CHAPTER 3 Alluvial Sediments

J. D. COLLINSON

3.1 INTRODUCTION

Although rivers have long been recognized as major transporters of sediment, the appreciation that they contribute directly to the rock record is a comparatively recent development. Up to the 1960s sediments were interpreted as fluvial, or more likely fluvio-deltaic, only if channel forms were recognized, as in coal measure 'wash-outs', whilst many other sequences lacking channel forms, which we now interpret as fluvial, were regarded as lake deposits.

Geomorphological work on channel types and processes in the 1950s (e.g. Sundborg, 1956; Leopold and Wolman, 1957) was not immediately applied to ancient deposits. However, in pioneering studies of the ancient, Bersier (1959), Bernard and Major (1963) and Allen (1964) recognized the occurrence of fining-upwards sandstone units often of wide lateral extent and equated these with the migration of river channels. This early emphasis on the importance of lateral accretion, particularly by migrating point bars, meant that other channel processes were largely ignored. Fining-upwards sequences overlying sharp bases were equated almost uncritically with point bar migration and, with a few exceptions (e.g. Ore, 1963), little attempt was made to recognize the deposits of other types of channel.

With an increasing appreciation of the hydrodynamic significance of bedforms, particularly those formed in sand, and with increasing study by geologists of low sinuosity streams, the complexities of channel deposition have become more apparent. Meandering streams are now seen to be capable of producing a whole range of different sequences some of which grade into and overlap with the deposits of low sinuosity streams. In addition, anastomosing river systems characterized by non-migrating channels, are now being studied after a history of almost total neglect. Soil-forming processes which operate in interchannel or floodplain areas have been seen to offer not only an insight into the prevailing climate but also a means of correlation and of understanding the longer term development of river systems (e.g. Allen and Williams, 1982).

The wider and more balanced view of alluvial sediments which is now taken has led to a greater understanding of fluvial systems. This increasing appreciation of their complexity makes the application of simple models less and less satisfactory and leads to uncertainty and ambiguity in interpreting the ancient. Ancient sequences often provide greater insights than present-day systems into the long-term changes and controls on alluvial systems. Studies of the large-scale organization of successions have thrown light on processes of channel behaviour and have helped to build a basis for the prediction of sand body geometry.

In this chapter we deal first with the processes and products of present-day alluvial systems which are divided for convenience into four groups, namely *bedload streams*, *alluvial fans*, *meandering streams* and *anastomosing streams*. Interchannel processes and products which are similar in many systems are treated separately. We then examine the array of facies present in ancient alluvial sequences and explore the ways in which they are organized. This gives a basis for interpreting the rock record in terms of present-day systems and also points to ways in which some ancient systems must have differed from present-day examples. It also gives a basis for predicting the organization of ancient sequences, an aspect of interest to many applied and economic geologists.

3.2 PRESENT-DAY BEDLOAD STREAMS

Bedload streams are those in which the coarser grain sizes, gravels and sands transported mainly as bedload, dominate the deposits and in which fine-grained sediments transported in suspension do not contribute greatly to sediment accumulation even though such material may be abundant in the overall load. Such rivers contrast with mainly meandering suspended load streams (Sects 3.4, 3.5) where fine-grained sediment forms a significant proportion of the total deposit, occurring as overbank sediment and as the plugs of abandoned channels. Bedload streams commonly have channels characterized by low sinuosity and by considerable lateral mobility. The mobile channels are commonly subdivided internally into rapidly changing patterns of sub-channels and 'bars' giving a braided pattern which is most apparent at low stage. Some examples are more sinuous and grade into meandering types. Rivers of this type therefore form a continuum of grain size and sinuosity variation.

Within this major group it is possible to make a subdivision into pebbly and sandy types each with rather different bedforms and processes. In many natural systems pebbly streams occur in upstream reaches and grade downstream into sandy types, either within the confines of a valley or in less restricted settings of alluvial fans and outwash plains (Sect. 3.34).

3.2.1 Pebbly bedforms and processes

Most of our knowledge of pebbly braided streams comes from studies of proglacial outwash areas where the rather predictable pattern of seasonal discharge facilitates study (Fahnestock, 1963; Church, 1972). Similar streams in warm, semi-arid settings are susceptible to less predictable flash floods. Within the seasonal pattern of discharge seen in proglacial settings shorter term changes are superimposed, related to the prevailing weather. Temperature fluctuations on diurnal and longer time-scales lead to fluctuations of discharge when melting snow or ice is the primary source of water. In terms of sediment response, small-scale morphology responds fairly rapidly to changing discharge whilst the pattern of larger channels and bars may only change with major floods.

There have been many descriptions of glacial outwash areas mainly from the 'Sandur' (sing. 'sandar') of Iceland (Hjulström, 1952; Krigström, 1962; Bluck, 1974; Klimek, 1972) and from the sub-Arctic and mountain areas of North America (e.g. Williams and Rust, 1969; Boothroyd, 1972; Rust, 1972a; Boothroyd and Ashley, 1975; Boothroyd and Nummedal, 1978; Church, 1983; N.D. Smith, 1974). Such outwash fans and plains show complex patterns of channels and bars. The main flow at any one time is concentrated in a fairly well-defined zone or zones whilst the rest of the area is made up of abandoned channels and bars rather than a separate and discrete floodplain (Fig. 3.1).

The braided pattern results from a complex interaction of sediment supply and water discharge. High rates of supply of bedload sediment lead to overloading of streams and a steepening of slopes until a balance of transport and supply is roughly achieved. The bedforms and the pattern of channels contribute to the frictional resistance to flow in this complex balance. Along rivers of this sort, particularly ones confined within valleys, channels show alternations of stable and more mobile zones (Church, 1983) related to transport and storage of bedload sediment respectively. Within mobile zones, braiding is common and channel shifting fairly vigorous. In the less confined context of fans, the pattern of channels is usually entirely braided and unstable.

The braided pattern results from the development of bars which split the flow, commonly at several co-existing scales (cf. Rachocki, 1981). These bars have a variety of forms most of which have complex histories of erosion and deposition and which may evolve from one form into another. Leopold and Wolman (1957) showed experimentally that braiding may result from bar growth at steady discharge but in nature it is likely that much of the pattern results from emergence due to discharge fluctuations. Bars in braided rivers may be classified in various ways (e.g. Krigström, 1962; Church, 1972; N.D. Smith, 1974;

Fig. 3.1. Braided pattern of gravel bars and channels, Alakratiak Fjord Valley, Washington Land, Greenland. The main flow is concentrated in a zone flanked by areas of abandoned channels and bars. Bars and channels are superimposed at several scales.

Ferguson and Werritty, 1983). Whilst we discuss them here under three headings it should be borne in mind that these classes are arbitrary and intergradational and are able to evolve from one into the other through time.

LONGITUDINAL BARS

These are diamond or lozenge shaped and are the most obvious form in many pebbly braided streams (Fig. 3.1). They form initially by the segregation of coarse clasts as thin gravel sheets with a rhomboidal plan shape. Such sheet bars are common in the upstream parts of some outwash fans and probably grow into higher relief longitudinal bars by a combination of vertical gravel accretion, the development of a downstream slipface and by erosion and incision by lateral channels (Boothroyd, 1972). The upper surfaces of bars commonly have low relief 'transverse ribs' associated with the imbricate gravel (McDonald and Banerjee, 1971; Boothroyd and Ashley, 1975). Long axes of clasts are aligned transverse to the current. The upper surfaces of longitudinal bars also commonly show a gradual reduction in clast size from upstream to downstream. The clast size of the bar-top at its upstream end is the same as that of the flanking channel into which it grades but at the downstream end the

Fig. 3.2. A sand wedge at the lateral margin of a gravel bar has been dissected by water in minor channels draining off the bar during a fall in stage. The water in the adjacent channel is now stagnant and a thin film of mud drapes the sand ripples. Donjek River, Yukon, Canada (after Rust, 1972a).

bar-top sediments are markedly finer than the sediments of the adjacent channel floor. As bar-tops emerge, usually during falling stage, the gravel may be partially mantled with sand which occurs in rippled patches often elongated parallel to the flow as 'sand shadows' in the lee of larger clasts.

Downstream depositional margins are either avalanche slip faces or *riffles*, depending on the grain size and the water depth over the bar-front. Riffles are steeper sectors of channel floor where flow is rapid and more turbulent. Coarser grain sizes and shallow flows favour the development of riffles. They also develop at the confluence of two channels, which commonly occurs directly downstream of the apex of a longitudinal bar. At low water stage, bar margins may be draped with wedges of sand which grow under conditions of reduced shear stress after gravel movement has stopped (Fig. 3.2) (Rust, 1972a). Other bar margins are erosional where inter-bar channels shift laterally.

Internally, longitudinal bars show massive or horizontally bedded gravel with imbricated clasts. There may be an upward fining of clast size and the gravel framework may be filled with sandy matrix. Smith (1974) suggested that alternations of matrix-filled and open-work layers record normal (diurnal?) fluctuations of discharge. Avalanche slip faces lead to the development of sets of cross-bedded gravel which may be punctuated by reactivation surfaces (Collinson, 1970) reflecting episodic bar growth. Such discontinuities may also be recorded by wedges of sand interbedded with the gravel foresets. Horizontally bedded gravels should lie upstream of and progressively overlie cross-bedded gravel if a bar grew steadily. However, many bars show a complex internal structure reflecting many episodes of erosion and deposition.

BANK-ATTACHED BARS IN CURVED CHANNELS

Bars may be attached to either bank of a curved reach (Krigström, 1962). Some are extensions or modifications to the flanks of larger longitudinal bars and have downstream margins strongly oblique to the channel trend (Fig. 3.3). Such diagonal bars may evolve from or into longitudinal bars whilst others are gradational with point bars (Ferguson and Werritty, 1983). Where the upstream end of a bar is attached to the bank, flow is concentrated on the opposite side of the channel. As the bar-crest crosses the channel the flow switches to the other bank by crossing the bar-top. The downstream end of the bar is commonly a long, continuous riffle or a series of smaller riffles separated by sectors of emergent bar top which split the flow (Fig. 3.3B). Riffles are important sites of sediment accumulation giving units of gravel elongated parallel to the bar crest but of probably limited extent in the direction of growth (Bluck, 1976). Lee faces of some of these bars may be locally avalanche slip faces rather than riffles.

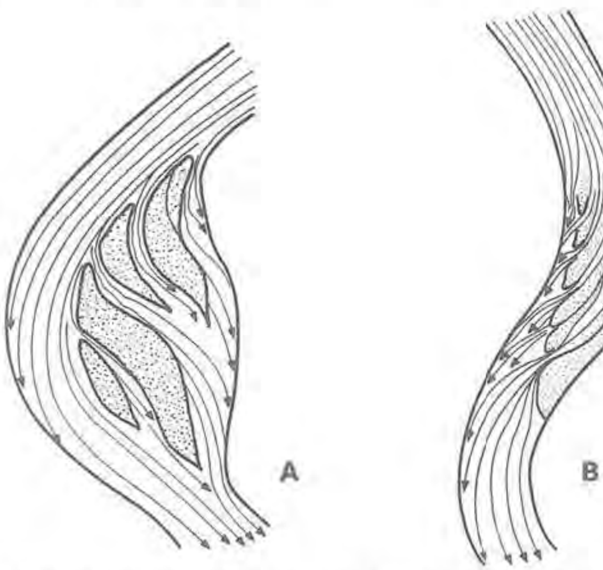

Fig. 3.3. Bank-attached bars in curved channels; A, attached to the inside bank and showing some characteristics of a point bar; B, attached to the outside bank but virtually crossing the channel as a continuous and highly skewed dissected riffle (after Krigström, 1962).

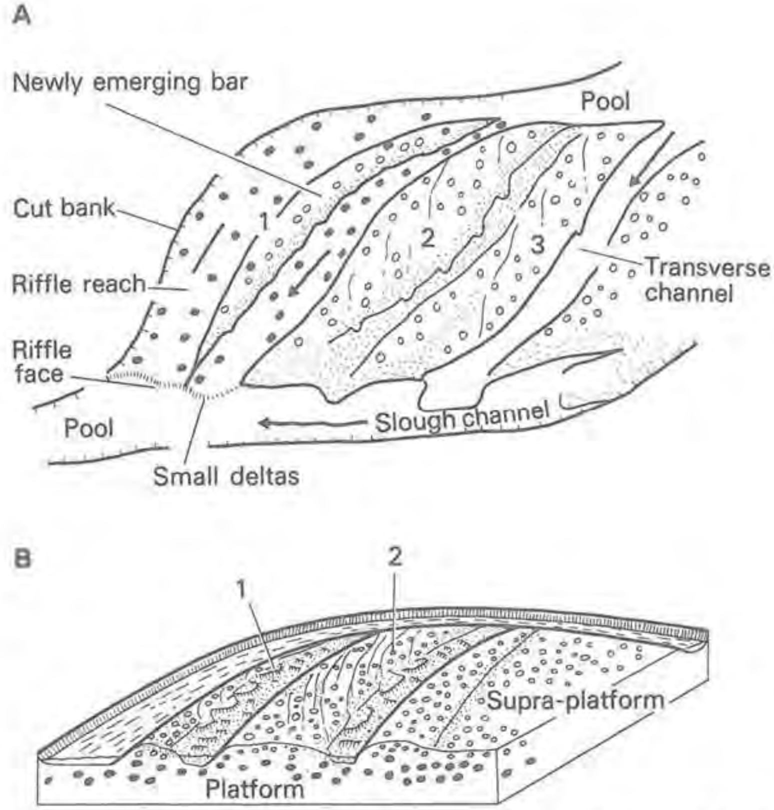

Fig. 3.4. Morphology, terminology and structure of a typical bank-attached bar, (1), (2) and (3) are individual units of the bar. (1) is the most recently emergent bar which will eventually accrete on to (2) (after Bluck, 1974).

Where the main part of the flow stays close to the outside of a bend and the gravel bar is confined to the convex bank, the situation is much more akin to a point bar in a meandering stream (Figs 3.3A, 3.4). However, in this case there is also flow over the bar-top, commonly in shallow sub-channels at high water stage. These flows feed into a channel close to the convex bank which carries a minor part of the overall flow often only at high stages. Such a channel commonly has a riffle at its downstream junction with the main channel. It is also likely to carry more gently flowing water during falling stage and to behave as a *slough channel*, that is, a site of quiet water where fine sediment is deposited from suspension. The residual flows over the bar-top may build small sand and gravel lobes into the channel whilst most of the channel floor accumulates a coating of fine-grained sediment on top of the high stage gravels.

Riffles are the main sites of accumulation in these cases and build up sheets or elongate bodies of gravel whose thickness reflects the relief developed across the riffle. Internally they are probably rather structureless, although they may show some overall upward grading in clast size and also pebble imbrication. In addition, they may show some low-angle inclined bedding dipping downstream or, where an avalanche slip face had developed, a set of normal, high angle foresets.

The generally mobile nature of gravel-bed streams and the frequent shifting of both bars and channels leads to a fragmentary preservation and it will normally not be possible to distinguish the products of longitudinal and bank-attached bars.

TRANSVERSE BARS

Transverse bars, as originally recognized by Ore (1963) and Smith (1970), are common features of sandy, low sinuosity streams but they also occur in gravel-bed streams where they may be transitional with the more continuous bars of curved reaches. In a pebbly proglacial stream, Smith (1974) recognized a subsidiary population of transverse bars with lobate and sinuous crests. Their downstream ends are commonly slip faces, a feature which they share with transverse bars of sandy streams. Transverse bars become more important downstream in a river at the expense of longitudinal bars, the change being associated with reductions of grain size and of gradient.

Transverse bars are likely to produce extensive sets of cross-bedding which may be punctuated by reactivation surfaces (see Sect. 3.2.2).

CHANNELS

Some comments have been made already about the relationship of channels to neighbouring bars. Channels are active for longer periods than bars and may erode into bars or be over-ridden by the advance of a bar. Active channels around bars are floored by the coarsest gravels which grade into the upstream ends of bar-tops. These are likely to be strongly imbricated where clasts are flattened. When channels are abandoned either temporarily during low water or more permanently by channel diversion, they may operate as slough channels. The gravel floor will then be commonly draped by finer sediment, first sand which, after filling the gravel framework, may continue to migrate as ripples or even dunes, and later by silt or mud from suspension (Williams and Rust, 1969). This may eventually dry out, forming clay clasts by cracking and curling and may also be subjected to wind deflation.

DIRECTIONAL PROPERTIES

Structures which record flow direction are formed at a variety of scales and in material of different grain sizes. The main structures with potential for preservation are channels, clast imbrication, cross-bedding, ripples, ripple cross-lamination and sand lineation. Transverse ribs in gravel may also be preserved (Rust and Gostin, 1981). Channels are the largest forms and have the lowest directional variability (Allen, 1966; Bluck, 1974). Of the smaller structures, clast imbrication, associated with the early development of bars, bar tops and the floors of channels, shows a unimodal trend of fairly low variability (Figs 3.5, 3.6) (Rust, 1972a; Bluck, 1974). Elongate pebbles, larger than 2–3 cm, tend to have long axes transverse to flow (McDonald and Banerjee, 1971; Rust 1972a and b). Cross-bedding produced at the downstream ends of bars is likely to be rather variable in orientation, probably bimodally distributed about the downstream direction as a result of the skewed trend of most bar-fronts. At lower water stage, the growth of sandy lobes on bar-fronts leads to even greater variability and other structures formed in sand, such as ripples and sand lineations,

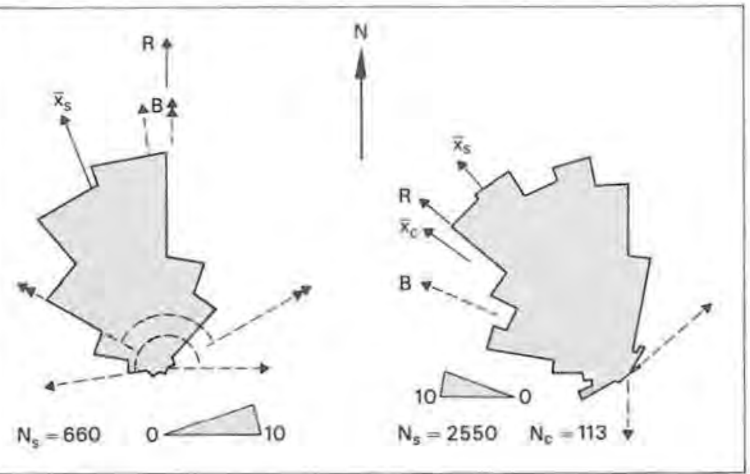

Fig. 3.5. Directional data from two areas of the braided pebbly Donjek River, Canada. The histograms are for small-scale structures. R = river trend, $\bar{x}_s$ = vector mean of small-scale structures; $\bar{x}_c$ = vector mean of channel orientations; B = channel arc bisector. Double arrows refer to channels; single arrows to small structures (after Rust, 1972a).

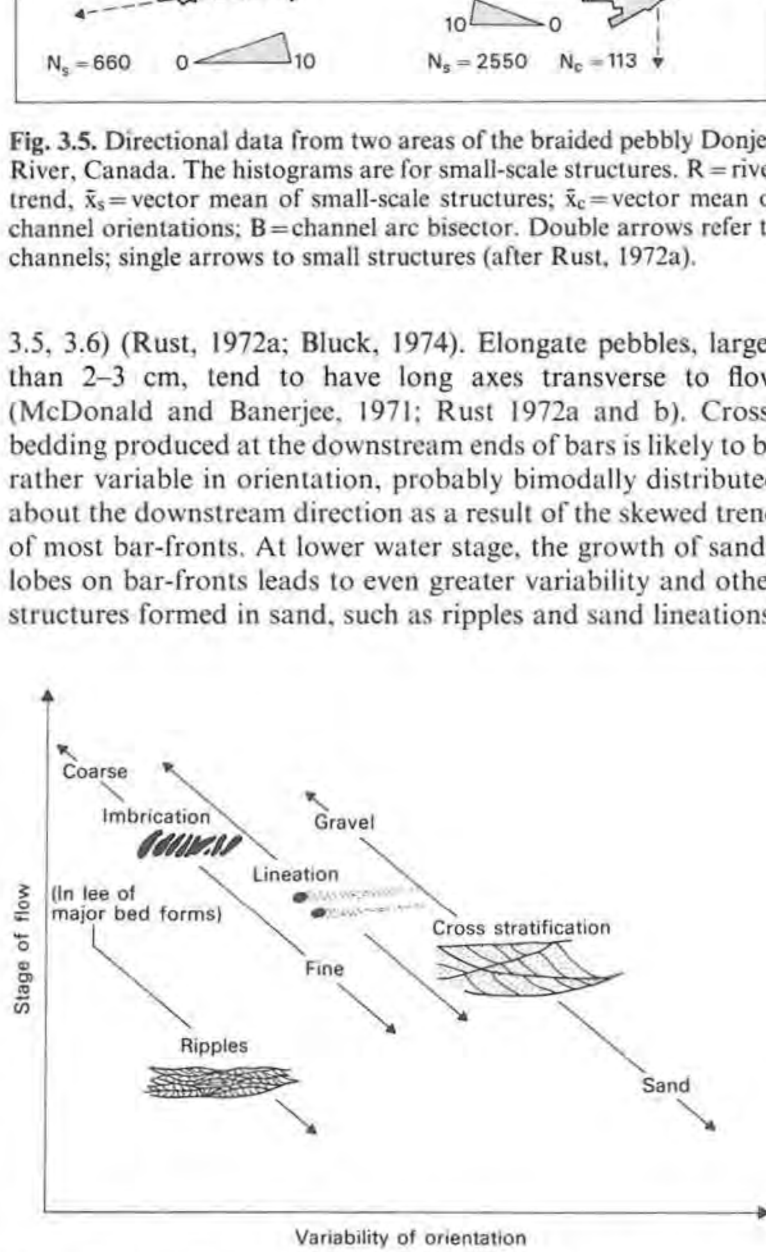

Fig. 3.6. Likely relationships between flow stage and the type and variability of orientation of active structures on a mixed sand and gravel channel floor. Note that ripples lie off the main trend. They tend to be caught in major channel troughs and their orientation is determined by them (after Bluck, 1974).

Fig. 3.7. Sandwaves (linguoid or transverse bars) in the Tana River, Finnmark, Norway, emerging at intermediate discharge during falling river stage. Together the bars make up a major composite sand flat. Contrast the smooth crestline of the submerged bar in the foreground with the more irregular plan of dissected emergent bars.

continue to move on submerged bar tops and on channel floors. These structures show high dispersion, reflecting the tortuous course of water at reduced discharge (Collinson, 1971b).

3.2.2 **Sandy bedforms and processes**

Sandy braided and low-sinuosity streams are gradational with the coarse-grained types described above and also with meandering streams. Many large, sandy low-sinuosity streams are braided owing to the development of mid-channel bars of various types. This braiding tends to become more pronounced at times of low discharge when more of the bed is exposed (Fig. 3.7). As with pebbly braided streams, rivers of this type are characterized by considerable channel mobility. A lack of the clay plugs which result from the infill of abandoned channels means that resistance to lateral erosion is low. Similarly the erodible nature of the sandy bed means that vertical scour and fill may be important over flood cycles. This is most spectacularly illustrated by the comparative echograms from the Brahmaputra presented by Coleman (1969) (Fig. 3.8). Similar changes occur in smaller streams, particularly during exceptional floods.

Most of our information on sandy low-sinuosity rivers comes from relatively small examples such as those of the mid-west USA (e.g. Loup (Brice, 1964); Platte (N.D. Smith, 1970, 1971; Blodgett and Stanley, 1980); Red (Schwarz, 1978)), and higher latitude settings (e.g. Lower Red Deer (Neill, 1969); Tana (Collinson, 1970); South Saskatchewan (Cant, 1978; Cant and

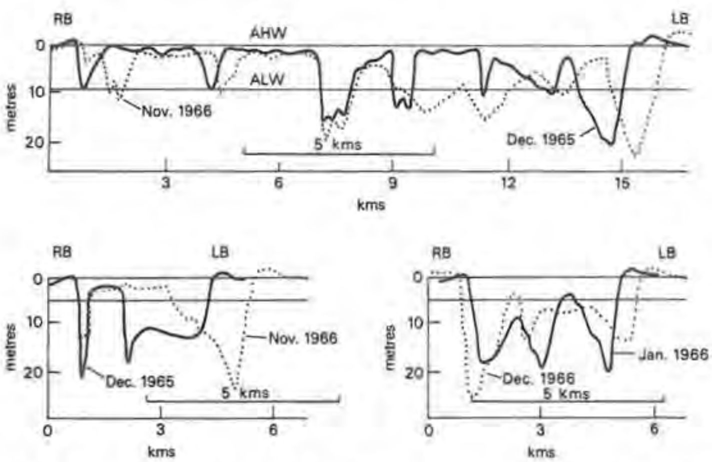

Fig. 3.8. Profiles of the bed of the Brahmaputra River transverse to the flow direction measured before and after a single monsoonal flood. Over this short interval, huge volumes of sediment were eroded and deposited and the channel pattern radically altered (after Coleman, 1969).

Walker, 1978)). However, there are also major rivers of this general type, such as the Niger-Benue (NEDECO, 1959), the Brahmaputra (Coleman, 1969), the Yellow River (Chien, 1961) and the major rivers draining the plains of South America and Siberia.

The smaller, better studied examples commonly occur in settings of net erosion and are confined by terraces or valley walls. However, rivers of this type also occur in settings with net deposition sometimes associated with major fans (e.g. Kosi River; Gole and Chitale, 1966), and it seems reasonable to apply our knowledge of the smaller examples to the rock record.

At first sight sandy rivers of low sinuosity show a bewildering range of type and scale of bedform and this has led to considerable confusion over nomenclature and the hydrodynamic status of the forms (N.D. Smith, 1978). In individual rivers it is usually possible to erect a hierarchy of forms but it is more difficult to devise a scheme of general applicability. The types described here are thought to cover most features. The hydrodynamic significance of the smaller repetitive forms is well established and is illustrated in Fig. 3.9.

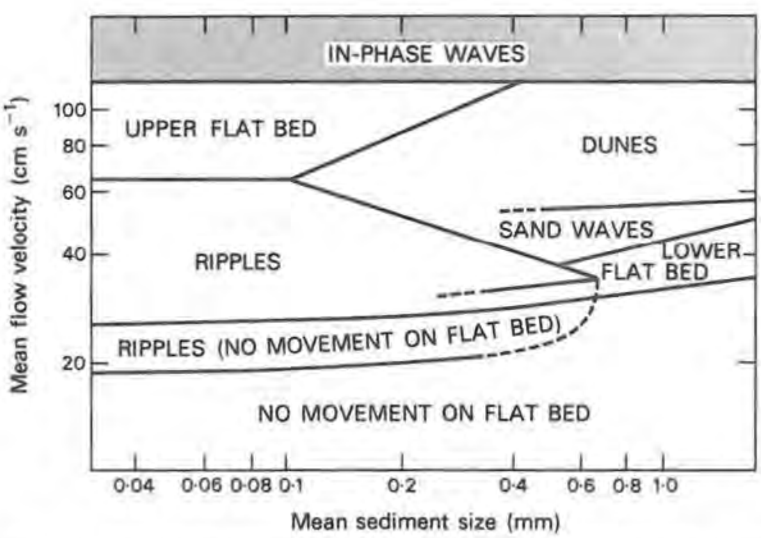

Fig. 3.9. The distribution of common sand bedforms plotted in the field of current velocity and sediment grain size. Note the pinch-outs of the ripple and dune fields at critical values of grain size (after Harms, Southard *et al.*, 1975).

RIPPLES

Ripples are almost ubiquitous on sand beds where the grain size is coarse sand or finer. Because of their small size, ripples quickly change their orientation in response to changing patterns of flow associated with discharge changes. On emergent areas they give a good idea of flow patterns just prior to emergence.

DUNES

These larger, repetitive structures, are common on river beds, particularly in deeper channels or sub-channels. They are less common on the topographically higher areas such as the backs of transverse bars and the tops of sand flats.

TRANSVERSE BARS

These are larger than dunes. They commonly have downstream slip faces and gently sloping tops, with a much lower height/length ratio than dunes. They are the sandy equivalents of the transverse bars described from pebbly braided streams and occur in a variety of intergradational forms defined on the shape and extent of their crestlines. All carry superimposed ripples on their upper surfaces and some carry dunes. They occur as isolated features and also in repetitive patterns.

