Faces Models

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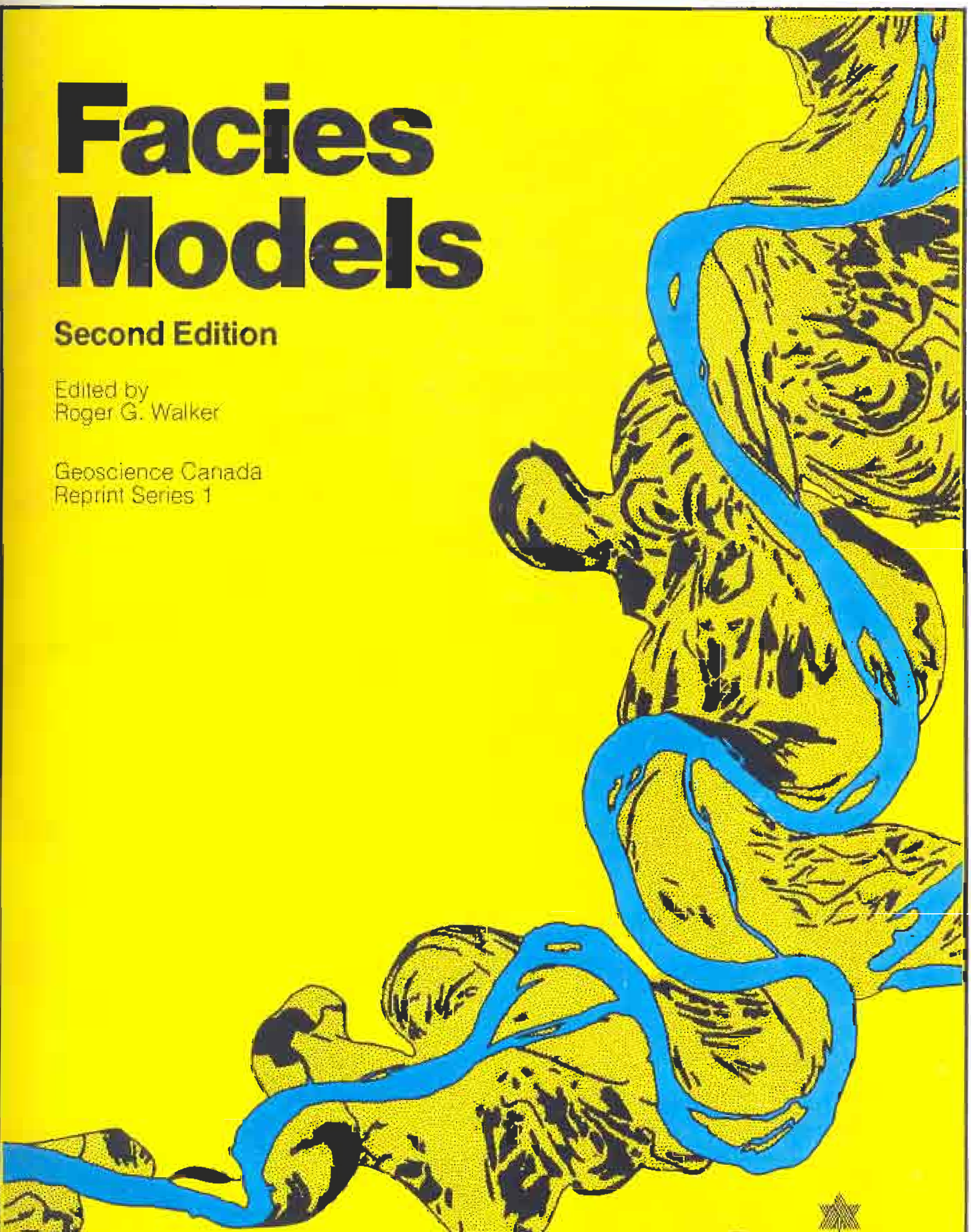

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Glacial Facies

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INTRODUCTION

Basin wide investigations of glacial facies based on integrated outcrop and subsurface data from modern and ancient sequences are as yet few in number, but the field is currently experiencing vigorous growth. This late development is probably because there are few significant Phanerozoic glacial sequences in North America or Western Europe apart from the late Cenozoic glacial record. Glacial sequences are also widely suspected to be non-productive and consequently glacial sedimentology has not caught the attention of petroleum geologists. This neglect contrasts with most other sedimentary environments, where assessment of petroleum potential and the search for modern analogs prompted much sedimentological research starting in the nineteen-fifties.

However, economic and potentially very important accumulations of glacial sediments occur in parts of Southern Asia and the Southern Hemisphere. These are of Permo-Carboniferous age, and represent a glacial episode within the former limits of the Gondwana supercontinent that lasted for about 90 million years. Representative deposits of this glaciation occur in successions up to several kilometres thick in South America, South Africa, India, Australia and parts of the Arabian Peninsula (Crowell, 1978; Hambrey and Harland, 1981; Visser, 1983). Petroleum is already produced from glaciated basins such as the Cooper Basin, and related basins of South Australia (Harris, 1981) and the Marsul Field, Oman (de la Grandville, 1982). Another thick glacial basin succession of potential hydrocarbon importance is the predominantly glaciomarine Late Cenozoic Yakataga Formation (20 to 2 Ma) of the Gulf of Alaska Tertiary Province (Plafker and Addicott, 1976; Armentrout, 1983). Exploration of these and other basins is creating a considerable stimulus for glacial basin studies and facies models.

Application of facies model concepts to glacial deposits is complicated by the fact that the only well-studied modern glacial environments are in continental settings, whereas because of selective preservation most ancient glacial sequences are glaciomarine in origin. Extensive Quaternary glaciomarine deposits exist, but with few exceptions these have not been studied using modern facies analysis methods. Considerable controversy is now arising over the interpretation of many Quaternary successions, much of which centres on the meaning and usefulness of the word "till". Space does not permit an examination of this problem here, and the reader is referred to discussions appearing elsewhere (Dreimanis, 1984; Eyles *et al.*, 1983, 1984). We do not use the term "till" in this paper. "Diamict", and its lithified equivalent "diamictite", are employed instead as non-genetic terms for poorly sorted gravel-sand-mud deposits.

Glacial facies models must be constructed from many sources, because so-called "glacial" sequences include the deposits of many distinct environments (rivers, lakes, continental shelves and margins, oceanic abyssal plains). In fact a case could be made for eliminating "glacial facies" as a distinct category and treating them as special sub-types of those other environments, characterized by the particular influence of grounded or floating ice and by the effects of rapid sea-level change.

In this paper we treat glacial facies under two main headings, continental and marine. The final part of the paper examines a number of glaciated continental margin depositional systems.

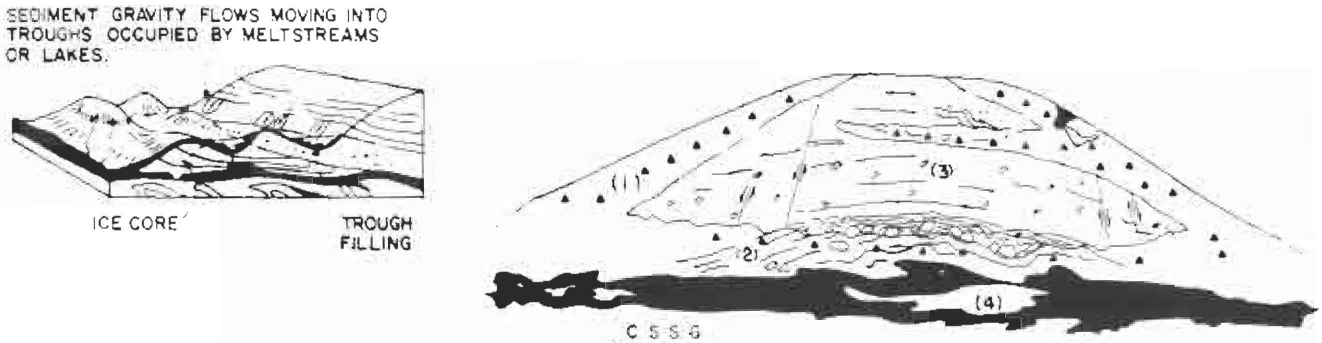
CONTINENTAL GLACIAL FACIES

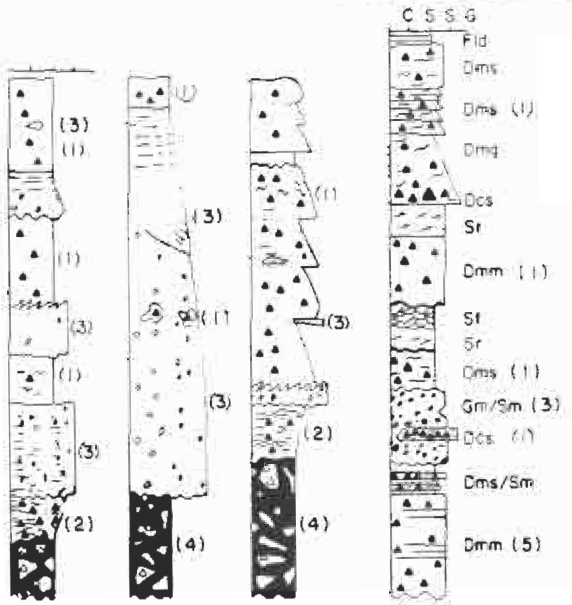
A fourfold division recognises grounded ice, glaciofluvial, glaciolacustrine and cold climate 'periglacial' facies.

Grounded Ice Facies

Under conditions of low temperature and sufficient precipitation, ice sheets grow to continent-size proportions. A complex zonation of subglacial erosion and sedimentation develops in response to different thermal conditions across the ice base (Andrews, 1982). Facies and associated bedforms exposed by retreat depend on whether the ice sheet margin is thin, inactive and frozen to the substrate, or actively sliding over the bed lubricated by basal meltwaters. In the first case (Fig. 1) thick englacial sequences of debris and ice are built up at the ice base either by refreezing of basal meltwaters or by intense folding of basal debris layers at the ice margin (Moran *et al.*, 1980; Paul, 1983). An inactive or retreating margin of this type becomes buried under a drape of diamict formed by the melt-out of englacial debris (Figs. 1 and 2). This mobile drape is resedimented down-slope by sediment gravity flow into local basins generated by the irregular melt of buried ice (Boulton, 1972; Lawson, 1982; Paul 1983: Fig. 3). Basin fills are exposed as hummocks when adjacent ice-cores, under a thinner sediment cover, melt-down more rapidly ('relief-inversion'). Typical vertical profiles through hummocks show uppermost sequences of resedimented massive, graded and stratified diamicts, variably reworked and interbedded with glaciofluvial and lacustrine facies, overlying crudely stratified diamicts which were aggregated *in situ* as the dirty ice base melted (basal melt-out; Fig. 1). These melt-out units may contain or drape over bedrock rafts that were present in the former ice base. Glacitectonised bedrock and incorporated substrate sediments may be an important component of the sedimentary sequence (Fig. 1; Moran *et al.*, 1980). Englacial structures (e.g., folded basal debris sequences) survive basal melt-out as ridge-like bedforms oriented transverse to former ice flow direction (e.g., Shaw, 1979). Broad belts of hummocky topography with complex internal stratigraphies, form extensive regional facies tracts in Quaternary glaciated terrains.

In contrast, Figures 4 and 5 show a wet-based ice margin where the glacier continues to slide over the bed even during retreat. Debris is transported within a crudely stratified basal layer less than 1 m thick, with intense abrasion between particles in the base of the





X-SECTION THROUGH A TYPICAL TROUGH-FILL EXPOSED AS A HUMMOCK

Figure 1
Facies deposited at the margin of a retreating or stagnant glacier with a thick englacial debris sequence (After Boulton, 1982; Paul, 1983). Five facies components can be identified within a complex hummocky topography (Fig. 2): 1) resedimented diamicts (Fig. 3) derived from melt-out of englacial debris and subsequent mass-movement over ice-cores; 2) diamicts formed in situ at the ice base by melt-out; 3) glaciofluvial, glaciolacustrine facies in ice-cored basins; 4) glaciotectonically deformed substrates; 5) lodged diamict (Fig. 4) from an earlier episode of glacier movement. For lithofacies code see Figure 15

Figure 2
Avaatsmarkbreen, Spitsbergen, showing a kilometre-wide arcuate belt of hummocky topography underlain by sequences such as shown in Figure 1. Photograph courtesy of T.E. Day

layer (the traction zone) and the substrate. Clasts that collide with the bed may beome lodged against the substrate. A characteristic and diagnostic glacially-faceted clast shape evolves as a result of dirty ice moving over the lodged clast (Boulton, 1978; Figs. 6 and 19). Continued lodgement gives rise to lenticluar beds of dense overconsolidated diamict (Fig. 4) which are massive but show many diagnostic structures indicating deposition under high basal shear stresses (Figures 4 and 7). Clasts show a strongly preferred direction of long axes aligned parallel to ice flow vectors. Measurement of a few 'shaped' clasts gives a rapid guide to former flow direction. Identification of lodged diamicts and former ice flow directions is of considerable importance in drift exploration programs in glaciated terrains where mineralised 'float' in glacial

◀ **Figure 3**
Lobate flow of resedimented diamict at Matanuska Glacier, Alaska. Resedimented diamicts become interbedded with glacio-lacustrine and glaciofluvial facies (e.g. Fig. 1).

Figure 4
Basal deposition at a thawed sliding ice margin (e.g., Breidamerkurjokull, Fig. 5), with typical vertical profile and diagnostic criteria for diamicts deposited by lodgement. Lithofacies code as in Figure 15 ▼

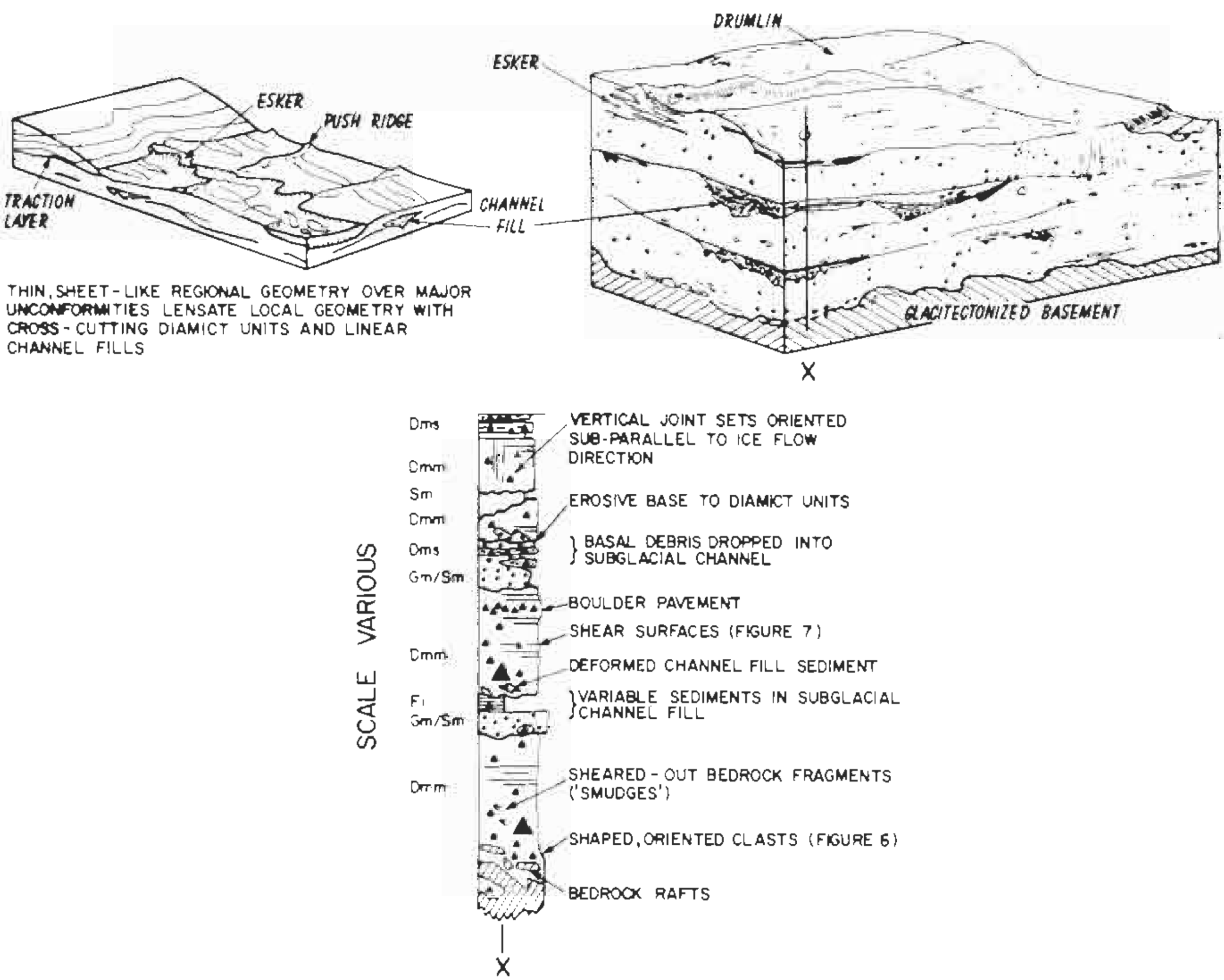


Figure 5
Glacier bed of lodged diamict exposed by the retreat of a sliding thawed glacier (Breidamerkurjokull, southeast Iceland). Small moraine ridges parallel to the ice margin are formed each winter by minor readvances of the margin. Low relief drumlins impart a streamlined surface. Section exposed in middle of picture is 8 m high.

Figure 6
Large glacially streamlined clast exhumed from diamictite deposited by lodgement below a Late Proterozoic sliding ice sheet: Mauretania, West Africa. Direction of ice flow from right to left. The presence of shaped clasts with a consistent long axis orientation is, in conjunction with other criteria (Fig. 4), an aid to identification of diamict(ite)s deposited by lodgement. In addition, it also enables a glacial source to be recognised where diamict(ite)s have been reworked or resedimented (Fig. 19).

sediment is traced to buried mineralised bedrock zones by reference to dispersal fans.

The geometry of lodged diamicts across a sedimentary basin is sheet-like lying above marked local and regional unconformities. Within this apparently simple geometry, are strongly lenticular diamict units in a cross-cutting and overlapping relationship as a result of intermittent erosion and changes in ice flow direction (Fig. 1). Lodged diamicts, containing clasts and matrix from contrasting bedrock lithologies, may be superimposed during a single glaciation emphasising the need for care in interpreting ice advance/retreat cycles from multiple stratigraphies. An integral component of the subglacial stratigraphy may be channel fills resulting from subglacial stream drainage (Davis and Mallett, 1981; Eyles *et al.*, 1982). These fills show a very wide range of lithofacies and may contain irregular diamict masses flushed from the former ice base. Channels have a planar upper surface truncated by overlying diamict, are oriented subparallel to ice flow direction and are genetically related to esker ridges (Fig. 4).

Eskers record either the infilling of

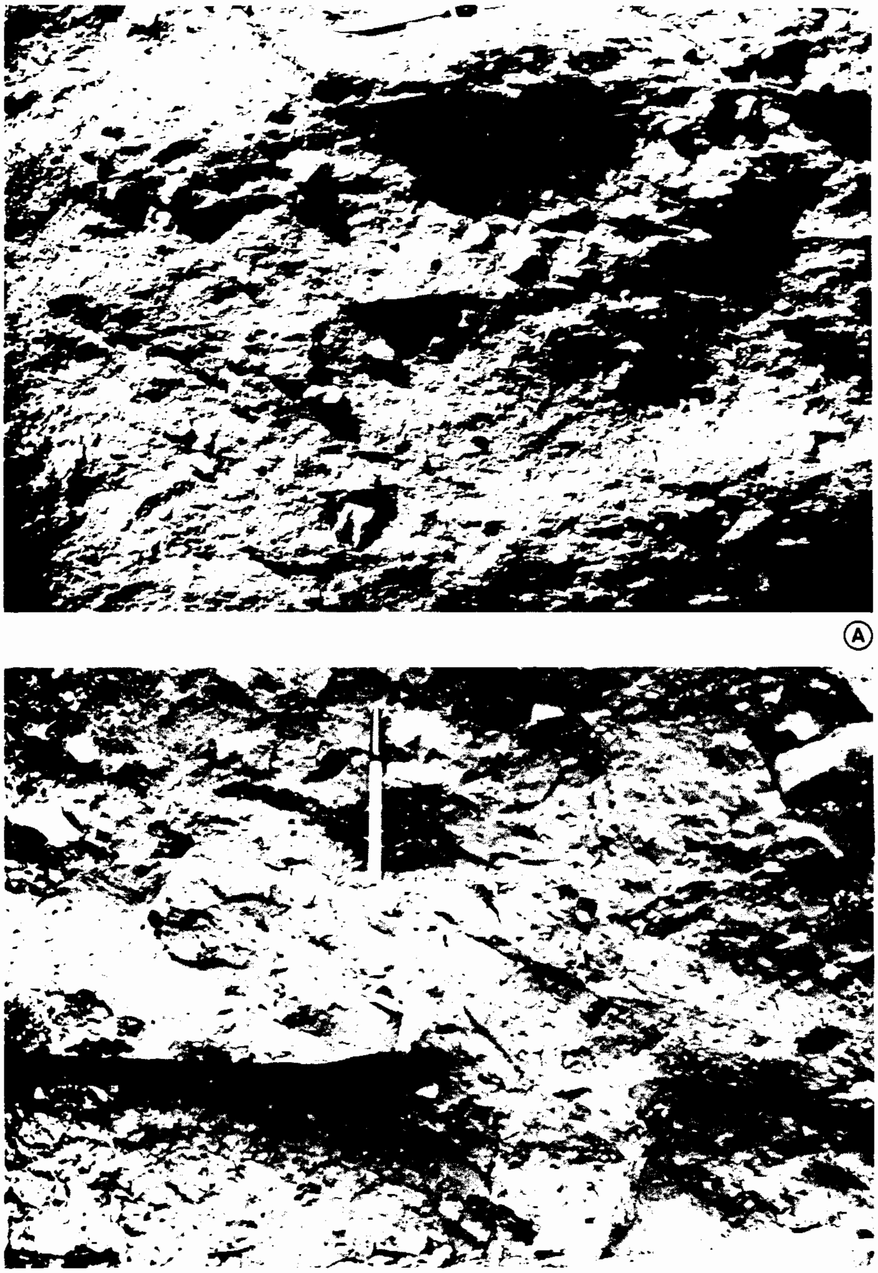


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Figure 7
Shear surfaces in lodged diamict, resulting from glacial shear stress at sliding ice base: a) diamictite of the Late Proterozoic Jbeliat Formation of Mauretania, West Africa; and b) diamict of Late Pleistocene age from Northumberland, Britain.

meltwater tunnels within and under the ice mass, confined braided stream sedimentation in open ice walled channels or deposition as overlapping subaqueous fans in standing water as the ice margin retreats (Banerjee and McDonald, 1975). Tunnel-fill eskers are commonly exposed as steep sided ridges showing tabular and longitudinally extensive cross-bedded sandy lithofacies with restricted variance in paleocurrent directions. A beaded form is typical of eskers deposited in water. These show rapid downstream transitions into fine-grained lacustrine facies with considerable variation in paleocurrent directions. Eskers show characteristic deformation structures (faults, subsidence basins) resulting from the melt of underlying and buttressing ice cores and may drape over lodged diamicts having channeled upper surfaces.

Other bedforms exposed by the retreat of wet based glaciers are drumlins and more narrow flutes formed either by selective subglacial lodgement around obstacles or erosional streamlining of bed material (Boulton, 1979).

The most spectacular example of an exposed bed of a paleo-ice sheet, outside Quaternary glaciated terrains, occurs across the 1500 km broad Taoudeni Basin of Mauretania (Deynoux and Trompette, 1981) which represents the sedimentary cover of the West African Shield. Metamorphic effects are absent and extensive Late Proterozoic glaciated surfaces with streamlined highs have been exhumed. Continental glacial facies are less than 50 m thick and contain lodgement diamictites averaging 3 to 5 m thick covering glacitectonized surfaces (Figs. 5, 6 and 7). Lodgement diamictites are overlain by thick aeolian sequences recording arid cold climates following ice retreat. Whereas thicker sequences may be accumulated by repeated continental glaciation, the reduced regional thickness of individual lodged diamict(ite) units (generally < 15 m) is an important criterion for distinguishing continental facies from the thicker diamict(ite) sequences deposited in glaciolacustrine and glaciomarine environments.

Glaciofluvial Facies

Continental ice sheets release large quantities of meltwater that drain broad outwash plains, reworking glacial debris into a distinctive suite of glaciofluvial lithofacies.

Aggradation at the head of outwash fans is frequently rapid enough to bury portions of the adjacent ice margin; such ice-contact depositon generates widespread deformation structures in coarse-grained crudely-bedded or massive proximal outwash gravels. 'Pitted' or 'kettled' outwash surfaces may extend over many square kilometres, and are flanked by eskers or complex ice-contact diamict sequences (e.g., Fig. 1).

Beyond the immediate ice terminus lies the outwash plain or sandur (plural: sandar, Icelandic). Modern sandar are well developed in Alaska, Arctic Canada and Iceland (Rust, 1972; Church, 1982; Boothroyd and Ashley, 1975; Boothroyd and Nummedal, 1978). Glacial outwash rivers are typically of multiple-channel (braided) type, and are normally of low sinuosity. Their morphology and sedimentology are similar to those of other braided streams, (see "Coarse Alluvial Deposits", this volume) such as those occurring in arid environments (arctic to tropical), and their analysis forms part of general reviews of coarse-grained fluvial sedimentation that appear elsewhere (Miall, 1977, 1978; Bluck, 1979; "Coarse Alluvial Deposits", this volume). The braided morphology reflects an abundance of coarse bedload, variable discharge and non-cohesive channel banks. The rivers are unable to transport more than a fraction of the available bedload except during extreme floods (Ostrem, 1975), and at other times gravel and sand are deposited as a variety of bedforms and bars.

Modern sandar are up to about 100 km in length, and show gradients varying from 2 to 50 m/km. Typically there is a downstream decrease in grain size from coarse gravel to sand. Sandar commonly terminate as coastal fan deltas (e.g., Galloway, 1977) which may be flanked by barrier ridges created by seasonal pack ice-push. Deflation may strip abandoned fan surfaces and deposit blankets of wind blown coarse silt (loess) downwind. Under more humid conditions, swamps of subarctic vegetation may become established in the proximity of water bodies. These may generate thick peat accumulations and ultimately cold-climate coals such as those of the Permo-Carboniferous coal bearing glaciated basins of the southern continents (e.g., Le Blanc Smith and Eriksson, 1979).

Glaciolacustrine Facies

Overdeepening by glacial erosion, glacial derangement of drainage and the release of large volumes of meltwater results in frequent lacustrine ponding. Lake basins vary from narrow 'alpine' basins in areas of high relief and isostatically-depressed continental interiors evacuated by ice sheets. Lake Agassiz is the most famous example of the latter, extending over a total area of some 1,000,000 km^2 in interior North America (Teller and Clayton, 1983). Many complex classifications of lake basins exist, but for the sedimentologist working from outcrop and subsurface lithofacies data a simple distinction between 'periglacial' and 'proglacial' lake bodies may be the most useful. Periglacial lakes are not in direct contact with an ice margin and are fed by braided stream systems. In contrast proglacial lakes form in direct contact with the ice margin and receive a substantial volume of sediment direct from meltwater conduits and subaqueous fans with an additional component supplied by ice rafting (Fig. 8).

Periglacial Lakes

Sedimentation is dominated by the rapid growth of arcuate delta lobes. Incoming sediment laden and higher density meltwaters, therefore, move down these lobes as density underflows. A distinct sequence of sandy lithofacies is deposited with each melt season (Fig. 9) and records the start, later increase and ultimate decline of density underflow activity (Ashley, 1975). The sequence is bounded top

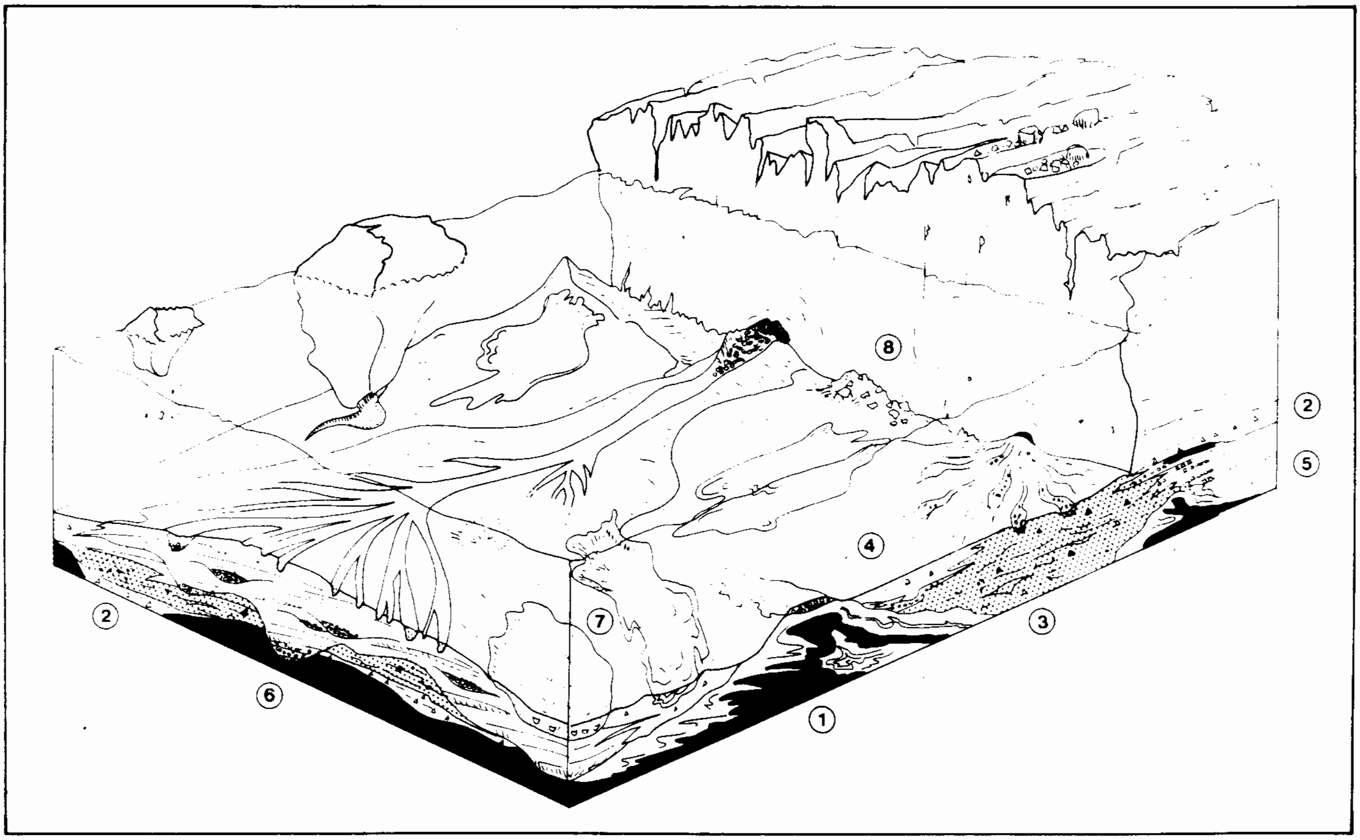


Figure 8

Proximal subaqueous sedimentation; 1) glacitectonized marine sediments; 2) lensate lodged diamict units; 3) coarse-grained stratified diamicts (Figs. 12 and 15); 4) pelagic muds and diamicts (Fig. 12); 5) coarse-grained proximal outwash; 6) interchannel cross-stratified sands with channel gravels; 7) resedimented facies (debris flow, slides and turbidites); and 8) supraglacial debris. Deformation results from ice advances (Fig. 15), melt of buried ice and iceberg turbation. Suspended sediment plumes not shown. The same model may apply with modifications to sedimentation adjacent to grounding lines of large ice shelves.

and bottom by a 'winter' clay layer. Distal equivalents of these sandy lithofacies can be identified as isolated 'starved' ripples and rare ripple-drift cross-lamination within clay/silt couplets (varves). A varve is an annually-produced couplet of silt and clay laminae and shows a sharp division between the clay and silt components (Fig. 10), each deposited in a different season and by different sedimentary processes. The light coloured silt unit (summer layer) may be graded and can show multiple laminations representing deposition by a single pulsating or intermittent density underflow with a minor contribution of pelagic material from interflows or overflow 'plumes' of suspended sediment (Banerjee, 1973). The dark clay unit (winter layer) may show normal grading recording principally the deposition of suspended sediment under a closed lake ice cover. Bioturbation and trace fossils are commonly present. Clay layer thicknesses are generally uniform across the basin but may contain massive or cross-stratified sands and laminated silts recording the winter drawdown of lake levels, delta foreslope slumping and the generation of grainflows and turbidity currents (Shaw, 1977; Fig. 9).

The term 'varve' (or 'varvite') is used unfortunately as a routine descriptive term for sediments that more correctly should be termed rhythmically-laminated, or simply laminated, sediments (rhythmites, laminites). The real significance of many 'varved' sequences as indicators of seasonally controlled sedimentation remains to be assessed rigorously in view of observations showing that many such units are not annual couplets but single or multiple graded units (silt – clay; units C,D and E of a Bouma sequence; see "Turbidites and Associated Coarse Clastic Deposits", this volume) deposited by discrete-event turbidity currents with no seasonal control (e.g., Lambert and Hsü, 1979; Fig. 10). In Pre-Quaternary sequences 'varves' are used to infer glaciolacustrine, and therefore continental environments and seasonality of climate, but new data increasingly show that many of these are discrete-event turbidites of glaciomarine origin (Fig. 10).

Diamict lithofacies are a very minor component of reported periglacial lake sequences. Graded and stratified diamict lithofacies may, however, occur as thin channelised lenses within deltaic sequences and result from the mixing of fine and coarse-grained sediments by downslope resedimentation (e.g., Cohen, 1983).

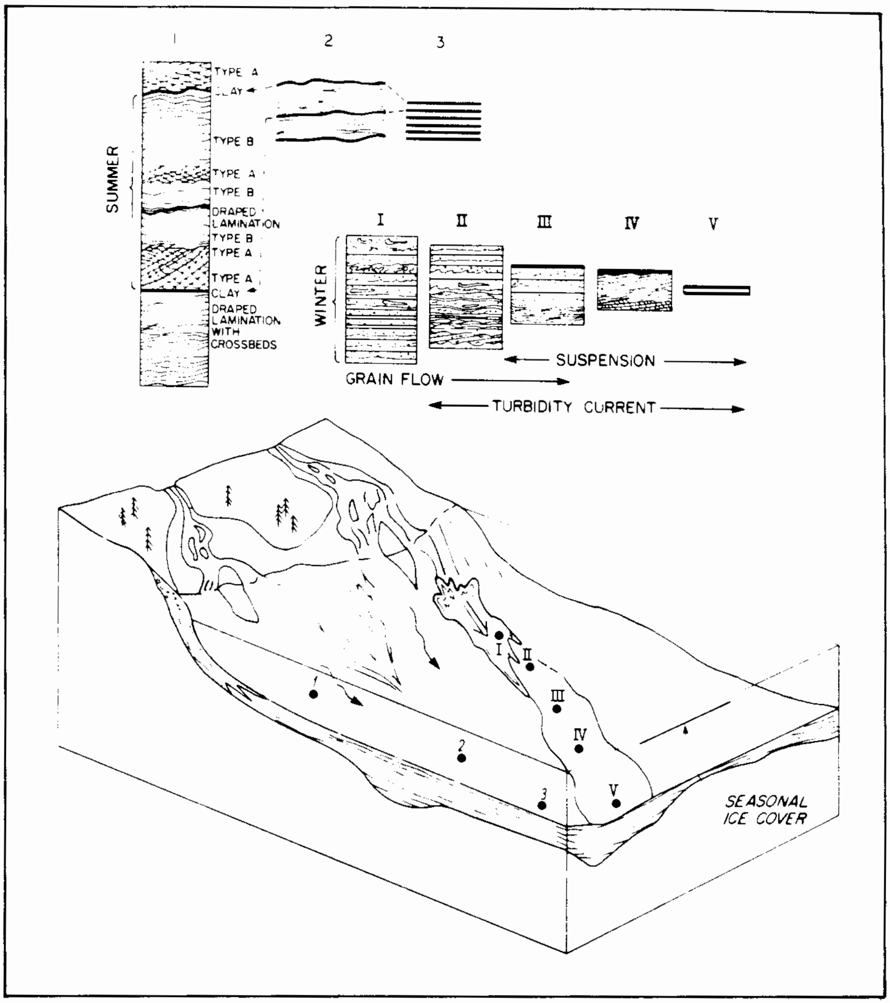
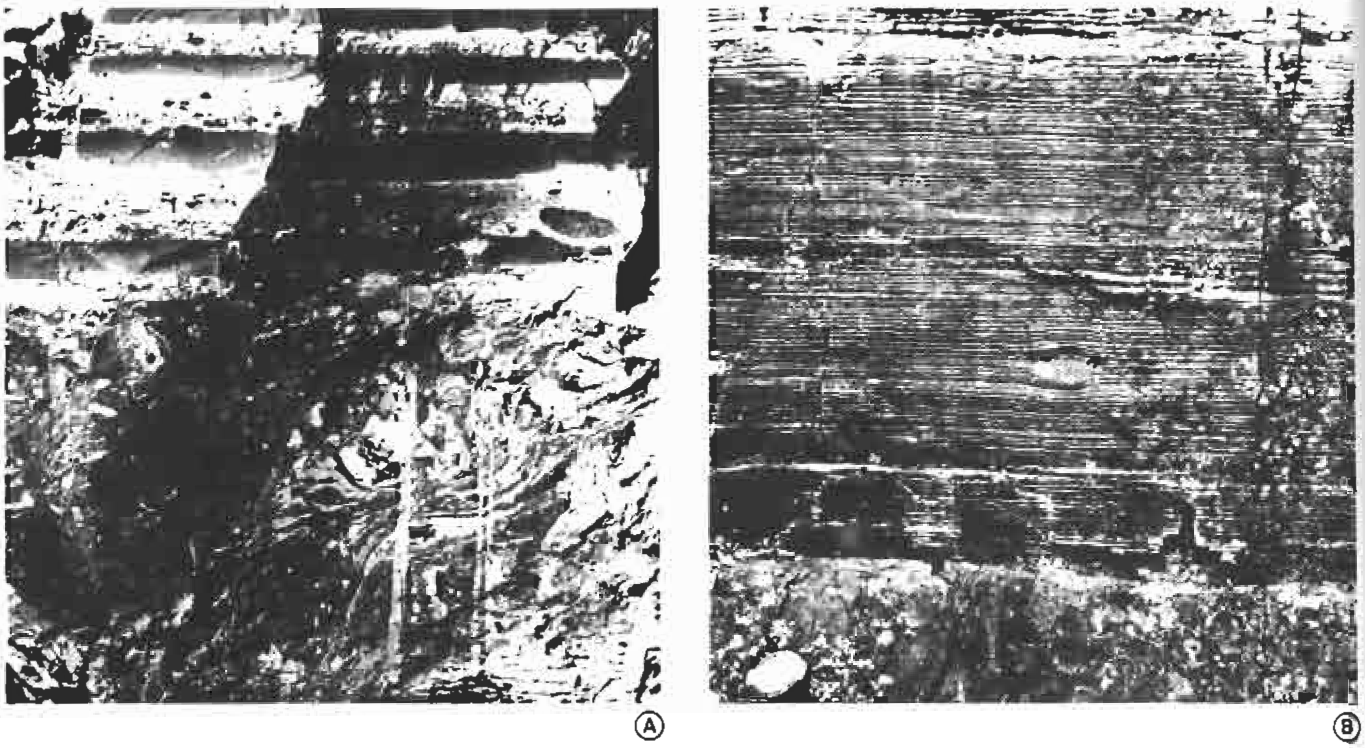


Figure 9
Annual cycles of sedimentation in a periglacial lake. 1 to 3 show lateral facies variation in varved sediments (After Ashley, 1975). I to V show sand lithofacies within winter clay layers that result from delta foreslope slumping (after Shaw, 1977).

Proglacial Lakes

Facies modelling in proglacial lacustrine sediments is frustrated by logistic difficulty of working on modern proglacial lakes, (Gustavson, 1975) and the small size of these lake basins compared to Pleistocene and older examples. Extensive Pleistocene proglacial sequences exposed around the modern Great Lakes in North American are of considerable significance therefore to modelling studies. Figure 11 shows a 10 km long section of a Late Pleistocene glaciolacustrine sequence from the Lake Ontario Basin, divided into three repeated lithofacies associations. Diamicts at the base of each association have a blanket-like geometry, thickening in topographgic lows and thinning over highs, and internally are complex assemblages of massive and stratified lithofacies. Massive facies result from the basin floor accumulation of suspended sediment and ice rafted debris with stratification produced by variable reworking by downslope resedimentation and traction currents, as is seen in glaciomarine environments (Figs. 12 and 13 and below). Diamict assemblages are overlain conformably, frequently with a gradational contact, by laminated silty clays of probably turbidite origin, containing dropstones. Laminated silty clays occupy broad channels (Fig. 11) illustrating the control on



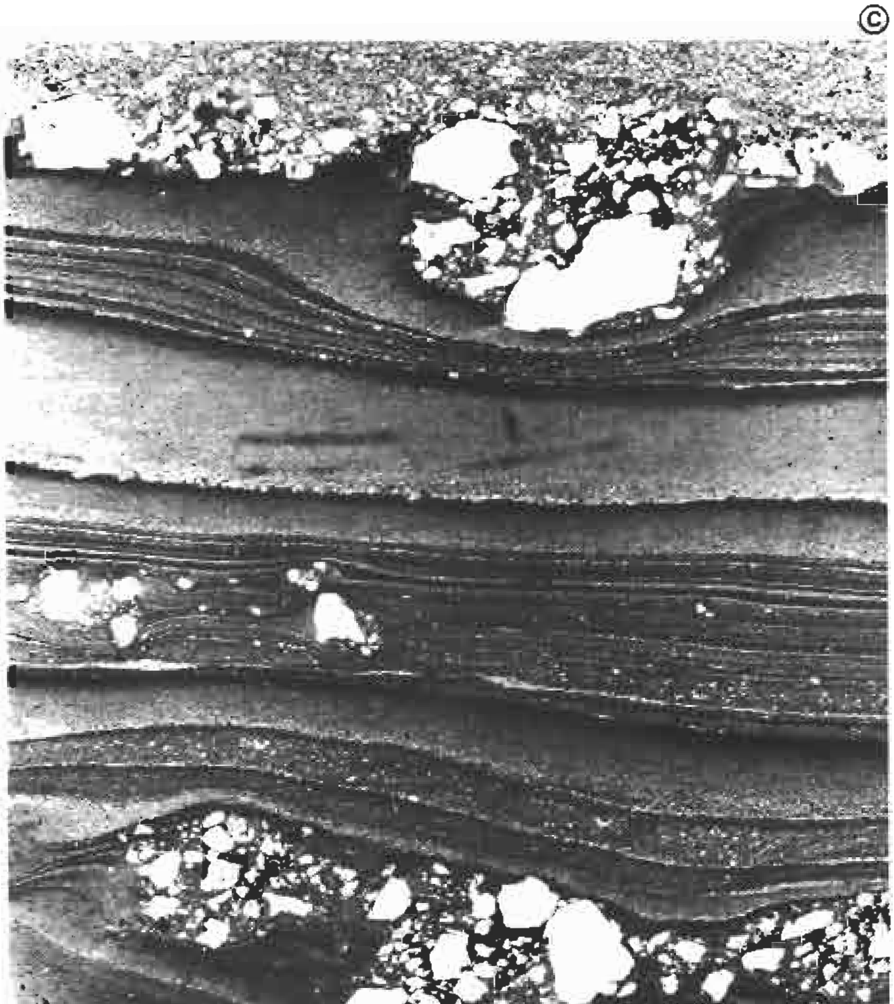
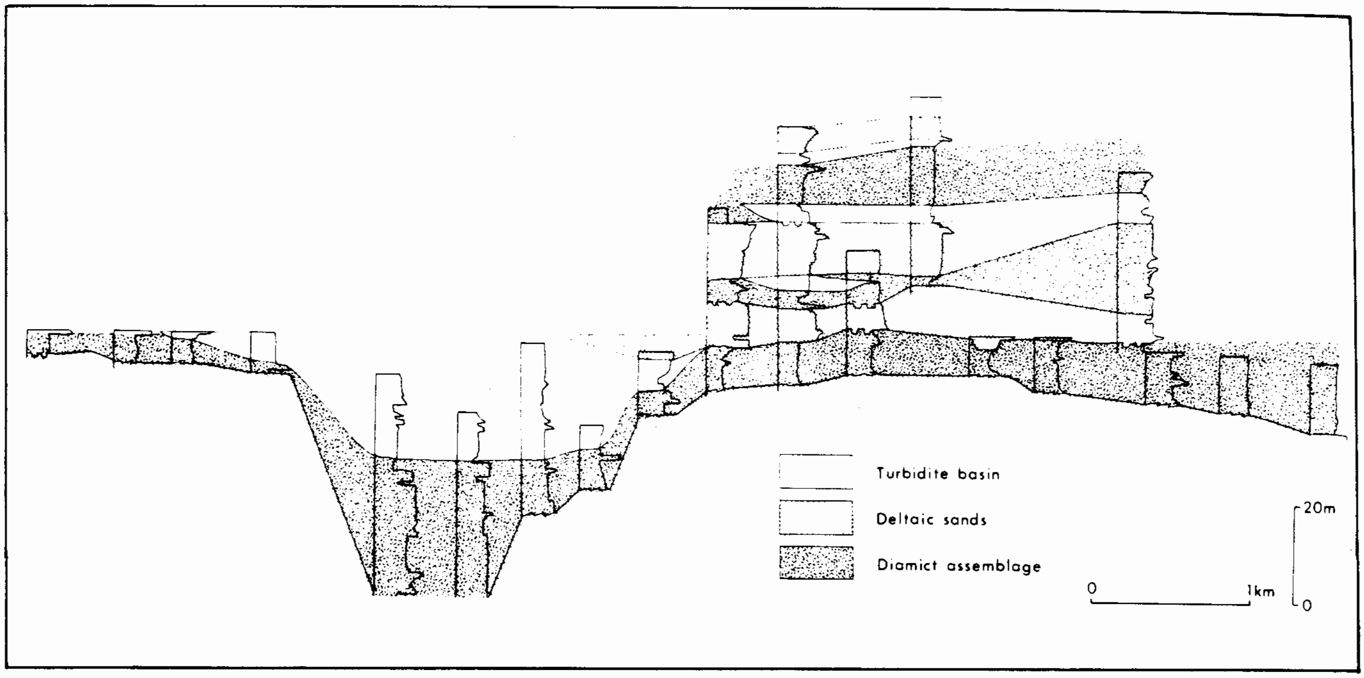
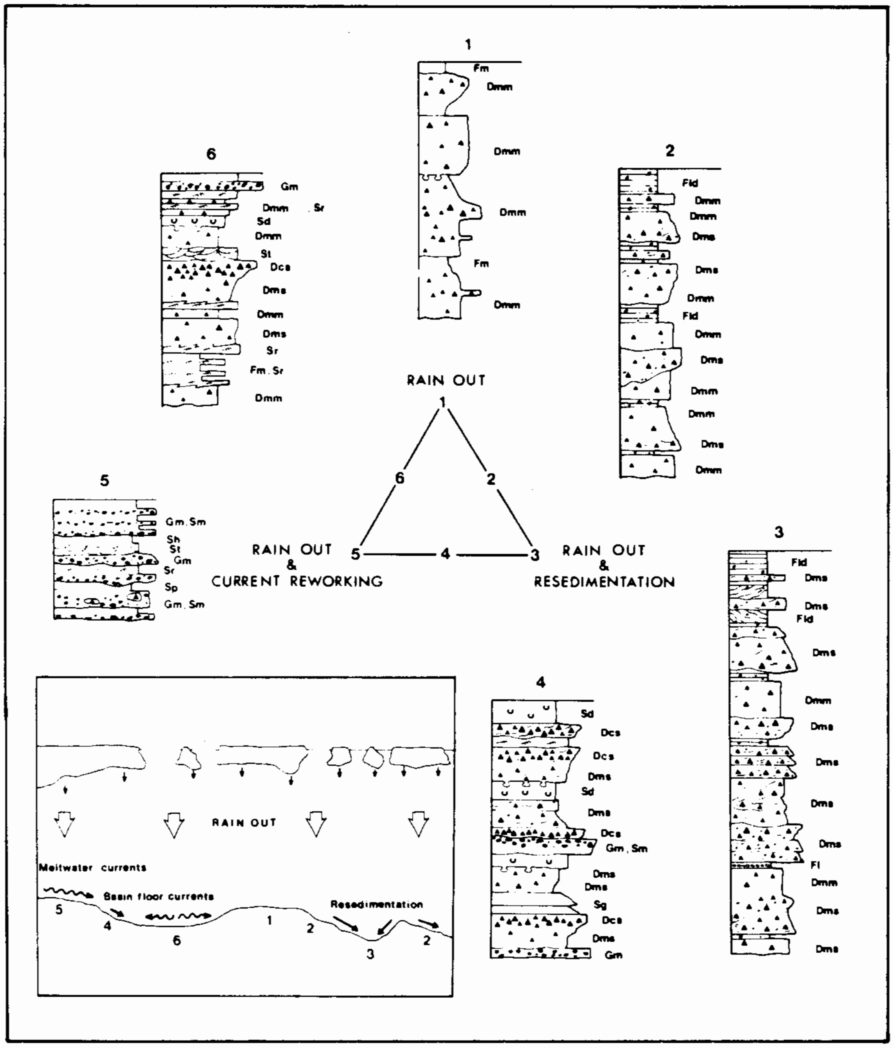


Figure 10
a) Laminated and graded silty clays described as 'diamictic varves' by Banerjee (1973) with abundant ice-rafted lithic clasts, diamict pellets and angular rip-up clasts of silt and clay. The thick lower unit was probably transported as a debris flow. Don Valley, Ontario. b) Laminated argillites, consisting mainly of the deposits of dilute turbidity currents. Note scattered dropstones, and a sandy sediment gravity flow with a "flow nose" at the base of the picture. Early Proterozoic Gowganda Formation, Ontario. c) Laminated graded argillites of turbidite origin from the Gowganda Formation. Width of photograph is 1 cm; photograph courtesy of P. Fralick.

density underflow location of basin floor relief, and probably record the release of flows from a prograding delta. Coarsening-upwards ripple laminated, planar and trough cross-bedded sands complete the lithofacies association recording delta progradation over sites of diamict accumulation. The most common deltaic facies is a crudely bedded silty sand with abundant liquefaction structures indicating rapid subaqueous deposition (e.g., Rust, 1977). Sands are often loaded into diamict upper surfaces as a result of progradation across a wet diamict substrate.

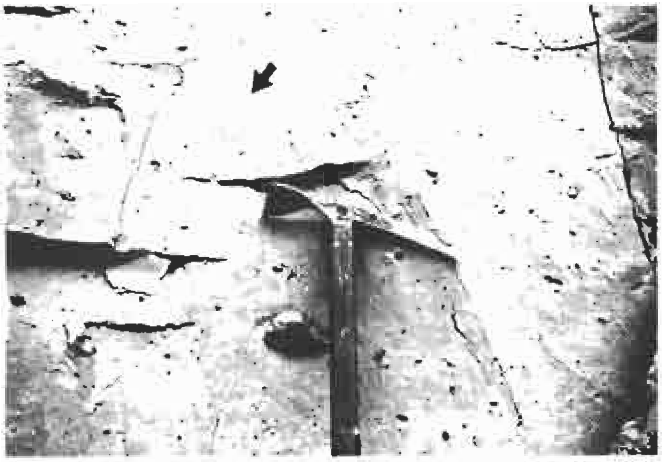
The geometry of the lithofacies asso-





▲ Figure 11
Glaciolacustrine lithofacies associations of Late Pleistocene age exposed along the northern margin of the Lake Ontario Basin, Ontario (from Eyles et al., 1984).

◀ Figure 12
a) Diamict lithofacies produced by ice rafting and suspension deposition on basin floors. Rain-out refers to suspension deposition and ice rafting. By allowing for changes in the relative importance of end member processes, the model integrates a wide variety of diamict assemblages deposited on marine and lake floors away from ice-proximal environments (e.g., Fig. 8). Lithofacies codes as in Figure 15.



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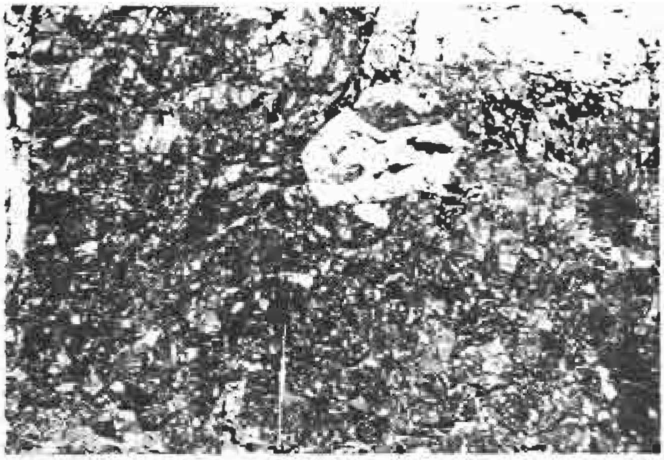

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Figure 12

b) Massive diamict with starved ripples (arrowed) indicating minor traction current activity during deposition. Late Pleistocene of the Great Lakes. c) Massive diamictite with dispersed clasts and occasional out-sized boulders. Scale is 1.5 m long. Late Proterozoic Gowganda Formation, Ontario (Fig. 19). d) Stratified diamictite with traction current deposited sand stringers penetrated and deformed by ice-rafted clasts. Late Proterozoic Port Askaig formation of Scotland (Fig. 18). e) Stratified sandy diamict resulting from suspension deposition, traction current activity and ice-rafting. Late Pleistocene, Irish Sea Basin (from Eyles and Eyles, 1984, Fig. 15). f) Lithofacies continuum of sands and sandy diamicts indicating episodic traction current activity, ice rafting and suspension deposition on a proglacial subaqueous fan (Fig. 8). Late Pleistocene, Irish Sea Basin (Fig. 15).

ciations in Figure 11 is distinctly different from that shown by lithfacies deposited under and at the margin of grounded ice sheets (Figures 1 and 4). Glaciolacustrine diamicts sit in conformable sequence context with deltaic or deeper water muddy lithofacies recording sedimentation on a low relief lake floor.

The architecture of the infills of other extensive lake bodies of Pleistocene age (e.g., Agassiz, Copper River Basin, many Great Lake Basins) still awaits documentation using surface and subsurface facies analysis techniques. Such data would be a significant aid to interpretation of the infills of intracratonic basins of similar dimensions dating from the Permo-Carboniferous glaciation of the Gondwana Supercontinent (e.g., Wopfner, 1972; Thornton, 1974; Harris, 1981). These provide examples of glacial overdeepening in continental interiors and the selective preservation of continental glacial facies. We suspect that many thick sequences in other Quaternary basins in Europe and North America interpreted as grounded ice deposits

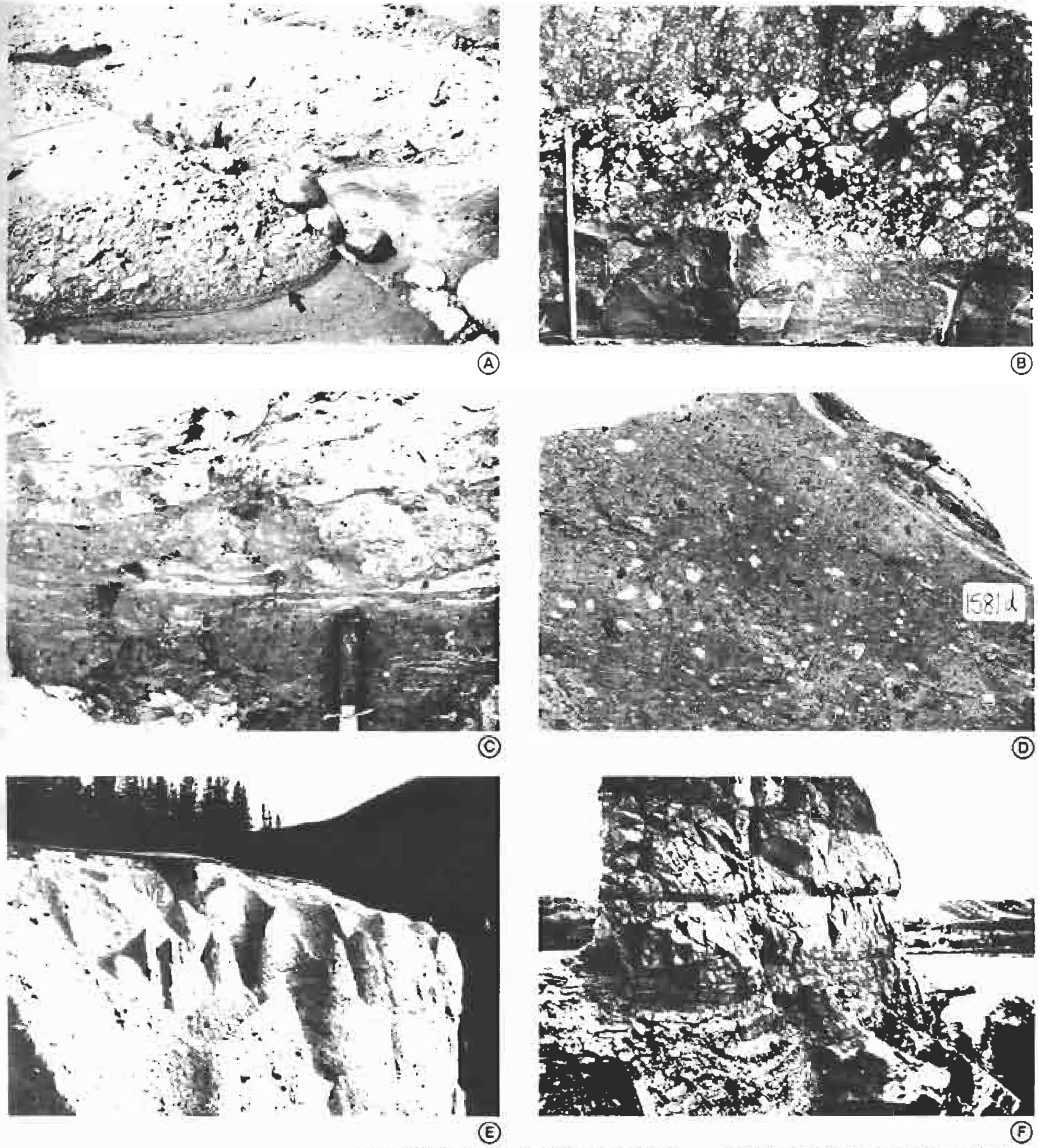


Figure 13
Resedimented subaqueous diamict lithofacies. a) Resedimented diamict lithofacies unit (2 m thick) showing inverse grading at base (arrowed) passing up into normally graded clast-supported diamict (at hammer level) and normally graded coarse sands. Early Pleistocene glacially-influenced submarine channel. Upper Yakataga Formation, Alaska. b) Diamictite deposited as a subaqueous debris flow. Note inverse grading at base and imbrication of large clasts. The flow rests on a massive to faintly bedded, well sorted sandstone of probable liquified/fluidized flow origin. Scale is 1 m long. Early Proterozoic Gowganda Formation, Ontario. c) Stratified, resedimented glaciolacustrine diamict with graded silty clay laminations at top left. The latter are turbidites and are genetically associated with subaqueous resedimented diamicts (Fig. 12). The section above the knife contains brecciated diamict and laminations. Late Pleistocene of Great Lakes. d) Cut slab of stratified resedimented glaciolacustrine diamict with abundant silt and clay clasts. Label is 2 cm long. e) 30 m section through 'paraglacial' fan; stratified diamicts and gravel lithofacies deposited by debris flow from valley sides during deglaciation. Late Pleistocene. Canadian Rockies. f) graded diamict units up to 1m thick recording downslope subaqueous resedimentation. Pleistocene Copper River Basin, Alaska.

are in fact of similar glaciolacustrine origin. Their stratigraphy reflects a very complex response to changing ice volumes, lake levels, and basin subsidence, and simplistic interpretations relating "till sheets" to climatic ice advance-retreat cycles need to be re-evaluated.

Cold Climate ('Periglacial') Structures and Facies

The term 'periglacial' is used in this section in its broadest sense identifying glacial and non-glacial cold climates. Definitions vary (Washburn, 1980).

Cold climate landscapes are characterised by aeolian activity, mass wasting processes, mechanical distrubance of surface sediments and rocks by ground icings and braided stream deposition. Periglacial structures and facies are most commonly associated with major unconformities, weathered zones and depostional hiatuses separating stratigraphic successions, and indicate subaerial exposure. Certain structures develop only under a restricted climatic range and identify 'permafrost' conditions where the ground remains perennially frozen to perhaps depths of 500 m or more, except for a seasonal surface thaw layer, under a mean annual air isothern of below -4 to -1° C.

Modern structures diagnostic of permafrozen ground are penetrative ice wedges, formed by continued refreezing of waters in open contraction cracks, and 'pingo' mounds formed essentially as hydrolaccoliths by the upwards injection of water along a migrating freezing front. Other structures include involutions and flame structures, formed as refreezing generates high pore water pressures in summer thaw layers (cryoturbation), and collapse structures that follow the irregular melt of ground ice (thermokarst). None of these has been systematically described in section using modern facies treatments and so the interpretation of former permafrost climates from the rock record is hazardous.

Many workers report polygonal networks of fossil ice wedge casts infilled by a variety of sediments, in Pre-Quaternary diamictite sequences (Hambrey and Harland, 1981; Deynoux, 1982; Fig. 14). Possible pingos up to 350 m in diameter have been identified from exhumed surfaces of Ordovician sandar of the central Sahara (Biju-Duval *et al.*, 1981). Whilst 'periglacial' structures are identified and used to infer subaerial continental permafrost climates, similar structures also occur as a result of subaqueous loading, liquefaction, thermal contraction, syneresis and desiccation (Black, 1976; Mills, 1983). Identification of rapidly deposited subaqueous facies, seismically active episodes in basin histories and associated cold climate continental facies is an aid to interpretation of 'periglacial' structures.

A distinctive periglacial facies is windblown silt (loess) derived from deflation of large braided stream networks. Extensive loess blankets are a marked feature of Quaternary periglaciated continental interiors (e.g., North America: 800,000 km²). Whereas total thicknesses of Quaternary loess approach 300 m in some areas of China, reports of loessite are few. Edwards (1979) reports Late Proterozoic loessite from Norway and Svalbard. There are many descriptions of the action of seasonal ice in rafting clasts, grooving and deforming soft sediments along temperate and cold climate shorelines (Dionne, 1974). These have a greater

Figure 14
a) Polygonal network of small sand-filled wedge structures in the Late Proterozoic Port Askaig Formation, Scotland, argued to be of permafrost ice-wedge origin by Spencer (1971) and discussed by Eyles and Clark, (1985). b) Intersection of large fossil sand-wedge polygons developed in diamictite; Late Proterozoic of Maurentiania, West Africa. Note v-shaped furrow between polygons marking a former thermal contraction wedge filled by wind transport in a cold climate (Deynoux, 1982).

preservation potential (Rattigan, 1967; Dalland, 1977).

The predominant periglacial facies are probably coarse-grained fluvial sediments recording seasonal braided stream activity and access to large volumes of coarse debris. Other important periglacial facies are stratified diamicts, resulting from a wide variety of mass movement processes as seasonal thawed surface layers become waterlogged and are resedimented downslope. These facies frequently infill the lowest points of the topography, and are interbedded with fluvial lithofacies resulting from slopewash. It should be noted that the term 'paraglacial' has been used to refer a shortlived period immediately following deglaciation of high relief glaciated valleys, when glacial facies are resedimented downslope as debris flows over valleyside alluvial fans to be reworked along the valley floor by braided streams (Ryder, 1971; Fig. 13). Rockfall and slide debris is an associated stratigraphic component. Glacial landforms and facies deposited by valley glaciers seldom survive this episode and glaciated valleys are preserved in the rock record as finger-like fiord troughs filled with coarse resedimented facies, braided stream, glaciomarine and glaciolacustrine sequences (e.g., Visser, 1982).

MARINE GLACIAL FACIES

Most classifications of glaciomarine sedimentation contrast proximal facies belts characterized by strong bottom currents and subaqueous fan growth with distal locations where nonglacial marine processes predominate.

Proximal Facies

A number of distinct lithofacies can be recognized within submarine ice-contact fans (Fig. 8). Cobbles and gravels accumulate at the fan apex (Rust and Romanelli, 1975) with cross-stratified sands being typical of the main fan body away from bifurcating channels (Powell, 1981). Massive, horizontally stratified and inversely graded gravels and sands accumulate within steep sided channels by sediment gravity flow as pulses of dense sediment laden meltwater sweep down the fan (Rust, 1977; Cheel and Rust, 1982). Surface subsidence and associated faulting results from the melt of buried ice.

A wide range of diamict lithofacies (with geometries that vary from drapes and lenses to channel fills), accumulate under depositional regimes of suspended sediment deposition and berg rafting ('rain-out'), and reworking by downslope sediment gravity flow and traction current activity (Figs. 8, 12 and 13). Coarse-grained stratified diamicts with a muddy sand matrix accumulate in areas of episodic-traction current activity and rain-out. These contain ice-rafted 'outsized' clasts, abundant traction current structures and rounded clasts rolled along as bedload, and occur as part of a lithofacies continuum with pebbly sands and poorly sorted gravels (Figs. 12 and 15). Coarse grained, angular, ice rafted debris in areas of high basement relief represents former supraglacial debris derived by rockfall from valleysides. Former basal debris components may be preserved as cohesive diamict pellets (Ovenshine,

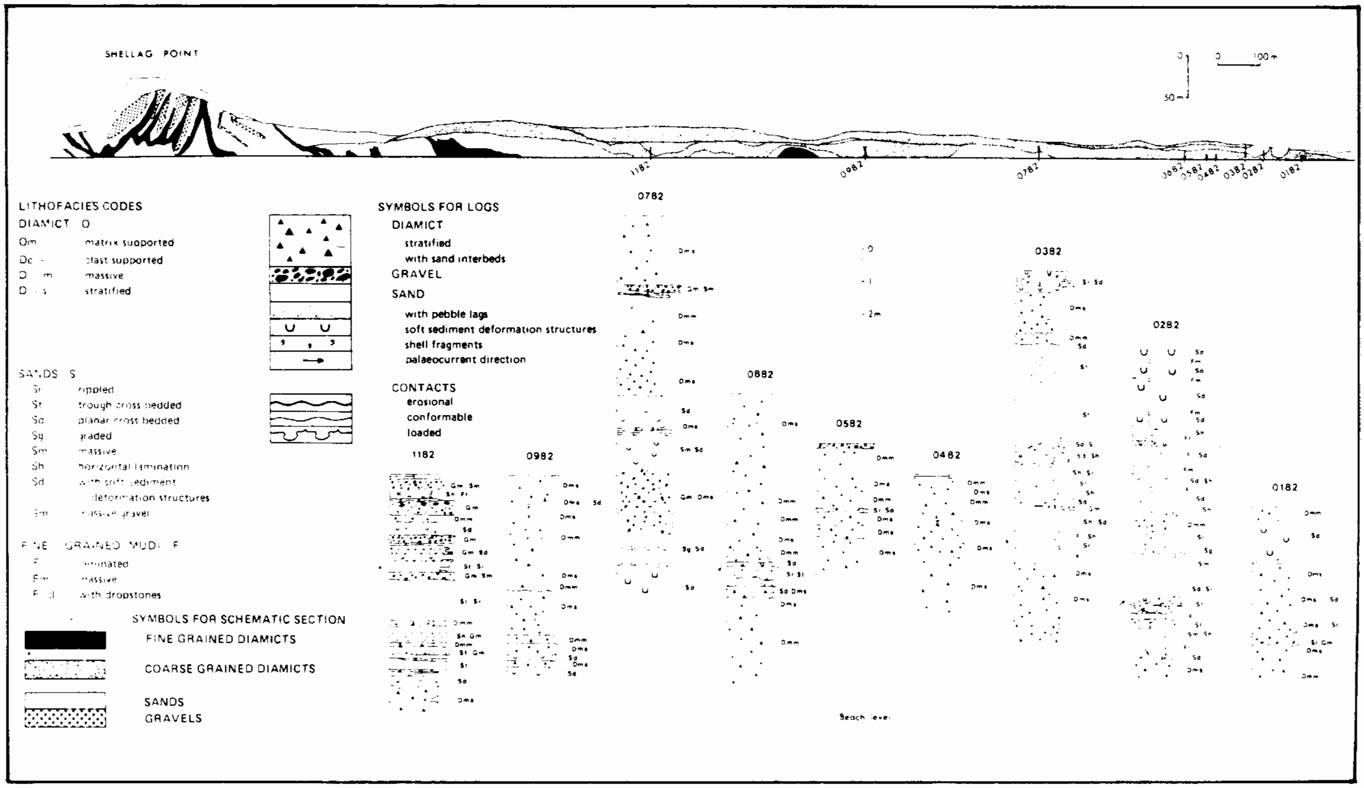


Figure 15

Lithofacies logs, from Quaternary glaciomarine sediments of the Irish Sea Basin, of sandy stratified diamicts draped over the back of a large push ridge (after Eyles and Eyles, 1984). Diamicts comprise lithofacies continuous with sand and gravel lithofacies showing ice-rafted clasts (Figs. 12e and 12f) indicating coeval traction current activity and diamict accumulation by suspension deposition and ice-rafting (Fig. 12a). Sedimentation probably occurred on a subaqueous fan (Fig. 8). Lithofacies codes from Eyles et al. *(1983).*

1970).

Mud belts occur where overflows or interflows dump large volumes of suspended sediment and where currents are sufficient to move ice bergs away from the ice margin (Powell, 1981, 1984; Elverhoi *et al.*, 1983). These muds may bury wedge-shaped fan accumulations as the ice margin retreats. Finely laminated and graded silty mud and sand lithofacies result from the repeated interaction of sediment plumes and tidal currents and comprise a very significant component of fiord fills (Mackiewicz *et al.*, 1984; Gilbert, 1983).

Downslope resedimentation of fan sediments is an important process. Gazdzicki *et al.*, (1982) describe a Pliocene example of sediment gravity flow deposition from the distal reaches of a subaqueous outwash fan merging with a basin plain from the South Shetland Islands. Deposition is dominated by slow sedimentation of mud from suspension interrupted by repeated incursion of pebbly sands and diamict by sediment gravity flow. The flows are defined by tabular or broadly lenticular geometries, matrix-supported fabric, laminated tops and sharp, sometimes erosive, bases. Processes range from debris flows through grain flows to turbidity currents. Similar sequences are described by Visser (1983) from a Permian fjord-head setting in the Karoo Basin of South Africa, and McCabe *et al.* (1984) from Late Pleistocene sequences in the Irish Sea Basin. Visser *et al.*, (1984) provide criteria for discriminating tectonic structures produced by compressional glacier overriding from those resulting from downslope (extensional) movement. Downslope resedimentation of subaqueous sediments is a common process (Nardin *et al.*, 1979) but it is the case that evidence for downslope resedimentation in glacial environments is still simplistically interpreted as indicating the release of debris direct from an ice margin.

Glacier retreat from the marine environment results in a fining-upward sequence with the upper part of the lithofacies sequences dominated by muds and fine-grained diamicts (Elverhoi *et al.*, 1983). However, isostatic emergence or rapid sediment accumulation may bring the sequence up above wave base, producing a coarsening-up beach or deltaic 'cap' (Andrews, 1978; Boulton and Deynoux, 1981; Boulton *et al.*, 1982).

Deposition from ice shelves represents a special case of proximal glaciomarine sedimentation. An ice shelf forms as a result of very high ice discharges into a marine environment with subsequent 'creep' thinning from a number of pinning points. Recent data show that large areas of ice shelves may undergo basal melting. This is of major importance because the bulk of basal debris is released close to the points at which the shelf begins to float. Consequently, icebergs calved from the ice front are fairly clean (Orheim and Elverhoi, 1981). In Antarctica, for example, ice shelves do not contribute significantly to continental margin sedimentation despite making up some 45% of the Antarctic coastline.

Many publications have maintained that extensive sequences of massive diamict(ite) in conformable sequence with marine sediments required continuous uninterrupted sedimentation from floating ice which, it was argued, could only be met below the closed cover of an ice shelf. A detailed facies model for the sub-ice shelf environment awaits documentation but a recent review (Eyles *et al.*, 1985) argues that the importance of such sedimentation in the rock record (as evidenced by the popularity of the ice shelf model) has been exaggerated beause of simplistic interpretation of massive diamictites. Such facies may be more likely to be of distal glaciomarine origin because sub-ice shelf sedimentation appears to be characterised by rapid exhaustion of basal debris, local deposition within fan-like accumulations (e.g., Drewry and Cooper, 1981) and widespread glaciotectonism resulting from migration of grounding lines. Lack of data from modern sub-ice shelf environments frustrates detailed facies modelling.

Away from ice-proximal environments, marine processes dictate the pattern of sediment accumulation and direct glacial influence is restricted to the supply of fine-grained suspended sediment, ice-rafted detritus and deformation and reworking by iceberg grounding (iceberg turbation; Vorren *et al.*, 1983). Sediment gravity flow and marine currents may be of considerable, even predominant, importance in generating the final depositional product.

There are problems in defining the outer limit of distal glaciomarine sedimentation which is commonly taken as the farthest extent of ice rafted debris. Definition is complicated because detritus is also rated by ice of various origins (e.g., seasonal ice on temperate and arctic coasts; see Piper, 1976; Dalland, 1977; Clarke *et al.*, 1980; and Andres and Matsch, 1983 for diagnostic criteria).

In mud belts where there is unrestricted supply of pelagic sediment, massive fine-grained diamicts accumulate where ice-rafted debris melts out (Miller, 1953; Ferrians, 1963; Miller, 1973; Plafker and Addicott, 1976; Armstrong, 1981). These show highly variable clast fabrics (Nystuen, 1976; Domack, 1983). Assemblages of massive and stratified diamicts result from spatial and/or temporal variation in the relative importance of: 1) pelagic rain-out and ice rafting, 2) traction current activity, and 3) downslope resedimentation (Figs. 12 and 13). Traction currents serve to winnow or supplement the fine-grained component derived from suspension. These currents vary from tidal, wind-driven and storm-generated currents on shallow marine shelves, contour currents of the upper continental slope, cold bottom currents released from ice shelves, thermohaline currents produced by the formation of pack ice, and mid-depth and bottom currents in abyssal plains. The effect of these is to maintain silt and clay in suspension, allowing selective deposition of coarser ice-rafted debris. Stratified 'residual' diamicts with a muddy sand matrix are widespread on the Antarctic continental shelf (Elverhoi and Roaldset, 1983; Elverhoi, 1984; Anderson *et al.*, 1983). The production of surface lags is promoted by iceberg turbation and re-suspension of fines (Vorren *et al.*, 1983). In areas of low current velocity silts may be deposited from waning traction currents to produce stratified and laminated diamicts with a silt-sized mode (Anderson *et al.*, 1982).

Subaqueously resedimented diamicts can be recognized by the variable presence of flow noses, flow banding and folding, creep structures, incorporated rafts of associated lithologies, frequent truncation of underlying lithologies, basal grooving and flutings (which can be confused with those associated with diamicts deposited by lodgement), abundant silt and clay clasts (Fig. 13) and preferential alignment of clasts. Silt and clay clasts are probably produced by brecciation of fine-grained laminated lithofacies during slumping, creep, subaqueous dewatering or sub-aerial exposure.

Lack of sorting, grading or stratification within resedimented diamict facies may indicate proximity to the flow source (c.f. "disorganized bed model for conglomerates", Walker, 1975). Increased internal sorting by either dispersive pressures, kinetic sieving, flow boundary effects or clay rheology, results in a downslope sequence of internal grading characteristics (Walker, 1975, and "Turbidites and Associated Coarse Clastic Deposits", this volume) and may ultimately result in the transformation of large and highly concentrated sediment gravity flows into turbidites (Nardin *et al.*, 1979; Wright and Anderson, 1982; Fig. 13).

Upper and lower contacts of glaciomarine (and glaciolacustrine) diamict assemblages with associated lithofacies are particluarly diagnostic (Figs. 15 and 18). Basal contacts are either interbedded over several centimetres to metres, recording episodic traction currents during suspension deposition, or are sharp, recording sudden changes in depositional regime. Contacts may also be transitional with underlying muds, showing increasing frequency of dropstones upwards in section. Upper contacts may be either loaded, or sharply eroded below shallowing-upwards marine sediments, frequently with a conglomeratic lag horizon. They may also be transitional to stone free muds or other fine grained lithofacies. Marine macro- or micro-fauna within a diamict(ite) may also help identification. The recognition of a conformable sequence context with marine sediments is critical. Diamicts formed by ice rafting and suspension deposition may also overlie with a sharp contact diamicts deposited in the grounded ice environment, recording flooding of the depositional site. A good example of the analysis of lithofacies relationships is provided by Mode *et al.* (1983) who also show how paleoecological and age data in Quaternary glaciomarine sediments can be integrated into a depositional model.

GLACIATED CONTINENTAL MARGIN DEPOSITIONAL SYSTEMS

Typical glaciomarine facies assemblages are probably best considered within the framework of a number of glaciated continental margin depositional systems. The latter can be regarded as being composed of glacially influenced varieties of the typical marine depositional systems, e.g., continental shelf, canyon, slope, rise and abyssal plain. The facies assemblages in these settings (described elsewhere in this volume), are modified by the two major glacial influences of sediment supply and sea level change. Abundant coarse debris and suspended sediment are fed into glacially influenced marine basins by meltwaters, by sediment gravity flows and from icebergs. Sea level changes result from variations in the volume of continental ice, and from the isostatic response to the ice load. Andrews (1978) and Boulton *et al.* (1982) show how the amplitude of these two effects varies diachronously with distance from the centres of ice loading, resulting in complex migration of facies belts across the depositional basin. Thus there are severe problems involved in the interpretation of multiple sequences of glaciomarine diamicts, muds and shallow marine sands in terms of synchronous climatically-driven glacier advance retreat cycles. Direct glacial influence in the marine environment is secondary to a number of other controls such as sea-level fluctuations and basin subsidence which are of pre-eminent significance in controlling alternations of diamict and other marine facies.

Continental Shelf Depositional System

Glacially modified continental shelves are underlain by subhorizontal, blanket-like units (e.g., Damuth, 1978; Josenhans, 1983) although this simple geometry may be complicated by basement tectonics or growth faults. A vigorous interglacial shelf circulation, scour by storms and grounded ice, and channeling by melt streams may reduce preserved thicknesses and generate widespread erosion surfaces and unconformities within stratigraphic successions (Fig. 16). A good example is the Quaternary infill of the current North Sea Basin where the sediment pile is

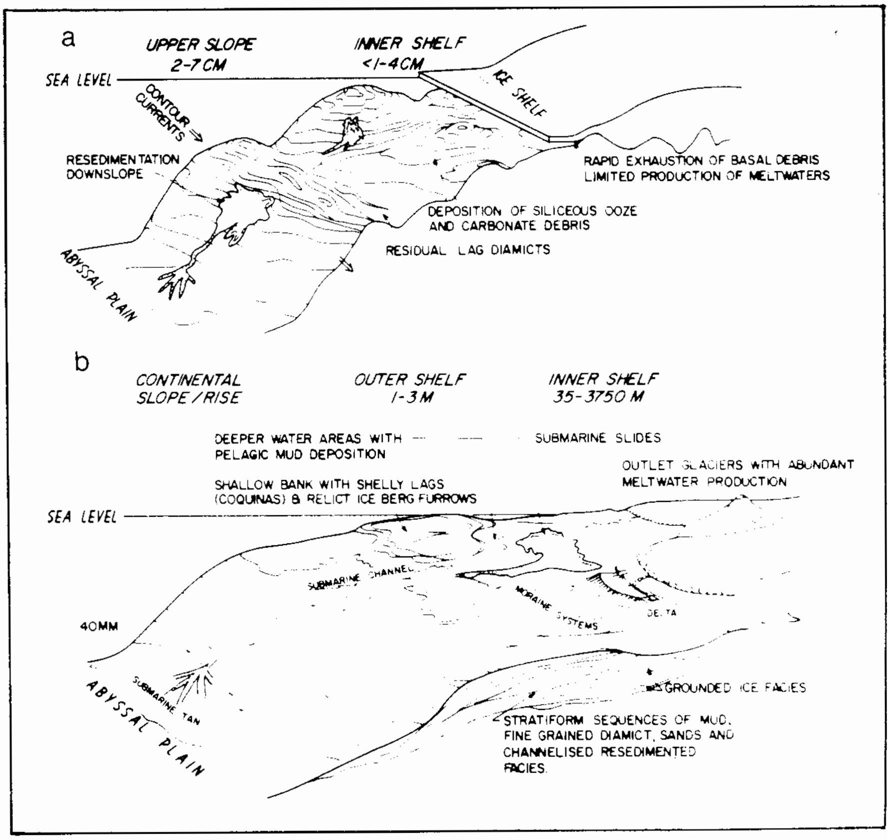
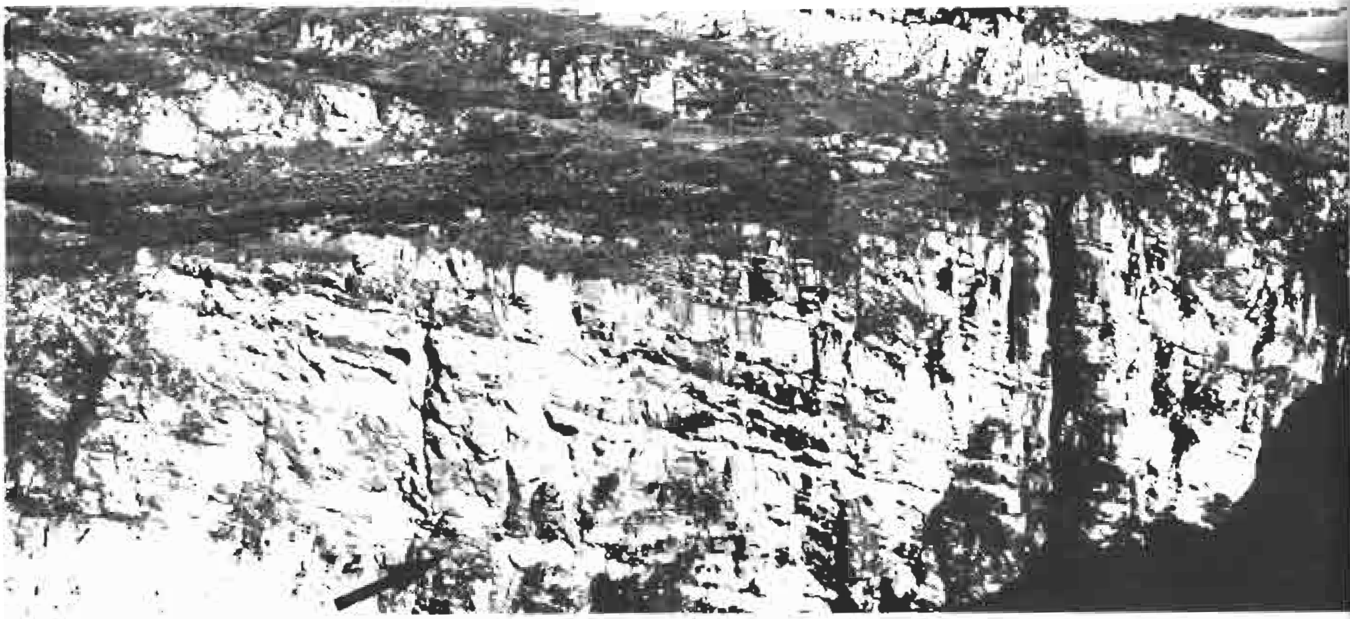


Figure 16

Contrasting types of glacially-influenced continental margin with reported deposition rates. a) 'sediment-starved' margin typical of Antarctica and other high relief margins (e.g., fiords) where cross-shelf transport is limited. b) glaciated continental margin with unrestricted sediment supply, e.g., Gulf of Alaska. Figures refer to average deposition rates per 1000 years.



(A)

(B)

Figure 17

a) Planar units of interbedded siltstone with dropstones, pebbly and sandy dolomites and diamictites deposited on a shallow marine shelf. Regional dip to right. Boudinage and necking structures (arrowed) indicate repeated downslope movement (Disrupted Beds and Great Breccia, Fig. 18). Late Proterozoic Port Askaig Formation, Scotland. Section about 35 m. b) Coquina bank of mollusc debris within glaciomarine diamicts deposited by suspension deposition and ice rafting (Fig. 12) of the Yakataga Formation, Alaska. Photograph courtesy of B. Kaye.

over 1 km thick as a result of continued subsidence (Caston, 1977).

The Late Miocene to present history of the Gulf of Alaska provides an excellent modern analog for glacially influenced shelf sedimentation. Adjacent glaciers are wet-based and deliver large volumes of suspended sediment to the Gulf. The tectonic setting of the area is a convergent plate margin, and continued uplift of an accretionary arc complex has exposed excellent sections through the 5 km thick Yakataga Formation which contains the most complete record of Late Cenozoic glacial activity in the world. Seismic traces show that the continental shelf is underlain by planar units of marine diamict formed by ice rafting and suspension deposition in water depths between 15 and 100 m (Plafker and Addicott, 1976; Molnia and Carlson, 1978) with associated muds, submarine channel sequences and coquina bands (Figs. 13, 16 and 17). The preservaiton of extensive coquina bands within the Yakataga Formation demonstrates that biogenic carbonates may be an integral part of Phanerozoic glaciomarine sequences. Cold water biogenic carbonates accumulate in many mid and high latitude areas at the present day (Björlykke *et al.*, 1978; Rao, 1981; Elverhoi and Roaldset, 1983).

Similar diamictite lithofacies with extensive planar geometries are well exposed over 700 km of strike in Scotland and Ireland within the 850 m thick Port Askaig Formation of the Late Proterozoic Dalradian Supergroup. Diamictite lithofacies are either massive or stratified, with traction current deposited interbeds containing dropstones (Figs. 12 and 18). Many diamictite surfaces are loaded by shallow marine conglomeratic lags or cross-bedded shallow marine sands.

Glacial influence on continental shelves may also restrict sediment supply and result in clastic starvation. The present day Antarctic continental shelf is the best studied example. Overdeepened submarine basins and fiords act as sediment traps and extreme cold prohibits meltwater transport of suspended sediment. Rapid loss of debris from basal melting ice shelves and isostatic subsidence below the Antarctic Ice Sheet also restricts deposition to landward margins, and limits cross shelf transport of sediment. As a result of sediment starvation, current-swept lag surfaces of relict ice-rafted debris occur over large areas. Siliceous oozes and biogenic carbonate accumulations (coquinas, foraminiferal debris, etc.) testify to low rates of clastic deposition. Significant sediment transport is limited to circum-continental contour current activity at the shelf break and sediment gravity flow processes in areas of irregular topography whereby older glaciomarine sediments are moved into slope

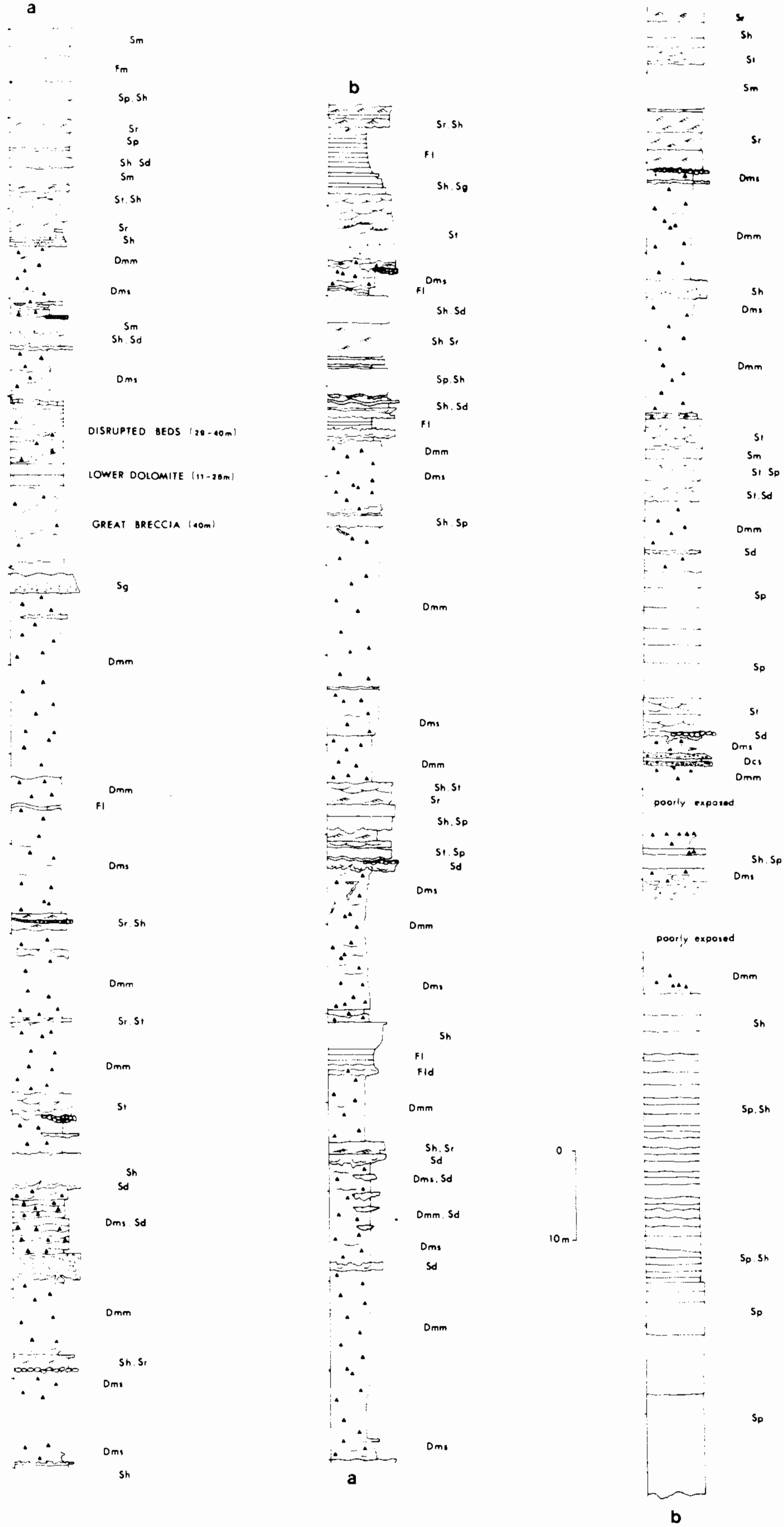


Figure 18
500 m lithofacies log from Late Proterozoic Port Askaig Formation of Scotland. Diamictites are conformably interbedded with shallow marine sandstones with a planar regional geometry (Fig. 17). Diamictite lithofacies have sharp, non-erosive lower contacts with dolomitic sandstones and loaded upper surfaces, consistent with distal glaciomarine deposition (Fig. 12d) on a shallow marine shelf undergoing episodic subsidence and subaerial exposure (Fig. 14a). The contrast in lithofacies sequences deposited by grounded glacier ice (Figs. 1 and 4) can be emphasized. From Eyles et al. *(1985).*

canyon and fan systems. Significant growth of slope canyon and fan systems only occurs during major glaciation when glaciers discharge directly into canyon systems or over the shelf-slope break (Anderson *et al.*, 1983).

The role of regional glaciation in producing sediment-starved continental shelves may have some bearing on the origin of mixed diamictite/carbonate successions. For example, the association of dolomite interbeds and cap rocks with Late Proterozoic diamictites, widely assumed to require either low latitude glaciation or rapid climatic change, remains a major problem following intense discussion in the mid-seventies (e.g., Schermerhorn, 1975). However, identification of detrital rather than primary dolomites (Fairchild, 1983; Fig. 17), recognition of the importance of diagnetic dolomitization of carbonates deposited on clastic-starved continental shelves following rapid post-glacial sea level rise (Deynoux and Trompette, 1976), and possible models of cold water dolomitisation all offer alternate cool climate mechanisms for dolomite formation. The debate has probably been defined too narrowly in the past because diamictite origins were ascribed simply either to glacial (from grounded ice sheets) or non-glacial origins (e.g., tectonically-triggered mudflows: Schermerhorn, 1974). Growing evidence of the importance of diamictite accumulation by ice-rafting and pelagic deposition on glacially-influenced marine shelves and reassessment of paleolatitude data (Stupavsky *et al.* 1982) broadens the scope for investigation.

Continental Slope, Rise and Abyssal Plain Depositional Systems

Modern slopes are cut by submarine canyons, which funnel sediment down to the continental rise and abyssal plain. Canyons may be enlarged to depths of up to 2 km by submarine mass wasting, or become plugged by sediment depending on the rate of sediment supply. Climatic changes during glaciation cause rapid changes in sea level, and may result in grounded ice extending far out on to the shelf to feed coarse debris directly to the outer shelf edge and canyon systems, e.g., northern Gulf of Alaska, Gulf of St. Lawrence, Antarctica (Piper *et al.*, 1973; Stow, 1981; Carlson *et al.*, 1982; Piper and Normark, 1982; Tucholke and Laine, 1982; Anderson *et al.*, 1983).

We cannot yet generalize about facies models for these glacially-influenced continental margin environments. A preponderance of submarine fans cut by extensive channel complexes is expected to comprise the major slope and rise facies assemblages. In the ancient record the only unequivocal way to distinguish such deposits from coarse non-glacial submarine fans may be the identification of ice-rafted debris (dropstones, diamict clots) in inter-fan argillites, and striated and facetted clasts within the channel fills. Parts of the Gowganda Formation (early Proterozoic) of northern Ontario, the Dwyka "Tillite" (Permian) of South Africa (Visser 1982) and the Yakataga Formation (Miocene to Recent) of the Gulf of Alaska show such facies assemblages (Figs. 10, 12, 13 and 19).

