The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers

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Abstract: Models of braided-river deposition must be detailed, fully 3D, and preferably quantitative to be of use in understanding and predicting the nature of ancient deposits. In order to construct and validate adequate predictive models it is necessary to have information on: (1) variation and interaction of channel geometry, water flow and sediment transport in time and space in modern channel belts, as these control erosion and deposition, the formation and migration of channels and bars, and channel abandonment and filling; (2) 3D variation of bed geometry, texture, sedimentary structures and paleocurrents throughout modern channel-belt deposits, including the age and spatial arrangement of preserved parts of bars and channel fills; (3) long-term (more than hundreds of years) trends in channel and floodplain geometry, flow and sedimentary processes in order to understand channel-belt movements such as avulsions, and the spatial arrangement of channel-belt deposits relative to overbank deposits. Such information is rare because: (1) it is difficult to study modern braided-river geometry, flow and sedimentary processes throughout a range of the all-important high discharges; (2) detailed reconstructions of braided channel and bar geometry and movement are only available for the past half-century and cannot readily be linked to causative mechanisms; (3) 3D documentation of modern deposits below the water table (especially large scale features like lateral-accretion bedding) requires extensive coring and dating of the deposits, and geophysical profiling. As a result of this lack of information, and because of the quality of analysis and presentation of the information available, existing braided-river facies models are virtually useless as interpretive and predictive tools. The nature of the information available is critically reviewed. Using information from recent detailed field and laboratory studies of the geometry, flow and sedimentary processes in braided rivers of simple geometry, in single river bends, in channel confluences, and using some theoretical reasoning, it has been possible to construct fully 3D qualitative and quantitative models of braided river deposits. These models can be used to provide sophisticated quantitative interpretations of palaeochannel geometry, hydraulics and migration, as illustrated by comparison with some particularly well described examples of ancient braided river deposits.

Channel geometry, water flow and sediment transport in braided rivers interact and vary in time and space, resulting in erosion and deposition, growth and migration of bars, and the formation, migration and filling of channel segments. An understanding of this interaction is important in modern environmental and engineering problems such as water supply and flood risk evaluation, dispersion of pollutants, disturbance of freshwater wildlife habitats, sedimentation in navigated channels and reservoirs, bank erosion and channel migration, construction of artificial channels, bridge piers, pipeline crossings, and stabilized banks, and sediment dredging operations.

A detailed knowledge of modern river deposits and their relationship to formative flow and sedimentary processes is critical for interpretation of ancient river deposits and predicting their subsurface geometry and facies. Braided river deposits are commonly associated with important resources such as water, oil, gas, coal, gold, sand and gravel. Fluvial reservoirs containing water, oil and gas are typically heterogeneous and definition of porosity and permeability variations requires, amongst other things, detailed and accurate facies models at a variety of scales. A particularly pressing concern is the widespread pollution of aquifers in Pleistocene braided river outwash deposits which blanket large areas of the northern hemisphere.

Depositional models of river channels abound in the sedimentological literature. Although they appear to be 3D, in reality the deposits are normally only represented in one or two vertical cuts through the channel belt; parallel and normal to the channel direction. The information shown in these 2D cuts lacks critical detail of the spatial variation of bed thickness and orientation, grain size, internal structures, paleo-

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currents, biological aspects and all the possible varieties of these features (Jackson 1978; Bridge 1985). The nature of preservation of superimposed bars and channel fills in modern channel belts is virtually unknown. Furthermore, the models are purely qualitative, and most do not contain even approximate scales. In short, existing facies models are of little use in interpreting ancient channel deposits to any reasonable degree of detail, and certainly cannot be used to predict spatial variations in porosity and permeability. This state of affairs is partly based on the method of analysis and presentation of the primary data available, but is mainly based on the lack of primary data from modern rivers.

It is common practice in building fluvial facies models to combine observations of grain sizes and bedforms on channel-bar surfaces with reconstructed modes of channel migration to predict the distribution of grain sizes and sedimentary structures in 1D or 2D vertical sections (e.g. Bluck 1971; Jackson 1976*a*; and many others). In most published studies, grain sizes and bedforms on bar surfaces are observed at relatively low flow stages (when they can be seen), whereas it is commonly not known what is present at relatively high flow stages when observation is much more difficult. But it is at these high flow stages when most erosional and depositional activity (including channel migration) takes place, and is most difficult to document. Understandably, studies of the interaction between flow, sediment transport and channel geometry in natural rivers over a range of high discharges are very rare. Modes and rates of channel and bar migration can be documented by direct observation for the duration of a field study, and for up to hundreds of years using maps, aerial photographs and radiocarbon dating of older surface deposits. However, the time resolution of these techniques decreases with increasing age. Long-term, large scale channel-belt movements like avulsion remain poorly known. Laboratory experiments have provided useful information about the nature and short-term evolution of braided rivers but direct application of this information to natural rivers must be treated with caution in view of simplified experimental conditions and scaling considerations (Ashmore 1982, 1991*b*).

Direct observations of sediment deposits in most river studies come from shallow trenches and cut-bank exposures, by necessity above the water table. Indeed, Bluck (1971, 1974, 1976, 1979) has divided river channel deposits into those that are above the low water table and can be observed (supraplatform) and those below that cannot (platform). In perennial rivers, trenches and cut-bank exposures represent a very limited sample of channel-belt deposits. Although it is possible to expose more deposits in ephemeral streams, there have been no studies (for obvious environmental reasons) in which a channel belt has been completely dissected. Long cores through channel belt deposits are not available in most studies, and where available are not normally spaced closely enough. Critical information that is usually lacking is the age of the deposits and direct documentation of large-scale bedding surfaces (e.g. 'lateral-accretion' surfaces, basal erosion surfaces). Thus, facies models based on bed-surface sedimentary features, presumed modes of channel migration and limited exposure of deposits, are largely hypothetical.

The inadequate documentation of modern river deposits probably does not come as a great surprise to those who try to interpret ancient river deposits using modern analogues. Bridge (1985) indicated that it is not particularly easy to distinguish deposits of ancient single- and multi-channel sinuous rivers unless large-scale bedding features associated with channel bars and fills can be documented and understood. It is precisely this information which is lacking from modern river deposits. However, the distinction between different ancient-channel patterns is critical for predicting the geometries and sedimentary characteristics of fluvial sandstone bodies, and for quantitative reconstruction of paleochannel geometry and hydraulics. Willis (1989) has clearly demonstrated that in order to understand any arbitrarily-oriented vertical section through a channel-bar deposit it is necessary to understand the distribution of grain size and bedforms over the whole surface of the bar during deposition, the mode and rate of bar migration, and the orientation of the cross-section relative to the direction of the channel and its migration. As an addendum to this, to fully understand the stacking and preservability of different bars and channel fills in channel belts, it is necessary to know more about the nature of channel migration (e.g. braided-channel diversions, meander cut-off, channel filling) in relation to net channel-belt aggradation.

Thus, in order to construct and validate adequate predictive models of channel-belt deposits it is necessary to have *all* of the following information for different channel-pattern types: (1) variation of flow, sediment transport and bed geometry in space and time, in order to understand and predict erosion and deposition associated with channel bars and fills; (2)

historic data on modes and rates of bank erosion, channel-bar migration, channel cutting and abandoned-channel filling; (3) 3D variation of bed geometry, grainsize, sedimentary structures, and palaeocurrents throughout the channel belt deposits, including the age and spatial arrangement of preserved bars and channel fills. Such data should preferably be quantitative, so that they can be incorporated into generalized mathematical models.

This paper is a review of current knowledge of the origin, geometry, flow, sediment transport, erosion and deposition of modern braided rivers, as gleaned from studies of modern rivers, laboratory experiments and theory. This information will be used to construct both qualitative and quantitative, 3D depositional models for braided rivers, and the use of such models in interpreting ancient braided river deposits is illustrated. The review will be limited to consideration of braided rivers at the scale of several bars over time periods of up to 10^2 years. Space and time limitations prohibit consideration of along-valley variations in braided rivers, associated floodplains or fans, long term ($<10^2$ years) processes associated with channel-belt avulsions, tectonism and climate change.

Definition and origin of braiding in rivers

Definition of braiding

The term 'braided river' denotes a channel pattern as seen in plan view. Leopold & Wolman (1957) defined a braided river as 'one which flows in two or more anastomosing channels around alluvial islands', whereas Lane (1957) stated 'a braided stream is characterized by having a number of alluvial channels with bars or islands between meeting and dividing again, and presenting from the air the intertwining effect of a braid'. Lane also used the term 'multiple-channel stream' to include both braided streams as defined above and anastomosing distributaries on deltas and alluvial fans (see also Chitale 1970). Brice (1964, 1984) also recognized the importance of defining the size of islands relative to channel width, and the difference between within-channel islands and those formed by river diversions. Schumm (1977) defined braided channels as single-channel bedload rivers which at low water have islands of sediment or relatively permanent vegetated islands, in contrast to multiple channel rivers (anastomosing or distributive) in which each branch may have its own individual pattern. The differences among these definitions raise a number of issues that require resolution. In particular: (1) the difference between 'bars' and 'islands' must be defined; (2) the appearance of bars or islands depends on flow stage; (3) there are different types of channel splitting, and the difference between the terms 'braided' and 'anastomosing' requires clarification; (4) the nature of the sediment load must be considered. Resolution of these issues will appear as it becomes known how the various channel patterns develop and why.

Formation of the continuum of channel patterns

Channel patterns and their associated flow and sedimentary processes form part of a continuum. This has been demonstrated in experimental studies of natural and laboratory channels (e.g. Leopold & Wolman 1957; Ackers & Charlton 1971; Schumm and Khan 1972; Ikeda 1973, 1975; Ashmore 1982, 1991*b*) and with theoretical models (e.g. Engelund & Skovgaard 1973; Parker 1976; Fredsoe 1978; Hayashi & Ozaki 1980; Blondeaux & Seminara 1985; Fukuoka 1989). As an illustration of this continuum, Fig. 1 shows how common types of channel pattern can evolve from a straight, erodible alluvial channel at constant water discharge, and from each other (see also Bridge 1985).

In the initial stages, the bed evolves towards a statistically constant geometry composed of single or multiple rows of bedwaves which are in equilibrium with the steady hydraulic and sediment transport conditions. In plan view, the water flow associated with these bedwaves follows a sinuous path, with a wavelength equivalent to that of the bedwaves in a particular row, and with a width equivalent to a bedwave width. Thus multiple rows of bedwaves have multiple rows of sinuous flow paths. Bedwave lengths are proportional to their widths (and thus to the widths of the sinuous flow paths), and their heights are comparable to flow depth. They are generally asymmetrical in alongstream cross section, may have an avalanche face on the downstream side, and generally migrate in the downstream direction.

The American Society of Civil Engineers Task Force on Bed Forms (1966) defined 'bedforms having lengths of the same order as channel width or greater, and heights comparable to the mean depth of the generating flow' as bars. The bedwaves under discussion are referred to as alternate bars because, within a given row, they occur on alternating sides of the channel with progression downstream. They are macroforms

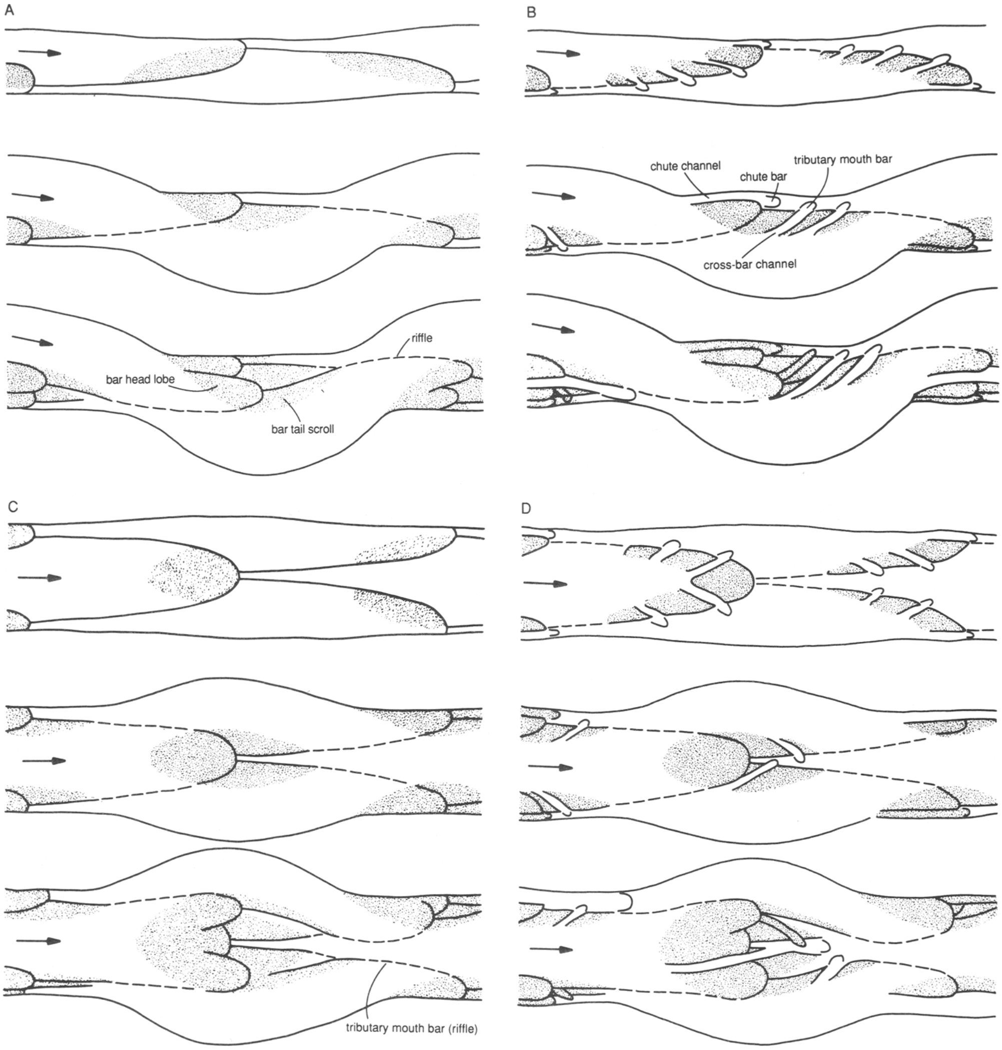


Fig. 1. Idealized evolution of channel patterns from straight alluvial channels resulting from bank erosion and channel widening. Deposition on evolving point or braid bars is shown as episodic accretion of unit bars (e.g. bar head lobes and bar tail scrolls), although such discrete features are not always present. The crestlines of unit bars (represented by solid or dashed lines) may or may not be associated with angle-of-repose avalanche faces. Arrows represent flow direction and stippled areas are topographic highs. (**A**) Single-row alternate bars (top) evolving into point bars (bottom). (**B**) As for (A) but with cross-bar channels and associated channel-mouth bars which may form at constant 'channel forming' discharge, or during falling stage. (**C**) Double-row alternate bars evolving into braid bars and point (side) bars. (**D**) As for (C) but with cross-bar channels and channel-mouth bars. See Bridge (1985) for models of the evolution of braid bars from multiple rows of alternate bars. This figure is based on numerous laboratory experiments and studies of rivers (e.g. Kinosita 1957, 1961; Krigström 1962; Stebbings 1964; Collinson 1970; Sukegawa 1970; Karcz 1971, 1972; Schumm & Khan 1972; H. Ikeda 1973, 1975, 1983; Bluck 1976, 1979; Lewin 1976; Parker 1976; Cant & Walker 1978; Ashmore 1982, 1991*b*; Mosley 1982*b*; Ferguson & Werritty 1983; S. Ikeda 1984; Jaeggi 1984; Bridge *et al.* 1986; Fujita 1989).

in the sense of Jackson (1975*b*) and Church & Jones (1982). Alternate bars or parts of them have also been referred to as unit bars, linguoid bars, side bars, transverse bars, cross-channel bars, and diagonal bars and riffles (Smith 1978; Church & Jones 1982). Some of the bedwaves which have been referred to as transverse bars or linguoid bars (e.g. Sundborg 1956; Collinson

1970; Boothroyd & Ashley 1975; Jackson 1976*b*) should probably be referred to as dunes (mesoforms) in that their equilibrium forms are apparently controlled by flow structures which scale with boundary-layer thickness (depth), and not flow width. Mesoforms such as dunes and bedload sheets are commonly superimposed on the alternate bars (macroforms), and during the initial stages of development may be the only bedwaves discernible (e.g. Fujita 1989; Ashmore 1991*b*).

The next stage in the evolution of channel patterns involves bank erosion and channel widening, which results in a drop in water level and emergence of the highest parts of alternate bars, all of which can be accomplished at constant discharge. Emergence of the highest parts of alternate bars and subsequent lee-side deposition allows recognition of three morphological units which grade into each other: bar tail; riffle; bar head (terminology partly after Bluck 1971, 1974, 1976, 1979). The units represent different positions on ancestral alternate bars.

If bank erosion is sufficiently rapid and intermittent, deposition may occur as recognizably distinct bar head and bar tail (scroll bar) units (Fig. 1). These accretionary units have been referred to as 'unit bars' (Smith 1974, 1978; Ashmore 1982, 1991*b*) and they appear to be directly analogous to the alternate bars that developed initially. Accretionary units can grow during steady flow stages in association with episodic erosion of an upstream thalweg and/or cut bank. The 'slug' of sediment thus produced migrates downstream and part of it may be deposited at a flow expansion as a discrete lobe with an avalanche face on the leading edge. If deposition is less rapid and more continuous, sheets of sediment without recognisable avalanche faces are produced.

The resulting accumulations of sediment in mid-channel or forming the banks on the inside of river bends are referred to as braid bars or point bars, respectively. Synonyms for braid bars include medial, longitudinal, crescentic, and transverse bars and sandflats, whereas point bars are also referred to as side bars and lateral bars (Smith 1978; Church & Jones 1982). Many refer to the incremental accretionary units on point or braid bars as bars also. As with the ancestral alternate bars, the basic geometry of point and braid bars is controlled by channel-forming discharge.

Bank erosion and bar deposition, still at constant discharge, create changing gradients of the bed and water surfaces, which in turn may lead to the formation of new channels which cut across braid bars and point bars (Fig. 1). Some of these channels may develop as a sheet of water starts to flow over a formerly emergent bar surface and progressively becomes concentrated into discrete channels which develop by headward erosion (e.g. Ashmore 1982). Other (chute) channels develop as the flow takes advantage of the low areas between adjacent bar head lobes or the 'slough' areas between adjacent bar tail scrolls. These cross-bar channels commonly develop their own bars (macroforms), the geometry of which is controlled by the flow and sediment transport conditions in these channels (Fig. 2). Where these channels join another channel, solitary delta-like deposits with avalanche faces commonly form (e.g. chute bars, tributary-mouth bars, Figs 1 and 2).

A cross-bar channel may be progressively enlarged at the expense of an adjacent channel, thus changing the location of the main channel segments. As such a channel develops, the previously described evolutionary sequence recurs, but in a different place. Chute cut off is an example of such behavior. In other cases a channel on one side of a braid bar may become enlarged at the expense of the channel on the other side. In this way, braided channel segments are commonly abandoned and filled, resulting in the accretion of a braid bar to the floodplain or to another bar. Channel diversions may also result in abandonment of a number of connected channel segments and bars. Previously abandoned channels are commonly reoccupied as the process of channel migration continues (Krigström 1962; Ferguson & Werrity 1983; Carson 1984*b*,*c*; secondary anastomosis of Church 1972).

Although these evolutionary patterns can be formed under constant discharge (Ashmore 1982, 1991*b*), they can all be found in natural rivers where discharge varies with time. However, some geometries arise specifically as a result of changing discharge, particularly the dissection and modification of emerging alternate, point and braid bars (Fig. 1 B, D). Such features may give a braided appearance to a river at low-flow which does not exist at high flow stage. It is therefore very important to study rivers over a large range of stage over an extended time span, because their appearance is controlled by the history of changing flow stage. Bluck (1974, 1976, 1979) assigned the formation of certain morphological features of bars to specific flow stages based on their topographic elevation and cross-cutting structures within their associated deposits. Thus the topographically high bar head was considered to be

Fig. 2. Example of a cross-bar channel through the downstream end of a side bar of the Brahmaputra River near Sirajganj. The main channel is in the far background, flowing to the right. The cross bar channel (where people are standing) has its own macroforms, and a dune-covered mouth bar (centre of photo) has built into the inner slough channel (foreground).

formed at high flow stage, whereas the lower bar tail and associated cross-bar channels were considered to form at lower flow stages. In general, the association of topographically low forms with low flow stage cannot be justified, and the high stage activity of bar tails and cross-bar channels can be demonstrated (e.g. Bristow 1987; Bridge & Gabel 1992).

In summary, the main braiding mechanisms that have been observed in laboratory and natural rivers are as follows.

(1) Development and emergence of individual or rows of alternate bars. This mechanism encompasses the mid-channel bar initiation described by Leopold & Wolman (1957) and cited in most standard texts which discuss braiding. However, initial deposition in mid-channel of relatively coarse grains is probably not associated with decrease in competence. Leopold & Wolman (1957) clearly show that the maximum bedload transport rate and grain size occurs over the crest of the developing mid-channel bar, as would be expected with an alternate bar under subcritical flow. This mechanism also encompasses the 'central bar initiation', 'transverse bar conversion' and 'multiple bar braiding' mechanisms of Ashmore (1991*b*). Bridge *et al.* (1986) and Ashmore (1991*b*) cite examples of this common type of braiding.

(2) Formation of cross-bar channels. This mechanism complicates the relatively simple patterns of braid bars and channels produced from emergence of alternate bars in (1). As clearly shown by Ashmore (1982, 1991*b*) this mechanism does not require changes in flow stage. However, there are numerous examples in the literature where channels cutting across alternate, point or braid bars are associated with falling flow stage (e.g. Krigström 1962; Collinson 1970; Smith 1970, 1971*a*, 1974; Hein & Walker 1977; Rundle 1985*a*,*b*). A variant of this mechanism is the chute cut off of point bars or single-row alternate bars (Ashmore 1991*b*). Chute cut off (and the equivalent cutting of a new channel through the central upstream part of a braid bar) is enhanced by the deposition of bar head lobes and lack of filling of the low areas between these lobes, as discussed previously. Chute cut-off of point bars has been described by many workers (e.g. Friedkin 1945; Kinoshita 1957; Krigström 1962, Hickin 1969; Ikeda 1973; Hong & Davies 1979; Ashmore 1982, 1991*b*; Teisseyre 1977*b*; Ferguson & Werrity 1983; Bridge, 1985; Bridge *et al.* 1986, Lewin 1976; Carson 1986). There is no reason why cross-bar channels could not develop a braided pattern also. If a cross-bar channel is enlarged it may become very difficult to distinguish a cross-bar channel from a 'main channel'.

In subsequent sections, details of the

geometry, flow and sedimentary processes in braided rivers will be considered. However, first it is necessary to discuss the implications of the braiding mechanisms above to the description and classification of channel patterns, and their hydraulic controls.

Description and classification of channel geometry in plan

Leopold & Wolman's (1957) classification of channel patterns as straight, meandering and braided is unsatisfactory because the classes are not mutually exclusive and different parameters are used to define the different patterns (Chitale 1970; Kellerhals *et al.* 1976; Rust 1978*a*; Knighton 1984). It is necessary to consider at least the nature of channel splitting around braid bars or islands, and the sinuosity of the channel(s) in a descriptive classification of channel patterns (Kellerhals *et al.* 1976; Rust 1978*a*; Brice 1984).

Bars or islands?

Brice (1964) defined mid-channel bars as being unvegetated and submerged at bankfull stage, whereas islands are vegetated and emergent at bankfull stage. The degree of development of vegetation on mid-channel bars is related to the amount of time the bar surface has been exposed above the seasonal low-water mark, the nature of the sediment surface exposed, and the types of vegetation available for colonization. These are in turn controlled by the history of erosion and deposition, the sediment available to the river, and the climate. Depending on these factors, freshly emergent unvegetated bars may become progressively vegetated as they accrete vertically and laterally (e.g. Brice 1964; Bridge *et al.* 1986), thus it is difficult to assess when a 'bar' becomes an 'island'. Such a distinction also artifically separates depositional forms which may have a common geometry and genesis.

Brice (1964) further classified bars as 'transient' as opposed to 'stabilized' islands. These terms are analogous to the terms 'unstable' and 'stable' which are used by geomorphologists and sedimentologists to imply the degree of erosion and deposition within the channel and the rate of channel migration. None of these terms have been defined objectively, and there is every gradation in the rates of channel and bar movement among different rivers, irrespective of whether the sediment surfaces are vegetated. Therefore, terms such as 'transient', 'unstable' and 'stable' should be replaced with quantitative measures of the lifespans, rates of creation, migration and destruction of bars.

Stage dependence of channel pattern

Bars generally exist for time periods in excess of a floodwave, and will therefore have experienced a complex history of erosional and depositional modification related partly to stage changes. Although the overall form of bars is controlled by near-bankfull flow patterns, certain geometrical features are related to falling flow stages (e.g. some cross-bar channels). It is essential to recognise that such falling stage features are genetically different from the overall, high-stage generated, bar morphology (see also Collinson 1970; Rust 1978*a*; Bristow 1987). Failure to do so will lead to errors in definition of the nature of braiding. Brice (1964) recognized that his 'transient' braiding index depends on flow stage, but the 'stabilized' index does not.

Ideally bars should be observed at their formative high discharges when falling-stage features are not present; however, this is very difficult to do. Kellerhals *et al.* (1976) recommended describing channel geometry at mean flow (noting also high and low flow geometries) because these have a high probability of occurring and being recorded on aerial photographs (see also Howard *et al.* 1970). Rust (1978*a*) similarly recommended definition of channel patterns at a stage approximately mid-way between bankfull and minimum discharge.

Hierarchies of bars and channels

Upon deciding at which flow stage(s) a channel pattern will be described, it is necessary to consider which of the channel segments and bars will be used to define the nature of channel splitting and channel sinuosity. For instance, should the relatively minor cross-bar channels be considered along with the relatively major channels around the main braid bars, even if all of the channels are active at the same flow stage? There appears to be no simple answer to this question in that any cross-bar channel may evolve into a 'main channel'. This dilemma has led to various attempts to assign different orders to channels and bars in multiple-channel rivers (e.g. Williams & Rust 1969; Rust 1978*a*; Bristow 1987; Fig. 3 A, B). Then description of channel geometry is referred to specific orders of channel.

The existing channel and bar ordering schemes are difficult to apply and are not defined consistently. For instance, in Rust's

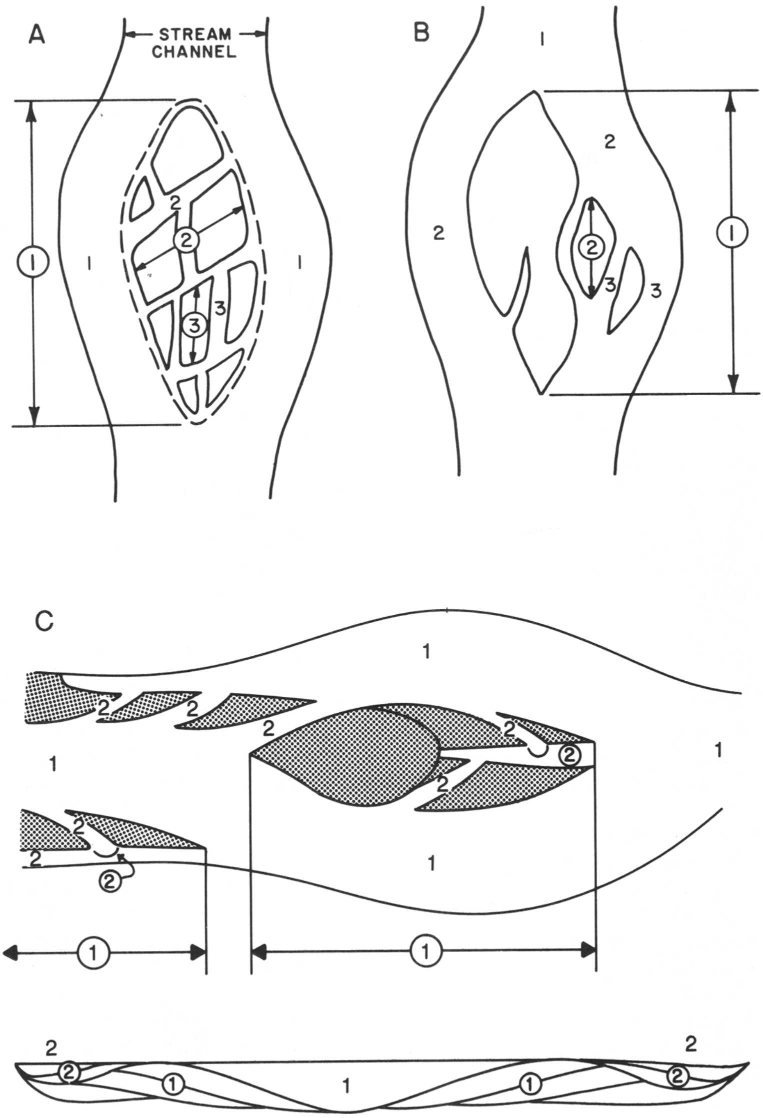


Fig. 3. Channel and bar ordering schemes of (**A**) Williams & Rust (1969), and (**B**) Bristow (1987). (**C**) Alternative channel and bar ordering scheme. Numbers in circles refer to bars, other numbers refer to channels. Cross section (lower figure) is from a confluence region, where a central channel (1) is bounded by side bars with cross bar channels (2).

scheme the difference between the second- and third-order (cross-bar) channels is not clear, and the second- and third-order bars are not bars in the true sense of the word; they are dissected segments of first order bars. Indeed, second- and third-order channels may have their own within-channel bars. In Bristow's scheme, all cross-bar channels are third-order because the main channels can be either first or second order. Really, the main channels that flow adjacent to and over the largest scale of bars in the river should all be of the same order. A simpler, compromise scheme is shown in Fig. 3 C, where the largest scales of bar and adjacent channels are first order (following precedent), but all channels cutting across these bars are second order. The segments of first-order bars bounded by the second-order channels are not second-order bars. They *may* be partly eroded depositional units (bar heads, bar tail, scrolls etc.) associated with episodic deposition on first-order bars. Second-order bars are those that form within and at the terminations of second-order channels. It should be remembered that second-order channels may evolve into first-order channels.

Distinction between 'braiding' and 'anastomosing'

The term 'braiding' is generally taken to mean the splitting of channels around bars (islands).