(a) *Cross-channel bars* are a type of transverse bar common in relatively narrow channels. At their simplest and in the early stages of their development their crests cross the channel from bank to bank, more or less normal to flow (Fig. 3.10). However, slight irregularities in height of the crest of a cross-channel bar lead to splitting of the flow around the high points, particularly during falling stage (Cant and Walker, 1978). Downstream extensions of the crest develop as 'wings' flanking the exposed sector, giving an increasing skewness and curvature to the crest and initiating growth of a mid-channel sand flat (see below).

(b) *Linguoid bars* are transverse bars with crest lines strongly curved in a downstream convex fashion (Allen, 1968; Collinson, 1970). They form repetitive patterns, particularly in deeper channel areas and are also seen to be the main accreting components of sand flats (Fig. 3.7). In the Tana River they are mostly 200–300 m long, 200 m wide and up to 2 m high (Collinson, 1970) whilst in smaller streams they are smaller (e.g. Blodgett and Stanley, 1980). During falling stage, these relatively large forms are unable to respond fully to the reduced

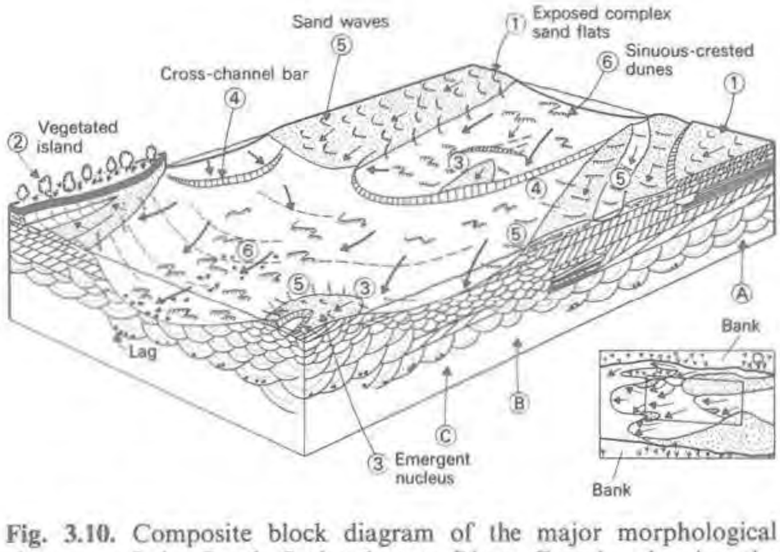

Fig. 3.10. Composite block diagram of the major morphological elements of the South Saskatchewan River, Canada, showing the predicted sequences of internal structure generated in different parts of the channel. Numbers refer to various examples of the same feature. A is the sequence typical of a major sandflat, B is transitional and C is the sequence generated by a channel setting. Inset shows the variations in direction of local flow and in the orientation of cross-channel bars (after Cant and Walker, 1978).

flow. They commonly emerge to split the flow and become dissected and modified (see below).

Repetitive, low relief forms of this type are similar in some respects to the 'sandwaves' described from some tidal flats but the hydrodynamic significance of the forms is not fully understood (cf. Crowley, 1983). In the Brahmaputra, very large bedforms up to 16 m high and 1 km long occur (Coleman, 1969) but their hydrodynamic similarity is not established.

(c) *Alternate bars* are features of rather straight, narrow channels and seem not to be very common. They are triangular in plan and are attached to alternate banks (Harms and Fahnestock, 1965; Maddock, 1969; Crowley, 1983). Similar bars also occur in the straighter channels of anastomosing streams (Smith and Smith, 1980).

All these transverse bars produce tabular sets of cross-bedding as a result of the downstream advance of their slip faces. The orientation of the cross-bedding will, however, be quite variable. The more skew-crested, cross-channel bars and the alternate bars give strongly bimodal divergence about the true downstream direction whilst the linguoid bars tend to give cross-bedding oriented downstream but with a wide unimodal spread. All will be susceptible to modification during emergence at falling or reduced stage (Collinson, 1970; Jones, 1977) (see below).

SAND FLATS

These composite forms are the largest morphological features of sandy stream beds (Fig. 3.10) (Cant and Walker, 1978). They occur in both mid-channel and marginal positions and have been called 'mid-channel bars' and 'side bars' (e.g. Collinson, 1970). They have no slip faces of their own, but gradually descend into flanking channels. They are composite bodies built up from accretion of smaller forms, mainly transverse bars and may grow from a nucleus produced by the emergence of a sector of a cross-channel bar (Cant and Walker, 1978). Once developed, such areas may persist for a long time and may become stabilized by vegetation. Some split the flow as an island which may then grow by vertical accretion of fine-grained sediment. In the Niger, such islands have survived for hundreds of years (NEDECO, 1959) and in the Volga, islands show signs of having grown by lateral accretion on their flanks (Shantzer, 1951). Such major accumulations will only be removed and replaced during major catastrophic floods.

The internal organization of sandflats is not known by direct observation, but their morphological development in the South Saskatchewan River suggests a plausible model (Fig. 3.10) (Cant and Walker, 1978). The accretion on to the sand flats of cross-channel bars with strongly skewed crest lines leads to tabular sets of cross-bedding oriented at quite high angles to the downstream direction, probably with a bimodal distribution symmetrical about the downstream direction. These tabular sets are interbedded with smaller scale tabular and trough sets whose directions are more closely grouped about the downstream direction. The pattern of cross-bedding of different scales, types and orientations contrasts with the unimodal pattern generated by the migration of dunes and linguoid bars in the channel areas between sand flats. The shifting of channels and flats through time will lead to the overall sequence illustrated in Fig. 3.10 but there will be a great deal of detailed horizontal variability.

WATER STAGE FLUCTUATIONS

Sandy low-sinuosity streams are commonly associated with quite large fluctuations of discharge at a variety of time-scales, some seasonal, some shorter and others longer and less predictable. The shape of the hydrograph reflects the climate of the catchment area and fluctuations may be reflected in the behaviour of the channels and of the bedforms. The large-scale scour and fill and shifting of channel position seen in the Brahmaputra (Fig. 3.8) (Coleman, 1969) have already been noted.

When areas of river bed are exposed at low water stage, it is commonly possible to observe several scales of bedform with superimposed relationships. These may, in some cases, be bedforms of different type, for example dunes and linguoid bars, whilst in other cases they may be of the same type but of different size. A dramatic illustration of the development of the second type of relationship is provided by sequences of echograms made over flood cycles (Fig. 3.11) (e.g. Pretious and Blench, 1951; Carey and Keller, 1957; Neill, 1969). The series illustrated in Fig. 3.11, shows how the bed responds to the

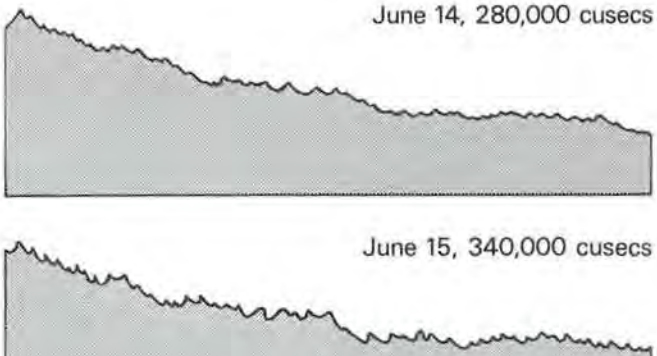

Fig. 3.11. Changing profile of bedforms in a reach of the Fraser River, British Columbia, during a flood event. Note the growth of forms beyond the flood peak and the superimposition of smaller forms during the falling stage. Length of reach is 670 m (after Pretious and Blench, 1951).

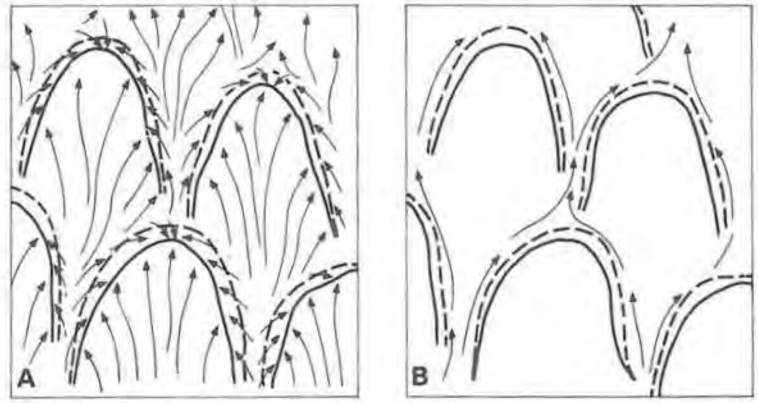

Fig. 3.12. Flow patterns associated with (A) high and (B) low discharge over a bed dominated by linguoid bars. At high stage, flow is over the bars whilst at low stage flow is concentrated between the bars (after Collinson, 1970).

changing flow with a significant lag period, a second set of smaller forms growing on the larger ones as stage falls. During a slow fall, larger forms may be obliterated by the growth of the smaller ones whilst a more rapid fall might leave the large forms emergent with few or no superimposed smaller forms.

The process of emergence may itself lead to modification of otherwise simple bedforms. As water level falls, partially emergent bars split the flow and the flow becomes concentrated in the topographic lows between bars, which function as sub-channels (Fig. 3.12). Ripples on the bar-tops and in the leeside areas are reoriented (Collinson, 1970); flow over the bar-tops may be split, leading to dissection of the bar and to the growth of small delta lobes in inter-bar areas (Fig. 3.7) (Collinson, 1970b; N.D. Smith 1971; Blodgett and Stanley, 1980). The emergent areas may also act as nuclei for the growth of mid-channel sand flats as outlined above (Cant and Walker, 1978).

As bars emerge, waves rework their slipfaces and build fringing ridges near the crestlines. Wave action may be erosive or it may reduce the slip face to a lower angle, sometimes associated with the concentration of heavy minerals. In sheltered areas along the fronts of bars and in inter-bar channels with low levels of flow, silt and clays may be deposited, in some cases veneering both the slip face and the rippled area immediately below it, a situation similar to that in slough channels of pebbly streams.

These modifications to the slip faces of bars are recorded in their internal structures (Collinson, 1970; N.D. Smith, 1971; Boothroyd and Ashley, 1975; Boothroyd, 1972; Blodgett and Stanley, 1980). The simple, tabular cross-bedding is complicated by low angle erosion surfaces (*reactivation surfaces*) which mainly record slipface reworking (Fig. 3.13). Similar structures may, however, result from the migration and overtaking of bedforms of different scales in a bedform population (McCabe and Jones, 1977). Normal cross-bedding which follows and overlies a reactivation surface results from a later high stage flow. Such reactivation surfaces are similar in their geometry to the third order bounding surfaces recognized in aeolian cross-bedding (Brookfield, 1977; Fig. 5.12). Flow parallel to the slip face of a fluvial bar during falling stage may lead to lateral accretion of ripple cross-laminated sand.

Details and extent of bedform modification depend upon both the regime of the river and upon the topographic level on the channel floor of a bedform. Rapid fall of water stage, as in the Tana River, favours the abandonment of bars and the development of simple reactivation surfaces. Slower rates, as in the Platte River (e.g. Blodgett and Stanley, 1980), favour bar dissection and the growth of delta lobes and skewed bar crests, producing more complex patterns of cross-bedding with a wider, possibly bimodal spread of directions (e.g. Cant and

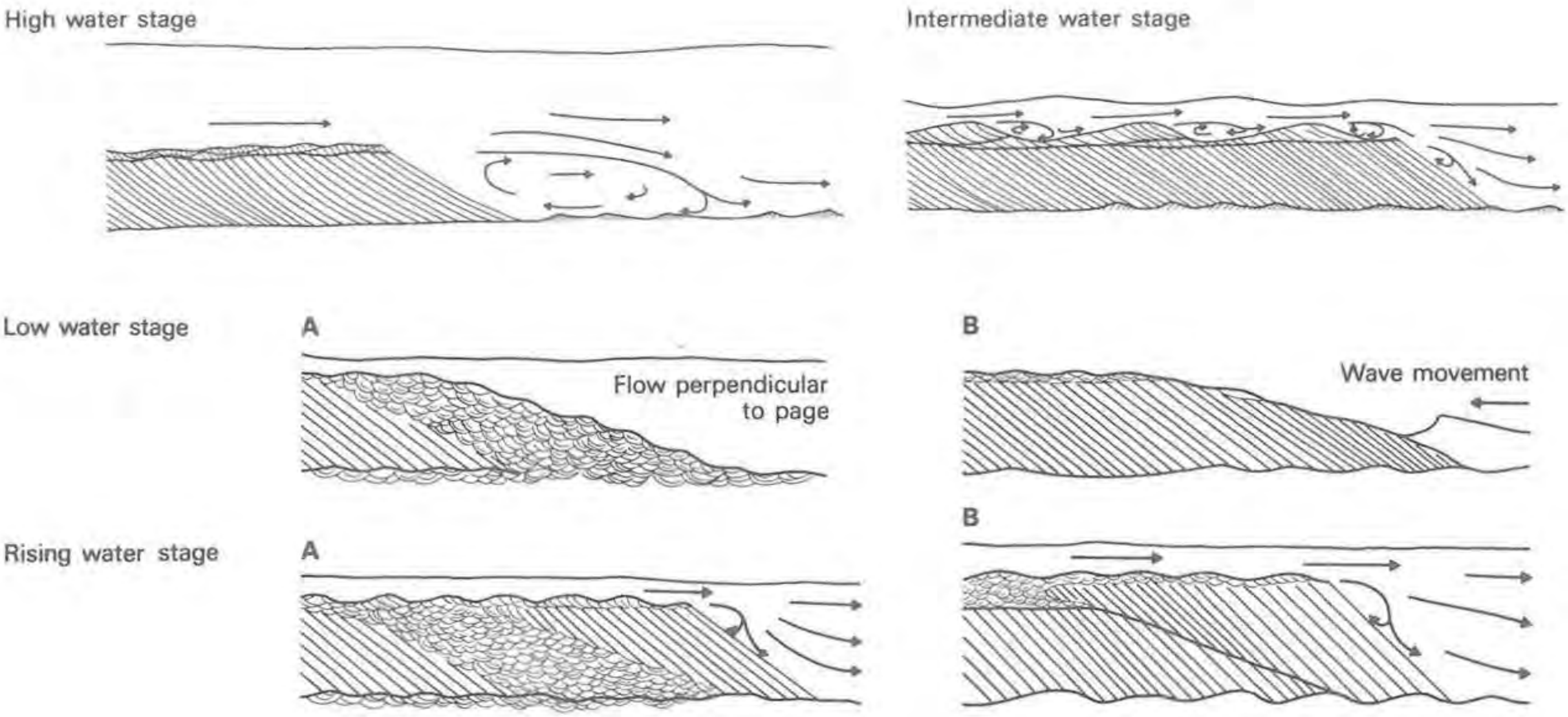

Fig. 3.13. Development of reactivation surfaces with changing water stage over a linguoid bar. Flow separation at high stage gives asymptotic foresets and counter-current ripples. Lowering of the water stage reduces the strength of the separation eddy, but does not immediately stop the advance of the lee face. Foresets due to avalanching have angular bases and bury inactive ripples. Further lowering may give currents parallel to the side of the bar, capable of depositing a laterally accreted unit (A) whilst wave activity during emergence can reduce the slope of the lee face (B). Rising water stage may reactivate the lee side deposition and cause burial and preservation of the falling stage feature with or without associated vertical accretion of the topset (after Collinson, 1970).

Walker, 1978). The detailed pattern of internal structure might therefore reflect a river's climatic regime (Jones, 1977). However, bedforms in topographically high areas of a river bed emerge and are abandoned more frequently than those in lower areas and attempts to infer a river's regime from internal structures in ancient deposits should proceed with careful regard to the position of the structures within the overall channel-fill sequence.

3.2.3 Semi-arid ephemeral streams

In many desert and desert margin settings, rainfall is mainly confined to short-lived and widely spaced storm events. The water discharge is therefore very high for short periods, but these are separated by long intervals when the sediment surface is exposed to the air. Infiltration of flood water into the bed is often an important factor in the dissipation of discharge and the streams commonly build terminal fans (see Sect. 3.3.1).

The stream beds show many of the features of more continuously flowing streams with bars, dunes, ripples and plane beds, all producing the expected internal structures (Fig. 3.9) (e.g. Williams, 1971; Karcz, 1972; Picard and High, 1973). Where a pre-existing channel is unable to cope with a major flood discharge, sheet flows may cover much wider areas flanking the channel. Where the stream flows in a valley, the sheet flows may totally cover the valley floor and deposit a sheet of sand. In Bijou Creek, Colorado, a sheet of sand 1–4 m thick was laid down on the valley floor (McKee, Crosby and Berryhill, 1967); it contained a high proportion of parallel lamination reflecting the upper flow regime conditions of the flood flow and ripple-drift cross-lamination recording a high rate of sediment accretion. Some sections showed tabular cross-bedding, probably the product of the late stage waning of the flood.

On some ephemeral stream beds and neighbouring areas, silt and clay may be laid down as a surface coating which, on drying, breaks up by polygonal cracking or curls up into mud flakes. Wind erosion leads to attrition of the clasts and to their wider dispersion as wind-borne dust. Sustained periods without floods may lead to the development of aeolian sand dunes on top of the stream deposits and these may block or divert channel patterns in subsequent floods. Intimate intermixing and mutual reworking of aeolian and fluvial sands may be a common feature of stream deposits of this type.

3.3 ALLUVIAL FANS

Alluvial fans are the larger scale morphological features built up by bedload streams and, more rarely, by streams with a high suspended load. In addition, they also occur in semi-arid settings where additional processes, in particular mass-flow, are important.

Fans of all types develop where the stream or mass-flow emerges from the confines of a valley or gorge into a basin. Lack of confinement allows horizontal expansion of the flow, deceleration and deposition of some or all of the sediment load. The emergence from a valley into a basin will commonly be associated with a reduction in gradient and this further favours deceleration and deposition.

Basins into which fans build are quite variable in character. They may be alluvial plains or valleys (e.g. Knight, 1975), inland drainage basins with or without tectonically active margins or bodies of standing water such as the sea or a lake. In the last case, the fans might be better termed 'fan deltas' (Sect. 12.4.3) (e.g. Wescott and Ethridge, 1980).

Overall, fans show a decrease in slope from the apex, close to the point of emergence, to the toe giving a concave upwards profile. Such a simple profile is, however, commonly broken into a series of segments. Each segment has a roughly even slope but the slopes of segments decrease sharply at particular points on the profile in a proximal to distal traverse (e.g. Bull, 1964). Such segmentation has been attributed to pulses of tectonic activity at the basin margin or to climatic changes. It may be associated with episodes of fan incision where the main feeder channel cuts into the upper part of the fan and emerges on to the fan surface at an '*intersection point*' (Fig. 3.14) (Hooke, 1967). Only below the intersection point does the flow expand and deposition take place.

The down-fan reduction in slope is commonly associated with a reduction in grain size, particularly the maximum particle size. There are also changes in the nature of the dominant bedforms in channels and in the dominant processes, depending on the type of fan. Two major types of fan have been distinguished, stream-dominated and semi-arid, though the division is arbitrary and gradational types occur.

3.3.1 Stream-dominated fans

Fans whose surface processes are dominated by streams flowing in channels have also been termed 'humid fans'. This is not a particularly apt term, however, as stream-dominated fans may also result from a lack of fine sediment in the source area and, as such, can occur in semi-arid settings, subject to sporadic floods. Stream-dominated fans are one of the main sites of deposition of low-sinuosity streams and have probably contributed a great deal to the geological record. They occur over a great range of scales from a few tens of metres up to hundreds of kilometres in radius. All tend to show a rather gradual reduction in slope and gradients are generally low compared with semi-arid fans. The largest described fan is that of the Kosi River which emerges from the Himalayan foothills to build a fan into the Ganges valley (Fig. 3.15) (Gole and Chitale, 1966). Historical records from this fan show the progressive shifting of the main channel across the surface of the fan over the past 230 years. We do not, at present, have sufficient data to know if such behaviour is typical of these fans. A more random switching of channel position through time is also known to occur on other fans through crevassing of banks because of constrictions in flow due to bars (e.g. Knight, 1975).

Where coarse material is supplied by the source area, as exemplified by proglacial outwash fans, there is commonly a downstream change from an upper fan with sheet bars through longitudinal bars as boulders die out to a more distal sandy channel with transverse bars in the lower fan (Fig. 3.16)

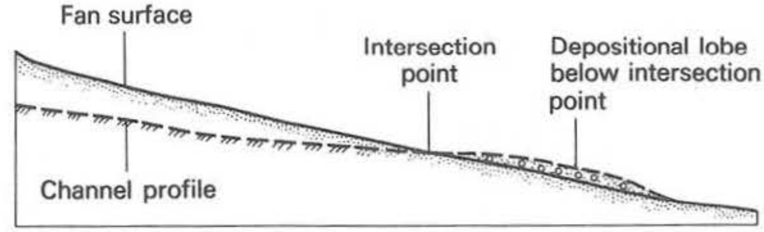

Fig. 3.14. Radial profile of an alluvial fan showing the position of the intersection point. This point will move up and down the fan surface in response to phases of incision and aggradation, probably related to tectonic activity (after Hooke, 1967).

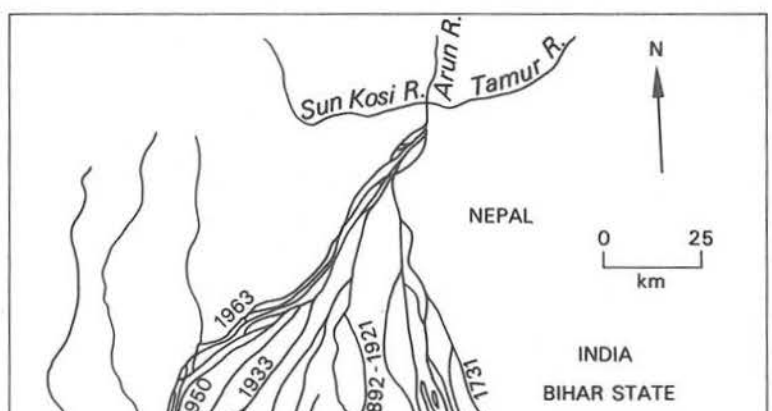

Fig. 3.15. The large, stream-dominated alluvial fan of the Kosi River on the southern flanks of the Himalayas; its channels have migrated from east to west over a period of 230 years (after Gole and Chitale, 1966).

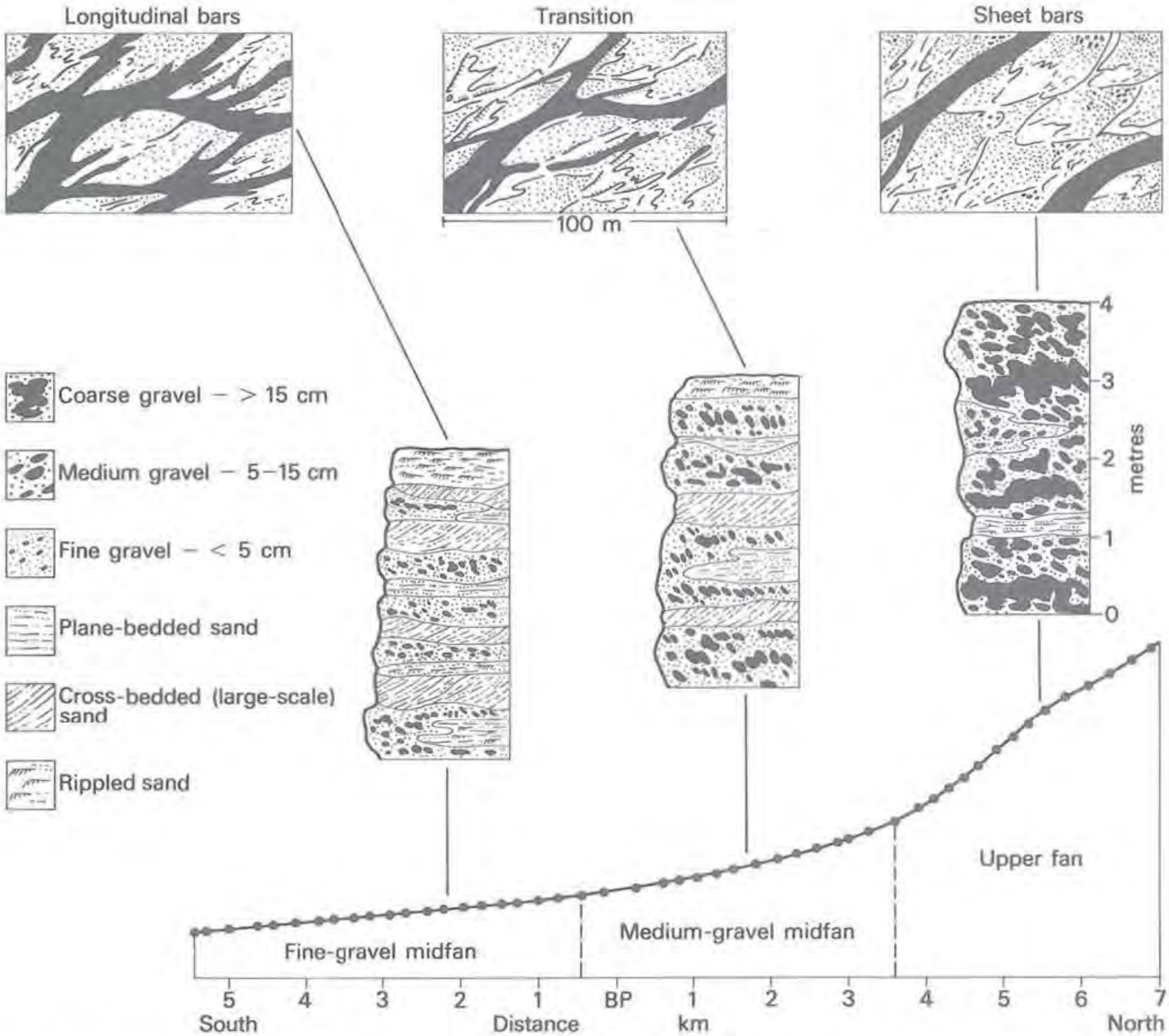

Fig. 3.16. Changes in slope, bar type and internal structure over a braided outwash fan in front of the Scott Glacier, Alaska (after Boothroyd, 1972).

(Boothroyd, 1972). The deposits of stream-dominated fans will be thick sequences of stacked channel deposits whose internal characteristics have already been described in terms of the more detailed channel processes (Sect. 3.2). Progradation and retreat of the fan for climatic or tectonic reasons may lead to changes of grain-size and structure in the vertical sequence.

TERMINAL FANS

A special case of stream-dominated fans is one in which the water discharge is progressively reduced down fan by a combination of evaporation and, more importantly, infiltration into the bed. The result is that no water exits from the system by surface flow. These terminal fans occur in arid basins of inland drainage where the stream flow is ephemeral (Mukherji, 1976; Parkash, Awasthi and Gohain, 1983). Channels split into networks of distributaries and subfans develop on the larger fan form (Fig. 3.17). Deposits are likely to be a mixture of channel deposits dominated by cross-bedding and sheet flood deposits of wider extent dominated by parallel lamination and ripple cross-lamination. Overall there will be a gradual proximal to distal decrease in the sand/mud ratio.

