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A Review of the Braided-River Depositional Environment

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ABSTRACT

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Generalized sedimentation models have been developed from a review of more than sixty recent papers on modern and ancient braided-stream deposits. Braided rivers consist of a series of broad, shallow channels and bars, with elevated areas active only during floods, and dry islands. There are three main bar types; longitudinal, comprising crudely bedded gravel sheets; transverse to linguoid, consisting of sand or gravel and formed by downstream avalanche-face progradation; and point or side bars, formed by bedform coalescence and chute and swale development in areas of low energy. Important sediment-forming processes include bar formation, channel-floor dune migration, low-water accretion and overbank sedimentation.

Braided-stream deposits consist of up to three gravel facies, five sand facies and two fine-grained facies. Vertical sequences recorded in modern and ancient deposits are of several types: flood-, channel fill-, valley fill-, channel re-occupation- and point bar-cycles. Some of these fine upward and could be confused with meandering-river sequences. Facies assemblages and vertical sequences fall into four main classes, which are proposed as sedimentation models for the interpretation of ancient braided-river deposits in the surface and subsurface:

(1) Scott type: consists mainly of longitudinal bar gravels with sand lenses formed by infill of channels and scour hollows during low water.

(2) Donjek type: may be dominated by sand or gravel; distinguished by fining-upward cycles caused by lateral point-bar accretion or vertical channel aggradation. Cycles commonly are less than 3 m thick, but cycles up to 60 m may be present, representing valley-fill sequences. Longitudinal and linguoid-bar deposits, channel-floor dune deposits, bar-top and overbank deposits all may be important.

(3) Platte type: characterized by an abundance of linguoid bar and dune deposits (planar and trough crossbedding). No well-developed cyclicity, probably owing to a lack of topographic differentiation in the river (no evidence of deep, primary channels, abandoned areas or overbank areas).

(4) Bijou Creek type: consists of horizontally laminated sand plus subordinate amounts of sand showing planar crossbedding and ripple marks. Formed during flash floods and may be most typical of ephemeral streams.

INTRODUCTION

Braided rivers and meandering rivers commonly are regarded as the two main river types in the geological literature. Meandering rivers, it is now widely understood, form deposits primarily by the action of lateral accretion on point bars within the concave sides of meanders, coupled with a lesser amount of vertical accretion on floodplains, resulting from overbank flooding. The rivers are of high sinuosity (sinuosity is the ratio of channel length to length of meander-belt axis) and their deposits typically form fining-upward cycles. Allen (1970) has provided the most complete description and interpretation of these sequences.

There is much less agreement in the literature about what constitutes a braided stream and what a typical braided-stream deposit looks like. Braided rivers are recognized as those comprising several or many channels are of lower sinuosity and carry a coarser sediment load than do meandering rivers; but no comprehensive model of braided-river sedimentation has yet been developed.

The only attempt at a generalized discussion is that by Ore (1964), and since it was published over sixty detailed papers on modern and ancient braided-river deposits have appeared in the English language. Nevertheless, the braided environment has not been thoroughly discussed in recent reviews of fluvial sedimentation, for example that by Visher (1972). A brief comparison between the characteristics of braided and meandering streams was given by Shelton and Noble (1974), Cant and Walker (1976) and Walker (1976).

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The purpose of the present paper is to review this subject and to develop a unified framework within which ancient braided-stream deposits can be interpreted. As a starting point it is necessary to define more precisely what is meant by the term "braided", in comparison to other river types.

CLASSIFICATION OF RIVER MORPHOLOGY

Various terms are used to describe different types of rivers: meandering, braided, high-sinuosity, low-sinuosity, suspension-load, bed-load, etc. A classification of rivers is given in Table I, based mainly on the work of Leopold and Wolman (1957) and Schumm (1963, 1968a,b). The morphology of the river types is illustrated in Fig. 1.

Meandering channels have received full treatment in the literature and will not be discussed in detail in this paper, except in so far as their deposits contrast with those of braided rivers.

Rivers with straight channels are commonly mixed- or suspension-load streams which occur on relatively gentle slopes (Schumm, 1968a, p. 1579). The channel sinuosity may be nearly equal to 1.0, but the thalweg, or line of maximum depth, meanders back and forth within the channel, and low mud or silt bars are deposited along the channel edges on the insides of the meanders (Leopold and Wolman, 1957, p. 53). Modern straight rivers are rare and little is known regarding their deposits. The Devonian low-sinuosity stream deposits of Moody-Stuart (1966) and "Facies association 2" of Leeder (1973a) may be of this type (see later section).

Schumm (1968a, p. 1580) proposed that the term "anastomosing" be restricted to rivers with relatively permanent and stable systems of high-sinuosity channels with cohesive banks, separated by large, stable, vegetation-covered islands. Such rivers are particularly common in parts of Southern Australia. D.G. Smith (1973) adopted the same definition in his studies of glacial outwash streams.

Braided streams are characterized by high width/depth ratios, possibly exceeding 300, steep slopes and, generally, low sinuosities. It has been found empirically that braided and meandering streams can be distinguished on the basis of slope and discharge, according to the equation:

$$S = 0.013Q_b{}^{-0.44}$$

where S is slope, and Q_b is bankfull discharge in m^3/sec (based on Leopold and Wolman, 1957). For a given discharge braided streams occur on slopes steeper than that given by the equation, whereas meandering streams occur on gentler slopes. Braided streams consist of a series of rapidly shifting channels and mid-channel bars. In some braided streams bars may be stable enough for vegetation to become established, converting them into islands, but even these tend to be flooded during periods of high discharge. The deposits of braided streams normally are coarser than those of the other river types, and are dominated by sand or gravel. There are some exceptions to these general-

TABLE I

Classification of river types

Type	Morphology	Sinuosity	Load type	Bedload per-cent (of total load)	Width/depth ratio	Erosive behaviour	Depositional behaviour
Meandering	single channels	>1.3	suspension or mixed load	<11	<40	channel incision, meander widening	point-bar formation
Braided	two or more channels with bars and small islands	<1.3	bedload	>11	>40	channel widening	channel aggradation, mid-channel bar formation
Straight	single channel with pools and riffles, meandering thalweg	<1.5	suspension, mixed or bedload	<11	<40	minor channel widening and incision	side-channel bar formation
Anastomosing	two or more channels with large, stable islands	>2.0	suspension load	< 3	<10	slow meander widening	slow bank accretion

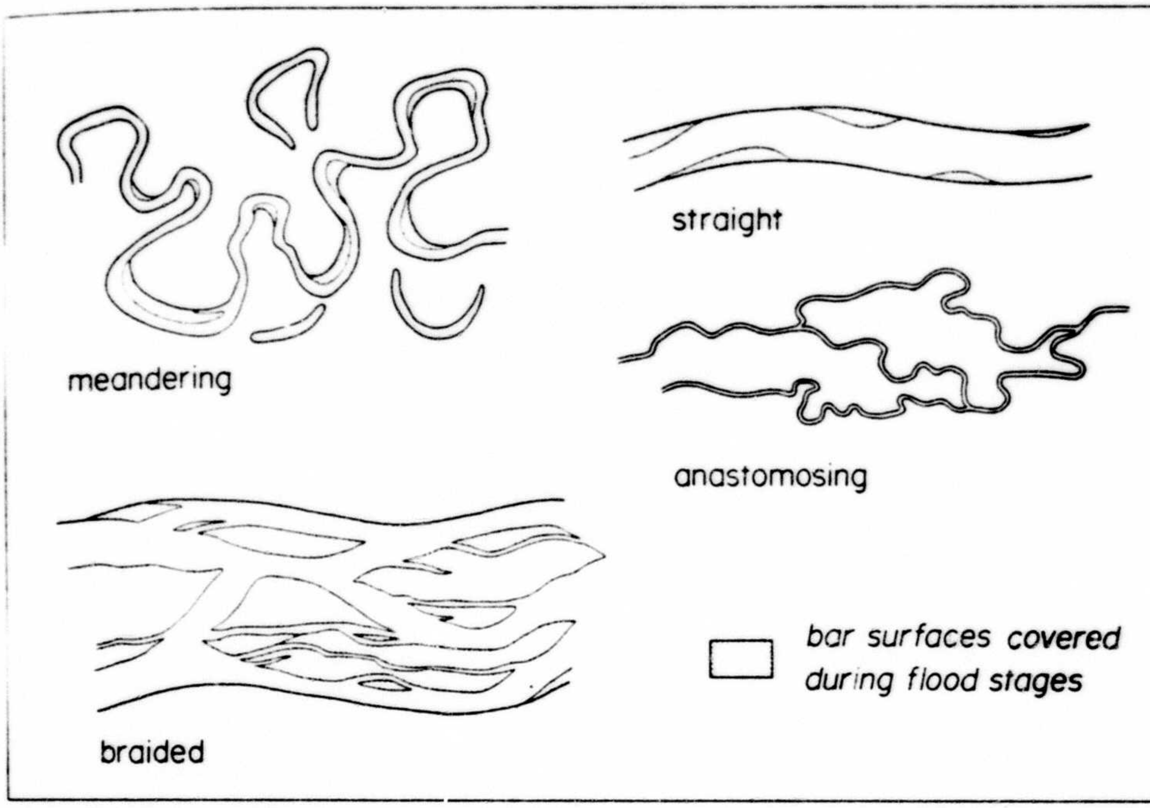


Fig. 1. Principal river types.

izations; for example, parts of the Donjek River (Yukon) show braided channels within a meandering river tract, the sinuosity of the main channel being 1.74 (Rust, 1972, Fig. 2C). The Amite River (Louisiana) is a bed-load river with sedimentary deposits and bedforms similar to those of typical braided rivers, yet it is a largely non-braided, meandering stream with a sinuosity of 1.4 to 1.7. The Endrick River (Scotland) is another example of a high-sinuosity, bed-load stream. It contains extensive gravel point bars (Bluck, 1971). D.G. Smith (1976) demonstrated experimentally the effectiveness of vegetation in stabilizing channel patterns. Where the banks are not easily erodible the development of the braiding pattern is inhibited, and this probably is the case in the Amite and Endrick rivers.

CAUSES OF BRAIDING

As shown by Leopold and Wolman (1957), at least nine variables interact to determine the nature of the resulting stream channel. They include discharge (amount and variability), sediment load (amount and grain size), width, depth, velocity, slope and bed roughness. Schumm (1968a) showed that the amount and type of vegetation growth also will affect stream type and, therefore, climatic and geological factors must also be considered. It is not yet possible to define the ranges of values that will invariably produce a river of a given type although, as noted below, certain interrelationships between the variables are now well enough understood for some generalizations to be made. Leopold and Wolman (1957, p. 72—73) stated:

> "Channel patterns, braided, meandering, and straight, each occurs in nature thoughout the whole range of possible discharges. Some of the largest rivers in the world are braided; for example, the lower Ganges and Amazon. More are meandering, of which the lower Mississippi is the best known example. Meanders are common in very small creeks and braids are common in many small ephemeral streams . . . [It has been observed that] a given channel can change in a short distance from a braid to a meander or vice versa, that the divided channels of a braid may meander, and that a meandering tributary may join a braided master stream. Such changes in a given channel or such different channels in juxtaposition can be attributed to variations in locally independent factors".

Wright et al. (1974) carried out multivariate statistical analyses on thirty-four modern alluvial-deltaic systems representing all climatic zones and a wide range of sizes. They performed cluster analyses on two groups of data pertaining to the alluvial valley. In the first, the relative abundance of seven flood-plain features were considered in a landform analysis: river scars, swamps (densely vegetated), lakes, barren flats (including evaporite flats), sand dunes, and braided and meandering channels. In the second analysis, four salient features of the discharge regime were compared: mean annual discharge, a coefficient of variability for mean monthly discharge, total discharge range and discharge flood peakedness.

The landform analysis revealed seven main clusters, in three of which (clusters 1, 3, 5 of Wright et al., 1974, Figs. 11, 12) braided channels were abundant. They tended to be associated with river scars, as would be expected. Wright et al. (1974, p. 36) stated that: "a comparison of the dendrograms for alluvial-valley morphology and discharge regime . . . suggests that, even though there are a few pairs of river systems which are similar to one another in both respects, the two sets of clusters generally exhibit minimal coincidence. The implication is simply that the composite alluvial valley landscape is not dependent solely on discharge regime". However, an examination of the discharge regime data (their Table 11) shows that those rivers dominated by braided, as opposed to meandering channels, have on average a higher flood peakedness, higher total discharge range and higher monthly discharge variability.

These correlations, which appear to have been missed by Wright et al. (1974), are of critical importance in understanding the braiding process. "Braiding is developed by sorting as the stream leaves behind those sizes of the load which it is incompetent to handle . . . if the stream is competent to move all sizes comprising the load but is unable to move the total quantity provided to it, then aggradation may take place without braiding" (Leopold and Wolman, 1957, p. 50). The action of depositing the coarser bed-load initiates mid-channel bar formation, the details of which are discussed in a later section. In rivers of highly variable discharge competency will be similarly variable, and there will be long periods of time throughout which the river will be unable to move at least the coarsest part of its bed-load. The incidence of bar initiation, flow diversion and the creation of new channels (braiding) will thus be high.

The conditions of abundant, coarse, non-cohesive bed-load, strongly fluctuating discharge and steep slope are not typical of any particular tectonic or climatic setting. An examination of the alluvial data tabulated by Coleman and Wright (1975, Table 1) shows that on a world-wide basis braided rivers are equally as common as meandering rivers in Arctic, temperate, dry tropical and humid tropical regions. There are two main reasons for this:

(1) The sediment load and discharge characteristics of a river may partially reflect the climate and relief of a source area many hundreds of kilometres distant; for example tropical rivers such as the Ganges, Brahmaputra and Mekong, all of which are strongly braided, have headwaters located in the Himalayan Mountains.

(2) The causes of a high bed-load and a fluctuating discharge are diverse. The following are some of the main ones:

(a) Alpine source areas provide strong relief and a predominance of mechanical over chemical weathering for the generation of coarse clastic debris. Discharge is markedly peaked during periods of spring snow melt. Glacial outwash streams are almost invariably braided. The deposits they form are referred to as sandurs (and have been the subject of several of the best recent studies of braided-stream sedimentation).

(b) Marked discharge fluctuations also are characteristic of Arctic, arid and monsoonal climatic areas.

(c) A lack of vegetation in a drainage basin means a lack of water and sediment-storage capacity, and consequent immediate response to storms in the form of flash floods. The degree of vegetation cover in an area is mainly controlled by climate, but removal of the cover by deforestation or fire can produce the same catastrophic flooding effects as characterize arid regions (Chawner, 1935).

(d) During most of geological time, until the Early or Middle Devonian, land vegetation had a very restricted distribution (Seward, 1959). Sediment transportation characteristics would, therefore, have been similar to those of modern arid regions — a predominance of bed-load rivers with strongly fluctuating discharges (Schumm, 1968a, p. 1583).

A given part of a river tract may change its morphology through time in response to climatic or other effects. Schumm (1968b) has documented in detail the change in the Murrumbidgee River (Australia) from a bed-load to a suspension-load type during the Late Quaternary, in response to a gradual increase in humidity. Steel (1974) and Karl (1976) recorded a similar evolution in some Permo-Triassic rocks in Scotland and some Cretaceous sediments in Nebraska, respectively. Conversely, in the Cimarron River, Kansas, a single major flood in 1914 changed the river morphology from a meander-

TABLE II

Hydrology of some modern braided rivers

River	d_b	w	Q_m	Q_{ma}	Q_{max}	v	Source
Brahmaputra	15	13	19000	40000	71000	2.4	Coleman (1969)
Ganga	<10	—	—	43300	—	3.4	Singh and Kumar (1974)
Donjek	3	0.4—2.2	—	1400	—	3.6	Rust (1972)
Durance	6.6	0.7	188	—	5200	6	Doeglas (1962)
Lower Platte	—	0.4—0.6	109	—	3200	—	N.D. Smith (1971a)
Slims	5	0.3—1.8	—	570	—	>2	Fahnestock (1969)
Amite	6.7	0.07	—	1400	—	—	McGowen and Garner (1970)
Tana	15	0.6—2	150	1800	3500	—	Collinson (1970)
Bijou Creek	2.4—3.6	0.2—0.7	—	—	13000	6	McKee et al. (1967)

d_b = average bankfull depth (m); w = bankfull width (km); Q_m = mean annual discharge (m^3/sec); Q_{ma} = mean annual flood (m^3/sec); Q_{max} = maximum recorded discharge (m^3/sec); v = flood velocity (m/sec).

ing, suspension-load stream to a broad, shallow, braided, bed-load stream. Channel widening continued until 1942, during a period of below-normal precipitation, which inhibited vegetation growth (Schumm and Lichty, 1963). An example of a change from suspension-load to bed-load sedimentation in the Devonian was described by Leeder (1973a). A series of complex regime changes in Devonian rocks was described by Allen (1974a).

To summarize, the controls of river morphology are complex and, as a result, the occurrence of a particular river type does not, by itself, lead to any unambiguous conclusions regarding climate or relief (contrary to the conclusions of Visher, 1976). Factors of both local and distant origin, acting over the short and long term, can have pronounced but delicately balanced effects on river regimen.

HYDROLOGY OF BRAIDED RIVERS

Some channel-size and discharge measurements from a range of modern braided rivers are given in Table II. Flood velocities are high, and the grade of material transported during peak flow may be limited only by the nature of the sediment available at the time. This point was well illustrated by a flood event in Bijou Creek, Colorado, the deposits of which were described by McKee et al. (1967). Up to 3.6 m of fine to coarse sand were deposited during this flood, without significant quantities of coarser material; yet the flood was sufficiently powerful to destroy highways and to carry a 11 m long bridge girder a distance of 330 m.

MORPHOLOGY OF BRAIDED RIVERS

Topographic levels

Braided rivers consist of two or more channels divided by bars and islands. In most examples "a single dominant channel can generally be distinguished within the overall braided pattern, although in some sections there are several principal channels" (Rust, 1972, p. 223). As described in detail by Williams and Rust (1969, pp. 650—652) and Kessler and Cooper (1970) several distinct topographic levels may be recognized in many braided rivers, ranging from the area of the deepest and most active channels, to elevated, abandoned areas, commonly vegetation-covered. These levels represent stages of progressive downcutting by the river, and will be most readily recognizeable in braided streams within valleys surrounded and fed by areas of strong relief, where degradation is active. The Donjek River (Yukon) is an example of this type. Brief descriptions of the four topographic levels recognized in the Donjek are given below (summarized from Williams and Rust, 1969, pp. 650—652) and an aerial view of the river showing these features is provided in Fig. 2.