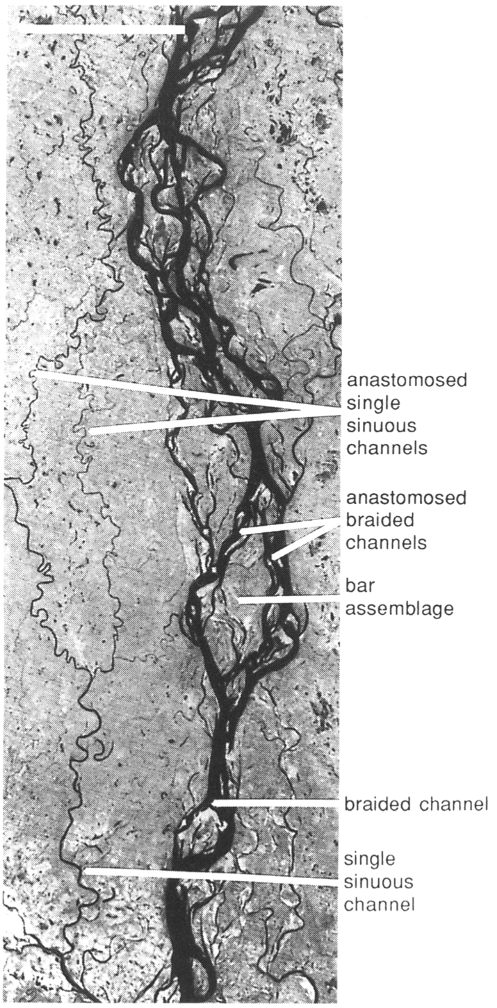


Fig. 4. Landsat photograph (courtesy of C. S. Bristow) of the Brahmaputra River immediately north of its confluence with the Ganges showing various kinds of channel pattern. Scale bar is 20 km.

A different type of channel splitting has been recognized and referred to as anastomosing (Lane 1957; Smith 1976) or anabranching (Brice 1964, 1984). The characteristic, definitive feature of anastomosing (anabranching) channel segments is that they are longer than a curved channel segment around a single braid or point bar and their width-scale flow patterns behave substantially independently of adjacent segments, in contrast to braided channel segments around bars (e.g. Fig. 4). Thus anastomosed channel segments contain their own bars in accordance with imposed discharge and sediment load, enabling definition of braiding index and sinuosity for each segment. Anastomosing is therefore more akin to terms like distributive and tributive (e.g. Schumm 1985).

The classification of a river as anastomosing is clear cut if, for instance, the individual segments are undivided, sinuous channels separated by areas of floodplain much larger than the largest bars present (e.g. Fig. 4). However, the distinction between anastomosing and braiding is not as clear cut in braided rivers where, for instance, some second-order channel segments may have the characteristics of anastomosed segments as defined above, but first-order segments do not. A solution to this problem is to define anastomosing channels as those where the length of channel segments exceeds the length of first order channels around individual first-order bars. Nevertheless, many braided rivers appear to be both braided and anastomosing (Fig. 4). In the Brahmaputra, anastomosis is associated with bar assemblages (Coleman 1969; Bristow 1987; Fig. 4), which are analogous to what Church & Jones (1982) refer to as megaforms or sedimentation zones. Such anastomosed reaches are commonly taken to be associated with relatively large deposition rates (Smith 1976, 1983; Smith & Smith 1980; Rust 1981; Church & Jones 1982). As the land areas between anastomosed channel segments are generally not braid bars, the term anastomosing cannot be used to define channel patterns based on braiding index, as done by Rust (1978*a*) and Miall (1981).

The degree and style of braiding

Measures of the degree of braiding (see Table 1) generally fall into two categories: those that consider the mean number of active channels or braid bars per transect across the channel belt, and those that consider the ratio of the sum of channel lengths in a reach to a measure of reach length (referred to here as 'total sinuosity'). The first type of braiding index is more desirable for

Table 1. *Braiding indices*

Author	Braiding index
Brice (1960, 1964)	$\dfrac{2\text{ (sum of lengths of bars or islands in a reach)}}{\text{centreline reach length}}$
Howard, Keetch & Vincent (1970)	Average number of anabranches bisected by several transects perpendicular to flow direction
Engelund & Skovgaard (1973), Parker (1876), Fujita (1989)	Mode = number of rows of alternate bars (and sinuous flow paths) = 2 times the number of braid bars and number of side (point) bars per transect
Rust (1978*a*)	Number of braids per mean curved channel wavelength = mode − 1 (see above)
Hong & Davies (1979)	Total sinuosity $= \dfrac{\text{length of channel segments}}{\text{channel belt length}}$ Number of braids or channels in cross-section
Mosley (1981)	Braiding index $= \dfrac{\text{total length of bankfull channels}}{\text{distance along main channel}}$
Richards (1982), Robertson-Rintoul & Richards (this vol.)	Total sinuosity $= \dfrac{\text{total active channel length}}{\text{valley length}}$
Ashmore (1991*a*)	Mean number of active channels per transect. Mean number of active channel links in braided network.
Friend & Sinha (this vol.)	Braid channel ratio $= \dfrac{\text{sum of mid-channel lengths of all channels in reach}}{\text{length of mid-line of widest channel}}$

two main reasons. First of all it is related to the 'mode' of (ancestral) alternate bars. In order to express the degree of braiding in terms of 'mode' it is necessary to count point (side) bars as well as braid bars. Secondly, the total sinuosity is a combined measure of channel-segment sinuosity and degree of braiding. Thus braided rivers with relatively large numbers of channel segments of low sinuosity can have a similar total sinuosity to those with fewer, higher sinuosity channel segments (Fig. 5). Therefore it is desirable to determine separately the braiding index and average sinuosity of curved channel segments around bars (see below).

Brice's (1960, 1964) braiding index (Table 1) is a measure of the sum of bar or island perimeters relative to reach length and is strongly dependent on flow stage. More recently, Brice (1984; see also Brice *et al.* 1978) abandoned his braiding index and classified the *degree* of braiding as the proportion of the channel length in a each that is divided by bars and islands and the *character* of braiding in terms of whether the bars or islands are dominant and the plan shapes of islands. The degree of braiding cannot be compared with the braiding indices discussed above, and the character of braiding cannot be defined objectively. Furthermore, in their classification of channel patterns, anabranched channels (where anabranching is arbitrarily defined as channel splitting where island width is greater than three times water width at average discharge) cannot clearly be distinguished from braided channels. Brice (1984) uses terms such as 'locally braided' and 'generally braided' based on arbitrary values of the degree of braiding. Rust (1978*a*) suggested the terms 'moderately braided' and 'highly braided' based on arbitrary values of his braiding parameter.

In Kellerhals *et al.*'s (1976) classification of channel splitting, the spatial distribution of islands in a reach is described but that of bars is not, and the shape of bars is described but not that of islands. Use of terms such as 'occasional', 'frequent', 'split', 'braided', to describe the distribution of islands mixes morphological terms with those having a time connotation. Some categories are objectively defined; others are not. Also, the terms do not have unambiguous meanings, in that 'braided' and 'split' could equally apply to all divided reaches.

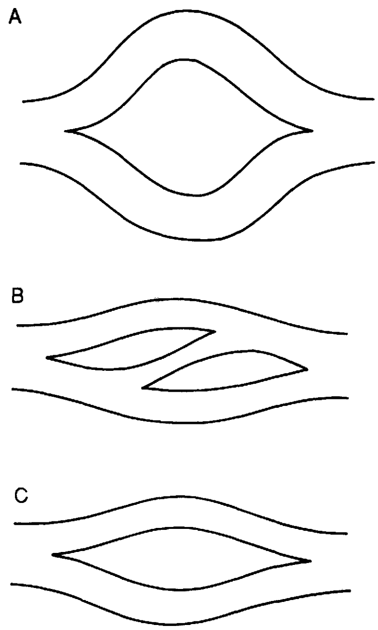


Fig. 5. Relationship between total sinuosity (ΣP), braiding index (Br) and mean sinuosity of channels (sn), where $\Sigma P = Br.sn$. (**A**) and (**B**) have the same ΣP but different sn and Br. (**A**) and (**C**) have the same Br but different sn and ΣP. (**B**) and (**C**) have the same sn, but different Br and ΣP.

Sinuosity of multiple-channel rivers

Sinuosity is defined as either channel thalweg length/valley length (Leopold & Wolman 1957; Rust 1978*a*), channel length/valley length (Brice 1984; Schumm 1985), or channel length/channel-belt axis length (Brice 1964), the latter being useful for defining channel-belt sinuosity. Definition of sinuosity for multiple-channel rivers, as suggested by Rust (1978*a*), requires definition of some average thalweg length of channel segments of a given order at a given flow stage. Friend & Sinha (1993) use the centreline length of the widest channel in a braided river to define sinuosity. Based on precedent, Rust (1978*a*) used a mean sinuosity of 1.5 to separate 'low sinuosity' from 'high sinuosity' rivers. Brice's (1984) classification of braided rivers into 'sinuous' and 'non-sinuous' is not based on objective criteria.

Hydraulic controls of channel pattern

Empirical approaches using discharge, slope and bed material size

There are many myths about what controls channel pattern. It is generally accepted that they are controlled by the supply of water and sediment and, in the short term, by valley slope. Some have also invoked bank erodibility, especially as influenced by vegetation. Empirically, the degree of braiding (i.e. braiding index) and overall channel width/depth increase as water-discharge is increased for a given slope, or as slope is increased for a given discharge (e.g. Lane 1957; Leopold & Wolman 1957; Howard *et al.* 1970; Ackers & Charlton 1971; Schumm & Khan 1972; Chitale 1973; Mosley 1981; Richards 1982; Ashmore 1991*a*). However, channel pattern also varies with bed-material size, such that braiding occurs at lower slopes and/or discharges as grainsize decreases (Henderson 1961, 1966; Osterkamp 1978; Carson 1984*a*; Ferguson 1984, 1987; Ferguson & Ashworth 1991). Positive correlations between width/depth and water discharge are shown by Chitale (1970) and Leeder (1973). Width/depth is also strongly negatively correlated with sinuosity for streams of a given discharge (Schumm 1963; Chitale 1970) except for those near the transition from straight to sinuous, undivided channels.

Discrimination between different channel patterns using discharge (Q) slope (S) and bed-material size (D) is generally poorer if a large range of experimental data is used than if a smaller, specific data set is used (Ferguson 1984, 1987; Carson 1984*a*). This is partly due to different definitions of the experimental parameters and partly due to inappropriate choice of parameters. Discharge measures used have included the mean annual, bankfull, and mean or median annual flood. The most appropriate discharge to use is a flood discharge that does not depend on a reference to channel geometry, such as a frequency-based measure like the one or two year flood (Ferguson 1987; Carson 1984*a*). Valley slope should be used instead of channel slope, to avoid biasing the plotting positions of sinuous streams which have a lower channel slope than valley slope. In general, the definition of channel patterns has not been objective, and different workers have classified rivers near the meandering-braided transition in different ways.

Discharge variability

It is commonly held that channel geometry of alluvial rivers is dominated by flow and sedimentary processes operating over a range of high discharges. At seasonally low flows, sediment transport rates are relatively diminished and modification of the high-stage adjusted channel pattern is expected to be minimal. However, the increases in discharge, width/depth, and sediment transport associated with exceptional floods may precipitate a major change in channel pattern (e.g. increasing degree of braiding; Schumm & Lichty 1963) which may remain more or less intact for many years. Subsequent, long-term (many flood periods) reduction in flood discharge may eventually lead to modification of channel pattern (e.g. from braided to undivided; Werritty & Ferguson 1980; Ferguson & Werritty 1983; Schumm 1985). It is therefore important to be aware of the history of major flood events when assessing the hydraulic controls on channel pattern (see also Howard *et al.* 1970; Mosley 1981). Discharge variability does not exert a major influence on the existence of the different channel patterns because they can all be formed in laboratory channels at constant discharge, and many rivers with a given discharge regime show alongstream variations in pattern. That braided channels are associated with greater discharge variability than undivided channels is one of the myths. However, as demonstrated below, discharge variability does influence the detail of sediment transport, erosion and deposition in braided rivers and all other alluvial river types.

Discharge, slope, bed material size and sediment transport rate

Discriminators between unbraided and braided rivers of the form $S = aQ^{-b}$ (see Table 2) have been interpreted to represent a constant value of a hydraulic property at the channel pattern threshold. Parameter a controls the constant value of the hydraulic property, and the hydraulic property depends on the value of the exponent, i.e. $b = 0.33$ for bed shear stress, $b = 0.5$ for stream power per unit bed area, $b = 1.0$ for stream power per unit channel length (Antropovskiy 1972; Ferguson 1981, 1984, 1987; Begin 1981; Begin & Schumm 1984; Carson 1984*a*). The stream power per unit channel length also correlates weakly with braiding index (Howard *et al.* 1970; Mosley 1981) and total sinuosity (Richards 1982; Robertson-Rintoul & Richards this volume). These hydraulic measures are all known to be related to sediment transport rate, but this also depends on what type of sediment is available and what proportion of the total bed shear stress or power is available to transport sediment in the presence of rough banks and bedforms. Indeed, parameter a above must include these effects and therefore cannot be constant.

Table 2. *Hydraulic controls on braided channel patterns: discharge, slope and bed material**

Equation	Comments	Author
$S = 0.0007Q_m{}^{-0.25}$	Meandering sand-bed channels	Lane (1957)
$S = 0.0041Q_m{}^{-0.25}$	Braided sand-bed channels	
$S = 0.0125Q_{bf}{}^{-0.44}$	Meandering $\rightarrow$ braided	Leopold & Wolman (1957)
$S = 0.000196D^{1.14}Q_{bf}{}^{0.44}$	Meandering $\rightarrow$ braided	Henderson (1961, 1966)
$S = 1.4Q_{maf}{}^{-1}$	Meandering $\rightarrow$ braided	Antropovsky (1972)
$S = 0.0009Q_m{}^{-0.25}$	Mainly meandering sand-bed rivers in Kansas	Osterkamp (1978)
$S = 0.0017Q_m{}^{-0.25}$	Braided sand bed rivers in Kansas	
$S = aQ_m{}^{-0.25}$	Meandering $\rightarrow$ braided	
$S = 0.0016Q_m{}^{-0.33}$	Meandering $\rightarrow$ braided	Begin (1981)
$S = 0.07Q_{2f}{}^{-0.44}$	Sinuosity > 1.25 and meandering $\rightarrow$ Braided for gravel-bed rivers	Bray (1982)
$S = 0.042Q^{-0.49}D_{50}{}^{0.09}$	Meandering $\rightarrow$ braided for gravel-bed	Ferguson (1984, 1987)
$S = 0.042Q^{-0.49}D_{90}{}^{0.27}$	rivers†	
$S = 0.0049Q^{-0.21}D_{50}{}^{0.52}$	Meandering $\rightarrow$ braided using Parker's theory and hydraulic geometry	
$S \approx aQ^{-0.5}D^{0.5}$	Meandering $\rightarrow$ braided	Chang (1985)
$\Sigma P = 1 + 5.52(QS_v)^{0.38}D_{84}{}^{-0.44}$	Gravel-bed rivers	Robertson-Rintoul & Richards (this vol.)
$\Sigma P = 1 + 2.64(QS_v)^{0.4}D_{84}{}^{-0.14}$	Sand-bed rivers	

* S.I. Units.
† D in mm.

Discriminators of the form $S = aQ^{-b} D^{c}$ (Table 2), where D is some measure of bed sediment size, explicitly recognize at least one aspect of sediment supply (and erodibility of bed and banks) but a is still not likely to be constant. Values of $b = 0.33$ and $c = 1$ imply that the theshold between braided and unbraided rivers occurs at a constant value of dimensionless bed shear stress if a is constant (Begin 1981; Ferguson 1986, 1987; Carson 1984*a*). Henderson's (1961) approach with $b = 0.46$ and $c = 1.15$ is based on the stability of channels at the threshold of bedload motion and therefore cannot be correct in view of the requirement of sediment transport for channel bars to form.

The stream power per unit channel length (ϱgQS), where ϱ is fluid density and g is gravitational acceleration, controls the width-integrated sediment transport rate, depending on the grain sizes available for transport and the bedform drag. Thus an increase in QS for a given available sediment should be associated with an increased total sediment transport rate and braiding index. However, in natural rivers it is common for bedload grain size to increase with QS, whereas changes in bedform drag are more difficult to predict (e.g. Leopold & Wolman 1957; Osterkamp 1978; Prestegaard 1983; Ferguson & Ashworth 1991). If bedload grainsize and width/depth increase as QS increases, sediment transport rate per unit bed area is likely to be much more conservative than total sediment transport rate. If grainsize is held constant as in laboratory experiments an increase in QS will result inevitably in an increase in sediment transport rate per unit bed area unless the banks can be eroded, thereby resulting in an increase in width/depth and a change in channel pattern. Thus, as total sediment transport rate must equal sediment supply in an equilibrium river channel, an abrupt change in sediment supply to a reach of a river may result in a change in channel pattern, an increase in supply tending to induce braiding (e.g. Smith & Smith 1984). Ashmore (1991*a*) has documented periodic changes in braiding index due to periodic changes in sediment supply while discharge, slope and bedload size remained constant. Also, Hoey & Sutherland (1991) associate increase in braiding index with periodic aggradation (sediment storage in bars), and vice versa during degradation.

Empirical approaches using other parameters

Japanese empirical approaches to the hydraulic controls on the mode of alternate bars and braids are summarized by Hayashi & Ozaki (1980), Fukuoka (1989), and Fujita (1989). These approaches fall into two main groups, based on different combinations of dimensionless quantities: u_*/u_{*c} (flow intensity) and wS/d (channel form index); d/D and w/D (Table 3). Here, u_* is shear velocity, u_{*c} is shear velocity at the threshold of bedload movement, w is channel width, and d is mean flow depth. Ikeda's (1973) criterion for braiding (Table 3) was based on laboratory experiments. Ikeda's (1975) natural river data agree reasonably well with the wS/d criterion but do not appear to show a clear dependence on u_*/u_{*c}. As u_*/u_{*c} is proportional to $(dS/D)^{1/2}$, this criterion involves essentially the same variables as the approaches mentioned above, since w and d can be expressed in terms of discharge. Although this criterion correctly predicts the transition to braiding as w/d and S increase, it incorrectly predicts that, for a given channel-form index, braiding is favoured by a decreasing u_*/u_{*c}, hence sediment transport rate. The alternative braiding criterion has no dependence on slope, but otherwise the controlling variables are the same as those used by Ikeda. Chien's (1961) braiding criterion is based on multiple regression analysis of data from Chinese sandy rivers, but the reasons for the choice of variables is not given.

Channel patterns and bank stability

It has been suggested (Schumm 1963, 1971, 1972, 1977, 1981, 1985; followed by many sedimentologists) that rivers which transport large proportions of bedload relative to suspended load tend to have relatively low sinuosity and high braiding index. Such 'bedload streams' were associated with relatively easily eroded banks of sand or gravel, large channel slope and stream power, such that they were laterally 'unstable'. In contrast, rivers with relatively large suspended loads were postulated to be characteristic of undivided rivers of higher sinuosity. Such 'suspended load' streams were associated with cohesive muddy banks, low stream gradient and power, and lateral stability. However, the correlation between channel sediment size, type of sediment load and channel pattern is not generally supported, as recognized by Schumm (1981, 1985) in his more recent classifications of channel pattern (but unfortunately still not recognized by many sedimentologists). In fact, many braided rivers are sandy and silty and many single-channel, sinuous rivers are sandy and gravelly (Rust 1978*b*; Jackson 1978; Bridge 1985). Braided

Table 3. *Hydraulic controls on braided channel patterns: non-dimensional criteria*

Equation	Comments	Author
$\frac{U_*}{U_{*c}} = 1.4\left(\frac{wS}{d}\right)^{1/3}$	Meandering → braided	Ikeda (1973, 1975)
$\frac{(w/D)^{2/3}}{(d/D)} = 6.7$ for $1 < \tau_0/\tau_c < 12$	Meandering → braided	Muramoto & Fujita (1977)
$\frac{(w/D)^{2/3}}{(d/D)} = 3.5$ *to* 6.7; $2.2m^{2/3} < \frac{(w/d)^{2/3}}{(d/D)} < 6.7m^{2/3}$	Meandering → braided; *m* is mode (degree of braiding)	Fujita (1989)
$\left(\frac{\Delta Q}{0.5TQ_{bf}}\right)\left(\frac{d_{bf}S}{D_{35}}\right)^{0.6}\left(\frac{Q_{\max}-Q_{\min}}{Q_{\max}+Q_{\min}}\right)^{0.6} \cdot \left(\frac{w_{rf}}{w_{bf}}\right)^{0.45}\left(\frac{w_{bf}}{d_{bf}}\right)^{0.3} = 5$	Transitional → braided*	Chien (1961)

* First term is dimensional, units days^{-1}.

channels have even been formed in muddy sediments on floodplains (Rust & Nanson 1986; Nanson *et al.* 1986), although the mud was probably transported in the form of sand-sized pellets.

Vegetation helps stabilize cut banks and bar surfaces (increasing tensile and shear strength) given adequate time and conditions for development (e.g. Brice 1964; Smith 1976; Witt 1985). Such stabilization allows the existence of relatively steep cut banks and may hinder lateral migration of channels. However, there is no conclusive evidence that vegetation has an important influence on the equilibrium channel pattern (contrary to the view held by Brice 1964). Early lithification of bank sediments (e.g. calcretes, silcretes) may have an effect on bank stability similar to vegetation (e.g. Gibling & Rust 1990).

Theoretical stability analyses

The theoretical stability analyses of Parker (1976) and Hayashi & Ozaki (1980) predict that channel pattern types are controlled by width, depth, slope and Froude number (*Fr*). Parker's (1976) criterion for braiding is $w/d \approx Fr/S$, which Ferguson (1984, 1987) reformulated as $S = Q/w^2(gd)^{1/2}$. Equations for *w* and *d* in terms of *Q* can then be substituted in order to produce the empirical criterion $S = aQ^{-b}$ where *a* must depend at least on bed-material size. However, even when Parker's (1979) hydraulic geometry equations for gravel-bed rivers are used, this braiding criterion does not agree very well with field data. Parker's (1976) braiding criterion also does not agree with data from the Calamus River (Bridge & Gabel 1992) and that given in Hayashi & Ozaki (1980, pp. 7-28 and 7-29). Hayashi & Ozaki's (1980) braiding criterion, $2(wS/d)^{1/2} \approx Fr$, does agree with Calamus River data. It can be reformulated as $S = Q^2/4gd^2w^3$.

The theoretical analyses of Engelund & Skovgaard (1973), Fredsoe (1978) and Fukuoka (1989) all indicate that the major control on braiding is *w*/*d* (being >50 for braiding to occur), with θ or (θ/θ_c) having a minor effect. θ is dimensionless bed shear stress, and θ_c is the value of θ at the threshold of bedload movement. In Fredsoe's analysis the type of bedform (i.e. flow resistance coefficient) also has a minor effect, and in Fukuoka's analysis slope has a minor effect (Table 4). Hayashi & Ozaki's braiding criterion can be recast as $w/d \approx 2/f$, where *f* is the Darcy-Weisbach friction coefficient. This agrees with Fredsoe's braiding criterion if *f* takes a value of approximately 0.04, and braiding can occur at lower *w*/*d* if flow resistance is increased. An unfortunate aspect of all of these theoretical approaches is that the main controlling variables (*w*,*d*) are not independent, but depend on the supply of water and sediment.

Table 4. *Hydraulic controls on braided channel patterns: theoretical stability analyses*

Equation	Comments	Author
$S/Fr \approx d/w$	Meandering → braided	Parker (1976)
$w/d \approx 50$	Meandering → braided Weak dependence on θ and f	Fredsoe (1978)
$2(wS/d)^{0.5} \approx Fr$	Meandering → braided	Hayashi & Ozaki (1980)
$\frac{S}{Fr^2}\left(\frac{w}{d}\right)^2 f(\theta) = constant$	Meandering → braided	Struiksma & Klaasen (1988)
$S^{0.2}w/d \approx 10$ to 20	Meandering → braided Weak dependence on θ/θ_c	Fukuoka (1989)

Minimum energy theories

Several approaches to the meandering-braided transition involve the concept of minimization of energy expenditure in transporting sediment (e.g. Kirkby 1972, 1977, 1980; Chang 1979, 1985; Bettess & White 1983; reviewed by Ferguson 1987). Both Chang and Bettess & White assume that discharge, valley slope and sediment supply are independent variables and then use regime theory (e.g. equations for flow resistance and sediment transport rate) to predict equilibrium channel slope, width, depth, and flow velocity. However, to predict channel pattern, the additional assumption of minimum stream power per unit channel length (or minimum S for a given Q) is also required.

According to Bettess & White, if the channel slope (S) equals the valley slope (S_v) the channel remains straight. If $S > S_v$ aggradation will occur in a non-equilibrium channel, whereas if $S_v > S$ the river is either meandering or braided. A braided pattern allows the slope of individual channels to approach the valley slope because the slope of small multiple channels is greater than that for a single large channel. Under these circumstances the choice between meandering and braiding is based on minimum stream power per unit channel length. Unfortunately, quantitative predictions with the model do not agree very well with natural data.

Chang's (1979) approach for sand-bed rivers results in four regions in plots of S against Q: (1) straight, (2) straight braided, (3) straight braided to meandering, (4) meandering to steep braided. Chang (1985; Fig. 6) slightly modified the positions and names of these four regions in

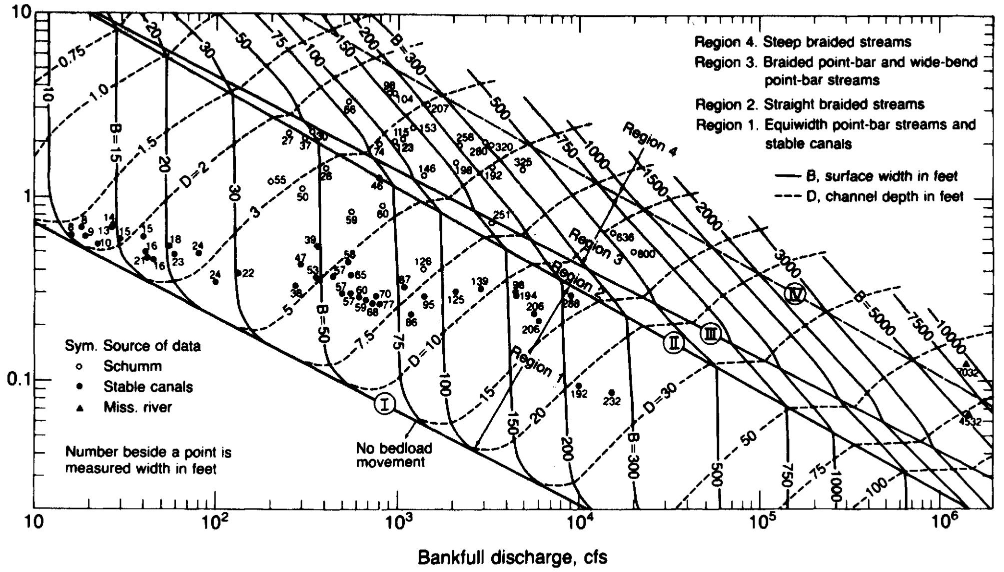


Fig. 6. Chang's (1985) theoretical prediction of channel patterns.

plots of $S/D^{1/2}$ versus Q: (1) stable channel and equiwidth point-bar streams, (2) straight braided, (3) braided point bar and wide-bend point bar streams, (4) steep braided. The transition from regions (1) to (4) as Q or S increases is clearly not as normally observed, the most glaring anomaly being region (2). The anomaly is probably based on Chang's criterion for braiding which is that an increase in S is associated with a large increase in w/d. Remarkably, however, Chang's (1985) region boundaries are lines of approximately equal w/d, and the middle of region (3) (which apparently marks the transition from meandering to braided rivers) has a w/d of 50 which is in close agreement with Fredsoe's (1978) criterion for braiding. Thus, if the discrimination between channel patterns is essentially in terms of w/d (as suggested by other theories) then the crux of the issue is finding out what independent variables control w and d. Chang's (1985) channel pattern discriminant lines have the approximate form $S \propto Q^{-0.5} D^{0.5}$, suggesting that the braiding theshold for sand-bed streams corresponds with some grain-size dependent threshold stream power per unit bed area (cf. Ferguson 1987). However, the exponents in this type of relationship are not likely to apply to gravel-bed streams, and there is much overlap in stream power per unit bed area for natural meandering and braided rivers (Carson 1984*b*,*c*; Ferguson 1987).

Geometry of braided rivers at the bar scale

Studies of the interaction between channel geometry, flow and sediment transport over a large discharge range for braided rivers are extremely rare. Those studies that do exist are only for the case of simple plan geometries. The essential features of flow in braided rivers are curved channel segments joined by zones of flow convergence (confluence) or divergence (diffluence). Flow around one side of a braid bar can be considered to be dynamically similar to flow in an undivided sinuous channel (Allen 1968, 1983; Bridge 1985; Bridge & Gabel 1992), and flow in river confluences can be considered to be dynamically similar to braided channel convergence zones (Best 1986). Therefore, it is possible to piece together a picture of the 3D geometry, flow and sediment transport at the bar scale using limited field and laboratory data from braided rivers (e.g. Straub 1935; Abdullayev 1973; Ashworth & Ferguson 1986; Bridge & Gabel 1992; Ferguson *et al.* 1992; Ashworth *et al.* 1992*a*,*b*), the more extensive experimental data from undivided curved channels (e.g. Bridge & Jarvis 1982; Dietrich & Smith 1983, 1984; Jackson 1975*a*, 1976*a*,*b*) and confluence zones (e.g. Best 1986, 1987, 1988; Roy & Bergeron 1990; Roy & De Serres 1989; Ashmore *et al.* 1992) and some theoretical reasoning (e.g. Bridge 1992).

Channel geometry associated with multiple-row alternate bars at constant discharge is illustrated in Fig. 7, as a reference for comparison with the geometry of braided channels that may have developed from such alternate bars. When considering the geometry of braided channels it must be remembered that the channels may be developing and widening, or becoming abandoned and filling. Channel geometry may also change cyclically, as opposed to progressively, in response to cyclical discharge variations. Therefore, the geometry may not necessarily be in equilibrium with the flow and sediment transport at any particular flow stage. Nevertheless, there are clearly systematic and understandable variations in geometry that can be related to certain discharges.

