DeCelles and Giles 1996). This effect can result in considerable widening of the basin (by more than 1000 km) and an additional subsidence of more than 1 km (Fig. 7.33). The broad, flooded continental crust behind the Sunda arc (Fig. 9.39) may be undergoing this type of additional subsidence (Gurnis 1992), and the importance of this "dynamic load" has also been demonstrated for the Alberta basin (Mitrovica et al. 1989; Catuneanu, et al. 1997; Sect. 7.4.2).

The basins of Sumatra and Java developed in a neutral-arc setting (Fig. 9.39), as described by Ben-Avraham and Emery (1973), Hamilton (1979), and Moss et al. (1997). The basement consists of Paleozoic to early-Tertiary cratonic rocks and arc complexes. Linear basins and ridges of the Sunda shelf are mainly controlled by these old alignments. Those basins whose origin can be related directly to Cenozoic retroarc subsidence behind the Java–Sumatra arc are the Madura, East Java, Tjeribon, West Java, and South Sumatra basins. Sediment thicknesses reach maxima of about 3–6 km in these basins. The Sunda and part of the West Java Basin are graben basins bounded by north–south faults. These faults have been active during the Cenozoic, suggesting that they have moved in response to arc-related intraplate stresses, but their trend suggests basement control. The West Java–Tjeribon and Madura basins are separated by a basement high. Sedimentation there began with marine transgression during the Oligocene. Bathyal conditions persisted until the later Neogene in the east, but thick deltaic clastics accumulated in the West Java Basin in the Oligocene. A widespread reef limestone developed in the Miocene. Regression occurred everywhere during the Neogene, and these basins are now largely continental. Basement rocks

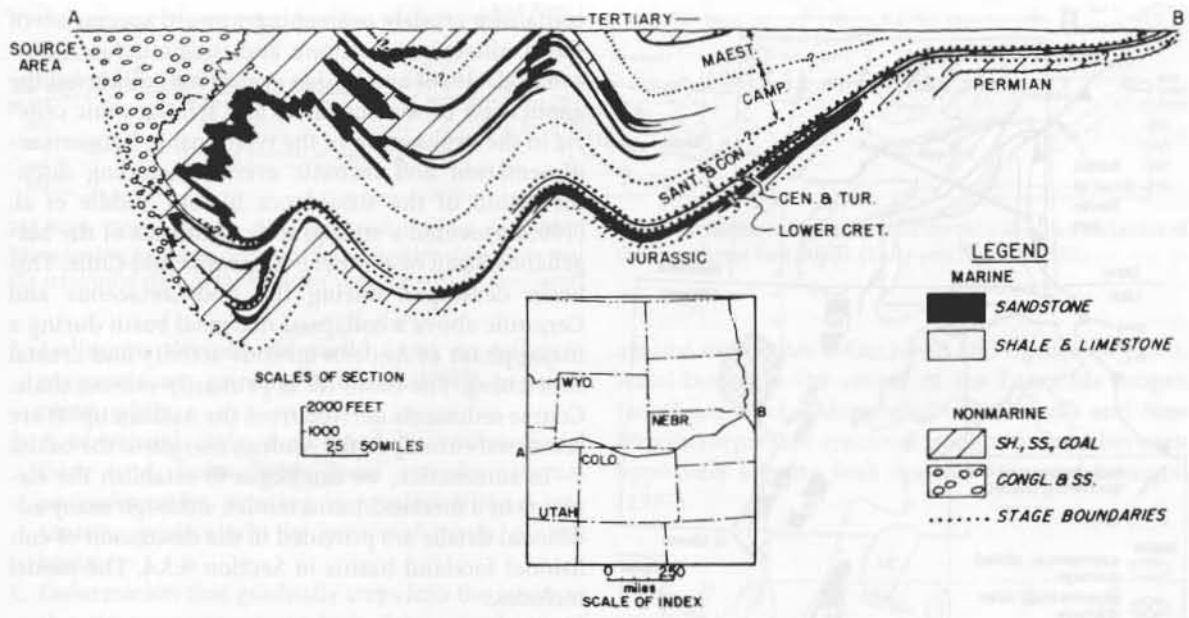


Fig. 9.40. Reconstructed stratigraphic cross section through the Cretaceous fill of the Rocky Mountain retroarc basin. (Weimer 1970)

contributed sediment to the basins at first, but contemporaneous volcanics became the main sediment source during the younger Neogene. Throughout this period, the magmatic arc gradually migrated northward. Syndepositional folding commenced first in the south and is probably continuing, resulting in deformation that decreases in intensity northward. Hamilton (1979) attributed the folding to lateral pressures of magmatic intrusion. The sedimentary basins probably owe their origin to downbowing of the crust beneath the weight of volcanic and plutonic rocks in the magmatic belt. A fold–thrust belt has not yet developed here. The geology of the Sumatra foreland is similar. Basement control by north to north–northwest trending structures had an important influence on sediment thicknesses and facies changes. Miocene strata include quartzose deltaic sandstones derived from the Malay Peninsula and Sunda Shelf to the north.

The retroarc zones of contractional arcs are characterized by thin-skinned fold–thrust belts. The load of each thrust sheet during crustal shortening results in structurally lower thrusts becoming locked by expulsion of pore waters, leading to an increase in friction. Continued shortening then results in the development of new folds and thrusts in the hitherto undeformed continental basement and cover rocks, so that deformation tends to gradually step basinward (Armstrong and Oriel 1965; Dahlstrom 1970). In this way, early basin-fill sediments are deformed and become the source of cannibalized or recycled sediment contributing to the younger basin fill (Steidtmann and Schmitt 1988). Sediment is usually also fed into the basin from the continental side. Exceptions to this general model of outward-stepping deformation and sedimentation are discussed in Section 7.4.2.

The Sevier foreland basin spanned the Jurassic to early Tertiary, but the main phase of rapid subsidence occurred during the Albian–Santonian, during a phase of rapid convergence between North America and the Farallon plate (Dickinson 1981; Cross 1986). This basin, therefore, is a contractional, retroarc type, in the sense of Dickinson (1974), Dewey (1980), and Hamilton (1985). Weimer (1970) showed that the Cretaceous section underwent a crude upward coarsening and a facies transition from marine mudstone and limestone to nonmarine sandstone and conglomerate. The clastic wedge thins and passes laterally into marine mudstones and limestones eastward, away from the orogen (Fig. 9.40). Syndepositional synclines developed progressively further from the source, in part as a response to sediment loading in shifting depocenters. Jordan (1981) showed that the isopach pattern of mid-Cretaceous strata could be explained by a model of flexural loading by thrust sheets, the mechanism described in Section 7.4. Campanian–Maastrichtian isopachs, however, do not correspond to a typical foreland-basin pattern, and Cross (1986) related them to a change in subduction patterns and subcrustal thermal events, the details of which are beyond the scope of this book.

Two studies of basins adjacent to the southern Andes also illustrate examples of retroarc foreland basins developed in the absence of collisional events. The Neogene Bermejo Basin of northwest Argentina

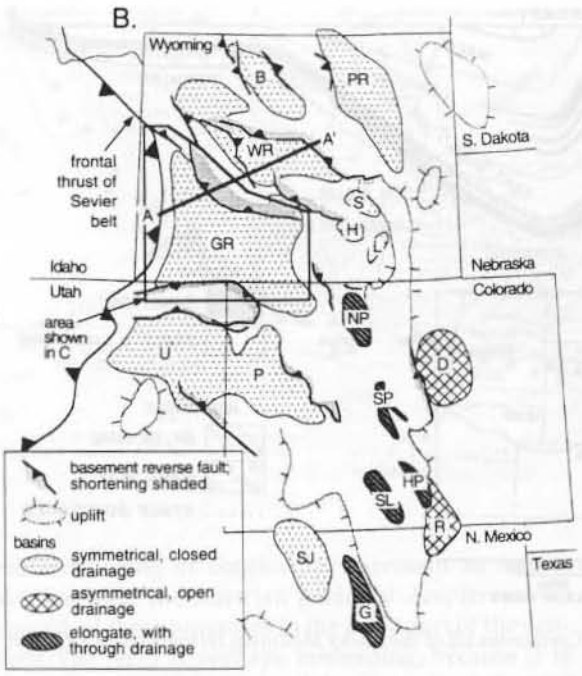
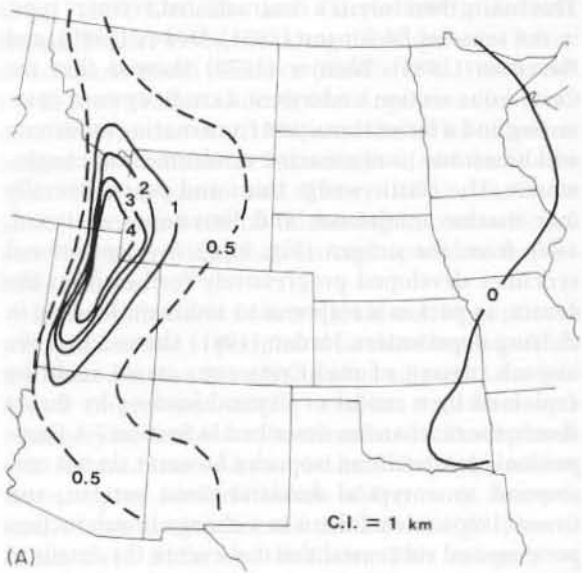


Fig. 9.41. Laramide basins of the US Rocky Mountains states. Basins include: *B* Big Horn; *PR* Powder River; *WR* Wind River; *GR* Green River; *H* Hanna; *S* Shirley; *L* Laramie; *U* Uinta; *P* Piceance; *NP* North Park; *SP* South Park and Echo Park; *D* Denver; *SL* San Luis; *SJ* San Juan; *HP* Huerfano Park; *R* Raton; *G* Galisteo. (Jordan 1995)

contains a crudely coarsening-upward succession of fluvial siltstone, sandstone, and conglomerate. Johnson et al. (1986) and Jordan et al. (1988) discussed the application of various facies and stratigraphic criteria in the exploration of the relationship between sedimentation and tectonic events, including documentation of the subsidence history. Biddle et al. (1986) provided a stratigraphic synthesis of the Magellanes Basin of southern Argentina and Chile. This basin developed during the mid-Cretaceous and Cenozoic above a collapsed marginal basin during a major phase of Andean igneous activity and crustal shortening. The basin fill is primarily marine shale. Coarse sediments derived from the Andean uplift are restricted entirely to the Andean margin of the basin.

To summarize, we can begin to establish the elements of a foreland-basin model, although many additional details are provided in the discussion of collisional foreland basins in Section 9.3.4. The model includes:

1. A basement of normal continental crust, the structural grain of which markedly affects isopach and facies patterns as a result of structural rejuvenation.
2. A dominantly shallow-marine to nonmarine, clastic fill that may contain an upward-coarsening transition from marine to nonmarine.



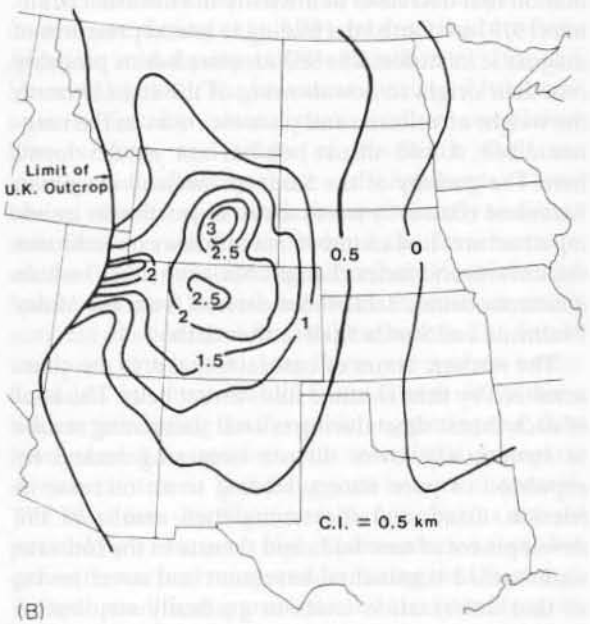


Fig. 9.42 A, B. Contrasting styles of subsidence in the foreland of the US Western Interior. A The typical pattern of foreland-basin subsidence generated by thin-skinned tectonics and supracrustal loading of the edge of the craton, as shown by isopachs of the upper Albian–Santonian strata adjacent to the Sevier orogen; B Isopach of Campanian–Maastrichtian strata of the same region, showing broad subsidence centered on Wyoming and Colorado, attributed to subcrustal loading by a cooling oceanic slab. (Cross 1986)

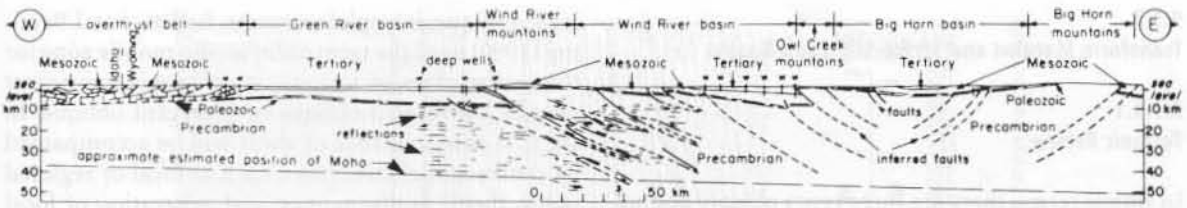


Fig. 9.43. Cross section through Powder River and Wind River basins, based on deep-seismic-reflection data. Note that the structural style of these basins is dominated by deep-seated thrust faults, contrasting with the thin-skinned tectonics of the overthrust belt at *left*. (Bally and Snelson 1980)

3. Sediments derived primarily from an adjacent magmatic arc, growing fold-thrust belt, or basement uplifts.
4. A structural style dominated by contractional folds and faults, with local intraformational unconformities attesting to syndepositional tectonism, particularly in proximal parts of the basin.
5. Deformation that gradually steps into the basin as thrusts are successively locked by expulsion of pore fluids. Older basin-fill sediments are themselves deformed, uplifted and cannibalized, with recycling into younger sediments.
6. Minor sediment input from the continental side of the basin, including the forebulge, is common.

A variant on the model described here is the **broken-foreland-basin** or **intermontane-foreland-basin** model. These basins are characterized by internal basement uplifts that divide the foreland into a series of smaller, isolated to semi-connected, fault-bounded basins (Rodgers 1987; Jordan 1995). The classic example is that of the Western Interior of the US Rocky Mountain states, the so-called Laramide basins (King 1977; Chapin and Cather 1981; Dickinson et al. 1988; Fig. 9.41). Jordan and Allmendinger (1986) described a comparable example in the Andean foreland of Argentina.

The consensus regarding the origins of these basins is that they developed in areas of flat to very shallow subduction (Cross and Pilger 1978; Dickinson and Snyder 1978; Cross 1986; Jordan 1995). The pattern of subsidence over the subducting slab is quite different from that of a typical foreland basin (Fig. 9.42). Isopachs of the sediments that immediately predate the generation of the Laramide basins show a broad area of subsidence generated by the thrust load. The basement uplifts that are the characteristic features of the Laramide basins consist of steep reverse faults that penetrate to the base of the crust (Fig. 9.43). These developed because the subducting slab remained mechanically coupled to the overlying crust, generating widespread contraction in the foreland. The foreland was fragmented into a series of broken "beams" that rotated against each other (MacQueen and Beaumont 1989). Axial basins in the center of the Laramide orogen (southern Rocky Mountains of Colorado and New Mexico) represent synclinal troughs formed between oppositely verging fault systems (Yin and Ingersoll (1997).

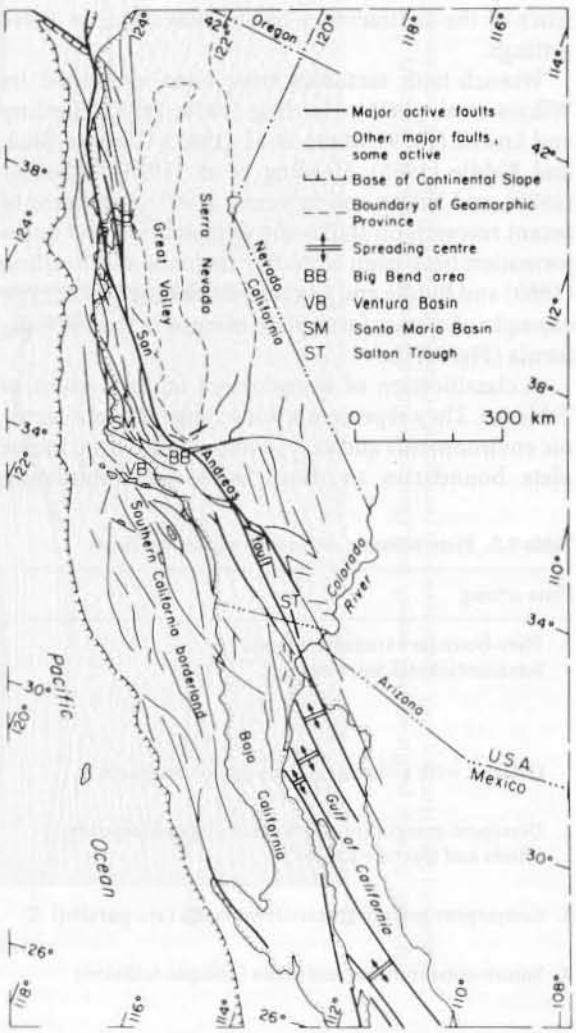


Fig. 9.44. Fault map of California, showing the relationship of the faults to Gulf of California spreading centers. (Based on Crowell 1979; Kelts 1981)

9.3.3 Transform Margins and Strike-Slip Fault Basins

9.3.3.1 *Tectonic Review*

In simple terms, there are three types of plate boundaries: convergent, divergent, and transform. However, as we have seen, convergent and divergent margins show simple, orthogonal plate trajectories only along short margin segments. Oblique motions are the rule, and the more oblique they are, the more important are strike-slip motions at plate junctions and in adjacent plate interiors. A case could be made for eliminating transform and strike-slip margins as a separate category for the purpose of this chapter and considering them as variations on extensional or contractional margins. However, a separate section such as this permits a focus on some distinctive characteristics of the sedimentary basins that occur in these settings.

Wrench-fault tectonics have been examined by Wilcox et al. (1973), Harding (1974, 1985), Harding and Lowell (1979), Mann et al. (1983), Christie-Blick and Biddle (1985), Harding et al. (1985), Sylvester (1988), and Nilsen and Sylvester (1995). Syntheses of recent research on strike-slip deformation and basin formation have been edited by Ballance and Reading (1980) and Biddle and Christie-Blick (1985). The type example of a transform plate margin is that of California (Fig. 9.44).

A classification of transcurrent faults is given in Table 9.5. They represent a wide range of plate-tectonic environments and vary in importance from major plate boundaries to short faults accommodating local, oblique, intraplate stresses. Ballance and Reading (1980) used the term **oblique-slip mobile zone** for this entire class of tectonic styles. Fault movement may be convergent oblique or divergent oblique. In the first case, evidence of shear will be accompanied by contractional structures, such as local or regional uplift, thrust faults, nappes, and generation of local unconformities in contemporaneous sediments. Sedimentary basins are of the supracrustal-loading (foreland) type (Sect. 7.4). Harland (1971) coined the term **transpression** for this type of stress. Divergent-oblique stress leads to the development of extensional faults and diverse types of extensional sedimentary basins (Sect. 7.3) and was termed **transtension** by Harland. In both cases, the presence of a shear component may be more difficult to prove than that of simple compression or extension, which accounts for the fact that geologists have been slow to recognize that strike-slip movements have been important in almost all orogenic belts. The terms transcurrent fault, strike-slip fault, and wrench fault are used more or less synonymously, although some authors suggest that only fundamental structures rupturing the entire lithosphere should be termed transcurrent faults (Harland 1971, in discussion of Norris et al. 1978; Reading 1980).

The structural style of strike-slip fault zones is distinctive. In cross section, they commonly show a splayed **flower structure** or **palm-tree structure**, the main fault splaying upward into several radiating fractures (Fig. 9.45). In plan view, careful mapping may reveal several kinds of associated structure, as summarized in Fig. 9.46. Synthetic strike-slip faults occur at an angle of 10–30° to the shear direction. Antithetic faults are oriented at 70–90° to the shear di-

Table 9.5. Plate-tectonic settings of strike-slip faults

Plate setting	Examples
1. Plate-boundary transform faults	
Intracontinental, intra oceanic	San Andreas, California; Alpine, New Zealand; Dead Sea, Middle East; El Pilar–Oca, Caribbean; Greater Antillean–Cayman, Caribbean; Magellan–North Scotia, Scotia Sea; Ninetyeast Ridge, Indian Ocean
Oceanic, with a continental margin or fragment	Bismark, Bismark Sea; New Guinea Molucca–Sorong, New Guinea; Fairweather–Queen Charlotte, western Canada
2. Divergent-margin transform faults (spreading-ridge offsets and fracture zones)	Spitzbergen (Hornsund Fault), Svalbard; Romanche, Atlantic Ocean; Falkland–Agulhas, Atlantic Ocean; Owen, Indian Ocean; Mendocino, Pacific; Pioneer, Pacific Ocean
3. Convergent-margin transcurrent faults (arc-parallel)	Sumatra, Sunda Arc; Philippine, Philippine Arc; Atacama, Chile; Median Tectonic Line, Japan
4. Suture-zone transcurrent faults (oblique collision)	Hornelen Basin faults, Norway; Cabot, Maritime Canada; Cobequid–Chedabucto, Maritime Canada; Rocky Mountain Trench–Tintina, Canada; Kaltag–Porcupine, Alaska; Altyn Tagh, Tibet; Kunlun, Tibet; Red River, Indochina; Quetta–Chaman, Pakistan–Afghanistan; North Anatolian, Turkey

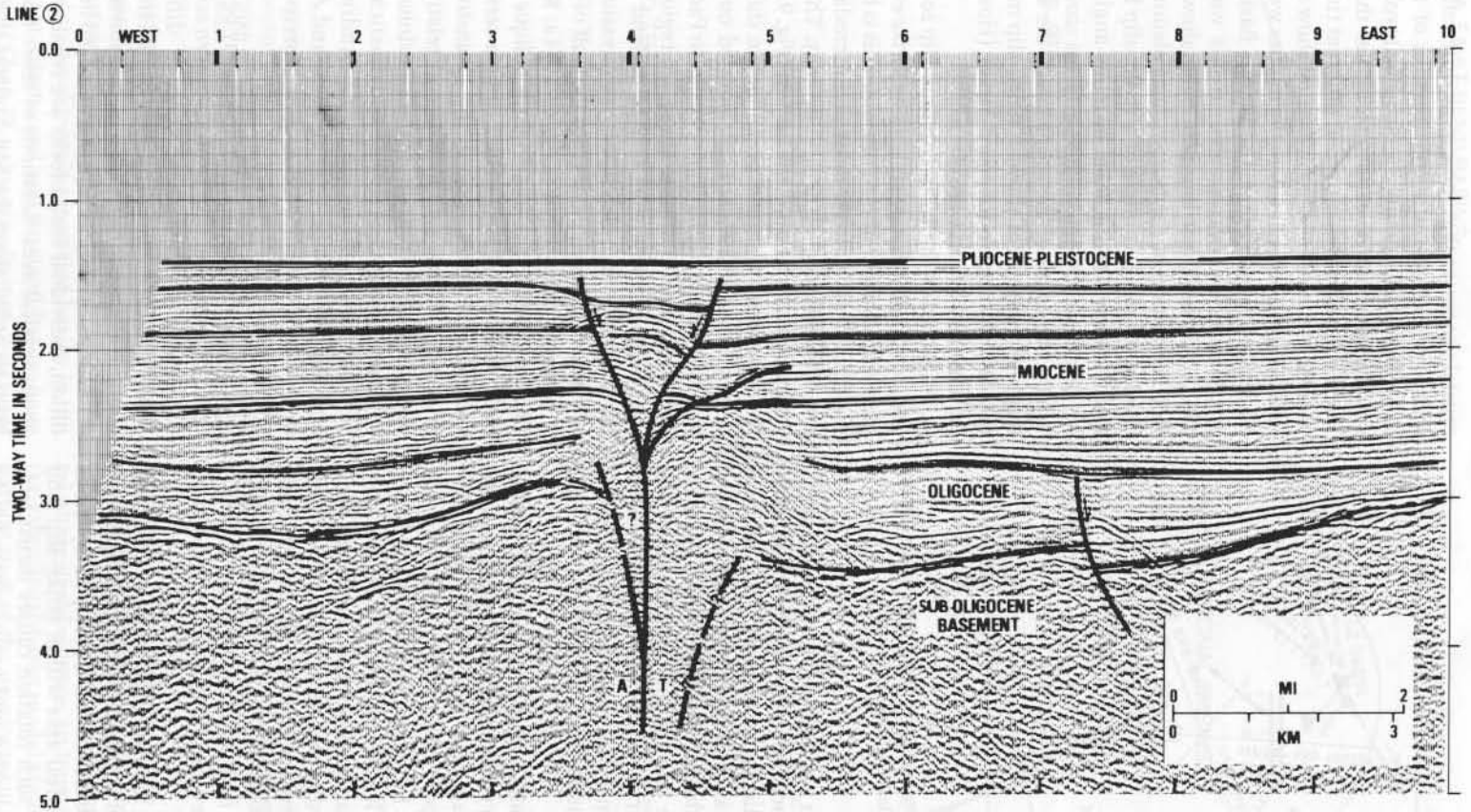


Fig. 9.45. Seismic profile across a wrench fault, showing typical development of a flower structure. *T, A* Displacement toward and away from the viewer, respectively. Andaman Sea. (Harding 1985, reproduced with permission of the American Association of Petroleum Geologists)

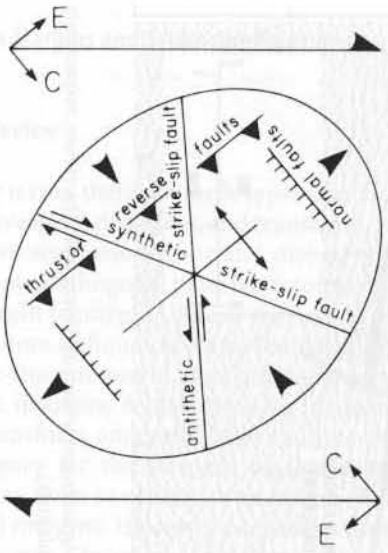


Fig. 9.46. Structural pattern resulting from simple dextral shear. (Harding 1974; Reading 1980)

rection and have an opposite sense of displacement. Normal faults may occur in early stages of fault movement, and perpendicular to these (15–45° to shear direction) are the axes of associated contractional features, including thrust faults and/or folds. All these features may rotate into the direction of shear. The normal faults, in particular, are susceptible to distortion and destruction, and the synthetic faults tend to merge into one master fault, possibly marked by a broad crush zone.

Transcurrent faults are associated with diverse types of sedimentary basins, as shown diagrammatically in Fig. 9.47. Diagram *B* illustrates the generation of a simple **pull-apart basin** where strike-slip displacement is offset. Initially, these may be only a few hundred meters or less in length and width, but continued faulting enlarges them into significant sedimentary basins. Aydin and Nur (1982) showed that basins tens of kilometers in width and length could be produced by coalescence of pull-apart structures on adjacent active fault strands. Mann et al. (1983) provided a useful review of the development of pull-apart basins.

Ingersoll (1988), following the work of Luyendyk and Hornafius (1987), proposed the definition of a class of **transrotational basins** developed by block rotation around vertical axes within strike-slip fault zones (Fig. 9.47C). Such rotation causes thrust faulting where blocks overlap and basins where blocks are rotated apart. Luyendyk and Hornafius (1987) demonstrated that many Neogene basins in southern California are of this type.

It is important to realize that strike-slip motion may, in fact, be a superficial representation of deep crustal processes that show quite different structural style. For example, the construction of the Carpathian Mountains during the Neogene involved differential movement of nappes along flat thrust faults. These nappes, which moved at different times, are separated in places by steep faults that show a history of strike-slip movement where the nappes were displaced relative to each other. The Vienna Basin is a pull-apart basin that was created in this way (Royden 1985). Conversely, it has now been shown that the Hornelen Basin of the Norwegian Caledonides (long cited as a classic example of a strike-slip basin; Steel and Gloppen 1980) is soled by a low-angle extensional detachment surface. This surface is scoop-shaped. Where the edges of the scoop reach the surface, extensional movement shows a strike-slip motion relative to adjacent, undisturbed strata (Hossack 1984; Seranne and Seguret 1987).

Transcurrent faults, and oblique-slip zones in general, commonly change style along strike even though plate trajectories remain constant. This is because the regional stress pattern must accommodate itself to the details of local structural geology. Thus, the San Andreas–Gulf of California system (Fig. 9.44) is both transtensional and transpressional in different segments. Crowell (1974a, 1979) pointed out that only three reaches of the fault are parallel to Pacific–North American plate motions. These are regions of pure strike-slip. Elsewhere, as at the Big Bend (Fig. 9.44), the fault zone is under severe compression. Conversely, where the fault passes into the Gulf of California, it becomes extensional (Crowell 1974a; Kelts 1981). The initial Gulf rift was slightly oblique to dextral plate motion and has opened along a series of short spreading centers linked by transform faults (Fig. 9.44). With a subtle change in relative rotation poles, regional contractional–extensional relationships on transcurrent faults could drastically change. This has happened on both the San Andreas (Nardin and Henyey 1978; Blake et al. 1978) and Alpine fault systems (Carter and Norris 1976; Norris et al. 1978; Spörli 1980).

In areas of regional shear, stress may be accommodated by a single master fault, several *en echelon* faults throughout a broad shear zone, or different fault strands that are active at different times. In its current configuration, the San Andreas transform system is an example of the last type, as discussed below. In addition, individual fault blocks are commonly rotated about vertical axes. Rotations of more than 90° have been recorded along the San Andreas transform system (Christie-Blick and Biddle 1985; Luyendyk and Hornafius 1987). Because of these complications, it is practically impossible to deduce regional stress vectors (plate motions) from the kinematics of local structural and basin histories or to predict local struc-

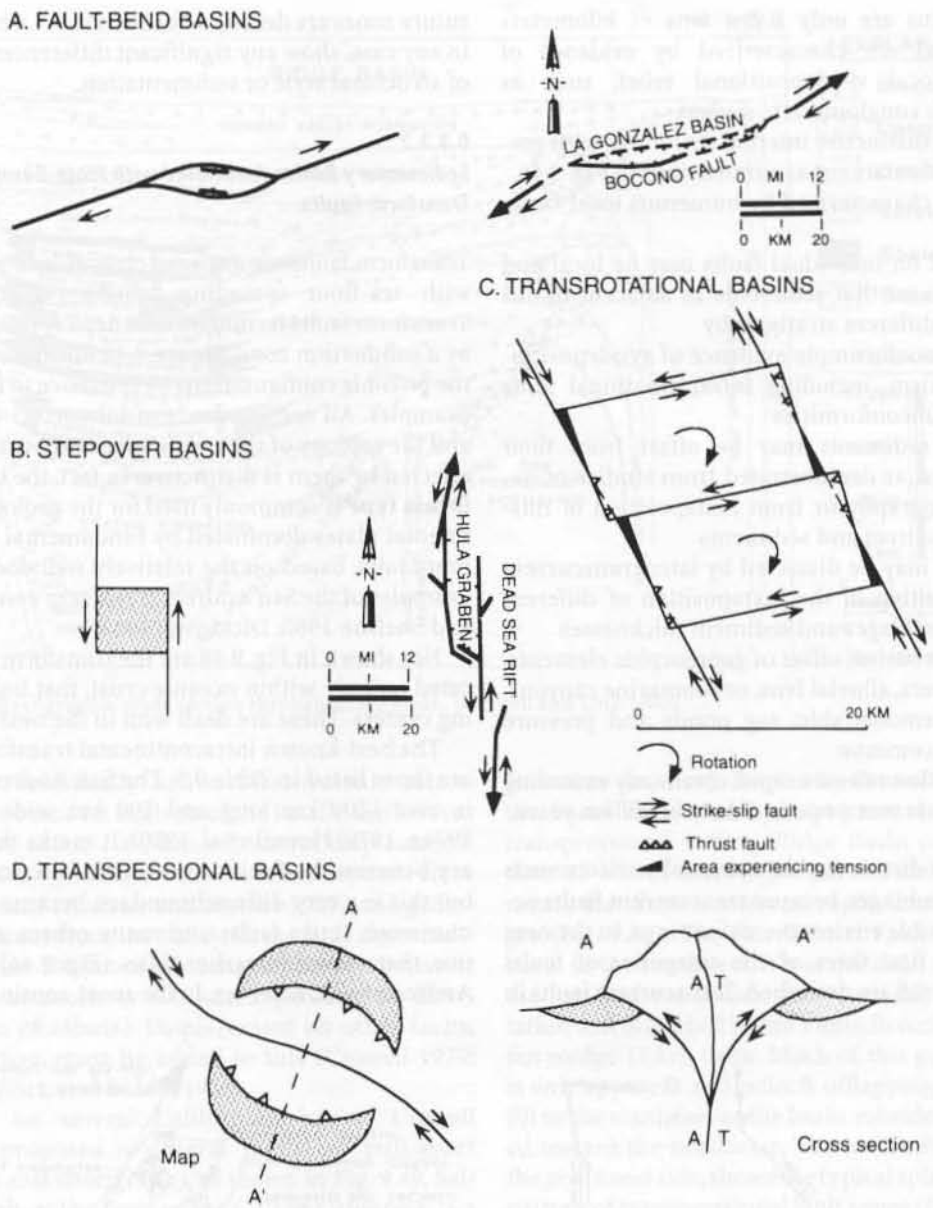


Fig. 9.47. Diagrammatic maps of some of the different types of basin associated with strike-slip faults. *A* Fault-bend basin (*left*) with a map of La Gonzalez Basin (*right*); *B* step-over (pull-apart) basin (*left*) with a map of part of the Dead Sea Rift (*right*); *C* transrotational basins (*black areas*); *D* transpressional basins in map view (*left*) and in cross section (*right*). (Nilsen and Sylvester 1995)

tural and stratigraphic patterns from known regional plate motions (Christie-Blick and Biddle 1985).

Several attempts have now been made to model the evolution of strike-slip-fault-associated basins in the same way that extensional and foreland basins have been modeled (Chap. 7), using simulations of flexural behavior, heat flow, isostasy, etc. Pitman and Andrews (1985) noted that for small, narrow basins, which these basins usually are, lateral heat loss through the margins of the basin is rapid and cannot be ignored. Taking this into account, it can be shown that, in pull-apart basins, thermal subsidence is likely to be more or less contemporaneous with stretching, leading to unusually rapid subsidence. A post-stretching, flexural phase of subsidence, an important component of the extensional-basin model (Sect. 7.3), is not an important part of the development of most pull-apart basins. Because rapid subsidence maintains sharp basin-margin relief, the supply of detritus and basin sedimentation rates are also expected to be very high.

Basin models for the transcurrent-fault setting have these elements in common:

1. Most basins are only a few tens of kilometers across and are characterized by evidence of marked local syndepositional relief, such as fault-flank conglomeratic wedges
2. They have distinctive internal structures with predictable orientations, as summarized in Fig. 9.46
3. Strata are characterized by numerous local facies changes
4. Movement on individual faults may be local and spasmodic, so that sediments in adjacent basins may have different stratigraphy
5. There is usually ample evidence of syndepositional tectonism, including intraformational folds and local unconformities
6. Basin-fill sediments may be offset from their source area, as demonstrated from studies of detrital petrography or from juxtaposition of mismatched sources and sediments
7. The basin may be dissected by later transcurrent faults, resulting in the juxtaposition of different facies assemblages and sediment thicknesses
8. In modern basins, offset of geomorphic elements, such as rivers, alluvial fans, or submarine canyons, may be demonstrable; sag ponds and pressure ridges are common
9. Sedimentation rates are rapid, commonly exceeding 1 m/10^3 years over periods of several million years.

There are no distinctive depositional environments or facies assemblages, because transcurrent faults occur in all possible environmental settings. In the next sections, the first three of the categories of faults listed in Table 9.5 are described. Transcurrent faults in suture zones are described in Section 9.3.4 but do not, in any case, show any significant differences in terms of structural style or sedimentation.

9.3.3.2 *Sedimentary Basins Associated with Plate-Boundary Transform Faults*

Transform faults are a special class of fault associated with sea-floor spreading centers (Wilson 1965). Transform faults terminate at either a spreading ridge or a subduction zone. Figure 9.48 illustrates some of the possible configurations by reference to five actual examples. All are hundreds of kilometers in length, and the geology of the continental and oceanic plates affected by them is distinctive. In fact, the term **California type** is commonly used for the geology of continental plates dominated by fundamental transcurrent faults, based on the relatively well documented examples of the San Andreas transform system (Bally and Snelson 1980; Dickinson 1981).

Not shown in Fig. 9.48 are the transform faults, located entirely within oceanic crust, that link spreading centers. These are dealt with in the next section.

