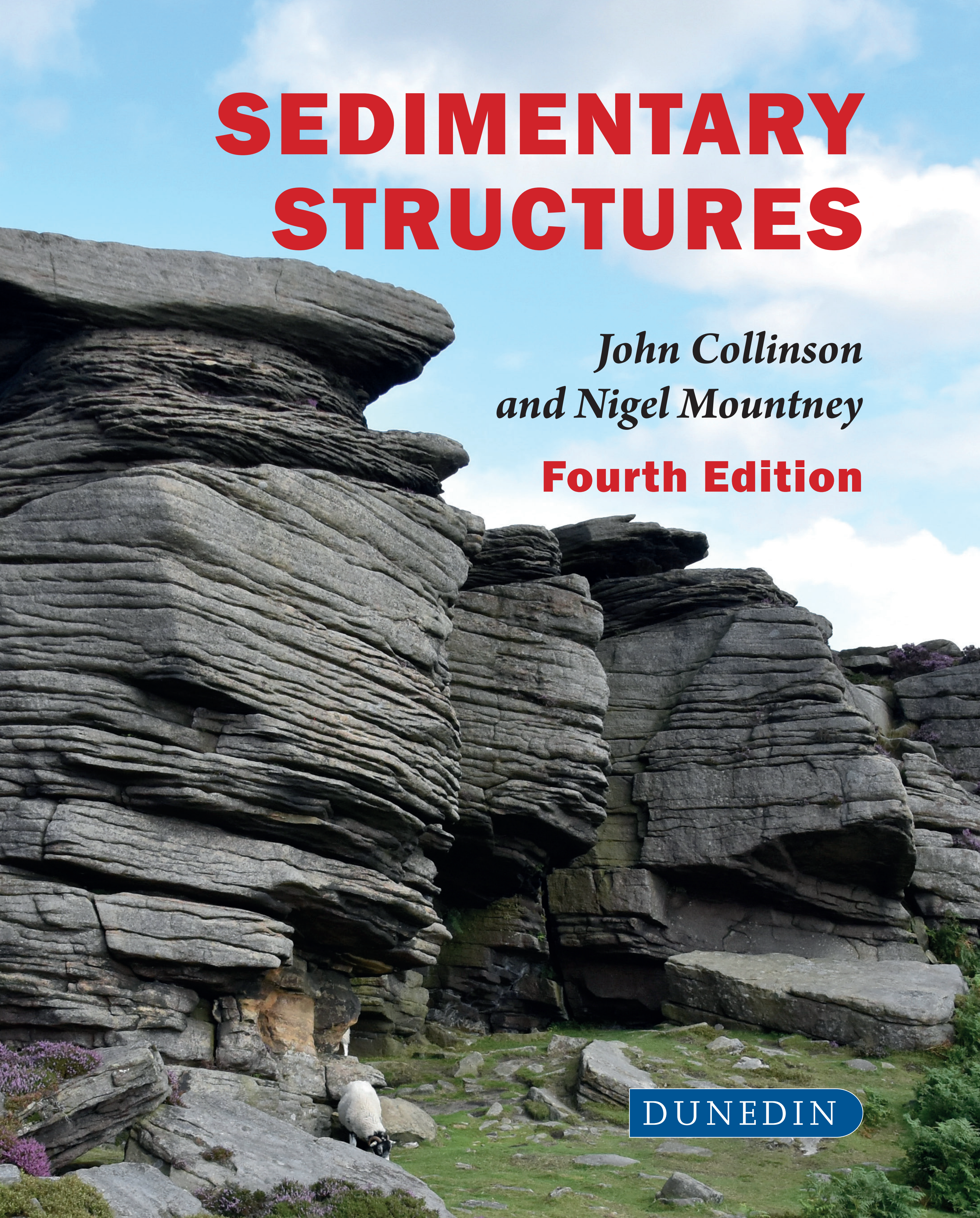
SEDIMENTARY STRUCTURES

John Collinson
and Nigel Mountney

Fourth Edition



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Fourth Edition

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Dedication: David B. Thompson

This new edition of *Sedimentary Structures* is dedicated to the memory of our late friend and colleague, David Thompson, who was a co-author of earlier editions of the book. David passed away in 2013, several years after the publication of the third edition of *Sedimentary Structures*, to which he made significant contributions, despite encroaching illness. David's exceptional experience of teaching geology, both as a school teacher and as a lecturer in Geological Education, led him to recognise the need for a textbook in basic sedimentary process and products that espoused a very direct, hands-on, practical approach. His pedagogic principles very much set the tone of the first two editions and continue to underpin this present edition.

David was an internationally acclaimed educationalist, widely recognised for his major contributions to the teaching of geology in schools, mainly carried out as a lecturer at Keele University, UK, but based on extensive experience of teaching at a secondary school in Manchester. At Keele, David was responsible for geological education in the Department of Education where many future teachers of geology benefitted from his enthusiasm, kindness and rigour. He was instrumental in the establishment of the Association of Teachers of Geology, latterly known as the Earth Science Teachers' Association (ESTA), and he remained an active officer of that body for many years. In addition to his educational responsibilities, David was a highly regarded sedimentologist in his own right having completed his MSc at the University of Manchester on the Triassic sedimentary successions of the Cheshire Basin, as a part-time supplement to his school teaching responsibilities. His interest in and enthusiasm for the Triassic sediments continued throughout his life and he continued to make original contributions into his retirement, publishing several notable papers in a variety of leading academic journals. He was a great support to many research students working in the area and an enthusiastic supporter of adult education and of amateur groups, especially the North Staffordshire Geological Association. He also developed strong interests in the industrial history of the Triassic sediments of Cheshire and Shropshire, especially the copper mineralisation and mining, and became a pioneering authority on the local history of several villages in North Staffordshire, Shropshire and Cheshire.

David was a supportive and thoughtful friend and colleague to both of us and we hope that this new edition is a fitting tribute to his memory.

Contents

Preface to the fourth edition

This Fourth Edition of *Sedimentary Structures* marks four important developments. First, this is the first edition to be published by Dunedin Academic Press (DAP), which took on the list of Terra Publishing following the untimely death of Roger Jones. We thank Anthony Kinahan of DAP for his patience and friendly encouragement throughout the preparation. Second, it is the first edition that has been prepared without any direct input from David Thompson who passed away in 2013 and to whom this edition is dedicated; much of David's philosophy and input remains. Third, this edition is the first to incorporate colour in both drawn and photographic figures. This has necessitated comprehensive re-drafting and we have also taken the opportunity to significantly expand the photographic content. Fourth, this edition makes *Sedimentary Structures* available as an ebook for the first time.

The book remains primarily a text that provides a basis for understanding the morphologies and processes of formation of sedimentary structures, both common and more obscure. Examples are taken both from present-day depositional settings and from ancient successions. The book is intended primarily for earth scientists, but also has a role for non-specialists from other subject areas where an understanding of the origins and forms of structures in sediments and sedimentary rocks would be valuable. It is hoped that the book will provide a sound basis for more advanced studies of sedimentary processes and for understanding ancient sedimentary environments through facies analysis.

Inevitably, any serious discussion of sedimentary processes requires some appreciation of basic physics and chemistry and, in the case of trace fossils, of basic biology. Whilst we have tried to minimise the use of equations and have included only those that we consider essential for description and understanding of key processes, we feel that even the most equation-shy reader will benefit from a careful working through of the modest number of important equations that we do present.

Compared with the Third Edition, this edition has undergone modest rewriting. There have been significant additions to reflect advances in our understanding of mass flows and of sediment instability. We have also reorganised Chapter 6, so that the sections dealing with aeolian processes are better integrated. The final chapter has been expanded to take the reader a little further along the road towards facies analysis and environmental interpretation. As such, this chapter serves as a primer to several other books that develop these concepts more fully.

Throughout the book we suggest simple experiments that may help to reinforce understanding of some of the processes discussed. We hope that these are mere starting points for further experimental studies in both field and laboratory. There is huge scope for imaginative and instructive interaction with on-going processes in several present-day sedimentary settings. Although the book is not explicitly intended as a field manual, it could be used in that way. However, we consider recognition and naming of particular structures as only a first step in understanding the physical, chemical and biological processes active in their development, and hope that this book will enable readers to carry forward that type of interpretation through their own bespoke inquiry. The reference lists and the bibliography have undergone revision to incorporate more recent literature, but we continue to avoid referring to websites as some are ephemeral and some, but by no means all, are of dubious accuracy.

As in earlier editions, where technical terms are first introduced and explained in the text, they are shown in bold type. These terms are also referenced in the Index and thus the need for a separate Glossary is avoided.

We hope that this new edition, with its colour presentation, provides a significant improvement on earlier editions. Several friends and colleagues have helped with advice during the preparation: Brian Rosen advised on reefs; Stephen Lokier provided guidance on numerous structures of chemical and biochemical origin; Jeff Peakall provided discussion of flow processes and controls on sedimentation; John Pollard's extensive work on the Third edition remains largely intact and much valued. We would also like to thank all those colleagues who have provided photographs (fully acknowledged elsewhere), and all authors and publishers who have allowed us to use illustrations from their publications.

John Collinson
Nigel Mountney
January 2019

Acknowledgements

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3.2, 3.5, 3.6: J. R. L. Allen (1968) 3.7: J. R. L. Allen (1982: Elsevier) 3.15: G. V. Middleton (1976: University of Chicago Press) 3.16: Å. Sundborg (1956) and P. Ljunggren & Å. Sundborg (1968), both from the Swedish Society for Anthropology and Geography 3.17: G. V. Middleton & J. B. Southard (1984: Society for Sedimentary Geology) 3.19: B. Kneller & M. J. Branney (1995: Blackwell Publishing) 3.20: G. V. Middleton & M. A. Hampton (1976: John Wiley) 3.21: W. Nemec & R. J. Steel (1984: Canadian Society of Petroleum Geologists) 3.22: T. Mulder & J. Alexander (2001: Blackwell Publishing) 3.24, 3.25: Y. K. Sohn (1997: Society for Sedimentary Geology) 3.26: J. E. Simpson (1987: John Wiley).

4.2: F. Ricci-Lucchi (1970: Zanichelli) 4.4: S. Sengupta (1966: Society for Sedimentary Geology) 4.5: J. R. L. Allen (1982: Elsevier) 4.7, 4.8, 4.9: J. R. L. Allen (1971: Elsevier) 4.10B: Gilbert Kelling 4.11: J. R. L. Allen (1971: Elsevier) 4.13B, 4.15: Gilbert Kelling 4.16: G. Y. Craig & E. K. Walton (1962: Geological Society of London) 4.21A: Gilbert Kelling 4.24: J. R. L. Allen (1964: Blackwell Publishing) 4.26: H. H. Roberts et al. (1976: Offshore Technical Conference, Houston, Texas) 4.28: D. B. Loope (1985: Geological Society of America).

6.1, 6.9: J. R. L. Allen (1968) 6.12: J. F. M. de Raaf et al. (1977: Blackwell Publishing) 6.13: J. R. Boersma (1970) 6.19, 6.20: H-E. Reineck & I. B. Singh (1973, 1980: Springer) 6.22: J. C. Harms et al. (1975: Society for Sedimentary Geology) 6.23, 6.24: J. R. L. Allen (1968, 1970: North Holland; Allen & Unwin) 6.25: D. L. Inman & A. J. Bowen (1963: American Society of Civil Engineers) 6.26: A. Kaneko (1980: Research Institute of Applied Mechanics, Kyushu University) 6.27: H. E. Clifton et al. (1971: Society for Sedimentary Geology) 6.29: J. R. L. Allen (1968) 6.35B & C: Gilbert Kelling 6.41: D. M. Rubin & R. E. Hunter (1983: Elsevier) and P. G. DeCelles et al. (1983: Society for Sedimentary Geology) 6.42B, 6.46B: Gilbert Kelling 6.48: A. V. Jopling (1965: Society for Sedimentary Geology) 6.49, 6.50, 6.51: J. D. Collinson (1970: Swedish Society for Anthropology and Geography) 6.52: J. R. L. Allen (1980: Elsevier) and M. E. Tucker (2001: Blackwell Publishing) 6.54: R. W. Dalrymple et al. (1991: Canadian Society of Petroleum Geologists) 6.56: E. S. Pretious & T. Blench (1951: National Research Council of Canada) 6.57: S. F. Leclair (2002: Blackwell Publishing) 6.59: P. J. McCabe (1977: Blackwell Publishing) 6.63: N. Yan et al. (2017: Elsevier) 6.65: Allen (1985b) 6.67: J. F. Kennedy (1961: W. M. Keck Laboratory of Hydraulics and Water Resources, California Institute of Technology) 6.68C, 6.71: Gilbert Kelling 6.73: J. R. L. Allen (1991: Society for Sedimentary Geology) and D. A. V. Stow et al. (1996: Blackwell Publishing) 6.74: Sarah Southern and P. Haughton et al. (2009 Wiley-Blackwell) 6.75 Cartigny et al. (2011: Wiley) and Covault et al. (2014: Wiley) 6.76: I. G. Wilson (1972: Blackwell Publishing) 6.81: R. E. Hunter (1977: Blackwell Publishing) 6.84E: Roy Fitzsimmons 6.85: R. E. Hunter et al. (1983: Geological Society of America) 6.88: S. G. Fryberger (1979: USGS) 6.89: R. J. Wasson & R. Hyde (1983: Nature) 6.92, 6.93, 6.94: E. D. McKee (1966: Blackwell Publishing) and E. D. McKee (1979: USGS) 6.95: R. A. Bagnold (1941: Dover) and E. D. McKee & G. C. Tibbitts (1964: Society for Sedimentary Geology) 6.96: M. E. Brookfield (1977: Blackwell Publishing) 6.97: R. E. Hunter (1977: Blackwell Publishing) and G. Kocurek & R. H. Dott (1981: Society for Sedimentary Geology) 6.98: R. E. Hunter (1977: Blackwell Publishing) 6.100: J. A. Howell (1992) 6.101: M. Crabaugh & G. Kocurek (1993: Geological Society of London).

7.2: D. P. Piper & P. J. Rogers (1980: British Geological Survey, NERC) 7.5: M. E. Tucker (1991: Blackwell Publishing) 7.7, 7.8: R.G. Walker in J. C. Harms et al. (1975: Society for Sedimentary Geology) 7.11: W. Nemec & R. J. Steel (1984: Canadian Society of Petroleum Geologists) 7.13: I. C. Davies & R. G. Walker (1974: Society for Sedimentary Geology) and R.G. Walker in J. C. Harms et al. (1975: Society for Sedimentary Geology) 7.14: E. Derbyshire et al. (1979: Butterworth) 7.16: W. Nemec & R. J. Steel (1984: Canadian Society of

Petroleum Geologists) 7.17: E. Derbyshire et al. (1979: Butterworth) 7.23: N. D. Smith (1974: University of Chicago Press) and F. J. Hein & R. G. Walker (1977: Canadian Society of Petroleum Geologists) 7.24: Y. K. Sohn (1997: Society for Sedimentary Geology) 7.25: W. Nemec & R. J. Steel (1984: Canadian Society of Petroleum Geologists) 7.26, 7.27: Y. K. Sohn et al. (1997: Society for Sedimentary Geology) 7.30: D. Carlisle (1963: University of Chicago Press), J. Lajoie (1979: Geological Association of Canada), R. S. J. Sparks et al. (1973: Geological Society of America), and R. S. Fiske & T. Matsuda (1964: American Journal of Science) 7.31: R.G. Walker in J. C. Harms et al. (1975: Society for Sedimentary Geology).

8.1: Stephen Lokier 8.2: B. C. Schreiber (1986: Blackwell Publishing) 8.7: Stephen Lokier 8.12: specimen provided courtesy of Jonathan Carrivick 8.13: specimen provided courtesy of Emily McMillan 8.15: NASA Earth Observatory 8.16: Brian Rosen 8.18: N. P. James (1979: Geological Association of Canada) 8.19: R. J. Dunham (1962: American Association of Petroleum Geologists) and M. E. Tucker (2001: Blackwell Publishing) 8.20: Stephen Lokier 8.21: E. B. Wolfenden (1958: Geological Association of America) and F. M. Broadhurst & I. M. Simpson (1967: Cambridge University Press) 8.23: Stephen Lokier 8.24: W. V. Preiss (1976: Elsevier).

9.1, 9.2: G. Owen (1987: Geological Society of London) 9.4B: Gilbert Kelling 9.5: H. C. Sorby (1908: Geological Society of London) 9.7: J. R. L. Allen (1982) 9.10: F. A. Audemard & F. de Santis (1991: Bulletin of the International Association of Engineering Geologists) 9.11: Martin Bochud 9.14: J. R. L. Allen (1982) 9.15: P. W. G. Tanner (1998: Blackwell Publishing) 9.21: R. Gruhn & A. L. Bryan (1969: Arctic Institute of North America) 9.23: D. R. Lowe (1975: Blackwell Publishing) 9.25: E. D. McKee (1979: USGS) 9.27B: Gilbert Kelling 9.28: J. D. Collinson in A. Maltman (1994: Chapman & Hall) 9.29: J. R. L. Allen (1982) 9.33: R. J. Cheel & B. R. Rust (1986: Elsevier) 9.34C: Gilbert Kelling 9.38: M. B. Edwards (1976: American Association of Petroleum Geologists) 9.40: P. Ringrose (1988: Blackwell Publishing) 9.44: W. C. Krumbein & R. M. Garrels (1952: University of Chicago Press) 9.45G: Brian Rosen 9.48: P. A. Allen (1997: Blackwell Publishing) 9.55: R. W. Frey et al. (1978: Elsevier) and B. G. Anderson & M. L. Droser (1998: Blackwell Publishing) 9.56: R. W. Frey (1975: Springer) 9.59P: Gilbert Kelling 9.60: R. G. Bromley (1996: Chapman & Hall) 9.61: A. Seilacher (1964: Blackwell Publishing) and R. W. Frey (1975: Springer) 9.63: R.G. Bromley in R. W. Frey (1975: Springer) 9.64: J.D. Howard in P. B. Basan (1978: Society for Sedimentary Geology) 9.65: R. G. Bromley (1996: Chapman& Hall) and R. Goldring (1999: Pearson Education) 9.66: R. W. Frey & S. G. Pemberton (1985: Bulletin of Canadian Petroleum Geologists) and S. G. Pemberton (1992: Society for Sedimentary Geology) 9.67: S. G. Pemberton (1992: Society for Sedimentary Geology) 9.70: R. W. Frey (1975: Springer) and P. B. Basan (1978: Society for Sedimentary Geology) 9.71: R. G. Bromley (1996: Chapman & Hall) 9.72: A. Wetzel (1984: Geological Society of London) 9.73: R. G. Bromley (1996: Chapman & Hall) 9.74: A. M. Taylor & R. Goldring (1993: Geological Society of London) and R. Goldring (1999: Pearson Education) 9.75: M. L. Droser & D. J. Bottjer (1986: Society for Sedimentary Geology) 9.76: R. W. Frey (1975: Springer) and J.D. Howard in P. B. Basan (1978: Society for Sedimentary Geology) 9.77: A. Seilacher (1967: Elsevier), R. W. Frey (1975: Springer), Chamberlian in P. B. Basan (1978: Society for Sedimentary Geology), S. G. Pemberton (1992: Society for Sedimentary Geology) and S. T. Hasiotis (2002: Society for Sedimentary Geology) 9.80: T.P. Crimes in R. W. Frey (1975: Springer).

10.2: S. A. Greer (1975: Senckenbergiana Maritima), J. M. Coleman & L. D. Wright (1975: Houston Geological Society) and F. Surlyk (1978: Geological Survey of Denmark and Greenland) 10.3: N. P. Mountney & D. B. Thompson (2002: Blackwell Publishing) 10.5A: Gilbert Kelling 10.7C: B. R. Pratt et al. (1992: Geological Association of Canada) 10.8: NASA Earth Observatory 10.9: S. Ghazi and N. P. Mountney (2009: Elsevier) 10.11, 10.12, 10.13: NASA Earth Observatory.

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CHAPTER 1

Introduction to the study of sedimentary structures

The study of sedimentary rocks has come a long way in the last 200 years. In the nineteenth century, sedimentary rocks were mainly regarded as the matrix in which fossils occurred and their study, as far as it went, was mainly tied up with the understanding of stratigraphy. Sedimentary rocks had clearly been deposited through time in various ways but little attention was paid to asking exactly how. There was a general appreciation of the idea that ancient processes and conditions of deposition were probably similar to those prevailing at the present day (actualism and uniformitarianism) but detailed study, with a few notable exceptions, concentrated on description of the rocks as materials, rather than as products of dynamic processes and environments. This attitude prevailed until the middle of the twentieth century though pioneering studies had, by then, used sedimentary structures as indicators of top and bottom (way-up) in deformed successions and as a means of deducing palaeocurrent directions.

The second half of the twentieth century saw the development of the distinct discipline of sedimentology. This sought to explain sedimentary rocks in considerable detail in terms of the processes of sediment transport and deposition, the environments in which the rocks were laid down, and the processes that had influenced post-depositional changes during burial. These developments, initially driven in part by the needs of the hydrocarbon industry in the exploration for oil and gas reserves, led to a much more detailed knowledge of the physical, chemical and biological processes of generation, transport and deposition of sedimentary materials. It also led to a greater understanding of the environments in which sediments were laid down and the development of models (facies models) for the characterization and prediction of the organization of sedimentary successions produced in different settings. At the same time, the effects of animal and plant life in modifying sediments and the role of chemical reactions between the sedimentary particles and their surrounding pore waters were all studied in great detail. Since the 1980s there has been an important integration of sedimentology and stratigraphy in the discipline of sequence stratigraphy. This seeks to explain sedimentary successions in terms of larger-scale controls, developing around the ideas of relative sea level and accommodation space. The emphasis that such an approach places on the identification of "key surfaces" of transgression (landward retreat of a shoreline) and/or deepening, or on regression (seaward outbuilding of a shoreline) and erosive incision, and on the vertical stacking patterns of sediments did, during the 1980s and 1990s, draw some attention away from the sediments themselves. This trend has now largely re-balanced and we live in a time when a full integration of sedimentological and stratigraphic skills can yield great insights into the history of sedimentary successions at all scales, from the basin fill to the pore space.

In this context, sedimentary structures have a key role to play in the interpretation of sedimentary processes, which, in turn, provides a starting point for the interpretation of depositional environments and palaeogeographies. In this book we consider the fundamental importance of sedimentary structures to virtually all interpretations of sedimentary rocks. Many sedimentary structures are fascinating and commonly beautiful features in their own right. Their study brings together diverse aspects of physics, chemistry and biology, often in unexpected and unique ways, and it demands a stimulating combination of observation, imagination, intuition, deduction and scientific understanding.

1.1 The nature of this book

To give you an idea of what this book is about, see to what extent you can describe and interpret, using your present experience, the series of geological structures and relationships shown in Figure 1.1. You might also think of what significance such structures could have for geologists exploring for and exploiting economic resources. Whatever experience you bring to bear on this exercise, it is likely that you will have followed many of the steps that an experienced sedimentologist would have taken in tackling the same problem. It is hoped that your ability to apply a more complete and detailed analysis will develop from reading this book.

But first, just what approaches might you have made in tackling Figure 1.1?

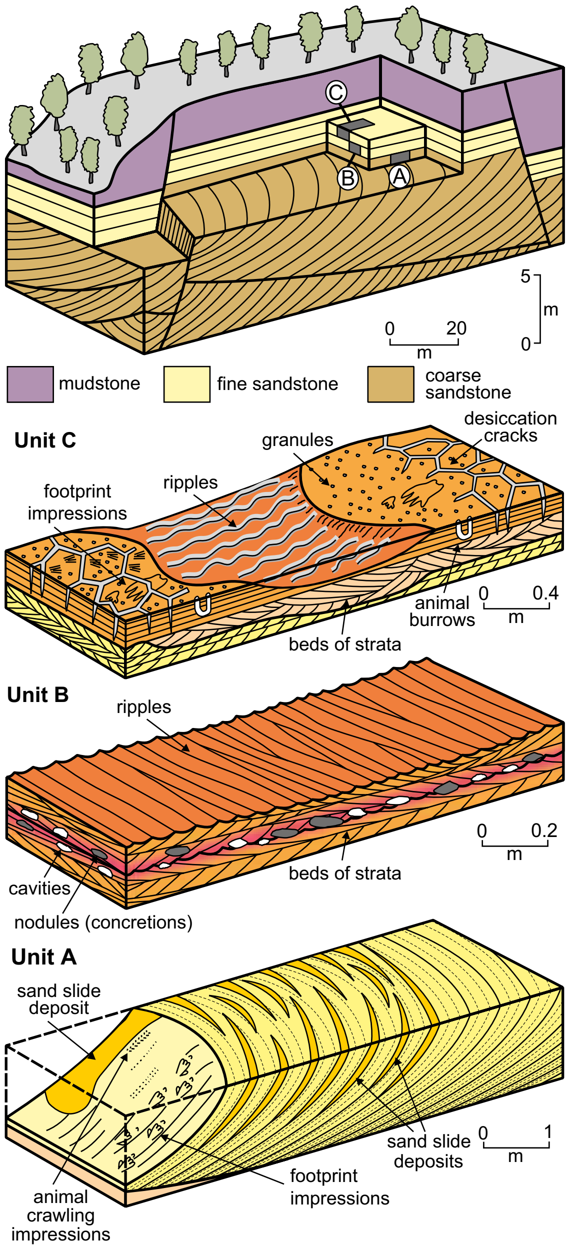


Figure 1.1 Sedimentary structures exposed in three blocks representative of units A, B and C in a hypothetical quarry. Note the scales of the blocks and their orientation within the quarry.

- You will have recognized and described several features on the basis of your everyday experience. This provides you with a valuable information base but it is clearly inadequate, on its own, to enable you to complete the task.
- You will have observed, compared, classified, inferred and possibly predicted certain features and relationships. This book should enable you to refine and enlarge this range of descriptive and interpretational skills and techniques.
- You may have tried to explain some of these features based on your understanding of physical, chemical and biological processes that you see operating today. In doing so, you will have applied a set of current beliefs about nature that suggest that it is orderly and uniform, in other words, applied the idea that the present is the key to explaining the past. This doctrine of **uniformitarianism** was promoted by Charles Lyell in the early 19th century and encapsulated the idea that uniformity in the laws of nature allowed present-day geological processes to be applied to the interpretation of ancient rocks through careful observation and extrapolation.
- You might ask yourself whether you first took in a great deal of information at a glance, produced one or more speculative explanations or hypotheses, and then tested these initial ideas by further, more critical examination of the evidence, or whether you first described each bit of the jigsaw and then came to a general idea of its meaning. In either case, working deductively (proving certain ideas false on the basis of critical evidence) or inductively (going from the particular to the general), you were applying fundamental processes and methods of scientific enquiry.
- You may have attempted to sort a great many features into time-space relationships: a process of historical ordering of events at a particular place, a technique which is at the heart of the geological sciences and which helps to distinguish them from the other sciences.

1.2 The wider geological context of sedimentary structures

Sedimentology is the study of the nature and origin of both present-day and ancient sedimentary deposits. It includes sedimentary petrography (the description of sediment composition, and sediment fabric and grain texture). Sedimentology is closely related to stratigraphy, particularly to sequence stratigraphy, which seeks to assign sedimentary deposits to genetically related units on a variety of scales, and to explain these units in terms of variations in the amount and rate of sediment supply and variations in the rate of change in the space made available for the accommodation (long-term storage) of sediment accumulations. In sequence stratigraphy, sedimentology contributes important criteria for the identification of key surfaces for correlation, through which one may develop a dynamic view of evolving palaeogeography. Sedimentology draws upon and contributes to geological subdisciplines such as geochemistry, geophysics, mineralogy, palaeontology and tectonics, and upon sciences such as biology, physics, chemistry, civil engineering, climatology, fluid dynamics, geomorphology, glaciology, oceanography and soil science.

Sedimentary structures, which may be understood by input from all these subdisciplines, are generated from materials of diverse compositions and are observably products of physical, chemical and biological processes. Whilst certain processes are common to many present-day environments, combinations of processes, often with particular directional properties, may be restricted to specific environments and can hence form a basis for palaeoenvironmental reconstructions. In present-day settings, combinations of processes vary laterally, in kind and intensity, from sub-environment to sub-environment. Comparable changes of processes can be inferred for the past if we learn to read the sedimentary structures that help to characterize different units in the rock record. However, the complete characterization of rock units is based on more than just the sedimentary structures and will commonly also involve palaeontological, compositional and textural features, some of which may only be identified by laboratory observation and analysis. Such characterization allows similar rock units to be grouped together as "lithofacies" (commonly just "facies") with the implication that different facies or a succession of facies will be interpretable in terms of a set of processes specific to a particular environment.

It is important to realize that the definition of facies and the assigning of rock units to facies are not determined by absolute criteria but will be determined by the aims and circumstances of any particular investigation. Although detailed discussion of facies analysis is beyond the scope of this book, Chapter 10 provides an introduction to some of the ways in which a study of sedimentary structures and processes may be developed into an appreciation of facies and environments, notably for the purpose of palaeoenvironmental reconstruction.

1.3 Sedimentary structures and science

From our example at the beginning of the chapter it is clear that the subdisciplines of geology differ from those of the basic theoretical sciences in that they are not necessarily concerned with generating and testing universal laws. In geology, established laws are commonly taken for granted and are used to find and hopefully solve particular problems relating to what happened sequentially at particular times and in particular places. Geology involves interpreting processes from products (rocks) that formed a long time ago, which almost certainly record only part of what actually happened and for which there is no way of testing the interpretation in absolute terms. In that sense the science can be regarded as "practical" rather than "theoretical", but is nonetheless satisfying for that. In some respects, geology has much in common with criminal detective work where a story has to be reconstructed from fragmentary evidence and where much of the procedure involves elimination of possible alternative explanations.

Sedimentologists work within a set of principles, held by all scientists, many of which are probably implicit in the way in which you tackled the initial exercise:

- Determinism: that Nature is constant with respect to its laws and that scientific laws are invariable with respect to time, space and circumstances. In other words, it is reasonable to believe that, for example, gravity, the principles of mechanics, the nature of chemical reactions have always operated in the same way. Application of these ideas is typically referred to as **actualism**.
- Uniformity of processes: that present-day processes, either directly observed or confidently inferred, are sufficient to explain phenomena that we observe in the rock record. There should be no need to invoke processes that cannot be seen today in order to explain rock units. This is the philosophy of **uniformitarianism**, an idea promoted particularly by Lyell in the early 19th century, and now regarded as somewhat flawed. Its most significant aspect is that it insisted that the *rates* of processes have remained constant through geological time and that the rock record must be explained in terms of gradual changes ("**gradualism**"). It is important to recognize that uniformitarianism is a more restricted sub-set of actualism, whose principles are universally accepted. Lyell's gradualist uniformitarianism was set against the idea of **catastrophism**, which suggested that major, sudden events had significantly altered the course of geological history, in particular the fossil record. The early widespread acceptance of Lyell's advocacy, whilst helpful in the solution of many problems, proved inhibiting in the longer term, and there has been a progressive acceptance, since Lyell's time, that the geological record is punctuated with "catastrophic" events. At the largest scale, these include asteroid impacts and major phases

of volcanic activity that have led, through associated secondary climatic consequences, to worldwide changes in the fossil record, in particular mass extinctions. Such major events are rare and widely spaced through the rock record. However, the recognition of catastrophic events at smaller and more local scales does impact on everyday sedimentology. Earthquakes, tsunamis, volcanic eruptions, submarine and subaerial landslides can all produce rare but distinctive deposits or discordances whilst, at a less spectacular scale, major floods or storms produce **"event beds".** Clearly there is a whole spectrum of scale and magnitude of "abnormal" events across which gradualist and catastrophist approaches converge.

- Continuity: that nature is continuous through space and time.
- Parsimony (**Ockham's principle** or razor): that the simplest hypothesis or theory offers the most likely explanation of the facts.

In addition, sedimentologists use procedures that perhaps were adopted intuitively by you in attempting the exercise associated with Figure 1.1. They attempt to develop:

- *Conjectures or speculations*: rapidly conceived, intuitive ideas about relationships of observed phenomena; hunches to be tested against the evidence at hand.
- *Hypotheses*: untested explanations of observations, logically developed and tentatively adopted. Against these you can deduce, on available evidence, whether critical aspects of your explanation are false and which can be accepted for the time being as worthy of further testing. Good hypotheses predict a great many consequences, many of which can be tested in a variety of ways, perhaps by experiments. Initial hypotheses are typically rather tentative and, in most investigations, it is necessary to set up multiple working hypotheses, (i.e. several possible ideas about the solution of a problem). This avoids becoming blinkered by a single explanation and thereby risking pursuing it into a dead end. The standing of any hypothesis changes as the evidence increases and as critical predictions of any explanation are tested against the available facts. Scientific "truths", in the form of well developed, long-standing hypotheses, are still subject to constant scrutiny and revision in the light of further data and are not absolute.
- *Theories*: coordinated sets of self-consistent hypotheses, each of which has been tested many times and remains a valid explanation of observations. A theory that remains valid should not be subject to any exceptions (or extremely few). However, beware of the fact that many theories are known to last longer for social reasons than is justifiable with hindsight (e.g. Lyellian-Darwinian gradualism or the fixity of continents). Theories may encompass and supersede one another.
- *Models*: idealized simplifications set up as an aid to understanding of, and communication about complex relationships between phenomena and processes, commonly employed to illustrate working hypotheses. We may draw up actual models based upon a modern environment (e.g. a desert) using data from a particular basin in the Sahara, or an inductive model (e.g. a sand-sea and sand-dune dominated desert) based on a synthesis of features of many basins in the Sahara. We may make scaled, experimental models to detect, under controlled conditions, say in a wind tunnel, the processes and variables responsible for particular structures, (e.g. aeolian sand ripples and flat beds). Mathematical models attempt to simulate complex geological processes. In our desert example, the effects of changed wind direction and strength, increased sand flow rate and change of grain size on the form of the sand sea, the sand dunes and the smaller structures therein, might be predictable from the developed model. Visual models, either diagrammatic or realistic, help us to see relationships and to picture processes, products and environments. Models may be static or dynamic. Static models are descriptive of a particular time in the past, yet are still predictive of many relationships, as in a palaeogeographic map. Dynamic models attempt to show a changing pattern or dynamic equilibrium of processes and environment over a period of time, or a steady-state equilibrium over the same period. Many sedimentary processes, ranging in scale from the movement of single particles to the infill of an entire basin, are now modelled by computer. These models increase in sophistication as physical and chemical processes become better understood and quantified. They allow the geologist to play complex "what-if" games in order to better understand the interactions of coexisting processes and often to demonstrate that a particular end product does not have a unique origin but may be produced from different combinations of circumstances.

The facies model is particularly important in arriving at an interpretation of a succession of sediments. This is a generalization and simplification of the observed vertical and lateral relationships of the facies observed in an accumulated sedimentary succession. It is typically an attempt to reduce the natural "noise" of the relationships and to reveal an underlying, commonly occurring pattern that can then be compared with predictive actual models derived from studies of present-day environments. Although this approach is mainly beyond the scope of this book, sedimentary structures are typically critical to the establishment of facies schemes, and thus are amongst the fundamental building blocks of predictive sedimentology used to propose facies models.

Scientists working in a particular field share values, practices (methodology) and beliefs (philosophy). These constitute a paradigm and help to build a scientific consensus. The general geological paradigm within which the proper interpretation of sedimentary structures could take place originated between 1785 and 1860 through the thinking of pioneers such as James Hutton and Charles Lyell. The basic principles of a specifically sedimentological paradigm can be traced to the work of Henry Clifton Sorby and Johannes Walther between 1850 and 1900, but it has only become fully developed since the 1950s.

Two major insights are associated with the names of Sorby and Walther. Henry Clifton Sorby (1826–1908) may truly be regarded as the father of sedimentology, for between 1850 and 1908 he pioneered most of the approaches that we develop in this book. He recognized the problems of understanding ancient sediments in terms of process; that critical questions had first to be identified by making acute field observations and careful records in the light of a thorough understanding of processes. As an aid to observation, he made thin sections and used the polarizing microscope. As a better guide to understanding processes and products, he performed experiments, for example by generating ripples and cross lamination by current action under laboratory conditions. He was the first to measure the orientation of structures such as cross bedding in the field and, above all, he used his understanding to make environmental reconstructions and put them in a palaeogeographic context.

By contrast, Johannes Walther (1860-1937), in his *Introduction to Geology as a Historical Science* (1890–93), drew together many scattered observations on modern sediments and processes, and demonstrated implicitly the power of the actualistic method as a basis for interpreting sedimentary rocks. In addition, he established a powerful stratigraphic principle: **Walther's principle of the succession of facies** (1894). This states that, unless the evidence indicates otherwise, we should expect the processes and environments that occur laterally adjacent to each other to be represented by facies that succeed each other non-erosively in a vertical geological column (see §10.3.3). Sadly, Sorby was ahead of his time, and worked somewhat in isolation whilst Walther wrote in a language not readily accessible to the Anglo-Saxon world, and it was not to be until the second half of the twentieth century that their ideas started to be developed and applied to sedimentary successions in a rigorous and widespread manner.

Study techniques

Recommended references

Allen, P. A. 1997. *Earth surface processes*. Discusses the physical basis for the main types of natural processes that act to erode, transport and deposit sediment.

Blatt, H., G. V. Middleton, R. C. Murray 1980. *Origin of sedimentary rocks* (2nd edn). A very sound and clear general textbook.

Boggs, S. (Jr), 2016. *Principles of sedimentology and stratigraphy*. A comprehensive and thoroughly updated text book.

Bridge, J. S., & R.V Demicco. 2008. *Earth surface processes, landforms and sediment deposits*. Detailed coverage of many aspects of sedimentology, including an overview of common sedimentary structures.

Friedman, G. M. & J. E. Sanders 1978. *Principles of sedimentology*. A very extensive reference list; if you cannot find an early reference, look here.

Jones, S. 2015. *Introducing sedimentology*. A succinct and simple introduction to the discipline of sedimentology and a useful book for those wanting to gain a basic understanding.

Middleton G. V. (ed.) 2003. *Encyclopedia of sediments and sedimentary rocks*. A series of short articles on most major topics in sedimentology; a good starting point for many investigations.

Leeder, M. R. 1999. *Sedimentology and sedimentary basins: from turbulence to tectonics*. A comprehensive textbook suitable for advanced-level undergraduates and covering sedimentology from the grain scale to the basin scale.

Nichols, G. 1999. *Sedimentology and stratigraphy*. An introductory text that covers the essential elements of the discipline at a basic level; illustrated with very simple diagrams suitable for first-year undergraduate students.

Prothero, D. R. & F. Schwab 2014. *Sedimentary geology*. An introduction to sedimentary rocks and stratigraphy.

Reineck, H. E. & I. B. Singh 1980. *Depositional sedimentary environments*. A beautifully illustrated book with good photographs of modern structures; a rather patchy treatment of sedimentary environments.

Selley, R. C. 1976. *An introduction to sedimentology*. A lively introduction to the subject; idiosyncratic in parts, but worth reading.

Stow, D. A. V., 2005. *Sedimentary rocks in the field.* A colour guide. An accessible and well illustrated introduction.

Tucker, M. E. 2001. *Sedimentary petrology*. A very sound introduction to petrography; particularly useful for developing more detailed ideas about carbonate sediments.

Tucker, M. E. 2003. *Sedimentary rocks in the field*. An excellent pocket guide for use in the field.

CHAPTER 2
Bedding

Bedding is one of the most distinctive features of sedimentary rocks and its occurrence is often associated with the development of many of the sedimentary structures that are dealt with in this book. Some broad understanding of the nature of bedding, its genesis and recognition is, therefore, an important starting point for studying sedimentary rocks, whether they are being considered from the point of view of their stratigraphy, their sedimentology or their post-depositional structural deformation. This chapter reviews some of these broader aspects of sedimentary successions, which can be important in understanding the context of particular sedimentary structures.

2.1 The nature of bedding

2.1.1 Where to start: recognizing sets of beds

When you approach any exposure of rock, you might usefully start by asking the following questions. You should not be discouraged if you cannot give clear answers to them all, especially if you have rather limited experience. However, it is hoped that by the time you have studied this book, you will be able to answer more of the questions with greater confidence.

- Can anything in the rocks be detected that suggests that they are bedded (i.e. arranged into layered accumulations)?
- Is there other evidence to suggest a sedimentary origin?
- If the evidence suggests that they are of sedimentary origin, is there evidence to suggest which is the top and which is the bottom of the succession? Note that, with very few exceptions, this will only be a relevant question where the rocks are clearly strongly deformed.
- Are there any features that are indicative of particular processes or environments of deposition, for example beds with erosive, channel-shaped bases?
- Are there vertical and lateral changes in the rocks that might suggest processes that changed over time and/or space, for example a distinctive vertical or lateral thinning or thickening of the beds? Might such changes be indicative of a particular environment of deposition?

2.1.2 The basis of this approach: the origins of bedding at the present day

In trying to answer some of these questions it can be helpful to think about simple experiments to examine sediment deposition, and about processes that can be observed in modern depositional environments. From simple observations it is possible to establish that if physical conditions and sediment supply remain steady (i.e. constant in time), then a body of sediment is deposited that is internally homogeneous in its composition and texture and in the nature of any internal lamination. Where physical conditions or sediment supply change through time, layers of sediment of somewhat different character are laid down. The boundaries between such layers may be sharply defined or they may be gradational, depending on the way in which processes or sediment supply changed and on the resulting textural characteristics of the sediments that make up the layers (Fig. 2.1). Many such layers of sediment possess effectively planar bottom and top surfaces, and are very extensive laterally in relation to their thickness; this might imply the distribution of sediment over a wide area and its deposition on a surface which itself was of low relief, for example on a fluvial (river) floodplain. Other

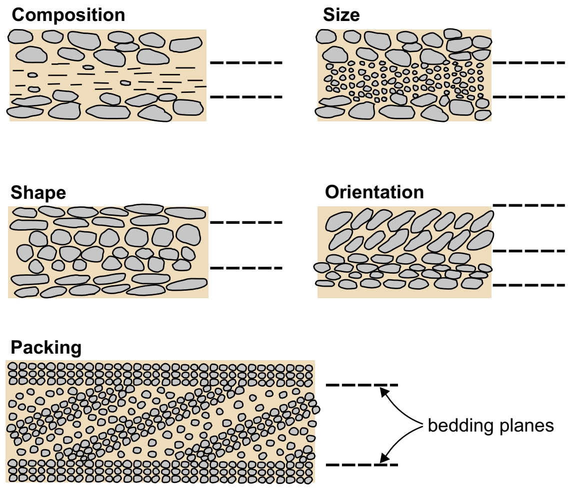


Figure 2.1 Bedding as the product of different combinations of grain composition, size, shape, orientation and packing. Modified after Pettijohn et al. (1972) and Griffiths (1961).

layers of sediment are more restricted laterally, possibly reflecting depositional processes that were not uniform (i.e. not constant in space), for example where confined as part of the fill of a relatively narrow fluvial channel.

Depositional units greater than 1 cm thick are known as **beds**. Where their boundaries are fairly sharply defined, these are known as **bedding** or **bounding planes**, the lower bounding surface often being referred to as the **sole** and the upper as the **upper bedding surface** (Fig. 2.2). Where boundaries are more gradational in character, bedding is defined less precisely. The terms **layers** and **strata** are sometimes used rather loosely as equivalents of "bedding" but "strata" may also be used at a larger scale to encompass a whole succession of constituent beds. Below 1 cm thickness, depositional units are termed **laminae**: the smallest units visible in a sequence. Layers and laminae that occur within beds and which are inclined at an angle to the main bedding surfaces are called **cross strata** (which include **cross laminae** and **cross beds**). The general phenomenon of inclined layers is termed **cross lamination** or **cross bedding**, depending on scale (see Ch. 6). Groups of similar beds may form **cosets** or **bedsets,** which may be **simple** or **composite** (Fig. 2.2).

In some ancient sedimentary successions, the rocks split along surfaces that are parallel to bedding but which occur within internally uniform beds. In such cases, the term **splitting** or **parting plane** should be used, as the surfaces may not necessarily correspond to bedding planes (Fig. 2.3).

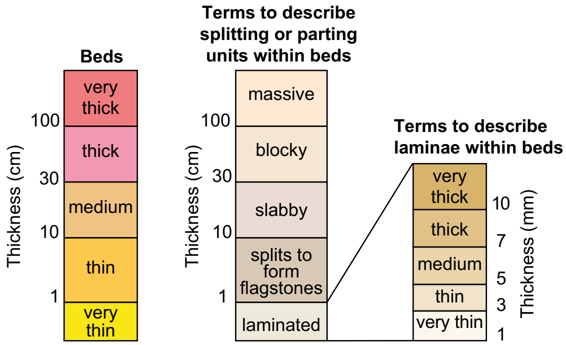


Figure 2.3 Terminology for thickness of beds and the description of units within beds due to splitting or parting, often after weathering. Modified from Ingram (1954), Campbell (1967) and Reineck and Singh (1973).

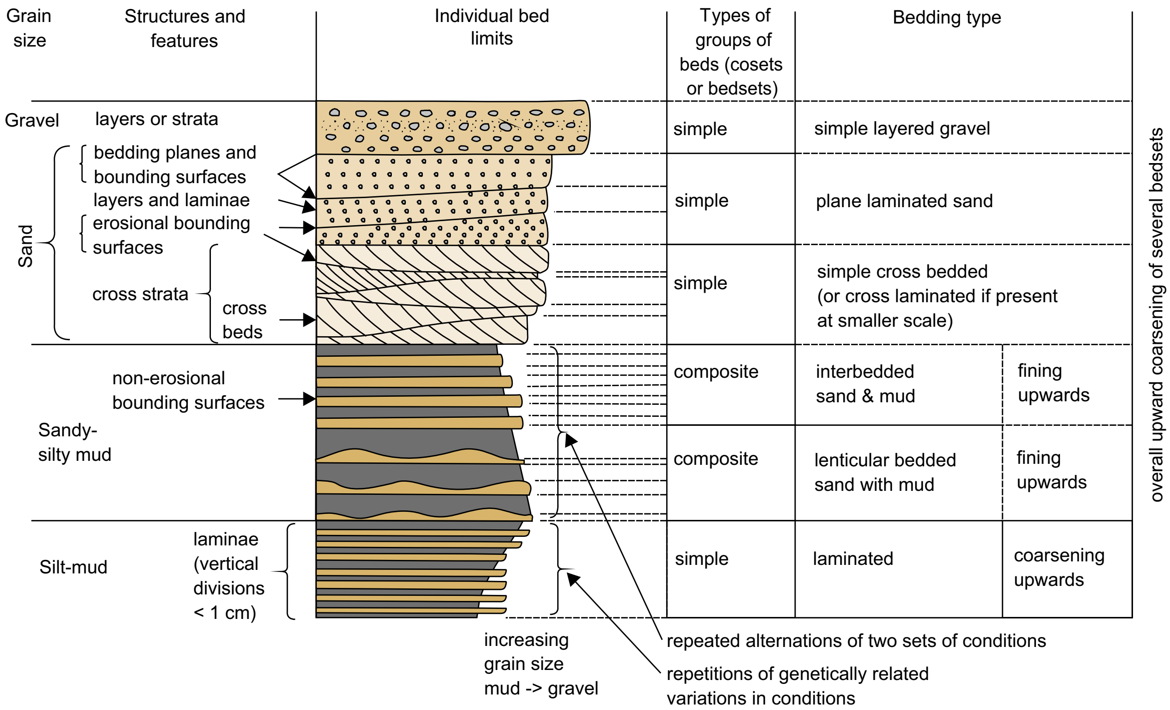


Figure 2.2 A scheme illustrating the terminology used to describe sedimentation units. Modified after McKee and Weir (1953), Campbell (1967) and Reineck and Singh (1973).

Many beds and bedsets maintain their thickness for considerable lateral distances (Fig. 2.4A), although all eventually thin out or change their nature, either gradationally or suddenly, when traced far enough. Vertical sections through deposits of river floodplains, estuarine flats or beaches, as seen in excavations or exposed in the erosive banks of migrating channels, typically show successions of beds, where each bed records a particular set of conditions. Vertical successions of beds commonly record systematic changes in bed thickness, composition or grain size (Fig. 2.4).

Any sedimentary structure that cross-cuts a bedding feature – for example a channel cutting down into horizontal layers (Fig. 2.5) – must have formed after that feature. Also in such a situation, fragments from an older bed could have fallen into and been incorporated within the younger bed. Both cross-cutting relationships and included fragments are evidence that might be used in distinguishing younger from older beds where they have been strongly disturbed by tectonic deformation.

2.1.3 Basic stratigraphic principles based on present-day phenomena

Observations of present-day processes and products allow us to establish several stratigraphic principles that help us recognize beds and understand their structural significance in ancient sedimentary successions that have been tectonically disturbed. These are:

- Original horizontality of beds: most beds are laid down either parallel to the Earth's horizontal or at very low angles to it. Exceptionally, beds deposited with primary depositional dips of up to 40° may be recognized by their associated sedimentary structures.
- Original continuity of beds: individual beds and groups of beds are commonly laterally extensive and maintain their thickness and continuity for great distances. Individual beds and layers are more likely than groups of beds to be lenticular over short distances.
- Superposition of beds: younger beds are deposited on top of older beds in an arrangement called a sequence or succession.
- Way-up; the tops and bottoms of beds can be recognized by their associated sedimentary structures, which are found either on upper and lower bed-bounding surfaces or within the beds.
- Included fragments (most obviously clasts): fragments of older sediment can be included in a younger deposit, but not vice versa.
- Cross-cutting relationships: a feature that cuts across (i.e. truncates) a bed must be younger than the bed itself.
- Strata identified and correlated by their included fossils: strata may be dated and correlated by the components and assemblages of the fossil flora and fauna within them, though with variable degrees of precision. Fossils may be present within beds or on bedding surfaces.

2.1.4 Applying stratigraphic principles to ancient sedimentary sequences

The principles of original horizontality, original continuity, superposition and cross-cutting relationships should form the basis of any initial investigation of rocks that have been tilted relatively little from their original attitude. Where rocks are suspected to be strongly deformed, the use of way-up features and sequences of fossils may permit the detection of overturned beds. In such cases, the term **younging** is used to indicate the direction of the top of the sequence. A succession could therefore be reported as "younging to the east", for example. Application of the principles of original continuity, cross-cutting relationships and included fragments may also enable features such as faults, unconformities and channels to be identified.

Identification of a similar succession of beds on either side of a structural discordance may allow the nature and displacement (throw) of a fault to be estimated. If changes in thickness of individual beds across the fault are apparent, this may indicate that the fault was active during deposition.

The truncation of otherwise continuous beds indicates later erosion. Where the erosion surface is flat and the beds above are concordant with the surface, an unconformity may be inferred; an inference that would be enhanced by the presence of fossil groups of significantly different ages above and below the surface. Where there is an angular discordance between beds below and above such an unconformity, a period of tilting and erosion must have occurred between deposition of the two groups of beds. The interval of time represented by an unconformity may, in many cases, be established from the fossils above and below the discordance.

Where truncation occurs at a surface showing relief, with beds above and below broadly parallel with one another, some form of channel may be present. No intervening tilting need be inferred and the time gap across the erosion surface may be small. However, that should not be assumed and a check using fossils should be attempted where possible.

2.1.5 Preliminary observation and recording of bedding

Several levels of observation and investigation of bedded sediments are possible, from the large-scale distant

Figure 2.4 Some examples of bedding in various sedimentary successions. A) Thick parallel-sided sandstone beds with thin mudstone interbeds. Hell's Mouth Grits, Cambrian, north Wales. B) Mudstones with interbedded thin, parallel-sided and laterally continuous sandstone beds. Aberystwyth Grits, Lower Silurian, west Wales. Photo courtesy of Gilbert Kelling. C) Broadly parallel, horizontal bedding developed in fluvial gravels as a result of textural differences (grain size and sorting) between layers. Modern, Kverkfjöll, central Iceland. D) Large scale trough cross-bedding in sandstones (foreground) overlain by horizontally bedded finer-grained sediments (background). The boundary between the two major units represents a marked change in depositional conditions. Cedar Mesa Sandstone and Organ Rock Formation, Permian; southern Utah. E) Mudstones with interbedded thin sandstone beds. Carboniferous, Broadhaven, South Wales. F) Thick, parallel-sided sandstone beds interbedded with thin, dark mudstones. Traced laterally, some thin mudstones die out. Ross Formation, Upper Carboniferous, western Ireland.

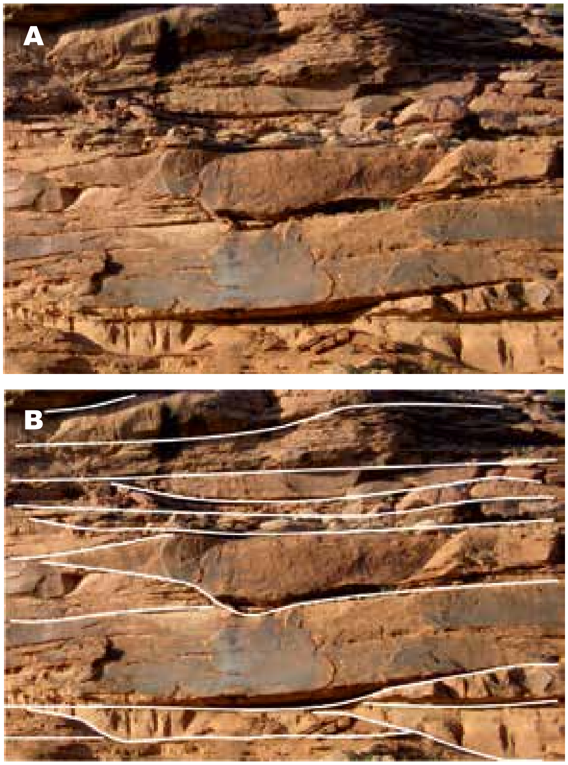


Figure 2.5 Sedimentary structures that cross-cut older bedding features. Channel elements with erosional bases that cut down into older, underlying strata. Width of view is ~80m. Jurassic Kayenta Formation, Utah, USA. White lines highlight erosional surfaces that bound the bases of channel bodies.

overview to small-scale scrutiny with a hand lens. At first it is useful to scan exposures from a distance in order to decide the general attitude and orientation of the beds. It is also helpful to work out, at an early stage, the way-up of the succession and where, in general, the older and younger strata are to be found. In many cases, beds will be only slightly tilted, if at all, from the horizontal and determining way-up will not be a serious problem. Where it is an issue, it may often be established at a distance by recognizing the bases of cross-cutting features such as large-scale channels. However, it is most likely to be confirmed by closer observation. From a distance it is possible to ask and commonly give preliminary answers to the following questions:

- Can way-up be determined? Are the bases of any beds very irregular on a large scale? Are there major channels cutting into underlying beds?
- Do the beds appear to form a conformable succession throughout the exposures? Are there groups of strata inclined at different angles or does one group have its lateral extent terminated by a second group? Hence, is there any likelihood of a major time gap, i.e. an unconformity?
- What is the approximate spacing of prominent bedding planes? Are the beds, so defined, of uniform thickness? Do successive beds thicken or thin upwards? Are there patterns of repetition of bed-thickness change vertically through the succession?
- Are there any suggestions of variation of grain size in the vertical succession? Are there beds, or groups of beds, that consistently "fine upwards" or "coarsen upwards'?
- Are vertical changes of lithology suspected, e.g. limestone – shale – sandstone – conglomerate? Are there patterns to such changes?
- Can the succession be divided into packages or units of contrasting aspect? Are there any systematic variations in the vertical sequences as characterized by bed thickness, grain size and composition?
- Do individual beds or groups of beds change thickness laterally and, if so, how? Are there any lateral changes of grain size and lithology?

It is a good idea to get into the habit of recording these preliminary observations and practising estimating the dimensions of the larger beds and successions of beds. From this kind of initial analysis, ideas will commonly emerge about where to start detailed work, where to sample, and where to locate key features and surfaces.

2.1.6 Detailed observation and recording of bedding: methodology

Although this section is mainly aimed at observations made at outcrop, most of the issues discussed are also applicable to observations of borehole cores.

Attitude of beds

Detailed work on the outcrop should begin with measurement of the attitude in space of beds (i.e. direction of strike and magnitude [angle] of dip) at a representative selection of places. If the dip is evidently very constant across the area of interest, a small number of measurements will suffice; if the structure is complex then it may be necessary to make more measurements, making sure to carefully locate their position on a map or on a preliminary sketch of the outcrop. These data, which will be recorded both on a map and in a notebook, are necessary for making any corrections to measured sections and for re-orientating measurements of inclination and alignment of sedimentary structures recorded in the field. Through these observations, the structures can be restored (on a stereonet) to their original depositional attitude and orientation (see Appendix 1). Usually such corrections are necessary only for dip, but sometimes plunge also needs to be considered.

Whilst measuring dip and strike, look out for structures that show way-up, and, where possible, test in detail the applicability of the principle of original horizontality. Some way-up structures are known as **geopetal** (spirit level) structures and, although quite rare, they can reveal the attitude of the original horizontal with some confidence, e.g. the boundary between sediment and crystalline infill of a brachiopod shell half-filled by sediment (e.g. Fig. 8.20). Such surfaces commonly coincide in attitude with bedding but, in some cases, they may diverge considerably. Beds formed on a reef front may, for example, have an original or initial dip of 30° or more and this might be detected by comparing the attitude of bedding with that of a geopetal surface.

Successions

An initial aim of work on most sedimentary successions should be to observe and describe them accurately and concisely, and to divide them into beds and bedsets. Where possible, measure and record several laterally equivalent vertical sections in the same interval, selected to document any lateral variability suspected from preliminary observation. This enables local variability to be distinguished from any regional variability and ensures that larger scale trends are seen in proper perspective. Be sure that measurements of thickness are made normal to the bedding or, where this is not possible due to restricted accessibility, make sure that any oblique measurements are adjusted to true vertical thickness (TVT) by simple trigonometry.

The division of the outcrop or core into measurable units may be rather arbitrary. The level of detail of the work will vary with the nature and scale of the questions being asked about the rocks and with the experience of the geologist. In all cases, an effort should be made to apply consistent criteria and to sample in a representative way (Fig. 2.6). It

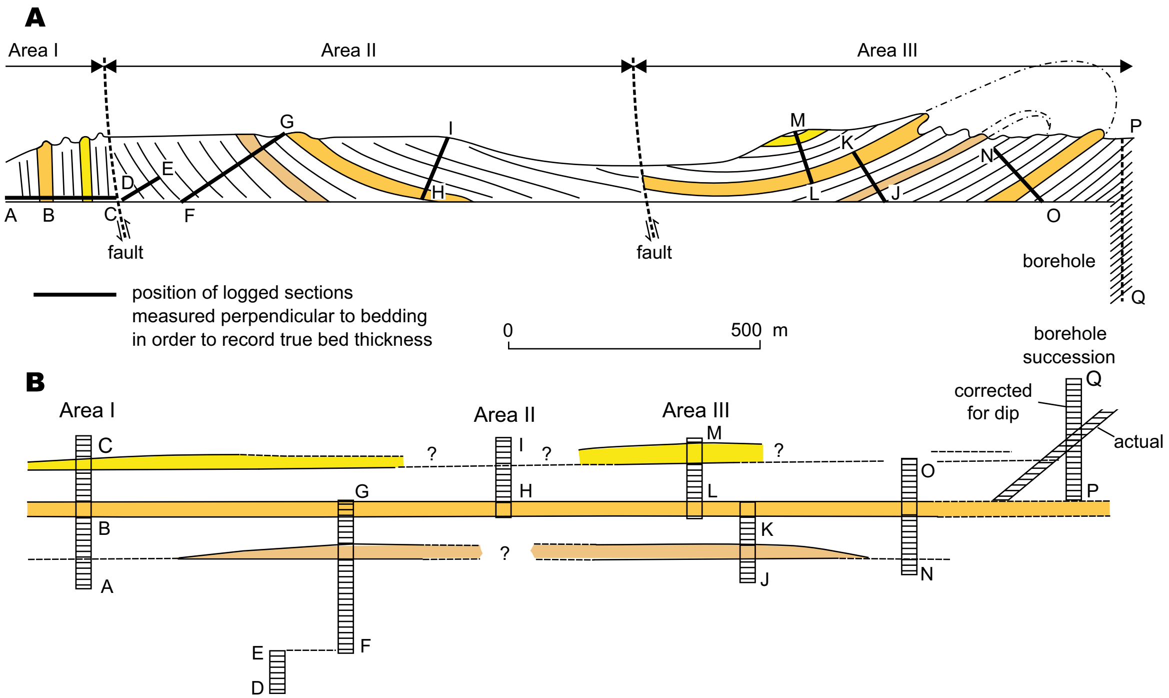


Figure 2.6 Illustration of a strategy for logging a sedimentary succession that has been tectonically deformed. A) Sketch of cliff section showing deformed beds. Note that the section has been divided into three areas, each bounded by a major fault. Also note the overturned strata at the right-hand end of the section. Sedimentary logs are measured perpendicular to the bedding so as to record the true bed thickness. Depending on exposure and degree of accessibility, it may be necessary to construct several log sections. In such cases, distinctive, laterally extensive marker beds can be used for correlation. B) Restoration of the folded succession to its pre-folded disposition. Notice how the absence of certain units in particular areas must be accounted for by lateral facies changes with the pinch-out of beds.

is important to employ systematic working procedures based on the stratigraphic principles already set down. Look laterally along any bed or layer to see what happens to continuity and thickness; look first beneath, then within, and then on top of the unit; investigate vertically, beginning with the oldest beds, and working systematically to the younger beds. Methodical habits of measuring and recording generally enhance powers of observation and make pattern recognition easier; the search for patterns within data then becomes second nature. Many of these points will be expanded and illustrated later, particularly in Chapter 10, where methods for the interpretation of measured successions are discussed.

In some circumstances, recognition of bedding and the definition of upper and lower bounding planes may be difficult. Bedding must be distinguished from tectonically induced cleavage, from joints and faults, and from colour banding due to diagenesis (post-depositional alteration) and weathering, all of which may cut across depositional features. Changes in composition or grain size are the best guide to identifying bedding and these are often more apparent on weathered surfaces. Differential weathering and erosion may accentuate differences that are virtually invisible in fresh rocks. In other cases, deep weathering obscures depositional structures. A few prominent subparallel splitting planes, or the suggestion of bedding planes in the shape of an outcrop, may provide the first clue to the orientation of bedding. Changes in colour, mineral composition, texture (grain size, grain-size variation, grain shape, porosity, packing, degree of cementation, hardness), internal structure (lamination, bedding) and orientation may serve to confirm or refute such an initial impression (Figs. 2.1, 2.2, 2.6, 2.7). In cores, colour is commonly a guide to grain size, with coarser sands commonly being pale and increasingly finer grained sediments progressively darker, though exceptions to this are not uncommon.

The approach illustrated in Figure 2.6 relates to a method of working that is essentially one-dimensional: the logging of a vertical section. It is often helpful, however, to photograph a group of beds in two dimensions from a series of positions equidistant from and

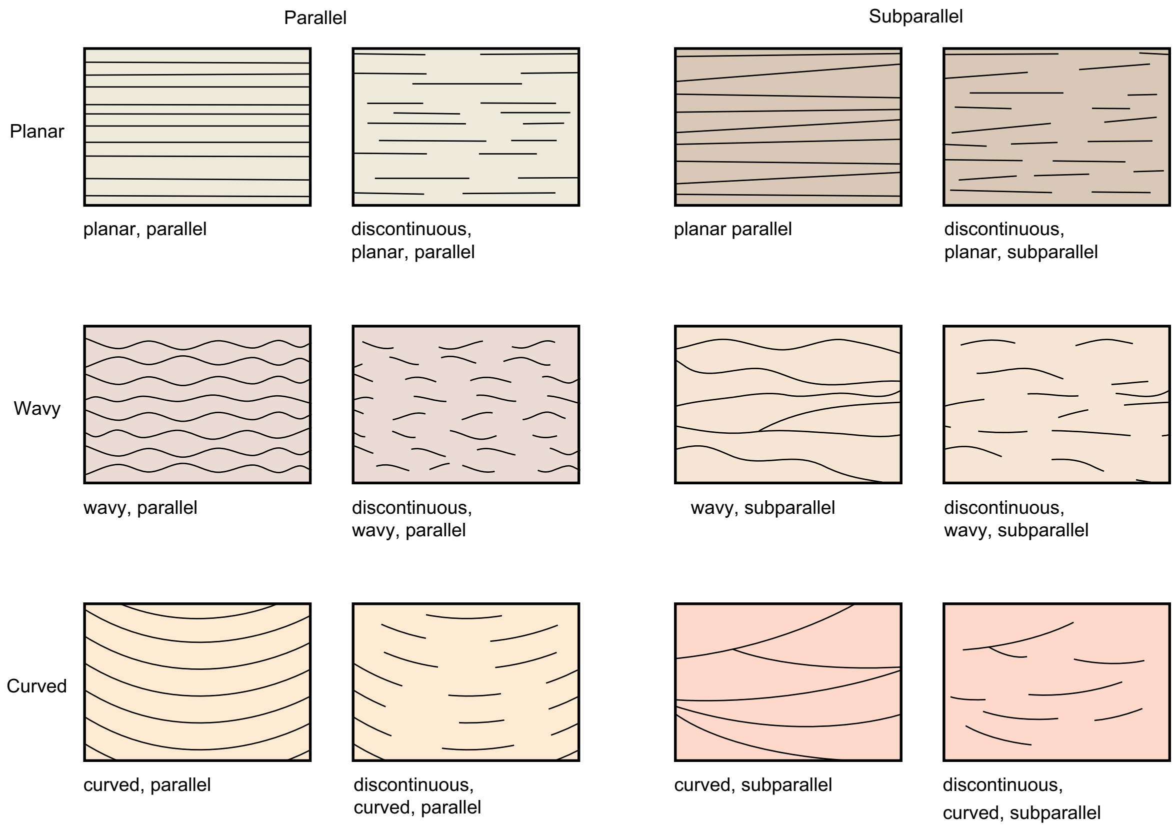


Figure 2.7 Useful bedding-lamination terminology. Modified after Campbell (1967) and Reineck and Singh (1973).

perpendicular to the outcrop. In this way a mosaic of overlapping photographs can be assembled and the more important bedding structures may be drawn from the resulting panorama. Subsequent checking against the outcrop helps to focus attention on critical details. This can be particularly useful where local lateral variability is apparent and where a one-dimensional approach is inadequate to fully characterize the succession. Where the exposure has promontories and recesses or is covered by vegetation, it is important to draw scaled diagrams which generalize the geometry of the outcrop. In some cases, the use of a digital laser scanning device (LiDAR – Light Detection And Ranging) can help with this exercise, as can stereo photogrammetry, in some cases acquired using a drone.

Thickness of sedimentary units

Measurement of thicknesses of beds, using Standard International Units (i.e. metres), should take place near to the vertical line (i.e. normal to bedding) selected. The position of this line is often determined by the accessibility of the exposure, as in steep, cliff-forming sections. In other outcrops the decision may be either selective or simply arbitrary. When measuring core, it is important to take into account both any tectonic dip and any well deviation from vertical in restoring thicknesses to their true value. Boundaries of units will normally be clear where bounding/bedding planes are sharp, but subjective decisions are often necessary where contacts are gradational. Where rocks are dipping and access is restricted, as at the foot of a cliff, then a series of short sections, whose thickness is determined by accessibility, may be compiled to give a dog-leg section. Although such a section is not strictly vertical at a single place, this should not matter provided there are no rapid lateral variations in lithology or thickness.

Lateral variations

The overall shape or geometry of a bed may be established by tracing it laterally, thereby testing the principle of original continuity. If beds terminate, they do so in one of four ways:

(a) by convergence and merging of their upper bedding surface and sole, whereupon they may be termed **lenticular**;
(b) by lateral gradation of the composition of the bed so that definition of the bounding surfaces dies away;
(c) by being truncated by a cross-cutting feature such as a channel;
(d) by meeting a cross-cutting feature such as a fault or an unconformity.

Lateral inspection will sometimes reveal that in case (a), and partly in case (b), beds lap onto and drape previous structures (for example a channel margin or an organic mound or bioherm). In some such cases the angle of drape may be very low and the relationship therefore subtle; in other cases it may be accentuated by post-depositional differential compaction. The upper bounding surface and the sole may be parallel or divergent, continuous or discontinuous, and either can be planar, wavy or curved (Fig. 2.7).

Features within a bed or bedset

Vertical variations within individual beds are typically due to changing grain size, composition, texture or internal structure. Beds and bedsets may be: (a) homogeneous or heterogeneous; (b) rhythmic; and (c) gradational. Their lithology may vary from homogeneous (e.g. uniform well-sorted sandstone or siltstone) to heterogeneous (e.g. silty-mudstone, pebbly sandstones). Homogeneous beds are sometimes apparently structureless but they may reveal unsuspected internal structure if special techniques (e.g. X-radiography) are applied to slabbed specimens in the laboratory. With core, wetting a cut surface typically reveals internal structure more clearly. Some beds are heterolithic due to sorting into repeated interlaminations of sediment of contrasting composition or grain size, e.g. silt and sand, silt and mud (Fig. 2.2). Systematic variation of composition and grain size together from, say, sand in the base to silt with interlamination of mud in the upper parts of the unit, is a feature of some beds deposited by episodic, decelerating currents. In addition to variations in grain size, composition and texture, beds typically are characterized internally by the scale and style of depositional lamination (Fig. 2.7) and by various types of organic and inorganic post-depositional disturbance. Any complete description of the internal features of a bed should include all the above properties. Together, these might be a starting point for the definition of **sedimentary facies** (see Ch. 10 for a development of this idea).

The nature of bed contacts

In recording a succession of beds, special attention should be paid to the nature of **bed contacts**, which can be sharp

(Fig. 2.4A) or gradational (Fig. 2.4C). They may be marked by subtle or abrupt changes of composition and colour, texture and structure. Sharp changes may be non-erosional or erosional. Erosional bed contacts are marked by cross-cutting relationships, as at the base of a channel, or where there are downward projections on the sole of the overlying bed.

Some bed contact surfaces may represent significant intervals of protracted non-deposition and may result in the development of cemented layers, highly burrowed horizons or mature soil profiles. Establishing the full significance of such surfaces may involve correlation over wide areas and it is usual that only provisional judgments can be made on the basis of a single section.

Relationships between groups of beds

Sometimes bedding relationships at outcrop and on the scale of seismic reflection profiles show angular discordances, as has already been noted for some unconformities. The general terms **toplap**, **offlap**, **onlap** and **downlap** can be used to describe different types of bedding relationships (Fig. 2.8) that result from various styles of accumulation (e.g. progradation, migration and infill) at a wide range of scales.

Patterns in sedimentary successions

Patterns of vertical change in groups of beds and bedsets occur where thickness, grain size, composition or sedimentary structure changes systematically (e.g. Fig. 2.2). The following kinds of pattern of vertical change are commonly seen, although this list is not exhaustive:

(a) Grain size becomes progressively finer upwards, for example from coarse sandstone at the base to siltstone and mudstone toward the top (as in some meandering-river successions).

(b) Grain size coarsens upwards from mudstone to coarse sandstone (as in some deltaic successions).

(c) Systematic and repeating patterns of vertical lithological change such as shale – sandstone – coal or limestone – shale – sandstone (as in other kinds of deltaic successions).

(d) Patterns involving units of carbonate-sulphate, hydrous and anhydrous sulphate, sulphate and halides (as in chemically precipitated evaporite successions).

(e) Patterns of bed thickness change (as in thickening- or thinning-upward successions).

These types of patterns occur on a variety of vertical scales from a few metres to several tens of metres or even hundreds of metres.

Developing the skills to usefully describe sedimentary successions in the field or in borehole cores requires not only organized working practices but also the acquisition and use of a new terminology. This can be quite daunting at first but practice and discussion with others can quickly clarify issues and engender the necessary confidence. If outcrops or cores are not readily available, it is often possible to make a lot of progress by describing photographs (e.g. Fig. 2.9). For some sediments, particularly limestones, it may be necessary to collect samples and examine them in thin section in the laboratory in order to identify the component grains, which are usually most diagnostic of depositional conditions.

However, description is seldom an end in itself but is usually a step towards understanding the processes and environment through which the sediments were deposited. This book explains sedimentary features in such a way that interpretations can move progressively from first understanding depositional processes and thence to deducing likely depositional environments.

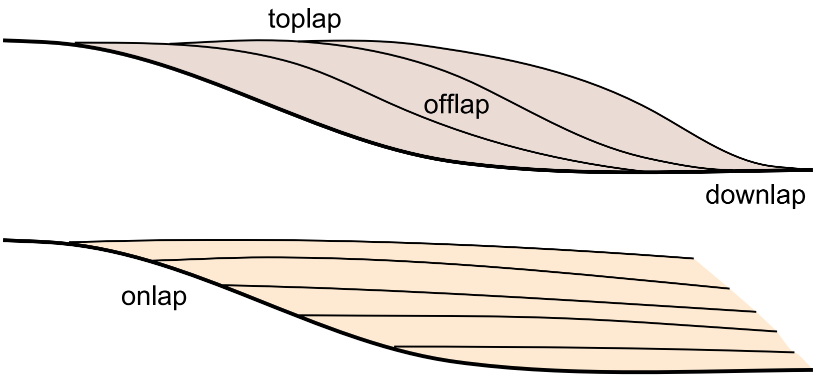


Figure 2.8 The terminology of discordant bedding relationships. These terms are used over a wide range of scales from that of a small outcrop to that of a seismic section. A) Terms associated with progradation. A conformable sequence of inclined strata deposited during progradation. Each stratum is succeeded laterally by progressively younger units, marking the direction in which the progradation took place. This relationship is commonly associated with regression whereby a shoreline (lake or sea) retreats basinward in response to progradation. B) Onlap of less steeply inclined strata onto a more steeply inclined margin.

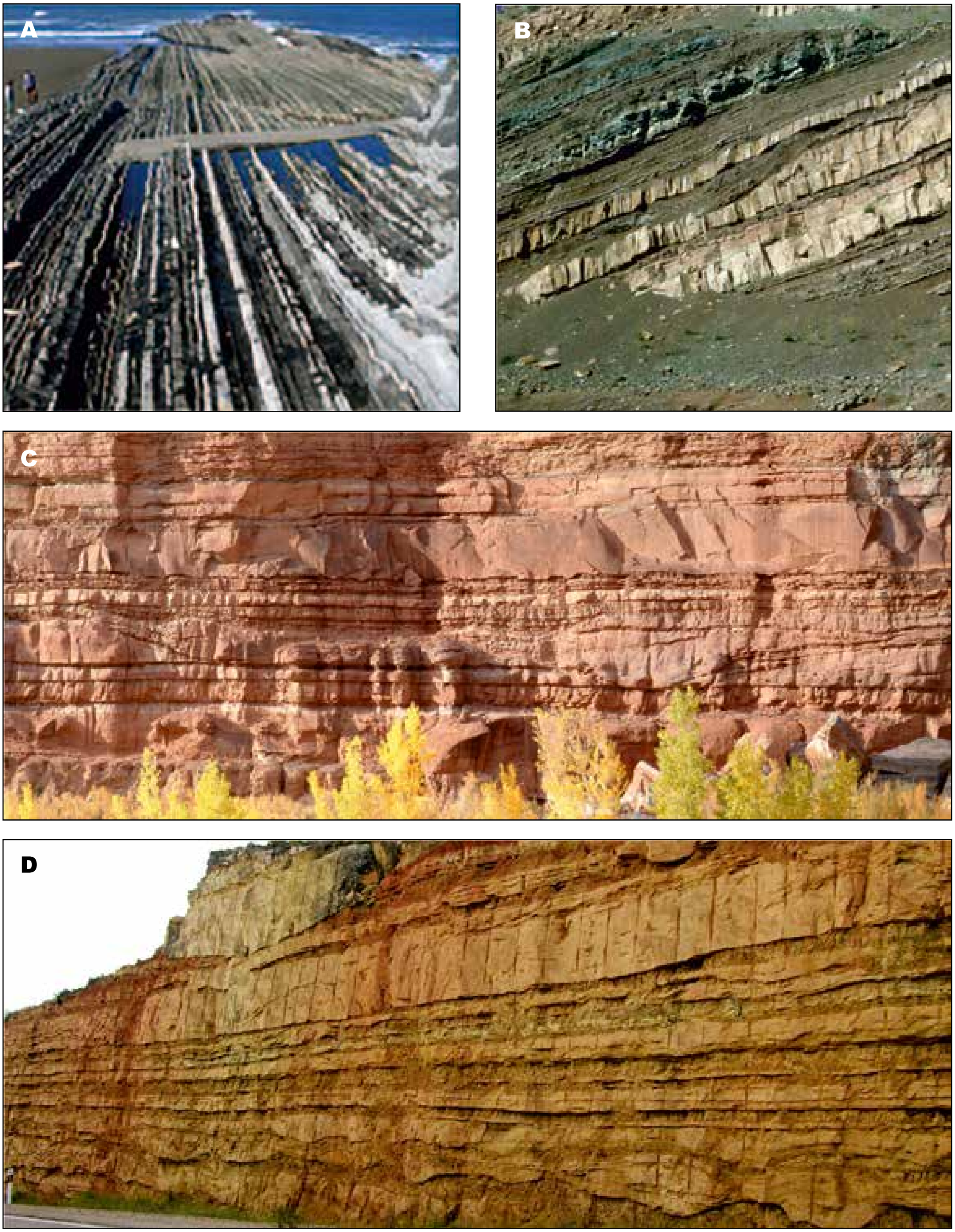


Figure 2.9 Try to describe the nature of the bedding in each of the four examples as fully as possible. See if you can draw any preliminary interpretations of the processes that were active during deposition of the sequences illustrated. The feature running across the centre of photo A is a concrete path. Thickest sandstone bed in B is ~5m thick. Cliff in C is ~25m high. Photo B courtesy of Gilbert Kelling.

2.2 The significance of bedding

2.2.1 Introduction

Individual beds are deposited either under essentially constant physical and chemical conditions or by systematic changes of process. Sharp bed contacts, true bedding planes and bounding planes represent changes in depositional conditions, in some cases involving non-deposition, erosion or a switch to a completely different regime. Other contacts, however, are more gradational and therefore likely reflect a more gradual change in conditions. Lamination typically results from minor fluctuations in quasi-steady conditions over a small area of the bed. Simple bedsets or cosets are the result of repetitions of genetically related variations in conditions. Composite bedsets (Fig. 2.2) represent repeated alternations of two contrasting sets of conditions. The preservation potential of beds and bedsets, i.e. their chance of becoming part of the geological record, is not necessarily related to the length of time over which they were generated. At the large scale, conditions of net subsidence must prevail and erosional activity in the area must be relatively subordinate in order for accumulation to occur. Periods of erosion may be sufficiently long for major time gaps (represented by unconformities) to be present in the succession, some of which may be caused by tectonic events. However, some depositional settings have within them features such as channels that are at least in part erosive. The erosion surfaces due to such features may punctuate the resulting sequence but are a natural part of the overall accumulation. Assigning time to such surfaces is seldom easy and it is, of course, in many cases impossible to know what type and thickness of sediments have been removed by an erosional episode. As a general rule, sediments deposited in the topographically higher parts of a depositional setting are more likely to be eroded than those laid down in the topographic lows (e.g. the bases of channels).

2.2.2 Basic processes of sedimentation

Several types of process contribute towards establishing the observed features of a bed. These are: physical, chemical, biological and diagenetic. Many of them are discussed in considerable detail in the later chapters, but they are briefly outlined here by way of introduction.

Physical processes

Most sedimentary rocks result from deposition of material transported as individual grains in suspension or near the bed (i.e. in traction) by water flows with low sediment concentration, although some are derived from denser flows or, at the extreme, from viscous mudflows. Others result from sediment being transported by wind. The nature and intensity of the transporting process depend on properties of both sediment and fluid. Grain size, shape and density are crucial properties of the sediment. The density and viscosity of the fluid, and the velocity of the flow, the strength of waves and the depth of flow are also important controls. A change in any of these parameters can influence the nature of the deposit, can initiate a new bed, or can trigger a phase of erosion.

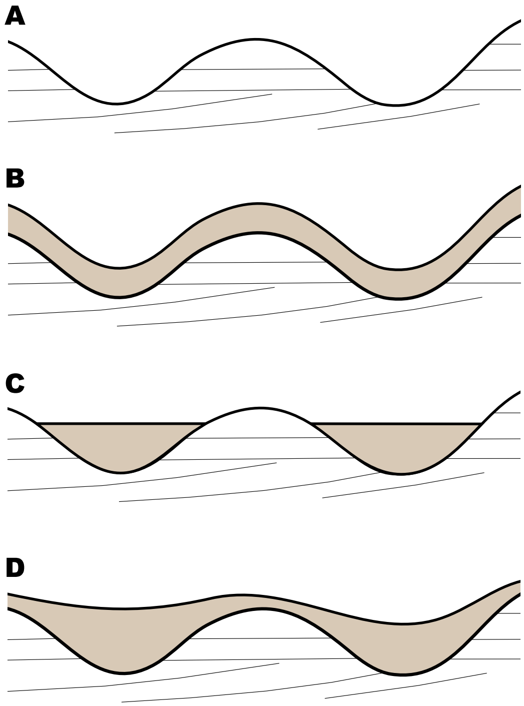


Figure 2.10 Variations in gross bed thicknesses produced by different types of volcanic processes. A) Pre-eruptive rock succession and topography. B) Pyroclastic airfall deposits mantle topography due to material falling vertically and evenly. C) Pyroclastic flow deposits infill the lower part of the topography due to the strong gravitational control on the flow. D) Pyroclastic surge deposits thicken into the topographic lows reflecting some gravitational influence. After Wright et al. (1980). Similar bedding geometries can be produced by other (non-volcaniclastic) types of sediment debris and gravity flows.

Explosive volcanic eruptions can throw a great deal of hot and possibly molten material into the atmosphere and can lead to a whole suite of complex processes that include ash falls and pyroclastic flows. The latter, like other density-driven flows, interact with the topography over which they flow so that zones of acceleration and deceleration determine the location of areas of erosion and deposition (e.g. Fig. 2.10).

In addition to transport by fluids, sediment grains can also be transported by ice in glaciers; other grains can move downslope solely under the influence of gravity, as in avalanches or rock falls.

Chemical processes

Much material, particularly in the sea and in some closed lakes (i.e. those with no outlet), is held in solution. In favourable conditions, brought about by changes in temperature, pressure of carbon dioxide, or concentration of ions, these solutes may be precipitated as minerals, either directly on the floor of the basin or as particles (crystals) in suspension. Such precipitates are susceptible to reworking by physical processes and to dissolution if the concentration of the solution is reduced by, for example, fresh-water influx.

Biological processes

Much of the calcium carbonate, which makes up limestones and present-day carbonate sediments, results from the activities of organisms (both animal and plant) which precipitate calcium carbonate as part of their metabolic processes. Other organisms secrete silica (e.g. sponges and diatoms) or phosphate (e.g. vertebrates), which may also contribute to sediments. Changes in the type of dominant organisms and their abundance may produce changes in the sediment and may lead to the generation of beds. A bed may be formed from the hard parts of organisms essentially remaining *in situ*, but skeletal material is most commonly redistributed by waves and currents before final deposition. Changes in organic activity commonly reflect changes in physical and/or chemical conditions. Some organisms such as corals and algae build large structures in their own right. In these mounds or reefs, the skeletons are essentially preserved where they lived with little or no physical reworking. As well as creating sedimentary material, both animals and plants can also destroy bedding and lamination and create secondary, post-depositional structures through burrowing and root disturbance.

Diagenetic processes

The final appearance of a bed in the stratigraphic column results not only from the conditions of its deposition and early disturbance but also from its history during subsequent burial. The processes of post-depositional change (**diagenesis**) vary with, for example, initial sediment composition, depth of burial (influencing pressure and temperature) and changing pore-water chemistry. As a result, some distinct "beds" may be due to diagenetic differences rather than to changing depositional processes.

In reality, many beds result from combinations of these various types of process. For example, biological activity may depend on chemical and physical conditions. Similarly, diagenetic processes may vary in different host beds, possibly reflecting original textural differences in the sediment related to the physical conditions of deposition. Study of bedding, therefore, should aim to understand the full assemblage of processes, both depositional and post-depositional, that were active in generating successive beds.

2.2.3 Vertical sequences, changes of process and bed generation

When we record a succession of beds and their contacts in a vertical sequence and understand that each bed records a change of process, we have the starting point for trying to deduce the depositional environment. Before attempting to use the vertical succession in this way, however, it is important to understand just how the vertical changes came about. A very simple view of sedimentation suggests that there are two main processes of change, which commonly act together but generally operate on different scales.

Changes due to lateral migration

If you observe an environment in which sediments are accumulating at the present day it is commonly apparent that different processes operate in different parts of the setting and give rise to different deposits. If it were feasible to map the distribution of these deposits over a long enough period of time, it might be possible to document gradual changes in the distribution pattern. An area that once accumulated sand, for example, might later be a site of mud deposition and, if net sediment accumulation prevails, we might expect to find a bed of mud overlying a bed of sand in a trench dug at the site. In other words, lateral migration of sub-environments under stable conditions can create changes (i.e. beds) in the vertical sequence. This idea, which is the essence of Walther's principle of the succession of facies (see §1.3), is discussed in more detail in Chapter 10.

Changes due to temporal fluctuations

In contrast to the steady-state, lateral migration mechanism outlined above, many beds reflect changes in time, due to processes active beyond the depositional setting. On a lake floor, for example, coarser silts and sands may reflect periods of high river discharge, whereas interbedded muds may be deposited during quieter periods. Similarly, on a shallow sea floor, grain sizes and sedimentary structures may fluctuate to record periods of stormy and calmer weather. In both these examples the environment is not changing, but rather the vertical sequence of sediments records the natural variability of process within that environment.

In some settings it is reasonable to distinguish *normal* deposits from *catastrophic* deposits. The classical example of such a setting is a deep-sea continental margin where, under normal conditions, fine-grained sediment accumulates steadily from suspension. Occasionally, however, this continuous process is punctuated by a catastrophic density current, carrying coarser sediment, generated on the continental slope or shelf edge. Such an event may deposit a substantial sheet of sand in a matter of hours but may not recur for hundreds or even thousands of years. In the resultant sedimentary succession, however, the sand beds may be more abundant and conspicuous than the normal interbedded muds. Short-lived, catastrophic events can, therefore, make a contribution to the rock record out of all proportion to their duration.

2.2.4 The importance of bedding planes

The importance of the boundaries between beds has been emphasized already. Some are gradational whilst others are sharp. Recording the nature of boundaries between beds is as important as recording the features of the beds themselves, if observations are to be used for environmental reconstruction. A gradational boundary may suggest lateral migration of processes under steady-state conditions, whereas sharp junctions between contrasting beds might suggest catastrophic events. Where large-scale relief due to channelling is observed, or where there is reason to suspect an unconformity, this may mark the beginning of a radically different pattern of sedimentation.

It is worth thinking of bedding in relation to time. It is possible to recognize sedimentary deposits that must have accumulated very quickly: vertical tree stumps covered and infilled by muds after they had partly rotted but before they had fallen; beds deposited from floods, dust storms and ash clouds. These beds often form as part of thick successions that are uniform at a larger scale of observation but which, when dated, appear to have accumulated over considerable periods of geological time, perhaps millions of years. This paradox of beds deposited relatively swiftly, making up successions representing many millions of years, has led to intuitive statements such as “98% of geological time must be represented by the bedding planes”, and highlights the importance of recognizing significant time gaps

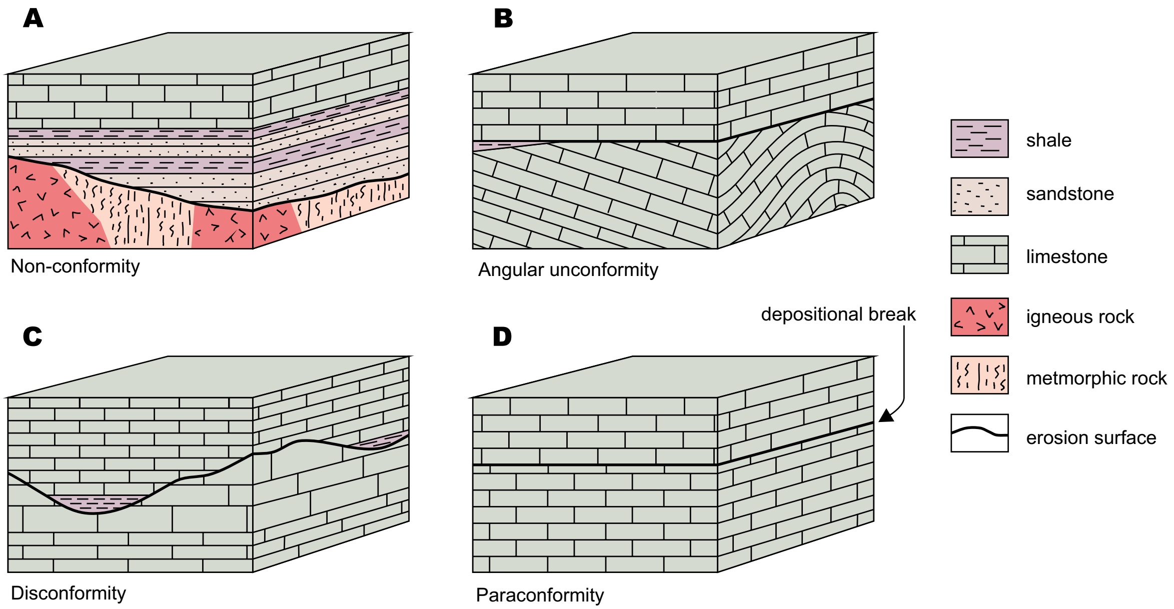


Figure 2.11 Four types of unconformity. A) non-conformity, B) angular unconformity, C) disconformity and D) paraconformity. Modified after Dunbar and Rogers (1957). Note that the terminology relating to discordant bedding relationships (e.g. onlap, apparent truncation) can be applied to beds above and below some unconformities.

in seemingly continuous sedimentary successions. The nature and distribution of certain trace fossils and of some early diagenetic concretions (palaeosols and hardgrounds) (see Ch. 9) may occasionally enable us to do this with some confidence.

2.2.5 Unconformities

An **unconformity** is a break in a stratigraphic succession resulting from a change in conditions that caused deposition to cease for a significant period of time. Various types of unconformity can be identified (Fig. 2.11). They are easiest to recognize and define where sedimentary rocks overlie igneous or metamorphic rocks (**non-conformity**) or where they rest upon previously folded and eroded strata (**angular unconformity**). In some cases, however, an angular unconformity may be wrongly suspected where flat bedding overlies a very large-scale set of cross bedding. In such cases, comparing the attitude of the beds underlying and overlying the cross-bed set should normally settle the issue. Time gaps are not easily defined, indeed they are often not recognized, without fossil evidence. Significant gaps can occur within a local succession that appears, at first sight, conformable where two sets of apparently conformable beds are distinguished only by a change of sediment type (**disconformity**), or even where the sediment types are virtually the same (**paraconformity**). In many cases, disconformity and paraconformity surfaces can only be satisfactorily identified where parts of a fossil sequence are proved to be absent. In other cases, such surfaces may be recognized by their association with intense burrowing, mature soils or with biogenic structures such as tree roots, which took significant time to develop during the period of non-deposition. Many successions, in which the preservation potential of fossils is not high, as in red beds, lack the means to be easily analysed in this way, although more elaborate methods such as magnetostratigraphy and chemostratigraphy may resolve this in some cases.

Some large erosional surfaces, for example the bases of channels cut by rivers on alluvial floodplains, or tidal-channels cut into mudflats, may be confused with true disconformities (i.e. one with a significant time gap). In the absence of fossils, a detailed sedimentological analysis may be necessary before the time significance of a surface becomes apparent. Hence, seemingly continuous successions may contain many time-gaps. Careful observation and recording of sedimentary features and relationships above and below any suspected "unconformity", and generation and evaluation of several working hypotheses are therefore important. Breaks in the record that represent time gaps so short that they are below the resolution of evolutionary changes in the fossils have been called **diastems** and it is sensible to regard potential disconformities and paraconformities as diastems until proved otherwise.

Study techniques

Field experience

Virtually any geological excursion will involve the observation and recognition of beds and bedding planes, and the measurement of their thickness and their attitude in three-dimensional space. The authors of field guides rarely suggest what to do about the study of bedding, and students may have to rely on general advice given in textbooks and on being taught by their tutors while in the field. Bed boundaries in coarse-grain successions are often characterized by subtle changes in clast composition, size, shape, orientation or packing; on initial examination in the field, such boundaries can commonly be difficult to discern (Fig. 2.4C). Bed boundaries in mixed sand–silt successions often exhibit distinct interlamination with bed boundaries characterized by either planar or irregular surfaces. Water-worn sections exposed in stream beds or along coastlines commonly provide excellent exposures for study of such beds of fine-grained lithology types.

Laboratory experience

Simple physical experiments can be devised to investigate bedding development and the form of bed boundaries. Thoroughly mix together 100–200g of loose sand with water in a jar, pour the mixture into a 1L measuring cylinder and allow to it settle. Repeat the process with a second sand/water mixture and add this to the contents of the measuring cylinder, taking care not to disturb the sediment added previously. Repeat the process with further sand/water mixtures, using sand with different grain size and sorting characteristics. Where adjacent sand layers are well sorted and characterized by similar mean grain sizes, then bed boundaries will be diffuse or difficult to recognize. Where adjacent sand layers are poorly sorted or characterized by different mean grain sizes, then bed boundaries will be readily identifiable. Try repeating the experiment with gravel samples that are characterized by the same overall mean grain size but whose clast shapes are markedly different.

Recommended references

Ager, D. V. 1981. *The nature of the stratigraphical record (2nd edn.)*. A concise and well-written book that considers the completeness (or otherwise) of the stratigraphic record.

Brenchley, P. J. & Williams B. J. P. (eds.) 1985. S*edimentology: recent developments and applied aspects*. A collection of review papers on many aspects of sedimentology and its applications.

Brookfield, M. E. 2003. *Principles of stratigraphy*. Oxford: Blackwell. A good discussion of the application of stratigraphical techniques in sedimentology.

Doyle, P., M. R Bennett &. A. N Baxter 2001. *The key to Earth history: an introduction to stratigraphy*. A well-illustrated introductory text that demonstrates the application of simple techniques in stratigraphy in the analysis of sedimentary successions.

Ingram, R. L. 1954. *Terminology for the thickness of stratification and parting units in sedimentary rocks*. A landmark paper from which much of present terminology flows.

McKee, E. D. & G. W. Weir 1953. T*erminology for stratification and cross stratification in sedimentary rocks*. The basis for much of the current terminology used in the classification of beds and bedding.

Tucker, M. E., 2011. *Sedimentary rocks in the field*. A practical guide. Considers how to describe, classify and interpret the environmental significance of bedding relationships.

CHAPTER 3

Basic properties of fluids, flows and sediment

3.1 Introduction

In order to understand the processes that produce many of the sedimentary structures observed in sedimentary deposits, it is necessary to have a basic understanding of the physical properties and mechanics of the fluids that erode, transport and deposit sediments. Most of these processes result directly from movement of a fluid, commonly water but also air and ice (which, although not a true fluid, does exhibit some similar behavioural properties). Exceptions are sediments emplaced by the direct action of gravity on loose particles and on sediment-water mixtures, usually moving on a slope. During gravity emplacement, water may be important as a lubricant or as an agent that acts to support the moving grains. In such cases, the moving mass of grains, with or without water, typically shows the behaviour of a plastic. The difference between fluidal and plastic behaviour is important and is explained later in this chapter.

It is also important to understand something of the physical properties of sedimentary particles themselves, both as individuals and as populations. The variation of size, shape and density found in natural sedimentary particles clearly influences their response to the flows that erode, move and deposit them.

This chapter, therefore, examines some of the properties of fluids and plastics, and shows how these properties influence the way in which they move. It also considers the physical properties of sediments and shows how particles and fluids interact during certain sedimentary processes.

This chapter may seem rather theoretical, but it mainly describes common phenomena. Many of the features can be illustrated by simple experiment and by experience of everyday events. Try wherever possible to develop a feel for the physical reality of the various processes described. We indicate where we think experiments might be helpful, and with a little imagination it may be possible to model features of fluids and flows other than those we suggest.

3.2 Properties of low-viscosity fluids and flows

3.2.1 Basic properties of fluids

The two simple fluids that account for the great majority of sediment movement on the surface of the Earth are water and air. Ice is also important in moving sediment because, when its behaviour is observed on a long timescale, it flows as a plastic. Additionally, mixtures of sediment and water, such as slurries and mudflows, move under gravity when on a slope and essentially show plastic deformation.

The media of water and air differ significantly in certain physical properties, in particular **density** and **viscosity**. The fluid density (ρ_f) determines the magnitude of forces such as shear stress, which act within the fluid and on the bed, particularly when the fluid moves down a slope under gravity. Density also determines the way in which waves are propagated through the fluid and controls the buoyant forces acting on sedimentary particles immersed in the fluid by influencing their **effective density** ($\rho_s - \rho_f$) (where ρ_s is the density of the solid particle). For example, quartz grains in water have an effective density of 1650 kg.m^{-3} compared with 2650 kg.m^{-3} in air, a difference that strongly influences the ability of the different fluids to move the grains.

The viscosity (μ) describes the ability of a fluid to flow. It is defined as the ratio of the shear stress (τ shearing force/unit area) to the rate of deformation (du/dy) sustained by that shear across the fluid:

$$\mu = \frac{\tau}{du/dy} \tag{3.1}$$

The viscosity of a fluid is not constant and its magnitude varies with temperature (compare, for example, hot and cold syrup).

At the simplest level, we can visualize flow by a model where a fluid is trapped between two parallel plates moving relative to one another. The fluid may then be envisaged as a stack of sheets parallel to the plates. These sheets move relative to one another at a uniform rate so that an initial straight line drawn perpendicular to the plates will deform

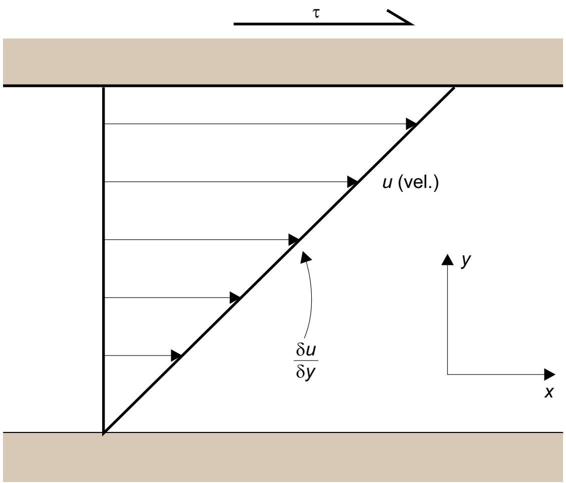


Figure 3.1 Definition diagram for viscosity. Two rigid parallel plates enclose the fluid. A shear stress (τ), acting parallel to 'sheets' within the flow sets up the steady-state velocity profile shown by the inclined line. The length of the arrows is proportional to velocity (*u*) relative to the lower plate.

into an inclined straight line, leaning in the direction of shear (Fig. 3.1). The viscosity reflects the force needed to produce a particular rate of deformation or sliding of the imaginary sheets. Increased viscosity demands a greater shear stress to produce the same rate of deformation.

As density and viscosity both play an important role in determining fluid behaviour, it is usual to combine them into a single term, the so-called kinematic viscosity (ν):

$$\nu = \frac{\mu}{\rho_f} \tag{3.2}$$

3.2.2 Laminar and turbulent flow

Some of the basic features of fluid flow can be investigated by means of a simple experiment. Inject a thin stream of dye into a very slowly moving flow of a viscous fluid, such as glycerine, in a narrow channel and carefully observe the form of the dye downstream of the injection point. Repeat the procedure at progressively increasing flow speeds, or with fluids of progressively lower viscosity. You will notice that, with low speeds and high viscosity, the dye persists as a fairly coherent and reasonably straight stream, whereas with increased velocity or decreased viscosity the stream breaks down and moves as a series of deforming masses within which there are components of movement perpendicular to the overall flow direction (Fig. 3.2).

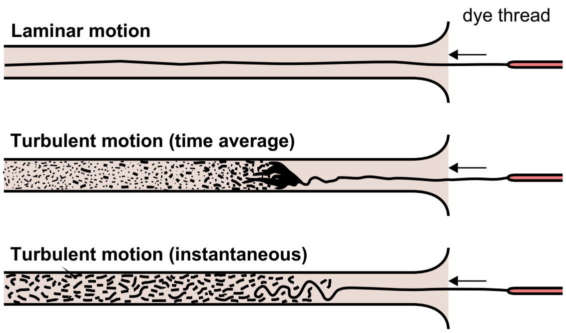


Figure 3.2 The Reynolds experiment to illustrate the difference between laminar and turbulent flow. Dye injected into the flow from a point source behaves in different ways depending on the velocity and viscosity of the flowing fluid. After Allen (1968).

With low velocity and high viscosity, the flow corresponds to the model outlined in §3.2.1 and the flow is said to be **laminar**. With more rapid flow or a lower fluid viscosity, the flow can no longer be visualized as a series of parallel sheets or filaments but clearly has some form of secondary motion superimposed upon the unidirectional flow. This motion is the very important phenomenon of **turbulence**.

3.2.3 Turbulence

An appreciation of turbulence is vital to understanding the origin and form of many of the sedimentary structures described later. The turbulence seen in the flow of water in a smooth-sided channel is an apparently random movement of parcels of fluid superimposed upon the overall flow. By slowing down the flow sufficiently or increasing the viscosity of the fluid, it is possible to eliminate this random motion and achieve conditions of laminar flow. However, in virtually all natural conditions involving air or water, turbulent flow is the norm (Fig. 3.2). Velocity measured at a point in a laminar flow is constant through time, whereas velocity at a point in turbulent flow will fluctuate, often widely, about a time-averaged value. This distinction between the two flow types suggests that it should be possible to use some combination of flow properties to predict the boundary conditions separating them. The factors that control the level of turbulence are usually combined to derive a **Reynolds number** (*Re*) for the flow. This is a dimensionless number that expresses the ratio between the inertial forces related to the scale and velocity of the flow (which will tend to promote turbulence) and the viscous forces (which tend to suppress turbulence):

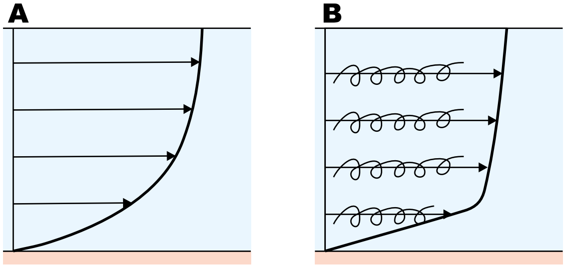


Figure 3.3 Vertical velocity profiles for A) laminar and B) turbulent flows in an open channel of great width. In the profile for turbulent flow, the velocities are time-averaged values and only close to the bed does a near-laminar pattern of movement occur.

$$Re = \frac{\overline{U} L \rho_f}{\mu} = \frac{\overline{U} L}{\nu} \tag{3.3}$$

where $\overline{U}$ is the mean velocity of the flow and L is some length which characterizes the scale of the flow (e.g. depth of flow, diameter of a pipe). Being dimensionless, the Reynolds number is useful when comparing different examples or designing scaled models. The transition from laminar to turbulent flow takes place at a critical value of the Reynolds number that will depend in each case upon the boundaries of the flow (e.g. the sides and bottom of a channel).

The existence of turbulence has important effects on flow properties. Because the generation and maintenance of eddies in a turbulent flow absorb energy, a greater shear stress is required to maintain a particular velocity gradient in turbulent flow than in laminar flow. Equation 3.1 has to be modified to account for the turbulence:

$$\tau = (\mu + \eta)\frac{d\bar{u}}{dy} \tag{3.4}$$

where η is the so-called **eddy viscosity**, an additional term which accounts for the extra shear needed to maintain turbulence and $\bar{u}$ is the time-averaged velocity. Eddy viscosity is not, however, a constant for the fluid but depends upon the level of turbulence in the flow; in other words it depends upon the Reynolds number. This makes calculation of shear stresses in a turbulent flow rather complex.

Another consequence of turbulence is that the velocity profile through a turbulent flow has a different shape from that through a laminar flow. Although the profile of laminar flow (Fig. 3.3A) is a realistic representation of the velocities at any instant, the profile for turbulent flow (Fig. 3.3B) is averaged over time to eliminate the fluctuations due to turbulence. For the same reason, the time-averaged velocity $\bar{u}$, rather than the instantaneous velocity u, is used in Equation 3.4. Instantaneous values of velocity in turbulent flows have components of direction and magnitude superimposed on the time-averaged velocity. The variations of velocity are commonly of the same order as the time-averaged value itself. This phenomenon accounts for the irregular buffeting experienced in trying to wade a rapidly flowing stream or when standing in a strong wind. Turbulence can commonly be seen on the water surface of rivers, particularly during floods: "boils" can be seen rising to the surface, particularly in subdued light and in rain, when reflections are reduced (Fig. 3.4A) and mixing zones of clear and turbid water can also be a good place to see the structure of eddies (Fig. 3.4B).

Where currents are strong enough to move sediments, flow will almost always be turbulent, and turbulence will influence the way in which the grains move. Turbulence is a crucial mechanism in the transport of sediment in suspension whereby upward components of the turbulent motion support the suspended grains (see §3.9.1). This implies that turbulence cannot be entirely random (statistically homogeneous) and that there must be a net upward energy flux. Some of this activity can be related to upward moving bodies of high-speed fluid bursting from the bed, and being generated from streaks of higher and lower velocity fluid close to the bed.

Such details of the structure of turbulent flows are occasionally important in understanding sedimentary structures but more important are the localized eddies, which are associated with the shape of the boundaries of the flow. Obstructions and irregularities fixed on the margins of flows generate eddies, the shape and organization of which closely relate to the shape of the obstruction and to the prevailing flow conditions. Sometimes a "captive" body of fluid rotates in the lee of an obstruction, whereas in other cases a spiral eddy is shed back into the main flow. In such cases, the flow is said to separate from the boundary at a **separation point** or **line** and to reattach itself downstream at a **reattachment point** or **line** (Fig. 3.5).

You can learn much about separated flows and the structure of eddies in water by simple experiments in laboratory channels or in small, natural, streams. The pattern of water movement can be seen from the movement of any small, suspended particles, but the best visualization method is to inject dye into the flow at selected points through a fine tube. A solution of potassium permanganate serves very well for this purpose. Place obstructions of different shapes

Figure 3.4 A) "Boils" on the upper surface of river water during high stage indicating the large scale general turbulence of the flow. River Trent, Staffordshire, England. B) Small-scale turbulent mixing between clear and turbid water at the junction of two streams. Note the discrete cells of water, which move at a high angle to the flow direction.

and sizes on the bed of the channel and carefully explore the local pattern of water movement. Try to determine the points or lines where flow separates from and reattaches to the bed. Try to determine the volumes and shapes of

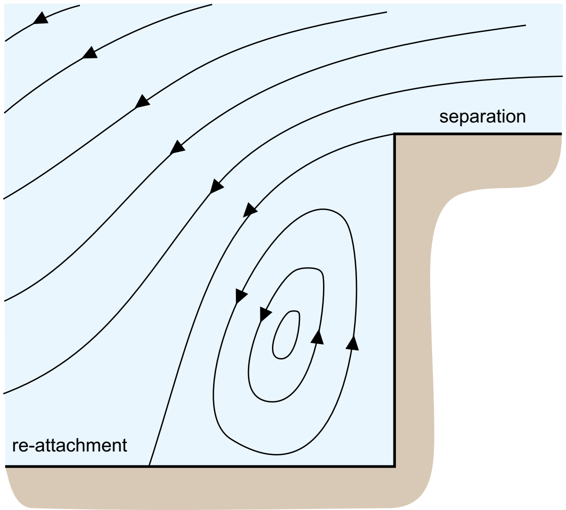


Figure 3.5 Separation and reattachment of flow at a negative step on the perimeter of a flow. A cell of rotating fluid is trapped within the separation eddy or 'bubble'. After Allen (1968).

eddies, and describe the patterns of water movement within them (compare your observations with Figures 3.6 and 3.7). If experimental facilities are not available, you can learn much from carefully watching the movement of water around bridge piers or large boulders in rivers, or the movement of snow, smoke or dry leaves on windy days. Try to develop a feel for the three-dimensional shape and organization of eddies in relation to the obstacles that create them. This will help you to understand how both erosional features and depositional bedforms seen on present-day sediment surfaces and in rocks at outcrop may have developed.

3.2.4 Bed roughness

Obstacles on the boundary of a flow generate eddies that influence the general level of turbulence. The larger and more abundant the obstacles, the more turbulence is generated and the more energy is absorbed, thus slowing the flow. This introduces the idea of **bed roughness**, which expresses the frictional effect that the boundary of the flow, for example a river bed, has on the flow. Roughness is made up of two components when the boundary consists of loose moveable grains (Fig. 3.8). The grains themselves constitute one component (**grain roughness**) and their frictional effect is a function of grain size. If the sediment is poorly sorted, however, large grains may be enveloped in finer material and their frictional effect reduced. Grain relief is the critical factor.

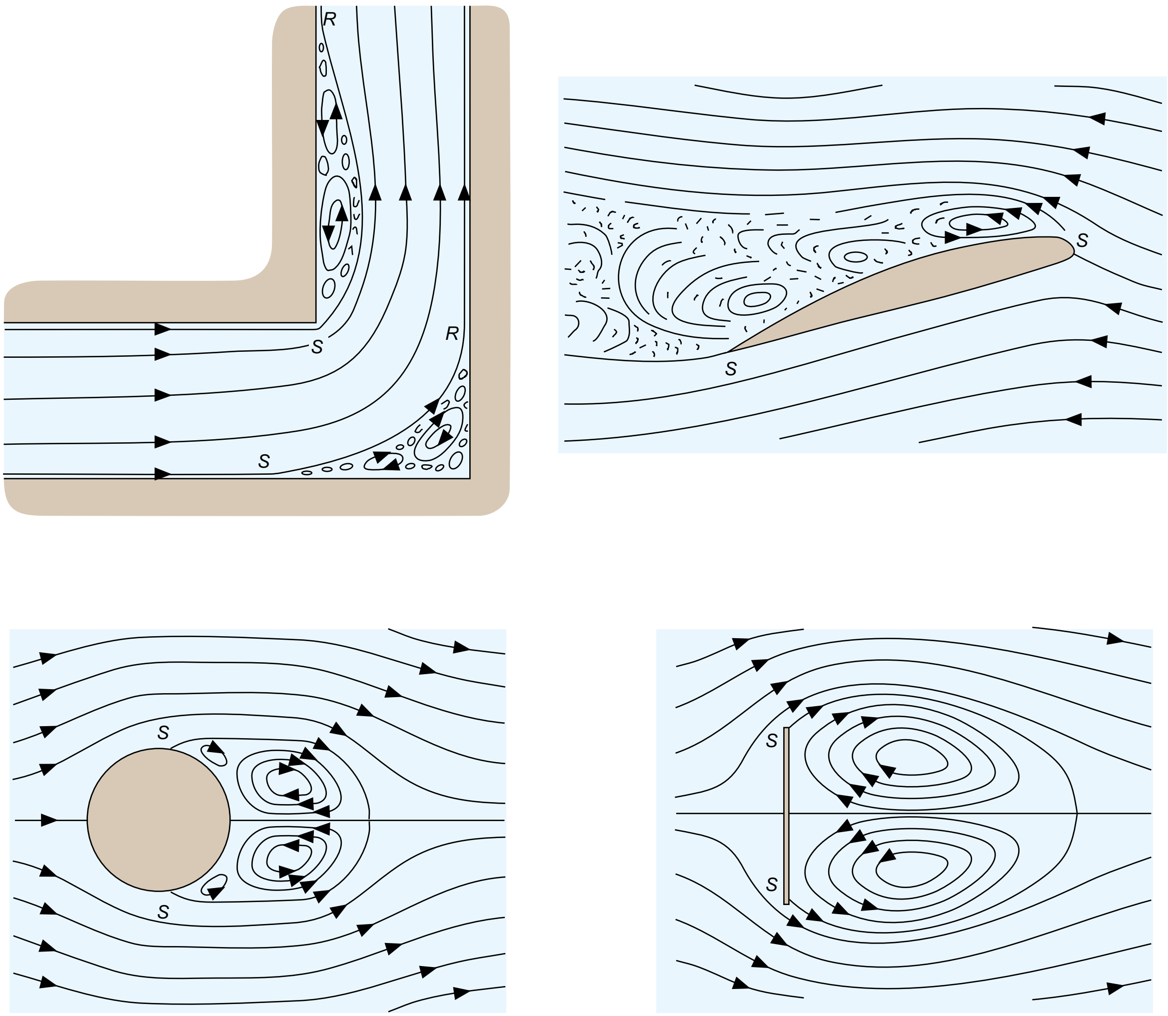


Figure 3.6 Examples of patterns of flow separation (*S*) and reattachment (*R*) around obstacles and features of different shapes. After Allen (1968).

The second roughness component is the bedforms into which the sediment may be moulded by the flow (**form roughness**). These bedforms depend very much upon the conditions of flow, which, in turn, depend to some extent on the bed roughness. The equilibrium established between the bedforms and the flow is, therefore, highly sensitive (see Ch. 6).

Where the relief at the boundary of a flow is very small, the roughness elements do not generate eddies and the bed is described as **smooth**. The critical relief for this condition to apply is determined by the prevailing flow conditions. Directly above the smooth bed and beneath the fully turbulent flow there is a thin layer within which the flow is much less turbulent, the **viscous sub-layer**, whose thickness depends upon the depth, velocity and viscosity of the total flow. If the bed relief exceeds this thickness, no viscous sub-layer can exist. The sub-layer is therefore only important with fine bed material. It is now known that the viscous sub-layer does not exhibit laminar flow, as had been thought previously; rather it is characterized by **streaks** of faster- and slower-moving fluid, aligned parallel to the flow. These periodically "burst" into the overriding turbulent flow. Streaks may be important in the initiation of grain movement and in the formation of ripples, in the production of some current lineation and in the development of lamination in fine-grained sediments deposited from suspension (see Chs. 5 & 6).

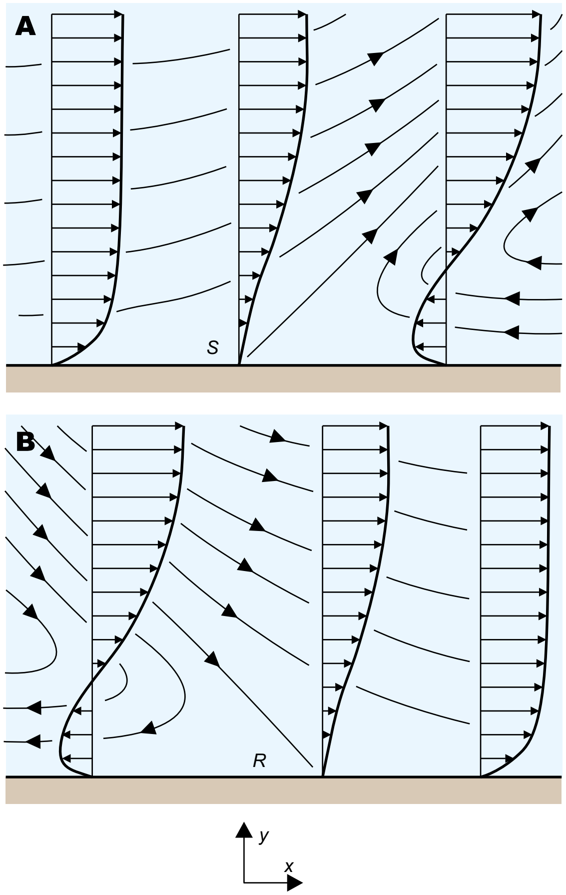


Figure 3.7 Flow velocity profiles associated with A) flow separation and B) flow reattachment. Modified after Allen (1982).

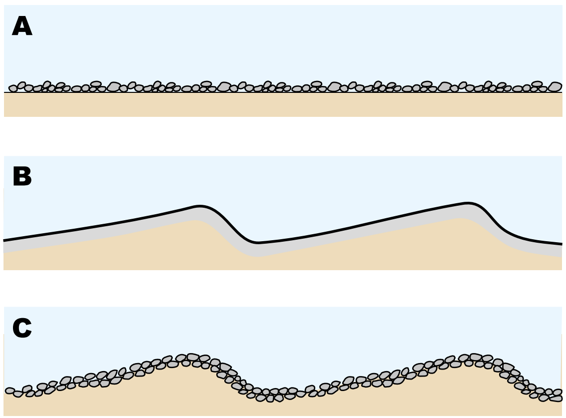


Figure 3.8 The distinction between grain roughness and form roughness. A) Purely grain roughness. B) Purely form roughness over a smooth artificial bedform. C) Grain and form roughness over a natural bedform.

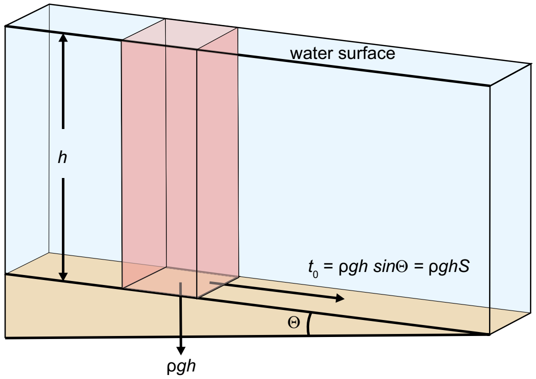


Figure 3.9 Definition diagram for the calculation of bed shear stress for water flowing downslope as an open channel flow. Forces acting on unit area of the bed are indicated. *g* is acceleration due to gravity. Other terms explained in main text.

3.2.5 Boundary shear stress

The behaviour of sediment on a bed below a flow is determined largely by the force that the flow is able to exert on the bed. The boundary shear stress (force/unit area parallel to the bed) is a function of depth (h), slope (S) and the nature of the fluid, and is indirectly a function of velocity of flow. Calculation of the boundary shear stress (τ_0) is complex. It depends upon the Reynolds number, the frictional characteristics of the bed and the shape of the velocity profile of the flow close to the bed.

A simple approximation of the boundary shear stress for a wide, open channel, where side effects are negligible, can be obtained from the idealized situation in Figure 3.9. For calculating the shear exerted by wind, this method is clearly inapplicable, since depth is indeterminate, as it also is in very deep water. Details of the shape of the velocity profile close to the bed are needed to estimate boundary shear stress more accurately.

3.2.6 The role of gravity: rapid and tranquil flows

In addition to the controls exerted by viscous and inertial forces on the character of a flow through their influence on turbulence, controls by gravitational forces are also important. In particular, gravity, being a body force acting on the fluid as a whole, influences the way in which the fluid transmits surface waves (see §3.3. for more detail). The speed at which a wave propagates in shallow water is given by the equation:

$$c = \sqrt{gh} \tag{3.5}$$

where c is the wave speed (celerity) and h is the water depth.

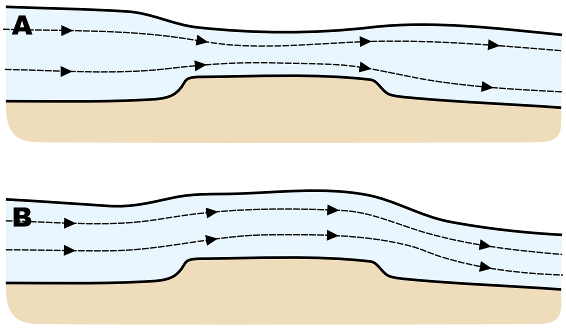


Figure 3.10 The form of the water surface and of time-averaged flow lines over a positive obstruction on the bed under conditions of A) tranquil and B) rapid flow.

Figure 3.11 Upstream facing breaking waves marking the position of hydraulic jumps in the flow regime. A) Jökulsá á Fjöllum, northern Iceland. B) Mýrdalssandur, southern Iceland.

It is clear that, for flowing water, there will be a velocity above which it will not be possible for waves to move upstream. This critical velocity separates two distinct types of flow: **tranquil** flow and **rapid** flow. The distinction is commonly drawn by reference to another dimensionless number, the **Froude number** (*Fr*), given by the ratio of inertial to gravitational forces in the flow:

$$Fr = \overline{U} / \sqrt{gh} \qquad (3.6)$$

For $Fr > 1$, we have **rapid flow** conditions in which waves cannot be propagated upstream and for $Fr < 1$, we have **tranquil flow** where this is possible.

The Froude number, and this distinction of flow types, applies only to liquids. In air, an analogy is provided by the Mach number and by subsonic and supersonic velocities, although in that case the wave motion involved is compressional and not gravitational.

In tranquil flow, the water surface is rather irregular as cells of turbulence move freely. In rapid flow, the water surface looks more glassy and the flow appears rather "streaked out" with turbulence somewhat suppressed. When these two types of flow encounter obstacles at their base, they react differently (Fig. 3.10).

Try to recognize which type of flow occurs in small streams, in rivers, in rainwater flow in gutters, or in laboratory channels. Rapid flow will be most likely where the gradient is high. It is quite common to see sharp transitions between the two states, when rapid flow passes down-stream into tranquil flow. The resulting breaking wave or **hydraulic jump** marks the sudden increase in depth and reduction of velocity (Fig. 3.11). A small-scale version of this phenomenon can be produced by directing a jet of water vertically downwards on to a flat, smooth, horizontal surface, as from a water tap on to the floor of a sink or bath. This distinction of flow type is independent of any sediment in the system. It is however related in a general way to the existence of upper and lower **flow regimes** defined by bedforms developed on a sand bed (see Ch. 6).

3.3 Waves

So far we have considered unidirectional currents in fluids but, as mentioned in the last section, water in particular also transfers energy through the movement of waves. You will likely have watched waves at the seashore or thrown stones into ponds and watched the concentric pattern of waves that result. Waves at their simplest involve localized vertical and horizontal movement of water without any net displacement of water taking place. The behaviour of waves in shallow water can be studied in a small laboratory wave tank. With simple waves, try to measure the length, height, speed and frequency of the waves, and the water depth. How do these properties relate to one another?

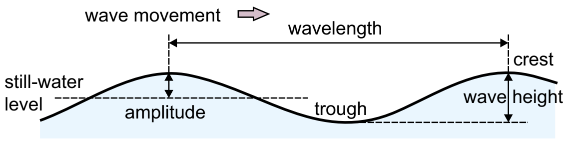


Figure 3.12 Definition diagram for the main physical properties of simple water surface waves.

The basic terminology used to describe water waves is shown in Figure 3.12. In addition, any wave is characterized by its **period** T, the time between the movement of successive wave crests past a point, and **wavelength** L, the distance apart of successive wave crests. From this it follows that:

$$c = L/T \tag{3.7}$$

However, it is important to know how wave speed c (propagation velocity or celerity) is controlled by other properties of the water body. Wave theory is mathematically complex and for our purposes it is sufficient to present two results. The theory recognizes two distinct types of waves which depend on the ratio of wave height to water depth. Shallow-water waves have lengths of at least twenty times the water depth, whereas deep-water waves have lengths of less than four times the water depth. In consequence, there are also intermediate forms. For the deep (d) and shallow (s) cases, wave speed is derived in the following ways:

$$c_s = \sqrt{gh} = 3.1h^{1/2} \tag{3.8}$$

$$c_d = \frac{gT}{2\pi} = 1.55T \tag{3.9}$$

Here c_s and c_d are in metres per second, g (acceleration due to gravity) is in metres per second squared, h (the water depth) is measured in metres, and T in seconds. For intermediate wave types, the relationships are rather more complex. The most obvious point about these relationships is that, for deep-water waves, the wave speed is independent of depth, indicating that wave behaviour is not influenced by the bed. For shallow waves, water depth is the prime control and the waves can be said to "feel" the bottom and react to it.

Associated with these differences in speed are differences in the pattern of water movement associated with the passage of a wave. Using beads, some of which just float and others of which just sink, it is possible to visualize the pattern of water movement as a wave passes a point.

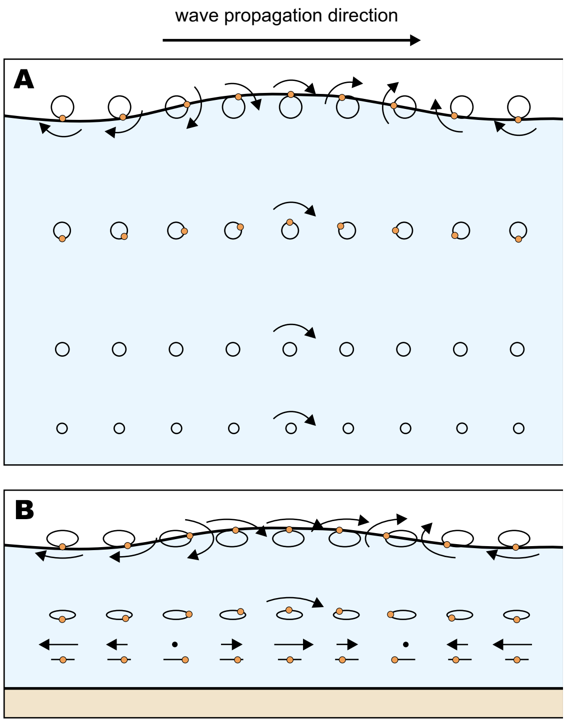


Figure 3.13 The pattern of movement of individual water particles associated with the passage of a surface wave for A) deep-water waves and B) shallow-water waves.

The beads have an elliptical pattern of movement near the water surface, but close to the bed the sinking beads may show a more or less linear pattern of to-and-fro movement. This so-called "orbital" motion characterizes all the molecules of water, but the nature of the orbital changes with depth and position in the water column. With deep-water waves, circular orbitals decrease in diameter with depth until movement dies out (Fig. 3.13). With shallow-water waves, the orbitals change their shape from nearly circular at the surface through elliptical to a linear (forwards and backwards) movement close to the bed. Such oscillatory motion is important in the movement of sediment and in the development of wave ripples (see §6.1.5).

As waves approach the shore, they become steeper and eventually become unstable and break. The style of breaking varies with the gradient of the beach. With steep beaches, waves **plunge** strongly close to the shore giving a short run-up (**swash**) and strong **backwash**. With more gentle slopes, breaking occurs further offshore and waves move as **surges**, often over long distances, and the strength of the backwash is less.

This account of wave motion is a simplification of what happens in most natural settings where several groups of waves of different length, and even different direction, may coexist. These may resolve into interference patterns in both the water movement and the sediment response.

3.4 Properties of sediments moved by flows

Grains that are moved by flows have their own physical properties and these influence their response to flows. The most important properties are size, shape and density. With respect to the grain size only a few points need emphasizing. If all sedimentary particles were spheres, cubes or some simple geometrical shape, then a dimension such as diameter or side length would be an appropriate measure of size. However, natural sedimentary particles are much more irregular and diverse. Grain size is usually measured by sieving through meshes of known spacing and, by using a series of graduated sieves, it is possible to derive, for any sediment, the percentage of grains (by weight) falling between any two mesh sizes.

The sieve mesh sizes are usually set on a $\log_2$ scale using 1mm as the starting point. Each sieve in the stack would thus have a mesh size half that of the overlying sieve, a typical range being 4mm to 0.068mm (1/16mm). The weights of each sieve fraction can be used to create histograms and other plots (e.g. cumulative curves) that allow the structure of the grain-size population to be visualized. Using a log scale leads to problems with reading off intermediate values from graphs as the grain-size scale is non-linear. To overcome this problem and also aid calculations of derivative properties such as median, sorting and skewness, the log scale is commonly transformed into a **phi (φ) scale** where

$$\varphi = -\log_2 \text{(grain diameter in mm)}. \tag{3.10}$$

Using this scale, 1mm has a value of 0φ, 2mm = −1φ, 4mm = −2φ, 0.5mm = 1φ, 0.25mm = 2φ etc. It is also possible to express mean, sorting, etc. in terms of the φ scale. A more complete account of the presentation and manipulation of grain-size data can be found in other texts on sedimentary petrography.

Sieving is a useful way of getting information on the grain-size structure of a sediment. In effect, it measures the smallest cross-sectional area of a particle and approximately records its intermediate axis. Although this can be helpful, what is really needed is a measure of size that reflects the behaviour of the particle in fluid flows. There is no reason why least cross-sectional area, or intermediate axis, should accurately reflect this behaviour, as these parameters can be identical for particles showing a whole range of different shapes.

A simple experiment shows how shape influences the behaviour of a grain in a fluid. Take two identical pieces of paper, screw up one of them into a ball and allow both to fall to the ground through the air. The unfolded one falls more slowly and with a pronounced side-to-side motion whereas the ball falls more or less directly, although both "particles" are of the same volume and mass. More realistic and controlled experiments along the same lines can also be made using real sedimentary particles. Take four glass cylinders (1000ml measuring cylinders are ideal) and fill two of them with water at different temperatures, one with glycerine and leave one empty (i.e. full of air). Using a selection of sands of varying size, shape and mineralogy, see how the speed with which grains fall through these various fluids is influenced by grain diameter, grain density, grain shape and fluid viscosity. The behaviour of any particle will be controlled by some combination of these variables. However, at present it is not possible to combine measurable physical parameters in such a way that they describe the hydrodynamic behaviour of a particle. Instead it is more productive and less time-consuming to investigate the hydrodynamic behaviour directly and to measure some value that directly reflects the combined effect of the physical parameters. The usual hydrodynamic parameter is the **fall velocity** of the particle; that is, the steady velocity with which it falls through a column of water at a fixed temperature (i.e. fixed viscosity), after an initial phase of acceleration.

We can illustrate the nature of this equilibrium by reference to a small spherical particle falling through a column of still water (Fig. 3.14). At the fall velocity V_0, the constant speed at which the sphere falls after its initial acceleration, the gravitational forces acting downwards on the sphere are balanced by the viscous drag that the fluid exerts on it, i.e.

$$\frac{4}{3}\pi\left(\frac{d}{2}\right)^3 g(\rho_s - \rho_l) = C_d\pi\left(\frac{d}{2}\right)^2 \rho_l \frac{V_0^2}{2} \tag{3.11}$$

and so

$$V_0^2 = \frac{4gd}{3C_d}\left(\frac{\rho_s - \rho_l}{\rho_l}\right) \tag{3.12}$$

Where ρ_s and ρ_l are the density of the solid and liquid, respectively, and C_d is a drag coefficient which depends upon particle shape and particle Reynolds number

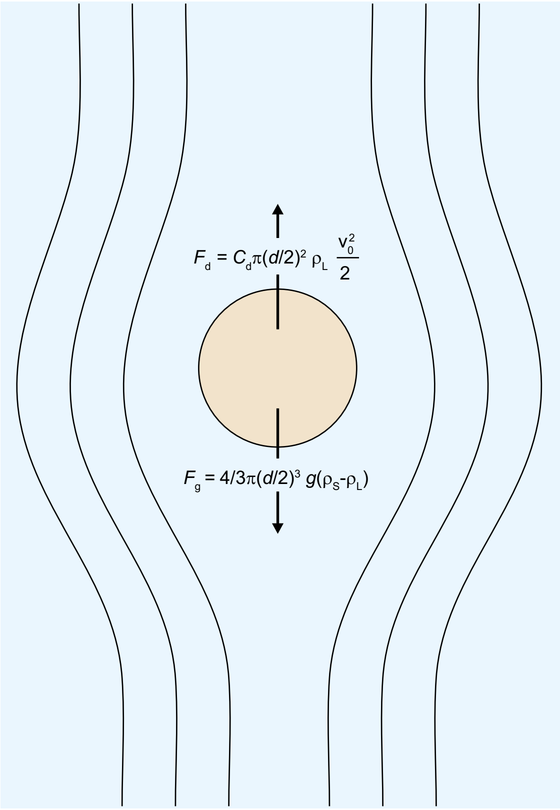


Figure 3.14 The forces acting on a spherical particle falling through a fluid. The streamlines indicate that flow around the particle is laminar and therefore either the viscosity of the fluid is high or the particle is small.

Re ($= V_0 d/\nu$) where d is particle diameter. For low particle concentrations and low Reynolds numbers:

$$C_d = \frac{24}{Re} = \frac{24}{V_0 d/\nu} \tag{3.13}$$

Thus we can write

$$V_0 = \frac{1}{18}\frac{(\rho_s - \rho_l)gd^2}{\mu} \tag{3.14}$$

which is the **Stokes' law of settling**. In other words, the fall velocity is proportional to the square of the grain diameter, and is therefore a reflection of grain size.

However, the Stokes' law of settling only applies for particles of small grain size (low Reynolds number) where laminar flow around the particle can be assumed. For larger particles, turbulence is generated and the equations must be modified. In general terms, the velocity is proportional to reducing powers of grain diameter with increasing turbulence. Also Equations 3.11–14 only apply to isolated or highly dispersed grains. With high grain concentrations there is interference between the falling grains, giving slower settling rates (so-called **hindered settling**). However, in spite of these drawbacks, the idea of using fall velocity as a way of describing grain size is useful and goes some way towards resolving the otherwise intractable relationships between size, shape, density and hydraulic behaviour.

It is a relatively simple experimental procedure to measure the fall velocity of any particle, whatever its shape, and from this measured velocity to calculate the diameter of the sphere that would fall at the same speed as that particle. If we standardize the density of the sphere (ρ_s) to that of quartz, it is possible to express the effective size of any grain, of whatever shape or density, in terms of the diameter of the equivalent quartz sphere (**hydraulic equivalence**).

In terms of sediment response to flow, it is usually the nature of the grains in bulk that is important, rather than properties of individual grains. The various statistical measures used to characterize grain-size populations (e.g. median, mean, sorting, skewness, etc.) have been reviewed at length in many standard texts on sedimentary petrography. These measures can often be useful in describing a sediment, but despite much effort, as yet, they provide only limited help in interpreting processes and environments of deposition. Work on cumulative grain-size distribution curves suggests that it is possible to recognize subpopulations within natural sediment and that these may correspond to different modes of transport such as suspension, intermittently suspended load and traction bedload (Fig. 3.15). However, other factors, such as grains of different compositions and with different shapes and densities, introduce significant additional complexity. As such, it is commonly not possible to discern sedimentary deposits that reflect particular processes or environments from analysis of grain-size distribution alone.

3.5 Erosion

The behaviour of grains when subjected to shear by a current is important for understanding sedimentary structures. At some point, with increasing boundary shear stress, grains on a bed begin to move, and **erosion** is then said to be taking place. It is important to understand in some detail the conditions under which grains actually begin to move.

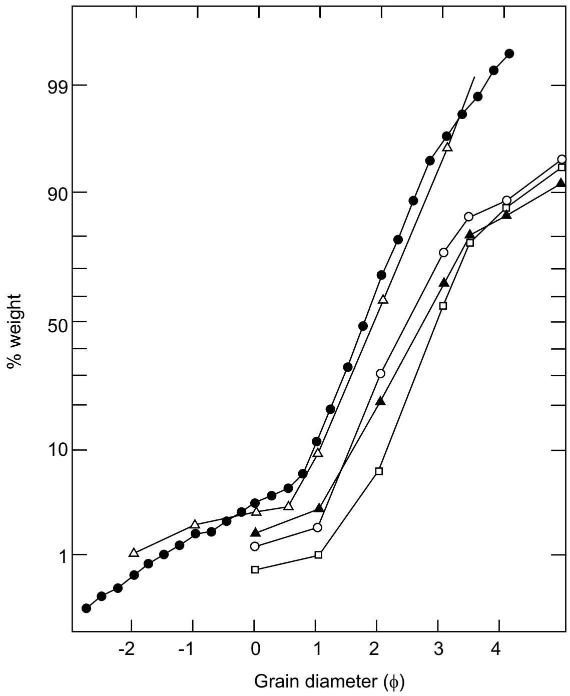


Figure 3.15 Cumulative grain size curves plotted on a probability scale show straight line segments likely to be due to the suspended sediment fraction, and the intermittently suspended bedload and traction load sediment fraction. Symbols relate to particular samples of sieved sand. Modified after Middleton (1976).

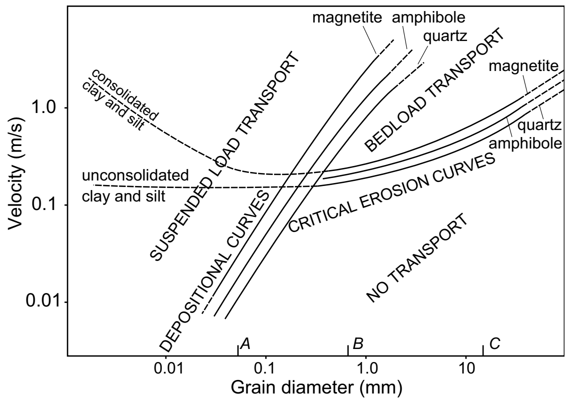


Figure 3.16 Hjulström-Sundborg plot of grain size against flow velocity 1m above the bed for initiation of movement of particles of different densities. The nature of the transport mechanism, once erosion has taken place, is also shown (after Sundborg, 1956 and Ljunggren and Sundborg, 1968). As an example, quartz grains of grain size B would begin to move as bedload at the critical erosion velocity of 0.12 m.s^{-1} and would go into suspension at a velocity of 1 m.s^{-1}. On deceleration, the particles would settle from suspension at a similar velocity. Try to predict the threshold conditions for changes of behaviour of particles of grain sizes A and C for similar cycles of acceleration and deceleration.

Consider the Hjulström-Sundborg curve, which plots grain size against critical erosion velocity (Fig. 3.16). From this diagram one can read off the velocity at which grains, of a given composition and size, begin to move if the flow velocity above the bed is gradually increased. However, we should not only try to understand the gross relationship between grain size and erosion threshold velocity but also to consider what is going on around individual grains when they are set in motion.

If we look at a rather idealized situation it is possible to isolate some of the factors involved. Consider a roughly equidimensional particle resting in a surface made up of similar particles (Fig. 3.17). When a fluid moves over this surface, four types of force act upon the particle: (a) the weight of the particle, (b) frictional forces between adjacent particles, (c) hydraulic lift forces, and (d) the tangential shear couple (drag).

Types (a) and (b) are forces that resist motion, whereas (c) and (d) are forces that encourage motion. To consider each of these in turn:

(a) The weight of the particle acting vertically downwards (F_g) will act as a moment trying to rotate the grain about the contact point with its underlying downstream neighbour or neighbours.

(b) Frictional forces between particles resist sliding motion and relate to the roughness of the particle surfaces and to the electrochemical forces between particles. The latter are only important with very small particle sizes (clay).

(c) Hydraulic lift forces, F_l, result from the flow accelerating over the upwards protruding grain on the bed. Low pressure above the particle and hydrostatic pressure below combine to give an upward-directed force, as on an aerofoil.

(d) The boundary shear stress (τ_0) of the flow acts on the exposed particle and can be envisaged as a horizontal force through the centre of the particle. This promotes rotation of the particle about the point of contact with the adjacent downstream particle. The force on the particle (F_d) depends both upon the boundary shear stress and upon the degree of exposure of the grain to the shear stress:

$$F_d = \frac{\tau_0}{N} \tag{3.15}$$

where N is the number of exposed grains per unit area.

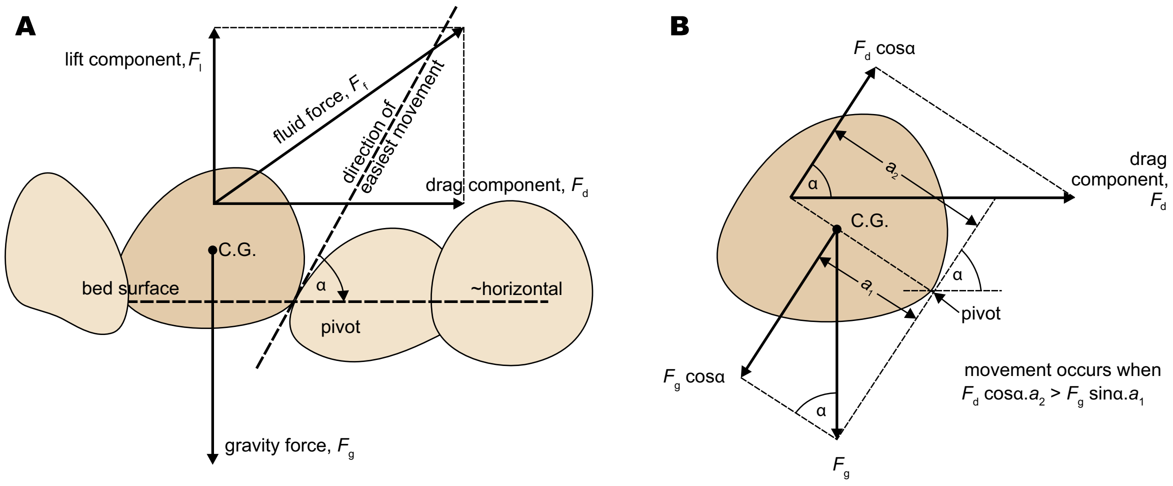


Figure 3.17 A) The forces acting on a particle in a bed of similar particles when subjected to the shear of an over-riding current. B) The movement forces which must be balanced for incipient motion; movement begins when $F_d\, a_2 \cos\alpha > F_g\, a_1 \sin\alpha$. C.G. is centre of gravity. After Middleton and Southard (1977, 1984).

If, as a gross simplification, frictional forces and hydraulic lift forces are ignored, then for grain movement to occur:

$$F_d\, a_2 \cos\alpha \geq F_g\, a_1 \sin\alpha \qquad (3.16)$$

In natural settings many other factors complicate matters. First, natural grains are not all equidimensional, nor are they of uniform size. Irregularly shaped grains have a stability that varies with their orientation with respect to the current and with the type of packing of grains on the bed. Naturally deposited grains tend to rest where drag and lift forces are at a minimum. Flattened grains are most stable when they are inclined upstream at angles in the range 10–20°, as seen in the phenomenon of **imbrication** of flat pebbles on beaches and river beds (see §7.4.4). Elongated grains are most stable if their long axes are parallel to the current; when aligned transverse to the current, they roll relatively easily. Measured experimental values of critical boundary shear stress tend therefore to diverge somewhat from the values predicted by simple models of erosion.

The packing of grains typically relates to the overall grain-size distribution. Well sorted sediments show a pattern of behaviour more closely related to the empirical relationships between grain size and velocity than do poorly sorted ones, where large grains are only partly exposed to the prevailing shear stress because of partial burial in finer-grained components and fine grains are protected by neighbouring larger grains. Simple relationships between grain size and critical boundary shear stress or critical erosion velocity do not apply over a full range of grain sizes as shown by the Hjulström-Sundborg diagram (Fig. 3.16). For grains over about 0.6mm diameter there is a gradual increase in critical shear stress and velocity, whereas below that grain size these parameters tend to increase with diminishing grain size. This rather unexpected result has been attributed to the increasing importance of intergranular forces in fine sediments, especially those that have been allowed to settle and compact under gravity for considerable periods. With decreasing grain size, the ratio of surface area to volume increases, particularly as many small particles have plate-like shapes. As a result, the surface forces become proportionally greater and grains show **cohesive** behaviour. In addition to raising the critical shear stress, cohesion also enables muds and silty sediments to remain stable on high-angle slopes, in some cases even going beyond the vertical. Commonly, this leads to them being eroded as larger aggregates and blocks (Fig. 3.18).

The protrusion of grains into the flow, giving lift forces and controlling the presence or absence of a viscous sublayer (see §3.2.4) will also help to determine just when a grain moves. An added complication is that the natural turbulence of any flow causes fluctuations in the instantaneous values of boundary shear stress. Some eddies may cause boundary shear stress and hydraulic lift forces to be temporarily and briefly large enough to entrain a particle which would not move under time-average conditions. Effects of

Figure 3.18 Example of cohesive behaviour: mud clasts in a sandy channel fill. The coherent nature of the mudclasts indicates their cohesive strength. Holocene, Finnmark, Norway.

this type make it difficult to satisfactorily define the onset of grain movement.

The streaks of high-velocity flow in the viscous sublayer may also localize the initial grain movement in fine-grained sediment and help to throw fine sediment into suspension as the streaks erupt into the overlying turbulent boundary layer. Wave action, when coexistent with a current, will give pulses of increased and diminished boundary shear stress and these may lead more readily to movement than would a steady unidirectional flow.

In natural settings, cohesive sediment in the size range of mud and silt may diverge considerably from the behaviour suggested by the Hjulström-Sundborg curves by virtue of erosion commonly taking place by removal of blocks or aggregates of grains (mud and silt intra-clasts) rather than of individual particles (Fig. 3.18). In addition, material already moving in the flow may help to promote or accelerate erosion by means of a "sand blast" (abrasion) effect when moving grains hit the bed.

The onset of grain movement due to wind is broadly similar to that with water. However, the occurrence of particles already in motion lowers the critical erosion velocity quite significantly. A sand bed that is stable in winds of a sub-critical speed can be set in motion if a few grains are thrown (seeded) on to the bed. The grain impacts trigger new grain movements and set off a chain reaction of movement down wind. The movement quickly ceases when seeding ends. There are therefore two critical shear stresses for wind erosion, an **impact threshold** where seeding is essential and a higher **grain threshold** above which movement takes place without any seeding of bed movement.

3.6 Modes of sediment transport

Having reviewed some of the factors involved in initiating grain movement, it is now necessary to outline the various ways in which movement of particles is sustained. These fall into two main groups: **suspension** and **bedload transport**.

3.6.1 Suspension

Sediment carried in a fluid without coming into contact with the bed is supported by fluid turbulence. The sediment moves at roughly the same rate as the fluid and the movement results from a balance between downward gravitational forces on the grains and upward forces derived from the fluid turbulence. This suggests that the turbulent flow has a net upward energy flux. In theory, any grain size of material can be carried in suspension if currents are strong enough, but in most natural situations it is usually the finer material of silt and mud grade that predominantly moves in this way. Indeed, below the grain size of about fine silt, grains, when eroded, go directly into suspension without an intermediate phase of bedload movement (Fig. 3.16). As the level of turbulence increases, the suspended-sediment carrying capacity and competence (maximum potentially transportable grain size) of the flow increase.

Increased load increases the viscosity and density of the flow so that larger grains can then be moved in suspension more readily. This process is, however, self-limiting as increased sediment concentration has a damping effect upon the turbulence due to the increase in viscosity until conditions approach those of a mudflow (see §3.7.1).

Sediment particles composed of low-density material can be carried in suspension in flows where the viscosity and density of the flow are relatively low. One such example is pebble-sized clasts of low-density pumice carried in suspension by stream flows.

3.6.2 Bedload transport

The movement of grains in continuous or intermittent contact with the bed may be by saltation, reptation, rolling or creep. **Saltation** describes the jumping and bouncing motion of grains close to the bed during vigorous bedload movement. Grains follow asymmetrical trajectories, which are commonly complicated in water by random, turbulence-induced fluctuations. There are gradations between true saltation and suspension as turbulence becomes more vigorous. As descending grains hit the bed, they may bounce back into the flow, dislodge stationary grains on the bed and help to set them in motion, or simply have their kinetic energy dispersed into the bed. In air, collisions on impact are more vigorous than in water because of the lower viscosity and higher effective density of the grains. **Reptation** occurs where grains that are too large to undergo saltation are temporarily lifted into the flow, usually as a consequence of incoming grain impacts, and hop short distances downstream before returning to rest on the bed. Mobilization of grains resting on the bed is then an important process. In water, a general damping of impacts takes place and hydraulic lift forces are probably more important in initiating grain movement.

Where a grain collides with the bed and does not bounce, its kinetic energy may be dispersed amongst several grains resting on the bed. As a result, some of these may be pushed a short distance down current or down wind. This is the phenomenon of **creep** and it can account for up to 25% of total bedload movement during wind transport. **Rolling** occurs when rather large or elongate clasts are set in motion. It will be favoured if a larger grain is moving over a relatively flat surface of smaller grains. There will be a much greater chance of a grain coming to rest if it is surrounded by, and it rests upon, grains of a similar size to itself. Rolling is additionally influenced by grain shape such that rounded equant-shaped clasts will roll more easily than flat disc-shaped clasts. All the modes of bedload transport can coexist to a greater or lesser extent. They will usually be associated with the development on the sediment surface of **bedforms**, which commonly occur as repetitive patterns on the bed at a variety of scales. When fully developed, they reflect an equilibrium between the strength of flow and the frictional drag of the sediment surface. These important sedimentary structures are described in Chapters 6 and 7.

Under very powerful currents, a thin layer close to the bed may have rapid grain transport with high grain concentrations to the extent that inter-granular collisions are a dominant process. Such layers have been termed **traction carpets** or **modified grain flows** and they can be important beneath flows carrying much sand-size material in suspension.

3.6.3 Accelerating and decelerating flows

In order for sediment to be eroded or deposited over significant periods of time it is generally necessary for the flow responsible to be either accelerating (erosion) or decelerating (deposition). Accelerating flows are sometimes referred to as being **accumulative** in that such flows tend to be underloaded with sediment and are therefore able to accumulate more material into motion. Decelerating flows, by contrast, are sometimes referred to as **depletive**, tending to be overloaded and therefore being depleted of their load. Flows that are neither accelerating nor decelerating should normally lead to a stable load with no net erosion or deposition.

Both acceleration and deceleration can take place in both time and space. Where changes take place through time, a current changes velocity, at a fixed place. Such currents are termed "**unsteady**" and a typical flow surge, such as a turbidity current, may have an early phase of acceleration (erosional) followed by a later phase of deceleration (depositional), commonly separated by a phase of relatively steady flow. Where changes in velocity take place in space, the flow may be steady at any fixed point, but changes take place upstream or downstream of that point, usually in response to the topography over or through which the flow is passing. Such flows are referred to as "**non-uniform**". For example, a steady flow that is moving down an increasingly steep slope will accelerate, whereas one that is encountering lower gradients at the foot of a slope will experience deceleration. Similarly, a steady flow that is being increasingly constricted laterally, perhaps as it moves into a valley or channel will experience acceleration whereas one that is expanding laterally, as at the downstream mouth of a channel, will decelerate and tend to deposit some of its sediment load (Fig. 3.19).

The situation is complicated by the fact that steady/unsteady and uniform/non-uniform distinctions are not mutually exclusive and so any one flow may be changing velocity (and hence sediment capacity) in both time (as a surge) and space (over complex topography), making prediction of erosion and deposition rather more complex (Fig. 3.19).

3.7 Sediment gravity flows

Mixtures of sediment and water, under appropriate conditions, move down slope as mass flows driven by gravity. In order to behave in a mobile fashion, the constituent sedimentary particles must be able to move relative to both one another and the inter-particle fluid, either water or air. Such particle separation or support involves a spectrum of processes, within which it is possible to isolate several theoretically distinct mechanisms. However, in reality, many mass flows have support mechanisms that are mixtures of these processes and the balance between them may change as a flow evolves. Processes that may be active to a greater or lesser degree include intergranular collision, intergranular friction, fluid turbulence, viscous shear and interaction

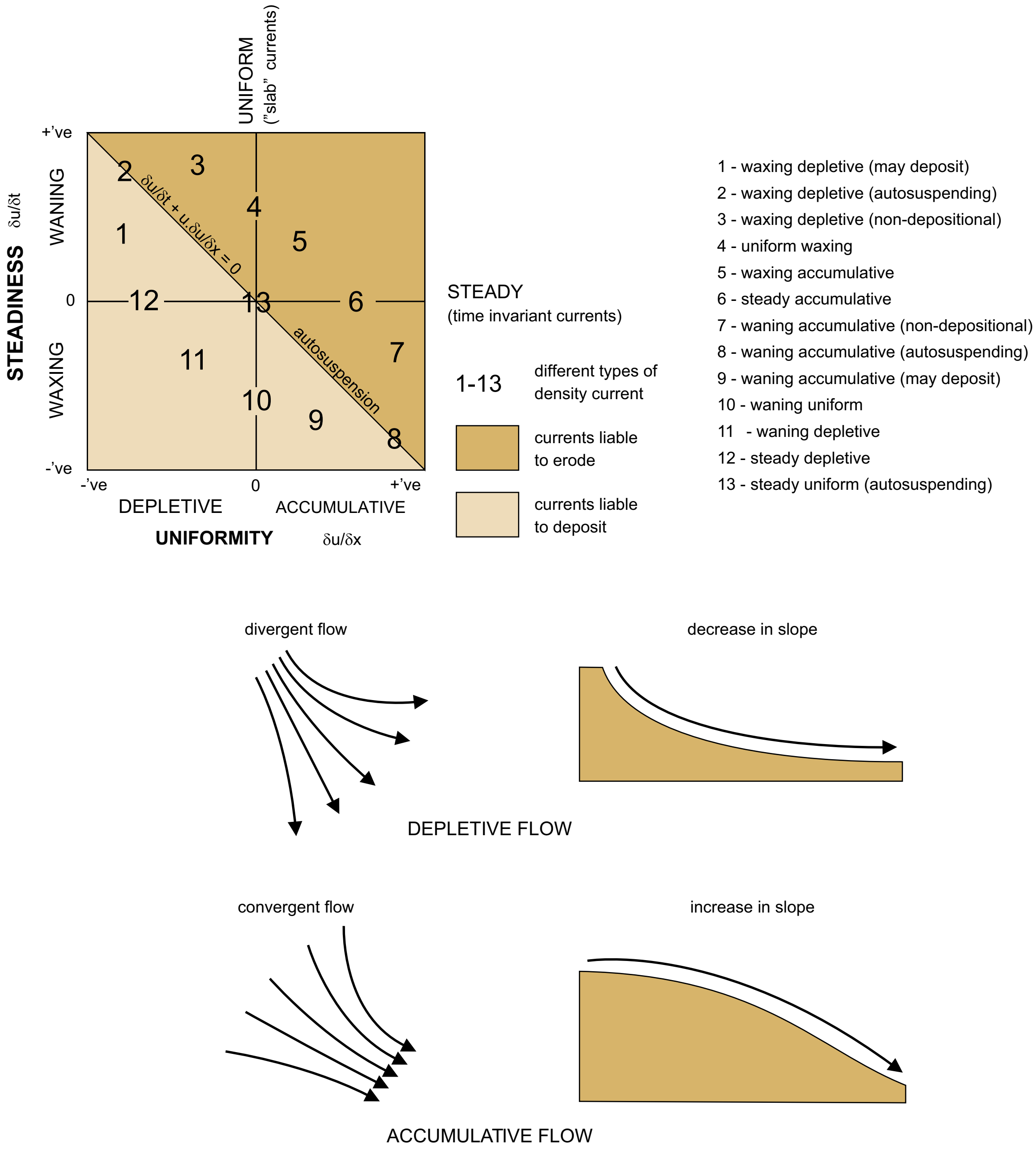


Figure 3.19 Thirteen different types of turbidity current classified according to their steadiness and uniformity in one spatial dimension (down current). The popular (lock-gate) model of a turbidity current is only one of the possible 13 types and lies in the bottom left-hand corner because it is strongly depletive and extremely unsteady. Many natural currents are hybrids and migrate from one field to another. Sketches of simple scenarios for depletive and accumulative flows. Modified after Kneller and Branney (1995).

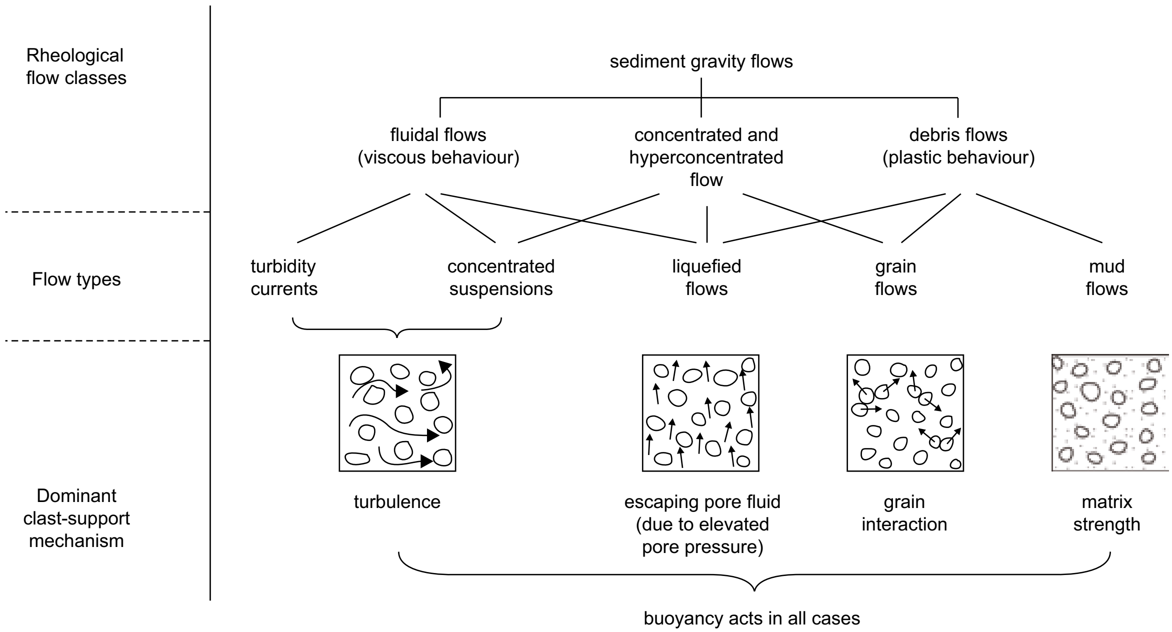


Figure 3.20 Various types of sediment gravity flow where different types of interaction between the water and the sedimentary particles create the mobility necessary for movement. Modified after Middleton & Hampton (1973).

between fluid and grains (Fig. 3.20). The complexity of the flow mechanisms and the problems of interpreting ancient deposits in terms of those processes have led to a bewildering and often contradictory terminology for both the flow types and their inferred deposits.

Here, in the first instance, we discuss four main grain-support mechanisms and then investigate the ways in which these change in relative importance across the spectrum of sediment gravity flows.

3.7.1 Particle support mechanisms

Matrix support

Where sediment-water mixtures have a significant content of clay, this imparts cohesive strength to the whole mixture and allows it to behave as a plastic. The high viscosity of the clay-rich matrix supports larger clasts in the mixture and allows the mixture to deform internally by shearing as the flow moves.

Mixtures of sediment and water that are dominated by viscous (cohesive) be processes have a yield (or shear) strength that must be first overcome for movement to begin. As such, they are plastics, in contrast with fluids, which have no yield strength and flow under minimal shear (Fig. 3.21). Movement of cohesive mixtures of sediment and water is usually initiated by the reduction of yield strength by addition of water to a mass of sediment through, for example, heavy rainfall. Conversely, as a flow decelerates, for example, on encountering a lower gradient or by gradually de-watering, there will be a point at which gravity-driven body forces that provide the applied shear no longer exceed the viscous strength. The flow will then stop abruptly, “freezing” the internal fabric of the flow, including that of the sediment contained therein.

Grain-to-grain support

Where muddy matrix is largely absent, larger particles may only move as a highly concentrated flow if they are kept apart and supported by inter-granular collision. Such non-cohesive flow requires an applied shear that can overcome the initial inter-granular friction. Pouring granulated sugar or tipping dry sand from a bucket or the back of a lorry are everyday examples of this behaviour. The particles are kept apart and hence free to move relative to one another by vigorous inter-particle collisions, which create a so-called **dispersive pressure**, whose magnitude will depend on the strength of the shear driving the flow. Once the applied shear falls below some critical value, the layer of dispersed grains will collapse and “freeze” as intergranular friction is re-established.

