A REEXAMINATION OF FACIES MODELS FOR CLASTIC SHORELINES

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Abstract: Currently accepted depositional models of coastal facies derive mostly from studies of modern depositional systems combined with interpretations of ancient successions. Two factors, however, can limit the efficacy of such facies models. First, Holocene coastal settings differ significantly from those in which many, if not most, ancient coastal sediments accumulated. Second, input to a model from the rock record commonly is specific to a particular tectonic and oceanographic setting (which may be poorly constrained) and is not fully exportable to other settings. This paper explores how these factors impact our interpretive capability.

Many, if not most, ancient shoreface deposits accumulated under conditions of progradation, a process relatively uncommon among the world's present shorelines. Instead, many modern postglacial coasts experience rising sea level and reduced sedimentation, which enhances barrier-island development and influences the shoreface-to-shelf bottom profile and sand–mud distributions. Ignoring these differences promotes inaccuracy in our facies models.

Often overlooked in the application of coastal facies models are variations imposed by texture, energy level, and tectonism. Sedimentary structures on fine-grained sandy coasts differ substantially from those on a coast underlain by coarse-grained sediment. Deposits on a high-energy coast are unlike those in a low-energy setting. Differing degrees of accommodation influence the nature of the preserved succession. Some of the problems inherent in current facies models can be obviated by considering them as end members within a continuum of models that incorporate different energy regimes, textural characters, and preservational modes.

INTRODUCTION

The basic model for open-coast clastic facies is not a complex three-dimensional model, as with deep-sea fan facies or deltaic facies, but rather a simple shallowing-upward facies succession that is perpetuated in a sheet sand to its landward and seaward pinchouts. The model follows the premise of Walther's Law, whereby the vertical ordering of facies reflects the lateral arrangement of facies in a conformably prograding system (Middleton, 1973). In such a case, beach and shoreface (used here as the relatively steep concave-up surface that lies between a beach foreshore and a shelf or basin platform) deposits prograde over adjacent shelf sediment. The resulting stratigraphic succession, where complete, is an upward progression of shelf–shoreface–foreshore–nonmarine facies (Fig. 1). Minor variations on the central theme exist in the various published iterations of the model (Figs. 2, 3, 4). All share a common motif of upward progression from bioturbated muddy sediment of the inner shelf to mixed mud and storm sand transitional to the sandy shoreface, which is dominated by storm structures in the lower part and by cross-bedded sand in the upper part. At the top lie flat-bedded beach foreshore deposits that are overlain by backshore or other nonmarine facies. More complex models have been proposed for open-coast accumulations in settings where base-level fluctuations impose patterns different from the laterally continuous systematic facies progression. These include the "forced regression" models of Plint (1988, 1991) and Posamentier et al. (1992) (Fig. 5) and the "transgressive incised shoreface" of Walker and Plint (1992) and Bergman and Walker (1999) (Fig. 6). In these models, stratigraphic associations differ from the basic model and the upward facies progression may be incomplete.

Although the basic concept of shallowing-upward facies succession has been widely applied to ancient open-coast accumulations, the specific models that exist do not cover all situations. Some include assumptions that are demonstrably erroneous. Much of the problem relates to limitations in the scope of studies on which the models are based. This paper reviews the origins and applications of the models and the impact of these problems. It concludes by suggesting an alternative approach that would allow a broader applicability of the models.

A DURABLE MODEL

The shallowing-up coastal classic facies succession in one or another of its various manifestations (Figs. 1, 2, 3) is familiar to virtually all students of shallow marine sandstones. The model has persisted with little modification for nearly 30 years, as depositional facies models for other systems, such as those for deep-sea fans, evolved dramatically (and some continue to evolve, as demonstrated in this collection of papers). The long-term effectiveness of this model derives in large part from a combination of simplicity and consistency in process, geometry, and preservation on wave-dominated coasts.

Although a complex array of processes influence coastal settings, the effects of shoaling waves are primarily responsible for shaping the nature of the clastic facies on open coasts, although tides and biogenic processes may be locally significant. Waves may exist as "seas", driven by local winds, or as "swell", generated by distant storms. Swell tends to have longer period and to influence the seabed to greater depths than do local sea waves. "High-energy" coasts are likely to be dominated by swell. "Low-energy" coasts receive smaller everyday waves but can experience very large waves during storms.

Waves move sediment by two mechanisms: by inducing water motion as they pass, and, at the shoreline, by generating sustained flow in the form of shore-parallel longshore currents and seaward-directed rip currents. The water motion induced by passing waves takes a consistent and predictable pattern that relates to shoaling changes in the wave form itself. Waves passing into shallow water change from a rounded, nearly sinusoidal form to one of sharp-crested peaks separated by broad, flat troughs. As the waves approach the beach they become increasingly asymmetric and ultimately break when they encounter water depths slightly (1.2–1.4 times) deeper than the wave height.

Facies Models Revisited

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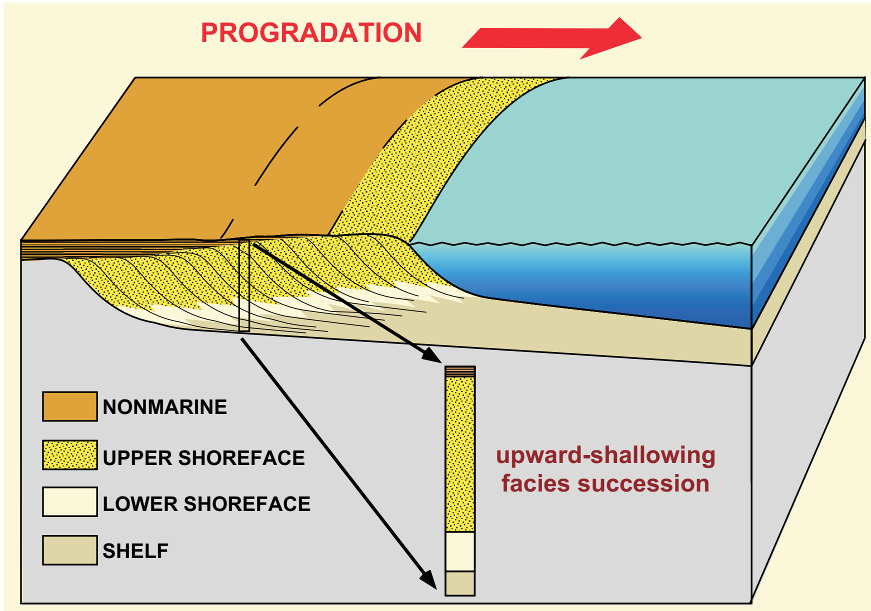

FIG. 1.—Characteristic mode of accumulation of shoreline deposits through progradation (shown here with a slight rise of relative sea level during progradation), whereby shallow-water facies build laterally over deeper-water counterparts, generating an upward-shallowing succession or, if sandwiched between transgressive episodes, a parasequence (see Harms et al., 1982).

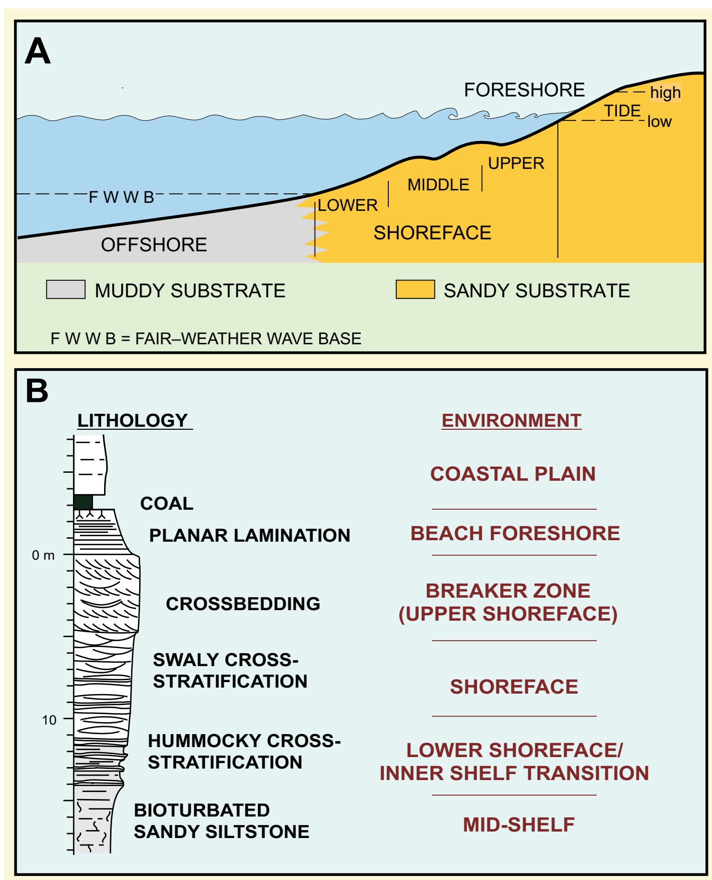

FIG. 2.—**A)** Beach-to-offshore profile in facies model of Walker and Plint (1992). Fair-weather wave base at base of shoreface. **B)** Shallowing-up facies succession in facies model of Walker and Plint (1992).

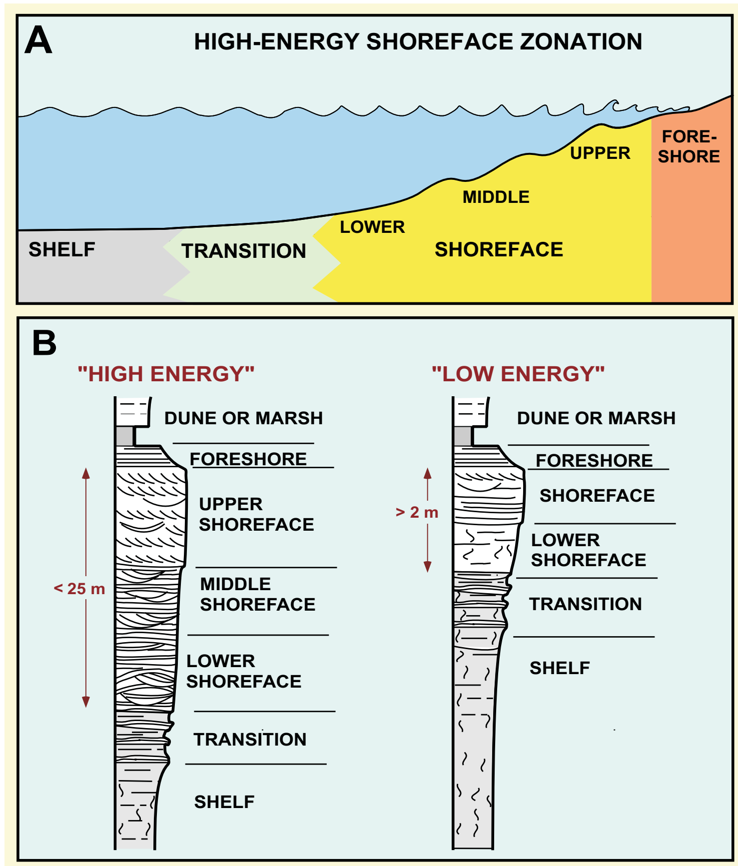

FIG. 3.—**A)** Beach-to-offshore profile (high-energy) in facies model of Galloway and Hobday (1996). Features a "transition zone" between shoreface and shelf. **B)** Shallowing-up facies succession in facies model of Galloway and Hobday (1996). Features a "transition zone" between shoreface and shelf in both high- and low-energy sequences.

The waves then pass through a breaker zone and end as swash and backwash on the beach foreshore.

Water moved by the passing of a wave follows a circular orbit, forward under the crest of the wave and backward under the trough, and the diameter of the circle diminishes with depth until the movement becomes insignificant ("wave base"). As a wave enters shallow water, the circle is deformed into an ellipse and the water movement just above the seabed is essentially horizontal (Fig. 7). As long as the wave form is nearly sinusoidal, the forward and backward movement of the water is symmetrical in both duration and velocity. But as the wave crests become peaked, the velocity profile of the orbital currents changes: the forward motion under the wave crests becomes stronger and of shorter duration than the seaward motion (Fig. 8).

The velocity asymmetry of oscillatory (orbital) motion has several geological ramifications. First, because bedload transport is on the order of the cube of the velocity, the asymmetry has the capability to drive sediment on the bed in a landward direction. Second, bedforms created by this flow face in a landward direction. Third, the stronger landward flow can overcome thresholds for movement of clasts that cannot be moved by the weaker seaward flow, and, where the grain-size range is sufficient, results in a preferential landward movement of the coarser clasts and textural sorting within the nearshore zone (Fig. 9).

All of this movement involves little or no mass transport of water. But as waves enter the breaker zone, water is carried forward by the waves. The result is a combination of setup and setdown, whereby the sea surface is elevated adjacent to the beach and depressed just seaward of the breaker line (Fig. 10). These changes in elevation create a hydraulic head that serves as the driving mechanism for longshore and rip currents, unidirectional flows that constitute nearshore circulation cells (Fig. 11). These simple processes encompass the significant forces that drive sedimentation in the nearshore, and they have prevailed since waves first came ashore early in the earth's history.