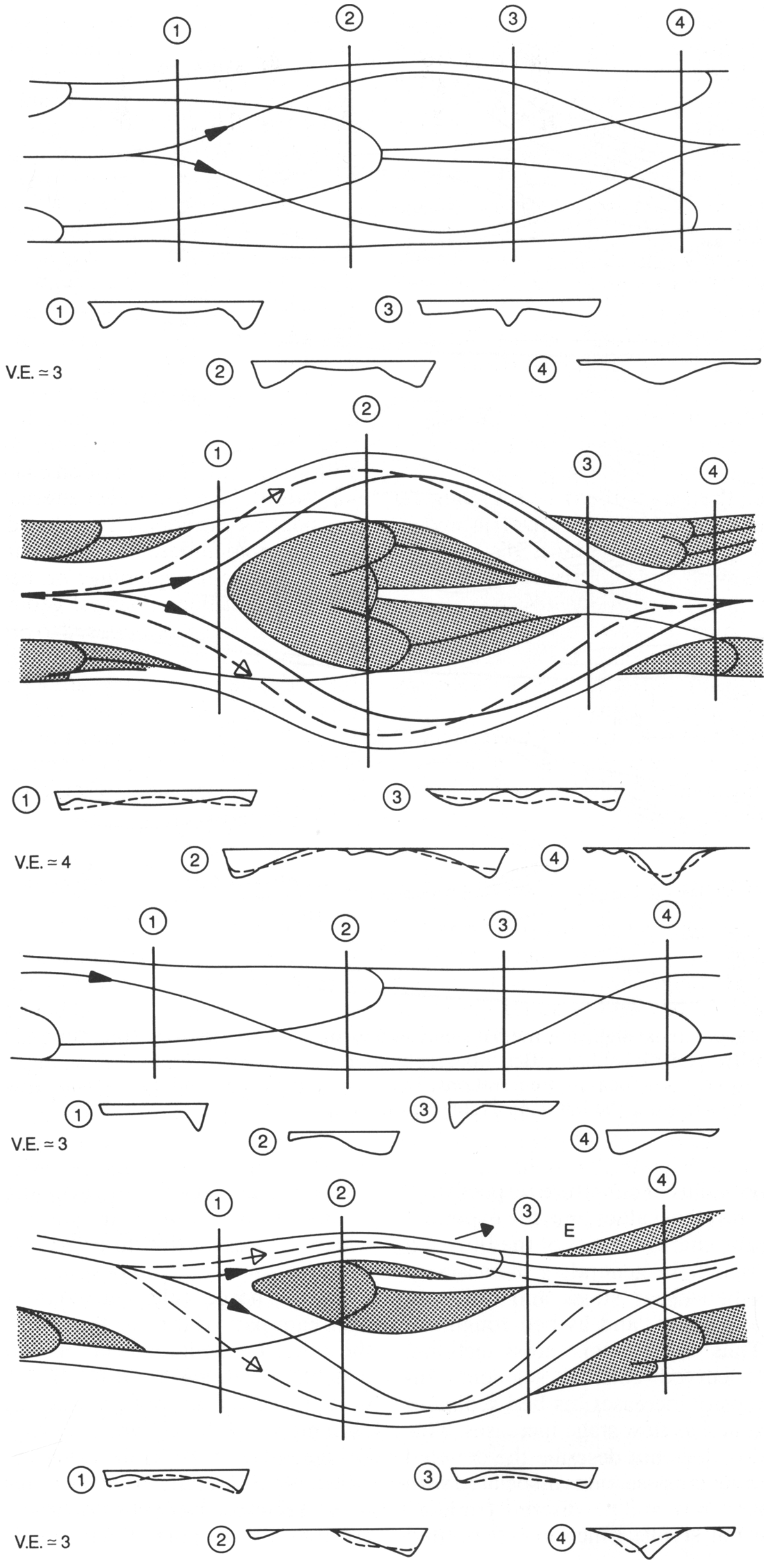
Cross-sectional geometry of curved channel segments

Upstream of a typical braid bar a mid-stream thalweg splits, and the loci of maximum depth move towards the outer, cut banks with progression around the braid bar, depending in detail on flow stage (Figs 7 and 8). At high flow stage the channel segments near the upstream tip of the bar are deepest close to the bar, whereas further downstream they are deepest near the outer cut banks. The depth variation in a cross-section within a bend can be approximated by

$$d_l/d_c = (r_l/r_c)^a \qquad (1)$$

as long as the geometry is in equilibrium with the flow (details in Bridge, 1992). Here d_l and r_l are

Fig. 7. Idealized channel geometry for two simple braided channel patterns. Equivalent alternate bar patterns in straight channels are shown (upper diagrams) for comparison. Bankfull cross sections are smoothed (no mesoforms or cross-bar channels shown) and have vertical exaggerations (V.E.) of approximately 3 or 4. Dashed lines in cross sections represent low-flow stage geometry. Lines with solid or open arrows represent loci of maximum flow velocity for high and low flow stage, respectively. Single arrow and E represent potential locations of flow diversion and erosion in response to encroachment of the tributary bar (riffle) of the major channel.


V.E. ≃ 3
V.E. ≃ 4
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E
V.E. ≃ 3

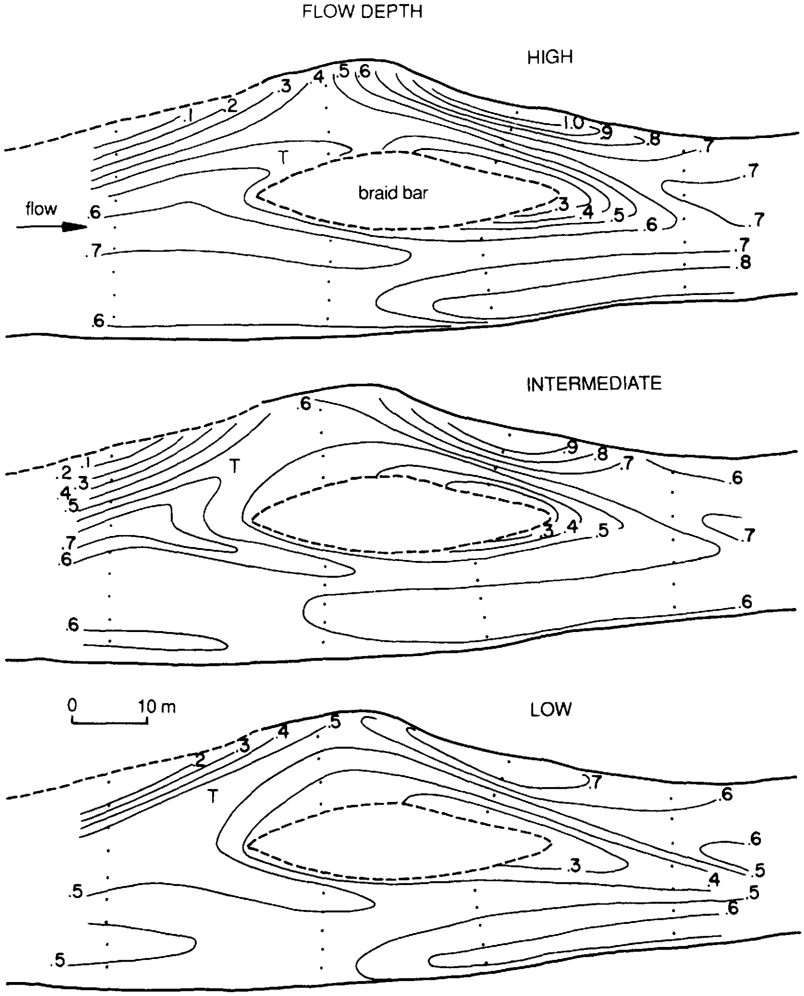


Fig. 8. Channel geometry at high, intermediate and low flow stages associated with a braid bar, Calamus River, Nebraska (from Bridge & Gabel 1992). Bed topography contours are in metres below a datum at approximately bankfull level. Dots are flow measurement stations. The thalweg of the left-hand channel near the upstream end of the bar (T) moves towards the left bank as stage falls.

local depth and radius of curvature respectively, d_c and r_c are centreline values, and the exponent a depends on the characteristics of the flow and sediment transport. In general, a increases as dimensionless bed shear stress and sediment transport rate increase, and as bed roughness decreases. Thus maximum cross-sectional asymmetry of curved channel segments increases as sinuosity increases, as bed sediment size decreases, and as flow stage increases. This kind of equation does not describe the detailed local relief of river cross-sections associated with mesoforms like dunes, and the discrete bar head lobes and bar-tail scrolls which are associated with episodic deposition. Also, if bank erosion occurs, it takes a finite amount of time for the widened cross-section to approach equilibrium with the flow.

As discharge falls the thalwegs near the upstream part of the bar tend to move towards the outer banks in association with deposition near the upstream tip of the bar, and resulting in a tendency to reverse the cross-sectional asymmetry of the channels here (Figs 7 and 8; Bridge & Gabel 1992). In the channel segments adjacent to the bar, cross-sectional asymmetry may be reduced by deposition in the thalweg and erosion of the topographically high areas of the bar (Fig. 7). Such falling stage modification in channel geometry depends on the ability of the flow to erode and deposit. This may not be possible if the bed becomes armored or if shallow parts of channel segments become emergent. The changes in braided channel geometry with falling discharge described above are equivalent to decreasing height and trimming margins of (ancestral) multiple-row alternate bars. Conversely, increasing discharge would result in increasing alternate bar height and the tendency to migrate downstream.

'Hydraulic geometry' of curved channel segments

The 'at-a-station hydraulic geometry' (normally represented as log-log plots of water surface width (w), cross-section average depth (d) or flow velocity (V) versus discharge (Q)) for individual curved channel segments in braided rivers is similar to that for undivided sinuous channels (e.g. Church & Gilbert 1975). Thus, hydraulic geometry varies along the length of a segment and between segments in response to varying cross-sectional geometry and flow resistance, and in general log-log plots of hydraulic geometry are nonlinear (Knighton 1972; Nordseth 1973; Cheetham 1979; Eschner 1983; Bridge & Gabel 1992). The mean geometry of individual channel sections cannot necessarily be related to whole-stream discharge because the individual channel discharge and the whole-stream discharge may not vary congruently.

At-a-station hydraulic geometry of the entire braided cross-section is discussed by Mosley (1982*b*, 1983) and Ergenzinger (1987). Mosley (1982*b*, 1983) collected at-a-station hydraulic geometry data from a number of braided rivers for whole-river transects and individual anabranches. As discharge increased with time existing channels enlarged and merged, and new channels became active. For all cases, width, depth and velocity increase with discharge and were expressed as simple power functions as observed in other studies. However, the variability in hydraulic geometry relationships among different rivers, reaches of the same river, transects in a given reach, and different anabranches is so great as to be of little predictive value. This variability is clearly because discharge is not the sole control of hydraulic geometry, and because log-log plots may not be linear (see also Eschner 1983). Furthermore, as channels become enlarged or filled, hydraulic geometry will change with time (even as it is being measured!).

Downstream hydraulic geometry may also be expressed in terms of individual segments (Fahnestock 1963; Church & Gilbert 1975; Rice 1982) or the entire river (Ashmore 1991*a*). Ashmore's (1991*a*) results from a laboratory channel are broadly similar to those from single-channel streams. However, it is difficult to make a strict comparison because in real rivers downstream increases in channel-forming discharge are usually accompanied by decreases in slope and bed grain size, parameters held constant in Ashmore's experiments. Ferguson & Ashworth (1991) show that mean channel width in braided rivers (with braiding index less than 2) varies with discharge, slope and median bed-material size, and that rational regime theories are generally better predictors of channel width than are empirical equations.

It is well known that in single sinuous channels the width and wavelength are proportional to the square root of the channel-forming discharge. Thus the wave length/width ratio tends to equal approximately 10, or the riffle-riffle spacing is approximately 5. This also seems to be the case for braided river segments (e.g. Bridge *et al.* 1986). Also, the relative sinuosities of channels on either side of braid bars is commonly such as to give a braid length/maximum width of approximately 3–4, which represents a streamlined form according to Komar (1983, 1984).

In many natural rivers, the riffle-riffle spacing, and overall bar dimensions do not change appreciably as discharge falls seasonally below 'channel-forming' discharge. However, if a braided channel segment experiences a reduction in discharge over a long enough term (relative to erosion and deposition rate), a series of bars and bends may develop that have a shorter wavelength than the original segment. This will give the appearance of an anastomosing channel segment that has nothing to do with avulsion or high sedimentation rates. Indeed, many sinuous channel belts of undivided channels which contain smaller wavelength bends are probably related to such long-term reductions in discharge.

Geometry of confluence zones

Figure 9 illustrates the important geometrical features of confluence zones (see also Mosley 1976; Best 1986, 1988; Bristow *et al.* 1993; Ashmore this volume). Features of particular concern are: (1) the confluence angle; (2) whether the incoming channels are oriented symmetrically or asymmetrically relative to the confluence direction; (3) relative size of

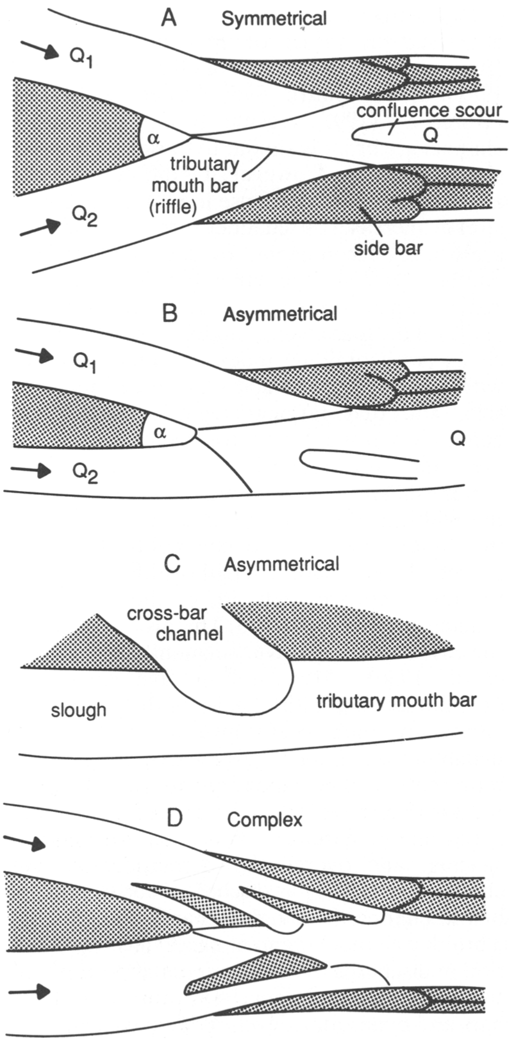


Fig. 9. Idealized plan geometry of different types of confluences at a given flow stage Q, Q_1 and Q_2 are discharges, and α is confluence angle.

incoming channels as measured by their widths, depths, cross-sectional areas, or discharges; (4) the maximum depth, width and length of the scour zone; (5) complications arising from the presence of more than two joining channels. These features are commonly difficult to define as they vary with time and discharge.

Relative discharge is defined in various ways, using different combinations of the incoming discharges and the combined discharge (Table 5). Changes in relative discharge with time can result during changes in overall discharge without any erosion or deposition in the joining channels (Bridge & Gabel 1992). They will also result if one channel segment is being erosionally enlarged while the other is being filled. One channel segment may also lose discharge due to a channel diversion (avulsion) upstream of the confluence. In major tributary junctions relative discharge changes may be associated with upstream avulsions or incongruous flood peaks, possibly associated with ice jams or differing hydrological nature of drainage basins (e.g. Best 1986; Bergeron & Roy 1988; Alam *et al.* 1985; Reid *et al.* 1989).

Confluence angles are difficult to measure in natural channels (as opposed to laboratory models) because channels vary in curvature as they join. Common values of confluence angle range from 15° to 110°. An estimate of the confluence angle in terms of the sinuosity (sn) of the joining channels can be derived if the channel segments can be represented by a sine-generated curve (Langbein & Leopold 1966). As confluences tend to be associated with reversal in curvature of the joining channel segments (i.e. crossovers), and the deviation angle of the local channel direction at the crossover from the mean down valley direction is given in degrees as $126°[(sn-1)/sn]^{0.5}$, the *maximum* confluence angle possible is

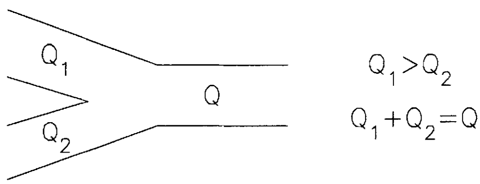
$$126°\left[\left(\frac{sn_1-1}{sn_1}\right)^{0.5}+\left(\frac{sn_2-1}{sn_2}\right)^{0.5}\right] \qquad (2)$$

where sn_1 and sn_2 are the sinuosities of the two channel segments. If the channel segments are assumed to be circular arcs, the constant in the above relationship is approximately 140°.

The geometry of confluence zones can be conveniently subdivided and described in terms of: (1) entering channels; (2) confluence scour zone (thalweg); (3) side (point) bars (Fig. 9). Braided channels entering confluence zones tend to be 'riffle' areas (e.g. Fig. 7) so that depth does not vary much across the channel width. However, with the channel geometry of Fig. 7, a thalweg is commonly present near the outer bank at high flow stage, but may be near the inner bank at low flow stage. Avalanche faces may be present where the entering channels pass into the confluence scour zone ('riffle' areas with avalanche faces are commonly referred to as tributary-mouth bars). Generally, their crestlines are oblique to the channel direction (as are the equivalent parts of ancestral alternate bars). This obliquity increases with cross-sectional asymmetry and flow stage and as one channel becomes dominant over the other. In the extreme case the avalanche face of the dominant channel may be almost parallel to its inner bank,

Table 5. *Measures of relative discharge of joining channels*

Measure	$Q_1 = Q_2$	$Q_1 \gg Q_2$	Author
$\frac{Q_1}{Q}$	0.5	1.0	Best & Reid (1984)
$\frac{Q_2}{Q}$	0.5	0.0	Best & Reid (1984)
$\frac{Q_2}{Q_1}$	1.0	0.0	Mosley (1976), Best (1986)
$\frac{(Q_1 - Q_2)}{Q/2}$	0.0	2.0	Ashmore & Parker (1983)
$\frac{(Q_1 - Q_2)}{Q}$	0.0	1.0	Bridge (This paper)



thereby blocking and/or migrating into the minor channel (e.g. Best 1986, 1987; Hein & Walker 1977; Krigström 1962; Lodina & Chalov 1971; Teisseyre 1975; Ashmore & Parker 1983; Bergeron & Roy 1988; Reid *et al.* 1989). This is an important process in the blocking of the downstream ends of channel segments which are becoming abandoned. If the channel being blocked is still active, this may lead to increased erosion of its outer bank or an upstream diversion (Fig. 9). Bluck (1979) referred to these channel blocking deposits as spits, erroneously assigning their origin to wave action. Where a cross-bar channel joins a larger main channel only one avalanche face may be present, associated with the tributary mouth bar of the cross-bar channel (Figs 1 and 2). According to Best (1987, p. 494) avalanche faces occur if confluence angle exceeds approximately 20°. In view of the above, the presence or absence of avalanche faces is also controlled at least by relative depth and discharge.

As discharge and relative discharge vary the tributary-mouth bars may grow forward or retreat, and change in crest height and orientation in plan (Best 1987; Bergeron & Roy 1988; Reid *et al.* 1989). The crests tend to increase in height and prograde during high flows, particularly near the outer banks where the flow velocities are highest, resulting in increasing obliquity of the crestlines relative to the flow direction (for the channel geometries of Figs 7 and 9). The crests tend to be eroded at low flow stages, and may become dissected, with higher parts emergent, resulting in a complicated confluence zone (Fig. 9D). These changes in channel geometry are entirely consistent with those that occur in single-channel rivers, where 'riffle' areas tend to be areas of deposition at high flow stages but areas of erosion at low flow stages (Lane & Borland 1954; and many others).

Confluence scour zones have received a lot of attention because the deep water in these zones is relevant to construction of bridge piers and pipeline crossings. The geometry and orientation of confluence scour zones are influenced once again by factors such as the confluence angle and the relative discharges of the entering channels. As the confluence angle increases the scour zone changes from trough shaped to more basin-like (Ashmore & Parker 1983; Best 1986). If the discharge of the entering channels are similar, the long axis of the scour tends to bisect the confluence angle (Best 1986). If one channel is dominant the scour zone tends to parallel the direction of this channel (Ashmore & Parker 1983; Best 1987; Figs 7 and 9).

The maximum depth of scour (d_s) is commonly expressed as a function of the average depth of the joining channels (d_1, d_2) at channel-forming discharge, i.e.

$$d_s^* = \frac{d_s}{(d_1 + d_2)/2}. \quad (3)$$

When expressed in this way the dimensionless scour depth (d_s^*) commonly ranges up to 4, and may be as large as 6 (Mosley 1976, 1982*a*; Ashmore & Parker 1983; Best 1986; Kjerfve *et al.* 1979). However, with field measurement it is difficult to locate the maximum depth, and to be sure this geometry is in equilibrium with the flow.

Dimensionless maximum scour depth has been related to the confluence angle and the relative discharge of the entering channels (Mosley 1976, 1982*a*; Ashmore & Parker 1983; Best 1986, 1988). For a given relative discharge, scour depth increases with confluence angle and appears to approach an upper limit asymptotically (i.e. the relationship is nonlinear). For a given confluence angle, d_s^* increases as the discharges in the entering channels tend to equality. If one channel is overwhelmingly dominant the maximum scour depth must be equivalent to that in a single channel bend and the dependence on confluence angle is lost. There is a lot of scatter on plots of scour depth versus confluence angle and relative discharge using field data, suggesting that not all of the controlling variables are included. Mosley (1976) related d_s^* to sediment transport rate, and Mosley (1982*a*) related d_s^* to discharge for the Ohau River. Ashmore & Parker (1983) also suggested that d_s^* might be related to a densimetric Froude number and the average water surface slopes of the entering channels, but field data did not support such correlations. It seems that the cross-sectional asymmetry of the entering channels, and the presence of multiple channels at the confluence may also influence d_s^*.

A simple, approximate model for the maximum equilibrium scour depth in confluence zones can be constructed if it is assumed that the joining channels can be represented by segments of sine-generated curves. In this model, the maximum scour depth is located approximately at the apex of the curved segments where the centreline radii of curvature (r_{c1} and r_{c2}) are at a minimum and the channels are orientated in the mean down valley direction. The upstream junction corner of the entering channels corresponds approximately to the crossovers in the sine-generated curves, where the channels have their maximum deviation angles (α_1 and α_2) from the down valley direction. The *maximum* confluence angle is given by

$$\alpha = \alpha_1 + \alpha_2 = \frac{1}{2\pi}\left(\frac{M_1}{r_{c1}} + \frac{M_2}{r_{c2}}\right) \quad \text{(radians)} \quad (4)$$

where M_1 and M_2 are the channel lengths in one wavelength, or four times the channel length from the crossover to the position of maximum scour depth. The deviation angles α_1 and α_2 are also related to channel segment sinuosity as

$$\alpha = 2.2\left(\frac{sn - 1}{sn}\right)^{0.5} \quad \text{(radians)}. \quad (5)$$

Using equation (1), the maximum dimensionless scour depth for each channel segment is given by

$$\frac{d_s}{d_c} = \left(\frac{r_c + w/2}{r_c}\right)^a \quad (6)$$

where d_c is the centreline depth of the segment, and w is its width. As the scour depth given by equation (6) will generally be different for each channel segment it is necessary to devise a way of determining a single value. In the absence of a more rigorous analysis, this value is assumed to be given by

$$d_s = \frac{q_1 d_{c1}}{Q}\left[1 + \frac{w_1}{2r_{c1}}\right]^a + \frac{q_2 d_{c2}}{Q}\left[1 + \frac{w_2}{2r_{c2}}\right]^a \quad (7)$$

where Q_1 and Q_2 are the discharges of the joining channels, Q is combined discharge, $r_{c1} = M_1/2\pi\alpha_1$ and $r_{c2} = M_2/2\pi\alpha_2$. Equation (7) indicates that the maximum scour depth must depend on at least the confluence angle and the relative discharges of the joining channels, but also M and a. However, if common values of w/r_c of 0.5 to 0.33 and a of 4 to 6 are used in the case of equal sized channels, values of d_s^* of between 1.85 and 3.8 are obtained which are in the same range as common field values. This method does not take into account the effects of the shear layer between the joining channels, nor of flow separation downstream of avalanche faces or the equivalent alternate bars. However, the heights of alternate bars with avalanche faces and point bars evolved from them will be comparable.

As confluence angles and relative discharges will change with flow stage, the orientation and geometry of the scour zone will also change (Ashmore & Parker 1983; Bergeron & Roy 1988, Ashmore this volume). The confluence angle may be larger at low flow stages relative to high flow stages, and the position of the confluence may be further upstream (Figs 7 and 9). Thus, at high flow stages deposition may occur in the upstream end of the low-flow scour zone, whereas at low flow stages these high stage deposits are eroded and deposited downstream in the high stage scour zone. If one channel becomes dominant the maximum scour depth may be decreased and the scour zone moved towards the outer bank of the subordinate channel, possibly inducing bank erosion (Ashmore & Parker 1983; Fig. 7). The position of the confluence scour zone will shift in response to erosion of the outer banks of both entering channels and concomitant deposition near the downstream tip of the upstream braid bar (Ashmore this volume). The confluence scour will also change in orientation and geometry if there is avulsive shift of an upstream channel (Ashmore & Parker 1983).

The confluence scour zone passes laterally into topographically high bars which are directly analogous to point bars in single curved channels and represent parts of alternate bars in straight channels. There is only one bar present if the confluence is asymmetrical (one straight-through channel; Fig. 9). Longitudinal ridges of sediment on these bars extending parallel to the edges of the scour zone (Ashmore 1982; Ashmore & Parker 1983; Fig. 9) result from episodic deposition and are dynamically equivalent to the bar tail scrolls mentioned previously.

'Hydraulic geometry' of confluence zones

The widths and cross-sectional areas of the single channels at and downstream of confluence zones are generally less than those of the entering channels combined. This must mean that mean flow velocities in confluence zones are relatively large. Lyell (1830) explained this phenomenon in the case of tributary junctions as due to less flow resistance in the single channel which has a smaller wetted perimeter than the two entering channels. The differences in hydraulic geometry between the entering channels and the confluence zone and downstream are discussed by Roy & Woldenberg (1986), Roy & Roy (1988), and Roy *et al.* (1988).

Geometry of diffluence zones

Downstream of confluence scours the maximum depth decreases and in braided rivers the channel commonly widens and splits around a braid bar. In straight channels a mid-channel alternate bar would occur in this position (Fig. 7). Migratory lobes of sediment ('unit bars', Ashmore 1982) occur in this zone, possibly related to episodic bank erosion within or upstream of the confluence scour zone. If the diffluence angle is large and/or one dividing channel is dominant, there is a tendency for the upstream side bar to build into the subordinate channel from the outer bank thus tending to block its entrance (Chalov 1974; Bridge *et al.* 1986; Best & Reid 1987; Kasthuri & Pundarikanthan 1987).

Flow in braided rivers at the bar scale

Flow in curved channel segments

The flow in individual curved channels around braid bars is broadly equivalent to that in single-channel sinuous rivers, and can be approximated by simple flow models (Bridge & Gabel 1992; Bridge 1992; Figs 10–12). Curved flow around any type of channel bar results in a flow-transverse component of water surface slope towards the inner, convex bank, a spiral flow pattern, and convective accelerations and decelerations of the depth-averaged flow. The spiral flow pattern arises primarily because of an inbalance through the flow depth of curvature-induced centrifugal forces and the pressure gradient associated with the transverse-sloping water surface. The convective accelerations and decelerations are associated with spatially-varying bed topography and channel curvature and, to a lesser extent, with the spiral flow. These flow patterns cause the maximum depth-averaged velocity to cross from the inner, convex bank at the bend entrance towards the outer, concave bank with progression around the bend. Associated with this flow pattern is a net outward flow in the upstream segments of bends, and a net inward flow in the downstream segments (Dietrich & Smith 1983; Bridge & Gabel 1992; Figs 10–12). Bed shear stress tends to vary in a similar way to depth-averaged flow velocity. However, as flow resistance coefficients tend to be controlled by local bed configuration rather than larger-scale bar topography, bed shear stress may not always have a simple relationship with depth-averaged flow velocity (Bridge & Jarvis 1982; Bridge & Gabel 1992).

In general, spatial variation in the magnitude of depth-averaged velocity, deviation of the depth-averaged velocity vector from the mean channel direction, magnitude of the spiral flow components, and transverse bed and water surface slopes all increase as the radius of curvature of the bend decreases, as dimensionless bed shear stress increases, and as the flow resistance

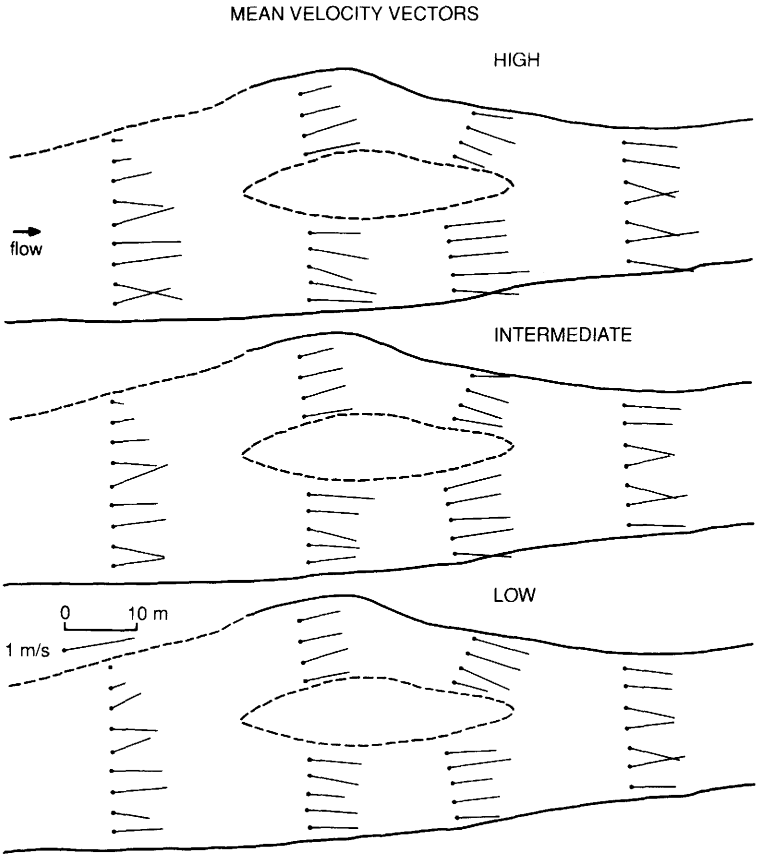


Fig. 10. Vertically-averaged flow velocity vectors for high, intermediate and low flow stages associated with a braid bar, Calamus River, Nebraska (from Bridge & Gabel 1992).

coefficient decreases. In the typically low sinuosity (<1.1) channel segments of braided rivers, the spiral flow and depth-averaged across-stream flow components are of the same order of magnitude, and are one or two orders of magnitude less than the along-stream flow components (Figs 11 and 12). At bend entrances the inward-directed near-bed spiral flow components tend to be counteracted by outward-directed depth-averaged flows, such that there is an overall outward flow increasing in magnitude from the bed upwards (Fig. 12). In the downstream parts of bends the inward-directed depth-averaged flow adds to the inward-directed spiral flow at the bed (Fig. 12). Deviation angles of these flow components from the mean downstream direction are generally only a few degrees, and normally less than 10°, which makes them very difficult to measure, especially in the presence of bedforms such as dunes (Bridge & Gabel 1992; Ashworth & Ferguson 1986).