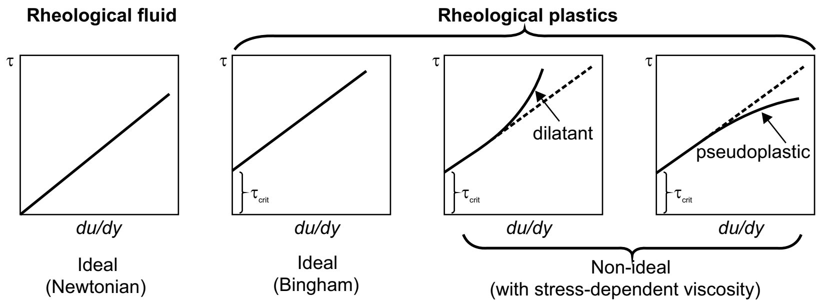


Figure 3.21 Graphs to show the relationship between applied shear stress (τ) and strain rate (du/dy) for flows with different rheological properties. τ_{crit} is the critical shear strength that must be overcome before deformation or movement can take place. After Nemec and Steel (1984).

Fluid turbulence

The support of sedimentary particles in suspension by the upwards components of fluid turbulence has been discussed in §3.2.3. Particles supported in this way constitute a suspended sediment load. Pure suspension operates at low sediment concentrations. As concentration increases, inter-granular collisions become more important, hindered settling occurs and Stokes' law ceases to apply strictly.

Buoyancy

In all types of flow, individual particles are present in a matrix made up of both other particles and fluid. In the case of high-concentration flows, a particle will "see" its surrounding sediment-fluid mixture as a matrix providing buoyant uplift and thereby reducing the effective density of the particle. For low-concentration aqueous flows, water will provide the main buoyant uplift whereas in a high-concentration flow the higher density of the matrix will further reduce the effective weight, allowing large particles to be rafted along within the flow.

3.7.2 Types of sediment gravity flow

The particle support mechanisms outlined above help to characterize differences within the spectrum of sediment gravity flows illustrated by Figure 3.22. On this basis, it is possible to deal with the spectrum of flow types under three broad headings though it should always be borne in mind both that intermediate types exist and that individual flow events typically evolve and transform in both space and time.

Debris flows

Flows with a high content of clay, and where matrix viscosity is the main particle support mechanism, are generally referred to as debris flows or mudflows. In some accounts, they are called "cohesive debris flows" to distinguish them from "non-cohesive debris flows", which are the hyperconcentrated flows of this chapter. Debris flows occur in both subaerial and subaqueous settings. In subaerial settings they are commonly initiated on steep slopes as a result of high water saturation following heavy rain or snow melt. In sub-aqueous settings, they may be triggered by shock or by progressively increasing sediment accumulation on a slope. Debris flows move as a result of viscous, near-laminar shear either dispersed through the flow or concentrated in a basal shearing layer (décollement). In the latter case, the upper part of the flow may move along as a non-deforming rigid plug. The high density of the deforming matrix means that buoyant uplift is very important in debris flows with the result that very large clasts can be rafted within the flow and even protrude from the top surface (Fig. 3.23B).

A general equation for describing the rate of shear strain in such deforming layers can be written as:

$$\frac{du}{dy} = \frac{1}{\mu}\left(\tau - \tau_{\mathrm{crit}}\right)^{k} \tag{3.17}$$

where μ is the apparent viscosity of the mixture, τ is the shear stress, and τ_{crit} is the critical shear stress. Yield strength or plastic limit (k) is a coefficient that describes the stress–strain relationship during deformation. For most debris flows, k is close to 1, approximating ideal (Bingham) plastic behaviour, but deviations from this value occur and their effects are illustrated in Figure 3.21.

In subaerial settings, debris flows may become diluted by the addition of more surface run-off to the extent that viscosity reduces to the point where intergranular contacts (**grain flows**), and even turbulence, become important and the flow transforms into a **hyperconcentrated flow**. In subaqueous settings, progressive dilution and transformation may take place through the incorporation of the sea or lake water by mixing at the front and top of the flow. In extreme cases, such dilution may lead to transformation into a turbidity current, involving the finer grain-size fractions of the initial debris flow (see 'Turbidity Currents' below).

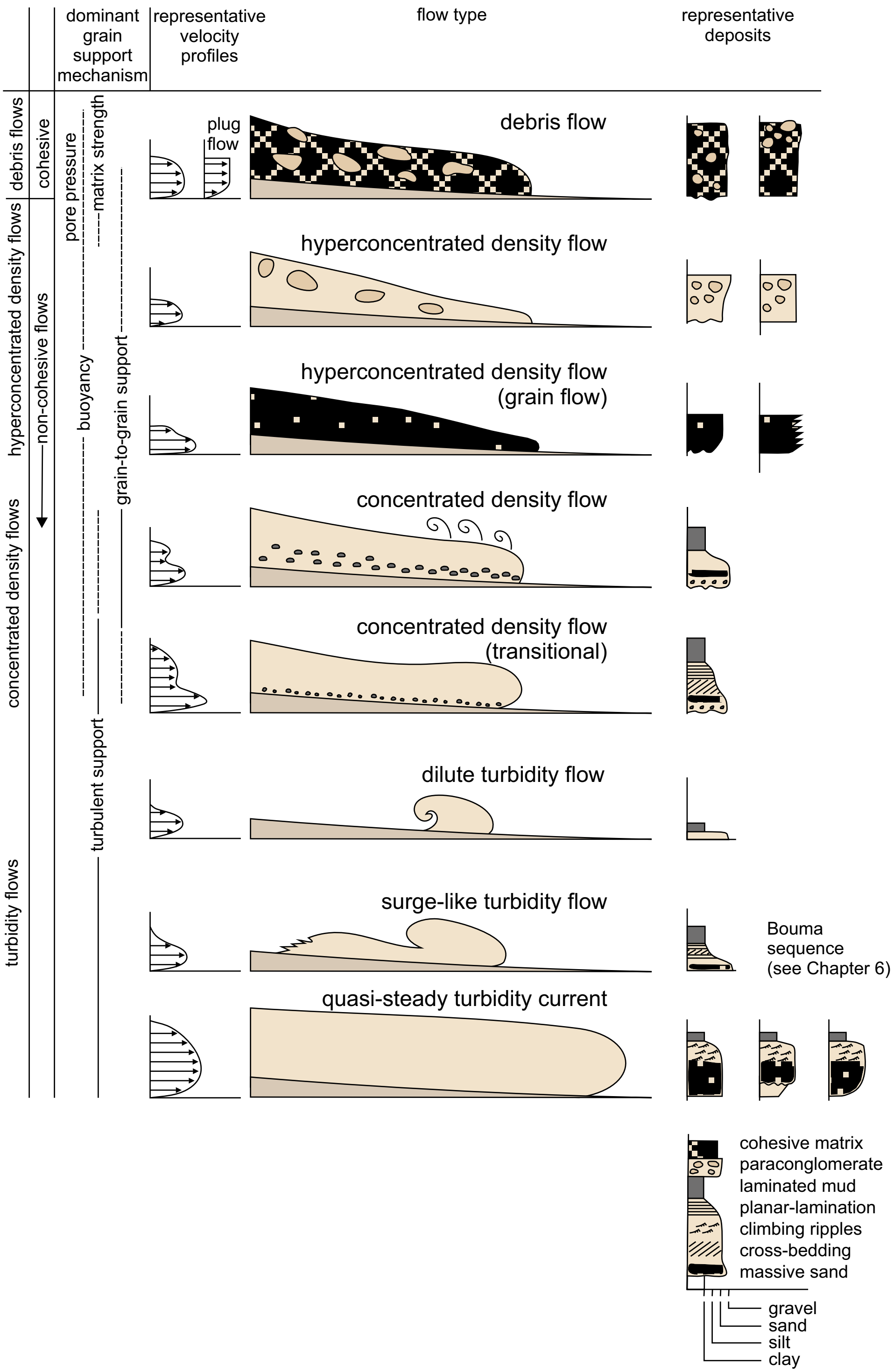


Figure 3.22 A schematic subdivision of gravity-driven mass flow types, showing the different particle support mechanisms and the resultant deposits. After Mulder and Alexander (2001).

Figure 3.23 Examples of sediment flow types and their deposits. A) Present-day mudflow, Svalbard. B) Deposits of a debris flow (a debrite), Gordo Megabed, Miocene, Tabernas, Spain. C) Grainflow on an aeolian dune lee slope, Namib Desert, Namibia.

Debris flows – a type of which are also known as **mudflows** – that maintain cohesive behaviour throughout their lives will eventually decelerate due either to a down-slope reduction in gradient or, in some subaerial cases, through progressive loss of pore water. When the shear due to the down-slope component of gravity falls below the prevailing yield strength of the sediment-water mixture, flow movement will cease and the textures and fabrics of the last stage of the flow will be frozen in the resultant deposit (Fig. 3.23A, B).

It is possible to investigate some of these features in at least a qualitative way by making experimental mudflows in the laboratory. Mix clay and sand with water to give different viscosities and see how these move on slopes of differing gradient and roughness, both in air and under water. The volumes, velocities, thicknesses and degree of internal deformation should all be noted. To preserve a record of internal deformation, make flows of Plaster of Paris injected with spots of dye or add Plaster of Paris to the matrix of a sandy flow. Careful slicing of the solidified mass will allow visualization of the internal deformation that acted in the latest stages of the flow.

In general, the deposits of debris flows and mudflows are characterized by a lack of stratification and of sorting of the particles within them. This is because all particles come to rest at more or less the same time when the flow stops. As we will see in later chapters, selective transport and deposition of grains of different sizes are much more effective in producing grading, stratification and lamination.

Hyperconcentrated and concentrated flows (including grain flows)

Natural concentrated and hyperconcentrated flows, acting on relatively low gradients, show a range of particle concentrations and an associated spectrum of particle-support mechanisms. Within such flows, turbulence and grain interaction together produce the particle support that allows movement. At the highest concentrations, intergranular collisions are most important whereas, at lower concentrations, the flows are transitional with turbidity currents where turbulence alone operates. Hindered settling, which occurs with all but very low sediment concentrations, also plays a role, as does buoyancy. In gravelly or sand-rich flows, intergranular collisions may dominate, resulting in the generation of **grain flows**.

Pure grain flows only occur on steep gradients. It is a common observation that a slope of loose dry sand is stable only below a certain gradient. Attempts to increase the gradient trigger a flowing movement of the sand after the angle has been increased by a few degrees (Fig. 3.23C). This movement reduces the gradient to one at which the slope is again stable. This type of avalanche movement and the existence of a particular **angle of rest** (repose) are important in depositing inclined laminae on the lee faces of ripples and larger bedforms in both air and water (see Ch. 6). Pure grain flows are most common on the steeply inclined flanks of aeolian sand dunes.

Some simple experiments can develop an understanding of this type of sediment behaviour. First of all, measure the angle of rest of sediments of different grain sizes, grain-size sorting and particle shapes under different conditions: dry, damp and saturated. See how the size and shape characteristics influence the angle. Try the experiments in air with dry sediment and, if possible, repeat the measurements under water. A second experiment is to steepen the slope of a pile of sand to see what angular difference exists between the angle of rest and the angle of slip. How does this angle vary with sediment type (e.g., grains of different shape) and conditions (e.g., dry sand versus sand with a small amount of water present)?

When the angle of rest is exceeded, internal shear stresses due to the down slope component of gravity overcome the intergranular friction. Once in motion, the expanded flowing layer is maintained by vigorous, intergranular shear and grain-to-grain collision creating a **dispersive pressure**, one of whose effects is to force larger particles upwards in the flowing layer. An everyday experience of this is the fact that shaking a bowl of sugar will bring any larger lumps to the surface. In addition, during the vigorous particle movements, smaller particles may filter downwards through the intergranular spaces of the dispersed layer, a process known as **kinetic sieving**. These two processes, either individually or combined, promote the development of an **inverse grading** in the shearing layer (see §6.8.2, §7.4.3). When the shear falls below a critical value, the flowing layer will "freeze" as it collapses upon itself as intergranular friction again dominates and grains resume a closer packing.

In concentrated and hyperconcentrated flows operating on lower gradients, other support mechanisms must also help to maintain particle support. During deposition from such flows, grain concentrations will be very high close to the bed and conditions akin to pure grain flow may be replicated. These so-called modified grain flows or traction carpets depend on the shear stress applied by an over-riding powerful current, probably a turbidity current with a high load of suspended sediment (Figs. 3.24, 3.25). Deposition from such a layer can occur gradually as sediment is added from suspension or it may freeze rapidly if the applied shear stress falls below a critical value. Inverse grading may develop within the moving layer where a suitable mix of grain sizes is available and this may itself be frozen where the layer freezes abruptly. In many cases however, the deposit will be of structureless sand and the bed said to be massive.

With lower sediment concentrations, turbulence will play an increasing role in particle support and the flows become transitional to high-concentration turbidity currents (see below).

Turbidity currents

Turbidity currents are the most important agents for transporting sand and silt-grade sediment into deeper-water settings. Their mobility depends upon the sedimentary particles being supported by the upwards components of

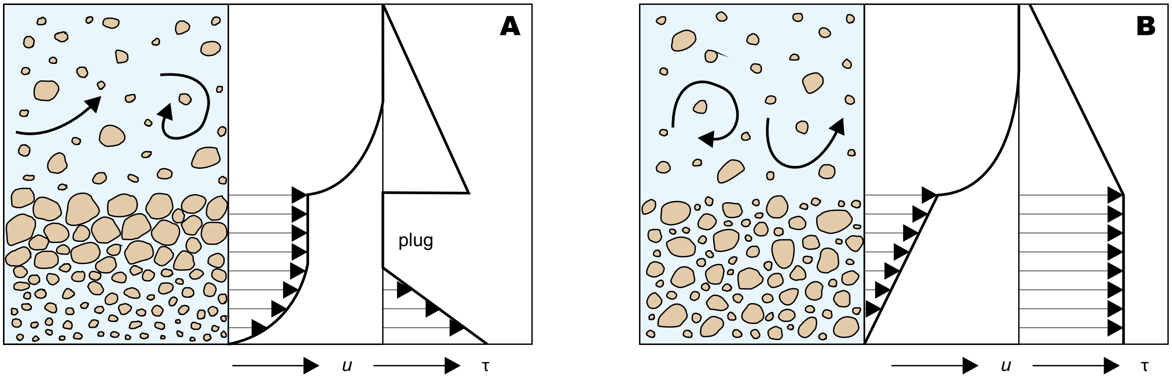


Figure 3.24 Comparison of the profiles of mean flow velocity u and shear stress τ for A) grainflow-like traction carpet (Lowe, 1982) with a non-deforming plug at the upper part and B) simple-shearing traction carpet (Hiscott, 1994) with constant shear stress throughout. After Sohn (1997).

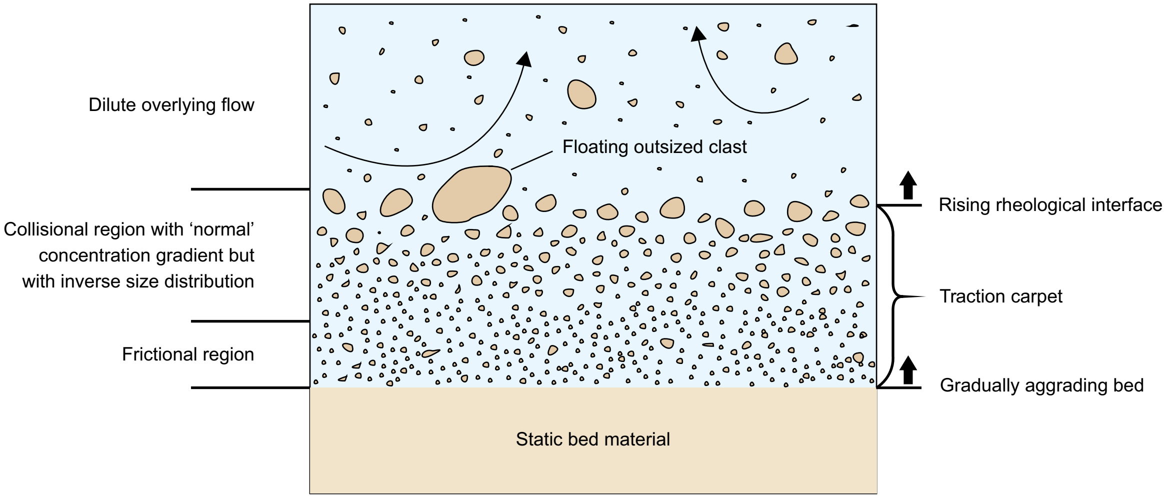
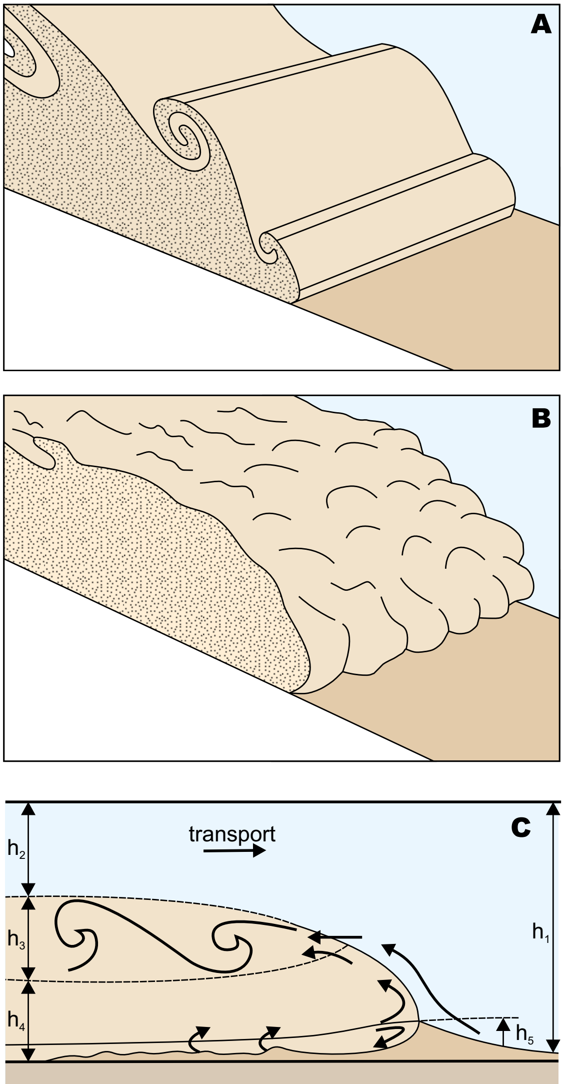


Figure 3.25 Deposition from a density-stratified and size-graded traction carpet via gradual aggradation of the bed. After Sohn (1997).

fluid turbulence. Turbidity currents are driven by gravity acting on the excess density of the suspension compared with that of the surrounding clear water. The various properties of the flow including thickness, concentration, velocity and gradient, are all highly interdependent. Changes to any one property will trigger changes in other properties. Turbidity currents, which are the principal flow mechanism by which large volumes of sand are carried out onto the deep ocean floor (submarine fans and abyssal plains), deposit their load when a reduction in gradient or in lateral confinement (as at the mouth of a channel) causes deceleration which in turn reduces turbulence and thereby the capacity of the flow to carry sediment. The currents also exert shear stresses on the bed, which may move, as bedload, sediment already deposited from suspension by the current (see §6.7.4). At their highest concentrations, turbidity currents grade into concentrated flows where other support mechanisms become increasingly important.

Large turbidity currents commonly originate by the slumping of poorly consolidated material near the top of a slope and are important in the transport of sand to the ocean floor, commonly via submarine canyons. The initial slumps mix with increasing volumes of sea water as they accelerate down slope and become more dilute until the sediment load is fully suspended, and the flow is a true turbidity current



h_1 - total height of the ambient fluid column
h_2 - height of ambient fluid column above flow
h_3 - height of mixing zone
h_4 - height of main body of flow
h_5 - height of foremost point of flow

Figure 3.26 Two kinds of instability at the front of a gravity current head which allow ambient water to be mixed into the body of the flow. A) Billows (Kelvin-Helmholtz waves). B) Brain-like lobes and clefts that develop at the contact of the overhanging head with a solid boundary. C) A section through the head of a turbidity current. Modified after Simpson (1987).

(Fig. 3.26). In the process, the flow may pass through stages involving movement as debris flows and as concentrated flows. Flows that are transitional between turbidity currents and mud flows are increasingly recognized in subaqueous environments. The deposits of these transitional flows are referred to as "hybrid-event beds (HEB)", "linked debrites" and "slurry beds" (see §6.7.7).

Because turbidity currents are fluidal flows, they have no yield strength and so decelerate gradually until the sediment load that drives them has been exhausted. As the flow drops its suspended load, the load may be reworked on the bed as a traction load and moulded into bedforms which may change character as the flow decelerates. Each bedform will produce its own distinctive lamination as reviewed in Chapter 6. The more rapid the deposition, the less time there is for bedforms and associated lamination to develop. If a range of grain sizes is present in the turbidity current, a graded bed may result through the coarser fraction falling out of suspension most rapidly on to its developing bed. Deceleration of a turbidity current, which causes it to deposit its load, can occur in either time or space. Deceleration in time occurs when the turbidity current is in the form of a surge that passes through a point and deposits its load as flow decelerates at that point. Deceleration in space occurs where a constant flow decelerates as it encounters lower gradients or expands from lateral confinement, for example at the mouth of a channel. In such a case, the flow velocity will decrease in a downstream direction even though it remains constant in time at any one place along its path. Similar considerations apply to accelerating currents which tend to be able to increase their capacity and have erosive potential. (Fig. 3.19).

Clear-water density currents, independent of any sediment load, also occur, commonly where cold river water enters a warmer lake. The higher-density cold water plunges below the lower-density warm water, forming a hyperpycnal, density-driven underflow and may cause sediment bedload transport on the lake floor. At a larger scale, oceanic water circulation is driven by thermo-haline density differences caused by the cooling of water near the poles and increased salinity due to the formation of pack ice.

Certain density currents, due to either temperature or very dilute sediment suspension, may have densities that fall between the extreme values in a stratified water column. In such cases, homopycnal **interflows** develop at density interfaces within the water column, and can carry fine material in suspension for long distances.

Density currents in general, and turbidity currents in particular, are suitable for simple laboratory-based

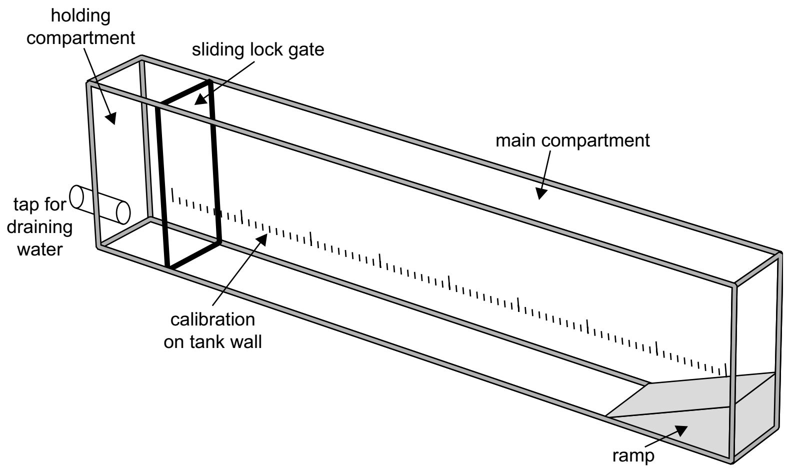


Figure 3.27 A simple clear plastic lock exchange tank for studying the influence of density on the mixing of different water bodies. Mixing is achieved by lifting the lock gate between the two compartments. A removable ramp can be used to study the effects of current reflections. Recommended dimensions are: length = 2-5m, width = 0.2-0.3m, height = 0.5-0.8m, water depth = 0.2-0.4m.

experimentation. Ideally, construct or obtain a narrow but deep glass- or clear plastic-sided tank at least 1m long and arrange a vertically sliding, fairly watertight gate near one end to create a small compartment (Fig. 3.27). Fill the whole tank with water and add dye to the water in the small compartment. Pull up the gate and observe how the coloured and the clear water interact. Now add a quantity of salt to the small chamber, add dye and repeat the experiment several times using different salt concentrations. How does the salinity of the introduced water influence the pattern and rate of the resultant flows? By using a small immersion heater or by allowing ice cubes to melt in the small compartment, repeat this experiment so that the released water is either warmer or colder than that in the main tank. How do the temperature differences influence the interaction of the water?

In a second series of experiments, add finely powdered, clay-sized mineral grains to the water in the compartment and stir them into a suspension before releasing it into the main tank. These experiments can be varied by using different concentrations of suspended material, by using minerals of different densities (e.g. kaolinite, calcite and barite) and by setting the tank at different gradients. If a tank is unavailable and cannot be improvised, it is still possible to appreciate some of these processes by stirring up the mud at the edge of a pond and watching the behaviour of the resultant suspension.

3.8 Pyroclastic density currents

Pyroclastic density currents are heterogeneous mixtures composed of volcanic particles and gas. The nature of their flow is determined by their density relative to the surrounding fluid (often air but sometimes water) and by the gradient over which they flow. They originate as a consequence of magmatic explosions, from which erupted material may be transported as a flow, surge or fall. **Pyroclastic flows** and **surges** are end-members of a spectrum of gas-rich gravity flows that range from concentrated laminar and plug flows to dilute turbulent currents. **Pyroclastic falls** involve the settling of sub-aerially erupted particles through air and then through water, which may be flowing or static. The behaviour of pyroclastic density currents is partly determined by the nature of the eruptive event, with gravitational collapse of vertical eruption columns, explosive disintegration of magma and rock, laterally inclined blasts and avalanches all being possible initiation mechanisms.

Many pyroclastic density currents are single-surge events generated by an individual, short-lived pulse of eruptive activity that waxes rapidly and then wanes rapidly. However, pyroclastic events resulting from longer-lived, fountaining eruptions may sustain pyroclastic density currents for several hours, during which time, the flow may vary between periods of quasi-steady conditions and periods of unsteady flow. Spatial changes in pyroclastic density current velocity are caused by factors such as downstream changes in slope, bed roughness and rate of ingestion of ambient fluid (air or water) into the flow. A pyroclastic density current is **accumulative** where it accelerates due to flow confinement or a downstream increase in gradient or both. It is **depletive** where it decelerates due to flow divergence or a downstream decrease in gradient or both (see §3.6.3; Fig. 3.19).

Study techniques

Field experience

It is important to generate a feel for the possible interactions of the many variables involved in sedimentary processes: different media (air and water in particular), mass, weight, density, effective density, temperature, size and shape of grains, laminar and turbulent flow, viscosity, eddy viscosity, bed roughness (grain and bedform roughness), shear stress, dispersive pressure, temperature, velocity, critical erosion velocity, celerity of waves, angle of slip and angle of rest. It is also important to develop an appreciation of certain dimensionless relationships: Reynolds number and Froude number.

Field programmes should include investigations, planned by both tutors and students, to observe and record some or all of the following processes in their natural settings:

Rivers, streams and estuarine channels Organized turbulence (around obstacles and bedforms; flow separation and attachment points; captive eddies); less-organized turbulence (e.g. boils); tranquil and rapid flow; hydraulic jumps and streaking.

Gutters and beaches Flow in very shallow currents; streaking in the viscous sub-layer.

Grainflows, mudflows, debris flows, soil creep (solifluction lobes) and avalanches Features on embankments of roads or cliffs in clays or tills commonly show the products of many of these processes. After wet weather, it may be possible to observe and document active processes, especially those operating at intermediate speeds. Documentation of slow processes (e.g. soil creep) require sustained periods of observation, whereas direct experience of rapid processes, such as avalanches will usually be by accident rather than design, and safety should be the priority. In observing both products and processes try to identify features of laminar flow and plastic and brittle deformation (e.g. crevasses, joints and rotational shear).

Flow of wind over dunes or obstacles Organized or less organized turbulence (see Ch. 6); transport mechanisms (suspension, saltation, creep, rolling); deflation.

Density and turbidity currents Flows of mud/silt at the edge of ponds; cold air or water beneath warm; salt water under fresh water.

Waves in ponds, lakes and at the seashore Wavelength, wave height, celerity, wave base, breaking waves, swash/backwash.

Laboratory experience

Many physical experiments can be devised so as to afford direct observation of the properties of fluids and flows. Three simple experiments are outlined here; many more are discussed by Allen (1985).

Hydraulic jump and the transition from rapid to tranquil flow Place a sheet of glass on a flat, horizontal surface and, using a hosepipe connected to a tap, allow a jet of water to fall vertically onto the glass sheet. Use the tap to control the water discharge. At low discharge, a uniform and comparatively thick flow covers the glass sheet. At higher discharge, the flow will become divisible into two regimes: an inner regime, close to where the jet strikes the glass, characterized by a small flow depth and a large velocity (rapid flow) and a contrasting outer flow characterized by deeper and slower tranquil flow. The abrupt transition from rapid to tranquil flow is represented by a hydraulic jump.

Stokes' law of particle settling Fill a 1L measuring cylinder with glycerine and add a grain of gravel (5–8mm diameter). Measure the rate of sinking of the grain through the glycerine column. Repeat the experiment with further grains of varying diameters. Next, repeat the experiment with high-density steel ball bearings. Shine a bright lamp on the measuring cylinder to increase the temperature of the glycerine and repeat the experiments. Note the effects that particle size, particle density and temperature of the fluid have on settling velocity. Relate these observations to Equation 3.14.

Fluid turbulence and the motion of non-cohesive particles Connect a rubber hose to one end of a transparent plastic tube (0.5m long, 0.05m diameter); leave the opposite end of the tube open. Stand the tube upright with the attached hose at the base and half-fill it with low-density polystyrene spheres (2–5 mm diameter). Pass an air supply down the hose and into the base of the tube (a regulated air bottle is ideal, although a person blowing on the end of the hose will also suffice). Note how, as air is blown through the polystyrene spheres at an increasing rate, a critical speed is reached at which the spheres become disengaged from each other and "float", such that the mixture becomes fluid-like.

Recommended references

Allen, J. R. L. 1970. *Physical processes of sedimentation*. A clear account of the basic physics of sediment transport and deposition.

Allen, J. R. L. 1982. *Sedimentary structures: their character and physical basis.* An encyclopaedic account of sedimentary structures and the physics of their development, as understood at the time.

Allen, J. R. L. 1985. *Experiments in physical sedimentology.* Outlines a series of experiments concerning flows and sediment transport that can easily be conducted in a standard laboratory.

Allen, P. A. 1997. *Earth surface processes*. A good general discussion of fluid dynamics in relation to sediment entrainment, transport and deposition.

Amy, L., W. D. McCaffrey & P. Talling (eds.) 2009. *Sediment gravity flows – recent insights into their dynamic and stratified/composite nature*. A thematic set of papers that consider the mechanics of sediment gravity flows and their resultant deposits.

Ashworth, P. J., S. J. Bennett, J. L. Best & S. J. McLelland (eds.) 1996. *Coherent flow structures in open channels*. Examines the behaviour of channelised flows and considers the nature of resultant deposits.

Bridge, J. S. 2003. *Rivers and floodplains: forms, processes and sedimentary record.* A thorough account of river sediments with a strong emphasis on hydrodynamics and computer modelling.

Leeder, M. R. 1999. *Sedimentology and sedimentary basins: from turbulence to tectonics*. A good general discussion of fluid dynamics in relation to sediment entrainment, transport and deposition.

Leeder, M. & M. Pérez-Arlucea 2006. *Physical processes in earth and environmental sciences*. Consider fundamental physical flow processes and their deposits.

McCaffrey, W. D., B. C. Kneller & J. Peakall (eds.), 2001. *Particulate gravity currents*. A thematic set of papers that consider processes associated with gravity current and their resultant deposits.

Middleton, G. V. & J. B. Southard 1984. *Mechanics of sediment movement*. A classic in its time and still one of the clearest accounts of the basic hydrodynamic principles.

Mulder, T. & J. Alexander 2001. *The physical character of subaqueous sedimentary density flows and their deposits*. A clear and logical account of this often confusing topic.

Pye, K. (ed.) 1994. *Sediment transport and depositional processes*. Explains the fundamental principles of sediment behaviour in flows.

Talling, P. J., D. G. Masson, E. J. Sumner & G. Malgesini 2012. *Subaqueous sediment density flows: depositional processes and deposit types*. A state-of-the-science journal article that brings together much of the thinking developed in the previous two decades and develops novel models that represent a major step forward.

Tennekes, H. & J. L. Lumley, 1972, *A first course in turbulence*.

Tritton, D. J. 1988. *Physical fluid dynamics*.

CHAPTER 4

Erosional structures

4.1 Introduction

Most areas of present-day sediment accumulation reflect complex interactions between erosion, transport and deposition. Even in areas of net long-term accumulation, deposition may be interrupted by periods of erosion. Similarly, most ancient successions are not the products of steady, continuous deposition but result from alternating periods of deposition, non-deposition and erosion. This chapter deals with features that indicate that erosion has taken place.

As with most depositional structures (Chs. 5–7), the chances of an erosional structure being preserved in the rock record are very small. For erosional structures to be preserved the eroded sediment has to be sufficiently cohesive and strong to maintain the erosional relief until it is buried by contrasting sediment, usually very shortly after the event responsible for the erosion. Small-scale erosional structures are almost always recognized as relief on the base of the bed immediately overlying the eroded surface. Erosion is also recognized in vertical sections by truncation of bedding or lamination in the sediment below the erosion surface. If erosion has been widespread, as at some unconformities, no discernible relief may be preserved and recognition of erosion may then depend upon indirect evidence. Where relief is observed, this may not reflect the total amount of erosion. Widespread erosion of a large thickness of sediment may result in preservation of only small-scale features. Observed relief therefore only reflects the minimum thickness of sediment removed.

Many erosional structures are valuable indicators of "way-up" and of palaeocurrent direction. They can, therefore, help in structural and palaeogeographical analysis, as well as giving insights into processes active during intervals of overall net sediment accumulation.

Classification of erosional structures has to be arbitrary as different types grade into one another. The scheme adopted here is based on both descriptive and genetic criteria (Fig. 4.1). Three broad categories are recognized, within which further subdivision is possible:

- sole marks on the bases of coarser beds in inter-bedded successions;
- small structures seen on modern sediment surfaces and more rarely on upper bedding surfaces in ancient strata;
- large structures normally recognized in vertical section in ancient sediments (i.e. channels and slump scars).

4.2 Sole marks

4.2.1 Preservation

Sole marks comprise a diverse group of structures found as casts on the bases of coarser-grained beds that are inter-bedded with mudstones. The coarser-grained sediments are commonly sandstones but exceptionally may be limestones or conglomerates. The sole marks result from the erosion of cohesive, fine-grained sediment, usually mud, which pass into suspension on erosion. The cohesive strength of the sediment allows details of the erosional relief to be maintained until they are buried by coarser-grained material (Fig. 4.2). Erosion of mud and deposition of coarser material may occur as different phases of the same current, separated by only a short period of time. Subsequent lithification usually renders the coarse-grained sediment more resistant than the finer material to eventual weathering, so that the fine-grained sediment is preferentially removed to expose a cast of the erosional relief on the base of the sandstone bed. Resumption of deposition of fine-grained sediment, similar to that eroded, without any deposition of coarse-grained sediment would not normally provide the lithological contrast needed to pick out the structures at the time of later exposure and weathering. It is very important to understand this mode of preservation and to recognize that the structures observed are negative impressions (i.e. natural casts) of the erosional relief.

Sole marks are typically the products of environments characterized by episodic sedimentation. Steady background deposition of mud is punctuated by sudden influxes of coarser sediment in high-energy events comprising an

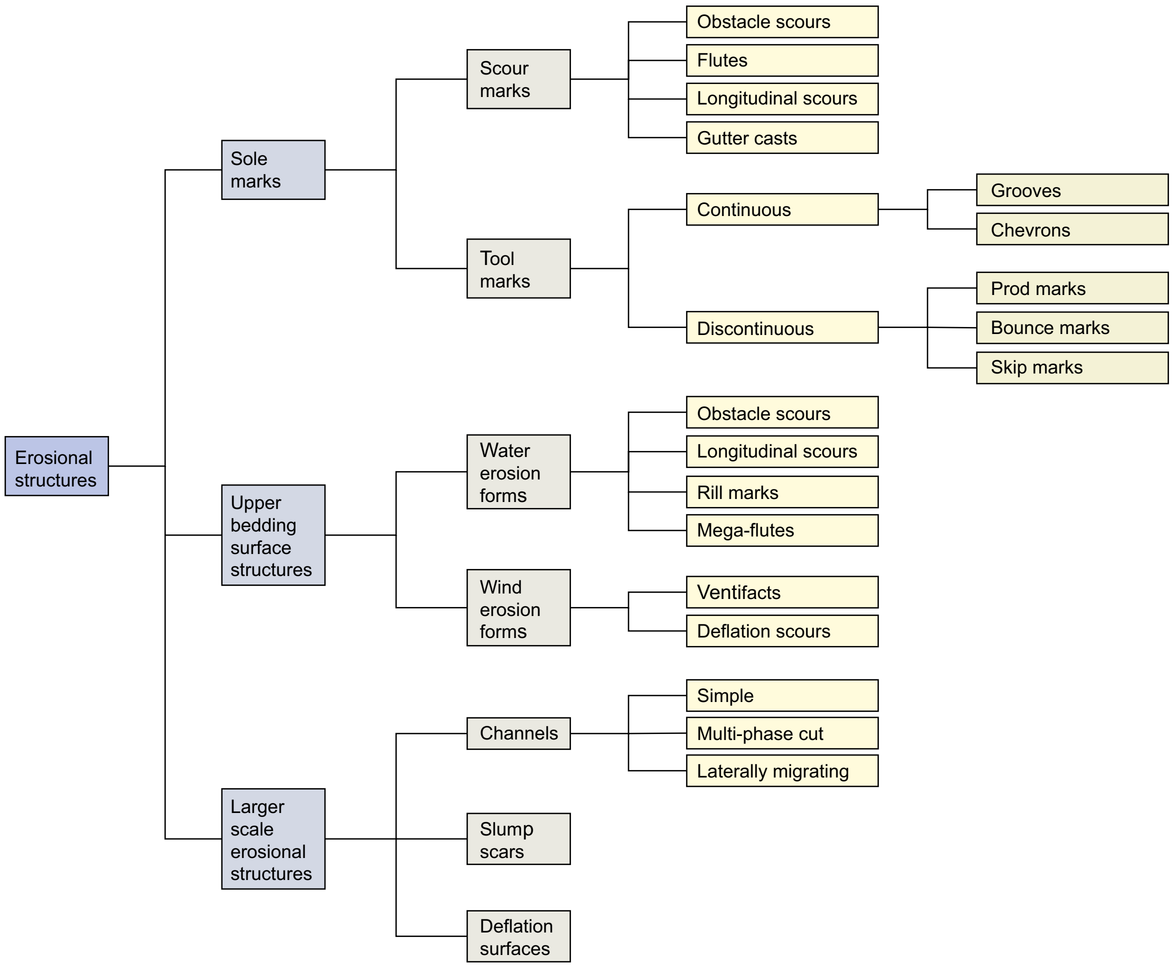


Figure 4.1 Scheme for the classification of erosional sedimentary structures.

early erosive phase and an immediately succeeding depositional phase. A common example of such an event is the turbidity current (see §3.7.2). It was once thought that sole marks were diagnostic of turbidites; however, storm surges in shallow seas, sheet-like flows in semi-arid environments, and crevasse surges into floodplains all have the necessary properties to produce such structures. Interpretation of sole marks should initially be restricted to the processes involved, rather than to the type of event or the environment until the full context of the structures is understood.

Sole marks are divided here into two broad classes which differ principally in the way the structures are generated: structures due to turbulent scour (scour marks); structures due to objects moved by the current (tool marks).

4.2.2 Scour marks

Scour marks are distinguished by their generally smooth shape and commonly by their streamlined appearance. They may occur as isolated casts or in groups that cover a bedding surface in organized patterns. A variety of shapes occur, amongst which it is possible to recognize groups that can be given a common name and described together. Four main groups cover the range of forms: obstacle scours, flutes, longitudinal scours and gutter casts.

Obstacle scours

Large clasts such as pebbles, fragments of wood and more robust types of fossils sometimes occur on the bases of sandstone beds (Fig. 4.3A) and are associated with distinctive ridges of sandstone that point down into the underlying bed. The ridges are commonly crescentic or horseshoe-shaped,

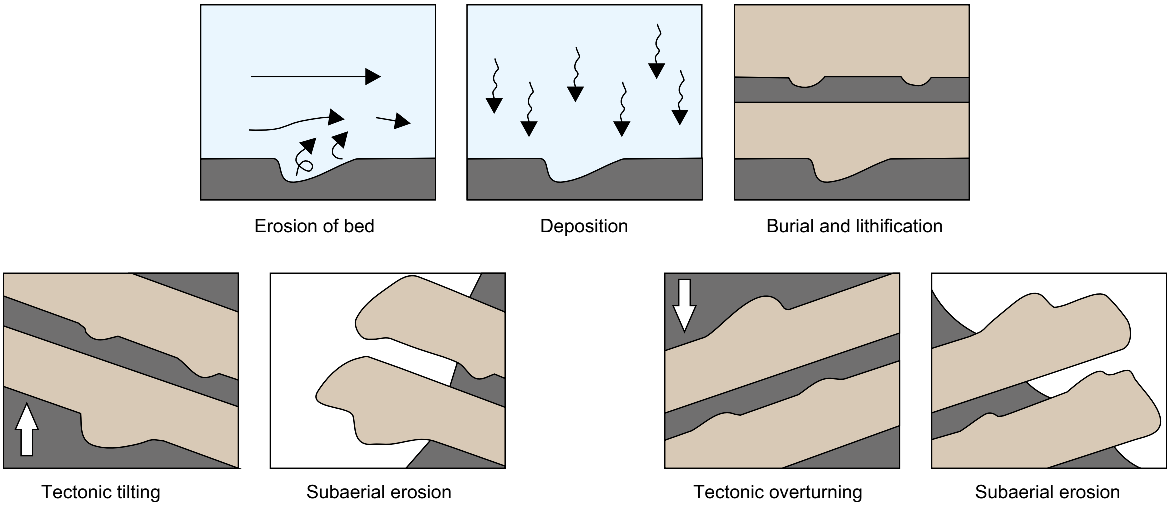


Figure 4.2 Stages in the development of a sole mark and its potential use as an indicator of 'way-up'. After Ricci Lucchi (1970).

partially encircling the large clast with tails dying away in one direction (Fig. 4.3B). These ridges are casts of troughs developed around the large clast. The development of these structures can be readily observed on a sandy beach or on sandy stream beds over which there is quite a strong flow of water. Place a pebble or some other obstruction on the bed and see what happens when the stream flow or the backwash of waves passes over the bed. A crescentic scour trough will commonly develop around the obstruction with the deepest part of the trough on the upstream side of the obstacle and the tails pointing downstream. The scour trough is caused by the accelerated flow around the obstacle and its shape relates to the pattern of eddies generated by this acceleration. The eddies are directed vigorously downwards onto the bed on the upstream side of the obstacle, whilst spiral eddies are shed on either side and die out downstream producing the tails (Fig. 4.4). The structure of these localized captive eddies can be picked out around obstructions using a stream of injected dye (see §3.2.3).

Obstacle scours are not very common as sole marks but, where present, they provide a good indication of current direction and of "way-up". In some cases, probable obstacle scours occur although the obstacle itself has been removed. In such cases, a horseshoe-shaped ridge occurs in isolation and the nature of the obstacle is left to the imagination.

Figure 4.3 Examples of obstacle scour marks on bedding surfaces. A) Obstacle scour around a pebble in the base of a sandstone bed. Several smaller pebbles also show their own scours. Current from top left to bottom right. San Vicente Formation, Eocene, Ainsa, Spain. B) Obstacle scour marks preserved as casts on a lower bedding surface. The original obstacles would have likely been pebbles but are not preserved. Flow would have been from the top to the base of the photo. Moenkopi Formation, Triassic, Utah, USA.

Flutes

Flutes are similar to obstacle scours and occur both as isolated features and as collective groups that share a common

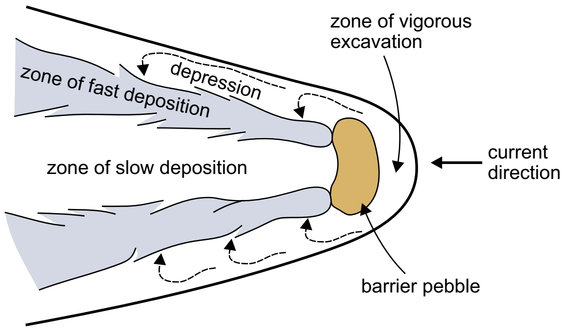


Figure 4.4 The pattern of eddies associated with an obstacle on a bed and its relationship to the formation of obstacle scours. After Sengupta (1966).

origin and form distinctive patterns. Individually they vary in shape and size, but on any one surface they tend to be rather similar. Flutes are characterized by a rounded, sometimes tightly curved "nose" at their up-stream end. The deepest part (i.e. maximum relief) occurs close to the nose, from which point the feature flares away and dies out. On any bedding surface the "noses" of all flutes will usually point in the same general direction. Flutes typically range in length from 5cm to 50cm, in width from 1cm to 20cm and in depth commonly a few centimetres, exceptionally up to 10cm. In shape they range from highly elongate forms, gradational with longitudinal scours, to very wide forms with gently curved noses that are termed **transverse scours** (Figs. 4.5, 4.6). Some flutes have highly twisted shapes (Fig. 4.6B), particularly close to the nose, whereas others have very simple streamlined forms (Fig. 4.6C, F). The sides of some flutes are unusual in showing a pattern of small-scale steps. These steps can be generally related to the lamination or thin bedding in the underlying sediments where slight differences in grain size have caused differential erosion.

Descriptions of flutes should include measurements of their dimensions, orientation and the direction in which they point, and comments on their overall shape, together with a note as to whether the flutes are distributed in a pattern on a bedding surface. Linear patterns occur where the flutes are arranged longitudinally, whereas other patterns are characterized by "en echelon" and "fish-scale" arrangements (Figs. 4.7, 4.8).

Flutes can be produced experimentally when water flows over a surface of cohesive sediment or over a slightly soluble substrate. Small bumps and depressions on the bed cause acceleration of the flow which gives rise to flow separation. The associated higher shear stresses lead to erosion which, in turn, emphasizes the relief near

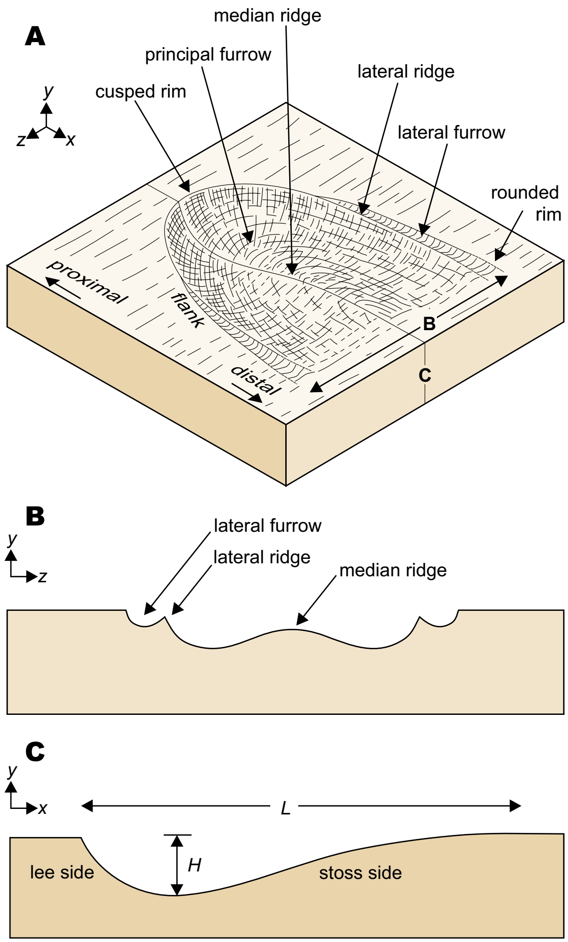


Figure 4.5 Morphological features of flute marks. A) Oblique view. B) Transverse section view. C) Longitudinal section view. Modified after Allen (1982).

the irregularity, and causes flute growth. The scale of separation and the erosional relief increase together and will continue to do so as long as suitable flow conditions are sustained. Eventually evidence of the initial irregularity will be destroyed. Erosion is most concentrated near the nose of the flute from where it dies out downstream as the eddies are absorbed into the body of the flow (Fig. 4.9). The shape of the flute is intimately related to the structure of eddies in the nose region.

The shape and pattern of flutes bear quite a close relationship to the shape and distribution of the initial irregularities if the scour and growth of the flutes did not last very long. Where erosion was sustained, the flutes may reflect the strength and duration of the current that eroded them. In addition to those flutes that occur

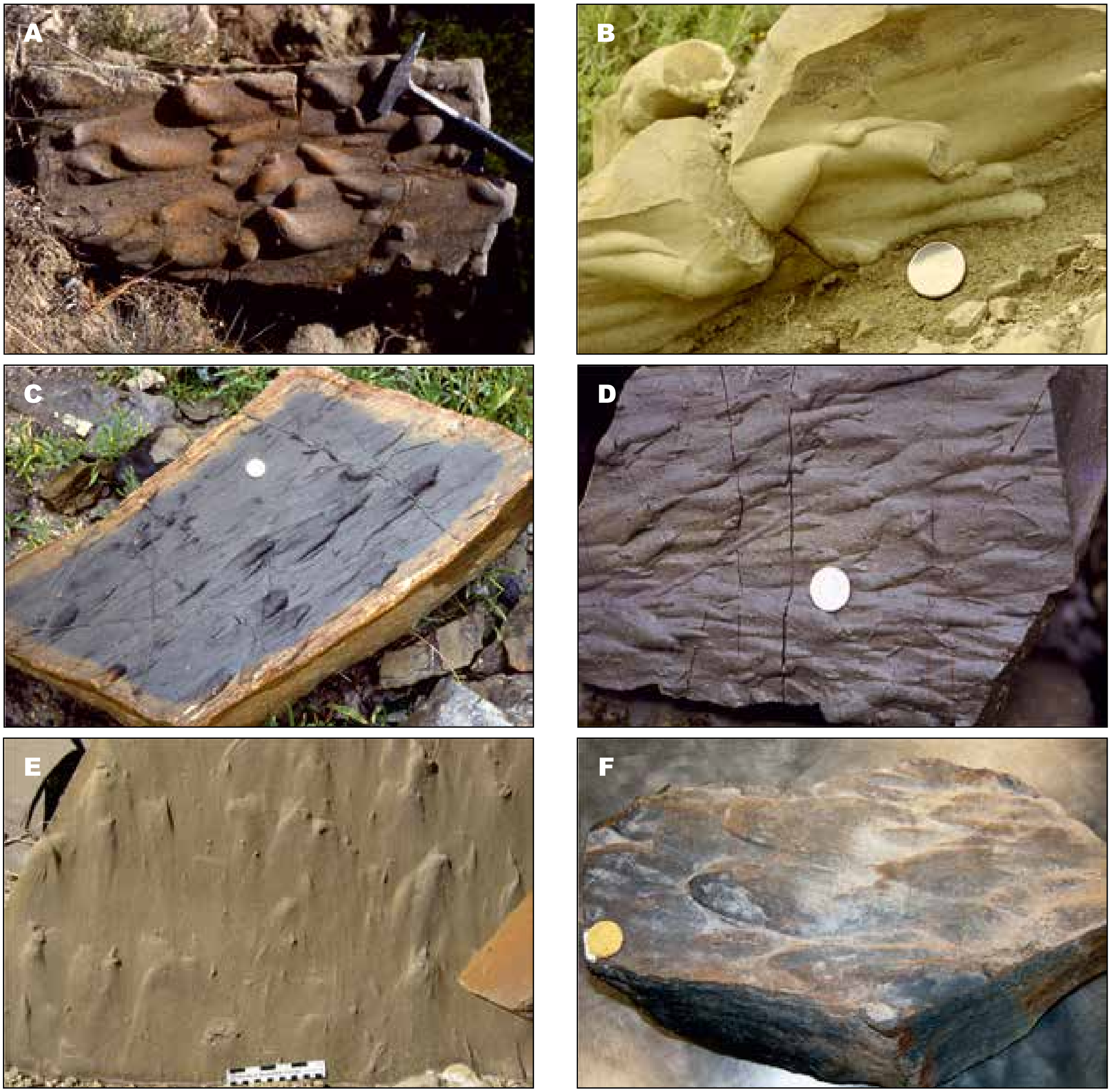


Figure 4.6 Examples of different flute forms on the bases of sandstone beds. Try to judge the palaeocurrent direction in each case. Examples C and E show both flutes and tool marks. Examples C and D also show cross-cutting elongate burrows. Photos A and B courtesy of Gilbert Kelling. Photo F courtesy of University of Leeds collection.

as casts of erosional forms on the bases of bed, flutes also occur on bedrock surfaces that have been subject to intense abrasion or dissolution by powerful currents on, for example, the walls of canyons and caves. These are particularly common where the bedrock is limestone and where differential dissolution reflects the patterns of local turbulence.

Not all flutes develop from initial irregularities of the bed. Some may develop from the lateral merging of longitudinal scours. As well as being a valuable indicator of "way-up" in deformed successions, flutes are amongst the most abundant and important indicators of palaeocurrent direction: the tightly curved (in some case bulbous) nose is located at the upstream end of the flute.

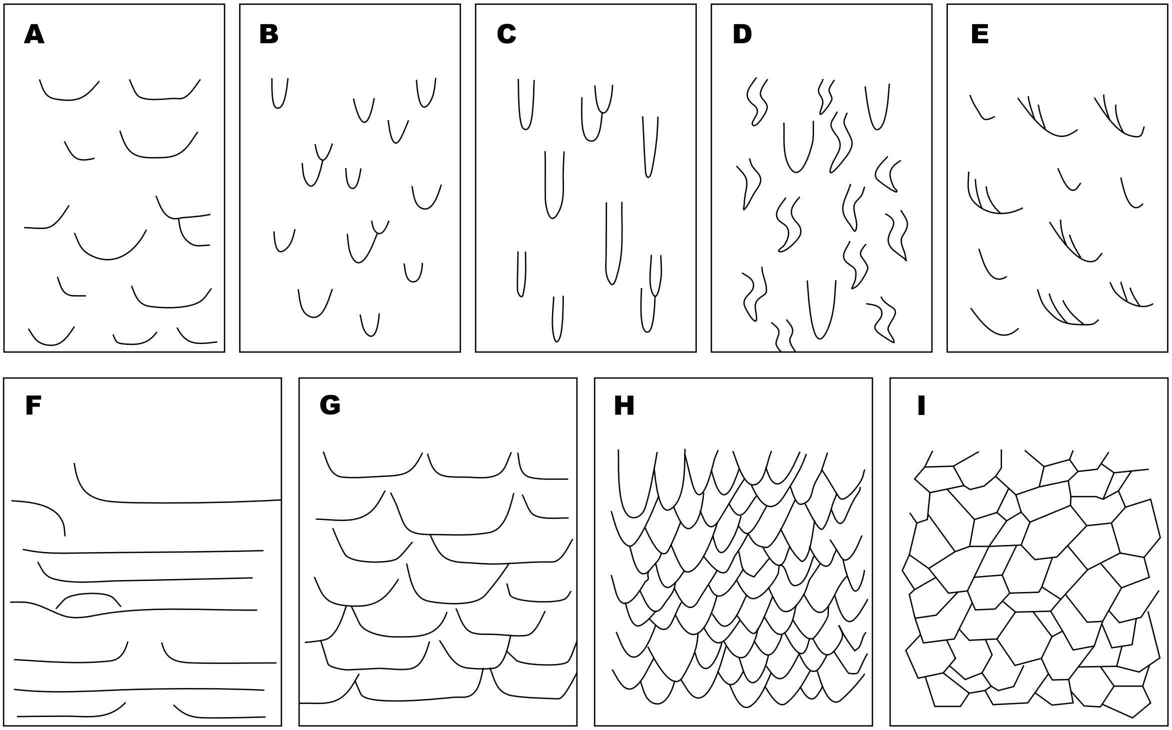


Figure 4.7 Schematic illustration of the range of shapes of flutes in plan view. Current from bottom to top in each case. After Allen (1971).

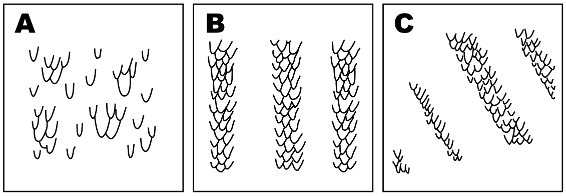


Figure 4.8 Heterogeneous patterns of distribution of flute casts on bedding surfaces. After Allen (1971).

Superficial examination could cause transverse scours to be confused with straight or sinuously crested ripples, leading to misinterpretation of both the way-up and the palaeocurrent direction. If there is any uncertainty, try to resolve it by looking for related internal structure. Ripples normally show an internal cross lamination (see §6.1.4) whereas transverse scours usually have no related internal structure.

Longitudinal scours (longitudinal ridges and furrows)

Longitudinal scours occur as patterns of closely spaced parallel ridges and furrows on the bases of sandstone beds. In transverse cross section, sandstone casts are characterized by rather rounded, downward-pointing ridges and intervening furrows that are rather sharp, reflecting round-bottomed troughs and sharp ridges on the surface of the eroded mudstone (Fig. 4.10). The spacing of the ridges is typically 0.5–1cm with a relief of a few millimetres. Although the overall pattern is one of parallelism, ridges do end if traced far enough along their length. Some die out by coalescing with a neighbour, whereas others show rounded ends reminiscent of the noses of flutes. Some patterns are parallel and continuous whereas others are markedly dendritic. Wider ridges with distinct noses are gradational to flutes.

Longitudinal scours result from patterns of small-scale eddying close to the bed where the axes of spiral eddies were parallel to the current. Adjacent eddies have opposite senses of rotation so that flow at the bed has alternating zones of upward- and downward-directed flow. The zones along which descending vortex limbs impinge on the bed will be sites of high stress and rapid erosion, whereas beneath ascending limbs stress will be at a minimum and erosion at its lowest (Fig. 4.11).

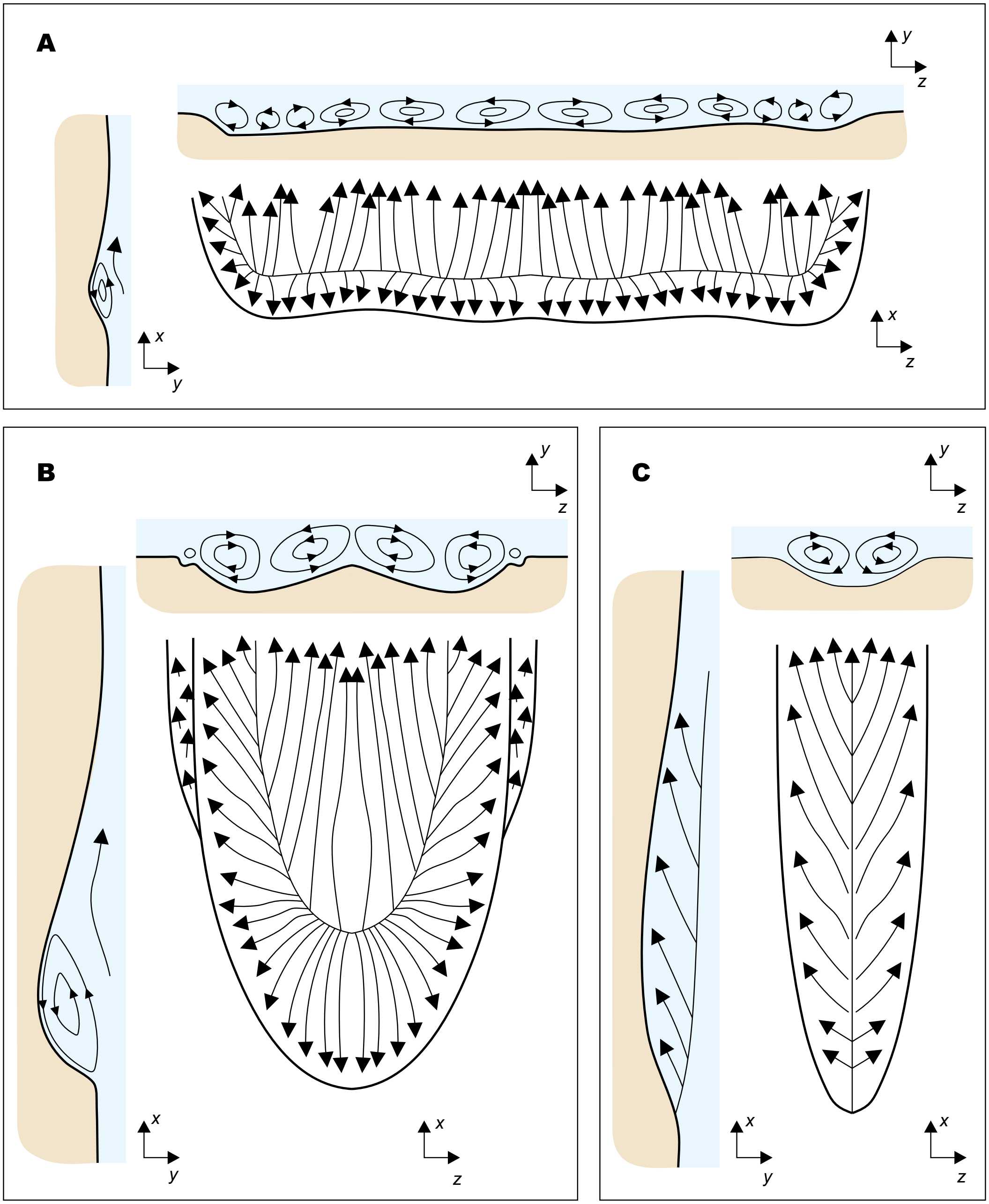


Figure 4.9 Simplified patterns of water motion (eddying) associated with erosional scours of different shapes. A) Transverse scour. B) and C) Parabolic flutes of differing width. Flow from bottom to top in all cases. x direction is downstream, y is vertical and z is transverse to flow. After Allen (1971).

Once localized on scour features, eddies become fixed and accentuate the relief. Where sandstone casts have rounded noses, these are convex up stream and they probably reflect a pattern of eddying similar to that occurring in flutes, with flow separation and a local transverse component to the eddy axis.

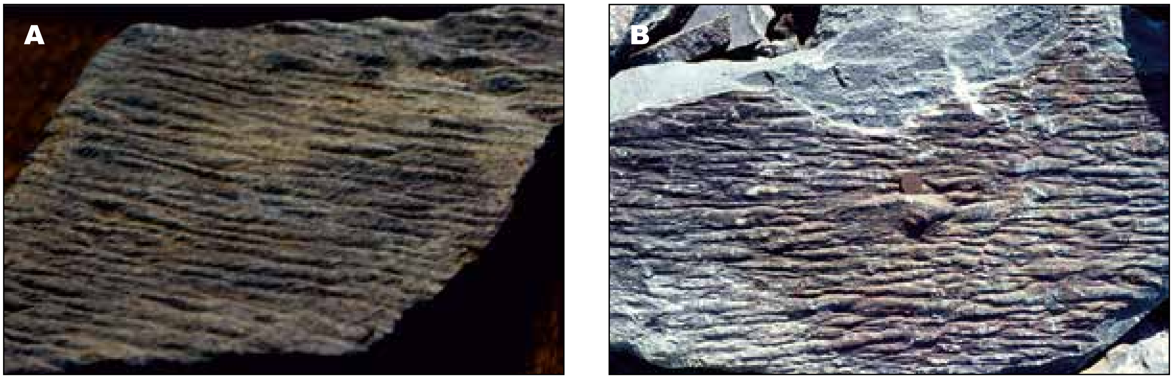


Figure 4.10 Examples of longitudinal scours on the bases of sandstone beds. In most cases it is only possible to tell the trend of flow and not its sense of direction. Examples with "noses" also enable the sense of movement to be judged. A) Mam Tor Formation, Upper Carboniferous, Derbyshire, England. B) Longitudinal scours that are transitional to flutes. Location unknown. Photo courtesy of Gilbert Kelling.

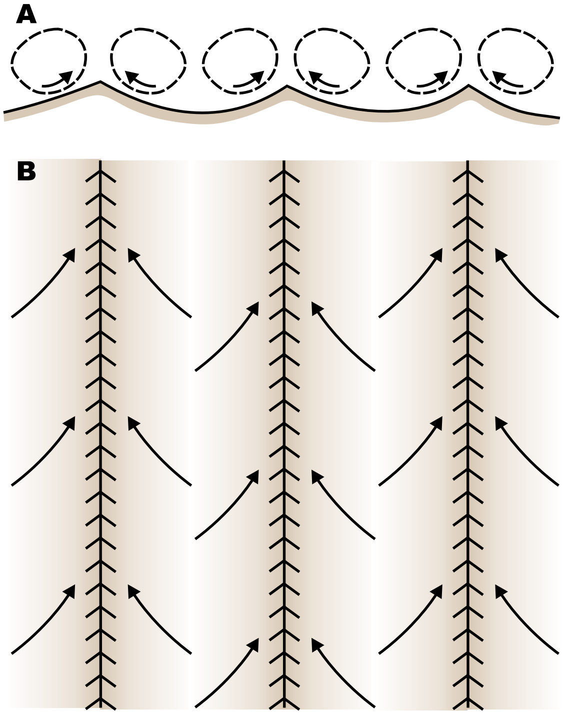


Figure 4.11 The pattern of water movement associated with the development of longitudinal scours. A) Vertical section normal to flow. B) Plan view on the bed. In part after Allen (1971).

Longitudinal scours are useful indicators of "way-up" and of palaeocurrent trend. However, only examples with flute-like noses can indicate the sense of palaeocurrent movement (i.e. upstream and downstream directions).

Gutter casts

These structures generally occur as isolated elongate ridges on the bases of sandstone or coarse-grained limestone beds. They protrude into the underlying finer-grained sediment from an otherwise rather flat bedding surface, and in vertical section they show U- or V-shaped profiles (Fig. 4.12). These are generally symmetrical and, more rarely, asymmetrical when one side is steeper than the other. They are commonly up to 10cm wide and of similar depth. Where the coarse-grained sediment does not give a continuous bed above the erosion surface, the coarse infills are preserved as isolated, elongate bodies in the finer-grained sediment.

In plan, the casts are commonly slightly to moderately sinuous and they extend for several metres. Sometimes their ends are seen and these may be quite steep, similar to flutes, or they may gradually die away. Some ridges are gently curved in plan, in which case it is common for the outer margin of the bend to be steeper. Smaller features (commonly tool marks) may be superimposed on the walls and floors of the gutter casts, showing preferred orientation parallel to the elongation of the gutter.

Gutter casts are the product of fluid scour, possibly aided by the "sand blast" effect of coarser grains carried by the flow. They appear to reflect a pattern of helical vortices with their horizontal axes parallel to the flow. Pairs of vortices are probably responsible, but on bends one may become dominant to give oversteepening of the outer wall, similar to the outer bank of a meandering river channel, albeit at a much smaller scale.

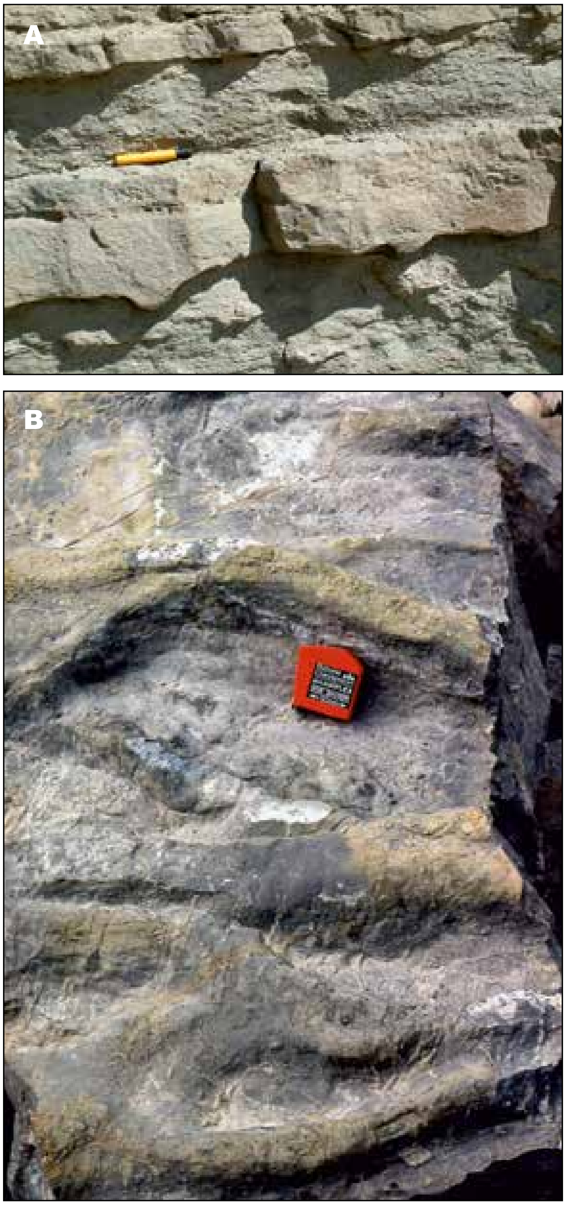


Figure 4.12 Examples of gutter casts. A) Gutter cast seen in end section at the base of a lower sandstone turbidite bed. Tabernas Basin, Miocene, southeast Spain. B) The lower surface of the fill of a gutter cast in a loose, overturned block of limestone. Note the sinuous shape of the gutter and the slightly anastomosing pattern. Campanuladal Formation, Proterozoic, north Greenland.

4.2.3 Tool marks

Tool marks differ from scour marks in being produced by objects carried by the flow rather than by the fluid itself. They also have rather more sharply defined shapes, and they often carry detailed patterns of small-scale relief. A simple morphological classification is:

Continuous	Sharp and irregular profile: **grooves**
	Smooth and crenulated: **chevrons**
Discontinuous	Single: **prod marks**, **bounce marks**
	Repeated: **skip marks**

Grooves

Groove casts are elongate ridges on the bases of sandstone beds. They occur in isolation or in groups composed of individual ridges all of which are parallel or near-parallel (i.e. have common orientations). In transverse vertical section they typically show a sharply defined "U"- or "V"-shaped relief. In some cases, the relief is irregular where smaller superimposed grooves and ridges are superimposed on the larger form (Fig. 4.13). The smaller, superimposed features tend to trend parallel to the larger ones, but sometimes they are twisted to give a corkscrew effect. Ends of groove casts are seldom seen but they may be gradual or quite sharp. Rarely, a mudflake, a plant fragment or a fossil may be found embedded at the end of the groove casts.

Most bedding planes with groove casts show only one trend of groove but in some cases, multiple trends may be apparent. It is important to carefully measure and record the orientation of groove casts. On surfaces where more than one trend is apparent, it can be valuable to record the various orientations and to try to put them into chronological order based on cross-cutting relationships.

Groove casts result from the infilling of erosional relief gouged by an object, or tool, being dragged through a cohesive substrate by a current (Fig. 4.14A). More rarely, it is possible that grooves may result from the rolling of disc-shaped tools leaving tracks similar to those of wheels in soft sand or mud. The twisted appearance of some grooves reflects rotation of the tools as they were dragged along the bed. The identity of the tool is usually unknown, though where an object is found at the end of the groove, the nature of the tool and the sense of movement are both established. However, with most grooves it is only possible to judge the trend of movement and measurements should therefore be recorded as lineations (e.g. 120–300°).

Chevrons

Rarer than grooves, chevrons are linear zones of "V"-shaped crenulations that consistently close in one direction to

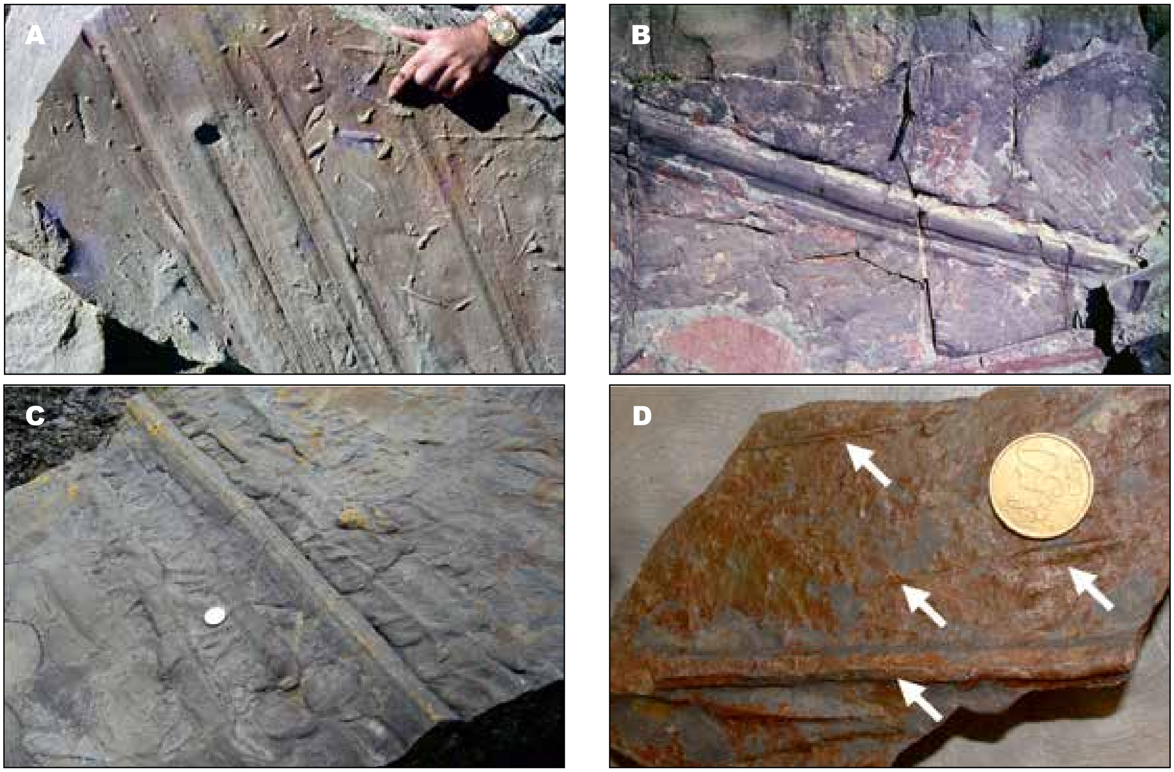


Figure 4.13 Examples of groove casts on the bases of sandstone beds. A) Rather low-relief grooves with later burrows. Hecho Group, Pyrenees, Spain. B) Large, well defined groove system, Silurian, southern Scotland. Photo courtesy of Gilbert Kelling. Note hammer for scale. C) Large groove probably generated by the dragging of a plant stem over a bedding surface while it was carried in a current. Mam Tor Sandstone Formation, Carboniferous, Derbyshire, England. D) Groove marks of different sizes but with similar trends on a bedding surface. Locality unknown.

produce a chevron pattern (Figs. 4.15, 4.16). The individual linear zones are seldom more than 3cm wide and the relief is generally less than 5mm. Dragging a stick through soft mud or any very viscous fluid produces a similar pattern. Chevrons record small-scale folding or “rucking-up” of the surface of weak but cohesive mud by the passage of a tool very close to the bed (Fig. 4.14B). The V-shaped ridges, which make up the chevron mark, close downstream, thus giving a sense as well as a trend to palaeocurrent measurements. Like other sole marks, chevrons are useful and reliable “way-up” indicators.

Prod marks and bounce marks

Sharply defined, discontinuous marks, usually elongate and with a preferred orientation, occur on the lower surfaces of many sandstone beds. Some, called prod marks, are notably asymmetrical along their length, with one end being deep and well defined whereas the other end is gradational and slopes gently to the bedding surface (Fig. 4.17A). Others, called bounce marks, are more symmetrical along their length, being gently sloping at both ends (Fig. 4.17B).

Both types vary in size from several centimetres wide and tens of centimetres long down to very delicate forms, less than 1cm in length and 1–2mm wide. Depths are roughly proportional to width, the smallest forms being only 1–2 mm deep. Larger examples may show the superimposition of delicate ribbed relief comparable to that seen on grooves.

In describing and recording these marks in the field try to measure their size and direction, making sure always to record the sense of any longitudinal asymmetry.

Prod and bounce marks record the impact of larger objects on the bed. With prod marks, the approach angle of these objects was rather large, so that on impact they dug deeply down into the mud before being pulled out steeply by the flow to leave a blunt “nose” at the downstream end of the mark (Fig. 4.14C). Note particularly that the asymmetry of prods is opposite to that of flutes whose “noses” are at their upstream ends. Bounce marks reflect a lower

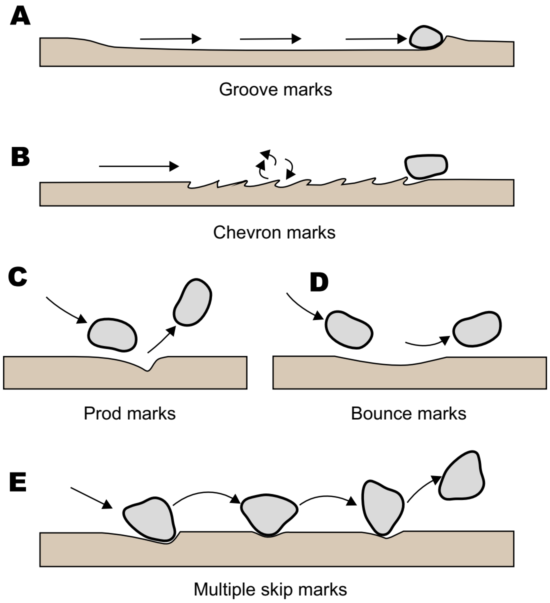


Figure 4.14 Schematic representation of the ways in which different types of tool mark are generated by different modes of behaviour of the tools.

Figure 4.15 Well-developed chevron mark on the base of a sandstone bed. Flow was from top to bottom of photo. Silurian, southern Scotland. Photo courtesy of Gilbert Kelling.

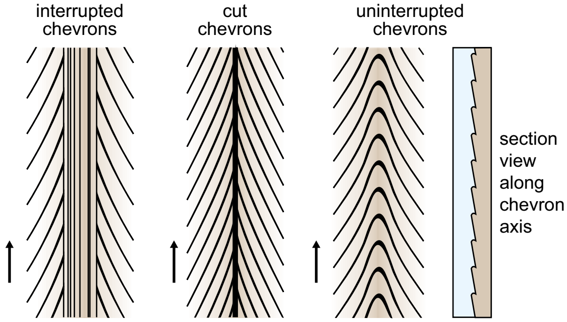


Figure 4.16 Common types of chevron marks. Modified after Craig and Walton (1962).

approach angle of the tool to the bed, so that on impact it digs more gently into the mud before lifting off again (Fig. 4.14D). No asymmetry is thereby produced and so only the trend of the current can be deduced. In most cases it is impossible to identify the tool though some exceptional examples are so distinctive that they can be related to, for example, fish vertebrae, ribbed shells or pieces of wood debris that leave a distinctive bark impression.

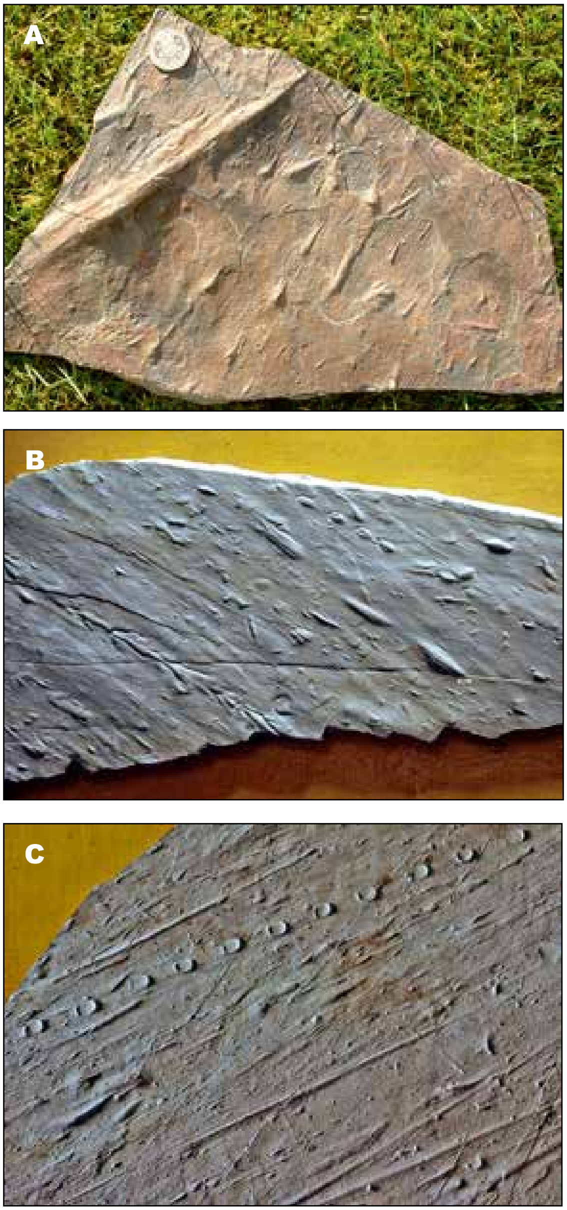
Skip marks

These are a series of genetically related bounce marks arranged linearly, usually with rather even spacing. The individual marks need not be identical, but they should be similar enough to suggest that they were produced by the same tool (Fig. 4.17C). In some cases skip marks may be very closely spaced and almost gradational with a groove.

Skip marks represent the repeated bouncing of the same tool above the bed. Differences in the shape of bounce marks can sometimes be related to rotation of the tool as it bounced along (Fig. 4.17E). Where the marks are almost continuous and approach grooves, the tool may have been disc-shaped and its behaviour analogous to that of a wheel rolling and bouncing downhill.

Distribution and association of tool marks

Tool marks generally occur in mixed assemblages with grooves and prods being found together. Grooves, being longer and larger, generally give a better measure of direction; prods, by their asymmetry, indicate the sense of the movement. In addition, tool marks are rarely seen on the same surfaces as scour marks. In many successions, particularly of turbidites, tool marks are more common on the bases of thin sandstone beds, whereas scours occur are more common on the bases of thicker beds. This suggests that tool marks represent superficial erosion by relatively weak or short-lived



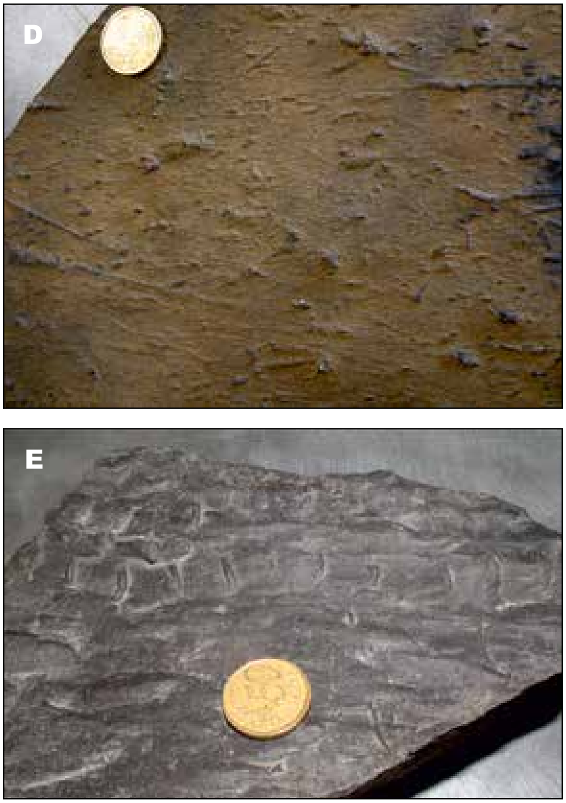


Figure 4.17 Examples of prod, bounce, skip and related tool marks revealed on the bases of sandstone beds. A) Prod marks, which are asymmetrical along their length with the deep end upstream. Location unknown. Keele University collection. B) Bounce marks, which are symmetrical along their length and generated by separate objects. Largest bounce marks are ~3cm long. Krosno Beds, Oligocene, Carpathians, Poland. C) Skip marks made by the repeated impact of the same object (probably a fish vertebra). Spacing between each skip is ~3cm. Krosno Beds. D) Prod marks likely generated by plant twigs jabbing into sediment substrate as they were carried by a flow. Locality unknown. University of Leeds collection. E) Skip marks made by the repeated impact of the same object (probably a shell). Locality unknown. University of Leeds collection.

currents, and scour marks record a more wholesale removal of the bed, possibly cutting down into more cohesive mud and suggesting stronger and more sustained currents.

4.3 Small-scale structures on modern and ancient upper surfaces

Erosional structures are fairly common on present-day sandy and muddy sediment surfaces but are rarely preserved on upper bedding surfaces in rocks. Both water and wind can erode sediment surfaces. Water erosion gives rise to obstacle and longitudinal scours, and rill marks; wind erosion leads primarily to erosional remnants, but may also etch out the internal parts of pre-existing, originally depositional structures in exposed sand.

4.3.1 Water erosion forms

Obstacle scours

The main features, and the nature and origin of obstacle scours, have been described with reference to sole marks.

They occur on both sandy and muddy surfaces and usually take the form of horseshoe-shaped troughs around a pebble, a block of ice or a shell or plant fragment, their size relating to the size of the obstacle (Fig. 4.18).

The trough is usually deepest along the upstream side or around the flanks of the obstacle and dies away downstream. On rippled sand, the scour trough and the obstacle may locally perturb the ripple pattern. If the mineralogy of the sand is varied, the scour may be accentuated by mineral sorting, notably where grain populations composed of minerals with different densities are present.

Longitudinal ridges and furrows

On flat muddy areas, especially on tidal flats, patterns of longitudinal ridges and furrows of gentle relief and variable length and spacing occur. They are usually very subtle features, best seen when looking towards the sun. They are parallel to dominant currents and are probably related to spiral patterns of secondary circulation in the water (cf. Fig. 4.11).

Rill marks

Rill marks are small-scale, dendrtic channels a few centimetres wide and are found on modern sand and silt surfaces, but rarely on bedding planes in ancient sediments (Fig. 4.19). They result from the emergence of pore water from within the sediment following a fall in water stage. They occur most commonly on beaches and on flanks of larger tidal bedforms at low tide, though they also occur on the flanks of large bedforms in rivers and at the edges of channels during low-stage flow. They are almost invariably destroyed by a rise of water level and they thus have a very low preservation potential. They have no palaeocurrent significance.

Megaflutes

Extensive upper bedding surfaces within some turbidite successions show larger-scale erosional features that have shapes similar to flutes but differ from them in several important respects. First they are found on the upper surfaces of thick sandstone beds and clearly reflect erosion of sand rather than mud. Second, the erosional relief is typically filled by mudstone, sometimes with a few thin interbedded sandstones. Third, they are much larger than ordinary flute marks, being typically several metres wide, many tens of metres long and of the order of 1 metre deep. In plan view, they have a curved margin, commonly quite sharply defined, which curves strongly around the upstream end (Fig. 4.20A). The sides of the scour are quite steep, approaching the angle of rest of sand in some cases. The lower parts of the slopes and the floors of the structures

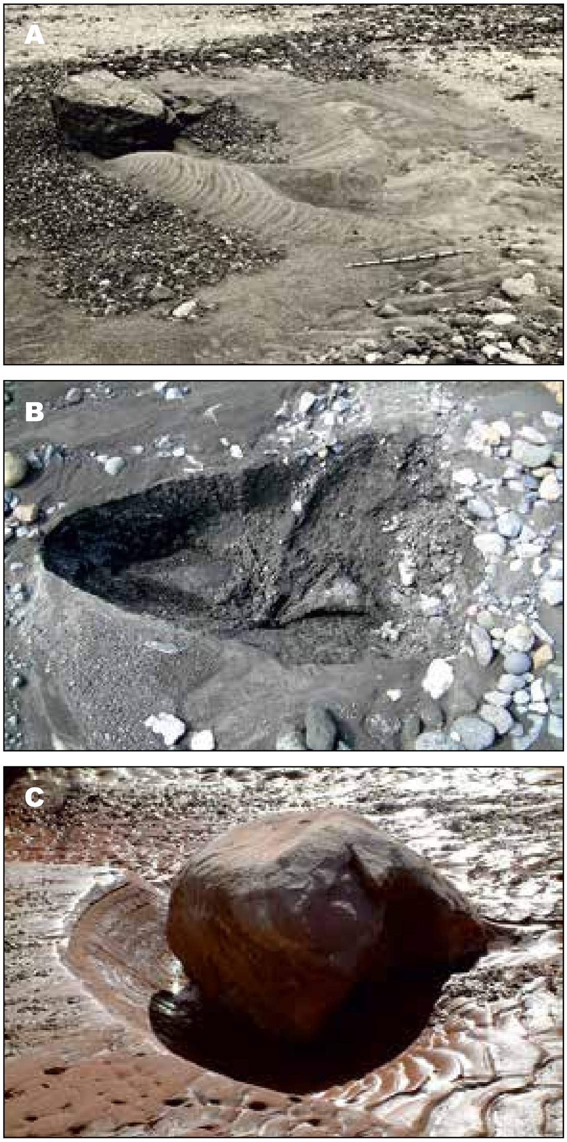


Figure 4.18 Examples of obstacle scours. A) Boulder on a river bed with associated scour around it. Note the deposition of sand ridges around the boulder, which indicate the down-current direction. Tana River, Finnmark, Norway. B) Obstacle scour on a fluvial outwash plain. In this case the obstacle was an ice block that has subsequently largely melted away. Penknife for scale. Mýrdalssandur, southern Iceland. C) A boulder with a well-defined scour trough on its upstream side, with ripples on the flanking sands. Ephemeral stream bed, Utah, USA.

Figure 4.19 Rill marks on the lower part of a sandy channel margin in an intertidal setting. The pattern of small dendritic channels is cut by water emerging from within the sand during falling and low-water stage. Tana Delta, Norway.

are commonly decorated with current ripples. Downstream the features die away gradually. In vertical section, the structures commonly have a channel-like form (see §4.4), especially when the section is at a high angle to the axis of the structure (Fig. 4.20B, C). The finer-grained fill may both onlap and drape the margins of the scour, typically repairing the structure so that bedding in the uppermost part of the fill is close to horizontal.

These rather uncommon features clearly record erosion of a sand surface by sustained fluid scour due to a current that did not deposit any sediment when it eventually waned. Instead, the erosional relief was abandoned in a quiet water setting where fine-grained sediments were deposited from suspension. The only known examples of such structures in present-day settings are on the surfaces of deep-water submarine fans where they have been mapped by deeply towed sonic imaging devices. In ancient successions, megaflutes are commonly associated with turbidite beds and are usually interpreted to have formed in deep-water submarine slope and fan settings. Where found, megaflutes help to characterize the suite of processes active in the overall environment of deposition and also act as useful palaeocurrent indicators.

4.3.2 Wind erosion forms

Strong winds blowing over damp or slightly cohesive sediment can lead to the development of erosional forms reminiscent of flutes but showing a positive relief on the upper surface. A blunt nose points upwind with a tail streaking

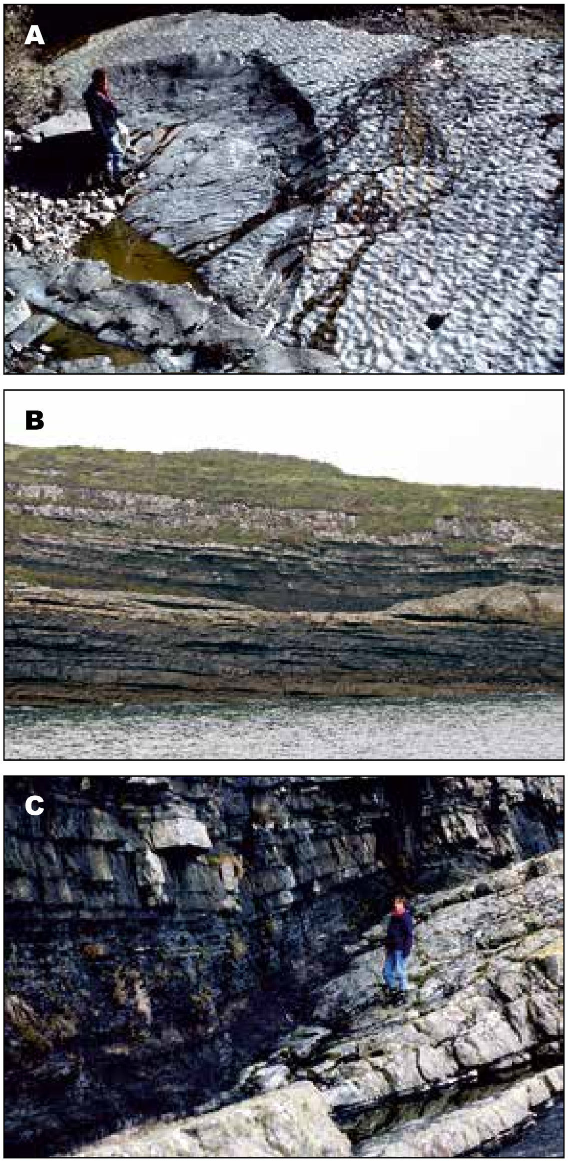


Figure 4.20 Examples of megaflutes. A) Upper bedding surface showing the strongly curved upstream end of a megaflute. Note the current ripples on both the surrounding bedding surface and on the floor of the megaflute. B) Cross-section through one flank of a megaflute. Note the massive nature of the eroded sandstone and the predominantly muddy nature of the fill of the megaflute. C) Detailed cross-section through megaflute shown in B. Cut is into thick-bedded sands and filled by draping and thickening muds. Note that the parallelism of the bedding is rapidly restored in the overlying sandstones. All from Ross Formation, Upper Carboniferous, western Ireland.

Figure 4.21 Examples of features associated with wind erosion. A) Deflation of sand grade material across a beach. Larger clasts (shells) act as obstacles, in the lee of which sand is protected from deflation. Ebro Delta, Spain. Photo courtesy of Gilbert Kelling. B) Deflation of sand grade material across an Icelandic sandur (outwash plain). Pebbles act as obstacles, in the lee of which sand is protected from deflation. Mýrdalssandur, southern Iceland. C) A wind deflation surface strewn with wind facetted cobbles. Huab Basin, Namibia. D) Wind facetted basalt boulder (ventifact) polished by sand abrasion. Askja Sandsheet, Iceland.

out downwind (Fig. 4.21A, B). Often, on modern surfaces, the erosional remnants are localized around pebbles or shell fragments. They are commonly up to a few centimetres wide and up to a few tens of centimetres long. They usually occur in groups rather than as isolated forms. They are rather uncommon in the rock record where they could be confused with flute marks, thereby providing misleading “way-up” criteria.

Where strong winds blow across loose sand surfaces they commonly strip away loose sand to generate erosional sand **deflation surfaces** (also known as **reg** or **serir**). Such surfaces may be planar, irregular and hummocky or covered with low-relief scallops arranged into regularly repeating patterns. Wind-facetted ventifacts commonly occur in large numbers strewn across low-relief sand deflation surfaces (Fig. 4.21C). Such deflation surfaces commonly have an associated lag of granules or pebbles upon them; these relatively coarse grains are the fraction of the original sediment that could not be deflated by the wind.

Another feature is that of wind-facetted pebbles and/or boulders, the faces of which have been abraded by the impact of wind-borne sand grains. Such **ventifacts** are characterized by one or several sharp ridges that separate smooth faces on individual clasts (Fig. 4.21D). The sharp ridges usually face in an upwind direction and can therefore be used to assess predominant wind direction. Ventifacts have a high preservation potential and, where observed in ancient successions, can be useful indicators of palaeowind processes and direction.

4.4 Erosional features in vertical section

The recognition that erosion has taken place during the accumulation of a sediment succession commonly depends on the occurrence of surfaces that truncate earlier lamination or bedding. On the larger scale, these features are best seen in vertical sections rather than on bedding planes. Clearly, the chances of recognizing large-scale erosional structures are much increased by large, laterally extensive exposures. In restricted exposures or in borehole core, this may be much less certain.

4.4.1 Downcutting relief

A surface that sharply truncates earlier bedding or lamination will commonly be inclined to the depositional horizontal and may be shown to be part of a larger structure if traced laterally for a sufficient distance. The form of the larger structure will depend upon the way in which the erosion took place. There are two main processes to consider when looking at any suspected erosion surface:

(a) erosion by scour creating a feature elongated in the direction of fluid movement, e.g. channels, megaflutes (see §4.3.1) or, in aeolian sediments, “blowouts”;

(b) erosion by mass movement down a slope, creating a feature of less definite shape and orientation but commonly arcuate along the slope, i.e. a slump scar.

However, the two processes can occur together. For example, a river channel, eroded mainly by fluid scour, may have slump scars as smaller-scale features on its banks.

Erosional features, of whatever origin, occur over a wide range of scales, up to hundreds of metres deep and kilometres wide. These largest forms require exceptional exposure for them to be seen in one outcrop and normally their existence has to be inferred from mapping, from the comparison of appropriately spaced, well-correlated sections, or from seismic reflection data in subsurface successions.

The erosional features most commonly recognized at outcrop usually show small- to medium-scale relief but may have a wide range of shapes, orientations and subsidiary features, all of which may be important for their correct interpretation (Fig. 4.22). Rather than trying to impose some scheme of classification, we suggest that any examples encountered should be described and measured with the following groups of questions in mind:

What is the overall three-dimensional shape of the erosion surface? Surfaces may have quite complex shapes with both flat and curved sectors and it is necessary to break the problem down in relation to more specific questions. Is the surface continuously concave-upwards or does it have a distinct base and sides? If it has sides, what is their maximum inclination? If both sides can be seen are they similar? In other words, is the cross section symmetrical or asymmetrical? Commonly, only one margin of a channel-like surface is seen and one can then only guess at the nature of the unseen margin. Is the apparent shape in the observed cross section the true shape or is it distorted by an oblique orientation of the exposure to the true cross section? Sections other than those perpendicular to a channel axis will have higher apparent width:depth ratios than the true cross section. In order to reconstruct the true width:depth ratio some means of estimating the orientation of the channel axis is needed and this is discussed below.

What are the dimensions of the erosion surface? Is it possible to measure the depth and width of channel-like forms? Remember that channels will appear wider when the exposure is oblique to the true cross section. Where exposure is incomplete, it is still valuable to record maximum observed values. Observed relief may in some cases represent the full depth of the erosional form but, in other cases, it may be only a fraction of total depth. Even a small and apparently complete channel could be superimposed on the floor of a much bigger form.

What is the orientation of the erosional form? If a channel form has been established, it will usually be important to know its orientation so that it can contribute to palaeogeographic reconstruction. If you can see a clear “channel” shape in an exposure, this tells you that the axis of the channel, or possibly a megaflute, makes a considerable angle with the face of the exposure. However, you should try to be more precise than that. Walking around outcrops, it is often possible to judge the orientations of small channels by eye with quite reasonable accuracy. This is much easier if steep channel sides are exposed. Their strike will commonly parallel the channel trend. Small-scale erosional structures superimposed on the floor or walls will also help. Erosional sole marks on a channel floor or ripples on the floor of a megaflute give a good indication of channel trend. However, the natural fluctuation of flow direction within a channel means that individual measurements may not be representative and, ideally, several measurements should be made from which a more representative mean can be calculated. In cases of extensive bedding-plane exposure, such as on a wave-cut platform or in conditions of semi-arid or desert

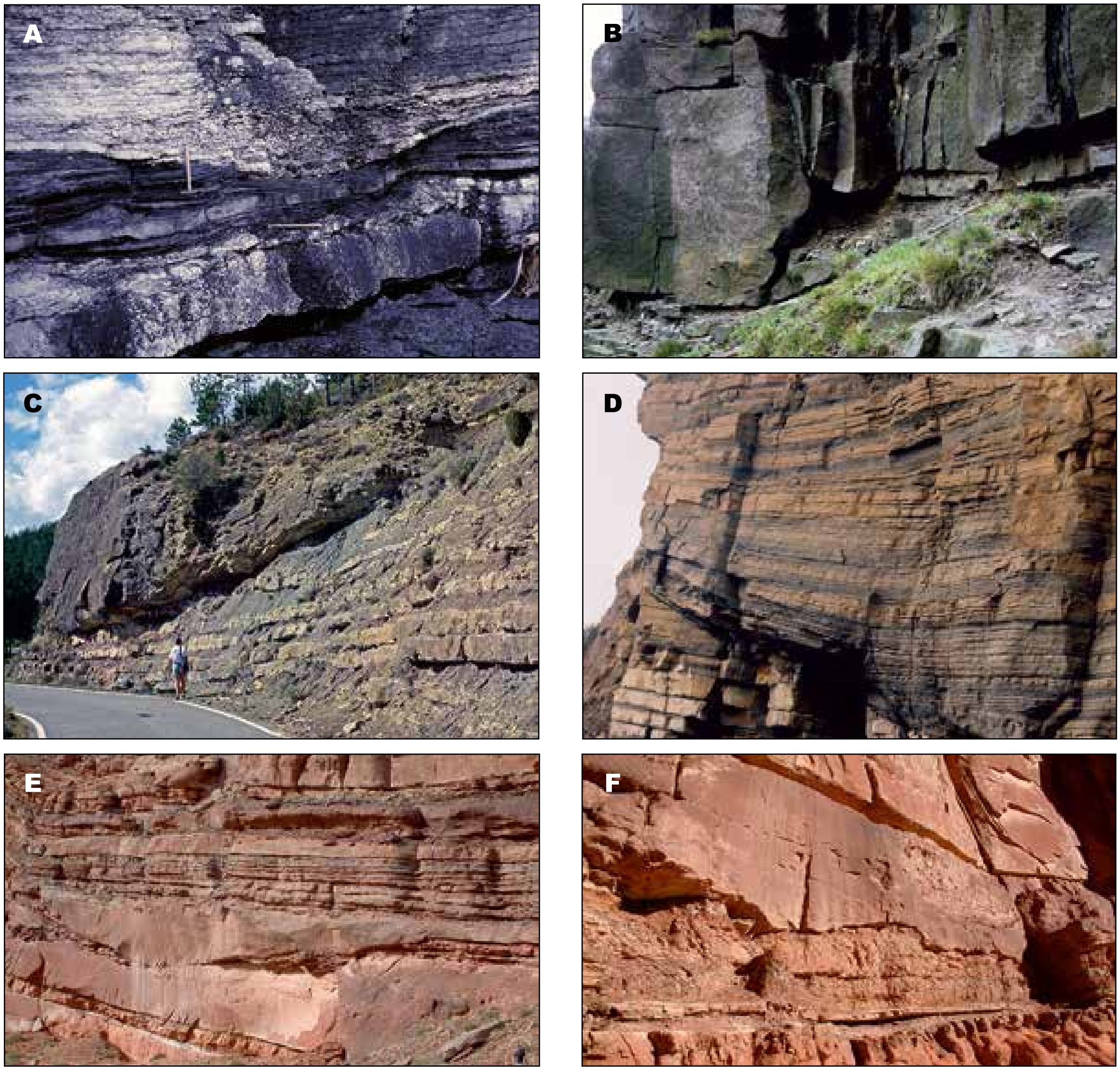


Figure 4.22 Examples of different channel margins and channel forms at a medium to large scale. A) Lower Carboniferous limestones, Anglesey, north Wales. B) Shale Grit, turbidites, Upper Carboniferous, Derbyshire, England. C) Sandstone-filled channel body. Campodarbe Group, Eocene-Oligocene, Pyrenees, Spain. D) Coal Measures, fluvial, Upper Carboniferous, south Wales. E) Lower Cutler beds, Pennsylvanian-Permian, Utah, USA. Note-erosive base, massive fill and person (circled) for scale. F) Sandstone filled fluvial channel eroding into laminated, nodular red mudstone. Lower Cutler beds, Pennsylvanian-Permian, Utah, USA. Channel fill is ~1m thick. G) Multiple stacked channel bodies filled with sandstone and separated by mudstone beds. Triassic Moenkopi Formation, Utah, USA.

weathering, it may be possible to trace the channel over long distances and thereby establish a very reliable trend. It may even be possible to judge its sinuosity. If possible, it is always worth looking down from cliff tops on to bedding surfaces if channel sand bodies are suspected (cf. Fig. 6.66A). Examination of aerial photographs can also be valuable.

4.4.2 Superimposed features on erosion surfaces

Not all channel cross sections have simple shapes, and the sides, in particular, often show subsidiary features such as steps, terraces and even overhangs. There are two main controls on the development of these features: the nature of the substrate and the history of erosion and infilling.

Steps and terraces on a channel side often relate closely to the lithology of the eroded sediment. Beds of varying composition or grain size respond differently to flowing water; some will be more readily eroded than others. More cohesive, generally fine-grained sediments commonly form more resistant features, whereas coarser, less cohesive sediment is preferentially eroded. In extreme cases, overhangs may develop and indicate either erosion of strongly cohesive or partly lithified (cemented) sediment.

Many channel forms seen in the field may have undergone a series of erosional and depositional phases. Figure 4.23 shows ways in which different sequences of "cut and fill" can produce a similar channel shape. The sequence of events can only be deduced from observing features of the channel fill as well as the bounding erosion surface. Erosion surfaces, which are obvious where there is a clear contrast between the lithologies of the substrate and the fill, will be less easily detected within the fill where the lithology is more homogeneous. An erosion surface within the fill may in some cases be a sharp parting and may commonly be associated with and accentuated by a thin conglomeratic layer of exotic or intraformational clasts (a pebble lag).

4.4.3 Problems and complications

There are at least three aspects of larger-scale erosional features that need separate discussion:

- recognition of erosion where no erosional relief is seen;
- relationship between preserved "channel" form and the instantaneous shape of the active channel;
- distinction between forms due to water scour (channels) and those formed by mass movement (slump scars).

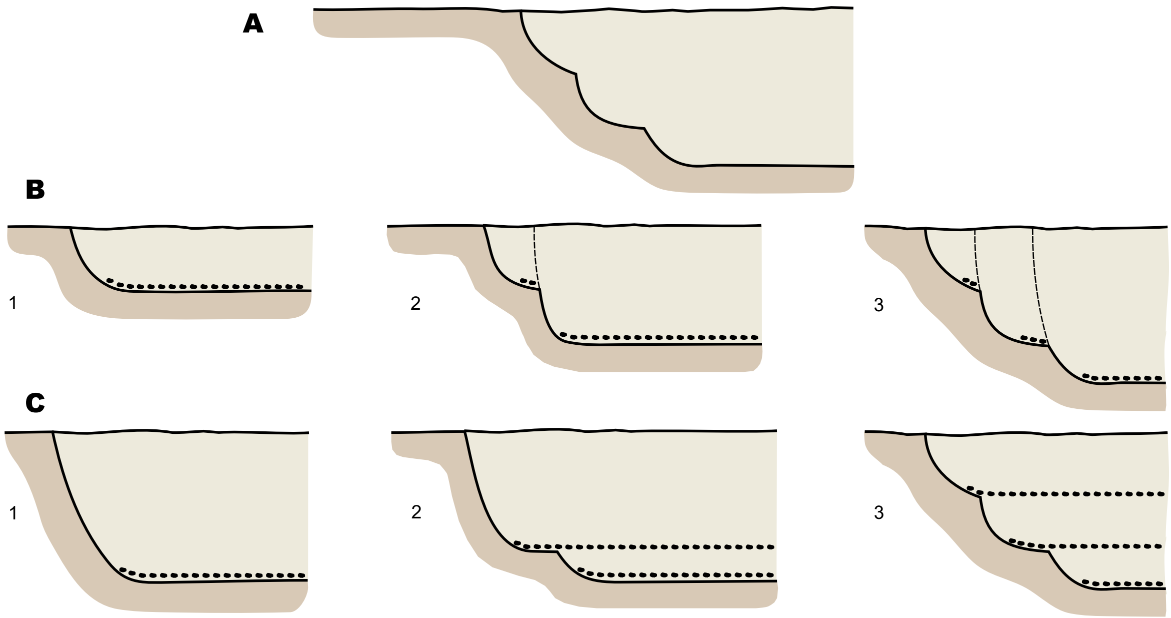


Figure 4.23 Possible stages in the erosion of a stepped channel margin by repeated episodes of cut and fill. A) Final channel shape. B) and C) Two different development histories by which the final form could be achieved. The ability to recognise erosion surfaces within the channel fill may be vital in understanding the full history of development of the channel.

Absence of distinct erosional relief

The absence of distinct relief on a surface does not necessarily mean that no erosion has taken place or even that overlying deposits are of non-channel origin. A very wide or very large channel may require an exceptionally large outcrop to establish its channel shape. An outcrop trending parallel to a channel axis may also prevent a channel-shaped cross section from being seen. However, there are clues that still may lead us to suggest a channel origin, even though none of them gives an entirely unambiguous answer.

A sandstone or conglomerate unit resting sharply on a unit of finer sediment may have been deposited in a channel. Such an explanation would be supported if a layer of coarse clasts occurs at or just above its base, particularly if the clasts are of intraformational (rip-up) origin such that they are composed of the same material that underlies the suspected channel. If the actual surface of contact is exposed, smaller-scale erosional structures such as flutes or grooves may give additional evidence of scour.

However, sheets of sand and conglomerate with slightly erosive bases and with coarse basal layers may also be deposited in non-channelized settings, for example by non-confined, sheet-like stream flows or by large turbidity currents. They may also result from widespread, non-channelized erosion due to a marine transgression whereby erosion takes place during the landward migration of a shoreline, with waves providing the erosional energy.

Preserved channel form and active channel shape

It should not be assumed that a channel shape needs to be observed in order to infer that a channelized flow was responsible for a particular rock unit. Clearly, if a channel form is seen, then a channelized flow was involved, but the absence of a channel form does not necessarily rule out channel activity during deposition. In many channel deposits, preserved channel margins are relatively rare features, as becomes apparent when you consider the medium-term behaviour of an active channel.

For the cross section of a preserved channel form to be identical to that of the active channel, the channel must have been eroded and infilled without shifting its position. This leads to the preservation of narrow "shoestring" sand bodies. If channels stay active for sustained periods, they commonly migrate laterally, possibly shifting position by several channel widths while still maintaining their cross-sectional shape. This generates an erosion surface which, although eventually ending in a channel margin, may be so laterally extensive that the chances of seeing either margin are small, and the chances of seeing both margins smaller still. The most common examples of this behaviour are provided by meandering river or tidal channels (Figs. 4.24, 4.25). Additionally, there are now several well documented examples of laterally migrating channels on submarine fans in deep-sea settings, where both high-quality outcrop and 3D seismic data have been used to demonstrate channel evolution through time.

Where the recognition of channelling depends upon criteria such as those outlined above, determination of the instantaneous shape of the active channel may be difficult or impossible. In some cases, internal features of the channel deposits above the erosion surface show lateral accretion surfaces which may suggest the shape of the channel cross section (see §6.2.10). In other cases, a laterally migrating channel body might be abruptly abandoned, for example due to cut-off via avulsion. Such abandoned channels commonly slowly fill with muddy sediments washed or blown in following abandonment. In cases where such muddy sediments contrast markedly with surrounding sediments, they form a channel "**plug**" which can reveal the geometry of the channel at the time of abandonment.

Distinguishing channels and slump scars

Not all erosional cross-cutting surfaces in rocks are the margins of channels scoured by flowing water or turbidity currents. Some result from slumping on sub-aqueous slopes, which leaves behind a slump scar, analogous to landslip back-scars on sub-aerial hillsides. The ability to distinguish slump scars from channels is important for interpreting processes and environments of deposition and

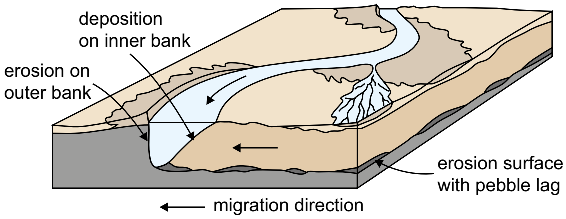


Figure 4.24 The lateral migration of a channel, in this example due to the development of meanders, may erode a near-horizontal erosion surface with little or no relief. The presence of a channel-body is inferred from the erosion surface and the nature of the infill represented by the tabular sandbody. The width of the formerly active channel can only be established where the clay- or mud-filled plug that represents abandonment is observed. See Chapters 6 (Figs. 6.60 to 6.63) and 10 (Figs. 10.7 to 10.9) for examples of the likely preserved succession. Based in part on Allen (1964).

Figure 4.25 Lateral migration of a tidal channel. The channel is migrating from right to left. The eroded bank shows blocks of cohesive, fine-grained bank material falling into the channel. The deepest part of the channel, below water, erodes a near-horizontal erosion surface as it migrates and is overlain by a sandy succession deposited on the gently sloping bank on the right. Loughor Estuary, south Wales.

for predicting the probable extent of an erosion surface and its relationship to the palaeoslope.

Slump scars are commonly broad curved features with their maximum horizontal extent along slope, whereas channels cut by currents are elongate down the slope (Fig. 4.26). In addition, all but the most strongly curved slump scars will appear as single-sided features; channels have two sides. However, partial preservation and poor exposure make it important to have other criteria for distinguishing channels and slump scars.

In vertical section slump scars are usually smooth, concave-upwards surfaces whose inclination may vary from near-vertical to near-horizontal. Sediments below the surface may show small, normal faults with a similar orientation to the surface, suggesting local horizontal stretching and secondary gravitational collapse. The surfaces lack both small-scale superimposed sole marks, such as flutes and grooves, and the steps and terraces that are common in many scoured channels. Another important criterion is the nature of the sediment above and below slump scars. Slumps typically originate due to instability of usually fine-grained sediment on a slope and they commonly move off spontaneously without any external trigger. Depositional conditions are unaltered, and later, similar sediments drape and gradually eliminate the topography of the slump scar. In contrast, the cutting of a channel implies the action of strong currents, and these will be reflected not only in the erosional surface but also in the coarser sediments that are commonly laid down above that surface. These coarser sediments may include intraformational conglomerates, which are not readily produced by slumping, and also depositional structures that reflect high-energy currents. However, if a channel is suddenly abandoned, fine sediments may infill it, making it difficult to distinguish from fine-grained bank material. With some slump scars, the mass of slumped sediment may not have moved far and a deformed, commonly chaotically bedded, slump deposit may be found close by. In other cases, a series of sub-parallel slip surfaces may occur with slices of slightly shifted but otherwise undisturbed sediment between them.

Although dealt with separately here, slumping and channel scouring quite commonly coexist. The undercutting of river banks by scour commonly leads to blocks or masses of bank material slumping into the channel and creating slump scars in the process (Fig. 4.25). The toe of the slip surface may extend below the floor of the channel. Abandonment of such a channel soon after a slump may lead to the preservation in the rock record of slump scars

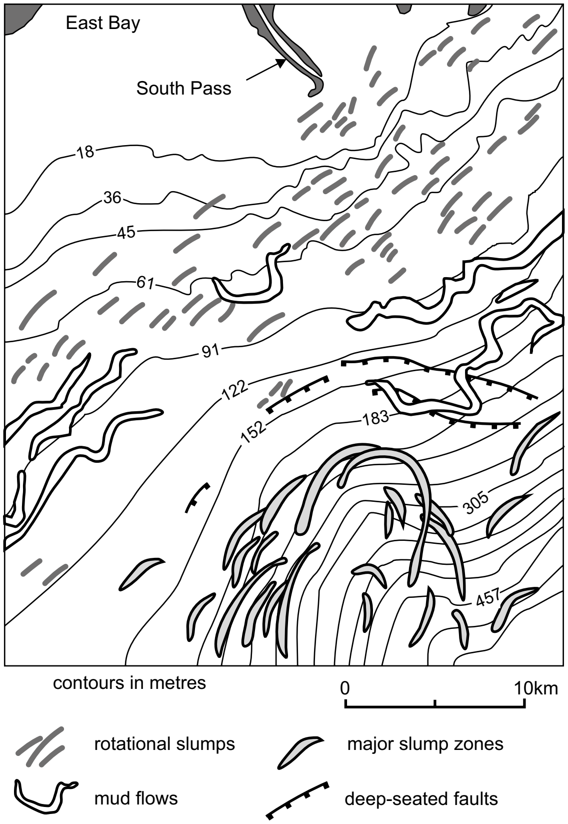


Figure 4.26 Pattern of major and minor slump scars on the slope of the present-day Mississippi Delta. Note that most of the slump scars are sub-parallel to the bathymetric contours. Modified after Roberts et al. (1976).

at the channel margin and of slumped, rotated blocks in or below the channel fill.

4.4.4 Wind erosion features in vertical section

Wind deflation surfaces and scours seen in section are characterized by either irregular, sharp-based scours with relief varying from a few centimetres to several metres (Fig 4.27A) or sharp-based, laterally extensive, planar surfaces often with associated features such as desiccation cracks, collections of wind-facetted pebbles, bioturbation or root traces extending down from the surface (Fig. 4.27B). Sand-deflation scours form where an airflow that is not fully saturated with respect to its potential sand-carrying capacity blows across a lose sandy substrate and net erosion occurs. Turbulent eddies within the airflow generate erosional scour pits. Where deflation is long-lived, the surface may be lowered until further erosion is no longer possible, either because the

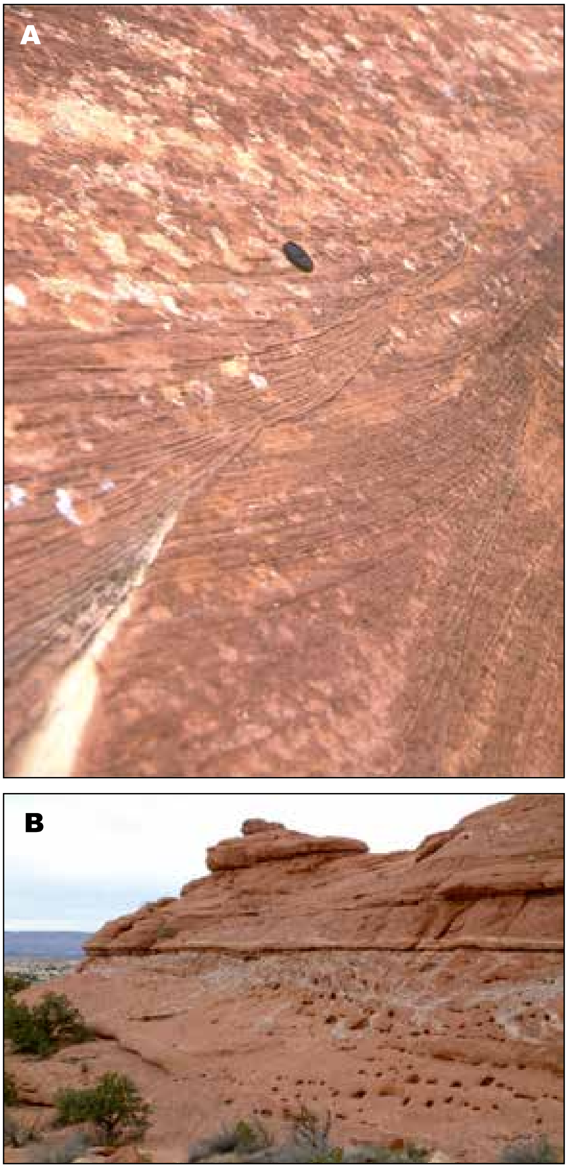


Figure 4.27 Examples of erosion surfaces in aeolian deposits. A) Erosional aeolian scour surface with cross stratified aeolian dune deposits beneath and above. B) Widespread wind deflation surface. Laterally extensive and planar aeolian deflation surface separating two aeolian dune accumulations. The white mottled horizon below the surface is a zone of rhizoliths (fossilised root traces). The surface represents a paraconformity or diastem. Both examples from the Cedar Mesa Sandstone, Permian, Utah, USA.

airflow becomes fully saturated with sediment, perhaps due to a reduction in wind speed, or because the surface is deflated down to the level of the water table thereby

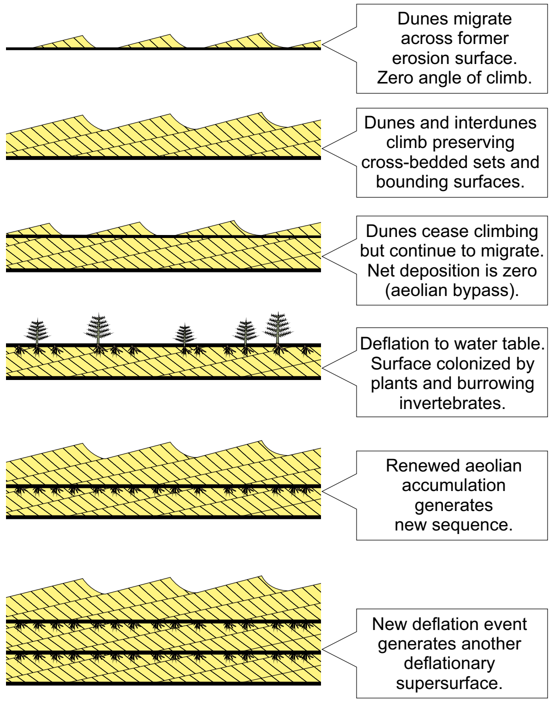


Figure 4.28 Model for the evolution of deflationary supersurfaces in aeolian systems whereby the wind becomes undersaturated with respect to its potential sand-carrying capacity and cannibalizes the existing sand surface. Sand deflation may result in the concentration of remaining larger clasts leading to the development of a coarse-grained 'armoured lag' that prevents further erosion. Alternatively, erosion may progress until the water table is reached, in which case deflationary supersurfaces may be characterised by widespread rhizolith (rooted) horizons, for example. Modified after Loope (1985).

restricting the availability of loose, dry sand for further erosion (Fig. 4.28).

Study techniques

Field experience

Present-day environments

Field programmes should include investigations of areas of interaction of erosion, transport and deposition which leave records of erosion intact. Scour marks, such as obstacle scours caused by water or wind, are observable on beaches, shallow sandy stream beds, estuarine sandflats, aeolian interdune areas and around everyday obstacles following where there has been strong winds driving loose snow. Flute marks, longitudinal ridges and furrows may be seen on cohesive mudflats. Tool marks (grooves, chevrons and prod marks) may be made by dragging an object such as a stick across a mudflat or a drying pond. Rill marks and dendritic channels are best seen on beaches. Larger channel forms are most easily examined on alluvial fans and in fluvial and intertidal areas. Aerial and satellite photographs can be useful in characterizing large channels in different types of setting.

Ancient environments

In the geological record, sole structures are most frequently displayed in the field in turbidite deposits. Channels are most commonly observed in rocks from alluvial fan, fluvial, intertidal, deltaic and some turbidite successions, where megaflutes may also exceptionally be present. Slump scars are most common in deltaic and turbidite facies. Larger erosional structures can commonly be identified by more complex investigations. These might involve measuring sections, recognizing marker beds and selecting a datum level, and then plotting successions in correlation panels. From these it may be possible to demonstrate that parts of a succession are missing, allowing a previously unremarkable bedding surface to assume wider significance as a major erosion surface.

Laboratory experience

In flumes and wind tunnels it is easy to observe the erosional features produced by pressure changes and eddies around obstacles placed in the flow. In the absence of a flume, try directing a jet of water from a hosepipe onto ground covered with a sediment mix of silt and sand grade. Place large obstacles, such as bricks, in the flow path and observe how scours develop as the silt and sand are washed around them.

Recommended references

Allen, J. R. L. 1982. *Sedimentary structures: their character and physical basis*. An encyclopaedic account of sedimentary structures and the physics of their development, as understood at the time.

Dzulynski, S. & E. K. Walton 1965. *Sedimentary features of flysch and greywackes*. In spite of its rather dated title, a well-illustrated compendium of sedimentary structures associated with turbidites, especially sole marks.

Julien, P. Y., 2010. *Erosion and sedimentation*.

Mutti, E. 1992. *Turbidite sandstones*. A lavish picture book of turbidites and their structures.

Pettijohn, F. J. & P. E. Potter 1964. *Atlas and glossary of sedimentary structures*. Beautifully illustrated with high-quality photographs of the main sedimentary structures, including sole marks.

Petts, G. &. P Calow (eds.) 1996. *River flows and channel forms.* Considers the nature of channelised flow and how channels undertake erosion.

Ricci-Lucchi, F. 1970. *Sedimentografia*. Good examples of erosional sedimentary structures.

CHAPTER 5

Depositional structures in muds, mudstones and shales

5.1 Introduction

The terminology of fine-grained siliciclastic sediments is rather confusing. A range of terms has been used in overlapping and sometimes ambiguous ways. These are discussed quite fully in most books dealing with sedimentary petrology and here we use the following loosely defined terms:

- **mud and mudstone** Unconsolidated and lithified (respectively) sediment in which grains of sand size (4φ or coarser) are absent or are an insignificant component. Where coarser grains are conspicuous the terms can be suitably qualified (e.g. sandy mud, pebbly mudstone). These terms include the more precisely defined terms silt, siltstone, clay and claystone and are useful in the field because of the difficulties of accurately judging the grain size of fine-grained sediments, especially where they have been deformed or metamorphosed.
- **silt and siltstone** These are rather more narrowly defined terms for sediments containing a dominance of grains in the range 4φ to 8φ. Rubbed against or between the teeth these sediments feel gritty. Grains are not generally visible to the naked eye, but may usually be distinguished with a lens.
- **clay and claystone** Unconsolidated and lithified (respectively) sediment where the dominant grain size is less than 8φ. Such sediments feel smooth and greasy to the touch, even between the teeth. Although many clays and claystones contain a high proportion of clay minerals (i.e. hydrated aluminosilicates), grain size rather than mineralogy is the basis of the definition.
- **shale** A widely and often loosely used field term for mudstone that often shows a conspicuous lamination and a fissility on weathering. The term is somewhat unsatisfactory in that weathering plays a part in its recognition and it cannot be consistently used in comparing rock at outcrop with, say, that of a borehole core.

Muds and mudstones are exceedingly abundant in both modern depositional environments and in the rock record, accounting for about 60% of the latter. They are derived from the products of chemical weathering of many unstable source rocks, (e.g. basic igneous rocks), and from extreme physical attrition. The fine-grained debris, produced by chemical weathering of silicate minerals other than quartz, comprises mainly clay minerals and chlorite, whereas physically derived sediment, for example in glacial "rock flour", has a mineral content dependent upon the rocks of the source area.

Although most mudstones were deposited from suspension, some may result from *in situ* weathering of unstable source material. In the latter case, the resultant soil profiles (**palaeosols**), where found within a rock succession, may be associated with depositional breaks or unconformities. Other mudstones may result from resedimentation as mudflows of original suspension muds (Figs. 5.1, 5.2). In many cases, this movement leads to the incorporation of coarser grains, which tend to "float"

Figure 5.1 A highly fluidised mud due to high pore-water content. Note the water-escape features formed where pipes carrying fluidised mud and emanating fromed deeper layers reach the surface. Note also the small rill marks around the margins of the depression. Modern, Jökulsá á Fjöllum, Iceland.

Figure 5.2 A small active mudflow where water-saturated muds have been re-mobilised through the addition of water. Note that the surface of the flow is highly irregular due to small aggregates and pebbles being rafted along with the flow. Present-day, Kong Karls Land, Svalbard.

within the predominantly muddy sediment (cf. Fig. 5.2) (see also §3.7.1).

In addition, fine-grained sediments are generated directly by explosive volcanic activity resulting in both **airfall** and **water-lain tuffs**, which may be subsequently reworked by currents or as mass flows (**lahars**). Such volcanic deposits are often recognized by their distinctive colour and weathering state. Confirmation of volcanic origin commonly requires laboratory analysis of clay minerals. High volcanic eruption columns (tens of kilometres in so-called Plinian eruptions) give very widespread sheets of **ash** through pyroclastic fall. After settling from the stratosphere, widespread distribution is achieved by winds in the upper atmosphere. Material may be transported worldwide; the most powerful processes may give rise to extremely thin but laterally widespread horizons in the geological record, which are useful for correlation and dating purposes. However, fine ash may also fall close to the volcanic centre as a result of a weak explosion or due to rain flushing grains from the eruption cloud. In the latter case, fine ash may occur as **accretionary lapilli**. Bed thickness will be controlled by the pattern of rainfall rather than by distance from the vent.

Many muds and mudstones are also rich in organic matter, which occurs either as finely divided organic (most commonly algal) debris or as organic molecules chemically attached to the clay mineral particles.

It is difficult to interpret the physical conditions of deposition of muds and mudstones in as much detail as those of coarser-grained sediments (Chs. 6 &, 7). There are two main reasons for this. First, the range of physical processes that operate during deposition of muds is more restricted. Second, fine-grained sediments, particularly those rich in clay minerals and organic matter, have a much higher initial porosity than most coarse-grained sediments and this makes them highly susceptible to compaction on burial. This typically distorts and compresses any depositional and organic structures, in some cases to the point where they are completely obliterated. The amount of compaction will vary with the composition of the sediment and with its burial history. Although some carbonate muds appear to have suffered little compaction, it is not uncommon for some clay- or organic-rich mudstones to have been compacted to a quarter or even an eighth of their initial depositional thickness. This effect can be observed by study of the internal structure of concretions that formed soon after deposition of the mud (see §9.3.1). Early-formed carbonate-cemented concretions may preserve relatively non-compacted depositional structures as well as uncrushed fossils. If concretions are present in a mudstone succession, it is always worth examining their internal structure as this may help in understanding the deposition of the mud (Fig. 5.3).

Tectonic movements have much more drastic effects on fine-grained sediments than on coarser ones. During folding, fine-grained sediments generally behave in an incompetent manner and also readily develop cleavage through rotation and recrystallization of clay minerals, thus obscuring and distorting depositional structures and fabrics. Where cleavage development has largely overprinted primary bedding features, colour changes may remain that serve as a proxy by which the attitude and thickness of the original bedding may be discerned (Fig. 5.4). Sedimentologically useful structures will be much more commonly found on cleavage planes than in sections perpendicular to them. The distortion of structures of known or assumed original

Figure 5.3 Concretions in mudstone. If they form early enough, they can preserve lamination in an uncompacted state. Upper Carboniferous, Pembrokeshire, south Wales. See Chapter 9 for discussion of concretions.

Figure 5.4 Cleaved mudstones interbedded with paler fine sandstones. The cleavage within the mudstone, which has formed at an angle to the bedding, is a tectonic foliation that has been superimposed on, and has obscured, the original depositional lamination. Old Red Sandstone, Devonian, Pembrokeshire, south Wales.

shapes in the cleaved mudstone may be used to estimate tectonic strain. For example, elliptical reduction spots are commonly assumed to have been circular (i.e. spherical) prior to deformation.

5.2 Structures and lamination

5.2.1 Detection of lamination

Cut and varnished slabs of fresh rock or naturally polished sections on coastal cliffs and foreshores or in stream beds provide the best opportunities for observing structures in mudstones. Structures are usually of small scale and are described in terms of different types of lamination. These types are intergradational and are described below under headings that suggest potentially useful criteria. The detection of grain-size differences in fine-grained sediments is usually based on differences in colour as the grains themselves are not normally visible. As a general rule, lighter colours indicate coarser-grained sediment, but there are cases where the opposite is true.

5.2.2 Very fine lamination and fissility

Very thin parallel lamination, which leads to fissility on weathering, is usually confined to claystones or to micaceous siltstones. On freshly cut surfaces perpendicular to the lamination it is usually impossible to see any colour banding that may reflect grain-size differences. The surfaces parallel to the fissility are commonly smooth and flat. When describing these mudstones, it is helpful to try to judge whether the rock will only split down to layers of a particular thickness or whether, given appropriate equipment and patience, it appears possible to go on splitting it indefinitely. If there seems to be a limiting thickness, it should be measured and recorded even though it must be accepted that fissility is a function of weathering history as well as being an intrinsic property of the rock. In splitting the rock, try to see if the surfaces of splitting correspond to mica- or organic-rich layers. The term **paper laminated** is sometimes used to describe shales that can be split apparently indefinitely.

The lack of any obvious grain-size differences in very-fine-grained fissile claystones suggests that grain orientation is responsible for the fissility. Clay minerals, chlorites and micas commonly occur as platy grains which, on compaction, are squeezed into a parallel orientation. Fine clay particles are carried in suspension by water and it requires a reduction in the level of turbulence for the grains to be deposited. This is usually achieved when a flow carrying a load of suspended sediment slows down on entering quieter water. In many cases, settling of clays from suspension is aided by a change in the salinity of the water as it enters the depositional basin, such as when a river enters the sea. In estuaries and other marginal marine settings, the higher salinity allows small clay particles to form aggregates known as **flocs** by a process of **flocculation**. Flocs are much larger than their constituent particles and they tend to settle out more quickly. The extent of flocculation is a function of particle concentration, fluid turbulence and of the chemistry of the particles and the receiving basin water. Highly turbulent conditions, such as wave agitation, tend to break up the flocs.

5.2.3 Fine lamination with grain-size differences

Close examination of artificially or naturally polished surfaces of some mudstones often reveals colour banding of paler and darker layers of the order of 1mm or less in thickness (Fig. 5.5A). This normally reflects slight grain-size differences that can sometimes be detected by close examination with a hand lens. If it is possible to see any differences in grain size, the coarser layers at least must fall within the silt-size class, and in such a case the individual layers are likely to be only a few grains thick. Try always to record the thickness of the laminae and to judge their lateral continuity and parallelism. With such thin layers, thickness is perhaps best indicated by quoting an average, calculated by counting the number of layers within a measured thickness, rather than by measuring thicknesses of individual laminae. Parallelism and continuity of lamination can be quite variable. Some examples show extreme continuity and others have laminae that pinch out laterally.

Once it is established that laminae are defined by grain-size differences, it follows that the processes responsible for deposition must have fluctuated (e.g., current strength), although it is typically difficult to estimate the time scale of this.

Two possible depositional processes must be considered. The first is that the sediment settles from suspension from the whole water column or from lighter turbid water floating near the surface. Fluctuation of the supply of suspended sediment will then give rise to the lamination. The second is that the coarser layers are the product of weak, dilute density currents flowing close to the bed, whereas the finer layers record the background settling of sediment from the water column above. Where the laminae are very thin, other features of the overall succession must be assessed. For example, if laminated mudstone occurs in a succession that has evidence of larger-scale density currents in the form of turbidite sandstones, the inference that the lamination in the mudstones is due to short-lived, weak density currents may be more reasonable.

Some fine lamination could also result from shorter-term fluctuations in more sustained currents. Sweep and burst processes in a viscous sub-layer (see §3.2.4) may sort sediment into coarser and finer layers, particularly in the silt-size range.

5.2.4 Thicker lamination or thin bedding with gradational boundaries

Many mudstones have a distinct "striped" appearance of alternating lighter and darker layers from a millimetre up to a few centimetres thick. Such layering is usually dominated by silt-grade material, although darker, finer-grained

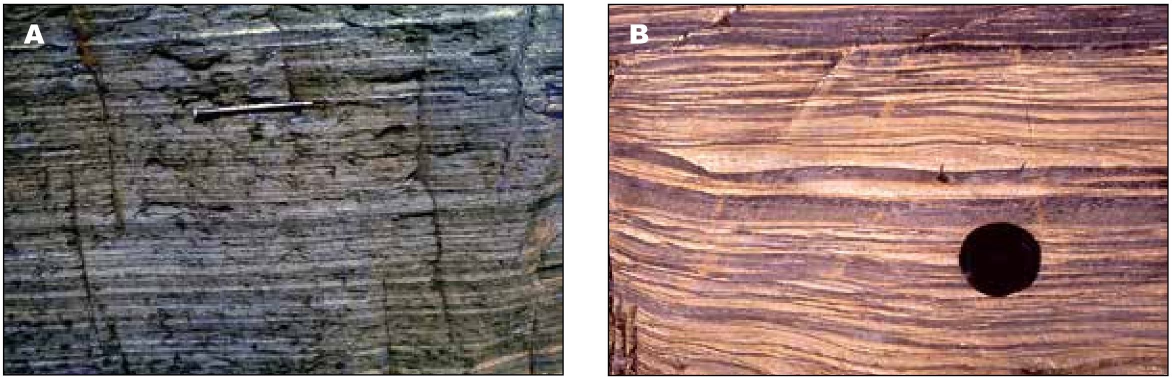


Figure 5.5 Examples of striped siltstones. A) Parallel lamination or thin bedding results from gradational grain-size changes suggesting long-term fluctuations in sediment load. B) Slightly finer and slightly coarser-grained siltstones are thinly interlaminated, probably as a result of regular (possibly seasonal) fluctuations in the suspended sediment load carried into the basin. Both examples from the Bishopston Mudstone Formation, Upper Carboniferous, Pembrokeshire, south Wales.

layers may have a substantial clay content, and paler, coarser-grained layers may contain very fine or even fine sand. It is important in such cases to try to check whether the beds or laminae are controlled mainly by overall grain size or whether they reflect fluctuations in a component such as mica, comminuted plant debris or siliceous microfossils such as diatoms or radiolaria.

These units are usually parallel-sided with gradational boundaries (Fig. 5.5B). Layer thickness can be estimated by counting layers over a measured interval or, in more detail, by measuring each layer. The second approach is important if the coarser and finer layers differ in thickness and also if it is suspected that some overall trend or pattern of thickness change occurs within the vertical sequence. In some instances, laminae may be arranged into regularly repeating groups (**rhythmites**) that occur in a specific order and which exhibit a characteristic thickness. **Couplets** are repeating pairs of laminae within a succession, whereas **triplets** consist of arrangements of three distinct laminae types. Sometimes cyclically repeating groups of laminae forming couplets or triplets may occur nested within larger repeating cycles.

These mudstones reflect fluctuations in the suspended sediment supply on a time scale too long to be attributed to sweep and burst mechanisms or to other short-period fluctuations in a statistically steady flow. Seasonal or other climatic factors may control sediment discharge to deltas, lakes and river basins where fine-grained sediments are common. The gradational contacts suggest gradually increasing and waning high discharge episodes rather than sudden "events", such as slump-triggered turbidity currents. It is not typically possible to tell if suspended sediment settled from the whole water column, from a floating plume or from a fluctuating but perhaps permanent density underflow. Alternating sets of laminae with gradational boundaries are particularly common in quiet lake environments where seasonal climatic variations control water and sediment influx. For example, lakes in proglacial settings commonly receive most of their sediment during summer months when rates of glacial meltwater influx are high, whereas during winter months, when surface waters are frozen, reduced rates of sedimentation occur from suspension settling. This example of seasonally-controlled sedimentation gives rise to **varves**, which form a particular type of rhythmite succession. Varves can also form through other mechanisms, for example climatically controlled, seasonal variations in the rate of production of organic matter. In larger lakes, seas and oceans, deep-water, bottom-hugging currents, which typically flow parallel to the base of slopes, act as effective mechanisms for the transport of fine-grained sediments and give rise to characteristic laminated **contourite** deposits.

5.2.5 Thin bedding with sharp-based, graded beds

Mudstones with sharply differentiated dark and pale layers are commonly characterized by the coarser-grained, paler layers having sharp bases and gradational tops (Fig. 5.6). The thicknesses of both the dark and pale layers are more varied than in gradationally striped mudstones. In these mudstones, beds tend to be laterally continuous and the coarser layers may show grain-size grading. Even if this cannot be directly observed, it may be inferred from the gradational tops of the beds. Bases of the coarser layers may sometimes be slightly irregular with relief of a few millimetres.

The sharp base and clear definition of the coarser layers suggest that they represent relatively sudden events superimposed upon the background of quieter, more constant sedimentation of the finer-grained, darker layers. The internal grading and the gradational tops of coarse-grained layers suggest waning of the suspended load during the more active episodes. The small-scale irregular morphology on their bases is probably mainly due to loading of the silts into soft, waterlogged clays (see §9.2.1), though minor erosive sole marks may also occur.

5.2.6 Structureless mudstones

Some mudstones show no obvious lamination, bedding or fissility irrespective of their weathering state. In some

Figure 5.6 Interlaminated mudstones and fine sandstones. The thin sandstones are sharp-based and have an internal lamination related to Bouma Sequence bedforms (see §6.7.4.) and a slightly lenticular form due to ripples on their upper surfaces. The sandstones are deposits of episodic, high-energy events. The mudstones between record quiet background sedimentation from suspension. Bude Formation, Upper Carboniferous, north Cornwall, England. Photo courtesy of Gilbert Kelling.

cases a rather blocky pattern of fracture is evident, but in others the sediment is completely massive and homogeneous, even to the point of breaking with a conchoidal fracture. Typically there is little to describe in these rocks, but it is still worth looking carefully at them as their lack of lamination may be due to one of several possible causes. It may reflect an original lack of depositional layering (i.e. continual steady deposition) or it could be due to later destruction of layering. On fresh surfaces of some apparently homogeneous mudstones it may be possible to see mud clasts in a similar mud matrix.

Lack of original layering in water-lain mudstones may be due to very homogeneous and possibly rather rapid deposition or to a lack of platy particles. Rapid deposition of muds from suspension is probably not uncommon; however direct evidence for it in the rock record is quite rare. Preservation of tree trunks in an upright growth position in some coal measure mudstones is one of the more compelling pieces of evidence. Sub-aerially deposited muds also commonly lack lamination. Thick and extensive accumulations of wind-blown silt (**loess**), which typify many proglacial areas, are examples of this. A mud that has been deposited as a mudflow may also lack structure if the large clasts that typically characterize mudflow deposits were unavailable in the source area (see §7.4.3). Careful examination of suspected mudflow deposits may reveal patches of poorly defined but highly deformed lamination that resisted homogenization.

Original layering may be destroyed later by a variety of agencies such as burrowing organisms, plant roots, evaporitic and soil-forming processes, although it is then in some cases possible to find recognizable burrow forms (see §9.4) or the remains of rootlets as in mudstone seat-earths (fireclays). Vertical colour variation or the development of concretions may support the idea of soil-forming processes.

Large-scale post-depositional movement of thick mud beds is common, particularly in rapidly deposited successions giving rise to slump deformation (see §9.2.2). Muds rapidly buried under younger sediments commonly flow both vertically and laterally to give diapiric structures (see §9.2.3), and this can lead to the development of massive, blocky or scaly fabrics.

5.2.7 Mud drapes

In certain sedimentary successions it is common to find very fine mud or mudstone closely interlaminated with very clean, well-sorted sand or sandstone (Fig. 5.7). The proportion of the two sharply contrasting lithologies varies from almost pure mud to almost pure sand but they each occur clearly and discretely with no mixing of materials. Where mud dominates, the sand occurs as thin laminae and lenses whereas where sand dominates, the mud occurs as continuous or discontinuous laminae. In many cases, the mud occurs as draping layers over sand ripple forms or on the foresets of cross-bedding within the sand.

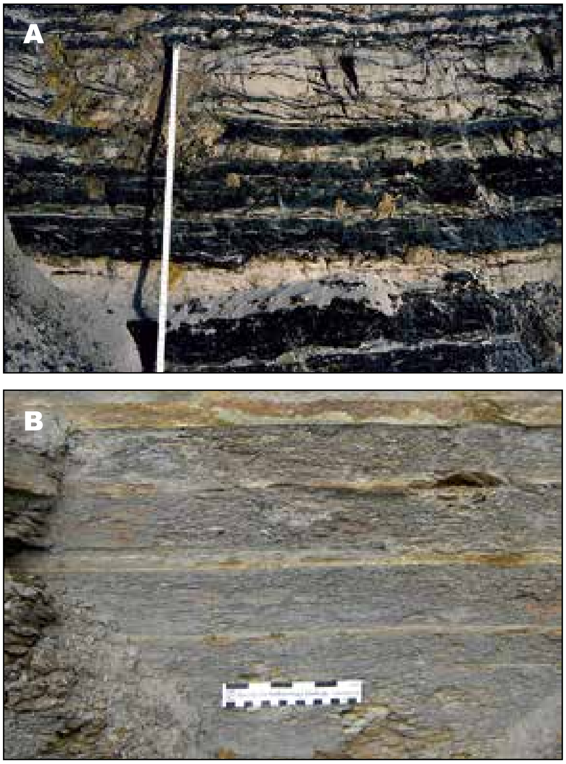


Figure 5.7 Examples of interlaminated mudstone and sandstone. A) The two components are very clearly different, fine mud and clean well-sorted sand. In the mud-dominated part: the sand occurs as thin laminae and lenses, whereas in the sand-dominated part the mud occurs as thin but discrete drapes. These patterns are typical of tidally influenced sediments, where the sand records the strong currents of the flooding and ebbing tide, and the mud records the quiet, slack-water periods when fine sediment can fall from suspension, probably aided by flocculation. Lower Greensand Group, Lower Cretaceous, Isle of Wight, England, B) Alternating thicker beds of mudstone and thinner beds of very fine-grained sandstone. The mudstone beds record settling of fine-grained sediment from suspension, whereas the sandstone beds record the deposition from traction load as weak flows carrying sediment passed over the bed as they slowed down. Edale Shale, Carboniferous, Derbyshire, England.

Such a clear separation of the products of high energy (rippled or cross-bedded sand) and very low energy (mud) is typically associated with a tidal regime wherein the sand represents the products of high energy ebb and flood currents whereas the mud represents the standstill between these higher energy episodes. The association is considered further in Chapter 6.

Study techniques

Field experience

Present-day environments

Study of present-day areas of deposition of mud can commonly be unrewarding, for suspended clay often obscures the depositional surface when the area is covered by water. After-the-event observation of muddy intertidal areas is often useful, but the dangers inherent in attempting to traverse mudflats and the physical effort of squelching through them must not be forgotten. Muddy density flows can often be generated at the edges of still, clear ponds by disturbing small masses of sediment at their edges. Mudflows can commonly be seen in cliffs and excavations in muddy sediments. They should be approached with caution, especially after heavy rain.

Ancient successions

Shales and mudstones from many environmental origins are commonly encountered during field excursions. Interpretation of the processes of origin of these rocks is typically limited. Ascribing a palaeoenvironment to the sediments often depends upon evidence of body and trace fossils, upon features in mudstones described in other chapters (e.g. post-depositional structures, Ch. 9) and upon the structures and features of interbedded sediments.

Laboratory experience

Useful experience may be gained by introducing mud suspensions and small amounts of somewhat coarser grain sizes into long settling tubes, so as to produce laminations and normally graded laminations. Attention should be given to controlling and measuring the effects of variables: the medium (air, water or even glycerine), grains size, temperature, viscosity, relative density of medium, relative density of grains, velocity of fall, etc.

Recommended references

Millot, G. 2014. *Geology of clays*. Weathering, sedimentology, geochemistry. A standard reference on the topic, and still very useful.

O'Brien, N. R., K. Nakazawa. & S. Tokuhashiu 1980. *Use of clay fabric to distinguish turbiditic and hemipelagic siltstones and silts.* An important attempt to interpret the deposition of clays in terms of process.

Potter, P. E., J. B. Maynard & W. A. Pryor 1980. *Sedimentology of shale*. A useful source book.

Stow, D. A. V. & D. J. W. Piper (eds.) 1984. *Fine grained sediments; deep-water processes and facies*. A valuable compilation related to a particular depositional context.

CHAPTER 6

Depositional structures of sands and sandstones

Structures developed in siliciclastic or carbonate sands and in sandstones and calcarenites reflect a variety of transport and depositional processes and they are our clearest indicators of the types and strengths of currents that move and deposit sediment. The transporting medium may be water or air. In this chapter, structures that results from aqueous transport and deposition are considered first, followed by an account of structures due to aeolian processes.

Deposition of sand by aqueous currents generally occurs through accumulation from bedload transport during steady or waning flow with excess sediment supply, or by fall-out from suspension from powerful, decelerating currents. Following deposition from suspension, sand may continue to move as bedload before it finally comes to rest. To classify structures in sand we have adopted a scheme that is partly descriptive and partly interpretive. Sand-size sediment of pyroclastic origin may also form many of the structures described in this chapter. Most pyroclastic deposits, however, are coarser-grained and are described in Chapter 7.

6.1 Aqueous ripples and cross lamination

6.1.1 Introduction

Ripples are relatively regularly spaced undulations on a sand surface or on a sandstone bedding plane. Their spacing (wavelength) is usually less than 0.5m and relief seldom exceeds 3cm. Bedforms with larger dimensions are referred to as **dunes** or **sandwaves** (see §6.2). Ripples show a wide variety of shapes, many of which relate to particular sedimentary processes and hence are useful in interpreting conditions of deposition.

Cross lamination is the pattern of internal lamination that develops within sand deposited by ripple migration. It can be seen both on bedding planes and on vertical surfaces. Some patterns of cross lamination are specific to particular types of ripple and so can aid interpretation.

6.1.2 Material

Although ripples and cross lamination are principally features of sand-grade sediment, they also occur in coarse silts. They are most common in fine- to medium-grained sand and are rare in material coarser than coarse sand, except where they are the result of wave action or of strong winds.

6.1.3 Ripple morphology

Ripples are characterized in terms of both their profile and their plan view (Fig. 6.1). The important distinction between symmetrical and asymmetrical ripples is based on their profile perpendicular to the crestline. Although there is some truth in the generalization that ripples with symmetrical profiles are the product of wave action and those with strongly asymmetrical profiles are due to current activity, the reality is rather more complex. The shape and continuity of ripple crestlines is at least as important for interpretation.

A broad range of patterns is seen in the rock record and on present-day beaches, river beds and tidal flats. Detailed measurement and description of ripple morphology can be very informative and should always be attempted in any serious study. Basic dimensions can be measured and their values combined to yield indices that point towards the dominant process, even if interpretation may still be ambiguous (Fig. 6.2).

The relationship between profile symmetry and crestline continuity and curvature is complicated. Whereas symmetrical ripples commonly have straight and rather continuous crests (Fig. 6.3), not all straight or continuously crested ripples are symmetrical. Some straight-crested ripples show a marked asymmetry (Fig. 6.4).

Ripples with highly sinuous crests (Fig. 6.5A) and those with a strongly three-dimensional plan-form shape, such as linguoid ripples (Fig. 6.5C), usually have asymmetrical profiles. They most commonly have steeper, concave-upwards lee faces and more gently sloping convex-upwards stoss sides. Such ripples result from currents flowing in one direction only (unidirectional). There is, however, a continuum of asymmetrical current ripples ranging from straight crested through sinuous crested to linguoid in plan-form shape (Fig. 6.5). Associated ridges and hollows on the stoss sides of ripples are aligned roughly parallel to flow. Scour pits may occur in the lee of ripples, commonly in front of a downstream embayment in the crestline or downstream of the gap between two linguoid ripples.

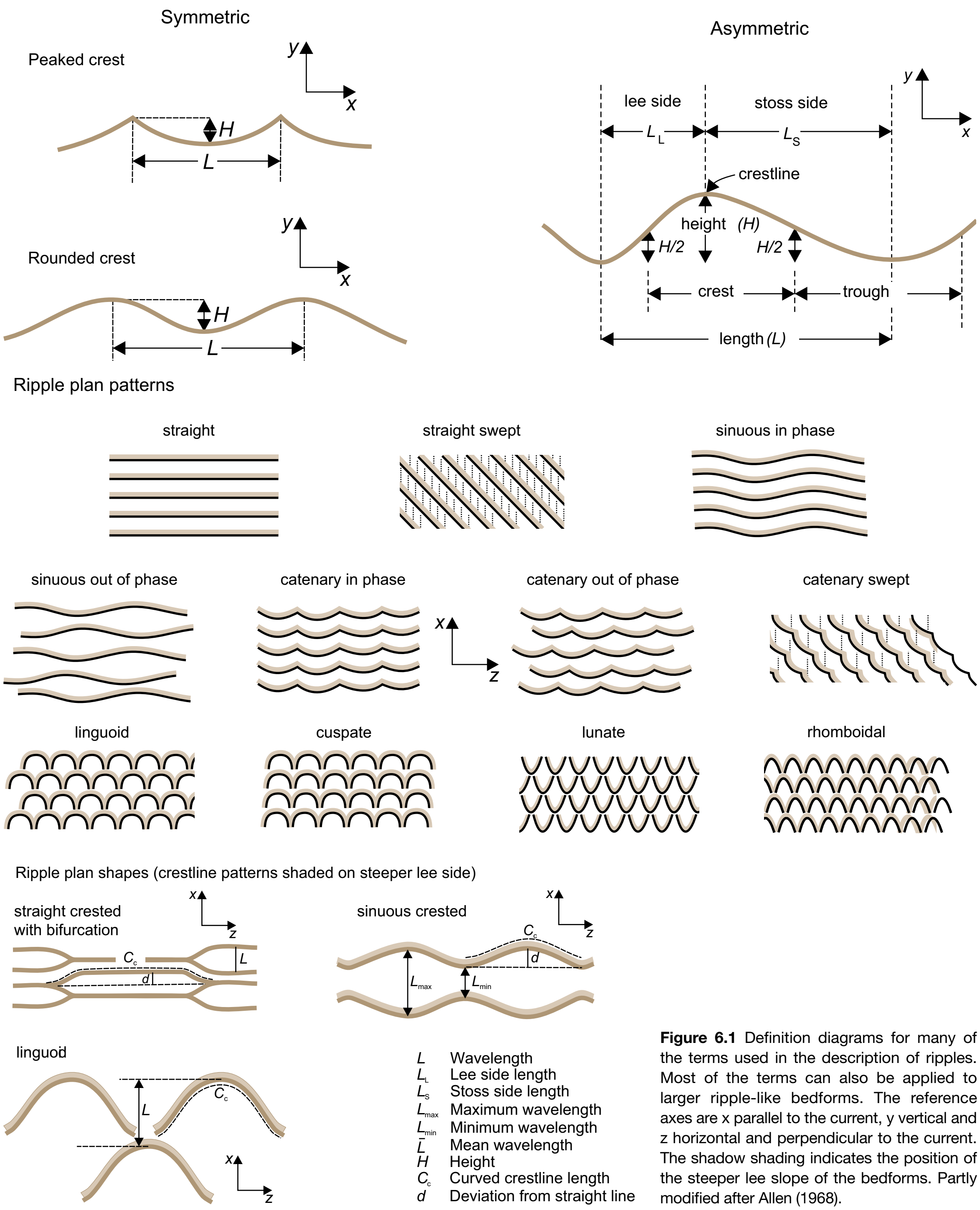


Figure 6.1 Definition diagrams for many of the terms used in the description of ripples. Most of the terms can also be applied to larger ripple-like bedforms. The reference axes are x parallel to the current, y vertical and z horizontal and perpendicular to the current. The shadow shading indicates the position of the steeper lee slope of the bedforms. Partly modified after Allen (1968).

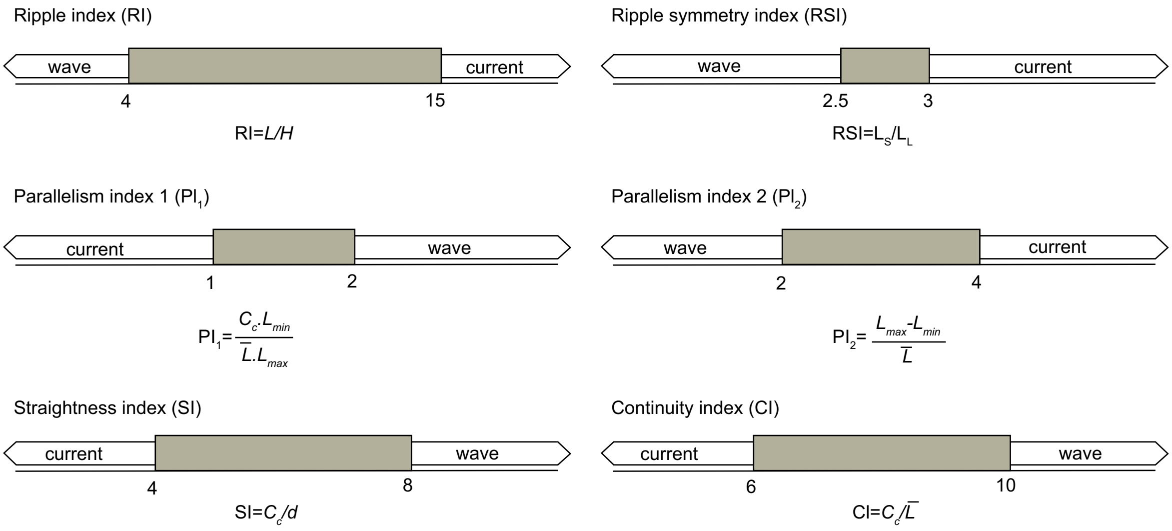


Figure 6.2 Simple ripple indices and their use as a means to discriminate between wave and current activity. The shaded areas indicate the range of values over which the indices are overlapping and non-discriminatory. See Fig. 6.1 for further explanation of the parameters.

Figure 6.3 Rounded, straight-crested symmetrical ripples, Tana delta, Finnmark, Norway.

Figure 6.4 Asymmetrical, straight-crested ripples, Kirkland, Scotland. Also note worm casts and crab tracks.

On some beaches, ripple forms with low relief have very marked repeated rhomboidal plan-form shapes that give a fish-scale pattern to the sediment surface. These **rhomboid ripples** are elongate parallel with the current (usually the backwash of waves on a beach) having spacings of the order of a few tens of centimetres and heights of less than 1cm. They are asymmetrical in profile, being highest at their downstream point. Examples in the rock record are rare.

Symmetrical ripples and those with very straight and continuous crestlines (Fig. 6.3) are associated with wave action. They generally lack scour pits and their crestlines show zig-zag junctions or so-called tuning-fork bifurcation (Fig. 6.6B). They may be either smoothly rounded (Fig. 6.5C) or quite sharply peaked in profile (Fig. 6.6E). More complex wave ripples show multiple crests (Fig. 6.6A). Others show flattened tops or small steps on their sides, usually produced by shallowing or emergence.

It is important to try to distinguish between current and wave ripples on the basis of symmetry and crestline continuity, but one should not expect complete success. Ripples with straight crestlines but with marked asymmetry may be caused by shoaling waves or by an interaction of waves and currents of similar direction (Fig. 6.4, see §6.1.5).

Figure 6.5 Examples of asymmetrical ripples with various morphologies. A) Strongly linguoid asymmetrical ripples. Sand with mud drape. Ephemeral stream bed, Utah, USA. B) Straight to sinuous-crested asymmetrical ripples, Tana River, Finnmark, Norway. C) Sinuous crested out-of-phase asymmetrical ripples, Mýrdalssandar, southern Iceland. D) Straight-crested asymmetric ripples with tuning-fork crest junction. Dee Estuary, England. E) Slightly sinuous crested ripple forms with weak asymmetry preserved on an upper bedding surface Ripple spacing is ~12cm. Moenkopi Formation, Triassic, Utah, USA. F) Linguoid asymmetrical ripple forms on bedding surface. Kayenta Formation, Jurassic, Utah, USA. G) Linguoid ripples on top of a thick turbidite sandstone. Ross Formation. Upper Carboniferous, western Ireland. H) Paired linguoid ripples, Scalby Formation, Middle Jurassic. east Yorkshire, England.