The best-known intracontinental transform faults are those listed in Table 9.5. The San Andreas system is over 1200 km long and 500 km wide (Crowell 1974a, 1979; Howell et al. 1980). It marks the boundary between the Pacific and North American plates, but this is a very diffuse boundary, because there are numerous active faults and many others, now inactive, that moved in earlier times (Fig. 9.44). The San Andreas Fault, however, is the most continuous and

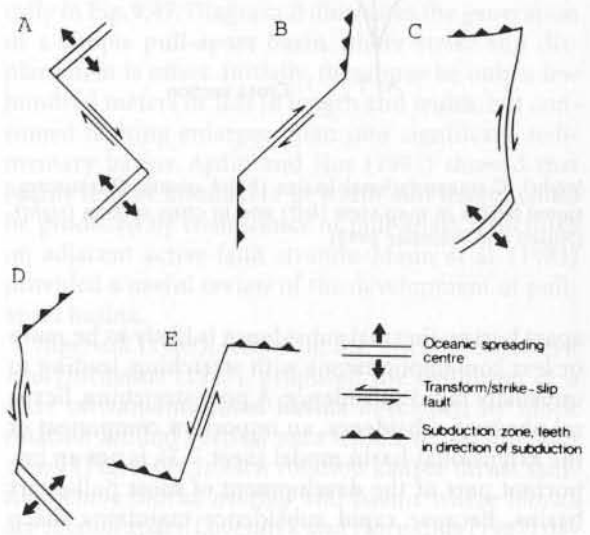


Fig. 9.48 A–E. Examples of plate-boundary transform faults. **A** San Andreas Fault; **B** Alpine Fault, New Zealand; **C** Chugach, Fairweather, Queen Charlotte Fault; **D** Dead Sea Fault; **E** Kirthar-Sulaiman Fault. Various scales. (Reading 1980)

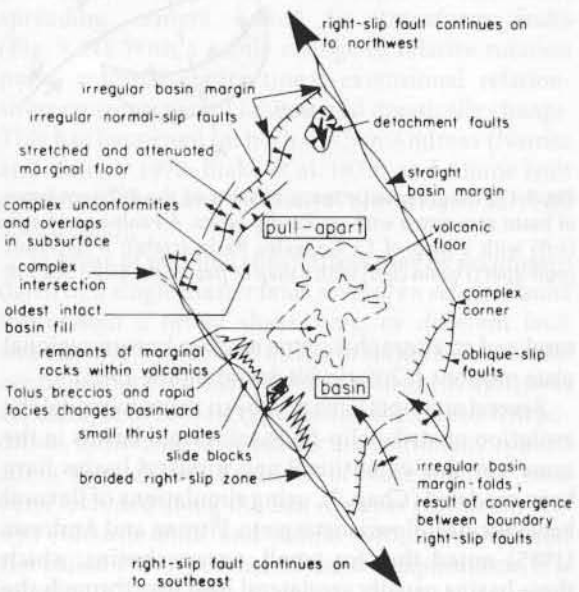


Fig. 9.49. A general model of pull-apart basins. (Crowell 1974b)

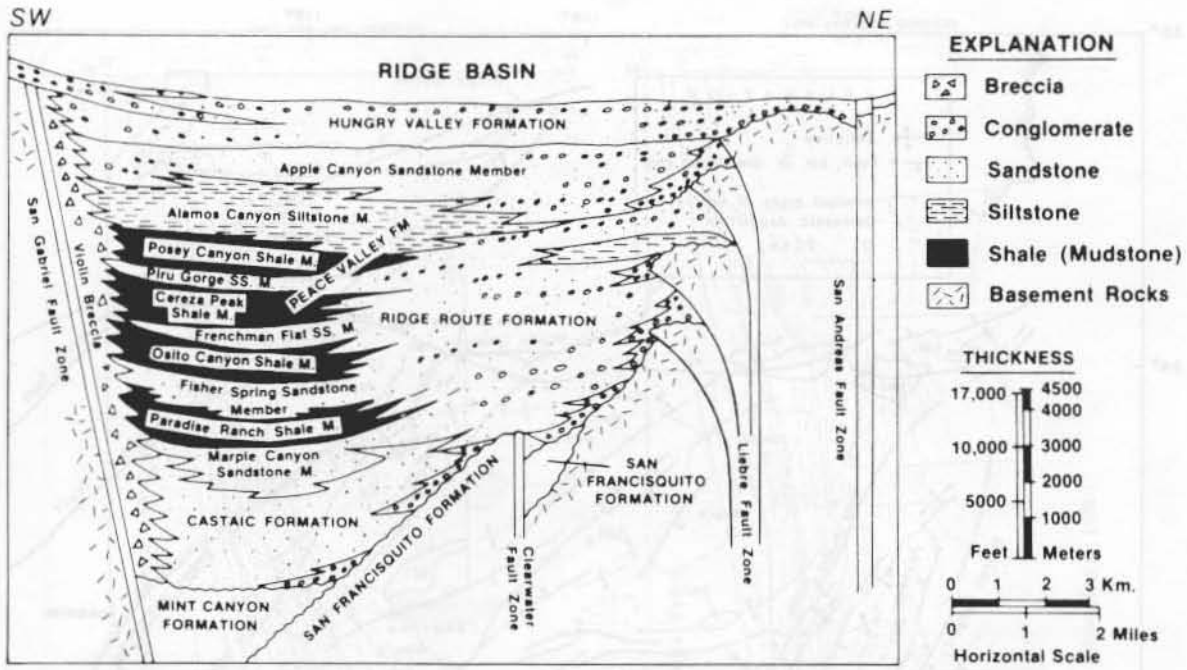


Fig. 9.50. Stratigraphic cross section through Ridge Basin. (Crowell and Link 1982)

the most seismically active. Plate-tectonic reconstructions based on analysis of the Pacific seafloor indicate that about 1000 km of dextral displacement should have occurred on the transform system since the Oligocene (Atwater and Molnar 1973), and geological evidence from the San Andreas fault alone indicates at least 300 km of Miocene to Recent movement (possibly as much as 600 km, depending on interpretations of offsets). Displacement on other faults, now inactive, must be added to this (Crowell 1979; Christie-Blick and Biddle 1985).

Based on several Californian basins, Crowell (1974b) proposed a general model of pull-apart structure and stratigraphy, as shown in Fig. 9.49. Salton Trough, at the head of the Gulf of California, is a typical example. It was initiated during the Miocene and is still active, so it has not been uplifted and dissected for our examination as has Ridge Basin near Los Angeles (discussed below). Volcanism, high heat flow, and tectonism in the Salton Trough suggest that it is an incipient spreading center (Crowell 1974a) and may eventually open fully to expose new oceanic crust, as in the Gulf of California (Kelts 1981). This would accord with what appears to be a pattern of oblique spreading in the area (a "leaky transform").

Ridge Basin is a well-exposed example of a continental strike-slip basin. Excellent studies by Crowell (1974b), Link and Osborne (1978), and Crowell and Link (1982) have turned it into a classic of its kind. The basin was active during Miocene and Pliocene times, when the San Gabriel Fault was the master fault (it is now inactive). Studies by Crowell showed that the basin developed by supracrustal loading in a transpressional setting. Ridge Basin contains more than 12 km of marine, lacustrine, and fluvial sediments that show numerous, rapid, local facies changes (Fig. 9.50). The San Gabriel Fault bounded the basin to the southwest. It was the main active fault of the San Andreas transform system during sedimentation and is flanked by the Violin Breccia, an alluvial-fan wedge 11 km thick. Much of this great thickness is only apparent and reflects offlapping of the basin-fill to the southeast as the basin subsided and enlarged toward the northwest. The Liebre Fault Zone, on the northeast side, shows the typical splayed or flower pattern of transpressional fault zones (Fig. 9.50; compare with Fig. 9.45). Uplift on various splays of this fault was intermittent and localized and generated local, short-lived basement-sediment sources. The basin center is filled with lacustrine mudstone, fluvial–deltaic sandstone, and minor conglomerate deposited by longitudinal and transverse depositional systems flowing from the north and northwest.

In southern California, west of the San Andreas fault, is the largely subsea continental borderland (Fig. 9.51). This area is underlain by a subduction complex formed during subduction of the Farallon plate in the Mesozoic and early Cenozoic (Howell and Vedder 1981). Extensive submarine-fan complexes formed in small, tectonically active basins along the coast during the Paleogene (Nilsen and Clarke 1975; Howell and Link 1979). Commencing in the early

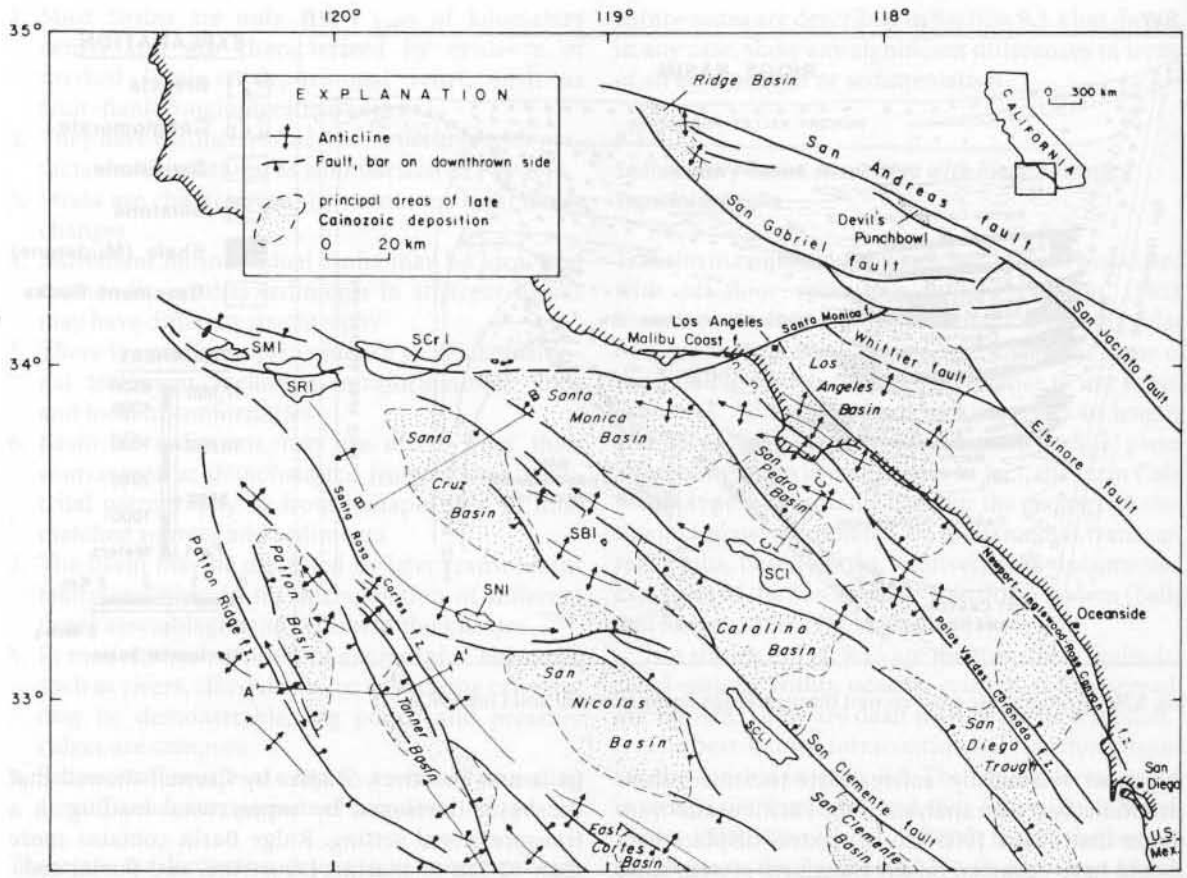


Fig. 9.51. Strike-slip faults and resulting sedimentary basins in the southern California Borderland. (Howell et al. 1980)

Miocene, the area was fragmented by dextral strike-slip faulting and has developed a topography of "*en echelon* linear to lens-shaped ridges and rhomboid-shaped basins" (Fig. 9.51; Howell et al. 1980) underlain by a structure of braided faults – a characteristic fault pattern for a broad zone of transcurrent shear. Sedimentation in the basins has been very rapid (>1 m/10^3 years) and, in those basins close to continental source areas (Ventura and Los Angeles), more than 8 km of Neogene sediment are present (Blake et al. 1978; Howell et al. 1980). Tectonic activity has been almost continuous in these basins during the last 10–15 m.y., so there are numerous syndepositional folds and faults and intraformational unconformities (Fig. 9.52). Fold axes are oriented slightly more westerly than fault trends, reflecting the dextral shear couple (Fig. 9.51). Sediments filling the basins are mainly turbidites, with coastal and nonmarine sediments along the margins. Paleocurrent directions shifted rapidly as local sediment sources were uplifted and then eroded or subsided. Similar sedimentary patterns persist (Malouta et al. 1981), and the entire borderland area seems likely to remain in a very active condition as long it is controlled by transform tectonics.

The Alpine Fault in New Zealand demonstrates a similar tectonic and stratigraphic history to that of California, although the evolution of the fault zone has not yet been clearly established (Carter and Norris 1976; Norris et al. 1978; Spörli 1980; Kamp 1986). During the Oligocene to mid-Miocene, it functioned as a transtensile or neutral strike-slip fault, with the development of fault-bounded turbidite basins. Toward the end of the Miocene, rotation poles between the Australian–Indian and Pacific plates changed slightly, and the fault became transpressive. Many of the sedimentary basins were deformed and uplifted to form the present mountainous backbone of South Island, and nonmarine gravel and sand were shed to fill new basins, such as the broad Canterbury Plains. This phase is continuing, and at least 480 km of dextral displacement have accumulated. At its north and south ends, the fault merges into opposite-facing subduction zones (Fig. 9.48 B). In North Island, the Hikurangi arc and accretionary complex provides an excellent example of oblique subduction well described by Lewis (1980).

The Dead Sea transform is an example of a transtensional fault zone. Sinistral displacement of 105 km

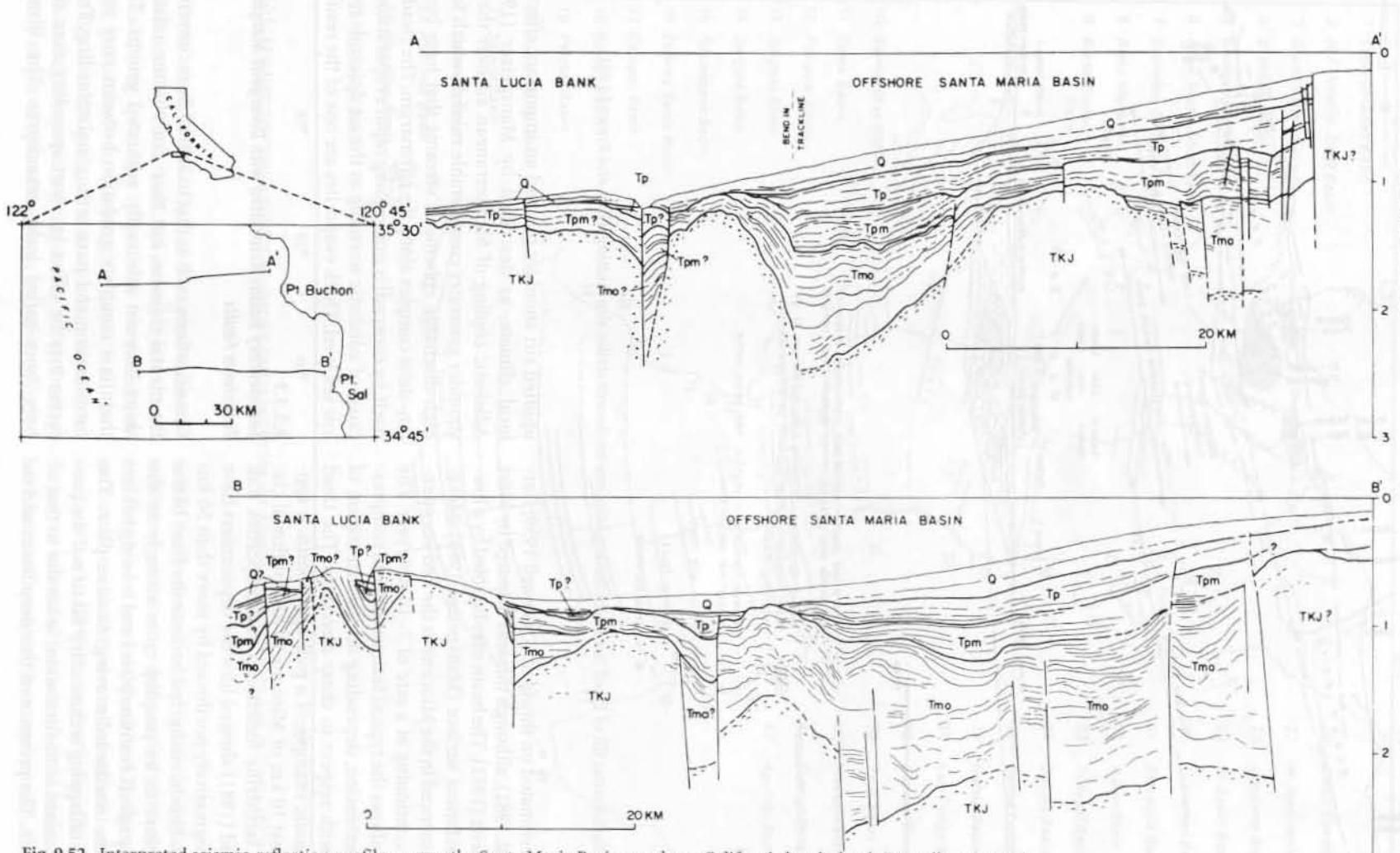


Fig. 9.52. Interpreted seismic-reflection profiles across the Santa Maria Basin, southern California borderland. (Howell et al. 1980)

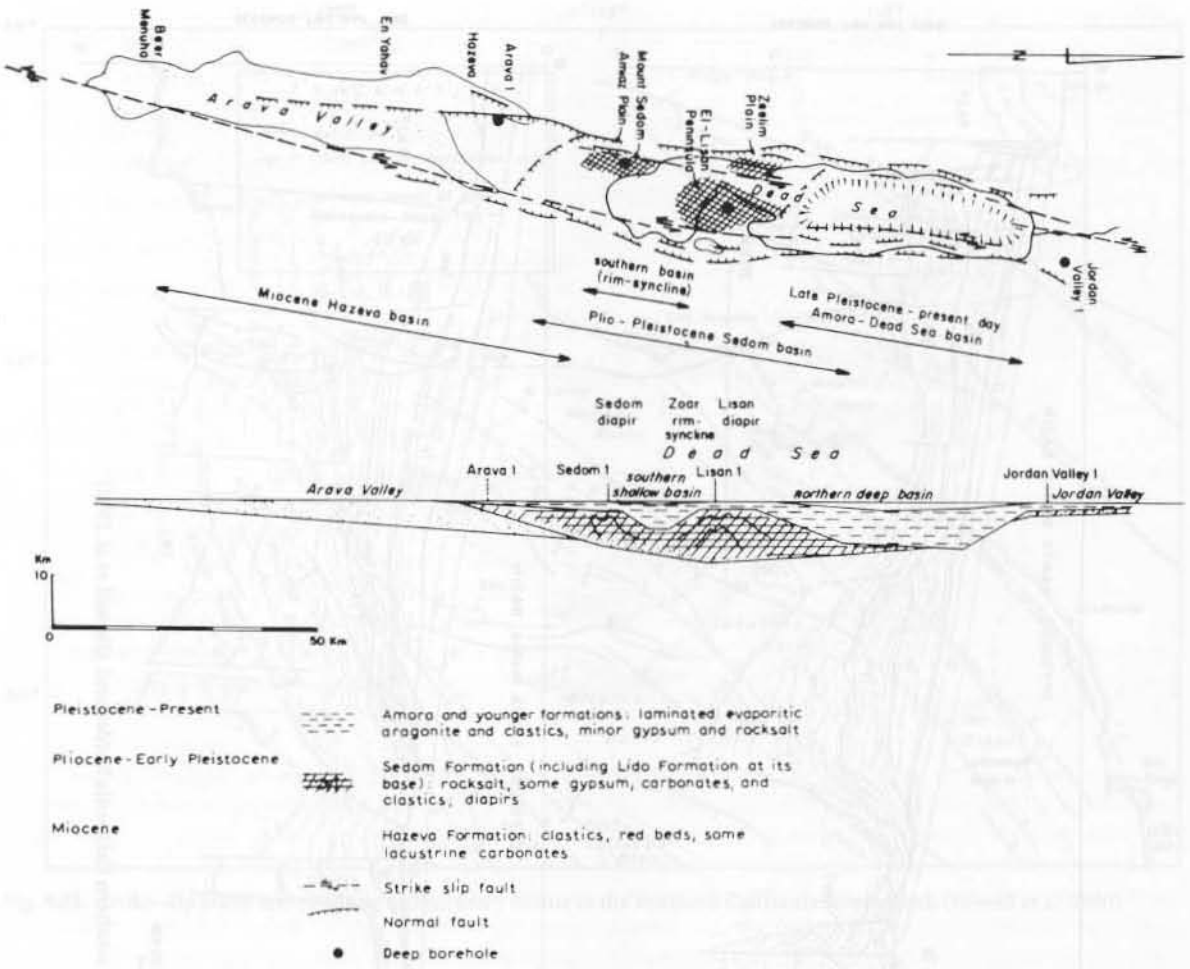


Fig. 9.53. Diachronous fill of Dead Sea Basin during Miocene-to-Recent strike-slip faulting. (Zak and Freund 1981)

has been estimated on this fault (Quennell 1959; Garfunkel et al. 1981), although this was disputed by Mart and Horowitz (1981). The basin may be soled by a low-angle detachment surface (Manspeizer 1985). Movement commenced in the Miocene as the Red Sea opened and is continuing at a rate of 7–10 mm/year. The fault zone shows the typical features of local compression and extension, depending on the orientation of the fault with respect to shear directions. The Dead Sea is a classic example of a pull-apart basin. It contains at least 10 km of Miocene to Recent fluvial clastics and lacustrine limestones and evaporites. Zak and Freund (1981) showed how the depocenters have shifted progressively northward by more than 50 km (Fig. 9.53). This has occurred because the floor of the basin appears to be coupled more strongly to the western margin (Levantine plate) and is being left behind by the northward-moving Arabian plate. The pattern of offlapping sedimentary fill (if not the precise mechanism of its formation) is similar to that of Ridge Basin. The presence of this deep basin and its uplifted rift shoulder has had an important effect on local climate, as described by Manspeizer (1985). Adiabatic cooling of Mediterranean air over the rift shoulder generates considerable rainfall, which feeds high-discharge ephemeral streams that have built a fan–delta complex along the rift margin. The Dead Sea itself, by contrast, is undergoing rapid evaporation because of adiabatic warming as the air descends to below sea level. Thick evaporites are one of the results.

9.3.3.3 *Sedimentary Basins Associated with Divergent-Margin Transform Faults*

Classifications such as that in Table 9.5 are convenient for mental tidiness, but they can obscure relationships between arbitrarily separated groups. Thus, there is a complete gradation between pure transform motion and pure orthogonal spreading. Transforms may be offset by short spreading axes; these have been called leaky transforms. The Bismark

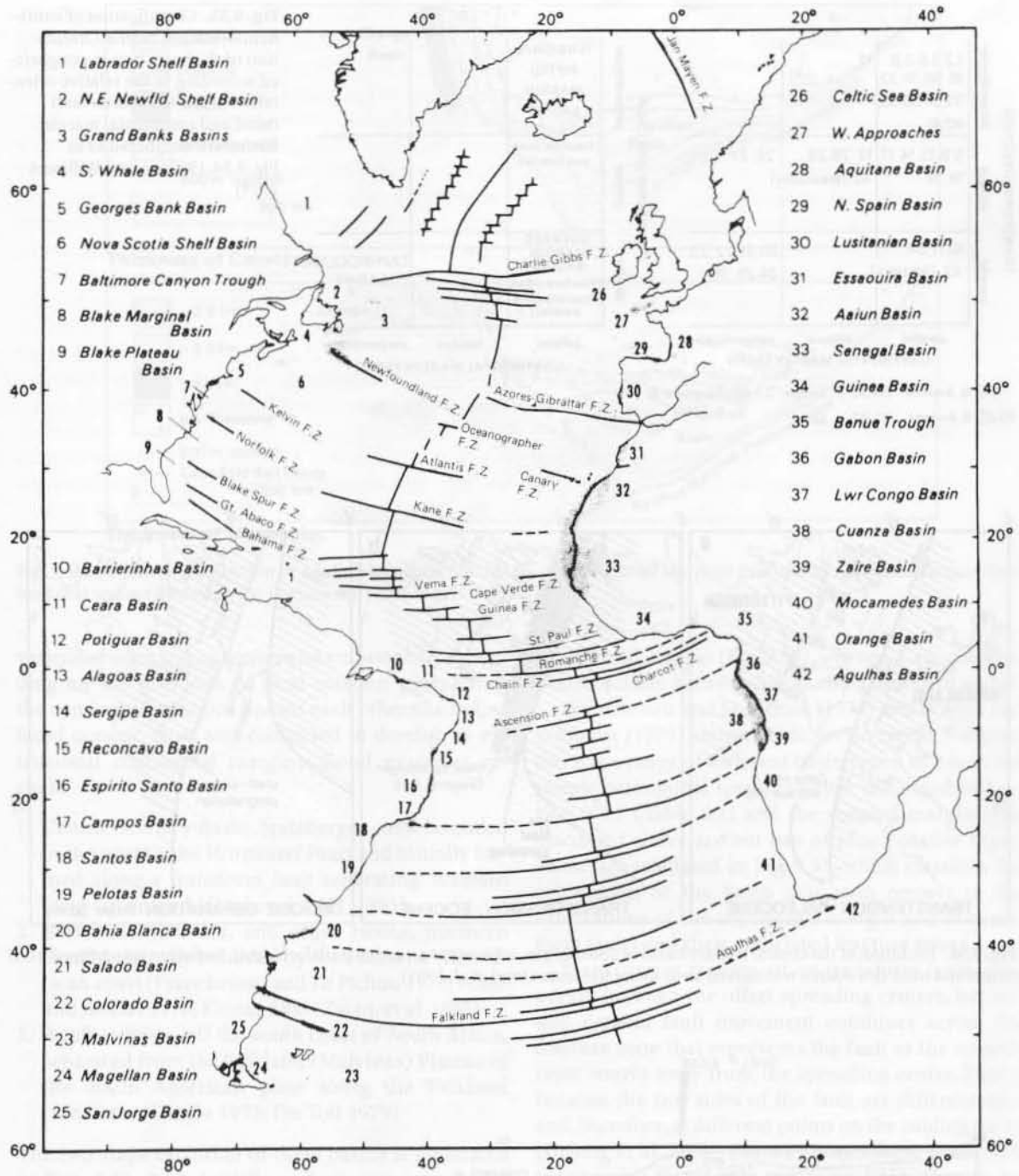


Fig. 9.54. Fracture zones and continental-margin basins bordering the Atlantic Ocean. (Wilson and Williams 1979)

transform north of New Guinea is a good example (Hamilton 1979). The Atlantic spreading center shows regions of long spreading ridges with few transform offsets, particularly in the North Atlantic and, between Brazil and Nigeria, an area with long transform offsets and short spreading ridges (Fig. 9.54). During the early stages of spreading, these transforms functioned as intracontinental plate boundaries, and structural and stratigraphic styles of resulting sedimentary basins were as described in the preceding section. These basins were called sheared-margin basins by Wilson and Williams (1979; Fig. 9.55). Initially, they were two-sided intracontinental basins and may have been either marine or nonmarine. Conditions

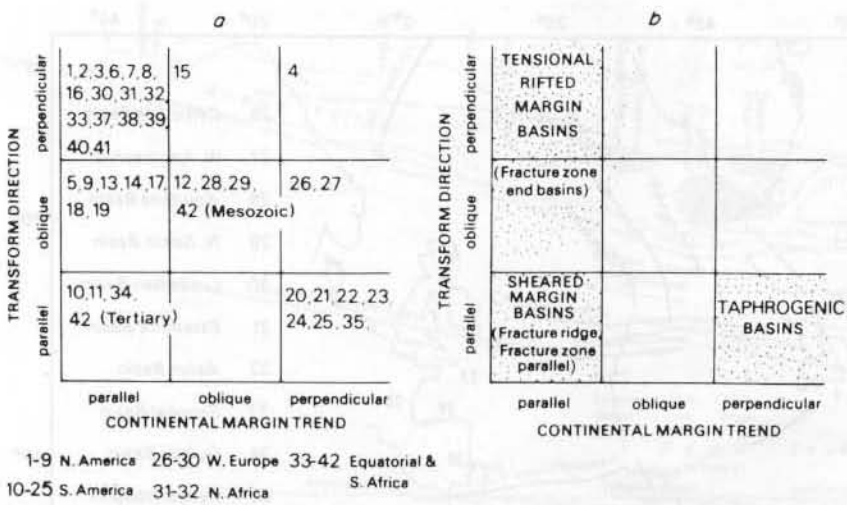


Fig. 9.55. Classification of continental-margin basins. Orientation of the basin axis is categorized according to the relative orientation of the transform-fault trend and continental margin. Basins are numbered as in Fig. 9.54. (Wilson and Williams 1979)

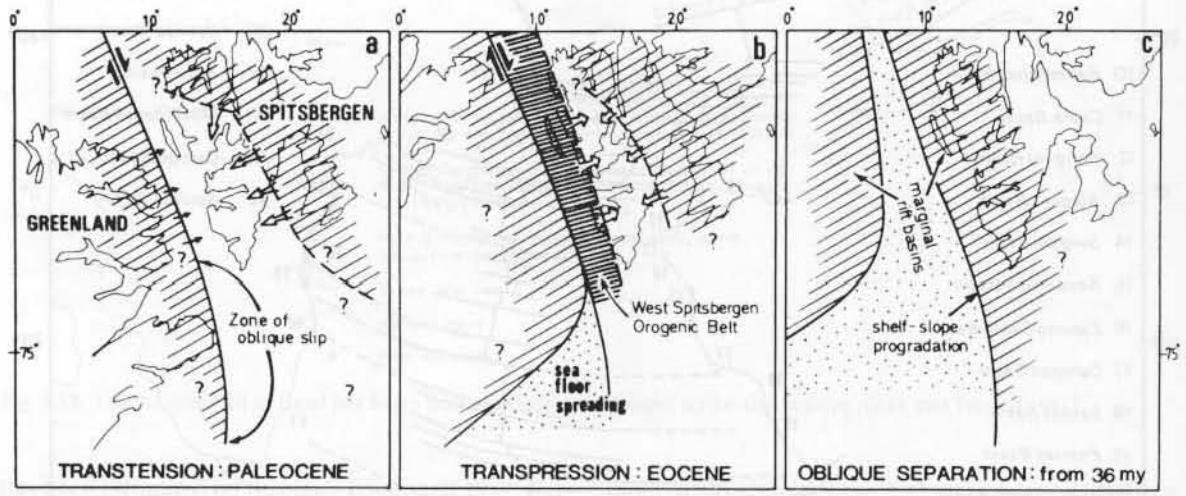


Fig. 9.56. Evolution of the central tertiary basin of Spitsbergen, adjacent to an active transform fault. Sediment was derived from Greenland until this source was moved away along the fault. (Steel et al. 1981)

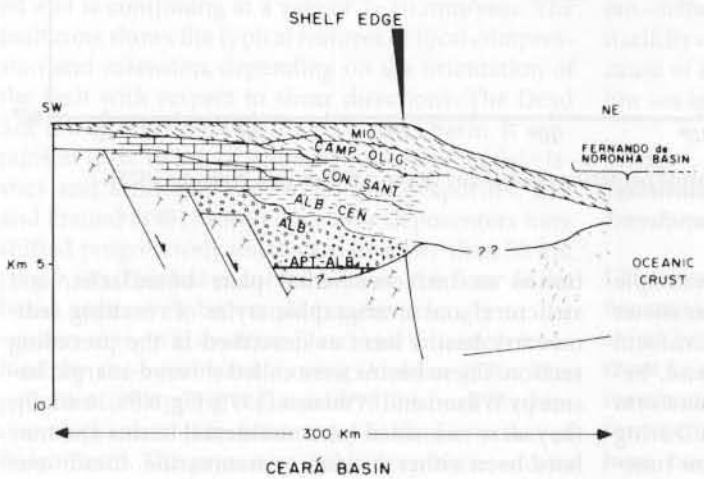


Fig. 9.57. Ceará Basin, Brazil, showing the position of the transform fault beneath the shelf edge, and its burial during the Cretaceous. (Kumar 1981)

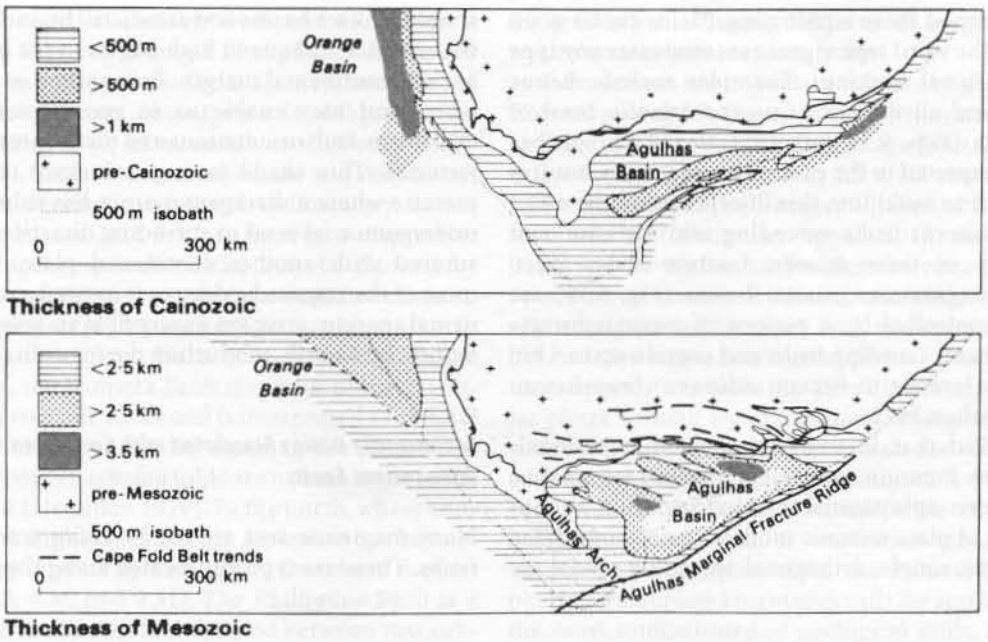


Fig. 9.58. Sediment distribution in Agulhas Basin, off South Africa. The Agulhas Marginal Fracture Ridge is a major transform fault that was active during the Cretaceous. (Wilson and Williams 1979)

varied between transpressive and transtensile, depending on the positions of local rotation poles. Once the continental margins passed each other, the basins faced oceanic crust and continued to develop as extensional continental margins. Good examples include:

1. Central Tertiary Basin, Spitsbergen, now bounded to the west by the Hornsund Fault and initially formed along a transform fault separating Svalbard from Greenland (Steel et al. 1981, 1985)
2. Barrierinhas, Ceará, and other basins, northern Brazil, originally bounded by the southern West African coast (Francheteau and Le Pichon 1972; Ponte and Asmus 1976; Kumar 1981; Zalan, et al. 1985)
3. Agulhas Basin, off the south coast of South Africa, separated from the Falkland (Malvinas) Plateau of the South American plate along the Falkland transform (Dingle 1973; Du Toit 1979)

The two-stage evolution of these basins is illustrated in Figs. 9.56–9.58. Initially, sedimentary sources on both sides of the basin were active (Fig. 9.56). Following separation, the marginal fracture ridge may have acted as a steep seaward-facing escarpment or as a sediment dam that ponded the sediment until it was buried (Fig. 9.57). Rift faults were reactivated as strike-slip faults. In the case of Agulhas Basin, dextral shearing reactivated old westnorthwest–eastsoutheast structures of the early-Mesozoic Cape Fold Belt, and Cretaceous isopachs show a series of depocenters and intervening sediment sources oriented oblique to the continental margin (Fig. 9.58), whereas Cenozoic isopachs indicate a seaward-thickening sediment wedge.

Francheteau and Le Pichon (1972) and Wilson and Williams (1979) showed that, on divergent margins, there is a range of different basin types, of which the simple orthogonal spreading type discussed in Section 9.3.1 (Table 9.2) and the sheared-margin type discussed above are but two of nine possible types. These are tabulated in Fig. 9.55, which classifies the orientation of the basin axis with respect to the orientations of the continental margin and of transform faults and their associated fracture zones.

Transcurrent movement on transform faults only occurs between the offset spreading centers, but minor, vertical fault movement continues across the fracture zone that represents the fault as the oceanic crust moves away from the spreading center. This is because the two sides of the fault are different ages and, therefore, at different points on the cooling curve (Delong et al. 1979). Fracture zones form distinctive topographic features on the ocean floor. They lie on small-circle trajectories about the rotation pole and serve a useful function as markers, preserving the tracks of the divergent margins as they move away from each other. Fracture zones may form transverse structural highs where they reach the continental margins and may influence sediment distribution (isopachs) and facies (Francheteau and Le Pichon 1972). This is particularly well seen where linear basins have formed parallel to transform faults at high angles to continental margins. Wilson and Williams

(1979) termed these taphrogenic basins (not a good term, as the word taphrogenic encompasses any type of extensional faulting). Examples include Benue Trough and all the basins on the Atlantic coast of Argentina (Figs. 9.54 and 9.55). Benue Trough has been interpreted in the past as an aulacogen, but it is important to build into this interpretation the effect of transcurrent faults extending into the continent from two or three Atlantic fracture zones (Sect. 9.3.1.4). Argentinian coastal basins (Fig. 9.59) are graben controlled by a pattern of westnorthwest–eastsoutheast-trending faults and contain up to 9 km of upper Jurassic to Recent sediment (Francheteau and Le Pichon 1972).

It is clear that, in analyzing ancient extensional-(divergent-) margin sedimentary basins, we must be much more sophisticated than in the past in our application of plate-tectonic models. Instead of relying on a single, simple, orthogonal-spreading model, we should look for basins and associated bounding faults that trend at oblique to high angles to the presumed ancient continental margin. Recognition of a particular trend may enable us to reconstruct ancient transform-fault orientations and plate-spreading trajectories. This would be a particularly instructive exercise where a divergent margin has subsequently undergone a reversal in spreading direction and has sutured with another continental plate. However, most of the required evidence is located at the continental margin, a region susceptible to severe deformation by A- or B-subduction during collision.

9.3.3.4
Sedimentary Basins Associated with Convergent-Margin Transcurrent Faults

Many magmatic arcs are cut by major transcurrent faults. These are typically located along the arc itself,

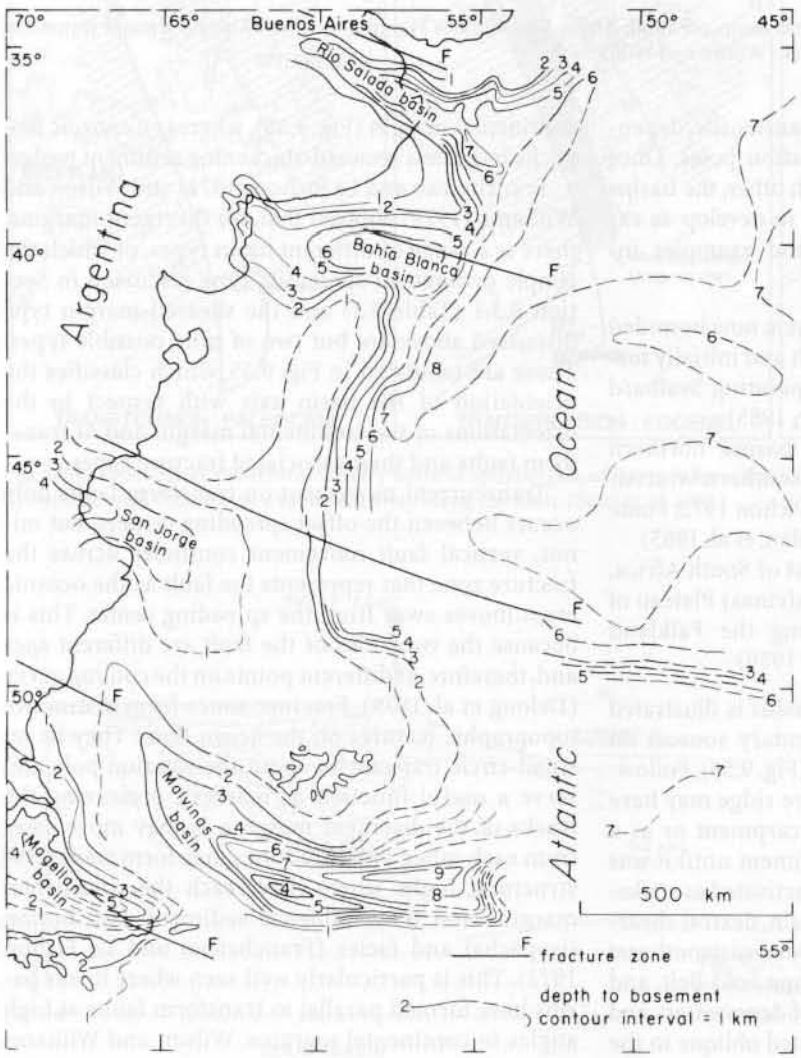


Fig. 9.59. Sediment distribution on the Atlantic margin of Argentina. Note the relationship of isopach trends to fracture zones. (Francheteau and Le Pichon 1972)

where the crust is relatively weak and movement is lubricated by rising magmas. Fitch (1972) and Dewey (1980) showed that these faults are a normal part of arc evolution where convergence is oblique. In the case of neutral or extensional arcs (Dewey 1980; Sect. 9.3.2), the faults are transtensile and may be expected to develop extensional sedimentary basins (intra-arc basins; Sect. 9.3.2.4). Even in transpressive systems, upthrust fault blocks may be flanked by isostatic depressions that could accumulate sediment.

