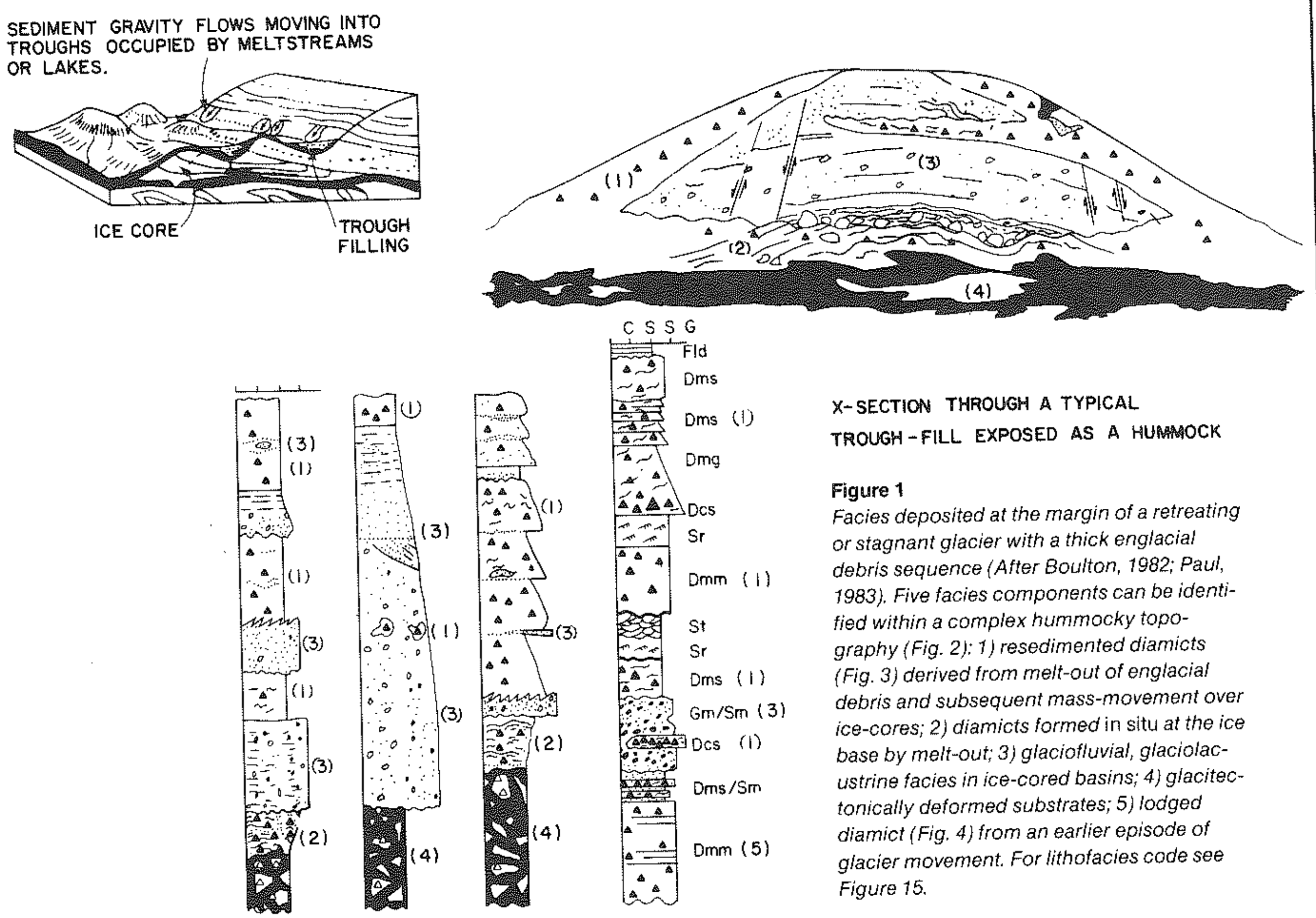

Figure 1
Facies deposited at the margin of a retreating or stagnant glacier with a thick englacial debris sequence (After Boulton, 1982; Paul, 1983). Five facies components can be identified within a complex hummocky topography (Fig. 2): 1) resedimented diamicts (Fig. 3) derived from melt-out of englacial 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) glacitectonically 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

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

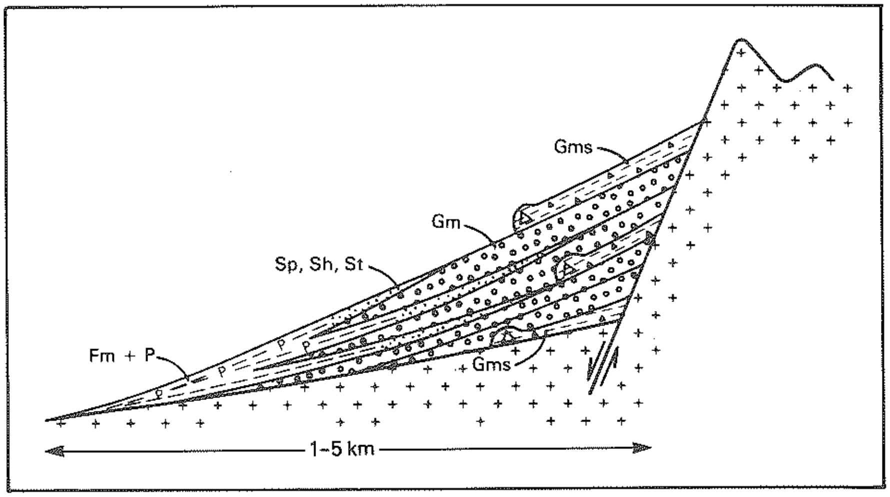

Figure 3
Diagrammatic cross-section of an alluvial fan, showing proximal-distal facies variation. See Table 1 and text for explanation of facies codes.

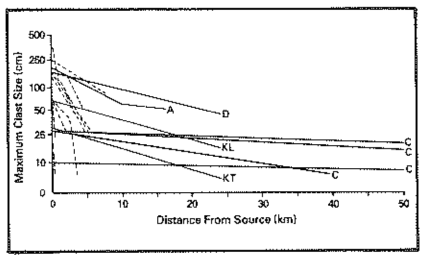

Figure 4
Variation in maximum grain size (i.e., mean of ten largest clasts at each site) versus distance downslope for various alluvial gravels. Modified after Figure 11 of Wilson (1970) and Figure 19 of Schultheis and Mountjoy (1978). Dashed lines are trends on alluvial fans, solid lines are trends 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.

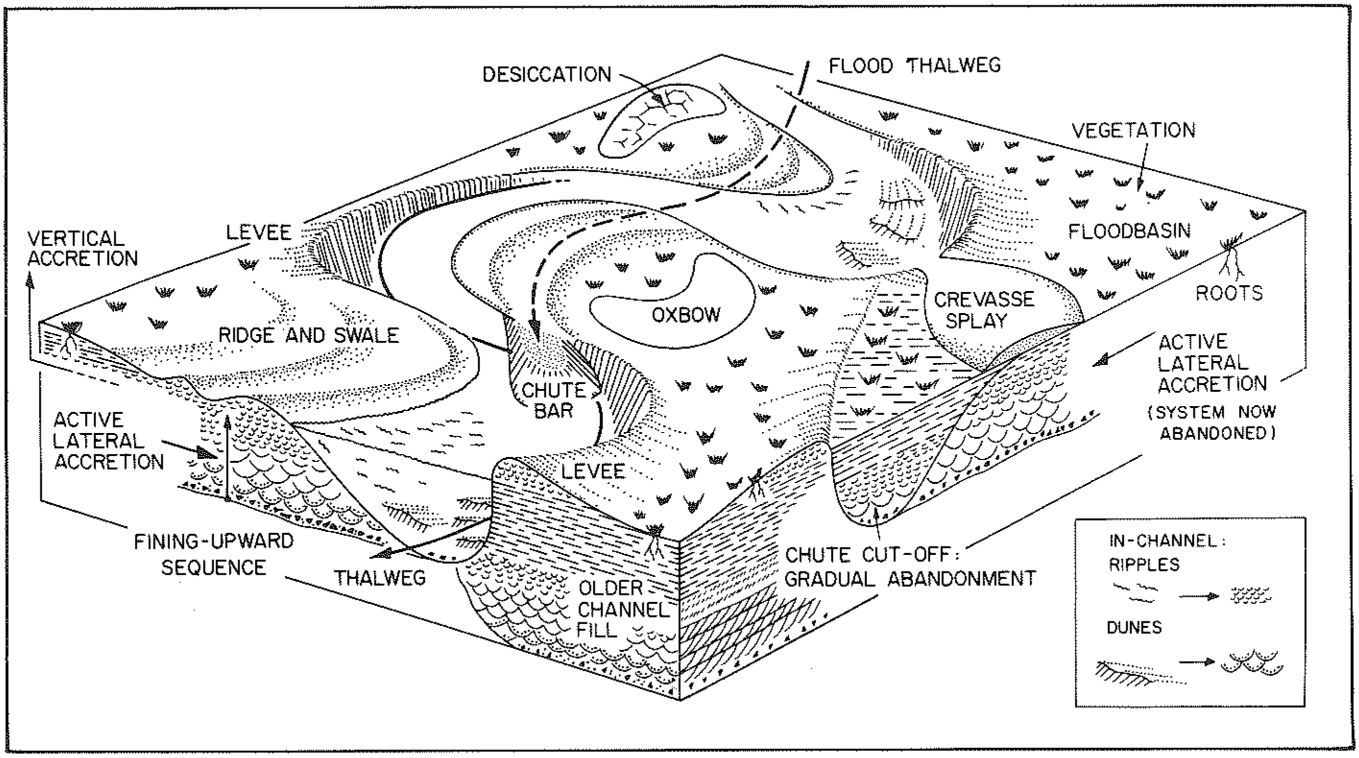

Figure 1
Block diagram showing morphological elements of a meandering river system. Erosion on the outside bend of a meander loop leads to lateral accretion on the opposite point bar. The dunes and ripples in the channel give rise to trough cross bedding and ripple cross lamination respectively (inset, 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

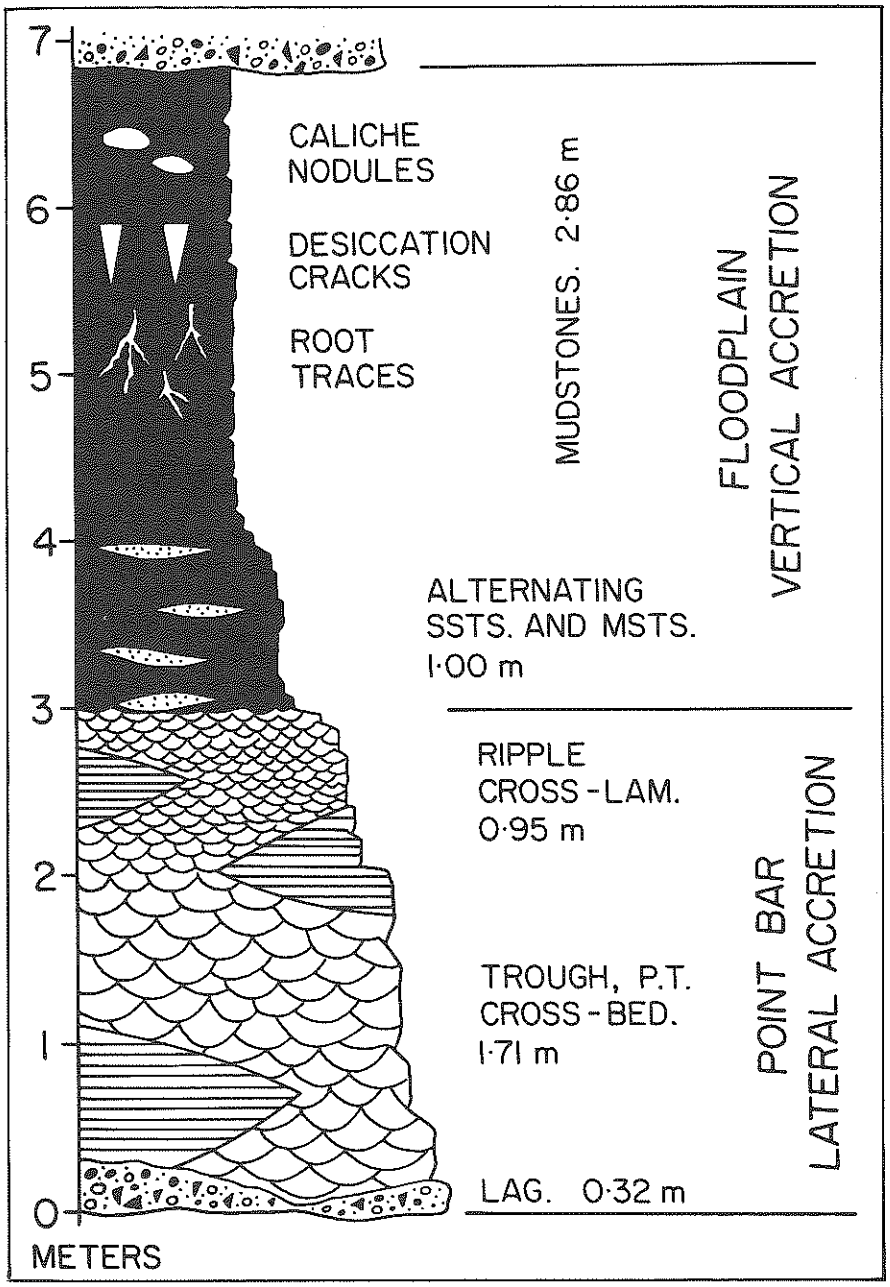

Figure 2
Model for lateral and vertical accretion deposits of meandering rivers. Data on facies sequence and fining upwards cycles summarized here are from the Devonian Old Red Sandstone of Britain and the Catskill rocks of the eastern U.S.A. (Allen, 1970). The average lateral accretion deposit is 2.98 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.

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.

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.

