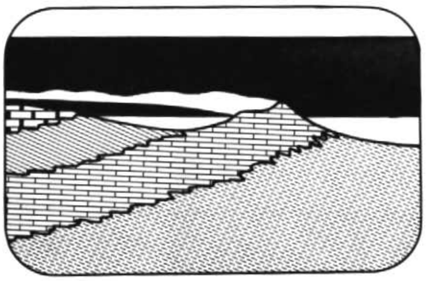

# Facies Models 14. Barrier Island Systems

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**Introduction**

The AGI *Glossary of Geology* defines a barrier island as a "long, low, narrow, wave-built sandy island representing a broadened barrier beach that is sufficiently above high tide and parallel to the shore, and that commonly has dunes, vegetated zones, and swampy terranes extending lagoonward from the beach." Inherent in this definition is the fact that there has to be three major geomorphic elements in a barrier-island system: 1) the sandy barrier-island chain itself, 2) the enclosed body of water behind it (lagoon or estuary) and, 3) the channels which cut through the barrier and connect the lagoon to the open sea (tidal inlets) (Fig. 1). This tripartite geomorphic framework clearly demonstrates that barrier-island systems are composites of three major clastic depositional environments: 1) the subtidal to subaerial barrier-beach complex, 2) the back-barrier region or subtidal-intertidal lagoon and, 3) the subtidal-intertidal delta and inlet-channel complex (Fig. 2). The view of the barrier island as a composite depositional system, until recently, has not been fully appreciated by geologists. This lack of appreciation is evident in most of the pre-1970 literature where there was an overwhelming preference for the use of just one barrier-island model (prograding Galveston Island Model) or "norm" for interpreting ancient rocks. If one recognizes the barrier-beach, lagoon, and tidal channel-delta scenario, it should be obvious that a single model for such a complex system is completely unrealistic.

Fortunately, within the last decade, there has been a renaissance in the interpretation of ancient barrier-island sequences. This has come about largely through the investigations of modern barrier-island systems by numerous workers including M.O. Hayes and J.C. Kraft (Hayes and Kana, 1976; Kraft, 1971, 1978). Because of these modern studies, we are now recognizing that Galveston Island is just one of at least


**Figure 1**
*Oblique aerial view of the barrier-island system at Tracadie, New Brunswick, showing the linear barrier-beach (B), the tidal inlets (I) through the barrier, and the lagoon (L) behind the barrier. Note the flood-tidal delta (F) in left foreground. In the upper part of photo, ice abuts against back-barrier marsh, May 3, 1977. (Photo by R. Belanger, AOL)*

three distinct barrier-island stratigraphic models.

In this review I will attempt to synthesize barrier-island stratigraphic sequences into three "end-member" depositional models for use in interpreting ancient rocks. As with deltas (Miall, 1976), our ideas on barrier-island rock deposits stem from the study of modern barrier systems. Consequently, the review draws heavily on examples of modern deposits to develop the "end-member" models.

**Origin and Occurrence**

Theories regarding the origin of barrier islands have been reviewed at length in the recent geological literature (Schwartz, 1973; Swift, 1975; Wanless, 1976; Field and Duane, 1976). The question of origin is controversial but there are three main hypotheses: 1) the building-up of submarine bars; 2) spit progradation parallel to the coast and segmentation by inlets; and 3) submergence of coastal beach ridges. The controversy remains largely unresolved because most of the evidence pertaining to origin has usually been destroyed by subsequent modification. Extensive modification and evolution of modern barrier islands has been occurring since the early Holocene through a combination of processes including inlet cut and fill, washover deposition, and longshore transport (Field and Duane, 1976). These processes have been enhanced by the progressive landward retreat of the barrier islands in response to the Holocene transgression (Swift, 1975).

Swift (1975) and Field and Duane (1976) consider that barrier formation by offshore bar emergence is insignificant compared to the other two mechanisms. Swift (1975) favors submergence of mainland beach ridges as the most important mode of formation. Considering the trend of sea level rise throughout the Holocene, it is certainly the most feasible mechanism for explaining the evolution, if not the initial origin of most of the extensive barrier-island regions existing today. However, spit progradation parallel to the coast cannot be completely dismissed as a significant mode of origin, because it is also readily observed to be initiating, as well as modifying, barriers at the present time. Many extensive barrier-island chains of the present day probably have had a composite mode of origin, by both spit progradation and coastal submergence. Variations in sediment supply and wave climate could easily induce periodic spit progradation at specific localities while submergence of coastal ridges was occurring on a more regional scale.

Barrier islands are most prevalent in coastal settings which have the following characteristics: 1) a low-gradient continental shelf adjacent to a low-relief coastal plain, 2) an abundant sediment supply, and 3) moderate to low tidal ranges (Glaeser, 1978). Both the shelf and the coastal plain are composed of unconsolidated sediments, which are the material source for the building of barrier islands by nearshore processes. Glaeser noted that only 10 per cent of the world's barrier-islands are present along coastlines where tidal ranges exceed three metres. However it was Hayes (1975,1976) who focused attention on the importance of tidal range in controlling the occurrence and morphology of barrier-island systems. Hayes observed not only that barrier islands were rare on macrotidal coastlines (greater than 4 m tidal range), but that there were geomorphological differences between barrier islands of microtidal regions (less than 2 m tidal range) and those of mesotidal regions (2 to 4 m tidal range). In general, microtidal barrier islands are long and linear with extensive storm washover features (Fig. 3), and tidal inlets and deltas are of relatively minor importance. Mesotidal barrier islands are short and stunted, and characterized by large tidal inlets and deltas. Microtidal barriers are overwashed frequently by storm waves because of the lack of large enough tidal inlets to allow storm surges to flow past the barrier, rather than overtopping it (Hayes, 1976). According to Hayes, microtidal barrier islands can be consi-

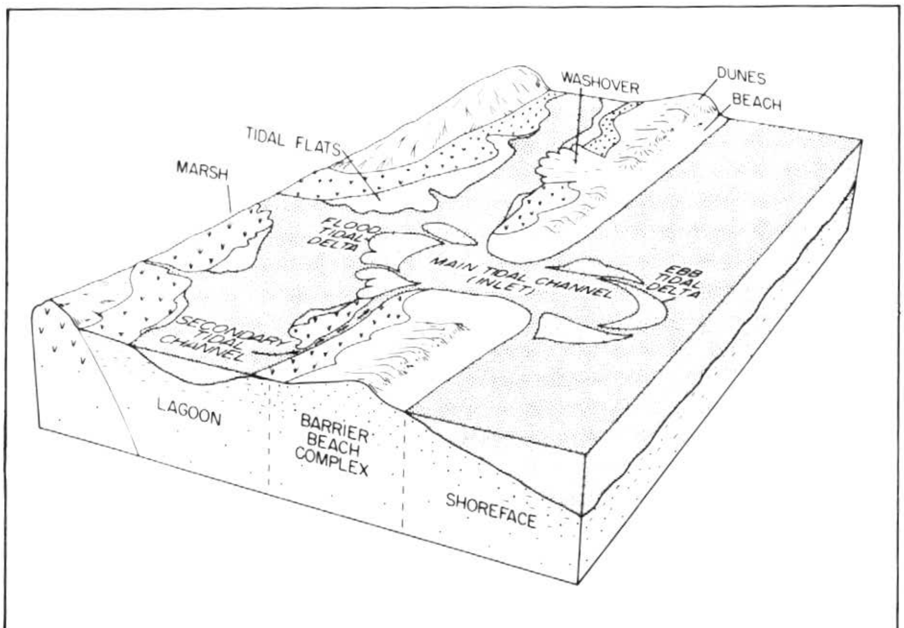


**Figure 2**
*Block diagram illustrating the various subenvironments in a barrier-island system.*


**Figure 3**
*Oblique aerial photo of the barrier beach at Tabusintac, New Brunswick, illustrating broad washover sand flats extending into the back-barrier marsh. The extent of the washover flats (W) is substantial relative to the dune ridge (D) (Photo by R. Belanger, May 3, 1977).*

dered to be wave dominated as opposed to mesotidal barriers which are affected by both wave and current processes.

**Depositional Environments and Lithofacies**

The three main environments of a barrier-island system (barrier beach, lagoon, tidal channel-delta complex) are made up of a number of subenvironments (Fig. 2), each of which is characterized by distinct lithofacies. Facies of the barrier-beach and channel-delta environments are mainly sand and gravel, whereas the lagoonal (back-barrier) deposits can consist of both mud and sand. Barrier-beach deposits are elongate bodies which parallel the strandline and enclose finer-grained deposits of the lagoon. Tidal-channel and delta sand deposits, on the other hand, are generally oriented perpendicular or oblique to the barrier complex, and can extend into the lagoon and seaward into the nearshore zone. The transition between lagoon deposits and barrier and channel-delta deposits occurs in the overlapping subenvironments of the back-barrier tidal flats, marsh, washover fans and flood-tidal deltas.

The lateral and vertical extent and the occurrence of specific facies within a barrier-island system is dependent upon tidal range and the relative importance of tidal-current versus wave-generated processes, as discussed previously. For example, tidal-flat deposits will not be an important facies in microtidal environments because of the limited tidal range, whereas they may be extensive in mesotidal environments. Similarly, tidal channel and delta deposits are likely to be more prevalent in mesotidal than in microtidal environments because of the stronger tidal currents generated by the larger tidal range. The following discussion covers all the depositional environments and corresponding deposits of barrier-island systems, but it should be kept in mind that all facies will not necessarily be present in every barrier-island deposit.

**Barrier Beach and Related Facies**

The depositional subenvironments of a barrier-beach complex include: 1) the subtidal zone or *shoreface*, 2) the intertidal zone or *beach (foreshore)*, 3) the subaerial zone or *back shore-dune* landward of the beachface, and 4) the supratidal to subaerial wave-and wind-formed *washover* flats which extend across the barrier into the lagoon. Shoreface deposits are discussed with the barrier-beach complex because they form the foundation for the barrier, and also are a major source of sediment for barrier-island accretion.

*Shoreface Deposits.* The shoreface environment is defined as the area seaward of the barrier from low tide mark to a depth of about 10 to 20 m (Fig. 4). The lower limits of the shoreface correspond to the position at which waves begin to 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.

Lower shoreface deposits occur seaward of the break in the shoreface slope at the toe of the barrier-island sediment prism. 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 (Howard, 1972).

Middle shoreface deposits extend over most of the shoreface slope (Fig. 4), in 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. 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. 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, especially in the lower parts, but the biogenic structures are generally less diverse than in deposits of the lower shoreface (Howard, 1972). The facies model proposed by Davidson-Arnott and Greenwood (1976) illustrates the complexity of sedimentary structures that can occur in a barred nearshore zone (Fig. 5). Vertical rock sequences of such deposits could

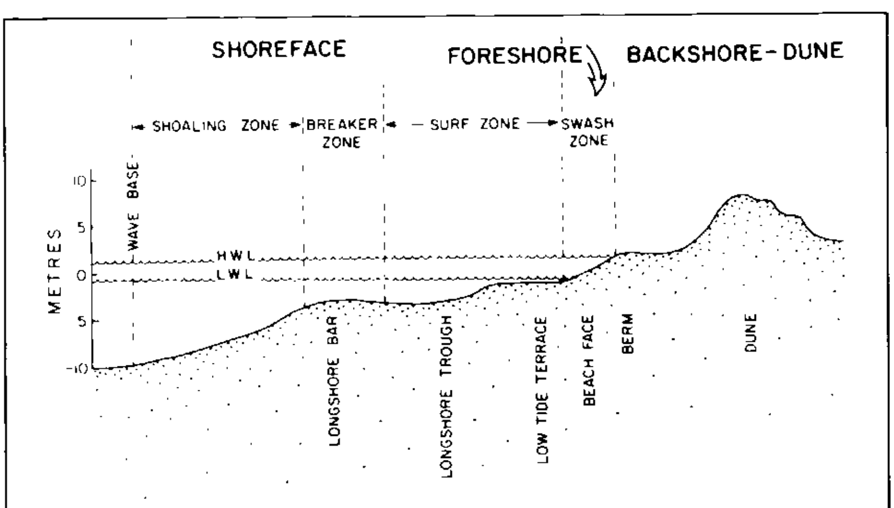


**Figure 4**
*Generalized profile of the barrier beach and shoreface environments.*

display interbedded sets of landward dipping ripple cross lamination, seaward dipping low-angle plane bedding, subhorizontal plane laminations and both landward- and seaward-dipping trough cross-bedded sets.