In most braided channel segments the discharge will change as overall discharge changes, but it may not change in the same sense. All of the flow patterns mentioned above will change character with discharge, and new ones arise as a result of discharge variation. With falling discharge, the water flows in a more sinuous course around the emerging bar, resulting in relatively strong across-stream components of depth-averaged flow near the bend entrance, possibly enhanced magnitude of spiral flow, and rapid movement of the maximum depth-averaged velocity from the inner to outer bank. However, the magnitude of the spiral flow also increases as flow resistance decreases, and changes in flow resistance with discharge depend on changes in the relative roughness of bed forms and bed grains. In the sand-bed Calamus river, resistance coefficients at a point vary little with flow stage because maximum-steepness dunes were always present and more or less in equilibrium with the

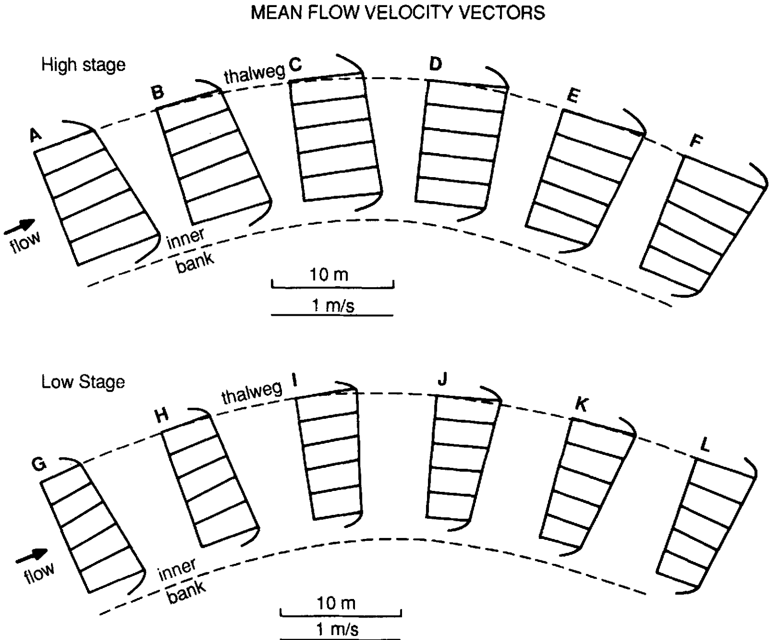


Fig. 11. Theoretical (see Bridge 1992) vertically-averaged flow velocity vectors at high and low flow stage for curved channel segments comparable to those in the Calamus River (Fig. 10). Channel cross-sections shown in Fig. 12.

flow (Bridge & Gabel 1992). In other sand-bed rivers large dunes in disequilibrium with falling flow, or a change from upper-stage plane beds to dunes as discharge falls, could increase resistance coefficients. In coarse sand-gravel rivers a transition from dunes to lower-stage plane beds as stage falls may decrease resistance coefficients. Alternatively, increasing relative roughness of falling flows over plane gravel beds may increase resistance coefficients. Thus, both flow velocity and bed shear stress patterns are likely to be difficult to predict in falling, disequilibrium flows.

New flow patterns may arise at high flow stages as new channels are cut, particularly where zones of high flow velocity can take advantage of topographically low areas. As an example, the relatively high water levels on the outside of curved channels may lead to spilling of water over a low section of bank, and possibly the development of a new channel transverse to the original channel (e.g. Ashmore 1982). Falling stage flow may also be diverted into low areas of the emerging bar topography, leading in some cases to incision of existing or new cross-bar channels.

As at-a-station hydraulic geometry normally varies among different channel segments (see previous section) cross-sectionally averaged flow velocity and depth will vary incongruently among channel segments as discharge varies (e.g. Cheetham 1979; Bridge & Gabel 1992). For example, higher velocities in chute channels relative to adjacent curved channels may occur at flood stage, but with similar or smaller velocities at lower stage. Such differences may lead to erosion of one channel and deposition in the other, which will result in a change in hydraulic geometry.

Flow in confluence zones

Very little detailed information is available on how the flow patterns in confluence zones vary in time and space. Most useful data come from flume experiments, and from river studies where measuring bridges were constructed (e.g. Mosley 1976, Ashmore 1982, this volume; Ashmore & Parker 1983; Best 1986, 1987, 1988; Best & Reid 1984, 1987; Best & Roy 1991; Roy *et al.* 1988; Roy & Bergeron 1990). However, other data are also available (e.g. Ashworth & Ferguson 1986; Davoren & Mosley 1986; Ashmore *et al.* 1992; Ferguson *et al.* 1992). Confluence flow has been compared to the flow

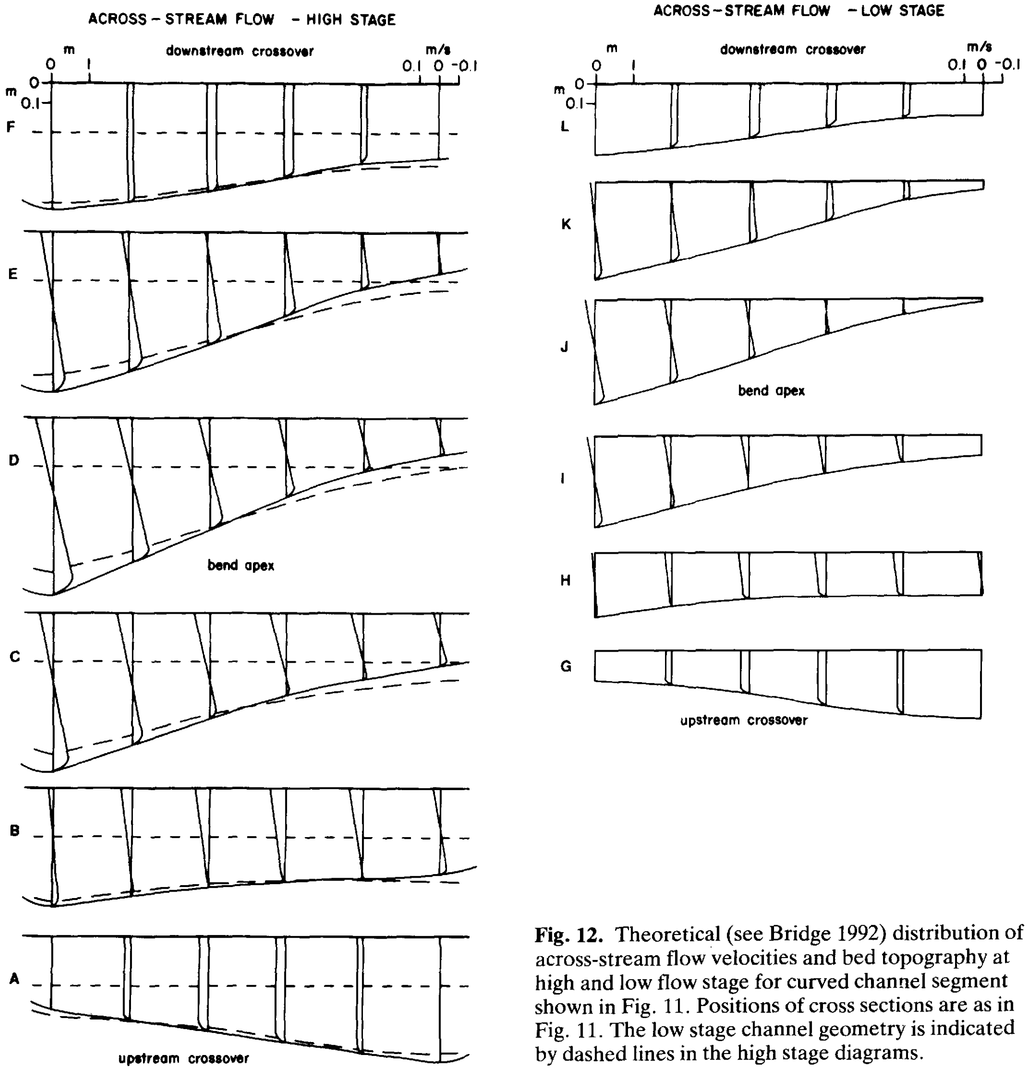


Fig. 12. Theoretical (see Bridge 1992) distribution of across-stream flow velocities and bed topography at high and low flow stage for curved channel segment shown in Fig. 11. Positions of cross sections are as in Fig. 11. The low stage channel geometry is indicated by dashed lines in the high stage diagrams.

in joining river bends and, because the cross-sectional area of the joined channels is less than the sum of the two joining channels, to wall jets (Mosley 1976; Ashmore & Parker 1983). Even in straight channels with double or multiple rows of alternate bars the water flows as a series of sinuous threads adjacent to each other (Fig. 1). The flow in confluences is discussed here in terms of three zones (cf. Best 1987): (1) entrance zones; (2) confluence mixing zone where the joining fluid streams become asymptotic and mix; (3) zones of flow separation with vertical axes near inner convex banks at the upstream tip of the confluence and associated with side bars adjacent to the confluence scour. Details of the flow in these zones depends at least on the confluence angle and the relative discharges of the joining channels, which vary with flow stage.

Entrance zones are equivalent to the 'riffle' zones of curved channels, and downstream-dipping avalanche faces (tributary-mouth bars, chute bars) may be present. Such avalanche faces are typically present in natural channels where the confluence angle exceeds 20°. However, in straight channels with small width/depth ratios, the confluence angles associated with the confluence between adjacent alternate bars with avalanche faces may be less than 20° (Fig. 1). At the crest of the avalanche faces at high flow stage the depth-averaged flow velocities are expected to be greatest near the outer banks, and there should be a net across-stream component of flow towards the centre of the confluence throughout the flow depth as curvature-induced secondary flows are negligible (Figs 11 and 12). However, as the crestlines of the avalanche faces are generally oblique to flow, the separated and reattached flow downstream of the crestline will have a component of near-bed flow towards the banks, whereas the surface flow will continue to be directed towards the centre of the confluence (Fig. 13). At low flow stages shifts in the maximum velocity loci (Fig. 7) may result in changes in crest height and obliquity. Dissection of the crests of the avalanche faces may produce a complex, multichannel confluence zone (Fig. 9). The distance downstream from the crest of the avalanche face to the reattachment zone is expected to be proportional to the height of the avalanche face, as with dunes. Thus, changes in the height and obliquity of the avalanche faces with discharge will influence the geometry of the scour zone associated with the reattaching boundary layer (Fig. 9).

The curvature of the joining channels in confluence zones should inevitably give rise to superelevation of the water surface in mid-channel and spiral flow with near-bed flow towards the outer banks (Mosley 1976, 1982*a*; Ashmore 1982; Ashmore & Parker 1983; Roy *et al.* 1988; Ashmore *et al.* 1992). The maximum water surface superelevation and the magnitude of the spiral flow tend to increase as the confluence angle (hence curvature) increase

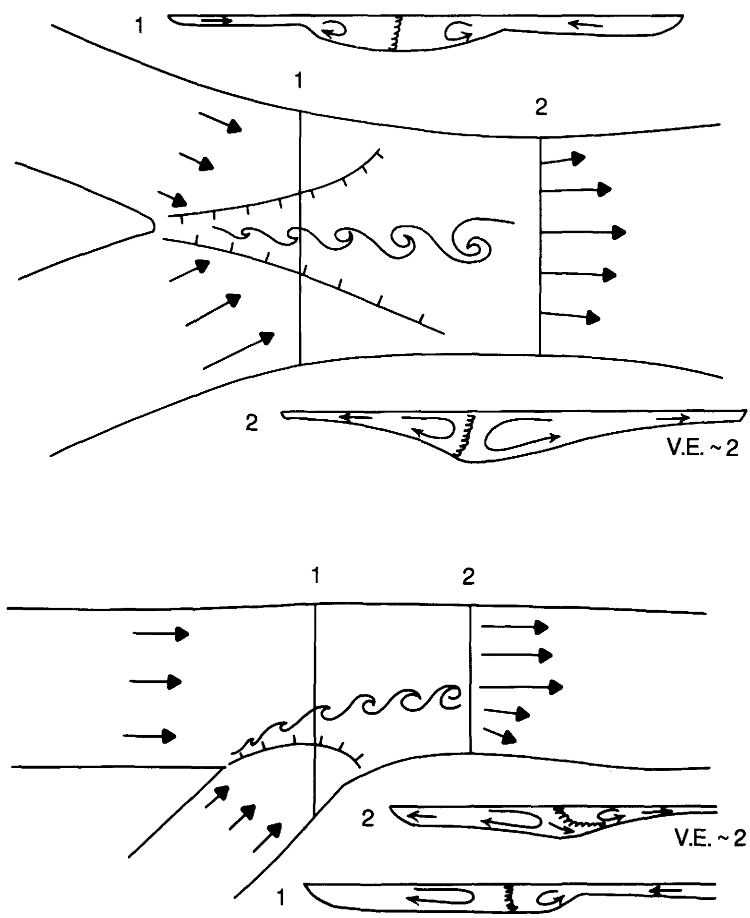


Fig. 13. Idealized high stage flow patterns at confluences showing tributary bars with avalanche faces, zones of Kelvin-Helmholtz instability at the interface of the joining streams, vertically-averaged flow velocity vectors (in plan) and cross-stream circulation patterns (in cross sections). Based on Best (1986) and Best & Roy (1991).

(Ashmore & Parker 1983) as expected. By analogy with flow in bends, the maximum high-stage flow velocity should occur near the centre of the confluence scour, with lower velocities near the outer banks (Fig. 11). The location and relative depth of the confluence scour zone is clearly influenced by the curvature-induced spiral flow, and migration of the loci of maximum depth-averaged velocity towards the centre of the channel. However, it is also influenced by: reattachment of the separated boundary layer downstream of the upstream avalanche faces; flow acceleration associated with a reduced cross-sectional area (possibly influenced by separation zones over side bars; Best & Reid 1984); enhanced turbulence in the mixing layer between the joining streams (Best 1986). At the time of writing, it is not possible to assess the relative importance of these factors.

The mixing zone between the two joining streams results from their velocity differential and can commonly be seen clearly if the suspended sediment loads of the two streams differ. Some mixing is inevitable in any turbulent flow. However, the existence of time-averaged relative velocity leads to a Kelvin-Helmholtz type instability and vortices with near-vertical axes form that become entrained in the faster moving flow (Fig. 13). The wavelength of these vortices is controlled by the relative velocity, densities and viscosities of the joining flows (i.e. dependent on Reynolds and Richardson numbers). If one stream is slower near the bed than the other, for instance due to flow separation in the lee of an upstream avalanche face, the mixing layer is distorted near the bed towards the slower stream (Fig. 13; Best & Roy 1991). This may result in upwelling of the fluid from one stream into the region occupied by the other (Fig. 13), which greatly affects the mixing of sediment from the joining streams.

Zones of flow separation with vertical axes of flow rotation are caused by adverse pressure gradients associated with curvature-induced transverse water surface gradients in combination with reduced downstream slope resulting from local flow expansion and deceleration. Such flows result in enhanced deposition on the downstream tips of braid bars and side bars adjacent to confluence scour zones, thereby reducing the effective local cross-sectional area of the channel. Note that flow separation *enhances* deposition on side bars rather than being the sole cause for their occurrence (cf. Bristow *et al.* 1993). These separation zones are most marked for large confluence angles. For the case of an asymmetrical junction the length and width of the separation zone increases with the confluence angle and the discharge of the tributary relative to the main channel (Best & Reid 1984).

During rising flow stages a partially abandoned braided channel segment may receive water from its downstream junction with an active channel. In other words, water flows upstream into the channel. This situation apparently occurs also in ephemeral rivers where water may start to flow in only part of a river system. This results in flow of water from an active tributary up the inactive tributary from their confluence (Alam *et al.* 1985; Reid *et al.* 1989).

Mathematical models of flow in confluences are not generally available because it is difficult to describe the complicated flow mechanics described above. Hager's (1984, 1987) flow model predicts well the geometry of the separated flow zone in Best & Reid's (1984) laboratory model of a tributary junction. However, some of the assumptions in this model (e.g. negligible friction loss, uniform flow, constant channel widths and depths) are unlikely to be applicable to natural rivers. Flow models of single river bends may be capable of giving at least a qualitative picture of the velocity field (e.g. Bridge & Gabel 1992; Bridge 1992). Indeed, as shown previously, such flow models predict confluence scour depths of the correct order of magnitude.

Flow in diffluence zones

Flow in diffluence zones is not known well (but see Straub 1935; Ashworth & Ferguson 1986; Davoren & Mosley 1986; Ferguson *et al.* 1992). By analogy with flow in curved channels, the maximum velocity locus, which is in mid-channel, starts to split in this zone such that each thread of maximum velocity is close to the upstream tip of the braid bar downstream. The mixing layer of the confluence zone does not exist as the relative velocity of the two streams is zero here (analogous to 'downstream recovery zone' of Best 1987). Bridge & Gabel (1992) observed relatively complicated patterns of convergence and divergence of local depth-averaged velocity in this zone, which were associated with bar-scale bed topography (see also Mosley 1976; Ashmore *et al.* 1992). In cases where the diffluence angle is large, as perhaps at the entrance of a channel that is being abandoned, a zone of flow separation with a vertical axis may occur near the outer bank of the channel entrance. Such a zone will enhance the deposition of sediment here. The width and length of this zone increase as the discharge into the filling channel decreases relative to the main channel (Kasthuri & Pundarikanthan 1987).

Sediment transport in braided rivers at the bar scale

It is very difficult to measure or calculate the total and local sediment transport in braided rivers because of its temporal and spatial variability, and obvious logistical problems. Some of the variability is associated with turbulence, the movement of sediment as bedwaves of various scales (e.g. ripples, dunes, unit bars, complete braid or point bars), random changes in sediment supply (e.g. bank slumping, breakup of 'armored' beds), local channel cutting and abandonment, weather-related changes in water discharge and sediment supply, and tectonically induced changes in sediment supply. Temporal scales of such variation range from seconds to many years, and spatial scales range from mm to km. Detection of any progressive or periodic changes in sediment transport must therefore depend on the time and space interval and extent of measurement. Most experimental studies of bedload transport are sufficient only to resolve time variations at a point on the order of hours (exceptionally minutes), and space variations on the bar and channel scale. Commonly, bedload transport samples taken over a minute or so at intervals of minutes or hours have a coefficient of variation of tens of percent (Bridge & Jarvis, 1982; Ashmore 1988; Gabel 1993; Bennett 1992). Typically, such measurements can reveal the growth and migration of dune-like bedforms, unit bars, and braid bar complexes (e.g. Griffiths 1979; Ashmore 1988, 1991*a*; Davoren & Mosley 1986; Kuhnle & Southard 1988; Bennett 1992; Ikeda 1983; Hoey 1992). The ensuing discussion is concerned mainly with grain size and rate of bedload transport at the bar scale over the lifespan of a bar. Then, the mode of sediment movement associated with smaller scale bedforms is considered.

Bedload transport at the bar scale

Studies of bedload transport in the curved channel segments adjacent to braid bars and point bars for the case of sandy rivers at high flow stage demonstrate that the loci of maximum bedload transport rate and grain size are similar to those of mean flow velocity and bed shear stress (e.g. Dietrich & Smith 1984; Bridge & Gabel 1992). With progression around a curved-channel segment the bedload grains larger than the average tend to move preferentially towards the outer bank, whereas those finer than the average tend to move towards the inner bank. In general, there is a small net component of bedload transport towards the outer banks. However, for streams with a substantial gravel fraction, although the locus of maximum grain size follows that of bed shear stress, the locus of maximum bedload transport rate stays close to the centre of the channel (Bridge & Jarvis 1982; Dietrich & Whiting 1989; Bridge 1992).

The reason for this difference between sandy and gravelly streams lies in the fact that both mean grainsize and transport rate of bedload depend on what proportion of the total bed shear stress is available for bedload transport relative to that associated with immobile bed friction and bedform drag. For example, in mixed sand-gravel streams, gravel is transported only where bed shear stress is greatest. Bedload transport rates tend to be low here, however, as the bed shear stress is just above the threshold of grain motion and the bed may be armored. Thus, higher bedload transport rates can occur where lower bed shear stresses act on sandy beds. In sand-bed streams an increase in bed shear stress may result in an increase or decrease in bedform height/length (hence form drag) such that the bed shear stress effective in bedload transport may decrease or increase.

Despite these complications, high stage bedload transport rates and mean grain sizes agree well with theoretical models for steady, equilibrium flow conditions (Bridge 1984, 1992; Bridge & Gabel 1992). It is not possible at present to predict the sediment transport vectors of individual grain size fractions as these are a complicated function of the near-bed flow field and bed slope over bars and oblique-crested bedforms like dunes (Bridge & Jarvis 1982; Dietrich & Smith 1984; Bridge 1992).

Bed material normally fines downstream on the tops of both braid bars and point bars in modern sandy and gravelly rivers, but is relatively coarse in the thalwegs of the downstream segments (Straub 1935; Bluck 1971, 1974, 1976, 1979, 1982; Smith 1974; Jackson 1975*a*, 1976*a*; Bridge & Jarvis 1976, 1982; Lewin 1976; Deitrich *et al.* 1979; Crowley 1983; Ferguson & Werritty 1983; Ashworth & Ferguson 1986, 1989; Ashworth *et al.* 1992*b*). Thus the mean grain size of bed material reflects the distribution of bed shear stress and bedload grain size at high (channel-forming) discharges, as would be expected (Bridge & Jarvis 1982; Bridge & Gabel 1992).

As discharge changes, the changing patterns of bed shear stress, bed configuration and available bed material will result in complicated patterns of transport rate and grain size of the bedload. Mixed-size bedload transport models (e.g. Bridge & Bennett, 1992) clearly show how bedload size is controlled by both the flow conditions and the available sediment. Over a flood event, the size distribution of the available sediment and the bedload may vary as well as the bedload transport rate (e.g. Vogel *et al.* 1992). In general, relatively diminished bed shear stresses at low flow stages will result in smaller grain sizes and transport rates of bedload, or bedload transport may cease altogether. Static armor layers may develop. Limited deposition at these low flow stages results in veneers and patches of relatively fine grained sediments (e.g. Rust 1972; Bluck 1979; Boothroyd & Ashley 1975) and filling of openwork gravels with a sandy matrix (e.g. Minter & Toens 1970; Smith 1974; Beschta & Jackson 1979; Frostick *et al.* 1984).

Studies of bedload transport rate in confluences are rare (e.g. Davoren & Mosley 1986; Ashworth *et al.* 1992*a*,*b*; Ferguson *et al.* 1992) but suggest that the largest bedload transport rates generally occur where flow velocities and bed shear stresses are also largest, as long as sediment is available for transport (e.g. bed is not armored). Tracing the paths of seeded particles of various sizes (Best 1986, 1987, 1988; Roy & Bergeron 1990) clearly demonstrates that bedload particles travel more-or-less parallel to the channel banks and bed contours as they pass through the confluence scour zone (see also Ashmore & Parker 1983). This implies that bedload from the two joining channels experience very little mixing as it passes through the confluence zone. Best (1986, 1987) observed in his flume experiments that bedload moves along the sides of the confluence scour but not in its deepest part, and this separation of bedload became more marked as the scour depth increased. In contrast, Roy & Bergeron's (1990) observations of low-flow bedload transport in a low-angle natural confluence indicate that some bedload moves laterally down the sides of the scour zone. By analogy with the movement of bedload through single curved channels, the movement of grains more or less parallel to channel banks implies an approximate balance between the transverse gravity force into the scour zone and the fluid drag out of the scour zone at channel-forming discharges. Grains coarser than the mean are expected to have a small transverse downslope component of motion, whereas finer-than-average grains move preferentially upslope. However, as the forces on bedload grains are expected to vary with discharge, there should be times and places where most grains are moving downslope into the scour zone (falling stage) or upslope out of it (rising stage).

Continuing the analogy with flow in bends, the mean grain size of bedload should increase with bed shear stress. Therefore, the largest mean grain sizes should occur in the base of the scour zone, whereas the finest grains should occur immediately upstream of the scour zone and near the banks adjacent to the downstream end of the scour zone (as observed by Best 1987, 1988). The absence of bedload in the deepest parts of the scour zone observed by Best (1987, 1988) may be due to armoring of the bed and/or the movement of all incoming bedload up the sides of the scour by spiral flows and intense turbulence.

Systematic detailed measurements of suspended sediment load have not been made in braided rivers. Suspended sediment distributions are likely to be very difficult to predict in confluence zones in view of the nature of the zones of mixing, upwelling and flow separation.

Bed configurations

Sediment is transported over braid bars and point bars as distinct bed configurations. These may be microforms, mesoforms or macroforms (terminology of Jackson 1975*b*). The occurrence and geometry of microforms (e.g. ripples) are controlled by inner-zone (viscous sublayer) boundary layer characteristics such as boundary Reynolds number or grain size. Mesoforms (e.g. dunes) are controlled by outer-zone boundary layer characteristics such as flow depth. Dunes include bed configurations that other authors have referred to as sand waves, megaripples, linguoid bars, transverse bars and so on (summary in Allen 1982; Bridge 1985; Ashley 1990). Macroforms (bars) scale with channel width, and their heights are comparable to mean depth of the formative flow. As discussed previously, solitary bedforms with or without slip facies occur on braid bars and point bars. These must be considered to be macroforms as well as the larger bars because they scale with the width of the channel segment in which they occur and have lifespans comparable with these channels. Thus solitary macroforms such as bar heads (Bluck 1971, 1976; Lewin 1976), riffles and tributary-mouth bars, and bar tail 'scroll' bars (Sundborg 1956; Nilsson & Martvall 1972; Jackson 1976*b*; Bridge & Jarvis 1982; Nanson 1980; Fig. 1) all represent various parts of alternate unit bars. The 'longitudinal ridges' adjacent to confluence scours described by Ashmore (1982) are also analogous to scroll bars. Chute bars and deltas occur at the end of channel segments as flow expands into a different segment (Collinson 1970; McGowen & Garner 1970; Smith 1974; Bluck 1976; Lewin 1976, 1978; Levey 1978; Gustavson 1978; Cant 1978; Ashmore 1982; Ferguson & Werritty 1983). The ephemeral 'chutes and lobes' of Southard *et al.* (1984) in shallow gravel-bed rivers are also macroforms. This section is concerned with mesoforms and microforms, the geometry, migration patterns and hydraulic controls of which are discussed at length in Allen (1982), Middleton & Southard (1984), and Southard & Boguchwal (1990).

At high flow stages the most common type of bed configuration on sandy braid bars or point bars is dunes with curved (sinuous and linguoid) crestlines (e.g. Harms & Fahnestock 1965; Coleman 1969; Bluck 1976; Smith 1971*a*; Jackson 1975*a*, 1976*b*; Blodgett & Stanley 1980; Cant 1978; Cant & Walker 1978; Bridge & Jarvis 1982; Crowley 1983; Bridge & Gabel 1992). Straight-crested forms are common on the higher parts of bars where flow velocity and depth are relatively low, and upper-stage plane beds occur locally in shallow areas of high velocity. Ripples are normally restricted to areas of slow-moving water near banks. In gravelly streams, high-flow bedforms include more straight-crested dunes, 'bedload sheets' (which are low-relief asymmetrical bedforms analogous to dunes; Hein & Walker 1977; Kuhnle & Southard 1988; Whiting *et al.* 1988; Dietrich *et al.* 1989; Bennett 1992), lower stage plane beds and transverse ribs (analogous to antidunes; Koster 1978; Allen 1982; Ferguson & Werritty 1983). Pebble clusters are common on lower stage plane gravel beds (Dal Cin 1968; Teisseyre 1977*a*; Brayshaw *et al.* 1983; Brayshaw 1984, 1985; Naden & Brayshaw 1987) and static and mobile armor layers commonly develop. Discoidal and tabular gravel grains on lower stage plane beds and the backs of bedforms like bedload sheets are normally imbricated (dip upstream), whereas those that accumulate on the avalanche faces of dunes commonly dip downstream (pseudoimbrication).

Dune crestlines at high flow stage are commonly oblique to local channel direction, being particularly obvious for straight- and sinuous-crested forms (Fig. 14). This is due to across-channel variation in bedload transport rate and dune height, in turn associated with the flow pattern over braid bars. The orientations of dune crestlines mimic the orientations of alternate unit bars (Figs 1 and 14), indicating that the pattern of flow and sediment transport over ancestral alternate bars has not been modified greatly in the development of braid or point bars. Thus the crestlines of these bedforms are not in general normal to directions of sediment transport or near-bed flow velocity (Fig. 1). However, the orientation of spurs in dune

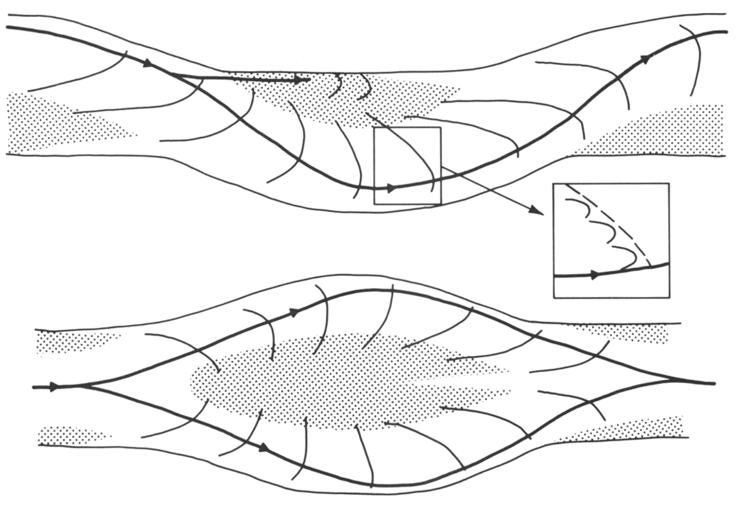


Fig. 14. Idealized pattern of dune crestlines and flow velocity in braided and unbraided channels at high flow stage, based on numerous studies of rivers (e.g. Hein & Walker 1977; Cant 1978; Cant & Walker 1978; Deitrich *et al.* 1979; Blodgett & Stanley 1980; Bridge & Jarvis 1982; Crowley 1983). Thick black lines with arrows are loci of maximum flow velocity and mean grain size. Dune crestlines are indicated schematically as sinuous to straight crested; however, the inset shows the equivalent orientation of linguoid dunes.

troughs is generally parallel to local flow direction. This has relevance to the orientation of cross-stratification, as discussed later.