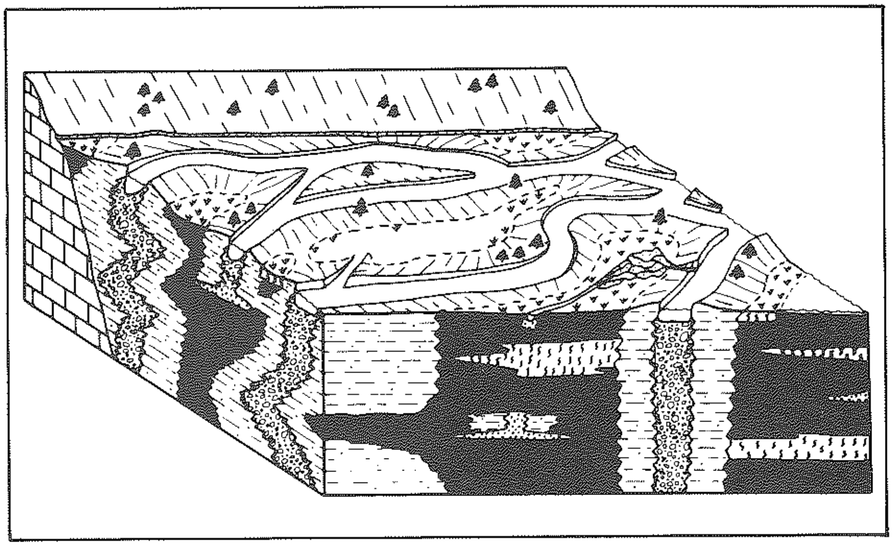

Figure 23
Block diagram of an anastomosed river, from Smith and Smith (1980). In this typical reach, channel sediments (gravel symbol) are bounded by sandy silt (dash, dot symbol) of the levees, which in turn grade into muds and silty muds of the wetlands (black). Peats are shown by small vertical wiggles. 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.

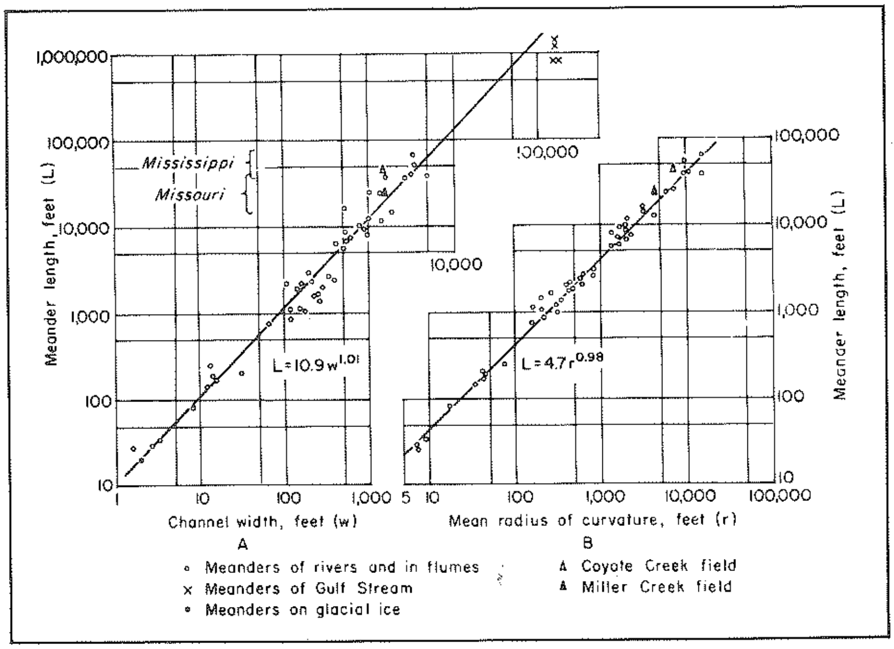

Figure 24
Relationships between meander length, channel width and meander loop radius of curvature. Original data for modern meandering streams from Leopold and Wolman (1960), with Coyote Creek and Miller Creek data plotted by Berg (1968). Note that both Coyote Creek and Miller Creek lie about as far 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).

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.

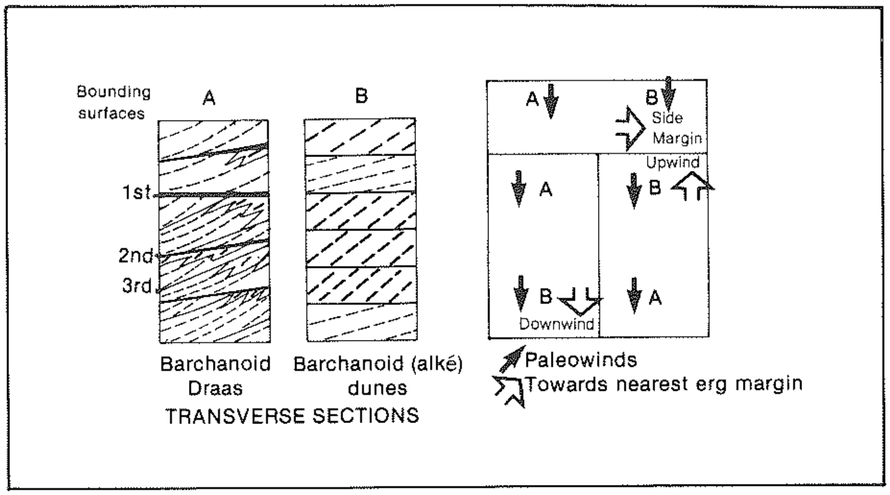

Figure 14
Two synthetic sections (A,B) (on left): inferred nearest erg margins based on locations of sections in relation to resultant wind direction (on right).

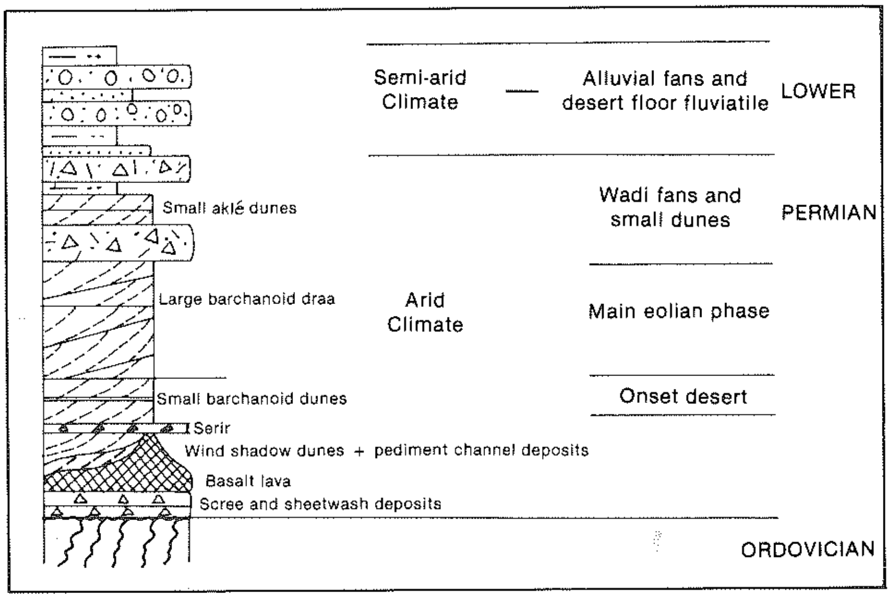

Figure 15
Section through Lower Permian deposits of the Thornhill and Dumfries intermontane basins, with interpretation of lithologies.

ACKNOWLEDGEMENTS

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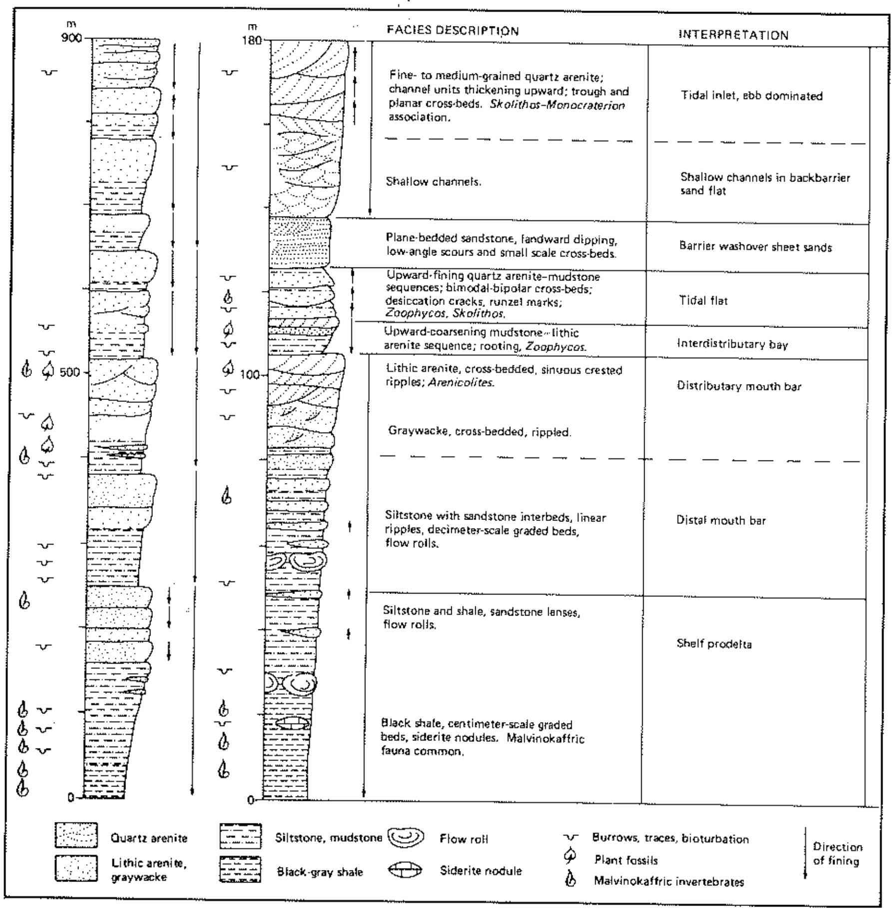


Figure 14
Stratigraphic section through the Bokkeveld Group, Cape Province, South Africa (left) and detailed lithofacies and interpretation of an idealized cyclic sequence (Tankard and Barwis, 1982).

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 cross-bedding. 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 pre-eminent 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

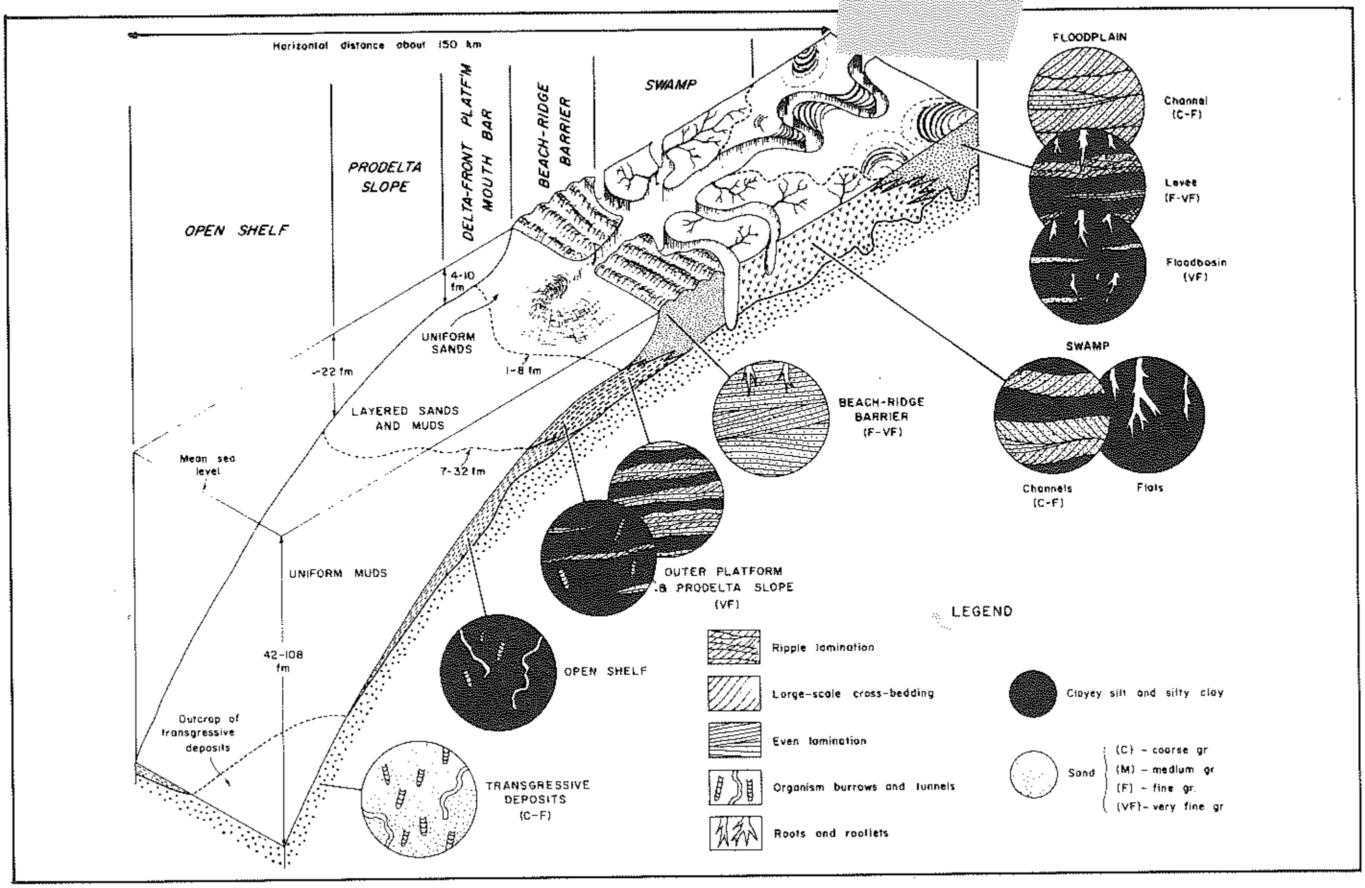


Figure 15
Block diagram model of the modern Niger delta (Allen, 1970).