The Gowganda Formation is a world famous "glacial" unit, traditionally interpreted as recording repeated episodes of subglacial sedimentation below grounded ice sheets and varved glaciolacustrine deposition (Lindsey, 1969, 1971). However, application of modern facies analysis methods has demonstrated that most of the formation is the product of submarine sedimentation on a continental margin (Miall, 1983, 1985; Eyles *et al.*, 1985). Sediment gravity flow and submarine fan models (Walker, this volume) are particluarly pertinent.

Many varieties of sediment gravity flows can be identified, including clast-rich bouldery debris flows, sandy fluidized/liquefied flows and thin-bedded Bouma-type turbidites (Fig. 10). These probably were derived by local slope failure of basin-floor accumulations of ice-rafted debris and muds (Fig. 12). Sandy sediment gravity flows may have been produced by winnowing of debris flows by the processes identified from the modern continental slope of Antarctica by Wright and Anderson (1982) or by resedimentation of beach deposits. Massive diamictite, with dispersed unsorted clasts (Fig. 12), is volumetrically the most abundant lithofacies and probably represents relatively rapid deposition of ice-rafted detritus and pelagic mud. Where the iceberg density was low, possibly as a result of dispersal by winds or marine currents, or because of climatic amelioration or where suspended sediment input to the marine environment was high (e.g., the situation described from the modern Gulf of Alaska by Molnia and Carlson, 1978), pelagic muds predominate over coarse ice-rafted debris. The result is argillite containing thin laminae of rafted silt and sand grains (commonly reworked into ripples), thin turbidites, and scattered dropstones (Fig. 19d).

The stratigraphic architecture of the Gowganda Formation can locally be constructed from networks of diamond drill holes (Fig. 19). Miall (1983) described a fining-upward megacycle interpreted as the deposit of a subaqueous outwash fan. Elsewhere a submarine channel, filled with laminted argillites and winnowed sandstones, can be recognized (Miall, 1985). Low-angle crossbed sets 4 to 5 m thick consist of well sorted sandstone and lenses of diamictite, and are interpreted as subaqueous point bars formed within meandering fan channels (Fig. 19).

Glacially-influenced abyssal plain sedimentation has not yet been recognised and a detailed facies model has yet to be established. The presence of extensive planar bodies of turbiditic silty clays and massive muds with ice-rafted clastics has been emphasized by Clark *et al.*, (1980) and Goldstein (1983) in studies of the Arctic Ocean Basin.

FUTURE PROSPECTS

Most problems in the interpretation of glacial sequences will revolve around the origin of diamict(ite) units. Diamict(ite)s continue to be simplistically described as till(ite)s often with the implicit assumption that the direct agent of deposition was glacier ice. The assumption that massive diamict(ite) necessarily indicates subglacial deposition from grounded glacier ice can also still be found in the literature; the student should be aware that massive lithofacies do not uniquely characterise subglacial environments. Detailed field examination, laboratory slabbing, thin sectioning and x-ray analysis frequently demonstrates that 'massive' diamicts reveal structures that aid diagnosis (e.g., Fig. 12b). Diamicts deposited at an ice base for example may be massive over short core or outcrop lengths but exhibit distinct structures (Figs. 1 and 4). There are now sufficient diagnostic criteria recognised to identify diamict lithofacies deposited at an active or stagnant glacier base by virtue of substrate

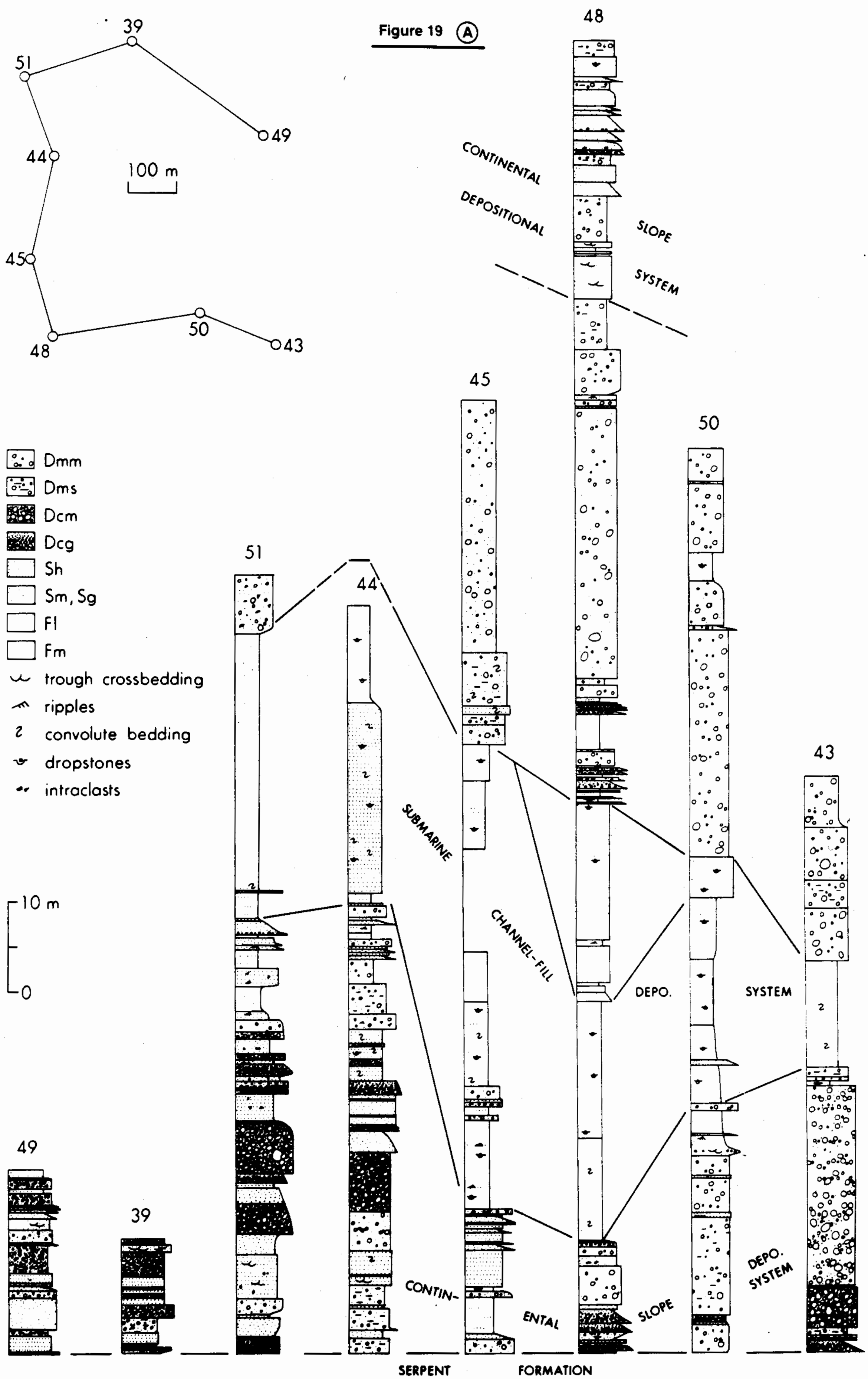


Figure 19 (A)
39
51
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100 m
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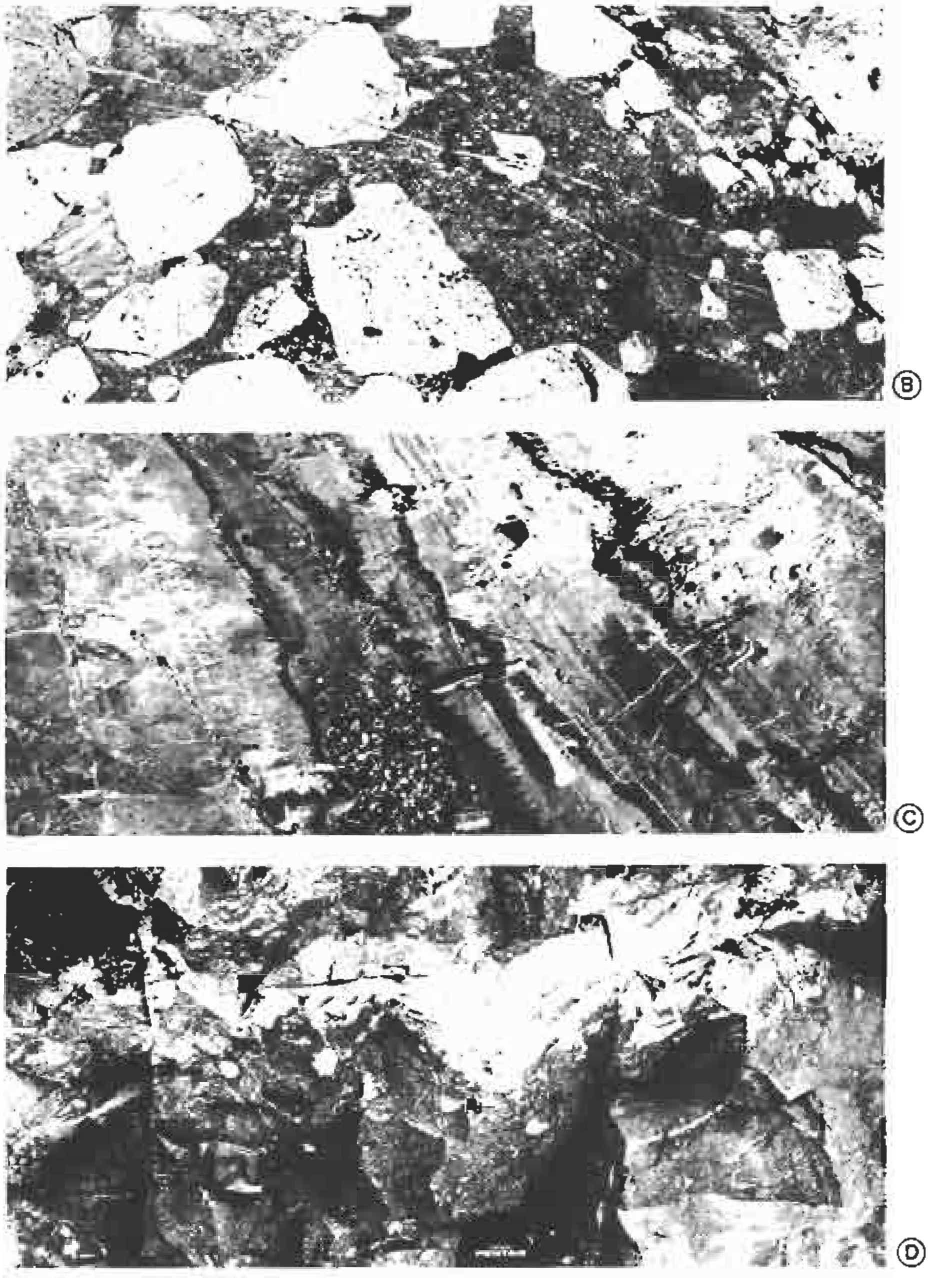

Figure 19
Lithofacies and structures from the Gowganda Formation, Ontario. a) Lithofacies logs through the lower part of the Gowganda Formation, from diamond drill holes near Elliot Lake, Ontario (Miall, 1985). b) Faceted boulders in the Gowganda Formation shaped by subglacial ice abrasion (Fig. 6) and resedimented as a debris flow. Near Elliot Lake. c) Grooves formed by clast rolling or scour by floating ice in a succession of argillites with thin-bedded turbidites. Cataract Falls, near Mississagi. d) Massive diamictite, probably deposited by ice-rafting and pelagic mud deposition (Fig. 12), containing wispy sandstone lenses. These were formed by current sorting or turbidity currents and have been deformed by loading into the soft substrate. Highway 108, near Elliot Lake.

deformation, associated lithofacies originating when the glacier moved over the site (Figs. 1 and 4), restricted sequence thickness across the depositional basin, and association with major regional unconformities. Glacitectonic structures may be absent locally at the scale of individual outcrops but will be clearly evident as data are assembled from adjacent sites.

Outside grounded ice environments, thicker diamict sequences are deposited in glacially-influenced environments (e.g., lacustrine, marine). These diamicts commonly occur therefore in conformable sequence context with lithofacies charateristic of these environments and, as a result the analysis of lithofacies relationships outlined by Walker ("Introduction" to this volume) is a major aid to interpretation. During the construction of lithofacies logs (e.g., Figs. 1, 4, 15, 18 and 19), the description of diamict and upper lower boundary relations with other lithofacies is of major importance. A critical need now exists for detailed field descriptions of modern and ancient glacial sequences from a range of environments and tectonic settings, using lithofacies criteria and facies analysis techniques.

A most important consideration with regard to future sedimentological investigations must be that most ancient glacial sequences accumulated and were preserved in oceanic and continental shelf marine environments that were influenced not by direct glaciation, but by the indirect glacial influence of sea-level and oceanographic change, ice-rafting and greatly enhanced sediment supply. In the Late Proterozoic for example there is a very clear association between diamictite sequences, continental rifting and the development of Atlantic-type trailing edge plate margins dominated by subsidence tectonics (Eyles *et al.*, 1985). In most cases continental glacial facies around basin margins were not preserved. As a result the data base from modern and Quaternary environments, overwhelmingly biased towards observations of continental glacial facies, is unrepresentative of much of the glacial rock record. However, knowledge of the offshore geology of continental margins affected by Late Cenozoic glaciations is expanding rapidly using sonar, seismic drill-core and down-hole data. If, as argued in this review, glaciomarine environments are regarded as special sub-types of the principal continental margin environments (e.g., abyssal plain, slope etc.) then this growing body of information from offshore exploration forms a firm foundation for sophisticated 'glacial' facies models applicable to modern and ancient sequences.

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REFERENCES

The following reading list emphasizes the growing stimulus to integrated facies models encompasing both modern and ancient examples, provided by recent process and subsurface data from modern glacial-influenced continental margins. It is this small but expanding literature that is of greatest application to the Earth's glacial record. References cited with regard to continental environments are examples where glaciological and hydrodynamic flow regime concepts have been integrated with descriptions of sedimentary products and basin studies.

CONTINENTAL GLACIAL FACIES

Andrews, J.T., 1982. On the reconstruction of Pleistocene ice sheets: a review: Quaternary Science Reviews, v. 2, p 1-30.

Boulton, G.S., 1972. Modern arctic glaciers as depositional models for former ice sheets. Journal of the Geological Society of London, v. 127, p. 361-393.

Boulton, G.S., 1978. Boulder shapes and grain size distributions of debris as indicators of transport paths through a glacier and till genesis. Sedimentology, v. 25, p. 773-799.

Cohen, J.M., 1983. Subaquatic mass flows in a high energy ice marginal deltaic environment and problems with the identification of flow tills. *In* Evenson E.B., *et al.*, eds., Tills and related deposits: Rotterdam, A.A. Balkema.

Eyles, C.H. and Eyles, N., 1983. Sedimentation in a large lake: a reinterpretation of the Late Pleistocene stratigraphy at Scarborough Bluffs, Ontario, Canada. Geology, v. 11, p. 146-152.

Eyles, N., 1979. Facies of supraglacial sedimentation on Icelandic and Alpine temperate glaciers. Canadian Journal of Earth Sciences, v. 16, p. 1341-1361.

Eyles, N., and Menzies, J., 1983. The sub glacial landsystem. *In* Eyles, N., ed., Glacial geology: an introduction for engineers and earth scientists. Oxford, Pergamon Press, p. 19-70.

Eyles, N., Sladen, J., and Gilroy, S., 1982. A depositional model for stratigraphic complexes and facies superimposition in lodgement tills. Boreas, v. 11, p. 317-333.

Eyles, N., *et al.*, 1985. The application of basin analysis techniques to glaciated terrains; an example from the Lake Ontario Basin. Geoscience Canada, v. 12, p. 22-32.

Gibbard, P., 1980. The origin of stratified Catfish Creek till by basal melting. Boreas, v. 9, p. 71-85.

Jorgensen, N.B., 1982. Turbidites and associated resedimented deposits from a tilted glaciodeltaic sequence, Denmark. Damn. Geol. Undrs., Arbog 1981, p. 47-72.

Jopling, A.V. and McDonald, B.C., eds., 1975. Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, 320 p.

Lawson, D.E., 1982. Mobilisation, movement and deposition of active subaerial sediment flows, Matanuska Glacier, Alaska. Journal of Geology, v. 90, p 279-300.

Miall, A.D., 1977. A review of the braided river depositional environment. Earth Science Reviews, v. 13, p. 1-62.

Moran, S.R., Clayton, R., Hooke, R. Le B., Fenton, M.M., and Andriashek, L.D. 1980. Glacier bed landforms of the Prairie region of North America. Journal of Glaciology, v. 25, p. 457-476.

Menzies, J., 1979. A review of the literature on the formation and location of drumlins. Earth Science Reviews, v. 14, p. 315-359.

Paterson, W.B., 1981. The physics of glaciers. Oxford, Pergamon Press.

Paul, M.A., 1983. The supraglacial landsystem. *In* Eyles, N., ed., Glacial geology: an introduction for engineers and earth scientists. Oxford, Pergamon Press, p. 71-90.

Postma, G., Roep, T.R., and Ruegg, G.H.J., 1983. Sandy gravelly mass flow deposits in an ice marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, the Netherlands) with emphasis on plug-flow deposits. Sedimentary Geology, v. 34, p. 59-82.

Schlüchter, C.H., ed., 1978. Moraines and varves. Rotterdam, A.A. Balkema.

Shaw, J., 1977a. Sedimentation in an alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada. Geografiska Annaler, v. 59A, p. 221-240.

Shaw, J., 1977b. Tills deposited in arid polar environments. Canadian Journal of Earth Sciences, v. 14, p. 1239-1245.

Shaw, J., 1979. Genesis of the Sveg tills and Rogen moraines of central Sweden; a model of basal melt-out. Boreas, v. 8, p. 409-426.

Symposium on Glacier Beds, 1979. The ice-rock interface. Journal of Glaciology, v. 23.

GLACIOMARINE: ICE PROXIMAL FACIES

Cheel, R.J. and Rust, B.R., 1982. Coarse-grained facies of glaciomarine deposits near Ottawa, Canada. *In* Davidson-Arnott, R., *et al.*, eds., Research in glacial, glacio-fluvial and glacio-lacustrine systems. Norwich, Geo-Books, p. 279-295.

Domack, E.W., 1983. Facies of Late Pleistocene glacial-marine sediments on Whidbey Island, Washington: an isostatic glacial-marine sequence. *In* Molnia, B.F., ed., Glacial-marine sedimentation: New York, Plenum Press, p. 535-570.

Elverhoi, A., Lonne, O., and Seland, R., 1983. Glaciomarine sedimentation in a modern fjord environment, Spitsbergen. Polar Research, v. 1, 127-149.

Eyles, C.H. and Eyles, N., 1984. Glaciomarine sediments of the Isle of Man as a key to Late Pleistocene stratigraphic investigations in the Irish Sea Basin. Geology, v. 12, in press.

Gilbert, R., 1983. Sedimentary processes of Canadian Arctic fyords. Sedimentary Geology, v. 36, p. 147-175.

Mackiewicz, N.E., Powell, R.D., and Molnia, B.F., 1984. Interlaminated Ice-Proximal glaciomarine sediments in Muir Inlet, Alaska. Marine Geology, v. 57, p. 113-147.

McCabe, A.M., Dardis, G.F. and Hanvey, P.M., 1984. Sediments of a late Pleistocene submarine moraine complex, County Down, Northern Ireland. Journal of Sedimentary Petrology, v. 54, p. 716-730.

Powell, R.D., 1981. A model for sedimentation by tidewater glaciers. Annals of Glaciology, v. 2, p. 129-134.

GLACIALLY-INFLUENCED CONTINENTAL MARGIN SEDIMENTATION

Anderson, J.B., Kurtz, D.D., and Weaver, F.M., 1982. Sedimentation on the West Antarctic continental margin. *In* Craddock, C., ed., Antarctic geoscience. Madison, University of Wisconsin, p. 1003-1012.

Anderson, J.B., Brake, C., Domack, E.W., Meyers, N., and Wright, R., 1983. Development of a polar glacial-marine sedimentation model from Antarctic Quaternary deposits and glaciological information. *In* Molnia, B.F., ed., Glacial-marine sedimentation: New York, Plenum Press, p. 233-264.

Crowell, J., 1978. Gondwana glaciation, cyclothems, continental positioning and climate change. American Journal of Science, v. 278, p. 1345-1372.

Domack, E.W. 1982. Sedimentology of glacial and glacial marine deposits on the George V - Adelie continental shelf. East Antarctica. Boreas, v. 11, p. 79-97

Elverhoi, A., 1984. Glacigenic and associated marine sediments in the Weddell Sea, fyords of Spitzbergen and the Barents Sea; a review. Marine Geology, v. 57, p 53-88.

Elverhoi, A., and Roaldset, E., 1983. Glaciomarine sediments and suspended particulate matter. Weddell Sea Shelf, Antartica. Polar Research, v. 1, p. 1-21.

Eyles, N., Eyles, C.H., and Miall, A.D., 1983. Lithofacies types and vertical profile analysis; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. Sedimentology, v. 30, p. 393-410.

Eyles, C.H., Eyles, N., and Miall, A.D., 1985. Models of glaciomarine sedimentation and their application to the interpretation of ancient glacial sequences: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 51, p. 15-84.

Goldstein, R.H., 1983. Stratigraphy and sedimentology of ice-rafted and turbidite sediment, Canada Basin, Arctic Ocean. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 367-400.

Minicucci, D.A. and Clark, D.L., 1983. A Late Cenozoic stratigraphy for glacial-marine sediments of the Eastern Alpha Cordillera, Central Arctic Ocean. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 331-365.

Mode, W.N., Nelson, A.R., and Brigham, J.K., 1983. A facies model of Quaternary glacial-marine cyclic sedimentation along eastern Baffin Island, Canada. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 495-534.

Plafker, G. and Addicott, W.O., 1976. Glaciomarine deposits of Miocene through Holocene age in the Yakataga Formation along the Gulf of Alaska margin, Alaska. *In* Miller, T.P., ed., Recent and ancient sedimentary environments in Alaska. Alaska Geological Society, p. 1-22.

Powell, R.B., 1984. Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. Marine Geology, v. 57, p. 1-52.

Stow, D., 1981. Laurentian fan; morphology, sediments, processes and growth patterns. American Association of Petroleum Geologists. Bulletin, v. 65, p. 375-393.

Tucholke, B.E. and Laine, E.P., 1982. Neogene and Quaternary development of the Lower Continental Rise off the central U.S. East Coast. *In* Watkins, J.S., and Drake, C.L., eds., Studies in continental margin geology. American Association of Petroleum Geologists Memoir 34, p. 295-306.

Vorren, T.O., Hald, M., Edvardsen, M., and Lind-Hansen, O.D., 1983. Glacigenic sediments and sedimentary environments on continental shelfs; general principles with a case study from the Norwegian shelf. *In* Ehlers, J., ed., Glacial deposits in northwest Europe. Rotterdam, A.A. Balkema, p. 61-73.

Weaver, P.P.E., and Kuijpers, A., 1983. Climatic control on turbidite deposition on the Madeira Abyssal Plain. Nature, v. 306, p. 360-363.

Wright, R. and Anderson, J.B., 1982. The importance of sediment gravity flow to sediment transport and sorting in a glacial marine environment, Eastern Weddell Sea, Antarctica. Geological Society of America Bulletin, v. 93, p. 951-963.

EXAMPLES OF PRE-LATE CENOZOIC GLACIAL FACIES

Early Proterozoic

Lindsey, D.A., 1969. Glacial sedimentology of the Precambrian Gowganda Formation, Ontario, Canada. Geological Society of America Bulletin, v. 80, p. 1685-1702.

Lindsey, D.A. 1971. Glacial marine sediments in the Precambrian Gowganda formation at Whitefish Falls, Ontario (Canada). Palaeogeography, Palaeoclimatology, Paleoecology, v. 9, p. 7-25.

Miall, A.D., 1983. Glaciomarine sedimentation the Gowganda Formation (Huronian), northern Ontario. Journal of Sedimentary Petrology, v. 53, p. 477-491.

Miall, A.D., 1985. Sedimentation on an early Proterozoic continental margin; the Gowganda Formation (Huronian), Elliot Lake area, Ontario, Canada. Sedimentology, v. 32, p. 763-788.

Late Proterozoic

Bjorlykke, K., Elvsborg, A., and Hoy, T., 1976. Late Precambrian sedimentation in the central sparagmite basin of south Norway. Norsk Geol. Tidsskrift, v. 56, p. 233-290.

Christie-Blick, N., 1983. Glacial-marine and subglacial sedimentation, Upper Proterozoic Mineral Fork Formtion, Utah. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 703-776.

Deynoux, H., 1982. Periglacial polygonal structures and sand wedges in the Late Precambrian glacial formations of the Taouden Basin in Adrar of Mauretania (West Africa). Palaeogeography, Paleoclimatology, Palaeoecology, v. 39, p. 55-70.

Edwards, M.B., 1976. Sedimentology of Late Precambrian Sveanor and Kappe Sparre Formations at Aldorsbreen, Wahlenbergfjorden, Nordaustlandet. Norsk Polarinst. Arbok 1974, p. 51-61.

Eisbacher, G.H., 1981. Sedimentary tectonics and glacial record in the Windermere Supergroup, Mackenzie Mountains, Northwestern Canada. Geological Survey of Canada, Paper 80-27.

Eyles, C.H. and Eyles, N., 1983. A glaciomarine model for Late Precambrian diamictites of the Port Askaig Formation, Scotland. Geology, v. 11, p. 692-696.

Eyles, C.H., Eyles N., and Miall, A.D., 1985. Models of glaciomarine sedimentation and their application to the interpretation of ancient glacial sequences. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 51, p. 15-84.

Eyles, N., and Clark, B.M., 1985. Gravity-induced soft sediment deformation in glacier-marine sequences of the Upper Proterozoic Port Askaig Formation, Scotland. Sedimentology, v. 32, p 789-814.

Hambrey, M.J., 1982. Late Precambrian diamictites of northeastern Svalbard. Geological Magazine, v. 119, p. 527-551.

Link, P.K. and Gostin, V.A., 1981. Facies and paleogeography of Sturtian glacial strata (Late Precambrian), South Australia. American Journal of Science, v. 281, p. 353-374.

Nystuen, J.P., 1976. Facies and sedimentation of the Late Precambrian Moelv Tillite in the eastern part of the Sparagmite Region southern Norway. Norges Geologiske Undersokelse, v. 329, p. 1-70.

Schermerhorn, L.J.G., 1975. Tectonic framework of Late Precambrian supposed glacials. *In* Wright, A.E., and Moseley, F., eds., Ice ages: ancient and modern. Liverpool, Seel House Press, p. 241-274.

Spencer, A.M., 1971. Late Precambrian glaciation in Scotland. Geological Society of London, Memoir 6.

Spencer, A.M., 1975. Late Precambrian glaciation in the North Atlantic region. *In* Wright, A.E. and Moseley, F., eds., Ice ages: ancient and modern. Liverpool, Seel House Press, p. 217-240.

Schermerhorn, L.J.G., 1974. Late Precambrian mixtites: glacial and/or nonglacial? American Journal of Science, v. 274, p. 673-824.

Young, G.M., 1982. The late Proterozoic Tindir Group, east-central Alaska: evolution of a continental margin. Geological Society of America Bulletin, v. 93, p. 759-783.

Phanerozoic

Casshyap, S.M., and Qidwai, H.A., 1974. Glacial sedimentation of Late Paleozoic Talchir Diamictite, Pench Valley Coalfield, Central India. Geological Society of America, Bulletin, v. 85, p. 749-760.

Davis, R.A. and Mallett, C.W. 1981. Sedimentation in a Permian subglacial channel. Journal of Sedimentary Petrology, v. 51, p. 185-190.

de la Grandville, B.F., 1982. Appraisal and development of a structural and stratigraphic trap oil field with reservoirs in glacial and periglacial clastics. *In* Halbouty, M.T., ed., The deliberate search for the subtle trap. American Association of Petroluem Geologists, Memoir 32, p. 267-286.

Harris, W.K., 1981. Permian diamictites in South Australia. *In* Hambrey, H., and Harland, W.B., eds., Earth's Pre-Pleistocene glacial record. Cambridge University Press, p. 469-473.

Visser, J.N.J., 1983. Glacial-marine sedimentation in the Late Paleozoic Karoo Basin, Southern Africa. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 667-702.

Visser, J.N.J., Collinson, W.P., and Terblanche, J.C., 1984. The origin of soft sediment deformation structures and related deposits in Permo-Carboniferous glacial and proglacial beds, South Africa. Journal of Sedimentary Petrology, in press.

OTHER REFERENCES CITED IN TEXT

Andrews, J.T., 1978. Sea level history of Arctic coasts during the Upper Quaternary. Progress in Physical Geography, v. 2, p. 375-407.

Andrews, J.T. and Matsch, C.L., 1983. Glacial marine sediments and sedimentation: an annotated bibliography. Norwich, U.K., Geo-Books.

Armentrout, J.M., 1983. Glacial lithofacies of the Neogene Yakataga Formation, Robinson Mountains, Southern Alaska Coast Range, Alaska. *In* Molnia, B.F., ed., Glacial-marine sedimentation. New York, Plenum Press, p. 629-666.

Armstrong, J.E., 1981. Post-Vashon Wisconsin Glaciation, Fraser Lowland, British Columbia. Geological Survey of Canada, Bulletin 322.

Ashley, G.M., 1975. Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 304-320.

Banerjee, I., 1973. Sedimentology of Pleistocene glacial varves in Ontario, Canada. Geological Survey of Canada Bulletin 226.

Banerjee, I. And McDonald, B.C., 1975. Nature of esker sedimentation. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 132-154.

Biju-Duval, B., Deynoux, M., and Rognon, P., 1981. Late Ordovician tillites of the Central Sahara. *In* Hambey, M.J., and Harland, W.B., eds., Earth's Pre-Pleistocene glacial record. Cambridge University Press, p. 99-107.

Björlykke, K., Bue, B., and Elverhoi, A., 1978. Quaternary sediments in the northwestern part of the Barents Sea and their relation to underlying Mesozoic bedrock. Sedimentology, v. 25, p. 227-246.

Black, R.F., 1976. Perglacial features indicative of permafrost: ice and soil wedges. Quaternary Research, v. 6, p. 3-26.

Bluck, B.J., 1979. Structure of coarse grained braided stream alluvium. Transactions of the Royal Society of Edinburgh, v. 70, p. 181-221.

Boothroyd, J.C. and Ashley, G.M., 1975. Process, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 193-222.

Boothroyd, J.C. and Nummedal, D., 1978. Proglacial braided outwash: a model for humid alluvial-fan deposits. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists Memoir 5, p. 641-668.

Boulton, G.S., 1979. Processes of glacier erosion on different substrata. Journal of Glaciology, v. 23, p. 15-38.

Boulton, G.S. and Deynoux, M., 1981. Sedimentation in glacial environments and the identification of tills and tillites in ancient sedimentary sequences. Precambrian Research, v. 15, p. 397-422.

Boulton, G.S., Baldwin, C.T., Peacock, J.P., McCabe, A.M., Miller, G., Jarvis, J., Horsfield, B., Worsley, P., Eyles, N., Chroston, P.N., Day, T.E., Gibbard, P., Hare, P.E., and

Von Brunn, V., 1982. A glacio-isostatic facies model and amino-acid stratigraphy for late Quaternary events in Spitsbergen and the Arctic. Nature, v. 298, p. 437-441.

Carson, P.R., Bruns, T.R., Molnia, B.F., and Schwab, W.C., 1982. Submarine valleys in the northeastern Gulf of Alaska: characteristics and probable origin. Marine Geology, v. 47, p. 217-242.

Caston, V.N.D., 1977. A new isopachyte map of the Quaternary of the North Sea and Quaternary deposits of the Forties Field, northern North Sea. Institute of Geological Sciences, Report 77/11.

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

Clark, D.L.,Whitman, R.R., Morgan, K.A., and MacKay, S.D., 1980. Stratigraphy and glacial-marine sediments of the Amerasian Basin, Central Arctic Ocean. Geological Society of America, Special Paper 181.

Dalland, A., 1977. Erratic clasts in the lower Tertiary deposits of Svalbard - evidence of transport by winter ice. Norsk Polarinst. Arbok. 1976, p. 151-165.

Damuth, J.E., 1978. Echo character of the Norwegian-Greenland Sea: relationship to Quaternary sedimentation. Marine Geology, v. 28, p. 1-36.

Deynoux, M. and Trompette, R., 1976. Late Precambrian mixtites: glacial and/or nonglacial? Dealing especially with the mixtites of West Africa. American Journal of Science, v. 276, 1302-1324.

Deynoux, M. and Trompette, R., 1981. Late Precambrian tillites of the Taoudeni Basin, West Africa. *In* Hambrey, M.J., and Harland, W.B., eds., Earth's Pre-Pleistocene glacial record. Cambridge University Press, p. 123-131.

Dionne, J-C., 1974. Polished and striated mud surfaces in the St. Lawrence Tidal Flats. Canadian Journal of Earth Sciences, v. 11, p. 489-494.

Dreimanis, A., 1984. Lithofacies types and vertical profile analysis: comments on the paper by Eyles, N., Eyles, C.H., and Miall, A.D. Sedimentology, v. 31, p. 885-886.

Drewry, D.J., and Cooper, A.P.R., 1981. Processes and models of Antarctic glaciomarine sedimentation. Annals of Glaciology, v. 2, p. 117-122.

Edwards, M.B., 1979. Late Precambrian glacial loessites from North Norway and Svalbard. Journal of Sedimentary Petrology, v. 49, p. 85-92.

Eyles, N., Eyles, C.H., and Miall, A.D., 1984a. Lithofacies types and vertical profile models: reply to Dreimanis *et al.*, Sedimentology, v. 31, p. 891-898.

Fairchild, I.J., 1983. Effects of glacial transport and neomorphism on Precambrian dolomite crystal sizes. Nature, v. 304, p. 714-716.

Ferrians, O.J., 1963. Glaciolacustrine diamicton deposits in the Copper River Fan-Delta, Alaska. United States Geological Survey, Professional Paper 475 C, p. C120-C125.

Galloway, W.E., 1977. Sediments and stratigraphic framework of the Copper River Fan-Delta, Alaska. Journal of Sedimentary Petrology, v. 47, p. 726-737.

Gazdzicki, A., Gradzinski, R., Porebski, S., and Wrona, R., 1982. Pholadid *Penitella* borings in glaciomarine sediments (Pliocene) of King George Isalnd, Antarctica. N. Jb. Geol. Palaont. Mh., v. 12, p. 723-735.

Hambrey, M.J. and Harland, W.B., eds., 1981. Earth's Pre-Pleistocene glacial record. Cambridge University Press.

Josenhans, H.W., 1983. Evidence of pre-late Wisconsinian glaciations on Labrador Shelf - Cartwright Saddle. Canadian Journal of Earth Sciences, v. 20, p. 225-235.

Lambert, A., and Hsu, K.J., 1979. Non-annual cycles of varve-like sedimentation in Walensee, Switzerland. Sedimentology, v. 26, p. 453-461.

Lindsey, D.A., 1969. Glacial sedimentology of the Precambrian Gowganda Formation, Ontario, Canada. Geological Society of America Bulletin, v. 80, p. 1685-1702.

Miall, A.D., 1977. A review of the braided river depositional environment. Earth Science Reviews, v. 13, p. 1-62.

Miall, A.D., 1978. Lithofacies types and vertical profile models in braided rivers: a summary. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists Memoir 5, p. 597-604.

Miller, J.D., 1953. Late Cenozoic marine glacial sediments and marine terraces of Middleton Island, Alaska. Journal of Geology, v. 61, p. 17-40.

Miller, R.D., 1973. Gastineau Channel Formation, a composite glaciomarine deposit near Juneau, Alaska. United States Geological Survey, Bulletin 1394C.

Mills, P.C., 1983. Genesis and diagnostic value of soft-sediment deformation structures - a review. Sedimentary Geology, v. 35, p. 83-104.

Molnia, B.F. and Carlson, P.R., 1978. Surface sedimentary units of Northern Gulf of Alaska continental shelf. American Association of Petroleum Geologists, Bulletin, v. 62, p. 633-643.

Nardin, T.R., Hein, F.J., Gorsline, D.S., and Edwards, B.D., 1979. A review of mass movement processes, sediment and acoustic characteristics and contrasts in slope and base of slope systems versus canyon-fan-basin floor systems. *In* Doyle, L.J., and Pilkey, O.H., eds., Geology of continental slopes. Society of Economic Paleontologists and Mineralogists, Special Publication 27, p. 61-73.

Ostrem, G., 1975. Sediment transport in glacial meltwater streams. *In* Jopling, A.V., and McDonald, B.C., eds. Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 101-122.

Ovenshine, A.T., 1970. Observations of iceberg rafting in Glacier Bay, Alaska and the identification of ancient ice-rafted deposits. Geological Society of America Bulletin, v. 81, p. 891-894.

Piper, D.J.W., Von Huene, R., and Duncan, J.R., 1973. Late Quaternary sedimentation in the active Eastern Aleutian Trench. Geology, v. 1, p. 19-22.

Piper, D.J.W., 1976. The use of ice rafted marine sediments in determining glacial conditions. La Revue de Geographie de Montreal, v. 30, p. 207-212.

Piper, D.J.W., and Normark, W.R., 1982. Acoustic interpretation of Quaternary sedimentation and erosion on the channelled upper Laurentian Fan, Atlantic margin of Canada. Canadian Journal of Earth Sciences, v. 19, p. 1974-1984.

Rao, C.P., 1981. Cementation in a cold-water bryozoan sand, Tasmania, Australia. Marine Geology, v. 40, p. 23-33.

Rattigan, T.H., 1967. Depositional, soft sediment and post-consolidation structures in a Palaeozoic aqueoglacial sequence. Journal of the Geological Society of Australia, v. 14, p. 5-18.

Rust, B.R., 1972. Structure and process in a braided river. Sedimentology, v. 18, p. 221-246.

Rust, B.R., 1977. Mass flow deposits in a Quaternary succession near Ottawa, Canada; diagnostic criteria for subaqueous outwash. Canadian Journal of Earth Sciences, v. 14, p. 175-184.

Rust, B.R. and Romanelli, R., 1975. Late Quaternary subaqueous outwash deposits near Ottawa, Canada. *In* Jopling, A.V., and McDonald, B.C. eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 172-192.

Ryder, J.M., 1971. The stratigraphy and morphology of paraglacial alluvial fans in south-central British Columbia. Canadian Journal of Earth Sciences, v. 8, p. 279-298.

Stupavsky, M., Symons, D.T.A., and Gravenor, C.P., 1982. Evidence for metamorphic remagnetisation of Upper Precambrian tillite in the Dalradian Supergroup of Scotland. Royal Society of Edinburgh Transactions, Earth Sciences, v. 73, p. 59-65.

Shaw, J., 1975. Sedimentary successions in Pleistocene ice-marginal lakes. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 281-303.

Teller, J.T., and Clayton, L., eds., 1983. Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

Thornton, R.C.N., 1974. Hydrocarbon potential of western Murray Basin and infrabasins. Department of Mines, Geological Survey, 5th Australian Report of Investigation, 41 p.

Visser, J.N.J., 1982. Upper Carboniferous glacial sedimentation in the Karoo Basin near Prieska: South Africa. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 38, p. 63-92.

Walker, R.G., 1975. Generalized facies models for resedimented conglomerates of turbidite association. Geological Society of America, Bulletin, v. 86, p. 737-748.

Washburn, A.L., 1980. Permafrost features as evidence of climatic change. Earth Science Reviews, v. 15, p. 327-402.

Wopfner, H., 1982. Depositional history and tectonics of South Australian sedimentary basins. Mineral Resources Review, South Australia, v. 144, p. 32-50.

REFERENCE ADDED TO SECOND PRINTING, JAN., 1986

Deynoux, M., ed., 1985. Paleogeography, Paleoclimatology, Paleoecology, v. 51. Special Issue: The Glacier.

Coarse Alluvial Deposits

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INTRODUCTION

Alluvial conglomerates are minor components of the stratigraphic record, but their tectonic and paleoclimatic significance give them an importance far greater than their abundance implies. They are indicators of the sharp terrestrial relief resulting from lithospheric uplift at continental margins and from intra-cratonic faulting. They are also indicative of climatic extremes, for the production of large lithic fragments is maximized on steep slopes in semi-arid or paraglacial/alpine settings (Wilson, 1973). The study of alluvial conglomerates can, therefore, reveal important tectonic or paleoclimatic influences on sedimentation and basin evolution.

Coarse-grained alluvial deposits also have economic importance, notably the Witwatersrand placer gold and uranium ores of South Africa (Minter, 1978; Smith and Minter, 1980) and the similar uraniferous conglomerates of the Blind River-Elliot Lake area, Canada (Pienaar, 1963). Robertson *et al.* (1978) noted that uranium placer deposits are confined to rocks between 3.0 and 2.2 billion years old, because their formation ceased when the atmosphere became oxygenic. Thick, laterally impersistent coals are associated with intermontane alluvial deposits, some conglomeratic, as described by Heward (1978) and Long (1981).

Compared with studies of sandy fluvial systems, those on alluvial gravels are hampered by problems of scale in flume work, and by the high energy and rarity of natural flows capable of transporting large clasts. In addition, clast size is not necessarily a function of flow competence alone; availability from the source terrane may also be a factor. For these reasons, experimental studies have not contributed greatly to gravel models, although notable exceptions are the work of Koster (1978b) and Southard *et al.* (1984). Studies of modern gravel systems are mainly geomorphic, because of the difficulty of observing active gravel transport, and of coring or trenching gravel. Another problem is the strong influence of Quaternary glaciation on present-day alluvial gravel transport in many parts of the world, an influence present only intermittently in the past. The recognition and interpretation of structures in gravels and conglomerates requires extensive, good quality exposure, which is rare in modern gravels. Both gravels and conglomerates are hard to interpret from cores, and commonly cannot be distinguished from sandstones using borehole logs (Cant, "Subsurface Facies Analysis", this volume).

From the discussion above it is evident that there are abundant data on the geomorphic features of alluvial gravels, but we rely heavily on the ancient record for evidence of stratal type and stratification sequence. This means that coarse alluvial facies models have rather limited use for hydrodynamic interpretation (Walker, 1979). Their function as guides and predictors relates to variations in external factors, largely climatic and tectonic. Recent reviews which provide useful insights are those of Miall (1977), Collinson (1978), Nilsen (1982) and Chapter 6 of Harms *et al.* (1982). Symposium volumes edited by Miall (1978a), Collinson and Lewin (1983) and Koster and Steel (1984) include additional reviews, case histories and discussions of alluvial models.

TERMINOLOGY

The facies terminology used here (Table 1) is that of Miall (1978b). We restrict the term coarse-grained to successions that contain at least 50% gravel, which are therefore dominated by facies prefixed G in Table 1. Sand facies are also present, but in subordinate amounts. In effect, this limits the discussion to braided alluvium, characterised by multiple, low-sinuosity, relatively unstable channels (Rust, 1978a). Various authors, for example McGowen and Garner (1970) and Jackson (1976, 1978) have described meandering-fluvial deposits containing gravel. However, the coarse sediment is restricted to lag accumulations within channels, constituting less than half the total succession. Exceptions are the deposits of streams like the Little Wind River (Jackson, 1978) and the Jarama River, Spain (Arche, 1983), which contains gravel in large-scale cross-strata formed by lateral migration of meander bars. Smith and Smith (1980) reported gravel-filled channels in the anastomosing Columbia River, but Smith (1983, p. 161) showed that these channel sediments are mainly coarse sand and granules. The valley fill is dominated by finer overbank deposits,

Table 1
Facies typical of fans and braidplain deposits (Miall, 1977; as modified by Miall, 1978 and Rust, 1978).

Major facies —	Gm:	Clast-supported, commonly imbricate gravel with poorly defined sub-horizontal bedding.
	Gms:	Muddy matrix-supported gravel without imbrication or internal stratification
	Gt:	Trough cross-bedded clast-supported gravel
	Gp:	Planar cross-bedded gravel, transitional from clast-supported gravel through sand matrix-supported gravel to sand (Sp)
Minor facies —	Sh:	Horizontally stratified sand
	St:	Trough cross-stratified sand
	Sp:	Planar cross-stratified sand
	Fm:	Massive fine sandy mud or mud
	Fl:	Laminated or cross-laminated very fine sand, silt or mud
	P:	Pedogenic concretionary carbonate

Table 2
Descriptive parameters for gravels/conglomerates.

- maximum clast size vs. bed thickness relationship
  - **MATRIX**
    - size, sorting
    - mineralogy
    - pedogenic modification
  - **COARSE FRACTION**
    - size, sorting; shape, sorting — textural maturity
    - fabric
    - lithotypes and compositional maturity
  - (MATRIX and COARSE FRACTION together:)
    - clast or matrix-supported
    - diagenetic changes
    - porosity and permeability
  - **INDIVIDUAL BEDS**
    - boundaries
    - distribution and thickness
    - preserved bedforms, stratification
    - grading
    - fossil content
- **SUCCESSION**
  - temporal trends in bed character
  - stratigraphic relationships with associated facies

Figure 1
Planar cross-stratified conglomerate (facies Gp) in the Middle Devonian Malbaie Formation, Pte. St-Pierre, Quebec. Notebook 19 cm long. Note sorting on cross-strata and sparry calcite cement filling voids in openwork conglomerate (arrowed). See Rust (in press, b).

so that the overall proportion of gravel is small (Smith, 1983, Fig. 5).

An important descriptive parameter for gravels and conglomerates is the relationship between framework (clasts greater than 2 mm in diameter) and matrix (sand- and mud-sized particles) (Table 2). Framework-supported gravel results from deposition of gravel bedload by an energetic aqueous flow that keeps sand in suspension. As flow velocity decreases, the sand infiltrates the spaces between the framework particles (Smith, 1974; Eynon and Walker, 1974; Beschta and Jackson, 1979). Openwork gravel is less common, and results from incomplete matrix infiltration, which ocurs mostly during the rapid accumulation of gravel on cross-strata (Fig. 1). There are two types of matrix-supported gravel: those with stratified sand matrix, and those with unstratified matrix, commonly of muddy sand. The former type indicates aqueous transport, but at an energy level lower than for framework gravel, so that sand and fine gravel are deposited together. The second type of matrix-supported gravel is typified by facies Gms (Table 1), which, in the alluvial context, forms mainly by debris flow deposition.

Other descriptive parameters include particle shape and fabric (Koster *et al.*, 1980) and stratification and stratal sequence (Table 2 and Harms *et al.*, 1982). Aqueous deposition commonly forms a fabric in which maximum projection *(ab)* planes dip upstream at moderate angles, and *a* axes are perpendicular to flow, due to rolling on the bed (Rust 1972b, 1975). Fabrics with *a* parallel to flow are less common, and apparently result from more energetic aqueous transport, in which elongate pebbles saltate longitudinally (Johansson, 1965). Stratification boundaries are commonly gradational, but may be abrupt or erosional. In contrast, debris flow deposits normally lack internal stratification and fabric.

DEPOSITIONAL ENVIRONMENTS

Alluvial gravels accumulate in two related depositional environments: 1) *fans*, and 2) *braided rivers and braidplains.*

Alluvial fans form where streams confined by narrow valleys emerge onto a plain or major trunk valley (Fig. 2). Related gravel-bearing landforms are the steep talus cones that accumulate below mountain gullies (Church *et al.*,

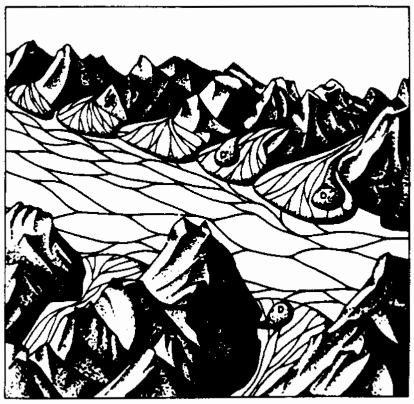

Figure 2
Block diagram of alluvial fans tributary to a braided river in a trunk valley. D) debris flows.

1979; White, 1981), and pediment mantles. Pediments are sloping surfaces cut on bedrock by streams emerging from mountain valleys, which are normally covered by a thin alluvial mantle (Denny, 1967; Twidale, 1979). Sedimentation under these conditions, particularly alluvial fan development, occurs in response to a sharp decrease in transport efficiency as the stream emerges from its confined valley. A semi-conical landform is built, with slopes and transport directions radiating from the mouth of the source valley. Grain size decreases rapidly down fan (Figs. 3 and 4), and roundness of gravels increases, whereas the proportion of finer facies increases distally (Fig. 3). Conditions on the steep valley slopes adjacent to alluvial fans commonly give rise to debris flows, particularly in proximal fan reaches (Fig. 2).

In contrast, *braided rivers and braidplains* (Allen, 1975) have two-dimensional depositional surfaces with lower slopes. Drainage patterns are essentially parallel, although they may radiate or converge locally due to increasing or decreasing space at the margins of the river or braidplain (Fig. 5). Downslope decrease in grain size and attendant facies changes occur over a considerable distance, generally an order of magnitude greater than that required for equivalent facies changes on alluvial fans (Fig. 4). Debris flows are rarely deposited, and if so, are unlikely to survive reworking by aqueous flows.

Some authors have extended the term fan to what are regarded here as braidplains, for example the Scott fan of Boothroyd and Ashley (1975). Boothroyd and Nummedal (1978) referred to coastal outwash plains in Iceland as alluvial fans. These landforms are morphologically unlike fans, and their sediment dispersal patterns and facies fit the braidplain model.

Alluvial fans that are tributary to braided rivers enter them perpendicularly, have significantly higher slopes and are readily distinguished from them (Fig. 5). For example, Spring Creek fan is one of the larger tributary fans of the Donjek River, Yukon, and has a slope of 0.019, whereas that of the trunk river at the same locality is 0.006 (Rust, 1972b). In contrast, a series of laterally contiguous fans formed adjacent to a mountain front may be transitional downslope to a braidplain on which the radiating flow

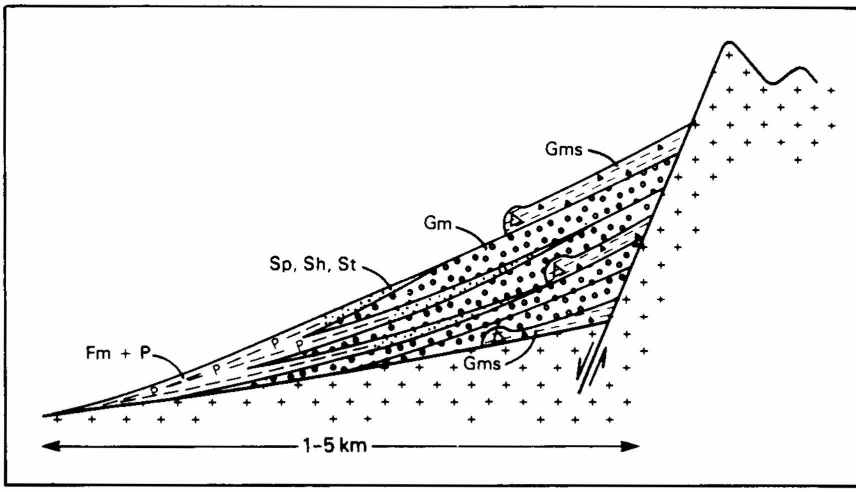

Figure 3
Diagrammatic cross-section of an alluvial fan, showing proximal-distal facies variation. See Table 1 and text for explanation of facies codes.

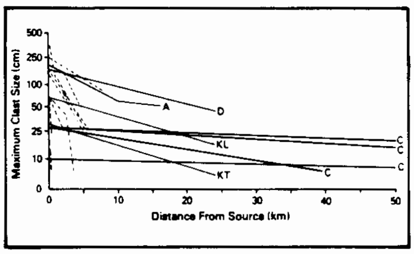

Figure 4
Variation in maximum grain size (i.e., mean of ten largest clasts at each site) versus distance downslope for various alluvial gravels. Modified after Figure 11 of Wilson (1970) and Figure 19 of Schultheis and Mountjoy (1978). Dashed lines are trends on alluvial fans, solid lines are trends on braided rivers and braidplains: A: Arroyo Seco (Krumbein, 1942), D: Donjek River (Rust, 1982a), KL: Knik River, lag gravel (Bradley et al., 1972), KT: Knik River, transported gravel, C: Cadomin Formation (McLean, 1977).

Figure 5
Vertical air photograph (A15517-19) of upper reach of Slims River, Yukon (61° 55'N, 138° 38'W), showing marked contrast between tributary alluvial fans and braided trunk river. Note the entrenchment features of the lower left fan (see Bull, 1977, Fig. 20b), and the constriction of the river by fans. Dark areas on fans are vegetated. Original photo supplied by the Surveys and Mapping Branch, Department of Energy, Mines and Resources, Canada. Width of view about 7.5 km, north toward top of photograph. Flow in trunk river left to right.

Figure 6
Block diagram of alluvial fans transitional downslope to a braidplain as the fans coalesce and lose their morphological identity. D) debris flow.

Figure 7
Horizontally stratified conglomerate (facies Gm) in Middle Devonian Malbaie Formation, Paradise Cove, Quebec. Notebook (arrowed) 19 cm long.

Figure 8
Bedding-plane view of imbricate, horizontally stratified conglomerate in Malbaie Formation, Belle Anse, Quebec. Scale 30 cm.

patterns of the individual fans are lost, but the mean drainage direction is the same (Fig. 6). The distinction between a braided river and a braidplain is one of confinement by a valley in the former case (Kraus, 1984). However, the width of the river is normally sufficient that the influence of the valley walls is minimal, so that the processes and sediments of braided rivers and braidplains are essentially identical.

ALLUVIAL FAN SEDIMENTATION

Modern Alluvial Fans

The classic descriptions of modern alluvial fans are mostly from the mountainous semi-arid regions of the southwestern United States (Bull, 1963, 1964, 1972, 1977; Hooke 1967). Fans of this type are uncommon in Canada, but paraglacial fans (those associated with retreat of valley glaciers) are relatively abundant (Ryder, 1981a,b; Church and Ryder, 1972). In each case the fans form adjacent to regions of high relief, which are rapidly denuded to provide the sediment which builds the fans. In semi-arid environments the relief is commonly a faulted mountain front, and denudation is promoted by sparse vegetation and occasional intense rainfall. Paraglacial fans form where tributary valleys enter a major glaciated valley (Figs. 2 and 5), in which case sediment production is promoted by seasonal temperature fluctuations and the high spring runoff. According to Ryder (1971b), paraglacial fans differ from arid-region fans by having steeper gradients and weaker correlations between fan and basin parameters. This is because their deposits derive from earlier glacially-eroded detritus, in contrast to the concurrent nature of denudation and sedimentation in tectonically-controlled systems. Fans in humid tropical settings are less common, because the climate induces chemical weathering rather than mechanical production of coarse detritus, and dense vegetation protects slopes. However, tropical storms in alpine areas can cause catastrophic mass movements, as illustrated by Bell's (1976) study of the effects of Cyclone Allison on South Island, New Zealand.

Most fans are dominated by water-laid deposits, predominantly facies Gm (horizontally stratified gravel, commonly imbricate) in proximal reaches (Figs. 7 and 8). Bull (1972, p. 66-9) divided water-laid gravels into stream channel, seive and sheetflood deposits, the latter defined in more detail by Hogg (1982). Sheetflood and stream channel deposits can be considered intergradational, for the surfaces on which sheetflood deposits accumulate are in fact composed of numerous shallow channels and bars. Sieve deposits are comparatively rare, forming as gravel lobes that receive little sand or mud from their source areas (Bull, 1977, Fig. 7). The broad, shallow channels on alluvial fans of humid climates commonly contain periodic accumulations of coarse, imbricate gravel, known as transverse ribs. They are interpreted as antidune bedforms, and can be used to estimate paleodepth, velocity and Froude number (Koster, 1978a). Rust and Gostin (1981) recognised fossil transverse ribs in Holocene fan gravels in semi-arid South Australia, and showed how paleohydraulic parameters could be estimated, using equations given by Koster (1978a).

On many fans the main channel is entrenched in the fanhead (proximal) region, but reaches the general level at a location downfan known as the intersection point (Hooke, 1967). Some authors attribute fanhead entrenchment to external causes such as climatic change (Lustig, 1965) or faulting (Bull, 1965), but others regard it as an intrinsic part of fan development. Wasson (1977b) suggested that both circumstances may prevail. Downcutting that contributes sediment to a lower part of the fan is part of fan construction, whereas down-cutting that results in sedimentation beyond, but not on the fan, constitutes destruction by external causes.

The nature of water-laid deposits shows progressive change down-fan. There is an increase in the abundance of cross-stratal sets, chiefly planar (Fig. 1) with transitions from coarse gravel through clast-supported fine-grained gravel, sand matrix-supported gravel to sand (Gm to Gp to Sp). These changes reflect gradual decrease in the particle size to water depth ratio as stream competence decreases down-fan. Minor deposits of horizontally laminated sand (facies Sh) and laminated or massive mud (Fl, Fm) also increase in abundance down-fan (Fig. 3). Fans formed entirely of sand and finer sediment are not part of our topic, but they are in any case rare, because they need a high-relief source of poorly consolidated sand or finer material, which is a short-lived feature of the landscape (Legget *et al.*, 1966).

Debris flow (or mudflow) deposits are the other principal component of most alluvial fan successions in both semi-arid and paraglacial environments (Fig. 3). Middleton and Hampton (1976) pointed out that debris flows are one member of a continuous range of sediment gravity flows. According to Bull (1977, p. 236) debris flows are promoted

Figure 9
Leveed edge of debris flow on west side of Donjek Valley between Spring Creek and Donjek Glacier. Pack (mid-ground, circled) and figure (behind) give scale.

by steep slopes, lack of vegetation, short periods of abundant water supply and a source providing debris with a muddy matrix. Johnson (1970) discussed debris flows, providing eyewitness accounts, as did Sharp and Nobles (1953), Curry (1966) and Winder (1965). The flows may be confined to channels, but commonly spread out as lobate sheets on lower reaches of fans. The lobate distal terminations are distinctive, and commonly concentrate the larger clasts at the steep outer margin of the flow, forming levees (Fig. 9). The flows lack internal stratification, but commonly show reverse or reverse-to-normal grading (Nilsen, 1982, Figs. 17 and 34C,D). In contrast with the imbricate fabric of waterlaid gravel, the clasts in debris flow deposits commonly lack an organised fabric. Bull (1963, p. 245) noted that more fluid (that is, proximal) debris flows may show subhorizontal orientation of megaclasts, whereas more viscous (distal) flows tend to have larger clasts in predominantly vertical orientations, due to matrix support. However, according to Shultz (1983), some flows of relatively low viscosity may be able to reach distal reaches of fans.

Schumm (1977, p. 246) recognized two types of alluvial fans: "... dry or mudflow fans formed by ephemeral stream flow, and wet fans formed by perennial stream flow". This implies that "wet" fans do not develop debris flows. It is true that evenly distributed rainfall favours steady erosional processes rather than mass movements, but short term fluctuation in precipitation can undoubtedly produce debris flows in humid areas (Curry, 1966; Broscoe and Thompson, 1969; Winder, 1965). Schumm (1973) suggested that initiation of debris flows requires accumulation of a threshold amount of loose detritus, a concept further elaborated by Heward (1978). Beaty (1970) estimated an average depositional rate of approximately 2400 m^3/yr on a debris flow-dominated fan of 4.4 km radius on the California-Nevada border.

Alluvial fans prograde into lakes or seas where high coastal topography causes alluvium to be shed directly into the water (Friedman and Sanders, 1978, Fig. 10-29; Gvirtzman and Buchbinder, 1978). These fans have been termed fan-deltas by several authors (Holmes and Holmes, 1978, p. 358-9; Wescott and Ethridge, 1980). However, because of the steep alluvial slope, the typical deltaic features of break in slope and facies change at the water plane are not always well developed, and the term coastal alluvial fan seems more appropriate (Hayward, 1983). Modification of coastal fans by marine processes was described by Ethridge and Wescott (1984), who noted that they form mainly along continental and island-arc collision zones where continental shelves are narrow and relatively steep.

Ancient Alluvial Fan Deposits

The principal features of modern alluvial fans - debris flow deposits and rapid downslope facies changes - are also recognisable in ancient fan successions. Many of these successions are thick, indicating formation in a tectonically-influenced setting, for example the Devonian Peel Sound Formation of Arctic Canada (Miall, 1970). The stratigraphic record also shows vertical changes in facies type. For example, when faulting gives rise to source elevation or basin subsidence, the alluvial system is rejuvenated, and the fan progrades. Areas on which proximal facies accumulate migrate downfan, so that the succession at any given location shows upward increase in grain size and bed thickness (Fig. 10a). This upward coarsening/thickening trend may be overlain by a thinner upward fining/thinning sequence (Fig. 10b) as the effects of rejuvenation wear off (Mack and Rasmussen, 1984). Repeated faulting gives rise to cyclic repetition of coarsening/thickening units or the asymmetric coarsening then fining units descirbed above. Heward (1978) suggested that simple fining-upward sequences may result when faulting causes retreat of the scarp front (Fig. 10c).

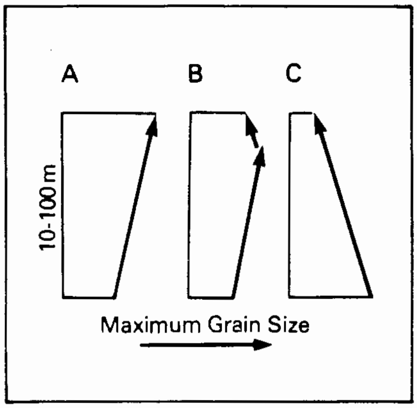

Figure 10

Allocyclic grain size/bed thickness trends in alluvial fan successions subject to periodic tectonic rejuvenation. a) Coarsening/thickening upward due to periodic fault uplift of source or subsidence of basin. b) Coarsening/thickening followed by fining/thinning in asymmetric cycles. Same mechanism as in a), but spacing between fault movements allows system to move toward equilibrium before renewed faulting. c) Fining/thining upward, ascribed to back-stepping of boundary fault (Heward, 1978).

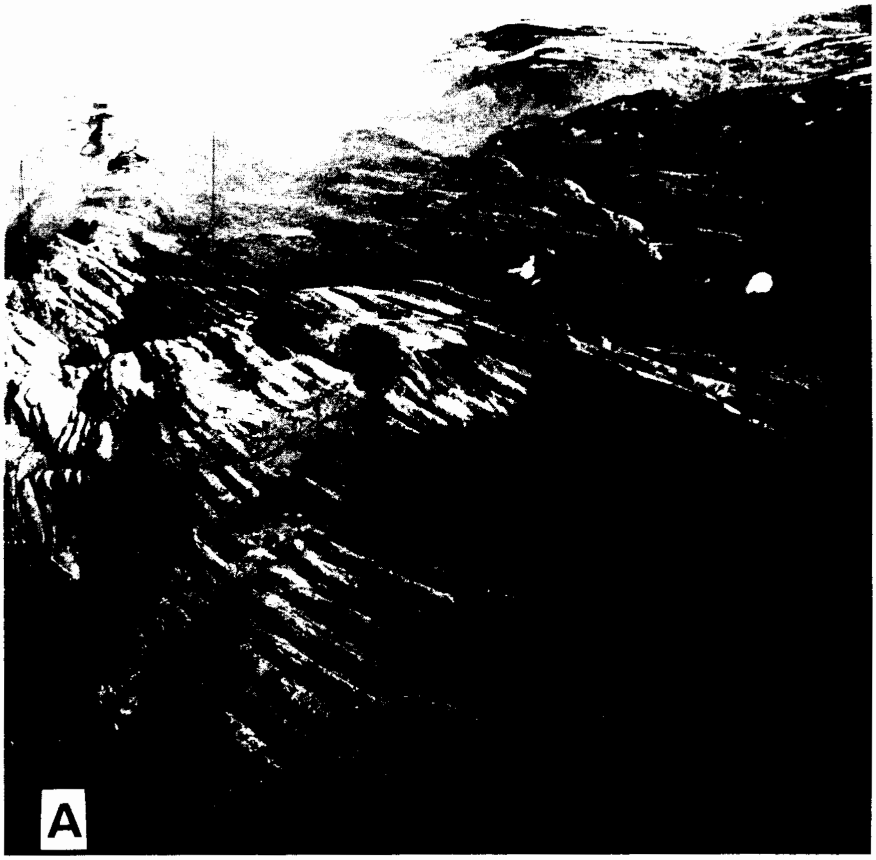

Figure 11

The Devonian Hornelen Basin, Norway, with its spectacular exposure of coarsening-upward cycles in alluvial fan deposits (Steel et al., 1977). A) Northern edge of the basin with successive scarps, ca. 100 m high, relating to each cycle. B) A steeply tilted, southern part of the basin showing three cycles which thin upwards (i.e., towards the basin centre) and interfinger with floodbasin/lake depostis. Photos kindly supplied by R.J. Steel who obtained 'A' from Fjellanger Wideroe A-5, Oslo.

Well exposed successions of ancient alluvial fan deposits were described from the Devonian of Norway by Steel *et al.* (1977), Steel and Gloppen (1980) and others (Fig. 11). They contain coarsening-upward sequences about 100 m thick, with coarsening-upward subcyles in the 10 to 25 m range, all attributed to allocyclic (tectonic) causes. Cycles of internal (autocyclic) origin are also present in these and other fan deposits. They result from major floods, or from the switching of deposition from one fan lobe to another. Repetitive units of this type are thinner, commonly a few metres thick, and fine upwards or coarsen then fine symmetrically (Muir and Rust, 1982).

Another feature which has been recognised mainly from studies of ancient fan successions is the relationship between maximum particle size (MPS, mean of 10 largest clasts) and bed thickness (BTh). Bluck (1967) demonstrated for both water-laid and debris flow conglomerates that these parameters commonly correlate, both decreasing downslope. Gloppen and Steel (1981) showed for debris flow deposits that the MPS:BTh ratio also decreases downfan, indicating that competence decreases more rapidly than other attributes of the flow. Bluck (1969) suggested that MPS:BTh ratios are higher for subaerial than for subaqueous flows, which would permit identification of submerged parts of coastal alluvial fans. In most cases, however, fossils and other subaqueous facies provide better environmental indicators.

Ancient debris flow deposits have been recognised in the Quaternary of Tasmania (Wasson, 1977a) and Spain (Harvey, 1984), the Miocene of Switzerland (Burgisser, 1984), the Permo-Triassic of Scotland (Steel, 1974) and in numerous other successions. Surface features such as lobate terminations are rarely preserved because of subsequent erosion or lack of exposure. However, Daily *et al.* (1980) recognised an example in Cambrian coastal fan deposits, in which a lobate termination with upward coarsening was apparent (Fig. 12). Commonly preserved features include a lack of internal stratification and imbrication, and a sheet-like form, in contrast with the common channel forms of water-laid deposits (Wasson, 1977a; Wells, 1984).

Figure 12
30 cm scale rests on top of coarsening-upward debris flow deposit with boulders emergent from its upper surface. The debris flow deposit forms a lobe, whose base is parallel to horizontal stratification in overlying shallow marine sandstones. Coastal alluvial fan deposits in the Lower Cambrian Boxing Bay Formation of Kangaroo Island, South Australia (see Daily et al., 1980).

Figure 13
Paleosol of nodular calcrete(P) in finer upper part of fining-upward sequence formed by flood deposition on an alluvial fan. Lower Member, Carboniferous Cannes de Roche Formation, Barachois-de-Malbaie, Quebec. Tape open 20 cm.

Red colouration and evaporitic paleosols (facies P, Table 1) occur in several ancient alluial fan deposits, and point to a semi-arid paleoclimate (Williams, 1973). Canadian examples are the Carboniferous Cannes de Roche Formation of eastern Gaspé (Rust, 1984a) and the Bonaventure Formation of Gaspé and New Brunswick (Zaitlin and Rust, 1983). The paleosols are predominantly nodular limestones (calcretes) within finer lithologies (Fig. 13). Other types of paleosol can provide paleoclimatic evidence, although with less confidence (Retallack, 1983). According to Bown and Kraus (1981), paleosols are the rule rather than the exception in ancient alluvial deposits, and therefore have considerable potential for paleoclimatic interpretation.

Ancient successions containing coastal fan deposits were recognised in the Miocene of Turkey (Hayward, 1983), the Devonian of Norway (Steel and Gloppen, 1980) and Arctic Canada (Muir and Rust, 1982), and the Cambrian of South Australia by Daily *et al.* (1980). In the Devonian example discussed by Steel and Gloppen (1980), the basin sediments are lacustrine. Repetitive cycles in marginal alluvial fan deposits were also recognised in the lacustrine succession, indicating that the cause of cyclicity was basin-wide subsidence. Daily *et al.* (1980) described a Cambrian fan succession that prograded across shallow marine environments. The unidirectional alluvial paleocurrents provide a clear indication of the orientation of the ancient coastline and therefore help in understanding the multipolar nature of the marine paleocurrents. For example, longshore currents can be distinguished from those induced by onshore wave attack, and by offshore flows such as rip currents.

Depositional Models for Fans

Miall (1977, 1978b) used a study of the modern Trollheim fan, California (Hooke, 1967) as the principal basis for his alluvial fan model. The Trollheim is a small fan with abundant debris flow deposits. Sedimentation is strongly influenced by two factors: the semi-arid climate and the active tectonic setting. It has been suggested that fans in humid climatic settings produce relatively fewer debris flows, and are therefore dominated by water-laid deposits (Schumm, 1977). This is probably true, but the concept has not been demonstrated quantitatively, and some humid-region fans contain abundant debris flow deposits (Broscoe and Thomson, 1969; Winder, 1965).

Other indicators of paleoclimatic influence in the ancient record are paleosols (discussed earlier), associated facies and biota. The facies indicative of a dry paleoclimate are evaporites in lacustrine or tidal flat deposits associated with coastal alluvial fans. Eolian deposits are less diagnostic, because they could be formed in paraglacial as well as arid climatic conditions (Brookfield, "Eolian Facies", this volume). However, the association of paraglacial alluvium with ancient glacigenic deposits has strong interpretive value. These include deposits from glacier ice (tillites), as well as characteristic facies assemblages of glaciomarine or glaciolacustrine environments (Eyles and Miall, "Glacial Facies", this volume). Paraglacial alluvial successions show upward-coarsening during periods of glacial advance and fining during retreat. The situation is complicated by the fact that coarse detritus results not only from proximity to glaciers, but also from the isostatic uplift consequent on the retreat of continental ice sheets. In general, alluvial successions generated in response to episodic glacial advance should show approximately symmetric coarsening-up, fining-up cycles, but such a situation has not been documented.

The preservation potential of fossils is not high in coarse-grained alluvium. However, Gostin and Rust (1983) described vertebrate and insect burrows and large upright tree trunks buried in Holocene alluvial fan gravels in South Australia. The preservation of xerophytic plant stems and associated faunal traces is likely to leave a permanent stratigraphic record of the semi-arid climatic conditions.

As described previously, tectonic influence can be recognised in the form of repetitive cycles of grain size and bed thickness variation with alluvial fan successions. However, the absence of such cyclicity should not be taken as evidence that tectonic influence was lacking. Small, frequent movements on faults maintain transport energy within fluvial systems, without being individually large enough to induce repeated trends.

Coastal alluvial fans are distinguished from their purely terrestrial counterparts largely by the presence of marine or lacustrine fossils. The effects of reworking by waves and subaqueous slumping may also be recognized (Kleinspehn *et al.*, 1984), but if the water body is small and protected, the resultant subaqueous deposits may be hard to detect.

BRAIDED RIVERS AND BRAIDPLAINS

Modern Examples

Gravelly braided rivers are common features of modern paraglacial environments (Rust, 1982a, 1975; Church and Gilbert, 1975). Gravelly braidplains are less common to-day, but their lateral extent gives them a high preservation potential, and they are more abundant in the ancient record.

The most abundant facies of coarse-grained proximal braided rivers and braidplains is horizontally bedded, imbricate gravel, which may appear massive where bedding is thick and texture uniform (Figs. 7, 8 and 14). This

Figure 14
Middle Devonian Malbaie Formation at Petite Pte. St-Pierre, Quebec, showing horizontally strafitied conglomerate (facies Gm), planar cross-stratified conglomerate (Gp) and horizontally stratified sandstone (Sh). Notebook (arrowed) is 19 cm long.

facies dominates proximal braided rivers of paraglacial environments (Boothroyd and Ashley, 1975; Church and Gilbert, 1975; Rust 1972a, 1975) as well as braided gravels not influenced by glacial melting (Ore, 1964, p. 9; Smith, 1970, p. 2999). The dominance of facies Gm reflects the low ratio of mean particle size to water depth, in turn a function of the relatively low relief of bars and channels in proximal reaches. The bars are mostly longitudinal, that is, elongate parallel to flow, with gentle slopes into surrounding low sinuosity channels. Diagonal bars are similar, but oblique to flow (Smith, 1974, p. 210). Leopold and Wolman (1957) proposed that longitudinal bars start as a nucleus of the coarsest bedload fractions deposited in mid-channel as flow diminishes, and grow by addition of finer sediment mainly downstream from the nucleus. Smith (1974) observed similar processes during diurnal stage fluctuations in the Kicking Horse River, British Columbia: lateral and downstream growth of 'unit bars' with predominantly depositional morphology. In the same river Hein and Walker (1977) observed an initial stage of bar formation as 'diffuse gravel sheets' a few pebble diameters thick. They postulated that the sheets evolve into longitudinal or diagonal bars with horizontal stratification, or transverse bars with cross-strata. The latter, however, are rare in gravel-bed braided streams (Smith, 1974, p. 218).

It is clear that falling-stage modifications of gravel bars occur by depositional and erosional processes, but observations during flood stage are hampered by turbid water and the impossibility of walking across bars, let alone channels. Remote sensing is also impracticable under these circumstances. Rust (1978b, p. 614-5) suggested that longitudinal bars are stable bedforms at flood stage, when all the bedload is in motion. An indication that this is so comes from the giant longitudinal bars (1.4 to 2.5 km long, 15 to 45 m high) of catastrophic Pleistocene floods in eastern Washington (Bretz *et al.*, 1956; Malde, 1968). Giant cross-bed sets with boulders up to 3 m diameter formed in estimated water depths up to 100 m (Malde, 1968), and longitudinal bar forms were preserved. Preservation of such large bars implies stability under the conditions prevailing. A possible analogy may be with the apparent bar forms in channels on Mars (Baker, 1978; Komar, 1983).

Sand facies are uncommon in proximal braided gravels, but where present include planar cross-stratified sand (facies Sp) and horizontally stratified sand (facies Sh, Fig 14.) Mud facies are rarely preserved. All these facies increase in abundance downstream, but unlike alluvial fan deposits the change is very gradual (Fig. 4). For example, clast-supported gravel is the principal lithotype of the Donjek River 50 km from its glacial source (Area 2, Rust 1972a).

Like alluvial fans, braidplains may also accumulate gravel in coastal environments (Leckie and Walker, 1982). Examples are the paraglacial coastal outwash plains of Alaska and Iceland (Boothroyd and Ashley, 1975; Boothroyd and Nummedal, 1978).

Ancient Braidplain Deposits

Like their modern counterparts, ancient successions formed in braided rivers and braidplains are characterised by gradual facies change and decreasing grain size in the downstream direction (Fig. 4). Examples of both confined and unconfined braided river deposits are known (Kraus, 1984), as well as close genetic relationships with neighbouring alluvial fan successions (Middleton and Trujillo, 1984; Rust, 1981). Although coarse-grained braidplain successions are generally thinner than those of alluvial fans, their extent parallel to paleslope may approach 500 km. Vonhof (1965) has documented cobble-grade Oligocene gravels at the Alberta-Saskatchewan border derived from the Montana area across a relatively steep foreland slope. Similarly, the Lower Cretaceous Cadomin Formation of the Cordilleran Foothills in Alberta and Montana (McLean, 1977; Schultheis and Mountjoy, 1978) is sheet-like, extending up to 300 km downslope from its source (McLean, 1977, Fig. 4). Maximum clast size decreases in the transport direction (northeastward) very gradually (McLean, 1977, Fig. 4). Plotted on Figure 4, these data are comparable with maximum clast size trends in modern braided rivers, but differ markedly from size trends on modern alluvail fans. In proximal reaches these deposits are characterised by an abundance of facies Gm (Fig. 7).

Other examples of proximal braidplain conglomerates in the ancient record include Triassic conglomerates in England (Steel and Thompson, 1983), conglomerate units within the Lower Paleozoic Piekenier Formation of South Africa (Vos and Tankard, 1980) and the Middle Devonian Malbaie Formation of Eastern Gaspé, Canada (Rust, 1978b, 1984b). Essentially continuous coastal sections of the Malbaie Formation expose braidplain conglomerates for about 4 km in the downslope direction and about 5.5 km across the slope of the plain. Within this area grain size does not vary appreciably, and paleocurrents determined from clast imbrication (Fig. 8) are essentially uniform (Rust, 1984b). This suggests that the rate of grain size reduction down the Malbaie braidplain was similar to that in the Cadomin conglomerate, and in modern equivalents. As with the other examples, Gm is the predominant facies of Malbaie conglomerate units. Planar cross-stratified conglomerate (facies Gp) makes up about 20% of the conglomerate, a much higher proportion than in the modern equivalents described above (Figs. 1 and 14).

Proximal braidplain deposits form in response to major glacial or tectonic events. However, they are not organised into smaller scale cycles that might represent individual tectonic episodes, as is the case with alluvial fans. It appears that the influence of individual tectonic episodes is lost when the resulting detritus has been transported away from the fans adjacent to the mountain front. Hence stratification and grain size changes are a response to major floods rather than tectonic events.

Distal gravelly braided fluvial deposits are not well represented in the ancient record. An example is the Upper Member of the Carboniferous Cannes de Roche Formation of Eastern Quebec (Rust, 1981, 1984a). Trough cross-stratified clast-supported conglomerate occurs in multiple sets above a sharp erosional base (Fig. 15). This facies (Gt) fines upwards to trough cross-stratified sandstone (St), commonly through intermediate units of horizontally bedded conglomerate (Gm) (Figs. 14 and 15). This succession is interpreted as a response to shallowing of water over bars and active gravelly channels as they accrete, accompanied by, or in response to migration of the active tract. The sequence ends with mudstone and organic material deposited as the tract

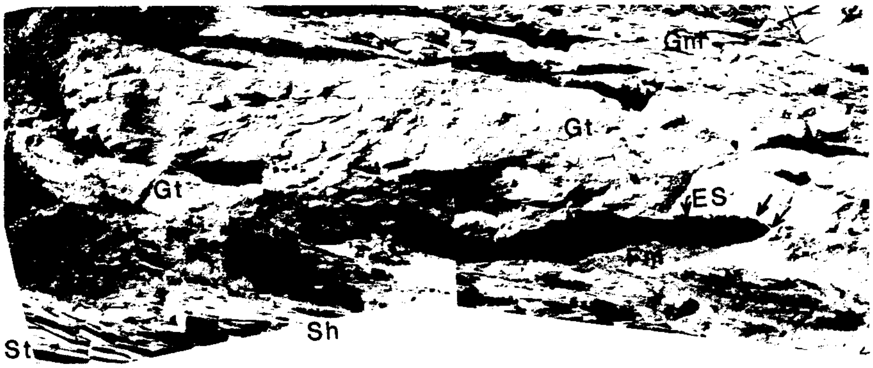

Figure 15
Repetitive stratification sequence in Upper Member of Carboniferous Cannes de Roche Formation, Coin-du-Banc, Quebec. Base of sequence is an irregular erosion surface (ES), overlain by trough cross-bedded conglomerate (Gt), comprising a coset in which some of the trough sets are sandstone-filled. These in turn are overlain by horizontally stratified conglomerate (Gm) and sandstone (Sh). Sandstone sets (Sh and shallow trough: St) are best seen at top of underlying sequence; in turn they pass upward into mudstone (Fm), cut into by erosional surface. Scale (arrowed) 1 m long.

became inactive and started to support vegetation (Figs. 16 and 17).

Models for Braided River and Braidplain Deposition

Two models, proximal and distal, are required to describe the sedimentary characteristics of braided rivers and braidplains.

Miall (1977, 1978b) based his proximal model on proximal reaches of the Scott outwash. The model is essentially the same as facies assemblage G_{II} of Rust (1978b). Imbricate, horizontally-stratified gravel is dominant (facies Gm), with minor amounts of planar cross-stratified gravel (Gp) and sand facies (Sp and Sh). This assemblage (Fig. 18a) characterises proximal outwash gravels, such as the Donjek

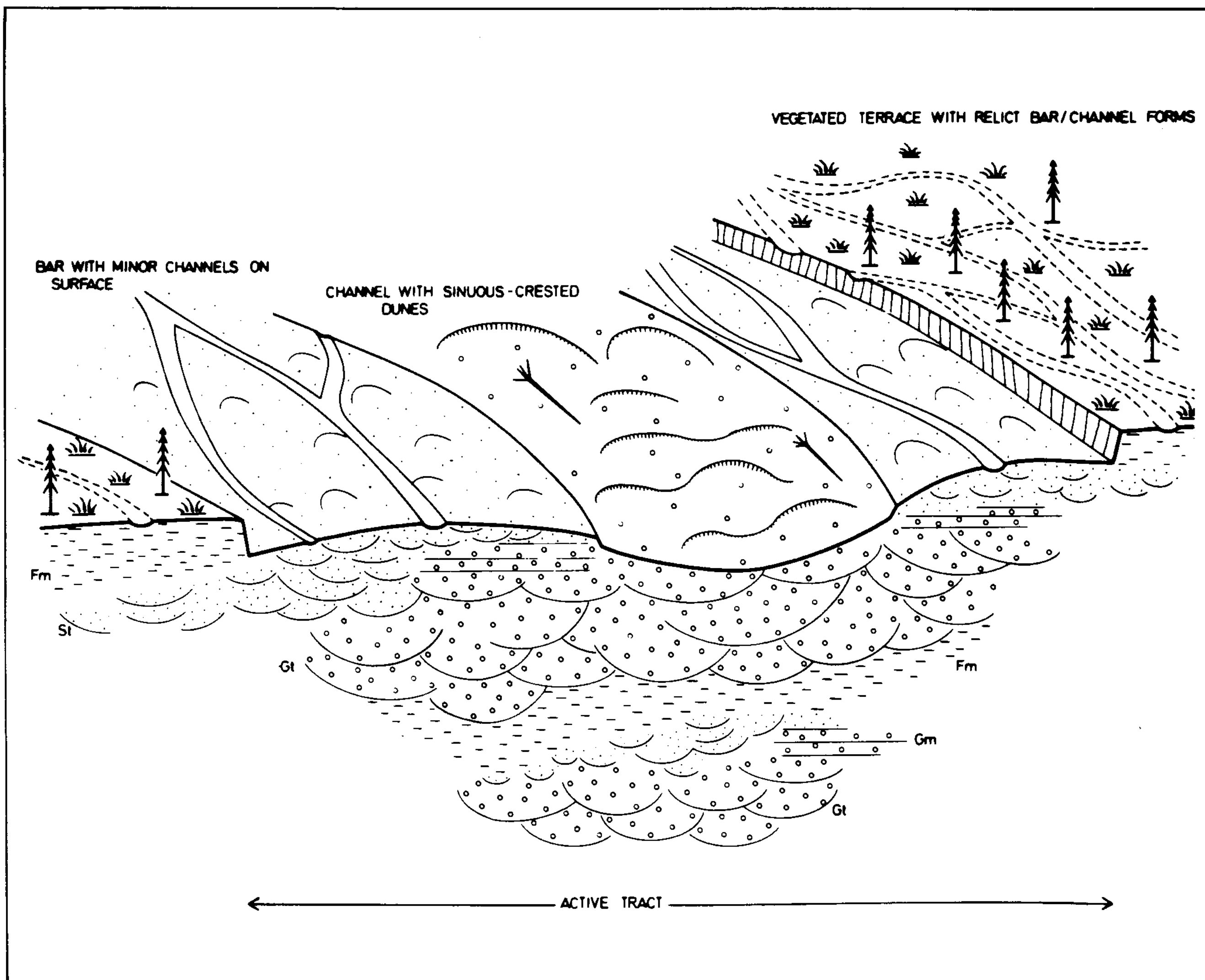

Figure 16
Repetitive stratification sequence in Upper Member of Cannes de Roche Formation, as related to facies assemblage G_{III} of Rust (1978b) and the Donjek model of Miall (1978b).

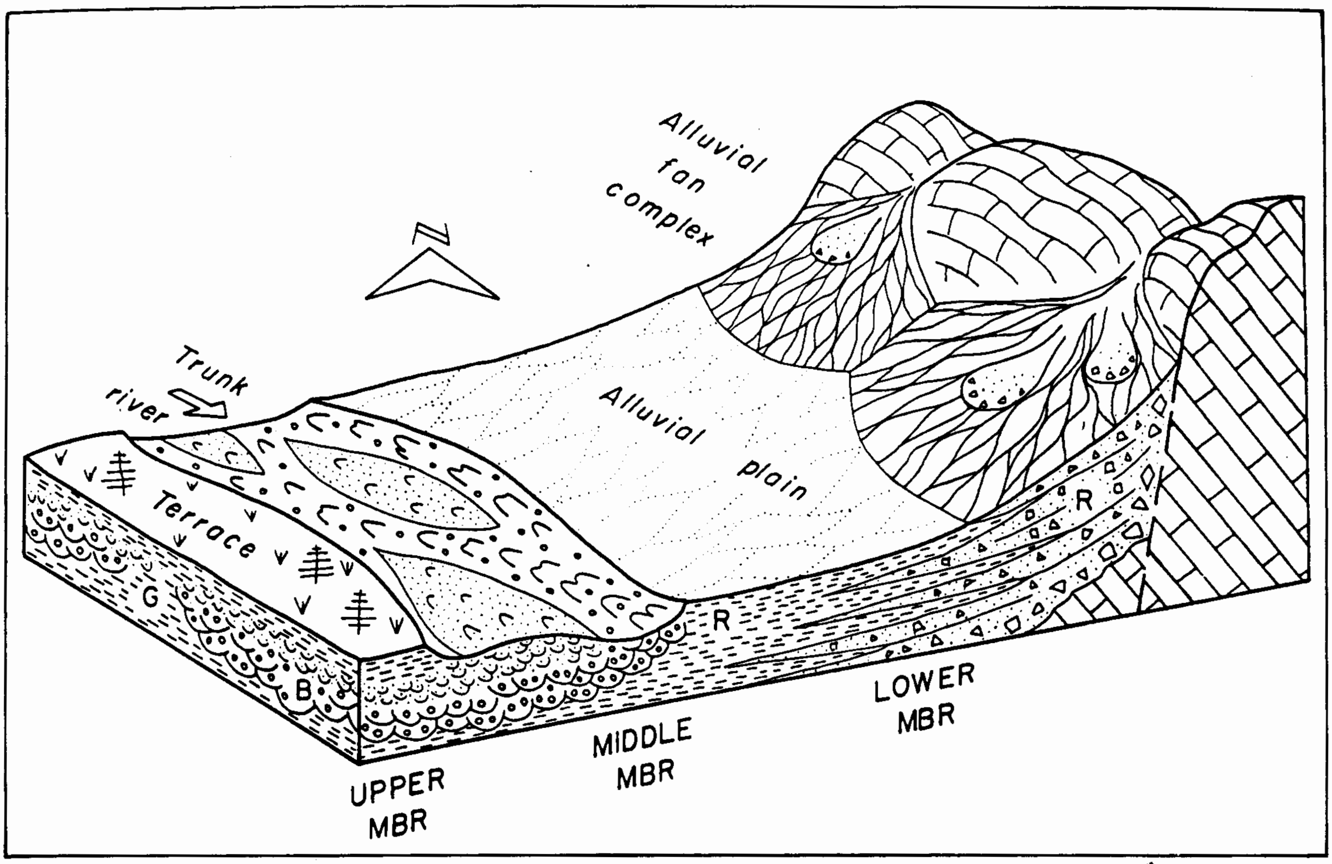

Figure 17
Depositional model for Cannes de Roche Formation, showing distal braided river deposits (Upper Member) in trunk valley confined by tributary alluvial fan deposits (Lower and Middle Members). R) red, B) buff, G) grey-green. See (Rust, 1984a).

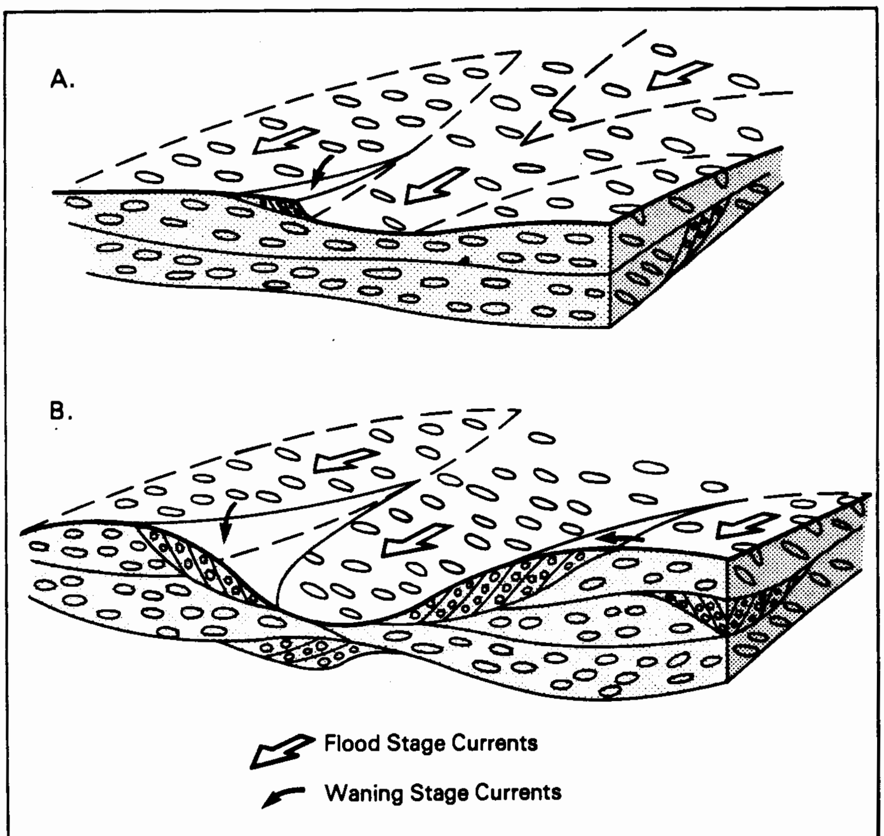

Figure 18
Depositional models for proximal gravelly braidplain/braided river deposits. A) Modern deposition, in which facies Gm is strongly dominant because of shallow flow in semi-arid or paraglacial settings. B) Ancient (Paleozoic) deposition, in which Gm is still the dominant facies, but Gp is more prominent, because the bar/channel relief is greater. This results from commonly recuring deep flood flows, due to formation under humid climatic environments.

(Rust, 1975) and gravelly braided rivers of non-glacial settings (Ore, 1964; Smith, 1970).

In the modern setting, deposits of this type are characteristic of climatic extremes (semi-arid or paraglacial), in which large amounts of coarse detritus are produced, and plant cover is sparse. In the Paleozoic, however, vegetation was confined to well-watered low-lying valleys or coastal plains. Hillslopes were unprotected, so upland gravels were generated abundantly in humid climates, forming extensive braidplains. Unlike modern counterparts, these floods were deeper and more frequent, so large-scale units of planar cross-stratified gravel (facies Gp) formed, together with falling-stage sequences of sand facies. In this sense the input into the model from modern and ancient deposits differs, but the difference is relatively minor (Fig. 18b).

The facies model for distal braided gravels discussed here (G_{III} of Rust, 1978b) is not as well established as that for proximal equivalents: it is offered as a guide for further investigation (Fig. 16). Miall (1977, 1978) chose the middle reaches of the Donjek River (Area 2 of Rust, 1972a, about 50km from its glacial source) as the type example of his facies assemblage for distal gravelly braided rivers. Clast-supported gravel is the dominant lithotype in this reach of the river, and is abundant in the active channel tract. In a similar setting in the Knik River, Alaska, Fahnestock and Bradley (1973, p. 241) identified dunes of fine gravel by echo sounding. They probably resemble the crescentic gravel bedforms observed in the North Saskatchewan River by Galay and Neill (1967), and would generate sets of trough cross-strata (facies Gt) by migration in flood. An alternative possibility in shallower rivers is the formation of transverse gravel bars (Hein and Walker, 1977, Fig. 3) which would generate principally planar cross-strata (facies Gp) on migration.

The mid reaches of the Donjek are also characterised by inactive tracts on levels or terraces above the active tract (Williams and Rust, 1969). The inactive tract is primarily subject to vertical accretion of fine sediment, and supports abundant vegetation. In flood, however, minor channels transport sand and fine gravel across the inactive tract. In time, migration of the active and inactive tracts can be expected to deposit a sequence which fines upward from trough cross-stratified clast-supported gravel through sand to mixtures containing mud and plant material, both transported and *in situ* (Fig. 16 and 17).

The depositional model described above is also representative of facies assemblage G_{III} of Rust (1978b). It has not been recognised in modern braidplain deposits, perhaps because confinement by valley walls is a necessary requirement for development of this facies sequence. The Upper Member of the Carboniferous Cannes de Roche Formation serves as an ancient example of facies assemblage G_{III} (Rust, 1981, 1984a). Paleogeographic reconstruction based on paleocurrents in alluvial fan deposits in the Lower and Middle Members suggests that the braided river deposits were confined to a paleovalley (Fig. 17) with tributary fans on either side. In this sense the Cannes de Roche Formation closely resembles the Donjek River, although the paleoclimatic setting is quite different. Abundant calcretes and red colouration in the Cannes de Roche fan deposits indicate a relatively dry paleoclimate.

TECTONISM AND ALLUVIAL FACIES MODELS

Miall (1981) recognised twelve plate tectonic settings for alluvial basins, in which most conglomeratic sequences were deposited as fault-bounded fan accumulations in forearc or foreland basins, or in intermontane successor or pull-apart basins. Most ancient fan successions show distinct cyclicity, ascribed to periodic tectonism, and ancient braidplain deposits also owe their origin to tectonic causes. The question remains: is there any fundamental difference between the tectonic framework of alluvial fan and braidplain successions?

Ancient braidplain successions are widespread deposits that form in response to major tectonic uplifts. For example, the Cadomin Conglomerate formed in response to Cretaceous uplift of the Cordillera (McLean, 1977; Schultheis and Mountjoy, 1978). The Malbaie Formation and other time-equivalent molasse deposits of the Appalachian and Caledonian belts accumulated in response to the mid-Paleozoic Acadian and late Caledonian Orogenies (Rust, 1981; Allen and Friend, 1968; Allen, Dineley and Friend, 1967). Such major episodes of deformation call for extensive compression, such as that caused by collision of an outboard terrane with a continental margin (Keppie, in press).