3.3.2 **Semi-arid fans**

These fans are the classical alluvial fans of tectonically active

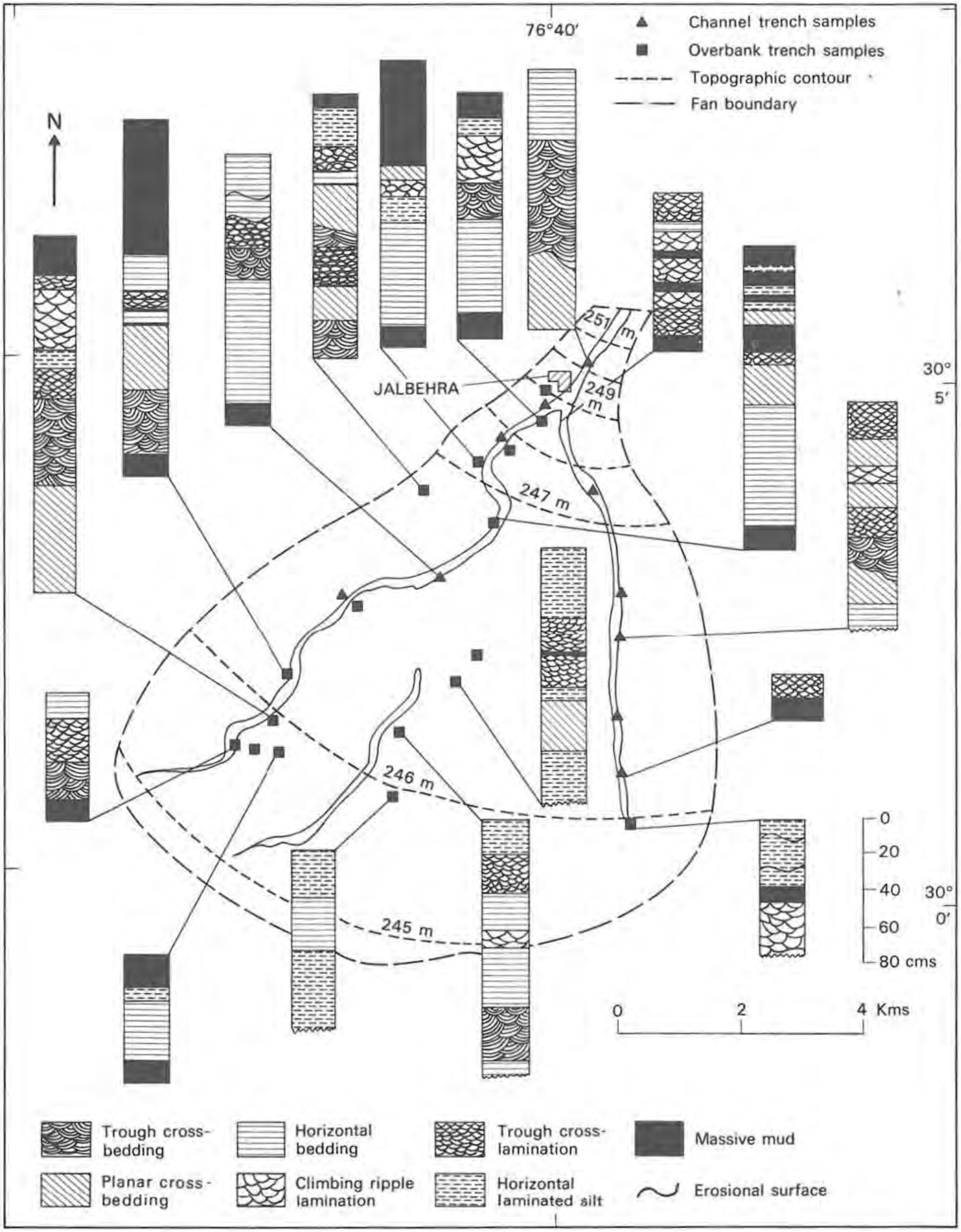

Fig. 3.17. Terminal fan of the Markanda River, India, showing the vertical sequences of structures encountered in shallow trenches. There appears to be a clearer differentiation of sand and mud in more proximal areas and an increase in silt distally (after Parkash, Awasthi and Gohain, 1983).

basin margins where mass-flow processes play a part in deposition. They have been most fully described from desert areas such as Death Valley (e.g. Bull, 1964, 1968; Denny, 1965, 1967; Hooke, 1967) but they may also occur in areas of high precipitation if there is abundant fine-grained material in the source area, low vegetation cover and high relief. The relief is commonly associated with an active fault line and, in such settings, adjacent fans may coalesce laterally as a sediment-accreting ramp, a 'bajada', within which it is possible to recognize fan sectors radiating from canyon mouths. The size of an individual fan is closely related to the size of its catchment area though lithology and climate also play a part. Where catchment areas are of similar size, mudstone-rich source areas tend to give fans which are considerably larger than those with sandstone-rich sources (Bull, 1964).

The relief on the surface of a fan varies with the size of the fan, large fans commonly having relief of hundreds of metres. The concave upwards radial profile which is common to all fans is particularly clear in small semi-arid fans. Smaller fans tend to have steeper slopes and those with a greater preponderance of mudstone in their source areas have 35–75% steeper slopes than those with sandstone for fans of similar size (Bull, 1964). Fan incision and segmentation are also particularly clear in semi-arid fans developed at fault lines.

PROCESSES AND DEPOSITS

Since most semi-arid fans are associated with intense ephemeral flood discharge, direct observations of processes are somewhat haphazard and reliance has often to be placed on after-the-event observations of the depositional products.

Four main types of depositional product have been recognized on modern semi-arid fans (Bull, 1972). These are inferred to be the products of distinct but intergradational processes based on the work of Blissenbach (1954), Bull (1964) and Hooke (1967).

Debris flow deposits (high viscosity)
Sheetflood deposits
Stream channel deposits } Fluid flow (low viscosity)
Sieve deposits

(a) *Debris flow deposits* are important on semi-arid alluvial fans (e.g. Blackwelder, 1928) and are the main reason for distinguishing these fans from stream-dominated ones. More recently the processes have been described in some detail but complete analysis is difficult owing to the unpredictable occurrence of the flows and the danger inherent in their close observation (Sharp and Nobles, 1953; Hooke, 1967; Johnson, 1970; Pierson, 1981). Debris flow and mudflow are here regarded as synonymous, though some authors suggest that debris flows should contain more coarse clasts.

The main prerequisites for debris flows are source rocks which weather to give some fine debris including clay, and steep slopes to promote rapid run-off and erosion. Debris flows move as dense, viscous masses in which strength of the matrix and buoyancy support clasts up to boulder size. The structures of flows are variable both between flows and within the same flow through time. More dilute, rapidly moving flows show some turbulence. More viscous, slower moving flows may be 'frozen' or 'rigid' at their edges and in a central plug where the applied shear is insufficient to overcome the shear strength of the deforming mass (Johnson, 1970; Middleton and Hampton, 1976; Pierson, 1981). The lateral edge zones may be preserved as levees where a flow has moved over the surface of a fan or as terraces on the sides of channels when the flow was confined above the intersection point. The whole flow stops when the zone of shearing around the central rigid plug ceases to be maintained. Some flows undergo a surging motion through time due to changes in fluidity (Sharp and Nobles 1953; Pierson, 1981). In the example described by Sharp and Nobles from southern California, surges moved at up to 4.4 m s^{-1} on slopes as low as 0.014 and laid down deposits up to 2 m thick as they decelerated. Material deposited by one surge may be remobilized by a later one.

The high viscosities of debris flows prevent them from sorting their load as they decelerate. All clast sizes are dumped together as the flow freezes, giving a very poorly-sorted deposit with larger clasts 'floating' in the finer matrix. Large clasts may actually protrude from the rather flat tops of the flow deposits. Debris flow deposits usually occur as rather narrow lobes and seldom give laterally extensive sheets. Only major catastrophic events which extend beyond a fan and onto a distal valley floor can be expected to give extensive sheets.

(b) *Sheetflood deposits* usually occur below the intersection point of a fan where flood flows, carrying sediment both as bedload and in suspension, expand laterally (Bull, 1972). The shallow sheet flows generally develop upper flow regime conditions and seldom persist far, largely due to infiltration (Rahn, 1967). Later stages of the flow split up into small channels which dissect the upper surface of the deposited sediment sheet. The result is a layer of fairly well-sorted sand or fine gravel with small-scale lenticularity and scouring. Cross-bedding and cross-lamination may occur but are not ubiquitous.

(c) *Stream channel deposits* tend to be more important on the upper parts of fans where flows are more likely to be confined. However, channels also occur lower on fans due to the emergence of groundwater. Channelized flow commonly occurs during the waning stages of floods when there is a tendency for the flow to rework and winnow earlier, less well-sorted deposits. The deposits are generally lenticular sands and gravels, the coarser layers commonly showing imbrication and the sands showing cross-bedding. Where these deposits are volumetrically important, the fan deposits will be gradational with those of stream-dominated fans (see Sect. 3.3.1).

(d) *Sieve deposits* commonly occur just below the intersection point when the sediment load of a flood is rather deficient in finer grained sediment. Permeable, earlier deposits allow rapid infiltration of water from the flow into the body of the fan. This results in the deposition of a clast-supported gravel lobe (Hooke, 1967). The lobes tend to have a clearly defined downstream margin and seem to develop from the earliest clear water stages of a flood (Wasson, 1974). The small sand lobes described by Carter (1975b) as debris flows are, in fact, small-scale sieve deposits resulting from rapid dewatering of sand slurries.

Sieve deposit gravels are probably rather well-sorted and poorly imbricated and may be composed of angular fragments. With burial, the interstices are slowly filled with finer, infiltrating sediment, giving the final sediment a markedly bimodal grain-size distribution.

POST-DEPOSITIONAL PROCESSES

At any particular time, only a small area of a fan receives sediment. Elsewhere, sediment-starved areas undergo a variety of post-depositional processes which may last for hundreds of years and substantially modify depositional features (Denny, 1967). Weathering, run-off and the wind are the main post-depositional agents. Chemically unstable clasts continue to break down and the products are washed either into the underlying sediment or over the fan surface into more distal environments. Some fine material might be moved by wind deflation and wind-blown sand might accumulate as dunes, commonly on the lower parts of a fan surface. Run-off leads to gullying whilst areas between gullies may develop a desert pavement of closely packed angular clasts commonly coated with desert varnish. Such pavements protect the fan surface from further deflation and the coarse clasts commonly overlie a silty layer. In semi-arid settings, fan sediments often become red as weathering breaks down ferro-magnesian minerals and biotite into clays and haematite (Walker, Waugh and Crone, 1978). Such transformations take thousands of years and are facilitated by wetting and drying. A fuller account of red-bed development is given in the discussion of inter-channel areas (Sect. 3.6.2).

DISTRIBUTION OF FAN PROCESSES AND PRODUCTS

Debris flow deposits tend to be more common on the upper parts of fans whilst sheetflood deposits occur more commonly on lower areas. Sieve deposits are concentrated around intersection points. Channel deposits can occur in almost any position either radiating from the fan apex or lower on the fan surface owing to the emergence of groundwater. The size of the largest clasts diminishes down-fan particularly in the deposits of debris flows compared with stream flows (Bluck, 1964). Maps of the distribution of deposits on present-day fans show highly variable and unpredictable patterns even between closely adjacent fans (Fig. 3.18) (Hooke, 1967). The unpredictable nature of the processes in detail and the shifting of the locus of deposition

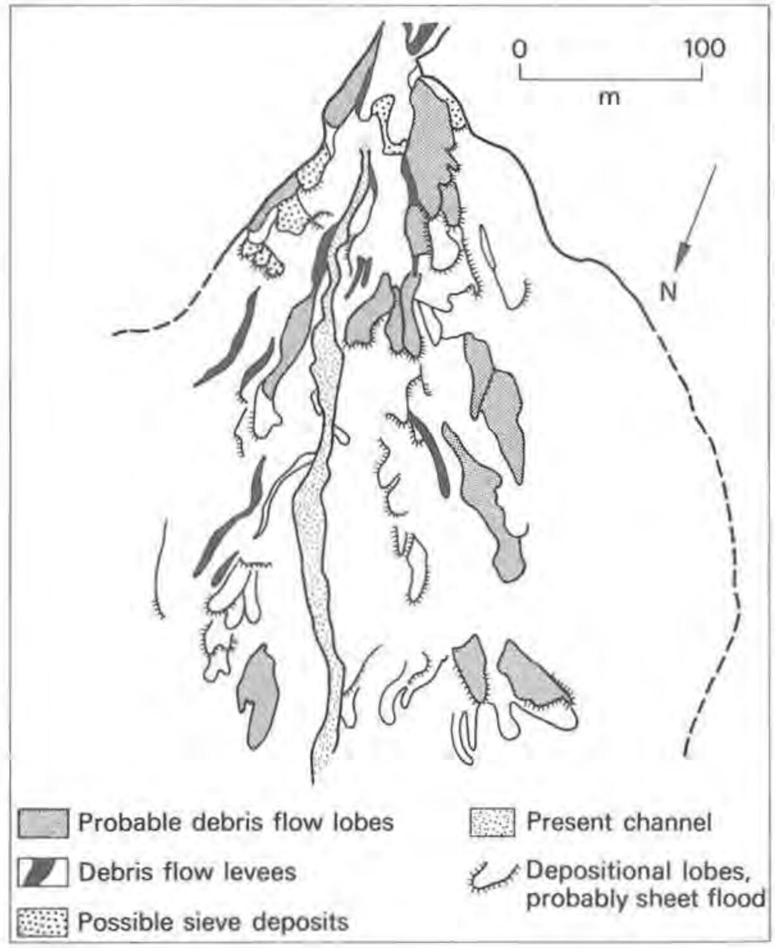

Fig. 3.18. The upper part of the surface of the Trollheim Fan, Death Valley, California, showing the distribution of recent debris flow lobes and levees. Other lobes are of sieve deposit or sheetflood origin. Undecorated areas are deposits of uncertain origin and fragments of older fan surfaces where post-depositional modification has masked the original character. The relief over the area shown is about 100 m (after Hooke, 1967).

over the surface of the fan lead to an essentially random interbedding of sheetflood, debris flow, channel and sieve deposits. Incision and segmentation of fans can lead to sequences which coarsen or fine upwards and if recognized in the ancient, these could form the basis for inferences about tectonic and climatic change during deposition (e.g. Heward, 1978).

3.4 PRESENT-DAY MEANDERING RIVERS

Meandering rivers are those where the channel has a markedly sinuous pattern. The sinuosity is commonly regular with a wavelength related to channel width. Meandering seems to be a feature of rather low slopes and is favoured by an abundance of fine-grained sediment both in the river banks and in the total sediment load. Meandering streams generally show a more

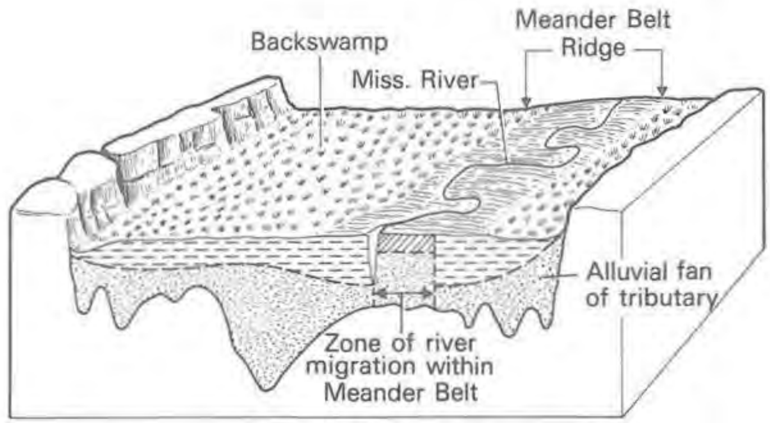

Fig. 3.19. The alluvial plain of the Mississippi River. The main channel occupies a meander belt on top of an alluvial ridge flanked by floodplains. The plain is about 150 km wide overall and the alluvial ridge about 15 km wide. Large vertical exaggeration (after Schumm, 1971b).

organized pattern of channel processes and a clearer separation of channel and overbank environments than is the case with low-sinuosity streams. They commonly occur on alluvial plains both within the confines of valleys or terraces or on more open tracts of country. The Gulf Coast plain of the southern USA is traversed by a suite of meandering streams each of which appears to have its zone of influence. Within the valley floor or the zone of influence the stream occupies only a small part at any one time. The channel lies within a meander belt which is a complex zone of active and abandoned channel environments and nearby overbank environments. Beyond the meander belt lie the more distant overbank or floodplain areas. When sinuosity is high, the position of the meander belt may be quite stable for a considerable period as the clay plugs generated by the infill of channel cut-offs prevent lateral migration. Throughout the life of a particular meander belt, sedimentation is most rapid near the belt close to the channel with the result that an alluvial ridge develops above the level of the more distant floodplain (Fig. 3.19) (Fisk, 1952b). This increasingly unstable situation is periodically relieved by *avulsion*. Probably during a flood, the channel bank will be breached and a new course will be established along the lowest route on the flood plain (Speight, 1965). This transfer commonly takes place gradually and it may be several years before all the discharge follows the new route. A new meander belt will then develop and a new cycle begin. The frequency of avulsion depends on the rate at which transverse slopes develop from alluvial ridge to flood-plain. This in turn is controlled by a complex interaction of hydrological and sediment properties unique to the river in question.

Meandering streams have a wide range of channel sediments, from gravels to muds. They are gradational with low-sinuosity and anastomosing channel patterns (Sect. 3.5). In order to deal with this spectrum, it is necessary to discuss high-sinuosity sand-bed streams as a norm and then to discuss other types in a comparative way. Floodplain and near-channel overbank deposits which are common to a wide variety of channel types are dealt with separately (Sect. 3.6.1).

3.4.1 Meander belts

Meandering is favoured by relatively low slopes, a high suspended load/bed load ratio and by cohesive bank materials (Leopold and Wolman, 1957; Schumm and Kahn, 1972). A relatively steady discharge regime may also help. However, these generalizations do not prevent meandering in coarse bed materials on high slopes and with quite flashy discharge and every channel pattern results from a subtle interplay of factors.

The relationships between meander geometry and other channel and discharge parameters have been studied both theoretically and empirically. The shape of a meander represents an adjustment of depth, velocity and slope to minimize the variance of shear and frictional resistance (Langbein and Leopold, 1966). Meander wavelength (λ) has a nearly linear relationship with channel width (w) and with the radius of curvature of the meander (r)

$$\lambda = 10.9\, w^{1.01}$$
$$\lambda = 4.7\, r^{0.98}$$

(Leopold and Wolman, 1960).

The relationship between wavelength and water discharge is complicated by the need to decide the most appropriate discharge parameter (Carlston, 1965). Mean annual discharge ($\bar{Q}$) and the mean of the month of maximum discharge ($\bar{Q}_{mm}$) give the closest correlations:

$$\lambda = 106.1\, \bar{Q}^{0.46} \quad \text{(standard error = 11.8\%)}$$
$$\lambda = 80.6\, \bar{Q}_{mm}{}^{0.46} \quad \text{(standard error = 15\%).}$$

This leads to the deduction that meander width (W_m) and channel width (w) are related to the mean annual discharge by the following equations:

$$W_m = 65.8\, \bar{Q}^{0.47}$$
$$w = 7\, \bar{Q}^{0.46}.$$

Several models for meander sedimentation are based on an assumption of bankfull discharge. Leeder (1973) showed that when sinuosity is greater than 1.7, bankfull channel depth (h) and bankfull width (w) are related by:

$$w = 6.8h^{1.54}$$

thereby giving a means of estimating widths from depth, a parameter which may be reasonably deduced from the rock record. By combining various empirical equations, Collinson (1978) derived a relationship between depth and meanderbelt width

$$W_m = 64.6h^{1.54}.$$

Equations of this type can give a guide in palaeohydraulic reconstructions but they reflect complex interplays of variables and have large variabilities. They also demand some pre-knowledge of the type of channel in question.

3.4.2 Channel processes

Deposition on point bars of meandering streams was the dominant theme in discussions of sedimentation in channels until quite recently and an elegant process/response model has been developed. Emphasis has now not only shifted towards an appreciation of low-sinuosity streams but also it has become obvious that point bars themselves are highly variable and complex. In this account, the classical model will first be described and the complexities of real point bars are then compared with it.

THE CLASSICAL POINT BAR MODEL

It has long been recognized that flow in meander bends is helicoidal with a component of surface flow towards the outer bank and bottom flow towards the inner bank (e.g. Fisk, 1947; van Bendegom, 1947; Sundborg, 1956). The locus of maximum depth in the channel, the *thalweg*, corresponds roughly with the zone of maximum velocity (e.g. Bridge and Jarvis, 1976, 1982; Geldof and de Vriend, 1983) with scour pools developing near the outside bank. In simple curved bends, the velocity, asymmetry and the position of the thalweg change over between bends as the helicoidal flow changes its sense of rotation. More complex bend shapes are associated with complex distribution patterns of both depth and velocity (Hooke and Harvey, 1983). As a result of the flow pattern, the outer concave bank is usually the site of erosion and the inner convex bank the site of deposition, the channel as a whole migrating transversely to the flow to deposit a unit of sediment by lateral accretion.

(a) *Erosion*. Erosion of the concave bank is influenced by the nature of the bank material. Floodplain silts and clays of high cohesive strength resist erosion unless they are underlain by channel sands. Thick cohesive sediments are eroded as blocks which founder into the channel by under-cutting and by the development of shear planes trending sub-parallel to the bank (Fig. 3.20) (Sundborg, 1956; Klimek, 1974a). Such planes are curved in plan and if they penetrate deeply they may allow the emplacement of bank material below the level of the channel base by rotational slumping (Turnbull, Krinitzsky and Weaver 1966; Laury, 1971). Sands in bank material, particularly where water saturated, are likely to slough into channels, and flowage towards the channel hastens undercutting.

The material eroded from the concave bank usually falls or slides into the deepest part of the channel where it is winnowed to give a lag conglomerate. Blocks emplaced below the level of the thalweg by large-scale basal sliding avoid this reworking.

(b) *Point bar deposition*. The sediment body enclosed by the meander loop is the *point bar*. It has an essentially horizontal top surface at about the level of the surrounding floodplain and it slopes from that surface level down to the thalweg of the channel. This *point bar surface* is the site of channel deposition and the classical model predicts the pattern of distribution of grain size and bedforms over this surface and, thereby, the vertical sequence of facies produced by lateral accretion (Sect. 3.9.4).

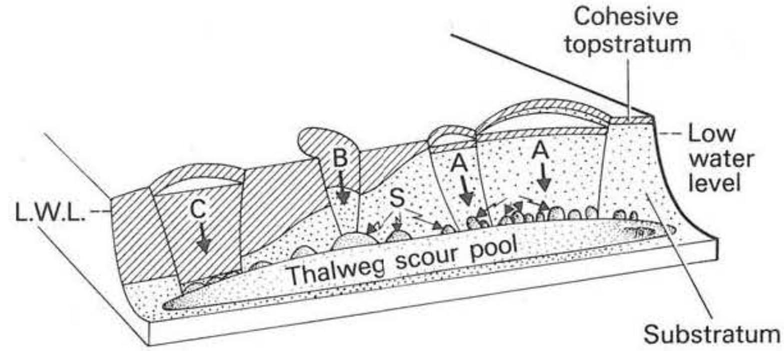

Fig. 3.20. The effect of the thickness of cohesive top stratum on channel bank failure. Accelerated scour at the thalweg and at the foot of the bank during high water stages is followed by subaqueous failures (either by shear or flow) in the relatively non-cohesive substratum sands or gravels. For thin (A) or very thick (C) top stratum, subaqueous failures (S) are numerous and small, initiating shallow upper bank failure by shear. Intermediate top stratum thickness (B) promotes small to large subaqueous failures followed by deeper, bowl-shaped upper bank failure (after Turnbull, Krinitzsky and Weaver, 1966).

The pattern of water flow around a meander bend is the key to understanding deposition on the point bar surface. The model, first outlined by van Bendegom (1947) was developed independently by Allen (1970a and b) and extended by Bridge (1975, 1977). It assumes a bankfull discharge and a fully developed helicoidal flow around the bend. Depth, velocity and boundary shear stress diminish away from the thalweg and, in combination with the upslope component of the helicoidal flow, lead to the near-bed flow operating as an elutriator giving an upslope reduction in grain size provided that a sufficiently mixed grain-size population is available for transport. This theoretical pattern of water movement and sedimentary response is borne out quite well by several natural rivers (e.g. Bridge and Jarvis, 1982; Geldof and de Vriend, 1983). Dunes tend to be the dominant bedforms on the lower parts of point bars whilst ripples and plane beds occur higher up. Sandwaves occur in a less predictable way (e.g. Sundborg, 1956; Frazier and Osanik, 1961; Harms, MacKenzie and McCubbin, 1963).

In addition to the pattern of transverse bedforms which occur on point bar surfaces, elongated ridges of sand (*scroll bars*) trend more or less parallel to the contours of the point bar. They originate low on the point bar surface and gradually migrate upslope until they reach bankfull level. They are here abandoned giving a series of ridges on top of the point bar (Fig. 3.21). The result is a pattern of roughly concentric swales and ridges from which it is possible to deduce the erosional path lines of individual meanders (Hickin, 1974).

(c) *The vertical sequence* predicted by the model of point bar sedimentation is fairly well established. Lateral migration of the channel gives a tabular sand unit overlying a near-horizontal

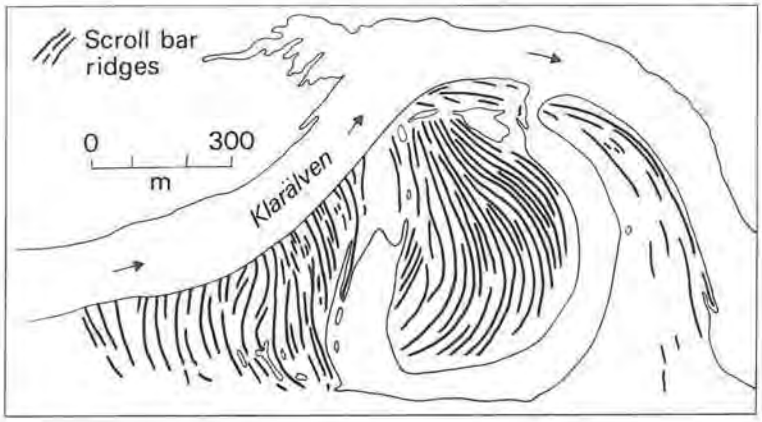

Fig. 3.21. Accretion topography of scroll bars on a point bar bordering Klarälven, Sweden (after Sundborg, 1956).

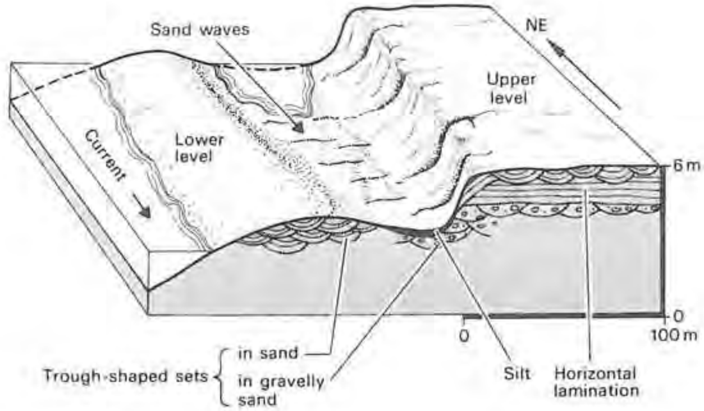

Fig. 3.23. Upstream part of the Beene point bar, Red River, Louisiana, showing the stepped profile and the associated internal structures observed in trenches (after Harms, MacKenzie and McCubbin, 1963).

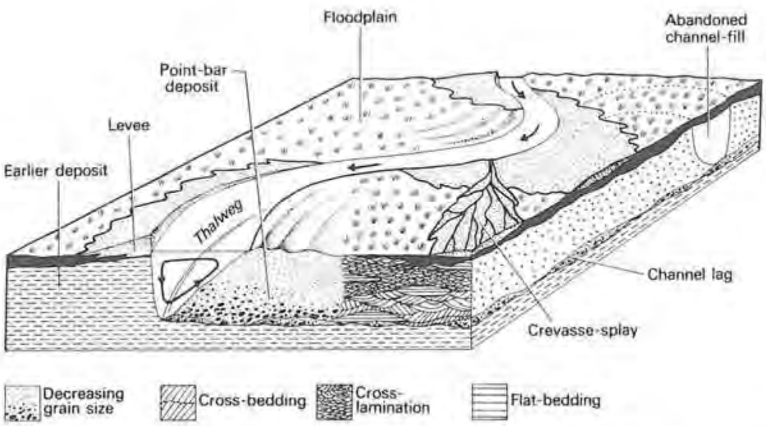

Fig. 3.22. The classical point bar model for a meandering stream (after Allen, 1964, 1970b).

erosion surface, with or without a lag conglomerate. The sands above the erosion surface show an upwards diminution both of grain size and of set size of cross-bedding. In the upper part, cross-bedding gives way to ripple cross-lamination and parallel lamination (Fig. 3.22) (Sundborg, 1956; Frazier and Osanik, 1961; Bernard and Major, 1963; Bridge and Jarvis, 1982). Tabular sets which result from the migration of scroll bars may be bigger than neighbouring trough sets and show a divergence of direction towards the convex bank (Sundborg, 1956; Jackson, 1975a). The thickness of the overall sand sequence compares closely with the depth of the channel and the relative abundances and distribution of the various structures are controlled by channel size and sinuosity (Allen, 1970a, b).