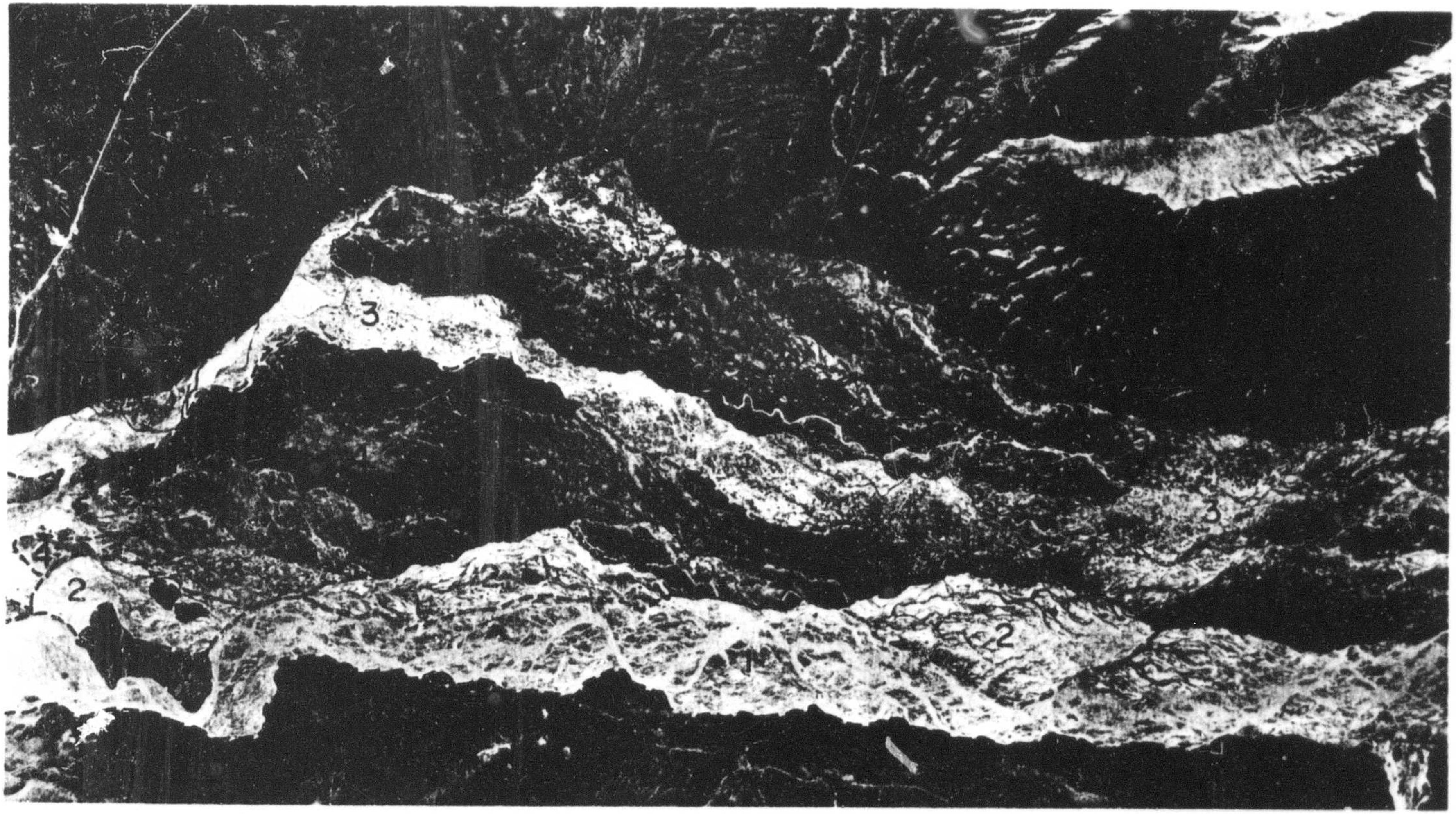


Fig. 2. Vertical airphoto of part of the Donjek River, Yukon Territory, showing the four topographic levels defined by Williams and Rust (1969) (part of photo A15728-89, National Airphoto Library, Canada).

Level 1: level of main channels and principal sediment dispersal route, little or no vegetation, bars exposed during low water.

Level 2: level active only during flood stage, with few active channels at other times, sparse vegetation cover.

Level 3: little continuous water movement, low-energy flow during flood stage, moderate vegetation cover.

Level 4: dry islands with dense vegetation.

Topographic differentiation is visible in other valley braided systems. for example the Scott and Yana glacial outwash rivers of southern Alaska (Boothroyd and Ashley, 1975, Fig. 3), the South Saskatchewan River (Cant, 1975), the South Canadian River, Texas (Kessler and Cooper, 1970), and in some sandur rivers described by Church and Gilbert (1975, p.62).

Beyond the mountain front, where degradation is less rapid, topographic differentiation will be less marked. For example in the Lower Platte River (Nebraska) permanent, vegetation-covered islands are present, but the remainder of the river tract is regularly covered by each spring flood (N.D. Smith, 1971a, p. 3409). The Tana River (Norway) is similar, although Collinson (1970, Fig. 4) illustrated a single large area of coalesced bar deposits with a light vegetation cover, which corresponds to levels 2 and 3 of the Donjek. Coleman (1969, p. 145) described similar features (called chars, locally) in the Brahmaputra. Most are unstable and change shape and position rapidly, but some become vegetated and evolve into semi-permanent islands.

The higher levels are eventually abandoned as terraces, such as those described by Church (1972, p. 19, Fig. 11) and Corner (1975). Any special features of the braiding process that lead to island aggradation and the enlargement of the higher levels before complete abandonment may be important geologically, because the removal of such areas from the region of active erosion could be the first step in the preservation of the deposits as part of the geological record. Modification of earlier deposits by later degradational events can produce complex facies relationships, such as are described by Eynon and Walker (1974).

Bars: Classification and origin

Numerous adjectival terms have been used to describe bars in braided streams, including scroll, spool, diamond, longitudinal, lobate, linguoid, transverse, side, lateral, medial, diagonal, alternating, meander, unit, cross channel, chute and point bars. A review of the terminological discussions contained in the literature (e.g., N.D. Smith, 1971a, p. 3410) would only add to the confusion and is omitted from this paper. Recent work by Hein (as reported in Harms et al., 1975, pp. 150—153) and Hein and Walker (1977) has indicated that certain bar types are members of evolutionary sequences and that the morphology of any particular bar type may be very ephemeral. From the geological viewpoint a bar definition is of use only if it can be recognized in the geological record by a distinctive lithologi sequence or assemblage of structures. An all-encompassing, but simple clas sification therefore is required, and this is proposed below. Illustrations o bar morphology based on examples described in the literature are given i Fig. 3.

Longitudinal bars

These are diamond- or lozenge-shaped in plan, and are elongated paralle to flow direction. They are bounded by active channels on both sides an may, as a result, have partially eroded margins. Bars formed in gravel ar most commonly of this type (Rust, 1972; N.D. Smith, 1974; Gustavson 1974; Boothroyd and Ashley, 1975) although some sand bars show simila morphology (Coleman, 1969, Fig. 20; N.D. Smith, 1970, p. 2999; G.E Williams, 1971, p. 15). In many instances the characteristic shape may b the result of erosion rather than deposition, the original bar being of quit different type (e.g., Boothroyd and Ashley, 1975, Fig. 19).

Longitudinal (= medial) bars are the classical braid bars of Leopold an Wolman (1957) and Allen (1968, p. 38). The process by which they forn was fully described by Leopold and Wolman (1957) on the basis of flum experiments and field observations. Similar observations have been repeate by Rust (1972), Schumm and Khan (1972), N.D. Smith (1974) and Churcl and Gilbert (1975, p. 75). The sequence of events is illustrated in Fig. 4 an is summarized below.

In an originally single or undivided channel the coarsest load is carrie along the deepest portion of the channel where competency is greatest Much of it accumulates as a lag in scour pools (D.G. Smith, 1973). Wanin flow, of whatever cause (waning flood, diurnal variation, etc.) or a reductio in competency where a channel widens will result in the deposition of par

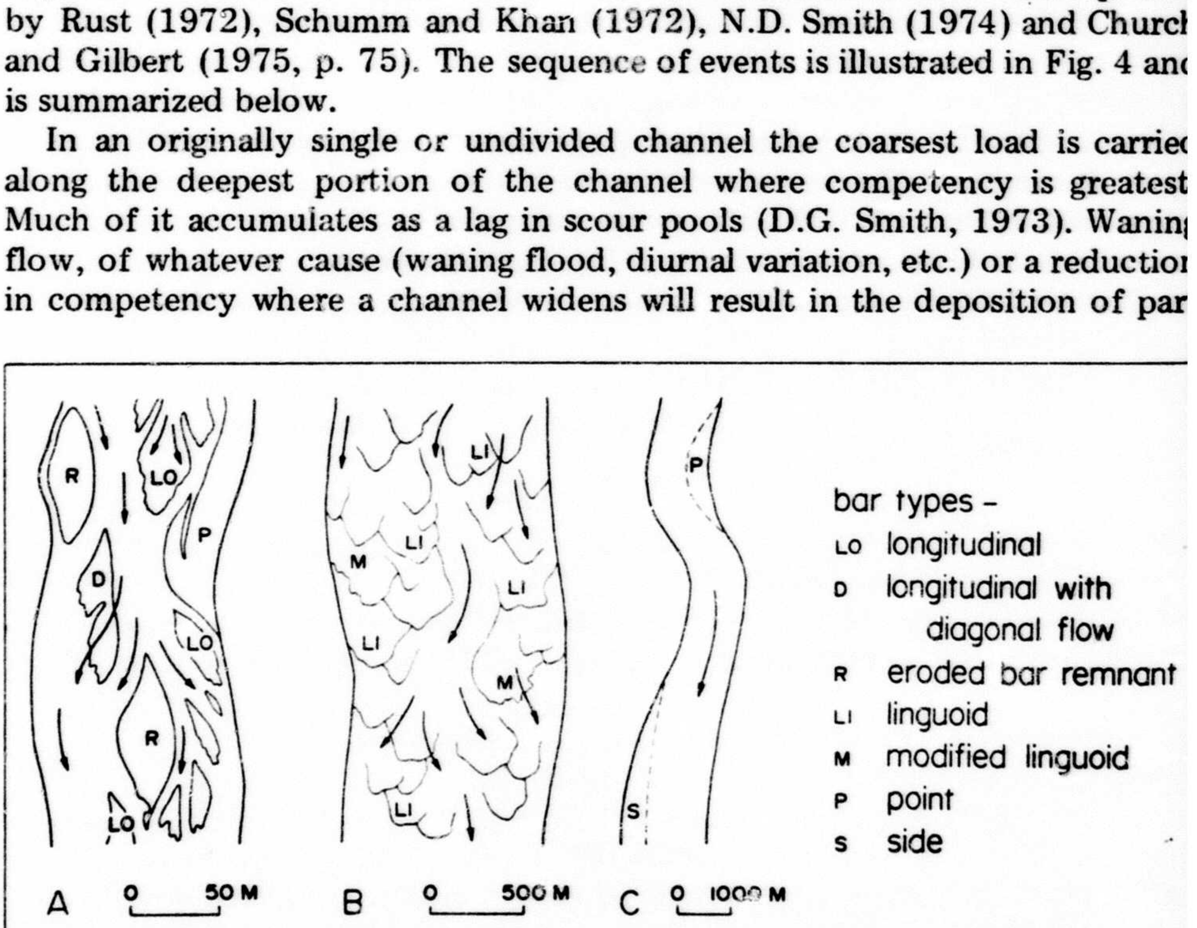


Fig. 3. Principal bar types.

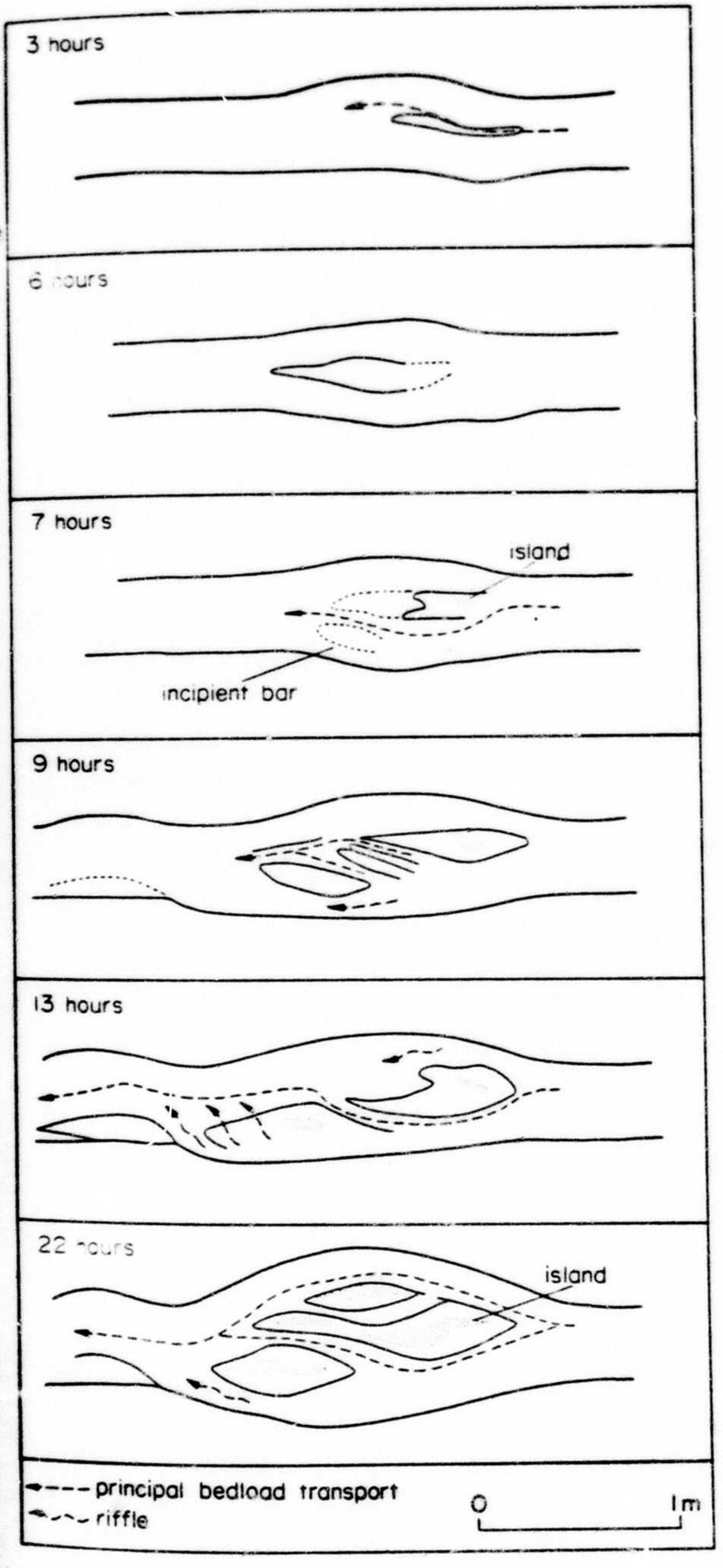


Fig. 4. The growth of longitudinal bars in a flume. Redrawn from Leopold and Wolman (1957).

of the coarsest bedload as a short, submerged, central bar — described as
"diffuse gravel sheet" by Hein (in Harms et al., 1975, p. 150) and Hein an
Walker (1977), over which the flow forms a riffle. Finer particles are trappe
in the interstices of the initial deposit, and more bedload is deposited down
stream, in the lee of the bar, so that growth continues. The flow diversio
causes the now divided channel to erode its banks (or an adjacent ba
deposit) and general channel widening results. The initial bar relief may b
no greater than the size of the largest clast (Boothroyd and Ashley, 1975, p
202) but as growth continues it may increase to as much as a metre (Rust
1972, Fig. 8). Bar length may reach several hundred metres, but the longe
examples probably represent composite, coalesced forms. The coarsest mate
rial is concentrated along the central bar axis (Boothroyd and Ashley, 1975
p. 202) and generally grain size diminishes upwards and downstream (N.D
Smith, 1974, p. 219). The internal structure of the bars is massive or crud
horizontal bedding (N.D. Smith, 1970, p. 2999), possibly indicating trans
portation in planar sheets, under very high flow energy (Rust, 1972, p. 243)
Longitudinal bars may also develop as stable bedforms analogous to dunes o
antidunes, under high flood conditions (Hein, in Harms et al., 1975, p. 150
Rust, 1975, p. 246; Hein and Walker, 1977) but the hydrodynamic mech
anisms of such a process have yet to be described.

Avalanche faces in longitudinal gravel bars are rare, and most longitudina
bars taper off downstream, except where their margins have been subjecte
to erosion. Church (1972), N.D. Smith (1974), Church and Gilbert (1975, p
58) and Hein and Walker (1977) recognized a distinct bar type, termed a dia
gonal bar, which is characterized by asymmetric flow, across the bar rathe
than parallel to the bar axis. This occurs anywhere in the channel, and mos
commonly at bends or junctions. Apart from the asymmetry there seems t
be little to distinguish the diagonal from the longitudinal bar. The two grad
into each other, as shown by Krigstrom (1962) and Boothroyd and Ashle
(1975, Fig. 12), and it is recommended that the more general, and mor
widely used term longitudinal bar be used for all such structures. A simila
point is made by Hein and Walker (1977).

Linguoid and transverse bars

These have been described and illustrated by Allen (1968), Collinso
(1970), N.D. Smith (1970, 1971a, 1972, 1974) and G.E. Williams (1971)
Lobate bar is an alternative term for the linguoid type. The chute bars o
McGowen and Garner (1970) and the meander bars of Karcz (1972) are als
of linguoid type.