The shoreface environment is subjected to extreme modification by storm processes because effective wave base can be lowered dramatically by larger than normal, storm-generated waves. Truncated laminated bed sets and eroded burrow tops are common in ancient middle shoreface deposits (Howard, 1972), as are thick units (2 m) of subhorizontal laminated sand overlying coarse lag layers (50 cm thick) in modern deposits (Kumar and Sanders, 1976; Reineck and Singh, 1972). These features indicate that very high-amplitude waves periodically scour the bottom, suspending and then redepositing the sediment as the storm wanes. Graded bedding has been documented in some shoreface sediments (Hayes, 1967) and is attributed to storm-generated turbidity currents. Kumar and Sanders (1976) and Davidson-Arnott and Greenwood (1976) suggest that the bulk of shoreface deposits preserved in the rock record may consist of storm deposits rather than fair-weather deposits.

Upper shoreface sediments are closely associated with foreshore deposits, because they are situated in the high-energy surf zone just seaward of the beachface and landward of the breaker zone (Fig. 4). Consequently they have been grouped with foreshore facies in some rock studies (i.e., Davies *et al.*, 1971) and have been considered to represent the shoreface-foreshore transition zone in others (Howard, 1972). The complex hydraulic environment of the surf zone (i.e., shore-normal currents generated by plunging waves superimposed on shore-parallel wave-driven currents) gives rise to the complex sequence of multidirectional sedimentary structures and variable sediment textures characteristic of these deposits. Textures range from fine sand to gravel, and biogenic structures are common but not abundant. The predominant depositional structures are multidirectional trough cross-bed sets (15 to 45 cm thick) (Fig. 6), but low-angle bidirectional planar cross-bedded sets and subhorizontal plane beds may also be present. The trough cross-beds are

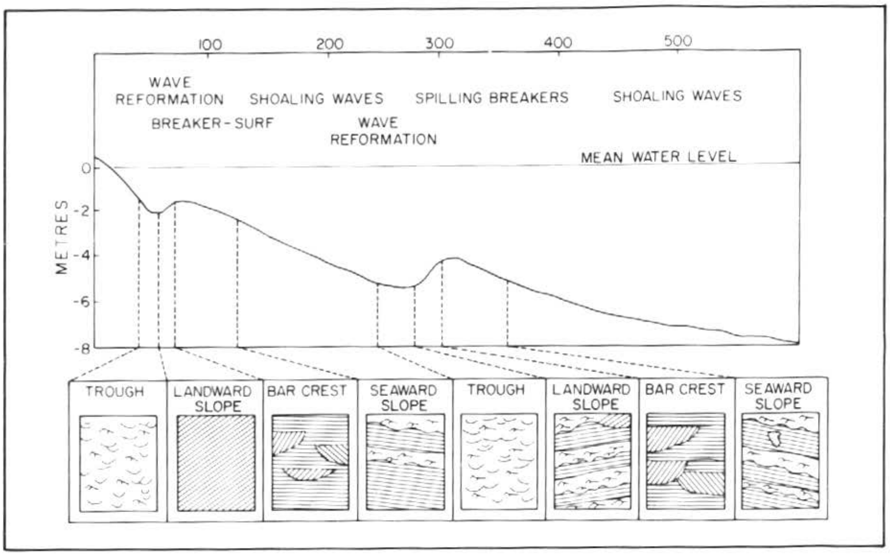


**Figure 5**
*Facies model of nearshore bars in Kouchibouguac Bay, New Brunswick, illustrating characteristic sedimentary structures and wave transformation zones (from Davidson-Arnott and Greenwood, 1976).*

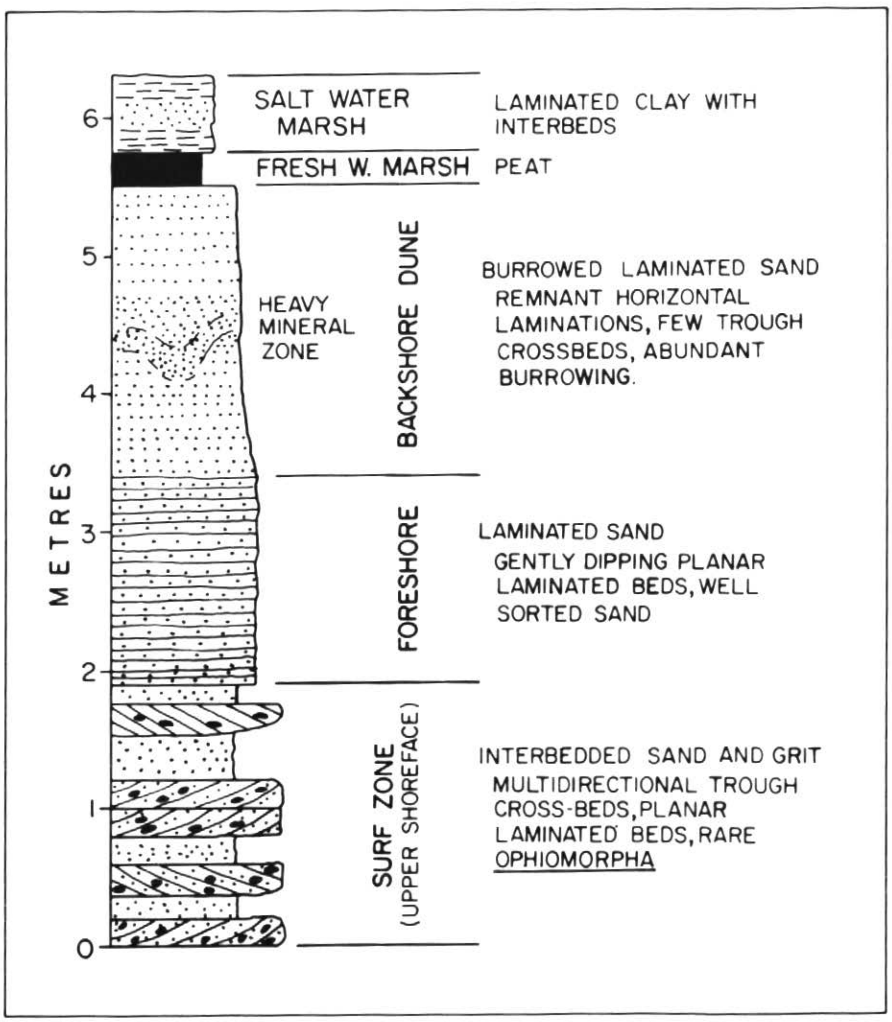


**Figure 6**
*Generalized barrier sequence in the Upper Tertiary Cohansey Sand of New Jersey (modified from Carter, 1978).*

thought to indicate the multidirectional current flow in the surf zone (Clifton *et al.*, 1971; Carter, 1978;). Predominantly bidirectional trough cross-beds oriented parallel to depositional strike are common in upper shoreface deposits, and may be indicative of deposition under strong longshore current conditions.

The effects of storm activity can be recorded in upper shoreface-foreshore deposits as well as in middle shoreface sediments. This is illustrated by the ridge and runnel sequence depicted in Figure 7. Such a sequence results when storm waves erode the beachface, removing sediment to the shoreface. The sediment is returned to the beach during the post-storm recovery, in the form of a ridge and runnel (bar and trough), which develops on the low tide terrace just seaward of the foreshore (Davis *et al.*, 1972; Owens and Frobel, 1977). The ridge migrates shoreward eventually welding onto the beachface and creating a distinctive sequence of upper shoreface-foreshore deposits.

*Foreshore Deposits.* The foreshore environment is confined to the intertidal zone, which is usually marked by a sharp change in slope, both at the base and at the top of the beachface (Fig. 4). The foreshore is the zone of wave swash, the surge of water caused by incoming plunging breakers in the surf zone. Swash runup occurs with each wave surge and backwash runoff between each surge. The swash-backwash mechanism is mainly responsible for the distinct subparallel to low-angle seaward-dipping, planar laminations (Fig. 8) which occur as wedge-shaped sets in most beach deposits. The boundaries between sets are generally not truncated, but rather mark the changing slope of the prograding beachface during the accretionary phase. Examples of foreshore deposits in the rock record include those illustrated in Figures 6 and 8 and those proposed by Campbell (1971), Howard (1972), Davies *et al.* (1971), and Land (1972).

*Backshore-Dune Deposits.* The backshore-dune environment is characterized by subaerial, predominantly wind-generated depositional processes. The backshore seaward of the dunes is a flat-lying to landward-sloping area called the berm; the seaward limit, called

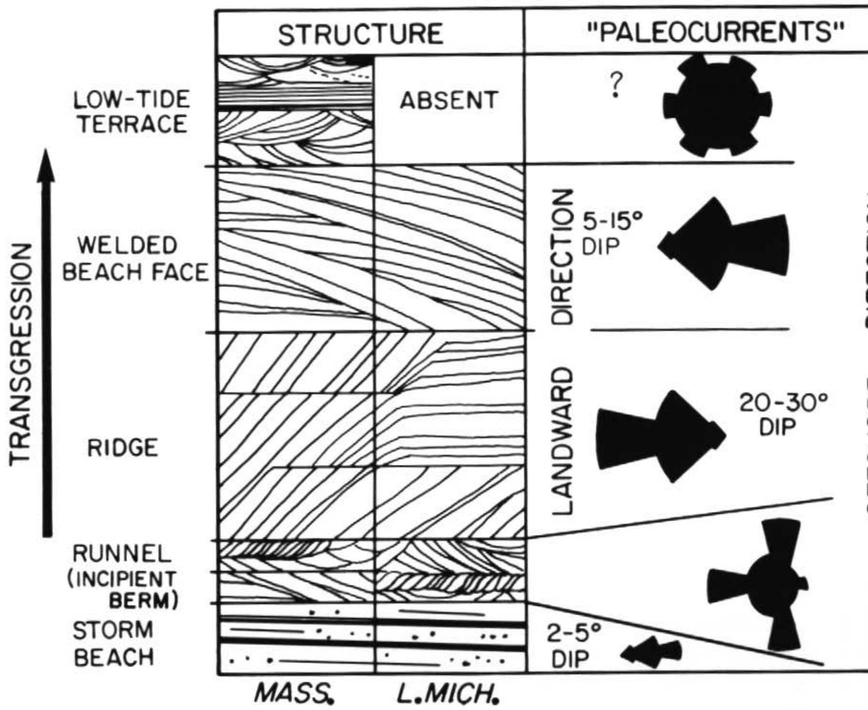


**Figure 7**
*Transgressive sequence formed by the landward migration of ridge-and-runnel during beach constructional phase. Vertical sequence would be about 1 m thick (from Davis* et al., *1972).*