Few sedimentary basins have, in fact, been recorded along transcurrent faults of modern arcs. In the Sunda arc, the Sumatra Fault displaces modern alluvium and volcanic rocks and is interpreted as a transtensile rift. Neogene rocks flanking the fault show characteristic *en echelon* folds at an acute angle to the fault trend (Hamilton 1979). To the north, where convergence is highly oblique, the Andaman Sea has developed by interarc spreading and transform motion (Figs. 9.18, 9.30, and 9.31). The Philippine Fault is a left-lateral strike-slip fault located between two subduction zones. It is flanked by several strike-slip basins (Ringebach et al. 1993; Smith and Landis 1995).

9.3.3.5
Sedimentary Basins Associated with Suture-Zone Transcurrent Faults

Sedimentary basins associated with suture-zone transcurrent faults are discussed below (Sect. 9.3.4).

9.3.4
Basins Associated with Suture Zones

9.3.4.1
Introduction

Sutures are complex orogenic zones where two buoyant crustal blocks have joined by subduction of intervening oceanic crust. They are characterized by intense structural deformation, regional metamorphism, plutonism, and distinctive styles of sedimentary basins. A simple classification of suture zone basins is impossible because of the variations inherent in the suturing process. Sutures do not only develop between major continental plates but also among continents, island arcs, and microcontinents. Convergence may be orthogonal or, as is more commonly the case, oblique, with the development of major strike-slip faults or intracontinental transforms. Arcs may be facing toward or away from the colliding continent. In either case, a backarc basin may collapse by subduction first, leading to an early orogenic episode predating the terminal collision between major continental masses (Fig. 9.21). Most important of all, collisions between irregular margins can lead to episodic or nearly continuous diachronous orogeny as the suture zone grinds inexorably shut.

Suture zones are important features of global continental geology. Burke et al. (1977) summarized the geology of 57 of them ranging in age from those still active back to some as old as 2.5×10^9 years. Examination of the more well-known Phanerozoic sutures, such as the Caledonian–Appalachian, the Cordilleran, and the Alpine sutures demonstrates that most suture zones consist of rotated, mangled, and partially amalgamated terranes and microplates caught between the larger continental plates. Only a few sutures appear to involve the direct interaction of major plates without significant intervening microplate activity. Examples include the Urals and the Indian continental block, which forms the south side of the Indus suture (the Asian continent, with which India collided, is composed of numerous terranes and microplates assembled since the Paleozoic). Deciphering microplate kinematics calls for application of the most sophisticated of geological skills, of which sedimentary-basin analysis is only one component. Some of the special problems of terrane analysis are considered in Section 9.3.5.

9.3.4.2
Basic Suture Model

A collision model for irregular continental margins was discussed by Graham et al. (1975), Şengör (1976), and Dewey (1977) and is illustrated in Fig. 9.60. Collision requires subduction on at least one margin (it could occur on both). The over-riding plate, therefore, commonly has a curving margin, reflecting the presence of the volcanic arc and subduction zone. Collision effects commence as the first projection of buoyant crust on the downgoing plate enters the subduction zone. Opposite the point of collision, extensional faults, mafic dykes, and alkaline volcanism extend into the over-riding plate perpendicular to the continental margin, and strike-slip faults develop as the rocks are squeezed and move laterally to compensate for the compression. The downgoing plate becomes intensely sheared and thrust faulted in the subduction zone and may be folded back on itself in the form of giant nappes. Ophiolite slabs and nappes are typically emplaced during early collision stages, followed by underthrusting of allochthons of continental cover rocks or basement as the downgoing plate is rammed into the subduction zone. Slivers of continental material are squeezed sideways along transcurrent faults, and a network of transcurrent faults also typically develops in either plate at an angle of about 45° to the suture zone. Crustal shortening may be on the order of several hundred kilometers, leading to crustal thickening, uplift, and the

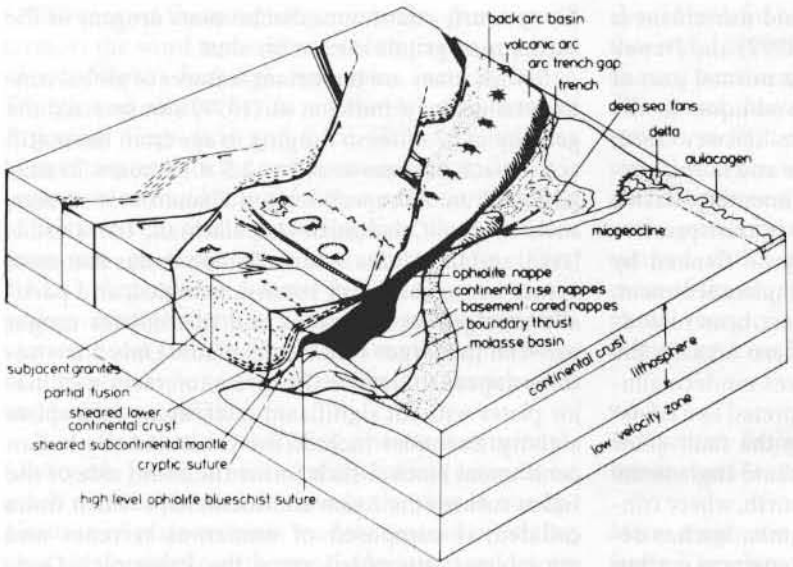


Fig. 9.60. Block-diagram model of tectonic and sedimentary evolution of a suture zone developed between irregular continental margins. (Dewey 1977)

production of vast quantities of clastic debris. The geometry of sutures is highly variable, depending on the broad architecture of the colliding masses (Coward and Ries 1986). Some sutures, such as the Indus suture in Pakistan, are subvertical (Andrews-Speed and Brookfield 1982). Others, such as the Iapetan suture of the southern Appalachians, are subhorizontal and hidden beneath allochthonous overthrust sheets (Cook et al. 1979; Cook and Oliver 1981). The differences may be partly due to the level of exposure.

Sediment shed from suture zones is deposited in four types of sedimentary basins:

1. **Proforeland** basins, also termed **peripheral** or **foredeep basins** in earlier literature (Dickinson 1974). Detritus shed from the rising orogen at the point of suture is shed onto the downgoing plate (Sect. 9.3.4.7). The latter becomes depressed into a moat by downbending and, later, by the weight of stacked nappes or thrust sheets. Initially, the basin may be under deep water and accumulates thick piles of olisthostromes and sediment-gravity-flow deposits. Later, as suturing continues, the foredeep becomes shallower, and depositional environments change to nonmarine. The rising orogen may be flanked by giant alluvial fans that become tilted and deformed. Longitudinal trunk rivers may flow away from the suture along the axis of the foredeep. Intrabasinal, syndepositional, tectonic features are characteristic of these basins.
2. **Remnant ocean basins.** Along strike, the foredeep may pass into embayments between approaching crustal blocks. These serve as sedimentary sinks and remain undeformed long after suturing has reached an advanced stage elsewhere (Sect. 9.3.4.8). In fact, they may never be deformed and may preserve a floor of oceanic crust representing the downgoing plate deeply buried beneath a thick blanket of sediment shed into the basin from all sides. The principal sediment source is typically the suture zone itself, from which submarine fans, deltas, and alluvial plains prograde in turn, longitudinally down the axis of the embayment (Fig. 9.60). Secondary sources may include the arc or subduction complex of the over-riding plate, and mature carbonates, clastics, or basement rocks from the downgoing plate, all of which feed transversely into the embayment. Terminal stages of embayment filling may be lacustrine.
3. Behind the volcanic arc, there may be a **retroforeland** or **retroarc basin** (Sect. 9.3.2), although this may be modified by the graben and transcurrent-fault tectonics of the over-riding plate. Backarc basins may develop in early stages of collision but typically are closed prior to terminal suture, as discussed in Section 9.3.4.4.
4. Associated graben and transcurrent fault basins. Strain may be transmitted hundreds or thousands of kilometers into continents on either side of the suture by systems of faults. These may develop graben similar to the rift basins discussed in Section 9.3.1, and transcurrent-fault basins similar to those described in Section 9.3.3.

At this point, we should say a few words about the widely used terms **flysch** and **molasse**. Both are old Swiss stratigraphic terms but acquired a general meaning in early geosynclinal theory. Flysch refers to deep-water clastic sediments deposited under what were described as preorogenic or early-orogenic conditions. It commonly passes up, stratigraphically, into molasse, a shallow-marine to nonmarine deposit

formed under late-orogenic to postorogenic conditions (Hsü 1970; Reading 1972; Van Houten 1973, 1981). With our knowledge of plate tectonics, we can now see that orogeny encompasses enormous variability in plate-tectonic settings and processes. Most flysch and molasse, including those in Switzerland, occur in what are now known to be remnant ocean basins and proforeland (peripheral) basins (Graham et al. 1975; Homewood and Caron 1982; Homewood et al. 1986; Ingersoll et al. 1995). However, the sedimentary facies comprising flysch can occur in many tectonic settings, including some that are not orogenic, in the original sense (Reading 1972). Similarly, Miall (1981b) showed that thick alluvial deposits similar to the original Swiss molasse can occur in at least 12 distinct, plate-tectonic settings, only two or three of which would be recognizable as synorogenic to postorogenic geosynclinal settings in pre-plate-tectonic terminology. Accordingly, most modern workers recommend abandoning the terms flysch and molasse (Bally and Snelson 1980; Miall 1984). In the following sections, we examine three variations of the basic suture model with reference to actual ancient examples.

9.3.4.3 *Arc–Arc Collision*

The Molucca Sea, between Indonesia and the Philippines, overlies one of the most curious permutations of arc geology. Here, two arcs, facing each other, are both subducting the same oceanic plate (Hamilton 1979). Figure 9.61 is based on a plot of earthquake activity and shows two downgoing slabs of oceanic crust. Seismic-reflection profiles show that the west-dipping Sangihe subduction zone is becoming inactive and that a new trench is developing on the west side of the arc, so that, shortly, the Sangihe plate will start to be subducted beneath the Halmahera plate.

Recognition of this sequence of events in an ancient orogen would require precise age determinations of tectonic events and sedimentary rocks within three parallel subduction complexes (Halmahera, old Sangihe, and new Sangihe). Late Cenozoic examples of similar events are exposed in the Japanese islands. For example, a collision between the central Honshu arc and the Izu–Bonin arc has taken place west of Tokyo. Resulting volcaniclastic fan–delta and submarine-fan successions have been uplifted and are well exposed in the region around Mount Fuji (Ito 1985, 1987; Ito and Masuda 1986).

9.3.4.4 *Collapse of Outboard Arcs*

Where subduction commences at the edge of a mature ocean, rapid sinking of old, cold, oceanic crust favors the development of extensional arcs and the opening of backarc basins. However, with time, younger and younger oceanic crust may be subducted, and the arc may change to a contractional-style arc (Dewey 1980). The mechanism is discussed in Section 9.3.2 and illustrated in Fig. 9.21. Note that this discussion relates to arcs that develop at the edge of the continental block, facing the ocean, as distinct from the oceanic arcs described in the next section.

The result of the subduction of progressively older oceanic crust is that marginal seas first open and then close, with the development of a major orogenic welt, and sedimentation patterns are profoundly altered. It

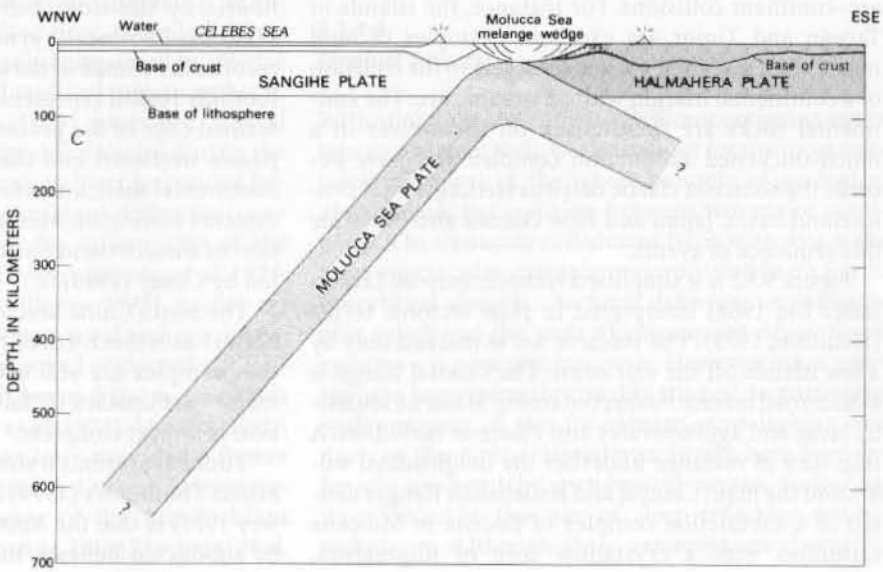


Fig. 9.61. Arc–arc collision in the Molucca Sea. (Hamilton 1979)

is now widely recognized that this process commonly occurs during the closing-ocean phase of a Wilson cycle. It leads to orogenic episodes that predate by tens or hundreds of millions of years the orogeny of the terminal suture. Ancient examples where this sequence of events has been proposed include the Paleozoic of the Western Cordillera, United States (Burchfiel and Davis 1972; Churkin 1974; Dickinson 1981), the northern Appalachians ("Proto-Dunnage back-arc basin"; Dewey 1974; St. Julien and Hubert 1975; Stockmal et al. 1987; Williams 1995), and the southern Appalachians (Hatcher and Odom 1980). In the southern Andes, a major backarc basin extending the full length of Peru and Chile opened in the Triassic–early Cretaceous and was then closed by compressive tectonism, possibly as a result of an increase in convergence rates between the Pacific and South American plates but without the collision of significant allochthonous masses with the Andean mainland (Dalziel et al. 1974; Dalziel 1986). In west China, remnant or relict backarc basins may be preserved within the complex of the Himalayan Orogen (Hsü 1988), as discussed briefly in Section 9.3.2.

Reconstruction of these backarc basins in orogens is commonly very difficult, because the process of final suturing results in extreme deformation, metamorphism, and igneous intrusion, possibly with the superimposition of major lateral displacement and the accretion of interposing, unrelated terranes. For this reason, most such interpretations, including all those mentioned above, are characterized by controversy.

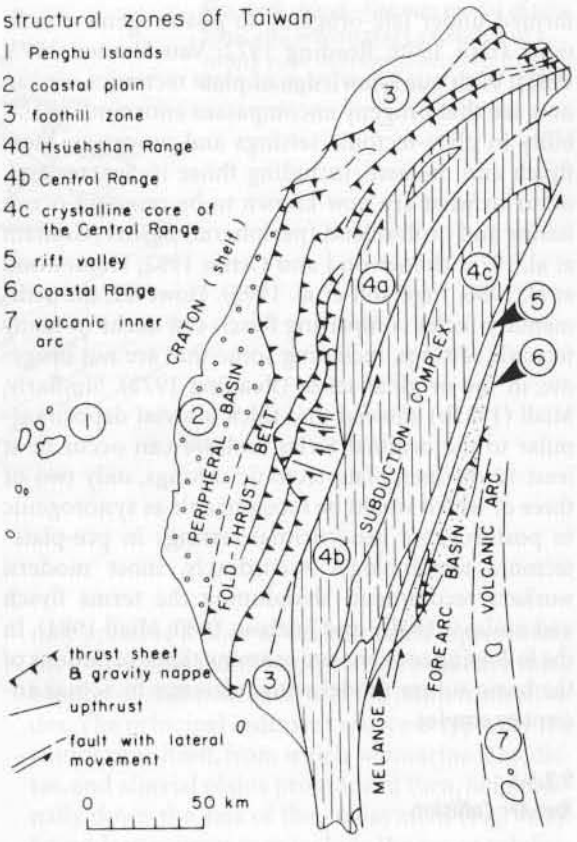


Fig. 9.62. Simplified tectonic map of Taiwan. (After Biq 1964; Hamilton 1979)

9.3.4.5 *Collisions of Continents with Arcs*

The west-Pacific region is replete with examples of arc–continent collisions. For instance, the islands of Taiwan and Timor are excellent examples of land masses that owe their existence largely to the collision of a continental margin with an oceanic arc. The continental rocks are thrust back on themselves in a much-thickened subduction complex and have become the source of clastic detritus feeding a local proforeland basin. Japan and New Guinea also illustrate this sequence of events.

Figure 9.62 is a simplified tectonic map of Taiwan (after Biq 1964) interpreted in plate-tectonic terms (Hamilton 1979). The volcanic arc is marked only by a few islands off the east coast. The Coastal Range is a deformed forearc basin containing Miocene andesitic lavas and agglomerates and Pliocene turbidites. A thin slice of melange underlies the longitudinal valley, and the main Central and Hsuehshan Ranges consist of a subduction complex of Eocene to Miocene turbidites with a crystalline core of migmatites, schists, marbles, and argillites of late-Paleozoic and Mesozoic age. The core represents deformed rocks of the Chinese continental margin. Along the western edge of the mountains is a fold–thrust belt consisting of Neogene shallow-marine to nonmarine sediments floored by Mesozoic shelf sediments and capped by a Pliocene–Pleistocene synorogenic conglomerate that records the climax of the arc–continent collision. This foothills region represents the syndepositionally deformed edge of the proforeland (peripheral) basin. It passes westward into the coastal plain and Chinese continental shelf, underlain by flat-lying Neogene sediments resting on Mesozoic shelf rocks. The evolution of this foreland basin has been examined in detail by Covey (1986).

The South China Sea, south of Taiwan, is a remnant ocean basin (Sect. 9.3.4.8). There, the arc and subduction complex are still active, and the occurrence of major earthquakes in Taiwan suggests that the collision is not yet complete.

Timor represents a similar plate-tectonic configuration. Hamilton's (1979) interpretation (see also Dewey 1977) is that the Australian plate is beginning to be subducted beneath the Sunda arc. At Timor, the

first projection of the continent entered the north-dipping subduction complex during the Neogene. Most of the island is underlain by imbricate masses and melange of Australian shelf sediments of Permian to early-Tertiary age. Faulted into this complex are Miocene to Pleistocene marine marl and shale that were probably deposited in small accretionary basins (Dickinson and Seely 1979) on the subduction complex. Subduction is still active, and Pleistocene reefs have been uplifted to heights of 800 m above sea level. Hamilton (1979) suggested that choking of the subduction zone by continental rocks may be leading to a reversal in polarity of the arc, so that a south-dipping zone may be developing in the Banda Sea to the north of the present volcanic arc. Within about 10 m.y., the Timor melange masses may have become the roots of south-verging nappes sliding southward into a retroforeland basin developed over the Australian continental shelf.

The Taconic Orogen of Quebec and Newfoundland is a well-exposed, ancient example of a continental margin involved in an arc–continent collision. Cambrian to lower Ordovician strata constitute a divergent-margin shelf–slope–rise assemblage facing the Iapetus Ocean (the Humber Terrane of Williams 1995). In the latter part of the early Ordovician, closure of the ocean commenced, with the development of a subduction complex to the southeast. Calc-alkaline volcanics fed tuffaceous material northwestward, and this marked the beginning of a reversal in sediment-transport directions. The divergent-margin prism was progressively covered by northwesterly prograding sediment-gravity flows and olisthostromes derived by submarine collapse of advancing thrust and nappe sheets consisting of ophiolites, arc volcanics, and basin-fill sediments (Dunnage Terrane). By the end of the Ordovician or early Silurian, closure and uplift of this basin was complete. The Taconic Orogeny, as summarized here, extended over a period of about 30 m.y. It was accompanied by intense structural deformation and regional metamorphism.

St. Julien and Hubert (1975) proposed several changes in arc configuration and polarity during the Ordovician. This is a difficult subject to resolve because of structural complexity. Many different reconstructions have been made for other parts of the Appalachian–Caledonian Orogen (Strong et al. 1974; Hatcher and Odom 1980; Williams 1995). Earlier reconstructions depended on structural and geochemical variations of the volcanics and obducted ophiolites, a discussion of which is beyond the scope of this book (Dewey 1974; Strong et al. 1974). More recently, deep-reflection-seismic data have provided a firmer basis for tectonic interpretation. Current interpretations of the Canadian sector of the Appalachians (Coleman-Sadd 1982; Keen et al. 1986; Stockmal et al. 1987) invoke the collision of a west-verging arc (proto-Dunnage terrane) with the North American divergent margin. In the Quebec area, the Paleozoic divergent-margin sediment prism, with its Precambrian (Grenville) basement, entered an east-dipping subduction zone. The uplift and erosion of this margin during the early Ordovician are "interpreted as effects of a cratonward migrating flexural bulge preceding an east-dipping subduction zone" (Stockmal et al. 1987). The Dunnage Terrane is represented, in part, by the Ascot–Weedon volcanics in Quebec, and is interpreted as the remnant of an Iapetan island arc-and-backarc complex (Strong et al. 1974; Stockmal et al. 1987). To return to the west Pacific for a modern analogy, in this reconstruction, the Dunnage arc is analogous to the Banda arc, and the continental margin of North America is comparable to the leading edge of the present Australian plate.

A general model of arc–continent collision can be developed from the examples described here. Collision results in a continentward migration of facies belts and tectonism. For example, water depths increase under the imposed load of the arc, slope failure and faulting of the continental margin begin, the forebulge migrates landward, and turbidites from the rising allochthon arrive, followed by thrust sheets of the allochthon itself. Documentation of these events by careful study of the structural and stratigraphic record can provide information regarding rates of plate convergence. For ancient orogens, this could add a useful component to a reconstruction of plate kinematics. Bradley and Kusky (1986) reviewed this topic and provided data documenting the convergence rate during the Taconic Orogeny of New York (Fig. 9.63). They were able to obtain reasonably consistent data demonstrating a convergence rate of 2–3 cm/year for the Taconic Orogeny.

9.3.4.6 *Terminal Sutures*

With all the arcs, terranes, and microcontinents swept into an orogenic welt, the stage is set for the most awe-inspiring event in the whole panoply of geological phenomena: the collision between two major continents. The Cenozoic collision of India with Asia is the most spectacular example presently visible on Earth. Everything about it – tectonic deformation, topographic relief, and the scale of consequent depositional systems – is on a gigantic scale. However, other sutures have been more thoroughly studied. In particular, such orogens as the Caledonian–Appalachian and parts of the Alpine–Himalayan suture have been intensely explored by geophysical means, including, most recently, the use of deep-reflection-seismic techniques. Although these examples are character-

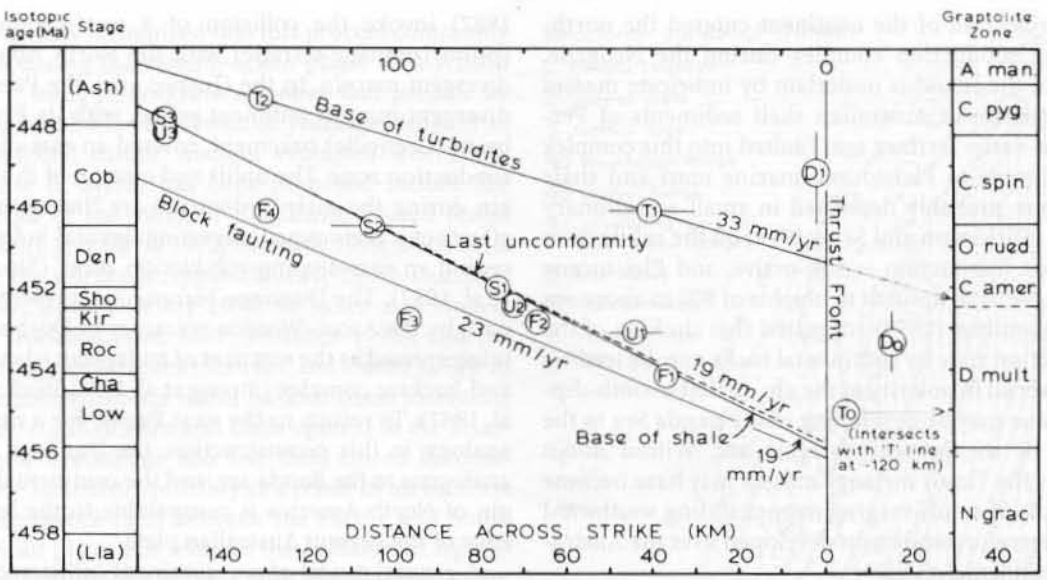


Fig. 9.63. Time-distance plot for events in the Taconic foreland of New York. *Circled letters* and *numbers* are data points. Block faulting is the normal faulting of the continental margin. The last unconformity is that associated with uplift of the migrating forebulge. Turbidites are those associated with development of a deep-water basin and derivation from the allochthon. (Bradley and Kusky 1986, reprinted with permission of University of Chicago Press)

ized by the welding of many small terranes prior to the welding of major continental blocks, many similarities between the resulting crustal fabrics are emerging.

A review of seismic data suggested to Allmendinger et al. (1987b) that continental crust involved in collision orogens is characterized by four basic types of crustal fabric, three of which are directly related to the orogenic process (Fig. 9.64). Zone 1 represents the little-deformed craton of continental interiors. Shallow, layered reflections represent Phanerozoic and possibly Proterozoic cover rocks. Zone 2 corresponds to the region of thin-skinned tectonism of the foreland fold–thrust belt. This may extend back into zone 3, a region of relatively steeply dipping reflections that penetrate deep into the crust at angles typically of about 35°. Most of the reflections dip away from the continental interior although, in some areas, the opposite sense of dip occurs. This zone commonly defines the outboard edge of the continental-plate overthrust during the collision and may become a region of detachment during subsequent extensional tectonism. Zone 4 is characterized by "numerous horizontal and gently-dipping reflections in the middle and lower crust and a high-amplitude, laterally continuous zone of reflections at or near the Moho" (Allmendinger et al. 1987b). The fabric develops by ductile extension (Fig. 9.38; Sect. 9.3.2.6).

The structural geology of the Himalayas and adjacent areas of southwest China is shown in Fig. 9.65 (also Fig. 9.1), and this orogen is shown in a broader context in Fig. 9.66. This region serves as a type example for all the major tectonic effects of terminal sutures, including the various classes of sedimentary basin referred to at the beginning of this section.

Reconstructions of the northward drift of the Indian subcontinent based on oceanic paleomagnetic evidence (McKenzie and Sclater 1971; Powell 1979) indicate that India began to collide with Asia in the latest Cretaceous and Paleocene; this is confirmed by the age of ophiolites and associated sediments along the Indus Suture (Gansser 1964, 1980; Powell 1979; Brookfield and Reynolds 1981; Andrews-Speed and Brookfield 1982; Tapponnier et al. 1986). North of the suture lie the poorly known mountains of Tibet and western China, which were assembled by the accretion of small plates during the Paleozoic and Mesozoic (Zhang et al. 1984). Subduction and underthrusting of the Indian plate and displacement of central and east Asia have accumulated at least 1000 km, possibly as much as 1500 km, of crustal shortening (Molnar and Tapponnier 1975; Powell 1979; Tapponnier et al. 1986) and, in the Miocene, this led to rapid Himalayan uplift, which continues.

The effects of the collision include the following:

1. Intense tectonic deformation, metamorphism, and igneous activity (most of the details of this are beyond the scope of this book, except where they are relevant to considerations of sediment age, detrital sources, and tectonic position)

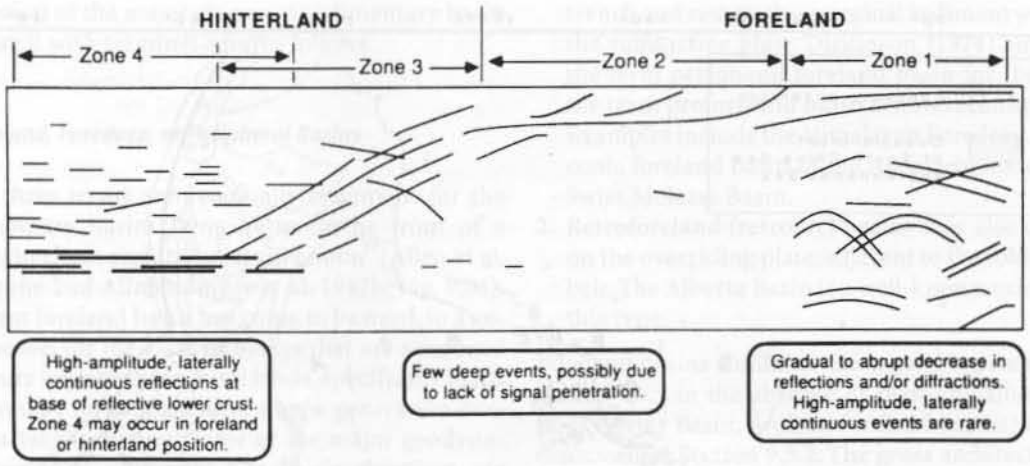


Fig. 9.64. A cartoon cross section through a typical continental megasuture, showing four characteristic crustal fabrics revealed by deep-seismic-reflection surveys. Zone 4 represents the structural style of the Basin-and-Range province of the United States. (Allmendinger et al. 1987b, reproduced with permission of Geological Society of America)

2. Crustal thickening and uplift of Tibet into a vast plateau
3. Deformation of island arc complex(es) at the suture zone
4. Development of a gigantic proforeland basin south of the fold–thrust belt, now occupied by the Indus, Brahmaputra, and Ganga rivers and their tributaries
5. Spillover of Himalayan detritus into remnant ocean basins southwest (Indus cone) and southeast (Bengal fan) of India
6. Extension of deformation far into China, Mongolia, and Siberia along major strike-slip faults and graben, possibly including major thrusts penetrating deep into the crust

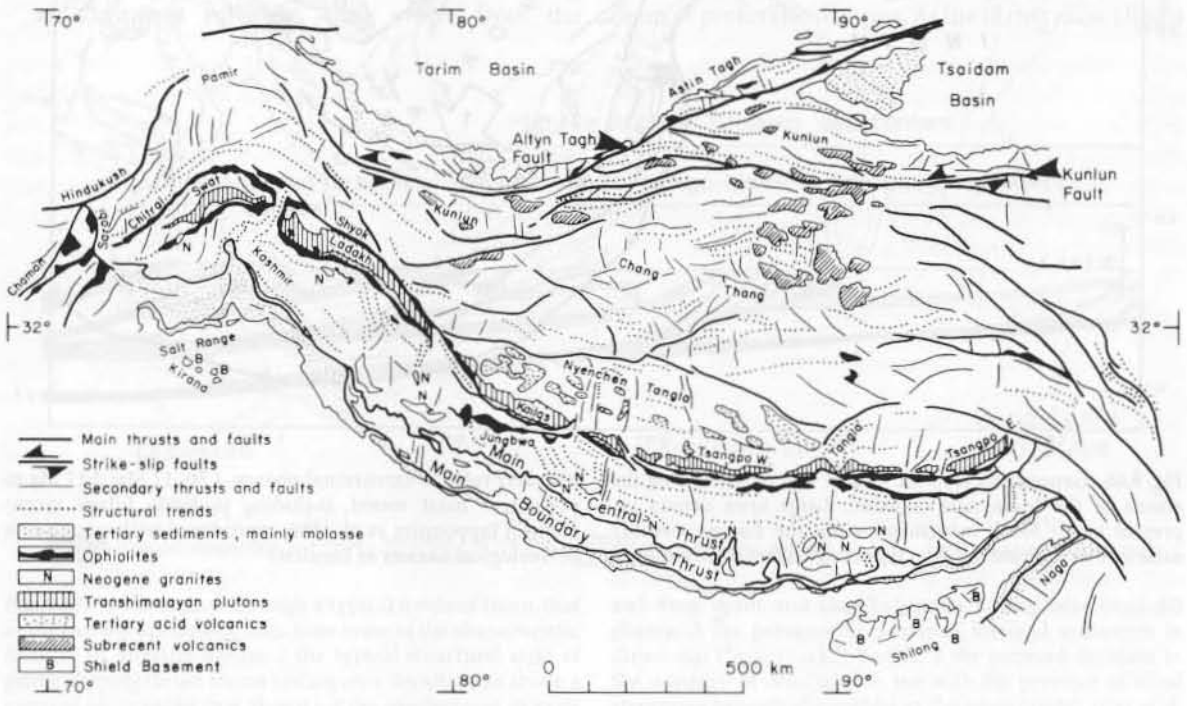


Fig. 9.65. Tectonic sketch map of the Himalayan region. (After Molnar and Tapponnier 1975; Gansser 1980)



Fig. 9.66. Generalized tectonic map of east Asia, showing the effects of the India–Asia collision. *Large open arrows* represent major block movements since the Eocene. *Smaller solid arrows* indicate the direction of extensional movements. *Numbers* refer to extensional phases: *1* 50–17 Ma; *2* 17 Ma to present; *3* most recent, including probable future movements. (Tapponnier et al. 1986, reproduced with permission of Geological Society of London)

Discussion of the major classes of sedimentary basin associated with terminal sutures follows.

9.3.4.7 *Proforeland, Foredeep, or Peripheral Basins*

These three terms are essentially synonyms for the "sedimentary basins lying between the front of a mountain chain and the adjacent craton" (Allen et al. 1986; zone 2 of Allmendinger et al. 1987b; Fig. 9.64). The term foreland basin has come to be used, in a generic sense, for the class of basins that are produced by flexure under supracrustal loads, specifically loads emplaced by thrusting and/or nappe generation during crustal shortening. Some of the major geodynamic principles of foreland-basin development are discussed in Section 7.4. The term **proforeland basin** was suggested for foreland basins situated on the downgoing plate by Johnson and Beaumont (1995) and seems likely to become the preferred name for this class of basin. An excellent recent research synthesis on foreland basins was published by Allen and Homewood (1986), and a more recent volume, containing several numerical-modeling studies, was compiled by Dorobek and Ross (1995). Figure 9.67 is a typical cross section through a foreland basin, illustrating most of the characteristic elements of this class of sedimentary basin.

There are two major types of collision-related foreland basins (Miall 1995).

1. There are those that develop in the forearc during arc–continent collision. They evolve from the trench and rest on the marginal sediment wedge of the subducting plate. Dickinson (1974) employed the term **peripheral foreland basin** for these, but the term **proforeland basin** is now recommended. Examples include the Himalayan Foredeep, the Taconic foreland basin of the Appalachians, and the Swiss Molasse Basin.
2. **Retroforeland (retroarc) basins** may also develop on the overriding plate, adjacent to the fold–thrust belt. The Alberta Basin is a well-known example of this type.

Foreland basins similar to the second type also form behind arcs in the absence of plate-collision events (e.g., Sevier Basin, Andean foreland basins) and are discussed in Section 9.3.2. The gross architectures of all types of foreland basins depend, in part, on the type of crust underlying the fold–thrust belt (Stockmal et al. 1986; Stockmal and Beaumont 1987). Where the continent consists of thick crust, its flexural strength is high, and a wide, shallow basin is produced. The Alberta Basin is a good example. It was built over Precambrian shield, on which rests a thick Paleozoic divergent-margin sediment wedge. During most of its history, the basin was under no more than a few hundred meters of water, and shelf and nonmarine sediments predominated. By contrast, where the thrust mass impinges on young, thin crust or a sediment-starved trench, there will at first be little subaerial relief, and the basin will remain under deep water, undergoing slow pelagic sedimentation. This condition is typical in the early stages of the development of proforeland basins. As the thrust mass climbs

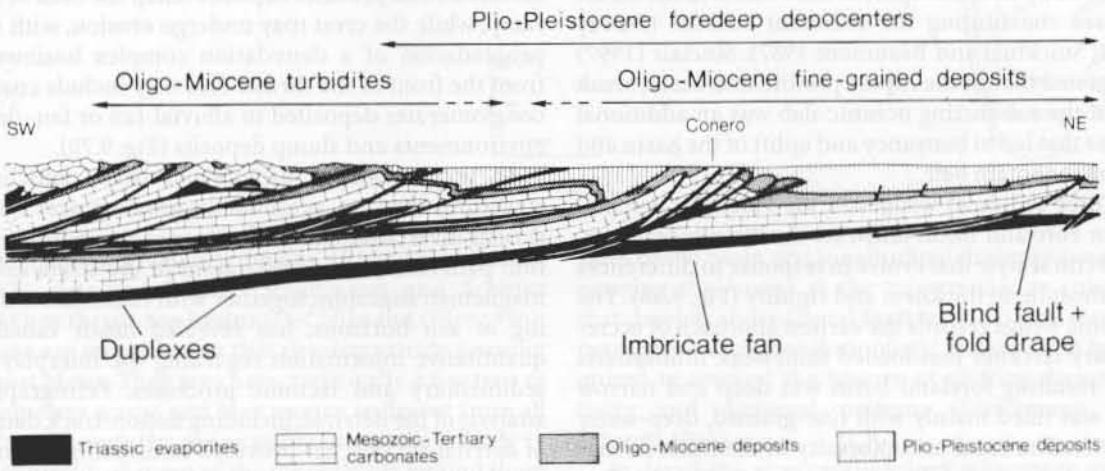


Fig. 9.67. Cross section through a typical foreland basin, that of the eastern Apennines, Italy. Note some of the characteristic features of foreland basins: *1* the typical structural style of gently dipping thrust sheets resting on a décollement above a ramp of older rocks (not shown); *2* the involvement of early basin-fill sediments (Oligocene–Miocene) in deformation, and their uplift and cannibalization during later basin-fill phases; *3* the presence of proximal foreland sediments in thrust-top ("piggyback") basins; *4* the outward decrease in the intensity of deformation, but with the presence of blind structures beneath the middle of the basin (*right*). (Ori et al. 1986)

the margin of the subducting plate, a subaerial landmass appears, and then terrigenous sediments start to flood into the basin, building a thick prism of deep-water sediments (one type of flysch). The collision of the northern Australian continental margin with the Indonesian arc is currently in this stage (Fig. 9.18). The island of Timor consists mainly of an allochthonous complex of Permian to Neogene rocks that originally formed part of the Australian continental margin and has been riding back over the Australian shelf and slope since the Miocene or Pliocene. The Timor Trough to the south is a growing proforeland basin that is presently receiving mainly fine-grained clastic sediments (Hamilton 1979; Audley-Charles 1986).

The isostatic response to thrusting of the supracrustal load up the ramp of the continental margin is the construction of a mountain range from which vast quantities of terrigenous debris are shed into the basin. The rate of sedimentation eventually exceeds the rate of flexural subsidence, and depositional environments become progressively shallower. Most foreland basins are ultimately filled by thick nonmarine sediment wedges (the classic molasse). Progradation and onlap onto the underlying plate are the characteristic stratigraphic patterns that result from these developments. The fold–thrust belt and foreland basin (coastal plain) of western Taiwan (Fig. 9.62) illustrate this later stage in proforeland evolution.

The transition from deep-marine to shallow-marine and nonmarine basins may record the passage of the colliding thrust wedge over the divergent margin of the under-riding continental plate, which would be followed by buoyant uplift of the basin and the thrust masses constituting the sediment sources (Dewey 1982; Stockmal and Beaumont 1987). Sinclair (1997) suggested that, in the Alpine proforeland basin, break off of the subducting oceanic slab was an additional factor that led to buoyancy and uplift of the basin and rising mountain belt.