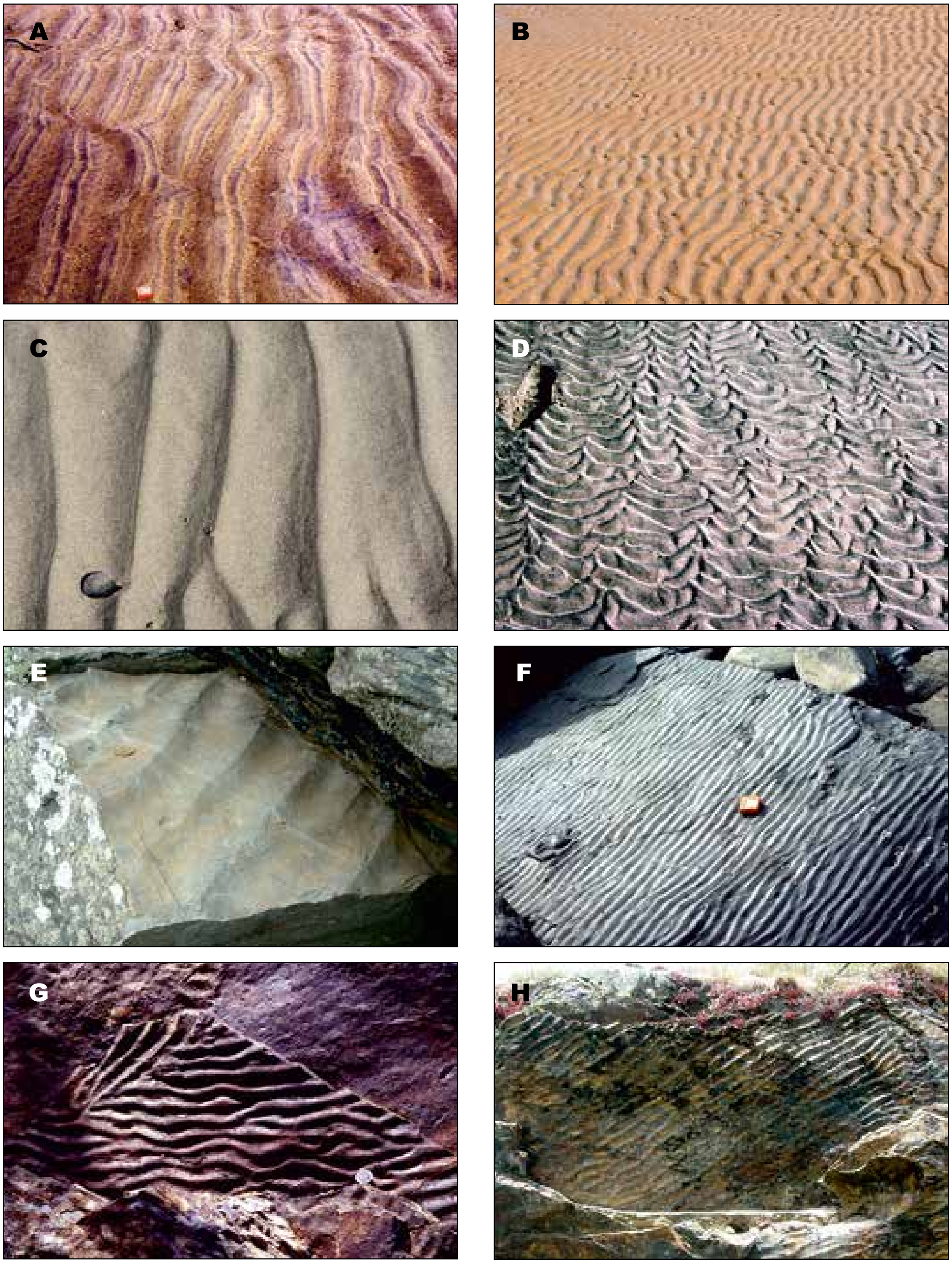


Figure 6.6 Examples of symmetrical ripples with various morphologies. A) Multiple peaked ripples, Tana delta, Finnmark, Norway. B) Rather straight-crested symmetrical ripples with good continuity, but some bifurcations. Barmouth, north Wales. C) Symmetrical ripples with rounded crests. Barmouth, west Wales. D) Rather symmetrical ripples but with cuspate crests. Tana delta, Finnmark, Norway. Ripple spacing is ~10cm. E) Symmetrical ripples with peaked crests. Central Clare Group, Upper Carboniferous, western Ireland. F) Straight crested wave ripples with tuning fork intersections. Coal Measures, Upper Carboniferous, south Wales. G) Sinuous crested symmetrical ripples, Silurian, Cantabria, northern Spain. H) Symmetrical ripples with rather straight crests. Silurian, Cantabria, northern Spain. Ripple spacing is ~6cm.

More complex patterns, resulting from interference between more than one wave set or between waves and currents with divergent directions, range from slight modification of one dominant ripple type to complex interference patterns (Fig. 6.7).

To develop descriptive powers and understanding, study photographs of rippled surfaces like those in Figure 6.8, or better still visit a modern, rippled, sand environment such as a beach or river bed and describe, measure and interpret the ripples found there.

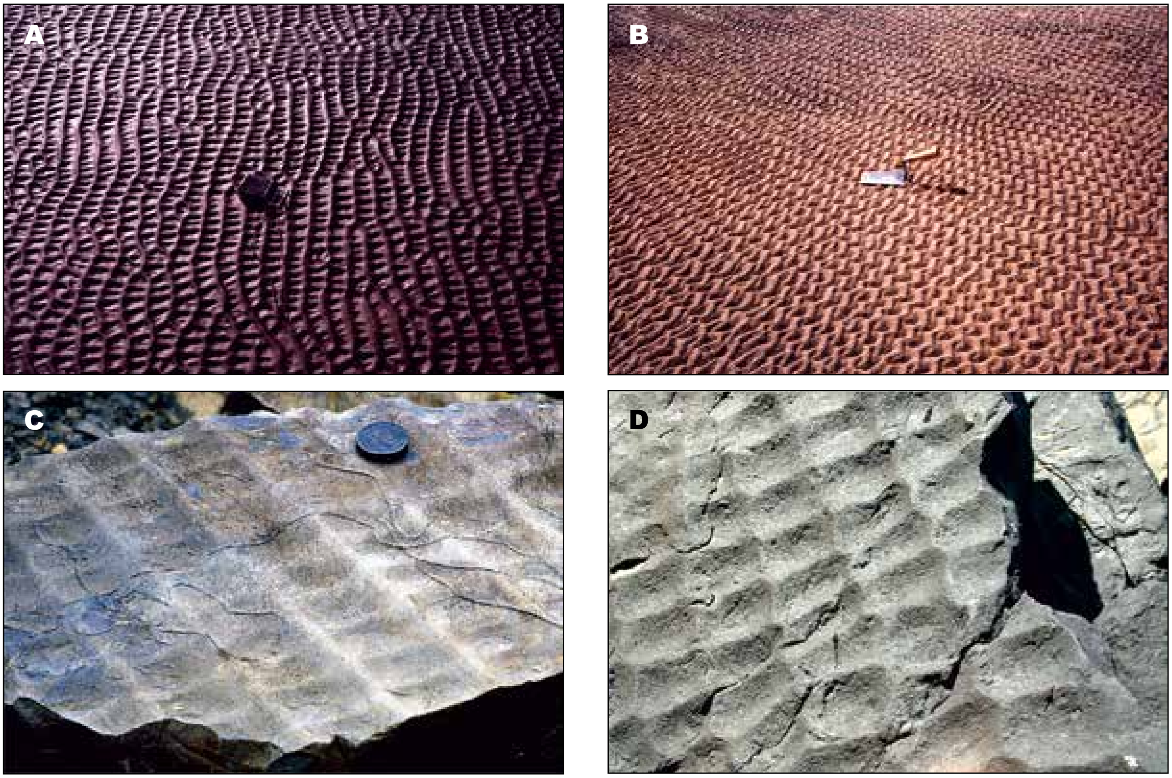


Figure 6.7 Examples of ripple interference patterns. A) Ladder ripples, Tana delta, Finnmark, Norway. B) Interference ripples, Tana delta, Finnmark, Norway. C) Interference ladder ripples with surface animal traces. Carolinafjell Formation. Lower Cretaceous, Spitzbergen. Svalbard. D) Interference ripples due to roughly equal and nearly orthogonal wave sets (~5cm spacing). Central Clare Group, Upper Carboniferous, western Ireland.

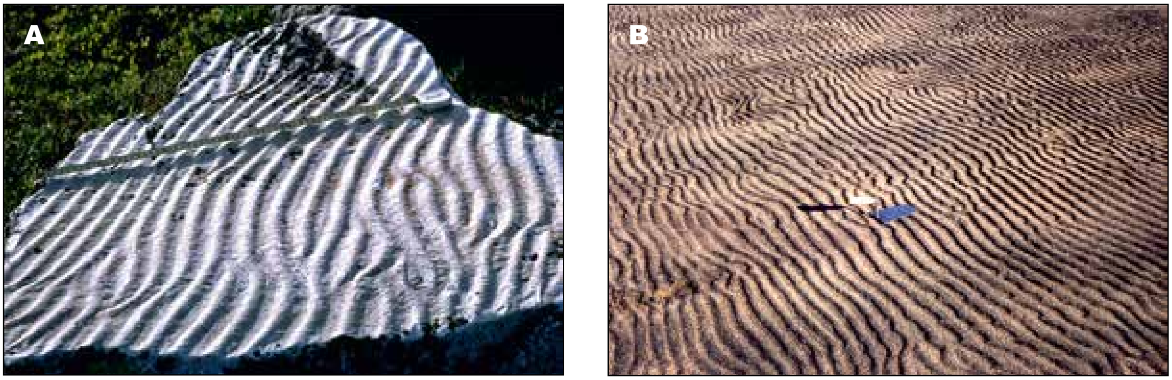


Figure 6.8 Examples of ripples from both an upper bedding surface of sandstone (A) and present-day sand surfaces (B-D). Try to describe the ripples as fully as possible and suggest what processes were responsible for generating them. In what directions did the currents and/or waves responsible operate?

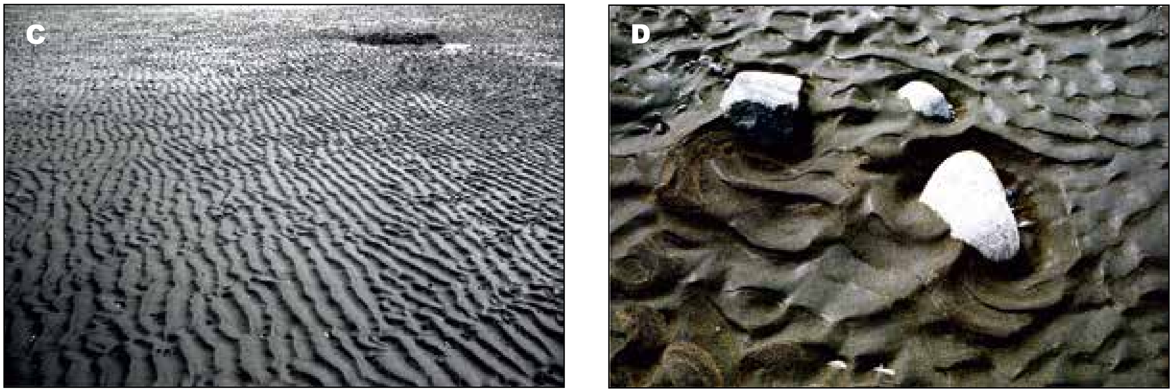


Figure 6.8 *Continued*

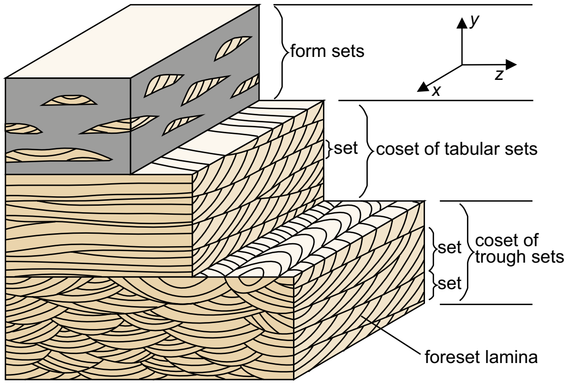


Figure 6.9 Definition diagram for the basic types of cross lamination. The same terms apply at a larger scale to cross bedding. After Allen (1968).

6.1.4 Internal structure: cross lamination

Where ripples occur on a bedding plane or a present-day surface, it is often possible to see associated patterns of internal lamination (Fig. 6.9). Recognition of such lamination is valuable in interpreting accumulated successions. In some successions of interbedded sand and finer sediment, sand ripples are isolated in the fine sediment or are preserved in relief on the top surfaces of thicker sand beds. Such units are termed **form sets**, and the relationship between the form and the internal lamination is usually clear (Figs. 6.9, 6.10). In many sandstones, however, only internal small-scale trough cross lamination occurs. This comprises units (sets) up to 2–3cm thick, each made up of inclined laminae (foresets or cross laminae) (Fig. 6.9). These are usually concave upwards with tangential lower contacts and sharp, truncated upper contacts. Bases of sets commonly have a trough or spoon shaped, being most strongly concave upwards transverse to the mean foreset dip and more gently curved parallel to the dip direction.

Figure 6.10 Examples of ripple cross laminated sands with ripple form sets preserved. A) Tana, Finnmark, Norway. Transport from right to left. B) Fish River Canyon, Namibia. Transport from left to right. In each example, the cross lamination dips down in the direction of bedform migration, reflecting the successive positions of the lee faces of the ripples. The set boundaries dip at low angles in the up-current direction and the sets 'climb' over one another to give 'ripple drift cross lamination'.

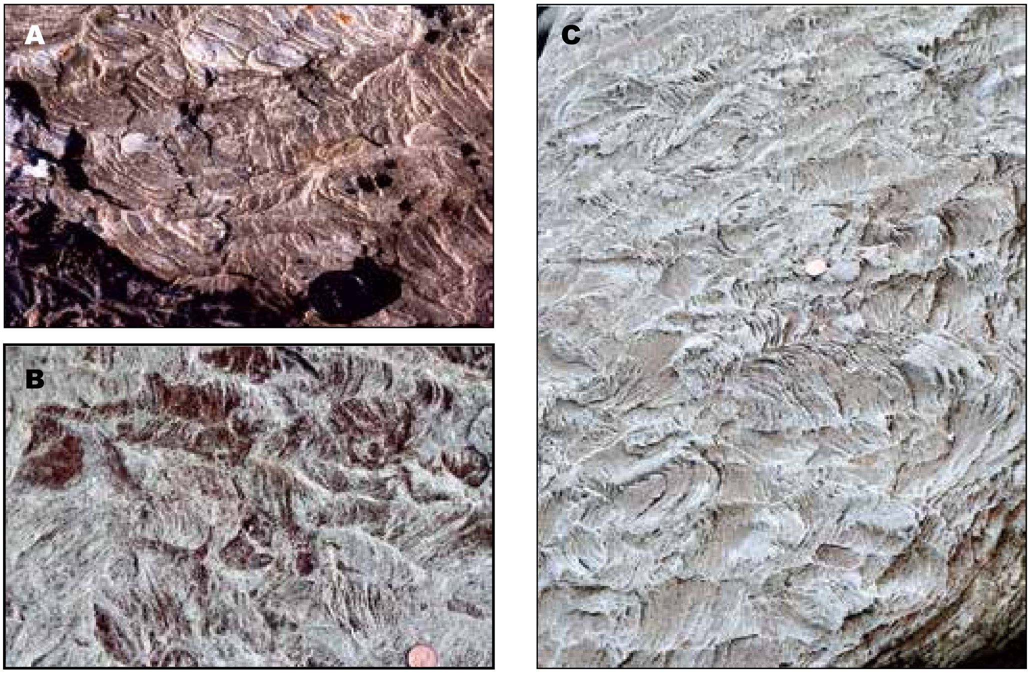


Figure 6.11 Upper bedding surface of sandstone showing 'rib and furrow', the horizontal expression of trough cross lamination. Compare this structure with the more idealized view shown on top of the lowest block in Figure 6.9, and hence determine current direction. A) Central Clare Group, Upper Carboniferous, western Ireland. B) Coal Measures, Upper Carboniferous, Northumberland, England. C) Coal Measures, Upper Carboniferous, Northumberland, England.

Exposed, wind-deflated sand surfaces and many bedding planes in ancient, medium- to fine-grained sandstones show a distinctive pattern of curved laminae dipping into the bed in parallel zones (Fig. 6.11). In plan view, the laminae are usually concave down dip and the zones are commonly up to 8cm wide and 20–30cm long, although in some cases they are longer. This pattern is termed **rib and furrow** and it is a horizontal expression of the trough cross lamination produced during migration of current ripples (see below). Less commonly, straighter cross laminae intersect bedding planes and may dip in opposed directions. This pattern of opposed cross lamination is generated by certain types of wave ripple. It is commonly accompanied by an interfingering of laminae at the ripple crest and by the draping of some laminae over the crest (Figs. 6.12, 6.13, 6.14). Cross lamination can only be understood by a full appreciation of its three-dimensional nature as different orientations of vertical section (i.e. of exposure) show rather different patterns of lamination. It is best understood by reference to block diagrams (Fig. 6.9), which should be studied with the aim of imagining how the lamination would appear in vertical sections with different orientations.

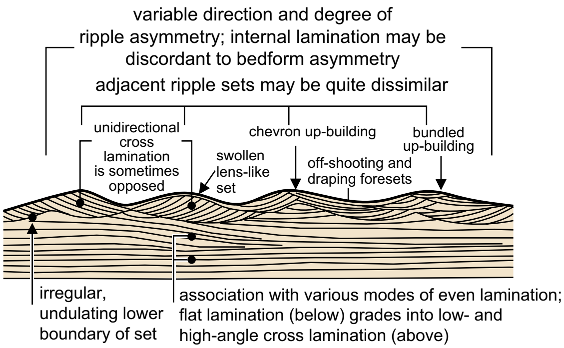


Figure 6.12 Some of the features which help in the diagnosis of wave-ripple cross lamination. After de Raaf et al. (1977).

Two important varieties of cross lamination warrant separate discussion.

Climbing ripple cross lamination (ripple drift)

In most cross-laminated sediment, boundaries between successive sets are erosive and roughly horizontal. In some examples, however, the boundaries between sets are inclined and not always erosive. They dip in the opposite direction to the dip of the cross laminae and at varying angles (Figs. 6.15, 6.16). This is **climbing ripple cross lamination** or **ripple drift**. The **angle of climb** of the ripple sets is determined by the ratio between the rate of downstream ripple migration and the rate of vertical aggradation of the bed (Fig. 6.15). In most cases, the accumulation rate is small compared to the migration rate and the resultant angle of climb is low so that **subcritical climbing** occurs. In that case, as ripples move down current, they truncate the upper parts of the preceding ripples in a train and only their basal part is preserved to form a set of climbing ripple cross lamination. **Critical climbing** occurs where the angle of climb exactly matches the angle of the stoss slope of the ripples such that the entire bedform is just preserved. **Supercritical climbing** occurs where the angle of climb is greater than the angle of the stoss slope of the ripples. In that case, both lee and stoss slope deposits accumulate so that laminae can be traced uninterrupted between successive sets.

It is useful to record the inclination of erosive set boundaries. Where stoss-side laminae are preserved, it is

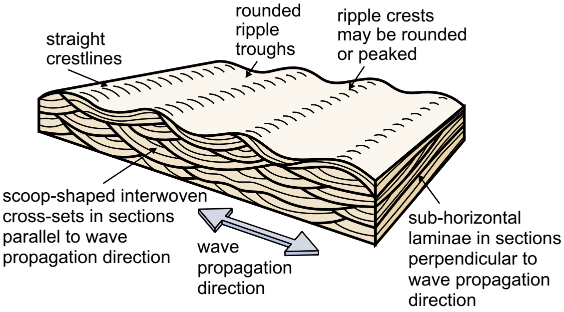


Figure 6.13 Idealized three-dimensional structure of stratification produced by wave oscillation. Modified after Boersma (1970).

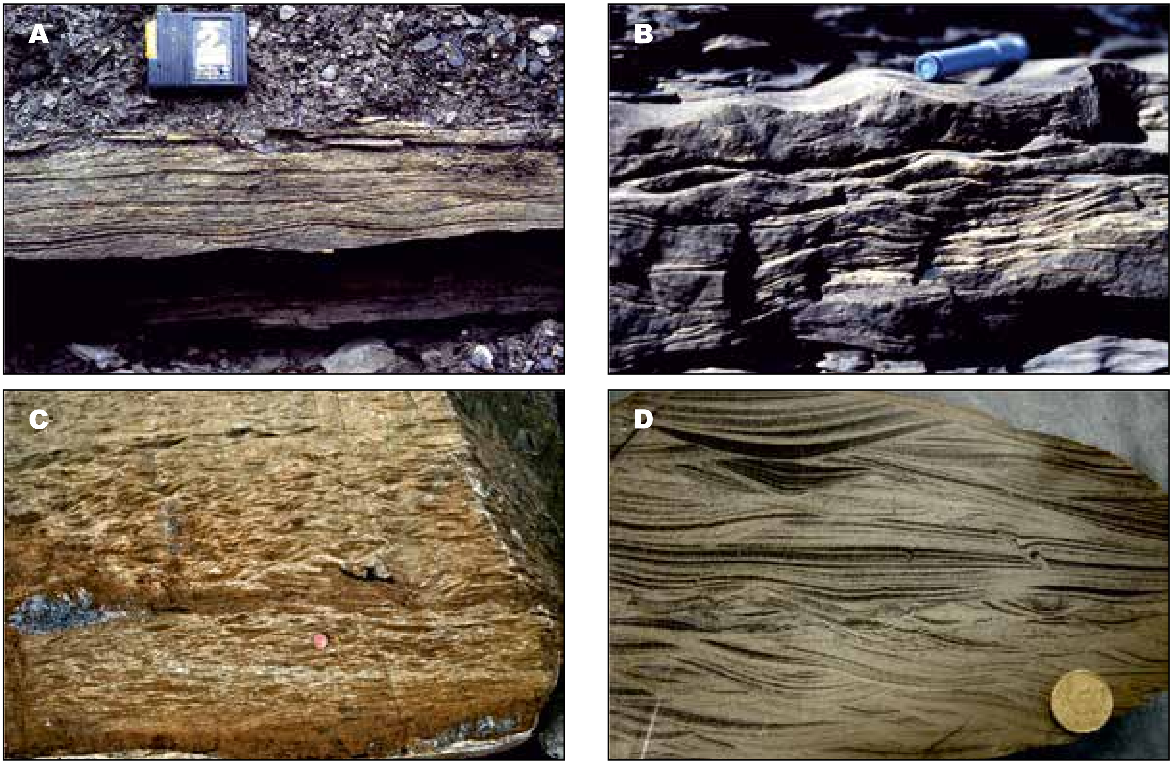


Figure 6.14 Patterns of internal lamination associated with wave action. A) Wave ripples showing internal lamination. A) Wave-ripple lamination occurring throughout a sharp-based, storm-generated sandstone interbedded with shallow marine mudstone. Carolinafjell Formation, Cretaceous, Spitsbergen, Svalbard. B) Wave ripples showing internal lamination and surface form. Central Clare Group, Upper Carboniferous, western Ireland. C) Thick unit of ripple cross-laminated sandstone with some wave influence. Coal Measures, Upper Carboniferous, Northumberland, England. D) Wave-ripple lamination with slight burrow disturbance. Locality unknown, University of Leeds collection.

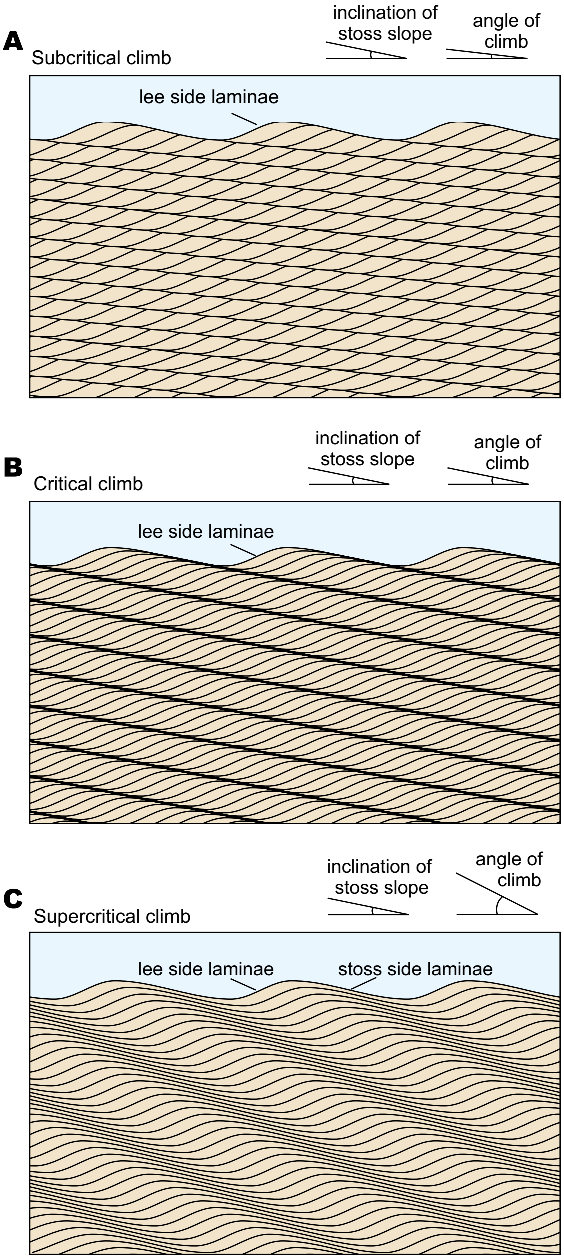


Figure 6.15 Schematic illustrations of different types of climbing ripple ('ripple-drift') cross lamination. A) Subcritical climb, whereby the angle of climb is less than the angle of the stoss slope of the ripples and erosion between sets occurs as a consequence of migration. B) Critical climb, whereby the angle of climb is equal to the angle of the stoss slope of the ripples. C) Supercritical climb, whereby the angle of climb is greater than the angle of the stoss slope, resulting in the preservation of stoss-slope laminae. In each case the ripples responsible for generating the cross lamination have the same geometry and morphology. The nomenclature for climbing ripples illustrated here also applies to larger bedforms.

important to record the inclination of a line through successive positions of the same ripple crest. These measurements record the trajectories of the ripples as they moved downstream whilst the bed accreted vertically (Fig. 6.15). The geometry of the cross-lamination is therefore an indication of the rate at which the bed aggraded vertically with steeper angles of climb (supercritical) associated with the highest rates of aggradation (Figs. 6.17, 6.18).

Flaser, lenticular and wavy bedding

In some units of ripple cross-laminated sand, the pattern is broken up by interlaminations and lenses of finer-grained sediment (silt and mud) (Figs. 6.19, 6.20). Where sand dominates, as in **flaser bedding** (see also §5.2.7**)**, the muddy sediment occurs as thin and often discontinuous laminae, which drape ripple forms or are confined to ripple troughs (Fig. 6.21A). Where the fine-grained sediment dominates, sand may occur as isolated ripple form sets (**lenticular bedding**: Fig. 6.21C). There is a continuous gradation in the proportions of sand and finer-grained sediment, with the general term "**wavy bedding**" often being used to describe units with intermediate proportions of sand and mud (Fig 6.21B). Any description of mixed intervals should try to estimate relative proportions of the different components.

6.1.5 Processes of ripple formation and deposition by water

Water movement over a sand bed, as unidirectional currents, as oscillatory waves or as a combination of both may give rise to ripples.

Unidirectional water currents

When the velocity of water flowing over a sand bed exceeds a certain critical value, grains begin to move (see §3.5). With widespread movement of grains finer than about 0.6mm in diameter, asymmetrical ripples begin to form almost immediately (Fig. 6.22). The earliest ripples are usually rather straight and continuously crested but with gradually increasing velocity the ripples transform into more three-dimensional patterns culminating in linguoid forms. Ridges and hollows parallel to flow become more common and more closely spaced on the stoss sides, and lee-side scour pits become more clearly defined. In plan, therefore, the shape of the ripples provides a rough qualitative guide to flow velocity (Fig. 6.23), although water depth also plays a part in the case of shallow flows.

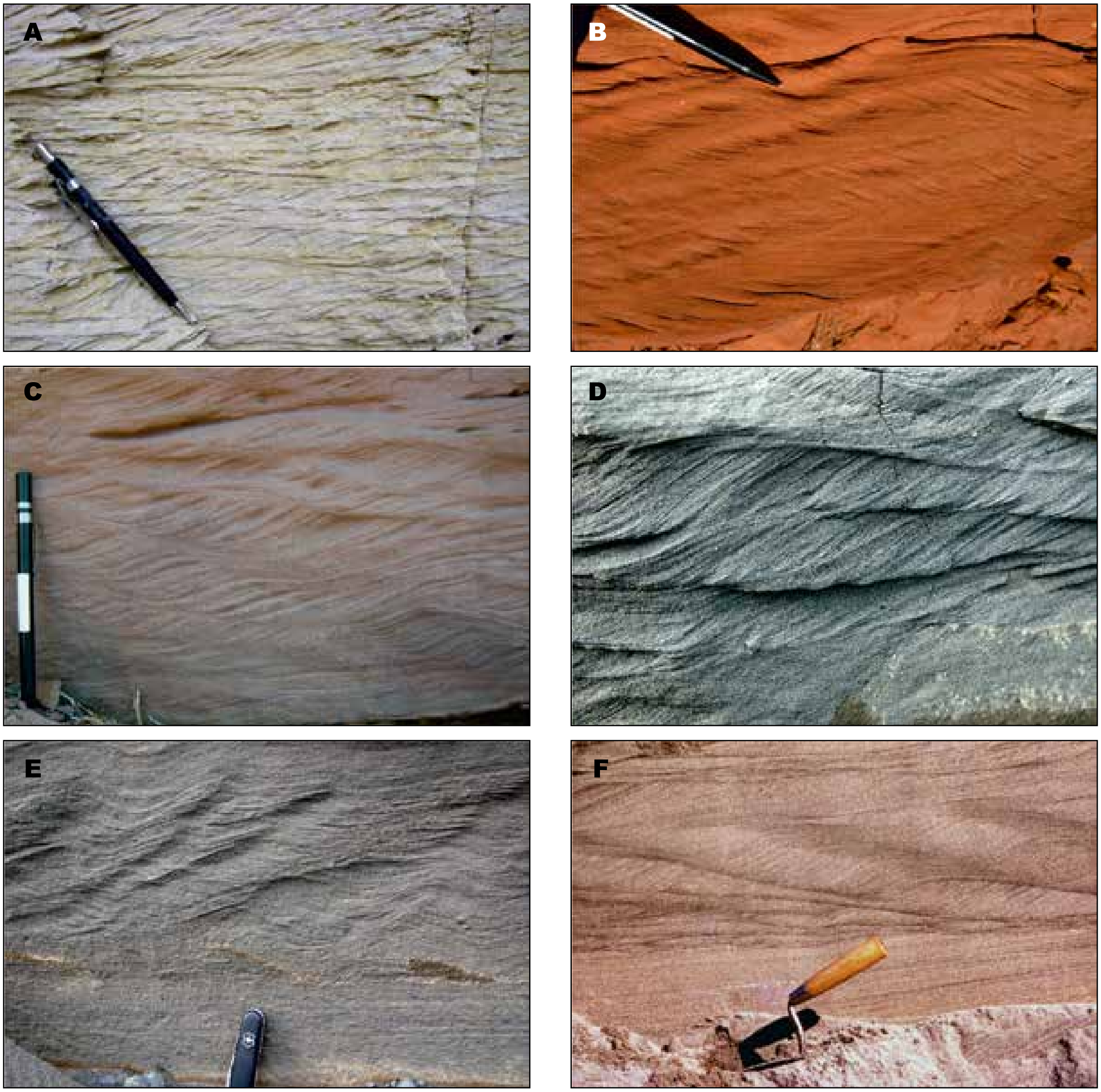


Figure 6.16 Examples of climbing ripple lamination produced by bedforms climbing at various angles. A) Subcritical climbing ripple cross lamination with erosion between sets and with sets thickening and thinning along current. Only the lower parts of the foresets are preserved. Sets at top appear to be descending. Kyenta Formation, Jurassic, Utah, USA. B) Near-critical climbing ripple cross lamination Moenkopi Formation, Triassic, Utah, USA. C) Supercritical climbing ripple cross lamination with preservation of stoss-side laminae in places. Jurassic Kayenta Formation, Utah, USA. D) Subcritical ripple climbing where only the lower part of the foreset is preserved. In this example, the structure passes to the left into ripple form sets (each ~6cm thick). Modern fluvial deposits, central Iceland. E) Supercritical ripple climbing where both the lee and the stoss slopes of the ripple forms are accumulated. Lee-slope deposits are shorter and dip down to the right as viewed. Ripple migration direction was to the right. Recent fluvial deposits, Iceland. F) Variable ripple climbing with subcritical broadly passing up into supercritical. Ripple migration direction was to the left. Recent fluvial deposits, Tana, Finnmark, Norway.

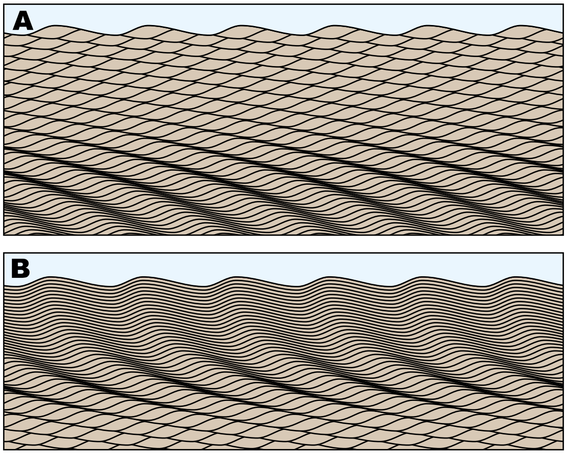


Figure 6.17 Computer generated patterns of ripple cross lamination showing the effects of changes in the angle of bedform climb on preserved stratification style. A) Decrease in angle of climb from supercritical at the base to subcritical at the top. B) Increase in angle of climb from subcritical at the base to supercritical at the top.

Figure 6.18 Example of ripple cross lamination that undergoes an increase in angle of climb from subcritical in the lower part of the section to supercritical in the upper part. Modern fluvial deposits, central Iceland.

Although there is a slight increase in wavelength with increasing flow velocity, the main control on ripple dimensions is the grain size of the sediment, coarser sand giving larger ripples. Current ripples are most readily envisaged as forming in shallow water, but they are also produced in deep water due to the action of ocean-bottom currents. Turbidity currents (see §3.7.2) can also give rise to ripples and cross lamination. During deceleration of such currents, sand and silt falling from suspension may be reworked on the bed into ripples.

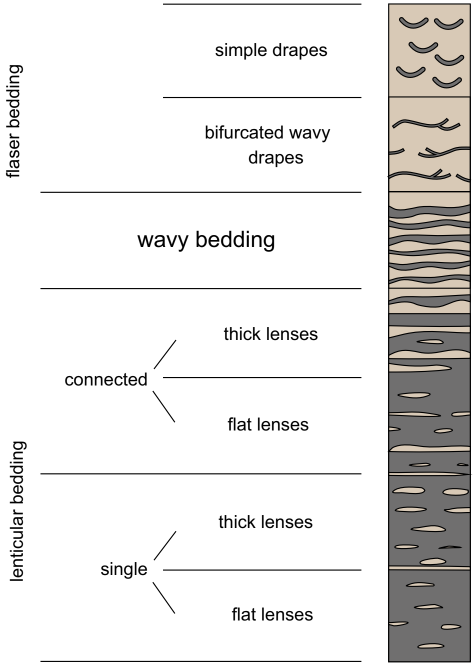


Figure 6.19 Variety of cross lamination and interlamination resulting from mixed lithologies of sand and mud. After Reineck and Singh (1973).

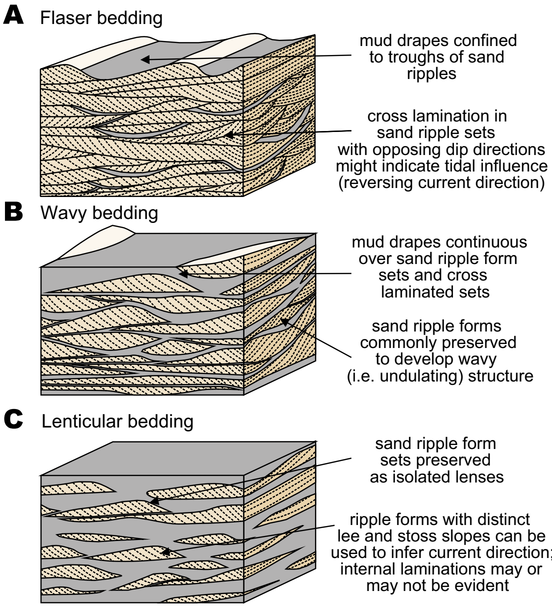


Figure 6.20 Styles of interbedding in mixed sand and mud lithologies. A) Flaser bedding. B) Wavy bedding. C) Lenticular bedding. The progression from A to C results from a net decrease in current speed and increased deposition and preservation of mud drapes. Modified after Reineck and Singh (1980).

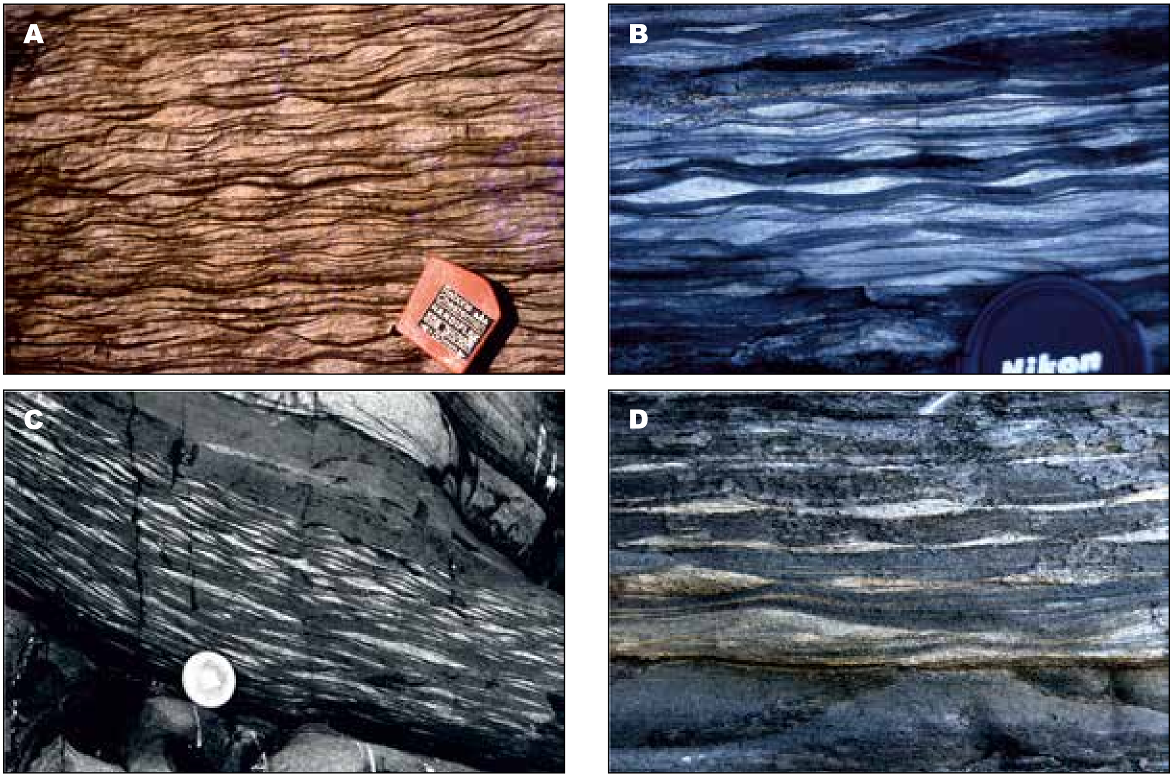


Figure 6.21 Examples of the variety of cross lamination structures resulting from mixed lithologies of sand and mud. A) Flaser and wavy bedding associated with wave modified ripples in heterolithic strata. Coal Measures, Upper Carboniferous, Northumberland, England. B) Wavy bedding with mud completely draping sandy, wave-influenced ripple forms. Central Clare Group, Upper Carboniferous, western Ireland. C) Lenticular bedding with rather peaked and isolated symmetrical ripple form sets. Northam Formation, Upper Carboniferous, north Devon, England. D) Dark siltstone with trains of wave-influenced ripples, with sharp crests and varying cross-lamination dip directions. Width depicted in photo is ~0.4m. Central Clare Group. Upper Carboniferous. western Ireland.

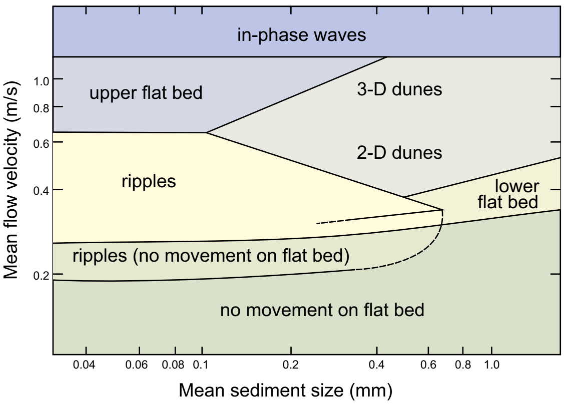


Figure 6.22 The stability fields of different bedforms in relation to flow velocity and grain size for water depth of 0.2m. Note that the plot is on a log-log scale. For other depths, the position of the lines are shifted and a three-dimensional plot is needed to illustrate fully the bedform distribution. After Harms et al. (1975).

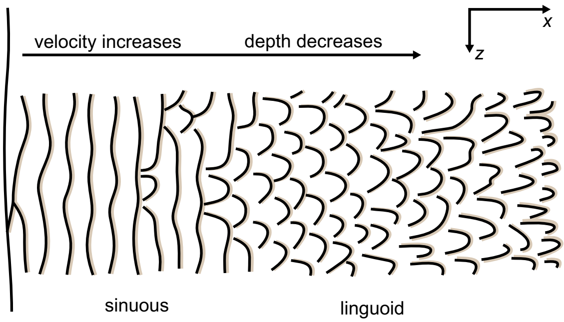


Figure 6.23 The shapes of asymmetrical current ripples, formed without wave influence, related to water depth and velocity. Plan view. After Allen (1968).

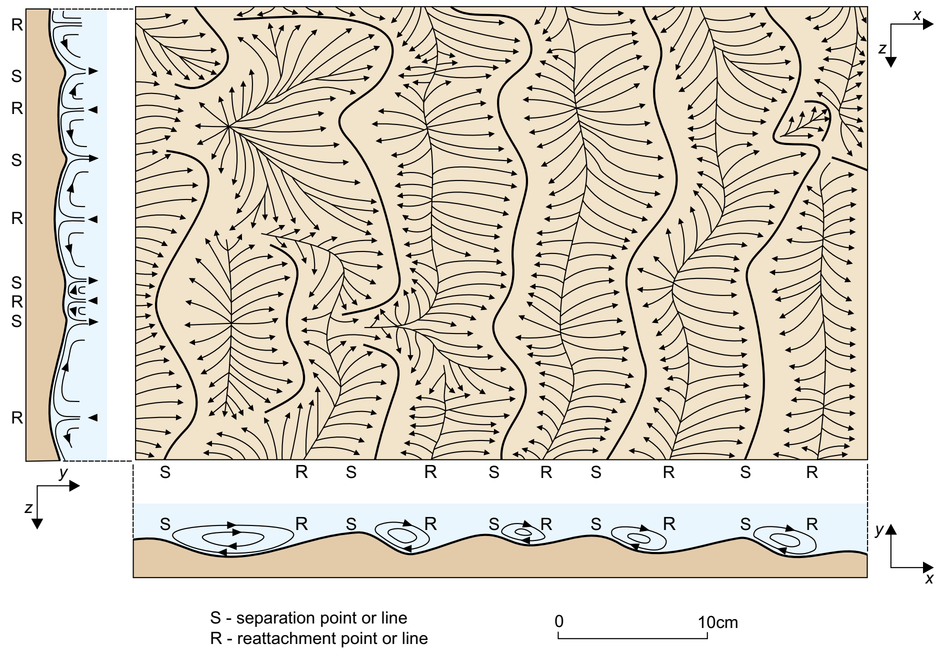


Figure 6.24 The pattern of water movement close to the bed over a field of asymmetrical current ripples. The plan shows flow directions at the bed; the heavier lines represent ripple crestlines. These directions will be similar to the directions of grain movement. Flow separates at the ripple crest and reattaches on the stoss side of the next ripple downstream. Note how the attachment is focused downstream of the concave sectors of the crestline. After Allen (1970).

The variety of ripple shape probably relates to the structure of water turbulence close to the bed. Straighter-crested ripples with crestlines transverse to the flow have rotating eddies in the lee of the ripples, with their axes of rotation parallel to the ripple crestlines (Fig. 6.24). With increasing flow velocity, eddies with axial components parallel to flow become more important, producing increasingly three-dimensional ripple shapes. Ridges and hollows on ripple stoss sides result from eddies with axes of rotation parallel to flow.

Rhomboid ripples, which have not been so extensively studied, appear to form under very shallow conditions close to the boundary between ripples and upper flat beds. The ripple crests appear to be associated with small-scale hydraulic jumps (see §3.2.6).

It is instructive to observe the movement of sand grains over ripples in a small stream or a laboratory channel. It should be possible to identify zones of separation and reattachment of the flow and to see how the flow reattaches to give an area of scour downstream of each ripple. With straight-crested ripples the reattachment is generally in a continuous zone, but with linguoid ripples reattachment is concentrated in scour pits from where grains are swept away in all directions. Some of the sand swept from the scour pit moves up stream to mix with sand being deposited in the lee of the upstream ripple. This mixing helps cause the lee face of the ripple to have a tangential base. Most of the sand swept from a scour pit moves down current to supply the next ripple downstream. Grains approach the crestline at different speeds. Those moving relatively slowly stop abruptly at the crestline and accumulate high on the lee face, oversteepening its gradient. As the angle of slip is exceeded, failure occurs and **grainflow** takes place (see §3.7.2). Grains moving more rapidly at the crestline are thrown out onto the lee side by a process of **grainfall**. The grains' trajectories are influenced by the strong eddies in a flow separation zone that developed directly downstream of a ripple crest. Together, these two processes generate the cross laminae which record the migration of the lee face of the ripple. The downstream movement of linguoid ripples with scour-pit lee-face couplets generates trough cross lamination (Figs. 6.9, 6.11). Straighter-crested ripples produce cross-laminated sets with less pronounced trough shapes.

Cosets of ripple cross lamination result from the migration of ripples combined with a net accumulation of sediment on the bed. With no bed aggradation, ripples migrate downstream and can then only be preserved when movement ceases, and then only as form sets. With a high rate of sediment supply, the bed will aggrade vertically as ripples migrate producing climbing ripple cross lamination (ripple drift). The angle of climb (trajectory) reflects the balance between the rates of vertical bed aggradation and of ripple migration. For a given downstream migration rate, the steeper the climb angle, the higher the rate of aggradation. When the angle of climb exceeds the slope of the ripple stoss side, supercritical climbing takes place and stoss-side laminae are preserved (Figs. 6.15, 6.16, 6.17, 6.18) (see also §6.1.4).

Surface wave processes

All symmetrical ripples, and many asymmetrical ones with straight and continuous crestlines, result from surface wave activity, sometimes acting in conjunction with a current. The ripple morphology and lamination are closely related to the pattern of water movement close to the bed. These can be understood by looking closely at gentle wave action on a beach or by trying simple experiments in a laboratory wave tank. Crystals of potassium permanganate on the bed of a wave tank will give a dye stream that shows the pattern of water movement close to the bed (see Fig. 3.13). Three main types of wave behaviour have a bearing on sediment response: **free gravity waves**, **forced waves** and **breaking waves**.

Free gravity waves move beyond the area where they were generated by wind action, and the pattern of movement for any water particle is an almost closed loop (see §3.3; Fig. 3.13). Close to the bed wave-orbitals become horizontally flattened, first as ellipses and eventually as linear movements with a to-and-fro movement. This oscillatory motion generates straight-crested ripples with crestlines parallel to the wave front.

The first ripples to form are **rolling grain ripples**, which are of low relief and reflect movement just above critical erosion conditions. With stronger waves, ripples generate eddies as each wave passes and **vortex ripples** develop (Figs. 6.3, 6.6, 6.25).

The size, spacing and symmetry of wave-generated ripples appear to be controlled by four principal factors which define boundary conditions: the maximum wave-orbital velocity at the bed, the asymmetry of orbital velocities at the bed, the mean grain size and the wave period. The last two factors mainly influence ripple size. Wave ripples occur when maximum orbital velocities fall between those that give no movement and those that give a plane bed

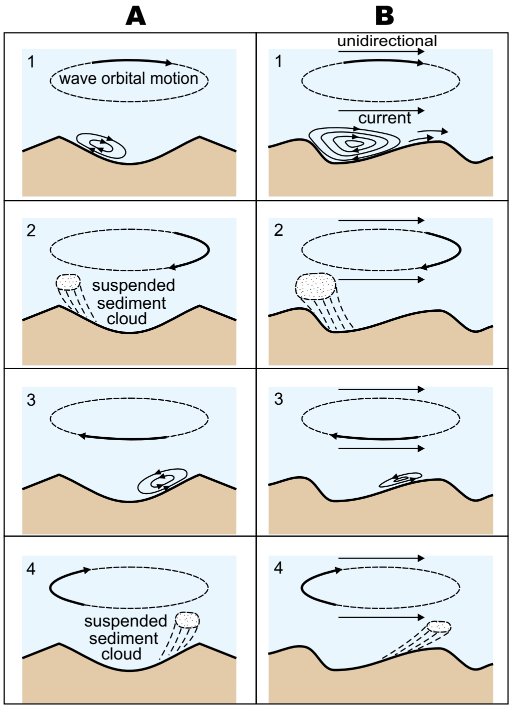


Figure 6.25 The patterns of water and sediment movement over vortex wave ripples. A) Where wave orbital velocities are similar ripples are symmetrical. B) Where there is an asymmetry in the orbital currents ripples adopt an asymmetrical form. After Inman and Bowen (1963).

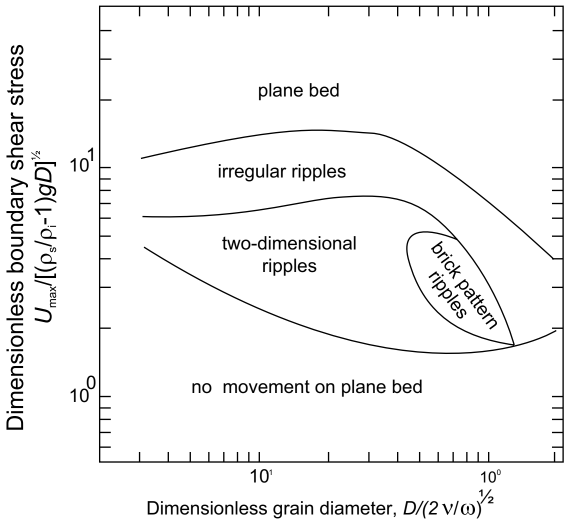


Figure 6.26 The occurrence of different types of bedform as a result of waves acting in a straight channel, under different conditions of wave strength and sediment grain size. U_{max} is the maximum orbital velocity close to the bed, D is particle diameter, ρ_s and ρ_i are solid and fluid densities, v is kinematic viscosity of the fluid and ω is the angular frequency of the waves. After Kaneko (1980).

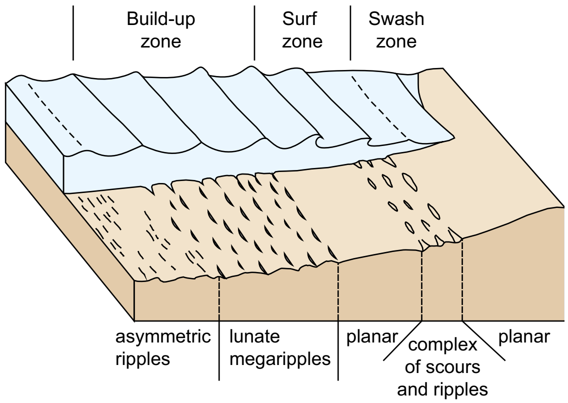


Figure 6.27 Zonation of wave-generated bedforms offshore from the beach on the high-energy coast of Oregon. Shorelines with different energy regimes have different patterns of zonation. After Clifton et al. (1971).

(Fig. 6.26). The asymmetry of the orbital velocity determines the boundary between symmetrical and asymmetrical ripples; greater velocity asymmetry gives rise to more asymmetrical ripples. In shallow offshore areas, where waves are shoaling, a zonation of ripple types can sometimes be recognized (Fig. 6.27).

Most waves result from the drag of wind on the water surface, but such processes are complicated and have little direct bearing on sediment response. In shallow water, however, the sediment surface may be strongly influenced by waves being actively driven by the wind (**forced waves**). Their pattern of water movement is more complex than that of free gravity waves, involving a combination of orbital motion and unidirectional flow. The resulting ripples are asymmetrical and may be difficult to distinguish from the products of shoaling waves or unidirectional currents (Fig. 6.5D & E).

Under **breaking waves**, flow is extremely confused. The surge and backwash of the swash zone will generate ripples only if the waves are gentle. Under more active conditions rhomboid ripples or a flat bed develops (see §6.1.2, §6.3, Fig. 6.26).

Shapes of symmetrical ripples are largely a function of water depth. Round-crested forms occur in rather deep water whereas strongly peaked ripples are more common in very shallow, near-emergent conditions.

Interference effects

In many settings, waves and currents or multiple wave sets may coexist and interact. With wave–current interaction, the ripple pattern produced depends on the relative strengths and directions of the two processes. If they act in similar directions, although not necessarily with the same sense of motion, straight-crested ripples result. These are difficult to distinguish from those produced by shoaling or forced waves. Waves straighten the crests of what might otherwise have been sinuously crested current ripples. When wave and current directions diverge, interference patterns develop (e.g. Fig. 6.7). However, not all interference patterns imply that the various processes operated at the same time. Separation of different processes in time is particularly common on tidal flats where conditions are continuously changing.

When wave motion is superimposed on currents, the water velocity close to the bed is instantaneously increased. Critical erosion velocity may then be exceeded and a bed may become rippled under a current whose time-averaged velocity is below the threshold required to initiate movement.

6.1.6 Uses of ripples and cross lamination

Ripple marks and cross lamination have three main uses.

Way up

Ripple marks and cross lamination are amongst the most reliable indicators of "way-up", although there are three ways in which they may be mistaken for structures whose "way-up" significance is ambiguous or opposite. In strongly deformed rocks, bedding-cleavage intersection sometimes causes a pattern of small-scale undulations on bedding surfaces that can look remarkably like ripples. Careful study of the cleavage and joint patterns may resolve the problem, but one should always be cautious when apparent ripple crests are closely parallel to fold axes or cleavage traces. Try to identify cross lamination associated with the "ripples" before asserting their sedimentary origin. In interbedded successions with preserved ripple morphologies, a lower bedding surface may sometimes preserve the ripples on top of an underlying bed as a cast. An examination of the internal structure of the beds either side of the rippled surface should show cross lamination beneath the rippled surface.

Superficially, transverse scours (see §4.2.2) can resemble current ripples. However, internally they have no cross lamination and may be associated with other types of sole mark.

Conditions of deposition

Ripples indicate deposition by currents and waves strong enough to exceed the critical erosion velocity but not strong enough to form dunes, sandwaves or a flat bed (see §6.2, §6.3). Ripple symmetry and crestline shape enable estimates of the relative strengths of currents and waves to be made. Hydrodynamic interpretations of preserved bedforms or internal structures are usually based on comparison with equilibrium conditions. By its very nature, all sedimentation demands non-equilibrium conditions with an excess of sediment supply due to either waning or expanding flow. Climbing ripple cross lamination can help to indicate sediment supply as the angle of climb varies with sedimentation rate.

Palaeocurrent and palaeowave direction

Both ripple morphology and cross lamination may indicate directions of waves and currents. Ripples respond quickly to local or short-term changes in flow direction, so they may record directions divergent from the overall palaeoslope or the high-stage flow direction.

Ideally, palaeocurrents are best measured from ripple forms or cross lamination on bedding planes rather than from cross lamination in vertical section. Remember that it is difficult to judge anything but a component of direction in a vertical section. Measure ripple crest orientation (for wave ripples), general ripple trend (for current ripples), or the axes of troughs on bedding planes showing rib and furrow structure. For symmetrical wave ripples, it may not be possible to judge the sense of wave movement.

Consider the examples of ripple stratification shown in Figure 6.28. Try to interpret the processes responsible for generating the types of stratification present.

6.2 Aqueous dunes, sandwaves, bars and cross bedding

6.2.1 Introduction

Many areas of sandy river beds, tidal flats and channels, and of sandy sea floor swept by tidal currents show bedforms many times larger than current ripples. These larger forms are separated from ripples by a distinct jump in both height and spacing even though many of the proportions and shape factors are often comparable (Fig. 6.29). It is relatively uncommon to find subaqueous current-generated

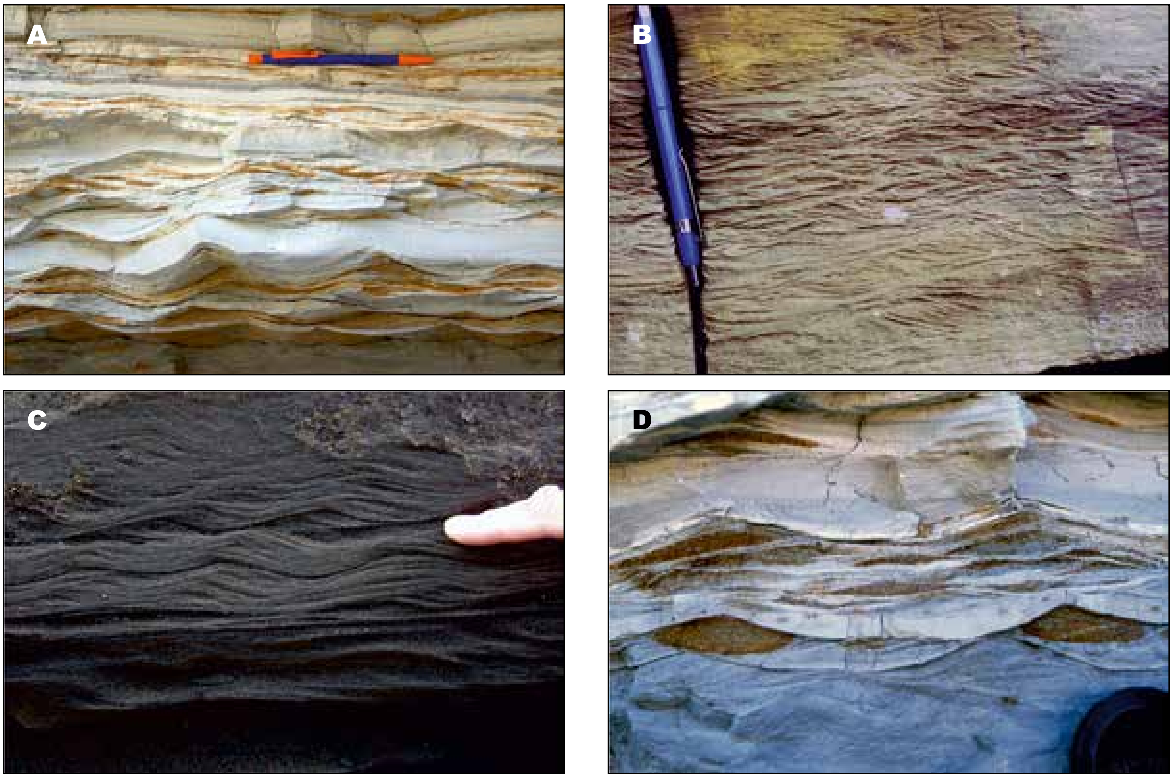


Figure 6.28 Examples of various types of ripple cross lamination. Suggest what has happened in terms of depositional process in each case.

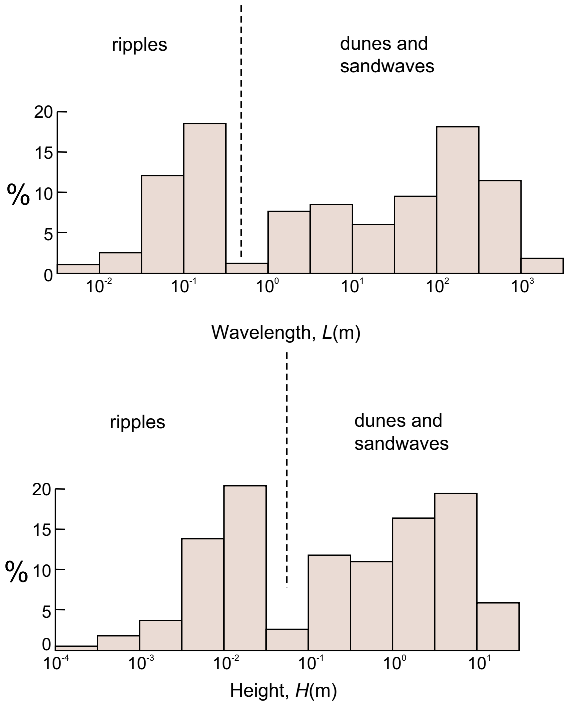


Figure 6.29 Histograms of wavelength and height of ripple-like sub-aqueous bedforms from various present-day environments. A conspicuous gap separates ripples from large forms. The population of larger forms probably includes representatives of both dunes and sandwaves. After Allen (1968).

bedforms in the height range 3–10cm and the spacing range 30cm –1m. Furthermore, the larger forms commonly have smaller current ripples superimposed upon their stoss sides. Where such superimposition is seen on a river bed or tidal flat, it is quite likely that it results, at least in part, from continued sediment movement during the falling river stage or the waning ebb tide, when the large forms were no longer active but the smaller ones were. However, in controlled experiments, superimposition of ripples on larger forms also occurs under equilibrium conditions.

In the geological record, only fairly small-scale examples of these larger bedforms occur as bedding surface features (form sets) and their occurrence is most commonly reconstructed from the patterns of cross bedding to which they give rise.

6.2.2 Material

Large-scale bedforms and the cross bedding which results from their migration most commonly occur in sediment of medium sand and coarser grain size. They may also occur in gravelly sands and fine gravels of any composition. Cross bedding is, in addition, found in many coarser conglomerates, where its origin may result from other processes (see Chapter 7). Certain types of large-scale cross bedding in sandstones also the result from the migration of large morphological features, not directly related to bedforms.

6.2.3 Size, shape and classification of large-scale bedforms

Large sandy bedforms, ranging upwards in size from 1m in wavelength have been variously described as dunes, sandwaves, megaripples, large-scale ripple marks and different types of bar. Lack of consistency in agreeing and applying terminology has led to some confusion, notably in older literature. Here we set out the features that seem most important in describing these forms. Most will be visible at low tide or low river level (Fig. 6.30), but similar criteria can be applied to the description of sub-aqueous features as revealed by echo sounders and other sonar imaging devices.

The first question to ask about large sandy forms is whether or not they form a repetitive pattern on the bed. Do they have a regular spacing and a more or less uniform height and is their plan form consistent? If so, the height and wavelength should be recorded and the plan form described using terminology similar to that for asymmetric ripples (cf. Fig. 6.1).

The second question to ask is whether there is only one size of bedform present (simple forms), or whether multiple sizes co-exist (compound forms). In the case of compound forms, it is important to judge if the smaller forms are confined to the stoss sides of the large forms or if they occur on both stoss and lee sides. It is also useful to record the dimensions of all scales of structure and to compare the orientations of smaller structures with those of the larger ones.

In some cases, the larger bedforms are repetitive and independent of the morphology of any channel in which they occur. In other cases, the larger forms may be related to bends in a channel, to sinuosity of the flow within a channel or to splitting and re-joining of the flow. This last judgement may, in the case of large rivers or estuaries, be rather difficult to make. Climbing a nearby hill, flying a drone (where permitted) or studying aerial photographs, satellite images or detailed topographic maps may be very informative (Fig. 6.31). In the largest systems, features of several different size groups may be superimposed.

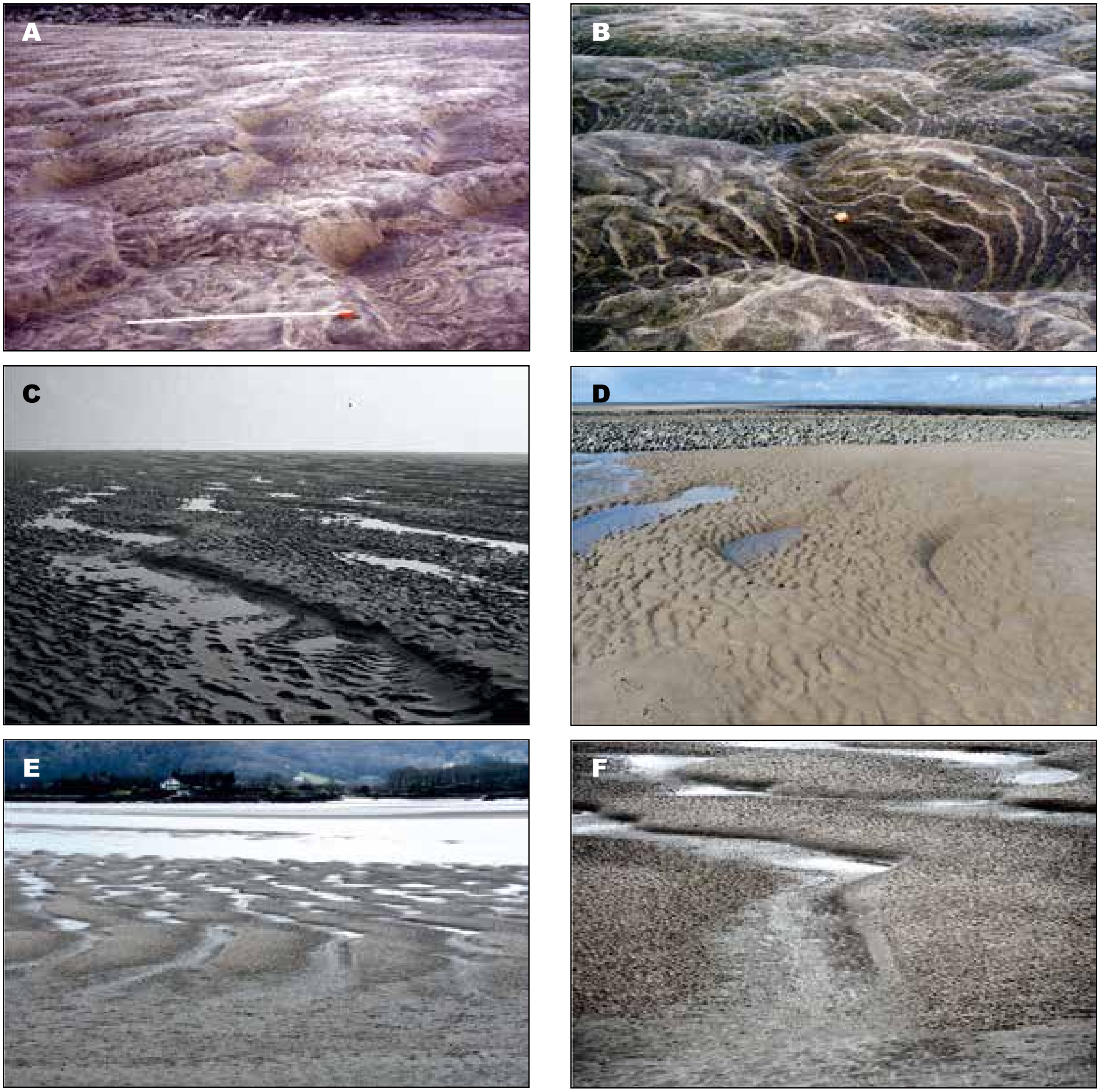


Figure 6.30 Examples of larger sub-aqueous bedforms that are broadly grouped as 'dunes'. A) Sinuous-crested to linguoid dunes showing superimposed current ripples, well-developed scour pits on the lee sides and lee-side avalanche surfaces. Tana River, Finnmark Norway. B) Downstream view of the same dunes showing the radial pattern of ripples around the lee-side scour pits due to the expansion of the reattached secondary flow. C) Rather straight-crested low-relief dunes with superimposed ripples on a tidal flat. Some of the ripple orientations probably record modifications during ebb-tide emergence. Haringvliet, Netherlands. D) Curved-crested low-relief dunes with superimposed ripples on a beach Barmouth, west Wales. E) Sinuously crested dunes with rather low relief and with superimposed current ripples that probably continued to move after the larger forms stopped moving. F) Close-up of E). Both Mawddach estuary, west Wales.

Simple, repetitive, strongly asymmetric forms, whose dimensions are independent of the width of the channel, are best referred to as **aqueous dunes**. These may be strongly three-dimensional with sinuous crest-lines and well-developed scour pits on their lee sides (Fig. 6.30A & B) or they may be gently curved or straight-crested without scour pits (Fig. 6.30C). Dunes commonly have small-scale current ripples on their stoss sides. These commonly face downstream, towards the dune crest. However, directly downstream of a dune lee face they may face up stream. In the scour pits of three-dimensional dunes, ripples are commonly present and may fan out from the centre of the pit (Fig. 6.30B).

In some areas subject to strong tidal currents, usually subtidal areas of the sea floor, large and apparently simple bedforms occur. These are up to several metres high and hundreds of metres in wavelength. They are most often seen on the records of echo-sounding or side-scan sonar surveys and are usually referred to as **sandwaves**. They are commonly asymmetric and have rather straight and continuous crestlines up to many hundreds of metres in length.

Sandwaves are oriented normal to the direction of tidal flow. The asymmetry, which can appear very obvious on foreshortened echo traces, is in reality commonly quite slight, although a whole spectrum exists from near symmetric forms to strongly asymmetric ones. With strong asymmetry, the steep (lee) side may be a slip face, but with progressively reducing asymmetry the angle of the steeper side declines.

With compound bedforms, which apparently are scaled independently of the width of the channel, a term such as **complex** or **compound sandwave** is appropriate, although the emergent top of such areas may be referred to as "sand flats" (Fig. 6.31).

Where a bedform is related in scale to the width of the channel, the general term **bar** is appropriate. Bars can be suitably qualified depending on their relationship to channel or thalweg curvature, on whether they are simple or compound, and on whether or not they have their own discrete slip faces (Figs. 6.31, 6.32).

Figure 6.31 Examples of larger-scale dune forms that are sometimes referred to as sandwaves or bars. A) and B). Large scale low-relief dunes with a broad linguoid shape that have been termed linguoid bars. The shapes of the crestlines have been somewhat modified as the bedforms emerged during falling river stage. Smaller dunes are present in some of the low areas between the larger forms. C) Part of a sandy river bed with repetitive, large-scale, linguoid dunes, offset en-echelon and making up a large bank-attached bar. Note the convoluted crestlines of the emergent forms at the top of the image. D) Aerial photograph of part of a sandy river bed showing large, linguoid repetitive dunes (linguoid bars) superimposed on larger composite sand accumulations that are best termed bars, some of which occur mid-channel, whereas others are attached to the banks. All from Tana River, Finnmark, Norway. River bed is around 1.5km wide.

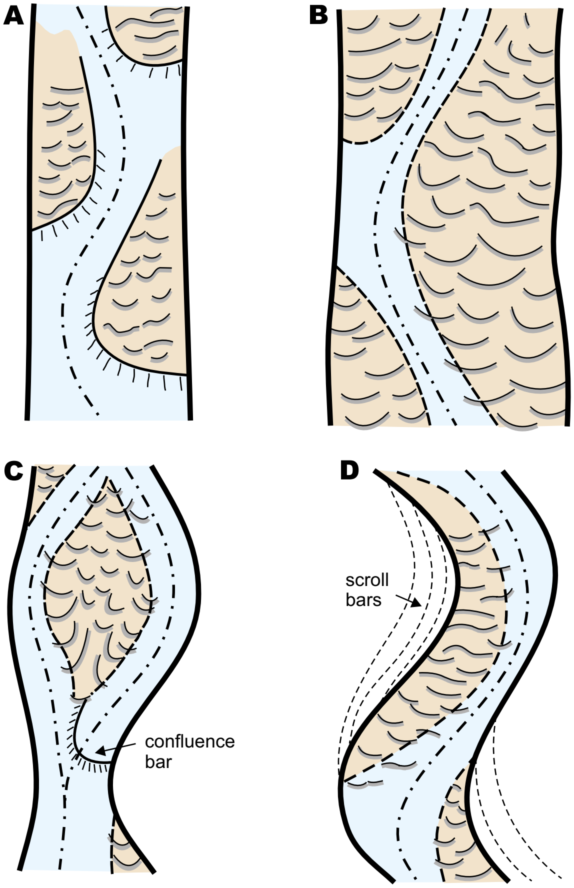


Figure 6.32 Definition diagram for different bar types on a sandy river bed. Note how the dimensions of the bars relate to the channel width. Surfaces of bars have superimposed repetitive bedforms (dunes) whose size is independent of channel width. A) Alternate bank attached (side) bars with their own large-scale slip faces. B) Side bars without their own slip faces. C) Mid-channel bars and possible related confluence bars. D) Point bars related to channel curvature. Lateral migration of the channel perpendicular to the mean flow direction gives rise to scroll bars.

Figure 6.33 Modification by emergence. Series of fan lobes developed as a consequence of waves breaching the crest of a bar and transporting sand onto the upstream side. Tana River, Finnmark, Norway.

6.2.4 Modification by emergence

Large-scale bedforms, exposed on a river bed at low water stage, commonly show superimposed features that were produced as they emerged. Such features occur at a variety of scales, and their development reflects the energy of the modifying processes and the rate at which emergence took place. With slow emergence, there is more time for modification to occur. With wave action, lee-side slip faces and crestlines of bedforms become rounded off and lobes of sand may extend up stream from the crestline as a result of wash-over by waves (Fig. 6.33). The same action may also reduce the slope of the slipface, concentrate heavy minerals and remould current ripples on the stoss side into wave or interference ripples. During falling water stage, the river or tidal flow may be split by emergent bedforms and the tops of large forms may become incised with the development of small delta lobes extending in front of the original, high-stage slip face (Fig. 6.31). Current ripples may be reoriented to reflect this flow and sand lobes may develop at the confluences of these flow threads. More rapid emergence causes these effects to be suppressed as there is less time for the processes to operate. On intertidal areas where emergence takes only a few hours, larger bedforms are commonly preserved in a relatively unmodified state (Fig. 6.30E).

6.2.5 Internal structures of dunes: medium-scale cross bedding

Excavation of trenches into dunes reveals patterns of inclined bedding similar to the cross lamination of ripples, but on a larger scale. This is called **cross bedding**, although the terms "current bedding" and "false bedding" are in some cases found in older literature. Although set thicknesses are usually greater than 10cm, much of the terminology of cross lamination (Fig. 6.9) still applies. Straight- or long-crested dunes generate **planar cross bedding** that forms tabular sets of wide lateral extent, whereas strongly three-dimensional dunes give **trough cross bedding** with scallop-shaped sets (Figs. 6.34, 6.35, 6.36). Both types are

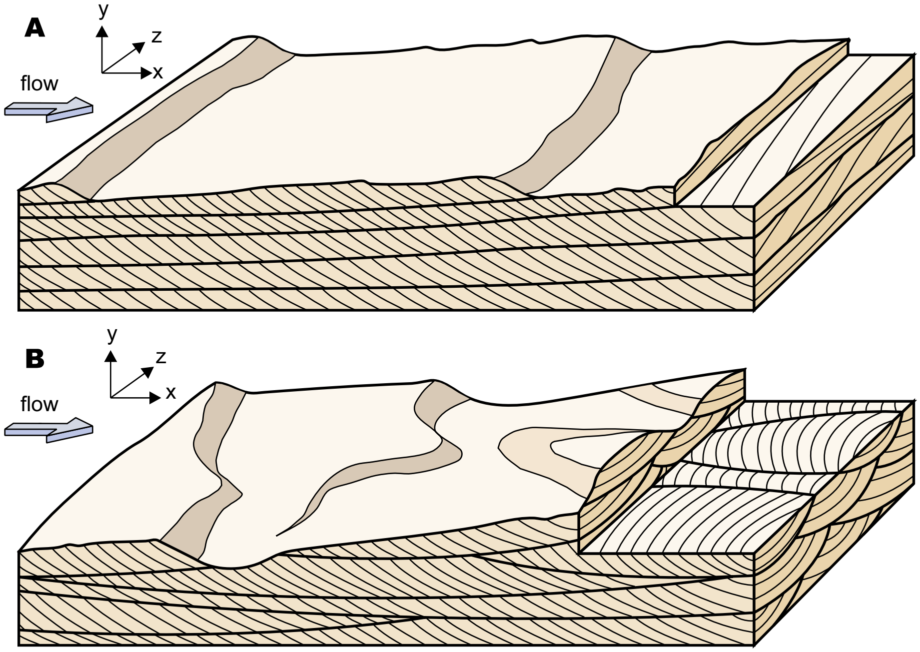


Figure 6.34 Patterns of cross bedding generated by the migration of dunes with different plan-form morphologies. A) Straight-crested (two- dimensional) dunes generate sets of planar cross bedding which exhibits relatively little variability in sections parallel to the dune crest. B) Sinuous-crested (three-dimensional) dunes generate sets of trough cross bedding in which trough-, scallop- or cylindrical-shaped scours filled with foresets are evident in sections parallel to the dune crest. These structures occur over a wide range of scales.

common in the rock record, particularly in medium- and coarse-grained sandstones.

Tabular sets have a wide range of sizes, although sets less than 1m thick are most common. Sets around 1m thick commonly extend laterally for tens of metres, often beyond the limits of exposure. Where several sets are stacked in a coset, they may be separated by thin layers of ripple cross lamination (Fig. 6.35A). Isolated, very thick, tabular sets, up to tens of metres thick, are probably not the product of dunes and are described later (see §6.2.9). The geometry of tabular sets of cross strata is relatively simple with relatively little along-strike variability and the foresets being straight or gently sinuous. However, as with ripples, bedforms with similar morphologies can preserve different patterns of cross stratification depending on their angle of climb (Fig. 6.37). The foresets of tabular sets are usually either asymptotic (tangentially based) or planar (angular-based); convex-up and sigmoidal foreset geometries are relatively uncommon (Fig. 6.38).

Trough sets are seldom more than 1.5m thick and are typically up to a few metres wide and a few tens of metres long. Most commonly, they are around 30cm thick, 1–2m wide and 5–10m long. Foresets of trough sets are usually concave upwards with tangential lower contacts. In plan view, trough cross bedding displays a larger version of "rib and furrow" (Fig. 6.36, cf. Fig. 6.11). Sections cut perpendicular to the current direction are sometimes described as showing **festoon cross bedding** (Fig. 6.35B & C). The geometry of trough-shaped sets of cross strata is more complex than that of tabular sets because the orientation of the foresets varies along the strike of the sets. This leads to significantly different patterns of cross stratification where bedforms with similar morphologies climb at different angles (Fig. 6.39). Furthermore, the plan-view alignment or **phase** of successive of crestline sinuosities also exerts a control on the geometry of resultant cross stratification, as does the amount of sinuosity (Fig. 6.40).

Care should be taken to correctly interpret the geometry of the cross strata that fill trough-shaped sets when seen in vertical profile. Bear in mind that outcrop sections oriented transverse to the trough axis will usually reveal foresets that are apparently concordant with the trough base (Fig. 6.41A),

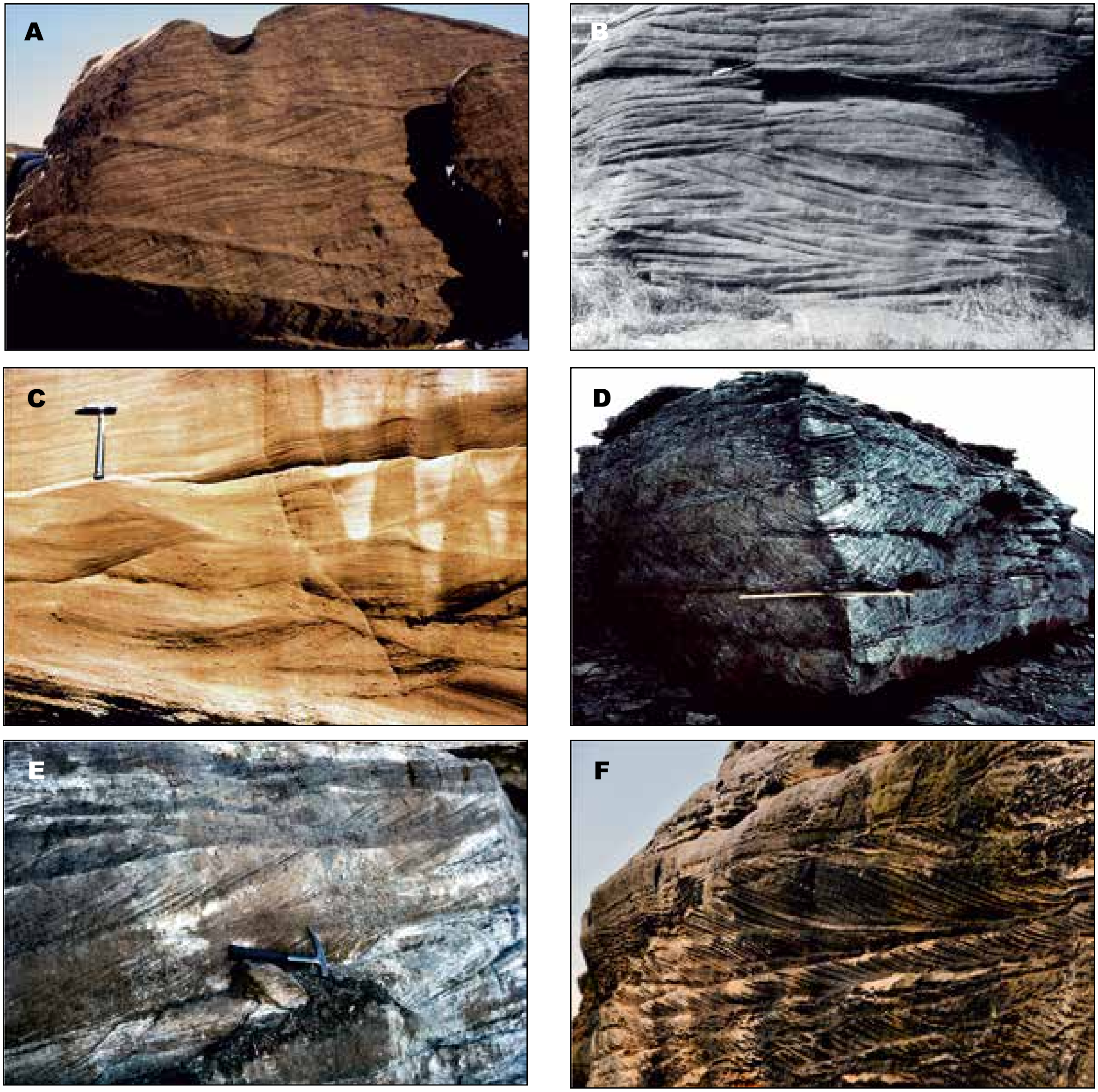


Figure 6.35 Types of medium scale cross bedding seen in section. A) Planar, tabular sets of cross-bedding view roughly parallel to palaeoflow. Sets are separated by thin units of ripple cross-laminated sandstone. Note the angular contact of the foresets at the bases of the sets. Light meter (8cm long) for scale. Roaches Grit, Upper Carboniferous, Staffordshire. B) Trough cross-bedded sandstone viewed normal to palaeoflow. Note the apparently opposed foreset directions due to their curved shape. Pen for scale. Pennant Sandstone Formation, Upper Carboniferous, south Wales. C) Trough cross-bedded sandstone viewed normal to palaeoflow. Note the asymmetric fills of the troughs. Upper Carboniferous, Lothian, Scotland. Photos B and C courtesy of Gilbert Kelling. D) Tabular cross bedding directed out of the corner of the exposure so that both the faces are oblique to the paleoflow. Scale 1m. Cretaceous, Spitzbergen, Svalbard. E) Trough cross bedding in a section sub-parallel with palaeoflow. Proterozoic. Inglefield Land. northwest Greenland. F) Descending sets of medium-scale cross bedding, in a section parallel to palaeoflow. The down-current dip of the set boundaries suggest accretion at the downstream end of a large compound bedform. Individual sets are 10-30cm thick. Fell Sandstone, Lower Carboniferous, Northumberland.

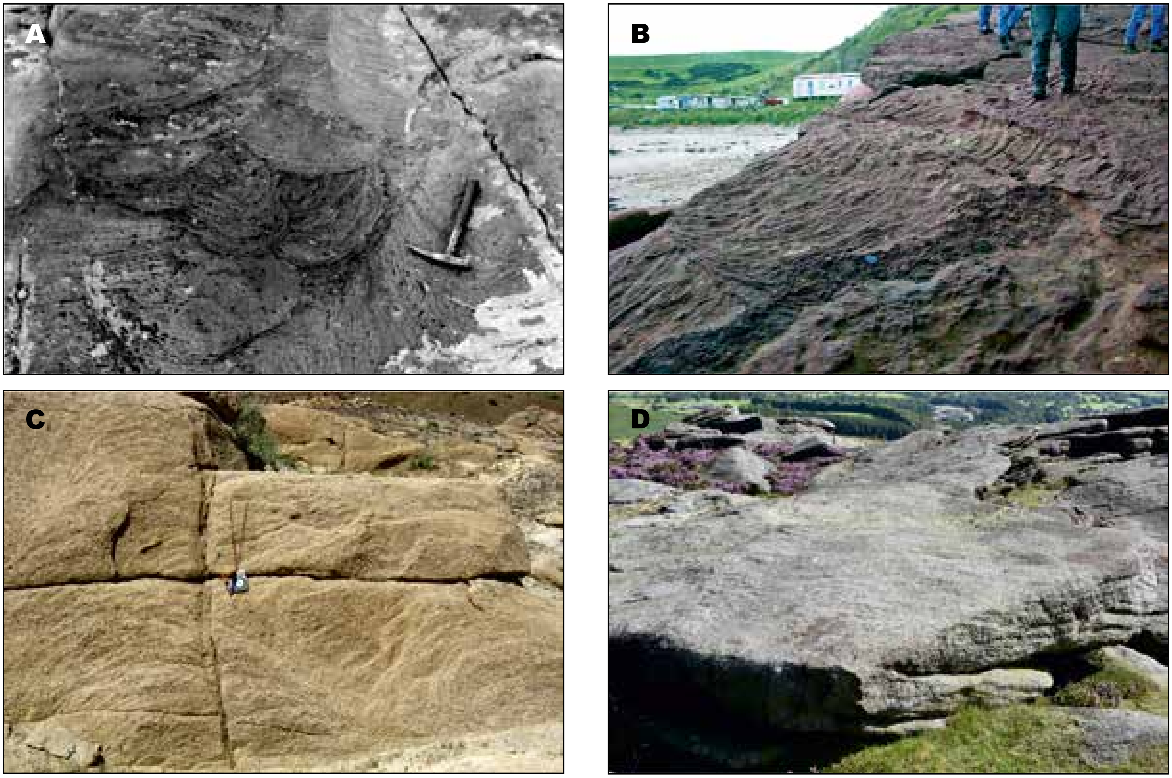


Figure 6.36 Types of cross bedding seen in plan. A) Small-scale trough cross-bedding on an upper bedding surface. Note the intersecting trough forms and the strongly curved foresets, concave down current. The axes of the troughs give the most reliable indicators of palaeoflow direction. Alston Formation, Upper Carboniferous, Northumberland, England. B) Old Red Sandstone, Devonian, Berwickshire, Scotland. C) Sainshand Formation, Cretaceous, Mongolia. D) Kinderscout Grit, Upper Carboniferous, Derbyshire, England.

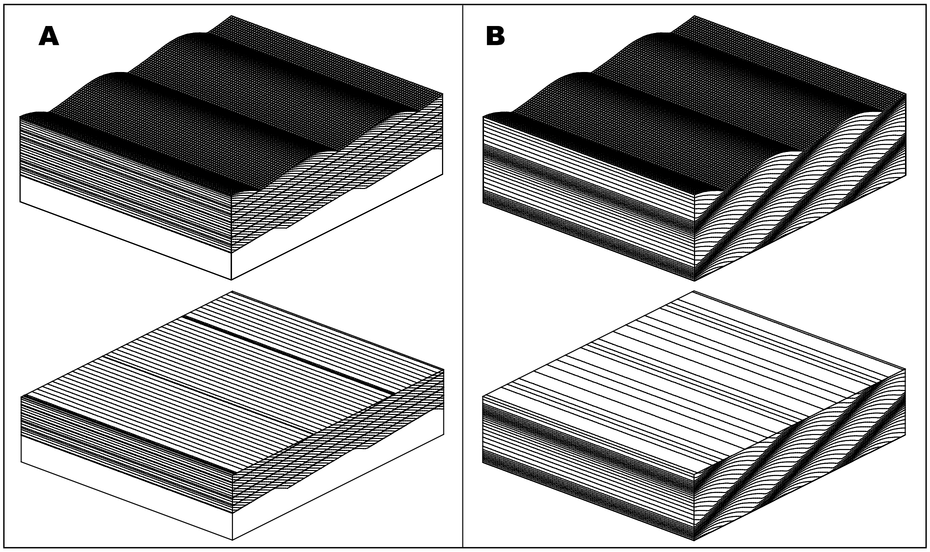


Figure 6.37 Schematic illustrations of the form of cross stratification generated by the migration of two dimensional (i.e. straight crested) bedforms. A) Subcritical angle of climb and horizontal section. B) Supercritical angle of climb and horizontal section. Computer models generated using the 'Bedforms' software of Rubin (1987) and Rubin and Carter (2006).

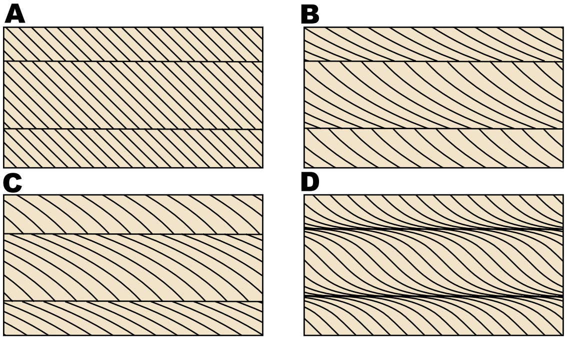


Figure 6.38 Idealized sections in tabular cross bedding, parallel to flow. A) Planar foresets with angular bases. B) Curved foresets with asymptotic or tangential bases. C) Convex up foresets. D) Sigmoidal foresets.

whereas sections oblique to the trough axis will show foresets that apparently fill the trough in an asymmetric manner with downlap onto the trough base (Fig. 6.41B).

Within cosets of cross bedding it is usual for the bounding surfaces between sets to be near-horizontal, although in small outcrops, it may be difficult to judge the orientation of the depositional horizontal. In some extensive exposures, however, bounding surfaces between sets are themselves seen to be inclined, defining larger-scale dipping units. If this is suspected, it can be very important to determine the magnitude and direction of this inclination in relation to the dip of the cross-bedded foresets and, if possible, to the true depositional horizontal. All types of relationship are possible of which ascending (upstream accretion), descending (downstream accretion) and along slope (lateral accretion) are end members.

In some cosets of tabular cross bedding, the directions of dip of foresets in adjacent sets are opposed. Some examples show alternation of direction from set to set, whereas in others only a small proportion of sets show an opposed direction (Fig. 6.42). Such **herring-bone cross bedding** is important in interpreting processes as it can be an important indicator of tidal activity, but care must be taken to distinguish it from festoon cross bedding seen when normal trough cross bedded sets are exposed in sections perpendicular to transport (Fig. 6.35B & C).

Some approximately tabular sets are unusual in showing sigmoidal foresets (Figs. 6.38D, 6.43) where convex-upwards foreset laminae at the top of the set pass up-dip into parallel lamination, which occurs as a "topset" unit (Fig. 6.43). Traced down dip, such foresets become concave upwards and sometimes show a gradual reduction in thickness and in foreset inclination.

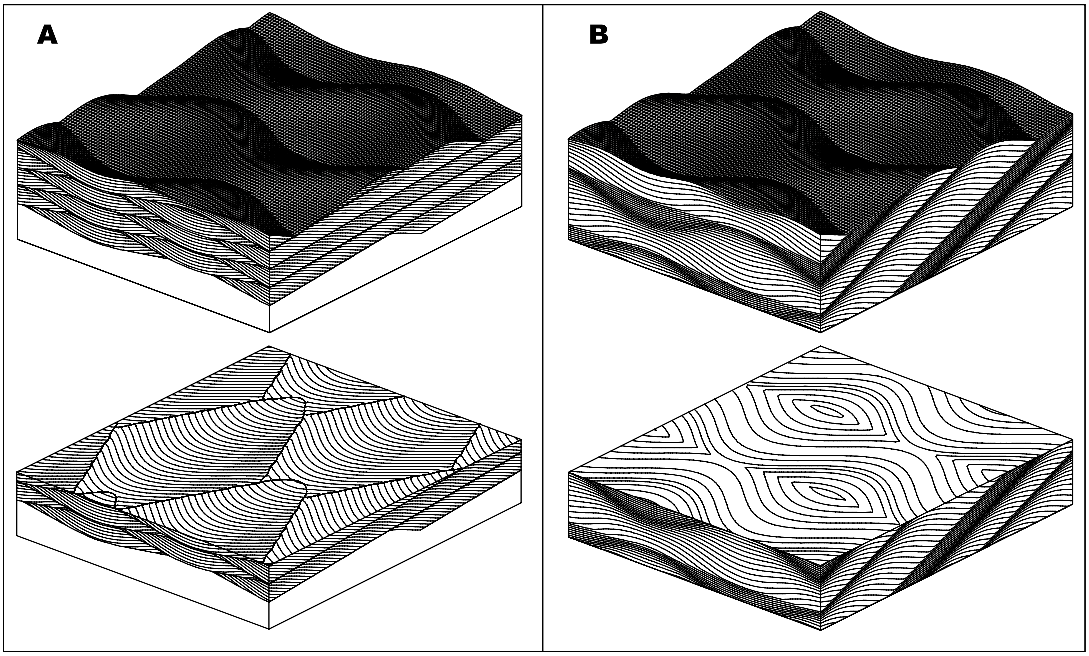


Figure 6.39 Schematic illustrations of the form of cross stratification generated by the migration of three dimensional (i.e. sinuous crested) bedforms. A) Subcritical angle of climb and horizontal section. B) Supercritical angle of climb and horizontal section. Computer models generated using the 'Bedforms' software of Rubin (1987) and Rubin and Carter (2006).

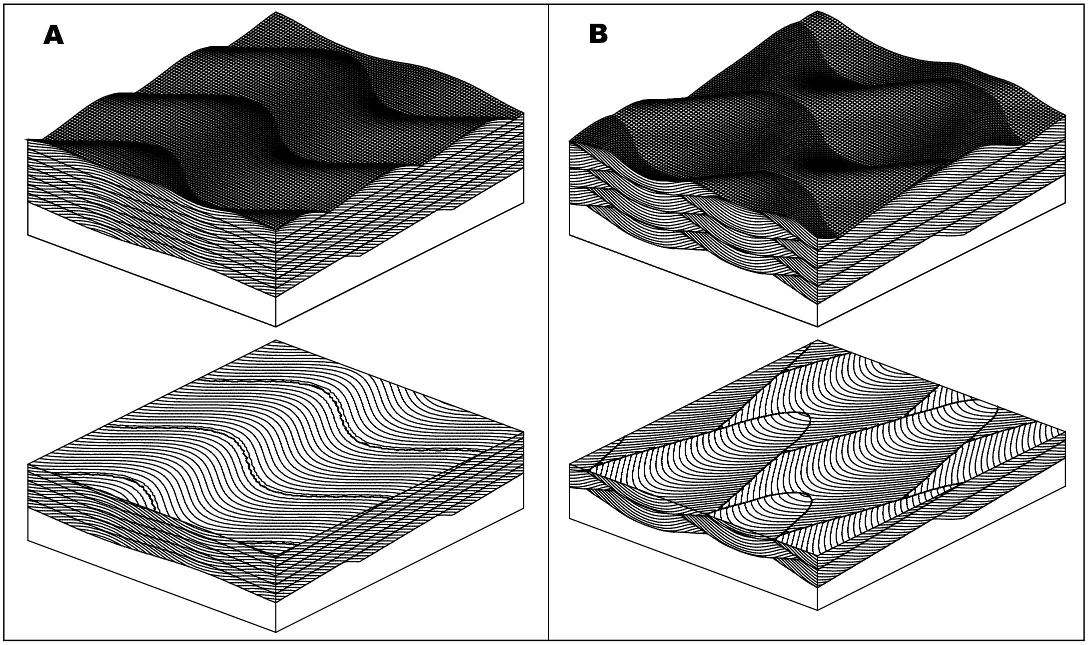


Figure 6.40 Schematic illustrations of the variety of forms of cross stratification generated by the subcritical migration of three dimensional (i.e. sinuous crested) bedforms. A) Train of bedforms with successive crestline sinuosities that are in-phase. B) Bedforms with high amplitude crestline sinuosities that are out-of-phase. Compare the patterns of cross stratification depicted here with those in Fig. 6.39. Computer models generated using the 'Bedforms' software of Rubin (1987) and Rubin and Carter (2006).

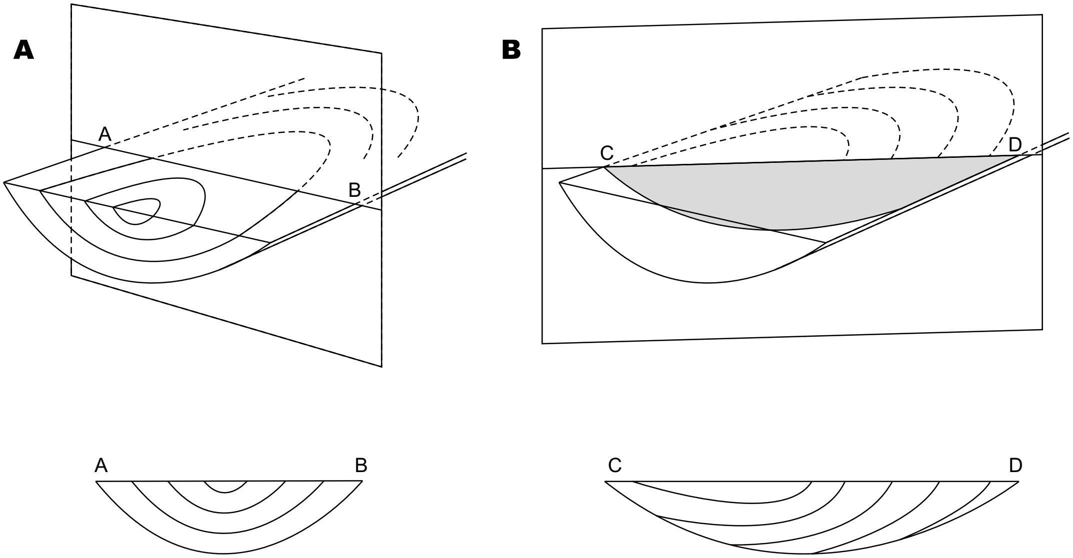


Figure 6.41 Schematic illustration of the geometric complexity of trough cross strata. A) A vertical section, oriented transverse to the trough axis reveals symmetrical cross stratification planes that are apparently concordant with the trough base. B) A vertical section oriented oblique to the same trough axis reveals cross stratification planes that apparently fill the trough asymmetrically and downlap onto its base. This illustrates the problems associated with the measurement of foreset dip azimuths for the purposes of establishing palaeo-transport direction from trough shaped sets of cross strata. Modified after Rubin and Hunter (1983) and DeCelles et al. (1983).

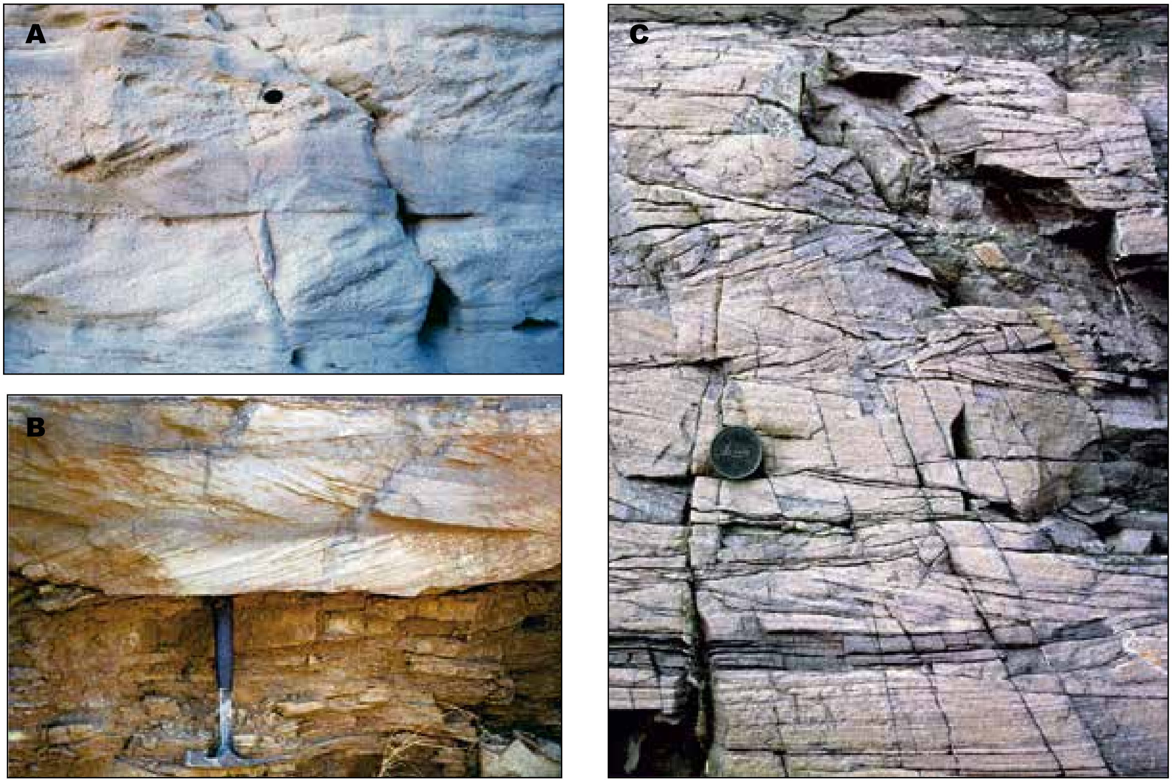


Figure 6.42 Examples of herringbone cross bedding, where successive sets show opposed palaeoflow directions and which is generally attributed to a bi-directional tidal flow regime. A) In gravelly sandstone, Sorbas Member, Miocene, Sorbas Basin, Spain. B) Late Pre-Cambrian, Bela Dam, Sagar, India. Photo courtesy of Gilbert Kelling. C) Smalfjord Formation. Late Proterozoic. Finnmark, north Norway.

6.2.6 Discontinuities and modifications in cross bedding

Cross bedding is not always simple, particularly in tabular sets. Complexity may take several forms. Small-scale ripple cross lamination may occur within foresets, particularly in the lowest parts of tangentially based foresets, the toe-set region. Such cross lamination is commonly directed up the slope of the larger-scale foresets and is termed **counter-current cross lamination** (Fig. 6.44). Discontinuities within foresets are sometimes seen in sections parallel to foreset dip. These erosion surfaces (**reactivation surfaces**) are less steeply inclined than the foresets on either side and they may occur in isolation or as multiple features within a set (Fig. 6.45). With trough cross bedding or with downstream-inclined bounding surfaces, it may be difficult to distinguish bounding surfaces of sets from reactivation surfaces.

In some cross bedding, mainly from tidal and fluvial environments where deposition from suspension is common, **clay** or **mud drapes** occur on the foresets (Fig. 6.46). Some examples occur in pairs, separating thicker and thinner sand foreset increments. In exceptional cases the spacing of the paired drapes increases and then decreases systematically when traced along the set, to define **foreset bundles**.

6.2.7 Process of formation of dunes, sandwaves, bars and cross bedding

Dunes and sandwaves are both responses of a sand bed to currents more powerful, and often deeper, than those that generate ripples. Differences in dune morphology result from differences in flow strength and depth, the three-dimensional forms reflecting deeper, more powerful flows. The pattern of eddying around dunes is closely related to the shape of the dune, with convergence of the reattaching flow in scour pits characterizing the three-dimensional types. In some low-relief dunes (long spacing and small height) the pattern of flow separation and reattachment is confined to the immediate lee-side area. In these cases, the

Figure 6.43 Sigmoidal cross bedding within a single tabular set, developed in medium-grained sandstone. Note that the foresets have tangential bases and the flatter-lying topset laminae are also inclined downstream. St Bees Sandstone Formation, Triassic, Cumbria, England.

Figure 6.44 Single tabular cross-bedded set with a unit of counter-current ripple lamination at the base. Note that the ripple lamination has a climbing geometry and that ripple forms are buried by the foresets. A) Trench in large low-relief aqueous dune. The small ripples on the top surface are wind impact ripples. Scale is 5cm. Present-day, Tana River, Finnmark, Norway. B) Units of upstream-facing (counter-current) ripple lamination (above lens cap) within the lower part of a tabular set of cross bedding in coarse sandstone. Roaches Grit, Upper Carboniferous, Staffordshire, England.

Figure 6.45 Upper part of trench shows a tabular set of cross bedding, within which the foresets are truncated by a lower angle erosion surface, a reactivation surface. Above and to the left of the discontinuity, foresets resume their normal downstream dip. The discontinuity results from falling river stage between successive flood events that caused the dune bedform to migrate. Below the set is a unit of ripple cross-laminated sand with rather variable directions and the base of the trench shows the top of a further tabular set with a rather different orientation. Recent river deposits, Tana River, Finnmark, Norway.

bedform heights are much closer to the flow depth and there is commonly a shallow flow over the crestline.

The separation eddy in the lee of a dune gives rise to a backflow component which helps produce a tangential lee face. The strongly focused eddying in front of concave sectors of the crest line leads to scour pits and to the associated ripple fans (Fig. 6.30). With straight-crested dunes, the strength of flow over the crest can strongly influence lee-side profile and hence foreset shape. With weak flows, grain flow (avalanching) dominates on the slip face, giving angular-based foresets. With stronger currents, flow separation and grain fall become more important, leading to tangential foresets, in some cases with counter-current ripples (Figs. 6.44, 6.47, 6.48, 6.49). Reactivation surfaces (Figs. 6.45, 6.50, 6.51) in river bedforms result from reworking by waves during emergence (see Fig. 6.33) and by flows around the margins of the emergent bedforms. The subordinate tide in both intertidal and subtidal settings is also able to produce similar discontinuities.

Tidal sandwaves have morphologies that reflect the imbalance between the two opposing tidal flows (Fig. 6.52). Very slight imbalance is not capable of generating a marked asymmetry, but in such cases migration rates are slow. With greater imbalance, migration rates are likely to be higher and bedform asymmetry is likely to be more marked. The imbalance is also likely to be reflected

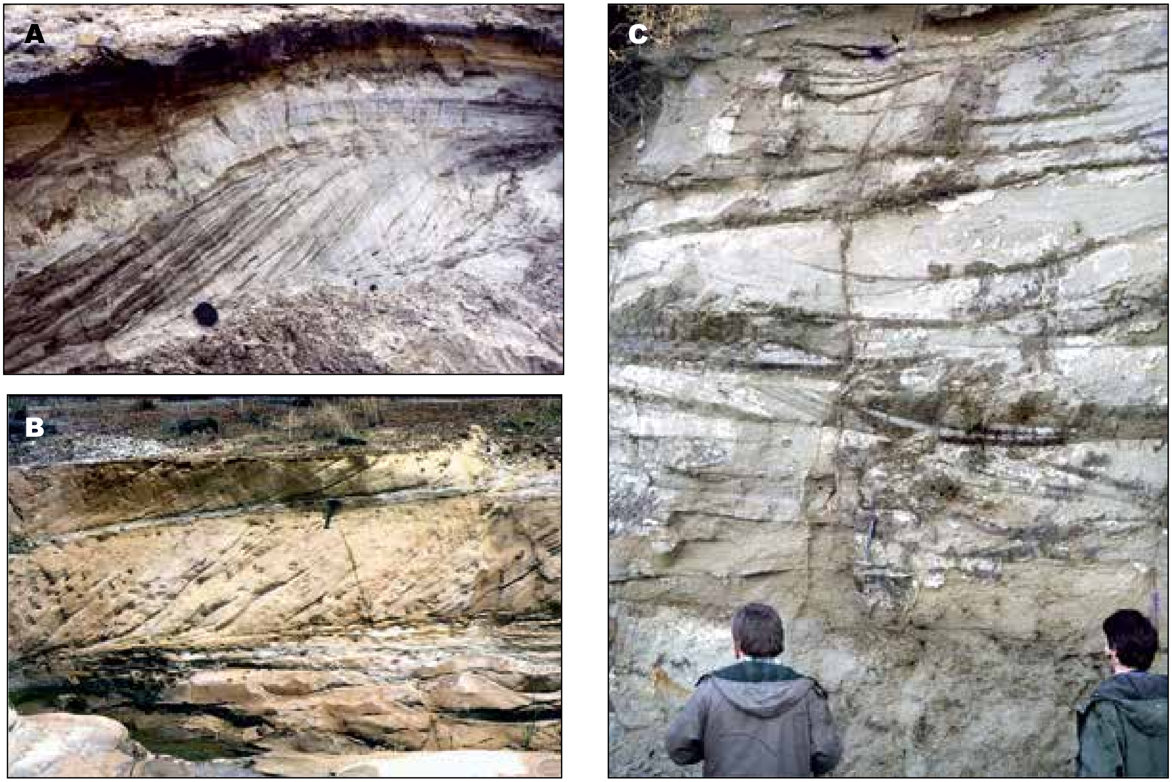


Figure 6.46 Sets of cross bedding with drapes of mud on the sandy foresets. A) The proportions of sand and mud vary along the set, possibly reflecting spring-neap cyclicity in a tidal regime. Lower Cretaceous, Kong Karls Land, Svalbard. B) Sets of medium-scale cross bedding with mud drapes on some foresets. The mud drapes thicken towards the toes of the forests in some cases. Cretaceous, Nigeria. Photo courtesy of Gilbert Kelling. C) Mud drapes between foresets and on trough bases in trough cross-bedded sands. Marine Molasse, Oligocene, Switzerland.

Figure 6.47 A tabular set of cross bedding with asymptotic (tangential) bases to the foresets. Vadsø Group, Late Precambrian, Finnmark, Norway.

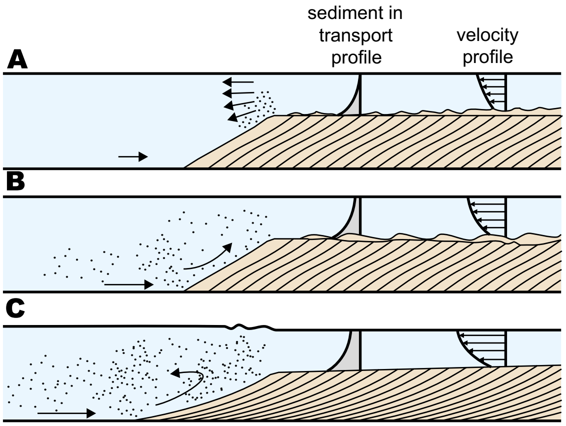


Figure 6.48 The changes in shape of the slip face of a small laboratory delta due to progressive increase in the velocity of flow over it from A to C. After Jopling (1965).

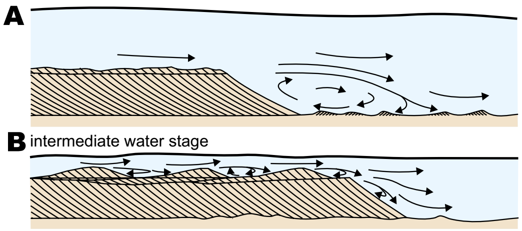


Figure 6.49 Changes in morphology and internal structure due to changing water stage over a sand wave. Note the counter current ripples developed in front of the bedform during high water-stage and compare the morphology with that in Fig. 6.44. Modified after Collinson (1970).

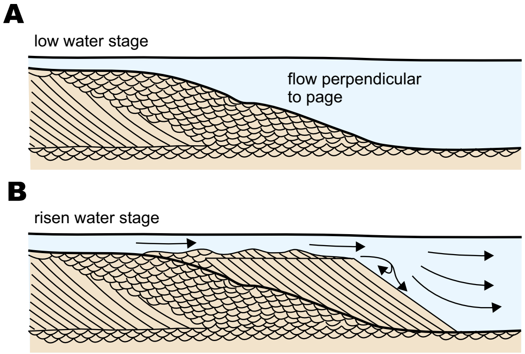


Figure 6.50 Changes in morphology and internal structure due to changing water stage over a sandwave. A) During low water stage when flow is sluggish, weak across channel currents may be aligned parallel with the front of a bedform, for example a linguoid sandwave, and can therefore deposit sand by lateral accretion. B) Downstream currents once again become dominant as the flow rises and deposition occurs by downstream accretion at the front of the bedform. Modified after Collinson (1970).

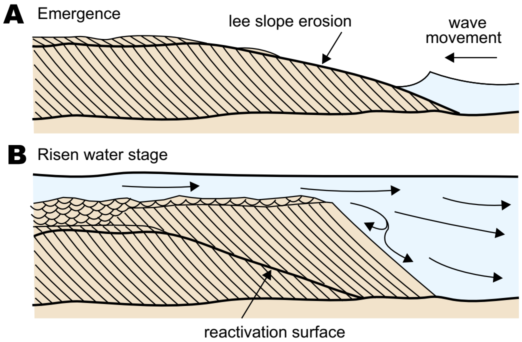


Figure 6.51 Changes in morphology and internal structure due to changing water stage over a sandwave. A) Emergence of bedform top and reworking by wave action to reduce angle of lee slope. Truncation of steeper foresets. B) Risen water stage and re-commencement of downstream bedform migration. Steeper foresets build over reactivation surface. Modified after Collinson (1970).

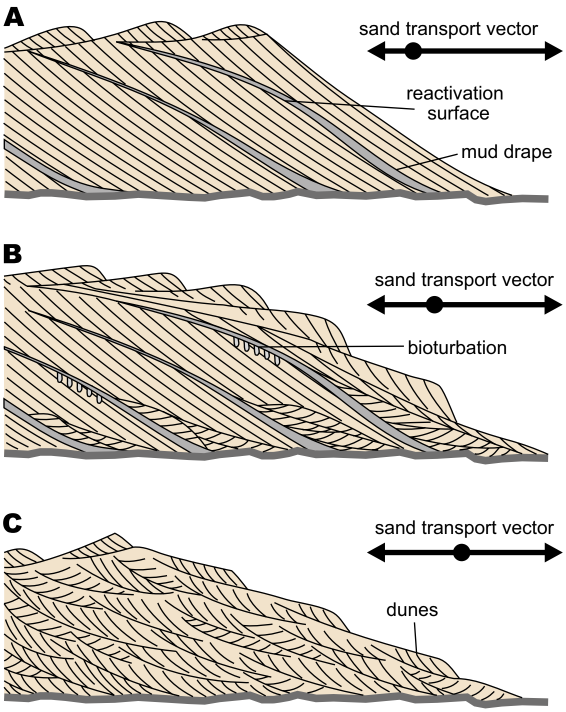


Figure 6.52 Models for the internal structure of tidal sandwaves influenced by bi-directional tidal flows of varying magnitude. A) Reactivation surfaces with mud drapes generated by a relatively weak current reversal. B) Reactivation surfaces with burrowing (bioturbation) and small dunes climbing back up the plinth of the sandwave during moderate current reversals. C) Sand wave influenced by strong current reversals, the product of ebb and flood tidal currents of equal strength. Modified after Allen (1980) and Tucker (2001).

in the internal cross bedding. Highly asymmetric bedforms are likely to show relatively simple, unidirectional cross bedding, reflecting the dominant tide, the activity of the subordinate tide being recorded only in reactivation surfaces in the upper parts of sets (Figs. 6.52A, 6.53). More symmetrical bedforms will have more complex cross bedding, showing a greater occurrence of reactivation and of sets of reversed foreset dips, leading to herringbone cross bedding (Figs. 6.52C, 6.54).

Herringbone cross bedding occurs predominantly in shallow sub-tidal areas where periodic reversals in the current direction occur due to tidal cycles (Fig. 6.54). Sets of herringbone cross bedding are sometimes separated by thin mud horizons that accumulate because sediment drops from suspension as the tidal current wanes to zero at high and/or low tide. The orientations of tidal flows are in large part controlled by the sea-floor morphology, as in the deeper part of an estuary or in an offshore strait.

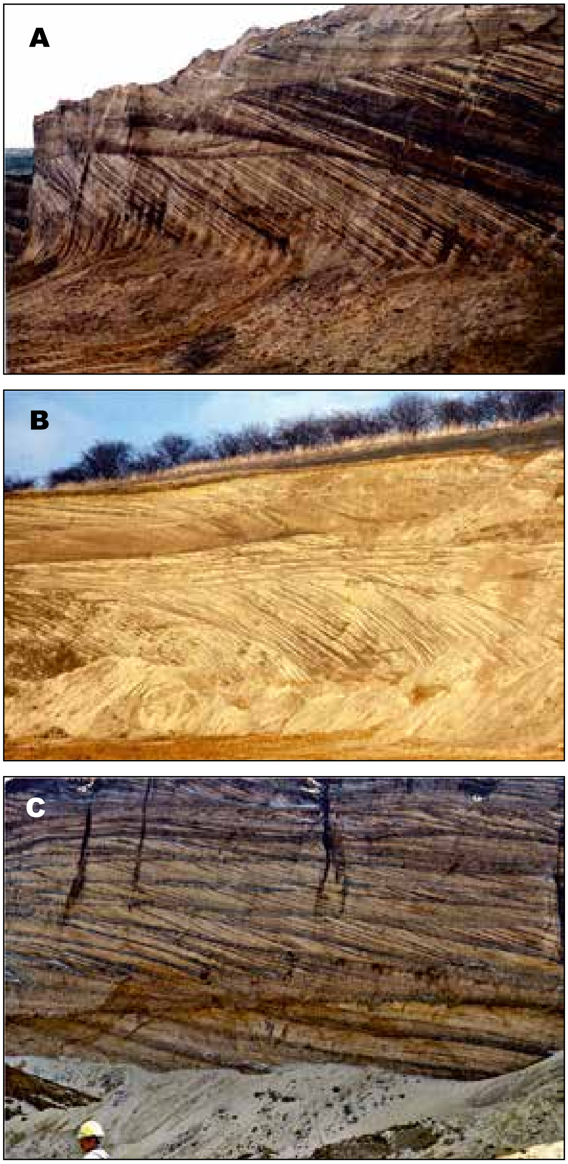


Figure 6.53 Cross bedding produced by tidal sandwaves. A) Broad trough-like sets with sweeping foresets. Internally, sets are broken by reactivation surfaces and there is apparent reversal between sets. Foresets are burrowed. Face about 6m high. B) Large scale cross bedding with multiple reactivation surfaces in large set and descending smaller sets at the right-hand side. C) Sets of cross bedding with low-angle foresets and set boundaries inclined in the same direction as the foresets. All examples from the Woburn Sands, Lower Cretaceous, Bedfordshire, England.

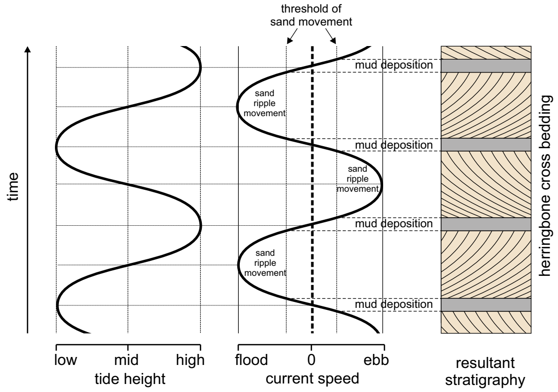


Figure 6.54 Variation in tidal current speed and direction over a tidal cycle and the resultant generation of tidal rhythmites for a position in the shallow sub-tidal realm. Deposition of mud occurs when the current speed is low (at both high and low tide as the tide turns). Deposition of cross stratified sands occurs when the current speed is high during mid-tides. The orientation of the cross strata is dictated by the current direction, which systematically reverses as the tide ebbs and floods, thereby generating sets of herringbone cross strata. The tidal currents in this example are of equal speed and strength in both directions, resulting in successive sand layers of equal thickness. As the difference in speed between the dominant and subordinate currents increases, the thickness of the two sand layers generated by each tidal cycle will become increasingly unequal. Modified after Dalrymple et al. (1991).

Mud drapes on bedform foresets result from fallout of suspended load at tidal slack water or during episodes of sluggish flow in rivers. This process is commonly accelerated by flocculation. When occurring in pairs in the shallow subtidal zone, mud or clay drapes record the slacks on either side of the subordinate tidal flow. The thicker sand layer, between drape pairs, is the product of the dominant tide, and its strength and duration may change through the lunar month to give bundles of systematic change as a consequence of spring-neap tidal cycles (Figs. 6.46, 6.55). The number of foreset increments per bundle broadly coincides with the tides of a lunar month and thus provides an internal indicator of bedform migration rate.

Sigmoidal foresets with preserved topsets (Fig. 6.43) record high rates of vertical bed accretion that was synchronous with bedform migration. Such forms are recorded from some fluvial as well as tidal settings. Parallel lamination on preserved topsets indicates upper flow regime conditions (see §6.4) on the top of the bedform. The whole assemblage may record a transition between the dunes and upper stage plane beds (Fig. 6.22).

A cross-bed set normally only preserves the lower part of the bedform that produced it. In some tabular sets, where there are preserved topset laminae, as in sigmoidal cross bedding, or ripple lamination that records ripples migrating on the top of the bedform, the set thickness may approximate the height of the bedform. For most examples, however, the set records only a proportion of the bedform height. This is particularly the case with trough cross beds where the foresets that fill the scour pit are those most likely to be preserved. Theoretical and experimental work suggests that, for sub-aqueous bedforms, about a third of the height of three-dimensional dunes is represented by the average cross-bed thickness, a figure largely controlled by the dynamics of migrating scour pits. The same analysis suggests that the average preserved length of trough cross-bed sets approximates to half the spacing (wavelength) of the bedforms.

Patterns of ascending, descending and laterally accreting sets of cross bedding reflect deposition on the flanks of major compound bedforms (bars) through the migration of dunes. Point bars, medial bars and side bars (Fig. 6.32) could all be sites of lateral accretion. Ascending sets (upstream accretion) are most likely to occur on the upstream sides of medial bars or side bars, whereas descending cross beds are likely at their downstream flanks. In addition, descending cross beds could reflect changing flow stage (see below and Fig. 6.51). A particular style of lateral accretion structure is dealt with in §6.2.10.

Bars, which form at the scale of the channel, probably result from the action of large-scale patterns of secondary water circulation in the channel flow. For example, the flow of water around a channel bend gives rise to a large-scale spiral vortex with surface flow directed towards the outer bank and near-bed flow directed towards the inner bank. This leads to movement of bedload sediment towards the inner bank and to the deposition of a point bar by lateral accretion. Similar processes are probably active along the flanks of medial bars. Converging and diverging flow at the upstream and downstream ends of medial bars also

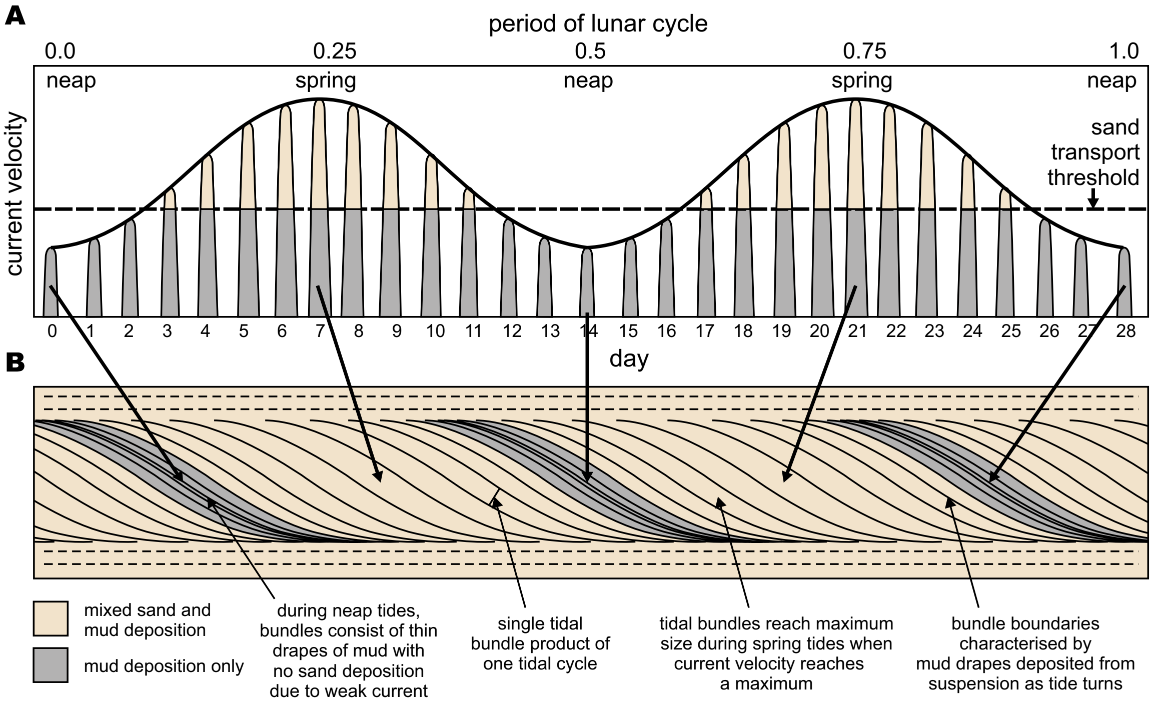


Figure 6.55 Schematic illustration of the mechanism by which sequences of tidal bundles are generated and vary in response to spring-neap tidal cycles. A) Changes in tidal current velocity over the period of a lunar month for a diurnal tidal system (i.e. one high and one low tide per day). Note how current velocity during neap tides does not exceed the sand transport threshold. B) The preservation of tidal bundles as foresets generated by the migration of a tidal sandwave. Note how the thickness of the individual bundles varies systematically in response to varying tidal current velocity. In this simple example, individual foresets represent single tidal cycles, each foreset being draped by mud deposited as the tide turns. In nature, sequences of tidal bundles are usually more complex than those depicted here, chiefly because most tidal systems are semi-diurnal (i.e. two high and two low tides per day), with one high tide being stronger than the other. Additionally, of the two spring tides in each lunar month, one is usually stronger than the other.

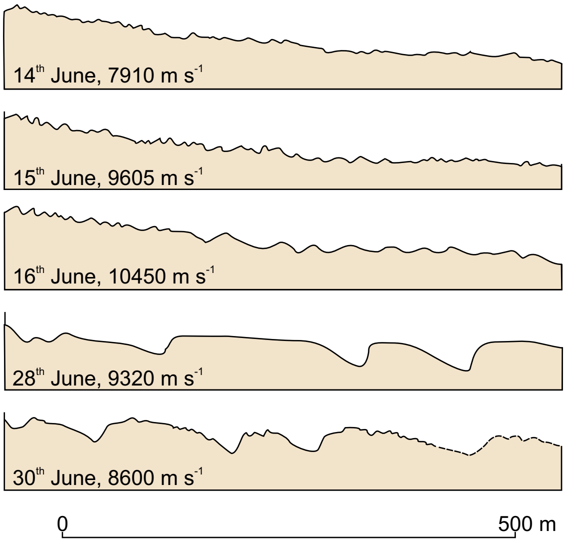


Figure 6.56 Echo sounding profiles made at different discharges of the flood on the Fraser River, British Colombia. The large bedforms increase in size during the rising discharge and continue to increase beyond the peak flood. During falling discharge, smaller forms develop superimposed on the backs of the large forms. Both these effects show how the bed response lags behind the prevailing flow due to the large volumes of sediment that must be reworked to modify large bedforms. Vertical exaggeration times 10. After Pretious and Blench (1951).

lead to deposition. Large alternate bars along the sides of straight channels develop in response to large-scale flow separation at their downstream ends, probably associated with development of large-scale spiral eddies.

6.2.8 Controls on bedform size

The sizes of dunes, unlike those of current ripples, are related to flow depth and are largely independent of grain size. Deep flows tend to generate higher and longer dunes. An approximate value of 1:6 has been suggested for the ratio of dune height to flow depth. This figure should be treated with caution as it results from a two-dimensional analysis when, in fact, many dunes are strongly three dimensional. Certain very low-relief dunes in sandy rivers show a much higher ratio (*c.*1:2) and the relationships are far from clear. The controls on dimensions of tidal sandwaves are not well established. Applying such relationships to the cross-bedding in the rock record to characterize flow conditions is fraught with uncertainties, not least because of the complications in translating cross-bed set thickness to bedform height, as discussed in §6.2.7.

A further complication is that dunes of different sizes may become superimposed as a result of changing flow

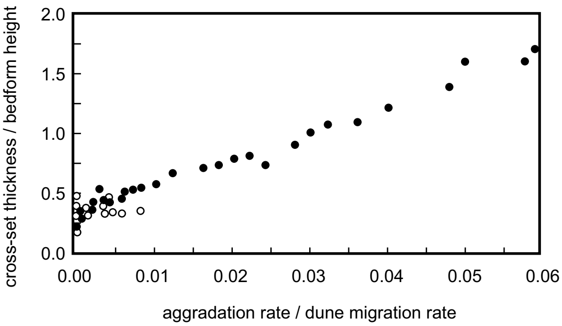


Figure 6.57 Relationship between the ratio of aggradation rate to mean dune migration rate, and the ratio of mean cross-set thickness to mean bedform height for subaqueous dunes. Open and black dots represent results from experimental and computer simulation runs, respectively. After Leclair (2002).

conditions (Fig. 6.56). When conditions change rapidly, large bedforms cannot adapt quickly enough to maintain equilibrium. The growth of smaller dunes on larger ones during falling stage could lead to descending cross bedding at the downstream end of the larger forms. The deposits of such dynamically changing settings are controlled by quite complex relationships between bedform dimensions, migration rates, bed aggradation rates and preserved cross-bed set dimensions (Fig. 6.57).

6.2.9 Isolated large-scale sets of cross bedding

Very large sets of cross bedding occur in both aeolian and water-lain sediments. Aeolian examples are discussed in §6.3 and care should be taken to establish this basic distinction. In water-lain sands, large, single tabular sets, several or even tens of metres thick are commonly overlain by a coset of smaller sets showing a similar current direction. Sets can be very extensive laterally: some sets 20–30m thick may be traced for several kilometres parallel to the dip direction (Fig. 6.58). Such sets are not explained by the migration of repetitive bedforms and at least two alternatives must be considered.

The first is by the advance of a delta formed by a stream that carried abundant bedload into quieter water such as a lake. Such deltas commonly have steep slopes at the angle of rest where avalanching causes the delta to advance and create a single, cross-bedded set. These deltas can easily be modelled in a laboratory tank (cf. Fig. 6.47) and were first recognized around lake margins by G. K. Gilbert in 1885. Examples of Gilbert-type deltas are increasingly recognized in the rock record and are especially well documented from Pleistocene and Holocene gravel-dominated

Figure 6.58 Different types of large-scale cross bedding. A) Large-scale set of cross-bedding, sharply overlain by a coset of medum-scale cross bedding, probably the result of the migration of a large-scale bar form in a major river channel. Set is about 8m thick. Kinderscout Grit, Upper Carboniferous, Derbyshire, England. B) A large isolated set of cross bedding with discontinuities amongst the foresets and overlain sharply by medium scale sets. The unit extends laterally for around 1km and is underlain and overlain by marine mudstones. It is thought to record the progradation of a sandy Gilbert-type delta into a shallow marine basin. Roda Sandstone, Eocene, Pyrenees, Spain.

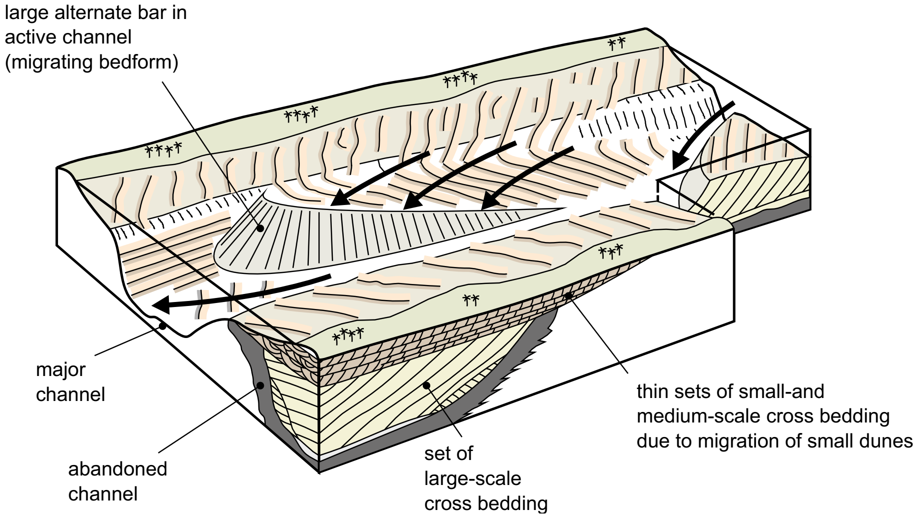


Figure 6.59 Possible model for the generation of very large sets of cross bedding by the migration of alternate bars in deep channels. This model was devised to explain large cross-bedded sets in the Upper Carboniferous rocks of Northern England. Modified after McCabe (1977).

successions where they form notable components in some pro-glacial settings. Their initial description, however, caused confusion as some geologists came to regard all cross bedding as diagnostic of deltaic sedimentation, a view quite commonly found in books from as late as the first half of the twentieth century. Although deltas dominated by coarse-grained sediment may indeed show cross bedding in their slope deposits, most large marine deltas are dominated by muds and silts whose deposition from suspension in front of river mouths leads to very low-angle delta-front gradients. Fan deltas form where alluvial fans with steep surface gradients pass directly across a shoreline into a lacustrine or marine water body. The dominant deposits of many such fan deltas are rudites (see §7.3.4), though bodies composed sand-grade deposits are also known.

The second possible origin of large-scale, isolated sets is by the advance of large bars either attached to alternating sides (so-called alternate bars) of very large, deep channels (Fig. 6.59) or in mid-channel or channel confluence settings. Bars of this sort occur on a small scale in modern streams and scaling up of these analogues, to sizes beyond anything known at the present day, is necessary to explain large-scale examples in the ancient record.

6.2.10 Epsilon cross bedding

This type of isolated, single set of inclined strata differs from ordinary cross bedding in several important respects. It consists of a tabular unit, usually between 1m and 5m thick, in which inclined beds dip at angles considerably less than the angle of rest, commonly 5° to 15°; the thicker the unit, the lower the inclination. The inclined strata, which typically extend over the full thickness of the unit, may be sigmoidal in profile and are commonly defined by conspicuous differences in grain size between adjacent layers (Fig. 6.60). In detail, the inclined beds usually contain smaller-scale internal structures, such as cross lamination and small-scale cross bedding, which indicate flow sub-parallel to the strike of the inclined beds. The near-horizontal basal surface of the set is erosional, commonly with a concentration of pebbles or intraformational clasts. There may be an overall upward-fining of grain size through the set to the extent that the upper part of the unit has interbedded inclined layers of sand and silt whereas the lower part is dominantly sandy. In rare examples, where upper bedding surfaces are extensively exposed, the inclined beds are seen to be strongly curved through several tens of degrees in plan view (Fig. 6.61). Sets commonly end laterally, at their down-dip ends, with erosion surfaces that dip steeply towards the inclined beds but are separated from them by a unit of fine grained sediment that may be post-depositionally disturbed.

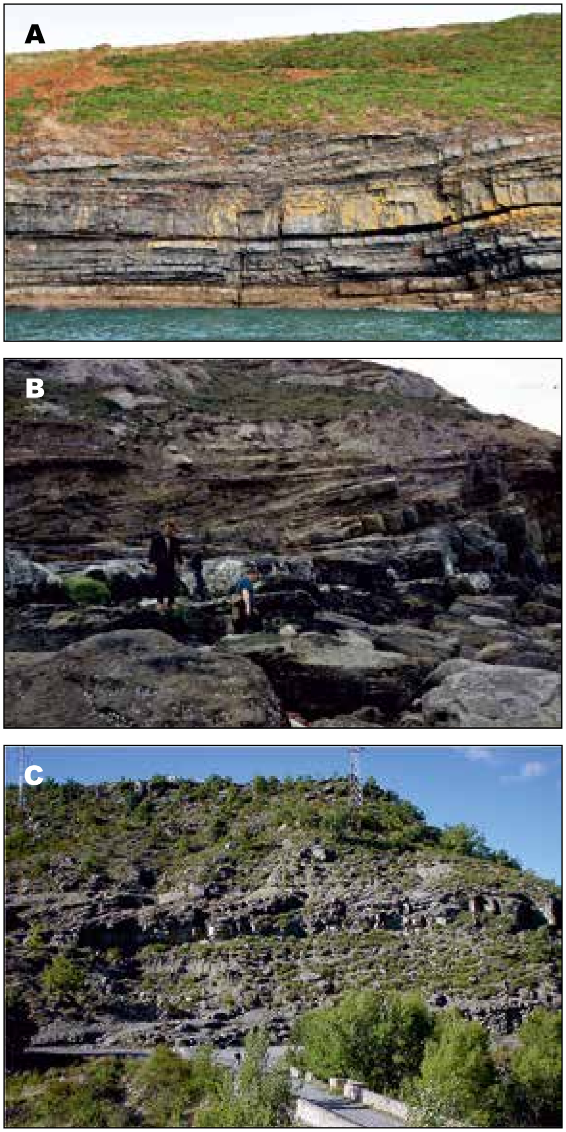


Figure 6.60 Examples of epsilon cross bedding. A) A unit of epsilon cross bedding. The base of the set is a roughly horizontal erosion surface. The surfaces dipping to the left represent successive positions of the depositional bank of a channel as it migrated laterally. Smaller structures within the inclined sand units indicate flow parallel to the strike of the inclined units. Cloughton Formation, Middle Jurassic, Yorkshire, England. B) Epsilon cross-bedding, in the upper part of the exposure, within a turbidite succession showing that deep-water channels can migrate laterally. Ross Formation, Upper Carboniferous, western Ireland. C) Epsilon cross bedding, Montañana Group, Eocene, Pyrenees, Spain.

Figure 6.61 Upper bedding surface of a sandstone unit made up of several laterally adjacent sets of epsilon cross bedding. The beds within each set are inclined in the direction of convex curvature, suggesting that the epsilon cross bedding is due to lateral accretion on the point bar of a meandering channel. Scalby Formation, Middle Jurassic, east Yorkshire, England.

This type of cross bedding is an organized assemblage of lithologies and structures. The similarity to the lateral accretion seen in more uniform cross-bedded sandstone is obvious (§6.2.7). The critical point in its interpretation is that the smaller-scale structures indicate flow sub-parallel to the strike of the inclined surfaces. From this, one can infer that an inclined depositional surface migrated laterally, transverse to the main current (Fig. 6.62). The basal erosion surface and the curved plan view of the dipping beds combine with this inference to suggest a curved channel side, most commonly in a meandering stream with a high-sinuosity channel. The inclined beds record successive positions of a laterally migrating point-bar surface or, more unusually, lateral migration of the flank of a medial bar. Present-day examples can often be seen in banks of rivers and tidal creeks.

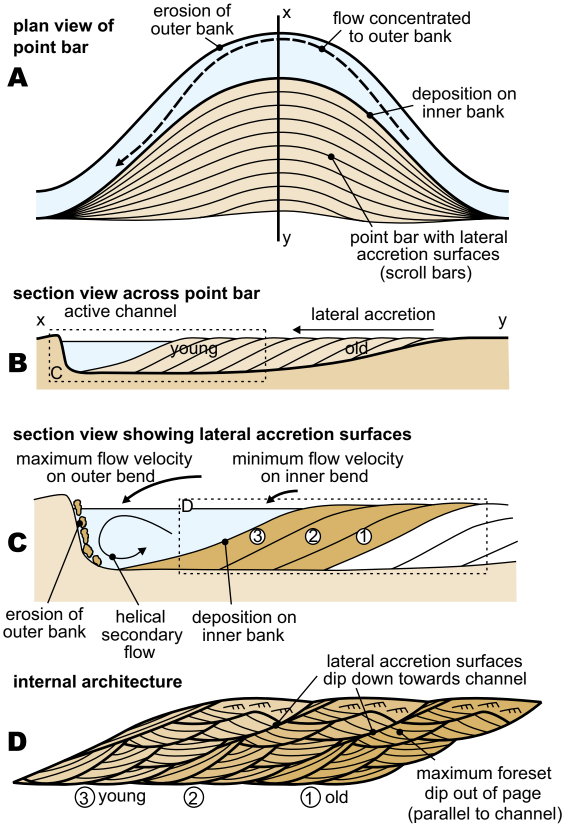


Figure 6.62 Schematic illustration of the development of lateral accretion surfaces through the lateral migration of a river channel. A) Plan view depicting a point bar with lateral accretion surfaces (scroll bars) revealing the former positions of the channel. B) and C) Section views depicting the lateral migration of the asymmetric channel through time by erosion from the outer bank and deposition on the inner bank. Flow velocity is greatest on the outer bend because water is slung to the outer bank by centrifugal force. D) The internal architecture of lateral accretion elements reveals sets of planar and trough cross-bedding deposited by sluggish flow on the inside bank of the river. Internal foresets are inclined in orientations at a high angle to that of the lateral accretion surfaces that bound them. This signifies that lateral accretion of the inner bank occurs at a high angle to flow direction. The lateral accretion process can also operate on the side-facing flanks of mid-channel bars.

6.2.11 The form and deposits of point bars in relation to channel bend evolution

In fluvial (river), tidal and submarine settings, the relationship between the form of meandering channel bends and their associated deposits record the transformation of a channel's position over time. Several fundamental types of transformations are recognized. Expansion occurs where a channel bend, which takes the form of a meander loop in plan-view, increases its amplitude over time by growing sideways relative to the overall trend of the channel system (Fig. 6.63). Translation occurs where a channel bend (meander) migrates in an orientation parallel to the overall trend of the channel system. Translation usually occurs in a downstream direction. Rotation occurs where a channel bend with the form of a meander loop rotates on its axis over time. Combinations of these transformation types can operate coevally or successively during the evolution of a channel bend. As such, the relationship between the migratory behaviour of an evolving meandering channel and the form of the sediment bodies accumulated as a consequence of that migration can be complicated. The growth of meander loops by simple expansion gives rise to point bars characterized by the epsilon cross bedding described

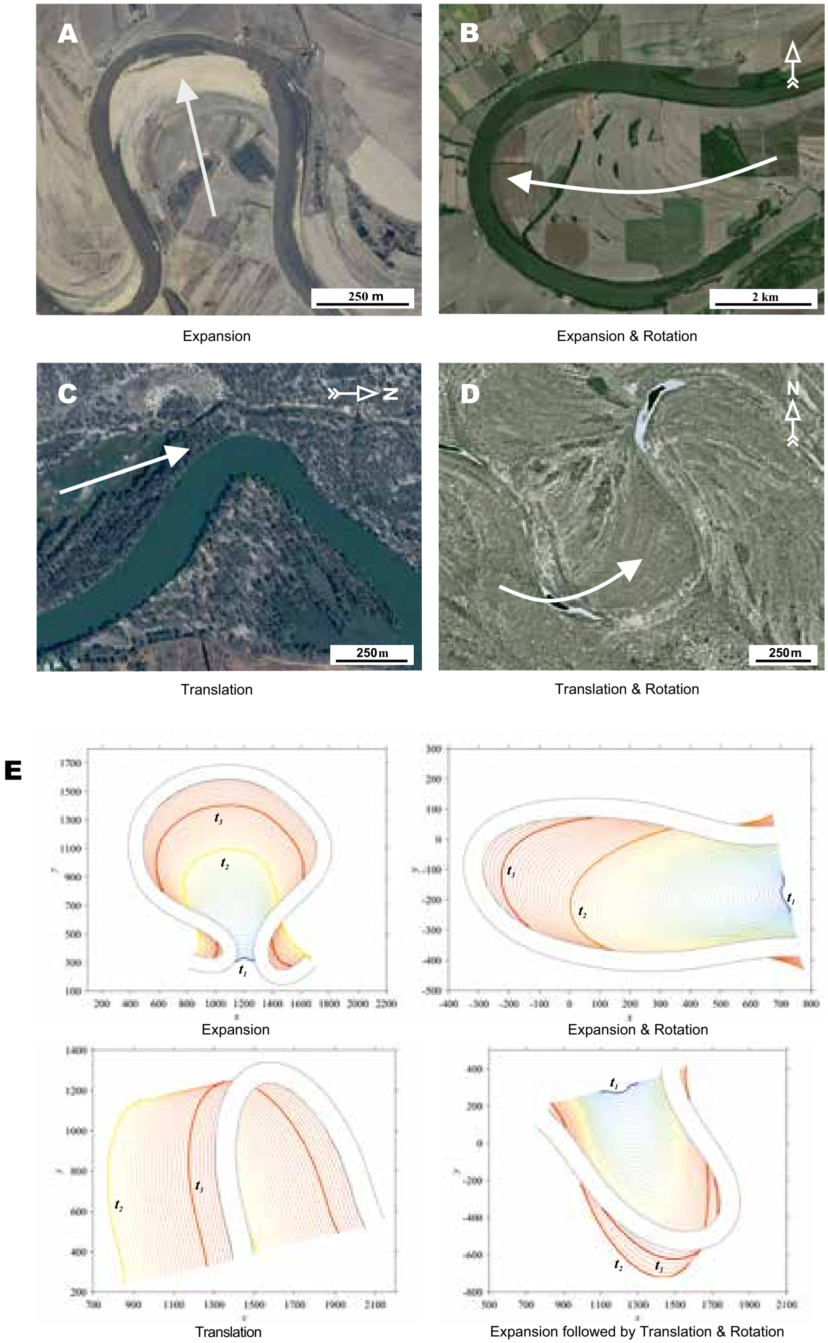


Figure 6.63 Spatio-temproal evolution of high-sinuosity meander bends. Four basic types of meander-bend transformations. The arrows show the migration direction of meander bends. Aerial photo examples from modern rivers (Google Earth™): A) Songhua River, China (46°38′ N, 126°30′ E); B) Mississippi River, USA (34°25′ N, 90°46′ W); C) Murray River, Australia (34°14′ S, 142°15′ E); D) Rio Negro, Argentina (39°49′ S, 64°56′ W). E) Numerical model output revealing meander bend development and resultant trajectories. The position of the meander bend at different times (t1 [earliest], t2, and t3 [latest]) are shown in each case. A colour spectrum (dark blue to dark red) is used to differentiate meander positions at different times. The dimension here is arbitrary, but the modelling results can be readily scaled to physical units by using data from field measurements or remote sensing. The shape of the modelled point bars is comparable with those found in the real world. Modified after Yan et al. (2017).

above (see §6.2.10, Fig. 6.62). By contrast, downstream translation of a meander loop favours the accumulation of so-called outerbank bars, also known as counter point bars or concave benches. Study of what are commonly large-scale sedimentary structures associated with channel bend evolution is not straightforward. However, a small number of exceptional outcrops expose both vertical and horizontal sections, though even these typically provide only fragmentary 'windows' that reveal only a minor part of large and complex three-dimensional geological bodies. More extended discussions of the processes involved are given in books dealing with sedimentary environments (see reading lists of Chapters 1 and 10).

6.2.12 The application of sandwaves, dunes and cross bedding

There are three main applications for the understanding of dunes, sandwaves and cross bedding. They can be used as indicators of way-up, of conditions of deposition, and of palaeocurrent direction.

Way-up

Where dunes are preserved on upper bedding surfaces or as form sets, they give a fairly certain indication of "way-up". Cross bedding is usually an even better indicator. In particular, the sharp cut-off of foresets at the top of many sets contrasts with their tangential bases, particularly in trough sets. Only in the case with sigmoidal foresets might there be scope for confusion.

Conditions of deposition

Dunes form under particular conditions of water depth, flow velocity and grain size (Fig. 6.22). It should therefore be possible to put limits on flow conditions based on the forms seen on present-day sand beds and on the cross bedding preserved in rocks. The shape of the foresets in tabular sets can indicate relative current strengths, and changes in foreset shape along a single tabular set could indicate fluctuation of current strength through time. These changes may be associated with reactivation surfaces if depth fluctuation was large. Complex and intensive reactivation, opposed foreset dip directions and clay drapes can all point towards tidal influence.

Although dune height relates, albeit loosely, to flow depth, there are problems in using the thicknesses of trough sets as indicators of flow depth. The average set thickness may correspond to roughly a third of bedform height but trough sets, by their very nature, may not always display their true thickness in vertical section. For example, in a borehole core or where a cliff section that provides exposure only clips the edge of a trough, the true thickness at the trough axis will not be evident. Great care is therefore needed in making reconstructions of palaeo-hydrodynamics. A systematic upwards reduction in set thickness through a coset may, however, suggest a shallowing flow.

Direction of palaeocurrents

Cross bedding is one of the most widely used palaeocurrent indicators. As large bedforms usually respond to a dominant flow and are not easily remoulded by low-stage flows they tend to give a good indication of the palaeoslope.

With tabular sets, the most valuable measurement is the direction of dip of the foresets (foreset azimuth), but it is also useful record the magnitude of dip as well, particularly if the succession is tectonically tilted. To reliably measure cross bedding in vertical sections, it is necessary to see faces with more than one orientation (e.g. Fig. 6.34). The apparent dip on a single face only shows a component of the true dip. A bedding surface view of the foresets will always give the most accurate measurement of foreset azimuth. Also bear in mind that the slip faces of many large-scale bedforms are strongly skewed and that foreset dips may diverge considerably from the true downstream direction. It is therefore important to collect measurements from several sets if a representative direction is needed.

With trough sets these problems are compounded by the curved nature of both the set boundaries and the foresets. To obtain the most reliable indictors of migration directions, it is best to measure the trend of trough axes on bedding planes (e.g. Figs. 6.36, 6.39). With experience, however, it is possible to judge the orientation of trough axes from vertical exposures to an accuracy of ±15°, which is adequate for many purposes.

Where cross bedding occurs between bounding surfaces which are themselves inclined, the relative orientation can provide evidence of the nature of accretion on larger bedforms. It is very important to recognize epsilon cross bedding and to distinguish it from normal cross bedding. An uncritical measurement of dip direction could suggest a palaeocurrent 90° divergent from the true trend.

6.3 Flat beds and parallel lamination

6.3.1 Introduction

Many sandy surfaces in modern aqueous settings and many bedding planes in sandstone are completely flat. These planar surfaces are usually related to parallel lamination within the underlying deposit.

Flat-bedding surfaces and parallel lamination occur mainly in sands and sandstones of fine-medium grain size, including those rich in mica. They can, however, occur in sediment up to very coarse sand size.

6.3.2 Flat-bed morphology and primary current lineation

On beaches with very little relief, on flat areas of a recently exposed river bed and on the bedding surfaces of parallel-laminated sandstone, it is common to see subtle small-scale relief with a distinct linear pattern. These features are particularly conspicuous with low-angle lighting. In very coarse-grained or slightly pebbly sandstone, a lineation is most commonly developed on the surfaces of slightly finer-grained layers, whereas in micaceous sandstone, mineral segregation often gives a colour lineation as well as a relief. This lineation is **primary current lineation** or **parting lineation** (Fig. 6.64).

The lineation comprises a series of closely spaced ridges and hollows. Typical spacing is a few millimetres and relief is of the order of the grain diameter. Individual ridges and hollows persist parallel to the lineation for a few centimetres or even tens of centimetres. There are no systematic differences between opposite ends of ridges and hollows and they cannot be used to determine the sense of current direction.

6.3.3 Internal structure: parallel lamination

Excavations into flat sand surfaces and vertical sections in sandstones with flat bedding planes usually show parallel lamination. Laminae are typically only a few grain diameters thick and, in coarse, less well-sorted sandstones, they may barely exceed the thickness of the coarsest grains.

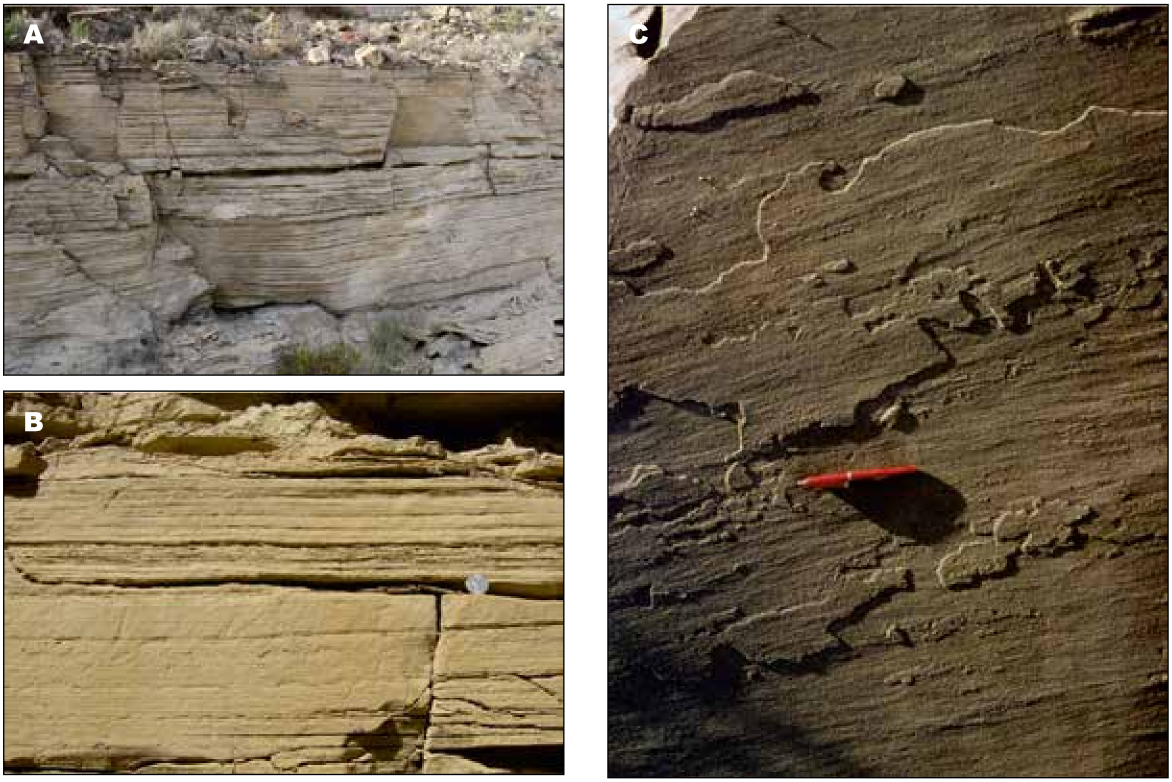


Figure 6.64 Parallel lamination and primary current lineation. A) Thin parallel horizontal lamination in fine-grained sandstone, reflecting deposition under upper flow-regime flat-bed conditions. Villar del Arzobispo Formation, Jurassic, eastern-Spain. B) Thinly, parallel laminated, fine-grained sandstone with hints of transition to ripple lamination. Neslen Formation, Cretaceous, Utah, USA. C) Primary current lineation or parting lineation on the bedding surface of parallel-laminated sandstone. The steps between adjacent laminae trend parallel to the lineation. Cloughton Formation, Middle Jurassic, east Yorkshire, England.

The laminae are defined by slight grain-size differences or by concentrations of mica and finely broken carbonaceous debris.

6.3.4 Process of formation

A flat bed is a distinct bedform, produced by particular flow conditions. Where the sand is medium- to fine-grained and reasonably free of mica, those conditions are of high flow velocity and shallow water depth, the so called "upper flow-regime flat bed" mode of transport (Fig. 6.22). As flows accelerate into these conditions ripples and dunes are destroyed and turbulence appears suppressed. The water surface takes on a smooth, glassy appearance and the conditions are similar to those of "rapid flow" (§3.2.6). Similar conditions also develop during the deposition of sand from turbidity currents, and parallel lamination is a feature of many turbidite sandstone beds.

The lineations, which run parallel to the flow probably relate to streaks of faster- and slower-moving water close to the bed, such as occur in the viscous sub-layer (§3.2.4). As such, a sub-layer only occurs with fine-grained sediments. By contrast, in coarse-grained sands, alternating spiral vortices close to the bed can be responsible for the lineations (Fig. 6.65).

Experimental work shows that ripples do not form in sands coarser than about 0.6mm, but that movement occurs on a so called "lower flow-regime flat bed" for currents above the critical erosion velocity and below those forming dunes (Fig. 6.22). Little is known about the ability of this mode of transport to produce parallel lamination and primary current lineation.

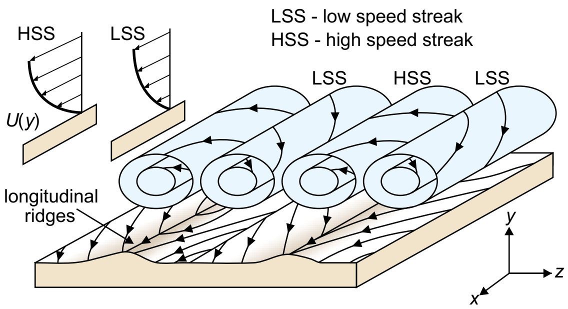


Figure 6.65 Idealized fluid motion associated with streakiness within the near-wall region of a turbulent boundary layer. Streaks of fluid moving at slightly different velocities initiate longitudinal vortices which cause locally converging flow where grains accumulate into longitudinal ridges. This longitudinal fabric produces primary current lineation on bedding planes. Modified after Allen (1985).

Other experiments suggest that abundant mica inhibits the formation of ripples whose existence depends, amongst other things, on the ability of grains to avalanche down the lee faces of the bedforms. The hydrodynamic interpretation of micaceous sands may therefore, be less straightforward. Some highly micaceous, parallel-laminated sands and sandstones could reflect deposition from suspension, but concentration of mica into layers implies sorting on the bed, possibly in conditions that might have produced ripples in the absence of mica. Lineation in such micaceous sands may be due to high- and low-velocity streaks in the viscous sub-layer.

Although the flat-bed mode of transport is most readily envisaged for unidirectional flow, waves can also lead to the development of a flat bed. The flat beds observed on exposed beaches relate to unidirectional backwash, but high-energy waves seaward of the surf zone can also give such a bed.

By its very nature, the formation of lamination demands some process of grain-size segregation on the bed. Fluctuations in flow strength may be involved, but it is also possible that grain segregation on the bed involves layers of moving grains with different size characteristics which deposit laminae when they stop. Another form of roughly parallel, horizontal lamination is produced by multiple freezing of a traction carpet, as discussed in §6.7.4. Parallel lamination in aeolian successions is dealt with in §6.8.4.

6.3.5 Uses of flat beds and parallel lamination

Flat beds and parallel lamination have little use as indicators of way-up, but they do provide useful information on (a) current strength and (b) palaeocurrent or palaeowave trend (though not usually absolute direction).