In contrast, alluvial fan deposits are more localised accumulations formed adjacent to active fault scarps. Fans may develop in relation to several tectonic situations. In extensional rifting the alluvial successions are relatively thin, because once the continental plates separated or the rift failed, the fault scarps were worn down and became inactive. An example of this scenario was described by Hobday and Von Brunn (1979). An alternative setting for alluvial fan conglomerates is a strike-slip plate margin. In this case, fan accumulations continue to form along the fault complex as it continues to slip, and may reach considerable thicknesses. For example, about 12,000 m of Cenozoic fan deposits accumulated in the Ridge Basin, California, in response to largely strike-slip movement on faults of the San Andreas System (Crowell, 1974). Crowell (1974, p. 300) cited the post-Acadian (Carboniferous) rift basins of Maritime Canada as a similar example in which thick alluvial fan deposits accumulated adjacent to a predominantly strike-slip fault complex. The Devonian deposits in Western Norway described by Steel and Gloppen (1980) represent a similar depositional situation. In terms of plate tetonics, these phenomena can be explained by the activity of a transform fault, causing strike-slip displacement of parts of two adjacent continental plates.

The coarse alluvial deposits of eastern Canada can be divided into pre-Upper Devonian rocks, dominated by braidplain deposits, and Upper Devonian and Carboniferous alluvial fan successions. Rust (1981) suggested that this change in alluvial style was a response to oblique continental collision during the Acadian (mid-Devonian) Orogeny, followed and partly overlapped by transcurrent shearing along the former continental margin during the Carboniferous. There is nothing unusual about such a deformation sequence, because continental collision, or docking of outboard terranes commonly takes place at an olique angle to the original continental margin. Hence initial compression is followed by transcurrent shearing, and the change in alluvial style described above is to be expected.

SUMMARY

Alluvial Fans

The basic model for alluvial fan deposition (the "norm" of the "General Introduction", this volume) is characterised by rapid fining in the downslope direction and by the presence of debris flow deposits (Fig. 3). An additional feature, which may not always be recognisable in the ancient record, is a radiating pattern of paleocurrents. This model can be used as a guide and predictor for understanding reasons for variations from the norm. These variations include:

i) Features associated with a semi-arid paleoclimate: arid-zone paleosols and biota, and association with evaporitic facies such as playa lake sediments.
ii) Features associated with a paraglacial setting: association with glacigenic sediments: tillites, glaciomarine and glaciolacustrine assemblages
iii) Coastal fan features: evidence of reworking by subaqueous processes; lacustrine or shallow marine biota.
iv) Sedimentary responses to tectonism: repetitive sequences of grain size and bed thickness trends on scales of around 10 to 100 m. Commonly they coarsen and thicken upwards, but asymmetric coarsening then fining, and thinning upward sequences are also encountered. A lack of such cyclicity does not necessarily indicate a lack of tectonic influence.

Braided Rivers and Braidplains

Deposits of these environments are essentially identical, and can be characterised by two models, which consititute norms for proximal and distal deposition, respectively (Fig. 17 and 18). The proximal model is simple, and is reasonably well established on several modern and ancient examples. The distal model is based on few case histories, and must be regarded as tentative.

1) *Proximal* deposits of gravel-dominated braided rivers and braidplains are characterised by an abundance of horizontally-stratified gravel deposited by vertical accretion on longitudinal bars. Planar cross-stratified gravel is the next most abundant facies, particularly in Paleozoic and older deposits. Debris flow deposits and tectonically-induced cyclicity are lacking, and the downstream transition to distal deposits typically occurs over a distance of several tens of kilometres.
2) *Distal* assemblages are characterised by autocyclic fining-upward sequences from gravel, chiefly trough cross-stratified, through sandstone to mudstone. The latter facies may include remains of *in situ* vegetation.

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REFERENCES

BASIC REFERENCES

Bull, W.B., 1977. The alluvial fan environment. Progress in Physical Geography, v. 1, p. 222-270.
The latest review by an author who has contributed much to our understanding of alluvial fans. Mainly deals with morphology and deposits of modern fans, but ancient equivalents are also discussed.

Collinson, J.D., 1978. Alluvial sedi ments. *In* Reading, H.G., ed., Sedimentary environments and facies. Oxford, Blackwell Scientific Publications, p. 15-60.
An excellent discussion of the whole spectrum of alluvial deposits.

Collinson, J.D., and Lewin, J., eds., 1983. Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, 575 p.
A volume of papers on fluvial sedimentology published following an international symposium held at Keele, U.K., 1982.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982. Structures and sequences in clastic rocks. Society of Economic Paleontologists and Mineralogists, Short Course No. 9.
Chapter 'Conglomerate, Emphasizing Fluvial and Alluvial Fan Environments' provides a good overview of descriptive features, processes and facies.

Koster, E.H., and Steel, R.J., eds., 1984. The sedimentology of gravels and conglomerates, Canadian Society of Petroleum Geologists, Memoir 10, 441 p.
A compilation of some of the most recent work on processes and facies in the complete range of gravel-dominant environments: several of its studies are reference material for this paper.

Miall, A.D., 1977. A review of the braided-river depositional environment. Earth Science Reviews, v. 13, p. 1-62.
A review of modern and ancient braided alluvial deposits, which introduces the facies code used in this paper.

Miall, A.D., ed., 1978a. Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, 859 p.
Papers resulting from the first international symposium on fluvial sedimentolgoy, held at Calgary in 1977.

Nilsen, T.H., 1982. Alluvial fan deposits. *In* Scholle, P.A., and Spearing, D.R., eds., Sandstone depositional environments. American Association of Petroleum Geologists, Memoir 31, p. 49-86.
A review of alluvial fan deposits with lavish colour illustrations. Despite the "sandstone" of the volume title, conglomerates are thoroughly discussed.

Rust, B.R., 1978b. Depositional models for braided alluvium. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 605-625.
A concise review of the facies sequences that characterise alluvial fan and braidplain environments and guide the basic treatment of the subject in this paper.

Steel, R.J., Maehle, S., Nilsen, H., Roe, S.L., and Spinnangr, A., 1977. Coarsening-upward cycles in the alluvium of Hornelen Basin (Devonian), Norway: sedimentary response to tectonic events. Geological Society of America, Bulletin, v. 88, p. 1124-1134.
A well illustrated account of a remarkable succession of ancient alluvial fan conglomerates and their tectonic setting.

GENERAL

Allen, J.R.L., Dineley, D.L., and Friend, P.F., 1967. Old Red Sandstone basins of North America and Northwest Europe. *In* Oswald, D.H., ed., International Symposium on Devonian System, v. 1, p. 69-98.

Arche, A., 1983. Coarse-grained meander lobe deposits in the Jarama River, Madrid. *In* Collinson, J.D., and Lewin, J.D., eds., Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, p. 313-321.

Beschta, R.L., and Jackson, W.L., 1979. The intrusion of fine sediments into a stable gravel bed. Journal of the Fisheries Research Board of Canada, v. 36, p. 204-210.

Bluck, B.J., 1967. Deposition of some upper Old Red Sandstone conglomerates in the Clyde Area. A study in the significance of bedding. Scottish Journal of Geology, v. 3, p. 139-167.

Bown, T.M. and Kraus, M.J., 1981. Lower Eocene palesols (Willwood Formation, Northwest Wyoming, U.S.A.) and their significance for paleoecology, paleoclimatology, and basin analysis, Palaeogeography, Palaeoclimatology, Palaeoecology, v. 34, p. 1-30

Friedman, G.M., and Sanders, J.E., 1978. Principles of sedimentology. New York, Wiley, 792 p.

Johansson, C.E., 1965. Structural studies of sedimentary deposits. Geologiska Föreningens i Stockholm Förhandlingar, v. 87, p. 3-61.

Keppie, J.K., in press. The Appalachian Collage. *In* The Caledonian Orogen, International Geological Correlation Project Uppsala.

Kormar, P.D., 1983. Shapes of streamlined islands on Earth and Mars: experiments and analysis of the minimum-drag form. Geology, v. 11, p. 651-654.

Koster, E.H., 1978b. A flume study of fluvial gravel fabric (Abst). *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 853.

Koster, E.H., Rust, B.R., and Gendzwill, D.J., 1980. The ellipsoidal form of clasts with practical applications to fabric and size analyses of fluvial gravels. Canadian Journal of Earth Sciences, v. 17, p. 1725-1739.

Long, D.G.F., 1981. Dextral strike-slip faults in the Canadian Cordillera and depositional environments of related fresh-water intermontane coal basins. *In* Miall, A.D., ed., Sedimentation and tectonics in alluvial basins. Geological Association of Canada, Special Paper 23, p. 153-186.

Miall, A.D., 1978b. Lithofacies types and vertical profile models in braided river deposits: a summary. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 597-604.

Miall, A.D., 1981. Alluvial sedimentary basins: tectonic setting and basin architecture. *In* Miall, A.D., ed., Sedimentation and tectonics in alluvial basins. Geological Association of Canada, Special Publication 23, p. 1-33.

A comprehensive classification of alluvial basins with respect to their plate tectonic setting.

Minter, W.E.L., 1978. A sedimentological synthesis of placer gold, uranium and pyrite concentrations in Proterozoic Witwatersrand sediments. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 801-829.

Pienaar, P.J., 1963. Stratigraphy, petrology and genesis of the Elliot Group, Blind River, Ontario, including the uraniferous conglomerate. Geological Survey of Canada, Bulletin, v. 83, 140 p.

Retallack, G.J., 1983. A paleopedological approach to the interpretation of terrestrial sedimentary rocks: the mid-Tertiary fossil soils of Badlands National Park, South Dakota. Geological Society of America, Bulletin, v. 94, p. 823-840.

Robertson, D.S., Tilsley, J.E., and Hogg, G.M., 1978. The time-bound character of uranium deposits. Economic Geology, v. 73, p. 1409-1419.

Rust, B.R., 1978a. A classification of alluvial channel systems. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 187-198.

Schumm, S.A., 1977. The fluvial system. New York, Wiley-Interscience, 335 p.

Smith, N.D., and Minter, W.E.L., 1980. Sedimentological controls of gold and uranium in two Witwatersrand paleoplacers. Economic Geology, v. 75, p. 1-14.

Walker, R.G., 1979. Facies models 1, General introduction. Geoscience Canada, v. 3, p. 21-24.

Wilson, L., 1973. Variations in mean annual sediment yield as a function of mean annual precipitation. American Journal of Science, v. 273, p. 335-349.

ALLUVIAL FANS

Beaty, C.B., 1970. Age and estimated rate of accumulation of an alluvial fan, White Mountains, California, U.S.A. American Journal of Science, v. 268, p. 50-77.

Bell, D.H., 1976. High intensity rainstorms and geological hazards: Cyclone Allison, March 1975. Kaikoura, New Zealand. Bulletin of the International Association of Engineering Geology, no. 14, p. 189-200.

Bluck, B.J., 1969. Old Red Sandstone and other Palaeozoic conglomerates of Scotland. American Association of Petroleum Geologists, Memoir 12, p. 609-629.

Broscoe, A.J. and Thomson, S., 1969. Observations on an alpine mudflow, Steele Creek, Yukon. Canadian Journal of Earth Sciences, v. 6, p. 219-229.

Bull, W.B., 1963. Alluvial fan deposits in Western Fresno County, California. Journal of Geology, v. 71, p. 243-251.

Bull, W.B., 1964. Alluvial fans and near-surface subsidence in Western Fresno County, California. United States Geological Survey, Professional Paper 437-A, 71 p.

Bull, W.B., 1972. Recognition of alluvial-fan deposits in the stratigraphic record. *In* Hamblin, W.K., and Rigby, J.K., eds., Recognition of ancient sedimentary environments. Society of Economic Paleontologists and Mineralogists, Special Publication 16, p. 63-83.

Bürgisser, H.M., 1984. A unique mass flow marker bed in a Miocene streamflow molasse sequence, Switzerland. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 147-163.

Church, M. and Ryder, J.M., 1972. Paraglacial sedimentation: a consideration of fluvial processes conditioned by glaciation. Geological Society of America, Bulletin, v. 83, p. 3059-3072.

Church, M., Stock, R.F., and Ryder, J.M., 1979. Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: debris slope accumulations. Arctic and Alpine Research, v. 2, p. 135-144.

Crowell, J.C., 1974. Origin of Late Cenozoic basins in Southern California. *In* Dickinson, W.R., ed., Tectonics and sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 22, p. 190-204.

Curry, R.C., 1966. Observation of alpine mudflows in the Tenmile Range, Central Colorado. Geological Society of America, Bulletin, v. 77, p. 771-776.

Daily, B., Moore, P.S., and Rust, B.R., 1980. Terrestrial-marine transition in the Cambrian rocks of Kangaroo Island, South Australia. Sedimentology, v. 27, p. 379-399.

Denny. C.S., 1967. Fans and pediments. American Journal of Science, v. 265, p. 81-105.

Ethridge, F.G., and Westcott, W.A., 1984. Tectonic setting, recognition and hydrocarbon reservoir potential of fan-delta deposits. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 217-235.

Gloppen, T.G., and Steel, F.J., 1981. The deposits, internal structure and geometry in six alluvial fan - fan delta bodies (Devonian, Norway) - a study in the significance of bedding sequence in conglomerates. *In* Ethridge, F.G., and Flores, R.M., eds., Recent and ancient nonmarine depositional environments: models for exploration. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 64-69.

Gostin, V.A., and Rust, B.R., 1983. Sedimentary features of some Quaternary alluvial fan successions, South Australia. *In* Williams, B.P.J., and Moore, P.J., eds., Fluvial sedimentology workshop. Australasian Sedimentologists Specialists Group, p. 37-55.

Gvirtzman, G. and Buchbinder, B., 1978. Recent and Pleistocene coral reefs and coastal sediments of the Gulf of Elat. International Congress of Sedimentologists, Postcongress Guidebook, p. 161-191.

Harvey, A.M., 1984. Debris flows and fluvial deposits in Spanish Quaternary alluvial fans: implications for fan morphology. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 123-132.

Hayward, A.B., 1983. Coastal alluvial fans and associated marine facies in the Miocene of S.W. Turkey. *In* Collinson, J.D., and Lewin, J., eds., Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, p. 323-336.

Heward, A.P., 178. Alluvial fan sequence and megasequence models: with examples from Westphalian D - Stephanian B coalfields, Northern Spain. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 669-702.

Hobday, D.K., and Von Brunn, V., 1979. Fluvial sedimentation and paleogeography of an Early Paleozoic failed rift, southeastern margin of Africa. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 28, p. 169-184.

Hogg, S.E., 1982. Sheetfloods, sheetwash, sheetflow, or ... ? Earth-Science Reviews, v. 18, p. 59-76.

Holmes, A., and Holmes, D.L., 1978. Principles of physical geology. Sunbury-on-Thames, Nelson, Third Edition, 730 p.

Hooke, R.L.B., 1967. Processes on arid-region alluvial fans. Journal of Geology, v. 75, p. 438-460.

Johnson, A.M., 1970. Physical processes in geology. San Francisco, Freeman, 575 p.

Kleinspehn, K.L., Steel, R.J., Johannessen, E. and Netland, A., 1984. Conglomeratic fan-delta sequences, Late Carboniferous - Early Permian, western Spitsbergen. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 279-294.

Koster, E.H., 1978a. Transverse ribs: their characteristics, origin and paleohydraulic significance. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 161-186.

Legget, R.F., Brown, F.J.E. and Johnson, G.H., 1966. Alluvial fan formation near Aklavik, Northwest Territories, Canada. Geological Society of America, Bulletin, v. 77, p. 15-30.

Lustig, L.K., 1965. Clastic sedimentation in Deep Springs Valley, California. United States Geological Survey, Professional Paper 352-F, p. 131-192.

Mack, G.H., and Rasmussen, K.A., 1984. Alluvial fan sedimentation of the Cutler Formation (Permo-Pennsylvanian) near Gateway, Colorado. Geological Society of America, Bulletin, v. 95, p. 109-116.

Miall, A.D., 1970. Devonian alluvial fans, Prince of Wales Island, Arctic Canada. Journal of Sedimentary Petrology, v. 40, p. 556-571.

Middleton, G.V., and Hampton, M.A., 1976. Subaqueous sediment transport and deposition by sediment gravity flows. *In* Stanley, D.J., and Swift, D.J.P., eds., Marine sediment transport and environmental management. New York, Wiley, p. 197-218.

Muir, A.D., and Rust, B.R., 1982. Sedimentology of a Lower Devonian coastal alluvial fan complex: the Snowblind Bay Formation of Cornwallis Island, Northwest Territories, Canada. Bulletin of Canadian Petroleum Geology, v. 30, p. 245-263.

Rust, B.R., and Gostin, V.A., 1981. Fossil transverse ribs in Holocene alluvial fan deposits, Depot Creek, South Australia. Journal of Sedimentary Petrology, v. 51, p. 441-444.

Ryder, J.M., 1971a. The stratigraphy and morphology of paraglacial alluvial fans in south-central British Columbia. Canadian Journal of Earth Sciences. v. 8, p. 279-298.

Ryder, J.M., 1971b. Some aspects of the morphometry of paraglacial fans in south-central British Columbia. Canadian Journal of Earth Sciences, v. 8., p. 1252-1264.

Shultz, A.W., 1983. The deposits, internal structure and geometry in six alluvial fan-fan delta bodies (Devonian, Norway) - a study in the significance of bedding sequence in conglomerates - discussion. Journal of Sedimentary Petrology, v. 53, p. 325-327.

Schumm, S.A., 1973. Geomorphic thresholds and complex response of drainage systems. *In* Morisawa, M., ed., Fluvial geomorphology. State University of New York, Binghampton, Publications in Geomorphology, v. 4, p. 299-310.

Steel, R.J., 1974. New Red Sandstone flood plain and piedmont sedimentation in the Hebridean Province, Scotland. Journal of Sedimentary Petrology, v. 44, p. 336-357.

Steel, R.J., and Gloppen, T.G., 1980. Late Caledonian (Devonian) basin formation, Western Norway: signs of strike-slip tectonics during infilling. *In* Ballance, P.F., and Reading, H.G., eds., Sedimentation in oblique-slip mobile zones. International Association of Sedimentologists, Special Publication 4, p. 79-103.

Twidale, C.R., 1979. The character and interpretation of some pediment mantles. Sedimentary Geology, v. 22, p. 1-20.

Wasson, R.J., 1977a. Last-glacial alluvial fan sedimentation in the Lower Derwent Valley, Tasmania. Sedimentology, v. 24, p. 781-799.

Wasson, R.J., 1977b. Catchment processes and the evolution of alluvial fans in the lower Derwent Valley, Tasmania. Zeitschrift für Geomorphologie, v. 21, p. 147-168.

Wells, A., 1984. Sheet debris flow and sheetflood conglomerates in Cretaceous cool-maritime alluvial fans, South Orkney Islands, Antarctica. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 133-145.

Wescott, W.A., and Ethridge, F.G., 1980. Fan-delta sedimentology and tectonic setting - Yallahs fan delta, Southeast Jamaica. American Association of Petroleum Geologists, Bulletin, v. 64, p. 374-399.

White, S.E., 1981. Alpine mass movement forms (noncatastrophic): classification, description and significance. Arctic and Alpine Research, v. 13, p. 127-137.

Williams, G.E., 1973. Late Quaternary piedmont sedimentation, soil formation and paleoclimates in arid South Australia. Zeitschrift für Geomorphologie, v. 17, p. 102-125.

Winder, C.G., 1965. Alluvial cone construction by alpine mudflow in a humid temperate region. Canadian Journal of Earth Sciences, v. 2, p. 270-277.

Zaitlin, B.A., and Rust, B.R., 1983. A spectrum of alluvial deposits in the Lower Carboniferous Bonaventure Formation of western Chaleur Bay area, Gaspé and New Brunswick, Canada. Canadian Journal of Earth Sciences, v. 20, p. 1098-1110.

RIVERS AND BRAIDPLAINS

Allen, J.R.L., and Friend, P.F., 1968. Deposition of the Catskill facies, Appalachian Region: with notes on some other Old Red Sandstone Basins. *In* Klein, G. de V., ed., Late Paleozoic and Mesozoic continental sedimentation, northeast North America. Geological Society of America, Special Paper 106, p. 21-74.

Allen, P., 1975. Wealden of the Weald: a new model. Proceedings of the Geological Association, v. 86, p. 389-437.

Baker, V.R., 1978. The Spokane flood controversy and the Martian outflow channels. Science, v. 202, p. 1249-1256.

Boothroyd, J.C. and Ashley G.M., 1975. Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 193-222.

Boothroyd, J.C., and Nummedal, D., 1978. Proglacial braided outwash: a model for humid alluvial fan deposits. *In* Miall, A.D., ed., Fluvial sedimentology, Canadian Society of Petroleum Geologists, Memoir 5, p. 641-668.

Bradley, W.C., Fahnestock, R.K. and Rowekamp, E.T., 1972. Coarse sediment transport by flood flows in Knik River, Alaska. Geological Society of America, Bulletin, v. 83, p. 1261-1284.

Bretz, J.H., Smith, H.T.U., and Neff, G.E., 1956. Channelled scabland of Washington: new data and interpretations. Geological Society of America, Bulletin, v. 67, p. 957-1049.

Church, M. and Gilbert, R., 1975. Proglacial fluvial and lacustrine environments. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 22-100.

Eynon, G., and Walker, R.G., 1974. Facies relationships in Pleistocene outwash gravels, southern Ontario: a model for bar growth in braided rivers. Sedimentology, v. 21, p. 43-70.

Fahnestock, R.K., and Bradley, W.C., 1973. Knik and Matanuska rivers, Alaska: a contrast in braiding. *In* Morisawa, M., ed., Fluvial geomorphology. State University of New York, Binghampton, Publications in Geomorphology, v. 4, p. 220-250.

Galay, V.J., and Neill, C.R., 1967. Discussion of "Nomenclature for bed forms in alluvial channels". Journal of Hydraulics Division, American Society of Civil Engineers, v. 93, p. 130-133.

Hein, F.J. and Walker, R.G., 1977. Bar evolution and development of stratification in the gravelly, braided, Kicking Horse River, British Columbia. Canadian Journal of Earth Sciences, v. 14, p. 562-570.

Jackson, R.G., 1976. Depositional model of point bars in the Lower Wabash River. Journal of Sedimentary Petrology, v. 46, p. 579-594.

Jackson, R.G., 1978. Preliminary evaluation of lithofacies models for meandering alluvial streams. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 543-576.

Kraus, M.J., 1984. Sedimentology and tectonic setting of early tertiary quartzite conglomerates, northwest Wyoming, U.S.A. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 203-216.

Krumbein, W.C., 1942. Flood deposits of Arroyo Seco, Los Angeles County, California. Geological Society of America, Bulletin, v. 53, p. 1355-1402.

Leckie, D.A., and Walker, R.G., 1982. Storm- and tide-dominated shorelines in Cretaceous Moosebar-Lower Gates interval - outcrop equivalents of Deep Basin gas trap in western Canada. American Association of Petroleum Geologists, Bulletin, v. 66, p. 138-157.

Leopold, L.B. and Wolman, M.G., 1957. River channel patterns: straight, meandering and braided. United States Geological Survey, Professional Paper 232-B, p. 39-85.

Malde, H.E., 1968. The catastrophic Late Pleistocene Bonneville flood in the Snake River Plain, Idaho. United Stated Geological Survey, Professional Paper 596.

McGowen, J.H., and Garner, L.E., 1970. Physiographic features and stratification types of coarse-grained point bars: modern and ancient examples. Sedimentology, v. 14, p. 77-111.

McLean, J.R., 1977. The Cadomin Formation: stratigraphy, sedimentology and tectonic implications. Bulletin of Canadian Petroleum Geology, v. 25, p. 792-827.

Middleton, L.T., and Trujillo, A.P., 1984. Sedimentology and depositional setting of the Upper Proterozoic Scanlan conglomerate (Central Arizona). *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 189-201.

Ore, H.T., 1964. Some criteria for recognition of braided stream deposits. University of Wyoming, Contributions to Geology, v. 3, p. 1-14.

Rust, B.R., 1972a. Structure and process in a braided river. Sedimentology, v. 18, p. 221-245.

Rust, B.R., 1972b. Pebble orientation in fluvial sediments. Journal of Sedimentary Petrology, v. 42, p. 384-388.

Rust, B.R., 1975. Fabric and structure in glaciofluvial gravels. *In* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation. Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 238-248.

Rust, B.R., 1981. Alluvial deposits and tectonic style: Devonian and Carboniferous successions in eastern Gaspé. *In* Miall, A.D., ed., Sedimentation and tectonics in alluvial basins. Geological Association of Canada, Special Paper 23, p. 49-76.

Rust, B.R., 1984a. The Cannes de Roche Formation: Carboniferous alluvial deposits in eastern Gaspé, Canada. Comptes Rendus, 12th International Carboniferous Congress, Urbana, Illinois, v. 3, p. 72-84.

Rust, B.R. 1984b. Proximal braidplain deposits in the Middle Devonian Malbaie Formation of eastern Gaspé, Canada. Sedimentology, v. 31, p. 675-695.

Schultheis, N.H., and Mountjoy, E.W., 1978. Cadomin conglomerate of western Alberta - a result of Early Cretaceous uplift of the Main Ranges. Bulletin of Canadian Petroleum Geology, v. 26, p. 297-342.

Southard, J.B., Smith, N.D., and Kuhnle, R.A., 1984. Chutes and lobes: newly identified elements of braiding in shallow gravelly streams. *In* Koster, E.H., and Steel, R.J., eds., The sedimentology of gravels and conglomerates. Canadian Society of Petroleum Geologists, Memoir 10, p. 51-59.

Smith, D.G., 1983. Anastomosed fluvial deposits: modern examples from western Canada. *In* Collinson, J.D., and Lewin, J., eds., Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, p. 155-168.

Smith, D.G., and Smith, N.D., 1980. Sedimentation in anastomosed river systems: examples from alluvial valleys near Banff, Alberta. Journal of Sedimentary Petrology, v. 50, p. 157-164.

Smith, N.D., 1970. The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, North-Central Apalachians. Geological Society of America, Bulletin, v. 81 p. 2993-3014.

Smith, N.D., 1974. Sedimentology and bar formation in the upper Kicking Horse River, a braided outwash stream. Journal of Geology, v. 82, p. 205-223.

Steel, R.J., and Thompson, D.B., 1983. Structures and textures in Triassic braided stream conglomerates ('Bunter' Pebble Beds) in the Sherwood Sandstone Group, North Staffordshire, England. Sedimentology, v. 30, p. 341-367.

Vonhof, J.A., 1965. The Oligocene Cypress Hills Formation and its reworked deposits in southwestern Saskatchewan. Alberta Society of Petroleum Geologists, 15th Annual Field Conference Guidebook Part 1, Cypress Hills Plateau, p. 142-161.

Vos, R.G., and Tankard, A.J., 1981. Braided fluvial sedimentation in the Lower Paleozoic Cape Basin, South Africa. Sedimentary Geology, v. 29, p. 171-193.

Williams, P.F., and Rust, B.R., 1969. The sedimentology of a braided river. Journal of Sedimentary Petrology, v. 39, p. 649-679.

Wilson, M.D., 1970. Upper Cretaceous-Paleocene synorogenic conglomerates of southwestern Montana. American Association of Petroleum Geologists, Bulletin, v. 54, p. 1843-1867.

Sandy Fluvial Systems

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INTRODUCTION

Sandy rivers can be subdivided into four types, straight, meandering, braided and anastomosed. Natural straight rivers are very uncommon, and there is probably a spectrum of types from meandering to braided. The anastomosed type has only recently been emphasized, and a model would probably be premature. We will therefore concentrate on the *meandering* and *braided* end members of the spectrum mentioned above. Comparison of new situations with our meandering and braided *norms* should help to establish the range of variation between the end members; it will also help to define better the end members themselves.

HISTORY OF SANDY FLUVIAL STUDIES

A full historical review has been given by Miall (1978a) - our purpose here is to introduce the reader to the development of ideas during the last 40 years. Before doing so, it is important to note that Barrell (1913, p. 458) had identified fining-upward sequences in the Devonian Catskill Formation of New York State, as had Dixon (1921, p. 32) in the South Wales coalfield. Dixon commented that the sequence: flaggy sandstone grading up into red marl is "repeated interminably and in all parts of the series".

The post-war period of about 1944-1960 was devoted mostly to the study of modern rivers, rather than the sediments therein. Classic work is that of Fisk (1944) on the Mississippi, and of the geomorphologists on various U.S. rivers (notably Leopold, Schumm and Wolman; references in Miall, 1978a).

In about 1960, two separate lines of study began to develop, namely recent sediments in rivers, and ancient fluvial sediments.

Recent sediments were studied by the Shell Oil Company in legendary but mostly unpublished work on Brazos River point bars in Texas (see Bernard *et al.*, 1970), where the classic fining-upward sequence was documented, along with vertical changes in sedimentary structures (giant ripple bedding, overlain by horizontal bedding, overlain by small ripple bedding).

Excavations in the Mississippi Old River Locksite were described by Frazier and Osanik (1961). In the early 1960s, the geometry of various sedimentary structures was still being determined, and this is the emphasis of Frazier and Osanik's work, as well as that of Harms *et al.* (1963),in the Red River of Louisiana. The relationship of fluvial bed forms, stratification and flow phenomena were further emphasized by Harms and Fahnestock (1965) from the Rio Grande near El Paso, Texas. The most significant addition to this work was probably that of Jackson (1976), who related in detail flow phenomena, bed forms and stratification sequences in meanders of the Wabash River.

Studies of *ancient meandering sandstones* blossomed along with the work on recent sediments. Fining-upward sequences were described in detail by Bersier (1968), and in classic work by Allen (1964) which has continued to this day (e.g., Allen, 1983). Low sinuosity streams were discussed by Moody-Stewart (1966) from Spitsbergen, but interpretations of *ancient braided sandstones* have been relatively few (Smith, 1970; Campbell, 1976; Cant and Walker, 1976; Allen, 1983). This may be due to the fact that relatively few *modern braided sandy rivers* have been described sedimentologically; classic studies include the Brahmaputra (Coleman, 1969), Platte (Smith, 1970; Blodgett and Stanley, 1980), Tana (north Norway, Collinson, 1970), and South Saskatchewan (Cant and Walker, 1978).

These studies constitute the core of the data base for the models we present below. For the meandering model, there seems to be much more data from ancient rocks than well-studied modern rivers. The emphasis on Devonian Old Red Sandstone/Catskill examples may bias the model toward smaller meandering streams. The data base for braided systems is much smaller - the four or five rivers mentioned above, together with a very small (but growing) number of interpretations of ancient braided sandstones.

MEANDERING SYSTEMS

The main elements of a modern meandering system (exemplified by the Mississippi or Brazos rivers) are shown in Figure 1. Sandy deposition is normally restricted to the main channel, or to partially or completely abandoned meander loops; deposition of fines (silt and clay) occurs on levees and in flood basins.

Basis For The Model

The model has been developed from both modern and ancient sediments. The most important papers on modern meandering streams used in developing the model are those of Sundborg (1956; River Klaralven), Harms *et al.* (1963; Red River), McGowen and Garner (1970; Colorado (Texas) and Amite rivers) and Jackson (1976; Wabash River), five in all.

Investigations of larger-scale processes such as meander loop migration, rates and patterns of channel switching and aggradation remain sparse. Many descriptions of ancient meandering stream deposits have also been integrated into the model, to the extent that the model is one of the most broadly based facies models in terms of numbers of modern and ancient examples used. However, much remains to be done in terms of documenting and understanding variations of the model.

The Channel And Point Bar

Meandering of the channel is maintained by erosion on the outer banks of meander loops, and deposition on the inner parts of the loops. The main depositional environment is the point bar, which builds laterally and downstream across the flood plain.

The channel floor commonly has a coarse "lag" deposit of material that the river can only move at peak flood time.

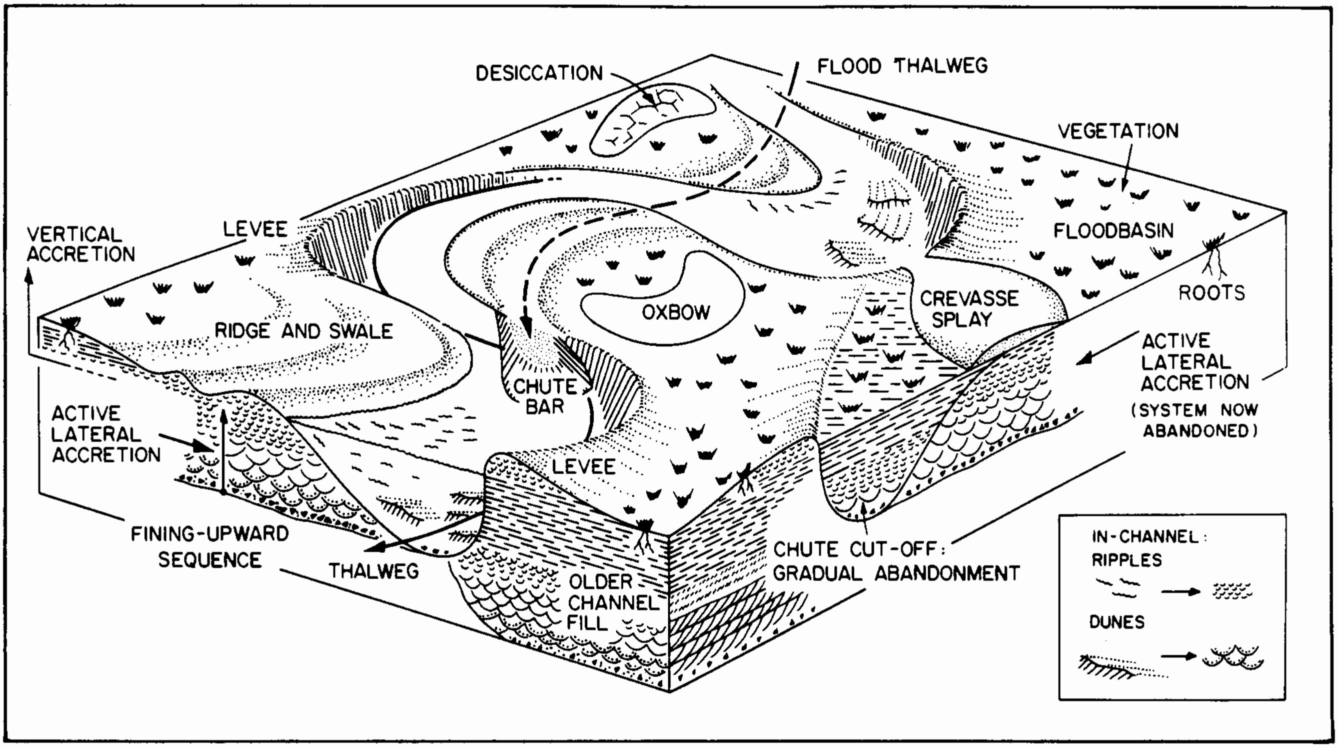


Figure 1
Block diagram showing morphological elements of a meandering river system. Erosion on the outside bend of a meander loop leads to lateral accretion on the opposite point bar. The dunes and ripples in the channel give rise to trough cross bedding and ripple cross lamination respectively (inset, lower right), which are preserved in a fining-upward sequence. See text for details.

This material includes the gravelly component of the clastic load, together with water-logged plant material and partly consolidated blocks of mud eroded locally from the channel wall. Above the lag, sand is transported through the system as bedload. During average discharge, the typical bedform on the channel floor consists of sinuous-crested dunes (Fig. 1) ranging in height from about 30 cm to one metre. Preservation of these dunes results in trough cross-stratification. In shallower parts of the flow, higher on the point bar, the bedform is commonly ripples (preserved as trough cross lamination, Fig. 1). As a broad generalization, we may propose that the preserved deposits of the active channel will pass from trough cross-bedded coarser sands to small scale, trough cross-laminated finer sands upward (Fig. 1).

The development of a plane bed (without ripples or dunes) is favoured by higher velocities, shallow depths and finer grain sizes. Deposition on the plane bed results in horizontal lamination. The particular combinations of depth and velocity required to produce a plane bed can occur at various river stages, and hence parallel lamination can be formed both low and high on the point bar. It can therefore be preserved interbedded with trough cross-bedding, or small scale trough cross-lamination (Figs. 2 and 3).

The sequence shown in Figure 2 is typical of Devonian Old Red Sandstone/Catskill deposits, but does not necessarily characterize deeper or flashier rivers. Very little attention has been given to the response of the sedimentary structure sequence to stage changes in meandering rivers. Also, many modern point bars appear to be terraced (Fig. 4), perhaps due to incision and erosion, or perhaps due to different levels of deposition at various flood stages. The relationship of structure sequence to terracing has not been investigated.

The fining-upward grain size change is a response to spiralling flow through the meander loop. Slightly higher water elevations on the cut-bank side drive a flow down toward the bed and up onto the point bar – the combination of this cross-channel flow with downchannel flow results in spiral flow. Gradually decelerating flow components up the point bar result in the transport of finer and finer sediment, and the general transition from sinous crested dunes (in channel) to small current ripples (near top of point bar).

Erosion of the cut bank and deposition on the point bar result in a gradual lateral and downcurrent shift in position of the point bar. The fining-upward sequence of grain sizes, and accompanying vertical sequence of sedimentary structures, is therefore preserved by LATERAL ACCRETION of the point part (Fig. 1). If the lateral accretion is episodic, or if there are periods of erosion during overall accretion, former positions of the point bar surface can be preserved within the sedimentary sequence. These surfaces are characteristically sigmoid (flat on top of the point bar, steepening down the point bar, and flattening again into the channel floor), with dips of a few degrees up to a maximum of around 15°. They are termed *lateral accretion surfaces* or *epsilon* cross beds (Fig. 5).

Channel Abandonment

Meander loops can be abandoned gradually (chute cut-off) or suddenly (neck cut-off) (Allen, 1965, p. 118-9, 156). During chute cut-off, the river gradually re-occupies an old swale, and

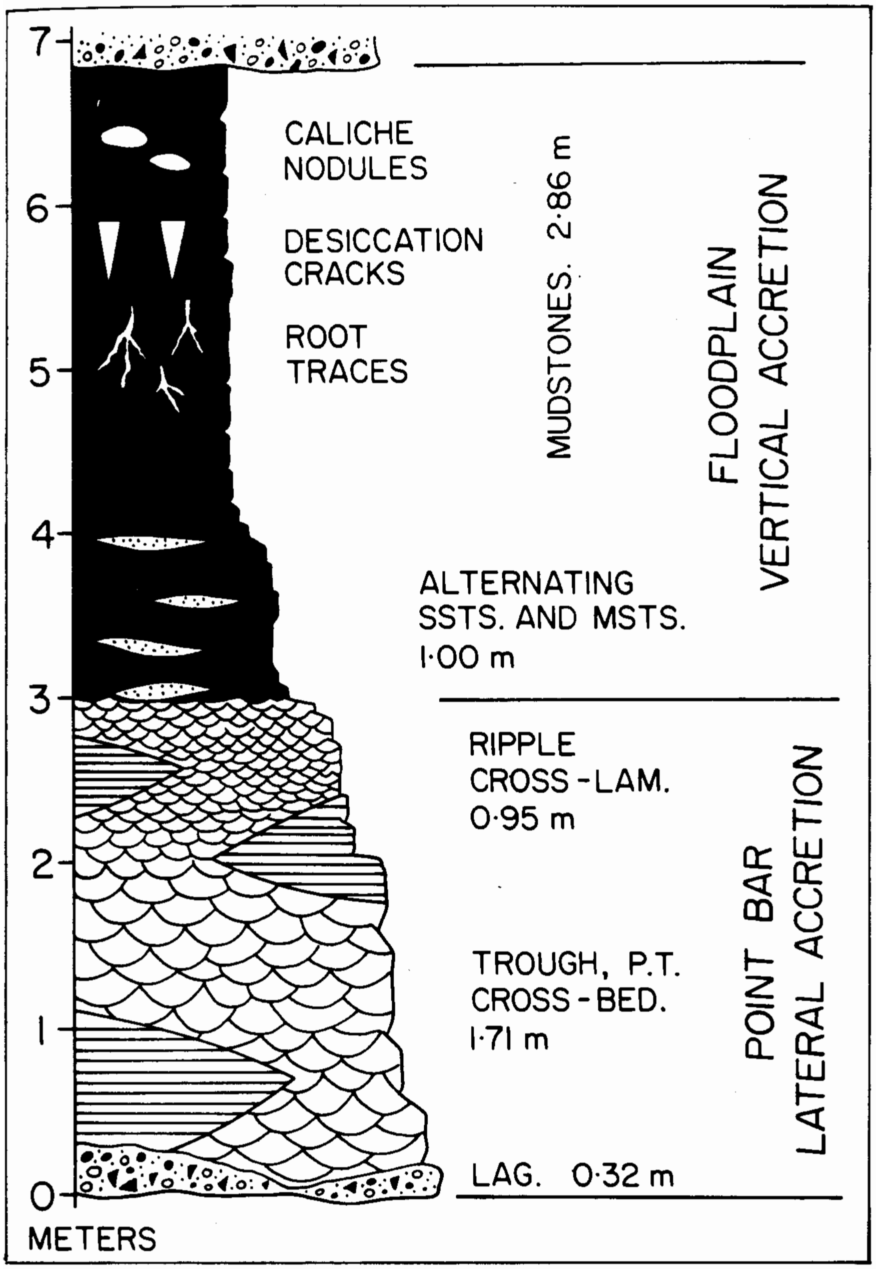


Figure 2
Model for lateral and vertical accretion deposits of meandering rivers. Data on facies sequence and fining upwards cycles summarized here are from the Devonian Old Red Sandstone of Britain and the Catskill rocks of the eastern U.S.A. (Allen, 1970). The average lateral accretion deposit is 2.98 m thick, and the vertical accretion deposit averages 3.86 m. Thus the average sequence is 6.84 m thick. Compare with braided stream sequences in Figures 17 and 19. Note that parallel lamination can replace trough cross bedding or ripple cross lamination, or both. The average thickness of parallel lamination is 1.30 m.

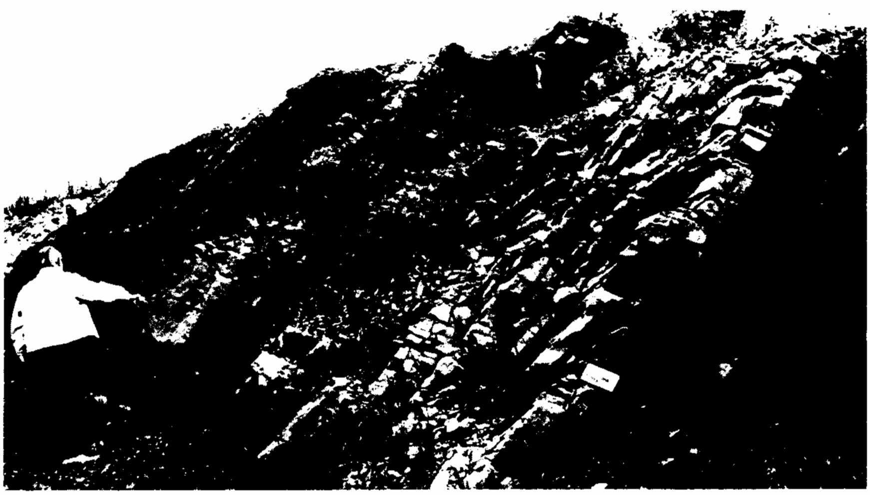


Figure 3
Fining-upward sequence, Cretaceous Belly River Formation, on Trans-Canada Highway between Calgary and Morley Road. Note sharp base to sand body, and cross bedding (by notebook) in lower part. Upper part of sand body (by geologist) is ripple cross laminated and overlain by fines.

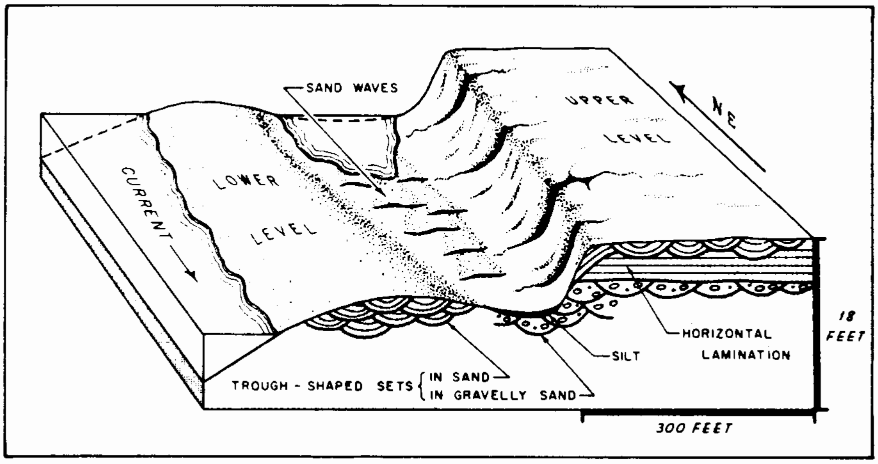


Figure 4
Block diagram showing upstream part of Beene Point Bar, Red River, Louisiana, from Harms et al. *(1963). Note terracing (2 levels) on the point bar, and interbedding of various sedimentary structures.*

Figure 5
Lateral accretion surfaces (epsilon cross strata), Gate Canyon (near Nine Mile Canyon, Utah). In these fluvial Eocene sandstones and shales, note erosive base of sand body. Superbly developed lateral accretion surfaces (L.A.), with overall aggradation during lateral accretion, and cut bank (C.B.) on opposite side. Main sand body about 5 m thick.

simultaneously flow gradually decreases in the main channel. Gradual abandonment thus results in gradual flow decrease, and this could be reflected in the sediments by the development of a thick sequence of low-flow sedimentary structures - essentially ripple cross-lamination (Fig. 6). After complete abandonment, forming an ox-bow lake, sedimentation would be restricted to fines (silt, mud) introduced into the ox-bow during overbank flooding from the main stream (Fig. 1).

Neck cut-off involves the breaching of a neck between two meanders, and the sudden cut-off of an entire meander loop. Both the entrance to and exit from the loop tend to be rapidly plugged with sand. Flow diminishes to zero rather quickly and the resulting sequence of deposits is dominated by later, flood-introduced silts and muds (Fig. 6).

Vertical Accretion Deposits

Outside the main channel, deposition in the flood basins, ox-bows and levees takes place by addition of sediment during flood stage when the river overtops its banks (Fig. 1). In contrast to the lateral accretion within the main channel, overbank deposition causes upbuilding of the flood plain, hence the term VERTICAL ACCRETION. Near to the main channel, where the flood waters sweep along as a stream, the vertical accretion deposits tend to be silty, and are commonly cross-laminated. In some cases, it appears that the levee is breached catastrophically, and the resulting splay bed has a sharp base and begins with parallel lamination. This may pass upward into ripple cross lamination, producing a bed which is descriptively and hydrodynamically akin to a turbidite. It can be distinguished from a deep marine turbidite (see "Turbidites and Associated Coarse Clastic Deposits", this volume) by its context, and the possible root traces in the top of the splay bed.

Farther from the river, flood waters may stagnate and only mud is deposited. After retreat of the flood, the mud and silt commonly dry out, and desiccation cracks are formed. The flood basins and levees of most river systems (post-Silurian) tend to be abundantly vegetated, and hence the deposits commonly contain root traces. In some climatic regimes, the vegetation may grow sufficiently abundantly to form coal seams. In semi-arid environments, the fluctuating water table and drying at the surface favour the formation of caliche-like nodules within the vertical accretion deposits.

The only other deposits that may rarely be preserved as part of the vertical accretion sequence are windblown, and may be either loess, or coarser sandy deposits blown in as large dunes (see "Eolian Facies", this volume).

Meandering River Facies Sequence

The distillation of observations from a large number of modern meandering streams, and from many ancient formations interpreted as meandering-fluvial, allows a general facies sequence to be formulated. One version of this sequence is shown in Figure 2; it was distilled statistically by Allen (1970) and is redrawn to scale here. In its simplest form, the sequence is FINING-UPWARD and consists of in-channel deposits (lateral accretion), followed by overbank fines (vertical accretion) (Figs. 7 and 8).

In this particular sequence, the facies relationships were determined statistically for a large number of Devonian outcrops in Britain and North America, but application of the model has demonstrated that it can be used appropriately in many other areas. The lag

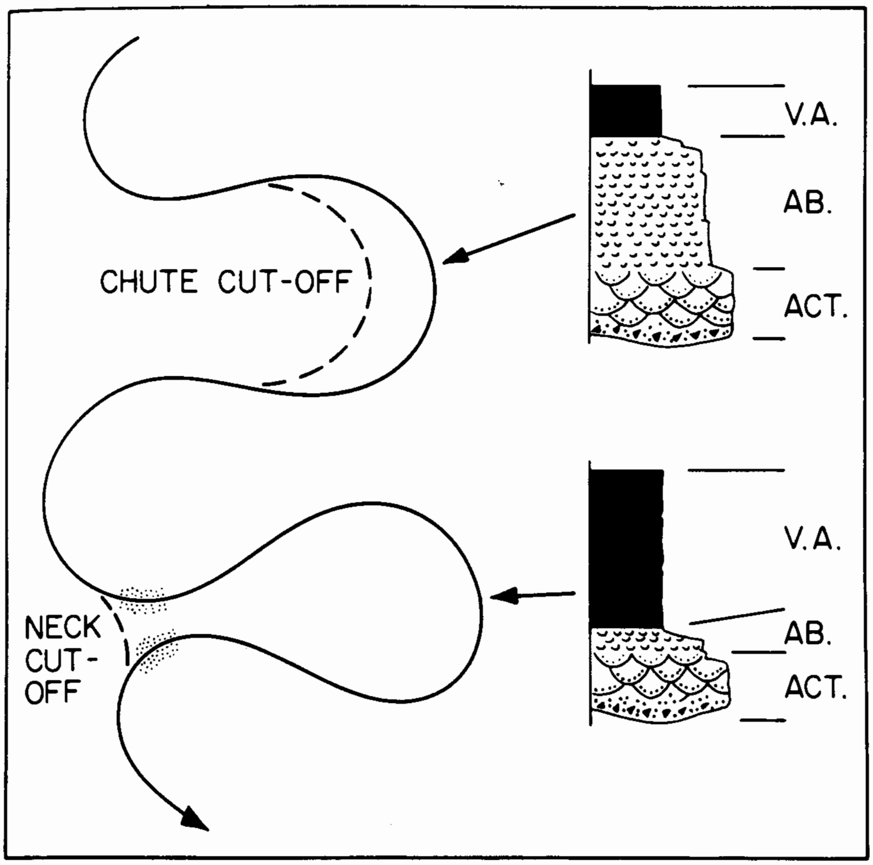


Figure 6
Meander loops can be abandoned by chute or neck cut-off. Old channel shown solid, new channels dashed. Chute cut-off involves reoccupation of an old swale and gradual abandonment of the main channel. The stratigraphic sequence will consist of some trough cross-bedded deposits of the active river (ACT) and a thick sequence of ripple cross-laminated fine sands representing gradual abandonment (AB). After cut-off, the sequence is completed by vertical-accretion (V.A.) deposits. By contrast, after neck cut-off, the meander loop is suddenly abandoned and sealed off by deposition of sand plugs (stipple). After the active deposits, the ripple cross-laminated fine sands representing low flow during abandonment (AB) are very thin, and the bulk of the sequence consists of vertical-accretion (V.A.) deposits washed into the abandoned loop at flood time. Compare with the active lateral-accretion sequence (Fig. 2).

Figure 7
A complete fining-upward sequence from the Pennsylvanian Maringouin Formation, Nova Scotia. Note sharp base, and interbedding of sandstones and shales toward top of sand body. Vertical accretion fines separate the interbedded sandstones and shales from the next sand body (top left). Directly above geologist's head is a small mud-filled channel that cuts out the interbedded sandstones and shales - it may represent an abandoned swale on a point bar.

Figure 8

Photomosaic of a multistory sand body with well-developed lateral accretion surfaces in lower part (LA). Sand body is about 15.5 m thick; Lower Cretaceous Gladstone Formation at Bighorn River, Alberta. The upper part of the sand body is inaccessible, but possibly contains smaller lateral accretion sets, or possibly huge wide shallow troughs dipping toward the reader in the left-hand half of the uppermost sand body.

deposits are overlain by trough cross-bedding, which is in turn overlain by small scale trough cross-lamination. Horizontal lamination can occur at several places within this sequence (Fig. 2), depending on the river stage at the time when the depth/velocity/grain size criteria for plane bed were met.

After the channel migrated away laterally, the facies sequence continued with vertical accretion deposits introduced at flood stage. The diagram (Fig. 2) shows root traces, desiccation cracks and caliche-like concretions. Using the data presented by Allen (1970, Table 9), it can be seen that the vertical and lateral accretion deposits in the meandering model are on average roughly equal in thickness.

Allen's model serves excellently as a norm with which to compare other fining-upward sequences. Comparison of sequences such as those in Figure 6 with Allen's norm immediately shows that the trough cross-bedding is very reduced in thickness, that the chute cut-off sequence contains an abnormal thickness of ripple trough cross-lamination, and that both contain unusual thicknesses of vertical accretion deposits. The comparison with Allen's model suggests the interpretations shown in Figure 6; without the model, we would not be so conscious that the sequences in Figure 6 differed significantly from the sequence developed by lateral accretion in an active channel.

The rocks in Figure 8 also emphasize the role of Allen's model as a norm. In Figure 8, both the sandstones and mudstones are much thicker than suggested by the norm. There are possibly two superimposed sets of lateral accretion deposits in the sandstone, suggesting two superimposed channel sands rather than one extremely thick sand body. The apparently unusual thickness of vertical accretion fines will be discussed again below.

Comparison of the models with some sequences developed for meandering rivers with slightly coarser loads also reveals some differences. Where coarse bedload is funnelled by flood waters through a swale on the top of a point bar, a much straighter thalweg is formed. The coarse material is dumped at the downstream end of the swale forming a chute bar adjacent to the point bar (Fig. 2; McGowen and Garner, 1970). In other rivers, coarser sediment and less mud deposition leads to a sequence without a really well developed fining-upward trend (Jackson, 1976). These variations from the standard point bar model must be recognized and allowed for in the study of ancient sediments.

Sand Body Geometry And Flood-Plain Aggradation

One of the essential components of a meandering model is the fact that meander loops are cut off, abandoned, and ultimately filled with fines - silt and clay. Through time, these clay plugs, along with thick back-swamp clays, may become abundant because the plugs are relatively hard to erode. Once confined, the entire meander belt may become raised above the general level of the flood plain by vertical accretion (Fig. 9A). This situation can persist until one catastrophic levee break results in the sudden switch of the entire river to a lower part of the floodplain ("avulsion", Fig. 9A). Thus the overall sand body geometry of a highly sinuous meandering stream will be essentially elongate ("shoestring"), bounded below and on both sides by flood-basin fines. The shoestring will also stand a good chance of being covered by overbank fines from the active river in its new position. Thus the *high sinuosity meandering model* predicts that, given continuing supply and basin subsidence, a series of point bar sand sheets interbedded with shales should be developed within the overall shoestring geometry. The internal structure of the point bar sands themselves should conform roughly to the pattern shown in Figure 1. A single-sequence sand body should be about as thick as the river was deep; however, if clay plugs restrain the river to the meander belt, multistory sand bodies should be common (Figs. 8 and 9A). Sequences of overbank fines thicker than those predicted by Allen's model could form if the river is confined in this way - compare the mudstones in Figure 8 with those in Figure 2. Conversely, using this model as a predictor, we suggest that unusually thick sequences of vertical accretion fines (10 m +) might predict (along strike) stacked sand bodies in a meader belt confined by clay plugs. The vertical scale in Figure 9A is considerably exaggerated, and individual shoestrings will probably be many times wider than they are thick.

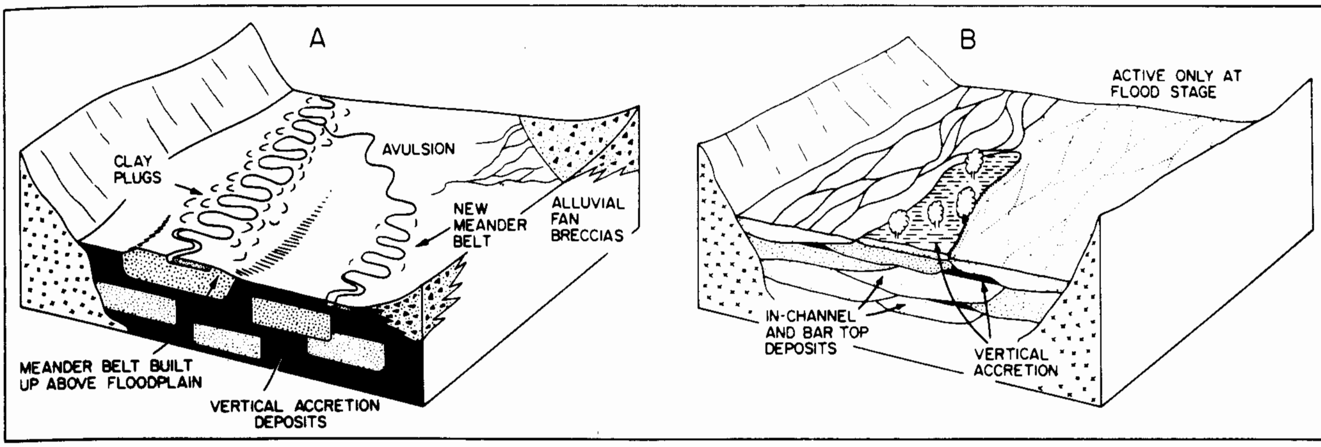
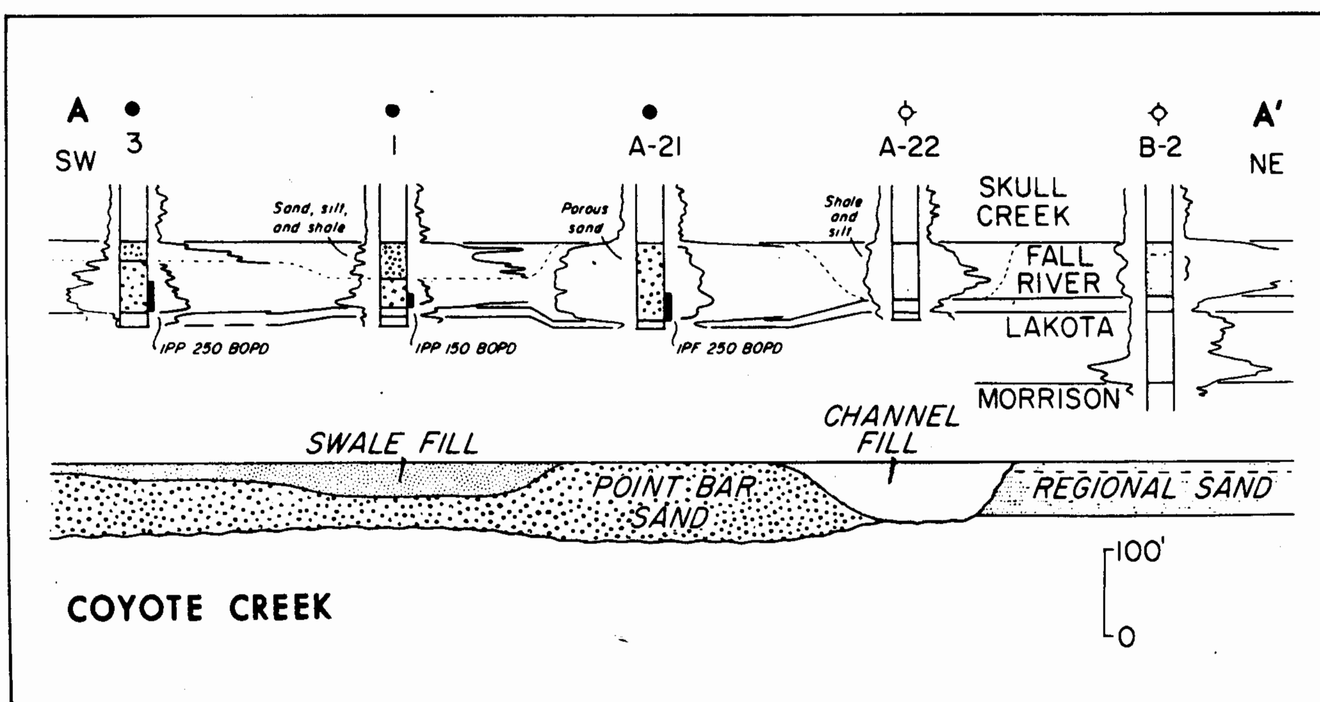


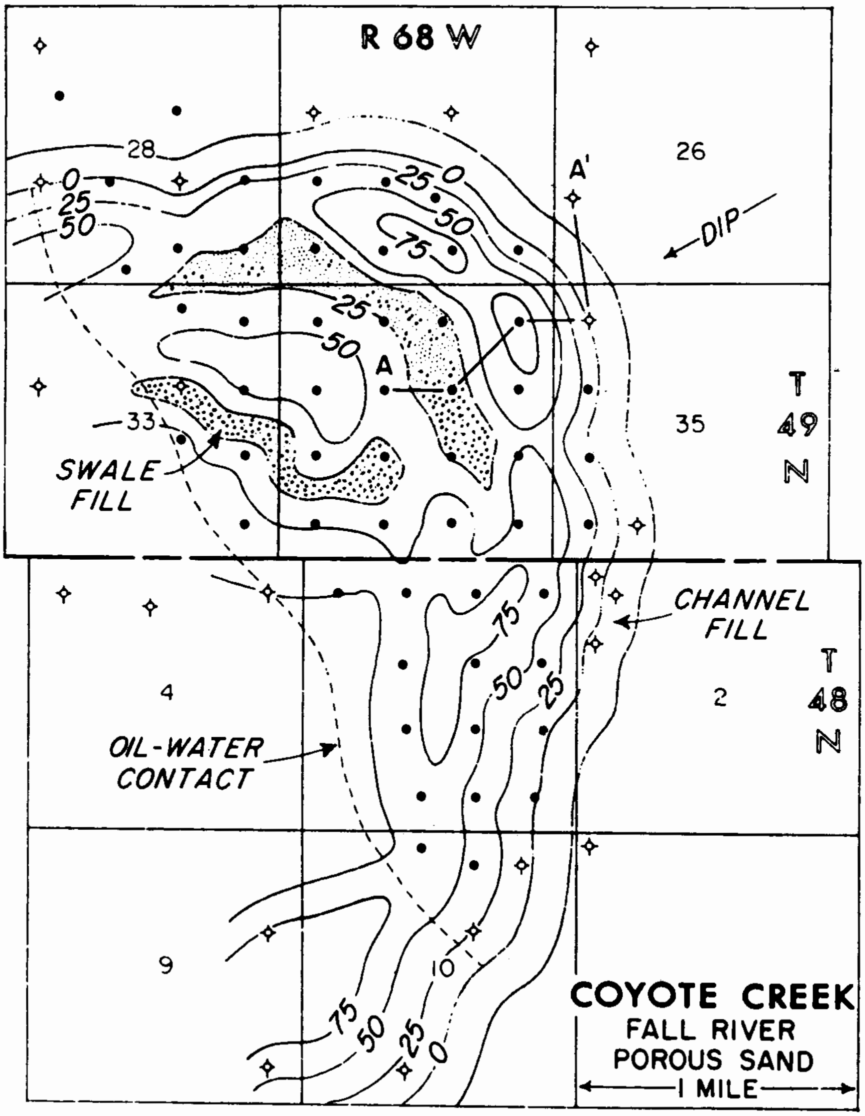
Figure 9
A) block diagram of flood plain aggradation with very sinuous rivers. Shoestring sands are preserved, and are surrounded by vertical accretion siltstones and mudstones. If the river is confined by clay plugs, very thick vertical accretion deposits can form without erosion. B) block diagram of sandy braided system with low sinuosity channels. Vertical accretion can occur during flood stage, but deposits are rarely preserved. Diagrams modified from Allen (1965).

Figure 10
Correlation of well logs (SP, resistivity) across Coyote Creek field (location in Fig. 11), with interpretation in terms of meander belt facies below. Note thickness of sand body - about 50 to 75' (15 to 23 m); channel width is about 1500 to 2000' (460 to 610 m). From Berg (1968).

Meandering Rivers In The Subsurface
There are many examples: we highlight two to illustrate how well-log data (see "Subsurface Facies Analysis", this volume) can be used in fluvial reconstructions. In the Lower Cretaceous sandstones in the Powder River basin, Wyoming, Berg (1968) has shown well-log correlations which can be interpreted in terms of: 1) porous point bar sands; 2) sands, silts and shales of swale-fill origin; and 3) shale and silt representing the fill of the abandoned channel (Fig. 10). Note the thickness of the point bar sand - up to about 75 feet, or 23 m. In the isopach map (Fig. 11), the shapes of the meander loop and swales are apparent. The isopach map does not suggest superimposed separate sand bodies - it appears to be one system 23 m deep which accreted laterally toward the NE. The size of this meander loop (radius of curvature about 2130 m) is comparable with loops in the modern Mississippi and Missouri rivers, which helps to explain the unusually great depth of this Cretaceous river.

In a second example, Hopkins *et al.* (1982) have presented subsurface data for Lower Cretaceous Mannville Group sands in southern Alberta (Figs. 12 and 13). Here, the Mannville channels are incised into older units (Ostracod Limestone and Bantry Shale, Fig. 13), and at least the lower ones were probably not freely meandering. Part of the line BB' (Fig. 12) is shown in Figure 13, and can be interpreted in terms of a laterally accreted sandstone about 20 m





◀ **Figure 11**
Isopach map of Coyote Creek field, northeastern Wyoming. Note location of cross-section AA' (Fig. 10). Swale fills show thinner sands, and the abandoned meander loop itself has no sand. This can be seen both in the sand isopach (values down to zero) and the location of dry holes (open circles) as opposed to oil wells (black circles). From this isopach map, Berg (1968) made the following estimates: channel width 1500 to 2000' (460 to 610 m), meander radius 7000' (2130 m) and meander wavelength 40000' (about 12,200 m). Note that the regional dip is southwestward, and hence the shaly fill of the abandoned meander loop is the updip stratigraphic trap. From Berg (1968).

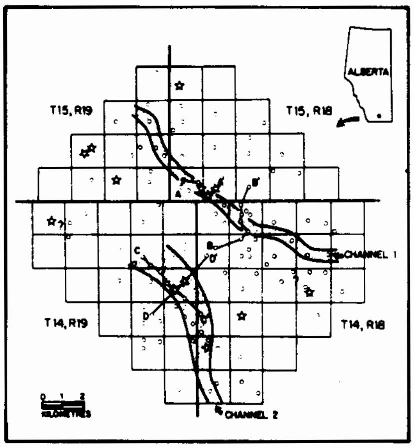


Figure 12
Location map of Little Bow area, southern Alberta. Location of two channels shown. Figure 13 shows cross section BB', omitting the wells at each end of the line. Glauconitic Member of the Mannville (Albian), from Hopkins et al., *1982.*

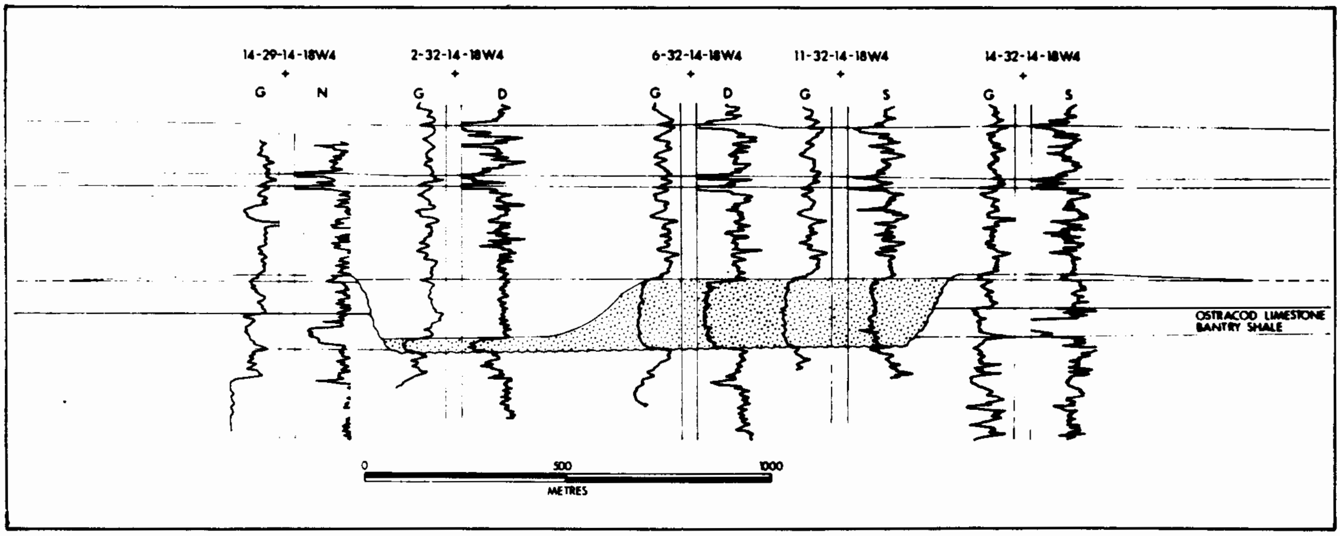


Figure 13
Part of cross-section BB' (see Fig. 12) from the Glauconitic member of the Mannville (Albian), from Hopkins et al. *(1982). From the correlations shown, note the laterally accreting sand body from well 11-32 to 6-32, and then abandonment of the system with subsequent mud filling of the empty channel (well 2-32). The laterally accreted sand is about 20 m thick, the empty channel about 500 m wide, subsequently filled with about 16 to 17 m of fines.*

thick (wells 6-32 and 11-32), with a channel some 500 m wide that filled with 16 to 17 m of mud after cut-off of this meander system (? by avulsion). The overall, incised channel width (Fig. 12) is about 1 km, suggesting that the 500 m quoted above (Fig. 13) represents the width of a meandering channel within the overall, straighter incised channel. This channel width (500 m) is comparable to the narrower parts of the present Missouri River.

In both examples, the simple models of Figures 1 and 2 must be use with caution in prediction, because the models are based upon much smaller rivers than those in Figures 11, 12 and 13, and because of the problem of river incision into older rocks.

Point Bar Reconstructions

Since earlier versions of this paper, several excellent reconstructions of ancient point bars have been made, particularly those of Nami (1976), Nami and Leeder (1978), Nijman and Puigdefabregas (1978) and Edwards *et al.* (1983). In some of these studies, the sand bodies are a little thicker than those in the sequence of Figure 2 (Nami and Leeder quote 3 to 9 m, and Nijman and Puigdefabregas quote a maximum of 11 m), although the 2 to 3 m thick epsilon cross beds quoted by Edwards *et al.* (1983) are very comparable with Figure 2. The importance of these studies is that three dimensional reconstructions of channel, point bar and swales can be made from excellent outcrops, allowing a direct link between modern point bars and preservability of facies in the geological record.

SANDY BRAIDED SYSTEMS

The fundamental processes that control whether a river has a braided or meandering pattern are not completely understood but we do know that braiding is favoured by rapid discharge fluctuations of a greater absolute magnitude than in meandering rivers. Braided rivers also tend to have higher slopes, a heavy load of coarse sediment, and more easily erodible banks. In combination these features would suggest that braiding is more characteristic of the upstream reaches of a river, with meandering becoming more common downstream as the slope and coarseness of load decreases. Braiding would also be more common in semi-arid or arid areas.

Basis For The Braided Model

In contrast to meandering rivers, sandy braided systems have received relatively little study. The best known rivers include the Durance and Ardeche (Doeglas, 1962), Brahmaputra (Coleman, 1969), Platte (Smith, 1970; Blodgett and Stanley, 1980), Tana (Collinson, 1970) and South Saskatchewan (Cant and Walker, 1978).

Few studies of braided rivers have been detailed and comprehensive enough to contribute to the models. The scale of the rivers studied varies enormously, from the Platte (1 to 2 m deep, hundreds of metres wide) to the Brahmaputra (up to 25 m deep, several kilometres wide). Relatively few ancient studies have been integrated as yet into braided stream models. The best documented studies include those of Moody-Stewart (1966, Devonian of Spitsbergen), Kelling (1968, Coal Measures, South Wales), Conaghan and

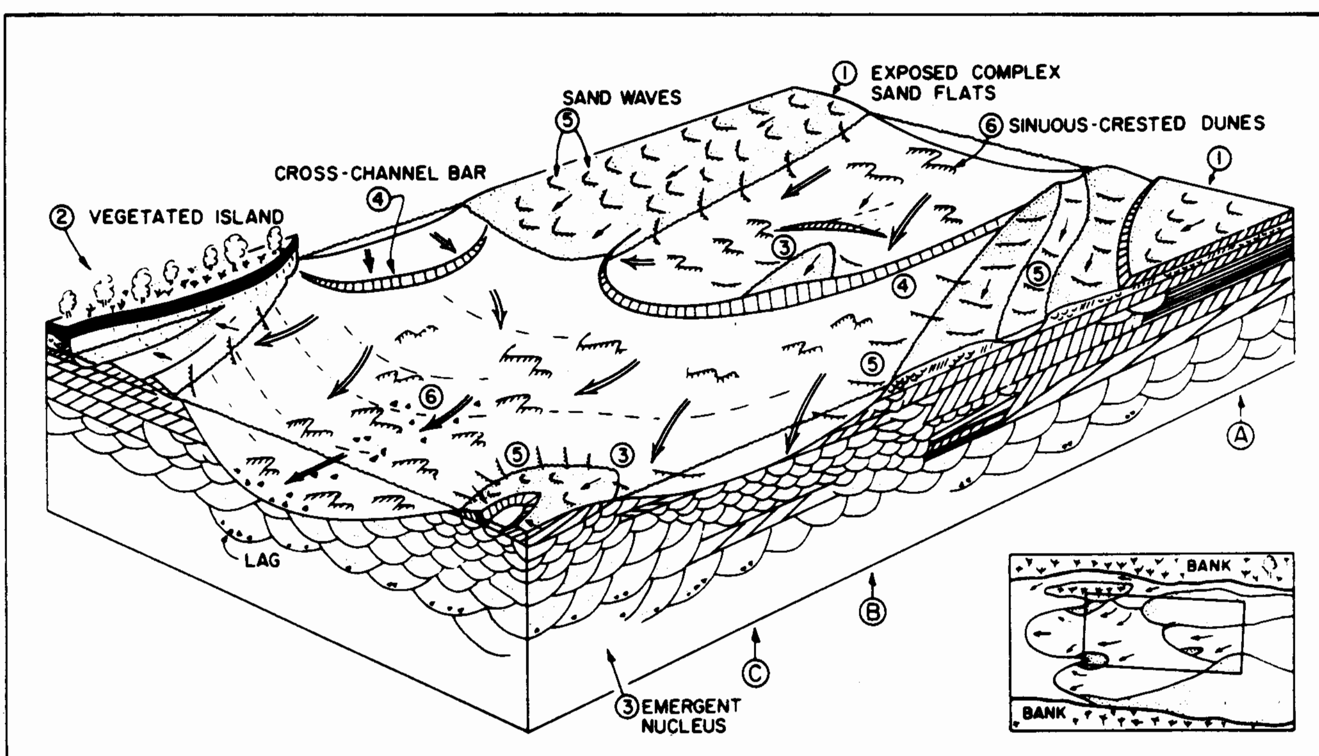


Figure 14
Block diagram showing elements (numbered) of a braided sandy river, based on the South Saskatchewan. Stippled areas exposed, all other areas underwater. Bar in left corner is being driven laterally against a vegetated island, but is separated form the island by a slough in which mud is being deposited. Large sandflats (e.g., right-hand corner) may develop by growth from an emergent nucleus on a major cross-channel bar (see Figs. 15 and 16). Vertical fining-upward sequences A, B and C are shown in Figure 17 and include in-channel and bar top' deposits. See text for details.

Jones (1975), and Jones and Rust (1983, both on the Hawkesbury Sandstone, Australia), Campbell (1976, Morrison Formation, New Mexico), Cant and Walker (1976, Devonian Battery Point Formation, Quebec), Allen (1983, Devonian Brownstones, Welsh Borderlands) and Haszeldine (1983, Upper Carboniferous, north-eastern England). However, the data presented in these studies are diverse, and modern and ancient studies cannot yet be fully integrated into a coherent model.

Braid Bars And Channels

The morphological elements of these rivers (Fig. 14) are complex, and include (in increasing scale) individual bedforms, small "unit" bars, bar complexes (or sandflats), and mature vegetated islands. The river itself flows over and between these sand accumulations in a constantly branching and rejoining braided pattern. The finer material (silt and clay) tends to be transported through the system without accumulation.

The channels tend to be very variable in depth and width, and do not conform to the simple pattern shown by meandering rivers. The channel floor commonly has a lag deposit, and above the lag, sand is transported through the system as bedload. Bedforms in the deeper channels (3 m or deeper) tend to be sinuous crested dunes that give rise to trough cross-bedding. Deposition within channels during waning flood stage can cause channel beds to aggrade, preserving flood stage sedimentary structures. In shallower channels, and on bar tops when they are submerged at flood stage, small dunes and straight-crested to rhomboid sandwaves (Harms *et al.* 1982) are common (Fig. 14, number 5).

Also in the channels are wedge-shaped foreset-bounded transverse or oblique bars. In the South Saskatchewan these can extend across the entire widths of channels, and are termed cross-channel bars (Fig. 15; Cant and Walker, 1978). They form where 1) a smaller channel discharges into a deeper one (as a microdelta), 2) where the flow spreads laterally, or 3) where the flow is forced by channel patterns upstream to flow obliquely across the main river system. This can result in a bar near the bank of the river, generally elongated parallel to the channel trend, but with foresets facing the channel bank (left hand end of Fig. 14). In the Platte and Tana rivers many bars have a more regular linguoid pattern, commonly in an en-echelon arrangement, but this may reflect only a more complete remolding of the bed by high stage floods than in the South Saskatchewan.

Many of the cross-channel bars in the South Saskatchewan have a higher area which is emergent at low stages (Figs. 14 and 15; Cant and Walker, 1978). This high area may act as a "nucleus" for further deposition. The nucleus grows by lengthening downstream as sand is swept around in two "horns" (Fig. 15),

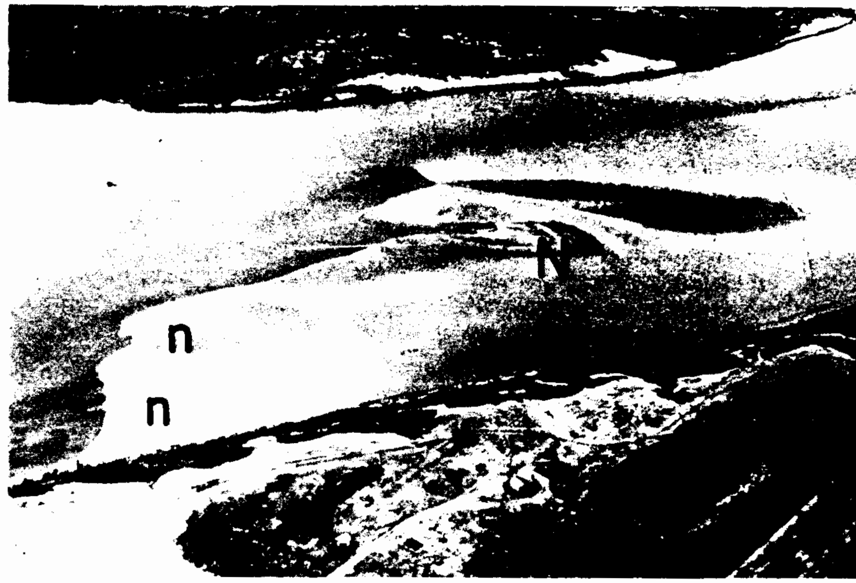


Figure 15
Cross channel bar linked to bank in foreground. Nucleus (N) has given rise to two simple horns, and the entire cross channel bar has apparently migrated downstream by a distance equal to the length of the horns. Note smaller nuclei (n) near bank in foreground.

Figure 16
Evolution of a large sand flat by the coalescing of two nuclei (N) with extensive horns. Note that even the low-stage channel (C) does not have erosive margins, and appears to be aggradational. South Saskatchewan River, flow toward bottom left.

and it also grows in the upstream direction as dunes and sand waves are driven up from the channel floor. As the nucleus grows, possibly with other bars coalescing onto it, the original unit bar expands into a large sandflat (Cant and Walker, 1978; Fig. 16). The South Saskatchewan sandflats are complex, and their original shape has been obscured by dissection and redeposition during changing river stage (Fig. 14). They are one to two km long in the South Saskatchewan, three km in the Tana (Collinson, 1970) and up to 10 km in the Brahmaputra (Coleman, 1969). In the South Saskatchewan, they remain constant for at least five to six years, and because of their size, they would seem likely parts of the braided system to be preserved in the stratigraphic record.

From our understanding of the South Saskatchewan (Cant and Walker, 1978), we propose a series of stratification sequences (Fig. 17) that might characterise the deposits of this type of river. The channel sequence (Fig. 17) would consist of a lag, overlain by trough cross stratification formed by migrating sinuous-crested dunes. Sandflat development appears to be initiated by the development and emergence (during falling stage) of a cross channel bar, which would be represented by a thick (0.5 to 2 m) set of planar-tabular cross bedding (Fig. 14, number 4). Nucleus aggradation, and horn growth and modification during a series of floods and falling stages (Fig. 18) would give rise to a complex set of small (tens of cm) planar tabular cross beds. A spectrum of sequences between channel aggradation and sandflat development probably exists, depending on where the sequence developed – in a deep channel, or in the immediate vicinity of a nucleus (compare Figs. 14 and 17).

The sandy tops of all of these sequences are composed of smaller planar and trough cross beds, and rippled sands, making up the feature termed bar top* in Figure 14. The bar top* (with asterisk) implies that deposition and modification are not restricted to the exposed bar tops, but may also take place in shallow dissection channels. The terminology of in-channel and bar top* was first used for ancient rocks (Cant and Walker, 1976; Fig. 19); it is important that the same terms be used for ancient and recent sediments where possible.

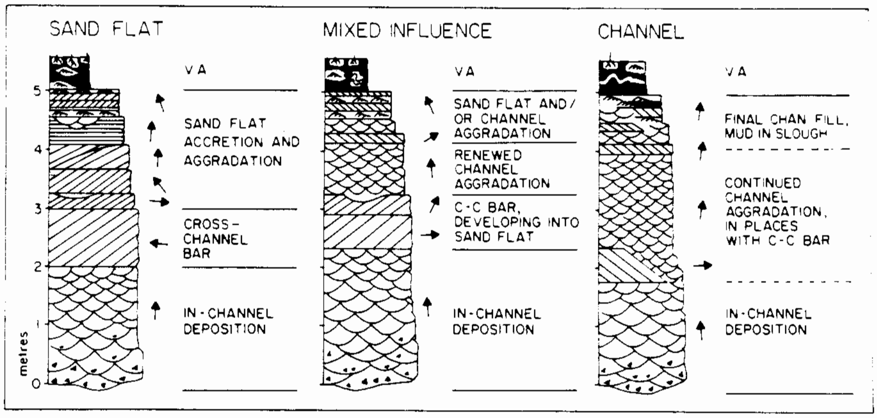


Figure 17
Three proposed sequences of sedimentary structures based on the South Saskatchewan. "Sand flat" corresponds to A (Fig. 14), "channel" to C, and "mixed influence" to B. Arrows indicate generalized paleoflow directions, and sequences are explained in the text.

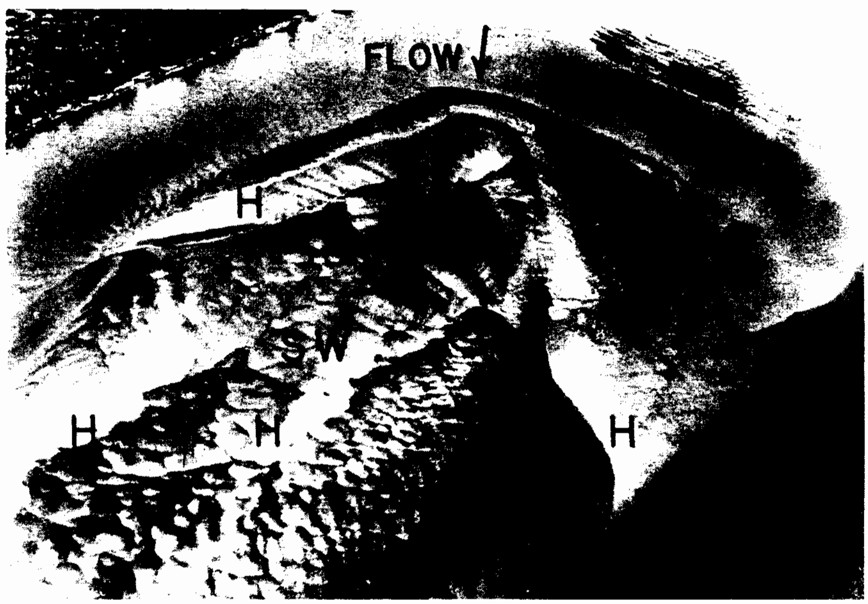


Figure 18
Large sand bar in the South Saskatchewan river; the original position of the nucleus (N) can only be estimated. Several horns (H) have developed from the nucleus, and relict positions of the inward facing sharp, steep crests of the horns can be seen. However, during stages of higher flow, the horns have been modified into a series of straight-crested sand waves (SW); the sand wave crests strike across the horns, and indicate downchannel flow at high stage, not flow expansion during falling stage in the lee of the nucleus (see Figs. 15 and 16).

Vertical Accretion Deposits

In contrast to meandering streams, the vertical accretion deposits of braided streams are less commonly deposited and only rarely preserved. Only during major floods does the river spill from its main channel system onto the surrounding flood plain. In the South Saskatchewan, the braided portion is essentially confined between Pleistocene bluffs. Consequently, the narrow flood plain and the vegetated islands can be relatively easily submerged and receive vertical accretion deposits. In the Platte River, a great deal of sand is swept into the overbank area.

The Brahmaputra spills into its flood basins every year, but the clays settling from the flood waters are deposited slowly, with thickness of two cm or less per annum. However, vegetation is abundant in these flood basins and peat deposits one to four metres in thickness are forming (Coleman, 1969, p. 232-3). These various sub-environments of the braided sandy system are sketched in

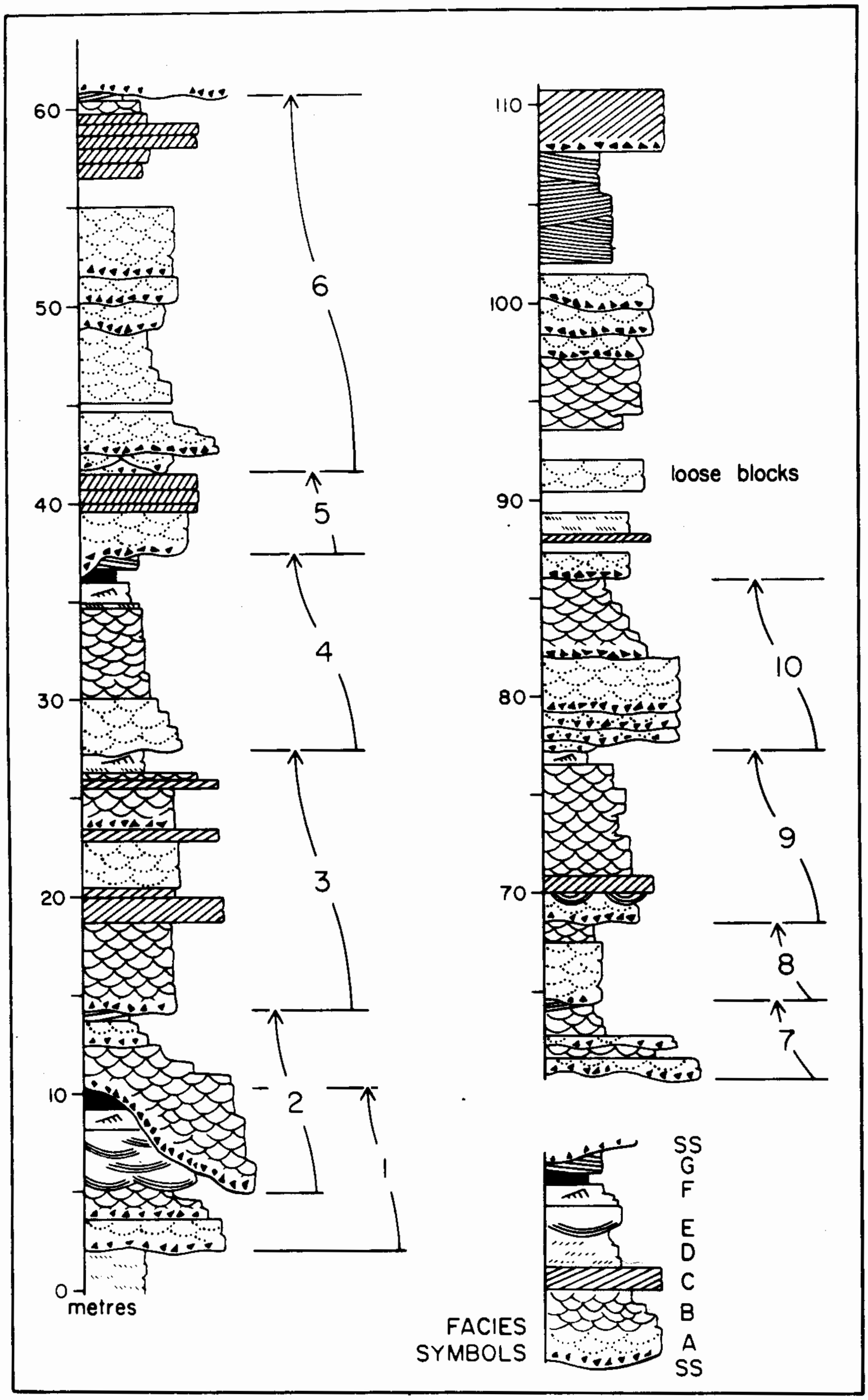


Figure 19
Measured section of the Devonian Battery Point Sandstone, Quebec (from Cant and Walker, 1976). Numbers indicate individual fluvial sequences, and letters define facies: SS = scoured surface; A = poorly-defined trough cross bedding; B = well-defined trough cross bedding; C = large planar tabular cross beds; D = small planar tabular cross beds; E = isolated scour fills; F = ripple cross laminated silts and muds; and G = low angle inclined stratification.

Figure 14, but there is certainly more complexity in the deposits than is indicated in the diagram.

Ancient Sandy Braided Fluvial Deposits
We will discuss in some detail the Battery Point Sandstone of Gaspé, Quebec and the Brownstones of Wales as examples of ancient sandy braided fluvial deposits.

The Battery Point section studied (Cant and Walker, 1976) could be subdivided into 8 facies based on sedimentary structures and grain sizes. At least 10 fluvial sequences of the type summarized in Figure 19 could be identified in the measured section. The order of occurrence of facies was "distilled" (Walker, "General Introduction" to this volume) in order to look for a general facies sequence that could act as a basis for interpretation - see also Miall (1973) and Cant and Walker, 1976, p. 111-114). The end result of the Battery Point distillation is the sequence shown here in Figure 20. It is *not* a model - it is only a summary of a local example that could, in the future, be re-distilled with local examples from other areas to produce a general facies model. In the Battery Point summary sequence, we identified a channel-floor lag overlain by poorly defined trough cross-bedding (Facies A, Fig. 20). The in-channel deposits consisted of well-defined trough cross-bedding (B) and large sets of planar-tabular cross-bedding (C) that commonly showed a large paleocurrent divergence from the trough cross-bedding (Figs. 14 and 20; Cant and Walker, 1976, Fig. 7). The bar-top* deposits consisted mainly of small sets of planar-tabular cross-bedding (D), and the thin record of vertical accretion included cross-laminated siltstones interbedded with mudstones (F), and some enigmatic low-angle cross-stratified sandstones (G).