Three sediment wedges in the Paleozoic Appalachian Foreland Basin illustrate the differences in architectural style that evolve in response to differences in lithospheric thickness and rigidity (Fig. 9.68). The Taconic wedge records the earliest approach of accretionary terranes that loaded thin, weak lithosphere. The resulting foreland basin was deep and narrow and was filled mainly with fine-grained, deep-water clastics. The basal unconformity in the Knox formation represents uplift of the forebulge. The Acadian Orogeny did not involve significant shortening. Accretionary terranes were thickened, and the lithosphere beneath the foreland underwent broad crustal flexure. The classic Catskill magnafacies was deposited at this time. Finally, during the Alleghenian Orogeny, thin thrust sheets were emplaced over strong, thick lithosphere, with resulting widespread deposition of relatively coarse clastics in shallow-water to nonmarine conditions (Tankard 1986).

This evolutionary pattern varies from basin to basin, but the general transition from deep-water pelagics through deep-water clastics (flysch) to shallow-water and nonmarine clastics (molasse) has been known for many years and formed part of the geosynclinal cycle of earlier authors (Dott 1974; Ingersoll et al. 1995). In detail, the succession may consist of several or many shallowing-up cycles, recording repeated episodes of tectonism (Miall 1978).

One of the unique characteristics of foreland basins is the style of syndepositional tectonism in most basins. Thrusts gradually step outward, resulting in widening of the fold–thrust belt (Dahlstrom 1970). The structural style varies in detail, depending on pre-existing basement fabric, including the presence or absence of suitable surfaces of décollement, such as salt beds (as in, for example, the Swiss molasse basin; Allen et al. 1986). The faults propagate from below the basin, gradually cutting through and uplifting the sediments of the basin itself. At the outermost limit of the fold–thrust belt, there may be a series of blind thrusts, which strongly deform the base of the basin fill but which may be invisible at the surface (Fig. 9.67). Many superb seismic transects across these structures have been obtained in some of the Alpine and Apennine foreland basins (Ori and Friend 1984; Ori et al. 1986; Ricci-Lucchi 1986). An example is illustrated in Fig. 9.69. The crests of the thrust ramps rise from deep to shallow water and eventually become exposed to subaerial erosion. Deep-water turbidites and prodelta deposits onlap the back of the ramp, while the crest may undergo erosion, with the progradation of a denudation complex basinward from the front of the thrust. This may include coarse conglomerates deposited in alluvial-fan or fan–delta environments and slump deposits (Fig. 9.70).

In the Himalayan proforeland basin, anticlinal structures that are, presumably, growing over blind thrusts divert drainage and influence channel-migration patterns. Careful correlation of the strata using magnetostratigraphy, together with radiometric dating of ash horizons, has revealed much valuable quantitative information regarding the interplay of sedimentary and tectonic processes. Petrographic analysis of the detritus, including fission-track dating of detrital zircons, has provided information regarding the uplift rates and unroofing history of the Himalayas (Burbank and Raynolds 1984; Johnson et al. 1985, 1988; Cerveny et al. 1988).

Another consequence of this syndepositional tectonism is the development of a class of small basins variously called **piggyback basins** (Ori and Friend

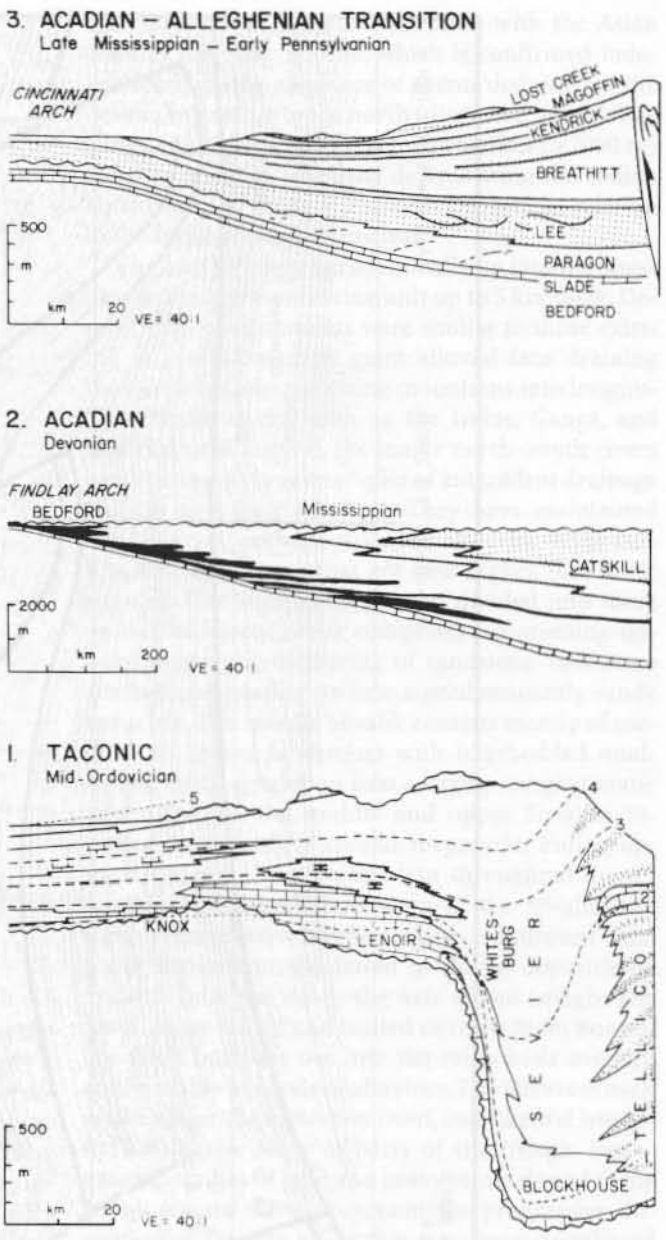


Fig. 9.68. Three clastic wedges in the Appalachian Foreland Basin, illustrating varying stratigraphic responses to the dynamics of foreland-basin construction. See text for discussion. (Tankard 1986)

1984), **floating satellite basins** (Ricci-Lucchi 1986, 1987), **detached basins** (Steidtmann and Schmitt 1988), or **thrust-top basins** (DeCelles and Giles 1996). These are minor basins that develop astride growing thrust plates. They are characteristically a few tens of kilometers across and may receive sediment from all sides, although the major sediment source is likely to be the uplifted ramp of the older thrust behind them (the fault block to the left in Fig. 9.69). Coarse submarine-fan, fan–delta, or fluvial sediments comprise the typical fill, depending on the structural elevation of the basin. Dispersal patterns are complex, with transverse drainage from the thrust ramps on both sides of the basin and longitudinal drainage possibly entering the basins at the topographic re-entrants that develop above lateral fault terminations. Careful petrographic and geochronologic research may be required to unravel the history of shifting thrust activity and dispersal patterns (Steidtmann and Schmitt 1988).

In detail, the structural geology adjacent to growing thrusts and drape folds may be extremely complex, with the development of several types of intraformational unconformities as the newly deposited sediments are subjected to continuing tectonism. Strata in the footwall of thrusts commonly undergo

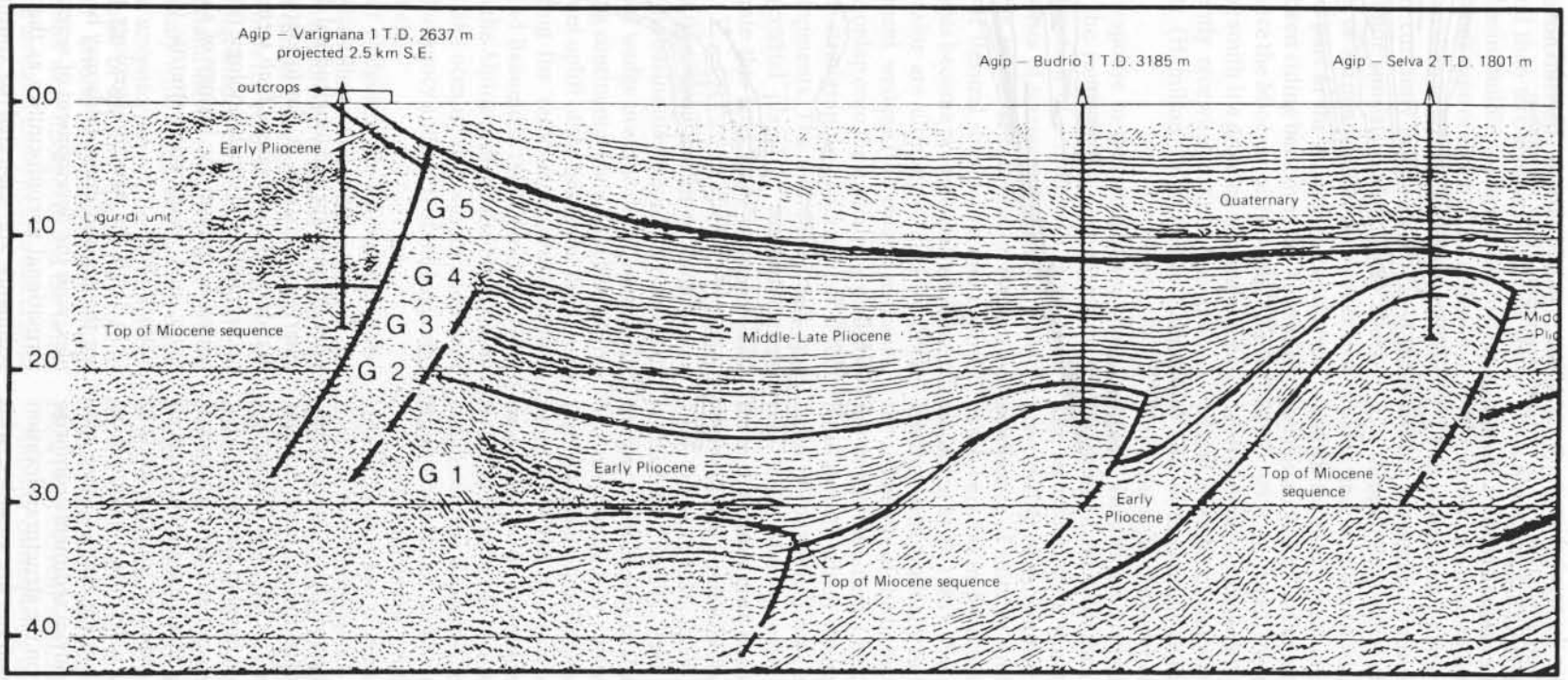


Fig. 9.69. Example of a seismic transect across the edge of a foreland basin, showing blind thrusts and sediment drape over them. The main basin, filled with Pliocene sediment at the *center*, is an example of a satellite, or piggyback, basin. (Ricci-Lucchi 1986)

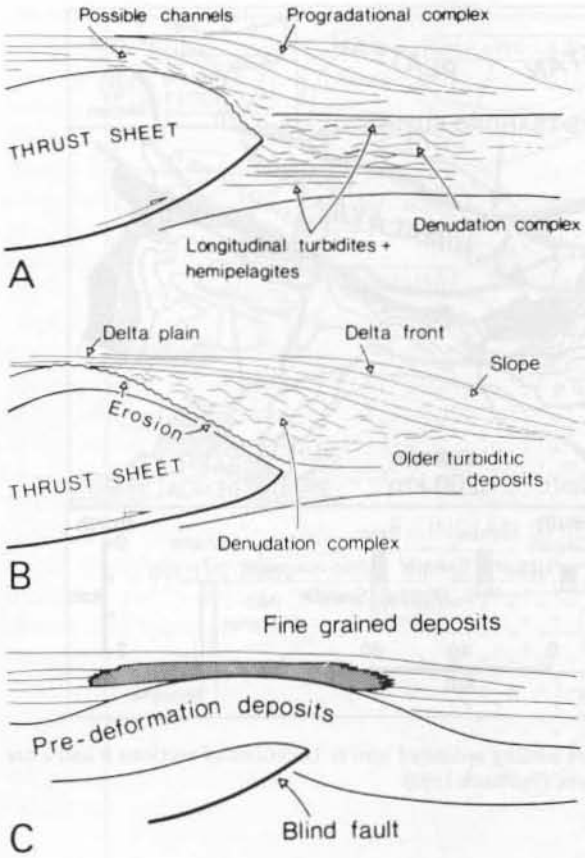


Fig. 9.70 A–C. Influence of structure on sedimentation. Blind thrusts in the Apennine–Adriatic Foreland Basin. (Ori et al. 1986)

folding even as fresh sediment is deposited across the top of the fault, leading to the development, in only a few million years, of markedly angular unconformities that pass down dip and along strike into perfectly conformable relationships over only a few hundred meters (Riba 1976; Anadón et al. 1986; Puigdefábregas et al. 1986). A quite different set of syndepositional unconformities develops over the forebulge, as discussed in Section 7.4 (Tankard 1986).

The Indian proforeland basin (Figs. 9.71, 9.72) is described briefly here as a case example. The structure, stratigraphy, and sedimentology of this region have been summarized by LeFort (1975) and Graham et al. (1975), and the sedimentology has been described in more detail by Johnson and Vondra (1972), Visser and Johnson (1978), Johnson et al. (1979), Parkash et al. (1980), Johnson et al. (1985), Burbank and Raynolds (1984, 1988), Burbank and Beck (1991), Burbank (1992), and Garzanti et al. (1996).

Cretaceous to Eocene strata of the northern Indian subcontinent are sandstones, mudstones, and limestones representing shallow-marine shelf environments. Faunal affinities of vertebrates suggest the establishment of a land connection with the Asian plate in the early Eocene, which is confirmed independently by the sequence of events deduced for the developing suture to the north (discussed below). The Oligocene to Miocene Murree Formation (> 2000 m) consists of deltaic deposits derived from the Indian Shield and the rising Himalayan Orogen. Its relation to the Indus Molasse is unclear.

The Murree passes gradationally up into the Siwaliks, an entirely nonmarine unit up to 5 km thick. Depositional environments were similar to those existing at present, namely giant alluvial fans draining transversely from the rising mountains into longitudinal trunk rivers, such as the Indus, Ganga, and Brahmaputra. Indeed, the major north–south rivers are regarded as type examples of antecedent drainage (Seeber and Gornitz 1983). They have maintained their courses, perhaps since the Miocene, while cutting through ranges that are now higher than their sources. The Siwaliks have been divided into three units. The lower Siwalik comprises a coarsening-upward megacycle consisting of sandstone–mudstone alternations passing up into a predominantly sandy sequence. The middle Siwalik consists mainly of medium to coarse sandstones with interbedded mudstones, and it grades up into a largely conglomeratic upper Siwalik, the middle and upper Siwaliks together comprising a second megacycle. Fining-upward cycles 1–35 m thick occur throughout the sequence. The three subdivisions of the Siwaliks are time transgressive, representing southward and southeastward progradation of fluvial depositional systems into and down the axis of the trough. The Siwaliks are folded and faulted near the Main Boundary Fault but pass out into flat-lying beds overlain conformably by modern alluvium. The alluvium itself is tilted near the mountain front, and Landsat images indicate active folds in parts of the trough. Petrographic studies of light and heavy minerals and of the conglomerate clasts document the progressive unroofing of Tethyan rocks that were metamorphosed deep in the collision zone.

9.3.4.8 *Remnant Ocean Basins*

The Bay of Bengal embayment is the type example of a remnant ocean basin (Graham et al. 1975; Ingersoll et al. 1995). Discharge from the delta of the Ganga and Brahmaputra rivers contains the largest sediment load of any river on earth (Stoddart 1971). It drains into the Bay of Bengal, an embayment along the Indian–Asian suture (Fig. 9.72). The sediment is deposited on the Bengal Fan, which is the world's largest clastic depositional system. In the Miocene, part of the sediment was dispersed southeastward to feed the

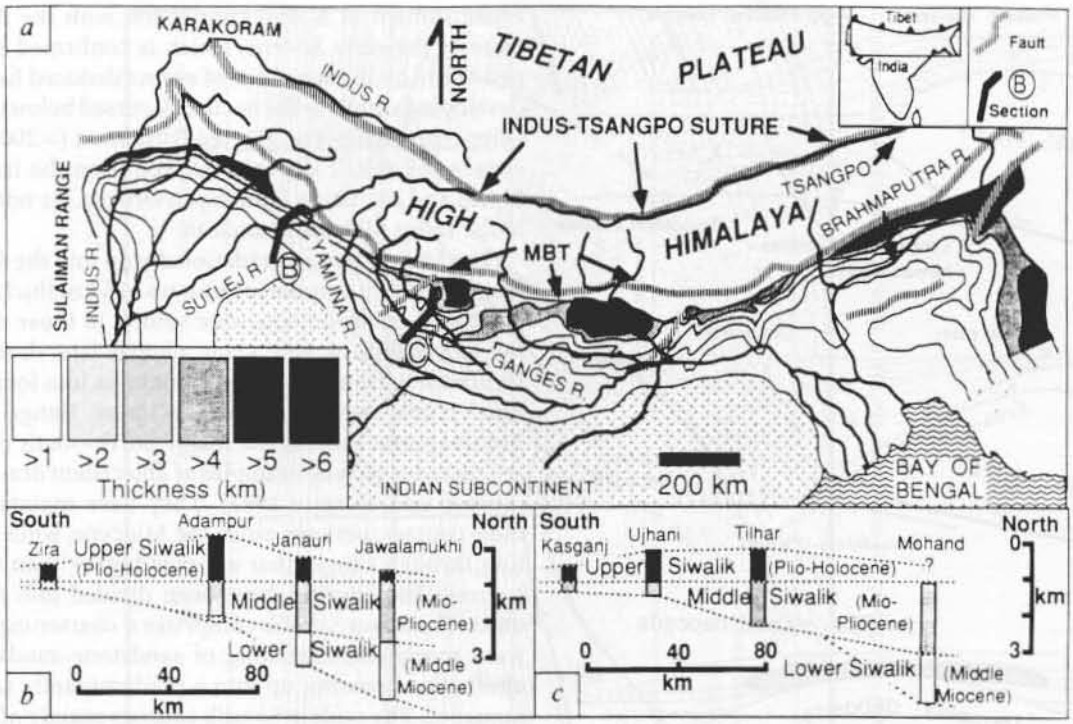


Fig. 9.71. The Indian proforeland basin and the major drainages feeding sediment into it. Locations of sections *b* and *c* are shown by *heavy solid lines* to the east and west of the Yamuna River. (Burbank 1992)

Nicobar Fan, which extends as far as the Java Trench. However, during the Pliocene or Pleistocene, northward subduction of the Indian–Australian plate beneath the Sunda arc brought the Ninetyeast Ridge (a volcanic ridge formed along a now-extinct leaky transform fault) into contact with the arc, cutting off the sediment supply, and the fan is now inactive (Curray and Moore 1974). The Indus fan is being deposited in a similar remnant ocean basin to the west of the Indian subcontinent (Fig. 9.72).

The general model for remnant ocean basins, first proposed by Graham et al. (1975), includes the concept of progressive closure – the "zipping-up" of a suture – with the uplifted orogen of the suture extending gradually along strike and shedding clastic detritus axially along the basin ahead of itself. They applied this concept to the development of the late-Paleozoic Ouachita system of Arkansas and Oklahoma, which represented a remnant ocean basin during the closure of the proto-Atlantic (Iapetus) Ocean.

In older orogens, the distinction between remnant ocean basins and remnant backarc basins may be difficult (Hsü 1988). The distinction may throw light on arc polarity (dip direction of the subduction zone) and the suturing history. The Black Sea and south Caspian Sea in the Alpine orogen are examples (Dewey et al. 1973). The Mediterranean Sea consists of many separate structural provinces but, in general, it must be classified as a remnant ocean basin, all that remains of the former Tethyan Ocean. These are all enclosed or nearly enclosed seas and are characterized by such facies as evaporites and euxinic shales (Degens and Ross 1974; Kojumdgieva 1983). The deposits of a remnant ocean are exposed within the Alpine foreland basin of Switzerland; in this ocean, thick, deep-water turbidites and associated facies (the original flysch formation) accumulated in the early Cretaceous–Paleogene, prior to terminal collision (Hesse 1982; Homewood and Caron 1982; Stockmal and Beaumont 1987).

Other modern and ancient examples of this class of basin are discussed by Ingersoll et al. (1995). The Markham Suture in Papua–New Guinea, discussed in the next section, includes a modern remnant ocean basin, the sedimentary history of which can be used as a basis for interpretation of the strata preserved in the suture.

Ingersoll and Busby (1995) proposed the term **dormant ocean basins** for a new class of basins floored by oceanic crust that is neither spreading nor subducting. These include basins where active spreading has ceased as a result of plate rearrangement. They cite the Gulf of Mexico as an example. A second type would include remnant ocean basins where subduc-

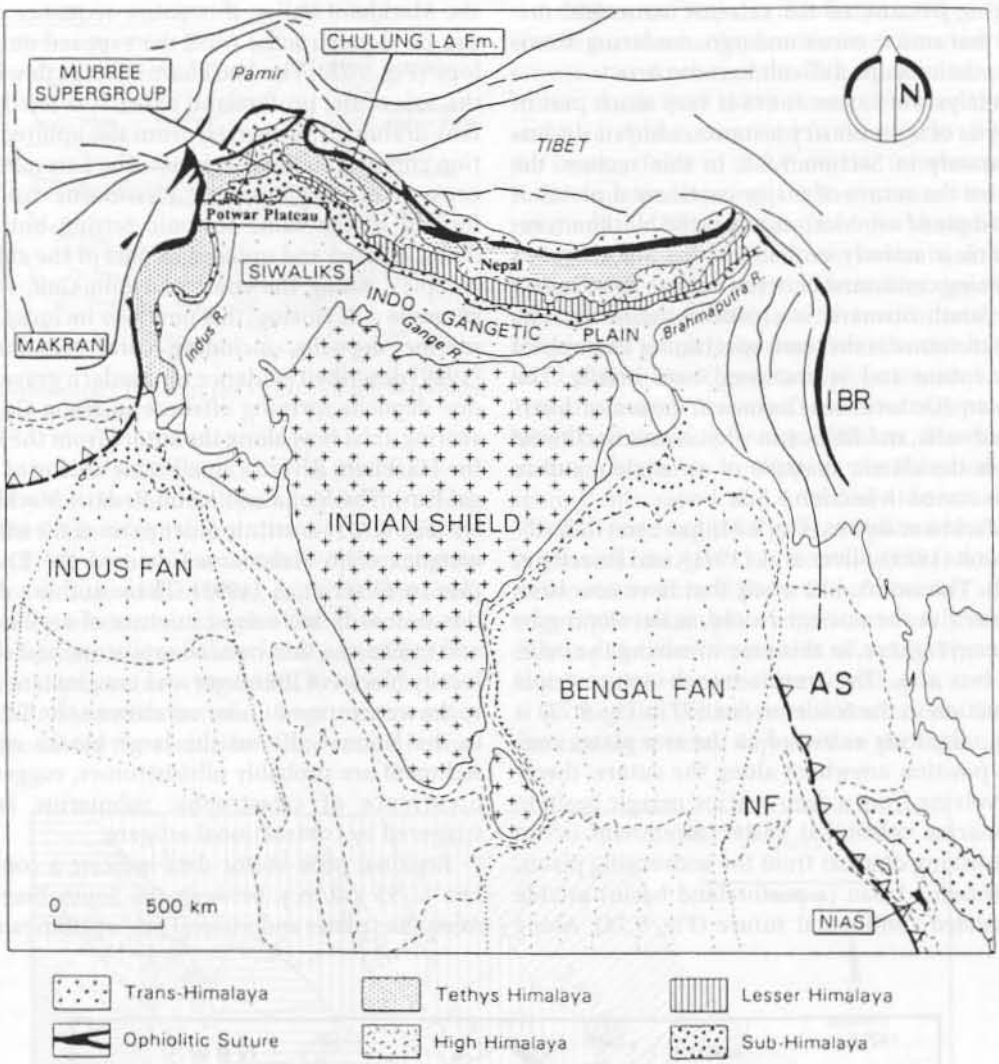


Fig. 9.72. The Indian–Asian collision zone. The Indus, Ganges, and Brahmaputra rivers and their tributaries occupy the proforeland (peripheral) basin that developed on the downgoing (Indian) plate following collision, probably in the Eocene. Proximal parts of this basin have been uplifted to form the Sub-Himalaya. The two major rivers draining this basin are discharging into remnant ocean basins to the southwest and southeast, where the giant Indus and Bengal fans (respectively) are being deposited. *IBR* Indoburman Ranges; *AS* Andaman Sea; *NF* Nicobar Fan. (Ingersoll et al. 1995)

tion has ceased, leaving an oceanic gap in the suture belt. Some of the basins in western China, such as the Tarim Basin, would fall into this class (assuming they are not floored by continental crust), as would the Caspian Sea Basin and the Black Sea, all of which are located within suture zones.

9.3.4.9 *Basin Analysis of Suture Zones*

These are highly complex regions, within which sedimentary rocks may be only minor components, although vast quantities of sediment may be shed into adjacent areas by uplift and erosion of the suture orogen. Reconstructing the geological history of suture zones requires the integration of data on structural geology, magmatism, metamorphism, and sedimentation. Facies analysis of the sedimentary record, coupled with evidence of stratigraphic age, can yield important information regarding major tectonic events, such as basinal subsidence due to crustal loading by the growing collisional orogen, or uplift and lateral displacement generated as the orogen accommodates crustal shortening. Petrographic analysis of detrital components and paleocurrent analysis of such facies as turbidities and fluvial deposits can yield information on source terranes and their uplift and unroofing histories. Such analyses commonly are

challenging because of the extreme structural disruption that suture zones undergo, rendering stratigraphic relationships difficult to reconstruct.

The analysis of suture zones is very much part of the analysis of accretionary terranes, which is discussed separately in Section 9.3.5. In this section, the focus is on the suture of major continental plates. A good example of a modern suture is the Markham suture, which is actively evolving as the Solomon Sea crust is being consumed between Papua–New Guinea and the South Bismark Sea plate to the north. The Himalayan suture is the most spectacular example of a major suture and is discussed here briefly. The Caledonian (Ordovician–Devonian) suture of Laurasia, Gondwana, and Baltica, in what is now northwest Europe, is the classic example of an ancient suture and is discussed in Section 9.3.5.

The Markham suture (Fig. 9.73) has been described by Crook (1989), Silver et al. (1991), and Brierley et al. (1993). The suture, like many that have now been documented in the ancient record, is developing by oblique convergence, in this case involving the collision of two arcs. The trench–trench–suture triple point junction in the Solomon Sea (*TJ* in Fig. 9.73) is unstable, migrating eastward as the two plates converge. A position anywhere along the suture, therefore, is evolving from a deep-marine pelagic basin to a deep-marine collisional basin (a remnant ocean basin) receiving detritus from the converging plates, to a nonmarine basin (a proforeland basin) astride the completed continental suture (Fig. 9.74). Along the Markham Valley, this entire sequence of events can be reconstructed from the exposed onland geology (Fig. 9.75). The Markham River is flowing along the axis of the proforeland basin. It is fed by alluvial fans draining transversely from the uplifted subduction complex to the northeast. The Leron Formation consists of conglomeratic Pleistocene fan deposits formed in the same tectonic setting but are now thrust-faulted and uplifted as part of the subduction complex. Along the coast of Huon Gulf, where the suture is still closing, this unit also includes shallow-marine deposits, including coral remains. Crook (1989) described evidence for modern gravelly turbidite deposits forming offshore in Huon Gulf, representing axial flow along the suture from the mouth of the Markham River (a small-scale analog of the Bengal Fan). The Mena and Mebu Beds of Markham Valley (Fig. 9.75) constitute older parts of the subduction complex (Plio-Pleistocene), named the Erap Complex by Silver et al. (1991). These authors described this unit as "a tectonized mixture of argillite, basalt, and sandstone, thin bedded sandstone and shale, and locally blocks of limestone and conglomerate." These rocks were formed under conditions similar to those in the Huon Gulf, but the large blocks of derived sediment are probably olistostromes, suggesting the occurrence of catastrophic submarine landslides triggered by contractional orogeny.

Regional plate-vector data indicate a convergence rate of 93 km/m.y. between the South Bismark and Australian plates and a lateral propagation rate of 120–

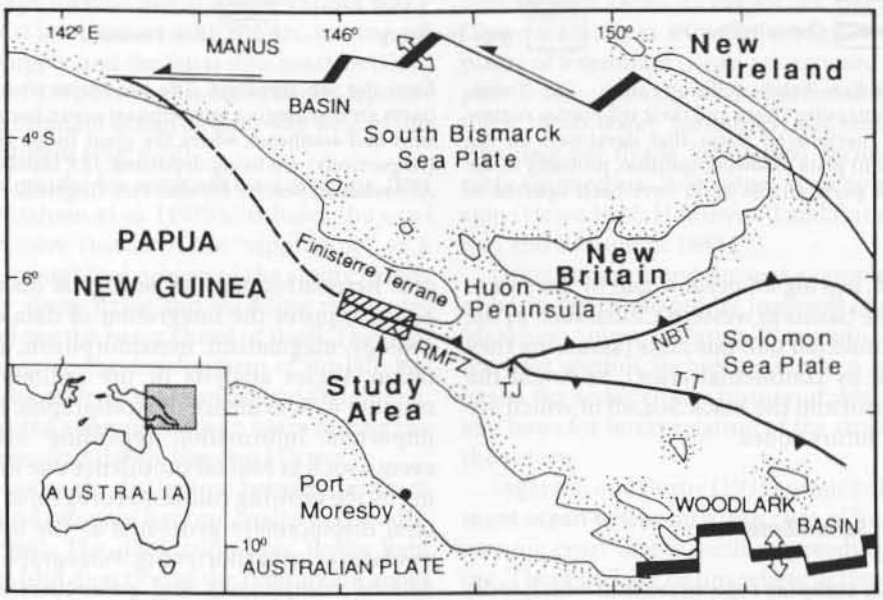


Fig. 9.73. Tectonic setting of the Markham Suture. *NBT* New Britain Trench; *TT* Trobriand Trench; *RMFZ* Ramu–Markham Fault Zone and Suture; *TJ* unstable triple junction. A geological map of the "study area" is shown in Fig. 9.75. (Brierley et al. 1993)

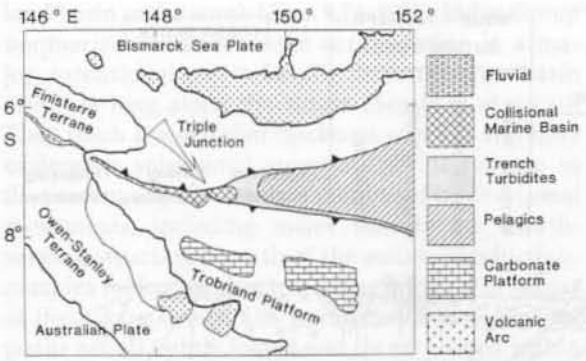


Fig. 9.74. Sedimentary facies in the vicinity of the Markham Suture. (Silver et al. 1991)

180 km/m.y. for the suture, depending on the orientation of the flanks of the suture. Crook (1989) and Silver et al. (1991) were able to confirm approximately this rate for closure of the suture, using ^{14}C and fossil dates from the subduction complex that show how the various facies have migrated eastward with time.

The Himalayan suture (Figs. 9.65, 9.66) is the classic example of a suture between large continents. A tectonic model for the evolution of the suture in southern Tibet and Nepal is illustrated in Fig. 9.76, based on international co-operative work between Chinese scientists and others from Britain, France, and elsewhere (Searle et al. 1987). Contact between the Indian and Asian plates occurred in the region of Nepal in the Eocene. Prior to that time, the Tethyan margin of India was undergoing shallow-marine sedimentation, and the Tibetan margin of the Lhasa block was a magmatic arc. In the remnant ocean basin between these plates, Indian-derived turbidites were deposited, deep-water radiolarian cherts accumulated in the trench, and granitic detritus from the arc shed southward to form the Xigase Group in a forearc basin. The Tethyan margin deepened and then collapsed in the latest Cretaceous prior to collision. The Yamdrock melange began to develop in the Paleocene, representing offscraped and tectonized trench deposits. The Tibetan margin became an uplifted Andean-type continental margin. Continental sedimentation within the suture zone (Qiuwa Conglomerate) and at the edge of the Indian plate began in the early Eocene, indicating the commencement of uplift. The main deformation north of the Himalaya began at this time.

In the mid-Tertiary, as the nonmarine Siwalik sediments were being deposited in the Indian profore-

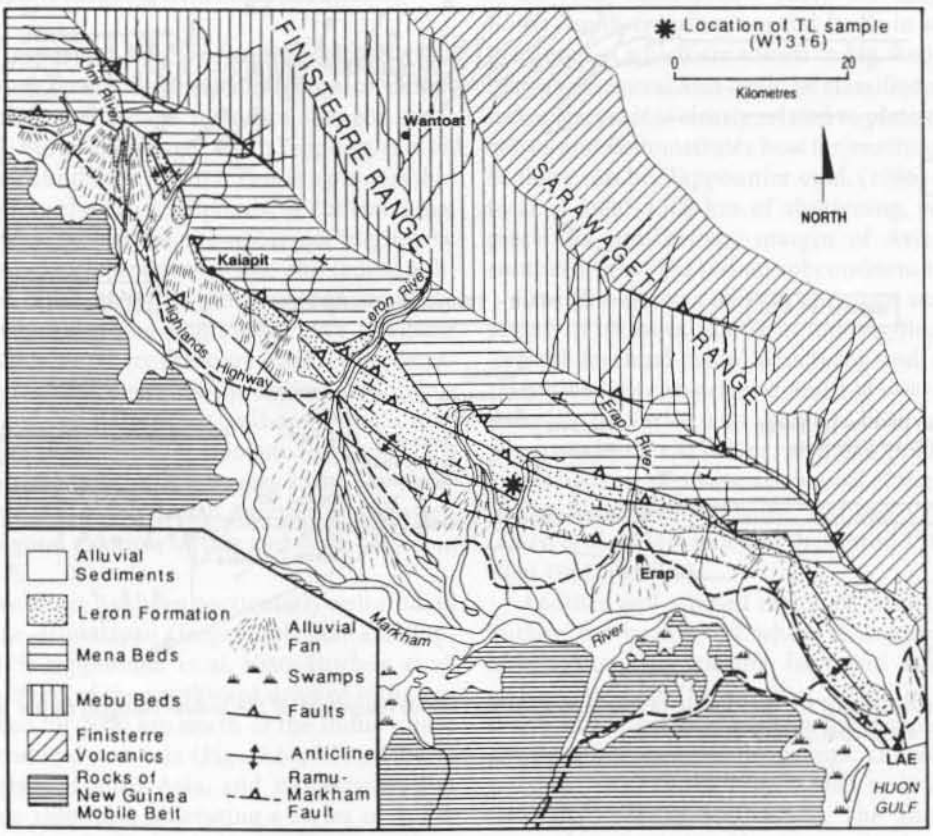


Fig. 9.75. Geology of the Markham Valley, Papua–New Guinea, showing the major lithostratigraphic units and major structures. (Brierley et al. 1993)

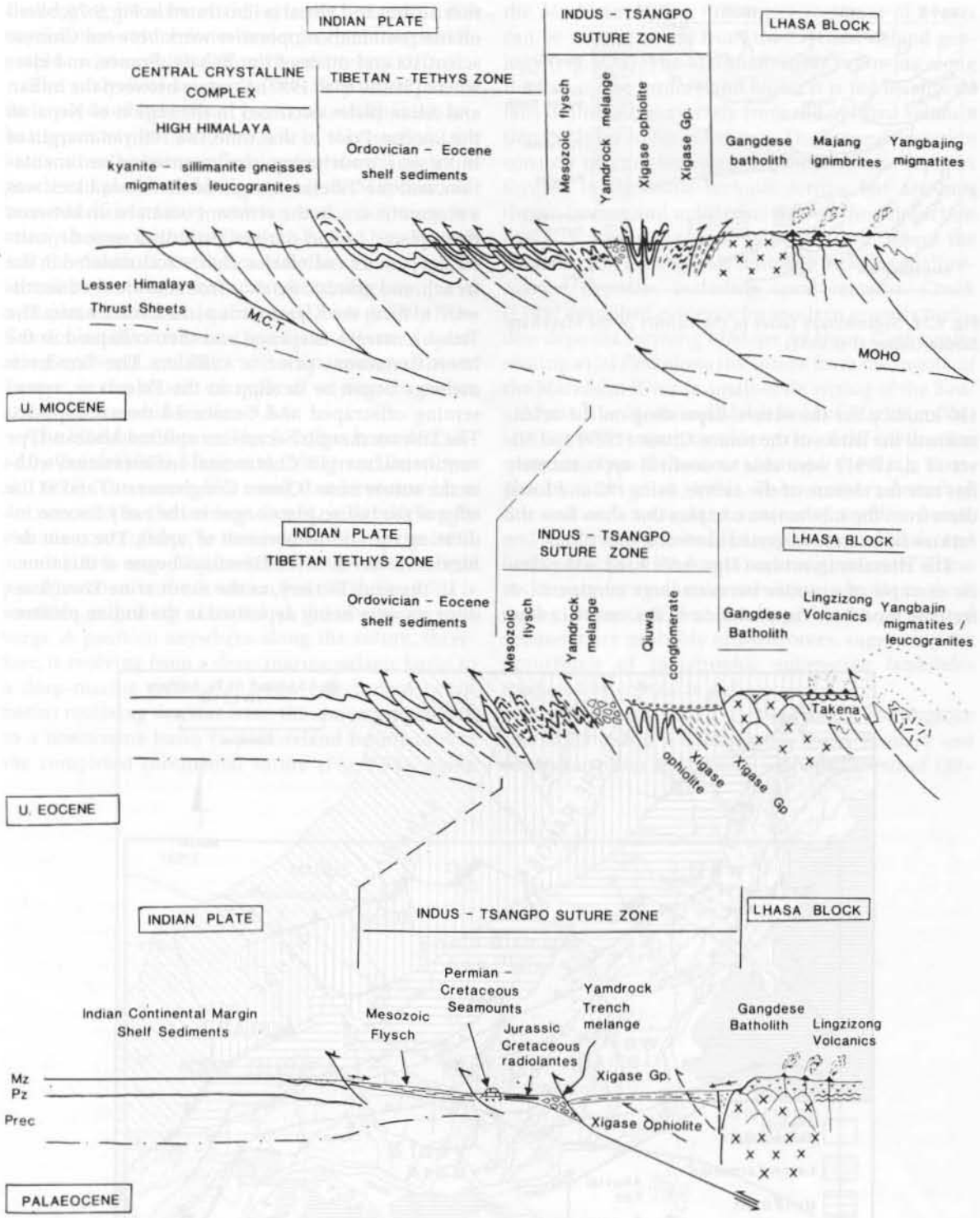


Fig. 9.76. Tectonic model for the closing of the Tethys Sea and the development of the Indus–Tsangpo suture in the area of southern Tibet and Nepal. (Searle et al. 1987)

land basin to the south (Figs. 9.71, 9.72), Indus Group nonmarine sediments were accumulating in a major extensional or strike-slip intermontane basin 2000 km long along the suture (Searle et al. 1990). They reach a maximum thickness of 4 km and have undergone substantial structural dislocation up to the present day as a result of continued contractional movements, including major backthrusts (north-verging structures) north of the suture. Subduction-complex rocks, Yamdrock melange, forearc turbidites of the Xigase Group, and postsuture nonmarine deposits are all tightly folded and thrust faulted within the suture zone (Fig. 9.76). Pleistocene glaciofluvial sediments and modern moraines and terraces are affected by thrust faults and extensional faults throughout the orogen (Searle et al. 1987). The stratigraphy and sedimentology of these and other syntectonic nonmarine deposits within the Himalayan orogen are now being unraveled, and the results are yielding fascinating insights into the timing of uplift and lateral movement of the various tectonic blocks in this complex region during the late-Cenozoic uplift of the region, as basins were created and uplifted and major rivers diverted (Brookfield 1993).