The orientation of bedform crestlines under zones of flow separation with vertical axes may be highly oblique to local channel orientation, and the bedforms might even migrate upstream. Upstream migration of bedforms might also be associated with the downstream parts of some abandoned channel segments. In these cases, the crestline orientations are controlled by the orientation of local flow vectors, and not necessarily by across-channel variations in bedload transport rate and bedform height, as above.

Longitudinal bedforms (crestlines parallel to local flow direction) also occur in rivers (reviewed by Allen 1982). The most common are ribbons of sand moving over lower-stage plane gravel beds, but longitudinal sand ridges have also been reported on upper-stage plane beds (e.g. Coleman 1969).

At flow stages lower than bankfull, dunes are generally shorter and lower, and the proportion of curved crested dunes decreases relative to straight-crested dunes, ripples and lower-stage plane beds (Coleman 1969; Singh & Kumar 1974; Jackson 1975*a*, 1976*b*; Bridge & Jarvis 1976, 1982; Gustavson 1978; Crowley 1983). Dune geometry may not be in equilibrium with changing flow stage. Gabel (1993) studied the geometry and dynamics of dunes over changing flow stage in a braided reach of the sandy Calamus River in Nebraska. Dune heights, lengths, migration rates, creations and destructions were measured concurrently with bedload transport rates, flow depth, flow velocity, and bed shear stress during a series of day-long surveys. These surveys were conducted sequentially during changing discharge at different sites around a braid bar. Within each survey, individual dune heights, lengths and migration rates as seen in streamwise cross sections varied greatly with time, due to local changes in flow conditions and the 3D shapes of the dunes. Over periods of several days, mean dune geometry and migration rate changed in response to changing flow conditions with a small time lag. Because this time lag was small, mean dune lengths and steepnesses did not deviate much from theoretical and empirical equilibrium values (e.g. Yalin 1977; Fredsoe 1982). Theoretical models for changes in dune height and length with changing discharge (e.g. Allen 1976; Fredsoe 1979) performed with varying degrees of success, pointing to a need to substantially improve such models.

Large dunes are commonly not in equilibrium with rapidly falling stages, and this is particularly true in shallow areas with small transverse bed slopes which can become exposed rapidly by small decreases in water level. There has been a tendency to define emergent dunes or solitary dune-like macroforms as braid bars, and the channels that dissect these bedforms as braided channels (e.g. Smith 1971*a*, 1974; Blodgett & Stanley 1980). Actually, dissected dunes may superficially resemble solitary macroforms because dunes in disequilibrium cannot be expected to show regular repetitive patterns (Allen 1982). Furthermore, straight-crested dunes normally have long wavelength/height and extremely variable sizes (e.g. Costello & Southard 1981).

The emergence of braid and point bar areas during falling flow stages, the modification of flow patterns, and the dissection of emerging areas by small channels, results in a diverse orientation of the superimposed bedforms. The dendritic channel systems that develop parallel to the local slopes of emerging bars may terminate in small deltas. Desiccation cracks occur in local areas of mud deposition (e.g. in bedforms troughs and chute channels), and the encroachment of animals and plants on to emerging bar surfaces results in root casts, burrows and trails.

Finally, distinct bed configurations may occur locally in areas where net erosion is proceeding, especially in the thalwegs of channels where cohesive mud is present. Such erosional marks include flutes, gutter casts, rill marks and potholes (reviewed by Allen 1982).

Grain size-density sorting associated with bedforms

The size sorting of sand and gravel fractions associated with bedload sheets and dunes is discussed by Smith (1972, 1974), Allen (1982), Ikeda (1983), Iseya & Ikeda (1987), Kuhnle & Southard (1988), Whiting *et al.* (1988), Dietrich *et al.* (1989), Wilcock & Southard (1989), Chiew (1991), and Bennett (1992). At low sediment transport and supply rates the trough areas of bedforms are commonly relatively coarse grained and armored, whereas the finer grains comprise the moving bedforms. As sediment transport rate increases the coarser grains can be transported to the bedform crests. If sand supply is abundant the pore spaces between gravel grains are occupied by sand, especially in the bedform trough areas. Gravel beds with sand filling the interstices are relatively smooth, which facilitates the movement of the larger grains (e.g. overpassing).

Heavy minerals transported as bedload and suspended load are commonly finer grained than associated light minerals. They may be concentrated where a flow which is powerful enough to entrain and transport all grain fractions decelerates and deposits some of them. The relatively small heavy minerals become protected from re-entrainment by the larger light minerals which can continue to move above them. Subsequent erosion may remove the larger light minerals but not the smaller heavies. Locations where such conditions prevail include the crestal and trough areas of a range of bedforms with or without flow separation (e.g. ripples, dunes, low-relief bedwaves such as bedload sheets, and bars) and in the lee of obstacles where deposition occurs (e.g. McQuivey & Keefer 1969; Brady & Jobson 1973; Slingerland 1977, 1984; Mosley & Schumm 1977; Minter 1978; Smith & Minter 1980; Buck 1983; Cheel 1985; Best & Brayshaw 1985; Slingerland & Smith 1986; Kuhnle & Southard 1990). Heavy mineral concentrations on the crests of bedforms are transient in view of the nature of bedform migration. However it is common to find heavy mineral concentrations above the erosion surfaces formed by migration of the trough regions of various scales of bedwave. At the bar scale, heavy mineral concentrations are common in channel thalwegs and bar heads (e.g. Smith & Minter 1980; Smith & Beukes 1983).

Erosion and deposition at the bar scale

Erosion and deposition at the bar scale are related to three main processes which are closely related to each other.

(1) Adjustments of the bed topography of braid and point bars as a result of seasonally changing discharge (e.g. Fig. 15). During rising flow stages erosion tends to occur in bend thalwegs, confluence scours and the upstream ends of bars, whereas these areas receive deposits during falling stages. In contrast, the downstream and topographically highest parts of bars tend to be areas of deposition at high flow stages, with erosion at low flow stages. It is important not to confuse the steep eroded downstream ends of bars with depositional avalanche faces.

(2) Bank erosion and deposition on adjacent bar margins (i.e. channel migration) are normally associated with (1). Such channel migration is episodic, and the deposition may be in the form of distinct unit bars (but not necessarily).

(3) Cutting of new channels, enlarging existing channels, abandonment and filling of others is closely associated with (2), as accreting bars commonly migrate into channel entrances, and low areas adjacent to discrete unit bars are common conduits for diverted discharge.

These processes are discussed below, followed by the nature of the deposited sediment.

Processes of erosion and deposition

In general, erosion and deposition are due to gradients of sediment transport rate, given by the sediment continuity equation:

$$-C_o \frac{\delta h}{\delta t} = \frac{\delta i_s}{\delta s} + \frac{\delta i_n}{\delta n} + \frac{1}{u_s}\frac{\delta i_s}{\delta t} + \frac{1}{u_n}\frac{\delta i_n}{\delta t} \tag{8}$$

where C_o is volume concentration of sediment in the bed, h is bed elevation, t is time, s and n are streamwise and across-stream space co-ordinates, i_s and i_n are streamwise and across-stream sediment transport rate by volume, and u_s and u_n are corresponding mean velocities of sediment grains. If bed topography is in equilibrium with a steady flow the term on the left hand side, and the last two terms on the right hand side, are zero when averaged over turbulence and bedforms of smaller scale than bars. As discharge varies, the bed topography is always potentially out of equilibrium with the flow. For example, as stage falls in curved channel segments, the cross-channel bed slope is too steep for the reduced flow and the transverse downslope gravity force on the bedload exceeds the upslope directed fluid drag. Thus, in the thalweg, the magnitude of $\delta i_n/\delta n$ exceeds $\delta i_n/\delta s$, the last two terms on the right hand side of equation (8) are generally negligible, and deposition occurs as described in (1) above. However, even in steady flows, $\delta i_s/\delta s$ and $\delta i_n/\delta n$ vary with time due to the passage of bedforms like unit bars, dunes, bank slumps, and so on.

The tendency for erosion in thalwegs near banks during floods leads to oversteepening and may induce failure (e.g. Turnbull *et al.* 1966; Thorne & Tovey 1981; Pizzuto 1984). The important factors governing bank erosion are the resistance of the banks to slumping and the ability of the flow to remove the slumped bank material. The shear strength of bank material, τ_s, is given by

$$\tau_s = C + P\mu_c \tag{9}$$

where C is cohesive strength, P is effective normal stress and ϕ_c is the static friction coefficient. The cohesive strength is increased by the presence of clay minerals, vegetation and cemented layers such as calcretes and silcretes. The effective normal stress depends on the weight of the potential slump and the excess

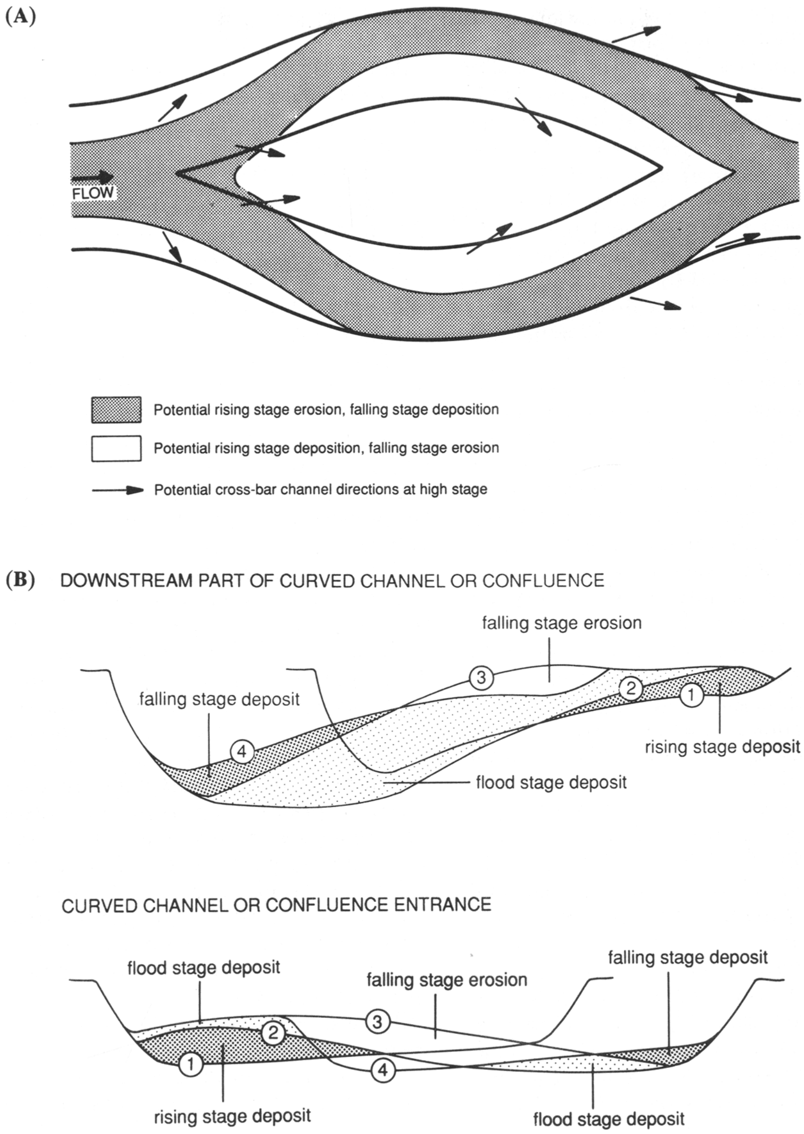


Fig. 15. (**A**) Theoretical locations of erosion and deposition in simple braided channel pattern during changing flow stage. (**B**) Cross-sections showing theoretical erosion and deposition during changing flow stage and channel migration. (1) low stage geometry; (2) flood stage geometry; (3) flood stage geometry following cut bank erosion and bar deposition; (4) final low stage geometry.

pore pressure. The excess pore pressure is increased (hence P is decreased) during falling flow stages when water flows out of the banks. The effect of undercutting is to increase the downslope component of the gravity force on bank material, which is the main force responsible for the downslope directed shear stress.

Slumps of bank material are typically on the order of 10^{-1}–10^{0} m wide, 10^{0}–10^{1} m long (i.e. order of 10^{0}–$10\,m^{3}$) and occur periodically in time and space. Upon failure, slumps may remain coherent or disintegrate and become grain flows. The slumped material may be entrained by the flow and, if not, will contribute to armoring of the bed and protection of the bank against further erosion. Thus rates of bank erosion must depend on magnitudes of $\delta i_s/\delta n$

and $\delta i_n/\delta n$ at the toe of the bank, which must in turn be controlled by the spatial gradients of $\tau_o - \tau_c$, where τ_o is bed shear stress and τ_c is the critical entrainment stress. In the case of banks with large shear strength, providing large cohesive slump blocks, τ_c is expected to be large, thereby inhibiting erosion. Banks of cohesionless sand and gravel are the most easily erodible, and bank erosion rates can be the order of channel widths per year (i.e. 10^1–10^3 m a^{-1}) (e.g. Chien 1961; Coleman 1969; Bluck 1979; Werritty & Ferguson 1980; Ferguson & Werritty 1983; Ashworth & Ferguson 1986; Bristow 1987; review in Allen 1982).

A number of theoretical models relate bank erosion and channel migration in single curved channels to the nature of flow and sedimentary processes (e.g. Ikeda *et al.* 1981; Parker & Andrews 1986; Johannesson & Parker 1989; Hasegawa 1989; Odgaard 1989; Howard 1992). Largest erosion rates are normally associated with the largest near-bank velocity or depth, which normally occur just downstream of the bend apex. Unfortunately, these theoretical approaches are overly simplistic in that they do not consider flow unsteadiness and the unpredictable entrainment stresses of slump blocks. Furthermore, they normally contain empirical coefficients, the values of which must be 'fitted' to field data.

As channel geometry tends to be in equilibrium with high stage flow and sediment transport (Bridge 1984, 1992; Bridge & Jarvis 1982; Bridge & Gabel 1992) cut-bank retreat and associated flow expansion will inevitably result in deposition on adjacent bar surfaces. Likewise, if deposition occurs on a bar due to a local increase in sediment supply from upstream, narrowing of the channel cross section may induce erosion of the adjacent cut bank. Such depositional activity may be distinctly episodic as in the accretion of 'unit bars' to braid or point bars, but more gradual accretion may occur which does not involve 'unit bars'. The familiar 'accretion topography' results from the sequential accretion of unit bars (e.g. scroll bars).

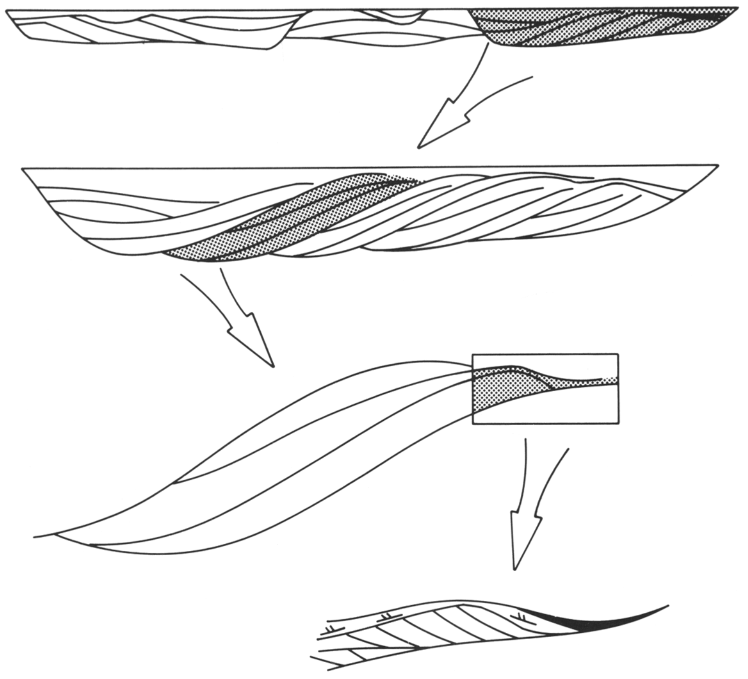
There have been many studies of the patterns of channel migration for single curved channels (reviewed in Allen 1982; see also Elliott 1984; Ikeda & Parker 1989) but fewer for braided rivers (Coleman 1969; Bluck 1974, 1976, 1979; Werritty & Ferguson, 1980; Ferguson & Werritty 1983; Ashworth & Ferguson 1986; Bridge *et al.* 1986; Bristow 1987; Ashmore this volume). The nature of channel migration is critical to understanding the preservation of channel deposits. As flow patterns in curved channel segments vary with flow stage, so also should channel migration patterns. For instance, if most erosion and deposition occurs around bankfull stage (e.g. gravel-bed streams), the large flow velocities near the outer banks of the downstream half of the bends should result in dominantly downstream bend migration. This would lead to preferential preservation of bar tail scrolls and erosion of bar heads. If substantial erosion and deposition are associated also with lower flow stages (e.g. sandy rivers) the largest flow velocities will also act upon cut bank locations all around the bend such that bend expansion may occur in addition to downstream translation. In this case, preservation of bar head units is possible, and abandonment of the curved channel by cut-off is more likely. If the upstream part of a bar is stabilized by vegetation, the bar may accrete and grow by lateral and downstream extension even if most erosion and deposition occur at high discharges (e.g. Calamus River; Bridge *et al.* 1986).

Confluence zones commonly migrate downstream in response to downstream migration of entering channels, especially for symmetrical confluences of similar channels (e.g. Ashmore, this volume). The lack of room for lateral migration of both side bars may lead to cutting of new (chute) channels, and the downstream migration of side bars may lead to blocking of the entrance of a downstream braid channel. Lateral migration of confluences is possible in the case of asymmetrical confluences, and those with entering channels of different discharge and/or geometry. Ashmore (this volume) also documents the expansion, rotation and obliteration of confluences in response to changes in the relative discharge or sediment supply of the entering channels.

Cutting and filling of channels is particularly characteristic of braided rivers although this occurs in all rivers to some degree. The processes are not well understood and no theoretical models are available. Channel diversion appears to be associated with high-stage scouring of thalwegs in curved channel segments and deposition of sediment as obstacles in braided channel entrances downstream (e.g. as riffles and bar heads) (Teisseyre 1977*b*; Ashmore 1982, 1991*b*; Bridge 1985). Once the process of channel abandonment is initiated the channel is progressively filled from the upstream end with bedload. Downstream migration of the adjacent enlarging channel causes blockage of the downstream end of the filling channel by the downstream tip of the enlarging channel's bar. In effect, this process represents the overriding of a stalled bar by an upstream bar, as happens with dune migration. The abandoned channel may be reopened later, especially where a concave channel bank (with high water level) approaches its entrance (e.g. Werritty & Ferguson 1980; Ferguson & Werritty 1983; Church's 1972 'secondary anastomosis').

According to Ferguson & Werritty (1983) and Bristow (1987) channel migration by bar erosion and accretion in braided rivers is much more common than by channel switching. Apparently, cross-bar channels are most susceptible to switching position (Bristow 1987). This is understandable as cross-bar channels commonly occur between the discrete unit-bar accretions to braid and point bars which form at high flow stage. Examination of the accretion topography in modern braided rivers and adjacent floodplain areas (e.g. see photos in Collinson, 1970; Mosley 1982*b*, Bridge *et al.* 1986; Cant & Walker 1978; Bluck 1979; Shelton & Noble 1974) clearly indicates that bar accretion deposits are dominant over channel-fill deposits, and that the downstream component of bar migration is dominant over the lateral and upstream components associated with bar growth (expansion).

(A)



(B)

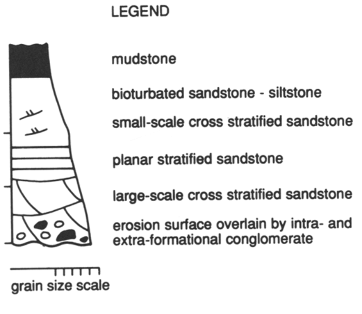


Fig. 16. (**A**) Different scales of braided river deposits. Top diagram is a cross section through the entire channel belt showing four separate channel bar deposits (storeys) and associated channel fills. A single channel bar and fill (stippled) is expanded to illustrate the seasonal accretionary units (bedsets). These inclined bedsets are commonly referred to as lateral-accretion deposits, although there is normally a component of downstream accretion also. See Fig. 15. The internal sedimentary structures associated with part of a seasonal bedset are shown in the lower diagram. The structures are associated with flood stage migration of dunes and falling stage migration of ripples. An organic rich mud drape occurs in the swale adjacent to the accretionary ridge. Symbols explained in legend. (**B**) Legend for Figs 16A, 18, 20, 24 to 26.

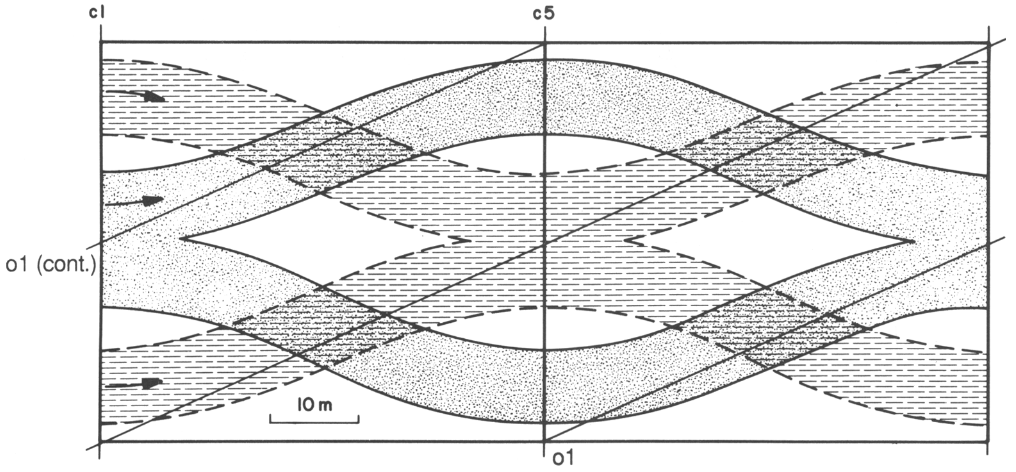


Fig. 17. Plan geometry of braided channel for the first model case. Initial channel position is stippled, whereas migrated channels are marked by dashed-line ornament. Cross-sections c1, c5 and o1 are shown in Fig. 18.

Quantitative depositional models

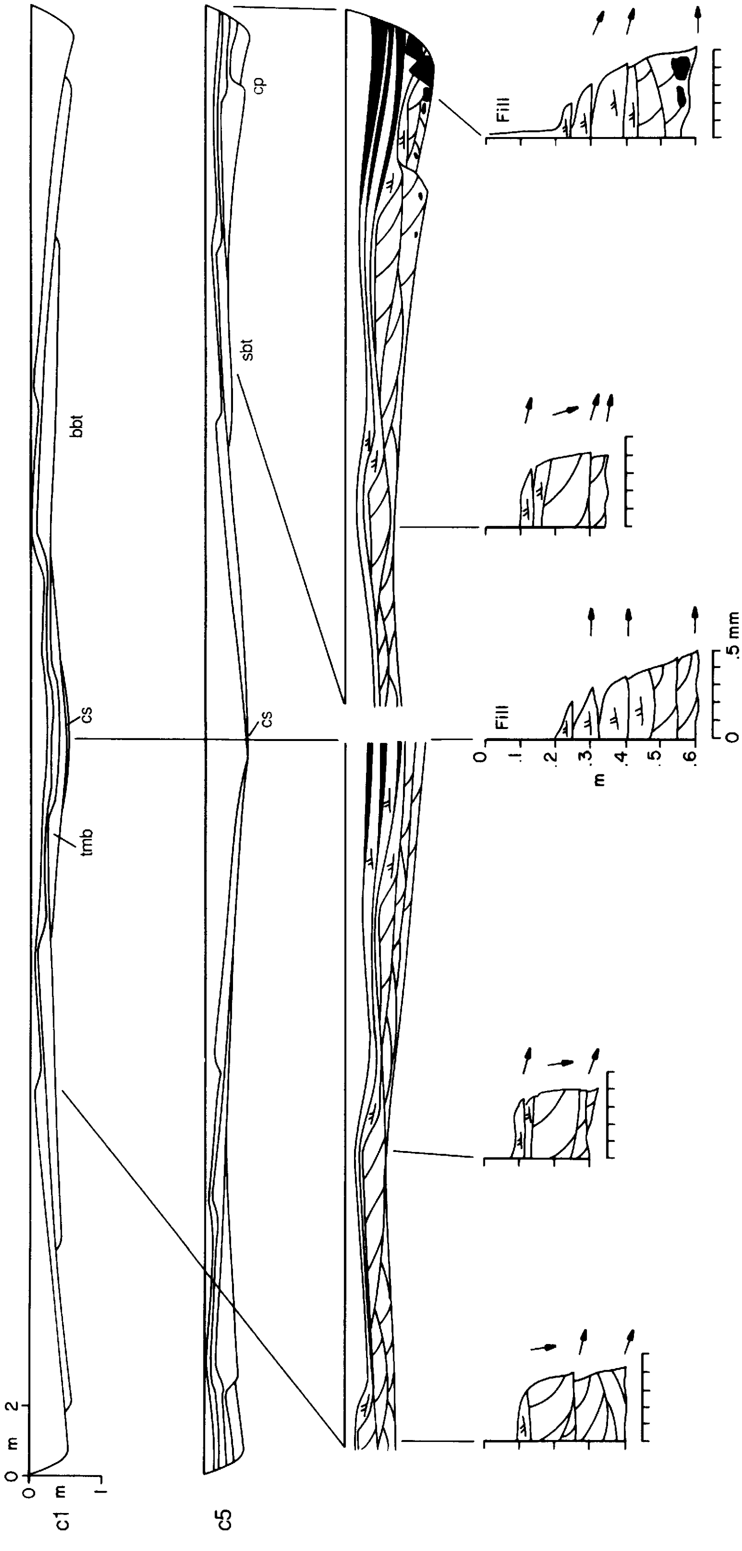
The previous discussion suggests that there should be different scales of depositional unit in braided rivers, depending on the scale of the associated topographic feature and the time over which deposition occurred (Fig. 16). For the purposes of this discussion four scales of deposition are recognized: (1) the complete channel belt; (2) the channel bar and adjacent fill deposits of major and minor channel segments; (3) seasonally controlled depositional increments on channel bars and fills; (4) increments of deposition associated with passage of discrete bed waves such as dunes, ripples, and bedload sheets.

An understanding of how these different scales of deposits are distributed in space requires detailed understanding of the bed geometry, flow, sediment transport and bed configurations during erosion and deposition, and the style of channel migration. Using the previously-discussed information on these topics, and the somewhat limited direct observations of channel deposits from modern rivers, it has been possible to start the construction of both quantitative and qualitative depositional models. Although these models are simplified and do not predict the full details of deposition in braided channel belts, they represent a fundamental advance over previous models in that they are 3D, and include most of the main sedimentological properties that can be observed.

Quantitative models at present can only be constructed for the simplest channel geometries and modes of channel migration. In the first case a single braid bar is bounded by two identical curved channels placed side by side (Fig. 17). The bed topography, flow and sediment transport in these channels are described using Bridge's (1992) model for the case of steady, equilibrium flow in single curved channels. Bankfull flow conditions are assumed (details in Table 6). Such channel geometry and flow conditions were chosen as an analogue of the

Table 6. *Geometry and flow conditions for model channels*

	Initial channels in both models	Final channel in second model	
Channel width from thalweg to inner bank (m)	7.5	10.0	5.0
Centreline wavelength in one wavelength (m)	103.0	110.0	103.0
Sinuosity	1.03	1.1	1.03
Mean depth (m)	0.38	0.52	0.25
Centreline water surface slope at crossovers	0.00097	0.00091	0.00097
Darcy-Weisbach friction coefficient	0.1	0.1	0.1
Mean flow velocity ($m\ s^{-1}$)	0.55	0.61	0.44


c1
c5
bbt
cs
tmb
sbt
cp
Fill
.5 mm
m

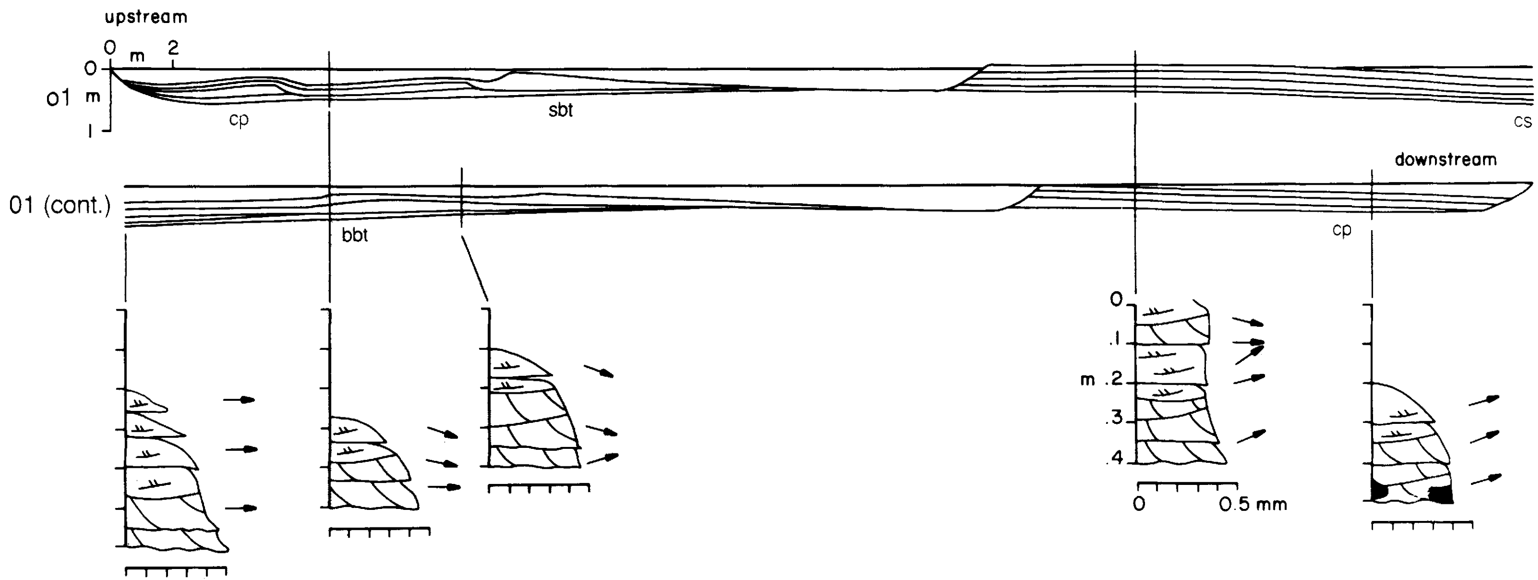


Fig. 18. Cross sections showing channel geometry and deposits formed by channel and bar migration shown in Fig. 17. The lines in the upper cross sections represent the high-stage bed surfaces at the end of each discrete increment of deposition (i.e. lateral and/or downstream accretion surfaces). For simplicity, these surfaces are smoothed (mesoform topography not shown), and cross-bar channels and falling-stage modifications are not shown. cs, confluence scour; tmb, tributary-mouth bar deposits; cp, counterpoint deposits; bbt, braid bar tail deposits; sbt, side bar tail deposits. More details of the spatial variation of mean grain sizes, internal structures and paleocurrent directions are shown in the lower parts of the diagrams. See Fig. 16B for legend.