(a) Flat beds with parallel lamination indicate upper flow-regime conditions, provided that the sand is in the medium to fine range and that it is not very micaceous. With coarser-grained or very micaceous sands, lower flow-regime conditions may have applied. Flat beds due to waves and currents cannot be differentiated by internal characteristics.

(b) To establish a palaeocurrent direction from flat beds it is necessary to see the primary current lineation. Because this is a simple, linear feature, it only indicates the trend and not the sense of flow.

6.4 Undulating smooth surfaces and lamination

6.4.1 Introduction

On some present-day sand surfaces and sandstone bedding surfaces, a gentle wave form is found associated with parallel but gently undulating lamination. This uncommon structure is probably confined to sands but its rarity prevents clear delineation of any grain-size limits. A similar style of bedding is sometimes associated with pyroclastic ash deposits of sub-aerial base-surge flows.

6.4.2 Morphology

The sand surface, or more commonly the sandstone bedding surface, is smooth with a gentle undulation, either two-dimensional waves or three-dimensional domes and hollows. The amplitude is commonly a few centimetres, and the wavelength tens of centimetres, sometimes being of the order of one metre (Fig. 6.66). The bedding surfaces commonly show a primary current lineation (see §6.3.2) which, in the case of the two-dimensional waves, is roughly normal to the wave crests. In vertical section, lamination is thin and roughly parallel to the surface undulation. Slight divergences in otherwise parallel-looking lamination may be related to this structure even where truly undulating surfaces are not seen. Low-angle cross bedding may sometimes be present below the waves or domes.

6.4.3 Processes of formation

The association of primary current lineation with this structure and its general similarity to parallel lamination suggests a related origin. If the velocity of water flowing over an upper flow-regime flat bed increases, a pattern of ephemeral wave forms develops on both the water and the underlying sediment surfaces.

It is common to see small examples of these waves in streams cutting across sandy beaches or in storm water flowing down gutters when gradients are quite high. They also occur at a larger scale during torrential flooding in rivers or in sheet-like flows. The dimensions of the water surface waves depend on the water depth and flow velocity. Where the waves are developed over a mobile bed, the waveform on the sediment surface is in phase with the water surface wave but is of lower amplitude (Fig. 6.67). When the position of the waves is relatively stable, they are known as **standing waves**. However, it is much more common for these waves to move upstream rather abruptly. In doing so the water surface wave may break and collapse giving a short-lived, flat water surface from which new waves quickly grow. When the waves move upstream in this way, they are called **antidunes**. The underlying sediment wave is usually destroyed during the breaking phase and it is clear that the chances of preservation of undulatory lamination formed under such flow conditions are very low. Antidunes may in rare cases give low-angle cross bedding that is inclined upstream.

6.5 Hummocky and swaley cross stratification

6.5.1 Introduction

Hummocky and swaley cross stratification is now widely recognized as an important and diagnostic structure

Figure 6.66 Undulating lamination associated with thin, roughly parallel lamination in sandstone. Primary current lineation occurs on the bedding surfaces. Moränesø Formation, Late Proterozoic, northeast Greenland.

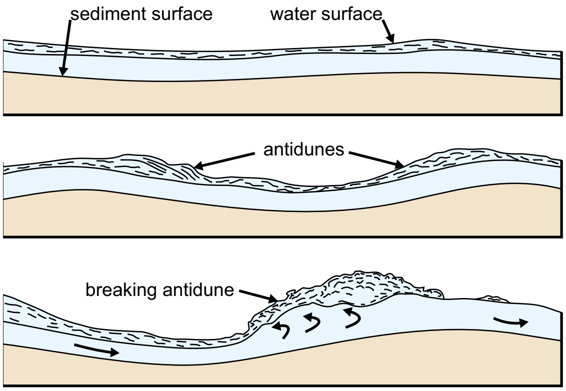


Figure 6.67 Development of standing waves and antidunes as observed in a laboratory flume experiment. Note how the water and sediment waves are in phase with one another and how the antidunes break in an upstream direction. Each section is 1 m long. Modified after Kennedy (1961).

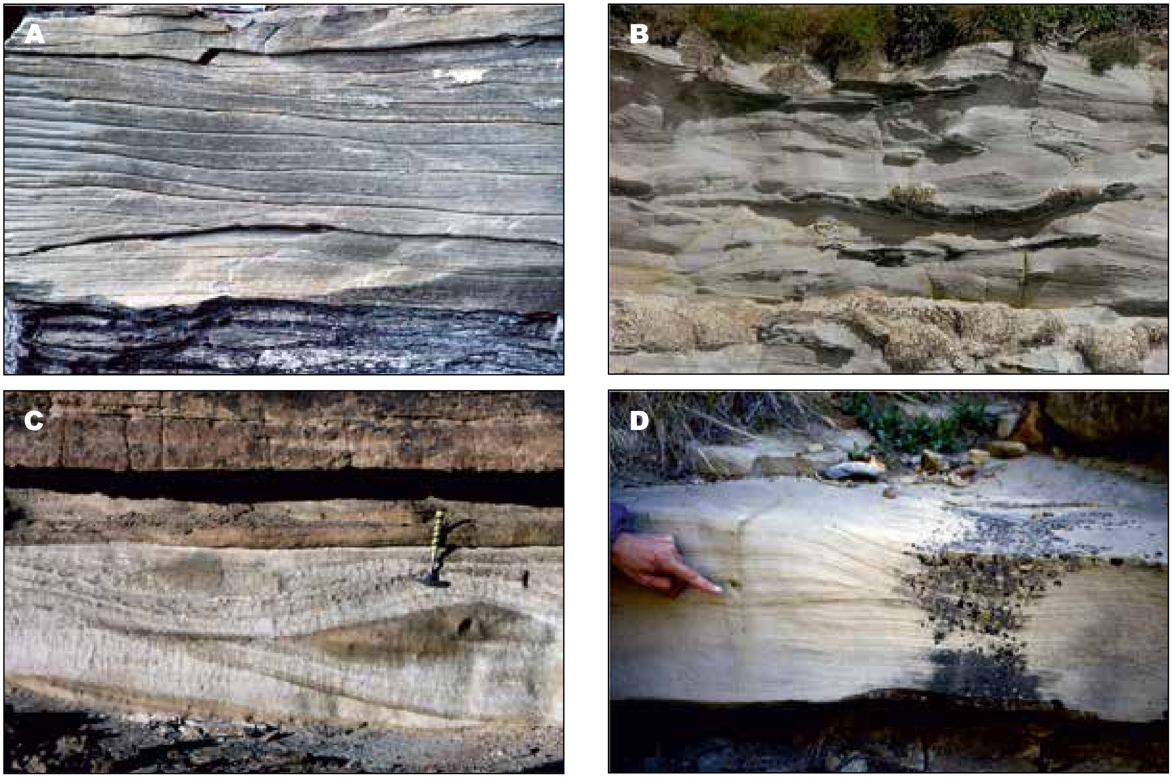


Figure 6.68 Examples of hummocky cross stratification (HCS) and swaley cross stratification (SCS). A) Low-angle undulating lamination in a sharp-based sandstone bed Width of view is ~0.8m. Carolinafjell Formation. Cretaceous, Spitzbergen, Svalbard. B) Thick interval of fine grained sandstone with pervasive mainly swaley cross stratification, probably deposited on a high energy shoreface Width of view is ~3m. Alston Formation, Lower-Carboniferous, Northumberland, England. C) Nubian Sandstone, Lower Palaeozoic, Sinai. Photo-courtesy of Gilbert-Kelling. D) Spring Canyon Member, Blackhawk Formation, Cretaceous, Book Cliffs, Utah, USA.

in ancient sandstones, although for a long time it was vaguely dismissed as "wavy", "irregular" or "undulating" bedding or lamination. These closely related structures are most common in fine- to medium-grained sandstone of shallow-marine or nearshore origin. They occur both within thicker sandstone units and within sharp-based sandstone beds of interbedded sandstone/mudstone successions.

6.5.2 Morphology

The structure comprises sets of curving lamination with both convex-up (**hummocks**) and concave-up (**swales**) sectors. The laminae seldom dip at more than 12° and sets intersect one another at low angles (Figs. 6.68, 6.69). Laminae may thicken into swales and thin over hummocks, so that the undulations gradually die out upwards. Heights of undulations seldom exceed 20cm and wavelengths are of the order of 1m. In many cases there is no apparent preferred orientation to the inclination of laminae, suggesting a more or less random three-dimensional pattern. In some cases however, a preferred orientation is apparent giving a form of low-angle cross bedding. Where the structure occurs in sharp-based sandstone beds, there is sometimes evidence of erosion in the form of sole marks. Upper contacts are broadly horizontal and are often characterized by wave ripples.

6.5.3 Process of formation

The occurrence of the structure in a shallow-marine setting (as deduced from associated fossils, trace fossils and overall context) and its clear association with wave action and an episodic style of deposition (as shown by interbedded sequences) have led to its interpretation as the product of strong and complex wave activity, mainly in areas below fair-weather wave base. In interbedded sandstones and mudstones its occurrence suggests a phase

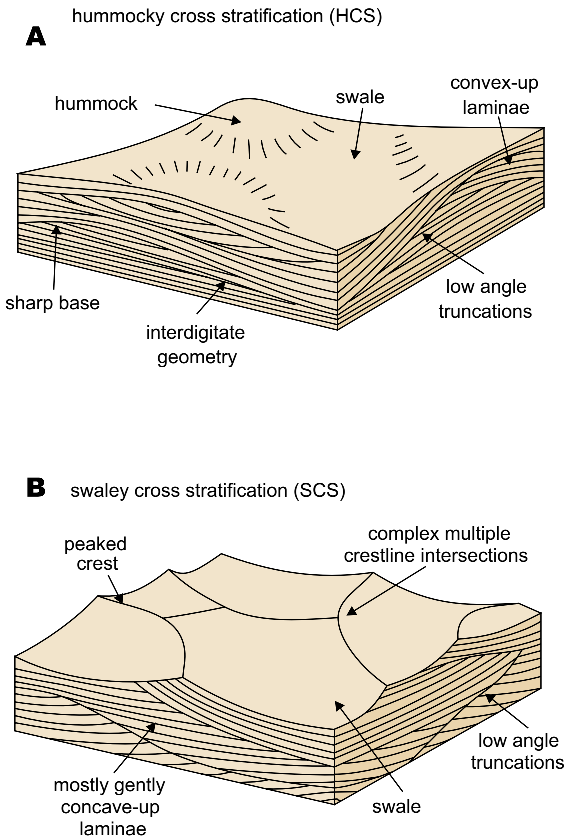


Figure 6.69 The internal architecture of hummocky cross stratification (HCS) and swaley cross stratification (SCS) and their relationship to external morphology. Laterally adjacent, alternating hummocks and swales are common in nature and generate mixed HCS-SCS structures.

of vigorous activity, which eventually decayed into lower energy wave oscillation. This is most likely in a storm, at the peak of which wave action is most vigorous. Storm bedforms are not well described for obvious reasons, but the inference from these ancient examples is that a complex pattern of erosion and rapid deposition occurs on an irregular undulating surface. Examples showing a recognizable preferred orientation to the inclined layers perhaps indicate a coexistence of strong wave action and a unidirectional current, a form of combined flow. Arguments about how the sand was transported to the site of deposition during storms are unresolved with wind-driven currents, storm surge and turbidity currents all being advocated. Such discussions demand a wider knowledge of the context of particular examples and are beyond our scope here.

6.6 Massive sand beds

6.6.1 Introduction

The structures of sands and sandstones described and discussed so far are all associated with well-defined lamination. There are, however, many sands and sandstones which lack recognizable lamination and which are therefore described as **structureless**, **massive** or **unlaminated**.

6.6.2 Description

These beds may occur in sand and sandstone of virtually any grain size or sorting. Some massive beds are lenticular, whereas others are parallel-sided and may be interbedded with finer-grained sediment. In general, it is much more difficult to establish the absence of a particular feature (in this case lamination) than its presence. In looking at apparently structureless sandstone, perhaps one is simply not seeing the lamination. Is it not weathering out in that particular exposure? Would some more sophisticated technique of observation reveal hidden lamination? Staining, etching and polishing of cut surfaces can indeed reveal previously unnoticed structures. X-radiography and MRI scanning of thin slabs cut normal to bedding can be even more effective. Despite these methods, however, there are still beds that lack detectable lamination. In the field, therefore, it is reasonable to apply terms like massive, unlaminated and structureless, especially where there is a clear contrast between the structureless beds and neighbouring laminated beds that have undergone similar weathering.

6.6.3 Processes of formation

Absence of lamination may reflect conditions of deposition or it may result from destruction of original lamination. A lack of primary lamination most commonly results from rapid deposition, most probably through the deceleration of a heavily sediment-laden current. Grains arrive at the bed so rapidly that they are buried before any bedload movement can occur and give rise to sorting into laminae. A “frozen” debris flow may also appear structureless, particularly if it comprises a fairly narrow range of grain sizes.

Destruction of depositional lamination can come about through intense reworking of sediment by organisms living within it and also by physical disruption of waterlogged sediment due to liquefaction and flowage. In the case of organic reworking, burrows may be visible in adjacent sediments or may be revealed by x-radiography (see §9.4). Where lamination has been destroyed by liquefaction, structures due to associated water-escape may be present in nearby beds (see §9.2.2). In aeolian sandstones, remnant blocks of brecciated sand or plastically folded patches occur within otherwise structureless sand and suggest a secondary destruction of lamination.

6.7 Normally graded beds, inverse grading, the Bouma sequence and hybrid-event beds

6.7.1 Introduction

Certain sharp-based sandstone beds, most commonly in interbedded sandstone/mudstone successions, show an assemblage of grain-size changes and sedimentary structures, which together are highly diagnostic of depositional processes. Such beds occur in a wide range of depositional settings and can involve sandstones ranging in grain size from very coarse and pebbly sand to very fine sand or even silt. The assemblage of features may involve both changes of grain size through the bed and a vertical sequence of different styles of lamination (the **Bouma sequence**).

6.7.2 Graded beds

A bed that shows a progressive upwards reduction in grain size from bottom to top is said to be **graded** or **normally graded** (Fig. 6.70). The grain size change can take one of two forms: **content grading** where the mean grain size of the sediment reduces upwards; and **coarse-tail grading** where the size of the coarsest grains diminishes, the rest of the population remaining roughly constant. It is not always possible to judge this difference in the field, especially in finer-graded sandstones, but one should attempt to discriminate where possible. In addition, it is instructive to record the range of size variation through a bed. Grading is also seen in finer-grained sediments, especially siltstones (see §5.2.5).

Graded beds are, in many cases, structureless or unlaminated. Where lamination is present it typically occurs in the upper part of the bed.

6.7.3 Inverse grading

In some beds, grain size may also show an upwards increase (**inverse grading**). Although this is rare in finer-grained sandstones, it is more common in coarser sandstones and conglomerates. Inverse grading rarely occurs as a thin layer at the base of an otherwise massive or normally graded bed. In some thick sand beds, a succession of thin inversely graded layers may give a crude horizontal stratification (cf. Fig. 7.24). Such patterns are quite a common feature of pyroclastic sediments. Inverse grading is also dealt with in Chapter 7.

6.7.4 The Bouma sequence

In addition to showing structureless and, in some cases, graded intervals, sharp-based sandstone beds may also include units of parallel lamination or ripple cross lamination. Where these are all present, they tend to occur in a particular vertical order (Figs. 6.71, 6.72), which has been called the **Bouma sequence** after its discoverer. In its complete development, structureless sand, which may or may not be graded (A division), is overlain by parallel lamination, which may show primary current lineation (B division). This in turn is overlain by ripple cross lamination (C division). The D division, the least often recognized, is also parallel laminated, but the sediment

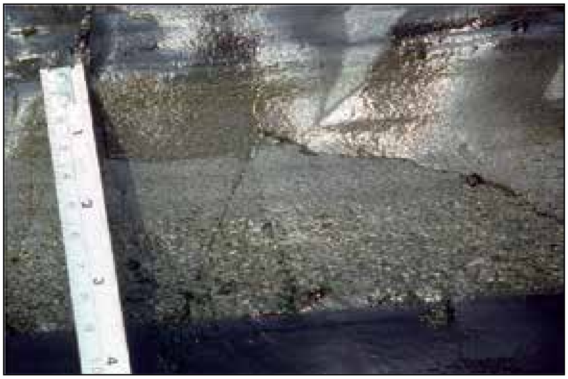
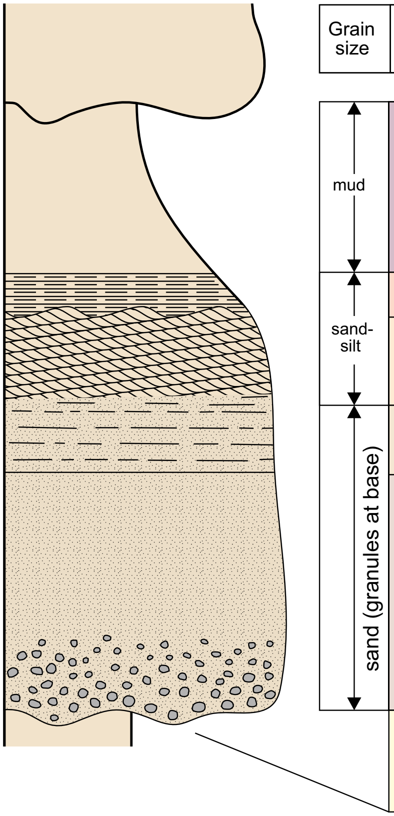


Figure 6.70 Normally graded sandstone bed. Cambrian, Whitesands Bay, Pembrokeshire, Wales.

Figure 6.71 Cyclical repetition of sand-dominated beds. Each sand bed is interpreted as a turbidite unit – the product of a single turbidity flow event. Aberystwyth Grits, Newquay, Wales. Photo courtesy of Gilbert Kelling.



Grain size	Bouma (1962) divisions	Hydrodynamic interpretation
mud	**E** usually mudstone	settling from suspension in quiet water or from slow currents with dilute suspensions
sand-silt	**D** upper parallel lamination	from suspension with some reworking on the bed
sand-silt	**C** ripple cross lamination, commonly climbing and with convolute lamination	ripple migration with net deposition; lower flow regime
sand (granules at base)	**B** lower parallel lamination; primary current lineation common	upper flow regime; plane bed
sand (granules at base)	**A** massive, sometimes normally graded; where present, granules are often confined to basal part; may contain mudstone rip-up clasts from bed beneath	rapid deposition from suspension with little or no bed transport; possible quick bed
	erosive base showing evidence of scour, sole structures such as flutes, grooves and chevrons common	passage of current capable of entraining sediment from underlying bed into flow

Figure 6.72 The Bouma Sequence of internal sedimentary structures which occur in sandstone beds generated by decelerating unidirectional currents. In most examples of such beds, one or more of the divisions may be missing.

is typically silty and the lamination is rather diffuse. The final interval, the E division, is fine-grained mudstone or siltstone, which may be difficult to separate from the fine-grained interbeds of the succession.

This complete sequence is very much an ideal development and, in reality, it is commonly the case that one or more division is missing. Only the vertical order remains constant. The relative thickness of the intervals also varies. Some beds are dominated by the laminated divisions, whereas others consist almost entirely of the A division, and may only have a thin capping of division B or C. Very thick beds, dominated by the A division may show signs of rapid dewatering in the form of water-escape structures (§9.2.2). Beds that end with division C commonly have preserved ripple morphology on their top surfaces, draped by the fine-grained interbedded sediment. Beds with a thick C division commonly show climbing ripple cross lamination (ripple drift), and it is also quite common for convolute lamination to occur within the C division (§9.2.2; Fig 9.23). These sharp-based beds commonly have sole marks on their lower surfaces in which case, it can be valuable to compare palaeocurrent directions derived from such marks with those derived from primary current lineation and ripple cross lamination in the B and C divisions.

6.7.5 Processes of formation

Sharp-based sandstones, in interbedded sandstone/mudstone successions suggest a pattern of episodic deposition, the sands recording high-energy events and the mudstones recording longer lasting intervals of deposition from suspension in quiet conditions. As discussed in §6.6, massive or unlaminated sand is attributable to rapid deposition from a heavily sediment-laden suspension. Associated with grading, this suggests a decelerating current, with coarsest particles falling to the bed first. Graded beds can be produced very simply in the laboratory by stirring up a suspension of mixed grain sizes in a beaker and allowing it to settle.

The formation of inverse grading is still a matter of some debate. For coarser, pebbly sandstones, it seems most likely the grain-size segregation results from inter-granular collision in a basal traction carpet layer where conditions like those of a grainflow prevail, driven by the shear of a powerful over-riding current (see §3.7.2; Figs. 3.24, 3.25). The development of a dispersive pressure within the layer forces larger clasts upwards and this process may be augmented by kinematic sieving whereby smaller particles preferentially fall through the mass of colliding grains. Such a layer may freeze if shear stress falls or if more grains are added from suspension, thus preserving the inverse grading. This process may repeat itself to give multiple inversely graded layers. An alternative process, which may be more appropriate for finer grained sandstones, is that grain-sizes become segregated horizontally in the active flow as coarser grains are transported more slowly. During deposition, the finer-grained sand at the head of the flow is deposited first with coarser material following.

The upward passage from structureless graded beds into laminated sand records a reduction in depositional rate, whereas the style of lamination records the flow strength when transport and sorting on the bed began. Parallel lamination (B division) records upper flow regime plane bed, whereas the C division records ripples migrating under a weaker current in the lower flow regime. The D and E divisions are mainly the result of direct deposition from suspension. The whole assemblage suggests a decelerating flow with material being deposited from suspension throughout and with phases of bedload transport (Fig. 6.73).

The Bouma sequence was first described from sediments of inferred deep-water origin where the currents were

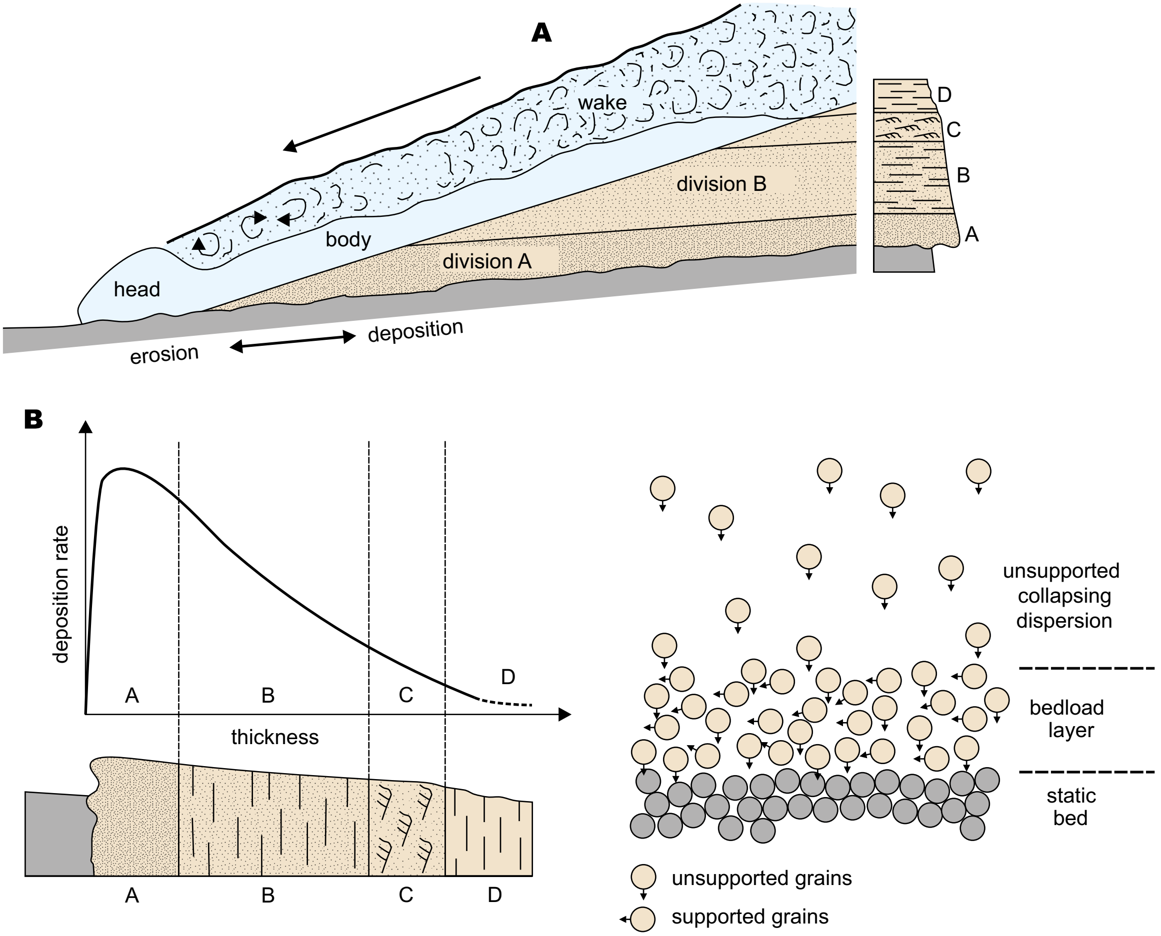


Figure 6.73 Features of a turbidity current and its deposits. A) Streamwise profile of a flow showing its major zones and the likely sites of deposition of the Bouma intervals (A, B, C, D). B) Depositional rate and its relationship to the Bouma sequence. C) Possible layer structure of the flow during deposition. Modified after Allen (1991) and Stow et al. (1996).

interpreted as turbidity currents (see §3.7.2). However, the sequence is in no way exclusive to such currents and similar beds can occur as a result of episodic decelerating flows in many settings. In most natural settings, the nature of the decelerating current can only be deduced from the context of the sediments. However, direct observations of large-scale turbidity flows have been made in recent years using sensitive monitoring equipment deployed in submarine canyons, for example. Much progress has been made in understanding the rheology and behaviour of turbidity currents and their resultant deposits through experimental study in laboratory-based tanks (see §3.7.2).

6.7.6 Uses of grading and the Bouma sequence

As well as containing valuable information about depositional processes, normally graded beds are valuable way-up indicators. The relative rarity of inverse grading in sandstones makes the use of grading quite reliable. However, it is best always to check the direction of grading in several beds before coming to a firm decision. The Bouma sequence, where present, is also a useful check on way-up as well as indicating the occurrence of a decelerating current. Both sole marks on the bases of beds and structures in the laminated intervals of the Bouma sequence provide valuable palaeocurrent data, as do ripple forms that are commonly preserved on the top of such sandstones.

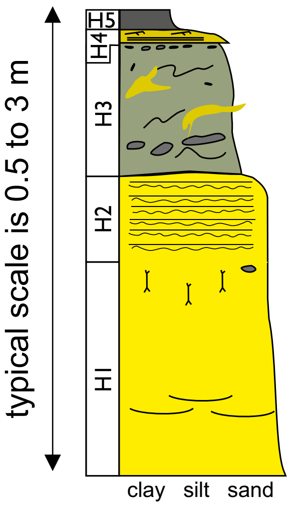
6.7.7 Hybrid-Event Beds (HEBs)

Flows that are transitional between turbidity currents and mud flows are increasingly recognized in subaqueous environments. The deposits of these transitional flows are referred to as hybrid-event beds (HEB), linked debrites and slurry beds. Mud flows, hyperconcentrated flows and concentrated flows may evolve down slope and over time into turbidity currents.

Hybrid event beds (HEB) in the ancient rock record are characterized by features indicative of both turbulence-supported (cohesionless) and mud-supported (cohesive) flow where no bed boundary is present between the two. Many HEBs are characterized in their lower part by a grain-supported texture that passes up into a bed characterized by a mud-supported texture, commonly with large mud rip-up clasts (Fig. 6.74). Debris flows and mud flows commonly evolve down slope into turbidity currents, and vice versa.

6.7.8 Sediment waves, associated scours and cyclic steps

Trains of long-wave, upstream-migrating bedforms are common in high-gradient subaqueous settings, notably on the slopes of continental margins. Changes in slope promote internal hydraulic jumps in turbidity currents, leading to the development of sediment waves and associated scours. Sediment waves are symmetrical or asymmetrical



Description	Interpretation
H5 - Mudstone.	Suspension fall-out.
H4 - Parallel and current ripple laminated sandstone.	Traction beneath dilute turbulent wake.
H3 - Muddy sand with or without mud-clasts, sheared sand patches, outsize granules. Carbonaceous material may segregate to top.	Quasi-cohesive debris flow locally modified by sand injection from beneath. Quasi-laminar plug flow.
H2 - Colour banded sandstone with loading at the base of lighter bands. Sheared dewatering pipes in light bands.	Transitional flow with cyclical turbulence suppression or cyclical turbulence suppression and scouring.
H1 - Low abundance of mud-clasts supported by clean sandstone in upper division. Graded to ungraded, unstratified, dewatered, matrix-poor sandstone.	Aggradation from non-cohesive high density turbulent flow with high sediment concentration and suspension fall-out.

Figure 6.74 Idealised model describing the vertical succession of deposits associated with emplacement of a hybrid event bed. Modified after Haughton et al. (2009). Image courtesy of Sarah Southern.

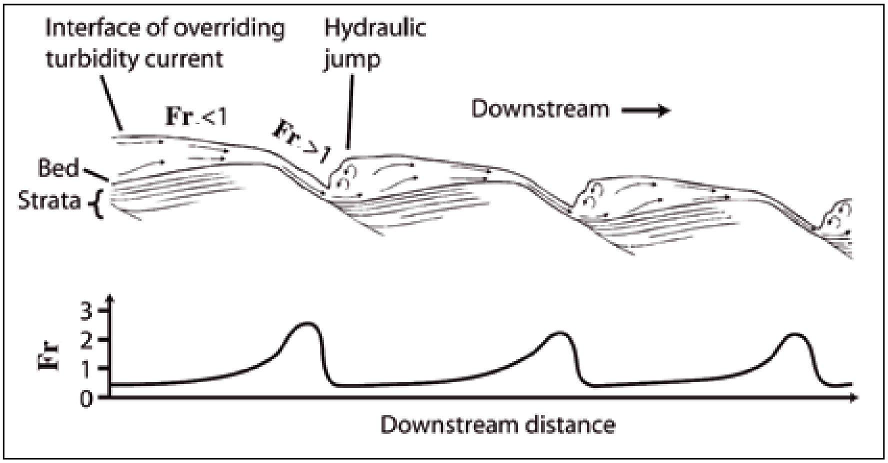


Figure 6.75 Schematic illustration depicting the development of a series of asymmetrical cyclic steps and their relationship to downstream variation in Froude number (Fr). Modified after Cartigny et al. (2011) and Covault et al. (2014).

bedforms typically with wavelengths of 10^2–10^3 m and heights of 10^0–10^2 m. As these bedforms develop, they migrate upstream. Sediment waves are commonly of turbidity-current origin. They are generally fine-grained sand, net-depositional bedforms, which may be preserved in the rock record as upstream-dipping backset beds accumulated as stoss-slope deposits, which are truncated by erosive surfaces on their lee (Fig. 6.75).

Associated scours take the form of erosional depressions, some with crescentic planform morphologies. These structures form linear trains of commonly asymmetrical waveforms. They have poor long-term preservation potential but are prominent features on many present-day sea floors.

Collectively, these long-wavelength bedforms are believed to develop in response to cyclic steps where high gradients and slope breaks act to promote internal hydraulic jumps within turbidity currents. Cyclic steps develop where internal hydraulic jumps (i.e., transition from Froude supercritical to subcritical flow) occur at the boundary between lee and stoss slopes such that supercritical to subcritical flow transformations arise. Cyclic steps are related to other supercritical bedform types, notably antidunes (see §6.4.3).

6.8 Aeolian bedforms and lamination

6.8.1 Introduction

Aeolian (i.e. wind-derived) deposits are associated with a range of sedimentary structures, including bedforms and plane beds. Three distinct scales of aeolian bedform are recognized: (i) ripples, (ii) dunes and (iii) megadunes (also known as draa) (Fig. 6.76). These three scales of structures represent a hierarchy within which similar features coexist at different sizes and spacings, suggesting the presence of equilibrium bedforms. Both dunes and megadunes invariably have ripples migrating across many parts of their slopes. Movement and growth rates of aeolian bedforms are related to the volumes of sand involved, as well as to wind intensity and duration, so that megadunes may take several thousands of years to develop and equilibrate, whereas ripples may respond almost instantaneously to changes in wind direction and strength.

Although small aeolian dunes of coastal belts or inland sand "seas" (**ergs**) are comparable in size with aqueous dunes or sandwaves, larger aeolian bedforms range up to significantly larger dimensions. Aeolian dunes also occur superimposed on and migrating across larger bedform structures (megadunes or draa), which are themselves migratory. Megadunes (draa) have no aqueous counterparts. Small aeolian ripples and horizontal beds are superimposed on both dunes and megadunes. The suggestion that ripples, dunes and megadunes form a hierarchy of equilibrium bedforms provides a basis for classification and description (Fig. 6.76). However, because of the low density and viscosity of air there is a high chance of aeolian processes frequently passing from equilibrium to gross disequilibrium, both in terms of the energy required to form structures

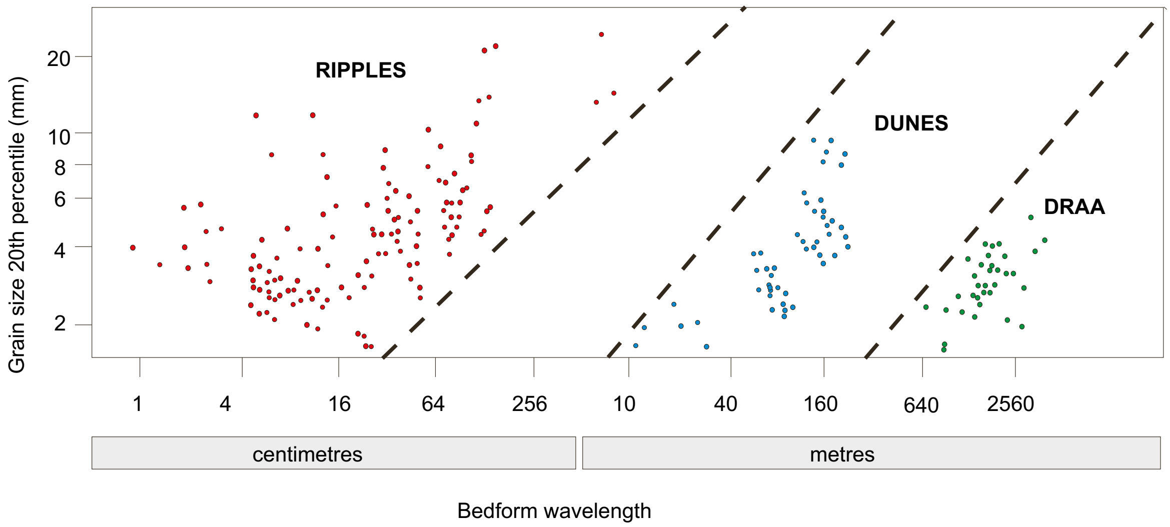


Figure 6.76 Grain size (coarsest twenty-percentile) versus wavelength for aeolian bedforms. Note the three distinct groups representing ripples, dunes and megadunes (draa). Modified after Wilson (1971).

and the direction of flow. Few thick aeolian deposits are forming today, and most large bedforms are currently not in equilibrium with the local wind regime. In the rock record, preservation of aeolian bedforms as relief features is now well documented. However, in the majority of aeolian successions, no original bedform relief is preserved and the former existence of dunes is deduced largely from internal structures preserved as sets of cross bedding. Be sceptical of books that give simple criteria for the interpretation of dune types from ancient strata; records of aeolian processes, structures and environments are amongst the most difficult to identify and explain in detail.

6.8.2 Material

Aeolian deposits occur mostly in sand, rarely extending into granule-sized gravels. The sand commonly exhibits distinctive grain size, shape and sorting characteristics because the wind is highly selective in terms of the grain sizes that it can carry for a given velocity. Saltation (see §3.6.2) is the dominant transport mechanism associated with aeolian dunes and inter-particle collisions result in high rates of grain abrasion such that dune sediments are commonly composed almost exclusively of highly resistant quartz, chert or lithic grains (especially metaquartzite). However, coastal dunes of carbonate sand are known and dunes of gypsum occur adjacent to inland evaporite lakes. Sands composed of resistant grains commonly develop a 'millet seed' texture characterized by highly rounded, high sphericity grains with surfaces that are dull (frosted) as a result of repeated grain collisions (abrasion). Although aeolian dunes composed of friable, cleavable sand grains of feldspar, mica or silt-clay aggregates may develop in areas close to the sediment source, they are virtually absent in settings more distant from the source. For example, dunes of dry, sand-size clay–silt aggregates are commonly seen, especially as parabolic dunes, forming immediately downwind of dried-out lake beds. On wetting, such dunes become solid masses but retain their cross bedded structure.

6.8.2 Wind ripples

Three morphologically distinct types of small-scale ripple occur on present-day wind-blown surfaces. Each are also recognized in vertical section where they produce different types of aeolian ripple stratification.

Impact ripples

Impact ripples are the most widespread ripple type developed in aeolian environments. They have low relief and they form from the coarser-grained fraction of the sand upon which they develop. They usually have wavelengths of 5–20cm and heights of 5–10mm. They exhibit a high ripple-form index (Fig. 6.2) and have straight or sinuous, continuous crestlines oriented transverse to the wind direction and upon which the coarsest grains are concentrated. These ripples are slightly asymmetrical in profile with their lee faces inclined at low gradients, below the angle of rest (Fig. 6.77).

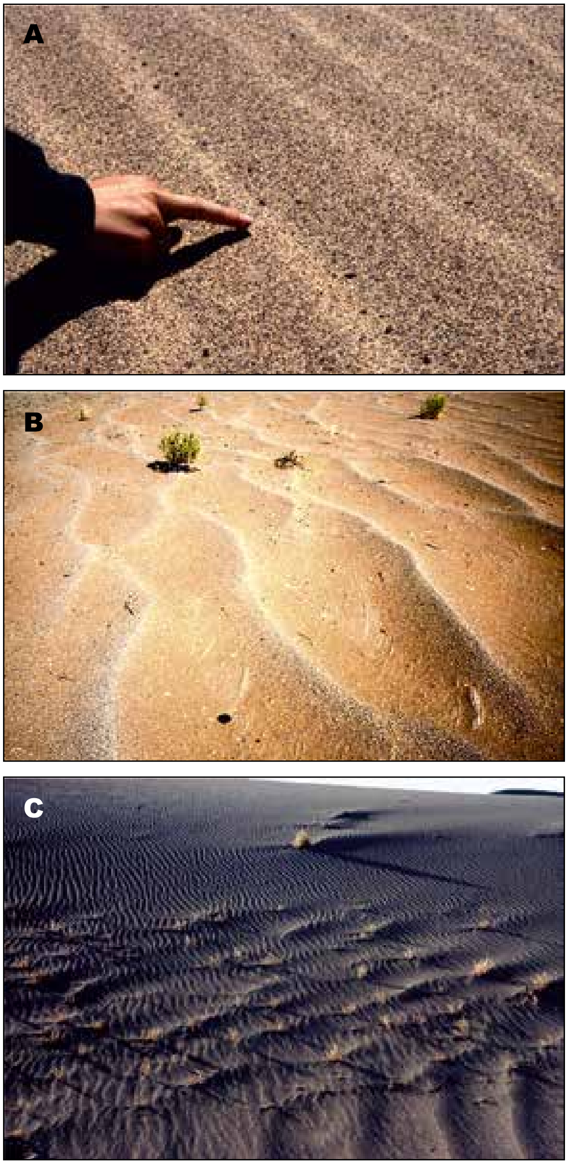


Figure 6.77 Examples of aeolian impact ripple morphology. A) Straight crested wind ripples in well sorted medium sand. Note occasional granules. Namib Sand Sea, Namibia. B) Sinuous-crested wind ripples with a distinct bimodal grain-where relatively few size distribution. Note how coarser grains are confined to ripple crests. Namib Sand Sea, Namibia. C) Aeolian ripples on the stoss slope of a dune bedform. Idaho, USA.

Impact ripples develop through a combination of saltation and reptation (see §3.6.2), where saltating sand grains act as high momentum impacting particles under the influence of wind shear and cause grains at rest on the bed to repate (hop) downwind. For a given wind velocity, reptating grains in motion are restricted to a narrow size range and the distance that they jump downwind (their **path length**) is similar for most of the sediment in transport. For this reason, impact ripples begin to form with spacings that are determined by path length (Fig. 6.78) and thus ripple spacing is proportional to wind velocity. Minor surface perturbations act as the catalyst required to initiate ripple development and, once initiated, the ripples themselves grow and steepen into bedforms because upwind facing stoss slopes act as an **impact zone** that catches incoming saltating grains, whereas downwind facing lee slopes act as a **shadow zone** where grain impacts are minimal (Fig. 6.78). The ballistic impact of grains landing in the impact zone causes other grains to creep up the stoss slope to the ripple crest until an impact causes them to launch into the airflow and reptate (hop) downwind to the next ripple. Coarser grains that are too large to reptate concentrate at ripple crests, whereas finer grains are preferentially trapped in ripple-trough shadow zones where the effects of wind shear are at a minimum. Air temperature, which influences

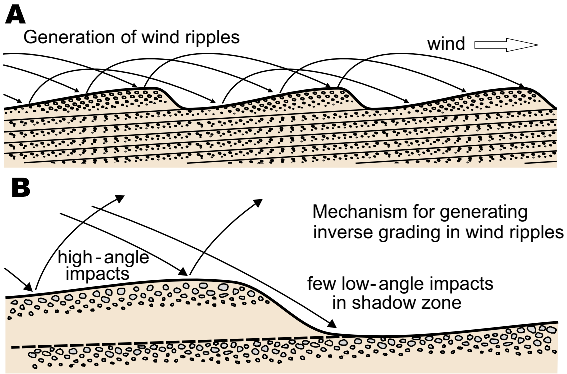


Figure 6.78 Wind ripples generated by ballistic impact of grains. The ripple spacing relates in a general way to the saltation path length, which is the characteristic distance that individual grains hop as a result of collision on the bed. The saltation path length varies with grain size, shape and density and mean wind velocity and gustiness close to the bed. A) The migration of wind ripples results in sub-parallel lamination. B) The impact angle of saltating sand grains differs between stoss and lee slopes. High-angle impacts on the stoss sides promotes creep of coarser grains towards the ripple crest. Lee slopes form a shadow zone where relatively few low-angle impacts occur, thus encouraging the accumulation of finer grains in ripple troughs. As ripples migrate downwind this sorting mechanism generates inversely graded laminae.

the viscosity of the airflow, will exert a limited influence on the grain size of ripples such that coarser sand ripples are more common in cold settings. Aeolian impact ripples may be differentiated from subaqueous-ripples because the former typically have high ripple-form index (ratio of wavelength : height) of 25-40+, and are commonly characterized by inverse grading that results from the migration of coarser-grained ripple crests over finer-grained ripple troughs. More complex ripple patterns may develop from the merging of smaller and larger ripples with different movement rates.

Aeolian mega-ripples

Where the supply of sediment for aeolian transport is restricted to coarse sand, gravel and small pebbles, and where the wind blows with sufficient intensity to move these grain sizes, aeolian mega-ripples (granule ripples) may develop. Although these bedforms also develop due to the impact of saltating and reptating grains, as described for impact ripples, they typically have more sinuous crestlines and wavelengths and heights up to 5m and 35cm, respectively (Fig. 6.79). A continuum of ripple sizes exists from impact ripples developed in fine sand to granule and pebble mega-ripples. However, aeolian ripples formed of pebble-grade material are rare and most are composed of clasts of low-density lithologies, such as pumice.

Adhesion ripples

When dry sand is blown across a wet sediment surface, some grains stick to the surface on impact. This process of grain adhesion results in the generation of a range of structures including adhesion ripples and adhesion warts (Fig. 6.80), which are characterized by low relief ridges and mounds that grow by adhesion to their upwind edge and thereby undergo upwind growth. Capillary rise of

Figure 6.79 Examples of aeolian granule and pebble mega-ripples. A) Skeleton Coast, northern Namibia. B) Trench dug through an aeolian granule mega-ripple to reveal the internal stratification. Note the foresets that dip down in the direction of the lee slope (to the right as viewed). C) Train of aeolian pebble mega-ripples with sinuous crests. D) Pebble grade aeolian mega-ripples showing grain-size differentiation between the troughs (finer) and the crests (coarser). Spacing between crests is approximately 3m. The exceptionally corase-grained fractions in examples C and D are composed of low-density pumice clasts. B), C) and D) are from the Askja region for Central Iceland.

Figure 6.80 Examples of wind adhesion 'ripples'. A) Adhesion structures generated on a surface of damp sand over which dry sand has been blown. Indian Creek, Utah, USA. B) Adhesion ripple structures preserved on a bedding surface. Permian Cedar Mesa Sandstone, Utah, USA. C) Upper bedding surface of sandstone showing an irregular small-scale morphology interpreted as wind adhesion 'ripples'. Independence Fjord Group, Proterozoic, north Greenland.

moisture helps to trap further grains by maintaining a damp surface. Although adhesion structures are found preserved both on bedding plane surfaces and in section, where strata form crinkly and wavy laminae, they typically have a low preservation potential since, on drying out, they tend to collapse and become reworked by the wind. Because adhesion strata require the accumulation surface to be damp, such structures are usually restricted to low-lying interdune and dune flank settings close to the water table.

Aeolian ripple stratification

The tractional processes that generate impact wind ripples give rise to various types of wind-ripple stratification. **Rippleform laminae** occur where grain size differentiation enables the internal foreset laminae of ripple sets to be distinguished (Fig. 6.81). However, the uniformity of grain size that typifies many aeolian sands means that internal laminae cannot necessarily be distinguished and **translatent rippleform stratification** results from ripple migration. Wind-ripple strata may exhibit a weak inverse grading, in part because the finest material tends to accumulate in sheltered ripple troughs whereas coarser grains concentrate on the ripple crests, and in part because finer grains tend to settle between coarser grains resulting in a **pour-in** texture. This means that the base of the ripple stratum is often characterized by a distinct surface defined by a thin lag of finer material. Where ripples preserve traces only one or two grains thick, a characteristic **pinstripe lamination** is preserved (Fig 6.82). As with subaqueous current ripples (see §6.1.4), the accumulation of migrating wind ripples results from **bedform climb** with respect to the accumulation surface. Aeolian ripple strata form widespread deposits in sandsheets, dry interdunes and on low to moderately inclined dune and megadune (draa) slopes (see §6.8.3).

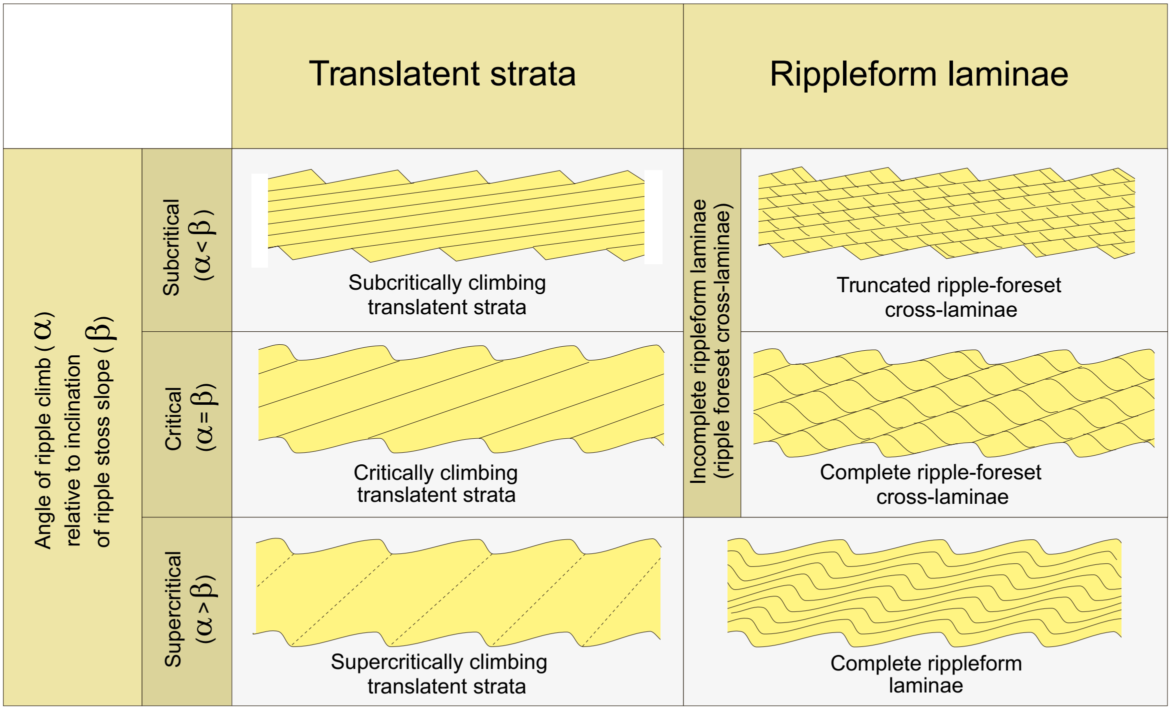


Figure 6.81 Classification of wind-ripple stratification according to the angle of ripple climb relative to the inclination of the stoss slope of the bedform and the presence or absence of cross lamination. Modified after Hunter (1977).

6.8.3 Size, shape and classification of large-scale aeolian bedforms

The classification of large-scale aeolian bedforms is based around data collected using a variety of techniques including ground observations, low-level aerial photography, remote sensing from space satellites, studies of internal structure (notably by examining natural exposures in blowouts, or via trenching or by using ground penetrating radar), the measurement of wind regimes and conceptual numerical models. As a consequence, dune-scale and larger aeolian bedforms are classified according to a number of separate criteria including their scale, morphology (shape, width, wavelength, height), spacing (frequency), orientation relative to net sand transport direction, style of migratory behaviour and form of any superimposed bedforms.

Aeolian dunes

Aeolian dunes have wavelengths of 5-250m and are commonly arranged into **trains** of regularly spaced bedforms. In terms of morphology, simple dunes are charactersied by a single windward **stoss slope** inclined at 8-16° and a **lee slope** inclined at 20-34°, whereas more complex forms may possess several stoss and/or lee slopes facing in various orientations. Lee slopes commonly comprise a **slip face** (i.e. a foreset slope), inclined at or close to the angle of rest and down which sand episodically avalanches. However, not all dunes possess such slip faces and some may be **slipfaceless** (i.e. generated at or degraded to lower angles). Dunes exhibit a wide variety of morphological forms that reflect the combined effects of a number of controlling factors including wind strength and directional variability on diurnal to seasonal (and longer) time-scales, sediment supply and the availability of sediment for transport. Additionally, dunes may be classified as mobile (actively migrating), active but anchored (for example, attached to a large boulder or area of vegetation), or stabilized.

Mobile dunes are classified according to their morphology based on the number of lee faces that they possess and according to the orientation of their crestlines relative to the predominant wind direction (Fig. 6.83). Common dune types classified according to these criteria include **transverse** dunes that possess a single, gently-inclined,

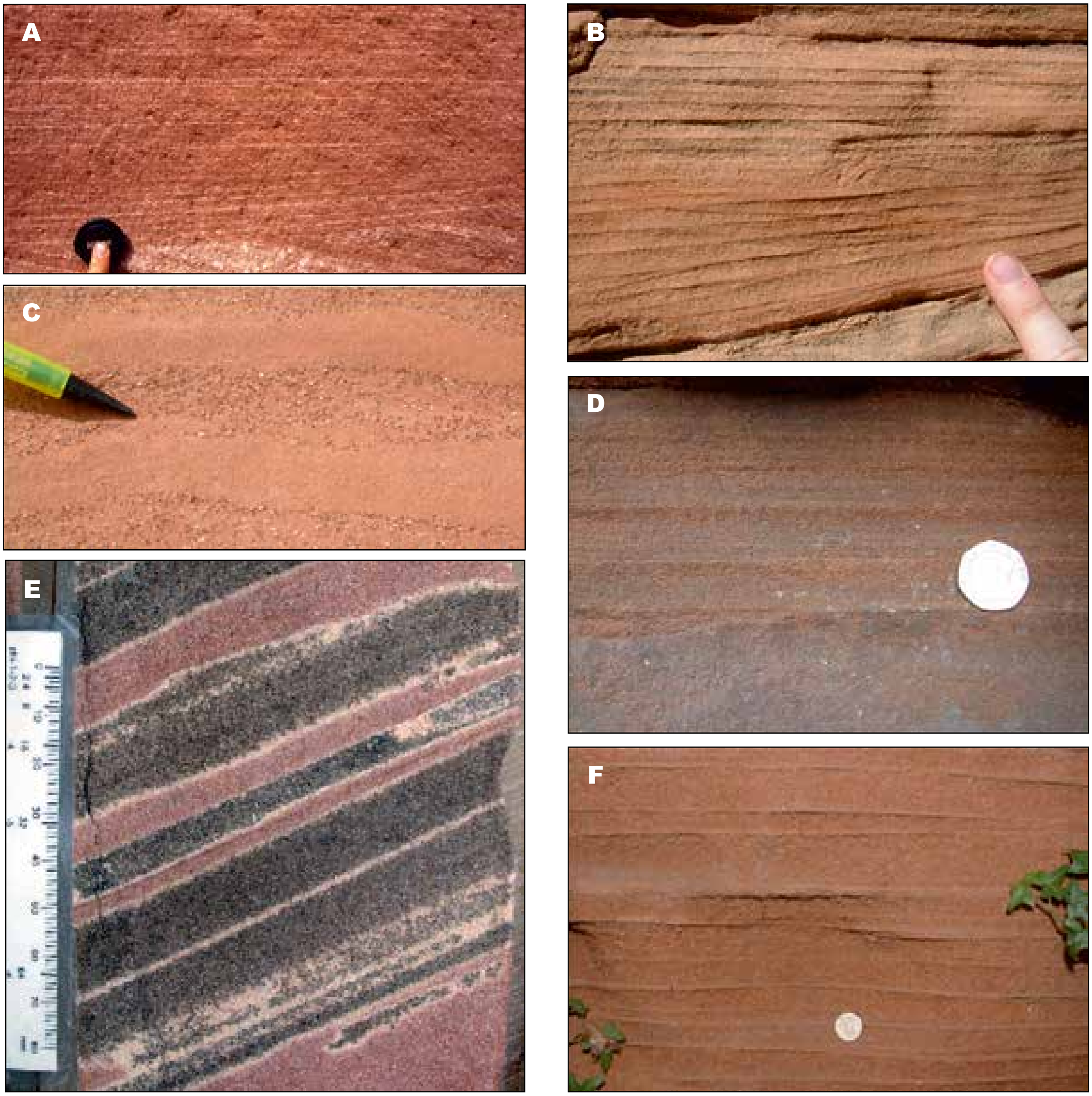


Figure 6.82 Common types of lamination in aeolian deposits. A) Pinstripe lamination represents the accumulated deposits of aeolian ripples Each stripe reflects the passage and accumulation of one ripple in a train, where by slightly finer grain fractions represent the passage of the ripple troughs and slightly coarser grain fractions represent the passage of the ripple crests. Etjo Sandstone Formation, Cretaceous, Namibia. B) Aeolian wind-ripple strata where the foresets of ripples are weakly preserved as truncated ripple foreset laminae (see Fig. 6.81). Jurassic Navajo Sandstone, Utah, USA. C) Bimodal grain sorting in aeolian deposits. Bridgnorth Formation, Permian, Shropshire, England. D) Weakly developed inverse grading in packages of grainflow strata. Bridgnorth Formation, Permian, Shropshire, England. E) Well-developed packages of aeolian grainflow strata observed in a cored section. The black colour is dues to residual oil in this hydrocarbon reservoir sandstone. The red colour is the original hematatite coating of the sand grains. The cream colour is due to leaching of iron from the surface of the grains due to contact with hydrocarbon fluids. The inclination of the strata reflects the slope of the original aeolian dune on which these grainflows developed. Auk Formation, Permian, Central North Sea, UK. F) Well-developed packages of aeolian grainflow strata observed in an outcrop section. The finer-grained fractions that delineate each package are thin accumulations of grainfall deposits between each grainflow deposit. Bridgnorth Formation, Permian, Shropshire, England.

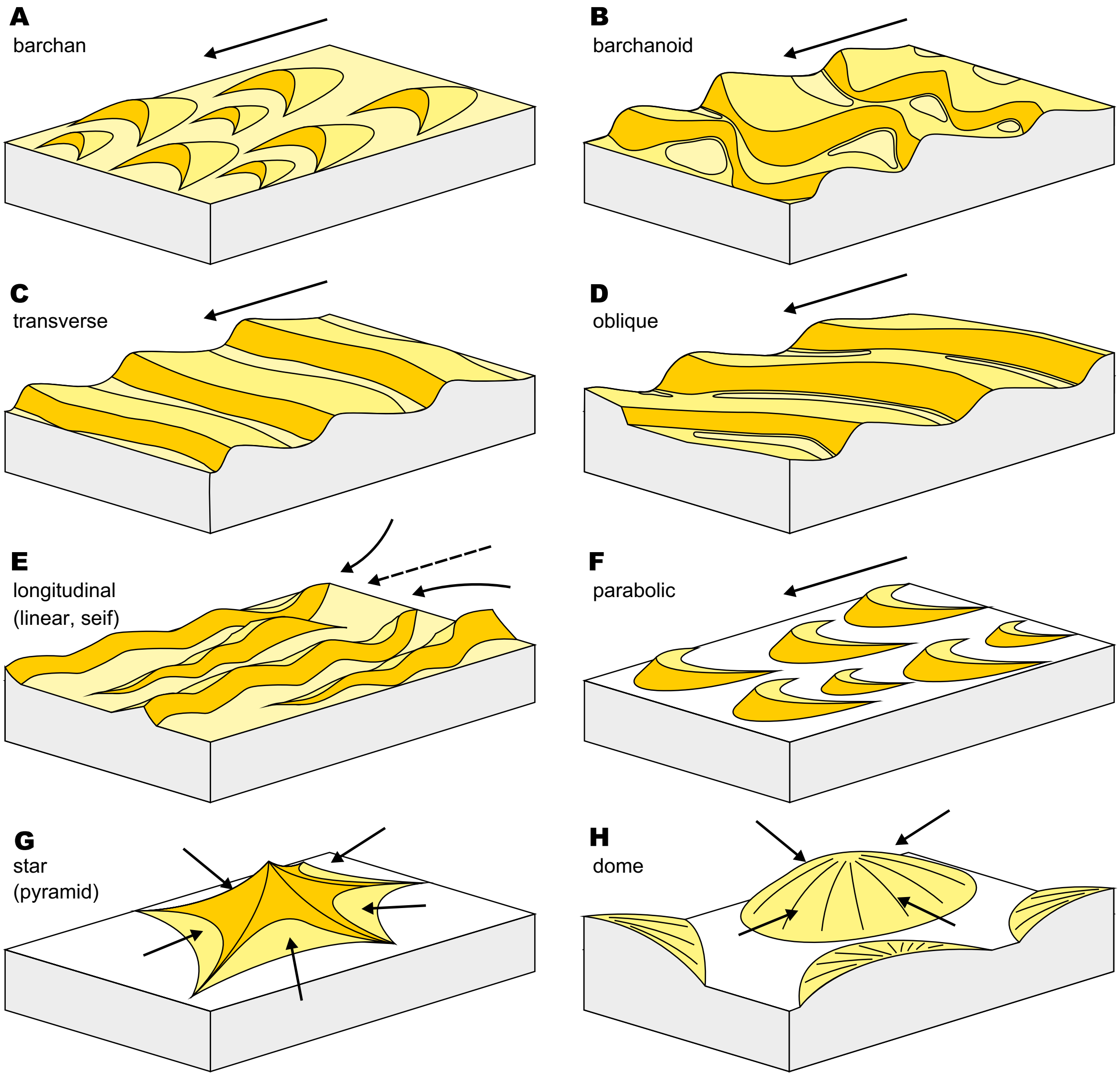


Figure 6.83 Three dimensional forms of some common dune types. The arrows mark the dominant directions of the effective winds and in case E, the dotted arrow indicates the resultant effective direction.

upwind-facing stoss slope, a steeper downwind lee slope and a crestline perpendicular to the prevailing wind. Longitudinally-oriented **linear** dunes (sand ridges) have 1 or 2 lee faces and a crestline parallel to the wind. **Star** (**pyramid**) dunes that have 3+ lee faces (Figs. 6.83, 6.84). Although the recognition of dunes as purely transverse or longitudinal is useful for simple classification schemes, it is potentially misleading because net sand transport direction across many dunes is oblique, resulting in the generation of **oblique** bedforms (Fig. 6.85).

At a more detailed level, dunes that possess straight crestlines are two-dimensional, whereas those with sinuous, cuspate or lobate crestlines are **three-dimensional** (Figs. 6.83, 6.86). Examples of three-dimensiona, transverse bedforms include spatially isolated, crescent-shaped **barchan** dunes that open downwind and commonly have corridors of sand-free ground all around. **Barchanoid** dune ridges are sinuous transverse ridges, where, in plan view, the crestline sinuosity of successive bedforms may be either in-phase

Figure 6.84 Examples of common aeolian dune morphologies. A) Lee face of a crescent-shaped barchan dune, Huab Basin, Namibia. B) Oblique aerial view of sinuous-crested transverse bedforms, western Namib Sand Sea, Namibia. Width of view in foreground is approximately 600m. C) Straight-crested linear dune, Huab Basin, Namibia. Bedform is 25m wide. D) Aerial photograph of stabilized linear dunes with wide interdune areas. What is the likely pattern of effective winds? Dunes are around 100m wide. Strzelecki Desert, South Australia. E) Aerial view of star draa, Central Namib desert. Width of view is approximately 1km. F) Star draa, Central Namib desert. Bedform is approximately 280m high.

Figure 6.84 *Continued*

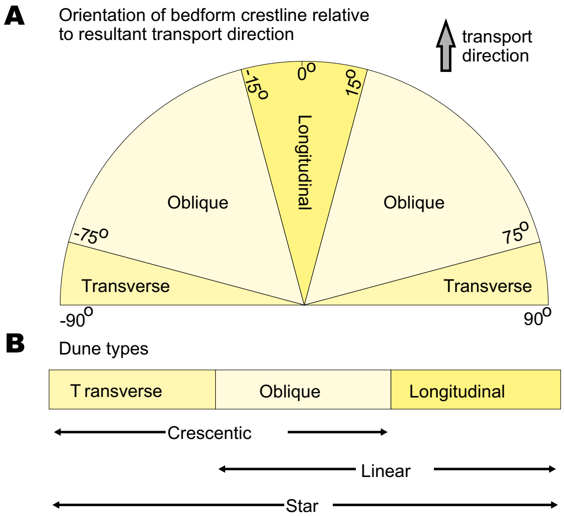


Figure 6.85 Classification of dunes. A) Morphodynamic dune types based on orientation of crestline relative to resultant transport direction. B) Probable range of morphological and morphodynamic dune types. Modified after Hunter et al. (1983).

or out-of-phase (Fig. 6.86). **Parabolic dunes** form U-shapes closing downwind and are common in areas where vegetation acts to anchor or stabilize the outer dune limbs but is insufficient to halt the migration of the central part of the dune. **Seif** dunes are a type of three-dimensional linear bedform in which the crestline sinuosities migrate downwind along the bedform crest in a motion similar to that of a snake (Fig. 6.86D). The spacing between seif dunes is typically approximately twice their mean width. The crests of such dunes exhibit a regular sinuosity with slip faces on alternative flanks. Upwind ends of ridges are rounded, and along their length Y-shaped forks show 30–50° angles to the flow and open up wind. At their downstream ends the ridges are pointed. Most seifs are parallel to the resultant vector of the effective winds. **Zibar** are low ridges of coarse-grained, hard-packed sand without slip faces that are usually aligned transverse to the wind and often occur in the corridors between seifs or as independent bedforms in sand sheets.

The style of migratory behaviour of mobile dunes can also be used for further classification. Dunes that migrate in a constant direction and at constant speed are **invariable**, whereas dunes that undergo temporal changes in migration direction, speed, asymmetry and/or steepness are **variable**. A temporal change in the style of migratory dune behaviour is one potential explanation for the origin of geometrically complex bed sets in the ancient record.

Dunes are commonly arranged into a network (**aklé**) in which there are transverse, longitudinal and oblique components. For example, sinuous, transverse ridges display alternating linguoid and barchanoid (i.e. concave downwind) sectors which are either in- or out-of-phase in relation to those in an adjacent ridge (Fig. 6.86A). Elsewhere there may be straight-crested dunes (or megadunes) transverse to the wind and, close to them, **dome-shaped** dunes with many minor slip faces and rounded flanks inclined at low angles. Given the morphological complexity of many aeolian bedforms it is useful to plot the orientation of lee slopes as dip-azimuth data on a stereogram (Fig. 6.87). This may enable subtle trends regarding bedform morphology and arrangement to be discerned.

Megadunes

Megadunes (draa) are larger scale bedforms than dunes with wavelengths of 500–5000m and heights in excess

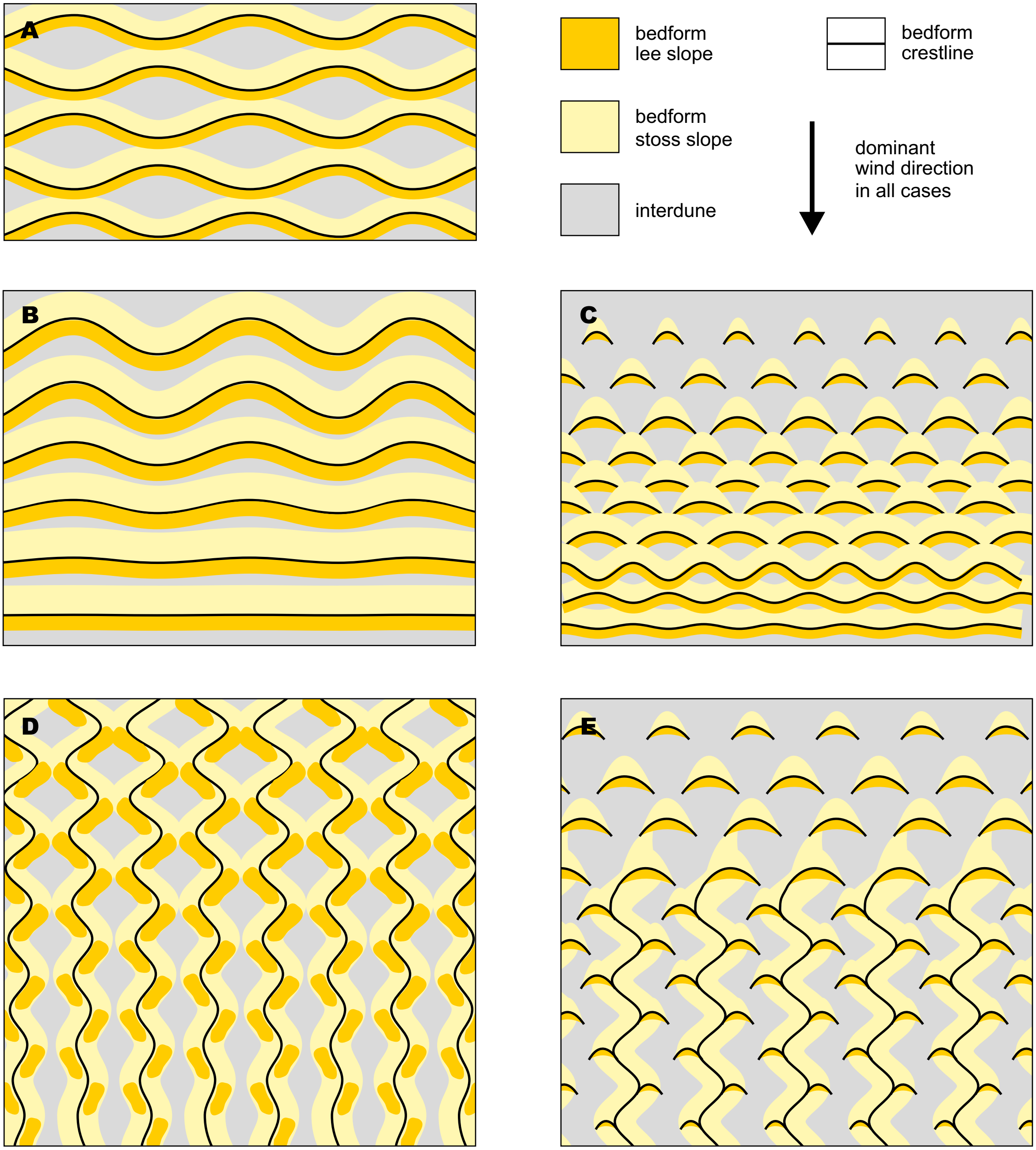


Figure 6.86 Schematic illustration of common bedform configurations in plan view. A) Sinuous-crested transverse bedforms with crestlines of adjoining bedforms 180° out-of-phase. Note how the interdune flats form spatially isolated depressions. B) Sinuous-crested transverse bedforms with crestlines of adjoining bedforms perfectly in-phase. Downwind decrease in amplitude of crestline sinuosity to zero. C) Downwind spatial transition from isolated barchan dunes, through a zone of laterally interconnected barchanoid dune ridges, to low-sinuosity transverse dunes. This pattern is a common configuration at upwind erg margins. D) Longitudinal (linear) dunes that undergo a downwind decrease in crestline sinuosity. Note the resultant increase in the degree of interconnectivity of the interdune flats. E) Downwind spatial transition from isolated barchan dunes to connected barchans that are transitional into sinuous-crested linear dune ridges with transverse spurs.

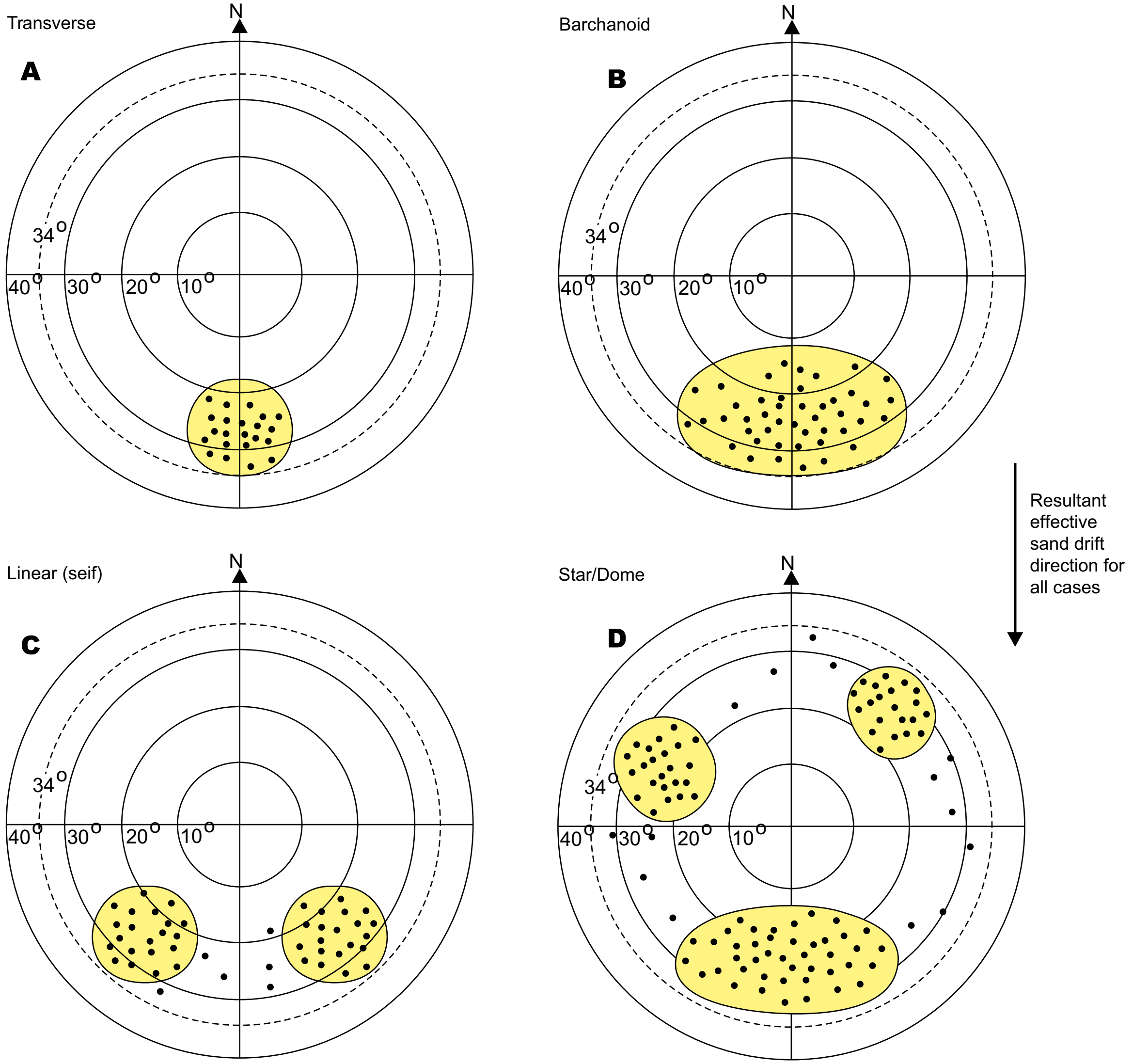


Figure 6.87 Schematic diagram illustrating palaeocurrent patterns predicted for common aeolian dune types, plotted as dip-azimuths of lee slope foreset laminae on lower hemisphere stereographic projections. The angle of repose in well sorted, loose dry aeolian sand is typically 34° and bedform lee slopes tend not to exceed this angle. Foreset laminae preserved in ancient successions tend not to exceed 26° due to the effects of compaction. Foreset dip azimuths from linear, star and dome-shaped dunes preserved in ancient successions tend not to exhibit pronounced bi- or multi-modal distributions like their modern counterparts, in part because over prolonged time periods these bedform types undergo net migration in a particular direction and, in doing so, preferentially preserve foreset laminae oriented in that direction.

of 50m, and in a few cases more than 300m. These large bedforms only occur in the largest sand seas where aeolian sediment supply and transport rates are high. Megadunes are described using the same terminology as for dunes but additionally may be characterized by the presence of superimposed dune-scale bedforms on their flanks. **Simple** megdunes lack superimposed dunes, whereas **compound** megdunes possess superimposed dunes of the same morphological type and **complex** megdunes possess superimposed dunes of a different type. Some compound and complex megdunes may be slipfaceless in that the megadunes themselves do not possess an active slip face, even if the dunes superimposed upon them do. Common examples of compound megdunes forms include longitudinal

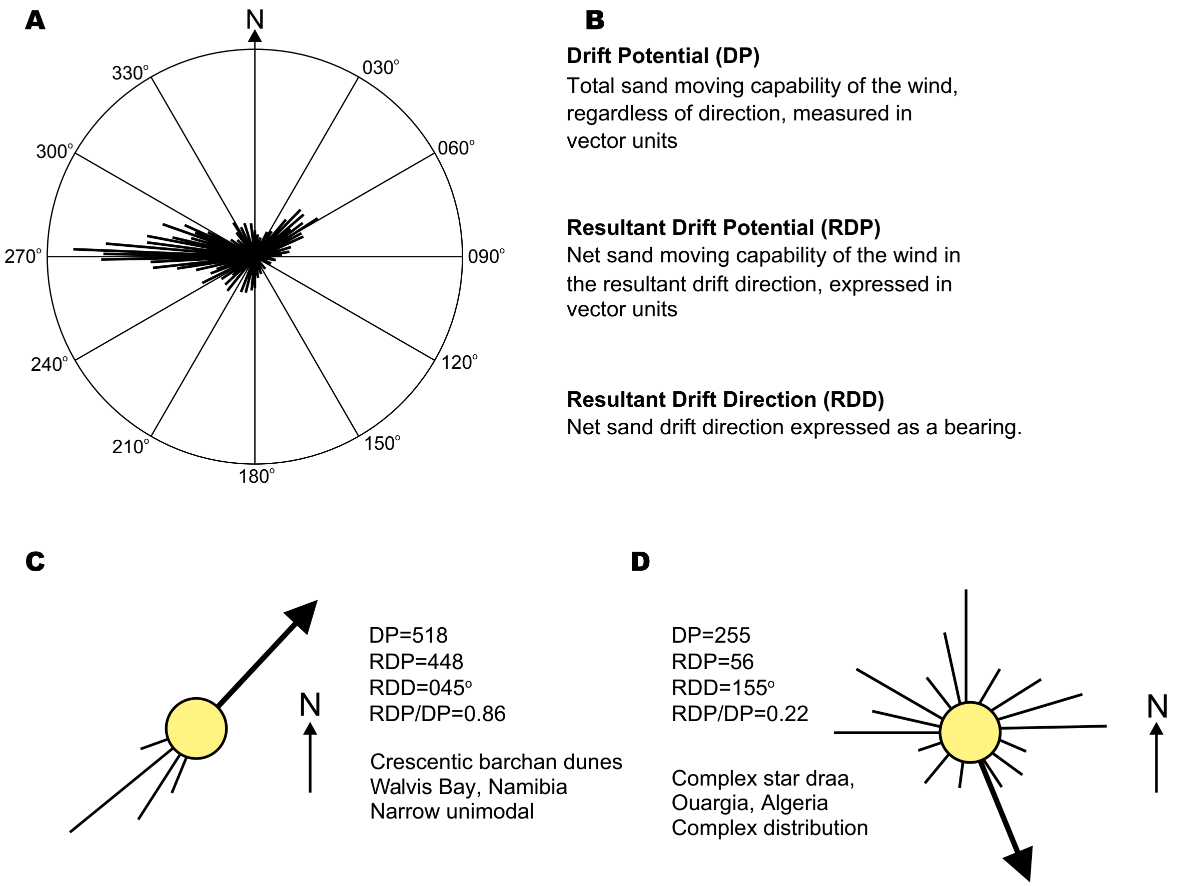


Figure 6.88 Techniques for the measurement and recording of wind variability and strength and corresponding sand mobility. A) A typical wind rose. Wind observations taken at 15 minute time intervals over a period of time. The lengths of the arms are proportional to the time the wind blew from a given direction. Winds from the west and northeast were dominant. B) Terms used in the quantification of sand mobility. C) and D) Example sand roses. The orientation of the finer lines (arms) records the direction of potential sand drift towards the centre. The length of the arms records the amount of potential sand drift (in vector units). The summation of the length of all the arms is the DP. The length of the heavier arrowed line indicates RDP. The orientation of this line indicates the RDD. RDP/DP is a measure of wind variability. Values greater than 0.75 indicate unimodal winds and favour the development of transverse or barchan dunes; values less than 0.2 indicate variable winds and favour the development of star or dome dunes. Modified in part after Fryberger (1979).

megdunes that have smaller seif dunes aligned across their slipfaceless flanks and barchan megdunes that support smaller, superimposed barchan dunes. Common examples of complex megdunes forms include star megdunes that have radiating arms that support superimposed three-dimensional transverse forms and linear megdunes that have superimposed transverse ridges that migrate along their flanks (Fig. 6.86E). The migration of superimposed dunes over larger, more slowly moving megdunes is a potential explanation for the origin of geometrically complex bed sets in the ancient record.

Four terms are commonly used to classify the energy and directional properties of the wind and to relate it to the construction of particular dune types (Fig. 6.88). Spread-out **Equivalent Sand Thickness** (EST) is a measure of the size of an aeolian bedform in terms of the thickness that the sand from which it is composed would reach if it were spread evenly over its basal area. **Drift Potential** (DP) is a measure of the total sand-moving capability of the wind without regard to wind direction. **Resultant Drift Potential** (RDP) is a measure of the resultant or net sand-moving capability of the wind in the **Resultant Drift Direction**

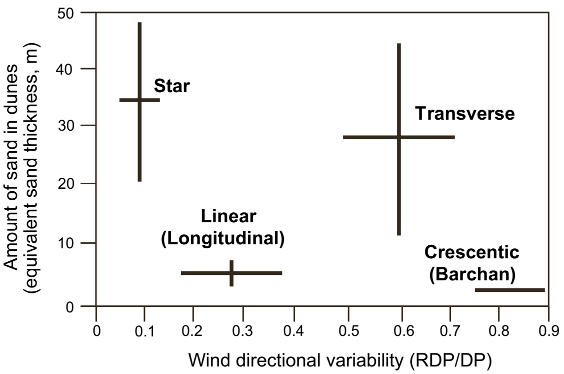


Figure 6.89 Application of the concept of Equivalent Sand Thickness, Drift Potential (DP) and Resultant Drift Potential (RDP). The graph depicts the relationship between dune type, wind regime and equivalent sand thickness. Transverse and barchan dunes develop under unimodal wind regimes (RDP/DP>0.5) and are sand transporting bedforms. Star dunes develop under multi-directional wind regimes (RDP/DP<0.2) and are sand storing bedforms. Modified after Wasson and Hyde (1983).

(RDD) and can be determined by plotting sand drift rose diagrams (Fig. 6.88C, D). RDP/DP (sometimes referred to as the Unidirectionality Index) provides an indicator of wind variability where values approaching unity (RDP/DP > 0.8) signify low variability (i.e. unidirectional winds) and low values (RDP/DP < 0.3) signify high variability. RDP/DP values plotted against EST yield a discriminant plot for different dune types (Fig. 6.89), which suggests that dune type is controlled by **wind regime** as well as by the availability of sand. Transverse dune forms tend to develop under conditions of unidirectional winds characterized by high RDP/DP values and are sand transporting bedforms (Fig. 6.89). By contrast, star dune forms tend to develop in response to variable winds (low RDP/DP values) and are sand accumulating bedforms, which migrate only very slowly, if at all. However, it is important to realise that factors such as the degree of vegetation cover may complicate the situation further.

In describing aeolian dunes and draa, measure bedform width, wavelength, height and geometry of crests (Fig. 6.83). Additionally, measure the angle of inclination of the top, middle and bottom of the stoss and lee slopes, evaluate the mean grain size and distribution of grain sizes, the coherence and porosity of the sand, and the nature, distribution and direction of superimposed smaller structures. Take lacquer peels of structures from both vertical and horizontal excavated surfaces (Appendix 2). Smoke pots may help to discern the local airflow over active dunes. Look for the presence or absence of a lee-side eddy, a cross wind and a reattachment point of the flow. The use of stakes may help to monitor the rate of advance of parts of a dune. Using simple sand traps it is be possible to measure sand flow into and off the dune and to plot a sand rose diagram (Fig. 6.88). Monitoring of wind direction and speed by a hand-held anemometer may be helpful locally. Regional climatological data should be plotted so that the weighting of the effective wind direction is properly represented. In describing networks of aeolian dunes and draa, note lateral variations in morphology between successive bedforms in a train in orientations both parallel and perpendicular to the dominant in direction. In particular, look for downwind changes in dune and interdune size, shape and spacing and note the distance over which any changes in form occur (Fig. 6.86).

6.8.4 Small-scale internal structures of modern and ancient aeolian sands

Small-scale aeolian stratification results from a distinct suite of processes that enables ancient strata of aeolian origin to be recognized. The cross-stratified sets that make up the interior of modern aeolian dune- and draa-scale bedforms and of ancient aeolian strata dominantly comprise four basic stratification types that occur in various configurations. These are **grainflow strata, grainfall strata** (Fig. 6.82), **wind-ripple strata** (see §6.8.2) and aeolian **plane bed** strata.

Aeolian grainflow stratification

Individual grainflow units occur on dune and draa lee slopes where an active slipface is developed (Fig. 6.90A). Grainflow avalanches generate foresets that are inclined at 18-34^{o} and are 2–5cm thick, although they often thin up-slope. They show a tongue- or cone-shaped structure which displays en-echelon, scalloped or tabular shapes in horizontal section in large dunes, and lenses in small dunes. The strata have sharp, concave-upwards erosional bases and are internally structureless (Fig. 6.90B), although they may show inverse grading normal to the base, as well as lateral grading along their length. Notably, at the foreset toe, coarse grains may occur at the top of the layer. Heavy minerals may line the upper parts of foresets. In modern dune sands, grain packing is usually very loose and porosity is very high (up to 45%). **Slump degradation** grainflows, where internal structure is destroyed as the flow travels downslope, are characterized by a chaotic wedge of loosely-packed sediment up to a few metres wide that typically thickens downslope before eventually pinching out. **Scarp recession**

Figure 6.90 Characteristic deposits of aeolian sand dunes. A) Grainflow tongues generated by gravitational collapse and avalanching of part of the lee face of a sand dune. Sossusvlei, Namib Sand Sea, Namibia. B) Grainflow strata seen in section. Individual grainflow tongues thin and pinch out at the base of the cross stratified set where they interfinger with lower-angle inclined wind-ripple strata. Grainflows are separated by thin grainfall deposits of finer grained sandstone. C) Oblique view of a cross stratified aeolian dune set with grainflow and grainfall strata interfingering with wind-ripple strata at the set base. Both from Cedar Mesa Sandstone, Permian, White Canyon, Utah, USA.

grainflows, where an initial point of failure generates a scarp that then retreats back upslope towards the brinkline, form tongue-like bodies that rarely exceed 0.5m in width but may extend almost the full height of the lee slope (Fig. 6.90A). Where developed in very well sorted sand, the boundaries between successive avalanches may not be evident and only amalgamated grainflow units are recognized.

Aeolian grainfall stratification

Grainfall strata form foresets that dip at 20–28° with indistinct but parallel internal laminae that lack grading and tend to follow pre-existing topography, though they may in some cases exhibit a wedge-shaped geometry that is thickest just leeward of the dune brinkline and that thins downslope. Grainfall units in modern dune sands are typically moderately packed and porosity may reach 40%. Grainfall strata have non-erosional contacts and individual units commonly blanket the upper parts of dune lee slopes for distances of tens of metres along slope, enabling them to be differentiated from individual avalanche strata. On small bedforms, grainfall strata commonly extend uninterrupted down to the base of the lee slope and may blanket grainflow strata (Fig. 6.90B). On larger bedforms, grainfall wedges tend to be cut out by grainflow strata.

Aeolian wind ripple stratification

Wind ripple strata are described fully in §6.8.2 and their additional description here relates solely to their occurrence on the lee and stoss slopes of larger aeolian bedforms. Aeolian ripples develop on dune and megadune slopes inclined at 0–25° where they form stratified units 1–15cm thick with the contacts between laminae either being erosional to non-erosional and sharp or gradational. Inverse grading of laminae is more common than normal grading. Cross lamination or in-phase waviness may be seen. Porosity in modern wind-rippled sands averages 39%, although this varies according to grain-size distribution. The strata appear on the stoss side, crest and convex-downwind projections of sinuous dunes and draa, the "horns" of barchans, the "noses" of parabolic dunes and on sand drifts in the lee of obstacles. The strata are commonly found interbedded with grainflow and grainfall strata, in many cases in rhythmic sequences (Fig. 6.90C).

Aeolian plane bed stratification

Aeolian plane-bed laminae on the slopes of dunes and draa dip at 0–15°, are parallel, even, and form sets of laminae 1–10cm thick that may be picked out by slight variations of grain type and size. Sharp or gradational non-erosional contacts are present and porosity in modern sands is typically less than 30%. These laminae occur only rarely and most commonly on upwind-facing stoss, or gently sloping, downwind-facing lee surfaces.

6.8.5 Large-scale internal structures of modern and ancient aeolian sands

Although the external forms of modern aeolian dunes and megdunes are readily apparent and well documented from aerial photographs and satellite images, the organization of their internal structure is less well known. Conversely, in the ancient rock record, bed sets of presumed aeolian origin are characterized by a variety of styles of cross stratification but the morphology of the bedforms responsible for generating them cannot usually be determined directly because the original bedform topography is typically not preserved. Thus, it is difficult to relate the external morphology and migratory behaviour of aeolian dunes and megdunes to their deposits, which are preserved as a product of bedform climbing, in the ancient record. Attempts to reveal the internal structure of modern aeolian bedforms have involved a variety of techniques. Trenches have been dug through stabilized dunes with a bulldozer in orientations both parallel and perpendicular to prevailing wind direction so as to provide a pseudo-three-dimensional view of their internal structure. In other studies, the upper parts of dunes have been planed down to a horizontal surface so as to reveal internal structure. In nature, wind reversals associated with major storm events may result in blowouts that leave exposed cliff sections. Where river systems flow into dune fields, flash floods may undercut dune flanks causing them to collapse and leave an exposed section. Since the early 1990s, geophysical techniques such as ground penetrating radar have been employed widely to image the internal structure of modern dunes. Although this non-destructive method has the potential to provide a three-dimensional view, it is restricted because the subsurface depth to which the technique can image rarely exceeds 10–20m. In the ancient record there are now several well documented examples where large-scale preservation of aeolian bedform topography reveals both the original bedform morphology and its internal structure, notably where the topography of the original bedforms has been preserved in-situ by a catastrophic event, such as rapid flooding and inundation by water or lava flows. Finally, because large-scale aeolian bedforms undergo relatively slow rates of growth and migration, their temporal evolution and migratory behaviour is difficult to study directly. One solution to this has been the development of computer models that simulate temporal bedform behaviour and predict how

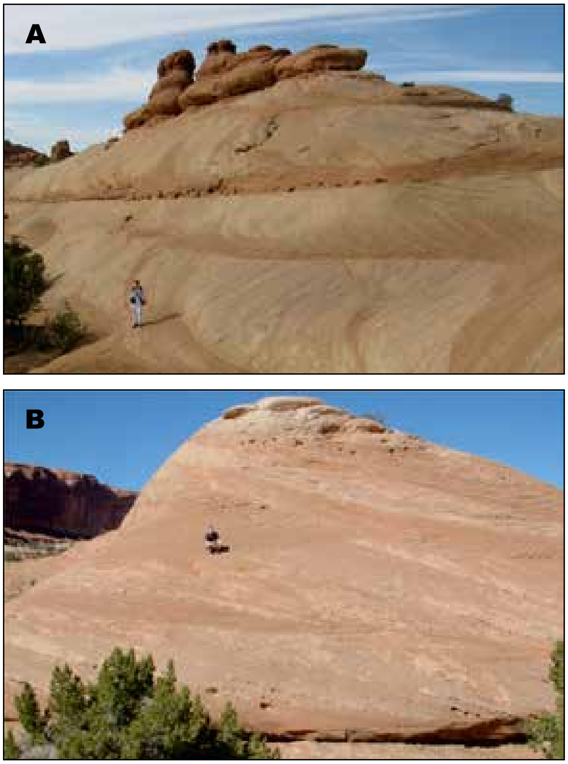


Figure 6.91 Bounding surfaces in aeolian sandstones that define large-scale sets of cross strata. A) Section aligned close to parallel to the original bedform migration direction. The laterally extensive, low-angle inclined bounding surfaces are interdune migration surfaces developed between dune sets. Minor reactivation bounding surfaces are also evident within sets. B) Section aligned close to perpendicular to the original bedform migration direction, revealing large-scale troughs indicative of original bedforms that had significant crestline sinuosities. Both examples from the Navajo Sandstone, Jurassic, Utah, USA.

different bedform types and styles of migration will generate and preserve different types of cross-bedded sets.

Cross bedding

Studies of modern cross-bedded aeolian sands suggest that a number of sedimentary features, when observed in association with one another, can be used as reliable indicators of large-scale aeolian bedform activity. Information from modern dunes is greatly augmented by studies of ancient aeolian sandstones (Fig. 6.91) wherein large-scale cross bedding is the most striking feature, with single large-scale sets commonly ranging in thickness up to 10m and occasionally attaining as much as 52m. Common sedimentary structures associated with aeolian dune activity include:

(a) A dominance of medium- and large-scale cross bedding composed of grainflow, grainfall and wind-ripple strata in varying arrangements (Fig. 6.90; see §6.8.4).
(b) The occurrence of cross bedding of several types, including tabular-planar, wedge-planar and trough types, the relative proportions of which vary between examples. Bundles of convex upward laminae may occur in several dune types.
(c) Stacked sets that are separated by erosive bounding surfaces (see below) which may be either near-horizontal or inclined at low to moderate angles in orientations that vary from sub-parallel to highly oblique relative to the foresets that they truncate.
(d) Trough cross beds that most commonly occur as solitary sets or thin cosets in the upper parts of dunes.
(e) Thin units of flat- and/or wind-ripple laminated bedding that commonly occur on the stoss sides and close to the crests of dunes.

The geometry of sets of cross strata and the orientation of foresets within aeolian bedforms varies with bedform type. Transverse dunes and megdunes are composed internally of planar and open trough cross-bedded sets with foresets inclined perpendicular to the trend of the bedform crestline (Fig. 6.92). Sinuous-crested barchanoid ridges are composed internally of trough cross-bedded sets that are typically rather complicated in terms of their internal architecture and with foresets inclined in a variety of directions (Fig. 6.93). Dome-shaped bedforms often exhibit sets composed of gently convex-up laminae (Fig. 6.94). The internal structure of linear and star-shaped bedforms is complicated, commonly characterized by a mosaic of overlapping and intersecting sets with foresets inclined in opposing directions (Fig. 6.95).

Bounding surfaces

Bounding surfaces are erosional surfaces that truncate sets of aeolian dune (or draa) cross strata. Four distinct types of bounding surface are recognized based on their shape, their orientation relative to the cross strata that they truncate, their lateral extent and their relation to one another (Fig. 6.96).

Reactivation surfaces occur within aeolian sets and are characterized by planar- or scalloped-shaped erosion planes that typically dip downwind at inclinations of 10-20°, somewhat less than the cross strata that they truncate. In sections perpendicular to aeolian transport, reactivation surfaces trend parallel to sub-parallel to the cross strata and can in some cases be traced for 10-100+m along strike, whereas in sections parallel to transport they may extend the full height of a set or may be restricted to its basal part,

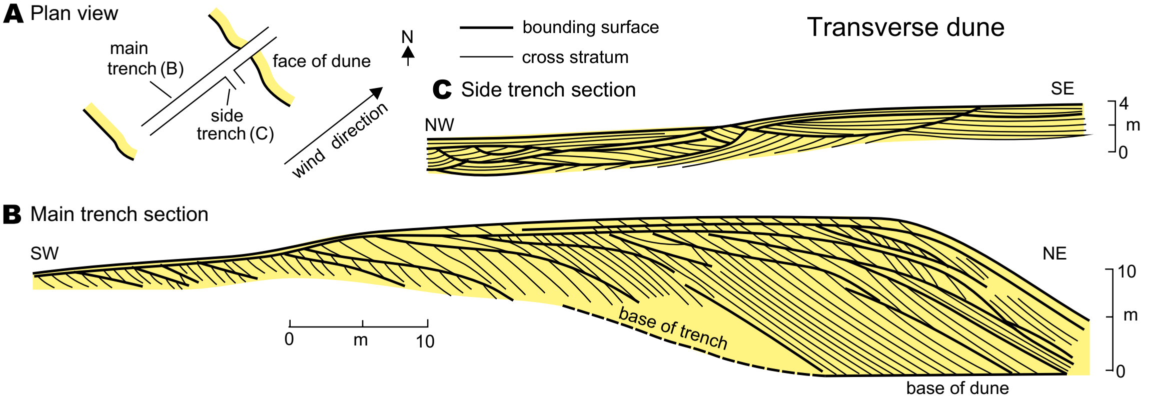


Figure 6.92 The structure of the interior of a transverse dune. A) Plan. B) Section SW-NE. C) Section NW-SE. Modified after McKee (1966) and McKee (1979).

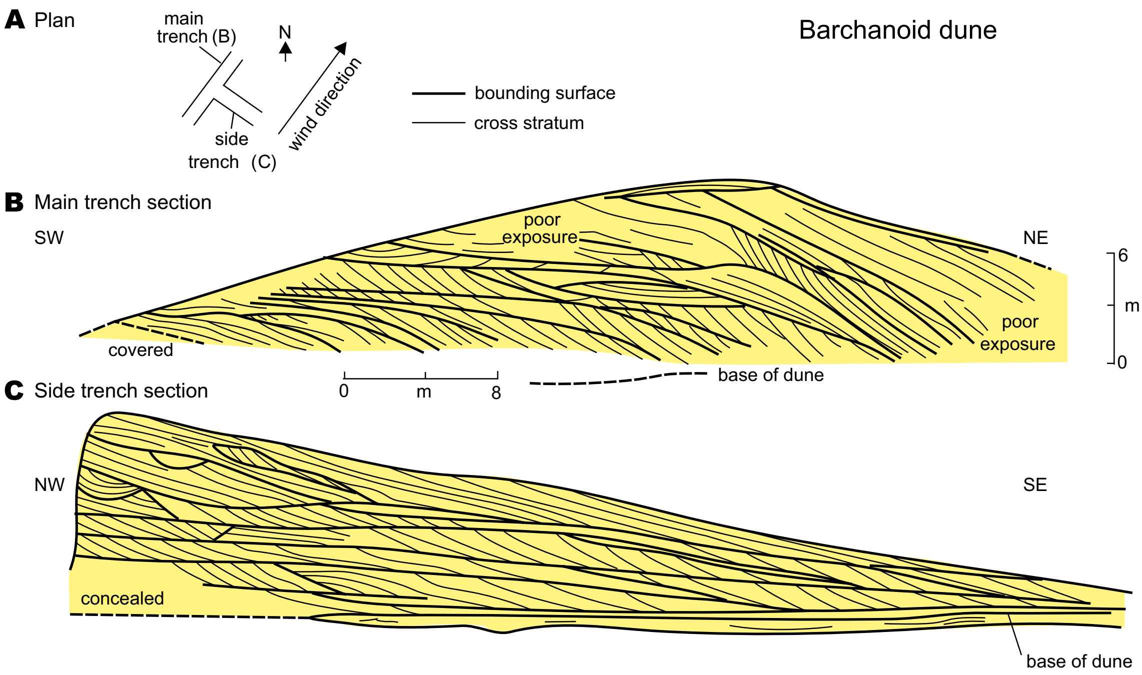


Figure 6.93 The structure of the interior of a barchanoid ridge dune. A) Plan. B) Wind-parallel section (SW-NE). C) Wind-perpendicular section (NW-SE). Modified after McKee (1966) and McKee (1979). This example shows a complexity of internal lamination that would not have been expected from the external morphology and suggests a complex evolution.

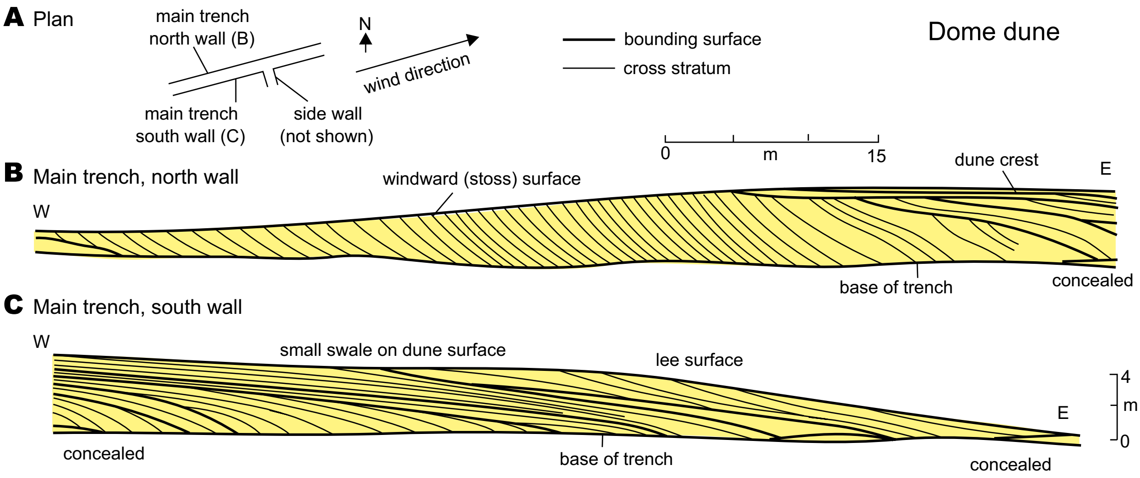


Figure 6.94 The structure of the interior of a dome shaped dune. A) Plan. B) and C) Wind parallel sections (W-E). Modified after McKee (1966) and McKee (1979).

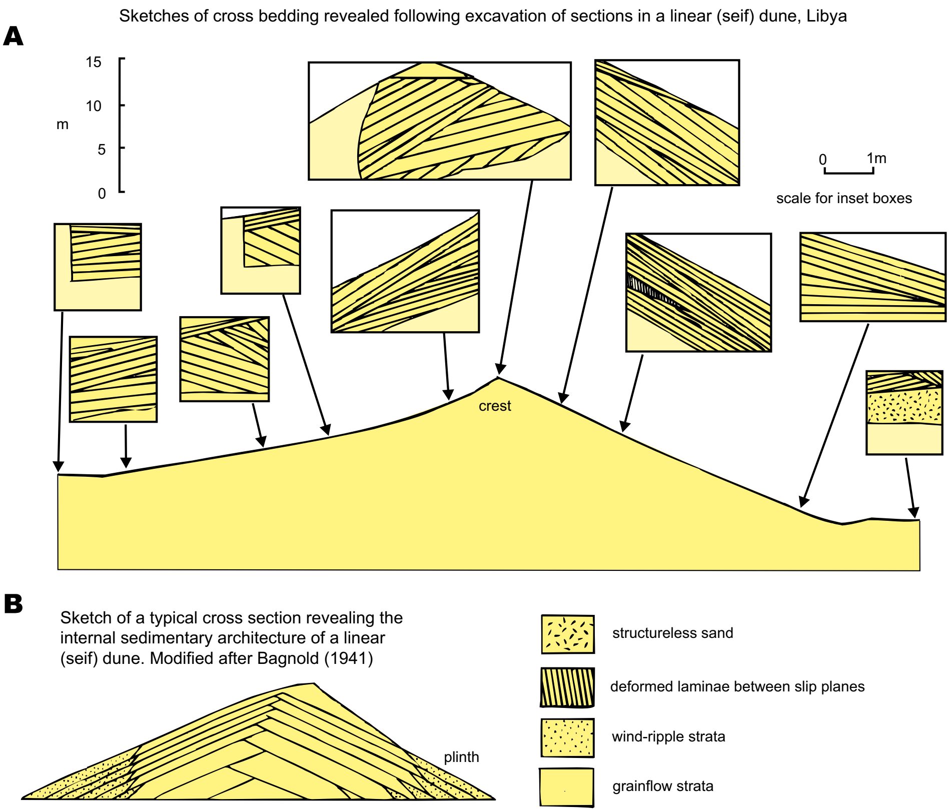


Figure 6.95 The structure of the interior of a linear (seif) dune. A) Bagnold's hypothetical model of a seif dune (1941) as modified by McKee (1979) to show foreset dips in opposed directions. B) Direct evidence of a real example, based on shallow excavations of a seif dune in Libya. After McKee & Tibbits (1964).

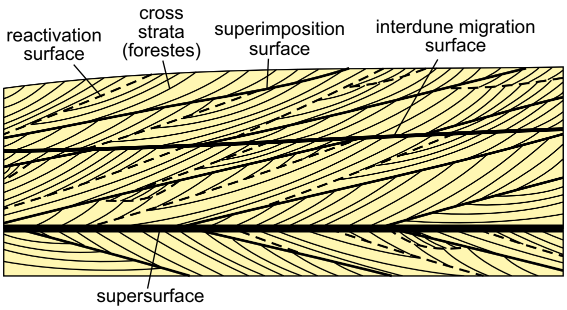


Figure 6.96 Definition diagram for the concept of bounding surfaces within compound cross-bedded sands and sandstones. Although this terminology was first proposed for aeolian sandstones, it can also be applied to water-lain sediments. Interdune migration surfaces arise as a consequence of dune migration. Superimposition surfaces represent the migration of superimposed bedforms and/or scour pits over a larger parent bedform. Reactivation surfaces represent partial deflation of a bedform lee slope and arise in response to periodic changes in bedform migration direction. Supersurfaces are usually regionally extensive and commonly bound entire erg accumulations. Modified after Brookfield (1977).

in which case they are commonly characterized by a sweeping (asymptotic) base. Although some reactivation surfaces occur apparently randomly within sets, others occur exhibit remarkably regular spacings indicative of a regularly repeating formative mechanism. Overlying cross strata exhibit either a concordant or downlapping relationship.

Superimposition surfaces occur within aeolian cosets and are characterized by planar to highly scallop-shaped erosion surfaces that dip in a wide range of orientations. In sections parallel to transport, these surfaces appear similar to reactivation surfaces and their identification can be problematic. However, in sections perpendicular to transport superimposition surfaces differ from reactivation surfaces because they are usually oriented oblique to the cross strata that they truncate. Where both reactivation and superimposition surfaces are developed, the latter always truncate the former.

Interdune migration surfaces are characterized by low-angle inclined erosive surfaces that typically extend downwind for distances of hundreds of metres to several kilometres. These surfaces, which bound sets or cosets, appear planar to slightly scalloped-shaped in sections parallel to aeolian transport, whereas in sections perpendicular to transport they may be moderately to highly scalloped. Interdune surfaces truncate both superimposition and reactivation surfaces.

Supersurfaces are erosive surfaces that truncate all other types of aeolian bounding surface and usually have great lateral extent and continuity such that they bound entire aeolian accumulations or significant parts thereof. Sedimentary features associated with supersurfaces include desiccation cracks and polygonal fractures, bioturbation, rhizoliths and halokinetic (salt) structures, all of which yield important paleoenvironmental information regarding the nature of the accumulation surface at the time of supersurface formation. Supersurfaces may be flat lying and planar or may exhibit considerable local relief, making their recognition problematic. It may be possible in some situations to correlate supersurfaces laterally into adjoining deposits of non-aeolian origin and to relate them to basin-wide events such as marine transgressions.

6.8.6 Processes of formation of dunes, megdunes and aeolian stratification

External form

The formation of dunes and megdunes requires the availability of abundant sand-size grains which are moved as bedload, whereas silt and clay are removed in suspension and gravel is left as a lag. Observations in modern deserts suggest that for a patch of sand to develop into a dune it has to be at least 4–6m long. The wind must be sufficiently retarded by the patch's saltation cloud for deposition to occur. Each patch grows to a critical height which depends on grain size and wind strength, whereupon a slip face may ultimately develop. Dunes and megdunes are regularly repeated bedforms which record the response of a sand bed to the shearing action of the wind. The bedforms record an attempt to reach a dynamic equilibrium in response to consistent fluctuation in the flow pattern.

Dunes form topographic obstacles that disrupt the **primary airflow** such that as the flow moves up the dune stoss slope it accelerates, thereby causing an increase in transport rate and promoting grain transport up the stoss slope to the dune crest. As the flow moves over the crest it passes into the lee-side depression, decelerates and causes a decrease in transport rate, thus promoting deposition on the lee slope. This provides the basis for a mechanism by which aeolian dunes advance downwind over time. **Flow separation** of the airflow from the bedform surface occurs beyond the crest, whereas **flow reattachment** occurs a distance of a few dune heights downwind. Thus, a **separation cell** exists in the dune lee within which turbulent **secondary airflow** occurs, which allows ripples and erosional scour hollows on the dune flanks (**plinth**) to develop and undergo complex migratory behaviour. Downwind of the reattachment point, renewed flow acceleration means that interdune sediments may potentially be eroded, thereby providing a local sediment supply for the next dune downwind in the train.

Historically, ideas on the controls of dune and megdunes shape have included the variable incidence of vegetation, the structure of organized turbulence in the lower atmosphere, the distribution of thermal convective plumes, the variability of effective winds and the amount of sand available for transport. Recent analysis suggests that the last two factors exert the greatest primary control (Fig. 6.89). Barchans, longitudinal and seif structures are associated with relatively low sand availability, whereas transverse barchanoid ridges and star forms develop where abundant sand is available. Barchans and transverse forms reflect low diversity of effective wind directions, whereas longitudinal and star dunes reflect high diversity.

Relating internal structures to external form

Relating the morphology and migratory behaviour of modern bedforms to the architecturally complex bed-set and bounding-surface geometries that they generate is important for understanding the environmental significance of ancient aeolian strata but is far from straight-forward. Although the external morphology of modern aeolian bedforms is readily apparent, their internal bed-set architecture is difficult to determine. Geometrical computer simulations demonstrate that bedforms of similar external morphologies can generate radically different patterns of cross bedding because they undertake different migratory behaviour through time. Furthermore, the amount of a bedform that is accumulated as a bed-set (i.e. not eroded) following the passage of subsequent bedforms in a train is typically only a small fraction (usually <10%) from the basal-most part of the entire bedform. As such, the reconstruction of bedform morphologies from bed-set architecture usually relies on the assumption that the preserved bottom sets adequately reflect the depositional processes that occurred on the upper (non-preserved) parts of the bedform lee slope.

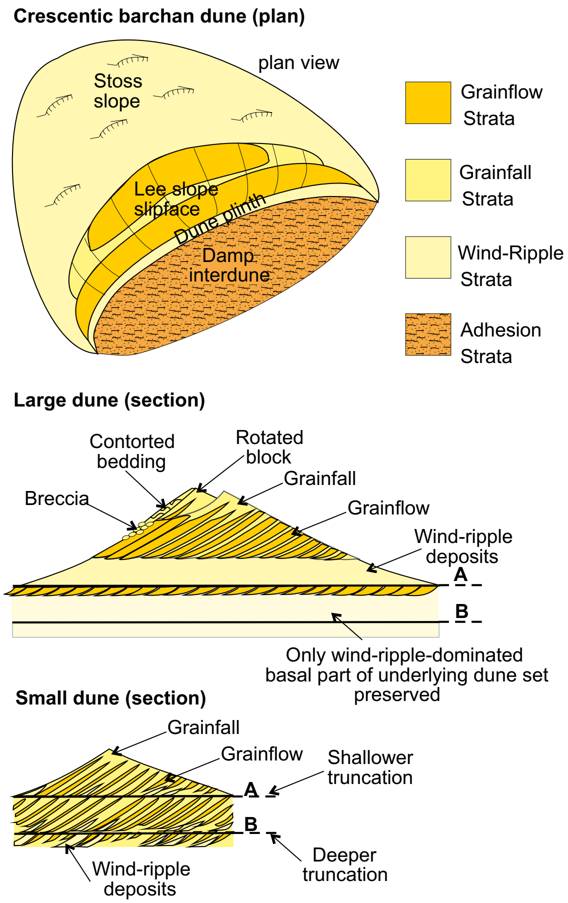


Figure 6.97 Examples of characteristic aeolian facies and their distribution on a simple crescentic (barchan) dune and on large and small scale aeolian dunes truncated to different levels (A, B). Level of truncation influences the preservation of facies types in the geological record, with features characteristic of the upper slipface lost. Modified after Hunter (1977) and Kocurek and Dott (1981).

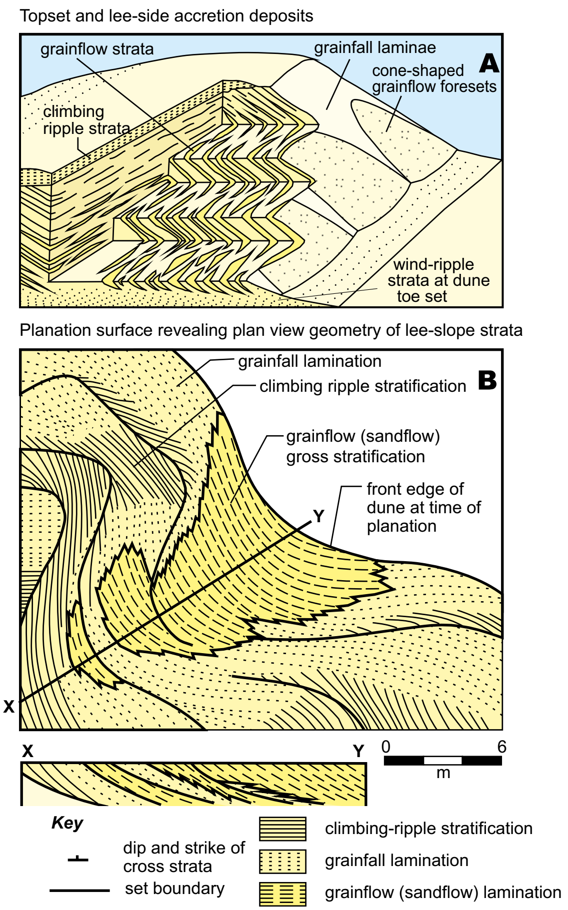


Figure 6.98 Typical distribution of small-scale lamination types in aeolian dune deposits. A) Schematic block diagram showing the different small-scale structures of foresets of different type: simple cone-shaped grainflow foresets, grainfall laminae and climbing-ripple strata. Plane-bed lamination, is commonly developed on exposed dune crests, but is not shown here. B) Map and cross section of dune foreset cross strata exposed on a planed-off sinuous transverse or barchanoid ridge dune, showing the distribution of small-scale foreset structures. Simplified from an exposure on Padre Island, Texas, USA. Both after Hunter (1977).

Formation of small-scale structures

The slip faces of aeolian bedforms generate the foresets of cross beds and studies of their small-scale structure show that these have a varied origin.

Grainflow strata represent the deposits of lee slope avalanches initiated by slumping when the lee slope of an aeolian dune exceeds the angle of repose (32-34^{o}) and an active slip face develops that is subject to gravity-driven collapse, resulting in the generation of various types of grainflow (sandflow or avalanche) strata (Figs. 6.97, 6.98). Their inverse grading is due to the dispersive pressures and kinetic sieving generated by colliding grains (see §3.7.2). The largest, roundest grains flow to the toe most rapidly and this accounts for the lateral grading and the vertical inverse grading in those parts.

Gravity-driven **grainfall** occurs as the wind carries clouds of saltating grains over a dune brink. Grainfall occurs because of a reduction in wind transport capacity usually as a consequence of airflow separation within the lee-side eddy on the upper part of a dune lee slope. Grainfall may additionally occur on the flanks or apron of a dune where secondary cross-wind airflow is not strong enough to form ripples (Figs. 6.97, 6.98). Distinct grainfall lamination occurs because of grain-size variations produced during fluctuations of transport power. Repeated grainfall deposition on the upper lee slope is the main mechanism by which the slope attains and exceeds the angle of repose, thus inducing reworking of grainfall strata by avalanche processes.

Lower-angle inclined wind-ripple strata that preferentially form in lower parts of dunes are produced by the migration of climbing ripples which may occasionally preserve internal foreset laminae. Generally, however, internal foresets are absent and the climbing ripple strata generate pseudo-laminae each consisting of the coarser grains at the top of the laminae which have migrated along the crests of the ripples and finer grains at the base which have migrated in the lower part of the ripple (Figs 6.22, 6.23).

Flat bedded lamination arises when the reptation and saltation which produces ripples is inhibited during gale-force winds. It arises in conditions equivalent to those of the lower part of the upper flow regime in aqueous flows.

Formation of cross bedding

Cross bedding is ubiquitous within aeolian dune sands and sandstones, and develops as a consequence of repeated and ongoing lee slope sedimentation whereby grainflow, grainfall and wind-ripple strata generate cross stratification (Figs. 6.97, 6.98). The interior of most aeolian bedforms is composed of cross-bedded sands and the stratification planes provide a record of the former positions and shape of the bedform lee slope and of the processes that operated on that slope. Where bedforms migrate over one another, cross strata are truncated and sets delineated by erosive bounding surfaces are generated.

The migration of simple, two-dimensional, invariable (i.e. constantly moving) transverse bedforms will generate cross strata characterized by constant foreset dip azimuths. The migration of three-dimensional transverse forms such as barchan and parabolic dunes, sinuous-crested barchanoid ridges and, to some extent, dome-shaped forms will generate cross strata that exhibit a range of foreset dip azimuths that vary around a mean that reflects the direction of the resultant effective wind. Variance will increase as the amount of crestline sinuosity increases, for example from straight-crested transverse to dome-shaped dunes. The predicted pattern of foreset dip azimuths for linear, especially seif, dunes is of two modes about 120° apart. However, cross-bedded sets with such foreset arrangements are rare in the ancient record, chiefly because most linear dunes have minor components of transverse motion causing them to shift laterally over prolonged periods and resulting in the preferential preservation of foresets with a more unimodal dip-azimuth arrangement. Where bedforms undergo temporal changes in migration direction, speed, asymmetry and/or steepness, more complicated patterns of cross strata are produced such that sets contain numerous erosive reactivation bounding surfaces. Wind reversals may locally preserve foresets with azimuths at 180° from the mode in the bases of larger scour troughs. Similarly, if deposition were to occur from time-to-time on stoss surfaces, then rare, less steep laminae with azimuths at 180° from the mode could occur. For complex and compound draa, where bedforms of different scales migrate over one another at different rates and usually in different directions, more complex patterns of cross bedding will be generated, as will be the case for bedforms that are out of equilibrium.

Cross bedding dip-azimuth data derived from ancient successions (both from outcrop and borehole dipmeter) can be interpreted through comparison with these predictions only with great caution, especially where complex or compound forms are suspected. One approach that has been attempted involves the use of computer simulations whereby the patterns of cross bedding generated by bedforms with various morphological forms and that undergo various styles of migratory behaviour are compared to the observed cross bedding patterns from ancient strata until a good match is found. However, solutions found using this technique could potentially be non-unique.

Formation of bounding surfaces

Erosional bounding surfaces are generated as an intrinsic product of aeolian dune migration whereby bedforms (or parts thereof) scour into pre-existing deposits as they move through space. One potential problem with the analysis of aeolian bounding surfaces is that geometrically similar bounding surface configurations can be produced by a variety of styles of bedform behaviour. Three-dimensional computer simulations have been successfully used to demonstrate how specific arrangements of bedforms can generate highly complex bounding-surface geometries. Reactivation, superimposition and interdune migration bounding surface are recognized to occur as a product intrinsic bedform migratory behaviour, whereas supersurfaces result from externally controlled changes to controlling parameters such as sand supply, sediment availability and the sediment carrying capacity of the wind.

Reactivation surfaces (see §6.3.5) result from episodic or periodic lee slope erosion followed by renewed sedimentation

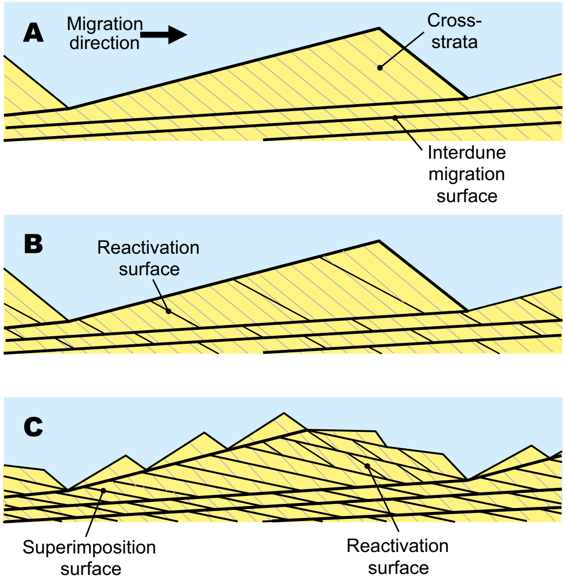


Figure 6.99 Generation of bounding surfaces in aeolian strata as a consequence of bedform migratory behaviour. A) Interdune migration surfaces. B) Reactivation surfaces. C) Superimposition surfaces arising from the migration of small bedforms over more slowly migrating parent bedforms. Reactivation surfaces may be nested within sets bounded by superimposition surfaces.

associated with a change in bedform migration direction, migration speed, asymmetry, or steepness (Fig. 6.99B, 6.100). The surfaces record erosion during anomalous wind intervals and compare with reactivation surfaces in aqueous cross bedding (see § 6.2.7). These changes are common because airflow on lee slopes is commonly subject to turbulent modification and is rarely steady. In some cases, the period of the flow fluctuation is regular and generates cyclic reactivation surfaces, as is the case for diurnal and seasonal wind reversals. Nested reactivation surfaces on two or more scales occur when cyclic cross bedding is generated by the interaction of two or more forcing parameters operating with different periodicities (Fig. 6.101).

Superimposition surfaces result from either the migration of superimposed dunes over a larger parent bedform, or the migration of scour troughs on the lee slope of a bedform (Figs. 6.99C, 6.100). The degree of curvature of these bounding surfaces gives some measure of the three-dimensional shape of the bedform responsible. Although superimposed dunes and scour troughs can theoretically migrate directly up or down the lee slope of a parent bedform, oblique migration is more common because secondary airflow tends to direct superimposed bedforms obliquely across the lee slope of the larger host bedform.

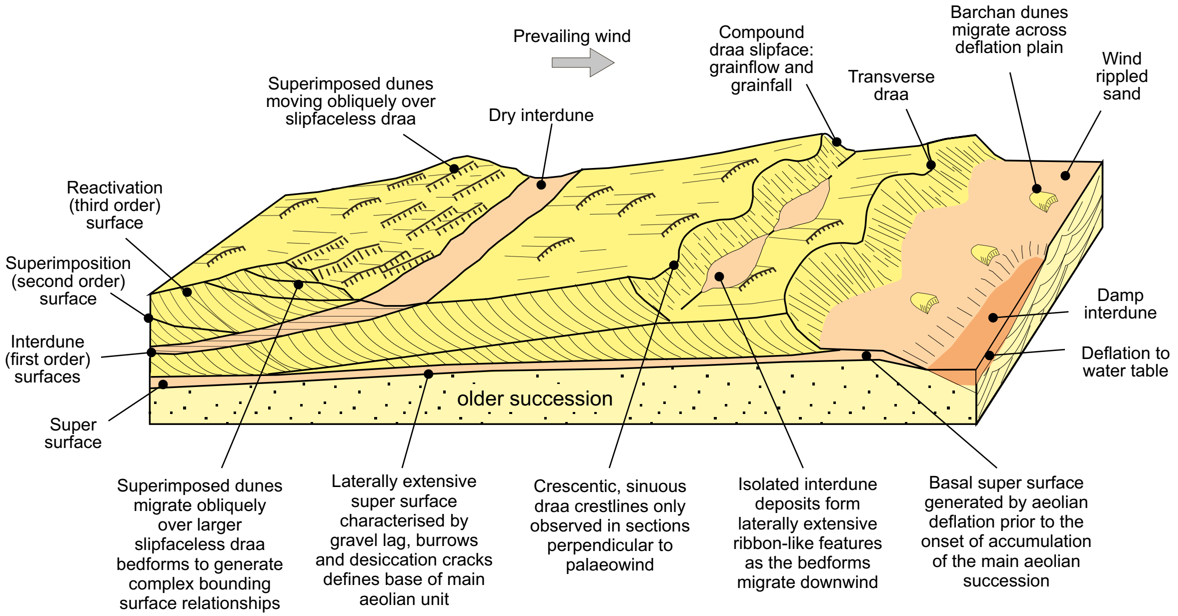


Figure 6.100 Schematic illustration of how various bedform scales and types migrating in various directions and at various rates, sometimes over one another, generate a hierarchy of bounding-surface types. Modified after Howell (1992).

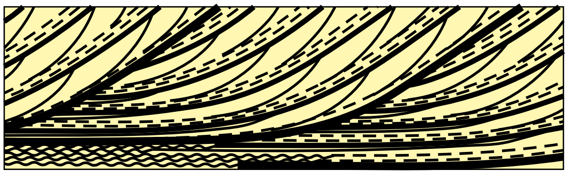


Figure 6.101 Schematic diagram illustrating coset of cross strata with internal cyclicity. Two distinct scales of bounding surface are evident within the coset. Note how bounding surfaces at the base of the sets pass down dip into wavy or corrugated surfaces. This relationship is indicative of aeolian dune migration that occurs synchronously with accumulation within damp, water table-controlled interdunes. This type of structure can potentially occur on a variety of scales from a few tens of centimetres to tens of metres. Based on observations from the Jurassic Entrada Sandstone, northeastern Utah. Modified from Crabaugh and Kocurek (1993).

Interdune migration surfaces result from the migration of bedforms separated by interdunes. The surfaces are carved by the passage of an erosive scour that defines the interdune trough between successive bedforms (Fig. 6.99A, 6.100). The depth to which the interdune trough scours as it migrates influences the extent to which deposits of the preceding bedform are eroded.

Supersurfaces are generated when aeolian accumulation ceases and is replaced either by bypass or deflation due to a switch from a positive to a neutral or negative sediment budget, respectively. Deflation may occur until either the net sediment flux once again becomes neutral or positive, or until further deflation is prevented because the accumulation has been deflated down to, for example, the water table. The rate of deflation may be controlled by the rate of water table fall. Where it is the water table that acts to limit the extent of aeolian deflation the resultant supersurface is sometimes called a **Stokes' Surface** (Fig. 4.28).

6.8.6 The uses of aeolian dunes, draa and cross bedding

The greatest use of these structures is in interpreting palaeo-environments. Aeolian rocks contribute to palaeo-environmental reconstructions at continental and global scales because they help to locate major climatological systems. Palaeowind studies are particularly helpful in this regard, but remember that dunes and draa reflect the effective winds, not necessarily weak prevailing and/or seasonal winds. Coarser-grained sand may be moved only by strong winds which may relate to only one season and/or direction. There is no reason why palaeowind directions should have a direct relationship with palaeoslope.

Although aeolian processes produce distinctive sedimentary textures, they can only be reliably used for palaeoenvironmental reconstruction when associated with additional diagnostic sedimentary structures. The intimate association of aeolian dune processes with fluvial, lacustrine and coastal processes means that sediments with aeolian textures are frequently reworked by non-aeolian processes.

Aeolian structures can help to establish way-up in highly dipping sequences, notably cross bedding for which lower toesets are asymptotic with the basal set bounding surface. Care is needed, however, for straight, angular-based foresets. Additionally, rare convex-upwards cross bedding may not be recognized to be the right way up without consideration of the whole succession and context.

Finally, thick aeolian successions of porous, permeable, cross bedded sandstone with few impermeable barriers are important as potential reservoirs for oil, gas and water, and as hosts for hydrothermal metalliferous deposits.

Study techniques

Field experience

Present-day environments

Processes, bedforms and structures in sands are open to study in most environments, sandy beaches, sandy tidal flats, dry river beds and areas of windblown sand being particularly well suited to detailed study by student groups. In many situations it will be possible to relate the surface form of bedforms to their internal sedimentary architecture by digging trenches or examining natural or quarried sections. More ambitious studies might make use of ground penetrating radar (GPR) techniques. Useful observations can additionally be made in driven snow. Large bedforms, for example aeolian dunes in desert environments, can be examined using remotely sensed data sets such as aerial photographs or satellite imagery, including data sets acquired from repeated surveys that yield a time series of images for a particular place, thereby enabling assessment of temporal rates of change in form or position of bedforms. Investigation may be enhanced by using aerial drones (where permitted) to collect images.

Ancient environments

The ancient record is equally easy to study, because sandstones are commonly well exposed and structures are easy to observe, measure and record. Although the surface form of small-scale structures such as ripples are commonly preserved on bedding surfaces, larger-scale structures can usually only be seen in sections that permit three-dimensional appreciation and analysis.

Laboratory experience

Structures developed in unidirectional currents

Many small-scale structures developed in sand, including asymmetrical current ripples and primary current lineation, can be generated in a small-scale flume tank. Experiments should employ sands with a variety of mean grain sizes and sorting characteristics, and should assess the significance of changes in flow discharge and speed. Note critical flow rates above which ripple development commences, measure ripple wavelength and height, and note planform morphology and internal structure.

Structures developed in bidirectional currents

Symmetrical wave ripples can be generated in a small water-filled tank with 0.05–0.08 m of fine- to medium-grain sand on its base. Place the tank on two cylindrical rollers (e.g. wooden rolling pins) and gently rock the tank back-and-forth to generate surface waves. Note how the waves agitate the sand on the bed and observe how symmetrical ripples develop after a few tens of seconds. Repeat the experiment with various water depths and vary the wave size by changing the intensity of the rocking motion.

Graded bedding

Add sediment composed of a variety grain sizes (silt to granules) to a 1L measuring cylinder and fill with water. Shake the cylinder to place the sediment into suspension and allow it to stand for 20–30 minutes. Note the resultant graded bed that develops.

Recommended references

Allen, J. R. L. 1982. *Sedimentary structures: their character and physical basis*. An encyclopaedic account of sedimentary structures and the physics of their development.

Kneller, B. 1995. *Beyond the turbidite paradigm: physical models for deposition of turbidites and their implications for reservoir prediction*. A landmark paper on deposition from turbidity currents in their wider setting.

Hunter, R. E. 1977. *Basic types of stratification in small aeolian dunes*. A detailed and widely cited account of aeolian lamination types with good illustrations.

King, C. H. J. 1980. *Experimental sedimentology for advanced level students using a motorized wave tank*. Useful techniques for designing laboratory experiments.

Lancaster, N. 1995. *Geomorphology of desert dunes*. Good discussion of the origin and geometry of stratification types associated with aeolian bedforms.

Leclair S. F. 2002. *Preservation of cross-strata due to the migration of subaqueous dunes: an experimental investigation*. An important account of the relationship between cross bedding and the dunes responsible for generating it.

McKee, E. D. (ed.) 1979. *A study of global sand seas*. A series of papers on the geomorphology of deserts and the sedimentology of desert dune deposits, well-illustrated with aerial and satellite images.

Mutti, E. 1992. *Turbidite sandstones*. A lavish picture book of turbidites and their structures.

Pettijohn, F. J. & P. E. Potter 1964. *Atlas and glossary of sedimentary structures*. A compilation of excellent photographs of the major types of erosional and depositional structures.

Pettijohn, F. J., P. E. Potter, R. Siever 1972. *Sand and sandstone*. Some useful material on sedimentary structures, as well as more general aspects of sedimentology.

Reineck, H. E. & I. B. Singh 1980. *Depositional sedimentary environments*. A beautifully illustrated book with good photographs of modern structures, but a rather patchy treatment of sedimentary environments.

Rubin, D. M. 1987. *Cross-bedding, bedforms and palaeocurrents*. Application of computer models to demonstrate how bedforms of varying shape, migration speed and migratory behaviour generate simple, compound and complex styles of cross bedding. Beautifully illustrated and well ahead of its time. At the time of writing, the computer modelling software could be downloaded free of charge from the USGS website.

Scholle, P. A. & D. Spearing (eds) 1982. *Sandstone depositional environments*. A well-illustrated compendium of environmental facies models with some illustration of individual structures.

CHAPTER 7

Depositional structures in gravels, conglomerates and breccias

7.1 Introduction

The general name for sediments containing a significant proportion of granule grade or coarser material is **rudites**. Studies of the depositional processes and structures of rudites are limited because the entrainment, transport and deposition of such sediments occur in high-energy environments where flow conditions make direct observations difficult. Direct-recording instruments, including the human body, tend to be severely damaged by the motion of large clasts and the situations in which they may be deployed for successful data collection are limited. However, as our knowledge of processes becomes more refined, so the features to be observed, measured and recorded become clearer. The transport and deposition of rudites is closely associated with a variety of both continental and marine environments including rivers, alluvial fans, reef talus slopes, storm beaches, submarine canyons, volcanic slopes and glacial settings, and as a result the compositions of rudites are extremely varied (Fig. 7.1). The installation of sediment traps in stream beds can give useful information on transport *rates* during floods but tells little of the *style* of transport and deposition. Some workers have attempted to overcome these problems by studying the day-by-day results of diurnal rise and fall of discharge on bedforms, for example in proglacial outwash areas. Processes are deduced from the products, revealed by both the surface morphology and the internal structures as revealed by trenches. Such methods are usually only applicable in accessible subaerially exposed settings. Laboratory experiments on gravels are increasingly attempted, but have been restricted by the need to build large and costly flumes or wave tanks. Even then, the scales of flows and structures are much smaller than the real phenomena. The study of conglomerates (rudites dominated by rounded clasts) and breccias (dominated by angular clasts) is an area of geology where detailed observation and interpretation of ancient deposits contributes significantly to understanding of present-day processes, particularly in deep-water settings so that the past becomes the key to the present. The careful analyses that allow these advances involve the recording of bed contacts, bed thicknesses, the style of framework or matrix support and the sizes and orientations of the larger clasts.

7.2 Classification

7.2.1 Defining rudites

There is no universal agreement on the percentage of clasts coarser than 2mm (–1φ) which must be present in a deposit before it is classified as a rudite (Fig. 7.2). Where there is a mixture of mud, sand and gravel, we recommend that the rock should contain more than 30 per cent by volume of clasts larger than 2mm before the terms **gravel**, **conglomerate** and **breccia** are used. In the field or in the laboratory, first try to estimate the percentages of gravel, sand and mud present, and refer the sediment to the classes shown in Figure 7.2.

7.2.2 Defining a sedimentary "structure" in rudites

The term sedimentary "structure" is here interpreted broadly to include several mass properties including textural features: (a) features based on composition; (b) features such as shape, roundness and surface morphology of the constituent clasts; (c) stratification and cross stratification; (d) features based on grain-size distribution, sorting and clast-support systems; (e) features based on fabric, packing and porosity; and (f) the presence and type of graded bedding. The first two properties should be recorded in any preliminary survey; they provide useful pointers concerning the provenance (i.e. the source regions) and the transportation history of the clasts. The last four properties demand particular attention if the aim is to understand processes and environments of deposition.

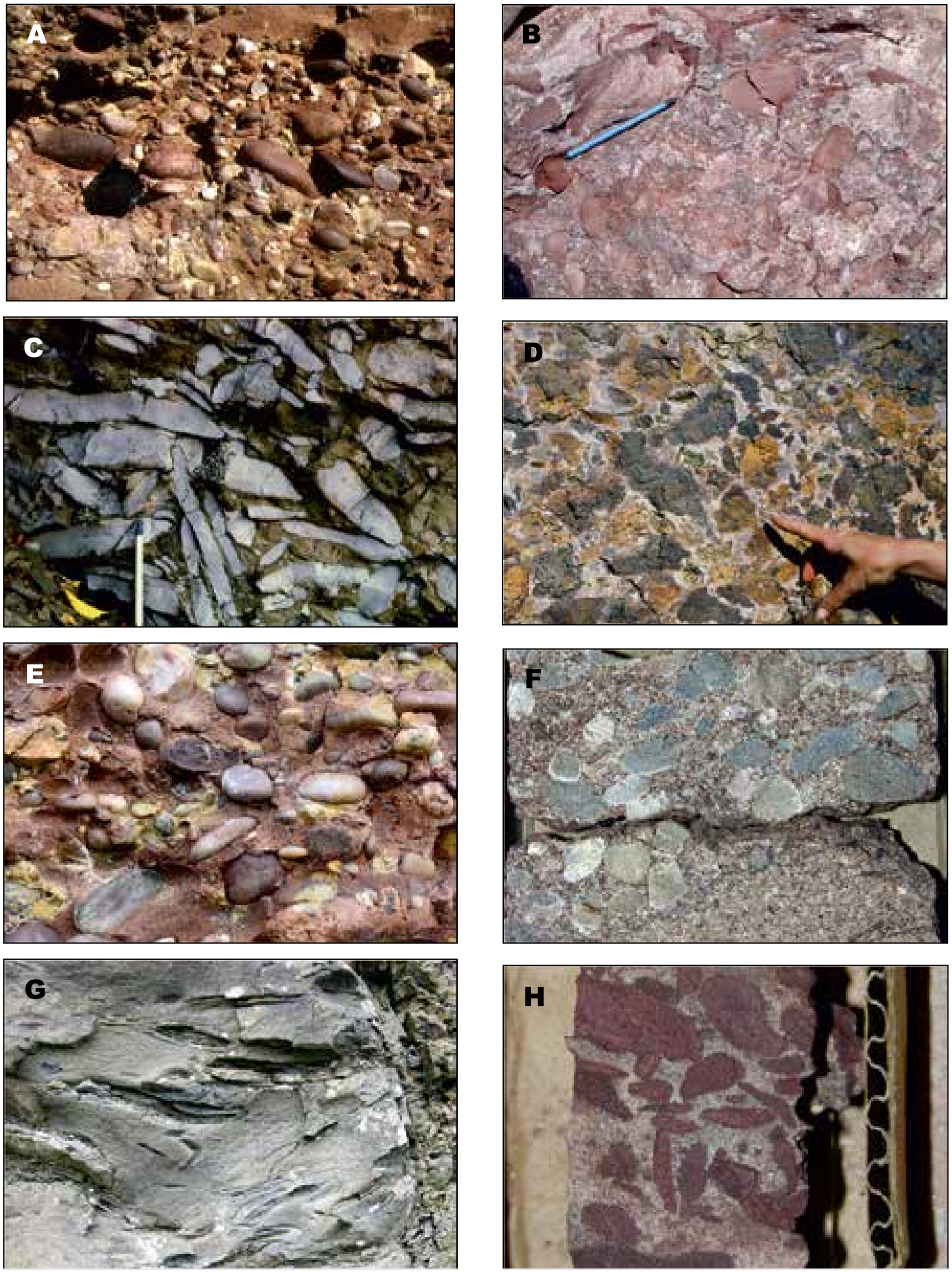


Figure 7.1 Examples of rudites of varying composition. A) Conglomerate with clasts composed of quartz, igneous and metamorphic pebbles and cobbles. Hawksmoor Sandstone Formation, Triassic, Staffordshire, England. B) Conglomerate composed of intraformational clasts of locally derived red mudstone. Old Red Sandstone, Devonian, South Wales. C) Conglomerate composed of intraclasts of limestone. Note the development of interlocking clasts as a consequence of partial dissolution under pressure. Silurian, New York State, USA. Photo courtesy of Gilbert Kelling. D) Basaltic clasts of volcanigenic origin within a matrix of aeolian sand grains. Etjo Formation, Cretaceous, NW Namibia. E) Well rounded pebbles of mixed affinity in a sandstone matrix. Hawksmoor Sandstone Formation, Triassic, Staffordshire, England Largest pebbles are ~8cm diameter. F) Clast supported framework in conglomerate at base of channel sandstone. Cored well, Upper Carboniferous, southern North Sea Core is ~12cm diameter. G) A mud clast intraformational conglomerate with a "slurried" sandstone texture. Hybrid mass flow. Mam Tor Sandstone, Upper Carboniferous, Derbyshire, England Clasts are each 1–2cm thick. H) Mud clast conglomerate with coarse sandstone matrix. Mud clasts reddened by later oxidation. Cored well, Upper Carboniferous, southern North Sea Core is ~12cm diameter.

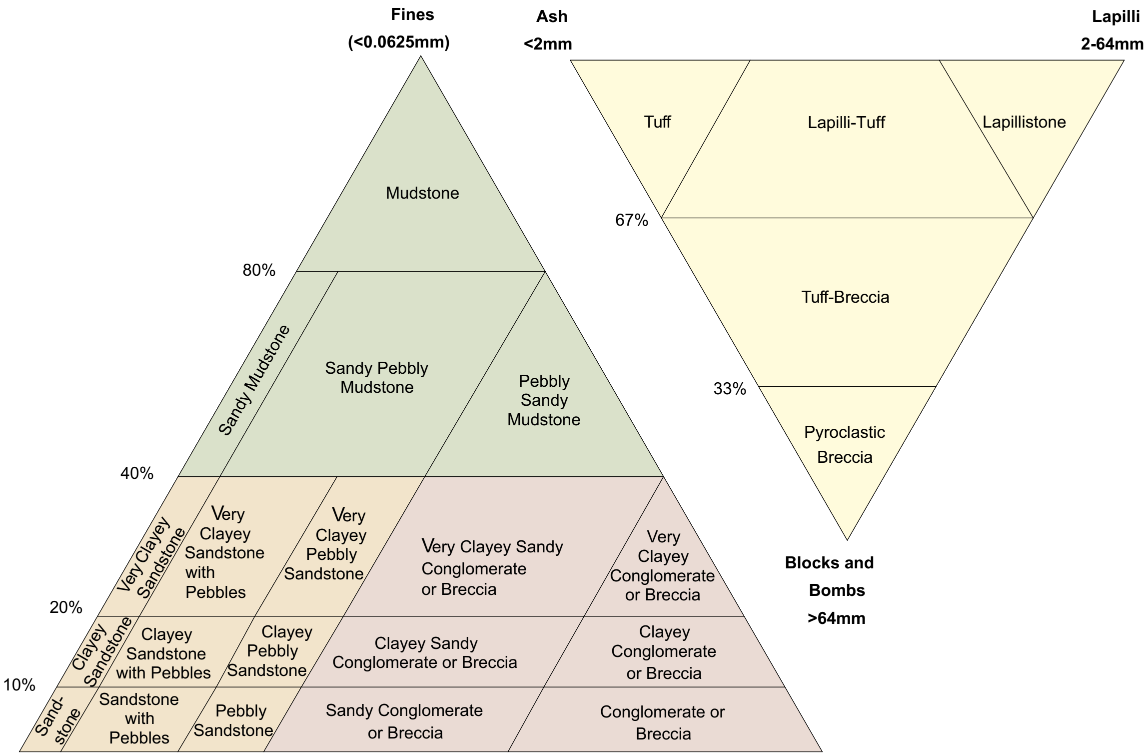


Figure 7.2 The nomenclature of rudites. A) Consolidated siliciclastic rudites defined by the proportions of the grain sizes of their component clasts. Modified after Piper and Rogers (1980). B) Pyroclastic rocks defined by the proportions of the grain sizes of their component clasts. Blocks consist of angular clasts, bombs of twisted, concentric solidified blobs of lava (sometimes partly molten in their centres at the time of eruption); precise percentage boundaries for mixtures vary depending on the prejudices of the user.

7.2.3 Composition and classification of rudites

One of the first properties to be recorded in the field is the composition of the large clasts (Fig. 7.1). This allows useful, preliminary conjectures concerning their possible provenance and their processes of origin, for example whether the sediments may be pyroclastic in origin. Observations should also try to establish whether some of the clasts originated from within the basin of deposition and were eroded from penecontemporaneous sediments, i.e. are **intraformational** (e.g. reef talus), or whether they come from a source area where older rocks outcrop, i.e. are **extraformational** or exotic.

7.2.4 Misidentification of rudites

Some rocks provisionally classified as rudites of sedimentary origin will, with further analysis and experience, be regarded as "pseudoconglomerates" or "pseudobreccias". These may originate from processes of *in situ*, post-depositional diagenetic change (e.g. concretions; see §9.3.1), tectonic disruption (e.g. fault breccias), or simply weathering (e.g. dolerite blocks that are undergoing spheroidal weathering are commonly surrounded by an altered clay-rich matrix). Likewise, features initially thought to be clasts may later prove to be trace fossils (e.g. rounded burrow fills, see §9.4).

7.3 Morphology and general settings of gravel deposition

7.3.1 Introduction

The large grain size of gravels prevents the formation of many of the bedforms and structures that are common in sands. For example, small-scale ripples will not form in sediments whose clast diameters approach the size of the ripples themselves. However, although rare, large ripples (some cases called mega-ripples) composed of granules and even small pebbles are recorded in some aeolian

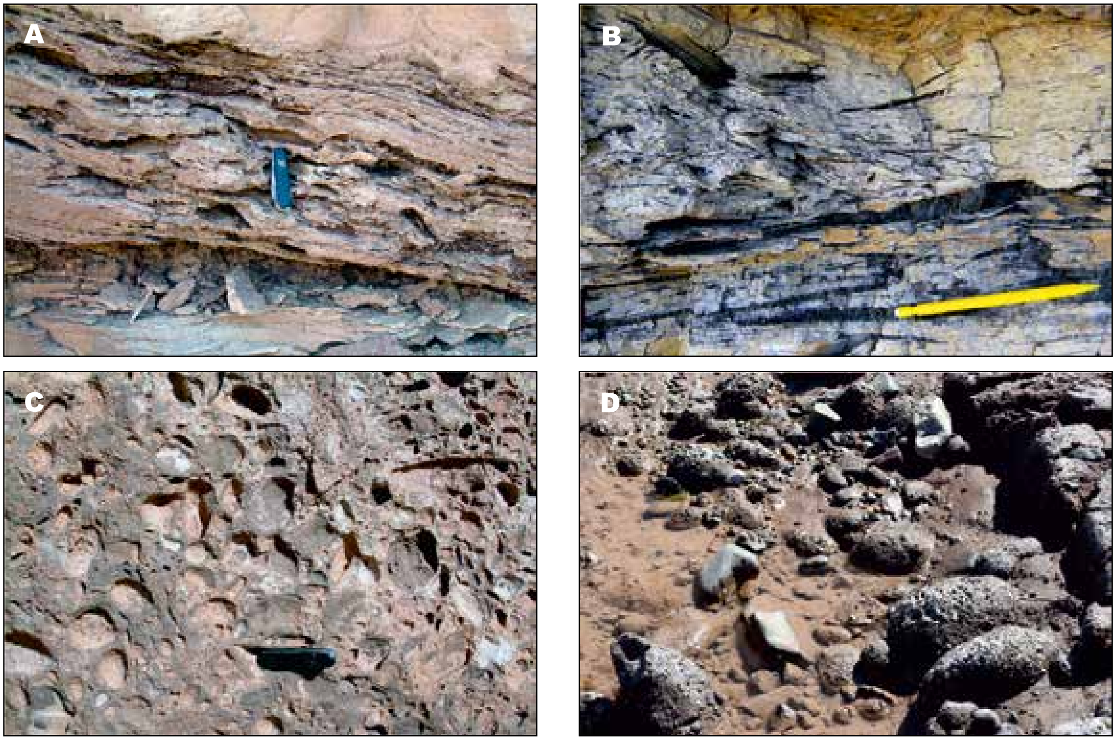


Figure 7.3 Conglomerates composed of different types of intraclasts – clasts that are locally eroded from the surrounding environment in many cases shortly prior to later deposition. A) Concentrations of mud clasts aligned parallel to bedding in a sandstone unit. Kayenta Formation, Jurassic, Utah, USA. B) Carbonaceous plant debris preserved as fragments of coal reworked as clasts. Neslen Formation, Cretaceous, Utah, USA. C) Flattened and rounded mud intraclasts on a bedding surface in sandstone. Some of the original mud clasts have been eroded out to leave rounded cavities (moulds). Kayenta Formation, Jurassic, Utah, USA. D) Rounded pebbles and cobbles of highly cohesive mud that have been armoured through the adhesion of smaller bedrock pebbles (up to 25cm diameter). Present day, Anglesey, north Wales.

environments (Fig. 6.79; see §6.8.2). Large dunes may be formed in coarse sand with a subsidiary component of gravel and subaqueous dunes formed entirely of fine gravel are known in some rivers. In exceptional cases of catastrophic flooding due, say, to the bursting of glacial lakes, very large dune-like forms may be developed in coarse gravels. The morphological features developed in gravel are relatively large and they may have distinctive packing fabrics developed within or superimposed upon them (see §7.4.4). Such fabrics commonly form the basis of interpretation of ancient gravels and conglomerates.

Unlike sands, there is no all-embracing scheme for gravels wherein the occurrence of particular bedforms can be related to specific flow conditions and grain sizes. However, in general, the larger and coarser the sediments the stronger the flow. These coarse deposits are increasingly interpreted as the products of two major sets of processes, those associated with bed-load transport and those associated with sediment gravity flows. Here we outline some gross structures and some environments of gravel deposition within which these processes may be active. In addition to the environments discussed below, rudites also develop through catastrophic events such as submarine slides and tsunamis.

7.3.2 River channels

Streams with gravel beds commonly show highly variable patterns of **bars** and **channels**. In braided rivers, bars elongated parallel to flow commonly split the stream on several co-existing scales. Deposition takes place on the flat bar tops, where pebbles lodge and may become interlocked in an imbricate packing (see §7.4.4), and at the down-stream ends, where avalanching gives rise to cross beds. Bars vary in relief from lobate sheets, only one clast thick, to forms several metres high. In more sinuous streams, gravel accumulates on **lateral** or **point bars**. In large rivers, bars are commonly composites of smaller features and the resultant gravel bodies are complex, tabular sheets. Channel fills are more restricted in extent and may appear lenticular. Deeper channel areas commonly have beds covered with the coarsest gravel, but this is commonly masked by finer sediment laid down from weaker currents during falling river stage. That is what is seen when the bed is sub-aerially exposed, but remember that some sand may also be wind-blown if the bed has been exposed for a significant period.

7.3.3 Sub-aerial fans and steep slopes

Episodic floods on alluvial fans commonly are characterized by a rapid rise to peak discharge, followed by a rapid waning stage. Such events promote the development of localized tongues and lobes of coarse sediment deposited by mudflows (cohesive debris flows) or generated as **sieve deposits**. In mudflows, the presence of abundant fine-grained sediment gives the flow a high viscosity and density (see §3.3) that enable large clasts to "float" in a finer matrix and resulting in poorly sorted, matrix-supported fabrics. Gravel lobes are produced by a sieving process whereby water transporting the gravel sinks into a permeable substrate leaving a clast-supported gravel on the surface.

On very steep slopes, both subaqueous and subaerial, loose blocks of rock fall down and form scree and talus cones. Such material tends to be angular and slopes of up to 35° (much steeper than the normal angle of rest) may develop where clasts interlock with each other. In subaerial settings, such deposits have a low preservation potential.

7.3.4 Fan deltas

Where a gravel-rich alluvial fan builds into a body of water such as a lake or the sea, a local wedge-shaped, gravel-rich fan delta may form. Because gravel is not readily redistributed by waves and currents, a steep delta slope may form at or close to the angle of rest, especially in low-energy lake settings. Some of the gravel may be transported into deeper water by processes of sediment gravity flow (see §3.7).

7.3.5 Shorelines

Debris at the foot of sea cliffs and eroding shorelines is quickly broken up and transported away, enabling wave attack to proceed further. The eroded material is typically transported along shore and gravel accumulation is a feature of many beaches, particularly storm beaches. These gravel ridges are broadly linear but some build out to produce sheet-like forms. Beach gravels are commonly very well sorted and the clasts are well rounded as a result of sustained abrasion. The clasts of such beach gravels commonly show subparallel inclination if they are flattened (i.e. imbrication: see §7.4.4).

7.3.6 Deep-water fans

At the foot of major submarine slopes, particularly close to the mouths of submarine canyons, turbidity currents and other sediment gravity flows deposit gravels, both as lobes and as components in channel fills. Knowledge of such settings comes mainly from the study of ancient sedimentary rocks.

7.3.7 Reefs and associated settings

On steep reef fronts, pounded by ocean waves, irregular carbonate clasts, sometimes of very large size and composed of reef material become detached, fall downslope and accumulate in steeply inclined wedges (up to 40°) which comprise interlocking frameworks with high initial porosities (up to 40%). Such talus deposits have high preservation potential and are well known from the ancient record.

7.3.8 Glacial and associated settings

Glaciers carry sediments of all sizes and deposit them in a variety of settings, with and without the aid of water. Some debris is dumped directly from melting ice as sheet-forming basal **lodgement till**, or more variably shaped, plastically deformed, surficial **flow till**; other coarse-grained material falls from floating icebergs and shelf ice to give isolated **dropstones**. Material deposited both subaerially and subaqueously from melting ice is commonly subjected to further movement and deposition by mudflows. Melt water flowing within, above, below or lateral to a glacier may sort, transport and deposit sand and gravel bodies to form geomorphic features such as **kames**, **eskers** and **outwash plains**. These gravels tend to be better sorted than those deposited directly from ice. The complex variety of morphological features associated with glacial and glaciofluvial activity is more fully covered in specialized textbooks.

7.3.9 Deposits associated with volcanoes

Coarse debris ejected during volcanic eruptions accumulates both as widespread sheets and as more restricted bodies. **Autoclastic** processes generate clasts through mechanical breakdown and gaseous explosion during the movement of magma and lava; **hyaloclastic** processes occur when lava is quenched and shattered by entry into water, water-saturated sediment or ice; **pyroclastic** debris is produced by explosion of magma and country rock. Pyroclastic processes operate in both subaqueous and subaerial settings. Airfall deposits result primarily from the settling from suspension of fine ash but are often interbedded with larger, typically randomly distributed dropstones of lapilli, blocks and/or bombs, which deform underlying strata. Such deposits typically drape pre-existing topography. Pyroclastic flow deposits are transported *en masse* as density currents that can carry a high proportion of large blocks. Because such flows are controlled by gravity, they are usually confined to valleys, which become progressively infilled. Some types of pyroclastic flow travel as base-surge density currents, which possess sufficient energy to rise up over topographic obstacles. Such surge deposits may be identified by being shown to thicken into topographic depressions and thin over topographic highs.

Many primary pyroclastic fragments become mixed during transport with "normal" siliciclastic and carbonate deposits, and are resedimented as **epiclastic** conglomerates.

Highly vesicular pumice, produced by some volcanoes may be sufficiently light to float for a while where is lands on water, sinking only when it is sufficiently water-logged. As water-logging acts more quickly on smaller clasts, this process has the potential to create inversely graded beds.

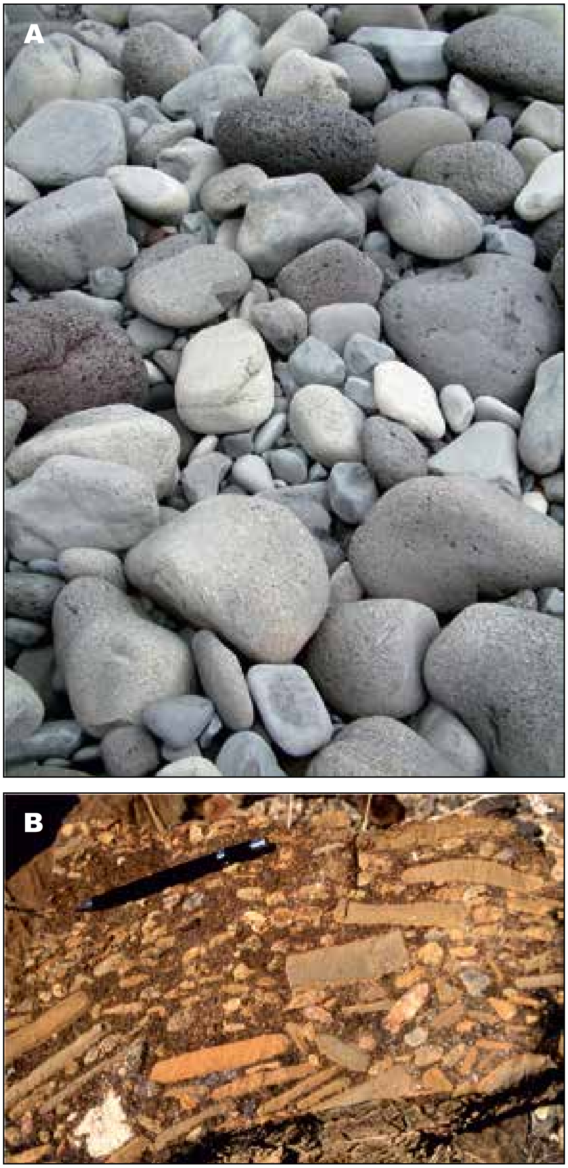


Figure 7.4 Examples of contrasting clast shape. A) Sub-rounded, moderately spherical, equant cobbles and boulders. Modern beach, southern Iceland. B) Extremely angular, low-sphericity, prolate clasts. Huab Formation, Permian, northwest Namibia.

7.4 Textures, fabrics, and structures: their mode of formation

In describing rudites, it is usually possible to make detailed observations of certain features directly in the field. These may be supplemented by observations and analysis of sediment samples in the laboratory. It is useful to note the following:

- the composition, colour, shape and surface features of the constituent clasts
- the sorting, grain-size distribution, porosity and clast-support characteristics of the sediment
- the nature of any vertical or lateral size grading within beds
- the presence of any preferred clast orientation
- the nature of any stratification and/or cross stratification
- bed thickness and grain-size relationships.

7.4.1 Shape (roundness and sphericity) and surface features of clasts

Shapes of clasts within rudites are extremely variable (Fig. 7.4). With experience, sphericity and roundness can be judged in the field by reference to visual comparators on a grain-size card. More detailed analysis can be performed in the laboratory by measuring sphericity and roundness parameters and by referring clasts to the categories of shape devised by Zingg (Fig. 7.5). Classes of roundness may be indicated by descriptive terms (from very angular to very rounded) or by using the numerical scales of Powers or Pettijohn (see reference list). Experience of measuring and handling individual clasts of various shapes is important in appreciating their three-dimensional properties, as clasts are commonly seen only in two dimensions in many rock exposures and borehole cores (see §7.4.4).

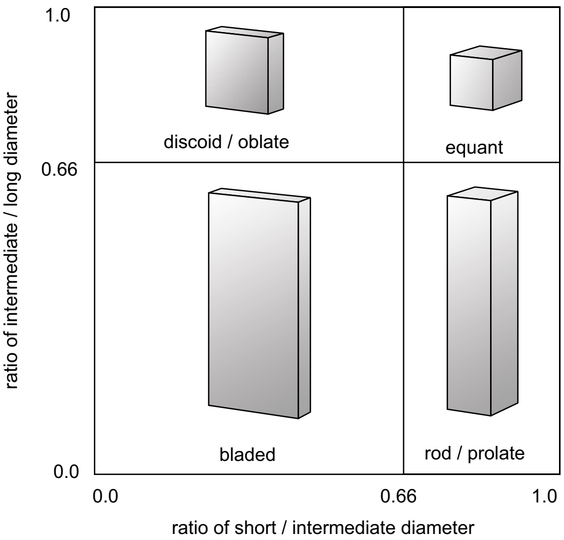


Figure 7.5 Scheme for the classification of clast shape. Equant- and disc-shaped clasts are most common. Modified after Tucker (1991).

Some shapes suggest processes of origin that allow one to make conjectures about the environment or transport history of the sediment which can then be tested by other means. Flat-sided clasts with two, three or four facets, the smooth surfaces of which may be pitted, fluted and polished, are known as **ventifacts** and represent wind-fashioned, sand-blasted objects. In most cases ventifacts will be rolled and reworked into later deposits, but rarely they may be found *in situ*. Similar, but unpolished, unfluted pebbles can be produced by wet-blasting. Tough, flat-iron pebbles bearing parallel or subparallel **striations** (scratches) and snubbed edges record glacial abrasion and attrition prior to deposition; softer material may be scratched in a range of environments. Concentrations of well worn, matt-surfaced, disc-shaped pebbles generally become more common as one goes from rivers, to low-energy, then high-energy beaches due to increased rates and duration of abrasion. Highly rounded, spheroidal pebbles arise through constant abrasion, attrition and reworking, sphericity reducing and roundness increasing from fluvial, to low-energy, then high-energy beach environments. The indices of sphericity and roundness provide measures of the textural maturity of a sediment. Note, however, that, given constant conditions, sand and small pebbles become rounded more slowly than do large clasts. Percussion rings due to high-velocity collisions are apparent on the surfaces of some clasts and should not be confused with pressure-solution pits.

7.4.2 Sorting, grain-size distribution, porosity and clast-support characteristics

Disaggregation and sieving of gravels, and the determination of their size distribution, can be performed in the laboratory, using a similar method to that for sands. However, when ancient conglomerates are well cemented, the approach is impossible. Nevertheless, it may be useful to estimate and describe these characteristics in an approximate way.

Initially determine the degree of sorting of individual beds and record whether there appear to be distinct grain-size modes. Very commonly there is a mode in the pebble-cobble size and another in the sand size; hence the sediment grain-size distribution is **bimodal**, the term **clasts** being applied to particles around the coarse mode and **matrix** to the fine components. Some deposits are **unimodal** and well sorted, lacking a well-defined matrix and hence having very high porosity (>40%). Other beds are unimodal and poorly sorted; yet others are **polymodal**.