9.3.4.10 *Basins and Uplifts Resulting from Intraplate Stresses*

One of the important effects of collision is what has variously been termed **indentation tectonics**, **extrusion tectonics**, and **escape tectonics** (Şengör 1976; Dewey 1977; Burke and Şengör 1986; Tapponnier et al. 1986; Hancock and Bevan 1987). This is a process that develops where the crust responds to contractional stress by lateral movement of continental fragments. The resulting development of strike-slip faults, pull-apart basins, and graben has been noted above. In complex orogens, where continents with irregular margins collide, or where there are intervening terranes and microplates, deep-marine sedimentation in a small remnant ocean basin or pull-apart basin may be contemporaneous with tectonism and uplift of a local suture a few tens of kilometers away. The Caledonian orogen of eastern North America and western Europe is a good example of this and is discussed in Section 9.3.5.

Escape tectonics has been particularly well studied in the Alpine–Himalayan Orogen. (Molnar and Tapponnier 1975; Tapponnier et al. 1986; Huchon et al. 1994). The effects of the northward drive of India can be recognized for 3000 km north of the Indus suture within central and east Asia (Figs. 9.66, 9.77). Tibet is being squeezed against Asia, and its response has been to slide sideways, generating a series of major left-lateral strike-slip faults that extend eastward into China and southeastward into Indochina, Thailand, and Malaysia. Displacements of hundreds of kilometers and block rotations in the order of 25° have been suggested for some terranes in southeast Asia (although the magnitude of the displacement has been disputed by Dewey et al. 1989). Many of the faults are transpressive and have developed typical overthrust foreland-basin margins flanking the major intermontane basins of west China (Chen and Dickinson 1986; Liu 1986; Zhai 1986). A smaller set of dextral faults trending north–northwest may represent the other pair of a conjugate set. Regions of extension and rift-basin development include the Baikal Rift, Shansi Graben, north China Sea, and Andaman Sea (Logatchev et al. 1978; Chen et al. 1984). Logatchev et al. (1978) demonstrated, from geophysical data, that the Baikal rift zone is underlain by attenuated continental crust, indicting a region of mantle upwelling in response to the regional stress pattern. There are numerous rift valleys in the area, Lake Baikal representing one of the oldest (mid-Cenozoic), with many others in the process of active formation and initiation (Fig. 9.78). Active rifting began in the Baikal Rift zone at about 35 Ma, according to Mats (1993). The rifts contain up to 6 km of lacustrine and fluvial sediments. In places, the Quaternary section alone, including basalt flows, is up to 400 m thick (Logatchev et al. 1978; Mats 1993). There are also numerous north–south-trending normal faults in southern Tibet, some of which are shown in Fig. 9.65. The deformation of central Asia could be classified as intraplate tectonics, yet it is clearly related to plate-margin processes and demonstrates how far-reaching the effects of these can be. Tapponnier et al. (1986) estimated a total of 1000–1500 km of shortening, which would place the pre-Tertiary margin of Asia where the southern tip of the Indian subcontinent is now.

The Himalayan collision zone can now be interpreted in terms of the basin models discussed here. Typical foreland-basin structures and nonmarine, clastic wedges characterize much of west China (e.g., margins of Tarim and Jiuquan basins; Fig. 9.77), whereas classic rift basins, with marginal fan wedges and lacustrine deposits, are typical of the extensional basins of the Baikal, Shansi, and east-China regions (Chen et al. 1984; Chen and Dickinson 1986; Liu 1986; Zhai 1986; Fig. 9.66).

Another well-studied region of escape tectonics is Turkey, the geology of which is dominated by the North Anatolian transfer fault and other offshoot strike-slip faults (Dewey et al. 1986). In this region, many pull-apart basins occur; these show structural, stratigraphic, and sedimentologic characteristics similar to those in California that were modeled by Crowell (1974a,b; Sect. 9.3.3). The Turkish basins were described by Hempton et al. (1983) and Hempton and Dunne (1984). Other examples of es-

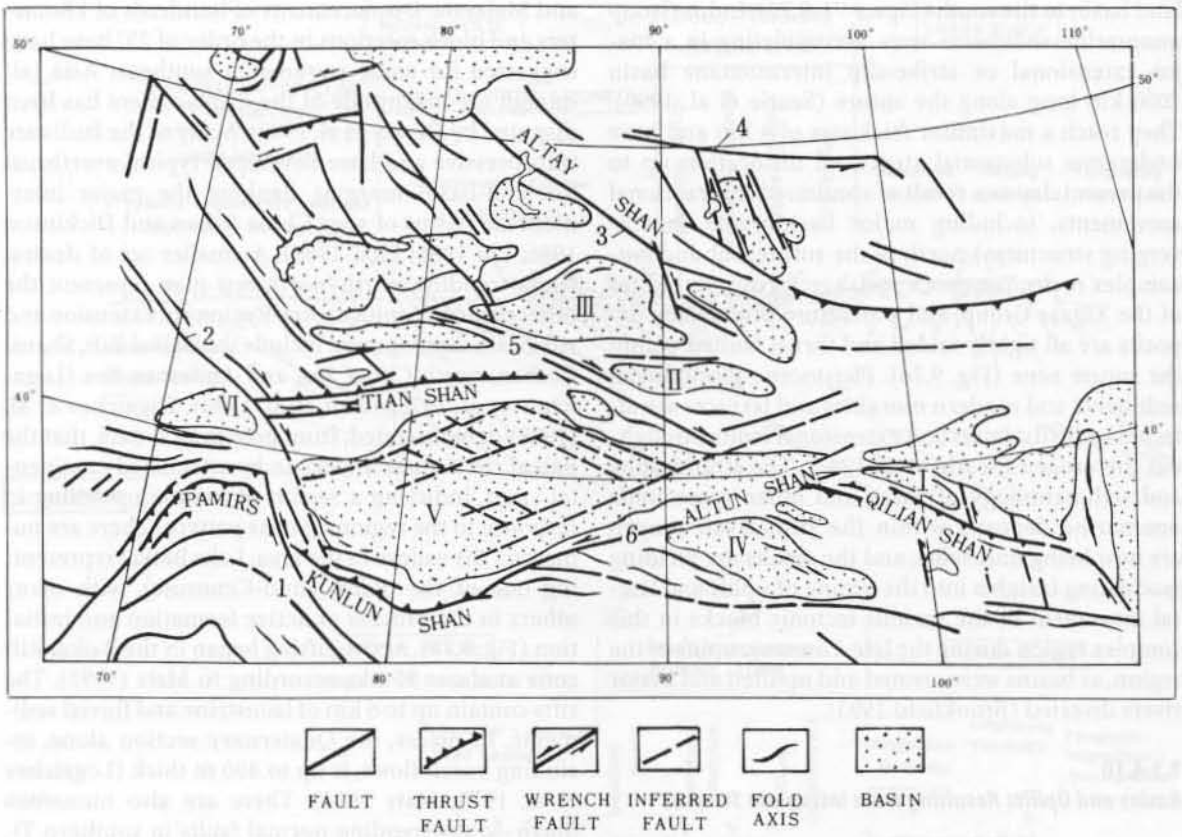


Fig. 9.77. Basins and major structures in west China and adjacent areas of Russia, Afghanistan, and the Indian subcontinent. Many of the basins are bounded by fold–thrust belts and have the character of foreland basins. Basins: *I* Jiuquan; *II* Turpan; *III* Junggar; *IV* Qaidam; *V* Tarim; *VI* Fergana. Faults: *1* Karakoram; *2* Talaso–Fergana; *3* Ertix; *4* Changajn; *5* Borohoro; *6* Altun Shan. (Liu 1986)

cape tectonics were noted by Burke and Şengör (1986).

The most well-known ancient megasuture is the Paleozoic Caledonian–Appalachian terminal suture. The gradual closure of Iapetus and other oceans from early Ordovician to late Devonian times is one of the more well-described Phanerozoic continental collisions. Dewey (1982) summarized the development of the orogen from a British standpoint, and summaries are contained in Windley (1984), Condie (1997), Blundell et al. (1992), and Stanley (1993). Closure was gradual, in part because of the presence of arc-related basins that were the first to collapse, but in part because subduction and convergence of the opposing continental plates was oblique and involved the suture of an irregular Laurentian margin against the (predominantly west-verging) arcs of the Iapetus Ocean (Phillips et al. 1976; Pickering et al. 1988; Stockmal et al. 1990). Terminal collision occurred across northern England in the late Ordovician, but not until the late Silurian in southwest Ireland. Convergence continued until the last oceanic hole, a back-arc basin in the Quebec–New Brunswick area, was closed, probably during the Devonian Acadian Orogeny (Stockmal et al. 1990). While convergence was continuing, those areas where collision had already occurred continued to deform along several major left-lateral transcurrent faults (Dewey 1982). This is, necessarily, a highly simplistic summary of an extremely complex sequence of events. The interested

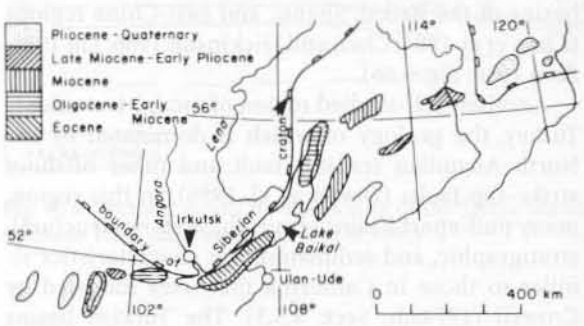


Fig. 9.78. Sequence of rift-valley development in the Baikal Rift System. (Logatchev et al. 1978)

reader should turn to the extensive compilation of research results recently compiled by Harris and Fettes (1988) for more complete details.

The Caledonian Orogen is characterized by the famous Devonian Old Red Sandstone, a fluvial and lacustrine unit that occurs in numerous late- to post-orogenic basins in Britain, Scandinavia, Svalbard, East Greenland, and northeast North America (Allen et al. 1968; Friend 1969, 1981; Fig. 9.79). It is by no means certain how all these basins fit into the plate-tectonic models discussed here. Such analysis is rendered particularly complex by the recognition that, like the Cordilleras and the Alpine–Himalayan orogen, the Caledonian–Appalachian Orogen is composed of a series of suspect terranes (Williams and Hatcher 1982; Stone and Kimbell 1995), the relative motions of which have yet to be fully resolved. Most of the Old Red Sandstone basins are considered to be late- to post-orogenic, in the sense that they postdate the major collisional events. Many of the basins may have been controlled by strike-slip faults that deformed the orogen during final closure (Friend 1981; Dewey 1982; Coward 1990),

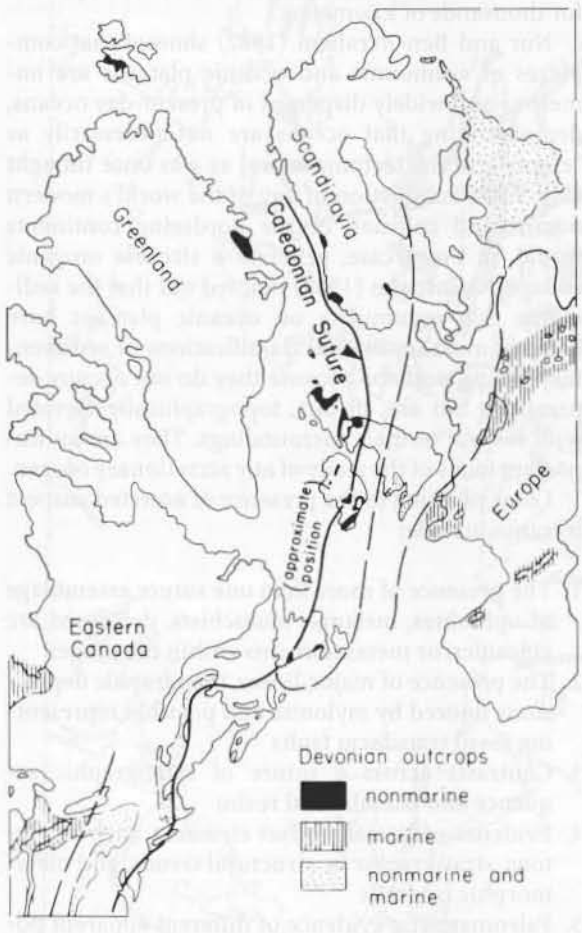


Fig. 9.79. The Caledonian terminal suture and resulting Old Red Sandstone basins. (After Friend 1969 and other sources)

similar to the escape tectonics of Tibet and Turkey referred to previously, although Friend (1985) stated that none of the resulting basins can be compared with the classic strike-slip-fault basins of California. McClay et al. (1986) compared the Caledonian Orogen during the Devonian to the Basin and Range province, in which extensional tectonism has been dominant and low-angle detachment faults and half-graben basins are typical (Sect. 9.3.2). McClay et al. (1986) regarded strike-slip faulting as being of secondary importance. The Hornelen basin in Norway has long been cited as an example of a basin developed by strike-slip faulting (Steel 1976; Steel and Gloppen 1980), but recent structural analysis suggests that the observed strike-slip faulting is a superficial effect of movement on deep-seated detachment surfaces during postorogenic extensional movements (Hossack 1984; Seranne and Seguret 1987). Bluck (1978, 1980, 1985) and Coward (1990) attributed the Devonian basin of the Midland Valley, Scotland, to strike-slip faulting modifying an earlier forearc basin. Other examples were listed by Friend (1981). External basins (Allen et al. 1968), namely those with coarse clastic facies passing laterally into marine-shelf deposits extending onto the adjacent craton (e.g., Catskill Delta of New York, Anglo Welsh Cuvette), represent retro- or proforeland basins.

Other examples of intraplate basement uplifts were reviewed by Rodgers (1987). To mention but one group of examples, Miall (1986) summarized data regarding a suite of late Silurian–early Devonian uplifts oriented transverse to regional fabrics in the Canadian Arctic Islands. He suggested that they were generated by intraplate stresses transmitted for more than 1500 km across North America from the Caledonian suture of Greenland and Scandinavia, and he proposed an analogy with the Laramide structures of the Rocky Mountains. Stratigraphic and structural data have now been shown to be consistent with this interpretation for at least one of these structures, Boothia Uplift, which is now interpreted as the upper, ramp-like plate of a steep reversed fault (Okulitch et al. 1986; de Freitas and Mayr 1993). Coarse clastic wedges were derived from this uplift and shed to both sides. The wedges are asymmetric, with alluvial-fan and fan-delta boulder conglomerates banked against the fault on the lower plate, with a much broader zone of finer-grained clastics extending away from the uplift across the upper plate (Miall and Gibling 1978; Miall 1986).

9.3.5 Basin Analysis of Accretionary Terranes

Early attempts at plate-tectonic reconstructions tended to treat the present continents as major pieces of the puzzle that moved and collided intact, albeit

sometimes with ragged tears across earlier sutures (e.g., Wilson's classic analysis of the Atlantic Ocean in 1966). However, modern work has demonstrated that many large orogenic areas are composed of unrelated **terranes** that have become juxtaposed and sutured (accreted) by complex processes of subduction and transform motion. A terrane is any body of rock that functioned as a geologic unit for a significant period of geologic time. Terranes range in size from a few square kilometers (e.g., an oceanic seamount) to the thousands of square kilometers of an individual continent. Typically, in orogens, terranes are bounded by major faults. The terms **allochthonous terrane** or **suspect terrane** are commonly used to suggest the transported and suspect affinities of the geology in complex orogens. An **orogenic collage** is a geologic province consisting of a series of terranes accreted by convergent and/or transform plate motion.

Several useful reviews of terrane geology have been published, among which are Ben-Avraham et al. (1981), Nur and Ben-Avraham (1982), and Ben-Avraham and Nur (1983), who illustrated the distribution of potential terranes in modern oceans and briefly discussed some of the world's existing orogenic collages. Schermer et al. (1984) traced the origins of the terrane concept, beginning with the work of W.P. Irwin in the Klamath Mountains of California in the 1950s and 1960s. Howell (1985) edited a major compilation of research results dealing with the terranes of the circum-Pacific region, and his book also contains several papers dealing with terrane-mapping methods and interpretation. Kleinspehn (1988) reviewed the application of basin-analysis concepts and methods to the documentation and interpretation of suspect terranes. Dewey et al. (1991) edited a collection of papers on terrane geology that were assembled for a discussion meeting, the purpose of which was to review the state of the subject some ten years after the first formal enunciation of the principles of terrane analysis, primarily by D.L. Jones, P. Coney, and J.W. Monger (Jones et al. 1977; Coney et al. 1980). Included in this collection is a critical review of the terrane concept by Şengör and Dewey (1991), who argued that little about the methods of terrane analysis is actually new, and that much terrane analysis is taxonomic (simply descriptive) rather than analytic.

Some suspect terranes are produced by fragmentation and internal readjustment of structural units at a plate margin, such as is occurring in the present western Pacific region, where the geology is characterized by short spreading centers, arcs and transform faults, and very complex internal motions (e.g., Indonesia; Fig. 9.18). Saleeby (1983) compared the Cordillera in the Paleogene with the present Melanesian re-entrant (the complex of arcs and small ocean basins extending from New Guinea to New Zealand). Silver and Smith (1983) pointed out some comparisons of the Cordillera with Indonesia. In other cases, paleomagnetic and faunal evidence indicates that terranes may have traveled thousands of kilometers, rafted on ocean floor that became completely subducted except, perhaps, for a belt of mangled ophiolite along the terrane suture (although most preserved ophiolites probably formed in backarc basins and were preserved by obduction during subsequent convergent motion and collapse of the basin). Some suspect terranes may be major microplates floored by Precambrian basement and carrying thick Proterozoic–Phanerozoic sedimentary sequences. Others may consist of volcanic seamounts and plateaus or arc assemblages and rifted continental fragments and may lack an exposed cratonic core. The Wrangellia terrane is a good example (see below). Each terrane functions as an independent continental fragment, because it cannot be readily swallowed by subduction but is accreted at a suture. Some terranes may have originated close to the continent against which they are accreted and may not, therefore, have been transported very far. Others may have been transported for thousands of kilometers.

Nur and Ben-Avraham (1982) showed that complexes of seamounts and oceanic plateaus are numerous and widely dispersed in present-day oceans, demonstrating that oceans are not necessarily as "empty" (in the tectonic sense) as was once thought (Fig. 9.80). Subduction of any of the world's modern oceans and collision of the bordering continents would, in every case, generate a sizeable orogenic collage. Kleinspehn (1988) pointed out that the sediments that accumulate on oceanic plateaus have escaped most traditional classifications of sedimentary basins, perhaps because they do not occupy depressions but are, in fact, topographically elevated with respect to their surroundings. They are an important focus of the study of any accretionary orogen.

Clues pointing to the presence of accreted suspect terranes include:

1. The presence of more than one suture assemblage of ophiolites, melange, blueschists, deformed arc volcanics, or metasediments within the orogen
2. The presence of major, linear, topographic depressions floored by mylonite and possibly representing fossil transform faults
3. Contrasts across a suture of stratigraphic sequence and faunal/floral realm
4. Evidence of laterally offset elements, such as plutons, stratigraphic or structural trends, and metamorphic isograds
5. Paleomagnetic evidence of different apparent polar-wander paths (substantial changes in declination and inclination) for adjacent terranes

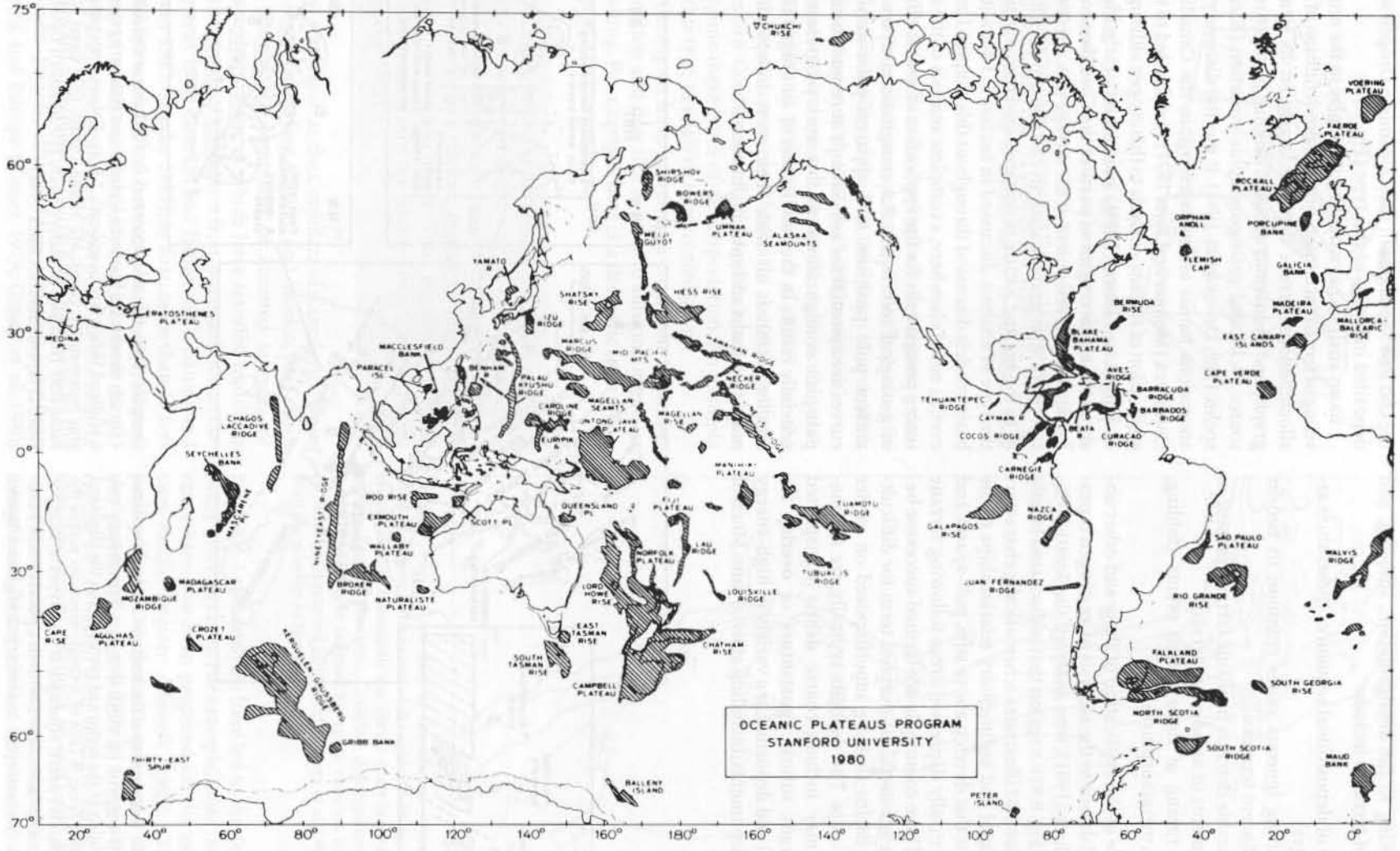


Fig. 9.80. Worldwide distribution of present-day oceanic plateaus. Some of these are continental fragments (e.g., Falkland Plateau and the terranes surrounding New Zealand); others are extinct arcs (e.g., Palau–Kyushu Ridge) and hot-spot traces (e.g., Hawaiian Ridge). All could become accreted terranes as the ocean on which they are located becomes consumed. (Nur and Ben-Avraham 1982)

Clues pointing to the amalgamation, suturing, and accretion of terranes include:

1. Overlap of depositional sequences (postsuture assemblages)
2. Crosscutting igneous suites common to two or more adjacent terranes
3. Recognizable detritus from one terrane present in the sediments of an adjacent terrane
4. Similar timing of deformation events (folding, faulting, metamorphism)

An example of the application of these and other observational details to the analysis of an orogen is provided by Bluck (1991), who analyzed the Scottish Caledonides (Fig. 9.81), a region that is discussed briefly below. Figure 9.82 illustrates schematically characteristic structural and sedimentary relationships of the small basins that develop above rifts, pull-aparts, and other structurally depressed areas following terrane assembly. These commonly are termed **successor basins**, which is a useful all-purpose term for difficult-to-classify basins that are superimposed on earlier structural units. The deposits typically are nonmarine and may include coarse debris transported across terrane sutures (**postsuture** or **overlap assemblages**) and deposited in a variety of high-energy fluvial environments, including alluvial fans. Much of the Old Red Sandstone of the Caledonian Orogen was deposited in basins of this type (Fig. 9.79).

Basin analysis has a major part to play in the unraveling of terrane accretion, because the definition of an allochthonous terrane depends largely on its stratigraphy and deductions about its original plate setting, source rocks, and paleogeographic evolution (Kleinspehn 1988; Dewey et al. 1991). It was the discovery of anomalous faunal assemblages within the Canadian Cordillera (Monger and Ross 1971) that first led to the application of terrane analysis to this orogen although, as Şengör and Dewey (1991) pointed out, early ideas about the exotic origins of parts of the Cordillera long predated the development of the terrane concept (Wilson 1968). Stratigraphic analysis can also provide clues as to the age and nature of terrane convergence, using the type of analysis discussed in Section 9.3.4 and the basin models discussed throughout this chapter. However, as noted elsewhere, a complete analysis of the suturing process calls for the application of a wide range of geological techniques. Paleomagnetic analyses of ancient pole positions and apparent-polar-wander curves are essential for any attempt at reconstructing past plate configurations. As this work progresses, it generally results in the subdivision of an orogen into smaller terranes, all with their own independent migration paths and suturing histories.

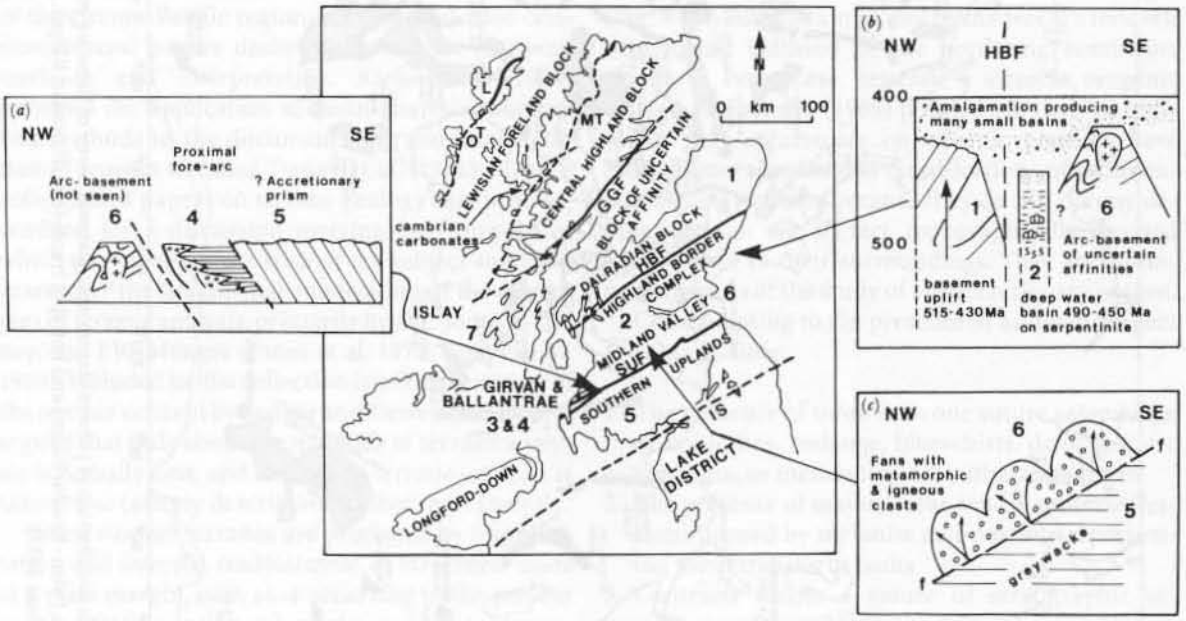


Fig. 9.81A–C. Terrane map of the Scottish Caledonides, showing some of the criteria by which they have been recognized. **A** Incompatibility of the history of events across the Highland Boundary Fault Zone; **B** Loss of tectonic elements across the Southern Uplands Fault; **C** Mismatch of clast types of local potential source areas across the Southern Uplands Fault. *1* Dalradian block (metamorphic basement); *2* Highland Border Complex (ophiolitic and other rocks); *3* Ballantrae Ophiolite; *4* Girvan (forearc); *5* Southern Uplands (accretionary prism); *6* Midland Valley (arc basement); *7* Islay–Colonsay metamorphic basement). *L* Lewis; *MT* Moine Thrust; *GGF* Great Glen Fault; *HBF* Highland Boundary Fault; *SUF* Southern Uplands Fault; *IS* Iapetus Suture. (Bluck 1991)

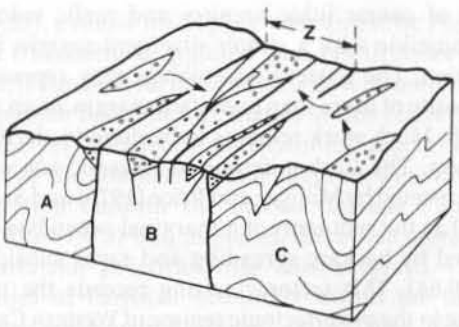


Fig. 9.82. The development of successor basins within orogens composed of accreted terranes, in response to late- to postsuture adjustments along terrane boundaries (rift and strike-slip faulting). *Arrows* show the direction of transport of sedimentary detritus, the analysis of which may be used to determine the timing of terrane-suture development. (Bluck 1991)

The tectonics of suturing have undergone a profound re-evaluation as a result of the deep-seismic-reflection studies of the COCORP team (Oliver 1982; Allmendinger et al. 1987a,b,c) and comparable studies elsewhere around the world, such as Canada's LITHOPROBE project (Cook et al. 1995) and the European Geotraverse (Blundell et al. 1992). Both the Western Cordillera and the Appalachians have undergone these types of reinterpretation. An example of this type of analysis is presented below.

Geologists have borrowed a term from space exploration and now like to refer to suturing events as docking. For example, here is a dramatic interpretation of Alaskan geology:

> The long history of successive collisions and terrane accretions around the rim of the northern Pacific suggests that a continental framework existed as a backstop or docking facility for migrating terranes, preventing them from bypassing Alaska and sweeping into the Arctic Basin (Churkin and Trexler 1981).

Vast areas, such as the Cordillera of Canada, the United States (including Alaska), and Mexico, plus much of Siberia, are composed of accreted terranes or microplates, some of which have traveled hundreds or thousands of kilometers from parts unknown (Coney et al. 1980; Churkin and Trexler 1981; Campa and Coney 1983; Saleeby 1983). At first, more than 50 terranes were identified between Alaska and Mexico, many of which were formerly scattered over the paleo-Pacific Ocean and were swept up by the relative westward movement of North America following the Triassic–Jurassic rifting of this continent away from Africa and Europe (Coney 1979; Coney et al. 1980;

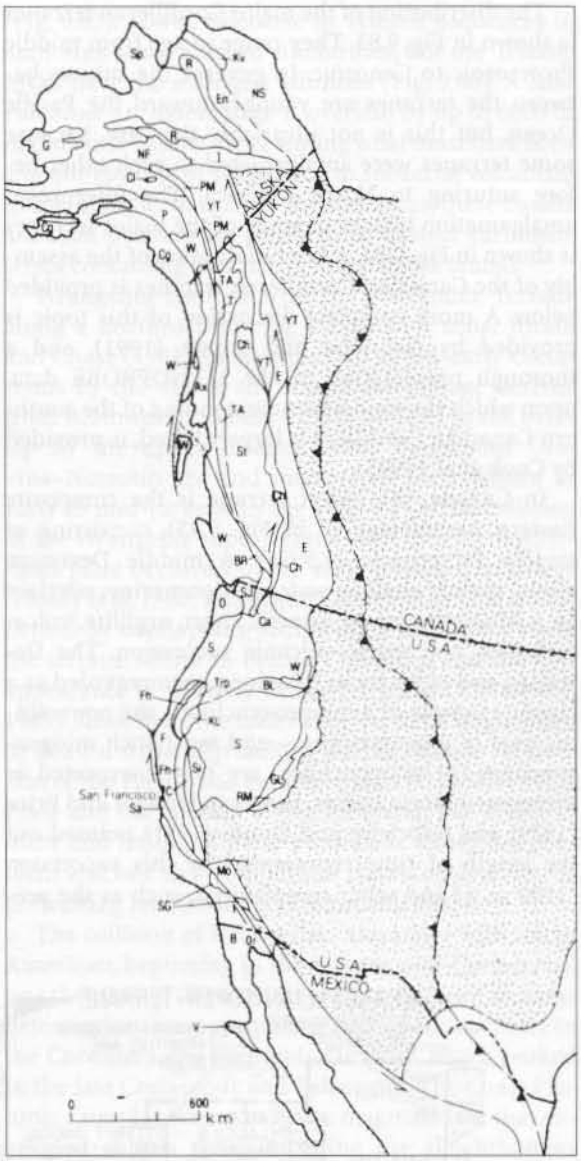


Fig. 9.83. Generalized map of Cordilleran suspect terranes. *Dashed pattern* is the extent of the autochthonous North American cratonic basement; the *thrust–fault symbol* marks the eastern limit of Mesozoic–Cenozoic deformation. The key to terrane symbols is omitted for space reasons, but some key terranes are discussed in the text. (Coney et al. 1980, reprinted with permission of Nature)

Fig. 9.83). Later analyses have led to the definition of many more terranes. For example, more than 350 terranes have now been proposed for the modern circum-Pacific region (Howell 1985). There has, perhaps, been a tendency to assign to a new terrane any problematic area within an orogen that cannot readily be correlated with adjacent areas, a practice that discourages attempts to carry out generalized correlations and to develop regional syntheses (Şengör and Dewey 1991).

The distribution of the major Cordilleran terranes is shown in Fig. 9.83. They range in age from middle Proterozoic to Cenozoic. In general, the sutures between the terranes are younger toward the Pacific Ocean, but this is not invariably the case, because some terranes were amalgamated to each other before suturing to North America. The interpreted amalgamation history of some of the major terranes is shown in Fig. 9.84. A brief discussion of the assembly of the Canadian Cordilleran terranes is provided below. A more complete discussion of this topic is provided by Gabrielse and Yorath (1991), and a thorough presentation of the LITHOPROBE data, upon which the modern understanding of the southern Canadian Cordillera is largely based, is provided by Cook et al. (1995).

In Canada, the oldest terrane is the composite Eastern Assemblage (E in Fig. 9.83), consisting of middle Proterozoic (1.5 Ga) to middle Devonian strata, mainly shallow-water to nonmarine, overlain by a Mississippian to Triassic chert, argillite, volcaniclastite, and mafic–volcanic succession. The Devonian and older rocks have long been regarded as a classic example of a miogeosyncline – the nonvolcanic part of a geosyncline – and most such miogeosynclines (or miogeoclines) are now interpreted as divergent-margin basins. However, Monger and Price (1979) and Gabrielse and Yorath (1991) pointed out the length of time represented by this succession (1100 m.y.) and other complexities, such as the presence of coarse lithic arenites and mafic volcanics incompatible with a simple divergent-margin interpretation. The Eastern Assemblage may represent a composite of more than one rifted margin, or an aborted rift. Much work remains to be done to clarify the problem. The Carboniferous-to-Triasssic succession is interpreted by Monger and Price (1979) and Saleeby (1983) as the remnants of a marginal ocean basin developed by backarc spreading and rapid subsidence (Fig. 9.84). This tectonic setting records the major change in the plate-tectonic regime of Western Canada from a divergent to a convergent environment.

Sutured against the Eastern Assemblage along the Teslin Suture is the Cache Creek or Atlin Terrane (*Ch* in Fig. 9.83), consisting of Mississippian to middle or upper Triassic, shallow-water limestones, highly disrupted radiolarian chert, argillite, basalt, and alpine-type ultramafics with local blueschist metamorphism (Coney et al. 1980; Saleeby 1983). Fusulinid faunas in the limestones are of tropical affinities, similar to those in Tethys and other circum-Pacific regions. This assemblage is a good example of a highly deformed remnant of ancient ocean floor. Mafic igneous rocks are interpreted as fragments of the late Paleozoic seafloor, including volcanic seamounts. The limestones were deposited in atolls developed over the seamounts. The terrane is caught between remnants of Triassic to Jurassic arcs (Takla, Hazelton, and Nicol arcs; Fig. 9.84), and the whole complex was accreted to the Cordillera when the Stikine Terrane (*St* in

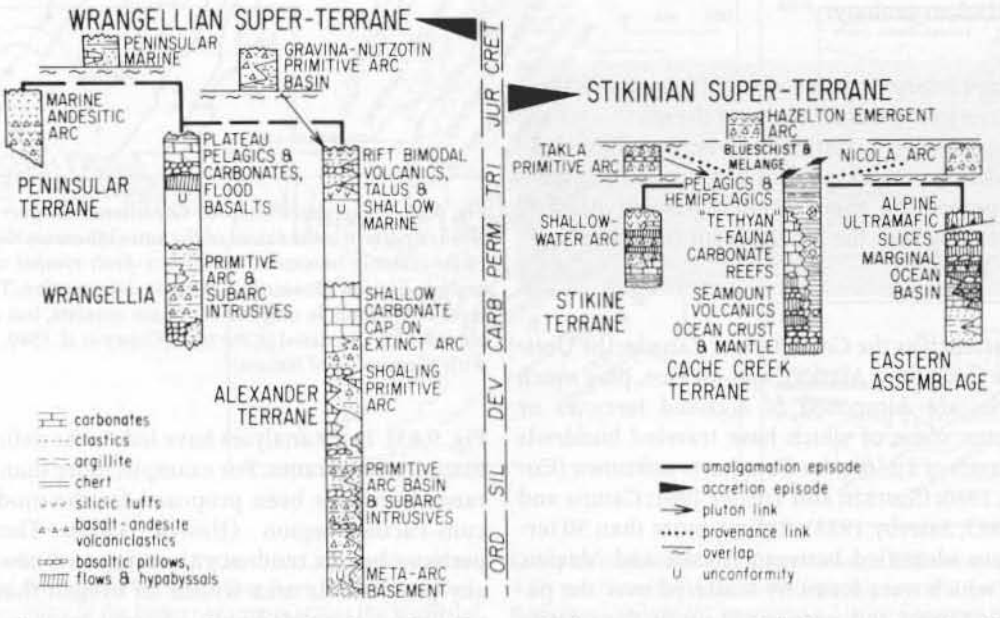


Fig. 9.84. The major terranes of British Columbia and southern Alaska, showing generalized stratigraphic columns and the history of amalgamation and accretion. Note the development of two large, amalgamated collages: the Wrangellian and Stikinian superterranes. (Saleeby 1983, reproduced with permission from the Annual Review of Earth and Planetary Sciences)

Fig. 9.83), a major microplate with an exotic Precambrian (?) basement, collided with North America in the early Jurassic, forming what is now termed the Intermontane Belt. The kinematics of this convergence are not yet completely understood. Structural evidence in the southern Yukon suggests west-dipping subduction beneath the Stikine Terrane (Tempelman-Kluit 1979) but, in places, there was more than one arc, and polarities may have reversed. Faunal affinities of Permian fusulinids within the Stikine Terrane are North American rather than Tethyan, suggesting that this terrane may have evolved outboard from North America, perhaps in fringing arc systems, and has not traveled far (Saleeby 1983). However, insertion of the Cache Creek Terrane between the Stikine and Eastern Assemblage terranes requires northward migration and subduction of an intervening ocean. Stikinia, with its overlapping arcs, may have assembled as a superterrane somewhere to the south and was, in part, emplaced against North America by dextral strike-slip (Saleeby 1983).