Wave-dominated shorelines typically have a simple geometry. In plan view, the shorelines tend to be two-dimensional and uncomplicated. In profile, shoaling waves create a relatively steep, concave-up shoreface (Fig. 12) that extends seaward from the beach foreshore and merges offshore with a much flatter shelf or basin platform (Johnson, 1919). On prograding shorelines, the shoreface is an equilibrium surface, probably reflecting a balance of seaward sediment transport during storms and the landward transport by shoaling waves between storms (Niedoroda et al., 1984). The sediment typically coarsens toward the upper part of the shoreface, the most energetic part of the system, where waves and longshore and rip currents typically shape the bed into dunes. The shoreface geometry may be complicated by the presence of breaker bars, but the overall system is one of general

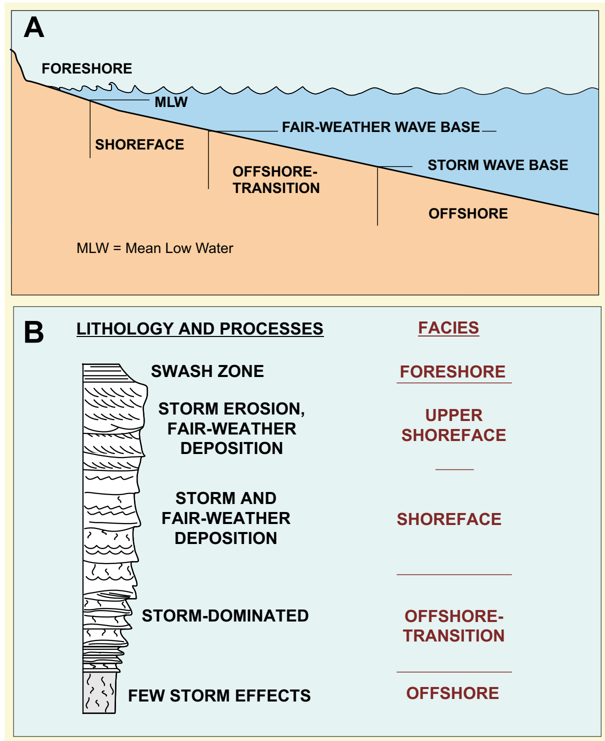

FIG. 4.—**A)** Beach-to-offshore profile in facies model of Reading and Collinson (1996). Fair-weather wave base defines base of shoreface. **B)** Shallowing-up facies succession in facies model of Reading and Collinson (1996).

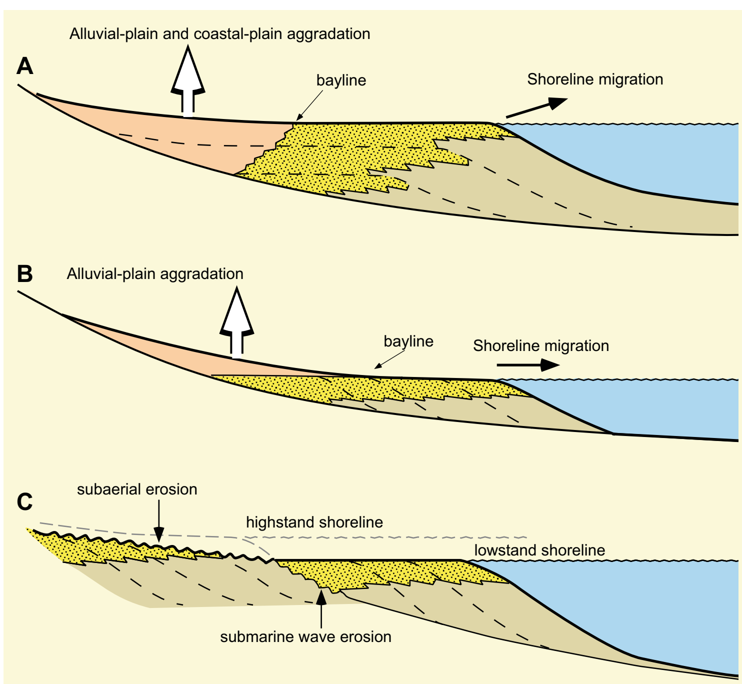

FIG. 5.—"Normal" and "forced" regressions. **A)** "Normal" regression under conditions of rising sea level. Shoreline deposits build seaward and upward. **B)** "Normal" regression under stable sea level. Shoreline deposits build laterally. **C)** "Forced" regression under conditions of falling sea level. Shoreline deposits become detached from their former position and can translate a substantial distance into the basin. Subaerial erosion occurs landward of the new shoreline deposits, which, because of wave erosion, display a sharp, erosional base over some portion of the previously deposited shelf facies. Modified from Posamentier et al. (1992).

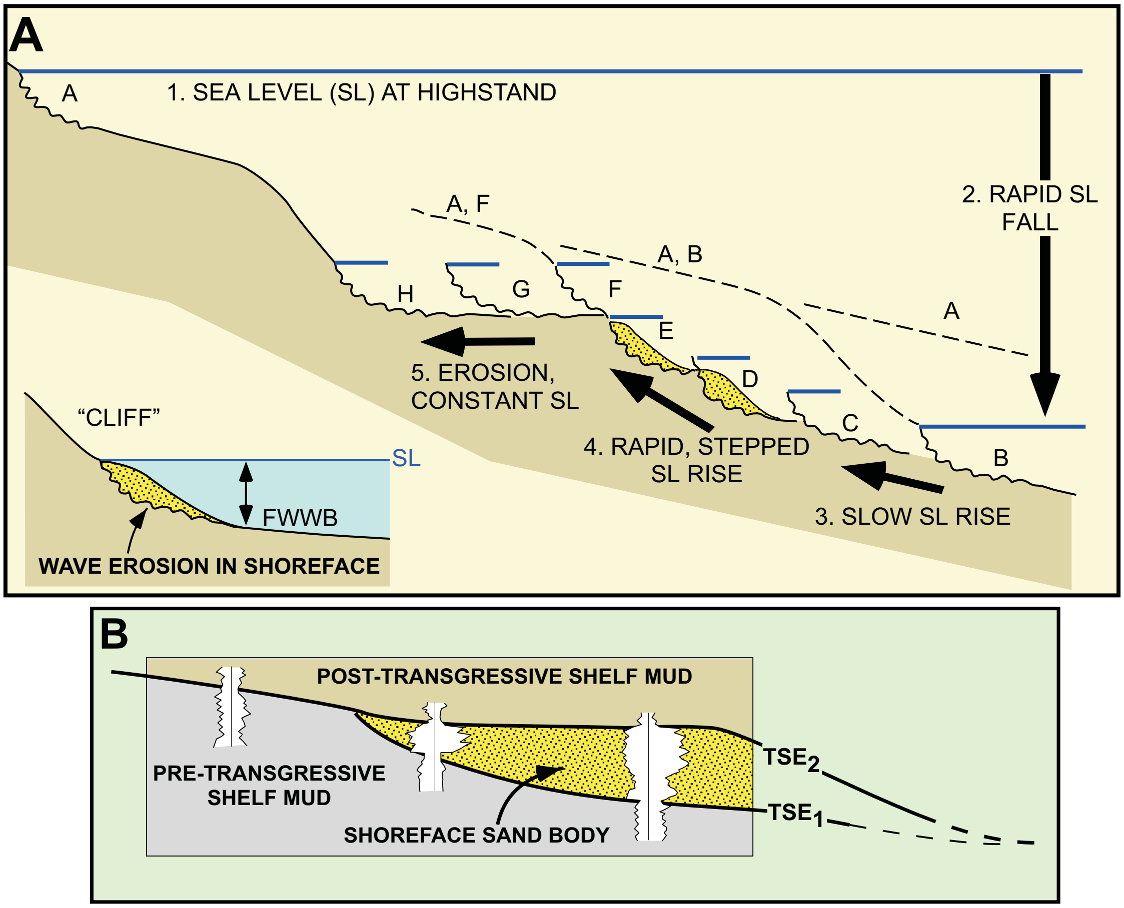

FIG. 6.—**A)** "Stepped transgressions" following a major fall in sea level (2). Rapid rise in sea level (4) can preserve shoreface deposits (E, D, formed during pauses in the transgression), which are "trapped" against a sea cliff, where they will be encased in offshore mud as transgression progresses. Erosion surfaces are marked by transgressive lags. Modified from Walker and Plint (1992). **B)** Template and log signatures for a shoreface deposit preserved after a stepped transgression (TSE1 = initial transgressive surface of erosion which cuts a "cliff", TSE2 = resumed transgressive surface of marine erosion). Shoreface sand body accumulates during brief stillstand of sea level between the two phases of transgression. After Bergman and Walker (1999).

simplicity (Fig. 13). Finally, because wave processes tend to straighten progradational shorelines, the process of progradation tends to produce simple sheet sand deposits, with minimal geometric complication.

There is also simplicity and consistency in preservation. Shoreline successions typically develop where the coast progrades through the addition and accumulation of sediment (Fig. 1). The result is the basic depositional facies model, reflecting the shallowing-upward succession of facies (typically coarsening-upward as well). Accumulations of coastal deposits are likely to be preserved only within a subsiding basin. Fluctuations in eustasy, sediment supply, or rates of subsidence typically generate alternations of marine regression (progradation) and transgression. During marine transgression, the landward migration of the shoreface equilibrium profile (Fig. 14) removes much of the previously deposited material and produces a surface of ravinement (transgressive surface of erosion). As a consequence, stacked sets of shoreline deposits tend to consist of a stacked set of progradational parasequences separated by erosional surfaces formed during intervening transgressions (Fig. 15). Erosional surfaces formed at the base of tidal inlet deposits during transgression may complicate the pattern (Clifton, 2003).

ORIGIN AND APPLICATION OF THE MODEL(S)

Among the earliest depositional facies models proposed for shallow marine sandbodies was that of the barrier island. Barriers are particularly prominent on the U.S. Gulf Coast and East Coast and on the southeastern coast of the North Sea, all sites of early coastal sedimentologic studies (e.g., Shepard and Moore, 1955; Van Straaten, 1959; Shepard et al., 1960; Bernard, et al., 1963; Reineck, 1963; Hoyt et al., 1964). Simultaneously, geologists began to recognize that depositional facies could be very useful in the exploration for petroleum hydrocarbons. As a result, "linear clastic shorelines" became an early analog for open-coast deposits, and barriers became almost synonymous with shallow-marine sand deposits. It was noted that barriers contained an upward-coarsening lithologic succession (Weimer, 1961; Bernard et al., 1963; Shelton, 1965, 1967; Berg and Davies, 1968), and for some time it was virtually assumed that all shallowing-up shallow-marine sandstones originated in a barrier. Selley (1969), in his survey of depositional environments, focuses almost exclusively on barriers, referring to nearshore deposits (even sheet sands) as "barrier beach" facies. In SEPM Special Publication 16, Recognition of Ancient Sedimentary Environments (Rigby and Hamblin, 1972),

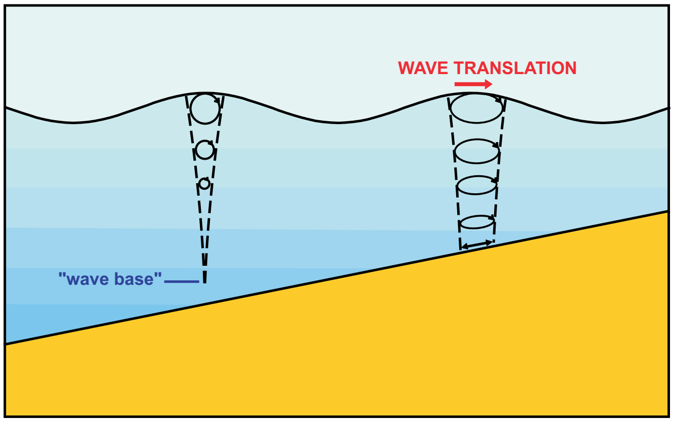

FIG. 7.—Water particles moved by passing waves in deep water follow a circular orbit that diminishes to zero at depth. In shallower water, the particles follow an elliptical orbit that flattens downward into a simple back-and-forth motion at the sea floor, forward under the wave crest and backward under the trough. The landward-most set of motions shown here is exaggerated for convenience.

