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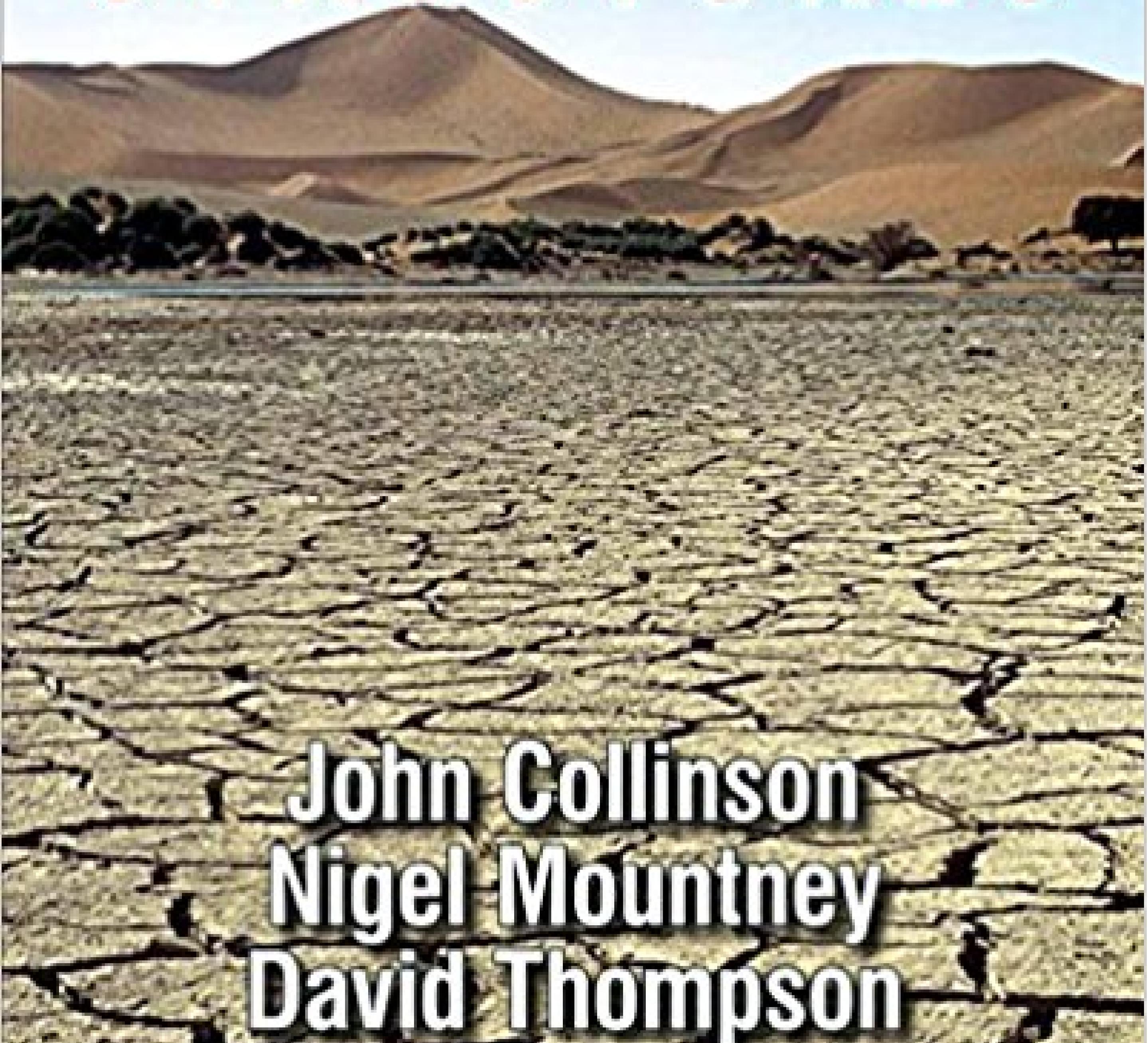

John Collinson
Nigel Mountney
David Thompson

THIRD EDITION

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Sedimentary structures

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TERRA

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Preface to the third edition

Since the publication of the second edition of this book in 1989, important advances have been made in many areas of sedimentology and, in view of frequent enquiries about the book's availability, we decided to prepare a third edition. This decision was made much easier by the inclusion of Nigel Mountney in the team, as he brings a fresh approach and a particular expertise in aeolian sediments, one of the main areas of advance in the past few years. The preparation of the book has also been encouraged by Roger Jones of Terra Publishing, who was also responsible for the publication of the first two editions.

The book is still envisaged primarily as an undergraduate text and it provides a starting point for understanding the morphology and process of formation of common sedimentary structures, with examples taken from both modern and ancient settings. The book is especially useful in both field and laboratory settings, and has been written for specialist Earth scientists and for non-specialists from a variety of educational backgrounds and subject areas who want to gain a basic understanding of the origin and form of structures in sediments and sedimentary rocks. It is hoped that this book will provide an introduction to more advanced topics in sedimentary processes and facies analysis

Although much of the book's content is based on basic physics and chemistry, we have tried to minimize the use of equations and have included only those that are essential for clear description and explanation of some of the key processes. We feel that even the most equation-shy reader will benefit from working through the basic explanations relating to important physical processes.

The whole book has been significantly rewritten and substantial changes have been made in the areas of aeolian sediments and trace fossils. Both of these topics have seen major advances in the past 20 years, with major textbooks and many scientific papers being published. In addition, new insights into gravitational mass movement of sediment have been developed. We hope that we have captured the essence of these advances within the confines and limitations of this relatively short book.

Throughout the book, we have tried to suggest ways in which simple experiments can help to reinforce understanding of some of the processes and ideas, and we hope that these will be seen as mere starting points for imaginative developments by teachers, tutors and students alike. The book is to some extent a field manual that allows structures to be recognized for a variety of geological purposes and, to some extent, a process-orientated account that allows students to use a basic experience of physics, chemistry and biology to explain the origins of structures. In this edition, we have simplified and updated the references and bibliographies. We have deliberately avoided referring to websites because many are ephemeral and others are of dubious accuracy.

We hope that this edition not only puts the book back into circulation but also provides a significant improvement on earlier editions. During its preparation, several colleagues have helped us in various ways. In particular we would like to thank John Pollard who has helped enormously with updating the section on trace fossils. We also thank Gilbert Kelling for providing several photographs, and all the authors and publishers who have allowed us to use illustrations from their publications.

John Collinson   Nigel Mountney   David Thompson
September 2006

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CHAPTER 1
Introduction

The study of sedimentary rocks has come a long way in the past 200 years. In the nineteenth century, they were 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 some way, 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, with a few notable exceptions, detailed study 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, although 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 to some extent by the needs of the oil industry in the exploration for hydrocarbon 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 to 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 involving 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 subdiscipline 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) associated with deepening, or on regression (seaward outbuilding of a shoreline) and erosive incision, and on the vertical stacking patterns of sediments drew attention away from the sediments themselves for a time. The balance is now nearly restored and we live in a time when a full integration of sedimentological and stratigraphical 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. We have, therefore, rewritten this book because of the fundamental importance of sedimentary structures to virtually all interpretations of sedimentary rocks and also because they are fascinating and often 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 and scientific understanding, which can give great intellectual satisfaction to those who enjoy asking questions of the world around them.

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 the series of

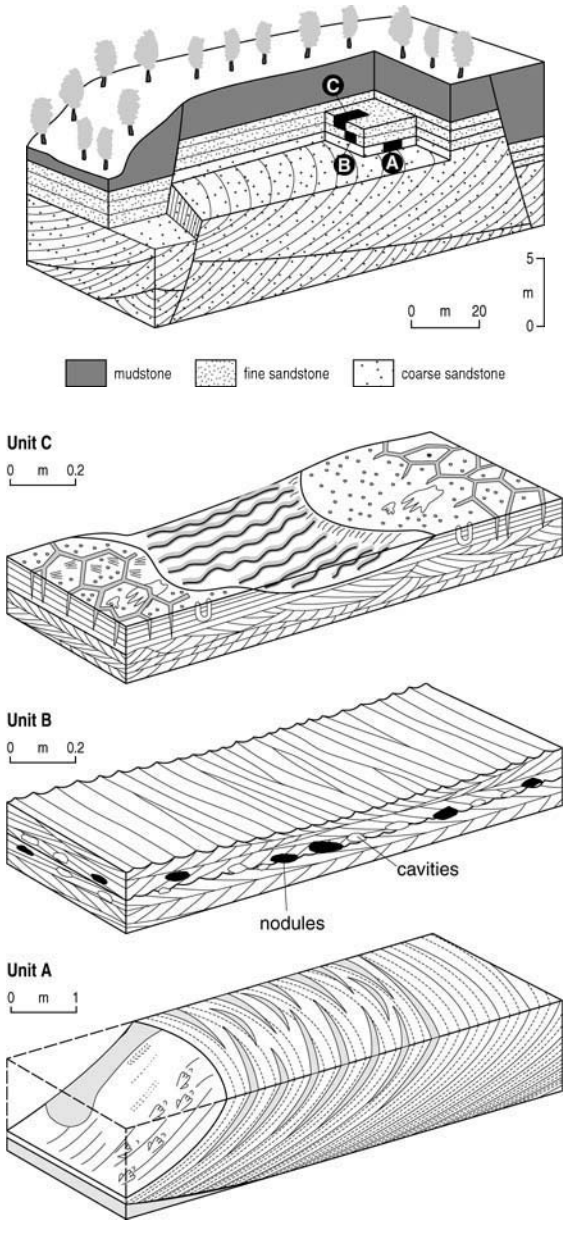


Figure 1.1 Sedimentary structures exposed in three blocks representative of units A, B and C in a hypothetical quarry. Note the scales of the blocks and their orientation within the quarry.

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. We hope that your ability to apply a more complete and detailed analysis will develop from reading this book.

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

- 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, and possibly classified certain features and perhaps have inferred and predicted relationships between them. 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, you have applied the idea that the present is the key to explaining the past. This doctrine of **uniformitarianism** was promoted by Charles Lyell in the mid-nineteenth century; it 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, critical, scutiny for examination of the evidence, or whether you first described each part 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 at the heart of the geological sciences and which helps to distinguish them from the other sciences.

1.2 The wider geological context

Sedimentology is the study of the nature and origin of both present-day and ancient sedimentary deposits. It includes sedimentary petrography (the description of composition and fabric) and is closely related to stratigraphy, particularly its most recent development, sequence stratigraphy, to which it contributes important criteria for the identification of key surfaces for correlation and 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 are best understood by input from all these subdisciplines, are generated from materials of diverse compositions and are observably products of physical, chemical and biological processes. Although certain processes are common to many present-day environments, combinations of processes, often with particular directional properties, may be unique to specific environments, and hence can 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 involve palaeontological, compositional and textural features, some of which may be identified only by laboratory observation and analysis. Such characterization allows similar rock units to be grouped together as “facies”, with the implication that different facies or successions 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.

1.3 Sedimentary structures and science

From the 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 successively at particular times and places. In particular, it involves interpreting processes from products 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 it is none the less satisfying for that. It 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 and the nature of chemical reactions have always operated in the same way. These ideas are 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 nineteenth 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. It is important to recognize that uniformitarianism is a subset of actualism, whose principles are universally accepted. Lyell's gradualist uniformitarianism was set against the idea of **catastrophism**, which suggested that sudden major events had significantly altered the course of geological history, in particular the fossil record. The early widespread acceptance of Lyell's advocacy, although helpful in the solution of many problems, proved inhibiting in the long run and, since Lyell's time, there has been a progressive acceptance 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, mass extinctions in particular. 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; 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): 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 present evidence, what 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 evidence. Scientific truths, in the form of well developed longstanding hypotheses, are still subject to constant scrutiny and revision in the light of further data and are therefore not absolute.
- Theories: coordinated sets of self-consistent hypotheses, each of which has been tested many times and remains a valid explanation of observations. There should not be exceptions (or extremely few) to any theory. 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 encompass and supersede one another.
- Models: idealized simplifications set up as an aid understanding of, and communication about, complex relationships between phenomena and processes, often employed to illustrate working hypotheses. We may draw up actual models based on 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 (e.g. 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 sandflow rate and change of grain size on the shape of the sand sea, the sand dunes and the smaller structures therein, could be predicted. 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 palaeogeographical 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 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 sequence of sediments. This is a generalization and simplification of the observed vertical and lateral relationships of the facies observed in a sequence. 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 commonly fundamental to the establishment of facies schemes and thus form the fundamentals of predictive sedimentology.

Scientists working in a particular field share values, practices (methodology) and beliefs (philosophy), which 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 Hutton and Lyell. The basic principles of a specifically sedimentological paradigm can be traced to the work of H. C. Sorby and J. Walther between 1850 and 1900, but it has become fully developed only 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 and 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. 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 palaeogeographical context.

By contrast, Johannes Walther (1860–1937), in his *Introduction to geology as an 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 the study of sedimentary rocks. In addition, he established a further and most powerful stratigraphical 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 gradationally in a vertical geological column (see §10.3.3). Sadly, Sorby was ahead of his time, and Walther wrote in a language not readily accessible to the Anglo-Saxon world, and it was not until the second half of the twentieth century that their ideas started to be applied to sedimentary successions in a rigorous and widespread manner.

Study techniques

Recommended references

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

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.

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

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.

Pettijohn, F. J. 1975. *Sedimentary rocks* (3rd edn). Rather old now, but a classic in its time and still valuable.

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.

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.

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 in terms 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.

- Can anything in the rocks be detected that suggests that they are bedded?
- Is there other evidence suggesting a sedimentary origin?
- If they do appear to be of sedimentary origin, is there evidence to suggest which is the top and which is the bottom of the succession observed? (N.B. With very few exceptions, this question will be relevant only where it is clear that the rocks are strongly deformed.)
- Are there any features that are characteristic of particular processes or environments of deposition, for example beds with erosive channel-shape bases?
- Are there any patterns of vertical and lateral change in the rocks that might suggest changing processes and thereby an environment of deposition, for example a distinctive vertical or lateral thinning or thickening of the beds?

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

In trying to answer some of the questions above, it can be helpful to think about simple laboratory experiments on sediment deposition and about processes seen 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 over time, layers of sediment somewhat different in character are laid down. The boundaries between such layers may be sharply defined or gradational, depending on the way in which processes or 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 more or less planar bottom and top surfaces, and are very extensive laterally in relation to their thickness. Others are more restricted laterally, possibly reflecting depositional processes that were not uniform (i.e. not constant in space). Depositional units greater than 1 cm thick are known as **beds**; their boundaries, where fairly sharply defined, 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 rather 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. At less than 1 cm thick, 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** or **cross beds**). The general phenomenon of inclined layers is termed **cross lamination** or **cross**

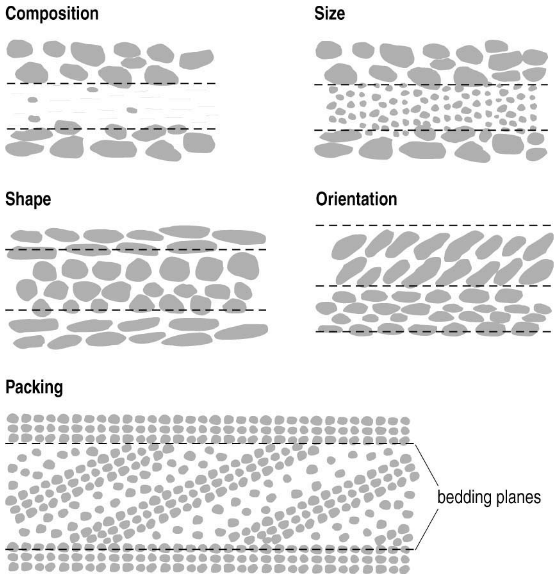

Composition
Size
Shape
Orientation
Packing
bedding planes

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

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).

Many beds and bedsets maintain their thickness for considerable lateral distances, although all eventually thin out or change their nature, either gradationally or suddenly, if 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, the oldest at the base, the most recent at the top, where each bed records a particular set of conditions.

Any sedimentary structure that cross cuts a bedding feature – for example, a channel cutting down into horizontal layers – 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 later

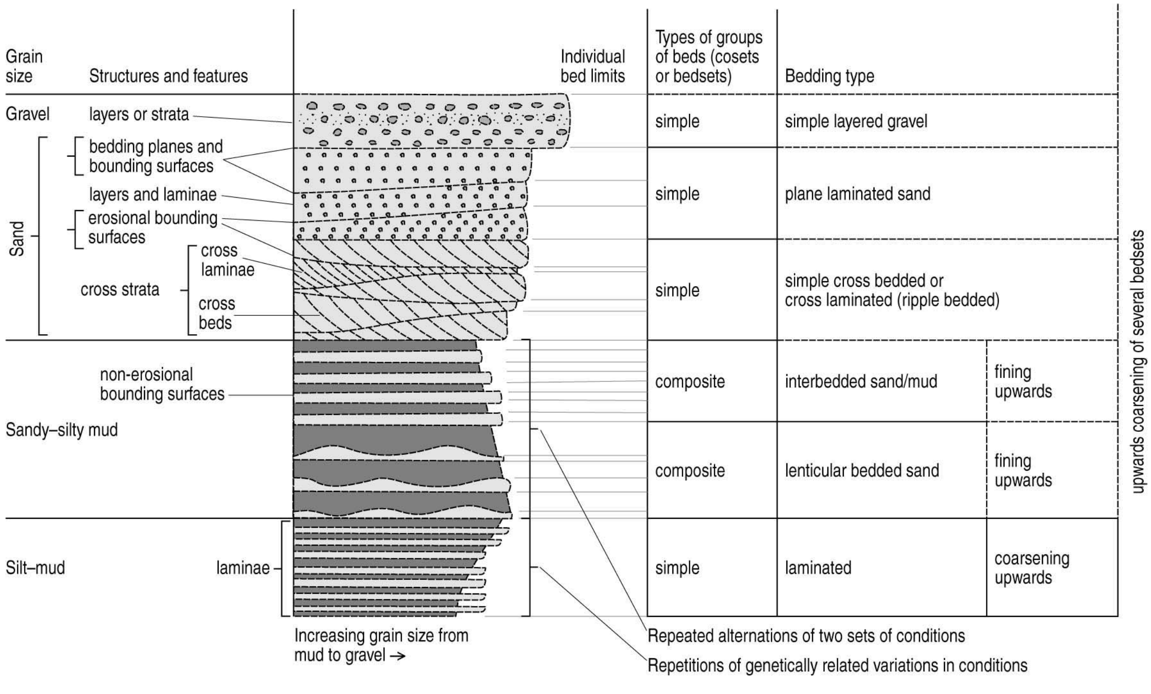

Grain size
Structures and features
Individual bed limits
Types of groups of beds (cosets or bedsets)
Bedding type
Gravel
layers or strata
simple
simple layered gravel
Sand
bedding planes and bounding surfaces
layers and laminae
erosional bounding surfaces
simple
plane laminated sand
cross laminae
cross strata
cross beds
simple
simple cross bedded or cross laminated (ripple bedded)
non-erosional bounding surfaces
composite
interbedded sand/mud
fining upwards
Sandy–silty mud
composite
lenticular bedded sand
fining upwards
Silt–mud
laminae
simple
laminated
coarsening upwards
upwards coarsening of several bedsets
Increasing grain size from mud to gravel →
Repeated alternations of two sets of conditions
Repetitions of genetically related variations in conditions

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

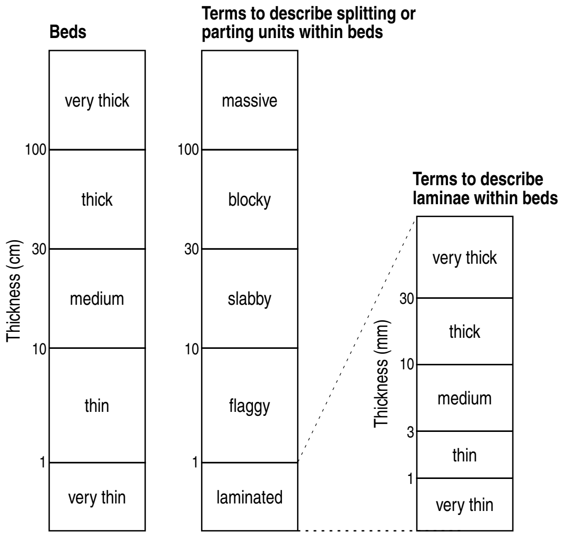


Figure 2.3 Terminology for thickness of beds and the description of units within beds created by splitting or parting, often after weathering. Modified from Ingram (1954), Campbell (1967) and Reineck & Singh (1973).

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. Figure 2.4 shows some bedding features that illustrate some of the points described here.

2.1.3 Basic stratigraphical principles derived from present-day phenomena

Observations of present-day processes and products allow us to establish several stratigraphical principles that help in recognizing beds and in understanding their structural significance in ancient sequences 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 they 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 a sequence.
- Way-up: the tops and bottoms of beds can be recognized by their associated sedimentary structures, which are found either on both the bed-bounding surfaces or within the beds.
- Included fragments: 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 it.
- 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.

2.1.4 Applying stratigraphical 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 the 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 to the top of the sequence. A sequence 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 recognized. If changes in thickness of individual beds across the fault are apparent, this may indicate that the fault was active during deposition. Where beds either side of a discordance do not have adequate bedding character to allow correlation, it may be necessary to study any fossil content in order to establish the nature of the displacement.

The truncation of otherwise parallel beds indicates later erosion. Where the erosion surface is flat and the beds above are concordant with the surface, an **unconformity** (see §2.2.5) may be inferred, an inference that would be enhanced by the presence of fossil groups of significantly different ages above and below the surface.

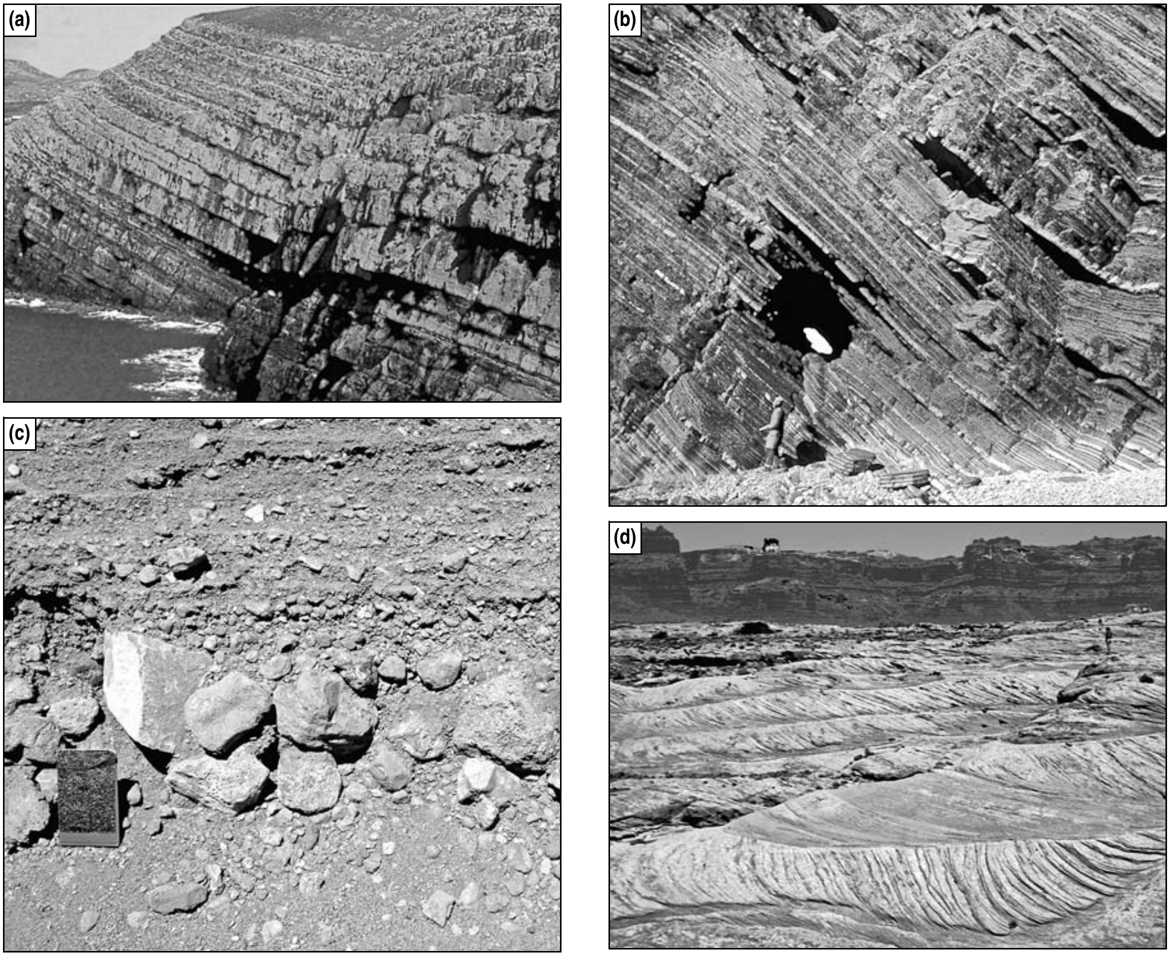


Figure 2.4 Some examples of bedding in various sedimentary successions. **(a)** Thick parallel-sided sandstone beds with thin mudstone interbeds; Hell's Mouth Grits, Cambrian, North Wales. **(b)** Mudstones with interbedded thin parallel-sided and laterally continuous sandstone beds; Aberystwyth Grits, Silurian, west Wales. **(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 in the foreground overlain by horizontally bedded finer-grain sediments in the background; the boundary between the two major units represents a very radical change in depositional conditions; Cedar Mesa Sandstone and Organ Rock Shale, Permian, southern Utah. (Parts (a) and (b) courtesy of Gilbert Kelling)

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 commonly be established from the fossils above and below the discordance.

Where truncation occurs at a surface showing relief, with the beds above and below concordant 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 overview to small-scale scrutiny with a hand lens. At first it is useful to scan exposures from a distance in order to determine 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 tilted only slightly, 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 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 sequence 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 the prominent bedding planes? Are the beds, so defined, of uniform thickness? Do successive beds thicken or become thinner upwards? Are there repeated patterns of bed-thickness change vertically through the succession?
- Are there any suggestions of variation of grain size in the vertical sequence? Are there beds, or groups of beds, that consistently "fine upwards" or "coarsen upwards"?
- Are vertical changes of overall composition suspected, for example, limestone/siltstone/sandstone/conglomerate? Are there patterns to such changes?
- Can the sequence be divided up 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 sequences of beds. From this kind of initial analysis, ideas will commonly emerge about where to start detailed work, where to sample, and where to find key features and surfaces.

2.1.6 Detailed observation and recording of bedding: methodology

Attitude of beds

Detailed work on the outcrop should begin with the 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 few will suffice; if the structure is complex, then it may be necessary to take more measurements, taking care to 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 reorientating measurements of inclination and alignment of sedimentary structures recorded in the field. Through these observations, the structures can be restored (on a stereogram) 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.

While 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, for example, the boundary between sediment and crystalline infill of a brachiopod shell half-filled by sediment (e.g. see Fig. 8.6). Such a surface will commonly coincide in attitude with bedding, but in some cases it 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 the geopetal surface.

Successions

A basic 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 succession, selected to document any lateral variability suspected from preliminary observation. This enables local variability to be distinguished from any regional variability

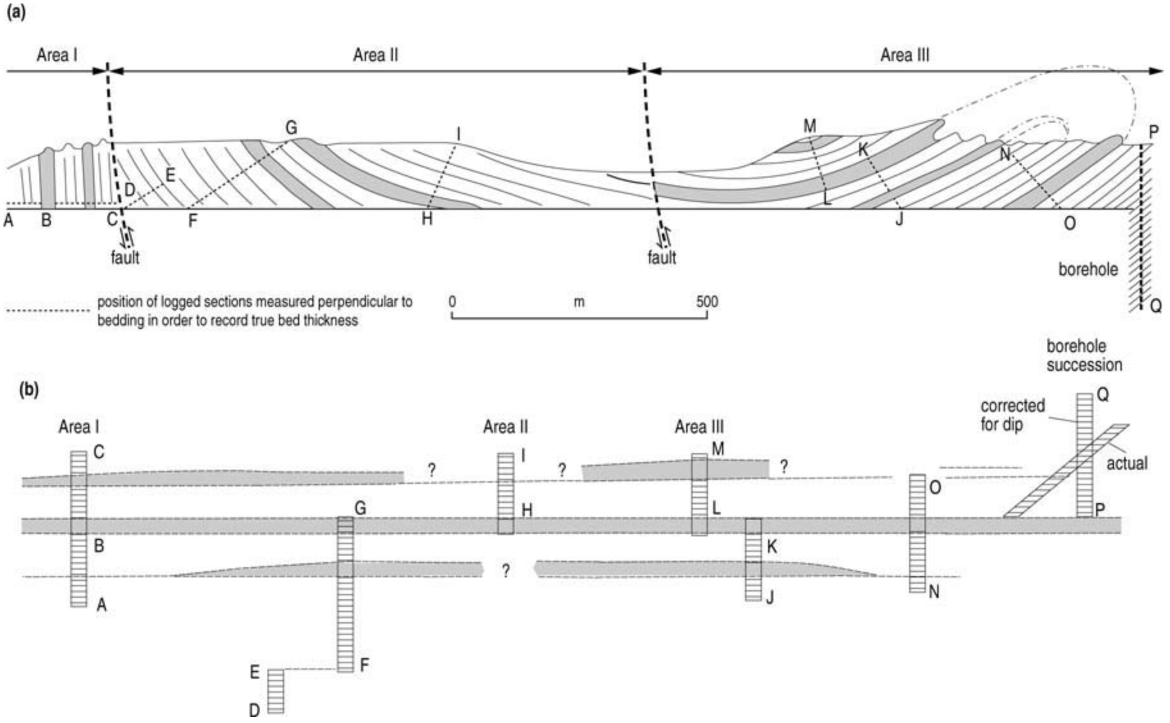


Figure 2.5 Illustration of a strategy for logging a sedimentary succession that has been tectonically deformed. **(a)** Sketch of cliff section, showing deformed beds. Note that the section has been divided into three areas, each bounded by a major fault. Also note the overturned strata at the right-hand 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.

and it 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 because of restricted accessibility, make sure that any oblique measurements are corrected by simple trigonometry.

The division of the outcrop 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, but an effort should always be made to apply consistent criteria and to sample in a representative way (Fig. 2.5). It is important to employ systematic working procedures based on the stratigraphical 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. Methodical habits of measuring and recording generally enhance powers of observation and make pattern recognition easier; the search for patterns within field 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 has to be distinguished from tectonically induced cleavage, from joints and faults, and from colour banding as a result of diagenesis (i.e. 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 deny such an initial impression (Figs 2.1, 2.2, 2.5, 2.6).

The approach illustrated in Figure 2.5 relates to a method of working that is essentially one-dimensional: the logging of a vertical section. However, it is often helpful to photograph a group of beds in two dimensions from a series of positions equidistant from and 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 a one-dimensional approach is inadequate to characterize the succession fully. Where the exposure has promontories and recesses, or is covered by vegetation, it is important to draw scaled diagrams that generalize the geometry of the outcrop.

Thickness of sedimentary units

Measurement of thicknesses of beds should take place near to the vertical line (i.e. normal to bedding) selected. This line is often determined by access to the exposure, as in a stream bed. In other outcrops the decision may be either selective or simply arbitrary. Limits of units will normally be clear where bounding surfaces or bedding planes are sharp, but subjective decisions must be made 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 that there are no rapid lateral variations in lithology or thickness.

Lateral variations

The 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) convergence and merging of their upper bedding

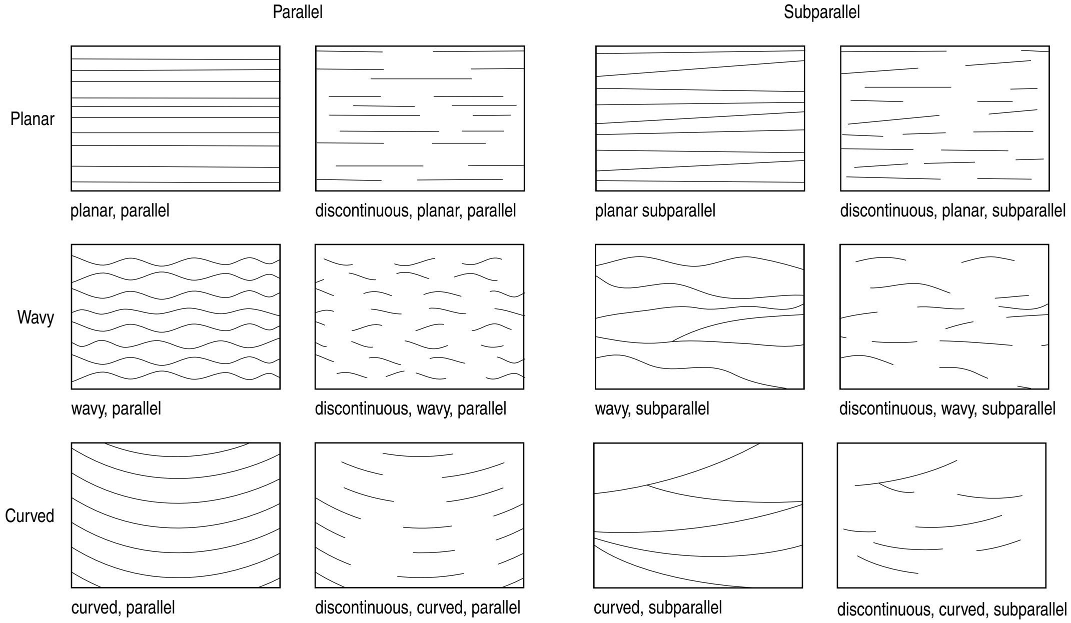

Parallel
Subparallel
Planar
planar, parallel
discontinuous, planar, parallel
planar subparallel
discontinuous, planar, subparallel
Wavy
wavy, parallel
discontinuous, wavy, parallel
wavy, subparallel
discontinuous, wavy, subparallel
Curved
curved, parallel
discontinuous, curved, parallel
curved, subparallel
discontinuous, curved, subparallel

Figure 2.6 Useful bedding-lamination terminology; modified after Campbell (1967) and Reineck & Singh (1973).

surface and sole, when they may be termed **lenticular**

(b) lateral gradation of the composition of the bed so that the bounding surfaces die away

(c) being intersected by a cross-cutting feature such as a channel

(d) 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 (e.g. a channel margin or an organic mound or bioherm), although the angle of drape 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. 2.6).

Features within a bed or bedset

Vertical variations within individual beds are typically attributable to changing grain size, composition, texture or internal structure (see Fig. 2.1). Beds and bedsets may be homogeneous or heterogeneous, rhythmic, or 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. Some beds are heterolithic (composed of different rocks) as a result of sorting into repeated interlaminations of sediment of contrasting composition or grain size (e.g. silt and sand, silt and mud; see 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 common in some beds deposited by episodic decelerating currents. In addition to variations in grain size, composition and texture, beds are typically characterized internally by the scale and style of depositional lamination (Fig. 2.6) 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 gradational or sharp (Fig. 2.4). 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, which are sometimes referred to as **bounding surfaces**, 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 periods of time, and protracted non-deposition may see 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 normal 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 that result from various styles of accumulation (e.g. progradation, migration and infill) at a wide range of scales (Fig. 2.7).

Patterns in sedimentary successions

Patterns of vertical change in groups of beds and bedsets may be identified where thickness, grain size, composition or sedimentary structure changes systematically (e.g. see Fig. 2.2). The following kinds of

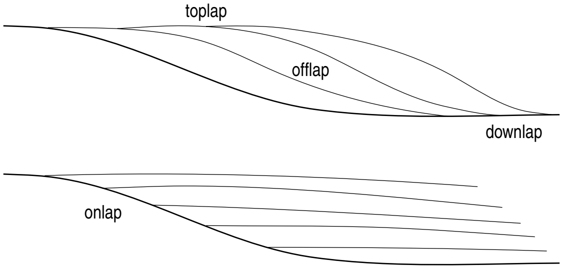


Figure 2.7 The terminology of discordant bedding relationships. These terms are used over a wide range of scales from small outcrop to seismic section.

pattern of vertical change are commonly recorded:

- Grain size becomes progressively finer upwards, from coarse sandstones at the base to siltstones and mudstones (as in some meandering-river successions).
- Grain size coarsens upwards, from shale to coarse sandstone (as in some deltaic successions).
- Systematic and repeating patterns of vertical lithological change such as shale–sandstone–coal or limestone–shale–sandstone, etc. (as in other kinds of deltaic successions).
- Patterns involving successive units of carbonate–sulphate, hydrous and anhydrous sulphate, sulphate and halides (as in chemically precipitated evaporite successions).
- Patterns of bed thickness change (as in thickening- or thinning-upwards successions).

These types of patterns may typically 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 describe sedimentary successions in the field or in borehole cores requires not only organized working practices but also the acquisition and application of a new terminology. That can be quite daunting at first, but practice and discussion with others can quickly clarify issues and lead to the necessary confidence. If outcrops or cores are not readily available, it is often possible to make good progress by describing photographs of outcrops (e.g. Fig. 2.8). 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 that are most diagnostic of depositional conditions.

It is important to remember that 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 understanding depositional processes to deducing the depositional environments.

2.2 The significance of bedding

2.2.1 Introduction

Individual beds are deposited under either essentially constant physical and chemical conditions or by systematic changes of process. 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. However, other contacts record gradational changes and probably reflect a more gradual change in conditions. Laminae typically result from minor fluctuations in statistically 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 (see 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 for accumulation to occur. Periods of erosion may be long enough for major time gaps (represented by unconformities) to be present in the sequence, 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 related to such features may punctuate the resulting sequence and are a natural part of the overall accumulation. Assigning time to such surfaces is seldom easy and it is, of course, often 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 topographical lows (e.g. the bases of channels).

2.2.2 Basic processes of sedimentation

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

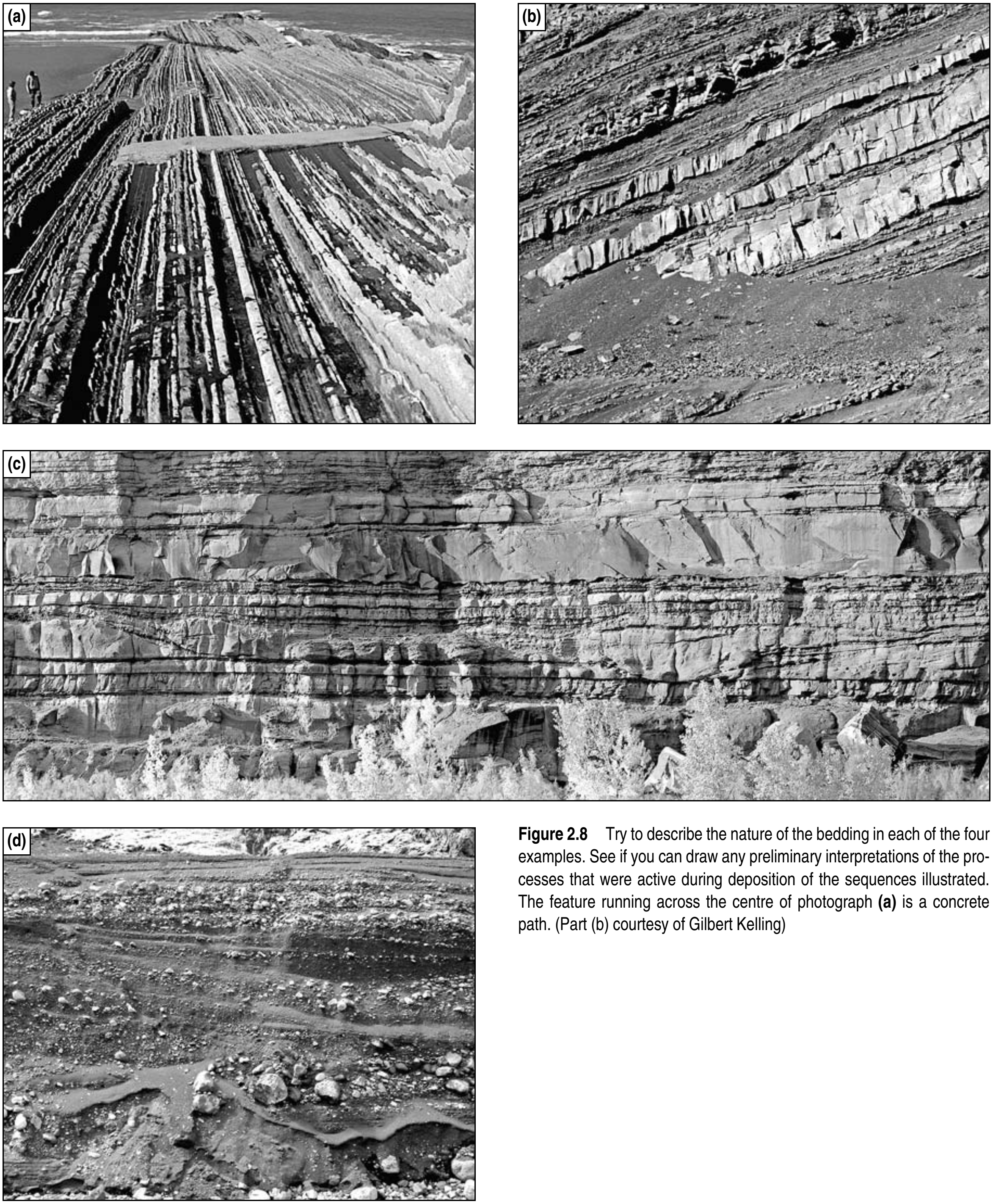


Figure 2.8 Try to describe the nature of the bedding in each of the four examples. See if you can draw any preliminary interpretations of the processes that were active during deposition of the sequences illustrated. The feature running across the centre of photograph **(a)** is a concrete path. (Part (b) courtesy of Gilbert Kelling)

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 mudflows. Others result from sediment transported by wind. The nature and intensity of the transporting process depend on properties of both sediment and fluid. Grain size and grain density are crucial properties of the sediment. The density, viscosity of the fluid and the velocity of the flow, the strength of waves and the depth of flow are also important. A change in any of these parameters can alter the nature of the deposit, initiate a new bed, or trigger a phase of erosion.

Explosive volcanic eruptions can throw a great deal of hot and possibly molten material into the atmosphere and may lead to a whole suite of complex processes that include ashfalls 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 respectively (e.g. Fig. 2.9).

Chemical processes

Much material, particularly in the sea and in some closed lakes (i.e. those with no outlet), is carried 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, freshwater influx.

Biological processes

Most of the calcium carbonate, which makes up limestones and present-day carbonate sediments, results from the activities of organisms (both animal and plant) that precipitate calcium carbonate as part of their metabolic processes. Other organisms secrete silica (e.g. sponges and diatoms) or phosphates (e.g. vertebrates), which may also contribute to sediments. Changes in the dominant organisms may produce changes in the sediment, leading 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 or chemical processes (or both). 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.

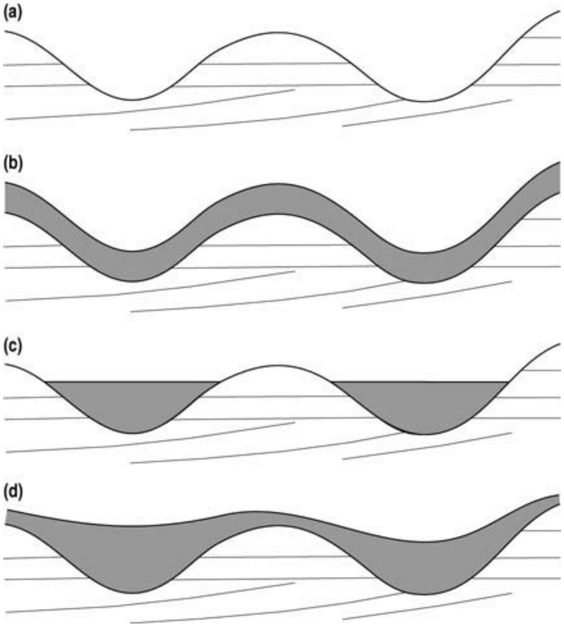


Figure 2.9 Variations in gross bed thicknesses produced by different types of volcanic processes. **(a)** Pre-eruptive rock succession and topography. **(b)** Pyroclastic airfall deposits mantle topography because material falls vertically and evenly. **(c)** Pyroclastic flow deposits infill the lower part of the topography because of the strong gravitational control on the flow. **(d)** Pyroclastic surge deposits thicken into the topographical lows, reflecting some gravitational influence. Similar bedding geometries can be produced by other (non-volcaniclastic) types of sediment debris and gravity flows. (After Wright et al. 1980)

Diagenetic processes

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

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. The study of bedding, therefore, should aim to interpret 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 we 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 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 attributable 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 a setting and give rise to different deposits. If it were feasible to map the distribution of these deposits over a long 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 in a trench dug at the site a bed of mud overlying a bed of sand. 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 attributable to temporal fluctuations

In contrast to the steady-state lateral migration mechanism outlined above, many beds reflect changes in time, because of 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, recording periods of stormy and calm weather. In both of these examples the environment is not changing, but rather the vertical sequence records the natural variation 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-grain sediment accumulates 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. However, in the resultant sedimentary sequence the sandbeds 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

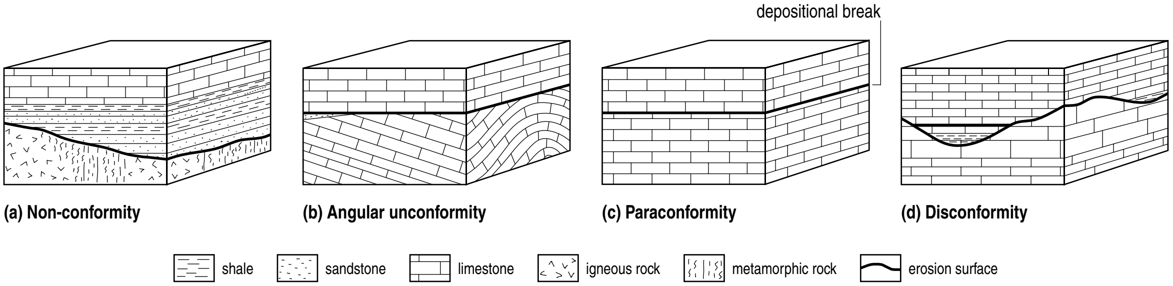
The importance of the boundaries between beds has been emphasized already. Some are gradational and others sharp. Recording the nature of boundaries between beds is as important as recording the features of the beds themselves, if observations are to be used for environmental reconstruction. A gradational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

Finally, 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 rotted but before they had fallen, and beds deposited from floods, dust storms and ash clouds. These beds often form part of thick uniform sequences, which, when dated, appear to have accumulated over considerable periods of geological time, often millions of years. This paradox of beds deposited relatively swiftly, making up sequences representing many millions of years, has led some geologists to coin intuitive statements such as “98 per cent of geological time must be represented by the bedding planes”, thus highlighting the importance of trying to recognize significant time gaps in seemingly continuous sedimentary sequences. The nature and distribution of certain trace fossils and of certain 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 stratigraphical sequence 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. 2.10). 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**). Sometimes, however, an angular unconformity may be wrongly suspected where flat bedding overlies a very large-scale set of cross bedding. In such cases, checking 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 be satisfactorily recognized 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 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 redbeds, lack the means to be easily analyzed in this way, although more elaborate methods such as magnetostratigraphy and chemostratigraphy may resolve this in some cases.

The largest erosional structures – for example, fossil channels cut by rivers into alluvial fans or floodplains, or fossil tidal channels cut into mudflats – may be confused with a true disconformity (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 any 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 discussion of several working hypotheses, are therefore very important. Breaks in the record that represent time gaps so short that they are not highlighted by evolutionary changes in the fossils have been called **diastems** and it is sensible to regard potential disconformities and paraconformities as diastems until proved otherwise.



(a) Non-conformity **(b) Angular unconformity** **(c) Paraconformity** **(d) Disconformity**

Figure 2.10 Four types of unconformity: **(a)** non-conformity, **(b)** angular unconformity, **(c)** paraconformity, **(d)** disconformity. Note that the terminology relating to discordant bedding relationships (e.g. onlap, apparent truncation) can be applied to beds above and below some unconformities. (After Dunbar & Rogers 1957)

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 often be difficult to discern. 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 often provide excellent exposures.

Laboratory experience

Simple physical experiments can be devised to investigate bedding development and the form of bed boundaries. Thoroughly mix together 100–200 g of loose sand with water in a jar, pour the mixture into a 1 L measuring cylinder and allow it to 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.

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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 sediments, it is necessary to have a basic understanding of the physical properties and mechanics of the fluids that erode, transport and deposit the sediment. 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 sediment/water mixtures, usually on a slope. During gravity emplacement, water may be important as a lubricant or as an agent that acts to support the moving grains. The moving mass of grains, with or without water, typically behaves as if it were 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, as both individuals and 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.

Therefore, this chapter examines some of the properties of fluids and plastics, and shows how these 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.

The 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 and observations of this type are helpful, but 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 virtually all 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, flow 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 that 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 $1650\,\mathrm{kg\,m^{-3}}$ compared with $2650\,\mathrm{kg\,m^{-3}}$ in air, a difference that strongly influences the ability of the different fluids to move the grains.

The viscosity (μ) describes the ability of the fluid to flow. It is defined as the ratio of the shear stress (τ shearing force/unit area) to the rate of deformation ($\mathrm{d}u/\mathrm{d}y$) sustained by that shear across the fluid:

$$\mu = \frac{\tau}{\mathrm{d}u/\mathrm{d}y} \tag{3.1}$$

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

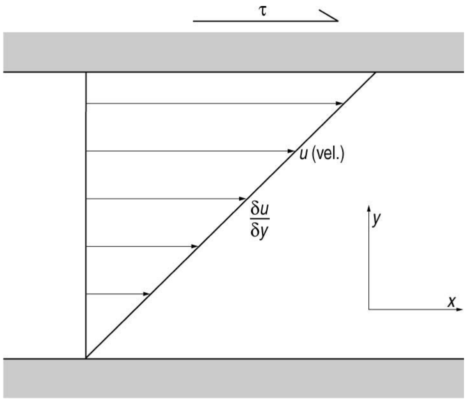


Figure 3.1 Definition diagram for viscosity. Two rigid parallel plates enclose the fluid. A shear stress (τ), acting parallel to the sheets, sets up the steady-state velocity profile shown by the inclined line. The length of the arrows is proportional to velocity (*u*) relative to the lower plate.

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 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):

$$\nu = \frac{\mu}{\rho_f} \qquad (3.2)$$

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 down stream 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).

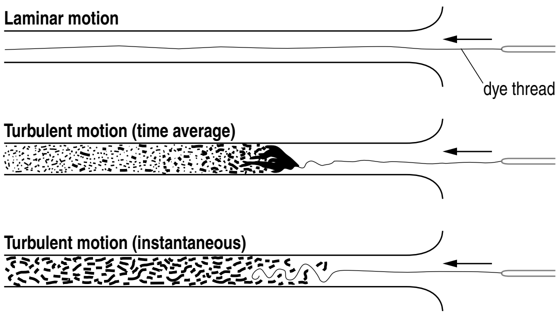

Laminar motion
dye thread
Turbulent motion (time average)
Turbulent motion (instantaneous)

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 a 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 dimensionless number 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):

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

where $\bar{U}$ is the mean velocity of the flow and L is some length that 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-down models. The transition from laminar to turbulent flow takes place at a critical value of the Reynolds number, which 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 that accounts for the extra shear needed to maintain turbulence and $\bar{u}$ is the time-averaged velocity. However, eddy viscosity is not 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 shape different 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 attributable 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 through 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).

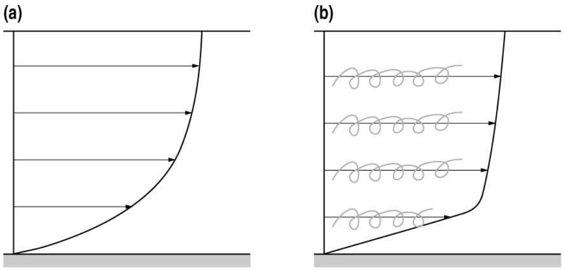


Figure 3.3 Vertical velocity profiles for **(a)** laminar and **(b)** turbulent flows in an open channel of great width. In the profile for turbulent flow, the velocities are time-averaged values and only close to the bed does a near-laminar pattern of movement occur.

Where currents are strong enough to move sediments, flow will almost always be turbulent, and that 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 upwards energy flux. Some of this activity can be related to upwards-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 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 or detach from the boundary at a **separation point** or **line**

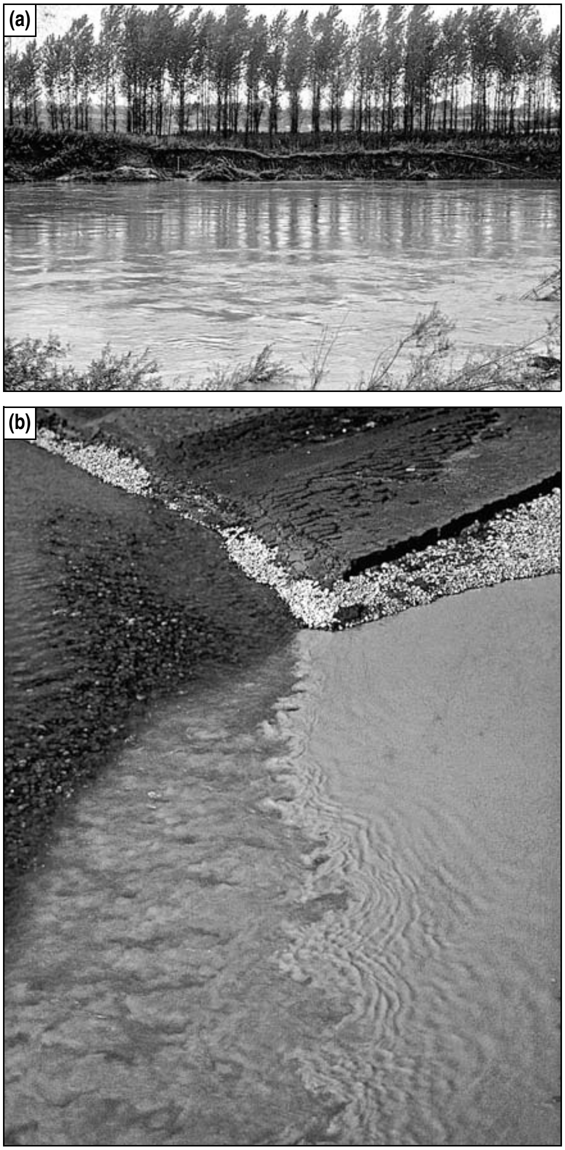


Figure 3.4 **(a)** "Boils" on the surface of a rapidly flowing river, showing the upwards movement of turbulent cells to the surface. **(b)** Small-scale turbulent mixing between clear and turbid water at the junction of two streams. Note the discrete cells of water, which move in directions highly divergent from the mean flow direction.

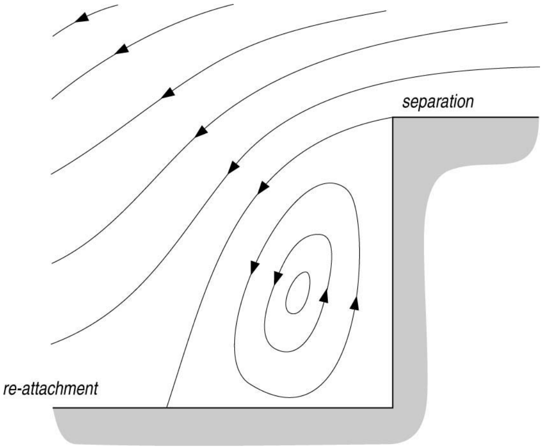


Figure 3.5 Separation and re-attachment of flow at a negative step on the perimeter of a flow. A cell of rotating fluid is trapped within the separation eddy. (After Allen 1968)

and to re-attach itself down stream at a **re-attachment 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 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 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 re-attaches to the bed. See if you can determine the volumes and shapes of eddies, and describe the patterns of water movement within them (compare your observations with Figs 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 structures and depositional bedforms seen on present-day 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**,

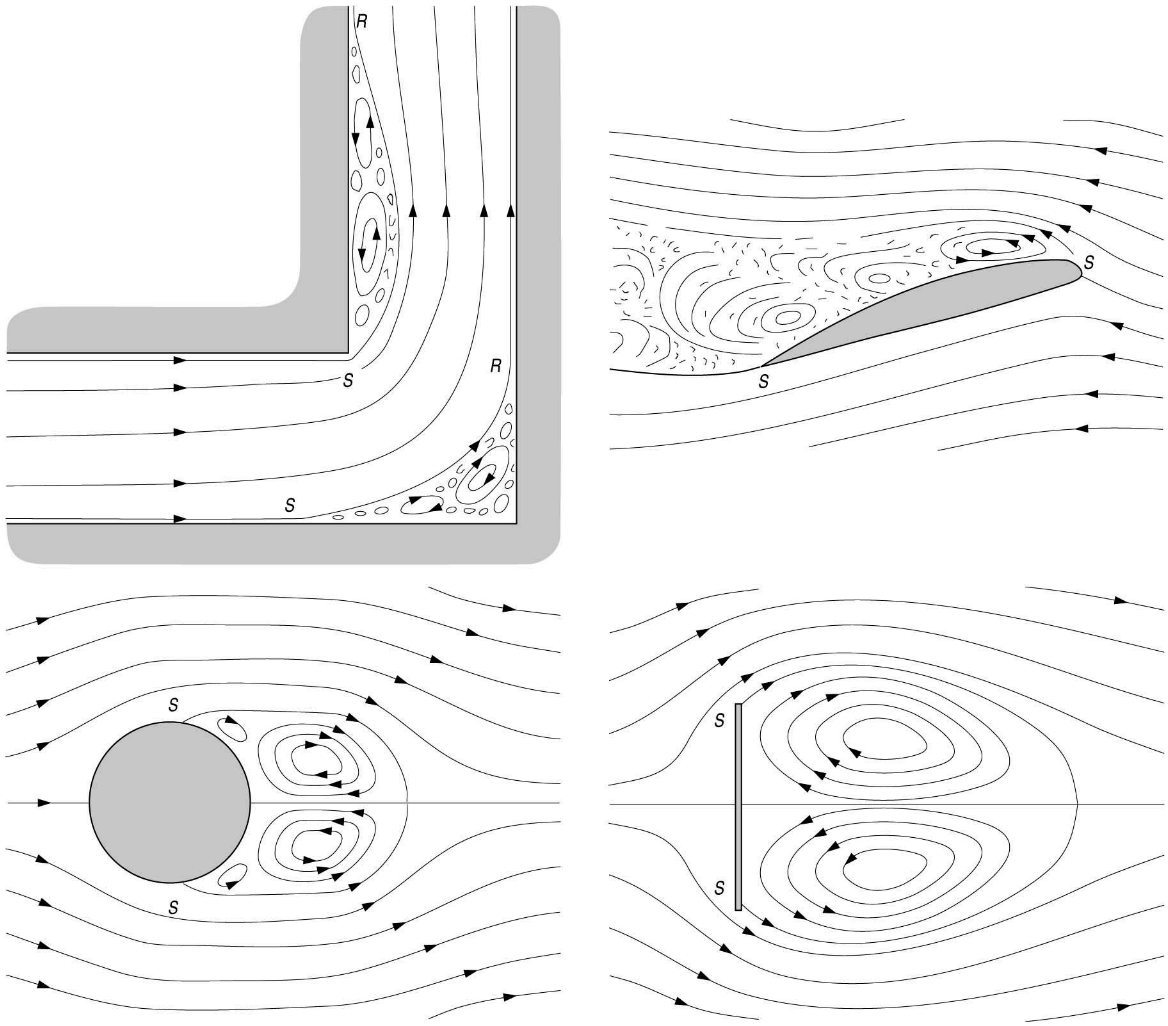


Figure 3.6 Examples of patterns of flow separation (*S*) and re-attachment (*R*) around obstacles and features of different shapes (after Allen 1968).

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 relates to 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 therefore the critical factor. The second 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).

When 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

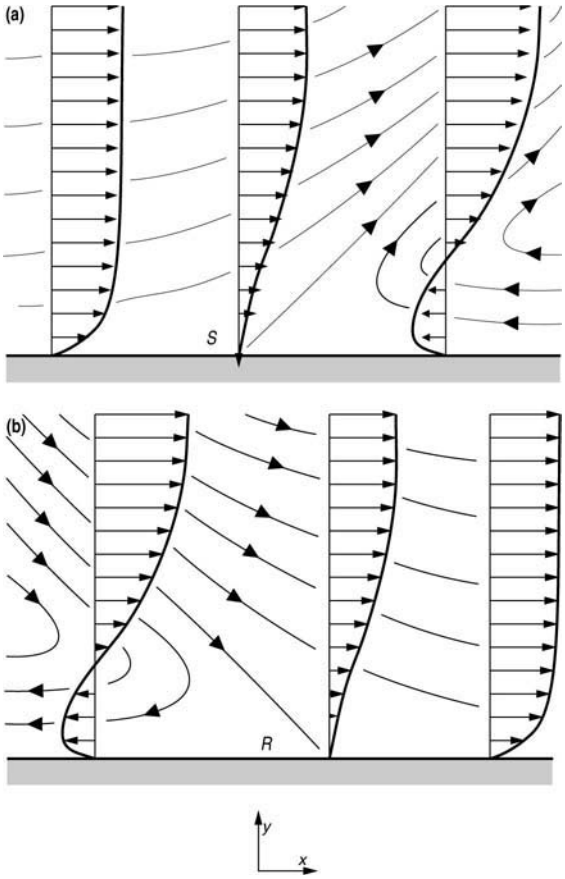


Figure 3.7 Flow velocity profiles associated with **(a)** flow separation and **(b)** flow re-attachment (modified after Allen 1982).

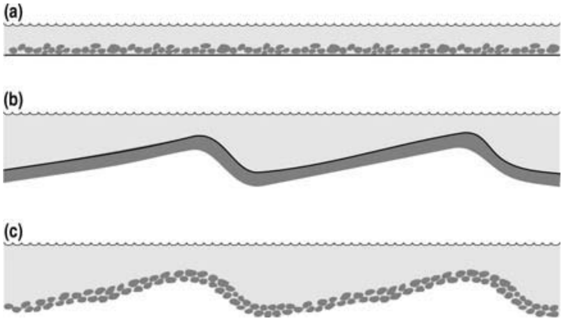


Figure 3.8 The distinction between grain roughness and form roughness: **(a)** purely grain roughness; **(b)** purely form roughness over a smooth artificial bedform; **(c)** grain and form roughness over a natural bedform.

exist. The sub-layer is therefore important only with fine bed material. It has now been shown 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 the formation of ripples, in the production of some current lineation, and in the development of lamination in fine-grain sediments deposited from suspension (see Chs 5, 6).

3.2.5 Boundary shear stress

The behaviour of sediment on a bed below a flow is determined mainly 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.

It is clear that in flowing water there will be a velocity

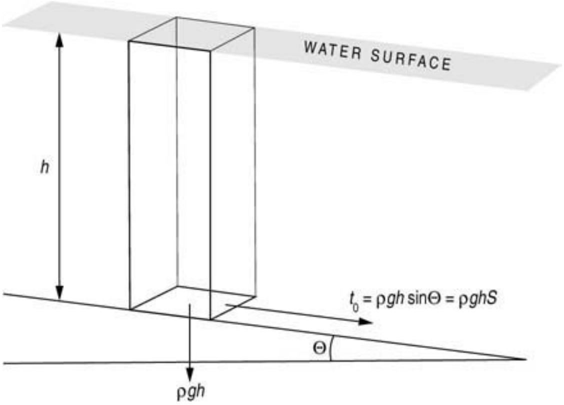


Figure 3.9 Definition diagram for the calculation of bed shear stress for water flowing down slope as an open-channel flow. Forces acting on unit area of the bed are indicated.

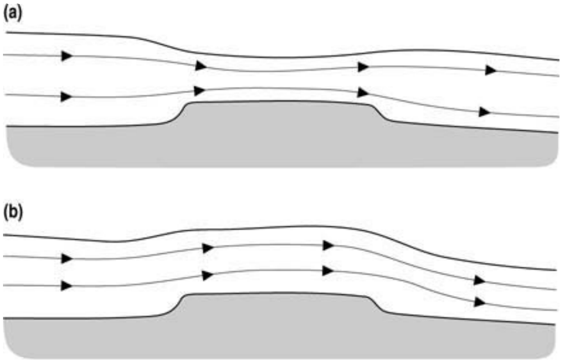


Figure 3.10 The form of the water surface and of time-averaged flow lines over a positive obstruction on the bed under conditions of **(a)** tranquil and **(b)** rapid flow.

above which it will not be possible for waves to move up stream. 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:

$$Fr = \bar{U}/\sqrt{gh} \tag{3.6}$$

For $Fr > 1$, we have rapid-flow conditions, in which waves cannot be propagated up stream, 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 then 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 onto a flat, smooth, horizontal surface, as from a water tap onto 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 sandbed (see Ch. 6).

3.3 Waves

So far, we have considered unidirectional currents in fluids, but, as mentioned in the previous section, water in particular also transfers energy through the movement of waves. 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 their length, height, speed and frequency, and the water depth. How do these properties relate to one another?

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:

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

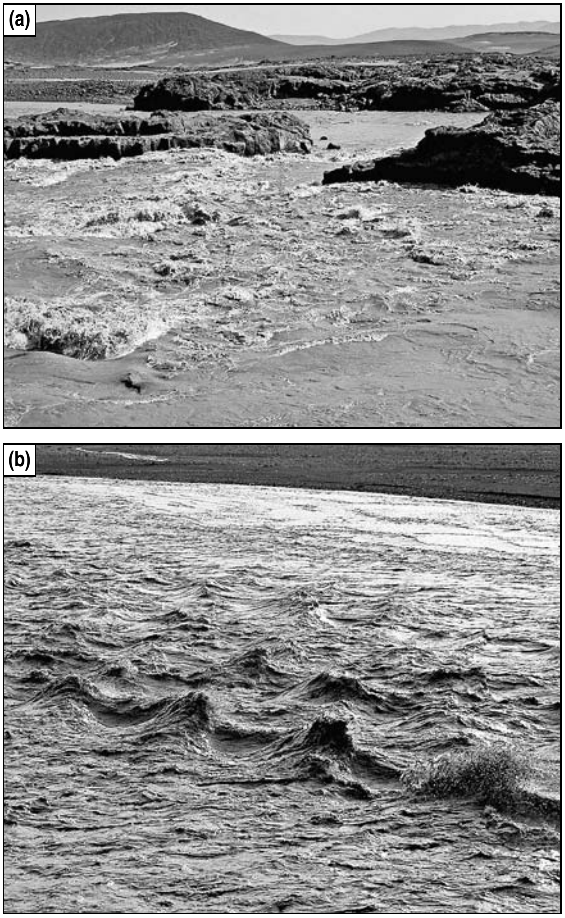


Figure 3.11 Upstream-facing breaking waves marking the position of hydraulic jumps in the flow regime: **(a)** Jökulsá á Fjöllum, northern Iceland; **(b)** Myrdalssandar, southern Iceland.

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 wave 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 deepwater 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:

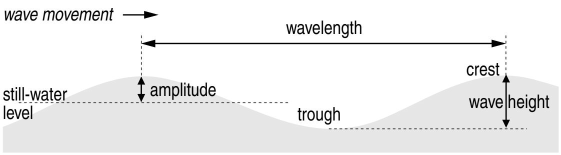


Figure 3.12 Definition diagram for the main physical properties of simple water-surface waves.

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

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

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 deepwater 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. 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 motion changes with depth and position in the water column. With deepwater 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 farther off shore, and waves move as **surges**, often over long

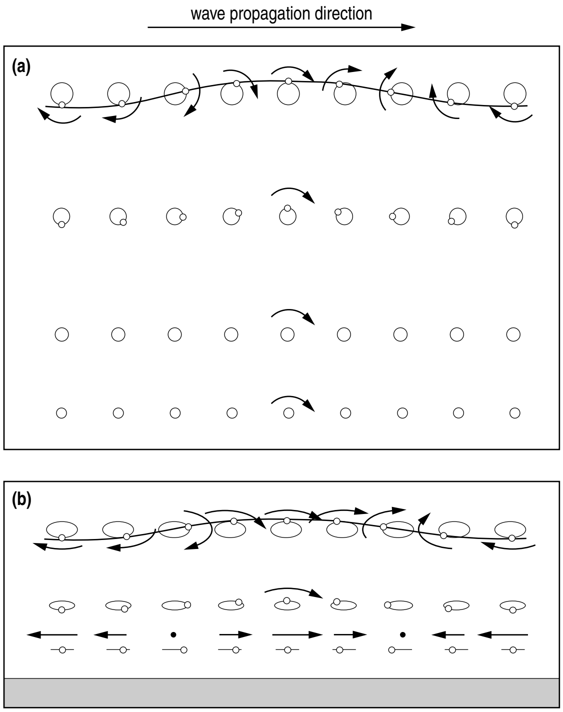


Figure 3.13 The pattern of movement of individual water particles associated with the passage of a surface wave for **(a)** deepwater waves and **(b)** shallow-water waves.

distances, and the strength of the backwash is lower.

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 1 mm 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 4.0–0.068 mm (1/16 mm). 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

$$\varphi = -\log_2 (\textit{grain diameter in mm}) \qquad (3.10)$$

Using this scale, 1 mm has a value of 0φ, $2\,\text{mm} = -1\varphi$, $4\,\text{mm} = -2\varphi$, $0.5\,\text{mm} = 1\varphi$, $0.25\,\text{mm} = 2\varphi$, and so on. It is also possible to express mean, sorting and skewness 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 obtaining 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, while 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

(1 L 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.

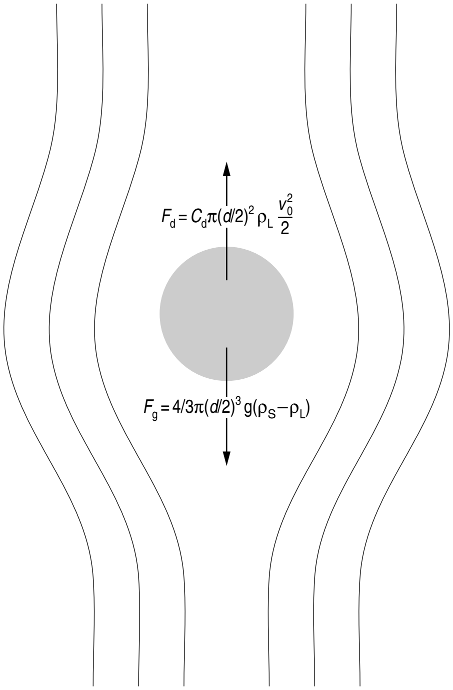

$F_d = C_d\pi(d/2)^2\rho_L\frac{v_0^2}{2}$
$F_g = 4/3\pi(d/2)^3 g(\rho_S-\rho_L)$

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, or both.

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:

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

and so

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

where ρ_s and ρ_l are the density of the solid and the liquid, respectively, and C_d is a drag coefficient that depends upon particle shape and particle Reynolds number Re (= V_0d/ν), 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 is valid only for particles of small grain size (low Reynolds number), for which 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 apply only 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 it goes some way towards resolving 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 the 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 scope for interpreting processes and environments of deposition. Work on cumulative 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).

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.

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 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 on 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:

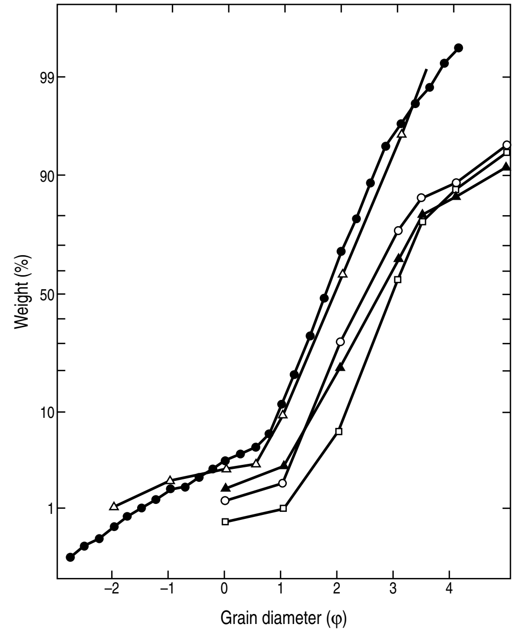


Figure 3.15 Cumulative grain-size curves plotted on a probability scale show straight-line segments thought to be attributable to suspended sediment, intermittently suspended bedload and traction load. Symbols relate to particular samples of sieved sand. (Modified after Middleton 1976)

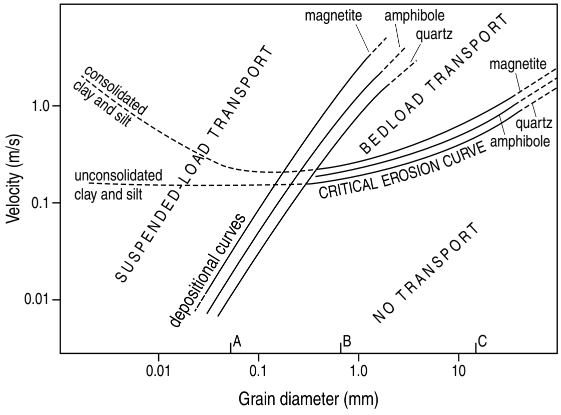


Figure 3.16 Hjulström–Sundborg plot of grain size against flow velocity 1 m above the bed for initiation of movement of particles of different densities. The nature of the transport mechanism, once erosion has taken place, is also shown (after Sundborg 1956 and Ljunggren & Sundborg 1968). As an example, quartz grains of grain size B would begin to move as bedload at the critical erosion velocity of $0.12\,\mathrm{m\,s^{-1}}$ and would go into suspension at a velocity of $1\,\mathrm{m\,s^{-1}}$. 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.

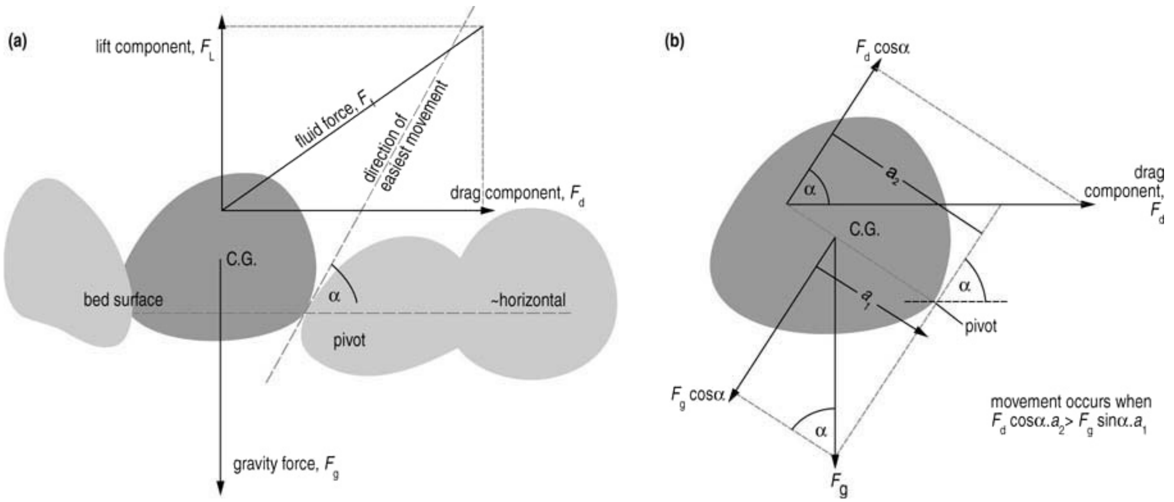


Figure 3.17 **(a)** The forces acting on a particle in a bed of similar particles when subjected to the shear of an overriding current. **(b)** The movement forces that must be balanced for incipient motion; movement begins when $F_d a_2 \cos\alpha > F_g a_1 \sin\alpha$. (After Middleton & Southard 1977, 1984)

(a) the weight of the particle
(b) frictional forces between adjacent particles
(c) hydraulic lift forces
(d) the tangential shear couple (drag).

Types (a) and (b) resist motion, whereas (c) and (d) 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 important only 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 upwards-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.

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

$$F_d a_2 \cos\alpha \geq F_g a_1 \sin\alpha \tag{3.16}$$

In natural settings many other factors complicate matters. First, natural grains are not all equidimensional, nor are they of uniform size. Grains of irregular shape have a stability that varies with their orientation with respect to the current. Naturally deposited grains tend to rest where drag and lift forces are at a minimum. Flattened grains are most stable when they are inclined up stream 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 often relates to the overall grain-size distribution. Well sorted sediments show a pattern of behaviour more closely related to the ideal than do poorly sorted grain populations, where large grains are exposed only partly to the prevailing shear stress because of partial burial in finer-grain components. 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 (see Fig. 3.16). For grains over about 0.6 mm diameter there is a gradual increase in critical shear stress and velocity; 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, sometimes even going beyond the vertical. Commonly, this leads to their 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 sub-layer (see §3.2.4), will also help to determine 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 that would not experience movement under time-averaged conditions. Effects of this type make it difficult to define the onset of grain movement satisfactorily.

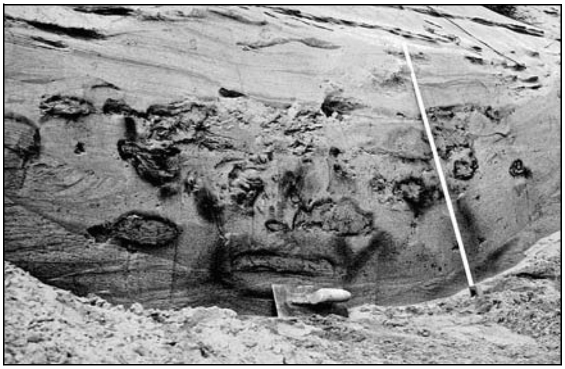


Figure 3.18 Example of cohesive behaviour: mudball clasts in a channel infill.

The streaks of high-velocity flow in the viscous sub-layer may also localize the initial grain movement in fine-grain sediment and help to throw fine sediment into suspension as the streaks erupt into the overlying 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 intraclasts), rather than of individual grains (Fig. 3.18). In addition, material already moving in the flow may help to promote or accelerate erosion by means of a "sandblast" (abrasion) effect when moving grains hit the bed.

The onset of grain movement because of wind is broadly similar to that with water. However, the occurrence of particles already in motion lowers the critical erosion velocity quite significantly. A sandbed that is stable in winds of a subcritical speed can be set in motion if a few grains are thrown (seeded) onto 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**, at which 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 the fluid turbulence of the flow. The sediment moves at roughly the same rate as the fluid, and the movement results from a balance between downwards gravitational forces on the grains and upwards forces derived from the fluid turbulence. This suggests that the turbulent flow has a net upwards 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 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 (see 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 be moved in suspension more readily. However, this process is self-limiting, as increased sediment concentration has a damping effect upon the turbulence because of the increase in viscosity until conditions approach those of a mudflow (see §3.7.1).

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 immersed weights 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 down stream 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 starting grain movement.

Where a grain collides with the bed and does not bounce, its kinetic energy may be dispersed among 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 per cent 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 size similar to itself. Rolling is additionally influenced by grain shape, such that rounded equant-shape clasts will roll more easily than flat disc-shape 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 intergranular collisions are a dominant process. Such layers have been termed **traction carpets** or **modified grainflows**, and they can be important beneath flows carrying much sand-size material in suspension.

3.7 Sediment gravity flows

Mixtures of sediment and water, under appropriate conditions, are able to 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 interparticle 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

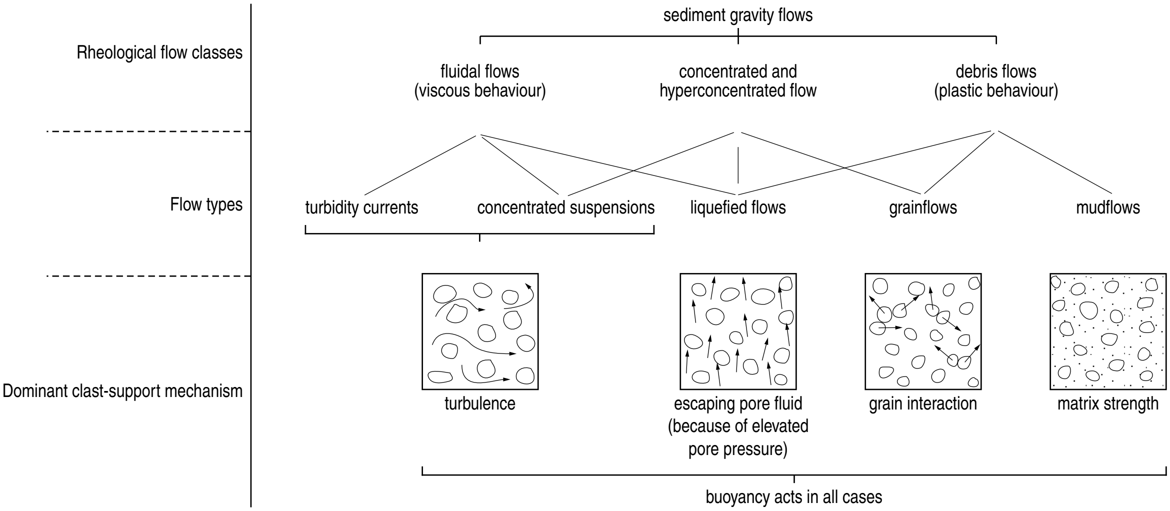

sediment gravity flows
Rheological flow classes
fluidal flows (viscous behaviour)
concentrated and hyperconcentrated flow
debris flows (plastic behaviour)
Flow types
turbidity currents
concentrated suspensions
liquefied flows
grainflows
mudflows
Dominant clast-support mechanism
turbulence
escaping pore fluid (because of elevated pore pressure)
grain interaction
matrix strength
buoyancy acts in all cases

Figure 3.19 Various types of sediment gravity flow, where different types of interaction between the water and the sedimentary particles create the mobility necessary for movement (Middleton & Hampton 1973).

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 between fluid and grains (Fig. 3.19). The complexity of the flow mechanisms, and the problems of interpreting ancient deposits in terms of those mechanisms, 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.

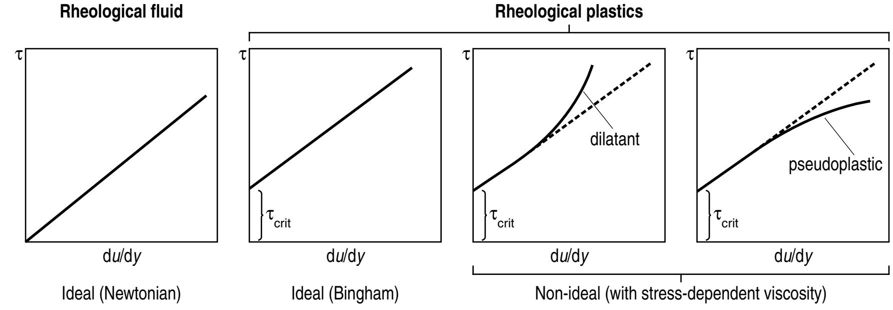
3.7.1 Particle support mechanisms

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 also allows the mixture to deform internally by shearing as the flow moves.

Mixtures of sediment and water that are dominated by viscous (cohesive) 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 which flow under minimal shear (Fig. 3.20). Movement of cohesive mixtures of sediment and water is usually initiated by the reduction of yield strength through the addition of water

Figure 3.20 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 & Steel 1984)



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 dewatering, there will be a point at which the 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.

Grain-to-grain support

Where a muddy matrix is mostly absent, larger particles may move as a highly concentrated flow only if they are kept apart and supported by intergranular collision. Such non-cohesive flow requires an applied shear that can overcome the initial intergranular friction. Pouring granulated sugar or tipping dry sand from a bucket or the back of a truck are everyday examples of this behaviour. The particles are kept apart and hence are free to move relative to one another by vigorous interparticle collisions, which create a so-called **dispersive pressure**, whose magnitude will depend on that 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.

Fluid turbulence

Support of sedimentary particles by the upwards components of fluid turbulence was 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, intergranular collisions become more important, hindered settling occurs and the 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 weight of the particle. For low-concentration aqueous flows, water will provide the main buoyant uplift, whereas in a high-concentration flow the 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 3.21. On this basis, it is possible to deal with the spectrum of flow types under three broad headings, although 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, in order to distinguish them from non-cohesive debris flows, which are the hyperconcentrated grainflows of this chapter. Debris flows occur in both sub-aerial and sub-aqueous settings. In sub-aerial settings they are commonly initiated on steep slopes as a result of high water saturation following heavy rain or snowmelt. 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 may even protrude from the top surface (Fig. 3.22a).

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

$$\frac{du}{dy} = \frac{1}{\mu}(\tau - \tau_{\text{crit}})^k \qquad (3.17)$$

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; their effects are illustrated in Figure 3.20.

In sub-aerial settings, debris flows may be diluted by the addition of more surface run-off, so that viscosity reduces to the point where **grainflows** (intergranular

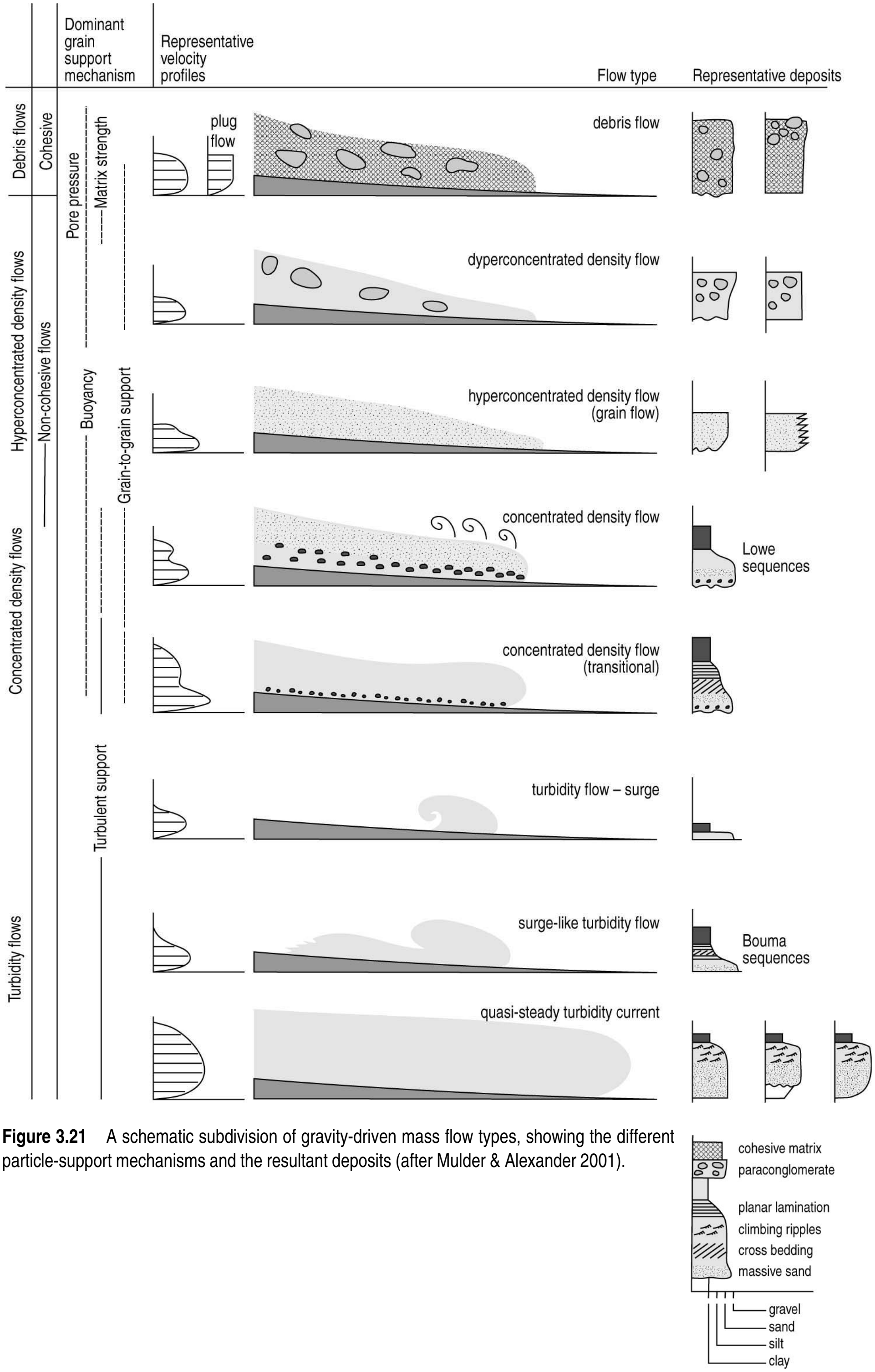

Dominant grain support mechanism
Representative velocity profiles
Flow type
Representative deposits
Debris flows
Cohesive
Hyperconcentrated density flows
Non-cohesive flows
Concentrated density flows
Turbidity flows
Pore pressure
Matrix strength
Buoyancy
Grain-to-grain support
Turbulent support
plug flow
debris flow
dyperconcentrated density flow
hyperconcentrated density flow (grain flow)
concentrated density flow
Lowe sequences
concentrated density flow (transitional)
turbidity flow – surge
surge-like turbidity flow
Bouma sequences
quasi-steady turbidity current
cohesive matrix
paraconglomerate
planar lamination
climbing ripples
cross bedding
massive sand
gravel
sand
silt
clay

Figure 3.21 A schematic subdivision of gravity-driven mass flow types, showing the different particle-support mechanisms and the resultant deposits (after Mulder & Alexander 2001).

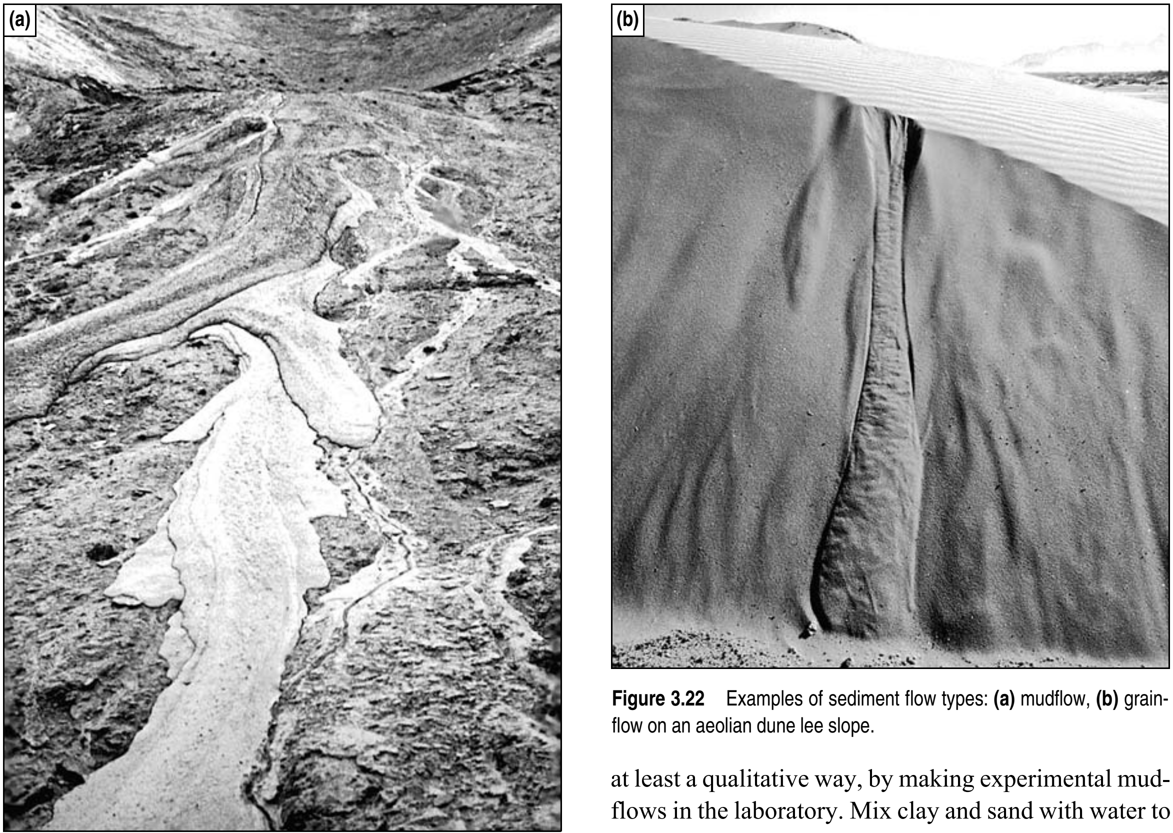


Figure 3.22 Examples of sediment flow types: **(a)** mudflow, **(b)** grainflow on an aeolian dune lee slope.

contacts), and even turbulence, become important and the flow transforms into a hyperconcentrated flow. In sub-aqueous 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 pp. 39–42).

Debris flows that maintain cohesive behaviour throughout their lives will eventually decelerate either because of a downslope reduction in gradient or, in some sub-aerial cases, through progressive loss of pore water. When the shear attributable to the downslope 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. 3.22a).

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 mudflows are characterized by a lack of stratification and of sorting of the particles within them, because all particles come to rest at more or less the same time when the flow stops. As we shall 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 grainflows)

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; at lower concentrations, the flows are transitional with turbidity currents, where only turbulence operates. Hindered settling, which occurs with all but 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 **grainflows**.

Pure grainflows occur only on steep gradients. It is a common observation that a slope of loose dry sand is stable only below a certain gradient. Attempts to steepen the gradient trigger a flowing movement of the sand after the angle has been increased by a few degrees (Fig. 3.22b). 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).

Some simple experiments can illuminate 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. Methods of doing this are outlined in some of the references given at the end of the chapter. 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 and conditions?

When the angle of rest is exceeded, internal shear stresses attributable to the downslope 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 pore 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 and the grains will resume a closer packing.

In concentrated and hyperconcentrated flows 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 grainflow may be replicated. These so-called modified grainflows or traction carpets depend on the shear stress applied by an overriding powerful current, probably a turbidity current with a high load of suspended sediment (Figs 3.23, 3.24). 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 are 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 fluid turbulence. Turbidity currents are driven by gravity acting on the excess density of the suspension over that of the surrounding clear water, and the various properties of the flow, including thickness, concentration, velocity and gradient, are highly interdependent. Alteration 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 (abyssal plain), deposit their load when a reduction in

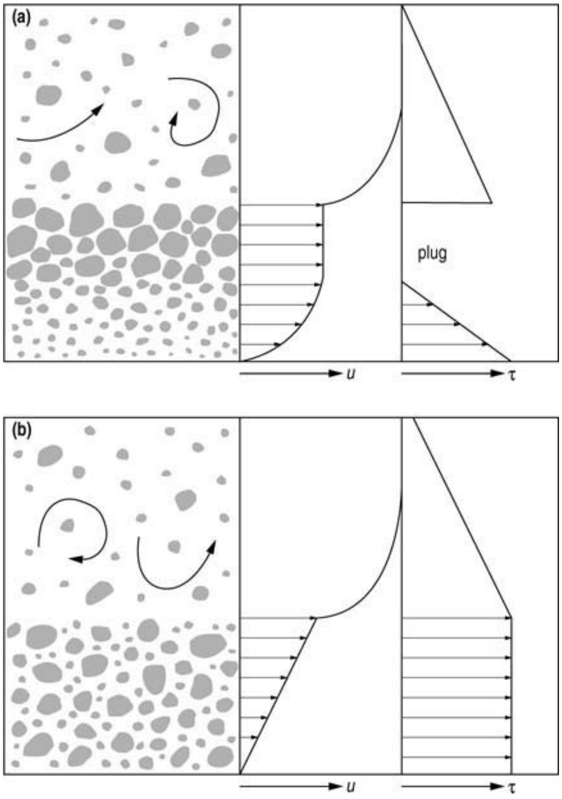

(a)
plug
u
τ
(b)
u
τ

Figure 3.23 Comparison of the profiles of mean flow velocity *u* and shear stress τ for **(a)** grainflow-like traction carpet with a non-deforming plug at the upper part (Lowe 1982) and **(b)** simple-shearing traction carpet with constant shear stress throughout (Hiscott 1994). (After Sohn 1997)

slope 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 §6.8.4). At their highest concentrations, turbidity currents grade into concentrated flows in which other support mechanisms become increasingly important.

Large turbidity currents commonly originate from the slumping of poorly consolidated material near the top of a slope and are important in the transport of sand to the abyssal regions of the ocean floor, usually via submarine canyons. The initial slumps tend to 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 (Fig. 3.25). In the process, the flow may pass through stages involving movement as debris flows and as concentrated flows. Because turbidity currents are fluidal flows, they have no yield strength and so are able to 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 that may change character as the flow decelerates. Each bedform will produce its own distinctive lamination (as reviewed in Ch. 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 most rapidly to its developing bed.

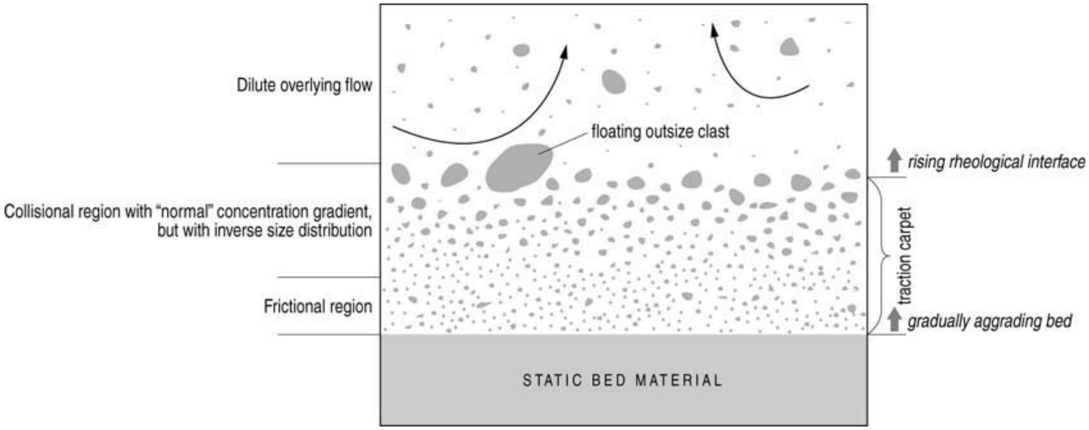

Dilute overlying flow
floating outsize clast
Collisional region with "normal" concentration gradient, but with inverse size distribution
Frictional region
rising rheological interface
traction carpet
gradually aggrading bed
STATIC BED MATERIAL

Figure 3.24 Deposition from a density-stratified and size-graded traction carpet via gradual aggradation of the bed (after Sohn 1997).

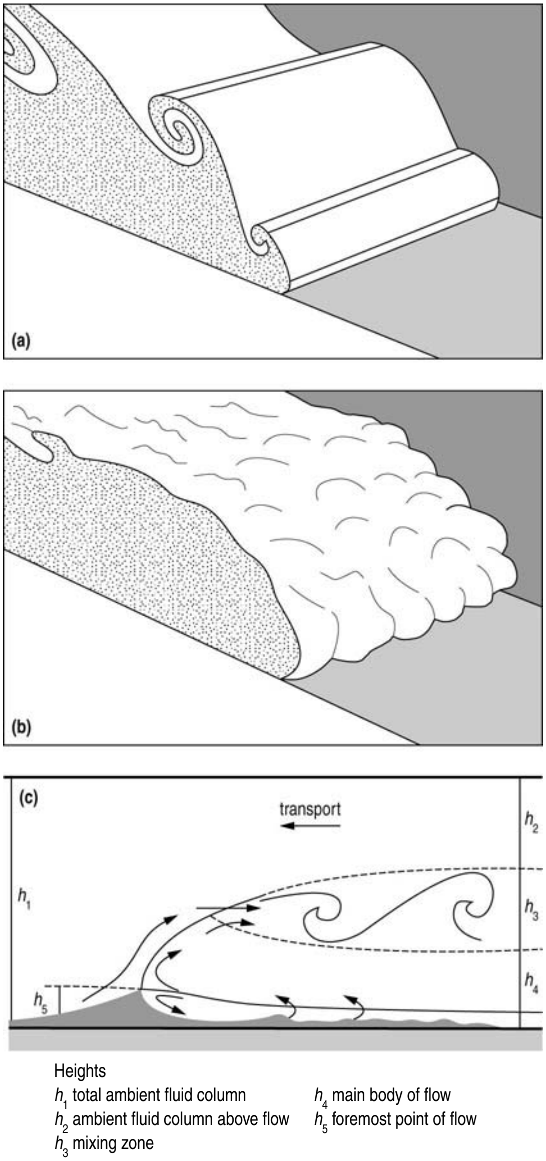

(a)
(b)
(c)
transport
h_1
h_2
h_3
h_4
h_5
Heights
h_1 total ambient fluid column
h_2 ambient fluid column above flow
h_3 mixing zone
h_4 main body of flow
h_5 foremost point of flow

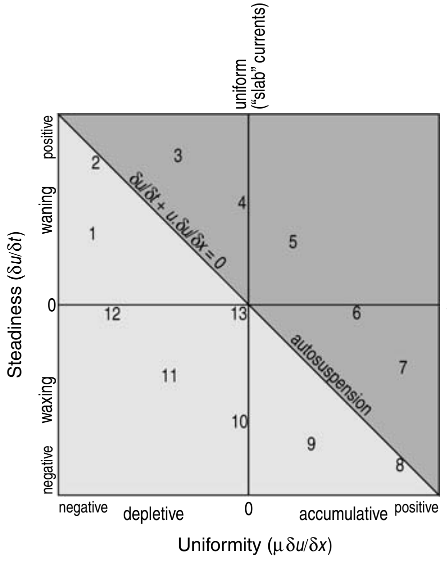
Figure 3.25 Two kinds of instability at the front of a gravity-current head, which allow ambient water to be mixed into the body of the flow. **(a)** Billows (Kelvin–Helmholtz waves); **(b)** brain-like lobes and clefts that develop at the contact of the overhanging head with a solid boundary; **(c)** a section through the head of a turbidity current. (Modified after Simpson 1987)

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 a constriction, 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 show erosive potential (Fig. 3.26).

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 thermohaline density differences caused by the cooling of water near the poles and increased salinity because of the formation of pack ice.

Certain density currents, because of 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 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 experimentation. Ideally, construct or obtain a narrow but deep tank with glass or clear plastic sides at least 1 m 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


uniform
("slab" currents)
Steadiness ($\delta u/\delta t$)
positive
waning
0
waxing
negative
1
2
3
4
5
6
7
8
9
10
11
12
13
$\delta u/\delta t + u\,\delta u/\delta x = 0$
autosuspension
negative
depletive
0
accumulative
positive
Uniformity ($\mu\,\delta u/\delta x$)
steady (time-invariant currents)

Key

1–13 different types of density current

▨ currents liable to erode

☐ currents liable to deposit

1 – waxing depletive (may deposit)
2 – waxing depletive (autosuspending)
3 – waxing depletive (non-depositional)
4 – uniform waxing
5 – waxing accumulative
6 – steady accumulative
7 – waning accumulative (non-depositional)
8 – waning accumulative (autosuspending)
9 – waning accumulative (may deposit)
10 – waning uniform
11 – waning depletive
12 – steady depletive
13 – steady uniform (autosuspending)

Figure 3.26 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 they migrate from one field to another. (Modified after Kneller & Branney 1995)

do the temperature differences influence the interaction of the water?

In a second series of experiments, add finely powdered clay-size 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 not available, 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 inhomogeneous 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 gravity. They originate as a consequence of

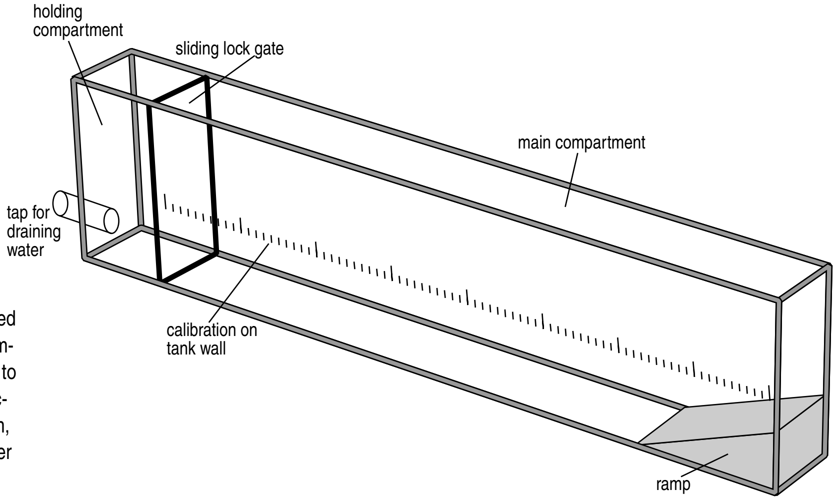

holding
compartment
sliding lock gate
main compartment
tap for
draining
water
calibration on
tank wall
ramp

Figure 3.27 A tank for studying density currents in the laboratory. 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 = 2–5 m, width = 0.2–0.3 m, height = 0.5–0.8 m, water depth = 0.2–0.4 m.

magmatic explosions, whereby 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 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 common flow-initiation mechanisms.

Many pyroclastic density currents are considered to be single-surge events generated by an individual short-lived pulse that waxes rapidly and then wanes rapidly. However, pyroclastic events resulting from 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. Waxing of a pyroclastic density current typically results from dilation of a volcanic conduit or vent, often associated with an increase in gas pressure, whereas waning typically results from the progressive depletion of volatiles in a volcanic conduit, often associated with eruptive withdrawal. 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 because of flow convergence or a downstream increase in gradient, or both; it is **depletive** where it decelerates because of flow divergence or a downstream decrease in gradient, or both (Fig. 3.26).

Study techniques

Field experience

It is important to generate a feel for the interaction 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) requires sustained periods of observation, whereas direct experience of rapid processes, such as avalanches, will usually be by accident rather than design. 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 or 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 (1985a).

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's law of particle settling Fill a 1 L measuring cylinder with glycerine and add a grain of gravel (5–8 mm 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 then 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 (0.5 m long, 0.05 m 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

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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 sequences 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 almost immediately. Small-scale erosional structures are almost always recognized as relief on the base of the bed immediately overlying the erosion surface. Erosion is also recognized in vertical sections by truncation of bedding or lamination in the sediment below the erosion surface. However, if erosion has been widespread, 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 reflects only the minimum thickness of sediment removed.

Many erosional structures are valuable indicators of both way-up and palaeocurrent direction. They can, therefore, help in structural and palaeogeographical analysis, as well as giving insights into processes active during 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 interbedded sequences
- small structures seen on modern surfaces and more rarely on upper bedding surfaces in ancient strata
- large structures normally recognized in vertical section in ancient sediments (i.e. channels, 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-grain beds interbedded with mudstones. The coarser-grain sediments are commonly sandstones, but exceptionally may be limestones or conglomerates. The sole marks result from the erosion of cohesive fine-grain sediment, usually mud, which passes on erosion directly into suspension. The cohesive strength of sediment allows details of the erosional relief to be maintained until they are buried by coarser-grain material (Fig. 4.2). Erosion of mud and deposition of coarser material can often be phases of the same current, separated by only a short period of time. Subsequent lithification usually renders the coarse-grain sediment more resistant than the finer material to eventual weathering, so that the fine-grain sediment is preferentially removed to expose a cast of the erosional relief on the base of the sandstone bed. Resumption of deposition of fine-grain sediment, similar to that eroded, without any deposition of coarse-grain sediment, would not normally provide the lithological contrast needed to pick out the structures on weathering. It is very important to understand this mode of preservation and to recognize that the structures observed are negative impressions of the erosional relief.

Sole marks are typically the products of environments characterized by episodic sedimentation. Background deposition of mud is punctuated by sudden

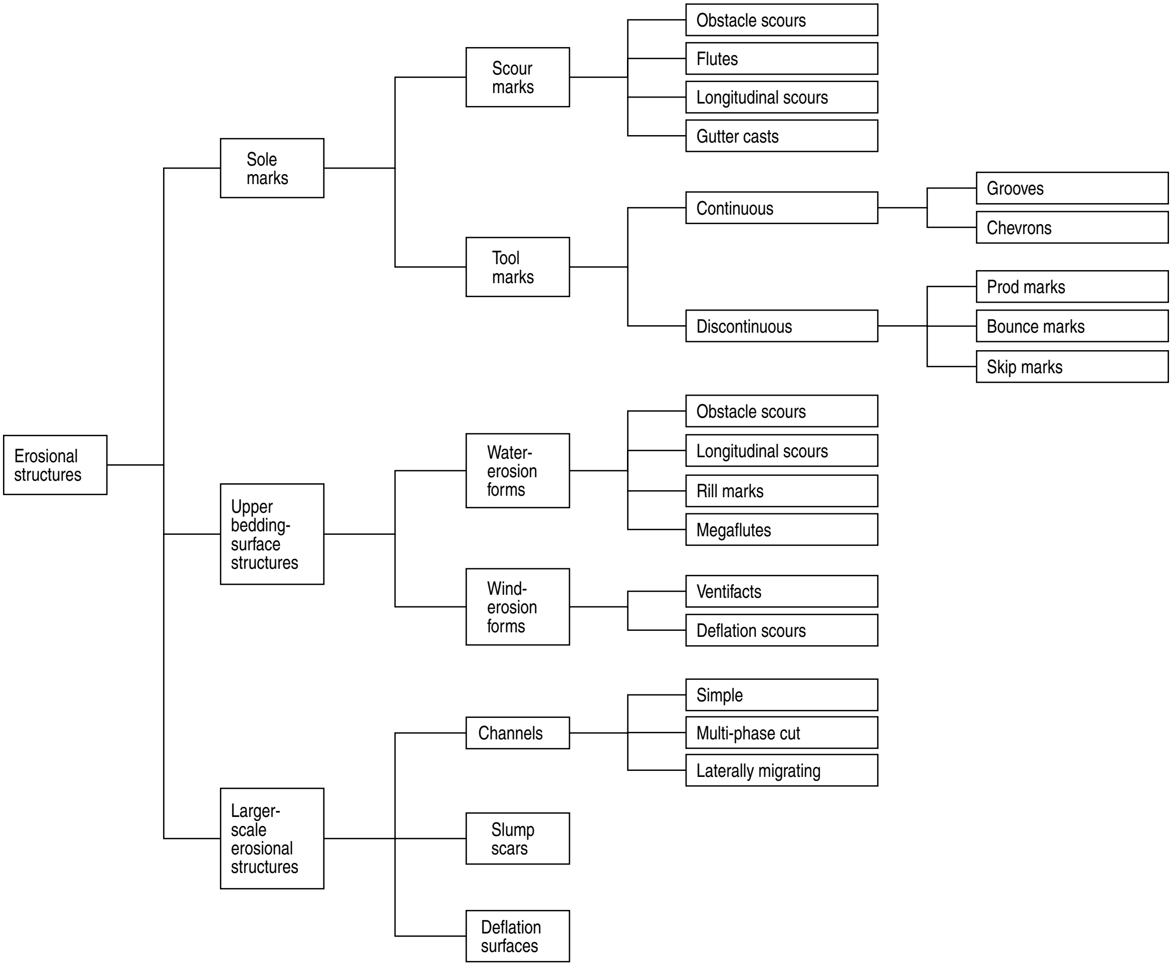

Erosional structures
Sole marks
Scour marks
Obstacle scours
Flutes
Longitudinal scours
Gutter casts
Tool marks
Continuous
Grooves
Chevrons
Discontinuous
Prod marks
Bounce marks
Skip marks
Upper bedding-surface structures
Water-erosion forms
Obstacle scours
Longitudinal scours
Rill marks
Megaflutes
Wind-erosion forms
Ventifacts
Deflation scours
Larger-scale erosional structures
Channels
Simple
Multi-phase cut
Laterally migrating
Slump scars
Deflation surfaces

Figure 4.1 Scheme for the classification of erosional sedimentary structures.

influxes of coarser sediment in events comprising an 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 floods in semi-arid environments, and crevasse surges into floodplains all have the properties necessary 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 that differ principally in the way the structures are generated: turbulent scour (scour marks) and objects moved by the current (tool marks).