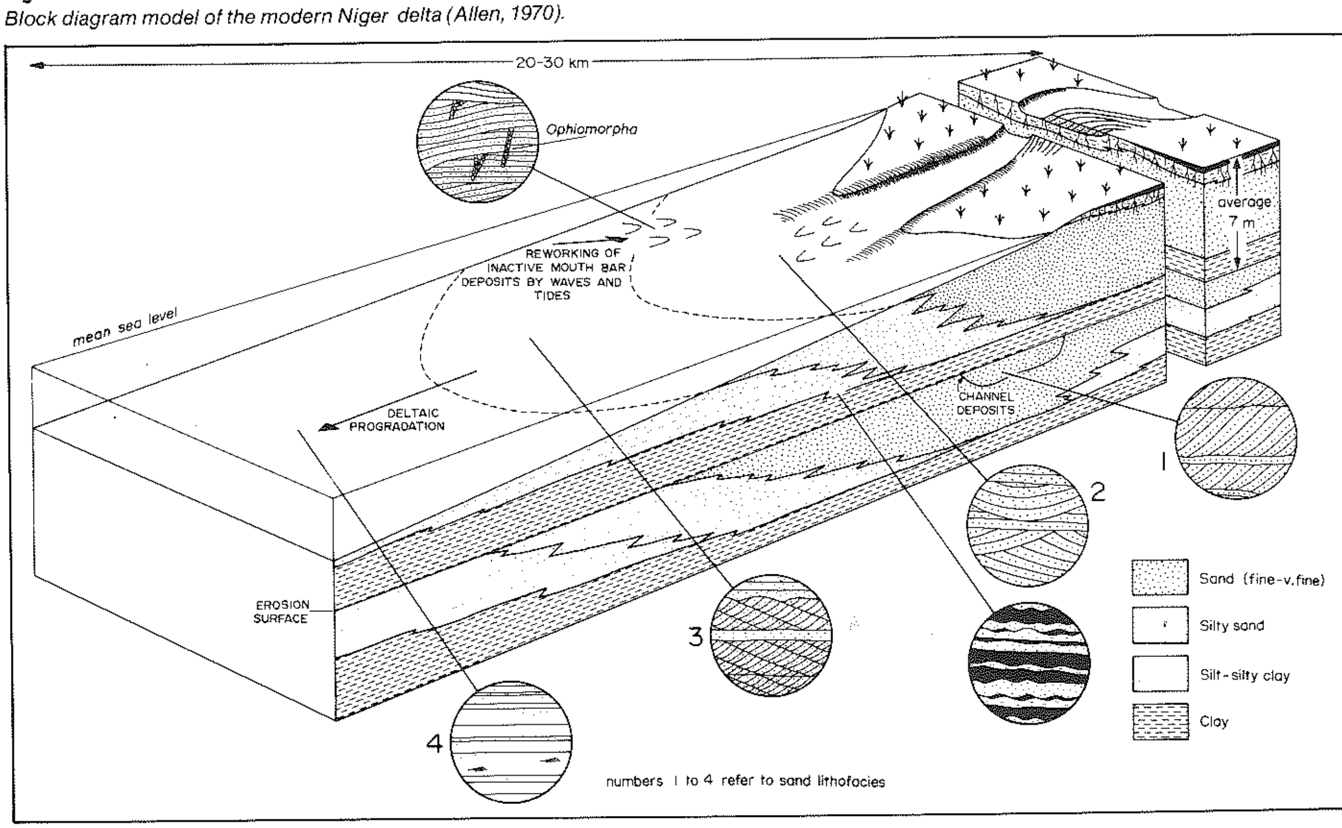


Figure 16
Block diagram model of the Eureka Sound Formation (Tertiary), Banks Island, Arctic Canada, showing interpretation of coarsening upward cycles in terms of distributary mouth bar sands (Miall, 1976b).

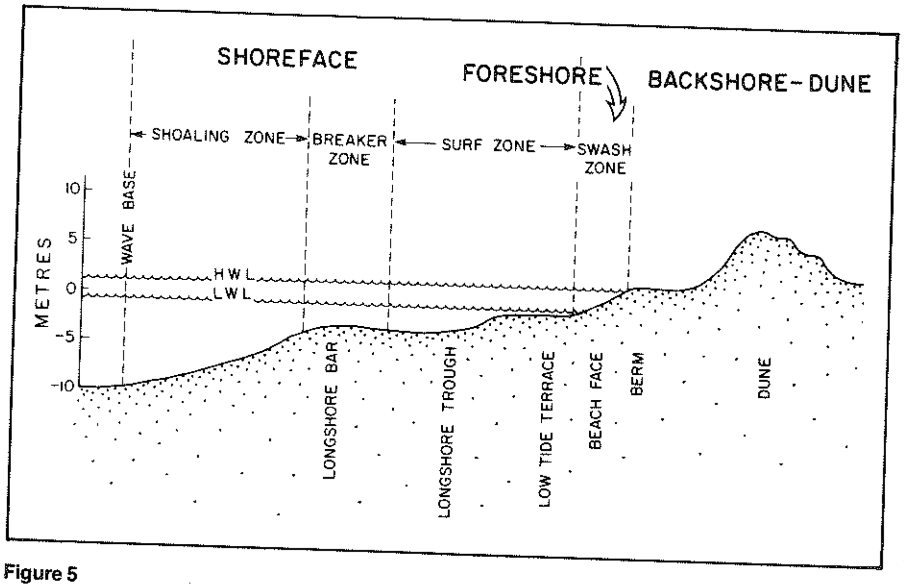


Figure 5
Generalized profile of the barrier beach and shoreface environments.

affect the sea bed. Hence the shoreface is an environment in which depositional processes are governed by wave energy. The amount of wave energy dissipated on the bottom decreases with increased water depth, and this inverse relationship governs the range of textures and sedimentary structures observed in shoreface deposits.

The shoreface is usually divided into three interrelated zones whose boundaries are not oftern sharply-defined. These zones are termed 'lower', 'middle' and 'upper' in some studies, and 'transitional', 'lower' and 'upper' in others. The former division, adhered to in this review, was generally utilized until very recently, in studies of both modern and ancient deposits. For examples of the recent trends, compare the zonal divisions of Howard (1972) with those of Howard and Reineck (1981), and the interpretation of the Galveston borehole sequence in Figure 12 with that of McCubbin (1982, Fig. 22). This inconsistency in terminology may reflect the findings of some recent studies, which suggest that the shoreface environment and resultant deposits can be much more variable than realized previously (this is discussed further in a later section).

Lower shoreface deposits occur seaward of the break in the shoreface slope at the toe of the barrier-island sediment prism. Under normal conditions, the lower shoreface is a relatively low-energy transitional zone, where waves begin to affect the bottom, but where offshore shelf or basinal depositional processes also occur. This is reflected in the sediments which consist generally of very fine to fine-grained sands with intercalated layers of silt and sandy mud. Physical sedimentary structures include mainly planar laminated beds, which are often almost completely obliterated by bioturbation. Trace-fossil assemblages are abundant in lower shoreface sediments (see "Trace Fossil Facies Models", this volume; Howard, 1972).

Middle shoreface deposits extend over the zone of shoaling and breaking waves. This zone is subjected to high wave energy relative to the lower shoreface and is characterized generally by one or more longshore bars (Fig. 4). The occurrence of longshore bars is related to a low-gradient shoreface and abundant sediment supply (Davis, 1978); both these conditions favor the landward movement and build-up of linear sand bars by shoaling and breaking waves.

Middle shoreface deposits can be highly variable in terms of sedimentary structures and textures, depending on whether nearshore bars are present or absent (Hunter *et al.*, 1979). Generally fine- to medium-grained, clean sands predominate, with minor amounts of silt and shell layers. Depositional structures include low-angle wedge-shaped sets of planar laminae, but ripple laminae and trough cross laminae are common (Campbell, 1971; Howard, 1972; Land, 1972). Middle shoreface deposits may be extensively bioturbated (Fig. 6), especially in the lower parts, but the biogenic structures are generally less diverse than in deposits of the lower shoreface (see "Trace Fossil Facies Models", this volume; and Howard, 1972). The model proposed by Davidson-Arnott and Greenwood (1976) illustrates the complexity of sedimentary structures that can occur in a barred nearshore zone (Fig. 7). Vertical

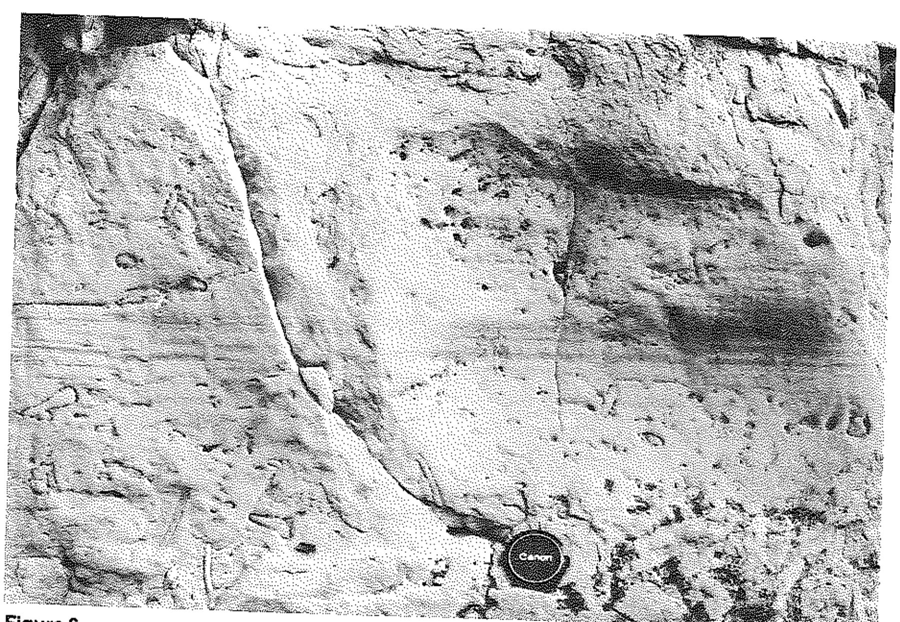
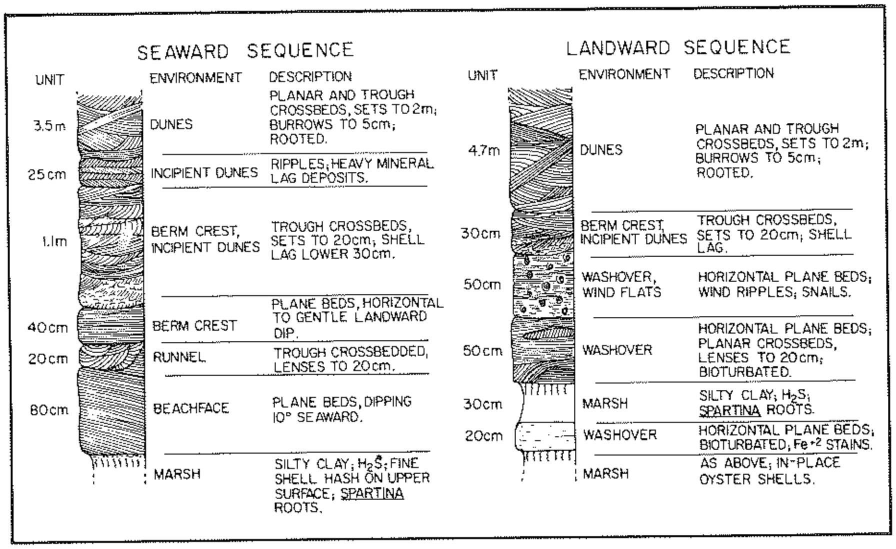
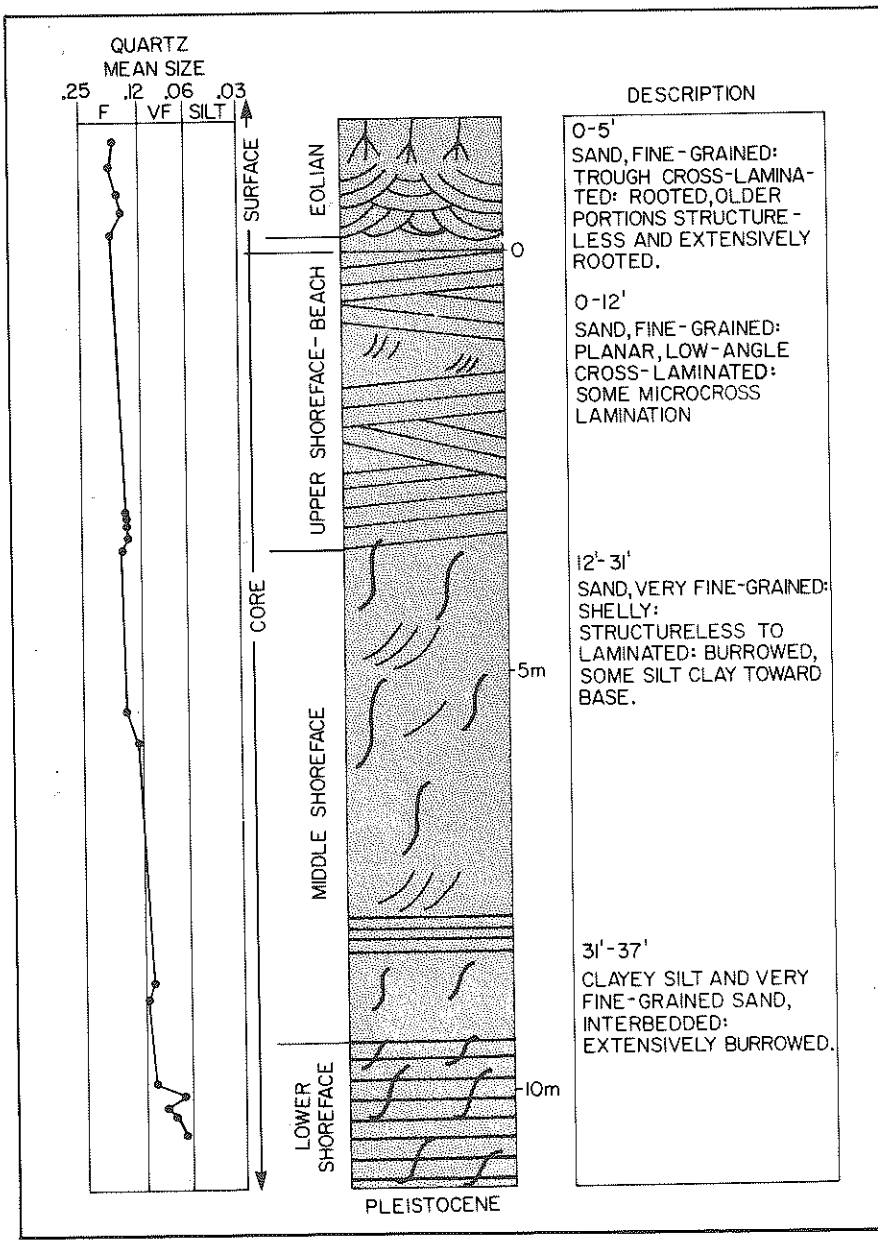


Figure 6
Alternating laminated-to-burrowed (Ophiomorpha) beds in lower to middle shoreface deposits of the Upper Cretaceous Blood Reserve Sandstone, southern Alberta (photo courtesy of Monti Lerand).



◀ **Figure 11**
Lithologic sequences observed in the landward and seaward edges of a Kiawah Island (South Carolina) beach ridge (from Barwis, 1978, Figs. 2 and 3).