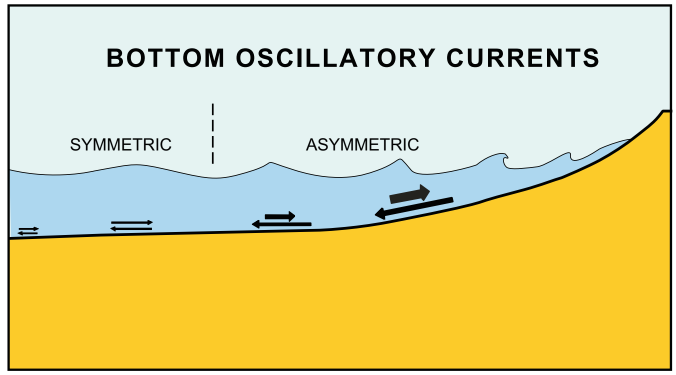

FIG. 8.—Schematic representation of relative velocity (arrow thickness) and duration (arrow length) of oscillatory flow at the bottom beneath shoaling waves. Flow is symmetric under sinusoidal waves and asymmetric under sharp-crested waves. Flow is landward under wave crests and seaward under wave troughs. Stronger landward velocities (1) generate landward bed-load transport, (2) create landward-directed cross-bedding and ripple lamination, and (3) overcome threshold velocities of clasts too large to be moved seaward, thereby contributing to selective landward transport of coarser clasts and overall textural sorting on the beach and shoreface.

a chapter is dedicated to criteria for recognizing ancient barrier coastline (Dickinsen et al., 1972), and scant attention is given to strand plains. As late as 1982, McCubbin, in a discussion of sandy coastal environments, discusses strand plains but devotes most of his text to barriers. In my experience in the petroleum industry in the 1990s, I found a surprising number of shallow-marine sandstones interpreted as linear barriers, even where the geometry did not support a barrier-island interpretation.

Early depositional facies studies of shallow-marine sandbodies encountered difficulties with the barrier interpretation. Harms et al. (1965), in a description of the Fox Hills Sandstone in Wyoming, could not accept an earlier interpretation of the unit as a barrier deposit (Weimer, 1961). They noted that although the sandstone showed the upward lithologic progression attributed to barrier-island deposits (upward coarsening, upward transition from marine to nonmarine environments), the lateral facies relations were inconsistent with a barrier interpretation. They could not, however, offer an alternative interpretation.

The description by Curray et al. (1969) of a prograding strandplain at Nayarit, Mexico, provided an alternative to the barrier-island model: a sheet sand with an internal shallowing-up facies succession. The facies models shown in Figures 1–3 are consistent with either prograding strand plains or barrier islands, where the primary differences lie in the geometries and facies

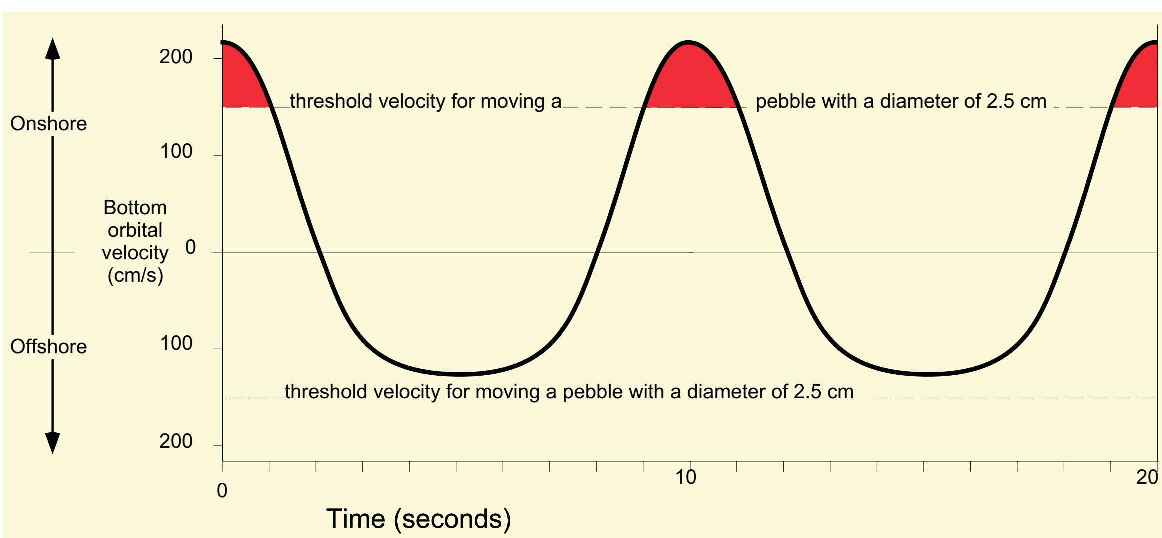

FIG. 9.—Selective shoreward transport of a 2.5 cm pebble under shoaling waves in which the landward oscillatory flow under the wave crest has a greater velocity than the seaward flow under the trough. Dashed lines indicate threshold velocity, areas in red indicate intervals of transport (adapted from Komar, 1976). Threshold velocity is derived from Komar and Miller (1973).

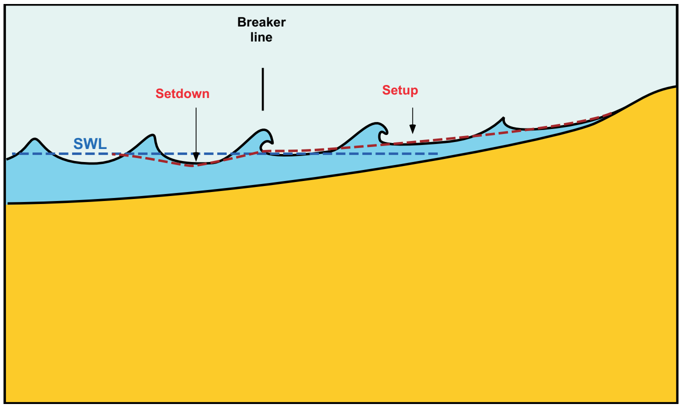

FIG. 10.—Setup and setdown of the sea surface (departure from the still-water line (SWL)) owing to the landward transport of water within the surf zone. Depression of the sea surface (setdown) just outside the breaker line and elevation of the sea surface adjacent to the beach provides the hydraulic head for nearshore circulation cells.

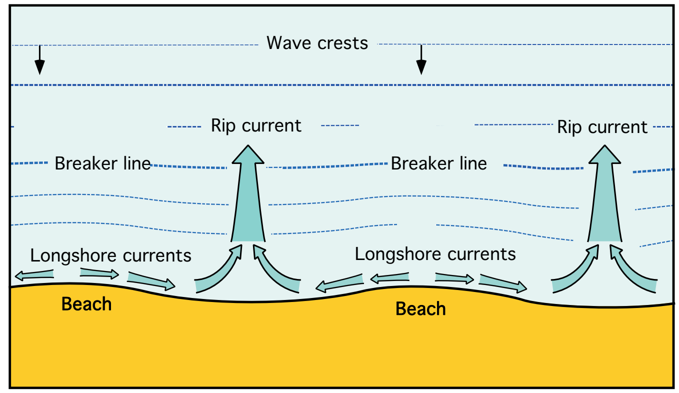

FIG. 11.—Nearshore circulation cells where wave incidence is parallel to coast. Circulation consists of unidirectional rip and longshore currents generated by the hydraulic head caused by setup and setdown of the sea surface in and near the breaker zone.

associations. Ryer (1977) noted the asymmetry of shoreface successions, in which the shallowing-up facies progression is typically capped by an erosional surface. He attributed the erosion and the accompanying abrupt return to facies deposited in deeper water to a landward shift in shoreface profile during transgression (Bruun, 1962).

The prograding strand-plain model, which may have first been proposed by Harms et al. (1975) and was expanded upon by Harms et al. (1982), has been applied successfully in many basins throughout the world. It predicts that prograding shoreline sandstones become progressively coarser and / or cleaner in an upsection direction, an attribute readily identifiable in well logs in areas where rock data are sparse or absent.

Despite the success of the prograding shoreline models, they did not explain several categories of deposits, such as sharp-based shoreface deposits resting abruptly on shelf mudstone, or isolated sandstone bodies encased in shelf mudstone or shale (Snedden and Bergman, 1999). Although these latter deposits commonly share similar characteristics with shoreline successions (upward coarsening and upward increase in depositional energy), they lack the sheet geometry and lateral facies relations normally found in a progradational unit.

The advent of sequence stratigraphy created a new awareness of the potential importance of sea-level change on facies associations. The progradational-shoreline model is premised on constant or slightly rising relative sea level (Fig. 1). New models were proposed involving a fall of relative sea level, or "forced regression" (Plint, 1988; Posamentier et al., 1992). In these models, a relatively rapid fall in sea level induces a seaward "jump" of the shoreline deposits to a new, topographically lower position (Figs. 5, 6, 16, 17). The models differ in their extent of wave erosion and

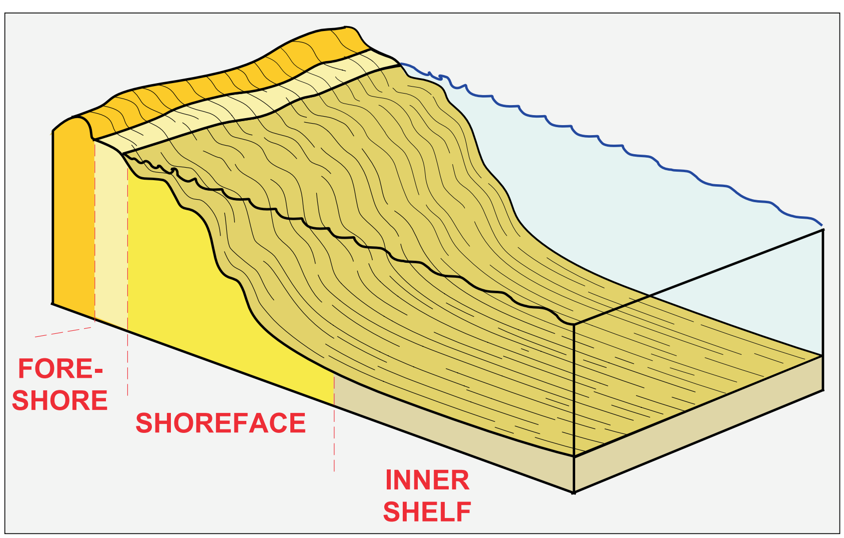

FIG. 12.—The shoreface is best defined as a morphologic feature that attends nearly all clastic shorelines. It has a relatively steep concave-up surface that extends seaward from the beach foreshore and merges with the much flatter shelf or basin platform. Breaker bars, as shown, may cover the upper part of the shoreface here.

FIG. 13.—Two-dimensional wave-dominated open coast, South Carolina, U.S.A. Two sets of breaker bars are clearly visible: a continuous inner bar, and an outer bar on which waves break more sporadically. Ridge-and-runnel systems occupy lower beach foreshore, particularly in the distance. Although a barrier presently occupies this coast, progradation would produce a shallowing-upward sheet sand.

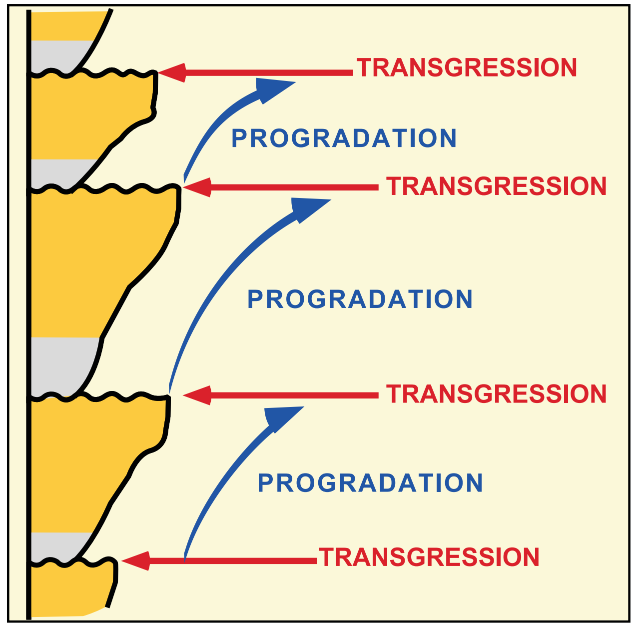

FIG. 15.—Typical pattern of stacked shoreline successions. Shallowing-upward progradational parasequences meters to tens of meters thick are separated by transgressive surfaces of erosion (ravinement surfaces). Falls in sea level prior to transgressions can produce erosional sequence boundaries and incised-valley-fill deposits at the top of the parasequences.

nature of subsequent transgression. Plint (1988) and Snedden and Bergman (1999) envision wave erosion by fair-weather waves across the shelf platform in advance of the new shoreline position, producing an extensive erosional base to the advancing shoreface deposits (Figs. 16, 17). Posamentier et al. (1992) envision wave erosion limited to the position of the newly established shoreline, and a gradational shoreface–shelf transition as progradation ensues (Fig. 5). Posamentier et al. (1992) and Snedden and Bergman (1999) invoke a steady transgression that isolates the lowstand deposits, whereas Bergman and Walker (1988) and Walker and Plint (1992) call on a stepped transgression, featuring sporadic stillstands of the sea. During the stillstands, shoreline deposits can accumulate against a wave-cut sea cliff (Fig. 6), thereby forming narrow, shore-parallel linear sandbodies encased in marine shale as transgression proceeds.