4.2.2 Scour marks

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

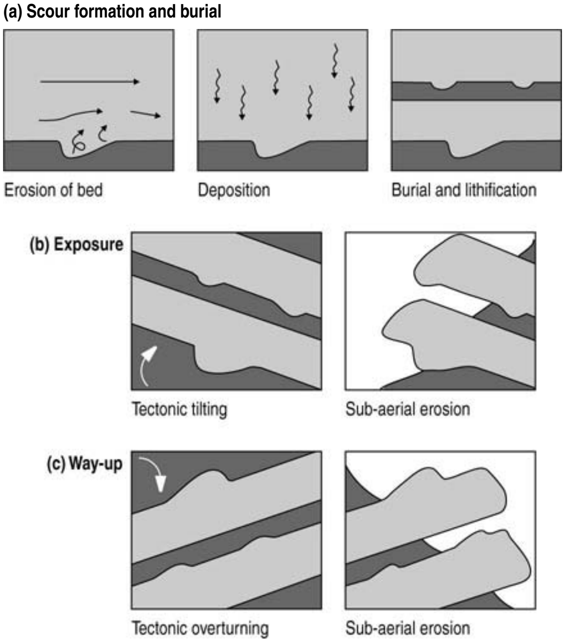

(a) Scour formation and burial
Erosion of bed
Deposition
Burial and lithification
(b) Exposure
Tectonic tilting
Sub-aerial erosion
(c) Way-up
Tectonic overturning
Sub-aerial erosion

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).

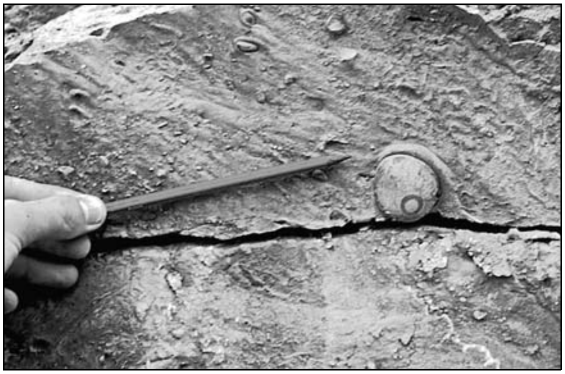


Figure 4.3 Obstacle scour around a pebble in the base of a sandstone bed. Several smaller pebbles also show their own scours. Current from top left to bottom right. Hecho Group, Eocene, Pyrenees, Spain.

Obstacle scours

Large clasts such as pebbles, fragments of wood and more robust types of fossils sometimes occur on the bases of sandstone beds and are associated with distinctive ridges of sandstone that point down into the underlying bed. The ridges are commonly crescentic or of horseshoe shape, partially encircling the large clast, with tails dying away in one direction (Fig. 4.3). 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 streamflow 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 down stream. 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; spiral eddies are shed on either side and die out down stream, 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, when 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-shape ridge occurs in isolation and the nature of the obstacle is left to the imagination of the observer.

Flutes

Flutes are similar to obstacle scours. They occur both as isolated features and as collective groups that share a common origin and form distinctive patterns. Individually they vary in shape and size, but on any one surface

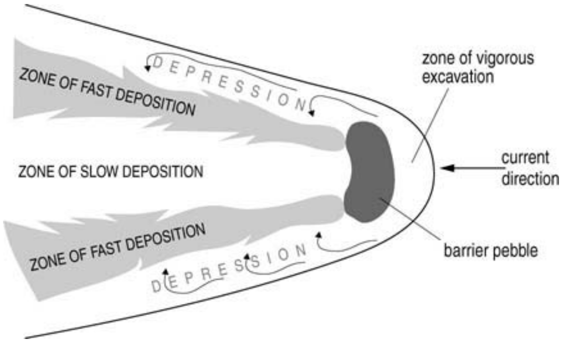

DEPRESSION
ZONE OF FAST DEPOSITION
zone of vigorous excavation
ZONE OF SLOW DEPOSITION
current direction
ZONE OF FAST DEPOSITION
barrier pebble
DEPRESSION

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

they tend to be rather similar. Flutes are characterized by a rounded, sometimes tightly curved, “nose” at their upstream 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 5 cm to 50 cm, in width from 1 cm to 20 cm, and in depth commonly a few centimetres, exceptionally up to 10 cm. In shape they range from highly elongate forms, gradational with longitudinal scours, to very wide forms with gently curved noses called **transverse scours** (Figs 4.5, 4.6). Some flutes have highly twisted shapes particularly close to the nose (Fig. 4.6d); others have very simple streamlined forms (Fig. 4.6b,c). 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.

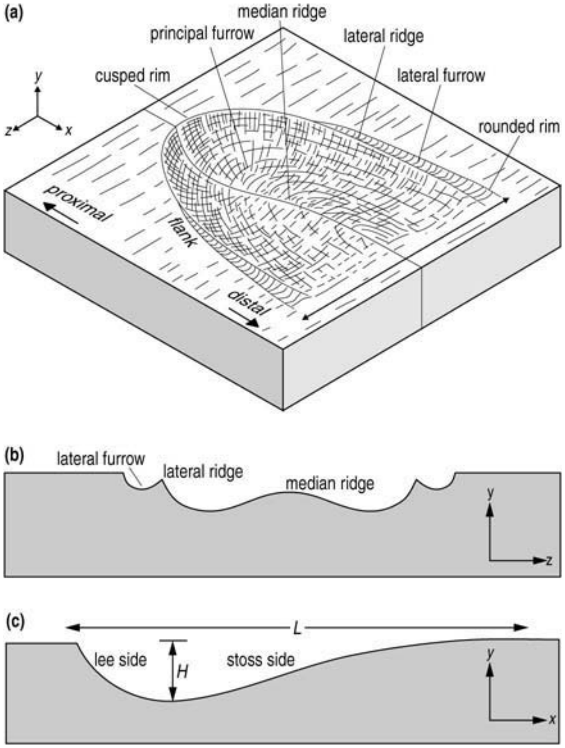

(a)
median ridge
principal furrow
lateral ridge
cusped rim
lateral furrow
y
z
x
rounded rim
proximal
flank
distal
(b)
lateral furrow
lateral ridge
median ridge
y
z
(c)
L
lee side
H
stoss side
y
x

Figure 4.5 Morphological features of flute marks: **(a)** oblique view, **(b)** transverse section view, **(c)** longitudinal section view (modified after Allen 1982).

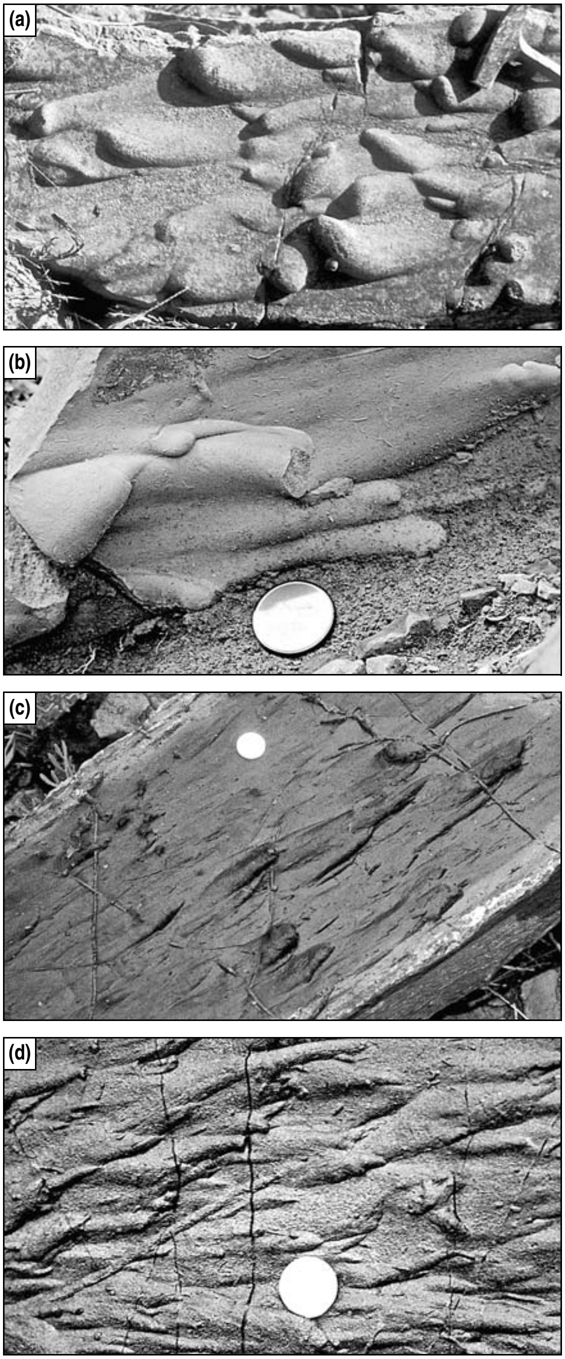


Figure 4.6 Examples of different flute forms on the bases of sandstone beds. Try to judge the palaeocurrent direction in each case. Example **(c)** is unusual in showing both flutes and tool marks, as well as later, cross-cutting elongate burrows. (Photos (a) and (b) courtesy of Gilbert Kelling.)

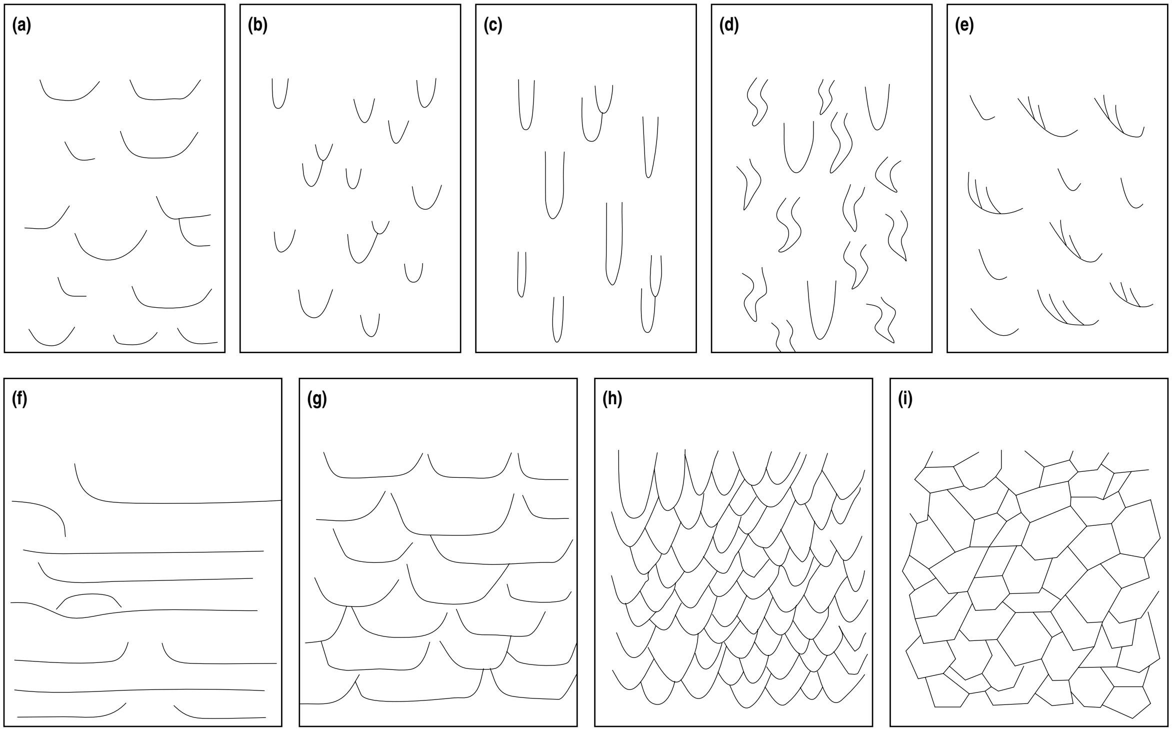


Figure 4.7 Schematic illustration of the range of shapes of flutes in plan view; current from bottom to top in each case (after Allen 1971).

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; 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 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 down stream 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.

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 sequences, flutes are among

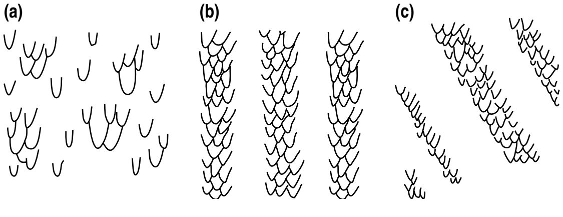


Figure 4.8 Heterogeneous patterns of distribution of flute casts on bedding surfaces (after Allen 1971).

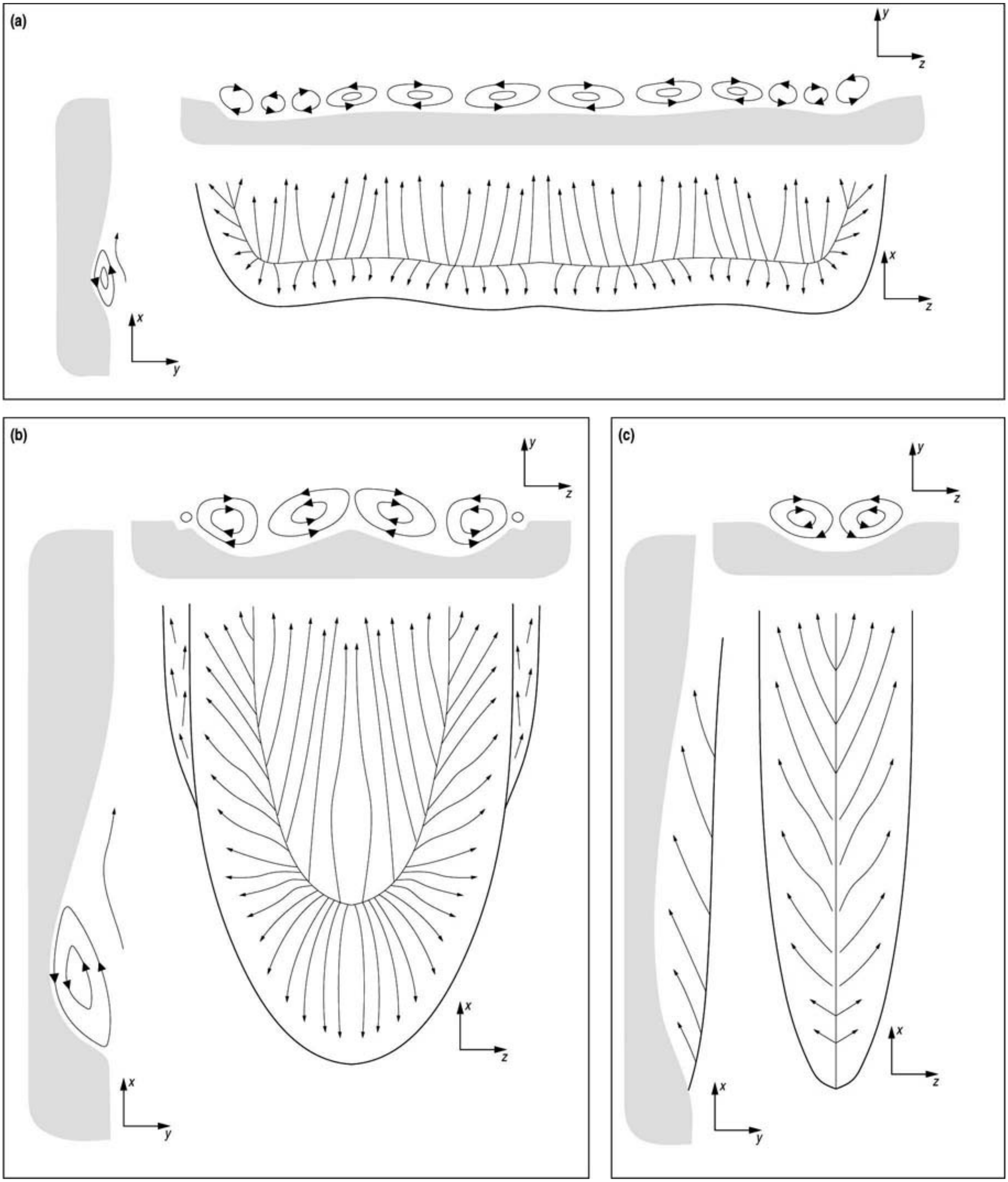


Figure 4.9 Simplified patterns of water motion (eddying) associated with erosional scours of different shapes: **(a)** transverse scour, **(b, c)** parabolic flutes of differing width; flow from bottom to top in all cases. Dimensions: *x* direction is down stream, *y* is vertical and *z* is transverse to flow. (After Allen 1971)

the most abundant and important indicators of palaeocurrent direction.

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 downward-pointing ridges that are rather rounded and intervening furrows that are rather sharp, reflecting round-bottom troughs and sharp ridges on the surface of the eroded mudstone (Fig. 4.10). The spacing of the ridges is typically 0.5–1 cm, 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 flow-parallel zones of alternating zones of upwards- and downwards-directed components of flow. The lineation along which descending vortex limbs impinge on the bed will be sites of high stress and rapid erosion, and, beneath ascending limbs, stress will be at a minimum and erosion at its lowest (Fig. 4.11). Once localized on scour features, eddies become fixed and they 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 axis of the eddy.

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).

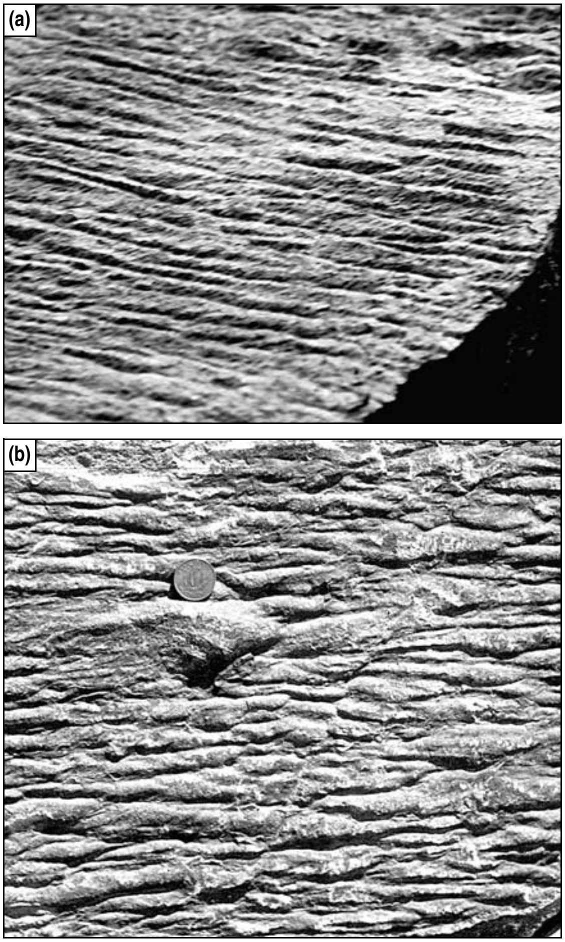


Figure 4.10 Examples of longitudinal scours. **(a)** Longitudinal scours on the bases of sandstone beds. In most cases it is possible to tell only the trend of flow and not its sense of direction. Examples with “noses” also enable the sense of movement to be judged. Mam Tor Formation, Upper Carboniferous,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

Gutter casts

Gutter casts generally occur as isolated elongate ridges on the bases of sandstone or coarse-grain limestone beds. They protrude into the underlying finer-grain sediment from an otherwise rather flat bedding surface, and in vertical section they show U- or V-shape profiles (Fig. 4.12). These are generally symmetrical, but more rarely may be asymmetrical, with one side steeper than the other. They are commonly up to 10 cm wide and almost as deep. Where the coarse-grain sediment does

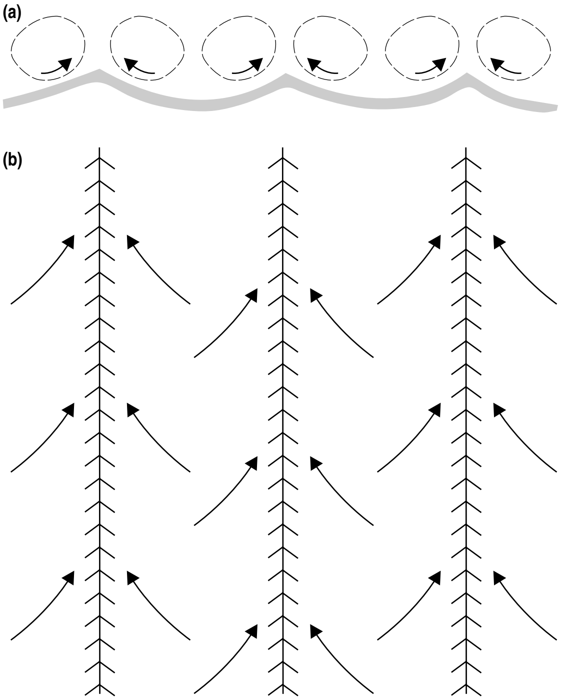


Figure 4.11 The pattern of water movement associated with the development of longitudinal scours: **(a)** vertical section normal to flow, **(b)** plan view on the bed (in part after Allen 1971).

not give a continuous bed above the erosion surface, the coarse infills are preserved as isolated elongate bodies in the finer-grain 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 outside 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 "sandblast" 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 river-channel meander bend.

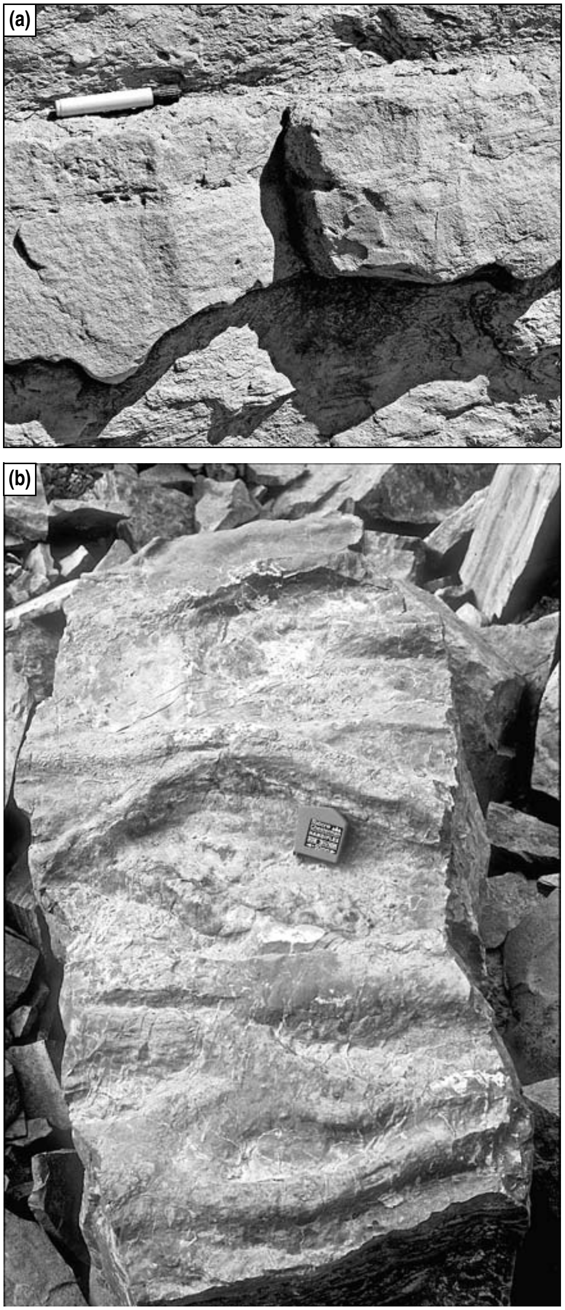


Figure 4.12 Examples of gutter casts. **(a)** Gutter cast seen in end section at the base of a lower sandstone turbidite bed, Tabernas Basin, Miocene, southeast Spain. **(b)** The lower surface of the fill of a gutter cast in a loose, overturned block of limestone. Note the sinuous shape of the gutter 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 flow 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 parallel groups. In transverse vertical section they show an irregular sharply defined relief, usually related to smaller superimposed grooves and ridges (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, plant fragment or 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 take care in measuring and recording the orientation of groove casts. On surfaces where more than one trend is apparent, it is usual to record the various orientations and also 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-shape 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 will usually be unknown, although, where an object is found at the end of the groove, the nature of the tool and the sense of movement of the current are both established. However, with most grooves it is possible to judge only the trend of movement, and measurements should therefore be recorded as undirected lineations (e.g. 120–300°).

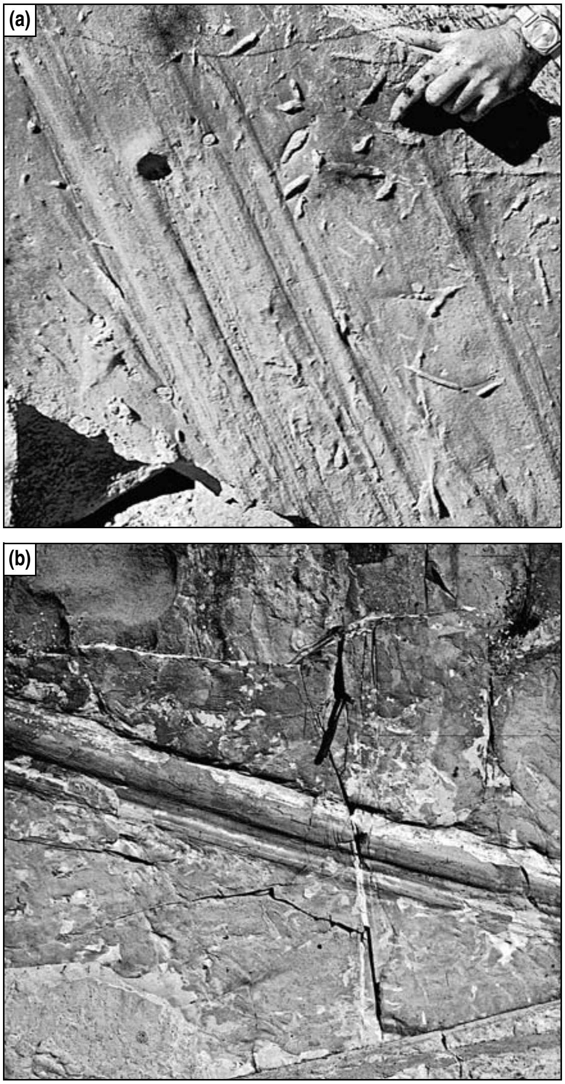


Figure 4.13 Examples of groove casts on the bases of sandstone beds. **(a)** Rather weakly defined grooves with later burrows; Ainsa Basin, Spain. **(b)** Large, well defined groove system; Silurian, southern Scotland (photo courtesy of Gilbert Kelling).

Chevrons

Rarer than grooves, chevrons are linear zones of V-shape crenulations that consistently close in one direction to produce a chevron pattern (Figs 4.15, 4.16). The individual linear zones are seldom more than 3 cm wide and the relief is generally less than 5 mm. Dragging a stick through soft mud or any very viscous liquid

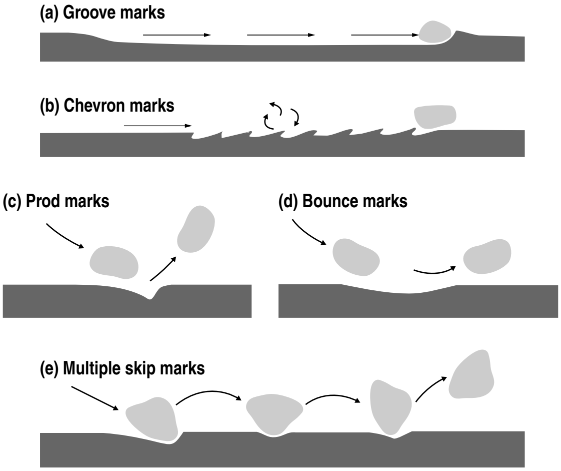


Figure 4.14 Schematic representation of the ways in which different types of tool mark are generated by different modes of behaviour of the tools.

produces a similar pattern. Chevrons record small-scale folding or "rucking-up" of the surface of rather weak but cohesive mud by the passage of a tool very close to the bed (Fig. 4.14b). The V-shape ridges, which make up the chevron mark, close down stream, thus giving a sense as well as a trend to palaeocurrent measurements.

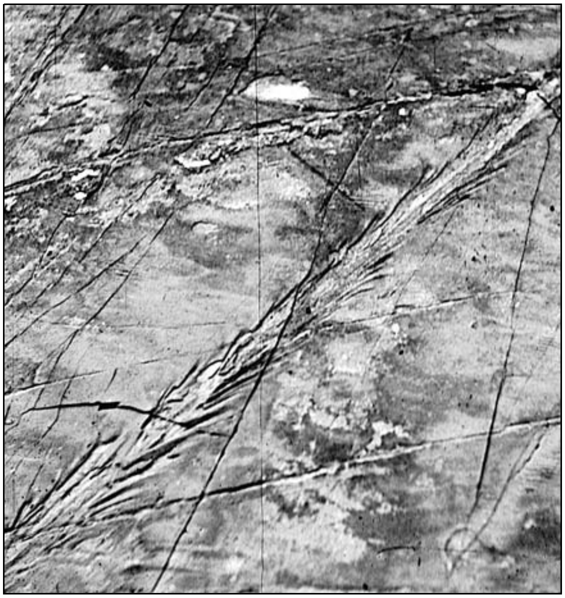


Figure 4.15 Well developed chevron mark on base of sandstone, Silurian, southern Scotland (photo courtesy of Gilbert Kelling).

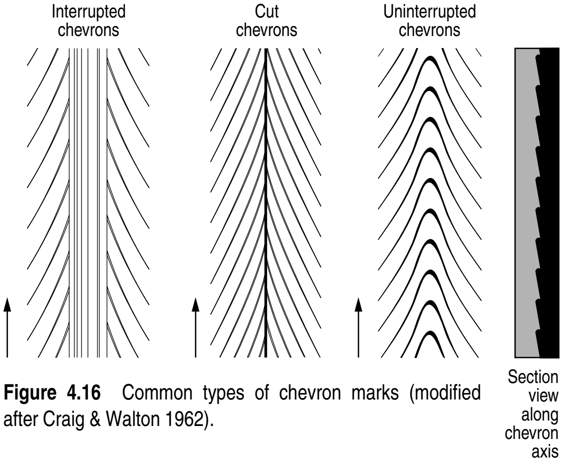


Figure 4.16 Common types of chevron marks (modified after Craig & Walton 1962).

Like other sole marks, chevrons are good way-up indicators.

Prod marks and bounce marks

Sharply defined discontinuous marks, often 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, the other end gradational and sloping 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 1 cm in length and 1–2 mm 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 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, although 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 (Fig. 4.14e). Differences in the shape of bounce marks can sometimes be related to rotation of the tool as it bounced along. Where the marks are almost continuous and approach the appearance of grooves, the tool may have been of disc shape and its behaviour analogous to that of a wheel rolling and bouncing down hill.

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, provide the sense of the movement. In addition, tool marks are rarely seen on the same surfaces as scour marks. In many sequences, particularly of turbidites, tool marks are more common on the bases of thin sandstone beds, whereas scours occur on the bases of thicker beds. This suggests that tool marks represent superficial erosion by relatively weak or short-lived currents, and scour marks record a more wholesale removal of the bed and possibly a cutting down into more cohesive mud.

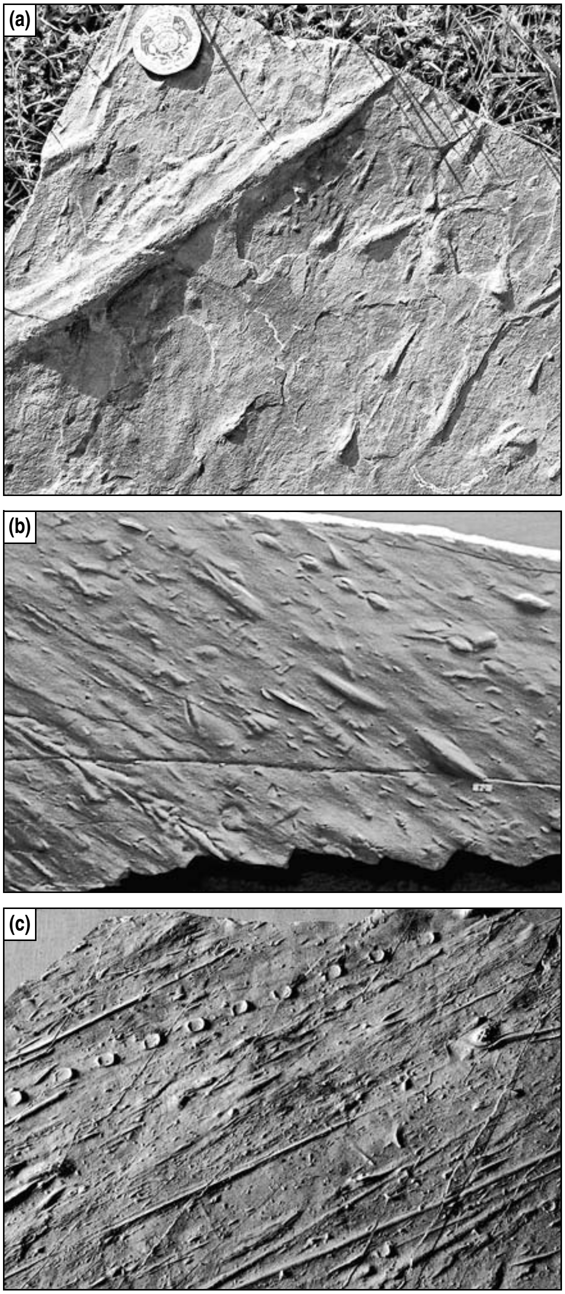


Figure 4.17 Examples of prod, bounce, skip and related tool marks revealed on the bases of sandstone beds. **(a)** Prod marks, which are asymmetrical along their length with the deep end down stream; location unknown. **(b)** Bounce marks, which are symmetrical along their length and generated by separate objects; Krosno Formation, Oligocene, Carpathians, Poland. **(c)** Skip marks made by the repeated impact of the same object (probably a fish vertebra); Krosno Beds.

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 act erosively on 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 internal 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-shape 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 it dies away down stream. On rippled sand, the scour trough and the obstacle may locally perturb the ripple pattern. If the mineral composition of the sand is varied, the scour may be accentuated by mineral sorting.

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 often 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, dendritic channels a few centimetres wide. They 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 level. They occur most commonly on beaches and on flanks of larger tidal bedforms at low tide, although they also occur on the flanks of large bedforms in rivers. They are almost invariably destroyed by a rise of water level and thus they have a very low preservation potential. They have no palaeocurrent significance.

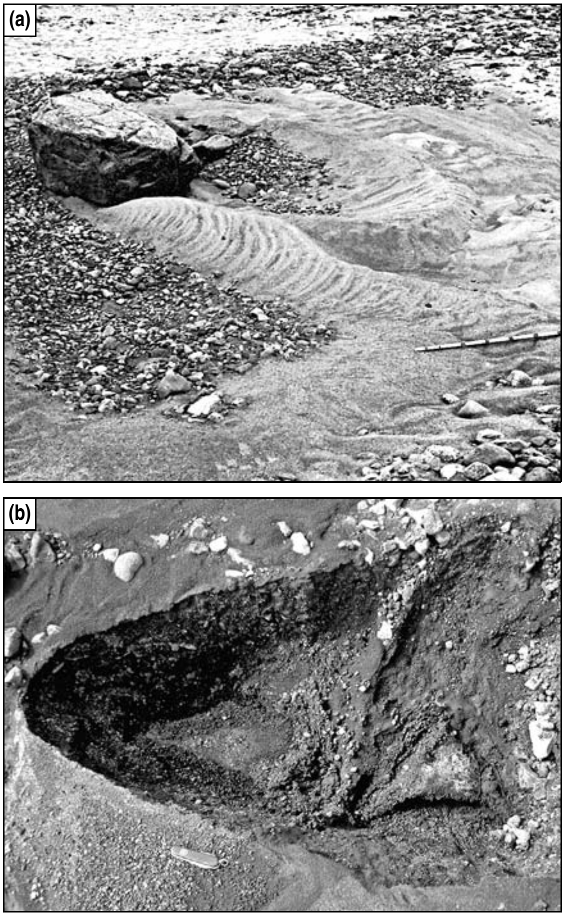


Figure 4.18 Examples of obstacle scours. **(a)** Boulder on a beach with associated scour around it. Note the deposition of sand ridges in the lee of the boulder, which indicate the downcurrent direction. **(b)** Obstacle scour on a fluvial outwash plain. In this case the obstacle was an ice block that has subsequently largely melted away. Myrdalssandar, Iceland.

Megaflutes

Extensive upper bedding surfaces within some turbidite successions show larger-scale erosional features that have shapes similar to flutes, but they differ from them in several important respects. First, they are found on the upper surfaces of thick sandstone beds and they clearly reflect erosion of sand. Secondly, the erosional relief is typically filled by mudstone, sometimes with a

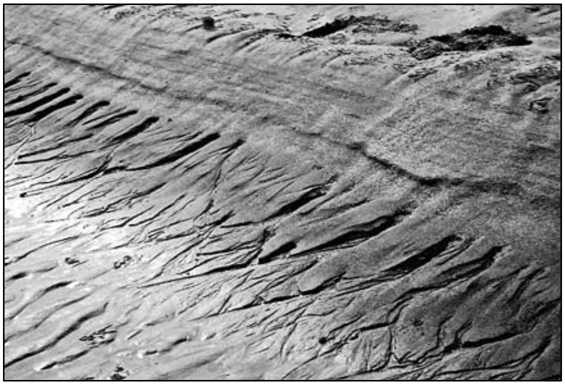


Figure 4.19 Rill marks on the lower part of a sandy channel margin in an intertidal setting. The pattern of small dendritic channels is cut by water emerging from within the sand during falling and low-water level. Tana Delta, Norway.

few thin interbedded sandstones. Thirdly, they are much larger than ordinary flute marks, being typically several metres wide, many tens of metres long and of the order of 1 m deep. In plan view, they have a curved margin, often 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 are commonly decorated with current ripples. Down stream 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). Those of finer grain 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 by a current that did not deposit any sediment as it eventually waned. Instead, the erosional relief was abandoned in a quiet water setting, where fine-grain sediments were deposited from suspension. The only examples of such structures in present-day settings are on the surfaces of deepwater submarine fans, where they have been mapped by deeply towed sonic imaging devices. When found, megaflutes help to characterize the suite of processes active in the overall environment of deposition and also act as useful palaeocurrent indicators.

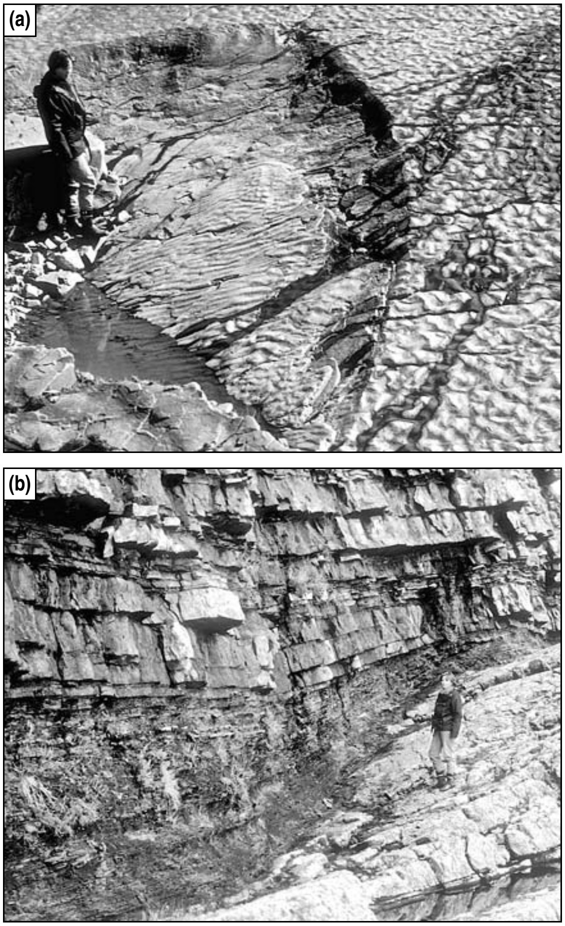


Figure 4.20 Examples of megaflutes. **(a)** Upper bedding surface showing the strongly curved upstream end of the megaflute. Note the current ripples on both the surrounding bedding surface and the floor of the megaflute. **(b)** Cross section through one flank of a megaflute. Note the massive nature of the eroded sandstone and the predominantly muddy nature of the fill of the megaflute. Ross Formation, Namurian, western Ireland.

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 up wind with a tail streaking out down wind (Fig. 4.21a). 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, and they occur in groups rather than as

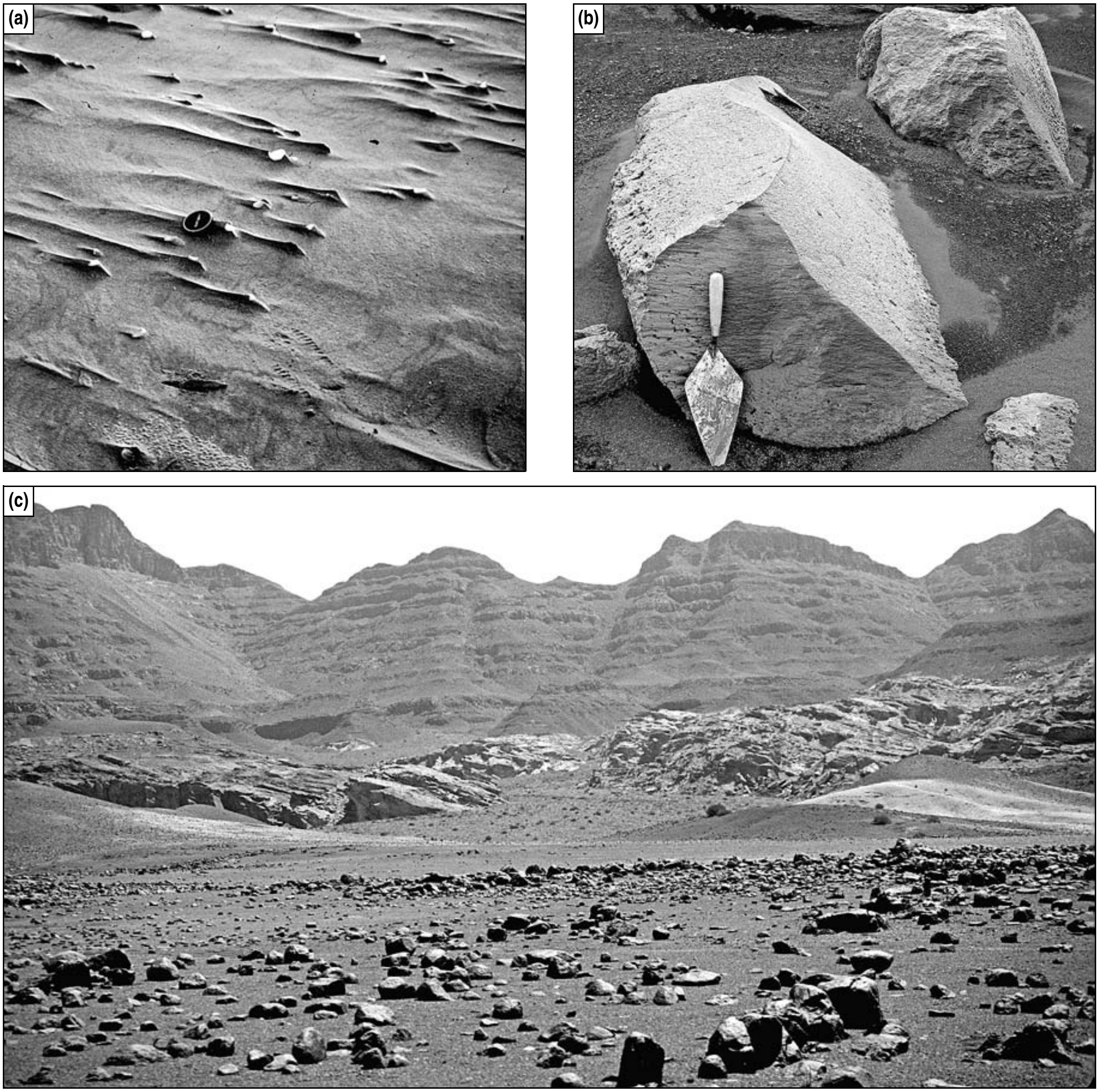


Figure 4.21 Examples of features associated with wind erosion. **(a)** Deflation of sand-grade material across a beach. Larger clasts (shells) act as obstacles, in the lee of which sand is protected from deflation. Ebro Delta, Spain; photo courtesy of Gilbert Kelling. **(b)** Wind-faceted basalt boulder (ventifact) polished by sand abrasion. Askja Sandsheet, Iceland. **(c)** A wind-deflation surface strewn with wind-faceted cobbles. Huab Basin, Namibia.

isolated forms. They are rather uncommon in the rock record, where they could be confused with flute marks, thereby providing misleading way-up criteria.

Another feature is that of wind-faceted pebbles or boulders (or both), 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 bound smooth faces on individual clasts (Fig. 4.21b). 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.

Where strong winds blow across loose sand surfaces, they often strip away sand to generate erosional sand **deflation surfaces**. Such surfaces may be planar, irregular and hummocky, or covered with low-relief scallops arranged into regularly repeating patterns. Wind-faceted ventifacts often occur in large numbers strewn across low-relief sand-deflation surfaces (Fig. 4.21c).

4.4 Erosional features in vertical section

The recognition that erosion has taken place during the accumulation of a sediment sequence 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 laterally extensive exposures. In restricted exposures or in borehole core, this may be impossible.

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 far enough laterally. 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:

- 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"
- erosion by mass movement down a slope, creating a feature of less definite shape and orientation but commonly arcuate along 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 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.

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 are 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. 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? Bear in mind the distortion of width 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 small fraction of total depth. Even a small and apparently complete channel could be superimposed on the floor of a much larger 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 help to establish a palaeogeography. If you can see a clear "channel" shape in an exposure, this tells you that the axis of the channel, or possibly a megaflute,

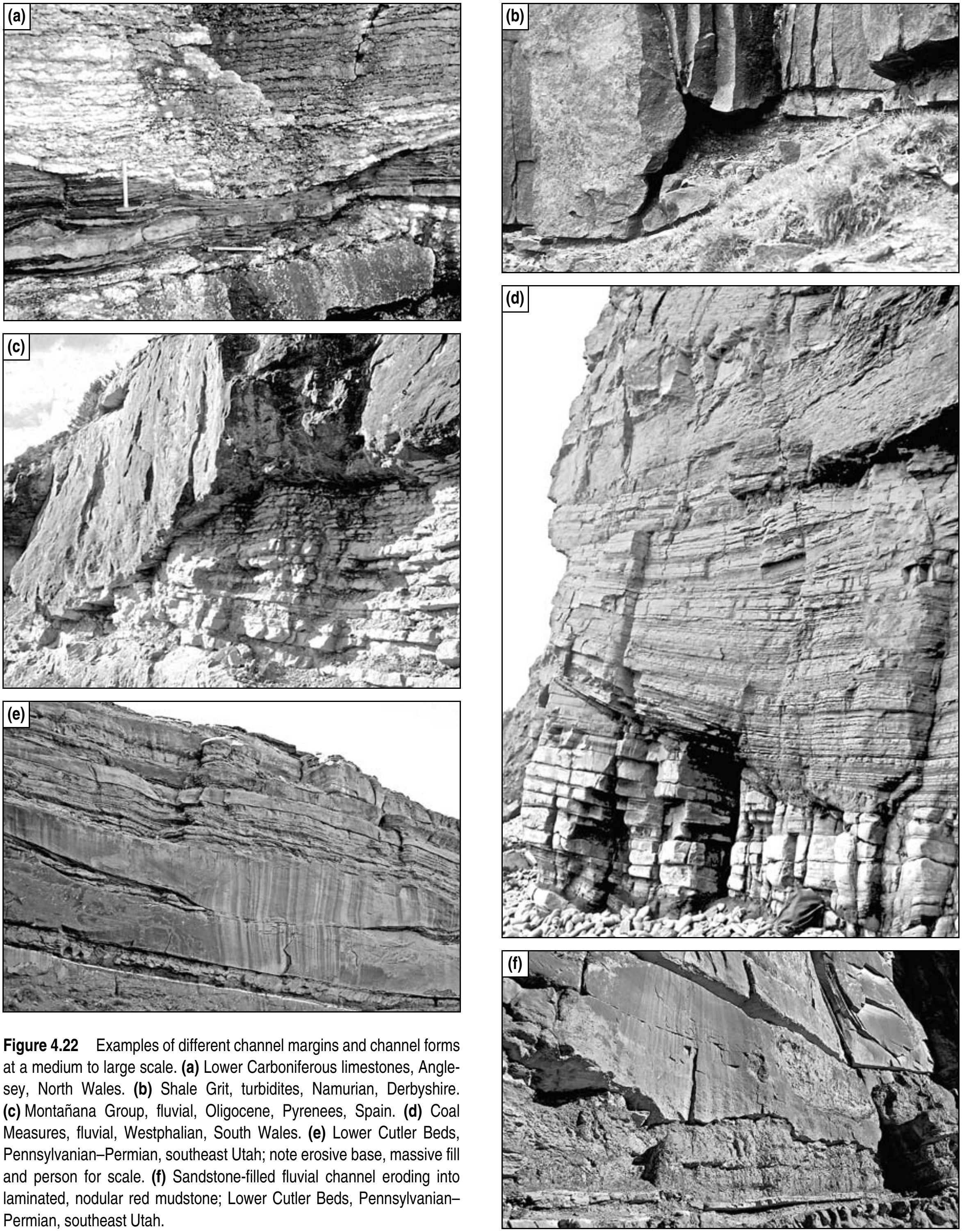


Figure 4.22 Examples of different channel margins and channel forms at a medium to large scale. **(a)** Lower Carboniferous limestones, Anglesey, North Wales. **(b)** Shale Grit, turbidites, Namurian, Derbyshire. **(c)** Montañana Group, fluvial, Oligocene, Pyrenees, Spain. **(d)** Coal Measures, fluvial, Westphalian, South Wales. **(e)** Lower Cutler Beds, Pennsylvanian–Permian, southeast Utah; note erosive base, massive fill and person for scale. **(f)** Sandstone-filled fluvial channel eroding into laminated, nodular red mudstone; Lower Cutler Beds, Pennsylvanian–Permian, southeast Utah.

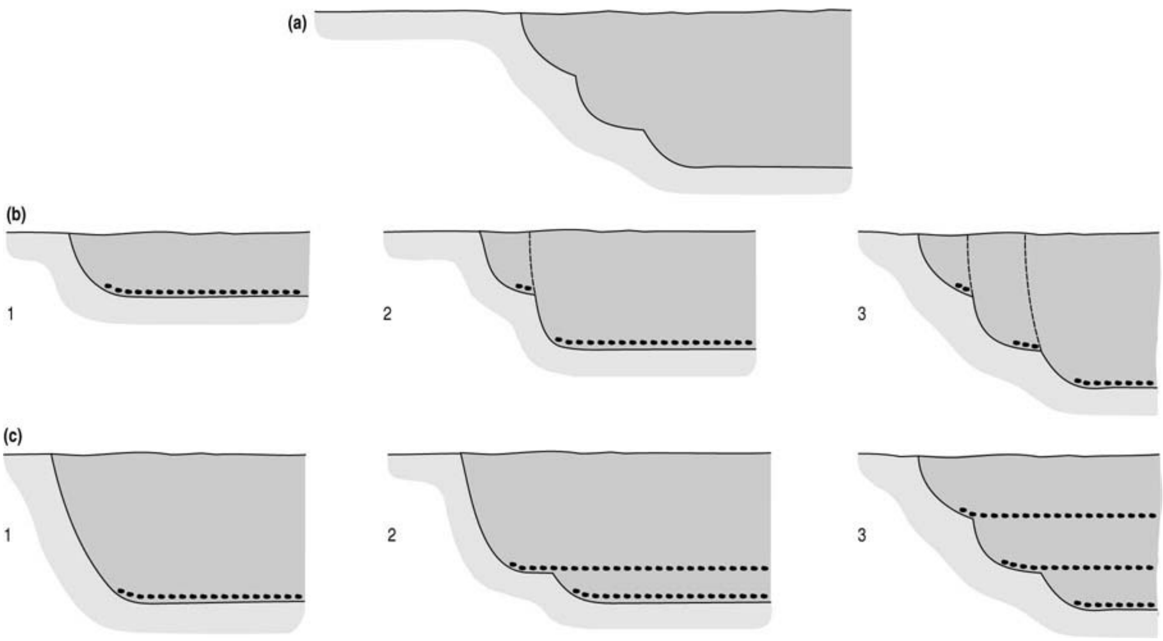


Figure 4.23 Possible stages in the erosion of a stepped channel margin by repeated episodes of cut and fill. **(a)** Final channel shape. **(b, c)** Two different stages by which the final form could be achieved. The ability to recognize erosion surfaces within the channel fill may be vital in understanding the full history of development of the channel.

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 be parallel to the channel trend. Small-scale structures superimposed on the floor or walls will also give a clue. Erosional sole marks on a channel floor or ripples on the floor of a megaflute give a good indication of channel trend, even though the natural fluctuation of flow direction in the channel will reduce the accuracy of any measurement. In cases of extensive bedding-plane exposure, such as on a wavecut platform or in conditions of semi-arid or desert weathering, it may be possible to trace the channel over long distances and thereby establish a more reliable trend. It may even be possible to judge its sinuosity. If possible, it is always worth looking down from clifftops onto 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-grain sediments most commonly form more resistant features, whereas coarser, less cohesive sediment is preferentially eroded. In extreme cases, overhangs may develop and indicate either erosion of cohesive sediment or at least partly lithified non-cohesive sediment.

Many channel forms seen in the field may have undergone a series of erosional and depositional events. Figure 4.23 shows ways in which different sequences of "cut and fill" can produce a similar channel shape. The sequence of events can be deduced only from observing the channel fill as well as the erosion surface. Erosion surfaces, which are obvious when 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 sometimes be a sharp parting and may often 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).

Absence of distinct erosional relief

The absence of distinct erosional relief does not necessarily mean that no erosion has taken place or even that 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 will also prevent a channel-shape cross section from being seen. Clearly, we must be aware of 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 resting sharply on a unit of finer sediment will often have been deposited in a channel. This explanation is supported if a layer of coarse clasts occurs at or just above its base, particularly if the clasts are of intraformational (rip-up) origin. 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 sheet floods or by large turbidity currents. They may also result from widespread non-channelized erosion as a result of a marine transgression, whereby erosion takes place as a result of 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 one must observe a channel shape in order to infer that a channelized flow was responsible for the deposition of a particular rock unit. Clearly, if a channel form is seen, a channelized flow was involved, but the absence of a channel form does not necessarily rule out the presence of a channel during deposition. In many channel deposits, preserved channel margins are relatively rare features, as becomes apparent if 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 maintaining their cross-sectional shape. This generates an erosion surface that, although eventually ending in a channel margin, may be laterally so extensive that the chances of seeing the margin are small. The most common examples of this behaviour are in meandering rivers or tidal channels (Figs 4.24, 4.25). Additionally, there are now several well documented examples of laterally migrating submarine fan channels in deep-sea settings, where both high-quality outcrop and 3-D seismic data have been used to demonstrate channel evolution through time.

In a case 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,

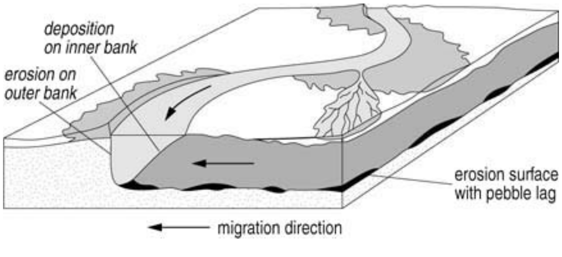


Figure 4.24 The lateral migration of a channel, in this example because of the development of meanders, may erode a horizontal erosion surface with little or no relief. The presence of a channel can only be inferred unless its final position, prior to its abandonment, is observed at outcrop. See Chapter 10 for an example of the likely preserved sequence. (Based on Allen 1964)

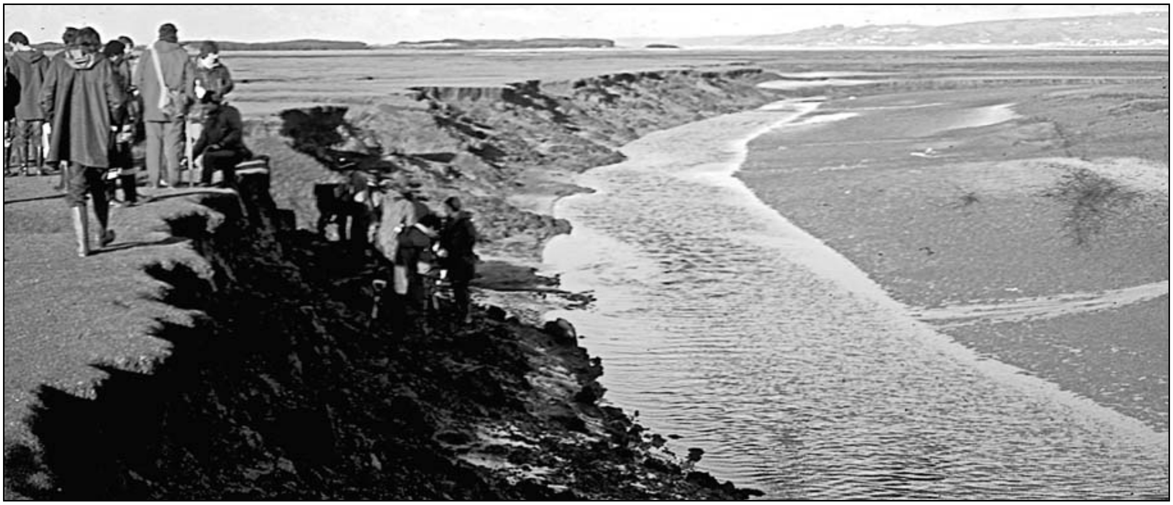


Figure 4.25 Lateral migration of a tidal channel. The channel is migrating from right to left. The eroded bank shows blocks of cohesive fine-grain bank material falling into the channel. The deepest part of the channel, below water, erodes a near-horizontal erosion surface as it migrates and is overlain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

internal features of the channel deposits above the erosion surface show lateral accretion surfaces that may suggest the shape of the channel cross section (see §6.2.10).

Distinguishing channels and slump scars

Not all erosional cross-cutting surfaces in rocks are the margins of channels scoured by flowing water. Some are products of slumping on sub-aqueous slopes, which leaves behind a slump scar. The ability to distinguish slump scars from channels is important for interpreting processes and environments of deposition and 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 an orientation similar to that of the surface, suggesting local

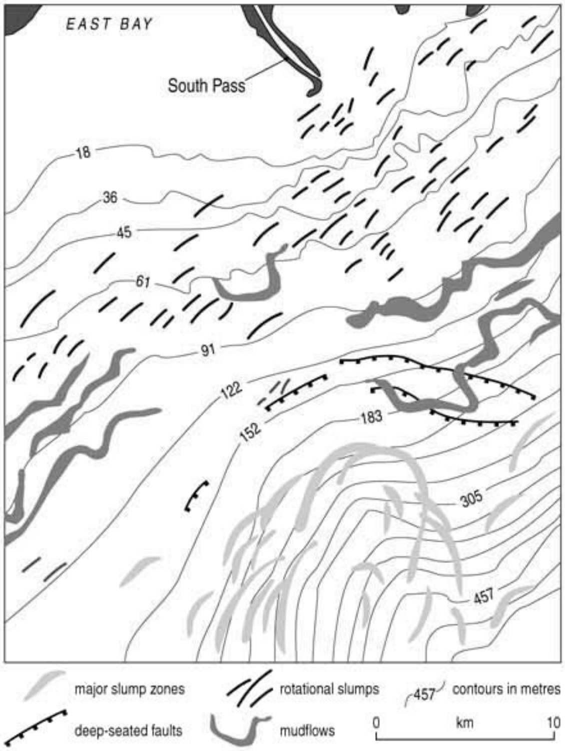


Figure 4.26 Pattern of major and minor slump scars on the slope of the present-day Mississippi Delta. Note that most of the slump scars are sub-parallel to the bathymetric contours. (Modified after Roberts et al. 1976)

horizontal stretching. 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 originate because of instability of usually fine-grain sediment on a slope and they commonly move off spontaneously without any external trigger. Depositional conditions in the area are unaltered; later, sediments similar to those below the surface 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 in not only 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-grain bank material. With some slump scars, the mass of slumped sediment may not have moved far, and a deformed, often chaotically bedded, slump deposit may be found close by. In other cases, a series of subparallel slip surfaces may occur, with slices of slightly shifted but otherwise undisturbed sediment between them. In the unusual case of megaflutes (§4.3.1), it is the coarser sand that is eroded, with a fill or drape of finer material.

Slumping and channel scouring do, in addition, 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 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-faceted pebbles

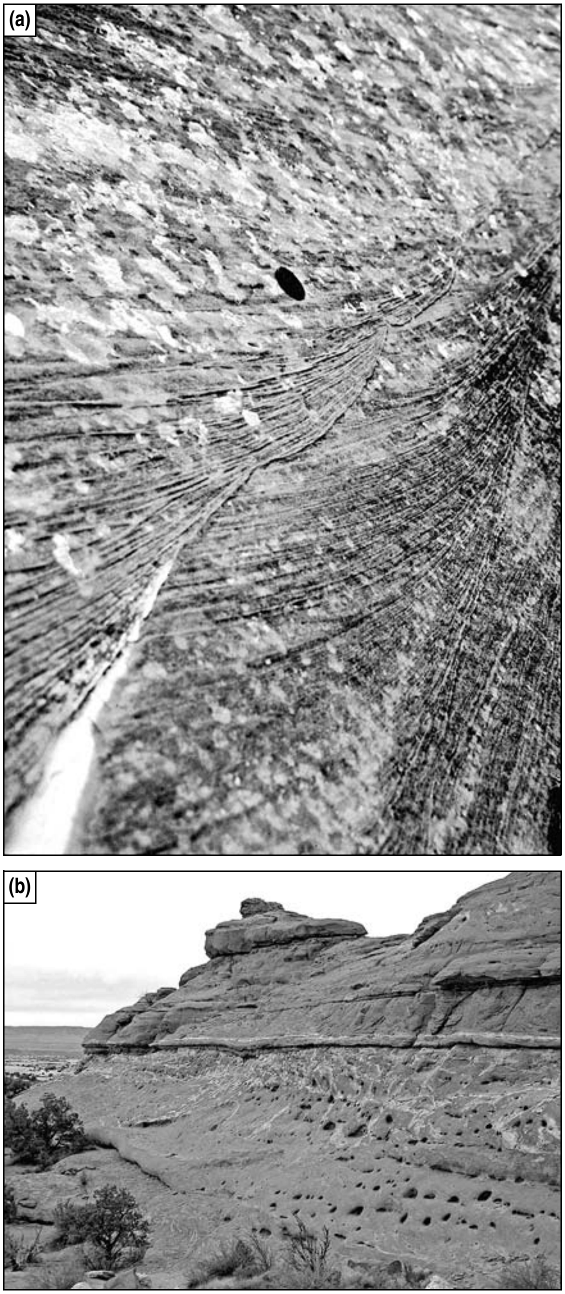


Figure 4.27 Widespread wind-deflation surfaces. **(a)** Erosive aeolian scour surface with cross-stratified aeolian dune deposits beneath and above. **(b)** Laterally extensive and planar aeolian deflation surface separating two aeolian dune accumulations. The white mottled horizon emanating from the surface is a zone of rhizoliths (fossilized root traces). The surface represents a paraconformity or diastem. Both examples from the Cedar Mesa Sandstone, Permian, southeast Utah.

or bioturbation 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 loose 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 airflow becomes fully saturated with sediment, perhaps because of a reduction in windspeed, or because the surface is deflated down to the level of the water table, thereby restricting the availability of loose dry sand for further erosion (Fig. 4.28).

Study techniques

Field experience

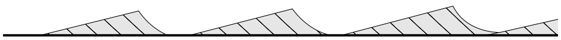
Present-day environments

Field programmes should include investigations of areas of interaction of erosion, transport and deposition that 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 where there have 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.

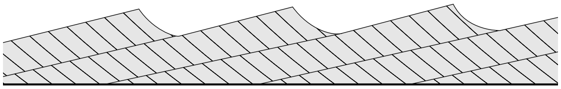
Ancient environments

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 situations, where megaflutes may also exceptionally be present. Slump scars are most common in deltaic and turbidite deposits. Larger erosional structures can often be identified only 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.

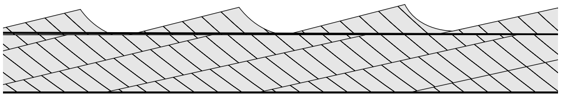
(a) Dunes migrate across former erosion surface; zero angle of climb



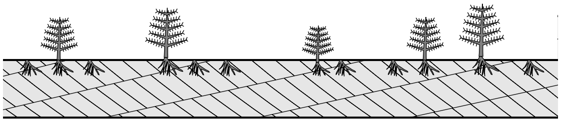
(b) Dunes and interdunes climb preserving cross-bedded sets and bounding surfaces



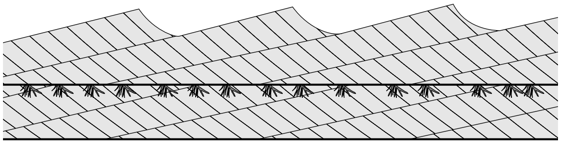
(c) Dunes cease climbing but continue to migrate; net deposition is zero (aeolian bypass)



(d) Deflation to water table; surface colonized by plants and burrowing invertebrates



(e) Renewed aeolian accumulation generates new sequence



(f) New deflation event generates another deflationary supersurface

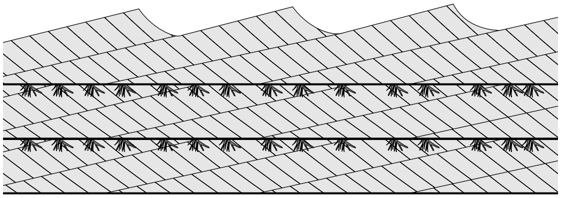


Figure 4.28 Model for the evolution of deflationary supersurfaces in aeolian systems, whereby the wind becomes undersaturated with respect to its potential sand-carrying capacity and cannibalizes the existing sand surface. Sand deflation may result in the concentration of remaining larger clasts, leading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

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.

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CHAPTER 5

Depositional structures in muds, mudstones and shales

5.1 Introduction

The terminology of fine-grain siliciclastic sediments is rather confusing. A range of terms have been used in overlapping and sometimes ambiguous ways. These are discussed quite fully in most books dealing with sedimentary petrology; 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-grain 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 alumino-silicates), grain size rather than mineral composition 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. It 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 the rock record, accounting for about 60 per cent 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-grain 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**), when found within a rock sequence, may be associated with depositional breaks or even unconformities. Other mudstones may result directly from resedimentation of original suspension muds as mudflows (Figs 5.1, 5.2). In many cases, this

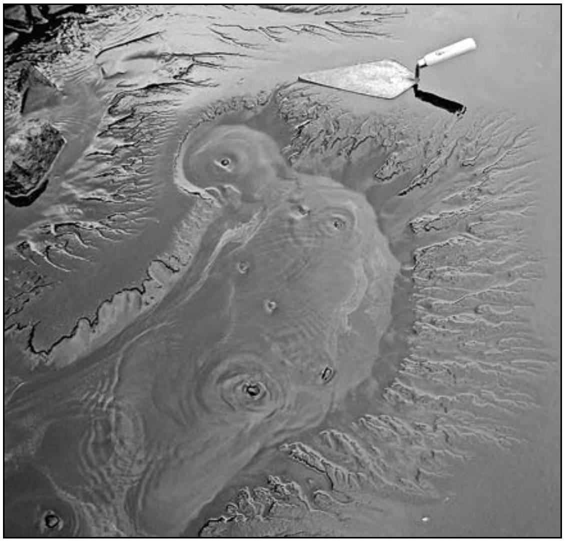

Figure 5.1 A highly fluidized mud explained by high pore-water content; note the water escape features. Jökulsá á Fjöllum, Iceland.

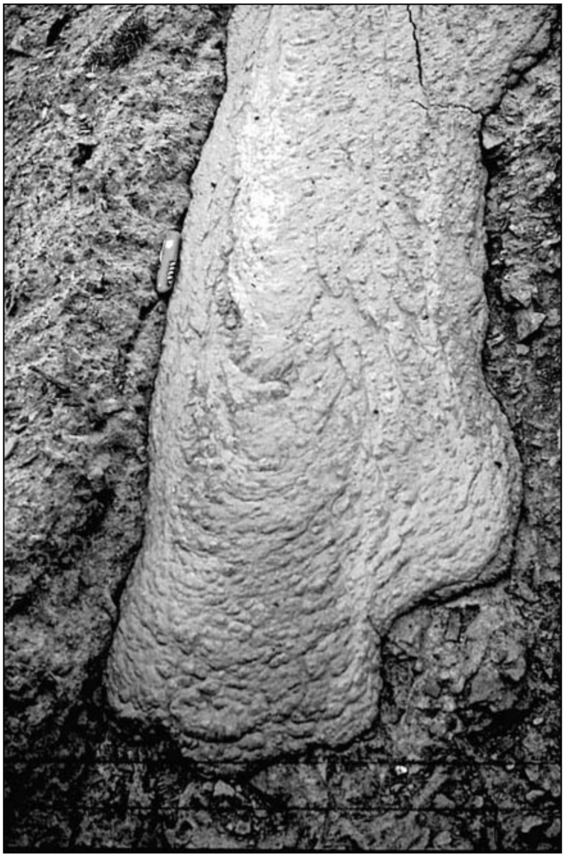

Figure 5.2 A small active mudflow in which water-saturated muds have been remobilized through the addition of water. Note that the surface of the flow is highly irregular because of small aggregates and pebbles being rafted along with the flow. Modern, Svalbard.

movement leads to the incorporation of coarser grains, which tend to "float" within the predominantly muddy sediment (Fig. 5.2; see also §3.7.1).

In addition, fine-grain sediments are generated directly by explosive volcanic activity resulting in both **airfall** and **water-lain tuffs**, which may be prone to subsequent reworking 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 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 world wide, with the paradox that the most powerful processes give rise to extremely thin but laterally widespread horizons in the geological record, which are often used for correlation and dating purposes. However, fine ash may fall close to the volcanic centre as a result of a weak explosion or because of 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 as either 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 compared with those of coarser-grain 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 than for coarser-grain sediments. Secondly, fine-grain sediments, particularly those rich in clay minerals and organic matter, have a much higher initial porosity than most coarse-grain sediments, and this makes them highly susceptible to compaction on burial. This has the effect of distorting and compressing any depositional and organic structures, sometimes 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 organics-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). Carbonate-cemented concretions that formed soon after deposition, before significant burial, sometimes preserve relatively uncompacted depositional structures as well as uncrushed fossils. If concretions occur in a mudstone sequence, 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-grain sediments than on coarser ones. During folding, fine-grain sediments generally behave in an incompetent manner and also readily develop cleavage through rotation and recrystallization of clay minerals,

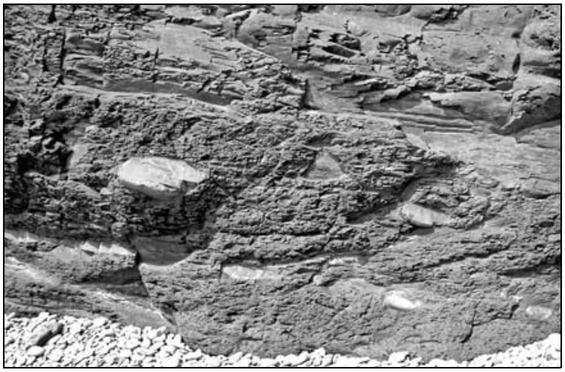

Figure 5.3 Concretions in mudstone. Upper Carboniferous, Amroth, Pembrokeshire, South Wales.

thus obscuring and distorting any original structures and fabrics. Where cleavage development has mostly overprinted any 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 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 of seeing 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-grain sediments is usually based on differences in colour, as the grains themselves are not normally visible. As a general rule, lighter colours indicate coarser-grain 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 would be 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

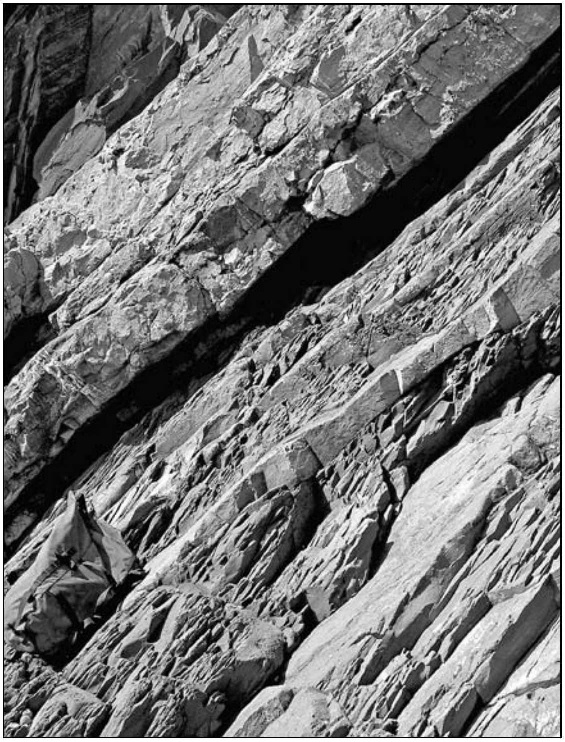

Figure 5.4 Cleaved mudstones interbedded with fine sandstones. The cleavage within the mudstone, which has formed at an angle to the bedding, is a tectonic foliation that has been superimposed on, and has obscured, the original depositional lamination. Old Red Sandstone, Devonian, Pembrokeshire, South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

describe shales that can be split apparently indefinitely.

The lack of any obvious grain-size differences in very fine-grain 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 the chemistry of the particles and the receiving basin. Highly turbulent conditions 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 1 mm 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 differences in grain size, it follows that the process responsible for deposition must have fluctuated in strength, although it is usually difficult to estimate its timescale.

Two possible depositional processes must be considered. The first is that the sediment settles from

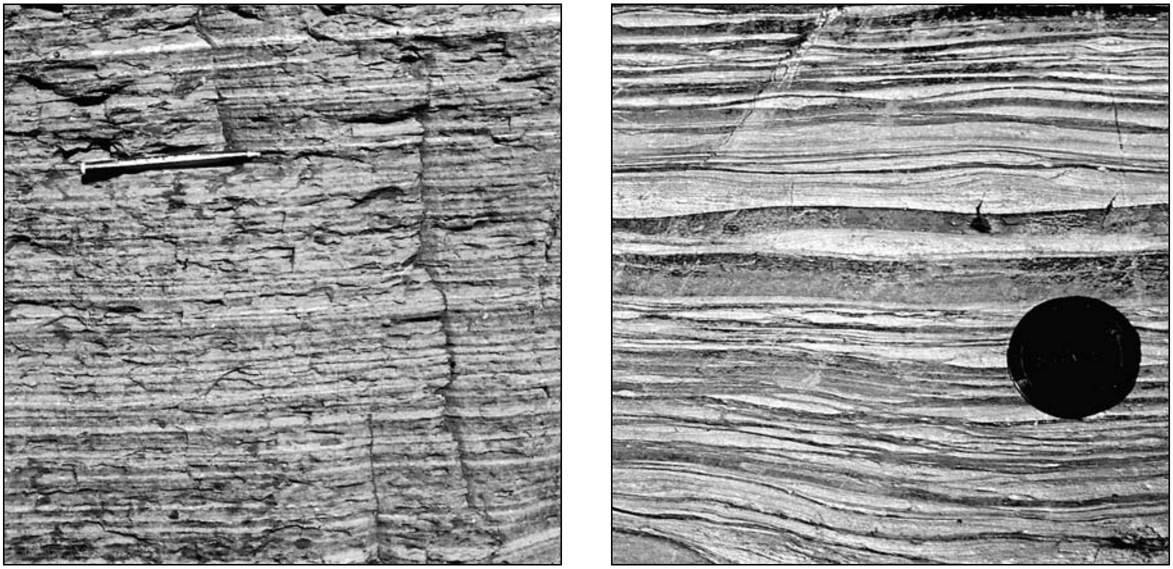

Figure 5.5 Examples of striped siltstones. **(a)** Parallel lamination or thin bedding results from gradational grain-size changes suggesting long-term fluctuations in sediment load. **(b)** Slightly finer and slightly coarser-grain siltstones are thinly interlaminated, probably as a result of regular (perhaps seasonal) fluctuations in the suspended sediment load carried into the basin. Both examples from the Middle Shales, Namurian, Pembrokeshire, South Wales.

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 products of weak, dilute density currents flowing close to the bed, and the finer layers record the background settling of sediment from the water column above. When the laminae are very thin, other features of the overall sequence must be assessed. For example, if lamination occurs in a sequence that has evidence of larger-scale density currents in the form of turbidite sandstones, the inference that the lamination in the mudstones is attributable 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-grain layers may have a substantial clay content, whereas paler coarser-grain layers may contain fine or even very 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 comminuted plant debris or microfossils such as diatoms or radiolaria.

These units usually have parallel sides 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 timescale too large to be attributed to sweep and burst mechanisms or to other short-period fluctuations in an otherwise steady flow. Seasonal or other climatic factors may control sediment discharge to deltas, lakes and river basins, where fine-grain sediments are common. The gradational contacts suggest gradually increasing and waning high-discharge episodes rather than sudden "events", for example slump-triggered turbidity currents. It is again impossible 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 influx during summer months when rates of glacial meltwater influx are high. By contrast, such lakes typically have a frozen surface during winter months, when significantly reduced sedimentation rates are determined by the rate of fallout of mud from suspension in the deeper (non-frozen) lake waters. This example of seasonally controlled sedimentation gives rise to **varves**, which form a particular type of rhythmite succession. In larger lakes, seas and oceans, deep bottom-hugging currents, which typically flow parallel to the base of slopes, act as effective mechanisms for the transport of fine-grain 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-grain 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

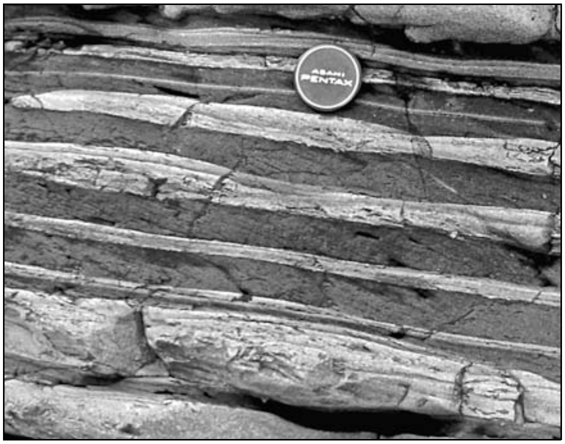

Figure 5.6 Interlaminated mudstones and fine sandstones. The thin sandstones are sharply based and have an internal lamination related to Bouma sequence bedforms (see §6.8) and a slightly lenticular form because of ripples on their upper surfaces. The sandstones are deposits of episodic high-energy events. The mudstones between record quiet background sedimentation from suspension. Bude Formation, Namurian, north Cornwall. (Photo courtesy of Gilbert Kelling)

layers suggest that they represent relatively sudden events superimposed upon the background of quieter, more constant sedimentation of the finer-grain darker layers. The internal grading and the gradational tops of coarse-grain layers suggest waning of the suspended load during more active episodes. The small-scale irregular morphology on their bases is probably mainly explained by loading of the silts into soft waterlogged clays (see §9.2.1), although minor erosive surfaces may also occur.

5.2.6 Structureless mudstones

Some mudstones show no obvious lamination, bedding or fissility, irrespective of their weathering state. In some 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. Often, there is little to describe in these rocks, but they still merit careful examination, as their lack of lamination may be attributable to one of several possible causes. It may reflect an original lack of depositional layering (i.e. continual steady deposition) or it could be the result of later destruction of layering. On fresh surfaces of some apparently homogeneous mudstones, mud clasts in a similar matrix are sometimes evident.

Lack of original layering in water-lain mudstones may be attributable to a very homogeneous and possibly rather rapid depositional process or to a lack of platy grains. 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 accumulations of windblown 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 can sometimes 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. It is often 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 be attributable to soil-forming processes.

Large-scale post-depositional movement of thick mud beds is common, particularly in rapidly deposited sequences giving rise to slump deformation (§9.2.2). Muds buried under younger sediments commonly flow both vertically and laterally to give diapiric structures (see Ch. 9), 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 with a domination of sand 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.

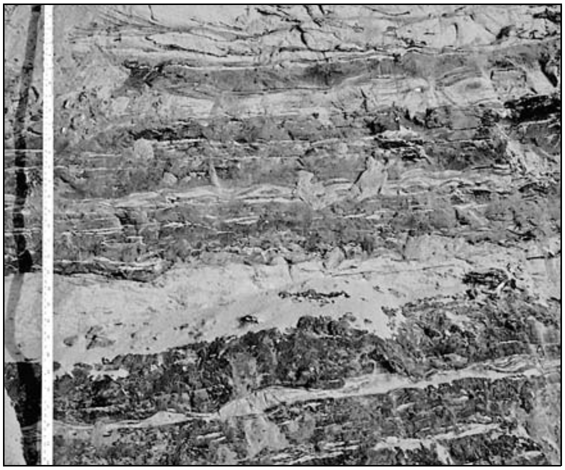

Figure 5.7 Interlaminated mudstone and sandstone. The two components are very clearly different: fine mud and clean well sorted sand. In the mud-dominated part, the sand occurs as thin laminae and lenses; in the sand-dominated part, the mud occurs as thin but discrete drapes. These patterns are typical of tidally influenced sediments, where the sand records the strong currents of the flooding and ebbing tide, and the muds record the quiet slack-water periods when mud can fall from suspension, probably aided by flocculation. Lower Greensand, Aptian–Albian, Isle of Wight.

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 often be unrewarding, because 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 (turbidity) flows can often be generated at the edges of still, clear ponds by disturbing small masses of sediment at their edges. Mudflows can often be seen in cliffs and excavations in muddy sediments. They should be approached with caution, especially after heavy rain.

Ancient environments

Shales and mudstones of various environmental origins are commonly encountered during field excursions. Interpretation of the processes of origin of these rocks is often limited. Ascribing a palaeoenvironment to the sediments often depends upon evidence of body and trace fossils (Ch. 9), upon physical features in mudstones described in other chapters (e.g. post-depositional structures, Ch. 8) 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), grain size, temperature, viscosity, relative density of medium, relative density of grain, velocity of fall, etc.

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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 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. Deposition of sand generally occurs through accumulation from bedload transport during steady flow with excess sediment supply, or by fall-out from suspension from powerful decelerating currents. After 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. However, most pyroclastic deposits are of coarser grain and are described in Chapter 7.

6.1 Ripples and cross lamination

6.1.1 Introduction

Ripples are quite regularly spaced undulations on a sand surface or on a sandstone bedding plane. Their spacing (wavelength) is usually less than 0.5 m and relief seldom exceeds 3 cm. 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 on both bedding planes and vertical surfaces. Patterns of cross lamination are often 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-grain 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 profile and 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 attributable to current activity, the reality is rather more complex. The shape and continuity of ripple crestlines is at least as important for interpretation.

A whole 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. Although 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. 6.5c) and those with a strongly three-dimensional shape (e.g. linguoid ripples, Fig. 6.5d) usually have asymmetrical profiles. They 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). However, there is a continuum of asymmetrical current ripples ranging in shape from straight crested through sinuous crested to linguoid

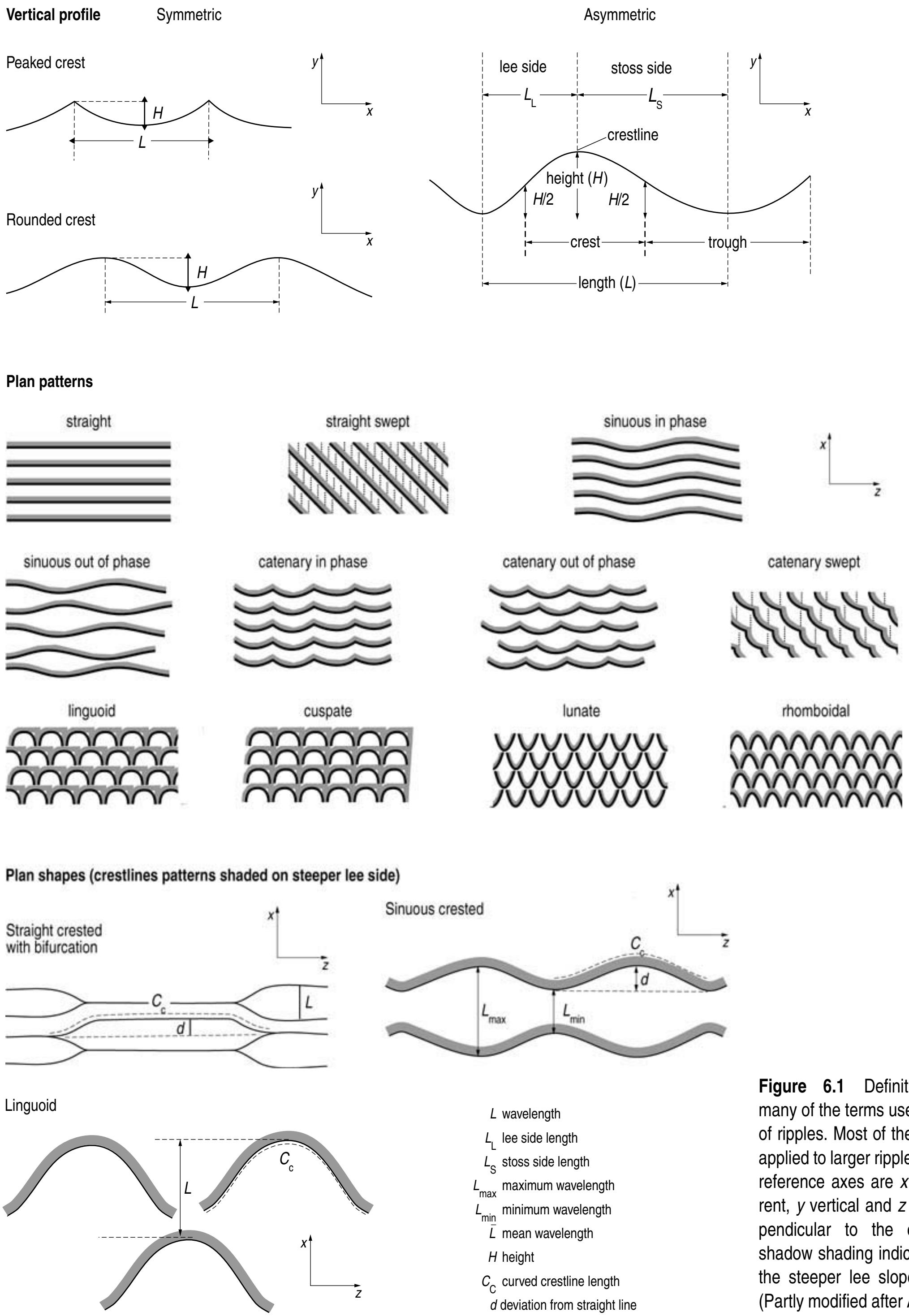

Vertical profile
Symmetric
Asymmetric
Peaked crest
H
L
y
x
Rounded crest
lee side
stoss side
L_L
L_S
crestline
height (H)
H/2
H/2
crest
trough
length (L)
Plan patterns
straight
straight swept
sinuous in phase
x
z
sinuous out of phase
catenary in phase
catenary out of phase
catenary swept
linguoid
cuspate
lunate
rhomboidal
Plan shapes (crestlines patterns shaded on steeper lee side)
Straight crested
with bifurcation
Sinuous crested
C_c
d
L
L_{max}
L_{min}
Linguoid
L wavelength
L_L lee side length
L_S stoss side length
L_{max} maximum wavelength
L_{min} minimum wavelength
$\bar{L}$ mean wavelength
H height
C_C curved crestline length
d deviation from straight line

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 grey shadow shading indicates the position of the steeper lee slope of the bedforms. (Partly modified after Allen 1968)

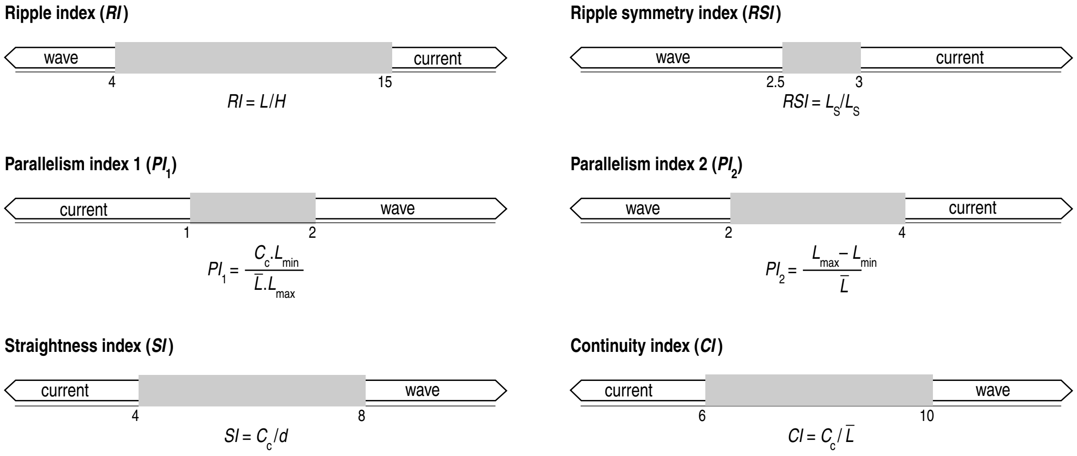

Ripple index (RI)
wave
current
4
15
$RI = L/H$
Ripple symmetry index (RSI)
wave
current
2.5
3
$RSI = L_S/L_S$
Parallelism index 1 (PI_1)
current
wave
1
2
$PI_1 = \frac{C_c.L_{min}}{\bar{L}.L_{max}}$
Parallelism index 2 (PI_2)
wave
current
2
4
$PI_2 = \frac{L_{max} - L_{min}}{\bar{L}}$
Straightness index (SI)
current
wave
4
8
$SI = C_c/d$
Continuity index (CI)
current
wave
6
10
$CI = C_c/\bar{L}$

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 Figure 6.1 for definition of parameters.

(Fig. 6.5). Associated ridges and hollows on the stoss sides of ripples are aligned roughly parallel to flow. Scour pits occur in the lee of ripples, commonly in front of a downstream embayment in the crestline or down stream of the gap between two linguoid ripples.

On some beaches, ripple forms with low relief have very marked repeated rhomboidal 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 1 cm. They are asymmetrical in profile, being highest at their downstream point.

Symmetrical ripples and those with very straight and continuous crestlines are associated with wave action (Fig. 6.3). They generally lack scour pits, and their crestlines show zig-zag junctions or bifurcation (Fig. 6.6b). They may be either smoothly rounded (Fig. 6.5a) or quite sharply peaked in profile (Fig. 6.6b). 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.

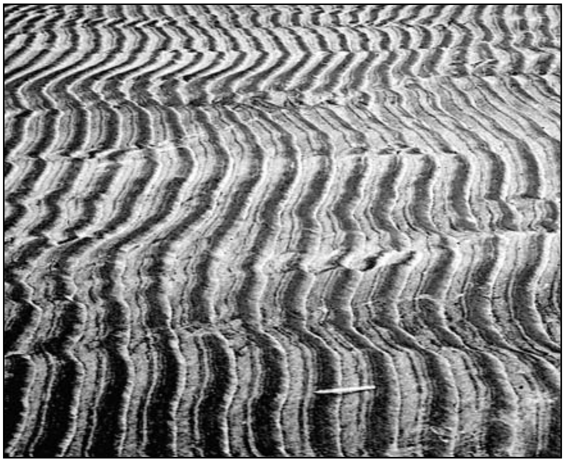


Figure 6.3 Rounded straight-crested symmetrical ripples, Tana, Norway.

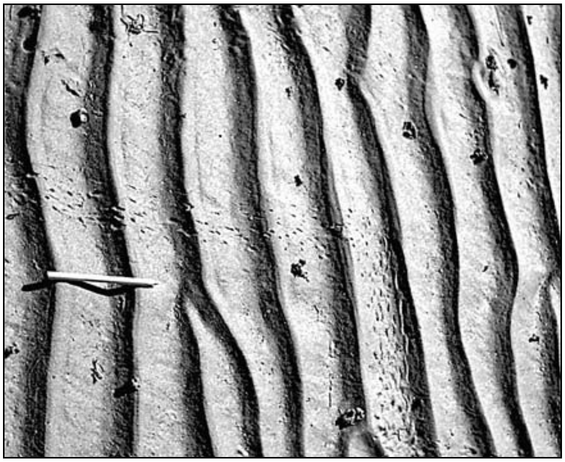


Figure 6.4 Asymmetrical straight-crested ripples, Kirkland, Scotland. Also note worm tubes and crab tracks.

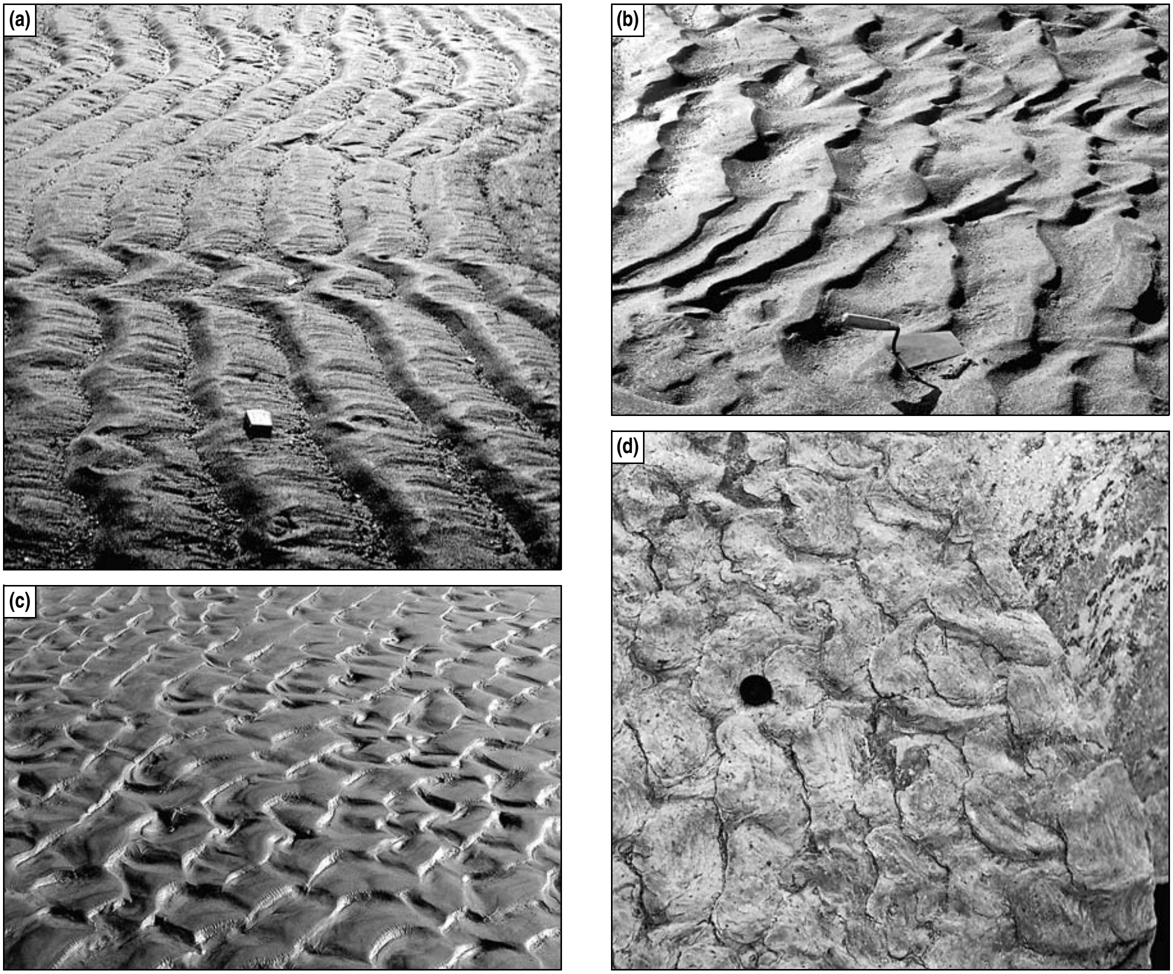


Figure 6.5 Examples of asymmetrical ripples with various morphologies. **(a)** Asymmetrical wave ripples with straight crestlines, Tana, Norway. **(b)** Straight to sinuous crested asymmetrical ripples, Tana, Norway. **(c)** Sinuous crested out-of-phase asymmetrical ripples, Myrdalssandar, southern Iceland. **(d)** Paired linguoid ripples, Scalby Formation, Middle Jurassic, east Yorkshire.

It is important to try to distinguish between current and wave ripples on the basis of symmetry and crestline continuity, without expecting 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).

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).

In order to develop descriptive powers and understanding, it can be valuable to study photographs of ripple surfaces like those in Figure 6.8, or better still to visit a modern rippled sand environment and describe, measure and interpret the ripples found there.

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). The recognition of such lamination is valuable in interpreting rock sequences. In some sequences of interbedded sand and finer sediment,

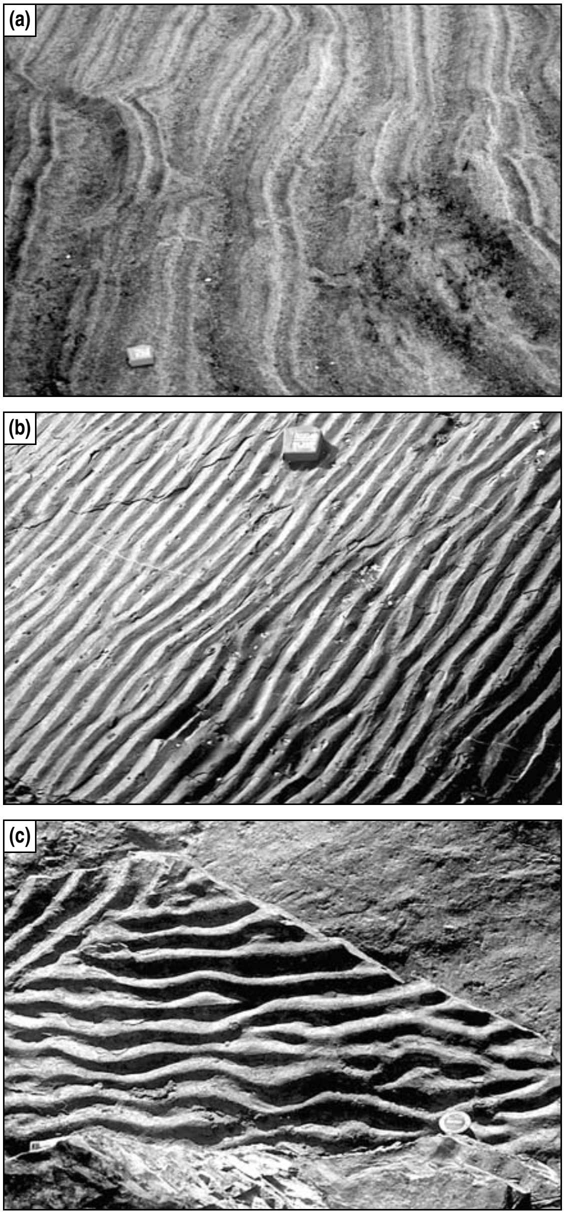


Figure 6.6 Examples of symmetrical ripples with various morphologies. **(a)** Multiple peaked ripples, Tana, Norway. **(b)** Straight crested wave ripples with tuning-fork intersections, Coal Measures, Westphalian, Pembrokeshire. **(c)** Sinuous crested symmetrical ripples, Silurian, Cantabria, northern Spain.

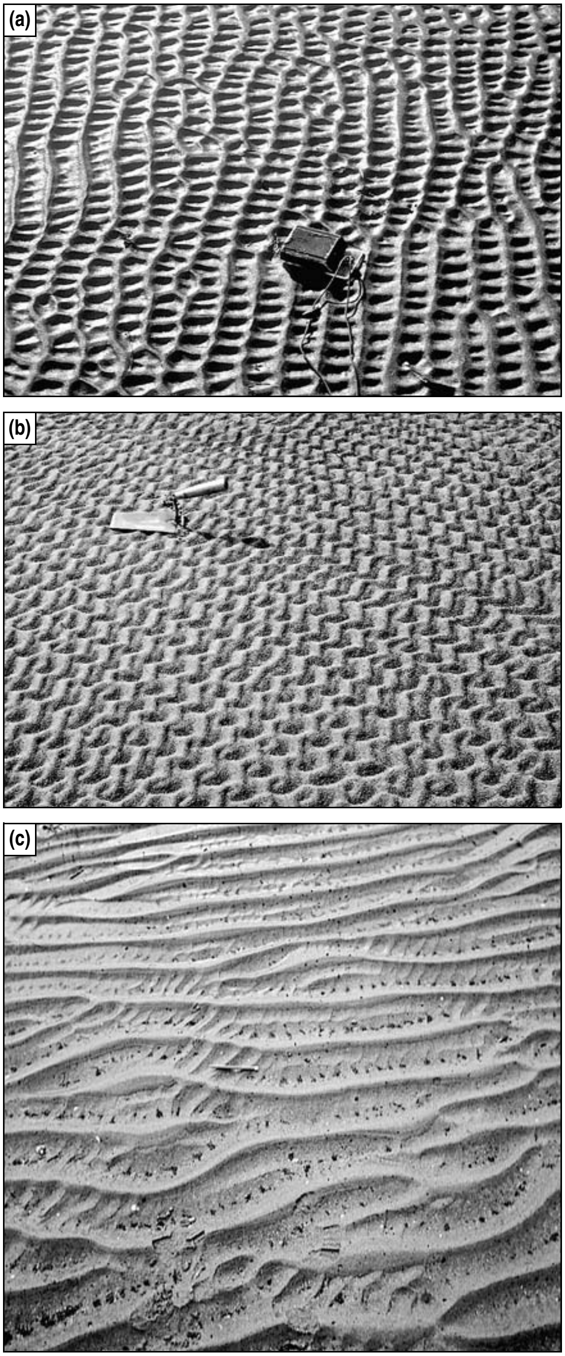


Figure 6.7 Examples of ripple interference patterns. **(a)** Ladder ripples, Tana, Norway. **(b)** Interference ripples, Tana, Norway. **(c)** Wave ripples with a secondary perpendicular set developed in the troughs (ladder ripples), Aberlady Bay, East Lothian, Scotland.

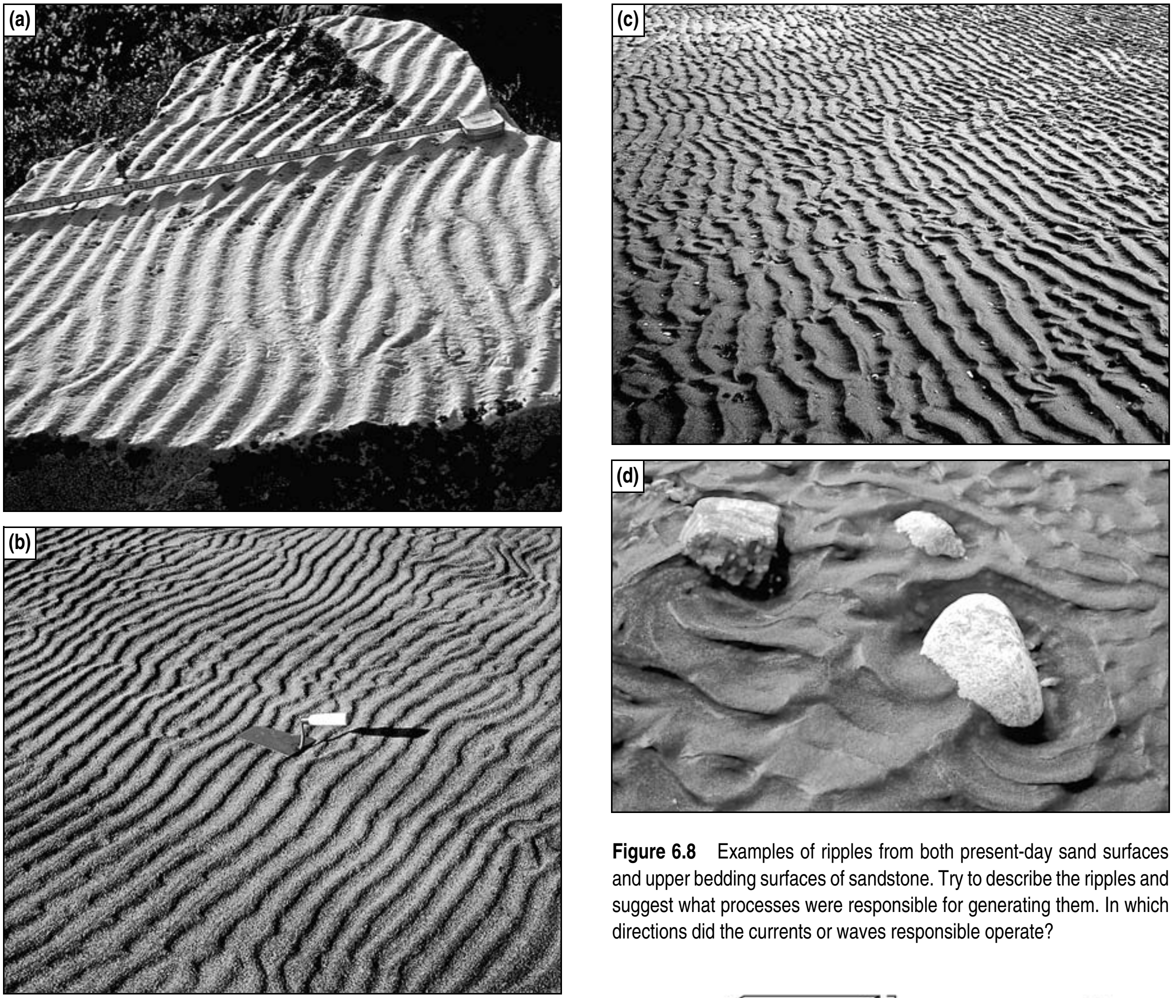


Figure 6.8 Examples of ripples from both present-day sand surfaces and upper bedding surfaces of sandstone. Try to describe the ripples and suggest what processes were responsible for generating them. In which directions did the currents or waves responsible operate?

sand ripples are isolated in mud or silt or are preserved as morphological features on the top surfaces of thicker sandbeds. 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 2–3 cm 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 are commonly of trough or spoon shape, being most strongly concave upwards transverse to the mean foreset dip and more gently curved parallel to the dip.

Both exposed wind-deflated sand surfaces and many

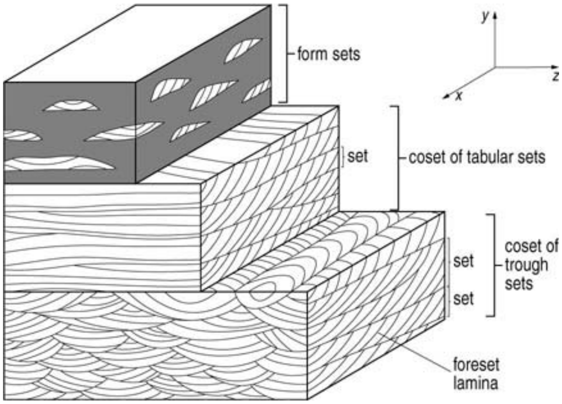


Figure 6.9 Definition diagram for the basic types of cross lamination. The same terms apply at a larger scale to cross bedding. (After Allen 1968)

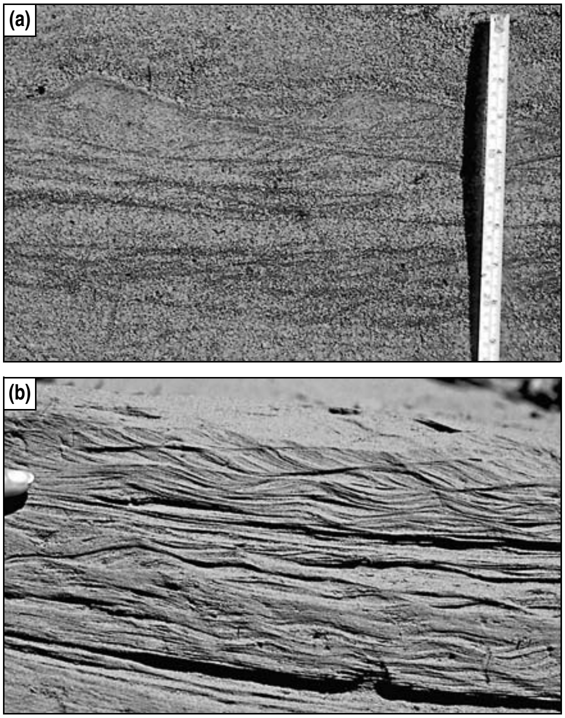


Figure 6.10 Examples of ripple cross-laminated sands with ripple form sets preserved. **(a)** Tana, Norway; transport from right to left. **(b)** Fish River Canyon, Namibia; transport from left to right. In each example, the cross lamination dips down in the direction of bedform migration, reflecting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

bedding planes in ancient, medium- to fine-grain 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 8 cm wide and 20–30 cm long, although sometimes 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 often 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 be understood only by a full appreciation of its three-dimensional nature, as different orientations of vertical

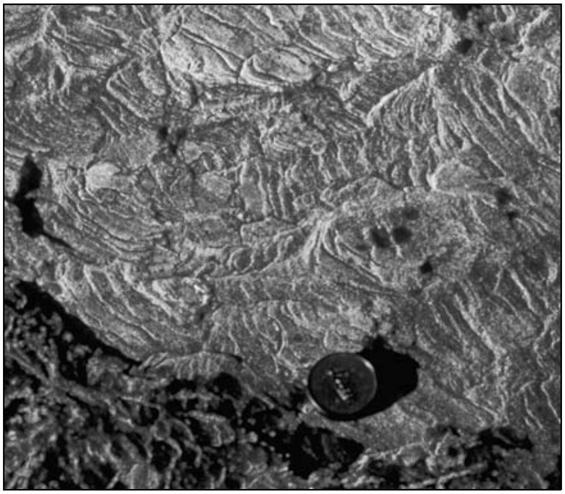


Figure 6.11 Upper bedding surface of sandstone showing "rib and furrow", the horizontal expression of trough cross lamination. Compare this structure with the more idealized view shown on top of the lowest block in Figure 6.9, and hence determine current direction. Central Clare Group, Namurian, County Clare, Ireland.