Calamus River (Bridge *et al.* 1986; Bridge & Gabel 1992). The channels are then allowed to migrate downvalley in four discrete steps for a total distance of half a bend wavelength. During this downvalley migration channel geometry is held constant, and it is assumed that most deposition occurs during the constant bankfull flow conditions. In order to simplify the model, no cross-bar channels were included, and no net vertical deposition occurred.

The deposits are represented in Fig. 18 by two cross-stream sections and one oblique section which is almost alongstream. The deposits are primarily represented by the bounding surfaces which separate the four discrete depositional increments. The local waviness of these surfaces reflects the preservation of unit bars. The internal features of these depositional units are indicated in places by vertical logs or 2D sections. Internal structures reflect the mesoforms or microforms present during deposition, which were determined using Southard & Boguchwal's (1990) diagrams. Palaeocurrent orientations for trough cross strata are taken as parallel to local channel orientation whereas those for planar cross-strata reflect the crest-line obliquity of the formative straight crested bedform (e.g. scroll bars).

Section c1 (Fig. 18) represents the migration of the widest braid bar section into a confluence. The lowest, central deposits lie above the confluence scour, and the uppermost and marginal deposits are associated with the braid bar apex. Tributary-mouth bar deposits occur immediately above and to the side of the maximum confluence scour (Fig. 18) and braid-bar tail (including scroll-bar) deposits occur above and to the side of these deposits (compare with Bristow *et al.*'s (1993) model of confluence deposits). The thickness of the bar deposits can vary across the section by up to a factor of 2. The inclination of the major bedding surfaces increases outwards as predicted in Willis' (1989) point bar models. The widths of these surfaces vary as the projected channel widths vary. In general, the thickest depositional units bounded by these bedding surfaces are those where the change in bed elevation in a migration step is the greatest (e.g. associated with the downstream tip of the bar). If the migration rate was less there would be more depositional increments in a section, and the increments would be thinner. The vertical logs show either little vertical change in mean grain size or fining upwards.

Section c5 (Fig. 18) is where a confluence migrated into a widest braid bar section. The central parts of the section represent the confluence scour, whereas the outer parts are side bar deposits. Although not shown in Fig. 18, these deposits may show evidence of large-scale flow separation (e.g. upstream-dipping cross strata). The outermost, lowest deposits are counterpoint deposits (e.g. Carey 1969; Taylor & Woodyer 1978; Hickin 1979; Lewin 1982, 1983; Page & Nanson 1982; Nanson & Page 1983; Smith 1987) which represent the most downstream parts of scroll bars.

In section o1 (Fig. 18) the lowest surface represents the downstream transition from channel margin at a bend apex through a cross-over to a confluence zone to another bend-apex channel margin on the opposite side of the channel belt. The deposits on the left hand side of the section represent the downstream migration of a side bar into the bend-apex section, and thus comprise mainly fining upward 'counter-point' deposits. The central parts of the cross section represent the migration of the braid bar into the confluence. The vertical section in this region that fines upward then coarsens upward represents the upward progression from cross-over (riffle) to bar tail to bar head deposits. The right hand side of the section represents counter-point deposits as seen in a view almost parallel to local flow direction.

If all of the palaeocurrent orientations in all of these vertical logs are considered (excluding those from straight crested bedforms like scroll bars) they give an accurate record of the range of channel orientations. However, the range of palaeocurrent orientations in single vertical logs does not.

In the second quantitative model, the initial conditions are the same as the first case. However, as the channels migrate downvalley by a quarter bend wavelength in two steps, one channel increases in sinuosity and width, whereas the other decreases in width (Fig. 19). The large-scale bedding patterns in section c3 (Fig. 20) are fairly similar to the younger deposits of section c1 in Fig. 18. The increase in inclination of the bedding surfaces and thickening of the deposits on the left hand side of section c3 reflect a change from a bar tail to a bar apex section and an increase in sinuosity and channel size. The increasing channel width may not be apparent in the widths of successive inclined bedding surfaces as these are apparent widths. The vertical log here shows little change in mean grain size. On the right hand side of section c3 the reduction in channel size here is reflected in a decreasing deposit thickness and inclination of large-scale bedding surfaces, and a fining upwards trend, even though a bend apex migrated over a bar tail region. Section c7 (Fig. 20) is similar to the central parts of section c5

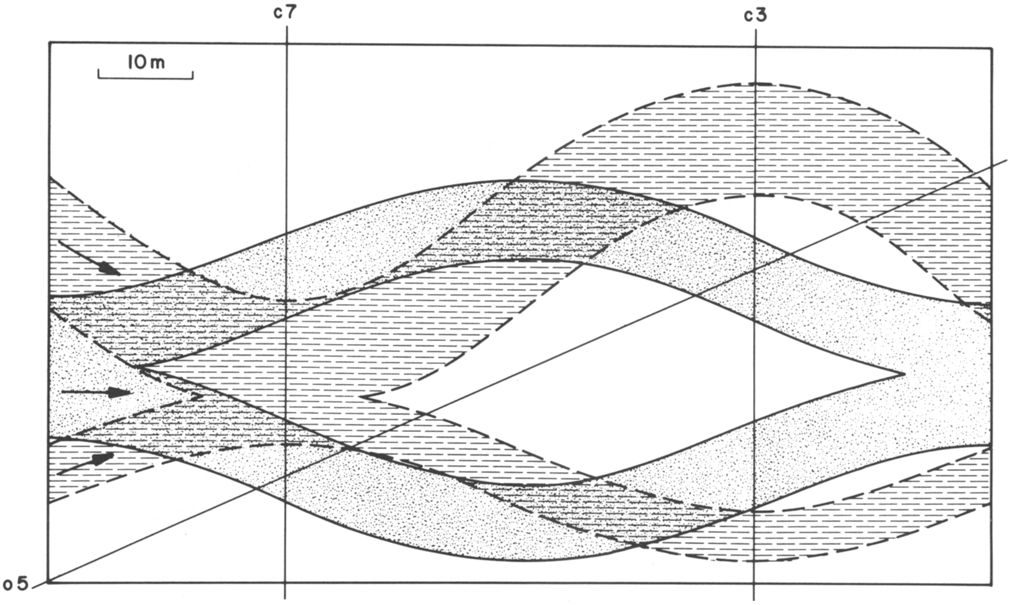


Fig. 19. Plan geometry of braided channel for the second model case. Initial channel pattern is stippled, whereas migrated channels are marked by dashed-line ornament. Cross-sections c3, c7 and o5 are shown in Fig. 20.

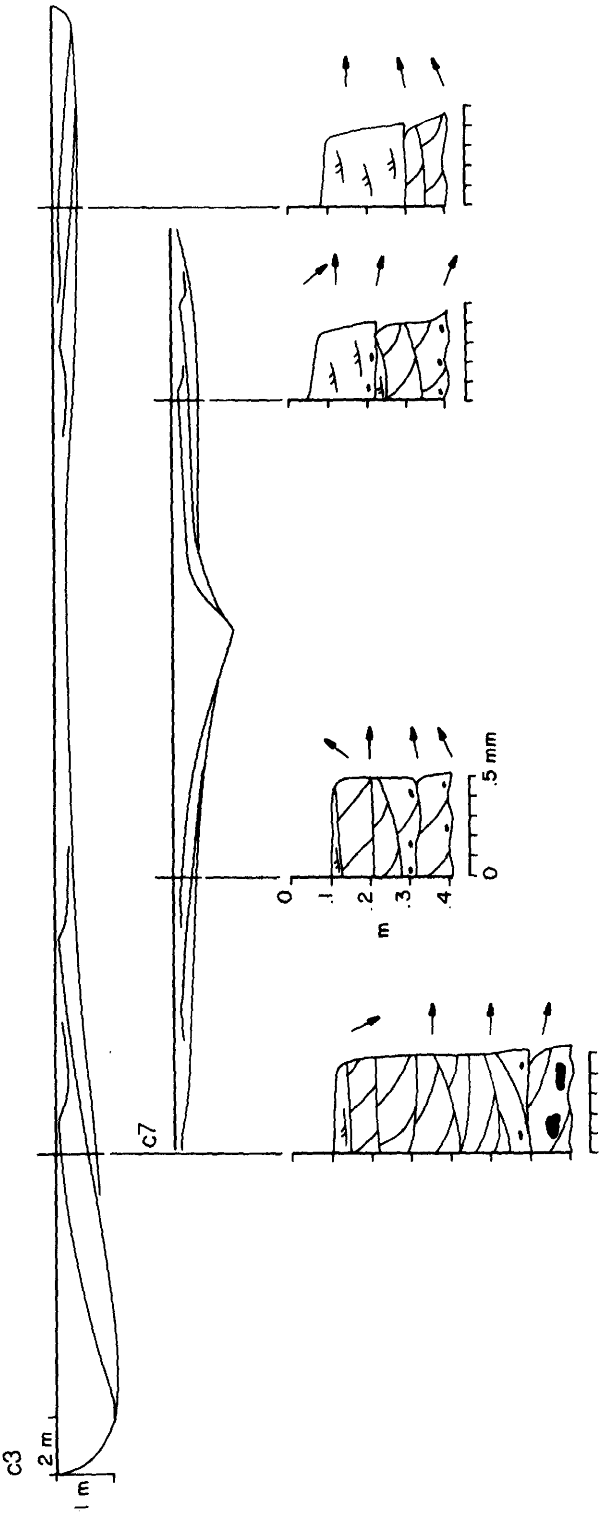
(Fig. 18) except for the markedly asymmetrical confluence scour in c7.

Qualitative depositional models

The two quantitative models are very specific, and more generality can only be achieved at present with qualitative models. The models in Figs 21 to 23 emphasize large-scale bedding geometry associated with episodic seasonal accretion on bars and in channel fills. For simplicity, cross-bar channels have not been included, there is no net vertical accretion, and the later stages of vertical accretion and channel filling (associated with channel-belt abandonment) have not been included. Figure 21 shows how downstream translation of bars results in preferential preservation of bar tail deposits, and erosional truncation of bar head deposits. With bar expansion and limited downstream translation (Fig. 22) bar-head deposits can be preserved, and erosional truncation of deposits is not as marked. Figure 23 shows how channel enlargement and filling results in subtle erosional truncations and changes in apparent channel geometry. In this example, the filling channel contains small bars and is being blocked at both ends by bar deposits. Detailed facies of these deposits are discussed below.

By reference to the quantitative models of Figs 17–20, fining upward sequences within bar deposits are expected to be most common and occur where point (side) bar tails build into bend apices, and where braid bar tails build into confluences (e.g. Shelton & Noble 1974). The vertical variation in mean grain size increases as channel sinuosity increases. Bar sequences with little vertical variation in mean grain size occur where bend apex areas build over bar tails. Such sequences may coarsen at the top if the bar head migrates over bar tail deposits (e.g. Bluck 1976).

The internal structure of the discrete accretionary units in sandy braid and point bars should be dominated by large-scale trough cross-stratification in view of the ubiquitous presence of curved-crested dunes mentioned previously (e.g. Harms *et al.* 1963; Allen 1965, 1982; Harms & Fahnestock 1965; McGowen & Garner 1970; Blodgett & Stanley 1980; Jackson 1976*a*; Bridge & Jarvis 1982; Bridge *et al.* 1986; Sarkar & Basummalick 1968). Gravelly deposits may have relatively more planar stratification (with imbrication) and planar cross-stratification (Williams & Rust 1969; Smith 1970, 1974; Rust 1972; Boothroyd & Ashley 1975; Bluck 1976, 1979; Boothroyd & Nummedal 1978; Gustavson 1978; Ferguson & Werritty 1983). Planar cross-stratification arising from straight-crested dunes is commonly described from the upper parts of braid bars and point bars (Sundborg 1956; Sarkar & Basumallick 1968; Smith 1970, 1971*a*, *b*, 1972; Jackson 1976*b*; Cant 1978; Cant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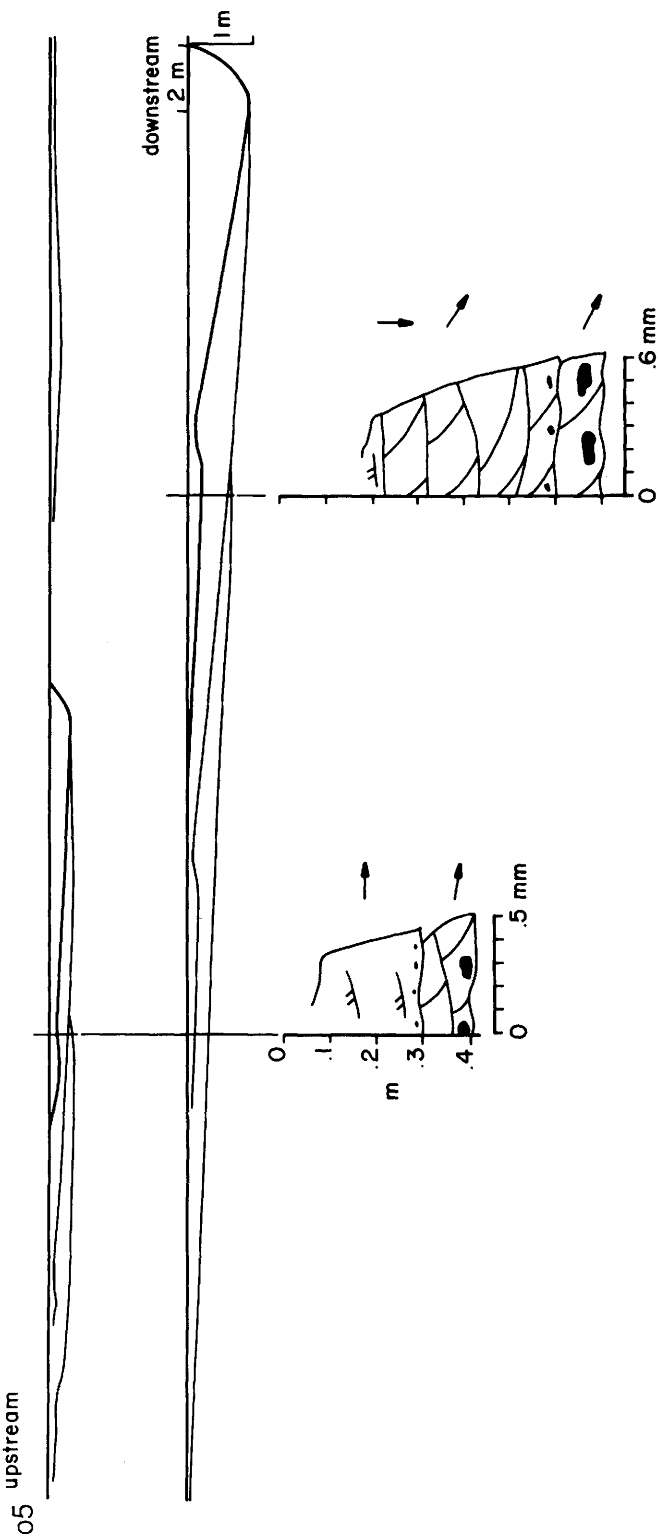

Fig. 20. Cross sections showing channel geometry and deposits formed by channel and bar migration shown in Fig. 19. See caption of Fig. 18 for explanation.

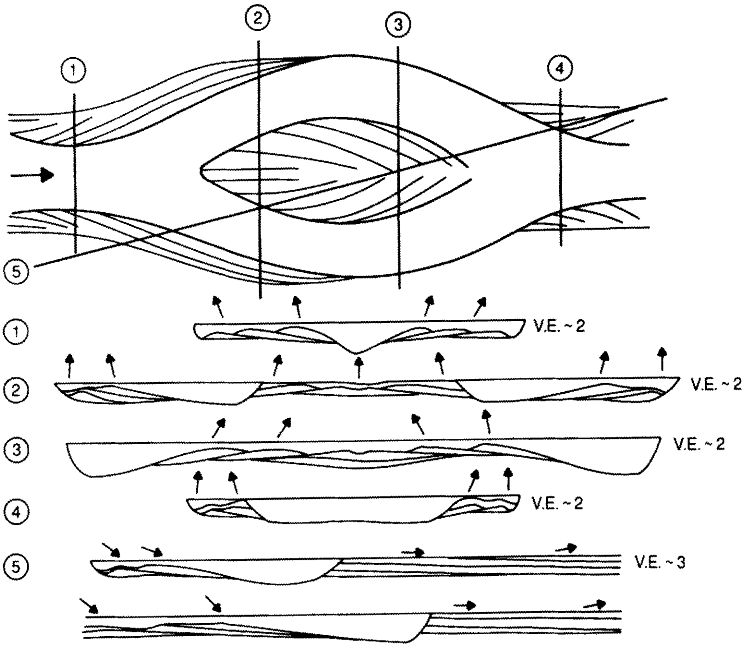

Fig. 21. Qualitative depositional model associated with a simple braided channel pattern and downstream migration, but no vertical deposition. Lines in plan and cross section are smoothed boundaries of seasonal accretionary bedsets (no mesoforms or cross-bar channels shown). Arrows represent channel orientation during deposition of uppermost bedset in the cross section (lower bedsets will have different orientations). The spatial variations in mean grain size can be reconstructed from Figs 17 to 20. Vertical exaggerations (V.E.) are approximately 2 to 3, and channel belt widths in nature will range from tens to thousands of meters.

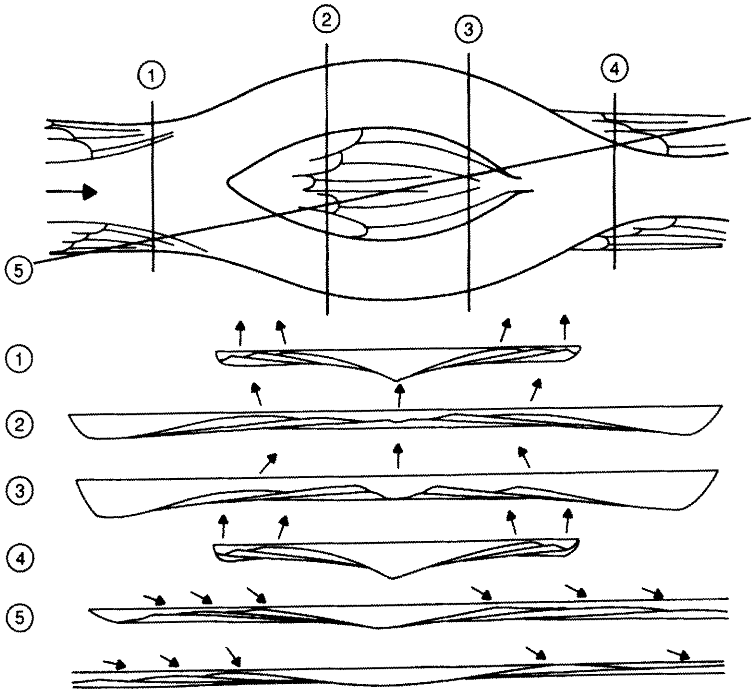

Fig. 22. Qualitative depositional model associated with a simple braided channel pattern, downstream migration and bend expansion. See caption for Fig. 21.

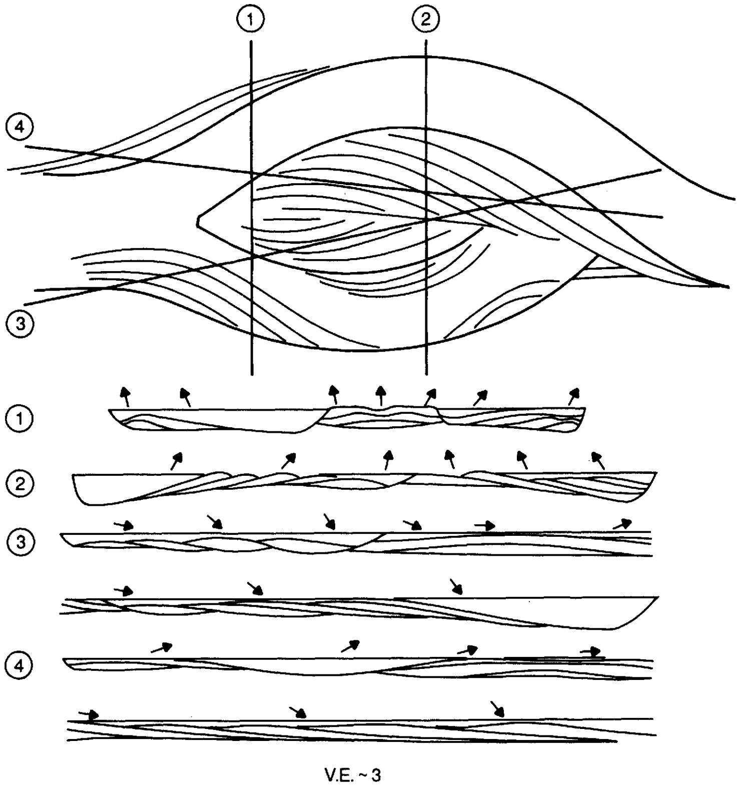

Fig. 23. Qualitative depositional model associated with a simple braided channel pattern, downstream migration, bend expansion, and channel switching. The lower channel has started to fill, and the upper channel is expanding. See caption for Fig. 21.

& Walker 1978; Crowley 1983) but it is not necessarily restricted to this position.

Exceptionally large scale, isolated sets of planar cross strata which comprise most of the thickness of an accretionary unit may represent parts of unit bars (i.e. bar heads, scroll bars) or chute and tributary-mouth bars. Examples of these structures where they have been dissected on the upper parts of bars are given by Collinson (1970), Jackson (1976*b*), Bluck (1976, 1979), Cant & Walker (1978), Blodgett & Stanley (1980), and Crowley (1983). Bar-head and scroll-bar planar sets tend to occur near the top of a bar sequence, whereas tributary-mouth bar (riffle) planar sets should occur nearer the base (e.g. Fig. 18, but compare with Bristow *et al.* 1993). If deposition is associated with downstream migration of alternate bars with avalanche faces in straight channels it is expected that most of the deposits will be composed of a single set of planar cross-strata (e.g. Smith 1970, 1971*a*, 1972; Blodgett & Stanley 1980; Crowley 1983).

Planar strata are common in the tops of sandy braid and point bars (e.g. Harms *et al.* 1963; Sarkar & Basumallick 1968; Coleman 1969; Smith 1971*b*; Shelton & Noble 1974; Cant & Walker 1978; Bristow this volume) but there is little information on their occurrence elsewhere in a bar sequence. Cross laminated sand from ripple migration tends to be restricted to high areas near banks, and in other positions as falling stage deposits. Near banks, cross-laminated, bioturbated sand commonly occurs interbedded with vegetation-rich mud as cm-thick units ('accretionary bank deposits' of Bluck 1971).

The deposits of channel fills are poorly known from modern rivers and details are not included in the depositional models presented here. If the angle between the enlarging and filling channel segment is relatively small, as in low sinuosity rivers, flow is maintained in the filling channel so that bedload can be deposited particularly at the channel entrance. Although bedload may extend a considerable way into such filling channels the downstream ends will receive mainly fine grained, suspended sediment and organic matter from slowly moving water (e.g. Fisk 1947; Teisseyre 1977*b*; Bridge *et al.* 1986). With larger angles of divergence, both ends of the abandoned channel are quickly blocked so that most of the channel fill is relatively fine grained and organic-rich due to suspension deposition from ponded water (Fisk 1947).

Channel fills should generally fine upwards reflecting progressively weaker flows (e.g. Williams & Rust 1969; Bridge *et al.* 1986), and fine downstream reflecting progressive blockage at the entrance. The bedload deposits at the upstream end of a channel fill tend to fine upwards as they represent the progradation of bar tail deposits into the channel entrance (e.g. Bridge *et al.* 1986). Bedload deposits in channel fills may also show evidence of accretion on progressively smaller bars as discharge is reduced (Fig. 23). Deltaic deposits may be formed at the entrances of abandoned channels with ponded water (e.g. Gagliano & Howard 1984), and sediment gravity flows from cut banks may accumulate in thalwegs (Bridge *et al.* 1986). If suspended sediment loads are low, peat may accumulate in the ponded water of channel fills in humid climates. In arid climates, evaporitic tufas may form.

Evidence of changing stage during seasonal deposition on bars and channel fills includes fining of grain size and associated changes in sedimentary structure in the upper parts of accretionary units (e.g. Collinson 1970; Rust 1972; Bluck 1974, 1976, 1979; review in Allen 1982). Particularly common are reactivation surfaces in cross-stratified sets with current and wave ripples superimposed, and possibly mud drapes with abundant plant debris. Rill marks orientated parallel to depositional slopes represent falling stage drainage channels, and cross-stratified sand wedges represent the small deltas that form as these channels flow into standing water. Desiccation cracks occur in emergent mud drapes, and rooted plants can colonise areas exposed at low flow stage. Indeed the level of features such as these in channel sequences gives an indication of the low stage level. Burrowing and surface-browsing animals are most active above and below the water table following major erosional and depositional events, although escape burrows may occur within the main flood deposits. A detailed treatment of organic activity in braided rivers is not possible here.

Palaeocurrents

Palaeocurrent orientations observed in channel deposits must depend on: (1) orientation of the bedform (and associated sedimentary structure) relative to channel orientation, which varies with type of bedform, its position in the channel, and with river stage, and: (2) what part of the channel bar or fill is preserved. The orientations of structures like imbrication and various scales of cross-stratification from emergent parts of modern river bars correspond closely with *local* water flow directions (e.g. Harms *et al.* 1963; Steinmetz 1967; Coleman 1969; Bluck 1971, 1974, 1976, 1979; Shelton *et al.* 1974; Teisseyre 1975; Tandon & Kumar 1981) except in the case of planar cross stratification from straight crested dunes and unit bars (e.g. scroll bars) which are highly oblique to local flow directions (e.g. Rust 1972; Bluck 1971, 1974, 1976, 1979; Smith 1972; Jackson 1976*b*; Cant & Walker 1978; Bridge & Jarvis 1982). Also, imbrication near banks may be oblique to *local* channel direction (Teisseyre 1975) and downstream-dipping imbrication on bedform slip faces (pseudoimbrication) must be recognized as such. However, these local palaeoflow directions may be associated with deposition at a range of palaeoflow stages from flows of a range of strengths, and will not necessarily be parallel to the orientation of the high stage channels. As a result of this, the overall mean palaeocurrent azimuth for any particular structure may not be parallel to the mean channel orientation and the range of azimuths will probably greatly exceed the range of local channel orientations (e.g. Williams & Rust 1969; Bluck 1971, 1974, 1976, 1979; Schwartz 1978).

By virtue of the nature of channel migration discussed above, it is expected that flood stage palaeocurrents from the downstream parts of channel bars will be preferentially preserved in ancient deposits, which will reduce the variability evident in the studies of modern emergent channel bars. Thus observations of palaeocurrents from various sedimentary structures throughout a single, well-exposed braided channel-belt deposit will probably indicate the mean and range of channel orientations if analyzed correctly, but cannot generally be used to indicate sinuosity of individual segments (cf. Miall 1976). The sinuosity of individual segments and the mode of channel migration can, however, be reconstructed from very detailed observations from large outcrops and using 3D depositional models of the kind given previously (e.g. Willis 1993).

Discussion

The models described above are for braided channel belts with low braiding parameter; however, it would be conceptually simple to repeat the main features in the across valley direction, but practically difficult. Bridge (1985) speculated that the proportion of channel-fill deposits relative to bar-accretion deposits within a channel belt should increase with braiding parameter. Bristow (1987) challenged this based on the proportion of bar migration relative to channel cutting and filling in the Brahmaputra. In retrospect, if braided rivers can be looked upon as a series of single curved rivers placed side by side, there is no reason to think that my earlier speculation is correct. Indeed, the only definitive depositional evidence for braiding appears to be cross sections through braid bars with coeval channels, and confluences. Braided river facies models that show a series of overlapping channel fills associated with different scales of channels (e.g. Williams & Rust 1969) cannot be justified.

Vertical superposition of channel bar and fill deposits in single channel belts can be accomplished by superposition of a cross-bar channel on a larger bar and migration of one main channel bar over another. In the latter case, the degree of preservation of the overridden bar depends on the relative elevations of the two superimposed basal erosion surfaces. As is the case with dunes, the likelihood of preservation of the lower parts of the eroded bar increases with the vertical deposition rate relative to the lateral migration rate of the superimposed bar, and the variability of bar thicknesses (e.g. Allen 1982; Paola & Borgman 1991; Best & Bridge 1992). This kind of superposition of bars and fills could not be included easily in the models above because it is very difficult to predict how individual channel segments and bars will migrate and become superimposed on others within channel belts. Note that vertical superposition of channel bar and fill deposits can also result from superposition of distinct channel belts.

Other important aspects of braided river deposition which cannot be considered here include: (1) alongvalley variation of braided river geometry and processes (including on alluvial fans) associated with varying discharge, sediment supply and valley slope; (2) floodplain geometry, processes and deposits; (3) periodic avulsions of channel belts to other floodplain locations; (4) temporal changes in braided channels and floodplains associated with local and regional tectonism, climate and sea-level change.

Finally, the models presented here are largely hypothetical for reasons explained previously. In order to validate and extend these models it is necessary to undertake a detailed program of coring and geophysical profiling of modern channel belts, possibly supplemented by scale modelling of braided-river deposition. In the meantime, the next section is an examination of how and whether these models can further our interpretation and understanding of ancient river deposits.