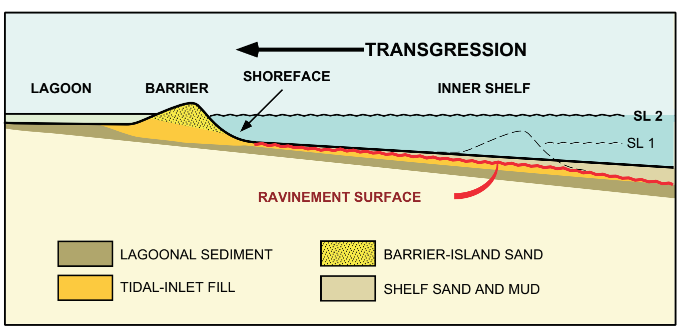

FIG. 14.—Process of marine transgression is commonly associated with the landward migration of a barrier island or barrier spit. Landward translation of the shoreface profile as sea level rises (here from SL 1 to SL 2) creates an erosional surface (surface of ravinement or transgressive surface of erosion) cut into previously deposited sediment. Because deeper-water sediments (here inner-shelf sand / mud) accumulate on the ravinement surface, it marks an abrupt upward change to deeper-water deposits.

The forced-regression models have been much applied since their inception (Posamentier and Chamberlain, 1993; Ainsworth and Crowley, 1994, to cite a few). The Bergman and Walker (1988) transgressive-incised-shoreface model for the Cardium has been used to interpret other less well-documented shorefaces (Pattison and Walker, 1992; Walker and Wiseman, 1995; Le Roux and Elgueta, 1997; Bergman, 1999; MacEachern et al. 1999), including, somewhat controversially, the Shannon Sandstone (Bergman and Walker, 1999). As will be discussed, however, not all sharp-based shoreface deposits are necessarily produced by forced regression.

Ichnofacies models, based on associations of trace fossils, have also been proposed for wave-dominated coastal successions. Early students of these deposits recognized that traces could be useful in the interpretation of shallow-marine sandstone (Weimer and Hoyt, 1964; Harms et al., 1965; Howard, 1966). The trace *Ophiomorpha* was thought initially to be indicative of a shallow-marine environment (Weimer and Hoyt, 1964; Harms et al., 1965), but subsequent studies found it in a variety of other marine environments, including those at bathyal depths (Kern and Warme, 1974). The trace *Macaronichnus* occurs in many beach foreshore facies (Saunders and Pemberton. 1986; MacEachern and Pemberton, 1992), but the trace is present in other paralic environments as well, including upper shoreface (MacEachern and Pemberton 1992; Male, 1992), lower shoreface (Clifton, 1981), and tidal flats and channels (Clifton and Thompson, 1978). A problem with trace fossils is that, for most, we have no knowledge of the physical or chemical factors that limit their distribution.

Several attempts have been made to associate ichnologic trends to lithologic trends in nearshore sediment. Howard (1966) identified patterns in the trace assemblages that corresponded to the

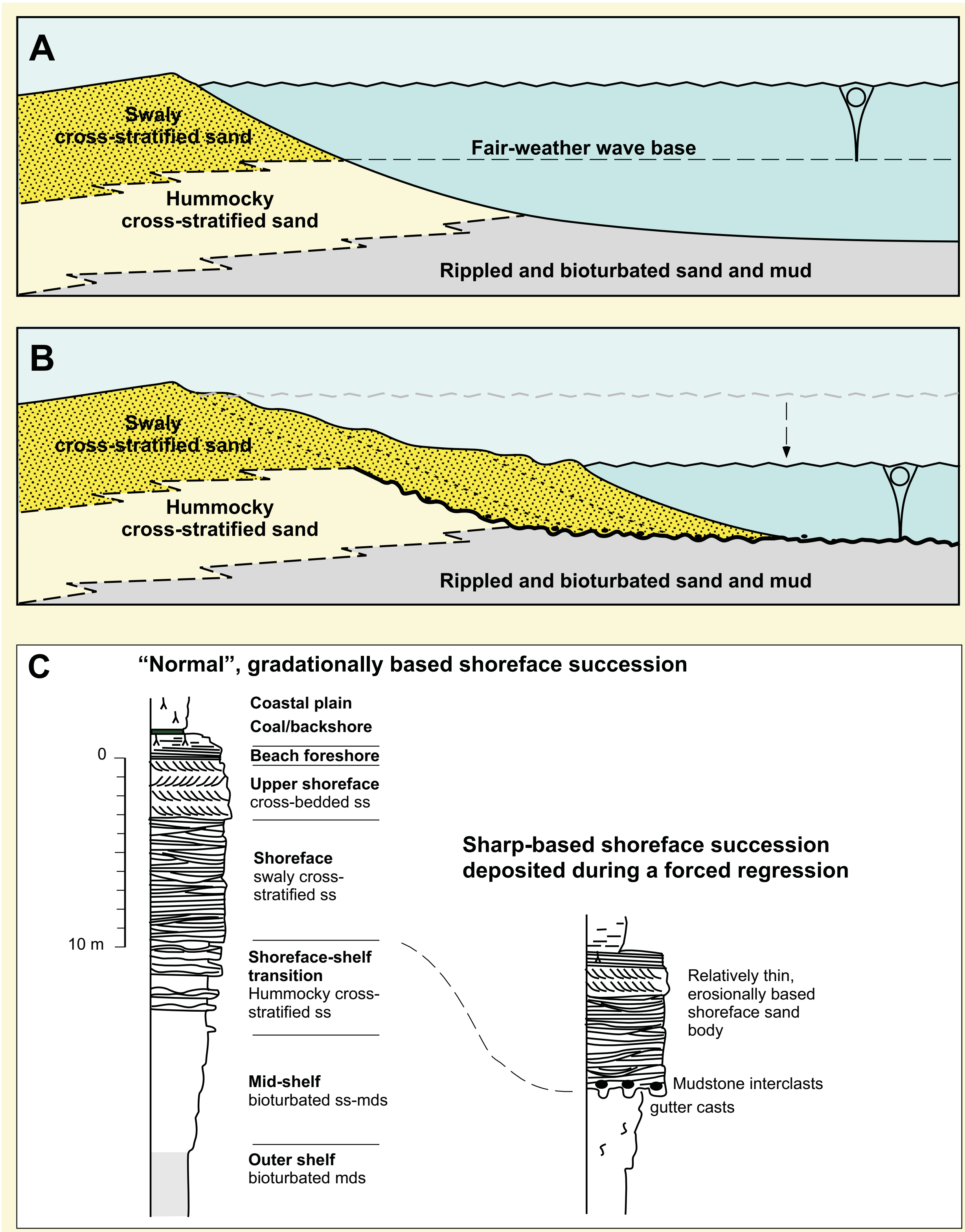

FIG. 16.—"Forced regression" as visualized by Plint (1988). **A)** Prograding shoreline in which swaly cross-stratification defines the lower shoreface, hummocky cross-stratification defines the shoreface–shelf transition, and rippled and bioturbated sand and mud typify the mid-shelf facies. **B)** Result of "forced regression", where falling sea level induces mid-shelf wave erosion and shoreface sands accumulate on the erosional surface. **C)** Vertical succession of a "forced regressive" deposit compared to a "normal" progression.

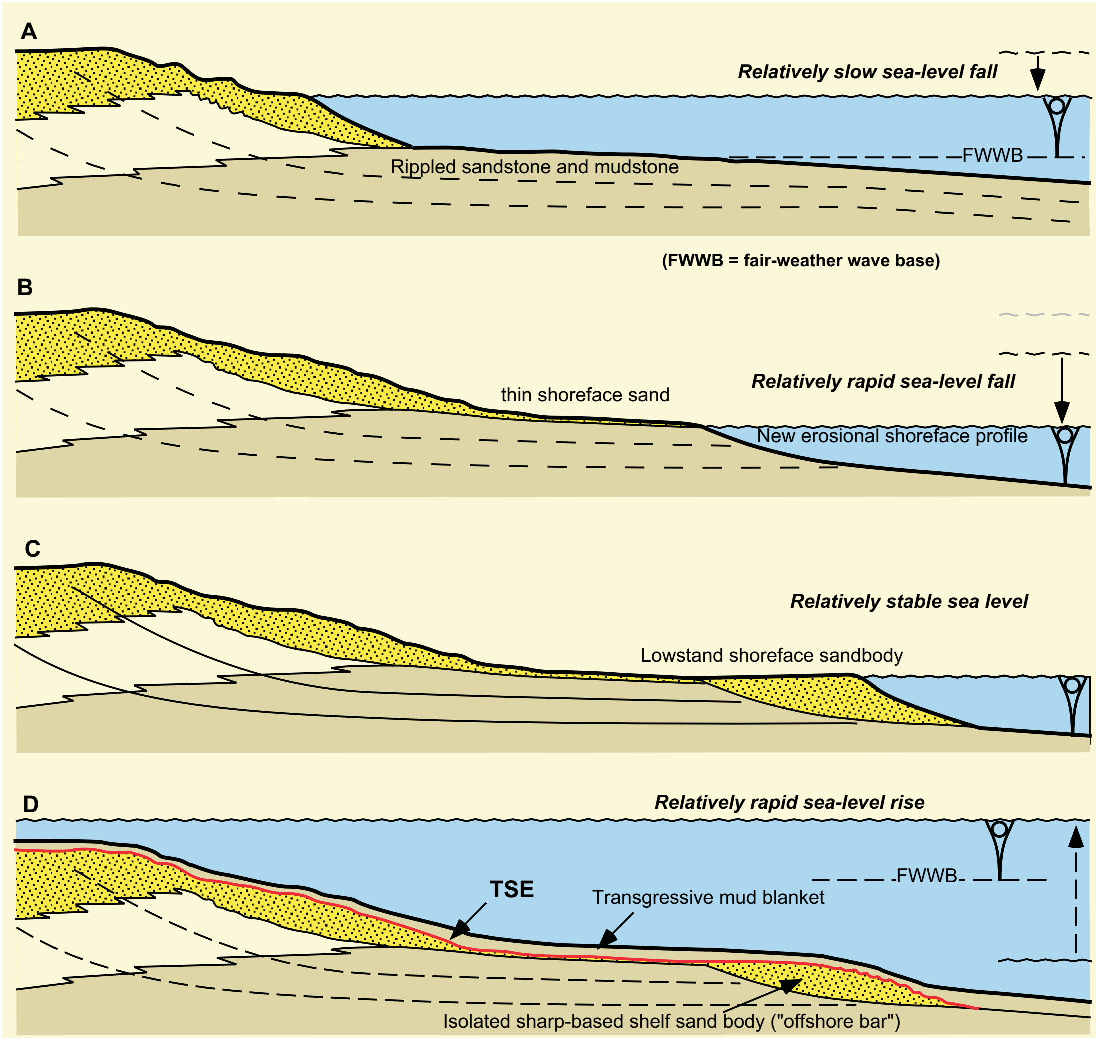

FIG. 17.—Model of formation of isolated shallow marine sandbodies by relative sea-level change. **A)** Relatively slow sea-level fall; shoreline moves seaward relatively slowly. Reduced accommodation results in a relatively thin shoreface sandbody. **B)** Faster relative sea-level fall; shoreline moves seaward at an increasing rate. Rate of sea-level fall exceeds subsidence, accommodation is reduced to nil, and the shelf becomes emergent. **C)** Relatively stable sea level; shoreline incises in seaward position. Erosion plus subsidence create space into which a new shoreface sandbody can prograde. **D)** Relatively rapid sea-level rise; shoreline shifts rapidly landward. Sandbody is isolated in shelf mud. After Snedden and Bergman (1999), following Plint's (1988) model (Fig. 16).

lithologic changes in shallowing-up Cretaceous sandstones in the Book Cliffs, Utah, U.S.A. (Fig. 18), and Howard et al. (1972) and Howard and Reineck (1981) describe the distribution of physical and biological structures in beach-to-offshore transects at Sapelo Island, Georgia, and Port Hueneme, California, U.S.A. More recently, Pemberton et al. (1992) followed up on Howard's work in the Book Cliffs with a more detailed analysis of the traces associated with the lithologic succession (Fig. 19), and in 1992 MacEachern and Pemberton (1992) proposed an ichnofacies model for Cretaceous shoreface successions in the western interior basin (Fig. 20).

The ichnofacies model is a useful adjunct to models based on texture and sedimentary structures, and may provide a basis for subtle environmental interpretations not possible on the basis of physical structures alone (MacEachern and Pemberton, 1992). Unlike models based on physical features, ichnofacies models are subject to biological evolutionary trends and may therefore be somewhat time-specific. Comparisons with modern analogs are also inherently difficult owing to the typically limited view of the sub-sea-floor section on modern coasts. The studies of modern biological structures by Howard et al. (1972) and Howard and Reineck (1981), for example, provide little data that bear on the model developed by MacEachern and Pemberton (1992).