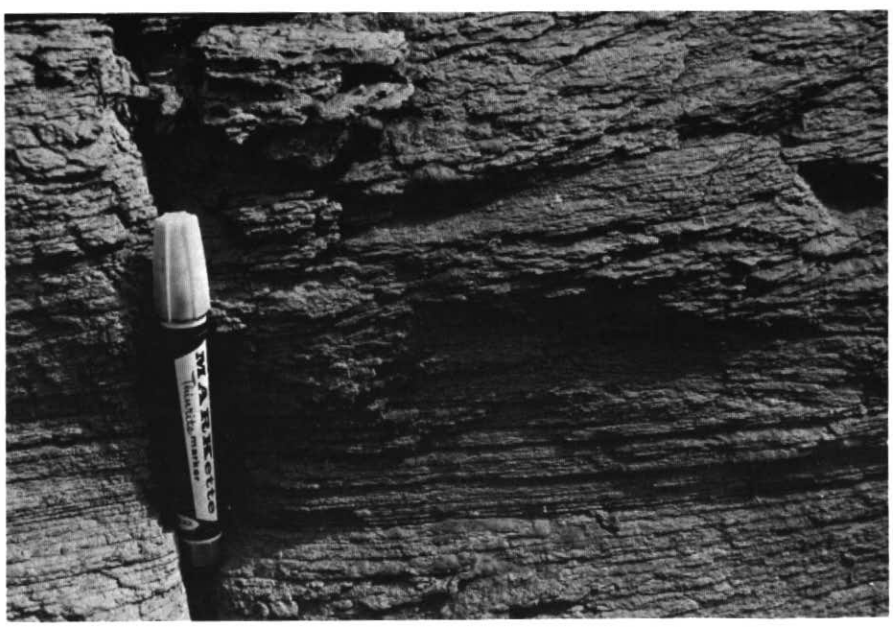

**Figure 8**
*Horizontal even parallel laminae and low-angle inclined laminae in beach deposits of the Foremost Formation, southern Alberta (from Ogunyomi and Hills, 1977, plate 1C).*

the berm crest, is well defined by the marked change in slope at the top of the beachface (Fig. 4). Sediment is transported to the berm crest by high spring tides or storms and is distributed over the backshore area by winds and washover (discussed below). Subhorizontal to landward-dipping plane beds characterize the backshore (Fig. 9) and may be interbedded or overlain by small- to medium-scale trough cross beds of incipient dune origin. Trough cross-stratified sets, up to 2 m in thickness, are characteristic of dune deposits, but planar cross-stratified sets are also common. The trough cross strata may be multidirectional in orientation and bounded by curved bedding surfaces (Campbell, 1971). Dune beds are commonly extensively disturbed by root growth (Figs. 9, 10) and may contain small paleosol horizons and isolated organic debris. Other biogenic structures such as decapod burrows may also occur in backshore-foredune deposits (Fig. 6).

*Washover Deposits.* Washover deposits result when wind-generated storm surges overtop and cut through barriers creating lobate or sheet deposits of sand which extend into the lagoon (Figs. 2, 3). These washover flats then provide corridors for transferring wind-transported sand across the foredune belt to form back-barrier sand flats (Fig. 9). This mechanism increases the width of the barrier, providing environments favorable for stabilization by marsh growth.

Modern studies of washover deposits indicate that there are two dominant sedimentary structures, subhorizontal (planar) stratification, and small- to medium-scale delta foreset strata where the washover detritus protrudes into the lagoon (Fig. 11). Textural and heavy mineral laminations and graded bedding can also occur (Andrews, 1970; Schwartz, 1975) 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.

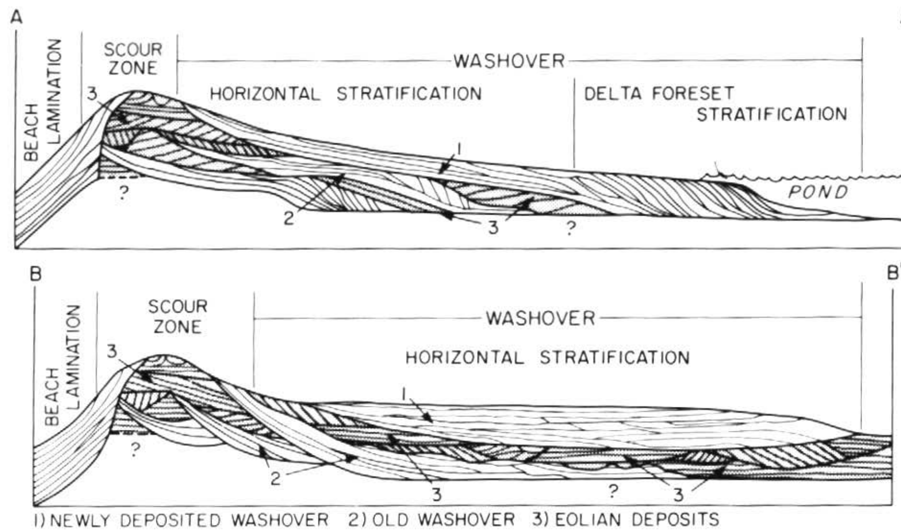


**Figure 11**
*Schematic cross-sections through two washover fans showing sequences of sedimentary structures. A "transgressive" situation is depicted. The sedimentary structures represent the upper few meters or less of the sand-body complex. Section A-A' shows a horizontal stratification to delta-foreset structural sequence resulting from flow across a subaerial surface into a body of standing water. Section B-B' shows occurrence of horizontal stratification resulting from flow across a subaerial surface. Eolian processes may modify or bury washover deposits to various degrees (from Schwartz, 1975).*

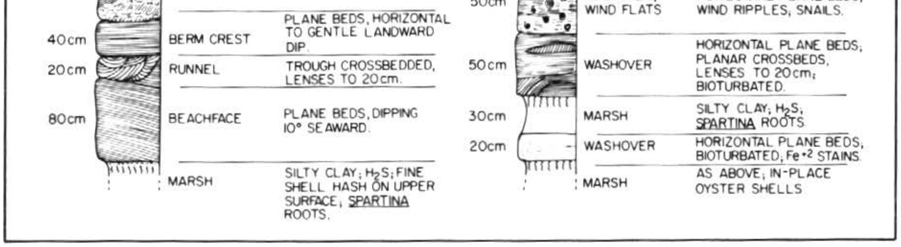


**Figure 9**
*Lithologic sequences observed in the landward edge, and in the seaward edge of a Kiawah Island beach ridge (from Barwis, 1976, Figs. 2 and 3).*

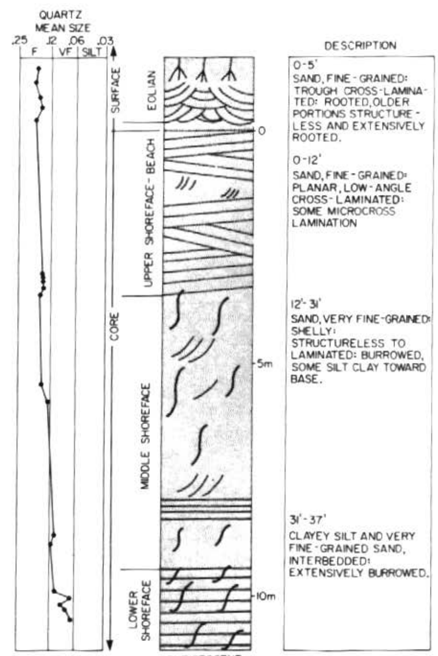


**Figure 10**
*Sequence of sedimentary structures, textures and lithology in a core through Galveston Island (From Davies* et al.*, 1971, Fig. 6).*

Washover deposits are generally thin, ranging from a few centimeters to two metres for each overwash event. In plan form they form elongate, semi-circular, sheet-like or tabular bodies a few hundred metres in width and oriented normal to the shoreline. Coalescing washover fans can be in the order of kilometres in width (Fig 3), 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 probably is 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.

**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 their internal sedimentary structures and textures. This is because their formation is governed primarily by tidal-current processes directed normal or oblique to the sand barrier. The *ebb-tidal delta*, the sand accumulation formed seaward of the barrier by ebb-tidal currents, is affected by longshore and wave-generated currents, whereas the *flood-tidal delta*, the sand body deposited landward of the barrier by flood-tidal currents, is little influenced by wave and wind-generated processes (Figs. 2, 12).

There are two types of tidal-channel environments, the main channels, or tidal inlets connecting the lagoon to the ocean, and the secondary channels located adjacent to the tidal deltas and back barrier lagoon margins. Secondary tidal channels are sometimes so closely related to the formation of tidal delta complexes that the resultant facies are difficult to differentiate. Tidal channel and tidal delta deposits are separated here mainly for ease in discussion. However, this separation serves also to emphasize the fact that channel facies can occur independent of tidal deltas, whereas the occurrrence of tidal delta facies is dependent on the presence of tidal channels.

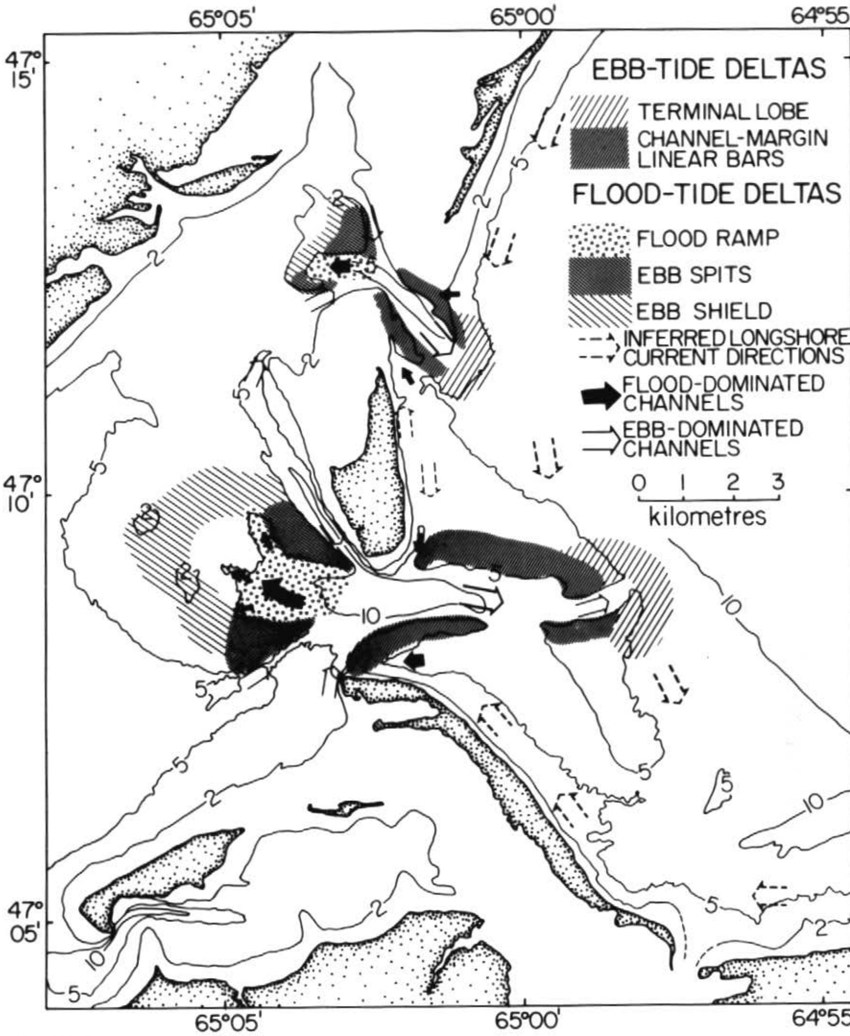


**Figure 12**

*Morphology of tidal deltas and inferred tidal- and longshore-current patterns at the mouth of the Miramichi estuary, New Brunswick (from Reinson, 1977).*