◀ **Figure 12**
Sequence of sedimentary structures, textures and lithology in a core through Galveston Island (from Davies et al., 1971).

sedimentary structures, subhorizontal (planar) stratification, and small to medium-scale delta foreset strata where the washover detritus protrudes into the lagoon (Fig. 13). Textural and heavy mineral laminations and graded bedding can also occur (Andrews, 1970; Schwartz, 1982) depending on the nature of the source material. Textures may range from fine sand to gravel, but generally fine- to medium-grained sand forms the bulk of washover deposits. Washover deposits are generally thin, ranging from a few centimetres to two metres for each overwash event. In plan, they form elongate, semi-circular, sheet-like or tabular bodies a few hundred metres in width and oriented normal to the shoreline (Fig. 4). Coalescing washover fans can be in the order of kilometres in width, creating extensive washover flats which cover large tracts of the barrier.

Recent studies on modern barrier-island systems have illustrated that washover deposits form a significant portion of barrier sand bodies, especially in microtidal regions. Under transgressing conditions washover is one of the main processes by which the barrier island migrates landward, and scouring associated with washover is probably one of the main mechanisms responsible for the initiation of new tidal inlets. It is likely that washover deposits are more prevalent in ancient barrier sequences than has been recognized to date. Examples of ancient washover deposits include those documented by Bridges (1976), Horne and Ferm (1978), Hobday and Tankard (1978), and Hobday and Jackson (1979).

Tidal Channel (Inlet) and Tidal-Delta Facies

Tidal channel and tidal-delta sand bodies are intricately associated facies both with respect to their close proximity to one another, and with regard to

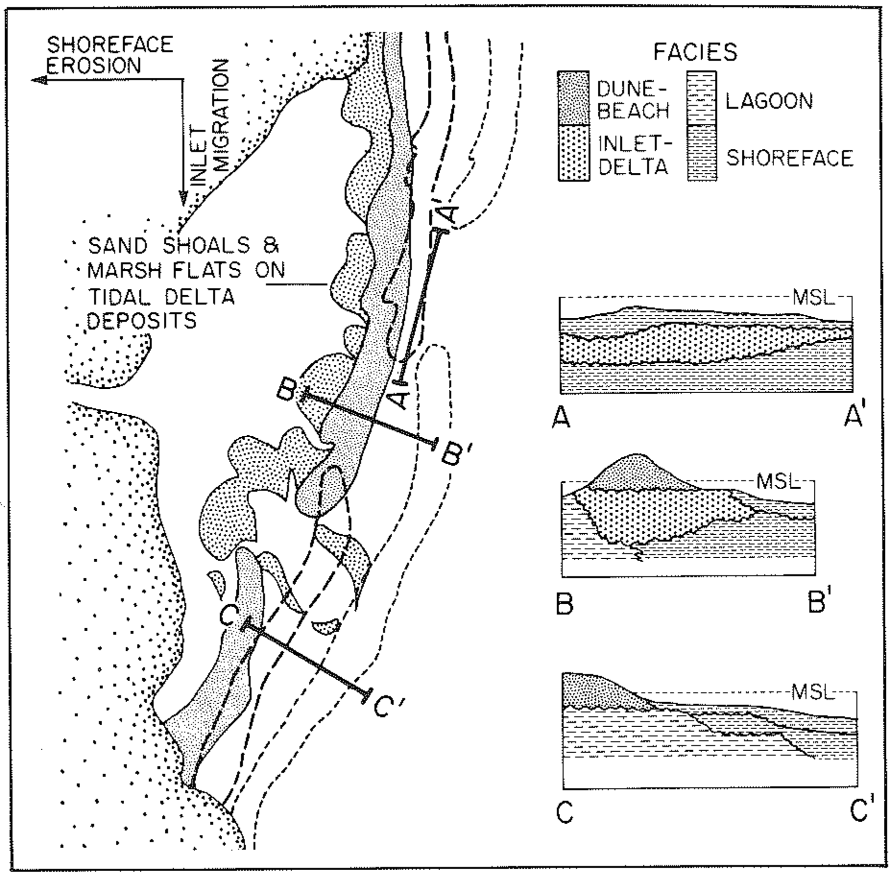


Figure 25
Schematic map and cross-sections illustrating how an inlet-delta sand body bounded by disconformities could be formed under conditions of inlet migration and concomitant shoreface erosion. The occurrence and shape of the sand body, and the nature of the enclosing sediments, vary with position in the barrier-island complex. (The heavy and light dashed lines on the map delineate former positions of the subaerial barrier.)

(Dixon, 1982), the Lower Cretaceous Glauconitic Sandstone of southern Alberta (Tilley and Longstaffe, 1984), and the Jurassic Borden Island Formation of the Canadian Arctic (Douglas and Oliver, 1979; see also Reinson, 1975). Some of these subsurface studies depict the coarsening upward, funnel-shaped, mechanical log response considered to be indicative of a prograding shoreline sequence (see "Subsurface Facies Analysis", this volume).

It is noteworthy that many ancient documented examples of prograding barrier – strand plain sequences are shown to be associated with deltas, either in interdeltaic positions, or in superposition but alternating with episodes of deltaic progradation. The primary requisite for development of prograding sequences is an abundance of sediment supply, and deltaic and interdeltaic coastal settings provide this.

Transgressive Model

The distillation of the generalized "end-member" sequence for the transgressive facies model comes also from modern examples, such as that depicted in Figure 23. This facies model is more complicated than the regressive model in terms of interbedding of facies and alternating lithologies. It is characterized by subtidal and intertidal back-barrier facies and does not show a fining-upwards or coarsening-upwards trend. The contact between some facies may be sharp or erosional. Many ancient sequences will deviate substantially from the normative model, because the facies stacking in transgressive sequences is quite variable,

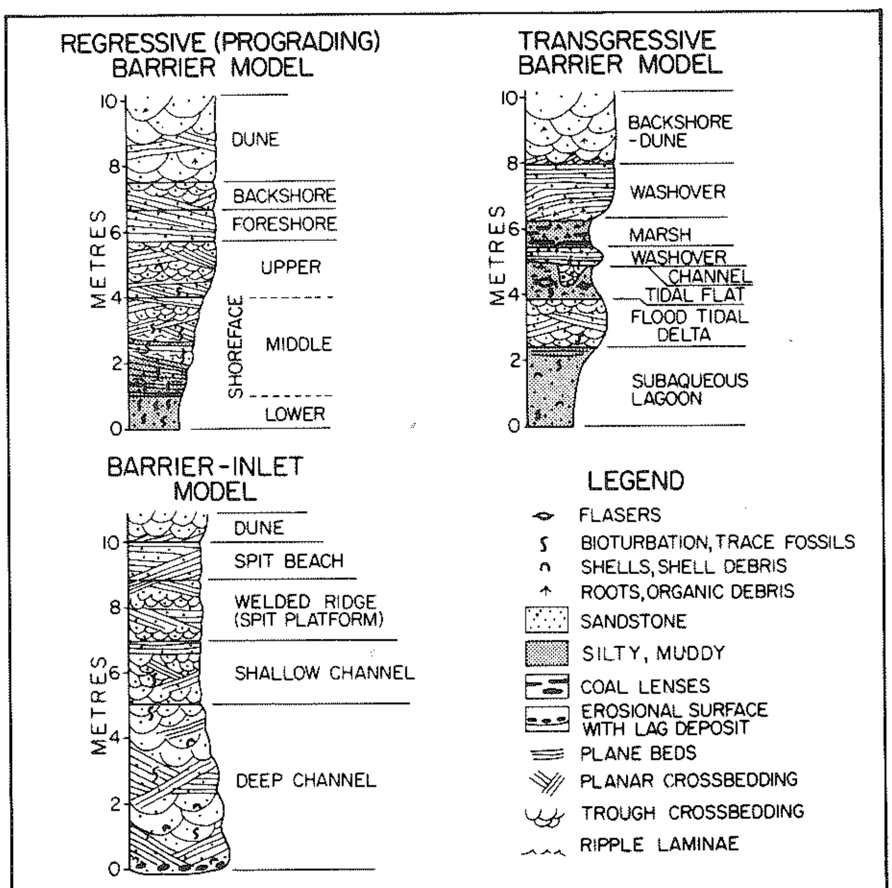


Figure 26
The three "end-member" facies models of barrier island stratigraphic sequences (a standard 10-metre unit is shown, but thickness could range up to a few tens of metres).

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, 1984b). 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 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.

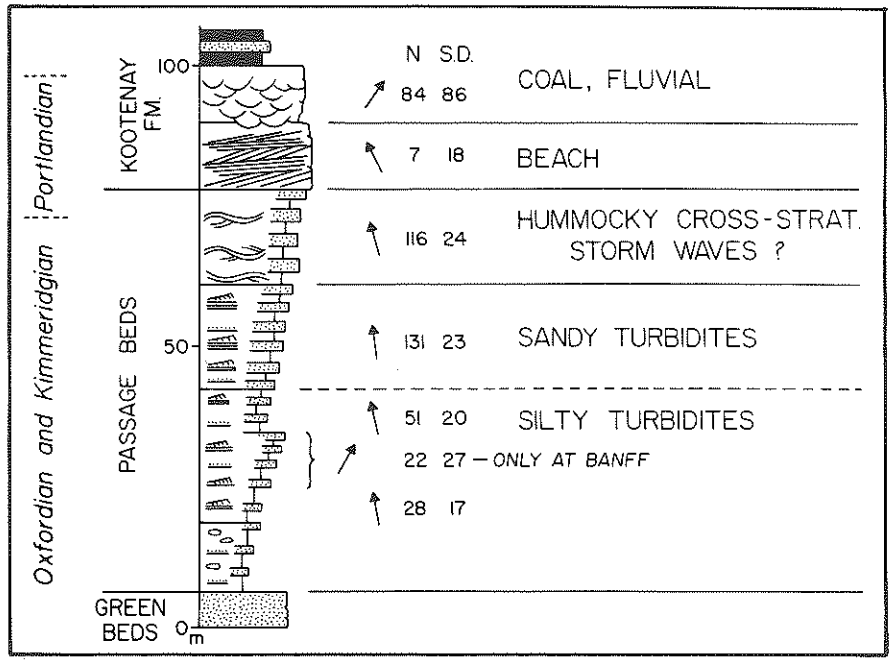


Figure 14
Generalized stratigraphic sequence for the Upper Jurassic "Passage Beds", Crowsnest Pass to Banff area, Alberta. Arrows show paleoflow vector means, N = sample size, SD = standard deviation. Turbidites show Bouma sequences, and form a thickening upward sequence at Banff. Note the same paleoflow 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.

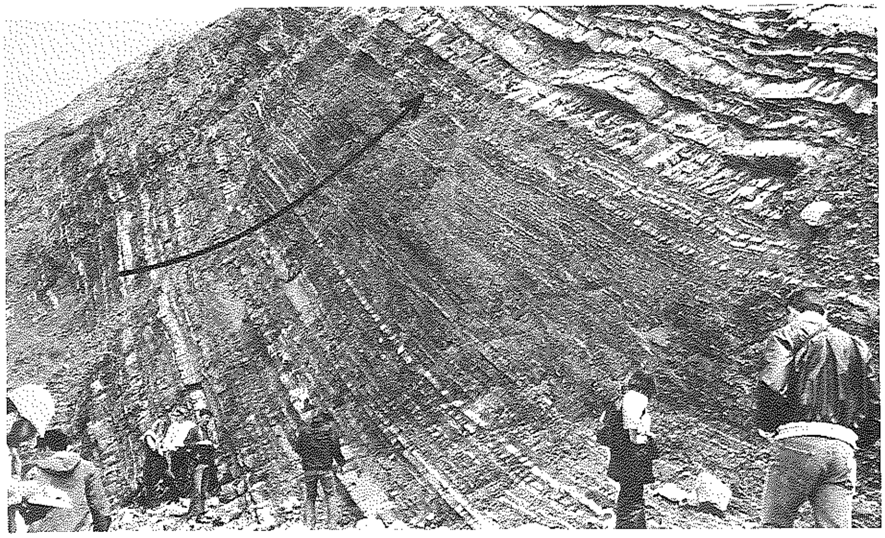


Figure 18
Thinning-upward sequence (arrowed) in the Paleocene turbidites at Shelter Cove, Point San Pedro, California. The spectacular continuity of the turbidites, and absence of a massive sandstone facies suggest that this sequence is not necessarily a channel fill. It may result from lateral lobe 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.

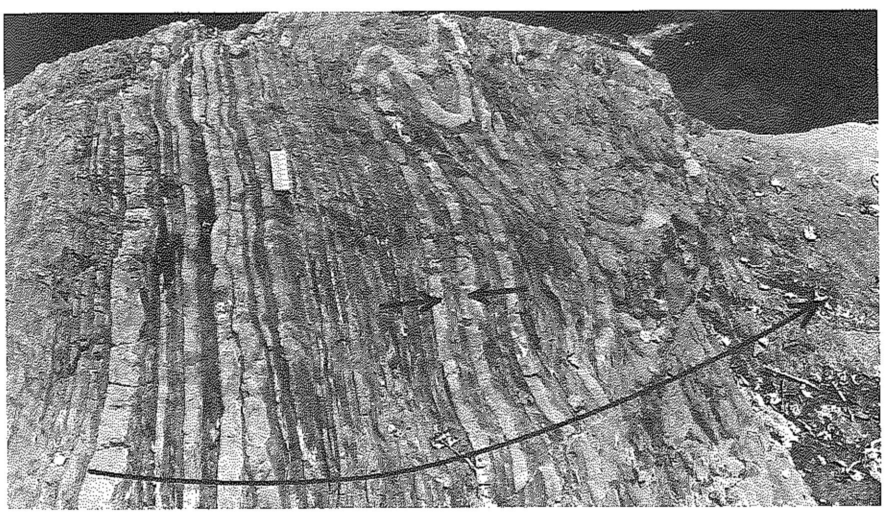


Figure 19
Thinning-upward sequence (arrowed) from turbidites, via a soft sediment slump (small arrows show way up) into mudstones. Stratigraphic top to right; Cretaceous tubidites at Wheeler Gorge, California. Compare this sequence with those in Figures 17 and 18. It probably represents neither channel 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 (in press).

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.