Interpretation of ancient deposits: some examples

Willis (1993) has done a remarkable job of describing some Miocene fluvial deposits from the Siwaliks of northern Pakistan, where the rocks are exposed continuously along strike for many kilometers. He has also performed perhaps the most sophisticated qualitative and quantitative palaeoenvironmental interpretation of channel deposits to date. It is clear that to accomplish this considerable task the following are required: (1) very detailed description of large outcrops; (2) thorough understanding of the geometry, flow and sedimentary processes associated with modern channel bars and fills, and; (3) knowledge of how channel bar and fill deposits resulting from various modes of channel migration appear in variously orientated 2D sections. Regarding (3), Willis (1993) made use of diagrams similar to, but simpler than, those in Figs 21–23. Willis was able to quantitatively reconstruct the width, depth, mean velocity, slope, wavelength and sinuosity of individual channel segments, and remarkably, to estimate channel belt widths and braiding index. Channel bars migrated by downstream translation (mainly) and bend expansion, and by channel switching (cutting and filling of channels within the channel belt). Rates of channel migration could be estimated at up to the order of a channel width per seasonal flood period. Figures 24 to 26 show examples of some of Willis' channel-belt deposits, in order to illustrate the usefulness of the facies models produced here. As indicated in the captions of Figs 24–26, all of the various channel bar and fill deposits in these examples can be explained by direct comparison with Figs 21–23. In most cases, Willis' interpretations are supported, but not always. It is commonly not easy to tell if a major bedding truncation in a braid-bar deposit is due to seasonal discharge changes or to channel switching.

Unfortunately, the majority of other studies of ancient river deposits do not show large enough outcrops with *details* of large scale bedding geometry, grain textures, internal structures and palaeocurrents to allow detailed

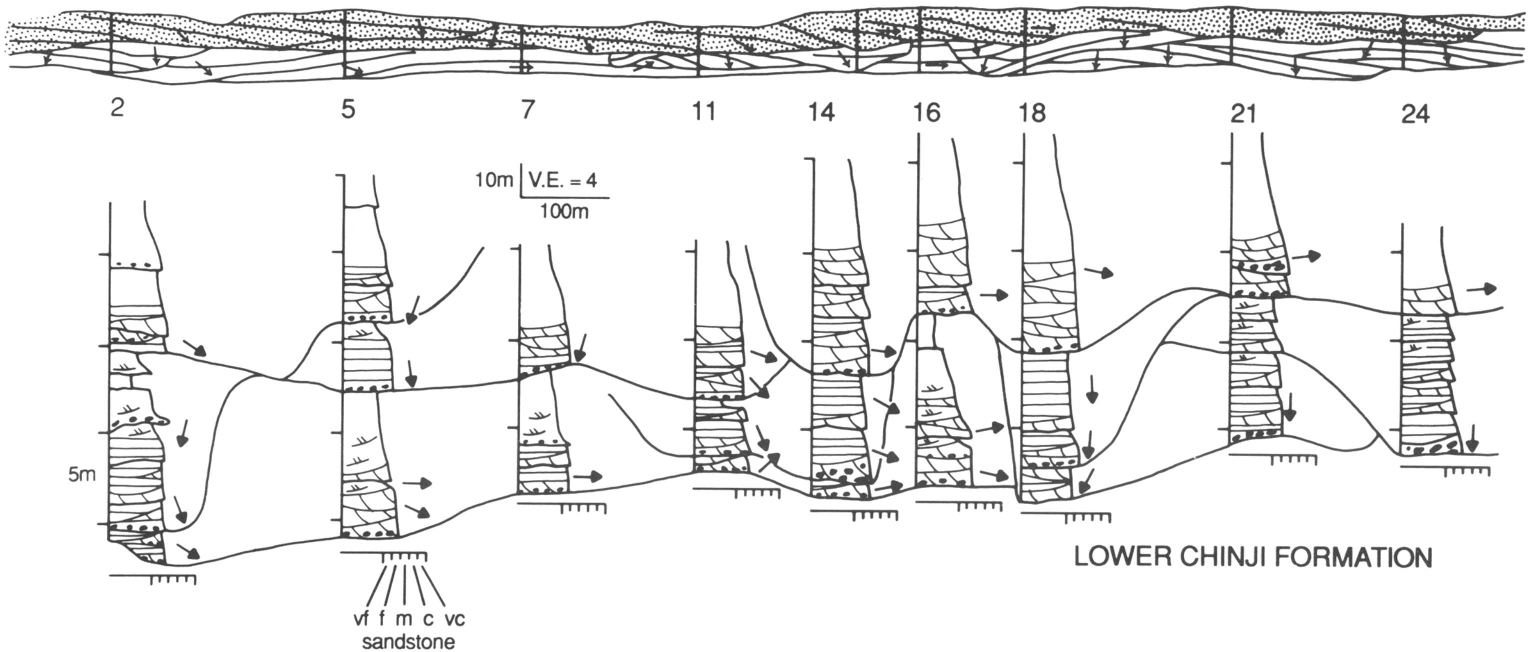

Fig. 24. Redrawn part of Willis' (1993) 2D bedding diagrams and selected vertical logs comprising a sandstone body from the Lower Chinji Formation, Pakistan Siwaliks. The bedding diagram shows major bedset and storey boundaries, and arrows represent paleocurrents relative to the outcrop (a downward pointing arrows indicates paleocurrents normal to, and out of, the outcrop). Symbols for vertical logs are explained in the legend (Fig. 16B), and arrows represent palaeocurrent directions relative to north (which is upwards). The storeys in the upper half of the sandstone body (shaded), and the lowest storeys on the left hand side are comparable to Fig. 23, section 3. The lowest storeys on the right hand side are comparable to parts of Fig. 21, section 3 and Fig. 23, section 2.

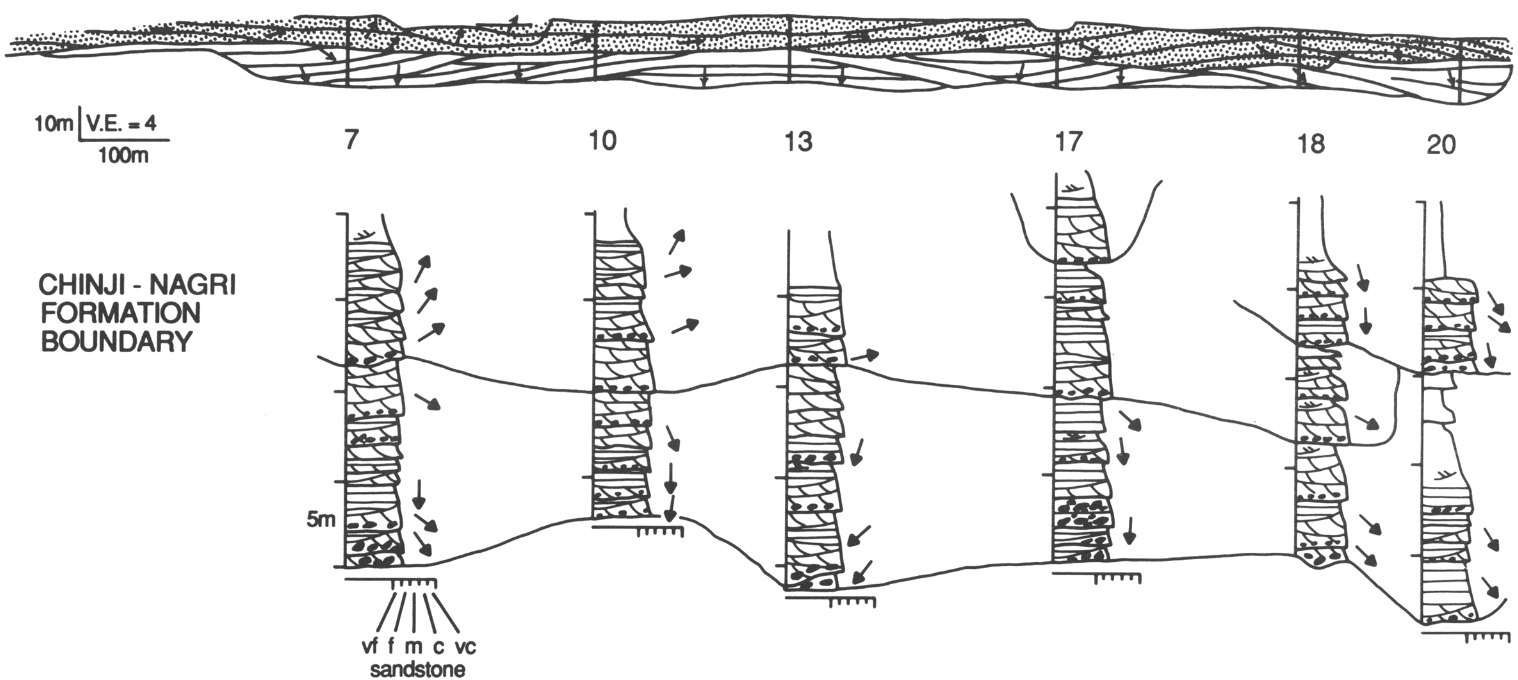

Fig. 25. Redrawn part of Willis' (1993) 2D bedding diagram and selected vertical logs for a sandstone body at the Chinji-Nagri formation boundary. See Fig. 24 caption for explanation. The upper storeys are comparable to Fig. 22, section 5 and Fig. 23, section 4. The lower storeys are comparable to Fig. 21, section 3 and Fig. 23, section 2.

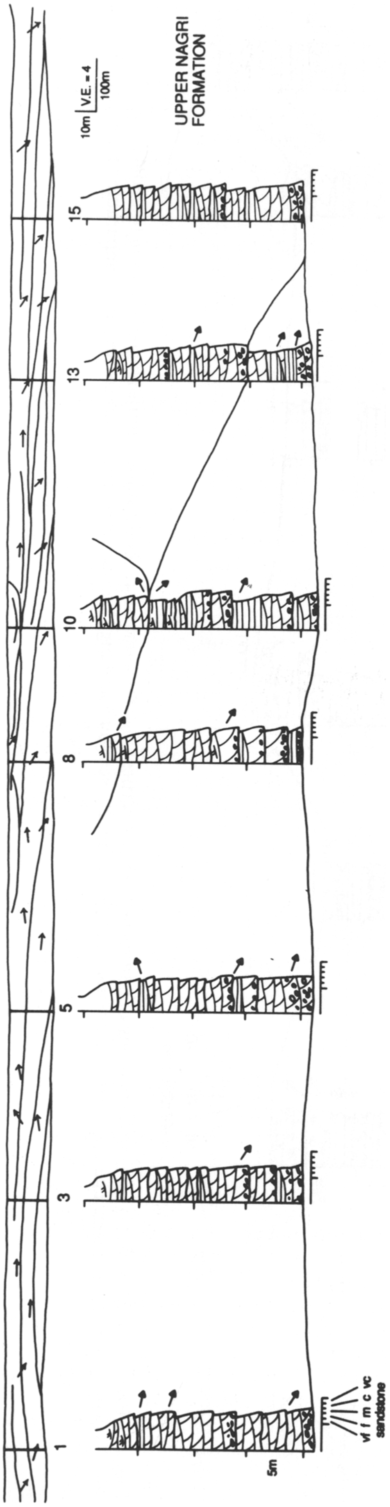

Fig. 26. Redrawn part of Willis' (1993) 2D bedding diagram and selected vertical logs for a sandstone body from the Upper Nagri formation. See Fig. 24 caption for explanation. This outcrop is comparable to Fig. 23, section 3.

interpretations of the kind shown here. Indeed, many interpretations of ancient fluvial deposits as due to braided rivers have relied upon criteria which have been shown here to be mythical.

Concluding remarks

Substantial progress in our understanding of braided river processes will only come from a combination of: (1) field and laboratory studies of the variation and interaction of channel geometry, flow, sediment transport, erosion and deposition, and channel migration over spatial scales of several bars and time scales of decades, including sequential observations over large discharge ranges; (2) quantitative documentation of the 3D variation of all sediment properties of channel-belt deposits (including the age and spatial arrangement of preserved parts of bars and channel fills) and their relation to the causative processes; (3) development of theoretical models which can be tested using extensive data from braided rivers.

Such an ambitious undertaking would require a substantial commitment of resources over a relatively long period. In particular, collaboration of personnel with a wide range of expertise is essential. Innovative measuring equipment and techniques need to be developed. For instance, it is desirable to operate flow and sediment transport measuring equipment from stable positions, with automated deployment and recording. Channel belt deposits should be described using a combination of closely-spaced cores and a geophysical profiling method such as ground-penetrating radar or high-frequency seismic. A long-term program of sequential aerial photography and satellite imagery is required. Clearly, such a large scale, long term undertaking could not possibly be financed under the current mode of operation of government and private-sector funding agencies.

List of symbols

a	coefficient or exponent
b	exponent
c	exponent
C	cohesive strength

C_o	volume concentration of bed sediment
d	mean flow depth
d_{bf}	bankfull mean flow depth
d_c	centerline depth
d_l	local flow depth
d_s	maximum confluence scour depth
d_s^*	dimensionless d_s
d_1,d_2	mean depths of channels entering a confluence
D	grain size
D_{35}	35th percentile grain size
D_{50}	median grain size
D_{84}	84th percentile grain size
D_{90}	90th percentile grain size
f	Darcy-Weisbach friction coefficient
Fr	Froude number
g	gravitational acceleration
h	bed elevation
i_s,i_n	volumetric sediment transport rate in s and n directions
m	mode of alternate bars
M	channel length in one curved channel wavelength
M_1,M_2	M values for channels entering a confluence
n	across stream spatial coordinate
P	effective normal stress
Q	discharge
Q_{bf}	bankfull discharge
Q_m	mean annual discharge
Q_{maf}	mean annual flood discharge
Q_{max}	maximum discharge in flood season
Q_{min}	minimum discharge in floor season
Q_1,Q_2	discharges of channels entering a confluence
Q_{2f}	two-year flood discharge
r_c	centreline radius of curvature
r_{c1},r_{c2}	r_c values of channels entering a confluence
r_l	local radius of curvature
s	streamwise spatial coordinate
sn	sinuosity
sn_1,sn_2	sinuosity of channels entering a confluence
S	channel slope
S_v	valley slope
t	time
T	duration of flood
u_s,u_n	mean grain velocity in s and n directions
u_*	shear velocity
u_{*c}	critical u_* for sediment entrainment
V	cross-section average flow velocity
w	channel width
w_{bf}	bankfull channel width
w_{rf}	rare flood channel width
w_1,w_2	width of channels entering a confluence
α	confluence angle
ΔQ	rise in discharge during flood
θ	dimensionless bed shear stress
θ_c	value of θ at threshold of sediment motion
μ_c	static grain resistance coefficient
ϱ	fluid density
ΣP	total sinuosity
τ_0	bed shear stress
τ_c	value of τ_0 at threshold of sediment motion
τ_s	shear strength

References

ABDULLAYEV, E. A. 1973. Experimental investigation of midstream bars on an erosible model. *Transactions State Hydrological Institute Trudy GGI*, **209**, 103–116.

ACKERS, P. & CHARLTON, F. G. 1971. The geometry of small meandering streams. *Proceedings of the Institution of Civil Engineers, Supplement* **XII**, 289–317.

ALAM, M. M., CROOK, K. A. W. & TAYLOR, G. 1985. Fluvial herring-bone cross-stratification in a modern tributary mouth bar, Coonamble, New South Wales, Australia. *Sedimentology*, **32**, 235–244.

ALLEN, J. R. L. 1965. A review of the origin and characteristics of recent alluvial sediments. *Sedimentology*, **5**, 89–191.

—— 1968. *Current ripples*. North Holland, Amsterdam.

—— 1976. Computational models of dune time-lag: an alternative boundary condition. *Sedimentary Geology*, **16**, 255–279.

—— 1982. *Sedimentary structures: their character and physical basis, vol 1*, Elsevier, Amsterdam.

—— 1983. Studies in fluviatile sedimentation: bars, bar-complexes and sandstone sheets (low-sinuosity braided streams) in the Brownstones (L. Devonian), Welsh Borders. *Sedimentary Geology*, **33**, 237–293.

AMERICAN SOCIETY OF CIVIL ENGINEERS TASK FORCE ON BED FORMS IN ALLUVIAL CHANNELS, 1966. Nomenclature for bedforms in alluvial channels. *American Society of Civil Engineers Journal of the Hydraulics Division*, **92**, 51–64.

ANTROPOVSKIY, V. I. 1972. Quantitative criteria of channel macroforms. *Soviet Hydrology*, 477–484.

ASHLEY, G. M. 1990. Classification of large-scale subaqueous bed-forms: a new look at an old problem. *Journal of Sedimentary Petrology*, **60**, 160–172.

ASHMORE, P. E. 1982. Laboratory modelling of gravel braided stream morphology. *Earth Surface Processes*, **7**, 201–225.

—— 1988. Bed load transport in braided gravel-bed stream models. *Earth Surface Processes and Landforms*, **13**, 677–695.

—— 1991*a*. Channel morpology and bedload pulses in braided, gravel-bed streams. *Geografiska Annaler*, **73A**, 37–52.

—— 1991*b*. How do gravel-bed rivers braid? *Canadian Journal of Earth Sciences*, **28**, 326–341.

—— 1993. Anabranch confluence kinetics and sedimentation processes in gravel braided streams. *This volume*.

—— & PARKER, G. 1983. Confluence scour in coarse braided streams. *Water Resources Research*, **19**, 392–402.

——, FERGUSON, R. I., PRESTEGAARD, K. L., ASHWORTH, P. J. & PAOLA, C. 1992. Secondary flow in anabranch confluences of a braided, gravel-bed stream. *Earth Surface Processes and Landforms*, **17**, 299–311.

ASHWORTH, P. J. & FERGUSON, R. I. 1986. Interrelationships of channel processes, changes and sediments in a proglacial braided river. *Geografiska Annaler*, **68A**, 361–371.

—— & —— 1989. Size-selective entrainment of bed load in gravel bed streams. *Water Resources Research*, **25**, 627–634.

——, —— ASHMORE, P. E., PAOLA, C., POWELL, D. M. & PRESTEGAARD, K. L. 1992a. Measurements in a braided river chute and lobe: 2. Sorting of bedload during entrainment, transport and deposition. *Water Resources Research*, **28**, 1887–1896.

——, —— & POWELL, M. D. 1992*b*. Bedload transport and sorting in braided channels. *In*: BILLI, P., HEY, C. R. THORNE, C. R. & TACCONI, P. (eds) *Dynamics of Gravel-bed Rivers*. Wiley, 497–515.

BEGIN, Z. B. 1981. The relationship between flow shear stress and stream pattern. *Journal of Hydrology*, **52**, 307–319.

—— & SCHUMM, S. A. 1984. Gradational thresholds and landform singularity: significance for Quaternary studies. *Quaternary Research*, **21**, 267–274.

BENNETT, S. J. 1992. *A theoretical and experimental study of bedload transport of heterogeneous sediment*. PhD thesis, State University of New York at Binghamton.

BERGERON, N. & ROY, A. G. 1988. Les effets d'un embacle sur la morphologie du lit d'une confluence de cours d'eau. *Geographie physique et Quaternaire*, **42**, 191–196.

BESCHTA, R. L. & JACKSON, W. L. 1979. The intrusion of fine sediments into a stable gravel bed. *Journal of the Fisheries Research Board of Canada*, **36**, 204–210.

BEST, J. L. 1986. The morphology of river channel confluences. *Progress in Physical Geography*, **10**, 157–174.

—— 1987. Flow dynamics at river channel confluences: implications for sediment transport and bed morphology. *In*: ETHRIDGE, F. G., FLORES, R. M. & HARVEY, M. D. (eds) *Recent Developments in Fluvial Sedimentology*. SEPM Special Publication, **39**, 27–35.

—— 1988. Sediment transport and bed morphology at river channel confluences. *Sedimentology*, **35**, 481–498.

—— & BRAYSHAW, A. C. 1985. Flow separation — a physical process for the concentration of heavy minerals within alluvial channels. *Journal of the Geological Society, London*, **142**, 767–755.

—— & BRIDGE, J. S. 1992. The morphology and dynamics of low-amplitude bedwaves upon upper-stage plane beds and the preservation of planar laminae. *Sedimentology*, **39**, 737–752.

—— & REID, I. 1984. Separation zone at open-channel junctions. *Journal of Hydraulic Engineering, ASCE*, **110**, 1588–1594.

—— & —— 1987. Separation zone at open-channel junctions: closure. *Journal of Hydraulic Engineering, ASCE*, **113**, 545–548.

—— & ROY, A. G. 1991. Mixing-layer distortion at the confluence of channels of different depth. *Nature*, **350**, 411–413.

BETTESS, R. & WHITE, W. R. 1983. Meandering and braiding of alluvial channels. *Proceedings of the Institution of Civil Engineers*, **75**, 525–538.

BLODGETT, R. H. & STANLEY, K. O. 1980. Stratification, bedforms and discharge relations of the Platte braided river system, Nebraska. *Journal of Sedimentary Petrology*, **50**, 139–148.

BLONDEAUX, P. & SEMINARA, G. 1985. A unified barbend theory of river meanders. *Journal of Fluid Mechanics*, **157**, 449–470.

BLUCK, B. J. 1971. Sedimentation in the meandering River Endrick. *Scottish Journal of Geology*, **7**, 93–138.

—— 1974. Structure and directional properties of some valley sandur deposits in southern Iceland. *Sedimentology*, **21**, 533–554.

—— 1976. Sedimentation in some Scottish rivers of low sinuosity. *Transactions Royal Society of Edinburgh*, **69**, 425–456.

—— 1979. Structure of coarse grained braided alluvium. *Transactions Royal Society of Edinburgh*, **70**, 181–221.

—— 1982. Texture of gravel bars in braided streams. *In*: HEY, R. C., BATHURST, J. C. & THORNE, C. R. (eds) *Gravel-bed rivers*. Wiley, 339–355.

BOOTHROYD, J. D. & ASHLEY, G. M. 1975. Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. *In*: JOPLING, A. V. & MCDONALD, B. C. (eds) *Glaciofluvial and Glaciolacustrine Sedimentation*. SEPM Special Publication, **23**, 193–222.

—— & NUMMEDAL, D. 1978. Proglacial braided outwash: a model for humid alluvial fan deposits. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 641–668.

BRADY, L. L. & JOBSON, H. E. 1973. *An experimental study of heavy mineral segregation under alluvial flow condition*. US Geological Survey Professional Paper 562.

BRAY, D. I. 1982. Regime equations for gravel-bed rivers. *In*: HEY, R. D., BATHURST, J. D. & THORNE, C. R. (eds) *Gravel-bed rivers*. Wiley, 517–552.

BRAYSHAW, A. D. 1984. Characteristics and origin of cluster bedforms in coarse-grained alluvial channels. *In*: KOSTER, E. H. & STEEL, R. J. (eds) *Sedimentology of Gravels and Conglomerates*. Canadian Society of Petroleum Geologists Memoirs, **10**, 77–85.

—— 1985. Bed microtopography and entrainment thresholds in gravel-bed rivers. *Geological Society of America Bulletin*, **96**, 218–223.

——, FROSTICK, L. E. & REID, I. 1983. The hydrodynamics of particle clusters and sediment entrainment in coarse alluvial channels. *Sedimentology*, **30**, 137–143.

BRICE, J. C. 1960. Index for description of channel braiding. *Geological Society of America Bulletin*, **71**, 1833.

—— 1964. *Channel patterns and terraces of the Loup rivers in Nebraska*. US Geological Survey Professional Paper 422D.

—— 1984. Planform properties of meandering rivers. *In*: ELLIOTT, C. M. (ed.) *River Meandering*. American Society of Civil Engineers, 1–15.

——, BLODGETT, J. D. *ET AL*. 1978. *Countermeasures for hydraulic problems at bridges*. Federal Highway Administration Report FHWA-RD-78-162, Washington, D.C.

BRIDGE, J. S. 1984. Flow and sedimentary processes in river bends: comparison of field observations and theory. *In*: ELLIOTT, C. M. (ed.) *River meandering*. American Society of Civil Engineers, 857–872.

—— 1985. Paleochannels inferred from alluvial deposits: a critical evaluation. *Journal of Sedimentary Petrology*, **55**, 579–589.

—— 1992. A revised model for water flow, sediment transport, bed topography, and grainsize sorting in natural river bends. *Water Resources Research*, **28**, 999–1013.

—— & BENNETT, S. J. 1992. A model for the entrainment and transport of sediment grains of mixed sizes, shapes and densities. *Water Resources Research*, **28**, 337–363.

—— & GABEL, S. L. 1992. Flow and sediment dynamics in a low-sinuosity river: Calamus River, Nebraska Sandhills. *Sedimentology*, **39**, 125–142.

—— & JARVIS, J. 1976. Flow and sedimentary processes in the meandering River South Esk, Glen Clova, Scotland. *Earth Surface Processes*, **1**, 303–336.

—— & —— 1982. The dynamics of a river bend: a study in flow and sedimentary processes. *Sedimentology*, **29**, 499–550.

——, SMITH, N. D., TRENT, F., GABEL, S. L. & BERNSTEIN, P. 1986. Sedimentology and morphology of a low-sinuosity river: Calamus River, Nebraska Sandhills. *Sedimentology*, **33**, 851–870.

BRISTOW, C. S. 1987. Brahmaputra River: channel migration and deposition. *In*: ETHRIDGE, F. G., FLORES, R. M. & HARVEY, M. D. (eds) *Recent developments in Fluvial Sedimentology*. SEPM Special Publication, **39**, 63–74.

—— 1993. Sedimentary structures exposed in bar tops in the Brahmaputra River, Bangladesh. *This volume*.

——, BEST, J. L. & ROY, A. G. 1993. Morphology and facies models of channel confluences. *In*: MARZO, M. & PUIDEFABREGES, C. (eds) *Alluvial Sedimentation*. International Association of Sedimentologists Special Publication, **17**, 91–100.

BUCK, S. G. 1983. The Saaiplaas Quartzite member: a braided system of gold- and uranium-bearing channel placers within the Proterozoic Witwatersrand Supergroup of South Africa. *In*: COLLINSON, J. D. & LEWIN, J. (eds) *Modern and Ancient Fluvial Systems*, International Association of Sedimentologists Special Publications, **6**, 549–562.

CANT, D. J. 1978. Bedforms and bar types in the South Saskatchewan River. *Journal of Sedimentary Petrology*, **48**, 1321–1330.

—— & WALKER, R. G. 1978. Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. *Sedimentology*, **25**, 625–648.

CAREY, W. C. 1969. Formation of floodplain lands. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **95**, 981–994.

CARSON, M. A. 1984*a*. The meandering-braided threshold: a reappraisal. *Journal of Hydrology*, **73**, 315–334.

—— 1984*b*. Observations on the meandering-braided river transition, the Canterbury Plains, New Zealand: part one. *New Zealand Geographer*, **40**, 12–17.

—— 1984*c*. Observations on the meandering-braided river transition, the Canterbury Plains, New Zealand: part two. *New Zealand Geographer*, **40**, 89–99.

—— 1986. Characteristics of high-energy 'meandering' rivers: the Canterbury Plains, New Zealand. *Geological Society of America Bulletin*, **97**, 886–895.

CHALOV, R. S. 1974. Effect of side bars in riffles on river bank dynamics. *Soviet Hydrology: Selected Papers*, **1974 (4)**, 270–273.

CHANG, H. H. 1979. Minimum stream power and river channel patterns. *Journal of Hydrology*, **41**, 303–327.

—— 1985. River morphology and thresholds. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **111**, 503–519.

CHEEL, R. J. 1985. Heavy mineral shadows, a new sedimentary structure formed under upper-flow regime conditions: its directional and hydraulic significance. *Journal of Sedimentary Petrology*, **54**, 1175–1182.

CHEETHAM, G. H. 1979. Flow competence in relation to stream channel form and braiding. *Geological Society of America Bulletin*, **90**, 877–886.

CHIEN NING. 1961. The braided river of the Lower Yellow River. *Scientia Sinica*, **10**, 734–754.

CHIEW, Y. M. 1991. Bed features in nonuniform sediments. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **117**, 116–120.

CHITALE, S. V. 1970. River channel patterns. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **96**, 201–221.

—— 1973. Theories and relationships of river channel patterns. *Journal of Hydrology*, **19**, 285–308.

CHURCH, M. 1972. *Baffin Island sandurs: a study of arctic fluvial processes*. Geological Survey of Canada Bulletin 216.

—— & GILBERT, R. 1975. Proglacial fluvial and lacustrine sedimentation. *In*: JOPLING, A. V. & MCDONALD, B. C. (eds) *Glaciofluvial and glaciolacustrine sedimentation*, SEPM Special Publications, **23**, 22–100.

—— & JONES, D. 1982. Channel bars in gravel-bed rivers. *In*: HEY, R. D., BATHURST, J. C. & THORNE, C. R. (eds) *Gravel-bed rivers*, Wiley, Chichester, 291–324.

COLEMAN, J. M. 1969. Brahmaputra River: channel processes and sedimentation. *Sedimentary Geology*, **3**, 129–239.

COLLINSON, J. D. 1970. Bedforms of the Tana River, Norway. *Geografiska Annaler*, **52A**, 31–56.

COSTELLO, W. R. & SOUTHARD, J. B. 1981. Flume experiments on lower-flow-regime bedforms in coarse sand. *Journal of Sedimentary Petrology*, **51**, 849–864.

CROWLEY, K. D. 1983. Large-scale bed configurations (macroforms), Platte River Basin, Colorado and Nebraska: primary structures and formative processes. *Geological Society of America Bulletin*, **94**, 117–133.

DAL CIN, R. 1968. Pebble clusters: their interpretation and utilization in the study of paleocurrents. *Sedimentary Geology*, **2**, 233–241.

DAVOREN, A. & MOSLEY, M. P. 1986. Observations of bedload movement, bar development and sediment supply in the braided Ohau River. *Earth Surface Processes and Landforms*, **11**, 643–652.

DIETRICH, W. E. & SMITH, J. D. 1983. Influence of the point bar on flow in curved channels. *Water Resources Research*, **19**, 1173–1192.

—— & —— 1984. Bedload transport in a river meander. *Water Resources Research*, **20**, 1355–1380.

—— & WHITING, P. J. 1989. Boundary shear stress and sediment transport in river meanders of sand and gravel. *In*: IKEDA, S. & PARKER, G. (eds) *River Meandering*. American Geophysical Union Water Resources Monographs, **12**, 1–50.

——, KIRCHNER, J. W., IKEDA, H. & ISEYA, F. 1989. Sediment supply and the development of the coarse surface layer in gravel-bedded rivers. *Nature*, **340**, 215–217.

——, SMITH, J. D. & DUNNE, T. 1979. Flow and sediment transport in a sand-bedded meander. *Journal of Geology*, **87**, 305–315.

ELLIOTT, C. M. (ed.) 1984. *River Meandering*. American Society of Civil Engineers.

ENGELUND, F. & SKOVGAARD, O. 1973. On the origin of meandering and braiding in alluvial streams. *Journal of Fluid Mechanics*, **57**, 289–302.

ERGENZINGER, P. 1987. Chaos and order: the channel geometry of gravel bed braided rivers. *Catena Supplement*, **10**, 85–98.