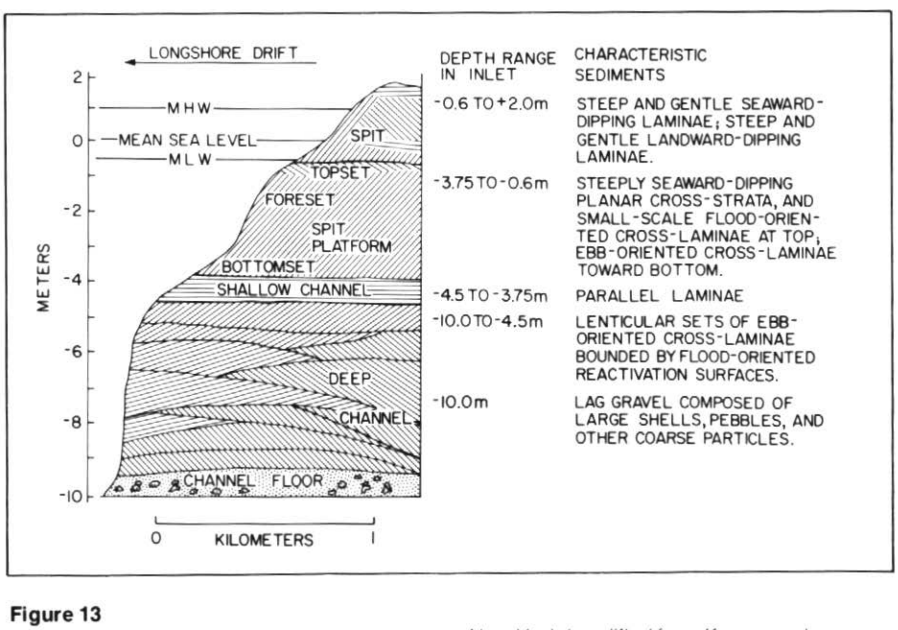


**Figure 13**

*Vertical sequence of sedimentary structures formed by the migration of Fire Island Inlet, New York (modified from Kumar and Sanders, 1974).*

*Tidal Channel Deposits.* Tidal channel deposits form mainly by lateral migration, as in a meander bend in a river. The best known and most important channel deposits are tidal-inlet fill sequences, which result from the shore-parallel migration of tidal inlets (Fig. 13). The direction and rate of inlet-channel migration is controlled by the magnitude of net longshore sediment supply. Barrier extension occurs by spit accretion on the updrift side of an inlet, with a corresponding erosion of the downdrift channel margin (Fig. 14). The shifting of the main inlet through a barrier causes the tidal channels both landward and seaward of the barrier, and the tidal deltas, to shift position also. The sand body that is deposited by inlet migration will be elongated parallel to the barrier island, having a length equal to the distance the inlet has migrated (Fig. 14). The thickness of the inlet lithosome will be equal to the depth of the inlet, if no subsequent erosion of the upper boundary occurs during deposition of dune, beach and washover deposits of the overlying accreting barrier.

The studies of Land (1972), Kumar and Sanders (1974), Hubbard and Barwis (1976), Hayes (1976), Barwis and Makurath (1978), and Carter (1978) indicate that channel-fill sequences resulting from barrier-inlet (or tidal channel) migration have the following general characteristics: 1) an erosional base often marked by a coarse lag deposit; 2) a deep channel facies consisting of bidirectional large-scale planar and/or medium-scale trough cross-beds; 3) a shallow channel facies consisting of bidirectional small- to medium-scale trough cross-beds and/or plane-beds and "washed-out" ripple laminae; 4) a fining-upward textural trend and a thinning-upward of cross-bed set thickness. The difference in size, orientation and type of sedimentary structures in the deep channel and shallow channel deposits generally reflects an increase in current-flow conditions in the shallow channel relative to the deep-channel environment.

The modern inlet-fill sequence (Fig. 13) described by Kumar and Sanders (1974) has a deep channel facies characterized by ebb-oriented planar cross-laminae; this reflects the predominance of sand-wave bedforms deposited under lower flow regime conditions in an ebb-current dominated environment. The overlying shallow channel facies is characterized by plane-parallel laminae and "washed-out" ripple laminae, reflecting plane bed deposition under "transitional" or upper flow regime conditions.

The studies of Barwis and Makurath (1978) and Land (1972) serve as comparative rock analogs to the Fire Island Inlet deposits (Fig. 13). The Silurian inlet sequence of Barwis and Makurath consists of a channel lag deposit overlain by 4.1 m of bidirectional trough and planar cross-bedded, medium-grained sandstone, with set thickness averaging 15 cm. The cross-bed orientations reflect deposition under tidal-current transport reversals along an axis oblique to the paleostrand. This deep channel facies is overlain gradationally by a fine-grained sandstone unit (1.8 m thick) dominated by bidirectional trough cross-bed sets averaging 2.5 cm in thickness, and "washed-out" ripples. The tidal channel sequence described by Land (1972) averages 8 m in thickness and consists of bimodal to polymodal trough cross-bed sets (ranging from 10 cm to 90 cm in thickness) in the lower 5 to 6 m, and subparallel beds in the upper 2 to 3 m.

The hypothetical tidal-inlet sequences proposed by Hayes (1976) and Hubbard and Barwis (1976), based on their study of mesotidal inlets of South Carolina, differ slightly from the Kumar and Sanders model (Fig. 13) with regard to the vertical sequence of sedimentary structures. However, the inference of sequential deposition under increasing flow conditions is still evident. Their inlet sequence is as follows: 1) a basal lag or disconformable bottom; 2) a deep channel deposit consisting of bidirectional large-scale planar cross-beds that have a slight seaward dominance, interlayered with bidirectional medium-scale trough cross-beds; 3) a shallow channel deposit consisting predominantly of small- to medium-scale bidirectional trough cross-beds. The planar cross-beds are suggestive of sand-wave deposition under ebb-dominant channel flow, whereas the trough cross-beds record deposition as megaripples under stronger currents and alternating reversals of flow directions.

Rock sequences similar to the hypothetical inlet sequences of Hayes (1976) and Hubbard and Barwis (1976) are illustrated in Figures 15 and 16. Carter (1978) interprets the sequence in Figure 15 as a back-barrier tidal channel deposit, and the presence of the interbedded sand and clay facies seems to preclude an inlet-fill origin. This example illustrates the similarity between back-barrier tidal channel deposits and tidal inlet deposits, and also points to a similar mode of origin, that of lateral channel migration concomitant with barrier-inlet migration. The tidal channel sequence in Figure 15 could also be interpreted as part of a flood-tidal delta complex, and the reasons for this alternate hypothesis will become evident in the following discussion on tidal delta deposits.

*Tidal Delta Deposits.* Hayes (1975), based mainly on his work in New England and Alaska, recognized that tidal deltas display a common morphological pattern governed by segregated zones of ebb and flood flow. This recognition prompted him to propose generalized models for both ebb- and flood-tidal delta deposition. Subsequent studies by Hayes and co-workers (Hayes and Kana, 1976) and others (Reinson, 1977, Armon, 1979), indicate that the models are generally applicable elsewhere, in both microtidal and meso-

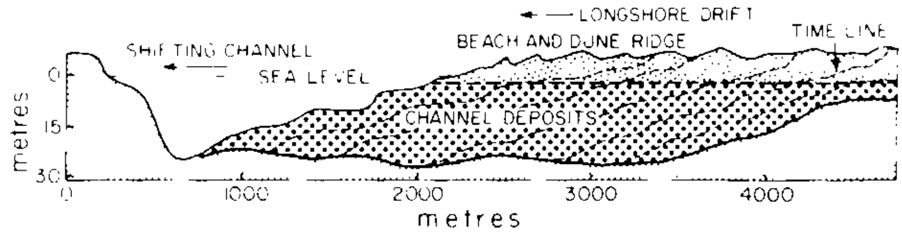


**Figure 14**
*Generalized cross-section parallel to shoreline illustrating the development of a barrier inlet sand body by lateral inlet migration (modified from Hoyt and Henry, 1965).*

tidal regions. Tidal deltas can occur in a variety of forms (from linear shoals to complex channel-shoal systems) depending on tidal range, wave climate, and sediment supply, but the basic morphological pattern as illustrated in Figure 12 is generally clearly evident.

The typical morphology of a flood-tidal delta, that of a seaward-opening parabola bounded by marginal channels, is related to the segregation of tidal-current flow paths during ebb and flood phases. This flow segregation results from the time-velocity asymmetry of the tidal currents; that is, maximum flood and ebb flows occur near high water and low water respectively. Maximum flood flow traverses through the flood ramp and over the shoal, whereas maximum ebb flow is diverted around the shoal because of the drop in water level. This flow segregation gives rise to a distinct pattern of bedform distribution (Boothroyd and Hubbard, 1975; Hubbard and Barwis, 1976; Reinson, 1979), with predominantly flood-oriented sand waves covering the flood ramp and centre of the shoal, bidirectional megaripples on the ebb-shield and ebb spits, and ebb-oriented sand waves in the adjacent channels.

The deposits resulting from flood-tidal delta formation will be characterized by a varied sequence of planar cross-beds and trough cross-beds. The preponderance of one bedform over the other, and their orientation and position in vertical sequence, will depend on the locality at which the sequence is located within the tidal-delta complex. Hubbard and Barwis (1976) proposed a lithologic sequence for a flood-tidal delta as follows: 1) basal bidirectional cross-strata (megaripples) - represents early phases of deposition; 2) interbedded seaward-oriented trough cross-strata (megaripples) and landward-oriented planar cross-beds (sand waves) - represents deposition prior to ebb-shield development; and 3) landward-oriented planar cross-strata with upward-decreasing set thickness (sand waves) - represents deposition on flood ramp. Deposits adjacent to this sequence would be characterized by bidirectional trough cross-strata (megaripples) - representing ebb-shield and ebb-spit deposition. The total thickness of such a sequence would be in the order of 10 m. Hayes (1976) proposed a stratigraphic sequence for a regressive flood-tidal delta situation (Fig. 17). This sequence is dominated by planar bidirectional cross-strata.