VARIATIONS AND COMPLICATIONS

The classical model of point bar deposition is based on assumptions of bankfull discharge, fully developed helicoidal flow and uniformity of conditions along the length of the point bar surface. It has become clear that while some rivers behave more or less in line with the model, many show features which reflect non-uniformity and unsteadiness of flow and the influence of lower discharges. Some of the divergences from the classical model become most apparent when the bed material contains a high proportion either of gravel or of fine, suspended-load sediment.

(a) *Two tier point bars*. Some point bars have distinctly stepped profiles and the steps appear to relate to recurrent discharges below bankfall. Few examples have been described in detail. On the sandy Beene point bar, trough cross-bedding dominates the sequence above and below the level of its step (Fig. 3.23) (Harms, MacKenzie and McCubbin, 1963). Silt, deposited during the falling stage, mantles the floor of a shallow channel cut into the step and the higher deposits are somewhat coarser than those below.

(b) *Coarse-grained point bars*. Gravelly streams tend to have somewhat lower sinuosities than sand-bed streams of similar size and at the extreme case they are gradational with low sinuosity streams with side bars (see Sect. 3.2.2). Several examples of coarse-grained point bars have been described but their variety makes generalization difficult if not dangerous. On gravel-rich point bars, the dominant bedforms are likely to be either flat pavements of imbricated clasts or transverse bars with straight or curved crest-lines (e.g. Gustavson, 1978). Gravel-bearing point bars do, however, tend to show a downstream diminution in grain size with gravels more abundant at the bar head (upstream end) (e.g. Bluck, 1971; Levey, 1978). In the Jarama River in central Spain (Arche, 1983) the gravel appears to form the bulk of the point bar, even contributing to the swale and ridge topography of the point bar top. This is then either buried by silt and sand during flood discharge without erosion or alternatively, a thicker sand unit with a scoured base overlies the gravels. This sand unit appears to result from the establishment of a high stage channel cut into the top of the gravel point

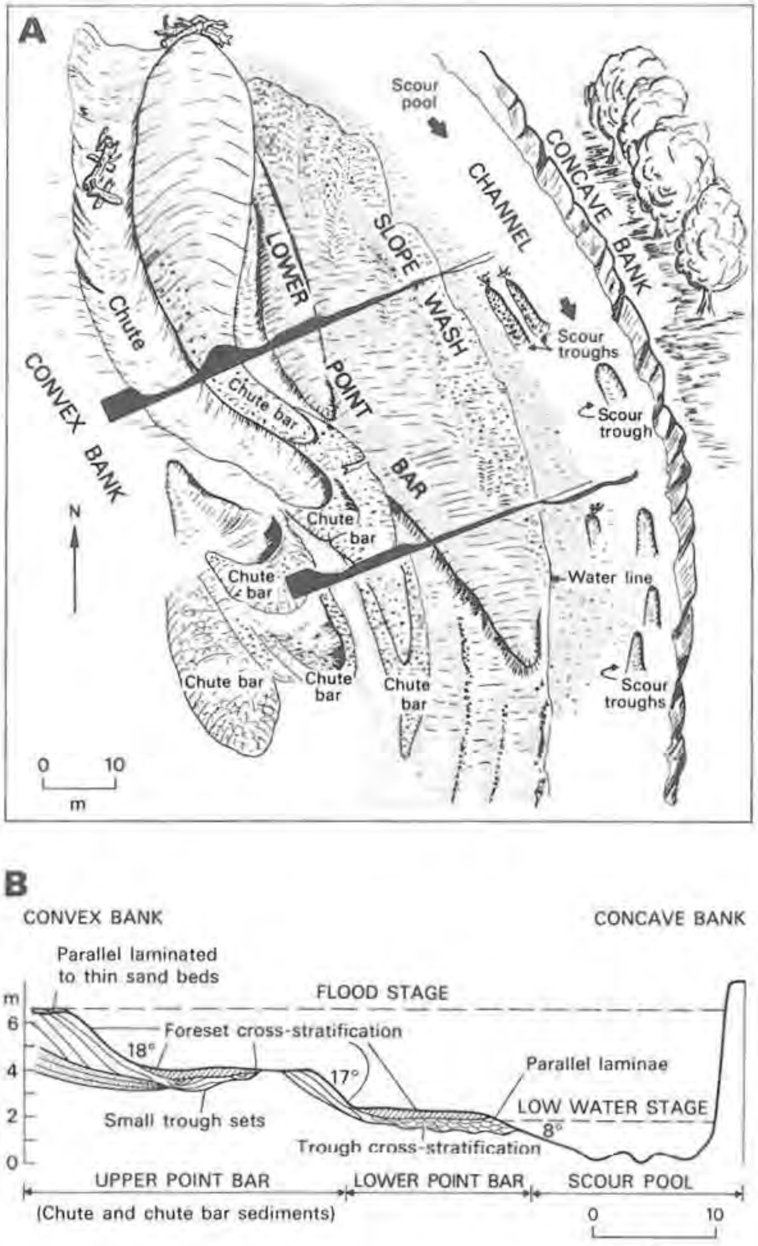

Fig. 3.24. Morphological features and internal structures of a coarse-grained point bar in (A) plan and (B) cross section, Amite River, Louisiana (after McGowen and Garner, 1970).

bar, possibly localized by the pre-existing ridge and swale pattern. Indeed, minor channels cut into the point bar, so called '*chutes*', seem to be quite common (e.g. McGowen and Garner, 1970; Levey, 1978). Floods tend to flow across the point bar surfaces, giving two major threads, one following the thalweg near the concave bank and one across the point bar surface. In the Colorado and Amite Rivers of Texas and Louisiana, channels ('*chutes*') not unlike those associated with two-tier point bars, are eroded into the upstream ends of the point bars and '*chute bars*' are deposited at their downstream ends (Fig. 3.24) (McGowen and Garner, 1970). Chutes vary in size and shape and are normally floored by gravel. At their downstream ends, their bases intersect the point bar and flow can expand to deposit chute bars. Successive floods may partially fill the chutes with graded beds a few tens of centimetres thick and ending in a vegetated mud drape.

Chute bars occur at various levels on the point bar surface and are most common near its downstream end (McGowen and Garner, 1970; Levey, 1978). They are commonly associated with the downstream end of a chute channel, but also appear to form as the result of convergent flow patterns at the downstream ends of unchannelled point bars (e.g. Gustavson, 1978). They grow both by the advance of a well-developed slip face and also by vertical accretion of their top surface. Slip faces are strongly convex downstream and may be 2–6 m high. The result is that unusually thick tabular sets of cross-bedding occur within sequences which may be otherwise made up of smaller trough and tabular sets. The chute bar foresets have a wider directional spread than associated smaller cross-bedding and may be punctuated internally by reactivation surfaces (Levey, 1978).

(c) *Muddy point bars*. Following Bagnold (1960), Leeder and Bridges (1975) cast doubt upon the general applicability of a simple helicoidal flow pattern for flow round a meander bend. At high curvatures, when r_m/w is less than about 2, large-scale flow separation occurs at the downstream end of the point bar surface (where r_m is the radius of curvature of the channel mid-line and w is channel width). When this curvature is exceeded in a developing meander, the growth direction changes from being transverse to down-valley. Such high sinuosities are most commonly developed in very muddy sediment, and as such are particularly important in tidal creeks. They also occur in very low slope, muddy rivers where the secondary flow patterns lead to the deposition of benches of material on the outer, concave banks of bends and to some degree of erosion on the inner convex banks (Fig. 3.25) (e.g. Taylor and Woodyer, 1978; Woodyer, Taylor and Crook, 1979; Page and Nanson, 1982).

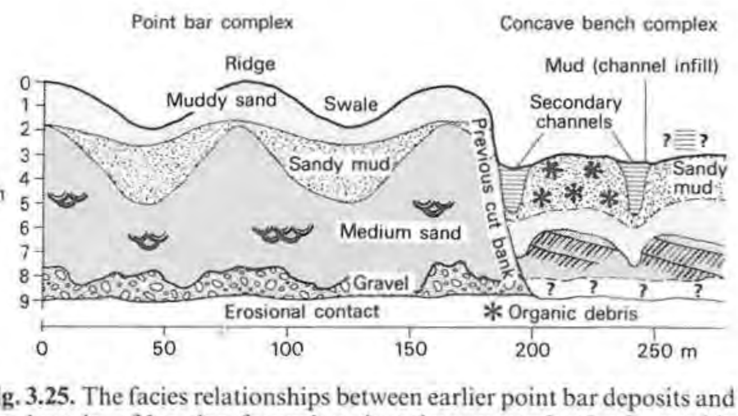

Fig. 3.25. The facies relationships between earlier point bar deposits and the deposits of benches formed against the concave bank of a muddy meandering river; Murrumbidgee River, Australia (after Nanson and Page, 1983).

These depositional benches do not reach the same height as the top of the main point bar and they rest with a laterally erosive contact against more normal point bar deposits. Their preservation potential and volumetric importance are likely to be rather low but such deposits could be confused with the infills of abandoned channels.

(d) *Out-of-phase flow patterns.* On virtually all meander bends the helicoidal flow is not immediately established at the inflection point but develops over a finite distance. In the upstream part of any meander bend, the flow pattern is largely inherited from the next bend upstream with rotation in the opposite sense (Jackson 1975a,b). In consequence, three gradational zones can be recognized around any bend: a 'transitional zone' where the influence of the upstream bend prevails; a 'fully developed zone' where the local bend dominates and the classical model prevails and an 'intermediate zone' where one pattern changes into the other. The extent of these zones varies with channel curvature and water stage. In the transitional zone,

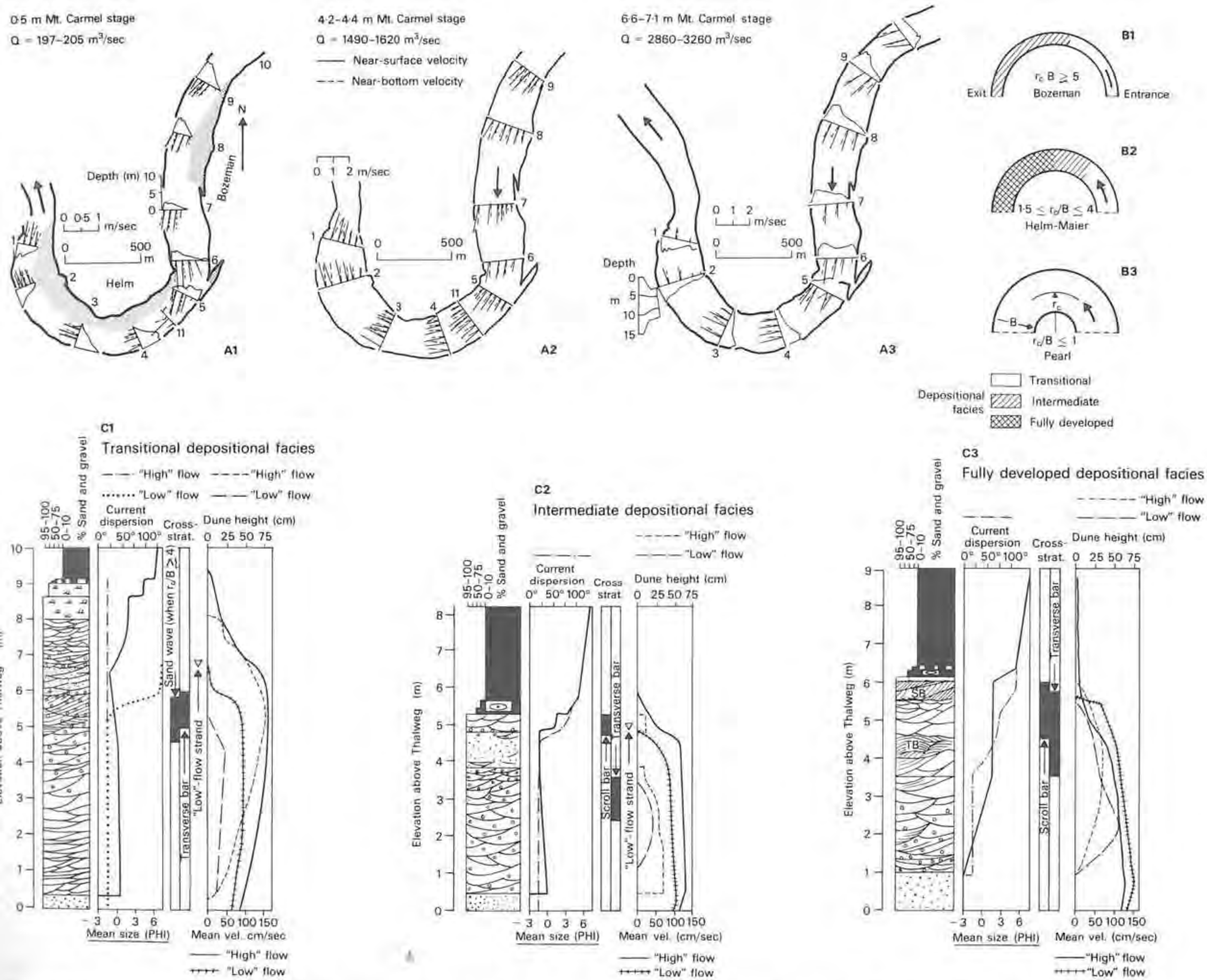

Fig. 3.26. The variation of water flow pattern and facies distribution on point bars in the Lower Wabash River, Illinois. A_1 to A_3: Cross-channel profiles and velocity distribution at Helm Bend at increasing values of discharge. B_1 to B_3: Distribution of depositional facies on point bars of increasing channel curvature. C_1 to C_3: Depositional facies sequences from upstream to downstream on a point bar of a curvature appropriate to the full facies development (after Jackson, 1975b, 1976).

velocity is highest near the inner bank, particularly at bankfull discharge, but the thalweg switches to the outer bank upstream of the point at which the velocity pattern and sense of rotation change (Fig. 3.26).

These down-channel changes are reflected in the bedforms on the point bar surface and in the vertical sequence of sediment produced. In the fully developed zone the classical model applies well enough. However, in the upstream transitional zone the pattern of grain-size change is unclear and the sequence may coarsen upwards. Also, the pattern of sedimentary structures is less well ordered with ripples rather than dunes in the deeper part of the channel (Fig. 3.26C).

The general rule which becomes clear from studies of modern point bars is that a whole variety of vertical sequences may be produced not only by different point bars but also within the deposits of a single point bar.

3.4.3 Channel cut-offs

A freely meandering channel is inherently unstable as differing rates of erosion in neighbouring meanders lead to periodic channel cut-offs. These are of two main but intergradational types: (a) chute cut-offs and (b) neck cut-offs (Fig. 3.27) (Fisk, 1947; Lewis and Lewin, 1983).

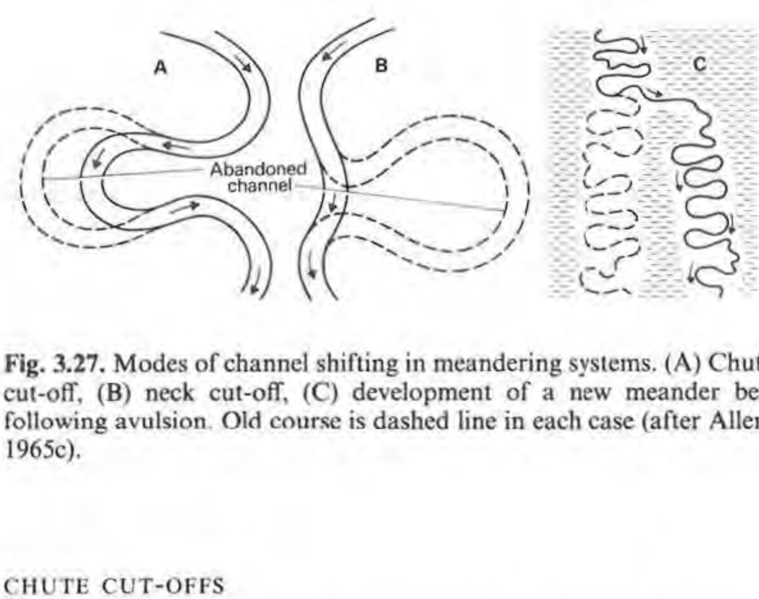

Fig. 3.27. Modes of channel shifting in meandering systems. (A) Chute cut-off, (B) neck cut-off, (C) development of a new meander belt following avulsion. Old course is dashed line in each case (after Allen, 1965c).

CHUTE CUT-OFFS

Streams tend to straighten the course of their main flow across meander bends during flood, cutting chutes into the point bar or deepening swales on the point-bar top. These channels may take an increasing proportion of the flow with the result that activity in the main channel is gradually reduced. Deposition in the channel is first by bedload and later by silts and clays from suspension as the ends of the cut-off reach are plugged with sediment.

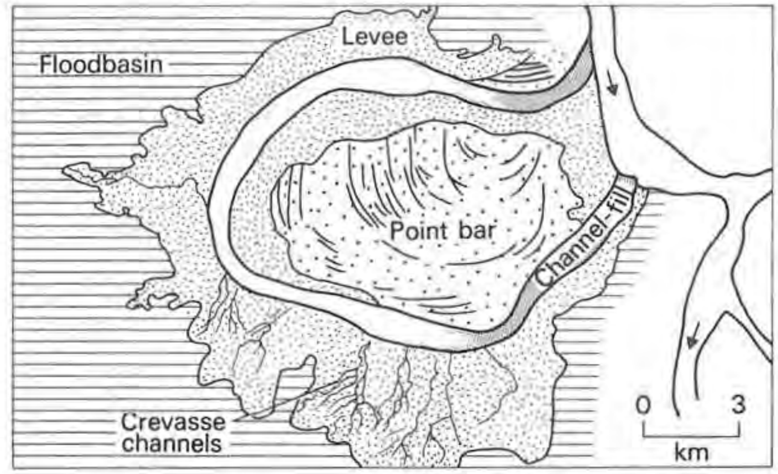

Fig. 3.28. Levee, crevasse and crevasse splay topography preserved around an ox-bow lake caused by neck cut-off. False River cut-off channel, Mississippi River (after Fisk, 1947).

NECK CUT-OFFS

Concave banks of adjacent meanders may sometimes erode towards one another, narrowing the area of point bar between them. If this neck is breached, the river will abandon the meander loop rather abruptly. The abandoned reach will then be rapidly plugged at both ends by material washed in by separated flow eddies and an oxbow lake will be created (Fig. 3.28). Such a lake will only receive sediment from suspension during flood and may exist as a lake for a long time. Bedforms, active on the point-bar surface immediately prior to cut-off, may be preserved below a drape of suspended sediment (McDowell, 1960).

Both chute and neck cut-off loops may be sites of dense vegetation growth and of accumulation of organic-rich muds and silts. They form restricted bodies of fine sediment curved in plan view and bounded on the outer side by an erosion surface and on the inner side by the inclined surface of the point bar sediments. The two types differ in the abruptness of the transition from bedload to suspended sediment.

3.5 ANASTOMOSING CHANNELS

In contrast with the highly mobile braided streams and the less mobile but still migrating meandering channels, anastomosing streams, once established, appear to be characterized by extremely stable channel positions (Smith and Smith, 1980). Anastomosing streams are those which are split into a series of sub-channels which divide and rejoin on a length scale many times the channel width. Each sub-channel may, in turn, be highly sinuous or relatively straight. These streams are characteristic of areas with very low down-stream slopes. They are common in swamps and marshes, on delta tops (e.g. Axelsson,

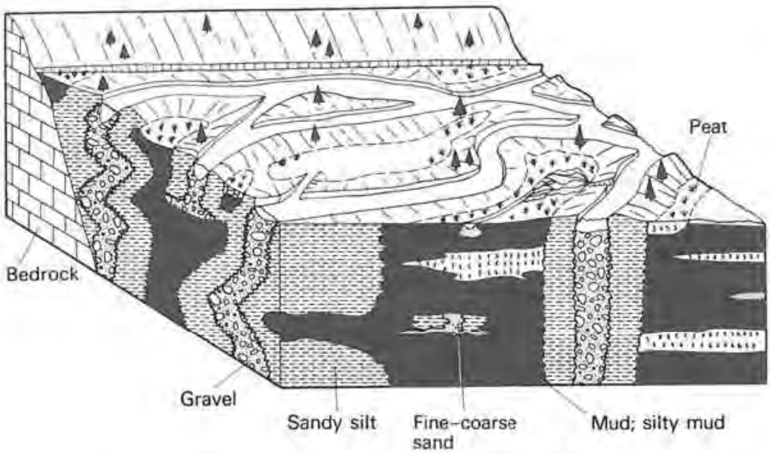

Fig. 3.29. Three dimensional distribution of facies beneath an anastomosing channel system, based on a series of close-spaced borings in the plains of the Alexandra and North Saskatchewan Rivers, Alberta, Canada (after Smith and Smith, 1980).

1967) and where valley floors are adjusted to a local base level such as a cross-valley barrier. In semi-arid settings, the anastomosing pattern may develop as a response to a reduced discharge regime, following a pluvial episode when low slopes were established by a high discharge braided system (Rust, 1981; Rust and Legun, 1983). Anastomosed channels are most commonly associated with highly stable banks, often fixed by vegetation, but they can also occur in more arid settings where vegetation is less important (e.g. Rust and Legun, 1983). Sedimentation is principally by vertical accretion of both channel floor and overbank. Well developed levees are a characteristic feature and channel migration when it occurs is by avulsion (Smith, 1983). The resulting channel sand bodies are characteristically of shoestring form, bounded laterally by levee and overbank deposits (Fig. 3.29). The nature of bedforms within the channels is not well documented but may involve normal transverse bedforms superimposed on larger forms such as alternating side bars.

3.6 INTER-CHANNEL AREAS

So far we have confined our attention to the processes and products of channels and have said little about what goes on between and beyond them. The inter-channel areas, however, are normally much larger than those of the channels themselves and their deposits play an important role in the overall alluvial sequence. Unlike the channels, inter-channel deposits have a strong climatically induced overprint, making them valuable indicators of palaeoclimate. Two types of inter-channel area can be distinguished: (a) those influenced by a channel, either adjacent (levee and crevasse splay) or distant (floodplain areas) and (b) those beyond the reach of direct river influence. These inter-channel deposits are common to a range of channel types but tend to be more abundant in association with high-sinuosity or anastomosing channels.

3.6.1 **Overbank Environments**

CREVASSE SPLAYS AND LEVEES

Levees are ridges which slope away from the channel into the floodplain and are particularly well-developed on the concave erosional banks of meanders. They are submerged only at the highest floods. During lesser floods they may be the only dry ground on the floodplain. As flood water overtops the channel banks there is a fall-off in the level of turbulence and suspended sediment is deposited, the coarser sands and silts close to the channel, the finer sediments further out on to the floodplain (e.g. Hughes and Lewin, 1982). Levees not only result from the fall-out of coarse components of the suspended load but also may be partly the result of the accretion of crevasse splays. These are more localized lobes of sediment laid down on the distal side of the levee at points where the levee crest has been locally breached in a crevasse. In some cases laterally coalescing crevasse splays may play the major role in construction of levees (Fig. 3.30) (e.g. Coleman, 1969). Some larger crevasse splays extend beyond the obvious distal limits of the levee to overlie and become interbedded with floodplain deposits (cf. Sect. 6.5.1). Such beds are not well documented but they are most

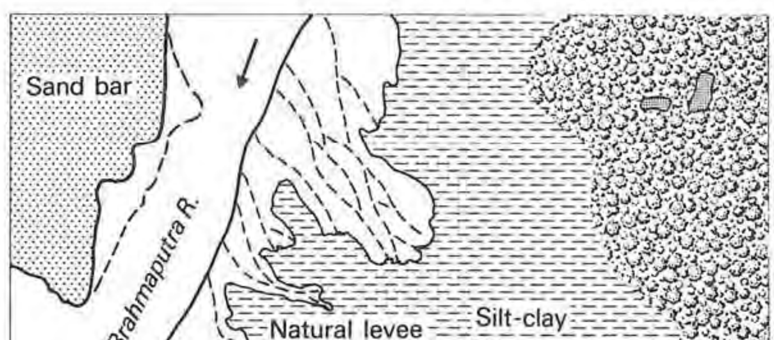

Fig. 3.30. Crevasse splays on the bank of the Brahmaputra build up a levee close to the channel (after Coleman, 1969).

likely to be sharp-based sand sheets with internal evidence of waning flow in the form of grading and a Bouma (1962) sequence of internal structures. Rapid deposition may be recorded in both levee and crevasse splay deposits by the occurrence of climbing ripple cross-lamination (e.g. Singh, 1972; Klimek, 1974b). Both are also likely to be sites of plant colonization and original sedimentary structures are likely to be disturbed or destroyed by roots. Directional structures are likely to be somewhat divergent from those derived from associated channel deposits. The more proximal, sand-rich levee deposits are of low preservation potential because of their susceptibility to erosion by channel migration.

The rate at which levees build up above the floodplain is one of the major factors determining the avulsion frequency of the channel.

FLOODPLAINS

Sedimentation and post-depositional changes on the floodplain depend on climate and on distance from the active channel. The floodplain is rarely inundated, the most common recurrence interval for overbank flooding being between one and two years (Wolman and Leopold, 1957). Overbank sedimentation rates are rather low, owing to the relatively high velocities of floodplain currents and low concentrations of suspended sediment at flood peak. Most sedimentation is from suspension and there is a tendency for deposits to fine away from the channel. Only major floods deposit more than a few centimetres of sediment and then only patchily. Vegetation may help to localize both sedimentation and erosional scour on the floodplain. Floodplain sediments dry out between floods and desiccation cracks or other features of subaerial exposure may develop.

In some humid or very low slope settings, where the river is flowing close to its base level, the floodplain may never dry out. Backswamps and lakes may form important elements in the overbank landscape as in the 'Sud' of Sudan (Rzoska, 1974) and the Atchafalaya River Basin (Coleman, 1966; Flores, 1981). Where vegetation is abundant, as in the Sud, peats will accumulate, often to considerable thicknesses. The plants serve to baffle overbank flows and cause sediment to be deposited near the swamp margins. Where vegetation is less luxuriant or where more sediment is available, thick peat accumulation is less important and a complex of lakes and swamps results. Lakes develop by compaction and subsidence of earlier organic-rich fine sediments and accumulate well laminated fine muds, often containing a fauna of non-marine bivalves. Switching of sediment distribution patterns leads to lakes being infilled by small deltas giving small-scale coarsening-upward units (see Chapter 6) and leading to the establishment of a new swamp. Swamps may be either poorly or well drained, depending on the proximity of a channel. Well-drained swamps have oxidizing near-surface conditions with less preservation of organic matter whilst poorly-drained swamps are reducing (Coleman, 1966). These differences are likely to be reflected in resultant soil profiles.

In semi-arid settings, where vegetation is less abundant, less organic matter is incorporated into the sediment and even that is likely to be oxidized. Disturbance by roots is less and the surface is more susceptible to aeolian deflation and reworking. Soil formation and reddening processes may be active though these are more extensive in areas beyond immediate river influence (Goudie, 1973). Whilst climate is a major control on floodplain development, large fluvial systems may show a discrepancy between the floodplain conditions and the hydrology of the river. The latter may be controlled by climate in a very distant catchment area and this may differ markedly from the local conditions which control floodplain processes.

Wind activity is important on some floodplains with aeolian dunes developing widely and finer material being winnowed and redeposited (e.g. Higgins, Ahmad and Brinkman, 1973). Large thicknesses of wind-blown silt may accumulate on floodplains and form the bulk of the deposits there (Lambrick, 1967). In ephemeral stream courses, aeolian dunes may block and divert channels and wind-transported and sorted sand may be reworked by subsequent stream activity.