The collision climaxed in the middle to late Jurassic with regional metamorphism of the Eastern Assemblage, which now began its evolution into the Omineca Crystalline belt. Uplift followed, producing a major mountain range – the first precursor of the modern Rocky Mountains – and the shedding of floods of detritus both east and west (Eisbacher et al. 1974; Eisbacher 1981). In Alberta, this detritus comprises the oldest molasse pulse of the Rocky Mountains foreland basin, a mainly shallow-marine sequence of late Jurassic and early Cretaceous age (Hamblin and Walker 1979). Toward the west, the sediments comprise the Bowser Lake and Skeena groups in Bowser Basin (Fig. 9.85). This basin had a long and complex history (Eisbacher 1981). It was initiated on the Stikine microplate following the Jurassic suturing event. Depending on the polarity of the subduction zone along the suture, the basin could be interpreted as either a retro- or proforeland basin at this time. Sedimentary environments ranged from deltaic in the east to basin-floor turbidites in the west.

Meanwhile, somewhere in the Pacific Ocean, two other terranes (microplates) were approaching each other, to collide at some time between the early Jurassic and early Cretaceous to form the Insular Composite Terrane. These are the Wrangellia and Alexander terranes (*W* and *Ax* in Fig. 9.83). Paleomagnetic evidence summarized by Yorath and Chase (1981) shows that both terranes lay several thousand kilometers to the south of their present position during the Triassic. Wrangellia was one of the first major allochthonous terranes to be recognized in the Cordillera (Jones et al. 1977). It is represented by five major fragments between the Wrangell Mountains, Alaska, and the Hells Canyon area of Washington–Oregon (Fig. 9.83). Lower Paleozoic to Triassic sequences in each area show marked similarities, but the Triassic rocks have the strongest affinities (Fig. 9.86). A late-Paleozoic arc assemblage is overlain by up to 6000 m of tholeiitic basalts, representing what must have been a major episode of submarine to subaerial volcanism in the Triassic. In each of the five areas, these basalts are then overlain by platform to basinal carbonate strata containing distinctive invertebrate faunas.

Wrangellia sutured with the Alexander Terrane along a northeast-dipping subduction zone. Yorath and Chase (1981) dated this event as pre-early Cretaceous by the age of the suture assemblage derived from it. However, Saleeby (1983) pointed to the overlap by the upper Jurassic–lower Cretaceous Gravina–Nutzotin arc and interpreted the collision as early to mid-Jurassic in age (Fig. 9.84). The collision of the Wrangellia superterrane with the North American plate occurred in the early to mid-Cretaceous (Coney et al. 1980; Yorath and Chase 1981). This is confirmed by overlapping sedimentary sequences along the terrane suture in British Columbia, and the first appearance of westerly derived detritus (Kleinspehn 1985). Since then, the entire area has been fragmented by dextral strike-slip faulting during oblique subduction of the Pacific and Farallon plates (Gabrielse 1985; Price and Carmichael 1986). Resolving the displacement and timing of these motions is a complex problem and one that complicates the reconstruction of prefaulting relationships (Kleinspehn 1985).

The collision of Wrangellia–Alexander with North American, beginning in the early or mid-Cretaceous, was the start of the most important phase of tectonic deformation, metamorphism, and igneous activity in the Cordillera, the Laramide Orogeny, which peaked in the late Cretaceous and Paleogene. The Coast Plutonic Complex (coast belt) is a magmatic arc that developed at this time, intruding the allochthonous Taku and Tracy Arm terranes (*T* and *TA* in Fig. 9.83). Convergence of the Pacific plate was probably oblique, and several major, dextral strike-slip faults in the Cordillera may have been initiated in this phase.

Part of the west coast of Canada is still an active, convergent plate margin. An active shallow-dipping subduction zone extends under Vancouver Island (the Queen Charlotte transform fault also forms a significant part of the continental margin), and recent seismic surveys reveal the youngest accretionary event, the subduction and underplating of Eocene basalts, which may represent a seamount chain, an arc, or a backarc basin (Clowes et al. 1986, 1987). Since the late Eocene, the subduction zone has developed an accretionary complex without the involvement of any allochthonous masses, and no terrane amalgamation events are anticipated for the immediate geological future. To the north, off Queen Charlotte Islands, the

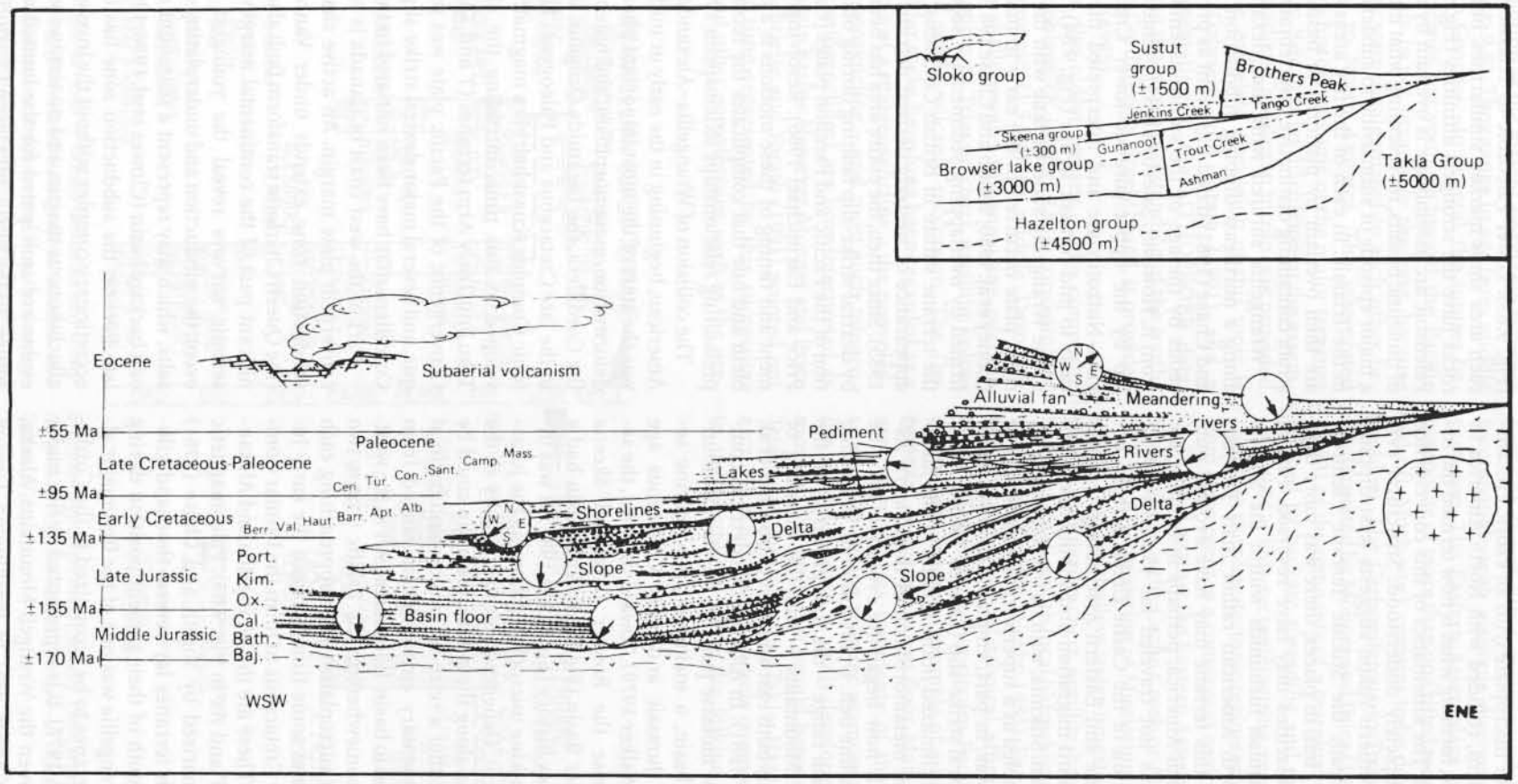


Fig. 9.85. Composite stratigraphic cross section of Bowser Basin, showing major depositional systems and transport directions. (Eisbacher 1981)

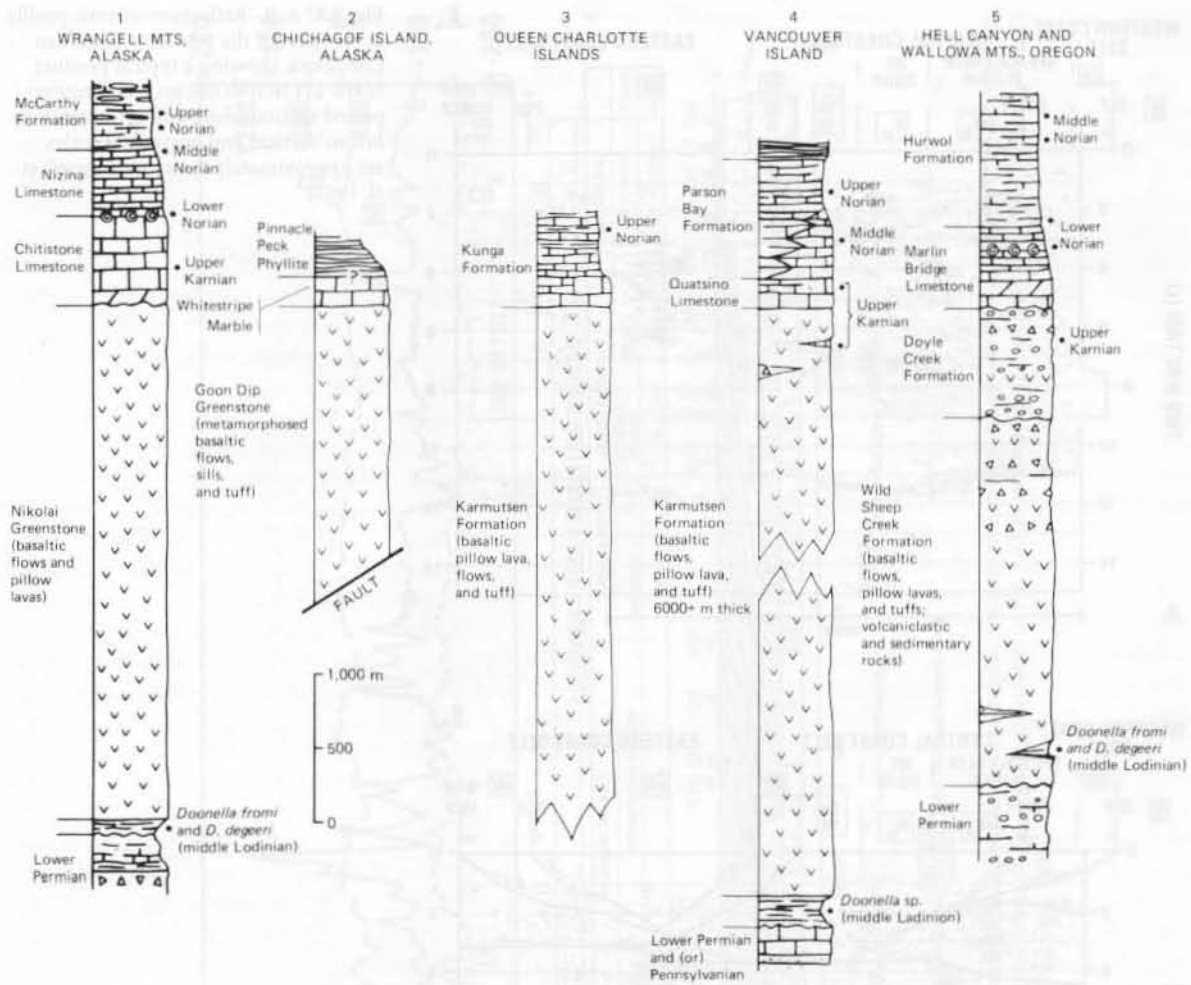


Fig. 9.86. Representative Triassic sections in each of the five fragments of the Wrangellia terrane (*W* in Fig. 9.83). The stratigraphic similarities between these sections were the main basis for proposing their original combination into a single microplate. (Jones et al. 1977)

subduction zone ends at a ridge–trench–transform triple junction, and the continental margin off northern British Columbia is a transform plate boundary, undergoing dextral strike-slip (Fig. 9.48C).

The welding on (obduction) of successive terranes resulted in extensive underplating of the crust, wedging of crustal slices into older terranes, and delamination of the crust from the autochthonous basement along major, nearly flat-lying décollement surfaces (Price 1986; Gabrielse and Yorath 1991). An example of a reflection-seismic profile across part of the southern Canadian Cordillera, showing the resulting structural complexity, is shown in Fig. 9.87, and a reconstruction of the entire Cordillera in southern Canada is shown in Fig. 9.88.

This account is necessarily highly simplified, and it ignores the chronology of terrane accretion in Alaska and areas south of the United States border. In fact, the story is by no means complete, and large teams of Canadian and American geologists are grappling with the many problems that the terrane model has raised, and probably will be doing so for many years to come.

Similar analyses have been undertaken for other orogens. Nur and Ben-Avraham (1982) reviewed many of the world's orogenic belts and demonstrated that most are constructed of accreted terranes. For example, Dewey et al. (1973) showed that the Alpine orogenic belt was constructed of many terranes welded together between the Jurassic and late Cenozoic (this work was actually carried out long before the terrane concept became fashionable). Nur and Ben-Avraham (1982) suggested that there never was a full-scale collision between Africa and Europe, the event that supposedly created the Alpine mountain chain. Further east, the Himalayan orogen of west China is now known to consist of many terranes and micro-

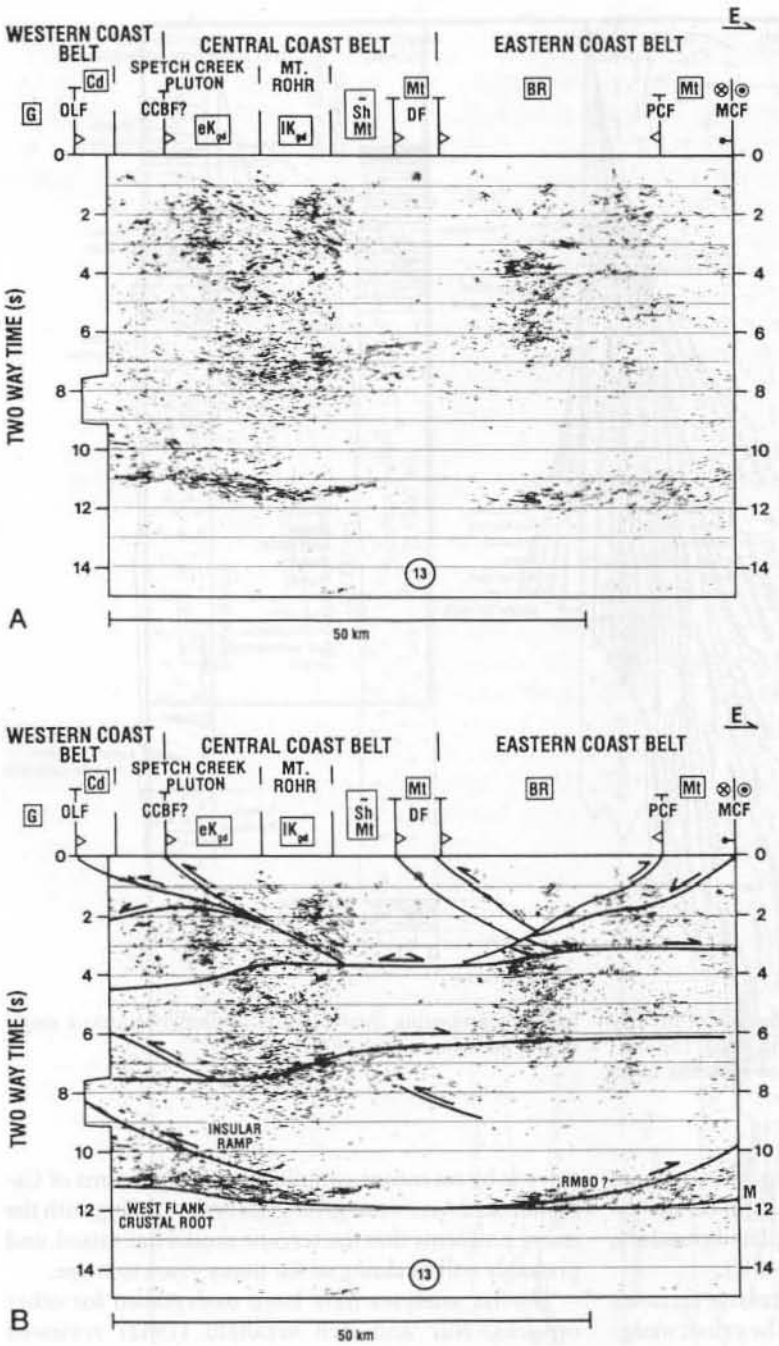


Fig. 9.87 A, B. Reflection-seismic profile across part of the southern Canadian Cordillera, showing a typical product of the LITHOPROBE project. Uninterpreted section *above* and interpretation *below*. Vertical and horizontal scales are approximately the same. (Varsek et al. 1993)

plates formerly scattered within the Tethyan Ocean (Zhang et al. 1984; Şengör 1986; Hsü 1988).

The British Isles may be a composite of fragments representing at least four major terranes: England–Wales–southern Ireland sutured to central and southern Scotland during the Ordovician–Devonian (Phillips et al. 1976; Pickering et al. 1988); southernmost England sutured on during the Hercynian collision of Europe with Africa (Dewey and Burke 1973); and northern Scotland was emplaced by sinistral transform faulting during the same Carboniferous event (Kent and Opdyke 1978; Van der Voo and Scotese 1981; Pickering et al. 1988). Dewey (1982) provided an eloquent discussion of this accretionary process and showed how the evolution of the British islands at several stages during the Phanerozoic could be compared with the present configuration of various parts of central and eastern Asia. For example, Britain in the middle Devonian, following the final closure of the Caledonian suture, is tectonically com-

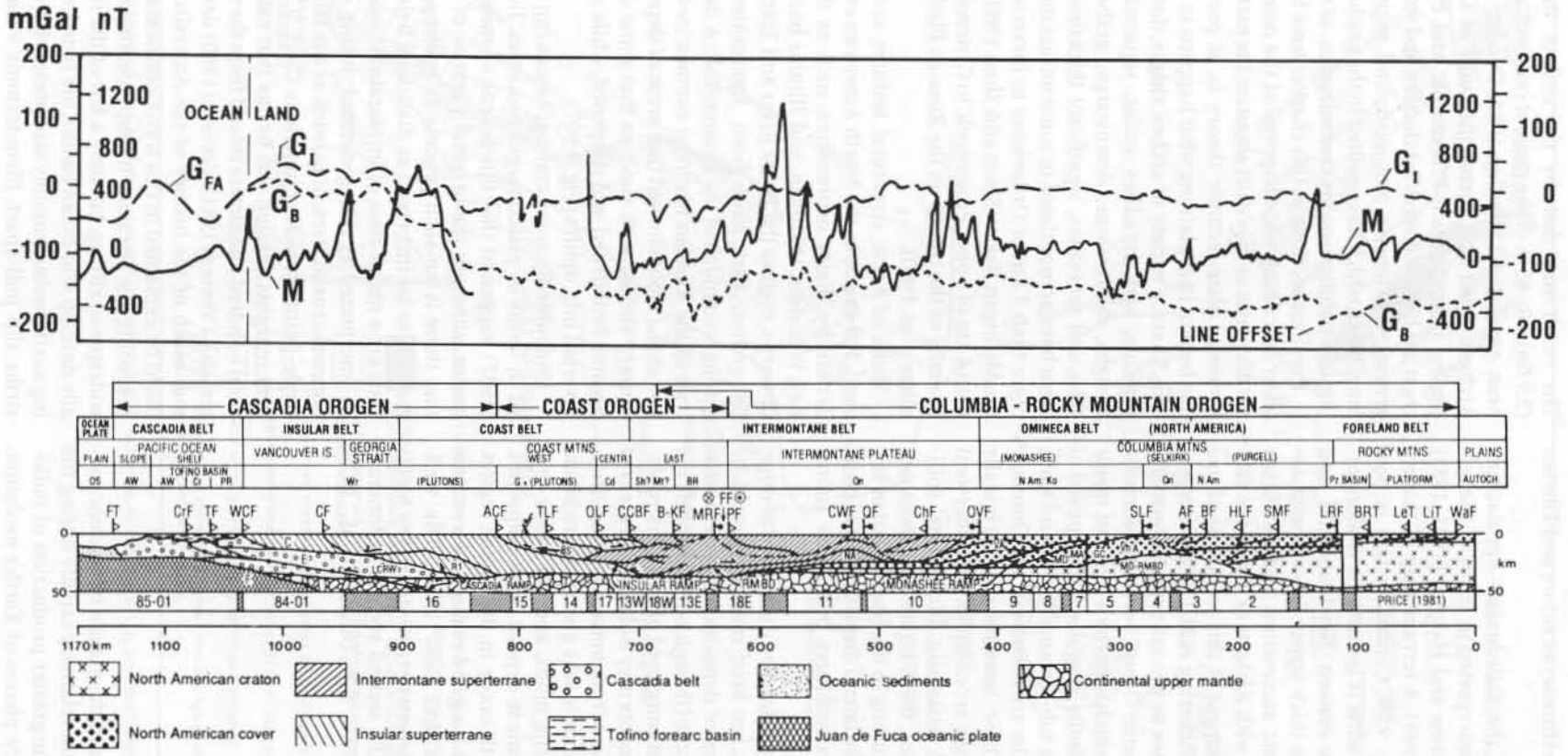


Fig. 9.88. Interpretation of a seismic transect across the southern Canadian Cordillera, from the modern subduction complex off Vancouver island to the plains of the Alberta Foreland Basin. Note the extensive underplating and delamination of all the accreted terranes. The profile illustrated in Fig. 9.87 shows the Coast Orogen at the center of the Cordillera. G_B Bouguer gravity; G_{FA} free-air gravity; G_I isostatic gravity; M magnetics. (Cook et al. 1995)

parable to the modern Himalayan suture and Tibetan plateau.

Britain is part of the Caledonian–Appalachian Orogen, which was reinterpreted in terms of the terrane concept by Williams and Hatcher (1982, 1983) and Williams (1984, 1985). A terrane map of this region is shown in Fig. 9.89. Central Britain is now interpreted as part of what is termed the Avalon terrane, which includes eastern Newfoundland and which is defined by its thick upper-Precambrian sedimentary and volcanic successions and Cambrian quartzites and shales with African or European faunal affinities. The basement of this terrane is of uncertain affinities, but Pickering et al. (1988) interpreted it as consisting of one or two microplates formerly located within the Iapetus Ocean, between Laurentia and Africa. Scotland, which is cut by a series of major northeast–southwest faults, is probably composed of at least five component sub-terranes (Stone and Kimbell 1995), including the arc massifs of the Dunnage terrane (Fig. 9.81). These terranes may originally have been part of a single arc complex but underwent substantial left-lateral strike-slip during final collision and closure during the Silurian and Devonian (Bluck 1985, 1991), resulting in the faulting out of key arc components and the lateral displacement of post-suture sedimentary assemblages from their source terranes (Fig. 9.81).

Space does not permit detailed discussion of other accretionary terranes, but brief mention should be made here of two types of depositional systems that may be expected to occur (though not exclusively) in a tectonic environment composed of small microplates, where plate movements may isolate small areas of an ocean or a backarc sea from normal marine circulation. These are (1) euxinic seas giving rise to black, organic, and sulphurous muds, and (2) saline seas that may totally desiccate to deposit thick evaporites. Good examples of both occur in the Alpine region. The modern Black Sea is a good example of a euxinic environment (Degens and Ross 1974). The well-known Messinian (Miocene) evaporites of the Mediterranean exemplify the second style of sedimentation (Hsü et al. 1973; Biju-Duval et al. 1978; Sects. 4.6.6, 6.7).

9.3.6 Cratonic Basins

9.3.6.1 *Tectonic Review*

Thermal fractionation of the mantle to produce continental crust took place relatively rapidly during the Archean because of the greater production of radioactive heat in the early phases of Earth's evolution. The result was that, by the end of the Archean (2.5 Ga), 50–65% of the present continental mass had been generated (Windley 1984; Condie 1997). This Archean crust now forms the cores of shield complexes, around which Proterozoic and Phanerozoic crust has been accreted. Subduction and suturing have generated extensive plutonism and regional metamorphism, which has resulted in the gradual thickening and stabilization, or **cratonization**, of the crust.

Previous sections of this chapter have been devoted to describing what happens at the margins of the earth's plates as they drift about on the earth's surface. However, plate-tectonic theory is, at present, much less helpful in explaining what happens in plate interiors. There, a variety of arches, ridges, domes, swells, plateaus, and anticlises occur, separating basins, troughs, depressions, downwarps, graben, aulacogens, and syneclises. Significant thicknesses of sediment have accumulated in some intracratonic basins; more than 3 km of Ordovician to Jurassic section in the Michigan Basin (Sleep and Sloss 1980) and more than 4 km of upper Proterozoic to Cenozoic sediment in many of the basins on the Russian Platform (Aleinikov et al. 1980).

Some of these structural features are very long lived. For example, the North American craton is underlain by major depressions, such as the Hudson Bay, Williston, Michigan, and Illinois basins, and by upwarps, such as the Peace River and Transcontinental arches, that have been intermittently active throughout much of the Phanerozoic. A detailed analysis of the Russian Platform summarized by Aleinikov et al. (1980) showed that areas of depression and upwarp shifted with time, so that some depressions became broadened and deepened, while others were inverted into uplifts (Fig. 9.90).

Undoubtedly, some cratonic basins can be explained in terms of plate-margin processes. In fact, Miall (1987) suggested that this is true of most intraplate basins, although the original premise of this paper, that there is no such process as epeirogeny, is now realized to be incorrect, as discussed below. Aulacogens are a major class of intracratonic basin but may be interpreted as rifts generated during episodes of regional crustal stretching, such as the failed arms of triple-point spreading junctions (Sect. 9.3.1). Divergent margins develop rim basins that extend for several hundred kilometers inland from the rifted-margin arch (Veevers 1981). Veevers (1980) demonstrated that much of the history of the Australian platform could be explained in this way. Continental extension preceding the breakup of supercontinents leads to widespread rifting, but only a few of the rifts eventually develop into ocean basins, the remainder surviving as rift basins, possibly also preserving evidence of strike-slip pull-apart. Plate-tectonic events, such as

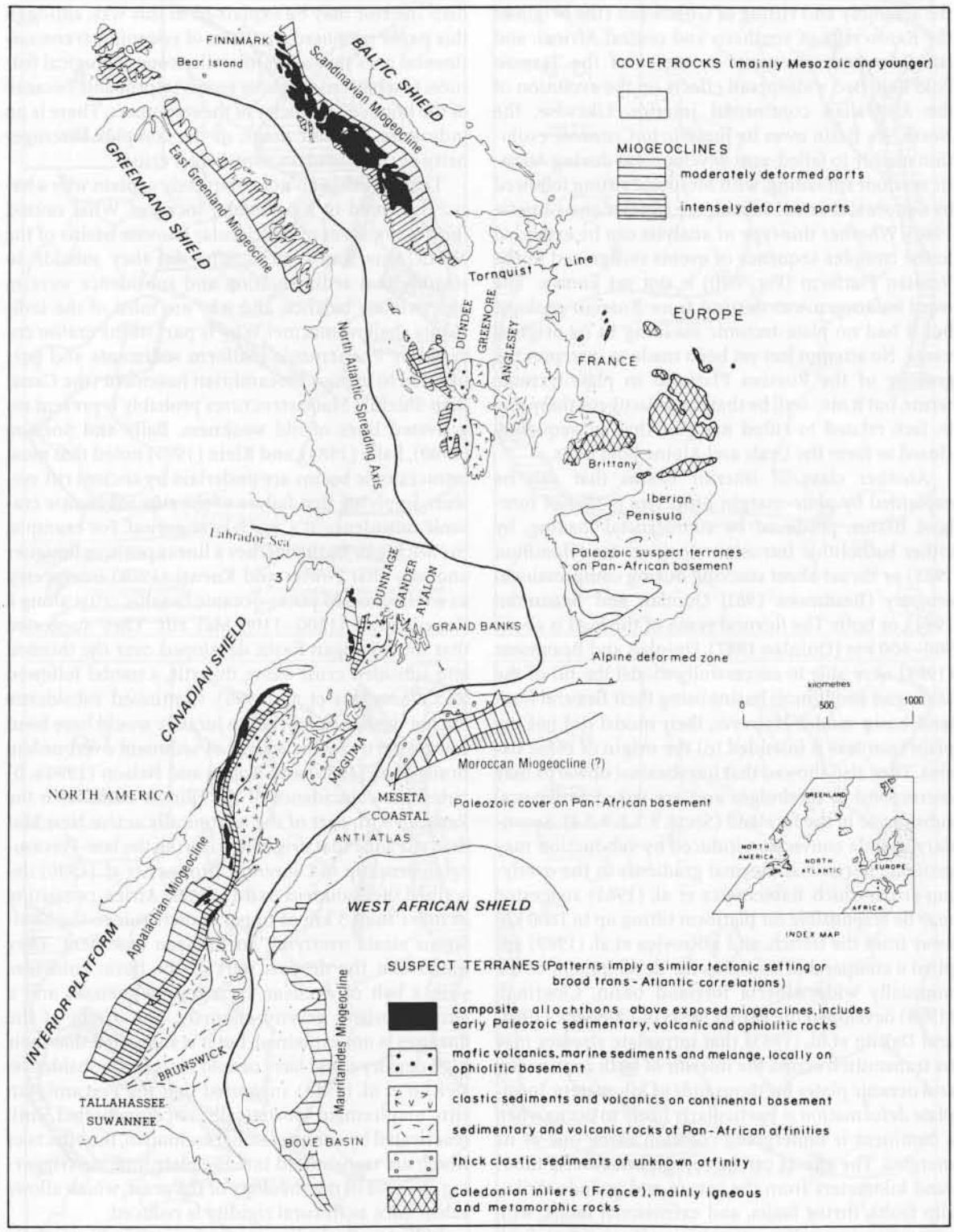


Fig. 9.89. Terranes of the Caledonian orogen of eastern North America, northern Africa, and northwest Europe. (Williams 1985)

the assembly and rifting of Gondwana (the origin of the Karoo rifts of southern and central Africa), and late-Paleozoic convergent tectonics of the Tasman Fold Belt, had widespread effects on the evolution of the Australian continental interior. Likewise, the North Sea Basin owes its Jurassic to Cenozoic evolution mainly to failed-arm development during Atlantic seafloor spreading, with Mesozoic rifting followed by Cenozoic thermal subsidence (Sclater and Christie 1980). Whether this type of analysis can be extended to the complex sequence of events recognized in the Russian Platform (Fig. 9.90) is not yet known. The word **aulacogen** was derived from Russian geology, but it had no plate-tectonic meaning in its original usage. No attempt has yet been made to interpret the geology of the Russian Platform in plate-tectonic terms, but it may well be that the aulacogens there are, in fact, related to rifted margins that subsequently closed to form the Urals and Alpine Fold belts.

Another class of interior basins that can be explained by plate-margin processes is that of foreland basins, produced by supracrustal loading, by either batholithic intrusion along an arc (Hamilton 1985) or thrust-sheet stacking during compressional orogeny (Beaumont 1981; Quinlan and Beaumont 1984), or both. The flexural reach of the load is about 500–600 km (Quinlan 1987). Quinlan and Beaumont (1984) were able to successfully model the fill of the Michigan and Illinois basins using their flexural foreland-basin model. However, their model did not explain (nor was it intended to) the origin of these basins. They also showed that intrabasinal upwarps may correspond to forebulges and are yoked to flexural subsidence in the foreland (Sects. 9.3.2, 9.3.4). Secondary mantle convection induced by subduction may generate horizontal thermal gradients in the overlying crust, which Rabinowicz et al. (1983) suggested may be responsible for platform tilting up to 1600 km away from the trench, and Mitrovica et al. (1989) applied a comparable model to the development of the unusually wide Alberta foreland basin. Cloetingh (1988) developed the earlier ideas of Lambeck (1983) and DeRito et al. (1983) that intraplate stresses may be transmitted across the interior of both continental and oceanic plates for thousands of kilometers. Intraplate deformation is particularly likely to occur when a continent is undergoing collision along one of its margins. The effects can be recognized several thousand kilometers from the suture and include strike-slip faults, thrust faults, and extensional faults, with accompanying sedimentary basins (Sect. 9.3.4). Such basins are intraplate in position, but are not intracratonic in a plate-tectonic sense, because their structure and stratigraphy may be explained with reference to plate-margin processes. Sanford et al. (1985) suggested that many of the cratonic structures in the Canadian interior may be explained in this way, although this paper proposed a pattern of sweeping, transcontinental axes that align miscellaneous geological features in a pattern that now seems improbable because of the implied simplicity of the structures. There is no indication, in their maps, of the complex heterogeneity of the Canadian continental crust.

These models do not completely explain why a basin is formed in a particular location. What caused the development of the circular to ovate basins of the North American craton? Why did they subside so steadily that sedimentation and subsidence were in approximate balance, and why are most of the sediments shallow marine? Why is part of the craton covered by Phanerozoic platform sediments and part uplifted to expose Precambrian basement (the Canadian Shield)? Many structures probably represent reactivated lines of old weakness. Bally and Snelson (1980), Bally (1984), and Klein (1995) noted that most intracratonic basins are underlain by ancient rift systems, implying that failure of the rifts may cause cratonic subsidence at a much later period. For example, the Michigan Basin overlies a linear positive Bouguer anomaly that Fowler and Kuenzi (1978) interpreted as a reflection of proto-oceanic basaltic crust along a Keweenawan (1200–1100 Ma) rift. They suggested that the Michigan Basin developed over the thinned and subsided crust along this rift, a model followed by Catacosinos et al. (1990). Continued subsidence during the late Cambrian to Jurassic would have been caused partly by the weight of sediment overburden. Braile et al. (1986) and Kolata and Nelson (1990a, b) noted the coincidence of the Illinois Basin with the Reelfoot Rift, part of the tectonically active New Madrid rift zone that originated during the late-Precambrian breakup of Laurentia. Bronner et al. (1980) described the Taoudeni Basin of West Africa, consisting of more than 3 km of upper Proterozoic to Carboniferous strata overlying an Archean basement. They noted that the deepest part of the basin coincides with a belt of Archean ferruginous gneisses and a strong positive gravity anomaly. The origin of the gneisses is not explained, but it is suggested that their high density could have caused increased subsidence. DeRito et al. (1983) suggested that the Precambrian rifts may remain isostatically uncompensated until reactivated by compressional tectonism, the effects of which are transmitted into the plate interior, triggering changes in the rheology of the crust, which allows subsidence as flexural rigidity is reduced.

Williston Basin lies across a possible Proterozoic suture that Camfield and Gough (1977) traced into the subsurface by geophysical methods. To the north, this anomaly lines up with the Wollaston Fold Belt in the Churchill Province of the Canadian Shield (Donaldson et al. 1976). Thomas et al. (1987) showed two

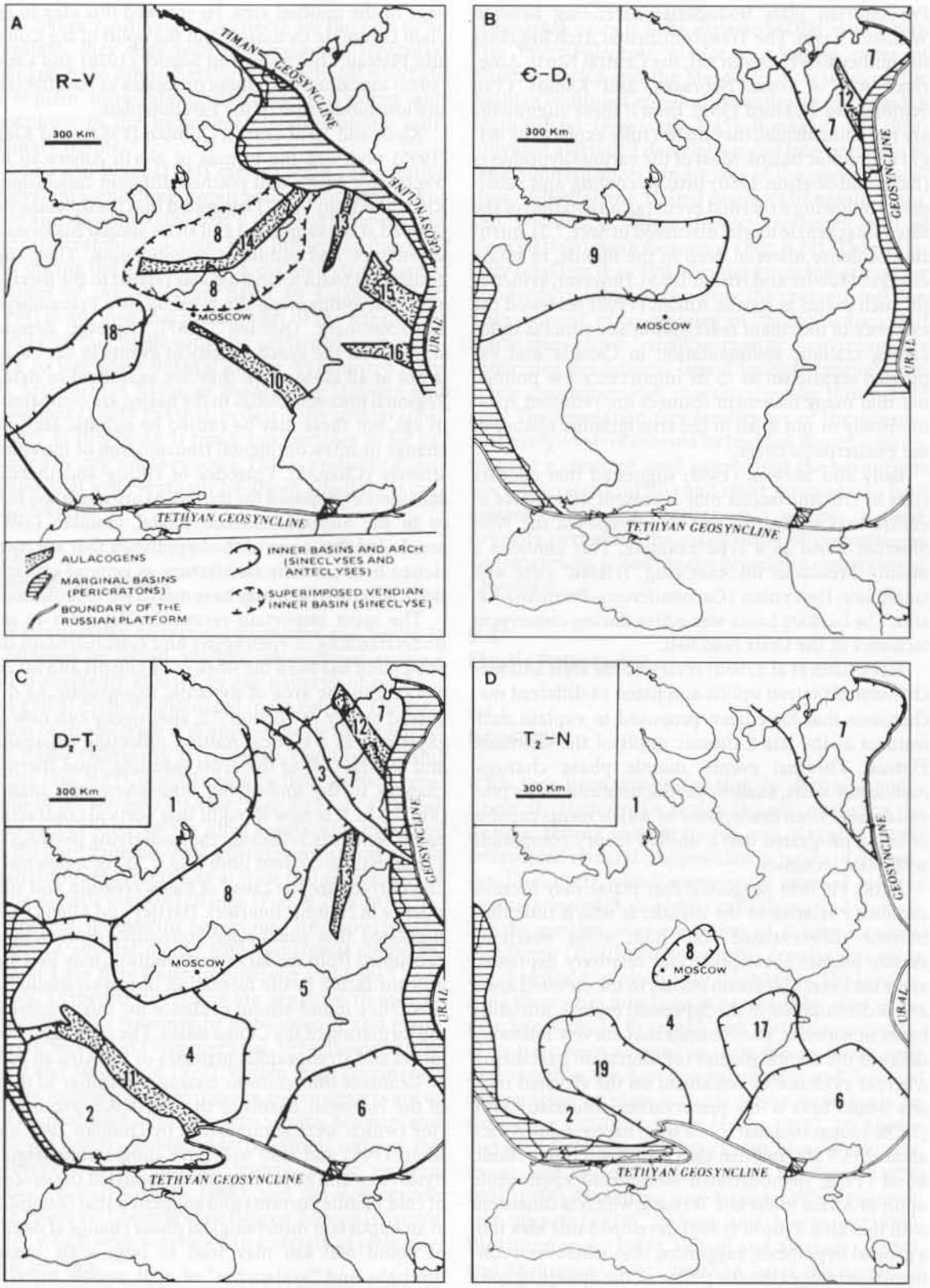


Fig. 9.90 A – D. Evolution of basins in the Russian Platform. **A** Late Proterozoic. **B** Early Paleozoic. **C** Middle and late Paleozoic. **D** Mesozoic and Cenozoic. (Aleinikov et al. 1980)

Precambrian plate boundaries extending beneath Williston Basin. The Transcontinental Arch lies close to another Keweenawan rift, the Central North American Rift of Iowa, Nebraska, and Kansas (Van Schmus and Bickford 1981). Even if these alignments are not coincidental, they do not fully explain the origin of cratonic basins. Most of the various hypotheses (Bally and Snelson 1980) involve cooling and subsidence following a thermal event (an adaptation of the Sleep–Mackenzie model discussed in Sect. 7.3), intrusion of dense material deep in the mantle, or phase changes (Fowler and Nisbet 1985). However, evidence for such events is sparse. Aitken (1993) reviewed the evidence of basement reactivation as a process influencing cratonic sedimentation in Canada and expressed scepticism as to its importance. He pointed out that many basement features are reflected spasmodically or not at all in the stratigraphic history of the Phanerozoic cover.