The morphology of ebb-tidal deltas is controlled largely by tidal-current segregation during different phases of the tidal cycle, but the interaction of waves with tidal currents is also important in the formation of ebb deltas. This interaction is reflected in the complex bedform distribution, which consists of ebb-oriented sand waves or megaripples in the main ebb-channel, with flood-oriented sand waves or megaripples in the marginal flood channels (Fig. 12). Channel-margin, linear bars and swash bars (areas of intense wave and current interaction) are characterized by multidirectional megaripples and plane beds. As in flood deltas, the vertical sequences resulting from ebb-delta formation would exhibit extreme variations in sedimentary structures from one locality to another, within a specific ebb-delta deposit. Ebb-delta deposits are so dependent on inlet conditions and wave climate that it is imposssible to characterize them in a specific sequence. Perhaps the major difference between ebb-tidal delta deposits and flood-delta deposits is the occurrence of multidirectional cross-beds in ebb delta sequen-

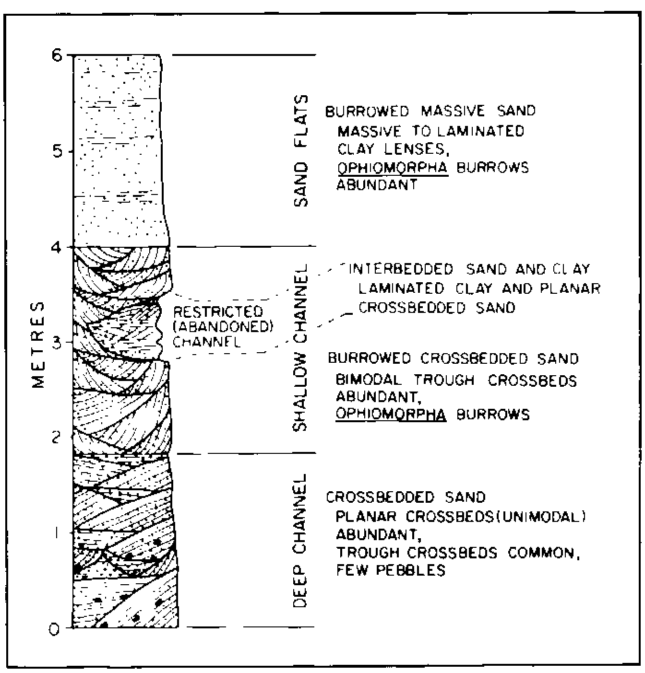


**Figure 15**
*Generalized barrier-protected sequence in the Upper Tertiary Cohansey Sand of New Jersey (modified from Carter, 1978).*

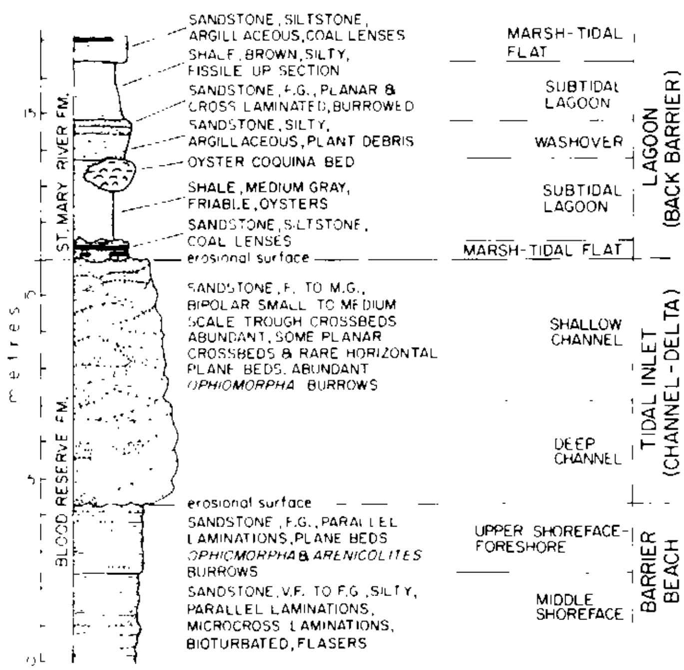


**Figure 16**
*Composite stratigraphic section of the Blood Reserve - St Mary River Formations, southern Alberta (modified from Young and Reinson, 1975)*

ces, as opposed to the predominantly flood-oriented or bidirectional cross-beds of flood-tidal delta sequences.

As mentioned earlier, flood- and ebb-delta deposits have textures and sedimentary structures similar to inlet fill sequences, and therefore the identification of delta sand bodies in the rock record may depend more on their geometry and stratigraphic position relative to surrounding facies. In modern barrier systems, tidal inlet-delta associated deposits are integral parts of barrier-island sand bodies. By analogy such deposits shoud be expected to occur in ancient barrier sequences, yet they have been little recognized up until very recently. The studies of Barwis and Makurath (1978), Horne and Ferm (1978), Hobday and Horne (1977) and Land (1972) amply illustrate the importance of channel-delta deposits in ancient barrier sequences, and also lead one to suspect that such deposits have been misinterpreted in many rock sequences in the past.

**Lagoonal (Back Barrier) Facies**

Lagoonal sequences generally consist of interbedded and interfingering sandstone, shale, siltstone and coal facies characteristic of a number of overlapping subenvironments (Figs. 16, 18). Sand facies include *washover* sheet deposits and sheet and channel-fill deposits of *flood-tidal delta* origin. Fine-grained facies include those of the *subaqueous lagoon* and the *tidal flats,* which are situated adjacent to the barrier or on the landward side of the lagoon abutting the hinterland marsh and swamp flatland (Fig. 2). Organic deposits of coal, peat, etc., record *marsh* and *swamp* environments, and usually are very thin, having formed on sand and mud flats of the lagoonal margin, and on emergent washover flats. Abandoned or mature flood-tidal deltas can become stabilized by marsh vegetation also; this situation and that of the vegetated washover flat can lead to the presence of very thin coal lenses overlying organic-rich sheet sandstones in the rock record (Fig 18). Subaqueous shale and siltstone facies are often characterized by brackish water macroinvertebrate shells, and in Cretaceous lagoonal deposits, coquinid oyster beds up to 1 m thick, are common (Fig 16 and Land, 1972). Disseminated carbonaceous material, imprints of plant remains, and root and reed fragments are common in some shale beds, indicating the interfingering of proximal marsh and subaqueous lagoonal environments.

The topic of tidal-flats cannot be given justice here, but some mention is made of these deposits because they do occur in the barrier-island setting. The extent of tidal flat environments in a barrier-island system is a function of tidal range, the greater the tidal range, the more extensive are the flats So in mesotidal barrier systems we may expect to find sequences similar to the classical tidal flat deposits described by van Straaten (1961), Evans (1965), Reineck and Singh (1975) and Klein (1977). The low tidal flats would be characterized by fine- to medium-grained ripple-laminated sand, the mid flats by interbedded sand and mud containing flasers and lenticular layers, and the high tidal flats by layered mud. The high tidal flats would be succeeded landward (and upwards in a prograding situation) by salt marsh. In most microtidal and mesotidal barrier-island systems the above tidal-flat sequence is attenuated because of the limiting conditions of tidal range.

Lagoonal or back-barrier sequences present a marked contrast to the predominantly clean sandstone sequences of the barrier-beach and inlet-delta environments (Fig. 19). Although the sandstone deposits interfinger with the fine-grained lagoonal deposits in the back-barrier marsh, tidal flat, washover, and flood-tidal environments, this lateral facies change from sandstone to siltstone and shale is still relatively abrupt.

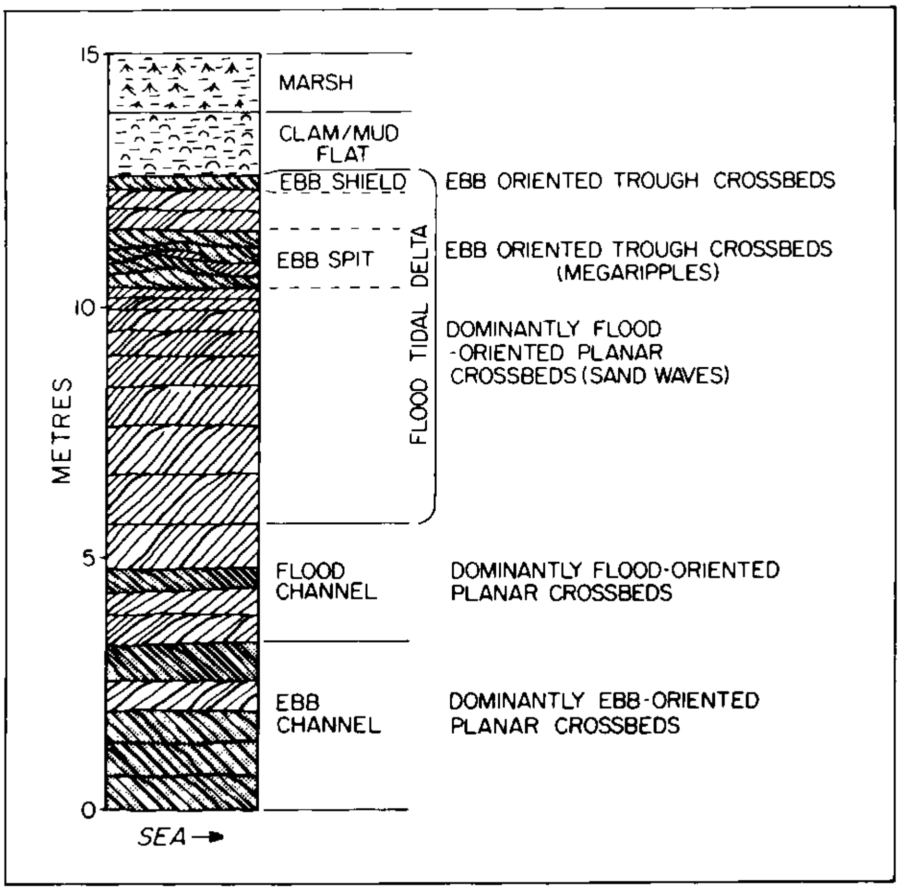


**Figure 17**
*Hypothetical regressive sequence for a mesotidal flood-tidal delta complex (modified from Hayes, 1976)*

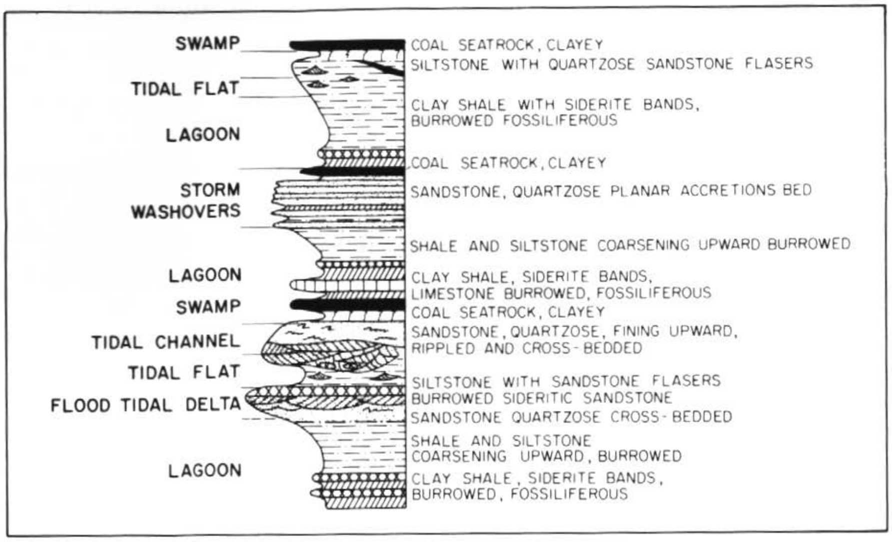


**Figure 18**
*Generalized lagoonal sequence through back-barrier deposits in the Carboniferous of eastern Kentucky and southern West Virginia. Such sequences range from 7.5 to 24 m thick (from Horne and Ferm, 1978, Fig. 11).*