3.6.2 **Areas beyond river influence**

These areas encompass a whole range of settings, both erosional and depositional. Here we confine our attention to those areas where alluvial sediment has accumulated and where it may accumulate again in the future.

TERRACES

In many present-day alluvial settings terraces are important elements in the landscape. They may be caused by lowering of a local or a more general base-level, by depletion of sediment or by complex responses to climatic and tectonic change (Schumm, 1977). Base-level changes may be due to eustatic, isostatic or tectonic causes. Rapid climatic and vegetation changes, operating at a very local scale have caused streams to be incised as 'arroyos' in alluvial plains in the south-western USA (Haynes, 1968; Cooke and Warren, 1973). The important processes which modify sediments on terraces are reworking by wind and *in situ* processes of soil formation, both of which are strongly influenced by climate.

WIND ACTIVITY

Wind erosion and reworking will only be significant if vegetation cover is low. It is therefore more common in arid or semi-arid settings or at high latitudes. Erosion removes finer particles leading to deflation of the terrace top. Coarse particles on the pavement may be facetted into ventifacts and may acquire a coat of desert varnish (Glennie, 1970). The material removed by the wind may accumulate locally as sand dunes or it may be transported in suspension to be laid down as loess in more distant areas (Lambrick, 1967; Higgins, Ahmad and Brinkman, 1973; Yaalon and Dan, 1974) (see Sect. 5.2.8). Wind not only deposits fine-grained clastic debris but also may introduce material such as carbonate which is important for soil formation.

SOILS

Soil formation is the second main influence in inter-fluvial areas and again climate is the most important control. In spite of their great importance, particularly in the investigation of problems of Quaternary geology, it is beyond the scope of this book to discuss soils in detail. Their study is a subject in its own right and several books are devoted exclusively to them (e.g. Bunting, 1967; Bridges, 1970; Hunt, 1972; Goudie, 1973; Birkelund, 1974; Duchaufour, 1982). Only a few points of potentially wider geological significance will be dealt with here.

Soils develop on both recently deposited sediments and poorly consolidated or weathered bedrock. Although the latter group may sometimes be important in understanding major unconformities in the stratigraphic record, the former group is our main concern. Many schemes of classification of soils have been developed, commonly relating soil type to climate. Whilst a broad climatic control undoubtedly prevails it is important to realize that, because most soil-forming processes are slow acting, soil profiles observed at the present day may show overprinting of more than one climatic regime.

Most modern soils show a vertical profile which involves several horizons, the details of which and their position in the profile are controlled by both climate and the nature of the starting material. Most modern soils have an organic-rich upper layer which grades down into more mineral-dominated layers. The extent and importance of the organic-rich layer is controlled by the prevailing rates of organic production and decay which in turn reflect climate and the position of the water table. Preservation potential of the organic layer is small except in swamps and peat bogs, but it has an important influence on the rest of the soil profile.

Rainwater passing through the organic layer becomes enriched in CO_2 and in various organic acids with a resultant decrease in pH. In addition, the roots of plants growing at the surface may extend into the lower mineral layer, causing physical disturbance and abstracting material in solution.

In the underlying mineral layers, rain water moves mainly downwards. As it is usually acidic, it removes alkali and alkali earth ions in solution and helps the breakdown of various silicate minerals to clays. Both clays and dissolved ions are transferred to lower levels in the profile where they may be deposited. In humid settings, however, the more soluble ions may be removed in solution from the soil system by transfer to the ground-water. Clays and the normally insoluble iron which is mobilized by organic complexing are deposited in lower levels of the profile. Under intense humid tropical conditions, silica may be leached from the soil giving a residual concentration of the oxides of iron and aluminium as laterites and bauxites. Such intensely leached soils commonly show cavernous and pisolitic textures. In cold humid settings limonite may cement the upper layers of the soil profile as a hardpan, precipitation being aided by organic activity.

In semi-arid and arid areas, precipitation is too low to allow the development of an important organic layer and the pH of soil waters is consequently high. This, combined with the smaller volumes of percolating water, means that solution rates are relatively low and that water, rather than escaping into the groundwater system, is retained in the soil profile and eventually lost through evaporation or transpiration. This results in precipitation of material from solution within the profile, under alkaline conditions. The depth at which material is precipitated relates to the rainfall and to the permeability of the host material. The mineral which most commonly precipitates in this way is calcite, though silica also occurs. Profiles which involve carbonate layers or nodules are called 'calcrete' or 'caliche'; those with secondary silica are called 'silcrete'. Goudie (1973) has extensively reviewed the formation of calcretes and has shown that whilst downward movement of dissolved material is the main process, it is difficult to account for the supply of calcium carbonate to the sediment surface. Possible sources are carbonate-rich loess (Reeves, 1970; Yaalon and Dan, 1974) or leaf fall and plant drip. Areas downwind of sites of gypsum precipitation have accelerated rates of caliche development, presumably due to the introduction of calcium sulphate by the wind and the precipitation of calcite by the common ion effect (Lattman and Lauffenberger, 1974). Rates of calcrete formation are variable but it seems that periods of the order of 10^3–10^5 years are involved in the development of mature caliche profiles (Gardner, 1972).

Silcretes seem to be the product of warm humid settings and occur in areas of very mature soil development (Twidale, 1983). They are often associated with deeply weathered intermediate and basic igneous rocks or with sandy substrate, but also develop rapidly where pyroclastic rocks are abundant (Flach, Nettleton *et al.* 1969). Silica is mobilized in solution and reprecipitated locally, though some may be introduced over greater distances. Precipitation may sometimes occur where upward-moving, silica-rich solutions meet downward percolating water rich in dissolved salts (Smale, 1973), particularly sodium salts (e.g. Frankel and Kent, 1938) and as such are likely to be found in association with lake sediments.

RED COLORATION

The origin of 'red beds' has been a subject of heated controversy with the main argument centring on whether the red pigment is of detrital or diagenetic origin (see Glennie, 1970, for a concise discussion). Observations of alluvial fan and tidal flat sediments of Pliocene to Recent age in Baja California show clearly that here the reddening is essentially diagenetic (T.R. Walker, 1967; Walker, Waugh and Crone, 1978). The processes seem to be quite slow acting; the Recent sediments, freshly derived from adjacent granites, are grey, the Pleistocene ones yellow and only those of Pliocene age show a good red coloration. The processes seem to involve the breakdown of biotite and hornblende to give clays and immature iron oxides and hydroxides which are washed to lower levels in the weathering profile and coat grains. With time the oxides mature to haematite. These changes will be aided by elevated temperatures but their main prerequisite is water provided by ephemeral rainfall and run-off. Rates of reddening vary with lithology, clay-rich sediments altering more slowly due to lower permeabilities.

The rock record clearly shows that some ancient red-beds were deposited in humid regimes, but no present-day red-beds are known in such settings. Instead, present-day, warm, seasonally humid source areas supply detritus which is mainly grey or brown, with iron present as hydrated oxides (Van Houten, 1973; Turner, 1980). If these are deposited in settings which are then subjected to an oxidizing ground-water regime, probably favoured by a lowered water table, then the amorphous hydroxides and oxides may mature *in situ* to give eventually a red deposit (e.g. Paijmans, Blake *et al.*, 1971).

3.7 ANCIENT ALLUVIAL SEDIMENTS

Alluvial deposits are generally recognized by an absence of marine fossils, by the presence of red coloration and channels, by broadly unidirectional palaeocurrents, particularly in the coarser sandstone or conglomerate units, and by evidence of emergence such as palaeosols and desiccation mudcracks particularly in finer grained deposits. However, none of these features is diagnostic by itself since all may occur in other environments. In Precambrian rocks in which a fauna is lacking and soils seem much less common, it may often be difficult to distinguish between fluvial and shallow marine deposits in thick sandstone formations. Even in the Phanerozoic where, for example, thick conglomerate units are associated with active tectonics, the distinction between fluvial and basinal conglomerates may not be obvious (e.g. Harms, Tackenberg *et al.*, 1981; Stow, Bishop and Mills, 1982).

Once a broadly alluvial interpretation is established, a sequence can be discussed in terms of more specific alluvial settings and attempts made to place the deposits in the continuum of river types recognized at the present day. However, comparisons with the present are seldom direct since land plants and their role in covering the land surface have changed through time. For example, pre-Devonian sequences are much less likely to have meandering river deposits than are younger deposits since there were then no land plants to stabilize overbank sediments (e.g. Cotter, 1978).

To simplify discussion, it is thought best to recognize two rather loosely defined types of alluvial sequence, one dominated by sandstones and conglomerates with little fine-grained sediment and the other dominated by sandstones and finer sediments with relatively little conglomerate. The first group may be compared with deposits of present-day bedload streams, both sandy and pebbly and commonly occurring in the wider setting of an alluvial fan. The second group finds modern analogues in suspended-load streams, commonly meandering but also including anastoming streams and in the distal parts of some terminal fans. These two broad facies associations are, like their modern counterparts, intergradational.

3.8 ANCIENT PEBBLY ALLUVIUM

Thick sequences of pebbly alluvium are generated and preserved where there is topographic relief and this commonly implies tectonic activity during or immediately prior to deposition.

Many alluvial fans, both stream-dominated and semi-arid are associated with normal faults and the development of graben, half graben and pull-apart basins (Sects 14.4.2, 14.8) (e.g. Collinson, 1972; Steel, 1974, 1976). Other successions are not so directly related to fault lines but are the fills of basins flanking recently uplifted source areas, sometimes a consequence of continental collision. These are the typical molasse deposits of many mountain belt foreland basins (Sect. 14.9.2). Where fans developed in response to normal extensional faulting, mass-flow deposits may be abundant if climate and source rocks were appropriate. Relief was maintained and the position of the fan apex either remained fairly fixed allowing a thick wedge of sediment to accumulate and be preserved on the down-thrown side of the fault (Fig. 3.31A) (e.g. Collinson, 1972) or it migrated backwards along a series of successively active listric normal faults (Fig. 3.31B) (e.g. Steel and Wilson, 1975). In strike-slip, pull-apart basins, the position of the active fault may migrate through time either laterally along the marginal strike-slip fault or backwards towards the end of the basin. In the latter case, a series of proximal-distal wedges are stacked in an imbricated style along the length of the basin resulting in huge aggregate thicknesses (Sect. 14.8.2) (Steel and Gloppen, 1980).

In foreland molasse basins, the deposits are commonly very widespread and tend to be dominated by stream deposits as a result of larger catchment areas and lower slopes. Sediments may be laid down during active tectonism and be overthrust and deformed soon after deposition as the foreland-thrust belt overrides and disrupts the foreland basin (Sect. 14.9.2).

3.8.1 Facies

Whilst conglomerates may make up a large proportion of

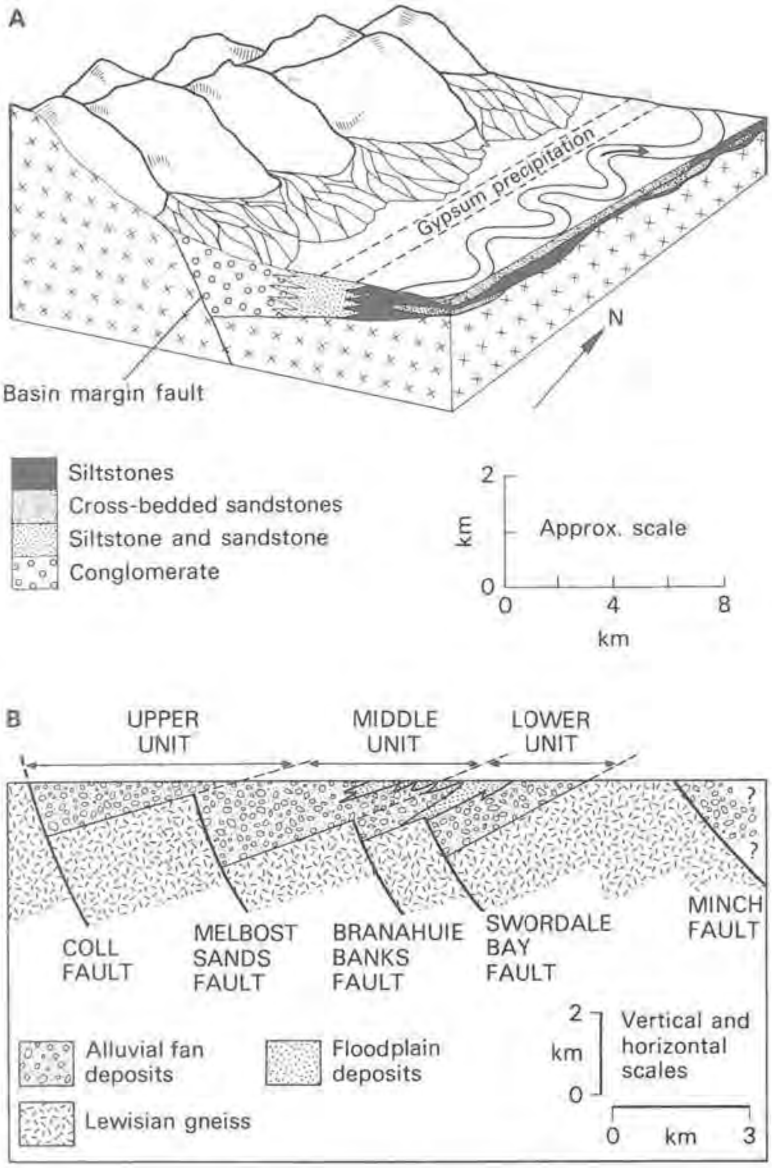

Fig. 3.31. (A) Postulated environmental setting for the deposition of the Røde Ø Conglomerate and associated sediments, East Greenland, controlled by one active normal fault (after Collinson, 1972). (B) Schematic distribution of conglomerate wedges associated with a succession of backward stepping normal faults, Permo-Triassic of the Outer Hebrides (after Steel and Wilson, 1975).

ancient coarse alluvial successions, it is unusual for sandstones and siltstones to be entirely absent and most sequences show a range of facies. The conglomerates are commonly divisible into facies on the basis of texture and style of stratification. The simplest subdivision distinguishes unstratified paraconglomerates with a matrix-supported texture from commonly stratified orthoconglomerates where the texture is a framework of larger clasts usually infilled by finer matrix. The stratified conglomerates grade through stratified pebbly sandstones into variously bedded sandstones (Fig. 3.32). Several schemes of more detailed facies discrimination have been applied to specific sequences by

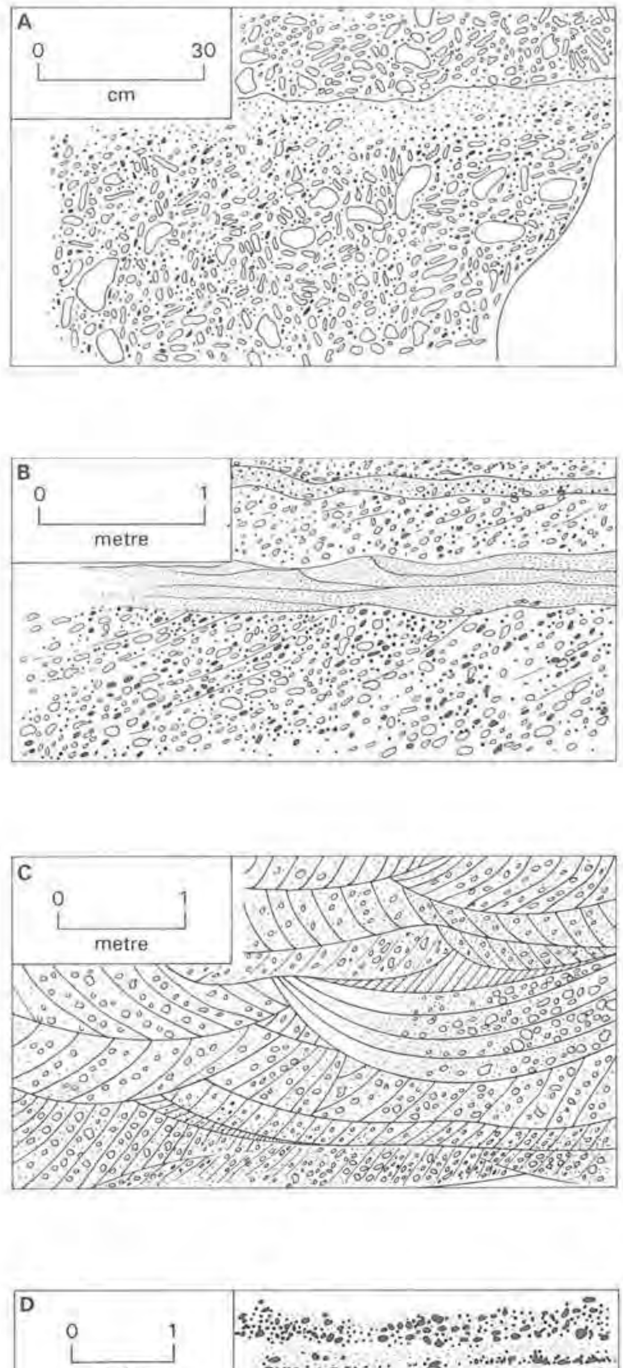

Fig. 3.32. Four conglomeratic facies recognized in Old Red Sandstone alluvial fan sediments, Firth of Clyde, Scotland. (A) is a matrix-supported paraconglomerate and is attributed to mudflow deposition. (B), (C) and (D) are all orthoconglomerates. (B) is characterized by interbedded fine units and is interpreted as sheetflood; (C) and (D) represent points in a spectrum of variation and are stream channel deposits (after Bluck, 1967).

individual authors (e.g. Bluck, 1967; McGowen and Groat, 1971; Steel, 1974; Steel and Thompson, 1983). A scheme which seems widely applicable and which is becoming increasingly used as a convenient shorthand is that of Miall (1977). This recognizes three major grain-size classes, gravel, sand and fines (G, S, F) and several modes of bedding (e.g.: m, massive; t, trough cross-bedded; p, planar (i.e. tabular) cross-bedded; r, rippled; h, horizontal laminated; s, shallow scours, etc.) which may be combined in various ways (e.g. St = trough cross-bedded sandstone). Such pigeon-holing is convenient when some form of numerical analysis of data is planned (e.g. Markhov chain analysis) but it can lead to a somewhat uncritical approach to primary observation. The scheme is a useful starting point but one which must be adapted and extended to the needs of a particular sequence (e.g. Massari, 1983). Other workers have incorporated aspects of bed shape and lateral variation directly into their schemes of facies classification (e.g. Ramos and Sopeña, 1983). These authors distinguish sheet-like units from those filling channels and they also recognize the existence of low-angle lateral accretion surfaces within conglomeratic sequences.

The various facies schemes mainly allow interpretation in terms of process. The matrix-supported conglomerates (paraconglomerates) which commonly lack internal structure or even clast imbrication are generally ascribed to high viscosity mass-flow processes. Correlation of bed thickness with maximum clast size is thought to reflect the positive relationship between competence and size of flow (Bluck, 1967; Larsen and Steel, 1978). Some units of this type are overlain by thin beds of sandstone showing parallel and low angle bedding, interpreted as the product of waning flood stages (Steel, 1974).

Conglomerates with clast-supported frameworks and pebbly sandstones with clear stratification are the result of bedload deposition. Horizontally stratified or unstratified conglomerates with clast imbrication record deposition on a flat bed with vigorous grain transport such as might occur on the top of a longitudinal bar or on a channel floor and textural variations may relate to water stage changes (Fig. 3.33) (Steel and Thompson, 1983). Laming (1966), in a pioneering study of the Permian conglomerates of south-west England, recognized two types of imbricated and parallel-bedded conglomerate. One had 'clast imbrication' where clast/matrix ratio is high and clasts are tightly packed in a framework and the other had 'isolate imbrication' where similarly inclined clasts floated in a more abundant matrix. Where such conglomerates have a sheet-like form and are interbedded with finer sediment (sand or silt) they probably result from sheet-floods. This interpretation would be supported by a correlation between bed thickness and clast size (e.g. Bluck, 1967).

Other conglomerates show a more lenticular or channelized geometry with cross-bedded fills of scours, close-packed lag conglomerates on scour bases and fine drapes both to bed tops and to the floors of scours. Sections parallel to the palaeocurrent

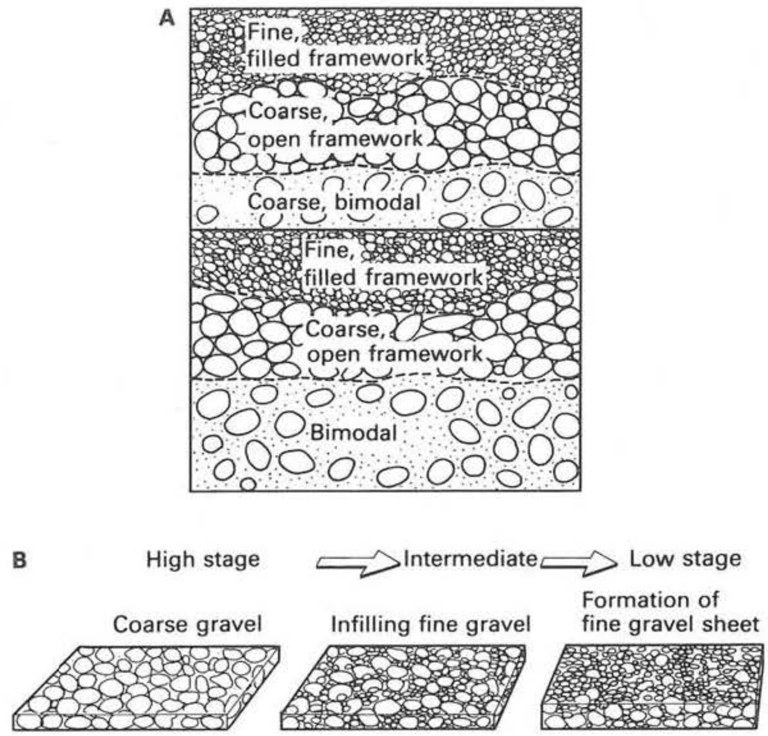

Fig. 3.33. (A) Variable textures in bedded conglomerates from the Triassic of Cheshire, England. The variation is interpreted (B) in terms of varying transport populations during waning water stage (after Steel and Thompson, 1983).

show individual cross-bedded sets which change laterally in grain-size from framework gravel to pebbly sandstone and back again, presumably reflecting discharge fluctuations in the flow over the bedform's (bar's) slip face. Sequences with lenticular bedding and channels do not show any correlation between clast size and bed thickness and record a more sustained though probably variable flow such as might be found in a river rather than a series of episodic sheet flood events.

In gravels or conglomerates where channel forms are suspected but are not readily apparent, it may be possible to detect them from a detailed study of clast imbrication. In the Carboniferous of the Intra-Sudetic Basin, Teisseyre (1975) showed that pebbles have systematic changes in transverse profile across channels. In the axial area of a channel, clasts dip directly upstream whilst near the margins they dip with a component towards the centre line. Such an approach could be of great value in the economic exploitation of alluvial heavy minerals such as gold which can occur as intergranular fines in coarse orthoconglomerates. These tend to be most common in channels and in sites of active reworking on the more proximal parts of a fan (e.g. Sestini, 1973).

3.8.2 Lateral facies distributions

Within wedges of alluvial conglomerates both facies and thickness change laterally. In isolated fans, changes take place radially from the fan apex whilst in laterally extensive wedges against faults, changes will be broadly away from the fault line. The rates of change depend on the scale of the system and the nature of the dominant processes. Along with a general proximal to distal thinning there is a diminution in the thickness of individual beds and in the size of the largest clasts (Bluck, 1967; Nilsen, 1969; Miall, 1970) as is seen on present-day fans and outwash areas (Bluck, 1964; Boothroyd, 1972). In deposits of some small Triassic fans in South Wales, maximum clast size falls off exponentially with distance (Bluck, 1964). On semi-arid fans there is commonly a distal decline in mudflow deposits and a proportionate increase in streamflood and channel deposits (Bluck, 1967; Nilsen, 1969; Steel, 1974; Steel, Nicholson and Kalander, 1975). However, mudflows may occur interbedded with distal playa sediments laid down beyond the normal range of fan processes (Fig. 3.35) (Bluck, 1967). The large travel distances of modern mudflows make this quite possible (e.g. Sharp and Nobles, 1953). Mudflow conglomerates may be interbedded distally with flood basin fines. Through loading and infiltration these become intimately mixed so that textural inversion takes place with a distal deterioration in sorting (Larsen and Steel, 1978). Resedimentation as granule sandstone beds may occur to give inversely and normally graded beds.

In areas of exceptional exposure it may be possible to demonstrate lateral changes in the deposits of much larger alluvial fan systems usually dominated by stream processes (i.e. 'humid' fans). The Van Horn Sandstone of Texas which can be traced laterally for tens of kilometres has been interpreted in such a way (McGowen and Groat, 1971) (Fig. 3.34). The proximal facies is characterized by thick, framework conglomerates in units which have flat bases and convex upwards tops in sections transverse to the palaeocurrent. The units are elongated parallel to the current and are flanked by cross-

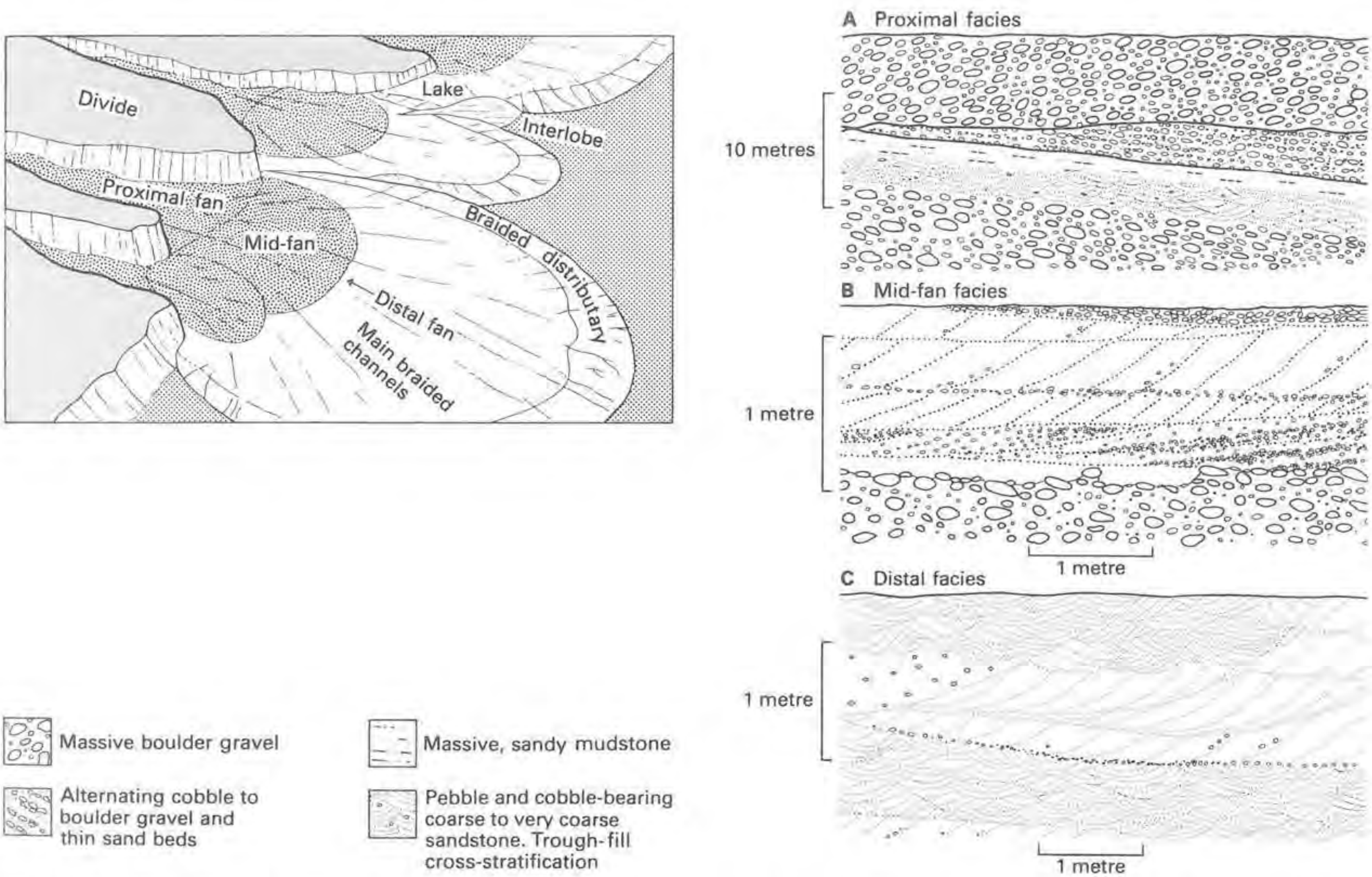

Fig. 3.34. Distribution of facies and environments in the stream dominated fan deposits of the Van Horn Sandstone, probable Precambrian age, Texas. Conglomerates of the proximal fan area have boulders up to 1 m diameter. In the midfan, conglomerates are interbedded with pebbly sandstones and in the distal fan, tabular and trough cross-bedded sandstones dominate. Proximal to distal facies changes are gradational over a distance of 30–40 km (after McGowen and Groat, 1971).

bedded pebbly sandstone, a relationship similar to that observed in Pleistocene outwash gravels by Eynon and Walker (1974). These proximal facies record the dominance of longitudinal gravel bars. Inter-bar channels are filled with finer sediment at low stages or as they are gradually abandoned. In the mid-fan area there is less gravel and more bed bases are scoured, though some of the elongated, convex-upwards bodies persist. Parallel-bedded gravels are interpreted as longitudinal bars which grew mainly by vertical accretion on the bar-top whilst flanking sandstones show cross-bedding and are interpreted as transverse bars which migrated down inter-bar channels (cf. N.D. Smith, 1974). In the distal areas gravels are confined to thin beds and lenses scattered in tabular and trough cross-bedded sandstones as in modern outwash fans (Fig. 3.16) (cf. Boothroyd, 1972).