An important distinction amongst rudites is the separation of **orthoconglomerates** and **paraconglomerates** (Figs. 7.6, 7.7). Orthoconglomerates have well sorted, clast-supported, openwork or matrix-filled frameworks, whereas paraconglomerates are poorly sorted, matrix-supported and with very variable grain-size distributions. Hence it is important to determine whether a deposit is clast- or matrix-supported. Are larger clasts in contact with each other, forming a framework, or are they dispersed in a finer-grained matrix (Fig. 7.6)? This is not always easy to judge in a two-dimensional outcrop without disaggregating the rock. A search for indentations of one clast by another (i.e. pressure pits) might help. Matrix-free, clast-supported, framework conglomerates commonly have a high porosity and permeability, and are described as having **openwork** textures. Alternatively, if the clasts are dispersed and supported by the matrix, try to establish whether the supporting matrix is of sand (e.g. in a sandy conglomerate) in which case the rock may be very porous and permeable, or of mud (e.g. in a pebbly mudstone), in which case the rock may lack significant porosity and have low permeability.

Rudites with a matrix dominated by mud are commonly mudflow deposits or possibly glacial till, with clasts that tend to be very poorly sorted and that may be striated. Volcaniclastic deposits are often characterized by blocks and bombs floating in a fine ash matrix. Dropstones of both glacial and volcaniclastic origin may be found within marine or lacustrine muds.

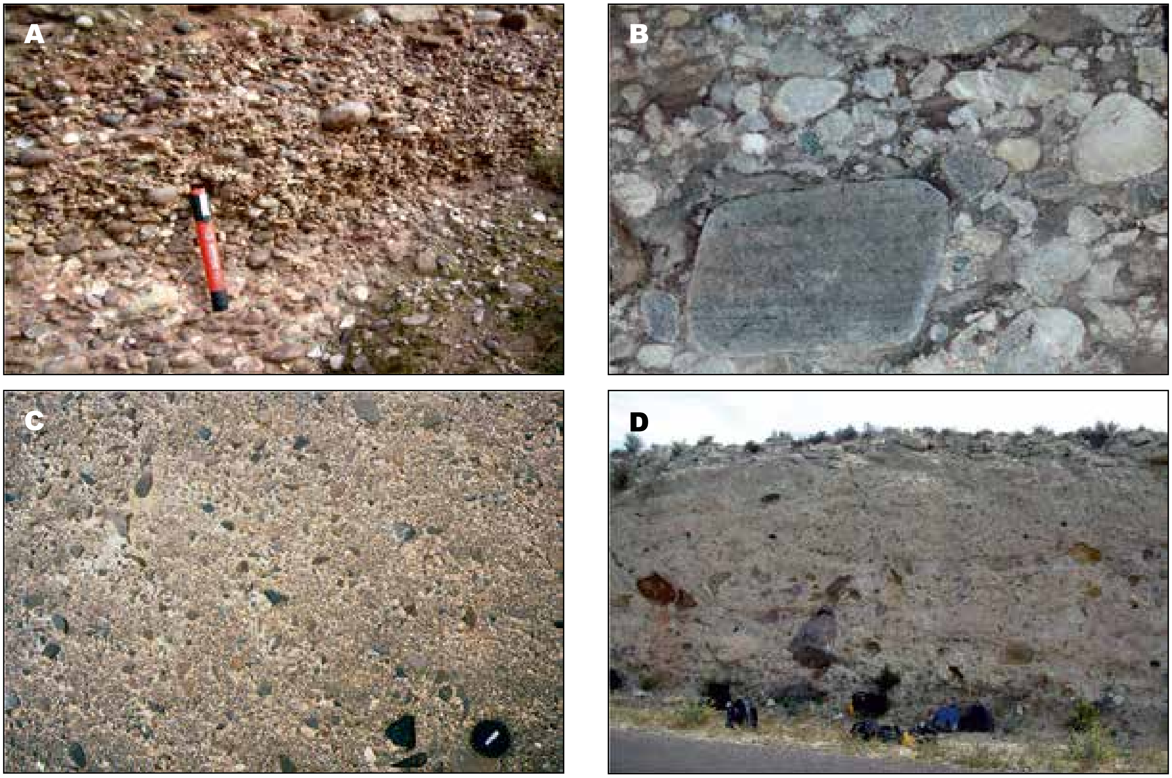


Figure 7.6 Examples of various styles of clast support within a finer-grained matrix. A) Orthoconglomerate with tight packing of pebbles and only a minor component of sand matrix. Hawksmoor Sandstone Formation, Triassic, Staffordshire, England. B) Tightly packed, clast-supported orthoconglomerate in a channel deposit. Well core, Upper Carboniferous, southern North Sea Large clast is ~4cm diameter. C) Paraconglomerate with angular clasts 'floating' in a poorly sorted sand matrix. Cretaceous, Namibia. D) Paraconglomerate of volcaniclastic origin with blocks and bombs floating in a lapillistone matrix. Cabo de Gata, Miocene, Almeria Province, southeast Spain.

Following these initial distinctions, further questions can help to generate ideas about modes of transport and depositional processes: the questions may not necessarily all have ready answers in some cases but they help to focus attention on potentially important issues.

- If the rudite is clast supported and has a matrix, was the finer-grained interstitial material deposited along with the larger clasts or did it fill the framework later?
- If the texture is clast supported but retains an open-work structure, did the current winnow away potential matrix-grade sediment or maintain it in suspension while rolling and sorting larger clasts on the bed?
- What does the bed indicate about the abundance of fine-grained sediment during transport and deposition?

Although the possible threshold velocities for the erosion and transport of each mode may be estimated from the Hjulström–Sundborg graph (Fig. 3.16), bear in mind that these data relate mainly to rivers with low suspended-sediment concentrations, and with beds of relatively uniform material. Figure 7.8 relates the sizes of clasts transported by rolling to the size of sand that could be suspended by the same flow. Where the conglomerate has a pebble framework which is filled with sand, however, these graphical relationships do not indicate whether the infilling was largely contemporaneous with deposition of the gravel or if it largely post-dated gravel deposition. In the former case, part of what originated as a heavy suspended load might have become trapped in the quieter interstices of the gravel bed. In the latter case, a river that in flood carried less sand in suspension could still have transported sand as bed load during lower-stage flows, some of which could infiltrate the gravel framework. In most cases it is difficult to judge these issues of timing and the process of emplacement of the matrix. Furthermore, wind-blown sand could fill a water-lain but subaerially emergent gravel framework without leaving much evidence of the process.

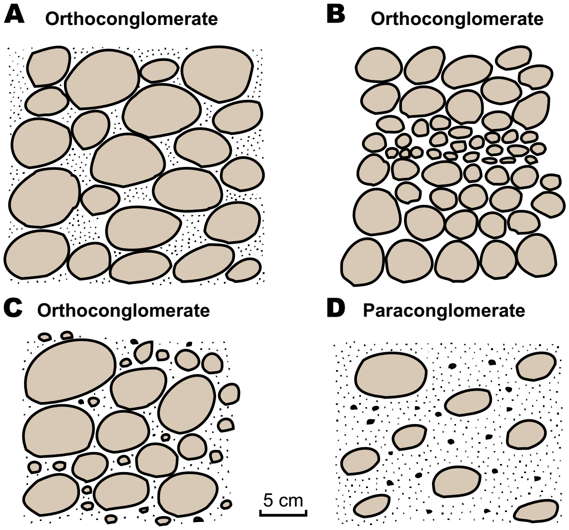


Figure 7.7 Descriptive features of sorting and size distribution in rudites. There is a spectrum in nature between types A and D. A) Bimodal, clast-supported framework; well sorted matrix (orthoconglomerate). B) Clast-supported, open-work framework (orthoconglomerate). C) Polymodal, clast-supported framework, poorly sorted matrix (orthoconglomerate). D) Polymodal, matrix-supported, lacking a framework (paraconglomerate). Modified after Walker in Harms et al. (1975).

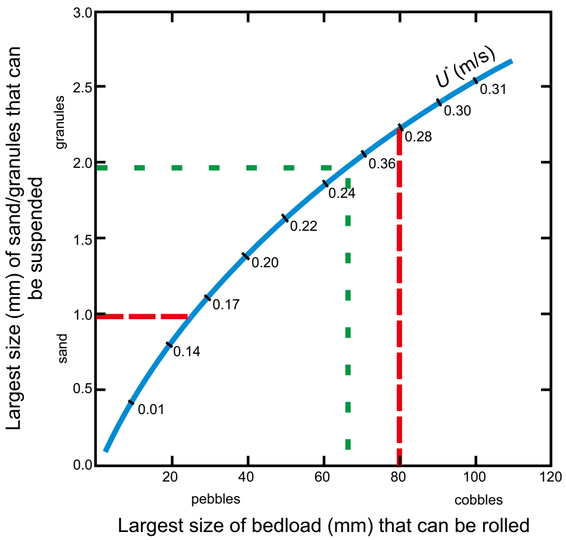


Figure 7.8 Graphical representation of size of clasts transported by rolling, compared with the size of grains suspended by the same flow. Numbers on the curve refer to critical values of shear velocity U* for bed rolling. The dotted green line shows that the co-existence in a conglomerate of pebbles of 63 mm and sand of 1.95 mm diameter relates to slight velocity fluctuations around a mean value of U* = 0.25 m/s. The dashed red lines show that the co-existence of clasts in a bi-modal, matrix-filled framework orthoconglomerate composed of cobbles of 80 mm and sand of 1 mm would probably result from the deposition from bed load of cobbles at high velocity (0.28 m/s), whereas sand of 1 mm was still in vigorous suspension. With diminishing velocity, the sand would fall into the framework and fill it. Modified after Walker (1975).

If the rudite is matrix-supported, consider whether the large clasts were transported along with the finer-grained matrix as a high-viscosity flow or whether they were dropped into already deposited finer-grained sediment. The occurrence of disturbed laminae in the matrix may help to decide the case. If the clasts are transported with the matrix, for example in a debris flow, there may be no internal indication of the setting in which the flow took place. Alluvial fans, deep-water turbidite fans and channels, the surfaces of glaciers and the sides of volcanoes are all possible settings. If the clasts were dropped into a deposit, we do not always know whether this took place from a melting iceberg, from a floating and rotting tree root, or as the result of a volcanic explosion. In all cases, this could have taken place in lacustrine or marine settings. Clasts of pumice can float for long distances before becoming waterlogged and sinking; they tend not to greatly disturb the underlying sediment when they settle. Normally, evidence on the depositional context of the rudites derived from surrounding sediments will help constrain such problems of interpretation.

Glaciers typically carry clasts that possess a broad range of range of sizes (and shapes). Glacial deposits in the form of tillite are, in almost all cases, poorly sorted and commonly form matrix-supported paraconglomerates as a result of the melting of ice that typically indiscriminately deposits any sediment that was being carried (Fig. 7.9).

Figure 7.9 A paraconglomerate with a wide range of clast size and a highly dispersed fabric characteristic of many glacially emplaced sediments. Upper Tillite, Late Precambrian, Tanafjord, north Norway.

Figure 7.10 Examples of normally graded beds in rudites. A) Fluvial braidplain gravels, modern, southern Iceland. B) Pyroclastic flow deposit with blocks and bombs. Cabo de Gata, Miocene, Almeria Province, SE Spain.

7.4.3 Grading within beds

The main styles of grading described in §6.7. for sands apply equally well in gravels and rudites (Figs 7.10, 7.11). Normal grading usually suggests that turbulence was developed during deposition (see §3.7). Content or distribution grading and coarse-tail grading are both common in normally graded rudites. Inverse grading is much more common in rudites than in sands. This can be explained by large clasts rising upwards due to dispersive pressures or to progressive loss of larger clasts from the lower, more strongly sheared part (traction carpet) of a powerful flow or in a grain flow. Inverse grading may also be a product of along-flow sorting whereby finer particles arrive first at the depositional site. Ungraded beds may indicate high shear strength or high viscosity, thereby preventing turbulence and effective grain interaction; in such circumstances there is little scope for differential settling to generate normal grading, or for dispersive pressure to generate inverse grading. **Lateral grading** of grain size is evident in some cases, especially in orthoconglomerates. In volcanic rocks, **density grading** occurs whereby clasts of different composition (and density) become vertically or laterally differentiated. This should be considered separately from size grading, especially where pumice or other vesicular clasts are involved. In water, smaller pumice clasts become waterlogged and sink more quickly and this can lead to inverse grading.

7.4.4 Fabric

Clast fabric analysis is used to describe the orientation of particular dimensions of the constituent clasts. In some deposits a strongly preferred orientation of clast long axes may occur; in others the short axes may be aligned; and in others no pattern may be discerned (Figs. 7.12, 7.13).

In lithified conglomerates measurement of the three-dimensional orientation of clasts may not be practicable, and a statistical resultant may have to be estimated from measurements of apparent axes in two-dimensional horizontal and vertical surfaces (Fig. 7.13B). In the laboratory, three-dimensional data can be plotted on a stereonet

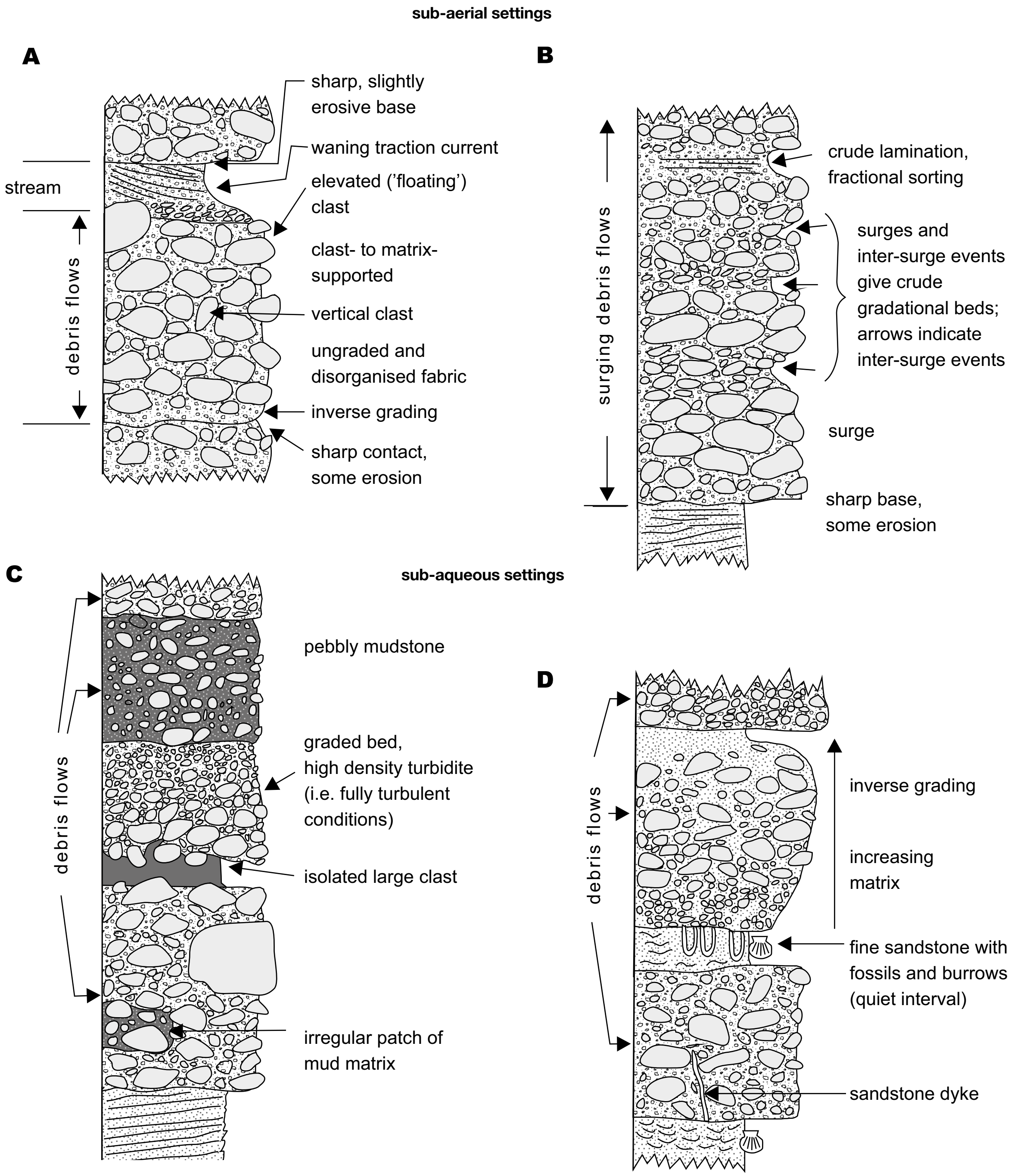


Figure 7.11 Examples of different styles of grading in conglomerates due to different depositional processes acting in a variety of typical sub-aerial (A, B) and sub-aqueous (C, D) settings. Bed thickness ranges from a few decimetres to metres. After Nemec and Steel (1984).

Figure 7.12 Examples of imbricated clast fabrics developed in rudites. A) Interlocking disc-shaped clasts. Transport from left to right. Present-day alluvial fan, Tabernas, southeast Spain. B) Rounded mud clasts above an erosion surface showing clear imbrication indicating transport from left to right as viewed (largest 2cm long). Location unknown. University of Leeds Collection. C) Imbrication indicating transport from right to left. Pleistocene, Gregory Rift, Kenya. D) Imbrication indicating transport from left to right. Quaternary, Siensas, Spanish Pyrenees.

(Fig. 7.14) or orientations measured in two dimensions can be plotted as rose diagrams (Fig. 7.13B). See also Appendix 1.

The most commonly occurring fabric, which arises when the rudite is rich in disc- and blade-shaped clasts, is **imbrication** (Figs. 7.12, 7.15). Here the flattened surfaces all dip in an up-current direction. This is common in horizontally bedded, clast-supported conglomerates, and two variants may occur. In one, the long a-axis is transverse to the dip direction of the clasts and the intermediate b-axis is parallel to the dip. In the other the b-axis is transverse and the a-axis is parallel to dip. Imbrication of either sort should be carefully distinguished from the preferred orientation of clasts in cross-bedded gravels. Flattened clasts on avalanche faces are oriented parallel to the face and may therefore dip down current, opposite to that of genuine imbrication. Other clasts roll down slope on slip faces with their long a-axes parallel to the slope.

Rudites with flattened clasts lacking preferred orientation suggest transport by processes in which the clasts were not completely free to move relative to one another, possibly by virtue of higher flow viscosity or a rapid rate of accumulation, which did not allow time for an organized fabric to develop.

Well organized fabrics suggest that clasts have been free to move individually and independently of one another above the bed and that they have been selectively incorporated onto the bed when they landed in a stable position. Clasts that landed in positions other than the stable, upstream-dipping orientation are prone to re-entrainment in most cases.

Theoretically, the attitude of each clast on the bed is its response to the combined forces that acted on it during and shortly after deposition. Gravity tends to keep the clasts in place, a lift force may or may not act upwards, and the drag force will try to roll the clast (Fig. 3.17). The most stable

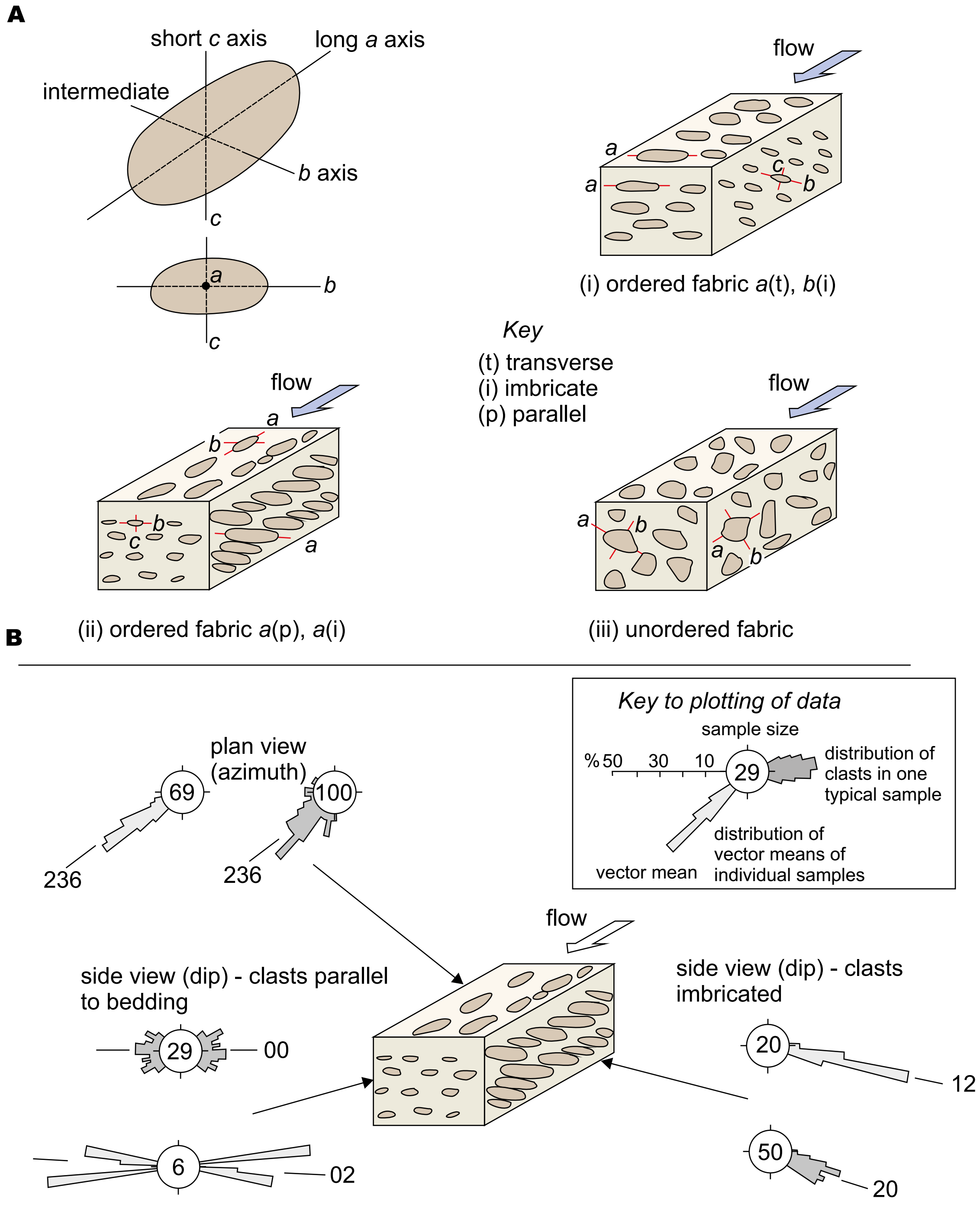
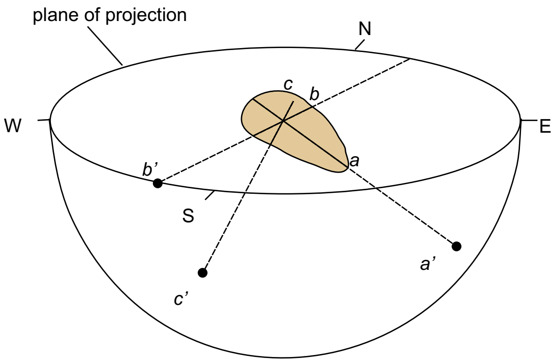


Figure 7.13 The fabric of rudites. A) The nomenclature of ordered and unordered fabrics. B) The plotting of fabric measurements for well cemented ordered rocks. Data collected from orthogonal faces not from three-dimensional measurements of clasts. Plan views of beds measured at 69 localities, 100 clasts at each place; imbrication measured at 20 localities on faces parallel to flow and perpendicular to bedding; measurements at 6 localities with facies perpendicular to flow and bedding. After Davies and Walker (1974) and Walker in Harms et al. (1975).



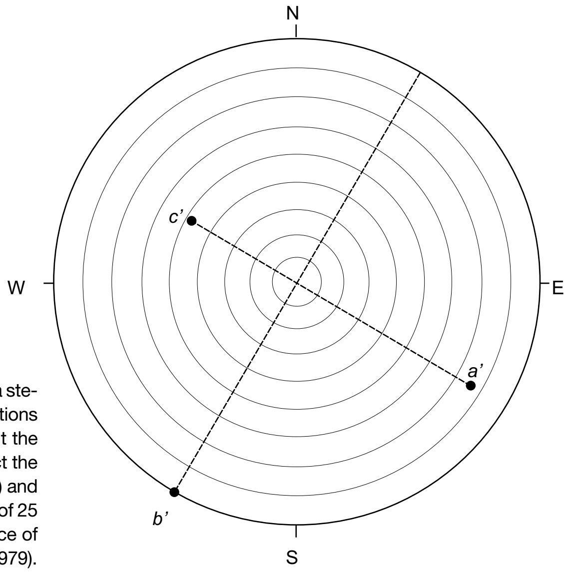


Figure 7.14 The three-dimensional disposition of rock particles shown on a stereonet projection of a lower hemisphere. In the left hand diagram the elongations of the a, b and c axes of a clast centred on the plane of the projection cut the hemisphere at a' and c' and at the periphery b'. It is conventional to project the axial data onto the equatorial plane of the hemisphere (right hand diagram) and in particular to plot and contour the density of the projections of the a axes of 25 or more clasts from each locality in order to reveal the presence or absence of ordered fabrics and preferred orientations. Modified after Derbyshire et al. (1979).

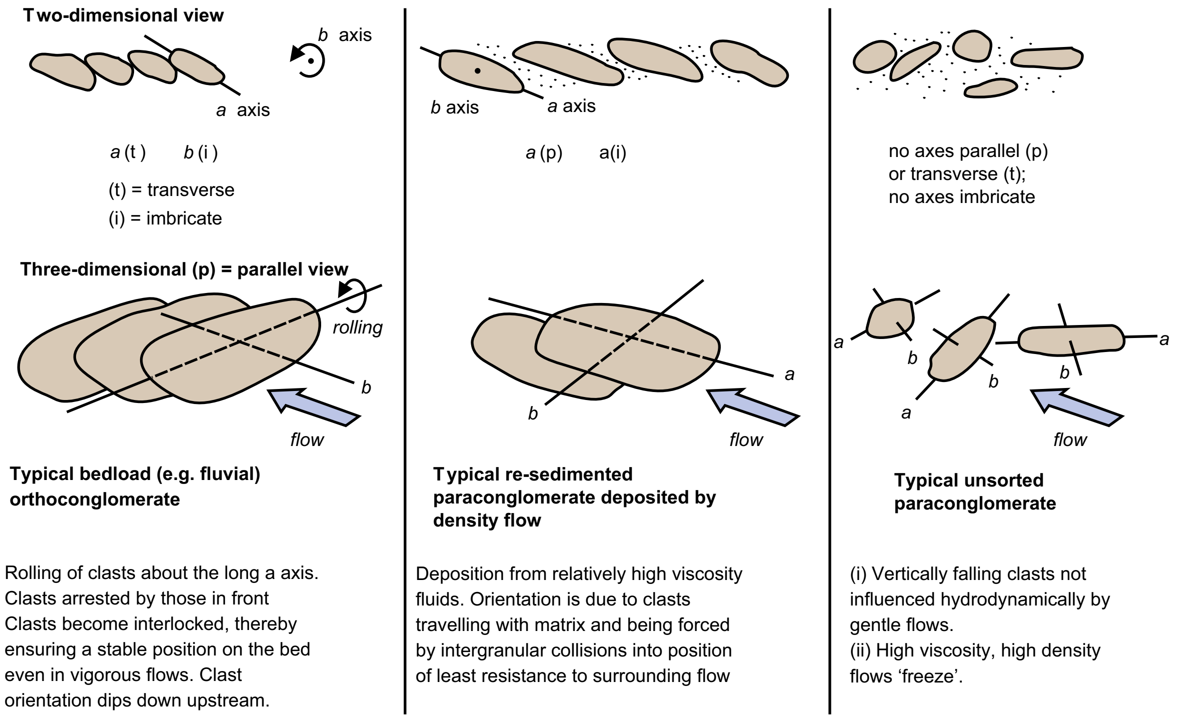


Figure 7.15 The nature and processes of origin of imbricated disc- and blade-shaped clasts.

position is when the forces of removal (lift and drag) are at a minimum. This occurs when the plane of flattening of the clast is tilted downwards at a small angle into the current. The drag is then minimized, the lift forces may even be negative, and the contact points of the clast lie to the side and well forward of its centre of gravity.

Of the two variants of clast orientation, the one with the long a-axis transverse to flow has been ascribed, on the basis of experiments, to the rolling of clasts on the sediment surface. In view of its association with turbidites and density-flow deposits, the fabric with the long a-axis parallel to dip has been ascribed to flows that maintain large clasts above the bed, possibly as a result of intergranular collision (grain flow), up to the point of deposition. Clasts slide due to the shearing effects of the flow and their long axes become aligned parallel to the flow.

Well rounded and well sorted clasts in discrete but extensive beds may be formed by high-energy wave action. Where present on the upper and middle parts of the beach, imbrication commonly dips seawards (Fig. 7.16).

Fabrics in paraconglomerates are difficult to interpret, although they have been used in attempts to differentiate between glacially emplaced tills and the products of mass flows, and between different types of till, e.g. subglacial lodgement versus supraglacial flow till (§7.3.8; Figs. 7.17, 7.18). Although many tills show a preferred, clast long-axis orientation parallel to ice flow, the pattern is seldom well defined and few confident generalizations are possible. There will always be ambiguity in cases such as flow tills where later mass flow has reorganized the fabric of the original till.

7.4.5 Stratification and cross stratification

Detection of stratification in rudites, whether it be bedding or cross bedding, is often difficult, but it can sometimes be picked out by slight changes of colour, grain size, sorting and fabric in strata that are otherwise massive or crudely bedded (Figs. 2.1, 7.11, 7.16, 7.19, 7.20, 7.21). In such situations it may be important to measure and report bed and set thicknesses, but this may be difficult, and a rather

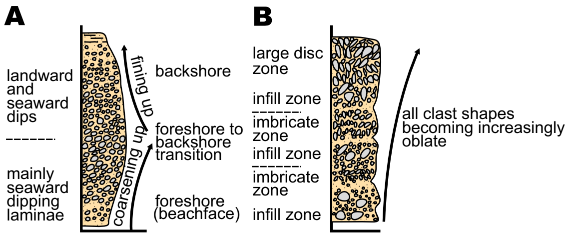


Figure 7.16 Examples of some systematic vertical changes in gravel fabric arising from the progradation of pebbly beaches. Note the patterns of change involve not only fabric but also grain size, sorting and grain shape. The changes of grain size typically occur across several beds because of gradually changing processes, and define coarsening-upwards and fining-upwards units. Modified after Nemec and Steel (1984), based on Bluck (1967) and Maejima (1982).

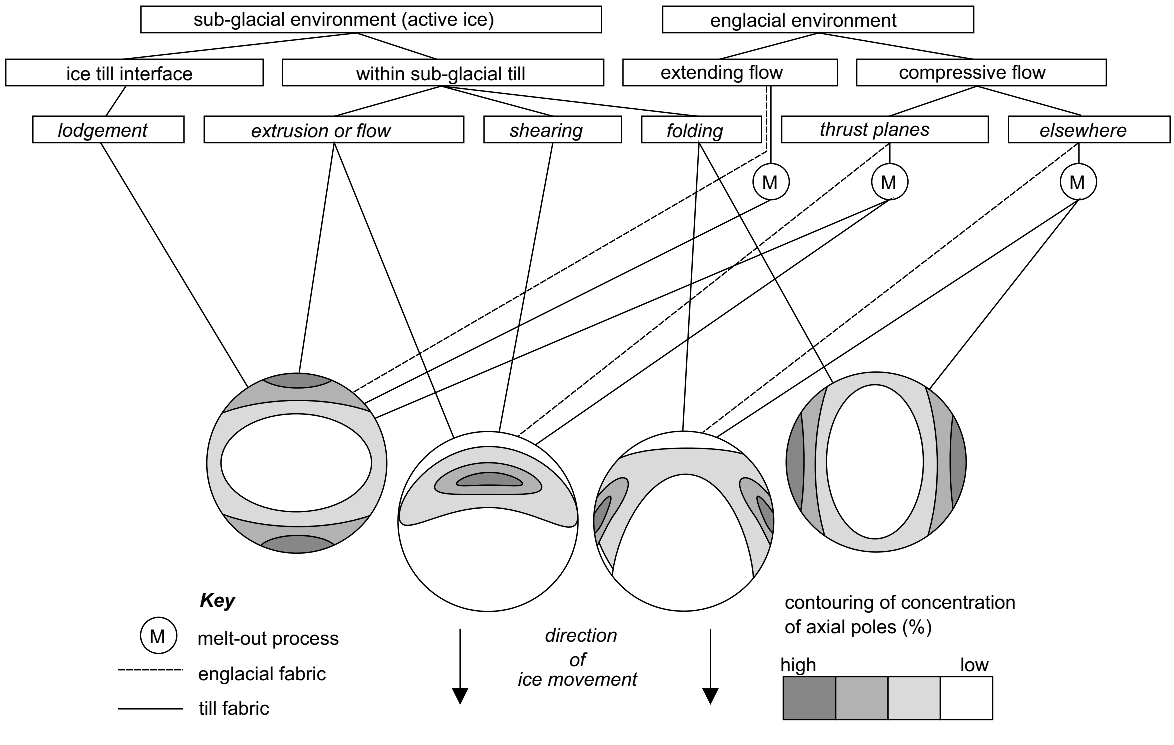


Figure 7.17 Possible relationships between fabric-forming processes and hypothetical fabric types in lodgement and melt-out tills as depicted on contoured plots on a stereonet. Modified after Derbyshire et al. (1979).

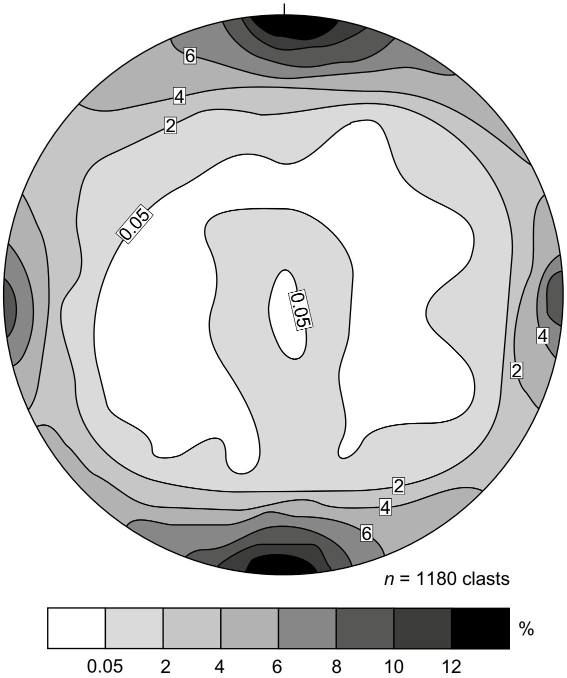


Figure 7.18 The plotting and contouring of a-axis fabric data (in the manner described in Fig. 7.14) for a large sample of clasts in a glacial till. Describe the pattern of organization of the till fabric; suggest the direction of transport of the clasts; suggest the type of conglomerate present.

Figure 7.19 Examples of stratification in rudites. A) Interbedding of conglomerate beds with different grain size. Coarsest units below measure have matrix support whereas finer-grained units are horizontally bedded and have some clast imbrication. Røde Ø Conglomerate, Permian, east Greenland. B) Crude horizontal stratification in rudite composed of reworked clasts of gypsum. The clasts were derived from a nearby exposed salt glacier (a namikir) and were reworked by fluvial flash-flood processes. Reworked clasts of gypsum are only likely to be preserved in arid or semi-arid climate settings. Moenkopi Formation, Triassic, Utah, USA. C) Horizontal stratified conglomerate with imbricated clasts that indicate transport from right to left Røde Ø Conglomerate (as in photo A). Permian, east Greenland. D) Low-angle inclined pebble foresets within a matrix-poor conglomerate. Krone Member of Etjo Formation, Cretaceous, northwest Namibia. E) Large scale, moderately inclined cross-stratified set. Modern fluvial deposits, southern Iceland. Width of view is 12 m. F) Cross-stratified conglomerate set with foresets alternately composed of sand and gravel, possibly indicating pulsed flow conditions. Hawksmoor Sandstone Formation, Triassic, Staffordshire, England. G) Tabular cross-bedded set of fine gravel between beds of coarser, clast-supported gravel. The dipping foresets indicate flow and beform migration to the right. Shainshand Formation, Cretaceous, Mongolia. H) Lenticular cross-bedded gravels. Holocene, Alberta, Canada.

Figure 7.19 *Continued*

Figure 7.20 Large scale, gravel-dominated foresets representing the deposits of a fan-delta. Modern, Skeiðarársandur, southern Iceland.

subjective estimate of thickness may be all that is possible. Bed contacts may be described as "gradational", where one bed merges with its neighbour, whereas other contacts between fairly distinct beds may be described as "amalgamated' where the adjacent beds have a similar matrix. Sharp bed contacts, commonly marked by changes in grain size, may coincide with an erosion surface (Fig. 7.22), for example at the base of a channel body. An extreme example is a basal conglomerate above an unconformity.

Multiple, distinct beds with sharp contacts are usually products of successive discrete depositional events. A basal conglomerate above an unconformity commonly marks the onset of deposition after a period of tectonic uplift and erosion. Gravel above an erosion surface in a conformable succession denotes a sharp increase in energy as when a river channel switches or migrates to an area of a river plain that had previously been abandoned. Gradational contacts suggest fluctuating and pulsating flow strength.

Amalgamated contacts between distinct beds may signify that two episodes of sedimentation were closely spaced in time, and that the energy available for the second episode partly reworked the particles of intervening beds before they achieved significant coherence. Massive and crude bedding may have involved rapidly fluctuating flow strength with high sediment concentration, whereby "freezing" of the load takes place, and the products of individual depositional events become hard to distinguish.

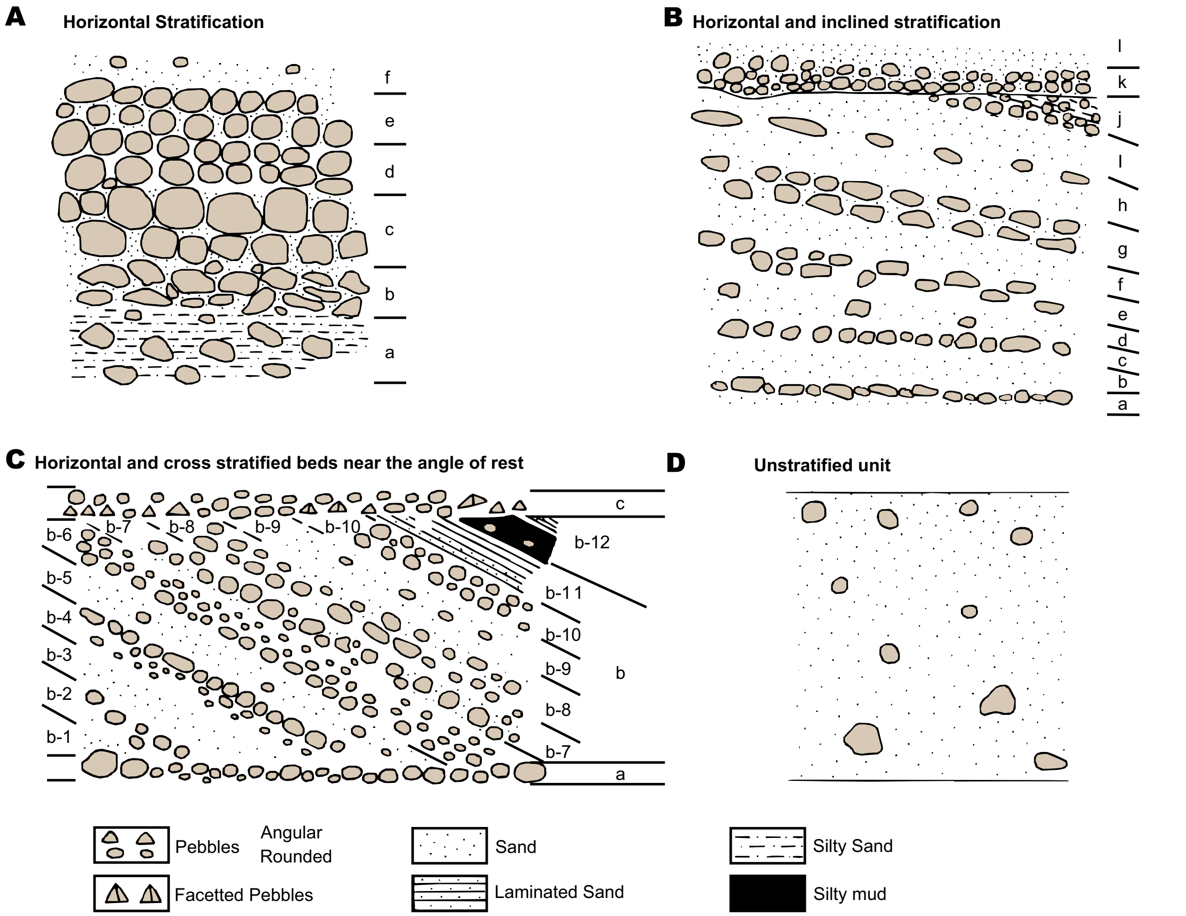


Figure 7.21 Stratification in rudites. A) Horizontal stratification with welded contacts. B) Horizontal stratification and inclined stratification. C) Horizontal and cross-stratified units near the angle of rest. D) Non-stratified unit.

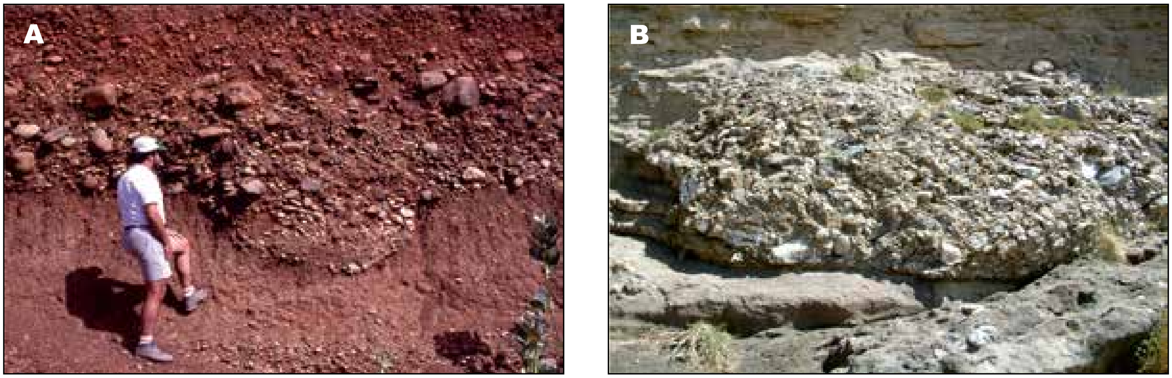


Figure 7.22 Examples of channels and scours filled with rudites. A) Fluvial channel, Pleistocene, Gregory Rift, Kenya. B) Submarine channel filled with a boulder-rich debris-flow deposit (~3m thick). Miocene, Tabernas, southeast Spain. C) Small conglomerate-filled channel of fluvial origin in otherwise fine-grained floodplain mudstone deposits. Bayanshiree Formation, Cretaceous, Mongolia. D) A lower unit of small- to medium-pebble conglomerate with lenticular shape overlain by finer-grained sediment. Above is a unit of cobble and boulder conglomerate showing imbrication in a channel unit. Pleistocene, Gregory Rift, Kenya.

Figure 7.22 *Continued*

In investigating rudites, pay particular attention to conglomerate beds that are only one clast thick. In many cases they are winnowed "lags" developed when a strong, erosive current winnowed a gravelly sand or shell bed and took the sand grains and hydrodynamically unstable pebble discs and shells into saltation or suspension. Pebbles and larger shells remained or "lagged" behind and were concentrated as a thin layer, which may have armoured the surface. A special case of lag conglomerate is that associated with channel migration (see §4.4.3, Figs. 4.24, 4.25). In contrast, a layer composed of ventifacts, the majority of which are the right way up (i.e. faceted surfaces uppermost) and therefore probably *in situ*, would record a setting where strong winds have winnowed away surface sand, and blasted and faceted the remaining pebbles, perhaps turning over a few in the process.

Once the likely attitude of the depositional horizontal has been established, various kinds of oblique or cross bedding may be identified (Fig. 7.19). Alternatively, the early identification of cross bedding may help to establish the attitude of the original horizontal. Indistinct low-angle bedding (e.g. Fig. 7.19D) has been noted in beach and fluvial gravels, as well as in deep-water conglomerates (see §7.5).

Cross bedding, with foresets at angles of 15–25°, is more common in conglomerates than in breccias (Fig. 7.19). Commonly, cross-bed sets in gravel are 1–2m thick, but isolated tabular and trough sets in excess of 10m thick have been reported. Compared with cross beds in sands, the sets in gravel, particularly of tabular type, are more commonly single and isolated. Single trough sets are known to fill erosion hollows, but many troughs form cosets. Very large isolated tabular "Gilbert"-type cross sets may also be developed in gravel (§6.2.9; Fig. 7.20). Internally, foresets may be indistinct or spectacularly rhythmic and normally graded with coarse clasts at the base of each foreset layer. Short axes of clasts tend to be oriented perpendicular to the foresets, and flattened surfaces therefore dip down current. Sandy foresets may also occur in dominantly gravelly sets, but mud-draped gravel foresets indicating temporary quiescence are rare. Reactivation surfaces (see §6.2.6) are commonly observed in larger sets.

The origin of cross beds in gravels is not always clear. By analogy with sand, the lee faces of bedforms are the most likely sites for foreset avalanching and grainfall. Echo-sounding in rivers in flood has revealed crescentic underwater dunes composed of fine-grained gravel, and their three-dimensional shape is now relatively well understood. Their migration and down-stream climb gives rise to cosets of cross beds. Large single sets of cross beds have been observed in lake deltas and in ponds in abandoned river channels. Longitudinal and diagonal bars in braided rivers are composed mainly of flat, sub-horizontal bedding, but slip faces at the downstream ends of bars may result in tabular sets of cross beds of limited extent (Fig. 7.23). Single troughs, many metres deep and filled with crude cross strata and gravelly cosets have been related to "catastrophic" discharges, associated with, for example, the breaking of natural dams during glacial melts. Some very large catastrophic flood episodes are known to have produced large repetitive bedforms with associated internal cross-bedding.

Flat sub-horizontal bedding in gravel is known from many environments. For example, it forms at high discharges and relatively shallow flow depths through vertical aggradation of the tops of longitudinal bars in braided rivers (Fig. 7.23).

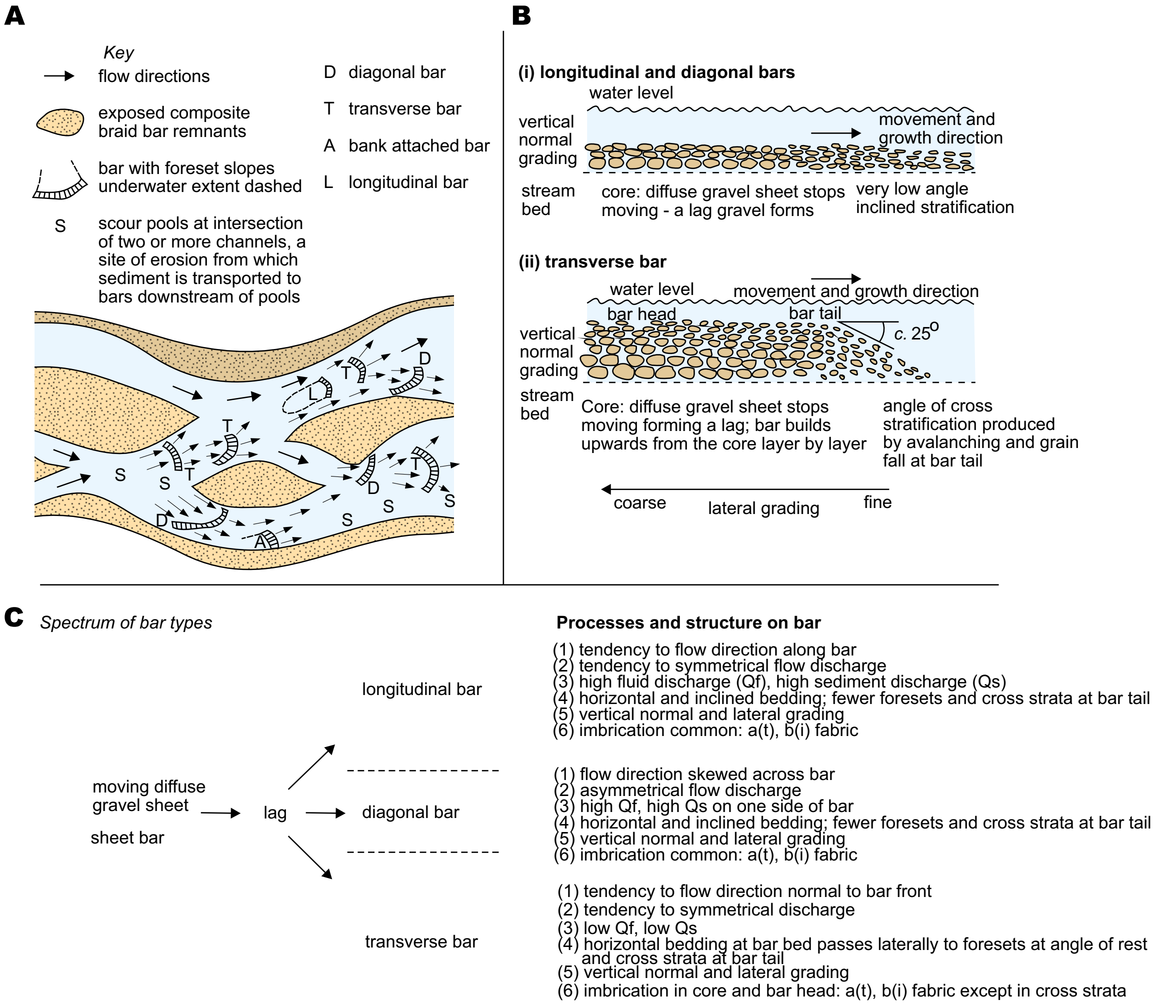


Figure 7.23 The distribution of flat-bedded, inclined and cross-bedded strata in relation to processes creating fluvial bars in braided gravelly streams. A) Gravel bar types in the Kicking Horse River of British Columbia, Canada. B) Cross-sections showing the association of stratification types and processes of gravel accumulation in longitudinal, diagonal and transverse bars. C) Model to account for the origin of sheet, longitudinal, diagonal and transverse bars. Modified after Smith (1974), Hein and Walker (1977). a and b refer to long and intermediate axes, respectively. t = transverse to flow; i = imbricate."

Deposits associated with flat and cross-bedded rudites and their significance

In successions dominated by rudites, it is important to record not only features of the rudites themselves but also those of interbedded units of sandstone, siltstone or mudstone, since these reflect important changes in the sedimentation regime and may provide the best indicators of the overall depositional environment. Thin units of cross-bedded or cross-laminated fine-grained sandstones are common as interbeds, and are typically the deposits of lower discharges and/or lower current velocities. However, the boundary shear stress needed to roll pebbles commonly coexists with levels of turbulence that are able to maintain sand grains in suspension (Fig. 7.8). This relationship, which is not valid for very high sediment concentrations or high transport rates, predicts that interbedded sand and gravel may occur without radical fluctuations in flow strength. Interbedded siltstones and mudstones in otherwise gravel-dominated successions may, however, represent periods of protracted settling from suspension and imply periods of quieter conditions. The presence of ripple marks, mud drapes, mud cracks, body fossils and particularly, established

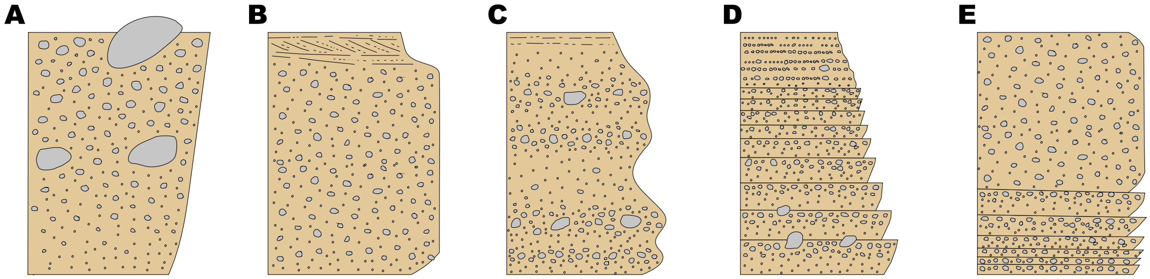


Figure 7.24 Variety of deposits that can be produced by traction carpet sedimentation. A) Thick-bedded and inversely graded. B) Thick-bedded and massive with only the basal part inversely graded. C) Diffusely stratified. D) Thinning-upwards stratified. E) Thickening-upwards stratified, depending on the duration and character of the traction carpet and the overall flow characteristics. The thicknesses are not to scale and can be very variable. Bed boundaries in these types of deposits are not necessarily easy to identify. After Sohn (1997).

populations of trace fossils (§9.4) provides more specific evidence of the environment in which the energy fluctuations took place.

7.4.6 Bed thickness and grain-size relationships

In thick rudite-dominated successions it can be useful to measure bed thickness and maximum particle size (commonly expressed as the arithmetic mean of the ten largest particles) within each bed. However, bed boundaries are not always easy to identify (Fig. 7.24) and, in many such successions, boundaries may be erosive. This can lead to difficulties in estimating bed thickness due to either amalgamation of beds (and thereby over-estimates) or removal of part of the bed (under-estimates).

Plots of bed thickness versus maximum particle size, in some cases, reveal a clear linear relationship (Fig. 7.25) whereas, in other cases, an apparently random scatter occurs. Cases of doubt can sometimes be resolved by calculating a correlation coefficient. Where a linear correlation is apparent it may be inferred that the beds were the products of discrete depositional events. The thickness of the bed is thought to be a crude proxy for the thickness of the flow, whereas the maximum particle size relates to the competence of the flow.

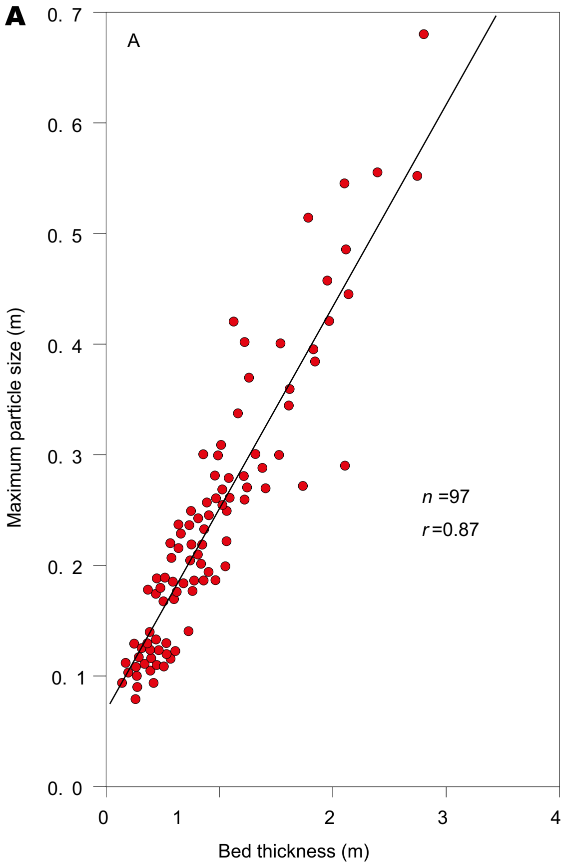
For carefully collected data, the correlation coefficient for a group of beds can give an approximate indication of the consistency of the physical behaviour of depositional events. It may also be possible to distinguish between deposition from predominantly cohesive (Bingham) flows from predominantly cohesionless (Newtonian) flows (Fig. 7.25 cf. Fig. 3.21), which are types of debris flow that occur in many environmental settings (see §3.7).

7.5 Processes of formation of mass properties and structures

When studied on a larger scale, many rudite successions exhibit lateral and vertical changes in their style of sedimentation (Figs. 7.26, 7.27). It is therefore important to assess how the parameters discussed in §7.4 vary and how they are combined in different ways in different parts of a succession. Some common associations and trends are recognized but further field observations and data from both modern and ancient successions are important to consolidate our understanding. Where the wider context of a particular succession is unclear, interpretation is best limited initially to inferring depositional process based on internal features so the rudites.

Cut banks of rivers and gravel pits commonly reveal short vertical successions of pebbly strata. To work most effectively on these and on extensive outcrops of ancient rudites, it may be useful to create a photomosaic from a set of overlapping photographs taken at fixed distances from the outcrop. This makes it easier to trace lateral changes in the distribution of features over many tens of metres. It can also be instructive to map the distribution of the features described in §7.4 on a horizontal surface in a suitable present-day environment, e.g. braided rivers at low water, beaches at low tide, debris flow units on abandoned parts of alluvial fans. Where possible, sample and observe from the proximal to the distal ends of extensive environments, as along a river, and axial and lateral parts of radial systems, such as a fan.

Large-scale coarsening- or fining-upward trends within ancient alluvial-fan successions can be interpreted in terms of temporal and spatial changes in energy regime that may, in turn, be related to changes in tectonic activity, relief and/



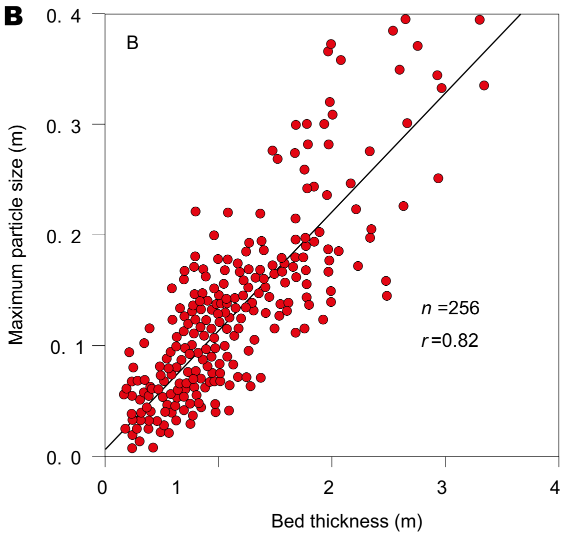


Figure 7.25 Examples demonstrating a relationship between maximum particle size and bed thickness. A) A subaerial fan deposit, Stornoway Group, Permo-Triassic, northwest Scotland. B) Submarine fan delta, Ksiaz Formation, Devonian, southwest Poland. Part (A) shows a higher correlation coefficient, and the correlation line intersects the maximum particle size axis, suggesting deposition by rather consistent, cohesive debris flows. Part (B) shows a lower correlation coefficient and the correlation line passing through the origin, suggesting slightly more variable, cohesionless behaviour (n = number of points in sample set; r = correlation coefficient). After Nemec and Steel (1984).

or climate. An overall increase in the size and strength of depositional events within alluvial environments, expressed as a coarsening-upward succession, has been associated with fan progradation. Independent evidence may relate such a trend to increasing tectonic activity, which enhances basin-margin relief. Alternatively, fan progradation may be a consequence of an increasingly wet climate. A fining-upward succession may similarly be interpreted in terms of fan recession, and may be related to waning tectonic activity or increasing aridity.

Limestone breccias occur quite commonly, in some cases associated with reefs where they occur as reef talus slope. They typically exhibit a characteristic down-slope decrease in maximum and mean clast size away from the reef crest, and a corresponding increasing in matrix content. In a vertical succession, an upward increase in clast size probably records deposition in successively higher positions on the talus slope and may signify a progradation of the reef fringe. Early carbonate lithification means that any post-depositional disturbance is likely to lead to *in situ* brecciation (Fig. 7.28) rather than slumping and folding (§9.4). Where limestones are interbedded with thick units of evaporites, spectacular limestone breccias may result from collapse following evaporite dissolution at outcrop or shallow burial (Fig. 7.29).

Volcaniclastic successions show varied combinations of structures and mass properties which are not always easy to interpret due to their complex and variable processes of origin (see §7.3.9), especially as volcaniclastic materials are commonly resedimented soon after initial deposition (Fig. 7.30). At a gross scale, pyroclastic flows, surges and falls all tend to thin and fine with increasing distance from the eruptive centre and thus exhibit lateral grading. Airfall deposits are strongly influenced by prevailing wind directions and their distribution may indicate palaeowind direction. Pyroclastic flow and surge deposits are controlled by pre-existing topography and may form thick beds in confined, narrow valleys, or may form thin sheets when spreading over widespread, low-relief plains. Most

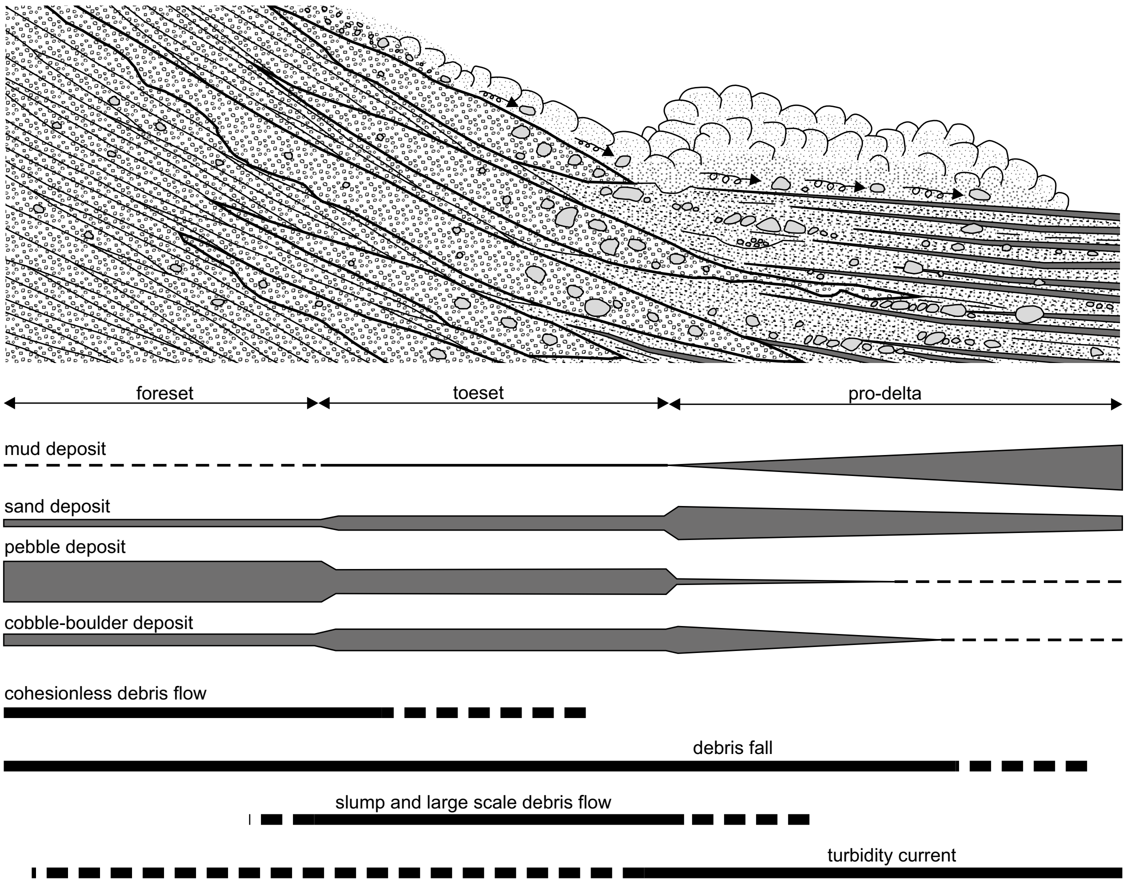


Figure 7.26 Schematic diagram showing the gross-scale geometry, internal architecture and textural variability of typical fan-delta deposits. The dominance of various depositional processes is indicated. Note the erosive bounding surfaces between the various depositional units within the large-scale prograding foresets, the sharp contrast in textural characteristics between the foreset-toeset region and the pro-delta, and the textural bimodality of the pro-delta deposits. Modified after Sohn et al. (1997).

importantly, volcaniclastic successions are influenced by the nature of the eruptive event. For example, a prolonged eruption that progressively increases in intensity might produce pyroclastic deposits that show vertical grading trends indicative of waxing and waning phases.

It may be useful to order descriptions, analyses and interpretations of rudites with the help of Figure 7.13 and to further refine the description and interpretation of observed relationships as experience increases.

7.6 Uses of structures

The obvious major use of structures in rudites is in identifying processes and environments of deposition. The interpretation of the environmental origin of a rudite, for example whether a very clay-rich conglomerate is a till of glacial origin, the product of a subaqueous or subaerial debris flow or an agglomerate, or lapilli-ash of volcanic origin, will depend on detailed description of its composition, the shape of its clasts, its relation to surrounding beds and consideration of its overall context.

Exploration geologists seeking oil, gas, water or metalliferous deposits may be attracted to thick successions of orthoconglomerates where initial porosity and permeability are likely to be high. Paraconglomerates, by contrast, are more likely to have low initial porosity and permeability due to the dominant matrix. Initial porosity and permeability, however, may be markedly reduced by

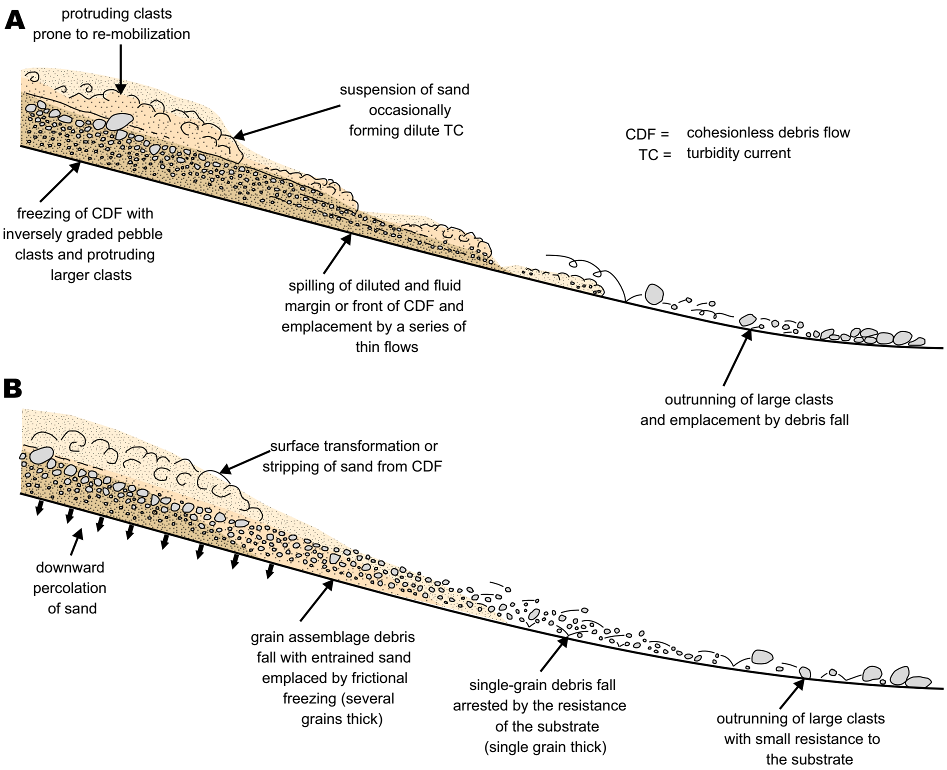


Figure 7.27 Schematic illustration of debris flow evolution. A) A cohesionless debris flow produces a massive, inversely graded bed. Spilling of diluted material over the front or sides of the flow may produce a sandy suspension from which thinner sandy layers may be deposited. Larger clasts 'float' to the top and front of the flow and may be further transported downslope because of their greater momentum and small resistance from the bed, forming debris fall deposits. B) Where a cohesionless debris flow experiences long-lived downslope transport, interstitial sand is progressively removed via surface transformation and percolation. The flow then transforms downslope into a grain-assemblage debris fall and then into a single-grain debris fall, producing gravel sheets. Coarser clasts are transported further downslope, resulting in lateral grading. Modified after Sohn et al. (1997).

Figure 7.28 Breccia of angular quartzite blocks at the base of a large slide block of thinly bedded quartzite, many hundreds of square metres in area. The slide detachment surface is just below the base of the photograph. The upper unit of quartzite, around 2m thick, remained intact during the sliding. Morsænesø Formation, Late Proterozoic, north Greenland.

Figure 7.29 Breccia of angular dolomite blocks chaotically dispersed in a clay-rich dolomitic matrix, formed as a result of gravitational collapse of the unit due to dissolution of an underlying evaporite interval. Seaham Residue, Upper Permian, County Durham, England.

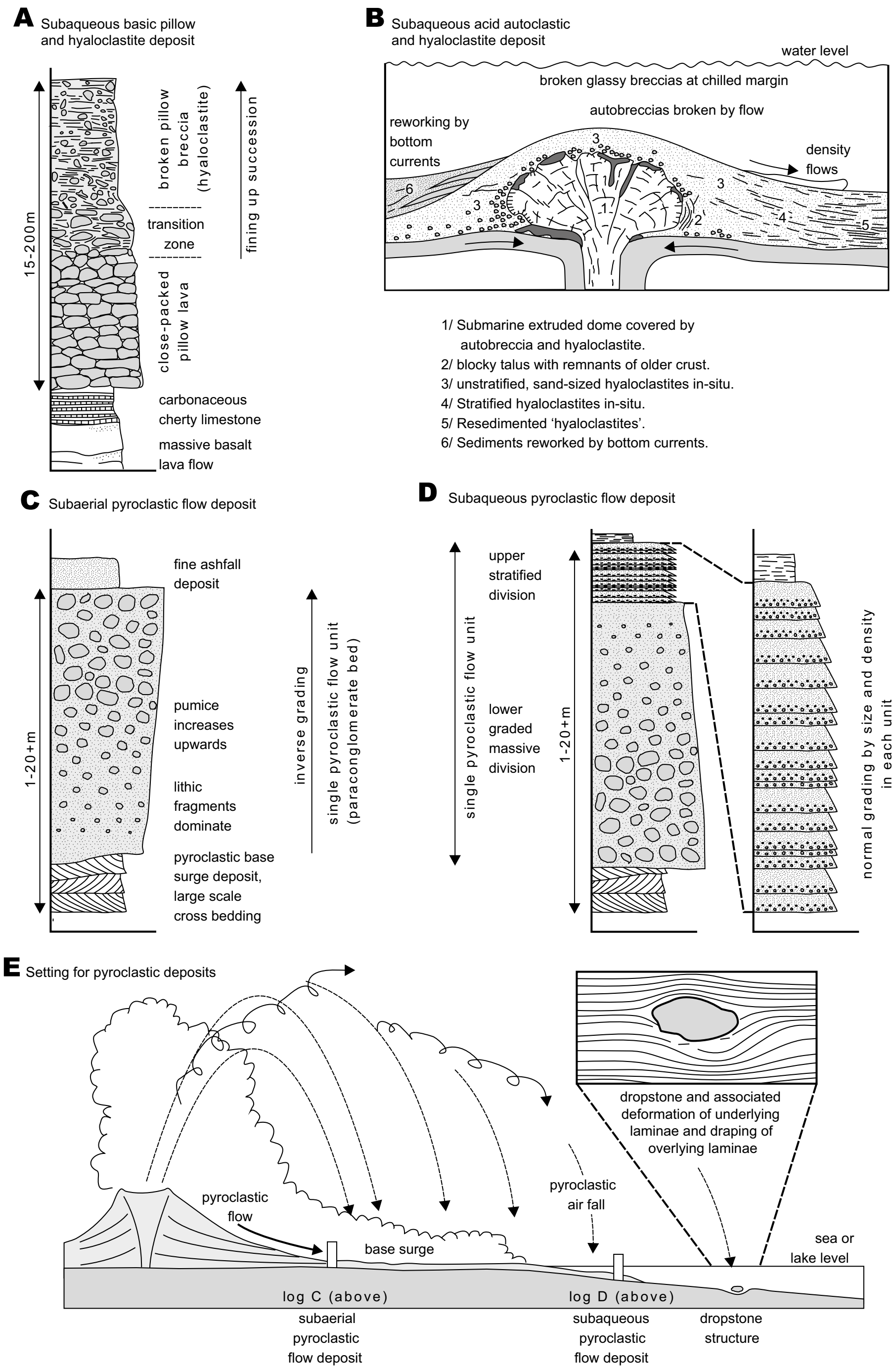


Figure 7.30 Volcaniclastic rudite successions and processes of formation. A) Sub-aqueous flow sequence in basic magma represented by pillow breccia and hyaloclastite succession. Modified after Carlisle (1963) and Lajoie in Walker (1979). B) Sub-aqueous autoclastic and hyaloclastic flow sequences in acid rhyolitic magmas with re-sedimentation. After Lajoie in Walker (1979). C) Subaerial pyroclastic flow deposit. D) Subaqueous pyroclastic flow deposit. Parts (C) and (D) show the vertical succession of primary structures commonly present in subaerial and subaqueous pyroclastic deposits. Modified after Sparks et al. (1973), Fiske and Matsuda (1964) and Lajoie in Walker (1979). E) Model for environmental occurrence of pyroclastic deposits.

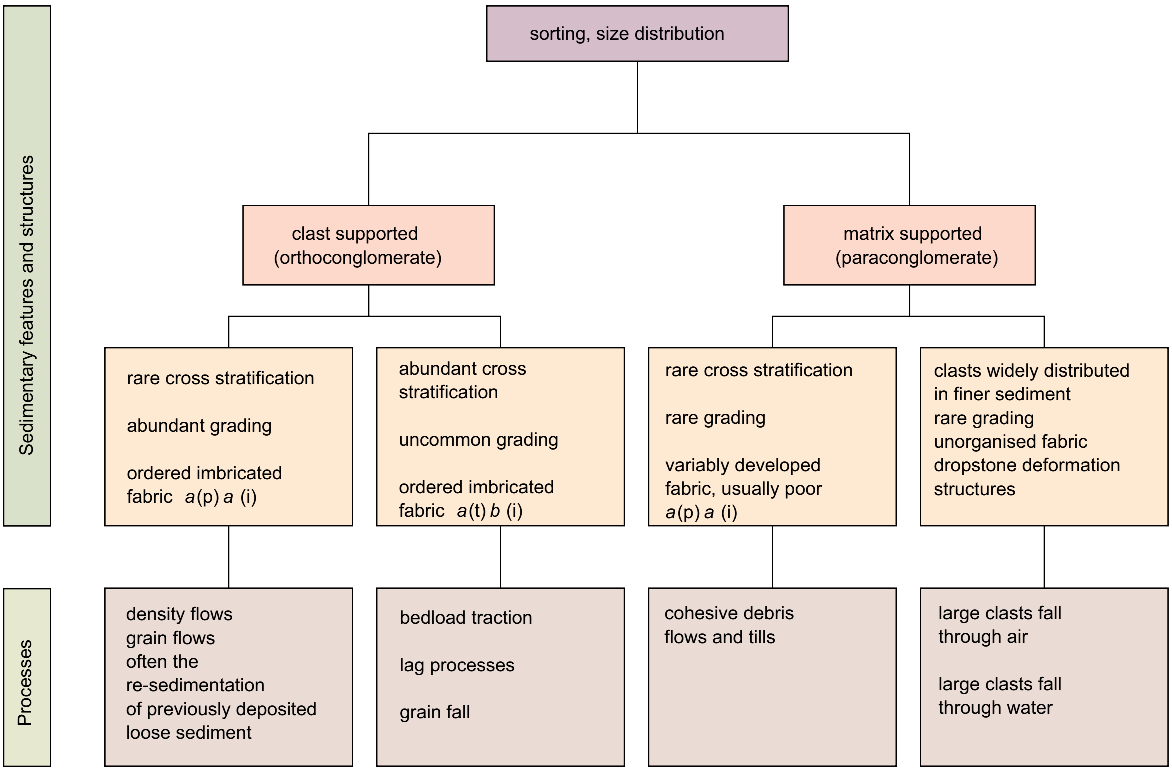


Figure 7.31 Common but not absolute associations of structures in rudites in relation to processes of origin. Consider how these structures and processes might arise in different environmental settings. Modified after Walker in Harms et al. (1975). a and b refer to long and intermediate axes, respectively. t = transverse to flow; p = parallel to flow; i = imbricate.

burial diagenesis. Autoclastic and pyroclastic breccias are commonly the host rocks for primary sulphide mineralization and zeolite formation. Studies of palaeocurrents may help in predicting the directions in which an interval might become thinner and finer and might also predict the direction in which porosity and permeability diminish. Cross bedding, clast imbrication and the orientations of channel margins can all contribute to such palaeocurrent analysis.

For the structural geologist, normal grading and cross stratification may help to determine "way-up" in rocks that are steeply dipping, though caution should be exercised given the relatively common occurrence of both inverse grading and planar-tabular cross bedding. Additionally, well rounded, mature conglomerates, with consistent and predictable sphericity and roundness values may be useful indicators of tectonic strain. Pebbles may become stretched and flattened, and comparison of the non-deformed and deformed pebbles may allow a measure of strain to be estimated.

Study techniques

Field experience

Present-day environments

Your field programme should include the observation and recording of some of the following processes in their natural settings:

River and stream courses Formation of frameworks, filling of frameworks, imbrication, development of lag deposits.

Beaches Distribution and orientation of differently shaped clasts; the formation of frameworks and their fills.

Debris flows, solifluction (soil creep) and debris avalanches Matrix-supported to clast-supported gravels; various types of grading.

Coastal or desert sand dunes in hot or cold climatic settings Deflation, the formation of ventifacts and pebblearmoured surfaces.

Ancient environments

The ancient record provides well exposed sequences showing a range of features and structures that extend the experience derivable from the present record. Gravel pits in Quaternary and older deposits are commonly available and might be worthy of investigation.

Laboratory experience

Films of continuing volcanic activity are many and there are numerous after-the-event shots of the resultant products, e.g. dropstones in ashfall deposits or imbrication in ashflow deposits. Experiments on the angle of initial slip and angle of rest of angular and rounded gravels are possible. The effects of introducing pebbles into otherwise sandy regimes are easily investigated in a flume. Descriptions of texture and fabric in the main types of orthoconglomerate and paraconglomerate are easily made from hand specimens in the laboratory and are a basic skill.

Recommended references

Brenchley, P. J. & B. J. P. Williams (eds) 1985. *Sedimentology: recent developments and applied aspects*. Good summary paper by Suthren on volcanic sediments.

Carling P. A. 1996. *Morphology, sedimentology and palaeo-hydraulic significance of large gravel dunes, Altai Mountains, Siberia*. A good account of large bedforms in gravels.

Cas, R. A. F. & J. V. Wright 1987. *Volcanic successions: modern and ancient*. Good examples of coarse-grain volcaniclastic successions.

Koster, E. H. & R. J. Steel (eds) 1984. *Sedimentology of gravels and conglomerates*. Probably the most wide-ranging descriptions of gravels in different depositional settings.

Nemec, W. & R. J. Steel 1984. *Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits*. An excellent example of the benefits of close study of depositional textures.

Steel, R. J. & D. B. Thompson 1983. *Structures and textures in Triassic braided stream conglomerates ("Bunter" Pebble Beds) in the Sherwood Sandstone Group, north Staffordshire, England*. A careful account of the use of textures in interpreting ancient river deposits

CHAPTER 8

Depositional structures of chemical and biological origin

8.1 Introduction

Much of the material weathered and eroded from land areas is transported to the seas as ions in solution. From geochemical studies, it is known that the composition of sea water has remained fairly constant throughout a large part of geological time and it follows that ions must have been taken out of solution by the precipitation of new minerals. This precipitation can be inorganic or it can be aided by, or due entirely, to organic agencies.

The most abundant minerals precipitated from sea water are aragonite and calcite, and most of this precipitation results from microbes, plants or animals. Although inorganic precipitation of carbonates is possible, most inorganic precipitates are evaporite minerals, the most abundant of which are gypsum, anhydrite and halite. In non-marine settings, such as saline lakes, the brine chemistry may be different from that of sea water, and different assemblages of evaporite minerals may form.

In this chapter, we deal first with the structures and textures produced by inorganic precipitation from bodies of saturated brine and then with structures due to organisms acting either to precipitate sediment or to bind existing particles.

8.2 Chemical precipitation

Inorganic precipitation of minerals from solution is largely confined to evaporite minerals, commonly gypsum and halite. For any mineral to be precipitated inorganically, an aqueous solution must be supersaturated with respect to that mineral's constituent ions. Irrespective of whether the water body is connected to the sea or is enclosed as a lake, conditions of net evaporation must occur and this commonly requires a hot, arid setting. When supersaturation is achieved, precipitation takes place providing that other ions in the solution do not interfere with crystal growth. Nucleation can occur spontaneously anywhere within the water column or on objects already on the floor of the basin. Crystals that nucleate at the water surface may float for a while, held by surface tension, and may exceptionally form surface rafts or crusts (Fig. 8.1). Eventually they sink to the floor of the basin where most precipitation and crystal growth takes place. For well-formed crystals to develop, both free space and an interval of time are required (Figs. 8.2, 8.3).

Processes of nucleation and crystal growth can be modelled in the laboratory, for example by allowing 1000 ml of

Figure 8.1 Halite precipitation. A) Clusters of small halite crystals floating on the surface of a hypersaline lake. United Arab Emirates. B) Halite crystals forming extensive floating mats on the surface of a lake. United Arab Emirates. Photos courtesy of Stephen Lokier.

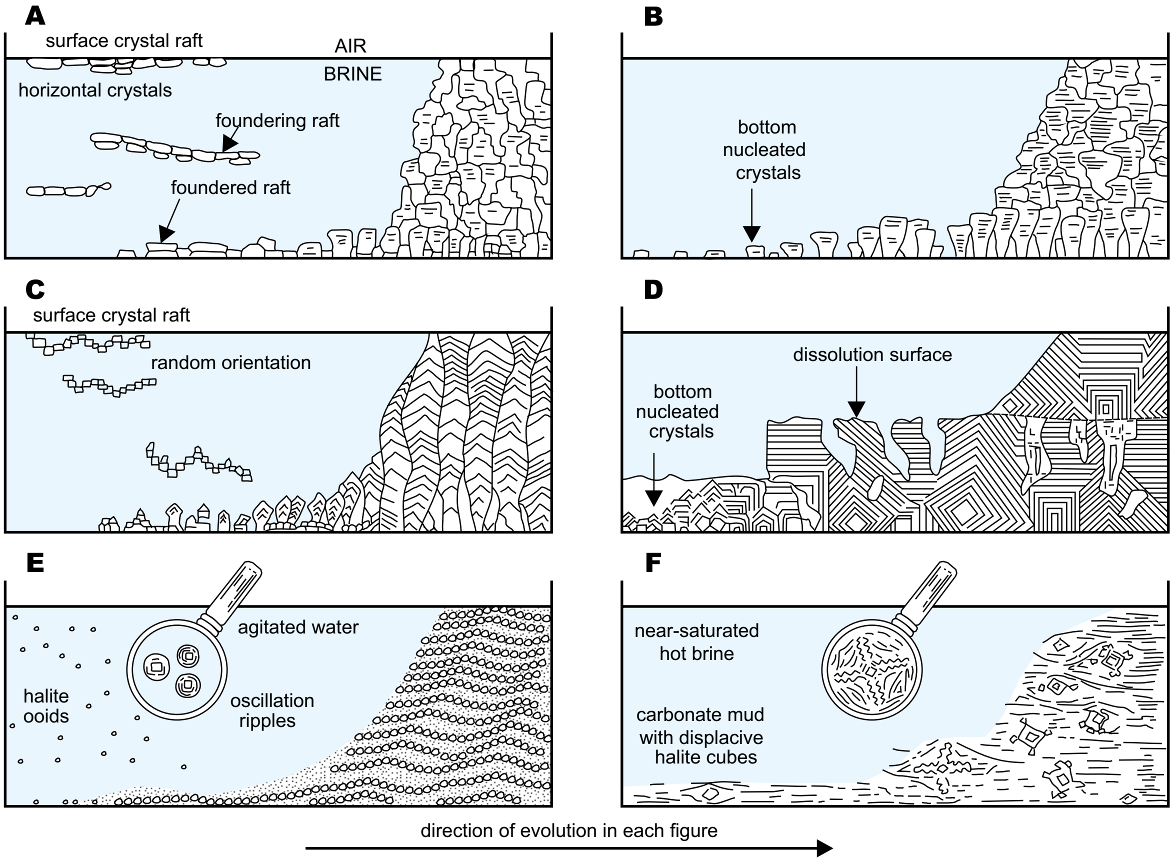


Figure 8.2 Idealized textures associated with the growth and emplacement of halite under differing conditions. Modified after Schreiber in Reading (1986), based on Arthurton (1973), Shearman (1971, 1978) and Weiler et al. (1974.)

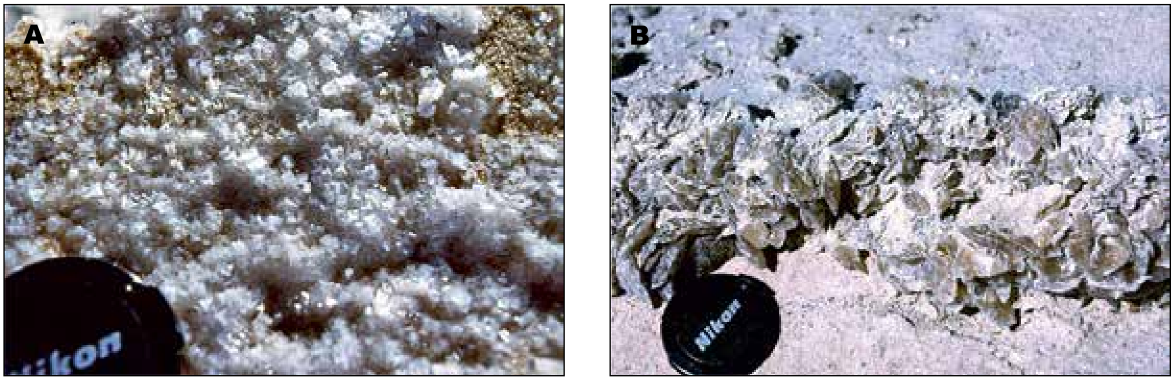


Figure 8.3 Crystal forms of evaporite minerals. A) Cubic "hopper" crystals of halite. B) Gypsum crystals. Both examples at the margin of present-day Lake Eyre, central Australia.

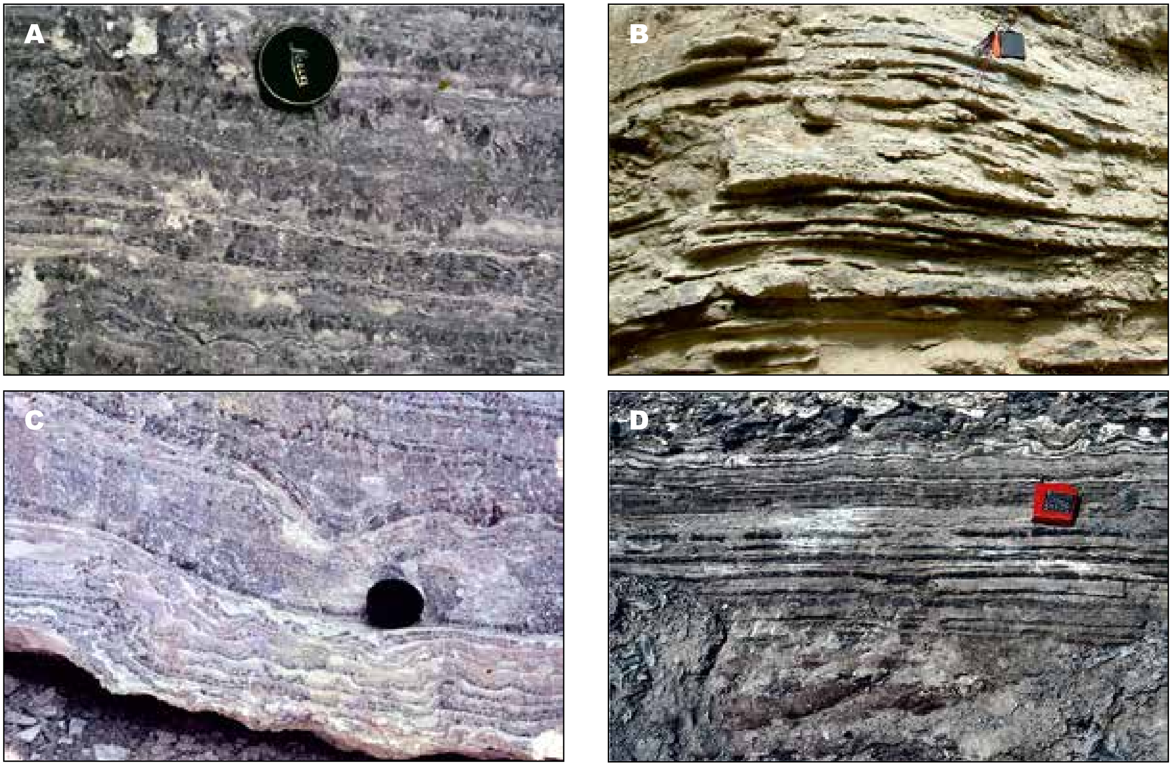


Figure 8.4 Examples of laminated and bedded evaporites. A) Thinly bedded gypsum with bedding defined by slight grain-size differences and impurities. Miocene, southern France. B) Interlaminated gypsum and mudstone, Yesares Member, Sorbas, SE Spain. C) Laminated and deformed gypsum, Miocene, southern France. D) Thinly interbedded gypsum, anhydrite and dark muds with some deformation due to evaporitic growth at the top of the profile. Poulsen Cliff Formation, Ordovician, Washington Land, northwest Greenland.

a saturated solution of sodium chloride to evaporate gradually in a suitable tank. A hand lens can be used to observe crystal growth both as they form at the water surface and after they have fallen to the bottom. Is it possible to distinguish crystals that nucleated at the water surface from those that nucleated on the floor of the tank? How do crystals continue to grow once they are on the floor? Try to monitor the temperature and rate of evaporation during the experiment. For a more elaborate experiment, try to do the same thing with about 4 litres of sea water. With the aid of some chemical analysis it may be possible to study the order of crystallization of different minerals and the changing chemistry of the remaining brine as evaporation proceeds.

8.2.1 Laminated evaporites

A common feature of many ancient evaporite-bearing successions is a fine, millimetre-scale interlamination of different mineral phases or of an evaporite mineral and organic-rich material. Where two minerals are present, these are commonly calcite ($CaCO_3$) and anhydrite ($CaSO_4$). Individual layers show great lateral continuity and may show grading, in terms of individual crystal size and/or mineral type or form (Fig. 8.4). Ungraded laminae probably record periods of settling of crystals precipitated at the water surface, possibly on a seasonal basis. In contrast, graded layers, especially those composed of randomly oriented crystals, might suggest reworking and re-sedimentation of previously precipitated crystals of various sizes (Fig. 8.5; see also §6.7). This suggests episodic high-energy events, such as storms, which stirred up crystals which then settled out as energy waned. Alternatively, grading of crystal size may reflect changes in the rate of evaporation due perhaps to seasonal or longer-term climatic variations. Laminated evaporites are commonly observed to have undergone plastic or brittle deformation. In profile, such features are evident as buckled layers or as layers that are faulted and overlapping (Fig. 8.6; see also Ch. 9). In plan view, such features may take the form of salt polygons,

Figure 8.5 Irregular and randomly oriented crystals of gypsum arranged into a normally graded bed, Paradox Formation, Pennsylvanian (Upper Carboniferous), Utah, USA. These crystals may have been reworked as clasts under the influence of a current.

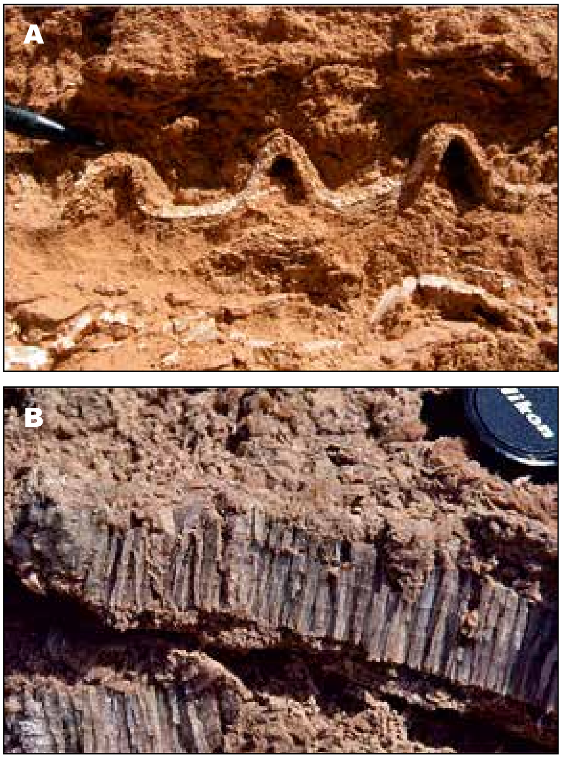


Figure 8.6 Vertical sections through structures resulting from evaporite precipitation. A) Tepee structures; folding due to the lateral expansion of gypsum layers as a result of crystal growth. Moenkopi Formation, Triassic, Utah, USA. B) Small-scale thrusting in a gypsum layer due to expansion during crustal growth. Present-day, Lake Eyre, central Australia.

salt ridges or salt curls (Fig. 8.7). These features develop where laminated evaporites have been precipitated such that they entirely cover a surface; continued precipitation induces deformation of the layer.

8.2.2 Fabrics due to vertical crystal growth

Growth of crystals on the basin floor commonly produces distinctive fabrics (Fig. 8.2A–C). Growth is most rapid parallel to certain crystallographic axes, commonly the c-axis. Crystals that precipitate within the water column fall to the bottom with different orientations. Those whose axis of preferred growth is oriented vertically will continue to grow most rapidly upwards, whereas growth of those with more inclined axes will eventually cease. The surviving, vertical crystals give rise to the tightly packed, columnar texture seen in many ancient deposits of halite and gypsum (Figs. 8.2C, 8.8A). Where crystal growth commences via nucleation at a specific site on the bed, distinctive crystal "trees" or "cones" may develop, the form of which can in some cases be preserved in the ancient record (Fig. 8.8B).

During growth, the sediment surface is made up of crystal faces upon which precipitation takes place. Records of the instantaneous position of this surface are in some cases picked out in ancient evaporites by thin layers or drapes of mud deposited mostly from suspension during events such as river floods into saline lakes or storms in lagoons. Particularly severe events may lead to partial dissolution of the minerals and to the truncation of the vertical crystal fabric. Above such surfaces, the pattern of vertical growth may be re-established (Fig. 8.2D).

8.2.3 Fabrics due to isolated crystal growth and the development of pseudomorphs

Under certain environmental conditions evaporite minerals grow slowly within host sediment either at the surface or in the shallow subsurface. This commonly leads to the

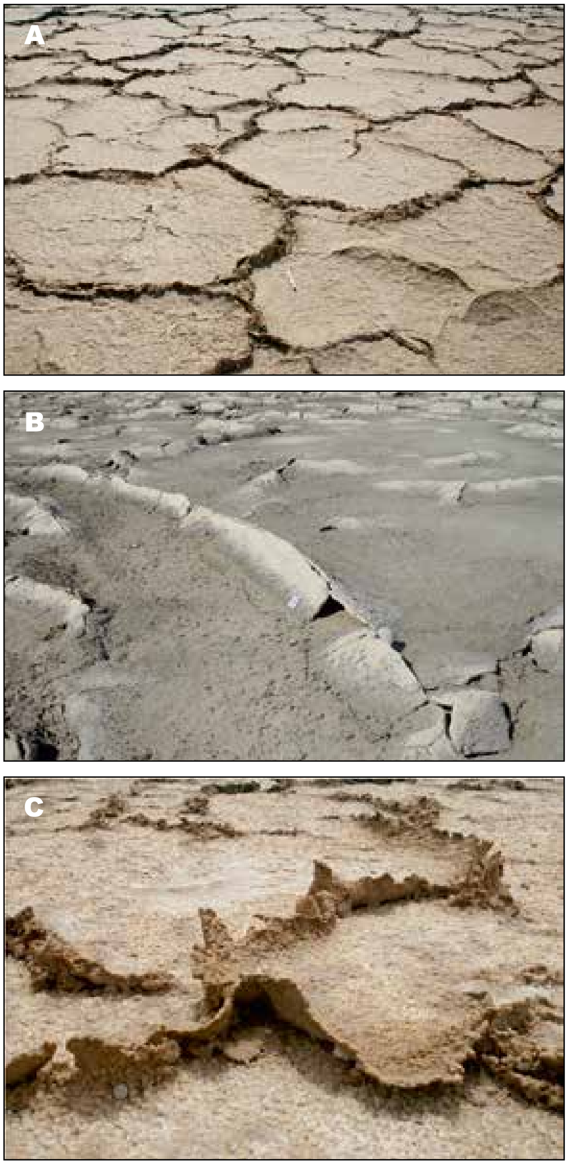


Figure 8.7 Polygonal structures resulting from evaporite precipitation. A) Polygonal pressure ridges due to isotropic expansion of a surface crust due to halite crystal growth. B) Tepee structure forming as a result of compression due to expansion of a surface layer during crystal growth. C) Close-up of a structure similar to that depicted in A) showing the curling of the thin surface layer. All three examples from the United Arab Emirates. Photos courtesy of Stephen Lokier.

Figure 8.8 Examples of twinned selenite crystals (a form of evaporitic gypsum) arranged into sub-vertical columns. A) Detail showing elongate blade-like crystal growth. B) A nucleation site from which a tree-like structure of selenite crystals has grown. Note the differential compaction of the underlying laminated mudstones. Both examples from the Yesares Member, Miocene, Sorbas, southeast Spain.

development of isolated groups of crystals that do not necessarily form a continuous layer. Where such evaporite precipitation is slow but sustained over a protracted period, the resultant crystals can to be large. One common evaporite mineral that is precipitated in this manner is halite which typically develops on an exposed sediment surface where ongoing evaporation draws moisture up from the subsurface through capillary action. Halite crystals are characterised by a distinctive cubic form up to 3cm in side length, commonly with stepped and indented

Figure 8.9 A large specimen of Desert Rose, a type of gypsum that grows in isolation within the shallow subsurface at the water table interface in arid settings. Recent, southern Namibia.

faces called “hoppers”. A second common type of isolated evaporite mineral is “desert rose” gypsum, which typically forms in the shallow subsurface, most commonly close to water table in arid settings (Fig. 8.9; see also §8.2.4). Such “roses”, which may be up to 15cm in diameter, are made up of groups of large, interlocking individual gypsum crystals.

The high solubility of evaporite minerals means that they are susceptible to dissolution when immersed in undersaturated water, for example by the flooding of supersaturated brine pools or by the influx of relatively fresh water into a lake. Even if the dissolution of evaporite minerals is total, all evidence of the minerals may not be lost. Some successions, usually of interbedded sandstone and mudstone or, less commonly, limestone and mudstone, show evidence of former evaporites, usually halite or gypsum, in the form of **pseudomorphs**. The original evaporite crystals have been replaced by sandstone or limestone but the mineralogy of the evaporite can still be deduced from the crystallographic shape of the pseudomorph. Halite pseudomorphs are cubic protrusions on the lower surfaces of sandstone or limestone beds. They occur in a variety of sizes, commonly up to about 1cm diameter side length, and they may be isolated or may

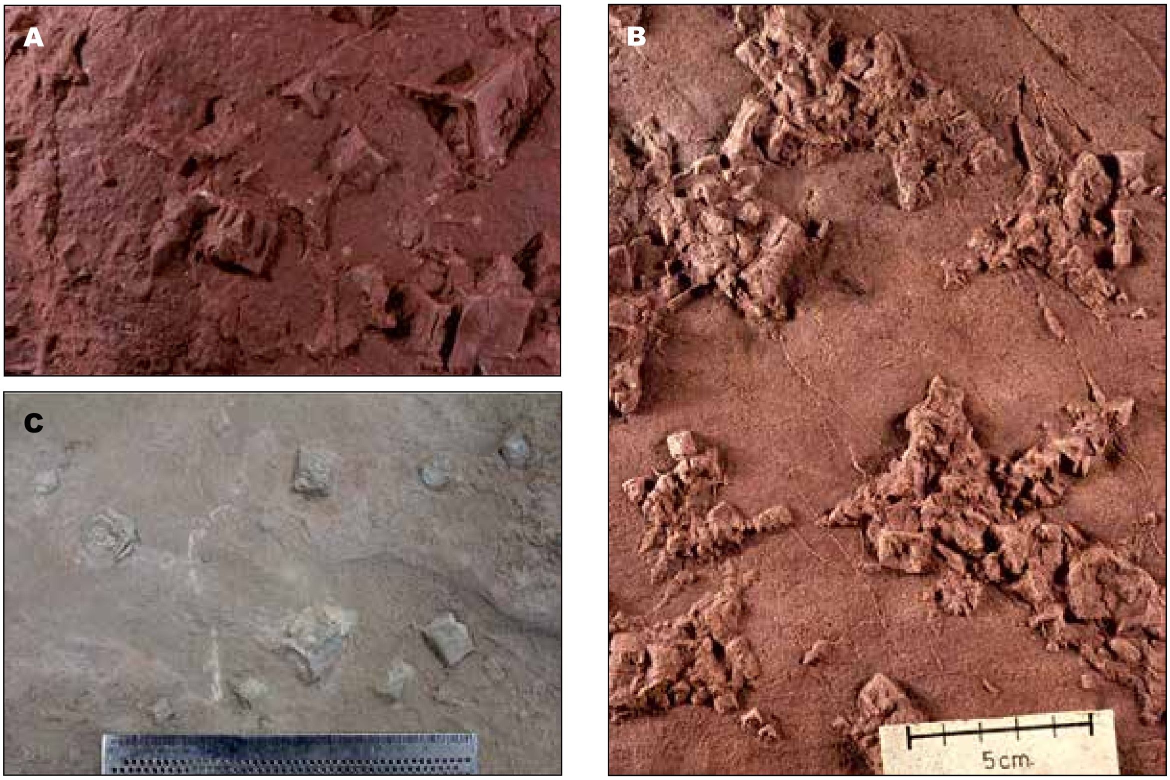


Figure 8.10 Examples of common sandstone pseudomorphs after halite. A) Halite pseudomorphs on the base of a sandstone bed from a thinly interbedded sandstone and siltstone sequence. Note that the larger pseudomorphs (~1.5cm diameter) have stepped ‘hopper’ faces. Independence Fjord Group, Proterozoic, north Greenland. B) Halite pseudomorphs on a sandstone bed. Locality unknown. University of Leeds Collection. C) Halite pseudomorphs on the base of a sandstone bed. Independence Fjord Group, Proterozoic, north Greenland.

occur in lines or clusters (Fig. 8.10). Some have one face of the cube parallel to bedding and have a square appearance but, more commonly, one corner of the cube protrudes so that triangular, pyramidal shapes occur. Larger pseudomorphs commonly show a diagnostic "hopper" form (indented surfaces to the cubes that reveal steps related to crystal growth). The best-shaped pseudomorphs are commonly those occurring isolated from their neighbours. Gypsum "desert rose" pseudomorphs are also common in sandstones (Fig. 8.11A) where dissolution may be controlled by changes in groundwater level and chemistry. Gypsum pseudomorphs can be confused with crack fills, particularly those discontinuous, lenticular cracks attributed to synaeresis (see §9.2.1).

Pseudomorphs usually record the former presence of evaporite crystals growing at or just below the muddy sediment surface from an overlying supersaturated brine (Fig. 8.2F). The occurrence of many, small pseudomorphs suggests rapid nucleation and possibly rapid evaporation, whereas a few large crystals may have resulted from slower, more sustained evaporation. Although pseudomorphs commonly occur in association with desiccation mudcracks, they do not themselves indicate emergence, but rather they record the existence of a shallow body of hypersaline brine that may or may not have dried out completely. A shallow water body is implied by the relatively small volumes of evaporite minerals involved.

Preservation of pseudomorphs results from a rapid influx of sediment-laden, low-salinity water that dissolves the crystals on the basin floor and fills the resulting spaces with sediment of a different character to the host material. This mechanism is sometimes associated with the occurrence of small erosional tool marks (§4.2.3) along with the pseudomorphs, possibly caused by the dragging of sharp-edged evaporite crystals as tools. Since pseudomorphs most commonly occur on the soles of sandstone beds, they are good indicators of "way-up".

Deeper bodies of water are likely to have produced thicker beds of evaporites as a result of sustained evaporation and they are therefore less likely to be prone to dissolution and the formation of pseudomorphs. Dissolution of thick beds of evaporites is more likely to be a result of much later access of fresh groundwater following tectonic uplift, the outcome of which is breccia beds in the overlying sediment (see §9.3.2).

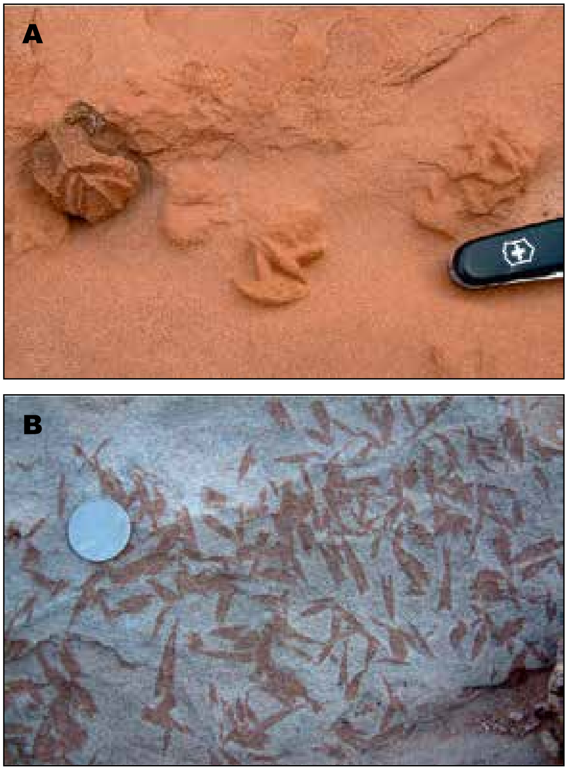


Figure 8.11 Examples of common sandstone pseudomorphs after gypsum. A) Gypsum 'desert rose' pseudomorphs within a sandstone bed. B) psudomorphs highlighting the former presence of blades of gypsum on a bedding surface. Both examples from the Cedar Mesa Sandstone, Permian, Utah, USA.

8.2.4 Diagenetic and reworked evaporites

Not all evaporite minerals occur as primary, basin-floor precipitates. Many occur as diagenetic concretions or nodules formed within a host sediment. The textures, fabrics and structures of such evaporite deposits are dealt with in §9.3.1. Desert roses are arguably diagenetic rather than primary precipitates.

Some evaporites show structures due to physical processes of erosion and deposition. Small-scale scours, ripples, cross lamination and cross bedding are all quite common (Fig. 8.2E). They record the reworking of primarily precipitated evaporite grains by currents, waves or even by the wind, as shown by large aeolian dunes of gypsum such as those found at White Sands in New Mexico (see §6.3.2).

8.2.5 Spring deposits: tufa, travertine and sinter

Around many present-day springs and caves, and in some deposits of Holocene and Pleistocene age, there are chemical precipitates in whose deposition evaporation played only a minor role. Two principal groups of deposit occur: calcium carbonate (usually calcite) and silica. Deposits of calcite are precipitated from both hot and cold springs, and the precipitation may be due to cooling, evaporation, loss of dissolved carbon dioxide, or to chemical reaction, all of which may be aided by the metabolism of algae and bacteria. One form of calcium carbonate is **tufa,** which commonly occurs as a coating on plants and plant debris. Its texture is often highly porous and spongy, and it commonly includes plant leaf and twig impressions within it. A second, more laminated and compact form of calcium carbonate is **travertine**, which occurs commonly in caves and also as the surface deposits of both hot and cold springs (Fig. 8.12). Cave travertines result from calcium carbonate, dissolved by percolating groundwater, being re-precipitated as calcite on emergence, probably as a result of de-gassing of dissolved CO_2. Elongate vertical columns (stalactites and stalagmites) develop on the roofs and floors of caves (Fig. 8.13), and steep surfaces are coated with dripstone layers. Spectacular terraces of hot springs, for example in Yellowstone Park, USA, and in New Zealand, are of travertine. Sections through these deposits show a fine lamination, some of which is columnar, and this could be confused with stromatolitic lamination of microbial origin (see §8.3.2). The lamination can be seen to be made up of layers of fibrous calcite crystals whose fibres are elongated normal to the layering.

Deposits of silica are confined to hot springs and geysers and are known as **sinter** or **geyserite**. These occur as

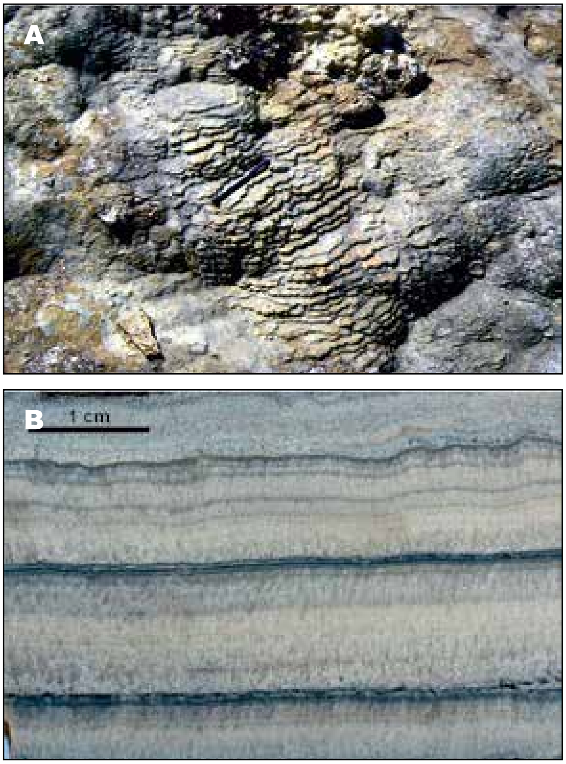


Figure 8.12 Travertine deposits associated with hot springs. A) Terraces developed around a natural hot spring. Recent, Tabernas, southeast Spain. B) Internal lamination of travertine deposits from Iceland. Specimen courtesy of Jonathan Carrivick.

Figure 8.13 Cave stalactite (speleothem), a type of travertine. A) External morphology. B) Internal lamination. Quaternary, SE Spain. Specimen courtesy of Emily McMillan. Speleothems must be left in-situ and undistributed; sampling for genuine scientific research requires special permission.

encrustations around geysers and springs, and they develop a wide variety of surface morphologies. Continued deposition leads to a variety of types of lamination, many of which compare quite closely with those of algal stromatolites (see §8.3.2).

8.3 Precipitation and binding of sediment by organisms

Organisms are active in both the precipitation of mineral matter and the binding of sedimentary particles. Here we deal, in general terms, with organically produced structures under two main headings – **reefs and bioherms**, due mainly to precipitation by a range of different organisms, and **stromatolites** and **oncolites**, due to the binding of sedimentary particles by algae and bacteria. In adopting this simple classification, it is important to note that stromatolites can themselves contribute to the construction of reef bodies.

8.3.1 Reefs and bioherms

Many animals and plants living in the sea and in freshwater settings produce aragonite and calcite as skeletal or other strengthening structures (e.g. corals, echinoids, bivalves and calcareous algae). After the death of the organisms, this skeletal material may become the main component of carbonate sediments. Some skeletons remain largely intact to become sedimentary particles in their own right, whereas others disintegrate or are broken up and abraded. The fine needles of aragonite, which make up much of the lime mud of present-day carbonate environments and which gave rise to many micritic (i.e. muddy) limestones, were precipitated chiefly by calcareous algae. Many carbonate-precipitating organisms, for example branching corals, contribute to the sedimentation process by building rigid framework structures that remain *in situ* following the death of the organism (Fig. 8.14). The form of these framework structures, as preserved in ancient successions, can yield important information about the palaeoecology of the organisms that contributed to building the framework.

Figure 8.14 Carbonate framework coral. This is an example of a framestone which, along with bindstones and bafflestones, represents one of the main reef-building structures in carbonate reefs. Lower Carboniferous, Derbyshire, England.

The detailed study of grains originating from organic processes is in part the province of the palaeontologist and the petrographer, and in this chapter we concentrate only on the larger features of limestone and dolomite successions that were produced by colonies of organisms. Where such features had topographic expression on the contemporaneous sea floor they have been termed **build-ups**, **reefs**, **mounds** or **bioherms**. They vary greatly in size, in morphology, in the nature of their internal framework, and in their organic make-up. Each of these features, but most notably the last one, has changed throughout geological time as different groups of organisms became important. At the present day, organic build-ups of carbonate on the sea floor are present in a wide range of water depths, although the most prolific and spectacular examples occur in shallow water (1–20m) of tropical and subtropical coasts.

Many reefs today occur as barriers, running for long distances parallel to a shoreline and separating the open ocean from more protected lagoonal areas (Fig. 8.15). Others are fringing reefs which encircle islands and may have a lagoon or reef flat between the reef edge and the land. Atolls are reefs that encircle a lagoon which lacks a central landmass at the present day. All these larger forms commonly have horizontal dimensions measured in kilometres and they may separate areas of deep and shallow water on opposite sides. Smaller organic build-ups or **patch reefs** (Fig. 8.16) occur within lagoons and in other shallow marine settings. A comparable variety

Figure 8.15 Pearl and Hermes Reef, northwest Hawaii. Photograph taken from the International Space Station at an altitude of 400km. Width across the image is ~20km. Image courtesy of the NASA Earth Observatory.

Figure 8.16 Small offshore patch reefs (bioherms), locally known as "boilers", constructed at the present day mainly by coralline algae and vermetid gastropods, in a very shallow, high-energy setting. South Shore, Bermuda. Photo courtesy of Brian Rosen.

occurs in the rock record, where the recognition of the largest forms can require extensive geological mapping, exceptionally large exposures or high-quality reflection seismic data. Smaller reef forms are commonly recognized in quarries and cliffs. In certain examples the present-day topography clearly reflects the palaeorelief of exhumed reef structures (Fig. 8.17). Throughout this discussion, the terms "reef" and "bioherm" are used interchangeably with no implications of water depth or other conditions of deposition.

Smaller bioherms and their associated sediments are commonly divided into **reef core**, **reef flank** and **inter-reef** components (Fig. 8.18A). Larger reefs, which are more likely to have acted as barriers between areas of contrasting water depth, are usually more appropriately divided into **fore-reef**, **reef** and **back-reef** components (Fig. 8.18B). The nature of the sediments that comprise these larger scale reef structures are commonly described with respect to their depositional texture (Fig. 8.19).

The reef core is usually a tightly bound mass of limestone, generally without any clear bedding. It may include a high proportion of framework-building skeletons (framestone) or it may be difficult to see any framework organisms where the reef mainly results from the binding effect of the organisms (bindstone). The framework organisms themselves are commonly encrusted with other organisms such as algae and bryozoa, and their surfaces bored by animals, recording early stability and lithification (see §9.4). Early formed cavities may be partially or completely filled with sediment (bafflestone), commonly in multiple generations, whereas others can have fills of fibrous or blocky calcite cement. Where a cavity is filled partly with sediment and the remaining overlying space by sparry calcite, the interface approximates to the

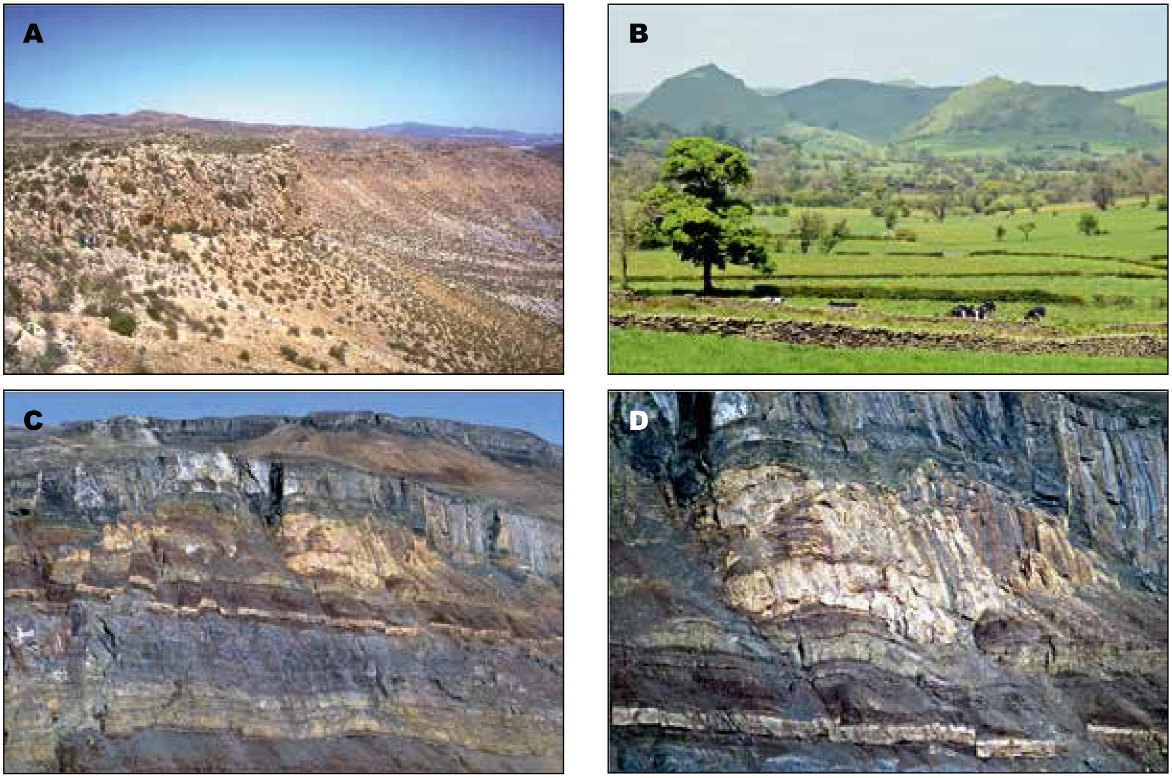


Figure 8.17 Examples of present-day surface topography revealing the palaeotopography of reef systems. A) A fringing reef exhumed in the present-day surface topography through the preferential erosion of softer overlying sediments. The prominent bench represents the transition from the reef crest to the reef front. The slope below is the fore-reef apron. Nijar, Miocene, SE Spain. Exhumation of reef bodies like these is common due to the removal by erosion of softer sediments that blanketed the reef. B) Exhumed pinnacle reefs at the edge of a limestone platform. The valley floor is underlain by mudstones that initially draped the reef topography. Bee Low Limestone, Lower Carboniferous, Derbyshire, England. C) Small stromatolitic limestone patch reef interfingering laterally with red mudstones. Maximum thickness around 50 m. Eleonore Bay Group, Late Proterozoic, east Greenland. D) Close-up view of C.

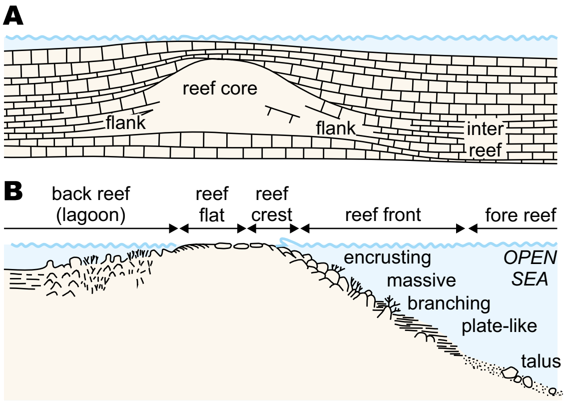


Figure 8.18 Definition diagrams for the major subdivisions of reefs. A) Isolated reef mound. B) Barrier reef at the boundary between deeper and shallower water showing the main growth forms of reef-building organisms. Modified after James (1979).

Figure 8.20 An example of a geopetal (fossil spirit level) structure. The internal cavity of the shell was partly filled with micritic mud shortly after the death of the animal. The micrite would have accumulated to a flat, palaeohorizontal surface. The remaining unfilled space in the upper part of the shell cavity was later filled by the growth of sparry calcite via diagenetic processes after burial. The level of the top of the micrite fill serves as a fossil spirit level to indicate the original attitude of the shell. Thamama Group, Early Cretaceous, United Arab Emirates. Photo courtesy of Stephen Lokier.

depositional horizontal. This can be used as a way-up indicator and for the measurement of dip in unbedded carbonates (see also §2.1.6). These **geopetal infills** can also occur in the body chambers of fossils (Fig. 8.20, 8.21). **Stromatactis** is a particular type of cavity infill developed in lime muds, commonly in reef settings. It is characterized by a flat floor, partially draped by sediment, and an irregular roof beneath which sparry calcite fills in the remaining space (Fig. 8.22). Its origin seems to relate to the compactional dewatering of the muds whereby water is trapped beneath slightly consolidated and organically bound layers. Not all the fossils of the reef core help to form a framework or to bind the sediment. Many are detached forms which simply lived within the general reef environment. Within the reef core, a vertical change in the organic content may record the progressive development of the reef. It may be possible to identify which organisms reflect stages of pioneer growth, colonization, diversification, domination, death and degradation. The flank or fore-reef deposits typically show clearer bedding, commonly with quite high depositional dips away from the reef core.

The back-reef or inter-reef sediments show more defined, horizontal bedding, although details may be obscured by burrowing. Differences in the sediments of the various zones are also reflected in the associated

original components not bound together during deposition				original components bound together	depositional texture not clear	original components not organically bound during deposition		original components organically bound during deposition		
contains lime mud			lacks mud and is grain supported			>10% grains >2mm		organisms act as baffles	organisms encrust and bind	organisms build a rigid framework
mud-supported		grain-supported				matrix supported	supported by >2mm components			
less than 10% grains	more than 10% grains				crystalline carbonate					
mudstone	wackestone	packstone	grainstone	boundstone	crystalline	floatstone	rudstone	bafflestone	bindstone	framestone

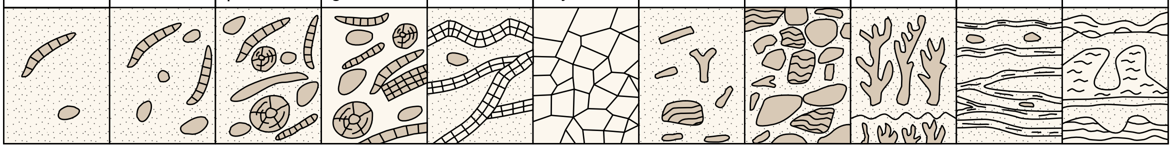


Figure 8.19 The classification of limestones based on depositional texture. Modified after Dunham (1962) and Tucker (2001).

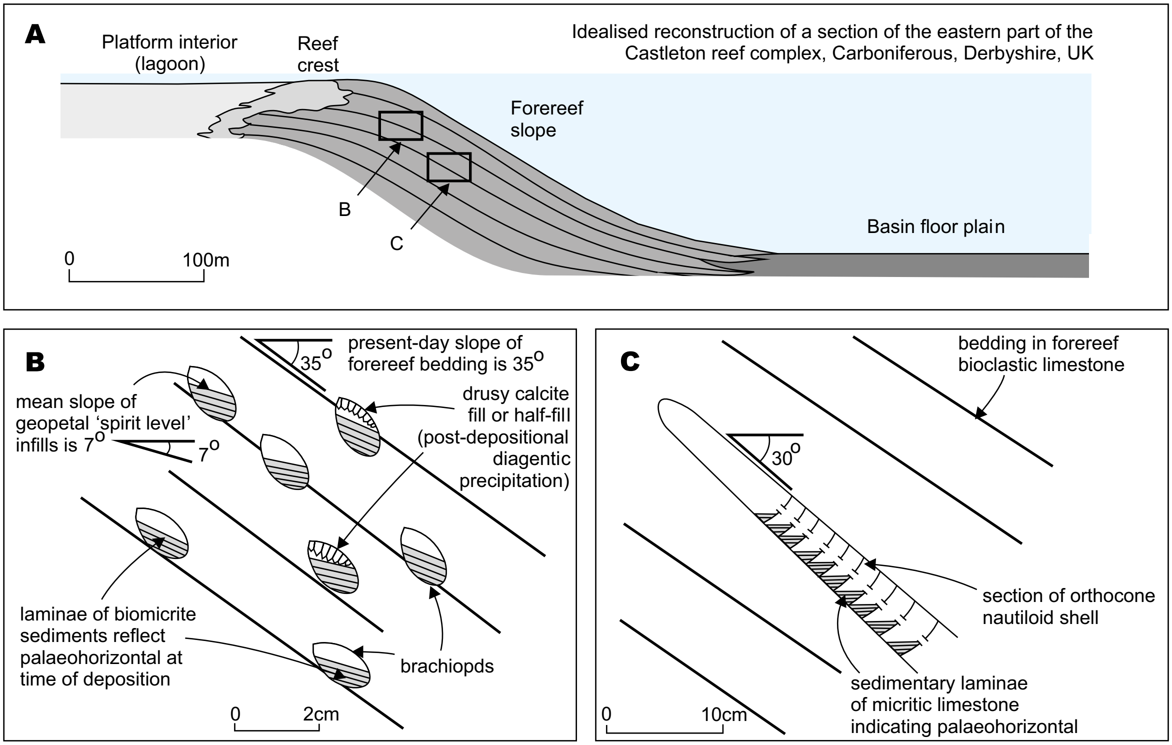


Figure 8.21 Illustration of the use of way-up and geopetal (fossil spirit level) structures for the determination of original bedding inclination. A) Idealised cross section of a carbonate rimmed shelf reef complex showing the relatively steep dip of the fore-reef slope and the inclination of the bedding developed therein. B) Geopetal infills of brachiopod shells found in sediments of the fore-reef slope. C) Geopetal infill of a nautiloid shell found on the fore-reef slope. Shell cavities are thought to have been partly infilled with lime mud soon after death of the organisms, when the shells settled to the sea bed. The laminae within the infills indicate the attitude of the palaeohorizontal at the time of deposition. In the case of the brachiopods, the inclination of the geopetal infills indicates that the reef complex has been subjected to approximately 7° of tectonic tilt. Thus, the original slope of the fore-reef was 28°, rather than 35° as indicated by the present-day attitude of the fore-reef bedding. In the part of the fore-reef where the nautiloid was found the original slope was about 30°. Data based on observations of a Carboniferous (Visean) age reef complex from Castleton, Derbyshire, England. Modified after Wolfenden (1953) and Broadhurst and Simpson (1967).

Figure 8.22 Stromatactis, an enigmatic structure in fine-grained limestone. Sparry calcite fills a flat-bottomed cavity which must have been kept open close to the sediment surface during the early stages of diagenesis. Pentamerus Bjerge Formation, Lower Silurian, Washington Land, northwest Greenland.

organisms, although the fore-reef or flank deposits may include both sediment and organisms which initially formed or lived on the reef or in the back reef. Fore-reef slopes are commonly covered with talus (rudstone) that represents debris shed down slope from the high-energy reef crest.

A full description of an ancient reef or bioherm complex should attempt to answer the following questions:

- Can the attitude of the original depositional horizontal be established (see §2.1.6)?
- How big was the reef? In particular, what topographic relief did the reef create when it was actively growing? Because sediment accumulates around a reef during its growth, assessment of contemporaneous relief in ancient examples may be difficult. With larger barrier or fringing reefs, there may have been significant

differences in water depth on opposite sides. It may in some cases be possible to trace marker horizons from the inter-reef sediments over the reef core and these might indicate topography. Bathymetry may also be suggested by larger-scale patterns of sediment distribution. Where the reef was buried by later sediments that have subsequently been preferentially eroded, the present-day, exhumed sub-aerial topography can closely match submarine topography (Figs. 8.15, 8.16, 8.17).

- What shape did the reef have in plan view? Present-day sub-aerial topography is particularly useful in some large ancient reefs, where the exhumed morphology is visible in three dimensions in the landscape. However, some inferences about reef shape can be made from more local observations. Clear differences in lithology between sediments on either side of a reef suggest that the reef acted as a barrier to water and sediment circulation, and fore-reef and back-reef areas may then be recognized. Small patch reefs or bioherms, recognizable in single outcrops, are usually flanked and surrounded on all sides by similar sediment.
- What organisms were involved in reef growth? Is it possible to identify a main frame-building organism? To what extent is the reef a result of frame building and to what extent is it due to sediment binding? Can any lateral or vertical variations be seen in the distribution of the types and abundances of different organisms? Recognition of such zonation may yield information regarding environmental parameters such as water depth and energy regime. These questions demand palaeontological expertise and are best approached after study of more specialized literature.
- To what extent did early lithification occur in the reef? This might be best judged from observation of fore-reef and reef-flank deposits. Do these consist of bioclastic debris or contain larger blocks of reef material? Bioclastic debris commonly occurs in steeply dipping beds, inclined away from the reef core. Large blocks, some of very large dimensions (some hundreds of metres in diameter), occur on some steep palaeoslopes associated with dipping fore-reef beds, and are interpreted as submarine scree (talus) made up of lithified blocks that fell from the reef front (see also olistostromes, §9.2.3). Geopetal infills within large blocks may indicate the degree of tilting of the blocks or their way-up if the infills formed prior to re-deposition (Fig. 8.21; see also §2.1.6).

8.3.2 Stromatolites, oncolites and mat-like structures due to algal or microbial binding

Stromatolites and **oncolites** are structures that show fine lamination caused by the trapping and binding of material by algae and cyanobacteria. Most occur in shallow-marine settings, but some examples are known from lake margins. Stromatolites occur as sheets, columns, hemispheres, domes and tabular forms on a variety of scales. Modern examples of stromatolites are rare, the world's best-known example being the columnar forms present in Shark's Bay, Western Australia (Fig. 8.23).

Stromatolites

Our knowledge of stromatolites derives largely from rocks of Precambrian age, although they are also quite common throughout the Phanerozoic. Stromatolitic lamination is commonly found in mud-size carbonate sediment although coarser-grained carbonate and detrital material can also be involved. Lamination is characteristically thin, usually 1mm or less, and has a rather delicate appearance. In some cases the lamination is irregular with small cavities filled by sparry calcite between the layers. This **birdseye structure** is the result of shrinkage of the algal layer and also the generation of gas from rotting algae. The shapes that the lamination takes are extremely varied and several taxonomic schemes have been proposed for their classification. There is, however, scope for confusion as stromatolitic lamination is in some cases rather similar to some non-biological lamination, such as that in travertine, tufa and the silica deposits which form around geysers and hot springs (see §8.2.5).

Figure 8.23 Present-day stromatolitic domes growing on the floor a shallow, hypersaline lagoon. Hamelin Pool, Shark's Bay, Western Australia. Photo courtesy of Stephen Lokier.

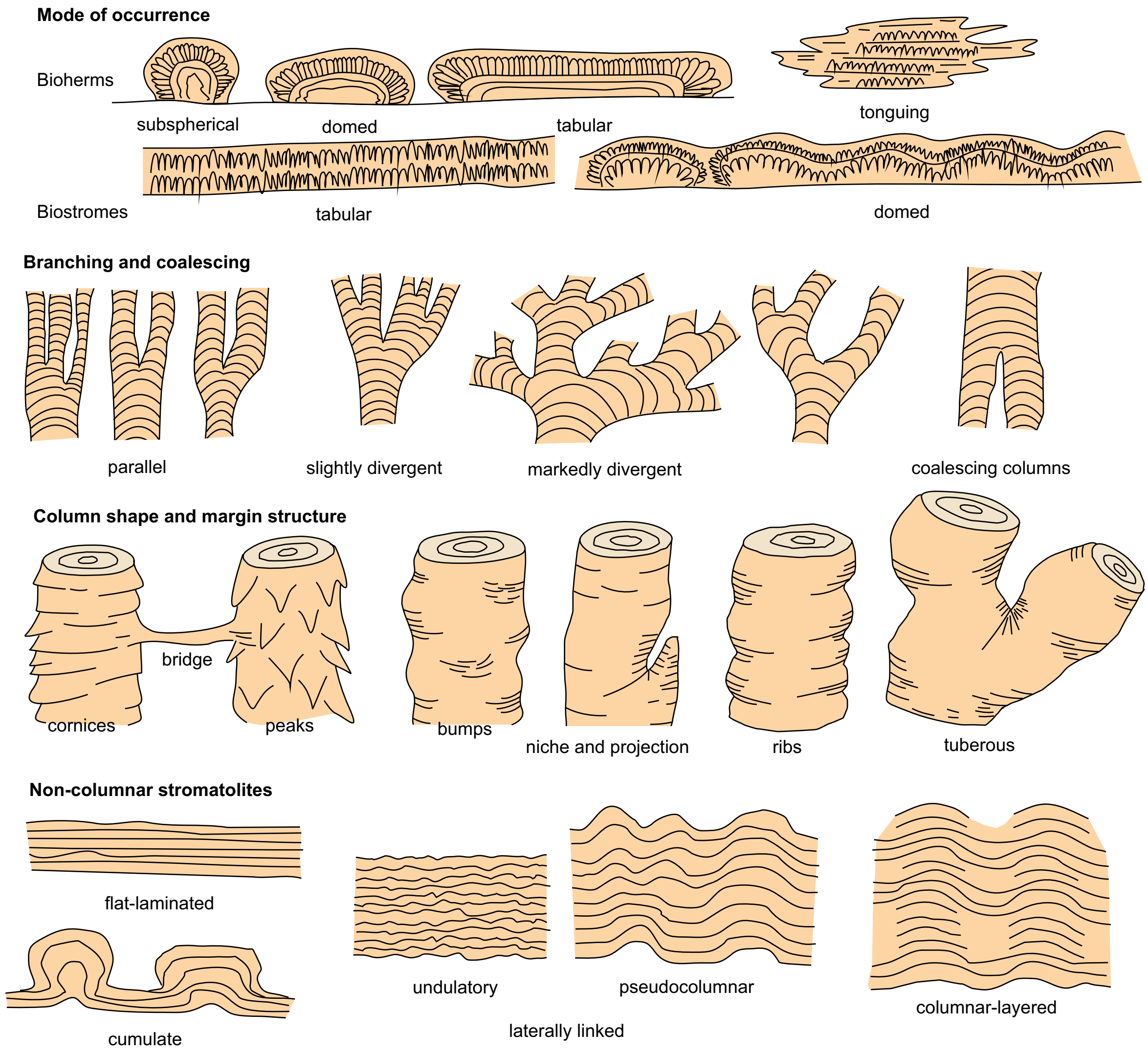


Figure 8.24 Definition diagram of the main terms used in the description of stromatolite bodies and stromatolitic lamination. Modified after Preiss (1976).

For most purposes, it is sufficient to recognize some of the large- and medium-scale features of stromatolites that contribute to their field description (Fig. 8.24). At the largest scale ("mode of occurrence") the broad shape of the stromatolitic unit is described. Such units can be of any thickness from a few centimetres to several metres, and stromatolitic biostromes may extend horizontally for many kilometres. Within bioherms or biostromes, the stromatolitic lamination may be organized into either columnar or non-columnar forms, and these in turn show a great variety of shape and scale. The types illustrated in Figure 8.16 do not depict all possible variations and a good, scaled field drawing or photograph will usually be more valuable than words. It is important to assess and document the three-dimensional form of stromatolites; in addition to their vertical section (Fig. 8.25), describe their appearance in horizontal section or on bedding surfaces wherever possible (Fig. 8.26). With columnar forms, try to establish their cross-sectional shape and size. Columns tend to be circular or elliptical in plan view, the latter type sometimes having a preferred orientation. Record the direction of any such orientation as it may give a guide to current or wave

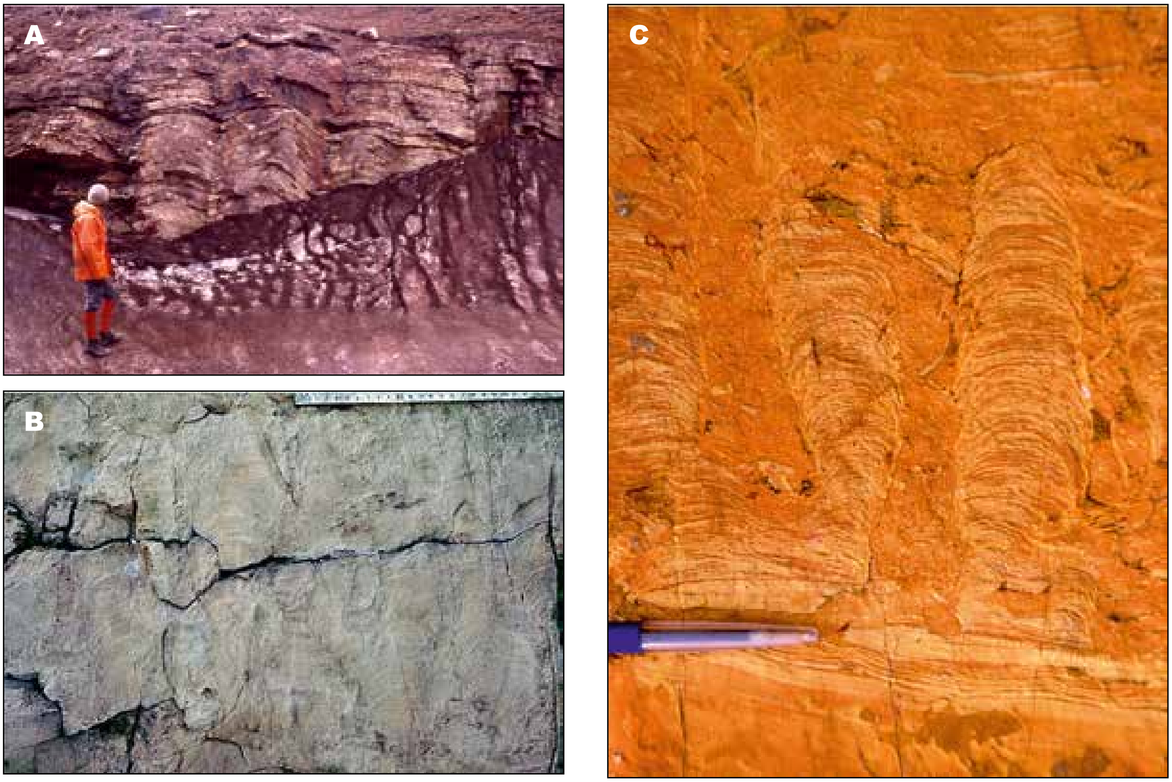


Figure 8.25 Examples of stromatolitic lamination seen in vertical section. A) Stromatolites with a columnar-layered structure, Morœnesø Formation, Proterozoic, northeast Greenland. B) Discontinuous columnar forms. Fyn Sø Formation, Late Proterozoic. northeast Greenland. C) Upright, non-branching stromatolite columns, Eleonore Bay Group. Proterozoic, east Greenland.

Figure 8.26 Horizontal sections through stromatolite columns. Fyn Sø Formation, Late Proterozoic, northeast Greenland.

directions and hence to the trend of a palaeoshoreline. Bedding surfaces may show a three-dimensional relief (Fig. 8.27) and this allows visualization of the morphology of the sediment surface during deposition.

Although much less widespread and varied than they appear to have been in the geological past, present-day algal mats show a range of morphological forms, from flat sheets through various crinkled and pustular types, to well-developed columns. Present-day types typically occur in shore -parallel zones, that reflect the physical and chemical conditions in which they develop. Columnar structures may be elongated in plan view and typically have a preferred elongation normal to the shore.

Stromatolitic lamination results from the trapping and binding of sediment by the mucilaginous filaments of algae and bacteria, which form mats growing on the sediment surface. Sediment settles from suspension onto the mats and is generally not precipitated by the algae or bacteria themselves. The lamination is produced by periodic variation in rates of sediment supply allowing the build-up of organic-rich layers. The algal or bacterial mat re-establishes itself after an episode of rapid deposition by growing through the sediment layer, thus binding the sediments to generate a bindstone.

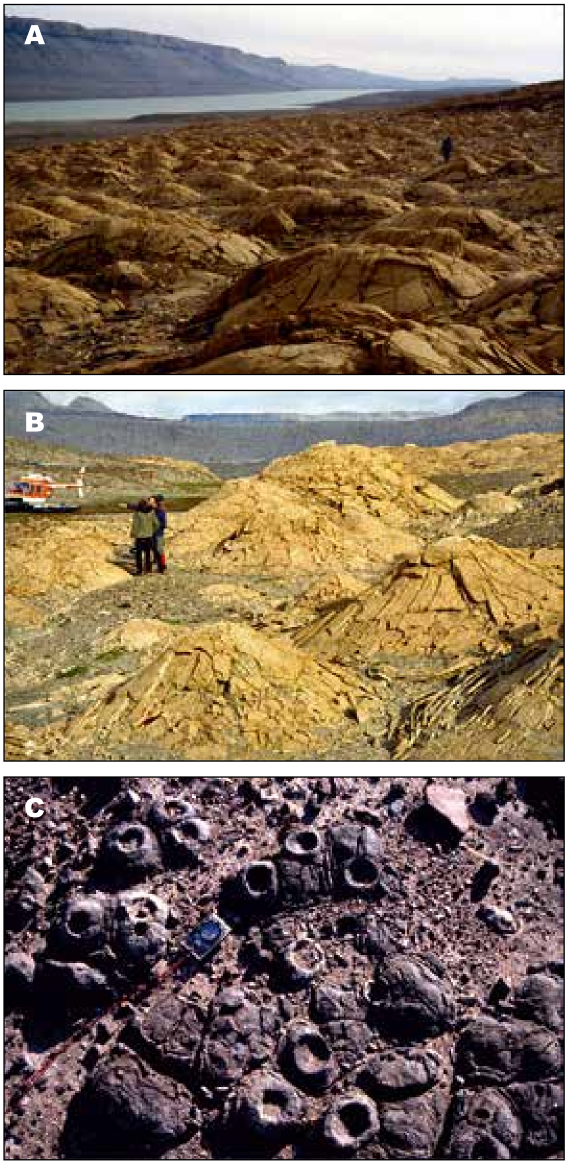


Figure 8.27 Stromatolite domes with their three-dimensional relief preserved on the upper bedding surface. A) and B) Morœnesø Formation, Proterozoic, northeast Greenland. C) Huab Formation, Permian, northeast Namibia.

Most present-day stromatolites are associated with particularly high salinities in intertidal and supratidal settings, although subtidal examples are also known. Thin developments are also known from the margins of lakes, as in East Africa. The more widespread occurrence of stromatolites in rocks of Precambrian age probably results from the absence of animals to graze upon the algae or bacteria.

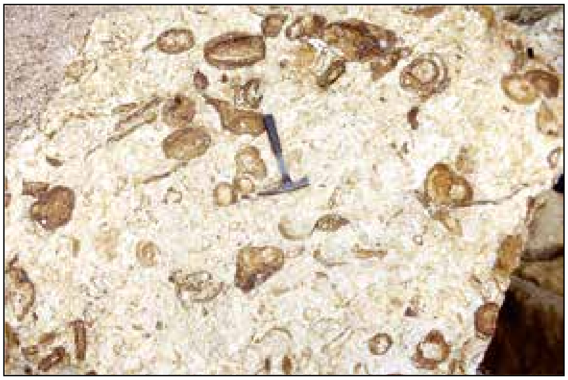


Figure 8.28 Cross section through large oncolites showing internal lamination. Photo courtesy of Gilbert Kelling. Oncolites are typically 1–2cm in diameter; these larger examples have been termed algal biscuits. Hammer is 0.3 m long. Burton Beds, Inferior Oolite, Middle Jurassic, Dorset, England.

The high salinities associated with many present-day examples create conditions hostile to those animals and thereby allow the mats to flourish. The commonly stated view that stromatolites are indicators of intertidal conditions is misleading and, for Precambrian examples particularly, there seem to be no environmental requirements other than the availability of water and sunlight.

Attempts to use stromatolites as a basis for the biostratigraphic zonation of otherwise unfossiliferous Proterozoic sediments have met with limited success and acceptance.

Oncolites

Oncolites (sometimes called **algal biscuits)** are spherical or less well-rounded structures, commonly up to 5mm in diameter but sometimes bigger, commonly with a rather flattened shape. Internally they have a roughly concentric pattern of fine lamination similar to that of stromatolites (Fig. 8.28). Careful examination of the lamination in cross section may reveal discontinuities in some cases.

Oncolites occur in both ancient limestones and in present-day lagoons and lakes. They result from the binding of sediment by algae onto isolated nuclei. Their discontinuous pattern of lamination, suggests relatively calm conditions, where growth took place on the upper surface, alternating with occasional higher-energy episodes that turned the oncolites over and allowed growth to proceed on the opposite side. Oncolites may be confused with diagenetically formed pisoliths which occur in some carbonate-rich soil profiles.

Figure 8.29 Microbial structures. A) Microbial mats on the surface of a saline lagoon giving a polygonal pattern of push ridges resulting from isotropic compressional stress. B) Vertical section through the pressure ridges seen in A, showing tight folding of the surface layer at the ridges. Polygons in A are ~0.5-1m diameter. Both examples from the United Arab Emirates. Photos courtesy of Stephen Lokier.

Microbial mat-like structures

Where microbial communities grow to cover sediment surfaces as a mat, their continued activity commonly leads to the development of ridge-like deformation features, many examples of which develop as tessellating polygonal forms in plan view. In section view, the mat-like structures may be interbedded with carbonate sediment to form laminations, commonly with upward-pointing deformed ridges (Fig. 8.29).

8.4 Early cementation

In addition to the binding activity of organisms, carbonate sediment on the sea floor may be subjected to early cementation through the precipitation of aragonite or high-magnesium calcite from sea water. Early cements are not particularly common in seafloor sediments and are favoured by slow depositional rates, a rather stable sea floor and relatively high levels of wave and current activity. Carbonate cementation also occurs on both carbonate and clastic beaches giving **beach rock.** Carbonate sediments which become sub-aerially exposed through a fall of relative sea level are also subjected to cementation through solution and re-precipitation because of the passage of fresh water. In that case, the cementing mineral is almost always calcite. Sustained sub-aerial exposure can lead to more extensive dissolution and to the development of karstified surfaces and carbonate soil profiles (see §8.4.2).

8.4.1 Seafloor lithification

Cementation close to the sea bed may be patchy or lead to the development of continuous layers. Early cemented patches take the form of **concretionary nodules** around which differential compaction can take place, especially if the host sediment is fine-grained. This is recognized in ancient sediments through tracing laminae from the surrounding compacted sediment into and across the nodule, a relationship that enables early cemented nodules to be distinguished from those of later diagenetic origin (see §9.3.1). When seen in the rock record, such nodules commonly follow particular bedding horizons.

Laterally continuous cemented layers commonly form a few centimetres below the sea floor and tend to have rather flat top surfaces and rather irregular bottoms. Continued precipitation may lead to buckling and breakage of the layers, the development of anticlines and the thrusting of layers one over the other to give carbonate **tepee structures** (see Figs. 8.6, 8.7 for examples formed of salt and 8.29 for a microbial example). In plan view the upwards-buckled ridges commonly have polygonal forms. These structures occur on the floors of present-day marine lagoons and sabkhas, as well as in the rock record.

When the cemented layers, which probably developed a few centimetres below the sea bed, are swept clean of loose sediment, the lithified surface may become encrusted and bored by marine organisms and be mineralized by phosphatic and manganese-rich minerals. These bored and encrusted surfaces are called **hardgrounds** and their occurrence in the rock record section provides evidence for sustained intervals of non-deposition. The recognition of borings is dealt with in §9.4. Partly or weakly lithified surfaces are called **firmgrounds** (Fig. 8.30) and these may be subject to particular types of burrowing by organisms that favour colonization of such substrates (see *Glossifungites* ichnofacies in §9.4.6 and Fig. 9.77).

Figure 8.30 Example of a present-day firmground. United Arab Emirates. Photo courtesy of Stephen Lokier.

Early cementation of carbonate sediments on submarine slopes may lead to the spectacular development of breccias through the secondary movement of the lithified or partially lithified material, for example by storm events that impact on shallow sea beds or as mass gravity flows. Breccias made up of tabular clasts are especially characteristic (Fig. 8.31).

More general comments on preferential cementation and the development of nodules and concretions are given in §9.3.

8.4.2 Sub-aerial exposure

Sub-aerial exposure of carbonate sediments leads to their rapid lithification in most circumstances. Percolating rainwater dissolves aragonite and high-Mg calcite and re-precipitates the ions as low-Mg calcite cement. Sustained exposure leads to further modification, which may take the form of either karstification or soil development.

Sustained dissolution associated with abundant fresh water will lead to the development of **karstic features**. Such features are not just the product of recently exposed carbonate sediments but also develop very extensively in older limestones which have been subjected to abundant rainfall or snow melt. Karstic features developed at the land surface give very characteristic morphologies and landscapes. They include patterns of sharp ridges and deep

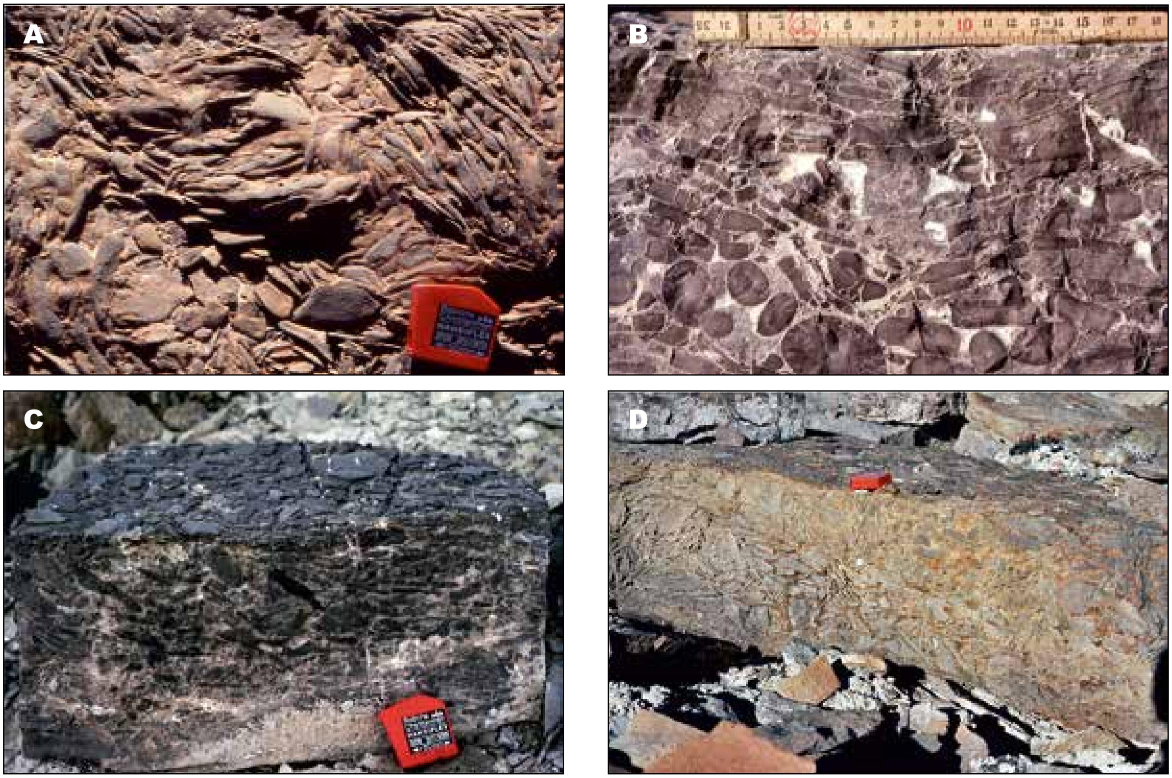


Figure 8.31 Breccias developed in thinly bedded limestones as a result of secondary movement following partial lithification. A) Cass Fjord Formation, Upper Cambrian, Washington Land, northwest Greenland. B) Broken slab of concretionary material. Brønlund Fjord Formation, Lower Cambrian, Peary Land, northeast Greenland. C) Same as A. D) Vertical profile through the breccia shown in A.

clefts (clints and grykes) due to the widening of joints by dissolution resulting in steep-sided funnels and pipes. Within limestone, solution produces cavities which may grow to the size of major caves.

Precipitation of carbonate as various types of travertine (see §8.2.5) may fill or partly fill voids and cavities. In some instances the cavities may collapse, leading to the addition of **collapse breccias** to the fill. Such units tend to be irregular in shape and bear no relationship to bedding. The clasts of such breccias are all of clearly local derivation, irregular and angular in shape, and with a very mixed range of sizes. Spaces between them may be partially or totally infilled with travertine deposits.

In studying any ancient limestone succession, it is very important to try to distinguish those karstic features due to relatively recent (Holocene or Pleistocene) activity from those which developed during the early post-depositional history of the limestone. This is usually far from easy and the careful application of principles of cross-cutting relationships will be needed (see §2.1.3). In some ancient successions, karstic dissolution surfaces occur repeatedly on the tops of many limestone beds and are typically draped by either thin layers of insoluble residue (usually clays) or by sandstones that infill the relief (Figs. 8.32, 8.33). Such occurrences are almost always palaeokarstic in origin (i.e. formed during the overall accumulation of the succession) and they typically record periods of lowered sea level when the sediment surface became emergent.

Continued solution and precipitation close to the surface, commonly under conditions of lower rainfall, may lead to the extensive and organized occurrence of calcite in **calcrete** or **caliche** soil profiles. Such profiles also occur in host sediments other than carbonates are dealt with in §9.3.1.

8.5 Other bedding phenomena in limestones

A large proportion of carbonate sediments are made up of sand, silt and mud-grade material, which is subjected to erosion, transport and deposition similar in most respects to that experienced by detrital particles. For such sediments, the resulting sedimentary structures are very similar therefore to those described in Chapters 3–6. For example, gutter casts occur on the bases of sharp-based calcarenites, probably related to storm events, wave ripples are quite common in bioclastic limestones and oolites, and cross bedding typically occurs in sand-grade limestones that have been subjected to tidal currents. Sand-grade limestones are commonly oolitic, comprising

Figure 8.32 Large-scale palaeokarstic solution pipe in limestone of shallows-water origin infilled by sandstone. The feature developed on the surface of the limestone, soon after its deposition and early lithification, during a period of lowered sea level and emergence. The base of the pipe is at the base of the sandstone and the cave below is of recent origin. Scale is 1m. Lower Carboniferous, Anglesey, Wales.

Figure 8.33 Small-scale relief at a limestone-limestone contact due to palaeokarstic dissolution during a period of emergence, augmented by pedogenic/solutional processes. Rocky Bay Formation overlying Belmont Formation. Pleistocene, Grape Bay. Bermuda. Photo courtesy of Brian Rosen.

grains coated in calcium carbonate as a result of inorganic precipitation in high energy, shallow water settings. The coated **oolith** grains are up to 5mm in diameter, though are more typically <1mm, and usually exhibit a fine concentric internal lamination when viewed in thin section.

In addition to the more common structures, a few bedding styles seem to be unique to carbonates and reflect either the different grain types that are present in lime sands or the fact that carbonate minerals are, in geological terms, readily dissolved and re-precipitated.

8.5.1 Large-scale sigmoidal cross bedding

Certain shallow-marine limestones, typically oolites or less commonly bioclastic limestone, show sigmoidal cross bedding at the scale of several metres vertical thickness and great lateral extent, commonly with complex smaller-scale structures superimposed. The cross-bedded structures are thought to reflect the growth of large oolite and lime-sand shoals, probably as a result of strong tidal currents in an area starved of detrital supply. Indeed, the agitation of carbonate sands grains by strong currents operating within well-oxygenated, shallow water is thought to be the most common mechanism for the inorganic precipitation of carbonate grain coatings that form ooliths.

8.5.2 Parallel-bedded limestones

Many limestones seen at outcrop show clear parallel and horizontal bedding. Close examination, however, commonly reveals no obvious control in the form of grain size and compositional differences between beds. The limestone seems to record very uniform conditions of deposition with no primary differentiation, yet bedding is apparent. In some cases bed-parting surfaces (bedding planes) may show some signs of dissolution in the form of stylolites (see §9.3.2), suggesting that the bedding is, at least in part, post-depositional in origin. A variety of detailed processes may have operated and not all cases need have a similar explanation. Careful documentation of the bed-parting surfaces in homogeneous, bedded limestones may lead to original insights into a poorly understood phenomenon.

Study techniques

Field experience

Present-day environments

First-hand investigation of structures in evaporites is possible in lagoons and in trenches cut in sabkhas (saline flats). Hot and cold springs, caves and lakes (with deposits of tufa and sinter) may be visited in many parts of the world. Field excursions are increasingly possible to present-day carbonate environments, wherein reefs, bioherms and stromatolites are found, but diving equipment and training are necessary in order to examine many subtidal processes. Equipment needed to monitor processes in present-day areas includes the normal field equipment plus Eh and pH meters, sampling bottles for salinity measurements and plankton counts, current velocity meters, boats, diving equipment, etc.

Ancient environments

Examination of ancient reefs, bioherms and stromatolites poses few problems, for they are frequently found in the geological record in many parts of the world and are generally well exposed. Outcrops of thick evaporite successions are rare in humid-climate settings because of their susceptibility to dissolution; they are, however, relatively common in arid and semi-arid climate settings. Visits to salt or gypsum mines may be possible.

Laboratory experience

Structures of evaporites may be available in borehole cores. Simple experiments to produce textures and structures from saturated solutions of sodium chloride or calcium sulphate are relatively simple to carry out.

Recommended references

Busson, G., 1980. *Evaporites*. Illustration and interpretation of some environmental sequences. A well-illustrated text showing many examples of differing types of evaporate deposits.

Dunham, R. J. 1962. *Classification of carbonate rocks according to depositional texture*. The classification scheme most widely used in discussing limestones.

James, N. P. & B. Jones 2015. *Origin of carbonate sedimentary rocks*. Considers carbonate sedimentary structures, lithofacies and environments in detail.

Nissenbaum, A. (ed.) 1980. *Hyperslaine brines and evaporitic environments*. Considers chemical sedimentation in hypersaline settings.

Preiss, W. V. 1976. *Basic field and laboratory methods for the study of stromatolites*. A good introductory text if you have to deal with stromatolites in any detail.

Scholle, P. A., D. G. Bebout, C. H. Moore (eds) 1983. *Carbonate depositional environments*. A very wide-ranging and well-illustrated compilation.

Scoffin, T. P. 1987. *An introduction to carbonate sediments and rocks*. A very good starting point for studying carbonate sediments.

Tucker M. E. & V. P. Wright 1990. *Carbonate sedimentology*. An excellent and accessible summary.

Walter, M. R. (ed.) 1976. *Stromatolites*. A comprehensive compilation of papers on the topic as appreciated at the time.

Warren, J. 1999. *Evaporites, their evolution and economics*. Considers chemical sedimentation in detail.

Wilson, J. L. 1975. *Carbonate facies in geologic history*. A classic of its time; carbonate sediments in their wider context, although little on sedimentary structures as such.

CHAPTER 9

Structures due to deformation and disturbance

9.1 Introduction

Any sediment may become disturbed after deposition, but disturbance is most common in sands and finer-grained material. Depositional structures may be disrupted and distinctive new structures may form as a result of physical, chemical and biological processes. It is often difficult to tell when physically and chemically induced disturbances took place. In some cases, they might have occurred soon after deposition at, or close to, the contemporaneous surface; in other cases, they might be associated with later burial and lithification.