Linguoid and transverse bars are most typical of sandy braided river
They are rare in the Donjek (Rust, 1972, p. 229), which is predominantly
pebbly river, but are common in the equally pebbly Kicking Horse River o
British Columbia (N.D. Smith, 1974; Hein and Walker, 1977). Other exam
ples consisting of gravel are listed by Rust (1975, pp. 245—246) who pointe
out that many occur in channels which are deep, and are confined betwee

relatively narrow banks. Such conditions are not typical of the aggrading braided environment (Table I), but they may occur in fan-head trenches, such as are described by McGowen and Groat (1971, p. 36). Hein and Walker (1977) suggested that gravel bars with foresets may develop in the same channel as longitudinal bars only under conditions of reduced sediment and water discharge.

The characteristic shape of linguoid bars is rhombic or lobate, with upper surfaces which dip gently upstream towards the preceding bar, and downstream facing, sinuous, avalanche-slope terminations. The bars vary in width from a few metres to 150 m and in length up to 300 m. The maximum height at the crest of the avalanche slope ranges from a few centimetres to 2 m but the most typical size range is 50 to 100 cm. Linguoid bars commonly occur in trains, and generally are in an out-of-phase relationship with one another, such that "the convex front of a bar tends to advance into the space between the two preceding bars" (Collinson, 1970, p. 39). Small delta-like bars grow at the mouths of chutes or prograde out from obstructions, and these also may have a linguoid morphology. The linguoid bars described by Church and Gilbert (1975, Fig. 32) appear to lack downstream avalanche slopes and would be termed longitudinal bars in the present classification.

An asymmetric bar type similar in internal structure to linguoid bars is present in the South Saskatchewan River (Walker, 1976). These bars form around initial nuclei, with foreset slopes at high angles to the channel direction. Their development may be related to the confined nature of the flow in a deeply incised channel. Walker (1976) refers to them as "horns" or "wings".

Linguoid bars that are exposed to view in modern rivers commonly are covered by dunes and ripples (Collinson, 1970; N.D. Smith, 1971a, Figs. 3, 8), but it seems probable that these structures are simply superimposed on the bar surface during low water stages and that, in fact, the bars are themselves large-scale bedforms generated during flood stages (Collinson, 1970, p. 41). N.D. Smith (1971a, p. 3411) suggested that they form by the coalescence of dune fields, and this would appear to be confirmed by the observations of Culbertson and Scott (1970). They observed that bars were generated only at a discharge in excess of 28 m^3/sec., whereas the bar-modification processes documented by N.D. Smith (1971a, Fig. 8) occurred when discharge over the bar was less than 4 m^3/sec. Internally the predominant structure of linguoid bars is planar-tabular crossbedding representing avalanche-slope progradation.

Published descriptions of transverse bars (N.D. Smith, 1971a, 1972, p. 625; G.E. Williams, 1971, pp. 15—18) indicate that they are genetically similar to linguoid bars, as described above, except that they tend to have straighter crests. They may represent coalesced linguoid bars or solitary bars that extend completely across a channel or the space between obstructions, such as trees or bushes. The two bar types probably could not be distinguished in the geological record except, perhaps, by very detailed palaeocurrent work (their recognition in modern streams tends to be arbitrary, in the opinion of the writer). Diagonal transverse bars develop in areas of asymmetric flow, as in gravelly rivers but, again, their recognition in the ancient record may be impossible. Perhaps a generalized structural as opposed to morphological term such as "*sandy foreset bars*" would be adequate to describe this entire class of structures in the ancient record. They are mesoforms in Jackson's (1975) terminology.

Point bars, side bars, lateral bars

Genetically these bars are similar. They form in areas of relatively low fluvial energy, such as the inside of a meander where the main current strength is diverted against the opposite, outer bank. Point bars tend to be thought of as typical of meandering streams, but they also occur in braided environments, such as in the Kicking Horse River (N.D. Smith, 1974, Fig. 7c), and some Scottish rivers (Bluck, 1976). Collinson (1970, Fig. 5) illustrated a side bar (or lateral bar) 6 km in length, occupying one complete side of the Tana River tract. Alternating bars form in straight channels within the meanders formed by the meandering thalweg. As such they are analogous to point bars. The "side-flats" and "compound flats" of Cant (1975) appear to belong in this category of bar deposits.

This group of bar structures is of a larger order of magnitude than the types discussed above. They tend to form in braided rivers by the coalescence of smaller bedforms, such as dunes and linguoid bars (the "unit bars" of N.D. Smith, 1974, and Walker, 1976), as is well illustrated by the Tana side bar of Collinson (1970, Fig. 5), and the point bars of the Colorado (Texas) and Amite (Louisiana) rivers, as described by McGowen and Garner (1970). Vertical and lateral accretion are both important. Many of the islands in braided streams probably develop in the same way (e.g., Sarkar and Basumallick, 1968). The internal structures are correspondingly complex, and may include planar-tabular crossbedding of linguoid-bar origin, trough crossbedding of dune or scour origin, ripple marks of various types, coarse-grained lag deposits formed in chutes, and fine-grained drape and fill deposits formed in swales. Recognition of the specific point-bar or lateral-bar setting of such deposits may not be possible in the ancient record, unless the edge of the river tract to which the bar is attached is preserved, as a large-scale, cross-cutting scour surface. A general term, "*compound bar*", is suggested for these structures. They are macroforms in Jackson's (1975) terminology.

Other large-scale bar forms

Most of the rivers which have been discussed above are of small to average size, viewed on a world scale, and their bar structures and bedforms are of a size that generally can be recognized readily in the relatively small outcrops that normally are available to us. However, much larger rivers undoubtedly existed in the past, as they do today, and the scale of structures they created may be so large that exceptionally good exposure or very careful mapping may be needed to reveal their existence. Coleman (1969, p. 190) described sand waves in the Brahmaputra River that are 8 to 15 m high and 180 to well over 900 m in wavelength. Such waves may migrate as much as 600 m in 24 hours. Internally they comprise extremely large-scale sets of trough and planar crossbedding (Coleman, 1969, Fig. 38A). P. McCabe (personal communication, 1976) has mapped crossbedding 40 m high in what he interprets as a delta distributary deposit in the Carboniferous of England. Conaghan and Jones (1975) describe trough and planar crossbed sets more than 5 m thick which they interpret as the deposits of Brahmaputra-type sand waves in a fluvial setting.

Channel characteristics

Observations on the morphology of modern braid channels are limited, because the most active are constantly water-filled, and bottom structures are obscured by turbidity. Profile measurements by Coleman (1969) on the Brahmaputra, Neill (1969) on the channel bottom of the Lower Red Deer River (Alberta) and Culbertson and Scott (1970) on a conveyance channel in New Mexico show that dunes and bars (probably of linguoid type) of a variety of scales are present. Their internal structures are not known, but there is no reason to doubt that they are the same as those observed on exposed levels of braided-river tracts. The record of ancient braided-stream deposits shows that channel deposits rest on scour surfaces and commonly contain a basal lag of gravel (Miall, 1970a; N.D. Smith, 1970; Cant and Walker, 1976). Cant (1976) stated that the lower parts of the sedimentary structures created by large bedforms will tend to be preserved because smaller bedforms cannot cause deep erosion. The deposits of infrequent large floods should, therefore, be preserved selectively.

Morphological modifications due to waning flow

"The large and rather sudden changes in water discharge throughout the summer mean that the bed is seldom, if ever, in equilibrium with the flow" (Collinson, 1970, p. 43). The response to waning flow is twofold: higher relief structures may be eroded or dissected, and smaller-scale structures may be superimposed on or built out from the larger bedforms. Linguoid and transverse bars may continue to accrete as dunes and ripples migrate over the upper surface, carrying sand to the crest of the avalanche slope (Collinson, 1970, Fig. 27; N.D. Smith, 1971a, Fig. 8; 1972, Fig. 4). Elsewhere, new, shallow channels may dissect the bar surface (N.D. Smith, 1972, p. 3414; Karcz, 1972, p. 177). Wedge-shaped sand units may build out in the lee of longitudinal gravel bars (Rust, 1972, Fig. 4; Boothroyd and Ashley, 1975, Fig. 17A) or infill erosional channels (Boothroyd and Ashley, 1975, Fig. 14A). Bar relief tends to be smoothed over, as a result of sheet run-off or wave action in pools of standing water (Collinson, 1970), but channels that continue to be active may erode steep erosional banks on lateral bar-margins (Rust, 1972, p. 243). The last stage in a waning flood is the deposition of thin silt or mud drapes and channel fills in inactive areas.

The internal structures created by some of these processes are quite diagnostic of waning-stage origin, as will be discussed in the next section.

DEPOSITIONAL FACIES PART 1: DESCRIPTION

The first part of this section is a review of all the main lithofacies and sedimentary structure assemblages that have been recorded in modern braided-stream deposits, with a summary of who has recorded what in both the modern and ancient record (Table III). The second part of the section is intended to be a guide to the interpretation of these facies in terms of the local depositional environment. Such interpretations cannot by themselves provide conclusive identification of a braided-stream deposit in the ancient record, because many of the facies occur in other environments. The association of facies, particularly the vertical profile, also must be considered.

Gravelly facies

Individual clasts may reach more than 20 cm in diameter but the mean size generally is within the pebble range (2—64 mm). Sorting is variable.

Facies Gm: massive or crudely bedded gravel. This facies comprises pebble- or cobble-gravels in which crude horizontal stratification may or may not be apparent (Fig. 5). Most gravels are clast-supported, indicating that the matrix (sand and silt) filtered into the interstices following deposition. Some matrix-supported gravels also may be present, suggesting debris-flood deposition. Impersistent lenticles of clay, silt or crossbedded sand may be interbedded with the gravels. Typical dimensions for individual units range up 1 m in thickness and several hundred metres in lateral extent. Superimposed units may reach 4 m or more in thickness. Clasts commonly are imbricated and lateral grain-size variation may be apparent. The facies normally has an erosive base. Minor sedimentary structures associated with Facies Gm include small, dune-like accumulations of small, well-sorted pebbles (Martini and Ostler, 1973) and regularly spaced ribs of pebble- to boulder-sized material oriented perpendicular to the flow direction (McDonald and Banerjee, 1971; Gustavson, 1974; Boothroyd and Ashley, 1975). Neither has been recorded in any ancient braided-river deposits.

Facies Gt: trough-crossbedded gravel. These gravels are distinctly stratified, comprising beds of varying pebble grain size. They form broad, shallow channels, typically 20 cm to 3 m deep and 1 to 12 m wide. The channels commonly cut into each other both laterally and vertically. Each thus has an erosional base, and this may be followed by a lag deposit of coarser grain size

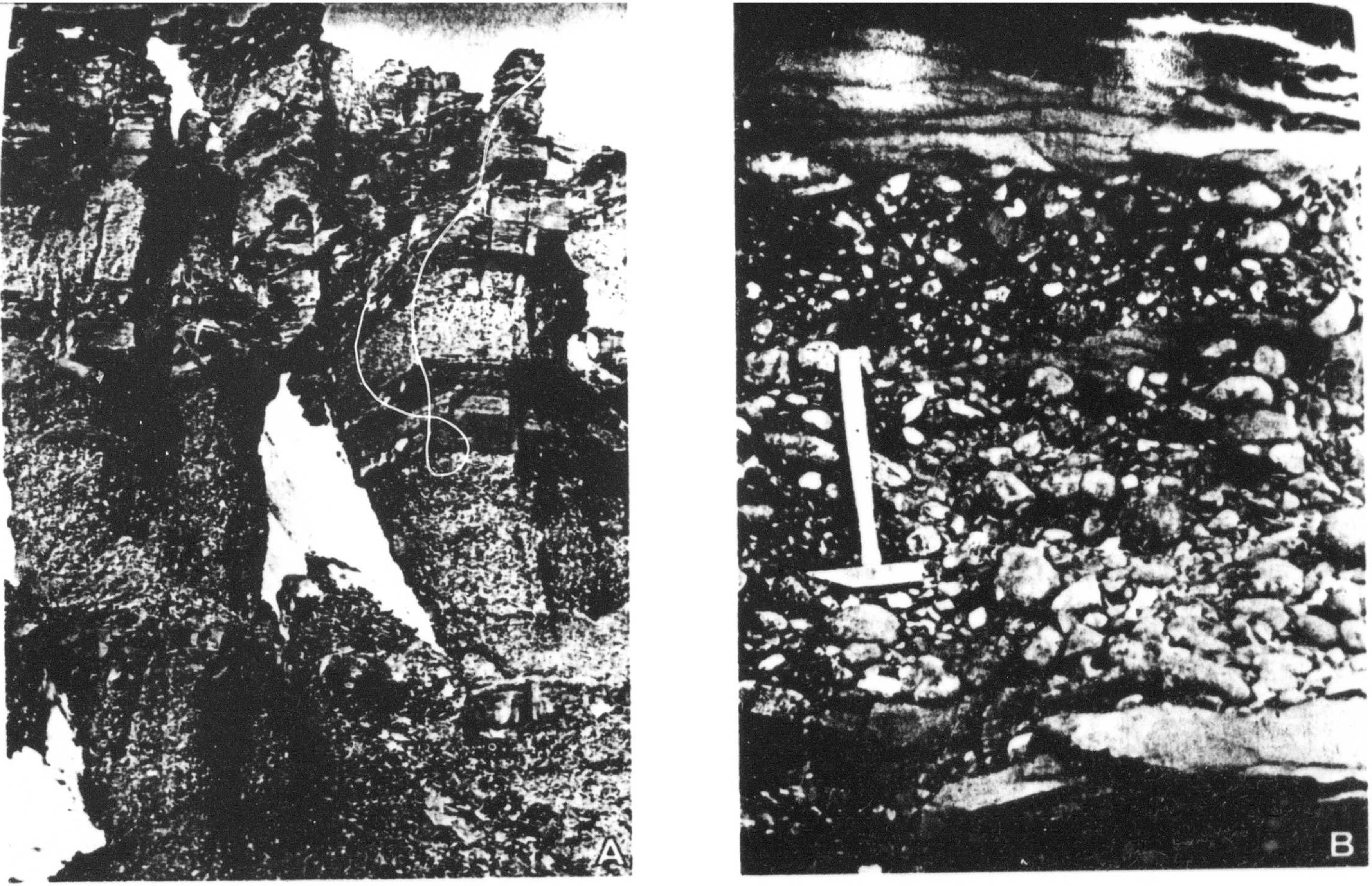


Fig. 5. Braided-stream deposits at the edge of an alluvial-fan system, Peel Sound Formation (Devonian), Canadian Arctic (Miall, 1970a, b). A. Interbedded conglomerates (including debris flood deposits, and bar deposits of Facies Gm), sandstones (Facies St, Sh) and sandy siltstones (Facies Sr, Fl). B. Conglomerate of longitudinal bar origin (Facies Gm) overlain by dune deposits (Facies St). Note the upward decrease in pebble grain size.

TABLE III

Lithofacies and sedimentary structures of modern and ancient braided-stream deposits

Facies identifier (this paper)	Lithofacies	Sedimentary structures	Interpretation
Gm	gravel, massive or crudely bedded, minor sand, silt or clay lenses	ripple marks, crossbeds in sand units, gravel imbrication	longitudinal bars, channel-lag deposits
Gt	gravel, stratified	broad, shallow trough cross-beds imbrication	minor channel fills
Gp	gravel, stratified	planar crossbeds	linguoid bars or deltaic growths from older bar remnants
St	sand, medium to very coarse, may be pebbly	solitary (theta) or grouped (pi) crossbeds	dunes (lower flow regime)
Sp	sand, medium to very coarse, may be pebbly	solitary (alpha) or grouped (omikron) planar crossbeds	linguoid bars, sand waves (upper and lower flow regime)
Sr	sand, very fine to coarse	ripple marks of all types, including climbing ripples	ripples (lower flow regime)
Sh	sand, very fine to very coarse, may be pebbly	horizontal lamination, parting or streaming lineation	planar bed flow (lower and upper flow regime)
Ss	sand, fine to coarse, may be pebbly	broad, shallow scours (including eta-cross-stratification)	minor channels or scour hollows
Fl	sand (very fine), silt, mud, interbedded	ripple marks, undulatory bedding, bioturbation, plant rootlets, caliche	deposits of waning floods, overbank deposits
Fm	mud, silt	rootlets, desiccation cracks	drape deposits formed in pools of standing wa

× = facies present; ? = facies probably present (description not definitive); — = facies no recorded. Other letters and numbers are facies codes assigned by the writer(s) named a the head of the corresponding column.

Modern Deposits							
Doeglas (1962)	McKee et al. (1967)	Coleman (1969)	Williams and Rust (1969)	Collinson (1970)	McGowen and Garner (1970)	N.D. Smith (1970, 1971a, 1972)	McDonald and Banerjee (1971)
x	—	—	F, G	—	x	x	b
x	—	—	F, G	—	—	—	?
?	—	—	?	—	—	—	c
x	—	x	D	x	x	x	c
x	x	x	D	x	x	x	c
x	x	x	C1, D	x	x	x	a
x	x	x	x	x	x	x	x
—	—	?	?	?	?	—	—
x	—	x	B, C1	x	?	?	—
x	—	—	A	—	x	x	—

Facies identifier (this paper)	Modern Deposits						
	G.E. Williams (1971)	Rust (1972)	Picard and High (1973)	Bluck (1974)	Gustavson (1974)	N.D. Smith (1974)	Boothroyd and Ashley (1975)
Gm	×	6	×	×	×	D	×
Gt	—	6?	—	—	—	—	—
Gp	—	—	—	—	×	D	—
St	×	2	×	×	—	C	×
Sp	×	2	×	×	—	C	×
Sr	×	3	×	×	?	B	×
Sh	×	—	×	?	×	C	×
Ss	—	—	?	—	—	—	—
Fl	—	1	?	—	×	B, E	×
Fm	×	1	×	×	×	A	?