In other systems lateral changes are recorded in directions normal to the general palaeocurrent. In the Messinian molasse of northern Italy, sequences dominated by vertically stacked conglomerates up to 15 m thick and with relatively few fines occur as lateral equivalents to tabular cross-bedded conglomerate beds up to 10 m thick separated by thick units of fine-grained sediments (Massari, 1983). The change takes place along depositional strike and suggests that a fan dominated by braided channels passed laterally into an interlobe area where more channelized, possibly sinuous streams were confined within muddy cohesive banks.

3.8.3 Vertical facies sequences

Small-scale vertical facies sequences in alluvial conglomerates tend to have a strong random element. Where episodic sedimentation is in the form of debris flows and sheetfloods, bed thickness and clast size vary randomly. The deposits of major depositional episodes are commonly separated by fines deposited by waning floods, but in many cases these are scarce owing to their removal by subsequent scour or to an inherent lack of fines in the sediment supply. When only a small proportion of sand and finer sediment is available, most or all of it is absorbed into the spaces of the gravel framework rather than giving fine interbeds.

Where stream processes are dominant, scour surfaces divide

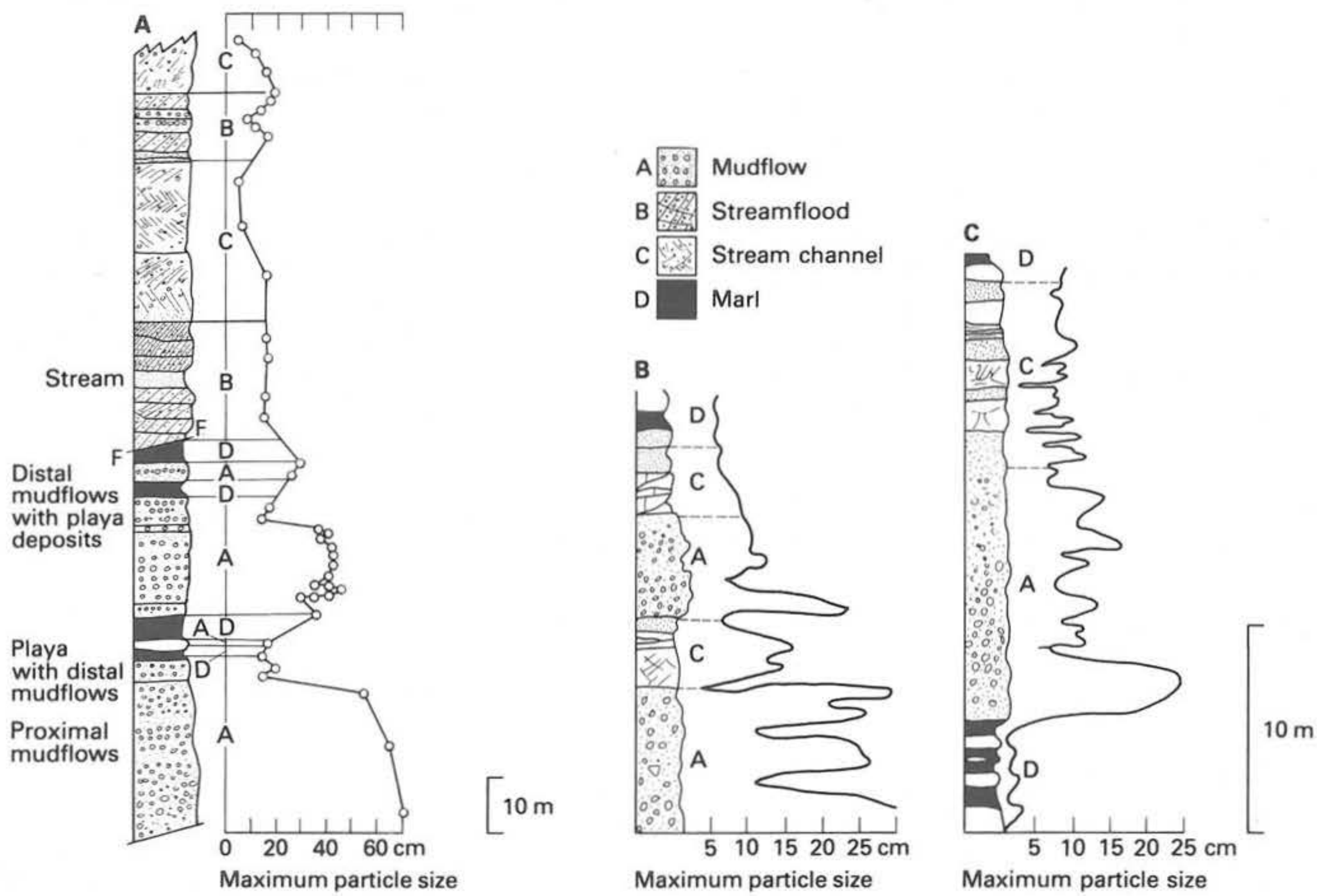

Fig. 3.35. Typical vertical sections through the deposits of semi-arid alluvial fans from (A) Old Red Sandstone of Scotland and (B and C) the New Red Sandstone (Permo-Triassic) of the Hebrides, Scotland. Variations in maximum particle size define coarsening- and fining-upward units and also show the high random component in the interbedding (after Bluck, 1967, (A) and Steel, 1974 (B, C)).

the sequence into units within which both lateral and vertical facies changes may be apparent. Both sheet-like units and more obvious channel forms are preserved (e.g. Ramos and Sopeña, 1983). Within the larger units of both sheet and channel type, more closely spaced minor scour surfaces reflect stage fluctuations. Elements of both upward fining and upward coarsening occur. The fining may be related to either channel abandonment or to lateral accretion. Additional evidence for the latter is seen in low-angle lateral accretion cross-bedding (Ramos and Sopeña, 1983). Upward coarsening in units a few metres thick in Pleistocene outwash gravels has been described by Costello and Walker (1972) and attributed to the gradual reactivation of a channel after an interval of temporary abandonment in the braided complex. Coarsening upward is also a pattern associated with the advance of a major mid-channel or lateral bar where the coarser grained bar head advances over the finer grained bar tail. The cross-bedding in the bar tail deposits may show greater variability than that in the head owing to the convergence of flows in the bar head area (Bluck, 1980). Upward coarsening at the scale of an individual bar may take place within a thicker channel unit which itself is broadly upward fining. Sequences of either type tend not to be preserved in their entirety and are commonly truncated by erosion.

At the larger scale, coarsening- and fining-upward trends in pebbly alluvium occur at the scale of tens or hundreds of metres. These are most commonly recognized on the basis of maximum clast size (Fig. 3.35) (e.g. Steel, 1974, 1976; Heward, 1978; Larsen and Steel, 1978). The grain-size changes may be accompanied by changes in the relative abundance of constituent facies. For example, upward fining may coincide with a change from mudflow dominance to channel dominance whilst a coarsening-upward sequence may show the reverse facies trend.

Changes at this scale are commonly attributed to tectonic causes. Coarsening- and fining-upward sequences may result respectively from tectonic uplift of the source area or from its subsequent wearing down during a quiescent phase (e.g. Heward, 1978; Larsen and Steel, 1978). In addition to tectonic controls, switching of the sediment distribution pattern on a fan, the establishment and decay of fan lobes and fan entrenchment all contribute to the sequence found at one point. Changes in climate and vegetation also lead to facies and grain-size changes, adding further to the complexity of interpretation (e.g. Croft, 1962; Lustig, 1965).

3.8.4 Palaeocurrents

Palaeocurrents from conglomeratic alluvium can be valuable in building up a picture of the syn-sedimentary topography and also in determining the channel type in the case of channel flow. Most ancient fan deposits show fairly tightly grouped palaeocurrents (e.g. Nilsen, 1968; Collinson, 1972) although palaeocurrent readings are difficult to obtain in the deposits of fans dominated by debris flows. It is sometimes possible to identify individual fans by the establishment of a radial pattern of palaeocurrents over an area (e.g. Bluck, 1965). In a large Torridonian fan of N.W. Scotland, dominated by stream processes, an upward decrease in the dispersion of palaeocurrent directions through the sequence was interpreted in terms of fan-head retreat (Williams, 1969). A sudden major change of direction at a particular level in a sequence might indicate that an adjacent fan took over deposition at that point. This would suggest a bajada made up of laterally interfering fans.

3.9 ANCIENT SANDY FLUVIAL SYSTEMS

3.9.1 Introduction

In most sandy fluvial sequences, two major facies associations can be distinguished. These are the '*coarse*' and '*fine*' *members* (Allen, 1965a) usually interpreted as channel deposits, commonly generated by lateral accretion and interchannel or overbank deposits dominated by vertical accretion. Precise demarcation of these facies associations may be difficult, for example, where a coarse member grades vertically into a fine member through a transitional facies such as might have formed close to the channel (e.g. levee) or where a channel has been abandoned and filled by fine material.

The coarse members, with their more appealing assemblage of sedimentary structures received great attention in the early days of research and their variety is now well-documented. The fine member sediments, in contrast, have only come to prominence more recently as it was realized that their more subtle and less easily studied variations can offer great insights to palaeoclimate and into the large scale geomorphology of ancient alluvial plains.

3.9.2 Fine member deposits

Fine member deposits can generally be separated into three primary depositional facies, all of which may be modified to a greater or lesser degree by post-depositional *in situ* processes of early diagenesis, pedogenesis and bioturbation.

SILTSTONES AND MUDSTONES

These are the most abundant facies and are generally horizontally laminated with a variable degree of fissility. The better laminated examples are often highly micaceous and rich in organic matter when the sediments are unoxidized. Some may carry a fauna of non-marine bivalves and ostracods (e.g. Scott, 1978; Gersib and McCabe, 1981) whilst others are homogenized by bioturbation and may show evidence of subaerial emergence in the form of mudcracks, rain pits and footprints (e.g.

Thompson, 1970). In some cases, the fine member units show small-scale coarsening upward units which pass from muds into siltstone or even fine sandstone. The upper parts of such units are likely to show evidence of the development of soils or of plant colonization. The sediments may be subdivided on the basis of colour into red and grey beds, a distinction which is usually also apparent in coarser grained interbeds. The fine members usually show a more intense development of red pigment compared with associated sands whilst the fines in grey sequences are darker than their associated sands due to a high organic content. Red units are more likely to show evidence of emergence whilst grey beds are more likely to be disturbed by plant rootlets and to be associated with coals. However, rootlet bioturbation is sometimes found in red beds, often associated with colour mottling.

The facies record the deposition of fine material from suspension. Such deposition is most common in inter-channel areas which receive sediment during floods. The areas may be floodplains which are predominantly subaerially exposed or may be perennial swamps or shallow lakes. Small-scale upward coarsening units probably record the infilling of shallow floodplain lakes by small deltas (Scott, 1978; Gersib and McCabe; 1981; Flores, 1981; cf. Coleman, 1966). Deposits of subaerially exposed floodplains are more likely to develop a red coloration whilst swamps and lakes are more likely to preserve organic matter and lead to a grey sequence. Colour can therefore offer a guide to the position of the normal water table during deposition. However, a post-depositional fall of water table could lead to draining of swamps and the reddening of deposits laid down in a permanent water body (Besly and Turner, 1983).

Some sequences are dominated by fine member siltstones and mudstones with only scattered thin sandstones. These present a problem in that it may be difficult to decide whether they are overbank sediments laid down in an area which seldom, if ever, was the site of channel activity or were laid down beyond the range of channel activity as on the most distal parts of a terminal alluvial fan or the central parts of an ephemeral lake.

It tends to be generally assumed that all fine member siltstones and mudstones were laid down by water. However, it is always possible that some may be wind lain as loess, particularly when structureless, homogenous and mainly red silts are abundant (cf. Lambrick, 1967) (Sect. 5.2.8).

SHARP-SIDED SANDSTONE BEDS

These commonly occur interbedded with siltstones and mudstones. They are usually thin, seldom more than a few tens of centimetres and exceptionally 1 m thick. They seem to commonly wedge out laterally with convex upwards tops (e.g. Leeder, 1974) though other examples are more nearly parallel sided (Tunbridge, 1981). They have many features in common with turbidites with sharp bases, solemarks, graded bedding and a Bouma (1962) sequence of internal structures with parallel lamination and ripple-drift cross-lamination particularly common (Fig. 3.36) (Steel and Aasheim 1978; Tunbridge, 1981). This similarity, which simply reflects the episodic and decelerating nature of the flows responsible, led to some discussion about the depths of deposition of 'flysch' when such beds in the Tertiary of the Pyrenees were found to have salt pseudomorphs and the casts of birds' footprints on their bases (Mangin, 1962; De Raaf, 1964). The general context and the occurrence of palaeosols and other evidence of emergence in the sequence suggests that in this case the decelerating currents were flood events. Where the sandstones are somewhat restricted laterally they are probably crevasse splay deposits, and where closely spaced they may represent a levee (Allen, 1964; Leeder, 1974). Where they are more extensive and parallel sided and removed from any obvious nearby channel sandstones, they are probably the deposits of sheet floods on the distal parts of a fan (Steel and Aasheim, 1978; Tunbridge, 1981; Hubert and Hyde, 1982) or of major sheet floods swamping the entire alluvial system (cf. McKee, Crosby and Berryhill, 1967).

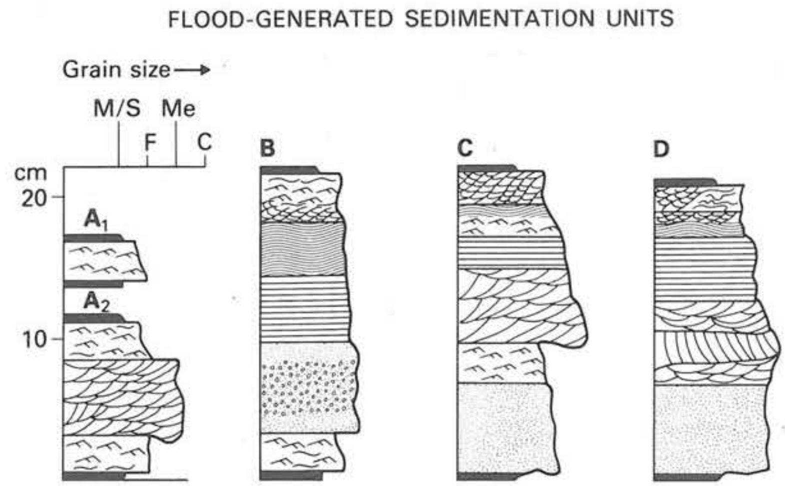

Fig. 3.36. Examples of sharp-based sandstone beds in a fine-grained alluvial succession. Such beds result from catastrophic overbank flows on floodplains or from sheet floods on the more distal areas of terminal alluvial fans (after Steel and Aasheim, 1978).

CROSS-BEDDED SANDSTONES

In some red, fine member sequences there are cross-bedded sandstones where grain size is more appropriate to the coarse member but which are not associated with basal erosion surfaces. Such units may involve sets up to several metres thick. The sand is commonly well-sorted, lacking platy minerals and the grains may be well rounded. Such sandstones are interpreted as the products of aeolian dunes which migrated in the interchannel areas. The interpretation is strengthened where the cross-bedding directions diverge widely from those in associated sands of clearly waterlain origin (e.g. Laming, 1966). Further discussion of aeolian dunes is given in Chapter 5.

In other examples, thin units of cross-bedded sandstone with rounded and polished grains occur in inter-channel siltstones of reddened alluvial sequences (Thompson, 1970). The nature of the grains suggest aeolian activity but with thin beds, only one set thick, it is difficult to know if the sands are the products of aeolian deposition or of aqueous reworking of aeolian sands on the floodplain.

PALAEOSOLS AND ASSOCIATED DEPOSITS

Inter-channel areas and areas of fans which are starved of coarse sediment supply for long periods may be sites of pedogenesis. In spite of the vast and detailed knowledge of the structures, mineralogy and textures of modern soils, only a fraction of these features are preservable in the rock record because of the diagenetic alteration and compaction which they suffer on burial (Roeschmann, 1971). In Quaternary and Tertiary deposits, where recognition and interpretation of soils are of crucial stratigraphical importance, many of the original features are preserved and quite sophisticated interpretation is possible (e.g. Yaalon, 1971; Buurman and Jongmans, 1975; Buurman, 1980; Watts, 1980; Retallack, 1983). In Mesozoic and older deposits, soil features are at present only recognized in a relatively crude way, but more subtle features are now being recognized and compared with modern soils.

(a) *Pedogenic concretions* are commonly developed in soils as calcite, siderite and quartz. Calcite is particularly common as a nodular mineral in red sequences. It occurs as isolated and coalescing sheets or layers, sometimes with a horizontal lamellar structure (Fig. 3.37). Carbonate-rich units range in thickness from a few centimetres to 2–3 m and are usually laterally continuous (Burgess, 1961; Allen, 1974a; Leeder, 1975). In some cases, a zone of isolated nodules pass up into a more continuous framework of calcite. More continuous calcite layers may be buckled into gentle folds and on bedding planes the more isolated nodules may follow a polygonal pattern.

The carbonate-rich layers compare with *caliche* soils (*calcretes*) of modern semi-arid regions (Allen, 1974a; Leeder, 1975). The pattern of vertical pipes and sheets seen in more isolated nodules may compare with modern rhizoliths (e.g. Klappa, 1980) and polygonal patterns and folding result from release of pressure due to the build-up of the carbonate layer.

As well as the insight which they offer into prevailing climate, calcrete profiles provide a useful guide to estimating the accretion rates of floodplains and to establishing at least local stratigraphy in unfossiliferous sequences. Mature profiles with lamellar upper parts (Fig. 3.37) require about 10,000 years to develop whilst less mature examples presently reflect shorter periods. Their occurrence therefore implies that for periods of this order of duration little or no deposition took place and that the water table was deep and the alluvial plain well drained

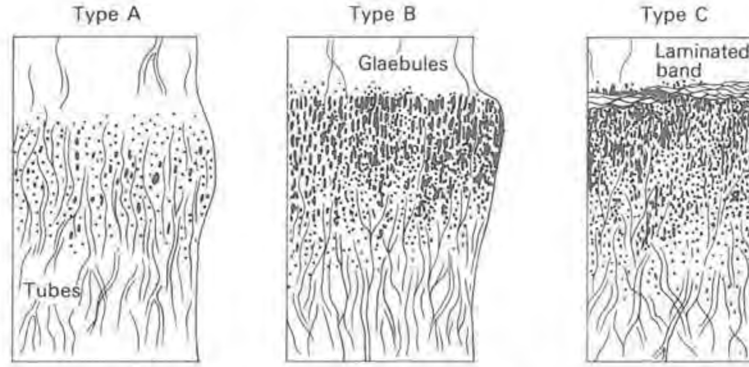

Fig. 3.37. Three intergradational stages in the development of a vertical profile in pedogenic carbonate units (caliche/calcrete) of the Old Red Sandstone of Wales (after Allen, 1974a).

(Leeder, 1975). Such sediment starvation and drainage may have resulted from migration of the river to a distant position but for the most mature soils a period of incision and terrace development seem more likely (cf. Tandon and Narayan, 1981).

Not all carbonate layers within fine member units are necessarily fossil caliches, however. Many micritic limestones formed in lakes (Sect. 4.6.3). In the Old Red Sandstone of Spitzbergen, laminated carbonates have ostracods and algal remains and are interpreted as lacustrine (Friend and Moody-Stuart, 1970). Other carbonate units might result from nearby springs of carbonate-rich water (Freytet, 1973; Ordóñez and García del Cura, 1983).

In grey fine members nodules are most commonly of siderite and tend particularly to be associated with rootlet-bearing beds below coal seams (Wilson, 1965; Retallack, 1976, 1977). Such nodules are commonly elongated normal to the bedding and are frequently associated with carbonaceous rootlets and plant fragments. They are thought to be precipitated from slightly reducing ground waters in a permanently saturated soil. The nodules may in some cases be pseudomorphing major root systems of plants growing at the surface of the developing soil. Less common than the usual grey seat-earths are brown ones which contain nodules of sphaerosiderite. These underlie thin coal seams and are thought to represent less saturated and partially emergent soils.

Syndepositional silica cements occur where the host sediment is sand. In red beds, sandstones, often of aeolian origin, show syntaxial overgrowths of quartz or, more rarely, orthoclase. This diagenesis is thought to be due to the precipitation of silica derived from the solution of quartz dust by alkaline ground water which is not a strictly pedogenic process. The formation and evaporation of desert dew may play an important role in giving an initial case hardening. In the Proterozoic of North Greenland siliceous nodules are associated with minor unconformities and compare with modern silcretes associated with alkaline waters (Collinson, 1983).

In grey beds, silica cementation results in quartz-rich *seat-*

earths or *ganisters*. Here, the silica concentration is thought to be due to leaching in a water-logged soil, with some contribution of silica to the matrix and cement by vegetation (Retallack 1976).

(b) *Mottling and clay coatings*. The migration in solution of ions of manganese and iron leads to the patchy accumulation of their oxides and hydroxides as grain coatings giving an overall mottled appearance to the sediment. Such mottling is a characteristic of gley soils where the movement of reducing pore waters is sluggish (Buurman, 1975).

At the microscopic level, clay coatings on sand-sized particles are found in rocks as old as Palaeozoic (Teruggi and Andreis, 1971). These are interpreted as products of soil formation by illuviation in the B horizon of clays transported downwards by water. The coatings therefore reflect pore-water conditions different from those associated with mottling. However, interpretation in detail is very complex, not least because of the overprinting of soil features within the same unit (Besly and Turner, 1983). Changing groundwater conditions may produce different features which are then preserved together, some recording the later stages of soil development and all subject to later diagenetic alteration on burial.

(c) *Rootlets and coal*. The penetration and disturbance of fine member sediments by rootlets is diagnostic of ancient soils. The sediments which may be mudstone, siltstone or sandstone, are usually grey but red mottled and variegated sediments are also sometimes found to have rootlets on close inspection (Besly and Turner, 1983). Rootlets penetrate the bedding at all angles and in Carboniferous Coal Measures are usually preserved as thin carbon films (Huddle and Patterson, 1961; Wilson, 1965). Major roots such as *Stigmaria* occur with smaller rootlets attached and the larger forms are in some cases preserved in full, uncrushed relief, usually with a sand infill. Rarely, larger roots are preserved in three dimensions in a coalified state (Baird and Woodland, 1982). *Stigmaria* usually lie in a horizontal plane reflecting the shallow rooting system of the plants. Other roots have vertical or sub-vertical orientations and they may be preserved by sideritic or calcite concretions (e.g. Klappa, 1980). An upward fining of host grain size is commonly observed in seatearths. This may reflect the decreasing energy caused by increasing plant colonization (Wilson, 1965), or may be due to chemical weathering within the soil profile. The presence of abundant rootlets may lead to the total destruction of original bedding and lamination and intense slickensiding can result from the collapse and compaction of roots (Huddle and Patterson, 1961).

Rootlet horizons are often overlain by coal seams of variable thicknesses which bear no relationship to the thickness of the underlying seatearth. Coals record prolific plant growth and preservation of the organic matter as a peat by acidic and reducing ground water and a high water table. Whilst the earliest plants rooted in the underlying sediment, later plants rooted in the accumulating peat mat. The coal seam can thus be regarded as a separate soil in its own right. In some red-bed successions where the reddening is of early diagenetic origin, seatearths are preserved as mottled and variegated units with poorly preserved root traces whilst any once-present coal seams are totally oxidized (Besly and Turner, 1983).

(d) *Clay mineralogy*. The chemical reactivity of clay minerals makes them sensitive to pedogenic processes. By the same token, however, they are susceptible to diagenetic alteration and more subtle properties have a low preservation potential. Watts (1976) recognized palygorskite in a Permo-Triassic sequence as the product of a semi-arid soil. The most clearly developed trend is towards the concentration of kaolinite in many fine-member seatearths at the expense of illite (Huddle and Patterson, 1961). This concentration, which is often accompanied by titanium enrichment, is interpreted as the result of *in situ* leaching in soil below growing vegetation. The mineralogy of seatearths and other soils not only reflects pedogenesis but is also due to the lithologies and weathering conditions of the source area.

(e) *Soil associations*. Soil features which may be preserved in palaeosols occur in combinations and sequences which show very subtle variation. Variation is due not only to the infinite range of combinations of host mineralogy, plant colonization, ground water chemistry, climate and topography, but also to the possibility of over-printing as conditions change through time. The red-bed caliche assemblage and the grey-bed seatearth coal assemblage are two end members of a spectrum of variation where intermediate stages are only just beginning to be explored (e.g. Retallack, 1976, 1977, 1983; Buurman, 1980; Bown and Kraus, 1981).

3.9.3 **Coarse member (channel) deposits**

Five principal facies can be distinguished.

CONGLOMERATES

These generally occur as thin beds, only a few clasts thick. When they are associated with major erosion surfaces they are interpreted as channel lags. However, when the bulk of the coarse member is pebbly sandstone, pebble concentrations are more widespread and may overlie more local scours such as the bases of trough cross-bedded sets. In the more extensive channel lag conglomerates, the clasts may be either extra-basinal or intra-basinal, derived from the erosion of inter-channel, commonly floodplain deposits. The most common clasts are mud-flakes, calcium carbonate or siderite concretions or large plant fragments and logs, depending on the nature of the inter-channel environment. Where a sequence is dominated by coarse members, to the virtual exclusion of fine member units, intraformational clasts may provide the only indication of the presence of inter-channel deposits.

In addition to pebble lags, major channel erosion surfaces sometimes have associated coarse breccias made up of angular blocks of material which fell or slid into the channel with minimal current reworking (e.g. Laury, 1971; Young, 1976). Some blocks show rotation due to sliding on a curved shear plane (e.g. Gersib and McCabe, 1981).