Many deformational structures are valuable as "way-up" indicators and most record something about conditions within the sediment or at its surface soon after deposition.

9.2 Physically induced soft-sediment deformation

This results from mechanical forces, commonly gravity, acting upon physically weak sediment, usually silts or sands, at the sediment surface or soon after burial (Figs. 9.1, 9.2). There is no universally accepted scheme for the classification of these structures. Here we use a broadly morphological scheme, based upon where they most commonly occur. However, several structures are observed in both vertical section and on bedding surfaces, and might, therefore, be placed in more than one category. They are described under only one heading, usually their most common mode of occurrence.

Most types of soft-sediment deformation depend on unconsolidated sediment being in a weak condition. The

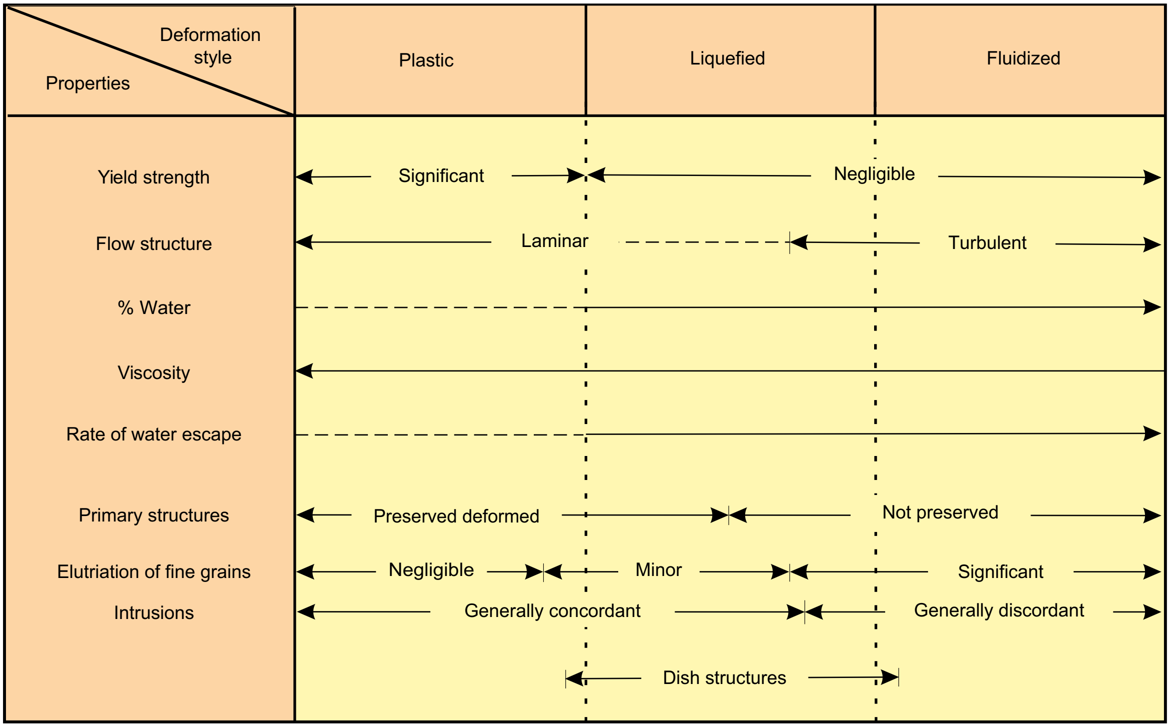


Figure 9.1 Characteristic properties of plastic, liquefied and fluidized styles of deformation. Modified after Owen (2005).

<table>
<tr><td colspan="2"></td><td colspan="5">Loss of Strength</td></tr>
<tr><td colspan="2" rowspan="2"></td><td colspan="3">Exceed Strength of Sediment</td><td colspan="2">Liquidize</td></tr>
<tr><td>Internal Tensile (Brittle)</td><td>Internal Cohesive (Plastic)</td><td>External Surface Cohesive (Plastic)</td><td>Liquefied</td><td>Fluidized</td></tr>
<tr><td colspan="2">Gravitational body force on slope</td><td>Slides</td><td>Slumps</td><td>Slumps and slides</td><td colspan="2">Debris flows</td></tr>
<tr><td colspan="2" rowspan="2">Unequal confining load</td><td rowspan="2">Growth faults</td><td colspan="2" rowspan="2">Loaded ripples
Shale ridges and diapirs</td><td colspan="2">Loaded ripples and sole marks</td></tr>
<tr><td></td><td>Clastic dykes
Sand volcanoes</td></tr>
<tr><td rowspan="4">Gravitationally unstable density gradient (Density inversion)</td><td>Continuous</td><td rowspan="6">Soft sediment faults</td><td colspan="2" rowspan="3"></td><td>Convolute lamination</td><td></td></tr>
<tr><td>Within a single layer</td><td>Dish structures</td><td>Water escape pipes and pillars</td></tr>
<tr><td>Multiple layer, not pierced</td><td colspan="2">Bedding surface load casts</td></tr>
<tr><td>Multiple layer, pierced</td><td colspan="2">Shale ridges and mud diapirs</td><td colspan="2">Ball and pillow/pseudonodules
Isolated load balls</td></tr>
<tr><td rowspan="2">Applied Shear stress</td><td>Current drag</td><td colspan="2" rowspan="2"></td><td>Overturned cross-bedding</td><td></td></tr>
<tr><td>Vertical</td><td></td><td>Water escape pipes and pillars</td></tr>
</table>

Figure 9.2 Types of physical deformation structures in relation to the nature of the deforming force. Modified after Owen (2005).

resistance of sediment to deformation is most commonly expressed by its shear strength τ, which is a function of grain cohesion C, intergranular friction and the effective pressure between the grains:

$$\tau = C + (\sigma - \rho)\tan\varphi \qquad (9.1)$$

where σ is pressure normal to shear, ρ is excess pore-fluid pressure and φ is the angle of internal friction.

For sediment to be deformed, its shear strength must be reduced or the applied shear stress increased. This can be achieved by loss of cohesion, by readjustment of grain packing to reduce $\tan\varphi$, or by increasing the pore-fluid pressure ρ. Cohesion is the least readily changed property as it is largely controlled by grain size. A shock applied to waterlogged, loosely packed sediment can change the packing and, in the process, increase the pore-fluid pressure to the extent that the sediment undergoes temporary **liquefaction** (Figs. 9.1, 9.2). In this condition, sediment and water together behave as a liquid, deforming very readily. This will continue until the pore-water pressure falls due to escape of the excess water, and the grains take on a closer packing and re-establish frictional contact with one another. The shocks that cause liquefaction may be widespread and external, for example earthquakes, or they may be local, for example a rise in water level or an episode of sudden deposition.

This effect is illustrated by jumping up and down on a sandy beach close to the water's edge. The surrounding sediment liquefies as frictional contacts break down and water escapes to the surface. Once this has happened, the same patch of sand is not easily liquefied again as a more closely packed grain fabric has been created.

In addition to shock and repacking, excess pore-fluid pressure is commonly produced during rapid deposition of fine-grained sediment. The low permeability of such sediments inhibits the escape of pore fluid and, thus, slows the rate of compaction of the sediment in response to the increasing overburden. **Overpressured** or **under-compacted** conditions are then said to occur, in which state the sediment is highly susceptible to deformation.

Liquefaction of sediment may be total, so that all grain contact is broken and the mass of sediment and water flows freely. In such cases, original lamination is destroyed, giving massive or "slurried" bedding. In other cases, where loss of strength is less comprehensive, deformation is limited and

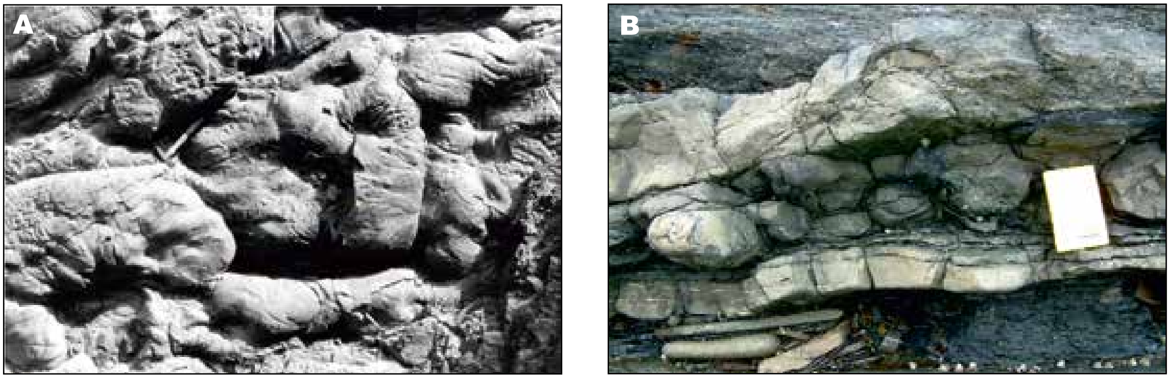


Figure 9.3 Load casts. A) Examples on the base of a sandstone bed from an interlaminated sandstone and mudstone succession. Bude Formation, Upper Carboniferous, north Cornwall, England. B) Large, almost detached load casts below a thick sandstone bed. Coal Measures. Upper Carboniferous, Pembrokeshire, Wales.

more plastic in nature so that original lamination is preserved, although distorted. A mass of liquefied sediment will remain mobile or weak until the excess pore-fluid pressure is dissipated either by general intergranular flow of pore water, usually upwards, or by water escape along localized pathways. If vigorous enough, the upward escape of fluid may lead to the **fluidization** of sediment within escape pathways (Figs. 9.1, 9.2). Rapid fluid movement between the grains causes a loss of strength and increased pore space. The relative movement of grains and fluid during fluidized flows allows some grain sorting to take place, usually by upwards removal of fines, thereby leaving cleaner sands behind. In liquefied sediment, fluid and grains move essentially together giving little scope for such sorting.

9.2.1 Features visible both on bedding surfaces and in vertical section

Load casts and flame structures

Load casts and flame structures occur most commonly on the lower surfaces of beds of sandstone that are interbedded with mudstones (i.e. they are a type of sole mark; see §4.2). They also occur within sandstone units and are commonly recognized in vertical section. **Load casts** on soles of sandstone beds are rounded, rather irregular lobes of variable size and relief. Small examples are measured in millimetres and large ones may be tens of centimetres or even metres in diameter. They seldom occur in isolation and usually cover large areas of a bedding surface (Fig. 9.3).

Upward-pointing fingers or wedges of the underlying unit occur between the sandy lobes. These are **flame structures** and they are an inevitable accompaniment of load casts (Figs. 9.4, 9.5). Although many load casts are simple protrusions on the sandstone base, some are more globular, being attached to an overlying sandstone by a thin neck or even being totally detached from it. In some cases, there is no sign of an overlying sandstone that could have been the source of the sand and the **isolated load balls** or **pseudo-nodules** "float" in the mudstone to give a **ball and pillow** structure. The mudstone surrounding the sandstone "balls" commonly shows a disturbed "slurried" texture (Figs. 9.6, 9.7).

Internally, load casts show contorted lamination. Close to the edges, lamination parallels the margins, but contortion is commonly more intense towards the centre. Where lamination is seen in the layer beneath the loaded surface, it tends to follow the margins of the flame structures, becoming contorted in the centre of the flame (Fig. 9.5). In relatively rare examples, load casts are centred on original ripples so that deformed laminae were originally cross laminae.

The mechanism of formation entails gravity acting on beds that were unstable due to their high porosity and lack of compaction and coherence, and to differences in density between the beds. Muds commonly have a high depositional porosity (60–70 %), much greater than that of even rapidly deposited sands (30–40 %). Thus, if a sand layer is rapidly deposited on a mud layer, it will be denser than the mud and, if both sediments are weak, the sand will tend to sink into the mud by loading.

In loading, the mud may have lost strength due to excess pore-fluid pressures generated by rapid deposition of the overlying sand layer. For isolated load balls or pseudo-nodules, a relatively thick mud bed must have lost its strength. The sudden deposition of the sand could

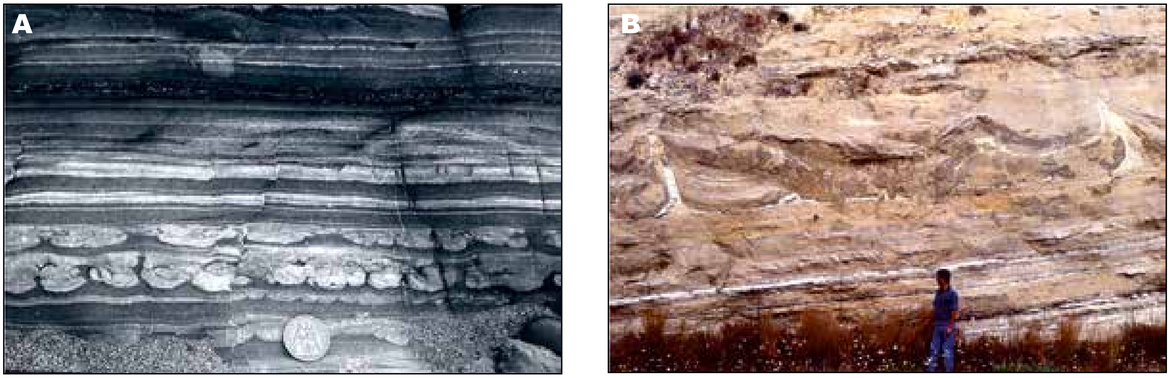


Figure 9.4 Examples of load casts and associated flame structures seen in section. A) Loads on the base of a sandstone bed with flames of mudstone squeezed upwards between them. Bude Formation, Upper Carboniferous, north Cornwall, England. B) Large loads and flames in sandstone beds of shallow-marine origin. Handere Formation, Pliocene, Adana Basin, Turkey. Photo courtesy of Gilbert Kelling.

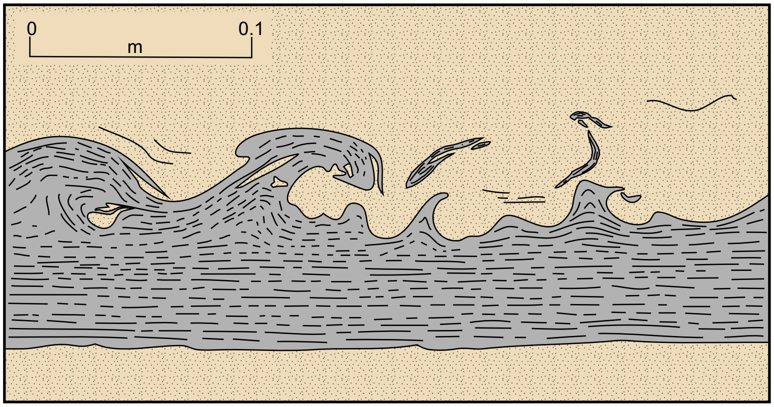


Figure 9.5 Vertical section revealing the geometry and interrelationship of load casts and flame structures resulting from the sudden emplacement of a volcanogenic sand bed on top of an unlithified mud unit. Langdale Slates, Ordovician, Cumbria, England. Modified after Sorby (1908).

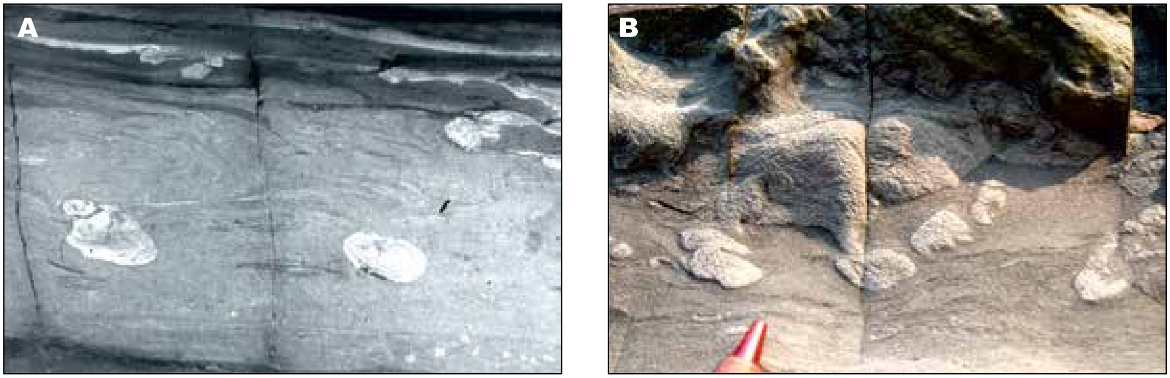


Figure 9.6 Examples of isolated load and flame structures. A) Isolated sandstone load balls ('pseudo-nodules') in a siltstone with a 'slurried' texture. The original sandstone bed has totally foundered and collapsed into the finer sediment. The load balls are about 5cm in diameter. Bude Formation, Upper Carboniferous, north Cornwall, England. B) Isolated load balls in a finer-grained silty matrix that shows a slurried fabric. Flames of the finer sediment penetrate upwards into the overlying sandstone. Clare Group, Upper Carboniferous, western Ireland.

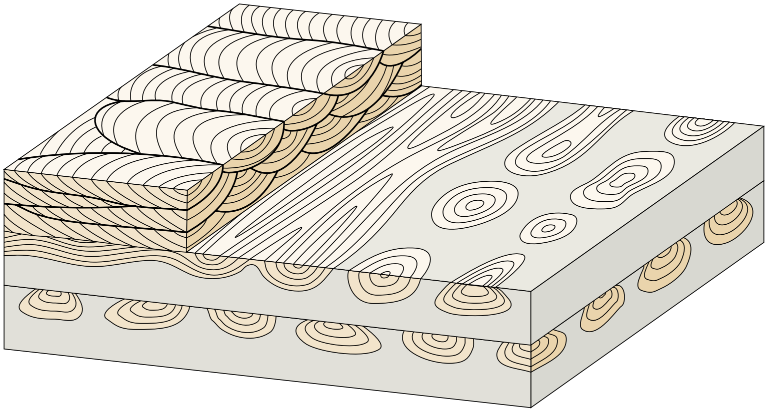


Figure 9.7 Schematic illustration of the generation of isolated load balls (pseudonodules) through the lateral passage of a continuous sand bed overlying mud and into a progressively more intensely deformed state, culminating in the isolation of sand 'nodules' that are completely enveloped in mud and give a ball and pillow structure. Note that the primary internal laminae are still evident, albeit in a deformed state. In this case, overlying troughcross stratified sets of sand have provided the loading mechanism necessary for generation of the pseudonodules. Modified after Allen (1982).

cause this, but an externally generated shock, such as an earthquake, is also possible. This effect can be simulated in the laboratory by violently jarring a waterlogged sand and mud succession.

Whatever the reason, the combination of density inversion and temporary weakness leads to the sinking of one bed into the other, either randomly or localized at pre-existing thickness differences (e.g. scours or ripples). The dimensions of load casts correlate roughly with the thickness of the sandstone bed. For isolated load balls, the whole sand bed must have sunk. Preservation of internal lamination within load balls implies that the sand was not fully liquefied.

As "way-up" indicators, load casts are generally unambiguous. Their downwards convexity and their association with flame structures are diagnostic. Contortions within large load balls can be similar to those of slumps (see §9.2.3), but the latter involve lateral movement, which is often indicated by a preferred orientation of folds. With loading, the dominantly vertical movement gives a random fold orientation. In addition, loading is normally confined to one pair of beds, whereas slumping may involve many or several beds that deformed together as they moved.

Sand and mud volcanoes

These relatively rare structures occur most commonly in sandstones, typically where they are interbedded with mudstones. The volcanoes themselves are usually of medium- or fine-grained sand, although in some cases they are of silt and mud. They occur on upper bedding surfaces and are also seen in vertical section. Volcanoes commonly overlie units showing extensive post-depositional disturbance such as loading, convolute bedding, sand- and mud-dyke intrusion and evidence of slumping and sliding.

On upper bedding surfaces, sand volcanoes are conical or dome-like, ranging in diameter between 10cm and several metres and being up to 50cm high. They commonly have a crater-like depression in their centre and their flanks may carry radial sand lobes up to a few centimetres wide and with rounded ends (Fig. 9.8). Vertical sections show internal inclined layering parallel to the flanks. In the central zone, a plug or pipe of structureless sand may underlie the crater and may link with a sand-filled dyke or tube below (Figs. 9.9, 9.10).

Sand volcanoes result from liquefied sand being extruded through a local vent at the sediment surface. The volcanoes are, in effect, localized equivalents of the transposed sand sheets associated with sandstone dykes (see later). Sand volcanoes commonly reflect release of excess pore-water pressure from a liquefied unit, possibly following a shock (e.g. seismic shaking).

Lobes on the flanks record the flowage of liquefied sand. The convex down-slope ends show that the sand flow must have stopped abruptly due to **dewatering** as it flowed. Preservation of lobes suggests that the extrusion of sand volcanoes took place in very low-energy settings, otherwise the sand would have been reworked by waves or currents. In some examples, extruded in higher-energy settings, reworking and partial erosion may be identified. In vertical section, the lenticular shapes of volcanoes and the inclined lamination could be mistaken for depositional bedforms and the inclined layering for cross lamination.

Mud volcanoes are features of the present-day surface in areas where muds have become over-pressured at depth and are typically features of active tectonic zones where sedimentation rates are high (Fig. 9.11). They are

Figure 9.8 Sand volcanoes of varying sizes on the upper surfaces of sandstone beds. Note the flow lobes in examples A, D and F, and the central crestal craters in most examples. H) Reveals the internal structure of a sand volcano in cross section; note the central plug. All examples from the Ross Formation, Upper Carboniferous, County Clare, western Ireland.

Figure 9.9 Internal structure of a sand volcano showing the central plug and the inclined bedding which could, without care, be mistaken for cross bedding. Drawn from a photograph. Ross Formation, Upper Carboniferous, western Ireland.

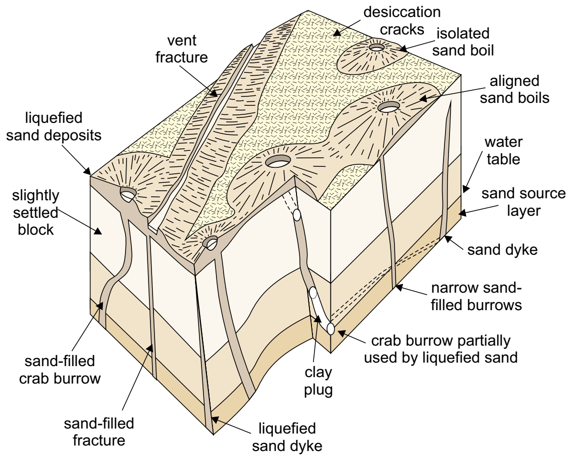


Figure 9.10 Schematic illustration of the range of liquefaction features associated with mixed sand and mud settings. Note the tendency for the liquefied sand to follow pre-existing fissures and conduits. Based on a model for the Tocuyo Delta. Modified after Audemard and Santis (1991).

ephemeral features with a very low potential for incorporation into the rock record.

Patterns of cracks

Patterns of cracks with a variety of scales, shapes and origins occur on present-day sediment surfaces, and on both upper and lower bedding surfaces in rocks. Four types of crack, each of a rather different origin, can be identified.

Desiccation mudcracks

These are common on the floors of dried-up ponds, lakes and playas, on river floodplains and on muddy intertidal and supratidal flats where they commonly take the form of open fissures or are only partially filled by other sediment (Fig. 9.12). In rocks they occur on the bedding surfaces of interbedded, sandstone-mudstone successions, and less commonly in thinly bedded carbonates.

In rocks, the cracks occur in muddy sediment and are infilled by coarser-grained material, usually sandstone. The cracks commonly form polygons from centimetres to metres in diameter, with different size populations sometimes present on the same surface (Fig. 9.12B). Although most crack patterns are broadly hexagonal, many polygons are quadrilaterals or triangles. The cracks are parallel-sided in plan, and in vertical section they usually taper downwards (Figs. 9.13, 9.14, 9.15), although this wedge shape is commonly complicated by later, compaction-induced, buckling or folding. Crack widths range up to several centimetres and depths up to several decimetres. On both present-day surfaces and on bedding surfaces, the areas between the cracks are commonly gently concave-upwards. The surface of a present-day mud layer may be curled up into a strongly concave shape (Fig. 9.16).

Drying out of a muddy sediment layer causes contraction, giving an isotropic, horizontal, tensional stress field that diminishes downwards from the surface. The stress is released by the development of vertical cracks that taper downwards to the level of no effective stress. On slopes steeper than about 5° the crack pattern tends to be rectangular, with one set of cracks parallel to the contours. Clearly, in this case, gravity generates a weakly anisotropic stress field. Cracks may be also localized around earlier disturbances such as footprints.

With homogeneous material, the depths of cracks and the diameters of polygons are directly related; the thicker the cracked layer, the larger the polygons. Thin, surface layers of mud, which dry out rapidly, commonly give small cracks superimposed upon the larger ones that develop during more sustained episodes of desiccation.

Filling of cracks takes place later, for example by the influx of a sediment-laden flood or by wind-blown sand becoming trapped in cracks (Fig. 9.15). Mudflakes, derived from fragmented surface mud layers, may become mixed with this sand. On burial, compaction of muds is greater than that of the sand infills, which respond by folding.

Sand-filled cracks are common as casts on the bases exposed sandstone beds where the underlying mudrock layer in which the original cracks were developed has been weathered away (Fig. 9.12D).

Cracks are common on bedding surfaces that are rippled (Fig. 9.17). The ripple forms are typically indicative of formation in shallow water, which then receded, allowing the exposed surface to dry out and crack.

Figure 9.11 Mud volcanoes, Azerbaijan. Photo courtesy of Martin Bochud.

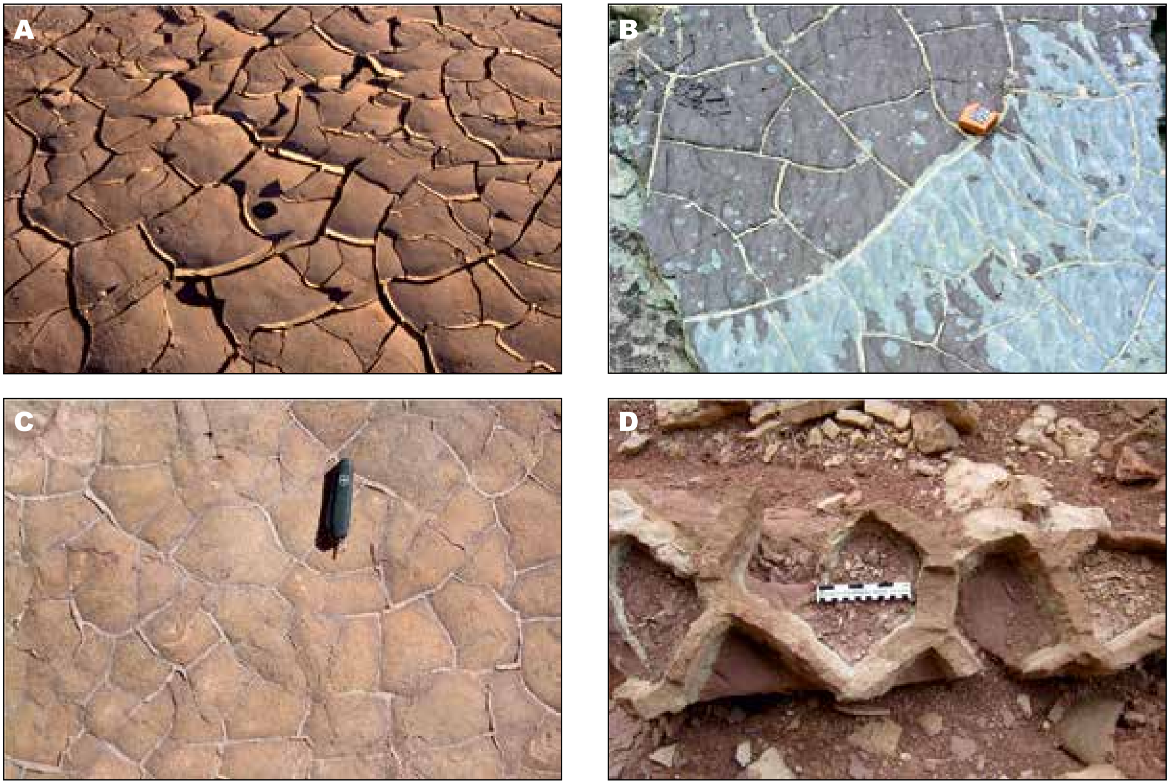


Figure 9.12 Desiccation cracks on bedding planes. A) A mud surface with crack patterns of different sizes, which have different depths of penetration. Modern, Sossusvlei, Namibia. B) Mud cracks on the top of a sandstone bed with weakly defined current ripples. Rensselaer Bay Formation. Late Proterozoic, Inglefield Land, northwest Greenland. C) Bedding surface revealing thinner sand-filled mud cracks in a mudstone bed. Jurassic Kayenta Sandstone, Utah, USA. D) Bedding surface revealing thicker sand-filled mud cracks in a mudstone bed, picked out by strong differential erosion whereby the fill of the cracks has positive relief. Kayenta Sandstone, Jurassic, Utah, USA.

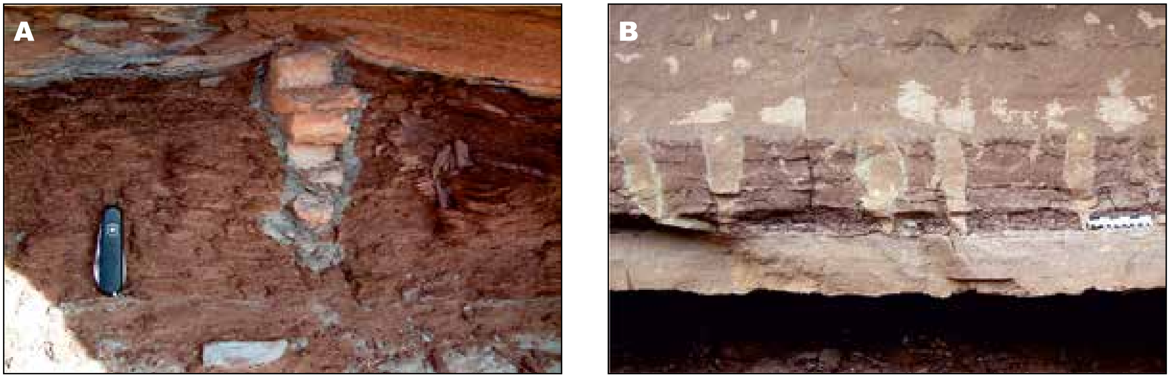


Figure 9.13 Desiccation cracks in vertical section. A) A crack developed in mud and subsequently filled by sand. Note the differential compaction of the mudstone around the crack. Cedar Mesa Sandstone, Permian, SE Utah, USA. B) Downward tapering sand-filled cracks in a mudstone bed. Note the change in colour of the mudstone from red to greenish-grey around the edge of the cracks due to reduction. Kayenta Formation, Jurassic, Utah, USA.

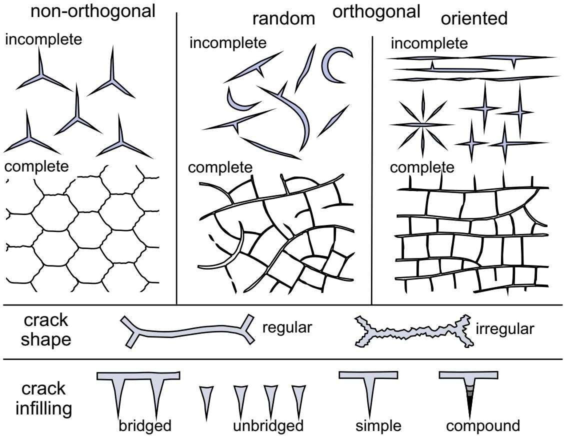


Figure 9.14 Classification of shrinkage cracks and their infillings. Modified after Allen (1982).

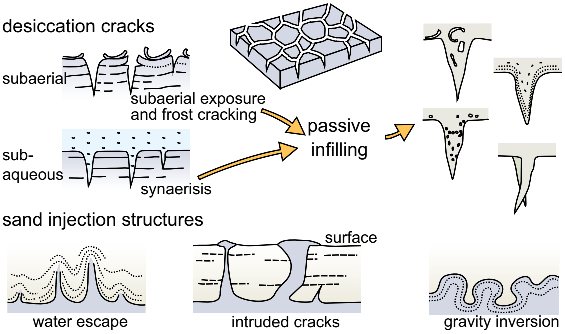


Figure 9.15 Profile sections showing the main features in a variety of naturally occurring unfilled and filled cracks in sediment. Modified after Tanner (1998).

Figure 9.16 Mud cracks in a thin mud layer which is curling at the edges of the cracks to generate potential mud flakes that might themselves be transported in the event of a later current flow over the bed. Note the rain pits on the mud surface. Modern river bed, Tabernas, Almeria Province, southeast Spain.

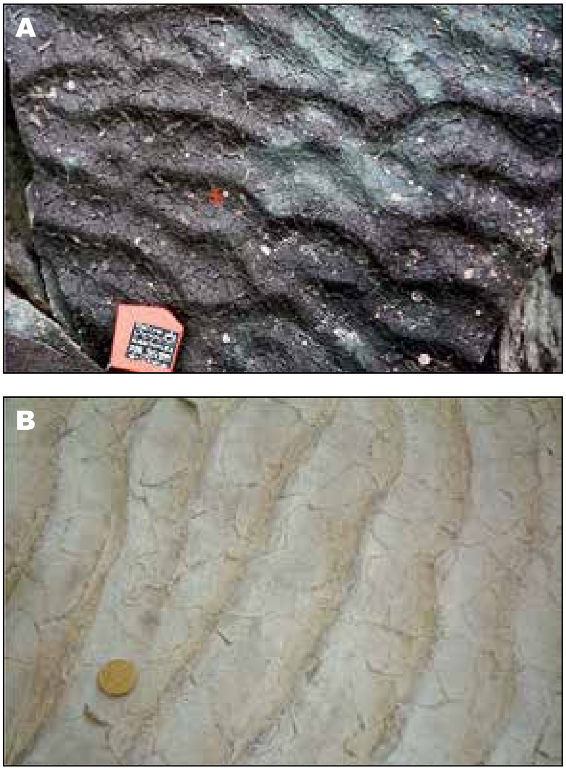


Figure 9.17 Rippled surfaces with mud cracks. A) Sinuous-crested ripple forms on a bedding surface with a pattern of small desiccation mud cracks superimposed. The size of the mud crack polygons suggests that the cracked mud layer deposited on top of the rippled sand layer was very thin. Rensselaer Bay Formation, Late Proterozoic, Inglefield Land, northwest Greenland. B) Near-straight-crested and symmetrical ripples developed in fine sands with a thin veneer of mud that dried and cracked. Note how the mud cracks pass over the ripple crests. Locality unknown. University of Leeds collection.

The concave-upwards surfaces between desiccation mudcracks and the downward tapering of the cracks themselves are useful indicators of "way-up".

Sub-aqueous shrinkage cracks (synaeresis cracks)

These cracks are common on the floors of shallow ponds in present-day salt marshes, but are not restricted solely to these settings. They also occur in mudstones interbedded with sandstone and in some clay-rich carbonate sediments, usually where the beds are thin. The cracks are most common either as positive relief features on bases of sandstone beds (Fig. 9.18A) or in vertical sections through muddy layers where their downward-tapering sandy fills may be contorted by small-scale compactional folding. They also occur as negative relief features on upper bedding surfaces.

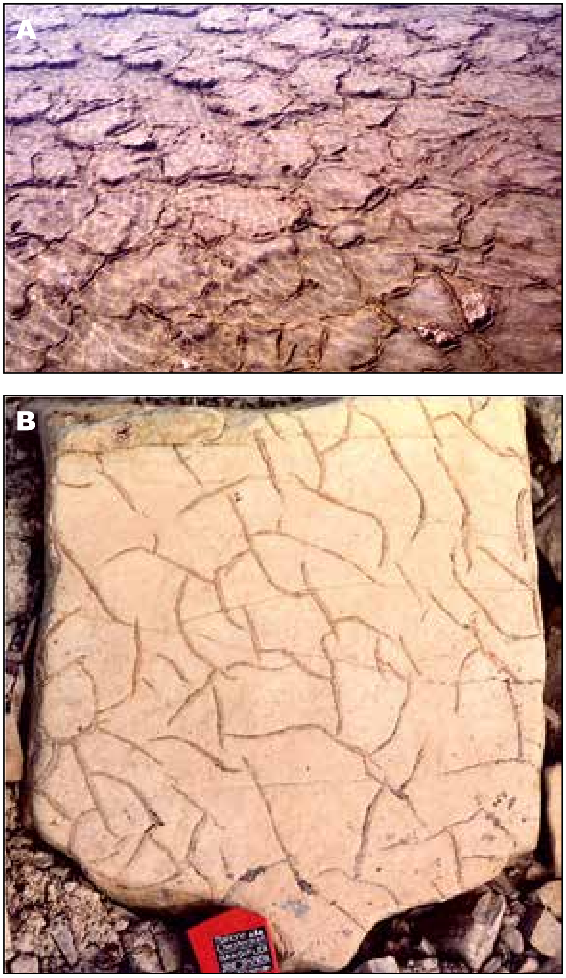


Figure 9.18 Examples of subaqueous shrinkage cracks due to synaeresis. A) On the floor of a shallow water lagoon (water depth is 0.2m). Modern, Las Salinas, Almeria Province, Spain. B) On the upper surface of a calcareous mudstone. Cass Fjord Formation, Cambrian, northwest Greenland.

In plan, sub-aqueous shrinkage cracks tend to have irregular or radiating patterns, in some cases cross-cutting one another. Individual cracks are lenticular, pinching out rather than joining with other cracks (Fig. 9.18B).

Sub-aqueous shrinkage cracks result from loss of pore water from the sediment because of a reorganization of originally highly porous clay particles, either due to flocculation, or because of salinity-induced changes of volume of certain clay minerals. These processes are referred to as **synaeresis**. The conditions under which they take place are not well known, but they occur in a variety of environmental settings and water depths. Marginal marine settings may be particularly favourable because clays there are likely to be subjected to changes in salinity. However, cracks of this type occur in ancient sediments of both marine and non-marine origin. The irregular distribution of cracks may sometimes be due to earlier inhomogeneities, such as burrows. Synaeresis cracks can be useful indicators of way-up and can suggest early post-depositional conditions. However, as way-up indicators, they are typically not as informative as desiccation mudcracks because inter-crack areas are flat.

Synaeresis cracks may be confused with desiccation cracks, with sandstone dykes and with elongate gypsum pseudomorphs. They have also commonly been mistaken for trace fossils, particularly in rocks of late Precambrian age. The most common misidentification is as a burrow, their lenticular shape in plan being similar to the base of a U-shaped burrow. Associated compactional folding has also led to confusion with organic traces.

Sandstone and mudstone dykes and transposed sand sheets

With these structures, crack-filling material is most commonly sand but the host sediment may be anything from mud to coarse gravel. Dykes, though relatively uncommon, occur in a wide variety of settings from deep-water sandstone and mudstone successions to sub-aerial mass-flow deposits. They are seen on both bedding surfaces and in vertical section. **Transposed sand sheets**, which are oriented parallel to bedding and are in some cases associated with the dykes, can easily be mistaken for normally interbedded sandstones or with sandstone sills and their certain identification is possible only with very good three-dimensional exposure.

Dykes occur in a range of sizes, with widths up to several tens of centimetres and vertical extents ranging up to several metres. As with other cracks, the fills may be folded, particularly if the host sediment is finer grained. Dykes can, in some cases, be traced downwards to link with underlying beds of similar lithology. Above they may be truncated by erosion or they may link with an overlying sandstone.

On horizontal surfaces, dykes can occur as positive or negative features depending upon the weathering of dyke and host lithologies. Dykes tend to be straight and rather parallel-sided (Fig. 9.19) and, although usually random in orientation where seen on bedding surfaces, they may show linear trends or polygonal patterns. Mostly they are near vertical but some may be oblique with diverging, cross-cutting and convergent patterns. Some may be sub-horizontal and are better described as sills. Internally they have a variably developed lamination parallel to their

Figure 9.19 Examples of large-scale sandstone dykes. A) Sandstone dyke developed in volcanic rocks and filled with aeolian sandstone. Etjo Formation, Cretaceous, Namibia. B) Sandstone dyke cutting through a poorly sorted conglomerate of probable mudflow origin. A weak lamination parallel to the walls of the dyke was produced by shearing of the liquefied sand during its intrusion from below. Moræneso Formation, Proterozoic, north Greenland.

sides and this tends to be less clearly developed towards the centre.

In good exposure some sandstone dykes may be traced upwards into horizontal sandstone sheets. These may be laterally extensive and, at first sight, can be mistaken for sandstone beds laid down above the dyke-intruded unit. These sheets tend to be rather featureless although they sometimes show gently undulating surfaces. Locally, more intense folding may occur. Internally the sandstones may show an irregular "slurried" texture of weakly defined and rather disturbed lamination.

Sediment dykes and sills result from the injection of sediment from an underlying, or more rarely overlying, source bed. This may have occurred during a short-lived, post-depositional event when both a buried source bed and the host layer were in a weakened condition. Where the intruded host sediment is fine-grained (e.g. sand or silt), and was laid down in relatively quiet conditions, an external shock may have been needed to liquefy temporarily the source bed. Where the host sediment is coarser-grained, its sudden emplacement by mass flow, for example, could have created the conditions necessary to liquefy the source bed. The host sediment commonly has a fairly high content of fine-grained material that would have reduced its permeability. Sediment and water would then be expelled from the liquefied layer through fissures with sufficient force to carry the liquefied sediment, in some cases, to the free surface.

In ancient deep-water sediments, and most commonly encountered in borehole cores, complex arrays of intruded sandstone sheets, involving both dykes and sills, occur in mudstones overlying thick bodies of massive sandstone, commonly channel fills. The intrusion of these **injectites** probably took place significantly later than initial sand deposition and it seems likely that liquefaction of the sand resulted from a gradual overpressuring, driven by the compactional dewatering of the surrounding muds through ongoing burial.

Lamination within the dyke-filling sandstone reflects shearing of liquefied sand as it moved along the fissures. When a fissure reaches the sediment surface, liquefied sand may be extruded and flow laterally before it loses excess water and hence mobility. Such transposed sand sheets occur on some modern debris flows and are occasionally preserved in the rock record. Their undulating bases reflect a tendency of the extruded sand to sink back into the still-mobile debris flow sediment by loading.

In poor exposure, sandstone dykes could be confused with shrinkage cracks, although this is less likely where the host sediment is coarse-grained. Intruded sills could be confused with extruded sands sheets or with normally interbedded sandstones. Internal lamination is good evidence of injection, but confusion could arise from similarity with ice wedges of periglacial areas.

Ice-wedge polygons

Large areas of present-day permafrost show patterns of polygonal cracks of variable plan shape. Similar patterns of cracks are also widely recognized on aerial photographs of areas subjected to permafrost conditions during earlier periods of Quaternary glaciation. Cracks are also seen in vertical sections through glacial and proglacial sediments in these areas. More rarely they are recognized in association with ancient glacial deposits (tillites) in the rock record.

In plan view, polygons range in diameter from about 3m up to several tens of metres and individual cracks are from a few centimetres up to several tens of centimetres wide. Cracks forming at the present day are commonly bordered by ramparts of host sediment that have been pushed upwards (Fig. 9.20).

In vertical section, ancient examples commonly penetrate vertically for several metres and show an upward-flaring, wedge shape which tapers downwards to a fairly sharp lower end. The cracks commonly show quite complex patterns of filling, many with several phases of contrasting lithology occurring in zones roughly parallel to the sides of the cracks (Fig. 9.21). The fill may also contrast to some extent with the host sediment, but coarse clasts, derived from the host, are commonly recognized in the crack fills. Where pebbles are included in the fills, they tend to have their long axes parallel to the sides of the cracks.

Figure 9.20 Present-day ice-wedge with ramparts on either side. Washington Land, northwest Greenland.

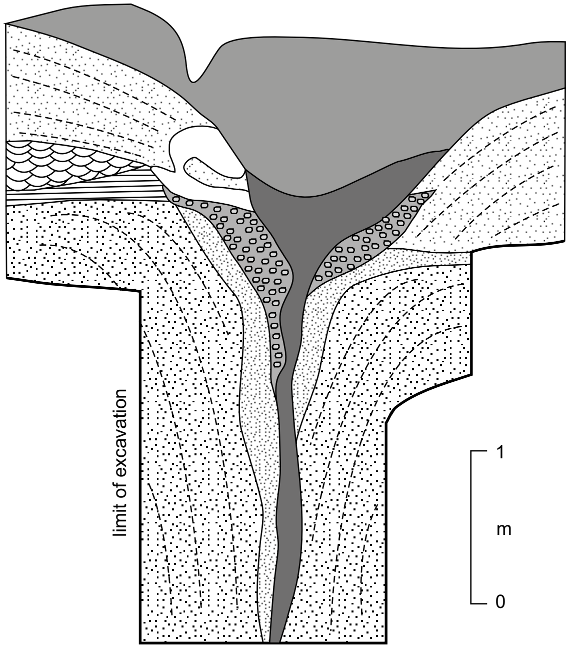


Figure 9.21 Cross section through an ice-wedge in Pleistocene sediments, exposed in a gravel pit. Note the downward tapering of the wedge and the various zones of infill. After Gruhn and Bryan (1969).

The cracks result from the thermal contraction of the host sediment in extremely cold conditions. The tensile strength of the host sediment is exceeded and a crack pattern develops. Water or loose sediment filters downwards into the crack from the active surface layer, which thaws out in the summer. Sediment may help to wedge open the crack, and water collecting there may help crack development by expanding on subsequent freezing. Many wedges in present-day settings have ice masses below them.

Ice wedges, where present, are clearly of considerable palaeoclimatic significance, and therefore great care should be taken to distinguish these features from other types of cracks.

Raindrop impressions

Some upper bedding surfaces in ancient mudstones, siltstones and sandstones and many present-day muddy or

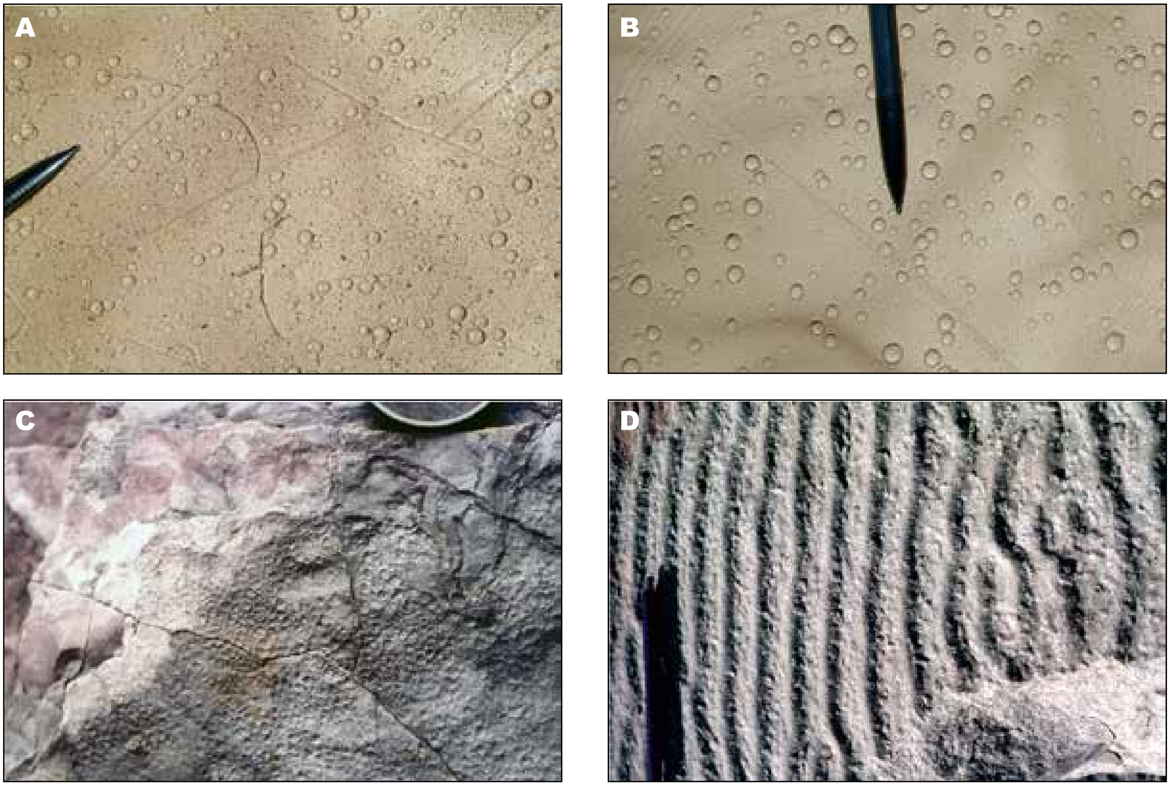


Figure 9.22 Examples of raindrop impressions (rain pits). A) and B) Modern rain pits on a mudstone surface. Tabernas, Almeria Province, southeast Spain. C) The upper bedding surface of a siltstone. West Bay Formation, Lower Carboniferous, Nova Scotia, Canada. D) Rain pits preserved on a rippled bedding surface. Location unknown.

sandy surfaces show patterns of shallow pits. These commonly occur in conjunction with desiccation mudcracks. Pits may be widely separated or may completely cover the surface. They are circular or, rarely, elliptical in shape, up to about 1cm in diameter and up to a few millimetres deep (Fig. 9.22). Where they completely cover a surface, they give a polygonal network. They have a slightly raised rim and the floor of each pit gives a smooth, concave-upwards crater-like form.

Large raindrops and hailstones impact with considerable force and produce small craters on damp sediment. Preservation in the rock record is only likely when the sediment is muddy, as this will have the cohesive strength to retain the impression when it dries out. Sandy surfaces, on drying, are typically reworked by the wind or water.

Rain pits could be confused with trace fossils or with gas-bubble escape features, but the underlying sediments show no traces of disturbed lamination.

9.2.2 Disturbance within individual beds

These structures are most commonly seen in vertical section, although some also have expression in plan view. The structures include those due to deformation of primary depositional lamination as well as new structures developed entirely by post-depositional processes. Thicknesses of disturbed units range from centimetres to metres, and all or only part of a unit may be affected. Both the sediment type and the rate of fluid escape control the style of deformation (Fig. 9.23).

Oversteepened and overturned cross bedding

This structure is common in medium- or finer-grained sandstones laid down in a variety of aqueous environments, although it is also known from aeolian settings.

Overturned or oversteepened cross bedding occurs mostly within single sets. The deformation ranges from foresets that dip more steeply than the angle of rest in the upper parts

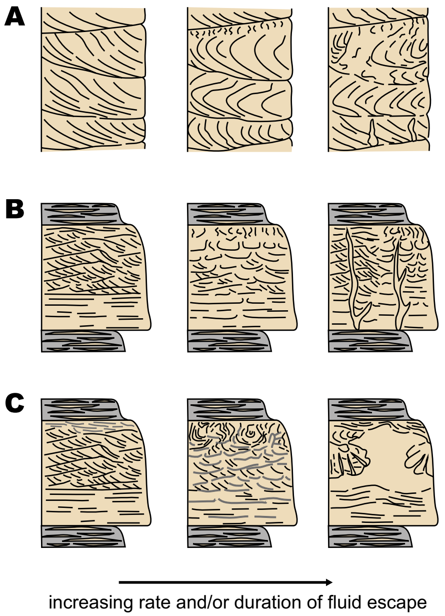


Figure 9.23 Schematic illustrations showing variations in water escape structures as a function of the original sediment characteristics and the rate of fluid escape. A) Medium- to coarse-grained cross-stratified sandstones. B) Alternating sandstones and silty mudstones where the sandstone intervals are non-cohesive. C) Alternating muddy sandstones and silty mudstones where the muddy sandstone intervals are cohesive. Examples B and C are common in turbidites, with the fluid escape structures being most pronounced in Bouma intervals C and D (ripple cross-laminated and planar-laminated sandstones and muddy sandstones). For each row, the rate of fluid escape increases from left to right, giving rise to a variety of structures ranging from minor disturbance (low fluid escape rates), through small scale dish structures, small water escape pipes and convolute bedding (moderate fluid escape rates), to more general liquefaction, loss of lamination and the development of strongly penetrative vertical fluid escape pillars (high fluid escape rates). Modified after Lowe (1975).

of sets (oversteepened) to extensive overturning of foresets (Fig. 9.24) into recumbent folds. The overfolding is always in the direction of the original foreset dip, and the intensity of oversteepening or overturning usually increases upwards through the set. The position of the fold axis within sets is variable and axial planes are inclined to the horizontal.

In addition to simple folds, some single sets show more complex folding which may also involve normal and reverse faulting, loss of definition of the foreset lamination, and enclosure of deformed blocks in structureless masses (Fig. 9.25).

For simple oversteepened or overturned foresets, the processes are fairly straightforward. A shear force acting on the upper surface of a bedform, and in the same direction as the current that produced the cross bedding, deformed the sediment which temporarily had lost strength. In subaqueous situations, the shear force seems, in most cases, to have been the water current that produced the cross bedding in the first place. The sand was probably weakened by partial liquefaction, which happens quite readily in rapidly deposited sands. If the whole set were partially liquefied either spontaneously or by shock, the repacking of grains into a more stable configuration would begin at the base and move upwards through the set as a **front of reconsolidation**. The upper parts of the set would, therefore, be susceptible to the deforming shear stress for a longer time than the lower parts and this would give rise to the observed fold shapes through essentially laminar flow.

Where the fold pattern is more complex, the above mechanism may have played a part, but forceful upward escape of water may also have caused deformation into more upright folds. Internal buckling, similar to that which produces convolute lamination (see below) may also have been involved. In cases where deformational structures include minor faulting and patches of structureless sand, the cross bedding may have originally been aeolian (Fig. 9.25).

Convolute bedding and lamination

These structures occur commonly in single beds of sand or silt in a wide range of environmental settings. They are most commonly recognized in vertical section, but are also seen on bedding surfaces and may be associated with water- and sediment-escape structures such as sand volcanoes.

Convolute bedding and convolute lamination are size-related terms for similar features, the former at the scale of decimetres or bigger, the latter at the scale of centimetres (Fig. 9.23). However, the terms are used loosely and there is no agreed or physically significant size limit. The structure involves folding of lamination, commonly into upright cuspate forms with sharp anticlines and more gentle synclines (Figs. 9.26, 9.27). Overturning of fold axes is seen in some cases, commonly with a preferred orientation. It is usually possible to trace laminae through the folds and it may sometimes be possible to detect original cross lamination within the folded sediment.

Convolution usually increases in intensity upwards through a bed from undisturbed lamination at the base (Fig. 9.27A). At the top it may either die out gradually or

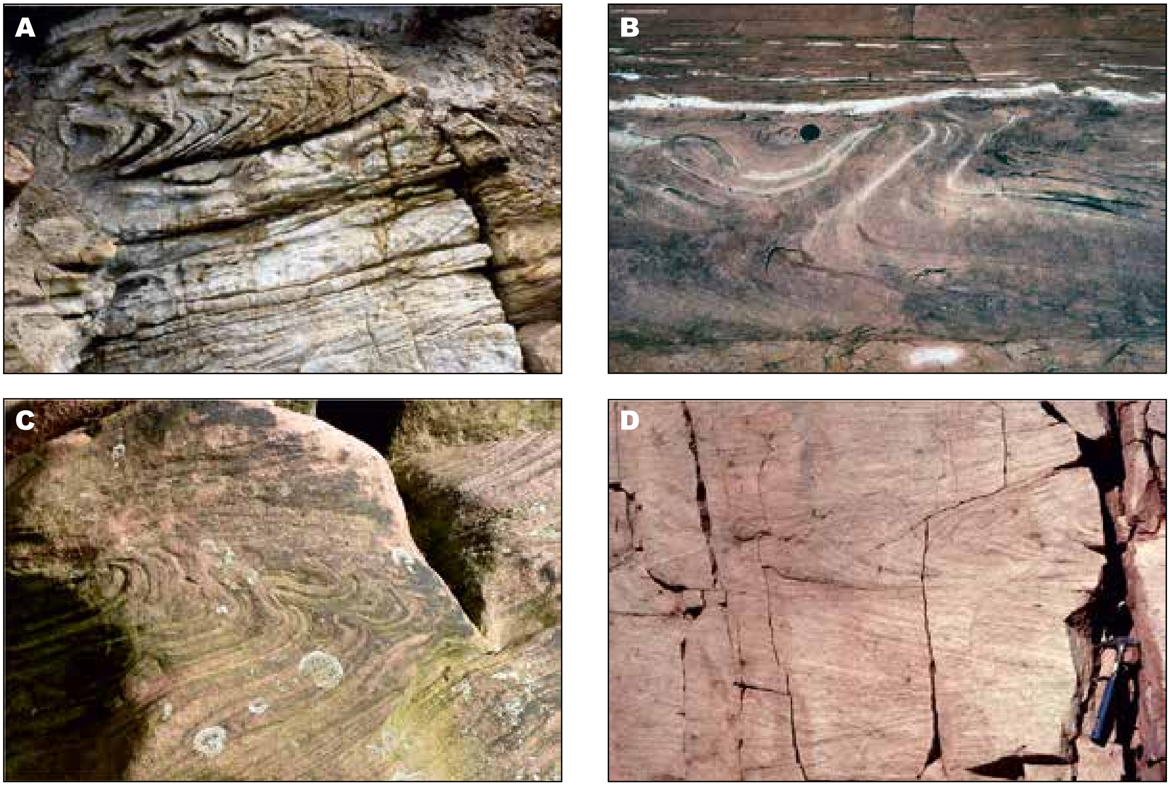


Figure 9.24 Examples of overturned bedding. A) Cross bedding with overturned foresets. Kap Holbæck Formation, Lower Cambrian, northeast Greenland. B) Overturned bedding in aeolian interdune strata due to loading by overlying dune. Cedar Mesa Sandstone, Permian, Utah, USA. C) Cross bedding with overturned foresets. Roaches Grit, Upper Carboniferous, Staffordshire, England. D) Set of overturned foresets. Independence Fjord Group, Proterozoic, northeast Greenland.

be sharply truncated. On upper bedding surfaces, convolute lamination commonly takes the form of a complex pattern of basins and ridges.

Convolution involves plastic deformation of partially liquefied sediment, usually occurring soon after deposition. The common presence of convolute lamination in turbidite sandstones, and just below the sediment surface in present-day river floodplains and tidal flats in seismically quiet areas, suggests that liquefaction can be spontaneous as well as externally triggered. On tidal flats liquefaction may be aided by breaking waves during emergence of the bed or by the rise and fall of the water table through the sediment. Where axial planes of folds have a preferred direction of inclination, this commonly coincides with the palaeocurrent, suggesting that convolution formed during deposition. In aeolian environments convoluted bedding may occur as water is squeezed out of sediment in front of an advancing dune (Fig. 9.28). Careful examination of the style of truncation of bedding or lamination within the deformed set may indicate whether the deformation occurred during or after accumulation (Figs. 9.29, 9.30).

Dish, pillar and sheet dewatering structures

The significance of these structures was recognized relatively recently. They were originally recorded from thick turbidite sandstones (and this remains their most common occurrence), but additionally they are now recognized in shallow-water sandstones and in volcanic ash layers. The host sediment ranges from coarse silts to coarse, even pebbly, sand and fine gravels. Dish structures also appear to require the presence of a certain amount of clay for their development. The sandstones and siltstones are usually discrete beds at least decimetres or even metres thick, in which dish and pillar structures are commonly the only structures present. The beds elsewhere appear massive. In some cases, the structures deform and cut across earlier lamination. Dish and pillar structures are usually poorly defined and they need rather exceptional exposure and weathering to

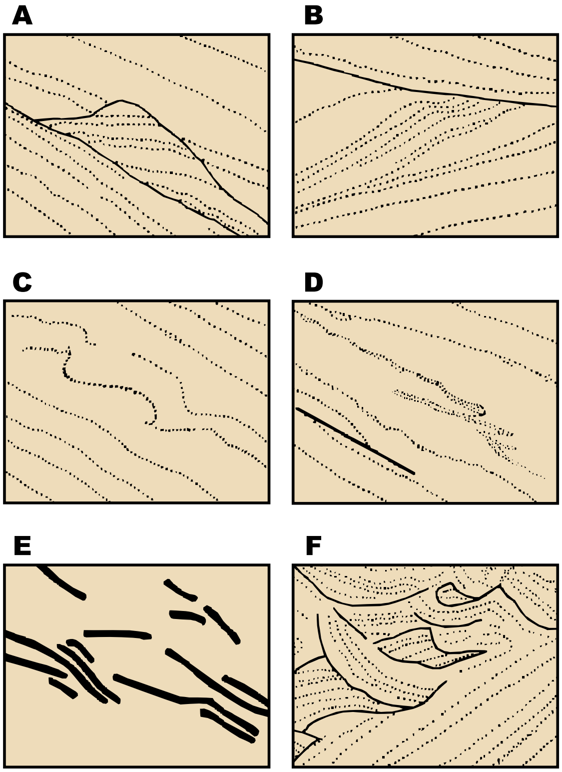


Figure 9.25 Principal types of deformation structures in the foresets of dunes. A) Rotated structures. B) Drag folds. C) High-angle asymmetric folds D) Overturned folds. E) Break-apart structures. F) Brecciated foresets. Modified after McKee (1979).

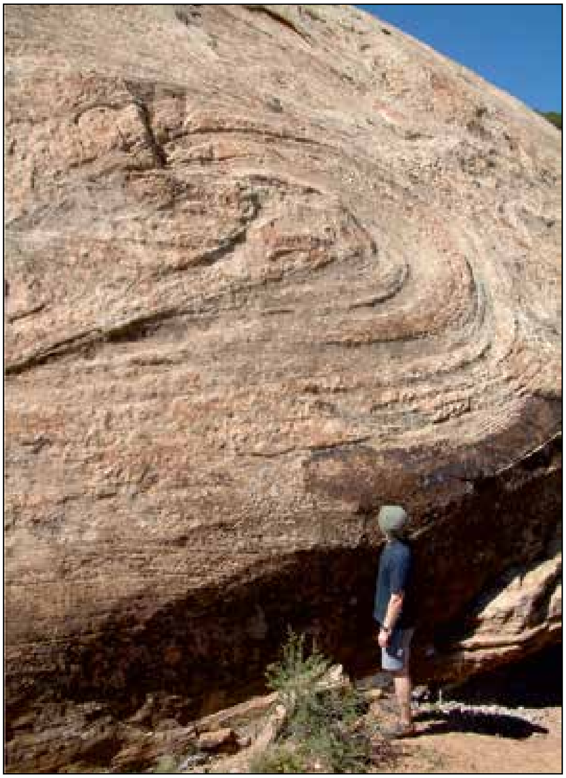


Figure 9.26 Large-scale set (~10m thick) of aeolian dune origin with cross-bedding that has been deformed such that the foresets are overturned. Navajo Sandstone, Jurassic, Utah, USA.

stand out clearly. Only after seeing good examples is it realistic to look for these structures in less favourable exposures. Sheet dewatering structures are rather more obvious and common.

Dish and pillar structures commonly occur together in the same bed. In vertical section, **dish structures** appear as thin, roughly horizontal zones, in some cases flat but more usually concave upwards (Fig. 9.31). Each is defined by a dark, clay-enriched zone up to around 2mm thick which contrasts with the paler host sandstone on either side. Plan views show that the three-dimensional shape consists of shallow dishes a few centimetres across, defined by concentrations of clay and platy mineral grains. The darker zones which define the dishes commonly have a vertical spacing of a few millimetres up to about 10cm.

Flat zones lacking the clear dish shape, have upturned ends associated with penetration of layers by **pillar structures** (Fig. 9.31B). These extend vertically for several centimetres and may pass through several of the horizontal zones and dishes. They have a core of cleaner sand with poorly defined, darker, clay-rich fringes and in plan are circular with a diameter of a few centimetres.

Sheet dewatering structures are sub-vertical sheets, up to a few millimetres wide, commonly arranged in a parallel fashion. They are continuous and more linear in plan and usually occur near the tops of structureless sandstone beds (Fig. 9.32). They are not necessarily associated with dish structures.

All these structures may coexist within a bed, and they are sometimes found in association with convolute lamination and ball and pillow structures. This suite of structures may sometimes show a crude vertical zonation (Fig. 9.33). The sequence of deformation events may commence with the development of mild convolute lamination, prior to dish, pillar and sheet structures developing higher in the set and more intensely-deformed ball and pillow structures overprinting the convolute bedding in the lower part of the set.

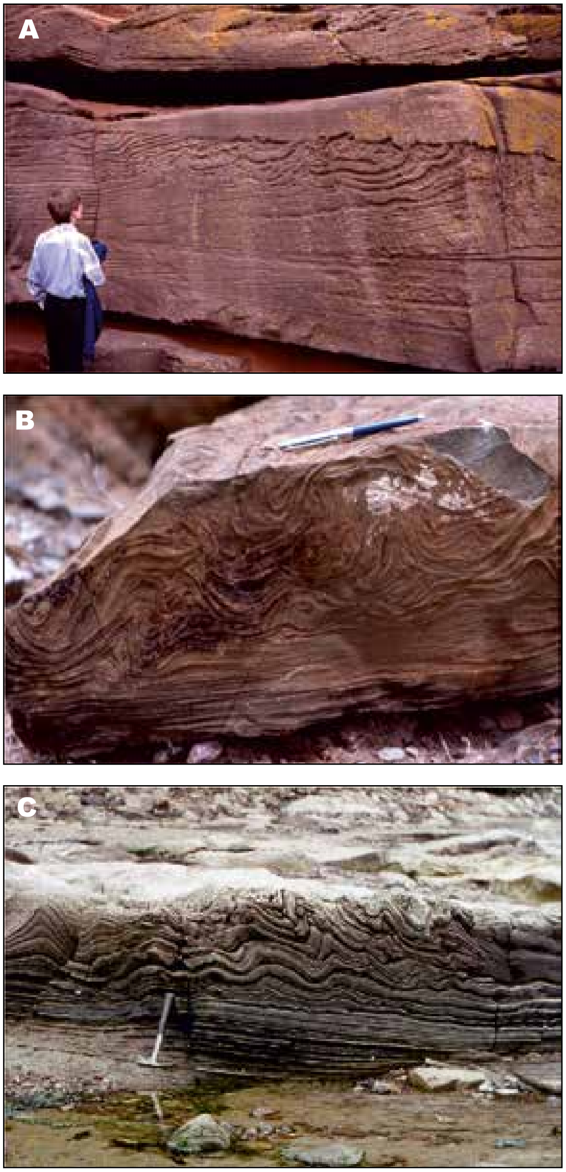


Figure 9.27 Examples of convolute bedding and lamination. A) Convolute bedding within the upper part of a cross-bedded sandstone set. St Bees Sandstone Formation, Triassic, Cumbria, England. B) Convolute lamination within ripple cross-laminated limestone. Izroutene Beds, Carboniferous, Morocco. Photo courtesy of Gilbert Kelling. C) Convolute bedding. Lower Carboniferous, Dunbar, Scotland.

The three types of structure are all due to the post-depositional escape of water from the sand. The vertical structures (the sheets and pillars) are conduits of water expulsion, similar to sandstone dykes and the feeder pipes

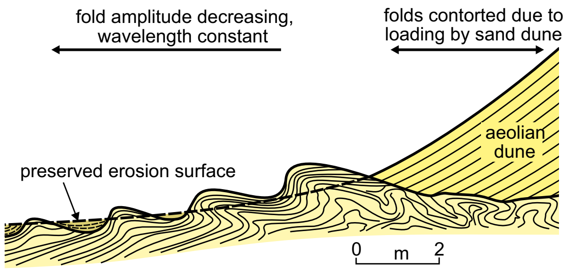


Figure 9.28 Schematic illustration of the development of contortion in a siltstone unit as the result of the advance of a large aeolian dune across its surface. Lnagra Formation, Upper Devonian, central Australia. Modified after Maltman (1994).

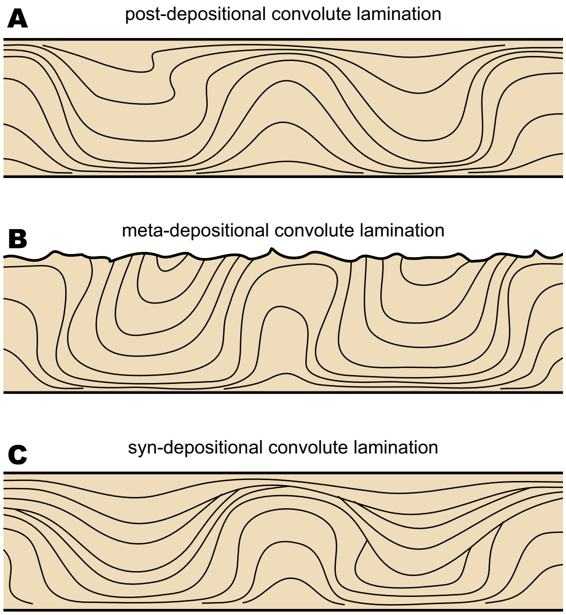


Figure 9.29 Types of convolute lamination. After Allen (1982).

of sand volcanoes but only restricted and selective movement of sediment (Fig. 9.10). The force of upward water movement causes local fluidization of the sediment through which it passes and also the selective winnowing of fine particles. This results in the slightly cleaner sand seen in the pillars and sheets.

The processes of formation of dish structures are less obvious and have to be inferred from the structures themselves. Their intimate association with pillars suggests a linked origin. Dish structures are thought to be produced by slower and, to some extent, horizontal water movement, restricted and controlled by semi-permeable, flat-lying barriers that probably started as weakly defined depositional

Figure 9.30 Package of chaotically deformed mixture of sandstone and mudstone that has undergone pervasive loss of strength and foundering through a combination of loading, mud injection and water escape. Kayenta Formation, Jurassic, Utah, USA.

laminations. The upward movement of water through the loosely packed sand soon after deposition is retarded by the laminations, and more fine particles are added to them, further reducing their permeability. Some of the escaping water is forced horizontally below barriers until it finds an easier route of upward escape, at a pillar or the upturned edge of a dish. This water movement probably takes place very soon after, or even during, rapid deposition that likely gave rise to very loose grain packing and a high initial porosity.

The relative rarity of these structures limits their use as way-up indicators although, where present, the upward concavity of dishes could be helpful. The lack of dish and pillar structures in the majority of massive sandstones probably reflects a lack of suitable clay rather than a lack of rapid de-watering.

Dish structures could be confused with trough cross lamination but foreset laminae are not present in each "trough" and the plan shape is different, so this pitfall usually can be avoided. Pillar and sheet structures could be confused with vertical burrows but the upward-turning of the flanking layers is not common in burrows.

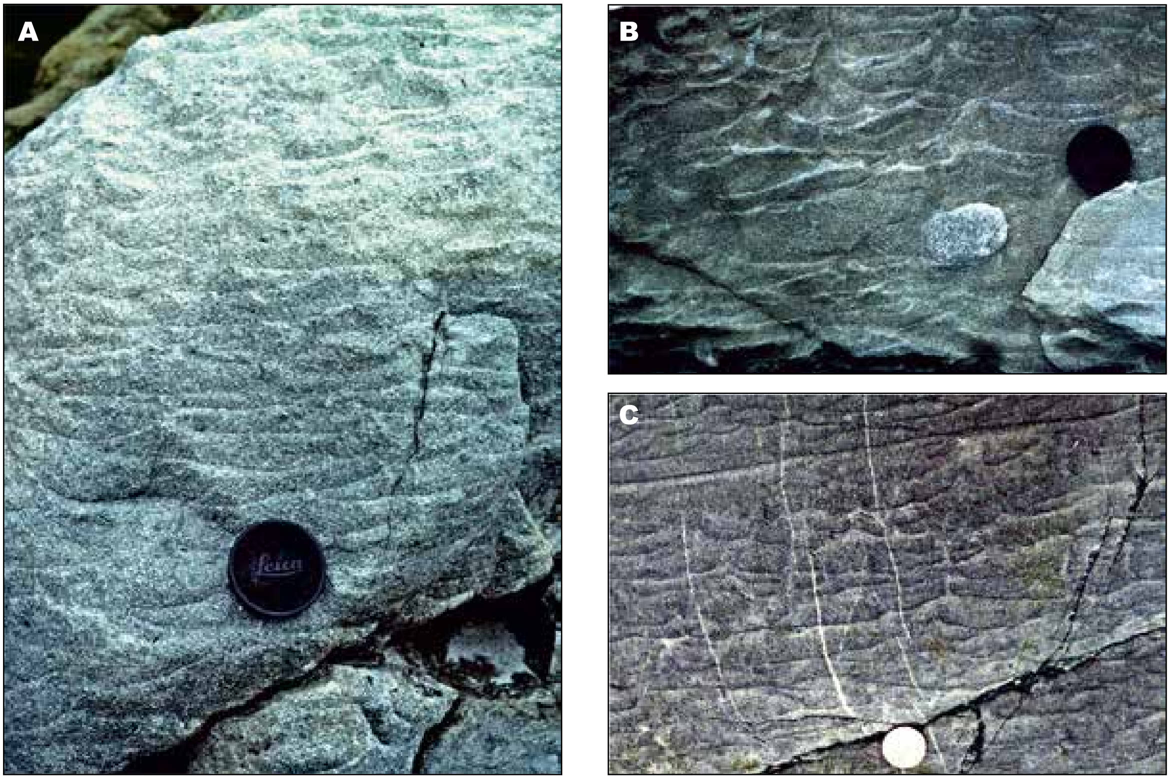


Figure 9.31 Examples of dish structures. A) and B) Dish structures in turbidite deposits, Eocene, Pasaia, northern Spain. C) Dish structures penetrated by vertical pillars and pipes which acted as conduits for vertical water escape. Ordovician, Gaspee, Canada.

Figure 9.32 Sheet dewatering structures in sandstone. Lower Cretaceous, Spitzbergen, Svalbard.

9.2.3 Disturbance affecting several beds

Some examples of larger-scale structures may be recognized at outcrop but others may require a larger scale of observation, possibly involving mapping. With structures such as slumps, which result from lateral mass movement, it would be equally valid to regard them as being depositional or, at least, "redepositional" in origin. It can be difficult to establish the timing of deformation from large-scale soft-sediment disturbance structures. It can also be difficult to distinguish tectonic from sediment-induced deformation in some cases.

Slumps: sedimentary folding

Units of folded sediment attributable to slumping usually occur in interbedded successions containing a substantial proportion of fine-grained sediment. The successions may be composed of detrital or carbonate material and the fine-grained sediment may be clay or lime mud. Slumped units typically occur on or at the bases of contemporaneous sub-aqueous slopes in a variety of environments.

Slump-folded units vary in thickness from less than one metre up to tens or even hundreds of metres. They are most commonly bounded above and below by undisturbed sediment and this helps to distinguish them from tectonically disturbed beds (Fig. 9.34). Within the slumped unit, folds may have preferred orientations which, when properly plotted on a stereonet (see Appendix 1), may help to indicate the direction of the palaeoslope. It is important to record the style and scale of the folding, the thickness of the deformed unit as a whole, and the thicknesses of beds within it. Systematic recording and comparison of the orientations of fold axes and axial planes with those of any tectonically produced structures (folds, cleavage, etc.) may

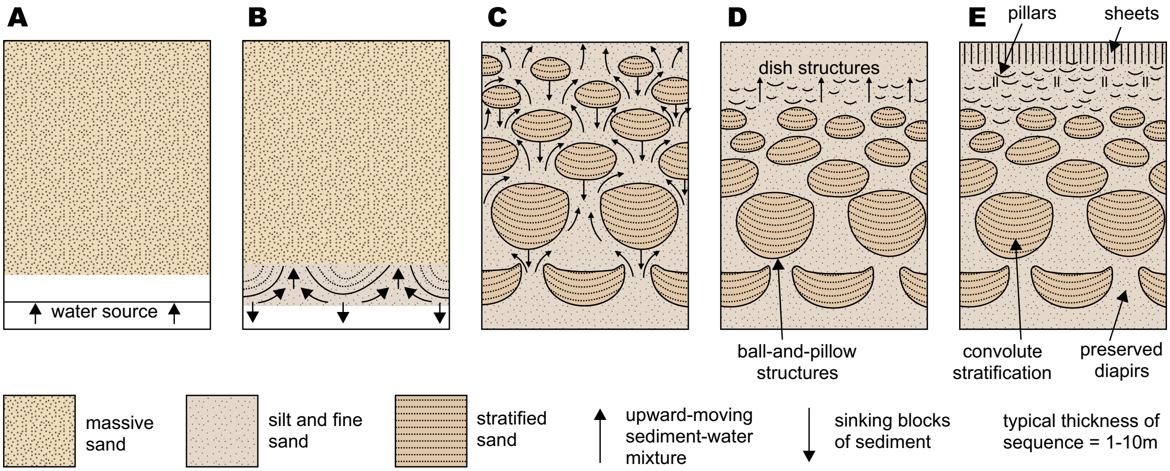


Figure 9.33 A model for the progressive development of convolute bedding, ball-and-pillow structures and dish structures. Excess pore fluid pressure drives the upward escape of sediment and water. Gravity assists with the downward movement of larger intact blocks. Modified after Cheel and Rust (1986).

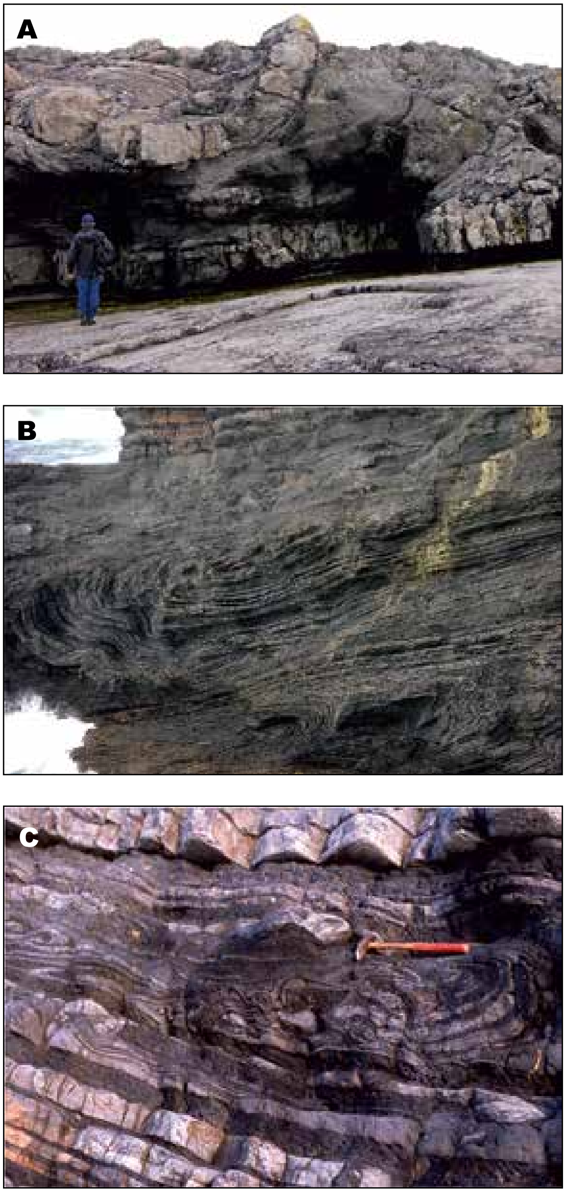


Figure 9.34 Examples of physical soft-sediment deformation affecting several beds. Interbedded sandstone and mudstone successions within which a group of beds have been folded as a result of slump movements. The folding is seen to be syndepositional by the undisturbed nature of the bedding above and/or below. A) Basal slip surface of slump at foot of face. Ross Formation, Upper Carboniferous, western Ireland. B). Intense folding in several stacked major slumps within thick silty sandstones. Gull Island Formation, Upper Carboniferous, western Ireland. Face about 30m high. C) Interbedded slump unit. Aberystwyth Grits, Lower Silurian, Wales. Photo courtesy of Gilbert Kelling.

help to differentiate between tectonic and soft-sediment disturbance.

Unconsolidated sediments, resting on a slope, may become unstable, possibly due to high pore-fluid pressure in a particular layer. Sediments above this weakened layer may then move down slope under gravity as a coherent mass. In some cases, an intact slab of material may detach and move. In other cases, the down-slope end may stay fixed as the up-slope end moves towards it. In both cases there are significant differences between the behaviour at the up-slope and down-slope ends. The up-slope end is subject to a dominantly tensional stress regime, whereas at the down-slope end compressive stress dominates. However, the deformation seen in slumped sediments is generally complex and must be interpreted with some caution. Slump folds commonly reflect a compressive regime in a down-slope position with fold axes normal to the direction of movement. However, folds also result from lateral compression and from internal shearing between components of the slump complex moving at different rates. Such shearing commonly leads to fold axes being rotated so that they lie parallel to the movement direction. Even where folds seem to be compressive, it is important to remember that most slumps are lobate in plan and that fold axes can be expected to show a spread of directions. Up-slope zones of slumps, dominated by tensional strain, commonly show a different suite of structures, as described below.

Slump folding helps our understanding of processes at or soon after deposition. Deformation produced by genuine slumping, involving lateral displacement of material, may be confused with that due both to vertical sinking (i.e. loading) and to tectonic deformation. Vertical sinking, as a rule, produces little or no preferred orientation of fold axes. With dominantly tectonic deformation, resulting folds commonly relate to or mirror larger-scale structures of undoubted tectonic origin within the local area.

Rotation and displacement of coherent blocks

Displaced relationships between blocks of internally coherent sediment occur on a variety of scales, in a variety of settings and in sediment of virtually any composition. Although many tectonic and sedimentary breccias could be placed under this heading, with many large-scale examples grading into tectonic structures, we deal here with small- and medium-scale examples where movement has been confined to discrete slip planes. Such structures can all be categorized as **synsedimentary faults** whose displacements can vary from centimetres to many hundreds of metres.

Features of this type are so variable that no comprehensive and systematic account is possible. A field description of any suspected example should be made with several questions in mind. However, in most cases, they will not all be answered fully and unambiguously.

- *Is the displacement of early, post-depositional or syndepositional origin or is it tectonically produced after lithification?* Where syndepositional, the disturbance may be confined between undisturbed units above and below, and the surfaces of movement will be sharp with little or no brecciation. Tectonic faults, produced after lithification, are usually accompanied by quartz or calcite veining and associated brecciation.
- *Did the movement take place close to the contemporaneous sediment surface and thereby create a topographic feature?* Draping of a topographic step, or onlap onto its flanks, by overlying sediment, lateral changes of thickness and the gradual upward elimination of relief should be looked for (Fig. 9.35).
- *What are the shape, size and spacing of the surfaces of displacement?* Pay particular attention to the vertical and lateral extents of the surfaces. Are they planar or concave upwards? Many small-scale synsedimentary faults are planar with small throws, whereas larger faults, particularly in mud- or silt-prone successions, are concave-upwards and pass down dip into bedding-plane faults.
- *Is there any evidence that movement took place during deposition or was the displacement a discrete event, followed by subsequent deposition?* Here it is necessary to look carefully at patterns of thickness change across the fault (Fig. 9.36).
- *Have the various blocks undergone any rotation in the course of displacement?* Careful comparison of dips on either side of displacement surfaces will help (e.g. Fig. 9.37A & B).
- *Are there any minor structures such as drag folds and smaller faults that help to elucidate the nature of the movement?*
- *Is there a hierarchy of faults of differing magnitudes and, if so, do minor faults show synthetic and/or antithetic geometries to the major faults?*

Be sure to distinguish those folds related to fault movement from those that pre-date the faulting. Some slumps, which develop folds during ductile deformation, are cut by faults due to later phases of brittle failure.

Figure 9.35 Examples of large-scale deformation whereby slumping is associated with sliding. A) Folding in slumped siltstone unit between intervals of undisturbed turbidite sandstones. Note the way that the upper surface of the slump is rapidly restored to a horizontal surface. Slide surface is on top of the thick sandstone at the base. Ross Formation, Upper Carboniferous, western Ireland. B) Thick, highly disturbed interval of thinly bedded siltstones and fine sandstone with major discontinuities related to the movement of several slumps and slides. Face about 40m high. Gull Island Formation, Upper Carboniferous, western Ireland.

Syn-sedimentary faults occur for a variety of reasons. Most have a normal throw indicating local tensional stress. It is not possible to give an exhaustive account of the ways in which these structures occur but the following illustrations may be helpful.

Slump scars

It was suggested earlier that the up-slope end of a slump sheet would be a zone of overall extension. If the sheet moved off in its entirety, it would leave an empty slump

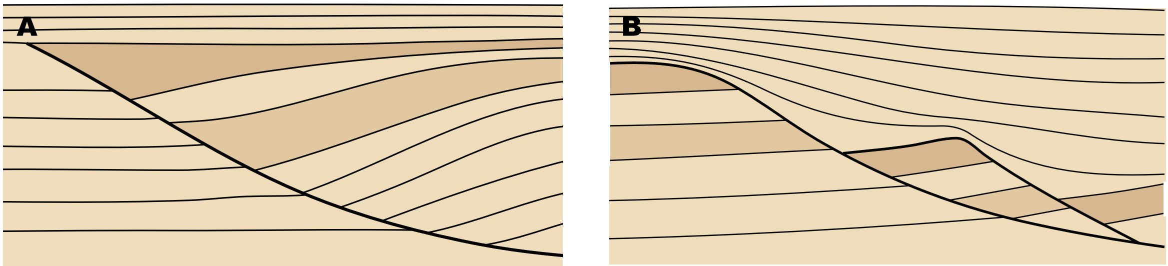


Figure 9.36 Schematic sections through synsedimentary faults. A) The fault movement has continued during deposition. B) Fault movement has occurred to the extent that a topography developed on the sediment surface. This was then draped by later deposits.

Figure 9.37 Syn-depositional faulting at a range of scales. A) Sedimentary growth fault within a deltaic succession. A wedge of sandstone that thickens and is folded in the hangingwall of the fault is absent on the footwall. Horizontal beds are re-established directly above, demonstrating the syn-depositional nature of the faulting. Central Clare Group, Upper Carboniferous, western Ireland. Cliff is ~35m high. B) Sandstones in the hangingwall of an inferred syn-depositional growth fault. The position of the inferred fault is just the left of the photograph. The well-bedded sandstones in the lower part of the face thicken towards the left (i.e. towards the fault) and show gentle folding, a pattern typical of the hangingwall of an active growth fault. The lens of thinly bedded sandstones in the upper part of the face thin away from the fault and show low-angle, large-scale cross bedding, recording the progradational infill of a hollow created by continued movement of the fault after the main sand supply ceased. Middleton Grit, Upper Carboniferous, Airedale, Yorkshire, England. Cliff is ~20m high. C) Curved trace of a syn-depositional normal fault on top of a sandstone bed. Such curved traces are typical of many listric faults that flatten with depth. Cretaceous, Spitzbergen, Svalbard. D) Small syn-sedimentary normal faults forming an anastomosing network, similar to that seen at larger, seismic scales associated with extensional rift basins. These sands were faulted as a result of extension on the flanks of a rising mud diapir. Central Clare Group, Upper Carboniferous, western Ireland.

scar, which would then be draped and filled by later sediment (Fig. 9.36B). However, at many slump scars a series of slip surfaces develop. Packets of sediment between these surfaces have only small displacements, and are commonly rotated to dip up, relative to the original slope (Fig. 9.37A).

Small-scale gravitational collapse features

Small-scale faulting in sands and sandstone can commonly arise due to the collapse of objects that were buried in the sand. Logs, masses of vegetation or blocks of ice may rot or melt and the sand then collapses into the resulting space, generating small extensional faults.

Growth faults

Many growth faults occur on a scale too large to see at outcrop and their recognition normally requires mapping. These features are commonly revealed by seismic reflection surveys. Small examples have, however, been recognized in extensive exposures (Fig. 9.37). The typical pattern of displacement and thickness change is shown schematically in Figure 9.36A. The gentle rotation and folding of the wedge of thickened sediment in the hangingwall are typical features. Faults of this type occur most commonly in deltaic deposits, particularly those with a high proportion of fine-grained sediment. High depositional rates lead to high pore-water pressure in buried muds with consequent loss of strength.

At the largest scale, where the over-pressured sediment is deeply buried beneath major deltas like the Niger or Mississippi, it may flow towards the ocean like toothpaste due to the weight of the overlying sediments. This ductile movement creates extensional stresses in the overlying sediments which respond in a brittle fashion to give deeply penetrating growth faults that flatten downwards into the ductile layer. At this scale, the entire depositional-deformational interplay operates at time scales of millions of years and the dimensions are such that the faults may only be studied using seismic reflection profiles.

Where overpressuring occurs at shallow depths, thinner bodies of overlying material move towards the basin and the situation is gradational with the extensional parts of slumps. Whatever the scale, displacement of the sediments takes place along curved (listric) fault surfaces that pass downwards into bedding-plane faults. At the depositional surface, differential movement across normal faults accommodates thickening of sediment on the downthrown hanging-wall side. When movement ceases, a fault line may be differentially draped before a more uniform thickness pattern is re-established (Figs. 9.37A & B, 9.38, 9.39A).

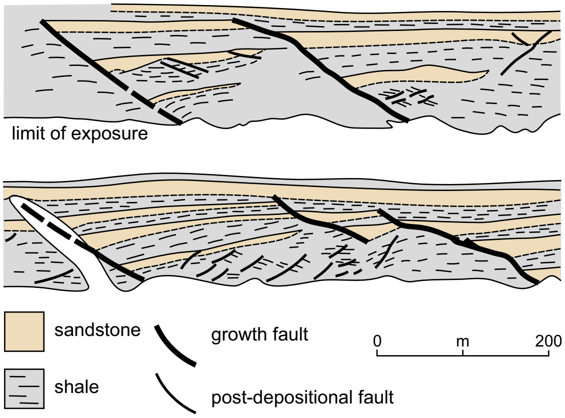


Figure 9.38 A system of growth faults exposed in cliff sections of Triassic sediments on Edgeøya, Svalbard. After Edwards (1976).

The displacement of coherent blocks of sediment through brittle failure along fault planes often occurs in close association with plastic, liquefied and/or fluidized styles of deformation, especially where the lithologies of interbedded successions have variable competence (e.g. sandstone–mudstone layers). Thus, brittle faulting may originate within or pass up into zones of convolute bedding or ball and pillow development (Fig. 9.40). Furthermore, faults may act as conduits for fluid escape and may develop into clastic dykes or feeder pipes for sand or mud volcanoes.

Large-scale gravitational collapse structures – ‘olistostromes’

Large-scale structures relating to gravitational collapse and down-slope movement are well documented from submarine slope settings, where the resultant deposits are often of sufficient extent to form mappable units called **olistostromes**. Olistostromes vary in thickness from less than a metre to a few hundreds of metres and the largest units range in extent up to hundreds of square kilometres. Olistostromes are composed of collections of very large blocks, commonly composed of limestone, with individual clasts ranging from a few metres to hundreds of metres in diameter (Fig. 9.39B). Blocks are commonly striated. These blocks, which occur in both clast-supported and matrix-supported frameworks, most commonly sit chaotically arranged within a fine-grained, typically muddy matrix. Olistostromes are most commonly found interbedded within fine-grained turbidite successions. Where the blocks are smaller, they may be arranged into a crude layering, for which the term **olistolite** is sometimes used. Rarely, units comprise single, very large-scale blocks called

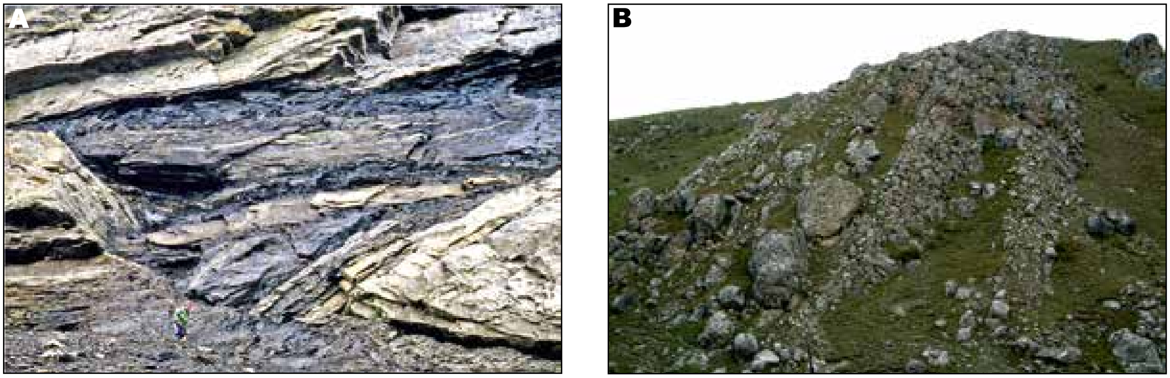


Figure 9.39 Examples of physical deformation due to the rotation and/or displacement of coherent blocks. A) Large rotational slide of a thick sandstone bed. The thick sandstone on the left and its underlying fine-grained sediments are *in situ*. The block on the right has slid along a narrow, concave-upwards shear zone and has rotated back into the zone. Person for scale. Lower Cretaceous, Spitsbergen, Svalbard. B) Limestone blocks forming an olistostrome megabreccia generated by the collapse of the leading edge of a carbonate platform. Lower Cretaceous, Greater Caucasus, Azerbaijan. Largest 2 blocks are each approximately 12m diameter.

olistoliths or **olistoplates** that have clearly undergone displacement from their site of original accumulation.

Olistostromes fall within a family of deposits that are the products of submarine slides driven by gravitational collapse. The large size of the blocks, together with their numerous striated surfaces shows that they were effectively lithified prior to movement. Their occurrence interbedded with fine-grained (commonly turbiditic) sediments suggests their development in marine slope or base-of-slope settings. Where the blocks are of limestone, they may contain shelly faunas and reef structures, indicative of an original shallow-water depositional setting. Thus, many olistostromes are considered to reflect gravitational collapse of the outer edge of a shallow marine shelf and the sliding of blocks of debris into deeper water. This commonly occurs in tectonically active areas, where tilting and earthquake activity are more pronounced.

Diapirs

Diapirs are large-scale structures that develop as a result of density inversions, where a low-density interval underlies a higher-density unit and where the lower unit has behaved in a plastic fashion due to loss of strength. The same principles apply to the development of load and flame structures as described in §9.2.1. However, with diapirs, it is the upward movement of the lower lighter layer that dominates the development, commonly giving rise to a dome-shaped structure that penetrates the overlying succession. The lower-density mobile units are of two contrasting types.

In deltaic successions, thick intervals of mud, rapidly buried by coarser grained sediments, may become over-pressured and lose strength and be less dense than the overlying sediments because of the greater retained pore water. This inherently unstable pattern may then be resolved by the mobile muds pushing upwards, often very rapidly, as mud diapirs, typically 10's of metres in width and height. In some modern deltaic settings mud diapirs penetrate to the surface and emerge as short-lived islands, offshore of the delta front. Where ancient examples are seen at outcrop (Fig. 9.41), the overlying sediments typically thicken into the flanks of the diapir, suggesting that they were being deposited at more or less the same time as the diapir was rising. Where seen in both present-day or ancient examples, the muds typically show pervasive centimetre-scale orthogonal fractures.

Other diapiric structures, usually at a larger scale, occur when thick intervals of evaporites are deeply buried. Such deposits are easily mobilized with increasing pressure and temperature and are almost always less dense than the overlying succession. The resulting density inversion is inherently unstable and the evaporites, most commonly halite, tend to rise to penetrate the overlying succession, usually resulting in diapiric structures that are typically kilometres wide and hundreds of metres high. Some have forms that are dome shaped; others are more wall-like. Where sedimentation took place during diapiric movement, strata may thicken into the flanking areas of the diapir and be folded as a salt-withdrawal syncline. These patterns are best studied in seismic section but in some fairly arid areas, salt domes penetrate the land surface (Fig. 9.42) where their internal structure can be seen to be highly contorted due to the intense plastic deformation that has occurred.

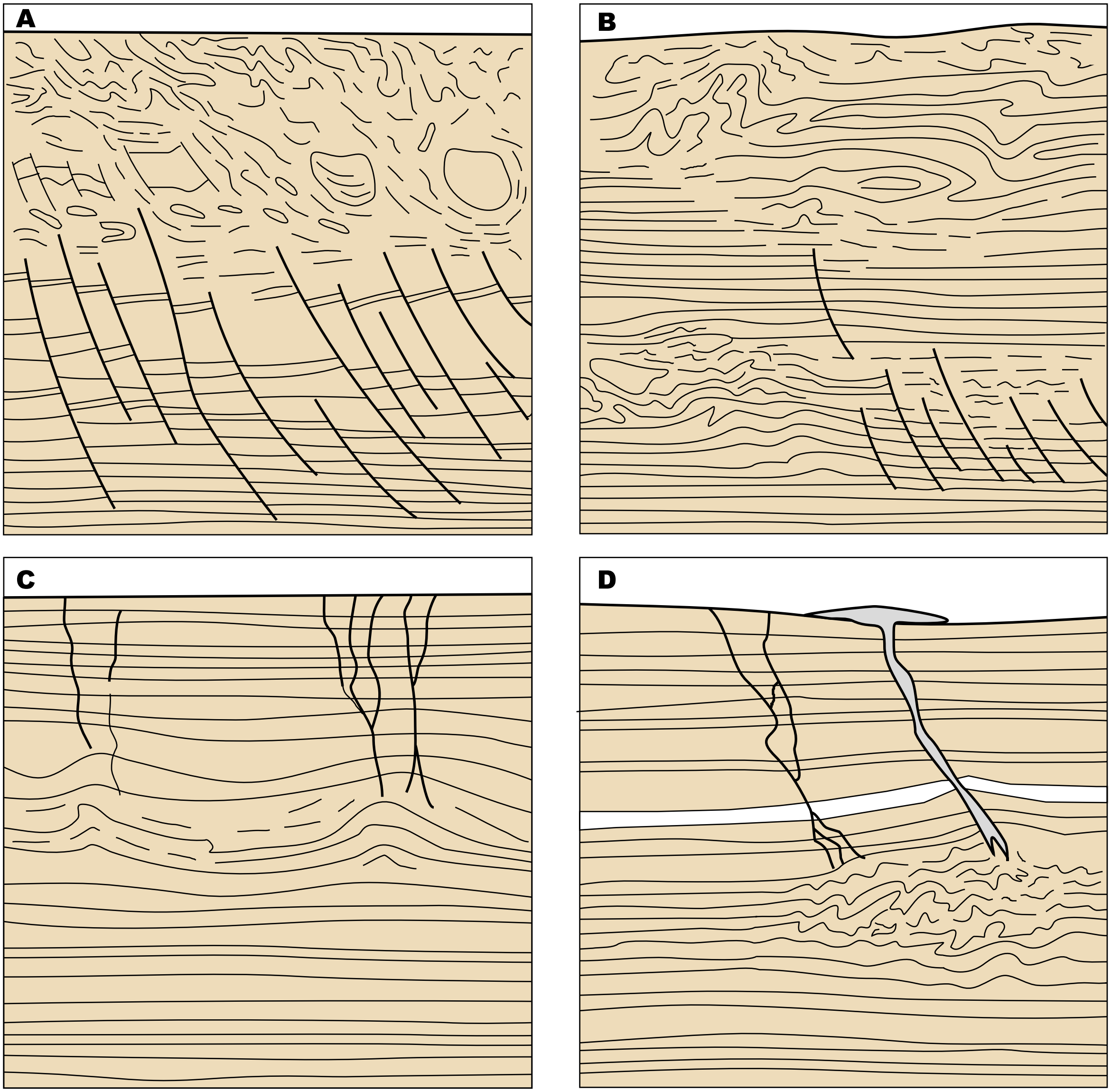


Figure 9.40 Soft-sediment deformational structures due to earthquake shock. These types of structures are typically developed in successions composed of laminated silts and sands. A) Closely spaced faults with throws that increase upwards and that pass upward into broken-up and then completely liquefied sediment. Ball-and-pillow structures are developed in the upper liquefied zone. B) Confined-layer deformation with ball-and-pillow structures and some faulting. Deformation near to the surface involves only mild contortion of layering. C) Incipient (discontinuous) confined-layer deformation with injection structures involving upward movement of liquefied material. D) Flame and fissure structures. The term fissure is used for small, near-vertical, irregular fractures that do not involve injection of sediment or significant displacement of layers. Modified after Ringrose (1988).

9.3 Chemically induced disturbance

Structures produced by post-depositional chemical activity are variable in their occurrence, mineralogy, and in the nature and timing of the chemical reactions. Three main types of process can be ideally envisaged: precipitation of minerals from pore waters, reactions between host sediment and pore waters, and dissolution of sediment by percolating water. Although this subdivision is theoretically acceptable, there are, in practice, real difficulties in distinguishing between the products of precipitation and

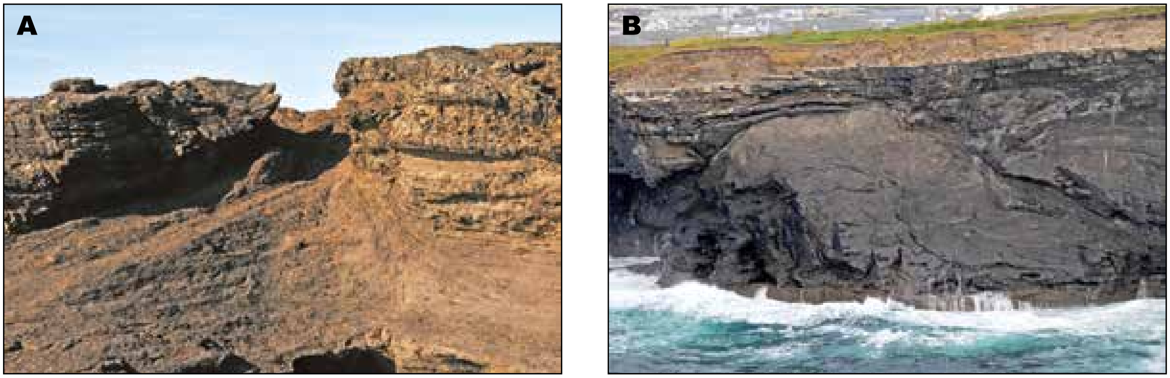


Figure 9.41 Mud diapirs. A) Small penetrative mud diapir that has punched through overlying delta-front sandstones. Overall thickness ca.15m. B) Small mud diapir penetrating delta-front mudstones and siltstones and overlain by mouth-bar sands. The flanks of the diapir behave rather like growth faults with roll-over folds. Both from Central Clare Group, Upper Carboniferous, County Clare, western Ireland. Cliff is ~25m high.

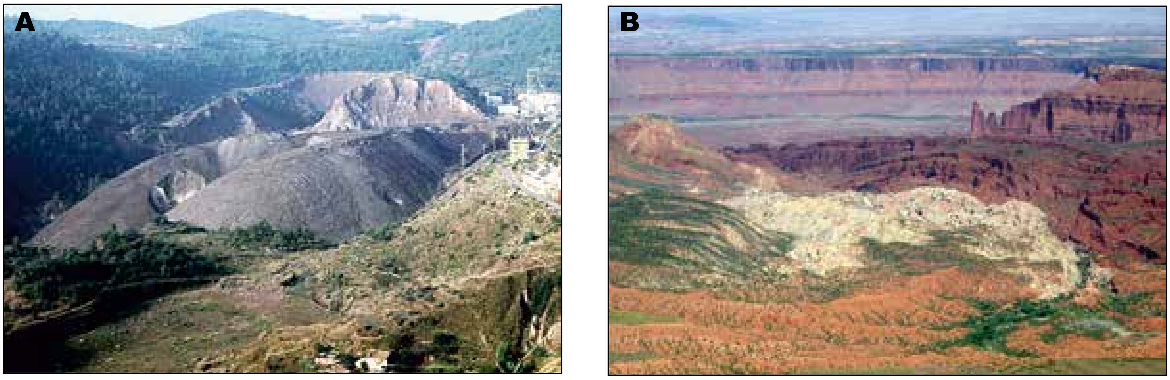


Figure 9.42 Examples of salt diapirs and associated sediments. A) An exposed salt dome, with overlying sediments stripped off. About 100m of relief seen on the top surface. The salt is of Miocene age, but the diapiric salt dome was emplaced more recently, probably in the Pliocene. Cardona, northern Spain. B) Upper surface of a salt dome, showing the folding of the overlying sediments. The white-coloured salt (a mix of halite and gypsum) is of Pennsylvanian (Upper Carboniferous) age. The overlying dark red sediments (background) are Permian to Jurassic. The orange sediments (foreground) are Pleistocene. Onion Creek Salt Diapir, Utah, USA. The prominent red sandstone tower (middle right) is ~200m high. Photo courtesy of Steven Banham. C) Intense and complex folding of deformed salt within a diapir, resulting from plastic flowage during emplacement. Cardona, northern Spain.

reaction, and the structures produced by these processes are best considered together.

9.3.1 Products of precipitation and reaction (nodules and concretions)

We are not concerned with general, large-scale lithification and cementation but with the local chemical precipitation and reactions that create structures commonly referred to as **nodules** or **concretions**. These two terms are used interchangeably. Nodules and concretions occur in host sediments of virtually any composition. They commonly stand out clearly because of a contrast in cementation or composition between the concretion and the host sediment (Fig. 9.43). They range in size from large masses, metres

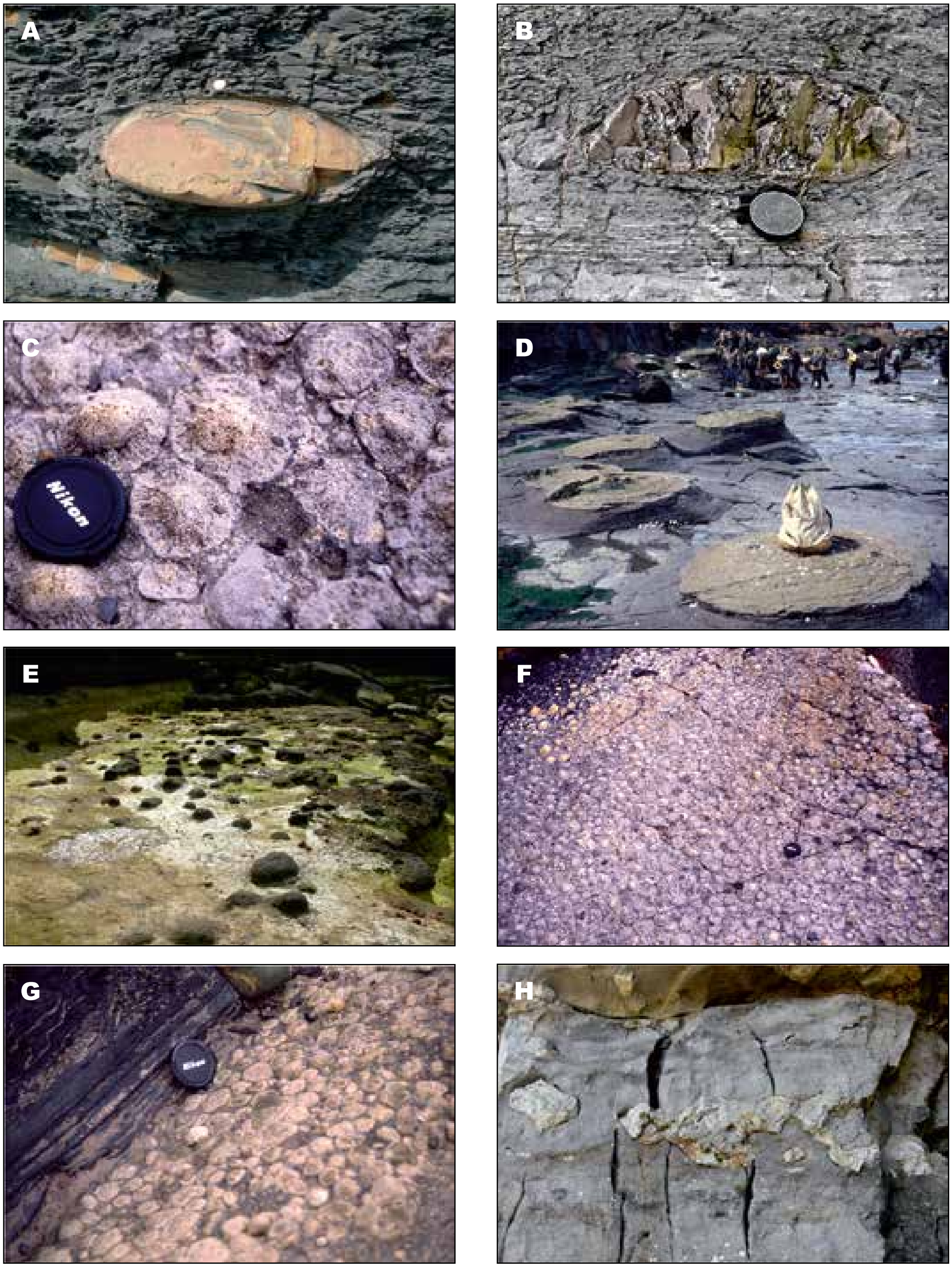


Figure 9.43 Examples of concretions and nodules. A) Large ironstone nodule (likely siderite) in black shale. Coal Measures, Upper Carboniferous, Amroth, Pembrokeshire, south Wales. B) Concretionary siderite in coal measures. Upper Carboniferous, Pembrokeshire, south Wales. C), F) and G) Concretionary dolomite nodules. Magnesian Limestone, Upper Permian, County Durham, England. D) and E) Calcite cemented concretions in siltstone seen in three-dimensional exposure. Lower Jurassic, Yorkshire, England. H) Chert nodules (5–10cm thick) in clean limestone of shallow-water origin. Clwyd Limestone Group, Lower Carboniferous, Anglesey, north Wales.

in diameter, to small bodies of 1mm or less. Shapes and patterns of distribution of concretions are highly variable. Understanding the processes of their formation may help elucidate the changing chemical conditions within a sediment following deposition.

There are five main sets of questions to have in mind when observing nodules and concretions:

- Is the nodule the product of direct precipitation into original pore spaces or is it the result of reaction between the host sediment and pore waters?
- What is the mineral composition of the nodule or concretion?
- What caused the concretion to be localized? Does it follow particular layers or beds? Is it associated with organic traces such as roots or burrows, or with body fossils? Are the concretions randomly distributed? Is there any systematic variation in the vertical distribution of the concretions?
- Has the concretion been precipitated to enclose grains of the host sediment (**poikilitic growth**)? Has it grown by pushing aside the host sediment (**displacive growth**)? Has it been precipitated in a large void space of primary or secondary origin?
- When did the processes occur relative to other post-depositional processes?

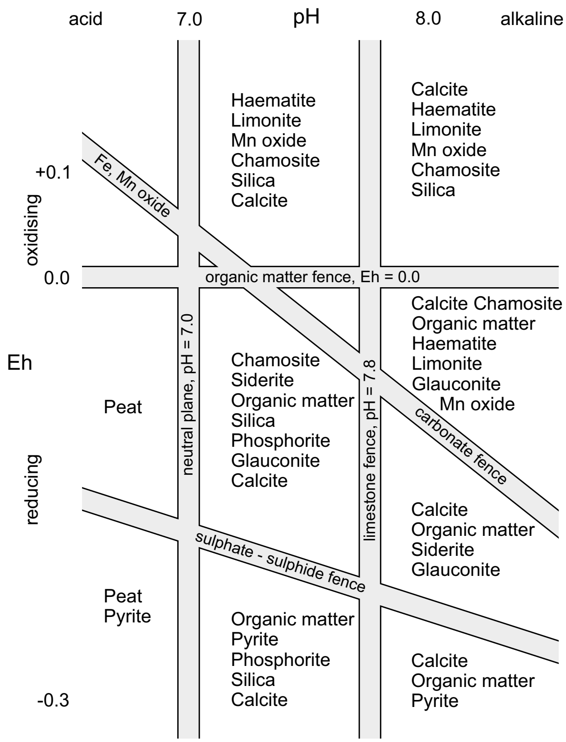


Figure 9.44 Stability fields of commonly occurring authigenic minerals found in sedimentary rocks, in terms of prevailing Eh and pH. For ground water with anionic composition comparable to that of sea water. After Krumbein and Garrels (1952).

Precipitation versus reaction

Are concretions or nodules formed by precipitation or reaction? Resolution of this question is not easy, particularly in the field, and in practice it may not always be important, as many concretions result from a combination of processes. Concretions or nodules in clastic successions are, on balance, more likely to be formed by precipitation, whereas those in carbonate and other chemical sediments (e.g. evaporites, ironstones) are more likely to have formed through reaction. There are few clear guidelines for making the distinction in the field, and laboratory examination of thin sections will nearly always be needed.

Mineralogy of concretions and nodules

The mineralogy of the material forming or cementing a concretion is the most important indication of the chemistry of pore water during nodule growth. For a particular mineral to be precipitated, the pore water must be supersaturated with the constituent ions. Other chemical conditions, notably acidity or alkalinity (pH) and oxidation or reduction potential (Eh) must also be appropriate. Figure 9.44 shows the general conditions under which different minerals form, but some caution is needed in applying it. Recently, the study of stable isotopes (especially of oxygen and carbon) has allowed the conditions of precipitation of some concretions to be more closely understood. Below are brief notes on the occurrence and significance of some common, concretion-forming minerals.

Calcite ($CaCO_3$)

Calcite is one of the most commonly occurring minerals. The calcium carbonate may already have been available within the sediment, for example, as shell fragments, or it may have been introduced from outside. Many calcite concretions in clastic sediments are of early diagenetic origin, forming from alkaline pore waters. Crystalline calcite also commonly infills larger cavities, particularly in limestone, as in geopetal infills (Fig. 8.14). Differential weathering may sometimes lead to early calcite concretions in sandstones being weathered out as cavities if the surrounding sandstone has been cemented by later silica cement, which is more resistant of weathering processes.

Dolomite–ankerite–siderite ($CaMg(CO_3)_2$–$Ca(MgFe)(CO_3)_2$–$FeCO_3$)

This family of carbonate minerals forms a continuous series with varying iron content and it commonly occurs in concretions in mudstones and siltstones. Alkaline conditions are required and precipitation of siderite and ankerite is favoured by reducing conditions (Fig. 9.44). Evidence of early diagenetic (pre-compaction) origin is common, suggesting that the concretions formed soon after deposition. In mudstones of some coal-measure successions, siderite forms a substantial proportion of the overall thickness, commonly occurring as laterally coalescing nodules or "beds" which were formerly mined as low-grade ores, the so-called "clay-ironstones" (e.g. Fig. 9.43B).

Pyrite and marcasite (FeS_2)

Pyrite and marcasite are very similar sulphide minerals that occur as nodules in both clastic and carbonate rocks. They are particularly common in dark fine-grained mudstones, usually associated with preserved organic matter.

Both minerals reflect strongly reducing conditions within the sediment, or even at the sediment surface in so-called "euxinic" environments. Within the sediment, restricted mixing of pore waters with overlying oxygenated water and the action of oxygen-reducing bacteria use up free oxygen. Sulphate-reducing bacteria then take over, producing free sulphide ions that are fixed by iron to give finely disseminated pyrite or marcasite. This gives a black colour to sediment only a few centimetres below the surface in finer-grained parts of present-day tidal flats. Larger pyrite concretions commonly show an internal radial pattern of crystal growth, and may replace fossils, preserving, for example, ammonites in three dimensions in mudstone intervals.

Silica (SiO_2)

Nodules or concretions of silica occur both in carbonates, where the silica is commonly crypto-crystalline chert or flint (Fig. 9.43H), and in sandstones, where it occurs as quartz overgrowths on detrital quartz grains. In limestones, chert is usually picked out by its darker colour and it is invariably a replacement of the carbonate, not a pore-filling precipitate. Flint nodules in chalk may follow particular horizons or have a more randomly scattered occurrence. Some may replace calcite fossils such as echinoids. In sandstones, silica-cemented concretions typically contrast strongly with less well cemented sandstone around them. Usually it is not easy to establish the timing of silica diagenesis, but some nodules in successions of continental origin compare with the **silcrete** nodules of some present-day soils and, by inference, are thought to have formed quite soon after deposition. As well as occurring as concretions, chert is also found in thinly bedded units due to lithification of primary siliceous oozes. The silica from which chert and flint is formed may be derived from the siliceous skeletons of marine planktonic organisms which are dissolved during diagenetic processes; silica is precipitated from the resulting solution. Chert commonly forms where an earlier material is replaced by the precipitation of silica. For example, petrified wood forms by replacement of the original organic matter by silica precipitated from fluids percolating through the dead wood whilst some calcite skeletons, for example, echinoids, can be precisely replaced by flint. Some types of sand-filled burrows (e.g. *Thalassinoides* burrows; see §9.4) are commonly sites of preferential chert precipitation especially where present in chalk deposits. Silica precipitation commonly requires weakly alkaline conditions.

Evaporites

Gypsum ($CaSO_4$, $2H_2O$) and anhydrite ($CaSO_4$) both occur at the present-day as nodules of early diagenetic origin in highly alkaline conditions, for example beneath the surfaces of ephemeral lakes and supratidal flats in hot, arid settings. Comparable structures are common in ancient sediments. The host sediment is usually carbonate, but evaporite nodules also occur in clastic sediments, particularly in muddy siltstones. With such diagenetic evaporites, the problem of precipitation versus reaction is particularly acute, as a carbonate host sediment is relatively reactive. Gypsum can occur poikilitically, enclosing grains of host sediment, whereas anhydrite typically develops displacively as nodules and layers that push aside the host. In some examples, gypsum precipitation follows plant roots, giving soil profiles called "gypcretes". In deserts, **sand roses** of gypsum form close to the sediment surface (Fig. 8.9).

In ancient successions, the textures, fabrics and structures developed by the growth of evaporite minerals may still be recognized even in cases where the evaporite minerals have been subsequently replaced by more stable phases such as chert.

Haematite (Fe_2O_3)

Haematite occurs in nodular forms as well as in its more familiar state as the pigment in red sediments. In each case its precipitation requires oxidising conditions although, once formed, it can persist in slightly reducing alkaline conditions (Fig. 9.44). Nodular haematite usually occurs in red or partially reduced successions, and commonly the nodules occur in profiles attributable to ancient soil development.

Barite ($BaSO_4$)

Barite is quite common as a localized cement, particularly in red sandstones where well-formed crystals poikilitically enclose the sand grains. Oxidising conditions and a supply of barium in solution are needed for barite to form in this way. Barite nodules commonly weather for form golf-ball sized spheres (Fig. 9.44). Such nodules may be recognized by their high specific gravity.

Limonite ($2Fe_2O_3.3H_2O$)

Although many iron minerals weather to limonite, the mineral also appears to form concretions and nodules directly, if oxidizing groundwater conditions prevail (Fig. 9.44). It most commonly occurs as concretions in sands and sandstones, often cementing the only lithified parts of otherwise unconsolidated sands. **Liesegang rings** are a colour banding, usually concentric and cutting across depositional lamination, that commonly involves limonite staining.

Form and location of concretions

A commonly recurring question is: Why has the concretion formed at this place and in this particular shape? In some cases the answer is obvious. In other cases, only general explanations are possible. The question implies that something in the host sediment caused chemical conditions to be suitable for a particular mineral at one place and not at another. Here we outline the most obvious controls, although many concretions appear to be randomly located.

Concretions that roughly follow bedding

In many fine-grained clastic successions, in chalk and in limestones, concretions or nodules occur in zones parallel to bedding. In some cases a slight difference in lithology is present between the concretionary horizons and those that contain fewer or no concretions. Individual concretions tend to be rather flattened parallel to the bedding and, in extreme cases, the concretions coalesce laterally into more or less continuous "beds". Common examples of this type are siderite concretions in siltstone and mudstone successions, flints in chalks and cherts in homogeneous limestones, possibly pseudomorphing early diagenetic evaporites (Fig. 9.43).

Subtle differences in, for example, organic content or permeability may control the development of such concretions. Compositional differences may allow particular chemical conditions to develop in pore-water, whilst permeability would control the rate at which pore water passes through the sediment. The nucleation of individual concretions must depend on even more subtle inhomogeneities in the sediments.

Concretions that follow burrows

Elongate and irregularly shaped concretions, particularly those that show branching patterns and cut across bedding, usually follow burrow traces and it may be possible to extract the cemented burrow from loose sediment (Fig. 9.45). Concretions centred on burrows are commonly of flint in chalk and of limonite in poorly consolidated sands.

Concretions that follow rootlets

In seat-earth palaeosols in coal measures, concretions are commonly associated with root systems that penetrate the bedding (Fig. 9.45). The concretions tend to be elongate normal to bedding. Most are composed of siderite, which weathers to limonite, but small pyrite crystals may fringe the concretion and be scattered within it. The concretions follow the traces of thicker roots, whereas the more common thin rootlets are preserved as carbonaceous films. The association with carbonaceous rootlets usually enables root concretions to be distinguished from those that follow burrows.

In other fossil soils, concretions of limonite, silica, gypsum calcite and/or haematite occur and these minerals may also contribute to a general colour mottling that follows the root disturbance (Fig. 9.45F & G).

Concretions centred on body fossils

There are three main ways in which fossils help to localize the development of concretions. First, the fossil itself can become a concretion when concretionary material replaces and exactly replicates the fossil as, for example, in pyritized bivalves, brachiopods, belemnites and ammonites (see Fig. 9.45D & E) or in flint echinoids in chalk. Secondly, the fossil forms a nucleus for precipitation. When broken, many ellipsoidal or more irregular carbonate concretions show fossils in their centres. The rotting organism provided local chemical conditions in the pore water that favoured precipitation. Thirdly, and more rarely, a concretion develops around the position occupied by the soft parts of an animal while the skeletal material remains undisturbed. For example, irregular masses of pyrite are sometimes found

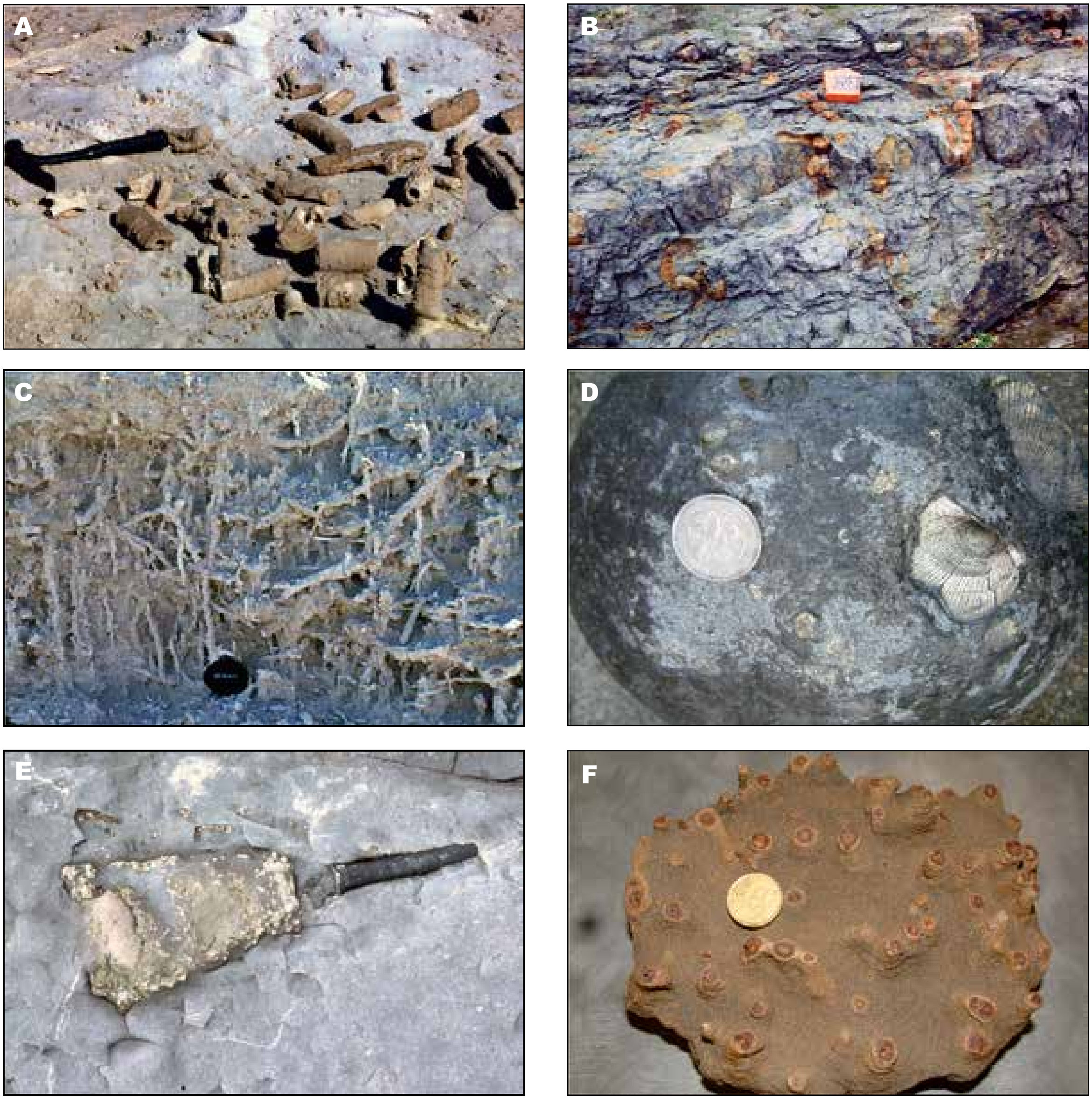


Figure 9.45 Examples of concretions around objects. A) Concretion (probably of limonite) around animal burrows. Lithification as part of the concretionary process has subsequently enabled the burrows to be exhumed intact. Locality unknown. Photo courtesy of Gilbert Kelling. B) Siderite concretions developed around root structures preserved in a seat-earth soil. Upper Carboniferous, Pembrokeshire, S Wales. C) Concretions around plant root structures. Plio-Pleistocene, Lake Eyre, Australia. D) Concretionary pyrite nodule developed around shell fragments which are themselves coated in pyrite. This indicates strongly reducing conditions after deposition. Hapton Valley Pit, Upper Carboniferous, Lancashire, England. E) Pyrite concretions in lime-rich mudstones concentrated at the end of a belemnite shell. The pyrite concentration is thought to result from strongly reducing conditions around the decaying soft parts of the belemnite. Upper Lias, Lower Jurassic, Yorkshire, Yorkshire, England.) Haematite concretions around plant root structures; the organic matter has subsequently decayed to leave fine tubes at the centres of each concretion. Holocene, Iceland. G) Concretions forming around the positions of former tree trunk, resting on a limestone palaeosol and formerly encased in carbonate aeolian sands. Pleistocene. Rocky Bay, Bermuda. Photo courtesy of Brian Rosen. *(continued)*

Figure 9.45 *(continued)*

at one end of a belemnite guard, in the position where the animal's soft parts would have been. Rapid deposition or a rather inhospitable sediment surface would have been needed to allow the soft parts to become buried before scavengers consumed them. Once buried, anaerobic rotting would lead to the reducing conditions necessary for pyrite precipitation.

Concretions in distinct vertical profiles

Concretions, particularly of calcite, may occur within red siltstones and sandstones in distinct vertical profiles ranging in thickness from decimetres to several metres. Similar profiles also occur in limestones, where they are commonly associated with karstic features (see §8.4.2). The concretions, which are commonly a few centimetres in dimensions, have irregular shapes but are, in many cases, elongated vertically. Some profiles have only scattered nodules, whereas others show an upwards increase in nodule size, abundance and coalescence, commonly as meshes and networks with linking veins (Figs. 9.46, 9.47). In rather rare examples, the upper part of the profile is a

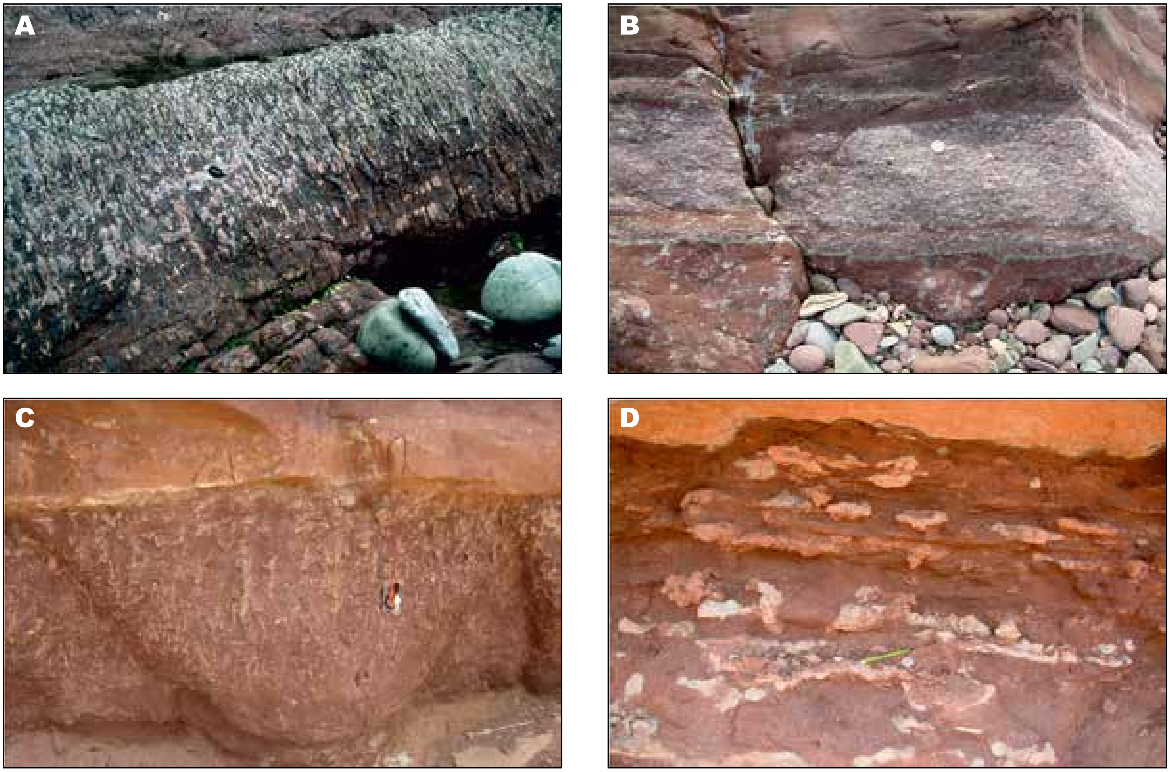


Figure 9.46 Carbonate concretions developed as a result of soil-forming processes to give calcrete or caliche profiles. A) A mature profile with concretions concentrated around root structures. This corresponds to stage B in Figure 9.47. B) Calcrete nodules developed to different degrees within different layers of the host sediment. Both examples from the Old Red Sandstone, Devonian, Pembrokeshire, south Wales. C) Carbonate concretions following vertical root systems below a sharp upper contact that was probably close to the contemporaneous exposed land surface. Note how the concretions become smaller downward, reflecting the form of the original root network within the palaeosol. Cedar Mesa Sandstone, Permian, Utah, USA. D) Calcrete nodules that have in places developed to an extent such that they have begun to form a laterally amalgamated 'hardpan' horizon. This corresponds to stage C in Figure 9.47. Undifferentiated Cutler Group, Permian, Utah, USA.

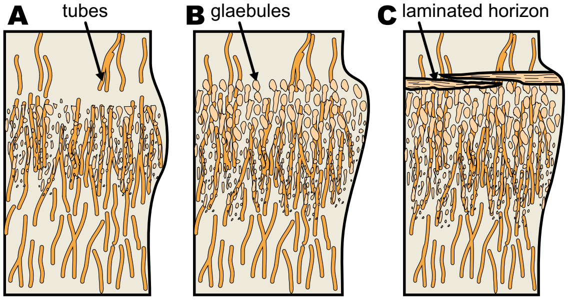


Figure 9.47 Carbonate concretions developed as a result of soil-forming processes to give calcrete or caliche profiles. Stages in the development of a mature profile. A) Incipient concretion growth. B) Development of large concretions, especially in the zone of water-table fluctuation. C) Intercalation of concretionary nodules to give a laminated 'hardpan' horizon.

more or less continuous bed of limestone with a crude horizontal lamination. These carbonate-rich layers tend to be laterally continuous and to maintain their character and they are hence often useful for correlation, at least over short and intermediate distances. In some examples, the upper surface of the profile shows vertical relief in the form of broad cuspate dishes. These may relate to radial patterns in the sub-vertical concretions and to convex-upwards veins within the network of concretions. Such structures have been called **pseudoanticlines**.

These profiles and associated features compare closely with those of present-day soils of some semi-arid areas, the so-called **calcrete** and **caliche** soils. These result from the vertical movement of water through the sediment due both to the downward movement of rainwater and the upward movement of ground water under dry evaporating conditions involving capillary action. Some of the vertical nodules (or **glaebules**) may reflect original plant roots (Fig. 9.47), whereas others may follow fractures. The more continuous limestone layers with lamination record the development of **hardpan** conditions in a fully mature profile. For calcretes developed in siltstones and sandstones, it is possible that the calcium was introduced as wind-blown dust, as either carbonate or sulphate, whereas an internal origin is clearly more likely in limestones. The development of calcretes requires that the sediment surface is subaerially exposed, with little or no sedimentation for a considerable period of time, probably thousands of years. The comparative maturities of profiles reflect the relative durations of non-depositional intervals within any particular succession. The identification of these profiles is clearly important in environmental and palaeoclimatic reconstruction, especially where the nodule-forming processes within soils are strongly controlled by climate (Fig. 9.48). Not only do the profiles indicate prevailing conditions but they also record fluctuations in sedimentation rate. Other similar profiles involve silica, haematite and gypsum nodules giving **silcrete**, **ferricrete** and **gypcrete** soils, respectively (Fig. 9.49).

Although many calcrete palaeosols occur in distinct profiles, other palaeosols show a more scattered distribution of carbonate nodules. These **accretionary palaeosols** record similar soil-forming processes and conditions but without the periods of sustained non-deposition that give clear profiles (Fig. 9.50).

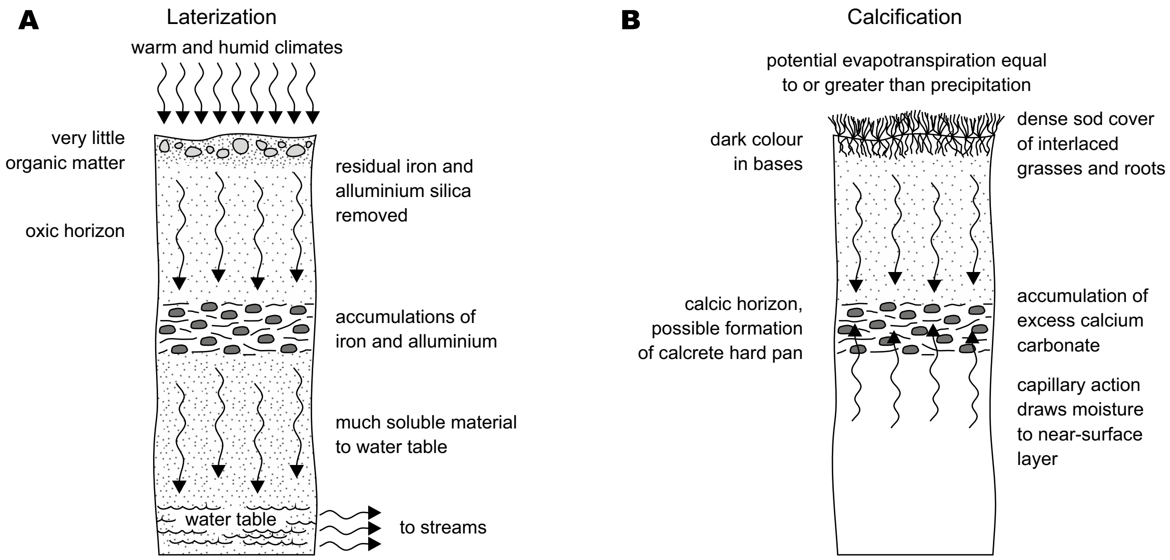
Mode and timing of growth of concretions

Inferences about the way in which concretions grew within the host sediment may be made by looking at internal features of the concretions. Such features may also give information on the timing of concretion development and on the degree of compaction that the sediment has subsequently experienced. There are four main ways in which concretions and nodules develop.

Primary pore filling

Evidence for primary pore filling is provided by the preservation of original sediment grains and bedding features within the concretion. Such concretions occur mainly in clastic sequences. In mudstones, it may be possible to see the original lamination running through the concretion and linking up with lamination in the host sediment on either side. If this is seen, it is useful to compare the thickness of the lamination inside the concretion with that in the host sediment and to note how the lamination in the host sediment behaves around the concretion. Lamination is commonly much thicker within a concretion than outside it and the lamination in the host sediment wraps around the concretion for some distance above and below. Clearly a certain amount of compaction has taken place after the concretion formed and comparison of lamina thicknesses can help estimate its magnitude. Where compaction is demonstrably large, it is reasonable to infer that the concretion formed soon after deposition.

In some concretions in mudstones, particularly those with a carbonate cement, the central parts of concretions show a pattern of irregular lenticular cracks filled with coarse calcite crystals. These are **septarian nodules** and the cracking reflects a synaeresis-type contraction due



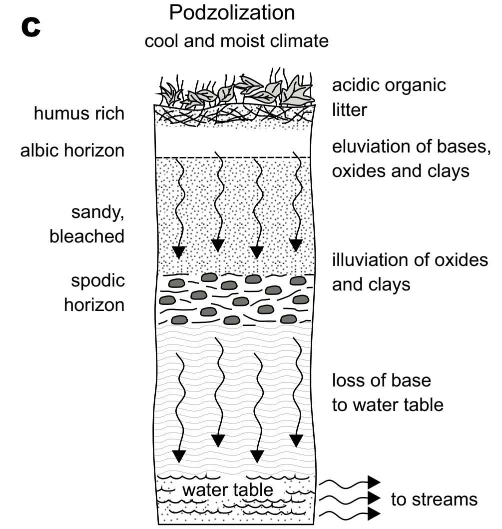


Figure 9.48 Models for the development of various types of soil profiles. A) Laterization is common in highly leached, humid tropical and subtropical climates. B) Calcification occurs in climates where evapotranspiration exceeds precipitation, producing aridisols and mollisols. C) Podzolization occurs in cool and moist climates. Modified after Allen (1997).

to the dewatering of a gel-like mass of clay minerals (Fig. 9.51). In other concretions a curious angular pattern of fractures is found around the margins involving stacked conical fracture surfaces. This is **cone-in-cone** structure (Fig. 9.52), which reflects a stress field set up by the growth of the concretionary cement.

In sandstones, early carbonate concretions may weather out as holes due to a later phase of general cementation by silica. This renders the concretions more susceptible to weathering than the host. It is often possible to obtain a clearer impression of the original shape and packing of the sand grains in partly weathered carbonate concretions where the silica cement is not present.

Displacive nodule growth

Here the host sediment is physically pushed aside as the nodule grows, and little or none of it, is incorporated within the nodule. Internally this may sometimes be apparent from the crystal structure of the nodule, as with those pyrite nodules with a radial fabric. One of the most common examples of displacive growth is that of gypsum and anhydrite nodules which occur both as layers and as more irregular forms. As the nodules grew, they pushed apart the host sediment until only thin remnants remained between the nodules, leading to so-called **chicken-wire texture** (Fig. 9.53). Growth of anhydrite as layers sometimes sets up stresses that produce highly contorted folding of the layers.

Nodule growth by replacement

This is not always easy to recognize. The replacement is, in some cases, associated with displacive growth, but in other cases, details of the internal structure of the host

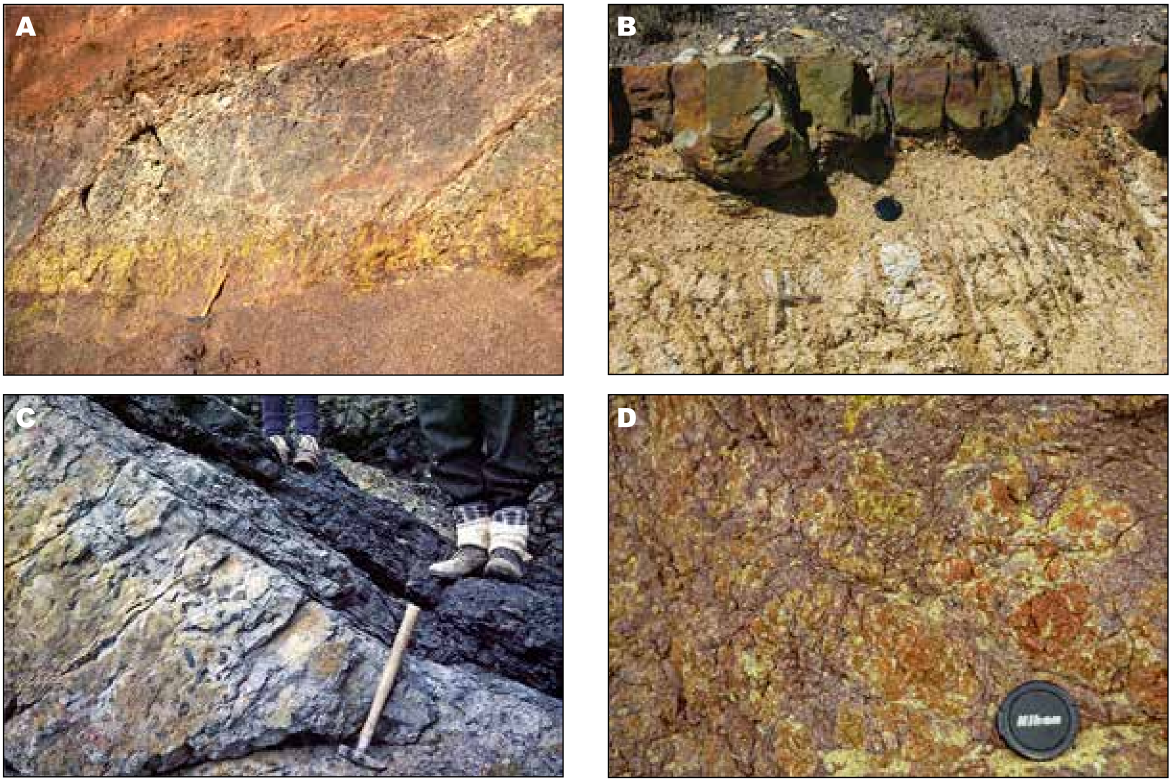


Figure 9.49 Colour patterns in palaeosols. A) Mudstone palaeosols showing a wide range of colours, organised into distinct zones, the colours reflecting subtle differences in the oxidation/reduction state of the pore waters during development of the soil. Such colour assemblages are typical of soils that formed with a fluctuating water table in humid settings. Etruria Formation, Upper Carboniferous, Staffordshire, England. B) A highly mature, strongly leached palaeosol in a sandstone host. The upper cemented layer has a strong silica cement whereas the more friable material below is enriched in quartz due to the dissolution of feldspars. Such palaeosols reflect prolonged subaerial exposure in a humid setting. Marsden Formation, Upper Carboniferous, Yorkshire, England. C) Seatearth palaeosol with an overlying coal seam. The sandstone host of the palaeosol is penetrated by carbonaceous rootlets and has siderite nodules that may follow larger roots. The preservation of carbonaceous material and the siderite nodules indicate that the soil formed under reducing, water-logged conditions. Coal Measures, Upper Carboniferous, Pembrokeshire, Wales D) A multi-coloured palaeosol showing a disorder assemblage of colours that are characteristic of soils formed in humid conditions but with a fluctuating water table that caused changing oxidizing/reducing pore waters. Wessex Formation, Lower Cretaceous, Isle of Wight, England.

Figure 9.50 Thick succession of multiple stacked palaeosol profiles developed in siltstone and mudstone with thin sandstone interbeds. The palaeosols developed in a fluvial floodplain setting. The thin sandstone beds represent crevasse splays whereby sand was introduced into the flood basin at times when nearby river channels burst their banks. The cliff is ~40m high. Morrison Formation, Jurassic, Utah, USA.

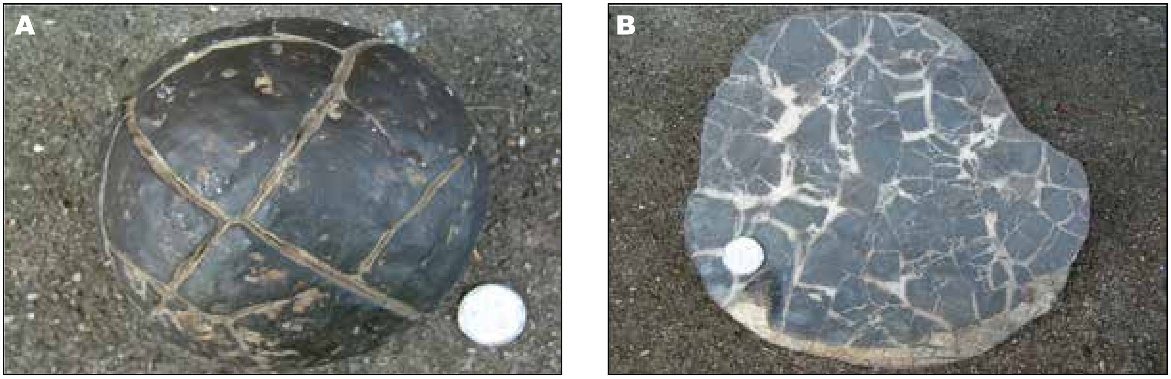


Figure 9.51 Examples of septarian nodules. A) External form. B) Internal structure. The lenticular cracks are filled with coarse calcite crystals. Silsden Formation, Upper Carboniferous, Lancashire, England. Keele University collection.

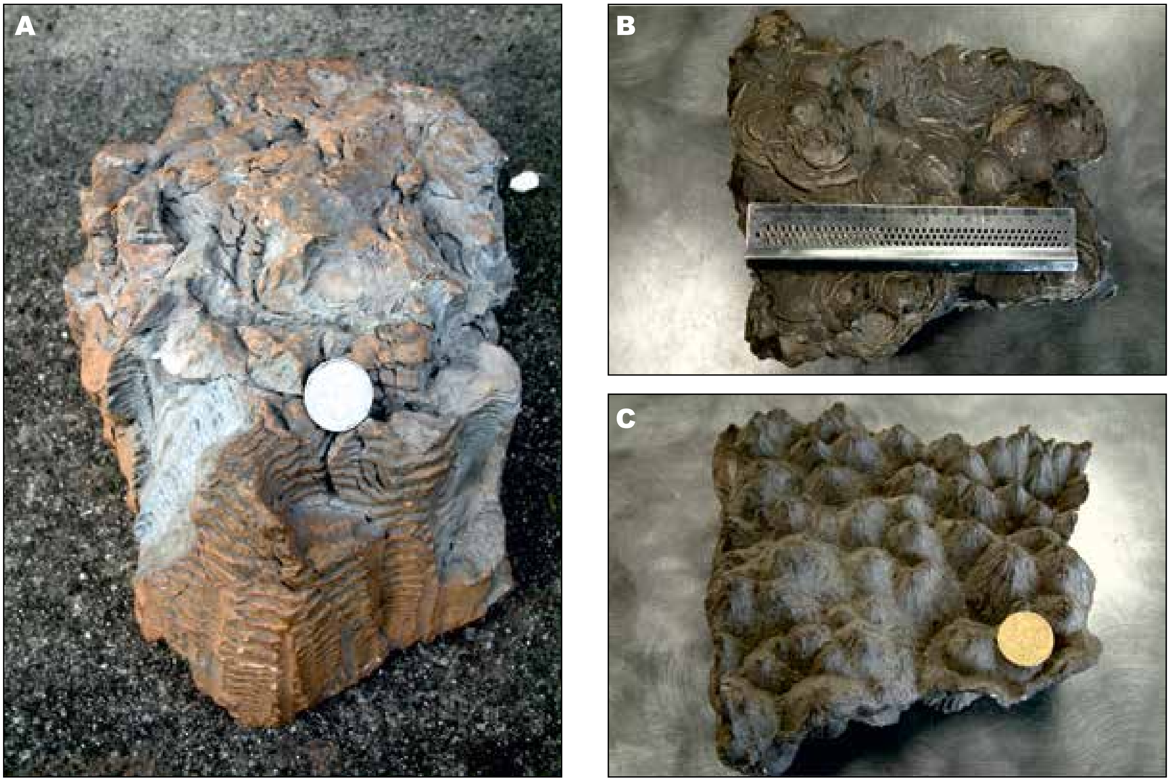


Figure 9.52 Examples of cone-in-cone structure in carbonate-cemented mudstone. Upper Carboniferous Coal Measures, England. A) Keele University collection. B) and C) University of Leeds collection.

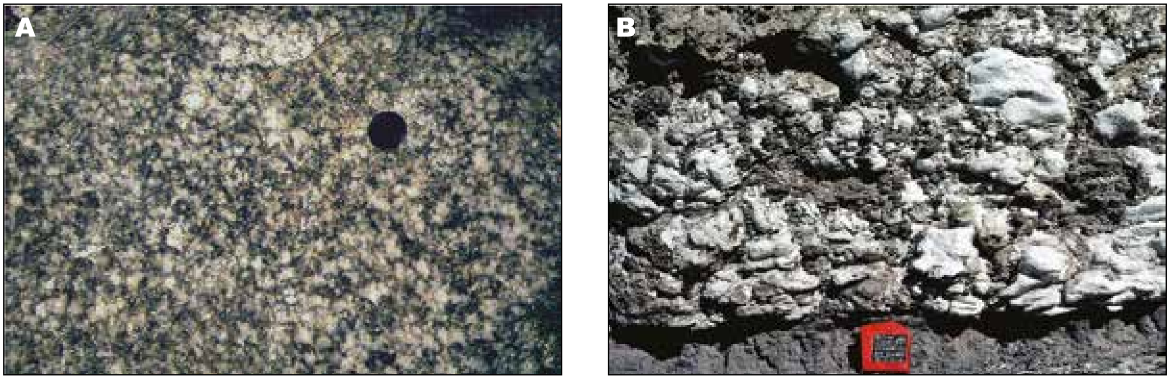


Figure 9.53 Anhydrite nodules which have grown displacively to push aside host sediment into thin veneers between the nodules, giving so-called 'chicken-wire' texture. A) Miocene, Mojacar, southeast Spain. B) Poulsen Cliff Formation, Ordovician, Washington Land, northwest Greenland.

sediment are preserved even though total replacement has taken place. In addition to showing vestiges of the original lamination, replacement nodules commonly have reaction rims where the process of replacement has not reached completion. If the replacement is complete, concretions formed by reaction between host sediment and pore water can have many of the attributes of displacive nodules and the two processes are not mutually exclusive.

Concretions due to cavity infill

Cavities within sediment may be of primary or secondary origin and are most common in limestones. An example of a primary cavity is the body chamber of a shelled organism where the strength of the shell maintained the cavity until the fill had been precipitated. A secondary cavity could result from the solution or rotting of some object, probably after the host sediment had become sufficiently lithified to support the cavity. Infills of voids are characterized by well-formed, pure crystals, growing inwards from the walls. Concentric zones may show crystals of increasing size from wall to centre. Two types of cavity fill of particular interest, **geopetal infills** and **stromatactis** were dealt with in Chapter 8 (see §8.3.1).

9.3.2 Products of dissolution

In addition to the large-scale dissolution features related to karstic surfaces (Figs. 8.32, 8.33), small-scale dissolution features also occur, most commonly in limestones.

Stylolites

On certain bed partings, particularly in limestones and less commonly in sandstones, a highly crenulated contact is seen (Fig. 9.54). In limestones a thin layer of clay often defines the surfaces, whereas in sandstones they are commonly coated with carbonaceous material. Relief is usually a few millimetres and seldom more than a few centimetres. Where such a surface is exposed in plan, the small-scale relief is seen to be highly irregular.

Stylolites result from dissolution of both upper and lower beds at a particular bedding surface. The beds grow into each other due to vertical compression. The relief of the irregularities gives a minimum measure of the thickness of material removed in solution and can in some cases represent a high proportion of the original rock. The clay and carbonaceous partings that occur on some surfaces represent the insoluble residue of the dissolution.

9.3.3 Palaeosols

Palaeosols are lithified soil profiles that originally developed at a subaerial surface or in the shallow subsurface. Palaeosols are important because they provide a valuable indication of past surface conditions, often related to climate. They are commonly arranged into horizons or layers which are typically aligned parallel to the palaeo-landscape surface on which the soil developed (Fig. 9.46). The uppermost horizon of many palaeosols is characterized by physically reworked rip-up clasts whilst the top of a palaeosol may, in some cases, be a sharp erosional surface.

Many palaeosols contain plant root traces, either as fossilized biogenic remnants or as chemical precipitates that developed around roots at the time of soil formation as plants took up water from the substrate (Fig. 9.45B & G). The composition and abundance of preserved plant root structures (if present) can indicate the drainage state of the soil. Preserved carbonaceous material (Figs. 9.45B, 9.49C)

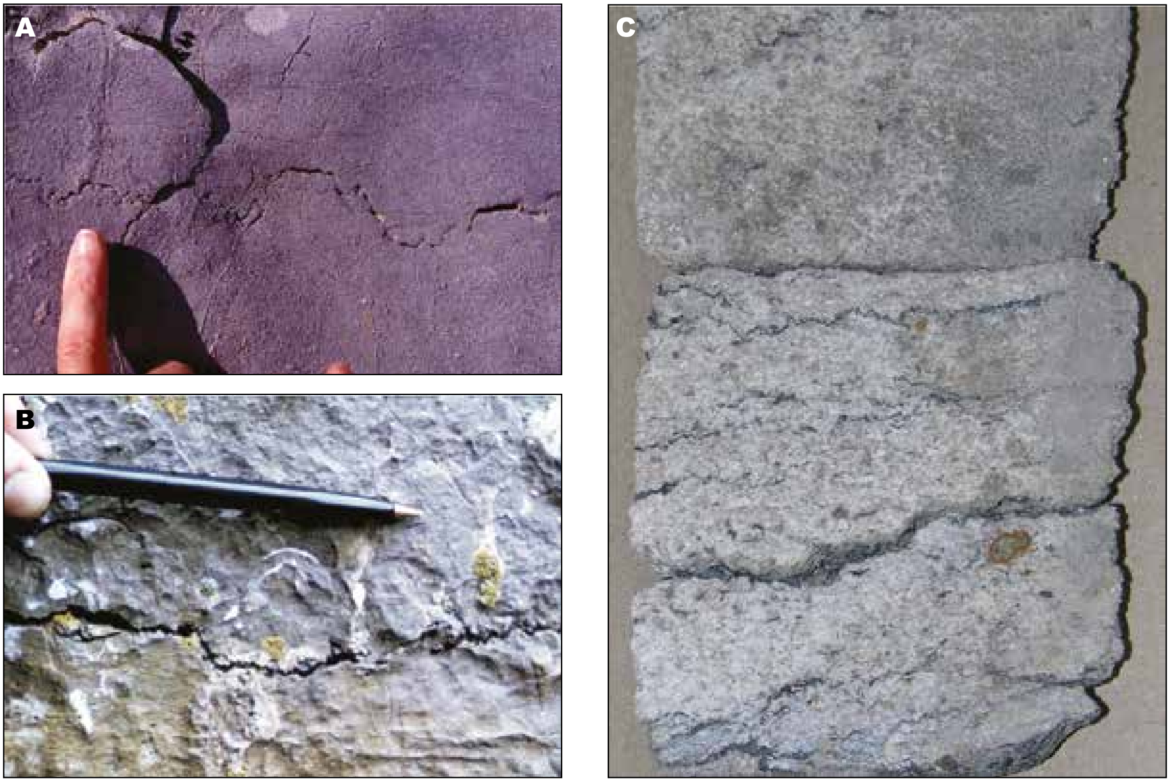


Figure 9.54 Stylolites. The crenulated surfaces are developed normal to the principal compressive stress and usually follow bedding. They commonly show concentration of insoluble residues. A) Devonian Limestone, La Vid, Cantabrian Mountains, Spain. B) Pembroke Limestone Group, Lower Carboniferous, Pembrokeshire, Wales. C) Small-scale stylolites in sandstone with the crenulated surfaces emphasised by concentrations of finely comminuted carbonaceous material. Cored well (~12cm diameter), Upper, Upper Carboniferous, southern North Sea.

indicates a water-logged state whereas preservation of calcified rhizoliths indicates well drained conditions, typically in semi-arid settings (Fig. 9.46).

Many palaeosols are characterized by abundant peds, which are soil structures formed from the repeated swelling and shrinking of clays as the water table varied during soil development. One common ped type with a granular texture represents an aggregate formed by binding of organic matter and clays. Another common ped type is characterized by a blocky form of shrinking and swelling of clays, in some cases with associated bioturbation. Another ped type is characterized by vertically oriented slickensides that formed from shrinking and swelling of clays. Platy peds oriented parallel to the land surface might record relic layering of the host sediment, possibly developed in response to freezing and thawing, action, or to the precipitation of layers of carbonate, silica, or iron oxides. In many palaeosols nodules or concretions are present, in the form of calcite, iron, silica, and opal, for example.

Illuviation is the progressive downward (or more rarely upward) movement of sediment through by mechanical, chemical or biogenic processes. Cutans, which are composed of clay, calcite, organic matter, silica, and iron oxides, are the mineral coatings of grains or peds in palaeosols produced by illuviation.

The colours evident in palaeosols are indicators of near-surface conditions that prevailed during soil development. Orange or reddish palaeosols typically indicate oxidizing conditions, typically associated with good drainage and a low water table (Fig. 9.46, 9.49A & D). Black or dark grey colours indicate anoxic conditions and are typically associated with the preservation of carbonaceous matter. A green-grey banded palaeosols are typical of reduced iron.

Descriptive terminology used in the classification of palaeosols has been derived mainly from schemes established for present-day soils. Many critical features would not survive burial diagenesis and so the classifications must be applied with care. Calcisols are enriched in calcium carbonate or dolomite from illuviation. Gypsisols are characterized by gypsum or anhydrite precipitated in the vadose zone. Vertisols exhibit homogenization of the profile by plant-root activity (pedoturbation), together with the shrinking and swelling of clays to form vertical slickenside surfaces. Gleysols preserve organic matter and are indicative of low redox conditions, for example due to anoxia and a permanently high water table, as in many swamps. Histosols are rich in preserved organic matter in the form of different types of coal. Oxisols record extensive chemical alteration of chemically unstable minerals to brown clays. Protosols are poorly developed soils with weakly developed horizons.

Different types of palaeosols record differences in factors such as climate (precipitation, temperature, humidity, moisture content and seasonality), host sediment (porosity, permeability, chemical composition), biogenic activity (plant root growth, bioturbation by animals, the action of mould and bacteria), groundwater hydrology and drainage state (redox conditions associated with water-table height), the duration, type and intensity of processes responsible for soil formation.

Many palaeosols show distinct vertical profiles wherein systematic changes in pedogenic features occur beneath a sharp upper boundary, which approximated to the exposed land surface (Fig. 9.50). The extent to which such a profile is developed depends upon prevailing conditions and upon the duration of the subaerial exposure before sedimentation resumed. The maturities of palaeosol profiles within a succession are, therefore, indicators of the duration of periods of non-deposition (Fig. 9.49B). Some successions of alluvial mudstones or siltstones show pedogenic features scattered more uniformly through the sediments. Such accretionary palaeosols reflect conditions where sedimentation and pedogenesis more or less kept pace with one another.