Ancient Deposits							
Seiley (1965)	Moody-Stuart (1966)	Bluck (1967)	Kelling (1968)	Nilsen (1968)	Miall (1970a, b)	N.D. Smith (1970)	Augustinus and Riezebos (1971)
–	–	×	A	×	×	×	×
–	–	?	–	–	–	–	–
–	–	–	–	–	–	–	–
×	×	×	B_2	×	×	×	×
×	–	?	?	×	–	×	×
–	×	×	$B_2 B_4$	×	×	×	–
×	×	×	$B_1 B_3$	×	×	×	×
–	–	–	×	×	–	–	–
×	×	×	C	–	×	?	–
–	×	?	C	–	–	×	–

Facies identifier (this paper)	Ancient Deposits						
	McGowen and Groat (1971)	Costello and Walker (1972)	Leeder (1973a)	Eynon and Walker (1974)	Mrakovich and Coogan (1974)	Steel (1974)	Allen (1974a) (Clee Fm
Gm	x	—	3	BC, SB	x	x	A
Gt	—	x	—	SB	—	x	A
Gp	x	x	—	BF, SB	x	—	A
St	x	x	3	SB, SC	x	x	B
Sp	x	x	3	BS	x	x	B
Sr	x	x	—	SB, SC	—	?	—
Sh	x	x	—	SB, SC	x	x	C
Ss	x	—	3	—	—	—	B
Fl	—	x	—	—	—	x	D
Fm	—	x	—	—	—	—	—

Allen (1974a) (Monkeys Fold Fm.)	Conaghan and Jones (1975)	Corner (1975)	Cant and Walker (1976)	Miall (1976a)
–	–	I	–	×
–	–	I	–	–
–	–	–	–	–
A	×	IV	A, B	×
A	×	IV	C, D	×
–	–	III, V, VI	F	×
B	×	II, III, VI	–	×
–	?	–	E	–
–	–	VII, IX	F	–
–	–	–	–	×

than the trough fill. Foreset dip is steeper in the smaller troughs, reachin maximum of about 30°.

Facies Gp: planar-crossbedded gravel. This facies also shows distinct strati cation, comprising individual sets or cosets of planar crossbedding. Set thic ness ranges from 25 cm to 4 m. Individual foreset beds may be as much as cm thick in the larger sets. Reactivation surfaces may be present. These surfaces of erosion which cut across the foresets with the same sense of d but a lower angle than the foresets (Collinson, 1970). As discussed later, tv types of sedimentation unit may be represented by this facies. Facies Gt a Gp may be 1 to 4 m in thickness.

Sandy facies

Included in the sandy facies are deposits ranging from very fine to ve coarse sand; the coarser beds commonly are pebbly. Sorting is extreme variable, and may be in part a reflection of the sorting in the source beds.

Facies St: trough-crossbedded sand. Trough crossbedding in this facies m consist of solitary scoops with an erosional relationship to underlying plan bedded units or other sedimentary structures (theta-cross-stratification Allen, 1963), or cosets of mutually cross-cutting troughs (festoon cro bedding; pi-cross-stratification of Allen, 1963). Set thickness typically rang from 5 to 60 cm; width shows a corresponding range, and may reach 3 Individual troughs may be traceable downcurrent for distances of up to 6 Grain size generally is medium to very coarse sand. Pebbles also may present. The base of the facies commonly consists of an erosion surfa although a thin unit of massive sand with or without intraclasts may occ immediately above the base of the unit. Deposits comprising this facies m range in thickness from that of an individual set up to cosets comprising much as 6 m of beds. Examples of Facies St are shown in Fig. 6.

Facies Sp: planar-crossbedded sand. Grain size and sorting characteristi generally are similar to those of Facies St. Planar-tabular crossbeds are alpha type (Allen, 1963). They have sharp, flat or slightly scoured bases a tops. Set thickness ranges from 5 cm to at least 5 m (typically less than 1 m and as many as ten sets may be superimposed within a single occurrence the facies, giving total thicknesses of up to a recorded maximum of 5 m (F 7). Individual sets may persist laterally and downcurrent for an observ maximum of 20 m, in the case of sets thinner than about 1 m, although co plete exposures are rare and the true maximum may be considerably excess of 20 m. Larger sets may persist downdip and along strike for seve hundred metres. Reactivation surfaces may or may not be present (Fig. Foresets represent avalanche slopes: they are defined by grain-size variatio of one or two ϕ classes, and dip angles commonly are between 15 and 35

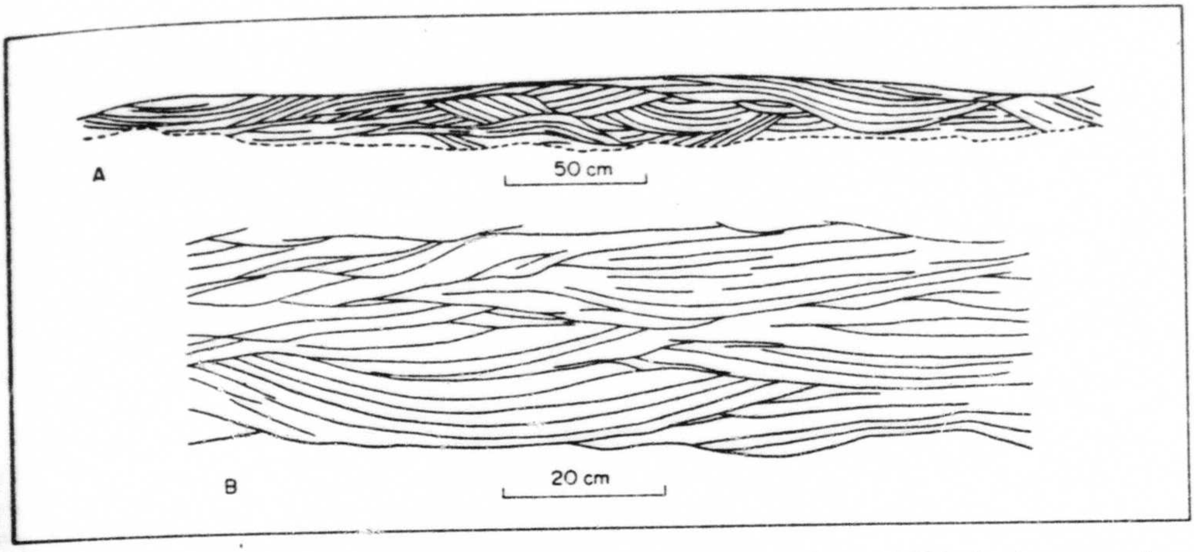


Fig. 6. Large-scale trough cosets (pi-cross-stratification of Allen, 1963). A. An example from a modern ephemeral stream (redrawn from G.E. Williams, 1971). B. An example from a Devonian braided-stream deposit (drawn from a photograph by Cant and Walker, 1976).

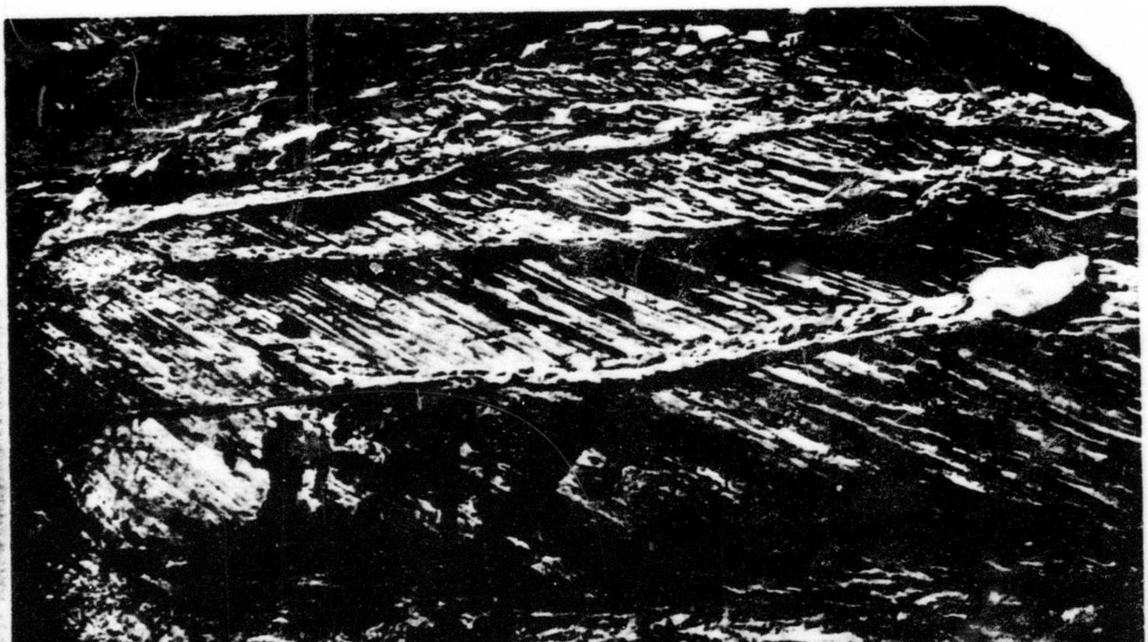


Fig. 7. Superimposed planar crossbeds of Facies Sp, Isachsen Formation (Cretaceous), Canadian Arctic (Miall, 1976a). These sets show mostly straight, rather than curved foresets, suggesting formation by currents with a weak separation eddy. Reactivation surfaces are absent. Shovel handle is 50 cm in length.

In these respects the crossbeds are quite different from epsilon cross-strati cation (Allen, 1963) which show much lower dip angles (generally less tha 15°) and comprise a variety of foreset lithologies, commonly including si or clay. This distinction is of considerable genetic significance, as is discusse in a later section.

Facies Sr: ripple cross-laminated sand. A variety of asymmetric ripple typ characterize Facies Sr. Ripple amplitude is less than 5 cm. Sand grain siz ranges from coarse to very fine, but medium sand is most typical. A wic variety of internal structures can be generated from ripple migratio depending on flow velocity and the rate of sediment supply (Allen, 196 Jopling and Walker, 1968); they include solitary ripple trains, small-sca trough and planar crossbedding and climbing ripples. Any of these types ma be present in Facies Sr. Individual occurrences of the facies range from a fe centimetres to a few metres.

Facies Sh: horizontally-bedded sand. The sand may be laminated to massiv very fine to very coarse-grained or (rarely) pebbly. Facies thickness rang from a few centimetres to several metres. Parting lineation may be we developed (Fig. 9) and very small-scale ripple marks with amplitudes of le than 5 mm may be present, particularly in the finer-grained sands.

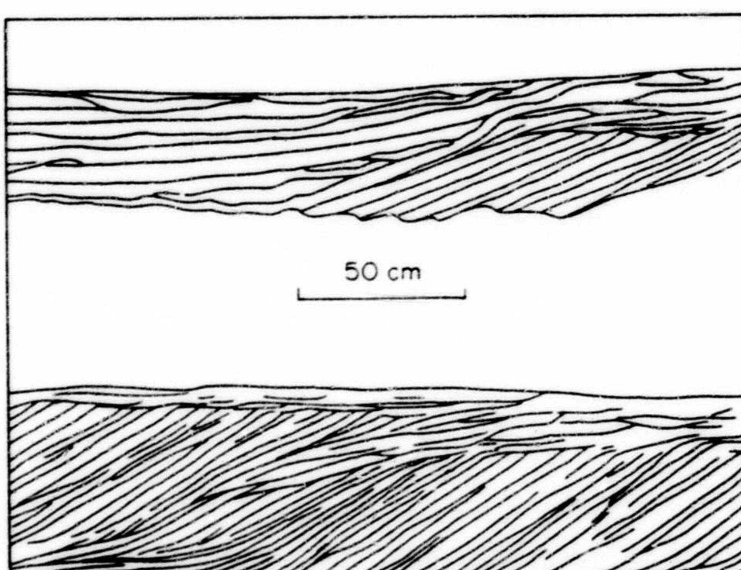


Fig. 8. Planar crossbed sets (alpha-cross-stratification of Allen, 1963) showing reactivati surfaces. Drawn from photographs by Collinson (1970).

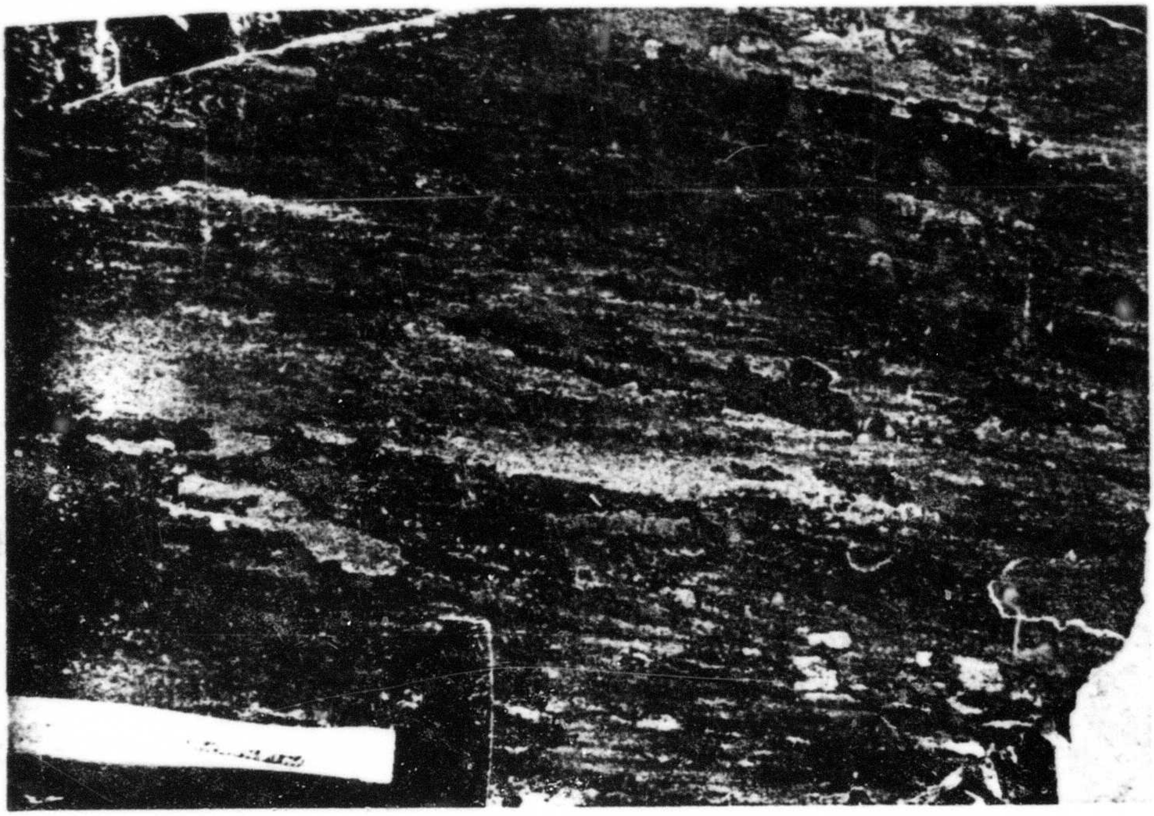


Fig. 9. Sandstone with parting lineation, Glenelg Formation (Proterozoic), Canadian Arctic (Young and Jefferson, 1975; Miall, 1976b). Facies Sh.

Facies Ss: scour-fill sand. This facies consists of large, asymmetric scours and scour fillings typically up to 45 cm deep and 3 m wide, although deeper channels (up to 5 m) have been recorded. They may be traceable downcurrent for several tens of metres. The scour fill comprises fine to coarse, commonly pebbly sand, and the stratification is at a very low angle to the basal erosion surface so that in exposures of limited lateral extent the beds appear to conform to the shape of the scour. Planar beds with parting lineation, trough crossbedding and small-scale cross-stratification may be present. The smaller scours conform in type to eta-cross-stratification of Allen (1963). The cutting and the filling of the structures were probably separate events.

Other sand facies. The five sand facies described above are the most common in modern and ancient braided-stream deposits. Several others remain to be described that have been recorded at few localities and do not constitute a major proportion of any given braided-stream deposit. They are not shown in Table III, but their origin is discussed briefly below.

Deposits showing low-amplitude ripples or radiating, plumose structures have been described by Williams and Rust (1969, pp. 667—668), (Rust (1972, pp. 230—232) and Collinson (1970, pp. 46—47) and have been ascribed to wind action. McGowen and Garner (1970, Fig. 6) and Singh and Kumar (1974, Fig. 5) illustrate deposition of windblown sand on or in the lee of bars. Such deposits are likely to be of minor importance in the sedimentary record.

Thixotropic silty sands (quicksands) have been described by Williams and Rust (1969, p. 664). Sand or silt volcanoes may be common and the upper surface of the deposit may be covered by small-scale ripple marks. Thixotropy probably is the cause of many of the convoluted structures observed by G.E. Williams (1971, p. 35), McKee et al. (1967), and Coleman (1969, p. 217).

Planar laminated sand (a variety of Facies Sh) showing strong size differentiation between the units, or small-scale grading, was observed in the Lower Platte River by N.D. Smith (1971b) and has been shown to be the product of migration of very low amplitude sand waves in shallow water (less than 1 cm). The sorting into discrete laminae takes place in the migrating wave foresets, which are normally less than 1 cm in amplitude and do not display internal cross-lamination.

Low-angle (<10°) cross-stratified sandstones have been described by Cant and Walker (1976, p. 110). Individual sets are 30 to 90 cm in thickness; as many as eight sets may comprise a coset. The sand is fine-grained. The sets resemble broad, shallow troughs and rest on flat or shallow, scoured surfaces. The origin of this facies is obscure, although a comparison with the low amplitude bedforms of Banks (1973a) is tentatively proposed.

Fine-grained facies

Silt and clay may comprise a very small percentage of a braided-stream deposit, but their presence is of importance because of the genetic implications that can be deduced from their position in the bed sequence.

Facies Fl: laminated sand, silt and mud. Sand in this facies is rarely coarser than very fine grained. Interbedding of sand, silt and mud on a fine scale is common. Very small-scale ripple marks, undulating bedding, bioturbation, plant rootlets and coal may all be abundant. Caliche nodules and other pedogenic features rarely are present, depending on climate. The thickness and lateral extent of this facies is variable. Typical thicknesses are a few centimetres to a few decimetres. Data on lateral extent are sparse.

Facies Fm: mud or silt drape. Dark-coloured, massive or laminated mud or silt occurs as lenticles a few millimetres to a few centimetres in thickness. Carbonaceous streaks, plant rootlets and desiccation cracks are common features. This facies occurs typically as a drape over underlying beds, its lower surface conforming to the shape of any underlying bedform, such as a ripple train.