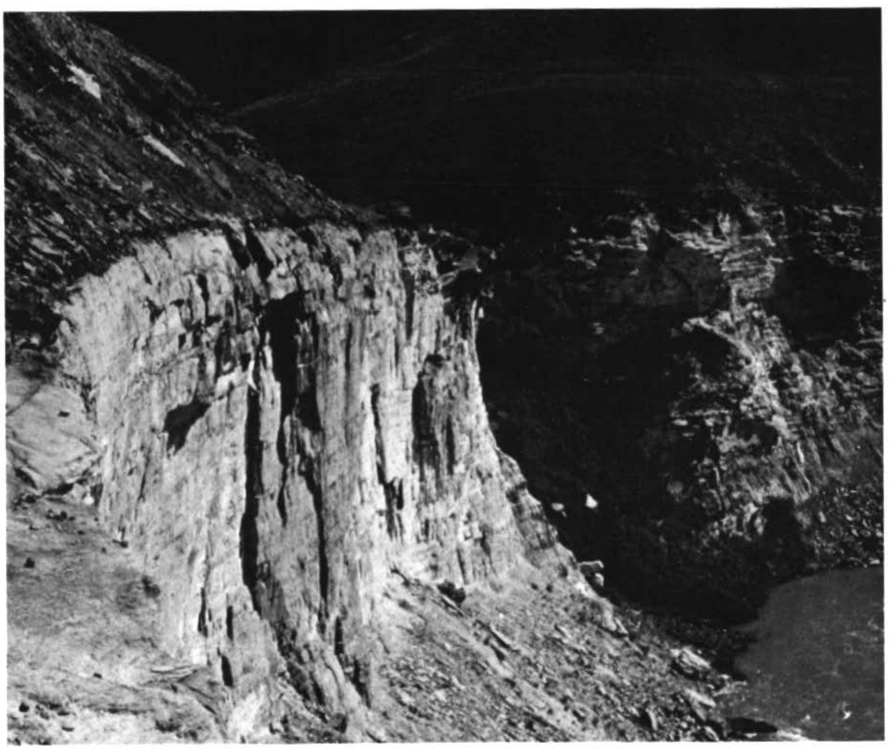

**Figure 19**
*Photo showing the sharp contact between the clean sandstones of the Blood Reserve Formation and the overlying finer-grained lagoonal deposits of the St. Mary River Formation (vertical sequence illustrated in Fig. 16).*

## Stratigraphic Sequences and Depositional Models

### Transgression, Regression, and Preservation of Facies

The preservation of specific barrier facies is dependent upon a number of factors including sea level fluctuations, sediment supply, inlet conditions (migrating or stable) and wave climate. The most important condition, dependent largely on sea-level fluctuations, is the nature of the shoreline in terms of transgression or regression. The concepts of transgression and regression as used by geologists usually refer to the overlapping of deeper water deposits over more landward or shallower-water deposits (transgressive sequence), or shallow water deposits over more marine or deep-water facies (regressive). The terms "transgression" and "regression" are also used to imply the process of migration of the shoreline of a water body, in a landward direction (transgression), or in a seaward direction (regression) (Curray, 1964). Generally, transgressive and regressive barrier-shoreline migrations produce corresponding simple transgressive and regressive overlapping sequences (Figs. 20, 21), but this is not always the case. This is because shoreline migrational trends can be "regional" or they can be "local". Régional transgressive shoreline trends can be caused by relative sea-level rise such as is now occurring on the Atlantic coast of the United States, or they can occur by shoreline erosion under relatively stable sea level conditions, in areas where sediment supply is cut off and wave attack is intensified (Kraft, 1978).

Klein (1974) suggests that transgressive sequences have a low preservation potential relative to regressive sequences. Klein bases his suggestion on the fact that along transgressive coasts a thin basal transgressive interval (ravinement deposit) is often preserved and buried by regressive sediments. However, Kraft (1971) contends that the possibility exists for a complete transgressive sequence to be preserved, and that the relative rate of sea level rise will govern the amount of preservation. If transgression occurs largely by shoreface erosion with little or no relative sea level rise, almost total loss will result (Fig. 22). Conversely, if sea level rise is rapid, almost total retention of the sedimentary sequence can be expect-

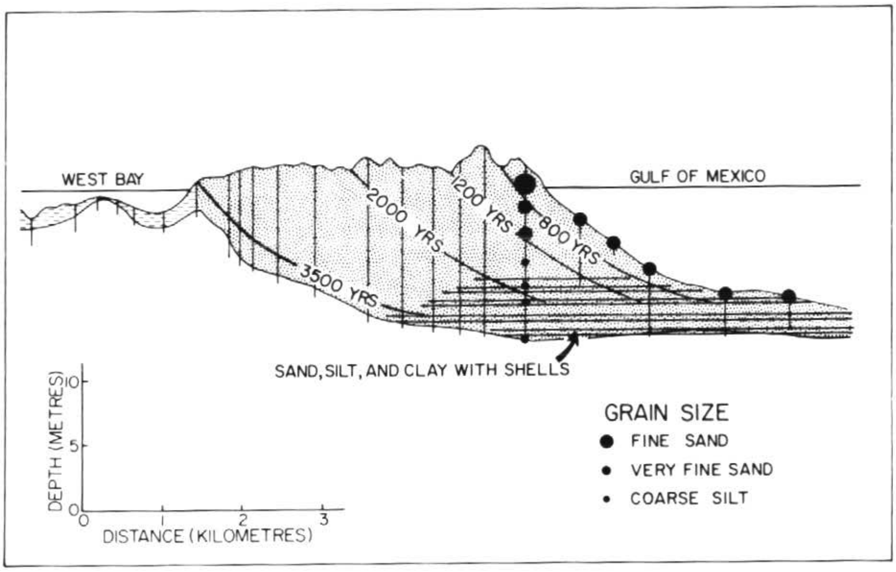


**Figure 20**
*Cross section of the prograding Galveston barrier island (from Bernard* et al., *1962).*

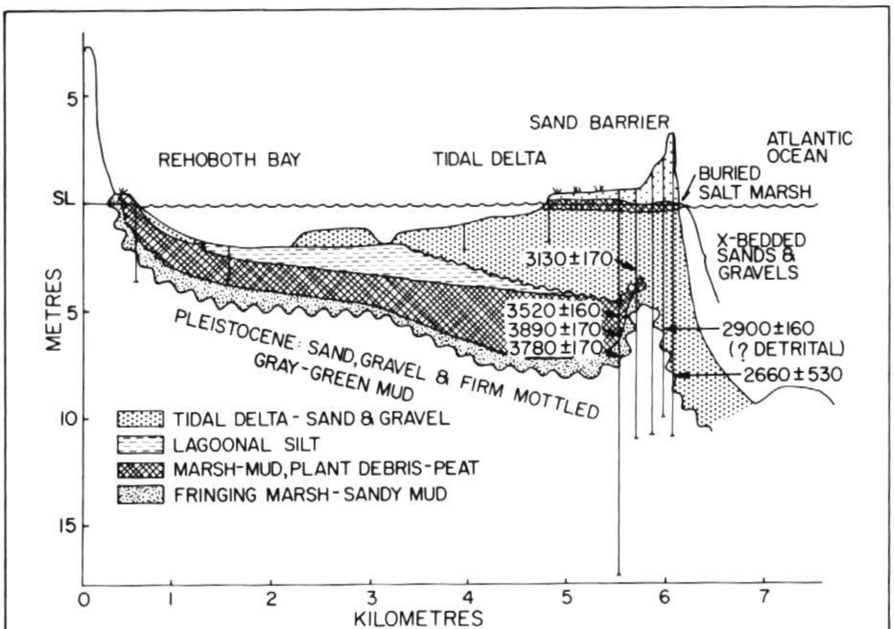


**Figure 21**
*Cross-section of the Delaware barrier coast in the vicinity of a tidal delta, showing the transgressive nature of the Holocene sequence (modified from Kraft, 1971, Fig. 16).*

ed. Therefore, under rapid relative sea level rise, it is conceivable that most of the facies, save for the upper portions of the back shore dunes, could be preserved. Under conditions of erosive shoreline retreat (transgression by erosion) most of the barrier facies will be destroyed, save for the reworked sediments of the ravinement lag deposit.

Dominantly transgressive shorelines can have "local" regressive segments within them. Such situations are caused by short-term temporal variations in depositional conditions along the barrier-island strandline. Longshore sediment supply, local wave climate, and number and location of tidal inlets are some of the conditions which can change significantly and can effect both progradational and erosional trends in near juxtaposition. This is illustrated by the beach sequences in Figure 9, the landward sequence being transgressive and the seaward sequence progradational or regressive. The Holocene studies of Kraft *et al.* (1978) also indicate that both transgressive and regressive shoreline trends could be inferred by two different vertical sequences in proximity. Given the presence of "local" regressive sequences in Holocene deposits, the possibility exists for their preservation in the rock record under conditions of rapid sea level rise. If such isolated stratigraphic sequences were encountered in the rock record they could be wrongly interpreted as being representative of the regional paleogeographic submergent or emergent conditions.

Certain facies have a higher potential for preservation than others because of their vertical position with respect to the intertidal zone (i.e., subtidal, intertidal), and their lateral position relative to the wave-dominated open coast or to a migratory inlet. Tidal inlet channel facies will probably have the highest preservation potential of all the sand facies because, depending on the depth of the inlet, they may extend well below low tide level, their basal portion thus being protected from shoreface erosion during transgressive conditions. In addition they would be protected above by the overlying beach-dune facies (Fig. 14). Flood-tidal delta deposits would have a high preservation potential as well because they are situated in the back barrier protected region (for the most part in the subtidal zone), and under

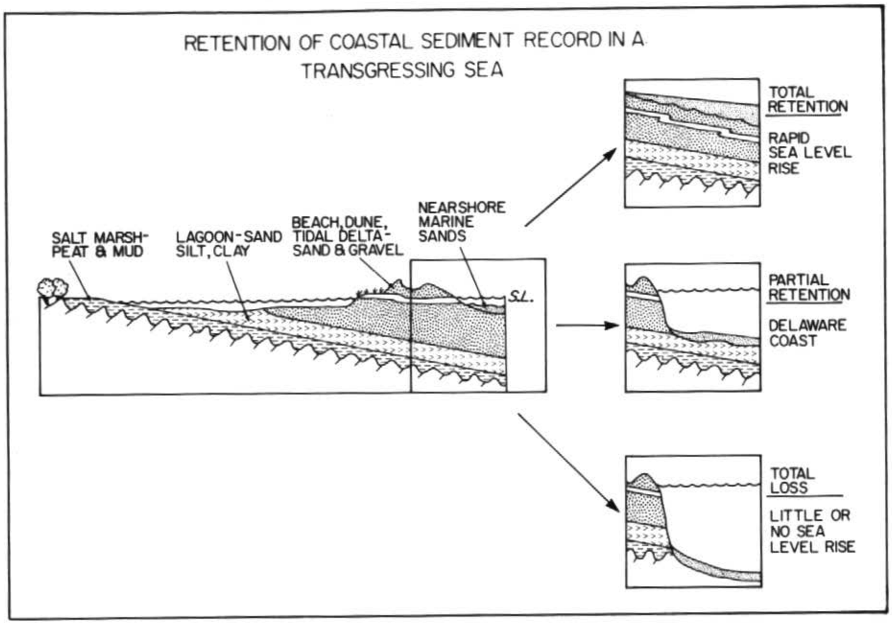


**Figure 22**
*Schematic diagram illustrating the variation in retention of the barrier-island sequence in a transgressing sea. Situations shown are those for rapid sea level rise, little or no sea level rise, and relative sea level rise-marine transgression as is occurring on the Delaware coast (from Kraft, 1971, Fig. 24).*