ESCHNER, T. R. 1983. *Hydraulic geometry of the Platte River near Overton, south-central Nebraska*. US Geological Survey Professional Paper 1277C.

FAHNESTOCK, R. K. 1963. *Morphology and hydrology of a glacial stream — White River, Mt. Rainier, Washington*. US Geological Survey Professional Paper 422A.

FERGUSON, R. I. 1981. Channel form and channel changes. *In*: LEWIN, J. (ed.) *British Rivers*. Allen & Unwin, 90–125.

—— 1984. The threshold between meandering and braiding. *In*: SMITH, K. V. H. (ed.) *Proceedings of the 1st International Conference on Hydraulic Design*. Springer, 6.15–6.29.

—— 1986. Hydraulics and hydraulic geometry. *Progress in Physical Geography*, **10**, 1–31.

—— 1987. Hydraulic and sedimentary controls of channel pattern. *In*: RICHARDS, K. S. (ed.) *River Channels: Environment and Process*. Blackwell, 125–158.

—— & ASHWORTH, P. 1991. Slope-induced changes in channel character along a gravel bed stream: the Allt Dubhaig, Scotland. *Earth Surface Processes and Landforms*, **16**, 65–82.

—— & WERRITTY, A. 1983. Bar development and channel changes in the gravelly River Feshie, Scotland. *In*: COLLINSON, J. D. & LEWIN, J. *Modern and Ancient Fluvial Systems*. International Associated of Sedimentologists Special Publications, **6**, 181–193.

——, ASHMORE, P. E., ASHWORTH, P. J., PAOLA, C. & PRESTEGAARD, K. L. 1992. Measurements in a braided river chute and lobe: I. Flow pattern, sediment transport and channel change. *Water Resources Research*, **28**, 1877–1886.

FISK, H. N. 1947. *Fine grained alluvial deposits and their effects on Mississippi River activity*. Mississippi River Commission, Vicksburg, Mississippi.

FREDSOE, J. 1978. Meandering and braiding of rivers. *Journal of Fluid Mechanics*, **84**, 609–624.

—— 1979. Unsteady flow in straight alluvial streams: modification of individual dunes. *Journal of Fluid Mechanics*, **91**, 497–512.

—— 1982. Shape and dimensions of stationary dunes in rivers. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **108**, 932–947.

FRIEDKIN, J. F. 1945. *A laboratory study of the meandering of alluvial rivers*. US Waterways Experimental Station, Vicksburg, Mississippi.

FRIEND, P. F. & SINHA, R. 1993. Braiding and meandering parameters. *This volume*.

FROSTICK, L. E., LUCAS, P. M. & REID, I. 1984. The infiltration of fine matrices into coarse-grained alluvial sediments and its implications for stratigraphical interpretation. *Journal of the Geological Society, London*, **141**, 955–965.

FUJITA, Y. 1989. Bar and channel formation in braided streams. *In*: IKEDA, S. & PARKER, G. (eds) *River Meandering*. American Geophysical Union, Water Resources Monographs, **12**, 417–462.

FUKUOKA, S. 1989. Finite amplitude development of alternate bars. *In*: IKEDA, S. & PARKER, G. (eds) *River Meandering*. American Geophysical Union, Water Resources Monographs, **12**, 237–265.

GABEL, S. L. 1993. Geometry and kinematics of dunes during steady and unsteady flows in the Calamus River, Nebraska. *Sedimentology*, in press.

GAGLIANO, S. M. & HOWARD, P. C. 1984. The neck cutoff oxbow lake cycle along the lower Mississippi River. *In*: ELLIOTT, C. M. (ed.) *River Meandering*. American Society of Civil Engineers, 147–158.

GIBLING, M. R. & RUST, B. R. 1990. Ribbon sandstones in the Pennsylvanian Waddens Cove Formation, Sydney Basin, Atlantic Canada: the influence of siliceous duricrusts on channel-body geometry. *Sedimentology*, **37**, 45–65.

GRIFFITHS, G. A. 1979. Recent sedimentation history of the Waimakariri River, New Zealand. *Journal of Hydrology (New Zealand)*, **18**, 6–28.

GUSTAVSON, T. C. 1978. Bedforms and stratification types of modern gravel meander lobes, Nueces River, Texas. *Sedimentology*, **25**, 401–426.

HAGER, W. H. 1984. An approximate treatment of flow in branches and bends. *Proceedings of the Institution of Mechanical Engineers*, **198C**, 63–69.

—— 1987. Separation zone at open-channel junctions: discussion. *Journal of Hydraulic Engineering, ASCE*, **113**, 539–543.

HARMS, J. C. & FAHNESTOCK, R. R. 1965. Stratification, bedforms and flow phenomena (with an example from the Rio Grande). *In*: MIDDLETON, G. V. (ed.) *Primary sedimentary structures and their hydrodynamic interpretation*. SEPM Special Publications, **12**, 84–115.

——, MCKENZIE, D. B. & MCCUBBIN, D. G. 1963. Stratification in modern sands of the Red River, Louisiana. *Journal of Geology*, **71**, 566–580.

HASEGAWA, K. 1989. Studies on qualitative and quantitative prediction of meander channel shift. *In*: IKEDA, S. & PARKER, G. (eds) *River Meandering*. American Geophysical Union Water Resources Monographs, **12**, 215–235.

HAYASHI, T. & OZAKI, S. 1980. Alluvial bedforms analysis. I. Formation of alternating bars and braids. *In*: SHEN, H. W. & KIKKAWA, H. (eds) *Application of Stochastic Processes in Sediment Transport*. Water Resources Publications, Colorado, Ch 7, 1–40.

HEIN, F. J. & WALKER, R. G. 1977. Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River, British Columbia. *Canadian Journal of Earth Science*, **14**, 562–570.

HENDERSON, F. M. 1961. Stability of alluvial channels. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **87**, 109–138.

—— 1966. *Open channel flow*. Macmillan, New York.

HICKIN, E. J. 1969. A newly identified process of point bar formation in natural streams. *American Journal of Science*, **267**, 999–1010.

—— 1979. Concave bank benches in the Squamish River, British Columbia. *Canadian Journal of Earth Sciences*, **16**, 200–203.

HOEY, T. 1992. Temporal variations in bedload transport rates and sediment storage in gravel-bed rivers. *Progress in Physical Geography*, **16**, 319–338.

—— & SUTHERLAND, A. J. 1991. Channel morphology and bedload pulses in braided rivers: a laboratory study. *Earth Surface Processes and Landforms*, **16**, 447–462.

HONG, L. B. & DAVIES, T. R. H. 1979. A study of stream braiding. *Geological Society of America Bulletin*, **79**, 391–394.

HOWARD, A. D., KEETCH, M. E. & VINCENT, C. L. 1970. Topological and geometrical properties of braided rivers. *Water Resources Research*, **6**, 1674–1688.

—— 1992. Modeling channel migration and floodplain sedimentation in meandering streams. *In*: CARLING, P. A. & PETTS, G. E. (eds) *Lowland Floodplain Rivers: Geomorphological Perspectives*. Wiley, 1–41.

IKEDA, H. 1973. A study of the formation of sand bars in an experimental flume. *Geographical Review of Japan*, **46**, 435–452.

—— 1975. On the bed configuration in alluvial channels; their types and condition of formation with reference to bars. *Geographical Review of Japan*, **48**, 712–730.

—— 1983. *Experiments on bedload transport, bedforms, and sedimentary structures using fine gravel in the 4-meter wide flume*. University of Tsukuba, Japan, Environmental Research Center Paper No. 2.

IKEDA, S. 1984. Prediction of alternate bar wavelength and height. *Journal of Hydraulic Engineering, American Society of Civil Engineers*, **110**, 371–386.

—— & PARKER, G. (eds) 1989. *River meandering*. American Geophysical Union, Water Resources Monographs, **12**.

——, —— & SAWAI, K. 1981. Bend theory of river meanders. 1. Linear development. *Journal of Fluid Mechanics*, **112**, 363–377.

ISEYA, S. & IKEDA, H. 1987. Pulsations in bedload transport rates induced by a longitudinal sediment sorting: a flume study of sand and gravel mixtures. *Geografiska Annaler*, **69A**, 15–27.

JACKSON, R. G. 1975*a*. Velocity-bedform-texture patterns of meander bends in the Lower Wabash River of Illinois and Indiana. *Geological Society of America Bulletin*, **86**, 1511–1522.

—— 1975*b*. Hierarchical attributes and a unifying model of bedforms composed of cohesionless sediment and produced by shearing flow. *Geological Society of America Bulletin*, **86**, 1523–1533.

—— 1976*a*. Depositional model of point bars in the Lower Wabash River. *Journal of Sedimentary Petrology*, **46**, 579–594.

—— 1976*b*. Largescale ripples of the Lower Wabash River. *Sedimentology*, **23**, 593–623.

—— 1978. Preliminary evaluation of lithofacies models for meandering alluvial streams. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 543–576.

JAEGGI, M. N. R. 1984. Formation and effects of alternate bars. *Journal of Hydraulic Engineering, American Society of Civil Engineers*, **110**, 142–156.

JOHANNESSON, H. & PARKER, G. 1989. Linear theory of river meanders. *In*: IKEDA, S. & PARKER, G. (eds) *River Meandering*. American Geophysical Union Water Resources Monographs, **12**, 181–213.

KARCZ, I. 1971. Development of a meandering thalweg in a straight, erodible laboratory channel. *Journal of Geology*, **79**, 234–240.

—— 1972. Sedimentary structures formed by flash floods in southern Israel. *Sedimentary Geology*, **7**, 161–182.

KASTHURI, B. & PUNDARIKANTHAN, N. V. 1987. Separation zone at open-channel junctions: discussion. *Journal of Hydraulic Engineering, ASCE*, **113**, 543–544.

KELLERHALS, R., CHURCH, M. & BRAY, D. I. 1976. Classification and analysis of river processes. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **102**, 813–829.

KINOSITA, R. 1957. Formation of dunes on river beds — an observation on the condition of river meandering. *Proceedings of the Japan Society of Civil Engineers*, **42**, 1–21.

KINOSITA, R. 1961. *Study of the channel evolution of the Isikari River*. Bureau of Resources, Department of Science and Technology, Japan.

KIRKBY, M. J. 1972. Alluvial and nonalluvial meanders. *Area*, **4**, 284–288.

—— 1977. Maximum sediment efficiency as a criterion for alluvial channels. *In*: GREGORY, K. J. (ed.) *River channel changes*. Wiley, 429–442.

—— 1980. The streamhead as a significant geomorphic threshold. *In*: COATES, D. R. & VITEK, J. (eds) *Thresholds in Geomorphology*. Allen & Unwin, London, 53–73.

KJERFVE, B., SHAO, C. C. & STAPOR, F. W. 1979. Formation of deep scour holes at the junctions of tidal creeks: an hypothesis. *Marine Geology*, **33**, M9–M14.

KNIGHTON, A. D. 1972. Changes in a braided reach. *Geological Society of America Bulletin*, **83**, 3813–3822.

—— 1984. *Fluvial forms and processes*. Edward Arnold, London.

KOMAR, P. D. 1983. Shapes of streamlined islands on Earth and Mars: experiments and analyses of the minimum-drag form. *Geology*, **11**, 651–654.

—— 1984. The Lemniscate Loop-comparisons with shapes of streamlined landforms. *Journal of Geology*, **92**, 133–145.

KOSTER, E. H. 1978. Transverse ribs: their characteristics, origin and paleohydraulic significance. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 161–186.

KRIGSTRÖM, A. 1962. Geomorphological studies of sandur plains and their braided rivers in Iceland. *Geografiska Annaler*, **44**, 328–346.

KUHNLE, R. A. & SOUTHARD, J. B. 1988. Bedload transport fluctuations in a gravel bed laboratory channel. *Water Resources Research*, **24**, 247–260.

—— & —— 1990. Flume experiments on the transport of heavy minerals in gravel-bed streams. *Journal of Sedimentary Petrology*, **60**, 687–696.

LANE, E. W. 1957. *A study of the shape of channels formed by natural streams flowing in erodible material*. Missouri River Division Sediment Series No. 9, U.S. Army Engineer Division, Missouri River, Corps of Engineers, Omaha, Nebraska.

—— & BORLAND, W. M. 1954. River bed scour during floods. *Transactions, American Society of Civil Engineers*, **119**, 1069–1080.

LANGBEIN, W. B. & LEOPOLD, L. B. 1966. *River meanders-theory of minimum variance*. US Geological Survey Professional Paper 422H.

LEEDER, M. R. 1973. Fluviatile fining-upwards cycles and the magnitude of paleochannels. *Geological Magazine*, **110**, 265–276.

LEOPOLD, L. B. & WOLMAN, M. G. 1957. River channel patterns: braided, meandering and straight. *US Geological Survey Professional Papers*, **262B**, 39–85.

LEVEY, R. A. 1978. Bed-form distribution and internal stratification of coarse-grained point bars, Upper Congaree River, South Carolina. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*, Canadian Society of Petroleum Geologists Memoirs, **5**, 105–127.

LEWIN, J. 1976. Initiation of bed forms and meanders in coarse-grained sediment. *Geological Society of America Bulletin*, **87**, 281–285.

—— 1978. Meander development and floodplain sedimentation: a case study from mid-Wales. *Geological Journal*, **13**, 25–36.

—— 1982. British floodplains. *In*: ADAMS, B. H., FENN, C. R. & MORRIS, L. (eds) *Papers in Earth Sciences*. Geo Books, Norwich, 21–37.

—— 1983. Changes in channel pattern and floodplain. *In*: GREGORY, K. J. (ed.) *Background to Palaeohydrology*. Wiley, Chichester, 303–319.

LODINA, R. V. & CHALOV, R. S. 1971. Effect of tributaries on the composition of river sediments and of deformations of the main river channel. *Soviet Hydrology; selected papers*, **4**, 370–374.

LYELL, C. 1830. *The Principles of Geology, volume 1*. John Murray, London.

MCGOWEN, J. H. & GARNER, L. E. 1970. Physiographic features and stratification types of coarse-grained point bars: modern and ancient examples. *Sedimentology*, **14**, 77–111.

MCQUIVEY, R. S. & KEEFER, T. N. 1969. The relation of turbulence to deposition of magnetite over ripples. *US Geological Survey Professional Paper*, **650D**, 244–247.

MIALL, A. D. 1976. Paleocurrent and paleohydrologic analysis of some vertical profiles through a Cretaceous braided stream deposit, Banks Island, Arctic Canada. *Sedimentology*, **23**, 459–483.

—— 1981. *Analysis of fluvial depositional systems*. American Association of Petroleum Geologists Education Course Note Series No. 20.

MIDDLETON, G. V. & SOUTHARD, J. B. 1984. *Mechanics of sediment movement.* SEPM Short Course No. 3.

MINTER, W. E. L. 1978. A sedimentological synthesis of placer gold, uranium and pyrite concentrations in Proterozoic Witwatersrand sediments. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 801–829.

MINTER, W. E. L. & TOENS, P. D. 1970. Experimental simulation of gold deposition in gravel beds. *Transactions of the Geological Society of South Africa*, **73**, 89–98.

MOSLEY, M. P. 1976. An experimental study of channel confluences. *Journal of Geology*, **84**, 535–562.

—— 1981. Semi-determinate hydraulic geometry of river channels, South Island, New Zealand. *Earth Surface Processes and Landforms*, **6**, 127–137.

—— 1982*a*. Scour depths in branch channel confluences, Ohao River, Otago, New Zealand. *Transactions, The Institution of Professional Engineers, New Zealand*, **9**, 17–24.

—— 1982*b*. Analysis of the effect of changing discharge on channel morphology and instream uses in a braided river, Ohau River, New Zealand. *Water Resources Research*, **18**, 800–812.

—— 1983. Response of braided rivers to changing discharge. *Journal of Hydrology (New Zealand)*, **22**, 18–67.

—— & SCHUMM, S. A. 1977. Stream junctions — a probable location for bedrock placers. *Economic Geology*, **72**, 691–694.

MURAMOTO, Y. & FUJITA, Y. 1977. Study on meso scale river bed configuration. *Annuals, Disaster Prevention Research Institute, Kyota Univ.*, **20B-2**, 243–258.

NADEN, P. S. & BRAYSHAW, A. C. 1987. Small and medium-scale bedforms in gravel-bed rivers. *In*: RICHARDS, K. S. (ed.) *River channels: environment and processes*. Blackwell, 249–271.

NANSON, G. C. 1980. Point bar and floodplain formation of the meandering Beatton River, northeastern British Columbia, Canada. *Sedimentology*, **27**, 3–29.

—— & PAGE, K. P. 1983. Lateral accretion of fine-grained concave bank benches in meandering rivers. *In*: COLLINSON, J. D. & LEWIN, J. (eds) Modern and Ancient Fluvial Systems, International Association of Sedimentologists Special Publication 6, 133–143.

——, RUST, B. R. & TAYLOR, G. 1986. Coexistent mud braids and anastomosing channels in an arid-zone river: Cooper Creek, central Australia. *Geology*, **14**, 175–178.

NILSSON, G. & MARTVALL, S. 1972. *The Ore River and its meanders.* Uppsala Universiteit Geologiska Institute Report 19, Uppsala, Sweden.

NORDSETH, K. 1973. Fluvial processes and adjustments on a braided river. The islands of Koppangsoyene on the River Glomma. Norsk Geografiska Tidsskrft, **27**, 77–108.

ODGAARD, A. J. 1989. River meander model. I: Development. II: Applications. *Journal of Hydraulic Engineering, American Society of Civil Engineers*, **112**, 1117–1150.

OSTERKAMP, W. R. 1978. Gradient, discharge, and particle size relations of alluvial channels in Kansas, with observations on braiding. *American Journal of Science*, **278**, 1253–1268.

PAGE, K. & NANSON, G. C. 1982. Concave bank formation and associated floodplain formation. *Earth Surface Processes and Landforms*, **7**, 529–543.

PAOLA, C. & BORGMAN, L. 1991. Reconstructing random topography from preserved stratification. *Sedimentology*, **38**, 553–565.

PARKER, G. 1976. On the cause and characteristic scales of meandering and braiding in rivers. *Journal of Fluid Mechanics*, **76**, 457–480.

—— 1979. Hydraulic geometry of active gravel rivers. *Journal of the Hydraulics Division, American Society of Civil Engineers*, **105**, 1185–1201.

—— & ANDREWS, E. A. 1986. On the time development of meander bends. *Journal of Fluid Mechanics*, **162**, 139–156.

PIZZUTO, J. E. 1984. Bank erodibility of shallow sand bed streams. *Earth Surface Processes and Landforms*, **9**, 113–124.

PRESTEGAARD, K. 1983. Variables influencing water-surface slopes in gravel-bed streams at bankfull stage. *Geological Society of America Bulletin*, **94**, 673–678.

REID, I., BEST, J. L. & FROSTICK, L. E. 1989. Floods and flood sediments at river confluences. *In*: BEVEN, K. & CARLING, P. (eds) *Floods: Hydrological, Sedimentological and Geomorphological Implications*. Wiley, 135–150.

RICE, R. J. 1982. The hydraulic geometry of the lower portion of the Sunwapta River valley train, Jasper National Park, Alberta. *In*: DAVIDSON-ARNOTT, R., NICKLING, W. & FAHEY, B. (eds) *Research in glacial, glacio-fluvial and glacio-lacustrine systems*. Proceedings of the 6th Guelph Symposium on Geomorphology, Geo Books, Norwich, U.K. 151–173.

RICHARDS, K. S. 1982. *Rivers: form and process in alluvial channels.* Methuen, London.

ROBERTSON-RINTOUL, M. S. E. & RICHARDS, K. S. 1993. Braided channel pattern and paleohydrology using an index of total sinuosity. *This volume.*

ROY, A. G. & BERGERON, N. 1990. Flow and particle paths at a natural river confluence with coarse bed material. *Geomorphology*, **3**, 99–112.

—— & DE SERRES, B. 1989. Morphologie du lit et dynamique des confluents de cours d'eau. *Bulletin de la Societe Geographique de Liege*, **25**, 113–127.

—— & ROY, R. 1988. Changes in channel size at river confluences with coarse bed material. *Earth Surface Processes and Landforms*, **13**, 77–84.

——, —— & BERGERON, N. 1988. Hydraulic geometry and changes in flow velocity at a river confluence with coarse bed material. *Earth Surface Processes and Landforms*, **13**, 583–598.

—— & WOLDENBERG, M. J. 1986. A model for

changes in channel form at a river confluence. *Journal of Geology*, **94**, 402–411.
RUNDLE, A. S. 1985*a*. The mechanism of braiding. *Zeitschrift fur Geomorphologie, Supplement Band*, **55**, 1–13.
—— 1985*b*. Braid morphology and the formation of multiple channels, the Rakaia, New Zealand. *Zeitchrift fur Geomorphologie, Supplement Band*, **55**, 15–37.
RUST, B. R. 1972. Structure and process in a braided river. *Sedimentology*, **18**, 221–245.
—— 1978*a*. A classification of alluvial channel systems. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 187–198.
—— 1978*b*. Depositional models for braided alluvium. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 605–626.
—— 1981. Sedimentation in an arid-zone anastomosing fluvial system: Cooper Creek, Central Australia. *Journal of Sedimentary Petrology*, **52**, 745–755.
—— & NANSON, G. C. 1986. Comtemporary and paleochannel patterns and the Late Quaternary stratigraphy of Cooper Creek, Southwest Queensland, Australia. *Earth Surface Processes and Landforms*, **11**, 581–590.
SARKAR, S. K. & BASUMALLICK, S. 1968. Morphology, structure and evolution of a channel island in the Barakar River, Barakar, West Bengal. *Journal of Sedimentary Petrology*, **38**, 747–754.
SCHUMM, S. A. 1963. Sinuosity of alluvial channels on the Great Plains. *Geological Society of America Bulletin*, **74**, 1089–1100.
—— 1971. Fluvial geomorphology. *In*: SHEN, H. W. (ed.) *River Mechanics I*. Water Resources Publications, Colorado, 1–22.
—— 1972. Fluvial paleochannels. *In*: RIGBY, J. K. & HAMBLIN, W. K. (eds) *Recognition of Ancient Sedimentary Environments*. SEPM Special Publications, **16**, 98–107.
—— 1977. *The Fluvial System*. Wiley, New York.
—— 1981. Evolution and response of the fluvial system, sedimentologic implications. *In*: ETHRIDGE, F. G. & FLORES, R. M. (eds) *Recent and ancient nonmarine depositional environments: models for exploration*. SEPM Special Publications, **31**, 19–29.
—— 1985. Patterns of alluvial rivers. *Annual Reviews of Earth and Planetary Science*, **13**, 5–27.
—— & KHAN, H. R. 1972. Experimental study of channel patterns. *Geological Society of America Bulletin*, **83**, 1755–1770.
—— & LICHTY, R. W. 1963. Channel widening and floodplain construction along Cimarron River in southwestern Kansas. *US Geological Survey Professional Paper*, **352D**, 71–88.
SCHWARTZ, D. E. 1978. Hydrology and current orientation analysis of a braided-to-meandering transition: the Red River in Oklahoma and Texas, U.S.A. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 105–127.
SHELTON, J. W. & NOBLE, R. L. 1974. Depositional features of braided-meandering stream. *American Association of Petroleum Geologists Bulletin*, **58**, 742–749.
——, BURMAN, H. R. & NOBLE, R. L. 1974. Directional features in braided-meandering deposits, Cimarron River, North Central Oklahoma. *Journal of Sedimentary Petrology*, **38**, 747–754.
SINGH, I. B. & KUMAR, S. 1974. Mega- and giant ripples in the Ganga, Yamuna and Son Rivers, Uttar Pradesh, India. *Sedimentary Geology*, **12**, 53–66.
SLINGERLAND, R. L. 1977. The effects of entrainment on the hydraulic equivalence relationships of light and heavy minerals in sand. *Journal of Sedimentary Petrology*, **47**, 753–770.
—— 1984. Role of hydraulic sorting in the origin of fluvial placers. *Journal of Sedimentary Petrology*, **54**, 137–150.
—— & SMITH, N. D. 1986. Occurrence and formation of waterlaid placers. *Annual Reviews in Earth and Planetary Sciences*, **14**, 113–147.
SMITH, D. G. 1976. Effect of vegetation on lateral migration of anastomosed channels of a glacial meltwater river. *Geological Society of America Bulletin*, **87**, 857–860.
—— 1983. Anastomosed fluvial deposits: modern examples from Western Canada. *In*: COLLINSON, J. D. & LEWIN, J. (eds) *Modern and Ancient Fluvial Systems*. International Association of Sedimentologists Special Publications, **6**, 155–168.
—— & SMITH, N. D. 1980. Sedimentation in anastomosed river systems: examples from alluvial valleys near Bauff, Alberta. *Journal of Sedimentary Petrology*, **50**, 157–164.
SMITH, N. D. 1970. The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. *Geological Society of America Bulletin*, **81**, 2993–3014.
—— 1971*a*. Transverse bars and braiding in the Lower Platte River, Nebraska. *Geological Society of America Bulletin*, **82**, 3407–3420.
—— 1971*b*. Pseudo-planar stratification produced by very low amplitude sand waves. *Journal of Sedimentary Petrology*, **41**, 624–634.
—— 1972. Some sedimentological aspects of planar cross-stratification in a sandy braided river. *Journal of Sedimentary Petrology*, **42**, 624–634.
—— 1974. Sedimentology and bar formation in the Upper Kicking Horse River, a braided outwash stream. *Journal of Geology*, **82**, 205–223.
—— 1978. Some comments on terminology for bars in shallow rivers. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 85–88.
—— & BEUKES, N. J. 1983. Bar to bank flow convergence zones: a contribution to the origin of alluvial placers. *Economic Geology*, **78**, 1342–1349.
—— & MINTER, W. E. L. 1980. Sedimentological control of gold and uranium in two Witwatersrand paleoplacers. *Economic Geology*, **75**, 1–14.
—— & SMITH, D. G. 1984. William River: an outstanding example of channel widening and

braiding caused by bedload addition. *Geology*, **12**, 78–82.

SMITH, S. A. 1987. Gravel counterpoint bars: examples from the River Tywi, South Wales. *In*: ETHRIDGE, F. G., FLORES, R. M. & HARVEY, M. D. (eds) *Recent developments in Fluvial Sedimentology*. SEPM Special Publications, **39**, 75–81.

STEBBINGS, J. 1964. The shapes of self-formed model alluvial channels. *Proceedings of the Institution of Civil Engineers*, **25**, 485–510.

STEINMETZ, R. 1967. *Depositional history, primary sedimentary structures, cross bed dips, and grain size of an Arkansas River point bar at Wekiwa, Oklahoma*. Report of the Pan American Petroleum Corporation Research Department. F67-G-3.

STRAUB, L. G. 1935. Some observations of sorting of river sediments. *Transactions of the American Geophysical Union*, **16**, 463–467.

STRUIKSMA, N. & KLAASSEN, G. J. 1988. On the threshold between meandering and braiding. *In*: WHITE, W. R. (ed.) *International Conference on River Regime*. Wiley, 107–120.

SOUTHARD, J. B. & BOGUCHWAL, L. A. 1990. Bed configurations in steady unidirectional flows. Part 2. Synthesis of flume data. *Journal of Sedimentary Petrology*, **60**, 658–679.

——, SMITH, N. D. & KUHNLE, R. A. 1984. Chutes and lobes: newly identified elements of braiding in shallow gravelly streams. *In*: KOSTER, E. H. & STEEL, R. J. (eds) *Sedimentology of gravels and conglomerates*. Canadian Society of Petroleum Geologists Memoirs, **10**, 51–59.

SUKEGAWA, N. 1970. Conditions for the occurrence of river meanders. *Journal of the Faculty of Engineering, University of Tokyo*, **30**, 289–306.

SUNDBORG, A. 1956. The River Klarälven: a study of fluvial processes. *Geografiska Annaler*, **38**, 127–316.

TANDON, S. K. & KUMAR, R. 1981. Gravel fabric in a sub-Himalayan braided stream. *Sedimentary Geology*, **28**, 133–152.

TAYLOR, G. & WOODYER, K. D. 1978. Bank deposition in suspended load streams. *In*: MIALL, A. D. (ed.) *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoirs, **5**, 257–275.

TEISSEYRE, A. K. 1975. Pebble fabric in braided stream deposits. *Geologia Sudetica*, **10**, 7–56.

—— 1977*a*. Pebble clusters as a directional structure in fluvial gravels: modern and ancient examples. *Geologia Sudetica*, **12**, 79–89.

—— 1977*b*. Meander degeneration in bed-load proximal stream: repeated chute cut-off due to bar-head gravel accretion — a hypothesis. *Geologica Sudetica*, **12**, 103–120.

THORNE, C. R. & TOVEY, N. K. 1981. Stability of composite river banks. *Earth Surface Processes and Landforms*, **6**, 409–484.

TURNBULL, W. J., KRINITSKY, G. L. & WEAVER, F. J. 1966. Bank erosion in soils of the Lower Mississippi Valley. *Proceedings of the American Society of Civil Engineers*, **92** (SM1), 121–136.

VOGEL, K. R. VAN NIEKERK, A., SLINGERLAND, R. L. & BRIDGE, J. S. 1992. Routing of heterogeneous sediments over moveable beds: model verification. *Journal of Hydraulic Engineering, American Society of Civil Engineers*, **118**, 263–279.

WERRITTY, A. & FERGUSON, R. I. 1980. Pattern changes in a Scottish braided river over 1, 30, and 200 years. *In*: CULLINGFORD, R. A., DAVIDSON, D. A. & LEWIN, J. (eds) *Timescales in Geomorphology*. Wiley, 53–68.

WHITING, P. J., DIETRICH, W. E., LEOPOLD, L. B., DRAKE, T. G. & SHREVE, R. L. 1988. Bedload sheets in heterogeneous sediment. *Geology*, **16**, 105–108.

WILCOCK, P. R. & SOUTHARD, J. B. 1989. Bed load transport of mixed size sediment: fractional transport rates, bedforms, and the development of a coarse bed surface layer. *Water Resources Research*, **25**, 1629–1641.

WILLIAMS, P. F. & RUST, B. R. 1969. The sedimentology of a braided river. *Journal of Sedimentary Petrology*, **39**, 649–679.

WILLIS, B. J. 1989. Paleochannel reconstructions from point bar deposits: a three-dimensional perspective. *Sedimentology*, **36**, 757–766.

—— 1993. Ancient river systems in the Himalayan foredeep, Chinji village area, northern Pakistan. *Sedimentary Geology*, in press.

WITT, A. 1985. Vegetational influences on intrachannel deposition: evidence from the Konczak stream, greater Poland lowlands, western Poland. *Quaestiones Geographicae*, **9**, 145–160.

YALIN, M. S. 1977. *Mechanics of Sediment Transport*. 2nd ed. Pergamon Press.