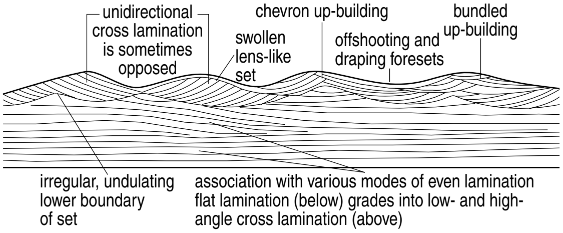

unidirectional cross lamination is sometimes opposed
swollen lens-like set
chevron up-building
offshooting and draping foresets
bundled up-building
irregular, undulating lower boundary of set
association with various modes of even lamination flat lamination (below) grades into low- and high-angle cross lamination (above)

Figure 6.12 Some of the features that help in the diagnosis of wave-ripple cross lamination (after de Raaf et al. 1977). Note that adjacent ripples may exhibit varying direction and degree of ripple asymmetry, and that adjacent ripple sets may be quite dissimilar.

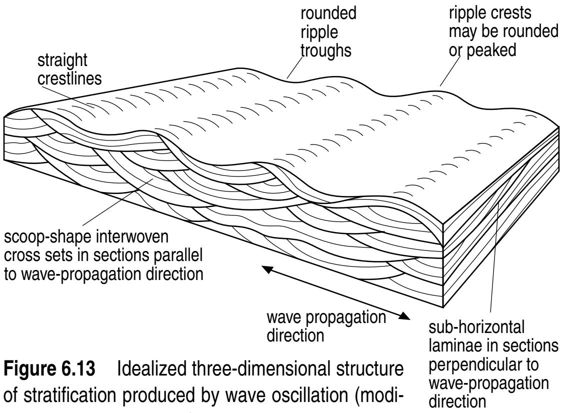

straight crestlines
rounded ripple troughs
ripple crests may be rounded or peaked
scoop-shape interwoven cross sets in sections parallel to wave-propagation direction
wave propagation direction
sub-horizontal laminae in sections perpendicular to wave-propagation direction

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

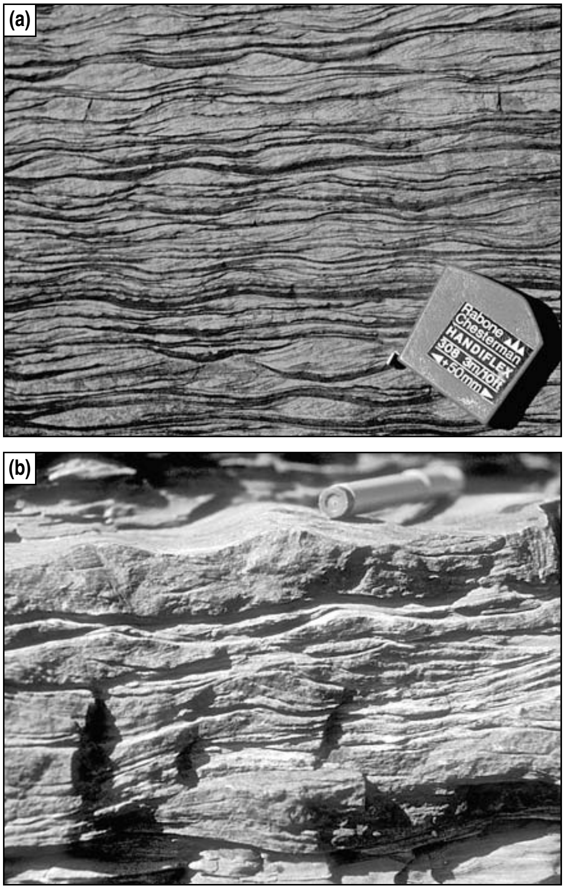

(a)
Rabone Chesterman HANDIFLEX 308 3m/10ft 50mm
(b)

Figure 6.14 Patterns of internal lamination associated with wave action. **(a)** Wave-modified ripples in heterolithic strata, Coal Measures, Westphalian, Hartley, Northumberland, England. **(b)** Wave ripples showing internal lamination and surface form, Central Clare Group, Namurian, County Clare, Ireland.

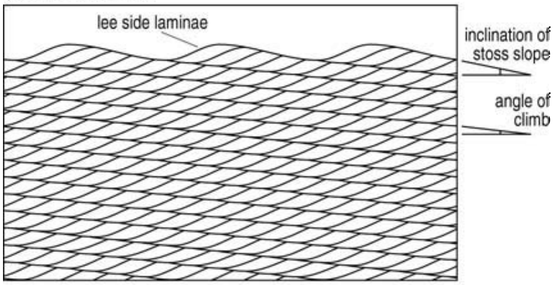
section (i.e. of exposure) give rather different patterns of lamination. It is best understood by reference to block diagrams (see Fig. 6.9), which should be studied with the aim of imagining how the lamination would appear in vertical sections with different orientations.

Two important varieties of cross lamination warrant separate discussion.

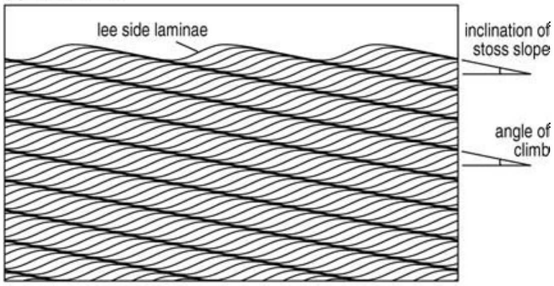
Climbing-ripple cross lamination (ripple drift)

In most cross-laminated sediment, boundaries between successive sets are erosive and roughly horizontal. In other 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 **ripple drift** or **climbing-ripple cross lamination**. The **angle of climb** of the ripple sets is determined by the ratio between the rate of downstream ripple migration and the rate of rise

(a) Subcritical climb


lee side laminae
inclination of stoss slope
angle of climb

(b) Critical climb


lee side laminae
inclination of stoss slope
angle of climb

(c) Supercritical climb

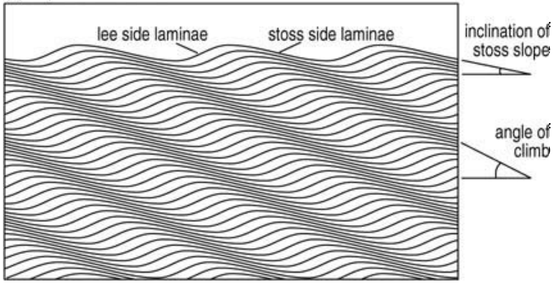

lee side laminae
stoss side laminae
inclination of stoss slope
angle of climb

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 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.

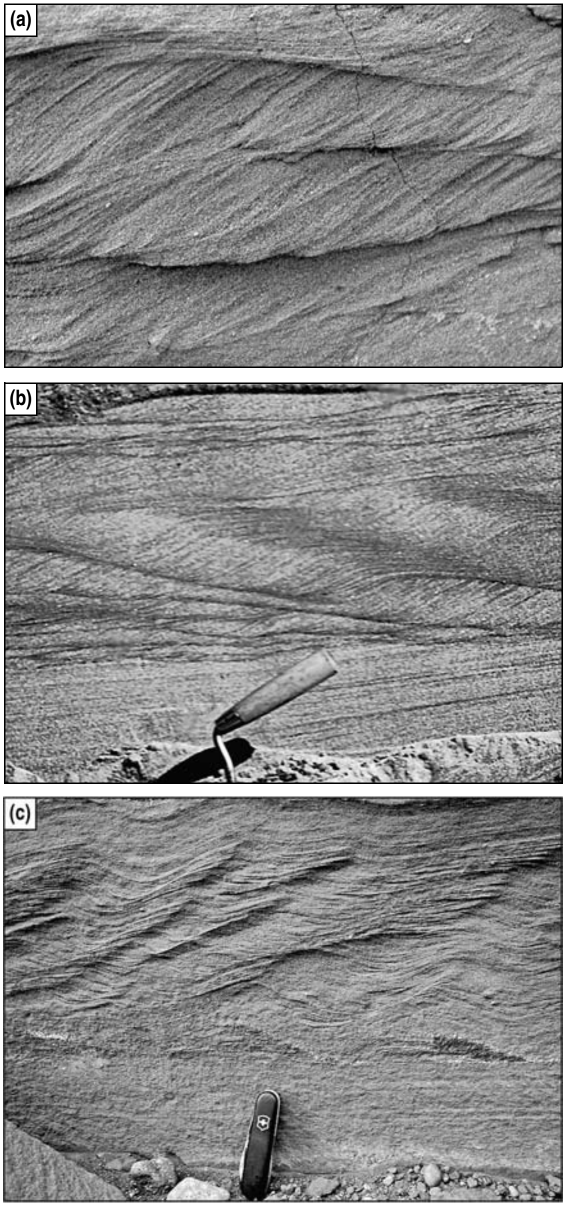


Figure 6.16 Examples of climbing ripple lamination produced by bedforms climbing at various angles. **(a)** Subcritical ripple climb, where only the lower part of the foreset is preserved. In this example, the structure passes to the left into ripple form sets; modern fluvial deposits, central Iceland. **(b)** Critical ripple climb, where the complete ripple form is just preserved, although with some supercritical climbing on the right; migration from right to left; modern fluvial deposits, Tana, Norway. **(c)** Supercritical ripple climb, where the climb angle is steep enough to preserve both the lee and the stoss slope of the bedform; migration and climb is from left to right; modern fluvial deposits, central Iceland.

of the accumulation surface (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 and only their basal part is preserved to form a set of climbing-ripple cross lamination. **Critical climbing** occurs where the angle of climb matches the angle of the stoss slope of the ripples exactly, so that the entire bedform is 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. When stoss-side laminae are preserved, it is 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 down stream while the bed accreted vertically (Fig. 6.15). The geometry of the cross-lamination therefore is an indication of the rate at which the bed aggraded vertically, with higher angles of climb (supercritical) being associated with the highest rates of aggradation (Figs 6.17, 6.18).

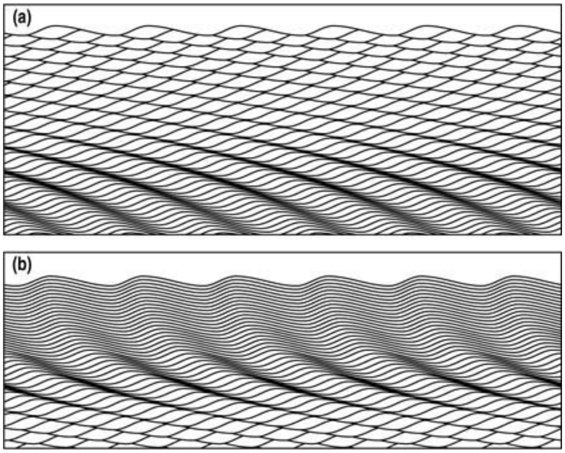


Figure 6.17 Computer-generated patterns of ripple cross lamination, showing the effects on preserved stratification style of changes in the angle of bedform climb. **(a)** Decrease in angle of climb from supercritical at the base to subcritical at the top. **(b)** Increase in angle of climb from subcritical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

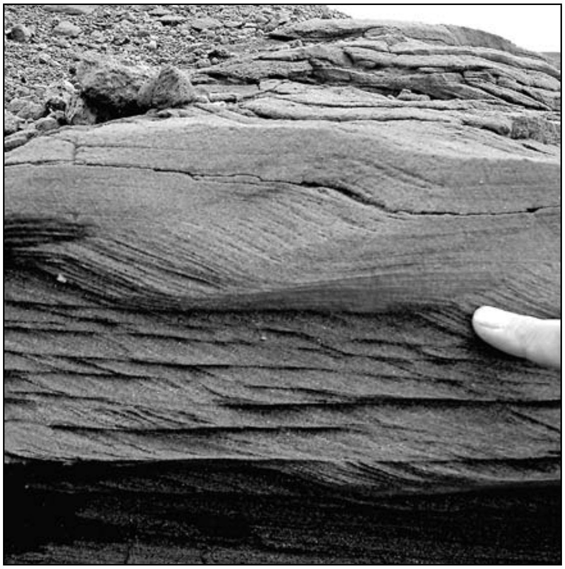


Figure 6.18 An example of ripple cross lamination that undergoes an increase in angle of climb from subcritical in the lower part of the section to supercritical in the upper part; modern fluvial deposits, central Iceland.

Flaser, lenticular and wavy bedding

In some units of ripple cross-laminated sand, the pattern is broken up by interlaminations and lenses of finer-grain sediment (silt and mud) (Figs 6.19, 6.20). Where sand dominates, as in **flaser bedding**, 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-grain 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-grain sediment, with the general term “**wavy bedding**” often being used for the description of ripple forms characterized by 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 sandbed, as unidirectional currents, as oscillatory waves or as a combination of both, may give rise to ripples.

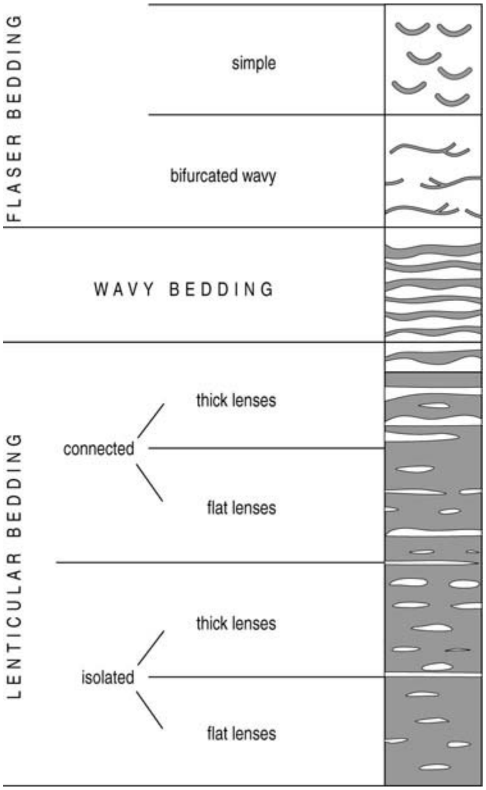


Figure 6.19 Variety interlamination resulting from mixed lithologies of sand and mud (after Reineck & Singh 1973).

Unidirectional water currents

When the velocity of water flowing over a sandbed exceeds a certain critical value, grains begin to move (see §3.5). With widespread movement of grains finer than about 0.6 mm 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.

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

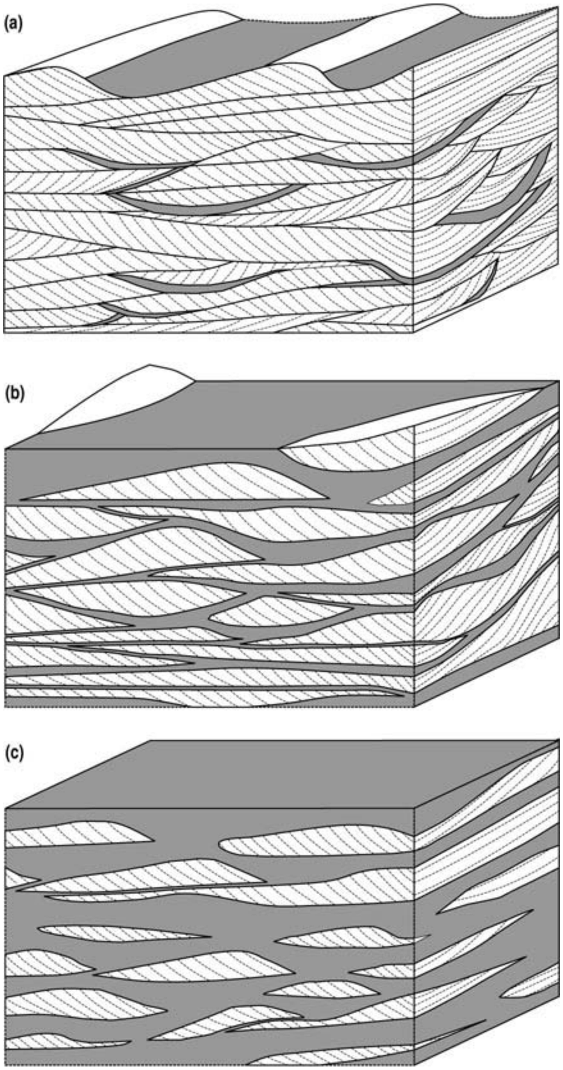


Figure 6.20 (above) Styles of interbedding in mixed sand and mud lithologies: **(a)** flaser bedding, **(b)** wavy bedding, **(c)** lenticular bedding. The progression from (a) to (c) results from a net decrease in current speed and increased deposition and preservation of mud drapes. (Modified after Reineck & Singh 1980)

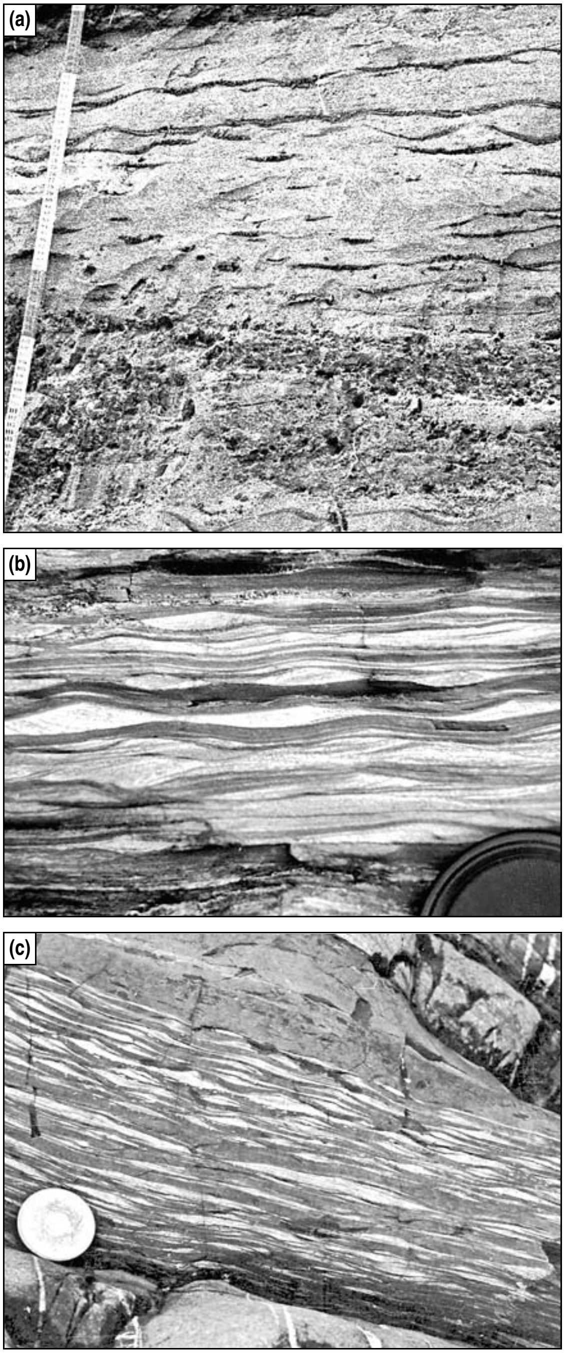


Figure 6.21 (right) Examples of the variety of cross-lamination structures resulting from mixed lithologies of sand and mud. **(a)** Flaser bedding, with mud drapes predominantly in the ripple troughs; note conglomerate of cohesive mud clasts in the lower part of the section; modern tidal flat, Haringvliet, Netherlands. **(b)** Wavy bedding with mud completely draping sandy, wave-influenced ripple forms; Central Clare Group, Namurian, County Clare, Ireland. **(c)** Lenticular bedding with rather peaked and isolated symmetrical ripple form sets. Northam Formation, Upper Carboniferous, North Devon, England.

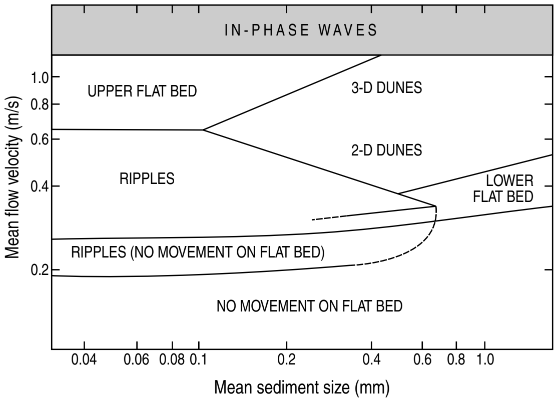


Figure 6.22 The stability fields of different bedforms in relation to flow velocity and grain size for water depth of 0.2 m. Note that the plot is on a log–log scale. For other depths, the positions of the lines are shifted and a three-dimensional plot is needed to illustrate the bedform distribution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

sand giving larger ripples. Current ripples are most readily envisaged as forming in shallow water, but they are also produced in deep water through the action of ocean-bottom currents. Turbidity currents (see §3.7.2) 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.

The variety of ripple shape probably relates to the structure of water turbulence close to the bed. Ripples with straighter crests and crestlines transverse to the flow have a rotating eddy in the lee of the ripple, with its axis of rotation parallel to the ripple crestline (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 also 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 re-attachment of the flow and to see how the flow re-attaches to give an area of scour down stream from each ripple. In the case of straight-crested ripples the re-attachment is generally in a continuous zone, but with linguoid ripples re-attachment is concentrated in scour pits from where grains move centrifugally. 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 to give the lee face of the ripple a tangential base. Most of the sand swept from a scour pit moves down current to supply the next ripple down stream. 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 farther out onto the lee side by a process of **grainfall**. The grains' trajectories are influenced by the strong eddies in the separation zone. Together, these two processes generate the cross laminae that record the migration of the lee face of the ripple. The downstream movement of linguoid ripples with scour-pit and lee-face couplets generates trough cross lamination (see 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 down stream, but are preserved only 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 ripple migration – the steeper the angle, the higher the rate of aggradation. When the

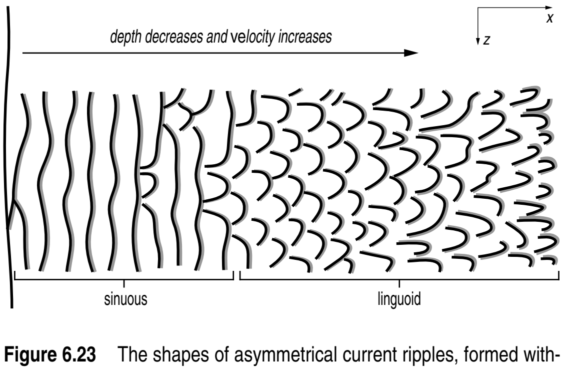


Figure 6.23 The shapes of asymmetrical current ripples, formed without wave influence, related to water depth and velocity; plan view (after Allen 1968).

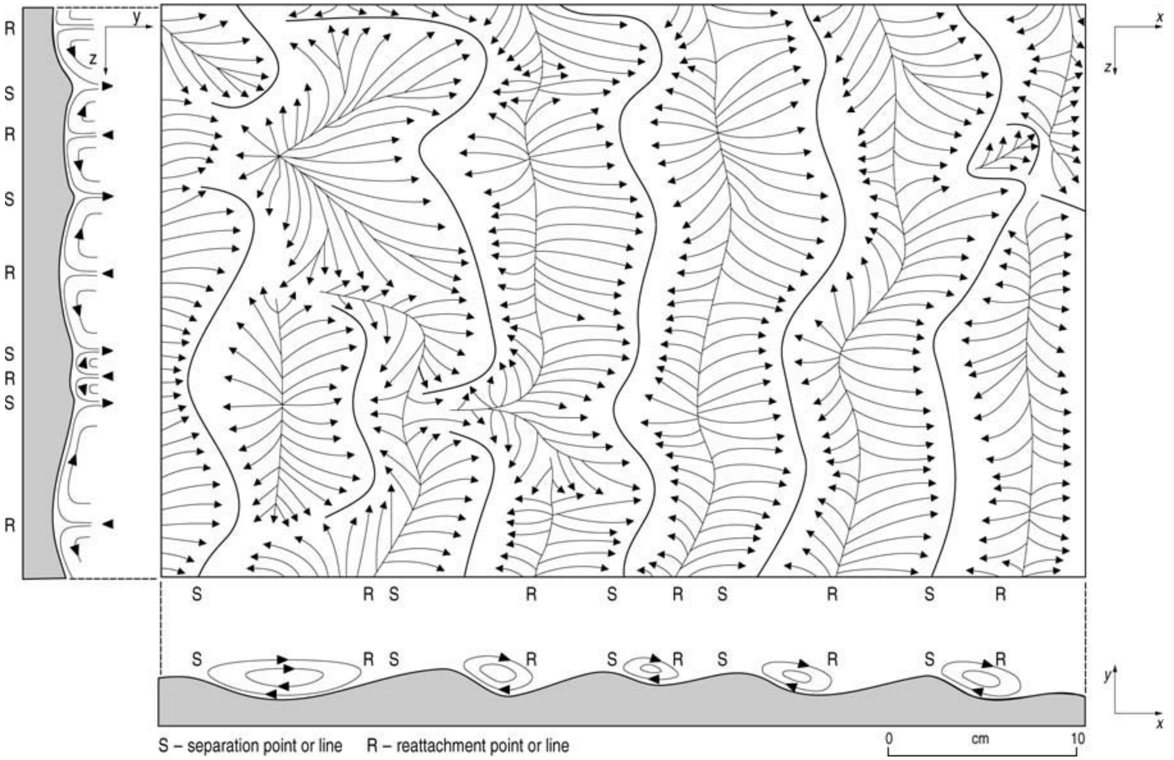


Figure 6.24 The pattern of water movement close to the bed over a field of asymmetrical current ripples, showing flow directions at the bed; heavier lines represent ripple crestlines. These directions will be similar to the grain-movement directions. Flow separates at the ripple crest and re-attaches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

angle of climb exceeds the slope of the ripple stoss side, supercritical climbing takes place and stoss-side laminae are preserved (see Figs 6.15, 6.16, 6.17, 6.18, 6.25).

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. 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). 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 in the water as each wave passes and **vortex ripples** develop (Figs 6.3, 6.6, 6.26).

The size, spacing and symmetry of wave-generated ripples appear to be controlled by four principal factors defining 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

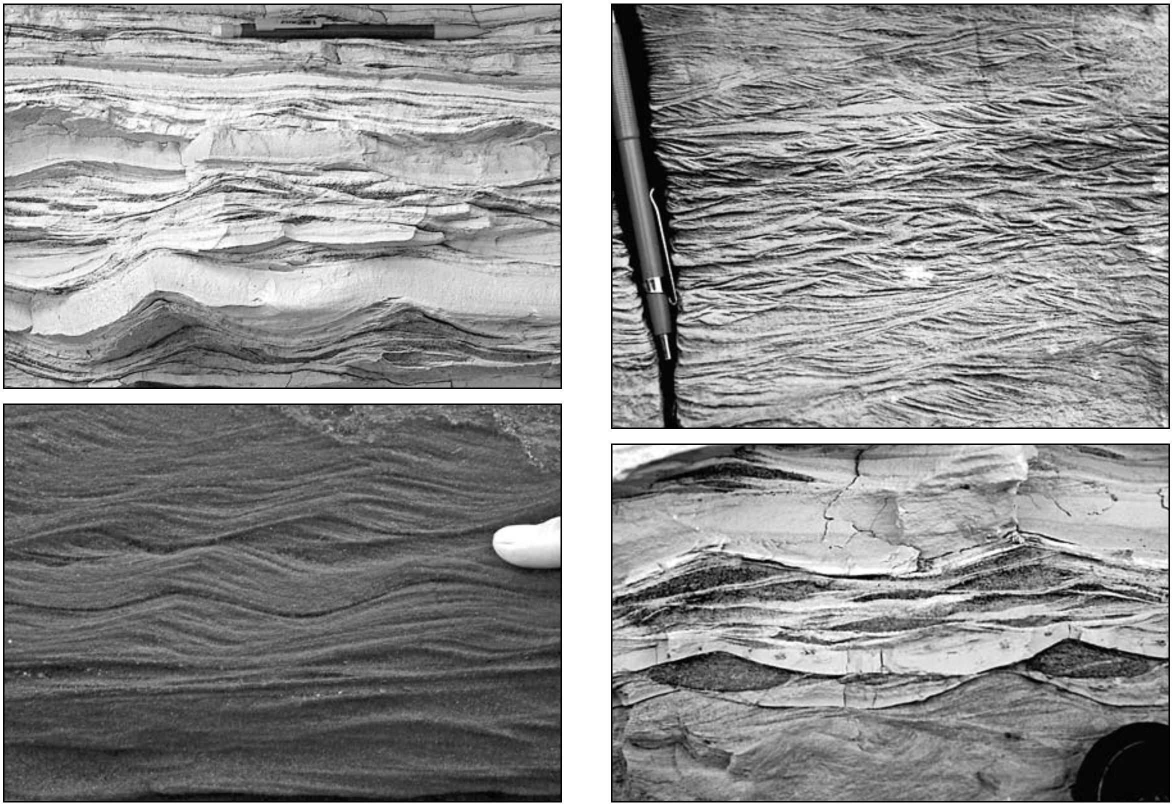


Figure 6.25 Examples of various types of ripple cross lamination. Suggest what has happened in terms of depositional process in each case.

fall between those that give no movement and those that give a flat bed (Fig. 6.27). The asymmetry of the orbital velocity determines the boundary between symmetrical and asymmetrical ripples, greater velocity asymmetry giving more asymmetrical ripples. In shallow offshore areas, where waves are shoaling, a zonation of ripple types can sometimes be recognized (Fig. 6.28).

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. 6.5a,b).

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, §6.4.4, Fig. 6.27).

The shapes of symmetrical ripples are mainly 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 crests of what might otherwise have been sinuously

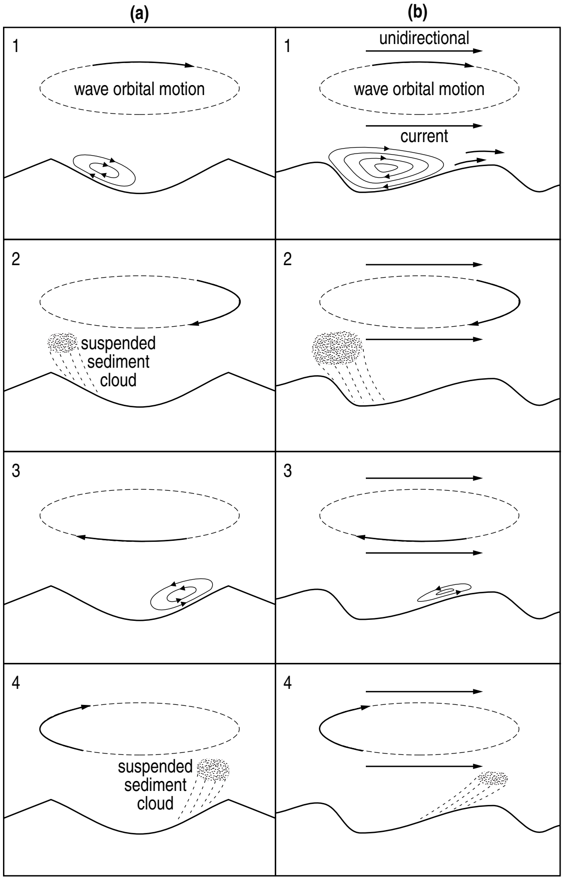

(a)
(b)
1
wave orbital motion
1
unidirectional
wave orbital motion
current
2
suspended sediment cloud
2
3
3
4
suspended sediment cloud
4

Figure 6.26 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 & Bowen 1963)

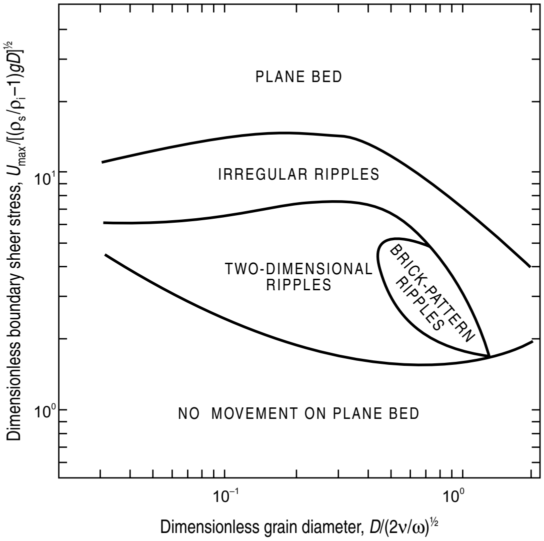


Figure 6.27 The occurrence of different types of bedform as a result of waves acting in a straight channel, under different conditions of wave strength and sediment grain size. U_{max} is the maximum orbital velocity close to the bed, D is particle diameter, ρ_s and ρ_i are solid and fluid densities, ν is kinematic viscosity of the fluid, and ω is the angular frequency of the waves. (After Kaneko 1980)

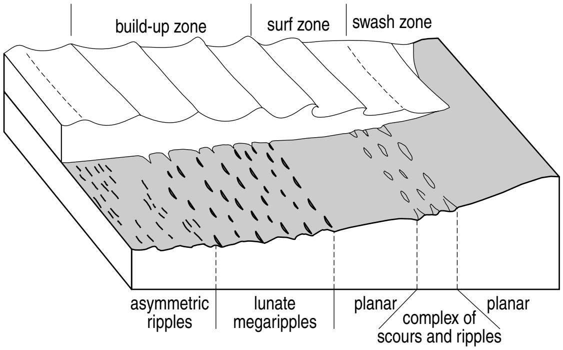

build-up zone
surf zone
swash zone
asymmetric ripples
lunate megaripples
planar
complex of scours and ripples
planar

Figure 6.28 Zonation of wave-generated bedforms off shore 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)

crested current ripples. When wave and current directions diverge, interference patterns develop (e.g. Fig. 6.7). 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 Wind ripples

Three morphologically distinct types of small-scale ripple occur on present-day windblown surfaces. Each is also recognized in vertical section where it produces 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-grain fraction of the sand upon which they develop. They usually have wavelengths of 5–20 cm and heights of 5–10 mm. They exhibit a high ripple-form index (see Fig. 6.2) and have straight or sinuous continuous crestlines 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 angles, below the angle of rest (Fig. 6.29).

Impact ripples develop through a combination of saltation and reptation (§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 reptate (hop) down wind. For a given wind velocity, reptating grains in motion are restricted to a narrow size range and the distance that they jump down wind (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. 6.30) and thus ripple spacing is proportional to wind velocity. Minor surface perturbations act as the seed 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. 6.30). 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 down wind 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 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 a high ripple form index (ratio of wavelength to height) of 25–40+, and are often characterized by inverse grading resulting from the migration of coarser-grain ripple crests over finer-grain ripple troughs. More complex ripple patterns may develop from the merging of smaller and larger ripples with different movement rates.

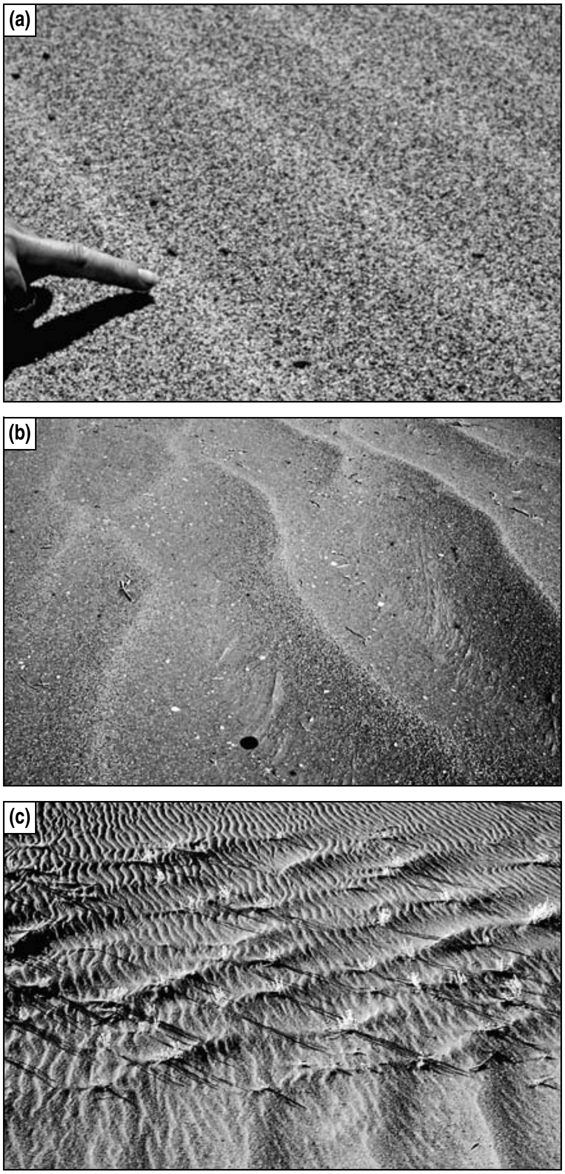


Figure 6.29 Some examples of aeolian impact ripple morphology. **(a)** Straight-crested wind ripples in well sorted medium sand; note occasional granules; Namib Sand Sea, Namibia. **(b)** Sinuous-crested wind ripples with a distinct bimodal grain-size distribution; note how coarser grains are confined to ripple crests; Namib Sand Sea, Namibia. **(c)** Aeolian ripples on the stoss slope of a dune bedform, Idaho.

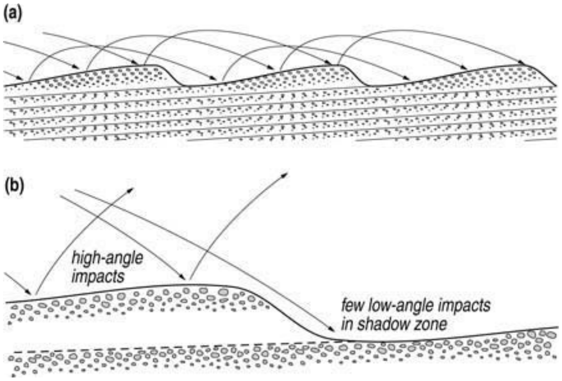


Figure 6.30 Wind ripples generated by ballistic impact of grains. The ripple spacing relates in a general way to the saltation path length, which is the characteristic distance that individual grains hop as a result of collision on the bed. The saltation path length varies with grain size, shape and density, and mean wind velocity and gustiness close to the bed. **(a)** The migration of wind ripples results in subparallel 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 down wind, this sorting mechanism generates inversely graded laminae.

Aeolian megaripples

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 megaripples (granule ripples) may develop. Although these bedforms also develop as a consequence of the impact of saltating and reptating grains, as described for impact ripples, they typically have more sinuous crestlines, and wavelengths and heights up to 5 m and 35 cm, respectively (Fig. 6.31). There is a continuum of sizes from impact ripples developed in fine sand to granule and pebble megaripples.

Adhesion ripples

When dry sand is blown across a wet sediment surface, some grains stick to the surface on impact. This process of adhesion results in the generation of a range of structures, including adhesion ripples and adhesion warts (Fig. 6.32), which are characterized by low relief ridges and mounds that grow by adhesion to their upwind edge and thereby undergo upwind migration. Capillary rise of moisture helps to trap further grains by maintaining a damp surface. Although adhesion structures are found

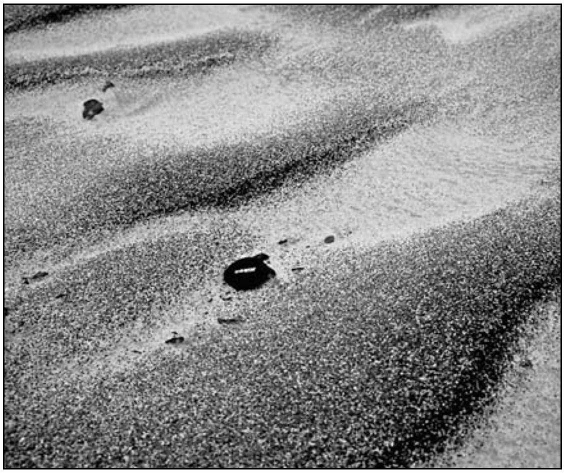


Figure 6.31 Aeolian granule megaripples, Skeleton Coast, northern Namibia.

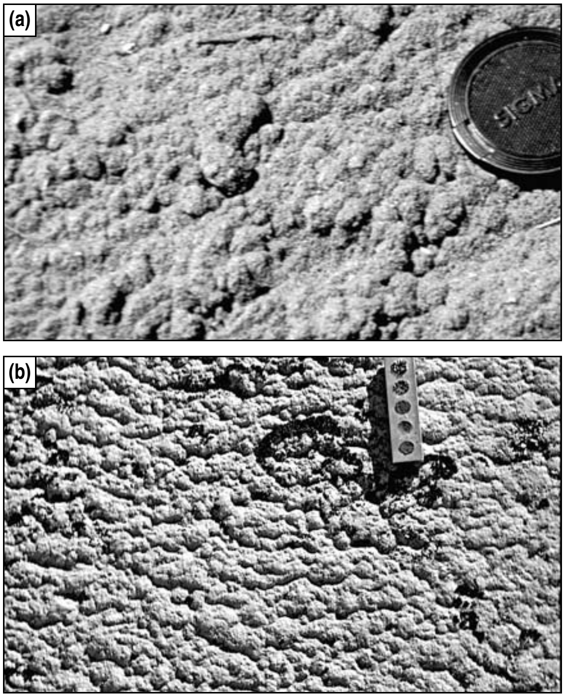


Figure 6.32 Examples of wind adhesion "ripples or warts". **(a)** Adhesion structures generated on a surface of damp sand, over which dry sand has been blown; Indian Creek, Utah. **(b)** Upper bedding surface of sandstone, showing an irregular small-scale morphology interpreted as wind adhesion ripples or warts; Independence Fjord Group, Proterozoic, north Greenland.

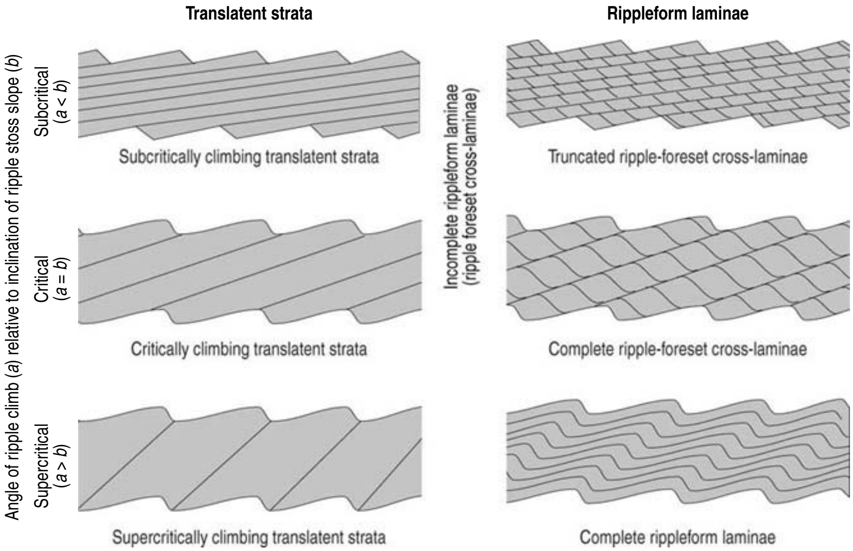

Translatent strata
Rippleform laminae
Angle of ripple climb (a) relative to inclination of ripple stoss slope (b)
Subcritical (a < b)
Critical (a = b)
Supercritical (a > b)
Incomplete rippleform laminae (ripple foreset cross-laminae)
Subcritically climbing translatent strata
Truncated ripple-foreset cross-laminae
Critically climbing translatent strata
Complete ripple-foreset cross-laminae
Supercritically climbing translatent strata
Complete rippleform laminae

Figure 6.33 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).

preserved both on bedding plane surfaces and in section, where strata form crinkly and wavy laminae, they often have a low preservation potential, since, on drying out, they tend to collapse and become reworked by the wind. Because the generation of adhesion strata requires the accumulation surface to be damp, such structures are often restricted to low-lying damp interdune and dune-flank settings.

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. 6.33). However, the uniformity of grain size that typifies many aeolian sands means that internal laminae often cannot be distinguished and **translatent rippleform stratification** results from ripple migration. Wind-ripple strata sometimes exhibit a weak inverse grading, in part because the finest material tends to accumulate in sheltered ripple troughs; 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. 6.34). As in the case of sub-aqueous current ripples (§6.1.4), the accumulation of migrating wind ripples occurs as a consequence of **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 draa slopes (§6.3).

6.1.7 Uses of ripples and cross lamination

Ripple marks and cross lamination have three main uses. Ripple marks and cross lamination are among 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

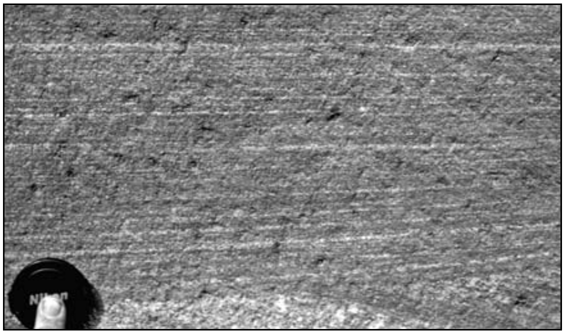


Figure 6.34 Pinstripe lamination is a common characteristic of wind-ripple strata in cross section, because such bedforms usually have a high ripple index (see Fig. 6.2) and often do not preserve internal laminae.

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 sequences with preserved ripple morphologies, a lower bedding surface may sometimes preserve the underlying rippled surface 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 can resemble current ripples. However, internally, they lack any cross lamination. Furthermore, they are commonly associated wth other types of erosive 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. 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 because of either waning or expanding flow. Climbing-ripple cross lamination can help to indicate changes in the rate of sediment supply, especially in cases where 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 surfaces, 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. For symmetrical wave ripples, it may not be possible to judge the sense of wave movement.

6.2 Aqueous dunes, sandwaves, bars and cross bedding

6.2.1 Introduction

Many areas of sandy river beds, tidal flats and channels, and 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. 6.35). It is relatively uncommon to find subaqueous current-generated bedforms in the height range 3–10 cm and the spacing range 30 cm to 1 m. The larger forms commonly have current ripples superimposed upon their stoss sides. When such superimposition is seen on a tidal flat or a river bed, it is quite likely that it results, at least in part, from continued sediment movement during the falling river level or the waning ebb tide, when large forms were no longer active. However,

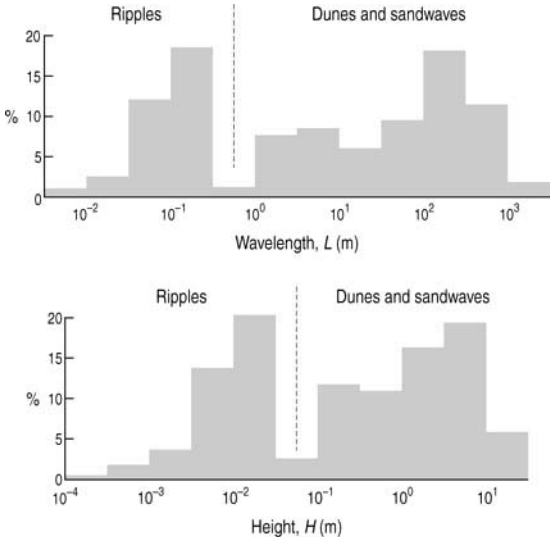


Figure 6.35 Histograms of wavelength and height of ripple-like subaqueous bedforms from various present-day environments. A conspicuous gap separates ripples from larger forms. The population of larger forms probably includes representatives of both dunes and sandwaves. (After Allen 1968)

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 presence is most often reconstructed from the patterns of cross bedding to which they give rise.

6.2.2 Material

Large-scale bedforms and the cross bedding that characterizes their internal structure 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 also found in many coarser conglomerates, where its origin may result from other processes (see Ch. 7). Certain types of large-scale sandstone cross bedding also result from 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 1 m in wavelength, have been described as dunes, sandwaves, megaripples, large-scale ripple marks and various types of bar. Lack of terminological consistency has led to some confusion. 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. 6.36),

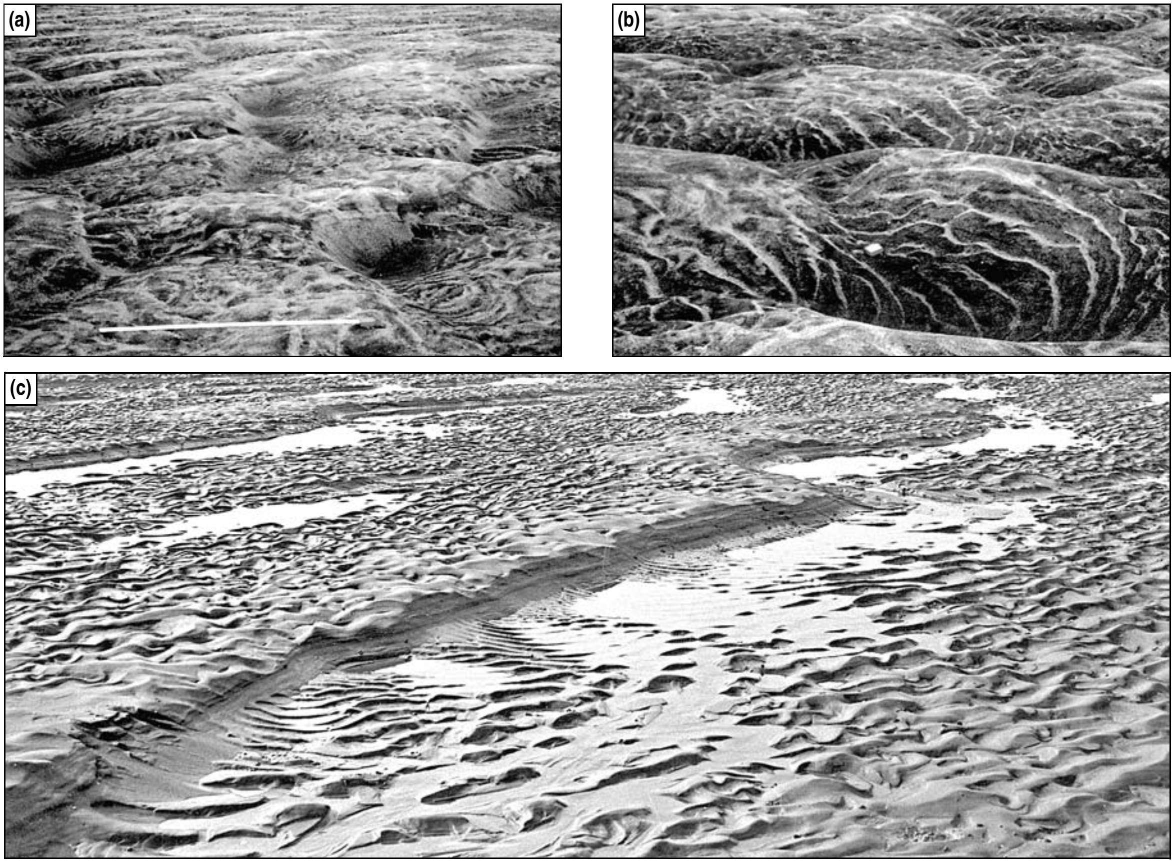


Figure 6.36 Examples of larger sub-aqueous bedforms that are broadly grouped as "dunes". **(a)** Sinuous-crested to linguoid dunes showing superimposed current ripples, well developed scour pits on the lee sides, and lee-side avalanche surfaces; Tana River, Norway. **(b)** Downstream view of the same dunes, showing the radial pattern of ripples around the lee-side scour pits caused by the expansion of the re-attached secondary flow; Tana River, Norway. **(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, the Netherlands.

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 coexist (compound forms). In the case of compound forms, it is important to judge whether the smaller forms are confined to the stoss sides of the large forms or whether 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 relative to 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 rejoining of the flow. In the case of large rivers or estuaries, this last judgement may be rather difficult to make. Climbing a nearby hill, or the study of aerial photographs or of detailed topographical maps, will often be very informative (Fig. 6.37). In the largest systems, features of several different size groups may be superimposed.

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 crestlines and well developed scour pits on their lee sides (Fig. 6.36a,b) or they may be gently curved or straight-crested without scour pits (Fig. 6.36c). Dunes commonly have small-scale current ripples on their stoss sides. These commonly face down stream, towards the dune crest. However, immediately down stream of a dune lee face they may face up stream. In the scour pits of three-dimensional dunes, the ripples commonly fan out from the centre of the pit (Fig. 6.36b).

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 orientated normal to the direction of tidal flow. The asymmetry, which may be very obvious on foreshortened echo traces, is in reality often quite slight, although a whole spectrum exists from near symmetric forms to strongly asymmetric ones. With strong asymmetry, the steep side may be a slipface, but with progressively reduced 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. 6.36).

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, to whether they are simple or compound, and whether or not they have their own discrete slipfaces (Fig. 6.38).

6.2.4 Modification by emergence

Large-scale bedforms, exposed on a river bed at low-water level, commonly show superimposed features that were produced as they emerged. These occur at a variety of scales, and the extent of 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 slipfaces 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. 6.39). 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 level, 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-level slipface (Fig. 6.37). Current ripples may be reorientated to reflect this flow and sand lobes may develop at the confluences of these flow threads. More rapid emergence causes these effects to

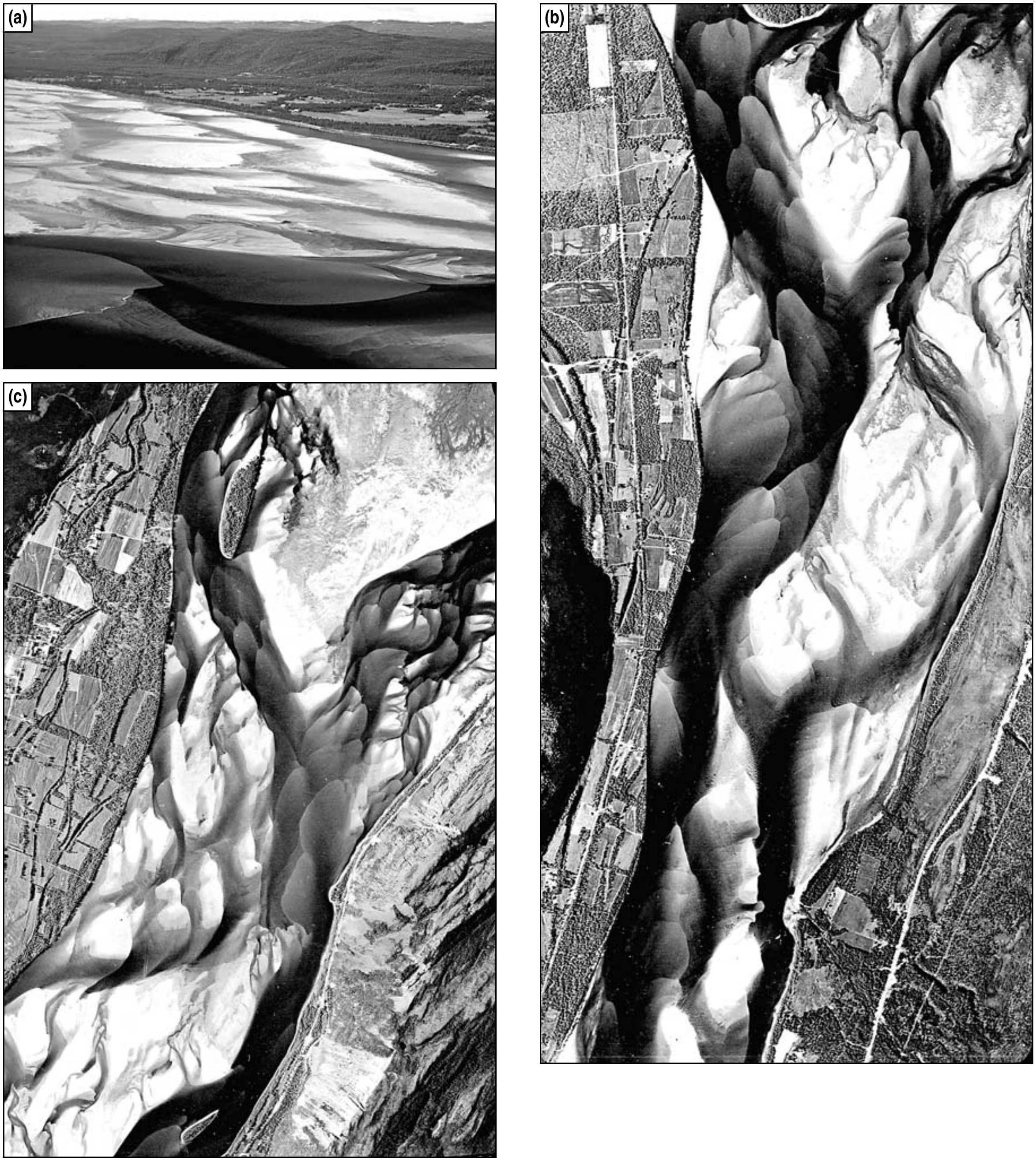


Figure 6.37 Examples of larger-scale dune forms that are sometimes referred to as sandwaves or bars. **(a)** Large-scale low-relief dunes with a broad linguoid shape that have been termed linguoid bars; the shapes of the crestlines have been somewhat modified as the bedforms emerged during falling river level; smaller dunes are present in some of the low areas between the larger forms. **(b)** 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. **(c)** Part of a sandy river bed with repetitive large-scale linguoid dunes, offset en echelon and making up a large bank-attached bar (cf. Fig. 6.38b); note the convoluted crestlines of the emergent forms at the top of the image. All in the Tana River, Norway.

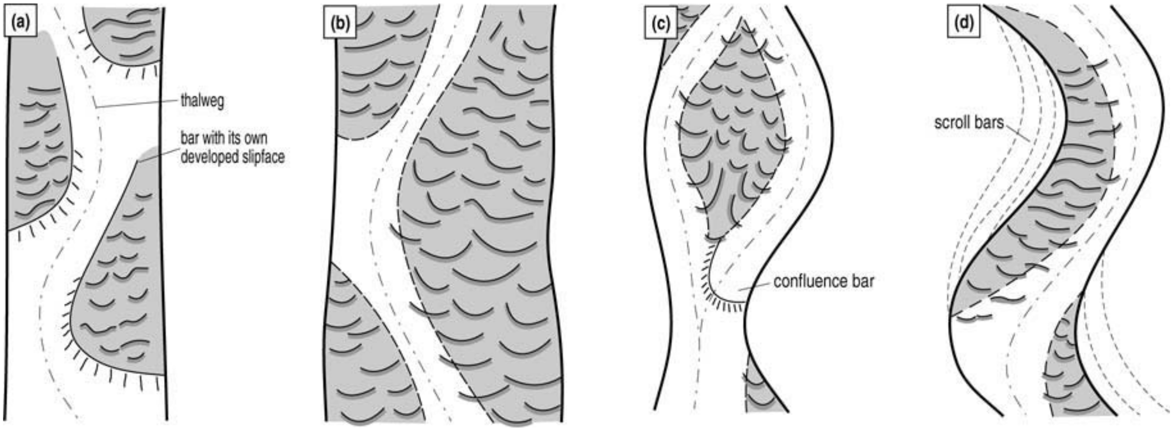


Figure 6.38 Definition diagram for different bar types on a sandy river bed. Note how the dimensions of the bars relate to the channel width. Surfaces of bars have superimposed repetitive bedforms (dunes), whose size is independent of channel width. **(a)** Alternate bank attached (side) bars with their own large-scale slipfaces. **(b)** Side bars without their own slipfaces. **(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.

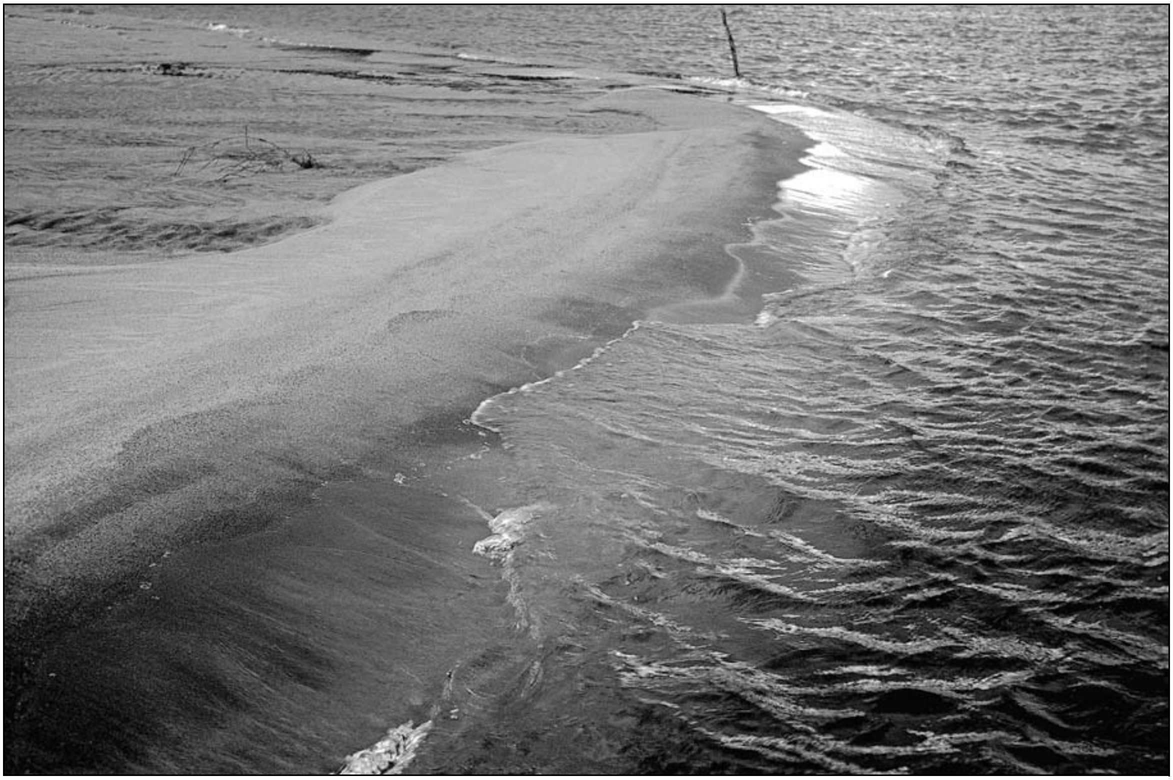


Figure 6.39 Modification by emergence. A series of fan lobes developed as a consequence of waves breaching the crest of a bar and transporting sand onto the upstream side.

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 often preserved in a relatively unmodified state.

6.2.5 Internal structures of dunes; medium-scale cross bedding

Excavation of trenches into dunes reveals patterns of inclined bedding similar to cross lamination, but on a larger scale. This is called **cross bedding**, although the terms "current bedding" and "false bedding" are sometimes found in older literature. Although set thicknesses are usually greater than 10 cm, much of the terminology of cross lamination can still be applied (Fig. 6.9). Straight- or long-crested dunes generate **tabular cross bedding** in sets with roughly parallel top and bottom boundaries and of wide lateral extent; strongly three-dimensional dunes give **trough cross bedding** with scallop-shape sets (Figs 6.40, 6.41, 6.42). Both types are common in the rock record, particularly in medium- and coarse-grain sandstones.

Tabular sets have a wide range of sizes, although sets less than 1 m thick are most common. Sets about 1 m 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. 6.41a). Isolated, very thick, single tabular sets, up to tens of metres thick, are probably not the product of dunes and are described later (§6.2.9). The geometry of tabular sets of cross strata is relatively simple because there is relatively little along-strike variability, the foresets being straight or gently sinuous. However, as is the case for ripples,

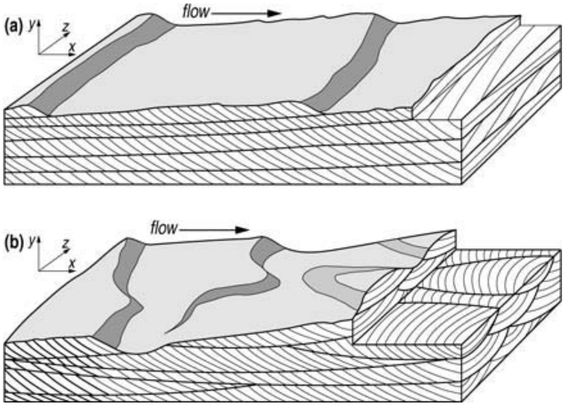


Figure 6.40 Patterns of cross bedding generated by the migration of dunes with different plan-form morphologies. **(a)** Straight-crested (two-dimensional) dunes generate sets of planar cross bedding, which exhibit relatively little variability in sections parallel to the dune crest. **(b)** Sinuous-crested (three-dimensional) dunes generate sets of trough cross bedding in which trough-, scallop-shape or cylindrical scours filled with foresets are evident in sections parallel to the dune crest. These structures occur over a wide range of scales.

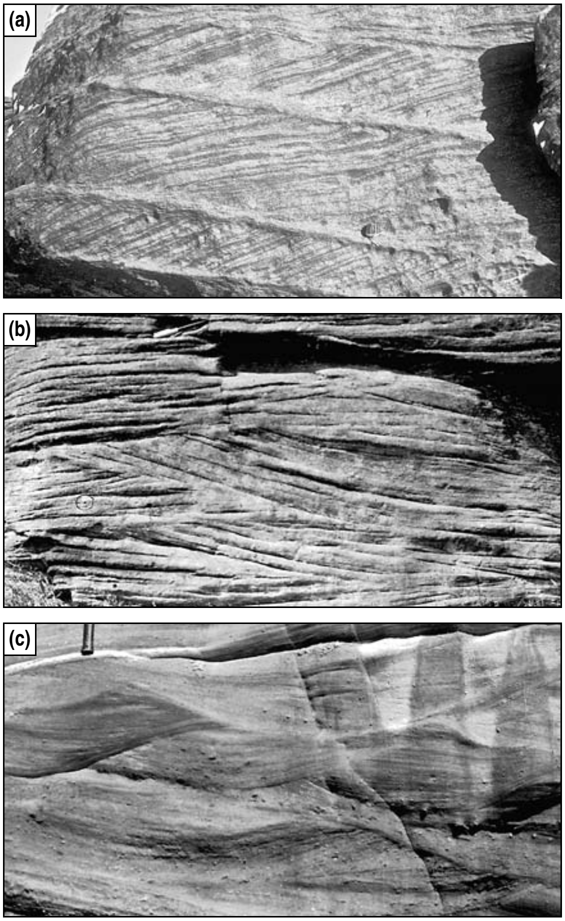


Figure 6.41 Types of medium-scale cross bedding seen in section. **(a)** Planar, tabular sets of cross bedding viewed roughly parallel to palaeoflow; sets are separated by thin units of ripple cross-laminated sandstone; note the angular contact of the foresets at the bases of the sets; light meter (8 cm long) for scale; Roaches Grit, Upper Carboniferous, Staffordshire. **(b)** Trough cross-bedded sandstone viewed normal to palaeoflow; note the apparently opposed foreset directions by dint of their curved shape; pen for scale; Pennant Sandstone, 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)

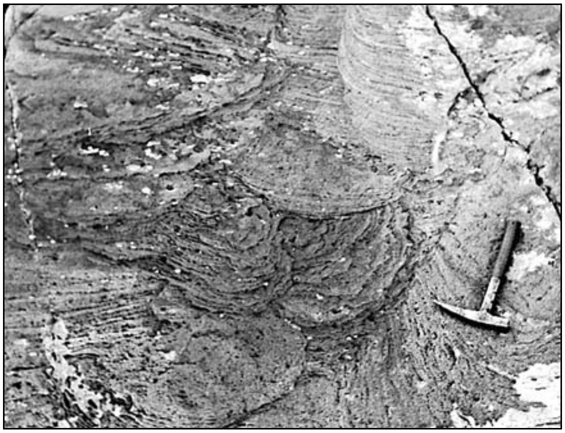


Figure 6.42 Small-scale trough cross bedding seen in plan on an upper bedding surface. Note the intersecting trough forms and the strongly curved foresets, concave down current. The axes of the troughs give the most reliable indicators of palaeoflow direction. (Locality unknown)

bedforms with similar morphologies can preserve different patterns of cross stratification depending on their angle of climb (Fig. 6.43). The foresets of tabular sets are usually either **asymptotic** (i.e. tangentially based) or **planar** (i.e. angular based); convex-up and sigmoidal foreset geometries are relatively uncommon (Fig. 6.44).

Trough sets are seldom more than 1.5 m thick and are typically up to a few metres wide and a few tens of metres long. Most commonly, they are about 30 cm thick, 1–2 m wide and 5–10 m 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. 6.42, cf. Fig. 6.11). Sections cut perpendicular to the current direction are sometimes described as showing **festoon cross bedding** (Fig. 6.41b,c). The geometry of trough-shape 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

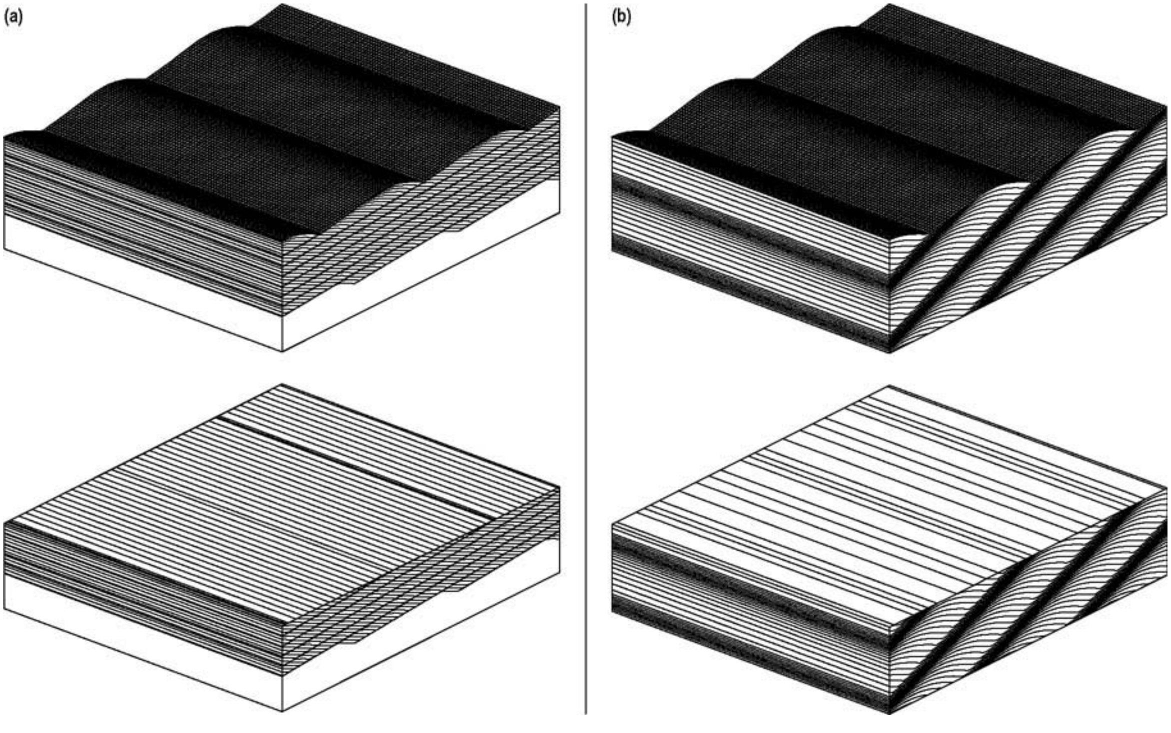


Figure 6.43 Schematic illustrations of the form of cross stratification generated by the migration of two-dimensional (i.e. straight-crested) bedforms. **(a)** Subcritical angle of climb and horizontal section. **(b)** Supercritical angle of climb and horizontal section. (Computer models generated using the Bedforms software of Rubin 1987)

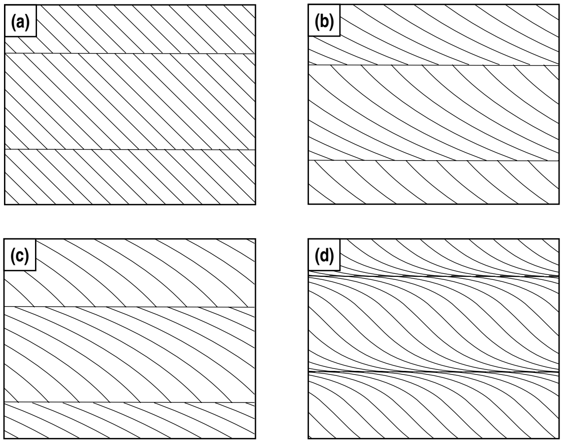


Figure 6.44 Idealized sections in tabular cross bedding, parallel to flow: **(a)** planar foresets with angular bases, **(b)** curved foresets with asymptotic or tangential bases, **(c)** convex up foresets, **(d)** sigmoidal foresets.

patterns of cross stratification where bedforms with similar morphologies climb at different angles (Fig. 6.45). 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. 6.46).

Care should be taken to interpret correctly the geometry of the cross strata that fill trough-shape sets when seen in vertical profile. Bear in mind that outcrop sections orientated transverse to the trend of the trough axis will usually reveal stratification planes that are apparently concordant with the trough base (Fig. 6.47a), whereas sections oblique to the trough axis will show stratification planes that apparently fill the trough in an asymmetric manner and downlap onto the trough base (Fig. 6.47b).

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

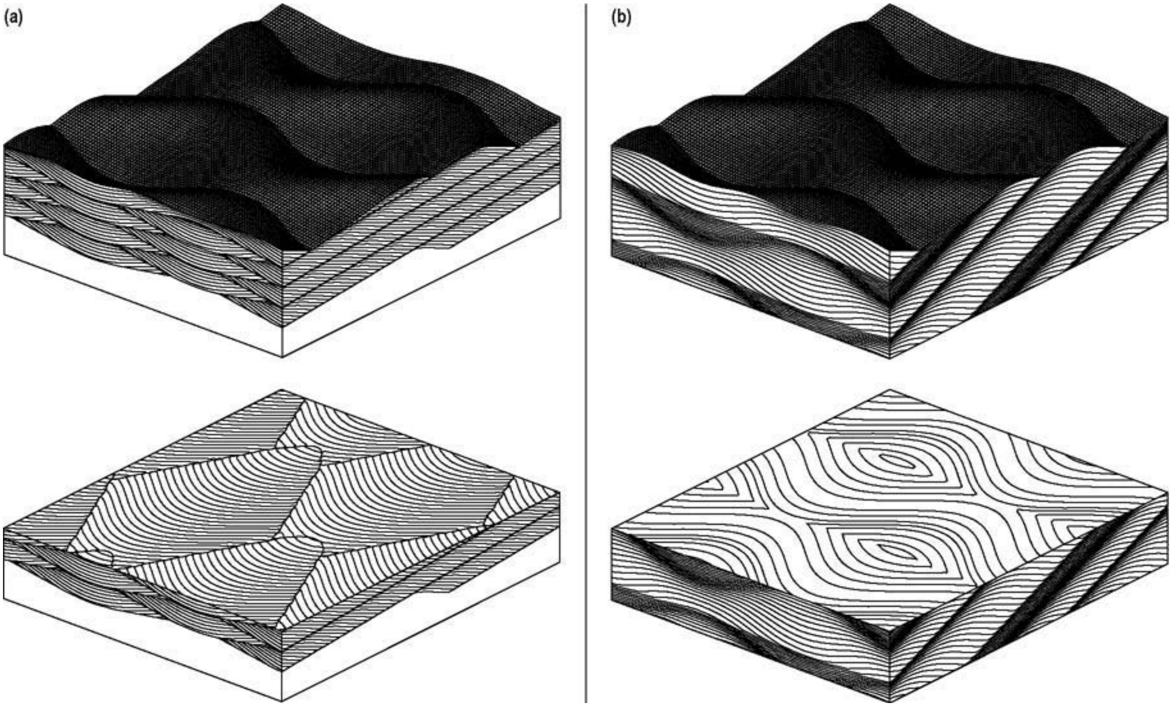


Figure 6.45 Schematic illustrations of the form of cross stratification generated by the migration of three-dimensional (i.e. sinuous-crested) bedforms: **(a)** subcritical angle of climb and horizontal section, **(b)** supercritical angle of climb and horizontal section (computer models generated using the Bedforms software of Rubin 1987).

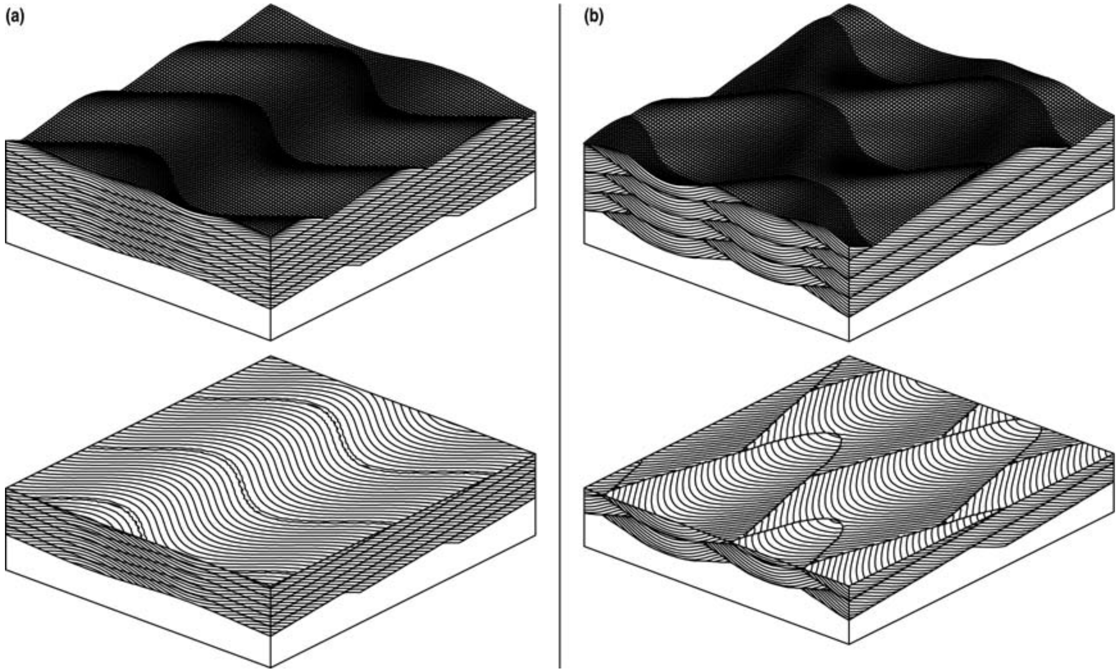


Figure 6.46 Schematic illustrations of the variety of forms of cross stratification generated by the subcritical migration of three dimensional (i.e. sinuous crested) bedforms: **(a)** train of bedforms with successive crestline sinuosities that are in phase; **(b)** bedforms with high amplitude crestline sinuosities that are out of phase. Compare the patterns of cross stratification depicted here with those in Figure 6.45. (Computer models generated using the Bedforms software of Rubin 1987)

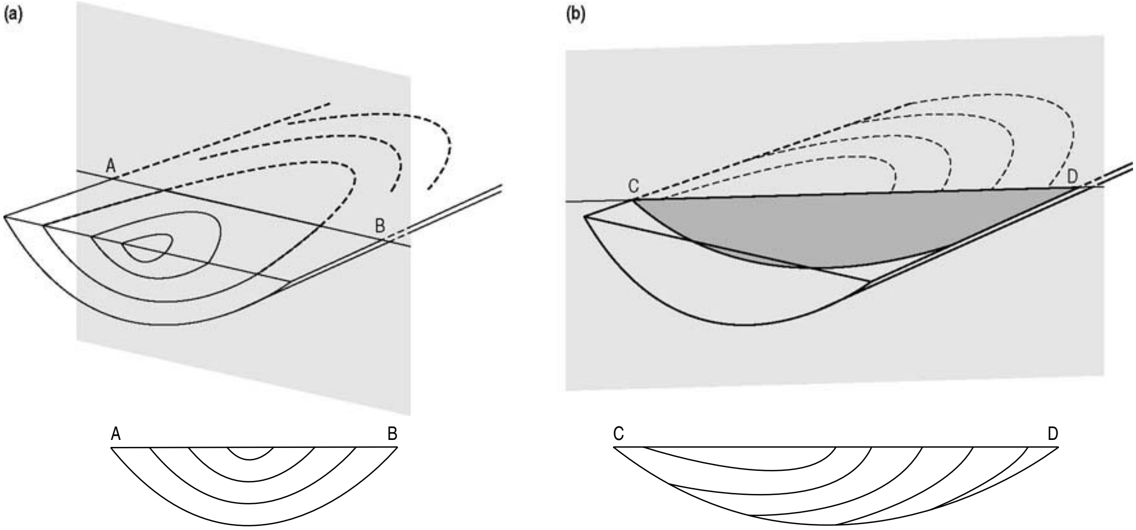


Figure 6.47 Schematic illustration of the geometric complexity of trough-cross strata: **(a)** a vertical section orientated transverse to the trough axis reveals symmetrical cross-stratification planes that are apparently concordant with the trough base; **(b)** a vertical section orientated oblique 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 palaeotransport direction from trough-shape cross strata. (Modified after Rubin & Hunter 1983 and DeCelles et al. 1983)

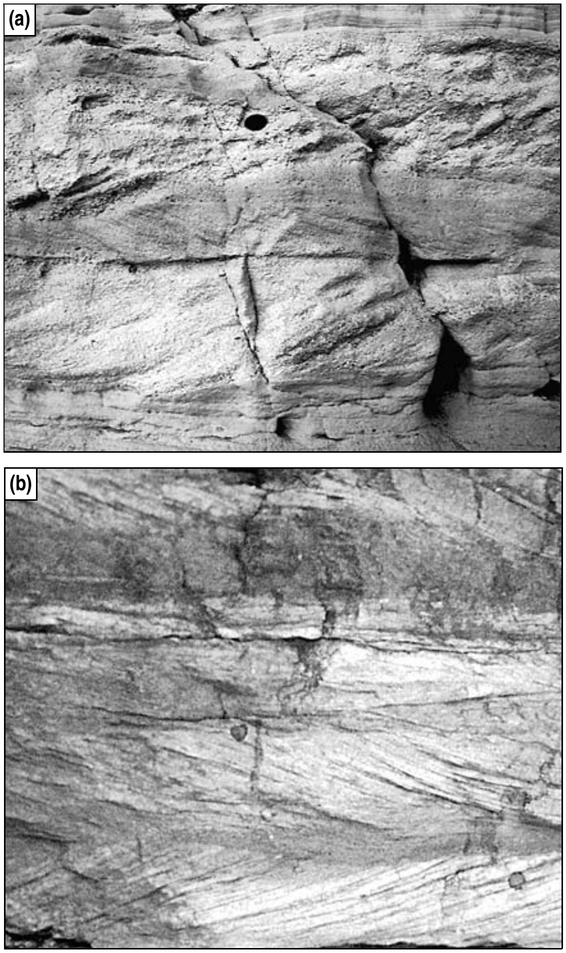


Figure 6.48 Examples of herringbone cross bedding, where successive sets show opposed palaeoflow directions and which is generally attributed to a bi-directional tidal flow regime: **(a)** in gravelly sandstone, Sorbas Member, Miocene, Sorbas Basin, Spain; **(b)** Late Precambrian, Bela Dam, Sagar, India (width of view: 0.6 m; photo courtesy of Gilbert Kelling).

the orientation of the depositional horizontal. However, in some extensive exposures it is apparent that bounding surfaces between sets are themselves 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. Many different relationships occur, of which ascending (upstream accretion), descending (downstream accretion) and along slope (lateral accretion) are end members.

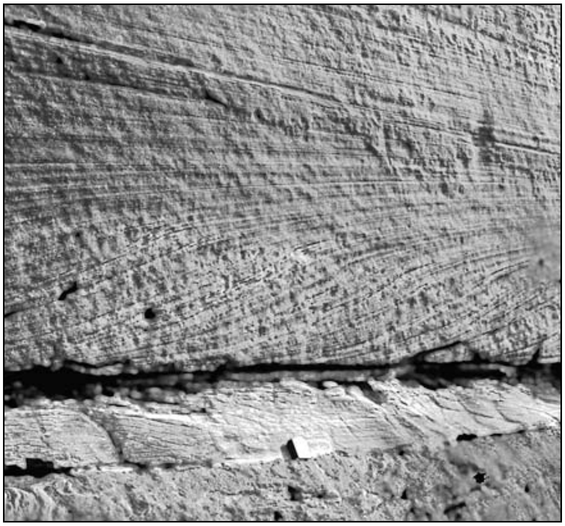


Figure 6.49 Sigmoidal cross bedding within a single tabular set, developed in medium-grain sandstone. Note that the foresets have tangential bases and the flatter-lying topset laminae are also inclined down stream. St Bees Sandstone Formation, Triassic, Cumbria, England.

In some cosets of tabular cross bedding, the dips of foresets in adjacent sets are in opposite directions. Some examples show alternation of direction from set to set, whereas in others only a small proportion of sets show an opposed direction (Fig. 6.48). Such **herringbone cross bedding** is important in the interpretation of processes, but care must be taken to distinguish it from festoon cross bedding generated by trough cross-bedded sets seen in sections perpendicular to transport, where foresets may *appear* to dip in opposite directions (Fig. 6.41b,c).

Some approximately tabular sets are unusual in showing sigmoidal foresets (Figs 6.44d, 6.49). In these cases, the convex-upwards foreset laminae at the top of the set may pass up-dip into parallel lamination, which occurs as a “topset” unit (Fig. 6.49). Traced down dip, such sets sometimes show a gradual reduction in thickness and in foreset inclination.

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 toeset region. Such cross lamination is

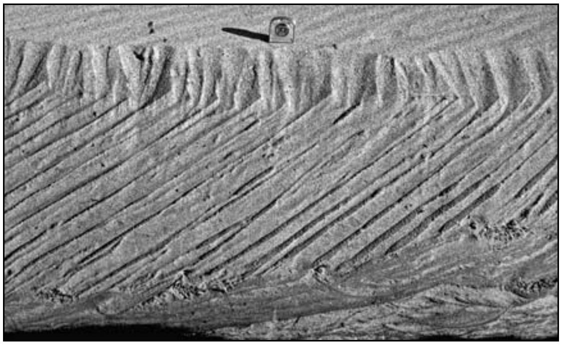


Figure 6.50 Single tabular cross-bedded set with a unit of countercurrent ripple lamination at the base. Note that the ripple lamination has a climbing geometry and that ripple forms are buried by the foresets. Trench in large low-relief aqueous dune. The small ripples on the top surface are wind-impact ripples. Scale given by tape case, 5 cm wide. Present day, Tana River, Norway.

commonly directed up the slope of the larger-scale foresets and is termed **countercurrent cross lamination** (Fig. 6.50). Discontinuities within foresets are sometimes seen in sections parallel to foreset dip. These erosion (**reactivation**) surfaces are less steeply inclined than the foresets on either side; they may occur in isolation or as multiple features within a set (Fig. 6.51). With trough cross bedding or with downstream-inclined bounding surfaces, it may be difficult to distinguish bounding surfaces of sets from reactivation surfaces.

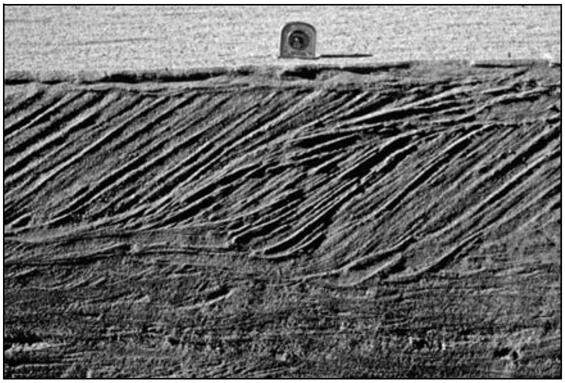


Figure 6.51 Upper part of trench shows a tabular set of cross bedding, within which the foresets are truncated by a lower-angle erosion surface, a reactivation surface. Above and to the left of the discontinuity, foresets resume their normal downstream dip. The discontinuity results from falling river level between the successive flood events that caused the dune bedform to migrate. Present day, Tana River, Norway.

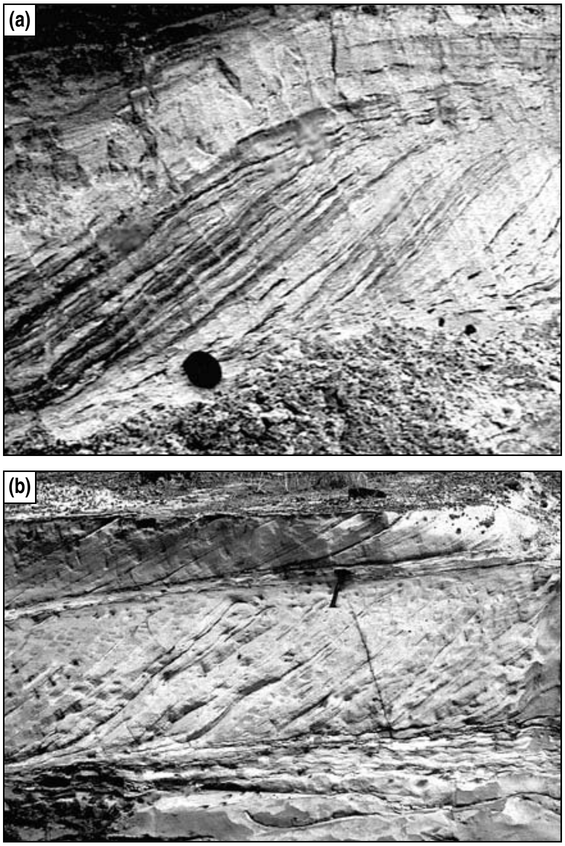


Figure 6.52 Sets of cross bedding with drapes of mud on the sandy foresets. **(a)** The proportions of sand and mud vary along the set, possibly reflecting spring/neap cyclicity in a tidal regime; Lower Cretaceous, Kong Karls Land, Svalbard. **(b)** Sets of medium-scale cross bedding with mud drapes on some foresets; the mud drapes thicken towards the toes of the foresets in some cases; Cretaceous, Nigeria (photo courtesy of Gilbert Kelling).

In some cross bedding, mainly from tidal environments, **clay drapes** may occur on the foresets (Fig. 6.52). In some examples these 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 sandbed to currents more powerful than those that generate ripples. Differences in dune morphology result from

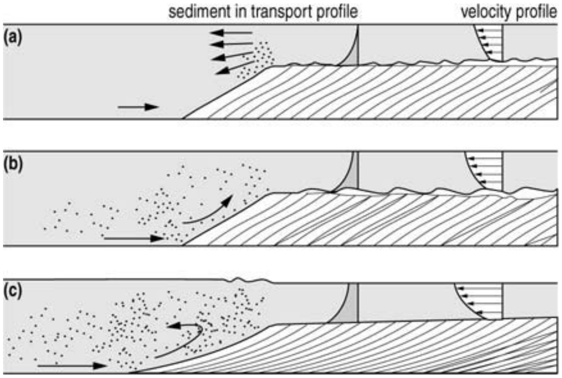
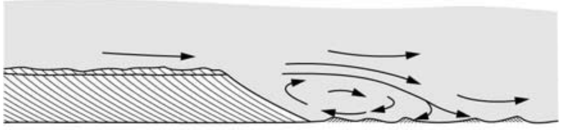


Figure 6.53 The changes in shape of the slipface of a small laboratory delta resulting from progressive increase in the velocity of flow over it from **(a)** to **(c)** (after Jopling 1965).

(a) High-water stage



(b) Intermediate-water stage

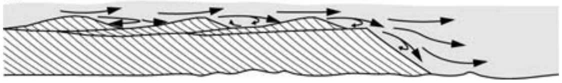
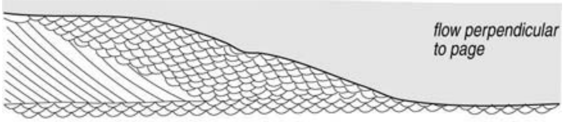


Figure 6.54 Changes in morphology and internal structure due to changing water level over a sandwave. Note the countercurrent ripples developed in front of the bedform during high water level and compare the morphology with that in Figure 6.50 (modified after Collinson 1970).

differences in flow strength and depth within the dune-field, 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 re-attaching flow in scour pits characterizing the three-dimensional types. In some low-relief dunes (long spacing and little height) the pattern of flow separation and re-attachment is confined to the immediate lee-side area. In these cases, the bedform heights are much closer to the flow depth and there is often a quite shallow flow over the crestline.

The separated flow in the lee of a dune gives rise to a backflow component that helps produce a tangential lee face. The strongly focused eddying in front of concave sectors of a lee face leads to the development of scour pits and to the associated ripple fans (see Fig. 6.36b). With straight-crested dunes, the strength of flow over the crest can strongly influence lee-side profile and hence foreset shape. With weak flows, grainflow (avalanching) dominates on the slipface, giving angular foresets. With stronger flows, flow separation and grainfall become more important, leading to tangential foresets, sometimes with countercurrent ripples (Figs 6.50, 6.53, 6.54). Reactivation surfaces (Figs 6.51, 6.55, 6.56) in river bedforms result from reworking by waves during emergence (see Fig. 6.39) and by flows around

(a) Low-water stage



(b) Risen-water stage

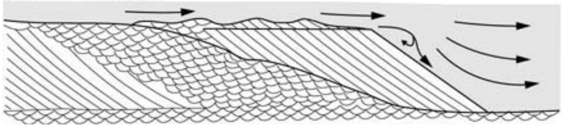
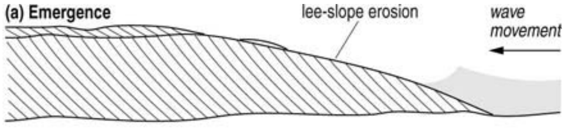


Figure 6.55 Changes in morphology and internal structure resulting from changing water level over a sandwave. **(a)** During low-water level when flow is sluggish, weak across-channel currents may be aligned parallel with the front of a bedform (e.g. a linguoid sandwave) and can therefore deposit sand 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)

(a) Emergence



(b) Risen-water stage

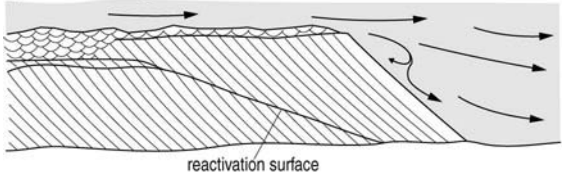


Figure 6.56 Changes in morphology and internal structure with changing water level over a sandwave. **(a)** Emergence of bedform top and reworking by wave action to reduce angle of lee slope; truncation of steeper foresets. **(b)** Risen water level and recommencement of downstream bedform migration; steeper foresets build over reactivation surface. (Modified after Collinson 1970)

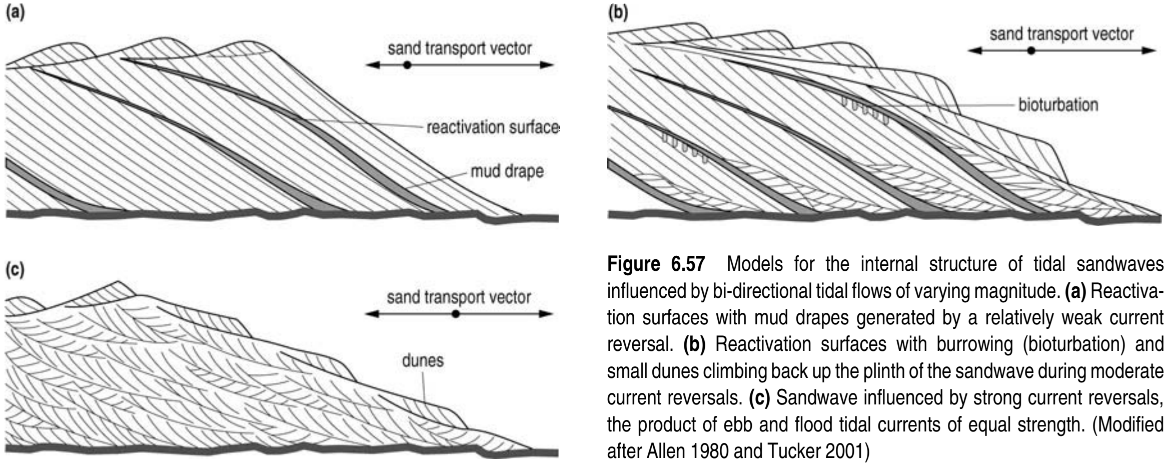


Figure 6.57 Models for the internal structure of tidal sandwaves influenced by bi-directional tidal flows of varying magnitude. **(a)** Reactivation surfaces with mud drapes generated by a relatively weak current reversal. **(b)** Reactivation surfaces with burrowing (bioturbation) and small dunes climbing back up the plinth of the sandwave during moderate current reversals. **(c)** Sandwave influenced by strong current reversals, the product of ebb and flood tidal currents of equal strength. (Modified after Allen 1980 and Tucker 2001)

the margins of the emergent bedforms. The subordinate tide in both intertidal and subtidal settings is also able to produce similar effects.

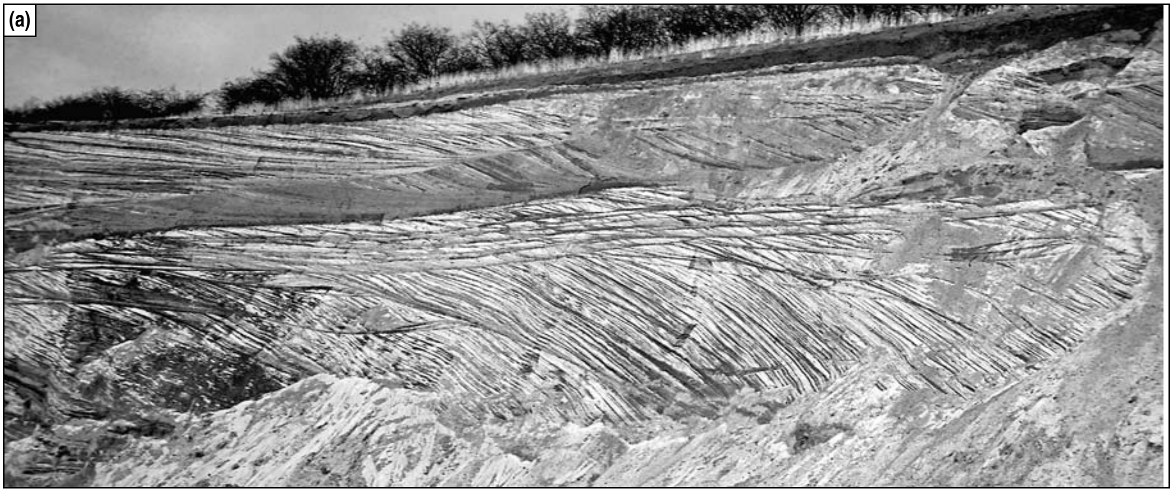
Tidal sandwaves have morphologies that reflect the imbalance between the two opposing tidal flows (Fig. 6.57). Very slight imbalance is not capable of generating a marked asymmetry, but in such cases migration rates are slow. With greater imbalance, migration rates will be higher and bedform asymmetry is likely to be more marked. The imbalance is also likely to be reflected in the internal cross bedding. Highly asymmetric bedforms may well show relatively simple 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 6.57a, 6.58). More symmetrical bedforms will have more complex cross bedding, showing a greater occurrence of reactivation and of sets of reversed cross bedding, leading to herringbone cross bedding (Figs 6.57c, 6.59).

Herringbone cross bedding occurs predominantly in the shallow subtidal zone because of periodic reversals in the current direction as a consequence of tidal cycles. Cross-stratified sets orientated towards the coast during the flood tide and away from the coast during the ebb tide (Fig. 6.59). 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 tide and low tide (or both).

Mud drapes on cross-bed foresets result from fallout of suspended load at tidal slack water, often accelerated by flocculation. When occurring in pairs in the shallow subtidal zone, they record the slacks on either side of the subordinate tidal flow. The thicker sand layer 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 6.52, 6.60). The number of foreset increments per bundle broadly coincides with the tides of a lunar month and thus provides an internal indicator of migration rate for the bedform responsible.

Sigmoidal foresets with parallel topsets (Fig. 6.49) record the condition of high rates of vertical bed accretion that was synchronous with bedform migration. The parallel lamination indicates upper flow-regime conditions on the top of the bedform (see §6.4). The whole assemblage may record a transition between the dunes and upper-level plane bed (see Fig. 6.22).

A cross-bed set normally preserves only 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. However, for most examples 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. Recent 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 mainly controlled by the



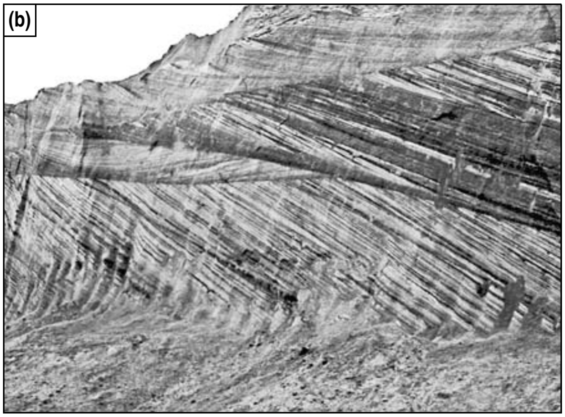


Figure 6.58 Cross bedding produced by tidal sandwaves. **(a)** Large-scale cross bedding with multiple reactivation surfaces in large set and descending smaller sets at the right hand side. **(b)** Broad trough-like sets with sweeping foresets. Internally, sets are broken by reactivation surfaces and there is apparent reversal between sets; foresets are burrowed; face about 6 m high. Both examples from the Woburn Sands, Lower Cretaceous, Bedfordshire.

dynamics of migrating scour pits. The same analysis suggests that the average preserved length of trough cross-bed sets approximates to half the spacing 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 could all be sites of lateral accretion (see Fig. 6.38). Ascending sets (upstream accretion) are most likely to occur on the upstream sides of medial bars or side bars; descending cross beds are likely at their downstream flanks. In addition, descending cross beds could reflect changing flow level (see below and Fig. 6.56). A particular style of lateral accretion structure is dealt with more thoroughly in §6.2.10.

Bars, which form at the scale of the channel, probably result from the action of large-scale patterns of water circulation in the channel flow. For example, flow of water around a channel bend leads to a large-scale spiral vortex, with near-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 lead to deposition. Large alternate bars along the sides of straight channels are not well documented or understood, but probably involve large-scale flow separation at their downstream ends, possibly associated with the 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 mainly independent of grain size. Deep flows 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, although many dunes are in

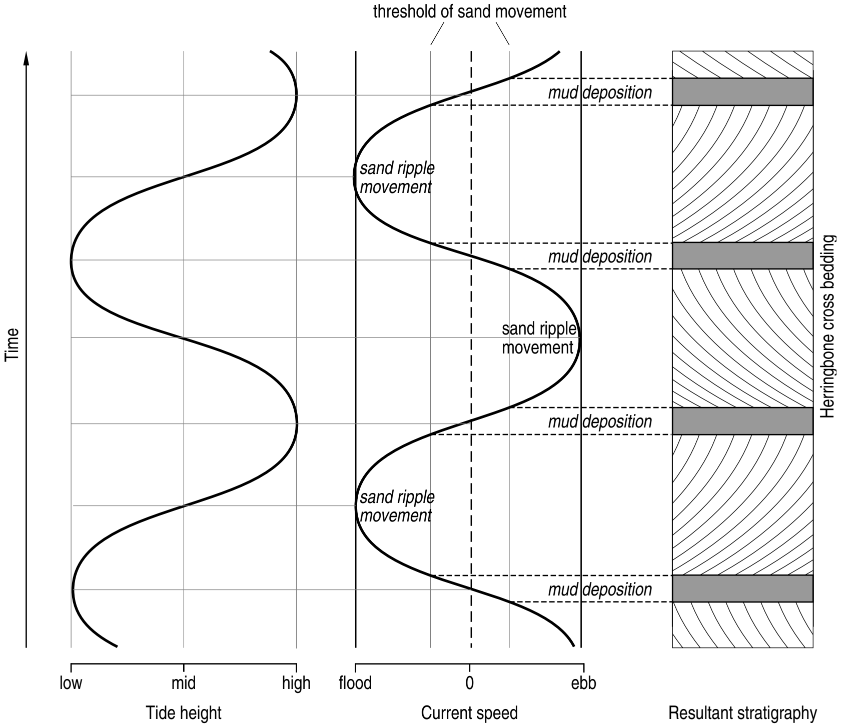


Figure 6.59 Variation in tidal current speed and direction over a tidal cycle and the resultant generation of tidal rhythmites for a position in the shallow subtidal realm. Deposition of mud occurs when the current speed is low (at both high and low tide as the tide turns). Deposition of cross-stratified 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 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)

fact 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 tidal sandwaves are not well established. In attempting to apply any such relationships to the cross bedding in the rock record, it is important to understand that there may be complications in translating cross-bed set thickness to bedform height, as discussed in §6.2.7.

One consequence of these general relationships is that dunes of different sizes may become superimposed as a result of changing flow conditions (Fig. 6.61). When conditions change rapidly, large bedforms cannot adapt quickly enough to maintain equilibrium. The growth of smaller dunes on larger ones during falling level 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. 6.62).

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). 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. Such sets can be very extensive laterally: some sets 20–30 m thick may be traced for several kilometres parallel to the dip direction. Such sets are not explained by the migration of repetitive bedforms and at least two alternatives must be considered.

One is by the advance of a delta formed by a stream that carried abundant bedload into quieter water. Such deltas commonly have steep slopes, and avalanching on such a slipface causes the delta to advance and create a single cross-bedded set. These deltas can easily be modelled in a laboratory tank (cf. Fig. 6.53) and were first recognized around lake margins by G. K. Gilbert in 1885. Examples of Gilbert-type deltas are quite rare in the rock record, except in Pleistocene and Holocene gravels. However, their initial description caused confusion, as some geologists came to regard all cross bedding

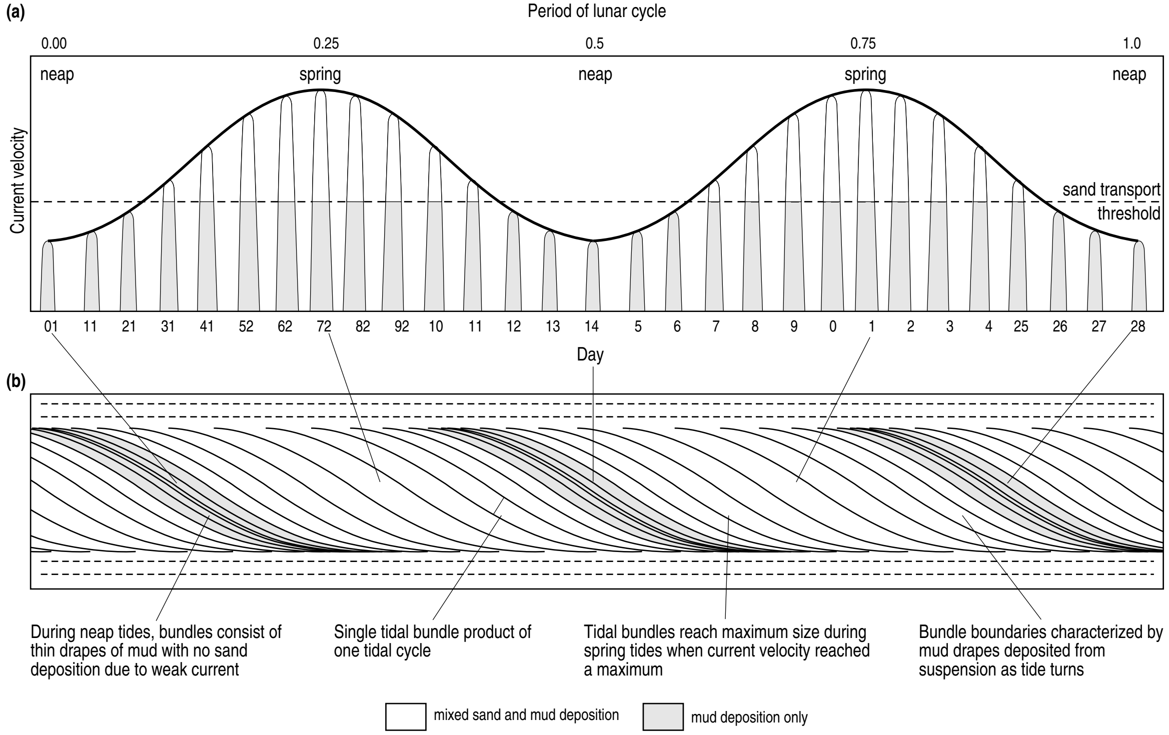

(a)
Period of lunar cycle
0.00
0.25
0.5
0.75
1.0
neap
spring
neap
spring
neap
Current velocity
sand transport threshold
01 11 21 31 41 52 62 72 82 92 10 11 12 13 14 5 6 7 8 9 0 1 2 3 4 25 26 27 28
Day
(b)
During neap tides, bundles consist of thin drapes of mud with no sand deposition due to weak current
Single tidal bundle product of one tidal cycle
Tidal bundles reach maximum size during spring tides when current velocity reached a maximum
Bundle boundaries characterized by mud drapes deposited from suspension as tide turns
mixed sand and mud deposition
mud deposition only

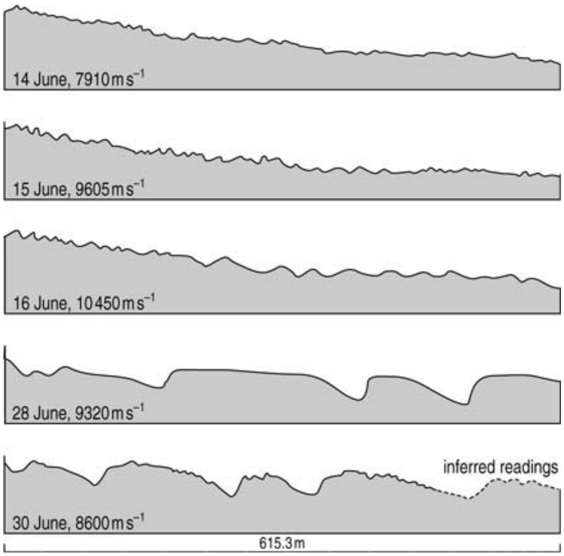
Figure 6.60 Schematic illustration of the mechanism by which sequences of tidal bundles are generated and vary in response to spring/neap tidal cycles. **(a)** Changes in tidal current velocity over the period of a lunar month for a diurnal tidal system (i.e. one high and one low tide per day). Note 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.

as diagnostic of deltaic sedimentation, a view common in books from the first half of the twentieth century. Although deltas dominated by coarse-grain 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 gradients.