THE HOLOCENE HERITAGE: ANALOGS FROM A NON-ANALOGOUS WORLD

At present, a glacio-eustatic highstand exists, following a rapid and large sea-level rise (that began about 17,000 years ago), a consequence of the melting of continental glaciers that developed during the Wisconsin glaciation. Sea level continues to rise

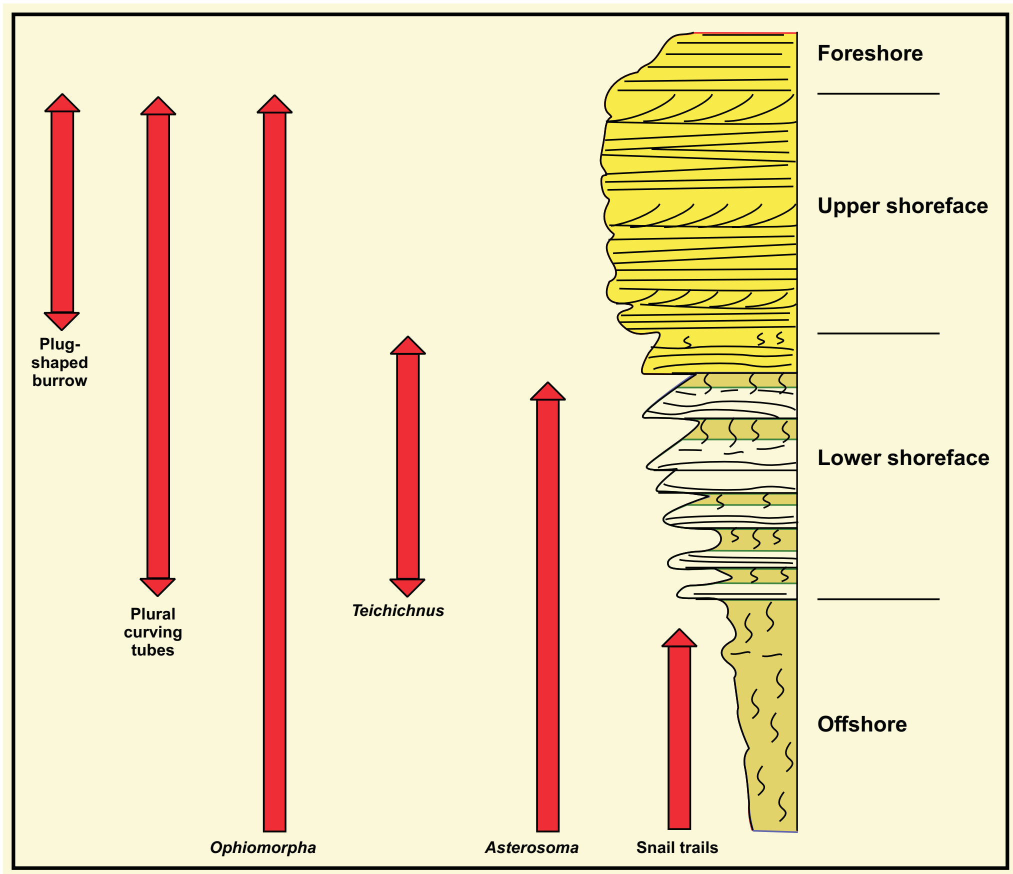

FIG. 18.—Distribution of traces in a shallowing-up succession in the Cretaceous Blackhawk Formation, Book Cliffs, Utah, U.S.A. Constructed from data provided by Howard (1966).

on many coasts (albeit at a greatly reduced rate), and much of the world's coastal area remains in a state of slow transgression. In contrast, the coastal successions to which the classic shoreface model applies are, by definition, progradational. The attempts to incorporate observations from a still somewhat transgressive world into a progradational model have created a number of problems for our models of open-coast clastic facies.

Barriers

Presently, about 15% of the world's coastline is fronted by sandy barrier islands or barrier spits (Glaeser, 1978), much of it on U.S. coasts, and it is unsurprising that the linear sandbodies represented therein provided an early, and widely used, exploration model. The origin of barrier islands has been much debated (see discussion in Davis, 1994), but most seem to form as a result of landward transport and upward accretion of sand (Davis, 1994). Many, if not most, modern barriers seem to be accumulations of sand that are migrating landward as part of a slow transgression (Kraft et al., 1973; Boyd, et al., 1992). A key factor in their development is probably a paucity of sand in the open-coast system. It is generally accepted that during a rise in relative sea level much sediment is trapped in rivers and estuaries and that the amount of sediment, particularly sand, delivered to the open coast in many settings is greatly reduced. Waves mobilize the available sand and concentrate it along the coast into a barrier, which exists because the sediment supply is insufficient to fill the basin landward of the barrier. Under this concept, the barrier continues to retreat to landward until a new progradational episode begins. As noted by Suter and Clifton (1999), the biggest pitfall to using modern analogs is preservation potential. The most common record of geologically preserved barrier islands may be as the landward-most part of a progradational sand sheet (Fig. 21).

Sand–Mud Distribution

The fact that many of the present-day coasts, particularly those with barriers, are sand-poor has also directed our thinking about the distribution of sand and mud in ancient coastal systems. Where sand is more or less confined to a barrier's shoreface, a transition from sand to mud is likely to coincide with the base of the shoreface. Because so many of the early studies of coastal facies focused on barrier islands, it became generally accepted that a transition from sand to mud defines the base of the shoreface. This concept persists today in our models of prograding coastal deposits. In nearly all of them, the base of the shoreface coincides with a downward textural transition from sandstone to shale (Figs. 2, 3, 4).

Under conditions of progradation, sand is likely to be far more abundant and can extend far out onto the shelf. North of the mouth of the Columbia River, for example, the transition from

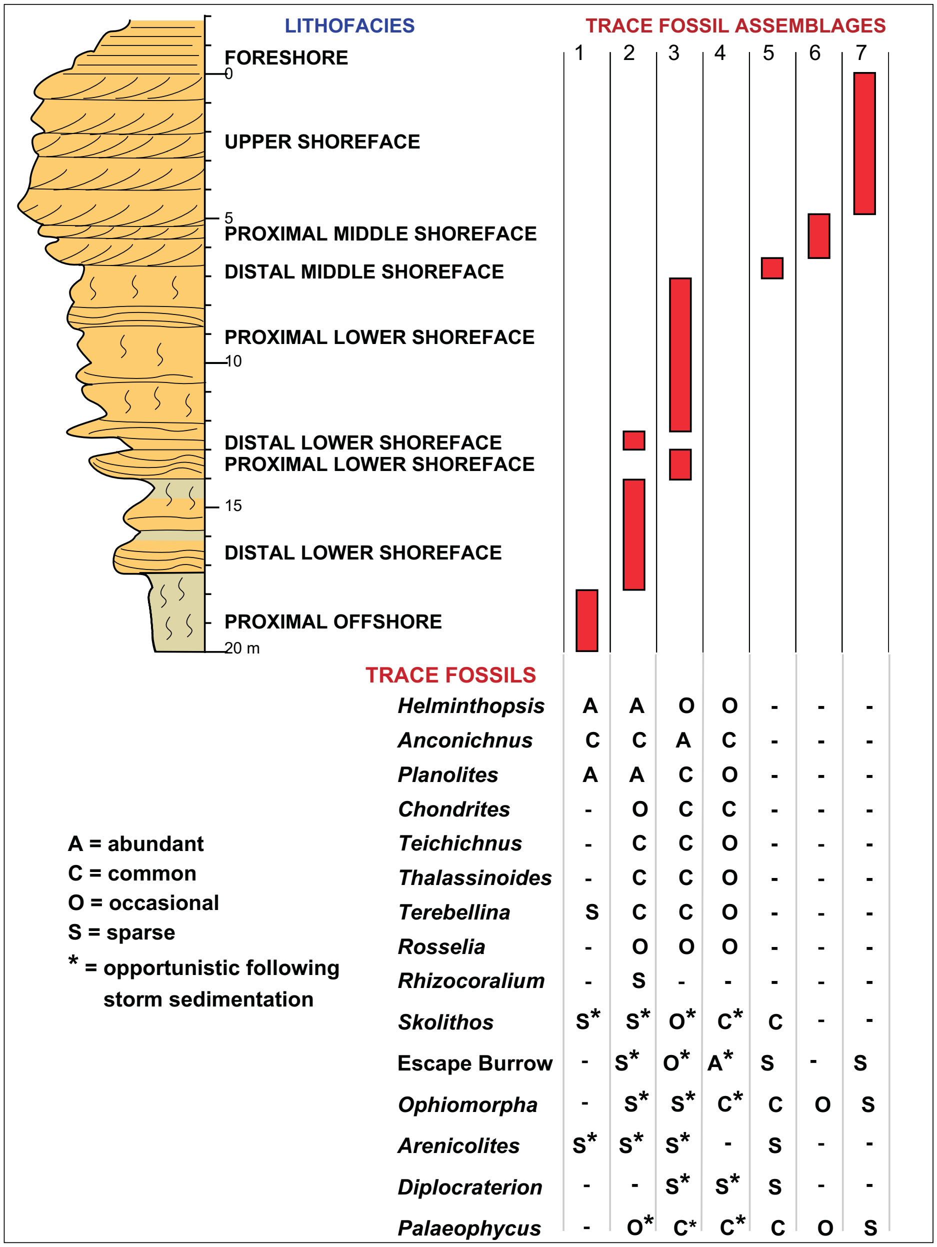

FIG. 19.—Trace-fossil distributions in core from the Spring Canyon Member of the Blackhawk Fm., Price, Utah, U.S.A. From Pemberton et al. (1992).

sand to mud lies in a water depth of about 40 m near the river mouth (Fig. 22). The transition extends to the north across the shelf in progressively deeper water, ultimately reaching about 90 m a few tens of kilometers north of the river mouth. Off north-central California, the sand–mud transition lies consistently at a water depth of about 60 m. (Fig. 23). Even off some barrier coasts, such as that of New Jersey, sand extends offshore well onto the shelf (Fig. 24). Such sands may not be coeval with those on the barrier but could be indistinguishable from them in the rock record. Depositional facies models that equate the sandstone–shale transition with the base of the shoreface may be valid for some deposits, but they do not provide an encompassing generalization.

Shoreface Profile

The relief of shoreface profiles in our present post-transgressive world is variable (Fig. 25), depending on energy level and the seafloor configuration prior to the transgression. The range in water depths at the base of modern shorefaces led Galloway and Hobday (1996) to conclude that that the thickness of shoreface facies successions spans some 2 to 25 meters (Fig. 3). This variation, however, occurs on erosional coasts. Prograding coasts, which provide the analog for nearly all ancient shoreline succession, show less variability. On the prograding high-energy coast of southern Washington state, U.S.A., the break in slope that defines the shoreface–shelf transition occurs at a water depth of about 10

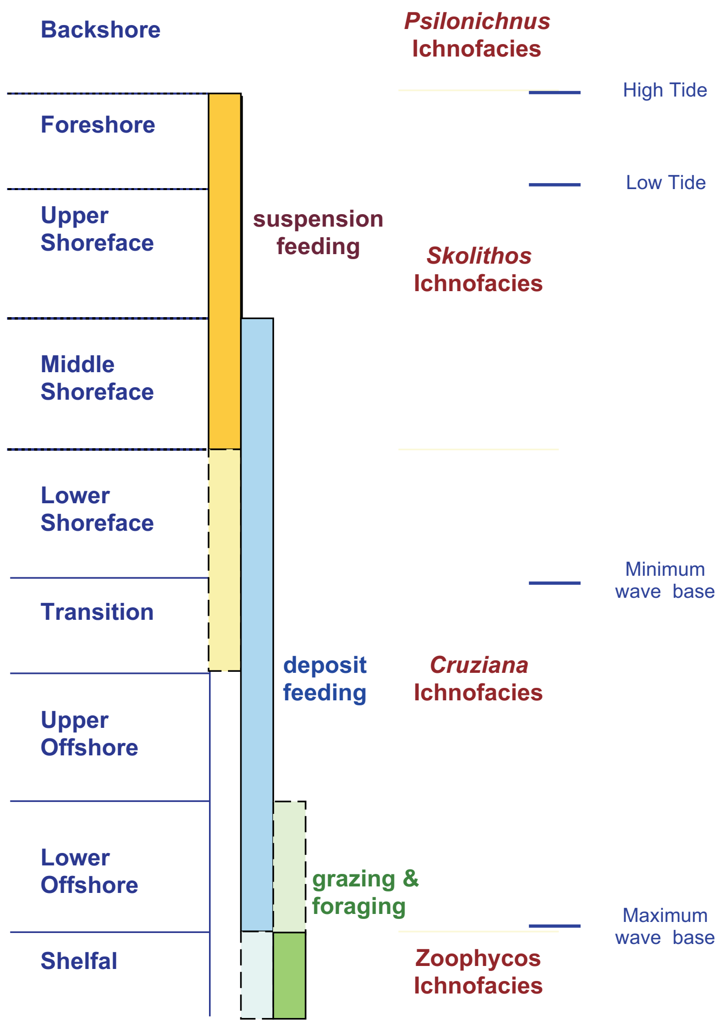

FIG. 20.—Ichnofacies model for the shoreface, after Pemberton et al. (1992).

m (Fig. 26). The progradational coast of Nayarit, Mexico, a region of somewhat lower wave energy, has the base of the shoreface at around 6–7 m (Fig. 27). The base of the shoreface on Galveston Island, a prograding part of the Gulf of Mexico also with relatively low wave energy (Morton, 1994), lies at a water depth of about 6 m (Fig. 28). In contrast, the base of the shoreface off Padre Island, an eroding part of the Gulf coastline (Morton, 1994), lies in water depths that approach 20 m (Fig. 25). On the basis of these observations it seems likely that, barring unusual rates of accommodation, the thickness of individual shoreface succession in the stratigraphic record is not likely to exceed 10–12 m.