Bally and Snelson (1980) suggested that another class of cratonic basins may represent subsidence of earlier backarc basins, and they proposed the West Siberian Basin as a type example. This contains a mainly Mesozoic fill overlying Triassic rifts and small, late-Hercynian (Carboniferous–Permian) basins. The backarc basin was active during convergent tectonics of the Urals Fold Belt.

McGetchin et al. (1980) reviewed the style and mechanism of plateau uplifts and listed 14 different mechanisms that have been proposed to explain such features as the late-Cenozoic uplift of the Colorado Plateau. Thermal events, mantle phase changes, mantle hot spots, shallow subduction, and other proposals have been made, none of which seems capable of being integrated into a unified theory compatible with plate tectonics.

Burke (1976b) suggested that plates may become stationary relative to the mantle, at which time they become differentiated into high areas overlying mantle plumes (hot spots) and relatively depressed areas between. Volcanism occurs in the elevated areas and sedimentation in the depressed regions, initiating basin subsidence. These basins may survive following decay of the mantle plumes (or migration past them), whereas evidence of volcanism on the elevated regions would have a low preservation potential. Burke (1976b) suggested that this is what happened to Africa about 25 ± 5 Ma, and that the Chad Basin is one result. Bond (1978) demonstrated widespread epeirogenic uplift of Africa in the late Tertiary, which is consistent with this idea. Wilson (1988) developed this idea into a general hypothesis, suggesting that, wherever a continent overrides a mantle plume or the upwelling zone along a spreading center, regional uplift develops. Gravitational sliding of the plate off the uplift may, then, be responsible for compressional tectonism at the edge of the uplifted area. He invoked this idea to explain Laramide tectonism and the uplift of the Colorado Plateau. Dickinson and Snyder (1978) and Cross (1986) explained both these processes as the effects of shallow subduction of the Farallon plate.

Klein and Hsui (1987), Quinlan (1987), and Klein (1995) reviewed the origins of North American intracratonic basins and reached different conclusions. Klein and Hsui (1987) suggested that these basins developed at the same time and share similar histories of subsidence and sediment accumulation. They suggested that basin formation was related to the thermal effects accompanying the breakup of a Precambrian supercontinent. Quinlan (1987), however, demonstrated that the synchronicity of events in the basins is not at all close when they are examined in detail. Regional unconformities in the basins are comparable in age, but these may be caused by eustatic sea-level change or intracontinental transmission of intraplate stresses (Chap. 8). Episodes of rifting and thermal subsidence proposed for the basins are much too long to fit the Sleep–Mackenzie model. Quinlan (1987) concluded that none of the hypotheses that are mentioned here are fully satisfactory as general explanations, although each may have merits for specific cases.

The most important recent development in our understanding of epeirogeny and cratonic-basin development has been the work of M. Gurnis and his coworkers in the area of dynamic topography. As discussed briefly in Section 7.5, epeirogeny can now be explained as a vertical motion reflecting expansion and contraction of the crust resulting from thermal changes in the underlying lithosphere and mantle (Fig. 7.32). It is now thought that vertical contractional movements, reflecting the underlying presence of a cold mantle-current limb (Fig. 7.33) or a downwelling current are the cause of basin creation and subsidence in cratonic interiors. Hartley and Allen (1994) suggested that small-scale convective downwelling, decoupled from the large-scale motion, may be a significant factor in the formation of intracratonic basins. They found strong evidence for this process in the formation of the Congo Basin. The structural evolution and stratigraphic histories of the African suite of Cenozoic intracratonic basins are similar to those of the Paleozoic basins of the North American interior (which were summarized by Quinlan 1987 and Klein 1995) and may well have comparable origins. Pysklywec and Mitrovica (1998) modeled the descent of cold mantle currents and suggested that instability at an important mineralogical phase change at depths of about 600 km may lead to large-scale breakthroughs and "avalanches" of cold mantle material deeper into the mantle. This is capable of generating surface deflections 100- to 1000-m deep over wavelengths of about 1000 km and with durations of

100–200 m.y. Although the effects are thought to be transient, they may lead to permanent structural changes in the lithosphere, and this process may help to explain intracratonic downwarping events and contribute to the formation of intracratonic sedimentary basins.

Loup and Wildi (1994) carried out a comparative analysis of some three-dozen basins in northwest Europe and demonstrated that a range of plate-margin and intracratonic processes have influenced these basins during the last 250 m.y. A detailed stratigraphic analysis of the Michigan Basin by Coakley et al. (1994) confirmed that at least part of the subsidence history of this basin could be explained with reference to intraplate flexural effects transmitted from the Appalachian basin, as modeled by Quinlan and Beaumont (1984), but that, in detail, the stratigraphic history of this basin is much too complex to be explained by any single process. This is probably a conclusion that will eventually be drawn about most cratonic basins.

9.3.6.2 *Sedimentation Patterns*

The stratigraphy of cratonic platforms and basins is dominated by broad, shallow depositional systems and is markedly affected by the patterns of regional and global cyclicity discussed in Chapter 8. The evolution of the Silurian reef and evaporite fill of Michigan Basin (Mesolella et al. 1974), Cambrian Grand Cycles (Fig. 8.21), and 10^6-year cyclicity in the Cretaceous of the Western Interior (Figs. 8.19) are excellent examples of this, but the most spectacular examples are the Carboniferous coal-bearing cyclothems, which retain internal evidence of at least two time scales of cyclic eustatic sea-level change (Figs. 8.23, 8.28).

Internal, nonmarine basins may be characterized by broad alluvial plains with centripetal drainage, such as Lake Eyre Basin in Australia and Lake Chad, Africa. Dispersal directions are sensitive to epeirogenic tilting in response to plate-interior and plate-margin tectonism (Wopfner et al. 1974; Doutch 1976; Ward and McDonald 1979). Within large continental masses in subequatorial regions, eolian deserts are to be expected, such as the modern Sahara, Arabian, and Simpson deserts, located in intracratonic settings, the Carboniferous to Jurassic deserts of the western United States (Kocurek 1988), and the Rotliegende (late Paleozoic) deserts of northwest Europe (Glennie 1983), which were developed in a variety of plate-tectonic settings.

Both marine and nonmarine basins may be characterized by layer-cake-type stratigraphy, with some individual lithosomes traceable for thousands of kilometers. Intimate interbedding of marine and nonmarine units is to be expected (Dott et al. 1986). The lower-Paleozoic platform-carbonate cover of North America and the Mesozoic succession on the Arabian continent are good examples, as discussed in Section 6.6.1 (Figs. 6.44, 6.45). Runkel et al. (1998) provided a detailed description and interpretation of the Cambrian sheet sandstones of the upper Mississippi Valley of the United States (Minnesota, Iowa, and Wisconsin). These deposits represent part of Sloss's (1963) Sauk Sequence (Fig. 8.14). They typically total less than 100 m in thickness but extend for hundreds of kilometers across the cratonic interior. They consist of a complex of fluvial, eolian, and storm-dominated shallow-marine environments containing gradational lateral facies transitions and subtle unconformities characterized by thin pebble lags. The vegetation-free landscapes of the nonmarine margins were crossed by braided-fluvial systems that fed a constant sand supply into the basin from low intracratonic uplifts and relict Precambrian land masses. A sequence-stratigraphic analysis of these deposits permitted the recognition of conventional systems tracts, following the principles elaborated in Chapter 8.

9.4 Clastic Petrofacies

9.4.1 Introduction

It has long been recognized that the detrital composition of clastic rocks is significantly related to the tectonic setting of their source area. Krynine (1942) interpreted mineral composition in terms of the geosynclinal cycle, which supposedly affected rates of source-area uplift and basin subsidence and controlled the geological composition of source terranes. Middleton (1960) reached similar conclusions using the chemical composition of sandstones. The geosyncline theory has now been superseded by plate tectonics, but many of Krynine's and Middleton's conclusions have been adapted to plate theory. Some early attempts to reinterpret detrital mineral and chemical composition in plate-tectonic terms were made by Dickinson (1970), Schwab (1971, 1975), Dickinson and Rich (1972), and Crook (1974). More recent syntheses and reviews include those by Potter (1978, 1984, 1986), Dickinson and Suczek (1979), Ingersoll and Suczek (1979), Valloni and Maynard (1981), Dickinson et al. (1983), Valloni and Mezzadri (1984), and Dickinson (1985, 1988). A brief discussion of modern ideas on the tectonic control of mineral composition is presented here.

It should be pointed out that detrital composition is controlled by several other important variables, in

addition to tectonic provenance. These include transport history (Suttner 1974; Franzinelli and Potter 1983), depositional environment (Davies and Ethridge 1975; Sect. 4.5.2), and paleoclimate (Basu 1976; Suttner et al. 1981). It should be noted that recycling, long transport distances, or tropical weathering will tend to increase the quartz content of the detritus at the expense of most other grains, which could considerably distort provenance interpretations (Mack 1984; Chandler 1988; Dickinson 1988), nor can diagenetic modifications of sand content be ignored (McBride 1985). Also, Schwab (1978) demonstrated gross changes in the chemical composition of sediments through geologic time. No attempt has yet been made to integrate the effects of all these controls, although Potter (1978) discussed the inter-relationship of climate and tectonics. Dickinson and Suczek (1979) did not deal with environmental, climatic, or time variables, and Valloni and Maynard (1981) restricted their discussion to modern deep-sea sands. Ingersoll (1978b, 1983) carried out statistical tests of the relationship between composition and depositional facies in the upper Mesozoic Great Valley Group of California and showed that, in this case, the relationship was weak and was far overshadowed by the control imposed by changes in source-area geology. It may be that this is a general rule in tectonically active settings, but far more study of the problem is required.

The value of petrofacies studies is that they help elucidate the tectonic setting for ancient detrital sequences, particularly those far removed from their original setting by some vagary of the Wilson cycle. Also, the appearance of a particular mineral assemblage in a stratigraphic succession may indicate an important tectonic event, such as uplift and erosion of an arc plutonic suite or an ophiolite assemblage along a suture, and this may help date the time of intrusion or of continental collision (Ingersoll 1978b, 1983; Eisbacher 1981; Schwab 1981; Dickinson et al. 1983; examples are discussed below.) Petrofacies studies are also of considerable importance in the evaluation of terrane-accretion processes. For example, the arrival of distinctive detritus from one terrane in a basin located on an adjacent terrane may help date the time of suturing, and the detrital composition of a suture assemblage or overlap assemblage may throw light on the nature of terrane displacement (Graham et al. 1984; Bluck 1985, 1991; Kleinspehn 1985; Dickinson 1988; Sect. 9.3.5). These applications of petrographic data are an extension of the petrographic-mapping technique discussed in Section 5.5, in that not only can the data be used to determine the source area and paleocurrent directions, but a plate-tectonic interpretation can be made from the same data. Refer to Section 5.5 for a discussion of grain-counting methods.

9.4.2 Sandstone Composition and Plate-Tectonic Setting

Dickinson and Suczek (1979), Valloni and Maynard (1981), and Ingersoll et al. (1984) found that the following modal quantities were useful in diagnosing the tectonic environment [Dickinson (1970) discussed the practical details of grain recognition and classification; Sect. 5.5]:

Q: Monocrystalline and polycrystalline quartz grains, including chert

$Q = Qm + Qp$

Qm: Monocrystalline quartz
Qp: Polycrystalline quartz, including chert
F: Monocrystalline feldspar

$F = P + K$

P: Plagioclase
K: Potassium feldspar
Lv: Volcanic (hypabyssal) fragments
Ls: Sedimentary fragments
Lm: Metamorphic fragments
L: Unstable polycrystalline lithic fragments

$L = Lv + Ls + Lm$

Lt: L+Qp
D: Dense minerals

Care must be taken to distinguish recrystallized detrital matrix (orthomatrix), diagenetic pore-filling matrix (epimatrix), and squashed lithic grains (pseudomatrix), not always an easy matter in immature lithic sandstones (Dickinson 1970). Care should also be taken to distinguish intrabasinal fragments (e.g., some carbonate particles) from extrabasinal detritus (Zuffa 1985).

Dickinson and Suczek (1979) distinguished nine provenance types based on tectonic setting:

1. Continental
   a. Craton interior
   b. Transitional
   c. Uplifted basement
2. Magmatic arc
   a. Dissected
   b. Transitional
   c. Undissected
3. Recyled orogen
   a. Subducted complex
   b. Collision orogen
   c. Foreland uplift

They compiled data from 88 suites of samples representing all these provenance types and ranging in age from Cambrian to modern. Four ternary diagrams were then used to demonstrate the distinctiveness of

the detritus derived from the nine provenance types (Fig. 9.91). The QFL and QmFLt plots both show the entire grain population but differ in the way in which Qp components (primarily chert) are plotted. The QpLvLs and QmPK plots show partial grain populations, and help discriminate some of the provenance types, mainly on the basis of the details of lithic and feldspathic composition.

Using all four plots, the provenance types are seen to yield distinctive detrital suites. The data support many of the ideas of Krynine and the other early workers, once allowance is made for the differences between geosycline theory and modern plate tectonics. Not surprisingly, geochemical data demonstrate the same trends (Bhatia 1983, 1985). Note that it is not necessary to use any formal sandstone classification here and, indeed, many of the provenance fields straddle the boundaries between some of the classic sandstone types, so the use of such terms as lithic wacke or arkose is not very helpful.

Continental provenances are distinguished mainly on the basis of the presence of abundant quartz and a paucity of lithic fragments. These are rather mature sediments and reflect derivation from crystalline Precambrian shield complexes and overlying platform sediments. The latter are typically polycyclic and may contain more than 95% monocrystalline quartz. Lithic fragments may be derived from ancient arcs or metamorphic complexes, the roots of which are commonly exposed in shield areas. K-feldspar is the feldspar most resistant to weathering. Its presence reflects the ultimate derivation from silicic (granitic or rhyolitic) igneous sources, although it can, like quartz, be recycled. The decrease in maturity in continentally derived sandstones as the provenance area becomes uplifted simply reflects the stripping away of the mature platform cover and the exposure of a variety of basement rocks. Sandstones derived from basement uplifts are the most feldspathic of any collected by Dickinson and Suczek (1979). This confirms a long-held impression that arkoses are typically local deposits related to block faulting or residual deposits above a granitic basement (Pettijohn et al. 1972; Potter 1978; Schwab 1981).

Arc-derived sandstones are typically immature, in that they are rich in volcanic, lithic fragments. Undissected arcs yield volcanic detritus containing abundant fine-grained volcanic fragments (Lv), and they tend to be rich in plagioclase. Dissected arcs expose the plutonic core and, in the case of continental (Andean-type) arcs, the local basement. Detrital composition, therefore, approaches that of sediment derived from uplifted continental basement, although the greater abundance of feldspar (particularly plagioclase) and volcanic fragments remains distinctive.

Recycled orogen sandstone is distinguished by an abundance of quartz and sedimentary–metasedimentary lithic fragments. Subduction complexes consist of metasediments and ophiolites and, therefore, yield abundant chert, argillite, and lithic arenite. The presence of abundant chert shows up as a marked shift in position of the data points between the QFL and QmFLt plots in Fig. 9.91. Suture zones and foreland uplifts expose thrust sheets and nappes of continental-margin sediments, together with deformed ophiolites and arc rocks (e.g., Figs. 9.1, 9.24, 9.60, 9.65, 9.76, 9.81). The abundance of sedimentary and metasedimentary rock fragments is, therefore, to be expected. High Qm contents may reflect derivation from deformed, mature platform sediments, whereas high Qp (chert) contents would indicate that the suture zone and/or fold–thrust belt are sources.

No characteristic detrital mineral assemblages have been proposed for basins in strike-slip fault regimes. It seems intuitively unlikely that they could ever be categorized, because they are by their very nature disruptive, juxtaposing source-area terranes of all kinds.

Care should be taken to differentiate between the tectonic setting of the provenance area and that of the basin. In some cases, a sedimentary basin may not receive the greater part of its detritus from a genetically related terrane. Thus, Valloni and Maynard (1981) classified their deep-sea sand samples according to the tectonic setting of the basin and, although most of their results confirm the findings of Dickinson and Suczek (1979), there are a few surprises. For example, modern offshore basins of Argentina are classified as divergent margin basins and would normally be expected to contain mature sands derived from a continental source. However, they proved to be quartz poor and rich in lithic fragments, falling in the arc field of the QFL plot (Fig. 9.91). This reflects transport by contour currents from the Scotia arc to the south, rather than offshore transport from the adjacent continental margin. Potter (1978) reported on a statistical study of the relationship between the tectonic setting of the mouths of major modern rivers and the petrographic composition of sands collected therefrom. In general, the results support those discussed here but, surprisingly, the Ganga, Brahmaputra, and Indus sands are classified as trailing-coast rather than collision-coast suites. These local anomalies should serve as a warning against too facile an interpretation of petrographic data. Ingersoll (1990) pointed out that interpretations of the plate-tectonic setting of sandstones should only be carried out at the largest (basin) scale. Ambiguous or misleading results can be obtained if the data are collected locally and reflect only part of a potential source terrane (Ingersoll et al. 1993).

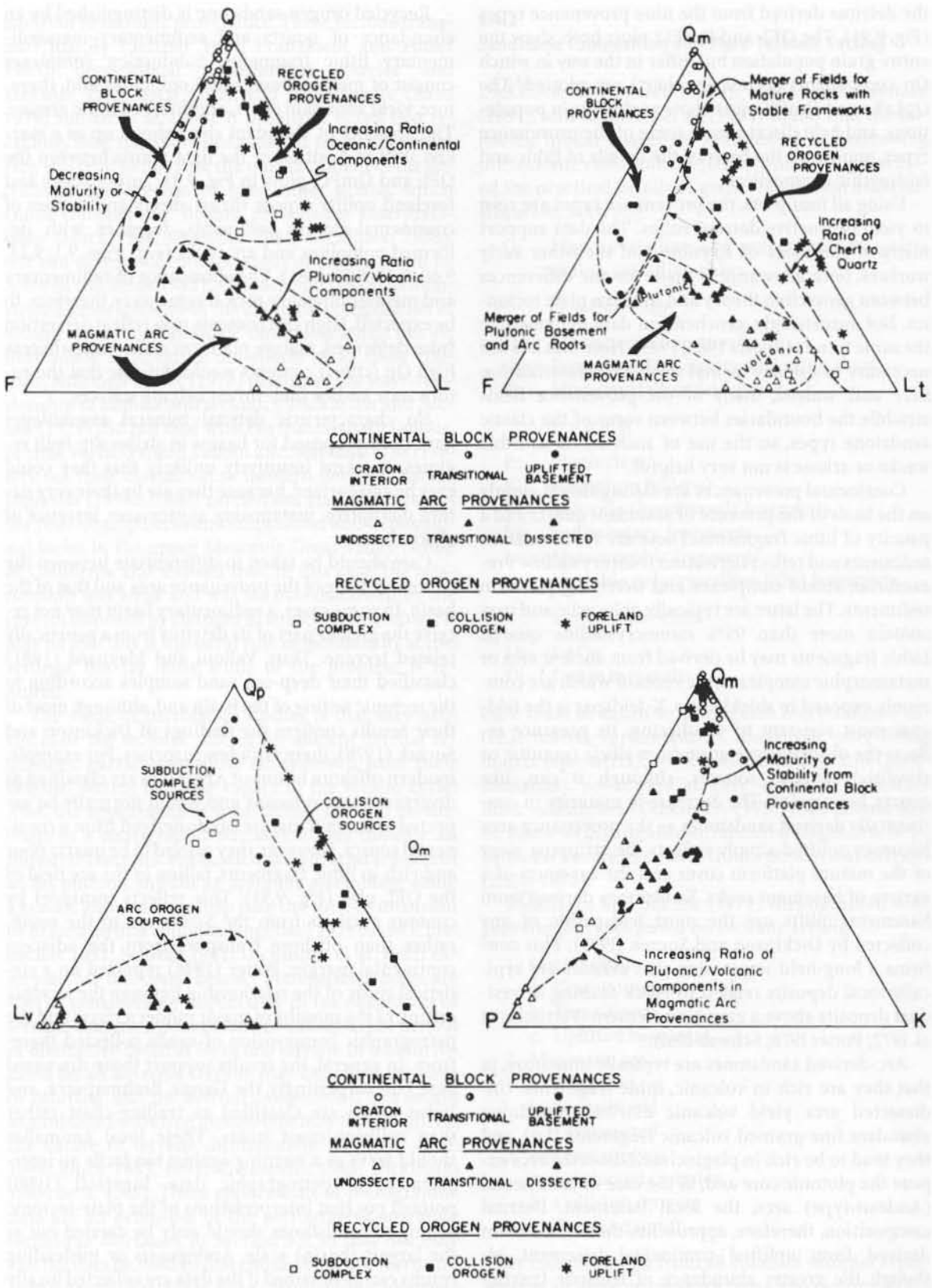


Fig. 9.91. Ternary plots of detrital sandstone composition of 88 sample suites from various tectonic settings. Carbonate grains and heavy minerals omitted. See text for explanation. (Dickinson and Suczek 1979)

Dickinson (1988) noted that there are several ways by which sandstones of mixed petrofacies might be produced. Long-distance (fluvial or marine) transport into a basin in a different tectonic setting and derivation from provenances of mixed petrofacies, such as accretionary collages, are the two principal sources of such mixed compositions. Graham et al. (1993) documented in detail examples of the second process. They studied Neogene to modern sandstones from the intermontane Cenozoic "successor" basins of western China – the basins within the accreted terranes north of the Himalayan orogen, which developed as a result of the India–Asia collision (Fig. 9.77). The source terranes indicated by these samples are highly varied and do not necessarily indicate the correct type of terrane in terms of contemporaneous tectonic setting. For example, sandstones in the Junggar Basin are "arc-derived", in the sense that they are poor in quartz and rich in plagioclase and volcanic rock fragments, but they were deposited in a foreland setting. Examination of the geology of the source area for these sands indicates that it is underlain by a relict arc.

9.4.3 Examples of Petrofacies Interpretation

Dickinson and Rich (1972) and Ingersoll (1978b, 1983) compiled what is probably the most detailed petrographic study of an arc-derived sequence yet attempted. They investigated the upper Jurassic and Cretaceous Great Valley Group of California. In the earlier study, Dickinson and Rich (1972) erected five petrofacies that they found to be locally reliable as chronostratigraphic marker units. They attempted to relate variations between lithic and more quartzofeldspathic petrofacies to magmatic events in the Sierra Nevada source terrane, the sandstones becoming more quartzofeldspathic following the temporary cessation of intrusion and volcanism and the consequent dissection of the arc. Ingersoll (1978b) challenged some of these conclusions on the basis of revised radiometric age scales for the sediments and igneous rocks. He was able to show a north-south variation in petrofacies along the strike of the forearc basin. This can be related to the way in which the Mesozoic arc truncates the late-Paleozoic continental margin. Northern segments of the arc and forearc lay across old oceanic crust, whereas the southern part of the arc was intruded into continental crust. Sediments in the north are, therefore, more lithic, reflecting a greater abundance of andesitic volcanism whereas, in the south, sediment sources consisted of felsic volcanics and metamorphic basement rocks. Intermediate plutonic sources were present throughout the arc. Ingersoll (1983) was also able to recognize the products of a Jurassic suturing by distinguishing suture-derived detritus, and documented an erosional unroofing history showing a change from volcanic to plutonic sources as the arc was uplifted and worn down.

Schwab (1981) sampled Carboniferous to Tertiary sandstones across the western French–Italian Alps (Valence to Turin) and related the detrital assemblages to tectonic evolution from a Tethyan rifted margin to a collisional suture (Figs. 9.92, 9.93). Unlike Dickinson and Suczek (1979), Schwab included carbonate sedimentary fragments in his study; they were assigned to the lithic category. His suite 1 is a typical feldspathic assemblage derived from local fault blocks; suite 2 is a mature cratonic suite deposited during a period of tectonic quiescence. Suite 3 was derived from the Brianconnais, which is interpreted as an isolated continental strip that originated much like Blake Plateau on the Atlantic coast. The mineralogy is typical of a mature continental suite. Suite 4 comprises fault breccias deposited during extensional breakup of the Jurassic carbonate platform. The lithic nature of the sands reflects the abundance of carbonate detritus and, because of this, the assemblage cannot be compared with the petrofacies classes of Dickinson and Suczek (1979). Suites 6 and 7 are from deep-water starved facies; they have the mature characteristics of continentally derived suites. Suites 7 and 8 are also mature and reflect a similar provenance.

Convergent tectonics commenced during the late Cretaceous. The resulting deep-water sands (suite 9) have a composition in the collision-orogen provenance class, rather than the magmatic arc category. Contemporaneous sands to the west (suites 10 and 11) are supermature, indicating that their provenance

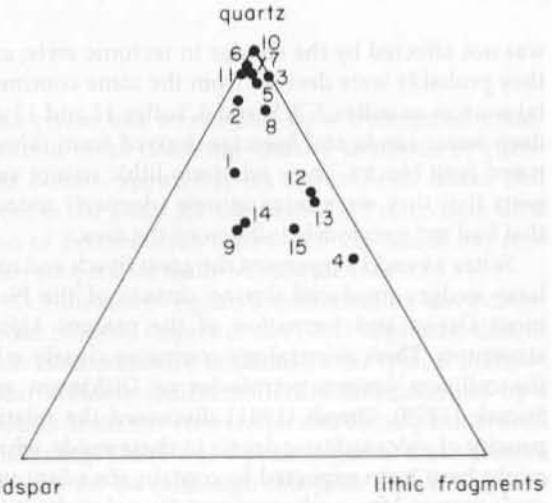


Fig. 9.92. Ternary plot of detrital composition of 15 sample suites from Carboniferous to Tertiary rocks of the western Alps. Carbonate grains are included with lithic fragments. (Schwab 1981)

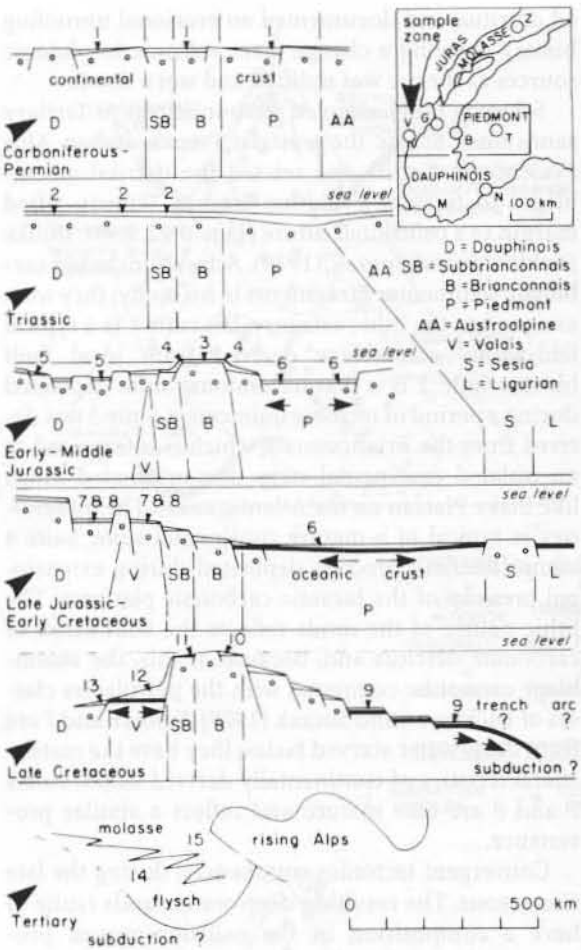


Fig. 9.93. Schematic structural cross sections through the western Alps, showing the interpreted tectonic evolution and location of sample suites plotted in Fig. 9.92. (Schwab 1981)

was not affected by the change in tectonic style, and they probably were derived from the same continental sources as suites 2, 3, 5, and 6. Suites 12 and 13 are deep-water sands and breccias derived from rejuvenated fault blocks. Their relatively lithic nature suggests that they were tapping new (deeper?) sources that had not previously influenced the area.

Suites 14 and 15 represent the great flysch and molasse wedges produced during closure of the Piedmont Ocean and formation of the present Alpine structures. Their mineralogy compares closely with the collision orogen petrofacies of Dickinson and Suczek (1979). Shwab (1981) discussed the relative paucity of volcaniclastic debris in these sands, which might have been expected to contain abundant volcanic material from adjacent arcs. He pointed out the localized distribution of volcanic detritus adjacent to modern arcs, which may result from local trapping or ponding of the debris.

9.5 Basin Models in the Precambrian

9.5.1 Precambrian Plate Tectonics?

The greater part of this book deals with the Phanerozoic record. Although this represents only about the last one-ninth of geological time, the record of the Phanerozoic is so much clearer and more complete than that of the Precambrian, it has provided virtually all the data on which modern basin-analysis methods are based. Previous sections of this chapter deal exclusively with the Phanerozoic record. This emphasis has been chosen deliberately, because the evidence for tectonic processes is so much more clearly preserved in the Phanerozoic record and because much of the economic interest in sedimentary basins revolves around occurrences of coal and petroleum, which are almost exclusively found in Phanerozoic hosts. It is clear that plate-tectonic processes were in operation far back into the Proterozoic, but a consensus now appears to be emerging that the tectonic processes during the Archean were quite different (Goodwin 1991; Rogers 1993; Hamilton 1998), although Condie (1997) summarized evidence that suggested to him that "some form of plate tectonics was in existence by 3.8 Ga."

The two distinctive types of Archean tectonic regime are greenstone–granite belts and granulite–gneiss belts. Greenstone belts (Fig. 9.94) are synformal, linear to curvilinear structures up to 800 km long, containing ultramafic and andesitic volcanics and various sedimentary facies. Windley (1981) and Nisbet (1984) compared the two assemblages to modern backarc-basin and plutonic-arc complexes, respectively. However, as Hamilton (1998) pointed out, these belts lack the evidence of large-scale horizontal plate movement that is characteristic of plate tectonics. Major rift faults, suture zones, ophiolites, etc., are lacking.

Goodwin (1981a), Windley (1981), and Kröner (1981) attempted syntheses of Precambrian tectonic and sedimentary evolution that form the basis for modern interpretations by most specialists. Goodwin (1981a) stated:

> The Precambrian record is viewed as moving in the direction of increasing continental stability from an early highly mobile microplate tectonics phase, through a more stable, largely intracratonic, ensialic, mobile belt phase, to the modern macroplate tectonics phase involving large, rigid lithospheric plates with modern continents and ocean basins.

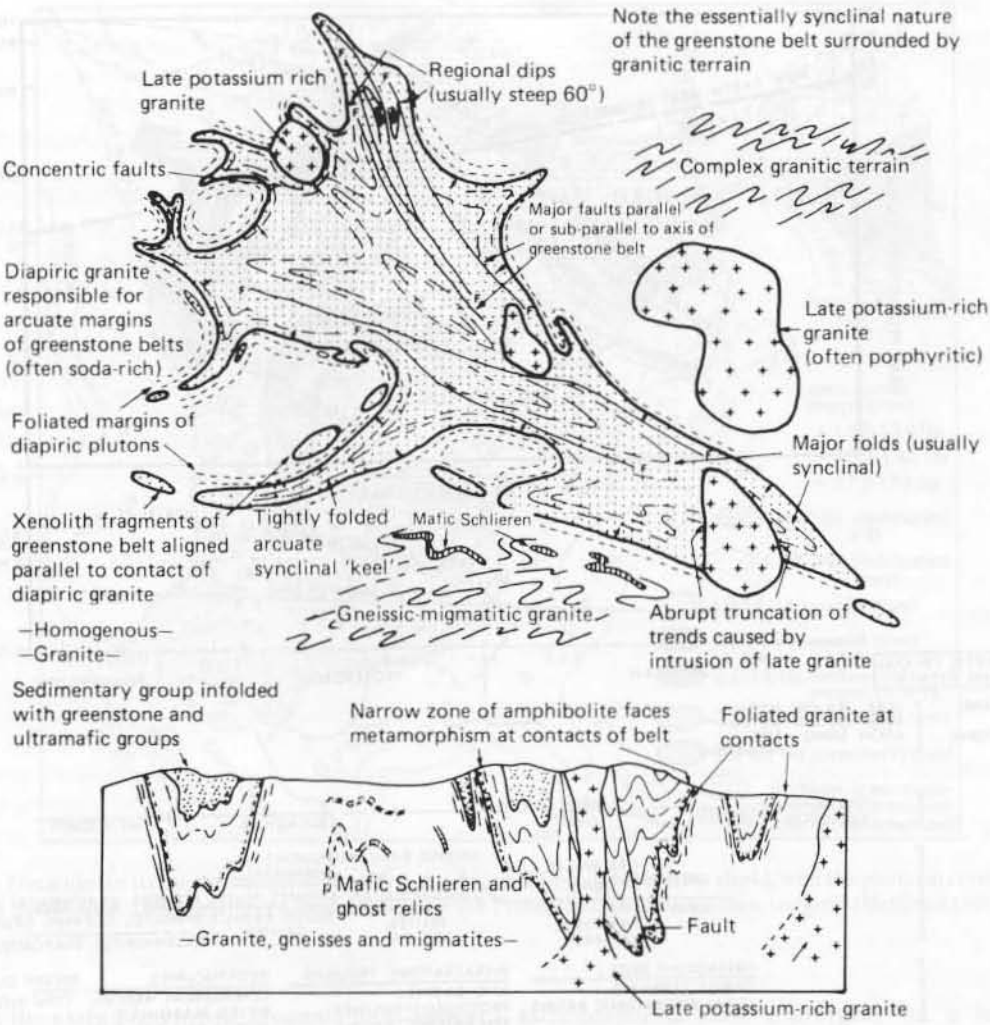


Fig. 9.94. Schematic map and section of a typical Archean greenstone–granite terrain, based on the Barberton Mountain area, Swaziland–South Africa. (Anhaeusser et al. 1969)

These and other evolutionary changes are summarized in Fig. 9.95. According to Goodwin (1981a), the microplate stage lasted from about 3.8 Ga to 2.6 Ga and resulted in the development of the characteristic shield pattern of cratonic blocks bordered by mobile belts. Heat flow was high and mantle convection fast during this early phase, leading to rapid overturn of small convection cells. Rogers (1993) stated:

> The oldest rocks in greenstone belts include mafic/ultramafic volcanic rocks and clayey sediments rich in mafic components. They did not evolve by erosion of continental crust, but they did require a mechanism of crustal downwarp to form the depositional basin and a mechanism of flank uplift to form the source terrane.

This phase was terminated by a widespread cratonization event involving massive intrusion of granitoid bodies, apparently on a worldwide scale. This event is the basis for the Archean–Proterozoic division of Precambrian time at 2.5 Ga, which has now been adopted virtually worldwide (Sect. 3.6.7).

As indicated in Fig. 9.95, cratonization was a continuous process. Rogers et al. (1984) suggested that, in individual cratons, it continued after typical shallow-water cratonic sedimentation had commenced by a process of mantle conversion and consequent crustal thickening. There was a peak in global cratonization at about 2.5 Ga (although this peak does not appear in some estimates illustrated in Fig. 9.95) and, in some areas, the development of Proterozoic-style crust and associated sediments took place during the Archean. The prime example of this is the Kaapvaal

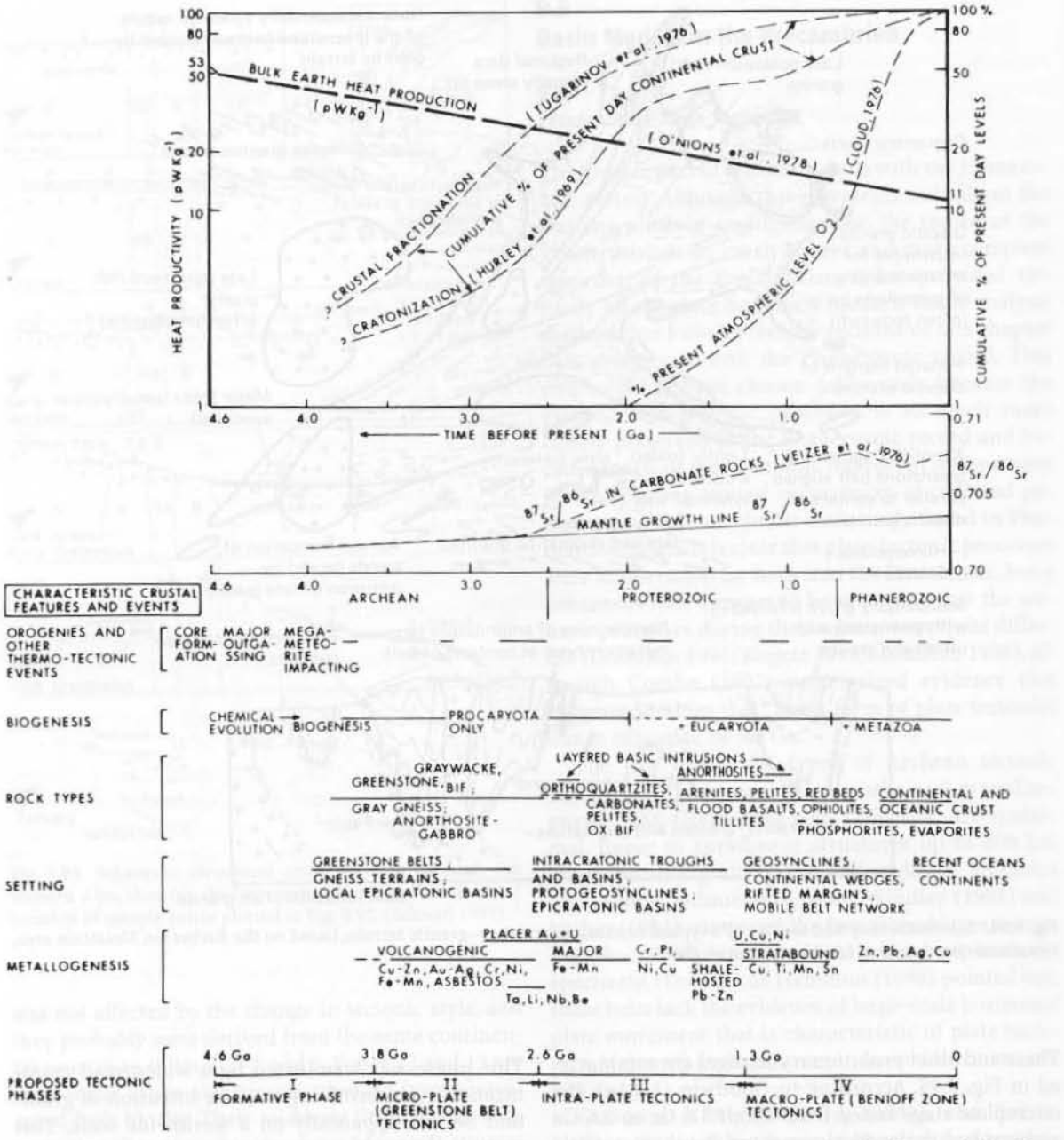


Fig. 9.95. A summary model for the evolution of the Precambrian earth. (Goodwin 1981a, reproduced with permission from the American Association for the Advancement of Science)

Craton of southern Africa. There, the Swaziland Supergroup (approximately 3.0–3.5 × 10^9 years old) contains the oldest example of continental-margin-style sedimentation. It is overlain by the Pongola Supergroup (2.9–3.1 × 10^9 years old), the oldest cratonic cover suite (Anhaeusser 1981; Tankard et al. 1982).