DEPOSITIONAL FACIES PART 2: INTERPRETATION

Each lithofacies can be interpreted in terms of its hydrodynamic origin and its position within the braided-river tract morphology. Interpretations are soundly based on observations in modern rivers in the case of the finer-grained lithofacies and smaller-scale bedforms, but many of the larger-scale structures and coarser deposits are formed in deep channels or during peak flood stages when direct observations are difficult or impossible and this may have introduced a bias into our interpretation of braided-river deposits.

The principal depositional facies of braided streams can be interpreted in terms of five main processes:

(1) Longitudinal bar formation.

(2) Bedform generation and migration (including linguoid and transverse bars, dunes, megaripples, etc.).

(3) Channel scour and fill.

(4) Low-water accretion and modification processes.

(5) Sedimentation in overbank areas.

Longitudinal bar formation (facies Gm)

The process of formation of longitudinal bars was summarized earlier in this paper. It consists of a clast-by-clast accretion over an obstruction or a channel lag deposit. The mechanism appears to occur only in pebbly braided rivers; in sandy rivers the bedload forms discrete bedforms such as linguoid bars, dunes or ripples, with quite different internal structures.

Longitudinal bars are of low amplitude, and this accounts for the low depositional dips observed in Facies Gm. The trapping of finer particles in the interstices between larger clasts gives rise to a clast-supported framework, and the process is not conducive to the development of well-defined stratification. Such stratification as there is probably relates to flood-stage discharge fluctuations.

Boothroyd and Ashley (1975, Fig. 5) provided measurements of flow depth, velocity and stream power for two glacial outwash streams taken during a falling flood stage (Table IV). They used the relationship:

$$\tau_0 = \rho g D S$$

to express shear stress, where ρ = fluid density (g/ml), g = acceleration due to gravity (cm/sec^2), D = flow depth (cm), and S = slope (expressed as a dimensionless number). Stream power then equals $\tau_0 V$, where V = mean velocity (cm/sec).

Stream power has been calculated for four other pebbly rivers from data provided by the authors (Table IV). Boothroyd and Ashley (1975, p. 219) state that observations and measurements indicate that stream power is greatest in deep channels. Coarse clasts can thus continue to be transported until the flow passes over a bar surface. At this point competency will be

TABLE IV

Flow characteristics of some modern braided-stream deposits

Locality	Principal facies	d_b (m)	S	V (m/sec)	$\tau_0 V$ (ergs/cm^2 per sec)	Author
Alaskan outwash streams (proximal)	Gm	0.2–1.0	0.005–0.018	1–3	30 000–400 000	Boothroyd and Ashley (1975)
Donjek *	Gm, Sp	3.0	0.0006	3.6	63 500	Rust (1972)
Kicking Horse	Gm, Gp, St, Sp	1.2	0.0034–0.0072	2	127 000	N.D. Smith (1974)
Durance *	Gm, Gt	6.6	0.0026	6	1 000 000	Doeglas (1962)
Slims	Gm	5.0	0.0028	>2	$>275\,000$	Fahnestock (1969)
Lake Eyre Basin *	St	0.8–2.5	0.00095–0.0024	0.8–2.8	6 000–165 000	G.E. Williams (1971)
Lake Eyre Basin*	Sh	2.8	0.00128	3.6	126 500	G.E. Williams (1971)
Brahmaputra *	Sp, St	15	0.00007	2.4	24 700	Coleman (1969)
Alaskan outwash streams (distal)	Sp, Sr	0.3	0.002	0.5	2 900	Boothroyd and Ashley (1975)
Canal experiment	Sp?	1.2	0.00053	1.4	9 150	Culbertson and Scott (1970)
Canal experiment	St?	1.9	0.00053	0.8	8 460	Culbertson and Scott (1970)
Lower Platte	St	0.2–0.5	0.0015	0.4–0.6	1 200–4 000	N.D. Smith (1970, 1971a)
Lower Platte	Sr	0.2–0.4	0.0015	0.2–0.4	150–500	N.D. Smith (1970, 1971a)

S = slope; $\tau_0 V$ = stream power; other symbols as in Table III. * Data pertain to flood conditions.

reduced and part of the coarsest bedload will be deposited. Hein and Walker (1977) record that in the Kicking Horse River bed load with a mean grain size of $-5.6\ \phi$ was moved only when the discharge exceeded 57 m^3/sec.

Interbedded sand or silt lenticles are the result of a low-water accretion process, as described later in this section.

Bedform generation and migration (Facies Gp, St, Sp, Sh)

Flume experiments and observations on modern rivers have shown that a wide variety of bedforms can be formed on non-cohesive sand beds, depending on grain size, flow depth and velocity and rate of sediment supply. These range in size from giant sand waves with amplitudes of up to 15 m (Coleman, 1969), to very small scale ripple marks with amplitudes of a few millimetres. During flood stage active bedforms of a variety of scales may be superimposed on each other, as shown by Neill (1969) and Singh and Kumar (1974). A variety of internal stratification types is produced by the migration of bedforms, some of which are common in ancient braided-stream deposits. The smaller-scale bedforms, those with amplitudes of less than 5 cm, are formed during periods of diminished flow and are discussed under the heading "low-water accretion processes" (pp. 35—37).

A certain nomenclature confusion exists in the literature regarding bedforms with amplitudes greater than 5 cm. The large-scale ripples of G.E. Williams (1971, p. 8), the dunes of Collinson (1970, p. 42) and N.D. Smith (1971a) and the megaripples of Karcz (1972) and Singh and Kumar (1974) appear to be virtually identical in morphology and internal stratification. They range in amplitude from 7 to 40 or, rarely, 70 cm and in wavelength (chord length) from 1 to 20 m. They generally are lobate in plan and commonly have scour hollows in their lees. In all cases the internal structure is trough crossbedding, either solitary (theta-cross-stratification of Allen, 1963) or grouped (festoon, or pi-cross-stratification), corresponding to sand Facies St of this paper. Sedimentation in straight channels in which bars are of minor importance may be dominated by dune deposits under the right hydrodynamic conditions, as tentatively suggested by Moody-Stuart (1966) and Leeder (1973a; his Facies 2). The "non-cyclic deposits" of Allen and Friend (1968, p. 56) also appear to be of this type. The sand waves described by Sundborg (1956, Fig. 45) and Harms et al. (1975, pp. 24, 47—49) are similar in amplitude to dunes (20—50 cm) but have straighter crests. They give rise to planar-tabular crossbedding.

No exact parallels appear to exist in the Brahmaputra. Coleman (1969, pp. 187—190) described lunate megaripples, which have heights of 30 to 150 cm and wavelengths of 30 to 150 m. The larger megaripples of the Brahmaputra are similar to the giant ripples of Singh and Kumar (1974) and may be of similar hydrodynamic origin to the linguoid (lobate, transverse) bars of Collinson (1970), G.E. Williams (1971), N.D. Smith (1971a, 1972) and Boothroyd and Ashley (1975), the geometry and scale of which were discussed earlier in this paper. Extensive trenching has shown that planar-tabula crossbedding of Facies Sp (alpha- and omikron-cross-stratification of Allen 1963) is the dominant internal structure (Collinson, 1970, Figs. 19—20; G.E Williams, 1971, Fig. 12; N.D. Smith, 1972, Figs. 3, 7; 1974, Fig. 16B). In th high water stage the foresets will have asymptotic (sigmoidal) shapes becaus of the presence of a strong separation eddy (Collinson, 1970, Fig. 27) Gravel Facies Gp (rare in braided-stream environments) also is generated by the migration of linguoid bars during flood stage but, as noted earlier, flood conditions or a confined channel, such as an alluvial fan-head trench, may b necessary (this is an area of hydrodynamics that needs further study).

An alternative origin for Facies Gp was discusssed by Eynon and Walke (1974) but it probably represents a special case. They described planar cross bedded gravels in cosets reaching 4 m in thickness and cut by numerous rea tivation surfaces. The beds were interpreted as the product of delta-lik growth from an eroded bar remnant in a relatively deep outwash channel.

Very large-scale crossbedding has been recorded in a few ancient braided stream deposits, such as the planar and trough crossbed sets 5 m or more i thickness in the Hawkesbury Sandstone near Sydney (Conaghan and Jones 1975). They undoubtedly represent deposition by the migration of th largest bedforms present in major river channels, such as the dunes and sand waves in the Brahmaputra. Dunes (in the terminology of Coleman, 1969, p 190) are 1.5 to 8 m in height; sand waves are defined as those bedform exceeding 8 m in amplitude, the largest recorded being 15 m high. Super imposed bedforms of different scales are described by Banks (1973a), but d not appear to be common in ancient braided-stream deposits.

Horizontal bedding or lamination (Facies Sh) can occur under two quit different conditions: in shallow water and during flood stage. As shown by Harms and Fahnestock (1965) a plane-bed condition is a specific respons to certain combinations of grain size and flow velocity, and thus is a "bed form" in the same sense as a dune or a ripple. During flood stage a plane bed condition develops during the upper flow regime (Harms and Fahne stock, 1965) when the channel floor becomes a traction carpet, with virtu ally continuous particle movement. Beds formed by this process may show poor sorting, strong or faint internal lamination, and parting or streamin lineation (McKee et al., 1967; G.E. Williams, 1971; Picard and High, 1973) Harms et al. (1975, Figs. 2—5) showed that a plane-bed condition also devel ops in sand coarser than 0.6 mm in lower flow regime conditions.

The bedforms present in Facies Gp, St, Sp, and Sh have, in most cases been reproduced in flume experiments, and can thus readily be interpreted in terms of stream flow velocity and stream power (Guy et al., 1966; G.P Williams, 1967; Allen, 1970, Fig. 13; Friend and Moody-Stuart, 1972, Fig 27). Rather than repeat this information, data regarding the hydrodynamic conditions of sedimentation based on measurements in modern braided streams are given in Table IV. The information is generalized, and there i some doubt as to whether the principal facies necessarily were deposited under the conditions specified. Where the data pertain to flood stage, the coarsest deposits and the largest bedforms probably were deposited under the given conditions. In other instances the data relate to lower water stages when smaller bedforms were being generated, and the flood deposits, such as gravel lenses and linguoid bars, were undergoing only minor modification. Stream power during the generation of the major sand bedforms is in the approximate range of 3000—160,000 ergs/cm^2 per sec.

Channel scour and fill (Facies Gt, Ss)

Shallow channels filled with sand or gravel have been described by Nilsen (1968, pp. 61—62), McGowen and Groat (1971), Costello and Walker (1972, p. 393), Steel (1974), Church and Gilbert (1975, Fig. 44) and Cant and Walker (1976, p. 109). Some of the smaller channels probably represent scour hollows such as have been observed in the Rio Grande (Harms and Fahnestock, 1965, Plate 2), the Brahmaputra (Coleman, 1969, Fig. 26c), the Donjek (Williams and Rust, 1969, p. 661) and the Amite (McGowen and Garner, 1970, p. 82). Scouring may be related to strong separation eddies in the lee of advancing dunes or sand waves (generation of Facies St, and excluded from this category of scour structure) or to local vortices developed around obstructions. The larger scours represent major and minor channels, formed by avulsion during high water stage or bar dissection during falling water conditions, as described by Collinson (1970, p. 43) and Williams and Rust (1969, p. 660). The larger channels may have composite fills that are capable of classification under one or more of the other coarse-grained facies described herein (such as those described by Steel, 1974, and McGowen and Groat, 1971), and should not be classed as Facies Gt or Ss.

In contrast to trough or festoon crossbeds, the erosion of the scour hollow and its subsequent infill do not represent virtually simultaneous events, as occurs during bedform migration. This is suggested by the fact that minor sedimentary structures, including parting lineation and large- or small-scale cross-lamination may be present within the channel fill, and would appear to be confirmed by the fact that such hollows persist long enough to be observed during low water stage in modern braided rivers. Facies Gt and Ss include eta-cross-stratification (Allen, 1963) in which the stratification of the fill is parallel to the lower bounding surface, which suggests formation under upper or lower flow regime, plane-bed conditions. Other types of channel fill formed exclusively during the lower flow regime, are discussed in the next section.

Low-water accretion processes (Facies Sr, Sh, Fl, Fm)

A variety of sub-facies develop during low water; most are characterized by small-scale cross-stratification in predominantly medium or finer-grained sand:

(1) Infill of minor channels and scour hollows, particularly on bar surfaces.

(2) Ripple and dune accretion on bar surfaces.

(3) Development of reactivation surfaces.

(4) Bar-front accretion.

(5) Drape deposits formed in pools of standing water.

As discussed in the previous section, some scour hollows fill with sediment during high-stage flow, forming eta-cross-stratification (Allen, 1963). Others are filled during waning floods or low water, generating lenticular sand or silt bodies characterized by medium or small-scale structures such as trough cosets or ripple marks (Facies Sr). Gravel beds of Facies Gm commonly contain thin beds of this type, as recorded in modern streams by Rust (1972, p. 229), and Boothroyd and Ashley (1975, Fig. 14A, B), and in ancient deposits by Miall (1970a, p. 128; 1970b, p. 560).

Ripples and dunes continue to migrate across bar surfaces during low water, even at extremely shallow depths. In so doing they form small-scale cross-stratification on the bar surface (Facies Sr) and the bedform migration results in a constant supply of sediment to the bar front, so that parts of the bar-front avalanche slope will continue to prograde. The process was fully described by N.D. Smith (1971a, 1972). Under water depths of less than a centimetre and flow velocities of less than 30 cm/sec, very low amplitude sand waves develop, resulting in horizontal lamination (Facies Sh) commonly characterized by fine grading (N.D. Smith, 1971b).

Bar-front accretion processes during low water stages have been described by Collinson (1970). Sheet runoff may generate small bar-top channels, and bar-front avalanche slopes develop straight foresets with angular relationships to the lower bounding surface as a result of the weakened separation eddy (Collinson, 1970, Fig. 27). If the bar is exposed, the bar front may be smoothed over as the result of sheet runoff. The erosion surface becomes a surface of reactivation during rising water stages. Small-scale ripples may accrete on it at first (Facies Sr) but at higher water levels avalanching and foreset progradation will be resumed. Small delta-like, lobate accumulations may build out from bar-top channels (Collinson, 1970, Fig. 11, 23; Singh and Kumar, 1974).

Bar-front sand wedges showing trough or planar crossbedding commonly develop at the margins of gravel bars during falling water stages (Rust, 1972, Fig. 4; Boothroyd and Ashley, 1975, Fig. 17A). The sand wedge may itself be dissected and/or covered by ripples as the bar becomes exposed, resulting in a complex internal structure.

At the lowest water levels pools of standing water will be left in abandoned channels, particularly on topographic levels 2 and 3 (see Fig. 2). Fine silt and mud will settle out, forming lenticular drape deposits, and when these in turn dry out they develop desiccation cracks (Facies Fm). The preservation potential of such beds is low (Cant, 1976), but they have been recorded in several ancient braided-stream deposits (Table III).

Stream power during low-water accretion is less than 4000 ergs/cm^2 per sec (Table IV).

Sedimentation in overbank areas (Facies Fl)

Braided streams are not characterized by large areas of floodplain, but many have abandoned areas with a sparse to thick vegetation cover, which are covered by water at only the highest flood stages. Flow velocity will tend to be slow in such areas because of the shallow depths and the friction induced by vegetation. Trapping of fine sediment is thus encouraged, small or very small-scale current structures are generated and the beds commonly are disturbed by bioturbation or root growth shortly after deposition (Doeglas, 1962; Williams and Rust, 1969; Stanley and Fagerstrom, 1974). Boothroyd and Ashley (1975, Fig. 25) showed that such deposits are commonest in the more distal parts of a braided-stream system. Kelling (1968) described overbank deposits which contain underclay and coal; Steel (1974) and Allen (1974a) recorded the occurrence of caliche nodules.

VERTICAL PROFILES AND PROXIMAL—DISTAL VARIATIONS

Many of the facies described in the previous sections are typically but not exclusively present in braided-river environments. The key to a positive identification of the environment thus lies in the nature of the facies assemblage, including the vertical profile and lateral lithologic variability. It is necessary to combine two approaches in an investigation of an overall depositional model: a study of the evolutionary processes taking place within modern braided rivers, and an investigation of facies assemblages and sequences in ancient rocks. Both approaches suffer from disadvantages: the problem of preservability hangs over any study of modern rivers, because it cannot be known with much certainty what fragments of the most recent deposits in the river will survive erosion long enough to enter the geological record. According to Cant (1976) the smaller the bedform the lower its preservation potential. Studies of ancient braided deposits provide only limited assistance in attacking the problem of model building, because at present there are so few of them containing the necessary level of detail. The variability in braided-river sediments calls for rigorous analysis, preferably using statistical techniques, but only a handful of such studies have been published at the present day. Picard and High (1973) studied the distribution of sedimentary structures in different parts of modern braided environments using factor analysis, but many of the structures they used were of low preservation potential, and the approach seems to be of limited usefulness.

Vertical profiles

Several types of repetitive vertical sequence, or cycle, can be envisaged in a braided-river environment:

(1) A flood cycle: a superimposition of beds formed at progressivel decreasing energy levels.

(2) A cycle due to lateral accretion: a cycle generated by side- or point-ba growth is possible, as in a meandering-river environment.

(3) A cycle due to channel aggradation: this cycle would represent the fi of a channel or a local channel system. Waning energy levels would occu during sedimentation, followed by channel abandonment as a result of avul sion.

(4) A cycle due to channel re-occupation: an abandoned, partially fille channel may be re-occupied by avulsion.

It is, of course, quite possible that in a given braided-stream deposit all o these cycle types might be present, rendering interpretation extremely diff cult. Another difficulty is the scale of thickness to expect from each cycl mode. The examples discussed below range in thickness from 15 cm to 60 m Lateral persistence of each bedding unit also is variable, although most bed can be traced for only a few metres or tens of metres. All interpretation must, therefore, be based on extremely careful field work and detailed loca analysis.

The best available technique for studying vertical cyclicity is Markov chai analysis. The technique was described fully by Miall (1973) and its mos recent application to braided-stream deposits was by Cant and Walke (1976). The analytical technique is, briefly, as follows. The lithologies in vertical sequence are classified into a limited number of facies states (fou to six states has been found to be the ideal number) and the number of verti cal transitions between each of the states is counted and tabulated. Th probability of occurrence of each transition can then be calculated, and can be demonstrated which of the transitions have occurred with greate than random frequency. The end-product of the analysis is a tree diagram showing the principal vertical facies relationships; such diagrams will be use extensively in the discussions below.