Many modern shorelines, particularly those on the U.S. Atlantic coast, are fringed by shoreface-attached ridges, linear bodies of sand, or sand ridges, that rise above the adjacent sea floor (Snedden et al., 1984; Hoogendoorn and Dalrymple, 1986; Antia et al., 1994; van de Meene, et al., 1996; Dalrymple and Hoogendoorn, 1997). The ridges can be tens of kilometers long, 0.7 to 8 km wide, and 5 to 40 m high, and they are composed of fine to coarse sand. The ridges typically lie oblique to the shoreline and tend to be asymmetric, with side slopes ranging from < 1° to a maximum of 7°. Sediment on the stoss sides of ridges is generally coarser than that on their lee sides. Most shoreface-attached ridges consist of an upward-coarsening accumulation of storm-event beds (Snedden et al., 1994; Hoogendoorn and Dalrymple, 1986; Rine et al., 1991; Dalrymple and Hoogendoorn, 1997), although some show evidence of both storm and tidal influences in their internal structures (Antia et al., 1994; van de Meene et al., 1996). Coring and/or high-resolution seismic profiling show that many contemporary ridges are compound features, composed of an upper part shaped by modern processes and a core derived in an earlier setting (Rine et al., 1991; Snedden et al., 1994).

Shoreface-attached ridges appear to be phenomena associated with a retreating shoreline (McBride and Moslow, 1991). As the shoreline shifts landward, some of the ridges are left behind as isolated features on the inner shelf (Fig. 29), where they may be further modified by shelf processes (Swift et al., 1986). Although shoreface-attached ridges can be imposing coastal features, their association with transgressive coasts minimizes any importance as part of a progadational shoreline model. Any preservation is most likely as isolated linear shelf sand bodies within a shelf succession (Swift and Parsons, 1999).

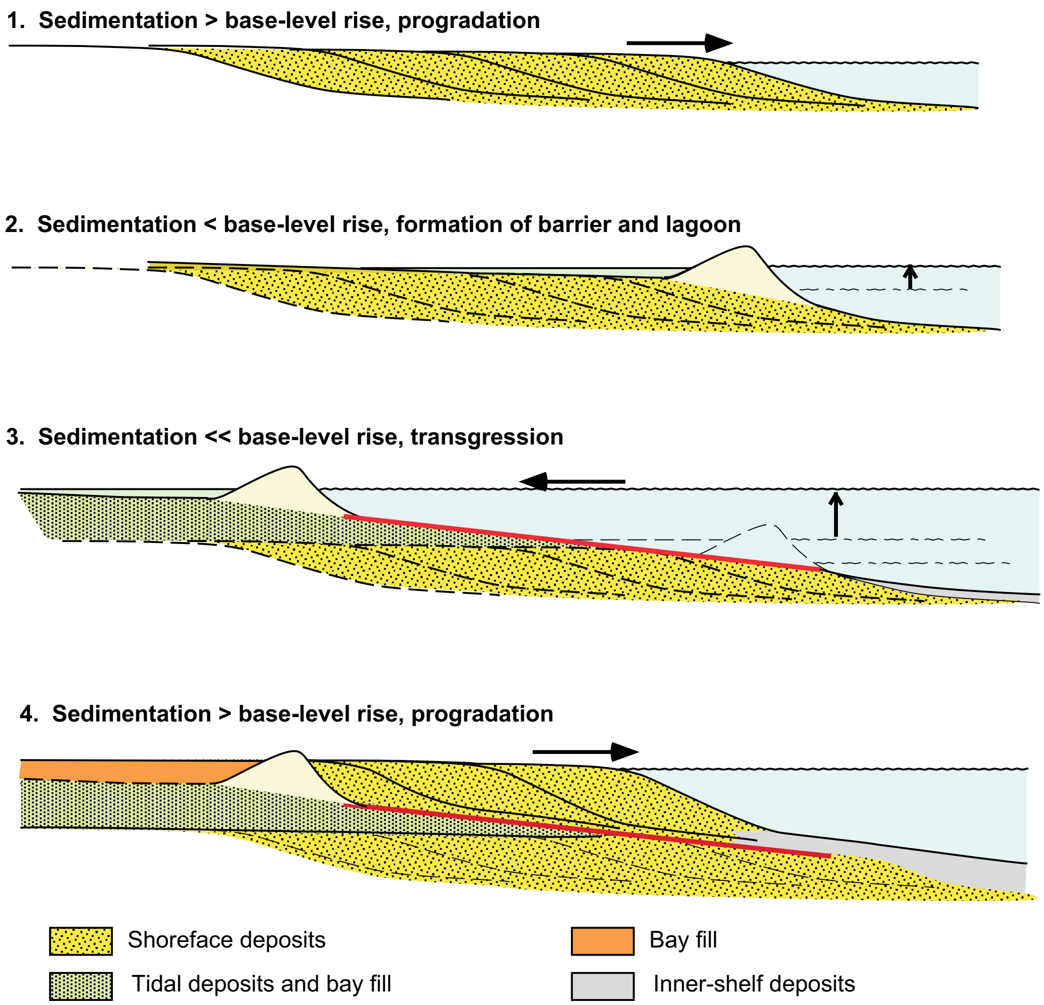

FIG. 21. Schematic diagram showing the development, landward migration, and stranding of a barrier in response to changes in the balance of sedimentation and relative sea-level fluctuation. During progradation in step 4, barrier maintains its identity until the embayment created by the barrier is filled, at which time the coast converts to a strand plain.

TECTONIC SETTING AND GRAIN SIZE

Most of the early studies of modern open-coast systems were conducted on tectonically passive margins of the U. S. Atlantic and Gulf coasts and the German and Dutch coasts of the North Sea, where rivers with low gradients cross broad coastal plains and deliver fine sand to the shoreline. As a result, the emerging models were premised on a nearshore system composed of uniformly fine-grained sand. These models were corroborated by studies in the Book Cliffs and elsewhere made on rocks of similar texture. The generalizations drawn in these studies, however, fail in varying degrees when applied to coarser-grained open-coast deposits, particularly those in tectonically active settings.

Sedimentologists have tended to consider grain-size distributions mostly to be reflective of processes of transport and deposition, hence the numerous, largely unsuccessful, attempts to reconstruct ancient depositional environments from textural parameters. While it is true that processes, in part, influence the texture of the sediment, sources and delivery systems also play a significant role, particularly in the marine environment. Waves can work only the sand population provided, and textural variation is a highly significant but commonly underrated parameter in determining the nature of coastal facies.

Grain size largely determines the slope of a beach–nearshore system. Coarse systems are steep, the waves break near the beach, and wave energy tends to be reflected back into the ocean (Wright et al., 1979). Fine beach–nearshore systems slope more gently, with the result that waves break farther from the shoreline and dissipate their energy across a wide surf zone, in which bars and troughs are likely to develop. It is commonly assumed that high-energy beaches are steeper than their lower-energy counterparts, but studies of modern beaches have demonstrated just the reverse: they are more gently inclined (Komar, 1976). The assumption probably derives from erroneously mentally associating coarse beaches with high energy.

Grain size is a major influence in the size and shape of bedforms. This has long been known for unidirectional flow, but the textural relation may be even greater for wave-generated structures. Figure 30 shows the spacing of symmetric ripples as a function of maximum bottom orbital velocity and grain size for a ten-second wave. Only very small ripples form in fine to very fine sand at the same orbital velocities that generate megaripples in

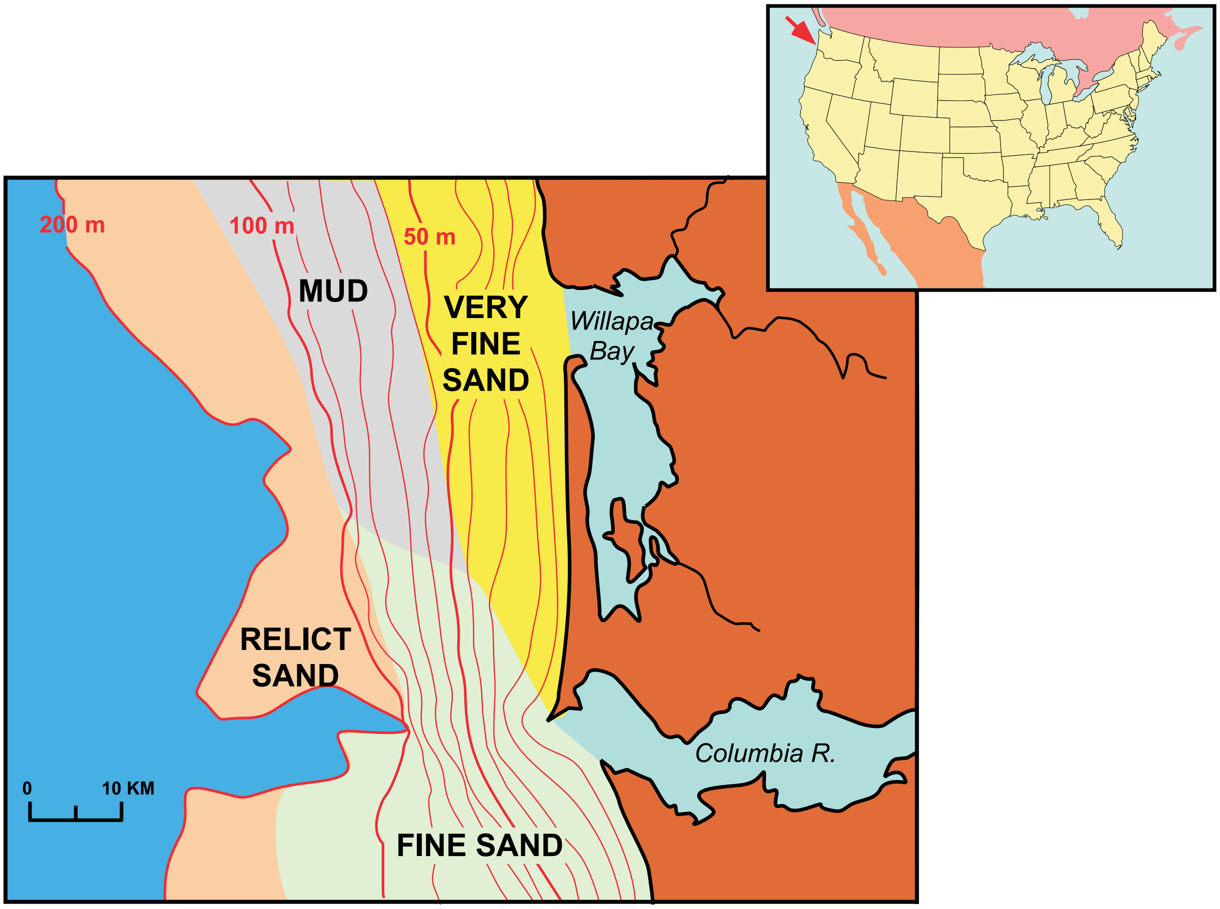

FIG. 22.—Distribution of sand and mud on the continental shelf off the mouth of the Columbia River, Pacific Northwest Coast of the United States. After Nittrouer et al. (1986).

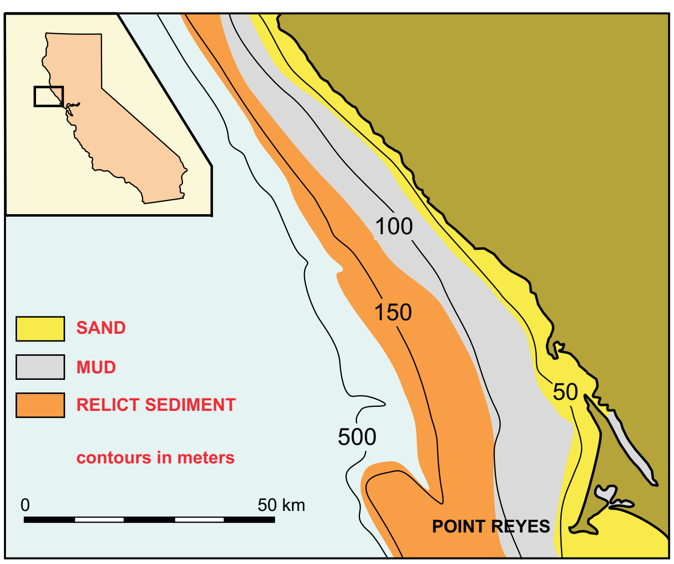

FIG. 23.—Distribution of sand and mud on the shelf off central California. The sand–mud transition lies at a water depth of about 60 m, well seaward of the shoreface. After Drake and Cacchione (1985).

medium to coarse sand. It is unlikely that bedforms capable of generating cross-bedding sets more than a few centimeters thick can develop in fine to very fine sand under purely oscillatory flow.