9.4 Biogenic sedimentary structures: trace fossils

9.4.1 Introduction

The study of trace fossils (**ichnology**) is concerned with understanding the disturbance of sediment by living organisms, i.e. with **biogenic** sedimentary structures. Apart from the consistent recognition of vertebrate (e.g. dinosaur) footprints from 1828 onwards, trace fossils were at first grouped as Fucoids (fossil seaweeds), and their algal origin was hotly debated. Until the 1970s, trace fossils were largely ignored in most geology courses. At outcrop they were commonly dismissed as "burrows" or "worm traces", suggesting that they have little to contribute to the understanding of Earth history. In fact, the reverse is commonly true. Trace fossils, in contrast to many body fossils, which are rolled and derived, are records of life and events that took place *in situ* during or soon after the deposition of the sediment. Trace fossils may be common in successions where no body fossils have been preserved, for example in non-marine red beds, or where any organisms were entirely soft bodied.

Trace fossils record behavioural, ecological and sedimentological events that body fossils and other sedimentary structures cannot record directly. Their study may, therefore, alter one's view of a problem or turn an investigation in a new direction. In some cases, they provide the key to what initially may appear to be a purely sedimentological problem, for example the origin of massive beds.

Even where body fossils and sedimentary structures are abundant, trace fossils should be studied with appropriate rigour, for they yield information against which to test a wide range of conjectures and speculations. They may inspire working hypotheses that would otherwise not be considered. They encourage the study of sediments from biological, ecological and biochemical standpoints, thus complementing the study of physical structures by providing a potentially valuable source of information and ideas. A variety of simple but effective techniques for observing and recording trace fossil structures is recommended in Appendix 4.

9.4.2 Classification of trace fossils

We emphasize a practical rather than a theoretical approach to the complex task of describing and interpreting sediments that contain trace fossils. Three main approaches to classification are considered; they are best used in combination:

- Taxonomic classification using morphological aspects.
- Preservational–sedimentological classification.
- Ethological classification through the consideration of behavioural and environmental aspects.

In introducing each of the approaches below, we pose a series of questions that can be used as a step-by-step guide to trace fossil classification. Question 1 is concerned with the description of trace fossil morphology, questions 2–8 consider modes and processes of preservation and questions 9–21 consider trace fossils in terms of the behaviour patterns of the organisms responsible.

Taxonomic (morphological) classification

Non-experts are often frustrated by not being able to identify trace fossils satisfactorily, even though they may have access to appropriate journals, recent books and the relevant treatises. Systematic ordering of trace fossils according to the taxonomy of the organisms producing them would be highly desirable, but even experts disagree on how to do this. The problem is that one trace fossil "genus" may cover traces made by several different organisms, and several "genera" may be made by the different activities of one organism. Recognising these difficulties, international codes of biological nomenclature have been modified to accommodate the naming of trace fossils with the introduction of new codes in 1985 and 1997 (see reference list). In this connection, the basic unit of nomenclature is the **ichnogenus**, defined morphologically. The lower category, the **ichnospecies**, typically reflects size, preservational or minor behavioural variations of the basic form (Fig. 9.55).

Given that the objects to be named are biogenic, five overall sets of characteristics or **ichnotaxobases** are commonly used for the morphological description and classification of trace fossils:

- The general form: describe the shape and orientation, the presence of shafts, networks, spiral meanders, concentric laminae (spreite) etc.
- The structures of the burrow boundary: describe the presence or absence of a lining; the presence of a dust film, a constructional lining or a zoned fill, the degree of wall compaction; the presence of diagenetic haloes, the type of wall ornament.
- Branching of burrows: describe the form and orientation of burrow splitting, the presence of true or false galleries, in the latter case due to reworking or intersection.
- The filling material and its structure: describe whether the structure has been passively filled by gravity settling of sediment or by the activity of the burrowing organism.
- Repetitive footprints and impressions: describe the form of any locomotion and related structures due to walking, crawling, bottom swimming by invertebrates and vertebrates; in the case of tracks measure track width, morphology, repetition modules, repeat distances, symmetry, obliquity, and the degree of continuity.

These guidelines suggest that the following questions should be considered:

Q.1 What is the morphology of the trace fossil? Are there identifiable shapes of organisms or parts of them? In particular, refer to the five ichnotaxobases discussed above and use these in conjunction with Figure 9.56 as a guide to description. A description should consider overall form, the nature of the wall, the style of branching (if any), the style of fill and the nature of any repetitive surface patterns. Is the trace best described as:

- Single shape (e.g. a print made by a foot; Fig. 9.57).
- Several similar shapes repeated to form a pattern (e.g. a track made during locomotion).
- A trail (i.e. a continuous groove made during locomotion).
- A radially symmetrical shape developed on a bedding surface (e.g. by the resting of a starfish).
- A tunnel or shaft, possibly caused by a burrower seeking food and/or refuge.
- A series of closely related, concentric laminae (spreiten), often U-shaped, caused by an animal shifting position within its burrow as it grows or moves upwards, downwards, forwards or backwards by excavating and backfilling (Fig. 9.58).

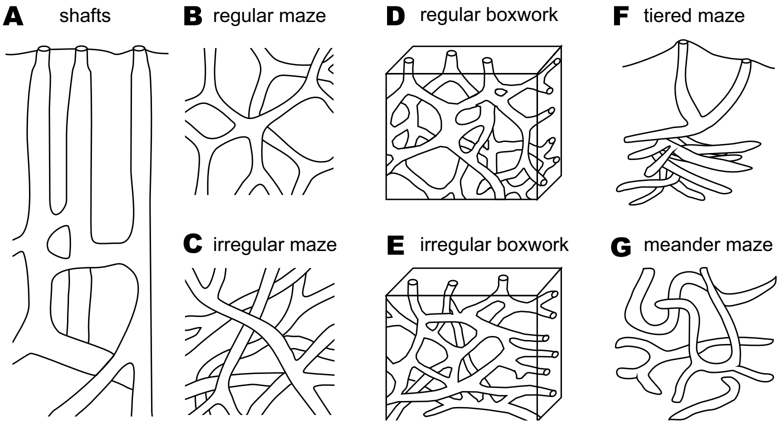


Figure 9.55 Illustration of the variability of burrow arrangements for a single ichnogenus. Classification scheme of *Ophiomorpha* burrow configurations and modern analogues. A) Vertical components are most dominant in the network. B) and C) Regular and irregular two-dimensional burrow networks consisting almost entirely of tunnels. D) Three-dimensional polygonal system of vertical, inclined and horizontal elements. E) Irregular three-dimensional burrow networks. F) Series of finite tunnels forming a main gallery connected to the substrate-water interface by shafts. G) Two-dimensional system of sinuous tunnels. *Ophiomorpha* is a dwelling structure characterized by pelleted wall linings. Modified after Anderson and Droser (1998) and Frey et al. (1978).

Rank IV	Rank III	Rank II	Rank I	Examples
Track-like trace on bedding plane		'Prods' or 'scratches' all alike	Clustered 'scratches'	_Paleohelcura_
			Rows of 'prods'	_Tasmanadia_
		'Prods' or 'scratches' of different kinds	Rows of 'prods'	_Kouphichnium_ phyllopod tracks*
Trail-like trace on bedding plane	Freely winding	Simple trail	No ornament	_Gordia_
			Transverse ornament	_Climatichnus_
		Bilobed trail	Transverse ornament	_Cruziana_*
		Trilobed trail	Transverse ornament	_Scolicia_*
	Windings in contact with one another; pattern on bedding plane	Simple trail	No ornament	_Helminthoida_*
		Bilobed trail	Transverse ornament	_Nereites_*
Radially symmetrical in a horizontal plane	Without axial vertical structure	Five-rayed	Rays are grooved	_Asteriacites_*
		Multirayed	Club-shaped rays	_Asterosoma_*
	With axial vertical 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_*
		Regular branching		_Chondrites_*
	Variable diameter	Irregular network		_Thalassinoides_*
Forms having a spreite		U-shaped	Vertical plane	_Diplocraterion_*
			Horizontal plane	_Rhizocorallium_*
		Spiral	Inclined plane	_Zoophycos_*
		Branched	Vertical	_Phycodes_*
Pouch shaped		Smooth surface		_Lockeia_*
		Transverse ornament		_Rusophycus_*
Miscellaneous		Net pattern		_Paleodictyon_*

Figure 9.56 A morphological classification of some common invertebrate trace fossils. Modified after Simpson in Frey (1975). Examples that are illustrated in the figures are marked by an asterisk*.

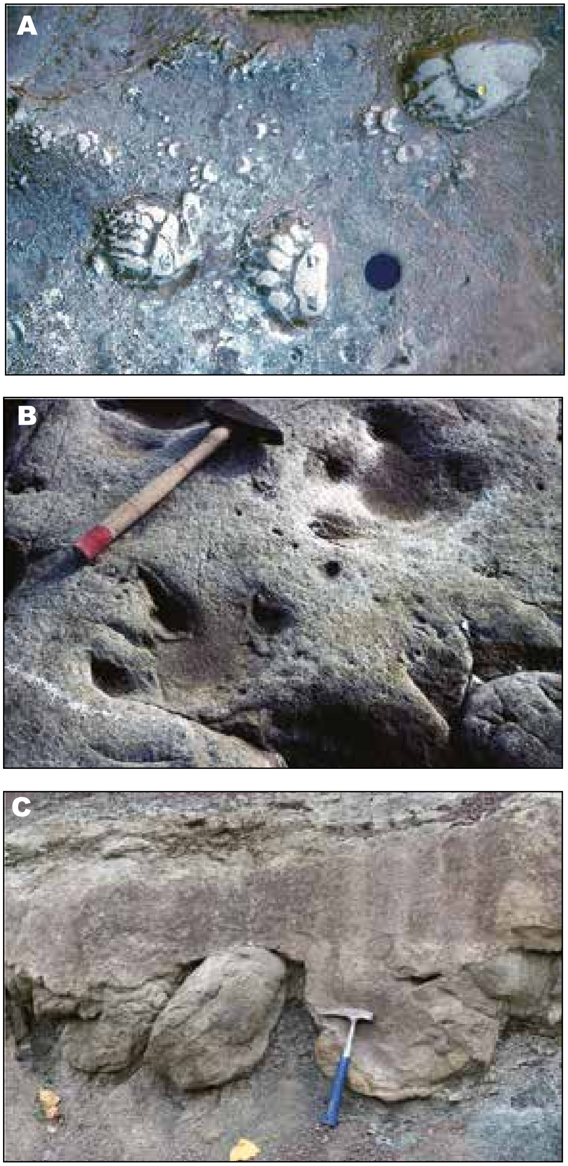


Figure 9.57 Footprints on originally soft sediment surfaces. A) Vertebrate (bear) footprints; two sets, parent and juvenile. Modern, White Canyon, Utah, USA. B) Dinosaur footprints on upper surface of a sandstone bed. Lower Cretaceous, Spitzbergen, Svalbard. C) Dinosaur (Sauropod) footprints seen in section where the animal's feet have sunk deeply into unconsolidated mud. Natural casts. Villar del Arzobispo Formation, Jurassic, eastern Spain.

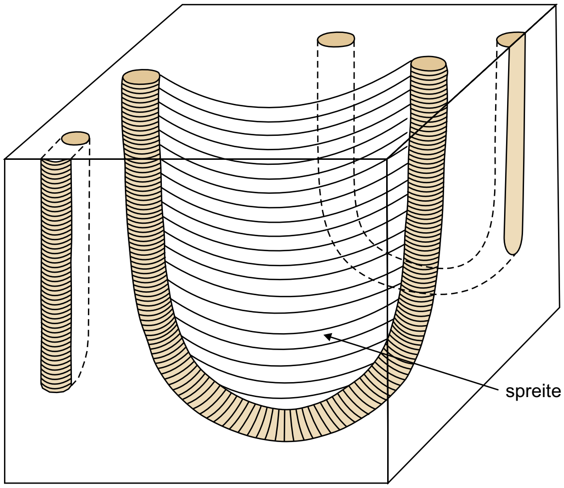


Figure 9.58 A trace fossil with spreiten. The U-tube is the dwelling burrow of an organism. The arms of the U indicate that the organism lengthened and deepened the tube by removing sediment from the floor of the burrow and plastering it against the ceiling, so forming a concentric 'spreite'. These spreiten are protrusive, the present burrow being the last-formed and underlying all previously formed ones.

- A pouch shape, for example caused by the resting of bivalves.
- A network pattern, perhaps due to some systematic activity such as grazing or farming of a sediment surface or interface.

Attempting to relate the trace fossils depicted in Figure 9.59 to these possibilities should help to illustrate how one might go about this categorization.

Classification according to mode of preservation

Diversity of forms of traces arises from the activities of differently shaped animals with different behaviour patterns living in different depositional settings. However, when trace fossils are analysed sedimentologically, it is clear that, although their morphologies typically reflect a wide range of animal behaviour patterns, their preservation usually results from a relatively small number of sedimentary and diagenetic processes. The study of the preservation of trace fossils in relationship to the host sediment is termed **toponomy**.

An awareness of the common modes of preservation is helpful in trying to interpret traces in terms of plant

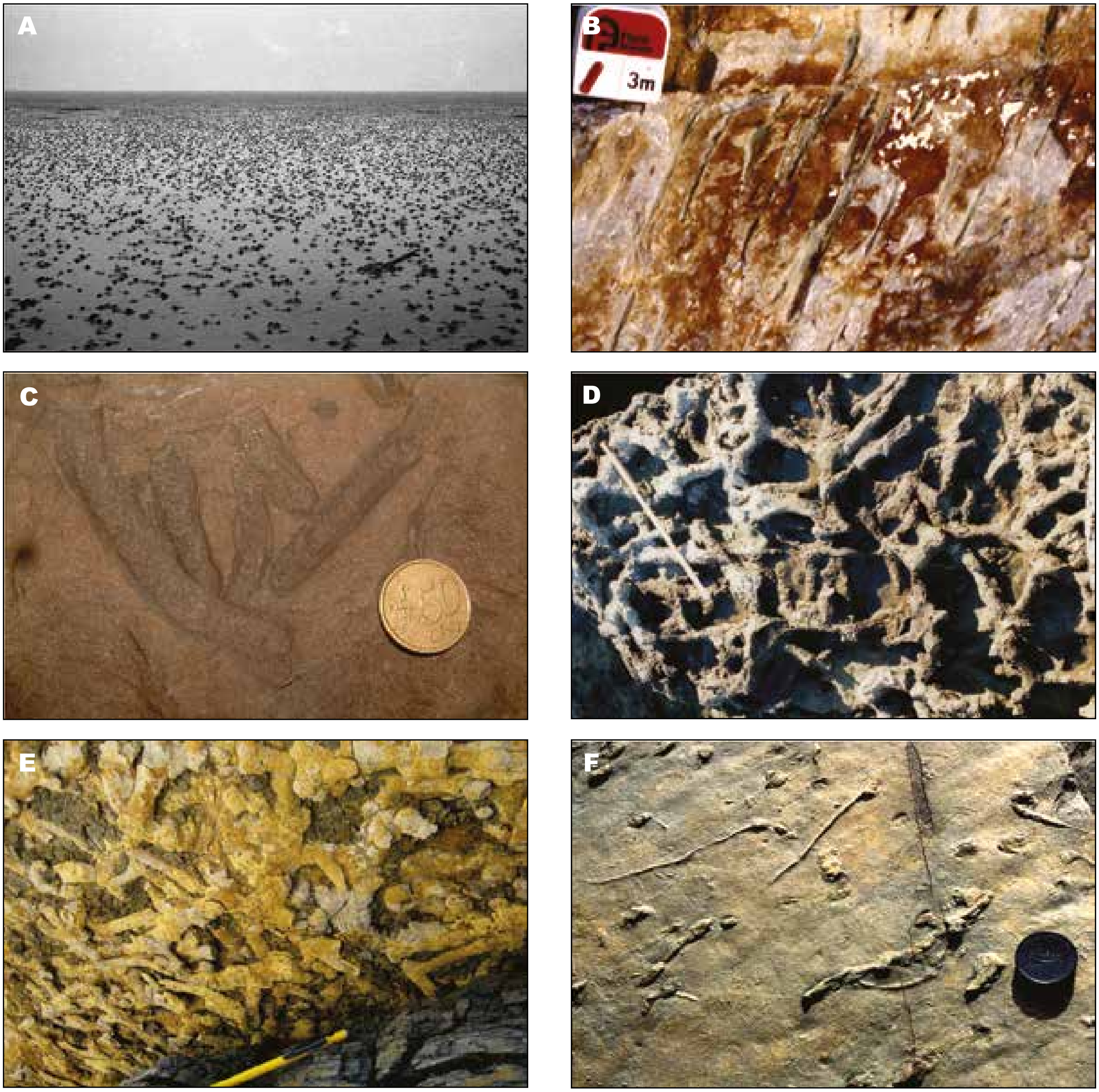
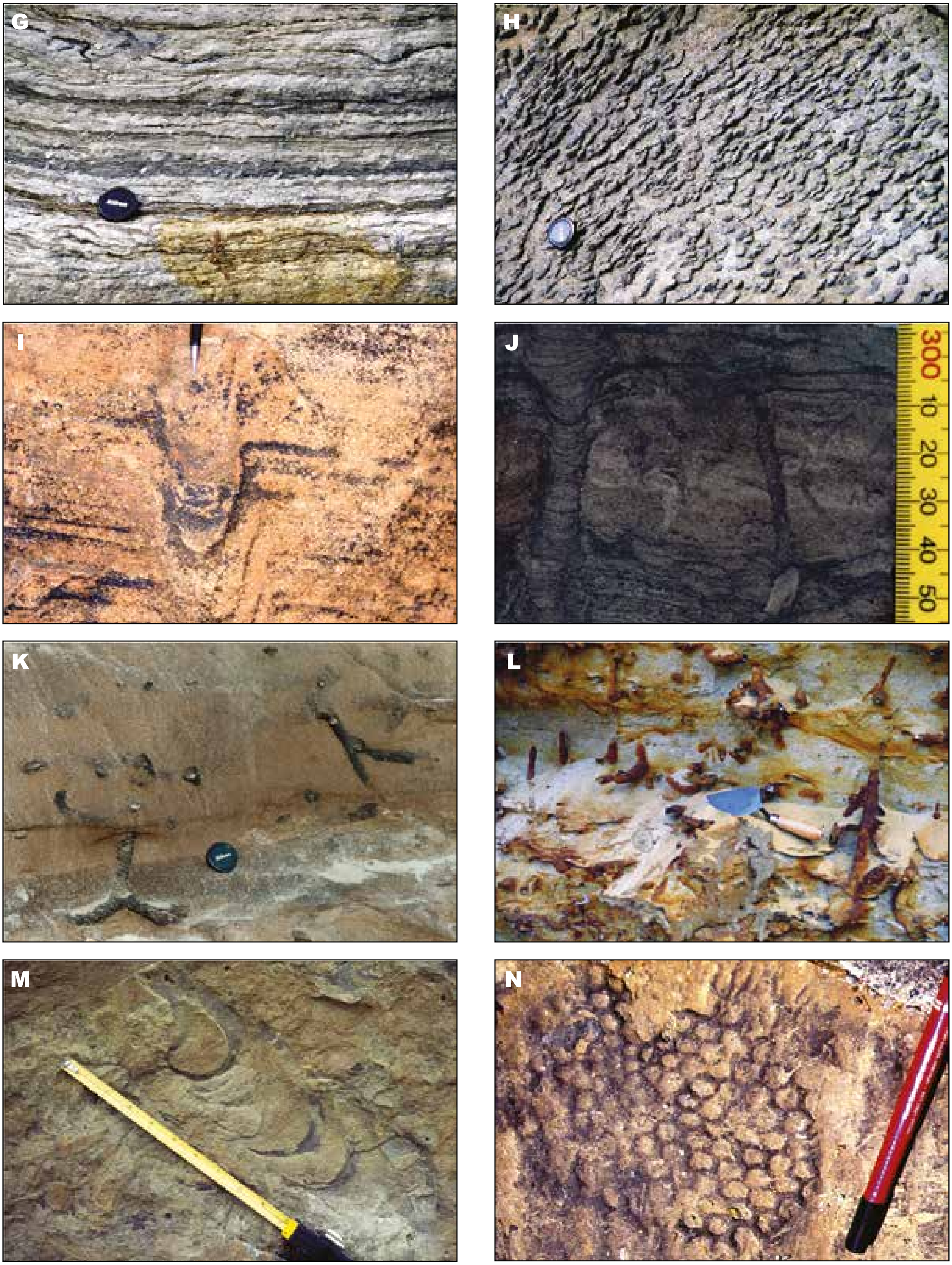


Figure 9.59 Examples of some common types of trace fossils. A) *Arenicola*, modern tidal flat. Haringvliet, Netherlands. B) *Skolithos* in sandstone. Lower Cambrian, northern Greenland. C) *Scoyenia* in sandstone. University of Leeds collection. Locality unknown. D) *Thalassinoides* in marl. Miocene, Nijar, southeast Spain. E) *Thalassinoides* preserved as natural casts on the underside of a sandstone bed. Neslen Formation, Cretaceous, Utah, USA. F) *Lockeia*, casts of a bivalve resting trace on a lower bedding plane surface. Haslingden Flags, Carboniferous, Lancashire, England. G) Interlaminated sandstones and siltstone with variable bioturbation due to vertical burrows; mainly *Lockeia*. Coal Measures. Upper Carboniferous. Northumberland, England. H) Intense bioturbation by closely spaced *Lockeia*. The strongly preferred orientation is parallel with palaeoflow, so that the traces are paleocurrent indicators. Haslingden Flags. Upper Carboniferous, Lancashire, England. I) *Diplocraterion* in shallow marine sandstones. Permian, Namibia. J) *Diplocraterion* with well-developed spreiten. Locality unknown. K) *Ophiomorpha*. Lower Cretaceous, Isle of Wight, England. L) *Ophiomorpha*. Burrows picked out by limonite cement that follows burrow walls. Woburn Sands, Lower Cretaceous. Bedfordshire, England. M) *Rhizocorallium*. Triassic. Spitzbergen, Svalbard. N) *Paleodictyon* in turbiditic calcarenities. Miocene, southeast Spain. O) *Urohelminthoida* in turbiditic calcarenities. Miocene, southeastern Spain. P) *Olivellites* (also known as *Scolicia*), a meandering back-filled burrow preserved as a cast on the underside of a sandstone bed, Central Clare Group, Upper Carboniferous, western Ireland. Photo courtesy of Gilbert Kelling. Q) *Phoebichnus*, a large radiating trace. Hecho Group, Eocene, Pyrenees, Spain. R) *Zoophycos*. Intense bioturbation of a bedding surface, probably recording a period of non-deposition. Alston Formation. Lower Carboniferous, Northumberland, England.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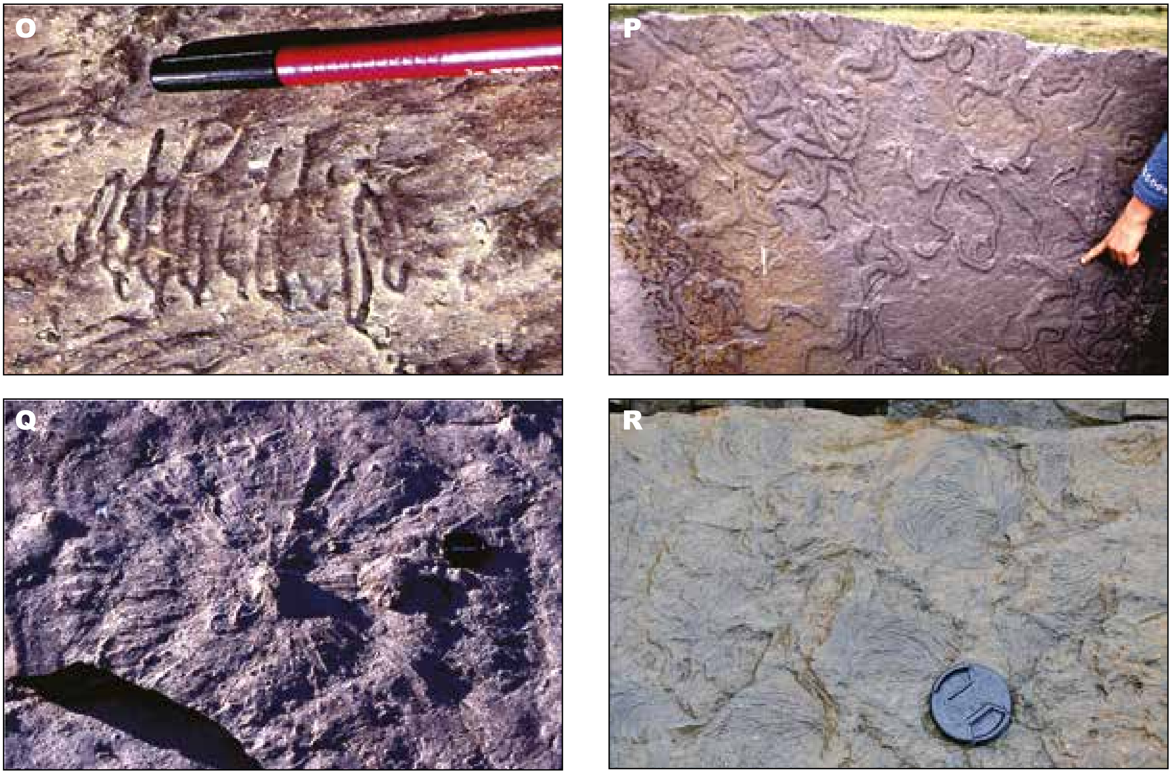


Figure 9.59 *(Continued)*

and animal activity. In trying to understand preservation, remember to look systematically under, within and on top of beds, and for features that cross-cut the bedding, having decided by independent means the way-up of the succession. Bear in mind, however, that the time relationships that hold for the generation of sedimentary structures by physical processes do not necessarily apply to trace fossils; features at the base of a bed or within it may be caused by burrowers that postdate the deposition of the bed by a significant period of time, and the timing of multiple episodes may be hard to establish (Figs. 9.60, 9.61).

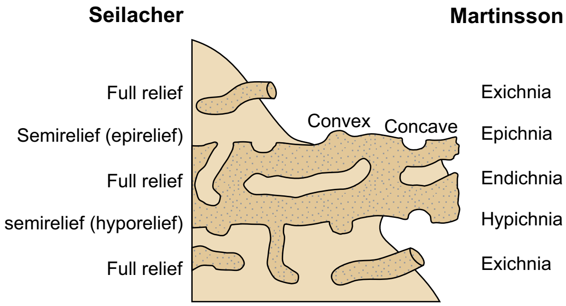


Figure 9.60 Classification of trace fossils in relation to their arenaceous casting medium (Toponomy). Comparison of the terminologies of Seilacher (1964) and Martinsson (1965). Modified after Bromley (1996).

Key questions about the mode of preservation of trace fossils

The questions below (Q. 2–8) serve as a guide to the systematic observation, description, measurement and recording of the style of preservation of trace fossils. Questions 2 and 3 relate to the form of trace fossil preservation, whereas questions 4–8 consider the position and process of preservation. While considering these questions, it is useful to speculate about the possible organisms responsible and processes that gave rise to the traces. The following questions can be used to help determine the mode of preservation of the trace fossils depicted in Figure 9.59.

Q.2 Is the trace fossil preserved as a cast or mould? An impression made in the surrounding sediment by the behaviour of an organism constitutes a mould. The filling of a mould by subsequent deposition, either by sediment

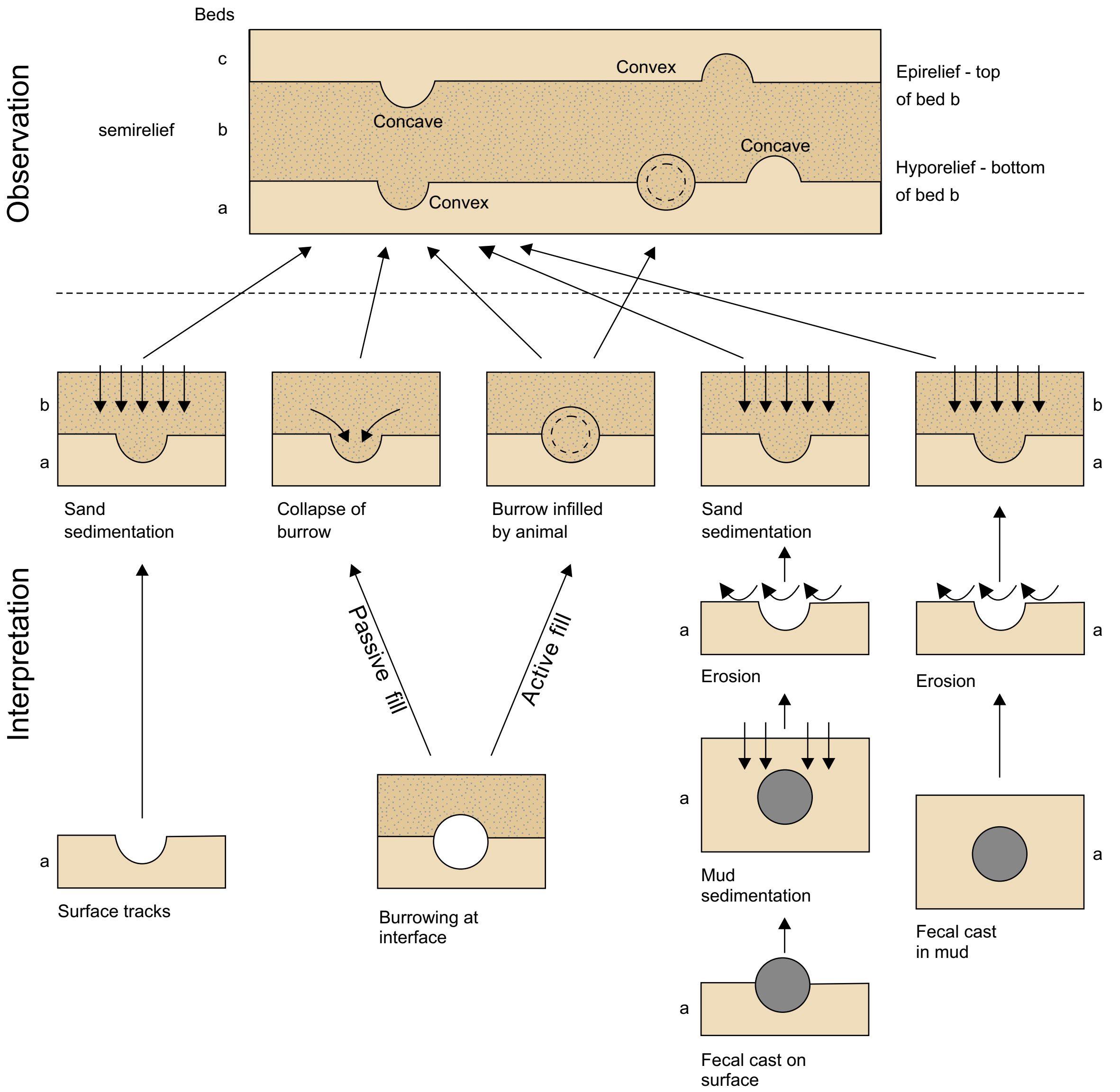


Figure 9.61 A classification of trace fossil preservation types and their interpretation. Modified after Seilacher (1964), Webby (1969) and Hallam in Frey (1975).

from an adjacent bed or by the precipitation of mineral matter, produces a natural cast. To appreciate the difference between moulds and casts try making examples in the laboratory (e.g. a mould and cast of a hand- or footprint, for example) using plasticine and a fill of dental plaster.

Q.3 Is the trace fossil accentuated because it is the site of a diagenetic concretion? *Chondrites*, *Rhizocorallium*, *Thalassinoides* and *Ophiomorpha* are commonly preserved as calcite and siderite nodules in shale or as limonite nodules in sand. Small burrows are commonly preserved as pyrite, which oxidizes to red-brown goethite, as flint or chert (e.g. crustacean burrows in chalk), or in collophane-cemented nodules (e.g. shrimp burrows in present-day conditions). Where diagenesis is less extreme, burrow margins may be conspicuous through having different chemical and physical compositions to the surrounding sediment (Fig. 9.59). Such styles of preservation are usually produced by the animal ingesting clay minerals and secreting colloidal organic compounds rich in Ca, Mg, Na and traces of Cu and Fe, which bind sand grains and faecal clay to the burrow wall. Alternatively, in dark shales produced in reducing conditions, the water-pumping activities of animals may give rise to lighter-coloured "haloes" around the nodular traces. It may be difficult to distinguish burrows preserved in this manner from the traces of plant roots.

Q.4 Is the trace fossil preserved at the top of a sandstone bed that is itself overlain a by another bed of a differing lithology? Trace fossils may be preserved in an interfacial position on the top of a sandstone bed as an **epichnial** trace like a ridge (positive feature) or a groove (negative feature) (Fig. 9.60)? What is its composition of the sediment representing the fill of the trace fossil and that of its surrounding or host sediment? Are there any marks on the top or bottom of the ridges and grooves?

Q.5 Is the trace fossil preserved in an interfacial position on the bottom of the casting medium as a **hypichnial** trace, e.g. a ridge or groove (Fig. 9.60)? If so, is there any evidence that this was a sediment/water interface when the trace was formed? Are only the sub-interface laminae deformed? Was the trace fossil preserved at a sediment/sediment interface (Figs. 9.61, 9.62), possibly between contrasting lithologies, possibly at a concealed junction? What is the composition of the underlying and overlying beds? Are the underlying *and* overlying laminae deformed?

Q.6 Is the trace fossil preserved within a bed but outside the main body of the casting medium as an **exichnial** trace (Fig. 9.60)? Here the traces of one lithology (e.g. sandstone) are isolated in a different lithology (e.g. shale). A sharp upward termination of the fill might suggest a former connection of the burrow fill to a bed of sand which has subsequently been removed by erosion, i.e. a concealed bed junction.

Q.7 Is the trace fossil preserved in an internal position within the main body of the casting medium as an **endichnial** trace (Fig. 9.60)? In such cases the full relief of the trace fossil is preserved. Examples of such features include mud-filled burrows within a sand substrate and burrows in a muddy substrate that are infilled with a contrasting lithology, possibly introduced by the activities of the trace-forming organism itself, or else filled later by physical process or by chemical precipitation processes.

Figure 9.62 Flattened, sediment-filled horizontal burrow systems (*Olivellites or Scolicia*), which show as both positive and negative features on the bedding surface. Liscannor Flags, Central Clare Group, Upper Carboniferous, western Ireland.

Q.8 Is the trace preserved by burial following erosion, i.e. is it a derived trace fossil? This arises when, after burrowing, erosion takes place and currents winnow away soft surrounding sediment to leave, for example, mucus-bound burrow linings as sediment-filled "gloves". These may be covered later by, possibly contrasting, sediments. Alternatively, currents may scour out burrows made in mud and afterwards fill them with sand. Bored pebbles and pieces of bored wood may be reworked as clasts into younger sediment.

Preservation potential

Processes of sedimentation strongly influence the trace-fossil assemblage that is preserved, thereby producing a physically induced bias. The majority of biogenic traces, particularly the epichnial ones on the upper surface of beds, have almost zero fossilization potential. Trace-fossil associations dominated by surface or near-surface traces may indicate near continuous sedimentation; those dominated by traces made at depth below the sediment surface may indicate discontinuous sedimentation with interspersed periods of erosion. However, caution in interpretation is required here since it is common for several ichnogenera to develop contemporaneously at different levels within the substrate (See tiering, §9.4.3). Many burrowing organisms progressively excavate and backfill their burrows as part of their life process. The backfilling process creates curved, sub-parallel laminae called **spreiten** (Fig. 9.58), which may be aligned in either a horizontal or a vertical plane and which record the passage of the organism through the substrate. **Protrusive spreiten**, marking sets of burrow fills where the last-formed burrow underlies all earlier ones (Fig. 9.58), reflect adjustment of the animal's position in response to erosion at the sediment surface or to growth. **Retrusive spreiten**, or nested cones marking sets of burrow fills where the last-formed burrow overlies all previous ones, reflect adjustment to ongoing sedimentation as the animal moves upwards. Preservation of delicate individual structures made by some organisms depends on a lack of later, more general, bioturbation. They may be referred to as 'elite' trace fossils. The preservation potential of shallow or near-surface traces is greatest in low-energy settings.

In order to consolidate the ideas developed so far, attempt to use combinations of the terms introduced above to describe and at least start to interpret the features depicted in Figure 9.63. Attempt to describe and interpret the broad sequence of events that gave rise to the preservation of the traces shown in Figure 9.64, including the recognition of protrusive and retrusive spreiten.

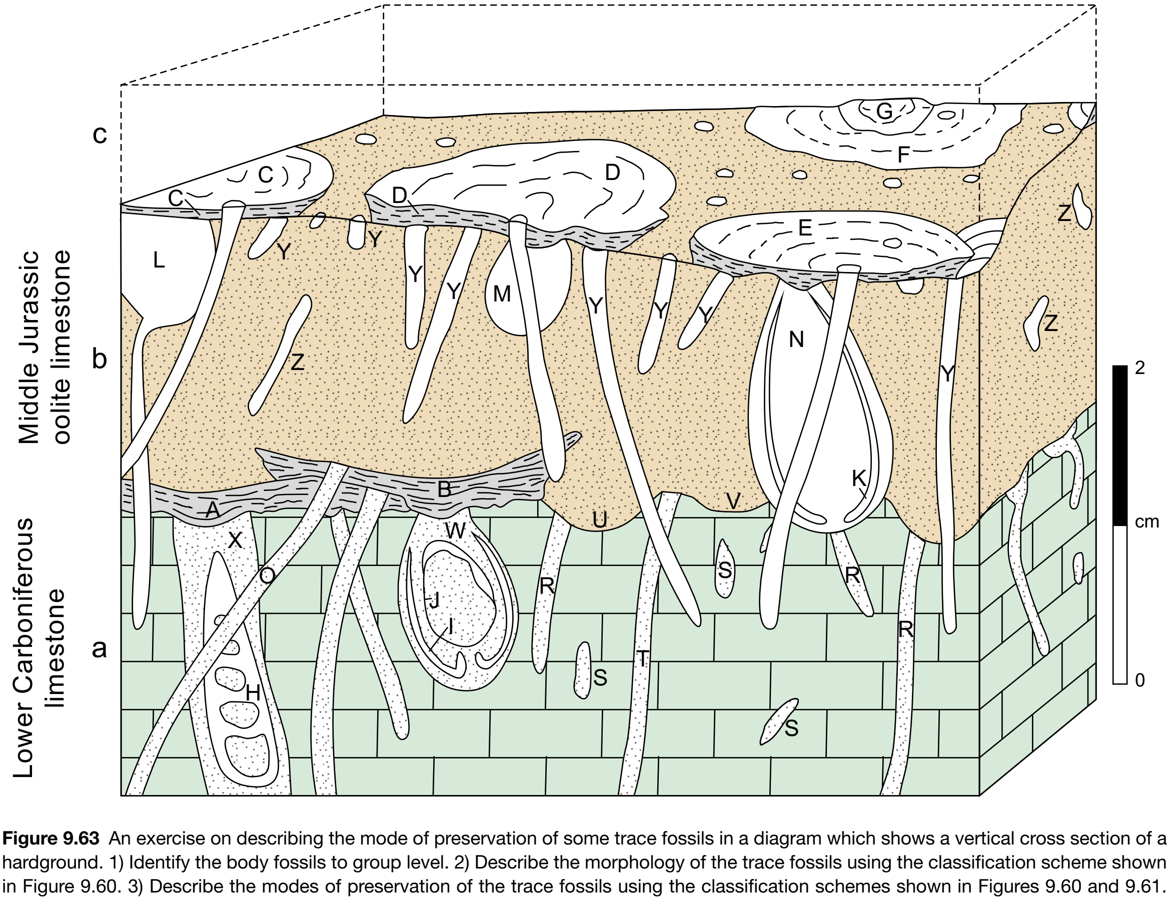


Figure 9.63 An exercise on describing the mode of preservation of some trace fossils in a diagram which shows a vertical cross section of a hardground. 1) Identify the body fossils to group level. 2) Describe the morphology of the trace fossils using the classification scheme shown in Figure 9.60. 3) Describe the modes of preservation of the trace fossils using the classification schemes shown in Figures 9.60 and 9.61. 4) Describe the mode of behaviour of the trace fossils using the classification schemes shown in Figures 9.64-9.67. Determine the stratigraphic history of the rocks depicted in the diagram using the letters a-c to identify the beds and A-Z to identify the body and trace fossils. Modified after Bromley in Frey (1975).

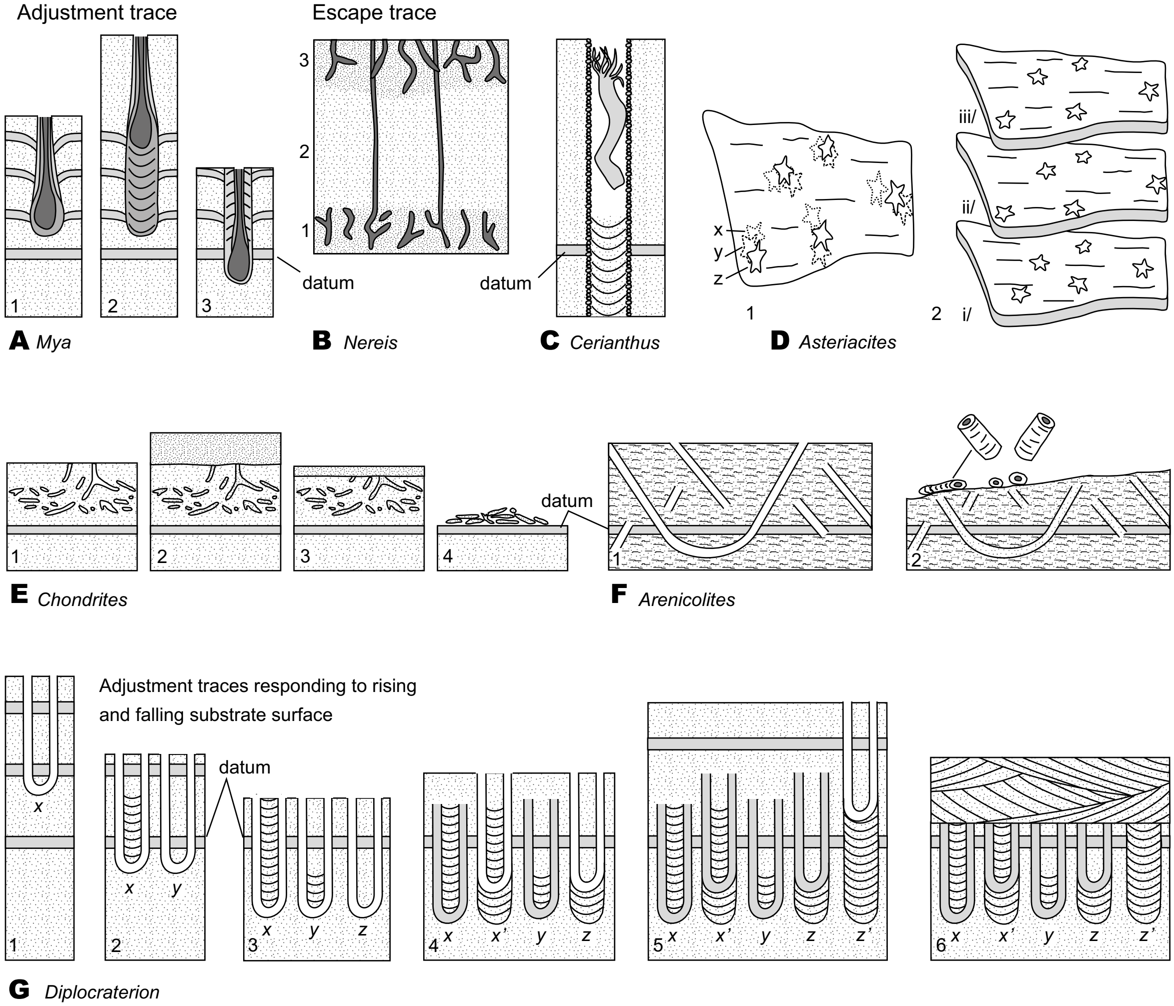


Figure 9.64 Exercise to demonstrate the use of trace fossils for determining amounts of sedimentation and erosion in clastic successions. The diagrams represent single or closely related sequences of vertical successions. For examples A-G, explain the sequence of events that has given rise to the preservation of each of the sets of trace fossils. In many cases a lithological datum plane (time-marker) is shown. The movement patterns relate to: A) the bivalve *Mya* which has a single siphon; B) the polychaete worm *Nereis*; C) the sea anenome *Cerianthus*, an organism that dwells in a single tube and produces a pattern similar to traces such as *Skolithos* or *Monocraterion*; D) a resting trace of a starfish (*Asteriacites*). Exposure 1: a bedding plane with groups of traces; exposure 2: three bedding planes (oldest = i); E) the preservation patterns of *Chondrites*, believed to be the work of sediment-feeding organisms; F) the preservation pattern of *Arenicolites curvatus*, believed to be the work of a suspension-feeding organism, possibly a worm; G) the movement pattern *Diplocraterion yoyo* (after Goldring, 1964). Modified after Howard in Basan (1978).

Classification according to behaviour (an ethological classification)

Certain types of trace-producing behaviour are common to several groups of organisms. Twelve general patterns of behaviour are recognized and these are shown diagrammatically in Figure 9.65 and defined in Figure 9.66, although there is some overlap between the basic categories (Fig. 9.67). None-the-less, behavioural considerations are useful in understanding the origins and interrelationships of both fossil and recent traces.

Key questions concerning the behaviours of trace-producing organisms

Questions 9–21 focus attention on possible behaviour patterns. These can be tested against experience of how present-day organisms graze, crawl, rest, burrow, feed at depth

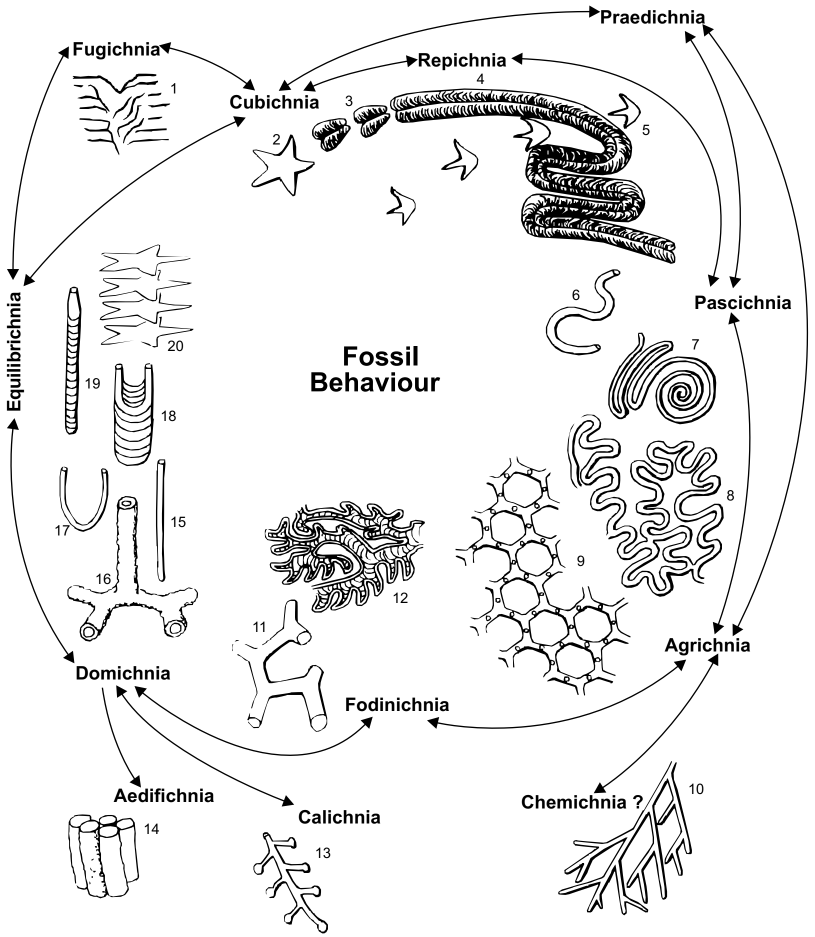


Figure 9.65 The ethological classification of trace fossils. Interrelationships between 12 major groups are suggested with arrows. It is debatable whether chemosymbiont structures should be grouped with Agrichnia, or be separated as suggested here (Chemichnia). Representative ichnogenera and other trace fossils indicated are: 1) escape structure; 2) *Asteriacites*; 3) *Rusophycus*; 4) *Cruziana*; 5) bipedal vertebrate trackway; 6) *Helminthopsis*; 7) *Helminthoida*; 8) *Cosmorhaphe*; 9) *Paleodictyon*; 10) *Chondrites*; 11) *Thalassinoides*; 12) *Phycosiphon*; 13) beetle brooding burrow; 14) sabellarian sand tubes; 15) *Skolithos*; 16) *Ophiomorpha*; 17) *Arenicolites*; 18) *Diplocraterion*; 19) bivalve adjustment trace; 20) vertically repeated *Asteriacites*. Modified after Bromley (1996) and Goldring (1999).

and escape in relation to a whole series of interacting controlling factors. Possible controlling factors include grain size, energy levels, packing, porosity, permeability, pH, Eh, salinity, degree of cementation, the availability of organic matter and nutrients, and the risk of predation.

Q.9 Could the structures be the legacy of plant roots in the form of root moulds, root casts, or petrified roots (rhizoliths) (Figs. 9.68, 9.69)? At first sight, some preserved plant-root structures may be confused with traces made by animals. Consider the palaeoenvironmental context of the host sediment in which the suspected root structures are found. Do the structures take the form of branching root networks? Is there physical and/or chemical evidence of pedogenesis (soil formation), including destratification, colour mottling, precipitation of calcrete, silcrete or gypcrete cements or nodules (see §9.3.1)? Root structures are commonly associated with carbon films and coal seams in waterlogged palaeosols. Are the structures found in association with leaf impressions?

Q.10 Could the traces be the result of animals temporarily interrupting their movement on or above the sediment surface to rest or seek refuge? Isolated, shallow, trough-like depressions may record the outline or morphology of

Definition	Characteristic morphology	Examples
Resting traces (Cubichnia) Shallow depressions made by animals that settle onto or dig into the substrate surface. Emphasis is on reclusion. May include shallow, ephemeral domiciles.	Trough-like relief, recording to some extent the morphology of the animal. Ideally structures are isolated but they may intergrade with crawling traces or escape structures.	*Asteriacites* *Lockeia* *Rusophycus*
Crawling traces (Repichnia) Trackways and epistratal or intrastratal trails made by travelling organisms. Emphasis is on locomotion, though secondary activities may be involved.	Linear or sinuous structures, some branched. Footprints or continuous grooves, commonly annulated. Complete form may be preserved.	*Aulichnites* *Cruziana* *Diplichnites* *Scolicia*
Grazing traces (Pascichnia) Grooves, patterned pits and furrows, many of them discontinuous, made by mobile deposit feeders or algal grazers at or under the substrate surface. Emphasis is on feeding behaviour analogous to 'strip mining'.	Unbranched, non-overlapping, curved to tightly coiled patterns or delicately constructed spreiten dominate. Patterns generally reflect maximum utilization of food resources. Complete structure may be preserved. Overall structure tends to be planar.	*Helminthoida* *Lophoctenium* *Nereites* *Spirophycus*
Feeding traces (Fodinichnia) Temporary burrows constructed by deposit feeders; the structures may also provide shelter for the organisms. Emphasis is on feeding behaviour analogous to 'underground mining'. May be gradational with dwelling structures.	Single branched or unbranched cylindrical to sinuous shafts or U-shaped burrows, or complex, parallel to concentric burrow repetitions (spreiten structures). Walls not commonly lined, unless by mucus. Oriented at various angles with respect to bedding.	*Gyrophyllites* *Phycodes* *Rosselia*
Dwelling structures (Domichnia) Burrows, borings or dwelling tubes providing permanent domiciles, mostly for suspension feeders and carnivores. Emphasis is on habitation, though secondary activities may be involved.	Simple, bifurcated or U-shaped structures perpendicular or inclined at various angles to bedding, or branched burrow or boring systems having vertical and horizontal components. Burrow walls typically lined.	*Diplocraterion* *Ophiomorpha* *Skolithos* *Trypanites*
Escape/Adjustment structures (Fugichnia/Equilibrichnia) Structures arising as a consequence of substrate degradation or aggradation. Emphasis is on escape or readjustment for the maintenance of an equilibrium between relative substrate position and the configuration of contained traces.	Vertically repetitive resting traces. Biogenic laminae either *en echelon* or as nested funnels or chevrons. U-in-U spreiten burrows and other structures reflecting displacement of animal upward or downward with respect to the original substrate surface. Complete form may be preserved, especially in aggraded substrates.	Nested funnels U-in-U spreite Down-warped laminae
Farming structures (Agrichnia) Regularly patterned burrow systems in which the activities of permanent dwelling and feeding are combined. Emphasis is on both habitation and feeding using farming or trapping strategies.	Horizontal tunnels organised in complex, regular geometric patterns such as meanders, spirals and hexagonal meshworks. Complete forms may be preserved.	*Belorhaphe* *Paleodictyon* *Spirorhaphe*
Predation traces (Praedichnia) Traces resulting from predation, usually bio-erosion structures produced on hard biological materials such as shell, bone or coral. Gnawings and other bite traces are included in this group. Emphasis is on feeding by predation. Predator-prey relationships may be discernable.	Simple, circular holes drilled in shells by carnivorous gastropods or cephalopods. Bite marks on ammonites made by aquatic reptiles. Distinctive patterns of chipped margins of gastropod and bivalve shells that have been attacked by crabs. Structures tend to be isolated.	*Centrichnus* *Oichnus*
Supra-substrate dwelling structures (Aedifichnia) Structures constructed above the original substrate and used as dwelling structures.	Simple tubes to complex multi-chambered structures. Wasp nests, termite mounds etc.	Sabellarian tubes Termite nests
Breeding/juvenile nursery structures (Calichnia) Structures for breeding and/or raising juveniles in a protected environment.	Simple to complex multi-component burrows with chambers for eggs, larvae or juveniles.	Beetle brooding burrows Bee and ant nests
Biochemical symbiosis structures (Chemichnia) Structures influenced by microbial-chemical interactions, often where low oxygen levels prevail.	Complex multi-branched burrows, often with a radiating architecture.	*Chondrites*

Figure 9.66 Ethological classification of invertebrate trace fossils. Modified in part from Frey and Pemberton (1985) and Pemberton et al. (1992).

the under-surface of an animal, or marks caused by its digging or temporarily settling into a stationary position in the substrate, or burying itself just under the sediment surface. Such resting traces, called **Cubichnia**, are made by epibenthic mobile animals (e.g. starfish, bivalves, arthropods like crabs, and flat fish). Trace fossil examples are *Asteriacites*, *Lockeia, Rusophycus* (Fig. 9.70). These traces are transitional to crawling, dwelling and escape structures (Fig. 9.67).

Q.11(a) Could the traces be produced by movement of animals on the sediment surface or along an interface? Features that might suggest this are footprints, trackways, grooves, trails, horizontal, epistratal or intrastratal burrows. Their linear, sinuous or branched forms reflect movement of crawling or walking limbs, bristles or other appendages, or the muscular movements of a body, or the dragging of a shell. Could different, but apparently related, traces be made by the same animals walking, running, galloping, hopping, crawling, half-walking, half-swimming, or fully-swimming? Could the traces be caused by an animal being drifted off course, whilst it first lost and then regained a direction of movement in the face of hostile currents? Examples of such **Repichnia** or locomotion traces are provided by arthropod and tetrapod tracks, e.g. *Diplichnites*, *Kouphichnium* and *Chirotherium* (Figs. 9.65–9.67).

Q.11(b) Could the traces be made just below the surface of sediment, especially in silt or sand? Such Repichnia are made by benthos and nektobenthos, predators, scavengers and deposit feeders (such as snails). Examples are *Gyrochorte* and *Cruziana* (Fig. 9.70E). Repichnial traces are often transitional to resting traces and surface grazing traces (Figs. 9.65, 9.67).

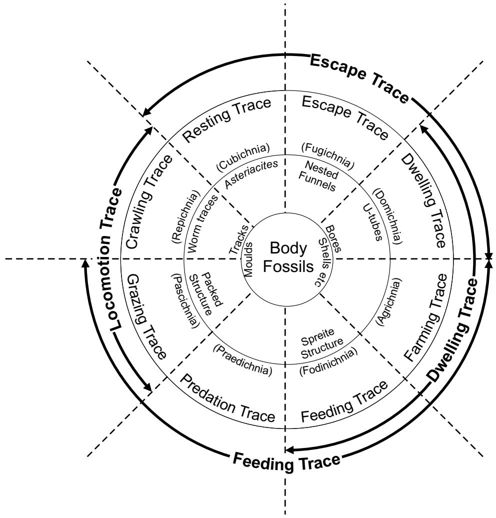


Figure 9.67 Ethological (behavioural) classification of trace fossils, and their relationship to body fossils. Overlap of categories acknowledges the integrations inherent in nature. Modified from Pemberton et al. (1992). Additional, more specialised behavioural groups not mentioned on this diagram include Aedifichnia, Calichnia, Chemichnia and Equilibrichnia (see text for details).

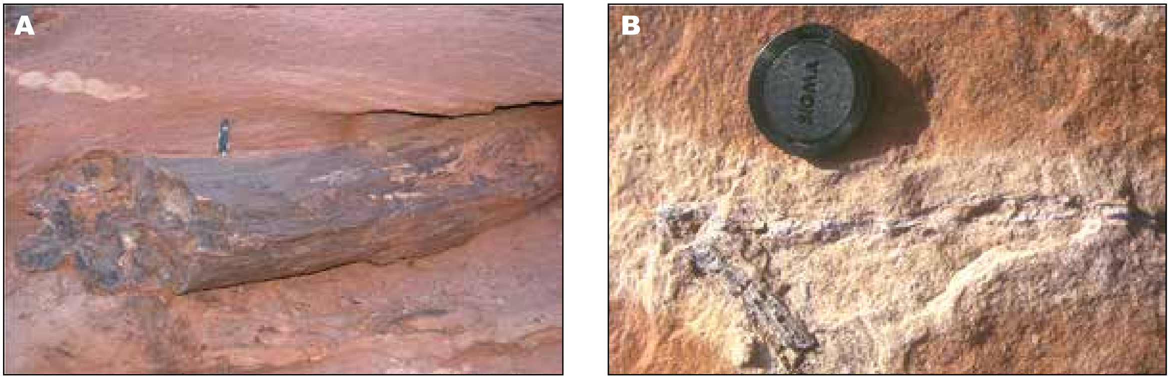


Figure 9.68 Biogenic plant structures. A) Silicified tree trunk preserved in fluvial channel strata, Lower Cutler Beds, Pennsylvanian, southeast Utah, USA. B) Calcified tree root preserved in aeolian strata, Cedar Mesa Sandstone, Permian, southeast Utah, USA. C) Plant root structures with iron concretions seen in plan-view, Holocene, Iceland. D) Tree stump with radiating roots, Marsden Formation, Upper Carboniferous, Lancashire, England. E) Silicified tree stump in sandstone of aeolian dune origin. Navajo Sandstone, Jurassic, Utah, USA. F) Silicified tree stump in siltstone of fluvial floodplain origin. Villar del Arzobispo Formation, Jurassic, Teruel, eastern Spain. G) Silicified plant stem debris forming a lag of intraformational clasts in the base of a fluvial channel-fill deposit. Undifferentiated Cutler Group, Permian, Utah, USA. H) *Neuropteris*, a seed fern, preserved in mudstone of fluvial floodplain origin. The delicate nature of the specimen indicates preservation in a very low-energy environment. Upper Carboniferous (Stephanian) Cantabria, Spain.

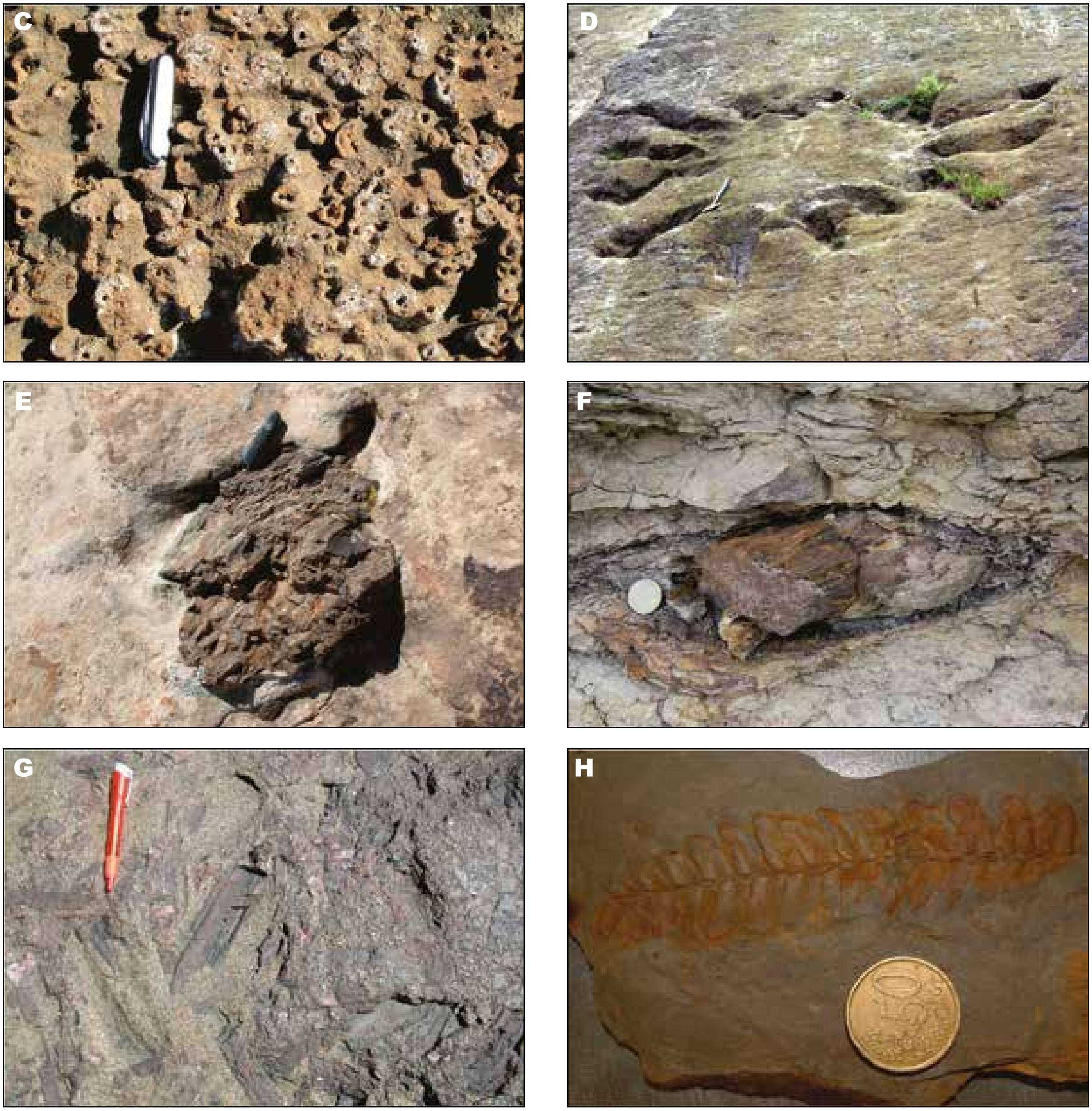


Figure 9.68 (Continued)

Q.12 Could the traces be due to systematically organized surface or horizontal-interface grazing? Are the traces planar and on the surface? Do they show discontinuous, systematically patterned meanders, loops, spirals and networks which are non-branched, or only occasionally branched patterns? Are they non-overlapping, curved to tightly coiled, grooves, pits and furrows that may relate to delicately constructed spreiten? Could they result from "strip mining" by a mobile deposit-feeding or algae-grazing animal to ensure the economical exploitation of food resources in the sediment? Such surface grazing traces are made by mobile epibenthos (e.g. grazing gastropods such as limpets, worms, echinoids and arthropods) and are known as **Pascichnia**. Examples are *Helminthoida*

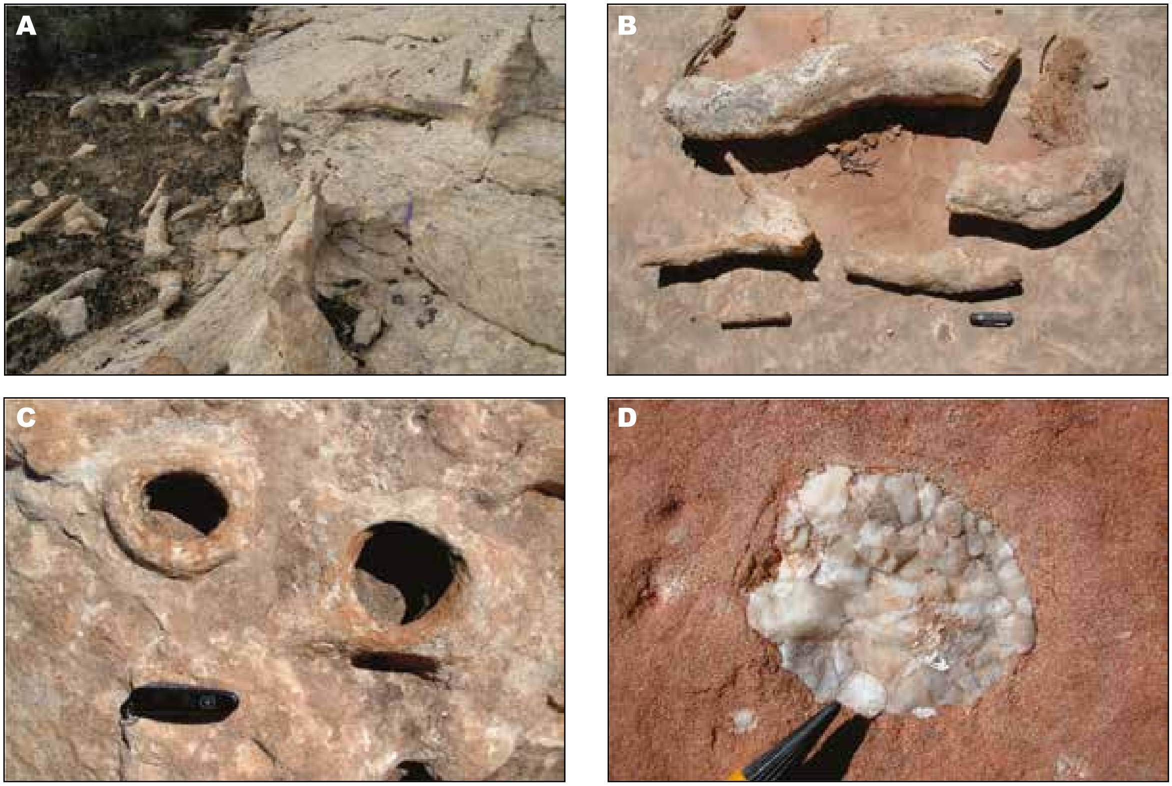


Figure 9.69 Biogenic plant structures preserved by chemical replacement. A) In-situ tree stumps preserved as calcite-cemented sandstone in otherwise less-well cemented sandstone of aeolian dune origin. Navajo Sandstone, Jurassic, Utah, USA. B) Weathered-out tree root structures preserved as calcite-cemented sandstone. C) In-situ tree root structures preserved as moulds with calcite-cemented sandstone rims. D) Smaller in-situ plant-root structures replaced by calcite crystals. End-section view. Examples B, C and D are from the Cedar Mesa Sandstone, Permian, Utah, USA.

(Fig. 9.59), *Lophoctenium, Nereites*, *Spirophycos* and planar types of *Zoophycos* (Fig. 9.70).

Q.13 Could the traces result from animals systematically farming the flora and fauna on the sides of a tunnel system? Do the traces form burrows that are systematically patterned and non-overlapping? Are the activities of permanent dwelling and feeding combined? Such traces are referred to as farming and trapping traces: **Agrichnia** (Figs. 9.65, 9.67). The complex, regular meanders, loops, spirals, nets and hexagonally, polygonally branched patterns (e.g. *Paleodictyon, Cosmorhaphe* and *Spirorhaphe*) are some examples (Figs. 9.59, 9.70). Relating to the mode of preservation, might such traces occur at the top surface of a stratum or on the base of an overlying event bed because subsequent erosion has been able to cut down only as far as a plane of indurated burrow fills?

Q.14 Could the traces be burrows excavated while animals searched for food within the sediment? Features that support such an idea are single, branched or unbranched, cylindrical to sinuous, radial shafts or U-shaped burrows, orientated at various angles to the bedding. There may also be complex, parallel to concentric, horizontal burrow repetitions (spreiten) with unlined walls (Fig. 9.65). The structures result from short-lived, ephemeral burrowing by animals that were "underground miners", i.e. were essentially deposit-feeders which, in addition, sought secure shelter and refuge. An important feature is that the radial or U-shaped burrows do not touch one another since animals generally avoid previously mined sediment on account of its toxicity. These traces are made by epibenthic and endobenthic deposit feeders like polychaete worms, and are known as **Fodinichnia**. These traces are related to

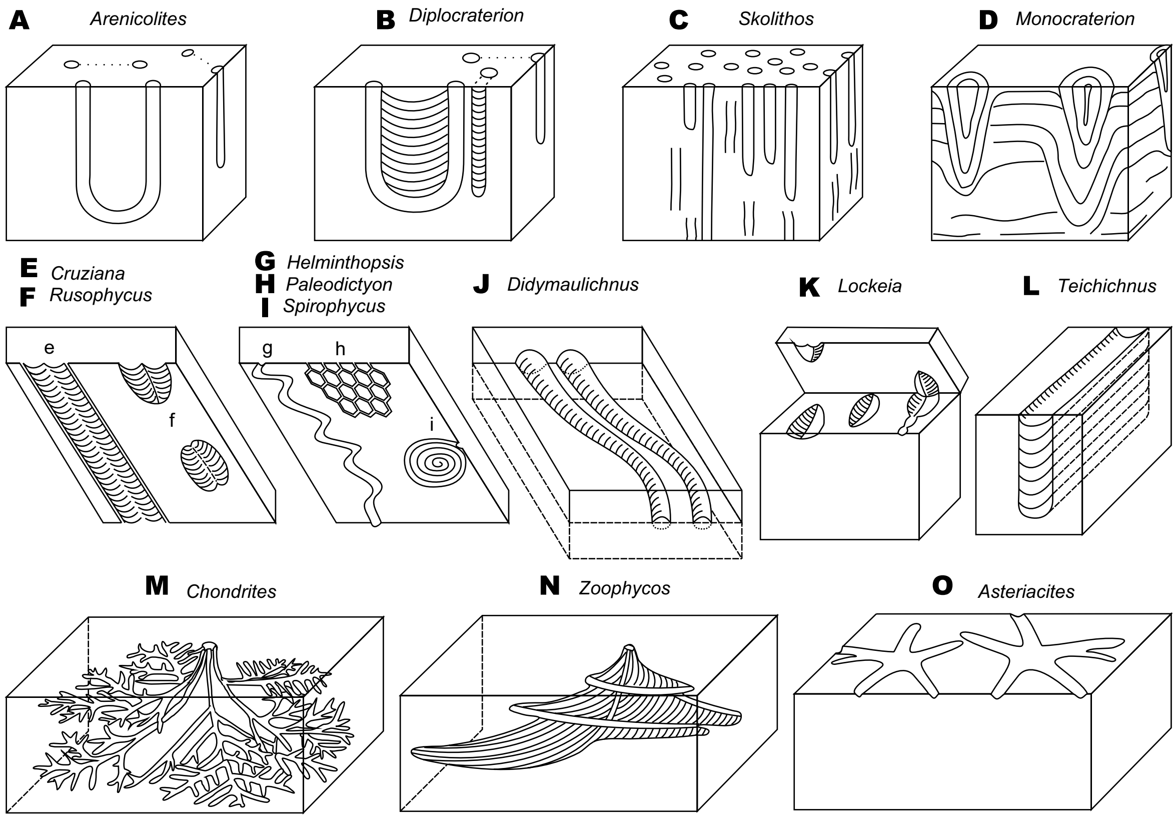


Figure 9.70 Exercise to work out the mode of behaviour of some common trace fossils. Describe the morphology of each of the trace fossils and their relationship to the substrate in or on which they are developed. Classify them into one or more of the ethological (behavioural) groups described in the text and in figures 9.65, 9.66 and 9.67. Suggest an animal group that might be responsible for generating each of the traces. Indicate whether each trace fossil could be used as a way-up indicator. Modified after diagrams by Crimes in Frey (1975) and Basan et al. (1978).

grazing (Pascichnia) and dwelling (Domichnia) structures (Figs. 9.65, 9.67). Examples are *Phycodes*, *Rhizocorallium*, *Zoophycos* (Fig. 9.70), *Gyrophyllites* and *Rosselia*.

Q.15 Could the structures be the dwellings of suspension feeders or carnivores? Features that suggest this are simple, bifurcated or U-shaped burrows perpendicular or inclined to the bedding, or branching burrows having vertical and horizontal components. Some traces lack spreiten and have mucus-cemented sand, silt or clay-lined walls. They may be borings or burrows that were more or less permanently occupied by suspension feeders or active carnivores. These animals strengthened the walls of their homes by lining them but did not backfill them. These forms are produced by suspension feeders (e.g. shrimps), or by predator tube-dwelling worms or arthropod scavengers (e.g. crabs) and are known as **Domichnia**. They are closely related to feeding (Fodinichnia) and resting (Cubichnia) traces (Figs. 9.65, 9.67). Examples are *Skolithos*, *Arenicolites*, *Diplocraterion*, *Thalassinoides, Trypanites* and *Ophiomorpha* (Figs. 9.59, 9.70).

Q.16 Could the structures have resulted from the upward or downward movement of an organism escaping adverse conditions at the sediment surface? Are the traces roughly cylindrical, sub-vertical and lacking a lining? Are they vertically repetitive resting traces with laminae concentric, *en echelon* or forming chevrons or nested funnels? Are there “U-in-U” spreiten in the burrows or other types of structures, which could reflect the displacement of semi-sessile suspension feeders (e.g. bivalves) upwards or downwards with respect to the original sediment surface as a response to erosion or sedimentation? Complete forms may be preserved in beds that record an episode of aggradation. These escape traces are classed as

Fugichnia (Figs. 9.65, 9.67) and there is overlap with resting (Cubichnia) and dwelling (Domichnia) burrows, and rare transitions to feeding and grazing structures (Fodinichnia and Pascichnia), though the last-named are unlikely to reveal evidence for rapid response to changes of interface position. Examples are found in retrusive *Diplocraterion* and *Lockeia*, in *Monocraterion*, in *Skolithos* which possess funnels or retrusive bases, and in elongate *Ophiomorpha* (see Figs 9.59, 9.60). This type of behaviour has overlap with that of trace makers that were seeking equilibrium, with an emphasis on re-adjustment of position in the substrate (see Q. 17 below). However, escape traces typically preserve evidence to indicate rapid evacuation.

Q.17 Could the traces result from animals within the substrate constantly adjusting their burrows to an equilibrium level in relation to gradually aggrading and degrading sediment and water levels of seas or lakes? Such accommodation traces are **Equilibrichnia** (Fig. 9.65) Examples include bivalve adjustment traces, *Skolithos* and *Diplocraterion* which display finely spaced retrusive or protrusive spreiten (Figs 9.59, 9.60, 9.64, 9.70), suggestive of repeated minor adjustment of the level of the organism in relation to the bed surface.

Q.18 Could the traces result from predator-prey relationships, i.e. by bio-erosion on hard materials e.g. bone, shell, coral etc? Could they represent gnawings and bitings, or circular holes drilled by gastropods or cephalopods through the shells of other living organisms such as bivalves? Bite marks on ammonites were probably made by aquatic reptiles, and the chipped margins of shells may result from attacks by organisms such as crabs. Might the traces have arisen from a predator stalking and then capturing its prey, for example two sets of foot prints, one recording the prey animal and the other recording the chasing predator? These are known as **Praedichnia** (Figs. 9.65, 9.67) and should be distinguished from Fodinichnia. Examples are *Centrichnus* and *Oichnus*.

Q.19 Have the structures been constructed above the original substrate e.g. as mud-dauber wasp nests, termite colonies, or as the structures of certain marine worms (polychaetes)? Such rare cases are **Aedifichnia** (Fig. 9.65). Ancient examples of such traces may be preserved where the structures (e.g. termite mounds) are well cemented such that they resist erosion as sedimentation takes place around them until they are buried.

Q.20 Could the structures be breeding places for protecting larvae or raising juveniles e.g. bee cells, dinosaur egg nest sites or beetle brooding burrows? These rare structures are known as **Calichnia** (Fig. 9.65).

Q.21 Could the structures be influenced by microbial-chemical interactions where low levels of oxygen prevailed? Such traces are known as **Chemichnia** (Fig. 9.65) and may be related to trapping and gardening (**Agrichnia**). Examples include complex feeding traces like *Chondrites* (Fig. 9.70).

To familiarize yourself with some of these terms, features and questions, try to sort the trace fossils depicted in Figures 9.63, 9.64 and 9.70 into one or more of these behavioural categories and, where appropriate, interpret a sequence of events. This is, however, no substitute for doing the same thing in the field or the laboratory.

9.4.3 Tiering

Several types of trace fossils are often found in close association with each other and indicate complex interactions between several groups of organisms. The plant and

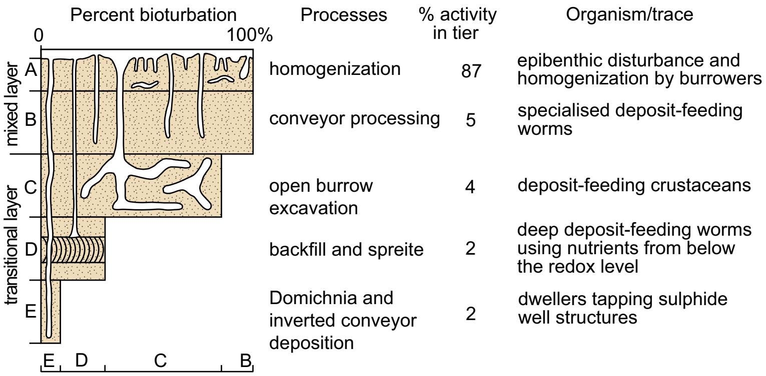


Figure 9.71 A generalized tiering bioturbation model indicating five levels of activity (A-E). For each tier the usual quantity and type of sediment turnover is suggested, together with the proportion of the whole community that may be active in each tier and the type of organism or trace present in each tier. The scale at the bottom indicates the expected representation of the tiers in the final preserved ichnofabric. Modified after Bromley (1996).

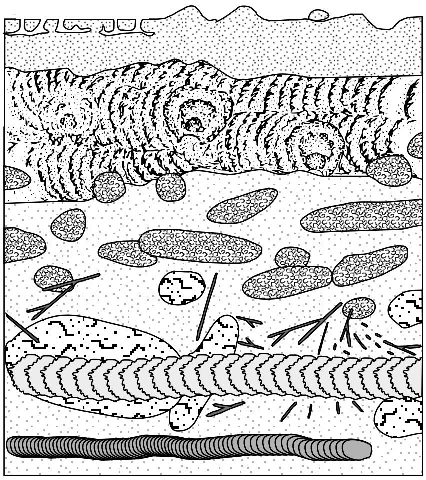


Figure 9.72 Tiering diagram of traces seen in box cores taken in a water depth of 2 to 3.5 km offshore northwest Africa. Surface trails and *Paleodictyon* isp. are indicated in the mixed layer. Modified after Wetzel (1984).

animal activity that takes place on and below the sediment surface within a particular environmental setting is typically characterized by an assemblage of trace-generating organisms which each occupy a particular level or **tier** on or within the substrate (Figs. 9.71, 9.72). The infauna occupies the host sediment in tiers in order to accommodate their different body sizes and numbers, and their differing modes of feeding, respiration and survival. When sediment aggradation occurs, organisms move upwards to re-establish life in relation to the new "equilibrium" surface and conditions. As a result, tiers of traces move upwards so that lower ones become superimposed upon higher, earlier formed tiers and thus potentially overprint those higher structures. Figure 9.73 shows the effects of tiered bioturbation on different hypothetical regimes. In comparing modern and ancient environments, it is important to realize that a biologist mainly sees traces of activity created by tiers on or just under the sediment surface. A geologist, by contrast, sees rock outcrops dominated by trace fossils mainly created in, and selectively preserved from, deeper tiers.

You should attempt to construct a tier diagram from evidence seen in outcrop by:

(a) Observing and carefully drawing a representative section of an outcrop, paying particular attention to all cross-cutting relationships.
(b) Photographing the outcrop and individual beds at various scales.
(c) Cleaning up outcropping beds and breaking up hand specimens (N.B. Not breaking up the outcrop).
(d) Selecting and securely labelling critical blocks with a known orientation for slabbing in the laboratory, but only as part of a serious research study, and with appropriate permission.

9.4.4 Ichnofabric and bioturbation

The presence and relationships of trace fossils in a sediment unit makes up its **ichnofabric. Bioturbation** is the alteration of the original sediment, which is inferred to result from the activity of animals and plants on, and within, the host sediment whilst living there. Because trace fossils are sedimentary structures that, in nearly all cases, formed exactly where they are found, they represent responses to the physical, chemical and organic nature of the substrate and are therefore sensitive indicators of prevailing and subsequent environmental conditions. Ichnofabrics therefore can provide valuable information regarding food supplies, the nature of the sediment and its deposition (grain size, sorting, permeability, porosity, sedimentation rates, especially of rare events), temperature, bathymetry, intensity of waves and currents, current directions, periods of episodic but temporary erosion, predators, skeletal degradation, oxygen levels, salinity values and subsequent diagenesis.

In adopting a systematic approach to describing the type of ichnofabric present, several general questions might be asked:

- Are the burrows very densely distributed and interpenetrating? If so, the sediment should be referred to as having a **bioturbated texture** or **ichnofabric**.
- Are the burrows common but indistinct? If so, the term **burrow mottling** may be more appropriate (see Fig. 9.59).
- Are the structures preserved in full relief (Fig. 9.60)?
- Is the wall of the cast of a different composition to the body of the cast, as when a burrow in sand is lined by a layer or layers of mucus and/or faecal pellets made of mud?
- Does the trace contain internal structures, e.g. spreiten (backfill laminae) (Fig. 9.58)?
- Is it possible to distinguish cross-cutting relationships and work out whether the fabric is best interpreted in terms of the evolution of the burrows with time or different communities living contemporaneously at different depths (tiers) (Fig. 9.73)?
- Is it possible to quantify the intensity of bioturbation through use of a bioturbation index (below) or the

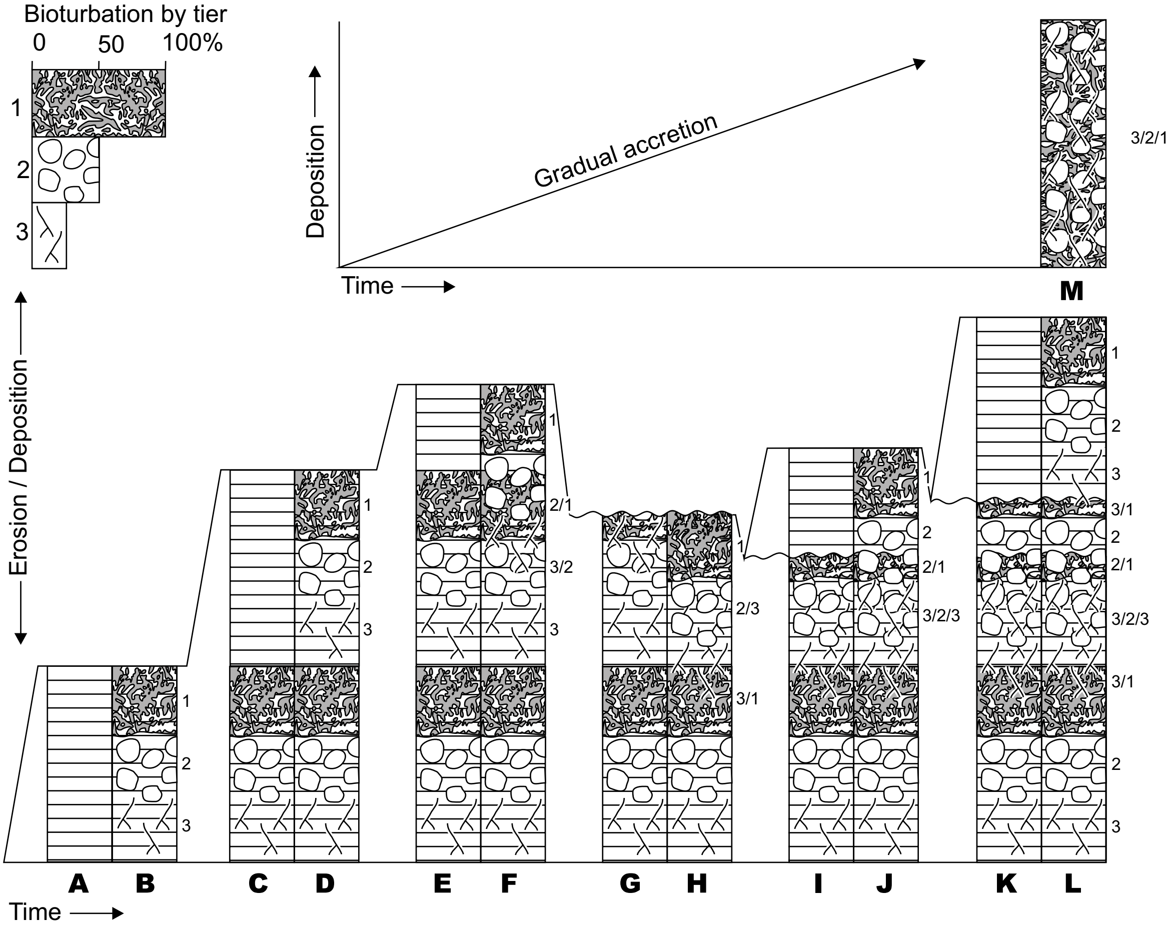


Figure 9.73 Effect of tiered bioturbation in different depositional regimes, based on the system shown top left. Deposition as a consequence of a series of discrete events produces many different effects. A) and B) Rapid deposition followed by non-deposition (omission) allows colonization. C) Rapid burial with a package thicker than the bioturbation zone enables preservation of the first bioturbated bed. D) New colonization. E) and F) Deposition of a thinner package leads to overprinting and produces a palimpsest fabric. G) and H) Erosion followed by non-deposition succeeded by colonization of the erosion surface. I) Slight erosion followed by deposition. J) The erosion surface is not a colonization surface and is largely obliterated by later bioturbation. K) and L) Tier overprinting indicated by numbers. M) The result of gradual accretion is the development of an ichnofabric having characteristic cross-cutting relationships. Modified after Bromley (1996).

numbers of individual trace fossil "species" (i.e. by using a measure of ichnodiversity – see Figure 9.73 and §9.4.5)?

In some sedimentary successions, it is not uncommon to encounter sediment showing few surviving primary sedimentary structures. Instead, a seemingly chaotic bioturbated ichnofabric may occur as a result of intense or protracted animal and plant activity (Fig. 9.59). To observe, record and interpret such apparent confusion it is worthwhile to first identify any remaining primary structures and then estimate the intensity of bioturbation at several key levels using a **bioturbation index** (Fig. 9.74) or ichnofabric index (Fig. 9.75), before describing and categorising the whole succession.

Interpretation of the bioturbation index is not easy. Total bioturbation suggests that the substrate provided a suitable environment for colonization and that the rate of biogenic reworking exceeded the rate of sedimentation, such that there was time to mix (churn) the sediments fully. Beyond this, many interrelated factors may have applied: the density of the population, water temperature, presence or absence of key bioturbator species, the type and rate of their activity, the tier in which the organisms were operating. Different types of trace belonging to different tiers and/or lifestyles may account for total mixing.

Bioturbation index (BI)[a]	Fraction bioturbated (%)[b]	Classification
0	0	No bioturbation
1	1-5	Sparse bioturbation: few discrete traces and/or escape structures
2	6-30	Low bioturbation: bedding distinct, low trace density, escape structures often common
3	31-60	Moderate bioturbation: bedding boundaries sharp, traces discrete
4	61-90	High bioturbation: bedding boundaries indistinct, high trace density with overlap common
5	91-99	Complete bioturbation: sediment reworking due to repeated overprinting
6	100	Intense bioturbation: bedding completely disturbed (just visible), limited reworking, later burrows discrete

[a] Each grade is described in terms of the sharpness of the primary sedimentary fabric, burrow abundance and amount of burrow overlap.
[b] Use these percentages as a guide, not as an absolute class division.

Figure 9.74 Bioturbation index for use in determining the extent to which biogenic activity has disturbed the primary (original) sedimentary texture/fabric. Modified after Taylor and Goldring (1993) and Goldring (1999).

Incomplete bioturbation suggests that stress factors prevented more complete reworking. These may be higher rates of sedimentation and/or a more energetic depositional regime (Fig. 9.76), although other factors may have been at work such as salinity variations affecting biodiversity, or the restricted availability of oxygen and/or nutrients. In order to identify the possible negative factors, it is necessary to identify and quantify the individual ichnogenera and evaluate the quality of the traces, as well as interpret the nature and origin of the sediment.

Total absence of bioturbation may result from an original lack of burrowing activity or a failure to preserve former biogenic structures. Undisturbed primary lamination may mean that no animals and plants were present. If intermittent erosion cuts deeper than any reworking, then no bioturbation will be preserved. In shallow seas, the substrate is commonly rapidly bioturbated during the summer but is completely reworked into primary physical structures by winter storms. Where storm layers are present, the most intense burrowing is commonly confined to the upper part of the layer, reflecting the period of non-deposition following the emplacement of the storm bed.

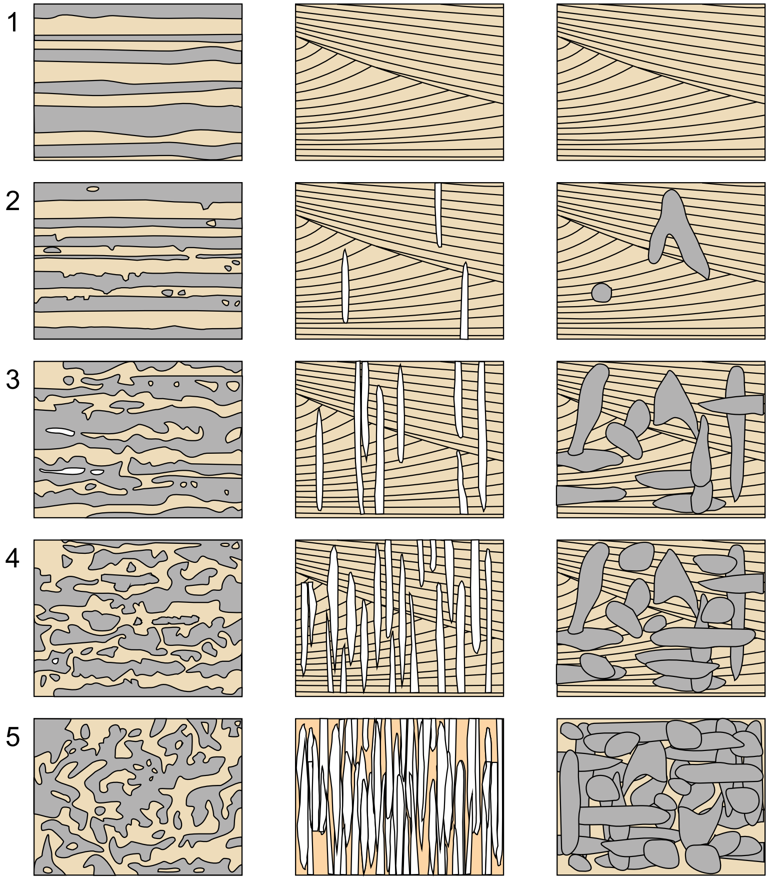


Figure 9.75 Bioturbation intensity and the graphic representation of ichnofabric indices (1–5) for various substrate types. Left: interlaminated sands and muds. Centre: trough cross-bedded sandstones. Right: trough cross-bedded sandstones dominated by *Ophiomorpha*. Modified after Droser and Bottjer (1986).

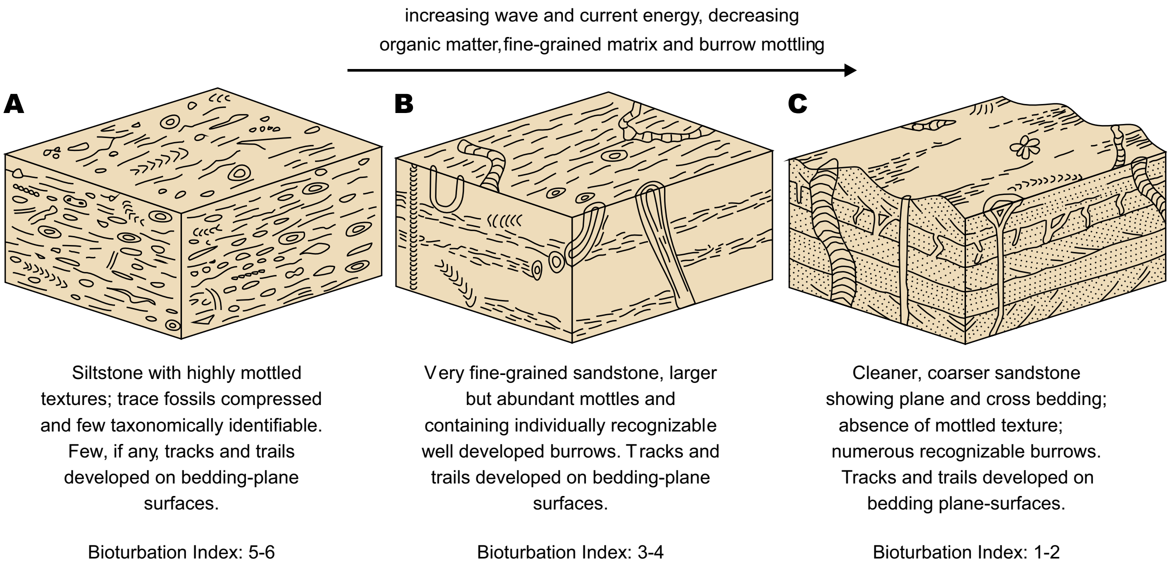


Figure 9.76 Diagrams to show the changing pattern of bioturbation in response to increasing energy in the depositional environment from A) to C). Modified from Howard in Frey (1975) and Howard in Basan (1978).

Some general environmental interpretations may be usefully drawn from these observations. Bioturbation tends to homogenize the grain-size segregation that helps define laminae (Fig. 9.76A). Destruction of preferred grain orientation by bioturbation can change the porosity and permeability structure of the sediment. If coarse and fine laminae become mixed, the sediment may retain a high water content and the onset of lithification may be delayed. If fines are moulded into faecal pellets, the mean grain size of the remaining sediment is increased and the sorting enhanced. Where sediment is bound by roots and certain burrow structures, it becomes stabilized. In cases where burrows are vertical they act as drainage channels, permeability normal to bedding increases and sediment tends to become firmer. Where burrows are parallel to bedding, horizontal permeability may be enhanced. If a burrow margin is lined, it influences preferred permeability pathways. Open burrows greatly extend the sediment-water interface and allow geochemical reactions and fluxes to increase within the substrate.

9.4.5 Ichnodiversity

In addition to considering the ichnofabric and the degree of bioturbation within a succession, it is also desirable to determine the **ichnodiversity** by counting the number of ichnogenera or ichnospecies within a given interval. Using this, an estimate may be made of the biodiversity of the original environment at and shortly after the time of deposition. However, it is important to remember that traces generated by different types of organisms and behaviour will have variable preservation potential so that the preserved ichnodiversity may not fully reflect the original biodiversity of the environment. Furthermore, bear in mind the fact that different behaviours by the same organism can generate different traces and, conversely, similar behaviour patterns of different organisms can result in the generation of only one trace.

9.4.6 Trace fossil ichnofacies and their environmental implications

The term **ichnofacies** characterizes associations of trace fossils that are repeated in time and space. These are thought to directly reflect environmental conditions such as water depth, salinity and substrate character (e.g. softgrounds, firmgrounds and hardgrounds). Several distinct ichnofacies are recognized (Fig. 9.77). Although each ichnofacies has been named after a representative ichnogenus, that particular ichnogenus does not necessarily have to be present in every example of that facies. A fundamental prediction of an actualistic approach is that these or similar traces would have formed wherever and whenever the particular set of conditions recurred, and the occurrence of trace fossils throughout the Phanerozoic rock record appears to support this.

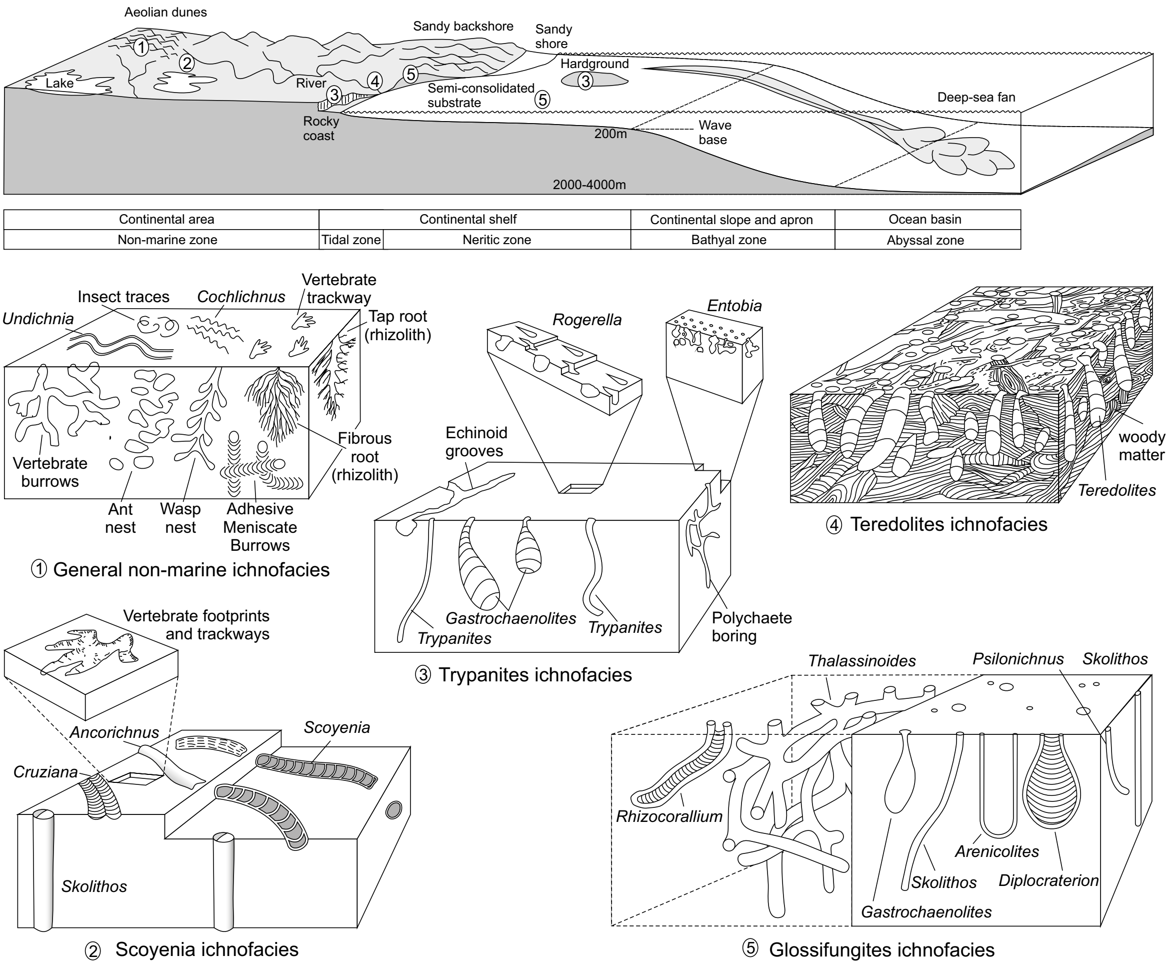


Figure 9.77 The occurrence of ichnofacies in relation to environmental setting. Modified after Seilacher (1967), Crimes (1975), Rhoads in Frey (1975), Chamberlain in Basan (1978), Frey and Pemberton (1975), Pollard (1982, pers. com.), Pemberton et al. (1992) and Hasitosis (2002).

Once processes relating to the formation of traces have been interpreted, it is possible to combine sedimentological and ichnological ideas to better determine the environment in which the traces were formed. Substrate consistency and energy levels, for instance, might be deduced quite confidently from traces of behaviour in the sediments, although biological and other environmental factors are less readily inferred.

Four commonly recognized ichnofacies are based on energy levels and loosely on water depth in the marine realm (Fig. 9.77, 7–10). The **Skolithos ichnofacies** reflects high-energy shoreface conditions, the **Cruziana ichnofacies** reflects medium-energy, sandy-silty lagoon/shelf-sea conditions, the **Zoophycos ichnofacies** reflects low-energy, muddy continental slope to abyssal plain conditions, and the **Nereites ichnofacies** reflects realms subject to the rapid deposition of sand in otherwise mud-dominated domains, for example as a result of turbidite deposition in base-of-slope and abyssal-plain settings. The **Psilonichnus ichnofacies** is primarily associated with sandy backshore and marginal marine environments whereas the **Glossifungites ichnofacies**, which is characteristic of firmground to hardground surfaces due to periods of non-deposition, occurs mostly in the marine realm. The **Trypanites ichnofacies** characterizes borings in rockground (i.e. lithified) substrates. The **Teredolites ichnofacies** is characteristic of woodgrounds; for example, log jams in fluvial and delta-plain settings, and in fluvial-marine transition zones.

Ichnofacies	General non-marine	Scoyenia	Trypanites	Teredolites	Glossifungites
Typical environments	alluvial fans, rivers and their floodplains, lakes, aeolian dunes and palaeosols	shoreline of ephemeral lakes, overbank areas of sluggish-flow rivers	hardground, reefs, rocky coasts, beach rock and other omission surfaces	Sites of accumulation of wood debris: rivers, deltas, lakes, coasts	environmentally wide-ranging, developed in firm, unlithified substrates
Energy (waves/wind)	low to high	low to moderate	high	moderate	moderate-high energy, frequent wave mixing
Eh	oxidising	oxidising	oxidising	oxidising	oxidising
Salinity	fresh water, though may rarely also be saline	fresh water, though may rarely also be slightly saline	normal	normal	normal
Temperature	variable, daily changes	variable, daily changes	variable, daily changes	variable, daily changes	daily changes
Light	daily changes	daily changes	daily changes	daily changes	daily changes
Substrate (sediment type and firmness)	sand and/or mud, softground, largely stable	sand and/or mud, softground to firmground, largely stable	lithified rock hardgrounds, stable. Walls of borings cut through hard substrate	woody or highly carbonaceous substrate (woodground)	sand = mud, firmground, reworked, eroded
Diversity	low to moderate	low	moderate	low	low
Abundance	low to moderate	low	moderate to high	low to high	high
Dominant organisms and traces	horizontal surface burrows, tracks and trails horizontal to near-vertical near-surface burrows multi-oriented to multi-component (chambered) sub-surface burrows root traces (rhizoliths)	small, horizontal-lined, back-filled burrows curved to tortuous feeding burrows vertical unlined cylindrical to irregular shafts sinuous crawling traces, tracks and trails	cylindrical to vase-, tear- or 'U'-shaped to irregular dwelling borings of suspension feeders or passive carnivores raspings and gnawings of algal grazers borings oriented perpendicular to the substrate	club-shaped borings with walls ornamented with the texture of the host substrate (e.g. tree-ring impressions) stumpy to elongate sub-cylindrical excavations in marine settings shallower, etchings in non-marine settings	arthropods, molluscs, echinoderms, corals, 'worms' vertical and inclined 'ear-shaped' burrows mostly suspension feeders protrusive spreiten due to animal growth

Figure 9.77 (Continued)

These ichnofacies, together with representative assemblages of ichnogenera by which they are identified, and the main environmental conditions that they record, are depicted in Figure 9.77.

The **Scoyenia ichnofacies** is now associated mainly with continental, freshwater lacustrine and fluviatile settings where sandy-silt and muddy firmgrounds are developed. Further non-marine, and much less common ichnofacies have been proposed; the **Mermia ichnofacies** for low-energy, loose- and softground freshwater turbidite environments, and the **Coprinisphaera ichnofacies** for insect burrows in palaeosols (Fig. 9.77, 1). Additionally, an **Arenicolites ichnofacies** is recognized for opportunistically colonized beds, which were the result of rare sand-silt-depositing events in either lacustrine or marine settings.

9.4.7 The uses of trace fossils

This section summarizes the main uses of trace fossils in understanding sediments and discusses their uses in palaeontology and palaeobiology, structural geology, geotectonics, stratigraphy and applied geology. Trace fossils contribute increasingly to the delineation of environmental conditions in many present-day settings because they reflect interactions between organisms and particular physical and chemical conditions. Traces reflect many of the processes that are the basis of the uniformitarian and actualistic models of distinct ecological and sedimentological settings.

Trace fossils can show whether sedimentation was continuous (at relatively slow or high rates) or discontinuous (at variable rates with or without erosion) (Fig. 9.64). They may give an indication of substrate consistency (Figs. 9.63, 9.77) and degree of aeration at the time of activity, as well as the degree to which the sediment has been reworked. Some trace fossils are valuable indicators of palaeocurrent where burrows and resting traces have a preferred orientation, whereby the organisms that generated them were aligned with the current, either for feeding purposes or as an aid to stability (e.g. oriented forms of *Lockeia*, Figs. 9.59H, 9.70K). Borings indicate a **hardground** into which organisms bored (Fig. 9.78). *Teredolites* records the boring of organisms (certain species of bivales) into a **woodground** (Fig. 9.79).

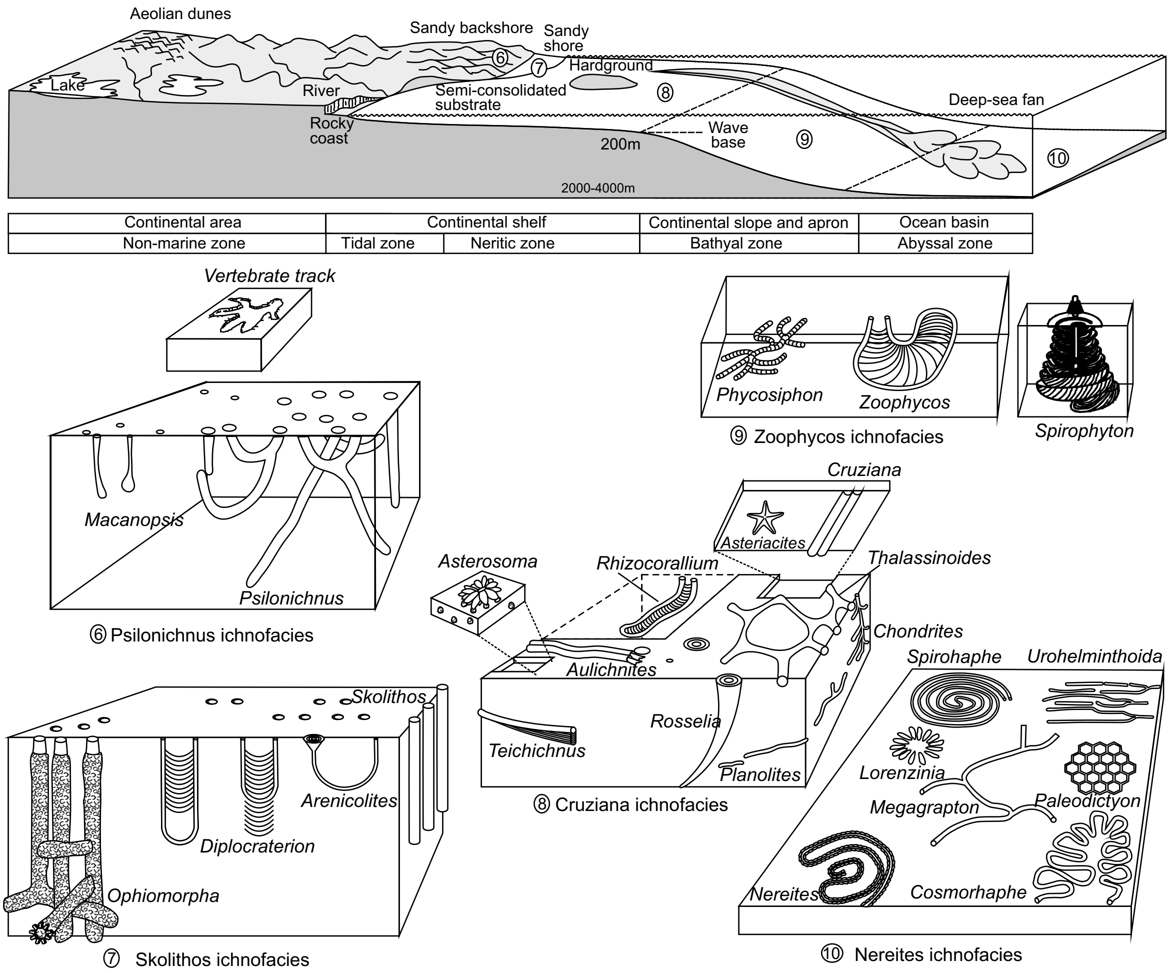


Figure 9.77 (Continued)

In characterizing **sedimentary environments**, trace fossils provide records of life *in situ* and, in many cases, the preserved ichnofacies is a valuable indicator of the depositional environment of the host sediments. Where organic activity has destroyed or masked primary depositional structures, trace fossils may provide the only clues to the nature of an environment. Traces are formed across almost the entire spectrum of sedimentary environments from continent to abyss, but the narrow environmental range of many forms reflects the preference of trace makers for particular ecological conditions and substrates (Fig. 9.77). Their abundance in clastic rocks which commonly lack body fossils – for example due to the dissolution of shells – means that the palaeo-environments can still be interpreted in the light of organic activity. The extensive time range of some traces throughout the Phanerozoic permits palaeoecological comparisons of rocks of different ages.

In **palaeontology** and **palaeobiology**, trace fossils record the behaviour patterns of extinct organisms, (e.g. the feeding, locomotive and protective activities of trilobites), information that cannot typically be gleaned from the body fossils themselves. Furthermore, trace fossils record the activities of organisms that had only soft parts and hence leave no other record. Thus, they increase the known diversity of fossil assemblages in the geological record. This provides a wider sample of former life forms and encourages the understanding of the evolution of fossil behaviour. Trace fossils also help to elucidate problems in late Precambrian rocks where body fossils are generally

Ichnofacies	Psilonichnus	Skolithos	Cruziana	Zoophycos	Nereites
Typical environments	backshore, coastal dunes washover fans, supratidal flats	foreshore & shoreface, bars and spits, some tidal deltas and submarine fans	shallow water below fair-weather wave base and offshore transition zone	variable, though often below storm wave base in areas largely free of turbidity flows	bathyal to abyssal, quiet but oxygenated environments
Energy (waves/wind)	extreme variations in energy levels	high energy, near constant wave mixing	lower energy, frequent wave mixing (storm action may introduce *Skolithos)*	low energy, very infrequent wave mixing, some rare density flows	no wave mixing, some density flows, some ocean-current flows
Eh	oxidising	oxidising	oxidising	Oxygen reduced, high organic content	limited oxygen (influx from density flows), not usually anoxic
Salinity	variable: fresh, brackish, normal marine, rarely hypersaline	mostly normal	normal	normal	normal
Temperature	daily changes	daily changes	seasonal changes	<10°C, no changes	2-10°C, no changes
Light	daily changes	daily changes	daily changes in upper part	none	none
Substrate (sediment type and firmness)	variable amounts of sand & mud, softground, reworked, highly mobile	sand > mud, softground, reworked, highly mobile loose particulate substrate	mud = sand, stable softground, rare reworking and ripples	mud, stable except in failure and density flow	pelagic mud dominant, mostly stable except for density & ocean currents
Diversity	low	low	high	low	moderate (higher than for *Zoophycos)*
Abundance	low	high	high	low to high	low but seems higher due to slow accumulation rates
Dominant organisms and traces	invertebrates (predators or scavengers), vertebrates (predators or herbivores) vertical shafts, bulbous-based, irregular, 'J' or 'U'-shaped dwelling structures invertebrate and vertebrate crawling & foraging traces, vertebrate tracks, coprolites	arthropods, molluscs, echinoderms, corals, 'worms' unbranched vertical burrows mostly suspension feeders	arthropods, molluscs, echinoderms, corals horizontal crawling and grazing traces more abundant than inclined burrows tiering common mostly sediment feeders	arthropods, polychaete worms, hemichordates (e.g. acorn worms), echinoderms complex horizontal grazing and shallow feeding traces, spreiten inclined in sheets, ribbons, spirals. Bioturbation sediment churners, feeders	arthropods, polychaete worms, hemichordates (e.g. acorn worms), echinoderms casts, crawling, grazing traces, spreiten planar and on surface sediment grazers, farmers

Figure 9.77 (Continued)

absent and where trace fossils record important events such as the appearance of the Metazoa. Recognition of distinctive trackways of terrestrial arthropods in continental sediments and burrows in palaeosols have contributed to recognition of such important events in the history of life as the invasion of land by animals in the early Palaeozoic. The presence of traces in coarse-grained sediments, laid down where abrasion and weathering (especially oxidization) were active in destroying both hard and soft parts of organisms, helps to bridge several palaeontological gaps. Vertebrates such as the producer of *Chirotherium* (the "hand beast" – probably an early thecodont archosaur of the Lower Triassic) are known only from their footprints and trackways, which offer important evidence of the evolutionary radiation and habits of the early reptiles. Indeed, the evidence of many reptiles is known only from their traces in aeolian sediments.

In **structural geology** the fact that the organisms react to gravity and light, grow, or move downwards and upwards, and are asymmetric about a horizontal plane, make their traces excellent small-scale, "way-up" indicators (Fig. 9.70). Furthermore, quantitative estimates of compaction and deformation depend upon the recognition of objects of known shape prior to deformation (Fig. 9.80). Burrows can be such objects, though great care is needed in their use. Some sand-filled burrows become flattened films on bedding planes when surrounding mud has been strongly compacted. U-shaped burrows are particularly useful because they allow the reconstruction of strain ellipses and aid quantitative estimates of strain due to pre-cleavage compaction, compression, rotation and cleavage distortion.

In **geotectonics**, associations of traces may help to define faunal provinces. Certain distinctive traces in successions of late Cambrian–early Ordovician age around the North Atlantic, for example, are attributable to trilobites of an Avalonian Province, others to a Laurentian Province. This evidence has contributed to current views concerning the separation of these provinces at that time by an Iapetus Ocean.

In **stratigraphy**, the extended time range of many trace fossils restricts their use for biostratigraphic or

Figure 9.78 Dolomite with intense borings by *Lithophaga*, a genus of marine bivalve molluscs that bore into carbonate rocks. The resultant trace fossil is *Gastrochaenolites*. Modern example of borings into dolomite of the Roker Formation, Permian, County Durham, England.

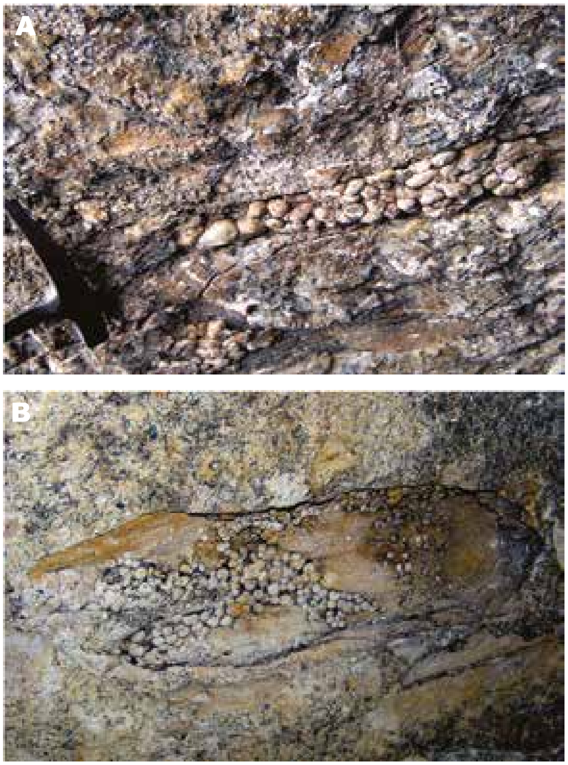


Figure 9.79 Fossilized wood with casts of animal borings, *Teredolites*, generated by Cretaceous to Recent marine bivalves that attach themselves to in-situ or rafted wood (e.g. drift wood). These trace fossils indicate infaunal boring into a woody substrate (woodgrounds) and are common in successions of shallow-marine, tidal and deltaic origin. A) Intensely bored fossilized wood. B) Larger piece of wood that has been partly bored. Wood is 0.6m long. Both examples from the Neslen Formation, Cretaceous, Utah, USA.

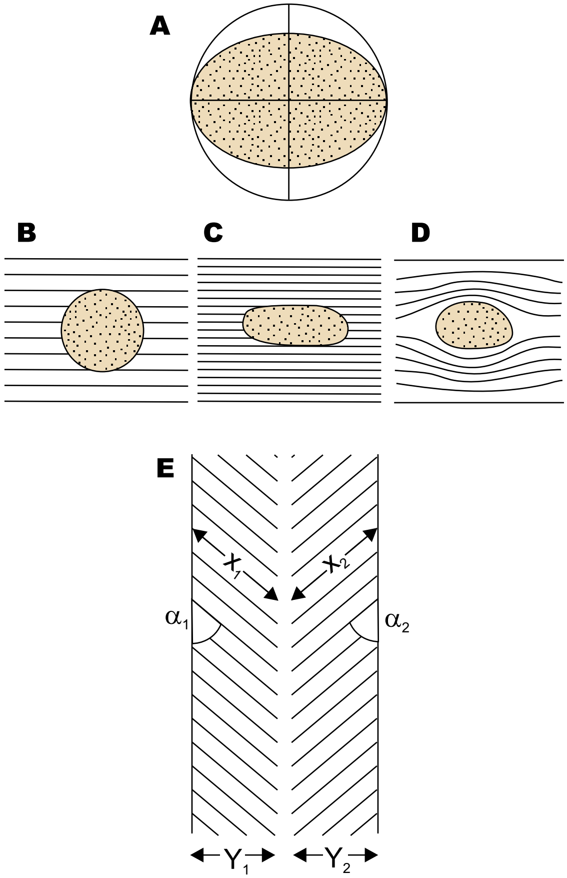


Figure 9.80 The use of trace fossils in deformational studies (after Crimes in Frey, 1975). A) The elliptical outline of a burrow (stippled) may be used to determine the percentage compaction by the construction of a circle to represent the outline of the undeformed burrow. A comparison of the axial lengths of the undeformed circle and the ellipse may then be made. This approach assumes that the shape of the original undeformed burrow was circular. B) No compaction after burrowing. C) Measurement of burrow compaction can be used to determine amount of compaction of the host sediment. D) Host sediment compacts more, the laminae being deformed around the burrow; compaction values for the burrow, therefore, are less than for the host sediment—as in the commonest cases where burrow infills are coarse-grained. E) Effects of tectonic deformation on *Cruziana*. In the undeformed state, normally x1 = x2; a1 = a2; Y1 = Y2. Any increment of compression or shear disturbs the equality and can be used to quantify deformation two-dimensionally in the plane of the trace. *Planolites*, *Chondrites* and *Beaconites* are the most frequently used traces for these studies.

chronostratigraphic purposes, but short-ranging forms such as certain types of *Cruziana* can be used to date poorly fossiliferous successions in the Lower Palaeozoic. Vertebrate footprints provide a worldwide basis for a stratigraphy of the Triassic. These two examples record rapid evolution during these times and hence the resultant correlations may have considerable reliability. Trace fossils also help to generate and control palaeogeographic reconstructions for individual stratigraphic time periods. Since the late 1970s the concepts and application of sequence stratigraphy have been greatly enhanced by trace-fossil analysis. Changes in ichnofabrics and ichnofacies permit the recognition of key stratal surfaces reflecting, say, a regionally-extensive depositional hiatus or a marine flooding event.

Lastly trace fossils are of considerable use in **applied geology**, particularly in the hydrocarbon industry. The understanding of the concepts and techniques outlined in §9.4 may be useful, and indeed some vital, if worthwhile predictions are to be made about depositional environments, and the extents of source, reservoir and cap rocks. Fortunately trace fossils show up well in cores where an appreciation of their three-dimensional nature is often easier than at outcrop. Ability to recognize bioturbation as the cause of massive sandstones with low porosity and permeability may be important in the prediction of reservoir quality and burrows may modify the directional permeability distribution of some reservoirs.

9.4.8 Confusion of traces with inorganic sedimentary structures

Traces may, in some cases, be confused with primary and secondary structures of inorganic origin. A mark made by an active trilobite is a trace fossil; a mark made by impact of a moulted or deceased adult carapace of a trilobite is an inorganic structure, i.e. a tool mark (skip cast). Confusion is particularly common where burrows are concerned, for they have been related to gas pits (feeding traces of polychaete worms), air-escape holes (burrows of amphipod crustacean sand-hoppers), conical fracture patterns (feeding burrows such as *Phycodes*), sand volcanoes and fulgurites – sediment structures generated by lightning strikes. In other cases, obstacle scours have been described as the trace fossil *Blastophycus*, rill marks as *Dendrophycus*, sinuous hierarchical mudcracks as *Manchuriophycus*, interference ripples as tadpole nests, and convolute or contorted lamination as bioturbation. The possibilities are many and varied; it is important to generate and test a wide range of working hypotheses when an apparently novel trace fossil is found.

Study techniques

Field experience

Structures produced by inorganic or organic disturbance may be found in almost any environment, but, when they are found, the processes that gave rise to them have often ceased to operate. Sandy beaches and fluvial and estuarine sand bars are good places for geologists to fluidize water-laden sand by stamping, thus changing the grain packing and the pore-water pressure so that the sediment becomes quick. With practice, sand volcanoes may be produced. Desiccation and synaeresis cracks may be observed in dried-out and still-water ponds, respectively (e.g. in supratidal or fluvial areas). Large-scale sub-aerial landslides are easy to study and may be usefully compared with slumps. Soil profiles in which concretions are forming may be studied in many temperate and tropical areas. Biogenic activity resulting in bioturbation is widespread on many tidal flats.

Laboratory experience

Rapid dewatering Physical deformation structures associated with dewatering can be effectively generated in the laboratory by part-filling a narrow glass tank with water-saturated fine-grain (ideally mud-dominated) sediment to a depth of 0.15–0.2m. The rapid addition of a 0.15–0.2m thick layer of dry or damp sand will increase the pressure on the already water-saturated fine-grain sediment and will induce de-watering through the generation of load and flame structures, mud volcanoes and, in some cases, sheet de-watering structures.

The development of desiccation cracks Fill a flat-bottom plastic tray with water-saturated silty-mud to a depth of 0.05–0.1m. Place the tray in a warm position (e.g. on a window sill) and allow the sediment to dry out over a period of several days. As the water evaporates, tension at the sediment surface will increase until desiccation cracks begin to develop. Photograph the sediment surface every 12 hours to build up a time-lapse image sequence as a record of crack growth. Repeat the experiment, but cut grooves into the sediment surface in order to ‘seed’ crack development. Repeat the experiment with silty mud of a different composition and note whether this affects the shape of the crack pattern network that develops.

Bioturbation Biogenic deformation structures (bioturbation) can be observed in the laboratory by filling a narrow (0.05m wide) glass tank with alternate layers of damp mud, silt and sand (each 0.01–0.03m thick and ideally of varying colour) and introducing live invertebrate organisms (earthworms are ideal). Over a period of a few days, observe how the burrowing and sediment-churning activity of the organisms deforms the originally horizontally bedded sediment layers. Measure the intensity of deformation using bioturbation and ichnofabric indices (Figs 9.74, 9.75). After about a week, the sediment will become completely homogenized. Photograph the side of the tank every 12 hours to build up a time-lapse image sequence as a record of the progression of bioturbation.

Recommended references

Bromley, R. G. 1996. *Trace fossils: biology and taphonomy*. A very useful place to start understanding the assemblages and relationships of trace fossils.

Frey, R. W. (ed.) 1975. *The study of trace fossils*. A classic of its time, with a series of well illustrated wide-ranging papers on all types of trace fossils.

Goldring, R. 1999. *Field palaeontology*. Useful advice on how to observe and record trace fossils and bioturbation in the field.

Hasiotis, S. T. 2002. *Continental trace fossils*. A simple and well illustrated guide to the study of non-marine trace fossils.

Jones M. E. & R. M. F. Preston (eds) 1987. *Deformation of sediments and sedimentary rocks*. Papers dealing with deformation at a wide range of scales, from microscopic fabrics to large-scale tectonics in particular settings.

Knaust, D. & R. G. Bromley 2012. *Trace fossils as indicators of sedimentary environments*. Detailed yet accessible consideration of trace fossils and their palaeoenvironmental significance.

Lowe, D. R. 1975. *Water escape structures in coarse grained sediments*. A good account of the mechanisms of formation, as understood at the time.

Maltman, A. (ed.) 1994. *The geological deformation of sediments*. A series of review chapters on major aspects of soft-sediment and later deformation.

Pemberton, S. G. 1992. *Applications of ichnology to petroleum exploration*: *A core workshop*. Possibly quite difficult to find, but a very good, practical guide to identification of trace fossils at a basic level.

Retallack, G. J. 1997. *A colour guide to palaeosols*. A beautifully illustrated atlas of palaeosols showing both outcrop profiles and microscopic textures.

Selles-Martinez, J. 1996. *Concretion morphology, classification, and genesis*. A comprehensive review.

Wright, V. P. (ed.) 1986. *Palaeosols: their recognition and interpretation*. A thematic set of papers.

Wright, V. P. & M. E. Tucker 1991. *Calcretes*. A detailed consideration of a type of carbonate soil that is common in sedimentary successions

CHAPTER 10

Assemblages of structures and environmental interpretation

10.1 Introduction

In earlier chapters we have shown how sedimentary structures relate to erosional, depositional and post-depositional processes. The ability to interpret sediments in these terms is clearly useful in its own right, but it is often more important to use that information as a step towards interpreting the depositional environment of sediments in the rock record. Earlier chapters made little mention of depositional environments. This omission was deliberate in order to highlight the fact that many structures and processes are common to a range of environmental settings. However, determining the processes responsible for generating a particular set of structures is a necessary first step in making an environmental interpretation. To move from an interpretation of processes to one of environment, further analysis is required. This usually seeks to establish the spatial and temporal relationships of the processes deduced from the sedimentary structures. These relationships can usually help to narrow the range of environmental possibilities. It is also useful to know something of the directional properties of sedimentary structures. With directional information we can test and refine our ideas because the relative directions of flows and wave movements help to characterize certain environments. Directional information also helps to orientate an inferred palaeoenvironment in space and thereby give it palaeogeographical significance.

In characterizing a modern environment or establishing an environmental interpretation for sedimentary rocks, therefore, it is important to record and present observations of sediments, their physical and chemical sedimentary structures, body and trace fossils, in addition to their directional properties and their positions in space or in measured sections in clear and well-structured ways.

10.2 Mapping of modern environments

The main aims of mapping sedimentary structures in modern environments are to learn something of the distribution of hydrodynamic or wind energy within the environment, and to predict the likely patterns of lithology and sedimentary structures, should deposits of the environment be preserved. The second aim has particular relevance to the application of uniformitarian principles to the interpretation of sedimentary rocks.

The most common method of investigating the distribution of water-generated bedforms, for example those encountered on intertidal areas or on river beds, is on foot and at low water. Notes on such methodology are presented in Appendix 3. Although the mapping is typically quite straightforward, interpretation is more complex, as the patterns observed on emergent surfaces probably result from a succession of flow conditions. All bedforms need time to respond to changes in flow. Large bedforms, produced under conditions of strong flow, may be stranded if the water level and flow strength fall rapidly. Small bedforms, such as ripples, adjust more quickly and many continue to respond to the flow almost to the point of emergence. Therefore, it is important to try to interpret exposed surfaces in terms of an evolving flow history rather than one specific set of flow conditions.

Predicting the vertical succession of sediment that will be generated by the processes operating in a given environment requires answers to several questions. Which of the observed bedforms is most likely to generate preserved internal structures? What is the distribution of such bedforms across the broader topography of the environment? How is the environment, as a whole, changing through time? In particular, is a systematic migration of sub-environments taking place over time? If so, it can be predicted that structures developed in topographically low areas will

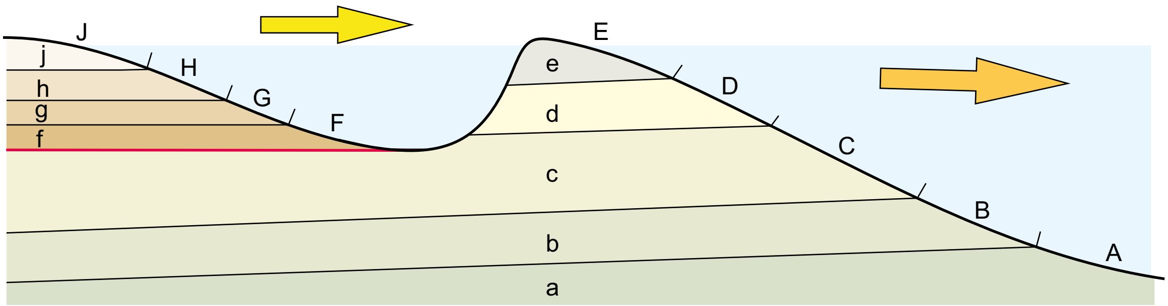


Figure 10.1 A schematic diagram to illustrate Walther's principle of succession of facies. Sub-environments A–E are on a sloping surface that is building out to the right, generating lithological units a–e. A channel which comprises sub-environments F–J is cut into the top of this topography and is migrating via lateral accretion in the same direction; this generates lithological units f–j. The boundary between lithological units c and f represents a break in deposition.

occur low in a vertical succession, with structures from successively higher topographic areas coming in above *in the same vertical order as their horizontal distribution* (Fig. 10.1). This method of relating the lateral distribution of surface features or sub-environments to a vertical succession of lithology and sedimentary structures is **Walther's principle of succession of facies** and is one of the fundamental starting points for any environmental interpretation of ancient sediments (see §1.3).

One complicating factor that is important in many environments, for example in intertidal settings, is the activity of burrowing animals. Animals that live below a depositional surface subjected to particular conditions of currents, waves or emergence may extend their burrows down into layers of sediment that were deposited under conditions quite different from those prevailing at the surface (see §9.4). By the time the burrowing takes place, these different conditions may have shifted some distance from the site of burrowing. In other words, burrows can cut across the vertical succession and the animals that produce burrows in a particular unit of sediment cannot be assumed to have lived under the conditions in which those sediments were laid down.

10.3 Measurement of sections in rock successions

Many environmental interpretations of sedimentary successions rely heavily on measured sections (Fig. 10.2). Such sections can record changing sedimentary processes through time, an important clue to the nature of the environment and to its evolution. In Chapter 2 we outlined the importance and some of the problems of section measurement and one or two points mentioned there warrant reflection and emphasis here. In logging a sedimentary, section it is important to decide upon its subdivision into significantly different units, where the differences may be based on grain size, on composition, and on combinations of sedimentary structures. The simplicity or complexity of the scheme chosen will depend upon the nature of the succession itself, the eventual aims of the investigation and the refinement or resolution of the interpretation being attempted.

Having established a basis for sedimentological sub-division, it is next necessary to describe and record the thickness and internal features of each unit and to determine the nature of its contact or boundary with units above and below. If beds conspicuously thicken and thin laterally within the extent of an exposure, record this either by noting it on the single measured section or by measuring and correlating more than one laterally equivalent section. When drawing up the section as a graphic log, remember to adjust the thicknesses of units so that the total thickness of the succession is accurately recorded. For example, where beds are conspicuously lenticular, it is important to record average thicknesses of units rather than maximum values as recording the latter would introduce a systematic error that would exaggerate the total thickness.

The features recorded will clearly vary with the nature of the succession and with the detail of interpretation required. It cannot be stressed too strongly that there is no absolute standard of description. Each investigation has its own aims and timetable, and these will determine the detail of the description and the criteria for subdivision.

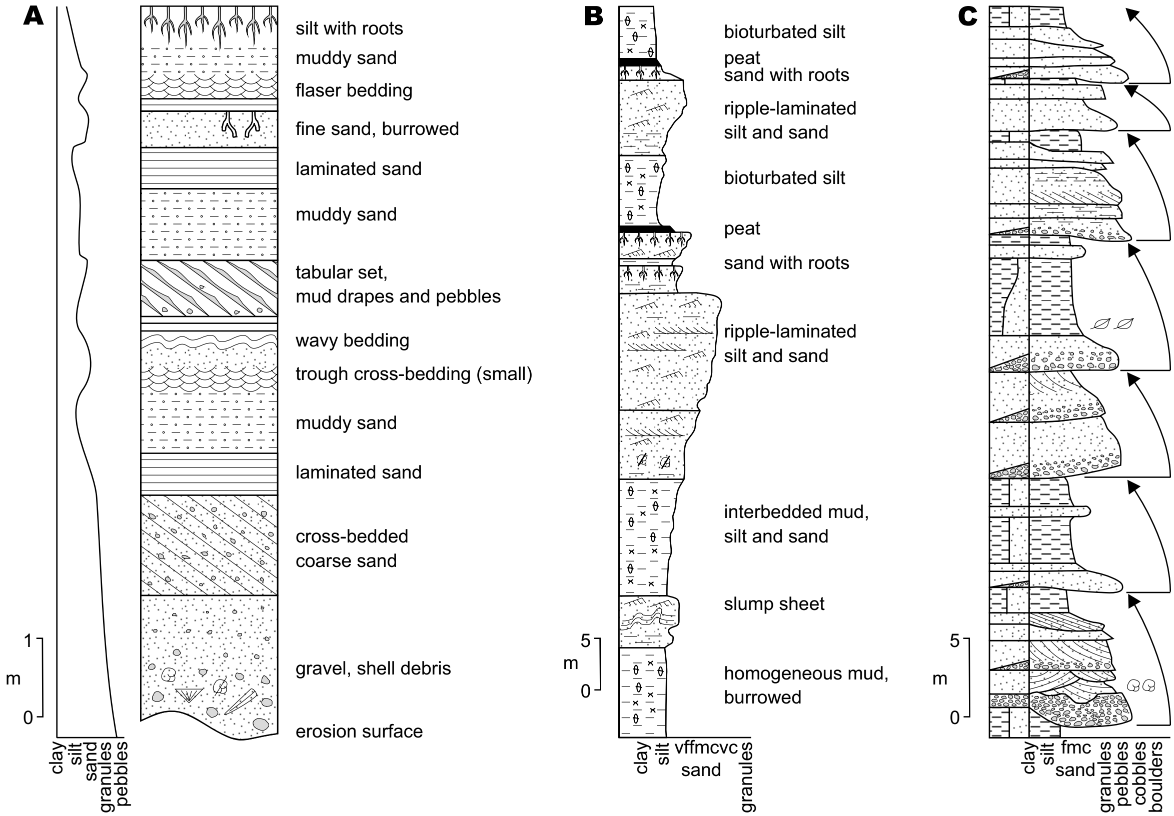


Figure 10.2 Examples of graphic logs of measured vertical successions through sediments. A) Structures depicted graphically, written description of lithologies and additional structures, separate column for grain size. Based in part on Greer (1975). B) Structures, lithologies and grain size all depicted in one column, with additional written comments where necessary. Based in part on Coleman and Wright (1975). C) Structures and grain size depicted graphically in the right-hand column, the proportion of sand and mud in the heterolithic (mixed) lithologies is indicated in the left-hand column, fining-upward units (cycles) within the overall succession are shown by arrows. Based in part on Surlyk (1978). See Figure A6.1 for a guide to the symbols used.

An important feature of measured sections that can sometimes be ignored is the nature of the contacts between units. Some contacts are gradational, sometimes to the extent that it is difficult to decide exactly where a boundary should be placed. Other contacts are sharp and some are clearly erosive, with conspicuous relief truncating underlying bedding or with erosional structures superimposed upon the surface. In §4.4.3 we suggest clues that may indicate an erosional contact even when such features are missing. Always consider the possibility of erosion wherever a sharp contact is seen, although, of course, not all sharp contacts are erosive.

Recording information as measured sections demands a disciplined method of working. Some geologists prefer to draw a graphic log directly in the field or core laboratory, either in notebooks or on specially prepared sheets (Appendix 5). Others use an essentially verbal description, supplemented with drawings and photographs where appropriate and leave the drawing of a graphic log for later. Drawing up an elaborate log whilst in the field can be time consuming, slowing down observation. However, it can be a useful aid to gaining understanding of relationships while in the field area. It is up to the individual to decide how to resolve this issue.

Where description in the field is mainly by a written log, it is important that it is organized so that information can be easily extracted later, allowing scaled, oriented diagrams, specimens and photographs to be easily related to it.

A system of columns, each devoted to separate features such as bed thickness, lithology and the nature of the contact with the overlying unit can work well. Other columns can be added for specimen and photograph numbers and for palaeocurrent measurements.

Presentation of measured sections as graphic logs is a matter of personal style. A glance through any sedimentological journal will show that there are almost as many styles of graphic log as there are authors. There is nothing wrong with this as it reflects the different aims and emphases of different pieces of work. Some styles of graphic logs are, however, more easily understood than others. Examples of commonly employed schemes are shown in Figure 10.2. In two of these, grain size is indicated schematically by column width and the symbols for lithology and sedimentary structures are, in many cases, self-evident (see Appendix 6 for a guide to the symbols used in these examples). The nature of the contacts between units is also well shown. Where thinly interbedded sandstone-mudstone lithologies occur, it is often useful to indicate their relative proportions in a separate "lithology" column (e.g. Fig. 10.2C).

Palaeocurrent measurements are best recorded alongside the units from which they were taken. A well-drawn graphic log incorporating all these features serves as a sound base for environmental interpretation and enables others to use the data to suggest alternative or more refined interpretations.

Where a succession shows conspicuous lateral variation, more than one vertical section may be needed. Location and spacing of the sections will depend on the complexity of the variation and on the aims and timetable of the study. With laterally continuous exposure, it may be appropriate to record the full two-dimensional form of the lithological units. Panoramic photographs of two-dimensional exposures often help in constructing suitable diagrams. Where erosion surfaces or bounding surfaces are apparent, it is important to try to establish if a hierarchy exists and to then assign each surface to an appropriate level. This is best achieved through the construction of scaled panels that depict the geometry and architecture of major units, and the form and interrelation of their bounding surfaces (Fig. 10.3).

With discontinuous exposure, as, for example, with a series of separated quarries, stream sections or boreholes, it is normally only possible to link the sections by correlating the most confidently identified bedding surfaces. It is important to take particular care when correlating sandstones or coarser units, particularly if there is any evidence that they may be lenticular (e.g. channelized). Correlation of similar sandstones at similar positions in a succession may give a misleading impression of lateral continuity. Beds could have died out laterally between observed sections and the true pattern may be one of shingled or offset lenses. As a general rule it is more reliable to correlate surfaces related to deepening or flooding, rather than erosion surfaces (See §10.5).

10.4 Interpretation of vertical sequences in rocks

The interpretation of sedimentary successions in terms of their environment of deposition is one of the main aims of sedimentology. We have seen how most sedimentary structures allow interpretation of processes of erosion, deposition or post-depositional alteration and how many of these structures and the processes responsible for their generation occur across a range of environments. One of the main starting points for moving the discussion towards an interpretation of depositional environment is the *succession of processes* deduced from the vertical sequence of lithology and sedimentary structures.

Before considering in detail the vertical sequence of lithology and sedimentary structures, several more general features of a sedimentary succession may suggest preliminary views about the environment of deposition. The most obvious of these is the presence or absence of body and trace fossils and, if they are present, their type. Their presence in grouped associations and their relative abundance, for instance, may tell us if a succession is of deep-marine, shallow-marine, marginal-marine or non-marine origin.

In some, cases, it may be possible to use body fossils to make preliminary inferences about water depth but before this is attempted it should be established if the fossils are *in situ* or have been transported after death. This may be judged by their state of articulation, their abrasion and the way in which they lie within the sediment (Fig. 10.4). If one is reasonably confident that the fauna, and occasionally the flora, is *in situ* or has not been transported far, then it may be appropriate to infer a shallower "shelf" setting for a certain shelly faunal assemblage and a deeper-water setting for a more pelagic

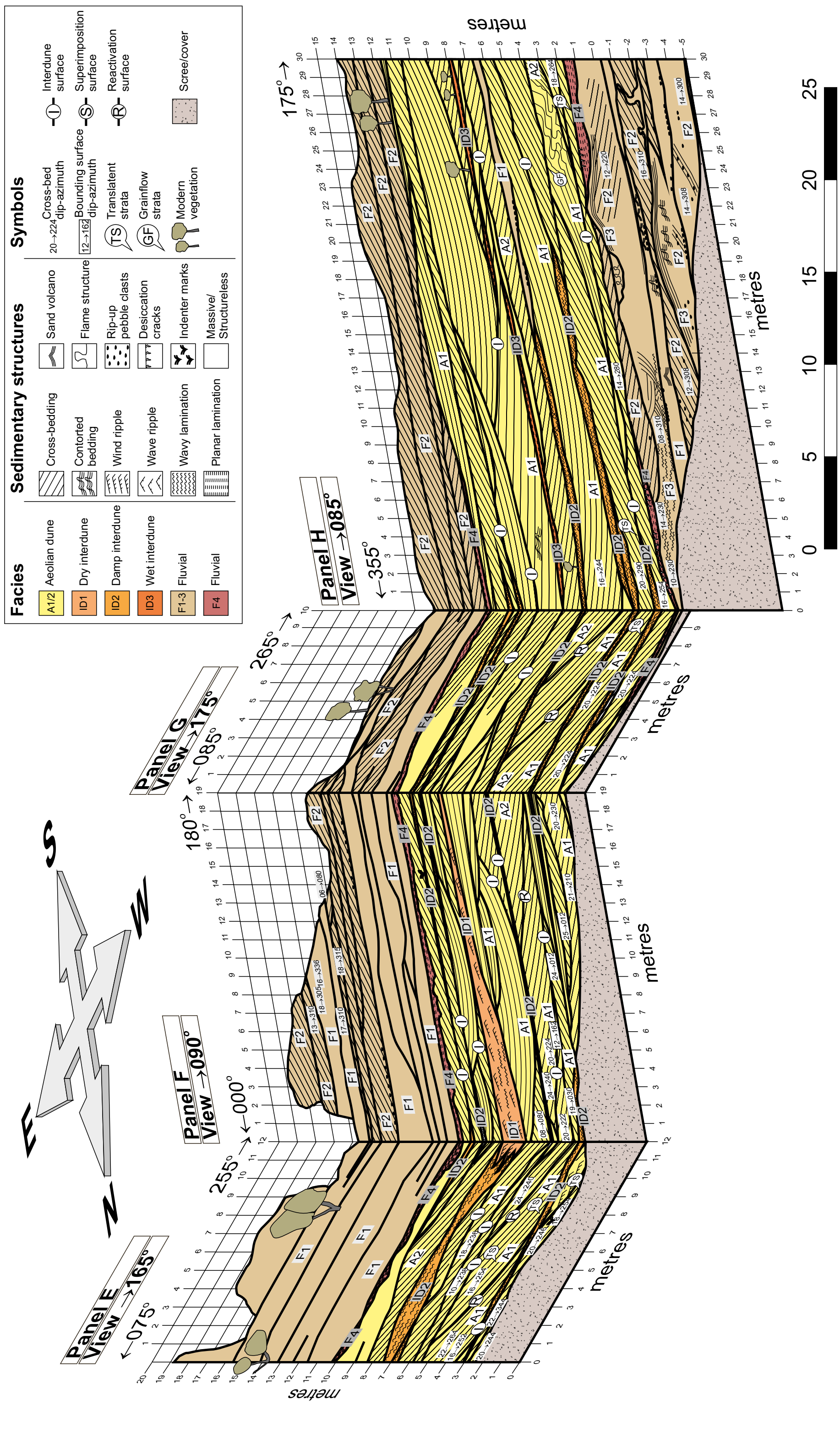


Figure 10.3 Example of a series of two-dimensional scaled drawings that depict the sedimentary architecture of a succession exposed in a cliff face. The four separate panels have been skewed with respect to one another in an attempt to depict their relative orientations in three-dimensional space. Although such panels are time consuming to construct, they are very useful for demonstrating lateral changes in strata geometry and the arrangement of bounding surfaces. Modified after Mountney and Thompson (2002).

Figure 10.4 Sandstone bed with a basal lag of shelly debris. Many of the shells are broken and randomly oriented (i.e. not in life position), indicating that they have been transported to the site of accumulation. Locality unknown.

faunal assemblage, for example. Palaeontological expertise is normally required to carry forward such arguments to a more sophisticated level. The environmental interpretation of Pre-Cambrian sediments is greatly restricted by the lack of this basic information and it is commonly difficult in some examples even to decide between a continental and a shallow-marine origin. Evidence of subaerial exposure of the sediment surface in the form of rain-pit casts, desiccation mudcracks and soil profiles (see Ch. 9) can help to narrow the range of possible environments.

When the broad environmental context has been narrowed down by such considerations, a more detailed analysis can proceed based on, amongst other things, the nature of the vertical sequence.

The law of superposition tells us that the vertical sequence of sediments records changes in depositional conditions *through time at that point*. Such changes occur for two fundamentally different reasons. In one case, the overall environment remains essentially unchanged, but within it changes in conditions take place *through time* to produce a succession of distinctive sediment units. For example, a deep basin normally receiving fine-grained sediment from suspension may have this background condition punctuated by the arrival of intermittent turbidity currents that deposit layers of coarser sediment (Fig. 10.5). The vertical change in lithology does not then record a change of environment but a temporary change in prevailing processes. The idea of temporary changes in conditions within a more or less stable environment means that the succession can be thought of as comprising the products of "normal" (background) and "catastrophic" (event) deposition (see §2.2.3).

In the second case, environmental conditions remain essentially constant through time, but there is a *spatial zonation* of processes and products within the overall environment. A gradual shifting of the environment through time thus leads to a vertical sequence of changing lithology and sedimentary structures (Fig. 10.6). This second case illustrates the application of Walther's principle of succession of facies introduced earlier in this chapter and illustrated here by a simple general model (Fig. 10.1). During migration of this system to the right, sub-environments A–E generate sediment units a–e in the same order and with non-erosive and probably gradational contacts. The channel system on top also generates its own gradational succession of sediment units f–j above an erosion surface, due to the migration of channel sub-environments F–J. If one interprets the succession on the left-hand side of the diagram without taking account of the nature of contacts between units, one would infer mistakenly that sub-environment F had been adjacent to sub-environment C. Recognition of erosion surfaces, therefore, is vital. When an erosion surface is identified, it is necessary to begin the application of Walther's principle afresh above that surface.

This idealized model can also be used to introduce another principle of environmental interpretation. As well as recognising the spatial relationships of the deposits and their depositional settings, it is also common practice to look for familiar (in some cases repeated) patterns of systematic vertical change in properties such as grain size, bed thickness and sedimentary structures. In the example shown in Figure 10.1, the earlier succession of units (A–E) could constitute an upwards-coarsening unit with the associated sedimentary structures showing evidence of progressively increasing energy. One possible interpretation could be that the succession was generated through the progradation of a shoreline, building out into a body of water, especially where other evidence, such as trace fossil assemblages (ichnofacies), supports a shallow-water setting. Characterizing the shoreline and the body of water more fully would depend on a consideration of, for example, the nature of the higher-energy processes (e.g. waves or currents) and the

Figure 10.5 Interbedded successions of sandstones and shales interpreted to be the result of episodic high-energy but short-lived turbidity flows carrying sand down a slope into normally quiet, deep-water environments. The successions are inferred to represent the products of alternating 'normal' and 'catastrophic' deposition. A) Aberystwyth Grits, Silurian, Wales. Cliff is approximately 30m high. Photo courtesy of Gilbert Kelling. B) Ross Formation, Carboniferous, western Ireland. Succession shown is approximately 15m thick.

assemblage of fossils and trace fossils (e.g. marine or freshwater).

Similarly, the upper succession in Figure 10.1 (F–J), above the erosion surface, may show a pattern of upwards-fining grain size and an associated diminution in the levels of energy as inferred from the sedimentary structures. The combination of the erosion surface and such a succession suggests the lateral migration of a channel system with stronger currents in the deeper part and weaker currents on the higher areas of the depositional surface. This could be a river channel, a delta distributary or a tidal inlet. However, such a refined interpretation would again depend on the specific details of the structures (e.g. the presence or absence of clay drapes), the directional properties of palaeocurrents (unidirectional or bi-polar) and the nature of fossils and trace fossils.

Several other predictable and ordered successions of beds or facies are widely recognized in the rock record and some common examples are illustrated in Figure 10.7. These ordered successions have been interpreted to be the products of sedimentary evolution in a range of depositional environments but, in each case, the techniques used in their recognition and interpretation remain the same.

10.5 Key stratigraphic surfaces

As well as local erosion surfaces, for example due to scour at the base of a channel, there are other surfaces that punctuate sedimentary successions and trigger the need to re-start an interpretation using Walther's principle. Some of these surfaces are extremely extensive and are often referred to as "**key surfaces**" that can

Figure 10.6 Examples of sedimentary successions that can be interpreted in terms of environment and underlying controls. A) Thick succession showing upwards-coarsening intervals in the lower part, each with a sharply defined surface that is a flooding (deepening) surface. The upper part is a major channel sandstone which has significant relief on its erosional base and internally contains multiple sandstone 'storeys' (i.e. is multi-storey). The succession overall records a major deltaic progradation, with minor progradations following each flooding event, and a major phase of erosional down-cutting, likely caused by a fall in relative sea level to create a palaeovalley. This was filled by coarse fluvial sandstone deposits as relative base level began to rise and create accommodation space. Central Clare Group, Upper Carboniferous, western Ireland. B) Succession of marine sediment showing changing grain size related to changing relative sea level. Coarser sediments are paler; finer sediments are darker. The lowest sands are estuarine channel deposits, overlain by a succession of interbedded offshore muds and sands, where the sands record advances of a shoreface in response to falling relative sea level. The thick yellow sand at the top contains hummocky cross stratification and is a high-energy shoreface deposit introduced rapidly by a major fall in sea level. Bracklesham and Barton Formations. Eocene, Dorset, England. C) Succession of mainly thick, massive, turbidite sandstone beds interbedded with thin mudstones or with packages of more thinly interbedded sandstones and mudstone (~6m thick). There is no obvious overall organization of the beds into patterns of systematic bed-thickness change. Subtle erosion and bed-thickness changes are present and the partings that separate some thick beds die out laterally through a process of bed amalgamation. Ross Formation, Upper Carboniferous, western Ireland. D) Interbedded sharp-based sandstones and mudstones in the lower part are sharply overlain by a thick sandstone unit with medium-scale cross bedding. Follifoot Grit, Upper Carboniferous, Yorkshire, England.

be used for widespread correlation. Key surfaces are usually generated as responses to external influences on sedimentation and commonly reflect the interactions of sea-level change, tectonic subsidence or uplift, sediment supply and, particularly in the case of non-marine environments, climate change. Together these factors control the production of "**accommodation space**", the capacity of an area to accumulate sediment, and the extent to which the available accommodation is filled by sediment. These interactions are complex and lie behind the theory and practice of "**sequence stratigraphy**", a thorough account of which is beyond the scope of this book.

A couple of examples will serve to illustrate the range of key surfaces. First, where accommodation space is being created rapidly through rising sea level or rapid subsidence, sediment supply may not be able to keep up and deepening and flooding of the depositional setting may occur over time. Such flooding surfaces are commonly expressed as a sudden reduction in grain size, from sands below to muds above, sometimes with evidence of deepening. Intense bioturbation at some flooding surfaces reflects a period of condensation due to reduced sedimentation or non-deposition (see §9.4). Flooding surfaces can have a wide range of lateral extents. Some are very extensive and are related to eustatic (global) changes in sea level. Others are more local, confined to embayments flooded as a result of localized subsidence or a switch in the location of sediment supply. Where accommodation space is destroyed, for example due to a fall in sea level or to tectonic uplift, erosion occurs and rivers, where present, cut deeply down into the landscape to produce incised valley systems. These valley-shaped erosion surfaces, which eventually become buried as new accommodation space is once again created, may be very extensive and are one expression of a "**sequence boundary**". Interactions of the various controlling variables can give a wide range of complex balances between sediment supply and accommodation space. The result is that the same controlling effects can produce different sedimentary expressions in different parts of a depositional area for a particular instant in time. The practical consequence is that any sedimentary surface across which grain-size changes abruptly should be considered carefully, its possible significance assessed and the stratigraphic implications thought through. These surfaces, if correctly identified and interpreted, can be the key to stratigraphic understanding, a basis for correlation and a way of understanding large-scale controls on deposition.

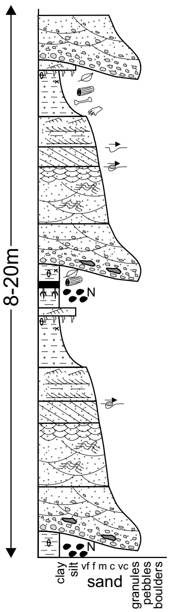
In sequence stratigraphy, trace-fossil ichnofacies are of great value in the recognition and interpretation of certain key stratal surfaces. In particular, unconformities associated with base-level fall (and the generation of sequence boundaries), surfaces associated with minor erosion as a result of rapid transgression (ravinement surfaces), surfaces of non-deposition that encompass considerable time (omission surfaces) and condensed sections generated by very slow rates of sediment accumulation are typically difficult to recognize from their physical characteristics alone. However, as these surfaces form, they are commonly colonized by organisms that leave characteristic traces both on the surfaces themselves and in the immediately underlying sediments. Although discontinuity surfaces may be generated in either sub-aerial or sub-aqueous settings, their colonization by animals typically occurs under marine influence. Many discontinuity surfaces result in the formation of firmgrounds and hardgrounds, which may be recognized as thin horizons dominated by the *Glossifungites* ichnofacies (see §9.4.6). An interpretation of the type of discontinuity can also be made by considering the type of ichnofacies immediately underlying and overlying the surface itself. For example, the occurrence of an omission surface in a deep-marine setting may be considered to be coincident with periods of high relative sea level when sedimentation is confined to the flooded shelf regions. In deep-water settings, the pre- and post-omission surface suites may be characterized by the *Zoophycos* ichnofacies, whilst the omission surface itself may be characterized by the *Glossifungites* ichnofacies. The duration of the time interval encapsulated by a discontinuity surface may control the intensity of any burrowing. Understanding the genetic significance of a discontinuity surface typically requires the integration of sedimentological, stratigraphical, ichnological and palaeontological techniques.

10.6 Interpretation of lateral relationships in sedimentary rocks

Although the vertical sequence is vital to the interpretation of rock successions, there are many cases where lateral relationships also play an important role. Where lateral

A

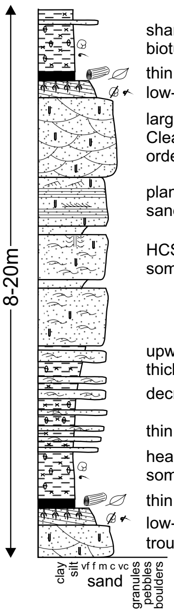
fluvial facies succession (sinuous, laterally-accreting river channel)



main facts	processes	environment
trough cross bedding with pebble lags erosive base non-marine fossils and trace fossils preserved ripple forms ripple cross-lamination flute and scour marks wood debris intraformational rip-up clasts thin coals rootlets and/or nodules desiccation cracks siltstone ripple cross lamination planar cross bedding smaller-scale trough cross bedding some deformed bedding larger-scale trough cross-bedding erosive base	erosion by fluid scour winnowing of finer sediment to leave a lag of intraformational clasts desiccation cracks indicate drying out of a formerly wet/damp surface post-depositional colonization by animals flute marks indicate a range of flow-transport directions intraclasts indicate erosion is taking place near-by deposition of silt from suspension post-depositional burrowing growth of concretions rapid deposition of sand during bedload transport as current ripples net accumulation of sand during migration of dunes post-depositional liquefaction suggests rapid deposition erosion by fluid scour winnowing of finer sediment to leave a lag of intraformational clasts	fluvial channel cut due to either lateral accretion or avulsion continental, overbank inter-channel setting accumulating sediment by vertical accretion concretions grow in calcrete soil profile laterally accreting channel with deposition on one side, from a point bar a broad range flow directions, as determined from palaeo-current indicators would imply a highly sinuous stream system rip-up clasts within the channel fill imply bank erosion as part of the lateral migration of the channel erosional channel scour

B

shallowing-upward shoreface facies succession

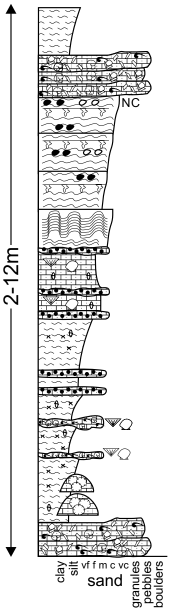


main facts	processes	environment
sharp transition to fossiliferous & bioturbated mudstones thin coals, rootlets & plant fossils low-angle laminated sandstones large-scale trough cross-bedding Clean sandstone arranged into ordered sets, vertical burrows planar bedded and rippled sandstone HCS-SCS some ripple cross stratification upward increase in sand bed thickness decreasing bioturbation upwards thin, silty sandstone beds heavily bioturbated mudstone some shelly fauna thin coals, rootlets & plant fossils low-angle laminated sandstones trough-cross bedded sandstone	deposition from suspension intense marine bioturbation marine flooding preservation of organic matter suggests low oxygen conditions dune migration, strong currents, erosion and winnowing transport by weak currents Storm-wave processes churning bed mixture of deposition from bedload and suspension deposition from suspension plus weak episodic decelerating flows deposition from suspension intense marine bioturbation marine flooding Dune migration, strong currents, erosion and winnowing	offshore marine shale below storm wave base marine transgression (flooding) coal swamp (?humid climate) emergence - foreshore (beach) upper shoreface, shallow water depth, within normal wave base (5-20m) middle shoreface lower shoreface, deeper water depth, within storm wave base (20-40m) offshore transition zone offshore marine shale below storm wave base marine transgression (flooding) coal swamp (?humid climate) emergence - foreshore (beach) upper shoreface

Figure 10.7 Examples of some commonly occurring vertical successions of sediment showing their interpretation in terms of process and environment by the application of Walther's principle. Part (C) is based in part on Pratt et al. (1992). See Figure A6.1 for a guide to the symbols used.