Modern deposits

Sequence studies in modern braided-stream deposits are limited by the dependency on trenching. The "outcrops" generally are few, as a result, an are of very short vertical extent. In addition, the sample of beds they expos is biased in favour of higher, drier parts of the river tract. Observations o the facies occurring in the major channels are very limited, and this may b why so few trough crossbeds (that originated as dunes migrating down chan nel floors) were observed by Williams and Rust (1969) and Boothroyd an Ashley (1975). Complete, statistically rigorous Markov chain analysis can not, therefore, be carried out. Fig. 10 shows some tree diagrams that have been constructed from published descriptions or illustrations of vertica sequences in various modern braided-stream deposits. They may, perhaps be subjective, but they serve to illustrate what appear to be some importan trends.

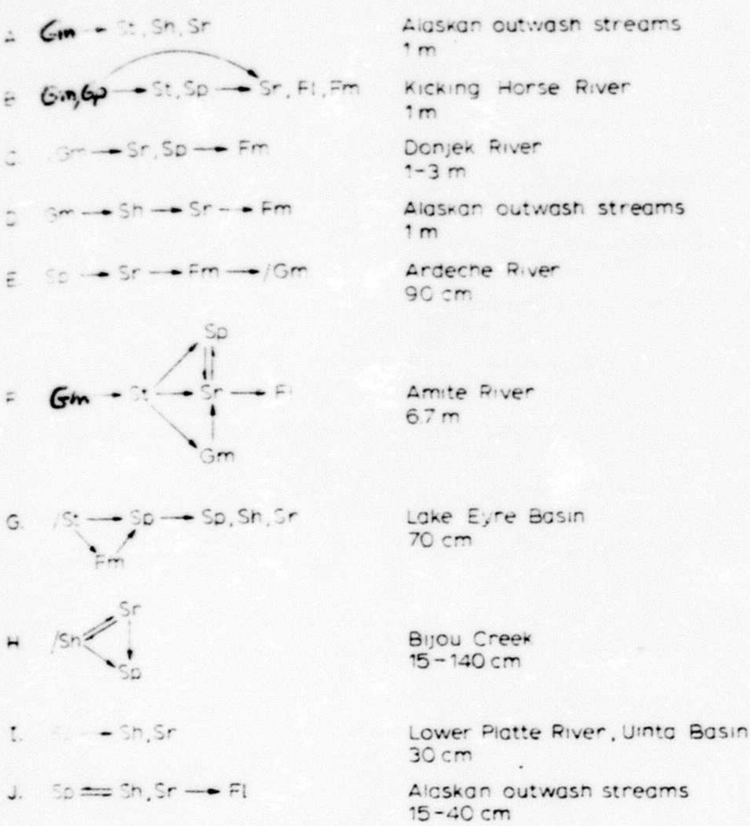


Fig. 10. Some vertical sequences observed in modern braided-stream deposits. Sequence thickness (approximate) is given at right. / = major scour surface; → = upward bed transition; facies separated by commas are alternatives in the sequence.

Fig. 10A (derived from Boothroyd and Ashley, 1975, p. 203) illustrates a typical sequence from a river dominated by pebble- or cobble-gravels. It represents an alternation between flood deposition and low-water accretion; the latter includes bar-edge sand wedges and the sand fill of minor channels and scour hollows. Where lateral accretion occurs, as in channel bends, a crude fining-upward cycle may occur in pebbly rivers, as shown by Figs. 10B (N.D. Smith, 1974, pp. 220—221), 10C (Williams and Rust, 1969, Fig. 21) and 10D (Boothroyd and Ashley, 1975, pp. 208—209). Alternatively these sequences may represent the cycles formed by waning floods, as in the very similar channel-fill sequences reported by Picard and High (1973, pp. 198—201). In either case the cycle thickness will be approximately the same as the channel depth during flood discharge although, where the river tract comprises several distinct topographic levels, this may be less than the river bank-full depth. Each cycle probably is floored by an erosion surface, although this has not been observed in all of the examples quoted. Fig. 10E shows a partial cycle observed in the Ardeche River by Doeglas (1962, p. 185, Fig. 29).

A composite point-bar sequence is given in Fig. 10F, based on the work by McGowen and Garner (1970, Fig. 4). It was derived from a meandering river, but the river transports a high proportion of its sediments as bedload and, therefore, the sequence is included for comparative purposes. Immediately above the basal scour surface is a thin gravel lag. The other occurrence of Facies Gm represents lag deposits in chutes, which are situated high on the point bar and are occupied only during floods. The relative preservation potentials of the various facies are difficult to assess, but similar sequences to that shown have been observed in ancient rocks interpreted to be of point-bar origin, as discussed in the next section.

The remaining tree diagrams (Figs. 10G—J) are derived from studies of sandy rivers. The Lake Eyre Basin (G.E. Williams, 1971, pp. 18, 29), and the Bijou Creek (McKee et al., 1967) sequences are flood cycles formed in ephemeral streams. Each represents the deposits of a single, recent, catastrophic flood. Locally, in Bijou Creek, the deposits attain a thickness of 3.6 m, but a range of 15 to 140 cm is more typical. The origin of the thin veneer of Facies Fm in the Lake Eyre Basin floods is obscure. Prograding bars are described by G.E. Williams (1971, p. 18) as having advanced over older bedforms (predominantly large-scale ripples or dunes), which themselves were veneered with silt and clay. It is unclear whether the ripples and their veneer represent, in fact, the deposits of a preceding flood cycle. Without Facies Fm the sequence is that of a waning flood: scour surface, dunes migrating over the channel floor, advancing linguoid bars, and low-water accretion deposits. The Bijou Creek deposits are noteworthy for the abundance of planar lamination formed under upper flow regime conditions. The flood discharge and velocity in this creek were exceptionally high (see Table II), which may be more characteristic of ephemeral than perennial streams. Horizontal lamination of upper flow regime origin, also is common in parts of the Lake Eyre Basin (G.E. Williams, 1971, p. 32).

The sequence from the Unita Basin (Fig. 10I) (Picard and High, 1973, pp. 194—197) also is from ephemeral streams. It is virtually identical in both thickness and facies succession to that from the Lower Platte River (N.D. Smith, 1970, p. 2999), which is a perennial stream. In both cases the sequence is related to linguoid-bar progradation followed by low-water accretion. Similar successions also have been observed in distal parts of some outwash fans (Fig. 10J) (Boothroyd and Ashley, 1975, p. 213), with the additional presence of a fine-grained overbank deposit capping the sequence.

Ancient deposits

Some braided-stream profiles derived from studies of ancient rocks are given in Fig. 11; a few are based on Markov chain analysis of measured stratigraphic sections.

Figs. 11A and B represent sediments dominated by gravel. The Peel Sound Formation profile is an alluvial fan succession, of Devonian age (Miall, 1970b, p. 560), and the Lake Louise profile a Pleistocene glacial outwash deposit (McDonald and Banerjee, 1971, pp. 1298—1299). The similarities between these two tree diagrams and that of Fig. 10A are marked, except

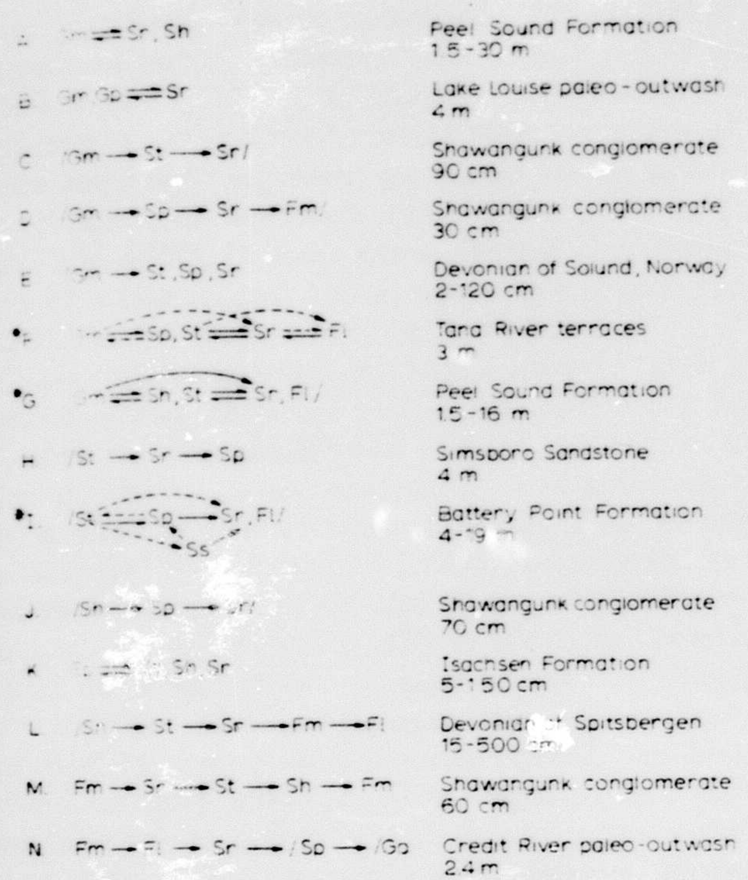


Fig. 11. Vertical sequences observed in some ancient braided-stream deposits. Those marked with an asterisk are based on Markov chain analysis; major transitions are shown by solid arrows, less frequent transitions are shown by dashed arrows. Other symbols as in Fig. 10.

that the cycles in the ancient rocks are considerably thicker. This point will be returned to later. The next five tree diagrams (Figs. 11C—G) are derived from sediments in which gravel and sand are of approximately equal importance. The Shawangunk conglomerate is a Lower Silurian rock unit outcropping in the north-central Appalachian Mountains (N.D. Smith, 1970, Fig. 17), the Devonian rocks of Norway represent alluvial-fan complexes deposited in intermontane basins (Nilsen, 1968, Fig. 41), and the Tana River terraces are post-glacial sediments in the Finnmark area of northern Norway (Corner, 1975, p. 249). The second Peel Sound profile (Fig. 11G) is derived from the distal parts of subaerial alluvial fans (Miall, 1970a, pp. 129—130; 1973, Table 6). An upward decrease in grain size and an upward diminution in the size of sedimentary structures are noteworthy features of all these cycles, although in both diagrams 11F and G Markov analysis indicates a marked tendency for an alternation within the cycles between sediments with large- and small-scale sedimentary structures (Facies Sp, St and Facies Sr, Fl).

Diagram 11H is interpreted as a point-bar sequence (of Eocene age) by

McGowen and Garner (1970, p. 107, Fig. 24), and by comparison with th modern point-bar sequence of Fig. 10F may represent a good example o how several facies are lost in the geological record as a result of low preserva tion potential. However, as pointed out earlier in this paper, there is no un equivocal way of recognizing a point-bar sequence in a braided-stream deposit and diagram 11H is similar in some respects to diagrams 11I and K which represent channel-fill and linguoid-bar sequences.

Diagrams 11I—L illustrate typical profiles in sandy braided rivers. Th Battery Point succession (Devonian of Quebec; Cant and Walker, 1976 Table 4) is the only one in which some statistical evidence for the position i the succession of Facies Ss is available. Scour-and-fill structures (includin minor channels) occur above channel-floor dune deposits, and are followe either by bar deposits or by the products of low-water sedimentation. Simi lar cycles are described by Kelling (1968) and Campbell (1976). The Isachse (Fig. 11K) is a Cretaceous unit (Miall, 1976a); the vertical profile is simila to that occurring in the modern Platte River and the ephemeral streams o the Uinta Basin (Fig. 10I), and consists almost entirely of planar crossbed formed by avalanching transverse or linguoid bars, alternating with bar-to sediments or, rarely, channel-bottom dune deposits. A very similar succes sion in a Pennsylvanian deposit has been described by Mrakovich and Cooga (1974).

The sequence shown in Fig. 11L was interpreted by Moody-Stuart (1966 as the product of a low-sinuosity stream in which bars and islands were o minor importance (low braiding index, in the terminology of Brice, 1964) The absence of Facies Sp and the abundance of Facies St, confirms thi interpretation. Facies Fm fills the entire width of a channel and represent the deposits which settle out after channel abandonment. Facies Fl repre sents flood-plain deposits. Strictly speaking, the sequence probably represent the deposits of a straight rather than a braided river, as defined in Table I.

The remaining diagrams (Fig. 11M, N) illustrate coarsening-upwar sequences. The Credit River succession was obtained in a post-glacial deposi in Ontario (Costello and Walker, 1972, Fig. 8). Both sequences indicate a increase in flow energy with time. Successions similar to that shown in Fig 10H for the Bijou Creek flood deposit are not given in Fig. 11, but man ancient rock sequences which show a predominance of Facies Sh with mino interbedded units of Facies Sp and Sr probably are of the same origin. Exam ples include parts of the Sandstone facies of the Peel Sound Formatio (Miall, 1970a, p. 134), and the Doko sandstone of Nigeria (Adeleye, 1974).

Discussion

In spite of the problem of preservability, referred to earlier, many similar ties exist between the modern and ancient vertical profiles. The main diffe ence is one of thickness; only small-scale cycles have been observed in mos modern deposits, whereas the more vertically extensive outcrops in som ancient deposits have enabled larger-scale cycles to be recorded in severa cases. Small-scale cycles commonly are contained within them, as in the alternations between Facies St and Sr in Fig. 11F and G, and between Facies St and Sp in Fig. 11I. A thorough statistical (Markov) evaluation of the other sequences would no doubt reveal other examples of such cycle nesting. Most of the sequences represent processes taking place *within* channels, such as lateral point-bar accretion (Fig. 10F, 11H) or linguoid-bar progradation (Figs. 10I, 11K). In some cases channel abandonment leads to the development of a fine-grained facies, either at the end of a flood (Fig. 10C), or as a result of lateral migration (Figs. 10E, 11L).

These processes may be inadequate to explain the thicker cycles, as pointed out by Schumm (1972, pp. 104–107). As shown in Tables II and IV, only the extremely large braided rivers, such as the Brahmaputra, have flood depths in excess of about 7 m (braided channels in the Brahmaputra have a maximum scour depth of about 25 m; meandering reaches commonly are deeper). Therefore, to produce some of the thicker Peel Sound Formation cycles (Fig. 11A, G) and those of the Rhondda Beds (Kelling, 1968; sequence not shown in Fig. 11) an additional mechanism may have operated. If they had been formed by channel migration in very large rivers it would be expected that some very large scale crossbedding would be preserved, as described by Conaghan and Jones (1975), but such structures are in fact not present in these examples. It is suggested that much the same process may have taken place as in alluvial fans and deltas: sedimentation in a given channel or channel complex results in aggradation above the surrounding area and a progressive loss of stream competency in response to the reduction of slope, this process being reflected in a gross upward fining. Eventually, probably during a flood event, the channel wall is breached and flow is diverted into topographically lower areas. The old channel or channel system is abandoned and the last sediments formed there will be fine-grained drape deposits. In deltas and meandering rivers this mechanism is responsible for the development of crevasse splay deposits (Coleman and Gagliano, 1964). The process has been described on modern alluvial fans (Denny, 1967) and is known to take place in the modern braided Kosi River, India, which has migrated laterally over its own deposits a distance of 112 km in 228 years (Gole and Chitale, 1966). Flow diversion would result in the catastrophic creation of new channels, and this probably is the origin of the scour surfaces, up to 5 m deep, at the base of the Battery Point cycles (Cant and Walker, 1976, p. 104). More research is needed to investigate this idea, but some support for it comes from a recent study by Campbell (1976), who described some exceptionally well-exposed Jurassic multi-storey channel-fill deposits interpreted to be of braided-river origin. They range from 6 to 60 m in thickness and generally contain several small-scale cycles 1 to 6 m thick similar to that shown in Fig. 11I. The fine-grained facies normally are truncated or removed, except in the uppermost cycle, and the complete multi-storey sequences show a gross upward fining. Each sequence probably represents several periods of channel aggradation (as described at the begin-

ning of this section) at decreasing energy levels, until the entire chann
system was abandoned for a new location by a process of avulsion.

It remains to discuss the origin of the coarsening-upward cycles. Th
shown in Fig. 11M is on average 60 cm thick and may result from re-occ
pation of a small abandoned channel. A similar origin was proposed for tl
paleo-outwash cycles (Fig. 11N) described by Costello and Walker (1972
but it is difficult to envisage a channel of the required depth (>2 m) bei
cut-off by levee development in an active braided system. Secondly, chann
shifting (and re-occupation), when it does occur, is the result of a floo
event, and the upward change in grain size is likely to be sudden rather th
gradual. An alternative explanation of these coarsening-upward cycles w
proposed by Rust (1975, p. 247) who related them to sedimentation in
depression developing over a mass of buried and slowly melting ice. Suppo
for such an interpretation would appear to be provided by the fact that tl
coarsening-upward sequences are confined to channel-like depressions co
monly about 6 m wide, but locally reaching a width of 60 m (Costello ar
Walker, 1972, p. 395). Small coarsening-upward cycles have been observed
the meandering, highly sinuous River Endrick by Bluck (1971). They for
in point bars where coarse bar sand and gravel migrate downstream ov
finer-grained lee-side sediments. But such cycles comprise only a minor pa
of the complete point-bar complex, which consists of an overall finin
upward succession. Bluck (1976) also postulated that coarsening-upwa
sequences were forming in a similar way in some low-sinuosity rivers in Sco
land, but he did not document this in detail. Costello and Walker (1972,
399) suggested that such sequences may be "highly indicative, if not dia
nostic, of braided situations, where a large number of dry disused channe
exist on the floodplain". However, Bluck's (1976) descriptions are the on
other record of coarsening-upward sequences in the braided environme
known to the writer, and their origin and relative importance remain u
certain.

Generalized vertical profile models

Three main parameters affect the nature of braided-river deposits, chann
depth, bedload grain size, and discharge (amount and variability). Examin
tion of the various profile types described in the foregoing sections indicat
that there are four principal ways in which these parameters combine
produce a given sedimentary succession and they are herein proposed as fo
empirical vertical profile models as a basis for the interpretation of braide
river deposits. The salient features of each model are summarized in Table
and hypothetical vertical sequences for each are given in Figs. 12 to 15. Ea
model is based mainly on descriptions of ancient rocks, but is named afte
modern river which appears to typify the interpreted depositional enviro
ment.