The shield area, termed Laurentia, that forms the core of the present North American continent, evolved by the amalgamation of a series of Archean microcontinents during the Proterozoic (Fig. 9.96):

> Laurentia owes its existence to a network of early Proterozoic orogenic belts … Many of the belts are collision zones preserving only the deformed margins of formerly independent microcontinents composed of Archean crust. Other belts contain accreted early Proterozoic island arcs and associated intraoceanic deposits. The assembly of Laurentia in the early Proterozoic may be compared with the assembly of Eurasia in the Phanerozoic (Hoffman 1988).

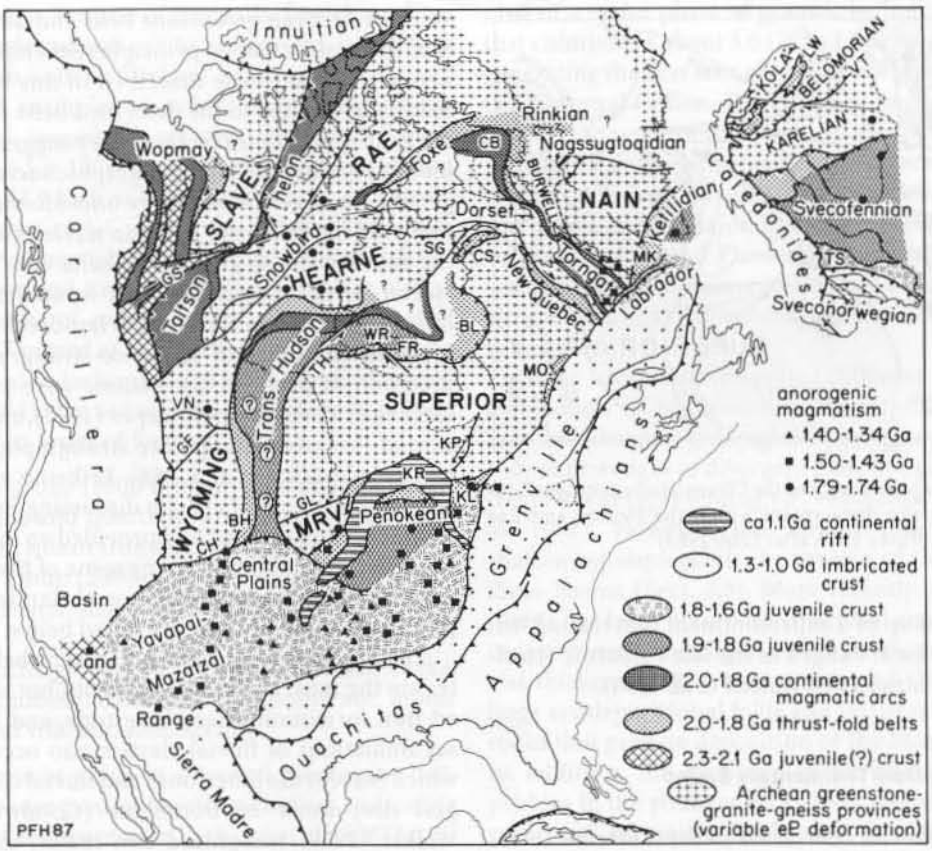


Fig. 9.96. Precambrian tectonic elements of the North American craton and the Baltic shield, with the platform cover removed. *Uppercase names* are Archean provinces, *lowercase names* are Proterozoic and Phanerozoic orogens. (Hoffman 1988)

Many of the early Proterozoic orogenic belts appear to represent collision zones between Archean provinces.

They are typically asymmetric, with one side characterized by a sedimentary prism overthrust toward the Archean foreland, and the other side characterized by a magmatic arc and by great reactivation of the Archean hinterland. The asymmetry presumably reflects the dominant polarity of subduction and consequent thermal regimes during ocean closure (Hoffman 1988).

Collision of irregular margins of the Archean microplates led to indentation and escape tectonics, with the development of large-scale strike-slip adjustments, similar to the evolution of the Himalayan orogen (Fig. 9.97; cf. Fig. 9.66). Assembly of the present core of Laurentia took place between 2.0 and 1.6 Ga, with the major Grenville province sutured on between 1.3 and 0.9 Ga. Hoffman (1989) speculated that this rapid aggregation may have been driven by asthenospheric flow converging on a large-scale mantle downwelling.

Because the mantle would be relatively cold in a downwelling region, continents there would stand low relative to sea level. The widespread deposition of platformal sediments during this interval, including the 1.7-Ga (Athabasca, Thelon, Sioux) cratonic basins, is consistent with such a hypothesis. As the interval from 1.9 to 1.7 Ga was one of intense orogenic activity and crust formation on other continents, it is conceivable that a supercontinent analogous to Pangea may have been assembled at that time. (Hoffman 1989).

As noted briefly in Section 8.3.1, the continent that had assembled by the late Proterozoic, with the addition of the Grenville orogen, began to rift and break up at about 0.8 Ga, with extensive rift and divergent-margin development occurring on all sides of Laurentia by the end of the Precambrian (Hoffman 1989, Fig. 47). This is now thought to have been part of a

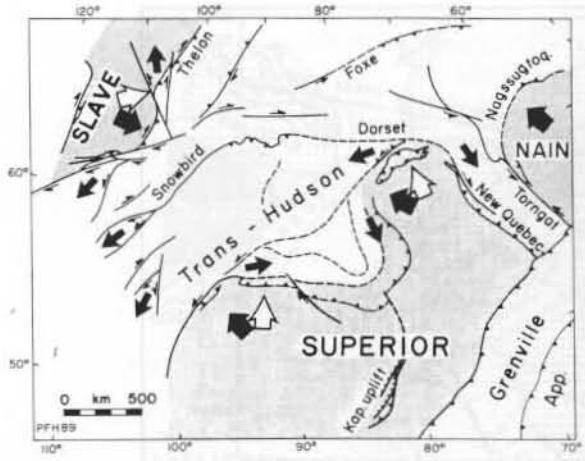


Fig. 9.97. Escape tectonics of the "Trans-Hudson orogen", involving strike-slip diplacement within the Hearne and Rae provinces. (Hoffman 1989; after Gibb 1983)

general breakup of a supercontinent that turned "inside-out" to form Pangea in the late Paleozoic (Hoffman 1991; Dalziel 1992; Dalziel et al. 1994).

9.5.2 The Evidence from Sedimentary Basins

Basin analysis can contribute to Precambrian tectonic studies in four main ways:

1. Stratigraphic and sedimentological analysis reveals the nature of depositional systems, which tells us much about crustal style, elevation, and paleoslope (Chaps. 4, 5, and 6)
2. Structural analysis of the sediments yields information on local contemporaneous tectonic control (Harding and Lowell 1979)
3. Clastic petrofacies data indicate the crustal composition of provenance areas, which can commonly be interpreted in plate-tectonic terms (Sect. 9.4)
4. Subsidence analysis provides data regarding crustal rheology and the rates of tectonic processes (Chap. 7)

Although many Precambrian sediments are severely deformed and metamorphosed, others are less so and are readily amenable to the types of sedimentological analysis discussed in this book. Walker (1988) discussed the application of basin-analysis techniques to deformed terranes, using Precambrian and Phanerozoic examples from the southwest Cordillera of the United States. Eriksson et al. (1988) discussed the application of basin-analysis techniques to Archean metamorphic terranes in southern Africa and western Australia. Eriksson et al. (1994) recommended the use of sequence-stratigraphic techniques for the study of Archean greenstone belts and demonstrated that the basin-analysis principles developed for Phanerozoic strata (those described in this book) could readily be applied to the deformed belts constituting Archean greenstone terranes. They suggested that the method of subdividing stratigraphic successions into genetic packages bounded by unconformities works well even for deformed Archean terranes and permits the subdivision of greenstone belts into "first-" and "second-order" sequences, that is, successions with interpreted durations of tens to hundreds of millions of years. Examples of sequence-stratigraphic interpretation of Archean successions were provided by Cheney et al. (1990) and Krapez (1993), and a discussion of Proterozoic sequence stratigraphy was given by Christie-Blick et al. (1988). Eriksson et al. (1998) edited a collection of articles discussing Precambrian sedimentation systems and provided an introductory overview article summarizing some of the important distinguishing features of Precambrian sedimentary processes. Some of these are noted below.

Turning first to Archean greenstone belts, turbidites are the most common sediments, but shale, banded iron formation, chert, limestone, and local, thick accumulations of fluvial clastics also occur, most of which (except the limestones and fluvial clastics) suggest deep-water environments (Goodwin 1981b). Walker (1978) recognized two major, clastic, sedimentary suites: (1) nonmarine sandstones and conglomerates and (2) resedimented conglomerates and sandstones of submarine-fan origin. Ojakangas (1985) confirmed and amplified this subdivision. He also pointed out the paucity of sediments indicating shelf environments. Most Archean sediments are associated with greenstone belts and, with the exception of the Kaapvaal Craton of South Africa (discussed below), their petrography and depositional environments suggest that stable cratons or platform areas at or near sea level had very limited distribution (Kröner 1981). Ayres (1983) showed by a petrographic analysis of lithic sandstones in a greenstone belt that they were largely pyroclastic in origin. Relatively high quartz contents do not indicate derivation from cratonic crust but from felsic volcanics. He concluded that greenstone volcanism was bimodal (basalt + rhyolite + rhyodacite), not predominantly basaltic, and that the felsic volcanics were typically removed by subaerial erosion and left a record only in the derived sediments. The bimodal nature of the volcanics indicates a thickened crust, probably developed by extensive mafic volcanism loading the crust.

Early Archean gneiss terranes were analyzed by Eriksson et al. (1988, 1994). The Limpopo Belt of southern Africa contains quartzite–marble successions that they compared to the quartz–arenite–carbonate suite of Phanerozoic cratonic basins. In the

Western Gneiss Terrane of Australia, fluvial sandstones and conglomerates can be recognized, suggesting deposition in a rift or foreland-basin setting.

Excellent examples of mildly deformed Archean supracrustal sequences are the Swaziland Supergroup ($3.0-3.5 \times 10^9$ years old) of southern Africa, the upper part of which still appears remarkably fresh, and the Gorge Creek Group of the Pilbara Block, Australia. The former has been studied sedimentologically by Lowe and Knauth (1977) and by K.A. Eriksson in a series of papers summarized in Eriksson (1982) and Tankard et al. (1982). Later data and ideas are contained in Jackson et al. (1987) and Eriksson et al. (1988, 1994). The sequence is subdivided as follows (numbered in order of formation):

3. Moodies group (3500 m): polymict conglomerate, feldspathic and quartzose arenite, siltstone, mudstone, and minor iron formation
2. Fig Tree group (2000 m): lithic arenite, tuff, agglomerate, minor shale, conglomerate, and iron formation
1. Onverwacht group (16,370 m): ultrabasic rocks, basalt, andesite, dacite, tuff, agglomerate, minor chert, and volcaniclastic arenite

The Onverwacht group is a typical greenstone belt. The arenites are predominantly turbidites, indicating subaqueous sedimentation. Jackson et al. (1987) concluded that the group was formed by shield volcanism and associated volcaniclastic sedimentation on an ocean floor far from continental crust. "The upper contact of the Onverwacht group marks the ... transition from exclusively oceanic sedimentation in a volcanic environment to sedimentation dominated by massive influx of extrabasinal terrigenous debris" (Jackson et al. 1987). The sediments of the Fig Tree and Moodies groups demonstrate an upward-coarsening character, from basinal mudstone, iron formation, and turbidites in the former to shallow-marine to nonmarine arenites in the latter. Sedimentary structures are well preserved, and traditional methods of facies analysis have been applied to these rocks (Erikkson 1982; Tankard et al. 1982). It is now known that, in the north, the three groups occur in conformable stratigraphic succession whereas, in the south, the Onverwacht and Fig Tree are involved in thrust faulting, and both are overlain unconformably by the Moodies group (Fig. 9.98). Petrographic and geochemical studies indicate an upward change from volcanic to plutonic (quartzofeldspathic) sources, indicating the uplifting and/or unroofing of a sialic plutonic or metamorphic terrane.

Anhaeusser (1981) interpreted this region in terms of a microplate "dome-and-sag" model. The vertical environmental and petrographic changes of the Fig Tree and Moodies groups are attributed to the beginning of a major phase of granitic batholith intrusion that culminated about 3.0 Ga and was responsible for generating the first area of (preserved) cratonic crust, the Kaapvaal Craton. The evidence for this model is beyond the scope of this book, except for the obvious fact that a convergent-margin plate-tectonic model seemed to be precluded by the absence of subduction complexes and by major differences between the Onverwacht group and Phanerozoic ophiolites. If each greenstone belt represents a separate backarc basin, it also implies a veritable festoon of bounding arcs, for which there is no evidence.

Other authors have applied different plate-tectonic models to the Swaziland Supergroup. The Fig Tree and Moodies groups have been compared to the sediment wedges of divergent plate margins, with the Onverwacht representing rift volcanism (Tankard et al. 1982). However, the upward-coarsening and shallowing sequence is the reverse of that found in these basins (Sect. 7.3). More recently, greater emphasis has been placed on structural analysis, which indicates major syndepositional shortening and crustal thickening. It is now recognized that there are large syndepositional folds and thrust faults in these rocks that predate deposition of the Moodies group. In addition, the presence of derived Onverwacht pebbles in the younger rocks indicates active syndepositional tectonism (Jackson et al. 1987). There is also geochemical evidence for the nature of Onverwacht volcanism and detrital sediment sources that is beyond the scope of this book. The current model for these rocks (Jackson et al. 1987; Eriksson et al. 1988) is that the Onverwacht group represents oceanic crust that was obducted onto sialic basement in a foreland setting. Crustal shortening and uplift then led to the development of the upward-coarsening Fig Tree–Moodies groups succession, which is thus comparable in a general way with the flysch–molasse succession of classic Phanerozoic remnant oceans and foreland basins.

Eriksson et al. (1994) used the sequence-stratigraphic study of Archean greenstones in the Pilbara Block of Western Australia by Krapez (1993) as the basis for a wide-ranging review of this association in Kaapvaal and Zimbabwe provinces of Africa, Pilbara Block, Australia, and the Superior Province of northern Ontario. They identified six main lithofacies associations: (1) a mafic-volcanic association containing subordinate sedimentary components, (2) a calc-alkaline volcanic association, (3) a bimodal volcanic association, (4) a quartz arenite–iron-formation association and/or a carbonate association, (5) a conglomerate–wacke association, and (6) a conglomerate–arenite association. Figure 9.99 illustrates a depositional model for one of these, the calc-alkaline volcanic association, based on studies in north-

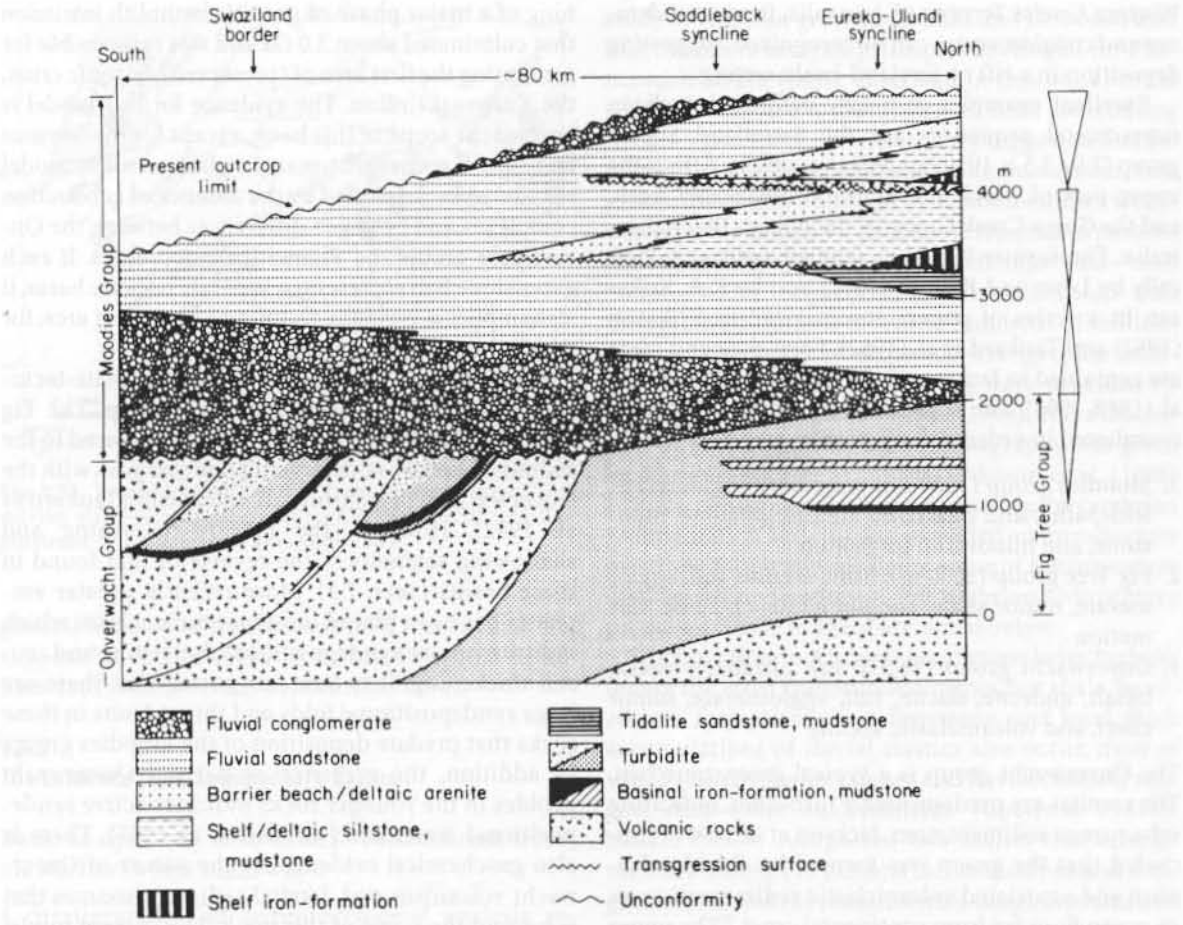


Fig. 9.98. Schematic stratigraphic reconstruction of the Swaziland Supergroup. (Jackson et al. 1987)

western Ontario. Alluvial fanglomerates pass laterally into shoreface sandstones exhibiting shallow-water sedimentary structures, and these, in turn, are transitional into turbidities and related facies deposited in a ramp setting. The basin is interpreted as a forearc flanking a felsic–volcanic arc to the north, with fault slices representing an accretionary subduction complex. However, as noted in the previous section, the strict application of plate-tectonic principles to the Archean is debatable.

What is probably the most detailed application of plate-tectonic principles to a Proterozoic orogen is the work of P. Hoffman and his coworkers on the Wopmay Orogen of northwest Canada. Their reconstruction of the stratigraphy of the eastern margin of this orogen is shown in Fig. 9.100. It consists of an extensional plate margin developed on Archean crust – the rifted western margin of the Slave Province, overlain by a foredeep succession shed from a suture zone that formed about 1.91 Ga when the Slave province collided with an arc terrane to the west. In this reconstruction, rift basins, the divergent-margin prism, and the foredeep wedge all show remarkable similarities to comparable sedimentary successions in Phanerozoic orogens.

Condie (1982a, b) explored the evidence for tectonic styles in the Proterozoic volcanic–sedimentary record. He examined 79 successions around the world and subdivided them into three assemblages, as summarized in Table 9.6. The data clearly show that Phanerozoic-style cratons were widespread by early Proterozoic time, because stratigraphically, sedimentologically, and petrographically, assemblages I and II can readily be compared with Phanerozoic cratonic and rift successions from the North American interior, Russia, Siberia, and elsewhere. A few Archean assemblages could also be fitted into the first two categories. For example, the major cratonization event of the Kaapvaal Craton about 3.0 Ga, noted previously, was followed by deposition of the Pongola Supergroup (2.9–3.1 Ga). This contains a basal volcanic unit up to 7000-m thick that Tankard et al. (1982) compared to Phanerozoic arc volcanics. It passes vertically and laterally into a clastic succession up

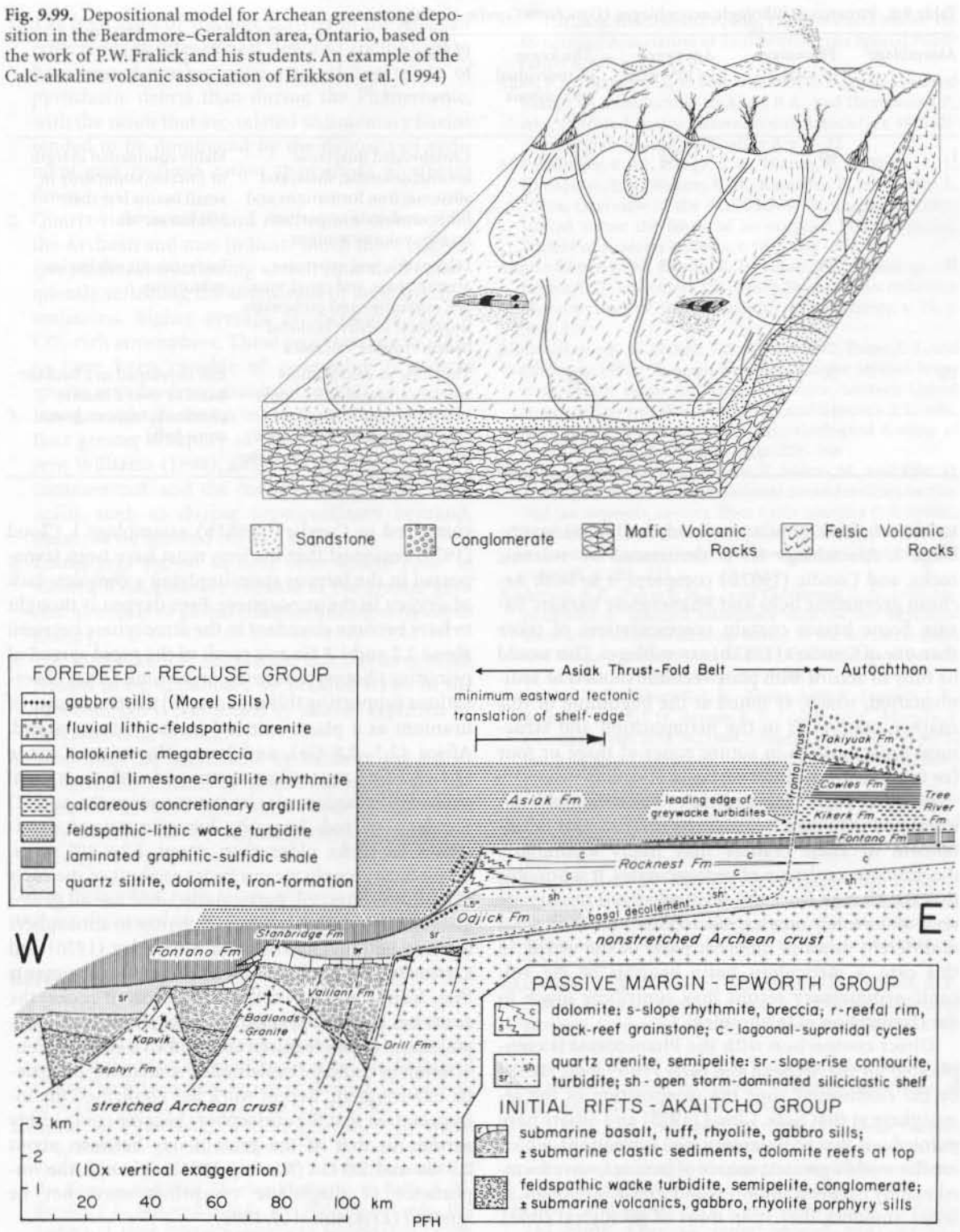


Fig. 9.99. Depositional model for Archean greenstone deposition in the Beardmore–Geraldton area, Ontario, based on the work of P.W. Fralick and his students. An example of the Calc-alkaline volcanic association of Erikkson et al. (1994)

Fig. 9.100. Stratigraphic reconstruction of the Coronation Supergroup of the Wopmay orogen, northwest Canada. The location of this orogen is shown in Fig. 9.96. (Hoffman 1989)

Table 9.6. Proterozoic lithologic assemblages. (Data from Condie 1982b)

Assemblage	Percentage of total	Age range ($\times 10^9$ years)	Thickness of individual successions (km)	Lithologies	Interpretation
I	60	1.3–2.5	>15	Crossbedded quartzose arenite, dolomite, shale, and siltstone; iron formations and lithic sandstone important locally; volcanics minor	Stable continental margin or interior, commonly in small basins less than 500 km across
II	20	1.0–2.5	1–10	Feldspathic and quartzose arenite, shale, and conglomerate; evaporites and carbonates important locally; bimodal basalt–rhyolite volcanics	Intracratonic rift basins, aulacogens
III	20	1.3–2.4	>15	Tholeiitic or calc-alkaline volcanics, pyroclastics, lithic sandstone, argillite, minor conglomerate, quartz arenite, iron formation, and carbonate	Rift developed in a backarc basin or over a mantle plume, cf. Archean greenstone belts

to 1800-m thick, similar to Condie's (1982b) assemblage I. Assemblage III is dominated by volcanic rocks, and Condie (1982b) compares it to both Archean greenstone belts and Phanerozoic backarc basins. Some basins contain representatives of more than one of Condie's (1982b) assemblages. This would be fully in accord with plate-tectonic models of sedimentation, which, as noted at the beginning of this chapter, may result in the juxtaposition and structural superposition in suture zones of three or four (or more) entirely different basin styles.

Clearly, each Precambrian basin should be examined with care, because the diachronous development of sialic cratons may imply a similarly diachronous evolution of tectonic styles. It is possible that the change from the relatively fixed microplate style to the freely moving macroplate style took place at different times in different parts of the world. In this case, a meticulous basin analysis of the volcanic–sedimentary record may contribute much to the clarification of the problem.

Direct comparison with the Phanerozoic is complicated for the Archean and early Proterozoic record by the controversy over the composition of the atmosphere at that time. Cloud (1973) and others have pointed out that sedimentary iron formations, which are the world's greatest source of iron ore, were formed mainly between about 1.8 and 2.6 Ga (although, in lesser amounts, they span most of geological time). Precambrian formations consist of iron oxides, carbonates, silicates, and sulfides, commonly interbanded with chert and other sediments, and were deposited in shallow- to deep-marine environments. The Superior type is typical of Proterozoic iron formations. It is associated with shelf sediments and can be compared to Condie's (1982b) assemblage I. Cloud (1973) reasoned that the iron must have been transported in the ferrous state, implying a complete lack of oxygen in the atmosphere. Free oxygen is thought to have become abundant in the atmosphere between about 2.2 and 1.8 Ga as a result of the rapid spread of primitive photosynthesizing organisms. Other observations supporting this idea are: (1) the abundance of uranium as a placer mineral in the Witwatersrand, Africa (2.3–2.8 Ga), and lower Huronian, Canada (2.1–2.7 Ga), successions compared with the virtual absence of uranium placers in younger rocks, and (2) the lack of red, hematite–limonite-stained sandstones in rocks older than about 2.2×10^9 years. These facts complicate our facies analysis of the early Precambrian record, particularly since not all geologists agree with the interpreted change in atmospheric composition. Dimroth and Kimberley (1976) and Clemmey and Badham (1982) presented arguments that, from the time of the earliest dated rocks, the earth has had an oxygenated atmosphere. For example, many of the features ascribed to a reducing environment could be explained by a reducing diagenetic environment. Recent work has confirmed the interpretation that the atmosphere became comparably aerobic to that of the present day between about 2.3 Ga and 2.0 Ga (Kasting 1991), although the importance of diagenetic overprints must not be ignored (Eriksson et al. 1998).

Other significant differences in sedimentary processes and environments between the Precambrian and the Phanerozoic were discussed in a special issue of *Sedimentary Geology* edited by Eriksson et al. (1998). Among the important points discussed are:

1. The high rates of volcanic activity in the Archean reflected high rates of heat flow and are thought to have caused much greater rates of generation of pyroclastic debris than during the Phanerozoic, with the result that arc-related sedimentary basins tended to be dominated by the deposits of sediment-gravity flows rather than those of stream flows.
2. Quartz-rich arenites and rudites are common in the Archean and may indicate much more aggressive chemical weathering at that time than subsequently, reflecting the abundance of toxic volcanic emissions, higher average temperatures, and a CO_2-rich atmosphere. These processes are thought to have been capable of generating first-cycle quartz arenites from granitoid rocks.
3. Low-latitude glaciation in the Proterozoic may reflect greater obliquity of the ecliptic than at present Williams (1998), although this idea is highly controversial, and the cooling effects of regional uplift, such as during supercontinent breakup, must not be ignored (Eyles 1993).
4. Eolian ergs appear to have been absent prior to about 1.8 Ga, possibly because of the greater general humidity of global climates during early earth history.
5. The long-standing suggestion that fluvial systems tended to be dominated by braided styles in the Precambrian is confirmed by studies reported by Eriksson et al. (1998).
6. Evaporites are dominated by halite in the Paleoproterozoic. Sulfates (gypsum, anhydrite) are not common in rocks older than about 1.7 Ga, probably because of low concentrations of sulfate in seawater before that time.

Other difficulties with facies analysis of the distant past are discussed in Section 4.4.4.

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Conclusions

One of the main messages of this book is the need for synthesis of many types of data, so that, in conclusion, it may be useful to re-emphasize some of the ideas I have tried to convey and perhaps clarify the structure of the book. The organization of the chapters is intended to reflect an increasing complexity of information and sophistication, corresponding to more advanced levels of basin analysis. However, it by no means reflects the way in which all basin analyses are carried out. In mature basins, the analyst may be faced, at the start, with a proliferation of old stratigraphic names. Sorting out the validity of these, their relative ages, facies interpretations, and paleogeographic evolution will consume the first round of analysis (Chaps. 3–6). Proceeding to backstripping analysis and models of crustal rheology (Chap. 7), interpretations of stratigraphic sequences and their causes (Chaps. 6 and 8), and plate-tectonic setting (Chap. 9) may come much later. For many basins, plate tectonics may seem irrelevant. For example, those working on the Alberta Oil Sands or the coal-bearing sequences of the Illinois Basin may regard plate tectonics as a very remote concern. Yet, this is not a wise approach to take. We have seen how plate-margin tectonism may be transmitted for thousands of kilometers across a cratonic interior, and we have seen how large-scale stratigraphic architecture can be controlled by eustatic sea-level changes that depend largely on spreading rates in far-away midoceanic ridges. It is a mistake to define too small an area as one's focus of interest. This can lead to serious problems, ranging from questions of stratigraphic correlation to erroneous interpretations of plate history. This point was convincingly argued by Dewey (1982) in his discussion of the evolution of the British Isles. His interpretive maps of this small area extend from Mexico to Russia.

Marine geologists and those involved with frontier petroleum exploration, particularly in offshore regions, tend to take the large-scale approach to basin analysis. They start with regional geophysical data: magnetic anomaly maps for seafloor-spreading history, regional crustal sections based on gravity and seismic data, and detailed seismic–stratigraphic interpretations (Chap. 5). These data provide a plate-tectonic framework and a broad structural–stratigraphic interpretation on which to base a drilling program. The next phase involves analysis of seismic sequences (Chap. 6) and subsidence history (Chap. 7). Interpretations of depositional systems and applications of facies-model concepts (Chap. 4) come much later, and only at advanced stages of exploration is there either the time to spend on or the data to justify the luxuries of refined lithostratigraphic nomenclature and formal biostratigraphic zonation (Chap. 3). For these types of basins, the chapters in this book would be used more or less in the reverse order. In much frontier exploration, the formalities of stratigraphy (Chap. 3) are dealt with in a rather relaxed way, as new material is constantly being added to the database. Informal stratigraphic names and biostratigraphic subdivisions provide adequate working tools until a late stage of basin exploration.

We can see, therefore, that the organization of a basin analysis must reflect the concerns of the analyst and the nature of the problem at hand. In many cases, different specialists will be working on various aspects of the work at the same time, as an industry project team, a government or academic group (e.g., the teams assembled for the Ocean-Drilling-Program legs, and the LITHOPROBE projects in Canada), or as separate individuals who may or may not communicate with each other.

One of the initial plans of this book was to include a series of complete, worked examples to illustrate the evolution of a basin synthesis. This idea was quickly abandoned, as it became obvious that any such series could only cover a few of the enormous variety of combinations of plate settings, stratigraphic architectures, and depositional systems. To deal with this complexity, it is necessary to touch on a few important details about many basins around the world, thus subjecting the reader to a surfeit of geography. Readers can amplify this treatment by studying the examples in Part 2 of the book. Details of stratigraphic correlation, facies analysis, etc., are dealt with very briefly here (for example, arguments about correlation and locations of data points are omitted), but enough information is provided that the reader should be able to see between the lines, and ample references are provided.

Basin-analysis methods have been in a stage of rapid development for at least the last 40 years and continue to evolve at a furious pace. The six major developments outlined in Chapter 1 are by no means complete; in fact, refinements are appearing so frequently that it was difficult to decide when to stop writing this book. Part 2 of this book, in particular, has required almost complete rewriting twice in the 15 years since it was written for the first edition.

The following are some of the major areas in which developments seem to me to have been the most rapid and in which continuing change seems likely:

1. Facies analysis and depositional systems of marine clastic environments, including shelf, slope, and rise areas (Sects. 4.6.7, 4.6.9, 6.5.2, and 6.5.3) are the most inaccessible for sedimentological study, and much remains to be learned about them. Shelf dynamics, bedform development, and submarine-fan evolution are undergoing intense study, both in modern environments and in appropriate ancient sequences. The increased use of deep-towed geophysical instruments, particularly high-resolution seismic and side-scan sonar, are having a profound effect on our knowledge of these environments.
2. An increased focus on the three-dimensional (3-D) complexity, or architecture, of complex clastic bodies (Sects. 4.3.4 and 4.5.9) is of particular interest in the area of enhanced petroleum recovery, and much industry effort is being expended on research in the area.
3. Tremendous strides have been made in the development of computer applications in basin analysis. Stratigraphic and seismic databases are becoming more reliable and more accessible, while the development of the workstation concept, with the use of color graphics and interactive software, has made mapping, modeling, and idea testing a highly sophisticated art (Sect. 8.6). A particularly interesting area to watch will be the development of sedimentological applications of 3-D seismic exploration.
4. Sequence stratigraphy. Sedimentary models for stratigraphic cycles are an area of considerable current interest (Chap. 6). One of the more fruitful outcomes of the seismic-stratigraphy revolution has been a renewed focus on the details of allogenic processes, including tectonics and sea-level change (Chap. 8). Applications of Milankovitch models of astronomical forcing are becoming increasingly convincing, especially in the area of short-term glacioeustasy (Sect. 8.2.4.1).
5. The initial enthusiasm for the Vail curves of sea-level change derived from seismic stratigraphy has been tempered with a more cautious, more sophisticated approach to the multiple causes of relative sea-level change, and a healthy scepticism is now apparent in the correlation of local curves to any assumed global standard (Sect. 8.5).
6. Quantitative physical models of basin formation and subsidence made enormous strides during the 1980s (Chap. 7), to such an extent that backstripping analysis, the plotting of subsidence curves, and their interpretation in terms of rheological models have become routine in many oil companies. There is, however, as with any new technique, a danger of overenthusiasm, and it seems likely that the more we learn, the more we will realize that the simple suite of models now available, despite the considerable computational complexity involved in their use, must be employed with great caution. For example, models are demonstrating that the response of the crust to stress may be highly variable, depending on local crustal history and the resulting heterogeneity of the crust. Future research will probably focus on the modeling of ever shorter and more precisely constrained events within basins, such as the link between individual tectonic episodes (thrust events) in foreland basins and the resulting subsidence pulses, as expressed in the stratigraphic record (Sect. 8.6).
7. Plate-tectonic interpretations (Chap. 9). Geophysical models of crustal behavior are rapidly improving in their ability to predict geological events (Chap. 7). At the same time, stratigraphic–structural basin models are providing increasingly sophisticated templates for the interpretation of ancient basins. New plate models for most areas of the globe will, therefore, continue to appear as incorporation of these new ideas becomes necessary. We have seen an evolution away from the early popularity of rift-and-drift models, such as divergent plate margins and failed rifts, to more sophisticated interpretations involving oblique slip and to reconstructions involving a wealth of terranes. Foreland basins of all types have received considerable attention and are perhaps currently "in vogue". At the same time, plate-tectonic interpretations of the Precambrian record are being carried out with increasing confidence.

As Dewey (1982) pointed out, the development of plate tectonics has, in many respects, made our work more difficult. We can no longer build *ad hoc* geosynclinal models but are severely constrained by a complex theoretical framework. It may never be possible to unravel completely the kinematics of many areas of repeated suturing and rifting or microplate accretion, where the spreading histories of large and small oceans have been destroyed by subduction. Even where these can be deduced, perhaps from motions of adjacent larger plates, rational interpretation may be

severely hampered by changes in plate kinematics, with resultant overprint of different styles and magnitudes of deformation at different orientations. Each phase of deformation develops its own crustal anisotropy, which may be reactivated later, with complex results, under a different plate regime. Good examples are the interaction of Caledonide and Hercynide trends in Britain, and the Hercynide–Alpine overprint through the Mediterranean area. Even the structurally and stratigraphically simple Alberta Basin shows the effects of the rejuvenation of Precambrian tectonic elements periodically throughout the Phanerozoic. Resolving the historical geology of areas such as these will continue to provide the greatest challenges for basin analysts in the years to come.

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Principles of Sedimentary Basin Analysis

This book incorporates developments in chronostratigraphy, sequence stratigraphy and sedimentation and tectonics that have taken place during the last decade. In particular, the wealth of case studies that has appeared since the late 1980s, focusing on sequence architecture and the relationships between sedimentation and tectonics, have required a virtually complete rewriting of Chapters 6 to 9. The global-eustasy model that was popular as a mechanism for sequence generation at the time of writing of the earlier editions of this book has been downplayed in this edition in the face of abundant new evidence pointing to the importance of tectonism as a process that may generate sequences over a wide range of temporal and physical scales. Amongst other additions, this edition contains new sections describing such topics as basin inversion and basement control of sedimentary basin development, and a range of new case studies of the plate tectonics of sedimentary basins.

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