A comparison of texturally disparate coasts that have similar wave climates illustrates the degree to which texture influences facies character. Study of medium- to coarse-grained sandy nonbarred nearshores in southern Oregon (Clifton et al., 1971) showed that the sedimentary structures were arrayed in simple shore-parallel patterns that reflected the transitions in the shoaling waves (Fig. 31A). Asymmetric ripples dominate in the offshore area, converting to decimeters-high lunate megaripples in the area of most intense wave buildup just seaward from the surf zone. Landward migration of these bedforms produces landward-dipping trough cross-bedding. Within the surf zone, the bed is essentially flat; small, low-amplitude, transitory ripples form between intervals of sheet flow as the waves passed overhead. Adjacent to the beach foreshore, the bottom again becomes irregular at the interface between surf and swash zones. Bedforms here faced seaward and produced seaward-dipping cross-bedding. Within the foreshore, the bed is planar, and the sediment contains gently inclined or planar parallel lamination.

The zones of sedimentary structures shift back and forth with changes in wave climate and tides, producing assemblages of structures. An offshore–nearshore transition zone contains ripple

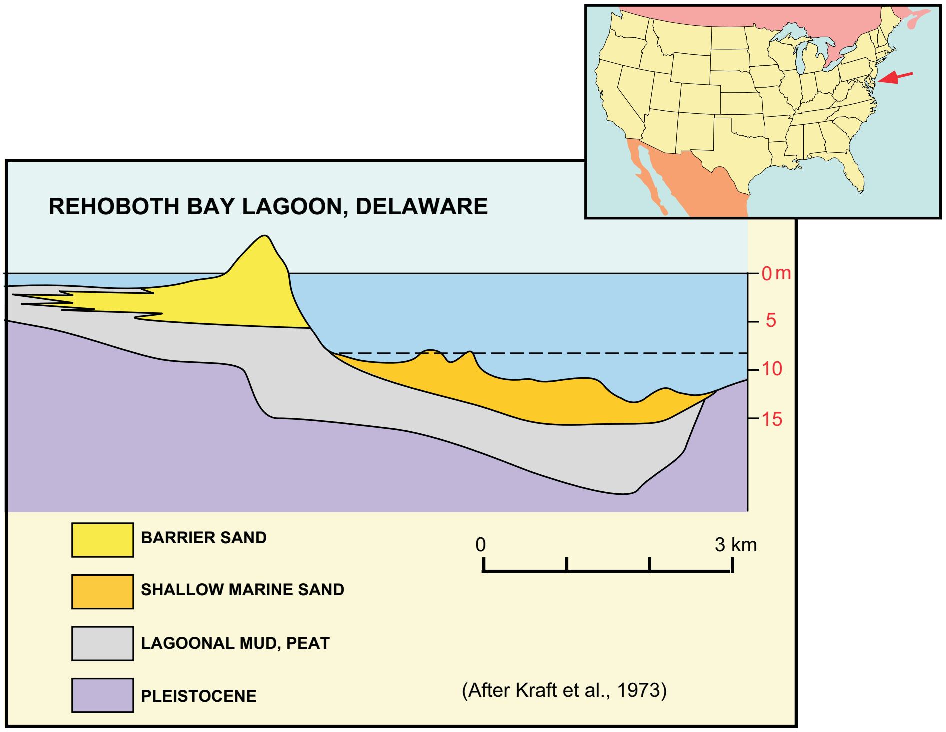

FIG. 24.—Distribution of sand on the shelf adjacent to the barrier at Rehoboth Bay, Delaware (arrow, above), after Kraft et al. (1973). Shallow marine sand may be an older palimpsest deposit, but it demonstrates the lack of correspondence of the base of the shoreface with the sand-mud transition.

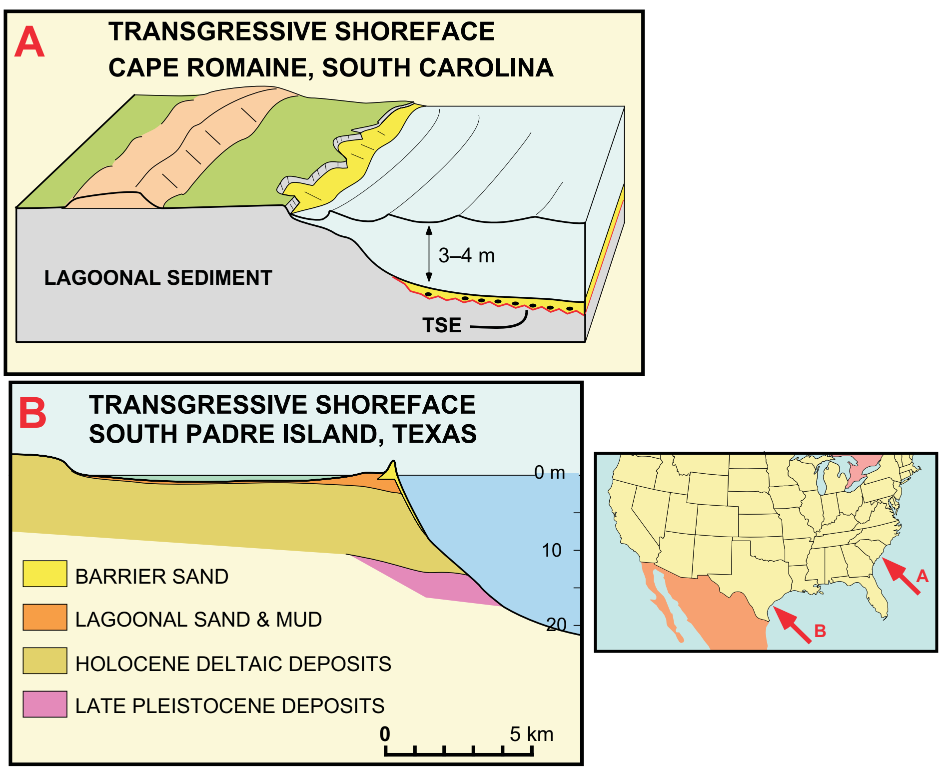

FIG. 25.—Variability in transgressive shoreface profiles, Cape Romaine, South Carolina, U.S.A. (after Hayes and Sexton, 1989) and South Padre Island, Texas, U.S.A. (after Morton, 1994).

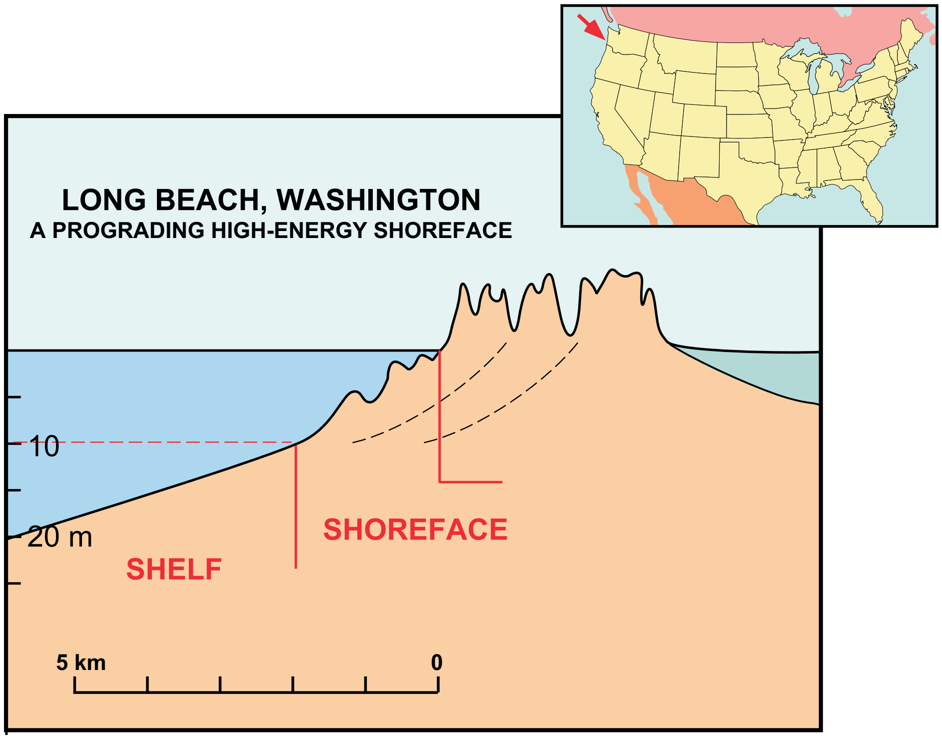

FIG. 26.—Shoreface profile on a high-energy, prograding coast, Long Beach, Washington, U.S.A. Break in slope defines the shoreface–shelf boundary lies a water depth of about 10 m. After Dingler and Clifton (1994).

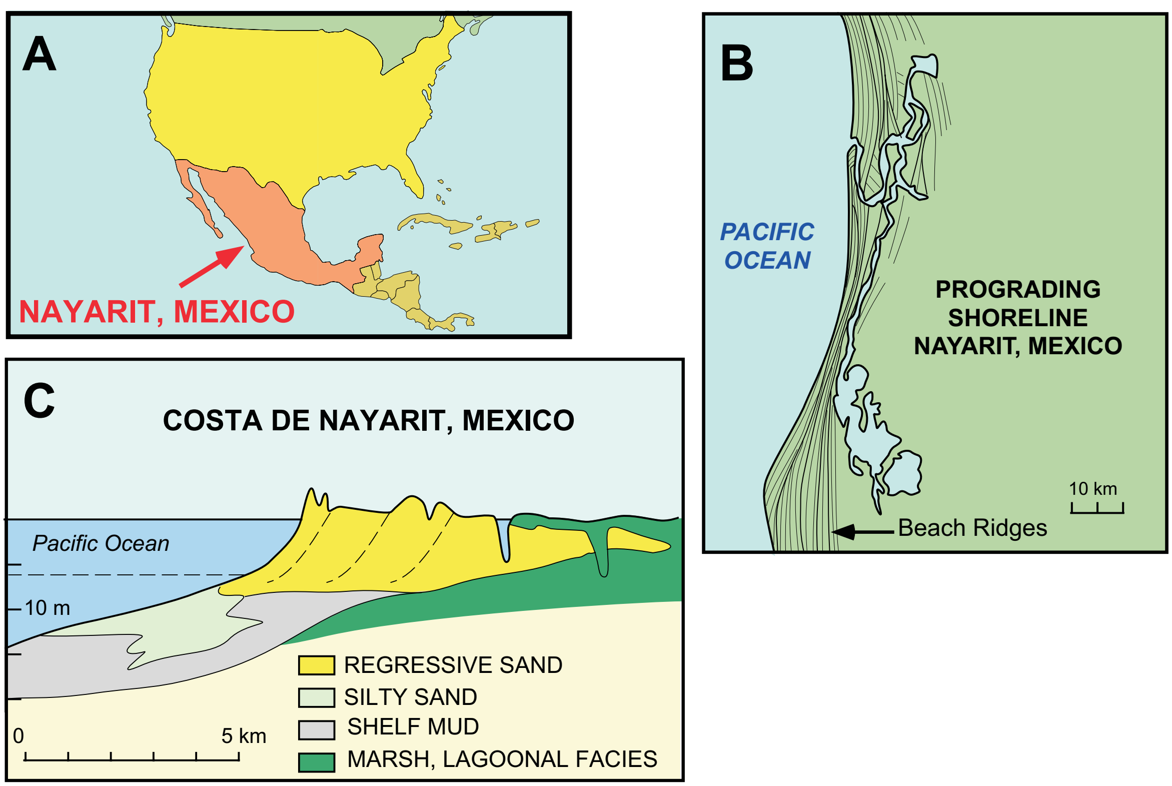

FIG. 27.—**A)** Location, **B)** plan view, and **C)** profile of the prograding coast at Nayarit, Mexico. Break in slope that defines shoreface–shelf boundary lies in about 6-7 m of water. After Curray et al. (1969).