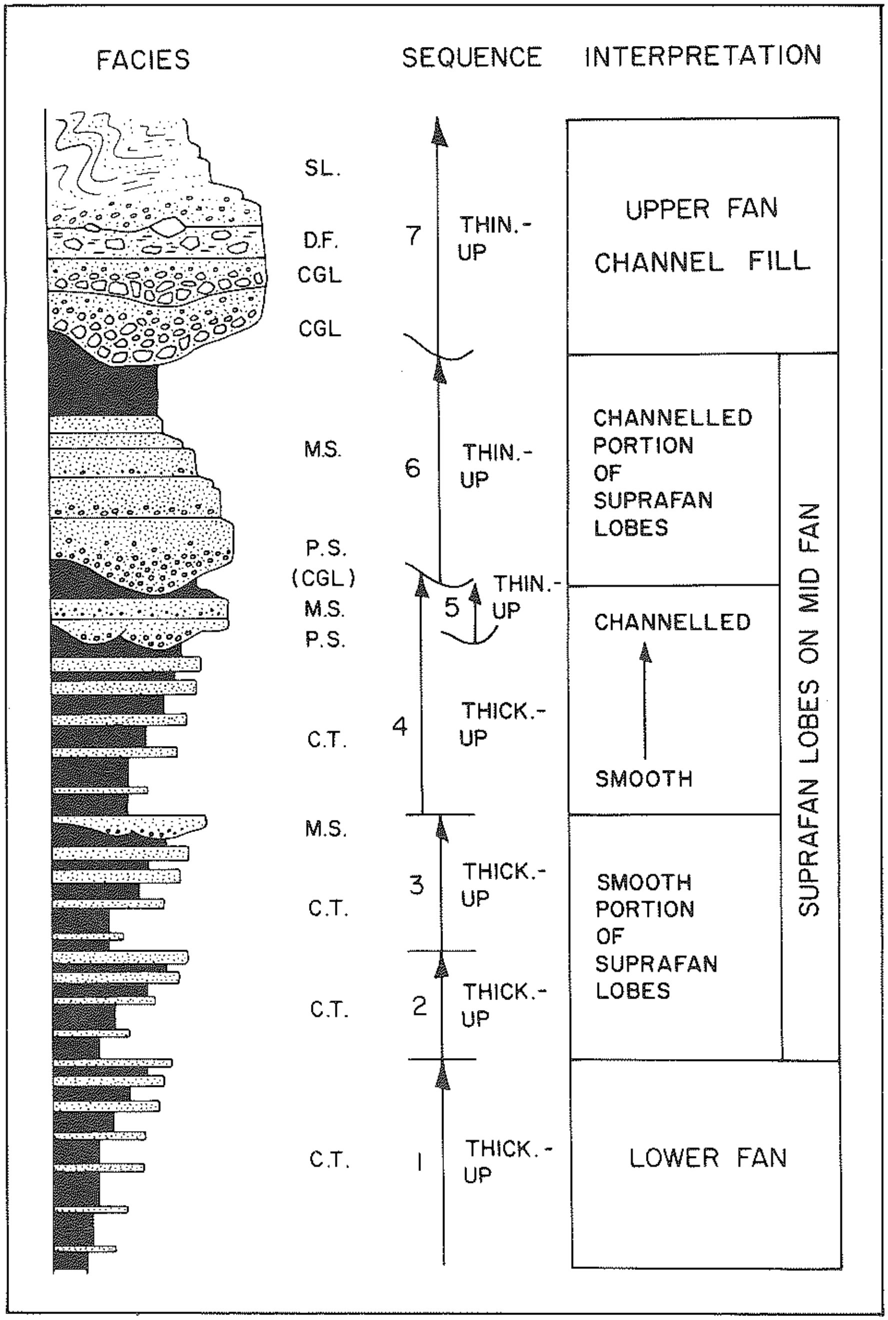


Figure 20

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-

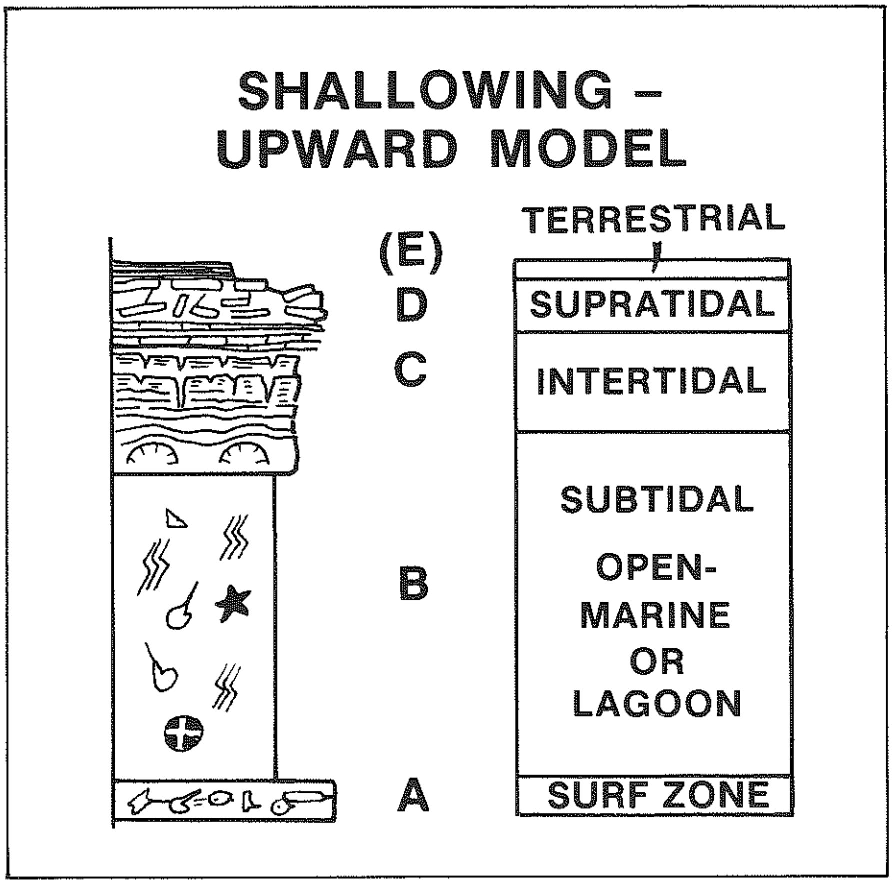


Figure 2
Five divisions of the shallowing-upward model for carbonates: A) lithoclast rich lime conglomerate or sand. B) fossiliferous limestone. C) stromatolitic, mud-cracked cryptalgal limestone or dolomite. D) well laminated dolomite or limestone, flat-pebble breccia. E) shale or calcrete, bracketed to emphasize that the unit is often missing - see text. Symbols used throughout are from Ginsburg (1975).

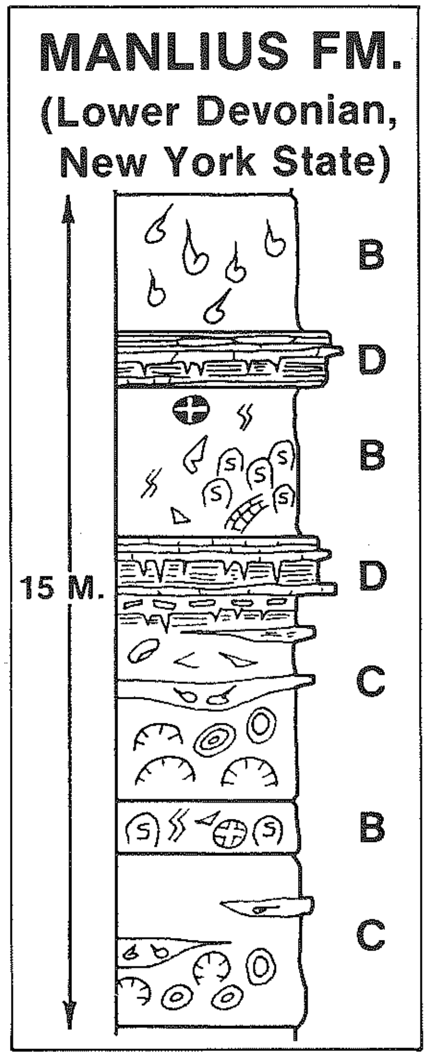


Figure 3
Actual sequence of several shallowing-upward sequences from the Manlius Fm., New York State (From Laporte, 1975).

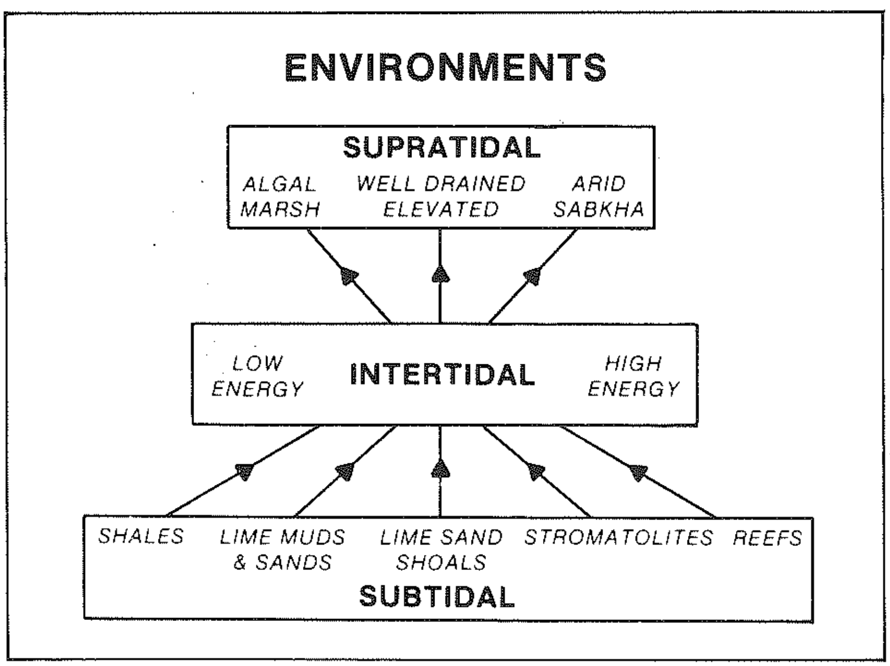


Figure 4
A flow diagram indicating the various possible environmental transitions present in a carbonate shallowing-upward sequence.

2) The Model as a Predictor. The thread that binds all such sequences together is the presence of the distinctive intertidal unit, which, once recognized, allows one to interpret the surrounding lithologies in some kind of logical sequence (Fig. 3), and thus predict what lithologies should occur in the rest of the succession under investigation.

First-order variation on the basic model revolves around the two main types of intertidal environment: 1) quiet, low-energy situations, commonly referred to as tidal flats, and 2) agitated, high-energy situations, or quite simply, beaches. Second-order variation involves the kind of subtidal units below and supratidal units above: the subtidal reflects the type of marine environment

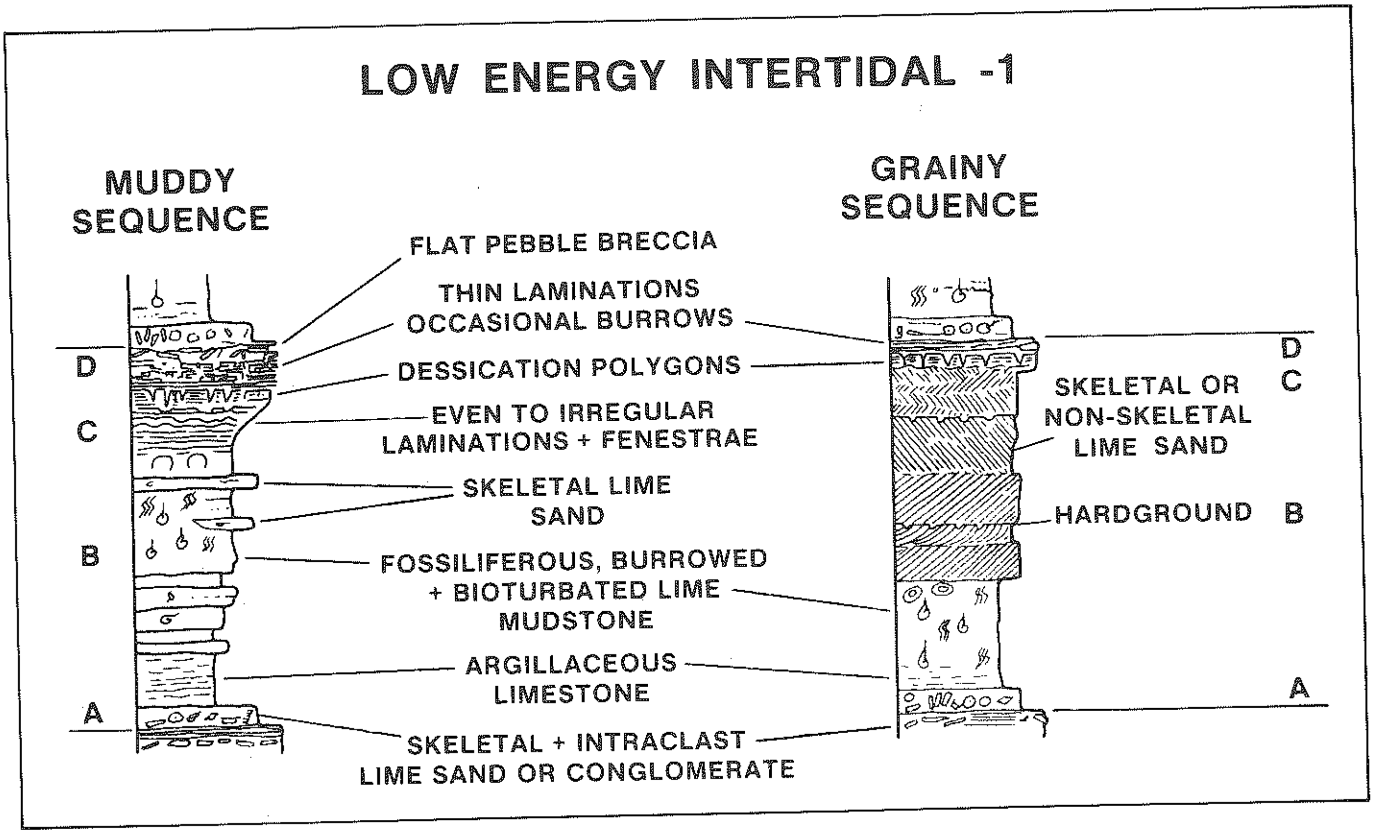


Figure 9
Two hypothetical sequences with a low energy tidal flat unit developed on a low energy subtidal unit (left) and a high energy lime sand unit (right).

If the creek levees in the intertidal zone have built up above normal high tide level, they consist of hard, finely to very finely laminated sediment, extremely regular and composed of alternating layers of sediment and thin algal mats with excellent fenestral porosity.

The landward parts of the supratidal zone may grade into various terrestrial environments, the end members of which are: 1) areas of elevated, pre-existing bedrock and no sedimentation in which the surface of the rock is characterized by intensive subaerial diagenesis, and the development of caliche (calcrete crusts): 2) areas of contemporaneous sedimentation which grade between: a) low-lying environments in regions of high rainfall occupied by algal marshes, b) low-lying environments in arid, desert regions, characterized by evaporite formation, and c) well-drained zones, often slightly elevated and with little deposition.