CROSS-BEDDED SANDSTONES

These are by far the most abundant facies within coarse members and in some cases they may be the only facies present. They encompass a wide range of grain sizes from pebbly sandstone down to fine sand and cross-bedding may be of several types and scales. Trough cross-bedding is the most abundant type and occurs in sets up to 3 m thick, though sets in the order of a few tens of centimetres thick are most common. Trough cross-bedding is attributed to the migration of dunes with sinuous crest-lines or sandwaves with a more three-dimensional, commonly linguoid form. Tabular cross-bedding is less common. Set thicknesses are commonly of the order of 1 m or less, but range up to 40 m. Tabular sets result from the migration of straight-crested megaripples, sandwaves, transverse bars and, less commonly, scroll bars and chute bars. The likely origin of any particular example is most commonly deduced from its context within the coarse member and not from any internal characteristics. Both types of cross-bedding may be deformed both by overturning of foresets (e.g. Beuf, Biju-Duval *et al.*, 1971; Banks, Edwards *et al.*, 1971; Hendry and Stauffer, 1977) and by more general convolution and water escape. Internal evidence of water stage fluctuation in the form of reactivation surfaces (Sect. 3.2.2) and mud-draped foresets also occur. Exceptionally, the down-stream faces of bars weather out in full relief showing gullying and terracing (Banks, 1973c) or the superimposition of ripples.

CROSS-LAMINATED SANDSTONES

These commonly form quite substantial parts of coarse members and may occasionally make up the whole member. They usually occur towards the top of a coarse member and involve finer grained and more micaceous or carbonaceous sand than that which forms underlying cross-bedding. The cross-lamination may show ripple-drift. The facies records the migration of small-scale ripples. Where ripple-drift occurs, it demonstrates a high rate of vertical bed accretion. Thick, coarse member units dominated by ripple cross-lamination may record partial abandonment of the channel by chute cut-off (Gersib and McCabe, 1981).

PARALLEL-LAMINATED SANDSTONES

These are normally of minor volumetric importance but in some cases form substantial thicknesses of coarse members. They are usually fine-grained with primary current (parting) lineation on bedding planes. They may occur at all levels in a coarse member but are most common towards the top. Where the sand is medium- to fine-grained and mica-free it may be interpreted as the product of upper phase plane bed transport which develops below flows of high velocity and low depth. Where the sand is highly micaceous, the deposit may result from slower currents as the presence of platy grains inhibits the formation of ripples (Manz, 1978).

LATERAL ACCRETION (EPSILON) CROSS-BEDDING

An important and increasingly recognized feature of coarse members is low-angle cross-bedding which extends as a single set over the whole thickness of a coarse member unit or over a substantial part of it. This 'epsilon' cross-bedding (Allen, 1963) dips at right angles to the palaeocurrent direction derived from smaller-scale structures within it, such as ripple cross-lamination and small-scale cross-bedding (Fig. 3.38). The inclined layers are defined by fluctuations in grain size, generally in the sandstone to siltstone range and there is commonly a tendency for there to be an overall upward fining so that sandy beds taper out upwards into silts, whilst the silts taper out downwards into sands (e.g. Nami and Leeder, 1978; Puigdefabregas and Van Vliet, 1978; Stewart, 1983). Where the epsilon cross-bedding does not occupy the full thickness of the coarse member, it tends to occur in the upper part above a unit of more normal cross-bedded sand (Fig. 3.39) (e.g. Puigdefabregas and Van Vliet, 1978). Coarse members with epsilon cross-bedding range

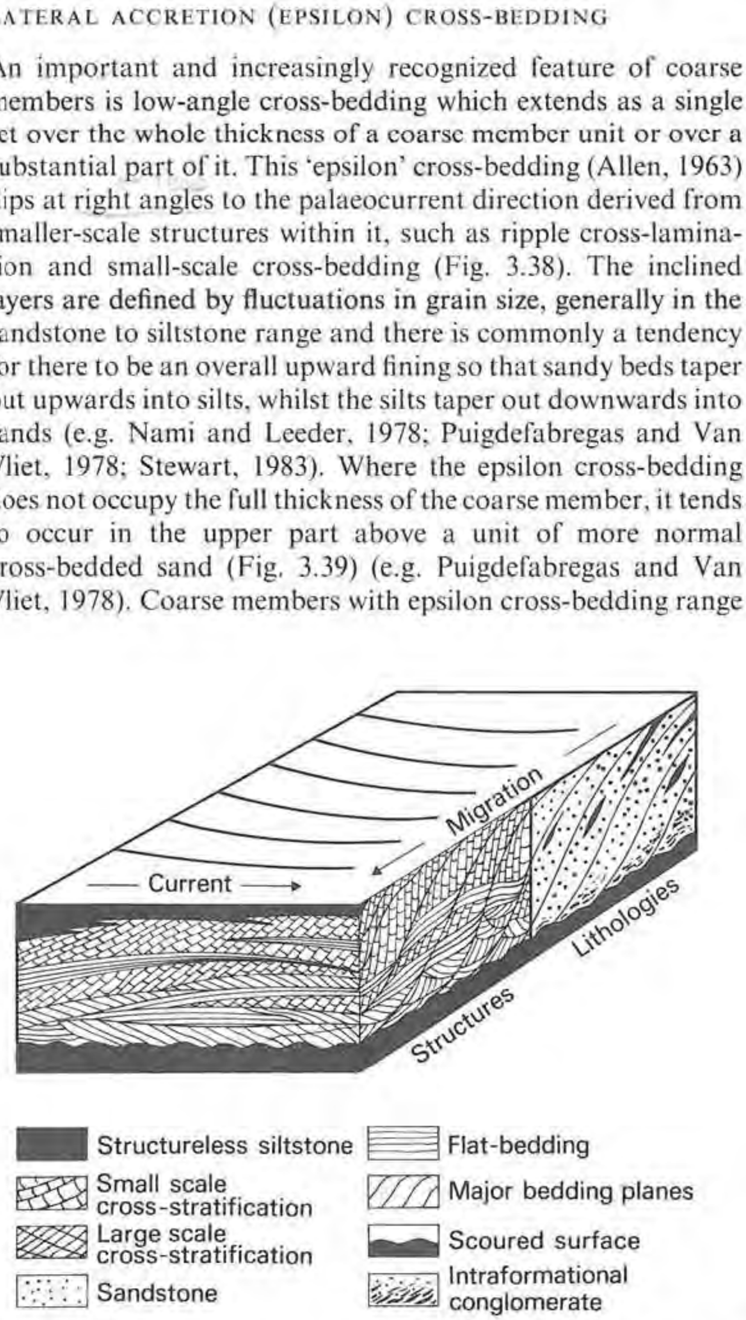

Fig. 3.38. Idealized model to show the main features of lateral accretion (epsilon) cross-bedding, based on examples from the Old Red Sandstone. Major bedding surfaces dip at between 4° and 14° depending on the thickness of the unit. Large vertical exaggeration (after Allen, 1965b).

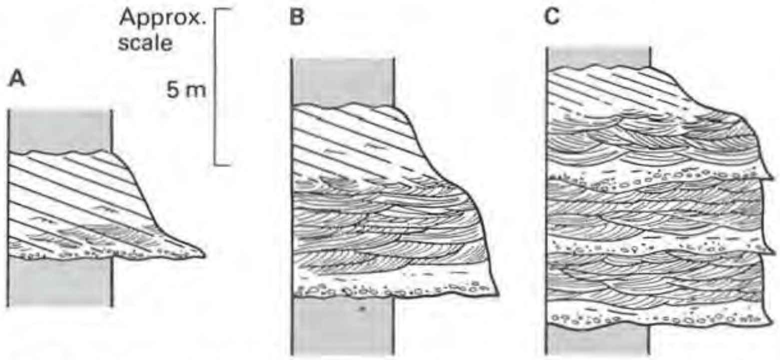

Fig. 3.39. Examples of different types of lateral accretion cross-bedding in small channel sandbodies in the Tertiary of the Pyrenees. (A) Lateral accretion bedding occupies the full thickness of the sand member. (B) Lateral accretion bedding is confined to the upper part of the sandbody, a situation probably related to the vertical range of water stage fluctuations. (C) Multistorey example of the type illustrated in (B); lateral accretion bedding is only apparent in the uppermost unit (after Puigdefabregas and Van Vliet, 1978).

in thickness up to about 5 m and sets may commonly be traced laterally over tens or even hundreds of metres. The cross-bedding is not always continuous and may be broken by internal erosion surfaces (e.g. Allen and Friend, 1968; Beuf, Biju-Duval *et al.*, 1971). Where seen, its down-dip termination is commonly a unit of fine-grained sediment bounded on the opposite side by a steep erosion surface (Fig. 3.40) (e.g. Puigdefabregas and Van Vliet, 1978; Nami and Leeder, 1978).

Where the upper bedding surfaces of epsilon cross-bedded units are seen, they are commonly characterized by a series of curved roughly concentric ridges (Fig. 3.41) (e.g. Puigdefabregas, 1973; Nami, 1976; Puigdefabregas and Van Vliet, 1978).

The whole unit is interpreted as the product of lateral accretion on an inclined surface which is usually taken to be a point bar of a meandering channel, but which may occur as a local element in a more braided system (e.g. Allen, 1983). The structure is not known from modern point bars, probably because of the logistical problem of excavation, but is well known from tidal creeks (e.g. Van Straaten, 1951). The fluctuation in grain size, necessary for the ready recognition of the structure, reflects stage fluctuation in the channel concerned. Stage fluctuation is also implied by the presence of erosion surfaces within the epsilon cross-bedding. There is a greater tendency for epsilon cross-bedding to show up in coarse members laid down by channels which had a high suspended load. The confinement of the structure to the upper parts of some coarse members may reflect the range of water stage fluctuation in the channel; the lower part of the coarse member, lacking epsilon cross-bedding was, presumably, subjected to vigorous currents throughout the year.

The down-dip termination of the structure in a clay-rich unit is thought to record the abandonment of the channel, probably by avulsion or neck cut-off and the eventual infilling by sediment from suspension. The patterns of curved ridges seen on upper surfaces clearly reflect the highly curved nature of the accretionary bank and imply a meandering channel. The ridges themselves are analogous with the scroll bar ridges which characterize modern point bars.

Fig. 3.41. Exposed scroll bars on the upper surface of a channel sandbody in the Scalby Formation (Middle Jurassic), Burniston, Yorkshire, England. The individual point bar accretion units have laterally erosive contacts with one another and a restricted lateral extent.

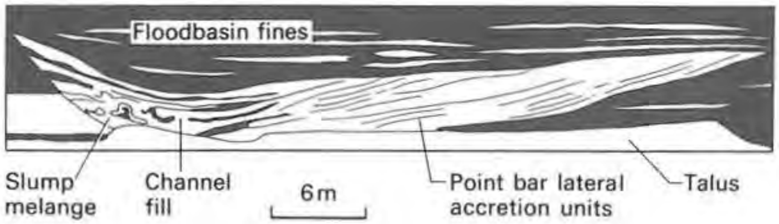

Fig. 3.40. Channel sandbody with restricted lateral accretion as shown by epsilon cross-bedding on the right hand side. Lateral movement ended with channel abandonment and the infilling of the residual channel by bank slumping and deposition of fine-grained sediment: Scalby Formation (Middle Jurassic), Yorkshire, England (after Nami and Leeder, 1978).

3.9.4 Patterns and organization in sandy alluvium

In order to deduce the nature of the alluvial system responsible for a particular sandy alluvial sequence, it is necessary to consider not only the constituent facies but also the facies organization and distribution of coarse and fine members both internally and with respect to one another. For convenience these are discussed under four headings, but these are not independent variables and are highly inter-related. Whenever possible evidence of all types should be integrated to give the least ambiguous indication of the alluvial system.

SANDBODY SHAPE

An important distinction must first be drawn between those sandbodies which develop during continuous accretion of a floodplain or fan system and those which infill a deeper valley cut during a period of incision. Incision can be into a recently deposited alluvial succession due to a fall in base level or a climatic or tectonic change (Allen and Williams, 1982). It can also occur at an unconformity of greater age significance (e.g. Sedimentation Seminar, 1978). In either case, the infill takes place as the base level begins to rise or the climatic or tectonic activity stabilizes or reverses (cf. Schumm, 1977).

Valley-fill sandbodies have highly variable shapes and sizes related to the erosional relief and the nature of the eroded substrate. It is not possible to generalize about them, though it is likely that internally they will show a complex stacking of channel sandbodies, commonly with little preserved fine member sediment. Sandbodies laid down as part of the more or less steady accumulation of an alluvial succession also differ greatly in their size and shape. They range from single channel units, isolated in fine member sediments to composite bodies made up of many channel units in lateral or vertical continuity. Sandbody shape depends partly on channel type and partly on channel behaviour (Fig. 3.42) (Friend, 1983). Where sandbodies are sheet-like with no observed lateral margins, they may result from major sheet floods (Collinson, 1978) or be the result of extensive channel migration. Where channelized flow is responsible for the sandbodies, it is important to differentiate between those formed by fixed channels and those resulting from a mobile channel belt (e.g. Nami and Leeder, 1978; Friend, 1983).

Fixed channels give laterally restricted and highly elongated sand ribbons, usually isolated in finer grained sediment. Where exceptional exposure allows the plan form of the channels to be seen, they may have straight or, more commonly, meandering traces as in the spectacular Tertiary exposures of the Ebro Basin, N. Spain (Friend, Marzo *et al*., 1981). The fact that these channels are filled with sand but are surrounded by fine member deposits argues for a mixed load stream with a high suspended load. It also suggests a gradual waning of flow in the channels so that bedload transport persisted almost to the point of abandonment. In addition, channels of this type tend to be rather steep sided with strongly concave upwards bases and their cross-section is close to that of the original channel.

Mobile channels deposit more laterally extensive sandbodies

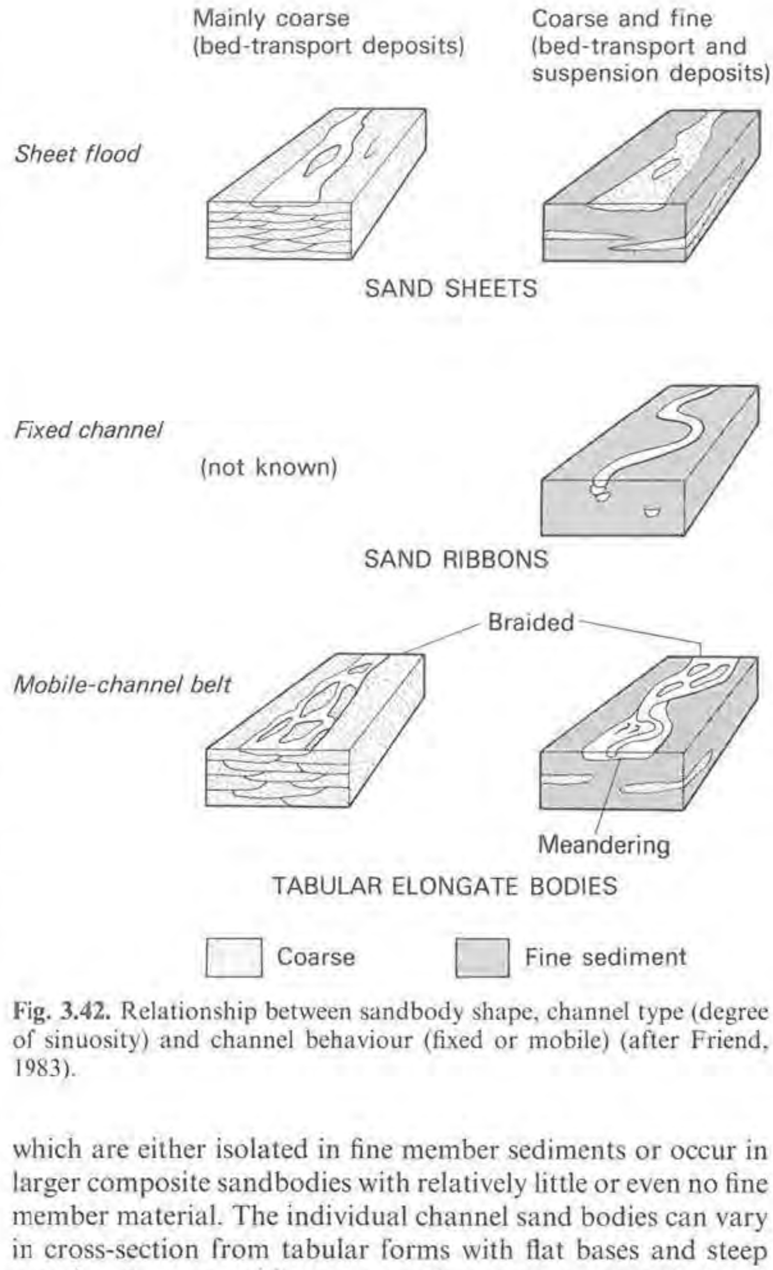

Fig. 3.42. Relationship between sandbody shape, channel type (degree of sinuosity) and channel behaviour (fixed or mobile) (after Friend, 1983).

which are either isolated in fine member sediments or occur in larger composite sandbodies with relatively little or even no fine member material. The individual channel sand bodies can vary in cross-section from tabular forms with flat bases and steep margins to ones with more gently concave upwards basal erosion surfaces (e.g. Moody-Stuart, 1966). Flat-based sandbodies are more commonly attributed to meandering streams where lateral migration of the scour pool erodes to a fairly constant level and where cut banks tend to be steep. Such an interpretation is supported by the presence of epsilon cross-bedding in the channel fill as in the Old Red Sandstone of Spitzbergen (Moody-Stuart, 1966; Fig. 3.43(A)) and the Tertiary of the Pyrenees (Puigdefabregas and Van Vliet, 1978). Such sandbodies produced by meandering streams are likely to have a more limited lateral extent than those produced by lower sinuosity streams owing to the tendency for clay abandonment plugs to limit their lateral migration (e.g. Collinson, 1978).

Concave-upwards based, lenticular channel bodies are commonly regarded as the products of relatively low-sinuosity

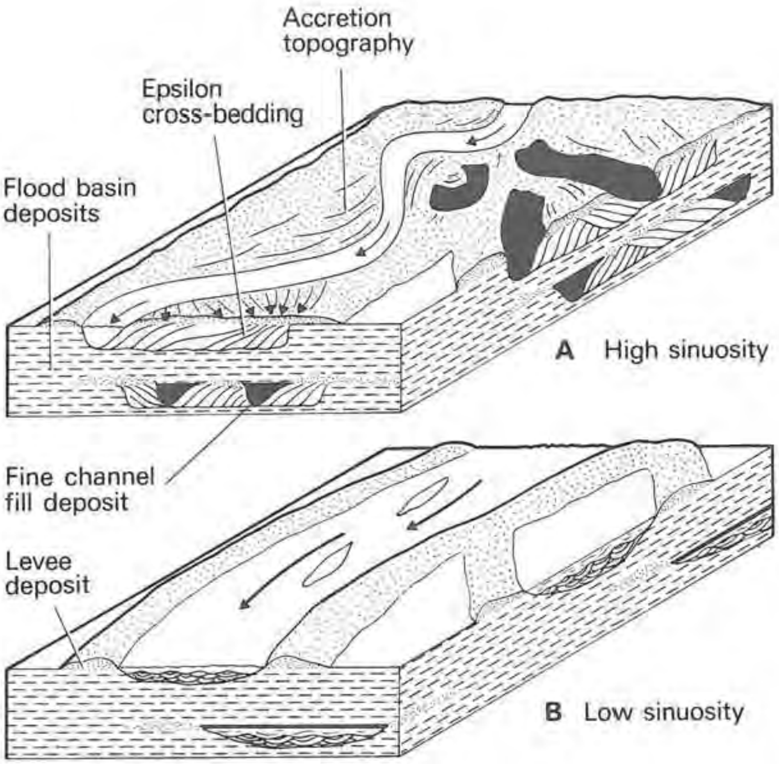

Fig. 3.43. Models suggesting the relationships between channel shape, internal structure and channel plan based on examples from the Old Red Sandstone of Spitzbergen (after Moody-Stuart, 1966).

streams which cut and then filled their channels by a mixture of vertical and lateral accretion (e.g. Moody-Stuart, 1966; Kelling, 1968; Thompson, 1970). Where these occur isolated in finer sediment as in the Old Red Sandstone of Spitzbergen (Moody-Stuart, 1966), it is relatively easy to be confident of their true cross-section (Fig. 3.43B). In composite units, made of many channel fills, the shape of the sandbody may be less clear and the status of particular erosion surfaces may be difficult to interpret. The Westwater Canyon Member of the Morrison Formation (Jurassic) of New Mexico is a composite sheet some 100 km wide, over 160 km long (parallel to the dominant palaeocurrent direction) and up to 60 m thick (Fig. 3.44; Campbell, 1976). This sheet is built up of a series of mutually erosive channel systems 1.5–34 km wide and 6–60 m deep each of which, in turn, is a composite of individual channels and channel fragments 30–350 m wide and 1–6 m deep. Whilst the channel systems are tabular in cross-section with fairly steep sides, the individual channels have concave upwards bases. Channel edges are sinuous but non-meandering in plan suggesting that the concave upwards channels were of low sinuosity within a braided channel system. Multiple cutting and filling were associated with channel switching within the tract. The system compares in scale with that of the present-day Kosi River (Gole and Chitale, 1966). In more laterally restricted outcrop, it would be very difficult to distinguish those erosion surfaces associated with channels from those associated with channel systems.

In general, sandbodies produced by low sinuosity streams are unlikely to show channel margins because of their greater tendency to migrate laterally, without the clay plugs which inhibit the migration of meandering streams.

ALLUVIAL ARCHITECTURE: COARSE MEMBER/FINE MEMBER RATIOS

The larger scale distribution of alluvial sandbodies within fine member sediments and their mutual relationships have come to be known as 'alluvial architecture' (Allen, 1978; Fig. 3.45). Some coarse member units are isolated within fine member sediments (e.g. Moody-Stuart, 1966; Wells, 1983); other coarse members

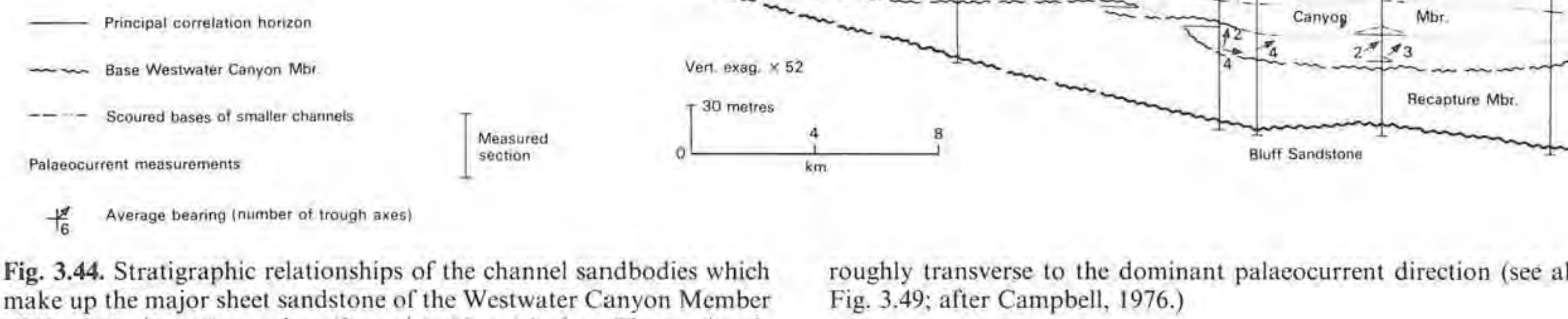

Fig. 3.44. Stratigraphic relationships of the channel sandbodies which make up the major sheet sandstone of the Westwater Canyon Member of the Morrison Formation (Jurassic), New Mexico. The section is roughly transverse to the dominant palaeocurrent direction (see also Fig. 3.49; after Campbell, 1976.)

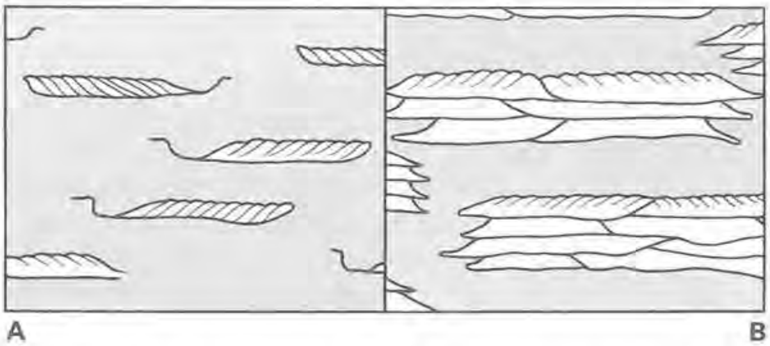

Fig. 3.45. Channel sandbodies in the Tertiary of the southern Pyrenees. (A) Isolated sandbodies with evidence of limited lateral accretion and a short residence period on the floodplain. (B) Multistorey sandbodies, individually showing evidence of lateral accretion and collectively showing the development of more long-lasting meander belts (after Puigdefabregas and Van Vliet, 1978).

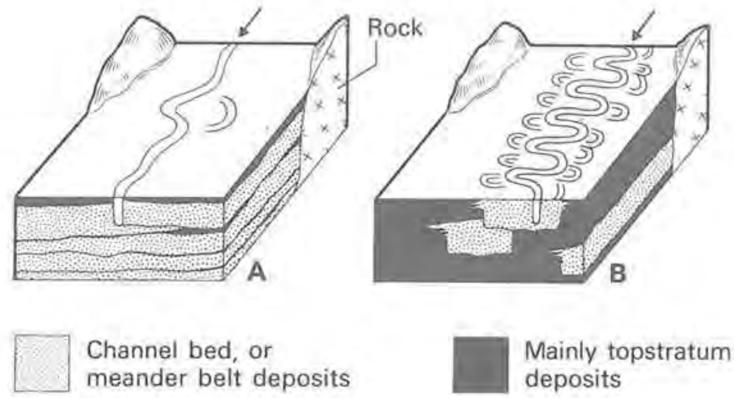

Fig. 3.46. Hypothetical models illustrating the broad facies relationships which might be produced by streams of (A) low sinuosity and high lateral mobility and (B) high sinuosity and restricted lateral mobility. The preservation of overbank fines depends on the relationship between migration frequency and subsidence rate in each case, cf. Fig. 3.40 (after Allen, 1965a).

are mutually erosive and combined into major sandstone sheets with virtually no preservation of overbank fines (Fig. 3.44) (e.g. Stokes, 1961; Campbell, 1976; Bhattacharyya and Lorenz, 1983). Between these extremes there is a whole spectrum of variation which expresses itself both as the *coarse member/fine member* ratio and the degree of *interconnectedness* of the coarse member sandbodies, a property of some economic significance.