The other possible origin of large-scale isolated sets is by the advance of large bars, either attached to alternating sides of very large deep channels (Fig. 6.63) or in mid-channel or channel-confluence settings. Bars of this sort occur on a small scale in modern streams, but a 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 1 m and 5 m thick, in which inclined beds dip at gradients considerably less than the angle of rest, commonly 5° to 15°; the thicker the unit, the lower the inclination. The inclined beds, which extend over the full thickness of the unit, may be sigmoidal in vertical section and may be defined by conspicuous differences in grain size between beds (Fig. 6.64). In detail, the inclined beds usually contain smaller-scale internal structures, such as cross lamination and small-scale cross bedding, which indicate flow subparallel to the strike of the inclined beds. The near-horizontal basal surface of the set is erosional, often with a concentration of pebbles or with



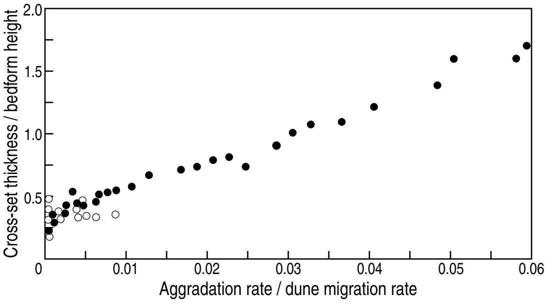


Figure 6.62 Relationship between the ratio of aggradation rate to mean dune migration rate, and the ratio of mean cross-set thickness to mean bedform height for sub-aqueous dunes. Open and filled circles represent results from experimental and computer simulation runs, respectively. (After Leclair 2002)

Figure 6.61 Echo-sounding profiles made at different discharges of the flood on the Fraser River, British Columbia. 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 because of the large volumes of sediment that must be reworked to modify large bedforms. Vertical exaggeration ten times. (After Pretious & Blench 1951)

intraformational clasts. There may be an overall upwards 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 extensive upper bedding planes are exposed, the inclined beds are seen to be strongly curved through several tens of degrees in plan view (Fig. 6.65). Sets often end laterally, at their down-dip ends, with erosion surfaces that dip steeply towards the inclined beds but which are separated from them by a unit of siltstone or of disturbed bedding.

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 (see §6.2.7). The critical point in its interpretation is that the smaller-scale structures indicate flow subparallel to the strike of the inclined surfaces. From this, one can infer that an inclined depositional surface migrated, transverse to the flow direction (Fig. 6.66). The basal erosion surface and the

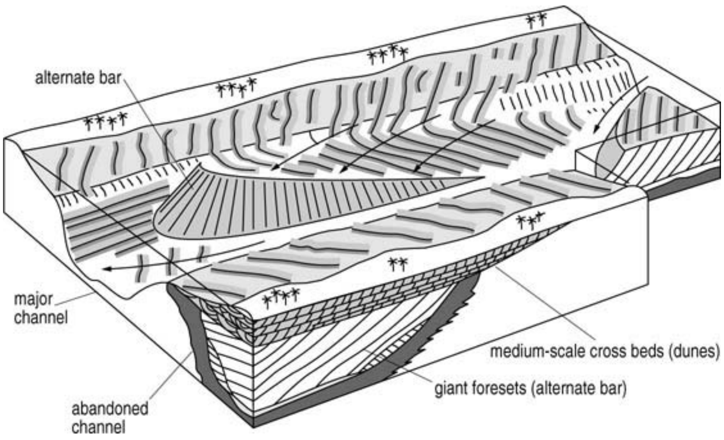

alternate bar
major channel
abandoned channel
medium-scale cross beds (dunes)
giant foresets (alternate bar)

Figure 6.63 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 Namurian rocks of northern England. (Modified after McCabe 1977)

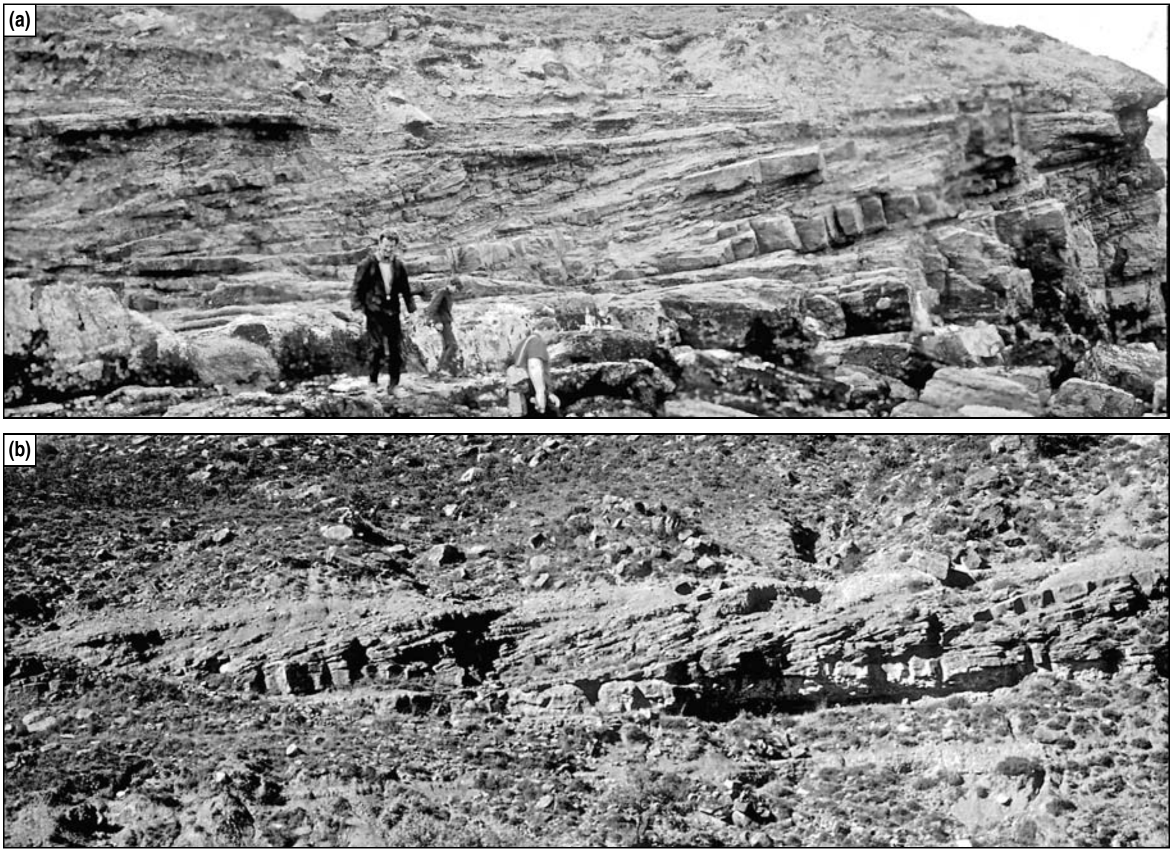


Figure 6.64 Examples of epsilon cross bedding. **(a)** A unit of epsilon cross bedding. The base of the set is a roughly horizontal erosion surface. The surfaces dipping to the left represent successive positions of the depositional bank of a channel as it migrated laterally. Smaller structures within the inclined sand units indicate flow parallel to the strike of the inclined units. Cloughton Formation, Middle Jurassic, Yorkshire. **(b)** Epsilon cross bedding, Montañana Group, Eocene, Spanish Pyrenees.

curved plan view of the dipping beds combine with this inference to suggest a curved channel side, probably in a meandering stream. 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 tidal creeks. More extended discussions of the processes involved are given in books dealing with sedimentary environments (see reading lists of Chs 1 and 10).

6.2.11 Uses of sandwaves, dunes and cross bedding

There are three main uses 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

When dunes are preserved on upper bedding surfaces or as form sets, they give a fairly positive 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 of sigmoidal sets 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 sandbeds and on the cross bedding preserved in sandstones. The shape of the

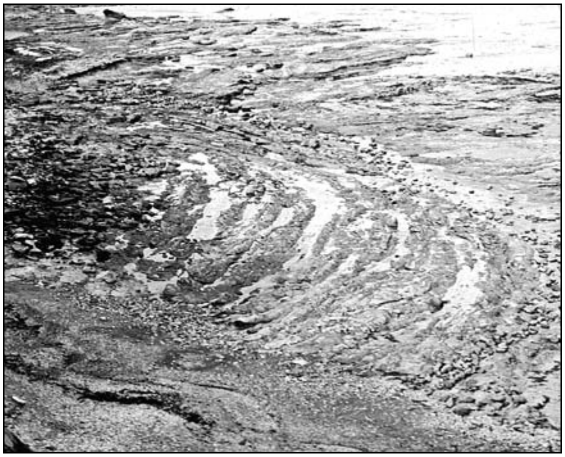


Figure 6.65 Upper bedding surface of a sandstone unit made up of several laterally adjacent sets of epsilon cross bedding. The beds within each set are inclined in the direction of convex curvature, suggesting that the epsilon cross bedding is explained by lateral accretion on the point bar of a meandering channel. Scalby Formation, Middle Jurassic, Yorkshire.

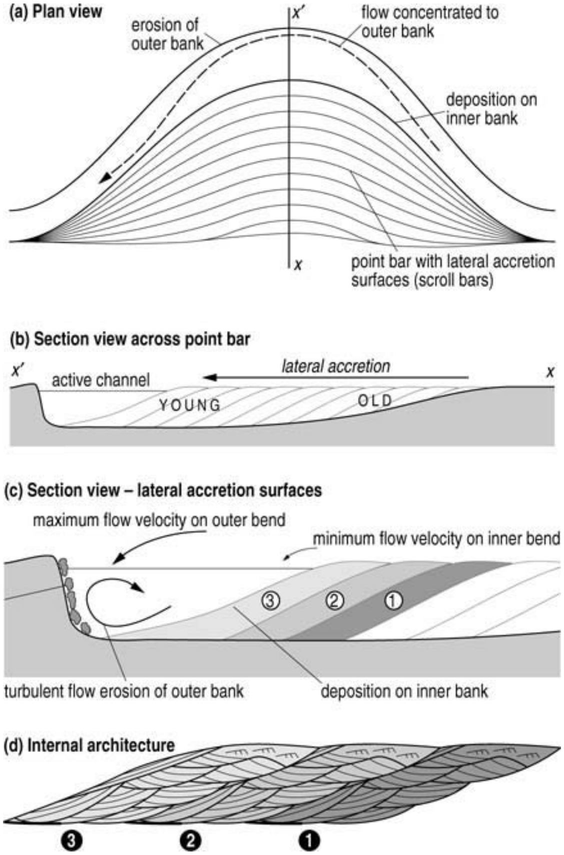

(a) Plan view
erosion of outer bank
x′
flow concentrated to outer bank
deposition on inner bank
x
point bar with lateral accretion surfaces (scroll bars)
(b) Section view across point bar
x′
active channel
lateral accretion
x
YOUNG
OLD
(c) Section view – lateral accretion surfaces
maximum flow velocity on outer bend
minimum flow velocity on inner bend
③
②
①
turbulent flow erosion of outer bank
deposition on inner bank
(d) Internal architecture
❸
❷
❶

Figure 6.66 Schematic illustration of the development of lateral accretion surfaces through the migration of a river channel transverse to flow. **(a)** Plan view depicting a point bar with lateral accretion surfaces (scroll bars) revealing the former positions of the channel. **(b, 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 forced 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 in mid-channel bars.

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 to flow depth, albeit roughly, 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. Great care is therefore needed in making reconstructions of palaeohydrodynamics. However, a systematic upwards reduction in set thickness through a coset may 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-level 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 to record the magnitude of dip, particularly if the succession is tectonically tilted. To measure cross bedding reliably in vertical sections, it is necessary to see faces with more than one orientation (e.g. Fig. 6.40). The apparent dip on a single face shows only a component of the true dip. A bedding-surface view of

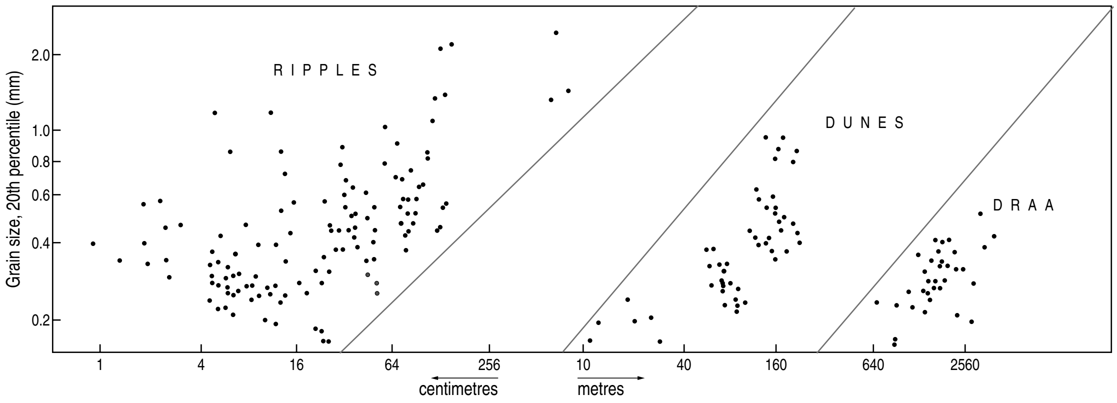


Figure 6.67 Grain size versus wavelength for aeolian bedforms; note the three distinct groups representing ripples, dunes and draa. There is a distinct gap in bedform size between the largest ripples and the smallest dunes. (Modified after Wilson 1972)

the foresets will always give the most accurate measurement of foreset azimuth. It is also important to bear in mind that the slipfaces 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 the set boundaries and the foresets. For really reliable directions, it is best to measure the directions of trough axes on bedding planes (e.g. Fig. 6.42). 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.

When cross bedding occurs between bounding surfaces that 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 Aeolian dunes and cross bedding

6.3.1 Introduction

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 called **draa**, which are themselves migratory. Draa are mega-bedforms that have no aqueous counterparts. Small aeolian ripples and horizontal beds are superimposed on both dunes and draa. The suggestion that ripples, dunes and draa form a hierarchy of equilibrium bedforms provides a basis for classification and description (Fig. 6.67). However, because of the low density and viscosity of air, there is a strong chance of aeolian processes frequently passing from equilibrium to gross disequilibrium, in terms of both the energy required to form structures and the direction of flow. No 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 rare, and the former existence of dunes is deduced mainly from internal structures. Be sceptical of sources that give simple criteria for the interpretation of dune types from ancient strata; records of aeolian processes, structures and environments are among the most difficult to identify and explain in detail.

6.3.2 Material

Aeolian dunes occur only in sand, rarely extending into granule-size gravels. The sand often 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 (§3.6.2) is the dominant transport mechanism associated with aeolian dunes, and interparticle collisions result in high rates of grain abrasion such that dune sediments are often 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 often 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 silt–clay aggregates are commonly seen, especially as parabolic dunes, forming immediately down wind of dried-out lake beds. On wetting, such dunes become solid masses but retain their cross-bedded structure.

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

The classification of 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, the measurement of wind regimes, and conceptual numerical models. As a consequence, aeolian bedforms are classified according to separate criteria, including their scale, morphology (shape, width, wavelength, height), spacing (frequency), orientation relative to net sand-transport direction, style of migratory behaviour and style of superimpositioning. Three distinct scales of aeolian bedform are recognized: ripples (see §6.1.6), dunes and draa (also known as megadunes) (Fig. 6.67). 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 draa 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 draa may take as long as 10 000 years to develop and equilibrate, whereas ripples may respond almost instantaneously to changes in wind direction and strength.

Aeolian dunes

Aeolian dunes have wavelengths of 5–250 m and are often arranged into **trains** of regularly spaced bedforms. In terms of morphology, simple dunes are characterized by a single windward **stoss slope** inclined at 8–16° and a **lee slope** inclined at 20–34°; more complex forms may have several stoss or lee slopes facing in various orientations. Lee slopes often comprise a **slipface** (i.e. a foreset slope), inclined at or close to the angle of rest, and down which sand periodically avalanches. However, not all dunes possess such slipfaces 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 controlling factors, including wind strength and directional variability on diurnal to seasonal (and longer) timescales, sediment supply and the availability of sediment for transport. Additionally, dunes may be classified as mobile (actively migrating), active but anchored (e.g. 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 and according to the orientation of their crestlines relative to the predominant wind direction (Fig. 6.68). Common dune types classified according to these criteria include **transverse** dunes, which have a single, gently inclined, upwind-facing stoss slope, a steeper downwind lee slope, and a crestline perpendicular to the prevailing wind. Longitudinally orientated **linear** dunes (sand ridges) have one or two lee faces and a crestline parallel to the wind, whereas **star** (**pyramid**) dunes have three or more lee faces (Figs 6.68, 6.69). 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. 6.70).

At a more detailed level, dunes with straight crestlines are **two dimensional**, whereas those with sinuous, cuspate or lobate crestlines are **three dimensional** (Figs 6.68, 6.71). Examples of 3-D transverse bedforms include spatially isolated crescent-shape **barchan** dunes, which open down wind and 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

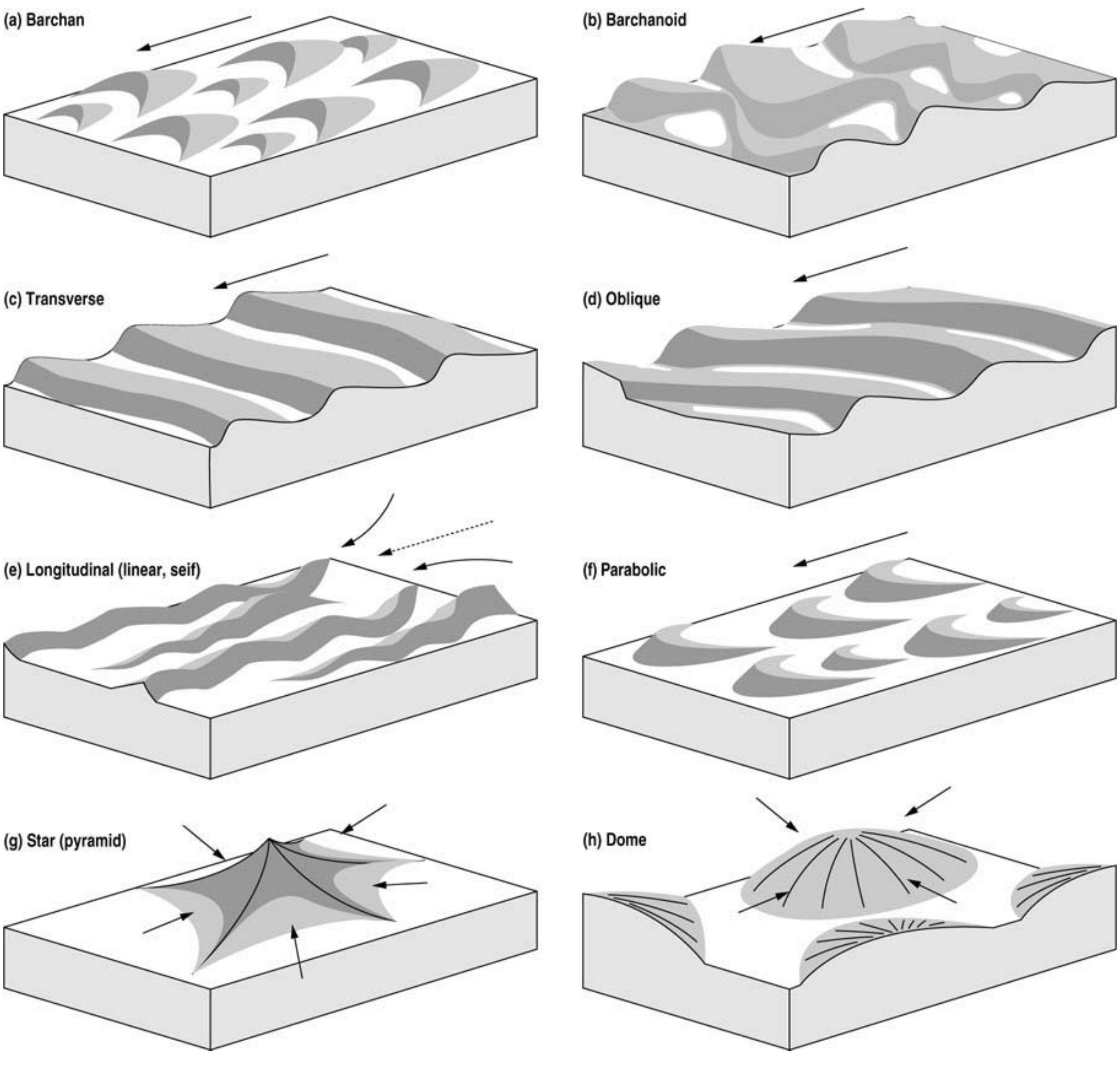


Figure 6.68 Three-dimensional forms of some common dune types. The arrows mark the dominant directions of the effective winds, and in **(e)** the dotted arrow indicates the resultant effective direction.

in phase or out of phase (Fig. 6.71). **Parabolic dunes** form U-shapes closing down wind 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 3-D linear bedform in which the crestline sinuosities migrate down wind along the bedform crest in a motion similar to that of a snake (Fig. 6.71d). The spacing between seif dunes is typically approximately twice their mean width. The crests of such dunes exhibit a regular sinuosity, with slipfaces on alternative flanks. Upwind ends of ridges are rounded, and along their length Y-shape 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. **Zibars** are low ridges of coarse-grain hard-packed sand without slipfaces, which are usually aligned transverse to the wind and often

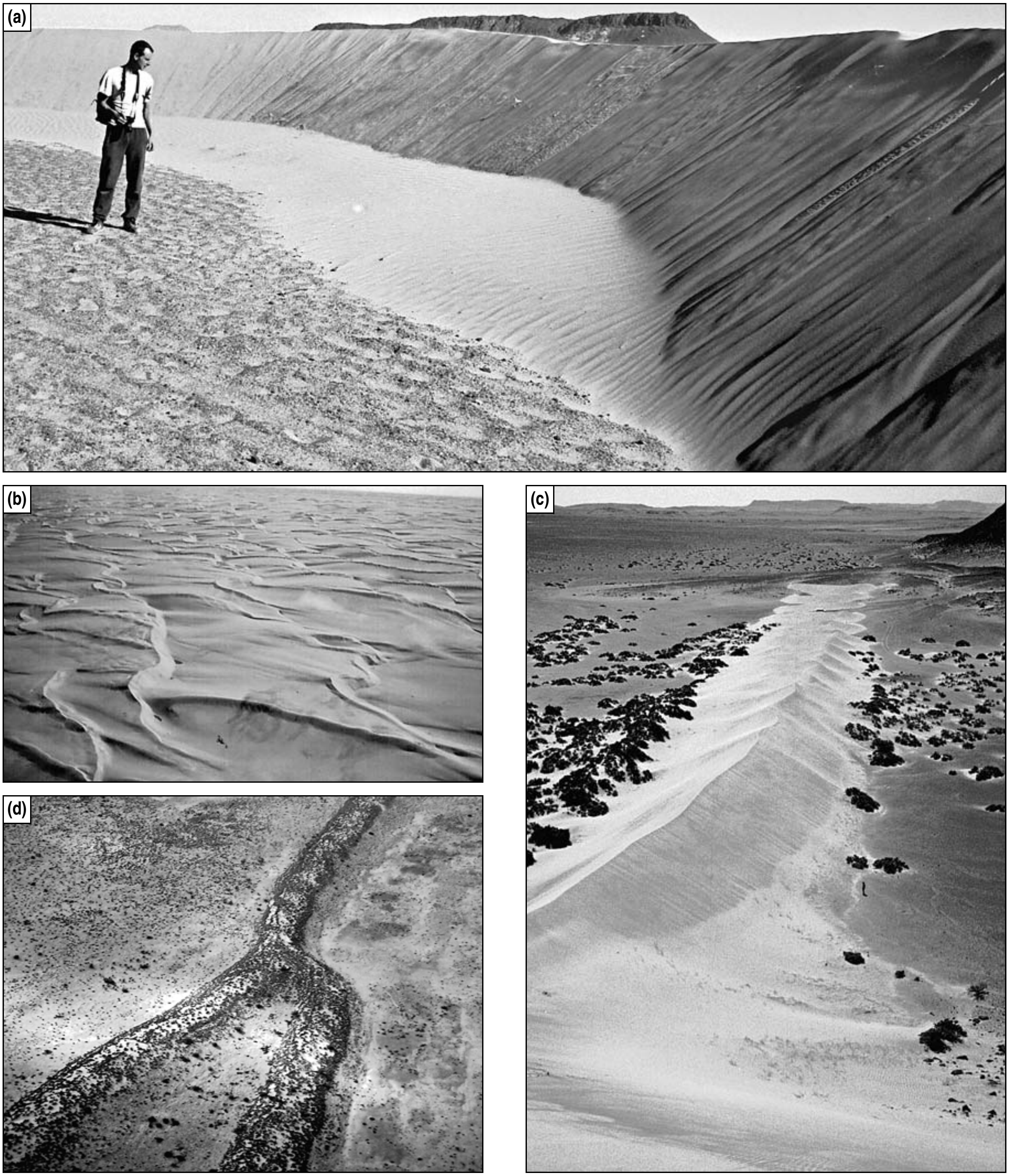


Figure 6.69 Examples of aeolian dune morphologies. **(a)** Lee face of a crescent-shape barchan dune, Huab Basin, Namibia. **(b)** Oblique aerial view of sinuous-crested transverse bedforms, western Namib Sand Sea, Namibia; foreground width is approximately 600 m. **(c)** Straight-crested linear dune, Huab Basin, Namibia. Bedform is 25 m wide. **(d)** Aerial photograph of stabilized linear dunes with wide interdune areas. What is the likely pattern of effective winds? Dunes are around 100 m wide. Strzelecki Desert, South Australia.

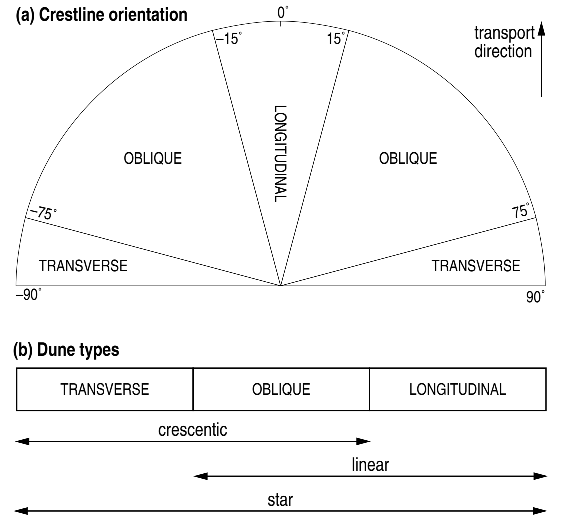


Figure 6.70 Classification of dunes: **(a)** morphodynamic dune types based on orientation of crestline relative to resultant transport direction; **(b)** probable range of morphological and morphodynamic dune types (modified after Hunter et al. 1983).

occur in the corridors between seifs or as independent bedforms in sandsheets.

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 change through time in migration direction, speed, asymmetry or steepness are **variable**. A temporal change in the style of migratory dune behaviour is one potential explanation for the origin of geometrically complex bedsets in the ancient record.

Dunes are often 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 down wind) sectors, which are either in or out of phase in relation to those in an adjacent ridge (Fig. 6.71a). Elsewhere there may be straight-crested dunes (draa) transverse to the wind and, close to them, **dome-shape** dunes with many minor slipfaces 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. 6.72); this often enables subtle trends regarding bedform morphology and arrangement to be discerned.

Draa

Draa are larger bedforms than dunes, with wavelengths of 500–5000 m and heights exceeding 50 m. These “megadunes” occur only in the largest sand seas, where aeolian sediment supply and transport rates are high. Draa are described using the same terminology as for dunes, but may also be characterized by the presence of superimposed dune-scale bedforms on their flanks. **Simple** draa lack superimposed dunes, but **compound** draa have superimposed dunes of the same morphological type, and **complex** draa have superimposed dunes of a different type. Some compound and complex draa may be slipfaceless, in that the megadunes themselves do not possess an active slipface, even if the dunes superimposed upon them do. Common examples of compound draa forms include longitudinal draa that have smaller seif dunes aligned across their slipfaceless flanks, and barchan draa that support smaller superimposed barchan dunes. Common examples of complex draa forms include star draa, which have radiating arms that support superimposed 3-D transverse forms, and linear draa, which have superimposed transverse ridges that migrate along their flanks (Fig. 6.71e). The migration of superimposed dunes over larger, more slowly moving draa results in the generation of complex bedsets in the stratigraphic 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. 6.73). 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** (RDD) and can be determined by plotting sand-drift rose diagrams (Fig. 6.73c,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 very effectively separate different dune types (Fig. 6.74), which suggests that dune type is controlled by wind **regime** as well as by the

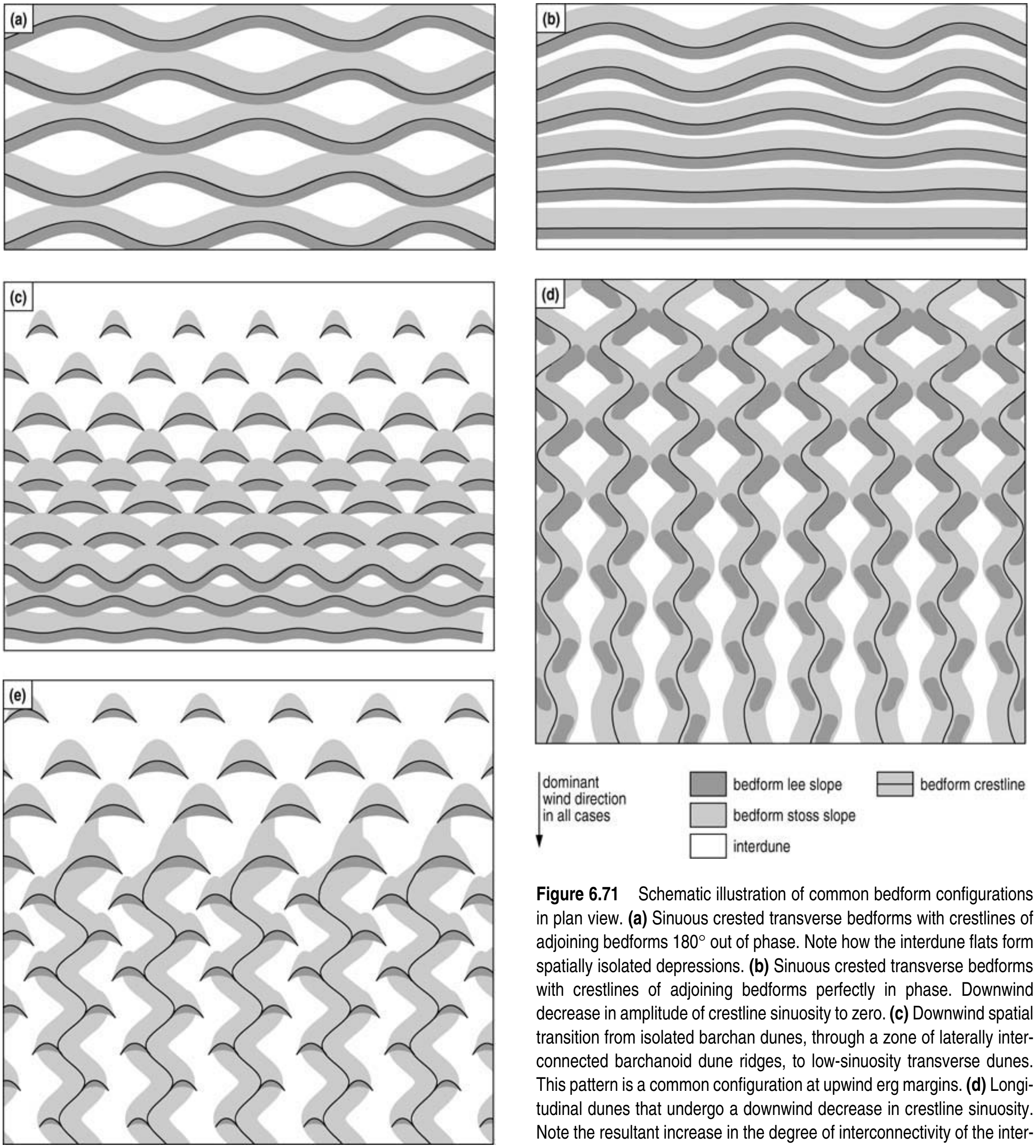

(a)
(b)
(c)
(d)
(e)
dominant wind direction in all cases
bedform lee slope
bedform stoss slope
interdune
bedform crestline

Figure 6.71 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 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.

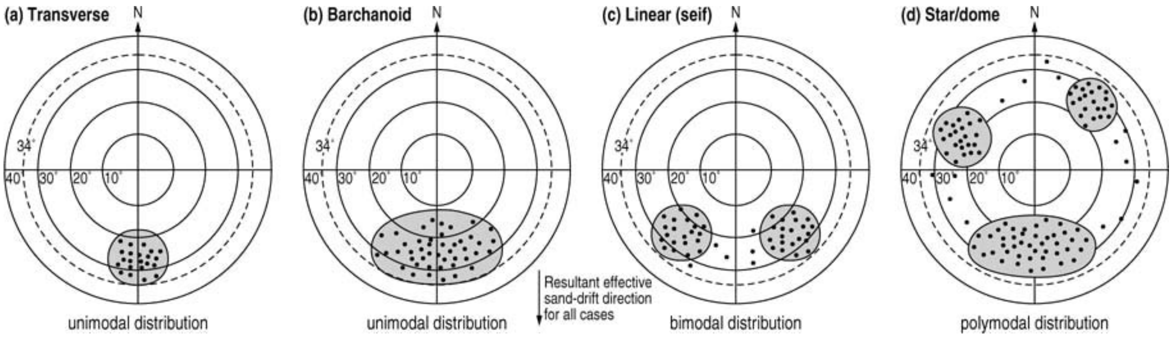
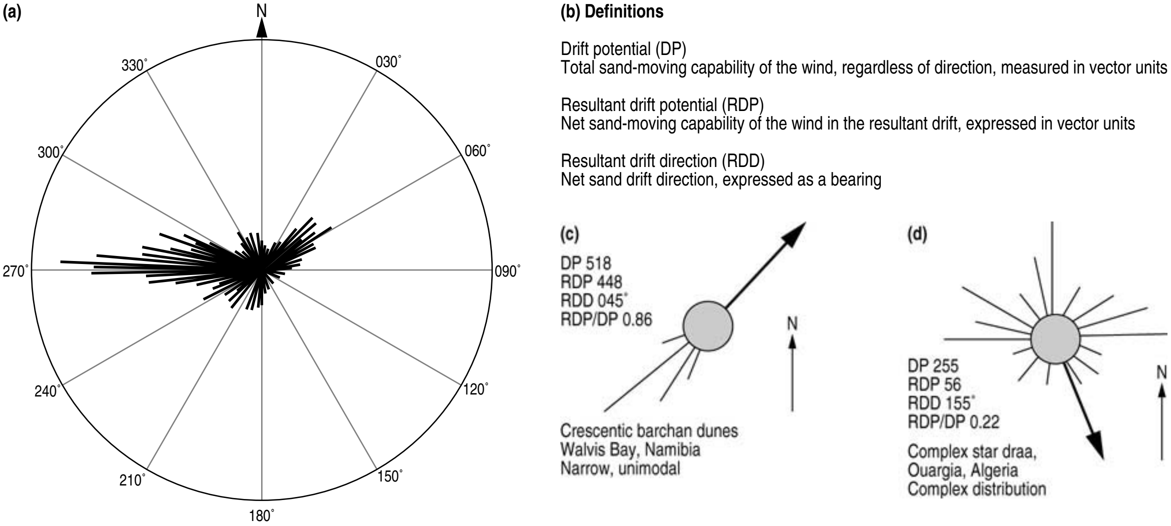


Figure 6.72 Schematic diagram illustrating palaeocurrent patterns predicted for common aeolian dune types, plotted as dip azimuths of lee slope foreset laminae. The angle of repose in well sorted, loose dry aeolian sand is typically 34°, and bedform lee slopes tend not to exceed this angle. Foreset laminae preserved in ancient successions tend not to exceed 26° because of the effects of compaction. Foreset dip azimuths from linear, star and dome-shape 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 orientated in that direction.



(b) Definitions

Drift potential (DP)
Total sand-moving capability of the wind, regardless of direction, measured in vector units

Resultant drift potential (RDP)
Net sand-moving capability of the wind in the resultant drift, expressed in vector units

Resultant drift direction (RDD)
Net sand drift direction, expressed as a bearing

(c)
DP 518
RDP 448
RDD 045°
RDP/DP 0.86

Crescentic barchan dunes
Walvis Bay, Namibia
Narrow, unimodal

(d)
DP 255
RDP 56
RDD 155°
RDP/DP 0.22

Complex star draa,
Ouargia, Algeria
Complex distribution

Figure 6.73 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 min 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: drift potential (DP) is a measure of the relative sand-moving capability of the wind, without regard to the direction of that movement; it is derived from surface wind data via a weighting equation; the resultant drift potential (RDP) averages drift potential values to derive a value for the net sand transport capability in the resultant drift direction (RDD). **(c, d)** Examples of 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 drift potential. The length of the heavier arrowed line indicates resultant drift potential. The orientation of this line indicates the resultant drift direction. 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 under 0.2 indicate variable winds and favour the development of star or dome dunes. (Modified in part after Fryberger 1979)

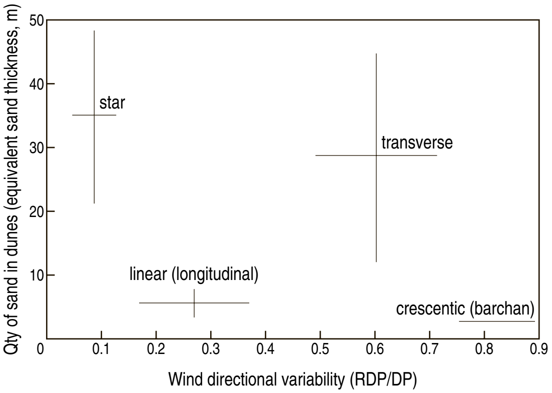


Figure 6.74 Application of the concept of equivalent sand thickness (EST), drift potential (DP) and resultant drift potential (RDP). The graph depicts the relationship between dune type, wind regime and equivalent sand thickness. Transverse and barchan dunes develop under unimodal wind regimes (RDP/DP > 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 & Hyde 1983)

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. 6.74), whereas star-dune forms tend to develop in response to variable winds (low RDP/DP values) and are sand-accumulating bedforms, which do not migrate great distances. However, it is important to realize that factors such as vegetation cover may complicate the situation further.

In describing aeolian dunes and draa, measure bedform width, wavelength, height and geometry of crests (see Fig. 6.68); 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 (see 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 crosswind and a re-attachment point of the flow. The use of stakes may help to monitor the rate of advance of parts of a dune. It may be possible to measure sandflow into and off the dune and to plot a sand-rose diagram (Fig. 6.73). 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 directions both parallel and perpendicular to the dominant wind 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. 6.71).

6.3.4 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 ancient aeolian strata, comprise four basic stratification types that occur in various configurations: **grainflow strata**, **grainfall strata** (Figs 6.55, 6.56), **wind-ripple strata** (see §6.1.6) 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. 6.75a). Grainflow avalanches generate foresets that are inclined at 18–34° and are 2–5 cm thick, although they often thin up slope. They show a tongue- or cone-shape 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. 6.75b), although they may show inverse grading normal to the base, and lateral grading along their length. At the foreset toe, coarse grains may occur at the top of the layer. Heavy minerals often line the upper parts of foresets. In modern dune sands, packing is very loose and porosity is very high (up to 45%). **Slump degradation** grainflows, where internal structure is destroyed as the flow travels down slope, are characterized by a chaotic wedge of loosely packed sediment up to a few metres wide that thickens down slope before eventually pinching out. **Scarp recession** grainflows, where an initial point of failure generates a scarp that then retreats back up slope towards the brinkline, form tongue-like bodies that rarely exceed 0.5 m in width but may extend almost the full height of the lee slope (Fig. 6.75a).

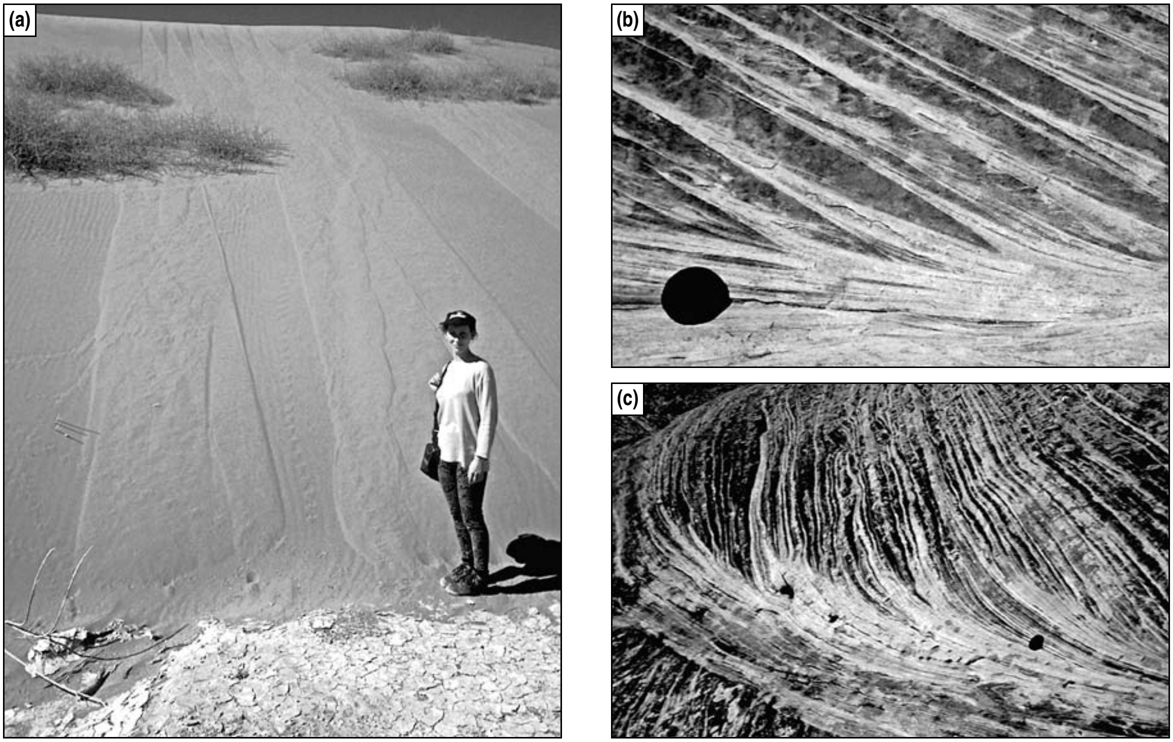


Figure 6.75 Characteristic deposits of aeolian sand dunes. **(a)** Grainflow tongues generated by gravitational collapse and avalanching of part of the lee face of a sand dune. Sossusvlei, Namib Sand Sea, Namibia. **(b)** Grainflow strata seen in section. Individual grainflow tongues thin and pinch out 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-grain sandstone. Cedar Mesa Sandstone, Permian, White Canyon, Utah. **(c)** Oblique view of a cross-stratified aeolian dune set, with grainflow and grainfall strata interfingering with wind-ripple strata at the set base. Cedar Mesa Sandstone.

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, although they may sometimes exhibit a wedge-shape geometry that is thickest just leeward of the dune brinkline and which thins down slope. Grainfall units in modern dune sands are often moderately packed, and porosity may reach 40 per cent. Grainfall strata have non-erosional contacts, and individual units often 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 often extend uninterrupted down to the base of the lee slope, commonly blanketing grainflow strata (Fig. 6.75b). On larger bedforms, grainfall wedges tend to be cut out by grainflow strata.

Aeolian wind-ripple stratification

Wind-ripple strata are described fully in §6.1.6 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 draa slopes inclined at 0–25°, where they form stratified units 1–15 cm thick. Contacts between laminae can be erosional, non-erosional, 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 per cent. 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, sometimes in rhythmic sequences (Fig. 6.75c).

Aeolian plane bed stratification

Aeolian flat-bedded laminae on the slopes of dunes and draa dip at 0–15°, are parallel, even, and form sets of laminae 1–10 cm 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 often less than 30 per cent. These laminae occur only rarely and always on upwind-facing stoss or gently sloping downwind-facing lee surfaces.

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

Although the external forms of modern aeolian dunes and draa are readily apparent and well documented from aerial photographs and satellite images, the organization of their internal structures is less well known. Conversely, in the ancient rock record, bedsets of presumed aeolian dune and draa 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 the migratory behaviour of aeolian dunes and draa 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-3-D 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 dunefields, 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 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 10–20 m. In the ancient record there are now a few documented examples where large-scale preservation of aeolian bedform topography reveals both the original bedform morphology and its internal structure. Finally, because large-scale aeolian bedforms undergo relatively slow rates of growth and migration, their temporal evolution and migratory behaviour are difficult to study directly. One solution to this has been the development of computer models that simulate temporal bedform behaviour and predict how 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 some 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. 6.76), wherein large-scale cross bedding is the most striking feature, with single large-scale sets commonly ranging in thickness up to 10 m and occasionally attaining as much as 52 m. Common sedimentary structures associated with aeolian dune activity include:

- A dominance of medium- and large-scale cross bedding composed of grainflow, grainfall and wind-ripple strata in varying arrangements (Fig. 6.75; §6.3.4).

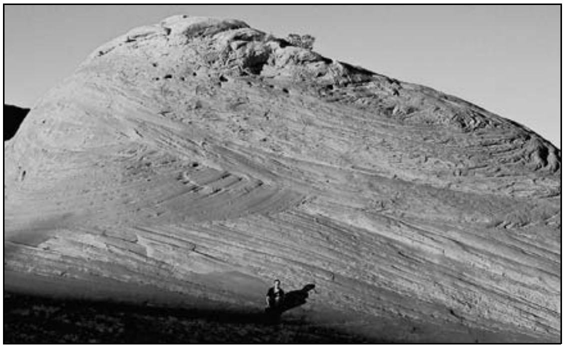


Figure 6.76 Bounding surfaces in aeolian sandstones that define large-scale trough-shape sets filled with cross strata; Navajo Sandstone, Jurassic, Utah.

- The occurrence of cross bedding of several types, including tabular-planar, wedge-planar and trough types, the relative proportions of which vary among examples. Bundles of convex upward laminae may occur in several dune types.
- 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 subparallel to highly oblique relative to the foresets that they truncate.
- Trough cross beds that often occur as solitary sets or thin cosets in the upper parts of dunes.
- Thin units of flat lamination or wind-ripple lamination (or both) that often occur on the stoss sides and crestal areas of dunes.

The geometry of sets of cross strata and the orientation of foresets within aeolian bedforms vary with bedform type. Transverse dunes and draa are composed internally of planar cross-bedded sets with foresets inclined perpendicular to the trend of the bedform crestline (Fig. 6.77). Sinuous-crested barchanoid ridges are composed internally of trough cross-bedded sets that are often rather complicated in terms of their internal architecture and with foresets inclined in a variety of directions (Fig. 6.78). Dome-shape bedforms often exhibit sets composed of gently convex-up laminae (Fig. 6.79). The internal structure of linear and star-shape bedforms is complicated, often characterized by a mosaic of overlapping and intersecting sets, with foresets inclined in opposing directions (Fig. 6.80).

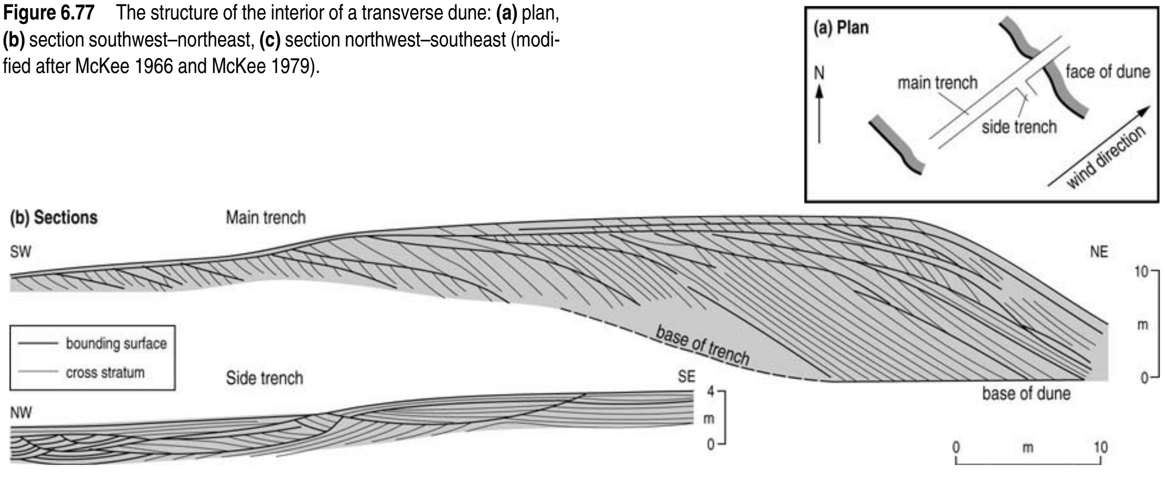
Bounding surfaces

Bounding surfaces are erosional surfaces that truncate sets of aeolian dune (or draa) cross strata. Four distinct types of bounding surface are recognized by their shape, their orientation relative to the cross strata that they truncate, their lateral extent and their relation to one another (Fig. 6.81).

Reactivation surfaces occur within aeolian sets and are characterized by planar or scallop-shape erosion surfaces that typically dip down wind 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 subparallel to the cross strata and can sometimes be traced for 10–100 m and more along strike; in sections parallel to transport they may extend the full height of a set or may be restricted to either its basal or upper part. Reactivation surfaces either occur randomly within sets or they may exhibit regular spacings. Overlying cross strata exhibit a relationship that will be either a concordant or downlapping.

Superimposition surfaces occur within aeolian cosets and are characterized by planar to highly scalloped 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 orientated oblique to the cross strata that they truncate. Where both reactivation and superimposition surfaces

Figure 6.77 The structure of the interior of a transverse dune: **(a)** plan, **(b)** section southwest–northeast, **(c)** section northwest–southeast (modified after McKee 1966 and McKee 1979).



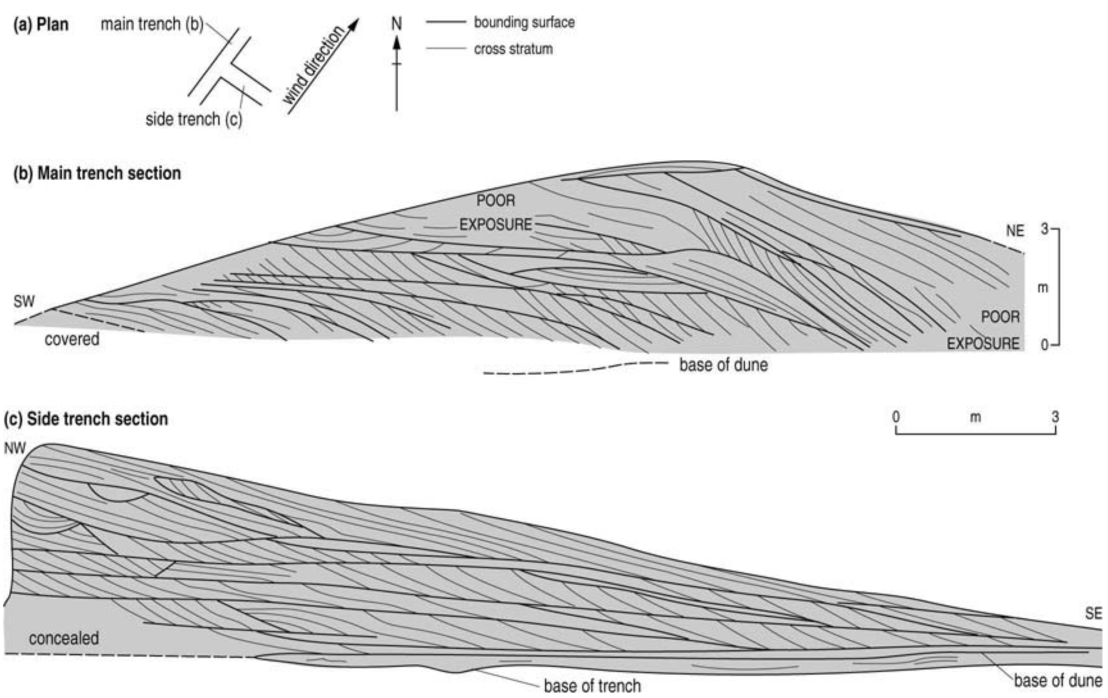

(a) Plan
main trench (b)
side trench (c)
wind direction
N
bounding surface
cross stratum
(b) Main trench section
POOR
EXPOSURE
SW
covered
NE
3
m
0
POOR
EXPOSURE
base of dune
(c) Side trench section
0
m
3
NW
concealed
SE
base of trench
base of dune

Figure 6.78 The structure of the interior of a barchanoid ridge dune: **(a)** plan, **(b)** wind-parallel section (southwest–northeast), **(c)** wind-perpendicular section (northwest–southeast). This example shows a complexity of internal lamination that would not have been expected from the external morphology and suggests a complex evolution. (Modified after McKee 1966 and McKee 1979)

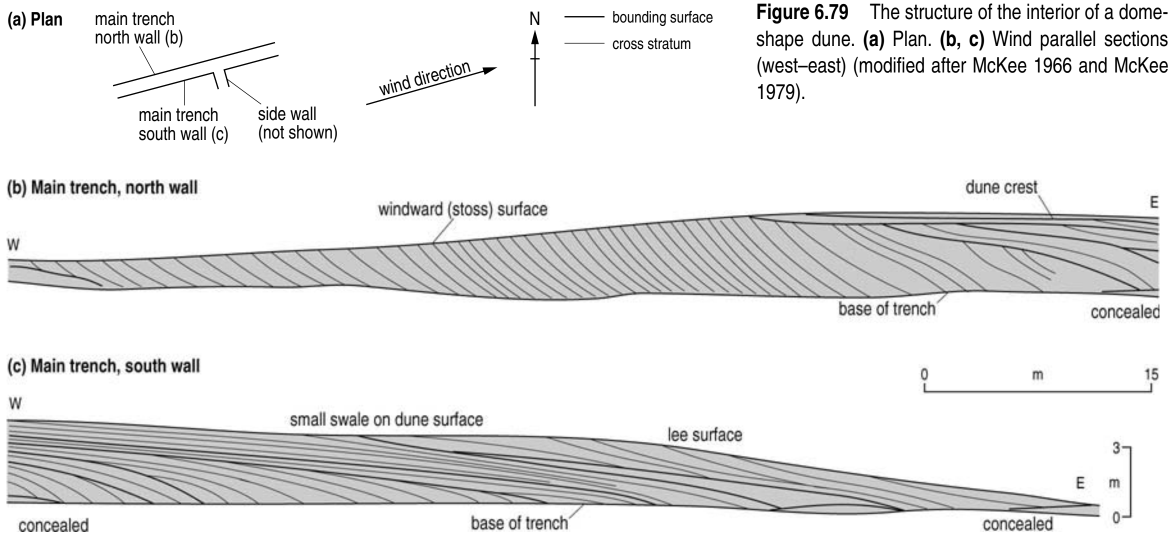
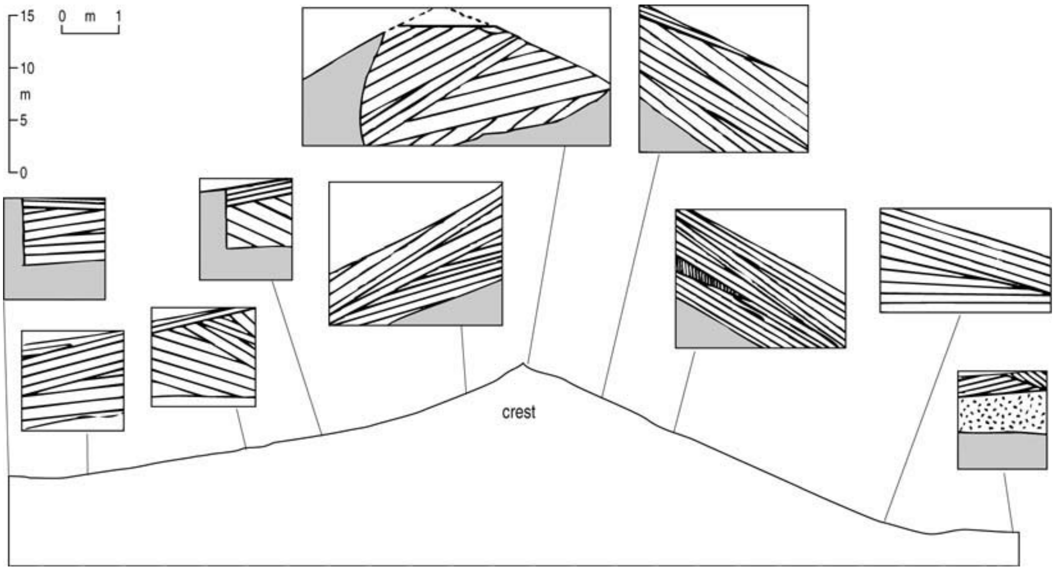

(a) Plan
main trench
north wall (b)
main trench
south wall (c)
side wall
(not shown)
wind direction
N
bounding surface
cross stratum
(b) Main trench, north wall
W
windward (stoss) surface
dune crest
E
base of trench
concealed
(c) Main trench, south wall
0
m
15
W
small swale on dune surface
lee surface
3
E
m
0
concealed
base of trench
concealed

Figure 6.79 The structure of the interior of a dome-shape dune. **(a)** Plan. **(b, c)** Wind parallel sections (west–east) (modified after McKee 1966 and McKee 1979).

Sketches of cross bedding revealed following excavation of sections in a linear (seif) dune, Libya



Sketch of a theoretical model for a linear (seif) dune after Bagnold (1941)

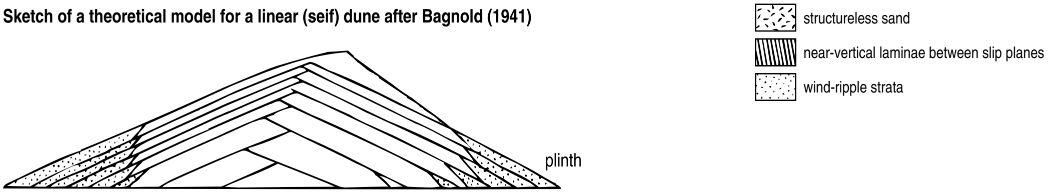


Figure 6.80 The structure of the interior of a linear (seif) dune: (above) direct evidence of a real example, based on shallow excavations of a seif dune in Libya (after McKee & Tibbits 1964); (below) Bagnold's hypothetical model (1941), as modified by McKee (1979) to show foreset dips in opposed directions.

are developed, the latter always truncate the former.

Interdune migration surfaces are characterized by low-angle inclined erosive surfaces that typically extend down wind for distances of hundreds of metres to several kilometres. These surfaces, which bound sets or cosets, appear to be of planar to slightly scalloped shape 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 often have great lateral extent and continuity, such that they bound entire aeolian accumulations or significant parts thereof. Sedimentary features present on 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 non-aeolian environments and relate them to basinwide events such as marine transgressions.

6.3.6 Processes of formation of dunes, draa and aeolian stratification

External form

The formation of dunes and draa 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

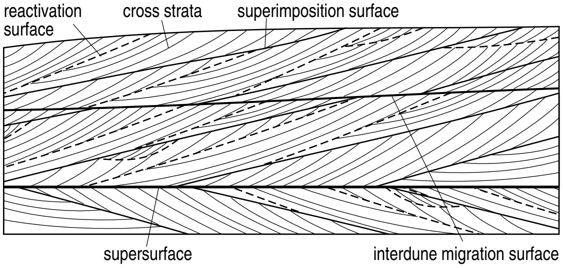


Figure 6.81 Definition diagram for the concept of bounding surfaces within compound cross-bedded sands and sandstones. Although this terminology was first proposed for aeolian sandstones, it can also be applied to water-lain sediments. Interdune migration surfaces arise as a consequence of dune migration. Superimposition surfaces represent the migration of superimposed bedforms or scour pits over a larger parent bedform. Reactivation surfaces represent partial deflation of a bedform lee slope; they arise in response to periodic changes in bedform migration direction. Supersurfaces are usually regionally extensive and often bound entire erg accumulations. (Modified after Brookfield 1977)

deserts suggest that for a patch of sand to develop into a dune it has to be at least 4–6 m 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 slipface may develop. Dunes and draa are regularly repeated bedforms that record the response of a sandbed 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 topographical obstacles that disrupt the **primary airflow**, such that, as the flow moves up the dune stoss slope, it is compressed and it accelerates, thereby causing an increase in transport rate and promoting transport up the stoss slope to the dune crest. As the flow moves over the crest it expands 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 down wind over time. **Flow separation** of the airflow from the bedform surface occurs beyond the crest, whereas **flow re-attachment** occurs a distance of a few dune heights down wind. Thus, a **separation cell** exists in the dune lee within which turbulent **secondary airflow** occurs. This allows ripples and erosional scour hollows on the dune flanks (**plinth**) to undergo complex migratory behaviour. Down wind of the re-attachment point, renewed flow acceleration means that interdune sediments may potentially be eroded, thereby providing a local sediment supply for the next dune down wind in the train.

Historically, ideas on the controls of dune or draa 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. 6.74). Barchans, longitudinal and seif structures are associated with low sand availability; transverse barchanoid ridges and star forms develop where abundant sand is available. Barchans and transverse forms reflect low diversity of effective wind directions; 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 straightforward. 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 bedset (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.

Formation of small-scale structures

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

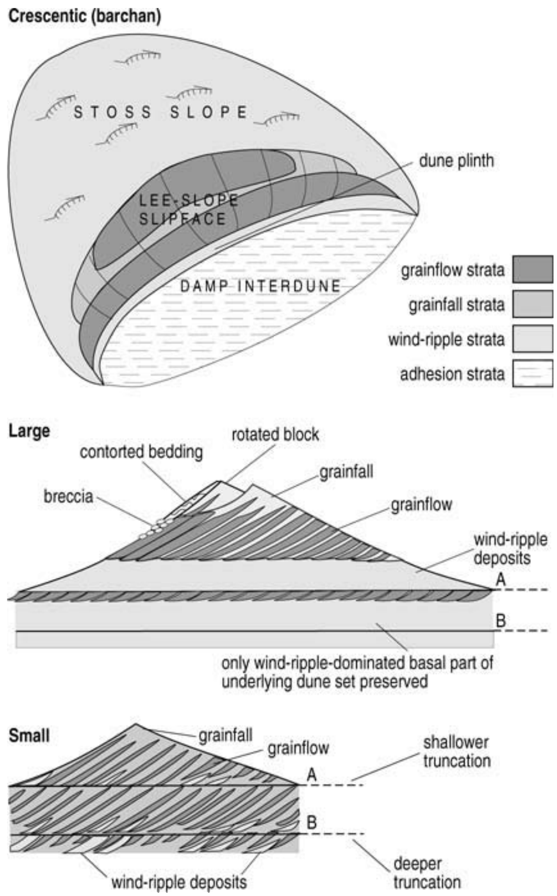

Crescentic (barchan)
STOSS SLOPE
dune plinth
LEE-SLOPE SLIPFACE
DAMP INTERDUNE
grainflow strata
grainfall strata
wind-ripple strata
adhesion strata
Large
rotated block
contorted bedding
grainfall
breccia
grainflow
wind-ripple deposits
A
B
only wind-ripple-dominated basal part of underlying dune set preserved
Small
grainfall
grainflow
shallower truncation
A
B
deeper truncation
wind-ripple deposits

Figure 6.82 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). The 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 & Dott 1981)

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°) and an active slipface develops that is subject to gravity-driven collapse, resulting in the generation of various types of grainflow (sandflow or avalanche) strata (Figs 6.82, 6.83). Their inverse grading is the result of 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 expansion and separation within the lee-side eddy on the upper part of a dune's lee slope. Grainfall may also occur on the

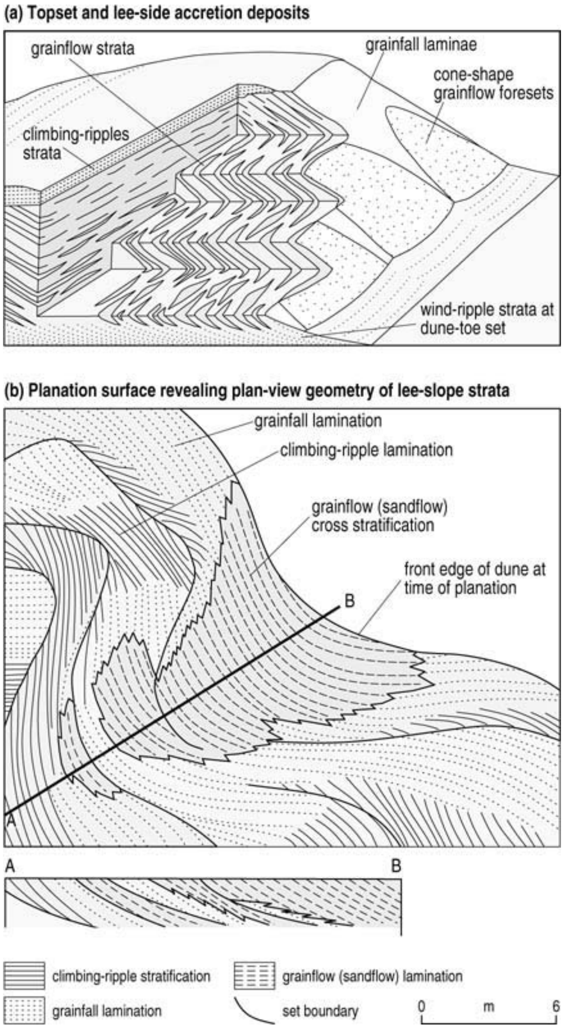

(a) Topset and lee-side accretion deposits
grainflow strata
grainfall laminae
cone-shape grainflow foresets
climbing-ripples strata
wind-ripple strata at dune-toe set
(b) Planation surface revealing plan-view geometry of lee-slope strata
grainfall lamination
climbing-ripple lamination
grainflow (sandflow) cross stratification
front edge of dune at time of planation
B
A
A
B
climbing-ripple stratification
grainflow (sandflow) lamination
grainfall lamination
set boundary
0
m
6

Figure 6.83 **(a)** Schematic block diagram showing the different small-scale structures of foresets of different type: simple cone-shape grainflow foresets, grainfall laminae and climbing-ripple strata. Plane bed lamination is often 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. (Both after Hunter 1977) flanks or apron of a dune where secondary crosswind airflow is not strong enough to form ripples (Figs 6.82, 6.83). Distinct grainfall lamination occurs because of grain-size variations produced during fluctuations of transport power. Porosity is high because of a lack of sustained saltation, which generates relatively tight packing. 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.

Less steeply 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 that 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 ripple-producing saltation 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 it develops through repeated and continuing lee-slope sedimentation, whereby grainflow, grainfall and wind-ripple strata generate cross stratification (Figs 6.82, 6.83). The interiors of most aeolian bedforms are composed of cross-bedded sands, and the cross strata provide a record of the successive 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-shape forms will generate cross strata exhibiting a range of foreset dip azimuths that vary around a mean reflecting 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-shape dunes. The predicted pattern of foreset dip azimuths for linear (especially seif) dunes is of two modes about 120° apart in a bimodal distribution (Fig. 6.72). 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 or steepness, more complicated patterns of cross strata are produced such that sets contain many reactivation 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 main 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 sequences (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. The usual approach involves the use of computer simulations, whereby the patterns of cross bedding generated by bedforms with various morphological forms, and which undergo various styles of migratory behaviour, are compared to the observed cross-bedding patterns from ancient strata until a good match is found.

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. One potential problem with the analysis of aeolian bounding surfaces is that geometrically similar bounding surfaces 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.

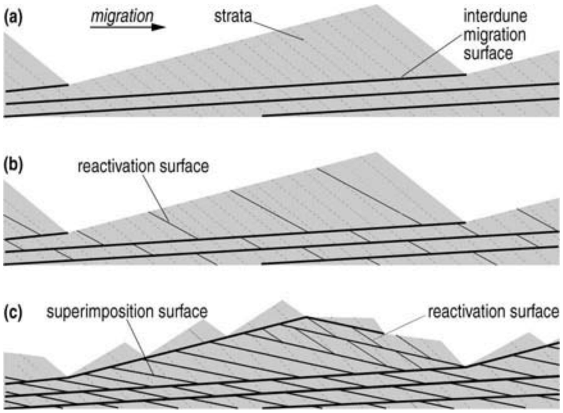

(a)
migration
strata
interdune migration surface
(b)
reactivation surface
(c)
superimposition surface
reactivation surface

Figure 6.84 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.

Reactivation, superimposition and interdune migration bounding surfaces are generated by dunes of different types and sizes, moving in different directions and at different rates relative to each other (Fig. 6.84). By contrast, supersurfaces result from externally controlled changes to key parameters such as sand supply, sediment availability and the sediment-carrying capacity of the wind.

Reactivation surfaces (§6.3.5) result from periodic lee-slope erosion followed by renewed sedimentation associated with a change in the direction and speed of bedform migration, asymmetry, or steepness (Fig. 6.84b, 6.85). The surfaces record erosion during anomalous wind intervals and compare with reactivation surfaces in aqueous cross bedding (see §6.26). These changes are common because airflow on lee slopes is often subject to turbulent modification and is rarely steady. In some cases, the period of the flow fluctuation is regular and it 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. 6.86).

Superimposition surfaces form when either superimposed dunes migrate over a larger parent bedform, or scour troughs migrate on the lee slope of a bedform (Fig. 6.84c, 6.85). 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

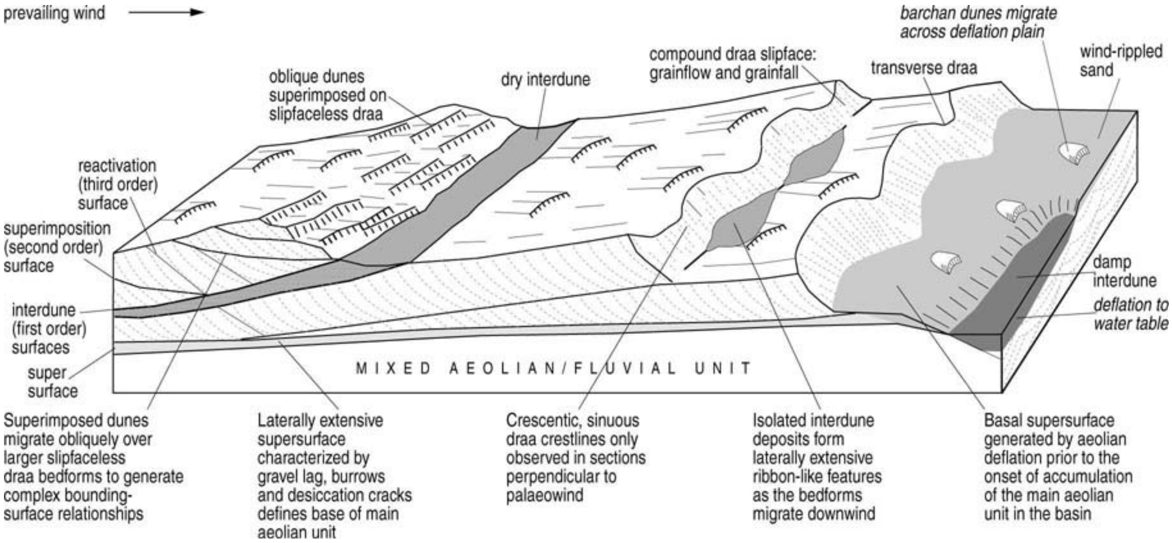

prevailing wind
oblique dunes superimposed on slipfaceless draa
dry interdune
compound draa slipface: grainflow and grainfall
transverse draa
barchan dunes migrate across deflation plain
wind-rippled sand
reactivation (third order) surface
superimposition (second order) surface
interdune (first order) surfaces
super surface
damp interdune
deflation to water table
MIXED AEOLIAN/FLUVIAL UNIT
Superimposed dunes migrate obliquely over larger slipfaceless draa bedforms to generate complex bounding-surface relationships
Laterally extensive supersurface characterized by gravel lag, burrows and desiccation cracks defines base of main aeolian unit
Crescentic, sinuous draa crestlines only observed in sections perpendicular to palaeowind
Isolated interdune deposits form laterally extensive ribbon-like features as the bedforms migrate downwind
Basal supersurface generated by aeolian deflation prior to the onset of accumulation of the main aeolian unit in the basin

Figure 6.85 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).

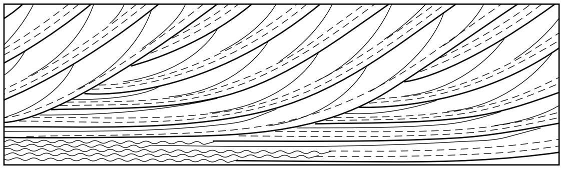


Figure 6.86 Schematic diagram illustrating a scalloped coset of cross strata with internal cyclicity. Two distinct scales of bounding surface are evident within the coset. Note how bounding surfaces at the base of the sets pass down dip into wavy or corrugated surfaces. This relationship is indicative of the 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, northeast Utah. (Modified from Crabaugh & Kocurek 1993)

of a parent bedform, oblique migration is more common because secondary airflow directs superimposed bedforms obliquely across the lee slope of the parent bedform.

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. 6.84a, 6.85). 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 by either bypass or deflation, because of 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 the water table acts to limit the extent of aeolian deflation, the resultant supersurface is sometimes called a **Stokes surface** (Fig. 4.28).

6.3.6 The uses of aeolian dunes, draa and cross bedding

The greatest use of these structures is in interpreting palaeoenvironments. Aeolian rocks contribute to palaeoenvironmental reconstructions at continental and global scales because they help to locate major meteorological systems. Palaeowind studies are particularly helpful in this regard, but remember that dunes and draa reflect the effective winds, not necessarily weak prevailing or seasonal winds (or both). Coarser-grain sand may be moved only by strong winds, which may relate to only one season or direction (or both). There is no reason why palaeowind directions should have any relationship with palaeoslope.

Although aeolian processes produce distinctive sedimentary textures, they can be reliably used for palaeoenvironmental reconstruction only 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. However, care is needed for straight angular-based foresets. Additionally, convex-upwards cross bedding may not be recognized to be the right way up unless the whole sequence and context are considered.

Finally, thick aeolian sequences of porous, permeable, cross-bedded sandstone with few impermeable barriers are important as potential reservoirs for oil, gas, water and hydrothermal metalliferous deposits.

6.4 Flat beds and parallel lamination

6.4.1 Introduction

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

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

6.4.2 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-grain or slightly pebbly sandstone, a lineation is most commonly developed on the surfaces of slightly finer-grain 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. 6.87).

The lineation comprises a set 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 flow.

6.4.3 Internal structure: parallel lamination

Excavations into flat sand surfaces and vertical sections in sandstones with flat bedding planes usually reveal parallel lamination. Laminae are usually only a few grain diameters thick and, in coarse, less well sorted sandstones, they may barely exceed the thickness of the coarsest grains. The laminae are defined by slight grain-size differences or by concentrations of mica.

6.4.4 Process of formation

A flat bed is a distinct bedform produced by particular flow conditions. Where the sand is of medium to fine grain 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 so-called rapid flow (see §3.2.6). Similar conditions also develop during the deposition of sand from turbidity currents, and parallel lamination is a common feature of 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 occurs only with fine-grain sediments, this process may not explain lineation in coarse-grain sands. It is possible that alternating spiral vortices close to the bed may then be responsible for the lineations (Fig. 6.88).

Experimental work shows that ripples do not form in sands coarser than about 0.6 mm, but that movement occurs on a so-called "lower flow-regime flat bed" for

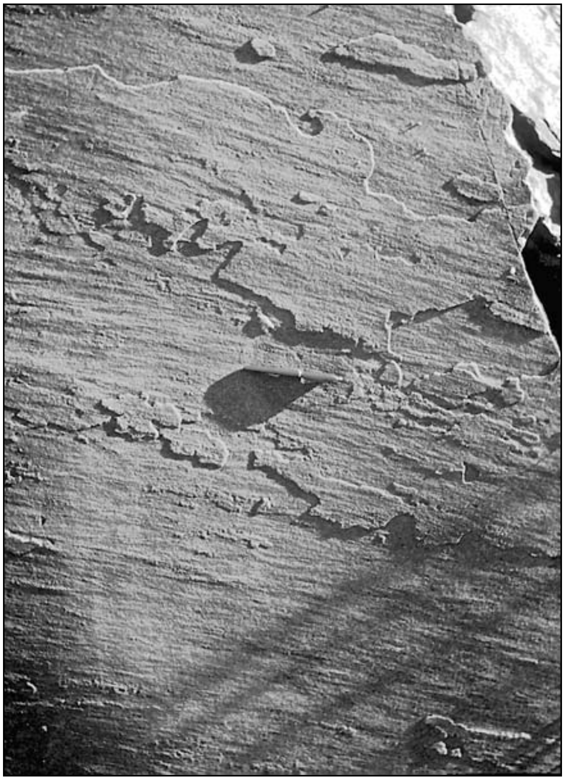


Figure 6.87 Primary current lineation or parting lineation on the bedding surface of parallel-laminated sandstone. The steps between adjacent laminae trend parallel to the lineation. Cloughton Formation, Middle Jurassic, Yorkshire.

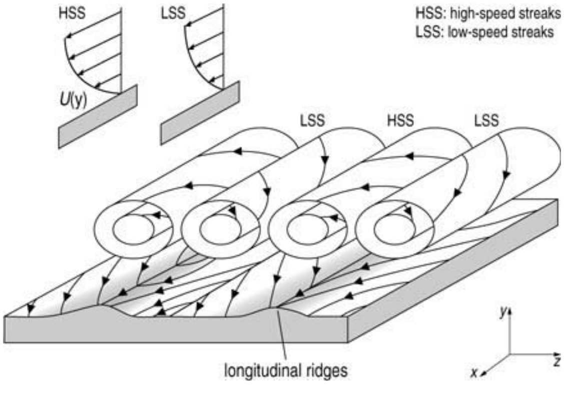


Figure 6.88 Idealized fluid motion associated with streakiness within the near-wall region of a turbulent boundary layer. Streaks of fluid moving at slightly different velocities initiate longitudinal vortices that cause locally converging flow where grains accumulate into longitudinal ridges. This longitudinal fabric produces primary current lineation on bedding planes. (Modified after Allen 1985b)

flows 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.

Other experiments suggest that abundant mica inhibits the formation of ripples, whose existence depends, among other things, on the ability of grains to avalanche down the lee faces of the bedforms. Some highly micaceous parallel-laminated sands and sandstones could reflect deposition from suspension, but concentration of mica into layers implies sorting on the bed. Lineation in such micaceous sands may be attributable 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 seawards 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 produces layers of moving grains with different size characteristics, which deposit laminae when they stop. Another form of roughly parallel horizontal lamination is that produced by multiple freezing of a traction carpet, as discussed in §6.8.5. Parallel lamination in aeolian successions is dealt with in §6.3.4 and §6.3.6.

6.4.5 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 current strength and palaeocurrent or palaeowave direction:

- It is fairly certain that flat beds and 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-grain or very micaceous sands, lower flow-regime conditions may have applied. Flat beds created by waves and currents cannot be differentiated by internal characteristics.
- 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 indicates only the trend and not the sense of flow.

6.5 Undulating smooth surfaces and lamination

6.5.1 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 a delineation of its grain-size limits. A similar style of bedding is sometimes associated with pyroclastic ash deposits of sub-aerial base-surge flows.

6.5.2 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 a metre (Fig. 6.89). The bedding surfaces show a primary current lineation (§6.4), 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.

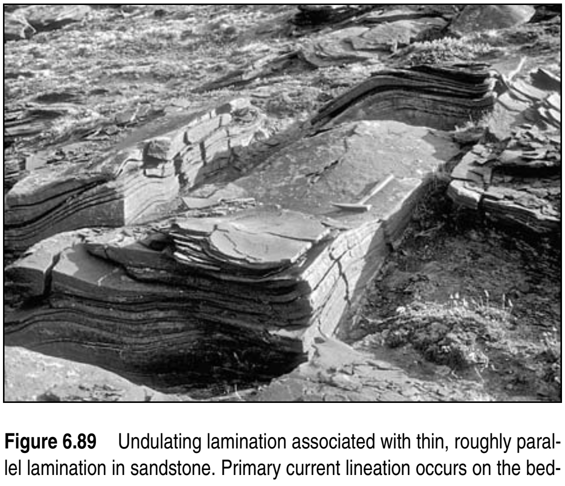


Figure 6.89 Undulating lamination associated with thin, roughly parallel lamination in sandstone. Primary current lineation occurs on the bedding surfaces. Moræneso Formation, Proterozoic, northeast Greenland.

6.5.3 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 stormwater flowing down gutters when gradients are quite high. They also occur at a larger scale during torrential flooding in rivers or in sheet floods (see Fig. 3.11). 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. 6.90). 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 up stream 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 up stream 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 cross bedding that is inclined up stream.

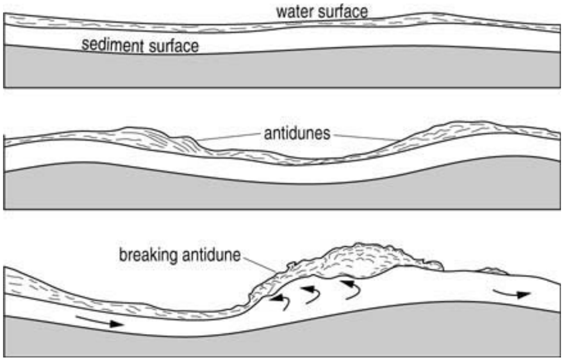


Figure 6.90 Development of standing waves and antidunes as observed in a laboratory flume experiment. Note how the water and sediment waves are in phase with one another and how the antidunes break in an upstream direction. Each section is 1 m long. (Modified after Kennedy 1961)

6.6 Hummocky and swaley cross stratification

6.6.1 Introduction

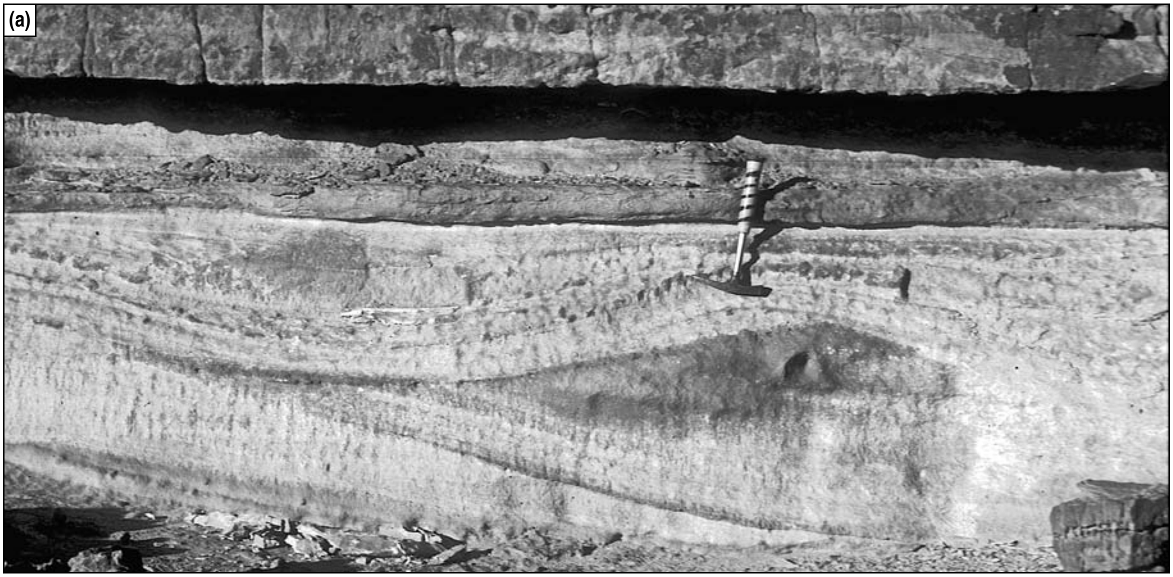
Hummocky and swaley cross stratification is now widely recognized as an important and diagnostic structure in ancient sandstones, although for a long time it was vaguely dismissed as "wavy", "irregular" or "undulating" bedding or lamination. It is most common in fine- to medium-grain sandstone of shallow-marine or nearshore origin. It occurs both within thicker sandstone units and within sharp-based sandstone beds of interbedded sandstone/mudstone sequences.

6.6.2 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 (Fig. 6.91, 6.92). Laminae may thicken into swales and thin over hummocks, so that the undulations gradually die out upwards. Heights of undulations seldom exceed 20 cm and wavelengths are of the order of 1 m. In many cases there is no apparent preferred orientation to the inclination of laminae, suggesting a more or less uniform 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 are commonly signs of erosion in the form of sole marks. Upper contacts are broadly horizontal, often characterized by wave ripples.

6.6.3 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 its occurrence suggests a phase 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 ancient examples suggest that a complex pattern of erosion and rapid deposition occurs. Examples



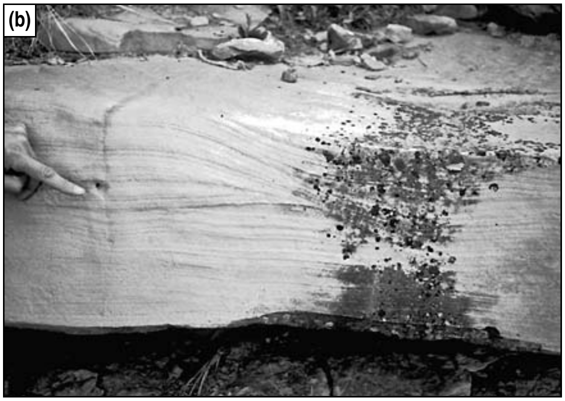


Figure 6.91 Examples of paired hummocky cross stratification (HCS) and swaley cross stratification (SCS). **(a)** Nubian Sandstone, Lower Palaeozoic, Sinai (photo courtesy of Gilbert Kelling). **(b)** Spring Canyon Member, Cretaceous, Book Cliffs, Utah.

showing a preferred orientation to the inclined layers perhaps indicate coexistence of strong wave action and a unidirectional current, a form of combined flow. Arguments remain about how the sand was transported to the site of deposition during storms, with wind-driven, 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.

6.7 Massive sand and sandstone beds

6.7.1 Introduction

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

6.7.2 Description

Massive beds may occur in sand and sandstone of virtually any grain size or sorting. Some massive beds are lenticular; others are parallel sided and may be interbedded with finer-grain sediment. As a rule, 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 such as massive, unlaminated and structureless, especially where there is a clear

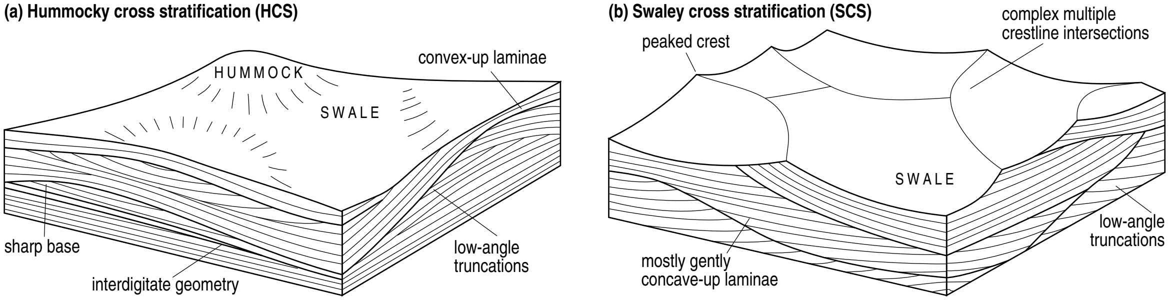

(a) Hummocky cross stratification (HCS)
HUMMOCK
SWALE
convex-up laminae
sharp base
interdigitate geometry
low-angle truncations
(b) Swaley cross stratification (SCS)
peaked crest
complex multiple crestline intersections
SWALE
low-angle truncations
mostly gently concave-up laminae

Figure 6.92 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.

contrast between the structureless beds and neighbouring laminated beds that have undergone similar weathering.

6.7.3 Processes of formation

Absence of lamination may reflect conditions of deposition or it may result from destruction of original lamination. A primary lack of 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 because of liquefaction and flowage. In the case of organic reworking, burrows may be visible in adjacent sediments or may be revealed by X-radiography (§9.4). Where lamination has been destroyed by liquefaction, structures attributable to associated water escape may be present in nearby beds (§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.

6.8 Normally graded beds, inverse grading and the Bouma sequence

6.8.1 Introduction

Certain sharp-based sandstone beds, most commonly in interbedded sandstone/mudstone sequences, 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 also the presence of a sequence of different styles of lamination (the **Bouma sequence**).

6.8.2 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. 6.93). 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 wherever possible. In addition, it is instructive to record the range of size variation through a bed. Grading is also seen in finer-grain sediments, especially siltstones (§5.2.5).

Graded beds are, in many cases, structureless or unlaminated, and where lamination is present it may occur solely in the upper part.

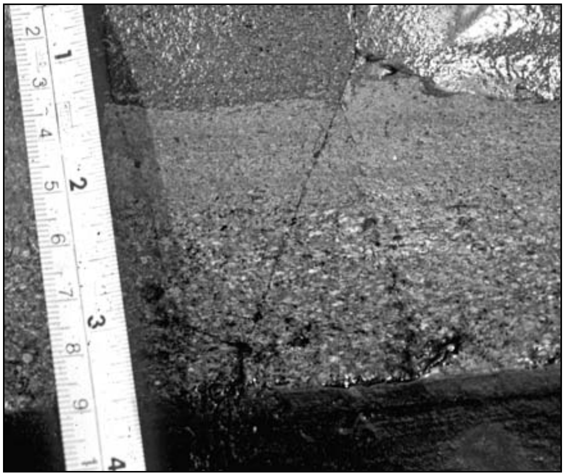


Figure 6.93 Normally graded sandstone bed. Cambrian, Whitesands Bay, Pembrokeshire, Wales.

6.8.3 Inverse grading

Grain size may also show an upwards increase through the bed (**inverse grading**), which, although rare in finer-grain sandstones, is fairly common in coarser sandstones and conglomerates. Inverse grading occasionally occurs as a thin layer at the base of an otherwise massive or normally graded bed. In some thick sandbeds, a succession of thin inversely graded layers may give a crude horizontal stratification to the beds. Such patterns are quite a common feature of pyroclastic sediments. Inverse grading is also dealt with in Chapter 7.

6.8.4 The Bouma sequence

In addition to showing structureless and sometimes graded intervals, sharp-based sandstone beds may also include units of parallel lamination or ripple cross lamination. When these are all present, they tend to occur in a particular vertical order (Figs 6.94, 6.95), 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 is often silty and the lamination is rather diffuse. The final interval, the E division, is fine-grain mudstone or siltstone, which may be difficult to separate from the fine-grain 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 interval is missing. Only the vertical order remains constant. The relative thickness of the intervals also varies. Some beds are dominated by the laminated divisions; others consist almost entirely of the A division; and may have only a thin capping of interval B or C. Very thick beds, dominated by the A division, may show signs of rapid de-watering in the form of water-escape structures (§9.2.2). Beds that end with interval C commonly preserve ripple morphology on their top surfaces, draped by the fine-grain 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 (§9.2.2) to occur within the C division. These sharp-based beds often have sole marks on their lower surfaces, in which case it can be valuable to compare palaeocurrents from such marks with those derived from primary current lineation and ripple cross lamination in the B and C divisions.

6.8.5 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 intervals of deposition from suspension in quiet conditions. As discussed in §6.7, 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

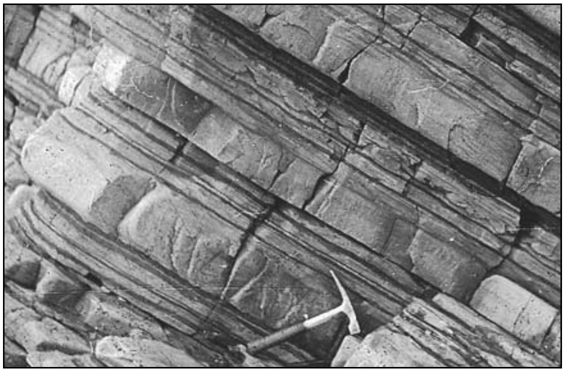
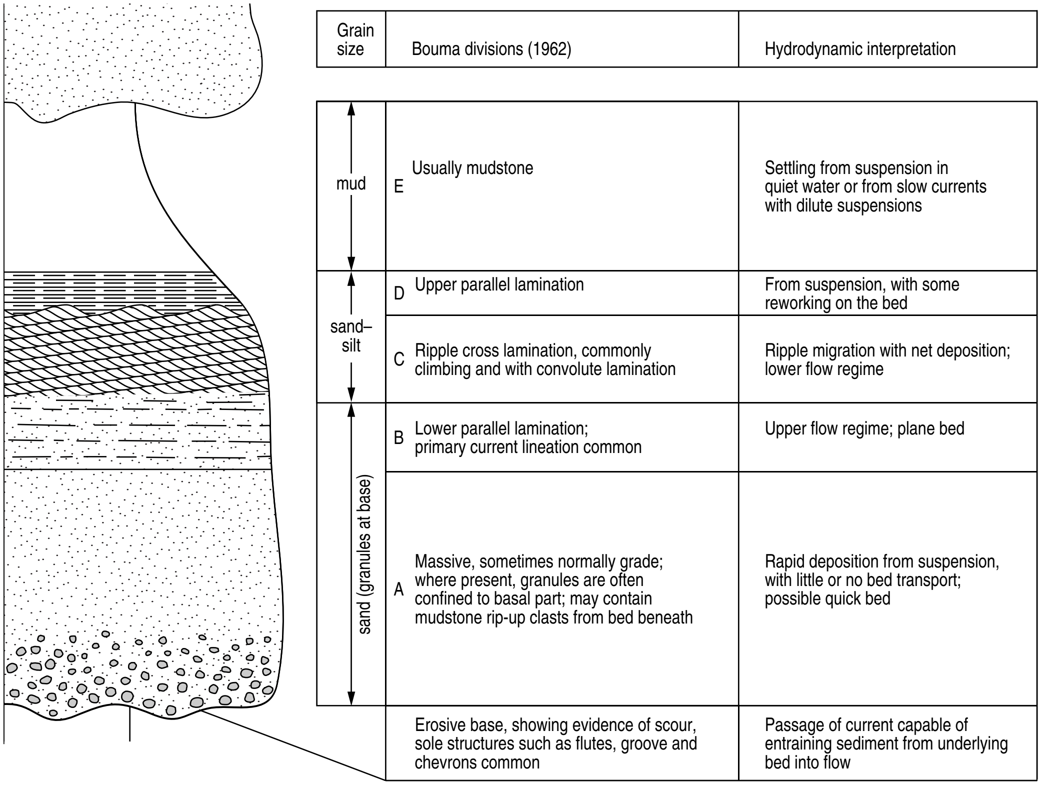


Figure 6.94 Cyclical repetition of sand-dominated beds. Each cycle is interpreted as a turbidite unit – the product of a single turbidity-flow event. Aberystwyth Grits, Newquay, Wales. (Photo courtesy of Gilbert Kelling)



Grain size		Bouma divisions (1962)	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
	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
	A	Massive, sometimes normally grade; 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, groove and chevrons common	Passage of current capable of entraining sediment from underlying bed into flow

Figure 6.95 The Bouma sequence of internal sedimentary structures that occur in sandstone beds generated by sudden decelerating unidirectional currents. In most examples of such beds, one or more of the divisions may be missing.

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 intergranular collision in a basal traction-carpet layer where conditions like those of a grainflow prevail, driven by the shear of a powerful overriding current. 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 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-grain sandstones, is that grain sizes become segregated horizontally in the active flow as coarser grains are transported more slowly. During deposition, the finer-grain 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, and 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 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 short-lived phases of bedload transport (Fig. 6.96).

The nature of the decelerating current can be deduced only from the context of the sediments. The Bouma sequence was first described from deepwater

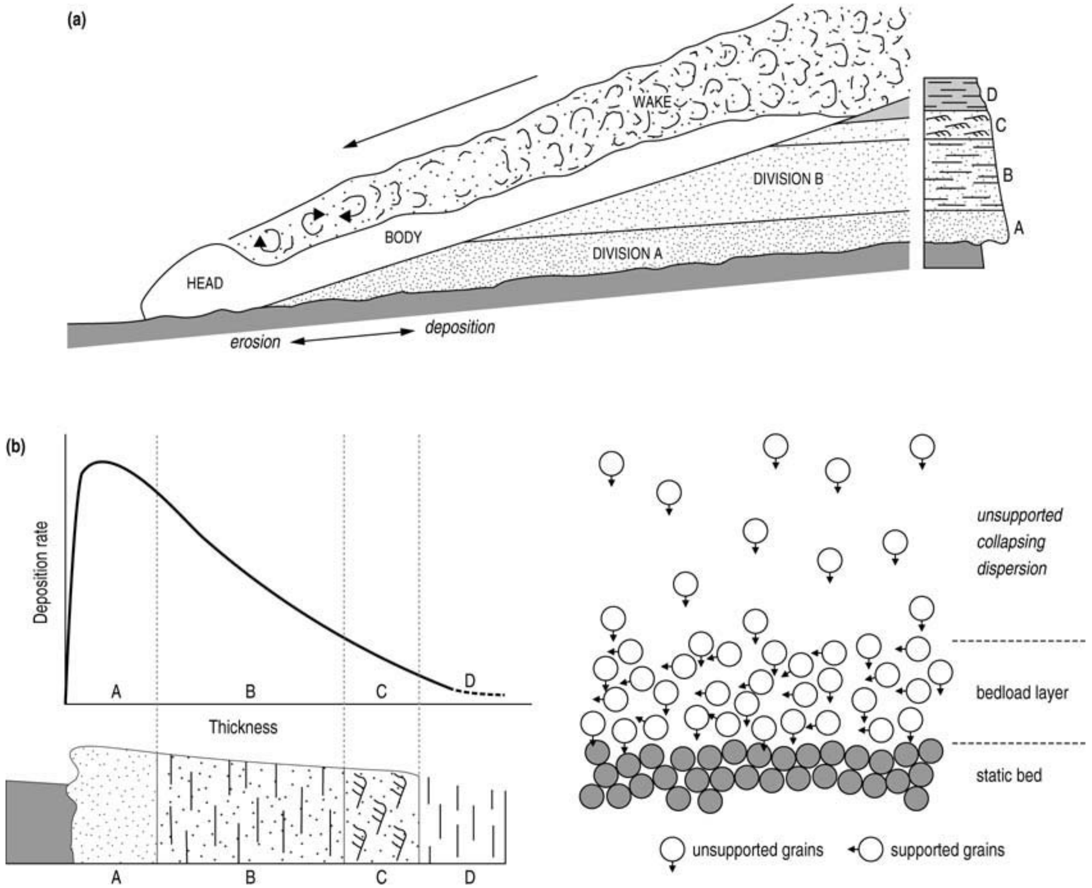

(a)
WAKE
D
C
B
A
DIVISION B
BODY
DIVISION A
HEAD
erosion
deposition
(b)
Deposition rate
A
B
C
D
Thickness
A
B
C
D
unsupported
collapsing
dispersion
bedload layer
static bed
unsupported grains
supported grains

Figure 6.96 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)

sediments, where the currents were interpreted as turbidity currents (§3.7.2), but the sequence is in no way exclusive to or diagnostic of such currents. Similar beds can occur as a result of episodic decelerating flows in many settings.

6.8.6 Uses of grading and the Bouma sequence

As well as containing valuable information about depositional processes, graded beds are one of the best way-up indicators. The relative rarity of inverse grading in sandstones makes the use of grading relatively 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 an episodic decelerating current. Both sole marks on the bases of beds and also structures in the laminated intervals of the Bouma sequence provide valuable palaeocurrent data.

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. Useful observations can additionally be made in driven snow.

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 be seen only 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 plan-form 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 1 L 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.

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CHAPTER 7

Depositional structures in gravels, conglomerates and breccias

7.1 Introduction

The general name for sediments containing a significant proportion of gravel 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 and volcanic slopes, 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 else 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 revealed by trenches. Such methods are usually applicable only in accessible sub-aerially 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 and breccias is yet another area of geology in which detailed observation and interpretation of ancient deposits can aid the better understanding of present-day processes, particularly in deepwater settings. The careful analyses that have allowed these advances have involved 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 Problems of classification

7.2.1 Defining rudites

There is no universal agreement on the percentage of clasts above 2 mm (–1φ) that need to 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 of clasts larger than 2 mm before the terms **gravel**, **conglomerate** and **breccia** are used. In the field or in the laboratory, first try to estimate the percentage of gravel, sand and mud present, and refer the sediment to the classes shown in Figure 7.2. 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 sub-aqueous or sub-aerial 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.

7.2.2 Defining a sedimentary "structure" in rudites

The term sedimentary "structure" is here interpreted broadly to include several mass properties that include textural features:

- features based on composition
- features such as shape, roundness and surface morphology of the constituent clasts

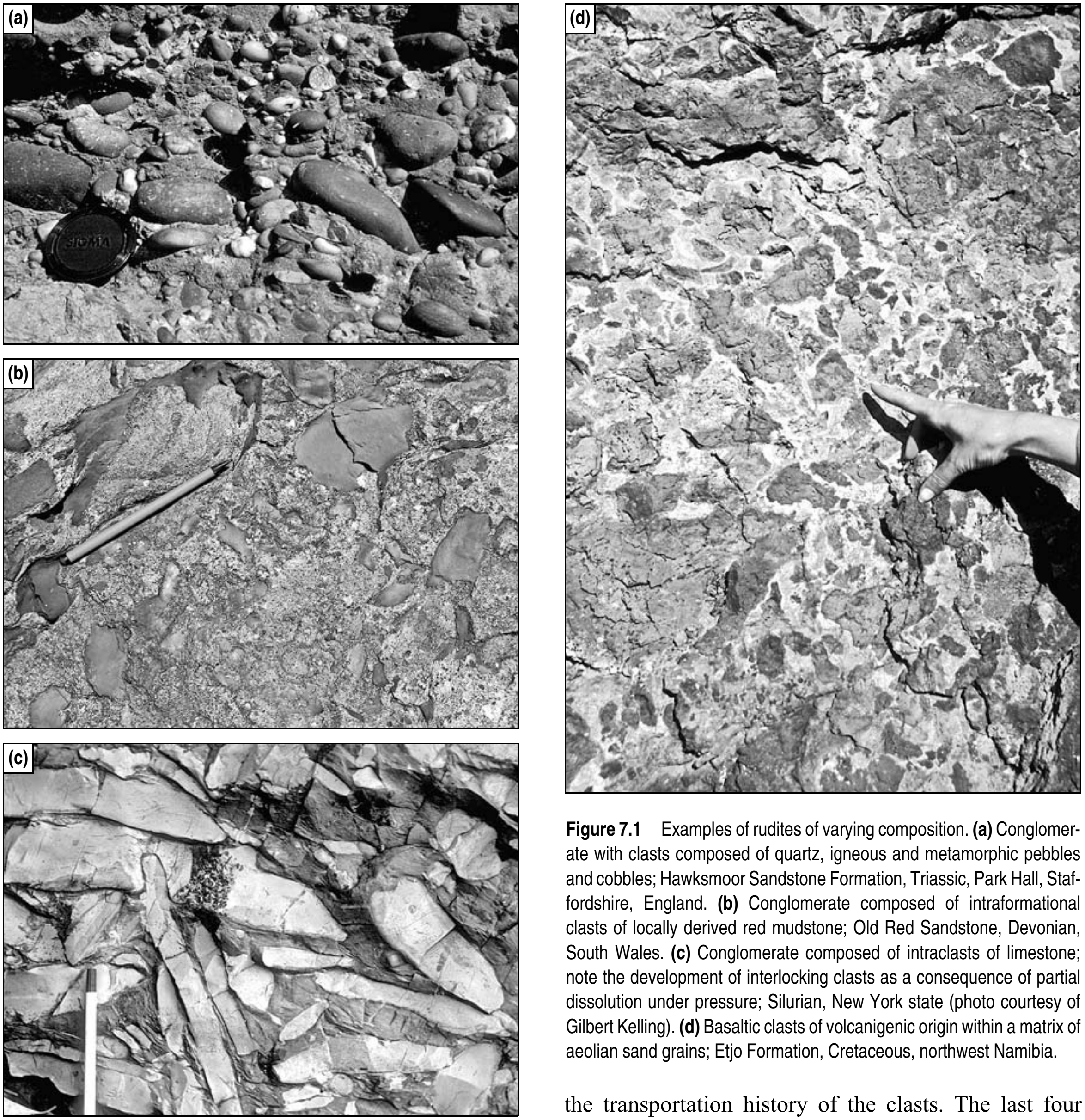


Figure 7.1 Examples of rudites of varying composition. **(a)** Conglomerate with clasts composed of quartz, igneous and metamorphic pebbles and cobbles; Hawksmoor Sandstone Formation, Triassic, Park Hall, Staffordshire, England. **(b)** Conglomerate composed of intraformational clasts of locally derived 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 (photo courtesy of Gilbert Kelling). **(d)** Basaltic clasts of volcanigenic origin within a matrix of aeolian sand grains; Etjo Formation, Cretaceous, northwest Namibia.

- stratification and cross stratification
- features based on grain-size distribution, sorting and clast-support systems
- features based on fabric, packing and porosity
- the presence and type of graded bedding.

The first two properties should be recorded in any preliminary survey and 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.

7.2.3 Composition and classification of rudites

One of the first properties to be recorded in the field is the composition of the larger clasts (Fig. 7.1). This allows useful preliminary conjectures concerning their possible provenance and their processes of origin, for example whether the rocks may be pyroclastic. Further

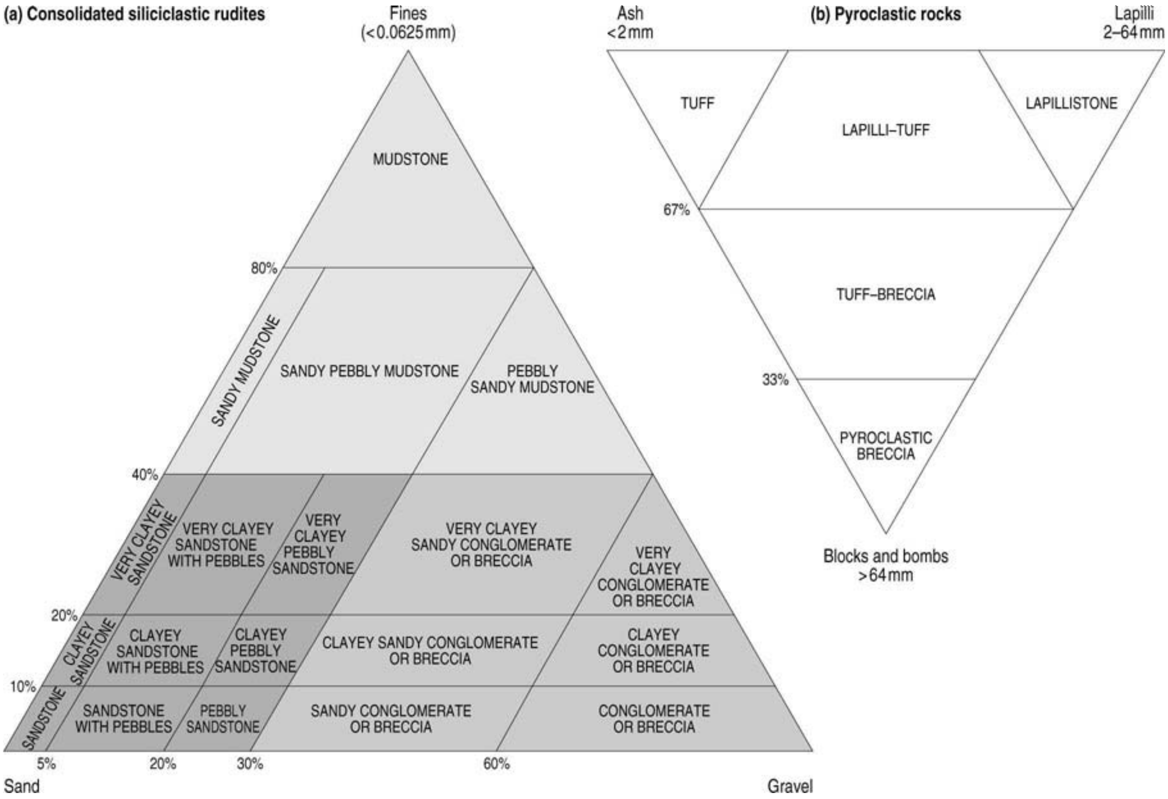

(a) Consolidated siliciclastic rudites
Fines (<0.0625 mm)
MUDSTONE
80%
SANDY MUDSTONE
SANDY PEBBLY MUDSTONE
PEBBLY SANDY MUDSTONE
40%
VERY CLAYEY SANDSTONE
VERY CLAYEY SANDSTONE WITH PEBBLES
VERY CLAYEY PEBBLY SANDSTONE
VERY CLAYEY SANDY CONGLOMERATE OR BRECCIA
VERY CLAYEY CONGLOMERATE OR BRECCIA
20%
CLAYEY SANDSTONE
CLAYEY SANDSTONE WITH PEBBLES
CLAYEY PEBBLY SANDSTONE
CLAYEY SANDY CONGLOMERATE OR BRECCIA
CLAYEY CONGLOMERATE OR BRECCIA
10%
SANDSTONE
SANDSTONE WITH PEBBLES
PEBBLY SANDSTONE
SANDY CONGLOMERATE OR BRECCIA
CONGLOMERATE OR BRECCIA
5%
20%
30%
60%
Sand
Gravel
Ash <2 mm
(b) Pyroclastic rocks
Lapilli 2–64 mm
TUFF
LAPILLI–TUFF
LAPILLISTONE
67%
TUFF–BRECCIA
33%
PYROCLASTIC BRECCIA
Blocks and bombs >64 mm

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 & 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; precise percentage boundaries for mixtures vary depending on the prejudices of the user.

observations relate to whether the materials originate from within the basin of deposition and are eroded from recently deposited sediment (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 the structures of rudites

Some rocks provisionally classified as rudites will, with further analysis and experience, be regarded as "pseudo-conglomerates" or "pseudo-breccias". These may have originated 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 undergoing spheroidal weathering are often 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, ripples will not form in clasts whose diameters approach the size of the ripples themselves. However, 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, resulting from, say, 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, as a general rule, 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 bedload 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 as a consequence of 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 often split the stream on several scales. Deposition takes place on the flat bar tops, where pebbles lodge in an imbricate packing (see §7.4.4), and at the downstream 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 lensoid. Deeper channel areas often have beds covered with the coarsest gravel, but this may be masked when the bed is sub-aerially exposed by finer sediment laid down during falling river level.