Upon developing this summary sequence, our first reaction was to compare it to the existing fluvial (meandering) norm (Fig. 2). Although both sequences showed channelled bases, followed by fining-upward sequences, there appeared to be sufficient differences that the norm would *not* act reliably as a basis for interpretation (Walker, "General Introduction" to this volume). In other words, the meandering model of Figure 2 seemed

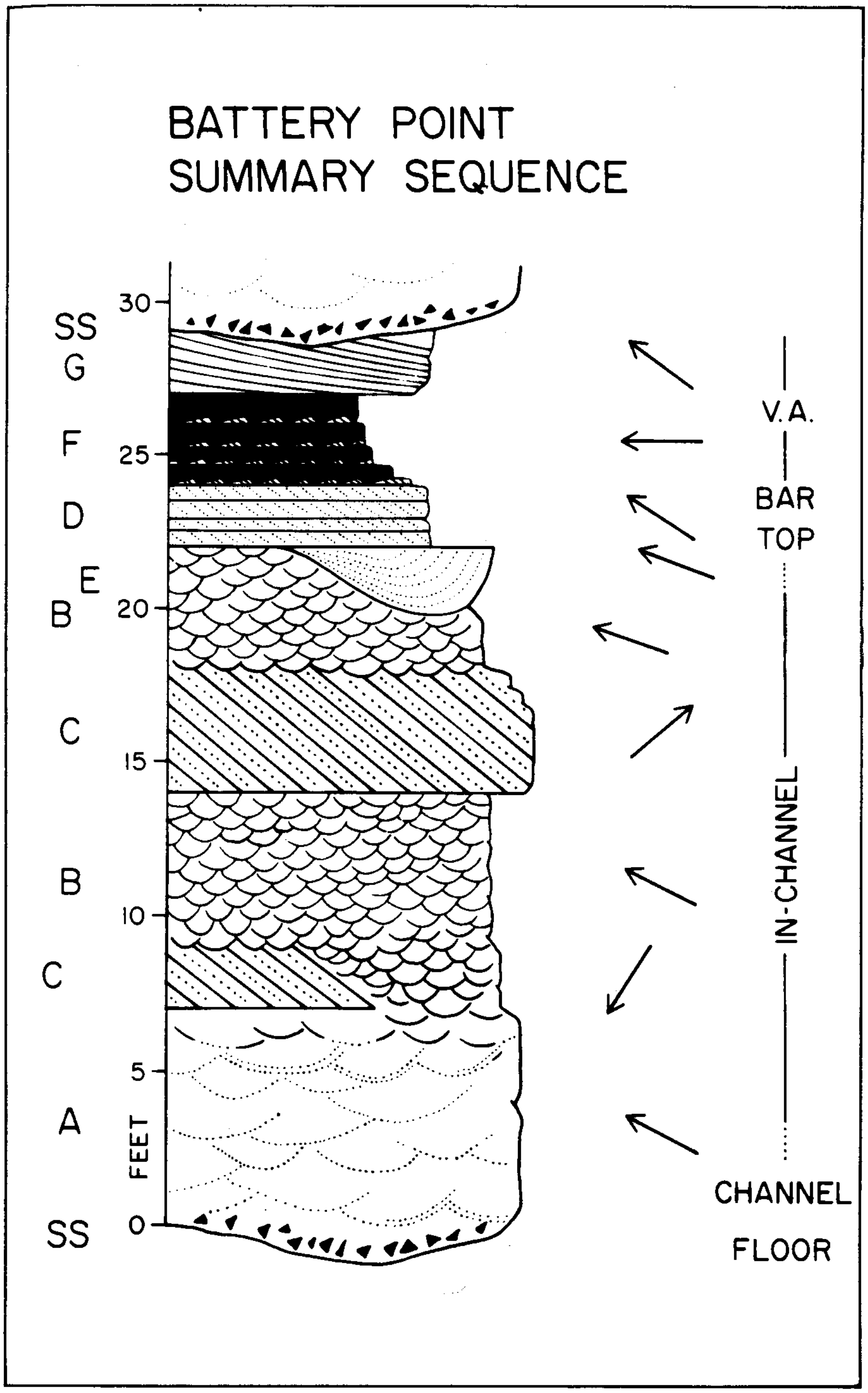


Figure 20
Summary sequence for the Devonian Battery Point Sandstone, Quebec. This sequence was developed by Markov analysis of the facies relationships (see "General Introduction" to this volume), and the preferred facies relationships were drawn as a stratigraphic column using average facies thicknesses. Arrows show paleoflow directions, letters indicate facies (see Cant and Walker, 1976). Compare the Battery Point system, interpreted as braided, with the South Saskatchewan (Fig. 17) and with Allen's (1970) sequence for meandering rivers (Fig. 2).

inappropriate for the Battery Point Sandstone of Figure 20.

Comparison with the norm nevertheless highlighted the major differences, and this gave us added understanding of the Battery Point. Similar comparisons of other systems with the squences in Figures 2 and 20 should also give added understanding. For example, the vertical-accretion deposits in the Battery Point are very thin compared with the meandering norm, both in absolute terms, and in proportion to the amount of in-channel sandstone. The in-channel sandstones do not contain parallel lamination, but planar-tabular sets of cross-bedding are common, and show high paleocurrent divergences from the main channel trend. All of these points of comparison aided in making our "braided" interpretation (Cant and Walker, 1976, p. 115-118). Refer also to annotated comments on the paper by Campbell (1976) in references.

Allen (1983) has interpreted the Devonian Brownstones of Wales as a series of sandy braided stream deposits. These are organized into a series of hierarchical units contained within intraclast-strewn scour surfaces. The 2 to 5 m thick units consist of parallel laminated and trough and planar cross-bedded sandstones and minor conglomerate. Traced laterally, the vertical sequence of sedimentary structures in each unit is extremely variable. The most striking aspect of these units is the almost ubiquitous inclination of planar cross-bed sets. Allen (1983) interprets this as slipface-bounded bars (Fig. 21) accreting laterally onto sandflats in a South Saskatchewan-like mechanism (Fig. 22). This study documents dramatically in 11 long profiles how much local lateral variability we should expect to find in braided stream sequences, as implied in studies of modern streams such as the South Saskatchewan.

Sand Body Geometry And Flood-Plain Aggradation

One major point of contrast with the meandering system is that braided rivers tend to have easily erodible banks, and no clay plugs. The area occupied by the braided river may therefore be very wide (see Campbell, 1976), and coalescing bars and sandflats will result in a laterally continuous and extensive sand sheet unconfined by shales (Fig.

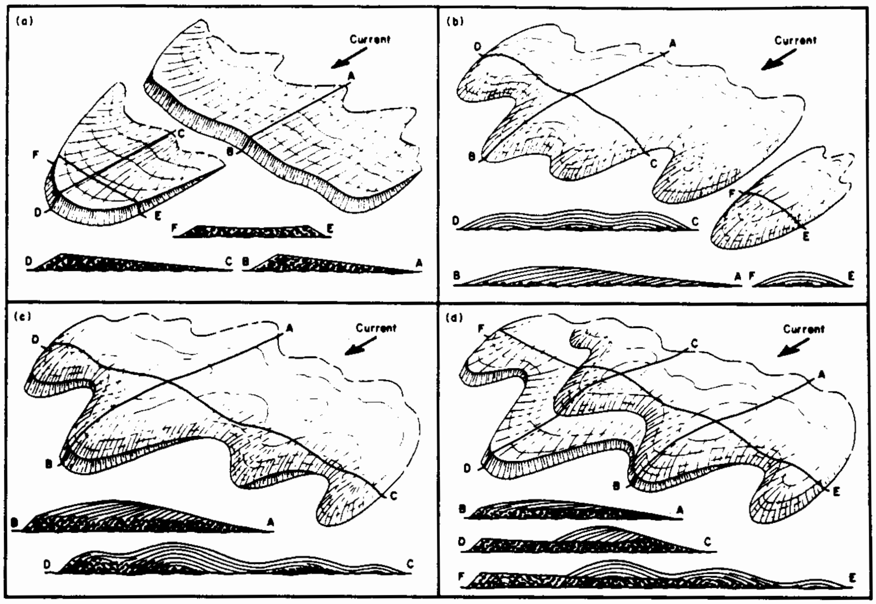
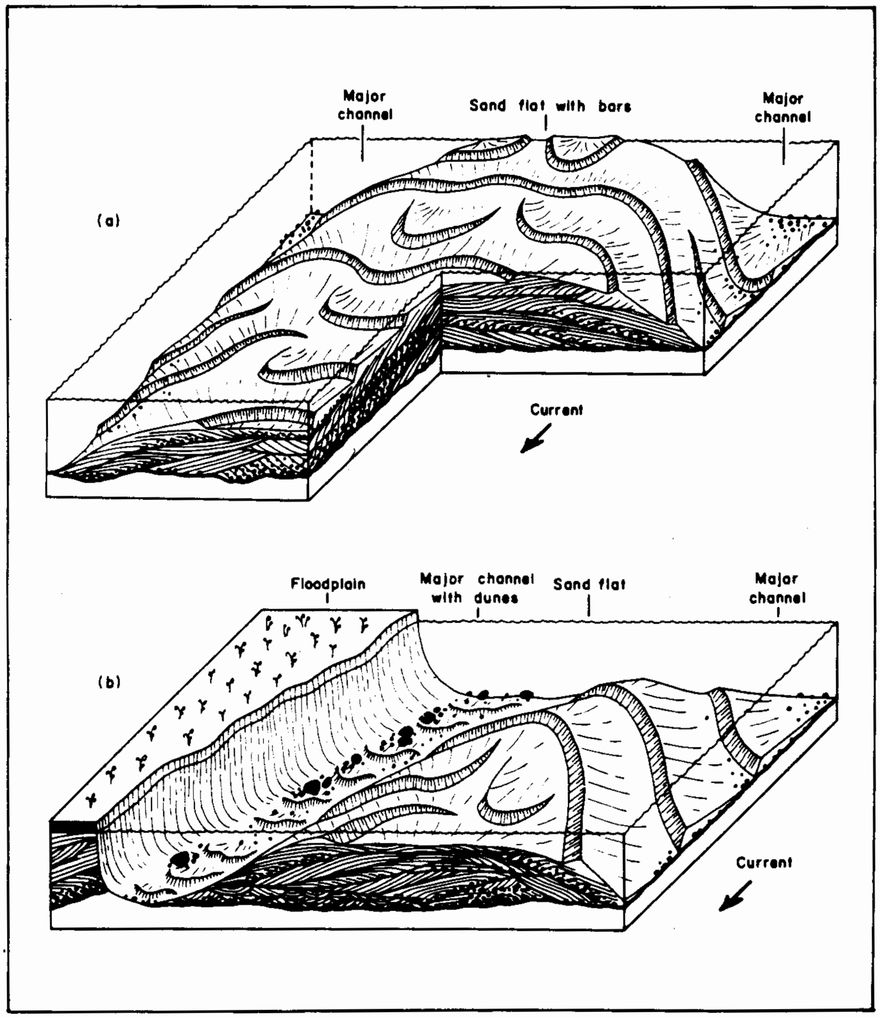


Figure 21
Superb outcrops in the Welsh Borderlands allowed Allen (1983) to reconstruct these four bar types for the Devonian Brownstones. A - cross bedded simple bars; B - plane bedded *simple bars; C* - a compound bar; D - a composite compound bar.



9B). Vertical accretion deposits (if formed) will tend to be quickly eroded because of the comparatively rapid lateral migration of channels. Consequently, any shales preserved in the section will tend to be patchy, laterally discontinuous, and relatively ineffective barriers to vertical hydrocarbon migration. This will not be the case for meandering systems.

Anastomosed Rivers

During the last five years, the anastomosed river has emerged as a type distinct from braided and meandering (Smith and Smith, 1980; Smith and Putnam, 1980; Putnam and Oliver, 1980; Putnam, 1982a; Smith, 1983). There are relatively few documented examples of modern anastomosed rivers (see Smith, 1983), and even fewer ancient examples (see Putnam and Oliver, 1981 and the discussion of this paper by Wightman *et al.*, 1981).

Smith and Smith (1980, p. 157) "use the term anastomosed river for an interconnected network of low-gradient, relatively deep and narrow, straight to sinuous channels with stable banks composed of fine-grained sediment (silt/clay) and vegetation . . . separating the channels are floodplains consisting of vegetated islands, natural levees, and wetlands".

Such streams differ from braided sandy rivers by having stable channel patterns and abundant areas in which fine-grained sediment is deposited and preserved.

Depositional Environments In Anastomosed Systems

Smith and Smith (1980) recognized six main facies in gravelly anastomosed rivers in western Canada (Mistaya, Alexandra, North Saskatchewan, in Alberta).

1) Peat bog facies, containing up to 98% vegetal matter, in layers a few cm to 1.5 m thick.

◀ **Figure 22**
Local facies models for sheet sandstones in the Devonian Brownstones, Welsh Borderlands, from Allen (1983). A - wide sand flat; minor channels related to falling stage may be expected to cross the top of the flat but are not shown. B - proximal part of a sand flat and its structure in strike section, together with a major channel and adjacent flood plain (vegetated island, valley flat). Vertical scales greatly exaggerated in both diagrams.

2) Backswamp facies, composed of silty mud or muddy silt, with variable amounts of organic debris.
3) Floodpond facies, consisting of laminated clay and silty clay with sparse vegetal material. Thickness is up to 6 m.
4) Levee facies, consisting of sandy silt and silty sand containing 10-22% roots by volume. This facies grades into the wetland facies (peat bog, backswamp and floodpond).
5) Crevasse splay facies, making up less than 5% of the vertical accretion facies (1 to 5, above), and consisting of thin layers of sand and/or fine gravel.
6) Channel facies, consisting of gravel and coarse sand, of unknown thickness due to limitations of augering.

These environments are controlled by a rapidly elevating base level at the downstream end of the anastomosed system, causing high rates of aggradation, deposition of fines, and stabilization of river channel patterns. In the geological record, thick vertically accreted sand bodies bounded by wetland facies would be predicted, and a block diagram emphasizing channel confinement and lack of lateral accretion is given in Figure 23. It is emphasized that the data base for this block diagram consists of augered holes a little more than 10 m deep; the aggradational history of these systems is not yet fully documented.

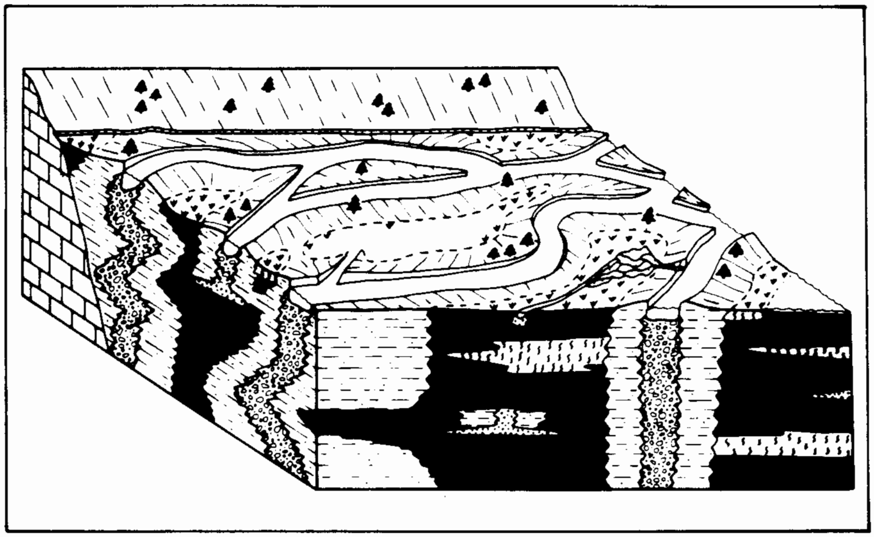


Figure 23
Block diagram of an anastomosed river, from Smith and Smith (1980). In this typical reach, channel sediments (gravel symbol) are bounded by sandy silt (dash, dot symbol) of the levees, which in turn grade into muds and silty muds of the wetlands (black). Peats are shown by small vertical wiggles. Note channel aggradation without significant lateral accretion, the channel pattern being stabilized by the muds and organic material of the wetlands, which are hard to erode.

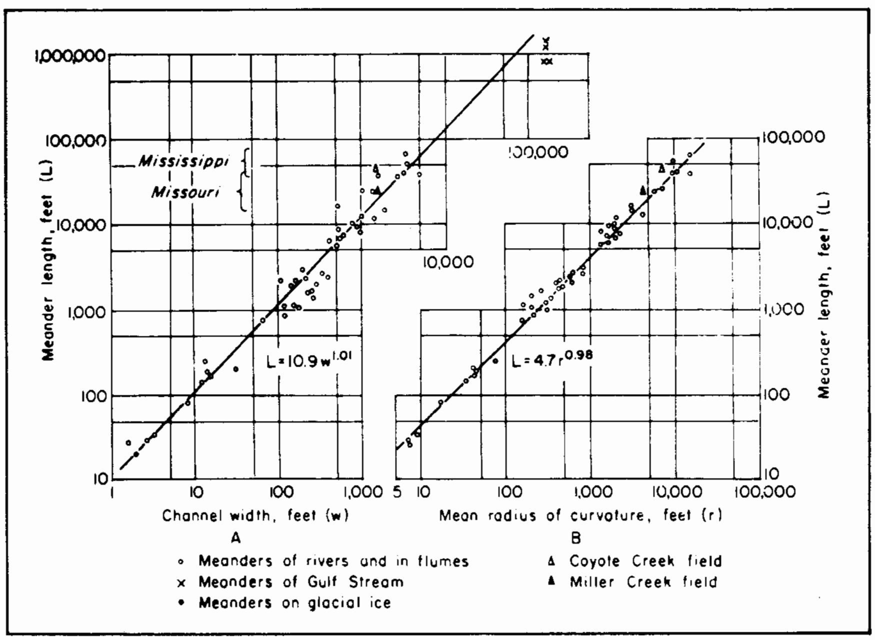


Figure 24
Relationships between meander length, channel width and meander loop radius of curvature. Original data for modern meandering streams from Leopold and Wolman (1960), with Coyote Creek and Miller Creek data plotted by Berg (1968). Note that both Coyote Creek and Miller Creek lie about as far from the regression line as any of the data, suggesting that these two fields differ somewhat from the meandering river "norm", which is represented by the regression line. From Berg (1968).

Ancient Anastomosed Systems

The only ancient system interpreted to be anastomosed is part of the Upper Mannville Group (Albian) of east-central Alberta (Putnam and Oliver, 1980; Putnam, 1980; Putnam, 1982a, 1982b). Here, a pattern of branching and rejoining channels has been illustrated (although it is not clear that all of the channels shown are exactly contemporaneous), and channel sandstones up to 35 m thick can be seen in cores and well-logs. Between the channel sands are "siltstones, shales, coals, and thin (generally less than 6 m thick) sheet-like sandstones which pinch out with increasing distance from the main channel fill" (Putnam, 1982a, p. 438). Some of the ideas and data presented by Putnam and Oliver (1980) have been challenged by Wightman *et al.* (1981), with a reply by Putnam and Oliver (1981). It seems clear that it is premature to propose a general anastomosed fluvial *model* on the basis of so few modern and ancient examples. However, it is important to bear in mind that this type of stream may help to explain or interpret as-yet-undescribed ancient examples that do not fit braided or meandering norms.

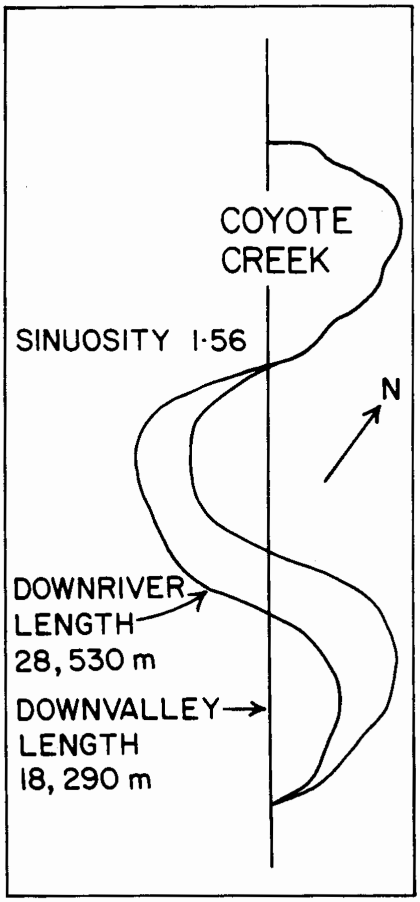


Figure 25
Extrapolation of Coyote Creek meander loop. Assuming a downvalley direction, a downvalley length of 18,290 m (60,000') was plotted. The calculated sinuosity is 1.56 (see text), so the equivalent downriver length is 28,530 m. Two possible river patterns are shown southeast of the Coyote Creek loop – many other patterns are possible, and can be estimated using a scale length of string laid over the diagram and bent into appropriate meander patterns. Details in text.

Fluvial Geomorphological Models And Paleohydraulic Reconstructions

There appears to be a direct and predictable relationship between many meandering river parameters, such as channel width (W), depth (D), sinuosity (P), meander wavelength (L), meander radius (R), slope (S) and mean annual discharge (Q). These relationships were developed by geomorphologists (especially Schumm and Leopold; references in Miall, 1978 and Ethridge and Schumm, 1978), and can be expressed by a series of regression equations (Fig. 24).

The regression equations state in a general way the relationship between parameters based on a large number of examples – that is, the regression lines are *norms*. The equations are likewise *predictors*, but beyond here, the analogy with facies models breaks down. The regression lines are not guides for future observations, nor can they be used as a basis for interpretations.

Earlier in this paper, we discussed the Coyote Creek subsurface point bar (Berg, 1968). Channel depth (roughly equal to point bar thickness) can be estimated at about 23 m (Fig. 10), and channel width (about 460 m), radius of curvature (about 2130 m) and meander length (about 12,200 m) can be estimated from Figure 11. It is important to see whether the Coyote Creek example lies close to the norm – if it does, it can be considered "typical", and one might have some confidence in using other regression equations. However, note in Figure 23 that the Coyote Creek data point lies far from the regression lines, implying that it does *not* closely resemble the "norm". The following calculations involve comparisons of Coyote Creek parameters with various other norms, and hence the results may not be too reliable. They are given here simply to illustrate the possibilities.

Mean annual discharge is estimated by

$$Q = W^{2.43}/18\ F^{1.13} \text{ (Imperial units)} \qquad (1)$$

where F is the width/depth ratio (here 460/23 = 20). The discharge of the Coyote Creek river is thus estimated as 98,280 cubic feet/second, or 2783 m^3/sec.

Sinuosity P is given by

$$P = 3.5\ F^{-0.27} \text{ (Imperial Units)} \qquad (2)$$

and works out to 1.56. This number could be very useful in predicting the position and size of the next meander loop upstream or downstream from Coyote Creek, which is of obvious significance in exploration for hydrocarbons; this prediction is attempted in Figure 25, again for the sole purpose of illustrating *how* a model can be used in prediction.

The reconstructed meander loops in Figure 25 were drawn in the following way:

1) using a downvalley distance of 60,000 feet, the river length is simply 60,000 x sinuousity = 93,600 feet (28,530 m),
2) a piece of string scaled to 93,600 feet was placed over the Coyote Creek meander loop. Its far end was placed at a downvalley distance of 60,000 feet, thus defining the rough size and position of the second and third meander loops,
3) apart from problems of whether it is valid to reconstruct Coyote Creek using the meandering river norm (Fig. 24), note that we also do not know the downvalley paleoslope orientation. We also assume that the Coyote Creek meander loop (Fig. 11) is of average size for this reach of the river, because the downriver distance used to measure sinuosity in the field should normally be taken through as many meanders as possible, not just one.

Meander wavelength (L) is estimated by

$$L = 18\ (F^{0.53}\ W^{0.69}) \text{ (Imperial units)} \qquad (3)$$

and is about 13,687 feet (4172 m).

In Figure 11, note that the estimated meander wavelength is actually 40,000 feet (12,192 m), very different from the 13,687 feet calculated above. There are at least three possible reasons for this discrepancy; 1) Coyote Creek is so different from the norm that this type of analysis is not valid; 2) the channel width has been underestimated, and is closer to 2000 feet (610 m) (Berg, 1968, p. 151); and 3) the point bar thickness is closer to 50 feet (15 m) and the 75 foot (23 m) isopachs represent unusually deep channel floor scours during lateral accretion of the point bar (Fig. 11). Thus with W = 2000 feet and D = 50 feet, equation (2) gives a sinuousity of 1.29 and equation (3) gives a value for L of 24,103 feet (7347 m). This is still much less than the 40,000 feet estimated in Figure 11, and suggests that the Coyote Creek river is too far from the meandering river norm (Fig. 24) for this type of analysis to be valid. The reader can pursue this problem by trying to reconstruct meander loops in Figure 25 using P = 1.29 and L = 24,103 feet.

CONCLUSIONS

The meandering model seems well established and is reasonably well understood. It is a good example of a facies model in that the relationships shown in a block diagram (Fig. 1) are well known from ancient and recent sediments. Also, the simplest vertical facies sequence (Fig. 2) is well established by Markov analysis. Furthermore, numerical predictions based on channel patterns may be possible using a series of regression equations - the equations themselves are a form of model.

Braided streams have a much smaller data base, both in recent and ancient sediments. There are few well-established ancient examples, and there appear to be no convincing and well-documented subsurface examples. Anastomosed systems are even less well understood, and the only proposed ancient example is somewhat controversial.

REFERENCES

BASIC READING

The papers cited here are all listed in the alphabetic reference list below. General review papers include those of Collinson (1978) and Cant (1982). The first papers to consult on meandering streams would probably be Allen (1965, 1970) and Jackson (1976); on braided sandy streams one should begin with Allen (1983) and Cant and Walker (1976, 1978). Miall's (1977) review of the braided system is also very useful, but considers gravelly systems as well as sandy ones.

For a historical review, the most useful paper is that of Miall (1978a), and for papers on exploration, and the subsurface significance of fluvial models, see Horne *et al.* (1978), Berg (1968), Hopkins *et al.* (1982) and Putnam (1982a, 1983).

There are three recent collections of papers which, although written at the technical research level, give an unsurpassed entry into the fluvial literature - see Miall (ed., 1978b), Collinson and Lewin (eds., 1982), and Ethridge and Flores (eds., 1981).

REFERENCES

Allen, J.R.L., 1964. Studies in fluviatile sedimentation: six cyclothems from the Lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, v. 3, p. 163-198.
Classic descriptions and interpretations of six fining-upward sequences.

Allen, J.R.L., 1965. A review of the origin and characteristics of Recent alluvial sediments. Sedimentology, v. 5, p. 89-191.
This remains a classic work, and forms the basis for most point bar models.

Allen, J.R.L., 1970. Studies in fluviatile sedimentation: a comparison of fining-upwards cyclothems, with special reference to coarse-member composition and interpretation. Journal of Sedimentary Petrology, v. 40, p. 298-323.
Presents the results of Markov analysis of sedimentary structure sequence in fining-upward cyclothems (see Fig. 2 of this paper), and continues with a full discussion of lateral accretion.

Allen, J.R.L., 1983. Studies in fluviatile sedimentation: bars, bar-complexes and sandstone sheets (low sinuosity braided streams) in the Brownstones (L. Devonian), Welsh Borders. Sedimentary Geology, v. 33, p. 237-293.
Thorough descriptions of excellent outcrops, with reconstructions of bar patterns within channels.

Barrell, J., 1913. The Upper Devonian delta of the Appalachian geocyncline. American Journal of Science, v. 36, p. 429-472.
Classic descriptions and interpretations of fluvial and deltaic rocks. Recognized fining-upward sequences and their repetitive nature.

Berg, R.R., 1968. Point-bar origin of Fall River sandstone reservoirs, northeastern Wyoming. American Association of Petroleum Geologists, Bulletin, v. 52, p. 2116-2122.
Excellent description of two subsurface point bars. See Figures 10, 11 and 23 of this paper.

Bernard, H.A., Major, C.F., Jr., Parrott, B.S., and LeBlanc, R.J., Sr., 1970. Recent sediments of southeast Texas. Texas Bureau of Economic Geology, Guidebook 11, variously paginated.
First part of guidebook presents some of the Shell work on the Brazos point bars. Illustrates fining-upward sequence and change in sedimentary structures.

Bersier, A., 1968. Sequences detritiques et divagations fluviales. Eclogae Geologicae Helvetiae, v. 51, p. 854-893.
Recognizes in-channel and overbank sequences, and discussed stacking of one channel sequence on top of another. A classic work that deserves more attention.

Blodgett, R.H., and Stanley, K.O., 1980. Stratification, bedforms and discharge relations of the Platte braided river system, Nebraska. Journal of Sedimentary Petrology, v. 50, p. 139-148.
Describes bar types and bar evolution and proposes a thin (1.5 m) idealized vertical structure sequence.

Campbell, C.V., 1976. Reservoir geometry of a fluvial sheet sandstone. American Association of Petroleum Geologists, Bulletin, v. 60, p. 1009-1020.
Describes lateral extent of sandbodies in part of the Morrison Formation. Proposes a sequence of structures that is dominated by trough cross bedding, and hence one that is very different from the South Saskatchewan sequences.

Cant, D.J., 1982. Fluvial facies models and their application. *In* Scholle, P.A., and Spearing, D.R., eds., Sandstone depositional environments. American Association of Petroleum Geologists, Memoir 31, p. 115-137.
A useful and well illustrated review, and a very good follow-up to this paper.

Cant, D.J., and Walker, R.G., 1976. Development of a braided fluvial facies model for the Devonian Battery Point Sandstone, Quebec. Canadian Journal of Earth Sciences, v. 13, p. 102-119.
Describes facies and establishes a summary facies sequence using Markov chain analysis. The data are used in the "General Introduction" of this volume, and the study is one of the better-documented braided fluvial examples.

Cant, D.J., and Walker, R.G., 1978. Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. Sedimentology, v. 25, p. 625-648.
Describes the morphological features of the river, and establishes a sequence of sand flat development from small bars. Proposes facies sequences for sand flat and channel aggradation (see Figs. 14-19, this paper).

Coleman, J.M., 1969. Brahmaputra River: channel processes and sedimentation. Sedimentary Geology, v. 3, p. 129-239.
Full description of the Brahmaputra, with preliminary discussion of sand flat development.

Collinson, J.D., 1970. Bedforms of the Tana River, Norway. Geografiska Annaler, v. 52A, p. 31-56.
Classic description of the Tana, emphasizing sand bars. Discusses changes due to reactivation, but does not propose an overall sedimentary structure sequence.

Collinson, J.D., 1978. Alluvial sediments. *In* Reading, H.G., ed., Sedimentary environments and facies. Oxford, Blackwell, p. 15-60.
An excellent general review of alluvial sediments.

Collinson, J.D., and Lewin, J., eds., 1982. Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 5, 535 p.
Very useful collection of papers, grouped into sections on Hydrodynamics and Bedforms, Present Day Channel Processes, Facies Models, and Economic Aspects. A very useful source of data and references.

Conaghan, P.J., and Jones, J.G., 1975. The Hawkesbury Sandstone and the Brahmaputra: a depositional model for continental sheet sandstones. Journal of the Geological Society of Australia, v. 22, p. 275-283.
Proposes a braided fluvial origin for the Hawkesbury, and gives an idealized vertical sedimentary structure sequence.

Doeglas, D.J., 1962. The structure of sedimentary deposits of braided rivers. Sedimentology, v. 1, p. 167-190.
Classic descriptions of the Durance and Ardeche Rivers in France. The paper has perhaps been overlooked in the last few years because it does not present overall sedimentary structure sequences - nevertheless, the descriptions remain extremely useful.

Dixon, E.E.L., 1921. The geology of the South Wales coalfield; Part XII: the country around Tenby. Geological Survey of Great Britain, Memoir.
In this paper, the repetitive nature of what we now call a fining-upward sequence was recognized and stated more clearly than by Barrell. This paper, therefore, contains the first explicit statement of repetitive, superimposed fluvial sequences.

Edwards, M.B., Eriksson, K.A., and Kier, R.S., 1983. Paleochannel geometry and flow patterns determined from exhumed Permian point bars in north-central Texas. Journal of Sedimentary Petrology, v. 53, p. 1261-1270.
Reconstruction of flow associated with obliquely-oriented ripples on point bar surfaces.

Ethridge, F.G., and Flores, R.M., eds., 1981. Recent and ancient nonmarine depositional environments: models for exploration. Society of Economic Paleontologists and Mineralogists, Special Publication 31, 349 p.
Contains a series of papers, written at the full technical level, grouped into: 1) recent and ancient nonmarine depositional models: 2) alluvial fan and fluvial deposits; 3) lacustrine deposits: and 4) eolian deposits.

Ethridge, F.G. and Schumm, S.A., 1978. Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations and assessment. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 703-721.
A very useful summary of paleohydraulic reconstruction methods, with all assumptions spelled out, and equations and their sources given.

Fisk, H.N., 1944. Geological investigation of the alluvial valley of the lower Mississippi River. Vicksburg, Miss., Mississippi River Commission, 78 p.
A classic description of the river pattern of the lower reaches of the Mississippi.

Frazier, D.E., and Osanik, A., 1961. Point bar deposits, Old River locksite, Louisiana. Gulf Coast Association of Geological Societies, Transactions, v. 11, p. 121-137.
One of the first large excavations within a fluvial system, with discussions of the geometry of stratification.

Harms, J.C., and Fahnestock, R.K., 1965. Stratification, bed forms and flow phenomena (with an example from the Rio Grande). *In* Middleton, G.V., ed., Primary sedimentary structures and their hydrodynamic interpretations. Society of Economic Paleontologists and Mineralogists, Special Publication 12, p. 84-115.
Classic paper relating flow phenomena to bed forms, and bed forms to the resulting sedimentary structures.

Harms, J.C., Mackenzie, D.B., and McCubbin, D.G., 1963. Stratification in modern sands of the Red River, Louisiana. Journal of Geology, v. 71, p. 566-580.
One of the first descriptions of the internal structure of point bars, emphasizing the development of stratification from bed forms.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982. Structures and sequences in clastic rocks. Society of Economic Paleontologists and Mineralogists, Short Course 9, variable pagination.
Revised version of Short Course 2, with important introductory chapters by Southard and Harms on flow, bed forms and stratification.

Haszeldine, R.S., 1983. Fluvial bars reconstructed from a deep, straight channel, Upper Carboniferous coalfield of northeast England. Journal of Sedimentary Petrology, v. 53, p. 1233-1247.
Excellent and detailed reconstruction of the internal form of stratification within a bar 10 m high and 200 m wide. Lobate sandwaves 20 m wide descended the lee face of the bar.

Hopkins, J.C., Hermanson, S.W., and Lawton, D.C., 1982. Morphology of channels and channel-sand bodies in the Glauconitic Sandstone Member (Upper Mannville), Little Bow area, Alberta. Bulletin of Canadian Petroleum Geology, v. 30, p. 274-285.
Reconstruction of channel patterns from subsurface data. See Figures 12 and 13 of this paper. Well logs supported by core photographs.

Horne, J.C., Ferm, J.C., Caruccio, F.T., and Baganz, B.P., 1978. Depositional models in coal exploration and mine planning in Appalachian region. American Association of Petroleum Geologists, Bulletin, v. 62, p. 2379-2411.
An excellent and well illustrated study of how fluvial (and marginal marine) models can be used in exploration.

Jackson, R.G., 1976. Depositional model of point bars in the lower Wabash River. Journal of Sedimentary Petrology, v. 46, p. 579-594.
Describes point bar migration, and relates grain size and sedimentary structure sequences to their position within the meander loop. Possibly the most detailed description of a North American meandering river since Fisk, 1944.

Jones, B.G. and Rust, B.R., 1983. Massive sandstone facies in the Hawkesbury Sandstone, a Triassic fluvial deposit near Sydney, Australia. Journal of Sedimentary Petrology, v. 53, p. 1249-1259.
Relates massive sandstone to liquefaction during foreset failure at low flow. At flood flows, the foresets may have been up to 15 m high.

Kelling, G., 1968. Patterns of sedimentation in Rhondda Beds of South Wales. American Association of Petroleum Geologists, Bulletin, v. 52, p. 2369-2386.
Describes two types of vertical stratification sequence, and interprets the Rhondda beds as having been deposited both from low and high sinuosity streams.

McGowen, J.H., and Garner, L.E., 1970. Phisiographic features and stratification types of coarse-grained point bars: modern and ancient examples. Sedimentology, v. 14, p. 77-111.
Very useful descriptions of the Amite (La.) and Colorado (Tx.) Rivers, and comparisons with Eocene and Pleistocene deposits. Stratification and internal structure sequences emphasized.

Miall, A.D., 1973. Markov chain analysis applied to an ancient alluvial plain succession. Sedimentology, v. 20, p. 347-364.
Applies Markov chain analysis to Devonian fluvial rocks in arctic Canada, and proposes a test of significance for the probability matrices as a whole. Harper's method tests each entry in the matrix (see "Improved Methods of Facies Sequence Analysis" in this volume).

Miall, A.D., 1977. A review of the braided river depositional environment. Earth Science Reviews, v. 13, p. 1-62.
Miall reviews sandy and gravelly braided systems, and proposes a standard set of facies for their description. Distillation of these facies into sequences suggests four main types - the Scott, Donjek, Platte and

Bijou Creek. Others have been added subsequently (see Miall, 1978b).

Miall, A.D., 1978a. Fluvial sedimentology: an historical review. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 1-47.
Undoubtedly the most thorough historical review of fluvial sedimentology, beginning with observations in prehistoric times and continuing to the present day. An invaluable source for older references.

Miall, A.D., ed., 1978b. Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, 859 p.
Collection of many papers, written at the technical research level, and grouped into Texture and Structure, Bedforms and Bars. Modern Rivers, Ancient Fluvial Systems, Fluvial Facies Models, Paleohydraulics, and Economic Applications. Overall an extremely useful source of data and references.

Moody-Stewart, M., 1966. High- and low-sinuosity stream deposits, with examples from the Devonian of Spitsbergen. Journal of Sedimentary Petrology, v. 36, p. 1102-1117.
Comparison of low and high sinuosity streams and their deposits. Does not go into details of stratification sequences.

Nami, M., 1976. An exhumed Jurassic meander belt from Yorkshire. Geological Magazine, v. 113, p. 47-52.
Brief description of an exhumed meander belt, with epsilon cross strata and plan views of the channel pattern. More detail in Nami and Leeder (1978), below.

Nami, M., and Leeder, M.R., 1978. Changing channel morphology and magnitude in the Scalby Formation (M. Jurassic) of Yorkshire, England. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 431-440.
Documents upward change from low to high sinuosity channels within one formation. Shows internal structure sequences and plan views of exhumed channels.

Nijman, W., and Puigdefabregas, C., 1978. Coarse-grained point bar structure in a molasse-type fluvial system, Eocene Castisent Sandstone Formation, south Pyrenean basin. *In* Miall, A.D., ed., Fluvial sedimentology. Canadian Society of Petroleum Geologists, Memoir 5, p. 487-510.
Describes a well-exhumed coarse grained point bar near Castisent, showing cut bank, point bar, chutes, and bed forms. Presents vertical stratification sequences and map views.

Putnam, P.E., 1982a. Fluvial channel sandstones within Upper Mannville (Albian) of Lloydminster area, Canada - geometry, petrography and paleogeographic implications. American Association of Petroleum Geologists, Bulletin, v. 66, p. 436-459.
Subsurface description of a possibly anastomosed fluvial system.

Putnam, P.E., 1982b. Fluvial deposits and hydrocarbon accumulations: examples from the Lloydminster area, Canada. *In* Collinson, J.D., and Lewin, J., eds., Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, p. 517-532.
Subsurface description of a possibly anastomosed fluvial system.

Putnam, P.E., 1980. Fluvial deposition within the Upper Mannville of west-central Saskatchewan: stratigraphic implications. *In* Beck, L.S., *et al.*, eds., Lloydminster and beyond: geology of Mannville hydrocarbon reservoirs. Regina, Saskatchewan Geological Society, Special Publication 5, p. 197-216.
Subsurface description of Mannville channels in Saskatchewan - a possible anastomosed system.

Putnam, P.E., and Oliver, T.A., 1980. Stratigraphic traps in channel sandstones in the upper Mannville (Albian) of east-central Alberta. Bulletin of Canadian Petroleum Geology, v. 28, p. 489-508.
Description of subsurface Mannville channels, and an explicit interpretation in terms of anastomosed channels. But see discussion by Wightman *et al.* (1981) and reply by Putnam and Oliver (1981).

Putnam, P.E., and Oliver, T.A., 1981. Stratigraphic traps in channel sandstones in the upper Mannville (Albian) of east-central Alberta: Reply. Bulletin of Canadian Petroleum Geology, v. 29, p. 626-629.
Authors reply here to the criticisms of Wightman *et al.* (1981).

Smith, D.G., 1983. Anastomosed fluvial deposits: modern examples from western Canada. *In* Collinson, J.D., and Lewin, J., eds., Modern and ancient fluvial systems. International Association of Sedimentologists, Special Publication 6, p. 155-168.
Comparison of the Columbia (B.C.) and Saskatchewan (Sask.) Rivers, with very useful data from coring (up to 15 m) the floodplains, and the facies relationships thus revealed.

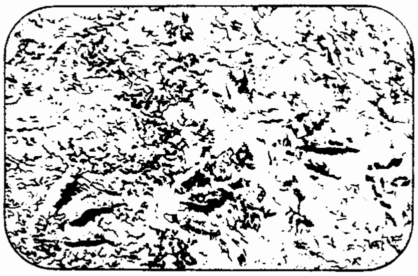
Smith, D.G., and Putnam, P.E. 1980. Anastomosed fluvial deposits, modern and ancient examples from Alberta, Canada. Canadian Journal of Earth Sciences, v. 17, p. 1396-1406.
Comparison of several ancient sandstones (Upper Mannville, Brazeau-Paskapoo, Kootenay) with the Alexandra (Alberta), Columbia (B.C.) and Saskatchewan (Sask.) rivers; emphasizes the anastomosed model.

Smith, D.G., and Smith, N.D., 1980. Sedimentation in anastomosed river systems: examples from alluvial valleys near Banff, Alberta. Journal of Sedimentary Petrology, v. 50, p. 157-164.
Describes three anastomosed rivers (Mistaya, N. Saskatchewan, Alexandra) and suggests six sedimentary facies.

Smith, N.D., 1970. The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians. Geological Society of America, Bulletin, v. 81, p. 2993-3014.
A classic study contrasting longitudinal and transverse bars in the Platte River, and comparing the results with stratification sequences in the Shawangunk Conglomerate.

Sundborg, A., 1956. The River Klaralven, a study of fluvial processes. Geografiska Annaler, v. 38, p. 217-316.
A classic study of fluvial processes and flood plain geomorphology. Predates the "sedimentological" approach to fluvial studies of the 1960s, so does not discuss in detail the internal structures or sequences.

Wightman, D.M., Tilley, B.J., and Last, W.M., 1981. Stratigraphic traps in channel sandstones in the upper Mannville (Albian) of east-central Alberta: Discussion. Bulletin of Canadian Petroleum Geology, v. 29, p. 622-625.
Criticism of Putnam and Oliver's (1980) paper, questioning the application of the anastomosed model to the Upper Mannville, and questioning the validity of the channel sandstone trends. See reply by Putnam and Oliver (1981).



Eolian Sands

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INTRODUCTION

The beautiful large-scale crossbedding of ancient wind-blown sandstones has often fascinated sedimentologists, but until recently no-one has attempted to construct facies models for eolian deposits in any detail. Perhaps this is due to the difficulty of connecting modern surface structures and morphology with the form of cross-bedding seen in sections of ancient eolian sandstones. However, similar problems have not hindered the development of turbidite fan models. Perhaps, as noted in an earlier version of this chapter (Walker and Middleton, 1979), there is no preferred vertical sequence or consistent lateral change. I do not believe that this is the case, and in this paper I will first note the basis of eolian facies models, including the major problems; then look at some features of modern eolian sands, develop some facies models, and see how they can be applied to ancient eolian sandstones.

BASIS FOR EOLIAN MODELS

The facies models are based on the idea of migrating hierarchies of eolian bedforms 'climbing' over one another (Allen, 1963; Banks, 1973; Brookfield, 1977; Rubin and Hunter, 1982). Different types of bedforms in modern deserts show different types and proportions of stratification (Hunter, 1977). These bedforms, when allowed to migrate, generate assemblages of strata whose sections can then be compared with actual sections of ancient eolian sandstones.

Unfortunately there are some major problems with this approach. Recent large eolian bedforms commonly rest on alluvium, and are usually post-Pleistocene in age (Wilson, 1973). No thick eolian deposits comparable to some ancient examples seem to be forming at present. Due to changes of wind circulation patterns during the Quaternary glacial epoch, and to the enormous lag time of the larger eolian bedforms, it seems likely that many, if not most, recent large bedforms are not in equilibrium with the present wind pattern (Wilson, 1973). The stratification types of Hunter (1977), which are used to determine the locus of deposition of an ancient eolian sandstone (described below), were described from small coastal dunes. McKee's (1966) trench sections through dunes, often used to interpret ancient eolian sandstone sections, were in small gypsum dunes which are easily stabilized by occasional wetting during rain. Thus, we do not know if the stratification types and cross-sections can be applied to the internal structures of large quartz sand eolian bedforms. Recent work (Clemmensen and Abrahamson, 1983) indicates that the stratification types are similar in recent and ancient quartz sand dunes, and that, with some modification, McKee's (1966) sections provide a guide to dune type - though not to bedform size or complexity. Direct comparison between recent dune trenches and ancient eolian sandstone sections suffers from an additional problem, namely, that only the lowest parts of eolian bedforms are usually preserved. Thus we have to reconstruct the large bedforms almost entirely from structures formed at their basal lee slopes.

MODERN EOLIAN SANDS

Modern eolian sands occur in two main settings: sandy deserts and coastal dunes. Deposits in deserts are by far the most extensive. Arid and semi-arid regions occupy about one third of the present land surface and include three main sedimentary environments: alluvial fans and ephemeral streams, inland sabkhas or playas, and sandy deserts - also called "sand seas" or ergs (Fig. 1).

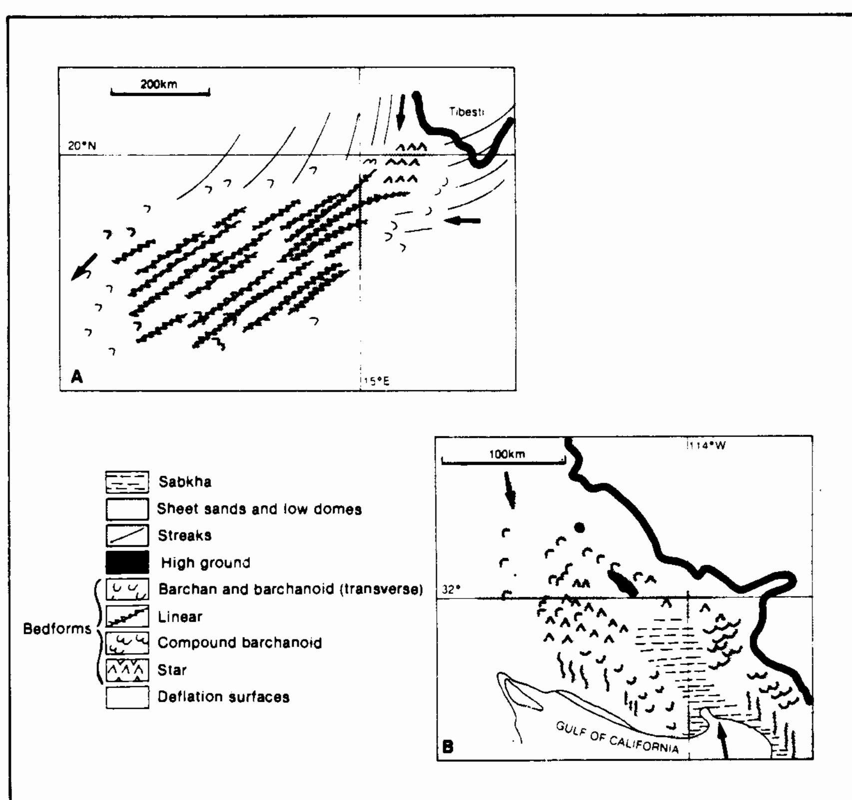


Figure 1
Bedform map of: A) the Fachi Bilma Erg, southeastern Sahara (after Mainguet and Callot, 1974; B) El Gran Desierto, Sonora, Mexico (M. Brookfield unpub.). Note difference in scale: Mexican desert is only one quarter the size of the Fachi Bilma erg. Arrows are dominant wind directions.

Sandy deserts form only about twenty per cent of the area of modern deserts. The rest is composed of eroding mountains (40%) and stony areas or serirs (10 to 20%) and desert flats (10 to 20%) with smaller areas of dry washes, volcanic cones and badlands, where erosion rather than deposition takes place (Cooke and Warren, 1973, p. 53). Since our ideas about modern environments tend to be strongly influenced by personal observation, it is worth noting that the desert areas of North America (with the exception of El Gran Desierto, Sonora, Mexico; Fig. 1b) are atypical. In America, alluvial fans are much more important (30%) and sandy deserts much less important (less than 1%) than in other major deserts.

The largest desert in the world is the Sahara (7 million km^2). It has several major ergs arranged in three belts. Individual ergs cover areas as large as 500,000 km^2 (twice the area of Nevada). They are generally located in physiographic or structural basins, with long histories of sediment accumulation, including extensive Tertiary and Pleistocene fluvial sediments. The modern eolian deposits, however, are rarely more than 100 m thick (see Mainguet, 1976; Mainguet and Callot, 1974; Wilson, 1971, 1972, 1973). The main reason for the accumulation of sand in an erg seems to be the presence of a topographic depression. Once formed, the dunes themselves can trap more sand (Bagnold, 1941). Mainguet and Callot (1974) made a study of the Fachi-Bilma erg, based on air and satelite photography, with some features checked in the field. This erg (Fig. 1a) is situated in the southern Sahara. The easterly trade winds are deflected by the Tibesti massif, and the erg is situated where the winds converge again. Within the erg, there is a definite spatial zonation of dune types. Barchans occur on all sides of the erg, marking zones of intermittent deposition. Inwards, these coalesce into larger, less mobile, sinuous longitudinal (seif) dunes and then into larger, fully developed, compound longitudinal draa (silks). In the upwind part of the erg (in the 'wind shadow' of the Tibesti massif, a zone of variable winds) is a zone of large pyramidal draas (star draa) over 100 m high and 0.5 to 1.5 km wide: these star draas are themselves arranged in regular geometric patterns or rows. This erg is fairly typical. For comparison, the only large American erg is shown on Figure 1b. We will use these two ergs as motifs for the development of facies models based on deserts with dominantly longitudinal and dominantly transverse bedforms. Breed *et al.* (1979) show bedform maps, based on satellite photography of many of the major ergs in the world, in which similar patterns can be seen.

Wilson's (1971, 1972) studies of Saharan bedforms led him to propose three main scales of eolian bedforms; ripples, dunes and draas. Ripples are flatter than those in water and usually have more regular crest lines (Sharp, 1963). Dunes have transverse and longitudinal morphologies and vary from 0.1 to 100 m in height. Draas are large sand bedforms between 20 and 450 m high, characterized by the superimposition of smaller dunes on them. Nevertheless, the dune-draa distinction is not universally accepted, and a descriptive classification based on form and complexity should probably be used (Table 1).

Dune form has been related by Fryberger (1979) to wind variability and sand transport ability (Fig. 2). Drift potential (D.P.) is the relative amount of sand *potentially* moved by the wind for a stated period of time, weighted for the wind velocity. Sand drift roses are then calculated for the different wind directions. The sum of these sand drifts for all wind directions is the drift potential. The resultant drift direction (R.D.D.) is the direction of the vector resultant of all drift potentials: similarly the resultant drift potential (R.D.P.) is the vector sum of the drift potentials. High R.D.P./D.P. values indicate low directional variabil-

Table 1
Morphology and classification of eolian bedorms (after McKee, 1979)

Morphology	Name	Associations
Sheet-like	Sheet sands	
Thin elongate strips	Streaks	COMPOUND - two or more of the same type combined by overlap or superimposition (Wilson's draa).
Circular to elliptical mound, dome-shaped	Dome	
Crescent in plan	Barchan	
Connected crescents	Barchanoid (akle)	COMPLEX - two different basic types occuring together, either superimposed (Wilson's draas), or adjacent.
Asymmetrical ridge	Transverse (Reversing)	
Symmetrical ridge	Linear (seif)	
Central peak with arms	Star (pyramidal)	
U-shaped	Parabolic	

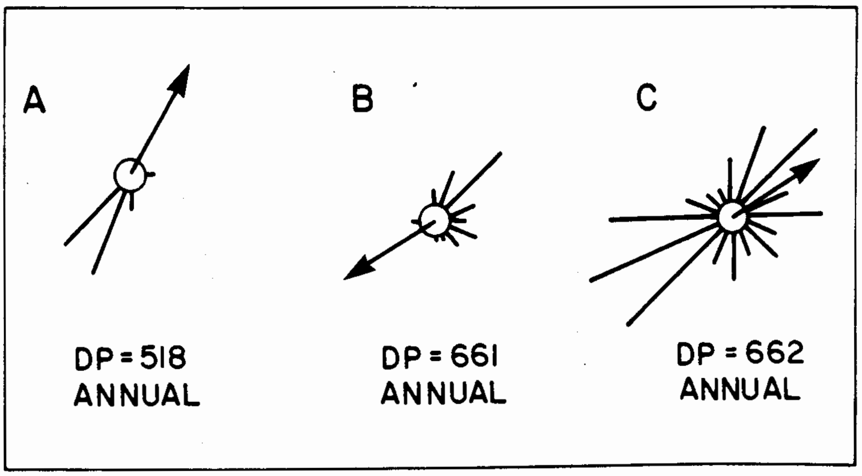


Figure 2
Characteristic high-energy wind regimes. A) narrow unimodal; barchanoid dunes, Peru. B) bimodal; linear dunes, Mauritania. C) complex; star dunes, Libya. Arrows indicate resultant drift direction (RDD) (After Fryberger, 1979).

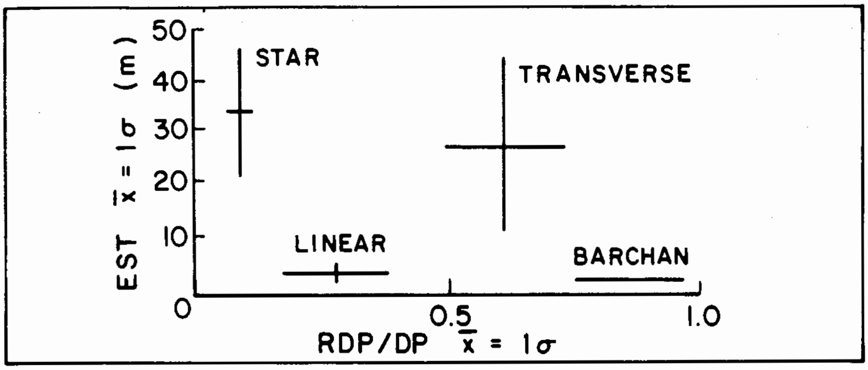


Figure 3
Statistically significant (P = 0.001) separation of the four elemental dune types by means of two variables: equivalent sand thickness (EST) and a measure of wind directional variability (RDP/DP). The $\bar{x} \pm 1\ \sigma$ for EST in barchans is 0.02 ± 0.005 m (n = 8) (from Wasson and Hyde, 1983a).

ity. Barchanoid and transverse dunes occur in areas of low directional variability; linear dunes in more variable wind regimes; and star bedforms in very variable regimes. Wasson and Hyde (1983a) found that various bedforms plotted in separate fields if wind variability was plotted against 'spread out' sand thickness (Fig. 3), indicating that the form of the dunes may be controlled both by wind regime and availability of sand.

Ahlbrandt (1979) summarized the textural parameters of eolian sand in terms of three sub-environments: well-sorted to very well-sorted fine coastal dune sands; moderately sorted to well-sorted fine to medium grained inland dune sands; and poorly-sorted interdune and serir sands. The dune samples vary in mean grain-size from 0.68 phi (1.6 mm) to 3.4 phi (0.1 mm). Most interdune and serir samples are bimodal in the sand fraction and have higher silt and clay contents when compared with adjacent dune samples from dunes formed under different wind regimes. In inland dune fields, with predominantly unidirectional winds, there is progressively improved sorting and finer mean grain-size downwind from the sand source in a sequence of dome, transverse, barchanoid and parabolic dunes. In reversing and multidirectional wind regimes, sand accumulates in dunes that have little net lateral migration. The clastic material in these dunes is a combination of available source material, and because of the fluctuating conditions crest and base tend to have more divergent means than in unidirectional wind regimes. Wilson (1972, 1983) attempted to show that differing bedform hierarchies could be separated by plotting the coarsest 20th percentile of the sand against the bedform wavelength – which would be nice if true, since we could then estimate the size of the ancient bedforms from grain-size and bounding surface data (discussed below). Unfortunately, his results have not been confirmed; and one study on Australian deserts shows no such relationship (Wasson and Hyde, 1983b).

Surface structures on eolian bedforms are among their more characteristic features and include ripple marks (really themselves a bedform), adhesion structures, animal tracks and trails, marks made by vegetation and rain or hail. All may be preserved by ripple climbing and grainfall deposition (see Kocurek and Fielder, 1982; Ahlbrandt and Fryberger, 1981; McKee, 1982; Steidtmann, 1973). Examples of these structures are shown in Figure 4.

Internal structures are difficult to study in modern dunes. Large-scale cross-bedding was formerly considered typical of eolian sandstones, but in fact also occurs in submarine bedforms, (commonly less than about 4 m; see "Shelf and Shallow Marine Sands", this volume) where it may closely resemble eolian cross-bedding (cf. Allen, 1982). The detailed structures of the laminae do, however, differ. Hunter (1977) proposed four main types of eolian laminae from study of small coastal dunes (Fig. 5).

Planebed lamination is produced by wind velocities too high for ripple formation and is analogous to upper flat bed in aqueous deposits.

Climbing ripple lamination closely resembles similar aqueous varieties, but because of the difficulty of recognizing the ripple foresets in eolian ripples (due to their low relief), Hunter (1977) distinguished two main types. *Translatent strata* occur where only the bounding surfaces between ripples are visible: *rippleform strata* occur when the ripple foresets can be identified. Both are inversely graded and relatively closely packed (porosity average 39%).

Grainfall lamination is produced in zones of flow separation by deposition from suspension. Grain segregation is relatively poor and laminae difficult to see. Packing is intermediate between the closely-packed traction deposits above and the loosely-packed sandflow strata below (average porosity 40%).

Sandflow cross-stratification (avalanche cross-bedding) is caused by slumping and consequent grain flow down slopes. Sandflow cross-strata are loosely packed (average porosity 45%), interfinger with grainfall laminae near their base, and form lenses in horizontal exposures.

Examples of all these stratification types have now been found in ancient eolian sandstones (Hunter, 1981; Fryberger and Schenk, 1981; Clemmensen and Abrahamsen, 1983). Each type is found in different parts of a dune, and can be recognized even where the dune has been deflated (Fig. 6). In this example, we can recognize the basic features of a simple barchanoid form; the slipfaces (sandflow cross-strata), the saddles (climbing translatent strata) and the passage between the slipfaces and saddles (grainfall laminae). Further information on stratification types and their relationship to dune morpholpgy is given by Hunter and Rubin (1983), Rubin and Hunter (1983) and Hunter (1981).

Our knowledge of the larger internal structures of modern dunes is still limited and due almost entirely to the work of McKee (McKee, 1966; McKee and Douglass, 1971; McKee and Tibbitts, 1964; McKee and Moiola, 1975) and Bigarella (Bigarella, 1972; Bigarella *et al.*, 1969). These studies are summarized in McKee (1979). The most detailed studies were done in the White Sands dune field of New Mexico which is not analogous to most ancient eolian deposits. The dunes are composed of gypsum and are easily stabilized by occasional wettings by rain. Although a close study of these sections is enlightening, space precludes their detailed consideration.

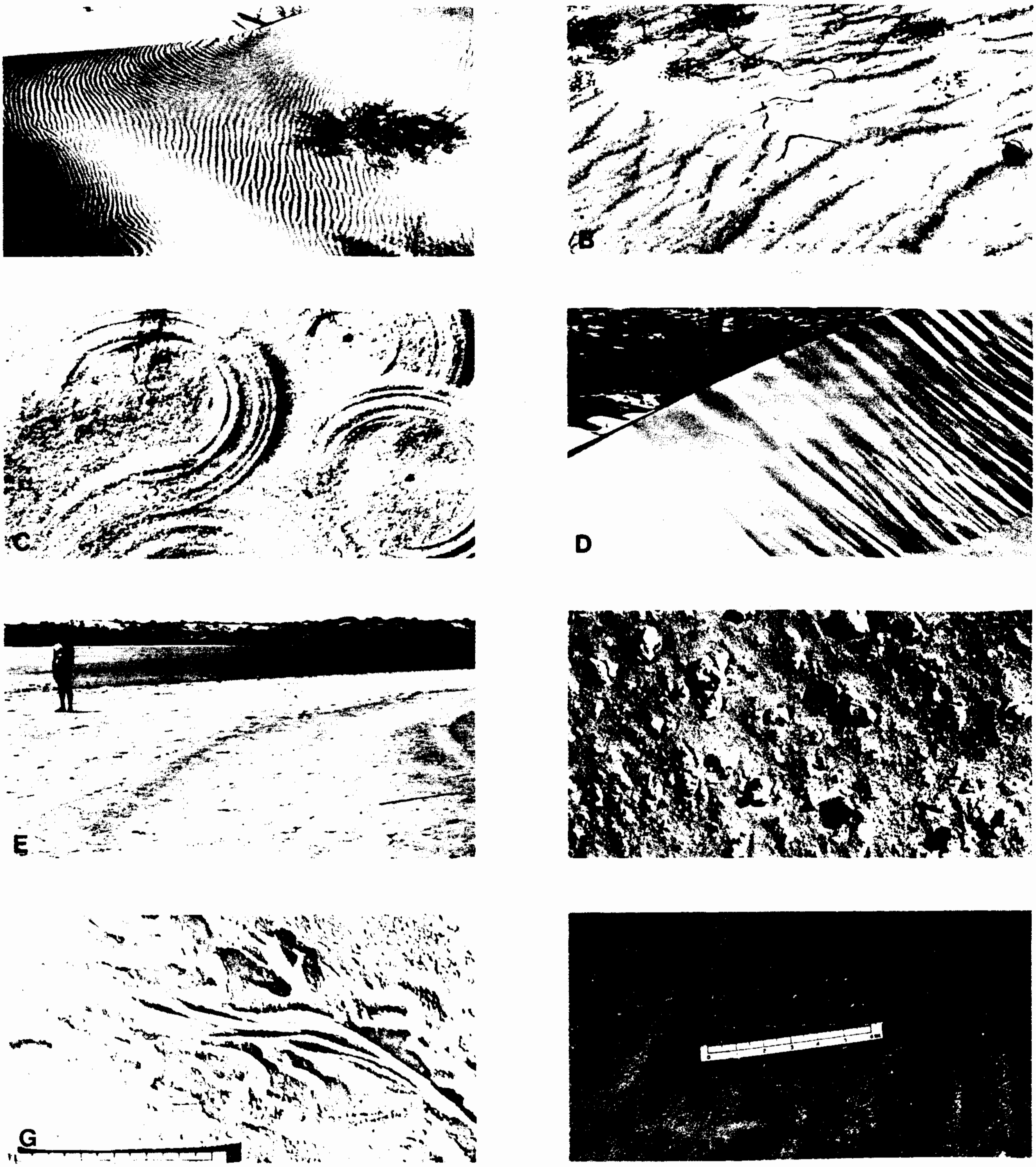
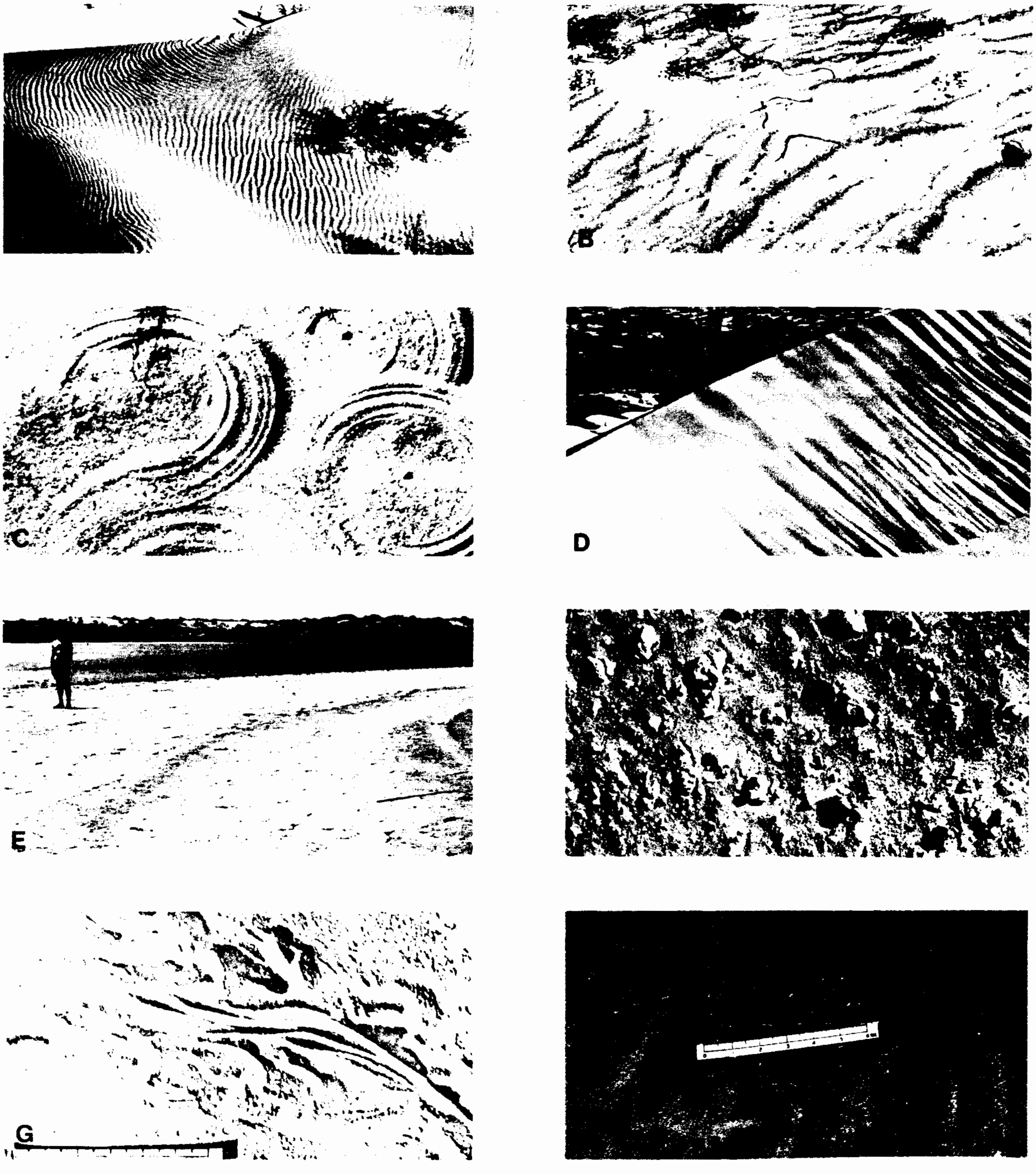


Figure 4
Surface structures on dunes. A) Fine sand ripples. B) Coarse sand ripples (compass is 5 cm across). C) Concentric grooves, 0.5 to 1.0 m in diameter created by plants, Cape Cod (courtesy of Rodney Stevens and Jour. Sed. Petrol.). D) Grainfall burying sand-flows on arm of star beform: slipface is 10 m high. E) Adhesion structures at margin of permanent saline Lake. F) Lag surfaces: pebbles are dominantly deflated caliche. Coin is 2 cm/diameter. G) Uphill track of chuckwalla lizard on dry foreset slope. Scale in inches. H) Pits of raindrops in modern sand. Scale in inches. (G and H, courtesy of E.D. McKee and United States Geological Survey) (A,B,D,E,F- from El Grand Desierto, Mexico).



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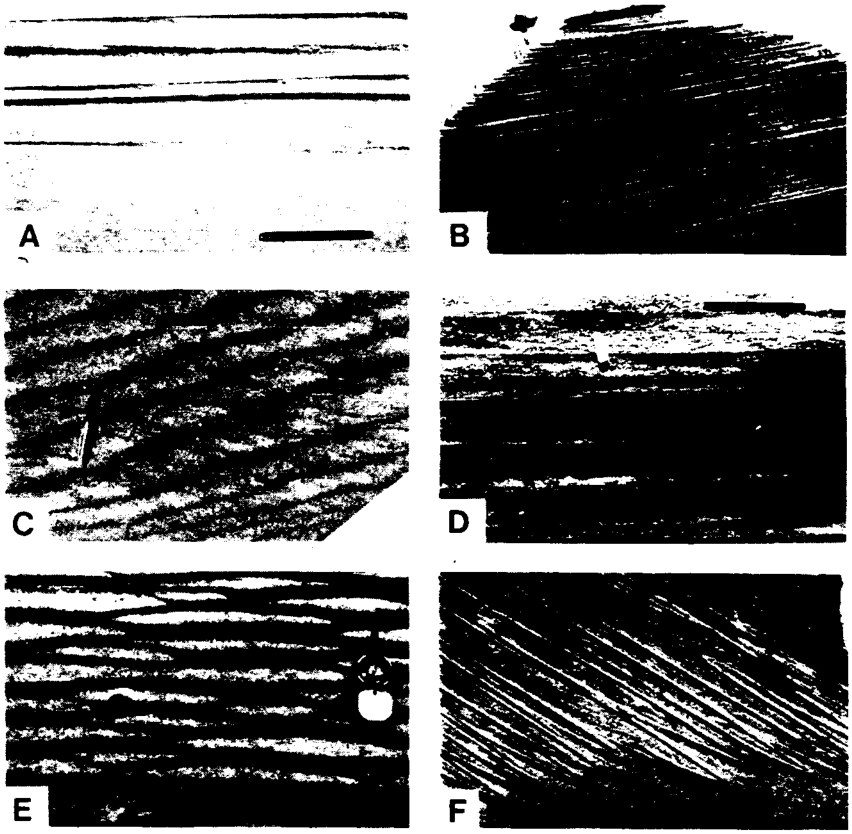


Figure 5
Stratification types.
A - Planebed lamination passing up into climbing ripple structure. Vertical section.
B - Typical subcritically climbing translatent stratification. Vertical section.
C - Subcritically to supercritically climbing translatent stratification: supercritical in centre of photo. Horizontal section.
D - Grainfall lamination, with interbedded sandflow at top. Vertical section.
E - Sandflow cross-strata: simple horizontal section.
F - Intertonguing of sandflow cross-strata and grainfall lamination near the base of a slipface. Vertical section.
Pen and bar scale all approx. 10 cm. All from Padre Island recent dunes (after Hunter, 1977, with permission), except D, which is from Permian of Arran (photo M.B., cf. Clemmensen and Abrahamson, 1983).

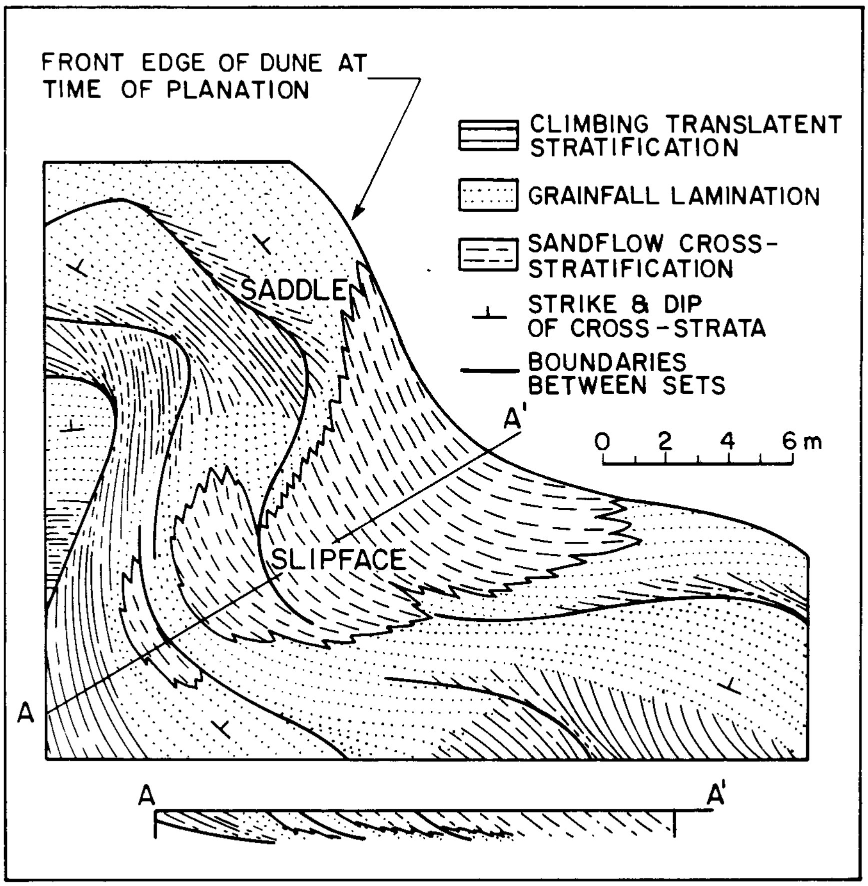


Figure 6
Map and cross-section of dune-foreset cross-strata exposed on a planed-off sinuous transverse or barchanoid dune, showing distribution of types of internal structure. Somewhat simplified from an exposure on Padre Island, Texas (from Hunter, 1977).

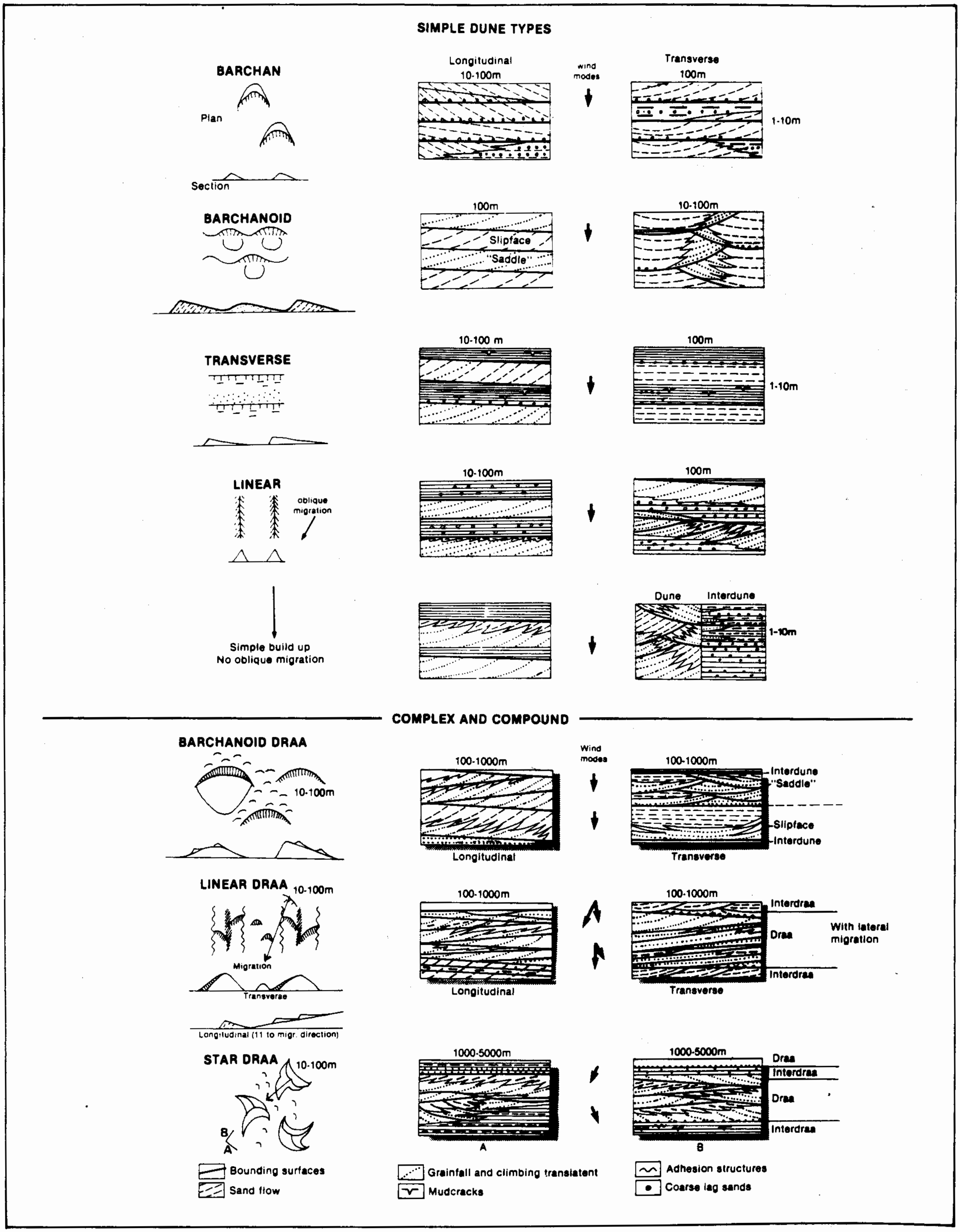


Figure 7
Stratification models for different dune types: both simple and complex/compound. Transverse sections are perpendicular to resultant wind direction: longitudinal sections are parallel to resultant wind directions.

Interdunes are an integral part of the eolian bedform system and must be taken into account when attempting to characterize, or determine the nature of the system (Ahlbrandt and Fryberger, 1981; Kocurek, 1981b). In deserts in which sand supply is limited, interdunes consist of lag deposits, coarse sand sheets, and small isolated dunes and sabhkas where the water table often reaches the surface (Stokes, 1964). Because of the way the sand is transported through the dune system, linear bedforms tend to have coarse lag and coarse sand sheets and dunes in the interdune areas; whereas transverse bedforms tend to have sabkhas and fine sand dunes between them (Glennie, 1970; Sharp, 1979). The size of the interdunes is also dependent on sand supply and on the stage of development of the erg in which they occur. Thus, most modern deserts have extensive interdunes because of their relatively recent development (Mainguet and Chemin, 1983).

Thus, as a basis for a facies model of ancient eolian sandstones, we have static features such as the textural parameters, distribution of stratification types on different parts and types of bedforms, and distributions of dunes and interdunes in modern deserts. We now need a dynamic dimension in order to see how thick, extensive ancient eolian sandstones may have formed.

STRATIFICATION MODELS BASED ON MIGRATING BEDFORMS

These are analogous to the Bouma sequence for turbidites, in that they attempt to describe the internal structures of the sand units. In this case, the internal structure is due to the migration of a packet of bedforms and not to an intermittent current.

Theoretical stratification models for transverse, longitudinal and star bedforms are shown in Figure 7. These assume *unidirectional climbing* of *ideal bedforms* in a fully developed draa-dune-interdune system. Migration of the dunes or dune-draa combination over interdunes leads to the development of bounding surfaces (Brookfield, 1977) (Figs. 8 and 9A).

First order surfaces are flat-lying bedding planes cutting across all other eolian structures and are attributed to the migration of draas.

Second order surfaces lie between first order surfaces, and usually dip downwind, though their inclination varies a great deal. These are attributed to dunes climbing down the lee slopes of draas, or to lateral migration of longitudinal dunes across the draa lee slope.

Third order surfaces bound bundles of laminae within cosets of cross laminae and are attributed to erosion followed by renewed deposition due to local fluctuations in wind direction and velocity. They are reactivation surfaces.

Simple dune systems should lack the second order surfaces: but in fact, dunes periodically overtaking dunes, and especially reversing dunes, are likely to show them (Fig. 9B). All dune-draa systems show a resultant sand drift, and unidirectional climbing is thus likely. But some star draa systems, due to their very slow rates of migration, could plausibly change their migration directions due to a change in wind regime over a long period of time. The models mostly show very low rates of bedform climbing, since most eolian sandstones seem to have preserved only a very small basal part of the original bedforms. Nevertheless, the models can be modified for supercritical climbing where necessary (Rubin and Hunter, 1982).

These stratification models are for uniform assemblages of one type of dune-interdune or dune-draa-interdune system. In deserts, many different systems are present. The next stage is to consider the possible stacking of different systems during the evolution of a desert. For this, we must consider the typical areal distribution of bedform types in a desert and note the vertical and lateral variations in stratification as they migrate and as the desert evolves.

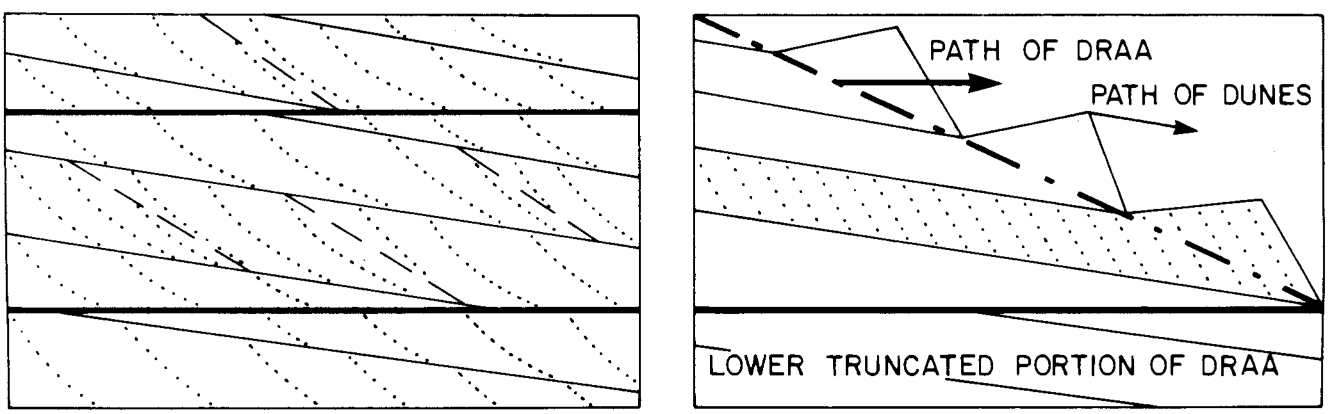
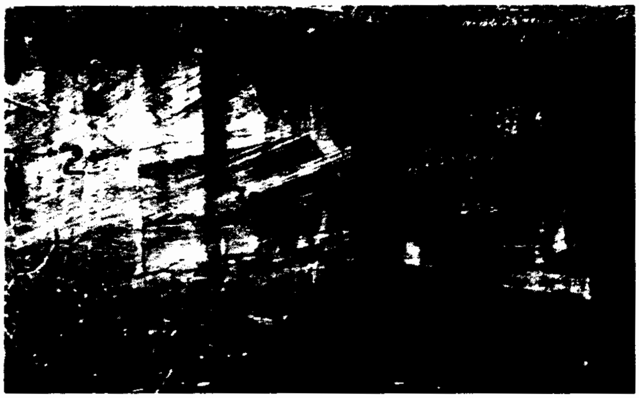
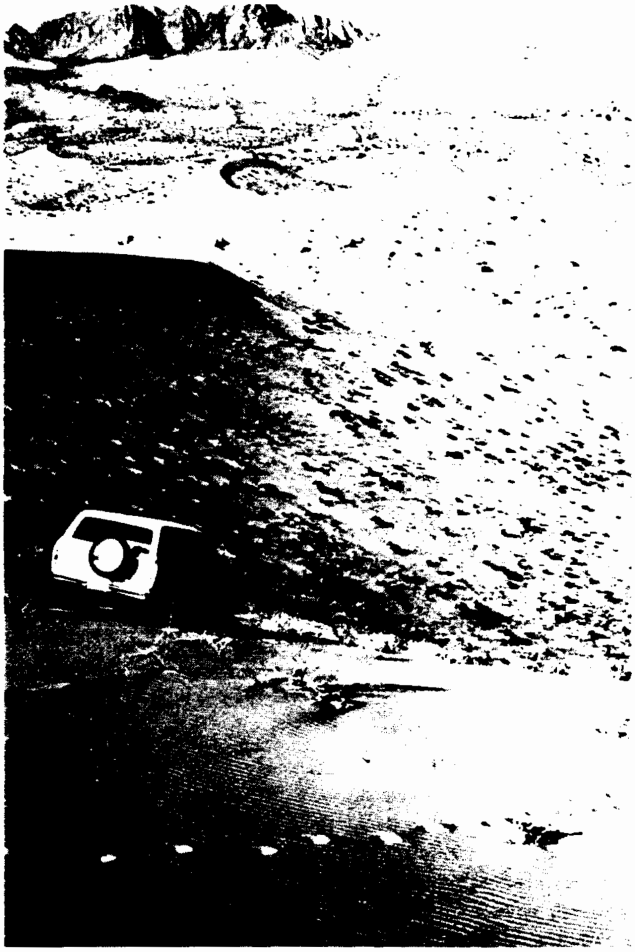


Figure 8
Relationship and origin or orders of bounding surface (after Brookfield, 1977). On left: synthetic section of ancient eolian sandstone parallel to bedform migration. Heavy lines - first order bounding surfaces; thin lines - second order bounding surfaces: dashed lines - third order bounding surfaces; dotted lines - eolian lamination. On right: origin of bounding surfaces by dunes climbing down draa lee slope. Bulk of draa is truncated by migration of succeeding draa.





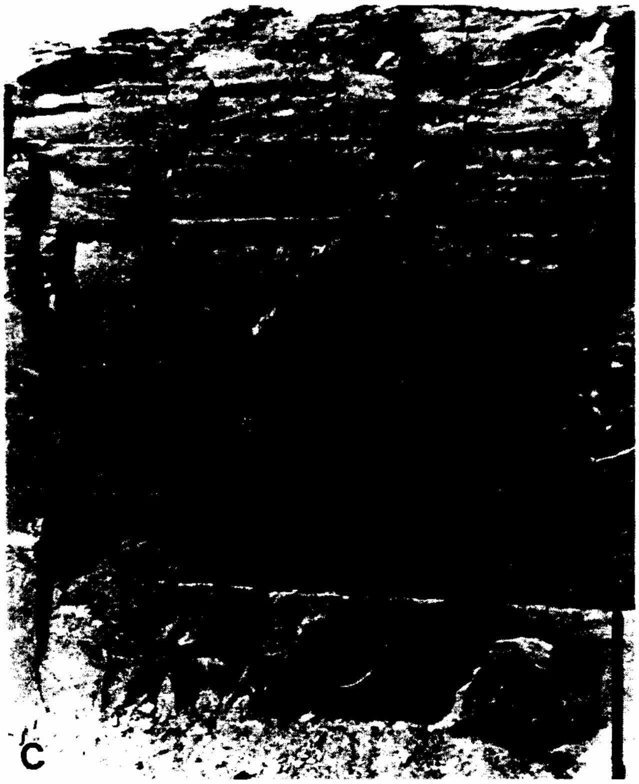


Figure 9

A) bounding surfaces in Permian eolian sandstone, Mauchline, Scotland. Scale - 1m.

B) reversing barchanoid dunes between rows of large star draa. One large draa being traversed by truck, others in distance. Note deflation surfaces on present lee slopes of both draa and dunes. These would form third and possibly second order surfaces in preserved section.

C) bounding surfaces (mostly first order), Navajo Sandstone (Jurassic), Page, Arizona. Note decrease in thickness upwards towards contact with overlying fluviatile Carmel Formation, suggesting decrease in size of bedforms with time. Scale bar, lower right, is 20m.

FACIES MODELS FOR EOLIAN SANDS

Two different models are shown on Figures 10 and 11, based on deserts with predominantly transverse and predominantly longitudinal bedforms respectively. These may be considered simplifications of the eastern part of El Gran Desierto (Fig. 1b) and of the Fachi-Bilma erg (Fig. 1a).

The transverse bedform model (Fig. 10) is based on unidirectional winds and upward and lateral accretion of sand within an enclosed basin with marginal fans. The fans allow us to keep the first order surfaces horizontal: in a wide extensive plain, these bounding surfaces would be convex upward on a large scale and resemble successive shells. The model involves the initial development of sand patches and barchan dunes with the onset of aridity, followed by the development of, successively, transverse dunes and compound transverse draa at the climax of aridity. With decreasing aridity, reduction in sand supply or both, there would be gradual contraction of the erg and eventual covering over by fluvial or lacustrine sediments.

The longitudinal bedform model follows the same pattern, except that in this case the early sand patches and barchans are followed by linear dunes and linear complex draas and eventually star draas. These longitudinal patterns have rarely been observed in ancient eolian sandstones, probably because such bedforms are characteristic of deserts with net through-flow of sand with little net deposition. The bedforms, as in the Fachi-Bilma erg, rest on

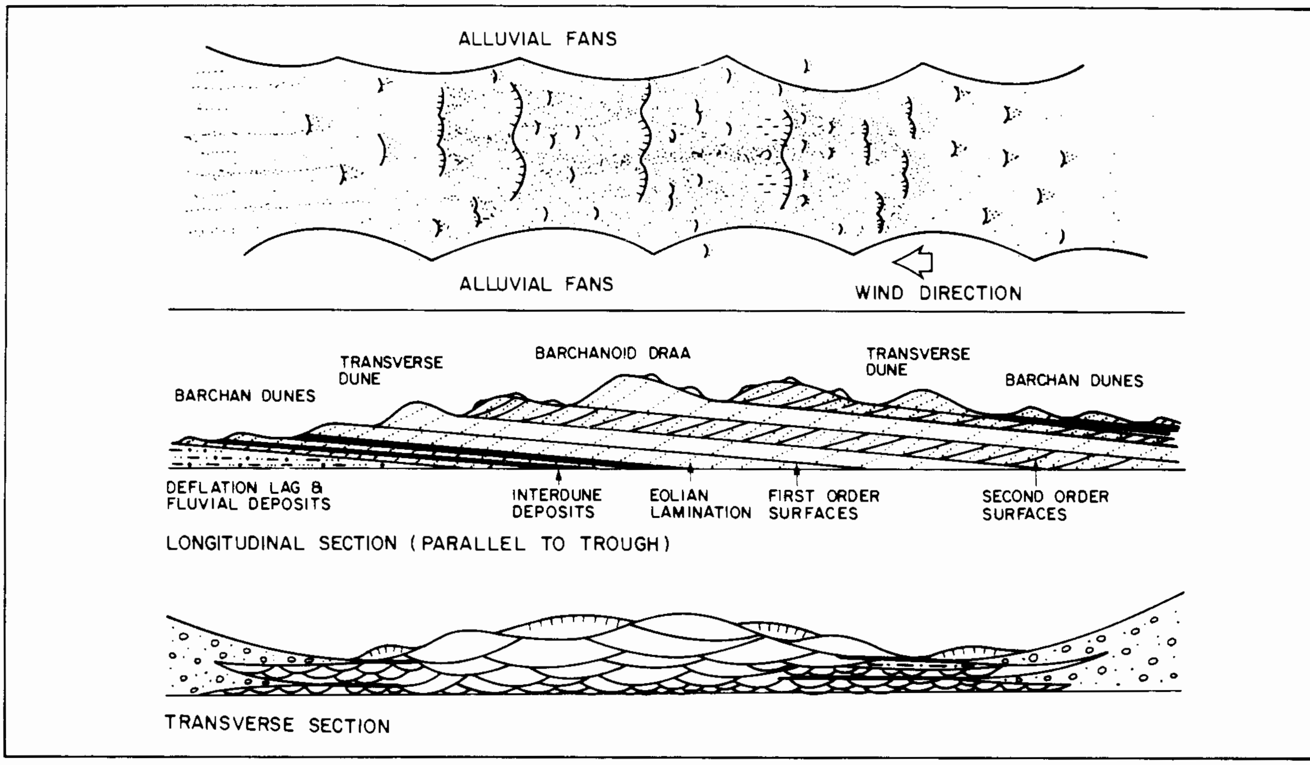


Figure 10
Synthetic model for desert with transverse bedforms. Note that bedforms are shown climbing far too steeply in the longitudinal cross-section, and that the troughs of the transverse section are too concave. Only first order surfaces are shown on the transverse section: some second order surfaces are shown on the longitudinal section. For details of stratification see Figure 7. Scale: basin is at least 5 km across.

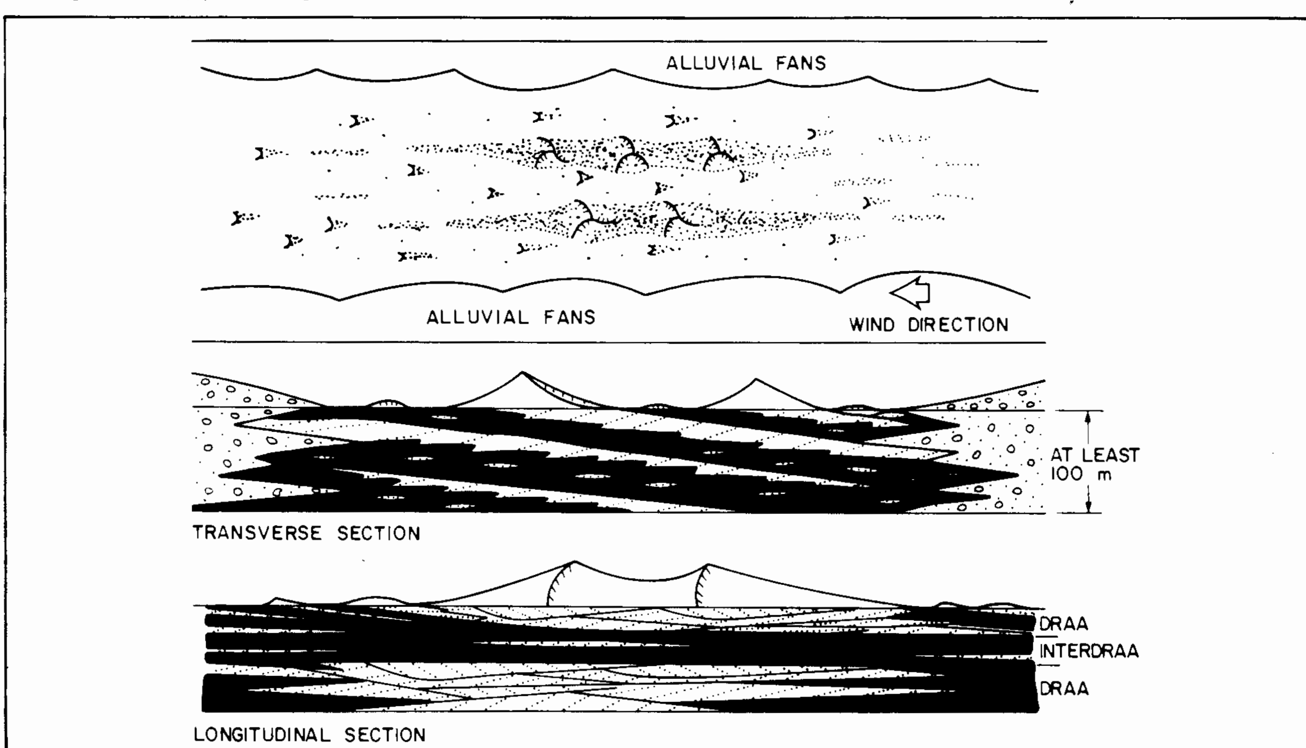


Figure 11
Synthetic model for desert with longitudinal bedforms in an enclosed basin. Note that some lateral migration of large longitudinal-star draas is permitted: without migration vertical stacks of longitudinal-star draa deposits would be separated by thick interdraa lag, barchan dune and fluviatile deposits. Black pattern includes all interdraa lag, small barchan dune, lacustrine and fluviatile sediment. For details of stratification see Figure 7. Scale: basin is at least 5 km across.

alluvium and lag deposits. Change in climatic conditions over a long period of time would probably lead to almost complete deflation, with perhaps isolated sand lenses left as the only remnant of the large star draas.

Note that the internal structures of the individual sets can not be shown at this scale: they are shown in Figure 7.

Similar models have been used by Kocurek (1981a) and Ross (1983) in their studies respectively of the Jurassic Entrada erg and the Precambrian Bigbear erg. Kocurek (1981a) used stratification types, bounding surfaces, foreset dip dispersion and interdune characteristics to reconstruct the central Entrada erg as shown in Figure 12. He estimated the wavelengths of the draas as between 900 and 1000 m from the intersection of first order bounding surfaces with supposed synchronous horizons. These wavelengths for the type of draa represented indicate draa heights approaching 100 m. At the edges of the Entrada erg, simple dune systems are indicated by the presence of only two orders of bounding surface.