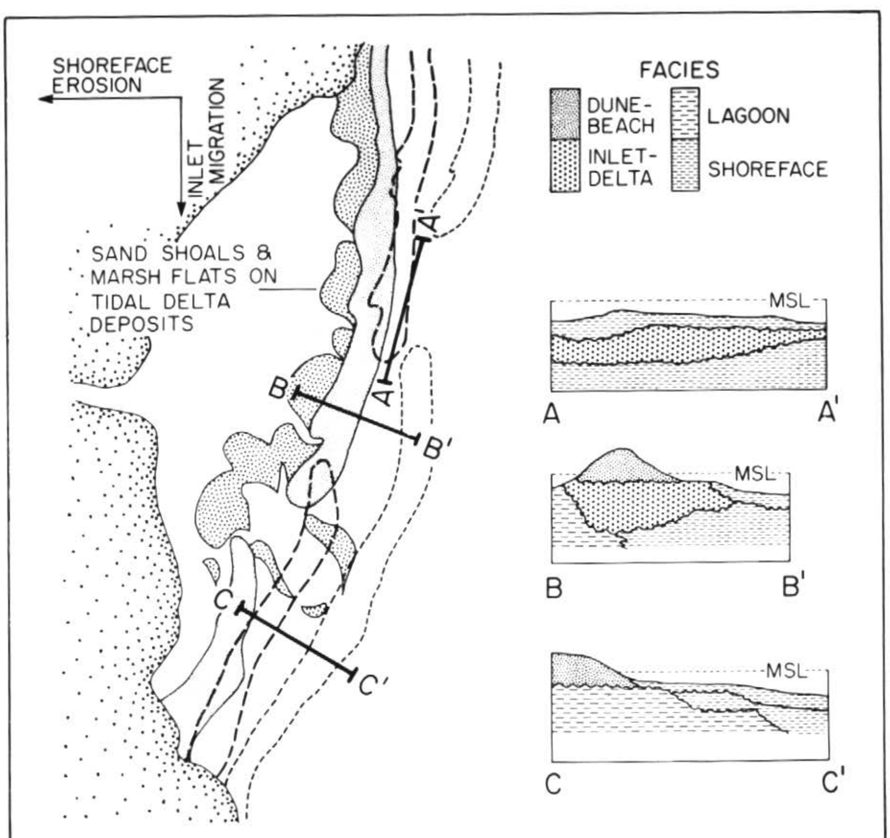


**Figure 23**
*Schematic diagram illustrating how an inlet-delta sand body bounded by disconformities could be formed under conditions of inlet migration and concomitant shoreface erosion. The occurrence and shape of the sandy body, and the nature of the enclosing sediments, vary with position in the barrier-island complex.*

migrating inlet conditions could form relict sand shoals disconnected permanently from the tidal-current conduit (Fig. 23). The distal portions of washover deposits, where they interfinger with lagoonal fine-grained facies, would have a high preservation potential, as would most of the lagoonal facies which are relatively protected by the seaward barrier. These deposits would be the last "to go" under intense shoreline retreat. Ebb-tidal delta deposits would have a low preservation potential under both migrating inlet conditions and transgression, because of exposure to reworking by longshore currents and onshore wave processes.

Regardless of potential for preservation, given the right combination of sediment supply, inlet stability, wave climate and sea-level fluctuations, any one of the barrier-island facies could be preserved in barrier-island stratigraphic sequences; and we should be prepared to encounter all of them in the rock record.

**Stratigraphic Models**

From the examples of modern and ancient sequences, it is obvious that there cannot be just one generalized facies model for barrier-island deposits. If we apply the facies model criteria of Walker (1976), three "end-member" models for barrier-island stratigraphic sequences can be recognized; the regressive (prograding) model, the transgressive model, and the migrating barrier-inlet model (Fig. 24).

*Regressive Model.* The distillation of the generalized regressive facies model has come from modern examples, particularly the Galveston Island model (Fig. 20). As mentioned previously, prior to 1970 this example was the "one and only" model accepted for use in interpreting ancient sequences. Such a situation arose because the study of Bernard *et al.* (1962) was one of the first to present a detailed stratigraphic model for a barrier-island system. The situation became analogous to the thinking on deltas; when the word "delta" was mentioned, geologists thought of the "Mississippi" (Miall, 1976). Similarly the "Galveston Island" model came to mind immediately at the mention of "barrier island". Some of the earlier literature depicted other stratigraphic models (i.e., Hoyt, 1967 - transgressive barrier, and

Hoyt and Henry, 1965 - migrating inlet barrier), but these were largely ignored in the wave of enthusiasm for the Galveston model. Galveston Island should be recognized for what it is, a good example of the regressive facies model. It is not adequate as a "norm", because it does not include the essential characteristics displayed by most modern middle and upper shoreface deposits (Figs. 5, 7).

The regressive facies model in Figure 24 serves as a norm for interpreting ancient regressive barrier sequences only. It is a gradational-based, coarsening upwards sequence, dominated by shoreface, foreshore, and backshore-dune facies of the barrier-beach complex. Ancient examples of regressive barriers include those of Davies *et al.* (1971), and Carter (1978) (Fig. 6).

*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 21. 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, due to the rapid response of depositional environments to change in sediment supply and inlet conditions in transgressive situations. A good example of an ancient transgressive sequence is the study of Bridges (1976). However, well-documented ancient examples of the transgressive facies model are few. This may reflect an overemphasis on the use of the regressive model in past literature.

One of the main differences between the regressive model and the transgressive model lies in the relationship with lagoonal facies. In the regressive sequence lagoonal deposits overlie the sand facies, whereas in the transgressive model lagoonal facies underlie, or are incorporated within the lower to middle portions, of the sequence (Fig. 24).

*Barrier-Inlet Model.* The distillation of the barrier-inlet model derives from very recent, well-documented modern and ancient examples (Figs. 13, 14, 16) including the studies of Land (1972), Hayes (1976), and Barwis and Makurath (1978). The barrier-inlet facies model is a fining-upwards sequence with a thinning-upwards trend in cross-bed set thickness (Fig. 24). It is characterized by an erosional base and dominated by sand facies of tidal-channel and marginal spit-beach environments.

*Examples of Hybrid Models.* Transgressive, regressive, and barrier-inlet depositional conditions can occur in combination to produce mixed sequences which have affinities with more than one "end-member" norm.

The so-called vertical build-up barrier of Padre Island is really a combination of the regressive and transgressive models, with the landward side of the barrier migrating into the lagoon by washover deposition, and the seaward side prograding outwards by beach-ridge accretion (Hayes, 1976; Dickinson *et al.*, 1972). Matagorda Island, situated near Galveston Island, has been shown by Wilkinson (1975) to have formed during both a transgressive phase and a subsequent regressive phase. Vertical sequences from the landward side of the Padre and Matagorda barriers are comparable to the transgressive model, whereas sequences from the seaward side are comparable to the regressive model.

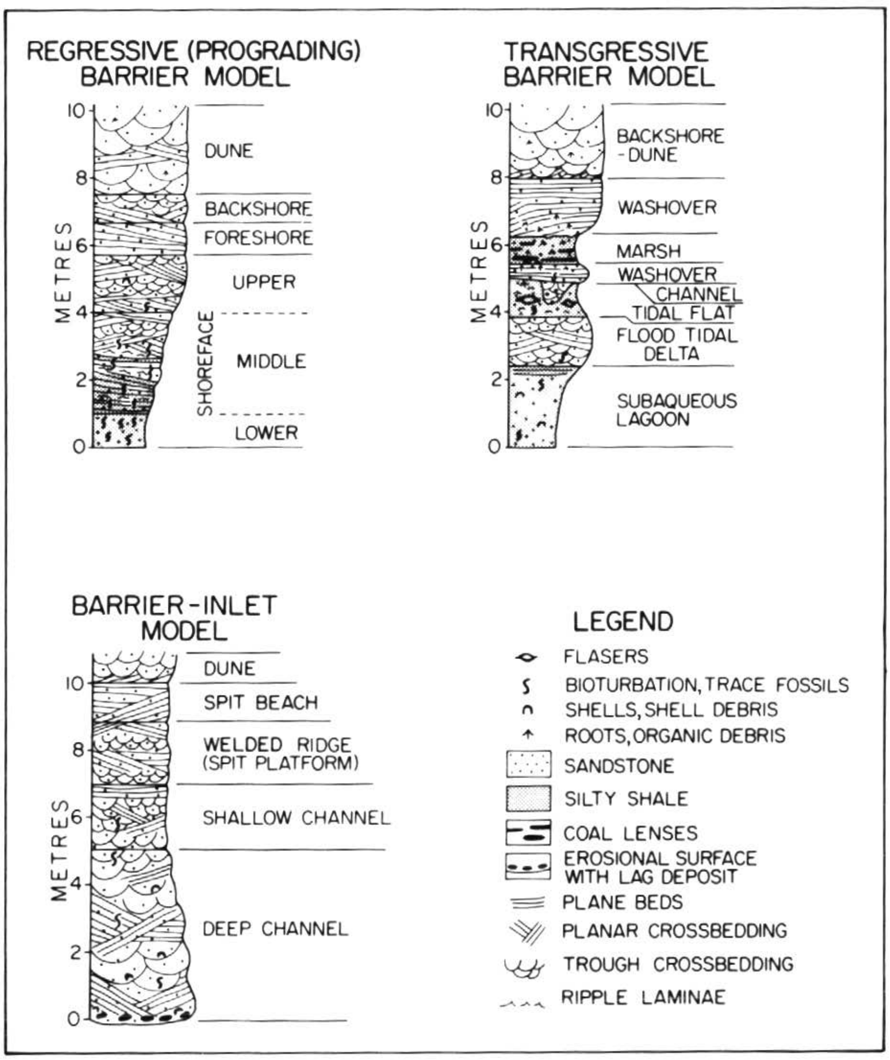


**Figure 24**
*The three "end-member" facies models of barrier-island stratigraphic sequences.*

An example of a combined regressive and barrier-inlet model may be the sequence illustrated in Figure 16. The top of the channel sequence is truncated by the overlying lagoonal deposits, and only the lower part of the barrier-inlet "end-member" model is preserved. The barrier-inlet "end-member" model (Fig. 24) may in itself be considered in part regressive because of the progradation of beach and dune facies over deeper-water channel deposits.

Perhaps the most complicated hybrid sequence that could occur is the situation of transgression concomitant with, or just after, barrier-inlet migration. Barwis and Makurath (1978) discuss the stratigraphic implications of such a setting in some detail. Basically, if transgression is occurring largely by shoreface erosion, and the migrating inlet is deep enough to produce an inlet deposit whose base is substantially below the foreshore-shoreface boundary, an inlet-delta sand body, bounded above and below by disconformities, could occur (Fig. 23). In vertical sequence such a deposit would be comparable to the lower part of the barrier-inlet "end-member" model, with an erosional surface similar to the basal lag, situated at the top. Kumar and Sanders (1970) consider that this dual migration setting could be the origin of many linear sand bodies on the inner shelf. They further suggest that submergence of migrating barrier-inlet shorelines may account for some of the basal transgressive sands in the geological record, the sands being of inlet-fill as opposed to barrier-beach or offshore bar origin.