7.3.3 Fans and steep slopes

Episodic floods on alluvial fans are often characterized by a rapid rise to peak discharge, followed by a rapid waning to a reduced level. 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 fine-grain sediment gives the flow a high viscosity and high density (see §3.3) that enables large clasts to "float" in a finer matrix, and results in the generation of matrix-supported deposits. Gravel lobes are produced by a sieving process when the water transporting the gravel sinks into a permeable substrate and so leaves clast-supported gravel on the surface.

On very steep slopes, both sub-aqueous and sub-aerial, 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. Such deposits have a low preservation potential in sub-aerial settings.

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-shape gravel-rich fan delta may be formed. Because gravel is not readily redistributed by waves and currents, a steep delta slope often forms 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 an important component of many beaches, particularly storm beaches. These gravel ridges are broadly linear, but they sometimes migrate 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 show subparallel inclination if they are flattened (i.e. imbrication: see §7.4.4).

7.3.6 Deepwater fans

At the foot of major submarine slopes, turbidity currents and other sediment gravity flows may deposit gravels, particularly close to the mouths of submarine canyons, 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 reefal material, become detached, fall down slope and accumulate in steeply inclined wedges (up to 40°) comprising interlocking frameworks with high initial porosities (up to 40%). Such 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-grain material falls from floating icebergs and shelf ice to give isolated **dropstones**. Material deposited both sub-aerially and sub-aqueously from melting ice is commonly subjected to further movement and deposition by mudflows. Meltwater flowing within, above, below or laterally to a glacier may sort, transport and deposit sand and gravel bodies to form geomorphological 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 also 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; and **pyroclastic** debris is produced by explosion of magma and country rock. Pyroclastic processes operate in both sub-aqueous and sub-aerial settings. Pyroclastic airfall deposits result primarily from the settling from suspension of fine volcanic ash, but are often interbedded with larger, often randomly distributed, dropstones of lapilli, blocks or bombs (or both), which deform underlying strata. Such deposits often 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 enough energy to rise up over topographical obstacles. Such surge deposits may be identified where they can be shown to thicken into topographical depressions and thin onto topographical rises.

Many primary pyroclastic fragments become mixed during transport with “normal” siliciclastic and carbonate deposits, and are resedimented as **epiclastic** conglomerates.

7.4 Structures and other descriptive features: 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 with further observations of sediment samples in the laboratory. It is useful to note:

- the composition, colour, shape and surface features of the clasts that make up the sediment
- the sorting, grain-size distribution, porosity and clast-support characteristics of the sediment
- the nature of any grading within beds
- the presence of any preferred clast orientation (fabric)
- the nature of any stratification or cross stratification (or both) present
- bed thickness and grain-size relationships.

7.4.1 Shape (roundness and sphericity) and surface features of the clasts

Clast shape within rudites is extremely variable (Fig. 7.3). With experience, sphericity and roundness can be judged in the field by comparison with 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. 7.4). 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 Bibliography). Experience of measuring and handling individual

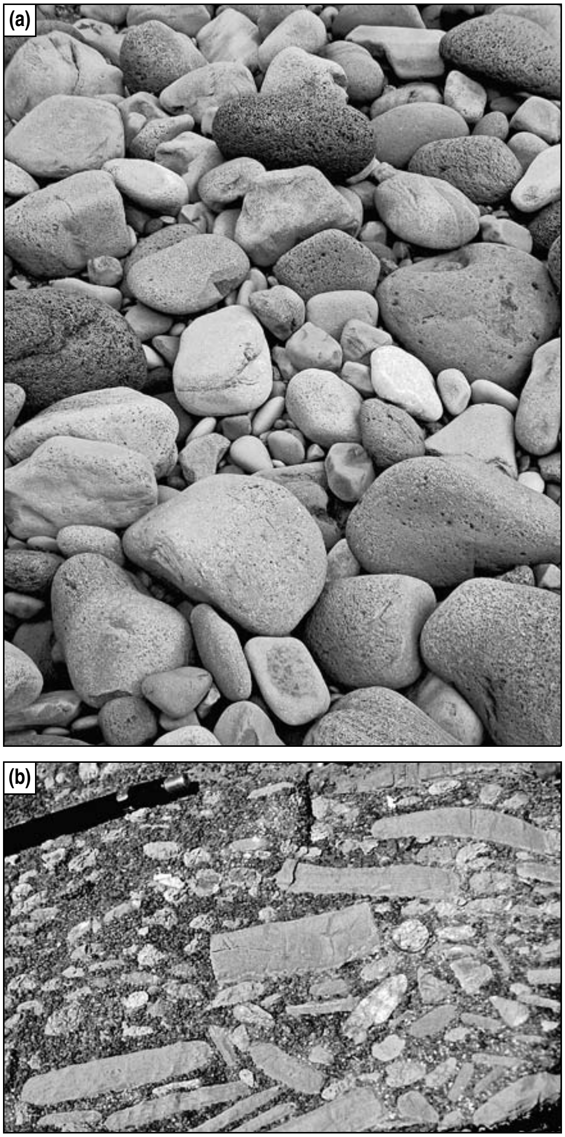


Figure 7.3 Examples of varying clast shape: **(a)** Sub-rounded, moderately spherical, equant cobbles and boulders; modern beach, southern Iceland. **(b)** Extremely angular, low-sphericity, prolate (rod-shape) clasts; Huab Formation, Permian, northwest Namibia.

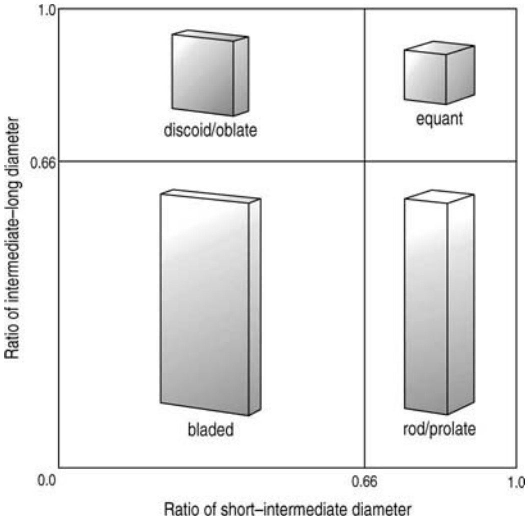

1.0
0.66
0.0
0.66
1.0
Ratio of intermediate–long diameter
Ratio of short–intermediate diameter
discoid/oblate
equant
bladed
rod/prolate

Figure 7.4 Scheme for the classification of clast shape. Equant- and disc-shape clasts are most common. (Modified after Tucker 1991)

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).

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** (see Fig. 4.21b); they represent wind-fashioned sand-blasted objects. In most cases ventifacts will be rolled and reworked into later deposits, but occasionally 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-shape pebbles generally become more common as one goes from rivers to low-energy then high-energy beaches. This is because of increased rates of abrasion by sliding. Highly rounded spheroidal pebbles arise through constant abrasion, attrition and reworking, sphericity falling and roundness increasing from fluvial to low- then high-energy beach environments. The indices of sphericity and roundness provide measures of the textural maturity of a sediment. Note, however, that it is more difficult, given constant conditions, for sand or small pebbles to become rounded than it is for large clasts. Percussion rings from high-velocity collisions are apparent on the surfaces of some clasts. These 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 method similar to that for sands. However, when ancient rudites are well cemented, the task is impossible. In such circumstances it is nevertheless important to estimate and describe these characteristics in an approximate way.

Initially, determine the degree of sorting of individual beds and record whether there are clear grain-size modes. Very commonly there is a mode in the pebble–cobble size and another in the sand size; hence the sediment is **bimodal**, the term **clasts** being applied to the coarse mode and **matrix** to the fine. Some deposits are **unimodal** and well sorted, lacking a well defined matrix and hence having very high porosity (>40%) and permeability. Other beds are unimodal and poorly sorted, and there are others that are **polymodal**.

Next, determine whether the deposit is clast or matrix supported. Are larger clasts in contact with each other, forming a framework, or are they dispersed in a finer-grain matrix (Fig. 7.5)? 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 pittings) might help. Matrix-free clast-supported framework conglomerates commonly have a high porosity and permeability, and the term **openwork** is used to describe them. Alternatively, if the clasts are dispersed and supported by the matrix, try to establish whether the supporting matrix is of sand (e.g.

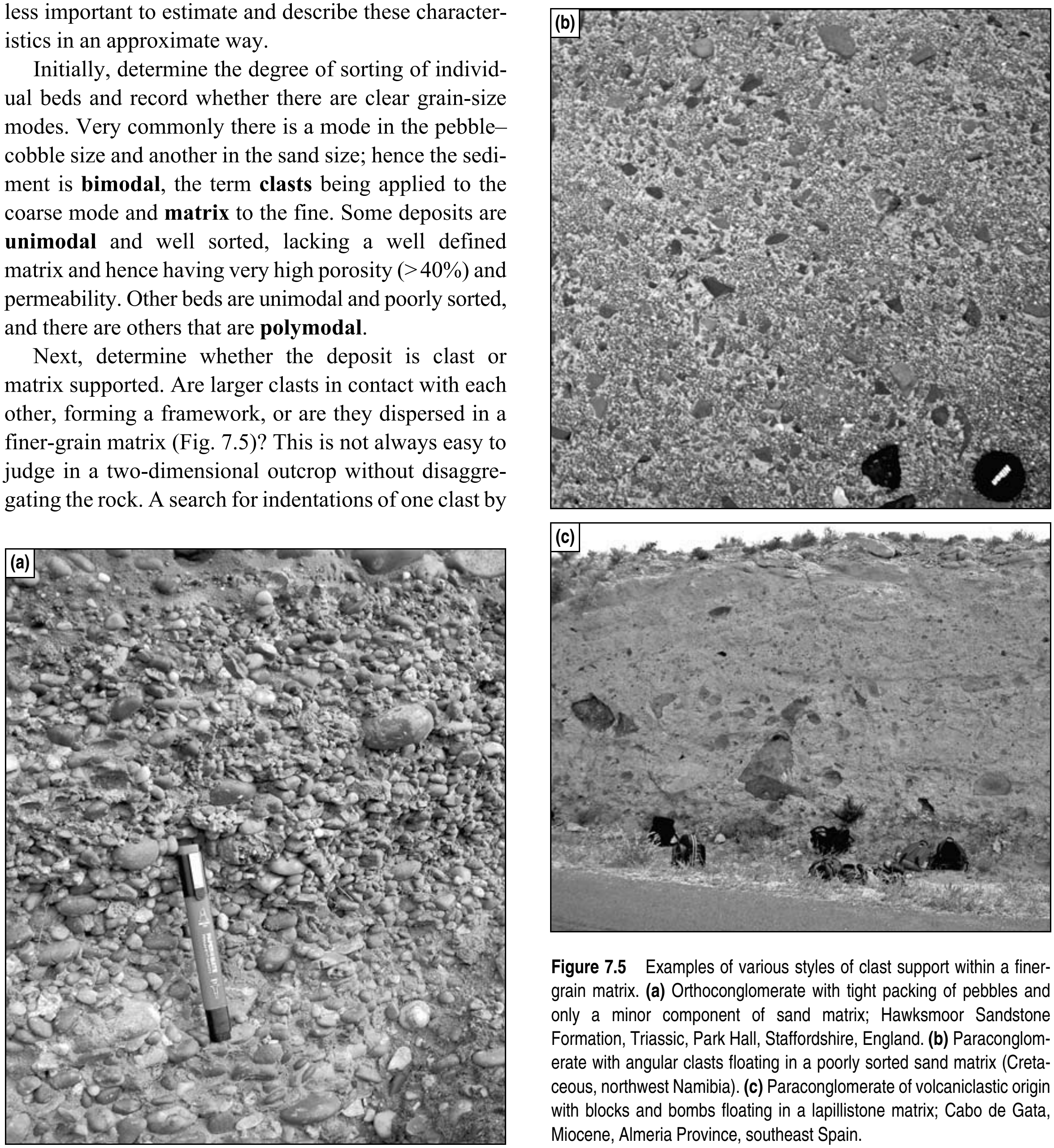


Figure 7.5 Examples of various styles of clast support within a finer-grain matrix. **(a)** Orthoconglomerate with tight packing of pebbles and only a minor component of sand matrix; Hawksmoor Sandstone Formation, Triassic, Park Hall, Staffordshire, England. **(b)** Paraconglomerate with angular clasts floating in a poorly sorted sand matrix (Cretaceous, northwest Namibia). **(c)** Paraconglomerate of volcaniclastic origin with blocks and bombs floating in a lapillistone matrix; Cabo de Gata, Miocene, Almeria Province, southeast Spain.

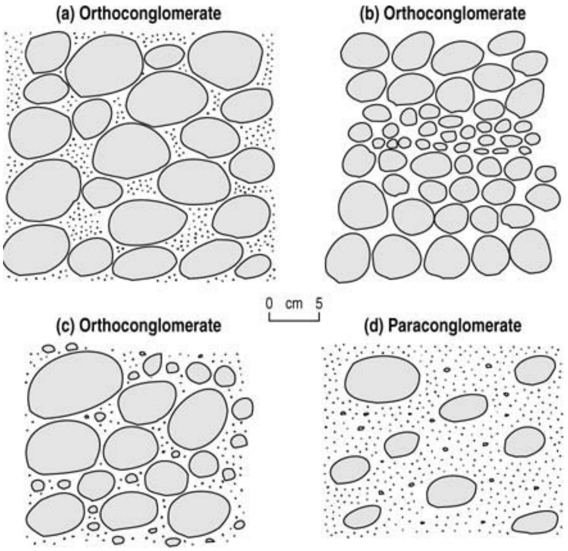


Figure 7.6 Descriptive features of sorting and size distribution in rudites. There is a spectrum in nature: **(a)** bimodal clast-supported framework, well sorted matrix; **(b)** clast-supported open-work framework; **(c)** polymodal clast-supported framework, poorly sorted matrix; **(d)** polymodal, matrix-supported, lacking a framework (paraconglomerate). (Modified after Walker in Harms et al. 1975)

in a sandy conglomerate), whence the rock may be very porous and permeable, or of mud (e.g. in a pebbly mudstone), whence the rock may lack significant pore space and have a low permeability. Rudites with a matrix dominated by mud are commonly glacial till deposits, with clasts that tend to be very poorly sorted and may possibly 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.

The distinctions made in the past two paragraphs provide the basis of a classification of rudites into the **orthoconglomerates** and the **paraconglomerates** (Fig. 7.6). In nature, conglomerates appear to range from orthoconglomerates, with well sorted, clast-supported, openwork or matrix-filled frameworks, to paraconglomerates, which are poorly sorted, matrix-supported and with very variable size distributions.

Following these considerations, further questions can be asked to help to generate ideas about modes of transport and depositional processes:

- If the bed is clast supported and has a matrix, was the finer-grain interstitial material deposited along with the larger clasts or did it fill the framework at a later time?
- If the bed is clast supported but retains an openwork structure, did the current winnow away potential matrix-grade sediment or did it maintain it in suspension while rolling and sorting larger clasts on the bed, or was some combination of these processes active?
- What is indicated about the abundance of fine-grain 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 7.7 relates the sizes of clasts transported by rolling to the size of sand suspended by the same flow. When the bed has a pebble

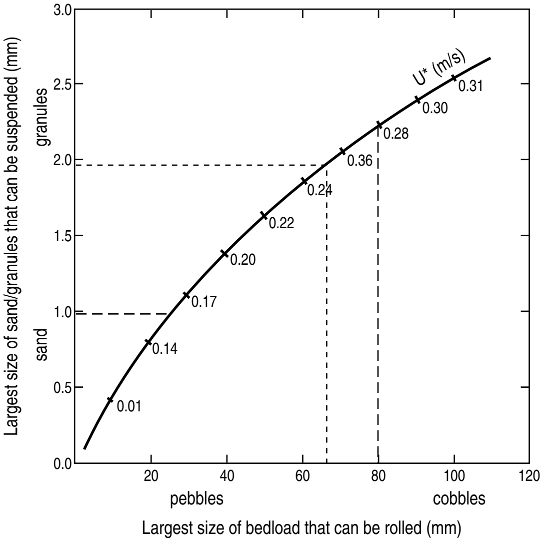


Figure 7.7 Graphical representation of size of clasts transported by rolling, compared with the size of grains suspended by the same flow. Numbers on the curve refer to critical values of shear velocity *U** for bed rolling. The dotted line shows that the coexistence in a conglomerate of pebbles of 63 mm and sand of 1.95 mm diameter relates to slight velocity fluctuations around a mean value of $U^* = 0.25\,\mathrm{m\,s^{-1}}$. The dashed lines show that the coexistence of clasts in a bimodal, matrix-filled framework orthoconglomerate composed of cobbles of 80 mm and sand of 1 mm would probably result from the deposition from bedload of cobbles at high shear velocity ($0.28\,\mathrm{m\,s^{-1}}$) while 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 in Harms et al. 1975)

framework filled with sand, however, these graphical relationships do not suggest whether the infilling was mainly contemporaneous with the deposition of the gravel, as when part of a heavy suspended load is trapped in the quieter interstices of the gravel bed, or if it mainly postdated the deposition of the gravel, as when a river previously in flood, and carrying relatively little sand in suspension, resumes a more leisurely flow with sand transported as bedload. In many cases it is difficult to judge both the timing and the process of emplacement of the matrix. Windblown sand, for example, could fill a water-lain but sub-aerially emergent gravel framework without leaving much evidence of the nature of the process.

If the rudite is matrix-supported, consider whether the large clasts were transported along with the finer-grain matrix as a high-viscosity flow or whether they were dropped into already deposited finer-grain sediment. The occurrence of disturbed laminae in the matrix may help to decide the case. If the clasts are transported with the matrix (e.g. in a debris flow), there may be no internal indication of the environment of the flow. Alluvial fans, deepwater 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 disturb the underlying sediment greatly when they settle. Only information on the context of the rudites derived from surrounding sediments will help resolve such problems of interpretation.

7.4.3 Grading within beds

The main styles of grading described in §6.8 for sands apply equally well in gravels and conglomerates (Figs 7.8, 7.9). Content or distribution grading and coarse-tail grading are both common in normally graded rudites. Inverse grading is much more common in rudites than

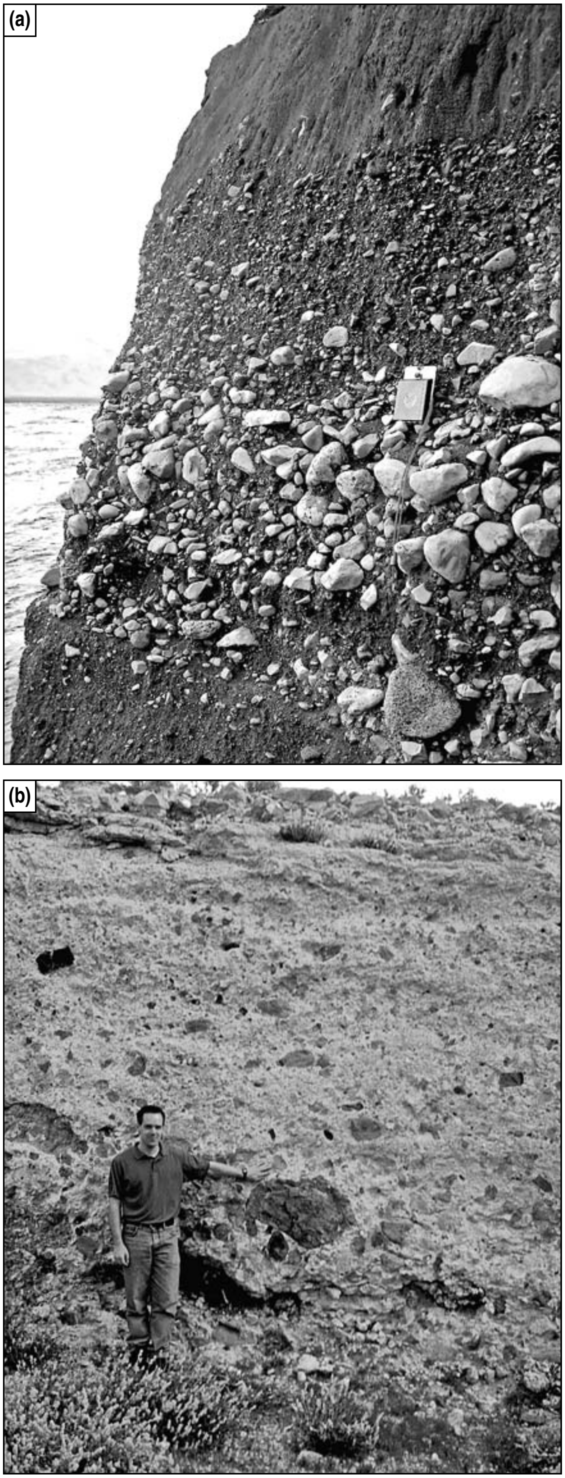


Figure 7.8 Examples of normally graded beds in rudites. **(a)** Fluvial braidplain gravels, modern, southern Iceland. **(b)** Pyroclastic flow deposit with blocks and bombs; Cabo de Gata, Miocene, Almeria Province, southeast Spain.

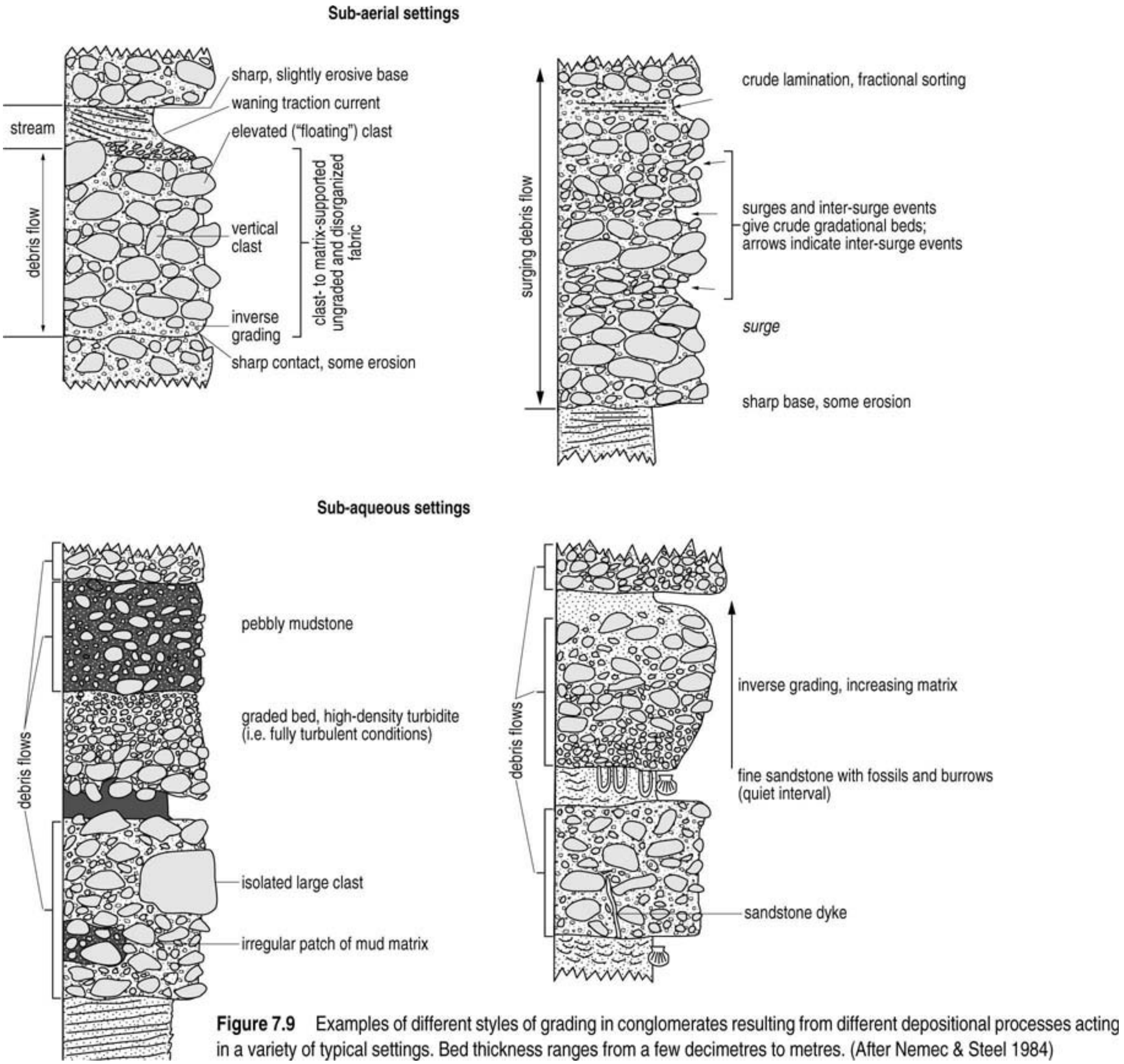

Sub-aerial settings
sharp, slightly erosive base
waning traction current
stream
elevated ("floating") clast
debris flow
vertical clast
clast- to matrix-supported ungraded and disorganized fabric
inverse grading
sharp contact, some erosion
surging debris flow
crude lamination, fractional sorting
surges and inter-surge events give crude gradational beds; arrows indicate inter-surge events
surge
sharp base, some erosion
Sub-aqueous settings
debris flows
pebbly mudstone
graded bed, high-density turbidite (i.e. fully turbulent conditions)
isolated large clast
irregular patch of mud matrix
debris flows
inverse grading, increasing matrix
fine sandstone with fossils and burrows (quiet interval)
sandstone dyke

Figure 7.9 Examples of different styles of grading in conglomerates resulting from different depositional processes acting in a variety of typical settings. Bed thickness ranges from a few decimetres to metres. (After Nemec & Steel 1984)

in sands. This can be explained by large clasts rising upwards because of dispersive pressures or by progressive loss of larger clasts from the lower, more strongly sheared part of the flow. It may also be a product of along-flow sorting, whereby finer particles arrive first at the depositional site. Ungraded beds usually indicate high shear strength or high viscosity, thereby preventing turbulence and effective grain interaction. Normal grading usually suggests that turbulence was developed during deposition (see §3.7). **Lateral grading** of grain size is sometimes noted, especially in orthoconglomerates. In some cases, especially in volcanic rocks, **density grading** occurs, whereby clasts of different composition (and density) become vertically differentiated. This should be considered separately from size grading, especially where pumice or other vesicular clasts are involved.

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

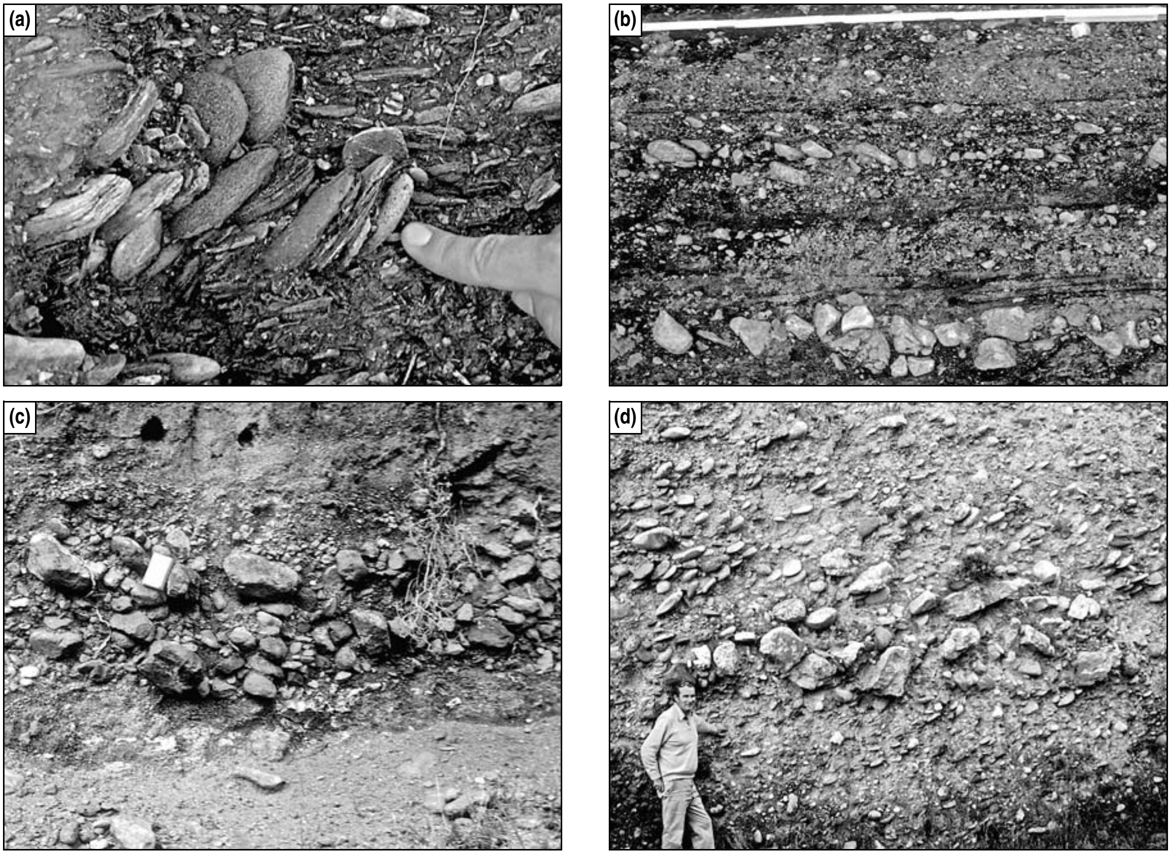


Figure 7.10 Examples of imbricated clast fabrics developed in rudites. **(a)** Interlocking disc-shape clasts; transport from left to right; modern alluvial fan, Tabernas, southeast Spain. **(b)** Imbrication indicating transport from right to left; Permian, east Greenland. **(c)** Imbrication indicating transport from right to left; Gregory Rift, Kenya. **(d)** Imbrication indicating transport from left to right; Siensas, Spanish Pyrenees.

long axes may occur; in others the short axes may be aligned; and in others no pattern may be discerned (Figs 7.10, 7.11).

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. 7.11b). In the laboratory, three-dimensional data can be plotted on a stereonet (Fig. 7.12) or orientations measured in two dimensions can be plotted as rose diagrams (Fig. 7.11b). See also Appendix 1.

The most commonly occurring fabric, which arises when the rudite is rich in disc- and blade-shape clasts, is **imbrication** (Figs 7.10, 7.13). Here the flattened surfaces dip in an upstream direction. This is common in horizontally bedded, clast-supported conglomerates, and two variants may occur. In one, the *a*-axis is transverse to the dip direction of the clasts and the *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 orientated parallel to the face and therefore dip down stream, opposite to that of genuine imbrication. Other clasts roll down slope on slipfaces 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

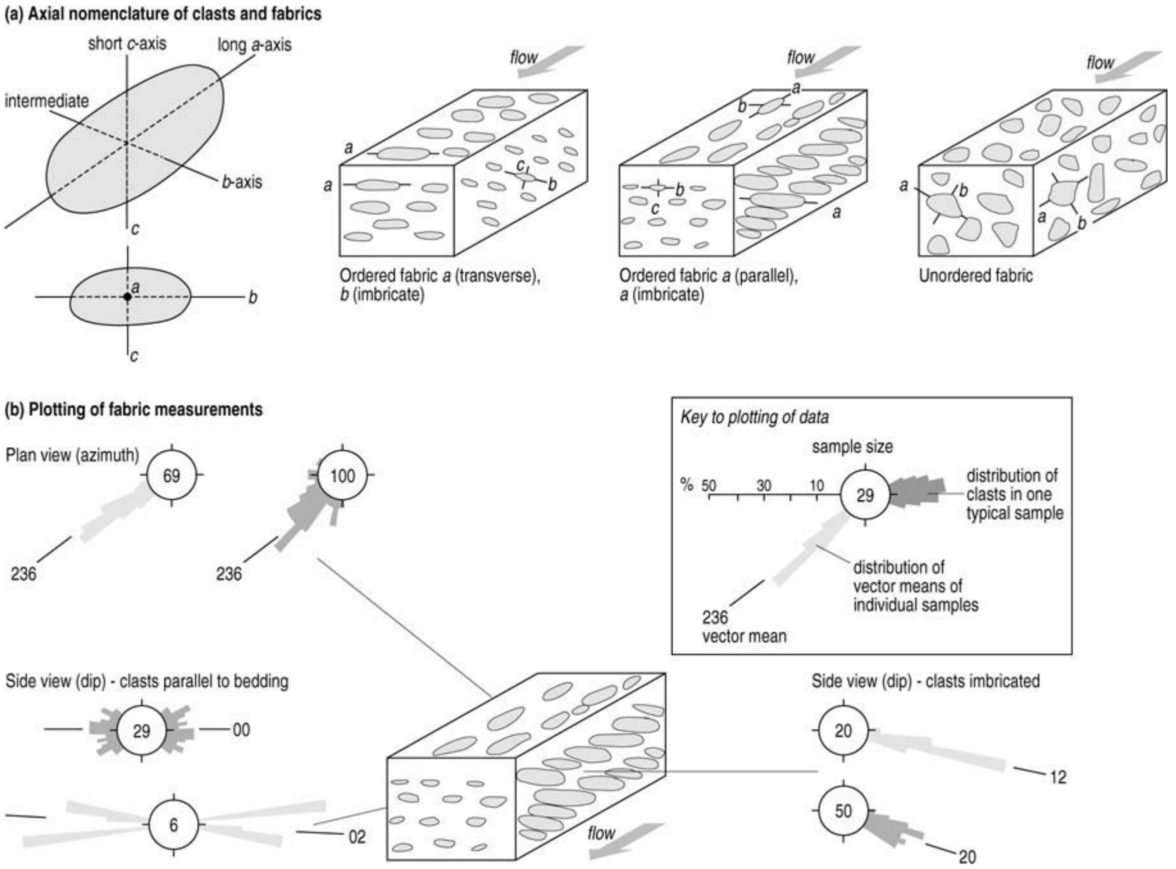

(a) Axial nomenclature of clasts and fabrics
short c-axis
long a-axis
intermediate
b-axis
c
a
b
c
flow
a
a
c
b
Ordered fabric a (transverse), b (imbricate)
flow
a
b
b
c
a
Ordered fabric a (parallel), a (imbricate)
flow
a
b
a
b
Unordered fabric
(b) Plotting of fabric measurements
Plan view (azimuth)
69
236
100
236
Key to plotting of data
sample size
% 50 30 10
29
distribution of clasts in one typical sample
distribution of vector means of individual samples
236 vector mean
Side view (dip) - clasts parallel to bedding
29
00
6
02
flow
Side view (dip) - clasts imbricated
20
12
50
20

Figure 7.11 The fabric of rudites. **(a)** The nomenclature of ordered and unordered fabrics. **(b)** The plotting of fabric measurements for well cemented ordered rocks. Data were collected from orthogonal faces, not from three-dimensional measurements of clasts. Plan views of beds were measured at 69 localities, 100 clasts at each place; imbrication was measured at 20 localities on faces parallel to flow and perpendicular to bedding; there were measurements at 6 localities with facies perpendicular to flow and bedding. (After Davies & Walker 1974 and Walker in Harms et al. 1975)

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 would be re-entrained 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 particles in place, the lift force may or may not act upwards, and the drag force will try to roll the particle. The most stable position is when the forces of removal (lift and drag) are at a minimum. This occurs when the plane of flattening of the particle 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 particle 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

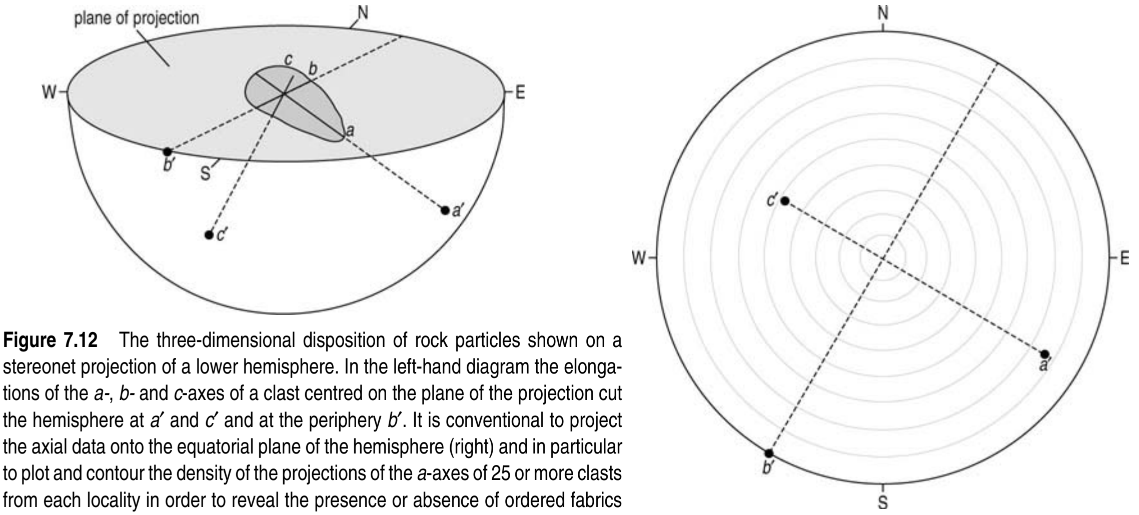

plane of projection
N
W
E
c
b
a
b′
S
a′
c′
N
c′
W
E
a′
b′
S

Figure 7.12 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) 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)

of intergranular collision, up to the point of deposition. Clasts slide because of 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. When present, imbrication commonly dips seawards on the upper and middle parts of the beach (Fig. 7.14).

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 also between different types of till (e.g. subglacial lodgement and supraglacial flow till: §7.3.8; Figs 7.15, 7.16). Although many tills show a preferred clast-long-axis orientation parallel to iceflow, the pattern is seldom well defined and few certain generalizations are possible. There will always be ambiguity in cases such as flow tills, where later mass flow has reorganized the material of the original till.

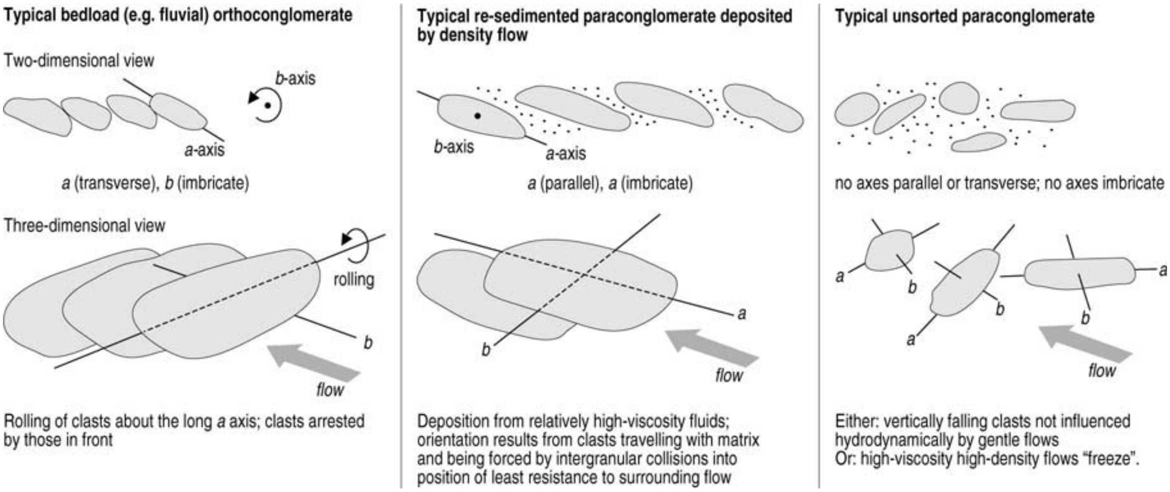

Typical bedload (e.g. fluvial) orthoconglomerate
Two-dimensional view
b-axis
a-axis
a (transverse), b (imbricate)
Three-dimensional view
rolling
b
flow
Rolling of clasts about the long a axis; clasts arrested by those in front
Typical re-sedimented paraconglomerate deposited by density flow
b-axis
a-axis
a (parallel), a (imbricate)
a
b
flow
Deposition from relatively high-viscosity fluids; orientation results from clasts travelling with matrix and being forced by intergranular collisions into position of least resistance to surrounding flow
Typical unsorted paraconglomerate
no axes parallel or transverse; no axes imbricate
a
b
b
a
a
b
flow
Either: vertically falling clasts not influenced hydrodynamically by gentle flows
Or: high-viscosity high-density flows “freeze”.

Figure 7.13 The nature and processes of origin of imbricated disc- and blade-shape clasts.

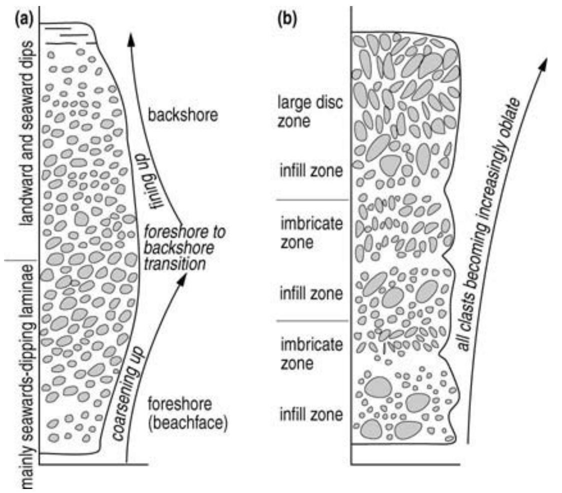

(a)
landward and seaward dips
mainly seawards-dipping laminae
backshore
fining up
foreshore to backshore transition
coarsening up
foreshore (beachface)
(b)
large disc zone
infill zone
imbricate zone
infill zone
imbricate zone
infill zone
all clasts becoming increasingly oblate

Figure 7.14 Examples of some systematic vertical changes in gravel fabric airising from the progradation of pebbly beaches. Note that 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 & Steel 1984, based on Bluck 1967 and Maejima 1982)

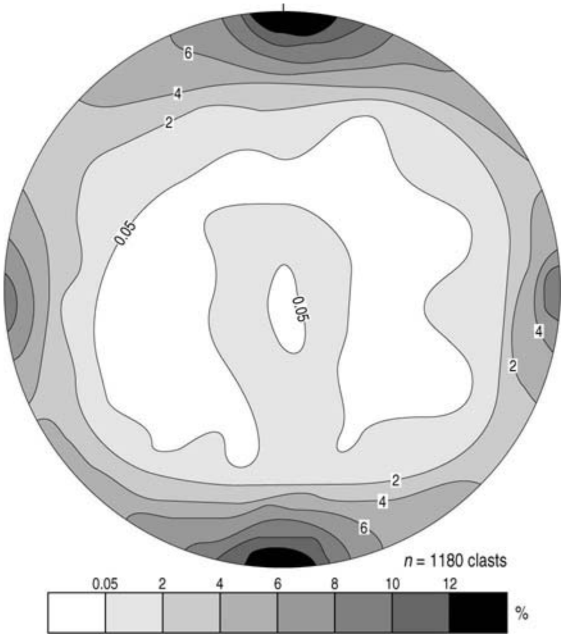


Figure 7.16 The plotting and contouring of *a*-axis fabric data (in the manner described in Fig. 7.12) for a large sample of clasts in a glacial till. Describe the pattern of organization of the till fabric; suggest the direction of flow of the clasts; suggest the type of conglomerate present.

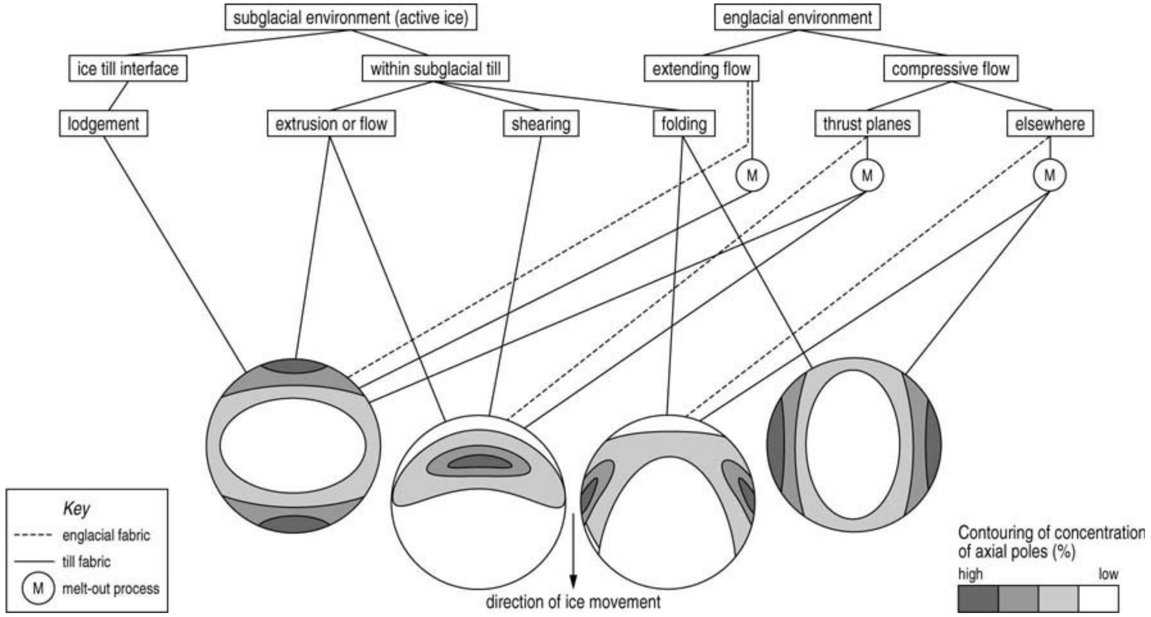

subglacial environment (active ice)
englacial environment
ice till interface
within subglacial till
extending flow
compressive flow
lodgement
extrusion or flow
shearing
folding
thrust planes
elsewhere
M
M
M
Key
englacial fabric
till fabric
M melt-out process
direction of ice movement
Contouring of concentration of axial poles (%)
high
low

Figure 7.15 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)

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, 7.17, 7.18). In such situations it may be important to measure and report bed and set thicknesses, but this is often difficult, and a rather 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 beds have a similar matrix. Sharp bed contacts, often marked by changes in grain size, may coincide with an erosion surface (Fig. 7.19), an extreme example being a basal conglomerate above an unconformity.

Multiple distinct beds with sharp contacts are the

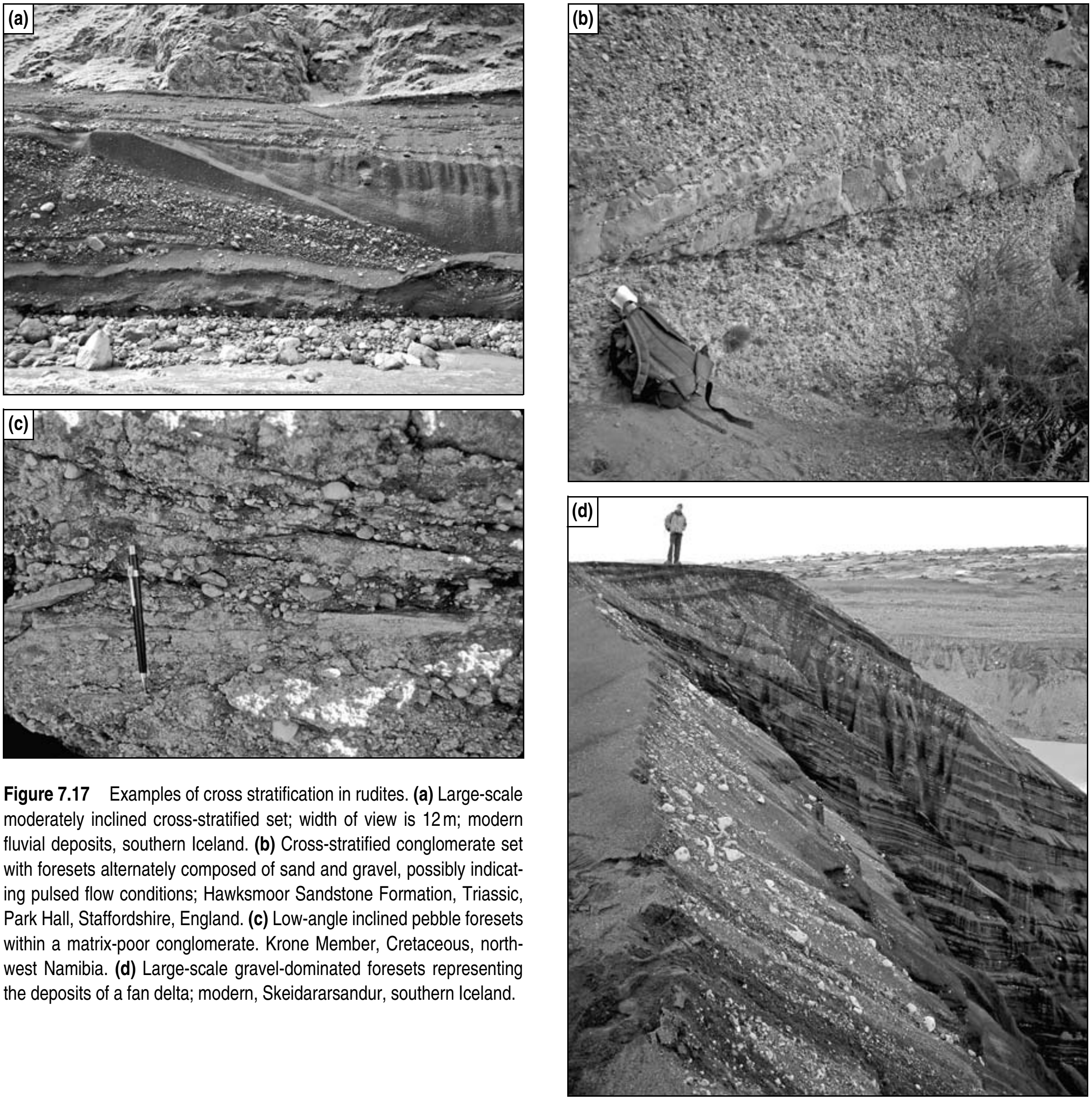
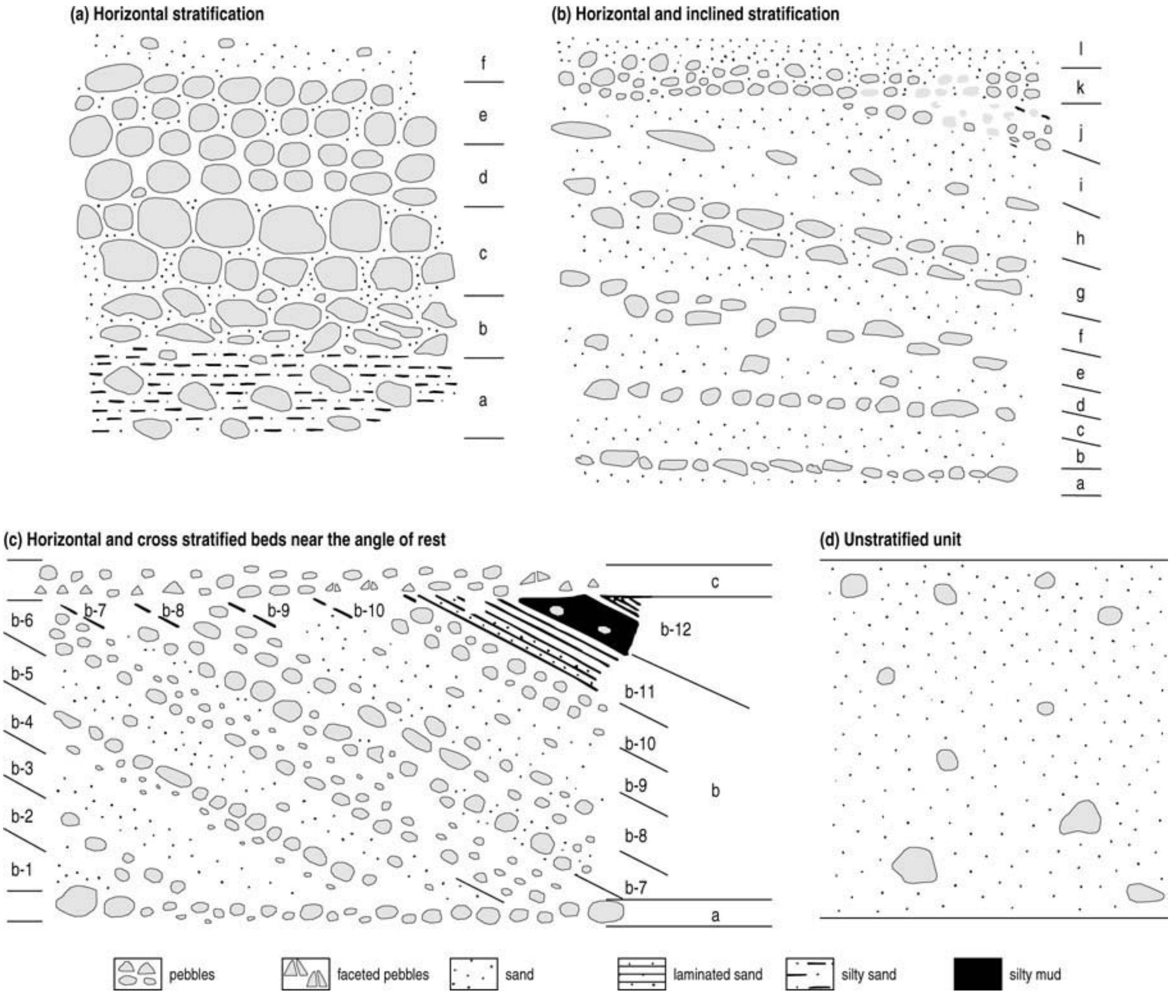


Figure 7.17 Examples of cross stratification in rudites. **(a)** Large-scale moderately inclined cross-stratified set; width of view is 12 m; modern fluvial deposits, southern Iceland. **(b)** Cross-stratified conglomerate set with foresets alternately composed of sand and gravel, possibly indicating pulsed flow conditions; Hawksmoor Sandstone Formation, Triassic, Park Hall, Staffordshire, England. **(c)** Low-angle inclined pebble foresets within a matrix-poor conglomerate. Krone Member, Cretaceous, northwest Namibia. **(d)** Large-scale gravel-dominated foresets representing the deposits of a fan delta; modern, Skeidararsandur, southern Iceland.


(a) Horizontal stratification

(b) Horizontal and inclined stratification

(c) Horizontal and cross stratified beds near the angle of rest

(d) Unstratified unit

Figure 7.18 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)** unstratified unit.

products of successive discrete depositional events. A basal conglomerate above an unconformity commonly marks the onset of deposition after a time interval dominated by tectonic uplift and erosion. Gravel above an erosion surface in a conformable sequence denotes a sharp change in energy or sub-environment, as when a river channel switches or migrates to an area of a river plain that had previously been abandoned or was inundated only at high flood. Gradational contacts suggest fluctuating and pulsating depositional processes.

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 underlying beds before they achieved significant coherence. Massive and crude bedding may involve rapidly fluctuating episodes of sedimentation, in which the sediment concentration is high, “freezing” of the load takes place, and individual depositional events are hard to distinguish.

In investigating rudites, pay particular attention to conglomerate beds that are only one clast thick. In many cases they are accumulated “lags” developed when a strong, erosive current winnowed a gravelly sand or

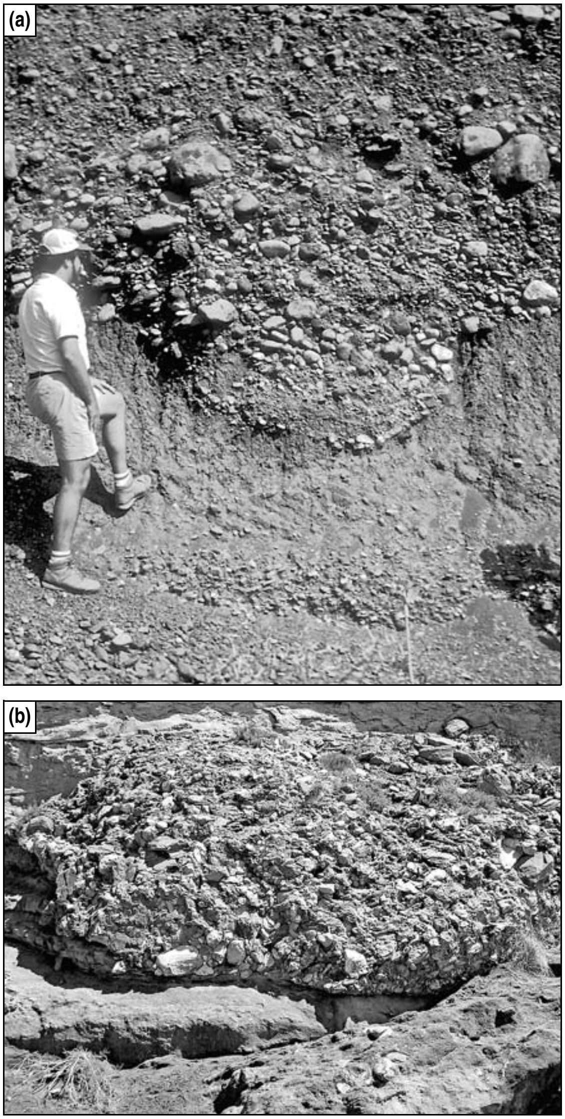


Figure 7.19 Examples of channels and scours filled with rudites: **(a)** fluvial channel, Pleistocene, Gregory Rift, Kenya; **(b)** submarine channel filled with a boulder-rich debris flow deposit, Miocene, Tabernas, southeast Spain.

shell bed and took the sand grains and hydrodynamically unstable disc-shape pebble and shells into saltation or suspension. Pebbles and larger shells remained or "lagged" behind and were concentrated as a thin layer, which may have eventually armoured the surface. A special case of lag conglomerate is that associated with channel migration (see §4.4.3; Figs 4.23, 4.24). 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*, records 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. 7.17). Alternatively, the early identification of cross bedding may help to establish the attitude of the original horizontal. Indistinct low-angle bedding (e.g. Fig. 7.17c) has been noted in beach and fluvial gravels (Fig. 7.20), as well as in deepwater conglomerates (see §7.5).

Cross stratification, with foresets at angles of 15–25°, is more common in conglomerates than in breccias (Fig. 7.17). Normally, cross-bed sets in gravel are 1–2 m thick, but isolated tabular and trough sets in excess of 10 m thick have been reported. Sets of tabular or trough type are difficult to identify in small exposures. 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 be developed in gravel (§6.2.9; Fig. 7.17d). Internally, foresets may be indistinct or spectacularly rhythmic and normally graded with coarse clasts at the base of each foreset layer (Fig. 7.17b). Short axes of clasts tend to be orientated perpendicular to the foresets, and flattened surfaces therefore dip down current. Sandy foresets are also seen in dominantly gravelly sets, but mud-draped gravel foresets indicating a temporary fall of level are rare. Reactivation surfaces (see §6.2.6) can be seen 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-grain gravel, and their three-dimensional shape is now relatively well understood. Their migration and downstream 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 the advance of slipfaces at the downstream ends of bars

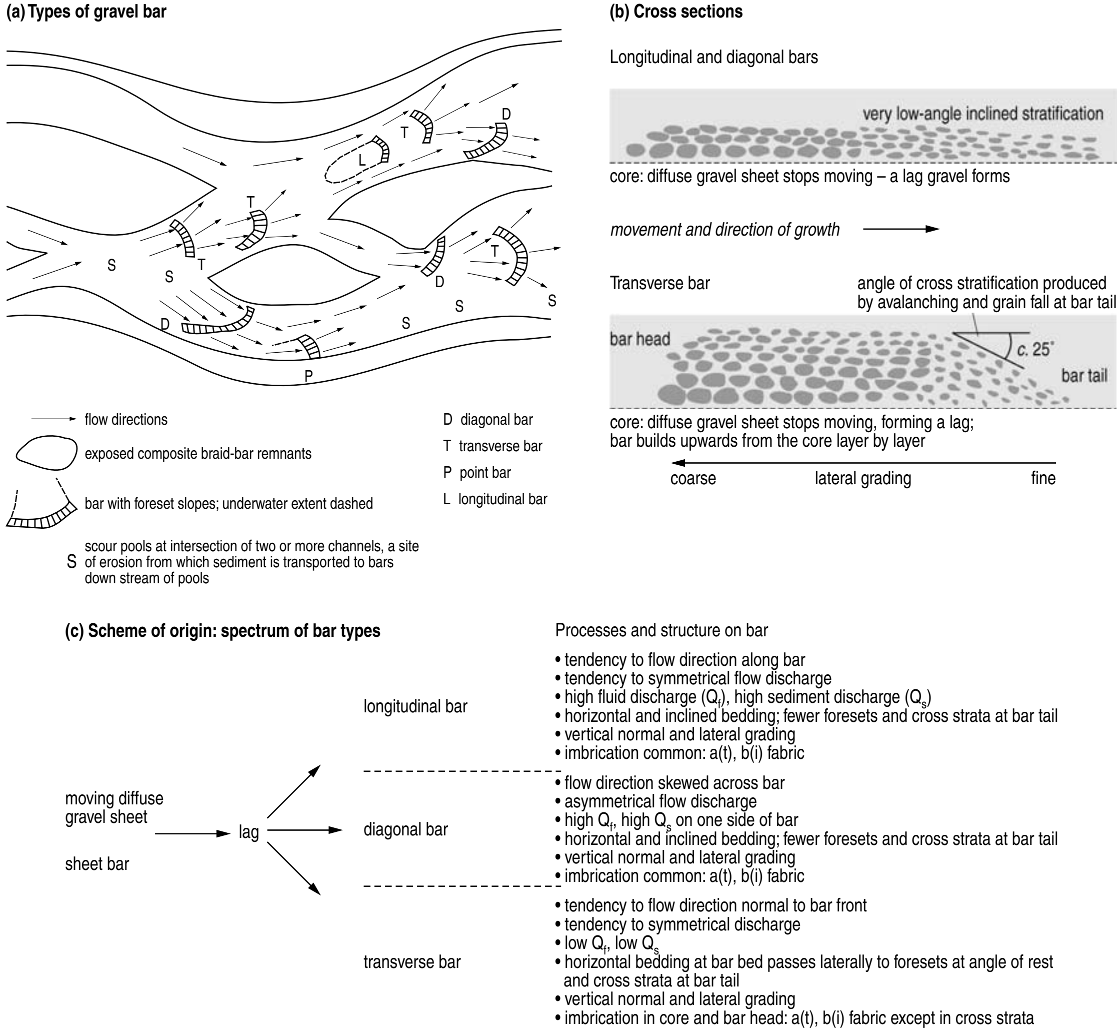

(a) Types of gravel bar
L
T
D
T
T
S
S
T
D
T
S
D
S
S
P
flow directions
exposed composite braid-bar remnants
bar with foreset slopes; underwater extent dashed
S scour pools at intersection of two or more channels, a site of erosion from which sediment is transported to bars down stream of pools
D diagonal bar
T transverse bar
P point bar
L longitudinal bar
(b) Cross sections
Longitudinal and diagonal bars
very low-angle inclined stratification
core: diffuse gravel sheet stops moving – a lag gravel forms
movement and direction of growth
Transverse bar
angle of cross stratification produced by avalanching and grain fall at bar tail
bar head
c. 25°
bar tail
core: diffuse gravel sheet stops moving, forming a lag; bar builds upwards from the core layer by layer
coarse
lateral grading
fine
(c) Scheme of origin: spectrum of bar types
moving diffuse gravel sheet
sheet bar
lag
longitudinal bar
diagonal bar
transverse bar
Processes and structure on bar
• tendency to flow direction along bar
• tendency to symmetrical flow discharge
• high fluid discharge (Q_f), high sediment discharge (Q_s)
• horizontal and inclined bedding; fewer foresets and cross strata at bar tail
• vertical normal and lateral grading
• imbrication common: a(t), b(i) fabric
• flow direction skewed across bar
• asymmetrical flow discharge
• high Q_f, high Q_s on one side of bar
• horizontal and inclined bedding; fewer foresets and cross strata at bar tail
• vertical normal and lateral grading
• imbrication common: a(t), b(i) fabric
• tendency to flow direction normal to bar front
• tendency to symmetrical discharge
• low Q_f, low Q_s
• horizontal bedding at bar bed passes laterally to foresets at angle of rest and cross strata at bar tail
• vertical normal and lateral grading
• imbrication in core and bar head: a(t), b(i) fabric except in cross strata

Figure 7.20 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. **(b)** Cross sections showing the association of stratification types and processes of gravel accumulation in longitudinal, diagonal and transverse bars. **(c)** Scheme of origin of sheet, longitudinal, diagonal and transverse bars. (Modified after Smith 1974 and Hein & Walker 1977)

may result in tabular sets of cross beds of limited extent (Fig. 7.20). 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. On wide open settings, such processes produce large repetitive bedforms, with associated internal cross bedding, that are thought to develop during waning of the flood.

Flat sub-horizontal bedding in gravel is known from many environments. For example, it forms at high discharges and relatively shallow flow depths during the development of longitudinal bars in braided rivers (Fig. 7.20).

Deposits associated with flat and cross-bedded rudites and their significance

In successions where rudites are dominant, it is important to record not only features of the rudites but also those of interbedded units of sandstone, siltstone or mudstone, since these reflect important changes in the sedimentation regime. Thin units of cross-bedded or cross-laminated fine-grain sandstones are not uncommon as interbeds, and are often the deposits of lower discharges. The boundary shear stress needed to roll pebbles commonly coexists with levels of turbulence needed to maintain sand grains in suspension (Fig. 7.7). This relationship, which is not valid for high sediment concentration or high transport rates, predicts that interbedded sand and gravel may occur without radical fluctuations in flow strength. However, interbedded siltstones and mudstones may represent periods of protracted settling from suspension during quieter current regimes. The presence of ripple marks, mud drapes, mud cracks, body fossils and, particularly, established populations of trace fossils (9.4) helps to characterize these periods and provides additional evidence for the setting in which the energy fluctuations were taking place.

7.4.6 Bed thickness and grain-size relationships

In thick rudite-rich sequences it has become useful to measure bed thickness and maximum particle size (often expressed as the arithmetic mean of the ten largest particles) within each bed. Bed boundaries are not always easy to identify (Fig. 7.21). In many such successions, boundaries may be erosive. This can sometimes lead to difficulties in estimating bed thickness on account of both amalgamation of beds (and thereby over-estimates) and removal of part of the bed (under-estimates). Particular care must therefore be taken to minimize this problem.

Plots of bed thickness versus maximum particle size will, in some cases, reveal a clear linear relationship (Fig. 7.22); in other cases a scatter occurs with no apparent order. Cases of doubt must be resolved by calculation of a correlation coefficient. Where a linear relationship is apparent, it may be inferred that the beds were the products of discrete depositional events. The thickness of the bed (BT) is thought to be a crude proxy for the thickness of the flow, while the maximum particle size (MPS) reflects the competence of the flow.

For carefully collected data, the value of the correlation coefficient for a group of beds can give an approximate measure of the consistency of the physical behaviour of depositional events. It also seems possible to distinguish deposition from predominantly cohesive (Bingham) from predominantly cohesionless (Newtonian) flows (Fig. 7.22; cf. Fig. 3.20), which are two 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 7.23, 7.24). It is therefore important to assess how the parameters discussed in §7.4 vary and how they are combined in different ways through the

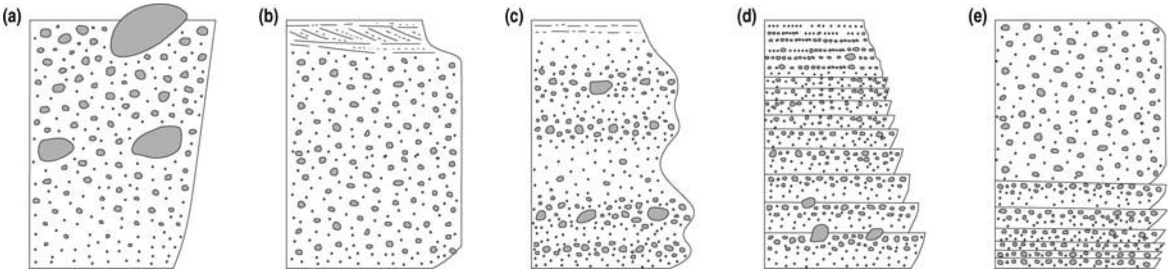
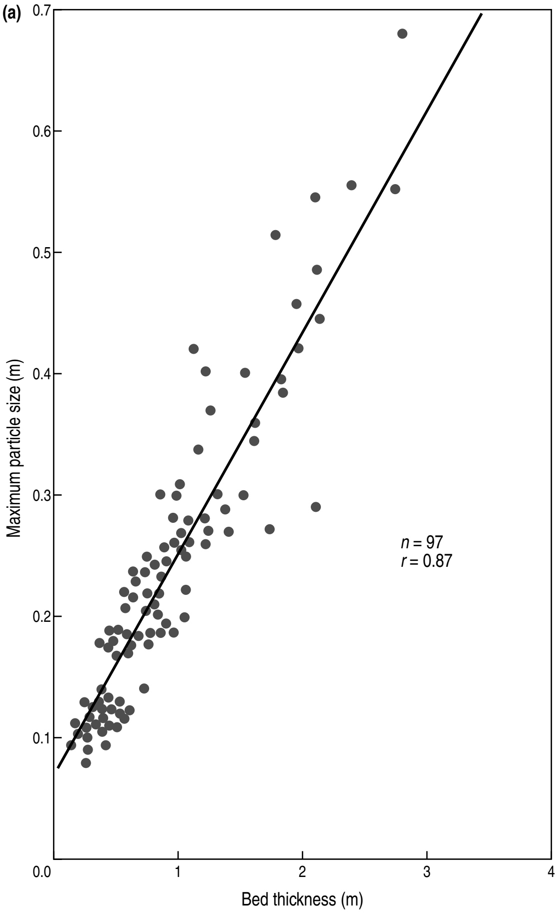


Figure 7.21 Variety of deposits that can be produced by traction-carpet sedimentation, depending on the duration and character of the traction carpet and the overall flow characteristics: **(a)** thick-bedded and inversely graded, **(b)** thick-bedded and massive with only the basal part inversely graded, **(c)** diffusely stratified, **(d)** thinning-upwards stratified, **(e)** thickening-upwards stratified. The thicknesses are not to scale and can be very variable. (After Sohn 1997)



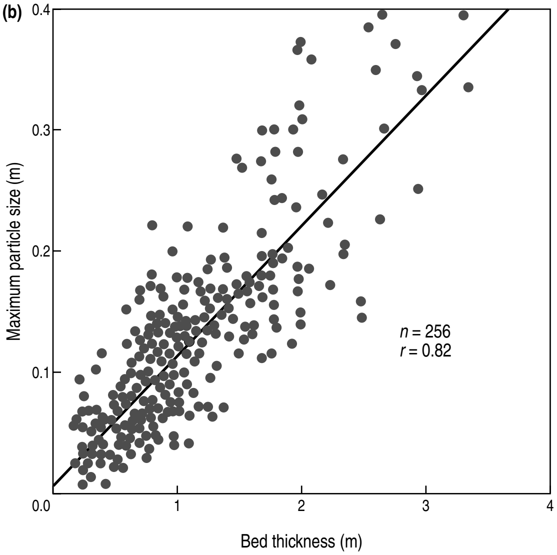


Figure 7.22 Examples demonstrating a relationship between maximum particle size and bed thickness: **(a)** a sub-aerial fan deposit, Stornoway Group, Permo-Triassic, northwest Scotland; **(b)** submarine fan delta, Ksiaz Formation, Devonian, southwest Poland. Part (a) shows a higher correlation coefficient (*r*), 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. (After Nemec & Steel 1984)

various parts of a succession. Some common associations and trends are now recognized, but further field observation and the collection of data from both modern and ancient successions continue to be needed. Without knowledge of the wider context of a particular succession, interpretation is best limited initially to depositional processes. The identification of environment may be successfully inferred in some cases (see Ch. 10), but will generally be suggested only by a broader analysis of all the components, discussion of which is beyond the scope of this book.

Cut banks of rivers and gravel pits often reveal short vertical successions of pebbly strata. To work most effectively on these, it is often useful to take a series of overlapping photographs at fixed distances from the outcrop face. From the resultant photomosaic, it is then possible to trace lateral changes in the distribution of features over many tens of metres. This technique is equally appropriate in ancient rudites.

It can be instructive to map variations of distribution of the features described in §7.4 on a horizontal surface in a suitable present-day environment, such as braided rivers at low water, beaches at low tide, or debris flow units on abandoned parts of alluvial fans. Useful insights can come from sampling and observing from the proximal to the distal ends of extensive environments, whether they be linear, as in a river, or radial, as on a fan.

Similar insights can come from the collection and plotting of a variety of data from rudite successions, the analysis of which can be useful in interpreting spatial and temporal trends. A simple example is the plotting of largest clast size against distance (it is common practice to use the mean size of the ten largest clasts), either across a present-day environment or down the observed

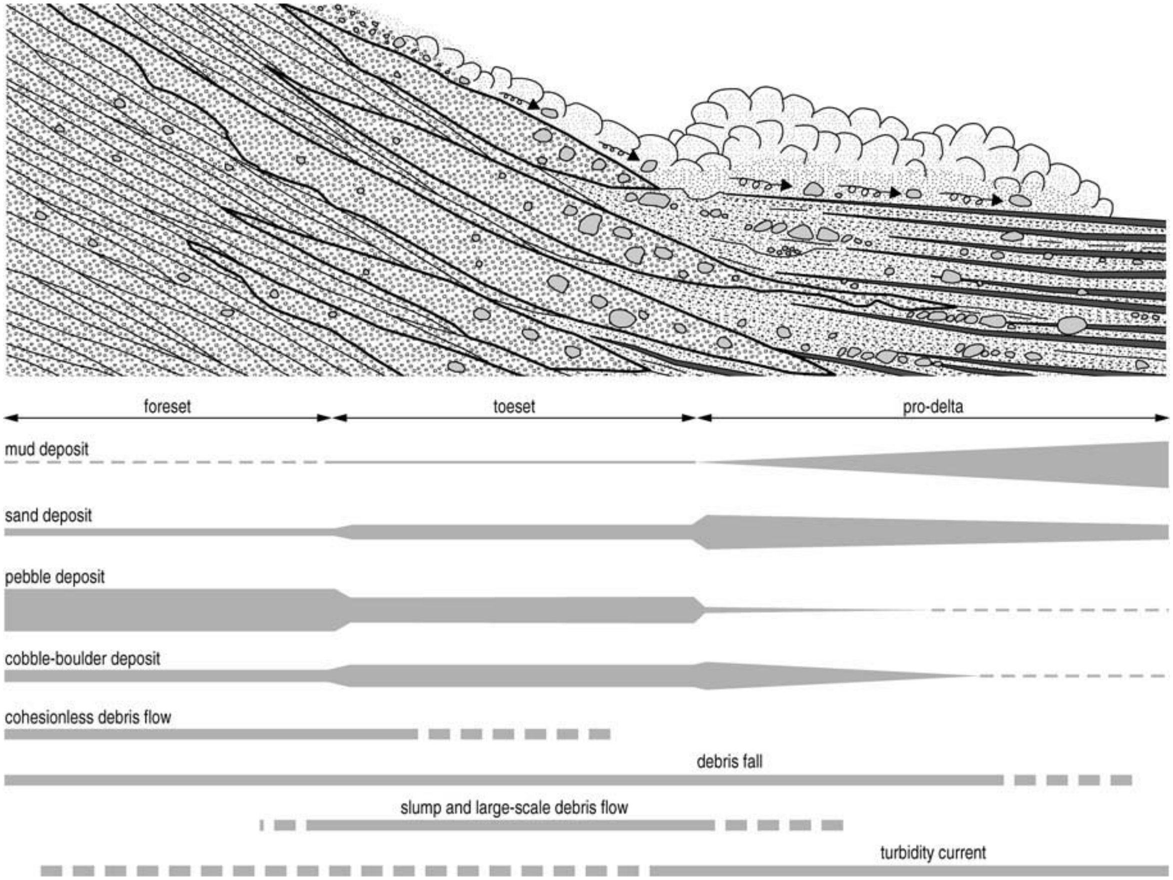


Figure 7.23 Schematic diagram showing the gross-scale geometry, internal architecture and textural variability of typical fan-delta deposits. The dominance of various depositional processes is indicated. Note the erosive bounding surfaces between the various depositional units within the large-scale 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)

or inferred palaeocurrent (e.g. in an alluvial fan succession). Another useful technique involves measuring bed thickness and comparing it with mean maximum clast size between various localities and stratigraphical positions within a succession (see §7.4.6).

Large-scale upward coarsening or fining of the component units of ancient alluvial fan successions can be interpreted in terms of temporal and spatial variations in energy regime that may be related to changes in tectonic activity, relief or climate. An increase in the energy regime of depositional events within alluvial environments is often associated with fan progradation and is expressed as a vertical coarsening-upwards succession. Such trends, when supplemented with independent evidence, may be attributed to increasing tectonic activity, which enhances basin-margin relief. Alternatively, fan progradation may occur as a consequence of the effects of an increasingly wet climate. A decrease in energy, leading to fan recession, is typically expressed as a fining-upwards succession, and may be related to a waning of tectonic activity or an increase in aridity.

Limestone breccias occur quite commonly, in some cases associated with reefs. Such breccias are often associated with reef talus-slope settings, where they exhibit a characteristic downslope decrease in maximum and mean clast size away from the reef crest, and a corresponding increase in matrix content. In vertical

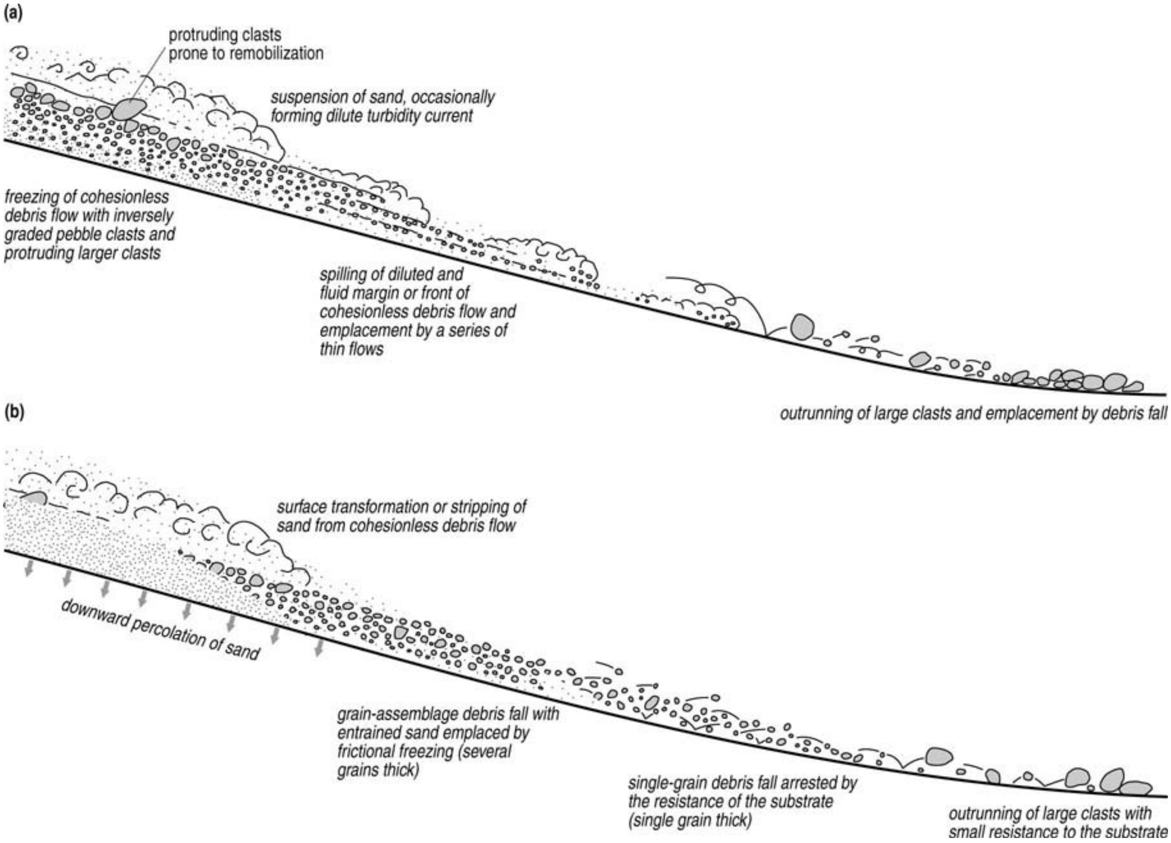
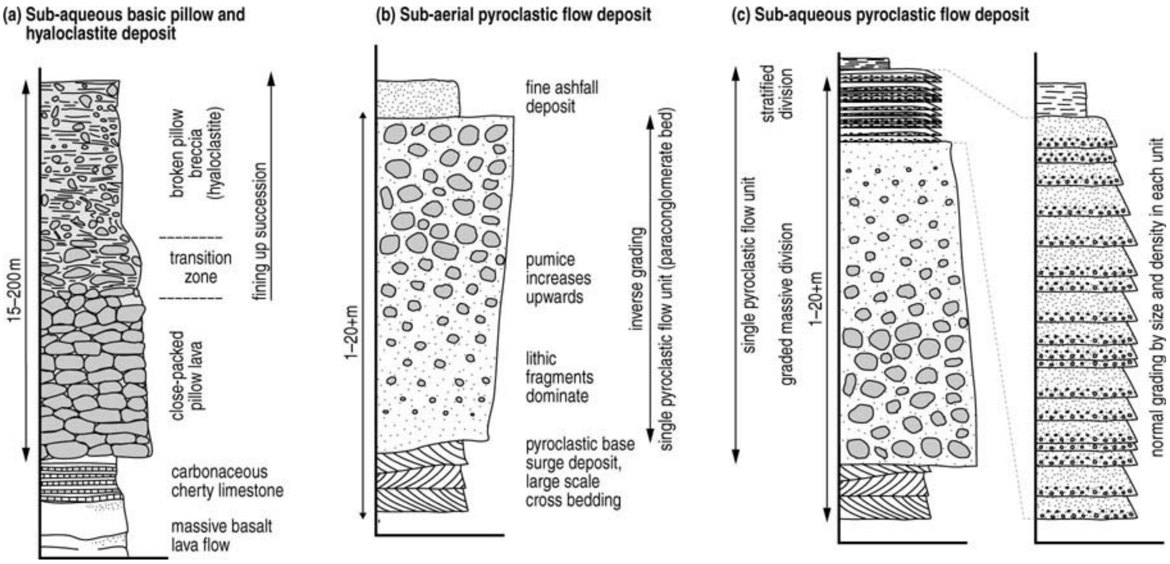

(a)
protruding clasts prone to remobilization
suspension of sand, occasionally forming dilute turbidity current
freezing of cohesionless debris flow with inversely graded pebble clasts and protruding larger clasts
spilling of diluted and fluid margin or front of cohesionless debris flow and emplacement by a series of thin flows
outrunning of large clasts and emplacement by debris fall
(b)
surface transformation or stripping of sand from cohesionless debris flow
downward percolation of sand
grain-assemblage debris fall with entrained sand emplaced by frictional freezing (several grains thick)
single-grain debris fall arrested by the resistance of the substrate (single grain thick)
outrunning of large clasts with small resistance to the substrate

Figure 7.24 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 down slope 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 down slope into a grain-assemblage debris fall and then into a single-grain debris fall, producing gravel sheets. Coarser clasts are transported farther down slope, resulting in lateral grading. (Modified after Sohn et al. 1997)

succession, an upward increase in reef talus clast size typically represents sites of deposition in successively higher positions up the talus slope and often signifies a progradation of the reef fringe. Early carbonate lithification means that any post-depositional disturbance is likely to lead to *in situ* brecciation rather than slumping and folding (see §8.4).

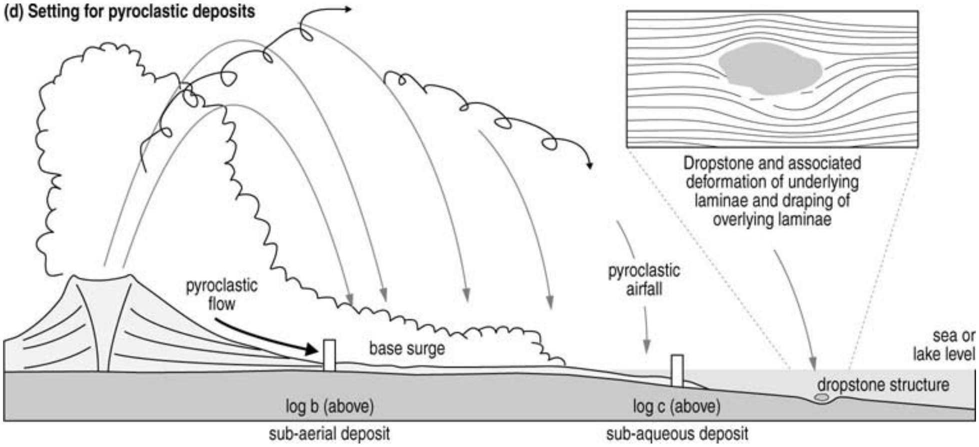
Volcaniclastic successions show varied combinations of structures and mass properties, which are not always easy to interpret in view of their specialized processes of origin (see §7.3.9), especially given that volcaniclastic materials are commonly resedimented soon after initial deposition (Fig. 7.25). 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 be an indicator of palaeowind. Pyroclastic flow and surge deposits are controlled by the form of topographical obstacles and may form thick beds in confined narrow valleys, or thin sheets that extend over widespread plains. Additionally, volcaniclastic successions are influenced by the nature of the eruptive event, such that, in the case of a prolonged eruption that progressively increases in intensity, pyroclastic deposits may show vertical grading trends that indicate a waxing phase. Careful analysis of both lateral and vertical trends can yield important information on


(a) Sub-aqueous basic pillow and hyaloclastite deposit

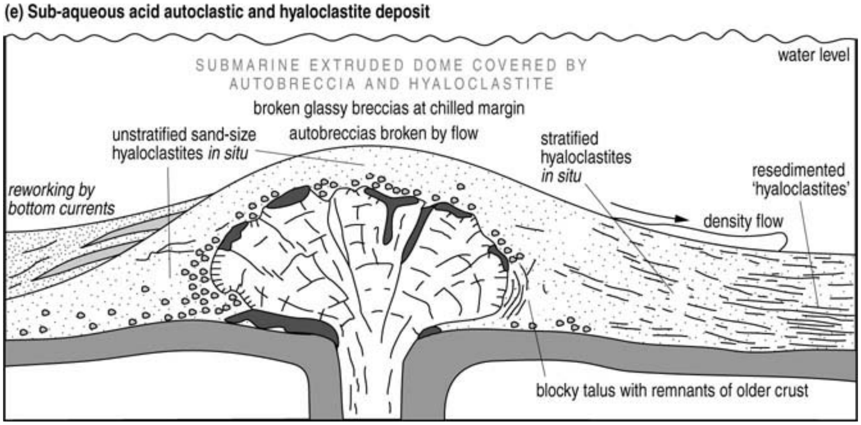
(b) Sub-aerial pyroclastic flow deposit

(c) Sub-aqueous pyroclastic flow deposit

15–200 m
broken pillow breccia (hyaloclastite)
transition zone
close-packed pillow lava
carbonaceous cherty limestone
massive basalt lava flow
fining up succession
1–20+m
fine ashfall deposit
pumice increases upwards
lithic fragments dominate
pyroclastic base surge deposit, large scale cross bedding
inverse grading
single pyroclastic flow unit (paraconglomerate bed)
single pyroclastic flow unit
stratified division
graded massive division
1–20+m
normal grading by size and density in each unit


(d) Setting for pyroclastic deposits

Dropstone and associated deformation of underlying laminae and draping of overlying laminae
pyroclastic flow
base surge
pyroclastic airfall
sea or lake level
dropstone structure
log b (above)
sub-aerial deposit
log c (above)
sub-aqueous deposit


(e) Sub-aqueous acid autoclastic and hyaloclastite deposit

SUBMARINE EXTRUDED DOME COVERED BY AUTOBRECCIA AND HYALOCLASTITE
water level
broken glassy breccias at chilled margin
unstratified sand-size hyaloclastites in situ
autobreccias broken by flow
stratified hyaloclastites in situ
resedimented 'hyaloclastites'
reworking by bottom currents
density flow
blocky talus with remnants of older crust

Figure 7.25 Volcaniclastic rudite successions and processes of formation. **(a)** Sub-aqueous flow deposit comprising pillow breccia and hyaloclastite (modified after Carlisle 1963 and Lajoie in Walker 1979). **(b)** Sub-aerial pyroclastic flow deposit. **(c)** Sub-aqueous pyroclastic flow deposit. **(d)** Model for the environmental occurrence of pyroclastic deposits. **(e)** Model for the environmental occurrence of sub-aqueous autoclastic and hyaloclastic flow sequences in acid rhyolite magmas with resedimentation (Lajoie in Walker 1979).

the nature of volcaniclastic processes and can help in palaeoenvironmental reconstructions.

It may be useful to order descriptions, analyses and interpretations of rudites with the help of Figure 7.26 and to improve these crude relationships as experience in field, laboratory and library grows.

7.6 Uses of rudite structures

The major use of structures in rudites is in identifying processes and determining environments of deposition. An obvious target for economic geologists seeking oil, gas, water or metalliferous deposits is a thick succession of orthoconglomerates in which porosity and permeability are high. Fluvial, shoreline and certain turbidite sediments will be more favourable in this regard than the deposits of environments where paraconglomerates are more common. Autoclastic and pyroclastic breccias are often the host rocks for primary sulphide mineralization and zeolite formation.

Studies of palaeocurrents based on the measurement of imbrication and cross bedding may help greatly in predicting the directions in which a succession should become thinner and finer, and hence the directions in which porosity and permeability diminish. It is easy to measure palaeocurrent directions at low water in gravelly streams or on fans, or beaches at low tide. Palaeocurrent measurements can be made on a variety of scales from imbrication, through cross bedding, to large features such as channels. The orientations of the sides of large-scale gravel-filled channels are likely to be parallel or subparallel to the overall palaeoslope; those of large cross beds are likely to show a unidirectional trend, although moderate variation about a mean is to be expected in most cases. In fluvial successions, the distribution of cross-stratal dip directions may be systematic. For example, it may be symmetrically bimodal about the downstream direction because of the growth of foresets at the ends of asymmetric diagonal bars (Fig. 7.20). Regional palaeocurrent patterns derived from imbrication in channels on submarine or alluvial fans often indicate a pattern of radial divergence. Those for beaches are mostly up or down slope, although a longshore orientation is noted in some instances. Where pebble imbrication coexists with minor structures developed in sand (e.g. on the exposed bed of a river), comparison of directions from sediments of different grain sizes can give valuable clues to changes in flow pattern as water level falls. Sand, being more mobile, is more easily reworked by reduced flow, whereas structures of gravels tend to preserve directions of high-discharge flows.

To the structural geologist, normal grading and cross stratification may help to determine way-up in rocks

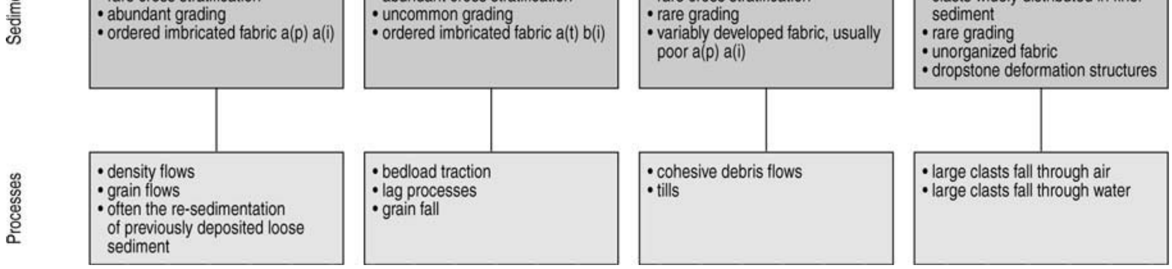

Sedimentary features and structures
sorting, size distribution
clast supported (orthoconglomerate)
matrix supported (paraconglomerate)
• rare cross stratification
• abundant grading
• ordered imbricated fabric a(p) a(i)
• abundant cross stratification
• uncommon grading
• ordered imbricated fabric a(t) b(i)
• rare cross stratification
• rare grading
• variably developed fabric, usually poor a(p) a(i)
• clasts widely distributed in finer sediment
• rare grading
• unorganized fabric
• dropstone deformation structures
Processes
• density flows
• grain flows
• often the re-sedimentation of previously deposited loose sediment
• bedload traction
• lag processes
• grain fall
• cohesive debris flows
• tills
• large clasts fall through air
• large clasts fall through water

Figure 7.26 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)

that are steeply dipping, although caution should be exercised, given the relatively common occurrence of both inverse grading and planar-tabular cross bedding, which make such successions less readily diagnosed in terms of way-up. 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 studies of the undeformed and deformed pebbles will enable a measure of strain to be determined.

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 pebble-armoured 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 often repay investigation.

Laboratory experience

Films of continuing volcanic activity are many and there are often 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 palaeohydraulic 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 much 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 is organic in nature. 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 resulting from organisms acting either to precipitate sediment or to bind existing particles.

8.2 Chemical precipitation

Inorganic precipitation of minerals from solution is mainly confined to evaporite minerals, commonly gypsum or halite. For any mineral to be precipitated inorganically, an aqueous solution must be supersaturated with respect to that mineral. 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 usually implies a hot arid setting. When supersaturation is achieved, precipitation takes place, provided 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. 8.1a). Eventually they sink to the floor of the basin

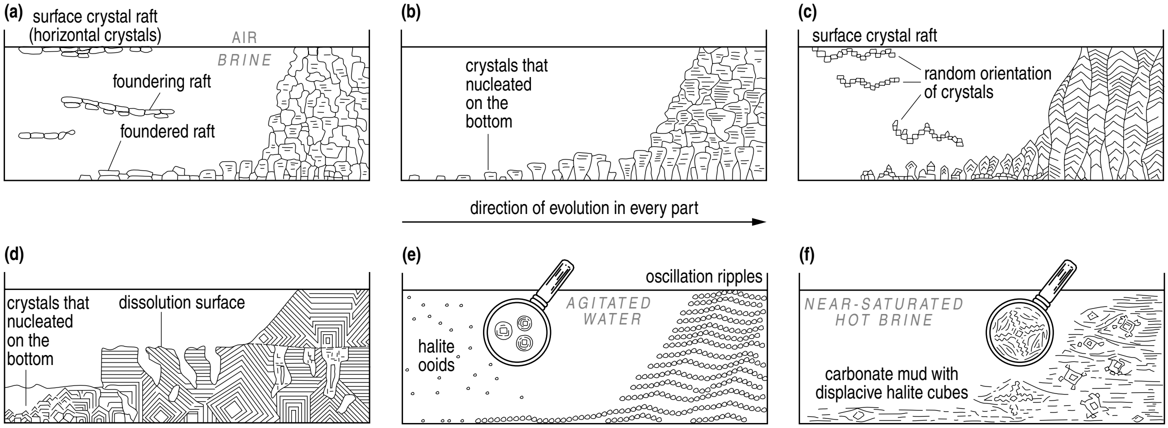

(a)
surface crystal raft
(horizontal crystals)
AIR
BRINE
foundering raft
foundered raft
(b)
crystals that
nucleated
on the
bottom
(c)
surface crystal raft
random orientation
of crystals
direction of evolution in every part
(d)
crystals that
nucleated
on the
bottom
dissolution surface
(e)
oscillation ripples
AGITATED
WATER
halite
ooids
(f)
NEAR-SATURATED
HOT BRINE
carbonate mud with
displacive halite cubes

Figure 8.1 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).

where most of the precipitation and crystal growth takes place. For well formed crystals to develop, both free space and an interval of time are required.

Processes of nucleation and crystal growth can be modelled in the laboratory, for example by allowing 1 L of a saturated solution of sodium chloride to evaporate gradually in a suitable tank. A hand lens can be used to observe the growth of crystals, 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 L 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 sequences 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 most commonly calcite ($CaCO_3$) and anhydrite ($CaSO_4$). Individual layers show great lateral continuity and may show grading, in terms of individual crystal size or mineral type or form (Fig. 8.2). 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 orientated crystals, suggest reworking and resedimentation of previously precipitated crystals

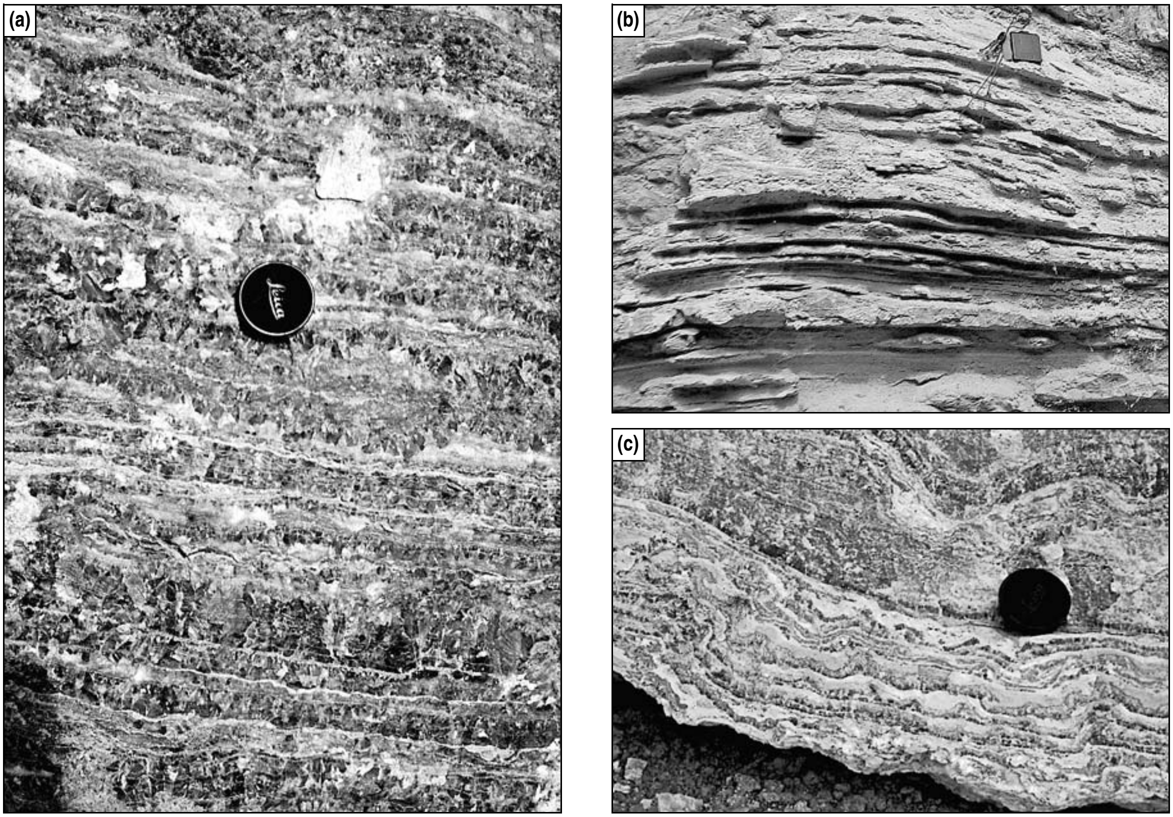


Figure 8.2 Examples of laminated and bedded evaporites: **(a)** Thinly bedded gypsum with bedding defined by slight grain-size differences and impurities; Miocene, southern France. **(b)** Interlaminated gypsum and mudstone; Yesares Member, Sorbas, southeast Spain; **(c)** laminated and deformed gypsum; Miocene, southern France.

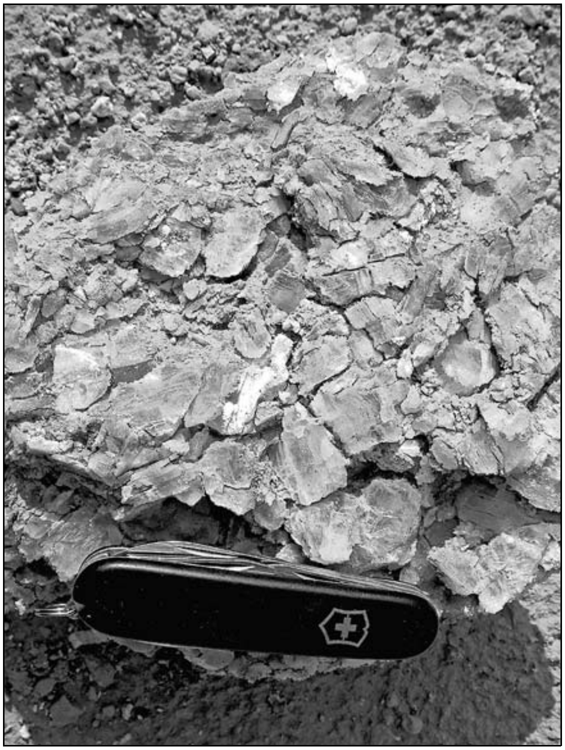


Figure 8.3 Irregular and randomly orientated crystals of gypsum arranged into a normally graded bed, Paradox Formation, Pennsylvanian, Utah. These crystals may have been reworked as clasts under the influence of a current.

of various sizes (Fig. 8.3; see also §6.7). This might imply episodic high-energy events, such as storms, which stirred up crystals and then allowed them to settle out as energy waned. Alternatively, grading of crystal size may reflect changes in the rate of evaporation, perhaps because of seasonal or longer-term climatic variations.

8.2.2 Fabrics due to vertical crystal growth

Growth of crystals on the basin floor commonly produces distinctive fabrics (Fig. 8.1a–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 with a preferred growth axis orientated vertically will grow most rapidly upwards, whereas growth of crystals 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 8.1c, 8.4a). Where crystal growth begins via nucleation at a specific site on the bed, distinctive crystal "trees" or "cones" may develop, the form of which can sometimes be preserved in the ancient record (Fig. 8.4b).

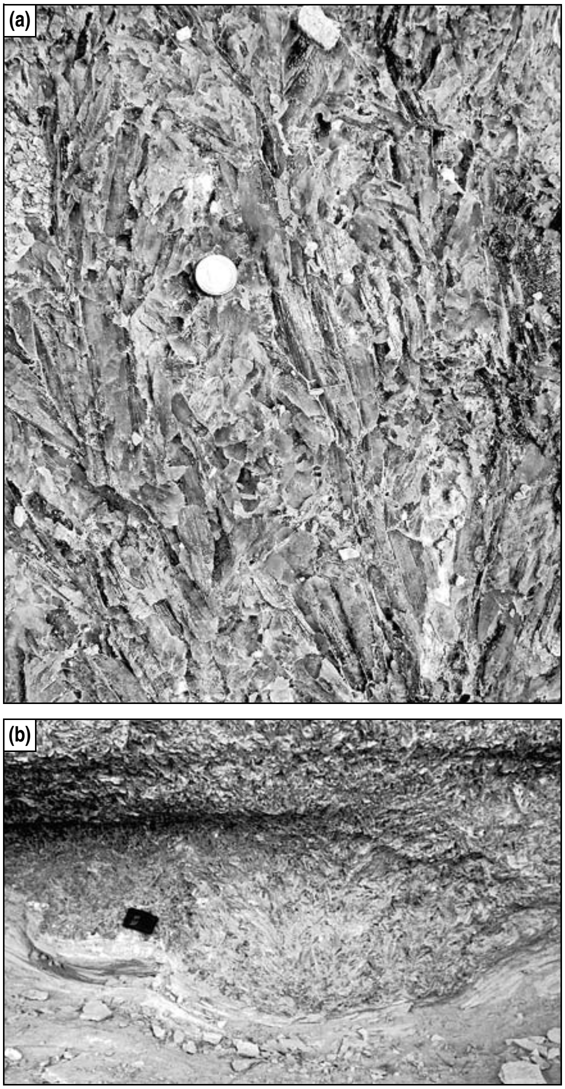


Figure 8.4 Examples of twinned selenite crystals (a form of evaporitic gypsum) arranged into subvertical columns. **(a)** Detail showing elongate blade-like crystal growth. **(b)** A nucleation site from which a tree-like structure of selenite crystals has grown; note the differential compaction of the underlying laminated mudstones. Both examples from the Yesares Member, Sorbas, southeast Spain.

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 sometimes 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. 8.1d).

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 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 tend to be large. One common evaporite mineral that is precipitated in this manner is halite, which typically develops on the sediment surface where continuing evaporation draws moisture up from the subsurface through capillary action. Halite crystals are characterized by a distinctive cubic form up to 3 cm in side length, often with stepped and indented faces called “hoppers”. A second common type of isolated evaporite mineral is “desert rose” gypsum, which usually forms in the shallow subsurface, often at the water-table interface in arid settings (Fig. 8.5). Such “roses”, which may be up to 15 cm in diameter, are made up of groups of large 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 sequences, 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 1 cm side length, and they may be isolated or may occur in lines or clusters (Fig. 8.6a). 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 shapes are apparent. Larger pseudomorphs commonly show a pattern of inwardly arranged steps on their faces, giving them an indented “hopper” form. The best-shaped pseudomorphs are commonly those occurring isolated from their neighbours. Gypsum “desert rose” pseudomorphs are also common in sandstones (Fig. 8.6b), especially those in arid settings where precipitation and subsequent dissolution are controlled by changes in groundwater level and chemistry. Gypsum pseudomorphs can be confused with crack fills, particularly discontinuous lenticular ones often attributed to synaeresis (see §9.2.1).

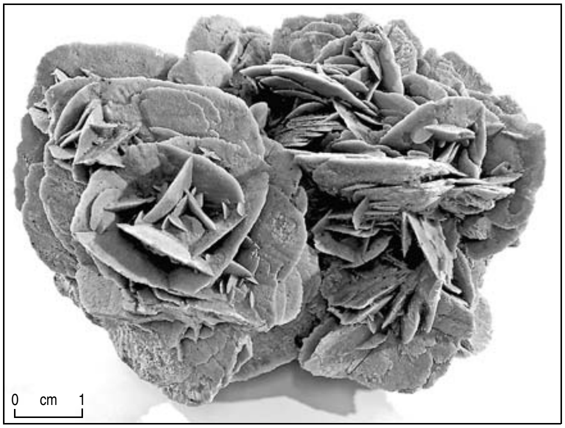


Figure 8.5 A large specimen of desert rose, a type of gypsum that grows in isolation within the shallow subsurface at the water table in arid settings; Recent, southern Namibia.

Pseudomorphs usually record the former presence of evaporite crystals growing at or just below the muddy sediment surface from an overlying supersaturated brine (Fig. 8.1f). 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

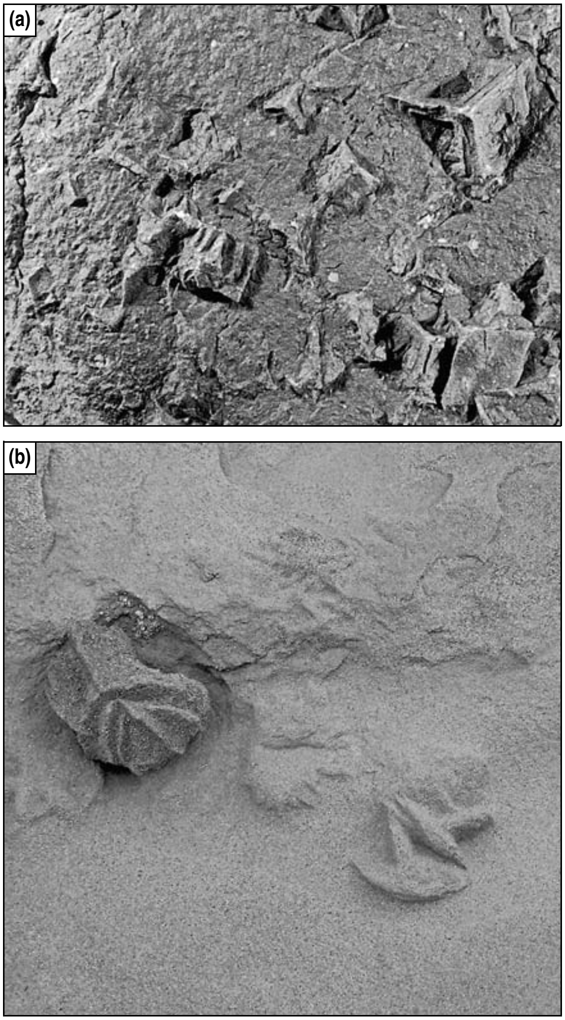


Figure 8.6 Examples of common sandstone pseudomorphs: **(a)** Halite pseudomorphs on the base of a sandstone bed from a thinly interbedded sandstone/siltstone sequence; note that the larger pseudomorphs have stepped "hopper" faces; Independence Fjord Group, Proterozoic, northern Greenland. **(b)** Gypsum **desert-rose** pseudomorphs within sandstone beds; Cedar Mesa Sandstone, Permian, Utah.

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 takes place as the result of a rapid influx of sediment-laden water, probably by a flood. This dissolves the crystals on the basin floor and fills the resulting spaces with sediment of a character different from that of the host material. This mechanism is sometimes borne out by 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 usually 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).

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.