Both models reflect growth and decay during slow environmental changes. In actual deserts, relatively sudden environmental change may cause major breaks in evolution. Thus, in the southern Sahara (Sahel), climatic change since the last ice age has led to vegetation overgrowing the dunes and preserving them intact in the full flower of their development (Talbot, 1980). Glennie and Buller (1983) studied the Permian Yellow Sands of northeast England which preserve much of a linear (seif) dune or draa system (see also Steele, 1983). This unusual preservation of a linear bedform assemblage was due to the extremely rapid inundation by the Zechstein Sea; so rapid in fact that the rounded profiles of individual dunes up to 50 m high are preserved below the overlying Marl Slate. Such feedback from ancient examples allows us to modify the models for specific cases. It would seem that such major breaks in the development of an erg – what might be called mega-bounding surfaces – should occur in the ancient; but I know of none yet described. They may be difficult to recognize except in very large continuous exposures.

Dynamic interpretations of ancient ergs are difficult, since we still do not know the exact cause for eolian bed-

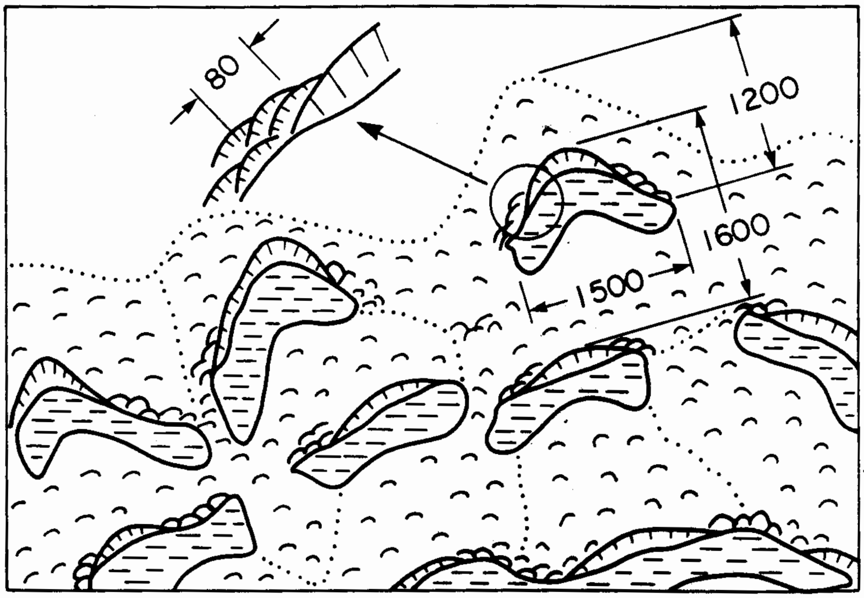


Figure 12
Interpretation of the overall structure and dimensions of bedforms and interdune areas in the central erg facies of the Entrada Sandstone (redrawn from Kocurek, 1981a). Draa slipfaces up to 110 m high: individual barchanoid dune slipfaces up to 8 m high. All measurements in metres.

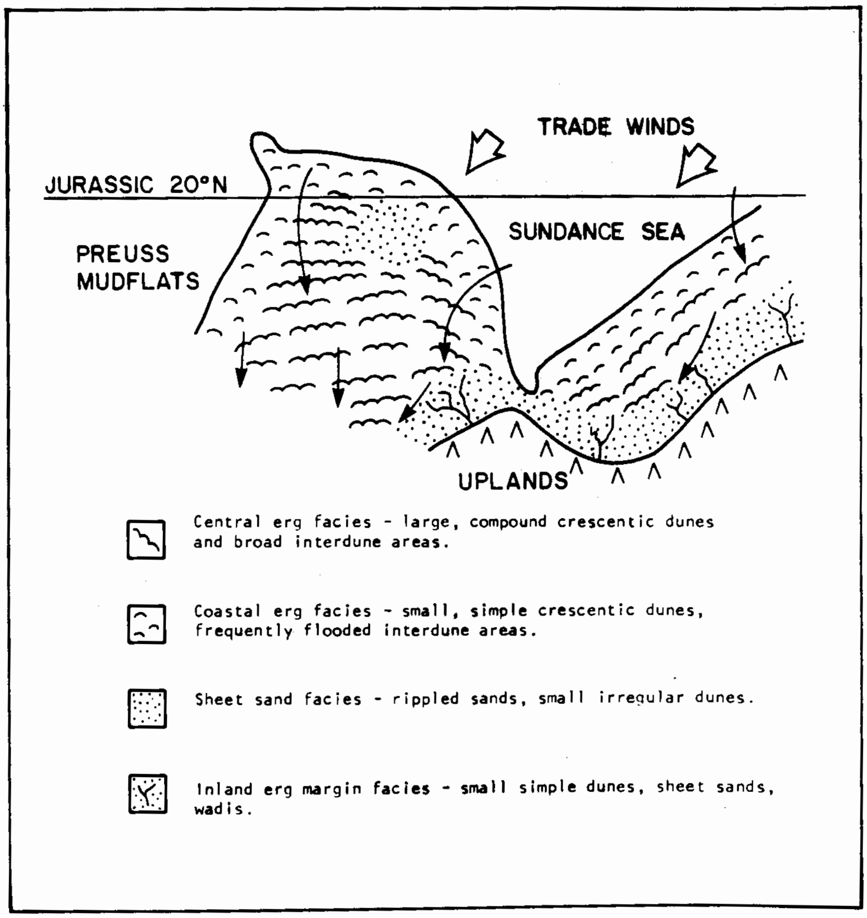


Figure 13
Reconstruction of the Entrada erg over the study area in N.E. Utah and N.W. Colorado. The region has been rotated to conform to the Jurassic paleolatitude position. Arrows indicate inferred erg circulation pattern (redrawn from Kocurek, 1981a).

form morphology and distribution in modern deserts. However, using the empirical results of bedform distribution from the Sahara, the concept of wind strength and directional variability and its effects on the nature of cross-stratification, some idea of wind regime for ancient ergs is possible. Glennie (1982) studied the paleowinds of the Permian Rotliegendes of the North Sea, and inferred that the Permian wind system was very simple. It matched that of the southern part of a modern northern hemisphere tradewind desert. A striking reversal of paleowinds around the Mid North Sea High was attributed to a 'wheel-around', similar to that seen in the Sahara and Arabia, and due to centres of high barometric pressure. A more detailed interpretation of the wind regime of the Jurassic Entrada erg is shown on Figure 13 (Kocurek, 1981 a; Kocurek and Dott, 1983).

THE EOLIAN MODEL AS A PREDICTOR

In an ideal desert, we would expect that thick compound superimposed cross-strata would characterize the interior. Towards the margins would be increasing numbers and thicker interdune deposits with smaller, simpler cross-strata, and possibly interbeds of fluvial deposits. In the two ideal sequences shown on Figure 14, we would expect "A" to have been deposited near the centre of the erg, and "B" closer to its margins. Depending on the paleowinds reconstructed from the cross-bedding, we could then try and predict in which directions the eolian sandstones would thicken and thin (Fig. 14C).

In terms of vertical sequence, I will show a possible interpretation based on a composite sequence in the Permian Thornhill and Dumfries basins of southwestern Scotland (Brookfield, 1977, 1980) shown on Figure 15. Note that the interpretation is based not only on the eolian sandstones, but also on the interdune and fluvial alluvial fan sequences with which they are interbedded.

These models do not take into account major changes, such as complete deflation of an erg due to climatic change, complete change in wind regime, or substantial lag in the response of large bedforms to changing conditions. I hope, nevertheless, that they at least illustrate that facies models for eolian sandstones are possible. Soon, eolian facies models will no doubt be as comprehensive as those for other sedimentary environments.

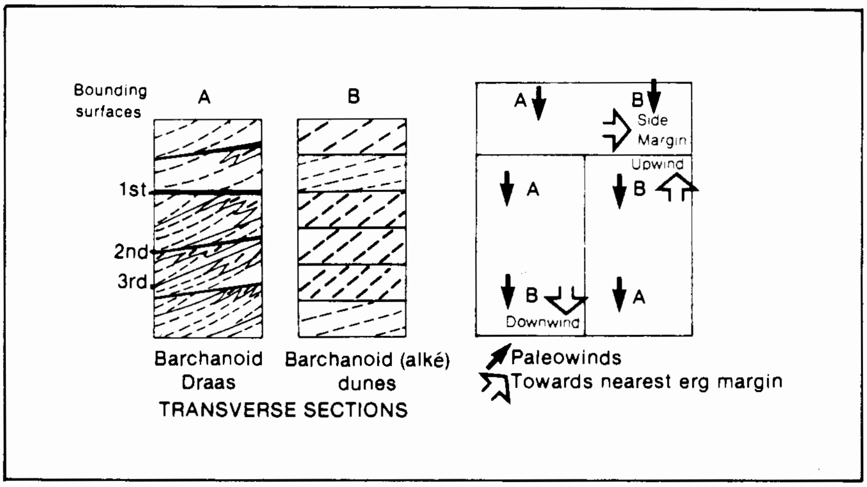


Figure 14
Two synthetic sections (A,B) (on left): inferred nearest erg margins based on locations of sections in relation to resultant wind direction (on right).

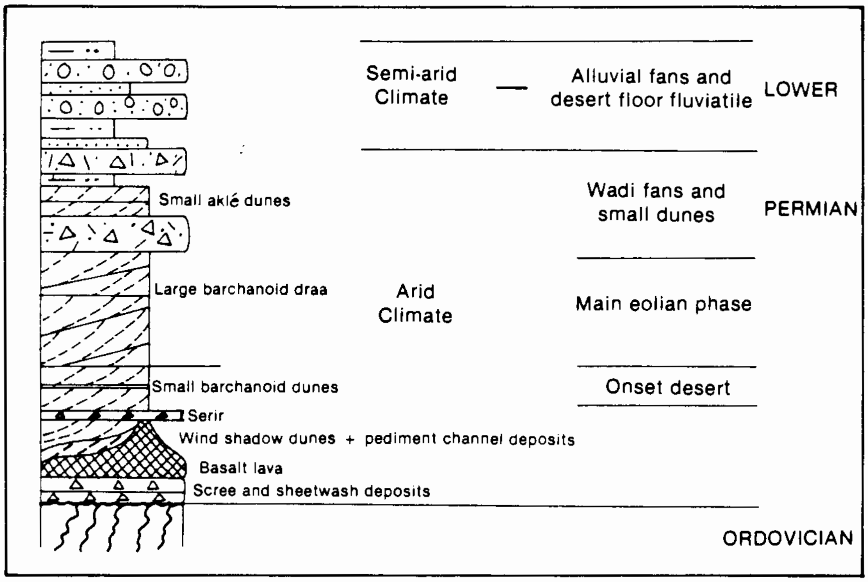


Figure 15
Section through Lower Permian deposits of the Thornhill and Dumfries intermontane basins, with interpretation of lithologies.

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REFERENCES

GENERAL AND VITAL

Bagnold, R.A., 1941. The physics of blown sand and desert dunes. London, Methuen, 265 p.
Still a marvellous, thought-provoking study of recent dunes and processes.

Bigarella, J.J., 1972. Eolian environments: their characteristics, recognition and importance. *In* Rigby, J.K., and Hamblin, W.K., eds., Recognition of ancient sedimentary environments. Society of Economic Paleontologists and Mineralogists, Special Publication 16, p. 12-62.
A bit dated but still useful.

Brookfield, M.E., and Ahlbrandt, T.S., 1983. Eolian sediments and processes. Amsterdam, Elsevier, 660 p.
A compilation of many diverse studies.

Cooke, R.V., and Warren, A., 1973. Geomorphology in deserts. London, Batsford, 394 p.

The best introduction to recent deserts.

Glennie, K.W., 1970. Desert sedimentary environments. Amsterdam, Elsevier, 222 p.

McKee, E.D., ed., 1979. A study of global sand seas. United States Geological Survey Professional Paper 1052, 429 p.
Lots of good stuff.

Walker, R.G. and Middleton, G.V., 1979. Facies Models 4. Eolian Sands. *In* Walker, R.G., ed., Facies models. Geoscience Canada Reprint Series 1, p. 33-41.

Walker, T.R. and Harms, J.C., 1972. Eolian origin of flagstone beds, Lyons Sandstone (Permian), type area, Boulder County, Colorado. Mountain Geologist, v. 9, p. 279-288.
A beautifully written paper, with good illustrations - start your eolian reading here.

Wislon, I.G., 1973. Ergs. Sedimentary Geology, v. 10, p. 77-106.
The origin of much of what is contained in this summary.

MODERN EOLIAN SANDS

Ahlbrandt, T.S., 1979. Textural parameters of eolian deposits. *In* McKee, E.D., ed., A study of global sand seas. United States Geological Survey Professional Paper 1052, p. 21-51.

Ahlbrandt, T.S. and Fryberger, S.G., 1981. Sedimentary features and significance of interdune deposits. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 293-314.

Andrews, S., 1981. Sedimentology of Great Sand Dunes, Colorado. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 279-291.

Bigarella, J.J., Becker, R.D. and Duarte, G.M., 1969. Coastal dune structures from Parana (Brazil). Marine Geology, v. 7, p. 5-55.

Breed, C.S., *et al.*, 1979. Regional studies of sand seas using Landsat (ERTS) imagery. United States Geological Survey Professional Paper 1052, p. 305-397.

Fryberger, S.G., 1979. Dune form and wind regime. *In* McKee, E.D., ed., A study of global sand seas. United States Geological Survey Professional Paper 1052, p. 137-169.

Fryberger, S.G. and Ahlbrandt, T.S., 1979. Mechanisms for the formation of eolian sand seas. Zeitschrift fur Geomorphologie N.F., v. 23, p. 440-460.

Fryberger, S.G., Ahlbrandt, T.S., and Andrews, S., 1979. Origin, sedimentary features and significance of low-angle eolian "sand-sheet" deposits, Great Sand Dunes National Monument and Vicinity, Colorado. Journal of Sedimentary Petrology, v. 49, p. 733-746.

Fryberger, G.S. and Schenk, C.J., 1981. Wind sedimentation tunnel experiments on the origin of eolian strata. Sedimentology, v. 28, p. 805-821.

Hanna, S.R., 1969. The formation of longitudinal sand dunes by large helical eddies in the atmosphere. Journal of Applied Meteorology, v. 8, p. 874-883.

Hunter, R.E., 1977. Basic types of stratification in small eolian dunes. Sedimentology, v. 24, p. 361-387.

Kocurek, G. and Fielder, G., 1982. Adhesion structures. Journal of Sedimentary Petrology, v. 52, p. 1229-1241.

Mainguet, M., 1976. Propositions pour une nouvelle classification des edifices sableux eoliens d'apres les images des satellites Landsat 1, Geminin, NOAA 3. Zeitschrifte fur Geomorphologie, v. 20, p. 275-296.

Mainguet, M. and Callot, Y. 1974. Air photo study of typology and interrelations between the texture and structure of dune patterns in the Fachi-Bilma Erg, Sahara. Zeitschrifte fur Geomorphologie, Suppl. Bd. 20, p. 62-69.

Mainguet, M. and Chemin, M.C., 1983. Sand Seas of the Sahara and Sahel - an explanation of their thickness and sand dune type by the sand budget principle. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., Eolian sediments and processes. Amsterdam, Elsevier, p. 353-364.

McKee, E.D., 1966. Structure of dunes at White Sands National Monument, New Mexico (and a comparison with structures of dunes from other selected areas). Sedimentology, v. 7, p. 1-70.

McKee, E.D., 1979. Introduction to a study of global sand seas. *In* McKee, E.D., ed., A study of global sand seas. United States Geogical Survey Professional Paper 1052, p. 1-19.

McKee, E.D., 1982. Sedimentary structures in dunes of the Namib desert, Southwest Africa. Geological Society of America, Special Paper 108, 64p.

McKee, E.D., and Douglas, J.R., 1971. Growth and movement of dunes at White Sands National Monument, New Mexico. United States Geological Survey Professional Paper 750-D, p. 108-114.

McKee, E.D. and Moiola, R.J., 1975. Geometry and growth of the White Sands dune field, New Mexico. Journal of Research, United States Geological Survey, v. 3, p. 59-66.

McKee, E.D. and Tibbits, G.C., Jr., 1964. Primary structures of a seif dune and associated deposits in Libya. Journal of Sedimentary Petrology, v. 34, p. 5-17.

Sharp, R.P., 1963. Wind ripples. Journal of Geology, v. 71, p. 617-636.

Sharp, R.P., 1979. Intradune flats of the Algodones Chain, Imperial Valley, California. Bulletin of the Geological Society of America, v. 90, p. 908-916.

Steidtmann, J.R., 1973. Ice and snow in eolian sand dunes of southwestern Wyoming. Science, v. 179, p. 796-798.

Talbot, M.R., 1980. Environmental repsonse to climatic change in the West African Sahel over the past 20,000 years. *In* Williams, M.A.J., and Faure, H., eds., The Sahara and the Nile; Quaternary Environments and prehistoric occupation in northern Africa. Rotterdam, Balkema, p. 37-62.

Tsoar, H., 1982. Internal structure and surface geometry of longitudinal (seif) dunes. Journal of Sedimentary Petrology, v. 52, p. 823-831.

Wasson, R.J. and Hyde, R., 1983a. Factors determining desert dune type. Nature, v. 304, p. 337-339.

Wasson, R.J. and Hyde, R., 1983b. A test of granulometric control on desert dune geometry. Earth Surface Processes and Landforms, v. 8, p. 301-312.

Wilson, I.G., 1971. Desert sandflow basins and a model for the origin of ergs. Geographical Journal, v. 137, p. 180-199.

Wilson, I.G., 1972. Aeolian bedforms - their development and origin. Sedimentology, v. 19, p. 173-210.

Wilson, I.G., 1973. Ergs. Sedimentary Geology, v. 10, p. 77-106.

ANCIENT EOLIAN SANDSTONES BEDFORM CLIMBING AND STRATIFICATION

Allen, J.R.L., 1963. Asymmetrical ripple marks and the origin of water-laid cosets of cross-strata. Liverpool Manchester Geological Journal, v. 3, p. 187-236.

Allen, J.R.L., 1982. Mud drapes in sand wave deposits: a physical model with application to the Folkstone Beds (Early Cretaceous, southeast England). Philosophical Transactions of the Royal Society of London, v. A306, p. 291-345.

Banks, N.L., 1973. The origin and significance of some downcurrent dipping cross-stratified sets. Journal of Sedimentary Petrology, v. 43, p. 423-427.

Brookfield, M.E., 1977. The origin of bounding surfaces in ancient eolian sandstones. Sedimentology, v. 24, p. 303-332.

Hunter, R.E., and Rubin, D.M., 1983. Interpreting cyclic cross-bedding with an example from the Navajo Sandstone. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., The best introduction to recent deserts.

Glennie, K.W., 1970. Desert sedimentary environments. Amsterdam, Elsevier, 222 p.

McKee, E.D., ed., 1979. A study of global sand seas. United States Geological Survey Professional Paper 1052, 429 p.
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Walker, R.G. and Middleton, G.V., 1979. Facies Models 4. Eolian Sands. *In* Walker, R.G., ed., Facies models. Geoscience Canada Reprint Series 1, p. 33-41.

Walker, T.R. and Harms, J.C., 1972. Eolian origin of flagstone beds, Lyons Sandstone (Permian), type area, Boulder County, Colorado. Mountain Geologist, v. 9, p. 279-288.
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Ahlbrandt, T.S., 1979. Textural parameters of eolian deposits. *In* McKee, E.D., ed., A study of global sand seas. United States Geological Survey Professional Paper 1052, p. 21-51.

Ahlbrandt, T.S. and Fryberger, S.G., 1981. Sedimentary features and significance of interdune deposits. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 293-314.

Andrews, S., 1981. Sedimentology of Great Sand Dunes, Colorado. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 279-291.

Bigarella, J.J., Becker, R.D. and Duarte, G.M., 1969. Coastal dune structures from Parana (Brazil). Marine Geology, v. 7, p. 5-55.

Breed, C.S., *et al.*, 1979. Regional studies of sand seas using Landsat (ERTS) imagery. United States Geological Survey Professional Paper 1052, p. 305-397.

Fryberger, S.G., 1979. Dune form and wind regime. *In* McKee, E.D., ed., A study of global sand seas. United States Geological Survey Professional Paper 1052, p. 137-169.

Fryberger, S.G. and Ahlbrandt, T.S., 1979. Mechanisms for the formation of eolian sand seas. Zeitschrift fur Geomorphologie N.F., v. 23, p. 440-460.

Fryberger, S.G., Ahlbrandt, T.S., and Andrews, S., 1979. Origin, sedimentary features and significance of low-angle eolian "sand-sheet" deposits, Great Sand Dunes National Monument and Vicinity, Colorado. Journal of Sedimentary Petrology, v. 49, p. 733-746.

Fryberger, G.S. and Schenk, C.J., 1981. Wind sedimentation tunnel experiments on the origin of eolian strata. Sedimentology, v. 28, p. 805-821.

Hanna, S.R., 1969. The formation of longitudinal sand dunes by large helical eddies in the atmosphere. Journal of Applied Meteorology, v. 8, p. 874-883.

Hunter, R.E., 1977. Basic types of stratification in small eolian dunes. Sedimentology, v. 24, p. 361-387.

Kocurek, G. and Fielder, G., 1982. Adhesion structures. Journal of Sedimentary Petrology, v. 52, p. 1229-1241.

Mainguet, M., 1976. Propositions pour une nouvelle classification des edifices sableux eoliens d'apres les images des satellites Landsat 1, Geminin, NOAA 3. Zeitschrifte fur Geomorphologie, v. 20, p. 275-296.

Mainguet, M. and Callot, Y. 1974. Air photo study of typology and interrelations between the texture and structure of dune patterns in the Fachi-Bilma Erg, Sahara. Zeitschrifte fur Geomorphologie, Suppl. Bd. 20, p. 62-69.

Mainguet, M. and Chemin, M.C., 1983. Sand Seas of the Sahara and Sahel - an explanation of their thickness and sand dune type by the sand budget principle. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., Eolian sediments and processes. Amsterdam, Elsevier, p. 353-364.

McKee, E.D., 1966. Structure of dunes at White Sands National Monument, New Mexico (and a comparison with structures of dunes from other selected areas). Sedimentology, v. 7, p. 1-70.

McKee, E.D., 1979. Introduction to a study of global sand seas. *In* McKee, E.D., ed., A study of global sand seas. United States Geogical Survey Professional Paper 1052, p. 1-19.

McKee, E.D., 1982. Sedimentary structures in dunes of the Namib desert, Southwest Africa. Geological Society of America, Special Paper 108, 64p.

McKee, E.D., and Douglas, J.R., 1971. Growth and movement of dunes at White Sands National Monument, New Mexico. United States Geological Survey Professional Paper 750-D, p. 108-114.

McKee, E.D. and Moiola, R.J., 1975. Geometry and growth of the White Sands dune field, New Mexico. Journal of Research, United States Geological Survey, v. 3, p. 59-66.

McKee, E.D. and Tibbits, G.C., Jr., 1964. Primary structures of a seif dune and associated deposits in Libya. Journal of Sedimentary Petrology, v. 34, p. 5-17.

Sharp, R.P., 1963. Wind ripples. Journal of Geology, v. 71, p. 617-636.

Sharp, R.P., 1979. Intradune flats of the Algodones Chain, Imperial Valley, California. Bulletin of the Geological Society of America, v. 90, p. 908-916.

Steidtmann, J.R., 1973. Ice and snow in eolian sand dunes of southwestern Wyoming. Science, v. 179, p. 796-798.

Talbot, M.R., 1980. Environmental repsonse to climatic change in the West African Sahel over the past 20,000 years. *In* Williams, M.A.J., and Faure, H., eds., The Sahara and the Nile; Quaternary Environments and prehistoric occupation in northern Africa. Rotterdam, Balkema, p. 37-62.

Tsoar, H., 1982. Internal structure and surface geometry of longitudinal (seif) dunes. Journal of Sedimentary Petrology, v. 52, p. 823-831.

Wasson, R.J. and Hyde, R., 1983a. Factors determining desert dune type. Nature, v. 304, p. 337-339.

Wasson, R.J. and Hyde, R., 1983b. A test of granulometric control on desert dune geometry. Earth Surface Processes and Landforms, v. 8, p. 301-312.

Wilson, I.G., 1971. Desert sandflow basins and a model for the origin of ergs. Geographical Journal, v. 137, p. 180-199.

Wilson, I.G., 1972. Aeolian bedforms - their development and origin. Sedimentology, v. 19, p. 173-210.

Wilson, I.G., 1973. Ergs. Sedimentary Geology, v. 10, p. 77-106.

ANCIENT EOLIAN SANDSTONES BEDFORM CLIMBING AND STRATIFICATION

Allen, J.R.L., 1963. Asymmetrical ripple marks and the origin of water-laid cosets of cross-strata. Liverpool Manchester Geological Journal, v. 3, p. 187-236.

Allen, J.R.L., 1982. Mud drapes in sand wave deposits: a physical model with application to the Folkstone Beds (Early Cretaceous, southeast England). Philosophical Transactions of the Royal Society of London. v. A306, p. 291-345.

Banks, N.L., 1973. The origin and significance of some downcurrent dipping cross-stratified sets. Journal of Sedimentary Petrology, v. 43, p. 423-427.

Brookfield, M.E., 1977. The origin of bounding surfaces in ancient eolian sandstones. Sedimentology, v. 24, p. 303-332.

Hunter, R.E., and Rubin, D.M., 1983. Interpreting cyclic cross-bedding with an example from the Navajo Sandstone. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., Eolian sediments and processes. Amsterdam, Elsevier, p. 429-454.

Kocurek, G., 1981b. Significance of interdune deposits and bounding surfaces in aeolian dune sands. Sedimentology, v. 28, p. 753-780.

Rubin, D.M. and Hunter, R.E., 1982. Bedform climbing in theory and nature. Sedimentology, v. 29, p. 121-138.

Rubin, D.M. and Hunter, R.E., 1983. Reconstructing bedform assemblages from compound cross-bedding. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., Eolian sediments and processes. Amsterdam, Elsevier, p. 407-427.

DESCRIPTIONS

Brookfield, M.E., 1980. Permian Intermontane basin sedimentation in Southern Scotland. Sedimentary Geology, v. 27, p. 167-194.

Clemmenson, L.B. and Abrahamsen, K., 1983. Aeolian stratification and facies association in desert sediemtns, Arran basin (Permian), Scotland. Sedimentology, v. 30, p. 311-339.

Glennie, K.W., 1982. Early Permian (Rotliegendes) paleowinds of the North Sea. Sedimentary Geology, v. 34, p. 245-265.

Glennie, K.W. and Buller, A.T., 1983. The Permain Weissliegend of N.W. Europe: the partial deformation of aeolian dune sands caused by the Zechstein transgression. Sedimentary Geology, v. 35, p. 43-81.

Hunter, R.E., 1981. Stratification styles in eolian sandstones; some Pennsylvanian to Jurassic examples from the western Interior, U.S.A. Society of Economic Paleontologists and Mineralogists, Special Publication 31, p. 315-329.

Kocurek, G., 1981a. Erg reconstruction: the Entrada Sandstone (Jurassic) of northern Utah and Colorado. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 36, p. 125-153.

Kocurek, G. and Dott, R.J., Jr., 1983. Jurassic paleogeography and paleoclimate of the central and southern Rocky Mountain Region. *In* Reynolds, M.W., and Dolly, E.D., eds., Mesozoic paleogeography of West-central United States, Rocky Mountain Section, Society of Economic Paleontologists and Mineralogists, Denver, Colorado. p. 101-116.

Ross, G.M., 1983. Proterozoic aeolian quartz arenites from the Hornby Bay Group, Northwest Territories, Canada: Implications for Precambrian aeolian processes. Precambrian Research, v. 20, p. 149-160.

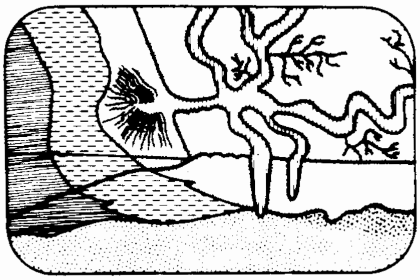
Sanderson, I.D., 1974. Sedimentary structures and their environmental significance in the Navajo Sandstone, San Rafael Swell, Utah. Brigham Young University Geological Studies, v. 21, p. 215-246.

Steidtmann, J.R., 1974. Evidence for eolian origin of cross-stratification in sandstone of the Casper Formation, southermost Laramie Basin, Wyoming. Geological Society of America, Bulletin, v. 85, p. 1835-1842.

Steele, R.P., 1983. Longitudinal draa in the Permian Yellow Sands of north-east England. *In* Brookfield, M.E., and Ahlbrandt, T.S., eds., Eolian sediments and processes. Amsterdam, Elsevier, p. 543-550.

Stewart, W.D., and Walker, R.G., 1980. Eolian coastal dune deposits and surrounding marine sandstone, Rocky Mountain Supergroup (Lower Pennsylvanian), southeastern British Columbia. Canadian Journal of Earth Sciences, v. 17, p. 1125-1140.

Thompson, D.B., 1969. Dome-shaped aeolian dunes in the Frodsham member of the so-called "Keuper" Sandstone Formation (Scythian-Anisian; Triassic) at Frodsham, Cheshire (England). Sedimentary Geology, v. 3, p. 263-289.



Deltas

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INTRODUCTION

Deltaic depositional models differ from most others in that their construction has not depended on a distillation of observations on ancient rocks but has arisen largely from a study of depositional processes on modern deltas. There are at least three distinct delta models or "norms" to consider in interpreting ancient rocks, but these are end members of a broad specturm of delta types, and many modern and ancient deltas combine features of all three.

DEFINITION

The concept of the delta is one of the oldest in geology, dating back to about 400 B.C. when Herodotus observed that the alluvial plain at the mouth of the Nile was similar in shape to the Greek letter △. The term has been used for similar geographic features ever since.

We now define a delta, geologically, as "a deposit, partly subaerial, built by a river into or against a permanent body of water" (Barrell, 1912). The result is an irregular progradation of the shoreline directly contolled by the river. The sediments are formed under subaerial and shallow marine or lacustrine environments and typically show a gradation into finer-grained offshore facies. A crucial part of the definition is that the influence of a river or rivers as the main sediment source should be recognized. In the ancient record this is best accomplished by mapping lithofacies distributions, which should show the presence of a significant thickening of the clastic succession close to presumed locations of riverine sediment input into the sedimentary basin. However, in many deltas the fluvial influence is masked strongly by waves, ocean currents, tidal currents or winds. Ancient deltaic deposits of this type may be hard to recognize, and it seems likely that many have been interpreted, in the past, in terms of these modifying processes as wave-formed beach complexes or tidal flat deposits.

A SHORT HISTORY OF DELTA STUDIES

Modern work in the English-speaking world commenced with the classic studies of Gilbert on the deltas in Lake Bonneville. Gilbert was the first to attempt a hydrodynamic explanation of delta formation, and his ideas dominated thinking on the subject for many years. A classic paper by Barrell (1912) on the ancient Catskill delta also had a far-ranging influence.

Since the 1920s interest in deltas has been stimulated by the fact that the sediments of many ancient deltas contain extremely large deposits of coal, oil and gas. Nowhere is this more true than in the hydrocarbon-rich Gulf Coast of Texas and Louisiana, and research into deltaic sedimentation during the last forty years has been overwhelmingly dominated by studies of Holocene Gulf Coast deltas and their Quaternary and Tertiary antecedents. Most attention became focused on the Mississippi, which rapidly replaced the Lake Bonneville deltas of Gilbert as the standard model delta in geology textbooks.

Sedimentological research into the Mississippi commenced with the monumental work of Fisk, who established the depositional framework of the modern delta with the aid of many thousands of shallow boreholes. Subsequently the American Petroleum Institute funded a major research effort (Project 51), the objective of which was the study of modern sediments along the northwest margin of the Gulf of Mexico. The publication which summarizes this work (Shepard *et al.*, 1960) contains landmark papers on depositional processes in the Mississippi by Shepard and by Scruton. Further publications on the depositional environments and cyclic sedimentation in the Mississippi were provided by Kolb and Van Lopik (1966), by Frazier (1967) and by Coleman and Gagliano (1964, 1965).

The other deltas that were studied extensively at this time were those of the Niger(Allen, 1970; Oomkens, 1974), the Orinoco (Van Andel, 1967) and the Rhône (Oomkens, 1970).

Useful compilations of papers on ancient and modern deltas include those of Morgan (1970), Broussard (1975) and Le Blanc (1976a, 1976b). The basis of the modern three-fold classification of deltas (Fig. 1) was established

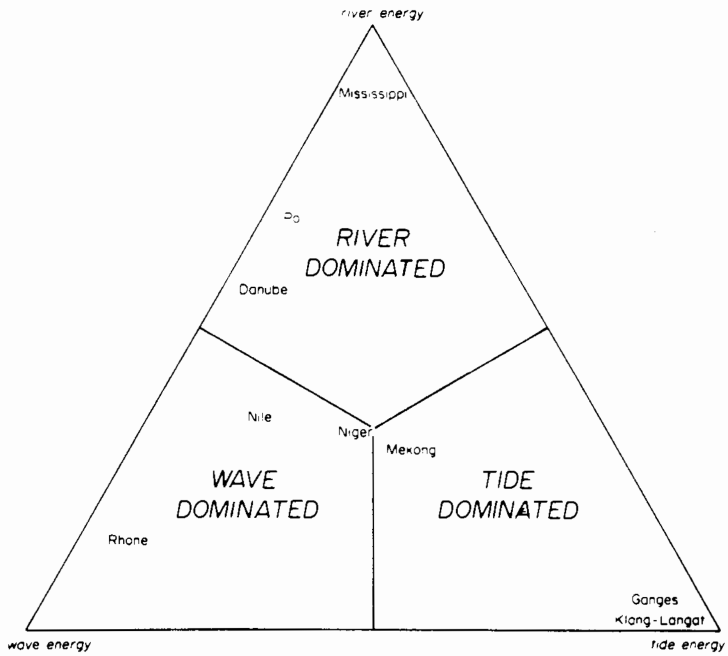

Figure 1
A classification of deltas based on variations in transportation patterns on the delta (after Galloway, 1975).

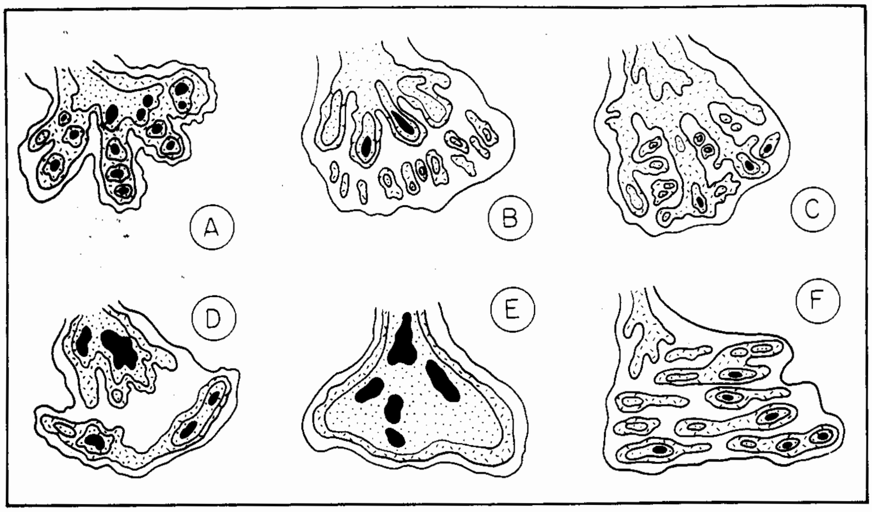

Figure 2
Delta models of Coleman and Wright (1975). A) River-dominated with low wave and tide energy, low littoral drift; B) River-dominated with low wave energy, high tide range, low littoral drift; C) Intermediate wave energy, high tide, low littoral drift; D) Intermediate wave energy, low tide range; E) High wave energy, low littoral drift; F) high wave energy, strong littoral drift.

by Fisher *et al.*, (1969; see also Galloway, 1975), who proposed a subdivision into river-, wave- and tide-dominated types (these are the three end members or "norms" referred to above). Wright *et al.*, (1974) elaborated this classification, showing that various combinations of the three main processes could form six principal delta types (Fig. 2). Useful summaries of this work are provided by Coleman (1981) and Coleman and Wright (1975), and it is discussed later in this paper. An excellent general summary of deltaic sedimentation is given by Elliot (1978).

The only major development in delta studies during the last ten years has been the increasing recognition of the importance of syndepositional deformation on delta front surfaces, particularly in river-dominated deltas. Slumps, slides and growth faults are pervasive in many modern and ancient deltas, and have a major effect on subsurface stratigraphy and lithofacies distributions (Coleman *et al.*, 1983; Winker and Edwards, 1983).

Most of the major developments in the understanding of deltas are attributed to Gulf Coast geologists, particularly the staff of the Coastal Studies Institute at Louisiana State University and the Bureau of Economic Geology at the University of Texas. The pre-eminence of this group is remarkable, and is mainly a reflection of the profound importance of modern and ancient Gulf Coast deltas to the economy of that region (petroleum, coal and uranium production, environmental geology). However, it has tended to bias geologists everywhere towards interpretations based on Gulf Coast models, particularly that of the Mississippi delta, although these are not everywhere appropriate and, to some extent, may even be unique.

Delta facies models seem now to have reached a mature phase of development, in contrast to those for other environments, particularly models of continental margin sedimentation and shelf sedimentation (Walker, "Shelf and Shallow Marine Sands", this volume), which are still undergoing rapid evolution. However, considerable work is needed to test the models by careful documentation of the ancient record. This is especially necessary for wave- and tide-influenced deltas, of which few well-described ancient examples exist.

DELTA FORMATION AND CLASSIFICATION

The distribution, orientation and internal geometry of deltaic deposits is controlled by a variety of factors, including climate, water discharge, sediment load, river-mouth processes, waves, tides, currents, winds, shelf width and slope, and the tectonics and geometry of the receiving basin (Wright *et al.*, 1974). In a brief paper such as this it is impossible to describe fully the inter-relationships between all these variables, but several generalizations are possible, such as those on which the principal classification of deltas are based (Figs. 1 and 2; discussed below).

Variations in Sediment Input

Climate, water discharge (rate and variability) and sediment load (quantity and grain-size) are to some extent interrelated. In humid, tropical regions precipitation normally is high relative to evapotranspiration; runoff tends to be high and steady. The predominance of chemical over mechanical weathering leads to high dissolved-load sediment yields. These factors give rise to relatively stable, meandering channel patterns.

In arctic or arid conditions precipitation is erratic, vegetation is sparse, and braided channel patterns with large bedloads tend to occur (Coleman, 1981, and Coleman and Wright, 1975 provide a more complete discussion of this topic).

These distinctions are most easily recognized in the fluvial delta plain deposits by the geometry and grain size of the distributary channel fill units (see "Coarse Alluvial Deposits", and "Sandy Fluvial Systems", this volume). However, where the delta is not significantly modified by processes there will be differences in the structure of the delta as a whole, as discussed below.

Variations in River-Mouth Flow Behaviour

When a sediment-laden river enters a body of standing water one of three types of flow dispersal may occur, depending on the density differences between the river water and that of the lake or sea into which it flows. Variations in temperature, salinity and sediment load can cause such differences in density.

A) Inflow More Dense. This is a common occurrence where sediment-laden streams enter fresh-water lakes (e.g., glacier-fed streams in Alpine regions). A narrow, arcuate zone of active deltaic progradation containing the coarse bedload may occur along the shore.
The delta which forms contains the distinct, steeply-dipping forests of the classical Gilbertian delta. The finer sediment fraction may be dispersed offshore as density interflows or underflows, forming repeated graded units.

B) Inflow Equally Dense. This is also a common occurrence in fresh-water deltas, and may also develop at the mouths of rivers entering brackish back-barrier lagoons. Sediment is dispersed radially and competency is lost rapidly. The bulk of the sediment is deposited on a Gilbertian delta.

C) Inflow Less Dense. Most marine deltas are formed under these conditions because freshwater is less dense than seawater, unless it is unusually cold or sediment laden. Lacustrine deltas formed at the mouths of suspended-load rivers are also of this type. The river effluent tends to form a discrete plume floating on the surface of the sea. The suspended sediment load is widely dispersed, resulting in a large active delta-front area, typically dipping at 1° or less, and contrasting with the 10 to 20° dip of typical Gilbertian deltas.

Marine waters beneath the effluent jet form a "salt-wedge", which may extend for tens of kilometres upstream, particularly during high tide. Marine faunas can thus be found well inland - a possible source of confusion in the study of ancient deltaic deposits.

These patterns can be radically modified by tide or current activity, as described below.

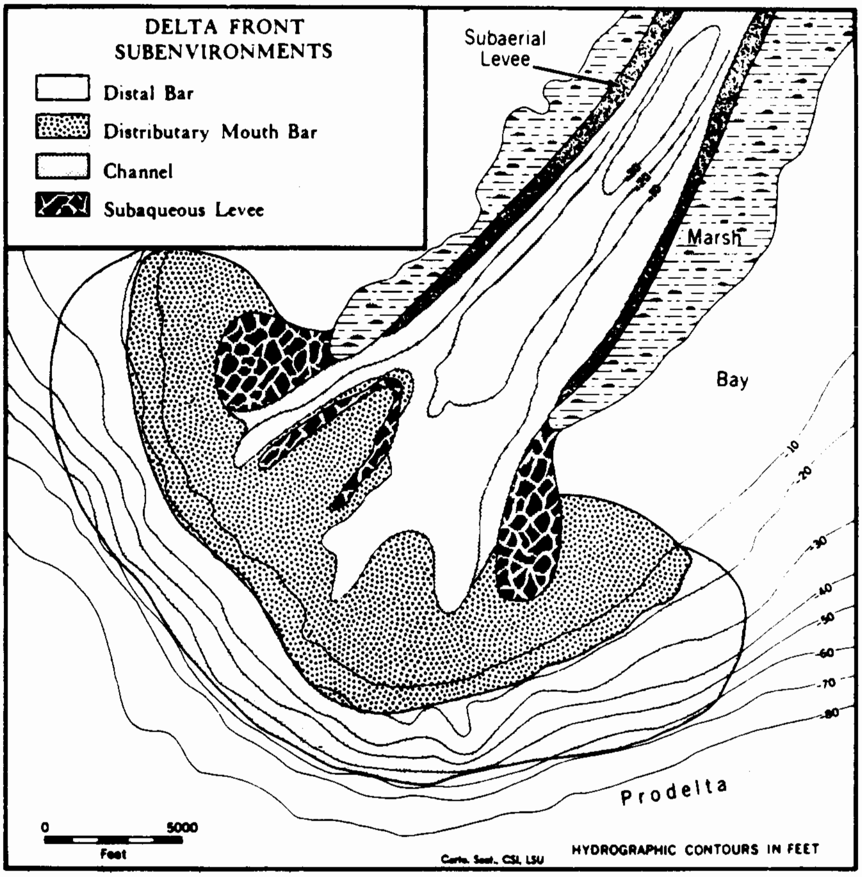

Figure 3
Subenvironments at a distributary mouth in a river-dominated, birdsfoot-type delta, South Pass, Mississippi delta. Progradation seaward leads to the development of elongate sand bodies called "bar-finger" or "shoestring" sands (Coleman and Gagliano, 1965).

Variations in Transport Pattern on the Delta

The type of energy conditions that exist in the sea at the river mouth are of fundamental importance in controlling depositional environments and the geometry of the resulting sediments. In fact the most useful classification of delta types is one based on the relative strengths of fluvial and marine processes (Fig. 1), as shown by Fisher *et al.* (1969), Coleman (1981), Galloway (1975) and Coleman and Wright (1975). Interrelationships between these processes form the main basis for recognizing three deltaic "norms".

A) River-Dominated Deltas. If waves, tidal currents and longshore currents are weak, rapid seaward progradation takes place, and a variety of characteristic, fluvially dominated depositional environments develops. At the mouth of each distributary subaqueous levees may form where the competence of the effluent jet is reduced by friction with the static sea water at the margins of the flow (Fig. 3). The main sediment load is deposited in a distributary mouth bar, which becomes finer grained toward the sea. The proximal mouth bar region is characterized by scour channels and by temporary bars and islands with abundant crossbedding, resulting from variations in flow conditions (changes in river discharge, tidal effects).

In the case of mixed- or suspended-load distributary channels, which are relatively stable in position, and in the absence of significant wave- or tide-induced scour, sedimentation gradually extends the mouths seaward. The resulting lithofacies assemblage, of which the mouth bar sand is the most important, tends to be oriented at a high angle to the coastline (basin margin), as in Figure 2A, a fact that can be of considerable importance in the understanding of an ancient deltaic deposit. "Bar-finger" or "shoestring" sands are a typical component of such a deltaic assemblage. The modern Mississippi delta is the best modern example of this pattern, showing the distinctive "birdsfoot" shape in plan view (Figs. 3 and 4).

Between the distributaries are interdistributary bays, which commonly are areas of low energy, muddy sedimentation and abundant organic activity. Shell beds and bioturbation are common. These bays eventually fill with sediment and become marshes. One of the most important ways in which this occurs is by the development of crevasse splays, which occurs in the following manner.

As progradation proceeds the river slope is flattened and flow becomes less competent. At this stage a breach in the subaerial levee may occur upstream during a period of high discharge. Such a breach is termed a crevasse. The shorter route it offers to the sea via an interdistributary bay generally is the cause of a major flow diversion, and a subdelta (crevasse-splay) deposit may develop rapidly. Eventually the crevasse may become a major distributary and the process is repeated.

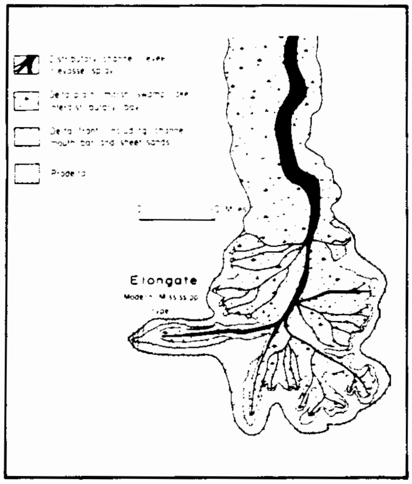

Figure 4
A birdsfoot-type river-dominated delta; the modern Mississippi delta (Fisher et al., *1969).*

Where delta distributaries are of the unstable, low sinuosity (braided) type, with shifting courses and numerous bars and islands, a different type of delta may develop. The outline tends to be lobate, and mouth bars merge laterally into a sheet sand. Crevasse splays may be absent, but sediment is distributed throughout the delta by distributary switching (avulsion), a process analogous to that of crevassing. The radiating pattern of distributaries is similar to that of alluvial fans, and the term fan-delta is commonly used to describe them (Fig. 5). Pebbly sands and gravels are common to dominant components of the delta plain and delta front environment. Good descriptions of modern fan delta sedimentation have been given by McGowen (1970), Galloway (1976), and Wescott and Ethridge (1980).

At present, fan deltas tend to occur in arctic or arid environments, where the abundance of coarse bedload and the variable river discharge favour unstable braided distributary networks. Fan deltas were probably the dominant type of river-dominated delta in pre-Devonian time because, until the advent of land vegetation, which tends to store rainfall and regulate runoff, braided channel networks were probably the main fluvial style.

B) Wave-Dominated Deltas. On most coastlines waves rework shoreline sediments and account for local distinctive facies. However, the contrasts between the minor wave activity in areas such as

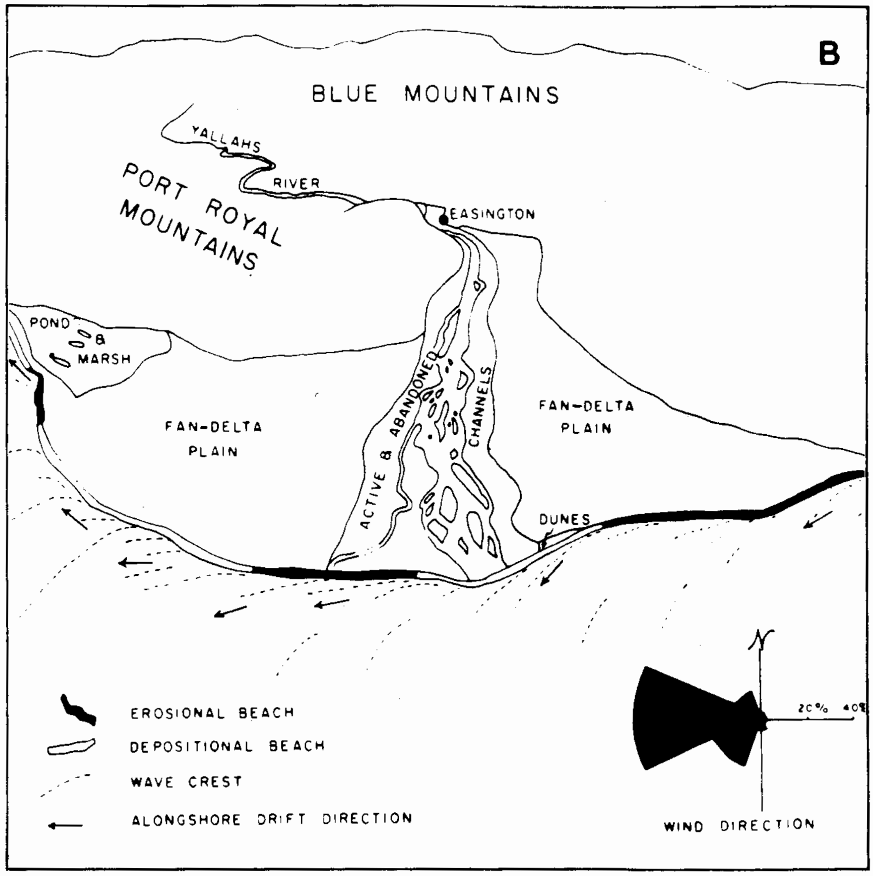

Figure 5
Interpretive sketch map of the modern Yallahs fan delta, Jamaica, drawn from an oblique air photo (Wescott and Ethridge, 1980).

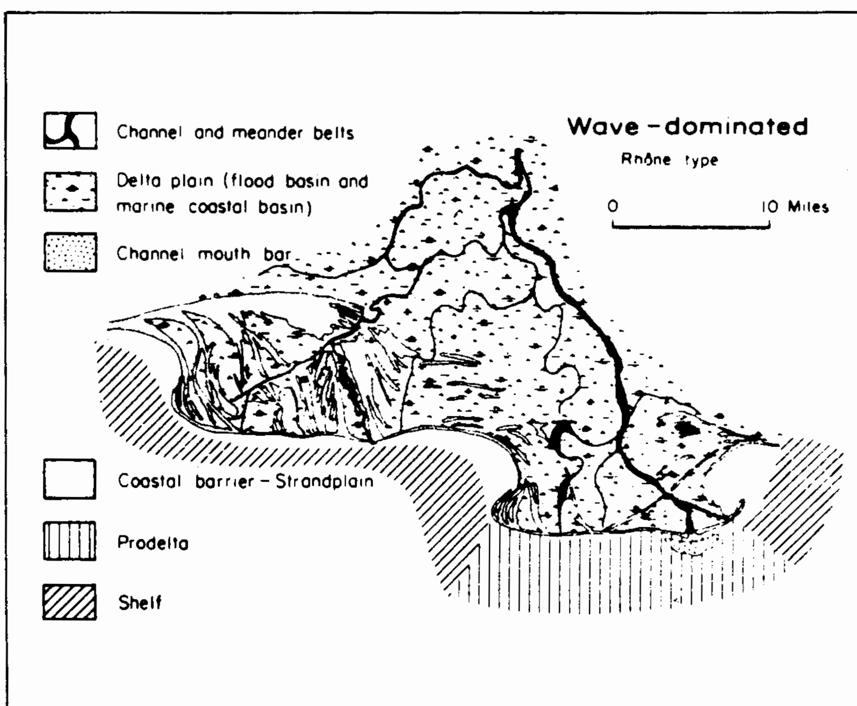

Figure 6
A wave-dominated delta; the modern Rhône delta (Fisher et al., *1969).*

the Gulf Coast, and the wave-dominated coastlines of much of the Atlantic and Pacific oceans, are dramatic. Coleman and Wright (1975) suggested that it takes a whole year of wave activity on the Mississippi delta to equal ten hours of wave energy expenditure on the São Francisco delta, Brazil.

In a wave-influenced delta (e.g., Figs. 2C, 2D, 6), mouth bar deposits are continually reworked into a series of curved beach ridges. If the winds are predominantly onshore, they may redistribute much of the beach sand as an eolian dune field capping the delta plain.

The geometry of the delta front beach complex depends largely on the nature of shoreline circulation patterns. An oblique angle of wave attack may develop a powerful longshore drift current, in which case the entire delta may become asymmetrical and skewed downcurrent (Fig. 2F). Beaches grow laterally and fill interdistributary bays by the develoment of curved spits, as in the modern Rhône delta (Fig. 6). Whether or not this longshore drift occurs, individual sand bodies tend to be oriented more or less parallel to the coastline in marked distinction to that of other delta types. The facies characteristics and mature petrography of these shoreline sand bodies are distinctive, as discussed elsewhere in this volume.

C) Tide-Dominated Deltas. Where the tidal range is high the reversing flow that occurs in the distributary channels during flood and ebb may become the principal source of sediment dispersal energy. Within and seaward of the distributary mouths the sediment may be reworked into a series of parallel, linear or digitate ridges parallel to the direction of tidal currents and separated from each other by linear scour channels (Fig. 7). The ridge-and-channel morphology, with a trend perpendicular to shoreline, is one of the most characteristic features of the tide-dominated delta, and may be readily detected by careful lithofacies mapping (Figs. 2B and 2C). The subaerial part of the delta consists largely of tidal flats comprising mainly fine-grained deposits. Distributaries may contain well sorted sands deposited under conditions of reversing flow, and large quantities of clay and silt will tend to be flushed into the delta marsh by overbank flooding during high tides.

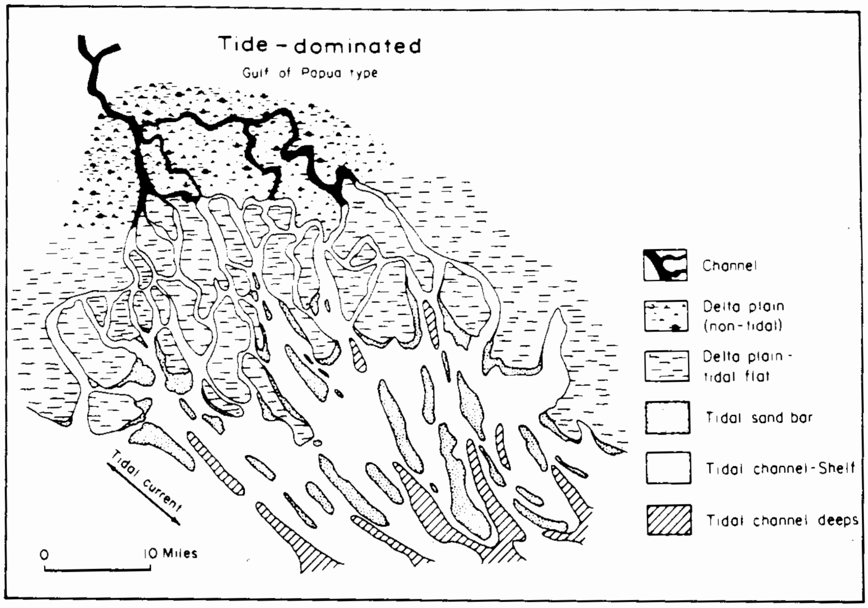

Figure 7
A tide-dominated delta; the modern Ganges-Brahmaputra delta (Fisher et al., *1969).*

As in the case of wave-dominated deltas, tidal currents may completely rework the deposits and redistribute them away from the river mouth. In such a case it may be difficult to recognize the deposits as deltaic. Many ancient beach or shallow marine deposits, with evidence of wave or tidal reworking, may have been misidentified as a result. The large volume of the deposit, or the presence of a landward facies change into a thick fluvial sequence, may be the only clues to a deltaic interpretation.

Syndepositional Deformation

Rapid sedimentation on deltas leads to gravitational instabilities and the generation of a variety of small to medium scale structures as a result of loading or slope failure (Fig. 8; see Coleman *et al.*, 1983). Such structures are likely to be more common on river-dominated deltas, where the rate of seaward growth tends to be more rapid. The most important of these structures are growth faults, formed by sediment loading and episodic failure on the seaward side of the fault plane. Sedimentary units

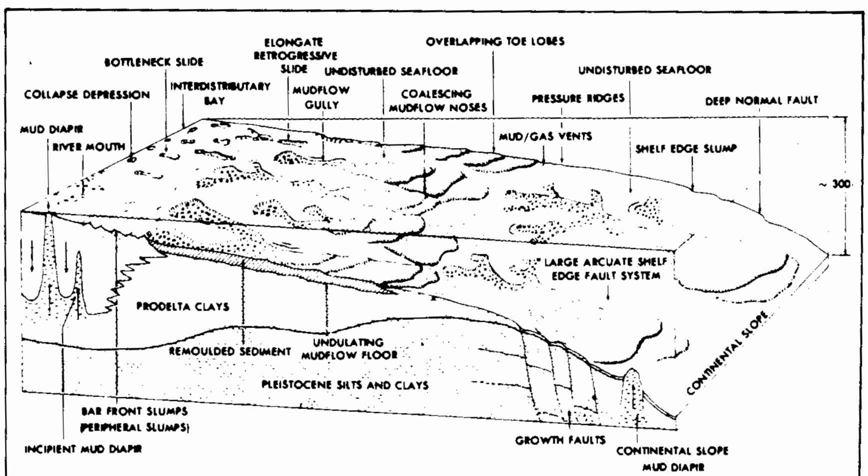

Figure 8
Schematic block diagram showing the various types of delta front to prodelta sediment instabilities off the modern Mississippi delta (Coleman et al., *1983).*

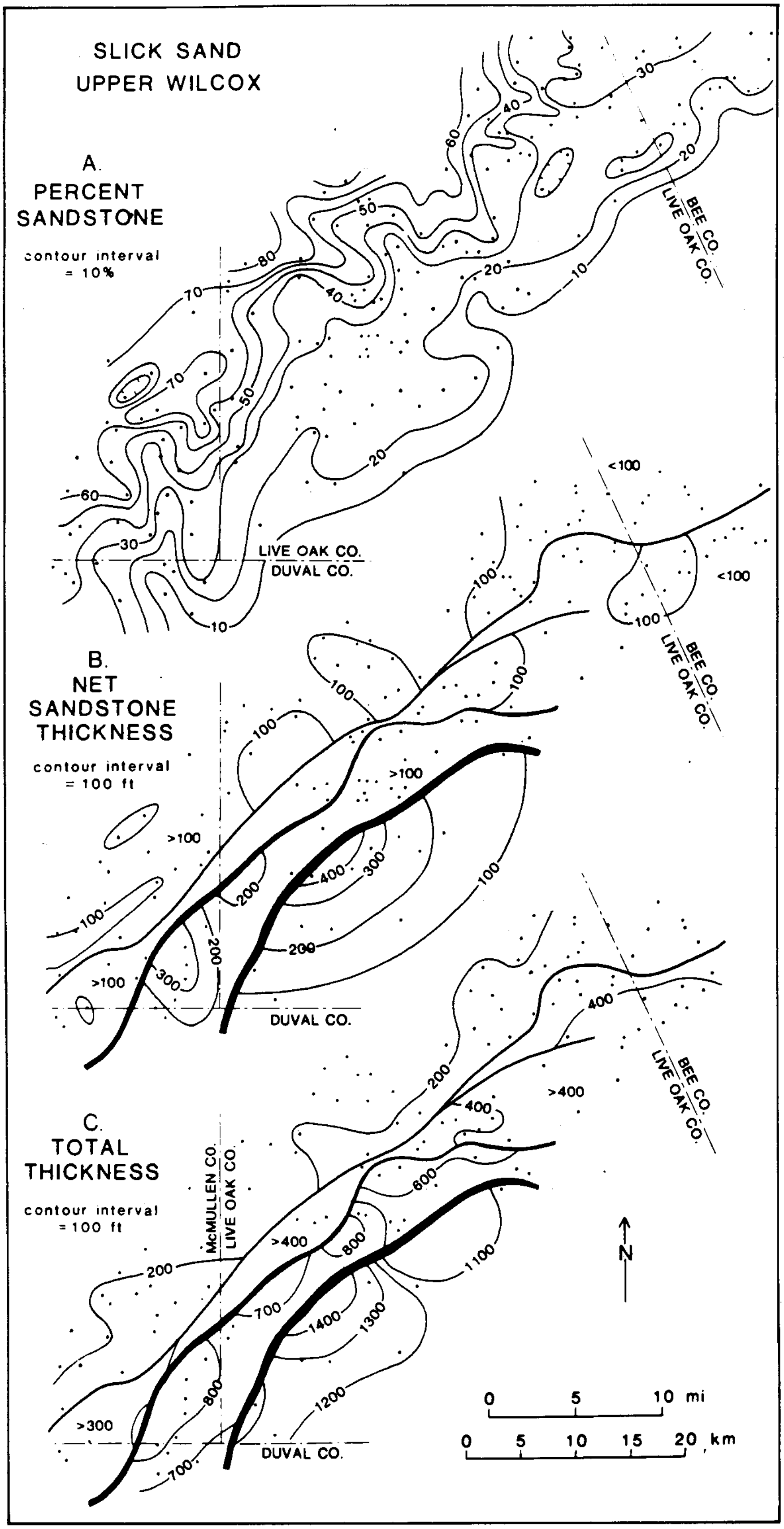

Figure 9
Lithofacies maps of an interval in the Wilcox Sand (Eocene), Gulf Coast. See text for discussion (Winker and Edwards, 1983).

thicken across the fault as a result of syndepositional movement. This occurs commonly particularly during deposition of denser sediment such as sand, and can result in the development of significantly thickened strike-parallel wedges of sandstone in the section (Fig. 9). These wedges could be misinterpreted as wave-modified sand bodies similar to those in Figures 2C and 2D, unless independent evidence or structure (e.g., seismic data) was available (Winker and Edwards, 1983).

The delta-front surface may be unstable because of sedimentary oversteepening and under-compaction. Slumps and slides commonly are the result, generating slide scars, large slide blocks, slump structures and convolute bedding. Diapiric intrusion of prodeltaic mud (or evaporite) into overlying deltaic facies is caused by rapid sediment loading. Growth of the diapirs tends to be long-lived, and they frequently rise to the surface to form sea-floor mounds, or even islands.

DELTAIC CYCLES

Scruton (1960) was one of the first to point out that the growth of a delta is cyclic. The process has now been described many times (e.g., Fisher *et al.*, 1969; Coleman and Wright, 1975; Elliot, 1978). There are two phases.

A) Progradational Phase. Active seaward progradation causes prodelta muds to be overlain by delta front silts and sands, and these in turn by distributary mouth deposits, mainly sands (and gravels, if present), and finally by topset delta marsh sediments, including fluvial facies and peats, mud, or eolian dunes, depending on local climate and sediment supply (Fig. 10).

B) Abandonment Phase. A delta lobe is eventually abandoned if crevassing generates a shorter route to the sea. The topmost beds are then attacked by wave and current activity and may be completely reworked. Compaction and/or subsidence may allow a local marine transgression to occur. The result typically is a thin to moderately thick unit of sands or clays containing a marine fauna, abundant bioturbation and possibly, glauconite. There may be abundant evidence of wave and tide reworking in the form of distinctive assemblages of sedimentary structures.

Lobe switching is probably more common in river-dominated deltas, resulting in a more frequent initiation of new progradational cycles. The overall mechanism probably is similar to wave-dominated deltas (e.g., the Rhône), but may not occur on tide-dominated deltas. Large-scale alternation between the two phases may reflect regional regression-transgression cycles caused by tectonism or eustatic sea level changes. An example is discussed below.

The complete delta cycle (sometimes termed a megacycle) may be about 50 to 150 m (or more) in thickness, but it may contain or pass laterally into numerous smaller cycles representing the progradation of individual distributaries or crevasse splays. As shown by Coleman and Gagliano (1964) and Elliott (1974) these can range from approximately 2 to 14 m in thickness. As in the case of the larger scale cycles they tend to coarsen upward, as described below.

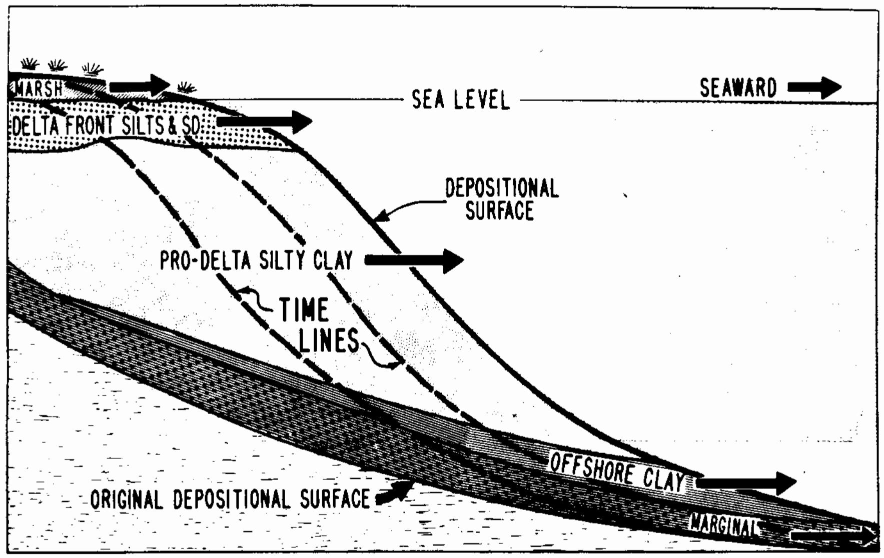

Figure 10
Development of the "clinoform"; depositional surface of the delta front and prodelta: the progradational phase of delta growth (Scruton, 1960).

The manner in which cyclic deltaic sequences are superimposed upon each other depends on the relative rates of sedimentation, subsidence (including compaction) and lobe switching. If the rates of sedimentation and subsidence are in approximate balance a delta will

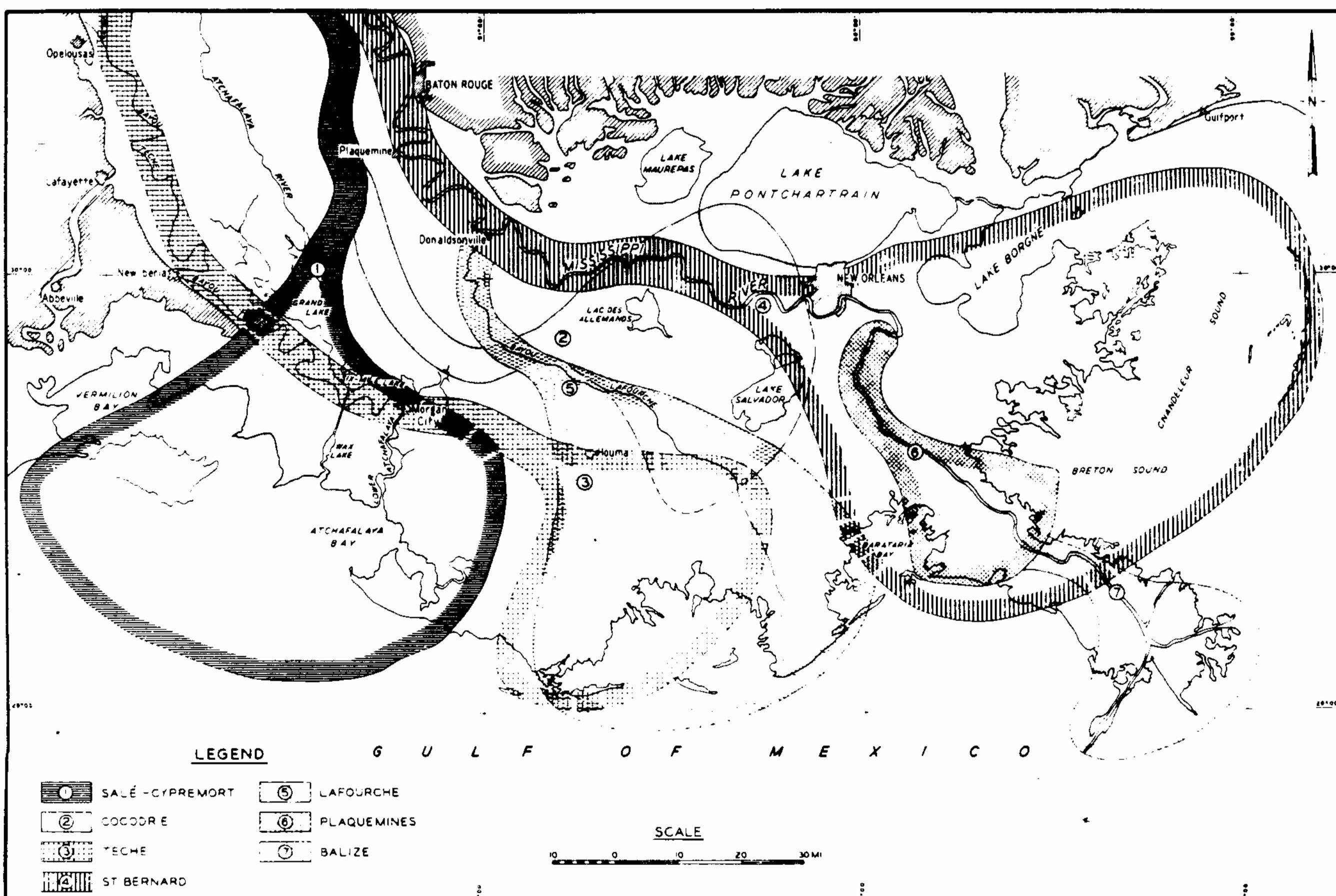

Figure 11
The seven partially overlapping lobes of the Mississippi delta which have developed during the last 5000 years (Kolb and Van Lopik, 1966). Sedimentation is now active again in the area of lobe 1 (Roberts et al., 1980) as well as on the main modern lobe (#7).

tend to build vertically; if subsidence is slower the delta will prograde seaward. As each part of the depositional basin becomes filled, successive progradational events will move laterally (Curtis, 1970, p. 293-297). This is demonstrated dramatically by the Mississippi delta. Here both subsidence and sedimentation have been rapid since the Pleistocene, but the enormous sediment supply has resulted in the development of a suite of seven separate but partially overlapping lobes at the mouth of the Mississippi during the last 5000 years (Fig. 11). The most recent lobe is itself in the process of forming several subdeltas, by similar processes of crevasse splay and distributary switching.

Given a broad shelf or a generally shallow basin a delta may continue to prograde basinward for many kilometres. The depositional surfaces representing each time horizon (Fig. 10) define gently-dipping, wedge-shaped stratigraphic units termed clinoforms. These are very distinctive on regional seismic cross-sections (Fig. 12, Brazos Delta; see Brown and Fisher, 1977; Winker and Edwards, 1983). In strike sections these same units show a large scale mounded or hummocky pattern, recording the lateral switching or offsetting of individual delta lobes.

RECOGNIZING ANCIENT DELTAS

Deltas contain no single distinctive lithofacies but consist of assemblages of lithofacies, each of which can occur in a variety of other environments. It is, necessary, therefore, to identify ancient deltas by a series of steps, eliminating other possibilities and using distinguishing characteristics of facies type, bed geometry and type of cyclic succession to focus in gradually on the correct delta model. This process is complicated by the existence of three end-member "norms", and by the fact that most natural modern and ancient deltas probably are combinations of all three, with added local complications of basin geometry and basin tectonics to be unravelled. In addition, very few good examples of ancient wave- and tide-dominated deltas are available for use as analogues.

The most useful overall indicator of a major deltaic deposit is the presence of a thick wedge or lobe of nonmarine to shallow marine lacustrine sediment, passing basinward into finer grained, deeper water facies, and landward into an entirely nonmarine (usually fluvial) facies (although the latter may have been removed by uplift and erosion of the basin margin). To detect such a deposit requires careful stratigraphic correlation and the application of lithofacies mapping techniques.

Attempts to correlate deltaic units must be carried out with care because the presence of numerous lateral facies changes can be the cause of many mistakes. Cant ("Subsurface Facies Analysis", this volume) describes the methods of subsurface correlation using geophysical logs, and Figure 12 (Southwest Pass) is an example of correlation of a Recent sand unit in the Mississippi delta. Note the typical coarsening-upward profile, and the interpretation of a locally thickened sand wedge in terms of growth fault. Figure 12 (Brazos Delta) is an example of the clinoform seismic facies so commonly recorded from deltaic deposits. This example is of a modern delta, in which the relationship of the dipping depositional surface to the clinoform stratigraphy is quite obvious (see Fig. 10). A word of caution is required, however, because clinoform reflections can be generated in other environments (alluvial fans, submarine fans, continental slopes, reef talus wedges) and so are not always reliable as a primary facies indicator of a deltaic environment.

If a network of well correlated surface or subsurface sections can be developed, the deltas can be delineated

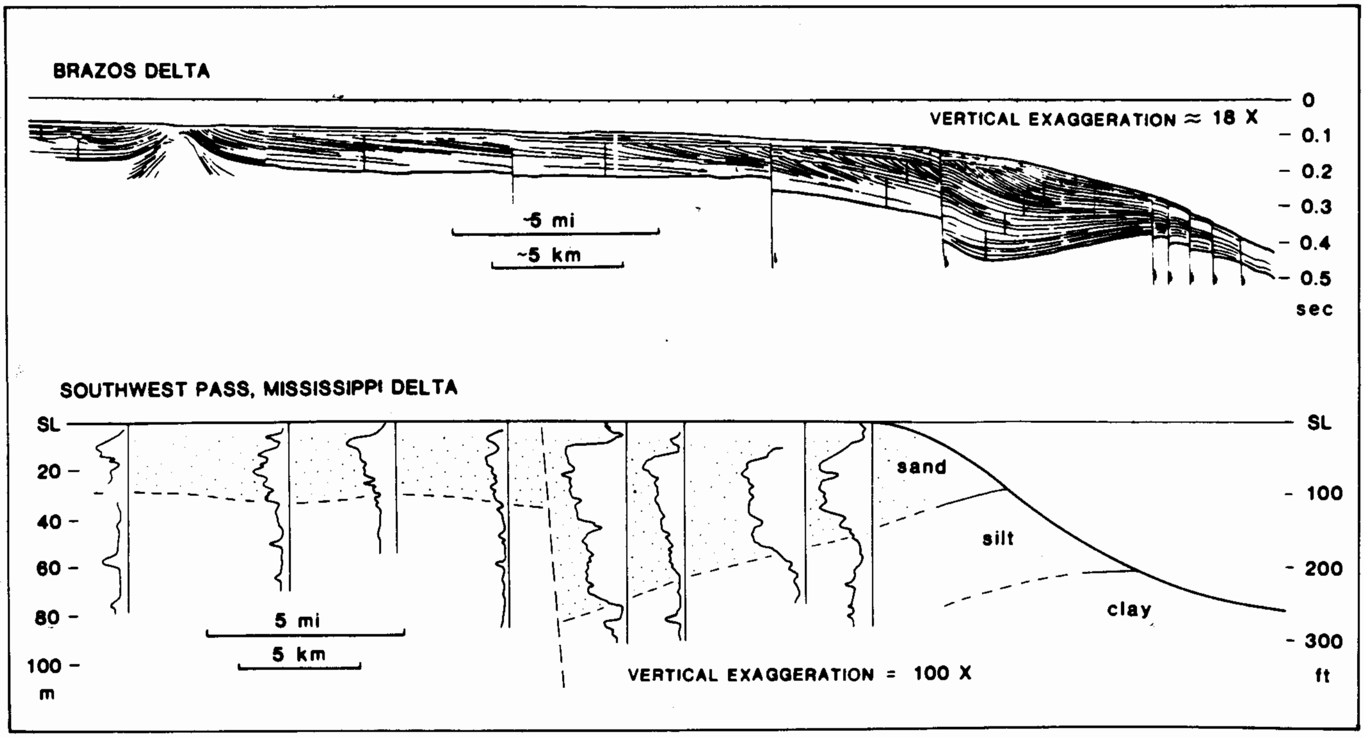

Figure 12
Seismic facies of the modern Brazos River delta (Winker and Edwards, 1983); and subsurface correlation of a deltaic sand-silt unit, in Southwest Pass, Mississippi delta, showing characteristic geophysical log profile of coarsening-upward cycles, and recognition of a growth fault.

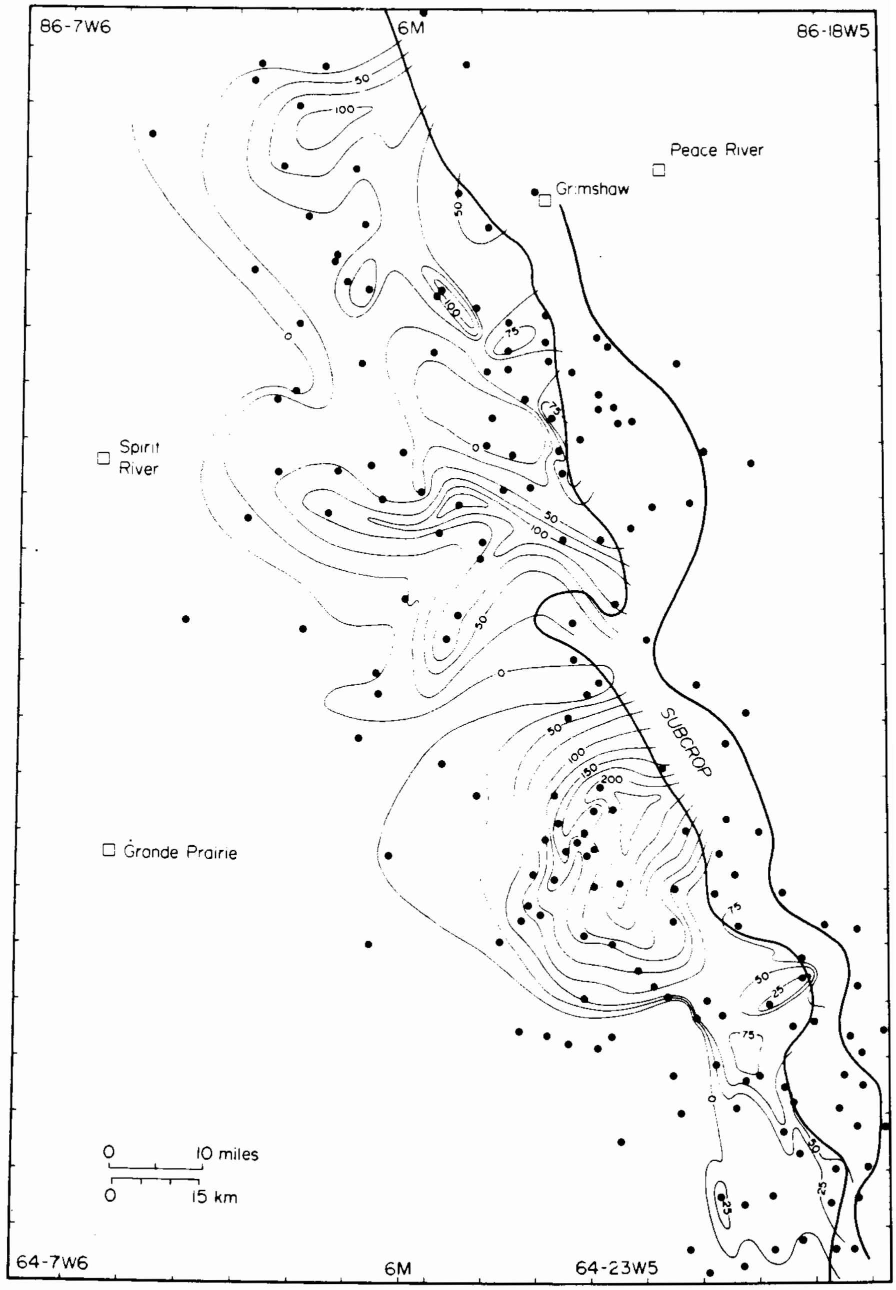

Figure 13
Lobate and birdsfoot deltas in a member of the Triassic Toad-Grayling Formation, northwest Alberta. Contours show the distribution of net porous section, in feet (Miall, 1976a).

using lithofacies mapping techniques. Various parameters may be used, including sand/shale ratio, total sand thickness, or sand thickness expressed as a per cent of a total section. The results may show important differences. For example, the same interval of Eocene sand on the Gulf Coast is mapped in three ways in Figure 9. Map A shows the characteristic lobate pattern of river-dominated deltas, with sand content diminishing distally toward the southeast. Maps B and C show a very different pattern. Lobes of thick sand are present, oriented parallel to strike, but are interpreted here in terms of locally increased subsidence and sedimentation rates along growth faults. The strike-parallel pattern of sand bodies could be confused with that of a wave-influenced delta (see Fig. 2C) if the researcher was not aware of the growth faults. Because the entire thickness of section increases across growth faults, maps of sand percentage (Fig. 9A) may not reveal the effects of syndepositional faulting. However, if allowance is made for these possibilities the outline of local deltaic depocentres revealed by lithofacies mapping techniques may yield useful clues about delta type. For example, Figure 13 shows a map of total porous section (mainly sandstone) in a member of the Toad Grayling Formation (Triassic) of northwest Alberta (Miall, 1976a). The shapes of the lobes and fingers of thick sandstone can be compared to idealized diagrams such as Figure 2. The subcrop of the Toad Grayling beneath the Jurassic is known to be approximately parallel to regional shoreline. The sandstone trends are more or less perpendicular to this shoreline, and have the shape of birdsfoot and lobate river-dominated deltas. Other excellent examples of such maps have been published by Busch (1971) and Wermund and Jenkins (1970).

Interpretations can be refined by detailed examination of vertical sections, using the characteristics of the three end-member delta types as "norms" and as guides for interpretation. For example, they may show the repeated coarsening-upward cycles characteristic of wave- and river-dominated deltas (Fig. 12, Southwest Pass). Cores and outcrops may reveal distinctive assemblages of lithofacies and sedimentary structures, and paleocurrent analysis may be employed (if suitable outcrops are available) in order to map dispersal patterns. Using these data the effects of fluvial and marine currents can be assessed and suitable comparisons with the appropriate deltaic norms (Figs. 1 and 2) can be suggested, and compared with the results of lithofacies mapping.

Figure 14 illustrates two outcrop profiles through the Bokkeveld Group (Early Devonian) of Cape Province, South Africa (Tankard and Barwis, 1982). The generalized section on the left illustrates repeated coarsening upward megacycles, while the detailed, interpreted section shows some of the subenvironments that can be recognized within individual megacycles. Smaller scale cycles up to 20 m thick record the progradation of mouth bars and some of the barrier and tidal sands produced by marine reworking. The lower 105 m of the detailed section is a typical product of river-dominated delta progradation, with a coarsening-

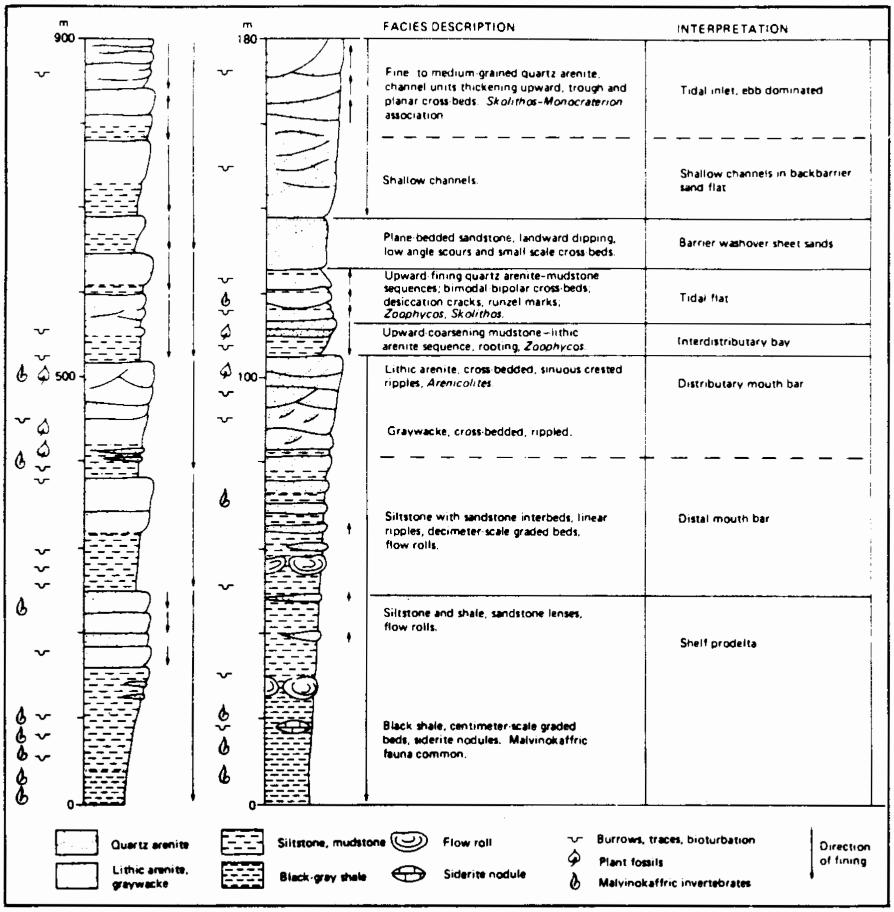

Figure 14
Stratigraphic section through the Bokkeveld Group, Cape Province, South Africa (left) and detailed lithofacies and interpretation of an idealized cyclic sequence (Tankard and Barwis, 1982).

upward succession of shales, siltstones and thin sands representing the build up of the prodelta to distal mouth bar sediments. The sequence is capped by scoured and crossbedded lithic arenites of the proximal mouth bar.

The progradational facies are overlain here by quartz arenites up to 70 m thick showing evidence of wave and tide reworking of the Bokkeveld deltas. Facies and structures are similar to those occurring in other wave- and tide-influenced coastlines (see "Barrier Island and Associated Strand Plain Systems", this volume) but their thickness and associations here suggest a deltaic origin. Barriers and washover sheets are indicated by flat to gently dipping planar sand sheets with a seaward oriented foreshore dip or with the landward dip of washover fans. Tidal inlet and associated delta deposits show polymodal, but commonly ebb-dominated paleocurrent patterns in medium scale crossbedding. Each facies contains a distinct ichnofacies (see "Trace Fossil Facies Models", this volume). The Bokkeveld deltas are interpreted as "wave-influenced" deltas, the lower part of each megacycle shows a predominant fluvial influence, while the reworked marine facies indicate strong wave activity and a moderate tidal influence. This alternation is probably the result of subsidence or sea level change periodically altering the subtle balance between fluvial and marine influences. Another similar example was described by Vos (1981b).

Examples of ancient tidally-influenced deltas have been described by Clemmensen (1976), Eriksson (1979), Verdier *et al.*, (1980) and Rahmani (1982). For example, Eriksson (1979) documented the presence of flood-dominated elongate sand shoals oriented perpendicular to the shoreline, and proposed a model of a non-barred estuary for part of the Archean Moodies Group of South Africa. Figure 15 illustrates a local facies model developed for the modern Niger River by Allen (1970). This river shows elements of all three deltaic end members or "norms", including well-developed beach ridges and active tidal channels undergoing vigorous reversing flow. The lithofacies characteristics shown in the circles around the block diagram illustrate the characteristic coarsening upward nature of the deposit, with distinctive beach-accretion sets and herringbone cross-bedding attesting to the strong marine influence.

Numerous examples of ancient river-dominated deltas have been described. Selected examples are listed in the bibliography. The presence of lobate or finer-shaped deltaic trends, radial paleocurrent patterns, and the characteristic lithofacies assemblages of shoe string sands, interdistributary bays, crevasse splays and mouth bars, are the main criteria for recognizing this type of delta. A Tertiary example is shown in Figure 16 exhibiting, in this case, most of the characteristic features of the river-dominated deltaic "norm".

Increasing attention is being paid to the fan-delta model, particularly by sedimentologists studying pre-Devonian (pre-vegetation) deltas. Another common paleogeographic environment in which fan deltas are found is at the mouths of short, steep rivers carrying abundant bedload. Fan deltas typically lack interdistributary bays and crevasse splays. They show a highly scoured and channelized transition between the coarse, commonly conglomeratic, delta plain and delta front deposits and the finer grained prodelta facies. Selected examples are listed in the bibliography.

CONCLUSIONS

The delta of the Mississippi is still preeminent in the minds of many geologists, for the historical and economic reasons described at the beginning of this paper. However, analyses of ancient deltas are becoming increasingly sophisticated, and the Mississippi is no longer the model automatically used in interpretations of the ancient record.

The next development in the interpretation of ancient deltas may be to interpret the alternations of progradational and abandonment phases in terms of regional changes in relative sea level, and to relate dispersal patterns to

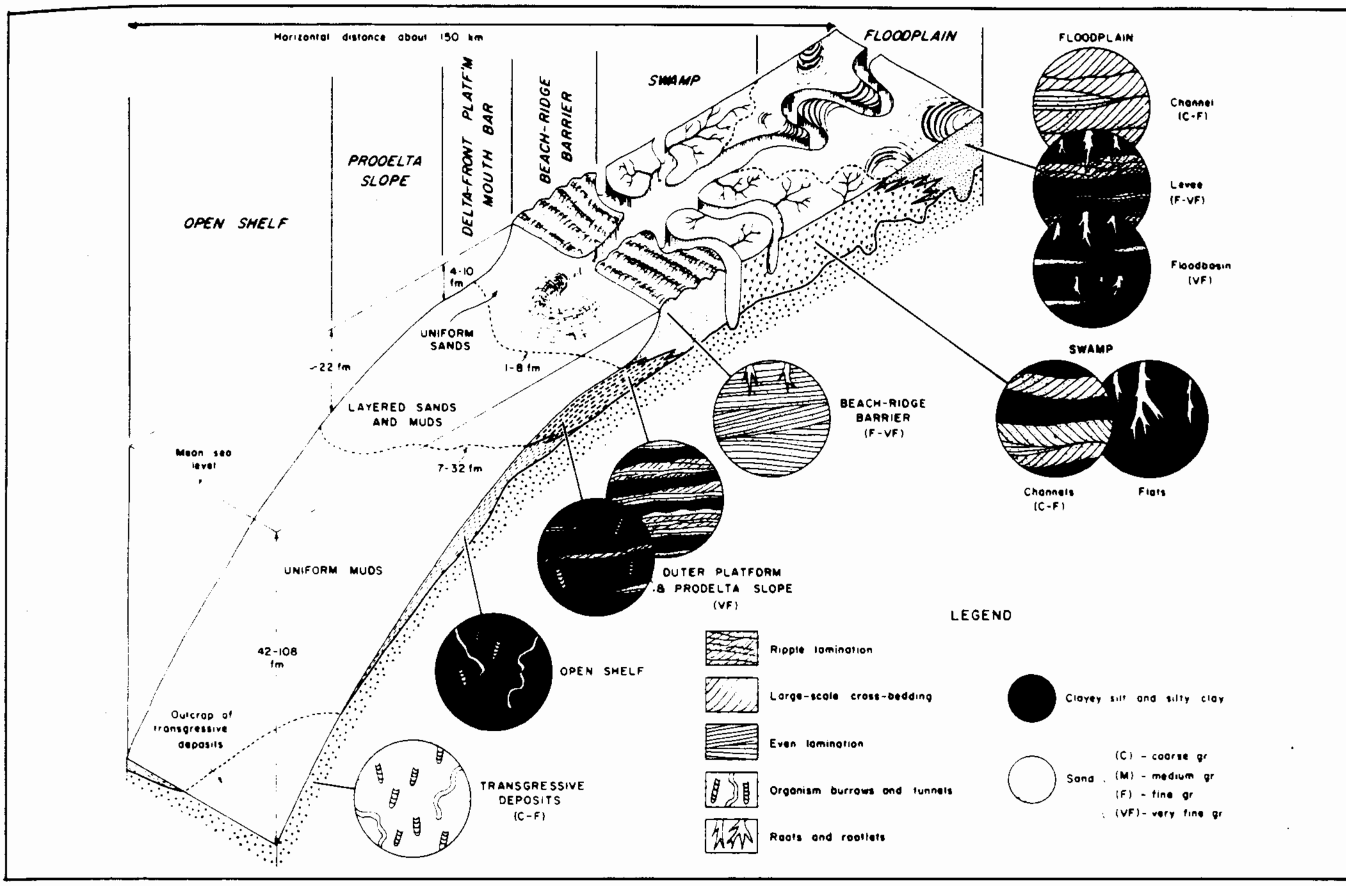

Figure 15
Block diagram model of the modern Niger delta (Allen, 1970).