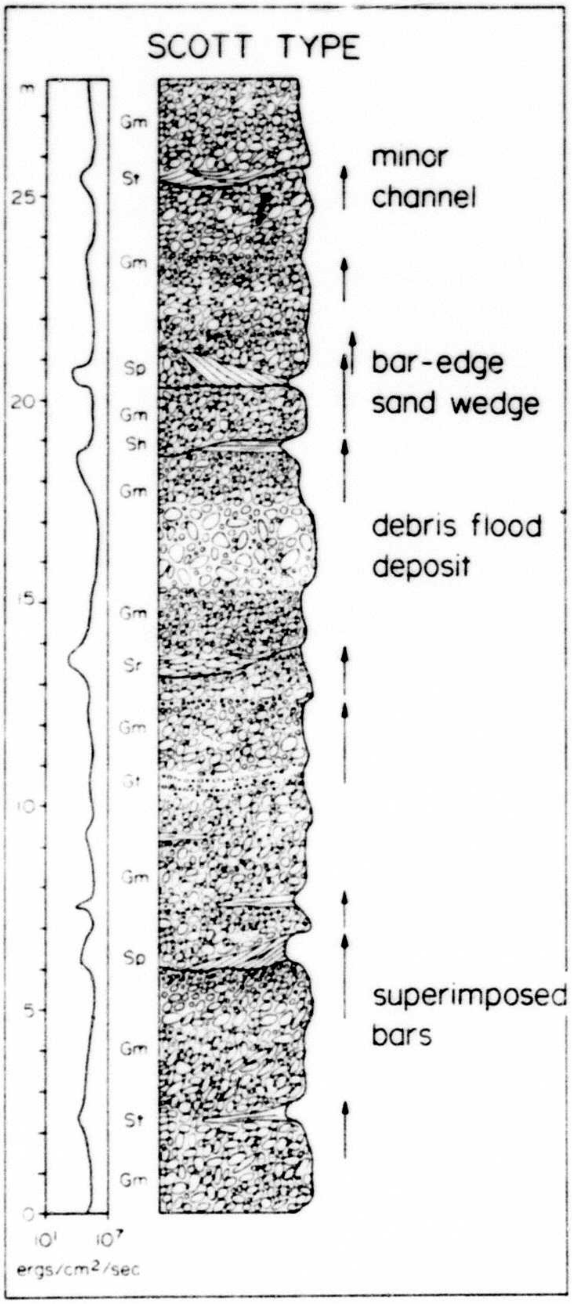
The Scott type (Scott fan of Boothroyd and Ashley, 1975), is character
tic of proximal, gravelly rivers (Fig. 12); it consists of small-scale gravel-sa

TABLE V

The four principal facies assemblages in braided-stream deposits

Name	Environmental controls	Main facies	Minor facies	Small-scale cycles * (within-channel)	Large-scale cycles *	Modern examples	Ancient examples
Scott type	proximal, gravelly river, may be ephemeral	Gm	Gp, Gt, Sp, St, Sr	Gm ↔ Sp, St, Sr 1 m	none	Boothroyd and Ashley (1975), Gustavson (1974)	Miall (1970b), Nilsen (1968), McGowen and Groat (1971)
Donjek type	braided with marked differentiation of topographic levels, perennial or ephemeral	Gm St, Sp, Sr, Fl	Gp, Sh, Ss, Fm	Gm → Sp, St, Sr → Fl 1–3 m Sp ↔ St 5–150 cm	aggradation of channel systems followed by channel switching, up to 35 m thick	Doeglas (1962), Williams and Rust (1969)	Cant and Walker (1976), Kelling (1968), Steel (1974) and Corner (1975)
Platte type	topographic differentiation less marked, may be braided only at low water, perennial or ephemeral	Sp, St	Sh, Sr, Ss, Gm, Fl	Sp ↔ St, Sh, Sr 5–150 cm	as above? (none recorded to date)	Collinson (1970), N.D. Smith (1970, 1971a, 1972), G.E. Williams (1971), Picard and High (1973), Boothroyd and Ashley (1975)	Miall (1976), Mrakovich and Coogan (1974), Asquith and Cramer (1975), Leeder (1973a), Adeleye (1974), McGowen and Groat (1971), Selley (1965)
Bijou Creek type	ephemeral stream subject to catastrophic floods	Sh	Sr, Sp	Sh ↔ Sr, Sp 15–140 cm	none	McKee et al. (1967)	Miall (1970a), Adeleye (1974)

* Including only "autocyclic" cycle types (Beerbower, 1964).



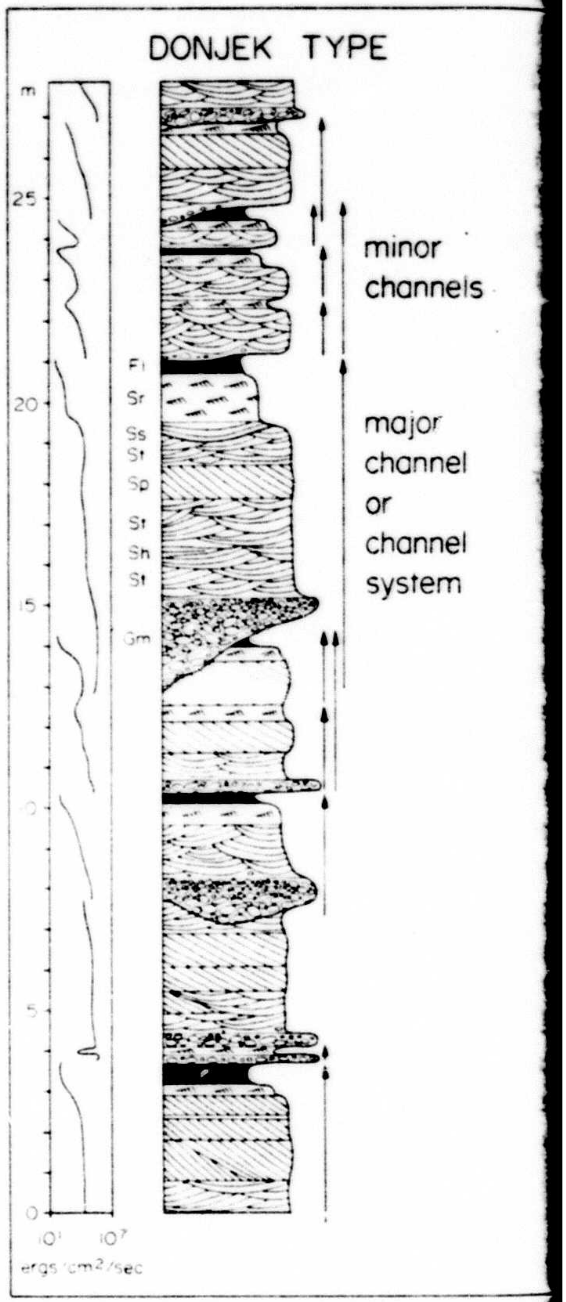
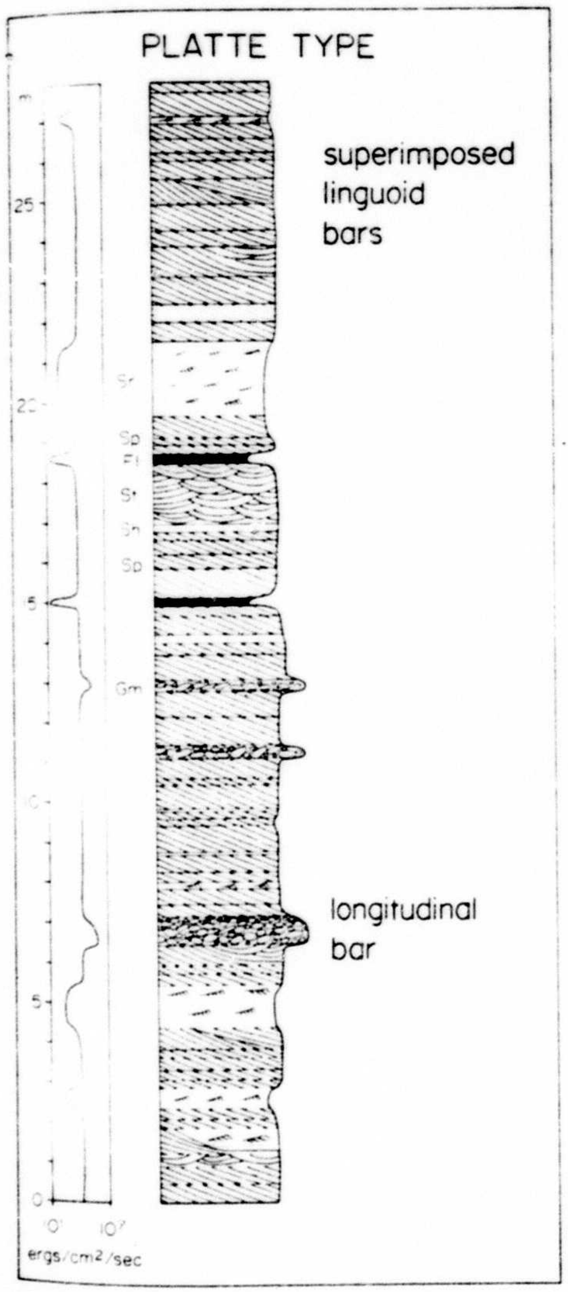


Fig. 12. Types of braided-river depositional profile: Scott type. Named to suggest a broad similarity with the proximal part of the Scott Glacier outwash river, Alaska (Boothroyd and Ashley, 1975). Column at left gives a generalized indication of changes in stream power. Arrows at right indicate cyclic sequences.

Fig. 13. Types of braided-river depositional profile: Donjek type. Named to suggest a broad similarity with the deposits of the Donjek River, Yukon Territory (Williams and Rust, 1969; Rust, 1972). This type of profile, characterized by cyclic sedimentation, is the most variable of the four. Gravel content varies widely (it is much more abundant in the Donjek than shown here) and cycle thickness may range from less than 1 m to more than 20 m.

cycles of waning-flood origin, and intervals of superimposed longitudinal-bar deposits. The Donjek type (Donjek River, described by Williams and Rust, 1969; Rust, 1972) is the most varied of the four models (Fig. 13). Fining-upward cycles of several scales may be present, the thicker cycles probably



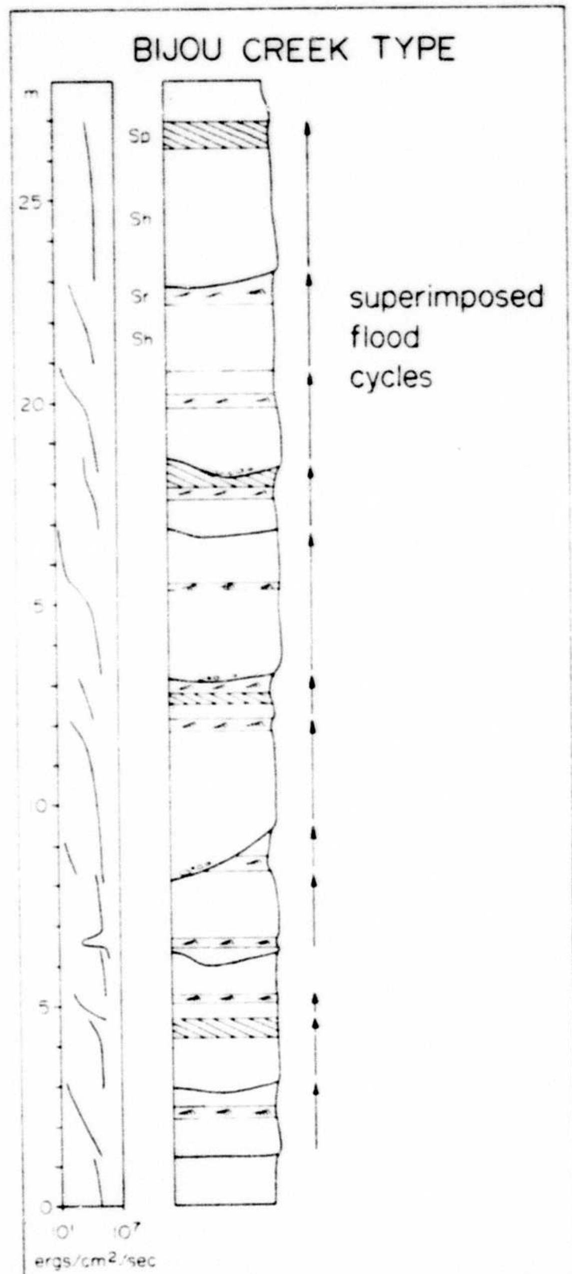


Fig. 14. Types of braided-river depositional profile: Platte type. Named after the Platte River, Colorado—Nebraska, the sediments of which have been described by N.D. Smith (1970, 1971a, b, 1972).

Fig. 15. Types of braided-river depositional profile: Bijou Creek type. Named after Bijou Creek, an ephemeral stream in Colorado, the sediments of which have been described by McKee et al. (1967).

reflecting sedimentation at different topographic levels within the channe system, or successive events of vertical aggradation followed by channe switching. The Platte type (Platte River, described by N.D. Smith, 1970 1971a, b, 1972) is thought to represent very shallow rivers or those withou marked topographic differentiation (Fig. 14). Linguoid, transverse or othe types of sand bar are the dominant depositional features. The Donjek an Platte types may merge imperceptibly into one another. The Bijou Cree type (described by McKee et al., 1967) represents the deposits of ephemera streams characterized by catastrophic floods (Fig. 15). Sand is transporte under upper flow regime conditions, and planar bedding with parting o streaming lineation is the dominant deposit.

Proximal—distal variations

There are no "absolute" indicators that can be used to estimate proximit to source in the braided-stream environment, but there are several param eters which show downstream changes, which are useful in providing relativ measurements of proximity. The most obvious of these is grain size (Fig. 16) Braided streams are dominated by bedload, which means a predominance o sand-grade or coarser material. Pebble or cobble gravel may be abundant But whatever the case, mean and maximum size tend to decrease down

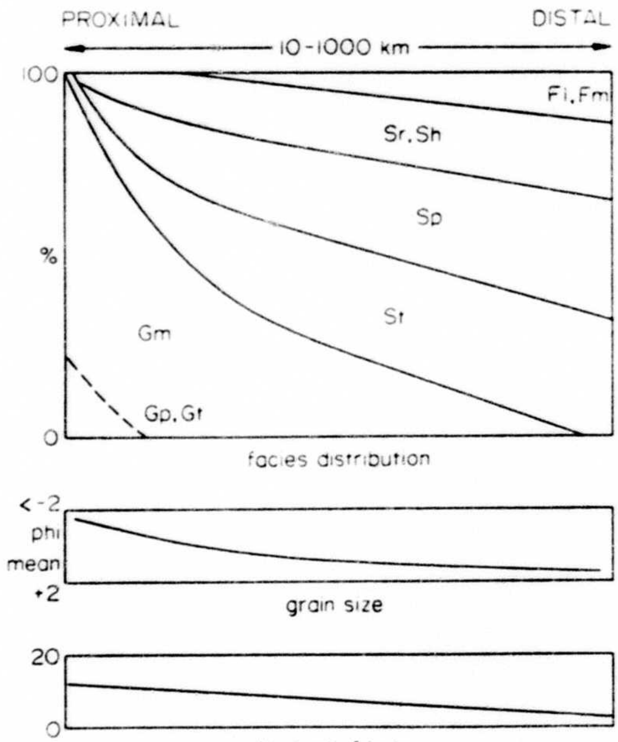


Fig. 16. Proximal—distal changes in braided-river deposits. Trends shown are approx mate, only.

stream. The rate of change varies depending on stream power. In the relatively small outwash fans studied by Boothroyd and Ashley (1975) gravel extends between 7.5 and 16 km from the glacier terminus. This compares with the Peel Sound Formation, in which pebble conglomerates extend for up to 16 km from the mountain front (Miall, 1970b, p. 569). In a recent debris flood deposit studied by Sharp and Nobles (1953) rare clasts up to 45 cm in diameter were carried 19 km from the source. In the Platte River gravel bars are abundant up to 250 km from the mountain front (N.D. Smith, 1970, Fig. 6).

Bar morphology and the types of internal stratification are dependent on grain size and, therefore, these also show downstream changes. Gravel bars commonly are longitudinal in form and are characterized by massive or crude horizontal stratification. Sand bars typically are linguoid and contain an abundance of planar crossbedding. N.D. Smith showed that in the Platte River the bar type changes systematically downstream, and the ratio of planar crossbedding to horizontal stratification shows a significant increase in the same direction. Both these parameters can be used in ancient rocks, as shown by N.D. Smith (1970), Asquith and Cramer (1975) and Gilbert and Asquith (1976).

Several authors (Schwarzacher, 1953; Momin Ul Hoque, 1968; Kumar and Bhandari, 1973) have noted a tendency for the thickness of planar crossbed units to decrease downstream in fluvial rocks. Pelletier (1965) has observed the same trend in some marine (deltaic?) deposits. The reason for this probably is that slope and velocity decrease in a downstream direction, resulting in a loss of competency and a decrease in the scale of bed forms. However, channel depth is also important in controlling bedform scale; depth generally increases downstream and would be expected to be reflected in larger bedforms. This is a difficulty that has yet to be resolved.

PALEOCURRENT MODELS

Langbein and Leopold (1966, p. H5) provide an empirical equation relating the sinuosity of a river to the range of angular orientations shown by the channel. Rewriting their equation using degrees instead of radians, as in the original, the relationship is as follows:

$$\text{sinuosity} = 1/[1 - (\theta/252)^2]$$

where θ is the maximum angular range of the mean azimuth. It follows from this that braided streams, which generally are of low sinuosity, should be characterized by lower directional variance than other more or less unimodal flow systems including meandering streams and many delta distributaries. That this has not been long since proven by field measurements is due to a variety of complicating factors (the following is modified from a discussion by Miall, 1974):

(1) Sedimentary structures form a hierarchy, ranging from a small-scale

ripple, through dunes, bars, channel reaches, etc. (conceptually an entire river may be considered as a large-scale sedimentary structure). Each scale of structures is characterized by a different range of directional variance, but many published paleocurrent studies have not taken this into consideration.