Some evaporites show structures resulting from erosion and deposition. Small-scale scours, ripples, cross lamination and cross bedding are all quite common (Fig. 8.1e). 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 and silica. Deposits of calcium carbonate are precipitated from both hot and cold springs, and the precipitation may be explained by 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 also bears plant impressions. A second, more laminated and compact, form of calcium carbonate is **travertine**, which occurs

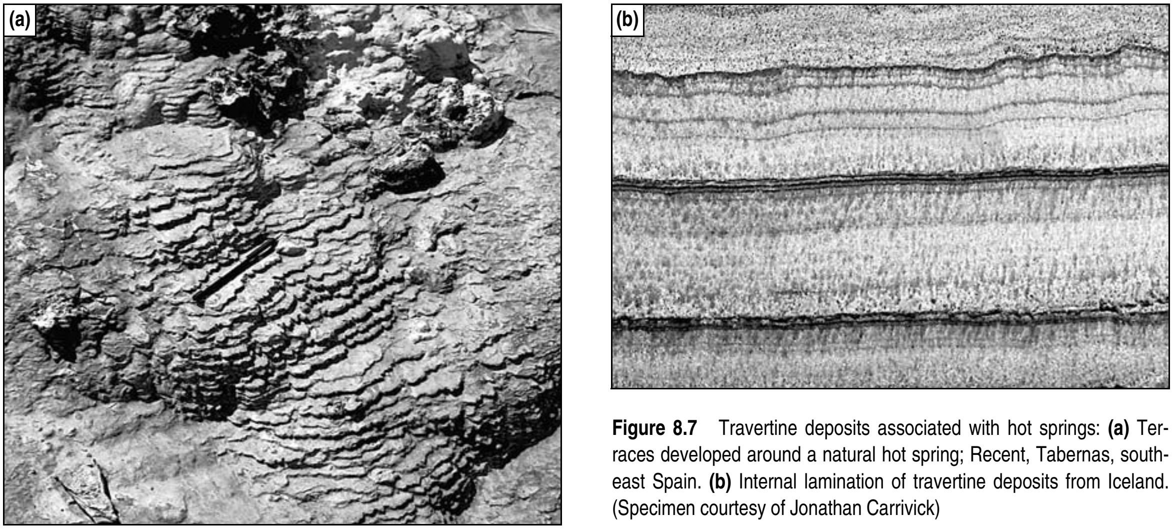


Figure 8.7 Travertine deposits associated with hot springs: **(a)** Terraces developed around a natural hot spring; Recent, Tabernas, southeast Spain. **(b)** Internal lamination of travertine deposits from Iceland. (Specimen courtesy of Jonathan Carrivick)

commonly in caves and also as the surface deposits of both hot and cold springs (Fig. 8.7). Cave travertines result from calcium carbonate, which is dissolved by percolating groundwater, being reprecipitated 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. 8.8), and steep surfaces are coated with dripstone layers. Spectacular terraces of hot springs, for example in Yellowstone Park (USA) or 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 microbially produced stromatolitic lamination (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 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**, resulting mainly from precipitation by a range of different organisms, and **stromatolites** and **oncolites**, produced by 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 fresh water 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 constitute the main component of carbonate sediments. Some skeletons remain mostly 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 (e.g. branching corals) contribute to the sedimentation process by building rigid framework structures that remain *in situ* upon the death of the organism (Fig. 8.9). 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.

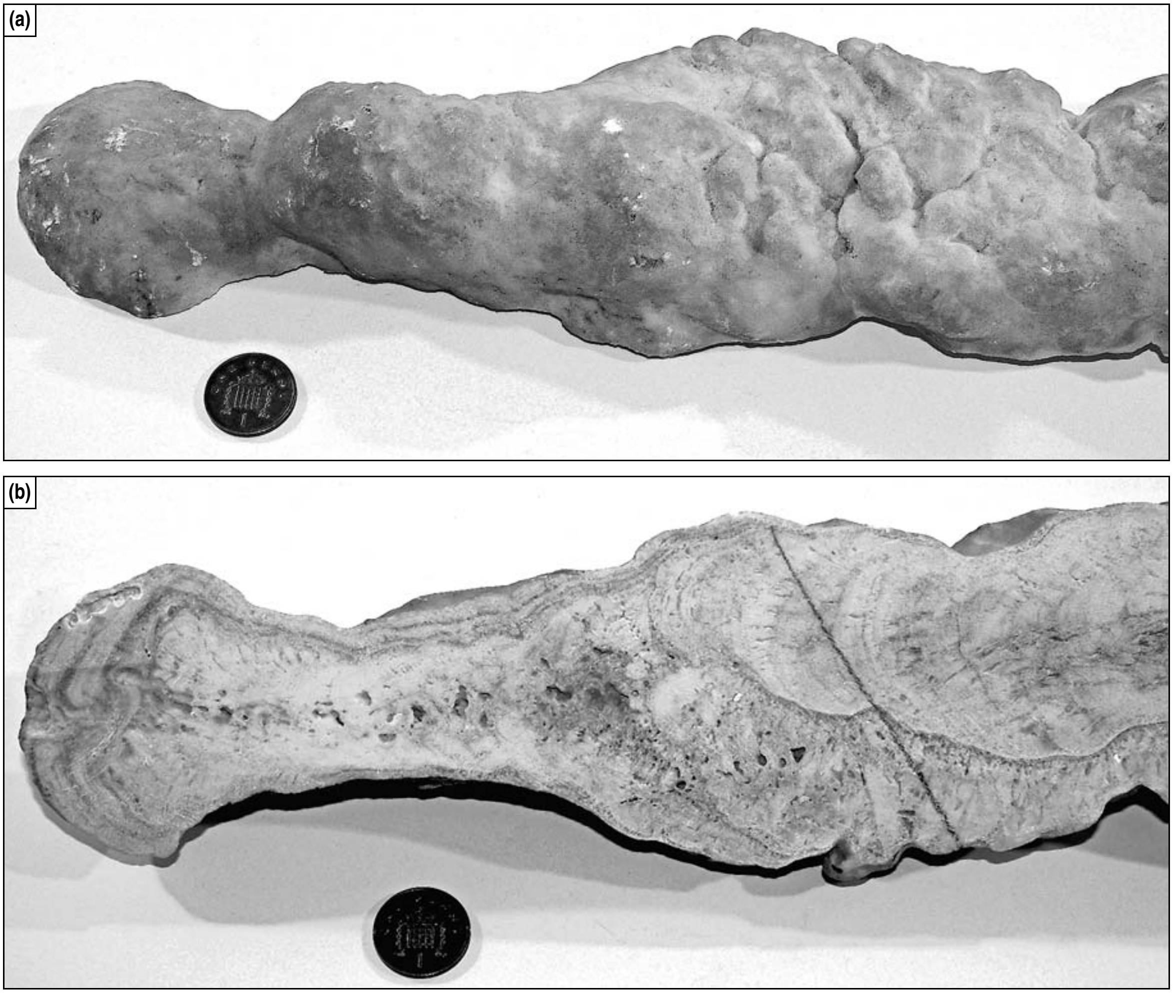


Figure 8.8 Cave stalactite (speleothem), a type of travertine: **(a)** external morphology, **(b)** internal lamination. Quaternary, southeast Spain. (Specimen courtesy of Emily McMillan)

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 sequences that were produced by colonies of organisms. Where such features had topographical 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, which has changed throughout geological time as different groups of organisms became important. Today, organic build-ups of carbonate on the sea floor occur in a wide range of water depths, although the most prolific and spectacular examples are in the fairly shallow water of tropical and subtropical coasts.

Reefs today occur as barriers, running for long distances parallel to a shoreline and separating the open ocean from more protected lagoonal areas (Fig. 8.10). 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 encircling a lagoon that lacks a central landmass at the present day. All these larger forms commonly have horizontal dimensions measured

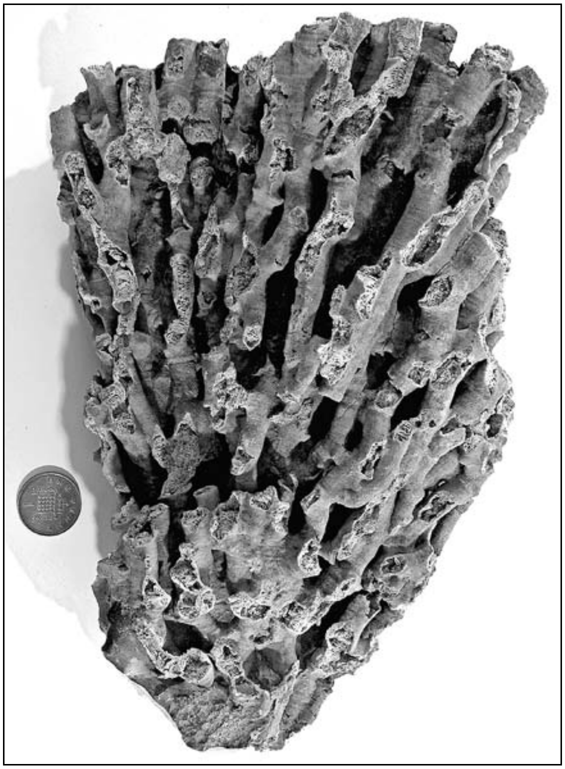


Figure 8.9 Carbonate framework coral. This is an example of a framestone, which, along with bindstones and bafflestones, represents one of the main reef-building structures in carbonate reefs. Lower Carboniferous, Derbyshire.

in kilometres and they may separate areas of deep and shallow water on opposite sides. At a more modest scale, small organic build-ups or **patch reefs** occur within lagoons and in other shallow marine settings. A comparable variety occurs in the rock record, where the recognition of the largest forms can require extensive geological mapping, exceptionally large exposures or high-quality seismic reflection data. Smaller reef forms can commonly be recognized in quarries and cliffs. In certain instances the present-day topography clearly reflects the palaeorelief of relict reef structures (Fig. 8.11). 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. 8.12a). Larger reefs, which are more likely to have acted as barriers between areas of contrasting water depth, are often more appropriately divided into **fore reef**, **reef** and **back reef** components (Fig. 8.12b). The nature of the sediments that comprise these large reef structures is commonly described with respect to their depositional texture (Fig. 8.13).

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), often in multiple generations; 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 depositional horizontal. This can be used as a way-up indicator and for the measurement of dip in unbedded carbonates. These **geopetal infills** can also occur in the body chambers of fossils (Fig. 8.14). **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. 8.15). Its origin seems to relate to the compactional de-watering 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 that 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 clearly defined bedding that is predominantly horizontal. Differences in the sediments of the various zones are also reflected in the associated organisms, although the fore-reef or flank deposits may include both sediment and organisms that

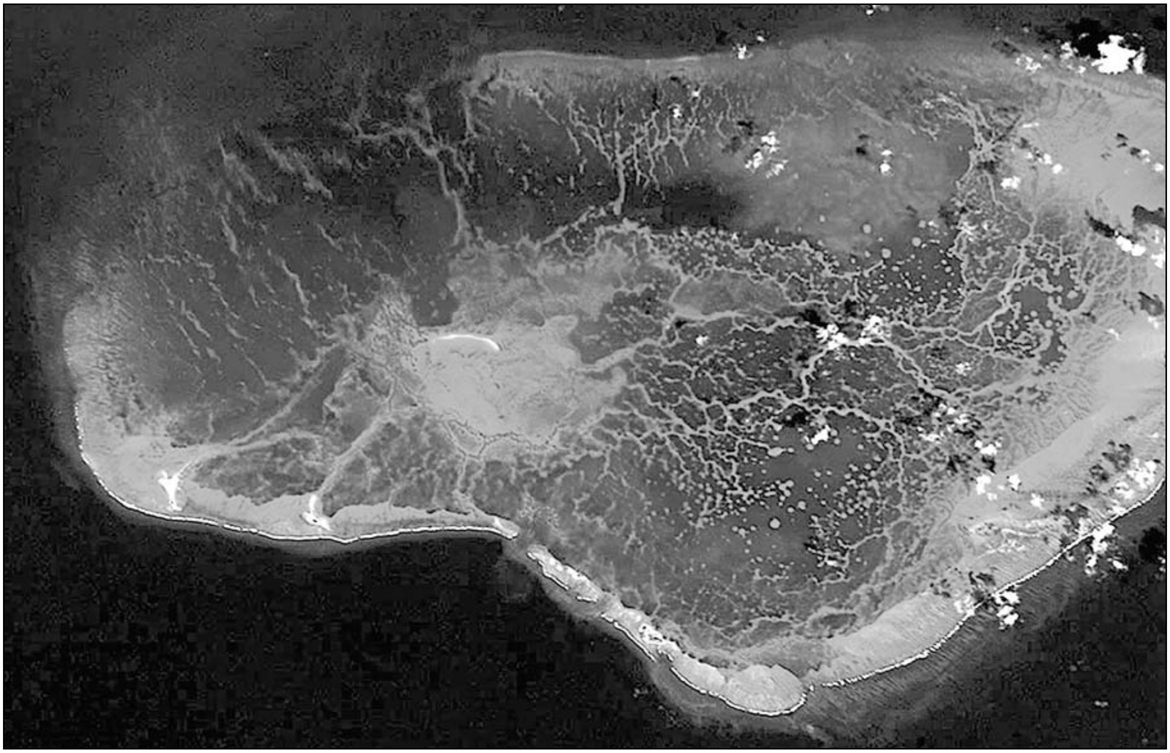


Figure 8.10 Pearl and Hermes Reef, northwest Hawaii; photograph taken from the International Space Station at an altitude of 400 km (image courtesy of the NASA Earth Observatory).

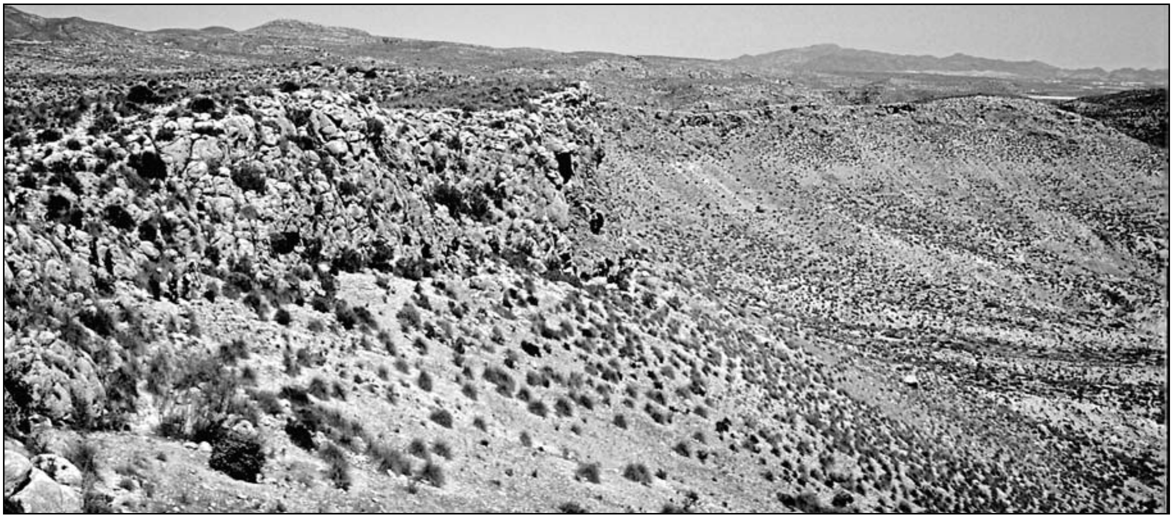


Figure 8.11 Example of present-day surface topography revealing the relict palaeotopographical form of a reef system. The prominent bench represents the transition from reef crest to the reef front; the slope represents the fore-reef apron. Exhumation of reef bodies such as these is common because of the removal by erosion of softer sediments that blanketed the reef. Nijar, Miocene, southeast Spain.

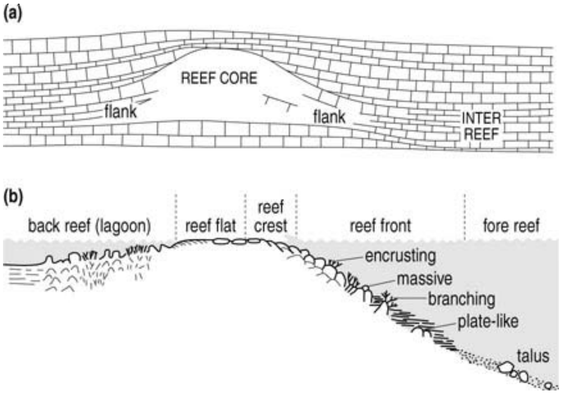
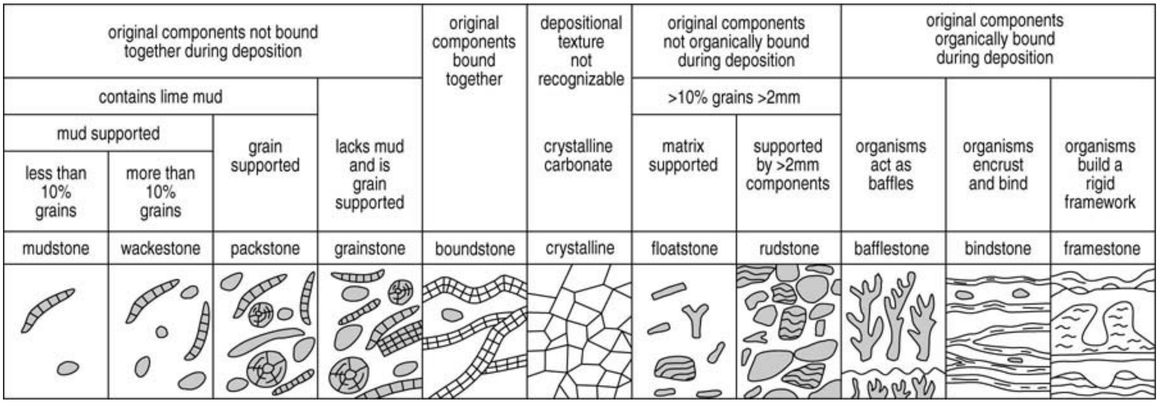


Figure 8.12 Definition diagrams for the major subdivisions of reefs: **(a)** isolated reef mound; **(b)** barrier reef at the boundary between deeper and shallower water, showing the main growth forms of reef-building organisms (modified after James 1979).

initially formed or lived on the reef or in the back reef. Fore-reef slopes are often covered with scree (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 large was the reef? In particular, what topographical 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 is often difficult. With larger barrier or fringing reefs, there may have been significant differences in water depth on opposite sides. It may sometimes be possible to trace marker horizons from the inter-reef sediments over the reef core, and these can give an indication of the topography. Bathymetry may also be suggested by larger-scale patterns of sediment distribution and by the present-day sub-aerial topography, which is commonly an exhumed submarine topography (Fig. 8.11).
- 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 may be visible in three dimensions in the present-day landscape. However, some inferences about reef shape can be made by observing at a more local scale. Clear differences in lithology between sediments on one side of a reef and on the other suggest that the reef acted as a barrier to water and sediment circulation, and the terms fore reef and back reef may then be appropriate. Small patch reefs or bioherms, recognizable in single outcrops are usually flanked and surrounded by similar sediment on all sides.
- 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



original components not bound together during deposition				original components bound together	depositional texture not recognizable	original components not organically bound during deposition		original components organically bound during deposition		
contains lime mud						>10% grains >2mm				
mud supported		grain supported	lacks mud and is grain supported		crystalline carbonate	matrix supported	supported by >2mm components	organisms act as baffles	organisms encrust and bind	organisms build a rigid framework
less than 10% grains	more than 10% grains									
mudstone	wackestone	packstone	grainstone	boundstone	crystalline	floatstone	rudstone	bafflestone	bindstone	framestone

Figure 8.13 The classification of limestones based on depositional texture (modified after Dunham 1962 and Tucker 2001).

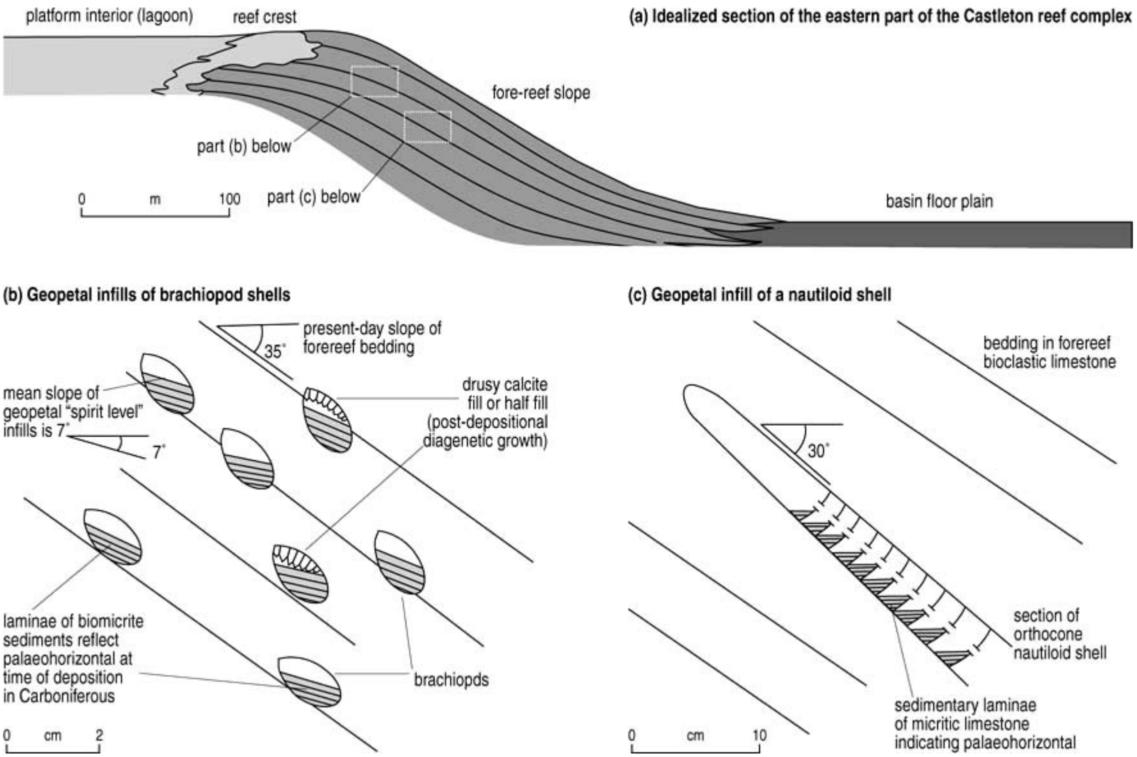

platform interior (lagoon)
reef crest
(a) Idealized section of the eastern part of the Castleton reef complex
fore-reef slope
part (b) below
0 m 100
part (c) below
basin floor plain
(b) Geopetal infills of brachiopod shells
35°
present-day slope of forereef bedding
mean slope of geopetal "spirit level" infills is 7°
7°
drusy calcite fill or half fill (post-depositional diagenetic growth)
laminae of biomicrite sediments reflect palaeohorizontal at time of deposition in Carboniferous
brachiopds
0 cm 2
(c) Geopetal infill of a nautiloid shell
bedding in forereef bioclastic limestone
30°
section of orthocone nautiloid shell
sedimentary laminae of micritic limestone indicating palaeohorizontal
0 cm 10

Figure 8.14 The use of way-up and geopetal (fossil spirit-level) structures for the determination of original bedding inclination. **(a)** 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 found in sediments of the fore-reef slope. **(c)** Geopetal infill found on the fore-reef slope. Shell cavities are thought to have been partly infilled with micritic limestone soon after death of the organisms, when the shells rested on the sea bed. The laminae within the infills indicate the attitude of the horizontal 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 (Dinantian) age reef complex from Castleton, Derbyshire, England. (Modified after Wolfenden 1953 and Broadhurst & Simpson 1967)

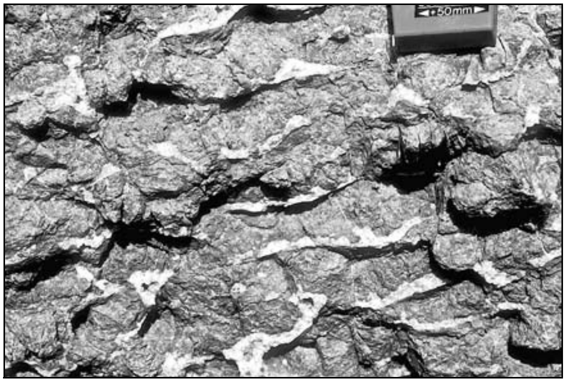


Figure 8.15 Stromatactis, an enigmatic structure in fine-grain limestone. Sparry calcite fills a flat-bottom cavity that must have been kept open close to the sediment surface during the early stages of diagenesis. Pentamerus Bjerge Formation, Lower Silurian, Washington Land, Greenland.

and are best approached after study of a book such as that by Wilson (1975).

- To what extent did early lithification occur in the reef? This is often best judged from observation of fore-reef and reef-flank deposits. Note whether these beds are made up of bioclastic debris or contain larger blocks of reef material. Bioclastic debris often occurs in steeply dipping beds, inclined away from the reef core and commonly showing normal or inverse grading. Large blocks, sometimes of huge dimensions, occur on some steep palaeoslopes associated with dipping fore-reef beds and are interpreted as submarine scree (talus) slopes, made up of lithified blocks that fell from the reef front. Geopetal infills within large blocks may help to establish the degree of tilting of the blocks or their way-up if the infills formed in the reef prior to redeposition (Fig. 8.14; see also §2.1.6).

8.3.2 Stromatolites and oncolites: structures due to algal binding

Stromatolites and **oncolites** are structures that show fine lamination caused by the trapping and binding of material by algae and cyanobacteria.

Stromatolites

Our knowledge of stromatolites derives mainly 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-grain carbonate and detrital material can also be involved. Lamination is characteristically thin, usually 1 mm 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.

For our purposes, it is enough to recognize some of the large- and medium-scale features of stromatolites that contribute to their field description. These are illustrated schematically in Figure 8.16. 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 develop a feel for the three-dimensional form of stromatolites; in addition to their vertical section (Fig. 8.17), describe their appearance in horizontal section or on bedding surfaces wherever possible (Fig. 8.18). 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 direction to their long axes. Record the direction of any such orientation, as it may give a guide to current or wave directions. Bedding surfaces may show a three-dimensional relief (Fig. 8.19) and this allows visualization of the morphology of the sediment surface during deposition.

Within a biostrome or bioherm there may be both lateral and vertical variation, and a full description should include not only the size and shape of the stromatolitic unit, but also the types, scale, orientation and distribution of different types of lamination.

A note of caution is necessary here in that stromatolitic lamination is rather readily confused with some non-biological lamination, such as that in travertine, tufa and the silica deposits, which form around geysers and hot springs (see §8.2.6).

Although much less widespread and varied than they appear to have been in the geological past, present-day algal mats display a range of morphological forms from flat sheets through various crinkled and pustular types to well developed columns. Present-day types occur in zonal patterns that relate to the physical and chemical conditions in which they develop; for example, columnar structures that are elongated in plan view 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 themselves. The lamination is produced by periodic variation in rates of sediment supply allowing the build-up of organic-rich layers. The algal mat re-establishes itself after an episode of rapid deposition by growing through the sediment layer, thus binding the sediments to generate a bindstone.

Most present-day stromatolites are associated with particularly high salinities in intertidal and supratidal settings, although subtidal examples are also known. The more widespread occurrence of stromatolites in rocks of Precambrian age probably results from the absence of animals that grazed upon the algae. The high

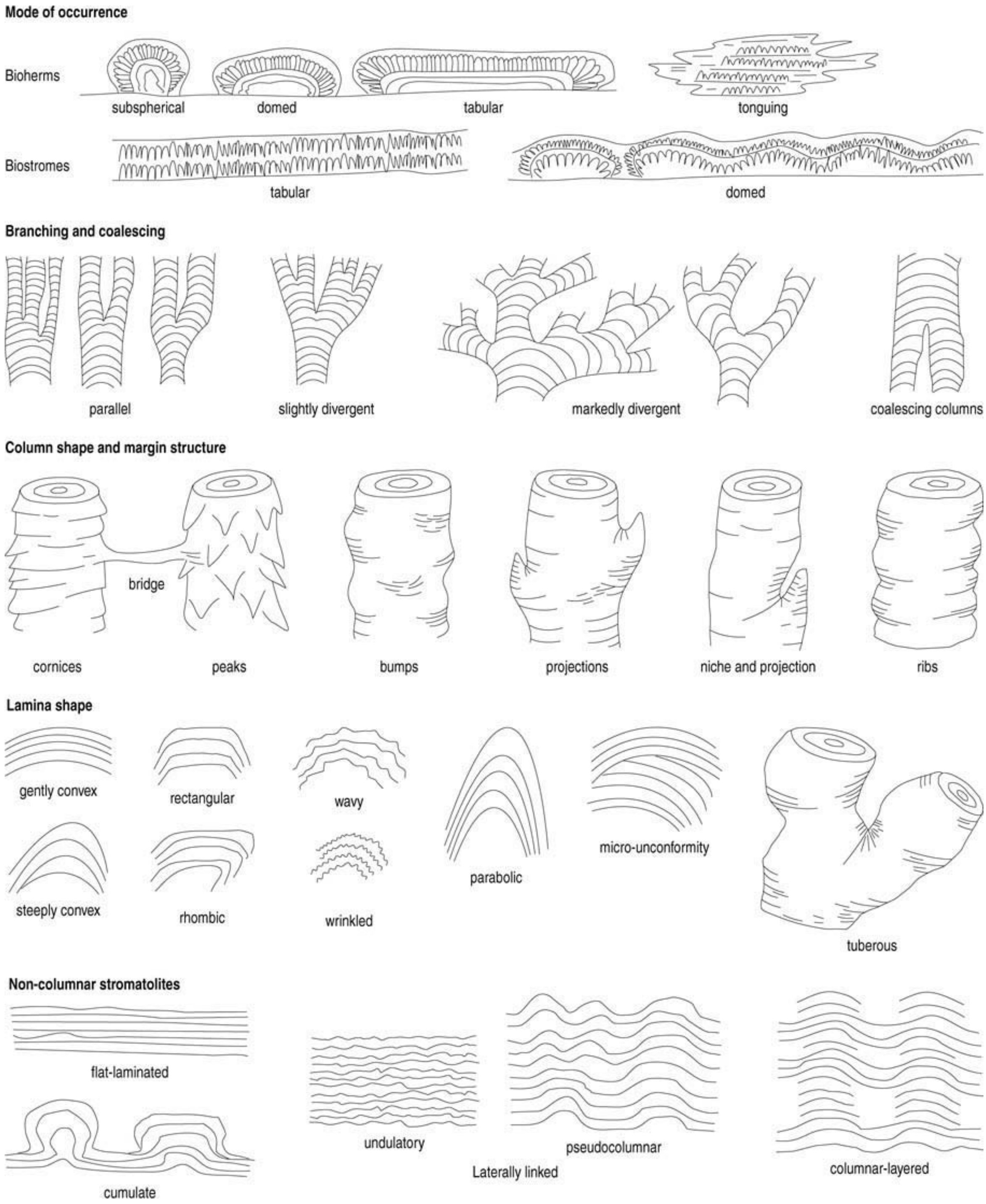

Mode of occurrence
Bioherms
subspherical
domed
tabular
tonguing
Biostromes
tabular
domed
Branching and coalescing
parallel
slightly divergent
markedly divergent
coalescing columns
Column shape and margin structure
bridge
cornices
peaks
bumps
projections
niche and projection
ribs
Lamina shape
gently convex
rectangular
wavy
steeply convex
rhombic
wrinkled
parabolic
micro-unconformity
tuberous
Non-columnar stromatolites
flat-laminated
cumulate
undulatory
pseudocolumnar
columnar-layered
Laterally linked

Figure 8.16 Definition diagram of the main terms used in the description of stromatolite bodies and stromatolitic lamination (modified after Preiss 1976).

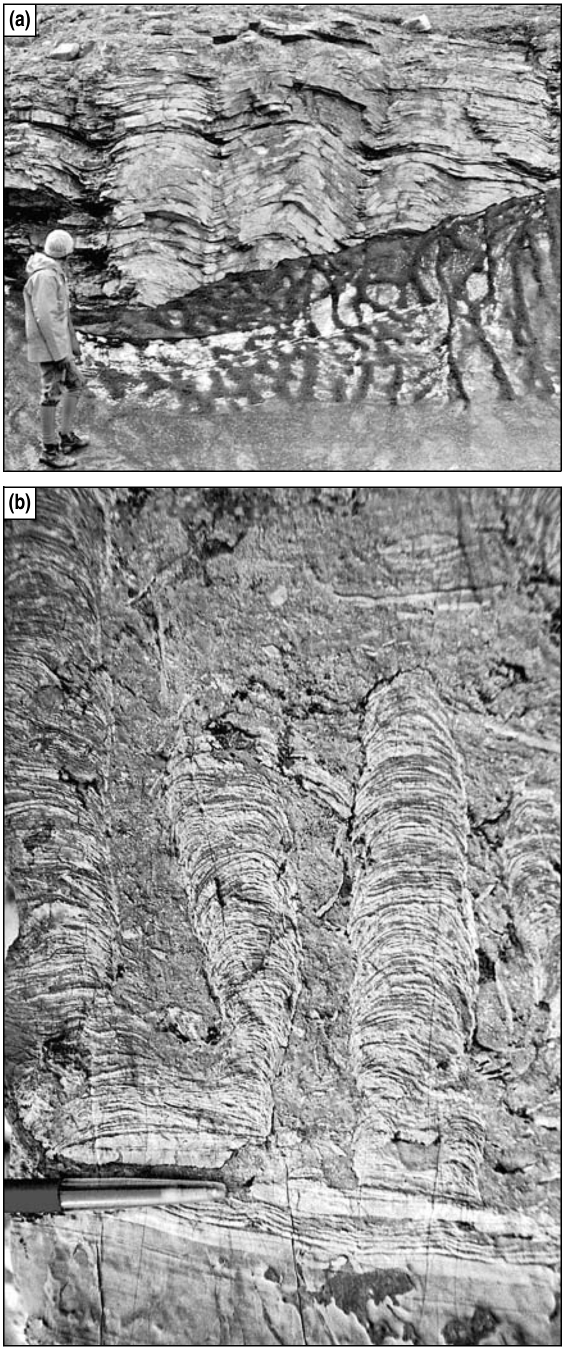


Figure 8.17 Examples of stromatolitic lamination seen in vertical section: **(a)** stromatolites with a columnar-layered structure, Moræneso Formation, Proterozoic, northeast Greenland; **(b)** upright, non-branching stromatolite columns, Eleanore Bay Group, Precambrian, eastern Greenland.

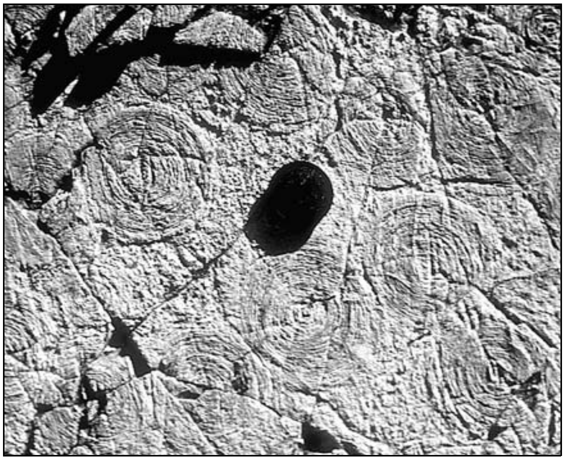


Figure 8.18 Horizontal sections through stromatolite columns; Fyn Sø Formation, Upper Proterozoic, northeast Greenland.

salinities associated with many present-day examples create conditions hostile to the animals that might normally graze on the algal mats 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 biostratigraphical zonation of otherwise unfossiliferous Proterozoic sediments have met with only limited success and acceptance.

Oncolites

Oncolites are spherical or less well rounded structures, commonly up to 5 mm in diameter but sometimes larger, often with a rather flattened shape. Internally they have a roughly concentric pattern of fine lamination similar to that present in stromatolites (Fig. 8.20). Careful examination of the lamination in cross section may sometimes show discontinuities.

These structures occur in both ancient limestones and in present-day lagoons and lakes. They result from the binding of sediment by algae onto isolated nuclei (sometimes called **algal biscuits).** Their growth, with its discontinuous pattern of lamination, suggests relatively calm conditions 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.

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, but 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 some beaches, giving rise to **beach rock**, a phenomenon also common in beaches of terrigenous sand. Carbonate sediments that become sub-aerially exposed through a fall of relative sea level are also subjected to cementation through solution and reprecipitation because of the passage of fresh water, when the cementing mineral is always calcite. Sustained sub-aerial exposure can lead to more extensive dissolution and to the development of karstified surfaces and carbonate soil profiles.

8.4.1 Seafloor lithification

Cementation close to the sea bed may be patchy or give rise to continuous layers. Early cemented patches take the form of **concretionary nodules**, around which differential compaction can take place, especially if the host sediment is of fine grain. 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 discerned 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 **tepee structures** (Fig. 8.21). In plan view the upwards-buckled ridges often have polygonal forms. These structures occur on the floors of present-day marine lagoons as well as in the rock record.

When the cemented layers, which probably formed 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 provides evidence for intervals of non-deposition. The recognition of borings is dealt with in §9.4.

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 as mass gravity flows. Breccias made up of rather tabular clasts are especially characteristic (Fig. 8.22).

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 reprecipitates it as low-Mg calcite cement. Sustained exposure leads to further modification, which may assume one of two forms: karstification and 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; they also developed very extensively in older limestones that have been subjected to abundant rainfall or snowmelt. Karstic features developed at the land surface give very characteristic morphologies and landscapes (Fig. 8.23). They include patterns of sharp ridges and deep clefts (clints and grykes), resulting from the widening of joints by dissolution resulting in steep-sided funnels and pipes. Within limestone, solution produces cavities that 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 breccia clasts are all of clearly local derivation, irregular and angular in shape and with a very mixed range of sizes. The spaces between them may

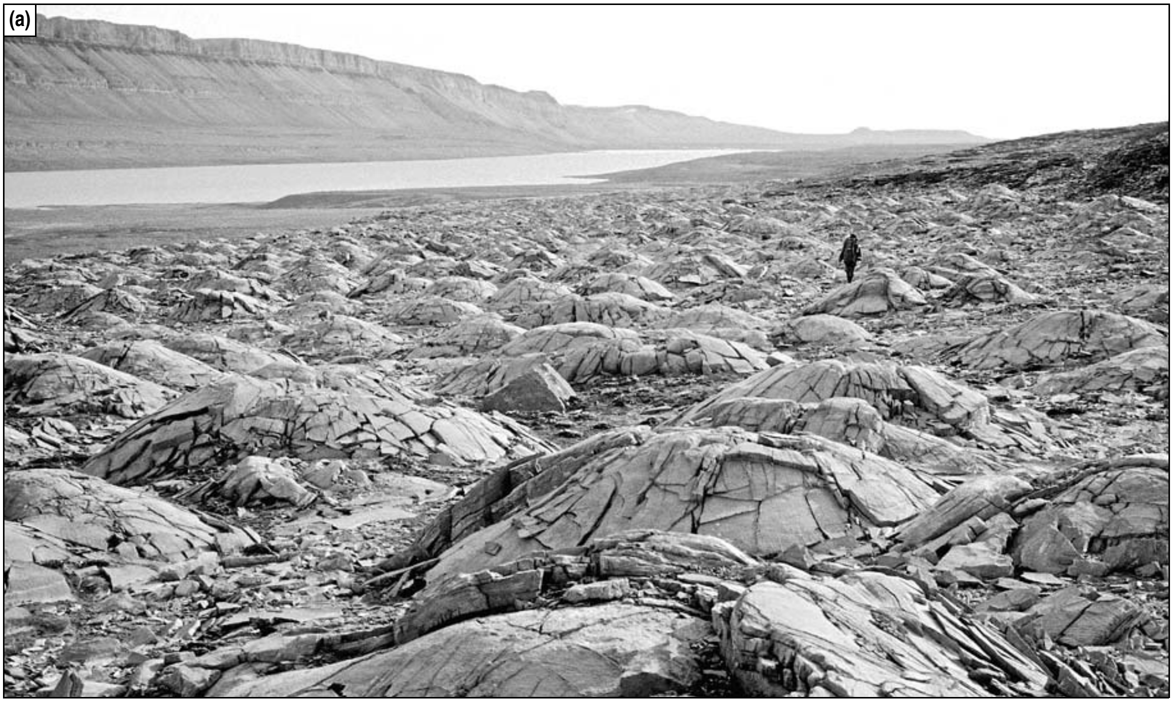
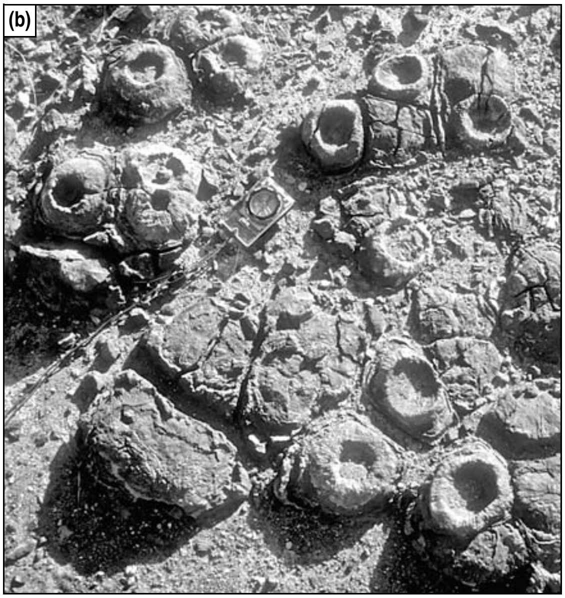


Figure 8.19 Stromatolite domes with their three-dimensional relief preserved on the upper bedding surface: **(a)** Morænesø Formation, Proterozoic, northeast Greenland; **(b)** Huab Formation, Permian, northeast Namibia.



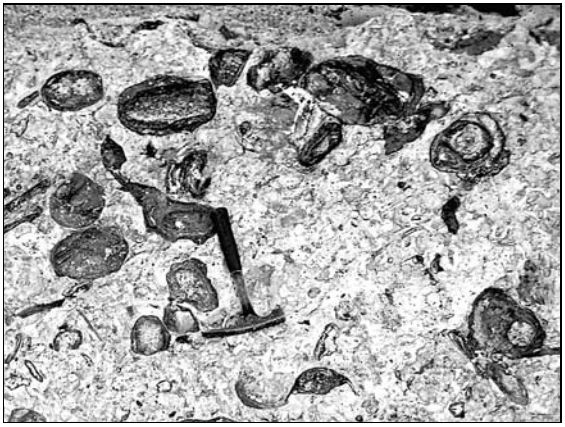


Figure 8.20 Cross section through large oncolites, showing internal lamination. Burton Beds, Inferior Oolite, Middle Jurassic, Dorset, England. Oncolites are typically 10–20 mm in diameter; these larger examples have been termed "algal biscuits". (Figure courtesy of Gilbert Kelling)

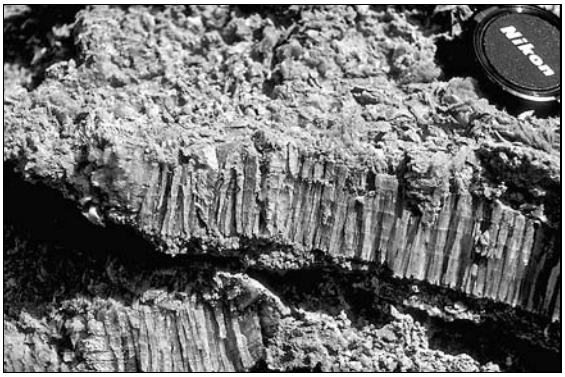


Figure 8.21 Gypsum teepee structure seen in section. Note the fibrous nature of the crystals, which have grown perpendicular to the depositional surface. As the gypsum crust grew to cover the surface, it cracked into a series of plates that, with continued growth, ramped over one another. Recent, Lake Eyre, Australia.

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 produced by relatively recent (Quaternary) activity from those that developed during the early post-depositional history of the limestone. This is often 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. Such occurrences are almost always palaeokarstic in origin (i.e. formed during the accumulation of the succession) and they represent 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, e.g. sulphates, and they are dealt with in §9.3.1).

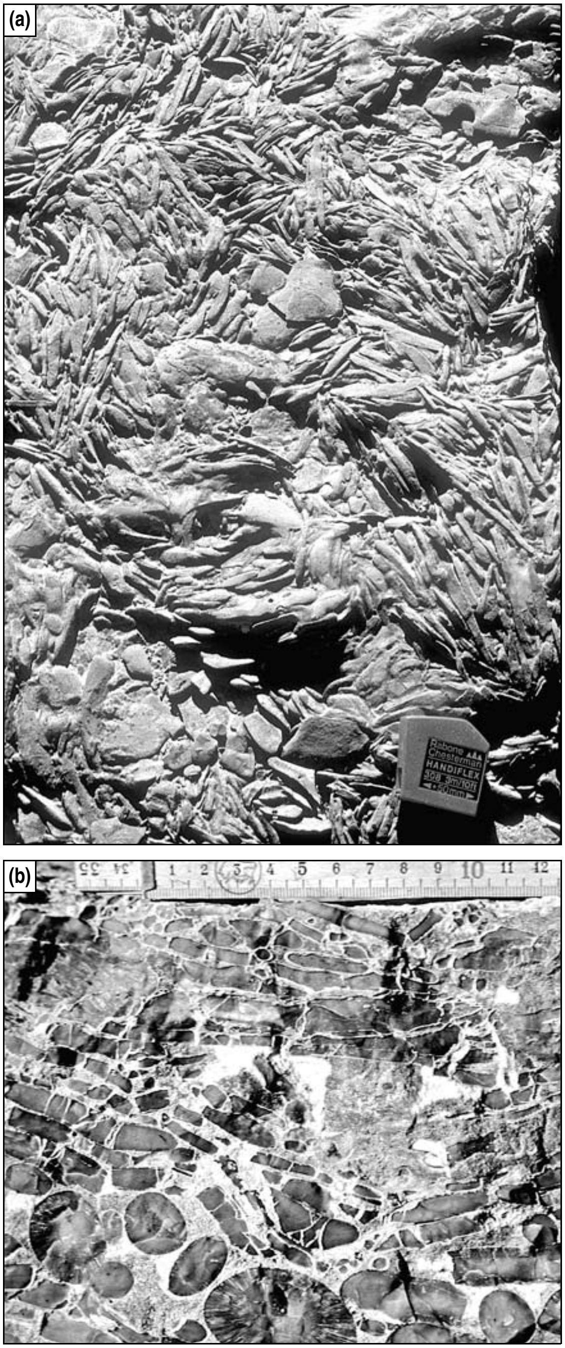


Figure 8.22 Breccias developed in thinly bedded limestones as a result of secondary movement following partial lithification: **(a)** Cass Fjord Formation, Upper Cambrian, Washington Land, northwest Greenland; **(b)** broken slab of concretionary material, Brønlund Fjord Formation, Lower Cambrian, Peary Land, northeast Greenland.

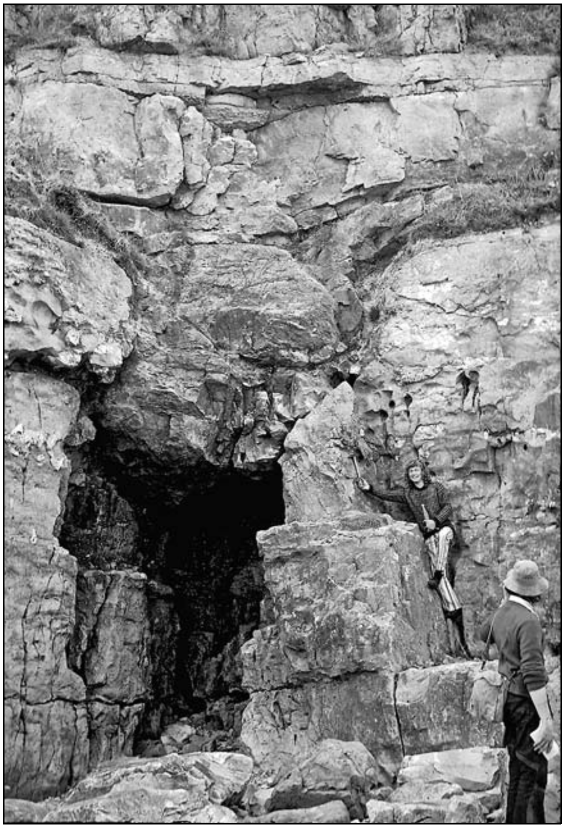


Figure 8.23 Large palaeokarstic solution pipe in limestone infilled by sandstone. The feature developed on the surface of the limestone soon after its deposition and early lithification, during a period of lowered sea level and emergence. The base of the pipe is at the level of the hammer; the cave below is of recent origin. Lower Carboniferous, Anglesey, Wales.

8.5 Other bedding phenomena in limestones

Many 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. The resulting sedimentary structures are very similar therefore to those described in Chapters 4–6. However, there are a few structures and bedding styles that seem to be unique to carbonates and which 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 precipitated.

8.5.1 Large-scale sigmoidal cross bedding

Certain shallow-marine limestones show sigmoidal cross bedding at the scale of several metres' vertical thickness and great lateral extent, often with complex smaller-scale structures superimposed. The sediment is often composed of oolitic limestone that is characterized by grains coated in calcium carbonate as a result of inorganic precipitation in high-energy shallow-water settings. The coated **oolith** grains are up to 5 mm in diameter and often exhibit a fine concentric internal lamination. In some cases, the cross-bedded units may alternatively be composed of bioclastic limestone. 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 terrigenous supply. Indeed, the agitation of carbonate particles of sand size 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. However, close examination 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. Visits to salt or gypsum mines may be possible.

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 environments because of their susceptibility to dissolution; they are, however, relatively common in arid and semi-arid climate settings.

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 well within the capability of most students.

Recommended references

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CHAPTER 9

Structures created by deformation and disturbance

9.1 Introduction

Any sediment may be disturbed after deposition, but disturbance is most common in sands and finer-grain 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 physical and chemical disturbance took place. Sometimes, it occurred soon after deposition at, or close to, the contemporaneous surface, and in other cases was associated with later burial and lithification.

Many deformational structures are valuable as way-up indicators and all record something about conditions within the sediment or at its surface after deposition.

9.2 Physically induced soft-sediment deformation

This results from mechanical forces, commonly gravity, acting upon weak sediment, usually silts or sands, at the sediment surface or soon after burial (Figs 9.1, 9.2). There is no neat way of classifying these structures. Here we use a broadly morphological scheme, based upon where they most commonly occur. However, several structures are seen in both vertical section and on bedding surfaces, and might, therefore, be placed in more than one class. They are described under only one heading, usually their most common occurrence.

Most types of soft-sediment deformation depend on unconsolidated sediment being in a weak condition. The 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 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 re-adjustment of grain packing to reduce tan φ, or by increasing the pore-fluid pressure ρ. Cohesion is the least readily changed property, as it is mainly 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 because of 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 either widespread and external (e.g. earthquakes) or local, for example a rise in water level or an episode of sudden deposition.

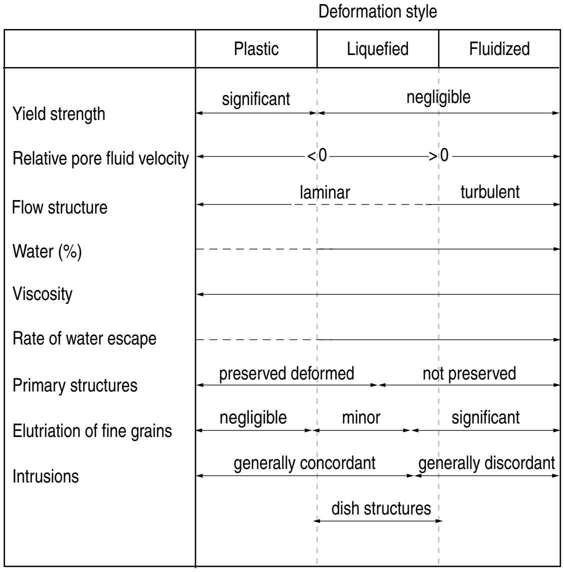

Deformation style
Plastic
Liquefied
Fluidized
Yield strength
significant
negligible
Relative pore fluid velocity
<0
>0
Flow structure
laminar
turbulent
Water (%)
Viscosity
Rate of water escape
Primary structures
preserved deformed
not preserved
Elutriation of fine grains
negligible
minor
significant
Intrusions
generally concordant
generally discordant
dish structures

Figure 9.1 Characteristic properties of plastic, liquefied and fluidized styles of deformation (modified after Owen 1987).

		LOSS OF STRENGTH				
		Exceed strength of sediment			Liquidize	
		Internal tensile (brittle)	Internal cohesive (plastic)	External surface cohesive (plastic)	Liquefied	Fluidized
Gravitational body force on slope		Slides	Slumps	Slumps and slides	Debris flows	
Unequal confining load		Growth faults	Loaded ripples; shale ridges and diapirs		Loaded ripples and sole marks	
						Clastic dykes Sand volcanoes
Gravitationally unstable density gradient (density inversion)	Continuous	Soft sediment faults			Convolute lamination	
	Within a single layer				Dish structures	Water-escape pipes and pillars
	Multiple layer, not pierced				Bedding-surface load clasts	
	Multiple layer, pierced		Shale ridges and mud diapirs		Ball and pillow/pseudo-nodules Isolated load balls	
Applied shear stress	Current drag				Overturned cross bedding	
	Vertical					Water-escape pipes and pillars

Figure 9.2 Types of physical deformation structures in relation to the nature of the deforming force (modified after Owen 1987).

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 closer grain packing has been created.

In addition to shock and repacking, excess pore-fluid pressure can be produced during rapid deposition of fine-grain sediment. The low permeability of such sediments prevents the escape of pore fluid and, thus, the compaction of the sediment at a rate that balances 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 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 restricted 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 fluidization allows some grain sorting to take place, usually by upward removal of fines. In liquefied sediment, fluid and grains move essentially together, giving little scope for 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 a whole bedding surface (Fig. 9.3).

Upwards-pointing fingers or wedges of the underlying unit occur between the sandy lobes. These are **flame**

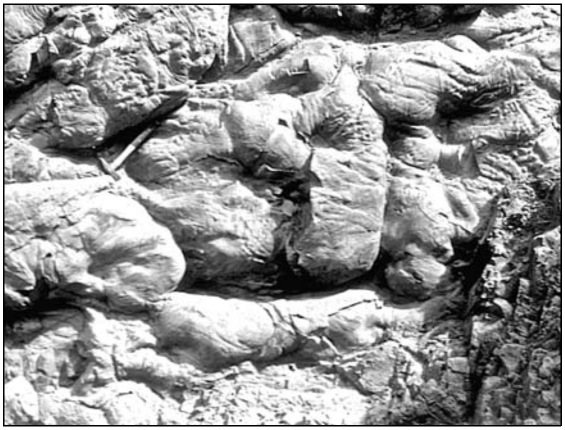


Figure 9.3 Load casts on the base of a sandstone bed from an interlaminated sandstone/mudstone succession; Bude Formation, Upper Carboniferous, north Cornwall, England.

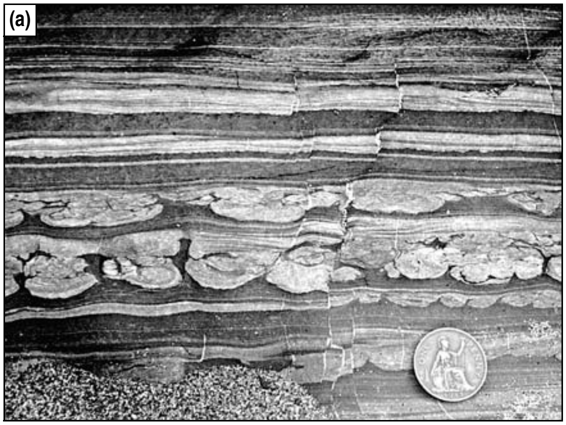
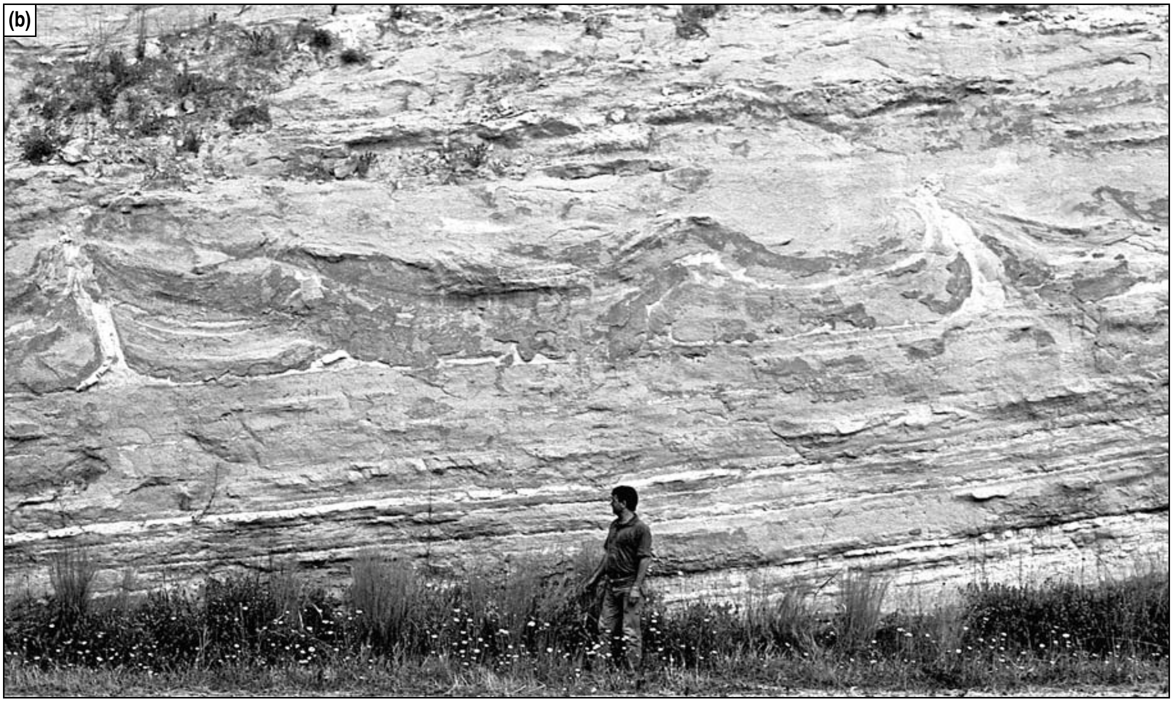


Figure 9.4 Examples of load casts and associated flame structures seen in section. **(a)** Loads on the base of a sandstone bed with flames of mudstone squeezed upwards between them; Bude Formation, Upper Carboniferous, north Cornwall, England. **(b)** Large loads and flames in shallow marine sandstone beds; Handere Formation, Pliocene, Adana Basin, Turkey (photo courtesy of Gilbert Kelling).



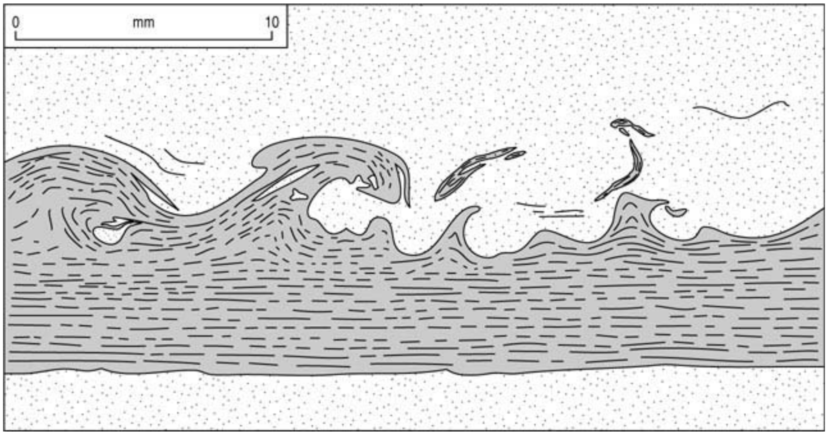


Figure 9.5 Vertical section revealing the geometry and interrelationship of load casts and flame structures resulting from the sudden emplacement of a volcanogenic sandbed on top of an unlithified mud unit. Langdale Slates, Ordovician, Cumbria, England (modified after Sorby 1908).

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 is parallel with 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 because of 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 because of excess pore-fluid pressures generated by 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 cause this, but externally generated shocks are equally possible. This effect can be simulated in the laboratory by violently jarring a waterlogged sand and mud sequence.

Whatever the reason, the combination of density inversion and temporary weakness leads to the sinking of one bed into the other, either randomly or at localized 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 sandbed must have sunk. Preservation of internal lamination within load balls implies that the sand was not totally 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 later), 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.

Sand and mud volcanoes

Sand and mud volcanoes are relatively rare structures that occur most commonly in sandstones, often interbedded with mudstones, the volcanoes themselves

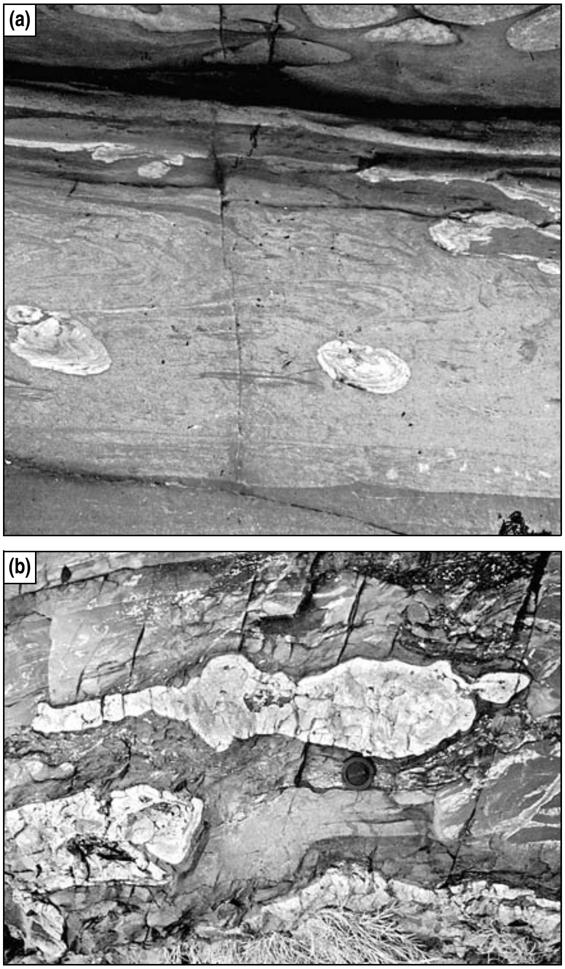


Figure 9.6 Examples of isolated load and flame structures. **(a)** Isolated sandstone load balls (pseudo-nodules) in a siltstone with a "slurried" texture. The original sandstone bed has totally foundered and collapsed into the finer sediment. Load balls are about 5 cm in diameter. Bude Formation, Upper Carboniferous, north Cornwall, England. **(b)** Isolated flame structure composed of siltstone that has been incorporated into surrounding volcanic rocks following their rapid emplacement; Strumble Volcanics, Ordovician, Strumble Head, Pembrokeshire, South Wales.

usually being of medium- or fine-grain sand, although in some cases of silt and mud. They occur on upper bedding surfaces and are also seen in vertical section. Volcanoes often 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 10 cm and several metres and being up to 50 cm high. They commonly have a crater-like depression in their centre, and their flanks may carry radially arranged 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).

They result from liquefied sand being extruded through a local vent at the sediment surface. The volcanoes are, in effect, localized equivalents of the transposed sandsheets associated with sandstone dykes (see pp. 190–193). Sand volcanoes commonly reflect release of pore-water pressure from a liquefied unit, possibly following a shock.

Lobes on the flanks record the flowage of liquefied sand. The convex downslope ends show that the sandflow must have stopped abruptly because of **de-watering** as it flowed. Preservation of lobes suggests that the extrusion of sand volcanoes took place in quiet conditions, otherwise the sand would have been reworked by waves or currents. In some examples, extruded in

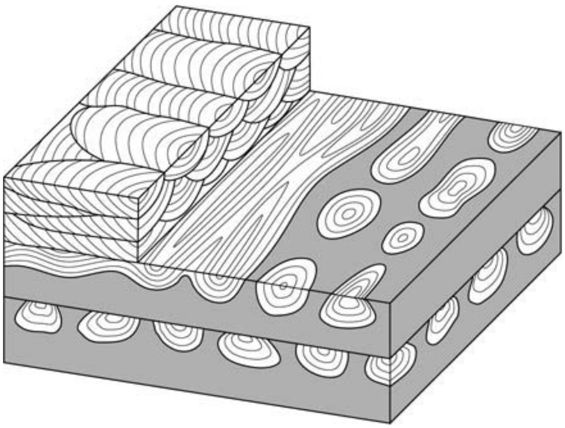


Figure 9.7 Schematic illustration of the generation of isolated load balls (pseudo-nodules) through the lateral passage of a continuous sand-bed overlying mud and into a progressively more intensely deformed state, culminating in the isolation of sand pseudo-nodules that are completely enveloped in mud and produce a ball and pillow structure. Note that the primary internal laminae are still evident, albeit in a deformed state. In this case, overlying trough cross-stratified sets of sand have provided the loading mechanism necessary for generation of the pseudo-nodules. (Modified after Allen 1982)

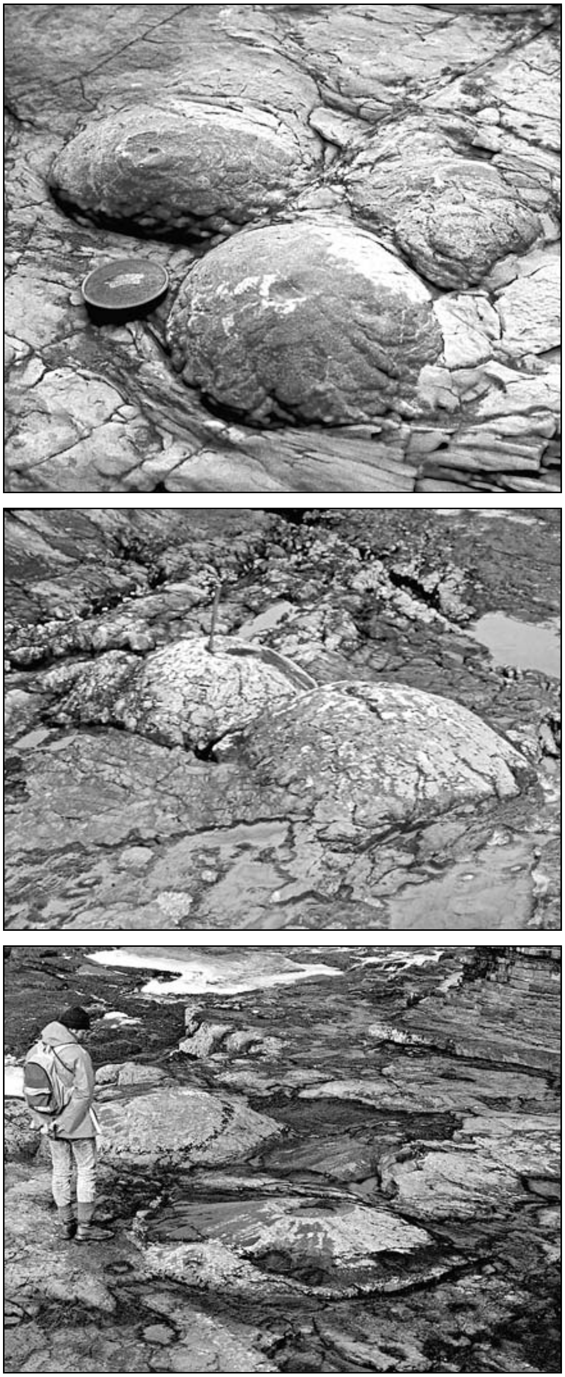


Figure 9.8 Sand volcanoes on the upper bedding surface of a sandstone bed; all examples from the Ross Formation, Namurian, County Clare, Ireland.

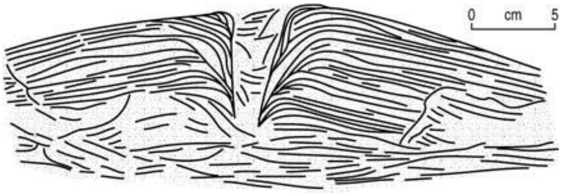


Figure 9.9 Internal structure of a sand volcano, showing the central plug and the inclined bedding that could, without care, be mistaken for cross bedding. Ross Formation, Upper Carboniferous, County Clare, Ireland.

settings with higher energy, reworking and partial erosion may be identified. In vertical section, the lenticular shapes of sand 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).

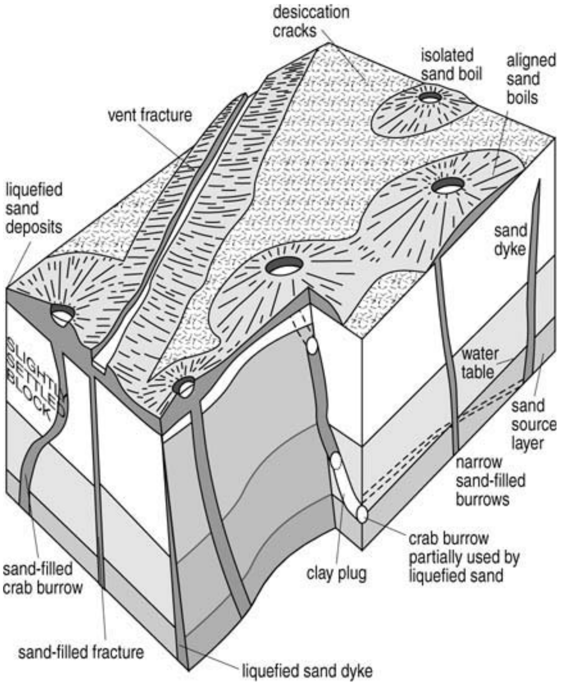

desiccation cracks
isolated sand boil
aligned sand boils
vent fracture
liquefied sand deposits
sand dyke
SLIGHTLY SETTLED BLOCK
water table
sand source layer
narrow sand-filled burrows
crab burrow partially used by liquefied sand
clay plug
sand-filled crab burrow
sand-filled fracture
liquefied sand dyke

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 & Santis 1991)

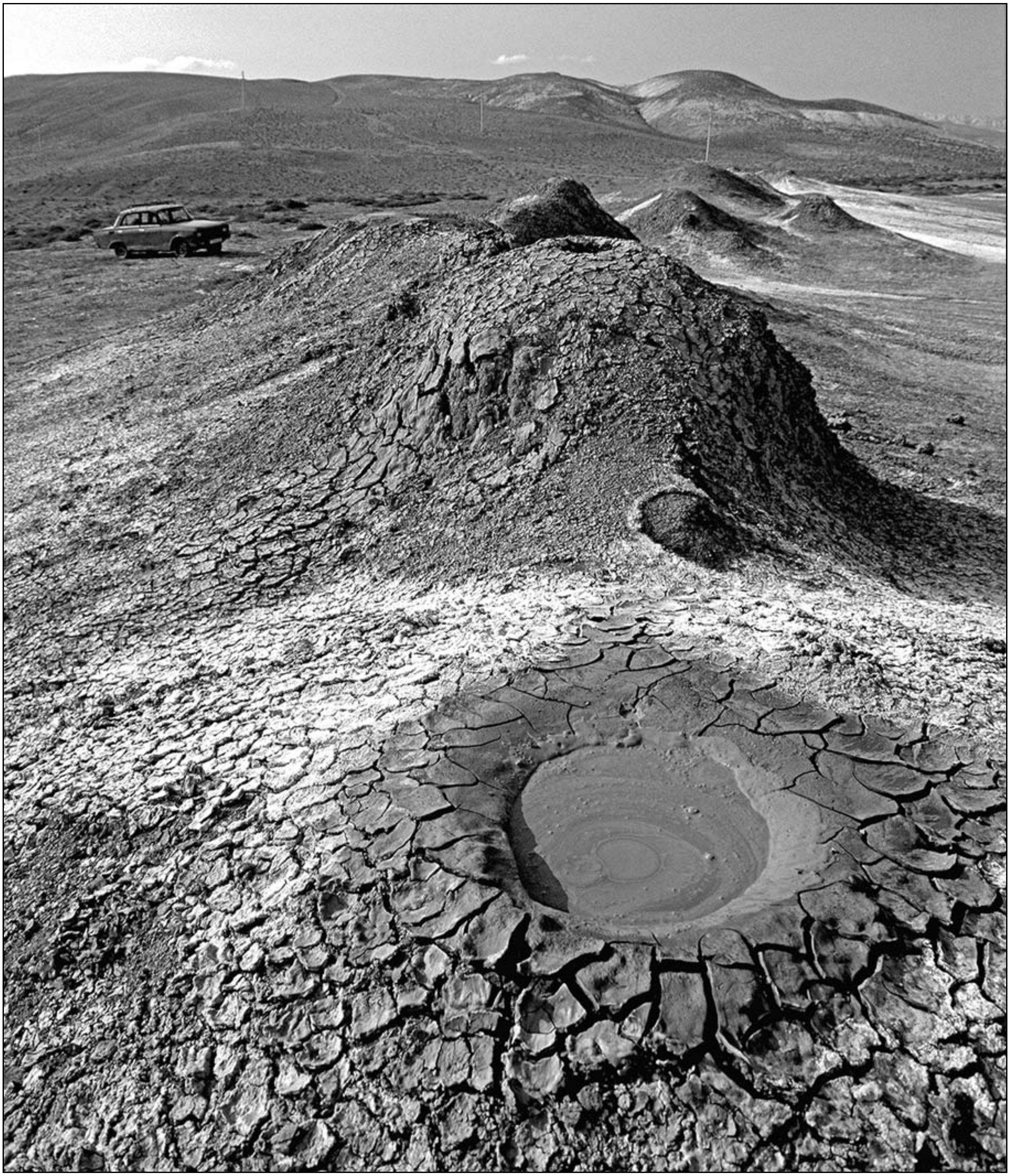


Figure 9.11 Mud volcanoes, Azerbaijan (photo courtesy of Martin Bochud).

They are 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

Desiccation mudcracks are common on the floors of dried-up ponds, lakes and playas, on river floodplains and on muddy intertidal and supratidal flats, where they are often open fissures or are filled only partially by other sediment (Fig. 9.12). In rocks they occur on the bedding surfaces of interbedded sandstone/mudstone sequences and less commonly in thinly bedded carbonates.

In rocks the cracks occur in muddy sediment and are infilled by coarser-grain material, usually sandstone. The cracks commonly form polygons from centimetres to metres in diameter, with different size populations sometimes present on the same surface. Although most crack patterns are broadly hexagonal, many polygons are quadrilaterals or triangles. The cracks in plan have parallel sides, and in vertical section they usually taper downwards (Figs 9.13, 9.14). In vertical section, this wedge shape is commonly complicated by folding. Crack widths range up to several centimetres, and depths up to several decimetres. On both present-day surfaces and 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 highly concave shape.

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 gives an 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

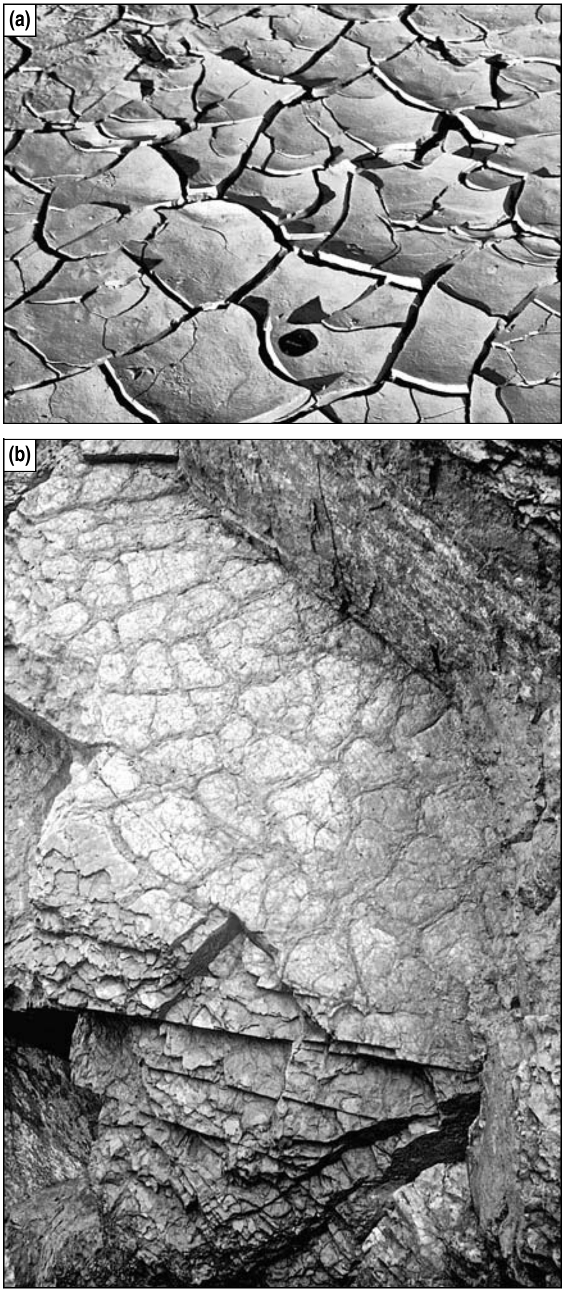


Figure 9.12 Desiccation cracks on bedding planes: **(a)** a mud surface with crack patterns of different sizes which have different depths of penetration; modern, central Australia; **(b)** ancient crack pattern on a bedding surface of interbedded sandstones and mudstones; polygons are 0.2–0.3 m diameter; West Bay Formation, Lower Carboniferous, Nova Scotia, Canada.

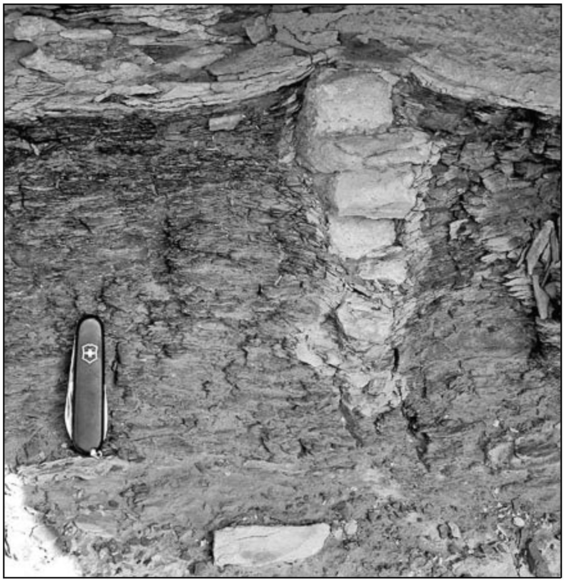


Figure 9.13 Desiccation crack in vertical section. The crack developed in mud and was subsequently filled by sand. Note the differential compaction of the mudstone around the crack. Cedar Mesa Sandstone, Permian, southeast Utah, USA.

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 desiccation.

Filling of cracks takes place later, for example by the influx of a sediment-laden flood or by windblown sand becoming trapped in cracks (Fig. 9.15). Mudflakes, derived from 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.

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 saltmarshes. They also occur in mudstones interbedded with sandstone and also 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. 9.16a) or in vertical sections through muddy layers, where their downward-tapering sandy fills may be contorted by small-scale compactional folding. They can also occur as negative relief features on upper bedding surfaces.

In plan, sub-aqueous shrinkage cracks tend to have irregular or radiating patterns, sometimes cross cutting one another. Individual cracks are lenticular, pinching out rather than joining with other cracks (Fig. 9.16b).

Sub-aqueous shrinkage cracks result from loss of pore water from the sediment, because of a reorganization of originally highly porous clay particles, either through 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 attributable to earlier inhomogeneities, such as burrows, in the host sediment. These cracks can be indicators of way-up and can suggest post-depositional conditions. As way-up indicators, they are not as good as desiccation mudcracks, because intercrack areas are flat.

Synaeresis cracks may be confused with desiccation cracks, sandstone dykes and elongate gypsum pseudomorphs. They have also 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 compared with the base of a U-shape burrow. The compactional folding of sandstone dykes has also led to confusion with organic traces.

Sandstone and mudstone dykes, and transposed sandsheets

With these structures, crack-filling material is most commonly sand, but the host sediment may be anything from mud to coarse-grain gravel. Although relatively uncommon, dykes occur in a wide variety of settings from deepwater sandstone/mudstone sequences to subaerial mass-flow conglomerates. They are seen on both bedding surfaces and in vertical section. **Transposed sandsheets**, which are sometimes 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.

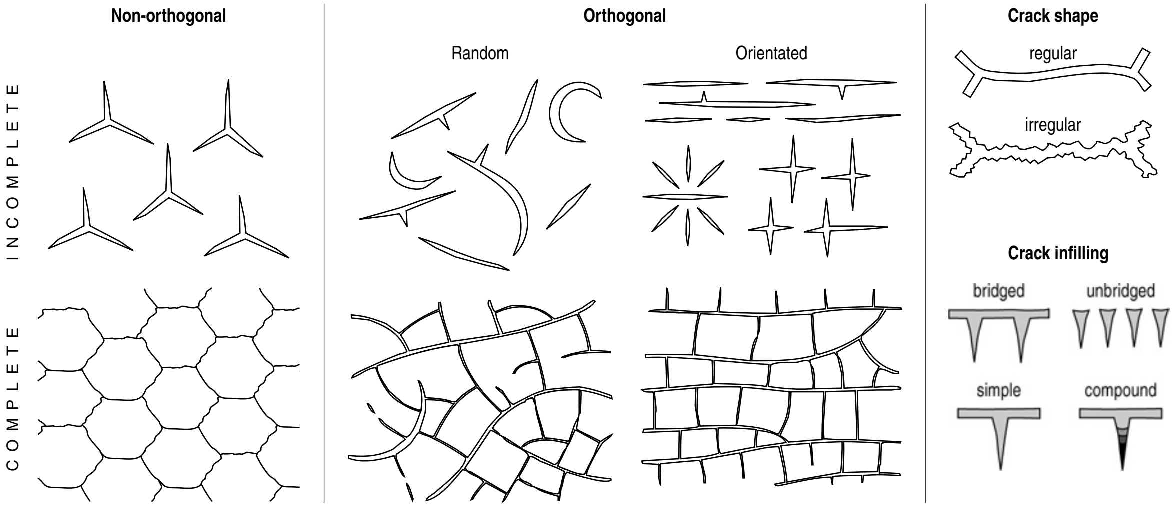

Non-orthogonal
Orthogonal
Random
Orientated
Crack shape
regular
irregular
Crack infilling
bridged
unbridged
simple
compound
INCOMPLETE
COMPLETE

Figure 9.14 Classification of shrinkage cracks and their infillings (modified after Allen 1982).

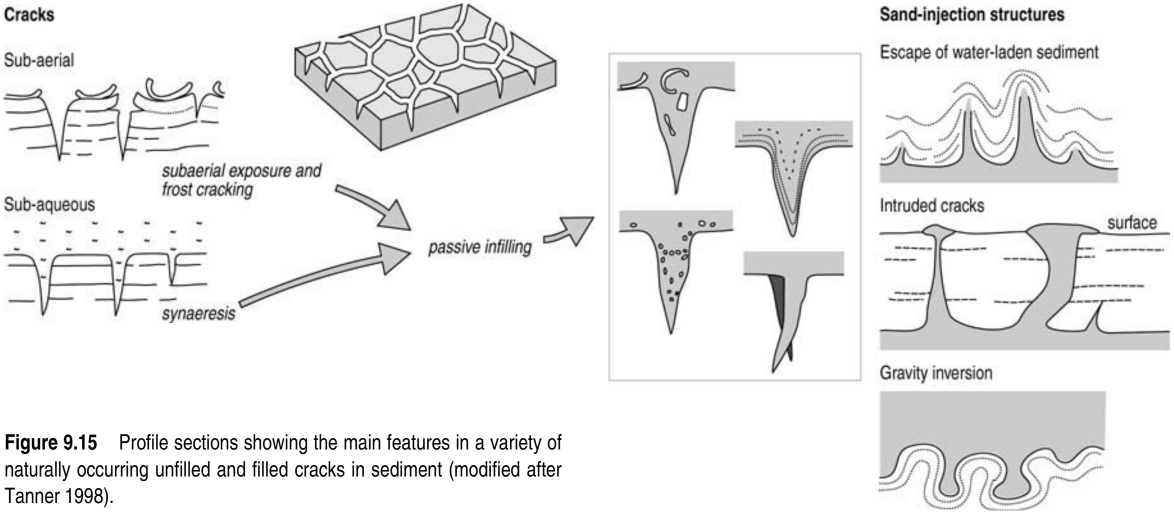
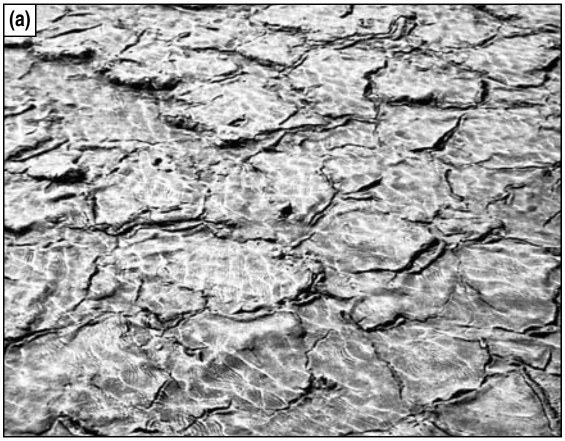

Cracks
Sub-aerial
subaerial exposure and frost cracking
Sub-aqueous
synaeresis
passive infilling
Sand-injection structures
Escape of water-laden sediment
Intruded cracks
surface
Gravity inversion

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).

Dykes vary in size, with widths up to several tens of centimetres and vertical extents up to several metres. As with other cracks the fills may be folded, particularly if the host sediment is of finer grain. Dykes can sometimes 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 weathering of the dyke and host lithologies. Dykes tend to be straight and have rather parallel sides (Fig. 9.17) and, although usually random in orientation when 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 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



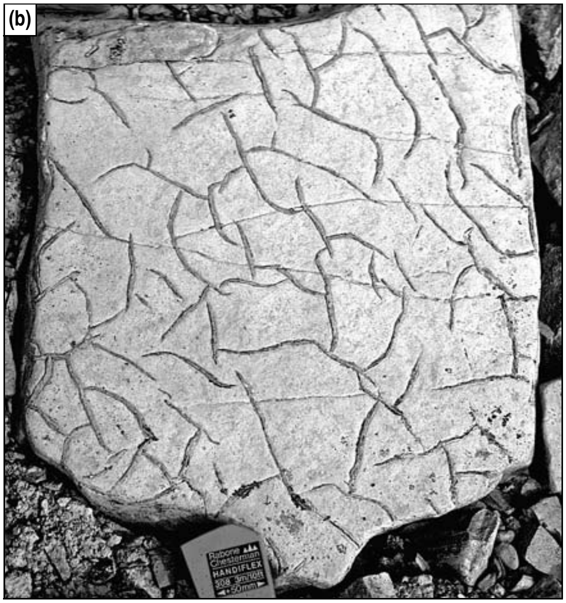
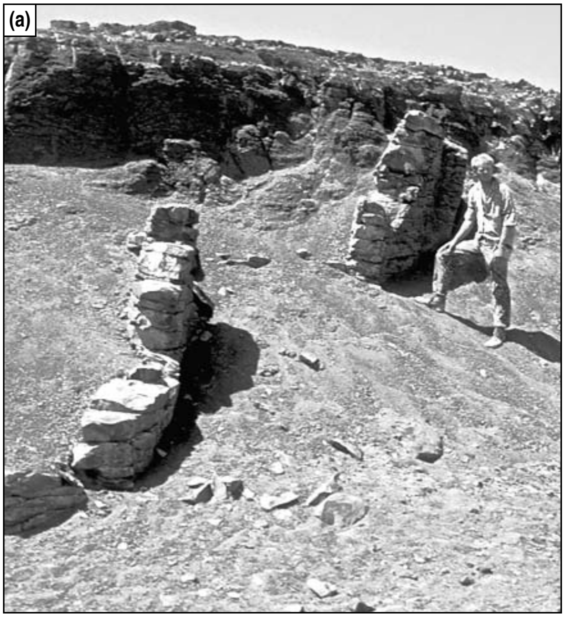


Figure 9.16 Examples of sub-aqueous shrinkage cracks resulting from synaeresis: **(a)** on the floor of a shallow-water lagoon (water depth 0.2 m); modern, Las Salinas, Almeria, Spain; **(b)** on the upper surface of a calcareous mudstone; Cass Fjord Formation, Cambrian, northwest Greenland.



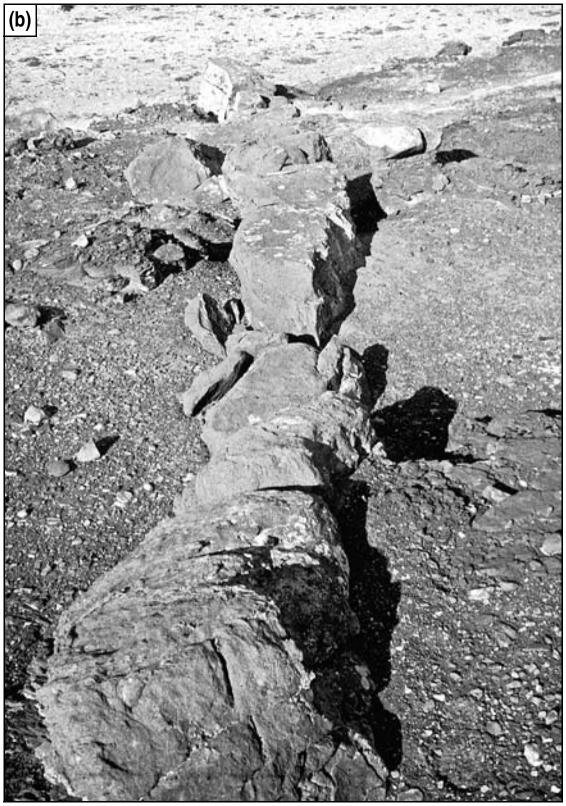


Figure 9.17 Examples of large-scale sandstone dykes: **(a)** developed in volcanic rocks and filled with aeolian sandstone derived from below; Etjo Formation, Cretaceous, Namibia; **(b)** cutting through a poorly sorted conglomerate of probable mudflow origin; a weak lamination parallel to the walls of the dyke is produced by shearing of the liquefied sand during its intrusion from below; Moræneso Formation, Proterozoic, north Greenland.

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 disturbed lamination.

Sediment dykes and sills result from the injection of sediment from an underlying, or more rarely overlying, source bed (Fig. 9.15). This appears often to 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 of fine grain (e.g. sand or silt) and was laid down in relatively quiet conditions, an external shock may have been needed to liquefy the source bed temporarily. Where the host sediment is of coarser grain, 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-grain 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.

Complex assemblages of intruded sandstone sheets, involving both dykes and sills intruded into mudstones, are found in some ancient deepwater sediments. These are most commonly encountered in borehole cores and mudstones typically overlie thick bodies of massive sandstone, often channel fills. These intrusions probably took place significantly later than initial sand deposition and it seems likely that the liquefaction of the sand resulted from a gradual overpressuring driven by the compactional de-watering of the surrounding muds.

Lamination within the sandstone dyke 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 sandsheets occur on some modern debris flows and are occasionally preserved in the rock record. Their undulating bases reflect a tendency of extruded sand to sink back into the still-mobile debris-flow sediment by loading.

In poor exposure, sandstone dykes can be confused with either type of shrinkage crack, although this is less likely if the host sediment is coarse grained. Intruded sills could be confused with extruded sandsheets or with normally interbedded sandstones. Internal lamination is good evidence of injection, but confusion can 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 crack are also widely recognized in 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 3 m 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. 9.18).

In vertical section, ancient examples commonly penetrate vertically for several metres and show an upwards-flaring wedge shape that tapers downwards to a fairly sharp lower end. The cracks usually show quite complex patterns of filling, often with several phases of contrasting lithology occurring in zones roughly parallel to the sides of the cracks (Fig. 9.19). 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.

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. The recognition of ice wedges is of considerable palaeoclimatic significance, and care should be taken to distinguish these features from other types of cracks.

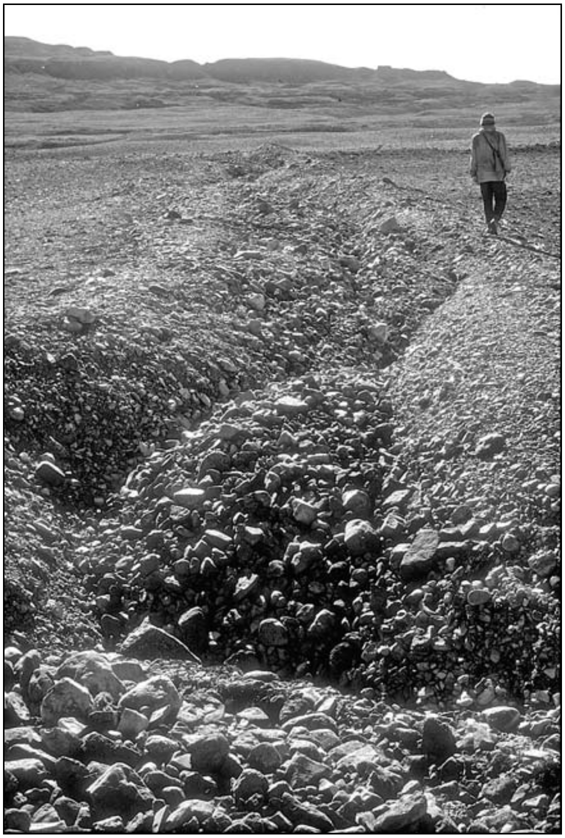


Figure 9.18 Present-day ice wedge with ramparts on either side. Washington Land, north Greenland.

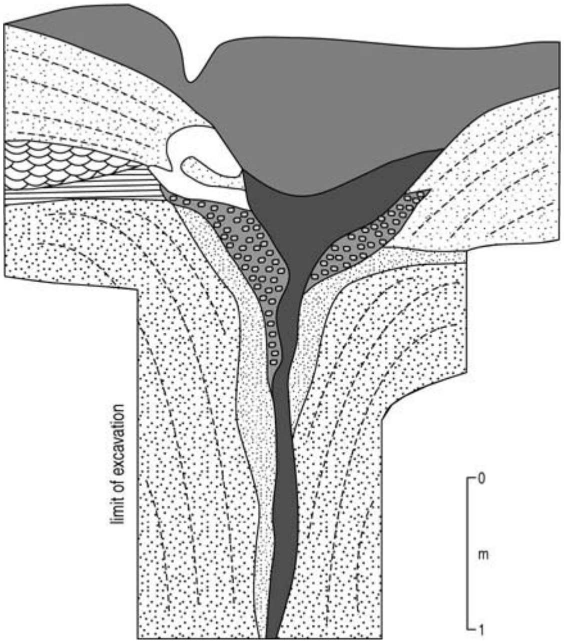
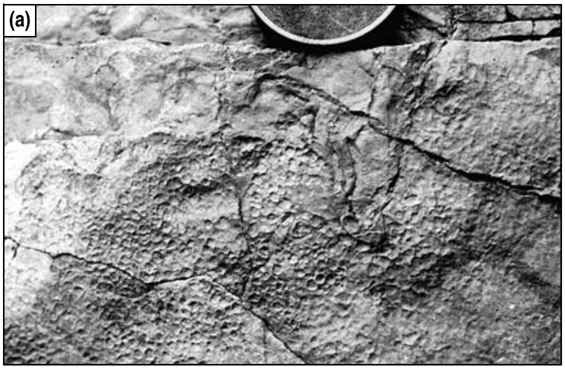


Figure 9.19 Cross section through an ice wedge in Pleistocene sediments, exposed in a gravel pit; note the downward tapering of the wedge and the various zones of infill (after Gruhn & Bryan 1969).



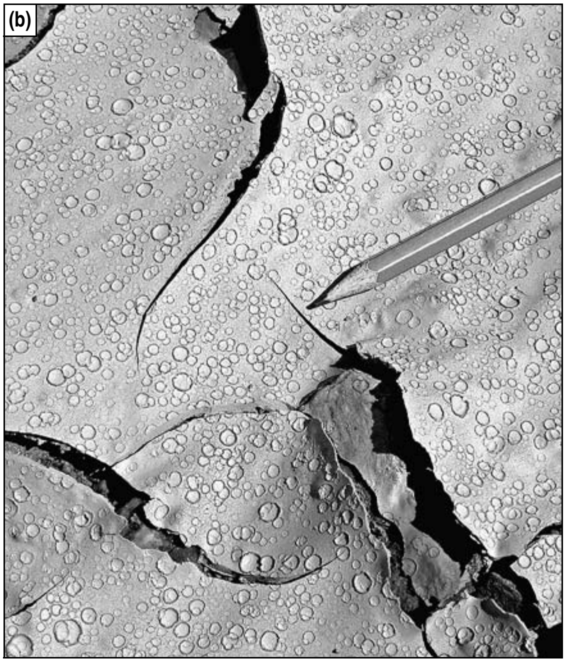


Figure 9.20 Examples of raindrop impressions (rainpits): **(a)** the upper bedding surface of a siltstone; West Bay Formation, Lower Carboniferous, Nova Scotia, Canada; **(b)** modern rainpits on a mudstone surface; Tabernas, southeast Spain.

Raindrop impressions

Some upper bedding surfaces in ancient mudstones, siltstones and sandstones, and also some present-day muddy or sandy surfaces, show patterns of shallow pits. These are commonly associated with desiccation mudcracks. Pits may be widely separated or may completely cover the surface. They are circular or, rarely, elliptical, up to about 1 cm in diameter and up to a few millimetres deep (Fig. 9.20). Where they completely cover a surface, they are polygonal. They have a slightly raised rim and the floor of each pit gives a smooth concave-upwards crater-like form.

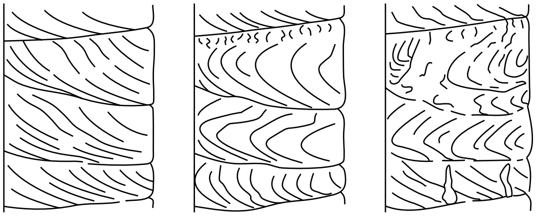
Large raindrops and hailstones impact with considerable force and produce small craters on damp sediment. Preservation in the rock record is likely only when the sediment is muddy, as this will have the cohesive strength to retain the impression when it dries out. On drying, sandy surfaces are reworked 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.

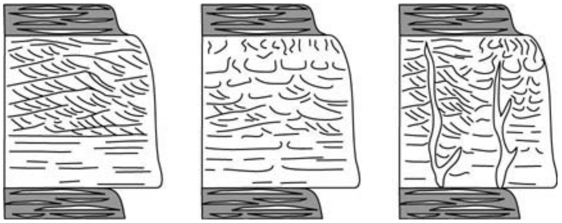
9.2.2 Disturbance within individual beds

Disturbance within individual beds is most commonly seen in vertical section, although some structures also have expression in plan view. The structures include those attributable to deformation of primary depositional lamination, as well as new structures developed entirely by post-depositional activity. Thickness of disturbed units ranges from centimetres to metres, and all or only part of a bed may be affected. Both the substrate type and the rate of fluid escape control the style of deformation (Fig. 9.21).

(a) Medium to coarse-grain cross-stratified sandstone



(b) Alternating beds of non-cohesive sandstone and silty mudstone



(c) Alternating beds of cohesive muddy sandstone and silty mudstone

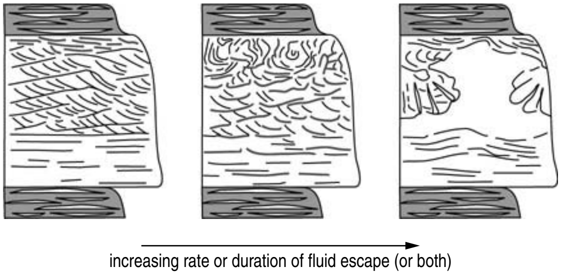


Figure 9.21 Schematic illustrations showing variations in water escape structures as a function of the original sediment characteristics and the rate of fluid escape. Examples **(b)** and **(c)** are common in turbidites, with the fluid escape structures being most pronounced in Bouma units C and D (ripple cross-laminated and planar-laminated sandstones and muddy sandstones). For each row, the rate of fluid escape increases from left to right, and gives 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).

Oversteepened and overturned cross bedding

Overturned or oversteepened cross bedding is common in medium- or finer-grain sandstones laid down in a variety of aqueous environments, although it is also known from aeolian settings. It occurs mostly within single sets. The deformation ranges from foresets that dip more steeply than the angle of rest in the upper parts of sets (oversteepened) to extensive overturning of foresets into recumbent folds (Fig. 9.22). 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. 9.23).

For simple oversteepened or overturned foresets, the processes are 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 had lost strength. In subaqueous situations, the shear force seems usually to have been the water current that produced the cross bedding in the first place. The sand was probably weakened by partial liquefaction, which is quite easily achieved 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 fold shapes observed 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 producing convolute lamination (see below) may also have contributed. When deformational structures include minor faulting and patches of structureless sand, the cross bedding may have originally been aeolian (Fig. 9.23).

Convolute bedding and convolute lamination

Convolute bedding and convolute lamination 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

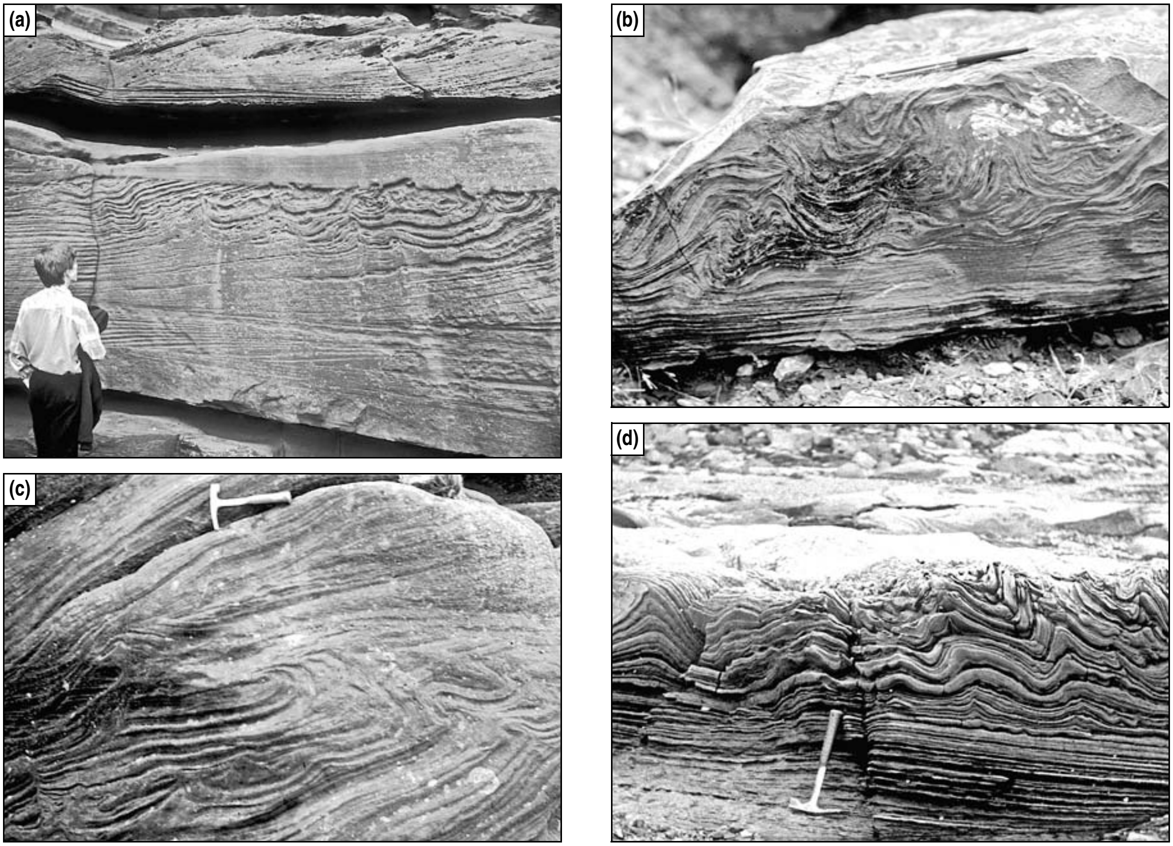


Figure 9.24 Examples of convolute bedding and lamination. **(a)** Convolute bedding within the upper part of a cross-bedded sandstone bed; St Bees Sandstone Formation, Triassic, Cumbria, England. **(b)** Convolute lamination within ripple cross-laminated limestone; Izroutene Beds, Carboniferous, Morocco (photo courtesy of Gilbert Kelling). **(c)** Convolute bedding within cross-bedded sandstone; the sharp, upwards-pointing folds suggest upwards fluid escape; Roaches Grit, Upper Carboniferous, Staffordshire, England. **(d)** Convolute bedding; Lower Carboniferous, Dunbar, Scotland.

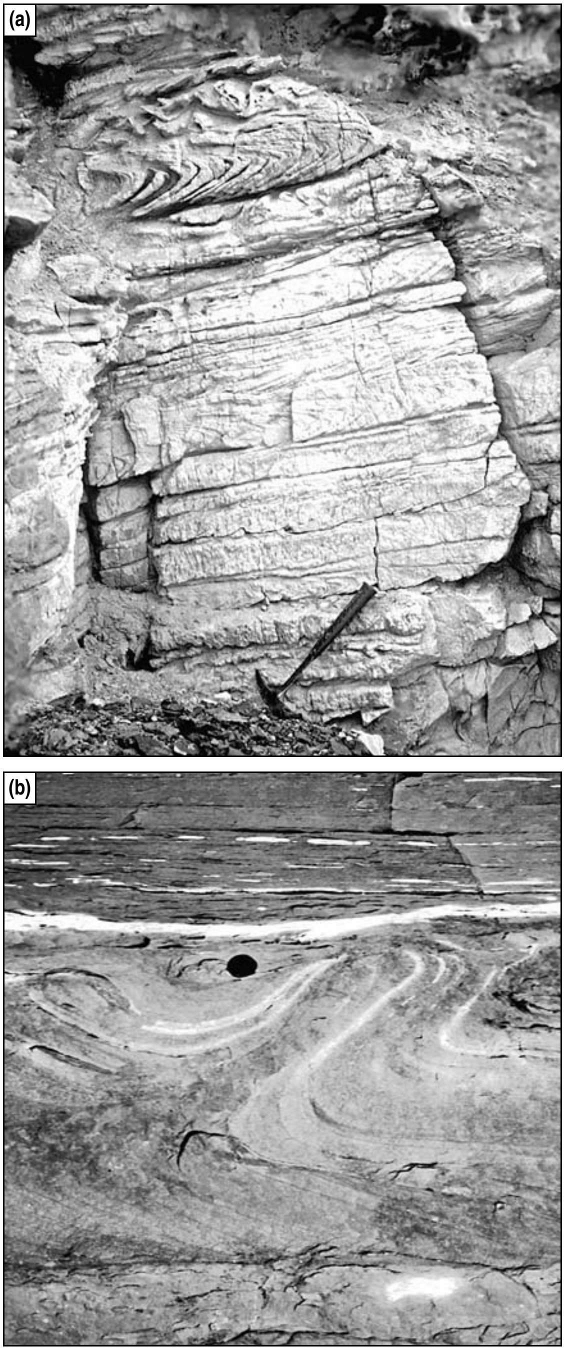
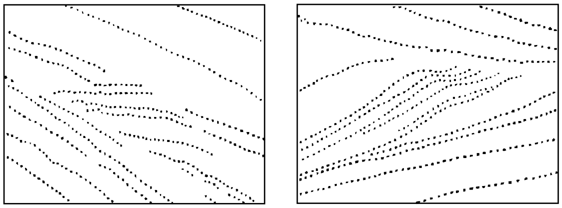


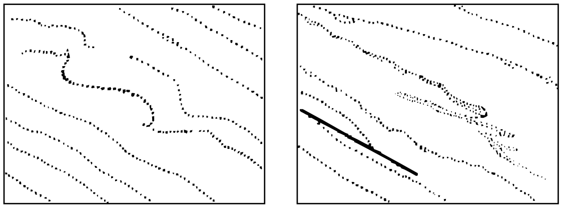
Figure 9.22 Examples of overturned bedding: **(a)** cross bedding with overturned foresets; Kap Holbæck Formation, Lower Cambrian, northeast Greenland; **(b)** overturned bedding in aeolian interdune strata attributable to loading by overlying dune; Cedar Mesa Sandstone, Permian, Utah.

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 larger, the latter at the scale of centimetres. 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 (Fig. 9.24). Overturning of fold axes is sometimes seen, often 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. At the top it may either die out gradually or be sharply truncated. On upper bedding surfaces, convolute lamination commonly takes the form of a complex pattern of basins and ridges (Fig. 9.24a).


(a) Rotated structures
(b) Drag folds


(c) High-angle asymmetric folds
(d) Overturned folds

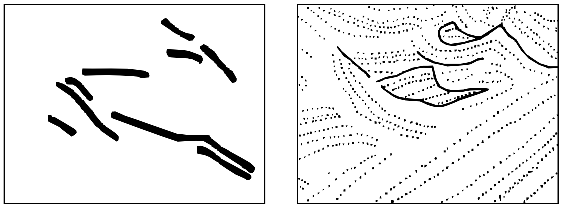

(e) Break-apart structures
(f) Brecciated foresets

Figure 9.23 Principal types of deformation structures in the foresets of dunes (modified after McKee 1979).

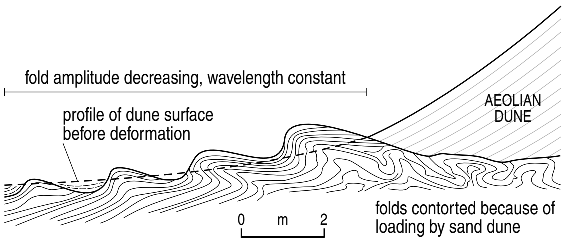

fold amplitude decreasing, wavelength constant
profile of dune surface before deformation
AEOLIAN DUNE
0 m 2
folds contorted because of loading by sand dune

Figure 9.25 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 Collinson in Maltman 1994).

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 often 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. 9.25). 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 (Fig. 9.26).

The main use of convolute lamination is as evidence of rapid deposition. It has limited potential as an indicator of way-up and as a rather uncertain palaeocurrent indicator when folds have a preferred direction of overturning.

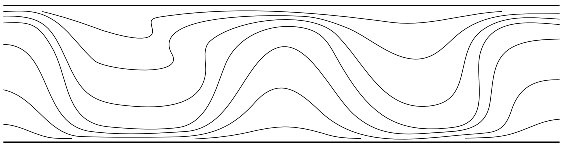
Dish, pillar and sheet de-watering structures

The significance of dish, pillar and sheet de-watering structures has been recognized only quite 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 require a certain amount of clay for their formation. 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 stand out clearly. Only after seeing good examples is it realistic to look for these structures in less favourable exposures. Sheet de-watering structures are rather more obvious and common.

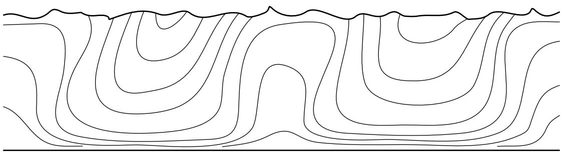
Dish and pillar structures commonly occur together in the same bed. In vertical section, **dish structures** appear as thin, roughly horizontal zones, often flat but more usually concave upwards (Fig. 9.27). Each is defined by a dark clay-rich zone 0.2–2 mm thick, which contrasts with the paler host 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 that define the dishes commonly have a vertical spacing of a few millimetres up to about 10 cm.

Flat zones, lacking the clear dish shape, have upturned ends associated with penetration of layers by **pillar structures** (Fig. 9.27b). These extend vertically

(a) Post-depositional convolute lamination



(b) Meta-depositional convolute lamination



(c) Syn-depositional convolute lamination

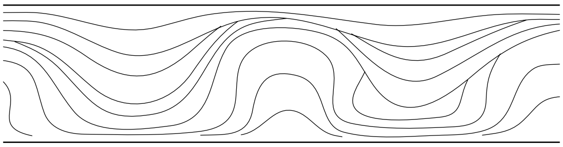


Figure 9.26 Types of convolute lamination (after Allen 1982).