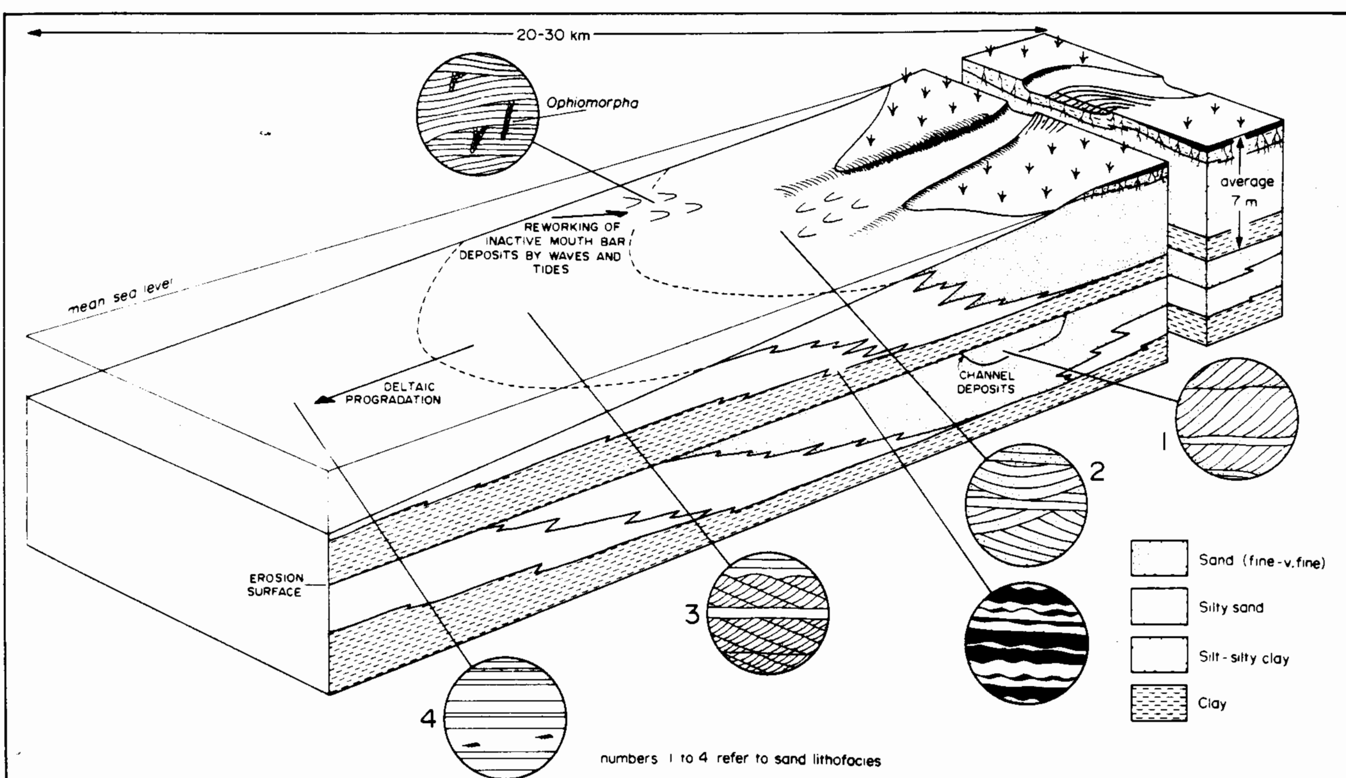

Figure 16
Block diagram model of the Eureka Sound Formation (Tertiary), Banks Island, Arctic Canada, showing interpretation of coarsening upward cycles in terms of distributary mouth bar sands (Miall, 1976b).

tectonic setting and structural grain. As suggested by Miall (1981) some useful information about local plate tectonic history may emerge from this type of analysis.

REFERENCES AND ADDITIONAL READINGS

PRIMARY REFERENCE LIST

Coleman, J.M., 1981. Deltas, processes of deposition and models for exploration. Minneapolis, Burgess Publishing Company, 2nd. Ed., 124 p.
A useful summary of the worldwide researches by the Coastal Studies Institute.

Coleman, J.M. and Wright, L.D., 1975. Modern river deltas: variability of processes and sand bodies. *In* Broussard, M.L., ed., Deltas, models for exploration. Houston Geological Society, p. 99-150.
Discussion of geomorphology and sedimentary processes of the six major delta types.

Eliott, T., 1978. Deltas. *In* Reading, H.G., ed., Sedimentary environments and facies. Oxford, Blackwell Scientific Publications, p. 97-142.
The best up to date study of modern deltas and ancient examples.

Fisher, W.L., Brown, L.F., Jr., Scott, A.J., and McGowen, J.H., 1969. Delta systems in the exploration for oil and gas. Texas Bureau of Economic Geology, 78 p.
Modern and ancient deltas are described and classified. The abundant illustrations are particulary useful.

Galloway, W.E., 1975. Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. *In* Broussard, M.L. ed., Deltas, models for exploration. Houston Geological Society, p. 87-98.
Establishes the by-now widely accepted tripartite classification of deltaic end members or "norms".

Le Blanc, R.J., ed., 1976a. Modern deltas. American Association of Petroleum Geologists, Reprint Series No. 18, 205 p.

Le Blanc, R.J., ed., 1976b. Ancient deltas. American Association of Petroleum Geologists, Reprint Series No. 19, 226 p.
Two collections of historically important and interesting papers.

IMPORTANT STUDIES OF MODERN DELTAS

Allen, J.R.L., 1970. Sediments of the modern Niger delta: a summary and review. *In* Morgan, J.P. ed., Deltaic sedimentation - modern and ancient. Society of Economic Paleontologists and Mineralogists, Special Publication 15, p. 138-151.

Bates, C.C., 1953. Rational theory of delta tion. American Association of Petroleum Geologists, Bulletin, v. 37, p. 2119-2162.
The first description of flow dispersal patterns at river-mouths.

Coleman, J.M. and Gagliano, S.M., 1964. Cyclic sedimentation in the Mississippi River deltaic plain. Gulf Coast Association of Geological Societies, Transactions, v. 14, p. 67-80.

Coleman, J.M. and Gagliano, S.M., 1965. Sedimentary structures: Mississippi River deltaic plain. *In* Middleton, G.V., ed., Primary sedimentary structures and their hydrodynamic interpretation. Society of Economic Paleontologists and Mineralogists, Special Publication 12, p. 133-148.
These two Coleman and Gagliano papers are important contributions on sediments and sedimentary processes in a river-dominated delta.

Coleman, J.M., Prior, D.B., and Lindsay, J.F., 1983. Deltaic influences of shelf edge instability processes. *In* Stanley, D.J., and Moore, G.T., eds., The shelfbreak; critical interface on continental margins. Society of Economic Paleontologists and Mineralogists, Special Publication 33, p. 121-137.
This promises to become a standard paper on the subject, based on many year's work on the Mississippi Delta.

Frazier, D.E., 1967. Recent deltaic deposits of the Mississippi River: their development and chronology. Gulf Coast Association of Geological Societies, Transactions, v. 17, p. 287-315.
A classic study of this birdsfoot delta.

Galloway, W.E., 1976. Sediments and stratigraphic framework of the Copper River fan-delta, Alaska. Journal of Sedimentary Petrology, v. 46, p. 726-737.
Brief description of environments and sedimentary processes.

Kolb, C.R. and van Lopik, J.R., 1966. Depositional environments of the Mississippi River deltaic plain - southeastern Louisiana. *In* Shirley, M.E., ed., Deltas. Houston Geological Society, p. 17-62.
Documents the development of the modern Mississippi delta as a series of overlapping lobes.

McGowen, J.H., 1970 Gum Hollow fan delta, Nueces Bay, Texas. Texas Bureau of Economic Geology, Report of Investigations 69, 91p.
Probably the first thorough study of a modern fan delta.

Oomkens, E., 1970. Depositional sequences and sand distribution in the post-glacial Rhône Delta complex. *In* Morgan, J.P., ed., Deltaic sedimentation - modern and ancient. Society of Economic Paleontologists and Mineralogists, Special Publication 15, p. 198-212.

Oomkens, E., 1974. Lithofacies relations in the Late Quaternary Niger Delta complex. Sedimentology, v. 21, p. 195-222.

Roberts, H.H., Adams, R.D., and Cunningham, R.H.W., 1980. Evolution of sand-dominant subaerial phase, Atchafalaya Delta. American Association of Petroleum Geologists, Bulletin, v. 364 p. 264-279.

Scruton, 1960. Delta building and the deltaic sequence. *In* Shepard, F.P., Phleger, F.B., and van Andel, T.H., eds., Recent sediments, northwest Gulf of Mexico. American Association of Petroleum Geologists, p. 82-102.
Recognition of the delta cycle, and what we now term the "clinoform" stratigraphy of deltaic accumulations.

Van Andel, Tj. H., 1967. The Orinoco Delta. Journal of Sedimentary Petrology, v. 37, p. 297-310

Wescott, W.A. and Ethridge, F.G., 1980. Fan-delta sedimentology and tectonic setting - Yallahs fan delta, southeast Jamaica. American Association of Petroleum Geologists, Bulletin, v. 64, p. 374-399.
Erection of a fan-delta depositional systems model.

Wright, L.D., Coleman, J.M., and Erickson, M.W., 1974. Analysis of major river systems and their deltas: morphologic and process comparisons. Baton Rouge, Louisiana, Coastal Studies Institute, Louisiana State University, Technical Report No. 156.
A study of 34 modern alluvial-deltaic systems using multivariate statistical techniques.

SELECTED EXAMPLES OF ANCIENT DELTAS

A) River-Dominated Deltas

Busch, D.A., 1971. Genetic units in delta prospecting. American Association of Petroleum Geologists, Bulletin, v. 55, p. 1137-1154.
Classic example of dendritic shoestring sands.

Collinson, J.D. and Banks, N.L., 1975. The Haslingden Flags (Namurian G_1) of southeast Lancashire; bar-finger sands in the Pennine Basin. Proceedings of the Yorkshire Geological Society, v. 40, p. 431-458.
Bar finger and distributary mouth deposits.

Dixon, J., 1981. Sedimentology of the Eocene Taglu Delta, Beaufort-Mackenzie Basin: example of a river-dominant delta. Geological Survey of Canada, Paper 80-11, 11p.

Edwards, M.B., 1981. Upper Wilcox Rosita delta system of South Texas: growth-faulted shelf-edge deltas. American Association of Petroleum Geologists, Bulletin, v. 65, p. 54-73.

Elliott, T., 1975. The sedimentary history of a delta lobe from a Yoredale (Carboniferous) cyclothem. Proceedings of the Yorkshire Geological Society, v. 40, p. 505-536.
A river-dominated delta, with progradational and abandonment phases.

Fisher, W.L. and McGowen, J.H., 1967. Depositional systems in the Wilcox Group of Texas and their relationship to occurrence of oil and gas. Gulf Coast Association of Geological Societies, Transactions, v. 17, p. 105-125.
A classic subsurface study.

Horne, J.C., Ferm, J.C., Caruccio, F.T., and Baganz, B.P., 1978. Depositional models in coal exploration and mine planning in Appalachian region. American Association of Petroleum Geologists, Bulletin, v. 62, p. 2379-2411.
Use of detailed outcrop studies to construct local facies models. Many good detailed maps and block diagrams of Carboniferous river-dominated deltas.

McCabe, P.J., 1977. Deep distributary channels and giant bedforms in the Upper Carboniferous of the Central Pennines, northern England. Sedimentology, v. 24, p. 271-290.
Giant bar deposits in distributary channels.

Miall, A.D., 1976a. The Triassic sediments of Sturgeon Lake South and adjacent areas. *In* Lerand, M., ed., The sedimentology of selected clastic oil and gas reservoirs in Alberta, Canadian Society of Petroleum Geologists, p. 25-43.
Subsurface study of river-dominated deltas.

Miall, A.D., 1976b. Sedimentary structures and paleocurrents in a Tertiary deltaic succession, Northern Banks Basin, Arctic Canada. Canadian Journal of Earth Sciences, v. 13, p. 1422-1432.
Facies analysis of a river-dominated delta system. Gross geometry of delta lobes can be outlined from scattered outcrop data.

Wermund, E.G. and Jenkins, W.A., Jr., 1970. Recognition of deltas by fitting trend surfaces to Upper Pennsylvanian sandstones in North-Central Texas. *In* Morgan, J.P., ed., Deltaic sedimentation - modern and ancient. Society of Economic Paleontologists and Mineralogists, Special Publication 15, p. 256-269.
Use of trend surface analysis to map a distributary sand complex in a major delta.

B) Fan Deltas

Dixon, J., 1979. The Lower Cretaceous Atkinson Point Formation (new name) on the Tuktoyaktuk Peninsula, N.W.T., a coastal fan-delta to marine sequence. Bulletin of Canadian Petroleum Geology, v. 27, p. 163-182.
Subsurface study of a small conglomeratic fan delta.

Dutton, S.P., 1982. Pennsylvanian fan-delta and carbonate deposition, Mobeetie Field, Texas Panhandle. American Association of Petroleum Geologists, Bulletin, v. 66, p. 389-407.

Flores, R.M., 1975. Short-headed stream delta: model for Pennsylvanian Haymond Formation, west Texas. American Association of Petroleum Geologists, Bulletin, v. 59, p. 2288-2301.
A coarse-grained, lobate delta.

Sykes, R.M., and Brand, R.P., 1976. Fan-delta sedimentation: an example from the Late Jurassic-Early Cretaceous of Milne Land, central East Greenland. Geologie en Mijnbouw, v. 55, p. 195-203.

Vos, R.G., 1981a. Sedimentology of an Ordovician fan delta complex, western Libya. Sedimentary Geology, v. 29, p. 153-170.

Wescott, W.A., and Ethridge, F.G., 1983. Eocene fan delta-submarine deposition in the Wagwater trough, east-central Jamaica. Sedimentology, v. 30, p. 235-248.

C) Wave-Influenced Deltas

Hubert, J.F., Butera, J.G., and Rice, R.F., 1972. Sedimentology of Upper Cretaceous Cody-Parkman delta, southwestern Powder River Basin, Wyoming. Geological Society of America, Bulletin, v. 83, p. 1649-1670.
A wave-dominated delta. Detailed paleocurrent studies, including the measurement and interpretation or oriented pillow structures.

Tankard, A.J. and Barwis, J.H., 1982. Wave-dominated deltaic sedimentation in the Devonian Bokkeveld Basin of South Africa. Journal of Sedimentary Petrology, v. 52, p. 959-974.
Alternation of river- and wave-influenced sequences, reflecting repeated transgression and regression.

Vos, R.G., 1981b. Deltaic sedimentation in the Devonian of Western Libya. Sedimentary Geology, v. 29, p. 67-88.
Alternation of river- and wave-influenced sequences.

D) Tide-Influenced Deltas

Clemmensen, L.B., 1976. Tidally influenced deltaic sequences from the Kap Stewart Formation (Rhaetic-Liassic), Scoresby Land, East Greenland. Bulletin of the Geological Society of Denmark, v. 25, p. 1-13.

Erkisson, K.A., 1979. Marginal marine depositional processes from the Archean Moodies Group, Barberton Mountain Land, South Africa: evidence and significance. Precambrian Research, v. 8, p. 153-182.
Probably the world's oldest well-described deltaic deposit.

Rahmani, R.A., 1982. Facies relationships and paleoenvironments of a Late Cretaceous tide-dominated delta, Drumheller, Alberta. *In* Walker R.G., ed., Clastic units of the Front Ranges, Foothills and Plains in the area between Field, B.C. and Drumheller, Alberta. International Association of Sedimentologists, 11th International Congress on Sedimentology (Hamilton, Canada), Excursion 21A, Guidebook, p. 31-60.

Verdier, A.D., Oki, T., and Atik, S., 1980. Geology of the Handil Field (East Kalimantan, Indonesia). *In* Halbouty, M., ed., Giant oil and gas fields of the decade 1968-1978.American Association of Petroleum Geologists, Memoir 30, p. 399-421.
Comparison of modern tide-dominated delta with ancient oil producing deltaic sediments.

OTHER REFERENCES CITED IN TEXT

Barrell, J., 1912. Criteria for the recognition of ancient delta deposits. Geological Society of America, Bulletin, v. 23, p. 377-446.
Classic paper dealing with the Catskill "delta".

Broussard, M.L., ed., 1975. Deltas, models for exploration. Houston Geological Society, 555 p.
Papers on process variability and delta classification by Galloway and by Coleman and Wright and a historical survey by LeBlanc are the most useful contributions.

Brown, L.F., Jr. and Fisher, W.L., 1977. Seismic stratigraphic interpretation of depositional systems: examples from Brazilian rift and pull-apart basisn. *In* Payton, C.E., ed., Seismic stratigraphy - applications to hydrocarbon exploration. American Association of Petroleum Geologists, Memoir 26, p. 213-248.
Many examples of deltas and related facies as observed in seismic records. Good discussion of seismic facies.

Curtis, D.M., 1970. Miocene deltaic sedimentation, Louisiana Gulf Coast. *In* Morgan, J.P., ed., Deltaic sedimentation - modern and ancient. Society of Economic Paleontologists and Mineralogists, Special Publication 15, p. 293-308.

Elliott, T., 1974. Interdistributary bay sequences and their genesis. Sedimentology, v. 21, p. 611-622.
A series of vertical profiles for recognizing some of the minor subenvironments in river-dominated deltas.

Miall, A.D., 1981. Alluvial sedimentary basins: tectonic setting and basin architecture. *In* Miall, A.D., ed., Sedimentation and tectonics in alluvial basins. Geological Association of Canada, Special Paper 23, p. 1-33.
Discussion of nine basin-fill architectural models and their relationship to plate tectonic setting.

Morgan, J.P., ed., 1970. Deltaic sedimentation - modern and ancient. Society of Economic Paleontologists and Mineralogists, Special Publication 15.
A somewhat mixed bag of papers, but including much data unavailable elsewhere.

Shepard, F.P., Phleger, F.B., and van Andel, T.H., eds., 1960. Recent sediments, Northwest Gulf of Mexico, 1951-1958. American Association of Petroleum Geologists, 394 p.

Shirley, M.L. and Ragsdale, J.A., eds., 1966. Deltas in their geologic framework. Houston Geological Society, 251 p.
Now rather dated, but containing much useful information.

Winker, C.D. and Edwards, M.B., 1983. Unstable progradational clastic shelf margins. *In* Stanley, D.J., and Moore, G.T., eds., The shelfbreak: critical interface on continental margins. Society of Economic Paleontologists and Mineralogists, Special Publication 33, p. 139-157.
An excellent up to date paper on the extensional deformation (including growth faulting) associated with major deltas, based on a wealth of subsurface data.

Shelf and Shallow Marine Sands

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INTRODUCTION

Of all the major clastic environments discussed in this volume, the shallow marine systems are probably the most complex, due in part to the interaction of many different processes, and in part to the effects of the Holocene rise of sea level on Recent shelf sediments. Research is very active in both ancient and modern shallow marine systems, and many of the ideas in this paper have been developed from promising but not yet fully worked out lines of research. Thus the reader is warned that ideas are in a particularly active state of flux, and that no neatly-packaged facies models have been developed yet. I will use the terms shelf and shallow marine interchangeably, ignoring the structural implications of the correct use of the term shelf.

HISTORY OF SHELF/SHALLOW MARINE STUDIES

One of the earlier ideas about the shelf concerned the grain size distribution. Johnson (1919) coined the term "graded shelf", implying a progressive decrease in grain size from coarse at the shoreline to very fine at the shelf edge. In 1932, Shepard commented that "many geologists have stated that marine sediments vary from coarse to fine as the distance from the shore and depth of water increase. Such gradation seems so reasonable in view of our knowledge of waves and currents that what is practically a geologic axiom has come into existence" (Shepard, 1932, p. 1017-1018). Shepard's point, however, was that the idea of a graded shelf was *incorrect*, and that sediment deposited during low stands of sea level by fluvial or glacial processes could be inundated by a rise of sea level without significant reworking. Such sediment was later termed "relict". Shepard (1932, p. 1038) concluded that "the most outstanding feature of the sediments on the Continental Shelves is the general scarcity of outward decreasing gradation of texture ... the broad shelves have for the most part a patchy arrangement of sediment without any apparent relation to either the shoreline or the other edge of the shelf".

The idea of "relict" sediments was most recently re-stated by Emery (1968). However, shortly afterward it was suggested by Swift *et al.* (1971) that during a rise of sea level, the older sediment could be partly or completely reworked and brought into partial or complete dynamic equilibrium with shelf processes - such sediments were termed "palimpsest". In the same paper, Swift *et al.* (1971, p. 324) first emphasized the different types of shelf currents, identifying: 1) intruding ocean currents, 2) tidal currents, 3) meteorological (storm) currents, and 4) density currents as the four main types. This classification, along with subsequent work, has suggested a division of shallow marine/shelf systems into three main types (Swift and Niedoroda, 1985): 1) tide dominated (17% of the world's shelves), 2) storm dominated (80%), and 3) shelves dominated by intruding ocean currents (3%).

History of Tide-Dominated-Shelf Studies

In these systems, large sand waves or megaripples are the characteristic and dominant bedforms. These were recognized by Van Veen (1935) and Hulsemann (1955) in the southern North Sea, but the important modern studies appeared in the early 1960s; Jordan (1962; Georges Bank, Altantic Shelf of USA), Stride (1963; British Coast), Reineck (1963; southern North Sea) and Off (1963; various localities). More recently, work done by Houbolt (1968) and McCave (1971) in the southern North Sea has led to the definition of tidal sand waves (up to about 7 m high) and tidal sand ridges (up to 40 m high). At about the same time, Klein (1970) described similar features in the Bay of Fundy.

Although many ancient nearshore tidal sandstones (tidal channels, tidal flats) had been described by the mid 1970s, there were remarkably few descriptions of open marine tidally-dominated sandstones. Only one or two are considered in the compilation of examples edited by Ginsburg (1975). In transgressive situations, sandwave complexes may be built and maintained by tidal currents, as discussed by Nio (1976), but there remain few examples of open marine ancient tidal sandstones.

History of Storm-Dominated-Shelf Studies.

The beginning of the present emphasis on storms can be traced to Hayes (1967), who discussed the effect of Hurricanes Carla and Cindy on the Texas coast and shelf. He suggested that during Hurricane Carla a density current had been generated which spread eroded sand from the barrier island across the shelf as a graded bed. This idea has been in the forefront of thought for 15 years, although there are now significant modifications. In the early 1970s, there were several storm interpretations of ancient rocks, for example by Ball (1971, Westphalia Limestone of US Midwest), and Hobday and Reading (1972, various stratigraphic units). However, the most important event was the description of "hummocky cross stratification" (HCS) by Harms (in Harms *et al.*, 1975), and its interpretation as a storm-formed sedimentary structure. Hummocky cross stratification has been used in the interpretation of ancient sandstones since 1979 (e.g., Hamblin and Walker, 1979; Bourgeois, 1980; plus over 100 other examples, Duke, 1985), and is currently the topic of extensive discussion and research. Also, during the last 10 years, there have been very important contributions from marine geologists and oceanographers concerning storm-generated flows on shelves (e.g., Swift and Niedoroda, 1985). However, the geological record gives a very different viewpoint on storms from that of marine geologists, a problem discussed by Swift (1984) and Walker (1984a), and considered again below.

It follows from these brief historical reviews that storm-dominated, tide-dominated, and ocean-current-dominated shelves should be consi-

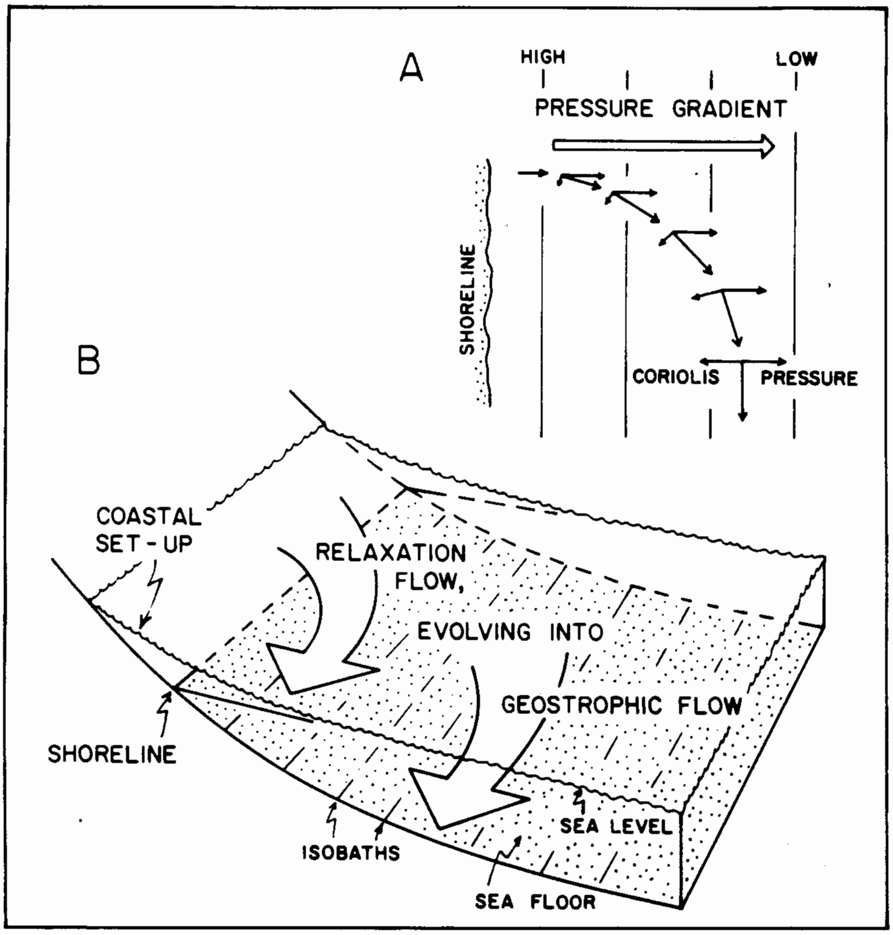

Figure 1
Coastal set-up ("storm surge") creates a seaward pressure gradient. Bottom water flows seaward as a result, but is deflected to the right (northern hemisphere) by Coriolis force to form a geostrophic flow parallel to isobaths. The relaxation, or seaward-directed flow is too transient to be measured, and neither sketch is to scale. After Swift and Niedoroda (1985), and Strahler (1963).

dered separately, although many interpretations of shelf sediments propose combinations of processes.

STORM DOMINATED SHELVES - THE OCEANOGRAPHIC VIEWPOINT

The oceanographer or marine geologist approaches the storm-dominated shelf in a very different manner from that of a land-based geologist, and acquires very different types of data. I will first present the oceanographer's view of shelf processes, and then the marine geologist's view of recent sediment accumulations. Finally, I will consider the land-based geologist's viewpoint, and see to what extent these various approaches converge toward a general model(s) for storm dominated shelves. I will begin by examining storm processes under the general headings of wind-forced currents, relaxation (storm-surge-ebb) currents, and turbidity currents.

Wind-forced currents are generated as the wind blows across the water surface, gradually entraining deeper and deeper ocean layers, until the moving water column may move sediment on the bed. In the simplest case (Fig. 1), the wind forces water onshore, creating an elevation of the water surface and hence a seaward pressure gradient. Water particles experience the pressure force, but also a Coriolis force, and the initial seaward flow will be deflected to the right (in the northern Hemisphere) and will evolve into a geostrophic flow moving parallel to isobaths. D.J.P. Swift (pers. commun., 1983) has suggested that the geostrophic discharge may be 2000 to 3000 times as great as the "relaxation" (seaward oriented) discharge for a 2-day storm. In general, Swift and Niedoroda (1985), note that on a north-south, east facing coast in the northern hemisphere, a wind from either the northeast or northwest quadrant will cause set-up. Similarly, a wind from either the southeast or southwest quadrant will cause set-down. Resulting flows are shown in Figure 2.

There is now some data on the magnitude of such flows close to the bed. Measurements by Swift and colleagues (see Swift *et al.*, 1979; Swift and Field, 1981) have shown near bottom flow velocites of up to about 60 cm/sec on the Atlantic Shelf in depths of 10 to 20 m (Fig. 3) - velocities of several tens of cm/sec occur several times per year as a result of northeasterly storms. These flows generate current ripples and megaripples, and the megaripples may appear in the geological record as medium scale, angle-of-repose cross bedding. The flows also deposit abundant graded beds, as observed in vibracores (Swift, pers. commun., 1984).

More catastropic storm flows have been monitored in the Gulf of Mexico. Tropical Storm Delia (Sept. 3-5, 1973; Forristall *et al.*, 1977) was monitored on a drilling platform 50 km offshore from Galveston Island, in about 21 m of water (Fig. 4). Alongshore flows reached nearly 2 m/sec and seaward-directed flows were between 50 and 75 cm/sec. The peak storm surge of 2 m at Galveston Island did not occur until 30 hours after storm landfall, whereas the maximum alongshore velocity occurred some 38 hours before peak storm surge. This suggests that the alongshore flow was due directly to wind stress, not to storm surge ebb. The geological effects of these flows can only be guessed, but the alongshore velocities suggest effective sand transport, and the creation of ripples, sinuous crested dunes and possible upper plane bed during the 30 or so hours of gradually increasing flows (Fig. 4). In wind-forced flows where the entire water column is moving, flow depths may be much too great for the formation of upper plane bed, although fine grain sizes (fine, very fine sand) even in deep flows favour upper flat bed rather than sinuous crested dunes. The stability fields of the various bed forms have been given by Southard (in Harms *et al.*, 1982), although there is little data for flows several metres deep. As the Tropical Storm Delia velocities decreased again during the following 18 hours (Fig. 4), the plane bed would have been reworked into dunes (if the sand were coarser than the middle of the fine

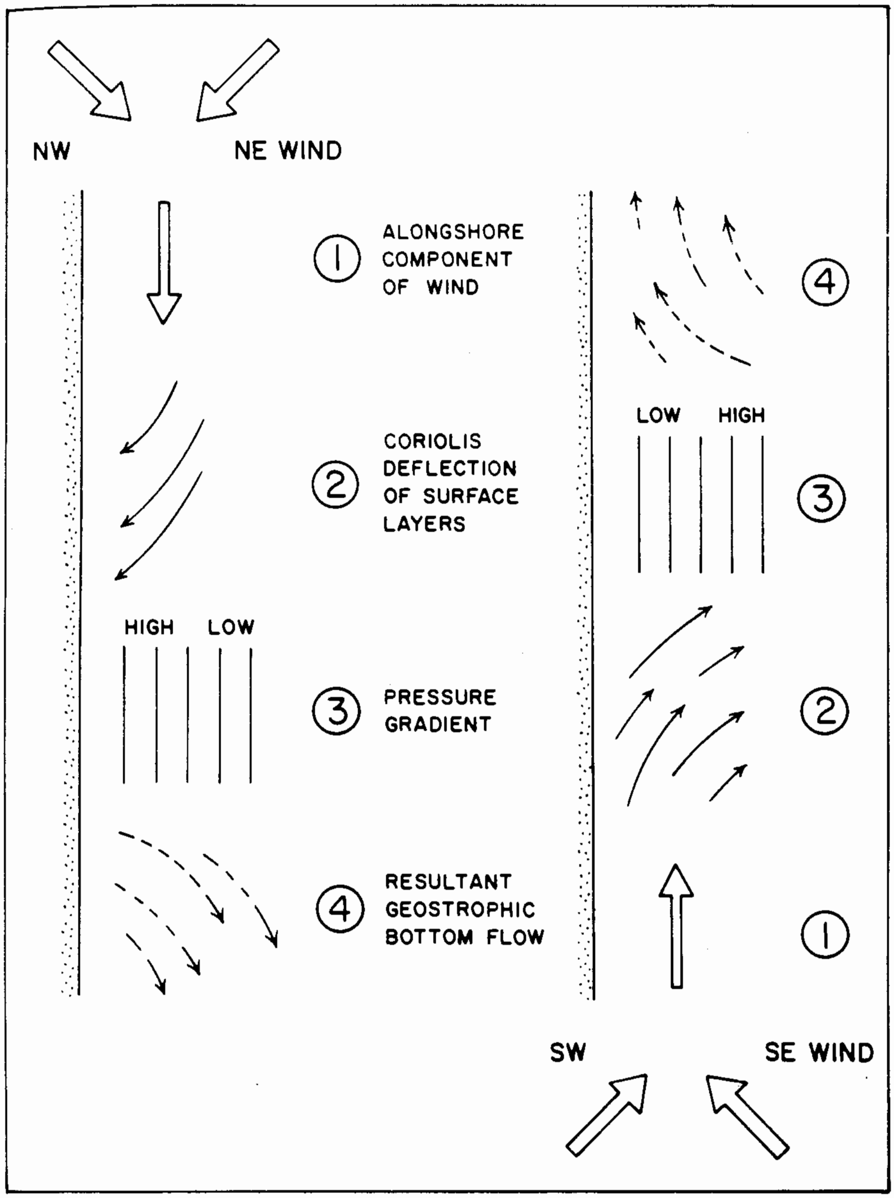

Figure 2
An east-facing coastline trends roughly N-S. Winds from the NW and NE will have a southward, alongshore component. This will entrain surface waters, which are deflected to the right (northern hemisphere), resulting in coastal set-up. Ebb of bottom waters, with Coriolis deflection, will produce a south-flowing geostrophic current. Similarly, winds from the SW or SE wil produce coastal set-down, and the resulting pressure gradient will produce a north-flowing geostrophic current. Note, importantly, that this step-wise analysis is presented only to clarify the processes. It is not observed in the field – what is measured is only the resultant geostrophic flow.

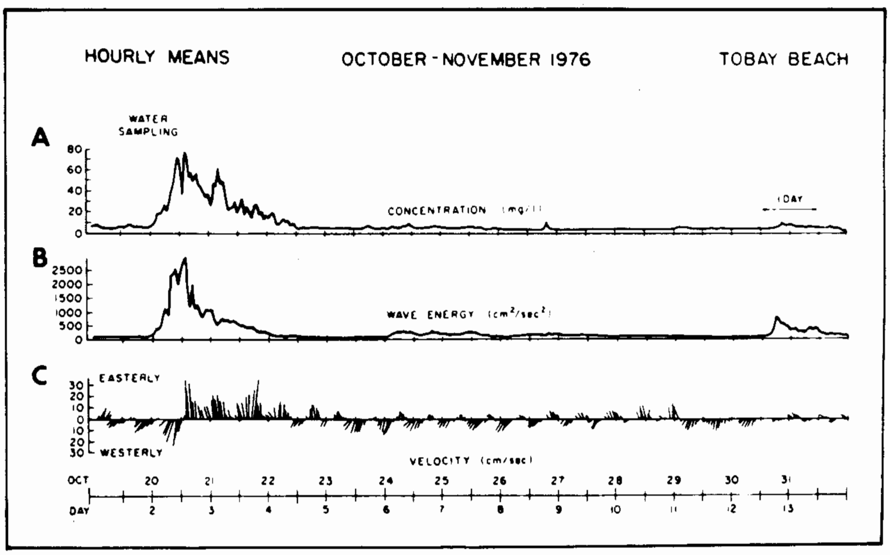

Figure 3
Sediment concentration, wave energy, and flow velocity measured at the 10 m isobath, almost 2 km off Tobay Beach, Long Island, New York. Note storm of October 20-22, and the very gentle, reversing tidal currents of October 22-29th. Data recorded once per second for an eight minute burst, once an hour. From Swift and Field, 1981.

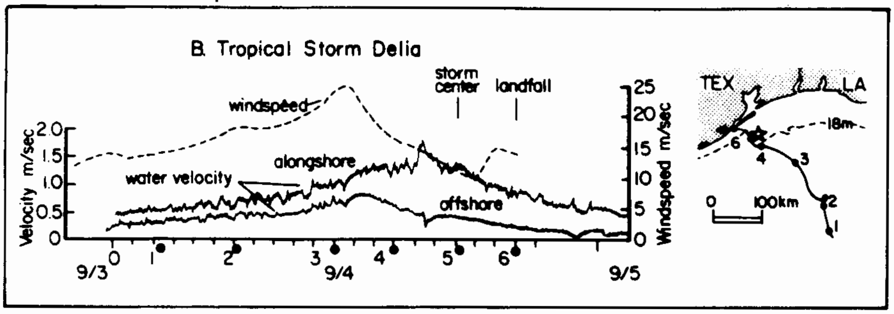

Figure 4
Windspeed, offshore and alongshore current velocities during Tropical Storm Delia, 3-5 September, 1973. Data recorded at site of drilling platform 50 km offshore of Galveston Island, Texas, in a depth of 21 m. From Forristall et al., *1977.*

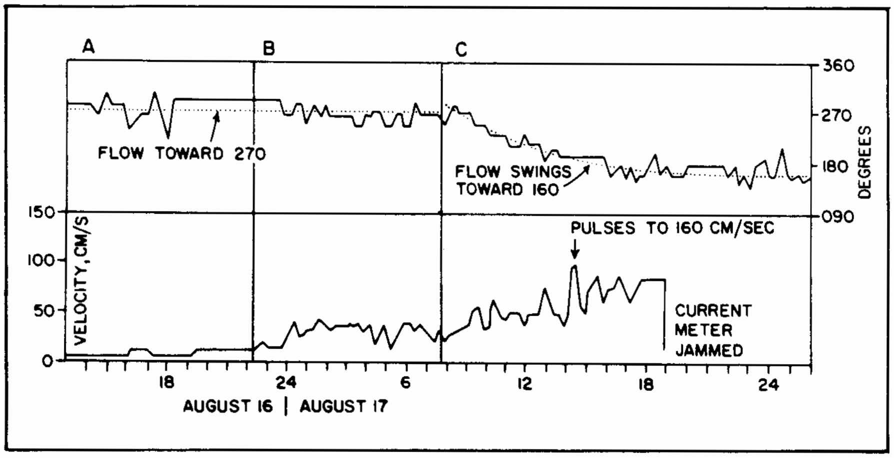

Figure 5
Current speed and directions during Hurricane Camille, 16-18 August, 1969. Recording site was 360 m offshore of the panhandle of Florida, in a depth of 6.3 m, about 160 km east of the path of the eye of the hurricane. Phases A, B and C discussed in text. From Murray, 1970.

sand range), and the dunes subsequently reworked into current ripples. The net result may therefore be an increment of sand transport, but no preservation of sedimentary structures other than ripple cross lamination.

Wind and current volocities were measured during Hurricane Camille (Aug. 16-18, 1969; Murray, 1970) using instruments installed in 6.3 m of water, 360 m offshore of the Florida coast. This location was about 160 km east of the path of the eye of the hurricane. During phase B (Murray, 1980; Fig. 5) flows gradually intensified, and were directed alongshore (toward 270°) - the wind was blowing from the east (080°). Murray's explanation for this flow was "that the longshore current generated in the surf zone set in motion by lateral friction the water beyond the outer bar" (1970, p. 4580). During phase C (Fig. 5), the flows intensified to about 1 m/sec, with pulses to 1.6 m/sec. The wind gradually changed from blowing from the east (080°) to blowing from the south (180°), and the bottom flow changed from 270° to 160° (offshore). The onshore wind piled water up against the shoreline, the resulting offshore (160°) flow can be ascribed to the seaward pressure gradient. This flow may have veered westward (to the right) due to Coriolis forces as it moved further offshore beyond the recording instruments. These storm flows effected an increment of sediment transport, but although dunes (and possibly upper flat beds) were formed locally, they were probably reworked into ripples during waning storm flows. Thus the effects of both Delia and Camille might be undetectable in the geological record of shelf deposits.

Relaxation (storm surge ebb) currents were emphasized by Hayes (1967), and have been subsequently invoked in many geological interpretations. Storm surges, or set-ups are unusually high water elevations due t storm winds blowing water onshore (Fig. 1). As coastal water levels rise, there is a bottom return flow which is normally a geostrophic flow (Figs. 1 a 2). In major storms, particularly hurricanes, the storm surge may reach ma imum elevations of 6 to 7 metres, but these heights are only maintained for to about five hours.

It has been assumed by geologists that the seaward flow driven by this hydraulic head, and the final ebbing c the head itself, would create a bottom current capable of transporting sand seaward below fairweather wave base This was first suggested by Hayes (1967), who observed a graded bed o the Texas Shelf off Padre Island follov ing Hurricane Carla in 1961. Hayes su gested that Carla had created a 7 m storm surge, which ebbed seaward as "density current" (i.e., a *turbidity* current, because the density would have been due to suspended sediment), de ositing the graded bed. However, the m surge was at Matagorda Bay, abou 185 km to the northeast, not at Padre Island (Hayes, 1967, Fig. 5; Morton, 1981). It now appears that the Carla graded bed was deposited by a geostrophic flow moving southward alon the Texas coast, and not directly from seaward moving storm surge ebb.

A second interpretation of recent marine sediments as storm surge ebb deposits was published by Nelson (1982). On the Bering Shelf off the Yukon River Delta, there are graded sand and silt layers with an idealized sequence of sedimentary structures including "basal parallel-laminated medium to fine sand (S_b), a centre sec tion of the sand layer with cross and convolute lamination (S_c), commonly containing laminated beds of epiclasti plant fragments (S_d), and an upper mu cap (S_e),that is absent in many places (Nelson, 1981, p. 539). Nelson also comments that "because of the similai ity to the vertical sequence of structur defined by Bouma (1962) for graded turbidite beds, I have chosen to designate shelf structures by sustituting the of Bouma's $T_{a\text{-}e}$ designation with a cap tal S, signifying shallow water, graded storm-sand beds". The beds can be traced for some 100 km from the Yukc Delta, but no single bed can unambiguously be correlated with a known storm surge ebb. Nelson (1982, p. 541) considers three mechanisms for formation of the layers – the "sudden high river discharge at the time of spring break up . . . wind forced currents and storm surge ebb-flow currents", and comments that "both [wind forced currents and storm-surge ebb-flow currents] appear to be responsible on the basis of limited oceanographic data". He does not discuss the obvious possibility that the "S_{b-e}" beds were actually deposited by turbidity currents. In view of the geological evidence for shelf/shallow marine turbidity currents discussed below, I suggest this is a more likely possibility for Nelson's beds than appealing to 100 km of bedload transport driven by "storm-surge ebb-flows". Swift (pers. commun., 1984) disagrees, and suggests incremental movement of the sand and silt by a geostrophic wind-forced flows. The sharp-based Bouma-like beds would therefore represent the *final* incremental flow.

Many of the ideas presented in this section, and the previous section on wind-forced currents, are presently (Spring of 1984) at the forefront of current research, and are controversial. This is at least partly due to the different types of data handled by geologists and oceanographers. I would suggest the following tentative conclusions:

1) for small coastal set-ups of a metre or two, the relaxation flow (Fig. 1) is deflected by Coriolis forces and evolves into a geostrophic flow. Sand is transported mostly parallel to isobaths, rather than directly seaward. This conclusion is based on extensive measurements during minor (annual) storms (Swift and Niedoroda, 1985).
2) for major coastal set-ups of several metres (5 to 7 m range), there is little oceanographic data on the distances of seaward or alongshore movement of the surge ebb current. Geologists have deduced from the paleocurrent patterns of sharp-based sandstones that sand *can* be transported directly seaward, and introduced suddenly into normally quiet depositional environments. Examples are discussed below. In many cases, they have appealed to "storm surge ebb" currents, perhaps of magnitudes not yet observed and documented by oceanographers.
3) the geological record contains some evidence that the sharp-based sandstones have been introduced by turbidity currents. Possible generating mechanisms are discussed later. If a turbidity current were generated, it would then be driven downslope by gravity, rather than being driven by a combination of pressure and Coriolis forces and therefore evolving into a geostrophic flow parallel to isobaths. The turbidity current possibility is considered below.

Turbidity currents are discussed in detail in a separate paper ("Turbidites and Associated Coarse Clastic Deposts") in this volume. It is emphasized that they are simply a special case of a density current, where the excess density of the flow is due to suspended sediment (rather than elevated salinities or lower temperatures). The turbidity current mechanism can operate in flumes and lakes, in shallow seas and on the continental shelf, and in deep ocean basins. For beds to be *recognized* as turbidites, they must retain the sedimentary structures imposed by the turbidity current, and this normally requires deposition below *storm* wave base.

Turbidity currents have not been directly observed on modern shelves. The gradient of the shelf averages 0° 07', or 0.002 (Shepard, 1963, p. 257), but this is probably no obstacle to turbidity current flow, because in the deep oceans they are known to travel hundreds of km across abyssal plains

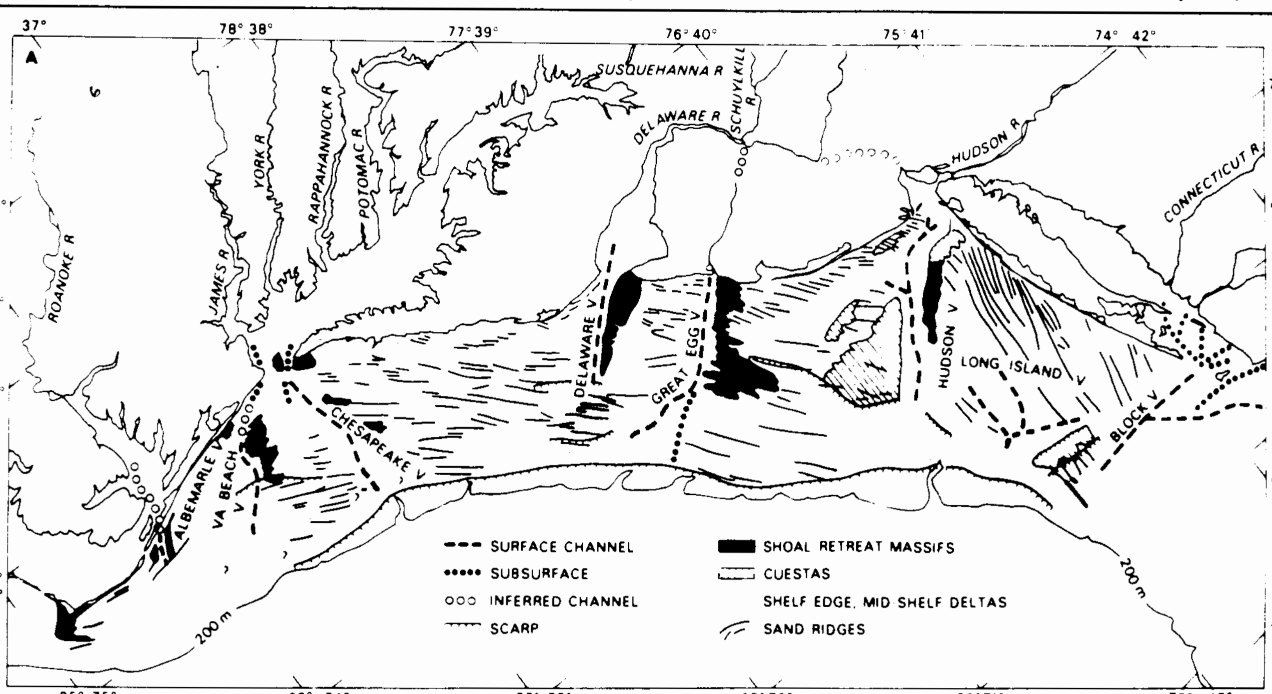

Figure 6.
Major morphological features of the Middle Atlantic Bight. Note relationship of shoal retreat massifs to present estuaries (and capes – e.g., Cape Hatteras, lower left corner of diagram). Note consistent 22° angle between linear sand ridges and the shoreline. From Johnson, 1978, after Swift et al., *1973.*

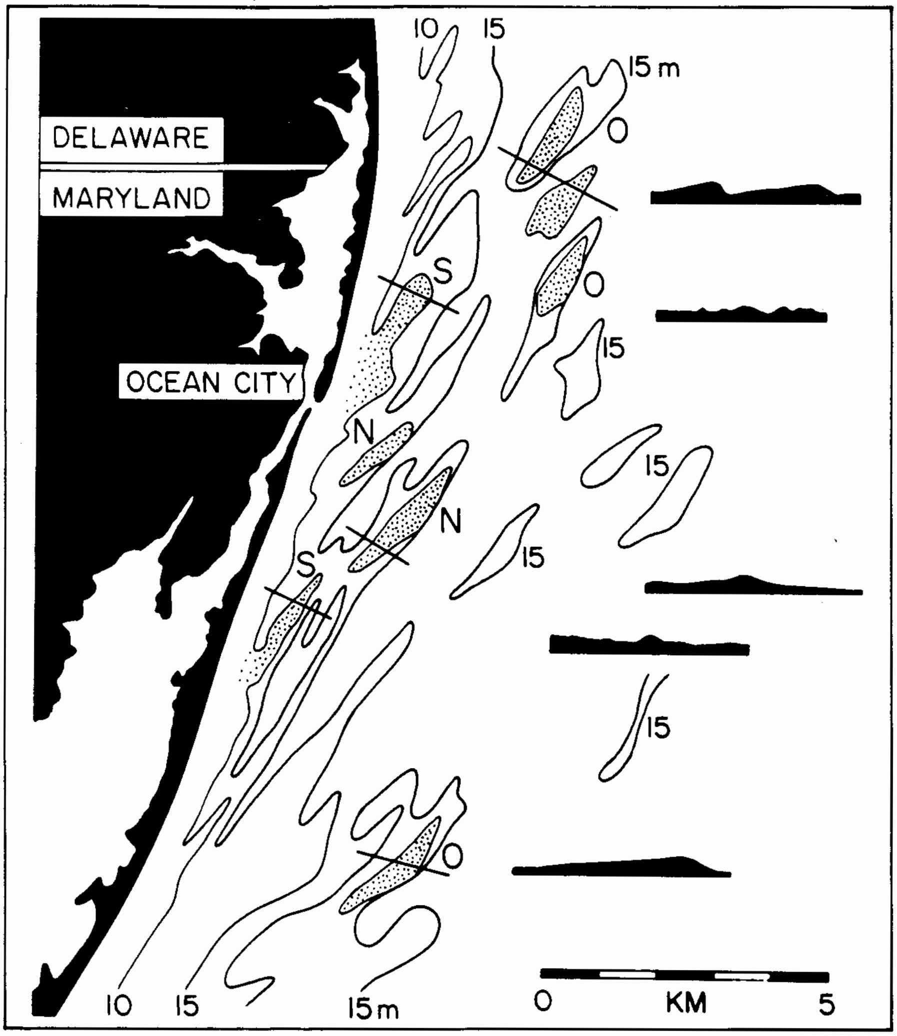

Figure 7
Bathymetry of the Assateague ridge field, offshore Maryland, U.S.A. Note typical shoreface (S), nearshore (N), and offshore (0) ridges, with typical bathymetric cross sections. The bathymetric cross sections are not to the same scale as the map of the ridges. Redrawn from Swift and Field (1981). Isobaths in metres.

with gradients of 1:1500 or less. Before travelling such distances, they have accelerated to high velocities on the continental slope. Such long transport distances would not be expected on the shelf. Here the problem is more one of *flow generation*, and *preservation of the deposit*. Because the "evidence" for shelf turbidity currents is mostly inference from the geological record, I will re-evalue turbidity currents after examining geological storm deposits.

STORM DOMINATED SHELVES - RECENT SEDIMENTS

The most extensive studies of recent shelf sediments are probably those of Swift and colleagues on the Atlantic shelf between Nantucket and South Carolina. Three scales of sand body can be identified:

1) shoal retreat massifs
2) linear sand ridges
3) ripple, megaripple and sandwave bedforms.

Shoal retreat massifs (Fig. 6) formed on the Atlantic shelf during the Holocene rise of sea level. Coastal depocentres were progressively transgressed, leaving a shore-perpendicular sand body on the shelf. Swift *et al.* (1972) identified *cape* retreat massifs, and *estuarine* retreat massifs (Fig. 6). Swift and co-workers have never been specific about the dimensions of shoal retreat massifs, but the Cape retreat massifs of the Carolina coast (Swift, 1975, Fig. 4) average about 10 km wid 30 km long perpendicular to the coas and about 2 m thick. Locally, howeve relief may be up to 20 m, as on Diamo Shoals. The estuarine retreat massifs the Middle Atlantic Bight (Fig. 6) average about 17 km wide and 45 km lon perpendicular to shoreline (Swift, 197 Fig. 3); thickness of the Delaware mas is about 15 m.

Linear sand ridges (Fig. 7) are supe imposed on the shoal retreat massifs but also occur in the areas between retreat massifs. They are up to about m high, 2 to 3 km wide, and have spa ings of about 2 to 6.5 km. They are u a few tens of km long, and make an average angle of 22° (Fig. 6) with the shoreline trend. Maximum slopes are few degrees or less. They are recognized on the Atlantic Shelf from the G of Maine to South Carolina, and have also been studied on the shelf off Argentina (Parker *et al.*, 1982) and in southern North Sea (Swift *et al.*, 197 The most detailed study is that of Sw and Field (1981), in the Delaware-Maryland sector of the Middle Atlanti Bight. Here, they recognize systemat morphological changes from *shorefa* ridges through *nearshore* ridges to o *shore* ridges (Fig. 7), which are summ ized in Table 1. The shoreface ridges can be traced into water as shallow a m before they lose their identity.

Swift and Field (1981, p. 480) concluded that "the sand ridges of the Maryland coast begin as bodies of sa extracted from the littoral drift and stored in the upper shoreface by hyd dynamic processes acting during sto events". Their original suggestion for the oblique orientation of the ridges with respect to the shoreline was a combination of erosional retreat of th upper shoreface combined with high sediment transport rates closer to sho line. Swift (pers. commun., 1984; and Figueiredo *et al.*, 1981; Parker *et al.*, 1982) now suggests that the stability model of Huthnance (1982) might be more appropriate (Fig. 8). In this mod formulated mathematically for tidally dominated ridges in the North Sea, slight original topographic elements tend to grow upward, at a rate propor tional to their spacing and orientation with respect to prevailing currents. Parker *et al.* (1982) note that "in Huthnance's scheme, an oblique orientati

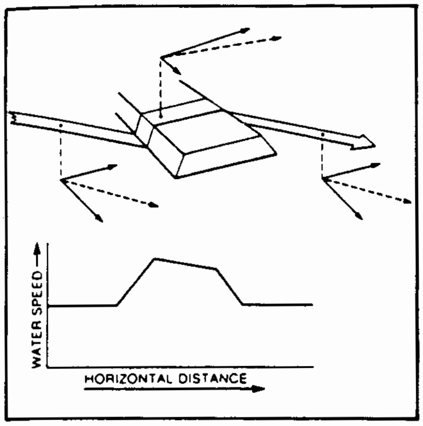

Figure 8
Huthnance model of flow-oblique sand ridge formation, from Parker et al., *1982. As flow moves obliquely up the ridge flank, the cross-ridge component accelerates due to decreasing cross-sectional area. Over the crest, the along-ridge component of velocity is reduced due to frictional drag. The cross-ridge component decreases to its original value.*

Table 1
Dimensions of shoreface, nearshore and offshore linear sand ridges

Ridges, General			
Parameter	**Extreme low**	**Extreme high**	**Typical**
Relief (m)	3.0	12.1	6.1 - 9.1
Length (km)	3.7	18.5	8.3 - 13.9
Width (km)	0.9	2.8	1.4 - 2.3
Spacing (km)	1.5	11.1	4.5 - 6.5
Angle to coast (°)	10	40	15 - 35
Max. slopes (°)	0.2	7.0	0.75 - 2.0

Comparison, Shoreface, Nearshore and Offshore Ridges			
Parameter	**Shoreface ridges**	**Nearshore ridges**	**Offshore ridges**
Mean slope	1.5°	1.0°	0.5°
Steepest slope	2.5°	2.0°	7.0°
Landward:seaward slope	1:1	1:2	1:5
Length:width	9:1	6:1	3:1

with respect to flow is essential because ridges parallel to flow can have no sand carried to their tops . . . ridges normal to flow must experience an acceleration across the crests in order to satisfy continuity, hence cannot deposit sand there . . . when flow is oblique with respect to the ridge crest, the cross-ridge component of flow must similarly accelerate. However, the along ridge components experience frictional retardation and the ensuing deceleration over the crestal area result in sand deposition". In essence, the alongshore geostrophic flows will favour the growth of those irregularities originally oriented obliquely to the shoreline; a rigourous mathematical treatment is given by Huthnance (1982).

As the transgression continues, the shoreface ridges begin to respond to deepening water, and evolve into nearshore ridges. The high frequency component of flow due to wave orbital motion is less important than wind-forced southward moving geostrophic flow, and hence unidirectional flow bedforms (especially sand waves) can form. Mud is deposited in the troughs between linear sand ridges. Over a period of time there is a tendency for erosion on the landward side, and deposition of fine sand on the seaward side, giving a net seaward lateral accretion. In even deeper water, the ridges become more asymmetrical, and adopt "some of the characteristics of flow-transverse bedforms" (Swift and Field, 1981, p. 480). Textural gradients associated with the ridges are preserved as far as 20 km from the beach, suggesting that these *offshore* ridges are still responding to flows – they are not relict.

If ridges of this type are formed by erosional retreat of the shoreface during transgression, we might expect to find them in transgressive situations in the geological record. They will tend to have sharp bases resting on a transgressive lag of coarse material. There are many linear sand bodies in the Cretaceous Western Interior Seaway, but they mostly grade downward into offshore bioturbated marine mudstones. I will show later that none of them has been interpreted in terms of erosional shoreface retreat as described by Swift and colleagues. See note added on p. 164 in second printing.

Ripples, Megaripples and Sandwaves are the smallest of the three types of sand bodies recognized on the Atlantic Shelf. *Ripples* occur from shoreline to shelf edge, and are current formed (asymmetrical) during peak storm flows. They are probably reworked to symmetrical wave ripples as the flow wanes (Swift *et al.*, 1979). The *megaripples* have low crestal continuity, and a spacing of 1 to 40 m. They form on the inner shelf during storm months (November to March) and may cover up to 15% of the area of the inner shelf in patches several km in diameter. They are modified or erased during the quiet summer months, with formation of rounded crests and slopes well below angle-of-repose. Swift *et al.* (1979, p. 401) suggest that this is due to the "ploughing and bulldozing action of benthic animals", although they have recently suggested (Swift *et al.*, 1983) that other rounded megaripples may be the bedform that gives rise to hummocky cross stratification (discussed below). Swift (pers. commun., 1984) notes that both degraded sharp-crested megaripples and true hummocky megaripples are present on the shelf. The *sandwaves* have spacings greater than 40 m, and heights normally less than 1 m. Swift *et al.* (1979, p. 401) note that "the virtual restriction of Middle Atlantic Bight sandwave fields to the inner shelf (<30 m depth) is probably a consequence of the greater frequency with which wind stresses are transmitted to the seafloor in this shallow region". Significantly, "portions of the downcurrent slopes of these sandwaves [are] frequently at the angle of repose" (Swift *et al.*, 1979, p. 401). Large sandwaves, up to about 7 m high "occur where the shelf cross-sectional area is constricted at Kitty Hawk, and on the cape exten-

Table 2
Recurrance intervals for preserved storm emplaced sharp based sandstones

Stratigraphic Unit	Author	Calculated Recurrance Interval, Years
"Passage Beds" (U. Jurassic) Alberta	Hamblin and Walker, 1979	3200
Devonian, Germany	Goldring and Langenstrassen, 1979	400 to 2000
Ordovician, Norway	Brenchley *et al.*, 1979	10000 to 15000
Triassic, Germany	Aigner, 1982	2500 to 5000, or 5000 to 10000
Ordovician, Virginia, USA	Kreisa, 1981	1200 to 3100

Approximate average 5000 years

sion shoals (Diamond Shoals, Frying Pan Shoals)" (Swift, pers. commun., 1984). These sandwaves are built by storm, not tidal currents.

In conclusion it would appear that the linear sand ridges form in the upper shoreface by a combination of oscillatory and unidirectional flow. Thus the internal structure should consist largely of current and wave ripples, and medium scale angle-of-repose cross bedding. Swift (pers. commun., 1984) comments that in vibracores, the most common primary structure after bioturbation is horizontal lamination. During deepening (transgression), ripples, megaripples and sandwaves may form on the flanks and in the troughs of the sand ridges, again forming wave ripples, angle-of-repose cross bedding (from sandwaves), and bioturbated low angle stratification (from megaripples). Continued transgression may allow the burial of the ridges by mud, and hence the preservation of their sedimentary structures and gross morphology in the geological record.

STORM DOMINATED SHELVES - THE GEOLOGICAL VIEWPOINT

The geologist's view of storm deposits is very different from that of marine geologists and oceanographers. This is probably due to the fact that many of the beds in the geological record represent the "thousand year event" (Table 2), and hence were deposited by processes which may not have been observed or monitored by marine geologists. In most shallow seas, we can identify two main areas of deposition. The first is a *shoreface zone* (Fig. 9) characterized by abundant day-to-day sand transport above fairweather water base. The depth of fairweather wave base varies, but normally lies in the 5 to 15 m range (Walker, 1985a). Below fairweather wave base there is the *normally quiet shallow shelf* (Fig. 9), commonly a site of quiet deposition of mud with extensive bioturbation by organisms that graze and mine the substrate (the *Cruziana*, and particularly the *Zoophycos* ichnofacies of Frey and Pemberton ("Trace Fossil Facies Models", this volume). It is of fundamental importance that sand cannot normally cross the interface between shoreface and open shelf, and that only storm-dominated processes can transport sand below fairweather wave base.

The first clue to the storm origin of some beds is therefore the occurrence of sandstones (and/or conglomerates) interbedded with bioturbated mudstones of the *Cruziana* or *Zoophycos* ichnofacies. Such sandstones commonly have sharp and/or erosive bases, and average 5 to 100 cm in thickness (Fig. 10). The tops of the beds may grade up into mudstone, or more commonly, the top contact is disrupted by bioturbation. In this interbedded sand-

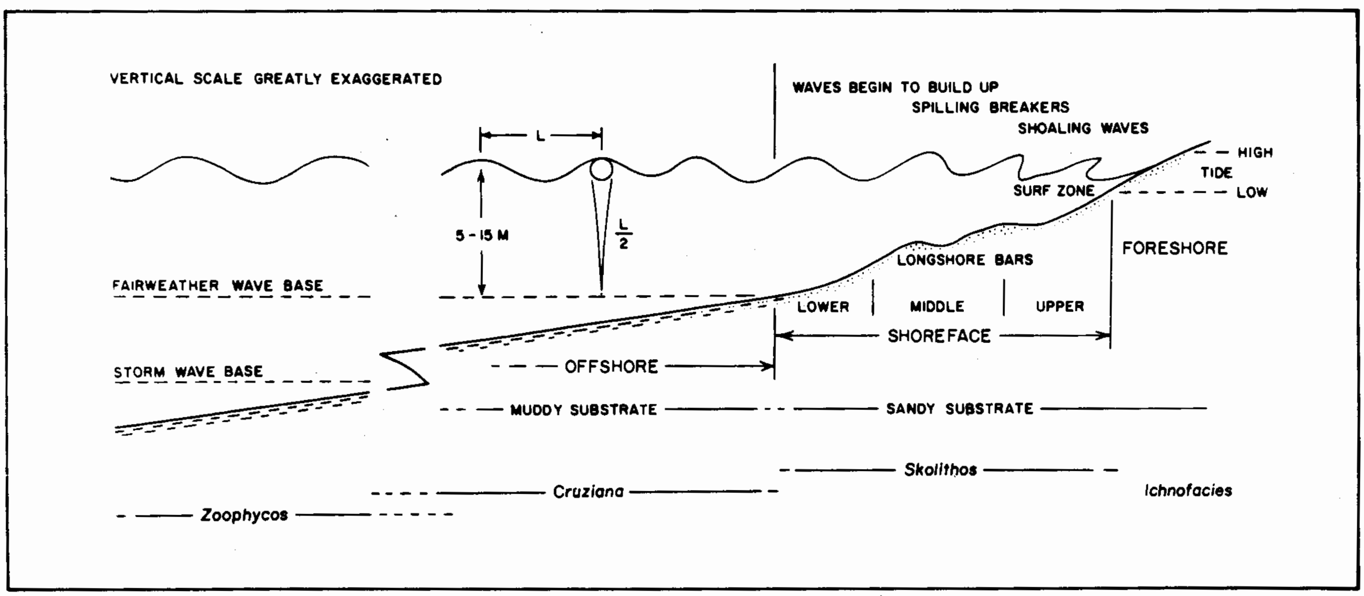

Figure 9
Offshore profile locating foreshore, shoreface and offshore, as well as fairweather wave base and ichnofacies occurrences. Note fairweather waves of wavelength L cannot agitate the bed at depths greater than roughly L/2. Fairweather wave base commonly lies at about 5-15 m depths.

stone/bioturbated mudstone facies, current ripple cross lamination and medium scale angle-of-repose cross bedding tend to be rare or absent, and the characteristic sedimentary structure is hummocky cross stratification (Fig. 11).

Hummocky Cross Stratification (HCS). The structure is shown in the block diagram of Figure 11, and in outcrop form the Cardium Formation (U. Cretaceous) at Seebe, Alberta (Fig. 10). The stratification consists of curving laminations, both convex-up (hummocks) and concave-up (swales), with maximum dips of about 12 to 15°. The laminations commonly intersect at low angles as the hummocks and swales migrate on the sea floor during aggradation. Laminae may thicken toward the centres of swales, but toward the tops of individual beds, hummocks and swales tend to lose their topography, flatten out, and pass into a thin veneer of small symmetrical oscillation ripples. The hummocky cross stratification looks the same regardless of the orientation of a vertical cross section through it; wavelengths are commonly of the order of 1 to 5 m, and heights are a few tens of centimetres. Full geometric descriptions have been given by Walker (1982), Dott and Bourgeois (1982, 1983) and Walker *et al.* (1983).

The term "hummocky cross stratification" was created by Harms (in Harms *et al.*, 1975), and there are now more than 100 stratigraphic units known to contain the structure (Duke, 1985). Harms suggested its formation by "strong surges of varying direction that are generated by relatively large storm waves of a rough sea", and most workers now agree that storm waves acting below fairweather wave base are one of the main agents forming hummocky cross stratification. The interpretation of wave action *below fairweather wave base* is based on the nature of the interbedded bioturbated mudstones, and the fact that in the interbedded HCS sandstone/-bioturbated mudstone facies (Fig. 10), medium scale angle-of-repose cross bedding is rare to absent. This implies that either grain size is consistently too fine for the formation of medium scale cross bedding, or that there has been no fairweather reworking of the storm-formed hummocky cross stratification, hence suggesting original formation above storm wave base but below fairweather wave base. The most recent discussions of the mechanics of HCS formation are by Dott and Bourgeois (1983), Walker *et al.* (1983) and Swift *et al.* (1983) - these latter workers suggest a combination of waning storm-generated unidirectional flows with superimposed oscillatory storm wave action.

Figure 10
Sharp-based hummocky cross stratified sandstones interbedded with bioturbated mudstones. Base of Cardium Formation at Seebe, Alberta. Hummocks (H) *and swales* (S) *are arrowed.*

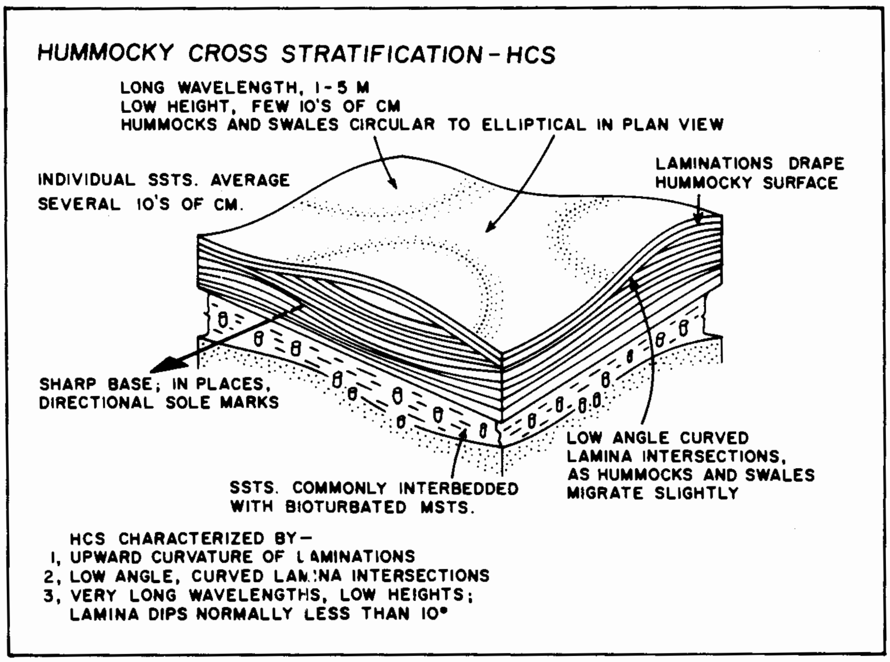

Figure 11
Block diagram of hummocky cross stratification, showing its typical occurrence in the interbedded HCS sandstone/bioturbated mudstone facies. From Walker, 1982.

Hummocky cross stratification has been recognized in Precambrian to Pleistocene deposits, in all parts of the world. It seems to be most commonly developed in fine to very fine sands, but also occurs in very coarse sand and granule gravels (Walker *et al.*, 1983, Fig. 4). It is common in the Cretaceous Western Interior Seaway of North America, particularly in the Gallup Sandstone (Harms *et al.*, 1975), the Cardium Formation (Wright and Walker, 1981; Walker, 1983b) and in parts of the Viking Formation. Hummocky cross stratification can be recognized in cores by the low angle lamina dips, the low angle lamina intersections, and the occasionally visible gentle curvature of the lami-

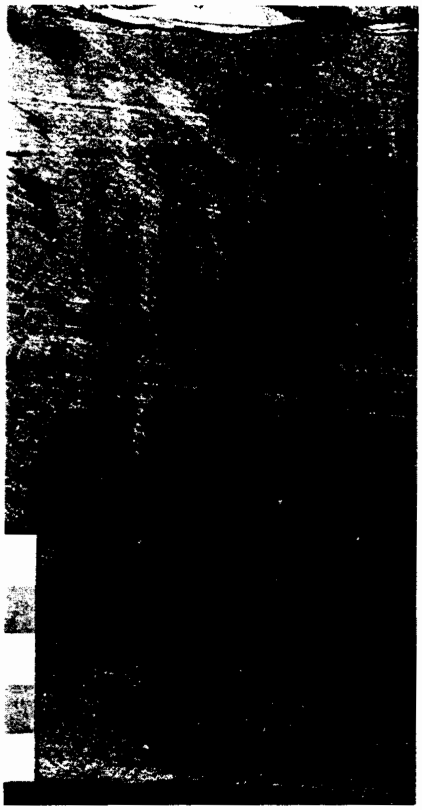

Figure 12
Core showing low angle lamination, with subtle low angle changes in dips of lamination, interpreted as hummocky cross stratification. Scale in cm. Base of sandstone rests sharply on bioturbated mudstone. Cardium Formation, well 4-22-35-8W5 between Caroline and Ricinus fields, southern Alberta, depth 8799 ft (2682 m).

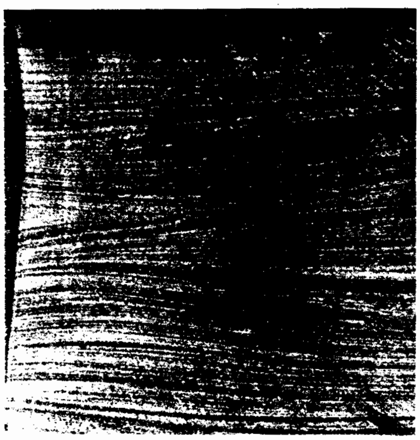

Figure 13
Core showing low angle lamination, with convex-up and concave-up curvature, and subtle low angle changes in dips of laminae. This is apparently not ripple cross lamination, nor angle-of-repose cross bedding, nor upper plane bed - it is interpreted as hummocky cross stratification. Nikanassin Formation, well 7-5-69-9W6. Elmworth Field, central Alberta; 8263 ft (2519 m). Core is about 7 cm wide.

nae (Figs. 12 and 13). The structure appears the same which ever way the core is viewed; part of the identification is on negative grounds, that the structure (Figs. 12 and 13) is neither ripple cross lamination, angle-of-repose cross bedding, nor plane lamination. In a context of bioturbated mudstones, hummocky cross stratification is the most likely alternative.

Emplacement of Hummocky Cross Stratified Sandstones

The identification of hummocky cross stratification as a storm formed structure below fairweather wave base is generally agreed upon, but there are still fundamental problems concerning *how* the sand was first transported to its depositional site before the hummocky cross stratification was formed. There appear to be three main possibilities:

1) wind-forced currents
2) relaxation (storm surge ebb) currents
3) turbidity currents

1) *Wind-Forced Currents.* As reviewed above, wind forced currents commonly achieve speeds of several tens of cm/sec on the Atlantic Shelf, with measured speeds of up to 2 m/sec as a result of hurricane winds. There is little doubt that these flows effectively transport sand below fairweather wave base, and it seems reasonable that superimposed oscillatory wave currents could form hummocky cross stratification. This method of sand transport is here termed *incremental,* in as much as the sand may not have moved to the final depositional site in one sudden flow. It is the incremental addition of sand transported during many storm events that finally creates the deposit. There are no examples of hummocky cross stratification in the literature which are *demonstrably* incremental. If there were, one might also expect to find evidence of geostrophic (parallel to isobaths) dispersal patterns, and some preserved ripple cross lamination or medium scale cross bedding, made by flows which moved sand but without superimposed storm waves to make hummocky cross stratification. Beds would probably tend to be relatively thin. Most workers agree that sharp-based sandstones similar to those in Figure 10 are single depositional events. They could represent the *last* of a series of incremental sand movements, or they may have been emplaced more catastrophically by one of the processes discussed below.

2) *Relaxation (Storm Surge Ebb) Currents.* This has been a popular geological interpretation of hummocky cross stratified beds, and the interpretation is based upon Hayes' (1967) study of hurricanes Carla and Cindy. As discussed above, Hayes' interpretation may be incorrect. Also, Swift *et al.* (in press) have emphasized that the "relaxation" (or seaward-directed) part of a storm surge ebb (Fig. 1) will be minor compared with the geostrophic flow parallel to isobaths. The geological record does not contribute solid evidence concerning seaward-directed bottom flows generated by storm surge ebb, so I will turn to a third mechanism, and then return to the problem of storm surges and their evolution into flows.

3) *Turbidity Currents.* There are two or three geological situations which strongly suggest that sharp based hummocky cross stratified sandstones have been initially transported seaward by turbidity currents.

a)Fernie-Kootenay (U. Jurassic) transition, Alberta (Hamblin and Walker, 1979). A composite stratigraphic section is shown in Figure 14. The lower part consists of classical turbidites mostly with Bouma BC sequences, and abundant sole marks indicating flow toward NNW (Fig. 14). There is a short transition zone where turbidites and hummocky cross stratified sandstones are interbedded (60 m level, Fig. 14), and then the upper part of the section is dominated by sharp-based hummocky cross stratified sandstones (Fig. 15). These have sole marks oriented in an identical direction to those of the turbidites (Fig. 14). The turbidity currents flowed down the paleoslope, driven by gravity acting on the density difference between the flow and the ambient sea water. The flows that emplaced the sharp-based hummocky cross stratified sandstones followed the same paleoslope, and it seems reasonable to suggest that they were also turbidity currents. This paleocurrent relationship would *not* be expected if the turbidity currents flowed downslope perpendicular to isobaths, and the hummocky cross stratified beds were emplaced incrementally by geostrophic flows *parallel* to isobaths.

Instead of Bouma sequences, the beds contain hummocky cross stratification, suggesting that they were deposited from turbidity currents *above* storm wave base, where the storm waves suppressed the Bouma divisions and formed hummocky cross stratification instead. This sequence is very well exposed in a continuous section just east of the traffic circle at Banff, Alberta.

b) Wapiabi-Chungo (U. Cretaceous) transition, southern Alberta (Walker and Hunter, 1982, p. 61-71). Sections at Trap Creek, and particularly in the Highwood River, southern Alberta, show sequences similar to the Fernie-Kootenay transition, namely classical turbidites with Bouma divisions passing upward into sharp-based hummocky cross stratified sandstones. There is not much paleocurrent data in the Trap Creek - Highwood River area, but the same transition at Lundbreck Falls (Bullock, 1981) shows northward flow both for the classical turbidites and for the hummocky cross stratified sandstones (Fig. 16). I deduce that the hummocky cross stratified beds were emplaced by flows responding to paleoslope dip orientation (as for the Fernie-Kootenay transition), making them much more likely to be turbidity currents than wind-forced geostrophic flows paralleling isobaths.

c) Cardium Formation (U. Cretaceous), Ricinus Field, Alberta (Walker, 1985). In the subsurface, dispersal of Cardium sand appears to be southeastward (Fig. 17). In Caroline and Garrington fields, sandstones with preserved sedimentary structures only show hummocky cross stratification. The sandstone at Ricinus clearly channels into the "A sand" at Caroline and Garrington (Walker, 1983a; Fig. 18). In some Ricinus cores, the sandstones are thick and structureless, but in others graded bedding (Fig. 19) and Bouma BC sequences are common. Beds are sharp based, and commonly contain ripped up mud clasts. Rarely, the complete Bouma ABC sequence is present (Fig. 20). In view of the fact that the Ricinus channel (about 45 km long, 4 to 5 km wide before palinspastic reconstruction, and 20 to 40 m deep) lies at least 100, and probably 140 to 200 km offshore (Fig. 17), and in view of the fact

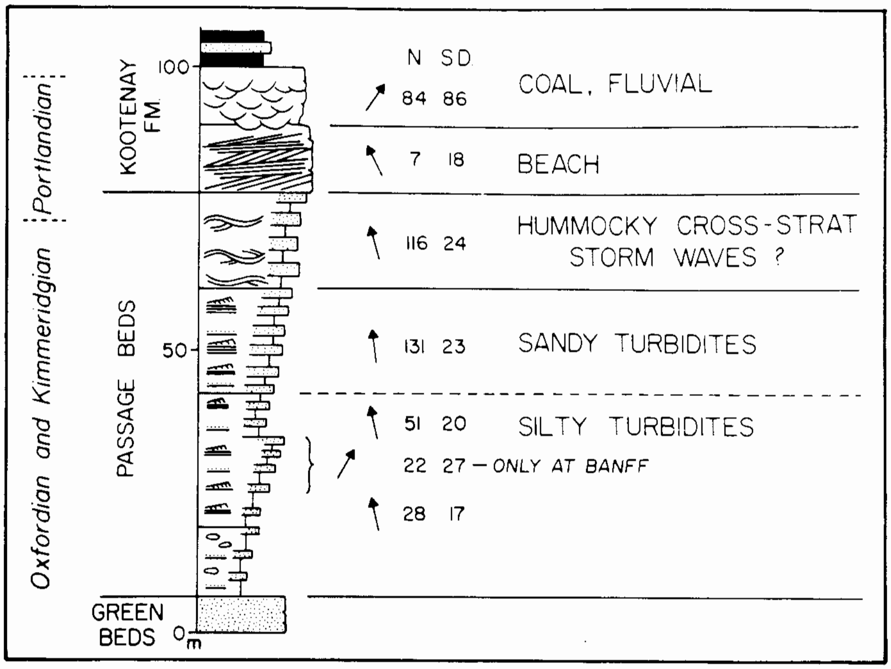

Figure 14
Generalized stratigraphic sequence for the Upper Jurassic "Passage Beds", Crowsnest Pass to Banff area, Alberta. Arrows show paleoflow vector means, N = sample size, SD = standard deviation. Turbidites show Bouma sequences, and form a thickening upward sequence at Banff. Note the same paleoflow directions for HCS beds and tubidites. The part of the section labelled "beach" may have to be reinterpreted due to highway construction and new exposures (1983) near Banff. Modified from Hamblin and Walker, 1979.

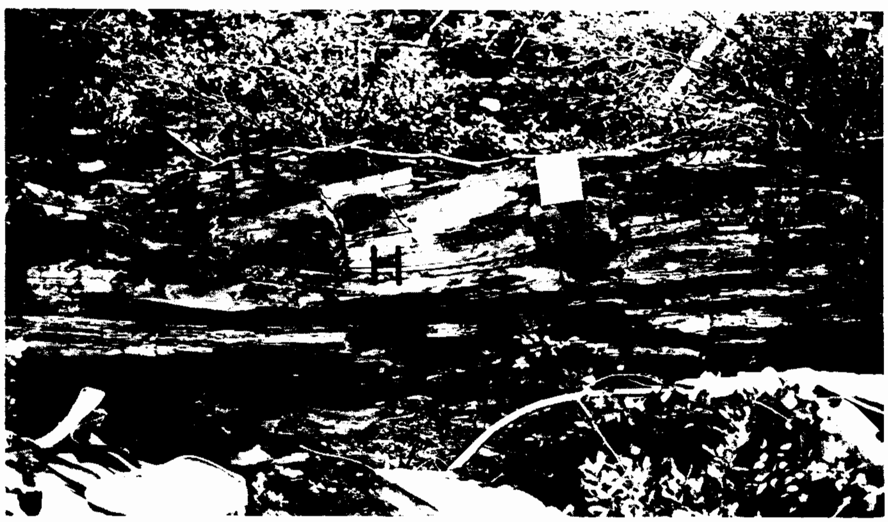

Figure 15
Hummocky cross stratified sandstone bed, intebedded with bioturbated mudstones, from the U. Jurassic "Passage Beds" at Banff, Alberta. The section is overturned, and the photo is shown with BEDDING UPSIDE DOWN. B = base of bed, H = hummock and G = gradational top of bed into mudstone.

that it contains graded beds, and Bouma BC and ABC sequences within the fill, it seems reasonable to conclude 1) that the fill was deposited by turbidity currents, and 2) that the channel itself was probably cut by turbidity currents. Because the A and B sands at Caroline and Garrington also lie similar distances offshore and have similar inferred dispersal directions, it seems reasonable to propose turbidity current emplacement of these sands, but with storm wave modification during deposition, to form hummocky cross stratification.

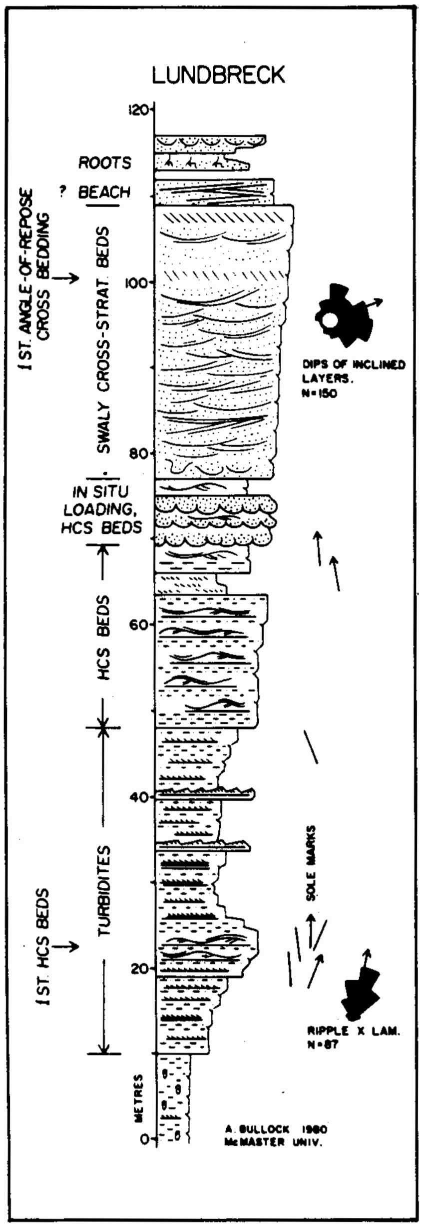

Figure 16
Measured section from Lundbreck Falls, Alberta, of the transition from Hanson (Santonian) to Chungo (Campanian) Members of Wapiabi Formation. Note similar northward paleoflow direction in turbidites and HCS sandstones, and overall progradational nature of the sequence. After Bullock, 1981.

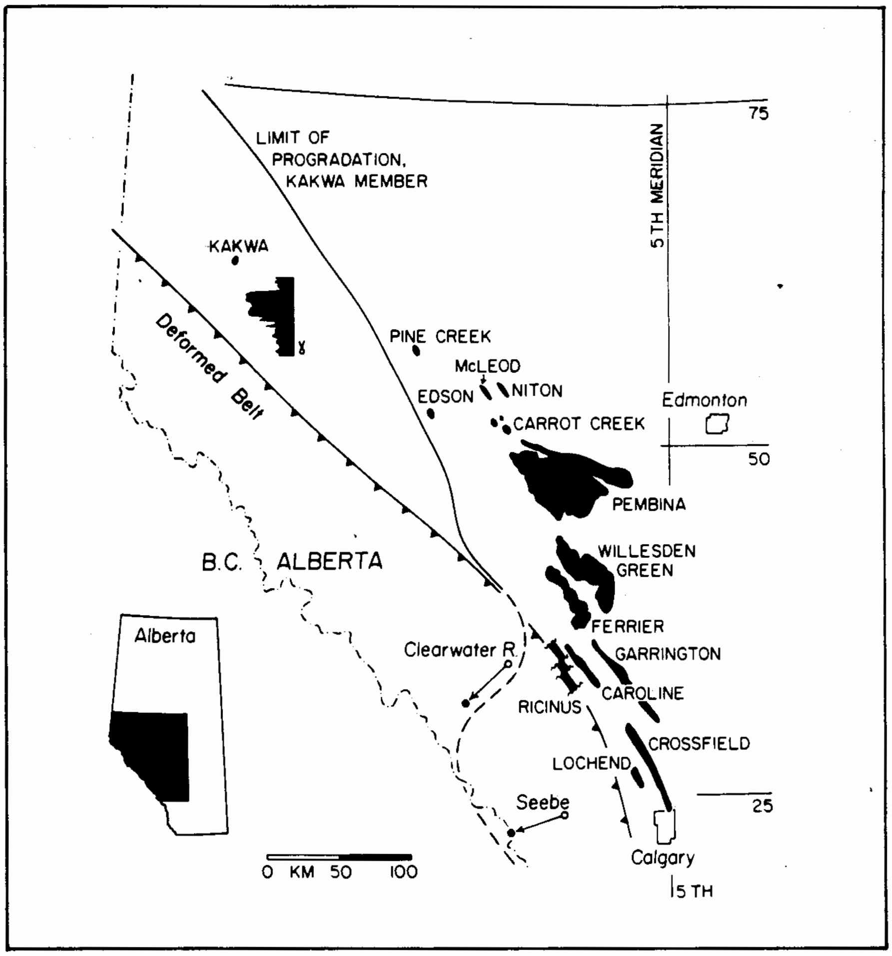

Figure 17
Cardium paleogeography, showing distribution of subsurface oilfields, and limit of northeasterly shoreface progradation of the Kakwa Member of the Cardium. The shoreface has a characteristic response on the gamma ray log. Sediment dispersal was generally southeastward. Numbers on right indicate townships. From Plint et al., *1986.*

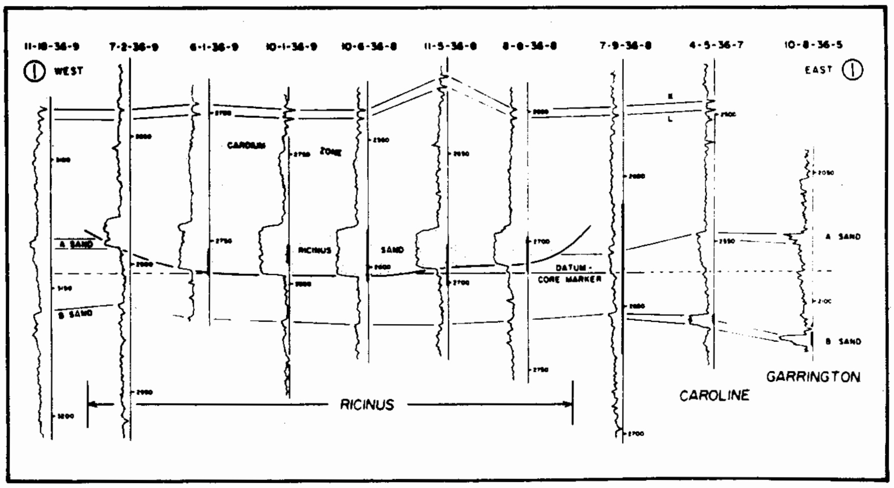

Figure 18
West to east cross section of Cardium sand (Turonian) in northern part of Ricinus field (Fig. 17). All logs are gamma rays – note the "cleaner upward" profiles for the A and B sands at Caroline and Garrington fields (Fig. 17), and the very blocky profile for the Ricinus channel sand. Datum is a core marker, cut out by the channel. From Walker, 1983a, 1985.

Figure 19
Graded beds from Cardium Formation (Turonian), Ricinus field, Alberta, well 3-29-34-8W5. 9384 ft (2860 m). Scale in cm. Dip is due to structural deformation within the field.

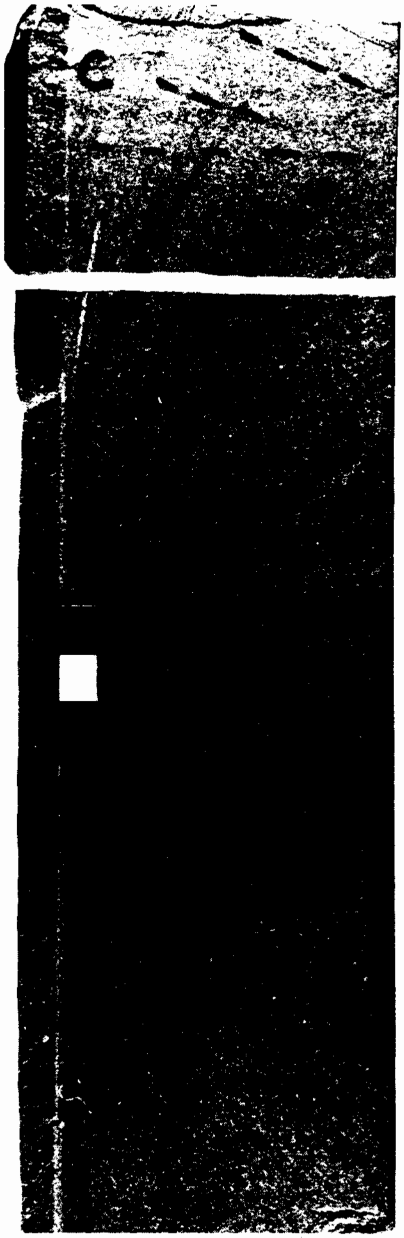

Figure 20
Complete Bouma ABC sequence from a Cardium sandstone in Ricinus field, Alberta, well 10-3-34-8W5, 8685 ft (2647 m). This bed (as well as the graded beds of Fig. 19) is interpreted as a turbidite. Scale in cm.

There is no space here to review the geological evidence for turbidity currents in very great detail, but the facts presented appear compelling. We are forced to ask the question, "is the geological evidence for turbidity currents so strong that we must hypothesize a generating mechanism, or must we re-evaluate the geological evidence because studies of modern storm-dominated shorelines rule out the possibility of turbidity current generation?" I suggest that the geological evidence steers us toward some form of turbidity current generation at the shoreline, and this is summarized in Figure 21.

This model is a development of that presented in the first edition of Facies Models. In the earlier version, it was difficult to envisage how large volumes of sand could be suspended at the shoreline. This is necessary to give sufficient density for offshore gravity flow, rather than alongshore geostrophic flow (Fig. 1). In the revised version (Fig. 21), it is proposed that storms produce cyclic wave-loading of rapidly-deposited fine sediment, which in turn causes liquefaction of the substrate. As this liquefied sediment flows downslope, the combination of flow acceleration and expulsion of pore fluid may be sufficient to suspend fine and very fine sand, and generate a turbidity current. There is an extensive literature on the effects of storm waves on the substrate, liquefaction, and sediment slumping. Readers are referred to the volume edited by Saxov and Nieuwenhuis (1982).

One of the most dramatic examples of substrate liquefaction occurred offshore of South Pass, Mississippi Delta, during Hurricane Camille in 1969, where Shell Oil Company's South Pass 70 Platform B overturned and sank (Sterling and Strohbeck, 1975; Bea and Audibert, 1980). Substrate borings before and after the accident showed a significant loss of cohesive shear strength in the upper 80 feet (24 m) of sediment. An area of at least 6.7×10^6 m^2 showed loss of sediment to depths averaging about one metre- this sediment flowed, and subsequently piled up in adjacent areas, raising the sea floor. Although a turbidity current does not appear to have been generated here, a slightly steeper deltaic slope than that of the Mississippi might have led to a moving liquefied sediment mass that could have accelerated to become a turbidity current (see Morgenstern, 1967). A slump of South Pass Block 70 size (roughly 6.7×10^7 m^3) could have formed a deposit averaging 20 cm thick and covering an area of 33.5 km^2. In the Cretaceous Western Interior Seaway, uplift of the Cordillera would give rapid sediment supply and relatively steep basin margins- perhaps the ingredients needed for storm wave loading to be effective in generating turbidity currents.

STORM-DOMINATED SHELVES-SYNTHESIS AND MODELS

It is presently impossible to synthesize observations from oceanography, marine geology, and the geological record into a single model. Two main "associations of features" are perhaps emerging: 1) an association of bioturbated mudstones with sharp based hummocky cross stratified sandstones

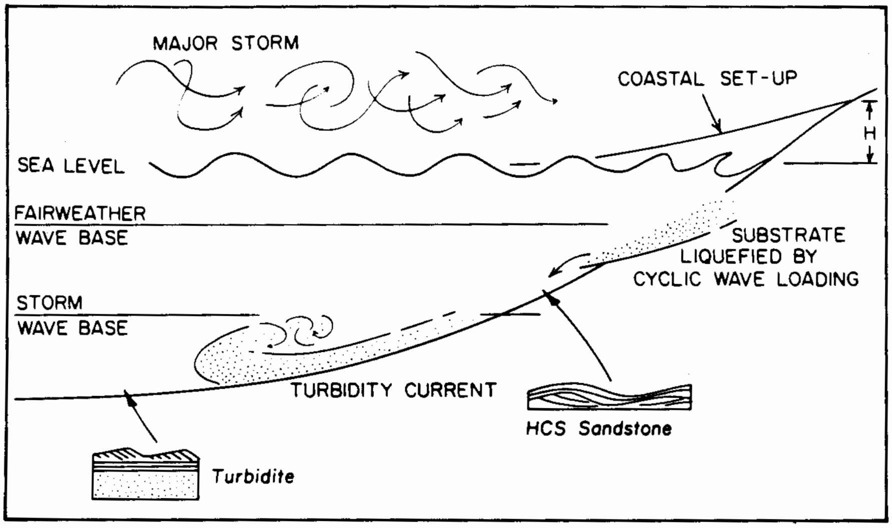

Figure 21
Storm winds create coastal set up, and cyclic loading of the substate by storm waves may liquefy the substrate. The liquefied sediment may flow and accelerate basinward, transforming into a turbidity current with all of the sediment in suspension. Deposition from this flow below storm wave base would result in turbidites with Bouma sequences. Above storm wave base, waves feeling the bottom would rework the turbidity current deposits into hummocky cross stratification. HCS could also form above fairweather wave base, but would probably be reworked into other sedimentary structures by the fairweather processes See text for details.

10-100 cm thick (Figs. 10 and 15), and 2) an association of various scales of cross-bedding in sharp based linear sand ridges a few km wide, several km long, and 3 to 12 m thick (Figs. 6 and 7; Table 1). There are no descriptions of shorefaced-detached, storm-maintained ridges (Fig. 7) in the geological record.

It is puzzling that the better known features of marine geology- the storm-maintained ridges- are not known in the geological record, whereas the abundant HCS sandstone/bioturbated mudstone facies of the geological record (Figs. 10 and 15) is almost unknown in recent sediment studies. The HCS sandstone/bioturbated mudstone facies functions as a model of limited scale, in a similar way to the Bouma sequence for turbidites (see "General Introduction", and "Turbidites and Associated Coarse Clastic Deposits", this volume). It is a point of reference (a "norm"- see especially Dott and Bourgeois, 1982 and Walker *et al.*, 1983), a guide to observations, and a basis for interpretation (Dott and Bourgeois, 1982). In its characteristic position in vertical facies sequences (Figs. 14 and 16), the HCS sandstone/bioturbated mudstone facies functions as a predictor of facies above and below.