Algal marshes, flooded by fresh water during the rainy season, are an ideal environment for the growth of algal mats and these mats are periodically buried by layers of sediment swept in during particularly intense storms. The preserved record is therefore one of thick algal mats alternating with storm layers. With progressive aridity the supratidal zone dries out. If the chlorinity of the groundwaters remains constantly above 39 ‰ cementation, particularly by aragonite, is common. Cementation is most common if there is minor but consistent input of fresh water from inland to dilute the hypersaline groundwaters somewhat. If the chlorinity of the groundwaters remains constantly about 65 ‰ then authigenic evaporites precipitate within the sediment below ground level. In this setting (called a supratidal sabkha, or salt flat in the Middle East; see "Continental and Supratidal (Sabkha) Evaporites", this volume) dolomitizaiton is also common in the subsurface, saline brine pools occur at the surface, and terrigenous wind-blown sand is common in the sediment.

In relatively well-drained zones the supratidal environment is a deflation surface occasionally cut by the upper reaches of tidal creeks, sometimes damp from rising capillary waters and covered by a thin film of algal mat. Scoured and rippled sediment is common and clasts are sometimes encrusted with algae to form oncolites.

COMMON SEQUENCES WITH A LOW—ENERGY INTERTIDAL UNIT

Muddy and Grainy Sequences. These sequences developed either by progradation of the wide continental tidal flat or by shoaling the lime sand bodies that formed the barrier offshore (Fig. 9). The climate in the region of deposition was generally too wet or the ground-water table too low or diluted by fresh water to permit precipitation of evaporites.

The muddy sequences, those in which skeletal lime muds or muddy lime sands are the main subtidal unit, are well developed today in well-drained areas of Shark Bay where salinities are too high to permit development of a normal marine fauna as well as browsing of the algal mats by gastropods. Muddy sequences are also well developed in the tidal creek and pond belt of the Bahamas. These sequences are generally regarded as the 'classic' tidal flat sequences. The basal unit, if present, records the initial incursion of the sea onto land and as such is commonly coarse-grained, composed of clasts: all diagnostic of surf-zone deposition. The cementation is characteristic, and so deposits contain many intraclasts of cemented lime sand, and bored surfaces. Once the shoals, or parts of the shoals are inactive they may be burrowed and much of the original cross-bedding may be destroyed.

The intertidal to supratidal units are similar to those described above but are generally relatively thin. If the shoal is exposed for a long time caliche and soil profiles commonly develop, reflected by brown irregular laminations, breccias, and thin shale zones.

An excellent description of muddy and grainy sequences can be found in Demicco and Mitchell (1982).

Stromatolite and Reef Sequences. One common variation on the model is the development of shoaling-upward sequences in association with abundant stromatolites in the lower Paleozoic/-Precambrian and with reefs in the Phanerozoic in general.

In Shark Bay, Western Australia, where all environments are hypersaline and so stromatolites abound, the interrelationship between stromatolite morphology and environment has only recently been documented (Hoffman, 1976). In the intertidal zone columnar to club-shaped forms up to one metre high are found rimming headlands. In relatively high energy, exposed environments the relief of the columns is proportional to the intensity of wave action. The columnar forms grade laterally away from the headlands to the lower energy bights, where the stromatolites are more prolate and elongate oriented normal to the shoreline. In tidal pools digitate columnar structures abound.

These growth forms are the result of active sediment movement; algal mats only grow on stablized substrate, thus columns are nucleated upon pieces of rock, or cemented sediment; growth is localized there and does not occur on the surrounding shifting sands. Early lithification of the numerous superimposed layers of mat and sediment turns the structure into resistant limestone. Moving sand continuously scours the bases of the stromatolites. The mounds or pillars are largest in subtidal or lower intertidal environments and decrease in synoptic relief upwards, finally merging with stratiform mats in upper intertidal zones, above the zone of active sediment movement.

The resulting model sequence, summarized in Figure 12, is integrated from the Shark Bay example and the summary sequence of 200 or more shoaling sequences present in the Rocknest Formation of middle Precambrian age near Great Slave Lake (Hoffman, 1976). In the intertidal zone deposits reflect higher energy than normal, indicating a more exposed shoreline. These sediments underlie and surround the domal (Fig. 13) to columnar stromatolites, which in turn grade up into more stratiform stromatolites, and finally into very evenly bedded structures. The supratidal unit of this sequence will be characterized by both desiccation polygons and flat-pebble breccias as well as occurrences of delicate branching stromatolites (Fig. 14), formed in supratidal ponds. Care should be taken in delineating this sequence because stromatolites that are similar to those in the intertidal zone also occur in the subtidal (Playford and Cockbain, 1976).

Shallowing-upward sequences are also common as the last stage of sedimentation in large bioherms, as numerous successions within the large back-reef or lagoonal areas of reef complexes, and as 'caps' on widespread biostromes. In this type of sequence the shoaling upward is first reflected in the subtidal unit itself, generally as a transition from large massive hemispherical colonial metazoans of the reef facies, to the more delicate, stick-like forms that are common in the shallow protected

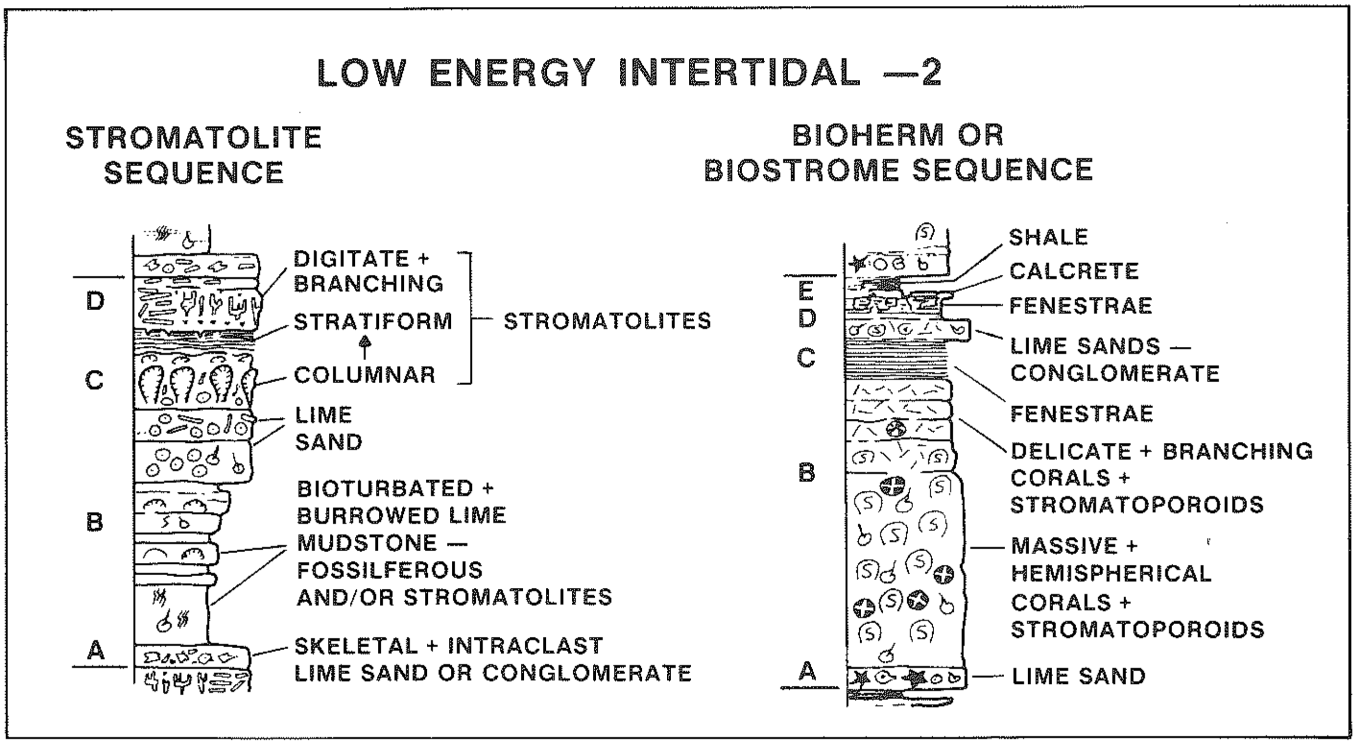


Figure 12
Two hypothetical sequences with a low energy intertidal unit developed in conjunction with stromatolites (left) and on top of a skeletal metazoan bioherm or biostrome (right).

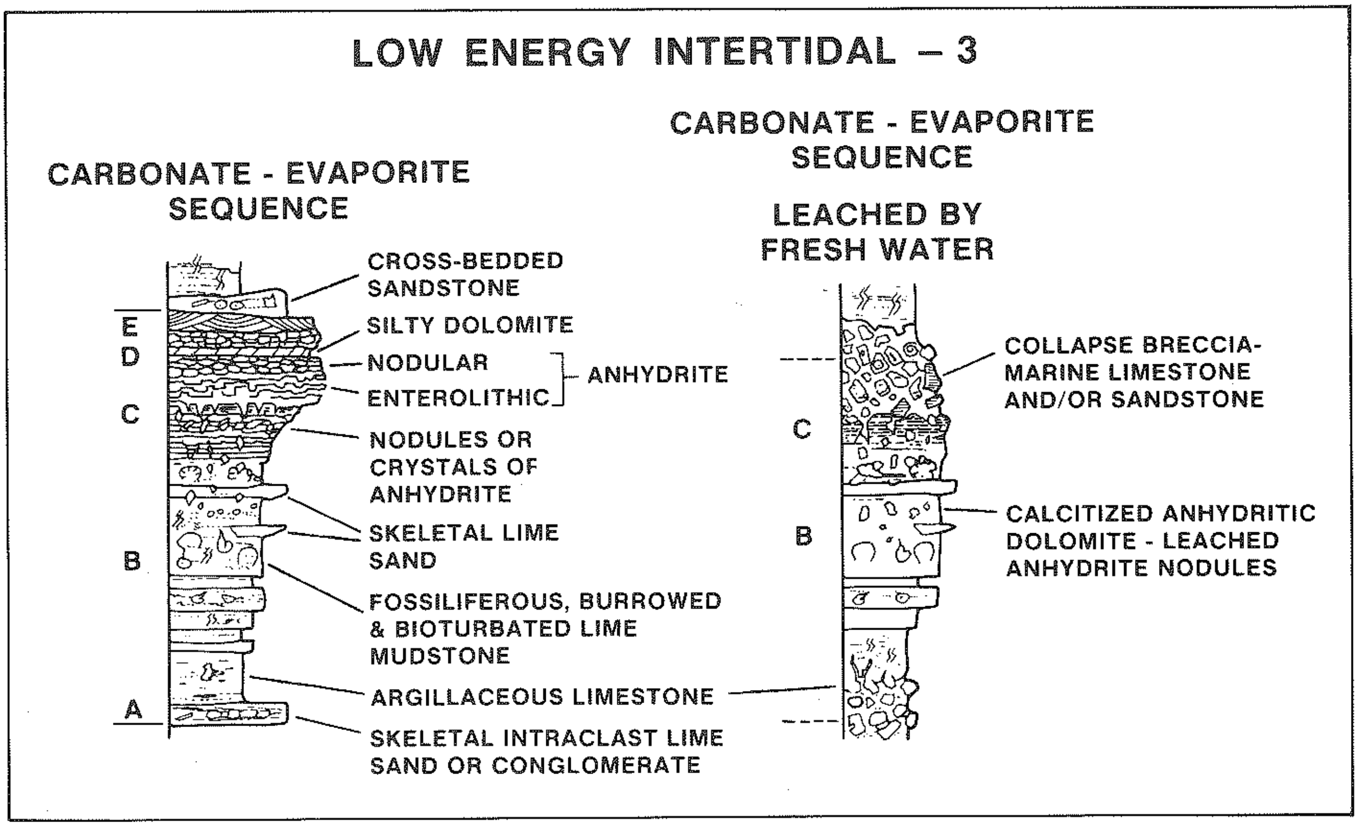


Figure 15
Two hypothetical sequences with a low-energy intertidal unit and a supratidal unit developed under arid conditions; on the right the evaporties have been dissolved by percolating fresh waters.

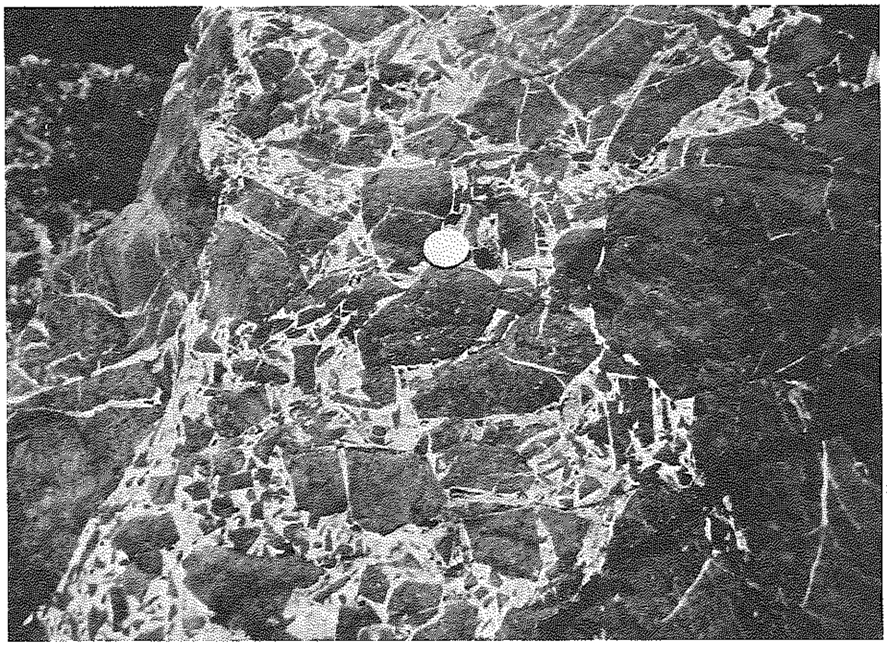
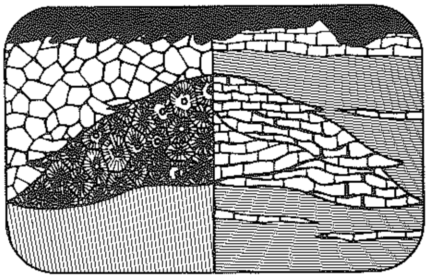


Figure 16
A collapse breccia of subtidal lime mudstone clasts in white calcite: caused by the solution of anhydrite at the top of a shallowing-upward sequence in the Shunda Fm. (Mississippian) at Cadomin, Alberta (Photo courtesy R.W. Macqueen).

upper part, is commoly altered to calcite, in the reverse of the dolomitization process (so-called "dedolomitization").