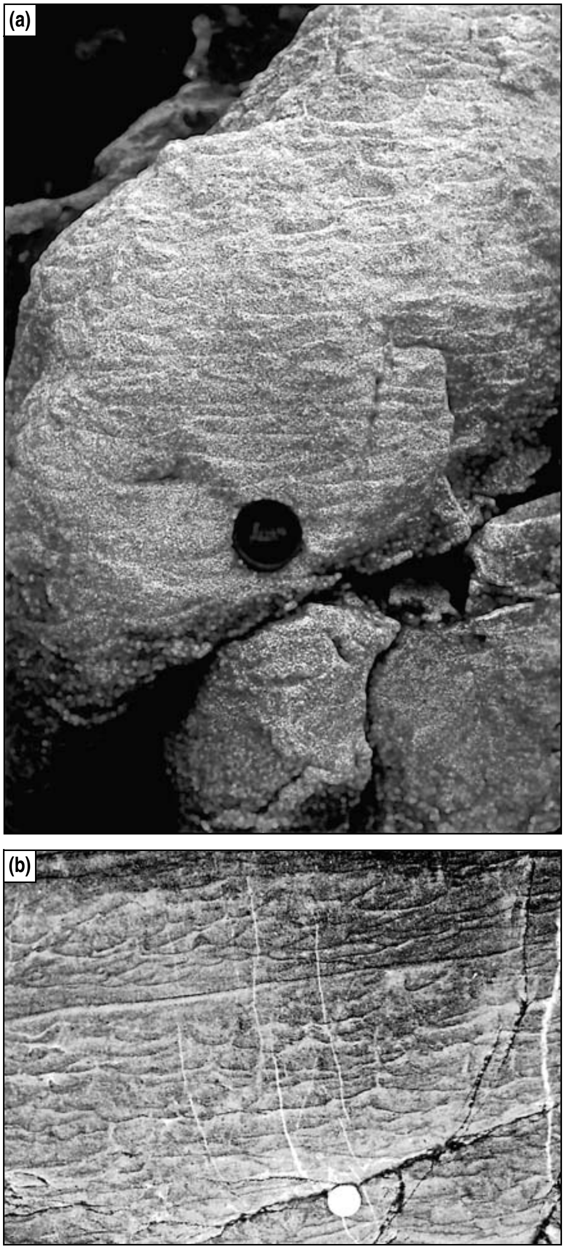


Figure 9.27 Examples of dish structures: **(a)** flysch deposits, Eocene, San Sebastian, Spain; **(b)** dish structures penetrated by vertical pillars and pipes that acted as conduits for vertical water escape; Ordovician, Gaspe, Canada.

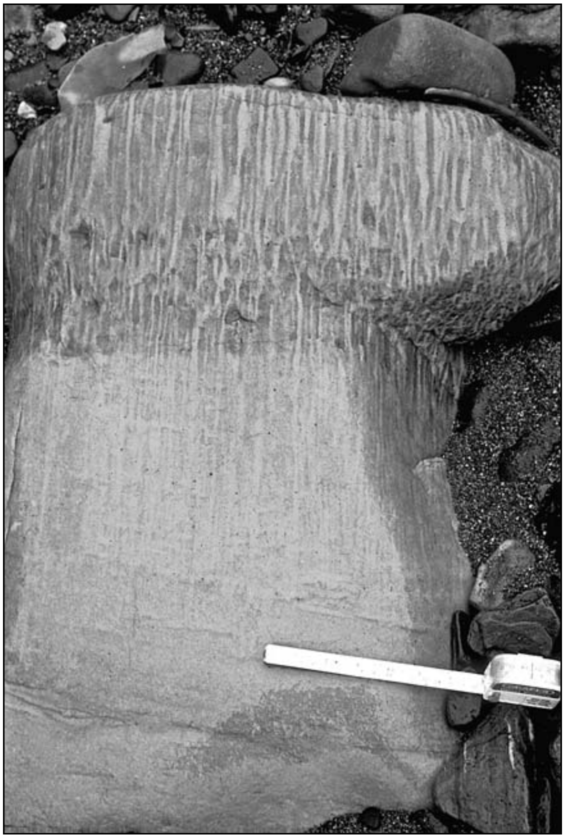


Figure 9.28 Sheet de-watering structures in sandstone; Lower Cretaceous, Spitzbergen.

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 de-watering structures are subvertical 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. 9.28). 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. 9.29). The sequence of deformation events may begin with the development of mild convolute

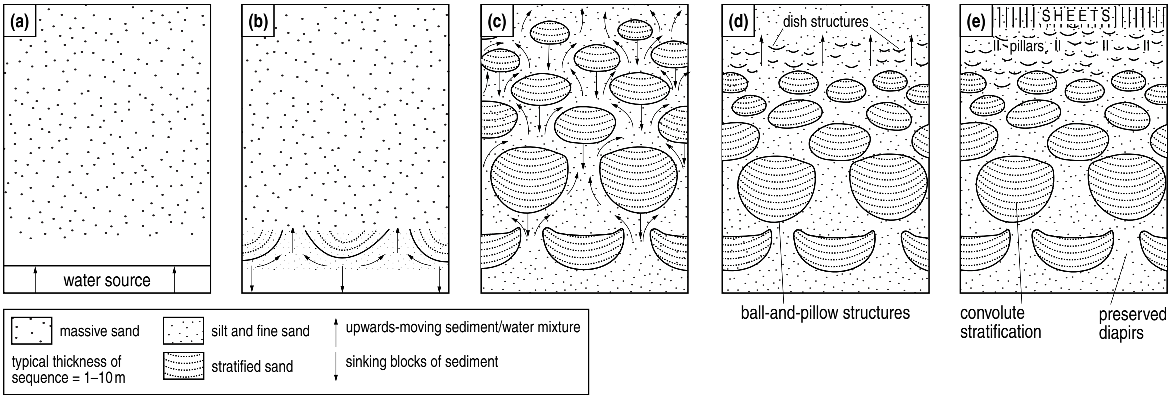


Figure 9.29 A model for the progressive development of convolute bedding, ball-and-pillow structures and dish structures. Excess pore fluid pressure drives the upward escape of sediment and water. Gravity assists with the downward movement of larger intact blocks. (Modified after Cheel & Rust 1986)

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.

The three types of structure are all the result of the post-depositional escape of water from the sand. The vertical structures (the sheets and pillars) are conduits of water expulsion, roughly similar to sandstone dykes and the feeder pipes of sand volcanoes (see Fig. 9.10). However, pillars and sheets record the movement of water with only restricted and selective movement of sediment. The force of upwards movement of water causes local fluidization of the sediment through which it passes and also the selective transport of fine particles; this results in the slightly cleaner sand of 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 laminations. The upwards movement of water through the loosely packed sand soon after deposition is retarded by the laminations, and further 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 upwards escape, at a pillar or the upturned edge of a dish. This water movement probably takes place very soon after, or even during, deposition. Rapid deposition probably gives 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, when present, the upward concavity of dishes could be helpful. The absence of dish and pillar structures in many massive sandstones probably reflects a lack of suitable clay, rather than little part having been played by 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 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.

9.2.3 Disturbance affecting several beds

Some examples of larger-scale deformation structures that affect several beds 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. With large-scale disturbance there is often difficulty in establishing the timing of deformation and in distinguishing tectonic from sediment-induced deformation.

Slumps: sedimentary folding

Units of folded sediment attributable to slumping usually occur in interbedded sequences containing a substantial proportion of fine-grain sediment. The sequences may be composed of detrital or carbonate material and the fine-grain sediment may be clay or lime mud. Slumped units occur on or at the bases of contemporaneous slopes in a variety of environments.

Slump-folded units vary from less than 1 m up to tens or even hundreds of metres in thickness. They are usually bounded above and below by undisturbed sediment; this helps to distinguish them from tectonically disturbed beds (Fig. 9.30). Within the slumped unit, folds may have preferred orientations that, when properly plotted on a stereonet (see Appendix 1), 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 the beds within the unit. Systematic recording and comparison of the orientations of fold axes and axial planes with those of any tectonically produced structures (folds, cleavage, etc.) should help to differentiate between tectonic and sedimentary disturbance.

Unconsolidated sediments resting on a slope may become unstable, possibly because of high pore-fluid pressure in a particular layer in the sediment pile. Sediments above the weakened layer may then move down slope under gravity as a coherent mass. In some cases a whole slab of material may detach and move. In other cases the downslope end may stay fixed as the upslope end moves towards it. In both cases there are significant differences between the behaviour at the upslope and downslope ends. The upslope end is subject to a dominantly tensional stress regime, while at the downslope end compressive stress dominates. However, the deformation recorded in slumped sediments is generally complex and must be interpreted with some caution. Slump folds often reflect a compressive regime in a downslope position with fold axes normal to the direction of movement. However, folds also result from lateral compression and also from internal shearing between parts of the slump moving at different rates. Internal 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

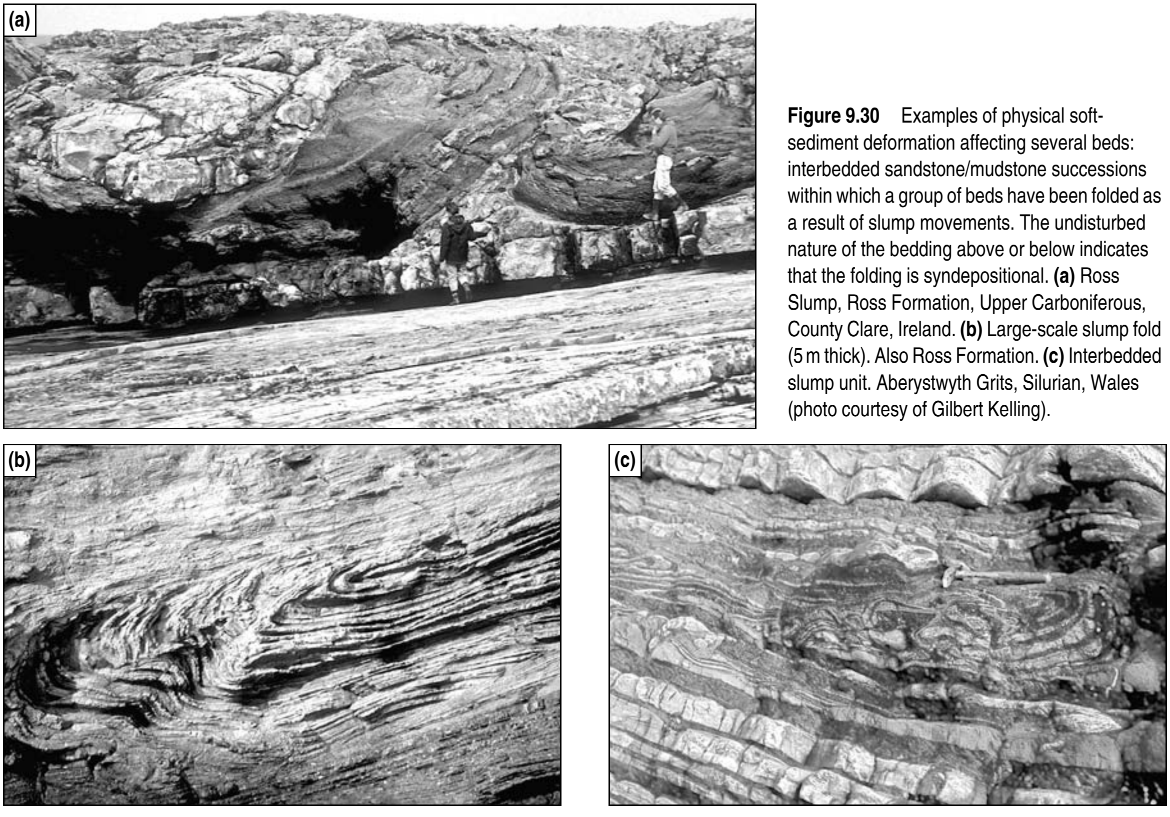


Figure 9.30 Examples of physical soft-sediment deformation affecting several beds: interbedded sandstone/mudstone successions within which a group of beds have been folded as a result of slump movements. The undisturbed nature of the bedding above or below indicates that the folding is syndepositional. **(a)** Ross Slump, Ross Formation, Upper Carboniferous, County Clare, Ireland. **(b)** Large-scale slump fold (5 m thick). Also Ross Formation. **(c)** Interbedded slump unit. Aberystwyth Grits, Silurian, Wales (photo courtesy of Gilbert Kelling).

important to remember that most slumps are lobate in plan and that fold axes can be expected to show a spread of directions. Upslope zones of slumps, dominated by tensional strain, commonly show a different suite of structures, described below.

Slump folding helps our understanding of processes at or soon after deposition. The deformation produced by genuine slumping, involving lateral displacement of material, may be confused with that attributable to both vertical sinking (i.e. loading) and tectonic deformation. As a rule, vertical sinking produces little or no preferred orientation of fold axes. With tectonic deformation, folds commonly relate to or mirror the larger-scale structures of undoubted tectonic origin.

Rotation and displacement of coherent blocks

Displaced relationships between blocks of internally coherent sediment occur at various 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.

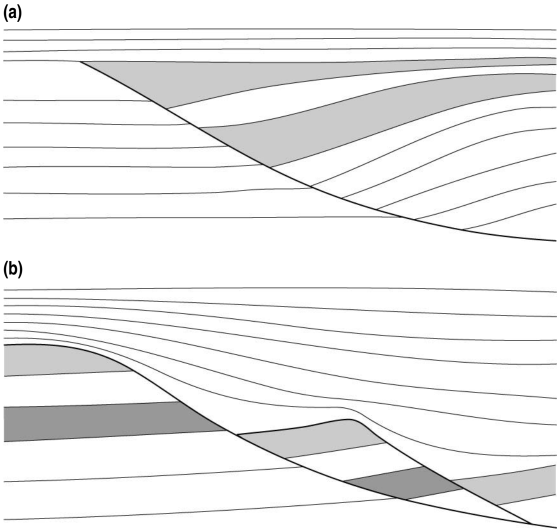


Figure 9.31 Schematic sections through synsedimentary faults. **(a)** The fault movement has continued during deposition. **(b)** Fault movement has occurred to the extent that a topography developed on the sediment surface. This was then draped by later deposits.

Features of this type are so variable that no comprehensive account is possible. A field description of any suspected example should be made systematically with several questions in mind. However, in most cases, all will not be answered fully and unambiguously.

- *Is the displacement of early, post-depositional or syndepositional origin or is it tectonically produced after lithification?* When syndepositional, the disturbance may be confined between undisturbed units above and below, and the planes of movement will be sharp with little or no brecciation. 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 topographical feature?* Draping of a topographical step by overlying sediment, lateral changes of thickness and the gradual upward elimination of relief should be looked for.
- *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 muddy or silty sequences, 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. 9.31).
- *Have the various blocks undergone any rotation in the course of displacement?* Careful comparison of dips on either side of surfaces of displacement will help (e.g. Fig. 9.32a,b).
- *Are there any minor structures such as drag folds and smaller faults that help to elucidate the nature of the movement?*

Be sure to distinguish those folds related to fault movement from those that pre-date the faulting. Some slumps that develop folds during ductile deformation are cut by faults caused by later brittle failure.

Synsedimentary faults occur for a variety of reasons. Most have a normal throw, indicating at least local tensional stress. It is not possible to give an exhaustive account of the ways in which these structures operate, but the following illustrations may be helpful.

Slump scars

It was suggested earlier that the upslope end of a slump sheet would be a zone of overall tension. If the sheet moved off in its entirety, it would leave an empty slump scar, which would then be draped by later sediment (Fig. 9.31b). However, at many slump scars a cluster of slip surfaces develops, and packets of sediment between these surfaces move only short distances, often rotating to dip up slope in the process (Fig. 9.32a).

Small-scale gravitational collapse features

Small-scale faulting in sands and sandstone can often be explained by the collapse of some buried objects in the sand. Logs, masses of vegetation or blocks of ice may rot or melt, and the sand then collapses into the resulting space.

Large-scale gravitational collapse structures – "olistostromes"

Large-scale structures relating to gravitational collapse and downslope movement are well documented from submarine slope settings, where the resultant deposits are often extensive enough 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, often of limestone, with individual clasts ranging from a few metres to hundreds of metres in diameter (Fig. 9.32c). These blocks, which are clast supported, usually sit chaotically arranged within a fine-grain, often muddy, matrix and commonly exhibit striated surfaces. Olistostromes are commonly found

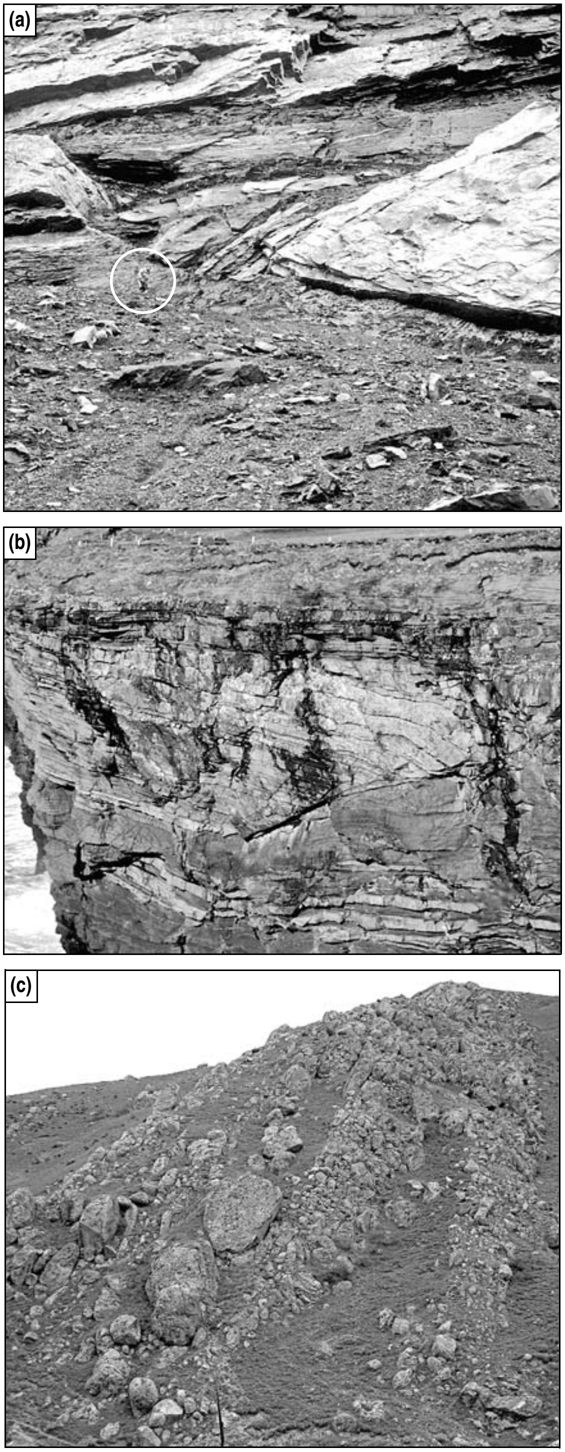


Figure 9.32 Examples of physical deformation due to the rotation or displacement of coherent blocks. **(a)** Large rotational slide of a thick sandstone bed. The thick sandstone on the left and its underlying fine-grain sediments are *in situ*. The block on the right has slid along a narrow, concave-upwards shear zone and has rotated back into the zone. Figure for scale. Lower Cretaceous, Kvalvågen, Spitsbergen. **(b)** Syndepositional growth fault; Central Clare Group, Upper Carboniferous, County Clare, Ireland. **(c)** Limestone blocks forming an olistostrome megabreccia generated by the collapse of the leading edge of a carbonate platform; Lower Cretaceous, Greater Caucasus, Azerbaijan. The largest two blocks are each approximately 12 m diameter.

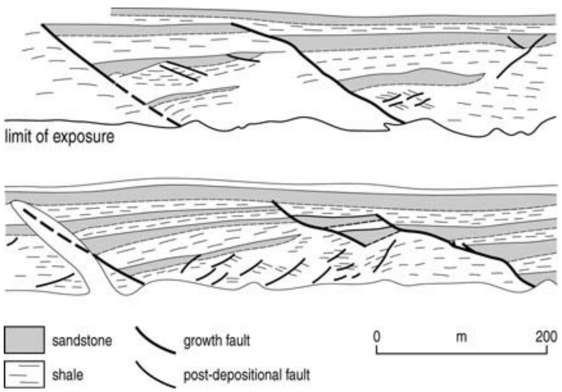


Figure 9.33 A system of growth faults exposed in two cliff sections on Edgeøya, Svalbard (after Edwards 1976).

interbedded within fine-grain turbidite successions. Where the blocks are smaller, they may sometimes be arranged into a crude layering, for which the term **olistolite** is sometimes used. Occasionally, units are composed of single very large-scale blocks called **olistoliths** or **olistoplates**, which have clearly undergone displacement from their site of original accumulation.

Olistostromes are a family of structures that represent the products of submarine slides driven by gravitational collapse. The large size of the blocks and their many striated surfaces show that they were effectively lithified prior to movement. Their occurrence interbedded with fine-grain turbiditic sediments suggests their development in marine slope or base-of-slope settings. The limestone composition of many of the blocks, which often contain shelly faunas and reef structures, indicates an original shallow-water depositional setting. Given the inferences above, 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.

Growth faults

Most growth faults occur on a scale too large to see at outcrop, and their recognition normally requires mapping or geophysical investigation. However, small examples have been recognized in extensive exposures (Figs 9.32b, 9.33). The typical pattern of displacement and thickness change is shown schematically in Figure 9.31a. Such faults are most commonly found in deltaic deposits, particularly those with a high proportion of fine-grain sediment. High depositional rates lead to high pore-water pressure in buried sediment and consequent loss of strength.

Where the over-pressured sediment is deeply buried, as beneath major deltas such as the Niger or Mississippi, it may flow towards the ocean like toothpaste, driven by 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. At this scale, the whole depositional–deformational interplay operates at timescales of millions of years and the dimensions are such that only seismic reflection profiles can be used to study the faults.

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 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 surface may be draped and a more uniform thickness pattern re-established.

The displacement of coherent blocks of sediment through brittle failure along fault planes often occurs in close association with plastic, liquefied 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. 9.34). Furthermore, faults may act as conduits for fluid escape and may develop into clastic dykes or feeder pipes for sand or mud volcanoes.

9.3 Chemically induced disturbance

Structures produced by post-depositional chemical activity are variable in their occurrence and mineralogy, and in the nature and timing of the chemical reactions. Three main types of process can be envisaged in ideal terms: precipitation of minerals from pore waters, reactions between host sediment and pore waters, and dissolution of sediment by percolating water. Although

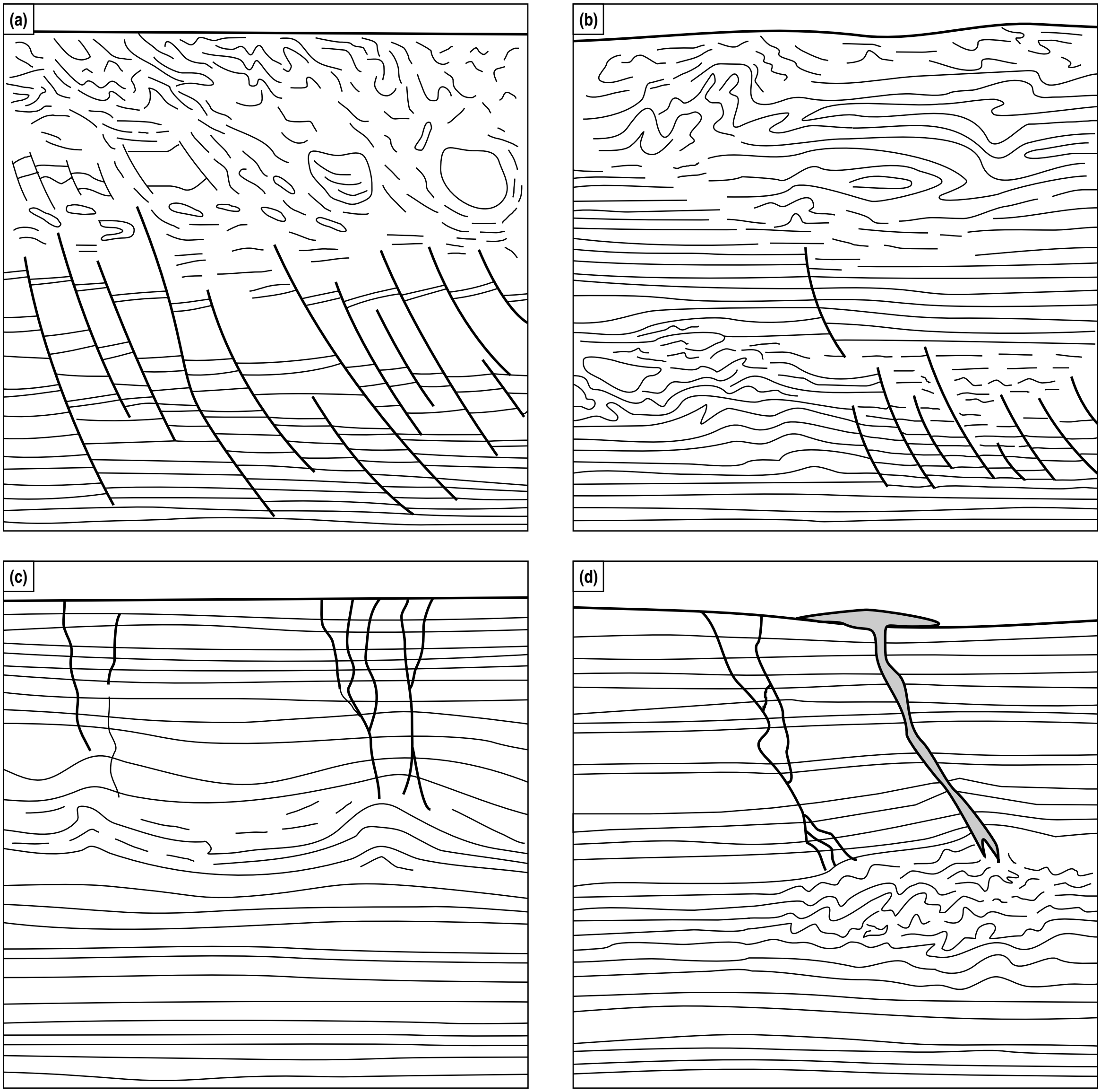


Figure 9.34 Soft-sediment deformational structures attributable to earthquake shock. These types of structures are typically developed in successions composed of laminated silts and sands. **(a)** Closely spaced faults with throws that increase upwards and that pass upwards into broken-up and then completely 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 upwards 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. Such structures could potentially occur on a variety of scales from a few centimetres to several metres. (Modified after Ringrose 1988)

this subdivision is theoretically acceptable, in practice there are real difficulties in distinguishing between the products of precipitation and reaction, and the structures that are produced by these processes are best considered together.

9.3.1 Products of precipitation and reaction (nodules and concretions)

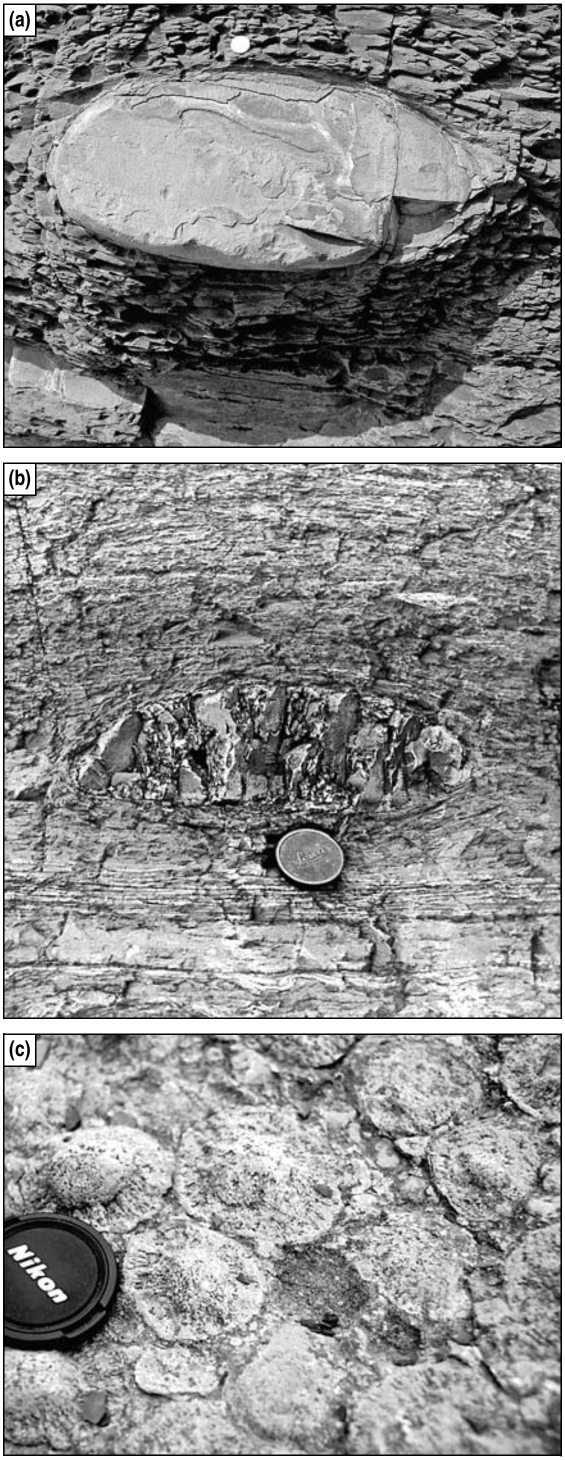
We are not concerned with general large-scale lithification and cementation, but with the local chemical precipitation and reactions that create structures commonly referred to as **nodules** or **concretions**. These two terms are used interchangeably. Nodules and concretions occur in host sediments 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. 9.35). They range in size from large masses, metres in diameter, down to small bodies of 1 mm 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 is causing the concretion to be localized? Is it following 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?

Precipitation versus reaction

Resolution of the precipitation versus reaction issue is not easy, particularly in the field, and in practice it may



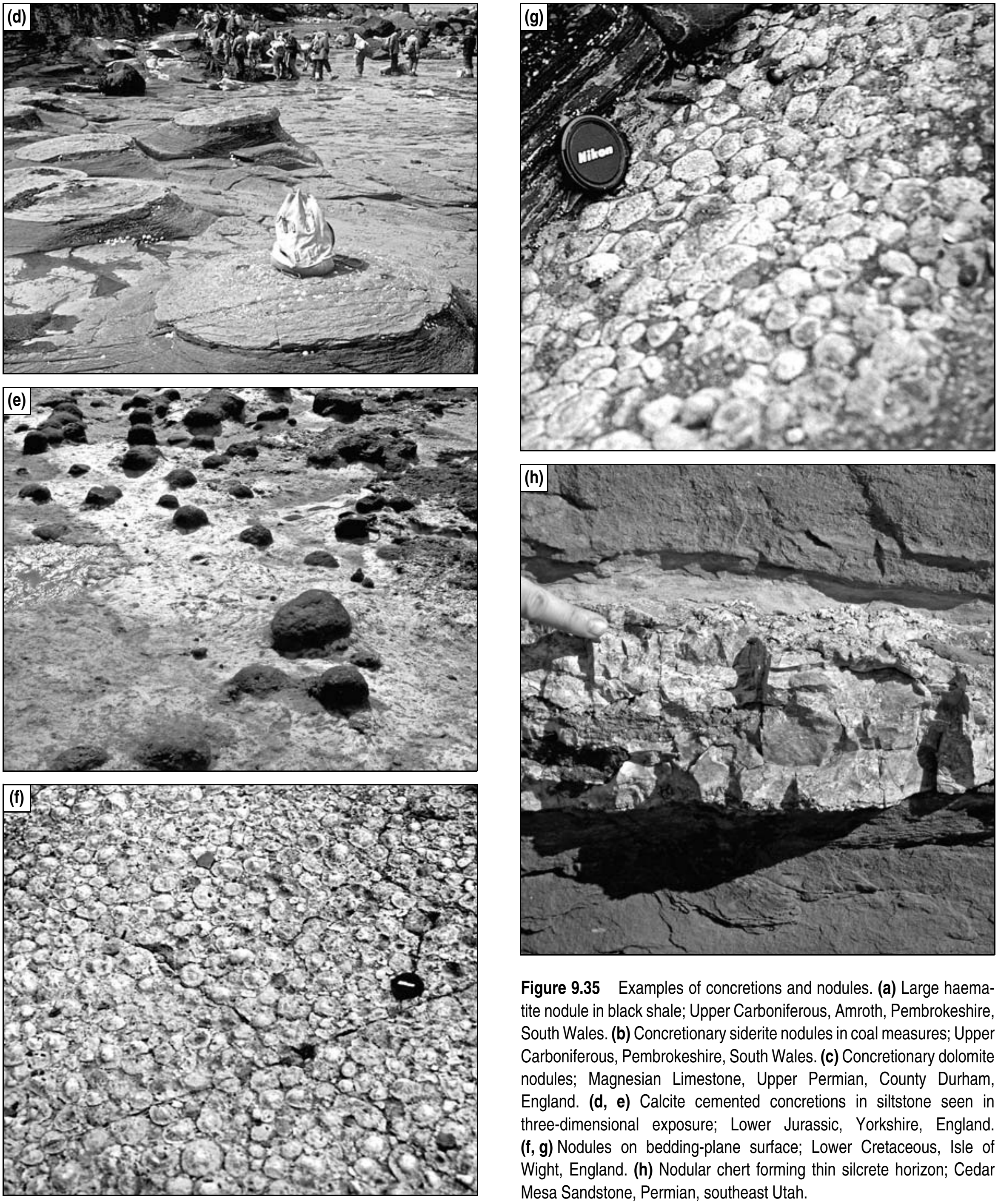


Figure 9.35 Examples of concretions and nodules. **(a)** Large haematite nodule in black shale; Upper Carboniferous, Amroth, Pembrokeshire, South Wales. **(b)** Concretionary siderite nodules in coal measures; Upper Carboniferous, Pembrokeshire, South Wales. **(c)** Concretionary dolomite nodules; Magnesian Limestone, Upper Permian, County Durham, England. **(d, e)** Calcite cemented concretions in siltstone seen in three-dimensional exposure; Lower Jurassic, Yorkshire, England. **(f, g)** Nodules on bedding-plane surface; Lower Cretaceous, Isle of Wight, England. **(h)** Nodular chert forming thin silcrete horizon; Cedar Mesa Sandstone, Permian, southeast Utah.

not always be important, as most concretions result from a combination of processes. Concretions or nodules in clastic sequences 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/alkalinity (pH) and oxidation/reduction potential (Eh) must also be appropriate. Figure 9.36 shows the general conditions under which different minerals form, but some caution should be used 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 determined. Below we comment briefly on the occurrence and significance of some common, concretion-forming minerals.

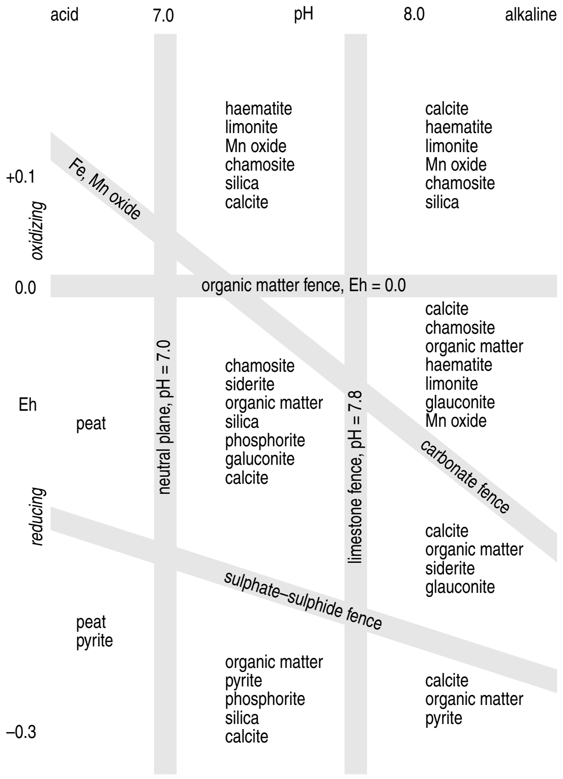

acid
7.0
pH
8.0
alkaline
+0.1
oxidizing
0.0
Eh
reducing
–0.3
Fe, Mn oxide
organic matter fence, Eh = 0.0
neutral plane, pH = 7.0
limestone fence, pH = 7.8
carbonate fence
sulphate–sulphide fence
haematite
limonite
Mn oxide
chamosite
silica
calcite
calcite
haematite
limonite
Mn oxide
chamosite
silica
calcite
chamosite
organic matter
haematite
limonite
glauconite
Mn oxide
peat
chamosite
siderite
organic matter
silica
phosphorite
galuconite
calcite
calcite
organic matter
siderite
glauconite
peat
pyrite
organic matter
pyrite
phosphorite
silica
calcite
calcite
organic matter
pyrite

Figure 9.36 Stability fields of commonly occurring authigenic minerals found in sedimentary rocks, in terms of prevailing Eh and pH. The chart applies for groundwater with an anionic composition close to that of sea water. (After Krumbein & Garrels 1952)

Calcite ($CaCO_3$)

Calcite is one of the most commonly occurring minerals. The calcium carbonate may 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.

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. 9.36). Evidence of early diagenetic (pre-compaction) origin is common, suggesting that the concretions formed soon after deposition. In mudstones of some coal-measure sequences, siderite occupies a substantial proportion of the 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. 9.35b).

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-grain mudstones, usually associated with preserved organic matter.

Both minerals reflect strongly reducing conditions within the sediment, or even at the sediment surface in a so-called "euxinic" environment. 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-grain parts of present-day tidal flats. Larger pyrite concretions commonly show an internal radial pattern of crystal growth.

Silica (SiO_2)

Nodules or concretions of silica occur both in carbonates, where the silica is commonly crypto-crystalline chert or flint (Fig. 9.35h), and in sandstones, where it occurs as quartz overgrowths on detrital quartz grains. Silica precipitation seems to require weakly alkaline conditions, but the precise controls and processes of flint and chert formation are poorly understood. In carbonates, chert is usually picked out by its darker colour and it is invariably a replacement of the carbonate host, rather than an intergranular precipitate. In sandstones, silica-cemented concretions usually 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 sequences of continental origin compare with the **silcrete** nodules of some present-day soils and, by inference, are thought to have formed soon after deposition. As well as occurring as concretions, chert is also found in thinly bedded units attributable to lithification of primary siliceous oozes.

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; anhydrite usually develops displacively as nodules and layers. In some examples, gypsum precipitation follows plant roots, giving soil profiles called "gypcretes".

In ancient sequences, the textures, fabrics and structures developed by the growth of evaporite minerals may still be recognized, even though 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 oxidizing conditions, although, once formed, it can persist in slightly reducing alkaline conditions (Fig. 9.36). Nodular haematite usually occurs in reduced or partially reduced sequences and commonly the nodules occur in profiles attributable to ancient soil development.

Barite ($BaSO_4$)

Barite is quite common as a local cement, particularly in red sandstones where well formed crystals poikilitically enclose the sand grains. In modern deserts sand roses form close to the sediment surface. Oxidizing conditions and a supply of barium in solution are needed for barite to form in this way.

Limonite ($2Fe_2O_3.3H_2O$)

Although many iron minerals weather to limonite, the mineral also appears to form concretions and nodules in its own right, if oxidizing groundwater conditions prevail (Fig. 9.36). It most commonly occurs as concretions in sands and sandstones, often cementing the only lithified parts of otherwise unconsolidated sands.

Form and location of concretions

A commonly recurring question is: Why has the concretion formed at this place and in this particular shape? Sometimes the answer is obvious. In other cases only general explanations are possible. The question implicitly assumes 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-grain clastic sequences, in chalk and in limestones, concretions or nodules occur in zones parallel to bedding. Sometimes a slight lithological contrast is seen 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 sequences, flints in chalks, and cherts in homogeneous limestones, possibly taking their form from earlier diagenetic evaporite nodules (see Ch. 8; Fig. 9.35).

Subtle differences in, for example, organic content or permeability may control the development of such concretions. Compositional differences may allow particular pore-water chemical conditions to develop, whereas permeability controls 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. 9.37a). 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 rootlets that penetrate the bedding (Fig. 9.37b,c). The concretions tend to be elongate normal to bedding. They are composed mostly of siderite, which weathers to limonite, but often small pyrite crystals fringe the concretion and are also 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 thin carbonaceous rootlets usually enables root concretions to be distinguished from those that follow burrows.

In other fossil soils, concretions of limonite, silica, gypsum or haematite occur and these minerals may also contribute to a general colour mottling that follows the root traces of plants.

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 and ammonites (Fig. 9.37d) 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 in the pore water local chemical conditions 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 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 is needed to allow the soft parts to become buried before scavengers consume them. Once buried, anaerobic rotting would lead to the reducing conditions necessary for pyrite precipitation.

Concretions in distinct vertical profiles

Concretions, particularly of calcite, sometimes 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; others show an upwards increase in nodule size, abundance and coalescence, often as meshes and networks with linking veins (Figs 9.38, 9.39). In rather rare examples, the upper part of the profile is a 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 subvertical concretions and to convex-upwards veins within the network of concretions. Such structures have been called **pseudo-anticlines**.

These profiles and associated features compare closely with those of present-day soils of certain semi-arid areas, the so-called **calcrete** and **caliche** soils. These result from the vertical movement of water

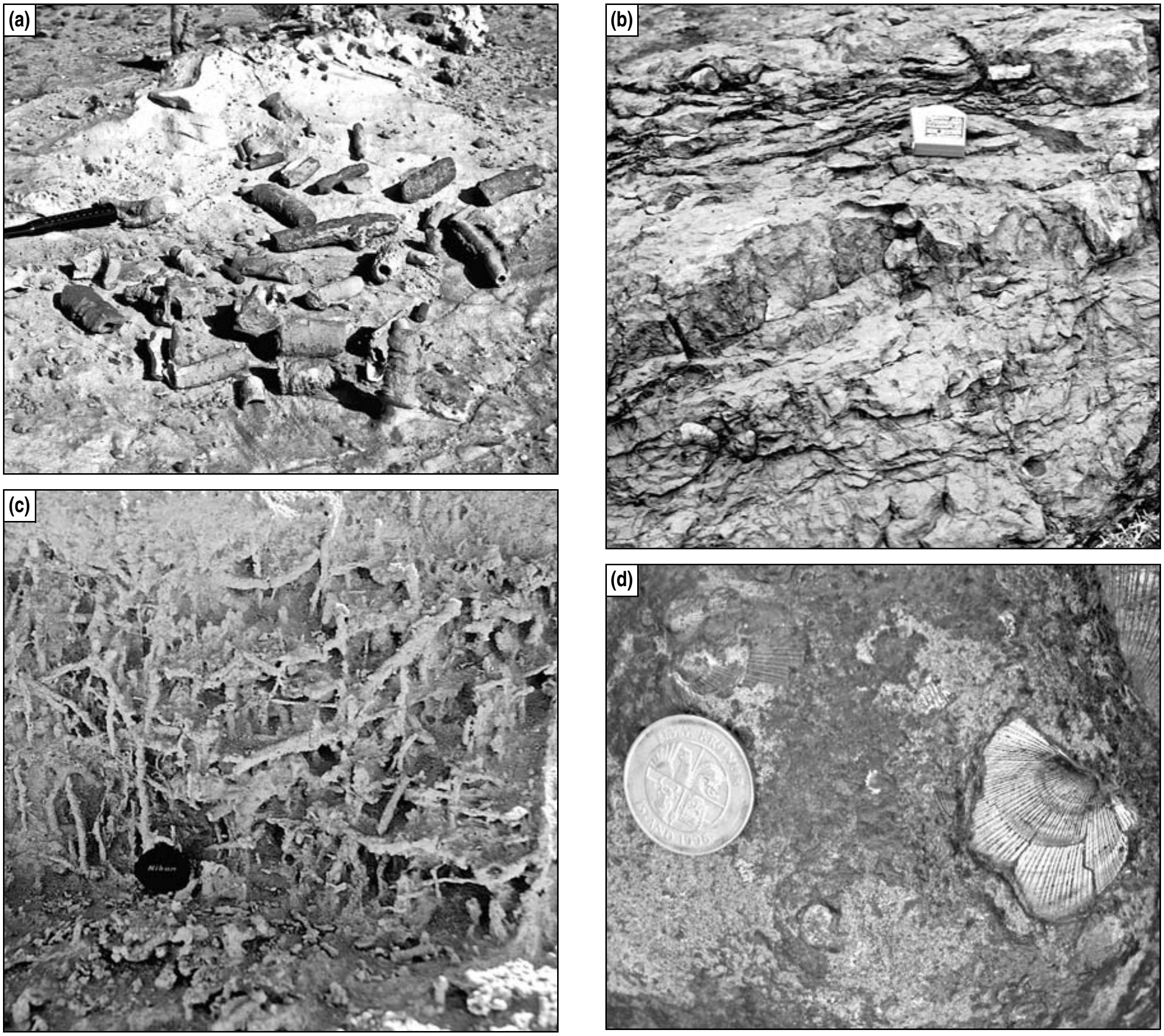


Figure 9.37 Examples of concretions around objects. **(a)** Concretion (probably of limonite) around animal burrows. Lithification as part of the concretionary process has subsequently enabled the burrows to be exhumed intact (photo courtesy of Gilbert Kelling). **(b)** Siderite concretions developed around root structures preserved in a seat-earth soil; Upper Carboniferous, Pembrokeshire, South 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, Carboniferous, Lancashire, England.

through the sediment, because of both the downwards movement of rainwater and the upwards movement of groundwater under dry evaporating conditions. Some of the vertical nodules (or **glaebules**) may reflect original plant roots (Fig. 9.39), and 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 windblown 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 be sub-aerially 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 sequence. The identification of these profiles is

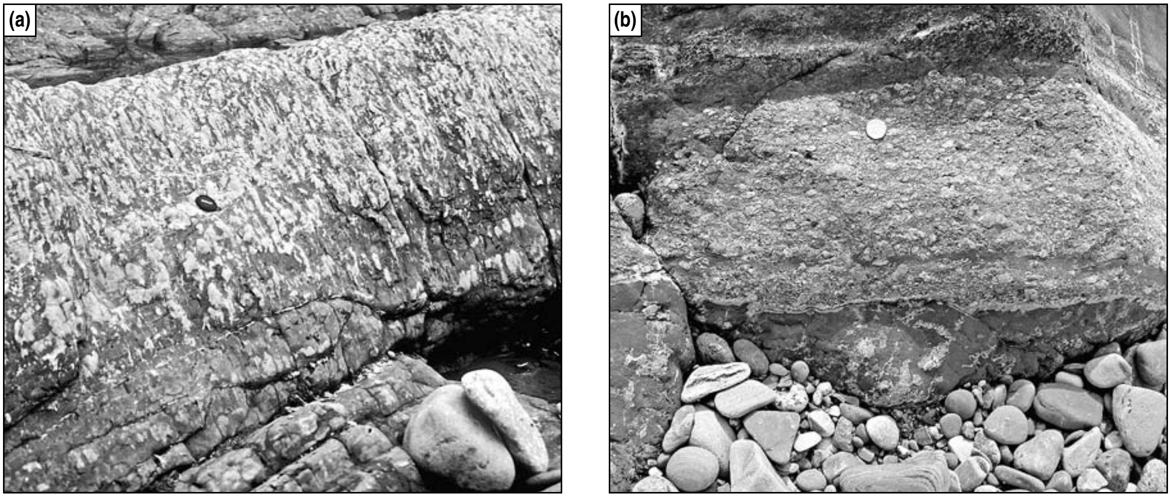


Figure 9.38 Carbonate concretions developed as a result of soil-forming processes to give calcrete or caliche profiles. **(a)** A fairly mature profile with concretions concentrated around root structures; this corresponds to stage b in Figure 9.39. **(b)** Calcrete nodules developed to different degrees within different layers of the host sediment. Both examples from the Old Red Sandstone, Devonian, Pembrokeshire, South Wales.

clearly important in environmental and palaeoclimatic reconstruction, especially where the nodule-forming processes within soils are strongly controlled by climate (Fig. 9.40). Not only do the profiles indicate prevailing conditions, but they also record fluctuations in depositional rate. Other similar profiles involve silica, haematite and gypsum nodules, giving **silcrete**, **ferricrete** and **gypcrete** soils respectively.

Mode and timing of growth of concretions

Inferences about the way in which concretions grow within the host sediment may be made by looking at their internal features. 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

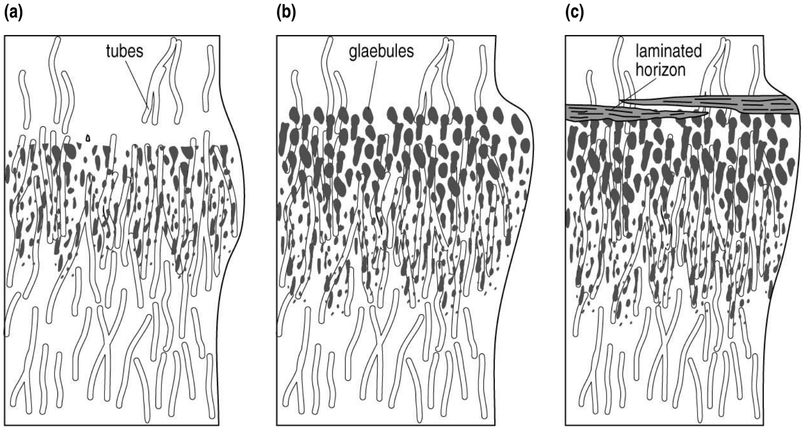


Figure 9.39 Carbonate concretions developed as a result of soil-forming processes to give calcrete or caliche profiles. Stages in the development of a mature profile: **(a)** incipient concretion growth; **(b)** development of large concretions, especially in the zone of water-table fluctuation; **(c)** intercalation of concretionary nodules to give a laminated hardpan horizon.

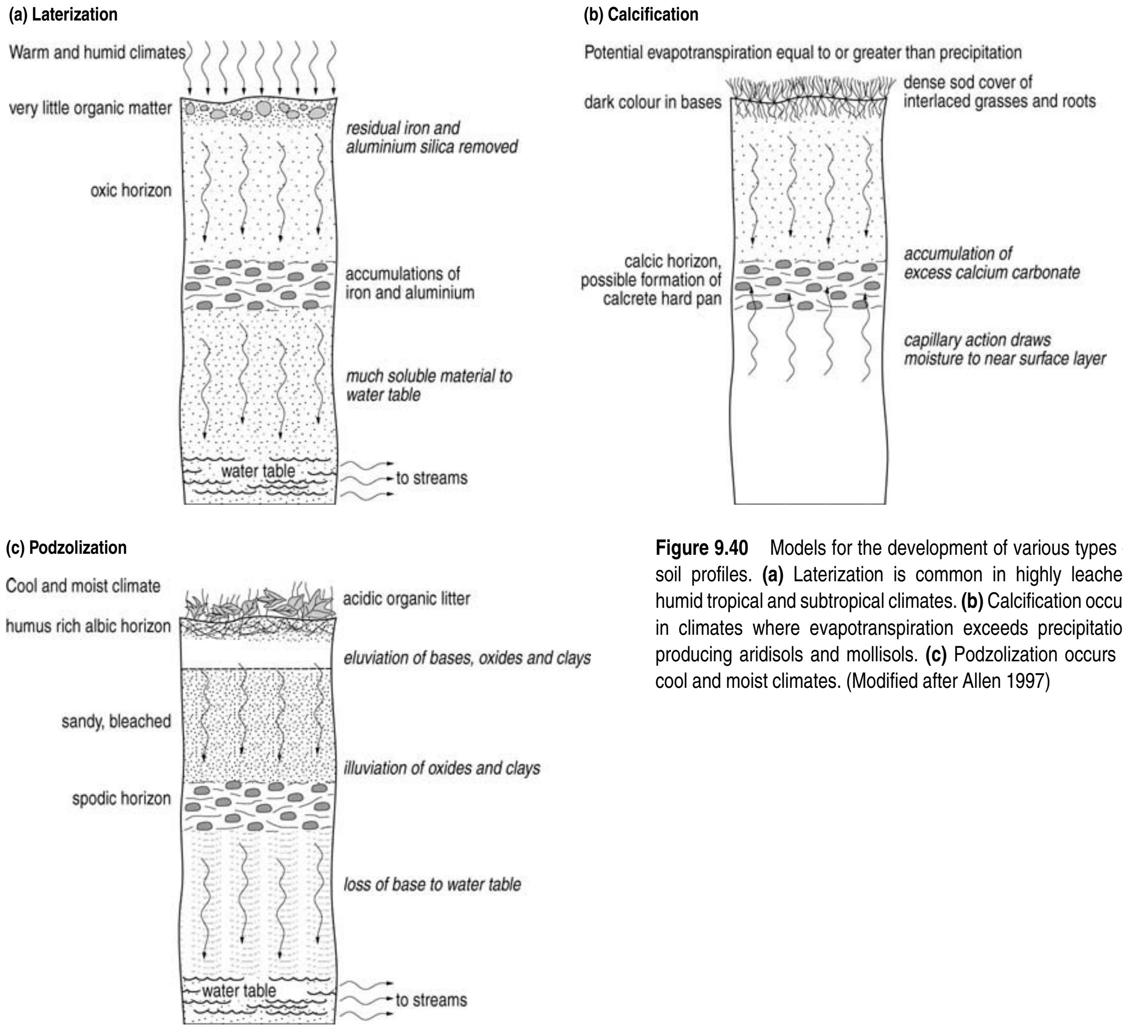

(a) Laterization
Warm and humid climates
very little organic matter
residual iron and aluminium silica removed
oxic horizon
accumulations of iron and aluminium
much soluble material to water table
water table
to streams
(b) Calcification
Potential evapotranspiration equal to or greater than precipitation
dark colour in bases
dense sod cover of interlaced grasses and roots
calcic horizon, possible formation of calcrete hard pan
accumulation of excess calcium carbonate
capillary action draws moisture to near surface layer
(c) Podzolization
Cool and moist climate
acidic organic litter
humus rich albic horizon
eluviation of bases, oxides and clays
sandy, bleached
illuviation of oxides and clays
spodic horizon
loss of base to water table
water table
to streams

Figure 9.40 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)

compare the thickness of the lamination inside the concretion with that in the host sediment, and also to note how the lamination in the host sediment behaves around the concretion. Lamination is commonly much thicker within the concretion than outside it, and the lamination in the host sediment wraps around the concretion both above and below. Clearly, a certain amount of compaction has taken place after the concretion formed, and comparison of lamina thicknesses can give a measure of its magnitude. When 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 cracks reflect a synaeresis-type contraction attributed to the de-watering of a gel-like mass of clay minerals (Fig. 9.41). 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. 9.42), which reflects the stress field set up by the growth of the concretionary cement.

In sandstones, early carbonate concretions often weather out as holes because of a later phase of general cementation by silica. This renders the carbonate more susceptible to weathering than the host. It is often possible to obtain a clearer impression of the original shape

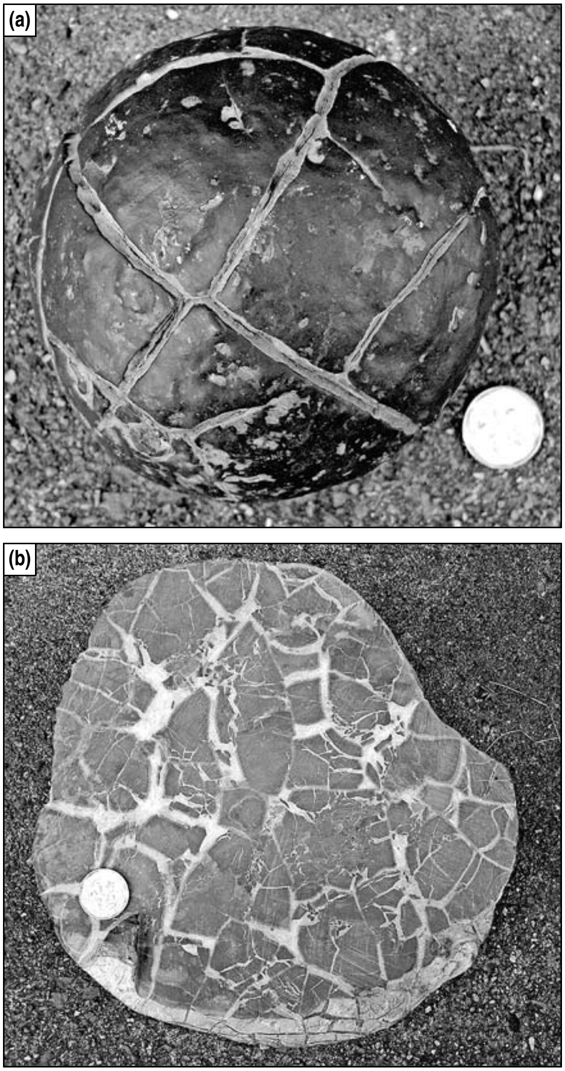


Figure 9.41 Examples of septarian nodules: **(a)** external form, **(b)** internal structure. The lenticular cracks are filled with coarse calcite crystals. Caton Shales, Carboniferous, Lancashire, England.

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. 9.43). Growth of anhydrite as layers sometimes sets up stresses that produce highly contorted folding within the layer.

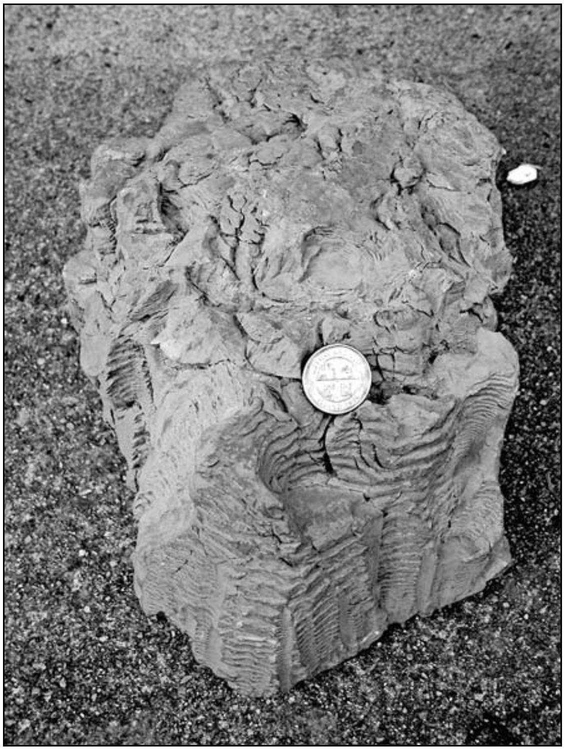


Figure 9.42 Cone-in-cone structure in carbonate-cemented mudstone. Probably Upper Carboniferous Coal Measures; location unknown.

Nodule growth by replacement

Nodule growth by replacement is not always easy to recognize. In some cases, the replacement is associated with displacive growth, but, in other cases, details of the internal structure of the host 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

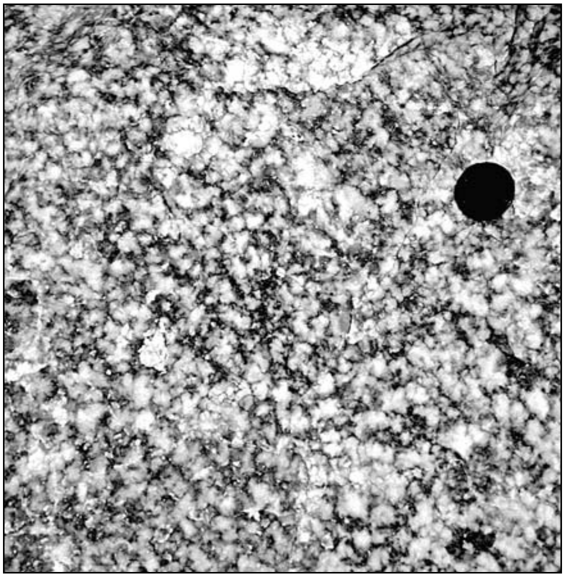


Figure 9.43 Anhydrite nodules that have grown displacively to push aside host sediment into thin veneers between the nodules, giving so-called chicken-wire texture; Miocene, Mojacar, southeast Spain.

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 shell organism, where the strength of the shell maintained the cavity until the fill has been precipitated. A secondary cavity could result from the solution or rotting of some object, probably after the host sediment had become partially lithified so as 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

Two types of dissolution product occur, commonly in limestone sequences.

Stylolites

On certain bed partings, particularly in limestones and less commonly in sandstones, a highly crenulated contact is seen (Fig. 9.44). 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. When 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 because of the vertical compression. The relief of the irregularities gives a minimum measure of the thickness of material removed in solution. This can sometimes 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.

Collapse breccias

In sequences that contain or may have contained evaporites, brecciated horizons sometimes occur. Angular blocks may be of any size, but they are similar in composition to other sediments in the sequence. Often, blocks will be shifted only slightly relative to one

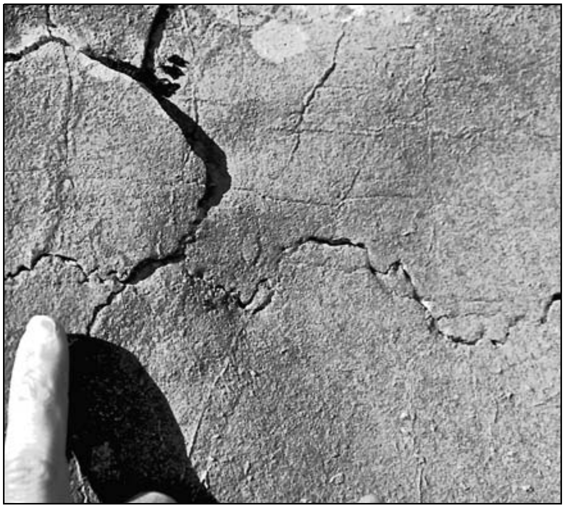


Figure 9.44 Stylolite in limestone. The crenulated surfaces are developed normal to the principal compressive stress and usually follow bedding. They commonly show concentration of insoluble residues. Devonian Limestone, La Vid, Cantabrian Mountains, Spain.

another, and it is possible to restore them mentally to their original positions, as in a jigsaw puzzle. Such breccias may often be the result of the wholesale removal in solution of a thick evaporite unit, so that the only evidence we have of its former existence is the breccia caused by the collapse of overlying strata.

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 mostly ignored in geology courses. At outcrop they were commonly dismissed as “burrows” or “worm traces”, suggesting that they had little to contribute to the elucidation of Earth history. In fact the opposite is often 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. They often occur where no body fossils have been preserved, for example in non-marine redbeds, or where organisms were entirely of the soft-body varieties.

Trace fossils record behavioural, ecological and sedimentological events that body fossils and other sedimentary structures cannot record directly. Therefore, their study may alter one’s view of a problem or turn an investigation in a new direction. Sometimes 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 equal 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. Students working on certain sedimentological projects should consult colleagues with experience of trace fossils, rather than ignore 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 outcrops that contain trace fossils. Three main approaches to classification are considered, which are best used in combination:

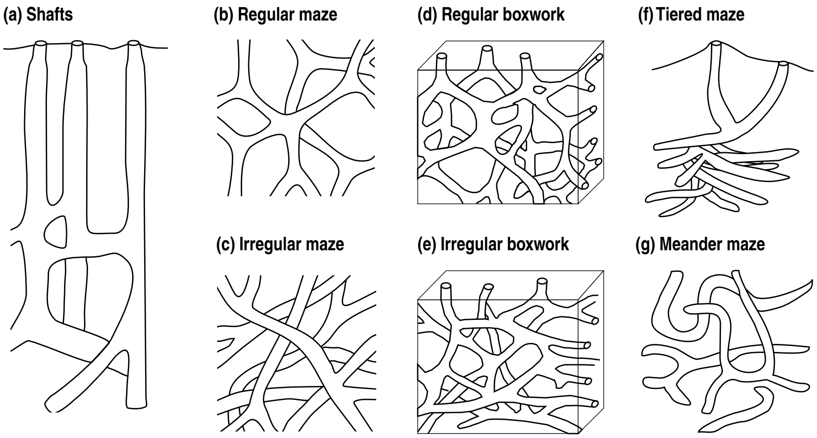
- taxonomic classification using morphological aspects
- preservational–sedimentological classification
- ethological classification through the consideration of behavioural and environmental aspects.

In introducing each of the schemes 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 for their generation.

Taxonomic (morphological) classification

Non-experts are often frustrated at 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. Recognizing these difficulties, international codes of biological nomenclature have been modified in order to accommodate the naming of trace fossils, with the introduction of new codes in 1985 and 1997 (see bibliography). In this connection, the basic unit of nomenclature is the **ichnogenus**, defined morphologically. The lower category, the **ichnospecies**, often reflects size, preservational or minor behavioural variations of the basic form (Fig. 9.45).

Given that the objects to be named are biogenic, five overall sets of characteristics or **ichnotaxobases** are



(a) Shafts

(b) Regular maze

(c) Irregular maze

(d) Regular boxwork

(e) Irregular boxwork

(f) Tiered maze

(g) Meander maze

Figure 9.45 The variability of burrow arrangements for a single ichnogenus, *Ophiomorpha*. **(a)** Vertical components are most dominant. **(b, 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)** A 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 & Droser 1998 and Frey et al. 1978)

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 multi-phase 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 because of 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 created by 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 the table in Figure 9.46 as a guide for description. A classification should involve a consideration of 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:

- a single shape (e.g. a print made by a foot)
- 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 or refuge (or both)
- a series of closely related concentric laminae (spreiten), often of U-shape, caused by an animal shifting position within its burrow as it grows or moves upwards, downwards, forwards or backwards by excavating and backfilling (Fig. 9.47)
- a pouch shape, for example caused by the resting of bivalves
- a network pattern, perhaps made by some systematic activity such as grazing or farming.

Attempting to relate the trace fossils depicted in Figure 9.48 to these categories should help to illustrate and answer these queries.

Classification according to mode of preservation

When trace fossils are analyzed sedimentologically, it is clear that, although their morphology typically reflects a wide range of animal-behaviour patterns, their preservation usually occurs as a result of relatively few

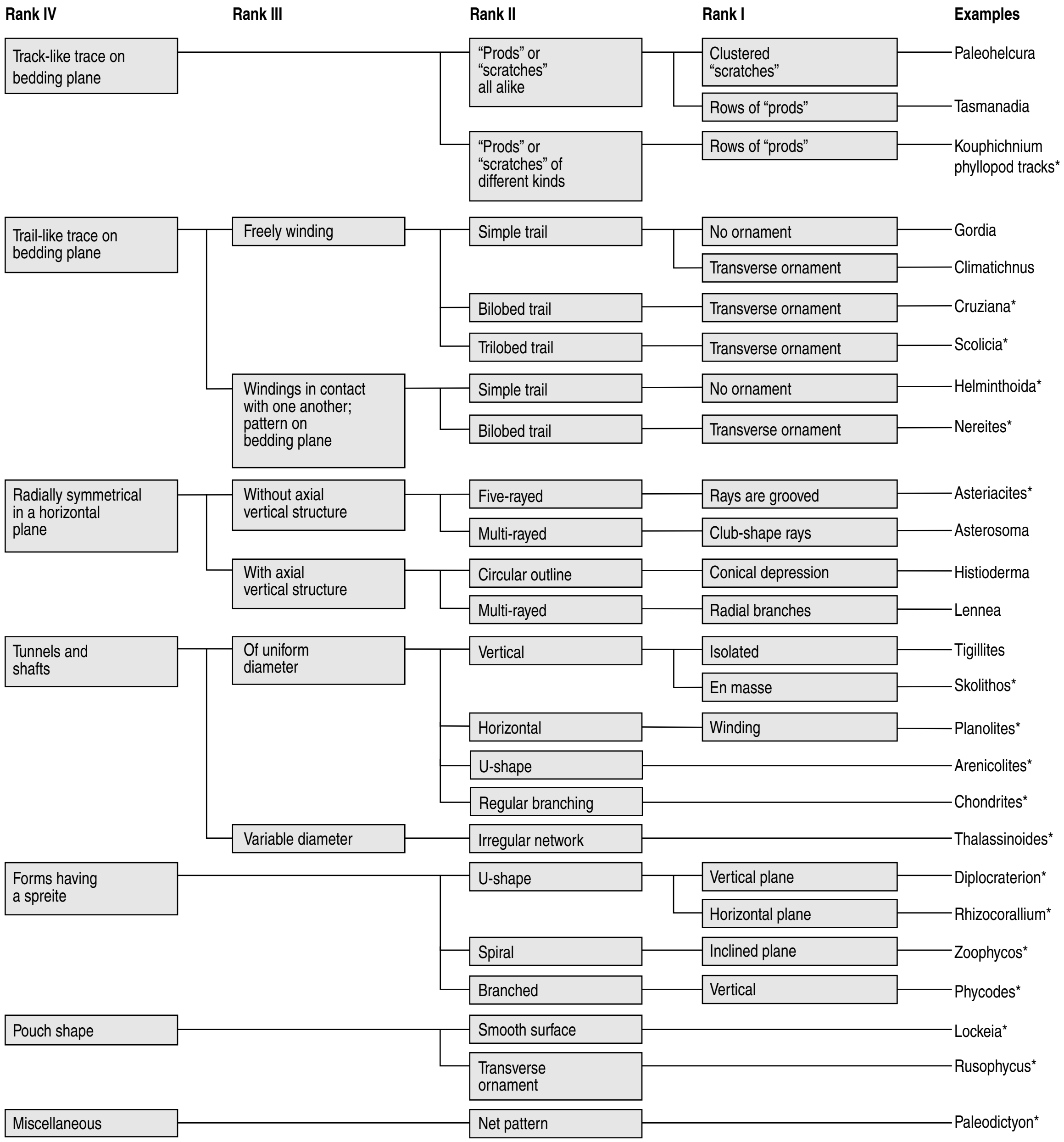

Rank IV
Rank III
Rank II
Rank I
Examples
Track-like trace on bedding plane
"Prods" or "scratches" all alike
Clustered "scratches"
Paleohelcura
Rows of "prods"
Tasmanadia
"Prods" or "scratches" of different kinds
Rows of "prods"
Kouphichnium phyllopod tracks*
Trail-like trace on bedding plane
Freely winding
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*
Multi-rayed
Club-shape rays
Asterosoma
With axial vertical structure
Circular outline
Conical depression
Histioderma
Multi-rayed
Radial branches
Lennea
Tunnels and shafts
Of uniform diameter
Vertical
Isolated
Tigillites
En masse
Skolithos*
Horizontal
Winding
Planolites*
U-shape
Arenicolites*
Regular branching
Chondrites*
Variable diameter
Irregular network
Thalassinoides*
Forms having a spreite
U-shape
Vertical plane
Diplocraterion*
Horizontal plane
Rhizocorallium*
Spiral
Inclined plane
Zoophycos*
Branched
Vertical
Phycodes*
Pouch shape
Smooth surface
Lockeia*
Transverse ornament
Rusophycus*
Miscellaneous
Net pattern
Paleodictyon*

Figure 9.46 A morphological classification of some common invertebrate trace fossils. Examples that are illustrated in the figures are marked by an asterisk. (Modified after Simpson in Frey 1975)

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 and animal activity. Diversity of forms of traces arises from the activities of animals of different shape and with different behaviour patterns living in different depositional settings. Remember to look systematically under, within and on top of beds, and for features that

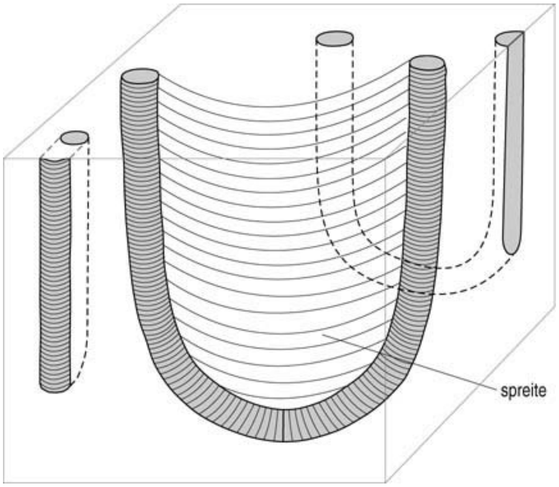


Figure 9.47 A trace fossil with spreiten. The U-tube is the dwelling burrow of an organism. The arms of the U indicate that the organism lengthened and deepened the tube by removing sediment from the floor of the burrow and plastering it against the ceiling, so forming a concentric *spreite*. These spreiten are protrusive, the present burrow being the last formed and underlying all previously formed ones.

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 hold for 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 events may be hard to establish (Figs 9.49, 9.50).

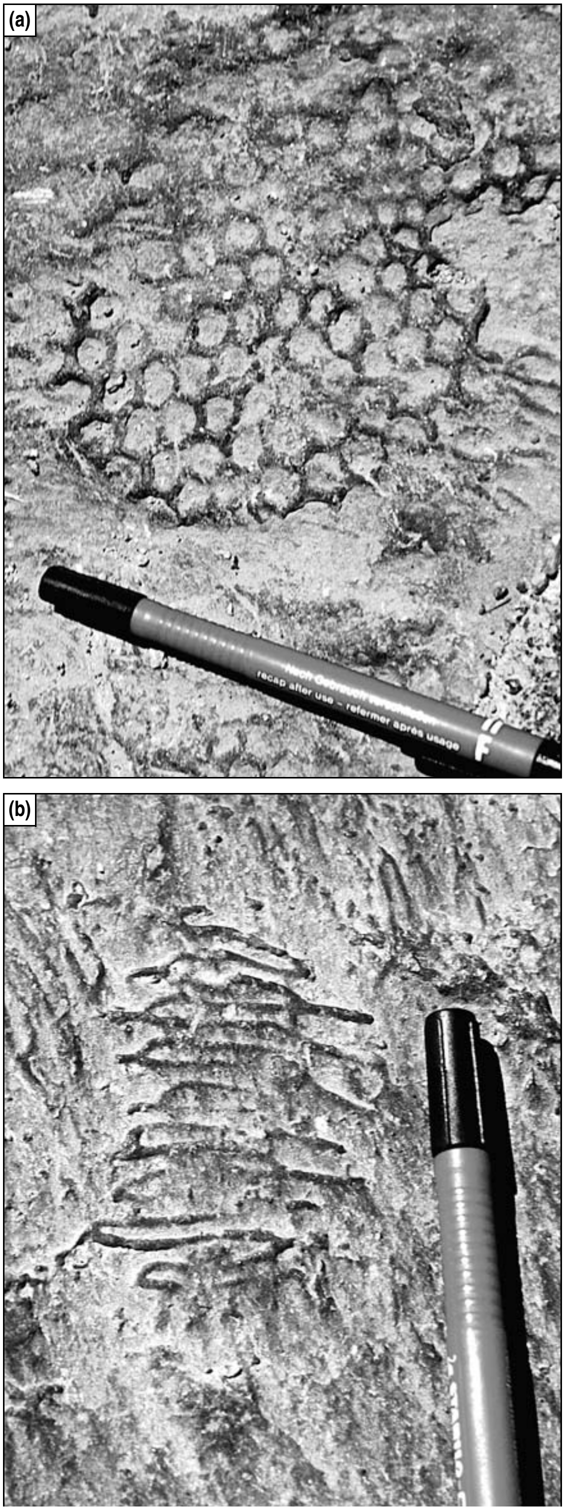
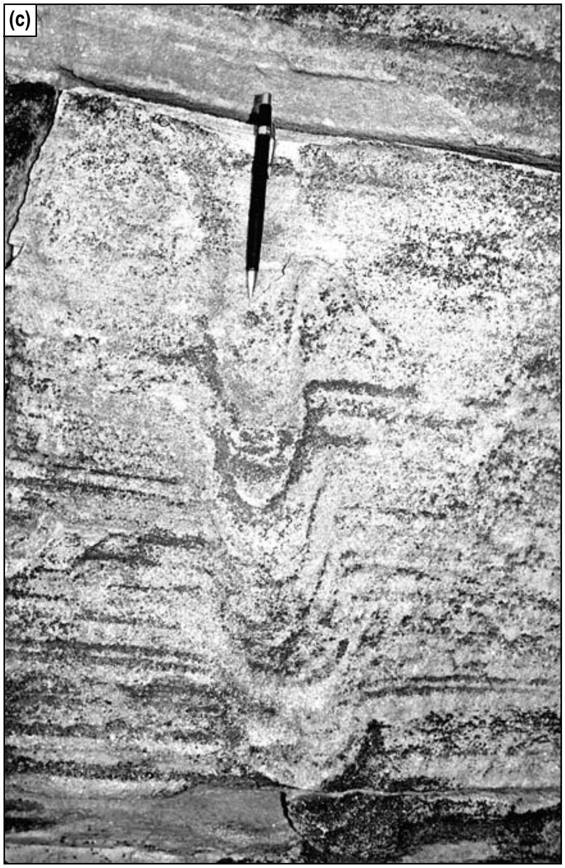
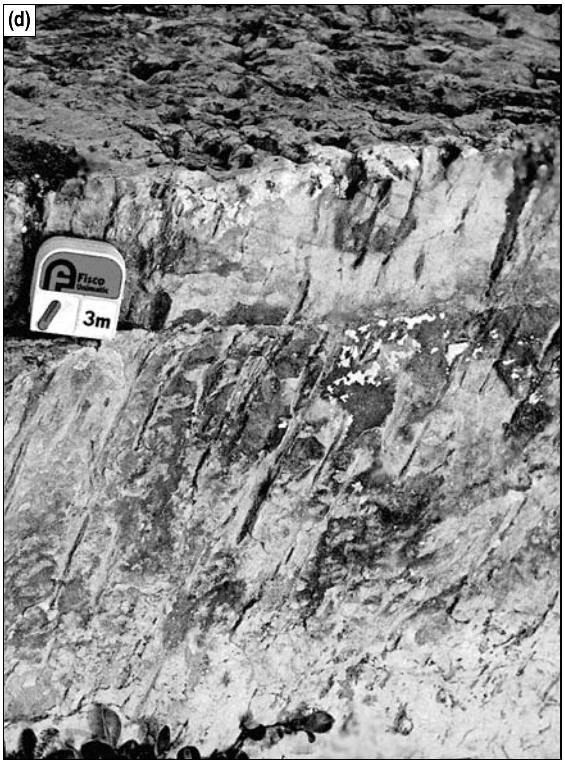
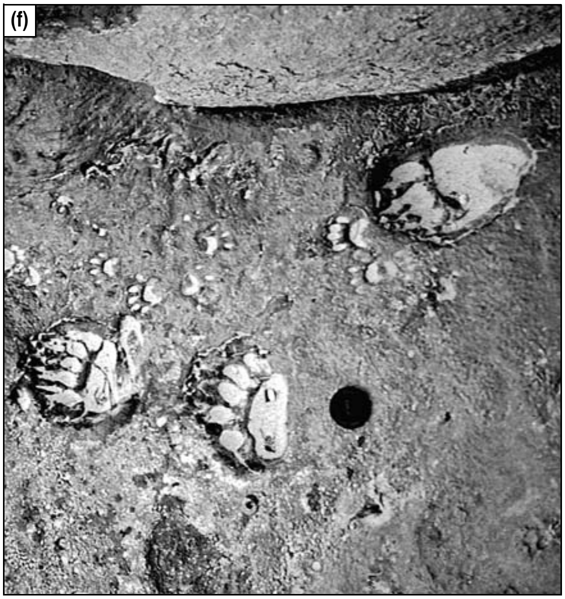


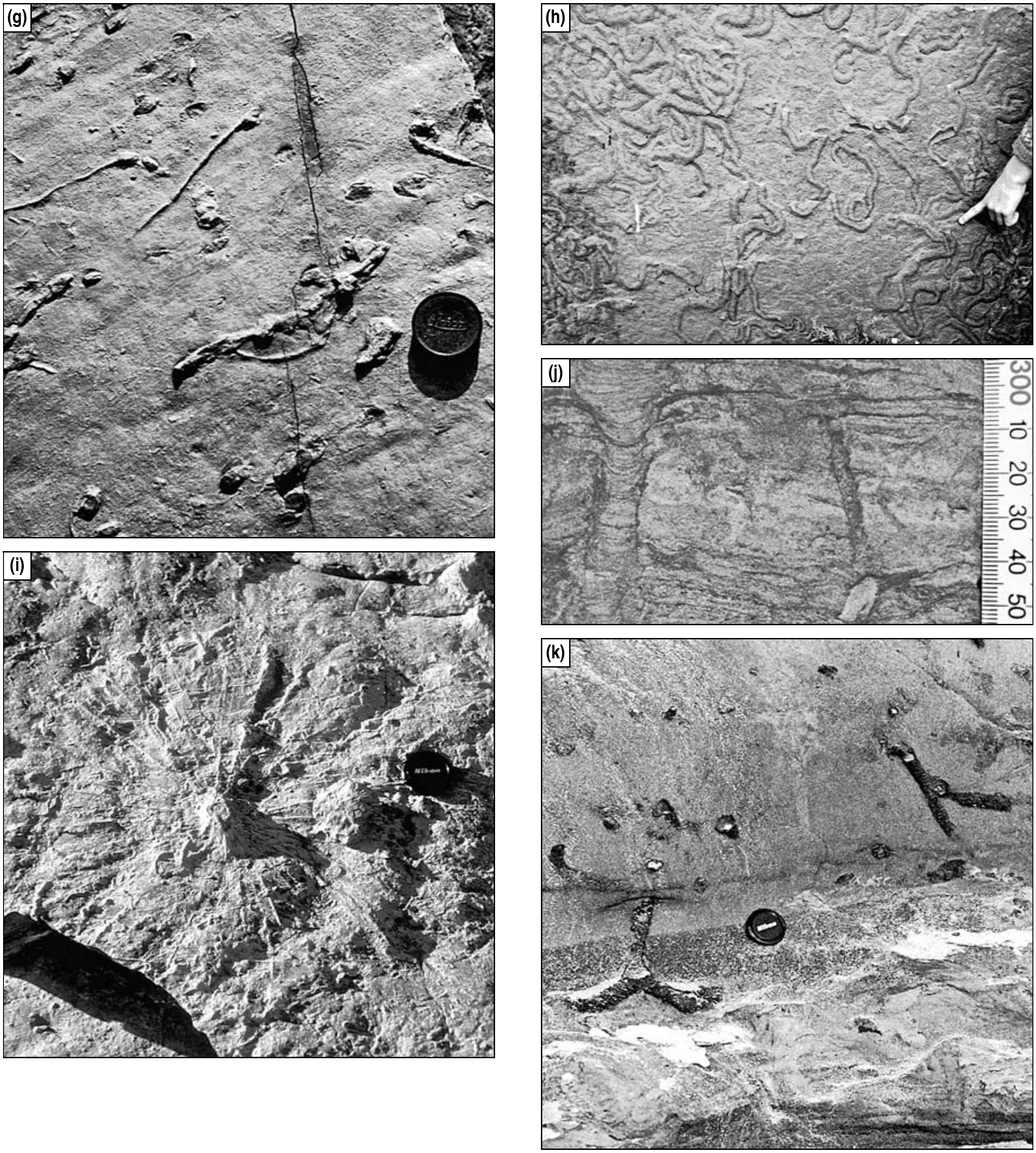
Figure 9.48 (continued overleaf) Examples of some common types of trace fossils. **(a)** *Paleodictyon* in turbiditic calcarenities, Miocene, southeast Spain. **(b)** *Urohelminthoida* in turbiditic calcarenities, Miocene, southeast Spain. **(c)** *Diplocraterion* in shallow marine sandstones, Permian, Namibia. **(d)** *Skolithos*, Lower Cambrian, northern Greenland. **(e)** *Arenicola*, modern tidal flat, Haringvliet, the Netherlands. **(f)** Vertebrate (bear) footprints, two sets – parent and juvenile, modern, White Canyon, Utah. **(g)** *Lockeia*, casts of a bivalve resting trace on a lower bedding-plane surface, Haslingden Flags, Carboniferous, Lancashire, England. **(h)** *Helminthopsis*, a back-filled burrow with a meandering course preserved as a cast within delta-front silty sandstones, Namurian, County Clare, Ireland (photo courtesy of Gilbert Kelling). **(i)** *Phoebichnus*, a large radiating trace, Hecho Group, Eocene, Pyrenees, Spain. **(j)** *Diplocraterion*, with well developed spreiten; Lower Cretaceous, Isle of Wight, England. **(k)** *Ophiomorpha*; locality unknown. **(l, p. 222)** *Thalassinoides*, in marl; Miocene, Nijar, southeast Spain.

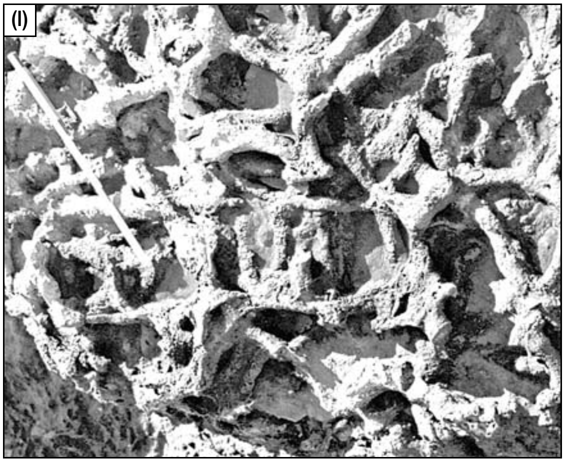

(c)


(d)
Fisco
Unimatic
3m

(e)


(f)


(g)
(h)
(j)
300
10
20
30
40
50
(i)
(k)



Key questions about the mode of preservation of trace fossils

The questions below (2–8) should be used 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; questions 4–8 consider the position and process of preservation. In considering these questions, it is useful to speculate and hypothesize about the possible organisms and processes that gave rise to them. Try using the following questions to help determine the mode of preservation of the trace fossils depicted in Figure 9.48.

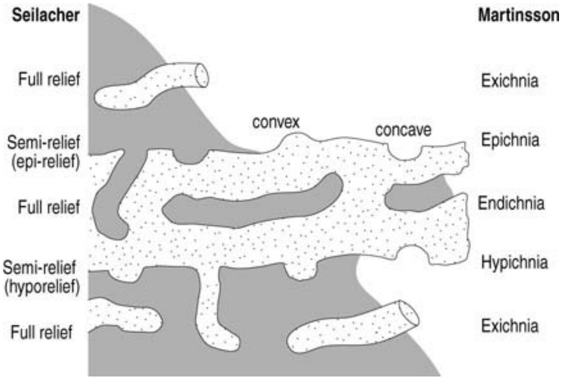


Figure 9.49 Classification of trace fossils in relation to their arenaceous casting medium (toponomy). Comparison of the terminologies of Seilacher (1964) and Martinsson (1965). (Modified after Bromley 1996)

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 from an adjacent bed or by the precipitation of mineral matter, produces a natural cast. The best way to appreciate the difference between moulds and casts is to make examples in the laboratory (e.g. a mould and cast of a hand- or footprint or burrow) 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 often preserved as calcite and siderite nodules in shale or as limonite nodules in sand. Small burrows are often preserved as pyrite, which already oxidizes to reddish-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 chemical and physical compositions different from those of the surrounding sediment (Fig. 9.48). 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 are then used to 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 paler "haloes" around the nodular traces. It is often difficult to distinguish burrows preserved in this manner from the traces of plant roots.

Q. 4 Is the trace fossil preserved in an interfacial position on the top of the casting medium as an epichnial trace like a ridge (positive feature) or a groove (negative feature) (Fig. 9.49)? What is its composition? 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. 9.49)? If so, is there any evidence that this was a sediment/water interface? Are only the sub-interface laminae deformed? Was the trace fossil preserved at a sediment/sediment interface, possibly between contrasting lithologies, possibly at a concealed

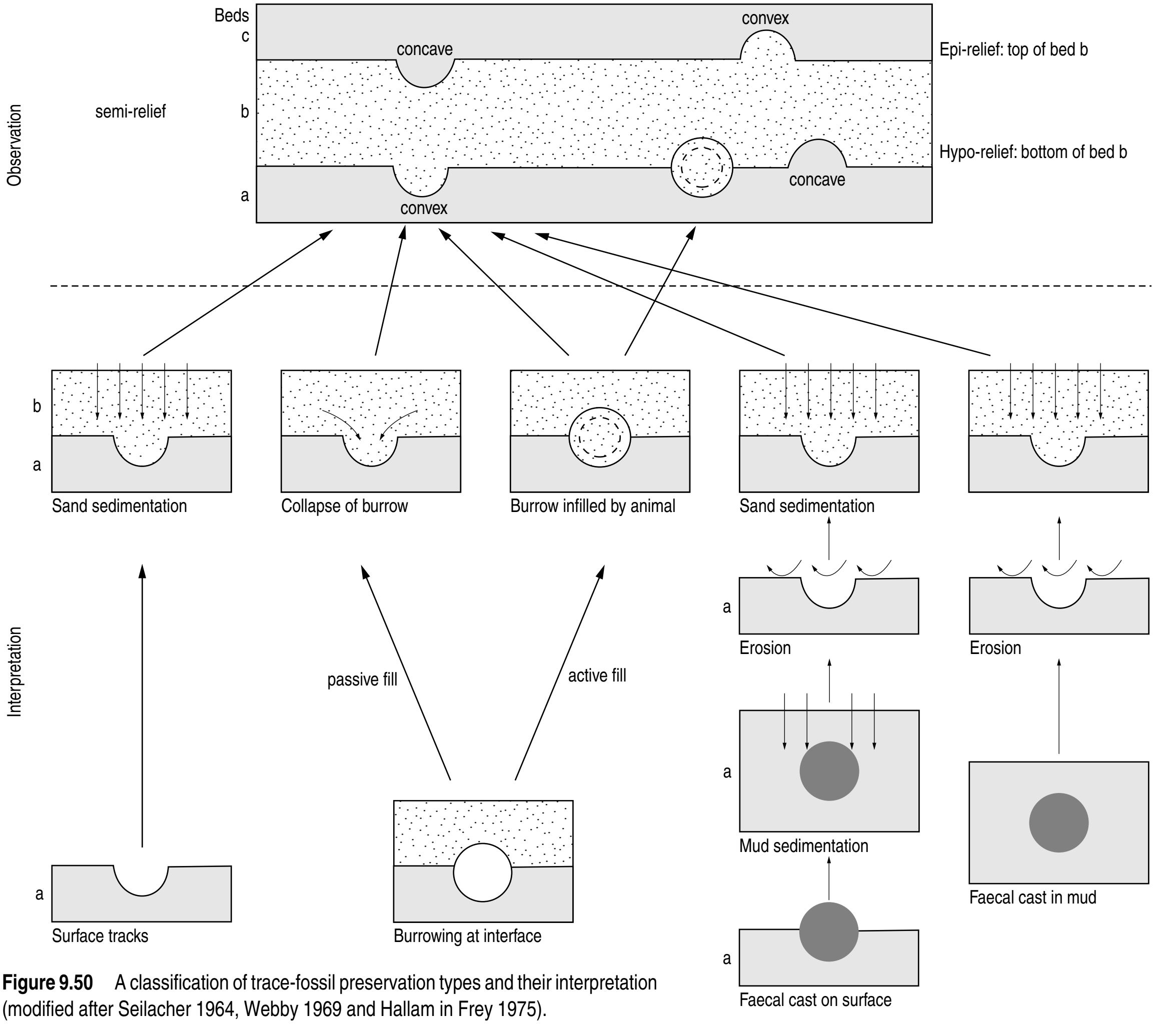

Observation
Beds
c
b
a
semi-relief
concave
convex
Epi-relief: top of bed b
Hypo-relief: bottom of bed b
concave
convex
Interpretation
b
a
Sand sedimentation
Collapse of burrow
Burrow infilled by animal
Sand sedimentation
Erosion
Erosion
passive fill
active fill
Mud sedimentation
Faecal cast in mud
Surface tracks
Burrowing at interface
Faecal cast on surface

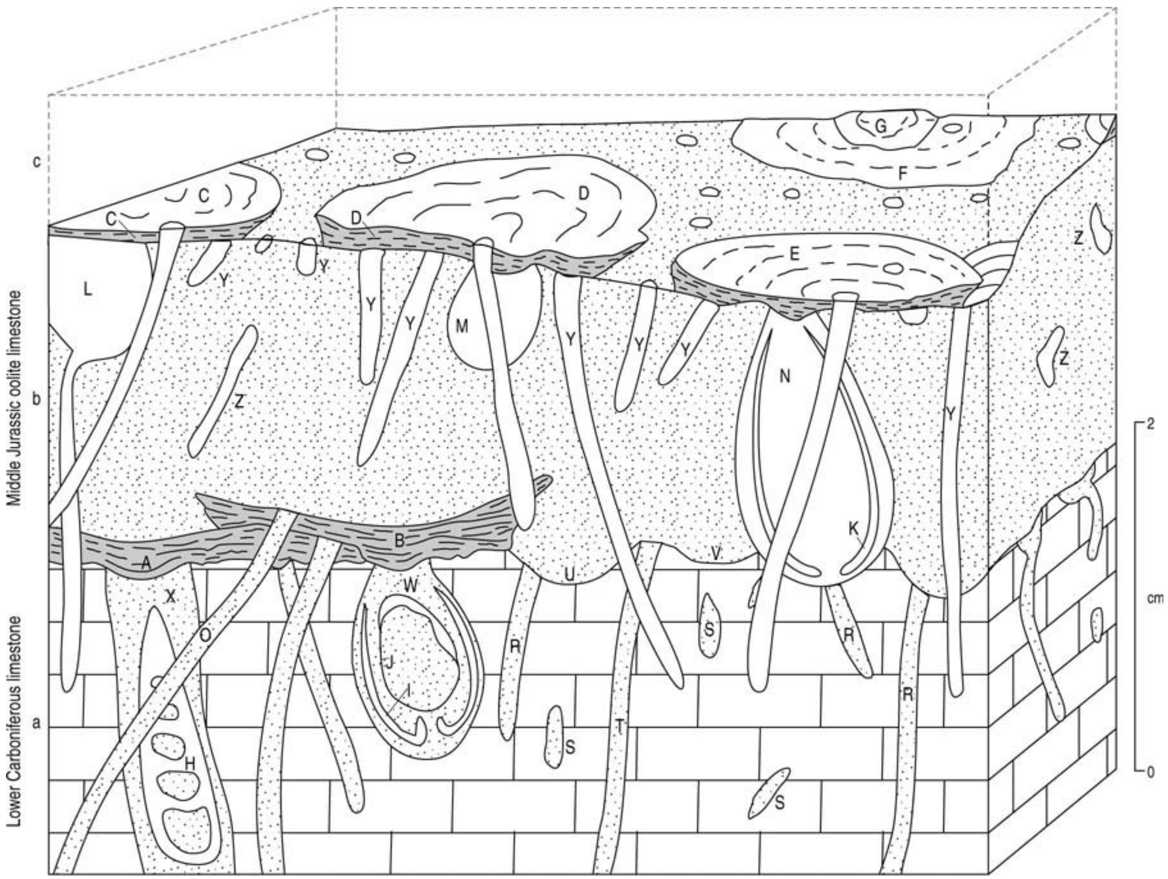
Figure 9.50 A classification of trace-fossil preservation types and their interpretation (modified after Seilacher 1964, Webby 1969 and Hallam in Frey 1975).

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. 9.49)? 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 that 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. 9.49)? 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 reworked.

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, and leave the mucus-bound burrow linings as sediment-filled “gloves”. These may be covered later by, possibly contrasting, sediments


c
b
a
Middle Jurassic oolite limestone
Lower Carboniferous limestone
G
F
C
C
D
D
E
Z
L
Y
Y
Y
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M
Y
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B
X
W
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I
R
S
R
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Figure 9.51 An exercise on describing the mode of preservation of some trace fossils in a diagram showing 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 9.49. 3. Describe the modes of preservation of the trace fossils using the classification schemes shown in Figures 9.53 and 9.54. 4. Describe the mode of behaviour of the trace fossils using the classification schemes shown in Figures 9.57, 9.58 and 9.59. Determine the stratigraphical 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)

(Fig. 9.48). 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 surfaces of beds, have almost no 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 subparallel laminae called **spreiten** (Fig. 9.47), 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. 9.47), 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 continuing sedimentation as the animal moves upwards. Preservation of delicate individual structures 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 9.51. Attempt to describe and interpret the broad sequence of events that gave rise to the preservation of the traces shown in Figure 9.52, including the recognition of protrusive and retrusive spreiten.

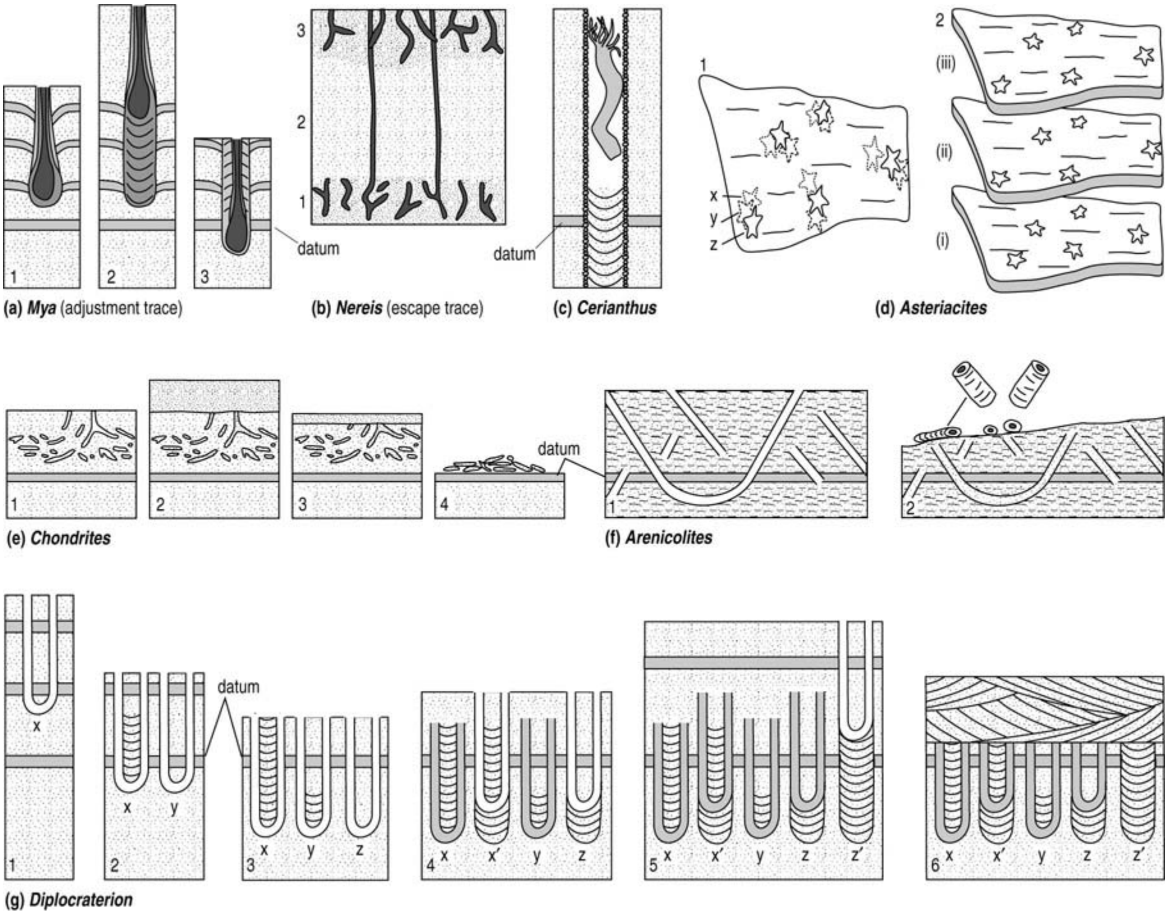


Figure 9.52 An exercise to demonstrate the use of trace fossils for determining amounts of sedimentation and erosion in clastic successions. The diagrams represent single or closely related sequences of vertical successions. For examples a–g, explain the sequence of events that has given rise to the 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, movement of which produces adjustment traces. **(b)** The polychaete worm *Nereis*, which produces escape traces when buried by sediment. **(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 (x, y, z); exposure 2 – three bedding planes (oldest = i, youngest = iii). **(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). Adjustment traces responding to varying substrate level relate to parts (e–g). (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 (Figs 9.53, 9.54). There is some overlap between the basic behavioural categories (Fig. 9.55). However, behavioural classification schemes 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 and escape in relation to a whole series of interacting controlling factors, such as grain size, energy levels, packing, porosity, permeability, pH, Eh, salinity, degree of cementation and the availability of organic matter and nutrients.

Q. 9 Could the structures be the legacy of plant roots in the form of root moulds, root casts, or petrified roots (rhizoliths) (Fig. 9.56)? At first sight, some of these 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 root networks? Is there physical or chemical evidence of pedogenesis (soil formation), including de-stratification, colour mottling, precipitation of calcrete, silcrete or gypcrete cements or nodules (see §9.3.1)? Root structures are often associated with carbon films and coal seams in waterlogged palaeosols. Are the structures found in association with leaf impressions? Is there evidence of downward bifurcation of the structure?

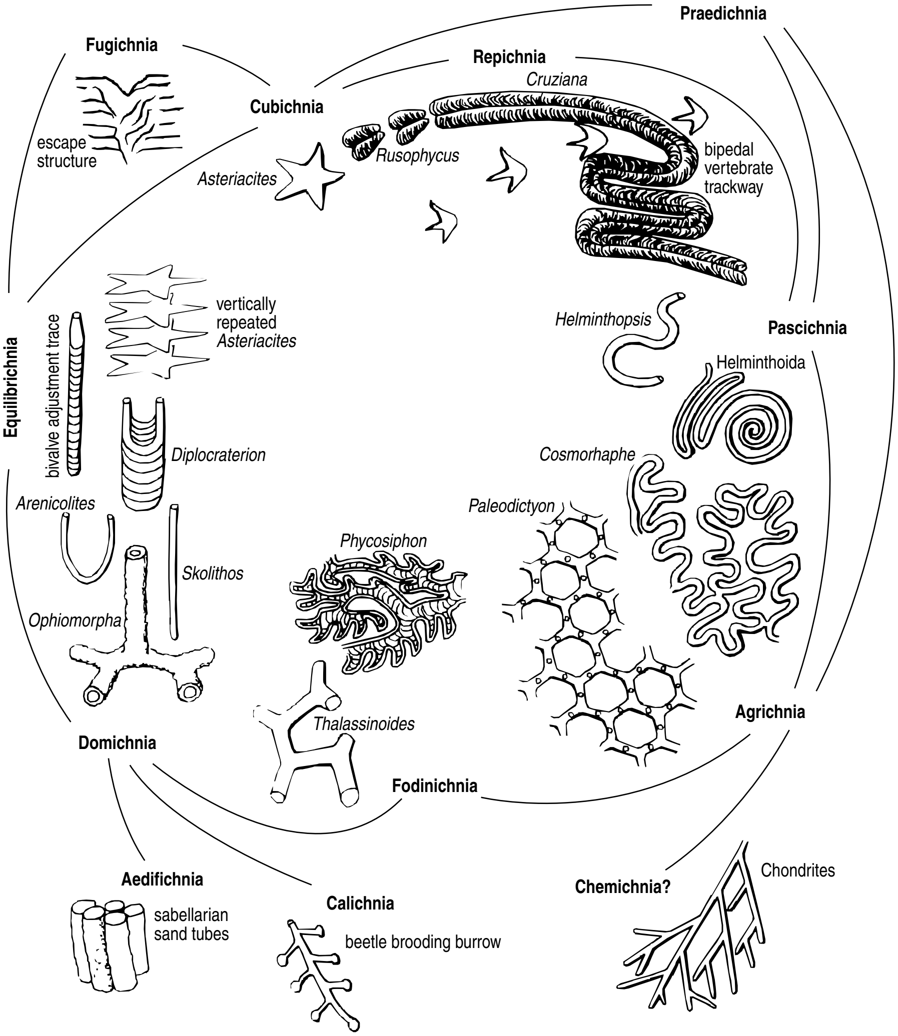
Q. 10 Could the traces be the result of animals temporarily interrupting their movement on or above the sediment surface to rest or seek refuge? Isolated, shallow, trough-like depressions may record the outline or morphology of the undersurface of an animal, or marks caused by its digging or temporarily 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 (starfish, bivalves, arthropods like crabs, and flat fish, among others). Trace fossil examples are *Asteriacites*, *Lockeia, Rusophycus* (Fig. 9.57). These traces are transitional to crawling, dwelling and escape structures (Fig. 9.55).

Q. 11a 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 animals walking, running, galloping, hopping, crawling, half-walking, half-swimming and fully swimming? Could the traces be caused by an animal being drifted by a cross current, while 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* (Fig. 9.48).

Q. 11b Could the traces be made just below the surface of the sediment, especially in silt–sand? Such repichnia are made by benthos and nektobenthos, predators, scavengers and deposit feeders (such as snails). Examples are *Gyrochorte* and *Cruziana* (Fig. 9.57e). Repichnial traces are often transitional to resting traces and surface-grazing traces (Figs 9.53, 9.55).

Q. 12 Could the traces be attributable to carefully 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 that 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 algal-grazing animal to ensure the economical exploitation of food resources in the sediment? Such surface-grazing traces are made by mobile bottom-dwelling organisms (epibenthos, e.g. grazing gastropods such as limpets, worms, echinoids and arthropods) and are known as **pascichnia**. Examples are *Helminthoida* (Fig. 9.48), *Lophoctenium, Nereites*, *Spirophycos* and planar types of *Zoophycos* (Fig. 9.57).

Figure 9.53 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). (Modified after Bromley 1996 and Goldring 1999)


Fugichnia
escape structure
Cubichnia
Asteriacites
Repichnia
Cruziana
Rusophycus
bipedal vertebrate trackway
Praedichnia
Equilibrichnia
bivalve adjustment trace
vertically repeated Asteriacites
Diplocraterion
Arenicolites
Skolithos
Ophiomorpha
Helminthopsis
Pascichnia
Helminthoida
Cosmorhaphe
Paleodictyon
Phycosiphon
Thalassinoides
Agrichnia
Domichnia
Fodinichnia
Aedifichnia
sabellarian sand tubes
Calichnia
beetle brooding burrow
Chemichnia?
Chondrites

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? Could they be preserved 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? The complex and regular meanders, loops, spirals, networks and hexagonally, polygonally branched patterns (e.g. *Paleodictyon, Cosmorhaphe* and *Spirorhaphe*) fit here (Figs 9.48, 9.57). Such traces are referred to as farming and trapping traces: **agrichnia** (Figs 9.53, 9.55).

Q. 14 Could the traces be burrows excavated while animals search for food within the sediment? Features that support such an idea are single, branched or unbranched, cylindrical to sinuous, radial shafts or U-shape burrows, orientated at various angles to the bedding. There may also be complex, parallel to concentric, horizontal burrow repetitions (spreiten) with unlined walls (Fig. 9.53). The structures result from short-lived ephemeral burrowing by animals that were "underground miners"; that is, they were essentially deposit feeders that, in addition, sought secure shelter and refuge. An important feature is that the radial or U-shape burrows do not touch, since the animals generally avoid previously mined sediment on account of its toxicity. These traces

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-shape 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, although secondary activities such as adjustment may also be involved.	Simple, bifurcated or U-shape 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 re-adjustment 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 upwards or downwards 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, with predator/prey relationships often being discernible.	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-branch burrows, often with a radiating architecture.	*Chondrites*

Figure 9.54 Ethological classification of invertebrate trace fossils. Modified in part from Frey and Pemberton (1985) and Pemberton (1992).

are made by epibenthic and endobenthic deposit feeders such as polychaete worms, and are known as **fodinichnia**. These traces are related to grazing (pascichnia) and dwelling (domichnia) structures (Figs 9.53, 9.55). Examples are *Phycodes*, *Rhizocorallium*, *Zoophycos* (Fig. 9.57), *Gyrophyllites* and *Rosselia*.

Q. 15 Could the structures be the dwellings of suspension feeders or carnivores? Features suggesting this are simple, bifurcated or U-shape 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 more or less permanently occupied by suspension feeders or

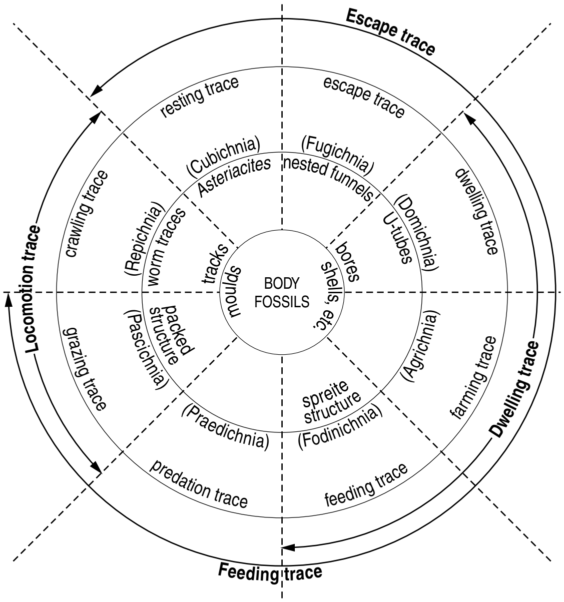
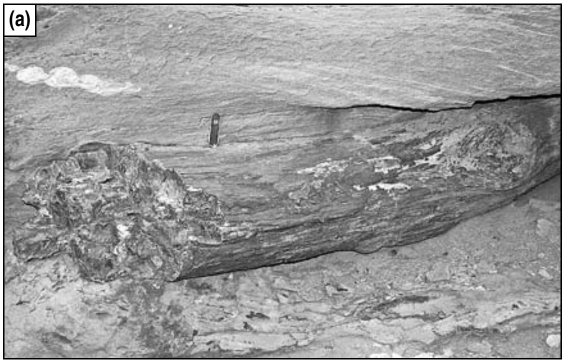

Escape trace
resting trace
escape trace
(Cubichnia)
Asteriacites
(Fugichnia)
nested funnels
(Domichnia)
U-tubes
dwelling trace
crawling trace
(Repichnia)
worm traces
tracks
bores
moulds
BODY FOSSILS
shells, etc.
Locomotion trace
grazing trace
(Pascichnia)
packed structure
(Agrichnia)
farming trace
Dwelling trace
(Praedichnia)
spreite structure
(Fodinichnia)
predation trace
feeding trace
Feeding trace

Figure 9.55 Ethological (behavioural) classification of trace fossils, and their relationship to body fossils; overlap of categories acknowledges the integrations inherent in nature (modified from Pemberton 1992). Additional, more specialized behavioural groups not mentioned on this diagram include aedifichnia, calichnia, chemichnia and equilibrichnia (see text for details).

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 9.53, 9.55). Examples are *Skolithos*, *Arenicolites*, *Diplocraterion*, *Thalassinoides*, *Trypanites* and *Ophiomorpha* (Figs 9.48, 9.57).