TIDE-DOMINATED SHELVES: TIDAL CURRENTS AND RECENT SEDIMENTS

Tides are generated by the gravitational attraction of the moon and the sun, interacting with the rotation of the earth (Coriolis force) and the geometry of ocean basins. The resulting tides may be *semidiurnal* (two high and two low tides per day), *diurnal* (one high and one low per day) or *mixed.* A good non-technical discussion of tidal current generation is given by Fox (1983), and a more technical account is presented by Howarth (1982).

Enclosed basins, or basins with small connections to open oceans, tend to have very low tidal ranges. Open seas (or enclosed seas with large connections to open seas) have higher tidal ranges, the range partly depending on the resonant oscillation period of the basin, which in turn depends upon configuration and depth. In tidal seas such as the North Sea, a progressive tidal wave moves parallel to the shoreline, circling an *amphidromic point* of zero tidal range (Fig. 22). At any one point, the tidal flow can be expressed as a vector showing magnitude and direction. These vectors can be enclosed in an ellipse ("tidal ellipse") which shows the change of magnitude and direction during one tidal cycle. In more open parts of the North Sea, the tidal ellipses locally approach circles, but in restricted areas, the ellipses are strongly elongated (Fig. 23). Here the velocities are for a height of 1 m above the bed, and they indicate that off Lowestoft, for example, flows are oriented strongly northward at about 1 m/sec for about half the tidal cyle, and strongly southward at the same velocity for the other half of the cycle.

By contrast, the Atlantic Shelf of North America has a low tidal range except in enclosed areas such as the Gulf of Maine and the Bay of Fundy, and on shallow banks such as Georges Bank. Near-bottom tidal currents measured by Swift *et al.* (1979; Fig. 3) off Tobay Beach (Long Island) in 10 m water depths barely reach 18 cm/sec, which is the threshold velocity for sediment movement in this area. It is noticeable that flow strengths are asymmetrical (Fig. 3) - this is a common pattern and results in unidirectional sediment movement even in areas of bi-directional tidal flows. The asymmetry may be one of magnitude and/or duration, or ebb and flood currents may follow different paths - this is particularly the case where bottom flows are strongly influenced by large sand waves or tidal sand ridges.

Sand Bodies in the North Sea

The origin of sand around the British Coast and in the North Sea is largely from the unmixing of pre-Holocene deposits during and after the Holocene transgression. Transport paths have been documented by Stride (1963) and Johnson *et al.* (1982). In Figure 24, it can be seen that sand is dispersing from the English Channel northeastward into the southern North Sea, and the down-transport path shows a progressive change from bare rock with gravel lag, through sand ribbons and scattered dunes into the main tidal sand wave accumulation. The sand waves off the Rhine-Meuse Estuary (McCave, 1971) cover an area of about 15000 km^2, and are up to 7 m high (Fig. 25), with most of the steep sides facing northeast. The echo-sounder profiles show smaller megaripples on the backs of the sand waves but the vertical scale is grossly exaggerated - the "steep"faces have dips of only 5 to 6°. Angle-of-repose slip faces develop only in areas where there

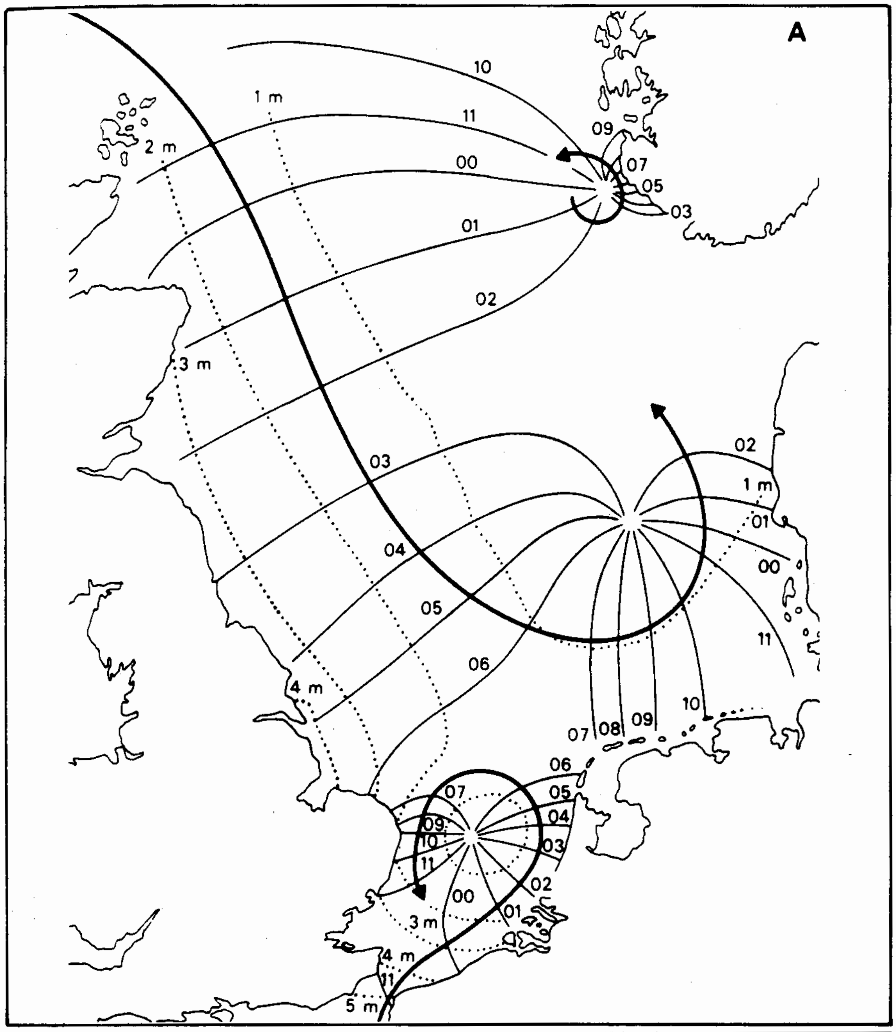

Figure 22
Amphidromic tidal system in the North Sea, showing amphidromic points of zero tidal range, and tidal co-range lines (dotted) showing increasing range from amphidromic points. Continuous co-tidal lines show times of high water in "lunar hours". From Johnson, 1978, after Houbolt, 1968.

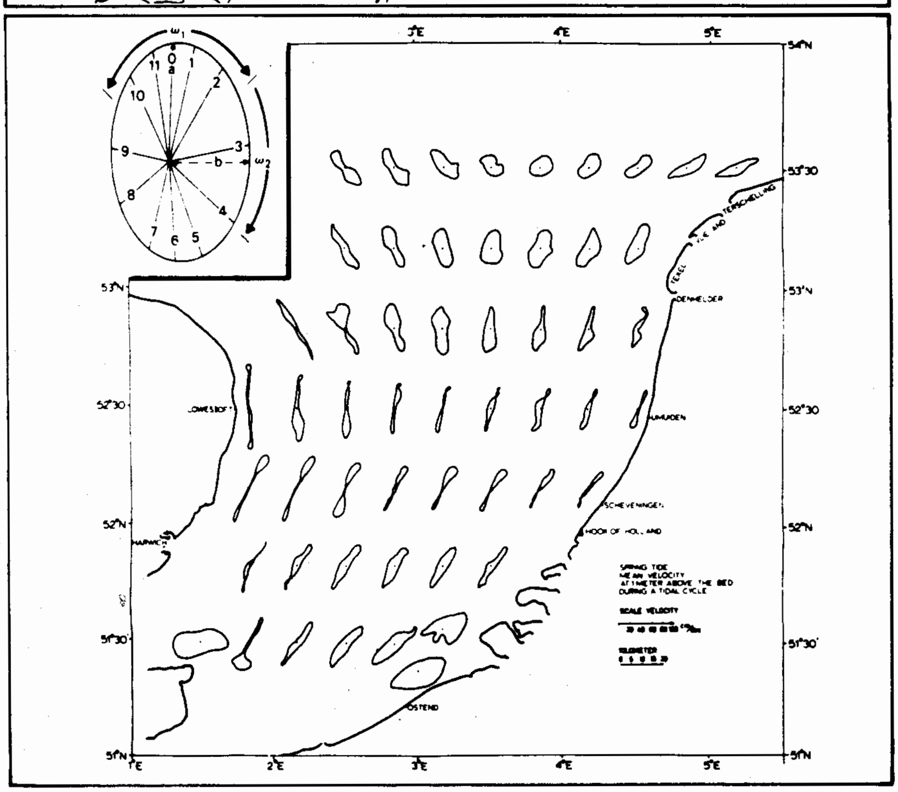

Figure 23
North Sea tidal current ellipses for computed velocities 1 m above bed. Note extreme elongation of ellipses due to flow constriction in southern North Sea, with northward and southward velocities of about 1 m/sec off Lowestoft. Inset – example of a tidal current ellipse with complex time-velocity asymmetry. Here, velocity a > velocity b and ω_1 (roughly northward flows) occupy 4.5 hours, and ω_2 (roughly eastward flows) occupy the other 1.5 hours of half the tidal cycle. From Johnson, 1978 and McCave, 1971 – original data source the Deutshces Hydrographisches Institut, 1963.

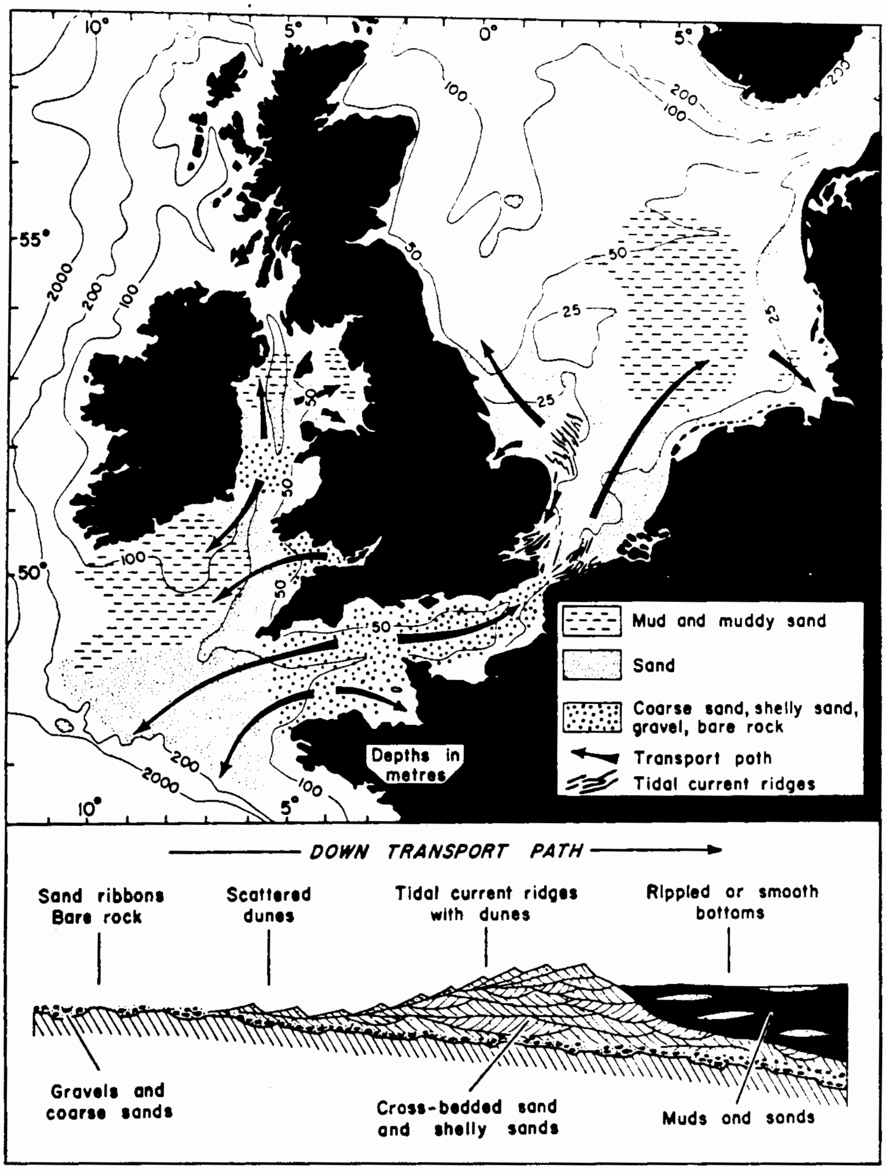

Figure 24
Dispersal of sand around the British coast, with sketch showing typical changes observed in a downcurrent direction, from bare rock and sand ribons, through scattered dunes into tidal current ridges and sandwave complexes. These grade finally into muds, with or without sand patches. From Allen, 1970, after Stride, 1963.

is a strong time-asymmetry of tidal flows (Allen, 1980; see Fig. 32 and discussion below). The internal structure of these low angle sand waves may therefore consist of sets of cross bedding perhaps up to a metre or so in thickness (a little less than the height of the megaripples), with bounding surfaces dipping downstream at 5 to 6° reflecting the dip of the "steep" face and the migration of megaripples down that face (Fig. 26).

Large tidal sand ridges up to 40 m high have been described by Houbolt (1968) off the Norfolk coast of Britain (Figs. 27 and 28). The ridges are "linear sand ridges" in the sense of Swift *et al.* (1973), and together they make up a shoal retreat massif (Swift, 1975, p. 128-9; Fig. 27). The area occupied by sand ridges is about 5000 km^2. There is no doubt that the ridges are presently being maintained by tidal flows (Houbolt, 1968), and they appear to have formed by shoreline detachment during the Holocene transgression (Swift, 1975, p. 126-128). Crests are in 10 to 20 m of water, and the ridges are about 5 km wide and up to 60 km long (Fig. 27). The "steep" side faces seaward, and dips at about 5 to 6° (Fig. 28). The internal reflectors were at one time mistakenly interpreted as angle-of-repose cross bedding, but presumably represent older positions of the "steep" face, implying a gradual seaward migration of the ridges. Houbolt (1968, p. 252) recognized that the sand "actually seems to go round the ridge" in what is now termed a "racetrack"pattern, and details of sediment transport and sand wave migration around the northern end of Haisborough Sand (Fig. 27) have been published by McCave and Langhorne (1982).

Sand Bodies on Georges Bank

Tidal sand ridges, sand waves, and megaripples are known from areas other than the North Sea. Nantucket Shoals (Mann *et al.*, 1981) and Georges Bank (Twichell, 1983) are good examples. On Georges Bank the sand ridges and sand waves cover an area of about 20000 km^2. They overlie the Holocene transgressive surface, and could have formed in the present tidal regime. Twichell (1983) describes the ridges as occurring in depths less than 60 m. They are 10 to 35 m high, 15 to 90 km long, and have wavelengths of 15 km. The "steep" (about 4°) side faces southwest. Superimposed sand waves also occur in depths less than 60 m; they are 1 to 15 m high with wavelengths of 50 to 100 m. Smaller megaripples have heights mostly less than 1 m, and wavelengths of 1 to 15 m.

In general, the sand ridges and sand waves line up with the long axes of the tidal current ellipses, indicating a tidal control. Sand wave migration "is oblique to the ridges and in opposite direction on the two sides of the ridges [indicating] that the ridges themselves may be active bed forms" (Twichell, 1983, p. 707). Although Twichell does not discuss lee face angles for the sand waves, measurements from his Figure 4A indicates "steep" slopes of about 1.4°. His Figure 4B indicates slopes of about 10°.

Stratification Produced by Tidal Sand Ridges and Sand Waves

The large tidal sand ridges and sand waves have not been cored systematically. The only studies that document the internal structure of these features are those of Reineck (1963; Fig. 29) and Houbolt (1968). From these two studies, and the work on sand wave and sand ridge morphology by McCave (1971) and Twichell (1983), we can argue that the internal structure might be characterized by low angle surfaces (the 5 to 6° "steep" faces) with smaller sets of cross bedding dipping down or up those low angle "master bedding surfaces" (Figs. 26, 29, 30, 31 and 32). Allen (1980) formalized this into a series of sand wave classes, suggesting that the controlling factors would be "(1) the tidal time-velocity pattern, (2) water depth, and (3) bed-material calibre" (Allen, 1980, p. 303). The six classes are shown in Figure 32. Classes 1 and 2 represent fluctuating unidirectional flows. Class 3 has a significant still-stand period between critical ebb and flood velocities for sand movement U_{CR}, during which time the foreset may begin to be bioturbated (Fig. 33). Class 4 shows major reactivation surfaces (E_2 denotes second order bedding surfaces), and classes 5 and 6 show almost symmetrical ebb and flood currents.

With Allen's model in mind, and recognizing its basis in theory and in observation of modern sand waves, we may proceed to look at the geological record.

TIDE-DOMINATED SHELVES - ANCIENT SEDIMENTS

Although there are several good examples of ancient tidal flat and tidal channel environments (see Ginsburg, 1975, and Reinson, this volume), there are remarkably few examples of open shelf tidal deposits. Of the latter, there are perhaps two categories - first, the very thick (100s or 1000s of m) cross bedded quartzites typical of the late Precambrian and Early Cambrian craton margins, and second, thin transgressive sandstones (a few 10s of metres). In some cases the cross bedding is abundant and/or ubiquitous but flow directions may be almost unidirectional, making a specifically tidal interpretation hard to establish. Many illustrations are given in a recent review by Walker (1985c).

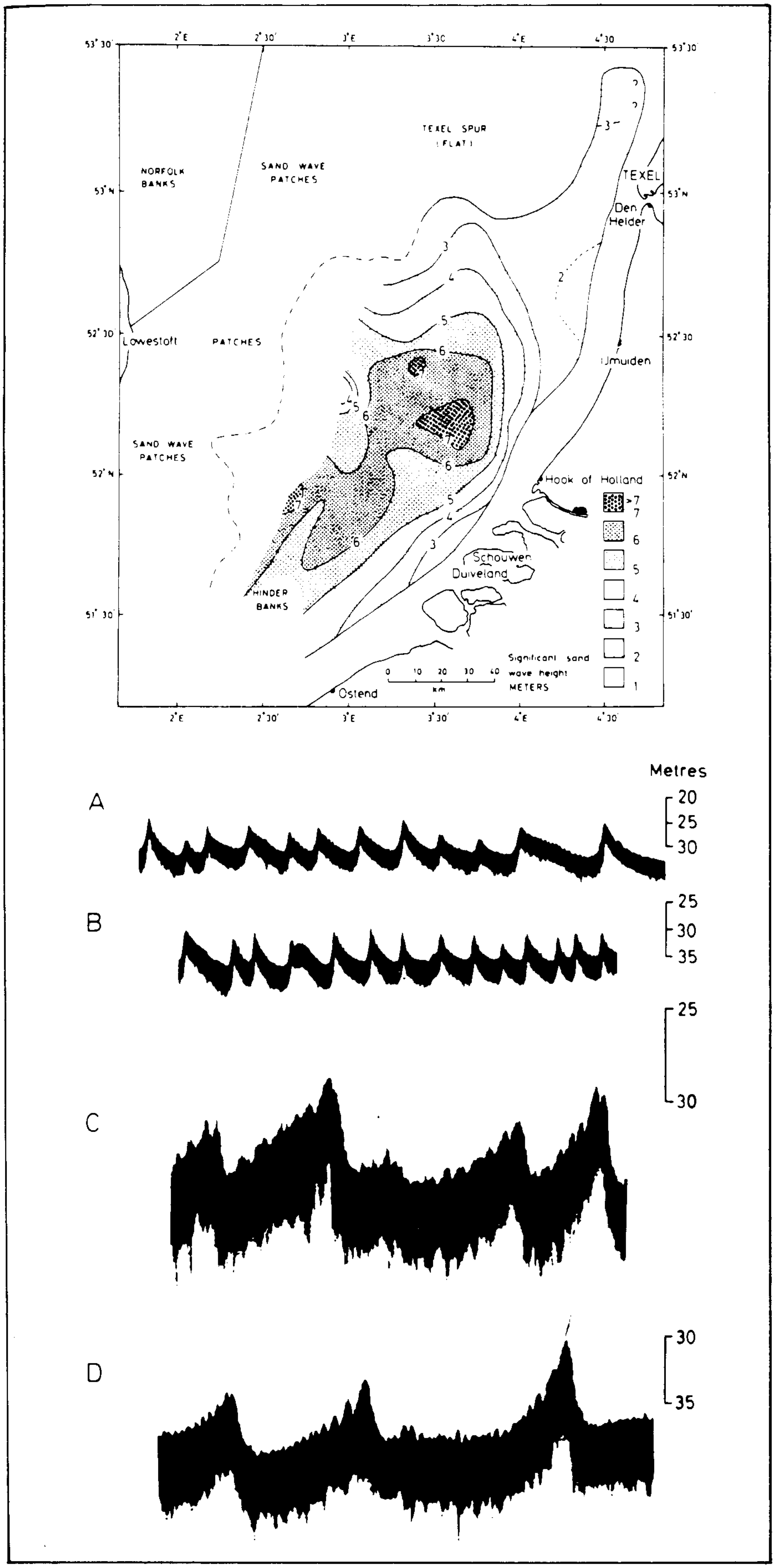

Figure 25

Sand wave heights, averaged over the period Aug. 1968 to May 1969. The echo sounder profiles show sand wave topography. Lengths of lines are: A) 3800 m; B) 2800 m; C) 900 m; and D) 1200 m. The vertical scale is grossly exaggerated, and calculated dips of lee faces from these data give 5 to 6°. From McCave, 1971.

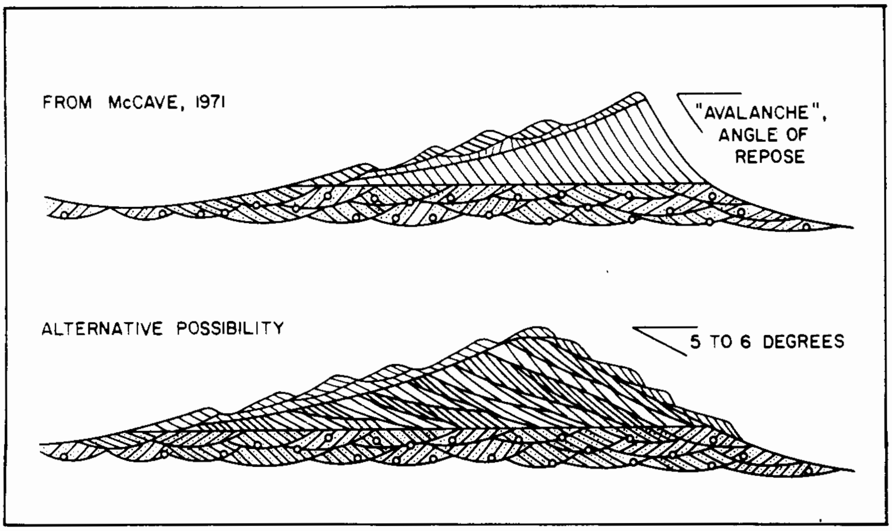

Figure 26
The upper model, from McCave 1971, shows the inferred internal geometry of the sand waves. However, it must be emphasized that most of the sand waves have lee faces in the 5 to 6° range (Fig. 25), and angle-of-repose cross bedding would not *be predicted. This diagram has been reproduced by Brenner (1980), Stride* et al., *(1982) and Galloway and Hobday (1983), with no comment concerning the style of cross bedding that would result when the exaggerated echo sounder profiles are reduced to real scale. I have made a guess (lower diagram), which shows megaripples moving down the lee (5 to 6°) side, forming compound cross stratification similar to that of Figures 31 and 32.*

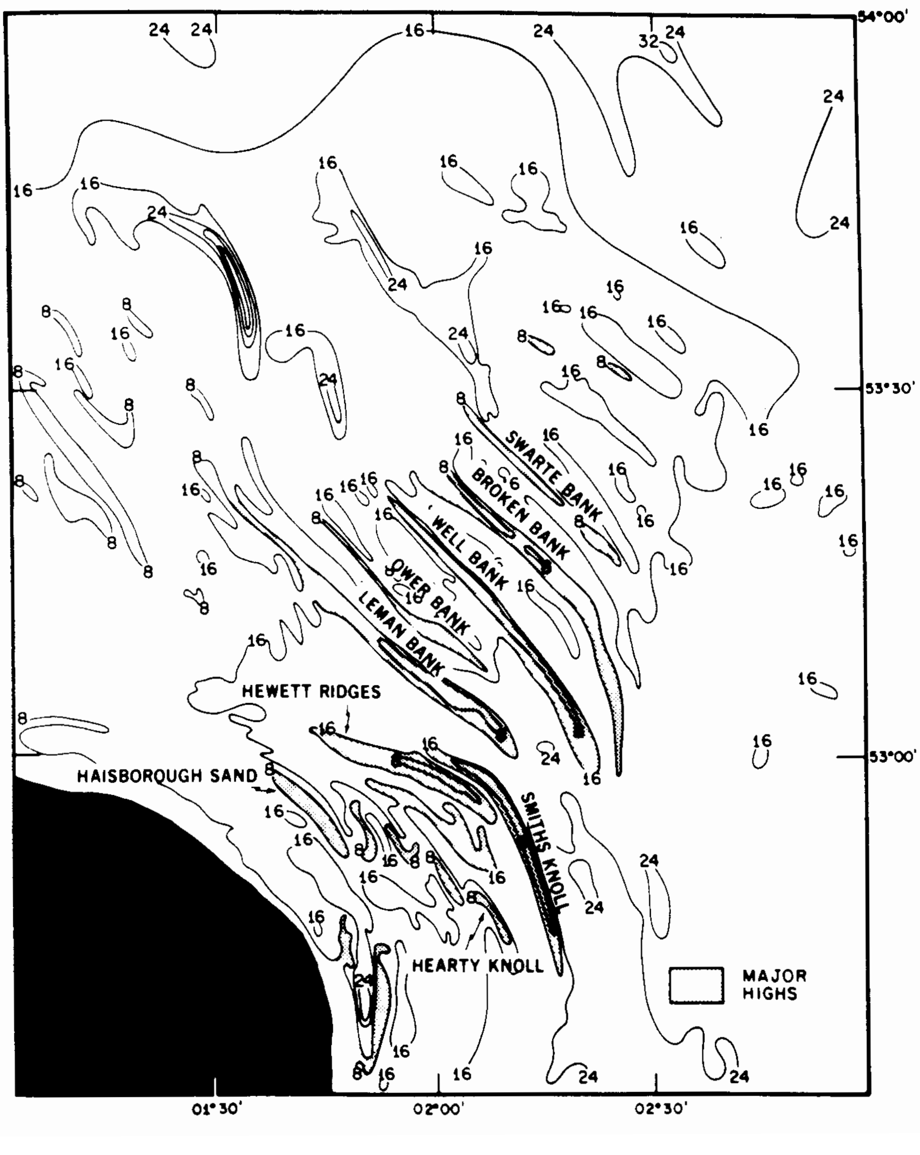

Figure 27
Map of sand ridges in the North Sea, from Swift (1975) after Houbolt (1968). The sparker profiles show that the ridges are up to 35-40 m high, and their crests are in about 10 to 15 m of water. "Steep" sides face northeast, but dip at only about 5° (see Fig. 28).

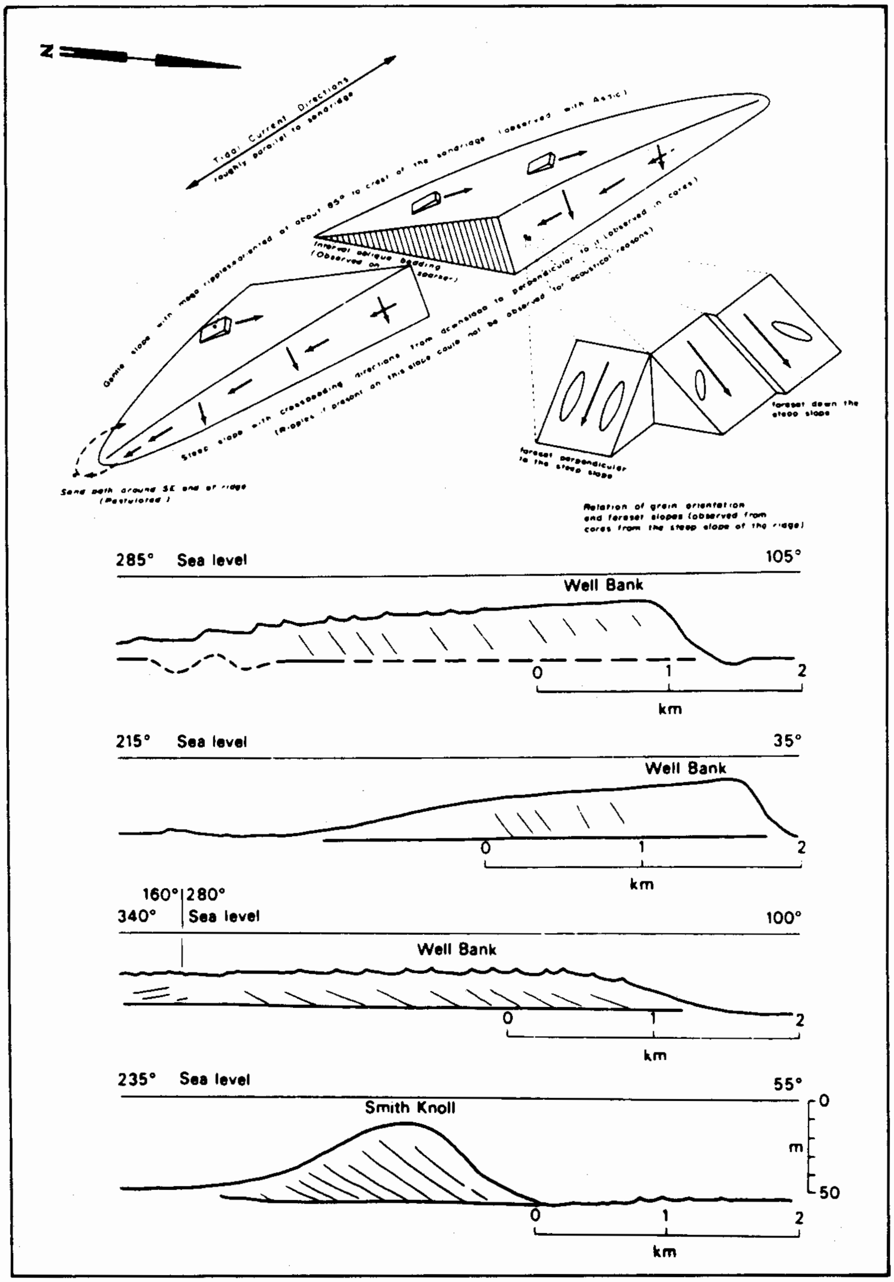

Figure 28
North Sea sand ridge as sketched by Houbolt (1968). Lower diagram shows sparker profiles of Well Bank (see Fig. 27), with great vertical exaggeration. Dips of the internal reflectors and the "steep" faces average about 5 to 6°.

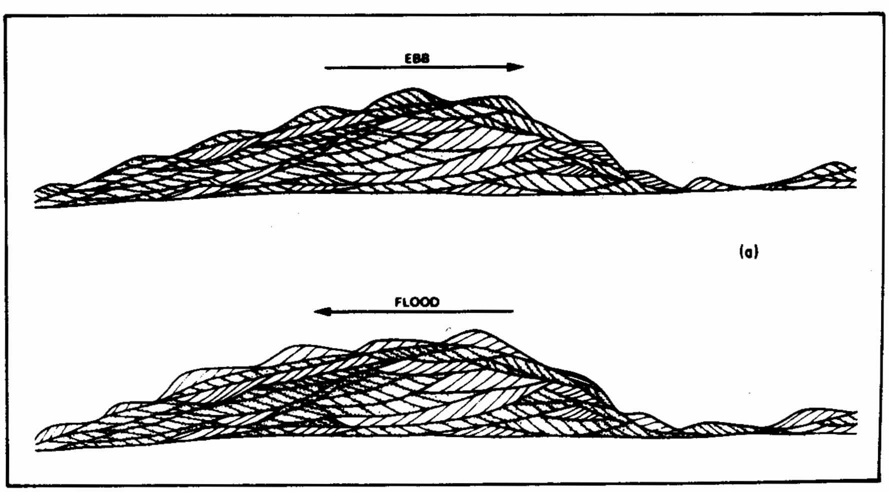

Figure 29
Internal structure of giant ripples from the outer Jade, German Bight, southern North Sea. Note "steep" face is on lee side; diagram has 10 x vertical exaggeration. In the outer Jade, the sand wave heights vary from 1.7 to 5.5 m. The internal structure is inferred from box coring studies of the uppermost parts of the sand waves. After Reineck, 1963.

Figure 30
Large sandwave in the Lower Greensand (L. Cretaceous) near Leighton Buzzard, England. Main set is about 5 m thick, but is truncated by thinner, horizontally-bounded sets at the top.

Figure 31
Compound cross bedding in the Lower Greensand (L. Cretaceous) at Leighton Buzzard, England. Regional bedding is horizontal – note the 5° dipping boundary surfaces of the cross bed sets. Bob Dalrymple is attempting to cope with field conditions.

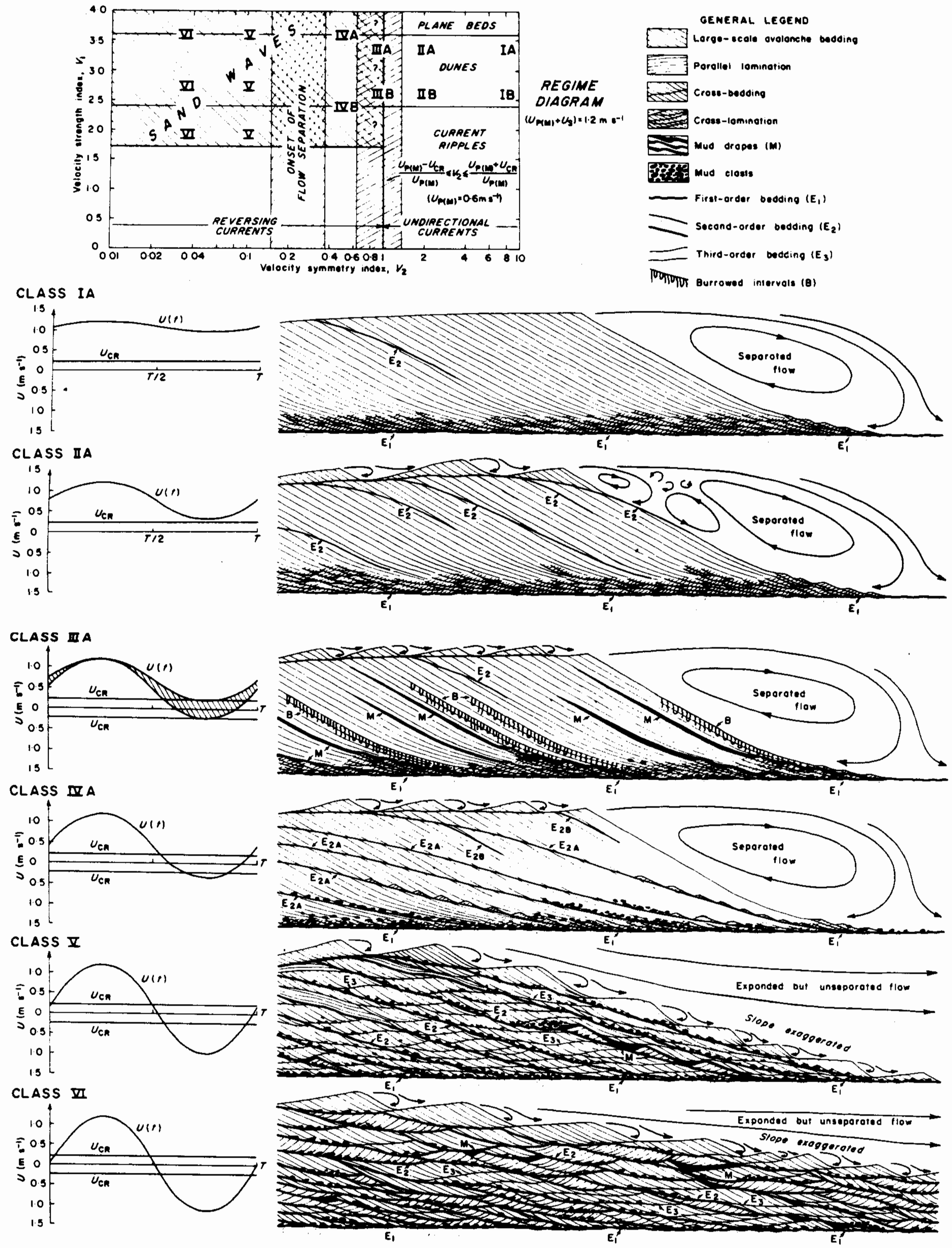

Figure 32
Regime diagram, and predicted categories of sand wave and dune internal structure. Note that the six classes are defined by the time/velocity asymmetry (or symmetry). Classes I to IV have large foresets (commonly with reactiviation surfaces, E_2) and flow separation, but classes V and VI have much lower slopes on the "steep" side, and no flow separation. From Allen, 1980.

Figure 33
Detail of large foresets shown in Figure 30. Note extensive mottling (bioturbation) of the darker foresets, and compare this figure and Figure 30 with Allen's class III in Figure 32.

Precambrian - Lower Cambrian Quartzites
The most detailed example of a thick Precambrian tidal quartzite is that of Anderton (1976) from the Jura Quartzite (Dalradian) of Scotland. The unit is about 5 km thick, and can be divided into a coarse facies and a fine facies. The coarse facies consists of various cross bedded sandstones, with sets from a few cm to 4.5 m thick. Locally, bimodal-bipolar paleoflow directions were observed, leading to an interpretation of the coarse cross bedded facies as various tidal sand wave deposits. The fine facies consist of interbedded mudstones and sharp-based sandstones with parallel lamination and ripple cross lamination. Beds over 10 cm thick. "show a sequence of structures similar to that found in some turbidites" (Anderton, 1976, p. 441 - the structure sequences referred to in Anderton's Fig. 12 are ABCE, BCE and CE in Bouma's (1962) terminology). However, Anderton (1976, p. 445) notes that the facies " could have been deposited by density currents", and later suggests "tidally-dispersed, storm generated suspension clouds". The possibility of finding turbidites associated with shelf deposits has been discussed earlier in this paper. One important point to emerge from Anderton's paper is that the various facies "alternate in vertical succession in an apparently random manner" (p. 447).

The Gog Quartzite (Lower Cambrian) of the Southern Rocky Mountains of B.C. and Alberta is up to about 3 km thick, and in the section along the Trans-Canada Highway in the Kicking Horse Pass (Hein, 1982), there is abundant cross bedding in sets up to almost 1 m thick. Paleocurrent patterns are complex, with some bimodal patterns suggesting tidally-influenced dune and bar systems. However, as with the Jura Quartzite, Hein has documented many other facies (some of which suggest a storm influence). It would appear that in general the thick quartzites have many depositional influences. Finally, Hein's (1982, p. 129-130) measured sections show no evidence of cyclicity or preferred facies sequence.

In another late Precambrian example from north Norway, Levell (1980) has documented the occurrence of simple and compound sets of cross bedding. The compound sets are made up of downcurrent-dipping bounding surfaces, with angle-of-repose cross bedding between the bounding surfaces. Levell (1980, p. 545) comments that "the compound cross bedding, in which each small-scale set is separated from its neighbours by a convex-upward surface could represent either a large bedform with megaripples superimposed on its lee face ... or a large bedform with extremely closely spaced, periodic, reactivation surfaces formed by reworking of a single angle-of-repose lee face". Levell's sketches and photographs closely resemble the model of Allen (Fig. 32, classes IIA and IVA). However, Levell notes that paleoflow directions are consistently eastward, and within his preferred tidal interpretation, suggests, 1) a regional transport dominance (as in the southern North Sea), 2) a local transport dominance (mutually evasive ebb and flood flows), and 3) a preservation factor that favours the lateral migration of large subtidal sand bodies.

Transgressive Sandwave Complexes
The general occurrence of sandwave complexes in transgressive situations was noted by Nio (1976) and Nio and Siegenthaler (1978), quoting particularly the Roda complex (lower Eocene, Spain) and the Lower Greensand (Lower Cretaceous, southern England). In the Roda, there are angle-of-repose cross bed sets up to 20 m thick; Nio's present interpretation is that these sandwaves are estuarine, and hence I can no longer use this example in a discussion of tidal shelf deposits. The Lower Greensand has been described breifly by Narayan (1971), de Raaf and Boersma (1971) and Bridges (1982). It is in an overall transgressive situation, and contains individual cross bed sets up to 4 or 5 m thick. There is evidence that smaller bedforms migrated both up and down low angle (1 to 5°) dipping surfaces (Figs. 30 and 31), in a manner illustrative of Allen's (1980) model (Fig. 32, classes IIA and IIIA). Paleoflow directions are essentially unidirectional, to the south in the Woburn area (Schwarzacher, 1953) and to the southeast in the Weald (Narayan, 1971; Bridges, 1982, p. 183-7). In commenting on earlier work, Bridges only notes that "a tidal interpretation is indeed attractive", but there are apparently no definitive criteria.

TIDE-DOMINATED SHELVES - SYNTHESIS AND MODELS
Ancient and recent examples suggest that tide-dominated shelves are characterized by sand waves and sand ridges. Most of these tend to have asymmetrical profiles, commonly with "steep" faces inclined at 10° or less. Consequently, the internal structure is characterized by gently dipping master bedding surfaces (commonly reactivation surfaces), with angle-of-repose cross bedding formed by migration of megaripples on the master bedding surfaces. The range of

possible structures is summarized well by Allen's model (Fig. 32).

The relatively thin transgressive sandwave complexes pose no major problems of sediment supply - it is essentially sediment reworked during transgression, as in the modern North Sea. The very thick late Precambrian/Cambrian quartzites pose two problems - first, how was the sediment transported from the shoreline out onto the shelf, and second, how did supply and subsidence remain so closely matched. If supply had lagged, we might expect deepening and mud deposition. If supply had exceeded subsidence we might expect shallowing and shoreline progradation. However, shorelines have rarely been recognized in these thick quartzites.

Finally, tidal shelf examples in the geological record tend to show dominantly unimodal paleoflow directions, as opposed to the common bimodal-bipolar patterns of very nearshore (barrier-associated) deposits. Levell (1980) addresses this problem, suggesting either regional transport dominance (southern North Sea), local transport dominance, or preservation over the long term *only* of master bedding surfaces inclined in one direction.

LONG, NARROW SHALLOW MARINE SAND BODIES

These are common in the Jurassic and Cretaceous of the Western Interior Seaway (Fig. 34). Details are given in Table 3, but it should be emphasized that linear bodies of several distinct origins may be tabulated here, and "average" dimensions should be interpreted with care. One point that they all have in common is a progressively coarsening-upward sequence, with marine mudstones passing up into bioturbated siltstones and thence into various types of cross bedded or hummocky cross stratified sandstones. Sand body orientation tends to be sub-parallel or somewhat oblique to regional shorelines, and where there is evidence of contemporaneous shorelines, the sand bodies appear to have formed many tens of km offshore (Table 3).

The Cardium examples in Table 3 probably do not belong - it has been shown that Garrington and Caroline are long, narrow zones of production within *sheet* sandstones (Walker, 1983a), and that Ricinus is probably a channel cut and filled by turbidity currents (Walker, 1985). However, average dimensions change only slightly if these three examples are deleted.

It will be noted in Table 3 that there are several "kitchen sink" interpretations - loose combinations of storms, tidal currents and oceanic circulations. However, it is also clear that almost all authors envisage storms as a major sand transporting process, recognizing that the sands in the lower part of the coarsening-upward sequences have to be transported long distances offshore into muddy environments. The fact of gradational bases for the sand bodies is a major reason why few of the authors strongly favour tidal currents, because the modern linear sand ridges and sand waves both rest at least in part unconformably on transgressive surfaces with gravel lags. It is commonly not stated whether sand transport is incremental or from relaxation (storm surge ebb) or turbidity currents, but in many cases it appears that the authors envisage a somewhat steady movement of sand parallel to isobaths, with special events sculpting this sand into distinct ridges. Thus there is some apparent opinion in favour of incremental dispersal parallel to isobaths, probably by storm-generated geostrophic flows (Fig. 34).

The suggestions of tidal modification are based on the presence of cross bedding in the upper parts of sand

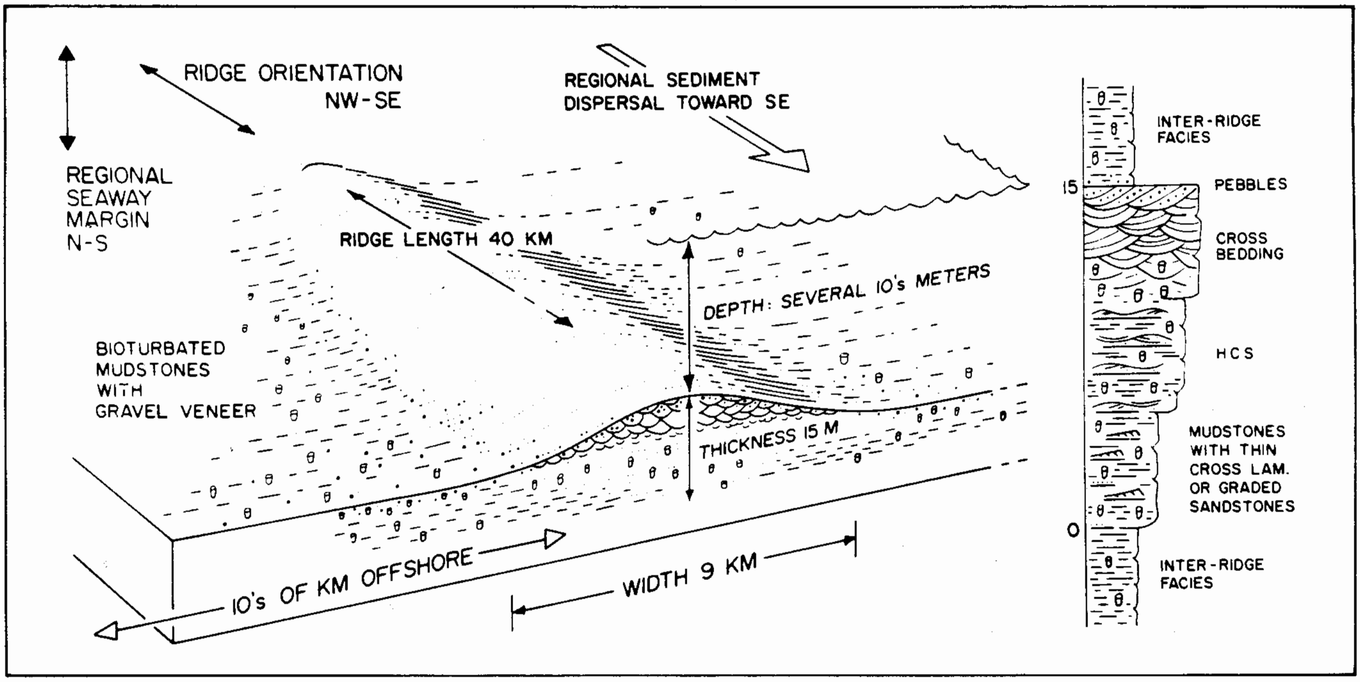

Figure 34
Diagrammtic summary of the main features of linear sand bars from the Cretaceous Western Interior Seaway. Data from Table 3.

Table 3

Dimensions of shallow marine sand bodies, western North America

Formation	Thickness of sand body (m)	Length (km)	Width (km)	Author	Interpretation, with (page references). Distance offshore quoted where possible.
Shannon,					
Upper ss.	15	50	30	Spearing, 1976	"storm system, superimposed on oceanic or tidal currents" (70). 100 km offshore.
Lower ss.	21	100	50		
Shannon	20	17.6	2.4	Seeling, 1978	"ancient hydraulic environment . . . was analogous in some important respects to those present day environments with prominent currents off the east coast of the United States and in the southern part of the North Sea" (133). 120 km offshore.
Sussex	12	40	1.6	Berg, 1975	"storm surge, density currents or other phenomena" (2109). 200 km from shoreline.
Sussex	c.30	c.50	up to 10	Brenner, 1978	"influenced by tidal and storm generated currents . . . neap and spring tides . . . intense storms" (195-7).
Sussex	12.2	45	1.6	Hobson *et al.*, 1982	"storm related and other marine currents" (697). > 160 km from shoreline.
Oxfordian ss.	c.15	>5	0.2 to 2	Brenner and Davies, 1974	"storms played a major role" (425). "Storm driven currents, normal tidal currents and regional circulation currents" (427).
Duffy Mountain	27	50+	8 - 16	Boyles and Scott, 1982	"fairweather shelf currents formed sand waves along the bar crest . . . transported sand . . . onto the back bar". "Interaction of fairweather and storm sediment transport" (505). At least 16 km offshore.
Semilla ss.					
Holy Ghost bar	21	20+	15	La Fon, 1981	"tidal origin doubtful" (720). "Sandstone bars were deposited by periodic storm events" (720), "storms, possibly augmented by weak tidal currents" (720).
Bernalillito Arroyo bar	12	?20	?8		
Frontier	13	10.4 - 12	3.2	Winn *et al.*, 1983	"storm generated shelf sand ridges . . . tides and permanent marine currents apparently were not important in transporting sand" (41).
Cardium					
Crossfield	6	96	1.6-2.8	Berven, 1966	"offshore bars . . . waves and nearshore currents" (208).
Garrington	c.5	71	6	Walker, 1983a, b	sheet sand emplaced by turbidity currents but modified by storms (Walker, 1983a). At least 100, probably 140-200 km offshore.
Caroline	c.5	48	5.4		
Ricinus	18	45	5	Walker, 1983a, b, 1985	turbidites in channel, channel probably cut by turbidity currents (see Walker, 1985). At least 100, probably 140-200 km offshore.
Carrot Creek	13.4	c.7	c.2.5	Swagor *et al.*, 1976	"pebbles were driven dominantly by storms across a shallow shelf" (94).
Viking					
Dodsland-Hoosier	8	up to 80	c.10	Evans, 1970	"Offshore environment affected by tidal currents" (484).
Gilby	c.2.5	30	3.3	Koldijk, 1976	"cannot be explained by normal marine conditions (76) . . . ephemeral currents, e.g. during severe storms" (77).
Joffre	9.5	40	1.6-3	Reinson *et al.*, 1983	"frequent storm events . . . tectonically induced strong density currents. Strong bottom traction currents (tidal current streams) redistributed the detritus" (104).
AVERAGE	14.0	43.5	9.0		
AVERAGE	14.9	41.4	9.7	(without Garrington, Caroline and Ricinus)	

Note: added in second printing, January 1986. It has now been shown that many of the Cardium sand bodies *do* rest on erosion surfaces. This places less emphasis on problems of sediment transport, but enhances problems of sea level change. The reader is referred to Plint *et al.*, (1986) and Bergman and Walker (1986).

bodies, which in places has apparently bimodal directions (Spearing, 1976, p. 76.) However, it is clear from the work of Swift and colleagues discussed earlier (Swift *et al.*, 1979) that megaripples up to about 1 m high can form as a response to storm flows, and can migrate and produce cross bedding. Thus the hard evidence for tidal flows is lacking. Seeling (1978) compares the Shannon morphologically with linear sand ridges of both the Atlantic Shelf (storm dominated) and the North Sea (tidally dominated), but does not preferentially suggest a tidal interpretation. In those examples where there is paleoflow evidence (Spearing, 1976; Brenner, 1978; Hobson *et al.*, 1982; Brenner and Davies, 1974; Boyles and Scott, 1982; Walker, 1983), dispersal is consistently south or southeastward. The only exception is apparently the Semilla Sandstone, where La Fon (1981, p. 719) reports trough cross beds that "generally dip northwest, paralleling the direction of elongation of the bar, although some sets dip southeast". Northwest and westerly flows, without supporting data, are also shown in La Fon's Figure 3.

In most of the examples in Table 3, there is little or no evidence of any relationship between the linear ridges and regional transgression. The major exception is the Sussex example in the interpretation of Hobson *et al.* (1982). Thus it is not possible to compare the ancient ridges with Swift's shoreface-attached ridges that become detached during transgression. Not only is the context wrong with respect to transgression; the ancient linear ridges coarsen upward from offshore marine mudstone, which would *not* be predicted for Swift's shoreface-attached ridges. The major problem, therefore, is not so much how the sand disperses southward down the seaway, but how these processes (incremental movement of sand by storms?) concentrate the sand into isolated bars and ridges that coarsen upward. If the flows were oblique to the crests of the linear ridges, Huthnance's (1982) stability model might apply (Fig. 8). Rate of ridge growth would then be a function of ridge/current obliquity, mean depth, a drag coefficient and the ridge spacing (see Figueiredo *et al.*, 1981, p. 188).

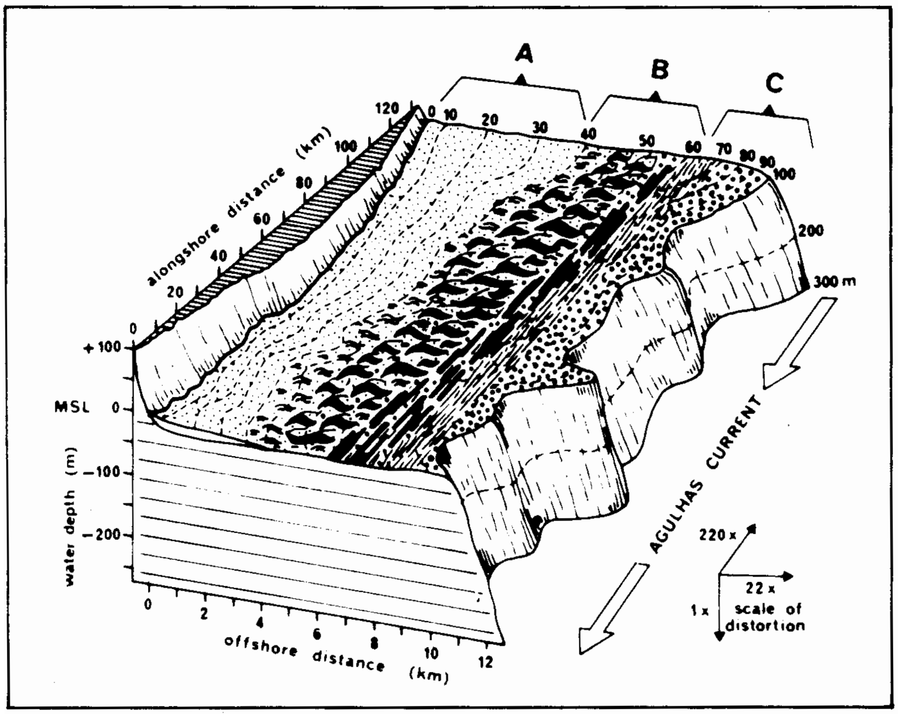

Figure 35.
Sandwaves, as observed on side scan sonar, on the shelf edge off the southeastern tip of Africa. The sand waves are driven by the Agulhas Current, which is a major Indian Ocean circulation that spills up onto the shelf. Stipple indicates coarse lag, black streaks indicate sand ribbons. Individual sand wave fields are up to 20 km long and 10 km wide. Sand waves are up to 17 m high, with angle-of-repose (25°) lee faces. From Flemming, 1980.

SHELVES DOMINATED BY INTRUDING OCEAN CURRENTS

This type makes up only about 3% of modern shelves, and the best example is that of Flemming (1978, 1980) from the southeastern shelf of South Africa (Fig. 35). Here, the intruding ocean current is the Agulhas Current of the Indian Ocean. The sand waves made by this current occur in 40 to 60 m of water, some 4 to 8 km offshore. They have been observed on side-scan sonar, and occur in continuous fields up to 20 km long and 10 km wide. Heights are variable, but the maximum is about 17 m. The largest sand waves have wavelengths of about 700 m and lee faces in excess of 25° (Flemming, pers. commun., 1979). Hence these sand waves might be expected to form angle-of-repose cross bedding in sets many metres thick.

The Agulhas Current is not contributing new sediment to the shelf, and hence the coarse lags (heavy stipple, Fig. 35) sand ribbons and sand waves (Fig. 35) represent the reworking of older sediment. It is unlikely in this situation (or in other situations where transgressed older sediments are reworked by intruding ocean currents) that a stratigraphic thickness of more than a few metres will build up, and this situation is probably rare in the geological record. I know of no ancient example that has been convincingly interpreted by this mechanism, although it may be a possibility for some transgressive sand wave complexes with consistent unidirectional flow (as described above).

CONCLUSIONS

The shelf is one of the most complex depositional environments because of the interaction of three different major processes - ocean currents, storm-generated currents and tidal currents. In some cases, modern and ancient studies converge toward the beginnings of a coherent model (as for tidal sand bodies). In other cases, there seem to be no modern equivalent of deposits seen in the geological record, particularly the gradationally-based linear sand ridges of the Western Interior Seaway. In yet other cases, the geological record suggests processes which have not been observed on modern shelves; here, I am referring particularly to the interbedded HCS sandstone/bioturbated mudstone facies with evidence that some of the beds were emplaced by turbidity currents.

On the other hand, there are well defined modern processes and deposits that have yet to be firmly identified in the geological record. For example there are no specific geological interpretations of incremental sand movement by storm-generated geostrophic flows (although this may be one way in which the linear sand ridges of the Western Interior Seaway were formed). Also, there are no examples of transgressive, shoreface-detached linear sand ridges similar to those described by Swift and colleagues.

These problems contribute to the present lively debate concerning shelf sands and sandstones, but really powerful models will not emerge until there is a closer synthesis of ancient and modern examples.

ACKNOWLEDGEMENTS

I am indebted to Don Swift for his careful and detailed review of this manuscript. Many of the ideas herein are new, and in a state of flux, and there is still considerable disagreement on interpretive matters. In marshalling a big body of literature into simplified facies models, any errors of interpretation which have crept in are mine. I also thank Bill Duke and Don Keith for their comments. The work has been supported by Strategic and Operating Grants from the Natural Sciences and Engineering Research Council of Canada.

REFERENCES

The references are presented in alphabetical order. For the newcomer to the subject, the most useful general reviews are in the SEPM Short Course Notes by Tillman, Swift and Walker 1985). At a more technical level, the compilation of papers edited by Tillman and Siemers (1984) is a very useful source of ideas and additional references. Most of the citations to hummocky cross stratification are given by Duke (1985), and citations to tidal processes and deposits are given by Allen (1980).

Aigner, T., 1982. Calcareous tempestites: storm-dominated stratification in Upper Muschelkalk limestones (Middle Trias, S.W. Germany). *In* Einsele, G. and Seilacher, A., eds., Cyclic and event stratification. New York, Springer, p. 180-198.

Allen, J.R.L., 1970. Physical processes of sedimentation. New York, American Elsevier, 248 p.

Allen, J.R.L., 1980. Sand waves: a model of origin and internal structures. Sedimentary Geology, v. 26, p. 281-328.
A documentation of field examples, and development of a theoretical model for sand waves. The model for stratification is based largely on the symmetry of flow (or lack of it) over the sand waves.

Anderton, R., 1976. Tidal shelf sedimentation: an example from the Scottish Dalradian. Sedimentology, v. 23, p. 429-458.
One of the most detailed descriptions and interpretations of tidal facies in a very thick Precambrian quartzite.

Ball, S.M., 1971. The Westphalia Limestone of the Northern Midcontinent: a possible ancient storm deposit. Journal of Sedimentary Petrology, v. 41, p. 217-232.
Description of transported foraminifera, with interesting calculations of volume of material moved. One of the earliest modern storm interpretations.

Bea, R.G., and Audibert, J.M.E., 1980. Offshore platforms and pipelines in Mississippi River delta. American Society of Civil Engineers, Proceedings; Journal of the Geotechnical Engineering Division, v. 106 (G.T. 8), p. 853-869.
A review of the effect of substrate conditions and substrate disturbance on platform and pipeline structures.

Berg, R.R., 1975. Depositional environment of Upper Cretaceous Sussex Sandstone, House Creek Field, Wyoming. American Association of Petroleum Geologists, Bulletin, v. 59, p. 2099-2110.

Bergman, K.M. and Walker, R.G., 1986. Cardium Formation 9. Conglomerates at Carrot Creek Field: offshore linear ridges or shoreface deposits. *In* Moslow, T.F. and Rhodes, E.G., eds., Modern and Ancient shelf clastics. Society of Economic Paleontologists and Mineralogists, Core Workshop, Atlanta, 1986.

Berven, R.J., 1966. Cardium sandstone bodies, Crossfield-Garrington area, Alberta. Bulletin of Canadian Petroluem Geology, v. 14, p. 208-240.

Bouma, A.H., 1962. Sedimentology of some flysch deposits. Amsterdam, Elsevier, 168 p.
The famous "Bouma sequence" for turbidites is established in this book.

Bourgeois, J., 1980. A transgressive shelf sequence exhibiting hummocky stratification: the Cape Sebastian Sandstone (Upper Cretaceous), southwestern Oregon. Journal of Sedimentary Petrology, v. 50, p. 681-702.
This is probably the best documented case of HCS in an overall *transgressive* situation.

Boyles, J.M., and Scott, A.D., 1982. A model for migrating shelf-bar sandstones in Upper Mancos Shale (Campanian), northwestern Colorado. American Association of Petroleum Geologists, Bulletin, v. 66, p. 491-508.
Elongate bars; sand source believed to be trangressed littoral deposits. Bars migrated by fairweather and storm processes.

Brenchley, P.J., Newall, G., and Stanistreet, I.G., 1979. A storm surge origin for sandstone beds in an epicontinental platform sequence, Ordovician, Norway. Sedimentary Geology, v. 22, p. 185-217.

Brenner, R.L., 1978. Sussex Sandstone of Wyoming - example of Cretaceous offshore sedimentation. American Association of Petroleum Geologists, Bulletin, v. 62, p. 181-200.
Outcrop and subsurface study which suggests ridge formation by storm and tidal currents, with ridge breaching by storms, the channels thus formed being subsequently used by tidal currents to form tidal deltas.

Brenner, R.L., 1980. Construction of process-response models for ancient epicontinental seaway depositional systems using partial analogs. American Association of Petroleum Geologists, Bulletin, v. 64, p. 1223-1244.
A comparison of modern shelves with ancient epicontinental situations.

Brenner, R. L., and Davies, D.K., 1974. Oxfordan sedimentation in Western Interior United States. American Association of Petroleum Geologists, Bulletin, v. 58, p. 407-428.
Outcrop study of various marine coarsening-upward sequences. Marine bars believed to result from interplay of storms, tidal currents and regional circulation.

Bridges, P.H., 1982. Ancient offshore tidal deposits. *In* Stride, A.H., ed., Offshore tidal sands. London, Chapman and Hall, p. 172-192.
The most recent review of offshore tidal sands. Emphasizes the Lower Greensand and Western Interior Seaway (quoting several coarsening-upward sequences which may not be tidally dominated).

Bullock, A., 1981. Sedimentation of the Wapiabi-Belly River transition (Upper Cretaceous) at Lundbreck Falls, Alberta. B.Sc. Thesis, McMaster University, Hamilton, Canada, 94 p.
Documentation of HCS sandstones overlying turbidites, both facies having identical northward-directed paleoflows.

de Raaf, J.F.M., and Boersma, J.R., 1971. Tidal deposits and their sedimentary structures. Geologie en Mijnbouw, v. 50, p. 479-504.
Briefly discusses seven examples, including the Lower Greensand. Good photos.

Dott, R.H. Jr., and Bourgeois, J., 1982. Hummocky stratification: significance of its variable bedding sequences. Bulletin of the Geological Society of America, v. 93, p. 663-680.

Duke, W.L., 1985. Hummocky cross stratification, tropical hurricanes and intense winter storms. Sedimentology, v. 32, p. 167-194.
A documentation of 105 occurrences of HCS, with interpretation of the origin of HCS from hurricanes or intense winter storms

Emery, K.O., 1968. Relict sediments on continental shelves of world. American Association of Petroleum Geologists, Bulletin, v. 52, p. 445-464.
This is the most recent statement of the "relict" idea - see Swift, Stanley and Curray, 1971 for modifications of relict sediments.

Evans, W.E., 1970. Imbricate linear sandstone bodies in Doosland-Hoosier area of Southwestern Saskatchewan, Canada. American Association of Petroleum Geologists, Bulletin, v. 54. p. 469-486.

Figueiredo, A.G., Swift, D.J.P., Stubblefield, W.L., and Clark, T.L., 1981. Sand ridges on the inner Atlantic Shelf of North America: morphometric comparisons with the Huthnance stability model. Geo-Marine Letters, v. 1, p. 187-191.
Points out the possibility that Huthnance's model for tidal current ridges might be modified to apply to storm maintained ridges.

Flemming, B.W., 1978. Underwater sand dunes along the southeast African continental margin - observations and implications. Marine Geology, v. 26, p. 177-198.
Side scan sonar pictures of sand waves, now being driven by the Agulhas current spilling up onto the Continental Shelf. This, along with Flemming (1980) is the best study of ocean current spillup on a shelf.

Flemming, B.W., 1980. Sand transport and bedform patterns on the Continental Shelf between Durban and Port Elizabeth (southeast Africa continental margin). Sedimentary Geology, v. 26, p. 179-205.
Follow up on Flemming (1978).

Forristall, G.Z., Hamilton, R.C., and Cardone, V.J., 1977. *Continental Shelf currents in* Tropical Storm Delia: observations and theory. Journal of Physical Oceanography, v. 87, p. 532-546.
Presentation of flow data resulting from a major Tropical Storm.

Fox, W.T., 1983. At the sea's edge,. Prentice Hall, Englewood Cliffs, N.J., 317 p.
Chapter 4, Tides (p. 93-124) is an excellent, non-technical discussion of tides, tide generating forces, and tidal currents.

Galloway, W.E., and Hobday, D.K., 1983. Terrigenous clastic depositional systems. New York, Springer-Verlag, 423 p.
Chapter 7, p. 143-165, gives an overview of terrigenous shelf systems.

Ginsburg, R.N., ed., 1975. Tidal deposits. New York, Springer-Verlag, 428 p.
A compilation of examples of modern and ancient clastic and carbonate tidal facies. The clastic examples (p. 5-195) consist of a series of short papers written in a standard format - a very useful standardized compilation of data.

Goldring, R., and Bridges, P., 1973. Sublittoral sheet sandstoes. Journal of Sedimentary Petrology, v. 43, p. 736-747.
Excellent description of what is now termed hummocky cross stratification, with comments on its facies associations and processes of formation. Still a very useful paper; originally it was some 5-6 years ahead of its time.

Goldring, R., and Langenstrassen, F., 1979. Open shelf and nearshore clastic facies in the Devonian. Special Papers in Paleontology, v. 23, p. 81-97.

Hamblin, A.P., and Walker, R.G., 1979. Storm dominated shallow marine deposits: the Fernie-Kootenay (Jurassic) transition, southern Rocky Mountains. Canadian Journal of Earth Sciences, v. 16 p. 1673-1690.
One of the first Journal papers to use HCS in the interpretation of an ancient sandstone. Established same paleoflow direction for HCS beds and classical turbidites below, hence introducing the possibility of sand transport by turbidity currents to emplace the HCS beds.

Harms, J.C., Southard, J.B., Spearing, D.R., and Walker, R.G., 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Society of Economic Paleontologists and Mineralogists, Short Course 2, 161 p.
In these notes, Harms introduced the term "hummocky cross stratification", described the deposits, and suggested a storm wave origin. Historically, an extremely important contribution.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982. *Structures and sequences in* clastic rocks. Society of Economic Paleontologists and Mineralogists, Short Course 9.
This volume is an update of SEPM Short Course 2, with an excellent section on flow, bed configurations, and stratification.

Hayes, M.O., 1967. Hurricanes as geological agents: case studies of Hurricanes Carla, 1961, and Cindy, 1963. Texas Bureau of Economic Geology, Report of Investigations No. 61, 54 p.
A classic paper which first emphasized storm surge ebb currents, and their ability to transport sediment onto the shelf and deposit graded beds. The paper has been very influential but some of Hayes' interpretations have recently been modified by Morton (1981). Nevertheless, it remains an important publication.

Hein, F.J., 1982. Slope to shelf transition: Precambrian Miette Group to Lower Cambrian Gog Group, Kicking Horse Pass/Spiral Tunnels, B.C. and Alberta. *In* Walker, R.G., ed., Clastic units of the Front Ranges, Foothills and Plains in the area between Field, B.C. and Drumheller, Alberta. International Association of Sedimentologists, 11th International Congress on Sedimentology (Hamilton, Canada). Guidebook to Excursion 21A, p. 117-136.
Measured sections and basic descriptions of the Gog Quartzite at Spiral Tunnels. Includes a basic facies classification and paleocurrent data.

Hobday, D.K., and Reading, H.G., 1972. Fairweather versus storm processes in shallow marine sand bar sequences in the late Precambrian of Finnmark, North Norway. Journal of Sedimentary Petrology, v. 41, p. 318-324.
Interpretation of major low angle dipping surfaces suggested alternations of storm and fairweather processes.

Hobson, J.P. Jr., Fowler, M.L., and Beaumont, E.A., 1982. Depositional and statistical exploration models, Upper Cretaceous offshore sandstone complex, Sussex Member, House Creek Field, Wyoming. American Association of Petroleum Geologists, Bulletin, v. 66, p. 689-707.
One of the few papers on the Western Interior Seaway to propose a *transgressive* origin for the sand bodies.

Houbolt, J.J.H.C., 1968. Recent sediments in the southern bight of the North Sea. Geologie en Mijnbouw, v. 47, p. 245-273.
Pioneering and classic account of tidal sand ridges off the British Coast. Internal structures demonstrated by Sparker surveys, and round-and-round sand movement first proposed for linear ridges.

Howarth, M.J., 1982. Tidal currents of the Continental Shelf. *In* Stride, A.H., ed., *Offshore tidal sands*. London, Chapman and Hall, p. 10-26.
Review of tidal processes on the world's shelves.

Hulsemann, J., 1955. Grossrippeln and Schragschichtungs Gefuge in Nordsee Watt und in der Molasse. Senckenbergiana Lethaea, v. 36, p. 359-388.
A discussion of "sandwaves" in the North Sea, mostly relatively small bedforms.

Huthnance, J.M., 1982. On one mechanism forming linear sand banks. Esturarine, Coastal, Shelf Science. v. 14, p. 79-99.
A mathematical treatment of sand bar growth and spacing.

Johnson, D.W., 1919. Shore processes and shoreline development. New York, Wiley, 584 p. (reprinted 1965, New York, Hafner Pub. Co.).
Introduced the idea of a graded shelf.

Johnson, M.A., Kenyon, N.H., Belderson, R.H. and Stride, A.H., 1982. Sand transport. *In* Stride, A.H., ed., Offshore tidal sands. London, Chapman and Hall, p. 58-94.
Very useful review of sand transport by tidal currents, with particular reference to the area around Britain and the southern North Sea.

Jordan, G.F., 1962. Large submarine sand waves. Science, v. 136, p. 839-848.
First major discussion of sand waves on the North American Atlantic Shelf, from the Gulf of Maine.

Klein, G. deV, 1970. Depositional and dispersal dynamics of intertidal sand bars. Journal of Sedimentary Petrology, v. 40, p. 1095-1127.
A thorough study of currents, bedforms, internal structures and sand dispersal for sand bars in the Bay of Fundy. Klein's data re-emphasized Houbolt's (1968) round-and-round "racetrack" model.

Kreisa, R.D., 1981. Storm-generated sedimentary structures in subtidal marine facies with examples from the middle and upper Ordovician of southwestern Virginia. Journal of Sedimentary Petrology, v. 51, p. 823-848.

Koldijk, W.S., 1976. Gilby Viking B: a storm deposit. *In* Lerand, M.M., ed., The sedimentology of selected clastic oil and gas reservoirs in Alberta. Canadian Society of Petroleum Geologists, p. 62-77.

La Fon, N.A., 1981. Offshore bar deposits of Semilla Sandstone Member of Mancos Shale (Upper Cretaceous), San Juan Basin, New Mexico. American Association of Petroleum Geologists, Bulletin, v. 65, p. 706-721.
Outcrop study of two bars which originally formed 50-80 km offshore. Storms more likely than tidal currents.

Levell, B.K., 1980. A late Precambrian tidal shelf deposit, the lower Sandfjord Formation, Finnmark, North Norway. Sedimentology, v. 27, p. 539-557.
Good facies descriptions of a thick Precambrian tidal quartzite with depositional summary relating the facies paleogeographically.

Mann, R.G., Swift, D.J.P., and Perry, R., 1981. Size classes of flow-transverse bedforms in a subtidal environment, Nantucket Shoals, North American Atlantic shelf. Geo-Marine Letters, v. 1, p. 39-43.

McCave, I.N., 1971. Sand waves in the North Sea off the coast of Holland. Marine Geology, v. 10, p. 199-225.
Excellent description of sandwaves up to 7 m high. Although the model of Figure 26 suggests large angle-of-repose cross bedding, the angles shown by echo sounding do not exceed 6 degrees.

McCave, I.N., and Langhorne, D.N., 1982. Sand waves and sediment transport around the end of a tidal sand bank. Sedimentology, v. 29, p. 95-110.
Detailed documentation of the orientation of smaller bedforms around the northern end of Haisborough Sand off the Norfolk Coast. Data suggests sand recirculation around the end of the sand bank.

Morgenstern, N.R., 1967. Submarine slumping and the initiation of turbidity currents. *In* Richards, A.F., ed., Marine geotechnique. Urbana, University of Illinois Press, p. 189-220.
This is still one of the best discussions of the initiation of turbidity currents.

Morton, R.A., 1981. Formation of storm deposits by wind-forced currents in the Gulf of Mexico and the North Sea. *In* Nio, S.D. *et al.*, eds., Holocene marine sedimentation in the North Sea Basin. International Association of Sedimentologists, Special Publication 5, p. 385-396.
Emphasizes the importance of wind forced currents rather than storm-surge-ebb currents for the Carla graded bed (Hayes, 1967).

Murray, S.P., 1970. Bottom currents near the coast during Hurricane Camille. Journal of Geophysical Research, v. 75, p. 4579-4582.
Brief presentation of flow data resulting from Hurricane Camille.

Narayan, J., 1971. Sedimentary structures in the Lower Greensand of the Weald, England and Bas-Boulonnais, France. Sedimentary Geology, v. 6, p. 73-109.

Nelson, C.H., 1982. Modern shallow water graded sand layers from storm surges, Bering Shelf: a mimic of Bouma sequences and turbidite systems. Journal of Sedimentary Petrology, v. 52, p. 537-545.
Describes graded beds with Bouma sequences on the Bering Shelf, but does not discuss a turbidity current origin. Hypothesizes a storm-surge-ebb origin.

Nio, S.D., 1976. Marine transgressions as a factor in the formation of sandwave complexes. Geologie en Mijnbouw, v. 55, p. 18-40.
Discusses large sandwaves in three ancient examples. The origin of some of these sandwave complexes is still controversial, but Nio's data is valuable.

Nio, S.D., and Siegenthaler, J.C., 1978. A lower Eocene estuarine-shelf complex in the Isabena Valley. State University of Utrecht, Sedimentology Group Report no. 18, p. 1-44.
Good descriptions and diagrams - a follow up to Nio, 1976.

Off, T., 1963. Rythmic linear sand bodies caused by tidal currents. American Association of Petroleum Geologists, Bulletin, v. 47, p. 324-341.
Off discusses the influence of tidal currents on linear ridges defined by bathymetric charts. There is no field data and little in the way of hydrodynamic data, but he pointed out the occurence of topographically similar features in many parts of the world.

Parker, G., Lanfredi, N.W., and Swift, D.J.P., 1982. Seafloor response to flow in a southern hemisphere sandridge field: Argentine inner shelf. Sedimentary Geology, v. 33, p. 195-216.
Description and interpretation of linear sand ridges on the Argentine shelf.

Plint, A.G., Walker, R.G., and Bergman, K.M., 1986. Cardium Formation 6. Stratigraphic framework of the Cardium in subsurface. Bulletin of Canadian Petroleum Geology, v. 34.

Reineck, H.E., 1963. Sedimentgefuge in Bereich der sudlichen Nordsee. Abhandlungen der Senckenbergischen Naturforschen des Gesellschaft, no. 505, 138 p.
Discussion of recent sediments and sandwaves in the southern North Sea.

Reinson, G.E. Foscolos, A.E., and Powell, T.G., 1983. Comparison of Viking sand stone sequences, Joffre and Caroline Fields. *In* McLean, J.R. and Reinson, G.E., eds., Sedimentology of selected Mesozoic clastic sequences. Canadian Society of Petroleum Geologists, p. 101-117.

Saxov, S., and Nieuwenhuis, J.K., eds., 1982. Marine slides and other mass movements. New York, Plenum Press, 353 p.
A collection of papers devoted to the topics named in the book title.

Seeling, A., 1978. The Shannon Sandstone, a further look at the environment of deposition at Heldt Draw Field, Wyoming. Mountain Geologist, v. 15, p. 133-144.
Compares the linear sand ridges with those in the North Sea and on the Atlantic Shelf. Does not suggest whether the main influence was tidal or storm.

Shepard, F.P., 1932. Sediments on the Continental Shelves. Geological Society of America. Bulletin. v. 43. p. 1017-1039.
Introduced the concept of relict sediments.

Shepard, F.P., 1963. Submarine geology. New York, Harper and Row. 557 p.

Spearing, D.R., 1976. Upper Cretaceous Shannon Sandstone: an offshore shallow marine sand body. Wyoming Geological Association, 28th Annual Field Conference, Guidebook, p. 65-72.
Brief description of facies and paleocurrent patterns, with interpretation in terms of tidal, storm and oceanic currents.

Strahler, A., 1963. The earth sceinces. New York, Harper and Row, 681 p.

Stride, A.H., 1963. Current-swept sea floors near the southern half of Great Britain. Quarterly Journal of the Geological Society of London, v. 119, p. 175-199.
Important and influential paper on the unmixing of older sediments inundated by the Holocene rise of sea level, and the dispersal of sand by tidal currents to form sand ribbons and sand waves.

Stride, A.H., Belderson, R.H., Kenyon, N.H.,and Johnson, M.A., 1982. Offshore tidal deposits: sand sheet and sand bank facies. *In* Stride, A.H., ed., Offshore tidal sands. London, Chapman and Hall, p. 95-125.
Excellent discussion of the origin and possible internal structure of large tidal sand bodies.

Swagor, N.S., Oliver, T.A., and Johnson, B.A., 1976. Carrot Creek Field, central Alberta. *In* Lerand, M.M., ed., The sedimentology of selected clastic oil and gas reservoirs in Alberta. Canadian Society of Petroluem Geologists, p. 78-95

Swift, D.J.P., 1975. Tidal sand ridges and shoal retreat massifs. Marine Geology, v. 18, p. 105-134.
Full discussion with many examples of the origin of these features - a classic paper.

Swift, D.J.P., 1984. Fluid and sediment dynamics on continental shelves. *In* Tillman, R.W. *et al.*, eds., Shelf sands and sand-stone reservoirs. Society of Economic Paleontologists and Mineralogists, Short Course.
A review of storm and tidal currents on the shelf, and fluid and sediment dynamics on the shoreface.

Swift, D.J., Duane, D.B., and McKinney, T.F., 1973. Ridge and swale topography of the Middle Atlantic Bight, North America: secular response to the Holocene hydraulic regime. Marine Geology, v. 15, p. 227-247.
Suggests that the ridge and swale topography of the Middle Atlantic Bight develops at the shoreline, and that growing ridges become detached during transgression. Paper contains examples and develops general mechanisms for ridge formation.

Swift, D.J.P., and Field, M.E., 1981. Evolution of a classic sand ridge field: Maryland sector, North American inner shelf. Sedimentology, v. 28, p. 461-481.
This is probably the best and most detailed study of the evolution of a field of linear sand ridges.

Swift, D.J.P., Figueiredo, A.G., Freeland, G.L., and Oertel, G.F., 1983. Hummocky cross stratification and megaripples: a geological double standard. Journal of Sedimentary Petrology, v. 53, p. 1295-1317.
A discussion of HCS, flows on the shelf and the possibility that HCS forms from combined flows. Rounded "hummocky" megaripples might be the modern HCS bedform.

Swift, D.J.P., Freeland, G.L., and Young, R.A., 1979. Time and space distribution of megaripples and associated bedforms, Middle Atlantic Bight, North American Atlantic Shelf. Sedimentology, v. 26, p. 389-406.
Best available discussion of the smaller bedforms on the Atlantic shelf.

Swift, D.J.P. and Niedoroda, A., 1985. Fluid and sediment dynamics on Continental Shelves. In R.W. Tillman *et al.*, Eds., Shelf sands and sandstone reservoirs. Society of Economic Paleontologists and Mineralogists, Short Course 13, p. 47-133.
A thorough review of processes on the modern Atlantic shelf. Careful reading of this paper will highlight for the reader the differing viewpoint of geologists and marine geologists.

Swift, D.J.P., Parker, G., Lanfredi, N.W., Perillo, G., and Figge, K. 1978. Shoreface connected sand ridges on American and European shelves: a comparison. Estuarine and Coastal Marine Science, v. 7, p. 257-273.
A useful comparison of data most of which had been previously published separately.

Swift, D.J.W., Stanley, D.J., and Curray, J.R., 1971. Relict sediments on continental shelves: a reconsideration. Journal of Geology, v. 79, p. 322-346.
This paper introduces the idea or reworking sediments during the Holocene marine transgression; historically, it ushers in modern ideas on dynamic equilibrium on shelves.

Tillman, R.W., Swift, D.J.P., and Walker, R.G., 1985. Shelf sands and sandstone reservoirs. Society of Economic Paleontologists and Mineralogists, Short Course 13, 708 p.
The volume contains much new research integrated into a series of review articles. An excellent source of current information.

Tillman, R.W. and Siemers, C.T., eds., 1984. Siliciclastic shelf sediments. Society of Economic Paleontologists and Mineralogists, Special Publication 34, 268 p.
This volume contains an excellent group of research contributions, hot off the press in May 1984.

Twichell, D.C., 1983. Bedform distribution and inferred sand transport on Georges Bank, United States Atlantic Continental Shelf. Sedimentology, v. 30, p. 695-710.
Excellent side scan sonar and echo sounding data for sand waves, sand ridges and megaripples, with inferences about sediment movement.

Van Veen, J., 1935. Sandwaves in the southern North Sea. Hydrographic Reviews, v. 12, p. 21-29.

Walker, R.G., 1982. Hummocky and swaley cross stratification. *In* Walker, R.G., ed., Clastic units of the Front Ranges, Foothills and Plains in the area between Field, B.C. and Drumheller, Alberta. International Association of Sedimentologists, 11th International Congress on Sedimentology (Hamilton, Canada), Guidebook to Excursion 21A, p. 22-30.
Full description of the geometry of HCS and SCS, and how these structures differ from angle-of-repose cross bedding.

Walker, R.G., 1983a. Cardium Formation 2. Sand body geometry and stratigraphy in Garrington-Caroline-Ricinus area, Alberta — the "ragged blanket" model. Bulletin of Canadian Petroleum Geology, v. 31, p. 14-26

Walker, R.G., 1983b. Cardium Formation 3. Sedimentology and stratigraphy in the Garrington-Caroline area. Bulletin of Canadian Petroleum Geology, v. 31, p. 213-230.
Description of coarsening upward sequences, with details of the various facies, in the Cardium "A" and "B" sands. Describes HCS in cores.

Walker, R.G., 1985. Cardium Formation at Ricinus Field, Alberta: a channel cut and filled by turbidity currents in Cretaceous Western Interior Seaway. American Association of Petroleum Geologists, Bulletin, v. 69, p. 1963-1981.

Walker, R.G., 1985a. Geological evidence for storm transportation and deposition on ancient shelves. *In* Tillman, R.W. *et al.*, eds., Shelf sands and sandstone reservoirs. Society of Economic Paleontologists and Short Courses 13, p. 243-302. These notes review in more detail than this paper the *geological* evidence for storms on shelves.

Walker, R.G., 1985b. Cardium Formation 4. Review of facies and depositional processes in the southern Foothills and Plains, Alberta. *In* Tillman, R.W. *et al.*, eds., Shelf sands and sandstone reservoirs. Society of Economic Paleontologists and Mineralogists, Short Course 13, p. 353-402.
Review of Cardium facies, depositional processes and environments in surface and subsurface. Ricinus field is interpreted as a channel which was both cut and filled by turbidity currents.

Walker, R.G., 1985c. Ancient examples of tidal sand bodies formed in open, shallow seas. *In* Tillman, R.W. *et al.*, eds., Shelf sands and sandstone reservoirs. Society of Economic Paleontologists and Mineralogists, Short Course 13, p. 303-341.
Review of many examples of ancient tidal sand bodies — a useful collection of illustrations.

Walker, R.G., Duke, W.L., and Leckie, D.A., 1983. Hummocky stratification: significance of its variable bedding sequences: discussion. Bulletin of the Geological Society of America, v. 94, p. 1245-1249.
Discussion of Dott and Bourgeois (1982), modifying their internal structure sequence and commenting further on the mode of formation HCS.

Walker, R.G., and Hunter, D.F., 1982. Transition, Wapiabi to Belly River Formation at Trap Creek, Alberta. *In* Walker, R.G., ed., Clastic units of the Front Ranges, Foothills and Plains in the area between Field, B.C. and Drumheller, Alberta. International Association of Sedimentologists, 11th International Congress on Sedimentology (Hamilton, Canada), Guidebook to Excursion 21A, p. 61-71.
Basic descriptions of the Trap Creek and Highwood River sections.
No interpretations.

Winn, R.D. Jr., Stonecipher, S.A., and Bishop, M.G., 1983. Depositional environments and diagenesis of offshore sand ridges, Frontier Formation, Spearhead Ranch Field, Wyoming. Mountain Geologist, v. 20, p. 41-58.
Subsurface study of storm-generated ridges with HCS in coarsening-upward sequences.

Wright, M.E., and Walker, R.G., 1981. Cardium Formation (U. Cretaceous) at Seebe, Alberta - storm transported sandstones and conglomerates in shallow marine depositional environments below fair-weather wave base. Canadian Journal of Earth Sciences, v. 18, p. 795-809.
Description of the classic outcrop at Seebe, and calculations pertaining to the mode and rate of gravel emplacement into the Cardium basin.

Turbidites and Associated Coarse Clastic Deposits

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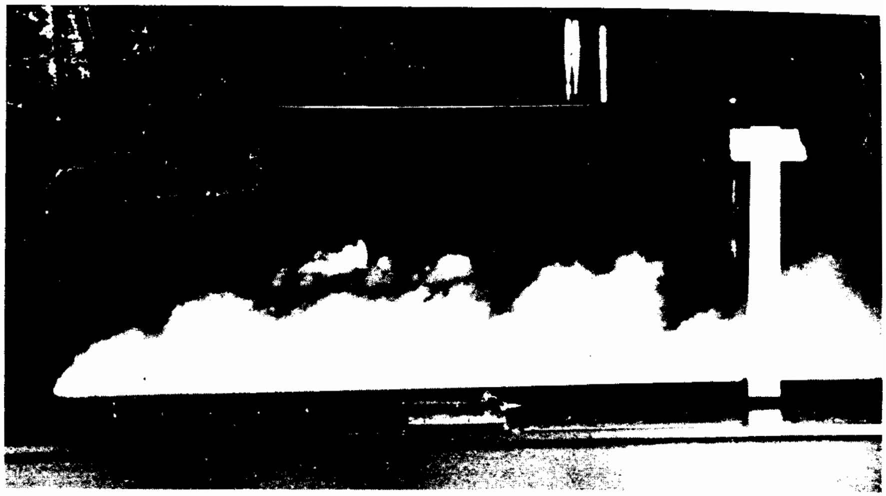

Figure 1
Experimental turbidity current in a flume at Caltech. Water depth is 28 cm. Note characteristic shape of the head of the current, and eddies behind the head. Sediment is thrown out of the main flow by these eddies - the main flow is only about half of the height of the head.

INTRODUCTION

The turbidity current concept is both simple and elegant. Each turbidite is the result of a single, short-lived event, and once deposited it is extremely unlikely to be reworked by other currents. The concept is elegant because it suggests that the deposition of thousands of graded sandstone beds, alternating with shales, is the result of a series of similar events. It can safely be stated that no similar volume of clastic rock can be interpreted so simply.

This review is presented in six parts, following closely the philosophy of facies models outlined in the "General Introduction" to this volume;

1) Introduction to turbidity currents and turbidites.
2) The variety of turbidites in the geologic record (the model as descriptor).
3) Turbidites in modern oceans (model as descriptor).
4) Combination of ancient and modern examples (to distill a model).
5) Use of the model (the model as predictor).
6) Feedback - facies sequences refining existing models and defining new ones.

TURBIDITY CURRENTS AND TURBIDITES

Density currents flow downslope on the ocean floor, being driven by gravity acting on the density difference between the current and the surrounding sea water. The density could be due to colder temperatures, higher salinities, or suspended sediment in the current. When the density is due to suspended sediment, the flow is termed a *turbidity current* (Fig. 1). A *turbidite* is defined as the deposit of a turbidity current.

The concept of turbidity currents was introduced to the geological profession in 1950. At that time, nobody had observed a modern turbidity current in the ocean, yet the evidence for turbidity currents had become overwhelming. The concept accounted for graded sandstone beds that lacked evidence of shallow water reworking, and it accounted for transported shallow water foraminifera in the sandstones, yet with bathyal or abyssal foraminifera in the interbedded shales. Low density currents were known in lakes and reservoirs, and they appeared to be competent to transport sediment fairly long distances. Many of these different kinds of evidence were pulled together by Kuenen and Migliorini (1950) when they published experimental and field observations in a now classic paper on "Turbidity currents as a cause of graded bedding". A full review of why and how the concept was established in geology was published by Walker (1973).

It is now known that turbidity currents operate on vast scales. In 1935, a slump removed 480 m of breakwater at the mouth of the Magdalena River in Colombia. The slump cut a channel 10 m deep through a bar, evolved into a turbidity current, and several hours later broke a submarine telegraph cable 24 km from the river mouth in 1400 m water depth (Heezen, 1956; Menard, 1964, p. 197). Surveys of the sea floor before and after the slump indicate that the minimum volume of sediment lost was $3x10^8$ m^3. This is an unimaginably large volume of sediment - it would require 2.14 million standard 50-foot box cars to transport this sediment by railroad. The resulting train would be 35,000 km long. However, the largest turbidite known makes the Magdalena flow seem small. The "black shell" turbidite, named for the distinctive corroded shells that it contains, covers an area about 500 km long and 200 km wide on the Hatteras Abyssal Plain off the eastern margin of North America. Its volume has been estimated at over 100 km^3 (Elmore *et al.*, 1979), about 333 times that of the Magdalena flow.

Relatively little is known about flow velocities. The classic data comes from the 1929 earthquake near the Grand Banks of Newfoundland, which triggered a flow that broke a sequence of submarine cables. Recently recalculated velocities of the head of the flow (Uchupi and Austin, 1979) give 20.3 m/sec at the cable broken 183 minutes after the quake, 14.4 m/sec (541 minute break), 12.8 m/sec (618 minute break), and 11.4 m/sec at the 797 (13 hours 17 minutes) minute break. At a velocity of 11.4 m/sec, the current could suspend by fluid turbulence alone low concentrations of quartz pebbles up to about 3 cm in diameter.

The Grand Banks flow appears to have travelled several hundred kilometres across the essentially flat Sohm Abyssal Plain. Similarly, the "black shell" flow must have travelled at least 500 to 600 km along the Hatteras Abyssal Plain.

Turbidity currents can be triggered by earthquakes (as in the Grand Banks), by rivers in flood (as in the Congo, Heezen *et al.*, 1964) and by spontaneous failure of rapidly deposited piles of sediment, commonly with relatively fine grain sizes and high pore pressures (the Magdalena flows [Heezen, 1956; Menard, 1964] seem to have been of this type). It has been suggested recently that cyclic wave loading by major storms can liquefy enough sediment near the shoreline to generate turbidity currents; see, for example, "Shelf and Shallow Marine Sands" in this volume.

Turbidity currents must be considered commonplace in modern seas and oceans. Their deposits are likely to be extensive and volumetrically important. To preserve the sedimentary structures made by the turbidity currents (i.e., to be able to *recognize* the beds as turbidites), deposition must take place below effective wave base. It has been suggested that turbidites have been deposited in some epeiric seas (see "Shelf and Shallow Marine Sands", in this volume), as well as in the more traditional "deep water" turbidite habitat of submarine fans and basin plains.

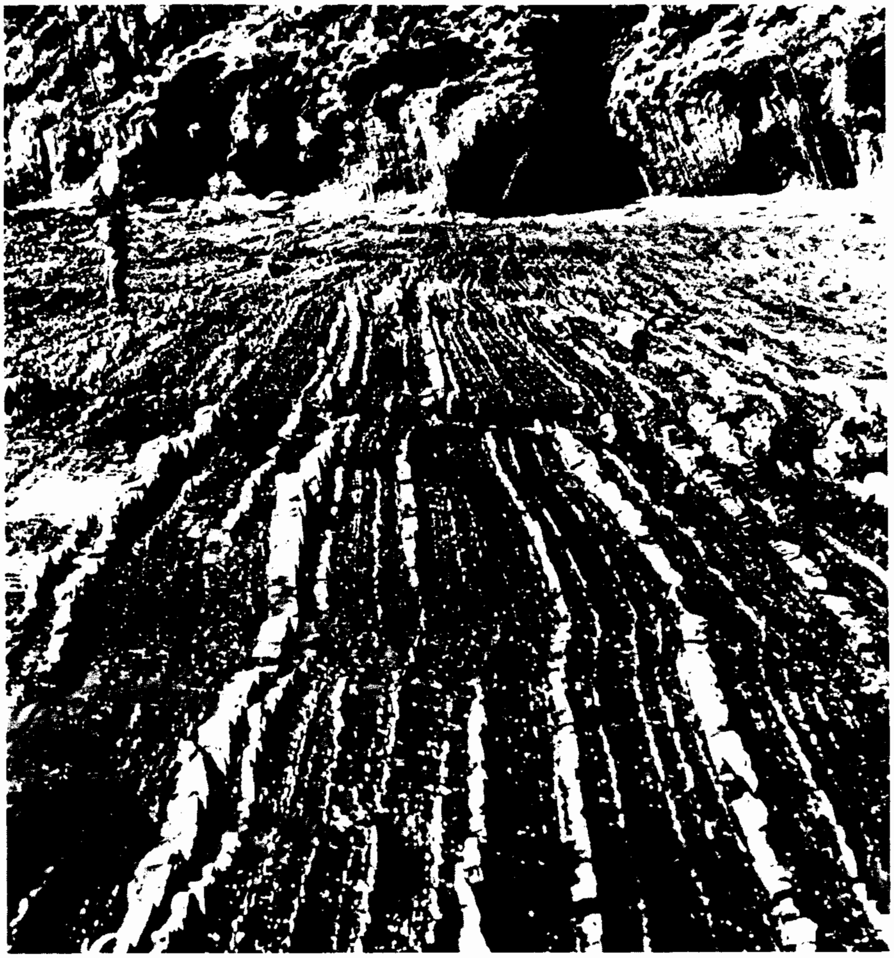

Figure 2
Monotonous interbedding of thin, sharp-based sandstones and mudstones. No sea-floor topography (channels, levees) visible. Stratigraphic top to right. Devonian, Cape Liptrap, South Australia.