SEQUENCES WITH A HIGH ENERGY INTERTIDAL UNIT
In contrast to the low-energy intertidal (the tidal flat) the higher energy beach zone is not commonly recognized in the rock record. This may be partly because it resembles many subtidal grainstone deposits and hence is not obviously distinctive. Also, it is relatively narrow compared to the tidal flat, and has a lower preservation potential. Finally, the beach deposits lack the distinctive sedimentary features of the tidal flat. These very reasons illustrate the value of the concept of a shoaling-upward sequence as a guide. Once the potential for such a sequence is recognized in the geologic record, then one can concentrate on the search for subtle features that characterize beach deposition, which otherwise might go unnoticed.

Modern Carbonate Beaches
The sedimentology of carbonate beaches is nicely illustrated by Inden and Moore (1983). The beach is characterized by two zones: 1) the lower foreshore, usually below the zone of wave swash, and 2) the upper foreshore,



Reefs

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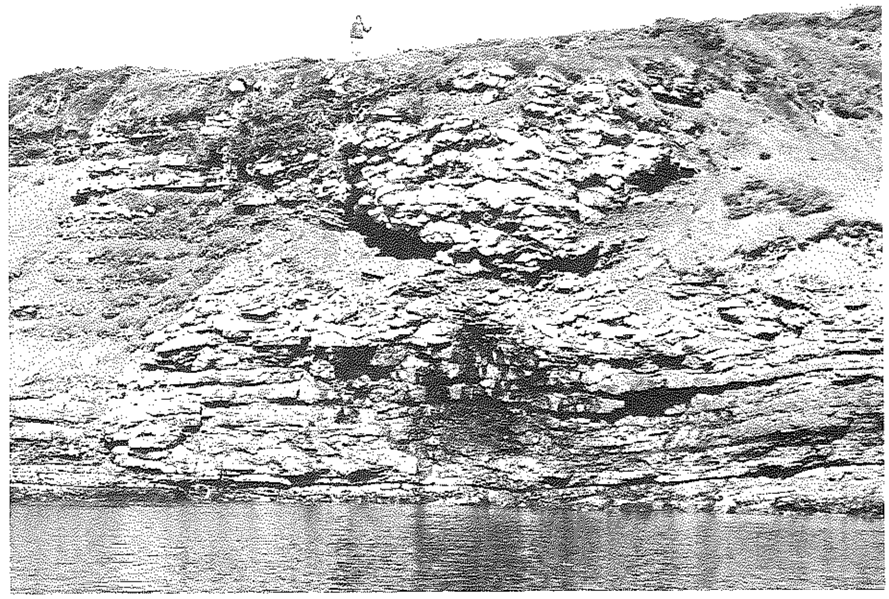

Figure 1
A patch reef of Lower Cambrian age exposed in sea cliffs along the northern shore of the Strait of Belle Isle, Southern Labrador.

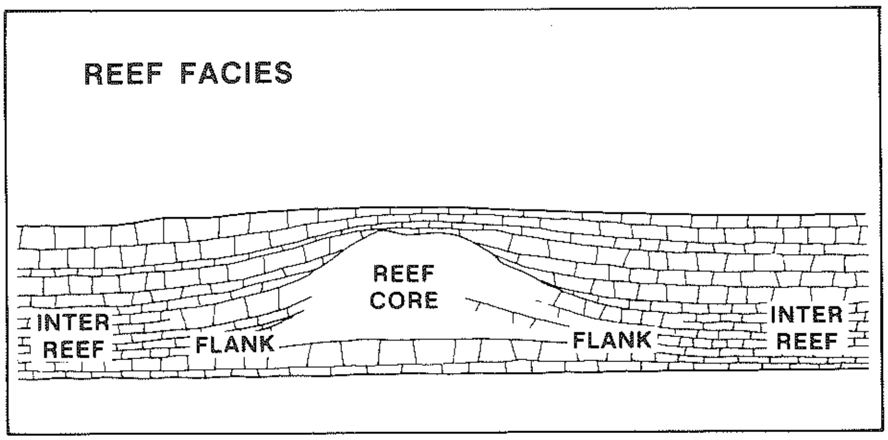

Figure 2
A sketch illustrating the three major reef facies in cross-section.

INTRODUCTION

A reef, rising above the sea floor, is an entity of its own making – a sedimentary system within itself. The numerous, large calcium carbonate secreting organisms stand upon the remains of their ancestors and are surrounded and often buried by the skeletal remains of the many small organisms that once lived on, beneath, and between them.

Because they are built by organisms, fossil reefs (Fig. 1) are storehouses of paleontological information and modern reefs are natural laboratories for the study of benthic marine ecology. Also, fossil reefs buried in the subsurface contain a disproportionately large amount of our oil and gas reserves compared to other types of sedimentary deposits. For these reasons, reefs have been studied in detail by paleontologists and sedimentologists, perhaps more intensely than any other single sedimentary deposit, yet from two very different viewpoints. This paper is an integration of these two viewpoints. I shall concentrate less on the familiar trinity of back-reef, reef, and fore-reef, but more on the complex facies of the reef proper.

Since the first edition of Facies Models, there has been much new information on both the sedimentology and paleontology of reefs. The model itself has been presented elsewhere (James, 1983) and amplified using numerous examples from the modern and fossil record. In this present version the model remains unchanged but many of the underlying concepts and implications that flow from it have been revised and/or enlarged.

THE ORGANISM-SEDIMENT MOSAIC

Reefs can generally be subdivided into three facies (Fig. 2).

1) *Reef-core facies* - massive, unbedded, frequently nodular and lenticular carbonate comprising skeletons of reef-building organisms and a matrix of lime mud.

2) *Reef-flank facies* - bedded lime conglomerates and lime sands of reef-derived material, dipping and thinning away from the core.

3) *Inter-reef facies* - normal shallow-water, subtidal limestone, unrelated to reef formation, or fine-grained siliciclastic sediments.

A useful, non-generic term for such a structure is "bioherm" – for discussion of this and other reef terminology the interested reader is referred to papers by Dunham (1970), Heckel (1974), Longman (1981), and James (1983).

Reef facies are best differentiated on the basis of several independent criteria including: 1) the relationship between, and relative abundance of large skeletons and sediments, i.e., the type of reef limestone, 2) the diversity of reef-building species, and 3) the growth form of the reef builders.

Types of Reef Limestone

The present state of any thriving reef is a delicate balance between the upward intolerant of fine sediment. The open ocean and windward locations are the only places in which fine sediment is continuously swept away.

Reefs form a natural breakwater when they grow into the zone of onshore waves and swells and create a relatively quiet environment in the lee of the reef crest. Commonly, this restriction significantly changes water circulation on the shelf, platform, or lagoon behind the reef. In such a marginal location, the symmetrical reef facies model comprising a reef-core facies surrounded on all sides by reef-flank facies is no longer discernable. Instead facies are more asymmetrically distributed with the reef-core facies flanked on the windward side by the fore-reef facies and on the leeward side by the platform facies (often called the back-reef facies).

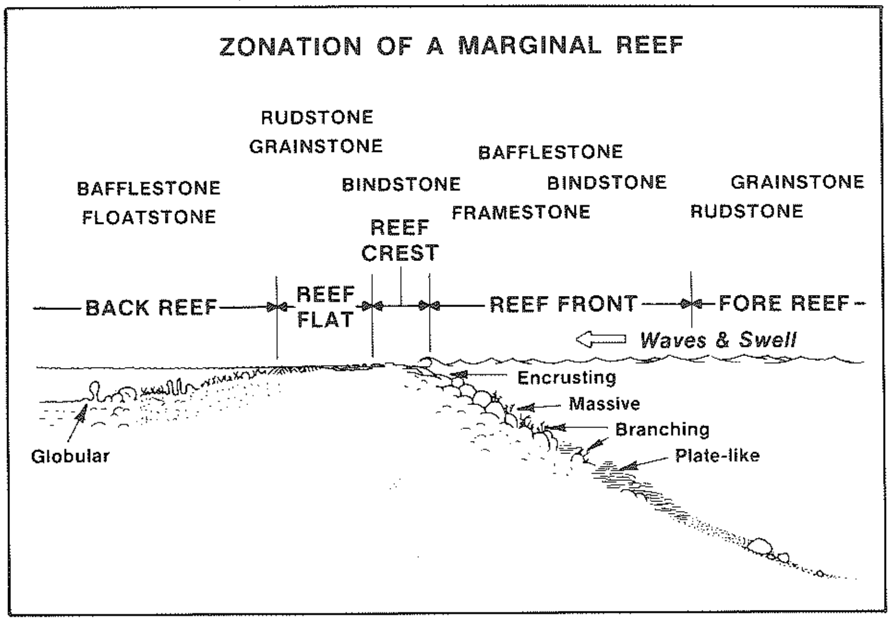

Figure 9
Cross-section through a hypothetical, zoned, marginal reef illustrating the different reef zones, spectrum of different limestones produced in each zone, and environment of different reef-building forms.

High-Energy Reefs

In these high energy settings the reef is distinctly zoned (Fig. 9)

Reef Crest Zone. This is the highest part of the reef at any stage in its growth, and if in shallow water, it is that part of the reef top that receives most of the wind and wave energy. Composition of the reef crest depends upon the degree of wind strength and swell. In areas where wind and swell are intense, only those organisms that can encrust, generally in sheet-like forms, are able to survive. When wave and swell intensity are only moderate to strong, encrusting forms still dominate but are commonly also bladed or possess short, stubby branches. In localities where wave energy is moderate, hemispherical to massive forms occur, with scattered clumps of branching reef-builders. The community is still of low diversity. The lithologies formed in these three cases would range from bindstones to framestones.

Reef Front Zone. This zone extends from the surf zone to an indeterminate depth, commonly less than 100 metres, where the zone of abundant skeletal growth grades into sediments of the fore-reef zone. Direct analogy between modern reefs, especially Caribbean reefs, and ancient reefs is difficult because today the sea floor from the surf zone to a depth of about 12 metres is commonly dominated by the robust branching form *Acropora palmata*, a species which developed only recently in the late Cenozoic. Such branching forms are rarely found in ancient reefs. Instead, the most abundant forms are massive, laminar to hemispherical skeletons, forming framestones and sometimes bindstones.

The main part of this zone supports a diverse fauna with reef-builders ranging in shape from hemispherical to branching to columnar to dendroid to sheet-like. Accessory organisms and various niche dwellers such as brachipods, bivalves, coralline algae, crinoids, and green segmented calcareous algae (Halimeda), are common. On modern reefs where the reef-builders are corals, this zone commonly extends to a depth of 30 metres or so. The most common rock type formed in this zone would still be framestone, but the variety of growth forms also leads to the formation of many bindstones and bafflestones.

Below about 30 metres wave intensity is almost non-existent and light is attenuated. The response of many reef-building metazoans is to increase their surface area, by having only a small basal attachment and a large but delicate plate-like shape. Rock types from this zone look like bindstones, but binding plays no role in the formation of these rocks and perhaps another term is needed.

The deepest zone of growth of coral and green calcareous algae on modern coral reefs is about 70 metres. The lower limit may depend upon many factors, perhaps one of the most important being sedimentation, especially in shale basins which border many reefs. This lower limit should therefore be used with caution in the interpretation of fossil reefs.

Sediments on the reef front are of two types: 1) internal sediments within the reef structure, generally lime mud giving the rocks a lime mudstone to wackestone matrix, and 2) coarse sands and gravels in channels running seaward between the reefs. These latter deposits have rarely been recognized in ancient reefs.

As a result of numerous observations on modern reefs it appears that most of the sediment generated on the upper part of the reef front and on the reef crest is transported episodically by storms up and over the top and accumulates in the lee of the reef crest. Sediments on the intermediate and lower regions of the reef front, however, are transported down to the fore-reef zone only when it is channelled by way of passes through the reef.

Reef Flat Zone. The reef flat varies from a pavement of cemented, large skeletal debris with scattered rubble and coral-

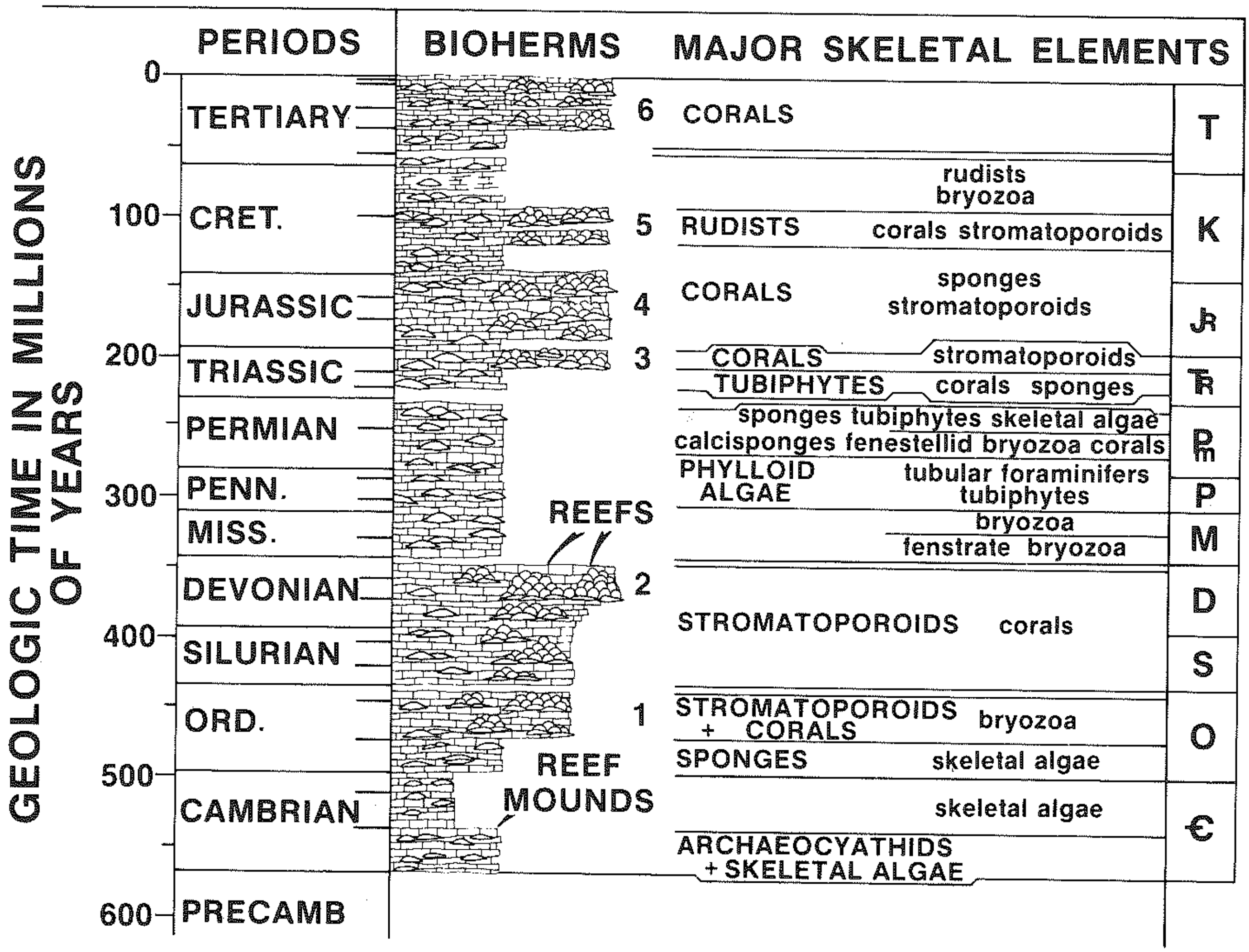

Figure 21

An idealized stratigraphic column representing the Phanerozoic and illustrating times when there appear to be no reefs or bioherms (gaps), times when there were only reef mounds and times when there were both reefs and reef mounds. The numbers indicate different associations of reef-building taxa discussed in the text. (From James, 1983).

depth, in front of the barrier reef down on the reef front or fore-reef.