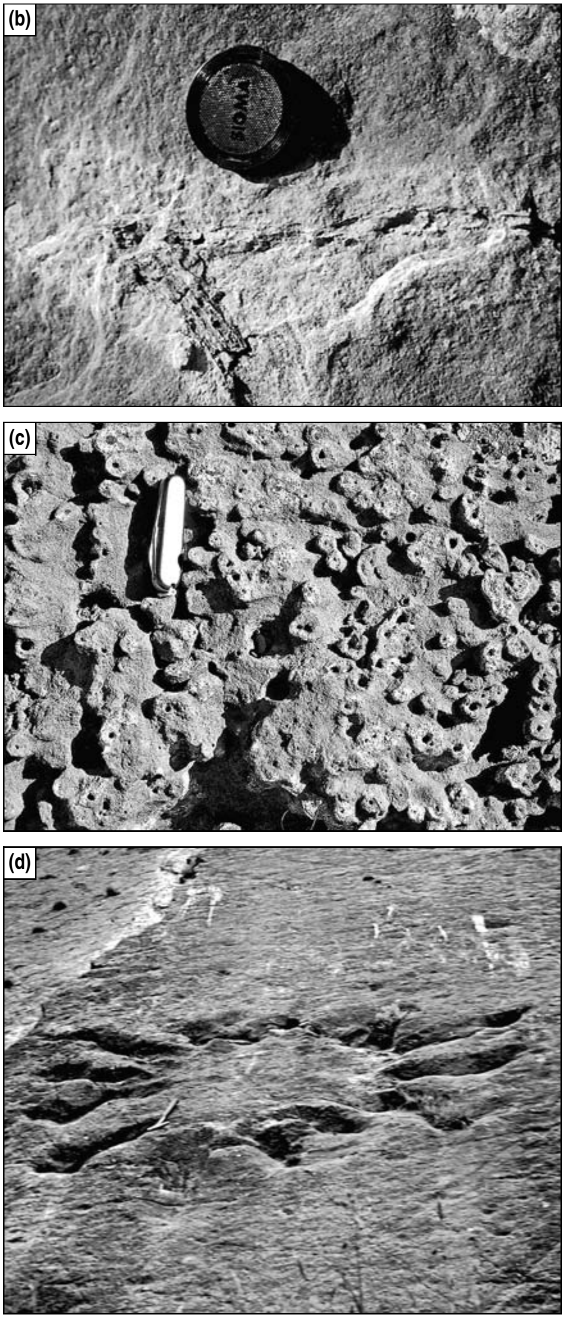


Figure 9.56 Biogenic plant structures: **(a)** tree trunk preserved in fluvial strata, Lower Cutler Beds, Pennsylvanian, southeast Utah; **(b)** calcified tree root preserved in aeolian strata, Cedar Mesa Sandstone, Permian, southeast Utah; **(c)** plant root structures with iron concretions seen in plan view, Holocene, Iceland; **(d)** tree stump with radiating roots, Namurian, Blackburn, northern England.

Q. 16 Could the structures have resulted from the upwards or downwards movement of escaping organisms? Are the traces roughly cylindrical, subvertical 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 aggrading sequences. These traces are classed as **fugichnia** (Figs 9.53, 9.55) and there is complete overlap with resting (cubichnia) and dwelling (domichnia) burrows, and rare transitions to feeding and grazing structures (fodinichnia and pascichnia), although the last named are less likely to show rapid response to changes of interface position. Be careful to distinguish these from trace makers that were seeking equilibrium, with an emphasis on re-adjustment of position in the substrate. Examples are found in retrusive *Diplocraterion* and *Lockeia*, in *Monocraterion*, in those *Skolithos* that possess funnels or retrusive bases, and in elongate *Ophiomorpha* (see Figs 9.48, 9.49).

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. 9.53) and can sometimes be distinguished from escape structures (fugichnia). Examples include bivalve adjustment traces, *Skolithos* and *Diplocraterion*, all of which display finely spaced retrusive or protrusive spreiten (Figs 9.47, 9.48, 9.52, 9.57).

Q. 18 Could the traces result from predator/prey relationships, i.e. by bio-erosion on hard materials, for example 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. These are known as **praedichnia** (Figs 9.53, 9.55) 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, the structures of certain marine worms (polychaetes)? Such rare cases are **aedifichnia** (Fig. 9.53).

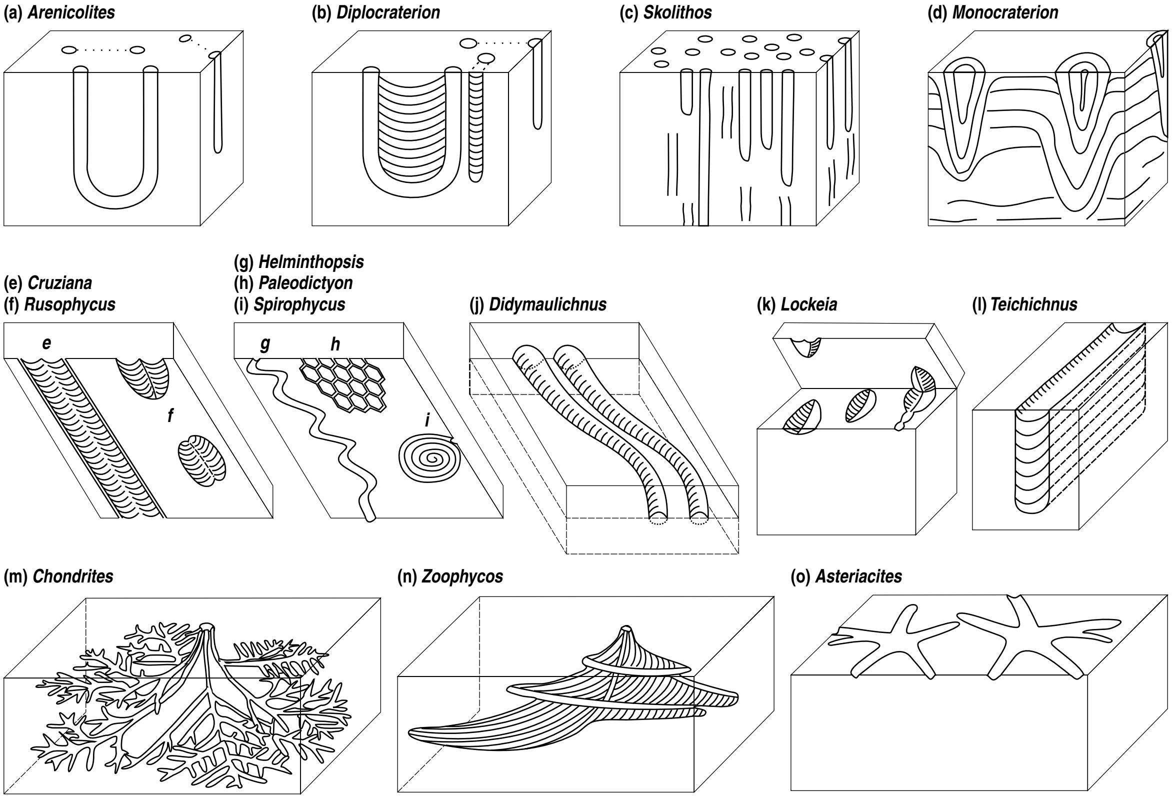
Q. 20 Could the structures be breeding places for protecting larvae or raising juveniles, for instance, bee cells, dinosaur nest egg sites or beetle brooding burrows? These rare structures are known as **calichnia** (Fig. 9.53).

Q. 21 Could the structures be influenced by microbial–chemical interactions where low levels of oxygen prevailed? Such traces are known as **chemichnia** (Fig. 9.53) and may be related to trapping and gardening (agrichnia). Examples include complex feeding traces such as *Chondrites* (Fig. 9.57).

To familiarize yourself with some of these terms, features and questions, try to sort the trace fossils depicted in Figures 9.51, 9.52 and 9.57 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

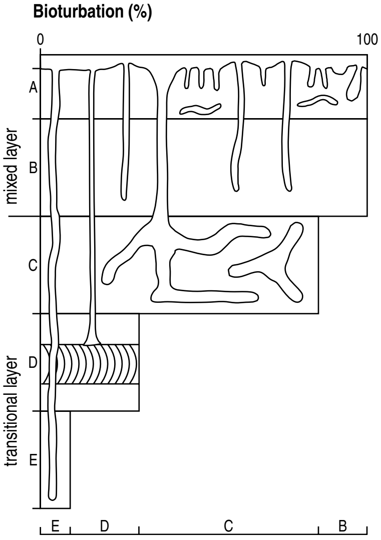
Several types of trace fossils are often found in close association with each other within the substrate, and indicate complex interactions between several groups of organisms. The plant and animal activity that takes place on and below the sediment surface within a particular environmental setting is typically characterized by a variety of trace-generating organisms that each occupy a particular level or **tier** on or within the substrate (Figs 9.58, 9.59). The infauna occupies the substrate 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 in order 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 tend to overprint those higher structures. Figure 9.60 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



(a) Arenicolites
(b) Diplocraterion
(c) Skolithos
(d) Monocraterion
(e) Cruziana
(f) Rusophycus
(g) Helminthopsis
(h) Paleodictyon
(i) Spirophycus
(j) Didymaulichnus
(k) Lockeia
(l) Teichichnus
e
f
g
h
i
(m) Chondrites
(n) Zoophycos
(o) Asteriacites

Figure 9.57 An 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 9.53, 9.54 and 9.55. 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 in Frey & Crimes in Frey 1975 and Basan et al. 1978)

Figure 9.58 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)



Bioturbation (%)
0
100
A
B
C
D
E
mixed layer
transitional layer
E
D
C
B

Processes	Activity in tier (%)	Organism/trace
Homogenization	87	Epibenthic disturbance and homogenization by burrowers
Conveyor processing	5	Specialized deposit-feeding worms
Open-burrow excavation	4	Deposit-feeding crustaceans
Backfill and spreite	2	Deep deposit-feeding worms using nutrients from below the redox level
Domichnia and inverted conveyor deposition	2	Dwellers tapping sulphide well structures

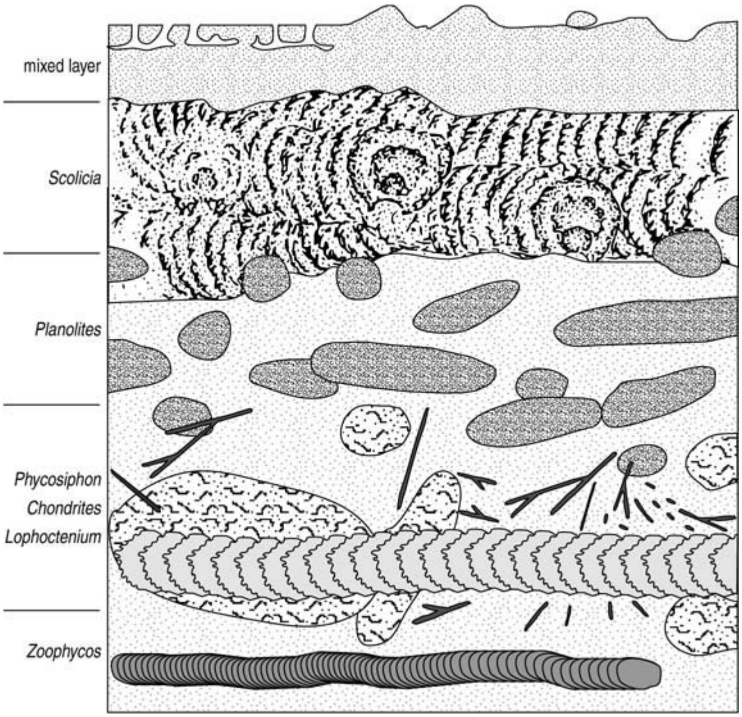


Figure 9.59 Tiering diagram of traces seen in box cores taken in 2–3.5 km water depths off northwest Africa. Surface trails and *Paleodictyon* isp. are indicated in the mixed layer. (Modified after Wetzel 1984)

activity created by tiers on or just under the sediment surface, whereas 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:

- observing and carefully drawing a representative section of an outcrop and in particular noting all cross-cutting relationships
- photographing the outcrop and individual beds at various scales
- cleaning up outcropping beds and, in extreme cases, breaking up hand specimens in order to gather further evidence
- selecting and securely labelling critical blocks with a known orientation for slabbing in the laboratory.

Breaking up and removal of material from the field should be done only in the context of a serious research programme and then only sparingly. Generally speaking, evidence is best left in place, as the context is usually critical to interpretation.

9.4.4 Ichnofabric and bioturbation

The presence and relationships of trace fossils in a rock sample make up its **ichnofabric**. **Bioturbation** is the alteration of the original sediment, which is inferred to be the result of the activity of animals and plants on, and within, the substrate while living there. Because trace fossils are sedimentary structures that, in the vast majority of 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 variations in prevailing and subsequent environmental factors. Ichnofabrics therefore 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 questions arise:

- Are the burrows very densely distributed and interpenetrating? If so, the sediment should be referred to as having a **bioturbated texture** or **ichnofabric**.

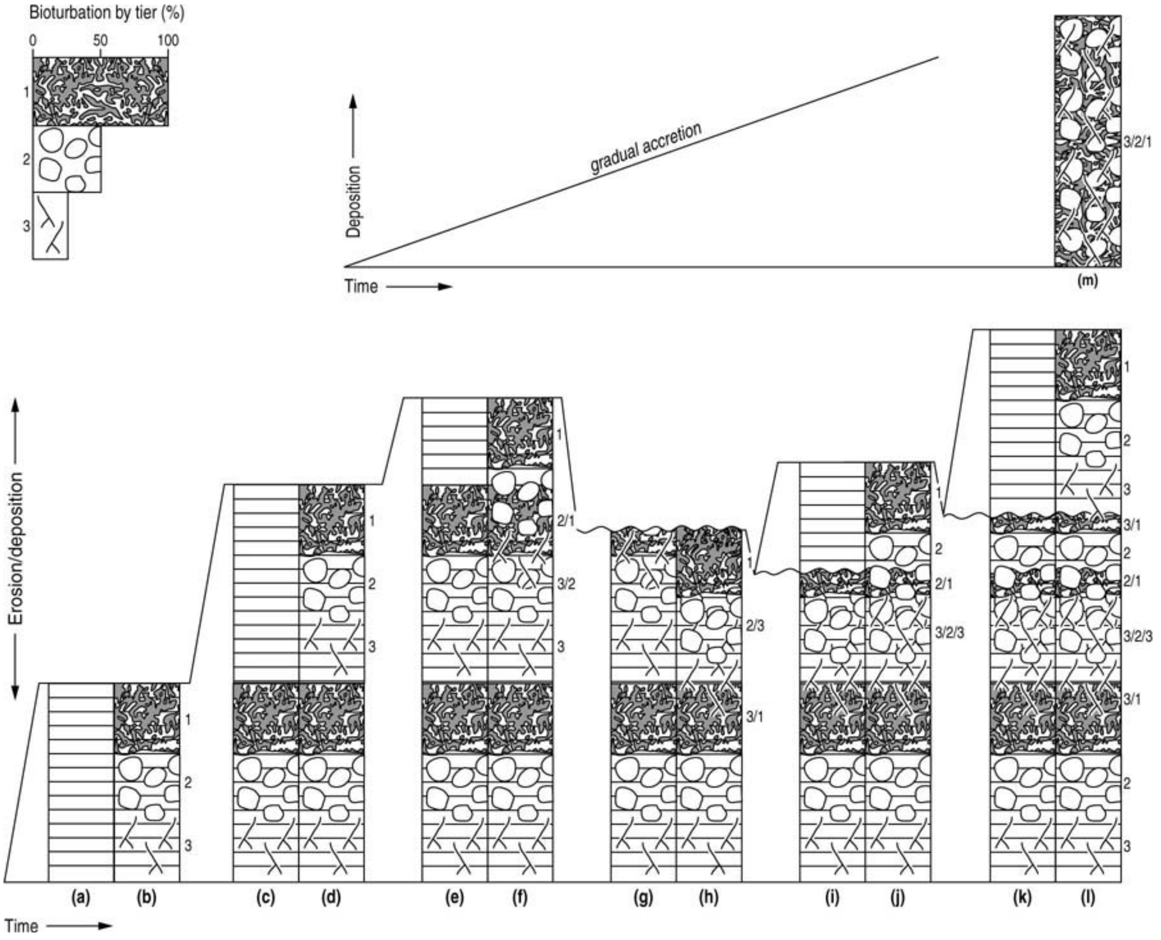


Figure 9.60 Effect of tiered bioturbation in different depositional regimes, based on the system shown top left. Deposition as a consequence of a series of discrete events produces many different effects. **(a, b)** Rapid deposition followed by non-deposition (omission) allows colonization. **(c)** Rapid burial with a package thicker than the bioturbation zone enables preservation of the first bioturbated bed. **(d)** New colonization. **(e, f)** Deposition of a thinner package leads to overprinting and produces a palimpsest fabric. **(g, 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, 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)

- Are the burrows common but indistinct? If so, the term **burrow mottling** may be more appropriate (see Fig. 9.48).
- Are the structures preserved in full relief (Fig. 9.49)?
- Is the wall of the cast different in composition from the body of the cast, as when a burrow in sand is lined by a layer or layers of mucus or faecal pellets made of mud?
- Does the trace contain internal structures, e.g. spreiten (backfill laminae) (Fig. 9.47)?
- 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. 9.60)?
- Is it possible to quantify the intensity of bioturbation through use of a bioturbation index (below) or the numbers of individual trace fossil “species” (i.e. by using a measure of ichnodiversity – see Fig. 9.60 and §9.4.5)?

Bioturbation index (BI)*	Fraction bioturbated (%)†	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

* Each grade is described in terms of the sharpness of the primary sedimentary fabric, burrow abundance and amount of burrow overlap.

† Use these percentages as a guide, not as an absolute class division.

Figure 9.61 Bioturbation index for use in determining the extent to which biogenic activity has disturbed the primary (original) sedimentary texture or fabric (modified after Taylor & Goldring 1993 and Goldring 1999).

In studying sedimentary successions, it is not uncommon to encounter sediment showing few primary sedimentary structures. Instead, a seemingly chaotic bioturbated ichnofabric may have been produced through intense or protracted animal and plant activity (Fig. 9.48). In order to observe, record and interpret such apparent confusion, it is worthwhile first to identify any remaining primary structures and then estimate the intensity of bioturbation at several key levels using a **bioturbation index** (Fig. 9.61) or **ichnofacies index** (Fig. 9.62), before describing and categorizing the whole outcrop.

Interpretation of the bioturbation index is not easy. Total bioturbation suggests that the rate of biogenic reworking exceeded that of sedimentation and that the substrate provided a suitable environment for colonization, 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 or lifestyles may account for total mixing.

Incomplete bioturbation suggests that prevented more complete reworking. This may be because of higher rates of sedimentation or a more energetic depositional regime (Fig. 9.63), although other factors may have been at work, such as salinity variations affecting biodiversity, or the non-availability of oxygen or nutrients. In order to distinguish the possible negative factors, it is necessary to identify and quantify the individual ichnogenera and to 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 erosion cuts deeper than reworking, then no bioturbation will be preserved. In shallow seas the substrate is commonly rapidly bioturbated in 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.

Some general environmental interpretations may be usefully drawn from these observations. Bioturbation tends to homogenize the grain-size differentiation that defines laminae (Fig. 9.63a). Destruction of preferred grain orientation by bioturbation can change the porosity and permeability structure of the sediment. If coarse and fine laminae become mixed, sediment maintains 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. Where sediment is bound by roots and certain burrow structures, it becomes stabilized. When 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 and ichnoguilds

In addition to considering the ichnofabric and the degree of bioturbation within a succession, it is also desirable to

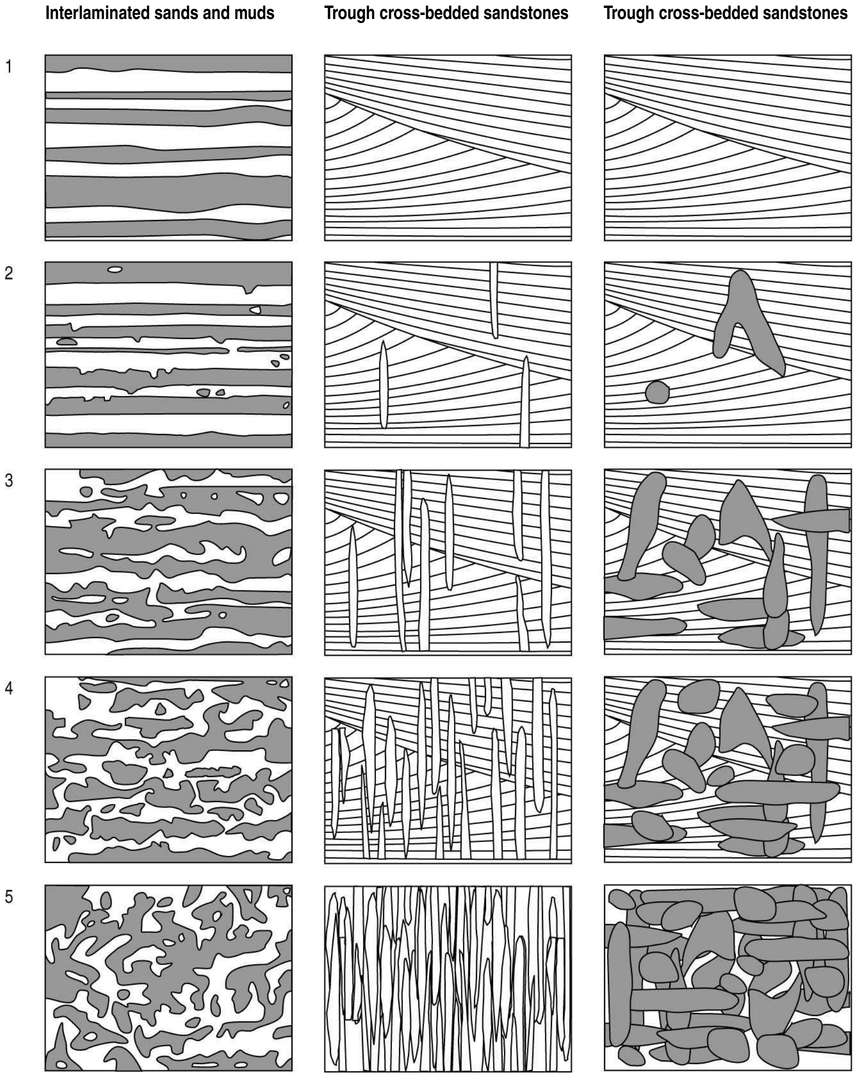
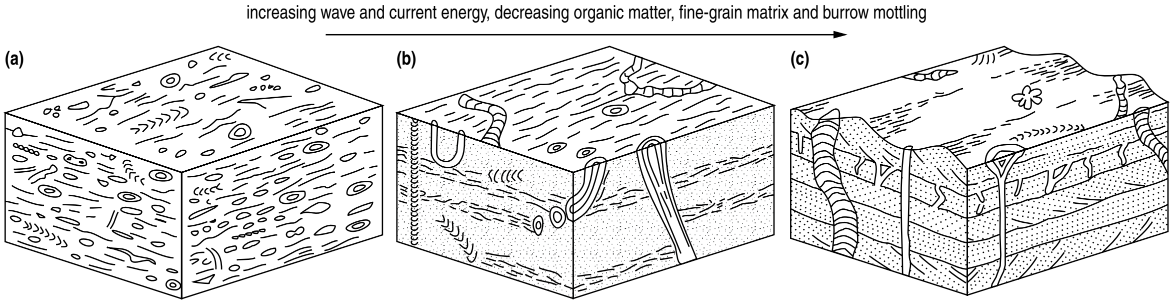

Interlaminated sands and muds
Trough cross-bedded sandstones
Trough cross-bedded sandstones
1
2
3
4
5

Figure 9.62 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 & Bottjer 1986)

Figure 9.63 (below) 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).


increasing wave and current energy, decreasing organic matter, fine-grain matrix and burrow mottling
(a)
(b)
(c)

Siltstone with highly mottled textures; trace fossils compressed and few taxonomically identifiable. Few, if any, tracks and trails developed on bedding-plane surfaces.
Bioturbation index: 5–6.

Very fine-grain sandstone, larger but abundant mottles and containing individually recognizable well developed burrows. Tracks and trails developed on bedding-plane surfaces.
Bioturbation index: 3–4.

Cleaner, coarser sandstone showing plane and cross bedding; absence of mottled texture and many recognizable borrows. Tracks and trails developed on bedding-plane surfaces.
Bioturbation index: 1–2.

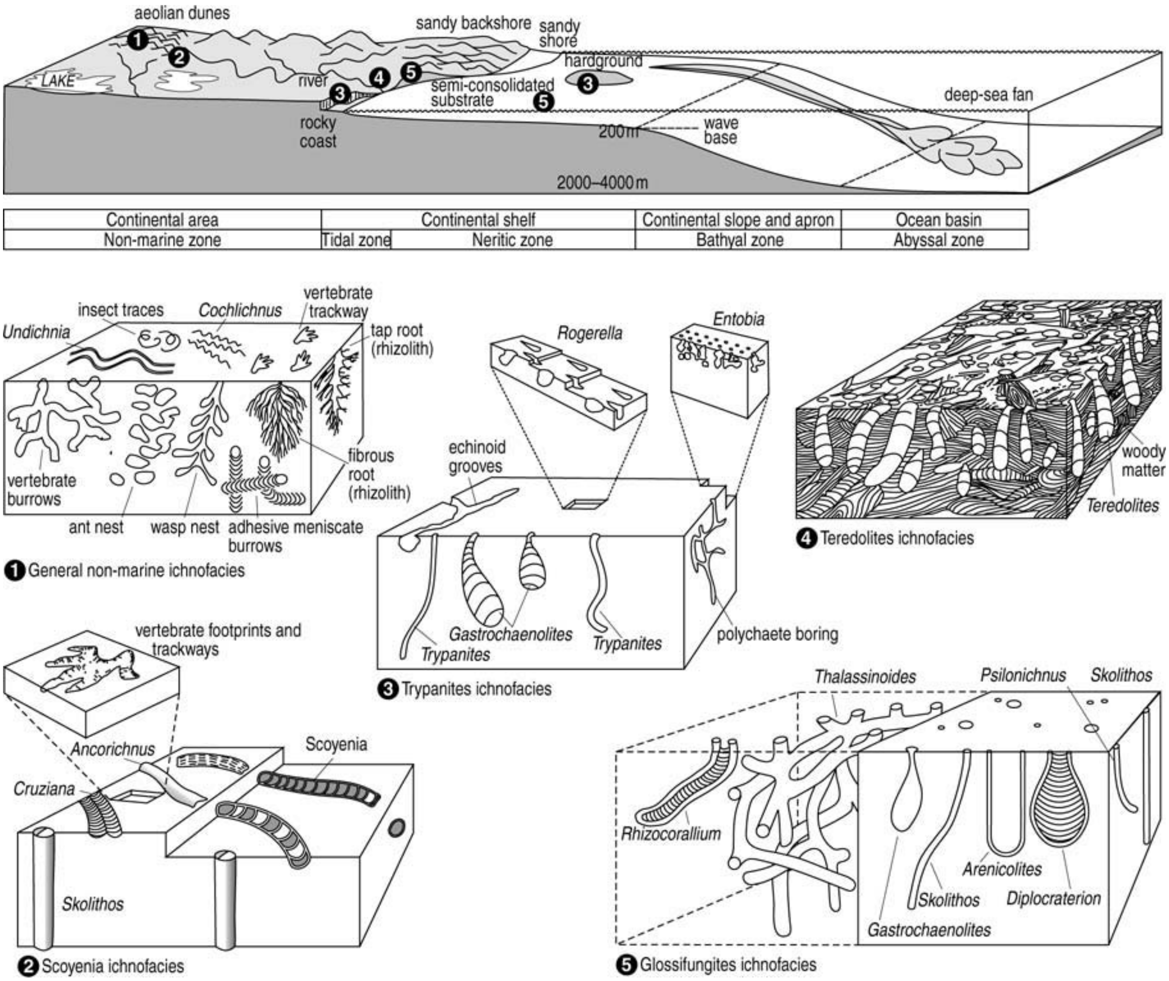

aeolian dunes
sandy backshore
sandy shore
hardground
LAKE
river
semi-consolidated substrate
rocky coast
200m
wave base
deep-sea fan
2000–4000m
Continental area
Non-marine zone
Continental shelf
Tidal zone
Neritic zone
Continental slope and apron
Bathyal zone
Ocean basin
Abyssal zone
insect traces
Cochlichnus
vertebrate trackway
Undichnia
tap root (rhizolith)
fibrous root (rhizolith)
vertebrate burrows
ant nest
wasp nest
adhesive meniscate burrows
1 General non-marine ichnofacies
Rogerella
Entobia
echinoid grooves
Gastrochaenolites
Trypanites
Trypanites
polychaete boring
3 Trypanites ichnofacies
woody matter
Teredolites
4 Teredolites ichnofacies
vertebrate footprints and trackways
Ancorichnus
Scoyenia
Cruziana
Skolithos
2 Scoyenia ichnofacies
Thalassinoides
Psilonichnus
Skolithos
Rhizocorallium
Arenicolites
Skolithos
Diplocraterion
Gastrochaenolites
5 Glossifungites ichnofacies

Figure 9.64 (a, above and opposite) 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. comm.), Pemberton 1992 and Hasitosis 2002).

determine the **ichnodiversity** of a rock unit 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 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 and the preserved ichnodiversity may not necessarily 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 several organisms can result in the generation of only one trace.

The description and analysis of ichnofabrics produced by tiered communities often reveals a series of repeated patterns that are defined by the types of trace fossils present. These reflect the activity level in the substrate and the feeding styles of their producers. The term **ichnoguild**, adapted from the word "guilds" as used in ecological studies, is used to encompass the grouping of such trace-fossil structures. It emphasizes the biogenic origin of trace fossils and highlights groups of ichnospecies that record exploitation of a particular environmental resource. The definition of ichnoguilds involves considering three aspects of the habitat structure of a community in space and time:

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	Oxidizing	Oxidizing	Oxidizing	Oxidizing	Oxidizing
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 ratio similar, 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) Subsurface 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-shape to irregular dwelling borings of suspension feeders or passive carnivores Raspings and gnawings of algal grazers Borings oriented perpendicular to the substrate	Club-shape borings with walls ornamented with the texture of the host substrate (e.g. tree-ring impressions) Stumpy to elongate subcylindrical excavations in marine settings Shallower, etchings in non-marine settings	Arthropods, molluscs, echinoderms, corals, “worms” Vertical and inclined ear-shape burrows Mostly suspension feeders Protrusive spreiten due to animal growth

Figure 9.64 (a, continued)

- the food sources of the likely organisms: whether they were deposit feeders, suspension feeders, gardeners, trappers or farmers
- their utilization of space in the tier represented
- whether the structures were produced: as semi-permanent or stationary branching burrows (as inhabited by static deposit feeders – fodinichnia; permanent dwellers – domichnia; or permanent burrow dwellers that trapped and gardened for food – agrichnia); or by more mobile organisms occupying temporary burrows (as produced by detritus and deposit feeders – pascichnia, and involving locomotory trails – repichnia, or resting traces – cubichnia).

Ichnoguilds are named after a characteristic ichnogenus, which does not necessarily need to be present in every instance. Identification of an ichnoguild is generally possible only when the tiering system of the deposit has been deduced. The ***Chondrites–Zoophycos* ichnoguild** is represented by deep-tier deposit-feeder structures generated by relatively static organisms. The ***Thalassinoides* ichnoguild** is represented by mid-tier deposit-feeder structures generated by more mobile organisms. The ***Planolites* ichnoguild** is recognized by the presence of shallow-tier deposit-feeder structures generated by mobile organisms. The ***Phycosiphon* ichnoguild** consists of deep deposit-feeder structures, cutting deeper than the *Thalassinoides* ichnoguild but less deep than the *Zoophycos* ichnoguild (it can occupy shallower tiers in opportunistic situations, for example where storm beds have been generated). The ***Skolithos–Ophiomorpha* ichnoguild** comprises suspension-feeder structures that penetrate relatively deeply into the unstable sandy substrates of high-energy environments.

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 be a direct reflection of environmental conditions such as water depth, salinity and substrate character (e.g. softgrounds, firmgrounds and hardgrounds). Several distinct ichnofacies are recognized, each of which is thought to record a particular set of environmental conditions (Fig. 9.64). Although each ichnofacies has been named after a representative ichnogenus, that particular ichnogenus does not necessarily have to be

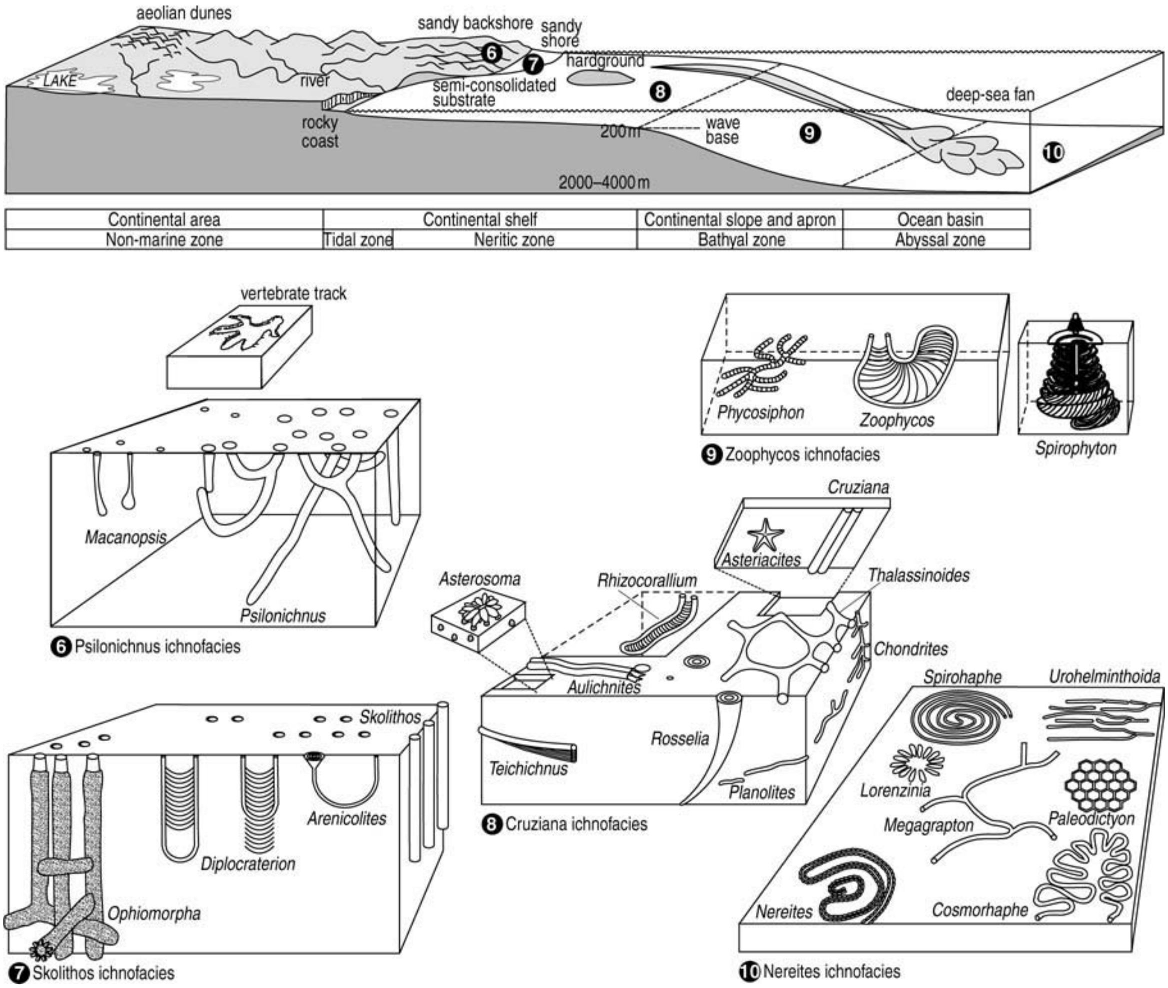

aeolian dunes
sandy backshore
sandy shore
6
7
hardground
LAKE
river
semi-consolidated substrate
8
deep-sea fan
rocky coast
200 m
wave base
9
10
2000–4000 m
Continental area
Continental shelf
Continental slope and apron
Ocean basin
Non-marine zone
Tidal zone
Neritic zone
Bathyal zone
Abyssal zone
vertebrate track
Macanopsis
Psilonichnus
6 Psilonichnus ichnofacies
Phycosiphon
Zoophycos
9 Zoophycos ichnofacies
Spirophyton
Cruziana
Asteriacites
Thalassinoides
Asterosoma
Rhizocorallium
Chondrites
Aulichnites
Rosselia
Teichichnus
Planolites
8 Cruziana ichnofacies
Skolithos
Arenicolites
Diplocraterion
Ophiomorpha
7 Skolithos ichnofacies
Spirohaphe
Urohelminthoida
Lorenzinia
Megagrapton
Paleodictyon
Nereites
Cosmorhaphe
10 Nereites ichnofacies

Figure 9.64 (b, above and opposite) 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. comm.), Pemberton 1992 and Hasitosis 2002).

present in every example of that facies. A fundamental prediction of an actualistic approach is that these or similar traces would have formed whenever the particular set of conditions recurred. This prediction appears to be justified by the occurrence of trace fossils throughout the Phanerozoic rock record.

Once processes relating to the formation of traces have been recognized, it is possible to combine sedimentological and ichnological ideas to determine more effectively the kind of environment and deposit in which the traces were formed. Substrate consistency and energy levels, in particular, can often be determined quite well from traces of behaviour in the sediments. However, other, more biological, environmental factors are less readily discerned.

Four commonly applied ichnofacies categories are based on energy levels and loosely on water depth in the marine realm (Fig. 9.64b, 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

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	Oxidizing	Oxidizing	Oxidizing	Oxygen reduced, highly organic content	Limited oxygen (influx from density flows), not usually anoxic
Salinity	Variable: fresh, brackish, normal marine, rarely hyper-saline	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 soft-ground, 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-shape 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 9.64 (b, continued)

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. Another commonly applied ichnofacies category is the **Glossifungites ichnofacies**, which is characteristic of firmground to hardground surfaces subject to periods of non-deposition, mostly in the marine realm. The **Trypanites ichnofacies** characterizes borings in rockground (i.e. lithified) substrates. The **Teredolites ichnofacies** is characteristic of woody substrates (woodgrounds); for example, logjams in fluvial delta-top settings. These ichnofacies, together with the typical assemblages of ichnogenera by which they are identified, and the main environmental conditions that they represent, are depicted in Figure 9.64.

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 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. 9.64a, 1). Additionally, an **Arenicolites ichnofacies** is recognized for opportunistically colonized beds, which have arisen as a result of some rare sand–silt-depositing events in either lacustrine or marine situations.

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 sets of physical and chemical conditions.

Traces reflect many of the processes that are the basis of the uniformitarian and actualistic models for 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. 9.52). They may give an indication of substrate consistency (Figs 9.51, 9.64) and degree of aeration at the time of activity, as well as the degree to which the sediment has been reworked (Fig. 9.56). Some trace fossils are valuable indicators of palaeocurrent where certain burrows and resting traces have a preferred orientation. The organisms that generated them were aligned with the current, either for feeding purposes or as an aid to stability (e.g. orientated forms of *Lockeia*, Fig. 9.57k).

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 inorganic depositional structures, trace fossils may provide the only clues about 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 sets of ecological conditions and substrates (Fig. 9.64). Their abundance in clastic rocks, which often lack body fossils because of the dissolution of shells, means that the palaeoenvironments of the rocks can be interpreted in the light of organic activity, which would not otherwise be recognized. The long 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. They increase the diversity of fossil groups known from the geological record. This provides a wider sample of former life forms and encourages the understanding of the evolution of fossil behaviour. They help to elucidate problems in late Precambrian rocks, where body fossils are generally 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 has contributed to the recognition of such important events in the history of life as invasion of land by animals in the early Palaeozoic. The presence of traces in coarse-grain 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 Trias), 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. 9.57). Quantitative estimates of compaction and deformation depend upon the recognition of objects of known shape prior to deformation (Fig. 9.65). Burrows can be such objects, although 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-shape burrows are particularly useful because they allow the reconstruction of strain ellipses and aid quantitative estimates of strain attributable to pre-cleavage compaction, compression, rotation and cleavage distortion.

In **geotectonics**, associations of traces may help to define faunal provinces. Certain distinctive traces in the late Cambrian to early Ordovician, for example, are attributable to trilobites of an Atlantic province, others to a Pacific province. This evidence has contributed to current views concerning the separation of these provinces at that time by an Iapetus Ocean.

In **stratigraphy**, the long time range of many trace fossils restricts their use for biostratigraphical or chronostratigraphical purposes, but short-ranging forms such as certain forms 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 Trias. These two examples testify to rapid evolution during these times and hence the correlations may be very reliable in some situations. Trace

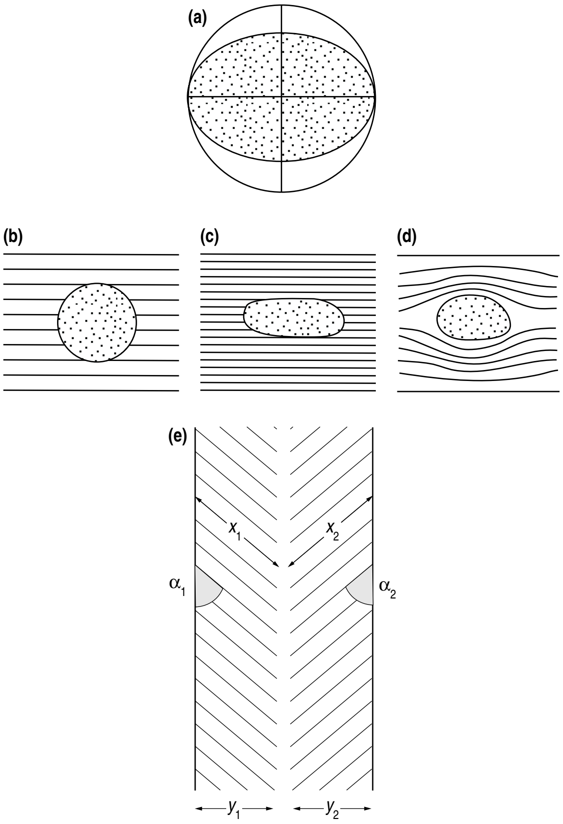


Figure 9.65 The use of trace fossils in deformational studies. **(a)** The elliptical outline of a burrow (stippled) may be used to determine the percentage compaction by the construction of a circle to represent the outline of the undeformed burrow. A comparison of the axial lengths of the undeformed 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 the amount of compaction of the host sediment. **(d)** The 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 of coarse grain. **(e)** The effects of tectonic deformation on *Cruziana*. In the undeformed state, normally $x_1 = x_2$; $\alpha_1 = \alpha_2$; $Y_1 = Y_2$. 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. (After Crimes in Frey 1975)

fossils also help to generate and control palaeogeographical reconstructions for individual stratigraphical time periods. Since the late 1970s the concepts and uses 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, tectonic setting, economic basement, and the extents of source, reservoir and caprocks. 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, somewhat argillaceous, 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. Trace-fossil studies have potential in relation to determining the mass properties of sediments required by engineers tackling problems of shifting coastal morphology.

9.4.8 Confusion of traces with inorganic sedimentary structures

Traces may easily be confused with a wide range of 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 or skip cast; see §4.2.3). 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 sandhoppers), conical fracture patterns (feeding burrows such as *Phycodes*), sand volcanoes, and fulgurites (sediment structures generated by lightning strikes). In other cases, current crescents have been described as the trace fossil *Blastophycus*, rill marks as *Dendrophycus*, sinuous hierarchical mudcracks as *Manchuriophycus*, small interference ripples as tadpole nests, and convolute or contorted lamination as bioturbation. The possibilities are endless and 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 de-watering Physical deformation structures associated with de-watering 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 0.15–0.2 m. The rapid addition of a 0.15–0.2 m 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 0.05–0.1 m. 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 induce 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 (0.05 m wide) glass tank with alternate layers of damp mud, silt and sand (each 0.01–0.03 m 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 9.61, 9.62). 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.

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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. Although the ability to interpret sediments in these terms is useful in its own right, it is often more important to use that information as a step towards interpreting the depositional environment of sediments found in the rock record. In earlier chapters little mention was made of environments. This omission was deliberate so as to highlight the fact that many structures and processes are common to a range of environmental settings. However, it is necessary to determine the processes responsible for generating a particular set of structures as a first step in making an environmental interpretation. In order to move from an interpretation of process to one of environment, further analysis is required. This involves trying to establish spatial and temporal relationships of the processes that can be deduced from the sedimentary structures, as these relationships can help to narrow the range of environmental possibilities. It is also useful to know something of the directional properties of sedimentary structures so that 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.

Therefore, in characterizing a modern environment or establishing an environmental interpretation for sedimentary rocks, 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 for 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 are probably the product of 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. 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 sequence of sediment that will be generated by a particular set of 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 occur low in a vertical

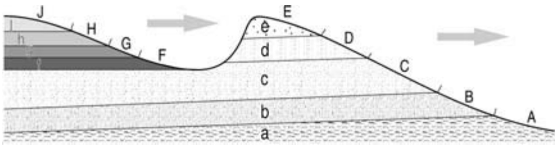


Figure 10.1 A schematic diagram to illustrate Walther's principle of succession of facies. Sub-environments A–E are on a sloping surface that is building out to the right, generating lithological units a–e. A channel comprising sub-environments F–J is cut into the top of this topography and is migrating via lateral accretion in the same direction, and generates lithological units f–j. The boundary between lithological units c and f represents a break in deposition.

sequence, with structures from successively higher topographical 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 sequence 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 particularly important in many environments (e.g. in intertidal settings) is the activity of burrowing animals. Animals that live below a 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 now 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 sequence, 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 sequences

Many environmental interpretations of sedimentary rock sequences rely heavily on measured sections through the sedimentary succession (Fig. 10.2). Such sections can give a record of changing sedimentary processes through time, an important clue to the nature of the environment and 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 lithological units, which may be based on grain size or on compositional differences. The simplicity or complexity of the scheme chosen will depend upon the nature of the succession itself, the eventual aims of the exercise and the refinement or resolution of the interpretation being attempted.

Having established a basis for lithological subdivision, 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 the 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 sequence is accurately recorded. For example, where beds are conspicuously lenticular, it is important to record their average thicknesses rather than maximum values, as recording the latter would introduce a systematic error that would exaggerate the total thickness of the succession.

The features recorded will vary with the nature of the sequence 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.

The feature of measured sections most commonly ignored is the nature of the contacts between units. Some contacts are gradational, sometimes to a degree 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 whenever a sharp contact is seen, although, of course, not all sharp contacts are erosive.

Recording your obervations from measured (optional) vertical sequences demands a disciplined method of working. Some geologists prefer to draw a graphic log while in the field, either in their notebooks or on

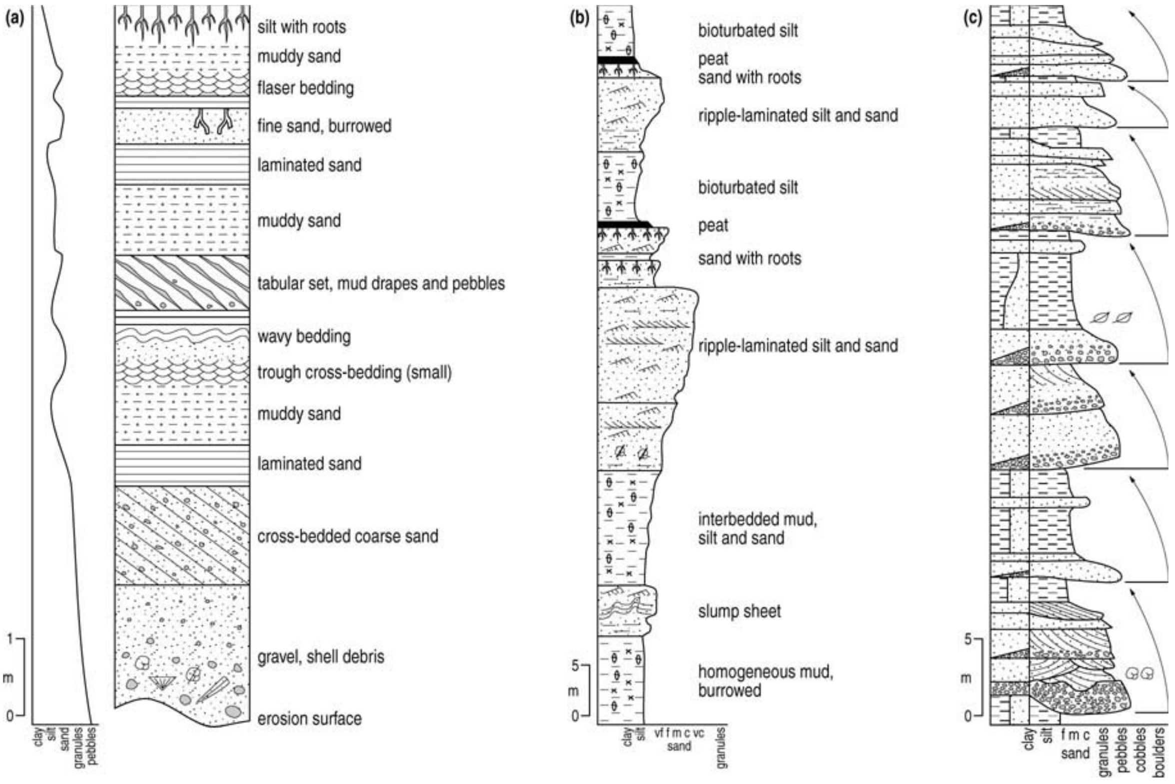

(a)
silt with roots
muddy sand
flaser bedding
fine sand, burrowed
laminated sand
muddy sand
tabular set, mud drapes and pebbles
wavy bedding
trough cross-bedding (small)
muddy sand
laminated sand
cross-bedded coarse sand
gravel, shell debris
erosion surface
1
m
0
clay
silt
sand
granules
pebbles
(b)
bioturbated silt
peat
sand with roots
ripple-laminated silt and sand
bioturbated silt
peat
sand with roots
ripple-laminated silt and sand
interbedded mud, silt and sand
slump sheet
homogeneous mud, burrowed
5
m
0
clay
silt
vf f m c vc
sand
granules
(c)
5
m
0
clay
silt
f m c
sand
granules
pebbles
cobbles
boulders

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 & 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-upwards units (cycles) within the overall succession are shown by arrows. Based in part on Surlyk (1978). See Appendix 6 for a guide to the symbols used.

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 while in the field can be time consuming, leaving less time for observing the rocks. 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, and scaled orientated diagrams, specimens and photographs can 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. However, some styles of graphic logs are more easily understood than others. Examples of fairly straightforward 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 clearly shown. Where units of thinly interbedded sandstone and mudstone occur, it is often useful to indicate the proportions of the components in a separate “lithology” column (e.g. Fig. 10.2c).

Palaeocurrent measurements are most effectively recorded opposite the units from which they were taken. A well drawn graphic log incorporating all these features serves as a sound basis for environmental interpretation and also enables others to use the data to suggest alternative or more refined interpretations.

Where a sequence 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 particular surfaces to an appropriate level. This is best achieved through the construction of scaled panels that depict the geometry and architecture of major stratal 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 possible to link the sections only by correlating the most confidently identified bedding surfaces. It is important to take particular care when correlating sandstones or coarser units, especially if there is any evidence that they may be lenticular (e.g. channelized). Correlation of similar sandstones at similar positions in a sequence 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.

10.4 Interpretation of vertical sequences in rocks

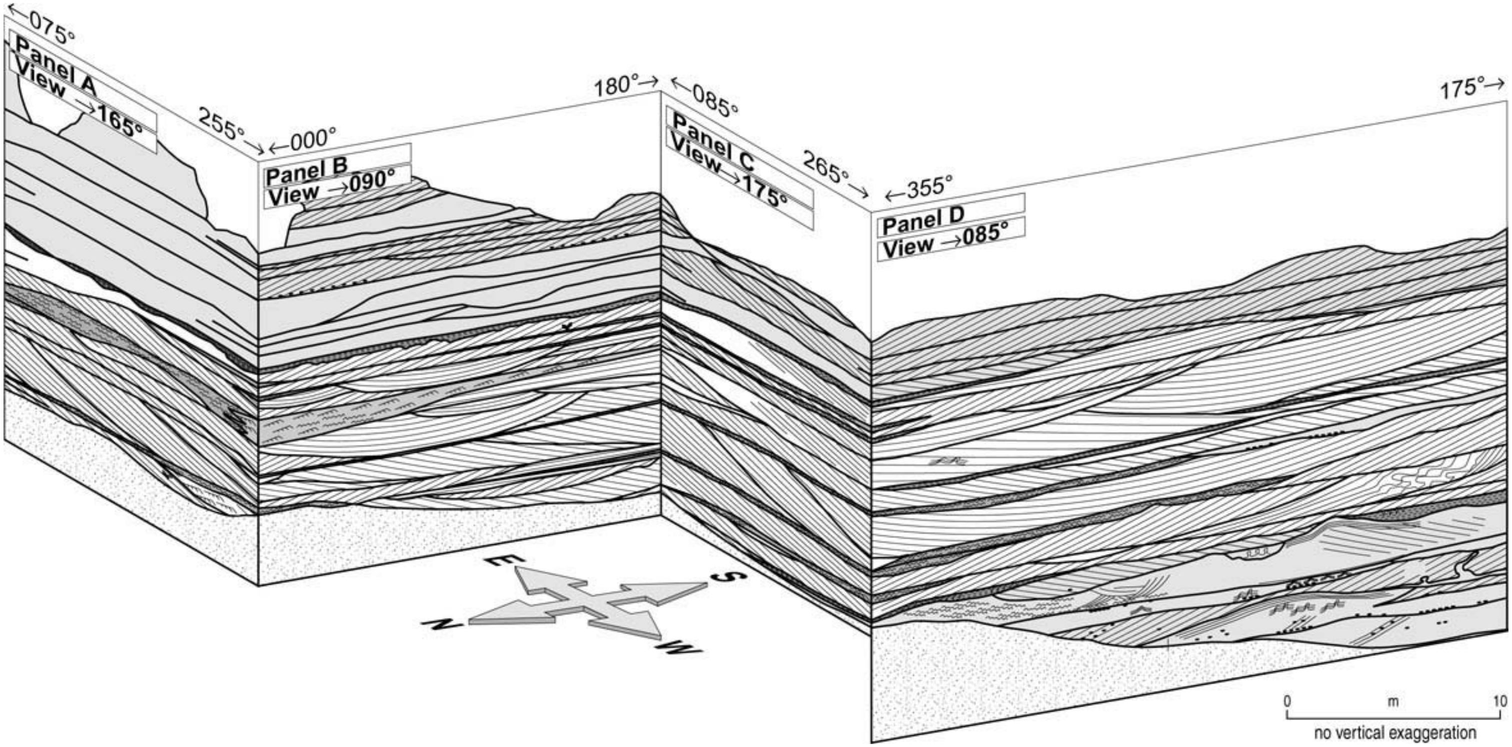
The interpretation of sedimentary successions in terms of their environment of deposition is one of the main aims of sedimentology. We have seen how most sedimentary structures allow interpretation of processes of erosion, deposition or post-depositional alteration, and how many of these structures and the processes responsible for their generation occur across a range of environments. One of the main starting points for moving the discussion towards inferring an environment of deposition is the *succession* of processes deduced from the vertical succession 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 allow us to develop 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 may tell us if a sequence is deep marine, shallow marine, marginal marine or non-marine.

In some, cases, it may be possible to use body fossils to make preliminary inferences about water depth. Before this is attempted, it is important to establish 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 possible to infer a shallower shelf setting for a shelly fauna and a deeper-water setting for a more pelagic fauna. Palaeontological expertise is normally required to carry such arguments forward to a more sophisticated level. The environmental interpretation of Precambrian sediments is greatly restricted by the lack of this basic information and it is often difficult in some examples even to decide between a continental and a shallow-marine origin. Evidence of sub-aerial exposure of the sediment surface in the form of rain-pit casts, desiccation mudcracks, seat earths and soil profiles (see Ch. 9) can help to limit the range of environmental possibilities.

When the broad environmental context has been narrowed down by such considerations, a more detailed analysis can proceed, based on the nature of the vertical sequence, among other things.

The law of superposition tells us that the vertical sequence of lithologies 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 distinctive sedimentary units. For example, a deep basin normally receiving fine-grain sediment from suspension may have this background condition punctuated by the arrival of



←075°
Panel A
View →165°
255°→
←000°
Panel B
View →090°
180°→
←085°
Panel C
View →175°
265°→
←355°
Panel D
View →085°
175°→
E
S
N
W
0
m
10
no vertical exaggeration

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 stratal geometry and the arrangement of bounding surfaces. (Modified after Mountney & Thompson 2002)

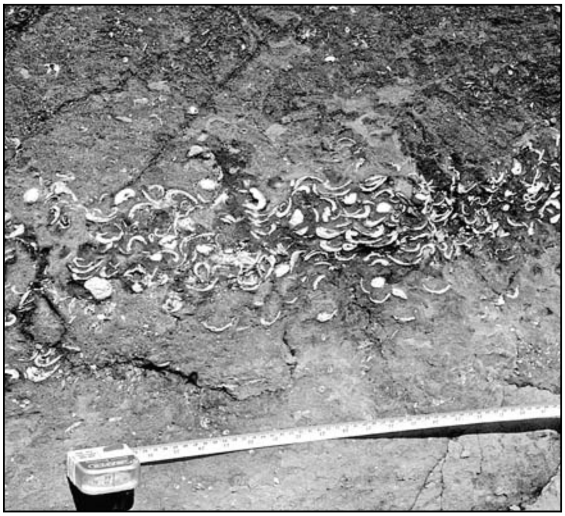


Figure 10.4 Sandstone bed with a basal lag of shelly debris. Many of the shells are broken and randomly orientated (i.e. not in life position), indicating that they have been transported to the site of accumulation. Locality unknown.

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" and "catastrophic" deposition (see §2.2.3).

In the second case, environmental conditions remain essentially constant through time, but there is a *spatial segregation* of processes and products within the environment. A gradual shifting of the environment through time thus leads to a vertical sequence of changing lithology and structure. 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 gradational non-erosive contacts. The channel system on top also generates its own gradational succession of lithologies f–j above an erosion surface, because of 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 recognizing the spatial relationships of the deposits and their depositional settings, it is also common practice to look for 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 as the product of a prograding 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 assemblage of fossils and trace fossils (e.g. marine or fresh water).

Similarly, the upper succession (Fig. 10.1f–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 and bar 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, this more precise interpretation again depending on the specific details of the structures (e.g. the presence or absence of clay drapes), the directional properties of palaeocurrents (unidirectional or bipolar) 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 10.6. These predictable successions have been interpreted to be the product of sedimentary evolution in a range of depositional environments, but in each case the techniques used in their recognition and interpretation remain the same.

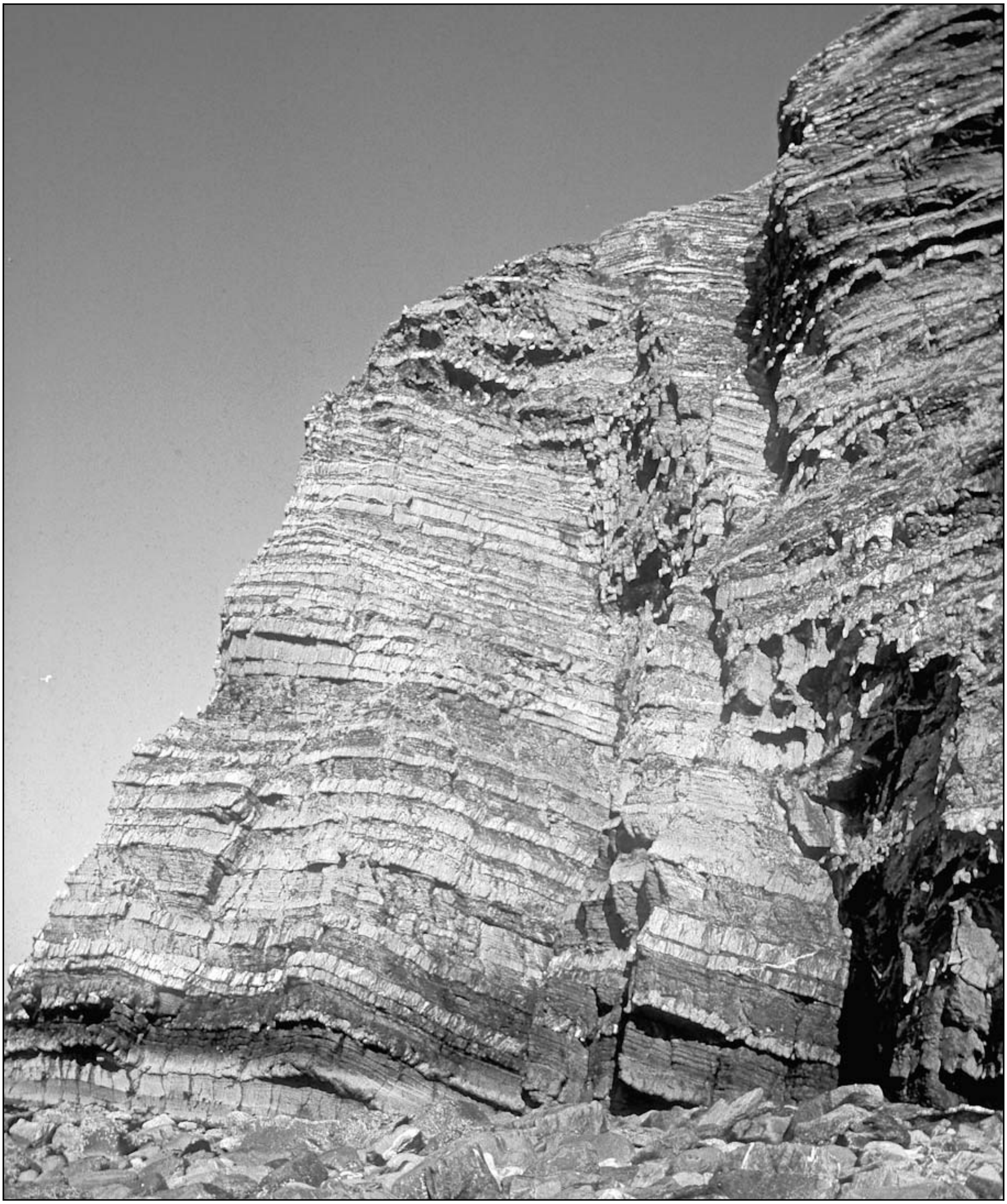
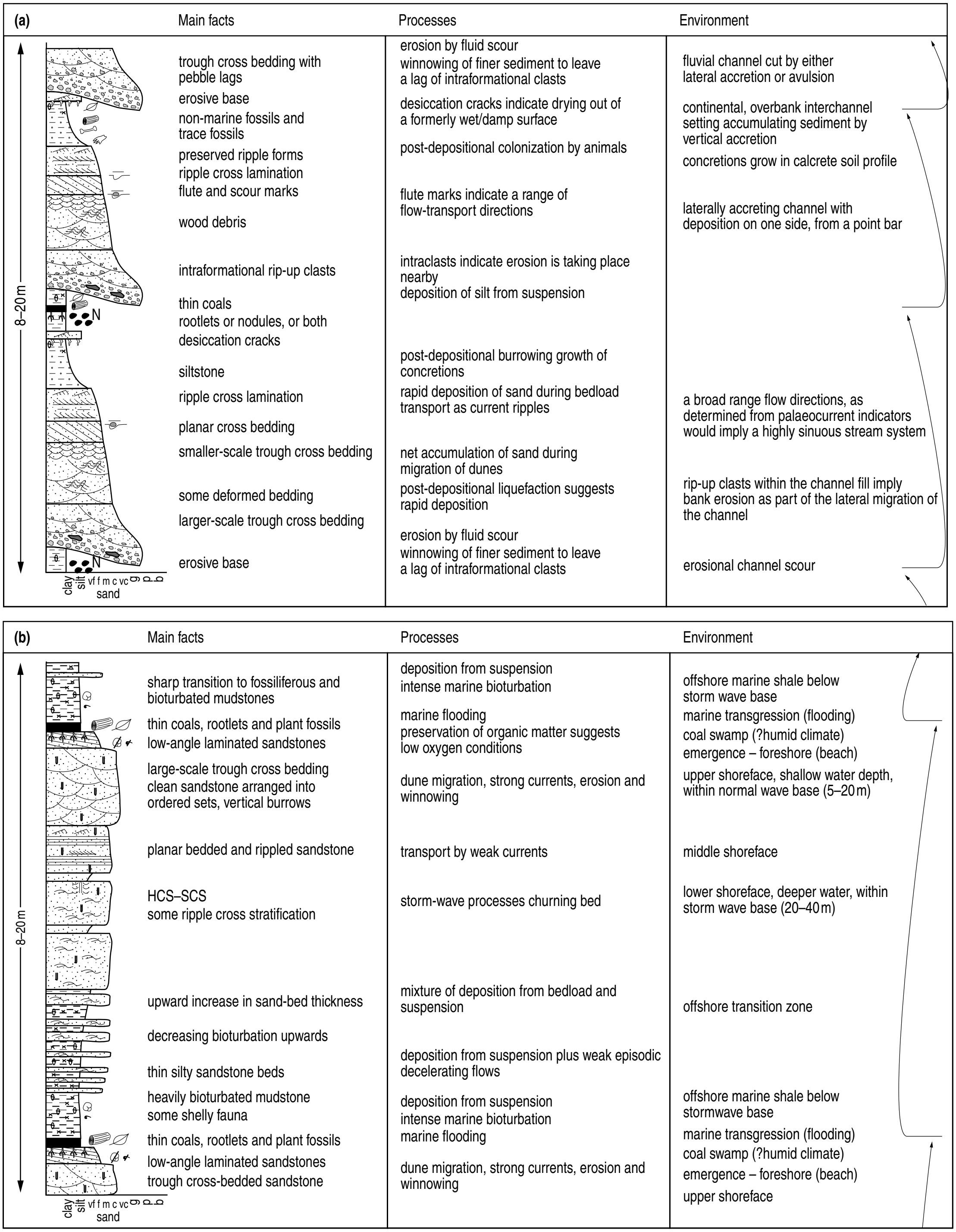


Figure 10.5 Interbedded succession of sandstones and shales, interpreted to be the result of periodic high-energy but short-lived turbidity flows carrying sand down a slope into a normally quiet deepwater environment. The succession is inferred to represent the products of alternating "normal" and "catastrophic" deposition. Aberystwyth Grits, Silurian, Wales. (Photo courtesy of Gilbert Kelling)



(a)

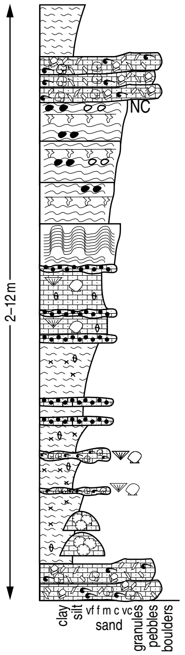
Main facts	Processes	Environment
	erosion by fluid scour	
trough cross bedding with pebble lags	winnowing of finer sediment to leave a lag of intraformational clasts	fluvial channel cut by either lateral accretion or avulsion
erosive base	desiccation cracks indicate drying out of a formerly wet/damp surface	continental, overbank interchannel setting accumulating sediment by vertical accretion
non-marine fossils and trace fossils	post-depositional colonization by animals	
preserved ripple forms		concretions grow in calcrete soil profile
ripple cross lamination		
flute and scour marks	flute marks indicate a range of flow-transport directions	laterally accreting channel with deposition on one side, from a point bar
wood debris		
intraformational rip-up clasts	intraclasts indicate erosion is taking place nearby deposition of silt from suspension	
thin coals		
rootlets or nodules, or both		
desiccation cracks	post-depositional burrowing growth of concretions	
siltstone		
ripple cross lamination	rapid deposition of sand during bedload transport as current ripples	a broad range flow directions, as determined from palaeocurrent indicators would imply a highly sinuous stream system
planar cross bedding		
smaller-scale trough cross bedding	net accumulation of sand during migration of dunes	
some deformed bedding	post-depositional liquefaction suggests rapid deposition	rip-up clasts within the channel fill imply bank erosion as part of the lateral migration of the channel
larger-scale trough cross bedding	erosion by fluid scour	
erosive base	winnowing of finer sediment to leave a lag of intraformational clasts	erosional channel scour

(b)

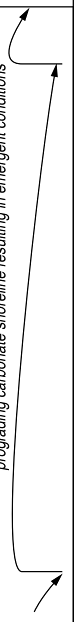
Main facts	Processes	Environment
	deposition from suspension	
sharp transition to fossiliferous and bioturbated mudstones	intense marine bioturbation	offshore marine shale below storm wave base
	marine flooding	marine transgression (flooding)
thin coals, rootlets and plant fossils	preservation of organic matter suggests low oxygen conditions	coal swamp (?humid climate)
low-angle laminated sandstones		emergence – foreshore (beach)
large-scale trough cross bedding clean sandstone arranged into ordered sets, vertical burrows	dune migration, strong currents, erosion and winnowing	upper shoreface, shallow water depth, within normal wave base (5–20 m)
planar bedded and rippled sandstone	transport by weak currents	middle shoreface
HCS–SCS some ripple cross stratification	storm-wave processes churning bed	lower shoreface, deeper water, within storm wave base (20–40 m)
upward increase in sand-bed thickness	mixture of deposition from bedload and suspension	offshore transition zone
decreasing bioturbation upwards		
thin silty sandstone beds	deposition from suspension plus weak episodic decelerating flows	
heavily bioturbated mudstone some shelly fauna	deposition from suspension intense marine bioturbation	offshore marine shale below stormwave base
thin coals, rootlets and plant fossils	marine flooding	marine transgression (flooding)
low-angle laminated sandstones		coal swamp (?humid climate)
trough cross-bedded sandstone	dune migration, strong currents, erosion and winnowing	emergence – foreshore (beach)
		upper shoreface

Figure 10.6 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. **(c)** based in part on Pratt et al. (1992). See Appendix 6 1 for a guide to the symbols used.

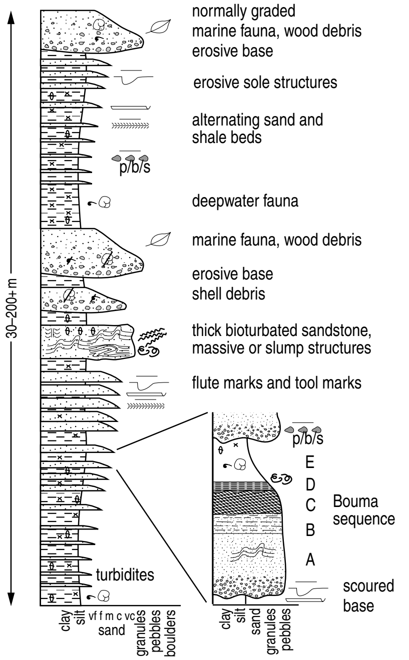
(c)



Main facts	Processes	Environment
wackestone and mudstone	low energy	
		flooding surface – marine transgression
flat pebble intraclasts	high-energy erosion and redeposition	
broken shell debris		
dissolution cavities	nodule growth – high-salinity dissolution	supratidal flat in arid environment
desiccation cracks	periodic drying-out of surface	microbial mat development around high tide level
	microbial colonization and growth	
nodular anhydrite	physical and chemical microbial–sediment interactions	
microbial laminates		
dome-shape stromatolites	colonization 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 low-energy lagoon setting
peloidal grainstone		
wackestone and mudstone	colonization of muddy substrate by shelly and soft-body fauna	
peloidal grainstone		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 support 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)



Main facts	Processes	Environment
normally graded		
marine fauna, wood debris	erosive channel cut and fill	inner-fan channel fill
erosive base		
	scour structures and toolmarks indicate moderate to strong unidirectional currents	mid-fan lobes
erosive sole structures		
alternating sand and shale beds		
	sedimentation from suspension and via dilute turbidity currents	distal outer fan
deepwater 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	
erosive base		inner-fan channel fill
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 down slope 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 (supra-fan lobes)
	sedimentation from suspension and via dilute turbidity currents (Bouma C–E)	
		distal outer fan

10.5 Key stratigraphical surfaces

As well as local erosion surfaces (e.g. resulting from scour at the base of a channel), there are other surfaces that punctuate sedimentary successions and also lead to the need to restart an interpretation along the lines of Walther's principle. These surfaces may be extremely extensive and are commonly referred to as **key surfaces** that can be used for widespread correlation. Key surfaces are usually generated as responses to external influences on sedimentation and they commonly reflect the interactions of sea-level change, tectonic subsidence, sediment supply and, particularly in the case of non-marine environments, climate change. Together these control the production of **accommodation space**, the capacity of an area to accumulate sediment. These interactions are complex and are encompassed in the field 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 the period of reduced sedimentation or condensation (§9.4). Flooding surfaces can have a wide range of 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 sediment supply. Secondly, 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 down deeply into the landscape to produce incised valley systems. These valley-shape 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 stratigraphical implications thought through. These surfaces, if correctly identified and interpreted, can be the key to stratigraphical understanding, a basis for correlation and a way of inferring 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 (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 usually difficult to recognize from their physical characteristics alone. However, it is often the case that, as these surfaces form, they are colonized by organisms that leave characteristic traces, both on the surfaces themselves and in the sediments immediately below the surfaces. 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 (§9.4.6). An interpretation of the type of discontinuity can 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 is often considered to be coincident with periods of high relative sea level, when sedimentation is confined to the flooded shelf regions. In deepwater settings, the pre- and post-omission surface suites may be characterized by the *Zoophycos* ichnofacies, whereas the omission surface itself may be characterized by the *Glossifungites* ichnofacies. The period of time encapsulated by a discontinuity surface may control the intensity of any burrowing. An 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 a vital role. Where lateral 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 floodplain 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.

As a second example, we can consider the lateral variability in sediments in the upper parts of a widespread upwards-coarsening deltaic succession (Fig. 10.7). 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. Therefore the delta was probably arcuate in form and was characterized by a laterally continuous prograding wave-dominated shoreface. 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 birdsfoot delta of the Mississippi seems more likely. Low basin energy leads to sands being deposited as local mouth bars with little reworking, whereas muddy shorelines prograde in interdistributary areas. Distributary channels cut across these elements to give the complex mosaic observed.

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 palaeogeographical 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 stratigraphical interval but separated laterally from each other by 100–200 m. 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 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.

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

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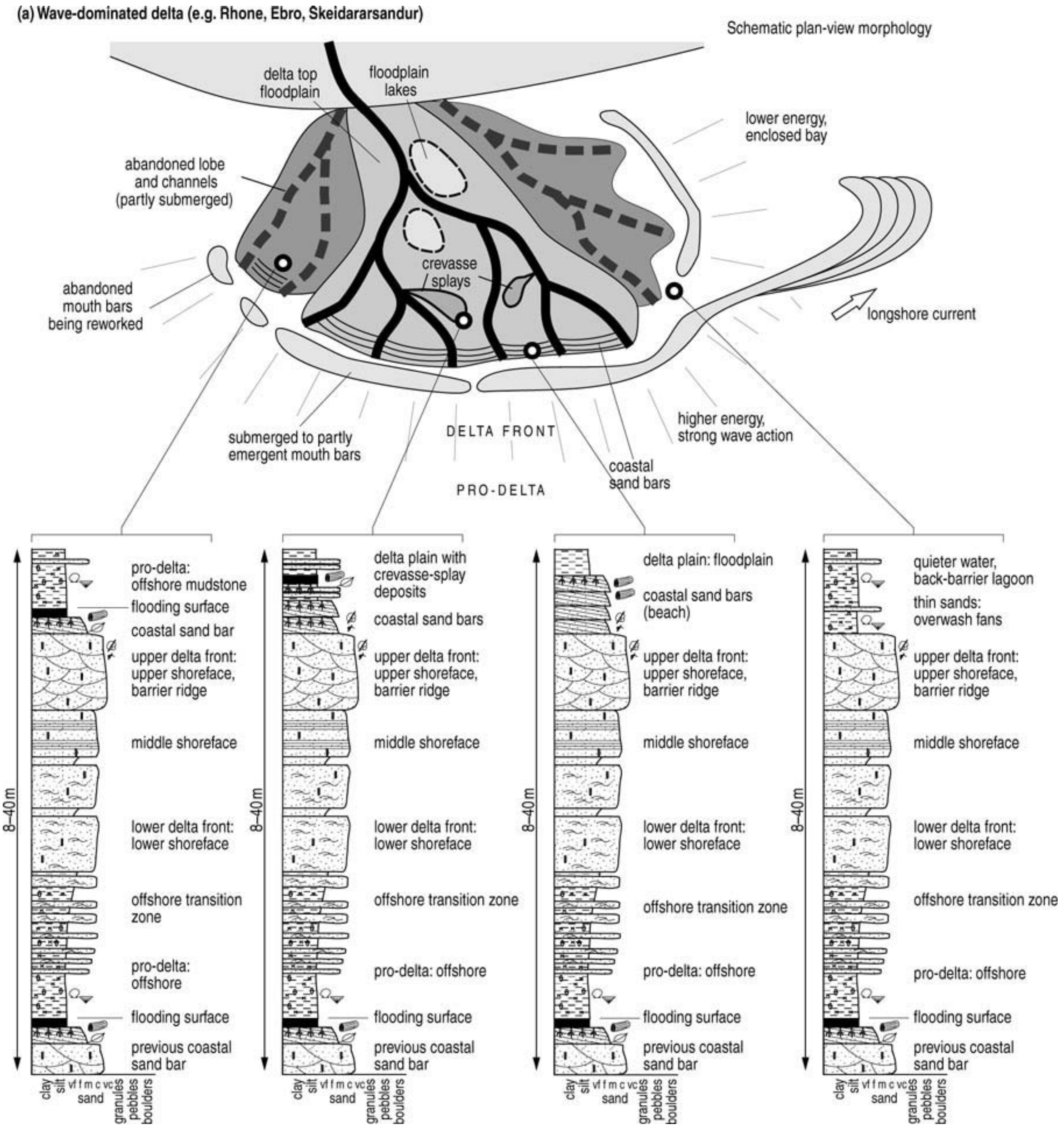
(a) Wave-dominated delta (e.g. Rhone, Ebro, Skeidararsandur)
Schematic plan-view morphology
delta top floodplain
floodplain lakes
lower energy, enclosed bay
abandoned lobe and channels (partly submerged)
crevasse splays
abandoned mouth bars being reworked
longshore current
submerged to partly emergent mouth bars
DELTA FRONT
higher energy, strong wave action
coastal sand bars
PRO-DELTA
8–40 m
pro-delta: offshore mudstone
flooding surface
coastal sand bar
upper delta front: upper shoreface, barrier ridge
middle shoreface
lower delta front: lower shoreface
offshore transition zone
pro-delta: offshore
flooding surface
previous coastal sand bar
delta plain with crevasse-splay deposits
coastal sand bars
upper delta front: upper shoreface, barrier ridge
middle shoreface
lower delta front: lower shoreface
offshore transition zone
pro-delta: offshore
flooding surface
previous coastal sand bar
delta plain: floodplain
coastal sand bars (beach)
upper delta front: upper shoreface, barrier ridge
middle shoreface
lower delta front: lower shoreface
offshore transition zone
pro-delta: offshore
flooding surface
previous coastal sand bar
quieter water, back-barrier lagoon
thin sands: overwash fans
upper delta front: upper shoreface, barrier ridge
middle shoreface
lower delta front: lower shoreface
offshore transition zone
pro-delta: offshore
flooding surface
previous coastal sand bar
clay silt vf f m c vc sand granules pebbles boulders

Figure 10.7 Example illustrating the importance of appreciating lateral variability in sedimentary successions in order to distinguish specific types of sedimentary environments. See Appendix 6 for a guide to the symbols used.

example of converting a geomorphological description into a facies model.

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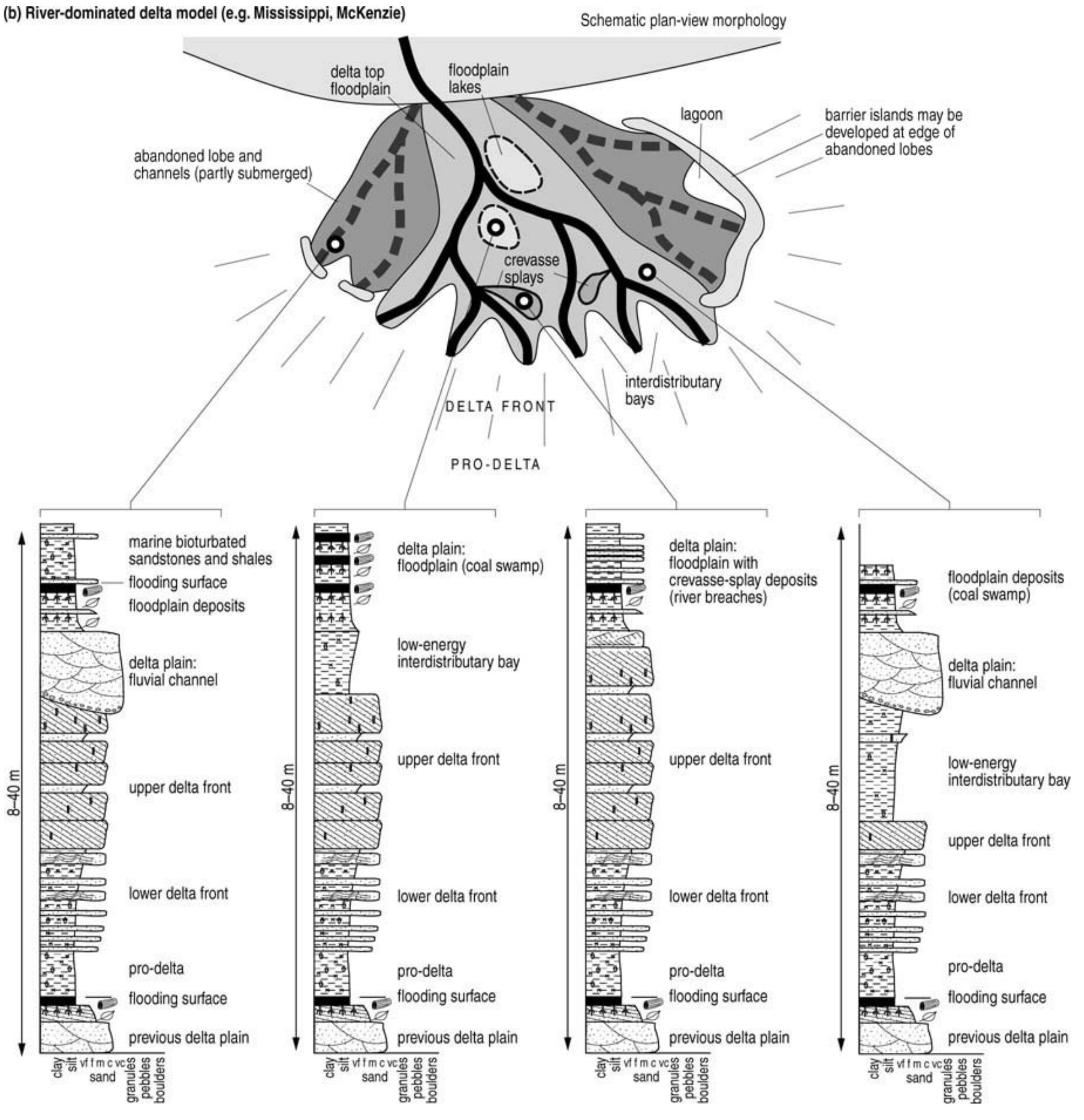
(b) River-dominated delta model (e.g. Mississippi, McKenzie)
Schematic plan-view morphology
delta top floodplain
floodplain lakes
lagoon
barrier islands may be developed at edge of abandoned lobes
abandoned lobe and channels (partly submerged)
crevasse splays
interdistributary bays
DELTA FRONT
PRO-DELTA
8–40 m
marine bioturbated sandstones and shales
flooding surface
floodplain deposits
delta plain: fluvial channel
upper delta front
lower delta front
pro-delta
flooding surface
previous delta plain
delta plain: floodplain (coal swamp)
low-energy interdistributary bay
upper delta front
lower delta front
pro-delta
flooding surface
previous delta plain
delta plain: floodplain with crevasse-splay deposits (river breaches)
upper delta front
lower delta front
pro-delta
flooding surface
previous delta plain
floodplain deposits (coal swamp)
delta plain: fluvial channel
low-energy interdistributary bay
upper delta front
lower delta front
pro-delta
flooding surface
previous delta plain
clay
silt
vf f m c vc
sand
granules
pebbles
boulders

Figure 10.7 Continued.

in the Lower Westphalian of north Devon, England. A true 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). Probably the most up-to-date and comprehensive 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.

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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 analyzed, 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 reorientate 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 small enough 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 applies only 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 normal to the bedding lies on a great circle of the projection (Fig. A1.1a,b). To restore the beds to horizontal, this point must be moved to the centre of the projection. As that shift is carried out, the normal 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 normal 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.

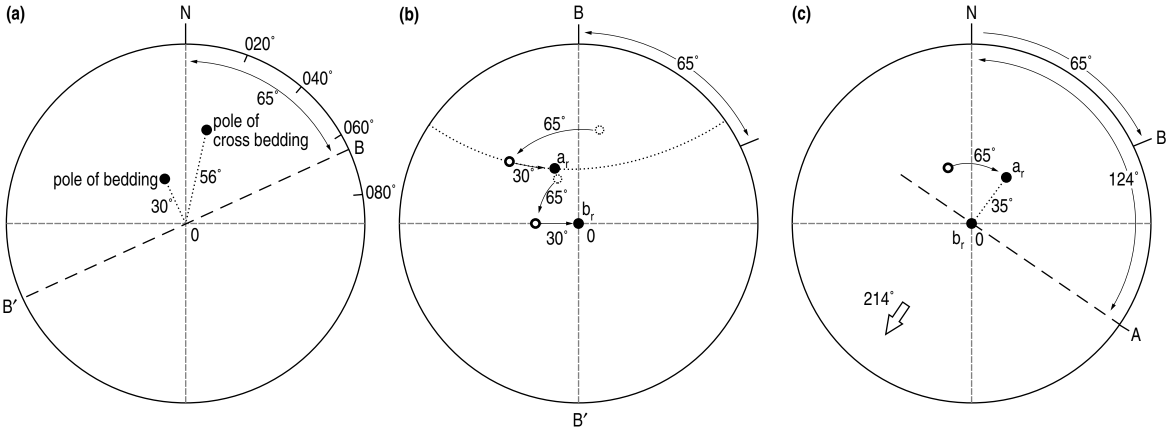
For linear data in steeply dipping beds, plot on the stereogram the attitude of the lineation in space (Fig. A1.2a) and rotate both the normal to bedding and the lineation as described above (Fig. A1.2b). This restores both the bedding and the lineation upon the bedding back to their original attitude 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 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 visualizing 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 they 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, whereas 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

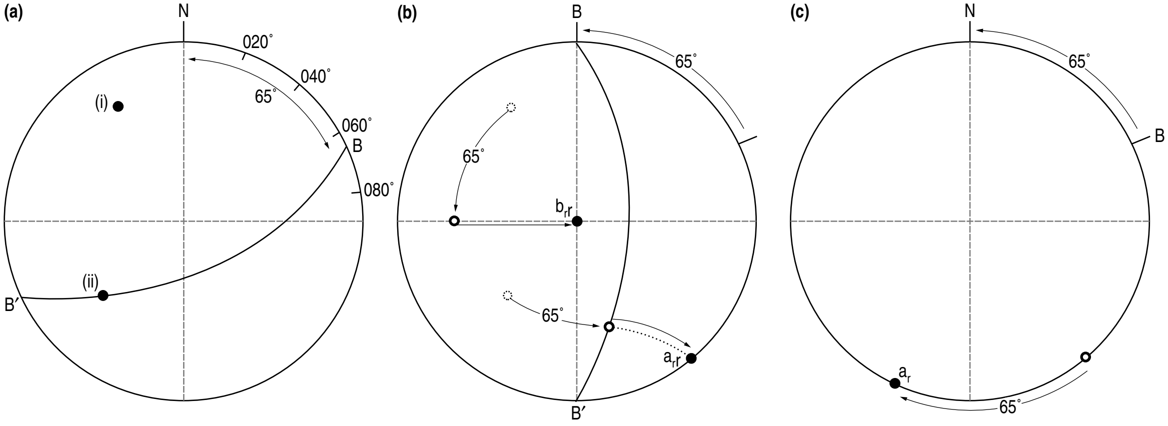


Observed data
Bedding strike = 065°, dip = 30° SE
X-bedding strike = 102°, dip = 56° SW
B–B′ = strike of bedding

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 steronet 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 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 a$_r$ 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 is plotted both as a plane and as a pole to that plane on a lower-hemisphere stereogram. Linear data is plotted as a plunge and a plunge direction (azimuth). See text for explanation.

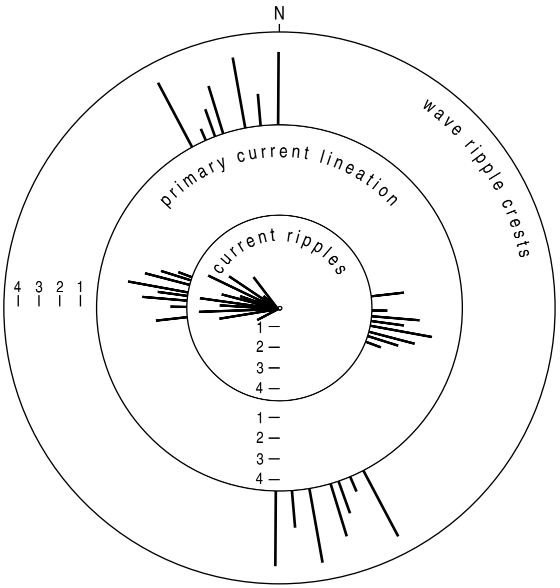
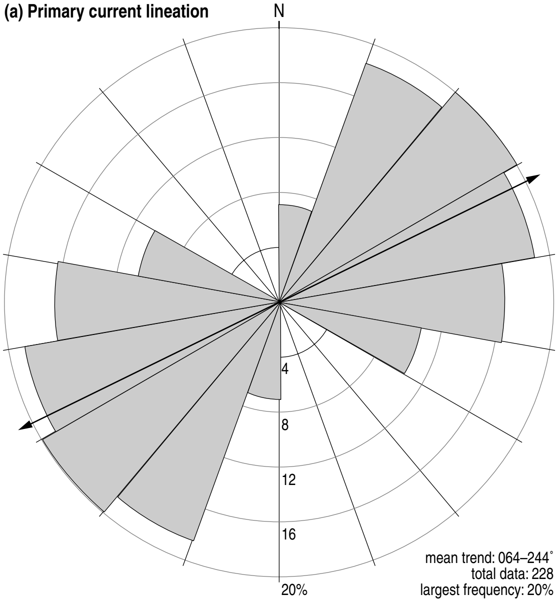


Figure A1.3 Current directions presented as a radial spoke diagram. Note how different types of structure are separated and how those structures that give only trend and not sense of movement are plotted as double-ended lines.

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 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 (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 the detailed structure of the population more clearly. 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), whereas those

(a) Primary current lineation



(b) Current ripples

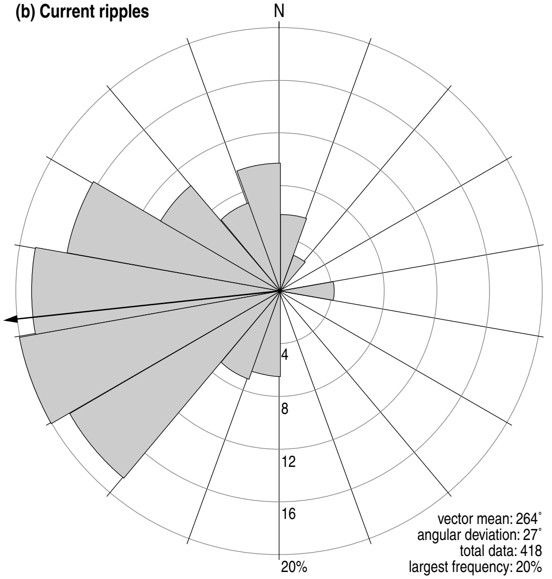


Figure A1.4 Current directions presented as rose diagrams with 20° class intervals. Data from different classes of structure are usually presented on separate diagrams, and structures that give only trend and not sense of movement are plotted as double-ended sectors. The rose may be scaled in terms of either a percentage of all observations (as here) or actual number of observations. Arrows indicate the vector mean directions.

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 diagram 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 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:

$$\tan\bar{\phi} = \frac{\Sigma\sin\phi}{\Sigma\cos\phi} \qquad \text{(A1.1)}$$

where $\bar{\phi}$ is the vector mean and $\Sigma\sin\phi$ and $\Sigma\cos\phi$ 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 also 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 is given by the equation:

$$S = \frac{\sqrt{(\Sigma\sin\phi)^2 + (\Sigma\cos\phi)^2}}{n} \qquad \text{(A1.2)}$$

where n is the number of readings. High values indicate narrow dispersion and low values a wider dispersion.

The most 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 coincides 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 topographical slope, and more than one modal wind direction may be recorded from cross bedding. 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 record 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 intertidal 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 level or peak flows and are quite narrowly dispersed. Associated smaller structures (typically ripples), formed during falling level 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, reorientate 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 reorientate 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 requires special techniques. These allow 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: 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 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-shape 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 glueing 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 that use volatile organic solvents such as acetone were often used formerly and, in such cases, the surface could be ignited 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, to remove the layer it must be strengthened by reinforcement. This is done by carefully plastering several layers of resin-soaked bandage or gauze onto 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. However, this preparation makes rather greater 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 to 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. When working on intertidal 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 flows 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 intertidal 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, establishing position along a traverse is no longer difficult in featureless terrain. However, if one has a topographical 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 will be useful for establishing position. In the absence of such a device, it will be necessary to take bearings or other angular measurements on surrounding fixed points. A sextant is an accurate and efficient tool for 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, so photographs should be taken at an early stage. 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-bottom 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 and so on 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 forms may be displayed on maps.

Examination of internal structures of modern sediments can be achieved by digging trenches or by taking shallow 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.

Figure A3.1 (overleaf) Example of a sample data sheet for the systematic collection of observations on a tidal flat.

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)	mean	mode(s)	mean	mode(s)
(2) Mineral composition (types by %)				
(3) Colour (Munsell or Goddard scale)				
(4) Fabric - texture		orientation		orientation
(5) pH				
(6) Eh - depth to reducing layer				
(7) Salinity				
(8) Temperature - air : substrate		water sample		water sample

Primary structures - bedforms		1	2	3	4	5	photo/ sketch no.	1	2	3	4	5	photo/ sketch no.
(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													

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				

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 have 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 or underwater cameras (or both) are needed. Estuaries provide accessible locations for a variety of case studies, but bear in mind the safety issues highlighted in Appendix 3. Exercises based in such settings can also develop skills such as plane tabling, aligning transects, siting quadrat surveys, sampling sub-environments for sediments as well as for organisms and the records of their activity, photographing evidence to scale, orientating data, drawing scaled diagrams and collecting and curating samples. Other useful techniques 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 under water (Appendix 2).

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). Delicate scratches and fine detail may be whitened with powdered chalk or ammonium chloride, and photographed in strong oblique light.

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 at least traces of structures where none appears to exist. Some apparently massive beds have intense bioturbation (i.e. maximum rather than minimum organic activity); but this, supplemented by diagenetic effects, enhances their apparent homogeneity.

- The making of peels from box cores.
- Staining of fine-grain carbonate and rocks rich in clay minerals 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 requires only a special film and filter, although the cutting of thinner slabs (0.5 cm), which give 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 sandblasting 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 × 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, often 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 organisms moving on the sediment surface, or the burrowing of given organisms placed upon a carefully prepared succession of particular composition and consistency, to ones that try to relate the functional morphology of the animal to its behaviour and to its burrow. More complex studies can try to match natural conditions more closely and describe 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

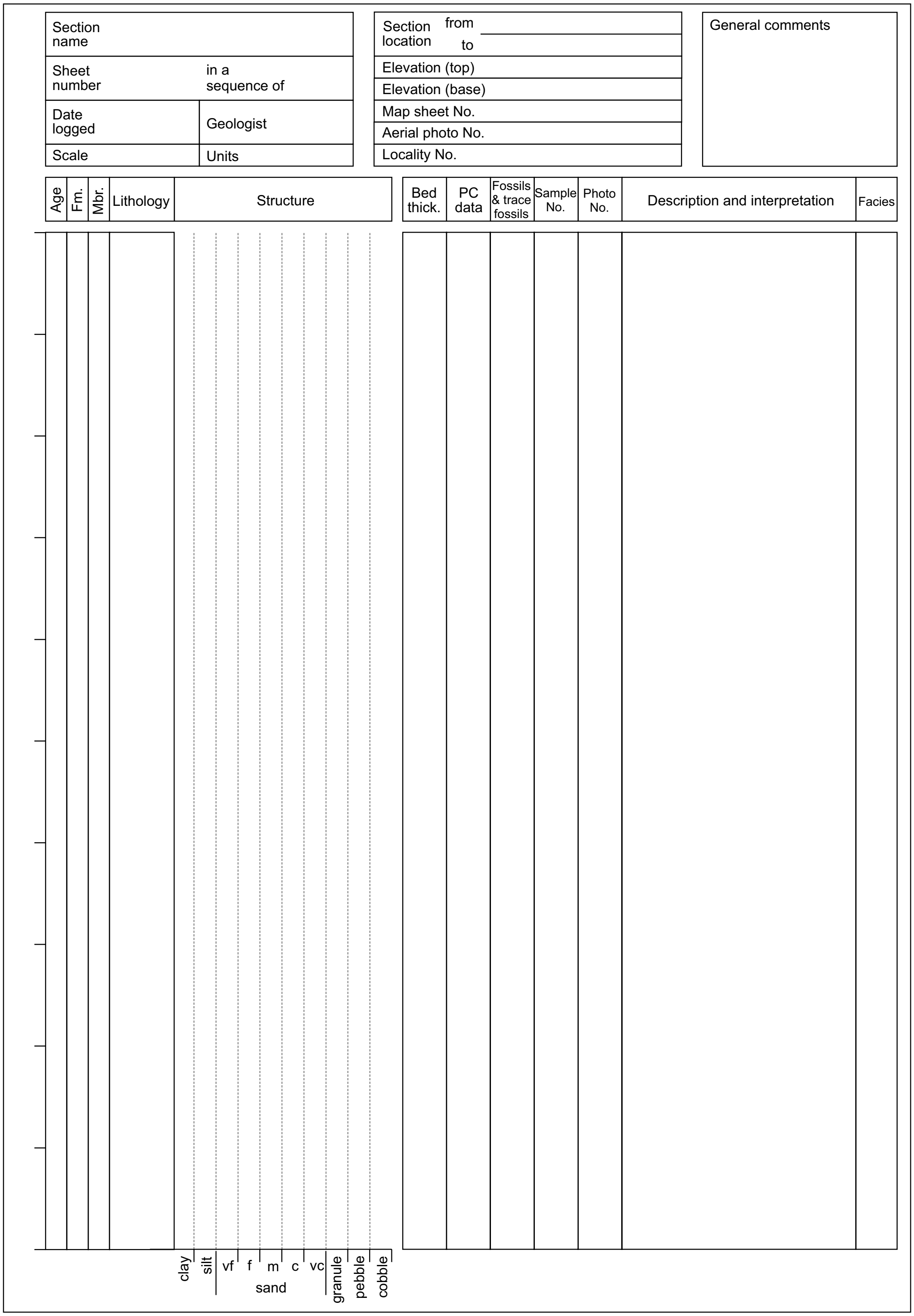
Sedimentary logs that give a bed-by-bed graphical depiction of the various lithologies and structures encountered within a succession of rocks are one of the primary methods that sedimentologists use to depict sedimentary data. Although there are many differing styles of sedimentary log, each with their 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 that are exposed.
2. Identify which part of the outcrop will be logged. Suitable sections need to be both well enough exposed to be able to generate a reasonably continuous log and also sufficiently accessible. Ideally, try to pick a location where a single continuous log can be made through the study section. However, bear in mind that this is not always possible and be prepared to construct several overlapping logs that are laterally offset from each other, in order to construct a complete run through the stratigraphical study section. Good logging sites include gulleys and ravines, dry stream beds, stepped hillsides, coastal cliffs and wavecut platforms. In regions where the beds have been tectonically tilted, good log sections can be constructed by traversing laterally along the base of cliff lines, and so on.
3. Decide on a scale for the logging exercise. This will be dictated by factors such as the complexity of the stratigraphy, the scale or thickness of the bedding, the outcrop quality, 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 characterize the study section adequately. One log may suffice for simple successions with little lateral variability, whereas more detailed studies of laterally complex and variable successions will need many logs.
5. In starting the logging exercise, try and choose a prominent bed as a start point and accurately record its geographical position and, where possible, its elevation above sea level.

The logging exercise

1. The thickness that you record on your log section for each bed should be the true bed thickness, which is not necessarily the same as the exposed bed thickness, especially when logging on a hillslope 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 5–10 cm, whereas for broader-scale logs it may be sufficient to group sets of similar beds together and record them as a single coset. If appropriate, you can 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-shape 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, although a note should be made if the bed clearly changes thickness when viewed along strike.
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 for every bed. 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.
5. For each bed look carefully for sedimentary structures, both in section and on exposed bedding surfaces, bearing in mind that they 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 a dip and dip 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 bedding, so that the palaeocurrent data can be restored at a later date (Appendix 1).
7. Make separate notes alongside the log, describing potentially significant features. In many instances you may also be able to infer something about the nature of the depositional process. Indeed, as the log is being constructed, you may even develop hypotheses about possible environments



Section name

Sheet number in a sequence of

Date logged

Geologist

Scale

Units

Section location from

to

Elevation (top)

Elevation (base)

Map sheet No.

Aerial photo No.

Locality No.

General comments

Age	Fm.	Mbr.	Lithology	Structure	Bed thick.	PC data	Fossils & trace fossils	Sample No.	Photo No.	Description and interpretation	Facies

clay
silt
vf
f
m
c
vc
sand
granule
pebble
cobble

Figure A5.1 Example of a logging sheet for the recording of vertical sequences in sedimentary rock successions.

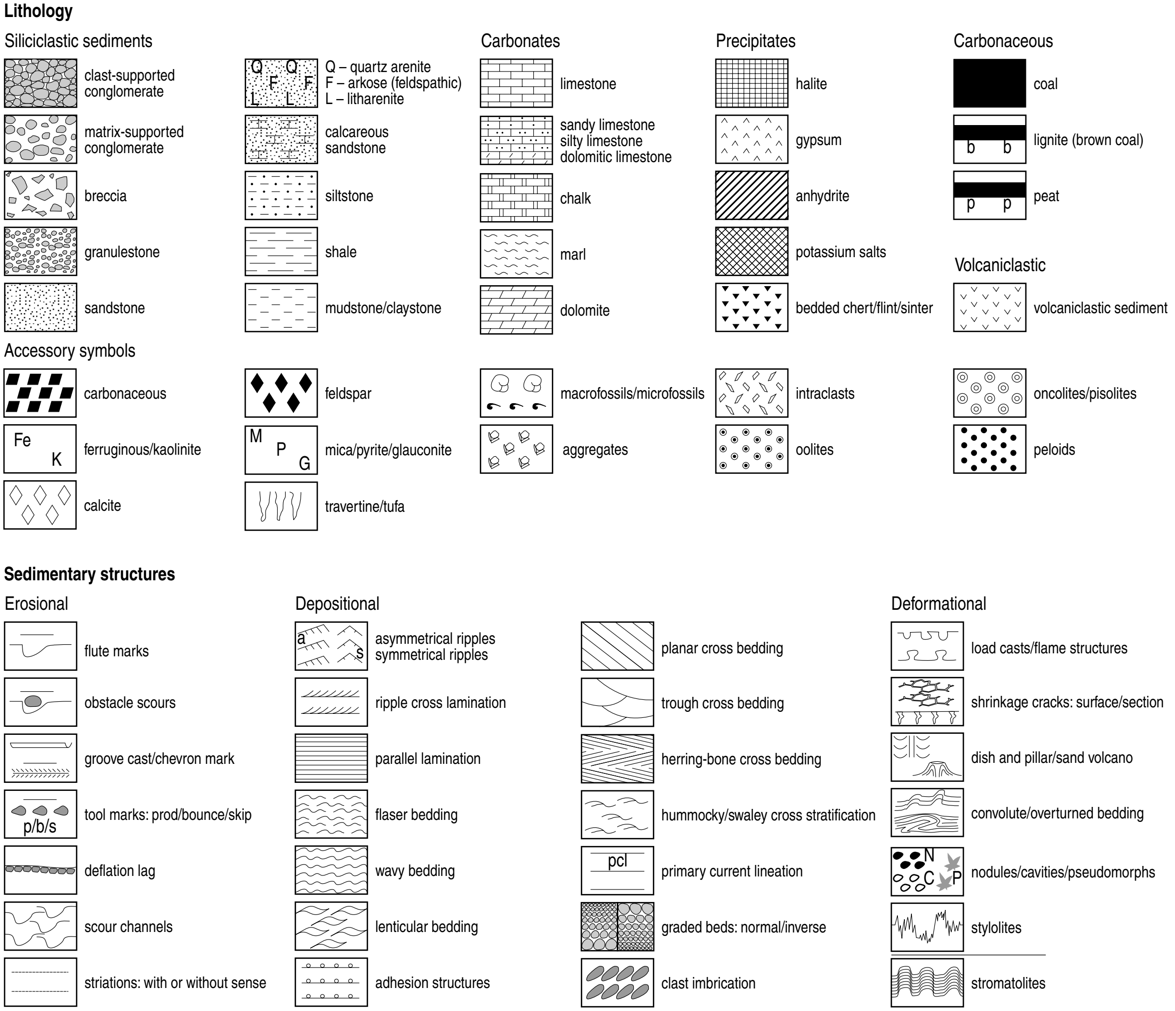
of deposition. However, although working ideas are useful, it is important not to become too focused on any one interpretation during the logging phase, as this may bias the way in which you make your observations.

8. In finishing the logging exercise, try to choose a prominent bed as an end point and accurately record its geographical 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 types of lithologies and 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 virtually standardized, and an example set of commonly used symbols is depicted in Figure A6.1. These symbols can be adapted to suit the suite of sediments or rocks being investigated; similarly, 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, where necessary, with written descriptive notes and preliminary interpretations of process or environment of deposition.



Lithology
Siliciclastic sediments
clast-supported conglomerate
matrix-supported conglomerate
breccia
granulestone
sandstone
Q F Q F L L
Q – quartz arenite
F – arkose (feldspathic)
L – litharenite
calcareous sandstone
siltstone
shale
mudstone/claystone
Carbonates
limestone
sandy limestone
silty limestone
dolomitic limestone
chalk
marl
dolomite
Precipitates
halite
gypsum
anhydrite
potassium salts
bedded chert/flint/sinter
Carbonaceous
coal
b b
lignite (brown coal)
p p
peat
Volcaniclastic
volcaniclastic sediment
Accessory symbols
carbonaceous
Fe K
ferruginous/kaolinite
calcite
feldspar
M P G
mica/pyrite/glauconite
travertine/tufa
macrofossils/microfossils
aggregates
intraclasts
oolites
oncolites/pisolites
peloids
Sedimentary structures
Erosional
flute marks
obstacle scours
groove cast/chevron mark
p/b/s
tool marks: prod/bounce/skip
deflation lag
scour channels
striations: with or without sense
Depositional
a s
asymmetrical ripples
symmetrical ripples
ripple cross lamination
parallel lamination
flaser bedding
wavy bedding
lenticular bedding
adhesion structures
planar cross bedding
trough cross bedding
herring-bone cross bedding
hummocky/swaley cross stratification
pcl
primary current lineation
graded beds: normal/inverse
clast imbrication
Deformational
load casts/flame structures
shrinkage cracks: surface/section
dish and pillar/sand volcano
convolute/overturned bedding
N C P
nodules/cavities/pseudomorphs
stylolites
stromatolites

Figure A6.1 Scheme for the graphic depiction of lithologies, sedimentary structures, fossils and trace fossils in sedimentary logs and panels.

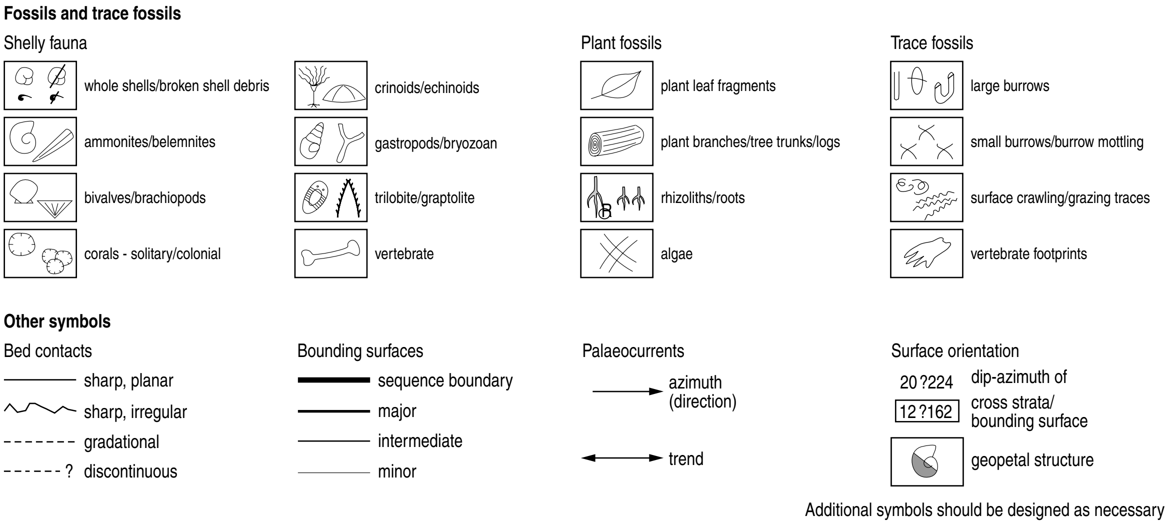

Fossils and trace fossils
Shelly fauna
whole shells/broken shell debris
ammonites/belemnites
bivalves/brachiopods
corals - solitary/colonial
crinoids/echinoids
gastropods/bryozoan
trilobite/graptolite
vertebrate
Plant fossils
plant leaf fragments
plant branches/tree trunks/logs
R
rhizoliths/roots
algae
Trace fossils
large burrows
small burrows/burrow mottling
surface crawling/grazing traces
vertebrate footprints
Other symbols
Bed contacts
sharp, planar
sharp, irregular
gradational
?
discontinuous
Bounding surfaces
sequence boundary
major
intermediate
minor
Palaeocurrents
azimuth (direction)
trend
Surface orientation
20?224
12?162
dip-azimuth of cross strata/ bounding surface
geopetal structure
Additional symbols should be designed as necessary

Figure A6.1 Continued.

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