C

shallowing upward peritidal carbonate succession

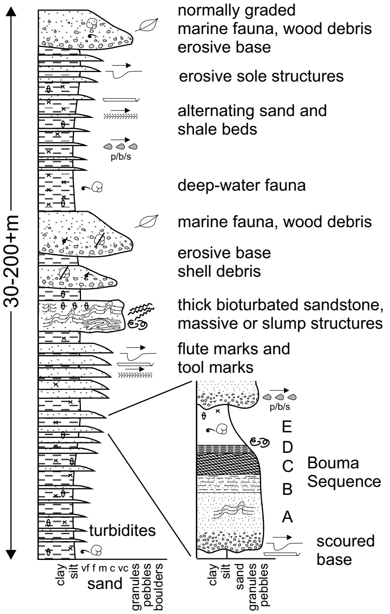


main facts	processes	environment
wackestone and mudstone	low energy	
flat pebble intraclasts broken shell debris	high-energy erosion and redeposition	flooding surface - marine transgression
dissolution cavities	nodule growth - high-salinity dissolution	supratidal flat in arid environment
desiccation cracks	periodic drying-out of surface	
nodular anhydrite	microbial colonisation and growth	microbial mat development around high tide level
microbial laminates	physical and chemical microbial-sediment interactions	
dome-shaped stromatolites	colonisation by microbial organisms (requires an absence of grazers which may signify high salinity in Phanerozoic successions)	intertidal zone
bioturbated lime mudstone with shelly fauna		shallow subtidal zone, probably a quiet energy lagoon setting
peloidal grainstone		
wackestone and mudstone peloidal grainstone	colonisation of muddy substrate by shelly and soft-bodied fauna	offshore shallow marine, predominantly low energy during fair weather but subject to episodic storm events
bioclastic grainstone	bioclastic debris may be concentrated by infrequent strong currents	
carbonate mudstone	generally low energy but with sufficient currents to supported filter feeders in patch-reef communities	
patch reefs		flooding surface - marine transgression
bioclastic and intraclastic grainstone	high-energy erosion and redeposition	high-energy beach

prograding carbonate shoreline resulting in emergent conditions

D

deep-water turbidite succession (progradation of a submarine fan system)



main facts	processes	environment
normally graded marine fauna, wood debris erosive base	erosive channel cut and fill	inner-fan channel fill
erosive sole structures	scour structures and tool marks indicate moderate to strong unidirectional currents	mid-fan lobes
alternating sand and shale beds	sedimentation from suspension and via dilute turbidity currents	distal outer fan
deep-water fauna	slow accumulation from suspension (pelagic settling)	condensed section deepening event (sea-level rise)
marine fauna, wood debris	waterlogged wood debris transported from up-dip coastal area	inner-fan channel fill
erosive base shell debris	repeated erosive channel cut and fill events	channelled part of mid-fan base-of-slope setting
thick bioturbated sandstone, massive or slump structures	chaotic sedimentation - deposit of mass flow	slump structure transitional downslope into debris-flow deposit
flute marks and tool marks	sedimentation from high-energy turbidity currents (Bouma A-C)	mid-fan lobes
	sedimentation from sand-prone turbidity currents (Bouma A-E)	smooth outer part of mid fan (suprafan lobes)
	sedimentation from suspension and via dilute turbidity currents (Bouma C-E)	distal outer fan

Figure 10.7 Continued.

Figure 10.8 ASTER satellite image of the Songhua River in northeastern China. The low gradient of the Manchurian Plain has resulted in the generation of a broad channel belt across which the river meanders. Meander loops have become amplified through swing and sweep processes as the channel has repeatedly 'combed' across the active channel belt. Lateral accretion surfaces on the inside of channel bends arise through the migration of point bars (scroll bars). Multiple avulsion and neck cut-off events have generated many abandoned reaches and oxbow lakes are numerous. The style of partial overprinting and reworking of multiple channel segments makes it possible to establish a relative chronology of events for some parts of the channel belt. Several larger floodplain lakes are evident. The scale and shape of many of the abandoned meander loops are similar, indicating that the river has repeatedly evolved to a similar form prior to neck cut-off and avulsion. Image courtesy of NASA Earth Observatory.

changes are recorded, they commonly allow refinement of the environmental model or the resolution of uncertainties that remain from consideration of a single vertical sequence. For example, a sequence of interbedded mudstones and sharp-based, graded sandstone beds can be interpreted in terms of normal and catastrophic deposition, but such processes may take place in a variety of environmental settings. If it was found that the sequence was laterally equivalent to a channel sandstone showing abundant unidirectional cross bedding, it might be a fair inference that the overall setting was fluvial. The interbedded sequence might then be interpreted in terms of crevasse splays (bank breaches) into an overbank flood plain or lake during floods. In contrast, if the interbedded sequence proved to be the lateral equivalent of highly bioturbated sandstones, then the setting might be a shallow-marine shelf with the high-energy events being storms. In many cases, it is the combined evidence of physical sedimentary structures, together with fossil and trace-fossil evidence that enables an environmental interpretation to be made with confidence.

The lateral migration of sinuous river channels as they migrate over their floodplain generates large point-bar deposits (Fig. 10.8; also see §2.2.3) The repeated abandonment of sinuous river reaches in response to meander growth, tightening and eventual cut-off causes many point-bar elements to accumulate over an alluvial plain. Gradual on-going subsidence causes later point-bar elements to be accumulated over older ones, thereby building a vertical accumulation. Stacked vertical successions of fining-upward point-bar deposits tend to occur as predicable successions of beds. In many cases point-bar elements are separated from each other in vertical section by finer-grained floodplain elements, which themselves generally record episodes lower-energy non-channelized deposition of fine-grained sediments. Understanding the relationship between surface form and process (Fig. 10.8), and the resultant stratigraphic arrangement of accumulated **architectural elements** (e.g. point bar, floodplain) enables the development of **facies models** (Fig. 10.9), which serve as a way to reconstruct sedimentary palaeoenvironments.

As a further example, we can consider the lateral variability in sediments in the upper parts of widespread, upwards-coarsening deltaic successions (Fig. 10.10). The nature of the variability may allow us to make more specific inferences about the type of progradation rather than just record a general deltaic interpretation. If the upper, sandy sediments are laterally very uniform in character with evidence of wave-generated structures, then that suggests that sand was distributed along the shore from river mouths and it implies a high level of basinal wave energy. The delta was therefore probably arcuate in form and characterized by a laterally continuous prograding, wave-dominated shoreface (Fig. 10.10A). By contrast, if the upper part of the progradational unit is very variable, with channel sandbodies in one place, upwards-coarsening sandstones with unidirectional current structures in others, and interbedded sandstones and mudstones elsewhere, a delta similar to the modern delta of the Mississippi might be thought more likely (Fig. 10.10B). Low basin energy leads to sands being deposited as local mouth bars with little reworking, whilst muddy shorelines prograde in interdistributary areas. Distributary channels cut across these elements to give the complex mosaic observed. Studies of modern deltas and the distribution of sedimentary structures within the deposits therein enables fundamental types of delta to be classified (Figs 10.11, 10.12, 10.13) and geologists can then seek to establish sedimentological criteria for the recognition of such types preserved in the ancient rock record.

These comments on the interpretation of vertical and lateral relationships in rock successions are beginning to carry the discussion beyond the stated scope of this book. They are intended merely as an introduction to the more complex field of **facies analysis** and palaeoenvironmental interpretation that leads, through stratigraphy, to the development of palaeogeographic and geotectonic reconstructions.

Study techniques

Field experience

Operating as a group or individually, students should practice constructing sedimentary logs from outcrop (natural shorelines and river-bank sections, together with road cuttings and disused quarries are ideal). Construct several logs from the same stratigraphic interval but separated laterally from each other by 100–200m. Compare each drawn log and identify similarities and differences between them. While in the field, try to trace prominent beds and bed boundaries laterally between logs by physically walking them out. Do the beds exhibit marked lateral variability or are they continuous and uniform? What does this imply about the nature of sedimentation across the area? Photograph key sections both vertically and horizontally.

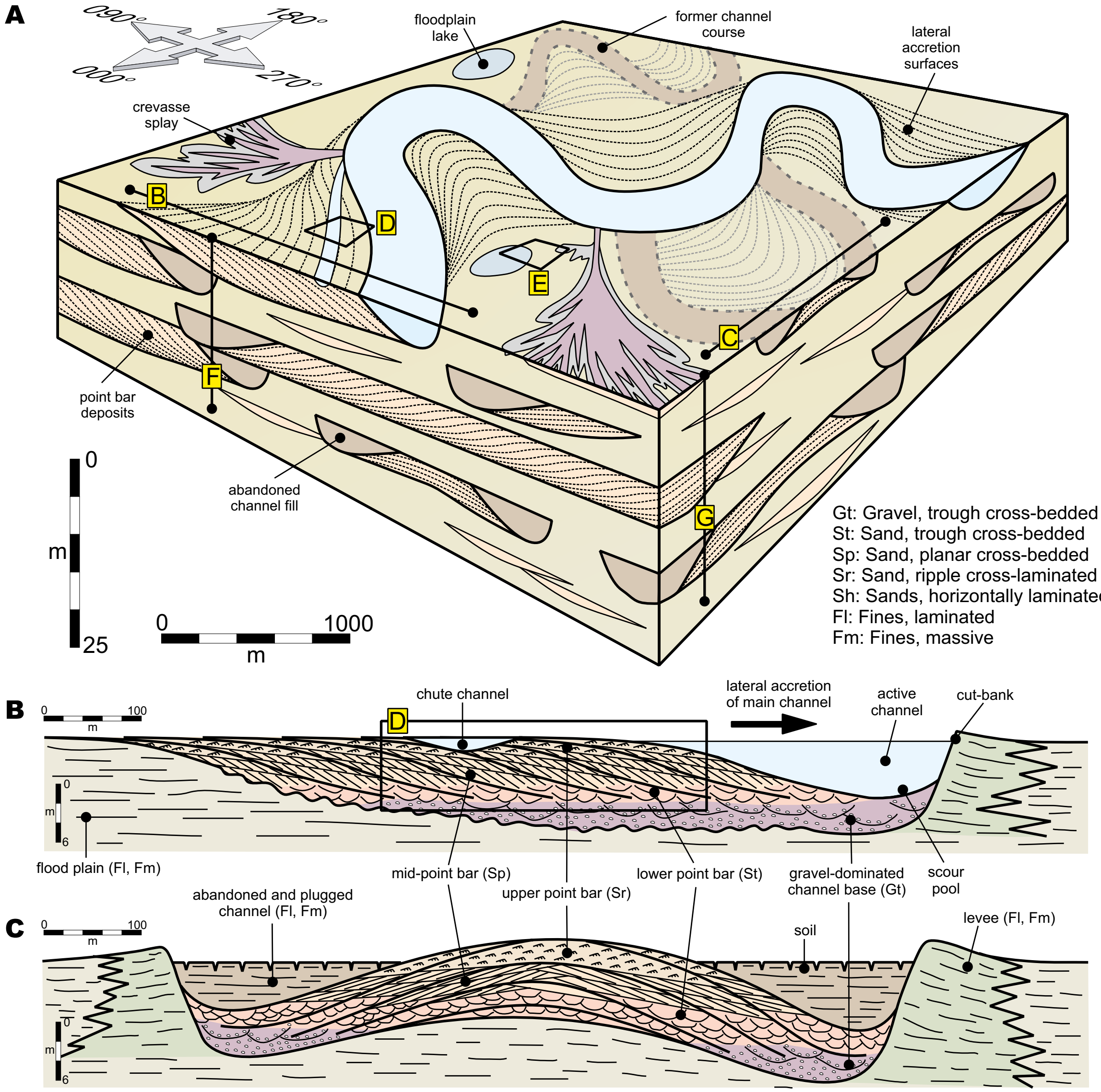
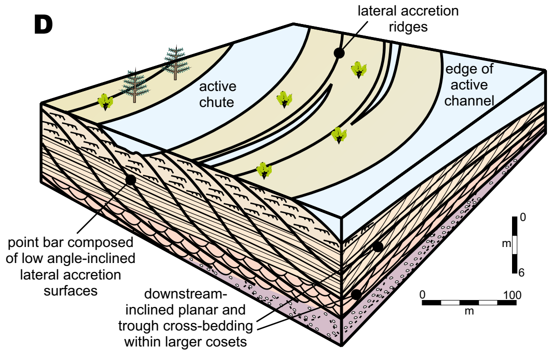


Figure 10.9 Facies model for a meandering river succession depicting the hierarchy and nature of stacking of facies, facies associations, architectural elements, and depositional elements (channel complex and floodplain) that make up the succession. Modified after Ghazi and Mountney (2009). The overall preserved architecture of meandering fluvial deposits is dictated by a number of factors. The depth of the main channel and the wavelength of the meanders determines the thickness and width of accumulated point-bar elements. The rate of swing and sweep controls how fast and how far the point bars migrate laterally over their lifetime. The longevity of meander-loop evolution from inception through to neck cut-off dictates how far a point bar can migrate laterally. The extent to which meander loops swing across their floodplain dictates the overall width of the active channel belt. The avulsion frequency dictates how long a single channel belt is active for before a new channel belt is initiated. The rate of subsidence dictates how successive channels and point-bar deposits stack vertically on top of each other. A fast rate of subsidence will favour preservation of vertically separated and spatially isolated point-bar deposits, whereas a slow rate of subsidence will favour preservation of vertically amalgamated, interconnected point bar deposits. Variation in all these controlling parameters gives rise to a complex range of possible architectures resulting from meandering fluvial river accumulation. As point bars progressively migrate, so the range of facies within them (each represented by diagnostic sedimentary textures and structures) migrate with them and this gives rise to a characteristic facies succession that accumulates and may become preserved in the geologic record. This provides a mechanism by which we can recognise the deposits of meandering rivers in the ancient rock record. Predicting these architectures is important for economic reasons. For example, effective prediction of the size and distribution of sandbodies (e.g. point-bar deposits) dictates reservoir quality in oil and gas provinces and the behaviour of groundwater aquifers.

D


lateral accretion ridges
active chute
edge of active channel
point bar composed of low angle-inclined lateral accretion surfaces
downstream-inclined planar and trough cross-bedding within larger cosets
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6
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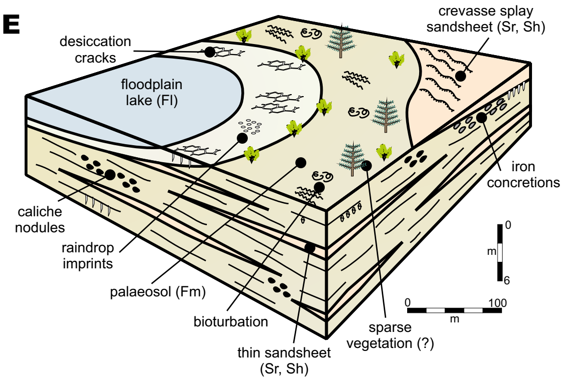
E


desiccation cracks
floodplain lake (Fl)
crevasse splay sandsheet (Sr, Sh)
iron concretions
caliche nodules
raindrop imprints
palaeosol (Fm)
bioturbation
thin sandsheet (Sr, Sh)
sparse vegetation (?)
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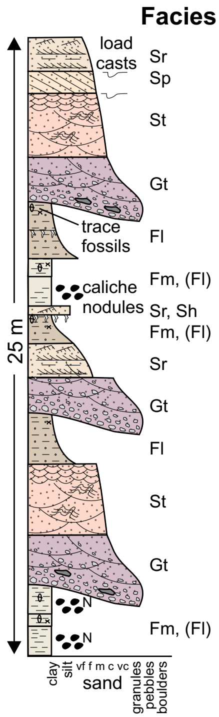
F


Facies
load casts
Sr
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Fm, (Fl)
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granules
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boulders

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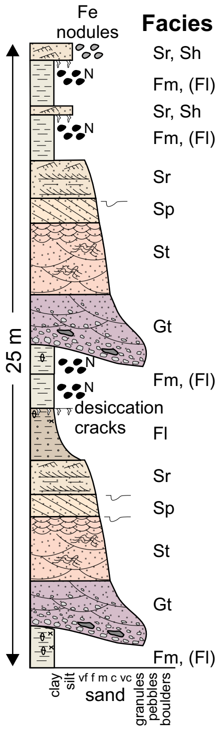
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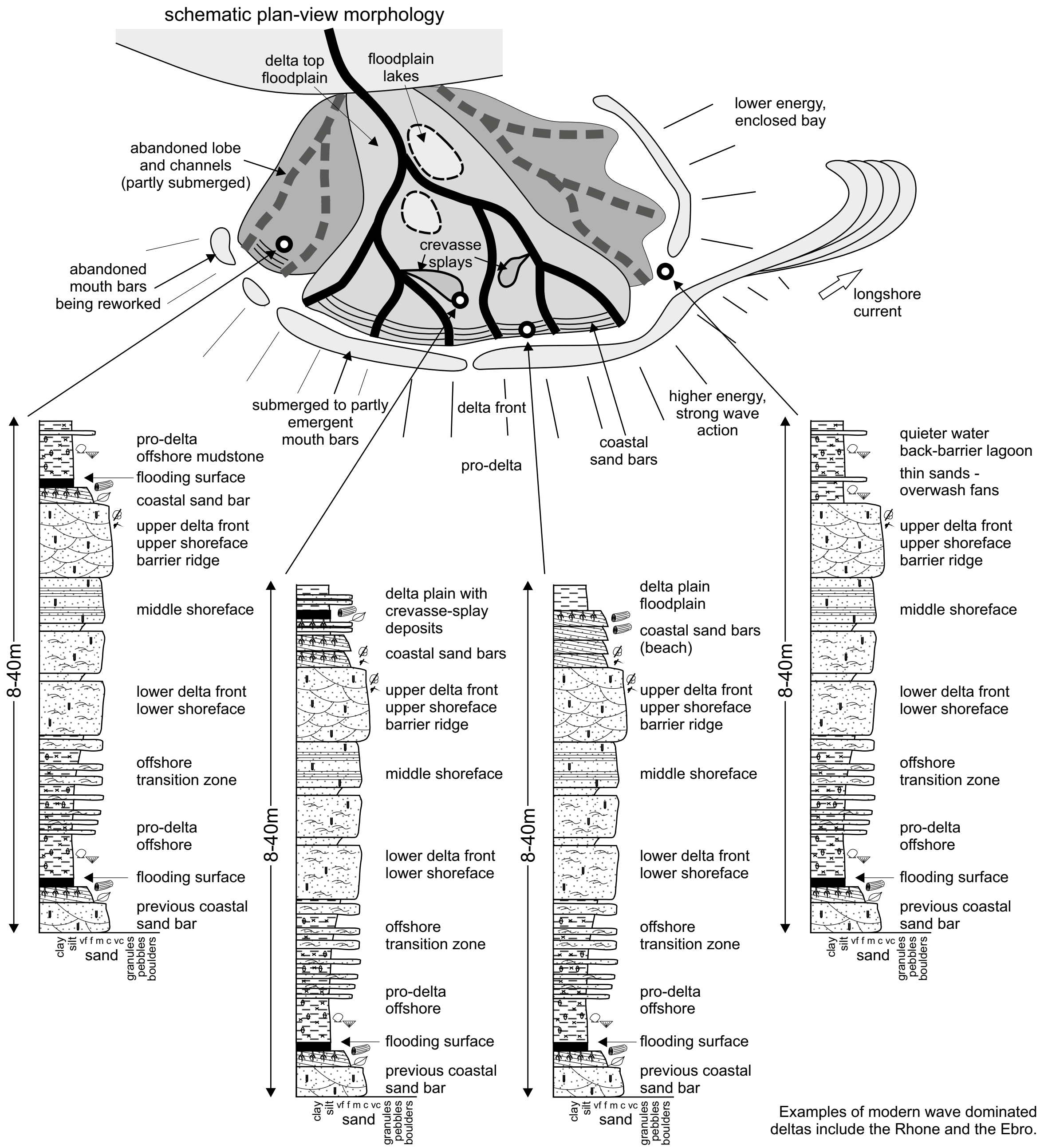


Figure 10.10A Example illustrating the importance of appreciating lateral variability in sedimentary successions in order to distinguish specific types of sedimentary environments. Wave-dominated delta model. See Figure A6.1 for a guide to the symbols used.

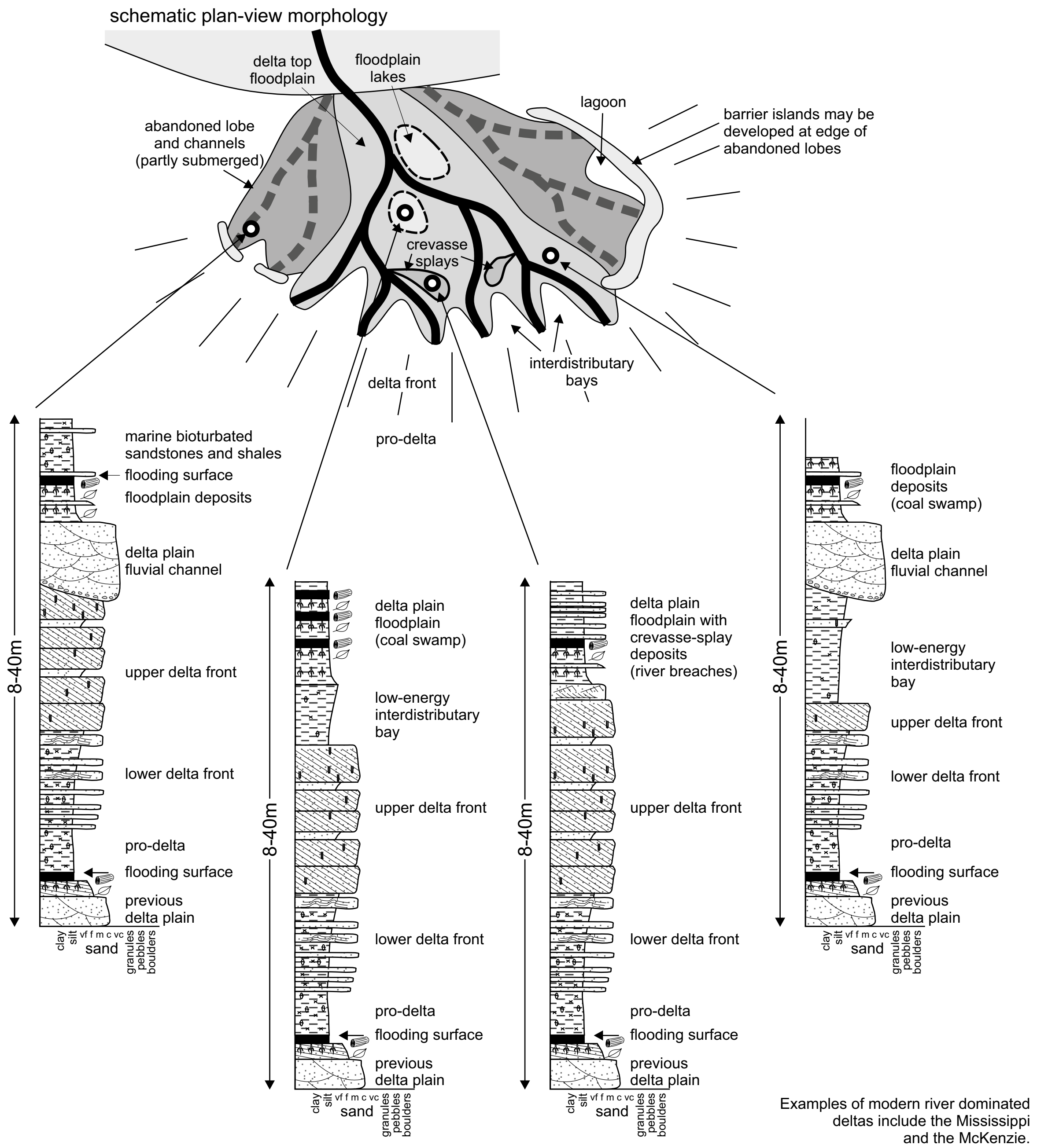
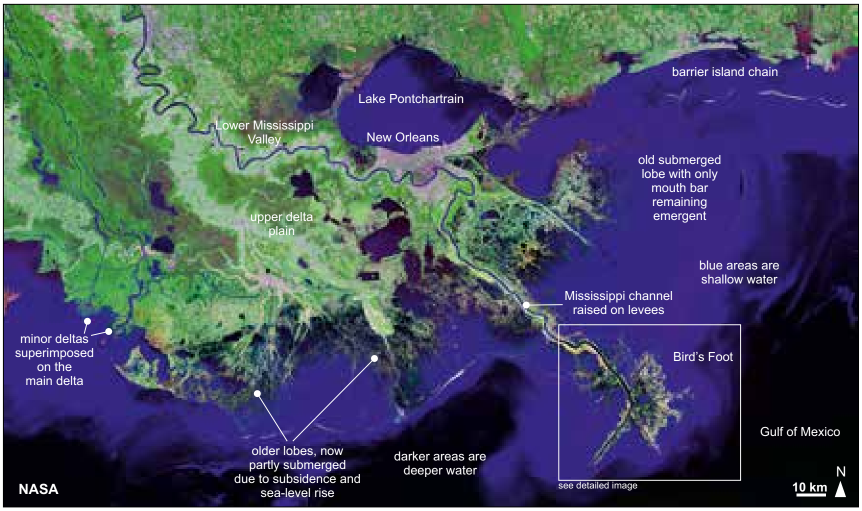


Figure 10.10B Example illustrating the importance of appreciating lateral variability in sedimentary successions in order to distinguish specific types of sedimentary environments. River-dominated delta model. See Figure A6.1 for a guide to the symbols used.



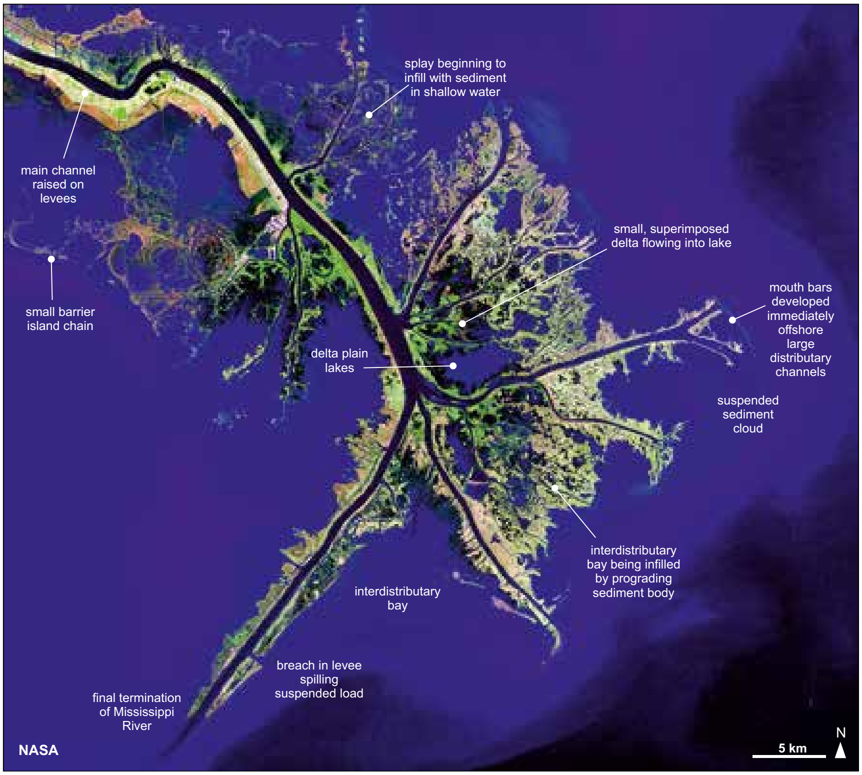


Figure 10.11 Main. Satellite image of the Mississippi Delta, Gulf of Mexico coast, USA, showing the presently active Bird's Foot lobe, a series of older, largely submerged lobes with only mouth bar deposits remaining emergent, and the upper delta plain around the city of New Orleans. Detail. Satellite image of the Mississippi Bird's Foot, which represents the currently active lobe of the Mississippi Delta and has developed in just the last 500-600 years. The finger-like, digitate morphology of the main distributary channels is typical of river-dominated deltas whose principal sedimentary load is a mix of both silt (dominantly carried in suspension) and fine sand (dominantly carried as bedload). Images courtesy of NASA Earth Observatory.

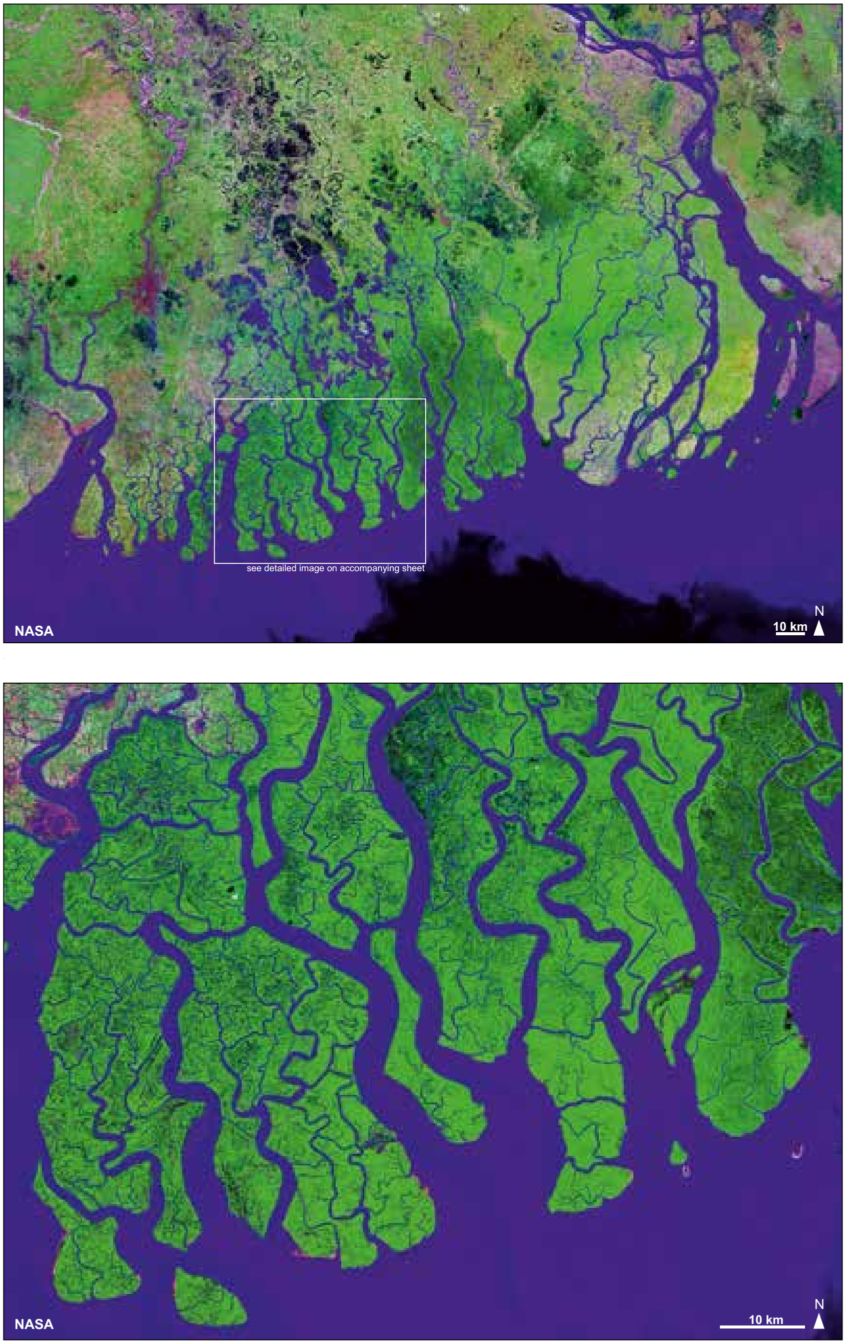
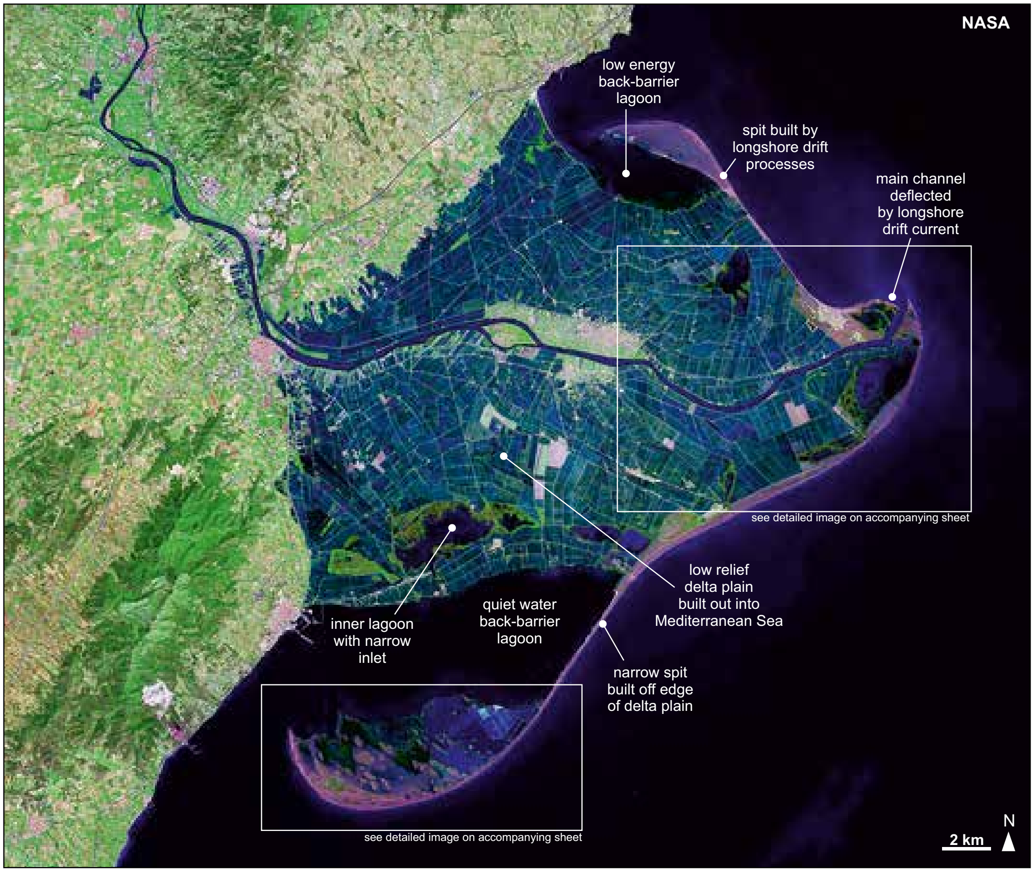


Figure 10.12 Main. Satellite image of the Ganges Delta, Bangladesh. Detail. Tidal channels of the lower delta plain, the so-called 'Mouths of the Ganges'. The channels exhibit a fractal geometry whereby successively smaller channels are nested inside parts of the delta plain bounded by larger channels. Channels have an anastomosing pattern and many are joined at both ends to the coast. The mid-blue colour highlights sediment suspended in the upper part of the water column. Images courtesy of NASA Earth Observatory.



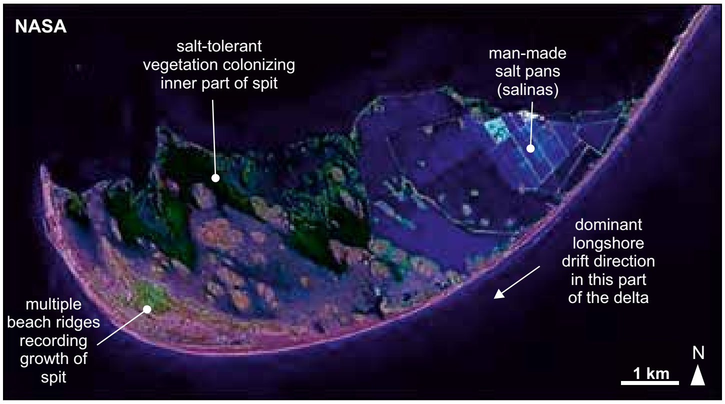


Figure 10.13 Main. Satellite image of the wave-influenced Ebro delta, Spain. Detail. The main Ebro River channel at its point of entry into the Mediterranean Sea. Note the absence of multiple distributary channels in this system, in part because wave action and longshore currents repeatedly rework sediment along the front of the delta. This prevents choking of the main channel by mouth-bar deposits, thereby reducing avulsion frequency. The channel is also managed to prevent avulsion. Detail. Wave-modified spit generated by longshore currents on the southern margin of the Ebro Delta. Note the multiple arcuate sand ridges that record the growth of the structure. Images courtesy of NASA Earth Observatory. Detail (overleaf). Termination of the main river channel with adjacent wave-dominated beach barrier system and back-barrier lagoon on its landward side.

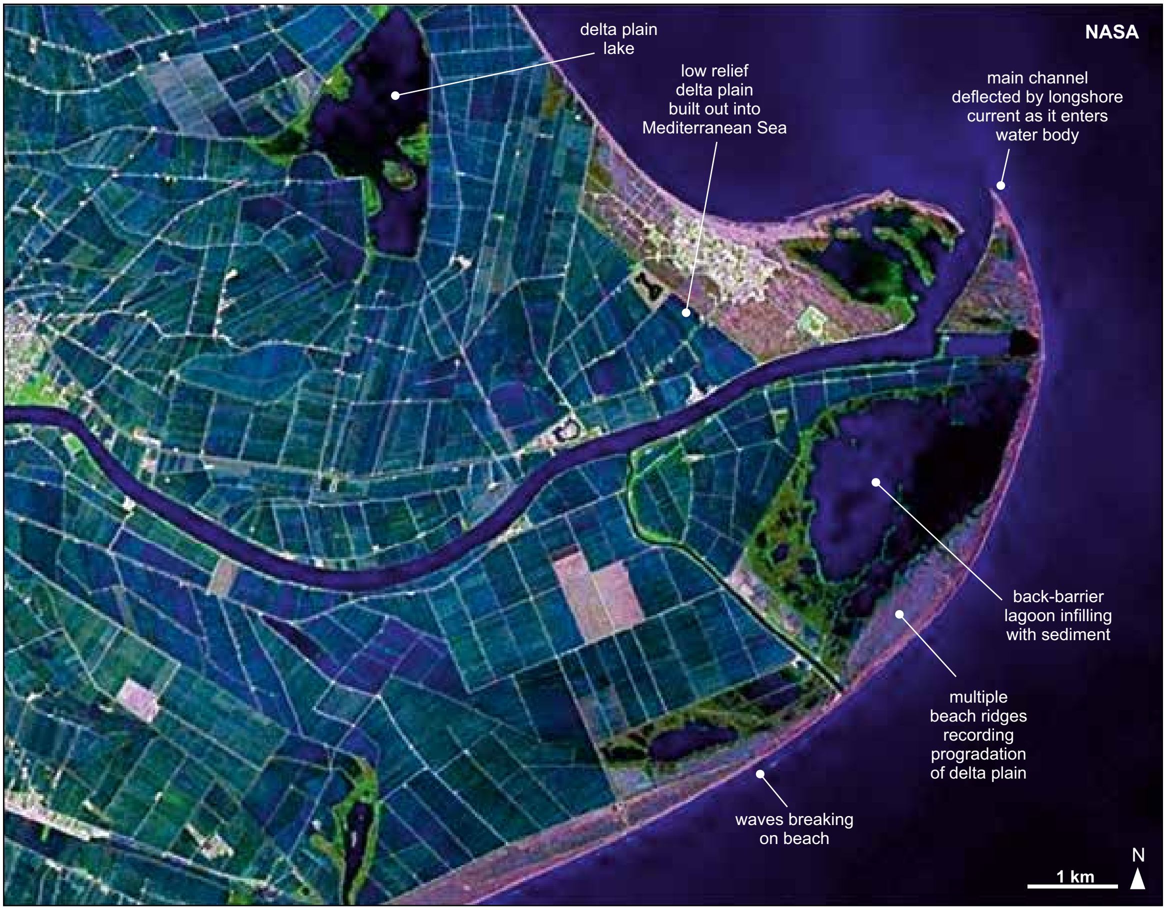


Figure 10.13 *Continued*

Laboratory experience

Suggestions for the following-up of fieldwork and the analysis of data are given in Appendices 1–6. The logging of borehole cores by groups of students can be particularly helpful. Some departments keep such cores for teaching purposes. After all the individuals in a group have completed their logs, comparisons between their efforts and those of their tutors, and discussion of differences and discrepancies can be very instructive.

Recommended references

Allen, P. A. & J. R. Allen 2005. *Basin analysis.* Principles and applications. Discusses how sediments accumulate to form larger basin-filling successions.

Brenchley, P. J. & B. J. P. Williams (eds.) 1985. *Sedimentology: recent developments and applied aspects*. An interesting mixture of papers with good reviews of clastic and carbonate facies models and on the application of sedimentology to the oil industry.

Bridge, J. S. 2003. *Rivers and floodplains: forms, processes and sedimentary record.* Well illustrated examples of facies models for various types of fluvial successions.

Brookfield, M. E. 2003. *Principles of stratigraphy*. Discussion of various stratigraphical techniques used in the development of facies models.

Cant, D. J. & R. G. Walker 1978. *Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada.* A good example of converting a geomorphological description into a facies model.

Coe, A. L. (ed.) 2010. *Geological field techniques*. Explains how to record sedimentological information in the field and how to best make use of it for mapping and palaeoenvironmental reconstruction. Well illustrated.

Coe, A. L. 2003. *The sedimentary record of sea-level change.* A well-illustrated introduction to explain how sedimentological principles can be effectively integrated with techniques in sequence stratigraphy.

Doyle, P., Bennett M. R. & A. N. Baxter 2001. *The key to Earth history: an introduction to stratigraphy*. A well-illustrated introductory text that demonstrates the application of simple techniques in stratigraphy in the analysis of sedimentary successions.

Emery, D. & K. J. Myers, 1996. *Sequence stratigraphy*. A simple introduction to fundamental techniques in sequence stratigraphy.

Galloway, W. E. & D. K. Hobday 1983. *Terrigenous clastic depositional systems.* A very good account of these systems including subsurface examples.

Hallam, A. 1981. *Facies interpretation and the stratigraphic record*. A succinct account explaining the nature of the stratigraphic record.

Hamblin, W. K. & J. D. Howard 1995. *Exercises in physical geology.* A beautifully illustrated set of examples of modern-day environments in a variety of settings. Other, later editions are also available.

James, N. P. & R. W. Dalrymple 2010. *Facies models 4.* An edited volume with separate chapters each considering a particular type of sedimentary environment in terms of facies analysis. The chapters are written by leading authors in the field.

Leeder, M. 2011. *Sedimentology and sedimentary basins.* Considers how sediment processes give rise to lithofacies and successions thereof, and how these are accumulated to form the fills of sedimentary basins.

Lindholm, R. 1987. *A practical approach to sedimentology*.

Posamentier, H. W. & R. G. Walker (eds.) 2006. *Facies models revisited.* A thematic set of chapters, each of which covers a specific type of sedimentary environment. A detailed explanation of how the principles of facies analysis are applied to sedimentary systems and their preserved successions.

Raaf, J. F. M. de, Reading H. G. & R. G Walker 1965. *Cyclic sedimentation in the Lower Westphalian of north Devon, England.* A classic pioneering paper of the development of facies schemes and on the interpretation of facies sequences.

Reading, H. G. (ed.) 1996. *Sedimentary environments; processes, facies and stratigraphy (3rd edn).* A wide ranging treatment of facies models and their sequence stratigraphical settings.

Scholle, P. A. & D. Spearing (eds.) 1982. *Sandstone depositional environments.* A well-illustrated compendium of facies models.

Tucker, M. E. 2003. *Sedimentary rocks in the field.* An excellent pocket guide for use in the field.

Walker, R. G. & N. P James (eds.) 1992. *Facies models: response to sea level change.* Excellent summaries of most depositional environments and their associated facies in a sequence stratigraphical context.

APPENDIX 1

Directional data: collection, display, analysis and interpretation

In earlier chapters, much has been made of the importance of certain sedimentary structures as palaeocurrent indicators. An individual measurement from a particular structure can, in most cases, have only local significance; in order to develop a feel for directions of wider significance, it is usually necessary to collect a considerable number of measurements. This appendix deals with some of the methods by which such data may be collected, displayed and analysed so that they give the most representative and reliable basis for interpretation.

Collection, restoration and presentation of data

Collection

From the various chapters in this book, it should be clear how directional data can be derived from particular structures. Palaeocurrent indicators revealed by sedimentary structures are of two basic types: **planar** features such as the foresets of cross-bedding and cross-lamination, and **linear** features such as groove marks, axes of trough cross-beds or primary current lineation. For modern sediments and for ancient ones that have undergone little or no tectonic displacement, the data can be collected and used directly.

Restoration

When the rocks have undergone considerable tectonic tilting, it may be necessary to re-orientate the directional measurements by removing the effects of the tilting and restoring the original bedding to horizontal. In doing this, the structures within the beds that act as the palaeocurrent indicators will themselves be restored (rotated) to their original attitude at the time of deposition. Restoration is performed using the following procedures.

For linear structures such as flutes, primary current lineation, the alignment of ripple crests or the axes of sets of trough cross bedding, deviations induced by tectonic dips of less than 20° are sufficiently small to be ignored. However, serious deviations occur when measurements are made on the foresets of cross-bedded sets. Tectonic dips of greater than only 5° then require reorientation.

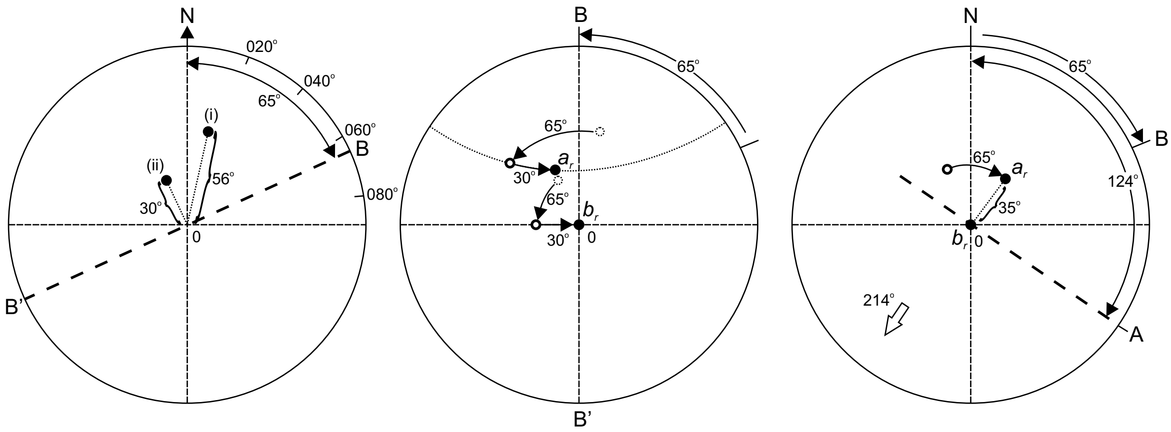
In order to restore cross beds to their original attitude, it is necessary to plot and manipulate the data on a stereogram or in a dedicated computer programme. This requires the magnitude and direction of dip both of the foresets as they now occur, and of the overall sequence (i.e. local tectonic dip). The procedure outlined here only applies if fold plunge is negligible. A more complex procedure is needed for plunging folds. Plot poles (normals) to both the foresets and the bedding on a stereographic projection (Fig. A1.1a). Rotate the points until the poles to the bedding lies on a great circle of the projection (Fig. A1.1a & b). In order that the beds may be restored to horizontal this point must be moved to the centre of the projection. As that shift is carried out, the pole to the foreset must be moved the same angular distance along the small circle upon which it lies (Fig. A1.1b). The new position of this point shows the pole to the foreset at the time of deposition and this can be converted back to a direction and magnitude of dip (Fig. A1.1c). This direction (foreset azimuth) may then be used as an indicator of palaeocurrent direction.

For linear data in steeply dipping beds, plot on the stereogram the attitude of the lineation in space (Fig. A1.2a) and rotate both the pole to bedding and the lineation as described above (Fig. A1.2b). This restores both the bedding and the lineation upon the bedding back to their original attitude prior to tectonic tilting. The orientation of the restored lineation may then be used to indicate palaeocurrent direction (A1.2c).

The principle of this restoration procedure is perhaps best grasped through practice and an example exercise is provided at the end of this appendix.

Presentation

Once directional measurements are restored to their original orientation, it is usually helpful to display them graphically. This can be done in several ways. The method chosen usually depends upon the quantity of data and the variety

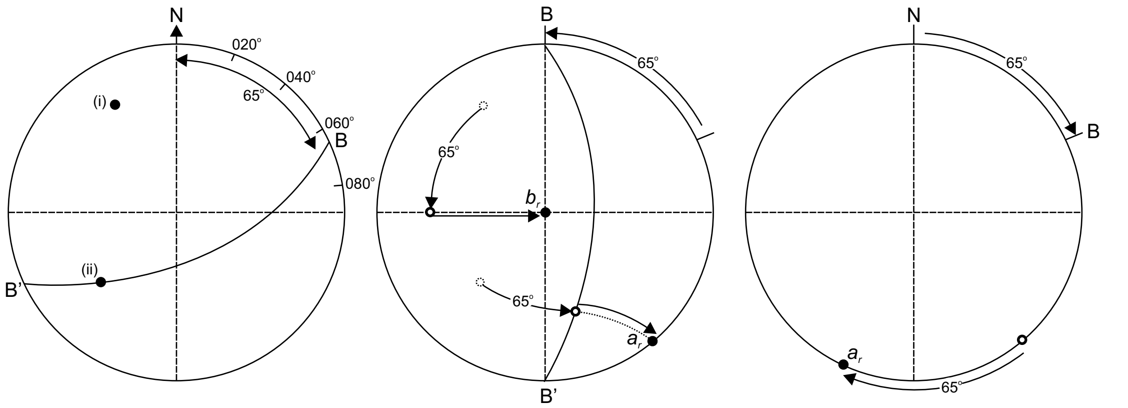


Observed data
Bedding strike = 065˚, dip = 30˚ SE
X-bedding strike = 102˚, dip = 56˚ SW
B–B′ = strike of bedding
(i) = pole of cross bedding
(ii) = pole of bedding

Rotation of bedding and cross bedding
b_r = pole to bedding rotated 30˚ to zero dip
a_r = pole to cross bedding rotated 30˚ in same direction along small circle

Restoration of a_r with respect to north to yield result
a_r = original dip of cross bedding 35˚ SW
A = original strike of cross bedding (124˚)
Original bedform migration to 214˚

Figure A1.1 Procedure for correcting planar cross-bedding data for tectonic tilt in order to determine original palaeocurrent direction. Bedding and cross bedding are plotted as poles to planes on a lower-hemisphere stereogram. In part B) the dotted circles represent the original placement of the poles to the bedding and to the cross bedding on the stereonet prior to calculation; the dashed circles represent a position midway through the calculation, and the filled black circles represent the position of the poles following completion of the rotation calculation. See text for fuller explanation.



Observed data
Bedding strike = 065˚, dip = 60˚ SE
Groove plunge = 34˚, azimuth = 224˚
B–B′ = strike of bedding
(i) = pole of cross bedding
(ii) = plunge and azimuth of groove

Rotation of bedding and groove mark
b_r = pole to bedding rotated 60˚ to zero dip
a_r = Lineation of groove mark rotated 60˚ in same direction along small circle to edge of stereonet

Restoration of ar with respect to north to yield result
a_r = original trend of groove mark on flat bedding surface (025˚–205˚)
The sense of direction cannot usually be discerned from groove marks.

Figure A1.2 Procedure for correcting linear data (e.g. groove marks) for tectonic tilt in order to determine original palaeocurrent direction or trend. For illustrative purposes, bedding data are plotted both as a plane and as a pole to that plane on a lower-hemisphere stereonet. Linear data are plotted as a plunge and a plunge direction (azimuth). See text for explanation.

of structures from which they were collected. Compilation inevitably leads to some loss of information; in particular the distribution of various directions, both laterally and vertically, within the sampled sequence. Compiling directional data is a useful way of visualising flow patterns but it is no substitute for relating directions to specific structures in a measured section when the aim of the exercise is to support environmental interpretation. Where compilation is carried out, it is important to produce plots that clearly distinguish the types of sedimentary structure from which the the palaeocurrent data are measured. This can be done either by producing separate plots for each type of structure or by using clearly distinguished symbols, colours or designs for each type of structure on a combined plot. It is also important to bear in mind that some structures can be recorded as a single direction to which flow is directed whilst others give only a trend along which flow could have been in either direction. Examples of the second group must be shown as double ended lines or sectors in any display.

Where data are few and have been collected from only a few types of structure, it is often convenient to plot each measurement as a radiating line of fixed length on a circular spoke diagram (Fig. A1.3). Where both dip direction (azimuth) and dip magnitude (inclination) have

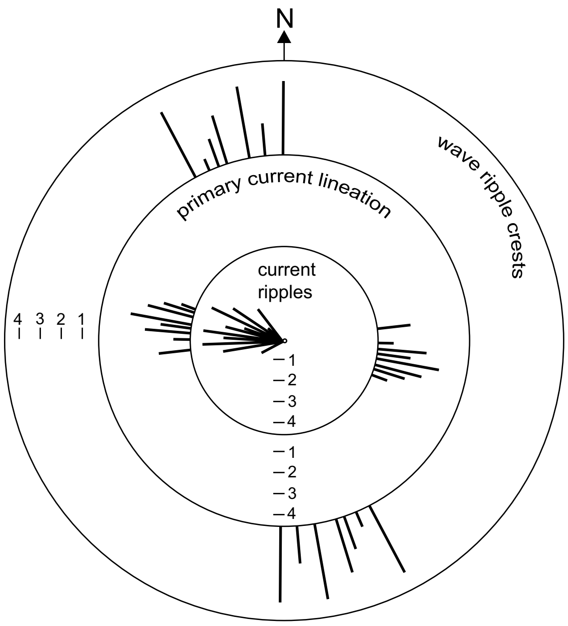


Figure A1.3 Current directions presented as a radial 'spoke' diagram. Note how different types of structure are separated and how those structures which give only trend and not sense of movement are plotted as double-ended lines.

been recorded from cross-bed foresets, the plotting of poles (normals) to the foresets on a stereogram may be preferred to plotting just the foreset azimuths.

When data are more numerous, spoke diagrams become very cluttered and data are then better grouped into classes in a circular histogram or "rose diagram" (Fig. A1.4). The class interval for such diagrams can be set at any size, although 10°, 15°, 20° and 30° are the most commonly used, depending on the volume and spread of directions. With abundant data, smaller class intervals show better the detailed structure of the population. However, with few data, small class intervals may give a false impression of complexity. In addition, the fact that rose diagrams usually employ a linear relationship between abundance and radius can lead to exaggeration of the apparent importance of more abundant classes. Modal classes may seem more abundant and well defined than they really are. There have been attempts to overcome this by designing schemes in which abundance relates to the area of the sectors but these are more difficult to apply and are not widely used. Where individual readings fall on a class boundary, it is good practice to allocate these equally either side of the boundary. Note that structures that yield only a palaeocurrent trend should be plotted as double-ended sectors (Fig. A1.4a), whilst those that yield a sense of direction can be plotted as single-ended sectors in the direction of transport (Fig. A1.4b).

Analysis and interpretation of data

When making an initial assessment of the pattern of current directions shown by a rose or radial line diagram there are at least four questions to have in mind:

- Is the pattern unimodal, polymodal or without any obvious preferred direction?
- What is the dominant or mean direction or directions?
- How widely scattered are the directions about the mean values (i.e. what is the spread of the directions)?
- Are there any systematic differences in the directions derived from different types of structure?

Mode of pattern

The most important feature of a population of directional data is its pattern of preferred directions. In rock sequences, this can often give information on both the nature of the depositional environment and the orientation of the regional palaeoslope. Although unimodal patterns are more common, certain environments generate bimodal or even

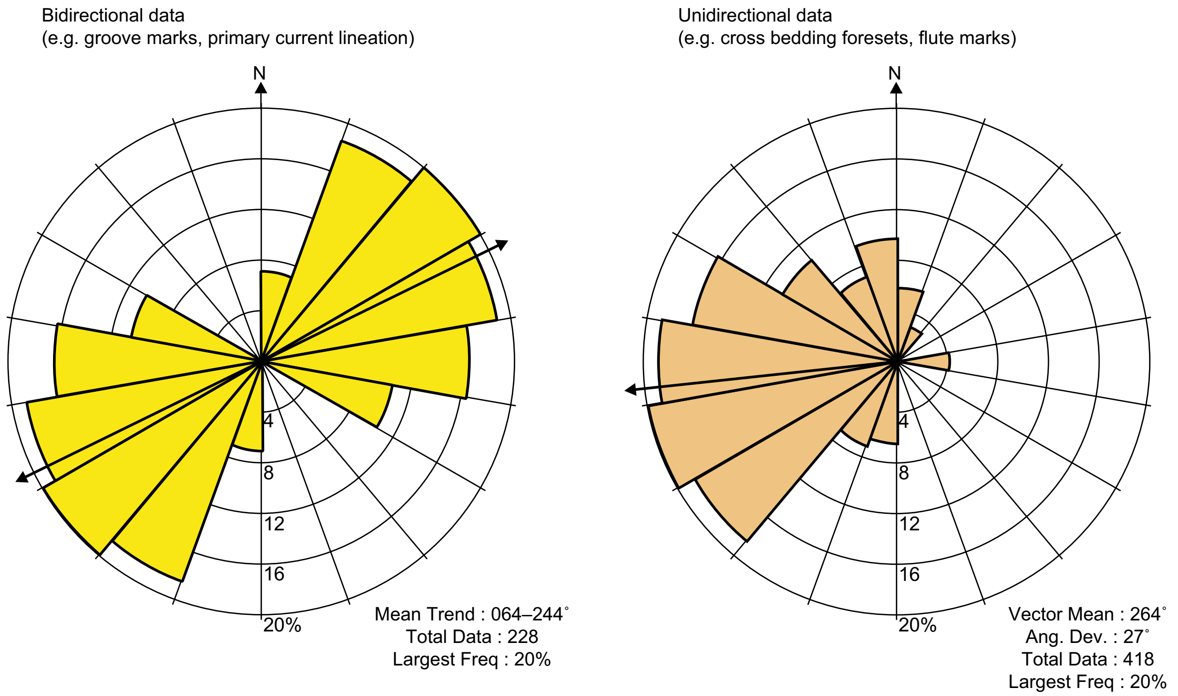


Figure A1.4 Current directions presented as 'rose' diagrams with 20 degree class intervals. Data from different classes of structure are usually presented on separate diagrams, and structures which give only trend and not sense of movement are plotted as double-ended sectors. The rose may be scaled in terms of either percentage all observations (as here) or actual number of observations. Arrows indicate the vector mean directions.

polymodal patterns. In some cases the modal directions derive from different types of structure (e.g. between sole marks and ripples in some turbidite sequences). In other cases structures from the same class of structure may be polymodal (e.g. tabular cross bedding in certain river deposits). An exceptional but highly diagnostic pattern is that of bipolar cross bedding (i.e. the two modes diametrically opposed), which is a strong indicator of tidal settings.

Mean direction

Where the distribution of directions is clearly unimodal, it is possible to calculate a mean direction. It would, however, be nonsense to calculate a single mean value for a bimodal population, especially one with a bipolar pattern. In that case, a single calculated mean could well be at right angles to both the dominant modal directions and hence be totally meaningless.

Because directional data are distributed around a circle so that 360° = 0°, calculation of mean is not a matter of simple averaging. For example, the mean of two directions close to but either side of north, say 350° and 10°, by simple averaging is 180° when clearly the sensible mean is 360°. This difficulty may be overcome at a simple level by using a false origin. After all, beginning the scale at north is a purely arbitrary convention. If an origin is chosen outside the range of recorded directions, a simple arithmetic average of the deviations from the false origin can be calculated. This can then be restored to a true bearing.

Such an approach is quite satisfactory for closely grouped data. However, a more widely applicable method treats the directions as vectors and resolves the vector components to give a **vector mean** by applying the formula:

$$tan\,\bar{\varphi} = \frac{\sum sin\,\varphi}{\sum cos\,\varphi} \qquad \text{(A1.1)}$$

where $\bar{\varphi}$ is the vector mean and $\sum sin\,\varphi$ and $\sum cos\,\varphi$ are the sums of the sines and cosines of the individual readings. In such calculations, it is essential to take account of the sign of the trigonometric functions and to have judged, by inspection, the general direction in which the mean is likely to lie.

In polymodal distributions, means can be calculated for data around each mode provided that the clustering is clear with no overlapping or ambiguous readings. Where there is overlap, calculations along these lines are probably meaningless and visual inspection of the rose diagram will be at least as good a guide to preferred directions.

Scatter of directions

The clustering of directions about a mean value may be quite close or more widely dispersed. In many cases it will be acceptable to describe dispersion qualitatively by inspection of the rose diagrams. In some circumstances a more quantitative expression of dispersion may be appropriate. The parameter most commonly used to express this is the **vector strength,** which for unidirectional palaeocurrent data (e.g. foreset dip azimuths) is given by the equation:

$$S = \frac{\sqrt{\left(\sum sin\varphi\right)^2 + \left(\sum cos\varphi\right)^2}}{n} \qquad \text{(A1.2)}$$

where n is the number of readings. High values indicate narrow dispersion and low values a wider dispersion.

One obvious example of environmentally significant modal patterns is the bipolarity that characterizes certain tidal deposits. However, caution is needed as not all tidal settings produce and preserve sedimentary structures with a symmetrical bipolar pattern. If either the ebb or the flood current dominates by even a small amount, an asymmetrical bipolar pattern or even a unimodal one may result.

In certain sandy braided river deposits, sets of tabular cross-bedding may show a bimodal pattern symmetrically distributed about a mean direction, which may coincide with the mean of a unimodal distribution derived from trough cross bedding.

If the mean directions of cross bedding from several fluvial sandstone units in an ancient sequence show a high dispersion, sinuous rivers are suggested; low dispersion suggests straighter channels.

In aeolian dune deposits, wind directions are not controlled by the topographic slope and more than one effective wind direction may be recorded as cross bedding modes. In particular, sets of cross bedding produced by linear (seif) dunes often show a bimodal pattern, whereas star dunes typically produce a polymodal distribution.

Differences in direction

Differences in the pattern of directions recorded from different types of structures in the same sequence can usually be detected by visual inspection of rose diagrams. It is seldom necessary to resort to statistical tests. The interpretation of such differences will clearly vary from case to case. In the rock record it may be possible to identify differences in wave and current directions or differences in movement pattern of sinuous-crested dunes (trough cross bedding) and straight-crested dunes and bars (tabular sets). A feature of some inter-tidal areas, of the beds of rivers with a high discharge range, and of large aeolian sand seas is that larger structures (typically dunes) tend to reflect the high stage or peak flows and are quite narrowly dispersed. Associated smaller structures (typically ripples), formed during falling stage have a much broader spread of directions, reflecting the tendency for those flows to be diverted around larger emerging bedforms.

Palaeocurrent restoration presentation and analysis exercise

In order familiarize yourself with the methods presented in this appendix, try the following exercise. Using the data in Table A1.1, re-orientate the cross-bedding and wave-ripple crestline measurements to their position before tectonic tilting. Plot the restored data as two rose diagrams, one showing the palaeocurrent direction indicated by the cross-bedding and one showing the trend of the wave-ripple crestlines. For the cross bedding, describe the spread of data; is the distribution unimodal, bimodal or polymodal? Calculate the vector mean using equation A1.1 and the dispersion of the data (vector strength) using equation A1.2.

Table A1.1 Dip and dip direction of cross-bedded foresets, and plunge and lineation of wave-ripple crestlines observed on a bedding plane that has been tectonically tilted (50/330 SW). Use the procedures outlined in Appendix 1 to re-orientate the cross-bedding and the ripple crests to their pre-tectonic orientations.

Dip and dip direction of cross bedding	Plunge/ lineation wave-ripple crestlines
70/320 SW	25→309
28/314 SW	27→310
67/317 SW	23→312
72/324 SW	25→308
30/316 SW	22→309
74/318 SW	24→305
68/322 SW	28→316
26/312 SW	26→315
32/317 SW	23→310
30/310 SW	22→309

APPENDIX 2

Sampling and preserving unconsolidated sediments

The collection and preservation of sedimentary structures from unconsolidated sediments for further study in the laboratory require special techniques. These involve the artificial consolidation of the sediment and often cause the lamination and bedding to be made more apparent. There are two main ways of doing this: by taking box cores and making lacquer peels. Some ideas on doing this are set out below but it is often possible to improvise if purpose-made equipment is not readily available.

Box cores

To take box cores, simple metal or plastic boxes are pushed into the sediment and then removed carefully to retrieve a relatively undisturbed sample. This sample can then be impregnated with glue or resin, either directly in the field or later in the laboratory. If the impregnating glue or resin is distributed evenly, it will penetrate to different depths according to slight differences in porosity and permeability between individual layers and laminae.

The simplest corer is the so-called “Senckenberg box”, a rectangular-shaped box with a removable sliding door panel on one side. On present-day surfaces it is pushed vertically into the sediment and then dug out after insertion of the cover. On a vertical face of a pit or trench it is pushed in horizontally in an upright position. The cover is then slid into place vertically after slight excavation of the top of the box.

A more complex, and slightly more difficult corer to use is the tapering “Reineck box”. This is valuable in shallow water or where the water table is too high to permit the use of a Senckenberg box. The corer is pushed vertically into the sediment and is followed by the cover. The flanges of the box and the grooves in the side of the cover hold the two parts of the corer together, but sediment can obstruct sliding of the flange. The box and the cover are then pulled vertically out of the sediment giving a downwards-tapering wedge-shaped core that can later be impregnated with resin, following careful removal of the cover After the resin has hardened, the sample can be further strengthened by gluing a sheet of hardboard or thick cardboard to the exposed surface. When this has set, the sample may be freed from the box, if necessary by cutting around the margins of the box. Any loose sediment can be removed from the newly exposed surface by gentle brushing or blowing. This should be done several times as the sample dries out. If permeability differences are present between laminae, the internal lamination should be picked out in relief.

Some glues are soluble in various solvents and this can be useful if, for example, you wish to investigate the grain-size distribution of particular laminae. Cutting out the laminae from the box core and dissolving the glue with a suitable solvent can give loose grains suitable for sieving and other grain size measurement. Be aware, however, that organic solvents can present health hazards and appropriate precautions should be taken.

Lacquer peels

Lacquer peels can be taken from the walls or floors of trenches. The surface should be carefully scraped flat and then sprayed, using a garden spray, with a dilute solution of an appropriate resin. Lacquers which use volatile organic solvents such as acetone were often used and, in such cases, the surface was ignited with appropriate precautions, after spraying, causing the sediment to dry out and the lacquer to penetrate more deeply. Epoxy resins are now more commonly used. The result is to cement and harden a surface layer. However, in order that the layer can be removed, it must be reinforced. This is done by carefully plastering several layers of resin-soaked bandage or gauze on to the surface. When the resin is thoroughly cured, the peel can be carefully removed, often with the support of a rigid board. Loose sediment can then be removed from the exposed surface and the surface fixed by further spraying. Peels have the advantage over box cores of allowing the sampling of larger areas and being lighter to carry. This preparation, however, makes rather bigger demands on field time.

APPENDIX 3

Methods for studying present-day environments

Many types of observation can be made and many methods for recording data can be applied on present-day sediment surfaces. In order to understand and characterize a tidal flat, a beach or an exposed river bed, for example, it can be useful first of all form a quick-look overall impression and then to carry out a systematic survey of a selected area that is thought to be "typical" or "representative". In some cases a single traverse will be appropriate, whereas elsewhere more detailed mapping might be called for. Generally the features to be recorded and mapped are predetermined or self-evident and the main problems relate to navigation and positioning, particularly on extensive, low and somewhat featureless areas such as tidal flats or wide beaches. These problems can sometimes be overcome through the use of a drone, either to take a series of aerial photographs at the outset of the study, or to use interactively during the study. Such a survey should be done with appropriate permission and following relevant regulations.

When working on inter-tidal areas, it is best, when possible to work during the falling tide. Not only are sedimentary features fresher but it is also safer. If working during a rising tide, make sure that you understand the way in which the tide floods and make one person responsible purely for safety, to the exclusion of participating in the field observations. Always allow a generous margin for safety and take local advice in unfamiliar areas. Never work alone in inter-tidal areas.

If observations are to be made along a straight-line traverse, two sighting posts placed some distance apart at one end of the traverse, and in line with it, are a great help. By keeping them in line it is possible to steer an accurate straight-line course on foot or by boat. With the advent and increasing availability of affordable Global Positioning System (GPS) receivers and drones with cameras, establishing position along a traverse is no longer as difficult in featureless terrain. However, if one has a topographic map at an appropriate scale, positioning should still be possible without a GPS. Compass bearings on nearby features of known position off the line of section can provide good fixes on long traverses. Measurement by tape or range-finder may be used over shorter distances or for more detailed work.

Mapping an area presents more complex problems. On a small scale it may be possible to mark out a measured grid; on a larger scale a series of cross-cutting traverse lines can be established by marker posts around the edge of the mapped area, like those set up for single traverses. A hand-held GPS receiver or a drone are both useful for establishing position. In the absence of such devices, it will be necessary to take bearings or other angular measurements on surrounding fixed points. A sextant is an accurate and efficient way of doing this. Two angles measured between any three fixed points establish position quite accurately.

When surveying on foot, remember that surface features on loose sediment are easily ruined by footprints and so photographs should be taken at an early stage, whether on foot or from a drone. For the same reason try to use a few strategic pathways. When working from boats, problems of disturbance are less acute, but observing the sediment surface can present problems. In shallow and reasonably clear water, a glass-bottomed box is very useful and polarizing sunglasses can help to reduce reflection. In deeper or turbid water, indirect methods of observation such as echo-sounding become essential.

Descriptions of sediment surfaces can be made at various levels of detail from the qualitative description of the type of bedform to detailed measurement of dimensions, orientations and distribution densities of particular structures. Systematic recording is often helped by a data sheet, which can be completed at each locality. Setting up an appropriate data sheet may necessitate a preliminary reconnaissance visit before the main study. An example for a tidal-flat setting is shown in Figure A3.1.

Presentation of directional data is discussed in Appendix 1. The resulting rose diagrams etc. can be shown on maps or profiles in a variety of ways. Rose diagrams can be superimposed on maps. Maps can be contoured for parameters such as height and spacing of bedforms, pebble size or distribution density of burrows. In addition, qualitative features such as types of organisms, the plan-view shape of bedforms or the superimposition of different types of bedforms may be displayed on maps.

Examination of internal structures of modern sediments can be achieved by digging trenches or by taking shallow

Locality:	**Sample study locality:**	**Map reference:**
Sub-environment number and type:		
Survey map no.	Plane table map no.	Sketch map no.
Scaled drawing nos.	Film nos.	Photograph nos.
Sample bag nos.	Water sample nos.	Notebook page nos.

Feature	**Observations at sample locality**	**Observations in adjacent areas**
Sediment types and conditions (1) Grain size (mm or phi) (2) Mineral composition (types by %) (3) Colour (Munsell or Goddard scale) (4) Fabric - texture (5) pH (6) Eh - depth to reducing layer (7) Salinity (8) Temperature - air : substrate	mean mode(s) orientation water sample	mean mode(s) orientation water sample
Primary structures - bedforms (1) Surface: spacing (wavelength), amplitude or height, ripple index, symmetry index, parallelism index 1, parallelism index 2, straightness index, continuity index, orientation (crest), inclination (lee), crestline shape, degree of bifurcation (2) Interior: foreset dip, foreset azimuth, drapes (3) Secondary and other structures	1 2 3 4 5 photo/sketch no.	1 2 3 4 5 photo/sketch no.
Life forms (1) Permanent forms - life assemblage (a) flora (b) epifauna (c) infauna: macrofauna, meiofauna (0.5-0.05mm), microfauna (2) temporary forms - life assemblage (a) marine (b) terrestrial obligate (dependent on intertidal zone) faculative (not dependent on intertidal zone) (3) Drifted forms - death assemblage	Types: depth, frequency, orientation, planktonic, nektonic, benthonic (vagrant, vagile, sessile, gregarious, solitary, epibiotic), attached, free, burrowing, boring, herbivore, carnivore, omnivore, deposit feeder, collector, swallower, suspension feeder, filterer, awaiter, grazer, scavenger. photo/sketch no.	photo/sketch no.
(4) Ethological relationships predator - prey, symbiosis etc food chain, food web		
(5) Ichnological observations morphological preservational behavioural	Track, trail, radial trace, tunnel, spreite, pouch, relief, epichnia, endichnia, hypichnia, exichnia, resting, dwelling, feeding, grazing, crawling, locomotion, escape, readjustment, orientation.	
General comments		

Figure A3.1 Example of a sample data sheet for the systematic recording of observations from an environment such as a tidal flat.

cores (Appendix 2). Allowing carefully cleaned sides of trenches to dry will often highlight lamination in more detail than on a freshly cut surface. When time is short or the water table is too high for trenching, cores of considerable length can be obtained by pushing boxes or tubes into the sediment. When taking cores be careful to record their orientation. In the laboratory, cores can be impregnated with resin to preserve them permanently and to show structures more clearly. Procedures for collecting and impregnating shallow cores are given in Appendix 2.

APPENDIX 4

Techniques for the study of trace fossils

The study of trace fossils requires one to try to relate fragmentary, usually two-dimensional patterns to complex three-dimensional records of behaviour left by a diverse range of organisms. Although a wide range of techniques has been developed, we concentrate here on cheaper, simpler techniques, which rapidly enlarge experience.

Observation and recording of trace fossils in the field and in the laboratory

In present-day sub-aerial and intertidal environments direct and "after-the-event" observation is possible. In sub-aqueous settings observation is more costly as diving equipment and expertise and/or underwater cameras are needed. Estuaries provide accessible locations for a variety of case studies but bear in mind the safety issues highlighted in Appendix 3.

In dealing with trace fossils in rocks, the drawing of scaled field diagrams and the photographing of traces may be helped by outlining inconspicuous features with chalk (not permanent ink). Burrows, along with other types of poorly defined lamination, may be accentuated by wetting a rock surface with water, glycerine, paraffin or light mineral oil (whereupon uptake of stain is controlled by differences in porosity). Structures with fine detail may be whitened with powdered chalk or ammonium chloride, and photographed in strong oblique light.

Many simple techniques can be applied to the study of the distribution of trace fossils (and indeed other sedimentary structures) on exposed bedding surfaces. These techniques can also be used for the study of animal traces on modern sediment surfaces, such as on beaches. Commonly applied techniques include plane-tabling, aligning transects, siting quadrat surveys, sampling of different sub-environments for assessment of changes in the abundance and/or diversity of forms, and to record types of activity, photographing evidence to scale, orientating data, drawing scaled diagrams and collecting and curating samples. Other useful techniques that can be applied especially well in modern sedimentary environments include the taking of box cores, vertical and horizontal peels using lacquer, polyester resin and epoxy resin, and the casting of burrows, both sub-aerially and underwater (Appendix 2).

Whatever the environment, it is important to define a problem and plan an appropriate programme of sampling, description and analysis. Graphic logs of sections should include data on occurrence and distribution of trace fossils in relation to other sedimentary features (Fig. 10.6b).

Methods for enhancing the visibility of structures

In the laboratory the following procedures may be appropriate, depending of facilities and the aims of the study. They may help to reveal traces of structures where none appears to exist. Some apparently massive beds have intense bioturbation, i.e. maximum rather than minimum organic activity. This, supplemented by diagenetic effects, may enhance their apparent homogeneity.

- The making of peels from box cores.
- Staining of fine-grained carbonate and clay mineral-rich rocks by organic dyes such as alizarin red, methylene blue, or Indian ink.
- Making acetate peels by polishing a cut surface and etching it with acid, then applying acetone and covering this with an acetate sheet which, when adherent, can be peeled off.
- Subjecting 1 cm-thick sawn blocks of sedimentary rocks, whether naturally cemented or impregnated, to X-radiography or infrared and ultraviolet photography. Ultraviolet photography is best applied to limestones that contain little iron.
- Infrared photography is cheap in that it only requires a special film and filter, although the cutting of thinner slabs (0.5 cm), which gives the best results, is difficult. Exposure time should be proportional to the organic content of the rocks, arenaceous ones being more transparent than argillaceous ones.
- Artificial weathering of apparently homogeneous rocks for a short period using sand-blasting equipment with an abrasive of unsorted sand slightly finer than the grain size of the rock.

- Making thin sections of impregnated sediment or rock. These should be made larger (about 5 x 5 cm) and slightly thicker (0.04 mm) than normal, whereupon they can be mounted in a slide projector or scanned into a computer. Thin sections may be stained to good effect (see above).

Experimental approaches to understanding the behavioural aspects of trace fossils

This approach involves the study in the field or the laboratory of the factors that influence the behaviour of organisms and the form of the resulting traces. Such an approach is mainly concerned with invertebrates rather than vertebrates or plants. Studies commonly focus on burrowing organisms, commonly bivalves, and the way in which they destroy primary sedimentary structures and form biogenic structures. Studies may vary from the simple observation of the marks made by moving organisms on the sediment surface, or the burrowing of given organisms placed upon a carefully prepared succession of particular composition and consistency (e.g. carefully prepared alternating layers of sand and mud), to ones which try to relate functional morphology of the animal to its behaviour and to its burrow. More complex studies can try to more closely match natural conditions and describe both the burrowing behaviour, its effect on the substrate, and the interaction with processes of erosion and sedimentation. See, for example, Bromley (1996) and Goldring (1999).

APPENDIX 5

Techniques for sedimentary logging

Sedimentary logs that give a bed-by-bed graphical depiction of the various lithologies and structures encountered within a succession are one of the primary means that sedimentologists use to record and communicate sedimentary data. Although there are many differing styles of sedimentary log (e.g. Fig. 10.2), each with their own relative merits, we offer here some general advice about how to represent a sedimentary succession in log form. A sedimentary log template is depicted in Figure A5.1, copies of which can be used in the field.

Before starting the logging exercise

1. Perform a reconnaissance of the outcrop to be logged in order to identify the younging direction of the succession and the lowest and highest points in the stratigraphy to be studied.
2. Identify which part of the outcrop will be logged. Suitable sections need to be both sufficiently well exposed to be able to generate a reasonably continuous log and sufficiently accessible. Ideally, try to pick a location where a single continuous log can be made through the study section. However, where this is not possible, be prepared to construct several overlapping logs that are laterally offset from each other in order to construct a composite log through the complete study section. Good logging sites include gullies and ravines, dry stream beds, stepped hillsides, coastal cliffs and wave-cut platforms. In regions where the beds have been tectonically tilted, good log sections can often be constructed by traversing laterally along the base of cliff lines etc.
3. Decide on a scale for the logging exercise. This will be dictated by factors such as the complexity of the stratigraphy, the scale and/or thickness of the bedding, the quality of the outcrop, the thickness of the section to be logged, the time available for the exercise and the overall aims of the project.
4. Decide how many log sections you are likely to need in order to reasonably characterize the study section. One log may suffice for simple successions with little lateral variability, whereas, more detailed studies of laterally complex and variable successions will need multiple logs.
5. In starting the logging exercise, try and choose a prominent bed as a start point and accurately record its geographic position and, where possible and appropriate, its elevation above sea level.

The logging exercise

1. The thickness recorded for each bed should be the true bed thickness. This is not necessarily the same as the exposed bed thickness, especially when logging on a hill-slope or when the beds have been tectonically tilted.
2. The amount of detail that you should include on your log will be dictated by the scale at which you are logging. For detailed logs, attempt to include individual beds down to 5-10cm, whereas for broader-scale logs, it may be sufficient to group sets of similar beds together and record them as a single unit. If appropriate, schematically sketch in any finer-scale details, such as laminae, between the major bed boundaries.
3. If beds have irregular bounding surfaces, these should be recorded graphically on the log section. For example, erosive, channel bases should be drawn cutting down into the underlying unit and lens-shaped bodies should be drawn tapering at their ends. For each bed, it is important to record its thickness at the point where you are logging, but with a note recording if the bed clearly changes thickness when traced laterally.
4. Pay careful attention to the grain size both within a single bed and between adjacent beds. Carry a grain size card and a hand lens and use them frequently. Look out for normally or inversely graded beds. Subtle grain-size changes between beds can be important in identifying gradual fining-up or coarsening-up successions over thicknesses of tens of metres, which may indicate something about gradual temporal changes in the energy regime during deposition.
5. For each bed look carefully for sedimentary structures, both in section and on exposed bedding surfaces,

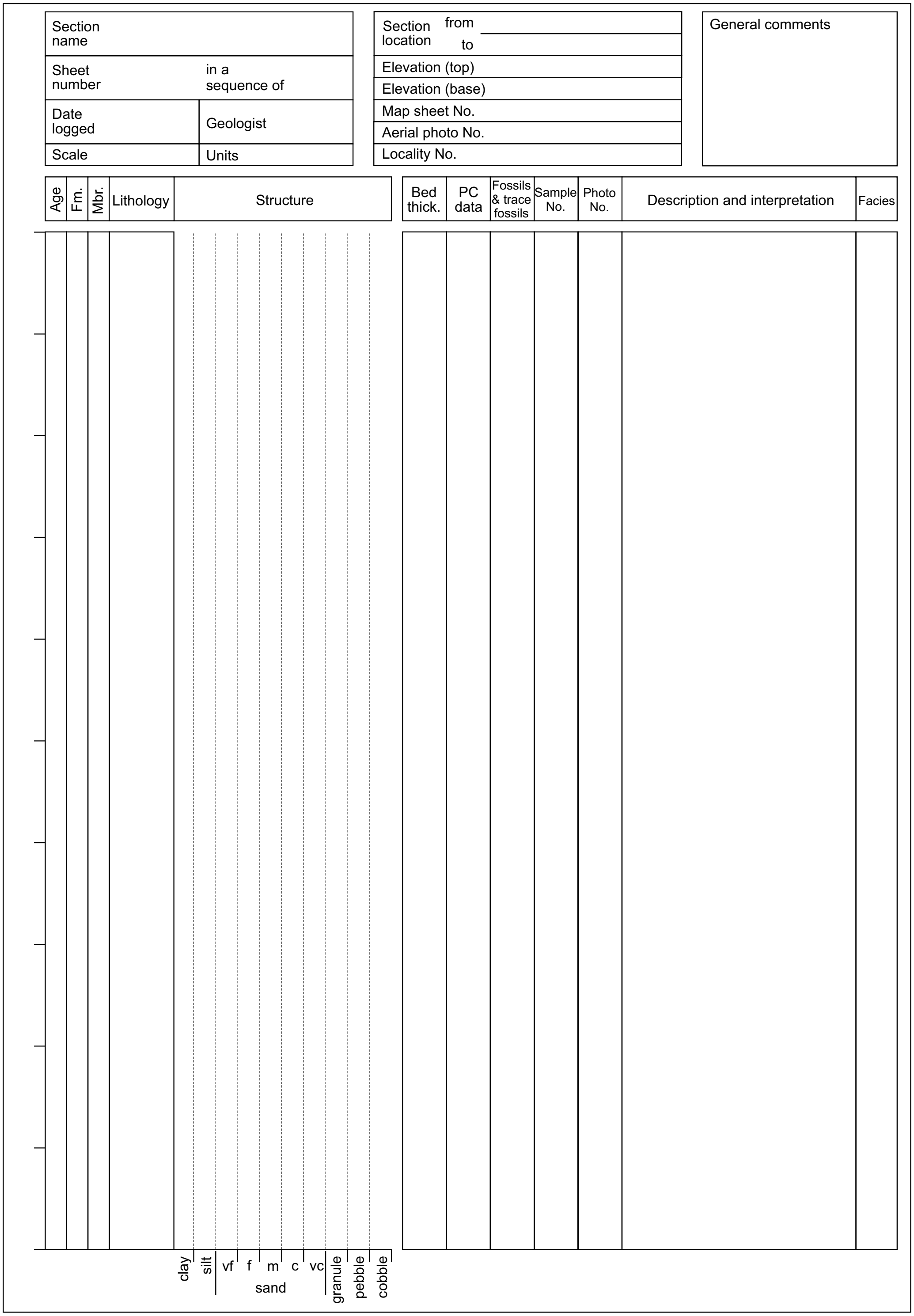


Figure A5.1 Example of a logging sheet for the recording of vertical sequences in sedimentary rock successions.

bearing in mind that some may be preserved on the undersides of beds. Adopt a systematic search approach for each bed. Where structures are evident they should be included graphically on the log using a standard set of symbols (Appendix 6). Additionally, record them in as much detail as possible, taking measurements, photographs and making sketches if necessary, especially if you are uncertain of their origin. Pay particular attention to fossils and trace fossils as these can be useful palaeoenvironmental indicators.

6. Make additional notes where structures can be used to identify way-up or palaeocurrent direction. Record palaeocurrent data in a separate column either as dip magnitude and direction (azimuth) for planar data such as cross bedding, or as plunge and plunge direction for linear features such as groove marks. In tectonically deformed successions, you should also record the dip and strike of the overall bedding so that the palaeocurrent data can be restored to true directions (Appendix 1).
7. Make separate notes alongside the log describing other potentially significant features, perhaps including provisional inferences about the nature of erosional or depositional processes. Indeed, as the log is being constructed you may develop ideas about possible environments of deposition. However, whilst working ideas are useful, it is important not to become too focussed on any one conjecture during the logging phase as this may bias the way in which you make your observations.
8. In finishing the logging exercise, try and choose a prominent bed as an end point and accurately record its geographic position and, where possible, its elevation above sea level.

APPENDIX 6

Key to common sedimentary lithologies and structures

When constructing sedimentary logs or panels, most sedimentologists augment their diagrams with symbols that represent the various lithologies and sedimentary structures encountered. Although there is no formal scheme for depicting these features, the symbols used to represent some of the more common lithologies and structures have become largely standardized and an example set of commonly used symbols is depicted in Figure A6.1. These symbols can be adapted to suit the sediments or rocks being investigated; additional symbols should be devised to represent those features that are not listed here. It is important when presenting graphical sedimentary data in the form of logs or panels always to include a full explanatory key to all the symbols used. Graphic symbols should be qualified on the log, where necessary, with written descriptive notes and preliminary interpretations of process and/or environment of deposition.

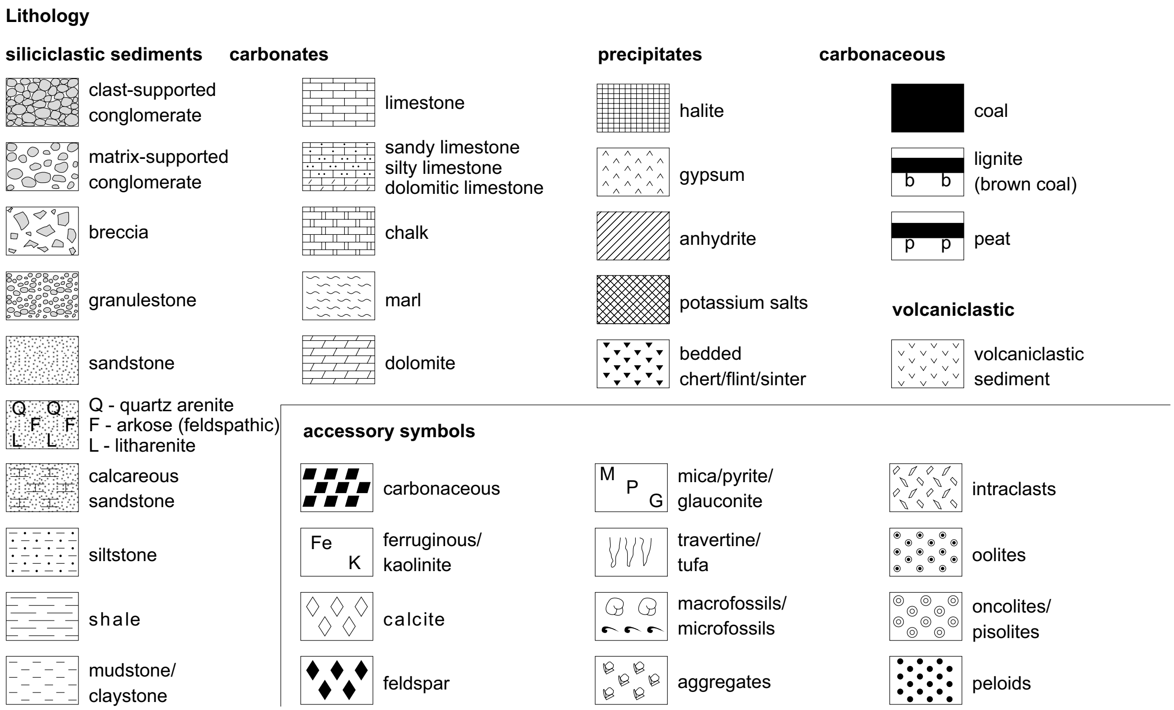


Figure A6.1 Scheme for the graphic depiction of lithologies, sedimentary structures, fossils and trace fossils in sedimentary logs and panels.

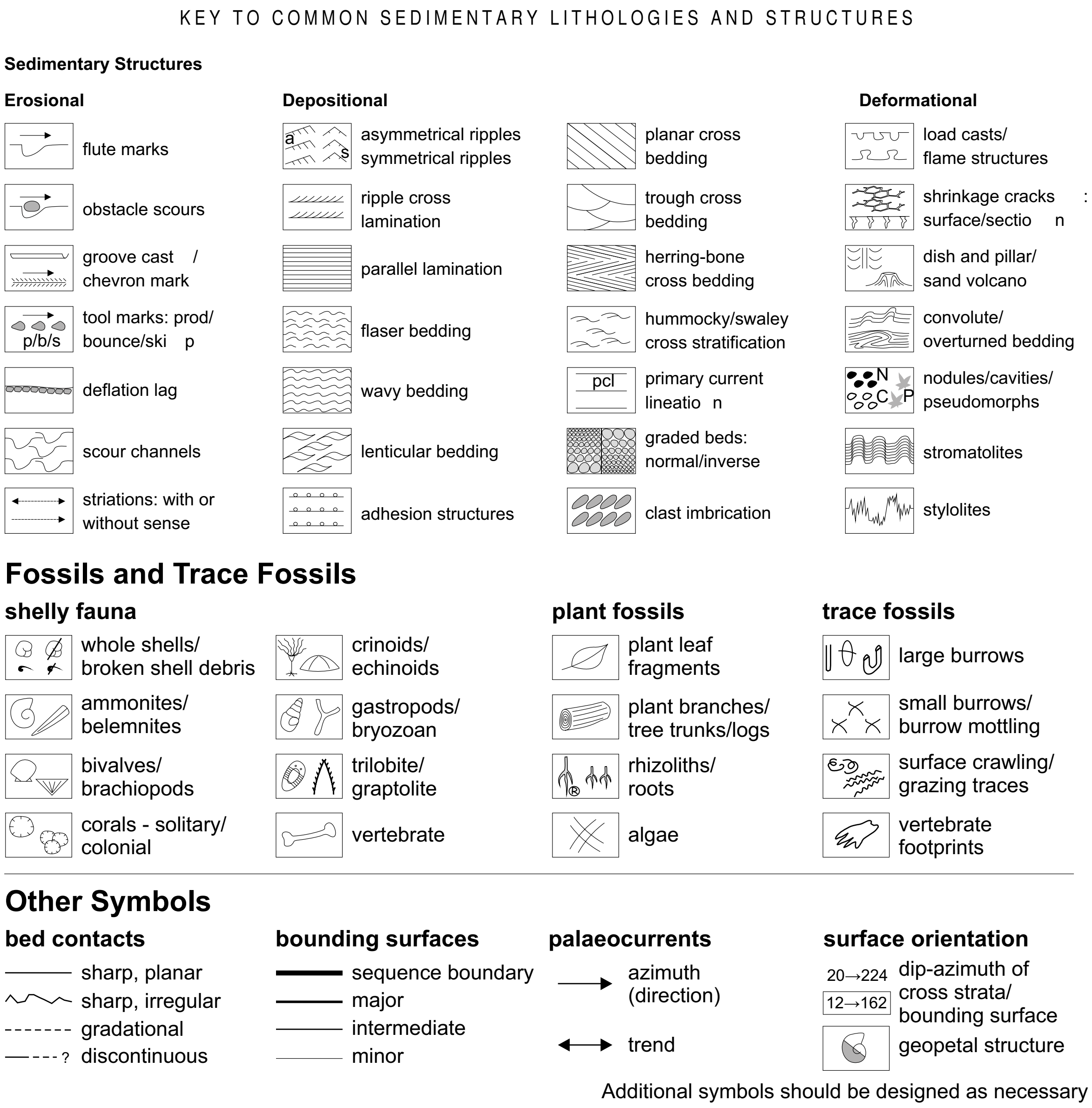


Figure A6.1 *Continued*

Bibliography

Adams, A.E., MacKenzie, W.S. and Guilford, C., 1984. *Atlas of sedimentary rocks under the microscope.* Longman.

Ager, D.V., 1981. *The nature of the stratigraphical record.* (2nd edn) McMillan Press.

Alexander J. & Fielding C. 1997. Gravel antidunes in the tropical Burdekin River, Queensland, Australia. *Sedimentology* **44**, 327-37. (7)

Alexander J., Bridge J.S., Cheel R.J. & Leclair S.F. 2001. Bedforms and associated sedimentary structures formed under supercritical water flows over aggrading sand beds. *Sedimentology* **48**, 133-52. (6)

Allen, J. R. L. 1964. Studies in fluviatile sedimentation: six cyclothems from the Old Red Sandstone. *Sedimentology* **3**, 163–98. (4, 10)

Allen, J. R. L. 1968. *Current ripples: their relations to patterns of water and sediment motion*. North Holland. (3, 6)

Allen, J. R. L. 1970. *Physical processes of sedimentation*. Allen & Unwin. (3)

Allen, J.R.L., 1977. Physical processes of sedimentation, an introduction. George, Allen and Unwin, Earth Science Series, 1, (4th Impression with Revised Readings)

Allen, J. R. L. 1971. Transverse erosional marks of mud and rock: their physical basis and geological significance. *Sedimentary Geology* **5**, 167-384.

Allen, J. R. L. 1977. The possible mechanics of convolute lamination in graded sand beds. *Journal of the Geological Society of London,* **134**, 19–31. (9)

Allen, J. R. L. 1980. Sand waves: A model of origin and internal structure. *Sedimentary Geology* **26**, 281-328.

Allen, J. R. L. 1982. *Sedimentary structures: their character and physical basis*. Developments in Sedimentology **30**. Elsevier. (3,4,6,7)

Allen, J. R. L. 1985a. *Experiments in physical sedimentology*. Allen & Unwin. (3)

Allen, J.R.L. 1985b. *Principles of physical sedimentology*. Allen & Unwin. (3)

Allen, J. R. L. 1991. The Bouma Division-A and the possible duration of turbidity currents. *Journal of Sedimentary Petrology* **61**, 291-95. (6)

Allen, J. R. L. & Banks N. L. 1972. An interpretation and analysis of recumbent-folded deformed cross-bedding. *Sedimentology* **19**, 257–83. (9)

Allen, P.A. 1997. *Earth Surface Processes*. Blackwell Science (3)

Allen, P.A. and Allen, J.R., 2005. *Basin analysis. Principles and applications*. (2nd edn) Blackwell Science

Amy, L., McCaffrey, W.D. and Talling, P. (Eds.), 2009. *Sediment gravity flows – recent insights into their dynamic and stratified/composite nature.* Marine and Petroleum Geology, Special Thematic Set, **26,** 1897-2043.

Anderson B.G. & Droser M.L. 1998. Ichnofabrics and geometric configurations of Ophiomorpha within a sequence stratigraphic framework: an example from the Upper Cretaceous US western interior. *Sedimentology* **45**, 379-96. (9)

Arnott R.W. & Southard J.B. 1990 Exploratory flow-duct experiments on combined-flow bed configurations and some implications for interpreting storm-event stratification. *Journal of Sedimentary Petrology* **60**, 211-19. (6)

Arthurton, R. S. 1973. Experimentally produced halite compared with Triassic layered halite-rock from Cheshire, England. *Sedimentology* **20**, 145–60. (8)

Ashley, G. M., Southard J. B. & Boothroyd J. C. 1982. Deposition of climbing ripple beds: a flume simulation. *Sedimentology* **29**, 67–79. (6)

Ashworth, P.J., Bennett, S.J., Best, J.L. and McLelland, S.J. (Eds.), 1996. *Coherent flow structures in open channels.* John Wiley and Sons.

Ashworth, P.J., Best, J.L. and Parsons, D.R., 2015. *Fluvial-tidal sedimentology*. Developments in Sedimentology, 68, Elsevier,

Assereto, R. L. A. M. & Kendall C. G. 1977. Nature, origin and classification of peritidal tepee structures and related breccias. *Sedimentology* **24**, 153–210. (9)

Audemard, F. A. & Santis F. de, 1991. Survey of liquefaction structures induced by recent moderate earthquakes. *Bulletin of the International Association of Engineering Geologists*, **44**, 5-16. (9)

Baas J.H. 1999. An empirical model for the development and equilibrium morphology of current ripples in fine sand. *Sedimentology* **46**, 123-38. (6)

Bagnold, R. A. 1941. *The physics of blown sand and desert dunes*. Methuen. (6)

Basan, P. B. (ed.) 1978. *Trace fossil concepts*. Short-course Notes **5**, Society of Economic Paleontologists and Mineralogists (9)

Bathurst, R. G. C. 1975. *Carbonate sediments and their diagenesis*, (2nd edn).: Elsevier. (8)

Best J.L. 1996. The fluid dynamics of small-scale alluvial bedforms. In *Advances in fluvial dynamics and stratigraphy.* P. A. Carling & M. R. Dawson (eds) 67-125, John Wiley & Sons. (6)

Best J.L., Ashworth P.J., Bristow C.S. & Roden J. 2003 Three-dimensional sedimentary architecture of a large, mid-channel sand braid bar, Jamuna River, Bangladesh. *Journal of Sedimentary Research* **73**, 516-30. (6)

Blatt, H., Middleton G. V. & Murray R. C. 1980. *Origin of sedimentary rocks*. (2nd edn). Prentice-Hall. (1,2,3,4,5,6,7,8,9)

Bluck, B. J. 1967. Sedimentation of beach gravels: examples from South Wales. *Journal of Sedimentary Petrology* **27**, 128–56. (7)

Boersma, J. R. 1970. Distinguishing features of wave-ripple cross-stratification and morphology. Doctoral thesis, University of Utrecht, The Netherlands, 65 pages. (Fig. 6.13)

Boggs, S. Jr. 1987. *Principles of sedimentology and stratigraphy*. Merrill.

Boggs, S. Jr. 2009. *Petrology of sedimentary rocks.* Cambridge University Press,

Boggs, S. Jr. 2016. *Principles of sedimentology and stratigraphy*. (5th edn), Pearson.

Bouma, A. H. 1962. *Sedimentology of some flysch deposits*. Elsevier. (4, 6)

Branney, M.J. and Kokelaar, P., 2002. *Pyroclastic density currents and the sedimentation of ignimbrites.* Geological Society of London, Memoir, **27**.

Brenchley, P. J. & Williams B. J. P. (eds) 1985. *Sedimentology: recent developments and applied aspects*. Geological Society of London Special Publication **18**, (2,10)

Bridge, J.S. 2003. *Rivers and floodplains: forms, processes and sedimentary record*. Blackwell Science. (3, 6, 10)

Bridge J. & Best J. 1997 Preservation of planar laminae due to migration of low-relief bed waves over aggrading upper-stage plane beds: comparison of experimental data with theory. *Sedimentology* **44**, 221-51. (6)

Bridge, J.S. and Demicco, R.V., 2008. *Earth surface processes, landforms and sediment deposits.* Cambridge University Press.

Bristow, C.S., Bailey S.D. & Lancaster N., (2000) The sedimentary structure of linear sand dunes. *Nature* **406**, 56-59. (6)

Broadhurst, F. M. & Simpson I. M. 1967. Sedimentary infilling of fossils and cavities in limestone at Treak Cliff, Derbyshire. *Geological Magazine* **104**, 443–8. (2,9)

Bromley, R.G., 1990. *Trace fossils. biology and taphonomy. Special Topics in Palaeontology*. Unwin Hyman.

Bromley, R.G., 1996. *Trace fossils: biology and taphonomy*. (2nd edn). Chapman and Hall. (9)

Bromley, R. G. and Ekdale A. A. 1986. Composite ichnofabrics and tiering of burrows. *Geological Magazine* **123**, 59–65. (9)

Brookfield, M. E. 1977. The origin of bounding surfaces in ancient aeolian sandstones. *Sedimentology* **24**, 303–32. (6)

Brookfield, M. E. 1992. Eolian Systems. In *Facies models: Response to Sea Level Change*, (3rd edn), R. G. Walker & N. P. James (eds), 143-156. Geological Association of Canada. (6)

Brookfield, M.E. 2003. *Principles of stratigraphy*. Blackwell Science. (10)

Brookfield, M. E. & Ahlbrandt T. S. 1983. *Eolian sediments and processes*. Elsevier. (6)

Buck, S. G. 1985. Sand-flow cross-strata in tidal sands of the Lower Greensand (Early Cretaceous), Southern England. *Journal of Sedimentary Petrology* **55**, 895–906. (6)

Busson, G., 1980. *Evaporites. Illustration and interpretation of some environmental sequences.* Éditions Technip.

Byers, C. W. 1974. Shale fissility: relation to bioturbation. *Sedimentology* **21**, 479–84. (5, 9)

Campbell, C. V. 1967. Laminae, lamina set, bed and bedset. *Sedimentology* **8**, 7–26. (2)

Cant, D. J. & Walker R. G. 1978. Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. *Sedimentology* **25**, 625–48 (10)

Carling P.A. 1996. Morphology, sedimentology and palaeohydraulic significance of large gravel dunes, Altai Mountains, Siberia. *Sedimentology* **43**, 647-64. (6, 7)

Cartigny, M.J.B., Postma, G., van den Berg, J.H. & Mastbergen, D.R., 2011. A comparative study of sediment waves and cyclic steps based on geometries, internal structures and numerical modelling. *Marine Geology,* **280**, 40-56. (6)

Cas, R. A. F. & Wright J. V. 1987. *Volcanic successions, modern and ancient*. Allen & Unwin. (7)

Chamberlain, C. K. 1978. Recognition of trace fossils in core. *Society of Economic Paleontologists and Mineralogists, Short Course No.* **5**, 119-66. (9)

Cheel, R. J. & Rust B. R. 1986. A sequence of soft-sediment deformation (dewatering) structures in Late Quaternary subaqueous outwash near Ottawa, Canada. *Sedimentary Geology* **47**, 77-93. (9)

Chowns, T. M. & Elkins J. E. 1974. The origin of quartz geodes and cauliflower cherts through the silicification of anhydrite modules. *Journal of Sedimentary Petrology* **44**, 885–903. (9)

Clari, G. & Ghibaudo G. 1979. Multiple slump scars in the Tortonian type area (Piedmont Basin, northwestern Italy). *Sedimentology* **26**, 719–30. (9)

Clemmensen, L. B. & Abrahamsen K. 1983. Aeolian stratification and facies association in desert sediments, Arran Basin (Permian), Scotland. *Sedimentology* **30**, 311–39.(6)

Clifton, H. E., Hunter R. E. & Phillips R. L. 1971. Depositional structures and processes in the non-barred, high-energy nearshore. *Journal of Sedimentary Petrology* **41**, 651–70. (6,10)

Coe, A.L. (ed.), 2010. *Geological field techniques.* Wiley-Blackwell.

Coe, A.L., 2003. *The Sedimentary record of sea-level change.* The Open University*,* Cambridge University Press,

Coleman, J. M. & Wright L. D. 1975. Modern river deltas: variability of processes and sand bodies. In *Deltas: models for exploration*, M. L. Broussard (ed.), 99–149. Houston Geol Soc. (Fig. 10.2)

Collinson, J. D. 1968. Deltaic sedimentation units in the Upper Carboniferous of northern England. *Sedimentology* **10**, 233–54. (6)

Collinson, J. D. 1970. Bedforms of the Tana River, Norway. *Geografiska Annaler* **52A**, 31–56. (6)

Collinson, J. D. & Lewin J. (eds) 1983. *Modern and ancient fluvial systems*. Special Publication **6**, International Association of Sedimentologists. (6, 7, 10)

Collinson, J.D. 1994. Sedimentary deformational structures. In: *The Geological Deformation of Sediments.* A. Maltman (ed). 95-125, Chapman and Hall. (9)

Conybeare, C. E. B. & Crook K. A. W. 1968. Manual of sedimentary structures. *Australian Dept. Nat. Dev. Bur. Mines Resources, Geol. & Geophys. Bull.* **102**. (2 ,4, 6, 7, 9)

Cooke, R., Warren A. & Goudie A. (1993). *Desert geomorphology*. UCL Press, (4, 6)

Covault, J.A., Kostic, S., Paull, C.K., Ryan, H.F. & Fildani, A., 2014. Submarine channel initiation, filling and maintenance from sea-floor geomorphology and morphodynamic modelling of cyclic steps. *Sedimentology*, **61**, 1031-1054.(6)

Crabaugh, M. & Kocurek G., (1993) Entrada Sandstone: An example of a wet aeolian system. In: *The Dynamics and environmental context of aeolian sedimentary systems*, K. Pye (ed.), 103-126. *Special Publication, Geological Society of London,* **72**. (6)

Crimes, T. P. & Harper J. C. (eds) 1970. *Trace fossils*. Special Issue **6**, *Geological Journal*, Seel House Press. (9)

Crimes, T. P. & Harper J. C. (eds) 1977. *Trace fossils 2*. Special Issue **9**, *Geological Journal*, Seel House Press. (9)

Crowe, B. M. & Fisher R. V. 1973. Sedimentary structures in base-surge deposits with special references to cross-bedding: Ubehebe craters, Death Valley, California. *Bulletin of the Geological Society of America,* **84**, 663–82. (7)

Curran, H. A. (ed.) 1985. *Biogenic structures: their use in interpreting depositional environments*. Society of Economic Paleontologists and Mineralogists, Special Publication **35**, Tulsa: (9)

Dalrymple, R. W. 1984. Morphology and internal structure of sandwaves in the Bay of Fundy. *Sedimentology* **31**, 365–82. (6)

Dalrymple, R. W., Makino Y. & Zaitlin B. A. 1991. Temporal and spatial patterns of rhythmite deposition on mud-flats in the macrotidal, Cobequid Bay-Salmon River estuary, Bay of Fundy, Canada. In: *Clastic Tidal Sedimentology,* D.G. Smith, G.E. Reinson B.A., Zaitlin & R.A. Rahmani (eds). *Canadian Society of Petroleum Geologists Memoir,* **16**, 137-160. (6)

Davies, I. C. & Walker R. G. 1974. Transport and deposition of re-sedimented conglomerates: the Cap Enrage Formation, Cambro-Ordovician, Gasp, Quebec. *Journal of Sedimentary Petrology* **44**, 1200–1216. (7)

Davies, J.C., Davis, P.K., and Davis, H. 2002. Statistics and data analysis in geology. (3rd edn) John Wiley and Sons.

De Boer, P. L., van Gelder A. & Nio S. D. 1988. *Tide-influenced sedimentary environments and facies (Sedimentology and petroleum geology)*. D. Reidel. (6)

DeCelles, P. G., Langford R. P. & Schwartz R. K., 1983. Two new methods of palaeocurrent determination from trough cross-stratification. *Journal of Sedimentary Petrology* **53**, 629-42. (6)

Degens, E. T. 1965. *Geochemistry of sediments*.: Prentice-Hall. (2, 8,)

Dennison, J.M. and Ettensohn, F.R. (eds.), 1994. *Tectonic and eustatic controls on sedimentary cycles.* Society for Sedimentary Geology (SEPM), Concepts in Sedimentology and Paleontology, **4.**

Derbyshire, E., Gregory K. J. & Hails J. R. 1979. *Geomorphological processes*.: Butterworth. (Figs. 7.12 & 7.15)

Donovan, R.N. & Foster R. J. 1972. Sub-aqueous shrinkage cracks from the Caithness flagstone series (Middle Devonian) of northeast Scotland. *Journal of Sedimentary Petrology* **42**, 309–17. (9)

Dowdswell, J. A. & Sharp M. 1986. Characterization of pebble fabrics in modern terrestrial glacigenic sediments. *Sedimentology* **33**, 699–710. (7)

Doyle, P., Bennett M.R. & Baxter A.N. 2001. *The key to Earth history: An introduction to stratigraphy*. (2nd edn). Wiley. (2, 10)

Drewry, D. 1986. *Glacial geologic processes*.: Edward Arnold. (7)

Droser, M. L. & Bottjer D. J. 1986. A semiquantitative field classification of ichnofabric. *Journal of Sedimentary Petrology* **56**, 558-9. (9)

Dunbar, C. O. & Rodgers J. 1957. *Principles of stratigraphy*. John Wiley. (1, 2)

Dunham, R. J. 1962. Classification of carbonate rocks according to depositional texture. In: *Classification of Carbonate Rocks,* W. E. Ham (ed). *American Association of Petroleum Geologists Memoir,* **1**, 108-21. (8)

Edwards, M. B. 1976. Growth faults in Upper Triassic deltaic sediments, Svalbard. *American Association of Petroleum Geologists, Bulletin* **60**, 341–55. (9)

Elders, C. A. 1975. Experimental approaches in neoichnology. In *The study of trace fossils*, R. W. Frey (ed.), 513–36. Springer. (9)

Ekdale, A. A., Bromley R. G. & Pemberton S. G. 1984. *Ichnology: the use of trace fossils in sedimentology and stratigraphy*. Short-course Notes **15**. Society of Economic Paleontologists and Mineralogists. (9)

Elliott, T. 1976. Upper Carboniferous sedimentary cycles produced by river-dominated, elongate deltas. *Journal of the Geological Society of London,* **132**, 199–208. (10)

Emery, D. and Myers, K.J., 1996. *Sequence stratigraphy*, Blackwell Science.

Eyles, N. & Clark B. M. 1985. Gravity-induced soft-sediment deformation in glaciomarine sequences in the Upper Proterozoic Port Askaig Formation, Scotland. *Sedimentology* **32**, 789–814. (9)

Fairbridge, R.W. and Bourgeois, J. (eds.), 1978. *The encyclopedia of sedimentology.* Dowden, Hutchinson and Ross.

Farrow, G. E. 1975. Techniques for the study of fossil and recent traces. In *Trace fossils*, R. W. Frey (ed.), 537–54. Springer. (9, Appendix)

Fisher, R. V. 1966. Rocks composed of volcanic fragments and their classification. *Earth Science Reviews* **1**, 287–9. (7)

Franks, P. C. 1969. Nature, origin and significance of cone-in-cone structures in the Kiowa Formation (early Cretaceous), north central Kansas. *Journal of Sedimentary Petrology* **39**, 1438–54. (9)

Frey, R. W. (ed.) 1975. *The study of trace fossils*. Springer. (9)

Frey, R. W. & Pemberton S. G. 1985. Biogenic structures in outcrops and cores. I. Approaches to ichnology. *Bulletin of Canadian Petroleum Geologists*, **33**, 72-115.(9)

Frey, R. W., Howard J. D. & Pryor W. A. 1978. *Ophiomorpha*: its morphologic, taxonomic and environmental significance. *Palaeogeography, Palaeoclimatology, Palaeoecology* **23**, 199-229.(9)

Friedman, G. M. & Sanders J. E. 1978. *Principles of sedimentology*. John Wiley. (1,2,10)

Frost, S. H., Weiss M. P. & Saunders J. B. 1977. *Reefs and related carbonates: ecology and sedimentology*. Studies in Geology **4**, American Association of Petroleum Geologists (8)

Fryberger, S.G. (1979) Dune forms and wind regime. In: *A study of global sand seas.* E.D. McKee (ed). *United States Geological Survey Professional Paper*, **1052**, 137-169. (6)

Fryberger, G. S. & Schenk C. J. 1981. Wind sedimentation tunnel experiments on the origin of eolian strata. *Sedimentology* **28**, 805-21. (6)

Fryberger, S.G., Al-Sari A.M. & Clisham T.J. (1983) Eolian dune, interdune, sand sheet, and siciliclastic sabkha sediments of an offshore prograding sand sea, Dharan Area, Saudi Arabia. *American Association of Petroleum Geologists, Bulletin* **67**, 280-312. (6, 8)

Ghazi, S. and Mountney, N.P., 2009. Facies and architectural element analysis of a meandering fluvial succession: the Permian Warchha Sandstone, Salt Range, Pakistan. *Sedimentary Geology*, **221**, 99-126.

Ghinassi, G., Colombera, L., Mountney, N.P. and Ressink, A.J.H. (eds.), 2019. *Fluvial meanders and their sedimentary products in the rock record.* International Association of Sedimentologists, Special Publication, **48.**

Gibling M.R. & Tandon S.K. 1997 Erosional marks on consolidated banks and slump blocks in the Rupen River, north-west India. *Sedimentology* **44**, 221-51. (3, 7)

Gilbert G.K. (1885) The topographic features of lake shores. *Annual Report, U.S. Geological Survey*. **5**. 75-123. (6)

Gilbert, G. K. 1914. Transportation of debris by running water. *United States Geological Survey Professional Paper* **86**. (3, 6)

Gingras M.K., Pemberton S.G. & Saunders T. 2000. Firmness profiles associated with tidal-creek deposits: the temporal significance of *Glossifungites* assemblages. *Journal of Sedimentary Research*, **70** 1017-25. (9)

Ginsburg, R. N. (ed.) 1975. *Tidal deposits: a casebook of recent examples and fossil counterparts*. Springer. (5, 6, 10)

Glennie, K. S. 1970. *Desert sedimentary environments*. Elsevier. (6, 7)

Gluyas, J. and R.E. Swarbrick, 2003. *Petroleum geoscience*. Blackwell Science. (1, 10)

Goldring, R. 1999. *Field palaeontology,* (2nd edn). Pearson Education Limited. (9)

Goudie, A. 1973. *Duricrusts in tropical and subtropical landscapes*. Oxford University Press. (9)

Greer, S. A. 1975. Sand body geometry and sedimentary facies at the estuary-marine transition zone, Ossabaw Sound, Georgia; a stratigraphic model. *Senckenbergiana Maritima* **7**, 105–35. (Fig. 10.2)

Griffiths, J. C. 1961. Measurements of the properties of sediments. *Journal of Geology* **69**, 487–98. (2, Appendix)

Gruhn, R. & Bryan A. L. 1969. Fossil ice wedge polygons in southeast Essex, England. In *The periglacial environment*, T. L. Péwé (ed.), 351–63. Arctic Institute of North America. (Fig. 9.19)

Hallam, A. 1967. Siderite- and calcite-bearing concretionary nodules in the Lias of Yorkshire. *Geological Magazine* **104**, 222–7. (9)

Hallam, A., 1981. *Facies interpretation and the stratigraphic record.* W.H. Freeman and Company,

Hamblin, W.K. and Howard, J.D., 1995. *Exercises in physical geology.* (10th edn) Prentice-Hall.

Hand B. 1997 Inverse grading resulting from coarse-sediment transport lag. *Journal of Sedimentary Research* **67**, 124-9. (6)

Hand, B. M., Wessel J. M. & Hayes M. O. 1969. Antidunes in the Mount Toby Conglomerate (Triassic), Massachusetts. *Journal of Sedimentary Petrology* **39**, 1310–16. (6)

Harms, J. C., Southard J. B., Spearing D. R. & Walker R. G. 1975. *Depositional environments as interpreted from primary sedimentary structures and stratification sequences*. Short-course Notes **2**, Society of Economic Paleontologists and Mineralogists (10)

Harrison, S. S. 1967. *Low-cost flume construction*. *Journal of Geological Education* **15**, 105-8. (3, 6)

Hasiotis, S. T. 2002. *Continental trace fossils*. SEPM Short Course Notes No. **51**. Society of Economic Paleontologists and Mineralogists.(9)

Haughton, P., Davis, C., McCaffrey, W. & Barker, S., 2009. Hybrid sediment gravity flow deposits - Classification, origin and significance. *Marine and Petroleum Geology*, **26**, 1900-1918. (6)

Hein, F. J. & Walker R. G. 1977. Bar evolution and development of stratification in the gravelly, braided Kicking Horse River, British Columbia. *Canadian Journal of Earth Science* **14**, 562–70. (7)

Hein, F. J. & Walker R. G. 1982. The Cambro-Ordovician Cap Enrage Formation, Quebec, Canada: conglomeratic deposits of a braided submarine channel with terraces. *Sedimentology* **29**, 309–29. (7)

Hendry, H. E. & Stauffer M. R. 1977. Penecontemporaneous folds in cross-bedding: inversion of facing criteria and mimicry of tectonic folds. *Bulletin of the Geological Society of America,* **88**, 809–12. (9)

Higham, N. 1963. *A very scientific gentleman: the major achievements of Henry Clifton Sorby*. Pergamon Press. (1)

Hiscott, R. N. 1994. Traction-carpet stratification in turbidites – fact or fiction? *Journal of Sedimentary Petrology* **64**, 204-8. (6)

Horowitz, D. H. 1982. Geometry and origin of large-scale deformation structures in some ancient wind-blown sand deposits. *Sedimentology* **29**, 155–80. (9)

Howard, J. D. 1978. Sedimentology and trace fossils. In: *Trace fossil concepts,* P. B. Basan (ed), *SEPM Short*

*Course No. **5**. Society of Economic Paleontologists and Mineralogists* 11-42.(9)

Howell, J.A. 1992. *Sedimentology of the Rotliegend Supergroup of the UK Southern*

North Sea. PhD thesis. University of Birmingham.

Hsu, K. J. & Jenkyns H. C. (eds) 1974. *Pelagic sediments on land and under the sea*. Special Publication **1**, International Association of Sedimentologists. (5)

Hunter, R. E. 1977. Basic types of stratification in small aeolian dunes. *Sedimentology* **24**, 361–87. (6)

Hunter, R. E. 1985. Sub-aqueous sand-flow cross-strata. *Journal of Sedimentary Petrology* **55**, 886–94. (6)

Hunter, R.E., Richmond B.M. & Alpha T.R. (1983) Storm-controlled oblique dunes of the Oregon Coast. *Bulletin of the Geological. Society of America,* **94**, 1450-65. (6)

Ingram, R. L. 1954. Terminology for the thickness of stratification and parting units in sedimentary rocks. *Bulletin of the Geological Society of America,* **65**, 937–8. (2)

Inman, D. L. & Bowen A. J. 1963. Flume experiments on sand transport by waves and currents. *8th Conference on Coastal Engineering, Proceedings*, 137–50. American Society of Civil Engineers, Reston, Virginia. (6)

James, N. P. 1979. Shallowing-upward sequences in carbonates (facies models 10). In *Facies models*, R. G. Walker (ed.), 109–19. Geological Association of Canada. (8)

James N.P. &. Kendall A.C, 1992. Introduction to carbonate and evaporite facies models. In *Facies models: Response to sea level change*, (3rd edn), R. G. Walker & N. P. James (eds), 265-275. Geological Association of Canada. (8)

James, N.P. and Dalrymple, R.W. (eds), 2010. *Facies models 4.* Canadian Sedimentology Research Group, Geological Association of Canada. GEOtext **6**,

James, N.P. and Jones, B., 2015. *Origin of carbonate sedimentary rocks.* Wiley-Blackwell,

Jerram, D.A. 2001. Visual comparators for degree of grain-size sorting in 2-D and 3-D. Computers in Geosciences **27**, 485-92. (2)

Johnson, H. D. 1977. Sedimentation and water escape structures in some late Precambrian shallow marine sandstones from Finnmark, North Norway. *Sedimentology* **24**, 389–411. (9)

Jones, C. M. 1979. Tabular cross-bedding in Upper Carboniferous fluvial channel sediments in the southern Pennines, England. *Sedimentary Geology* **24**, 85–104. (6)

Jones M. E. & Preston R. M. F. (eds) 1987. *Deformation of sediments and sedimentary rocks*. Special Publication **29**, Geological Society of London (9)

Jones, S., 2015. *Introducing sedimentology*. Dunedin Academic Press,

Jopling, A. V. 1965. Angle of repose box suitable for either class use or research. *Journal of Geological Education* **13**, 143–4. (6)

Jopling, A. V. 1965. Hydraulic factors controlling the shape of laminae in laboratory deltas. *Journal of Sedimentary Petrology* **35**, 777–91. (6)

Jopling, A. V. & Walker R. G. 1968. Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology* **38**, 971–84. (6)

Jopling, A. V. & McDonald B. C. (eds) 1975. *Glaciofluvial and glaciolacustrine sedimentation*. Society of Economic Paleontologists and Mineralogists, Special Publication **23**. (6, 7)

Julien, P.Y., 2010. Erosion and sedimentation. (2nd edn) Cambridge University Press.

Kaneko, A. 1980. The wavelength of oscillation sand ripples. Report of the Research Instituute of Applied Mechanics. Kyushu University. **28**, 57–71. (6)

Kawakami G. & Kawamura M. 2002 Sediment flow and deformation (SFD) layers: evidence for intrastratal flow in laminated muddy sediments of the Triassic Osawa Formation, Northeast Japan. *Journal of Sedimentary Research* **72,** 171-81. (9)

Kennedy, J. F. 1961. *Stationary waves and antidunes in alluvial channels*. Report KH-R-Z, W. M. Keck Laboratory of Hydraulics and Water Resources. California Institute of Technology. (Fig. 6.90)

Kennedy, W. J. & Juignet P. 1974. Carbonate banks and slump beds in the Upper Cretaceous (Upper Turonian Santonian) of Haute Normandie, France. *Sedimentology* **21**, 1–42. (8, 9)

King, C. A. M. (ed.) 1976. *Periglacial processes*. Dowden, Hutchinson & Ross. (7, 9)

King, C. J. H. 1980. A small cliff-bound estuarine environment; Sandyhaven Pill in South Wales. *Sedimentology* **27**, 93–105. (6, 10)

King, C. H. J. 1980. Experimental sedimentology for advanced level students using a motorized wave tank. *Geology Teaching* **5**, 44–52. (6)

Klein, G. de V. (ed.) 1976. *Holocene tidal sedimentation*. Dowden, Hutchinson & Ross. (6)

Kleinhans M.G., Wilbers A.W.E., Swaaf A.D. & Van den Berg J.H. 2002 Sediment supply-limited bedforms in sand-gravel bed rivers. *Journal of Sedimentary Research* **72,** 629-40. (6, 7)

Knaust, D. and Bromley, R.G., 2012. *Trace fossils as indicators of sedimentary environments.* Elsevier, Developments in Sedimentology, **64.**

Kneller B. 1995. Beyond the turbidite paradigm: physical models for deposition of turbidites and their implications for reservoir prediction. In: *Characterization of deep marine clastic systems*. A.J. Hartley & D.J. Prosser (eds), 31-50. *Special Publication Geological Society of London*. **94.** (6, 10)

Kneller B. & Branney M.J. 1995. Sustained high-density turbidity currents and the deposition of thick massive sands. *Sedimentology* **42**, 607-16. (6)

Kocurek, G. & Dott R.H. (1981) Distinctions and uses of stratification types in the interpretation of eolian sand. *Journal of Sedimentary Petrology* **51**, 579-95. (6)

Komar, P. D. 1976. *Beach processes and sedimentation*. Prentice-Hall. (3, 6, 7)

Koster, E. H. & Steel R. J. (eds) 1984. *Sedimentology of gravels and conglomerates*. Memoir **10**, Canadian Society of Petroleum Geologists. (7)

Kranck, K. 1975. Sediment deposition from flocculated suspensions. *Sedimentology* **22**, 111–23. (5)

Krumbein, W. C. & Garrels R. M. 1952. Origin and classification of chemical sediments in terms of pH and oxidation/reduction potentials. *Journal of Geology* **60**, 1–33. (8)

Kuenen, P. H. & Migliorini C. I. 1950. Turbidity currents as a cause of graded bedding. *Journal of Geology* **58**, 91–127. (6)

Lajoie, J. 1972. Slump fold axis orientations: an indication of palaeoslope? *Journal of Sedimentary Petrology* **42**, 584–6. (9)

Lajoie, J. 1979. Volcaniclastic rocks. In: *Facies models,* R.G. Walker (ed) (1st edn), 183-190. Geological Association of Canada. (7)

Lambert, A. & Hsu K. J. 1979. Non-annual cycles of varve-like sedimentation in Walensee, Switzerland. *Sedimentology* **26**, 453–61. (5)

Lancaster, N. (1988) Controls on eolian dune size and spacing. *Geology* **16**, 972-975. (6)

Lancaster, N. (1995) *Geomorphology of desert dunes*. Routledge. (6)

Langhorne, D. N. 1982. A study of the dynamics of a marine sandwave. *Sedimentology* **29**, 571–94. (6)

Larsen, F. D. 1968. A tank for demonstrating alluvial processes. *Journal of Geological Education* **16**, 53–5. (6, 7)

Leclair S.F. 2002. Preservation of cross-strata due to the migration of subaqueous dunes: an experimental investigation. *Sedimentology*. **49**, 1157-80. (6)

Leclair S.F., Bridge J.S. &Wang F. 1997 Preservation of cross-strata due to migration of subaqueous dunes over aggrading and non-aggrading beds: comparison of experimental data with theory. *Geoscience Canada*, **24**, 55-66. (6)

Leeder, M. R. 1980. On the stability of lower stage plane beds and the absence of current ripples in coarse sand. *Journal of the Geological Society of London,* **137**, 423–9. (3, 6)

Leeder, M.R., 1981. *Sedimentology, processes and product.* George, Allen and Unwin,

Leeder, M. R. 1999. *Sedimentology and sedimentary basins: From Turbulence to Tectonics*. Blackwell Science.

Leeder, M. R. 2011. *Sedimentology and sedimentary basins,* (2nd edn) Wiley-Blackwell,

Leeder, M. and Pérez-Arlucea, M., 2006. *Physical processes in earth and environmental sciences.* Blackwell Science.

Lindholm, R. 1987. *A practical approach to sedimentology*. Allen & Unwin.

Ljunggren, P. and Sundborg Å. 1968. Some aspects of fluvial sediments and fluvial morphology, II: a study of some heavy mineral deposits in the valley of the river Lule Älv. *Geografiska Annaler* **50A**, 121–35. (6)

Logan, A. 1975. A modified Bagnold-type wind tunnel for laboratory use. *Journal of Geological Education* **23**, 114–15. (6)

Loope, D.B. (1985) Episodic deposition and preservation of eolian sands - a late Paleozoic example from southeastern Utah. *Geology* **13**, 73-6. (4, 6)

Lowe, D. R. 1975. Water escape structures in coarse-grained sediments. *Sedimentology* **22**, 157–204. (9)

Lowe, D. R. 1976. Sub-aqueous liquified and fluidized sediment flows and their deposits. *Sedimentology* **23**, 285–308. (9)

Lowe, D. R. 1982. Sediment gravity flows: 2. Depositional models with special reference to the deposits of high-density turbidity currents. *Journal of Sedimentary Petrology*. **52**, 279-97. (3)

Lowe, D. R. & Lopiccolo L. D. 1974. The characteristics and origins of dish and pillar structure. *Journal of Sedimentary Petrology* **44**, 484–501. (9)

Lowe, J.J. and Walker, M.J.C., 1997. Reconstructing Quaternary environments. (2nd edn) Prentice-Hall.

Lowenstein, T. K. & Hardie L. A. 1985. Criteria for the recognition of salt-pan evaporites. *Sedimentology* **32**, 627–44. (8)

Lyell, C. 1830–1833. *Principles of geology, being an attempt to explain the former changes of the earth's surface by reference to causes now in operation*. John Murray. (1)

Maejima, W. 1982. Texture and stratification of gravelly beach sediments, Enju Beach, Kii Peninsula, Japan. *Osaka City University Journal of Geosciences* **25**, 35–51. (Fig. 7.14)

MacKenzie, W.S. and Adams, A.E., 1994. *A colour atlas of rocks and minerals in thin section.* CRC Press.

Major J.J. 2000 Gravity-driven consolidation of granular slurries: implications for debris-flow deposition and deposit characteristics. *Journal of Sedimentary Research* **70**, 64-83. (3, 6)

Maltman A. (ed.) 1994. *The geological deformation of sediments*. Chapman and Hall. (9)

Manz, P. A. 1978. Bedforms produced by fine, cohesionless, granular and flakey sediments under subcritical water flows. *Sedimentology* **25**, 83–103. (6)

Maroulis J.C. & Nanson G.C. 1996. Bedload transport of aggregated muddy alluvium from Cooper Creek, central Australia: a flume study. *Sedimentology* **43**, 771-90. (5, 6, 10)

Marszalek, D. S. & Hay W. W. 1968. Marine aquaria for palaeontology. *Journal of Geological Education* **16**, 159–63. (8, 9)

Martinsen, O. 1994. Mass movements. In: *The Geological Deformation of Sediments.* A. Maltman (ed).127-165. Chapman and Hall. (9)

Matthews, R.K., 1974. *Dynamic stratigraphy.* Prentice-Hall.

McCabe, P. J. 1977. Deep distributary channels and giant bedforms in the Upper Carboniferous of the central Pennines, northern England. *Sedimentology* **24**, 271–90. (4, 6)

McCaffrey, W.D., Kneller, B.C. and Peakall, J. (eds.), 2001. *Particulate gravity currents.* International Association of Sedimentologists, Special Publication, **31.**

McCave, I. N. & Geiser A. C. 1979. Megaripples, ridges and runnels on intertidal flats of the Wash, England. *Sedimentology* **26**, 353–69. (6)

McKee, E. D. 1966. Structure of dunes at White Sands National Monument, New Mexico. *Sedimentology* **7**, 1–70. (6)

McKee, E. D. (ed.) 1979. *A study of global sand seas*. United States Geological Survey Professional Paper **1052**, (6)

McKee, E. D. & Tibbitts G. C. 1964. Primary structures of a seif dune and associated deposits in Libya. *Journal of Sedimentary Petrology* **34**, 5–17. (6)

McKee, E. D. & Weir G. W. 1953. Terminology for stratification and cross-stratification in sedimentary rocks. *Geological Society of America, Bulletin* **64**, 381–90. (2, 6)

McKee, E. D., Douglass J. R. & Rittenhouse S. 1971. Deformation of lee-side laminae in eolian dunes. *Geological Society of America, Bulletin* **82**, 359–78. (9)

Miall, A. D. 1973. Markov chain analysis applied to an ancient alluvial plain succession. *Sedimentology* **20**, 345–64 (10)

Miall, A. D. 1977. A review of the braided river depositional environment. *Earth Science Reviews* **13**, 1–62. (6, 7, 10)

Miall, A. D. (ed.) 1978. *Fluvial sedimentology*. Memoir **5**, Canadian Society of Petroleum Geology, (6, 7,10)

Miall, A. D. 1982. Recent developments in facies models for siliciclastic sediments. *Journal of Geological Education* **30**, 222–40. (10)

Miall, A. D. 1996. *The geology of fluvial deposits: Sedimentary facies, basin analysis and petroleum geology.* Springer-Verlag. (4, 6, 7)

Miall, A.D., 2014. *Fluvial depositional systems,* Springer,

Miall A.D. & Jones B.G. 2003. Fluvial architecture of the Hawkesbury Sandstone (Triassic), near Sydney, Australia. *Journal of Sedimentary Research* **73**, 531-545. (6. 10)

Middleton, G. V. 1966–7. Experiments on density and turbidity currents. Parts I-III. *Canadian Journal of Earth Science* **3**, 523–46, 627–37; **4**, 475–505. (3, 6)

Middleton, G. V. 1973. Johannes Walther's law of correlation of facies. *Bulletin of the Geological Society of America,* **84**, 979–88. (1, 10)

Middleton, G. V. 1976. Hydraulic interpretation of sand size distribution. *Journal of Geology* **84**, 405–426. (3, 6)

Middleton G.V. (ed.) 2003. *Encyclopedia of sediments and sedimentary rocks*. Kluwer Academic Publishers.

Middleton, G. V. & Bouma A. H. (eds) 1973. *Turbidites and deepwater sedimentation*. SEPM Short-course Notes. Society of Economic Paleontologists and Mineralogists. (Pacific Section). (3, 5, 6, 10)

Middleton, G. V. & Hampton M. A. 1976. Sub-aqueous sediment transport and deposition by sediment gravity flows. In *Marine sediment transport and environment management*, D. J. Stanley & D. J. P. Swift (eds), 197–218. John Wiley. (3)

Middleton, G. V. & Southard J. B. 1984. *Mechanics of sediment movement*. Society of Economic Paleontologists and Mineralogists. Short-course Notes No. 3. SEPM (Eastern Section). (3, 4, 5, 6)

Midtgaard H. 1996 Inner shelf to lower shoreface hummocky sandstone bodies with evidence for geostrophic induced combined flow, Lower Cretaceous, West Greenland. *Journal of Sedimentary Research* **66,** 343-353. (6)

Miller M.C. & Komar P.D. 1980 Oscillation sand ripples generated by laboratory apparatus. *Journal of Sedimentary Petrology* **50**, 173-82. (6)

Miller, M. F., Ekdale A. A. & Pilard M. D. (eds) 1984. Trace fossils and paleoenvironments: marine carbonate, marginal marine terrigenous and continental terrigenous settings. *Journal of Paleontology* **58**, 283–598. (9)

Miller, R.L. and Kahn, J.S., 1962. *Statistical analysis in the geological sciences.* John Wiley and Sons.

Millot, G. 1970. *Geology of clays*. Springer. (5)

Millot, G., 2014. Geology of clays. (soft-cover reprint of 1979 edition), Springer (5)

Morton J.B. 1996 Morphology and paleoenvironmental significance of Quaternary sand veins, sand wedges, and composite wedges, Tuktoyaktuk coastlands, western Arctic Canada. *Journal of Sedimentary Research* **66,** 17-25. (9)

Moseley, F. 1981. *Methods in field geology*. W. H. Freeman. (2)

Mountney N. & Howell J. 2000. Aeolian architecture, bedform climbing and preservation space in the Cretaceous Etjo Formation, NW Namibia. *Sedimentology* **47**, 825-49. (6)

Mountney, N. P. & Jagger A.J. 2004. Stratigraphic evolution of an aeolian erg margin system: the Permian Cedar Mesa Sandstone, SE Utah, USA. *Sedimentology* **51**, 713-43. (6, 10)

Mountney, N.P & Thompson D.B. (2002) Stratigraphic evolution and preservation of aeolian dune and damp/wet interdune strata: an example from the Triassic Helsby Sandstone Formation, Cheshire Basin, UK. *Sedimentology* **49**, 805-34. (10)

Mulder, T. & Alexander J. 2001. The physical character of subaqueous sedimentary density flows and their deposits. *Sedimentology* **48**, 269-99. (3)

Mutti. E. 1992. *Turbidite sandstones*. San Donato Milanese: AGIP. (4, 6, 7, 9)

Myrow P.H., Fischer W. & Goodge J.W. 2002 Wave-modified turbidites: combined-flow shoreline and shelf deposits, Cambrian, Antarctica. *Journal of Sedimentary Research* **72**, 641-56. (6)

Nemec, W. & Steel R. J. 1984. Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits. In *Sedimentology of gravels and conglomerates*, E. H. Koster & R. J. Steel (eds), 1–31. *Memoir **10**, Canadian Society of Petroleum Geology, Calgary.* (3, 7)

Nichols, G. 1999. *Sedimentology and stratigraphy*. Blackwell Science

Nissenbaum, A. 1980. *Hypersaline brines and evaporitic environments*. Developments in Sedimentology **28**, Elsevier. (8)

Noffke N., Gerdes G., Klenke T. & Krumbein W.E. 2001. Microbially induced sedimentary structures - A new category within the classification of sedimentary structures. *Journal of Sedimentary Research* **71,** 649-56. (8, 9)

Nøttvedt A. & R.D. Kreisa 1987. Model for combined-flow origin of hummocky cross-stratification. *Geology*, **15**, 357-61. (6)

O'Brien, N. R. 1970. The fabric of shale – an electron microscope study. *Sedimentology* **15**, 229–46. (5)

Owen, G., 1987. Deformation processes in unconsolidated sands. In *Deformation of sediments and sedimentary rocks*, M.E. Jones and R.M.F. Patterson (eds.), *Special Publication, Geological Society of London* **29**, 11-24. (9)

Owen G. 1995 Soft-sediment deformation in Upper Proterozoic Torridonian sandstones (Applecross Formation) at Torridon Northwest Scotland. *Journal of Sedimentary Research*, **65,** 495-504. (9)

Owen G. 1996. Experimental soft-sediment deformation: structures formed by the liquefaction of unconsolidated sands and some ancient examples. *Sedimentology* **43**, 279-93. (9)

Paik I.S. & Kim H.J. 1998 Subaerial lenticular cracks in Cretaceous lacustrine deposits, Korea. *Journal of Sedimentary Research* **68,** 80-87. (9)

Park, R. K. 1976. A note on the significance of lamination in stromatolites. *Sedimentology* **23**, 379–93. (8)

Park, R. K. 1977. The preservation potential of some recent stromatolites. *Sedimentology* **24**, 485–506. (8)

Paru, W. C. & Schot E. H. 1968. Stylolites: their nature and origin. *Journal of Sedimentary Petrology* **38**, 175–91. (9)

Paterson, W.S.B. 1994. *The Physics of Glaciers*. (3rd ed). Pergamon Press (3)

Pemberton, S. G. 1992. *Applications of ichnology to petroleum exploration. A core workshop.* Society of Economic Paleontologists and Mineralogists. Core Workshop No. 17, Calgary, Canada. (9)

Peterson, J.A. and Osmond, J.C. (eds.), 1961. *Geometry of sandstone bodies.* American Association of Petroleum Geologists

Pettijohn, F. J. 1975. *Sedimentary rocks*, (3rd edn). Harper & Row.

Pettijohn, F. J. & Potter P. E. 1964. *Atlas and glossary of sedimentary structures*. Springer.

Pettijohn, F. J., Potter P. E. & Siever R. 1972. *Sand and sandstone*. Springer. (4, 6)

Petts, G. and Calow, P. (eds.), 1996. *River flows and channel forms*. Blackwell Science.

Picard, M.D. and High, L.R., 1973. *Sedimentary structures of ephemeral streams.* Elsevier, Developments in Sedimentology, **17**,

Piper, D. P. & Rogers P. J. 1980. *Procedure for the assessment of the conglomerate resources of the Sherwood Sandstone Group*. Mineral Assessment Report of the Institute of Geological Sciences. no. **56**. (Fig 7.2)

Plint, A. G. 1986. Slump blocks, intraformational conglomerates and associated structures in Pennsylvanian fluvial strata of eastern Canada. *Sedimentology* **33**, 387–99. (7)

Posamentier, H.W. and Allen, G.P., 1999. *Siliciclastic sequence stratigraphy – concepts and applications.* Society for Sedimentary Geology (SEPM), Concepts in Sedimentology and Paleontology, **7**.

Posamentier, H.W. and Walker, R.G. (eds.), 2006. *Facies models revisited.* Society for Sedimentary Geology (SEPM), Special Publication, **84**.

Postma, G. 1983. Water escape structures in the context of a depositional model of a mass flow dominated conglomeratic fan delta (Abrioja Formation, Pliocene; Almeria Basin, S. E. Spain). *Sedimentology* **30**, 91–103. (7, 9)

Potter, P.E. and Pettijohn, F.J., 1963. *Paleocurrents and basin analysis*. Springer,

Potter, P. E., Maynard J. B. & Pryor W. A. 1980. *Sedimentology of shale*. Springer. (5)

Pratt, B. R. & James N. P. 1982. Cryptalgal-metazoan bioherms of Early Ordovician age in the St George Group, western Newfoundland. *Sedimentology* **29**, 543–69. (8)

Pratt, B.R., James N.P. & Cowan C.A. 1992. Peritidal carbonates. In *Facies models: Response to Sea Level Change*, (3rd edn), R. G. Walker & N. P. James (eds), 303-322. Geological Association of Canada. (8, 10)

Pretious, E. S. & Blench T. 1951. *Final report on special observations of bed movement in the lower Fraser River at Ladner Beach during 1950 freshet*. National Research Council of Canada. (6)

Preiss, W. V. 1976. Basic field and laboratory methods for the study of stromatolites. In *Stromatolites*, M. R. Walter (ed.), 5–13. Elsevier. (8)

Price, R. J. 1973. *Glacial and fluvioglacial landforms*. Longman. (4, 7)

Prothero, D.R. and Schwab, F., 2014. Sedimentary geology. *An introduction to sedimentary rocks and stratigraphy.* (3rd edn) W.H. Freeman,

Pye, K. (ed.), 1994. *Sediment transport and depositional processes.* Blackwell Scientific Publications.

Raaf, J. F. M. de, Reading H. G. & Walker R. G. 1965. Cyclic sedimentation in the Lower Westphalian of north Devon, England. *Sedimentology* **4**, 1–52. (10)

Raaf, J. F. M. de, Boersma J. R. & Van Gelder A 1977. Wave generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. *Sedimentology* **24**, 451–83. (6)

Raiswell R. & Fisher Q.J. 2000. Mudrock-hosted carbonate concretions: a review of growth mechanisms and their influence on chemical and isotopic composition. *Journal of the Geological Society of London*. **157**, 239-51. (9)

Reading H.G. (ed.) 1986 *Sedimentary environments and facies*. (2nd edn), Blackwell Scientific.

Reading, H. G. (ed.) 1996. *Sedimentary environments; processes, facies and stratigraphy* (3rd edn) Blackwell Science. (10)

Reineck H.-E & Singh I.B. 1973 *Depositional sedimentary environments*, Berlin, Springer. (2)

Reineck, H. E. & Singh I. B. 1980. *Depositional sedimentary environments*, (2nd edn). Springer.

Reineck, H. E. & Wunderlich F. 1968. Classification and origin of flaser and lenticular bedding. *Sedimentology* **11**, 99–104. (5, 6)

Retallack, G.J. 1997. A colour guide to palaeosols. John Wiley and Sons. (9)

Ricci-Lucchi, F. 1970. *Sedimentografia*. Zanichelli. (4, 6, 7, 9)

Rider, M. H. 1978. Growth faults in Carboniferous of western Ireland. *Bulletin of the American Association of Petroleum Geologists,* **62**, 2191–213. (9)

Riding, R. 1979. Origin and diagenesis of lacustrine algal bioherms at the margin of the Ries crater, Upper Miocene, Southern Germany. *Sedimentology,* **26**, 645–80. (8)

Ringrose, P. 1988. Palaeoseismic (?) liquefaction event in late Quaternary lake sediment at Glen Roy, Scotland. *Terra Nova*, **1**, 57-62. (9)

Roberts, H. H., Cratsley D. W. & Whelant T. 1976. *Stability of Mississippi delta sediments as evaluated by analysis of structural features in sediment borings*. Paper OTC 2425, Offshore Technological Conference (4)

Rossetti D. de F. 1999. Soft-sediment deformation structures in late Albian to Cenomanian deposits, São Luís Basin, northern Brazil: evidence for palaeoseismicity. *Sedimentology* **46**, 1065-81. (9)

Rubin, D.M. 1987. Cross-bedding, bedforms and palaeocurrents. *SEPM Concepts in Sedimentology and Paleontology* **1** SEPM. (6)

Rubin, D. M. & Hunter R. E. 1982. Bedform climbing in theory and nature. *Sedimentology* **29**, 121–38. (6)

Rubin, D.M & Hunter R.E. 1983. Reconstructing bedform assemblages from compound crossbedding. In: *Eolian sediments and processes.* M.E. Brookfield & T.S. Ahlbrandt (eds), *Developments in Sedimentology*, **38**, 407-427. Elsevier. (6)

Rubin, D. M. & R. E. Hunter 1985. Why deposits of longitudinal dunes are rarely recognized in the geological record. *Sedimentology* **32**, 147–57. (6)

Rubin, D.M. and Carter, C.L., 2006. *Cross-bedding, bedforms and paleocurrents in animation*. Society for Sedimentary Geology (SPEM), Atlas, 2.

Ruffell A. & Wach G. 1998. Firmgrounds – key surfaces in the recognition of parasequences in the Aptian Lower Greensand Group, Isle of Wight (southern England). *Sedimentology* **45**, 91-107. (9)

Schaefer, W. 1972. *Ecology and paleoecology of marine environments*. University of Chicago Press. (8, 9)

Schmincke, H. U., Fisher R. V. & Waters A. C. 1973. Antidune and chute and pool structures in base-surge deposits of the Laacher See area, Germany. *Sedimentology* **20**, 553–74. (6, 7)

Scholle, P. A., Bebout D. G. & Moore C. H. (eds) 1983. *Carbonate depositional environments*. Memoir **33**. Tulsa: American Association of Petroleum Geologists. (8)

Scholle, P. A. & Spearing D. (eds) 1982. *Sandstone depositional environments*. Memoir **31**. Tulsa: American Association of Petroleum Geologists. (10)

Schreiber, B. C. 1978. Environments of sub-aqueous gypsum deposition. In *Marine evaporites*, W. E. Dean and B. C. Schreiber (eds), 43–73. *Society of Economic Paleontologists and Mineralogists, Short-course Notes* **4**. (8)

Schreiber, B. C. 1986. Arid shorelines and evaporites. In: *Sedimentary environments and facies* (2nd edn). H. G. Reading (ed.), 189-228. Blackwell Scientific Publications (8)

Scoffin, T. P. 1971. The conditions of growth of the Wenlock reefs of Shropshire, England. *Sedimentology* **17**, 173–219. (8)

Scoffin, T. P. 1987. *An introduction to carbonate sediments and rocks*. Blackie. (8)

Seilacher, A. 1964. Sedimentological classification and nomenclature of trace fossils. *Sedimentology* **3**, 253–6. (9)

Seilacher, A. 1967. Fossil behaviour. *Scientific American* **217**, 72–84. (9)

Seilacher, A. 1967. Bathymetry of trace fossils. *Marine Geology* **5**, 413–28. (9)

Selles-Martinez J. 1996. Concretion morphology, classification, and genesis. *Earth Science Reviews* **41**, 177-210 (9)

Selley, R. C. 1976. *An introduction to sedimentology*. Academic Press. (10)

Selley, R. C. 1988. *Applied sedimentology*. Academic Press. (10)

Selley, R.C., 1995. *Ancient sedimentary environments and their subsurface diagnosis.* Routledge.

Sengupta, S. 1966. Studies on orientation and imbrication of pebbles with respect to cross-stratification. *Journal of Sedimentary Petrology* **36**, 362-9. (7)

Shanley, K.W. and McCabe, P.J., 1998. *Relative role of eustasy, climate and tectonism in continental rocks.* Society for Sedimentary Geology (SEPM), Special Publication, **59**.

Shearman, D. J. 1971. *Marine evaporites: the calcium sulfate facies*. Seminar, American Association of Petroleum Geologists, University of Calgary, Canada. (8)

Shearman, D. J. 1978. Evaporites of coastal sabkhas. In: *Marine evaporites*, W. E. Dean & B. C. Schreiber (eds), 6–42. Society of Economic Paleontologists and Mineralogists, Short Course **4**, (8)

Simpson, J. 1985. Stylolite-controlled layering in a homogeneous limestone: pseudo-bedding produced by burial diagenesis. *Sedimentology* **32**, 495–505. (9)

Simpson, J.E. 1987. *Gravity currents: In the environment and the laboratory*. Ellis Horwood/Wiley. (3)

Smith, N. D. 1974. Sedimentology and bar formation in the Upper Kicking Horse River, a braided outwash stream. *Journal of Geology* **82**, 205-23. (7)

Smith R.M.H., T.R. Mason, J.D. Ward 1993. Flash-flood sediments and ichnofacies of the Late Pleistocene Homeb Silts, Kuiseb River, Namibia. *Sedimentary Geology* **85**, 579-599**.** (9)

Sorby, H. C. 1908. On the application of quantitative methods to the study of the structure and history of rocks. *Quarterly Journal of the Geological Society of London*, **64**, 171-233.

Sohn Y.K. 1997 On traction-carpet sedimentation. *Journal of Sedimentary Research* **67,** 502-9. (3, 6, 7)

Sohn Y.K., Kim S.B., Hwang I.G., Bahk J.J., Choe M.Y. & Chough S.K. 1997 Characteristics and depositional processes of large-scale gravely Gilbert-type foresets in the Miocene Doumsan fan delta, Pohang Basin, SE Korea. *Journal of Sedimentary Research*, **67,** 130-41. (7)

Southard J.B., Lambie J.M., Federico D.C., Pile H.T. & Weidman C.R. 1990. Experiments on bed configurations in fine sands under bi-directional purely oscillatory flow, and the origins of hummocky cross-stratification. *Journal of Sedimentary Petrology*. **53**, 1-17. (6)

Southgate, P. N. 1982. Cambrian skeletal halite crystals and experimental analogues. *Sedimentology* **29**, 391–407. (8)

Sparks, R. S. J., Self S.& Walker G. P. L. 1973. Products of ignimbrite eruptions. *Geology* **1**, 115-8. (7)

Steel, R. J. & Thompson D. B. 1983. Structures and textures in Triassic braided stream conglomerates (“Bunter” Pebble Beds) in the Sherwood Sandstone Group, North Staffordshire, England. *Sedimentology* **30**, 341–67. (7)

Stokes, S. L. 1968. Multiple parallel-truncation bedding planes – a feature of wind deposited sandstone formations. *Journal of Sedimentary Petrology* **38**, 510–15. (6)

Stow, D.A.V., 1992. *Deep-water turbidite systems.* International Association of Sedimentologists, Reprint series, 3,

Stow, D.A.V., 2005. *Sedimentary rocks in the field. A colour guide*. Manson Publishing.

Stow, D. A. V. & Piper D. J. W. (eds) 1984. *Fine grained sediments; deep-water processes and facies*. Special Publication **15**, Geological Society of London. (5)

Stow, D. A. V., Reading H. G. & Collinson J. D., 1996. Deep Seas. In: *Sedimentary environments: Processes, facies and stratigraphy* (3rd edn). H. G. Reading (ed), 395-453. Blackwell Science. (6)

Sugden, W. 1964. Origin of faceted pebbles in some recent desert sediments of southern Iraq. *Sedimentology* **3**, 65–74. (4)

Summerson, C. H. (ed.) 1976. *Sorby on sedimentology, a collection of papers from 1851 to 1908 by Henry Clifton Sorby*. Miami Comparative Sedimentology Lab. Miami University (1)

Sundborg, Å. 1956. The river Klarälven: a study of fluvial processes. *Geografiska Annaler* **38**, 127–316. (3, 6)

Surlyk, F. 1978. Submarine fan sedimentation along fault scarps on tilted fault blocks (Jurassic/Cretaceous boundary, east Greenland). *Grønlands Geol. Unders, Bull.* **128**. (Fig. 10.2)

Suthren, R. J. 1985. Facies analysis of volcaniclastic sediments: a review. In *Sedimentology: recent developments and applied aspects*, P. J. Brenchley & B. J. P. Williams (eds), 123–46. *Special Publication of the Geological Society of London.* **18**. (7)

Swift D.J.P., Figueiredo A.G.Jr., Freeland G.L. & Oertel G.F. 1983. Hummocky cross-stratification and megaripples: a geological double standard? *Journal of Sedimentary Petrology*. **53**, 1295-317. (6)

Talbot, M. R. 1985. Major bounding surfaces in aeolian sandstones – a climatic model. *Sedimentology* **32**, 257–65. (6)

Talling, P.J., Masson, D.G., Sumner, E.J. & Malgesini, G., 2012. Subaqueous sediment density flows: depositional processes and deposit types. *Sedimentology*, **59,** 1937-2003.

Tanner P.W.G. 1998. Interstratal dewatering origin for polygonal patterns of sand-filled cracks: a case study from late Proterozoic metasediments of Islay, Scotland. *Sedimentology*, **45**, 71-89. (9)

Tanner, W. F. 1962. Inexpensive models for study of helical flow in streams. *Journal of Geological Education* **10**, 116–8. (3)

Taylor, A.M. & Goldring, R. 1993. description and analysis of bioturbation and ichnofabric. *Journal of the Geological Society of London* **150**, 141-8. (9)

Tennekes, H. and Lumley, J.L., 1972, *A first course in turbulence*. MIT Press.

Thompson, D. B. 1969. Dome-shaped aeolian dunes in the Frodsham Member of the so-called "Keuper" Sandstone (Scythian–?Anisian: Triassic) at Frodsham, Cheshire (England). *Sedimentary Geology* **3**, 263–89. (6)

Thompson, D. B. 1975. Types of geological fieldwork in relation to the objectives of teaching science. *Geology* **6**, 52–61. (1, 2)

Till, R. 1974. *Statistical methods for the earth scientist: an introduction*. Macmillan. (10, Appendix)

Toomey, D. F. (ed.) 1981. *European fossil reef models*. SEPM Special Publication **30**, Tulsa: SEPM. (8)

Tritton, D.J., 1988. *Physical fluid dynamics.* (2nd edn) Oxford Science Publications.

Tucker, M. E. 1985. Shallow-marine carbonate facies and facies models. In *Sedimentology: recent developments and applied aspects*, P. J. Brenchley & B. P. J. Williams (eds), 147–69.

Special Publication of the Geological Society of London. **18**, (8)

Tucker, M. E. 1988. Techniques in sedimentology. Blackwell Scientific Publications. (10)

Tucker, M. E. 2001. *Sedimentary petrology*. Blackwell Science. (1, 2, 3, 4, 6, 7, 8)

Tucker, M. E. 2003. *Sedimentary rocks in the field*. Wiley. (1, 2, 10)

Tucker M.E. & V.P. Wright 1990. *Carbonate sedimentology*. Blackwell Scientific Publications. (8)

Visher, G. S. 1969. Grain size distributions and depositional processes. *Journal of Sedimentary Petrology* **39**, 1074–106. (3)

Walford, N., 2011. *Practical statistics for geographers and earth scientists.* Wiley-Blackwell.

Walker, G. P. L. 1971. Grain size characteristics of pyroclastic deposits. *Journal of Geology* **79**, 696–714. (7)

Walker, R. G. 1978. Deep-water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps. *Bulletin of the American Association of Petroleum Geologists,* **62**, 932–66. (10)

Walker, R.G. 1979. *Facies Models* (1st edn). Geological Association of Canada. (10)

Walker, R. G. & James N.P. (eds.) 1992. *Facies models: Response to sea level change.* (3rd edn). Geological Association of Canada. (10)

Walter, M. R. (ed.) 1976. *Stromatolites*. Elsevier. (8)

Waltham, D., 1999. *Mathematics: a simple tool for geologists.* Chapman and Hall.

Walther, J. 1894. *Einleitung in die Geologic als Historische Wissenschaft, Bd 3: Lithogenesis der Gegenwart* [Vol. 3: Modern lithogenesis]. Jena: Fisher, 535–1055. (1, 10)

Warren, J., 1999. *Evaporites, their evolution and economics*. Blackwell Science.

Wasson, R.J. & Hyde, R. (1983) Factors determining desert dune type. *Nature* **304**, 337-9. (6)

Webby, B. D. 1969. Trace fossils (Pascichnia) from the Silurian of New South Wales, Australia. *Paläontologische Zeitschrift* **43**, 81-94. (9)

Weiler, Y., Sass E. & Zak I. 1974. Halite oolites and ripples in the Dead Sea, Israel. *Sedimentology* **21**, 623-32. (8)

West, I. M. 1975. Evaporite and associated sediments of the basal Purbeck formation (Upper Jurassic) of Dorset. *Geologists' Association, Proceedings* **86**, 205–25. (8)

Wetzel, A. 1984. Bioturbation in deep-sea fine-grained sediments: influence of sediment texture, turbidite frequency and rates of environmental change. In *Fine-grained sediments: deep water processes and facies*. D. A. V. Stow & D. J. W. Piper (eds), 595-608. *Special Publication of the Geological Society of London.* **15.**

Whitaker, J. H. McD. 1974. "Guttercasts", a new name for scour and fill structures: with examples from the Llandoverian of Ringerike and Malmoya, Southern Norway. *Norsk Geologisk Tidsskrift* **53**, 403–17. (4)

Whitfield, W. B. 1979. The development and educational uses of a motorized wave tank. *Geology Teaching* **4**, 64–8. (3)

Williams, H. & McBirney A. R. 1979. *Volcanology*. Freeman, Cooper. (7)

Wilson, I. G. 1972. Aeolian bedform – their development and origins. *Sedimentology* **19**, 173–210. (6)

Wilson, I. G. 1973. Ergs. *Sedimentary Geology* **10**, 77–106. (6)

Wilson, J. L. 1975. *Carbonate facies in geologic history*. Springer. (8)

Wolfenden, E. B. 1958. Palaeo-ecology of the Carboniferous reef complex and shelf limestones in north-west Derbyshire, England. *Bulletin of the Geological Society of America,* **69**, 871–98. (8))

Woodcock, N. H. 1979. The use of slump structures as palaeoslope orientation estimators. *Sedimentology* **26**, 83–99. (9)

Wright, A. E. & F. Moseley (eds) 1975. *Ice ages, ancient and modern* Special issue **6**, Geological Journal. (7)

Wright, J. V., A. L. Smith, S. Self 1980. A working terminology of pyroclastic deposits. *Journal of Volcanology and Geothermal Research* **8**, 315–36. (7)

Wright, V. P. (ed.) 1986. *Palaeosols: their recognition and interpretation*. Blackwell Scientific Publications (9)

Wright, V.P. & M. E. Tucker, 1991. *Calcretes*. Reprint Series Volume 2 of the International Association of Sedimentologists. Blackwell Scientific Publications. (9)

Yan, N., Mountney, N.P., Colombera, L. and Dorrell, R.M., 2017. A 3D forward stratigraphic model of fluvial meander-bend evolution for prediction of point-bar lithofacies architecture. *Computers and Geosciences*, **105**, 65-80.

Yalin, M.S., 1977. *Mechanics of sediment transport.* Pergamon Press.

Yoxall, W. H. 1983. *Dynamic models in earth-science instruction*. Cambridge University Press. (2, 3)

Index

SEDIMENTARY STRUCTURES **Fourth Edition**

Now enhanced with colour photographs and illustrations, this thoroughly revised fourth edition of *Sedimentary Structures* enhances its status as a major textbook in geology. Since its first publication in 1982, *Sedimentary Structures* has established and maintained itself as a pre-eminent resource that sets out a clear methodology and philosophy for understanding present-day sediments and sedimentary rocks as the product of dynamic processes.

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***Sedimentary Structures* is designed principally for use in undergraduate settings and will be invaluable to students reading geology, earth sciences, physical geography and environmental sciences throughout their degree studies. It will also appeal to enthusiastic students at colleges and schools, as well as to amateur geologists who want to gain an understanding of sedimentary processes and products. Furthermore, the book is also valuable as a reference for both academic researchers and industry professionals alike. The fourth edition covers all major recent developments in the subject. It is characterized by an abundance of informative illustrations and photographic examples, and introduces colour figures for the first time.**

This edition, the first prepared without the direct input of the late David Thompson, builds on a major re-write that paid particular attention to recent advances in the understanding of aeolian processes and bedforms, and in the interpretation of trace fossils. The introduction to environmental interpretation has been further developed to reflect recent advances in stratigraphic thinking, thereby enabling sedimentologists to more readily relate the occurrence of assemblages of sedimentary structures to likely environments of deposition. *Sedimentary Structures* emphasizes a practical, hands-on approach. It remains indispensable to those with a serious interest in the study of sedimentary structures, not only as fascinating features in themselves but also as key indicators in the reconstruction of past environments.

John Collinson, formerly Professor of Sedimentology at the University of Bergen, Norway, and until recently ran his own geological consultancy. **Nigel Mountney** is Professor of Sedimentology in the School of Earth and Environment at the University of Leeds where he is the Director of the Fluvial & Eolian Research Group; additionally, he is a former Chief Editor of *Sedimentology*, a leading academic journal.

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