These larger scale properties are controlled by several variables which are not necessarily independent. The first control is the nature of the alluvial system. In a high bedload system where deposition of suspended load on inter-channel areas is small, channels tend to be mobile and to migrate so that little fine sediment is preserved (Figs 3.42, 3.46) (cf. Allen, 1965c;

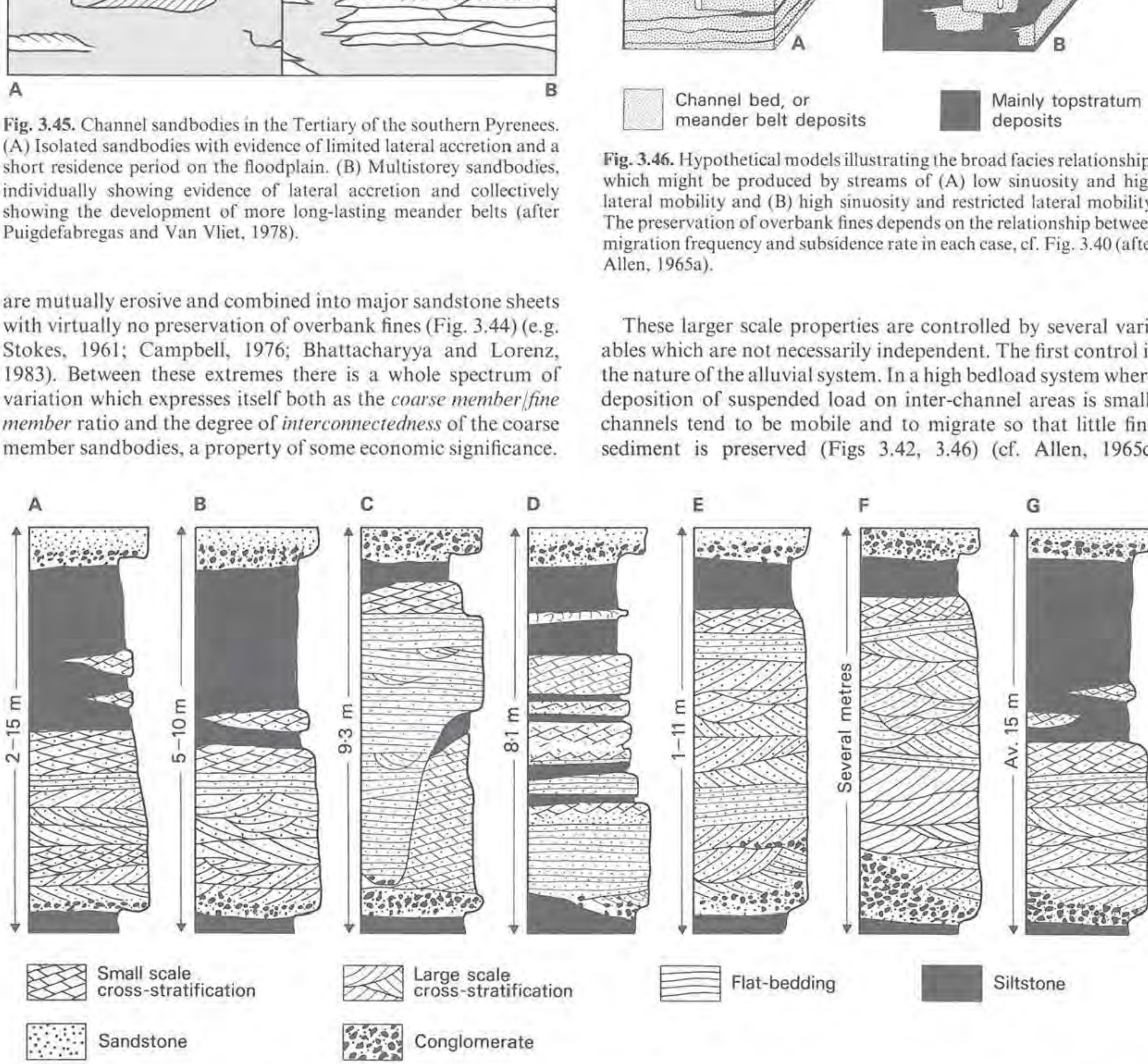

Fig. 3.47. Representative coarse member sandstones from the Old Red Sandstone. (A, B) Welsh Borders, (C) Tugford Clee Hills, Shropshire, (D) Mitcheldean, (E) Forest of Dean, (F) Clee Hills, (G) Spitzbergen. Note the variety of scale and of facies sequence. Where a range of thickness is given, the sequences are somewhat idealized standards (after Allen, 1965a).

Thompson, 1970). With a mixed load or a sand-poor system, overbank deposition of fines is more abundant and channels are more stable; channels shift by avulsion so that sandbodies stand a higher chance of being isolated in fine member deposits. Other controls, suggested by simulation studies, are subsidence rate, avulsion frequency and floodplain width (Leeder, 1978; Allen, 1978; Bridge and Leeder, 1979). Of these controls, avulsion frequency is not an independent variable but is the complex result of, amongst others, the sedimentation rate gradients across the floodplain (Crane, 1983). Interpretations of alluvial systems and particularly of channel type which rely heavily on coarse member/fine member ratios, as seen in borehole or other restricted vertical sections, should therefore be treated with great caution. In addition it is important to distinguish between those coarse members which result from infilling of incised valleys and those which are the result of continuing overall accumulation. In the Old Red Sandstone of South Wales, deeply incised channels whose margins cut through several palaeosol horizons are attributed to incision due to a lowered base-level. They contrast with the shallower coarse members which are the result of channels active during accumulation (Allen and Williams, 1982).

INTERNAL FACIES RELATIONSHIPS

The now classic fining-upwards sequence of Bersier (1959) and Bernard and Major (1963) has been extended and refined by many authors, notably Allen (1964, 1965a, 1965b, 1970a, 1970b, 1974b), Visher (1965), Jackson (1978), and Puigdefabregas and Van Vliet (1978). A sandstone overlying a horizontal erosion surface fines upwards and commonly shows a related upwards change from cross-bedding to parallel and/or ripple lamination before it grades into an overlying fine member. There may be a lag conglomerate at the base and, in thicker units, the cross-bedding may show an upward decrease in set thickness (Fig. 3.47). This simple pattern is, however, something of an idealization, akin to the Bouma sequence in turbidites. Whilst it does occur, many coarse members in fluvial sequences show different or more complex vertical facies sequences.

A direct interpretation of the classic fining-upwards sequence, using only the internal facies evidence, indicates a waning of flow power from an initial erosive phase. This waning may be accounted for by the steady state point bar model of lateral migration combined with the spatial separation of flow strengths over the point bar surface (Sect. 3.4.2). In this case, the thickness of the coarse member corresponds with the depth (bankfull) of the migrating channel. With this explanation, variations between coarse members in a sequence can be accounted for by differences in channel slope and channel curvature (Allen, 1970a, 1970b). The waning flow strength implicit in the fining-upwards sequence can, however, also be accounted for by a waning flow through time. Many coarse members thinner than, say, 2 m may not be channel deposits at all, but the results of catastrophic sandy sheet floods of wide lateral extent (cf. Fig. 3.36). They might, therefore, be better regarded as graded beds rather than fining-upwards units. This alternative interpretation is particularly attractive when the sandbodies are extensive laterally and have no observed erosive margins (Collinson, 1978). Allen's (1964) units in the Old Red Sandstone at Lydney (Fig. 3.48) and many of the coarse members in the Red Marls of Pembrokeshire (Allen, 1974b) can as readily be interpreted as sheet flood deposits of distal terminal fans or as crevasse splays (Collinson, 1978; Tunbridge, 1981; Hubert and Hyde, 1982). The problem of deciding which erosively based sheets result from channel migration and which result from episodic flood events is not easily resolved. Thickness of the coarse member gives a rough guide as it is difficult to imagine either sheet floods repeatedly generating sand units more than 2 or 3 m thick or laterally migrating channels

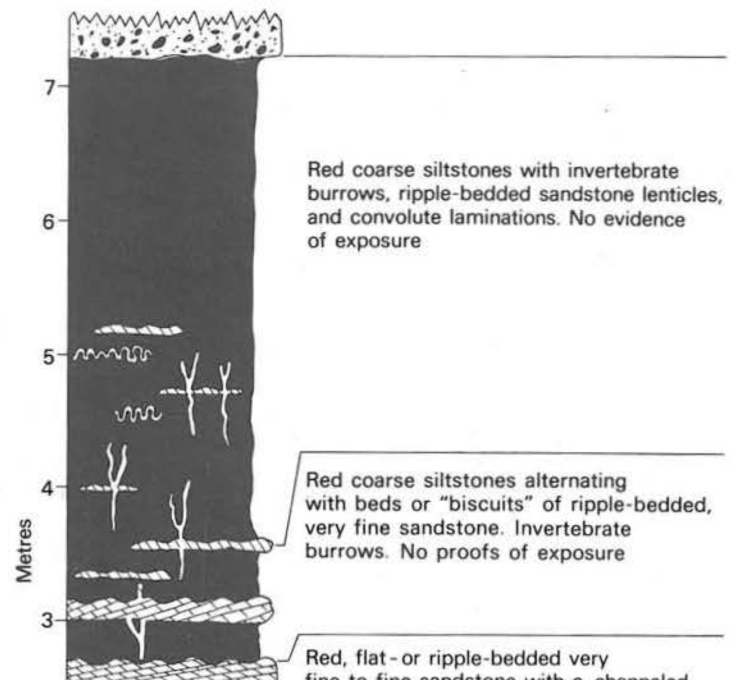

Fig. 3.48. Sequence from the Old Red Sandstone at Lydney, Gloucestershire. The sandbodies are all sufficiently thin for them to be interpreted as a series of sheetflood deposits, though the thickest sandstone could represent a shallow channel (after Allen, 1964).

generating sequences less than 1 m thick without observable channel margins. Between these thicknesses, however, either channel migration or sheet flooding may seem equally likely. The occurrence of lateral accretion (epsilon) cross-bedding is strongly suggestive of a meandering channel especially if the sandbody is of limited lateral extent and has steep margins. Top bedding surfaces showing curved scroll bar ridges confirm the interpretation when they are seen (Figs 3.38, 3.39) (Allen, 1965b; Puigdefabregas, 1973; Nami, 1976; Nami and Leeder, 1978).

Lateral accretionary bedding is not, however, exclusively confined to deposits of meandering systems. Allen (1983) has shown that within complex, sheet-like sandstones of the Devonian Brownstones of the Welsh Borders, elements of lateral accretion can be recognized along with other elements produced by downstream advance of transverse bars and the development of dunes, the whole reflecting a rather wandering, low-sinuosity system.

In addition, the absence of lateral accretion bedding should not be taken as indicating a non-meandering system. Epsilon cross-bedding appears to require a fluctuating discharge regime and probably a rather fine-grained load and its recognition in the field requires a section roughly perpendicular to flow (Puigdefabregas and Van Vliet, 1978; Plint, 1983; Stewart, 1983). The scour associated with the superimposed bedforms on the point bar surface may effectively obliterate any potential epsilon cross-bedding (e.g. Frazier and Osanik, 1961).

More sheet-like sandstones, and those with concave-upwards bases, both of which are more readily interpreted as the products of low sinuosity streams, tend, on the whole, to have a less well-ordered internal organization. Many show an upward fining, particularly in their upper parts. In some more isolated sandbodies with concave-upwards bases, the highest parts of the fill extend laterally beyond the confines of the channel to give 'wings' extending into the flanking fine member unit (Friend, Marzo *et al.*, 1981). The more extensive sheet-like coarse members are commonly composite units with a hierarchy of erosional surfaces (cf. Campbell, 1976; Allen, 1983) between which tabular and trough cross-bedded sandstone dominates but with little vertical ordering (Figs 3.49, 3.50). This probably reflects the less ordered spatial distribution of bedforms on the channel floor and also the more random patterns of channel shifting and migration of low sinuosity and braided sandy streams.

In an attempt to recognize some order in these sequences and to erect a vertical sequence model to compare with the classical fining-upwards model, Cant and Walker (1976) suggested a model sequence which integrated both facies and palaeocurrent information from the Devonian Battery Point Sandstone of

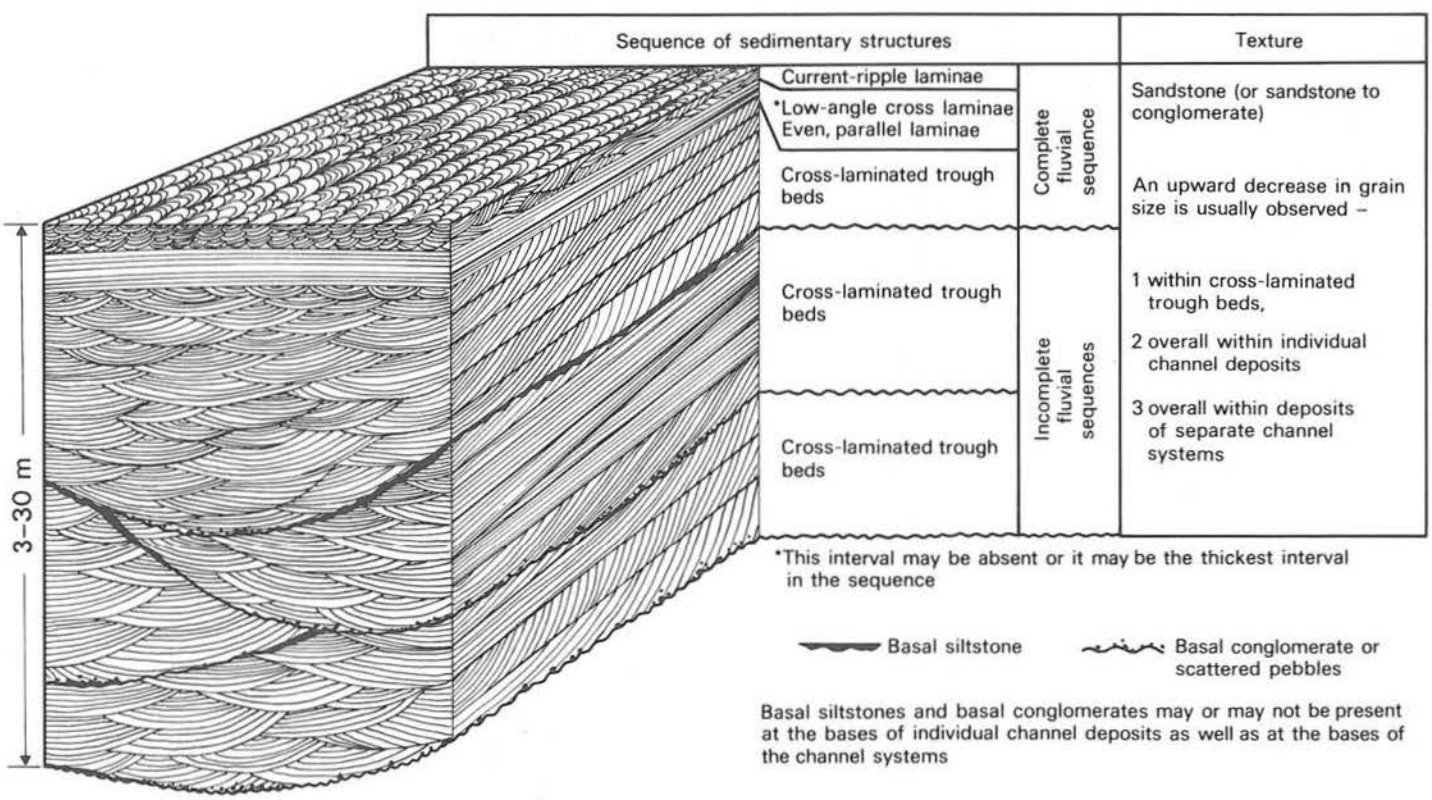

Fig. 3.49. Sequence of sedimentary structures and textures within a channel system of the Westwater Canyon Member of the Morrison Formation (Jurassic) of New Mexico. Note the concave upwards bases of the individual channels (see also Fig. 3.44; after Campbell, 1976).

Quebec (Fig. 3.51). This model is derived from the 'distillation' (*sensu* Walker, 1979) of sequences observed in a number of separate coarse member units (Sect. 2.1.2). An essential feature of the model is the presence of tabular sets whose dip azimuths diverge by around 60° (maximum 90°) to either side of the mean trough direction. These are interpreted as the results of the highly skewed crestlines and the oblique migration of mid-channel bars. The troughs result from dunes migrating in inter-bar areas and over the tops of the bars during the construction of sand flats.

Although the analysis is important in integrating the palaeocurrent and facies data, the 'distillation' process has effectively eliminated all information on lateral variability of facies within the coarse members. A comparison with the South Saskatchewan model (Fig. 3.10) where different vertical sequences are generated in different parts of the channel complex is thereby weakened. In addition, the single sequence model is open to alternative interpretation as divergent and anomalously large tabular sets might also reflect scroll bars on a point bar (cf. Jackson, 1976). Divergence of the tabular sets to both sides of the trough mean direction, *within the same channel unit* should ideally be demonstrated in order to apply confidently a South Saskatchewan model. With no directional information, anomalously large tabular sets could also reflect chute bars (Fig. 3.24). These qualifications clearly demonstrate the need not only to integrate facies description and sequence analysis with palaeocurrent data but also to use lateral variability to help interpretation rather than to filter it out as background noise.

Some thick channel sandstones are characterized by particularly large tabular sets of cross-bedding which make up a large proportion and in some cases nearly all the sandbody. In the Namurian of Northern England, large fluvial distributary channels in a delta top setting (Sect. 6.7.1) are up to 40 m deep and of the order of 1 km wide (Fig. 3.52). They have steep sides cut into finer sediments and are filled with coarse pebbly sandstone in four facies (McCabe, 1975, 1977):

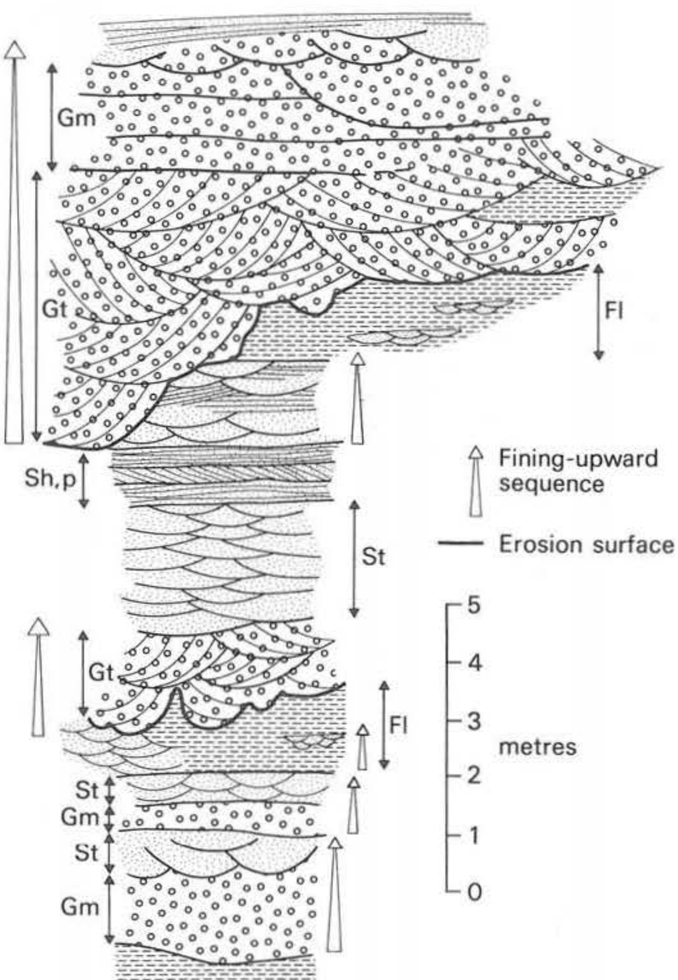

Fig. 3.50. Internal organization and larger scale erosional relationships between channel sandbodies attributed to sandy braided stream deposition. Letter code is that of Miall (1977) (see Sect. 3.8.1), Cannes de Roche Formation (Carboniferous) Gaspé, Quebec (after Rust, 1978).

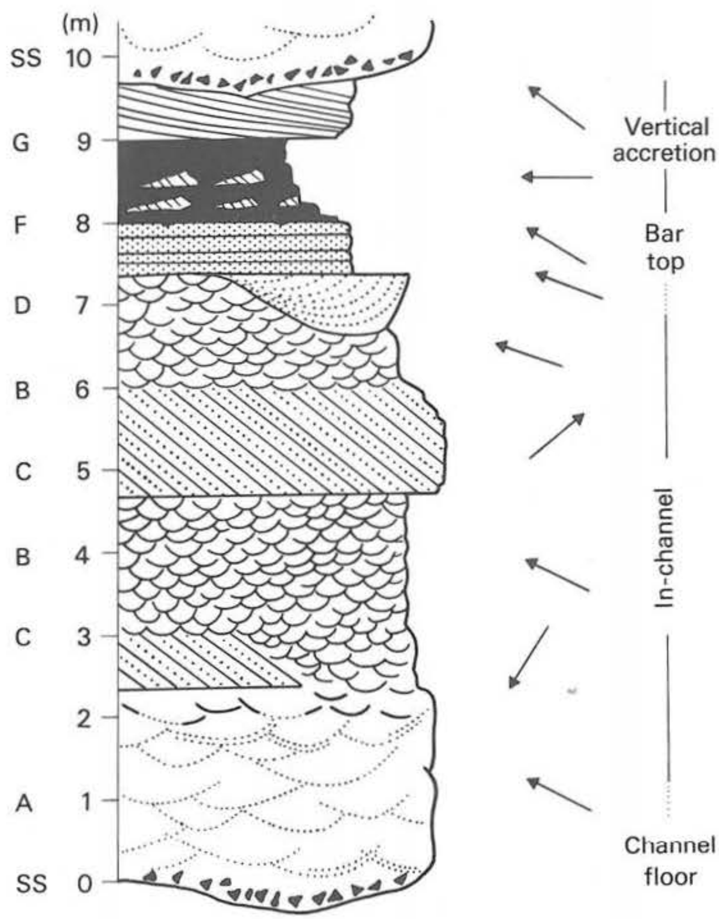

Fig. 3.51. Facies model for the Battery Point Sandstone (Devonian, Quebec) constructed by method described in Fig. 2.4, showing the relationship between vertical sequences of sedimentary structures and their palaeocurrents. The sequence has been compared with that predicted from the modern South Saskatchewan River (Fig. 3.10) (after Cant and Walker, 1976).

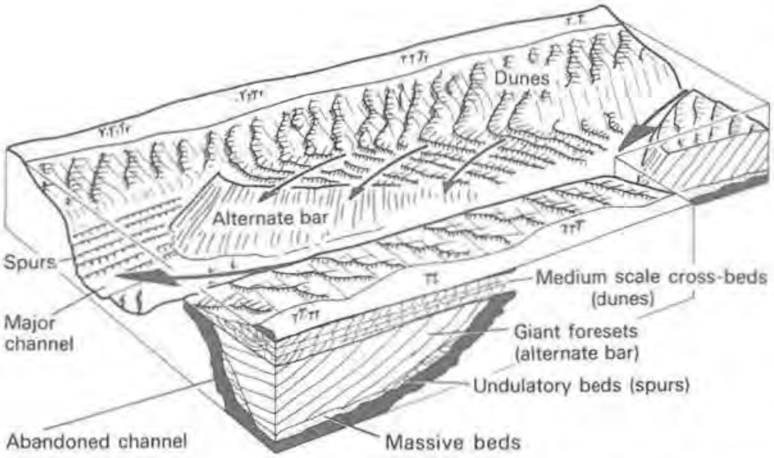

Fig. 3.52. Model for the large scale delta-top, fluvial channels of the Namurian Kinderscout Grit of northern England (after McCabe, 1975).

(a) *Massive sandstone beds* up to 2 m thick and roughly horizontally bedded rest directly on the basal erosion surface and pass up into (b) *undulatory beds*. These are about 10 cm thick and have a wavy form with a wavelength of between 10 and 20 m and a relief of about 1 m in sections perpendicular to the independently determined palaeocurrent. Individual beds rise gradually in height towards a channel margin with an overall rise of 7 m being recorded across a series of 6 undulations. This unusual facies is thought to be the product of vertical accretion on sand ridges aligned parallel to the current, similar to ones recorded from the Brahmaputra (Coleman, 1969).

(c) *Giant foresets* occur in large tabular sets of cross-bedding, up to 40 m thick though normally less than 25 m, which extend horizontally for more than 1 km both parallel and perpendicular to the foreset dip direction. In plan, foresets are convex downcurrent. Their original interpretation as Gilbert deltas (Collinson, 1968) is now superseded by the idea that, being channel-bound, they are the result of large alternate bars with slip faces which were active in a river channel and were attached to the banks. Resting directly on the large tabular sets are (d) *medium-scale, trough cross-beds* in sets of less than 1 m thick and with closely similar palaeocurrent directions.

The massive sandstone occupies the deepest part of the channels and the undulatory beds occur near channel margins. Both are overlain by the large tabular sets whose foresets dip in directions around 40° divergent from the inferred direction of elongation of the sand ridges. If the large foresets record the skewed slip faces of alternate bars, the ridges may have resulted from spiral eddies shed by these forms. The medium-scale cross-bedding records the migration of dunes over the alternate bars, feeding sediment to the major slip face, but also accreting vertically.

A similar sequence has been recorded from more sheet-like channel sandbodies in the Permo-Triassic Hawksbury Sandstone of New South Wales (Conaghan and Jones, 1975). Here massive and poorly bedded medium sandstones directly overlie irregular erosion surfaces and are in turn overlain by large single tabular sets up to 10 m thick developed in coarse sandstone. Above these are smaller, medium-scale sets, also dominantly tabular. In plan view the large and medium-scale sets show foresets which are mainly straight or concave downstream suggesting, for the large sets at least, lunate bedforms. Most interestingly, the large cross-bedded sets, when traced up current pass into a series of smaller sets whose bounding surfaces are inclined downstream. This suggests that the large forms which eventually developed a single major slip face began as composite forms by the stacking up of smaller bedforms, probably dunes and transverse bars. The descending sets on the downstream side of the composite bar accrete due to the expansion of flow there and eventually, as individual forms come progressively to a halt, a single major slip face develops. Such bar evolution might be related to flood cycles in the river as seen in the present-day Brahmaputra (Coleman, 1969). Similar descending sets are also reported from other channel sandstones of probable low-sinuosity origin (Fig. 3.53) (Banks, 1973d; Haszeldine, 1983).

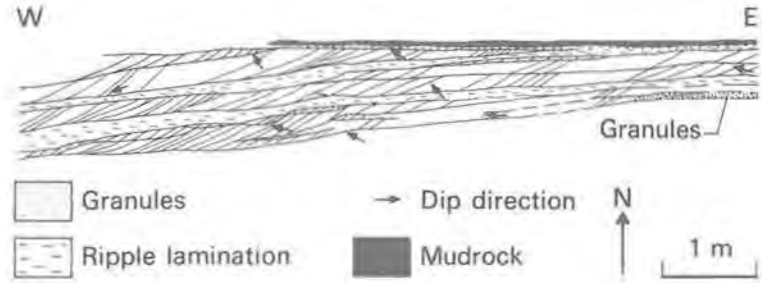

Fig. 3.53. Section parallel to the palaeocurrent through a bar complex within a fluviatile channel sandbody showing the compound nature of the bar unit made up of descending tabular cross-bedded sets. Upper Carboniferous Coal Measures, north-eastern England (after Haszeldine, 1983).

PALAEOCURRENT DISTRIBUTION

The idea that a wide dispersion of palaeocurrents is typical of deposits of meandering streams whilst lower dispersions characterize lower sinuosity streams is well established (e.g. Kelling, 1968; Thompson, 1970). However, this idea, which comes from an appreciation of the variation in channel orientation in present-day streams, is only appropriate when local vector means of cross-bedding or actual channel orientations are used. The individual bedforms which migrate in river channels and which give rise to cross-bedding are extremely complex in their behaviour. Cross-bedding directions most commonly relate to different patterns of bar movement rather than to channel type (Smith, 1972; Banks and Collinson, 1976). Palaeocurrents should therefore be carefully sampled with regard to the type of sedimentary structure and to their position in the channel sequence. They can then give some indication of detailed channel processes and of channel type (e.g. Cant and Walker, 1976). Their use in a more widespread regional sense is usually less diagnostic.

RÉSUMÉ

Few of the approaches outlined above give an unambiguous interpretation of channel type in alluvial successions. Our increasing appreciation of the complexities of modern alluvial processes shows that the application of a few simple models will never reveal the full variability of the ancient. Most interpretations will be based on a balanced appraisal of the total data. Whilst an interpretation of a particular local succession or group of closely spaced local successions may often be all that can be achieved, wherever possible it is very valuable to use the approaches outlined in a comparative way, either recording changes with time through a stratigraphic sequence or spatial changes from more widely spaced localities. Such comparisons can often give important clues to the nature of the alluvial system and thereby suggest the most likely channel types.

FURTHER READING

Collinson J.D. and Lewin J. (Eds) (1983) *Modern and Ancient Fluvial Systems*, pp. 575. *Spec. Publ. int. Ass. Sediment.*, **6.** Blackwell, Oxford.

Duchaufour P. (1982) *Pedology*, pp. 448. Allen and Unwin, London.

Goudie A. (1973) *Duricrusts in Tropical and Subtropical Landscapes*, pp. 174. Oxford University Press, Oxford.

Gregory K.J. (Ed.) (1977) *River Channel Changes*, pp. 448. Wiley, Chichester.

Leopold L.B., Wolman M.G. and Miller J.P. (1964) *Fluvial Processes in Geomorphology*, pp. 522. Freeman, San Francisco.

Miall A.D. (Ed.) (1978) *Fluvial Sedimentology*, pp. 859. *Mem. Can. Soc. Petrol. Geol.* **5**, Calgary.

Miall A.D. (Ed.) (1981) *Sedimentation and Tectonics in Alluvial Basins*, pp. 272. *Spec. Pap. geol. Ass. Can.* **21.** Waterloo.

Schumm S.A. (1977) *The Fluvial System*, pp. 338. Wiley, New York.