Patch reefs or reef mounds commonly form a very widespread lithofacies compared to the barrier reef. The stratigraphic thickness of these reefs is dependent upon the rate of subsidence: if subsidence rate is low, reefs are thin; if subsidence rate is high, reefs may be spectacular in their thickness.

Times When Only Delicate, Branching and Encrusting Metazoa Prevail

The margin of the shelf or platform is normally a complex of oolitic or skeletal (generally crinoidal) sand shoals and islands. The only reef structures are reef mounds which occur below the zone of active waves down on the seaward slopes of the shelf or platform and if conditions are relatively tranquil behind the barrier, on the shelf itself. Mounds may display a zonation, with ocean-facing sides in shallow water armoured with accumulations of fragmented and winnowed skeletal debris.

Not only can we predict adjacent facies with some confidence but given only small outcrops or pieces of core we can, within limits, predict the style of the reef in question. For example, a core composed of lime mudstone to floatstone with scattered bryozoans and stromatactis from a Mississippian sequence, when viewed from the point of view of the model clearly points to a reef mound, which since reefs proper did not form at this time, may be 10's to 100's of metres thick. A similar rock in a Silurian sequence, while predicting a similar structure if surrounded by slope or inner shelf strata, would in most other cases predict that the sample is from the basal part of a large reef. Alternatively a small Devonian outcrop, composed of a bafflestone of fasciculate stromatoporoids grading up into a framestone of large tabulate corals and many different stromatoporoids would predict that a massive reef should lie above.

The clear limitations of the model are, however, when using small samples. Reefs are generally large and heterogeneous structures so that a drill core for example may pass through areas of sediment between large skeletons, a very likely possibility since 30% to 40% of modern reefs are sediment or pore space, and so give a false picture of the true deposit.

they commonly lack any vestige of original fabrics, structures or mineralogy. They commonly are also the products of diagenetic/metamorphic processes rather than being primary sedimentary accumulates. There is a corresponding dearth of facies models for these deposits.

In the light of these four factors it is hardly surprising that basic disagreements exist about almost all aspects of evaporite genesis. Most significant amongst them are whether basin-central evaporites were deposited in deep or shallow water, and whether many evaporite structures and textures are of primary or post-depositional origin.

There is probably not a single Holocene depositional environment that is strictly analogous to those that generated most ancient evaporite sequences. Although there may be many similarities, there are always one or more aspects (commonly scale) that cannot be fitted into the analog scheme. The technique used to interpret ancient evaporites is to take items of information from several modern settings and relate them to their ancient counterparts.

No single facies model can be applied to so heterogeneous a grouping of rocks as the evaporites. A persistant dogma since the 1960s that evaporites represent deposits of supratidal or sabkha environments became much too one-sided. Yet today we face the danger of the other end of the pendulum, for some supratidal evaporites appear to be interpreted as subaqueous on insufficient grounds.

Given the correct environmental conditions, evaporites can mimic most other sediment types. There are evaporite turbidites and oolites; 'reefs' composed of huge gypsum crystals that formed mounds standing proud of the basin floor, and shallow-water clastic evaporites that resemble in texture and sedimentary structure their clastic or carbonate equivalents. Since evaporites may exhibit detrital as well as crystalline precipitate textures, these sediments constitute one of the most variable of sedimentary rock groups.

Evaporite minerals may form only a minor component of some deposits (isolated gypsum crystals in continental redbeds would be an example) and these are best considered part of other facies models.

At the present day, areas of sufficient evaporation (and negative water balance) to cause evaporite deposits to form are present between 10 and 40° north and south of the Equator (subtropical high-pressure atmospheric circulation zones). These zones have been wider in the past, particularly during periods of low sea level. The presence of evaporitic sodium sulfate deposits in seasonally-frozen lakes in Antartica and Western Canada deomonstrate that negative water balances are not exclusively a condition of low latitudes.

DEPOSITIONAL ENVIRONMENTS OF EVAPORITES

Evaporite sequences have been generated in a variety of geographic settings (Fig. 1): 1) coastal intertidal and supratidal environments (marine sabkhas), 2) lagoons on coasts, 3) large basins, with brine level below sea-level and marine inflow; brine level within the basin may be high (giving deep-water evaporites) or low, and 4) non-marine interior basins. These environments may be present in a wide spectrum of tectonic settings. They occur within 1) continental margins and shelves, 2) interior cratonic basins of varying

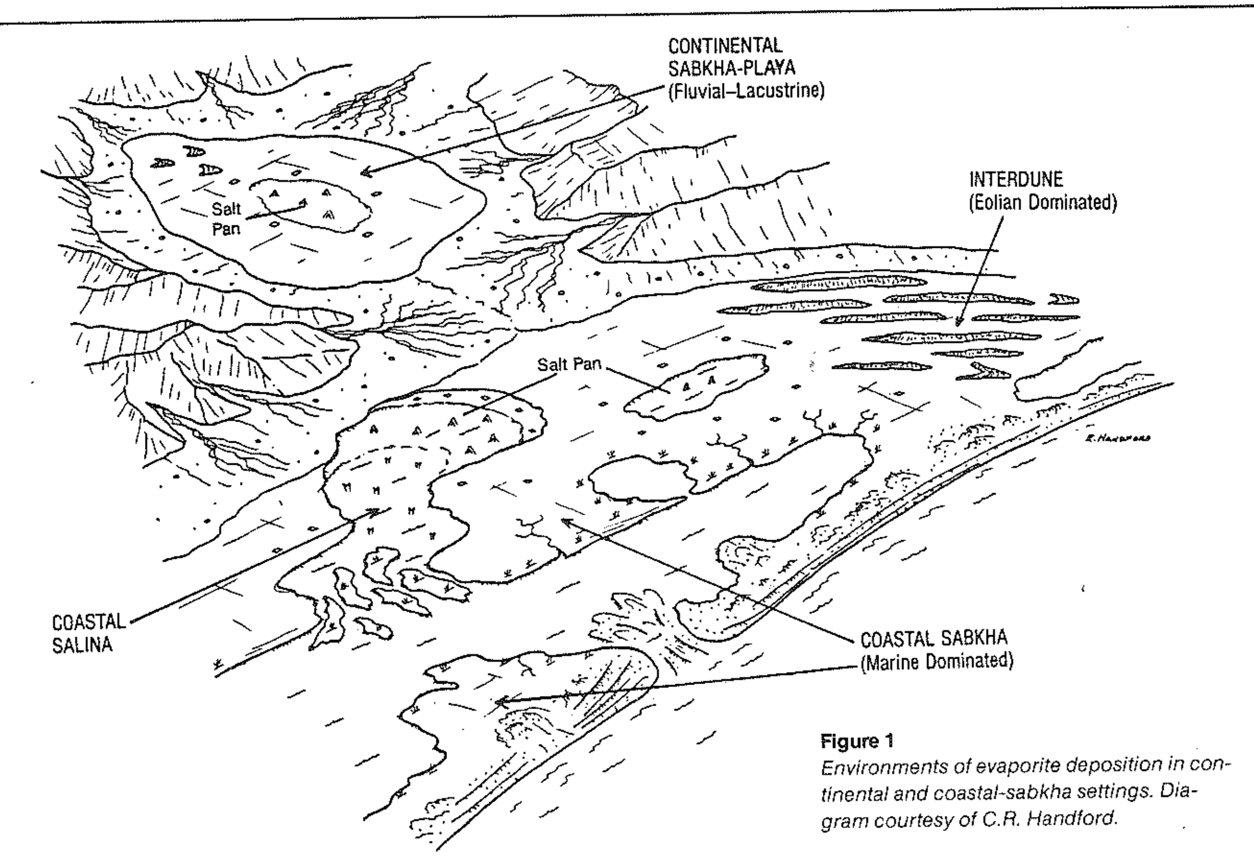

Figure 1
Environments of evaporite deposition in continental and coastal-sabkha settings. Diagram courtesy of C.R. Handford.

depths and 3) rifted continental margins – in which case evaporites may be present in areas that are now oceanic.

Of the many possible environments of evaporite precipitation, five major categories (or regimes) were identified by Schreiber *et al.* (1976) with a further subdivision in each category as to whether the evaporites are calcium sulphates or halides (with or without complex sulphates) (Fig. 2). Regimes grade into each other such that the identification may depend more upon associated facies than upon internal characteristics. Continental sabkha deposits commonly are internally identical with coastal sabkha deposits, differing only by being inserted within continental deposits. Furthermore, the degree of restriction required to generate halite and/or subaqueous sulphate deposits ensures that all these environments have some of the attributes of the continental regime. Distinction between large hypersaline inland lakes and partially desiccated small seas can be a somewhat academic exercise.

Three main environmental groupings are recognized here: continental, coastal sabkha, and subaqueous marine. Many ancient evaporites were deposited subaqueously within enclosed and hypersaline basins. The primary composition, textures and form of these subaqueous deposits are now only partially understood because, in part, so few hypersaline water bodies occur at the present day for study; none of them of large size.

The origins of small, thin evaporite deposits and marginal-marine evaporites composed of numerous superimposed sabkha cycles are readily discernable. In contrast, the formation of vast, thick, basin-central evaporites, some which cover millions of square kilometres and exceed several kilometres in thickness or which may directly overlie oceanic basement, present very different problems.

Some authors suggest that the enormous evaporite deposits form by lateral and vertical accretion within depositional environments similar to those of the present time (in supratidal flats, lagoons and salinas; Shearman, 1966; Friedman, 1972) whereas others consider that the great difference in scale between recent and ancient deposits requires explanations that are drastic departures from the present day settings of evaporites. They suggest either that precipitation occurred from vast bodies of hypersaline water (Schmalz, 1969; Hite, 1970; Matthews, 1974), or that evaporites were precipitated on the floors of desiccated seas (Hsu *et al.*, 1973).

Theoretical models, which were developed to answer the major compositional problems posed by large evaporite bodies, must be integrated with evidence from rock textures and structures (facies models). Unfortunately this integration is not yet possible because of basic disagreements concerning the depositional paleogeography of evaporites, and because many evaporite rock characteristics still have to be studied in detail or have disputed origins. Many evaporites are not just passive chemical precipitates or displacive growth structures, but are transported and reworked in the same ways as siliciclastic and carbonate deposits. For these sediments, sedimentary structures are a major key to unravelling the facies and will be emphasized here. Internal characteristics of evaporites alone can provide the necessary information about depositional environments. The most pressing environmental concern has been, and still is, the depth of water in which evaporites form.

Schreiber *et al.* (1976) recognize three main environmental settings for subaqueous evaporites. These are identified on the basis of sediment characteristics, believed to reflect the depth at which deposition occurs. Criteria used include: 1) structures indicative of wave and current activity, identifying an intertidal and shallow subtidal environment; 2) algal structures (in the absence of wave and current-induced structures), which are believed to identify a deeper environment but one that still resides with the photic zone; and 3) widespread evenly-laminated sediments (rhythmites) that lack evidence of current and algal activity (perhaps associated with gravity-displaced sediments), and characterize the deep, subphotic environment.

Considerable difficulty exists in using

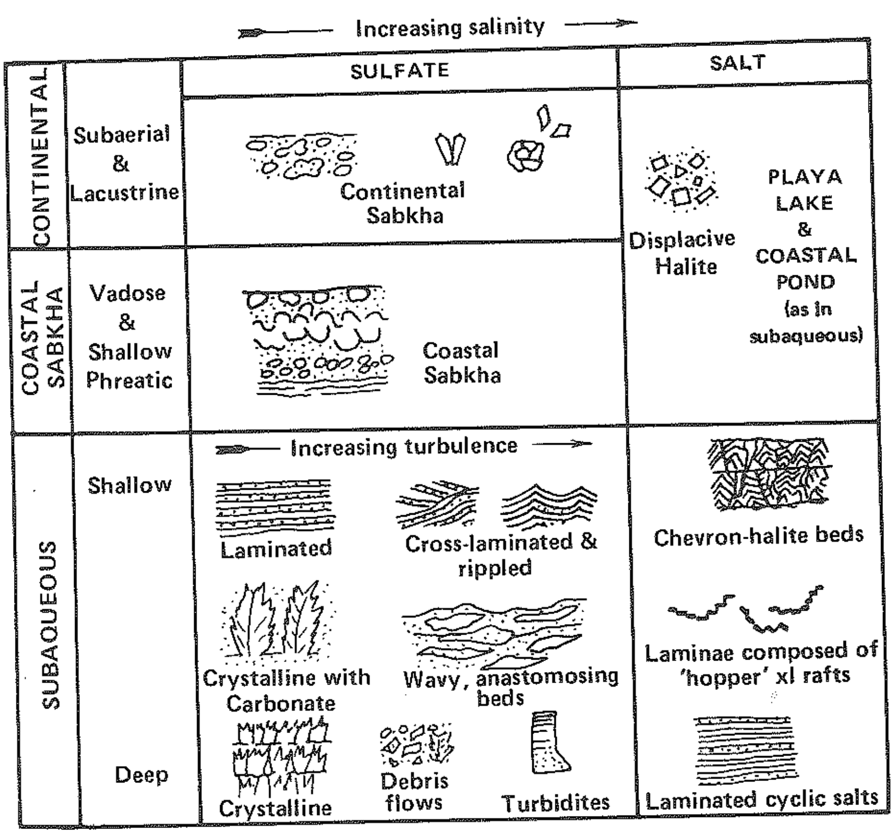

Figure 2
Summary of physical environments of evaporite deposition and the main facies present (modified from Schreiber et al.*, 1976).*