## Summary

Prior to the 1970s, the prograding Galveston Island depositional model was in the forefront in the minds of most geologists, as the "one and only" facies model for use in interpreting ancient barrier-island sequences. Studies conducted within the last decade on modern and ancient barrier-island deposits, indicate that the "regressive facies model" cannot be applied to a number of barrier-island sequences, and therefore the use of one normative model is unrealistic. Three generalized facies models or "end-member" norms can be recognized: the regressive barrier model, the transgressive barrier model, and the barrier-inlet model (Fig. 24). Most sequences can be explained through comparative analysis with individual "end-member" models, or combinations of them, and this is emphasized by the recent studies of Horne and Ferm (1978) (Fig. 25), Barwis and Makurath (1978), and others.



## Acknowledgements

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## Selected Bibliography

The following reference list purposely emphasizes the most recent and summary-type papers on barrier-island systems. The reference lists contained in some of these articles are exhaustive, so the reader can become familiar with nearly all of the barrier-island literature. Unfortunately, some of the excellent summary studies on both modern and ancient barrier-island deposits are contained in field and course guidebooks; these are not easily accessible to the interested geologist.

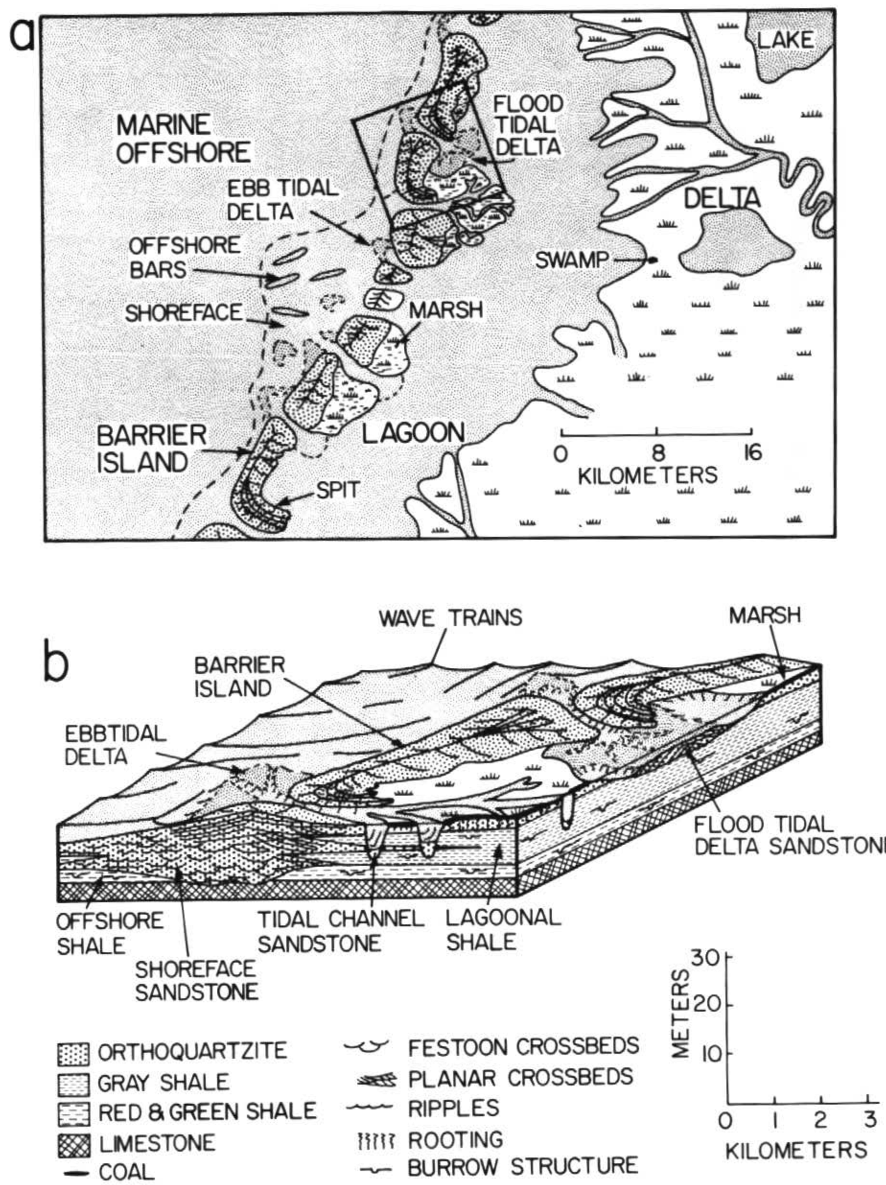


**Figure 25**
*Reconstruction of the depositional environments (from stratigraphic sequences and lateral relationships) in the Carboniferous Carter Caves Sandstone of Kentucky (from Horne and Ferm, 1978, composite of Figs. 21 and 22).*

**Essential Reference List**

**A. Holocene Barrier-Island Deposits**

Bernard, H.A., R.J. Leblanc, and C.F. Major, 1962, Recent and Pleistocene geology of southeast Texas: *in* Geology of the Gulf Coast and Central Texas and Guidebook of Excursions: Houston Geol Soc. - Geol. Soc. Amer. Ann Mtg., 1962, Houston, Texas, p. 175-205.

Documentation of the oft-cited (and illustrated) prograding Galveston barrier-island model. The material in this original paper is more accessible in the following summary article.

Bernard, H.A., and R.J. Leblanc, 1965, Resume of the Quaternary geology of the northwestern Gulf of Mexico Province: *in* H.E. Wright Jr., and D.G. Frey, eds., The Quaternary of the United States: Princeton University Press, Princeton, N.J., p. 137-185

Davis, R.A. Jr., ed., 1978, Coastal Sedimentary Environments: Springer-Verlag, New York, 420 p.

Contains excellent review chapters on modern beach deposits by R.A. Davis and Holocene coastal stratigraphic sequences by J.C. Kraft

Hayes, M.O., and T.W. Kana, ed., 1976, Terrigenous Clastic Depositional Environments, Some Modern Examples: Amer. Assoc. Petrol. Geol Field Course, Univ. South Carolina, Tech Rept. No. 11-CRD, 315 p

This field course guidebook is a must for geologists who wish to become aware of the latest thinking on the evolution of barrier-island deposits. Contains an excellent summary on transitional-coastal depositional environments by Hayes, an extensive bibliography on barrier-island and related environments, and some other useful papers (cited in text) on South Carolina barrier-island environments.

Kraft, J.C., 1971, Sedimentary facies patterns and geologic history of a Holocene marine transgression: Geol. Soc Amer Bull., v. 82, p. 2131-2158.

Discussion and documentation of the stratigraphy and facies distribution of a modern transgressive barrier-island system

Kumar, N., and J.E. Sanders, 1974, Inlet sequence: a vertical succession of sedimentary structures and textures created by the lateral migration of tidal inlets: Sedimentology, v. 21, p. 491-532.

An excellent study of an inlet sequence and development of an inlet-fill model

Schwartz, M.L., ed., 1973, Barrier Islands: Stroudsburg, Penn., Dowden, Hutchinson and Ross, 451 p.

A reprint collection of the classical historical and contemporary papers dealing with the origin and geomorphic development of barrier-island shorelines.

Swift, D.J.P., 1975, Barrier island genesis: Evidence from the Middle Atlantic Shelf of North America: Sediment. Geol., v. 14, p. 1-43

This paper discusses the origin and evolution of barrier islands within the context of the east-coast setting of the United States

**B. Ancient Barrier-Island Deposits**

Barwis, J.H., and J.H. Makurath, 1978, Recognition of ancient tidal inlet sequences: an example from the Upper Silurian Keyser Limestone in Virginia: Sedimentology, v. 25, p. 61-82.

Excellent documentation of an ancient tidal inlet-fill deposit. Also contains succinct reviews of modern tidal inlet studies and of published studies on ancient barrier deposits which appear to have tidal-inlet affinities.

Briggs, G., ed., 1974, Carboniferous of the Southeastern United States: Geol Soc. Amer. Spec. Paper 148.

Papers by D.K. Hobday, J.C. Ferm, J.C. Horne, Ferm and J.P. Swinchatt, and R.C. Milici deal with the interpretation of Carboniferous orthoquartzites and related deposits as being of barrier-island, tidal channel-delta, and back-barrier origin

Campbell, C.V., 1971, Depositional model - Upper Cretaceous Gallup beach shoreline, Ship Rock area, Northwestern New Mexico: Jour Sed. Petrology, v. 41, p. 395-409

Contains detailed descriptions of the geometry and sedimentary structures of Cretaceous barrier-beach facies

Carter, C.H., 1978, A regressive barrier and barrier-protected deposit: Depositional environment and geographic setting of the Late Tertiary Cohansey Sand: Jour Sed. Petrology, v. 48, p. 933-950

Includes ancient sequences interpreted as containing surf-zone deposits and tidal channel deposits.

Davies, D.K., F.G. Ethridge, and R.R. Berg, 1971, Recognition of barrier environments: Amer. Assoc. Petrol. Geol Bull., v. 55, p. 550-565.

Comparative study of the Galveston barrier model with Cretaceous and Jurassic sandstone deposits. Illustrates the use of the modern analogue in interpretation of ancient barrier-island deposits, and presents a model for barrier-island sedimentation. An excellent paper, but we must remember that the model presented is only one type of barrier island, the Galveston type. There are others.

Fisher, W.L., and J.H. McGowen, 1969, Depositional systems in Wilcox Group (Eocene) of Texas and their relation to occurrence of oil and gas: Amer. Assoc. Petrol. Geol. Bull., v. 53, p. 30-54.

Stratigraphic study using outcrop and subsurface data which proposes a regional depositional model consisting of deltaic environments flanked by barrier-island, lagoonal, and strandplain sediments.

Hobday, D.K. and J.C. Horne, 1977, Tidally influenced barrier-island and estuarine sedimentation in the Upper Carboniferous of southern West Virginia: Sedimentary Geol., v. 18, p. 97-122

Interpretation of Upper Carboniferous sandstone units as barrier-island deposits, with the recognition of tidal-delta, tidal-channel and washover facies.

Horne, J.C. and J.C. Ferm, 1978, Carboniferous Depositional Environments: Eastern Kentucky and Southern West Virginia: Dept of Geol., Univ. South Carolina, 151 p.

This excellent field guidebook summarizes the studies of Horne, Ferm and co-workers on the regional stratigraphy and paleoenvironmental interpretation of Carboniferous deltaic and barrier-island deposits of southeastern United States. Contains numerous outcrop illustrations and environmental reconstructions of various barrier-island facies and synthesizes these into a composite depositional framework

Some of this material has recently been published in the following Paper

Horne, J.C., J.C. Ferm, F.T. Caruccio, and B.P. Baganz, 1978, Depositional models in coal exploration and mine planning in Appalachian region: Amer. Assoc. Petrol. Geol Bull., v. 62, p. 2379-2411.

LeBlanc, R.J., 1972, Geometry of sandstone bodies: *in* T.D. Cook, ed., Underground Waste Management and Environmental Implications: Amer. Assoc. Petrol Geol., Mem 18, p. 133-189

One of the best of many articles dealing with the classification of clastic depositional environments, with a good introductory review of the barrier-island system

Land, C.B. Jr., 1972, Stratigraphy of Fox Hills Sandstone and associated formations, Rock Springs uplift and Wamsutter arch area, Sweetwater County, Wyoming: a shoreline-estuary sandstone model for the Late Cretaceous: Quart. Colorado School of Mines, v. 67, no. 2, 69 p.

Well-documented sedimentological and stratigraphic study of Upper Cretaceous tidal-channel, barrier-beach, and lagoonal deposits

**Secondary Reference List**

**A. Holocene Barrier-Island Deposits**

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**B. Ancient Barrier-Island Deposits**


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