(2) Even within structures of approximately similar scale (same rank in the hierarchy) different modes of origin lead to different variances. Thus, it is a common observation that in the same outcrop of a fluvial or deltaic unit planar crossbeds (of Facies Sp) tend to show higher directional variance than trough crossbeds (of Facies St) (Agterberg et al., 1967; High and Picard, 1973; McGowen and Garner, 1970; Cant and Walker, 1976). In the braided-river environment the reason for this is that trough crossbeds represent dunes migrating downstream, whereas planar crossbeds represent bar avalanche slopes that may be oriented at high angles to the channel direction. Directional variance of planar crossbeds may vary depending on bar type. Transverse bars, which tend to be straight-crested ("cross-channel bar" of Walker, 1976, Fig. 10), will show less variance than horn bars such as those observed in the South Saskatchewan River (Walker, 1976, Fig. 10). The variance may therefore be a useful indicator of bar type in the ancient record.

(3) Directional variability depends to a considerable degree on discharge levels within a given stream. At high water stage flow is directionally relatively uniform, but at low water it may be controlled by the topography of the larger bedforms (Collinson, 1970, Fig. 10; 1971; Bluck, 1974; Karcz, 1972, p. 177; Banks, 1973b; Bluck, 1976). Bluck (1974) showed that clast imbrication, a flood stage development, showed less variance than any other structure, in an Icelandic sandur deposit. Rust (1975, p. 244) indicated that directional variance in imbrication is inversely related to clast size, which may be expected to increase with stream power. Barret (1970) demonstrated that in a Triassic sandy braided-stream deposit parting lineation, an upper flow regime structure, showed the lowest directional variance.

(4) Braided streams are characterized by extremely variable discharge, and Allen (1973) has shown that in many cases the fluctuations of discharge in a unimodal flow environment may be too rapid for bedform equilibrium to be

On the second of these factors has been investigated in any detail. Most of the available field data pertaining to directional variability in structures of different size and type were summarized by Miall (1974, Table 1), and will not be repeated here. The only attempt to relate directional variability directly to sinuosity was made by Miall (1976a).

ASSOCIATED FACIES

Braided-stream deposits pass up into, or grade laterally into a variety of other facies. The following notes provide brief descriptions of these facies, and references to descriptions of such facies relationships.

Alluvial fans

The separation of a braided stream from an alluvial-fan facies may be arbitrary, because sediment dispersion on alluvial fans generally takes place by means of ephemeral, braided channels. Conversely, some modern perennial braided rivers are known to migrate laterally over a cone of their own deposits, a process identical to that of alluvial-fan generation (Gole and Chitale, 1966). Bluck (1967) and Steel (1974) distinguished three alluvial-fan depositional environments on the basis of sediment transport mechanisms, debris flood, stream flood and braided stream. In some vertical sections these three environments succeed one another, and Steel (1974) interpreted this to mean a progressive reduction in source-basin relief and/or a gradual change to a more humid climate. Miall (1970a, b) and McGowen and Groat (1971) described a proximal to distal change from boulder conglomerates deposited by debris flood on alluvial fans to a predominantly sandy facies formed by braided streams under upper flow regime conditions. Nilsen (1968, p. 69) described conglomeratic braided-stream deposits interbedded with a coarse breccia interpreted to be of rockslide origin. Rust (1972) described some modern braided-stream deposits that are fed by small alluvial fans.

Glacial environments

Eight of the publications on ancient and modern braided streams that are summarized in Table III describe glacial outwash deposits. A variety of other glacial facies may be interbedded with the fluvial sediments, including slide gravels, till and varved silts and clays (Costello and Walker, 1972), and these may provide good diagnostic evidence of the glacial environment. The coarsening-upward sequences described by Costello and Walker (1972) may be the result of sedimentation in depressions caused by melting ice (Rust, 1975, p. 247) in which case their presence is an additional criterion of the glacial environment.

Rust and Romanelli (1975) described outwash deposits formed in a subaqueous environment that superficially appear similar to subaerial braided-stream deposits. Facies Gm, Sp, St, Sr and Fl are common. Several characteristics are diagnostic of a glacial and/or subaqueous origin: masses of buried ice, when they melt, cause the development of small graben structures bounded by step faults. Channels up to 38 m wide and 7 m deep are present; they normally are filled with massive sand that probably was deposited by subaqueous mass flow during spring melt conditions. Finally, graded, climbing ripple sequences are present in which the angle of climb increases upwards and the grain size decreases. Rust and Romanelli (1975, p. 189) interpreted these structures as varves, related in their formation to the fluctuations in discharge.

Meandering rivers

As noted earlier in this paper, rivers can alternate between the meanderi and braided (or even straight) condition depending on local river slope, se ment load and flow velocity. The deposits of both environments may interbedded in the ancient record, or at least found in close spatial proxi ity. Such a situation has been described by several authors, including Mooc Stuart (1966), Bluck (1967), Kelling (1968), Friend and Moody-Stu (1972), Leeder (1973a), Steel (1974), Allen (1974a) and Karl (1976). It important to distinguish the two types of fluvial deposit because this c lead to refinements in paleogeographic interpretation, paleohydrolog reconstruction and even local tectonic or climatic history. Some morpholo cal differences between the river types were presented in Table I, and sor sedimentological criteria are given in Table VI. The facies states present meandering streams are similar to those described in this paper, but th relative abundance and vertical order are different.

TABLE VI

Differences between the deposits of braided and meandering rivers

Criterion	Braided rivers	Meandering rivers
Lateral accretion deposits	point bars, linguoid bars, low-water bar accretion	fining-upward point bars (in cluding epsilon crossbeds)
Vertical accretion deposits	channel-floor bedforms sheet flood deposits bar-top deposits minor overbank deposits	overbank deposits channel-lag deposits
Type of scour surface	channel erosion	meander widening
Channel abandonment behaviour	progressive, as a result of aggradational fill	sudden, as a result of meand neck cut-off
Channel abandonment deposit	fining-upward cycle	fine-grained fill
Facies occurrence		
Gm	common (longitudinal bar deposit)	rare to common (generally as a thin lag deposit)
Gt, Gp	rare to common	absent
St	common	common
Sp	common	generally rare
Sh, Sr	common	common
Ss	rare to common	absent
Fl, Fm	rare to common	common
Channel-fill sequences	rarely > 3 m	commonly > 3 m

In some instances the differences between the two river types and their deposits may be minimal. Thus, Allen (1970) emphasized the predominance of point-bar accretion in meandering-river deposition but, as we have seen, point (or side) bars can be common in the braided-river setting. Secondly, fining-upward cycles may develop in both environments, although their origin may be different.

Thirdly, rivers are ideally one of the four basic types described in Table I, but there are exceptions. For example, Rust (1972) described a tract of the Donjek River which is strongly meandering, yet the main channel is internally braided; other examples are given earlier in this paper.

Two difficulties have arisen in our attempts to decipher fluvial deposits: a controversy over the relative importance of vertical versus lateral accretion, and a confusion over the various types of autocyclic fluvial cycle that can develop.

Moody-Stuart (1966) was the first to attempt to distinguish the deposits of high- and low-sinuosity streams. These he interpreted as resulting predominantly from lateral and vertical accretion, respectively. To some extent this interpretation has introduced a semantic confusion, because most depositional mechanisms in rivers involve elements of both processes. The important point which Moody-Stuart (1966) and Allen (1970) emphasized, is that high-sinuosity rivers build up their deposits mainly by accretion on bars on the insides of meanders. High-energy deposits on channel floors are formed at the same time as low-energy deposits on the channel margins, and to this extent the point bar can be said to accrete laterally (Allen, 1970, Fig. 12). Epsilon-crossbedding (Allen, 1963) is the result. Moody-Stuart (1966) developed an alternative model, of low-sinuosity river sedimentation, which included the vertical facies sequence shown in Fig. 11L. This he interpreted as the product of vertical accretion due to the gradual fill of a channel, mainly by ripple and dune deposits. Allen (1970, pp. 311—314) disputed the importance of vertical accretion. He pointed out that the surface of lateral accretion may dip at such a low angle (generally less than 15° and commonly less than 2°) as to be missed in most outcrops, and that no studies of modern rivers up to 1970 had satisfactorily documented vertical-accretion processes. Even in straight channels, Allen (1970, p. 313) argued, deposition is by downstream migration of alternate bars within a gently meandering thalweg.

However, several recent studies have shown that significant thicknesses of sediment can be accumulated in rivers by vertical-accretion processes, and these processes appear to be particularly characteristic of braided rivers. For example, the flood deposits described by McKee et al. (1967) and G.E. Williams (1971) were formed in ephemeral streams following severe storms, and resulted in blanket-like accumulations of sand up to 3.6 m thick. Secondly, work by Collinson (1970) and N.D. Smith (1970, 1971a, 1972) has shown that in perennial, as well as ephemeral braided streams, one of the most important sedimentary processes is the creation and migration of transverse and linguoid bars (these advance by lateral accretion, but not in the sense of simultaneous high- and low-energy sedimentation which takes place on point bars). Studies of ancient deposits of the Platte type (Table V) have shown that tens of metres of sediment can accumulate, which consist largely of planar crossbedded sand (Facies Sp) identical in every way to the deposits of transverse and linguoid bars in modern rivers (N.D. Smith, 1970; Mrakovich and Coogan, 1974; Jima sandstone of Adeleye, 1974; Asquith and Cramer, 1975; Miall, 1976a), whereas Facies Sp generally constitutes only a minor part of the meandering river model (Cant and Walker, 1976, p. 116).

It is even questionable whether all the fining-upward cycles described by Allen (1970) are of simple point-bar origin. Their thickness, alone, is a cause for doubt. Five per cent of the cycles described in Allen's paper are thicker than 20 m, which would indicate deposition in very large rivers. In at least some of these large-scale bedforms such as those described by Coleman (1969) and Conaghan and Jones (1975) would have been expected, but such structures have not been recorded. Secondly, some of the fining-upward sequences are of greater complexity than would be anticipated on the basis of Allen's hydrodynamic model. The sand member of the cycles can be divided into four facies. If more than four facies representatives are present at least one of the states must be repeated, and this is the case in 35% of the cycles studied. A maximum of thirteen facies representatives in any one cycle was recorded by Allen (1970, Table 8). Facies repetition is indicative of a vertical repetition of hydrodynamic conditions, and this does not accord with the simple model of gradational stream power and grain size developed by Allen (1970). Cycles of this complex type are not illustrated in his summary hydrodynamic diagram (Allen, 1970, Fig. 18). They would seem to demand the existence of subordinate topographic features on the point-bar surface, such as swales, chutes, or additional dune or bar fields. Such a condition is approaching that of a braided environment, and may not represent lateral accretion. Alternatively, some thick cycles may include multistorey units, representing the over-riding of one meander by another (Allen, 1974b).

In summary, cycles of a variety of types may develop in fluvial environments: point-bar, braided channel-fill, valley-fill and flood cycles; all tend to fine upwards, and careful study is needed to distinguish each type. Paleocurrent analysis may be of considerable assistance in two ways, firstly, in determining sinuosity as an aid to investigating river type (Miall, 1976a) and secondly, in determining the orientation relationships of the various sedimentary structures to one another as an aid to reconstructing bedform dynamics (Cant and Walker, 1976).

There is a temptation to ascribe meandering and braided deposits in a fluvial sequence to different climatic settings (Visher, 1976) but, as pointed out in the section on causes of braiding, the controls on river morphology are very complex. The situation is confused in some instances by the fact that the red pigmentation that is so characteristic of many continental deposits may occur preferentially in the meandering-river deposits, braided-river deposits being drab in colour. An extensive literature on the subject of red beds (Van Houten, 1973) has shown that in some instances red colouration can be related to an arid climate but, where the two fluvial facies, of contrasting colours, are interbedded, the differences cannot be climatic in origin. Moody-Stuart (1966) describes such an example, and relates the colour differences to varying oxidation rates induced by differences in the level of the water table.

Marine rocks

Braided rivers which debouch directly into the sea are not common at the present day. Most pass through a deltaic setting (such as the Cretaceous example described by Van de Graaff, 1972), and many evolve downstream into meandering, suspension-load rivers as they approach base level. However, before the appearance of abundant land vegetation, deltas consisting of braided distributary channels probably were abundant (Schumm, 1968a). These have been termed fan deltas by McGowen (1970) and McGowen and Scott (1974). The Proterozoic Glenelg Formation of Banks and Victoria Islands, Arctic Canada (Miall, 1976b; Young, 1974; Young and Jefferson, 1975) may be an example of such a deposit. It contains an abundance of Facies Sp and Sh, as well as interbedded units of probable tidal origin. Flores (1975) described a Pennsylvanian example of a small delta with braided distributary channels. McGowen (1970) described a modern fan delta on the Texas Coast.

CONCLUDING REMARKS

The main purpose of this paper has been to develop a comprehensive review of the braided-river depositional environment which can be used as a framework within which to interpret ancient rocks. The recognition and interpretation of the principal facies types and their vertical sequence as summarized herein, should be of particular utility in this regard, both in surface outcrops and in subsurface cores.

Several aspects of fluvial sedimentology have been deliberately omitted from this review because they are not critical to the main theme. Generalized grain-size information has been included, but the analysis of cumulative probability distribution curves has not been included for two reasons. In spite of several comprehensive papers on the subject (Visher, 1969; Glaister and Nelson, 1974) curve analysis still does not appear to provide unambiguous environmental interpretations, and the technique is cumbersome in that it cannot be applied in the field but depends on extensive laboratory analysis.

Hydrodynamic analysis of bedforms is a subject about which much has yet to be learned. Some general remarks have been included herein and they serve to illustrate the great hydrodynamic variability that characterizes the braided environment.

Following from hydrodynamic interpretations of individual bedforms, th next step is the reconstruction of the hydrology of the river or rivers whic deposit a given fluvial unit. Estimation of river depth, width, discharge an drainage area can now be attempted, using the empirical relationships pro vided by Schumm (1972) amongst others (although Schumm's equatio are most accurate only for small streams). An example of such an attemp was provided by Miall (1976a). A difficulty in braided-stream deposits is th scarcity of clear indicators of water depth. Previous paleohydrologic analyse (Allen, 1965; Schumm, 1968b; Cotter, 1971; Friend and Moody-Stuar 1972; Leeder, 1973b) have used the thickness of sand units in point-ba fining-upward cycles as an estimator of flood depth. As has been shown i this report certain braided-stream deposits (particularly those of the Donje type) contain cycles the thickness of which depends on channel depth, s that at least in these cases the difficulty can be overcome.

A widely accepted definition of bars and bedforms has yet to be deve oped, and some confusion will continue to arise so long as this is the cas An attempt has been made here to use terms that are in the most general us and to avoid subdividing types into categories that could not readily b recognized in the ancient record given the types of outcrop normally avai able to us.

Many of the studies of modern rivers upon which this review is based ha as their subject relatively small rivers, and it may be thought that their geo logical significance is low. However, as Bluck (1971, p. 94) pointed ou "small streams are numerous on the flood plain areas of larger rivers, so tha they may contribute much sediment". Another important point is tha ancient fluvial deposits obviously record situations of net aggradatio whereas many of the modern rivers that have been investigated are located i actively degrading mountain belts. The significance of this difference i terms of sedimentary processes has yet to be determined.

New research in braided rivers should concentrate on refining our unde standing of their hydrodynamics. It would also be particularly useful if som modern deposits, particularly of large rivers such as the Kosi and Brahma putra, could be cored in order to gain some ideas concerning the preserva tion potential of some of the facies and sequences discussed in this paper.

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Book Reviews

PHILOSOPHICAL BACKGROUNDS OF GEOLOGY

C.C. Albritton (Editor), 1975. *Philosophy of Geohistory: 1785–1970.* Benchmark Papers in Geology. Dowden, Hutchinson and Ross, Stroudsburg, Pa. and Wiley, Chichester, xiii + 386 pp. £ 13.00.

This book is one in the well known and generally highly regarded "Benchmark Papers in Geology" series whose object is "to gather into single volumes the critical material needed to reconstruct the background to any and every major topic" within the Earth sciences. With such a wide-ranging aim, a volume on the philosophical-historical aspects of geology was only to be expected, although it is both a surprise and a pleasure to find it coming so early in the series. In it, the Editor has collected together 15 geo-philosophical writings originally published between 1785 and 1970.

The book opens with a scene-setting article on the discovery of time in which Toulmin (1962) discusses the earlier geological revolution initiated by Hutton and Lyell and relates it to the more general eighteenth century revival of interest in history. Next comes the long abstract in which Hutton himself (1785) anticipated (anonymously) most of the conclusions of his "Theory of the Earth" published three years later. In Albritton's view, the conclusions are stated more clearly in the abstract than in the later book; and it is certainly widely accepted that Hutton was not at his best in communication by the written word. Indeed, the early propagation of his ideas owes much to their retailing and development in Playfair's "Illustrations of the Huttonian Theory of the Earth" (1802), excerpts from which form the third article here.

The fourth article is an extract from Lyell's "Principles of Geology" in which he propounds the principle of uniformitarianism, and this is followed by Whewell's defence of catastrophism. This juxtaposition is all the more remarkable in that both statements were originally published in the same year (1872). Next comes Chamberlin's famous essay "The Method of Multiple Working Hypotheses" (1890) which is followed by Gilbert's application (1896) to the specific problem of the origin of Meteor Crater, Arizona. The eighth article is Davis's equally famous "The Value of Outrageous Geological Hypotheses" (1926) in which he stresses the importance of speculation in the Earth sciences. The following article is then provided by Johnson (1933), one of Davis's students, who returns to consider the method of multiple working hypotheses.

The final six articles are more modern. First, Mackin (1963) discusses rational and empirical methods of investigation in geology, and then Anderson (1963) analyses the use of the principle of simplicity in historical geology. Next are three essays on the debate between the uniformitarianists and catastrophists. Hubbert (1967) traces the history of uniformitarianism, Simpson (1970) discusses the same principle in some detail and Hooykaas (1970) re-examines the catastrophist viewpoint in a more sympa-