TURBIDITES IN THE GEOLOGIC RECORD

After its introduction in 1950, the turbidity current concept was applied to rocks of many different ages, in many different places. Emphasis was laid upon describing a vast and new assemblage of sedimentary structures, and using those structures to interpret paleocurrent directions. In the absence of a turbidite facies model, there was no norm with which to compare individual examples, no framework for organizing observations, no logical basis for prediction in new situations, and no basis for a consistent hydrodynamic interpretation. Yet gradually during the years 1950-1960, a relatively small but consistent set of sedimentary features began to be associated with turbidites. These are considered in the following list, and can now be taken as a set of descriptors for *classical turbidites:*

1) Sandstones and shales are monotonously interbedded through many tens or hundreds of metres of stratigraphic sections (Fig. 2). Beds tend to have flat tops and bottoms, with no scouring and channelling on a scale greater than a few centimetres.
2) Sandstone beds have sharp, abrupt bases, and tend to grade upward into finer sand, silt and mud. Much of the mud was brought into the basin by the turbidity current (it contains a shallow water transported faunal assemblage), but the uppermost very fine clay may contain a bathyal or abyssal benthonic fauna and hence represent slow hemipelagic deposition between turbidity current events.
3) On the undersurface (sole) of the sandstones there are abundant markings, now classified into three types; *tool* marks carved into the underlying mud by rigid objects (sticks, stones) in the turbidity current; *scour* marks cut into the underlying muds by fluid scour; and *organic* markings representing trails and burrows filled in by the turbidity current. Tool and scour marks give accurate indications of local paleoflow directions, and by now, many thousands have been measured to reconstruct paleoflow patterns in hundreds of turbidite basins.
4) Within the sandstone beds, combinations of parallel lamination (Fig. 3), ripple cross lamination (Fig. 3), climbing ripple cross lamination, convolute lamination and graded bedding (Fig. 3) have been noted by many authors. An ideal,

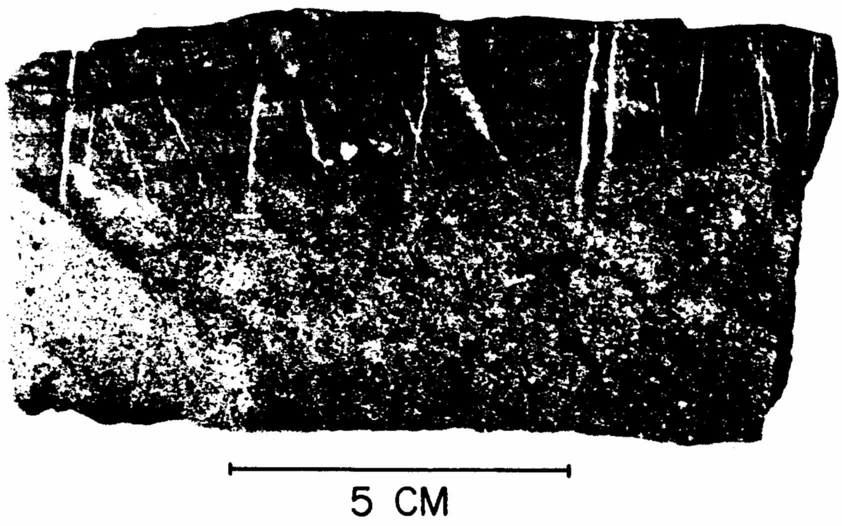

Figure 3
Complete Bouma sequence, beginning with a graded division A, overlain by parallel laminated division B and cross-laminated division C. Divisions D and E (see Fig. 4) broke off this specimen, which is from the Cote Frechette road cut, Levis Formation (Cambrian), Québec.

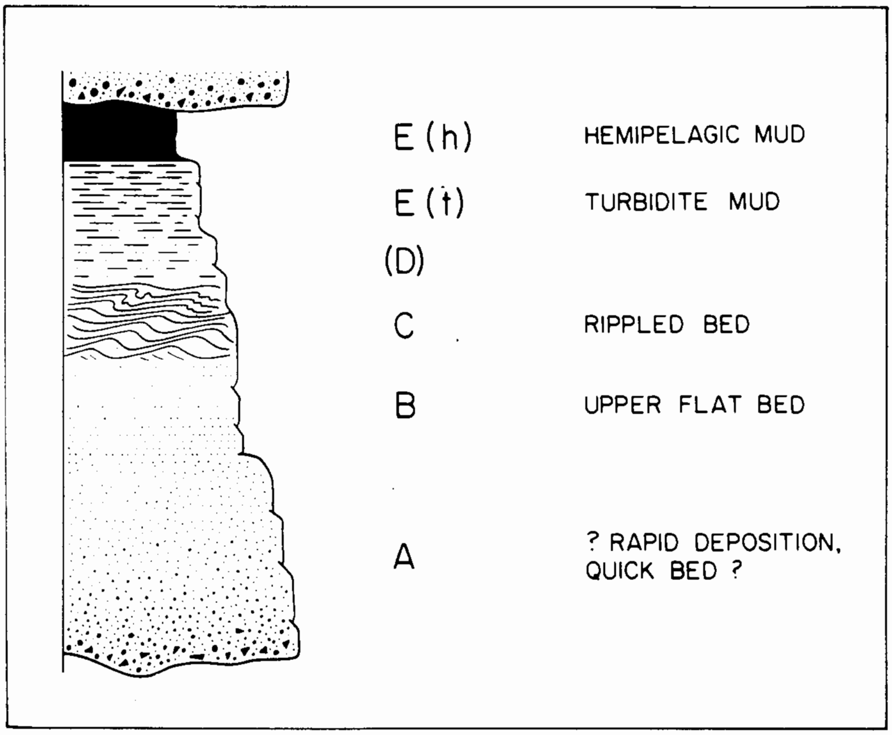

Figure 4
Five divisions of the Bouma sequence: A) massive or graded; B) sandy parallel laminations; C) rippled and/or convoluted; D) delicate parallel interlaminations of silt and mud; E(t)) mud introduced by the turbidity current and E(h)) the hemipelagic background mud of the basin. See text for details.

or generalized sequence was proposed by Arnold Bouma in 1962, and the Bouma sequence (Figs. 3 and 4) can be regarded as an excellent facies model for *classical* turbidites (see "General Introduction", especially Fig. 9).

THE BOUMA SEQUENCE AS A FACIES MODEL

On a small scale, the Bouma sequence has functioned so well as a facies model that I will digress briefly to illustrate some of the ideas developed in the introductory paper to this volume. First, the Bouma sequence has been distilled from a vast number of examples - literally thousands of individual beds. It can therefore be regarded as a homogeneous model of great generality. It functions well as a *norm* (Fig. 4), or point of comparison, and hence helps to explain those turbidites without the full sequence (Walker, 1967). For example, without a norm we would not know that BDE turbidites were any more or less common than ABCDE turbidites. The *norm* establishes a general point of reference. The model has acted well as a *guide* for further observations, making one aware both of the features presented by any one bed, and of features embodied in the model that might be missing in any specific bed.

The model has acted well as a *predictor*. For example, if an outcrop shows beds that begin only with Bouma's division C, the model predicts that these were deposited from slower turbidity currents, perhaps in a more distal geographic setting than beds which begin with Bouma's division A (Fig. 5; Walker, 1967). Alternatively, groups of beds beginning with division C might be proximal levee deposits, laterally adjacent to beds beginning with division A in a nearby channel.

Finally, the model has acted as a general basis for hydrodynamic interpretations. Before the Bouma sequence, varied interpretations were offered for individual beds or groups of beds. The Bouma sequence suggests a single coherent interpretation (Fig. 4), with division A suggesting very rapid settling of grains from suspension, possible in such quantities and at such a rate that water is rapidly expelled upward, and momentarily the grain/water mixture becomes fluidized. Fluidization would

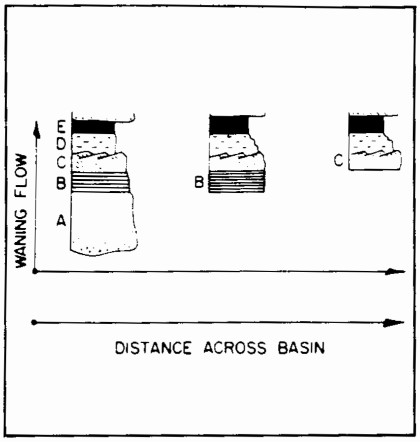

Figure 5
The ABCDE sequence in one individual turbidite suggests waning flow at the depositional site. Using the Bouma sequence as a predictor, we could suggest that groups of beds beginning with division B, and with division C, must represent deposition from progressively slower currents. Waning flow in the lateral sense can be correlated with distance flowed across the basin. There are limitations to this prediction - see text.

destroy any possible sedimentary structures except graded bedding. The second phase of deposition involves traction of grains on the bed, with division B representing the "upper plane bed" of experimental work (Harms *et al.*, 1982), and division C representing a rippled bedform. The upper flat bed passes directly into a rippled bed (with no formation of dunes) if the grain size is finer than about 0.15 mm (Southard, in Harms *et al.*, 1982, Figs. 2 to 5), as it is in many turbidites. If there is a high rate of deposition from suspension during rippling, climbing ripple cross lamination will form. Finally, as the flow dies away, turbidity current mud will blanket the bed (division (D) and E(t)), followed by hemipelagic mud E(h) (Fig. 4).

GENERAL TURBIDITE FACIES CLASSIFICATIONS

The above discussion was concerned with *classical turbidites*, those which consist of monotonous alternations of sandstones and shales, parallel bedded without significant scouring or channelling, and where all the beds can reasonably be described using the Bouma sequence.

It is interesting that the turbidite system was the first in which a universal facies scheme was proposed, by Mutti and Ricci Lucchi (1972). A universal scheme for fluvial deposits has recently been introduced and is discussed by Rust and Koster ("Coarse Alluvial Deposits", this volume). The Mutti and Ricci Lucchi scheme has been modified over the years (Mutti and Ricci Lucchi, 1975; Mutti, 1979), and is a more detailed scheme than is required here. I will use the simpler scheme introduced by Walker (1978), namely:

1) classical turbidites (discussed above),
2) massive sandstones,
3) pebbly sandstones,
4) conglomerates,
5) slumps, slides, debris flows and other exotic facies.

Both descriptive schemes serve their purposes well, and both can be related to deposition on various parts of submarine fans, as discussed below.

MASSIVE SANDSTONES

This facies consists of thick sandstones with thin (or absent) interbedded shales (Fig. 6). Individual sandstone beds range in thickness from about 50 cm to many metres, and the only Bouma division normally present is division A. A typical sequence of beds would be measured as A.A.A.A. using the Bouma model. However, I would consider this to be a mis-application of the Bouma model, because it is characteristically a five-part model being applied to beds that charateristically only contain one part. The functions of the model as norm, guide, predictor, and basis for interpretation are all seriously weakened to the point of uselessness if the beds only show an A.A.A.A. sequence.

The massive sandstones are commonly not so parallel sided as the classical turbidites; channelling is more common, and one flow may cut down and weld onto the previous one ("amalgamation") giving rise to a series of multiple sandstone beds.

The one common sedimentary structure found in the massive sandstones is termed "dish" structure, and is indicative of abundant fluid escape during deposition of the sandstone (Lowe, 1975). It indicates rapid deposition of a large amount of sand from a "fluidized flow" (akin to a flowing quicksand). This does not imply that the massive sandstone facies was transported all the way from source into the basin by a fluidized flow. However, it does imply that a turbidity current, which normally maintains its sand load in suspension by fluid turbulence, can pass through a stage of fluidized flow during the final few seconds or minutes of flow immediately preceding deposition. The massive sandstone facies is prominent in the Cambrian Charny Formation around Québec City and Lévis, and dish structures in massive sandstones are common in the Cambro-Ordovician Cap Enragé Formation (Hein,1982) near Rimouski, Québec. Massive sandstones are also well represented in many of the Cretaceous and Tertiary turbidite sequences of California and Oregon (e.g., Link and Nilsen, 1980; Link *et al.*, 1981; Nilsen and Abbott, 1981; Link and Welton, 1982; Chan and Dott, 1983).

Figure 6
Massive sandstone facies: the Upper Eocene Annot Sandstone, southern France. About 180 m of section can be seen in the photograph. Note thickness of individual sandstone beds, and absence of mudstone interbeds.

PEBBLY SANDSTONES

The pebbly sandstone facies cannot be described using the Bouma model, nor does it have much in common with the massive sandstone facies. Pebbly sandstones tend to be well graded (Fig. 7),

Figure 7
Graded bed of pebbly sandstone, followed abruptly by a second bed without a mudstone interbed. St. Damase Formation (Orodvician) near Kamouraska, Québec.

Figure 8
Pebbly sandstone facies showing medium scale cross bedding. In isolation, this photograph could easily be confused with one of fluvial gravels. In fact, it is from the Cambro-Ordovician Cap Enragé Formation, and the cross beds are interbedded with classical turbidites and graded pebbly sandstones.

and stratification is fairly abundant. It can either be a rather coarse, crude, horizontal stratification, or a well developed cross bedding of the trough, or planar-tabular type (Fig. 8). Imbrication of individual pebbles within the bed is common. At present, there is no "Bouma-like" model for the internal structures of pebbly sandstones; the sequence of structures, and their abundance and thickness has not yet been distilled into a general model. Models based on the characteristics of the Cap Enragé Formation have been proposed by Hein (1982). Pebbly sandstone beds are commonly channelled and laterally discontinuous, and interbedded shales are rare.

It is clear that with abundant channelling, and the presence of cross bedding in pebbly sandstones, this facies could easily be confused with a coarse fluvial facies (see "Coarse Alluvial Deposits", this volume). The differences are subtle and can be misleading to sedimentologists – the safest way to approach the interpretation of pebbly sandstones is to examine their context. If associated with, or interbedded with classical turbidites, the pebbly sandstone interpretation would be clear. Similarly, if associated with non-marine shales, root traces, caliche-like nodules, mud cracks, and other indicators of flood plain environments, the interpretation would also be clear. This facies highlights the fact that environmental interpretations cannot be based upon a "checklist" of features: the relative abundance and type of features, in their stratigraphic context, must always be the basis of interpretation.

Pebbly sandstones are particularly well exposed in the Cambro-Ordovician Cap Enragé Formation (Hein, 1982) at St. Simon (near Rimouski, Québec), where grading, stratification and cross bedding are prominent. The facies is also abundant in the Cambrian St. Damase Formation near Kamouraska, Québec, and in the Cambrian St. Roch Formation at L'Islet Wharf (near St-Jean-Port-Joli, Québec) (Walker, 1979). Many examples exist in the Cretaceous and Tertiary turbidite sequences of California and Oregon (e.g., Nilsen and Abbott, 1981).

CONGLOMERATES

Although volumetrically less abundant than classical turbidites, conglomerates are an important facies in deep water environments. They are abundant in California and Oregon (e.g., Walker, 1977; Nilsen and Abbott, 1981), and are particularly well exposed at many localities in the Gaspé Peninsula (Davies and Walker, 1974; Hendry, 1978; Johnson and Walker, 1979; Hein, 1982). Sedimentologists have tended to ignore conglomerates, probably because without a facies model, there has been no framework to guide observations, and hence the feeling of "not being quite sure what to measure in the field". I have proposed some generalized "Bouma-like" models for conglomerates (Walker, 1975a), but because the models are based upon

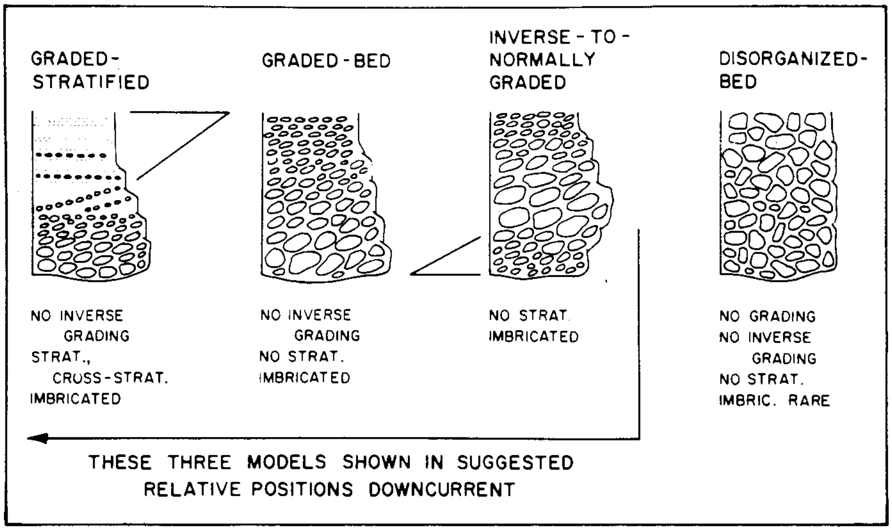

Figure 9
Four models for resedimented (deep water) conglomerates, shown in their inferred downcurrent relative positions. See text for details.

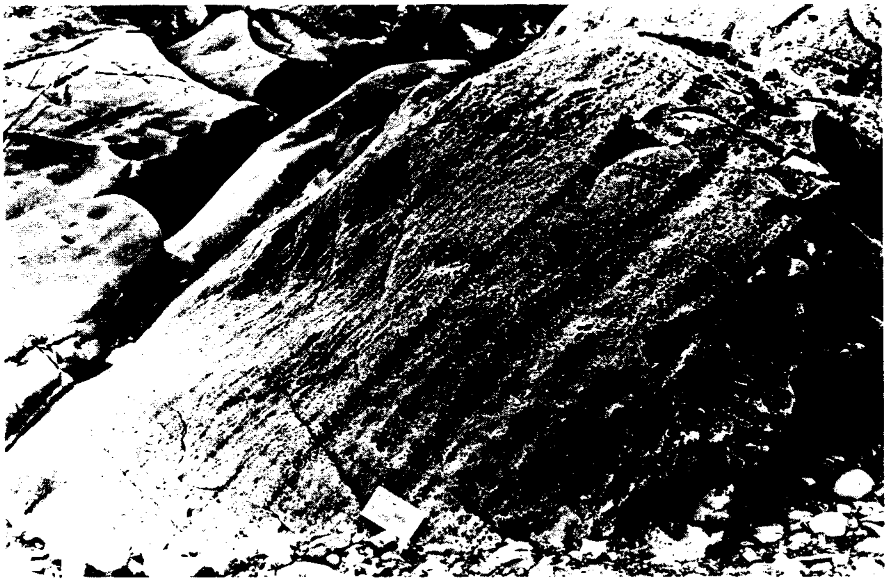

Figure 10
Normally-graded and stratified conglomerate. Cambro-Ordovician Cap Enragé Formation. Bic, Québec. Basal conglomerate rests on slates, and grades up into stratified conglomerate, very coarse sandstone with crude "dish structure", and finally into massive structureless sandstone.

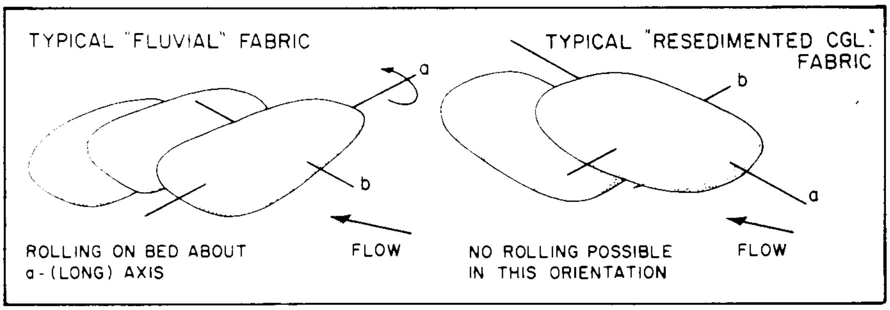

Figure 11
Contrast between a fabric produced by clast rolling (a-axis transverse), and a fabric characteristic of resedimented conglomerates (a-axis parallel to flow and dipping upstream). The a-axis-parallel fabric is incompatible with clast rolling, and is believed to form by clasts colliding in the flow, whilst dispersed above the bed.

fewer than thirty studies, they lack the universality and authority of the Bouma model for classical turbidites. The paper (Walker, 1975a) discusses the models, their relationships, and how they were established. In Figure 9, it can be seen that the descriptors include the type of grading (normal [Fig. 10] or inverse), stratification (Fig. 10), and fabric; in different combinations they give rise to three models which are probably intergradational, and a fourth (disorganized-bed) characterized only by the absence of descriptors.

One of the most important features of conglomerates is the type of fabric they possess. In fluvial situations, where pebbles and cobbles are rolled on the bed, the long (a-) axis is usually transverse to flow directions, and the intermediate (b-) axis dips upstream, characterizing the imbrication. However, for most conglomerates associated with turbidites, the fabric is quite different: the long axis is parallel to flow, and also dips upstream to define the imbrication (Fig. 11). This fabric is interpreted as indicating no bedload rolling of clasts. The only two reasonable alternatives involve mass movements (debris flows), or dispersion of the clasts in a fluid above the bed. Mass movements in which clasts are not free to move relative to each other do not produce abundant graded bedding, stratification, and cross-stratification, so I suggest the clasts were supported above the bed in a turbulent flow, The support mechanism may have been partly fluid turbulence, and partly clast collisions. Upon deposition, the clasts immediately stopped moving (no rolling), and the fabric was "frozen" into the deposit.

In the absence of experimental work on cobbles and boulders, the interpretation of the conglomerate models must be based largely on theory. I suggest a downcurrent trend from the inverse-to-normally-graded model, into the graded-stratified model. This trend does not necessarily exist in any one bed: rather, deposition from a particular current in one of the three downstream positions in Figure 9 will be of the type indicated in the figure.

Clast supported conglomerates are abundant in the Ordovician Grosses Roches Formation (Hendry, 1978) and Cambro-Ordovician Cap Enragé Formation (Hein, 1982; Hein and Walker, 1982), Gaspé Peninsula, Québec, and also make up part of the Cambrian St. Roch Formation east of Rivière-du-Loup, Québec. They are abundant in California and Oregon (Walker, 1977 and in press; Nilsen and Abbott, 1981).

SLUMPS, SLIDES, DEBRIS FLOWS AND EXOTIC FACIES

This facies includes a diverse group of rocks which are generally poorly to unstratified, which are commonly poorly sorted (blocks and boulders in a fine grained matrix), and which may show evidence of sedimentary deformation.

The debris flow deposits have clasts supported in a muddy matrix – they may show basal inverse grading and preferred clast alignment. Because the larger clasts in a debris flow are maintained above the bed by the strength of the matrix, the deposit commonly has large blocks projecting up above the top of the bed, or even resting almost entirely on top of the bed. The deposit shows no internal evidence of slumping.

By contrast, other exotic facies commonly show evidence of slumping, and represent the mixing of sediment within the depositional basin by post-depositional slumping. The deposits can range all the way from very cohesive slumps involving many beds, to very watery slumps generated by the deposition of coarse sediment on top of wet, poorly consolidated clays. The latter process gives rise to the classical pebbly mudstones (Crowell, 1957; Howell and Joyce, 1981).

Inasmuch as subaqueous debris flows, and slumps, require greater slopes than classical turbidity currents, the chaotic facies is most abundant at the foot of the slope into the basin. Very few examples have been described in Canada. Large scale slumps are known in Upper Ordovician turbidites in northeastern Newfoundland (Helwig, 1970), and pebbly mudstones are known in several units in western Newfoundland (Stevens, 1970). The best described debris flows are Devonian reef-margin examples adjacent to the Ancient Wall, Miette and Southesk-Cairn reef complexes in Alberta (Cook *et al.*, 1972; Srivastava *et al.*, 1972). Elsewhere, slumps have been described from the Tortonian of northwestern Italy (Clari and Ghibaudo, 1979), the Plio-Pleistocene of northern California (Piper *et al.*, 1976) and the Miocene of New Zealand (Gregory, 1969).

TURBIDITES IN MODERN OCEANS

Effective facies models must combine data from ancient and recent sediments, and hence it is necessary to review briefly what is known about turbidites in modern oceans. The main depositional environments are submarine fans (which may coalesce laterally to build up the Continental Rise) and basin plains. By far the greatest volume of modern turbidites occur in the submarine fans.

Many different submarine fans have now been described, and general models that try to summarize this work have been presented by Normark (1970, 1978). In his 1970 paper ("Growth patterns of deep-sea fans"), Normark proposed a general model, widely accepted by the profession, that was built essentially on data from only two California Borderland fans, La Jolla and San Lucas. The model consisted of three parts – a leveed valley on the upper fan, a mid-fan built up of suprafan lobes that periodically switched position, and a flat lower fan without channels. Many more examples were incorporated in Normark's (1978) later statement of the model (Fig. 12).

The most detailed study of the evolution of a small fan is that of Normark *et al.* (1979) and Piper and Normark (1983) for Navy Fan, California Borderlands. Based on precision echo sounding, seismic reflection profiling and side scan sonar, a three-dimensional physiographic map of Navy Fan was produced, showing the surface and subsurface locations of six suprafan lobes. By studying the way in which these lobes overlap, an evolutionary sequence was determined (Fig. 13).

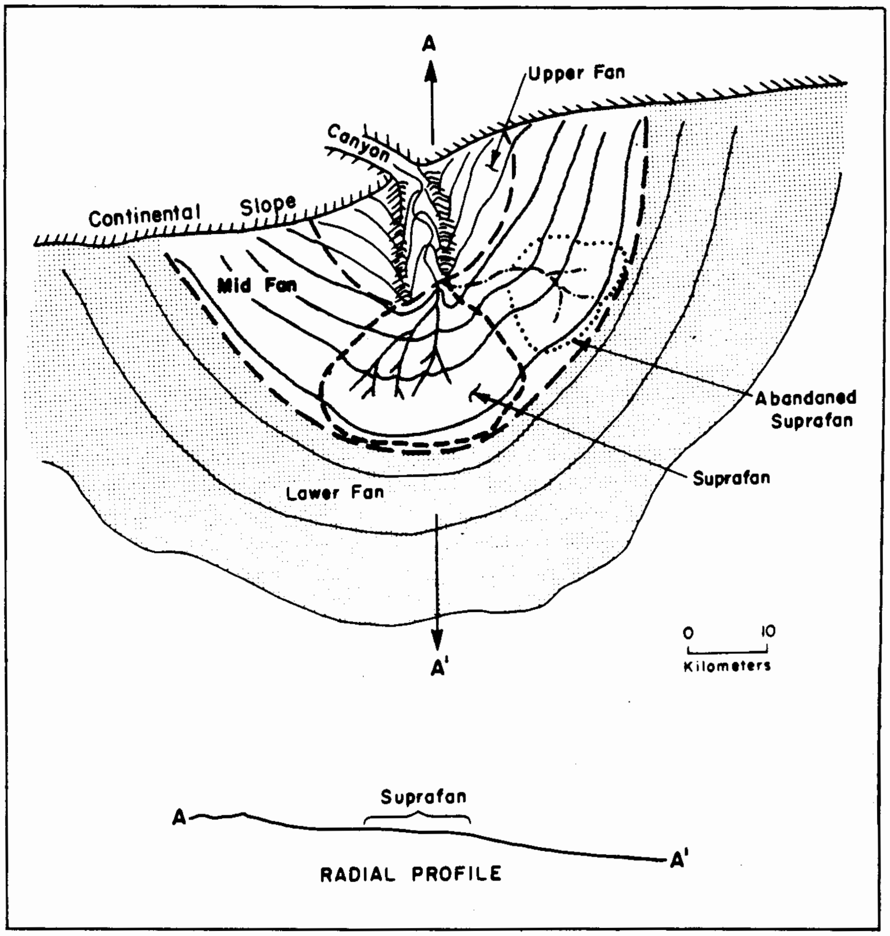

Figure 12
Submarine fan model of Normark (1978), based on several studies of modern fans. This model for fan growth emphasizes active and abandoned depositional lobes termed suprafans.

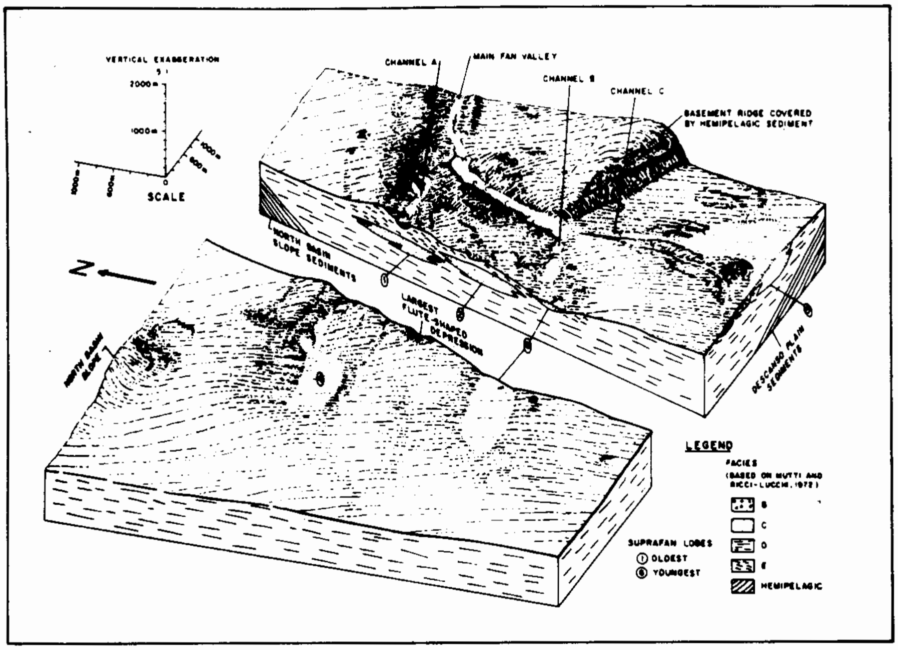

Figure 13
Block diagram of Navy Fan, California, based on seismic reflection profiling. Note suprafan lobes at ends of channels, with lobe 6 at the end of channel B being the youngest. From Normark et al. (1979) – these authors discuss in detail the pattern of channel and lobe switching. This fan strongly supports the ideas embodied in Normark's fan model (Fig. 12).

The upper fan has a single leveed channel that is about 400 m wide, but decreases in depth from 50 to 15 m over a length of 8 km. Only distributary channel B (Fig. 13) is presently continuous with the upper fan leveed channel. The pattern of channel switching suggests that as one lobe grows, and its feeder channel aggrades (or backfills), it eventually initiates a levee break and turbidity currents are diverted to a lower part of the mid-fan surface to begin construction of a new lobe. Thus on Navy Fan, the mid-fan is built up of a series of individual lobes formed by distinct jumps in the positions of distributary channels (not gradual lateral channel migration). The implication is that when lobe 5 (say) is active, lobes 1,2 and 4 receive only fine grained muds spilled over the distributary channel margins. In other words, when one suprafan lobe is active, other lobes and their former distributary channels are being blanketed by mud. This process is important in forming stratigraphic traps over potential suprafan oil and gas reservoirs. The geometry of individual beds, and their mode of deposition, has been studied in detail by Piper and Normark (1983).

Although Navy Fan has been studied in the most detail, Normark's (1978) later model was based particularly on Astoria, Monterey, Amazon and Bengal Fans (large fans built on oceanic crust), and Redondo, La Jolla, Navy and San Lucas Fans from the California Borderland. Since 1978, there have been detailed studies of Laurentian Fan (Uchupi and Austin, 1979; Stow, 1981; Piper and Normark, 1982; Normark *et al.*, 1983), Mozambique Fan (Kolla *et al.*, 1980), Amazon Cone (Damuth and Embley, 1981; Damuth *et al.*, 1983a, 1983b), Zodiac Fan (Aleutian Abyssal Plain, Stevenson *et al.*, 1983), Magdalena Fan (Colombia; Kolla *et al.*, 1984) and La Jolla Fan (Graham and Bachman, 1983).

These studies cannot easily be combined with Normark's (1978) model. For example, the Amazon Cone appears to have three huge "slump/debris flow complexes" which cover areas of 32,500 km^2, 28,850 km^2 and 21,200 km^2 (Damuth and Embley, 1981, p. 633-637). The thicknesses are less than 75 m, 10 to 50 m and up to 50 m, respectively. It is not clear how these complexes formed on the very low slopes of the fan. Also on Amazon Cone are a series of channel-levee complexes, with amazingly sinuous channel patterns (Damuth *et al.*, 1983a) which cannot be related to present concepts of turbidity current flow.

Remarkably little is known about the ages and thicknesses of major submarine fans. Few have been penetrated completely during the Deep Sea Drilling Program. Many appear "to have built out substantially since Miocene time, and especially during Pleistocene time" (Kelts and Arthur, 1981). Thicknesses range from about 300 m (margin of Astoria Fan) to well over 10 km (Bengal Fan). Rates of deposition can be incredibly high; at DSDP site 222 on the Indus Cone, sediment was deposited at 600 m/m.y. in the late Miocene, 135 to 350 m/m.y. in the Pliocene, and at less than 50 m/m.y. in the Quaternary (Whitmarsh *et al.*, 1974).

Fan sediments tend to pass distally into basin plain deposits. Here, the turbidites tend to be very extensive and continuous (Pilkey *et al.*, 1980), with more abundant and thicker sands close to points of entry onto the basin plains. Thicknesses on modern abyssal plains tend to be only a few hundred metres (Horn *et al.*, 1972). These few generalities constitute the essence of a "model" for abyssal, or basin plains.

The problem of coring deep sea facies, particularly the sands, and preserving long (several metres) sections with sedimentary structures, makes the comparison of modern and ancient turbidite facies rather difficult. Modern fan studies have contributed data on fan morphology on a rather large scale, whereas most studies of ancient rocks have been of a smaller scale. Studies such as those by Normark *et al.* (1979) and Damuth *et al.* (1983a, 1983b) are beginning to show the fine topographic details of modern fans – the problems of comparison are addressed below.

GENERAL FAN MODELS: COMPARISON OF ANCIENT AND RECENT SEDIMENTS

Interpretations of ancient sediments as fan deposits began in the early 1960s (Sullwold, 1960; Walker, 1966), but became more common as more modern fans were studied (Normark, 1970). Emiliano Mutti and his colleagues in Italy contributed extremely influential work (especially Mutti and Ricci Lucchi, 1972; Mutti and Ghibaudo, 1972), proposing a fan model based upon ancient rocks in Italy. This model was so similar

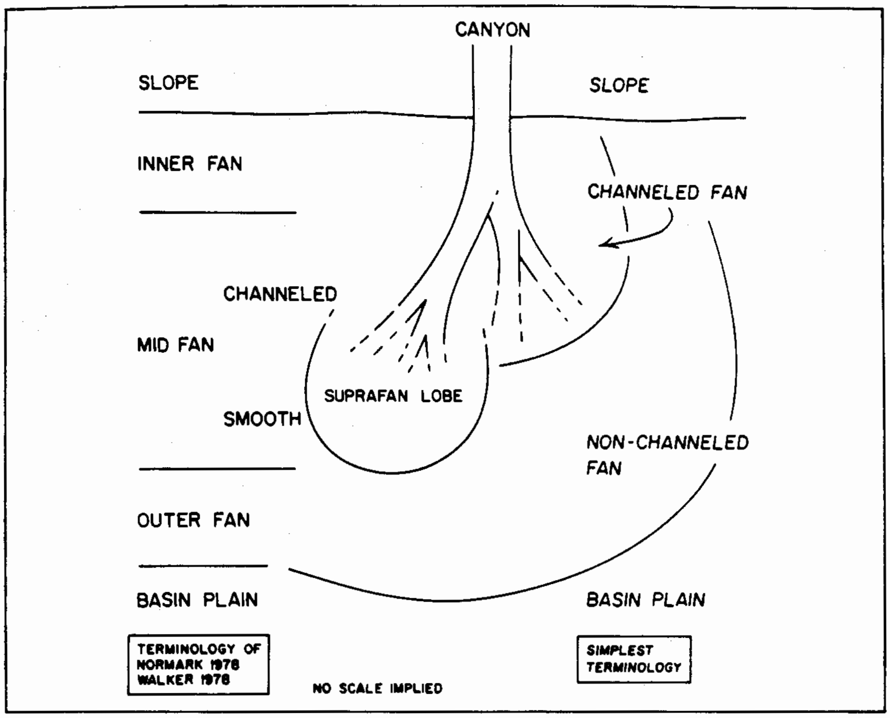

Figure 14
Simplified fan model. On the right is the simplest possible terminology, showing the absolute basics of almost all ancient and modern fans. On the left is a terminology which, although a little more complex, embodies the salient characteristics of Normark's model (Fig. 12) for modern fans along with the characteristics proposed by Mutti and colleagues for many inferred ancient fans.

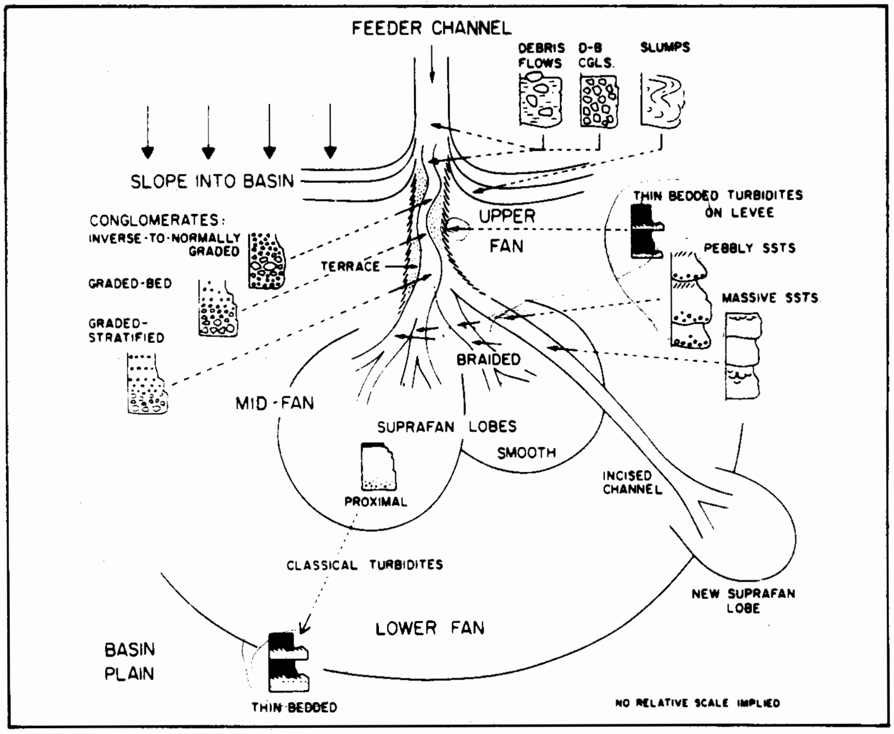

Figure 15
Fan model proposed by Walker (1978). Note that it incorporates features (terraces, inner fan meandering channel, levees, etc.) which although common, may not occur on all fans. Facies defined in ancient rocks are shown in their inferred positions on the fan. An incised channel is also shown, indicating a phase of downcutting, fan extension, and new lobe development (as in the modern La Jolla Fan of California).

to that of Normark's for modern fans that modern and ancient studies were distilled together to form the first modern-ancient integrated model (Walker and Mutti, 1973). As more work has been done, this model has evolved and diversified (see Walker, 1980, p. 1101-7), and I use the sketch in Figure 14 as a starting point. In the simplest possible terminology, almost all modern fans can be subdivided into a *channelled* fan, a *smooth* fan, and a basin plain. Many modern fans can be described by the terminology on the left (Fig. 14):

1) an inner (or upper) fan with a single channel
2) a mid fan, consisting of shallower branching channels which feed a depositional lobe (the "suprafan" of Normark, 1978)
3) a topographically smooth outer (lower) fan, which grades into
4) a basin plain.

Details can be added to this scheme; for example, the inner fan channels on some fans have prominent levees, and a flat aggradational floor with a smaller thalweg. This is typical of the inner fan channel of La Jolla Fan, where the main channel is 1 to 2 km wide, and about 140 m deep. The incised channel meanders within the main channel, and is about 200 to 300 m wide and 20 m deep. An example of a more complex fan model is given in Figure 15.

Currently, variations on this simple statement of the model (Fig. 14) include fans with prominent channel deposits but apparently no lobes, fans with abundant lobe deposits but fewer channels, and coalesced inner fan deposits that consist largely of channel-levee complexes. Because there are only one or two examples of each of these, it is probably premature to suggest many different fan *models*. It is important to understand how the existing general model(s) can be adapted to account for these new situations.

FACIES AND FAN MORPHOLOGY

Mutti and Ricci Lucchi (1972) assigned their turbidite facies to three associations - slope, fan, and basin plain. The schemes in Figures 14 and 15 are developed from this. The data base for the diagrams is a blend of ancient rock characteristics (especially grain size and observed frequency of channelling associated with the facies) and morphology of modern fans. The weakest part of the model is the relative lack of core data from modern fans.

Classical turbidites are spectacularly parallel bedded and unchannellized in the field, and hence are assigned to smooth fan environments - smooth outer parts of suprafan lobes, lower fan, and basin plain (Fig. 15). Beds change from being relatively coarse, thick-bedded and beginning with Bouma's division A to beds that are finer, thinner bedded, and beginning with divisions B and/or C with increasing distance from the ends of the distributary channels.

Massive and *pebbly sandstones* are commonly channellized in the field, and hence are assigned to channellized fan environments (Fig. 15). The finer facies occur in the distributary channels, and coarse facies occur higher on the fan surface, toward the inner fan channel.

Conglomerates, if supplied to the basin, tend to occur in the inner fan channel, or as a lag in some of the distributary channels (Fig. 15).

Major complications occur in the inner fan channel-levee complexes, where thin bedded classical turbidites (beginning with Bouma's division B or C) may occur on channel margins or levees, or in the basins behind the levees. Local facies changes may also involve finer beds on the terraces and coarser beds plugging the incised inner fan channel.

In general, these facies assignments are accepted by most workers, and they allow the development of specific facies sequences that seem to be characteristic of particular fan environments.

Figure 16
Ordovician Cloridorme Formation, Québec, showing beds slightly overturned with stratigraphic top to left. Note prominent thickening-upward sequence (arrowed), with abrupt return to thin-bedded turbidites and mudstones at the top.

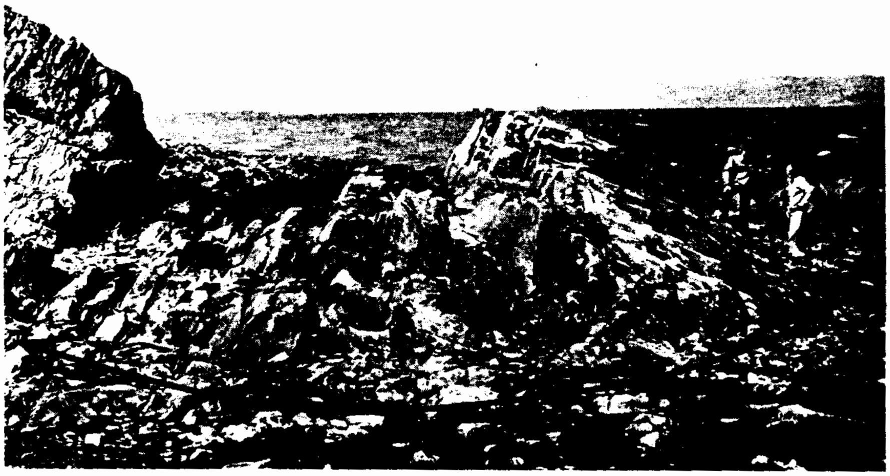

Figure 17
Cambrian St. Roch Formation near St. Jean-Port-Joli, Québec. Note prominent thinning-upward sequence (arrowed) which begins with a massive sandstone facies (compare with Fig. 18), and is interpreted as a channel fill.

FACIES SEQUENCES

Sequences are basically related to lobe progradation and aggradation, and to channel filling. The formative ideas are those of Mutti and Ghibaudo (1972), and it is now fair to state that:

1) most fan interpretations are based on facies sequences;
2) facies sequences are being used to modify fan models; and
3) some alleged facies sequences exist only in the eye of the beholder!

In depositional lobes, the marginal facies will tend to be thinner bedded than those near the apex of the lobe; consequently, if the lobe progrades it will produce a *thickening-upward sequence* (Fig. 16). As well as thickening-upward, the sequence may also become coarser grained upward, and beds that begin with Bouma's division C and B will tend to be replaced upward by beds beginning with division A. Thickening-upward sequences are very common in the geological record - prograding lobe fringes may form sequences only a few metres thick composed of relatively thin bedded turbidites, whereas an entire lobe may form a sequence typically a few tens of metres in thickness.

By contrast, *thinning-upward sequences* (Fig. 17) were interpreted by Mutti and Ghibaudo (1972) to represent gradual channel filling and abandonment. Thinning-upward sequences tend to be a few metres to about 50 m thick, and in those cases where a channel morphology can be observed in the field, the lower beds of the fill are commonly pebbly or massive sandstones, rather than classical turbidites. Many channels on modern fans are deeper than 50 m, especially on the inner fan, but systematic thinning-upward sequences more than 50 m thick are

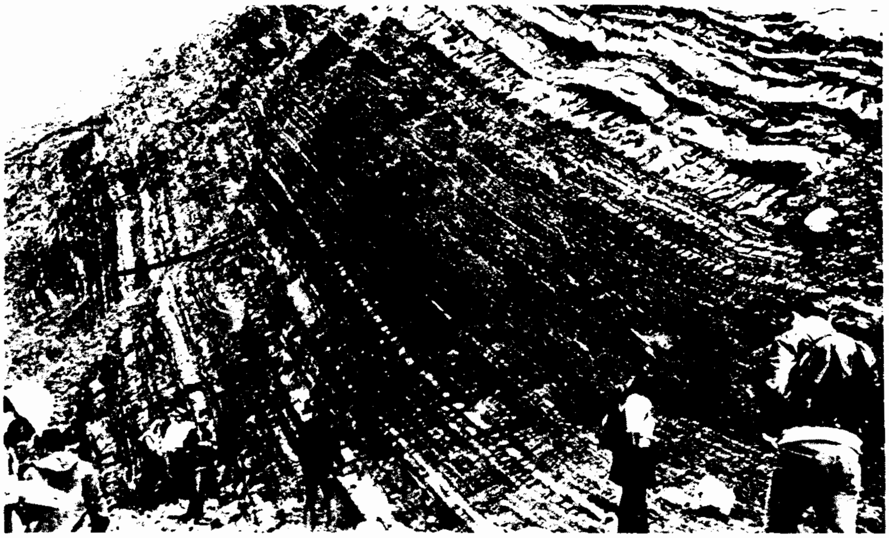

Figure 18
Thinning-upward sequence (arrowed) in the Paleocene turbidites at Shelter Cove, Point San Pedro, California. The spectacular continuity of the turbidites, and absence of a massive sandstone facies suggest that this sequence is not necessarily a channel fill. It may result from lateral lobe switching, with the thicker beds at the lobe centre, and the thinner beds representing lobe fringe. Alternatively, the turbidites may be interchannel, the thinning-upward sequence representing channel migration away from this area.

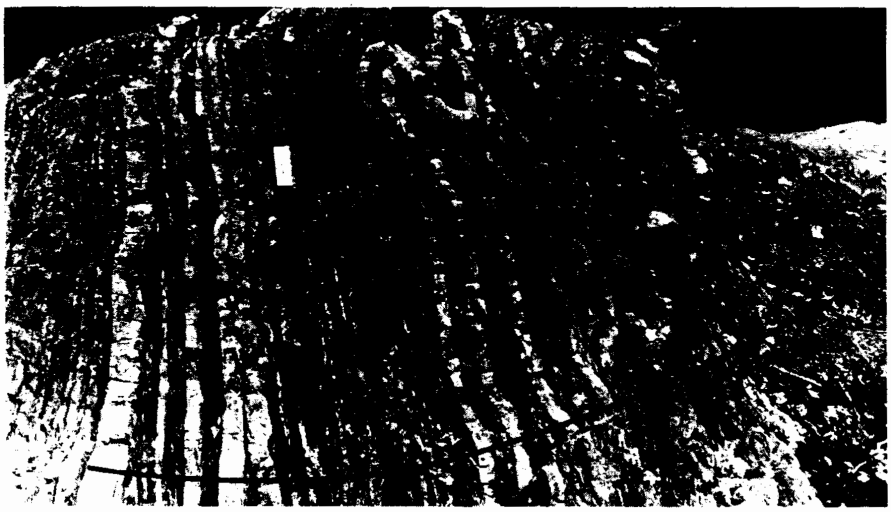

Figure 19
Thinning-upward sequence (arrowed) from turbidites, via a soft sediment slump (small arrows show way up) into mudstones. Stratigraphic top to right; Cretaceous turbidites at Wheeler Gorge, California. Compare this sequence with those in Figures 17 and 18. It probably represents neither channel filling nor lobe switching - its context, and especially the slumping, suggests deposition on a levee. The thinning-upward may reflect migration of the main turbidite channel away from the levee. See Walker (1985).

uncommon in the geological record. This suggests that deeper turbidite channels fill in a more complex way than the shallower ones. For example, well logs from the Miocene Rosedale Channel of California (Martin, 1963, Figs. 5, 6 and 7) indicate several packets of sand alternating with finer grained sediments in a channel fill up to about 400 m thick. There is certainly no suggestion of an overall fining-upward sequence. Similarly, channel-levee complexes in the Paleocene Frigg Fan of the North Sea (Heritier *et al.*, 1979, Figs. 9, 10, 12, and 13) may be 200 m or more in thickness, but are apparently compound, and not single channel fills. Relief on any single channel (floor to levee crest) is indicated to be about 60 m (Heritier *et al.*, 1979, Fig. 13).

The association of *thinning-upward sequences* with *channel filling* has become a standard part of the fan model, but it is becoming apparent that some thinning-upward sequences, especially those composed only of relatively thin-bedded classical turbidites, imply other processes. For example, the sequence shown in Figure 18 exhibits convincing thinning-upward, but the thin-bedded classical turbidites and smooth sea floor do not suggest channelling. The sequence could indicate gradual lateral lobe shifting, from a lobe centre to lobe fringe environment. Similarly, the sequence in Figure 19 also shows thinning-upward, but the soft-sediment slumping and thin-bedded turbidites could indicate deposition on the back of an inner fan levee, the sequence being due to gradual migration of the channel away from the depositional area.

If an entire fan complex were to prograde, an idealized sequence predicted by the fan model as its various lobe and channel sequences build up would be similar to that shown in Figure 20, which is largely self-explanatory. One of the most useful aspects of thickening- and thinning-upward sequences is that they can be recognized in sub-surface well logs. In Figure 21, I show one possible interpretation of an SP and resistivity log (see "Subsurface Facies Analysis", this volume) from Devonian turbidites in Pennsylvania. Again, the comparison of Figures 20 and 21 is obvious, and shows how the fan model can be used as a basis for interpretation. Many other examples are given by Walker (1978), and by Tillman and Ali (1982) who have compiled a very useful series of turbidite papers into a reprint volume of the American Association of Petroleum Geologists.

FEEDBACK: THE EVOLUTION OF FAN MODELS

In the first paper of this volume, it was emphasized that models are formed by the "distillation" of many local examples. It follows that as new examples are studied, there is more input, or feedback, into the model, and hence the possibility for the model to evolve. Currently, fan models are evolving as the result of more studies of ancient rocks, and better studies of modern fans.

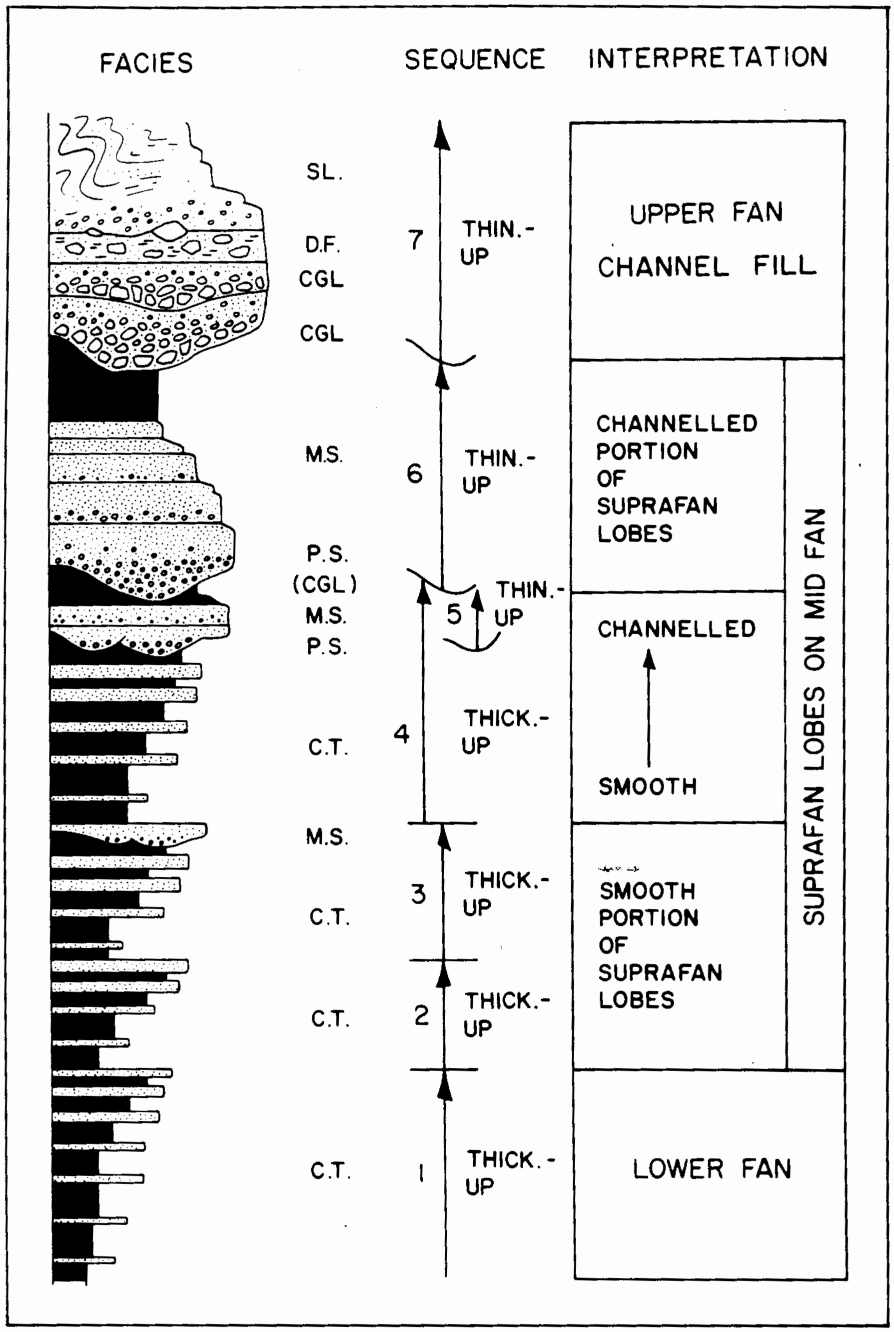

Figure 20
Generalized hypothetical sequence produced during over-all fan progradation. CT = classical turbidites; MS = massive sandstones; PS = pebbly sandstones; CGL = conglomerate; DF = debris flow; SL = slump. Sequences shown by arrows are THIN.-UP (thinning upward) or THICK.-UP (thickening upward).

The input from ancient rocks comes particularly from work on facies sequences. I mentioned earlier that some of these sequences perhaps exist only in the eye of the beholder – this is illustrated in Figure 22, and the general problem has been addressed by Hiscott (1981). One modification of the general fan model of Figure 14 is the "fan of low transport efficiency" (Mutti, 1979), in which most of the sand is deposited in channel complexes rather than on lobes. The evidence comes both from observations of channel contacts in the field, and from the abundance of thinning-upward sequences observed in units such as the Eocene Tyee Formation of Oregon (Chan and Dott, 1983) and The Rocks Sandstone of California (Link and Nilsen, 1980). A measured section of The Rocks is shown in Figure 23, where the thinning and fining-upward sequences are shown. Note that sequences average about 10 m in thickness; they mostly have erosional bases, and involve individual sandstone beds up to 7.5 m thick.

In using this example to modify existing models, note that two levels of interpretation are involved – first, the existence of the sequences themselves is an interpretation (and some of the sequences involve very few beds, and/or unconvincing thickness trends, Fig. 23), and second, it is an interpretation to suggest that the sequences necessarily involve channel filling. However, in view of the bed thickness and erosional bases, the latter interpretation seems reasonable and convincing.

A second modification of the basic fan model of Figure 14 is the recognition of channel-levee complexes, rather than a single channel with a levee. Channel-levee complexes exist on modern fans such as the Amazon (Damuth *et al.*, 1983a, 1983b), Indus, Laurentian (Stow, 1981) and Crati (Colella *et al.*, 1981), and are possibly present in ancient examples such as the Paleocene-Lower Eocene Frigg Fan (North Sea; Heritier *et al.*, 1979).

A channel-levee complex is essentially a central channel with fine grained sediment wedges on either side. The wedges stand up above the general topography of the basin floor, but gradually lose their relief away from the channel and merge with the basin floor. Most active levees appear to be constructed by spill-over from turbidity currents using the channel. Dimensions are extremely variable – on the Amazon Cone, mid-fan channel-levee complexes are up to 500 m thick and 25 km wide and those on the Laurentian Fan are a little thinner and perhaps a little wider (Stow, 1981). On a much smaller scale, the upper fan channel-levee complex on Crati Fan (Southern Italy; Colella *et al.*, 1981) is a total of 5 km wide, and is built up from at least 3 channels a little over 10 m deep.

Channel-levee complexes differ from fans of low efficiency, in that the latter are sand-rich, whereas the bulk of the channel-levee complex is relatively fine-grained. In the Amazon example, channels appear to switch abruptly by avul-

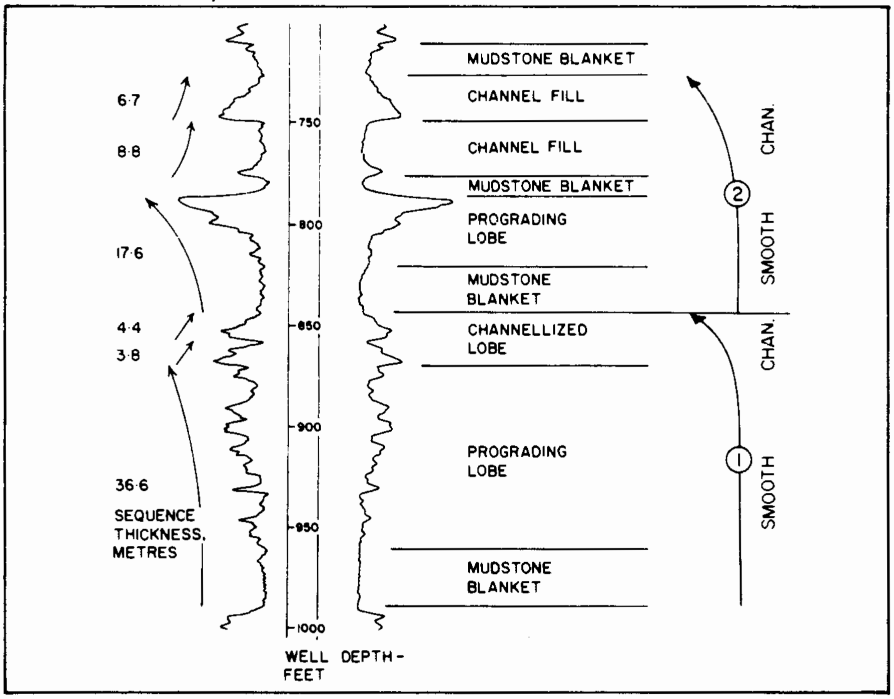

Figure 21
SP and resistivity logs from a turbidite formation in Pennsylvania. I have interpreted the shapes of the SP/resistivity trends as channel fill ("bell-shaped") or lobe progradation ("funnel-shaped"), with some mudstone blankets in between. Note that two overall lobe-to-channel sequences can be defined (see Fig. 20), the lower one having a greater proportion of lobe deposits (smooth fan surface), the upper one having a greater proportion of channel deposits.

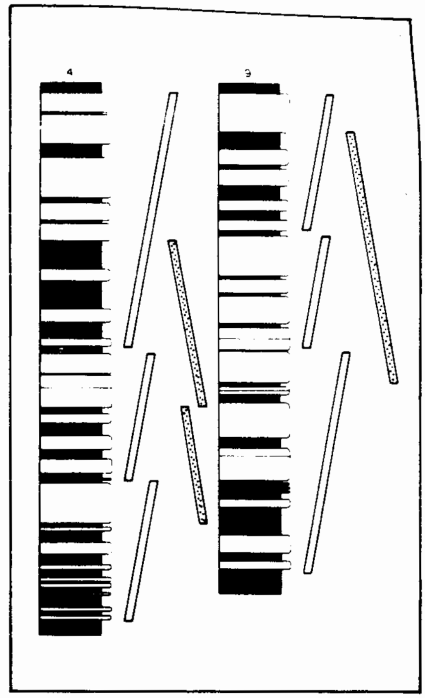

Figure 22
Some sequences in turbidites are "in the eye of the beholder". This diagram is from Mutti and Ghibaudo (1972) and shows part of their Figure 15 from the Miocene Arenarie di San Salvatore. Open bars show "negative megasequences", that is, thickening-upward sequences as proposed by Mutti and Ghibaudo. However, I have added some extra bars, stippled, which propose positive megasequences, or thinning-upward sequences. The reader should decide which sequences are preferable, remembering that one implies lobe fringe progradation, the other implies channel filling on a different part of a fan.

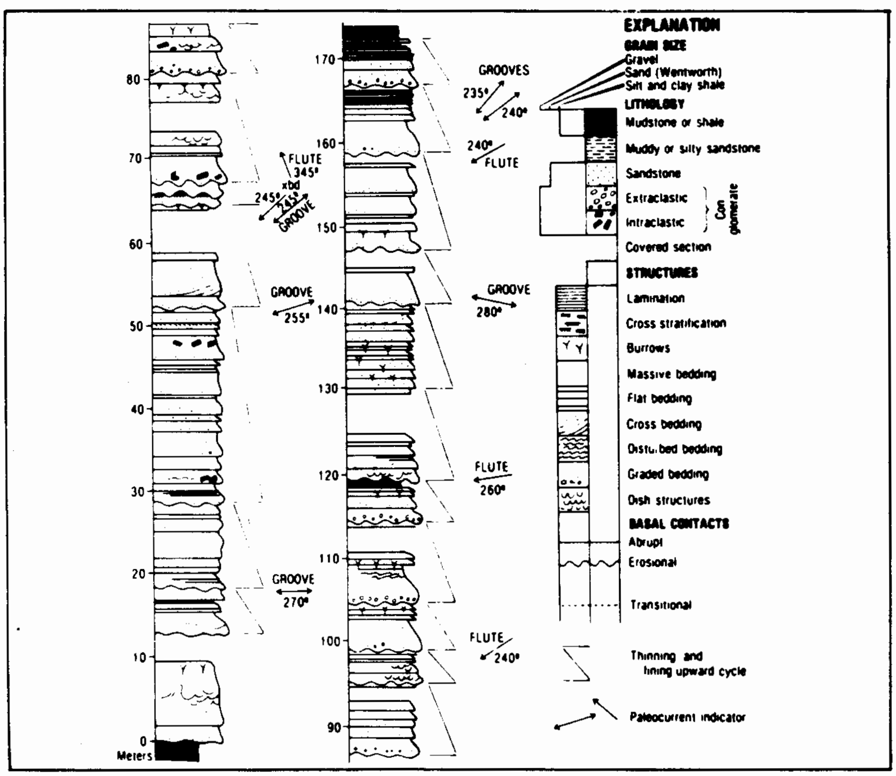

Figure 23
Measured section of The Rocks Sandstone (Eocene) in the Santa Lucia Range, California. Note the "thinning- and fining-upward cycles"; some are based on very few beds, or only show very weakly the proposed trends. Others seen well established with erosional (or channelled) bases. From Link and Nilsen, 1980.

sion (Damuth *et al.*, 1983a, 1983b), but in other fans (possibly the Indus is a good example) the channel appears to migrate progressively laterally, eroding into its older levee deposits as the whole system aggrades.

In the geological record the only well described candidate for a channel-levee complex is Frigg Fan in the Paleocene-Lower Eocene section of the North Sea (Heritier *et al.*, 1979). The fan is composed of four radiating sandy fingers apparently without well defined depositional lobes at the end. The topographically highest fingers are 2 to 4 km wide, and about 600 m thick. Within the fingers, there are interbedded channel turbidites, sandy levee deposits, and back-levee or interchannel shales. A second possible candidate is an Upper Cretaceous conglomerate channel complex at Wheeler Gorge, California (Walker, 1975b and 1985), where the conglomerates are overlain by several thinning-upward sequences of thin-bedded turbidites with abundant slump features. The vertical succession of conglomerates to thin-bedded turbidites with slumps suggests the migration of conglomerate-filled channels away from the area, with the slumped thin-bedded turbidites representing back-levee deposits of (? nearby) higher conglomerate-filled channels.

LIMITATIONS OF FAN MODELS

In 1976, when the "turbidite facies model" first appeared in Geoscience Canada, it was one of the better defined models. Many published studies in the last eight years have described systems that vary considerably from that model (low efficiency fans, channel-levee complexes), and descriptions of modern fans show variations from the summary model of Normark (1978). I believe that it is definitely premature to propose several different types of fan models - the data base for each proposed type is too scanty. It seems better to regard the "basic fan model" (Fig. 14 is but one example) as a norm, hence identifying, say, low efficiency fans as different from the norm. It is thus a fixed point for the comparison of many different fans, and can be modified to suppress the depositional lobes, and emphasize the channels, or channel-levee complexes, as appropriate. Once modified, it should still be possible to use the model in a predictive way, understanding how the modification of the model will affect channel and lobe distributions, sand body geometry, etc.

As presented, the fan model seems to be a useful framework for the investigation of small to medium scale ancient fans, with single points of input (single feeder channels) into the basin. The model loses much of its power if two separate fans overlap, because facies sequences may no longer be controlled by single processes (e.g., lobe progradation), but may be the result of irregular interbedding of beds from multiple sources.

The fan model may also lose some power when applied to long (hundreds of km) narrow "exogeosynclinal" troughs in which the paleoflow pattern is dominantly parallel to tectonic strike. Examples of turbidites in such troughs include the Middle Ordovician Cloridorme Formation (Gaspé Peninsula) and its time equivalent in the Central Appalachians, the Martinsburg Formation. The deposits consist dominantly of classical turbidites hundreds of metres thick, but showing no consistent proximal to distal change along the length of the trough in the downflow direction. It is commonly suggested that turbidity currents flowed downslope toward the trough axis, perhaps constructing fans at the trough margin. However, at the trough axis the flows turned and continued to flow parallel to the trough axis. The marginal fans were presumably destroyed by subsequent tectonics, and the absence of consistent proximal to distal changes along the trough axis is probably due to input from a whole series of fans along the trough margin. Thus any consistent changes developing from one source would be masked by input from adjacent sources up and down the trough. At present, there is no facies model that acts as a good predictor in this type of turbidite basin.

Even in short narrow elongate troughs, the fan models (Figs. 14 and 15) may be inapplicable. Hsu *et al.* (1980) have criticized the use of fan models in the Ventura Basin (California), and instead, they have emphasized the data for longitudinal flow along the axis of the basin. It follows that "the presence of most reservoir-sand beds [is] in the deepest part of the trough, not in a canyon or a fan environment on the basin flank" (Hsu *et al.*, 1980, p. 1050). This work combined with studies of small, tectonically active modern basins in the California Borderlands (e.g., Normark *et al.*, 1979; Field and Edwards, 1980; Underwood *et al.*, 1980; Graham and Bachman, 1983) emphasizes that basin geometry may greatly modify the way in which fans build into basins, adding another limitation to the use of fan facies relationships as predictors.

Recently, studies emphasizing facies and thickness sequences have been made of part of the Middle Ordovician Cloridorme Formation of Québec (Beeden, 1983). It was shown that thickening-upward sequences of classical turbidites (mostly relatively thin bedded) were present, and could be interpreted in terms of prograding lobe-fringe environments. In the same Middle Ordovician elongate trough between the rising Taconic Orogen and the craton, Belt and Bussieres (1981) have also described submarine fan facies (see their Fig. 8) in the area northeast of Québec City. Thus it may be that further studies of long narrow basins using techniques such as thickness sequence analysis will reveal more evidence of submarine fan environments.

REFERENCES

SOURCES OF INFORMATION

For a general background on some of the current arguments concerning ancient and modern fan models, the introductory papers in the reprint volume edited by Tillman and Ali (1982) are the best source. These papers include Walker (1978), Normark (1970, 1978), Nilsen (1980; a discussion of the earlier papers) and replies by Walker (1980) and Normark (1980). Another useful review is that of Howell and Normark (1982). For other aspects of models, consult the annotations of the general papers (below). As well as a general heading, I have subdivided the references into modern sediments and ancient rocks. Consult the annotations for details.

GENERAL PAPERS: MODELS

Bouma, A.H., 1962. Sedimentology of some flysch deposits. Amsterdam, Elsevier, 168 p.
This book contains a lot of useful but local data. It does, however, establish the classic Bouma sequence, and lays out some ideas on how the sequence might change laterally.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982. Structures and sequences clastic rocks. Society of Economic Paleontologists and Mineralogists, Short Course 9, variously paginated.
The introductory sections on flow, bed forms and stratification lay out the basis for interpreting sedimentary structures.

Hiscott, R.N., 1981. Deep sea fan deposits in the Macigno Formation (Middle-Upper Oligocene) of the Gordana Valley, Northern Appennines, Italy: Discussion. Journal of Sedimentary Petrology, v. 51, p. 1015-1021.
A very lucid discussion of the problems of identifying sequences in turbidites.

Howell, D.G., and Normark, W.R., 1982. Sedimentology of submarine fans. *In* Scholle, P.A., and Spearing, D.R., eds., Sandstone depositional environments. American Association of Petroleum Geologists, Memoir 31, p. 365-404.
A very useful review of all important aspects of fans, with many colour illustrations. The next paper to consult after reading this one.

Kelts, K., and Arthur, M.A., 1981. Turbidites

after ten years of deep-sea drilling - wringing out the mop? *In* Warme, J.E., Douglas, R.G., and Winterer, E.L., eds., The deep sea drilling project: a decade of progress. Society of Economic Paleontologists and Mineralogists, Special Publication 32, p. 91-127.
This is the only coherent source of information on the DSDP findings with respect to turbidites and submarine fans. It forms an invaluable entry into the vast library of DSDP reports. Read Walker (1973) to understand the significance of the title.

Kuenen, P.H., and Migliorini, C., 1950. Turbidity currents as a cause of graded bedding. Journal of Geology, v. 58, p. 91-127.
This paper established the turbidite concept, and is one of the most important papers in sedimentology this century. Required reading, not only for historical reasons.

Lowe, D.R., 1975. Water escape structures in coarse grained sediments. Sedimentology, v. 22, p. 157-204.
A thorough and very abundantly illustrated discussion of these structures, which are common in many turbidite facies.

Menard, H.W., 1964. Marine geology of the Pacific. New York, McGraw Hill, 271 p.
Chapter 9 on turbidity currents is still a useful review, and contains important quantitative data.

Mutti, E., 1979. Turbidites et cones sous-marins profonds. *In* Homewood, P., ed., Sedimentation detritique (fluviatile, littorale et marine). Institut de Geologie de l'University de Fribourg, Short Course 1979, p. 353-419.
Excellent review (in French) of submarine fan facies, facies sequences and facies models. Full discussion of channel mouth bars and by-passing, and introduces the idea of fans of "low and high transport efficiency". Required reading - a different point of view from Walker (1978).

Mutti, E., and Ghibaudo, G., 1972. Un esempio di torbiditi di conoide sottomarina esterna: le Arenarie di San Salvatore (Formazione di Bobbio, Miocene) nell'Appennino de Piacenza. Memorie dell'Accademia delle Scienze di Torino, Classe di Scienze Fisiche, Matematiche e Naturali, Series 4, No. 16, 40 p.
As well as describing the facies, this paper emphasizes facies sequences, and for the first time compares turbidite fining-up sequences with delta channels, and coarsening-up sequences with prograding deltaic lobes. Suggests a fan model with lobes at the ends of channels (see Mutti and Ricci Lucchi, 1972, below).

Mutti, E. and Ricci Lucchi, F., 1972. Le torbiditi dell'Appennino settentrionale: introduzione all'analisi di facies. Memorie dell Societa Geologica Italiana, v. 11, p. 161-199. An extremely important and influential paper that established a widely-used facies classification, grouped the facies into associations, and related the associations to fan depositional environments. Did not discuss sequences in as much detail as Mutti and Ghibaudo (1972), above, and proposed a fan model with channels but no lobes.
REQUIRED READING - fortunately, there is an English translation by T.H. Nilsen, 1978. Turbidites of the northern Appennines: introduction to facies analysis. International Geology Review, v. 20, p. 125-166.

Mutti, E., and Ricci Lucchi, F., 1975. Examples of turbidite facies and facies associations from selected formations of the northern Appennines. Nice, France, 9th International Congress of Sedimentology, Guidebook to Field Trip A 11 (Mutti, E., *et al.*, eds.), 120 p.

Nilsen, T.H., 1980. Modern and ancient submarine fans: discussion of papers by R.G. Walker and W.R. Normark. American Association of Petroleum Geologists, Bulletin, v. 64, p. 1094-1101.
Discussion of Walker (1978) and Normark (1978). An important paper which establishes many of the current topics of research and disagreement.

Normark, W.R., 1970. Growth patterns of deep sea fans. American Association of Petroleum Geologists, Bulletin, v. 54, p. 2170-2195.
This is the first generalization about fan morphology and growth patterns, now superceded by Normark, 1978.

Normark, W.R., 1978. Fan-valleys, channels and depositional lobes on modern submarine fans: characters for the recognition of sandy turbidite environments. American Association of Petroluem Geologists, Bulletin, v. 61, p. 912-931.
Summary of growth patterns of many modern fans, and still the best general summary. See discussion by Nilsen (1980) and Normark's reply (1980).

Normark, W.R., 1980. Modern and ancient submarine fans: reply. American Association of Petroleum Geologists, Bulletin, v. 64, p. 1108-1112.
Here, Normark replies to the discussion of his 1978 paper by Nilsen (above).

Tillman, R.W., and Ali, S.A., 1982. Deep water canyons, fans and facies: models for stratigraphic trap exploration. American Association of Petroleum Geologists, Reprint Series 26, 596 p.
A very useful collection of papers from the AAPG Bulletin. Pages 1-100 usefully group the papers by Walker and Normark, with Nilsen's discussion and authors' replies.

Walker, R.G., 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. Journal of Sedimentary Petrology, v. 37, p. 25-43.
Uses the Bouma sequence as starting point for an investigation of turbidite variablility. The proximal/distal ideas show how the sequence can be used as a predictor, but being "pre-fan", the predictions are now out of date.

Walker, R.G., 1973. Mopping-up the turbidite mess. *In* Ginsburg, R.N., ed., Evolving concepts in sedimentology. Baltimore, The John Hopkins University Press, p. 1-37.
A philosophical history of the turbidite concept, based on the ideas of Thomas H. Kuhn. Read the paper to understand the title!

Walker, R.G., 1975a. Generalized facies models for resedimented conglomerates of turbidite association. Geological Society of America, Bulletin, v. 86, p. 737-748.
Established four "Bouma-like" sequences for conglomerates - the scheme has been little-modified during the last 10 years but still has a rather slim data base.

Walker, R.G., 1978. Deep water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps.. American Association of Petroleum Geologists, Bulletin, v. 62, p. 932-966.
Review of facies, facies sequences and facies models. A very useful source of references, and ideas current to 1978. There is now a greater variety of modern fans and ancient rock studies that need to be blended or distilled into a general model.

Walker, R.G., 1980. Modern and ancient submarine fans: reply. American Association of Petroleum Geologists, v. 64, p. 1101-1108.
Reply to Nilsen's (1980) discussion of Walker (1978).

Walker, R.G. and Mutti, E., 1973. Turbidite facies and facies associations. *In* Middleton, G.V., and Bouma, A.H. eds., Turbidites and deep water sedimentation. Pacific Section, Society of Economic Paleontologists and Mineralogists, Short Course (Anaheim, 1973), p. 119-157.
This is the first attempt to blend modern fan studies (Normark, 1970) and ancient rock studies (Mutti and Ricci Lucchi, 1972) into a combined recent/ancient fan model. The evolution of this model has been discussed by Walker (1980).

RECENT SEDIMENTS

Colella, A., *et al.*, 1981. The Crati submarine fan, Ionian Sea. A preliminary report. International Association of Sedimentologists, Second European Meeting, Bologna, Abstracts, p. 34-39
Brief description of a complex of channels

and levees, passing basinward into elongate lobes and interlobe areas.

Damuth, J.E., and Embley, R.W., 1981. Mass-transport processes on Amazon Cone: Western Equatorial Atlantic. American Association of Petroleum Geologists, Bulletin, v. 65, p. 629-643.
General description of the Amazon Cone, emphasizing three huge slump/debris flow deposits.

Damuth, J.E. *et al.*, 1983a. Distributary channel meandering and bifurcation patterns on the Amazon deep sea fan as revealed by long-range side-scan sonar (GLORIA). Geology, v. 11, p. 94-98.
Beautiful and amazing side-scan sonar pictures of extremely sinuous channels within channel-levee complexes.

Damuth, J.E., *et al.*, 1983b. Age relationship of distributary channels on Amazon deep sea fan: implications for fan growth pattern. Geology, v. 11, p. 470-473.
Detailed description of the evolution of western and eastern channel-levee complexes.

Elmore, R.D., Pilkey, O.H., Cleary, W.J., and Curran, H.A., 1979. Black Shell turbidite, Hatteras abyssal plain, western Atlantic Ocean. Geological Society of America, Bulletin, v. 90, p 1165-1176.
Detailed description of the world's largest known turbidite, and its relationship to other Hatteras abyssal plain sediments.

Field, M.E., and Edwards, B.D., 1980. Slopes of the southern California continental borderland: a regime of mass transport. *In* Field, M.E., *et al.*, eds., Quaternary depositional environments of the Pacific coast. Pacific Coast Paleogeography Symposium 4, Los Angeles, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 169-184.
Shows the influence of basin topography on the way in which basins are filled.

Graham, S.A. and Bachman, S.B., 1983. Structural controls on submarine fan geometry and internal architecture: upper La Jolla Fan system, offshore southern California. American Association of Petroleum Geologists, Bulletin, v. 676, p. 83-96.
Shows the La Jolla Fan feeder channels (Newport, La Jolla, Loma) and the structural control of their evolution. Channel-levee complexes are well developed in the Newport system.

Heezen, B.C., 1956. Corrientes de turbidez del Rio Magdalena. Societa Geografica de Colombia, Boll., v. 52-2, p. 135-143.
This paper, quoted by Menard (1964), is apparently the data source for the Magdalena turbidity current of 1935. See Menard (1964) and Kolla *et al.* (1984).

Heezen, B.C., Menzies, R.J., Schneider, E.D., Ewing, W.M., and Granelli, N.C.L., 1964. Congo submarine canyon. American Association of Petroleum Geologists, Bulletin, v. 48, p. 1126-1149.
An important paper that establishes a direct relationship between turbidity current generation (cable breaking) and the behaviour of the Congo river – specifically *months* of high discharge and *years* when the river is establishing a new path through its estuarine sand bars. The paper is reprinted in Tillman and Ali (1982).

Horn, D.R., Ewing, J.I., and Ewing, M., 1972. Graded bed sequences emplaced by turbidity currents north of 20° N in the Pacific, Atlantic and Mediterranean. Sedimentology, v. 18, p. 247-275.
Describes abyssal plains, their sediments and sediment thicknesses, and points of input into the basins.

Kolla, V., Buffler, R.T., and Ladd, J.W., 1984. Seismic stratigraphy and sedimentation of Magdalena Fan, southern Colombia Basin, Caribbean Sea. American Association of Petroleum Geologists, Bulletin, v. 68, p. 316-332.
Establishes six seismic sequences, and models channellized and overbank turbidity current flow (upper fan) and unchannellized flow (down fan). A possible modern example of a "high efficiency" fan (see Mutti, 1979).

Kolla, V., Kostecki, J.A., Henderson, L., and Hess, L., 1980. Morphology and Quaternary sedimentation of the Mozambique Fan and environs, southwestern Indian Ocean. Sedimentology, v. 27, p. 357-378.
General description of Mozambique Fan.

Normark, W.R., Piper, D.J.W., and Hess, G.R., 1979. Distributary channels, sand lobes and mesotopography of Navy submarine fan, California Borderlands, with applications to ancient fan sediments. Sedimentology, v. 26, p. 749-774.
Establishes in detail the pattern of lobe/channel switching and sediment distribution. Contributes an important data base to fan models of the Normark type.

Normark, W.R., Piper, D.J.W., and Stow, D.A.V., 1983. Quaternary development of channels, levees and lobes on middle Laurentian Fan. American Association of Petroleum Geologists, Bulletin, v. 67, p. 1400-1409.
Establishes a pattern of channel switching and levee growth.

Pilkey, O.H., Locker, S.D., and Cleary, W.J., 1980. Comparison of sand-layer geometry on flat floors of 10 modern depositional basins. American Association of Petroleum Geologists, Bulletin, v. 64, p. 841-856.
Important and useful comparison of turbidite geometries on Atlantic and Caribbean abyssal plains (plus Santa Monica basin).

Piper, D.J.W., and Normark, W.R., 1982. Acoustic interpretation of Quaternary sedimentation and erosion on the channelled upper Laurentian fan. Atlantic margin of Canada. Canadian Journal of Earth Sciences, v. 19, p. 1974-1984.
Shows that previously suggested large slumps are in fact absent, but establishes a channel switching pattern.

Piper, D.J.W., and Normark, W.R., 1983. Turbidite depositional patterns and flow characteristics, Navy submarine fan, California Borderlands. Sedimentology, v. 30, p. 681-694.
Correlates beds on the fan, and introduces the concept of "flow stripping", whereby most of a flow can overtop a levee at a channel bend.

Stevenson, A.J., Scholl, D.W., and Vallier, T.L., 1983. Tectonic and geologic implications of the Zodiac Fan, Aleutian Abyssal Plain, northeast Pacific. Geological Society of America, Bulletin, v. 94, p. 259-273.
A large fan that differs from many others in that the channels continue essentially to the margins of the fan. Well developed channel-levee complexes.

Stow, D.A.V., 1981. Laurentian Fan: morphology, sediments, processes and growth patterns. American Association of Petroleum Geologists, Bulletin, v. 65, p. 375-393.
A good general description and discussion of all of the above topics.

Uchupi, E., and Austin, J.A., 1979. The stratigraphy and structure of the Laurentian Cone region. Canadian Journal of Earth Sciences, v. 16, p. 1726-1752.
Seismic reflection profiling shows two fan megasequences. Re-survey of the fan gives better definition of channel patterns, and allowed re-calculation of Grand Bands turbidity current velocities.

Underwood, M.B., Bachman, S.B., and Schweller, W.J., 1980. Sedimentary processes and facies associations within trench and trench-slope settings. *In* Field, M.E., *et al.*, eds., Quaternary depositional environments of the Pacific Coast. Pacific Coast Paleogeography Symposium 4. Los Angeles, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 211-229.
Describes and models the tectonic control of turbidite facies and channels in trench settings.

Whitmarsh, R.B., *et al.*, 1974. Initial reports of the Deep Sea Drilling Project, Volume 23. Washington, D.C., U.S. Government Printing Office, p. 211-289.
Documents DSDP Site 222, with comments on rates of sedimentation for this site (on the western margin of the Indus Cone).

ANCIENT ROCKS

Beeden. D.R., 1983. Sedimentology of some turbidites and related rocks from the Cloridorme Group, Ordovician, Québec. Hamilton, Ontario, McMaster University, M.Sc. Thesis, 256 p.
Documents sequences in turbidites, the turbidites occurring in a long narrow trough with all paleocurrent directions parallel to regional strike. Possible submarine fan facies in a long narrow trough.

Belt, E.S. and Bussieres, L., 1981. Upper Middle Ordovician submarine fans and associated facies, northeast of Québec City. Canadian Journal of Earth Sciences, v. 18, p. 981-994.
Documents submarine fans in a very long narrow basin between the Taconic Orogen and the Craton. Possibly a similar situation to that of Beeden (1983).

Chan, M.A. and Dott, R.H. Jr., 1983. Shelf and deep sea sedimentation in Eocene forearc basin, western Oregon - fan or non-fan? American Association of Petroleum Geologists, Bulletin, v. 67, p. 2100-2116.
Proposes a sand-rich (or "poorly efficient") fan interpretation for the Eocene Tyee Formation.

Clari, P. and Ghibaudo, G., 1979. Multiple slump scars in the Tortonian type area (Piedmont Basin, northwestern Italy). Sedimentology, v. 26, p. 719-730.
Excellent documentation of large slump scars, interpreted to have formed near the shelf-slope break.

Cook, H.E., McDaniel, P.N., Mountjoy, E., and Pray, L.C., 1972. Allochthonous carbonate debris flows at Devonian bank ("reef") margins, Alberta, Canada. Bulletin of Canadian Petroleum Geology, v. 20, p. 439-497.
Well documented example of huge debris flows.

Crowell, J.C., 1957. Origin of pebbly mudstones. Geological Society of America, Bulletin, v. 68, p. 993-1009.
The title says it all. For the Pigeon Point examples, see also Howell and Joyce (1981).

Davies, I.C. and Walker, R.G., 1974. Transport and depostion of resedimented conglomerates: the Cap Enragé Formation, Cambro-Ordovocian, Gaspe, Québec. Journal of Sedimentary Petrology, v. 44, p. 1200-1216.
Establishes conglomerate fabric types and uses the results to document paleoflow directions.

Gregory, M.R., 1969. Sedimentary features and penecontemporaneous slumping in the Waitemata Group, Whangaparaoa Peninsula, north Auckland, New Zealand. New Zealand Journal of Geology and Geophysics, v. 12, p. 248-282.
Contains excellent illustrations of some of the largest sedimentary slumps ever described.

Hein, F.J., 1982. Depositional mechanisms of deep sea coarse clastic sediments, Cap Enragé Formation, Québec. Canadian Journal of Earth Sciences, v. 19, p. 267-287.
Describes internal structures and structure sequences for massive and pebbly sandstones.

Hein, F.J., and Walker R.G., 1982. The Cambro-Ordovician Cap Enragé Formation, Québec, Canada: conglomeratic deposits of a braided submarine channel with terraces. Sedimentology, v. 29, p. 309-329.
Conglomerate facies sequences, both lateral and vertical, establish the nature of this deep water braided channel.

Helwig, J., 1970. Slump folds and early structures, northeastern Newfoundland, Appalachians. Journal of Geology, v. 78, p. 172-187.

Hendry, H.E., 1978. Cap des Rosiers Formation at Grosses Roches, Québec - deposits of the mid-fan region on an Ordovician submarine fan. Canadian Journal of Earth Sciences, v. 15, p. 1472-1488
Conglomerate-to-sandstone fining-upward sequences interpreted as backfilling a channel mouth in the mid-fan region.

Heritier, F.E., Lossel, P., and Wathne, E., 1979. Frigg Field - large submarine-fan trap in lower Eocene rocks of North Sea. American Association of Petroleum Geologists, Bulletin, v. 63, p. 1999-2020.
Subsurface data seems to indicate a fan made up of channel levee complexes without prominent smooth depositonal lobes. The authors supply little of their own fan interpretation, however. Reprinted in Tillman and Ali, 1982.

Howell, D.G., and Joyce, J.M., 1981. Field guide to the Upper Cretatceous Pigeon Point Formation. *In* Frizzell, V., ed., Upper Cretaceous and Paleocene turbidites, central California Coast. Los Angeles, Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 61-70
Describes the classic section where Crowell (1957) first studied and illustrated pebbly mudstones.

Hsu, K.J., Kelts, K., and Valentine, J.W., 1980. Resedimented facies in Ventura Basin, California, and model of longitudinal transport of turbidity currents. American Association of Petroleum Geologists, Bulletin, v. 64, p. 1034-1051.

Johnson, B.A., and Walker, R.G., 1979. Paleocurrents and depositional environments of deep water conglomerates in the Cambro-Orodovician Cap Enragé Formation, Québec Appalachians. Canadian Journal of Earth Sciences, v. 16, p. 1375-1387.
Establishes facies relationships among different types of conglomerate, and suggests initial channel model - see Hein and Walker, 1982.

Link, M.H., Squires, R.L., and Colburn, I.P., eds., 1981. Simi Hills Cretaceous turbidites, southern California. Los Angeles, Pacific Section, Society of Economic Paleontologists and Mineralogists, Field Trip Guide Book, 134 p.
Extremely useful guide book to a superbly exposed and very interesting group of turbidites, and massive and pebbly sandstones.

Link, M.H., and Nilsen, T.H., 1980. The Rocks Sandstone, an Eocene sand-rich deep-sea fan deposit, northern Santa Lucia Range, California. Journal of Sedimentary Petrology, v. 50, p. 583-601.
Sand-rich (channellized) fan based on abundant thinning- and fining-upward sequences.

Link, M.H., and Welton, J.E., 1982. Sedimentology and reservoir potential of Matilija Sandstone: an Eocene sand-rich deep-sea fan and shallow marine complex, southern California. American Association of Petroleum Geologists, Bulletin, v. 66, p. 1514-1534.
Well-documented example of a very sandy fan. Slope facies between fan and shallow marine parts of the section are very enigmatic.

Martin, B.D., 1963. Rosedale Channel - evidence for late Miocene submarine erosion in Great Valley of California. American Association of Petroleum Geologists, Bulletin, v. 47, p. 441-456.
Subsurface example of a partly sand-filled canyon. Reprinted in Tillman and Ali (1982).

Nilsen, T.H., and Abbott, P.L., 1981. Paleogeography and sedimentology of Upper Cretaceous turbidites, San Diego, California. American Association of Petroleum Geologists, Bulletin, v. 65, p. 1256-1284.
Well described and abundantly illustrated forearc basin fan deposits.

Piper, D.J.W., Normark, W.R., and Ingle, J.R., Jr., 1976. The Rio Dell Formation: a Plio-Pleistocene basin slope deposit in northern California. Sedimentology, v. 23, p. 309-328.
Describes slope deposits, with sketches (but no photos) of slumps.

Srivastava, P., Stearn, C.W., and Mountjoy, E.W., 1972. A Devonian megabreccia at the margin of the Ancient Wall carbonate complex, Alberta. Bulletin of Canadian

Petroleum Geology, v. 20, p. 412-438.
Good descriptions of debris flow/rockfall? megabreccias.

Stevens, R.K., 1970. Cambro-Orodovician flysch sedimentation and tectonics in west Newfoundland and their possible bearing on a Proto-Atlantic ocean. *In* Lajoie, J., ed., Flysh sedimentology in North America. Geological Association of Canada, Special Paper 7, p. 165-177.
Some sedimentological description, but the thrust of the paper is concerned with allochthonous flysch complexes.

Sullwold, H.H., 1960. Tarzana Fan, deep submarine fan of late Miocene age, Los Angeles County, California. American Association of Petroleum Geologists, Bulletin, v. 44, p. 433-457.
I believe this is the first interpretation of a turbidite sequence in terms of submarine fans. It is based on a radial spread of flow directions which, if traced upslope, converge at the suggested foot of a canyon. See next paper.

Walker, R.G., 1966. Shale Grit and Grindslow Shales: transition from turbidite to shallow water sediments in the Upper Carboniferous of northern England. Journal of Sedimentary Petrology, v. 36, p. 90-114.
A submarine fan interpretation based on variability of flow directions, facies relationships and, particularly, the abundance of channels. Interpretive block diagram redrawn in Walker (1978).

Walker R.G., 1975a. Generalized facies models for resedimented conglomerates of turbidite association. Geological Society of America, Bulletin, v. 86, p. 737-748.
Four "Bouma-like" models are established, with comments on transport mechanisms and proximal-distal implications.

Walker, R.G., 1975b. Upper Cretaceous resedimented conglomerates at Wheeler Gorge, California: description and field guide. Journal of Sedimentary Petrology, v. 45, p. 105-112.
Establishes three major thinning-upward sequences. This study is put in context by Walker (in press).

Walker, R.G., 1977. Deposition of upper Mesozoic resedimented conglomerates and associated turbidites in south western Oregon. Geological Society of America, Bulletin, v. 88, p. 273-285.
Establishes a series of thinning-upward sequences, and proposes a mechanism for channel blocking leading to channel fill.

Walker, R.G., 1979. Stop 1. L'Islet Wharf: an early Cambrian submarine channel complex. *In* Middleton, G.V., *et al.*, eds., Cambro-Ordovician submarine channels and fans, L'Islet to Sainte-Anne-des-Monts, Québec. Geological Association of Canada, Guidebook to Field Trip A-6 (Québec, 1979), p. 4-7.
Preliminary interpretation of a superbly exposed deep submarine fan channel.

Walker, R.G., 1985. Mudstones and thin bedded turbidites associated with the Upper Cretaceous Wheeler Gorge conglomerates, California: a possible channel-levee complex. Journal of Sedimentary Petrology, v. 55, p. 279-290.
Interprets mudstones below conglomerates as basin plain, and mudstones with turbidites above the conglomerates as levee.

Note: added to second printing, January 1986. A very useful new book has appeared, with descriptions of modern fans and ancient turbidites. References are up to date, and fans are described in a standardized way. Required reading: the reference is—Bouma, A.H., Normark, W.R. and Barnes, N.E., eds., 1985. Submarine fans and related turbidite systems. New York, Springer Verlag, 351 p.