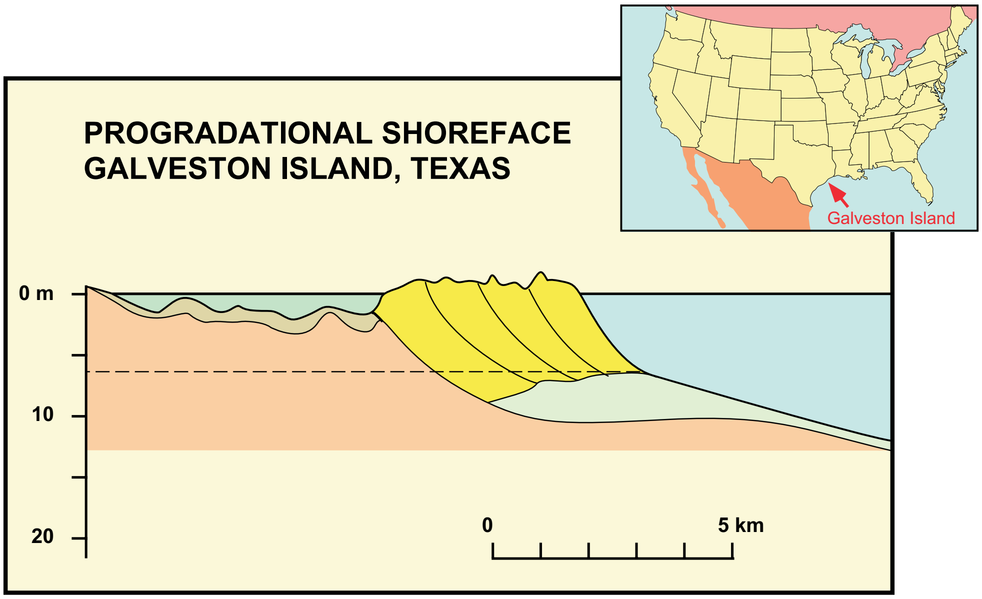

FIG. 28.—Profile across Galveston Island, Texas, U.S.A. Break in slope that defines shoreface–shelf boundary lies in about 6–7 m of water. After Morton (1994).

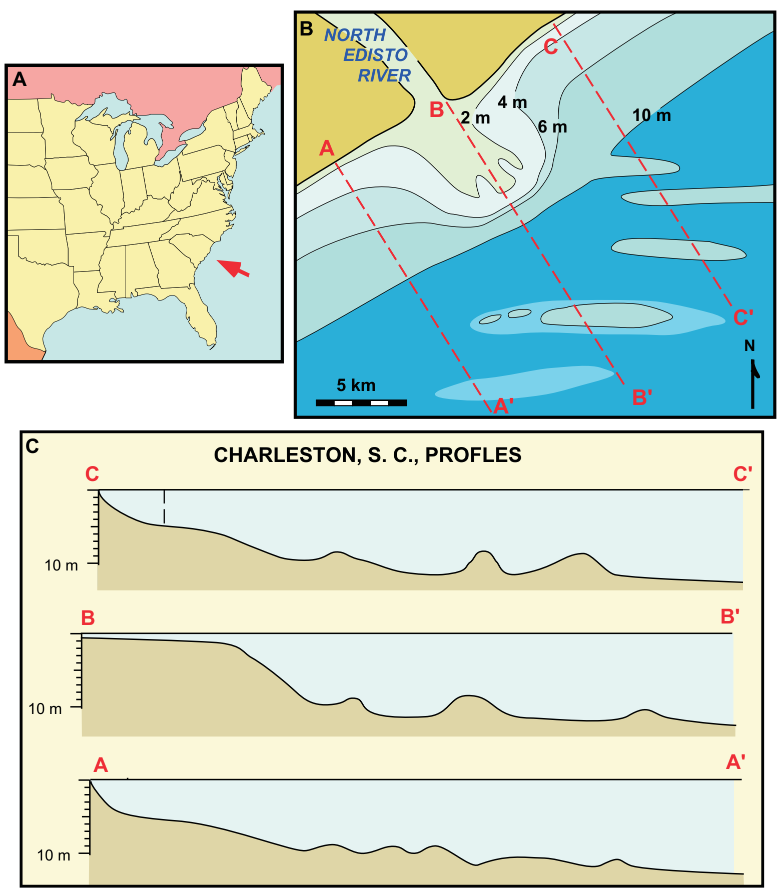

FIG. 29.—Shoreface-attached ridges and isolated ridges on inner shelf, off the North Edisto River, Charleston, South Carolina, U.S.A. **A)** Index map. **B)** Plan view. **C)** Cross section. Note break in slope at about 5 m on profiles A–A' and C–C', which is probably the wave-cut shoreface on this complex coast.

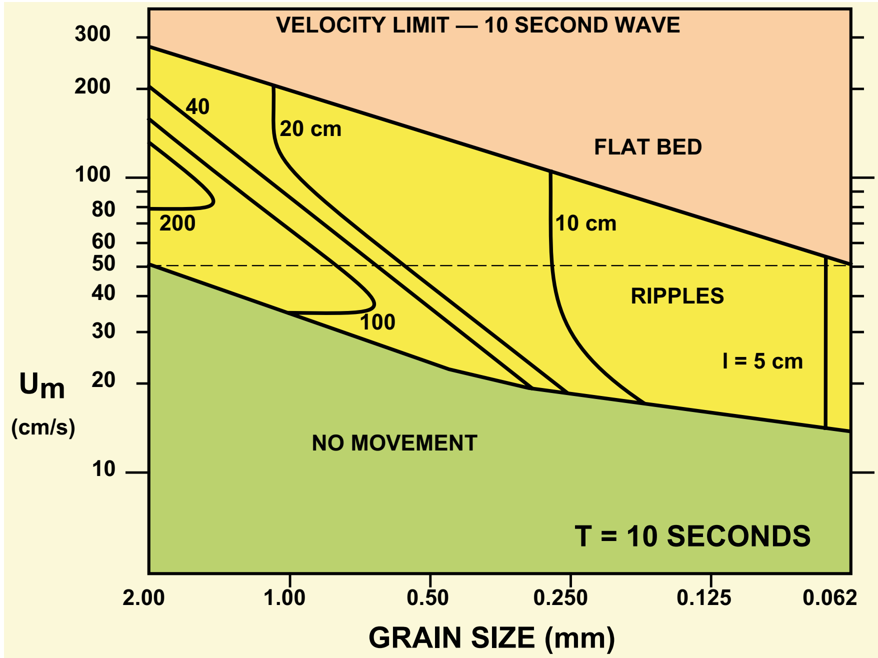

FIG. 30.—Variation in ripple size (spacing = l) as a function of grain size and maximum bottom orbital velocity (*Um*) under a wave with a period of 10 seconds. After Clifton (1976).

lamination, something similar to swaly cross-stratification, and landward-facing trough cross-bedding generated under conditions of heavy seas. A surf assemblage includes planar lamination and landward- and seaward-dipping trough cross-bedding. A surf–swash transition assemblage contains planar lamination interrupted by wedges or troughs of seaward-dipping cross-bedding. Progradation of a high-energy, non-barred, fair-weather, coarse sandy, wave-dominated shoreline produces a stacking of these assemblages in an upward-shallowing succession (Fig. 31B). In contrast, a fine sandy high-energy, nonbarred nearshore lacks the larger bedforms (Fig. 32A), and the vertical succession shows a dominance of planar or gently undulating stratification (Fig. 32B).

Where the nearshore is composed of gravel, large straight-crested, two-dimensional ripples or megaripples predominate (Fig. 33A). Stratification typically is difficult to delineate in the gravel beds, and the resulting nearshore succession consists of well-segregated layers and lenses of apparently structureless gravel interbedded with sand showing flat or inclined lamination and cross-bedding (Fig. 33B).

Sediment of different caliber can be transported in different directions under the same set of waves. The selective shoreward transport of the larger clasts by asymmetric orbital currents under shoaling waves has been noted (Fig. 9). Where megaripples or other large bedforms exist, the asymmetric flow may result in coarser sand moving landward as part of the bed load, and finer sand moving seaward as part of a suspended load (Inman and Bowen, 1963). While diving in the southern Oregon surf zone, we noted that fountains of suspended sand commonly erupted from the lee sides of megaripples as the landward surge of a wave diminished. The cloud of sediment would then drift seaward under the offshore component of the orbital motion. Observations of both the clouds of sediment and dye streams released as the sand fountains erupted indicated that the suspended sediment never settled landward of the point of origination. The process resulted in the finer sand moving offshore even as the coarser sand was being driven shoreward (Fig. 34). This process provides an effective means of textural segregation of sand on the upper shoreface (Komar, 1976).

Texture can also influence processes in the nearshore. A striking example occurs on the Surinam coast of South America, where suspended fine sediment discharged from the Amazon River accumulates in the nearshore area. A zone of fluid mud concentrated near the shoreline damps about 95% of the wave energy and transforms the incoming waves to a solitary wave form (Wells and Coleman, 1978; Rine and Ginsburg, 1985). Solitary waves are waves of translation that transport water (and mud) shoreward where the mud is trapped against the beach. Texture also controls rates of bioturbation, which tend to be most rapid in fine to very fine sand and diminish as the grain size either increases to coarse sand or gravel or decreases to mud.

EFFECTS OF BARS IN NEARSHORE SYSTEMS

Nearshore systems with breaker bars or other bars are inherently more difficult to study than are nonbarred coasts, particularly on high-energy coasts with intense longshore and rip currents. As a result several detailed open-coast studies that are applied to facies models were conducted on nonbarred coasts (Clifton et al. 1971; Howard and Reineck, 1981).

Bar–trough systems, however, are common on many, if not most, coastlines, and probably form part of nearly all ancient open-coast successions. The development of bars and troughs is commonly linked to nearshore circulation cells of longshore and rip currents (Fig. 11). Typically, on modern coasts, the location of offshore bars adjacent to a beach is readily seen from the breaking pattern of waves in the bar crests. The studies noted below have focused on the facies of bar–trough systems on modern open coasts.

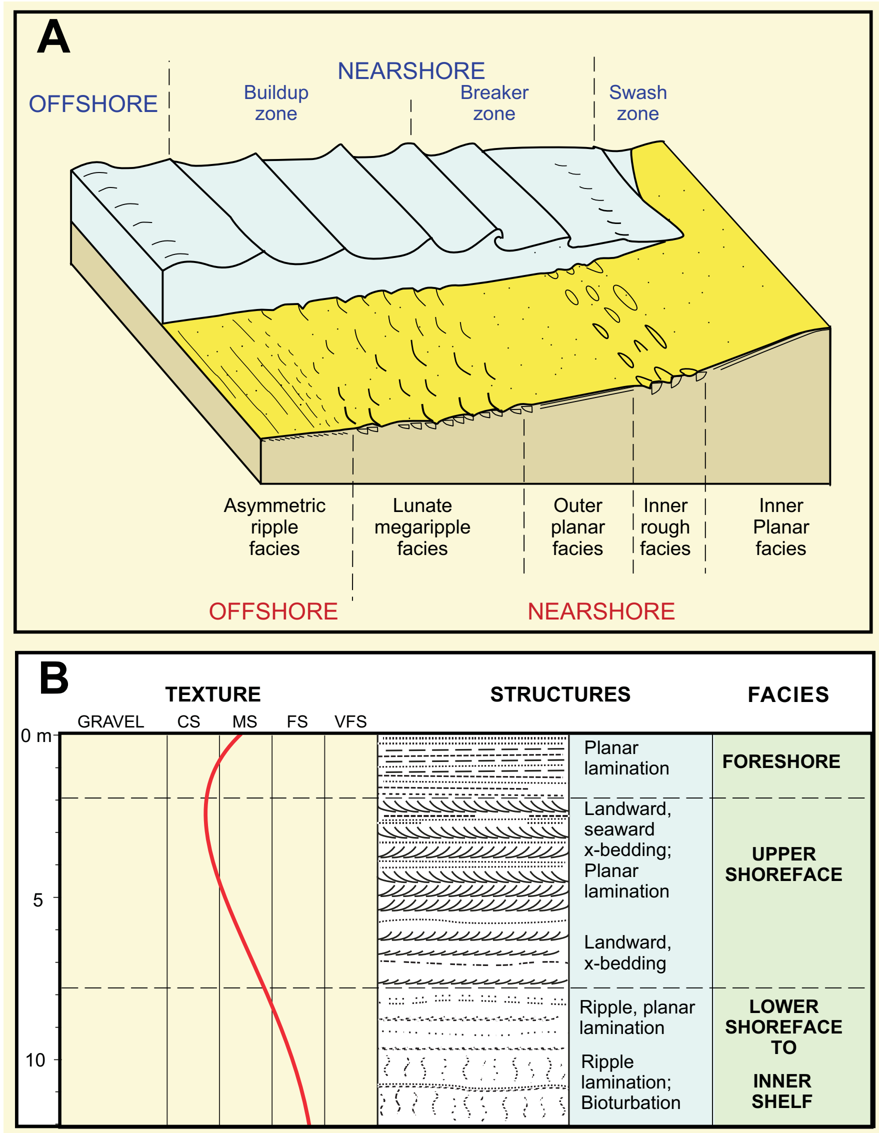

FIG. 31.—**A)** Sedimentary structural facies in the non-barred nearshore (upper shoreface) in coarse sandy sediment on the high-energy coast of southern Oregon, U.S.A., under fair-weather conditions. **B)** Vertical succession produced by progradation of such a system.

Bars and troughs may be parallel to the shoreline and can occur in multiple sets of two or three, such as those on the Texas Gulf Coast (Hill and Hunter, 1976). Here, breaking waves shape bar crests into a plane bed; the sea floor in deeper water is covered with wave ripples. Longshore currents stronger than about 0.5 m/s generate small dunes that produce medium-scale, shore-parallel cross-bedding. Hill and Hunter (1976) note that intense bioturbation destroys physical structures that lie 30 cm or more beneath the sediment–water interface.

Where waves approach a coast obliquely, the bars and troughs are likely to develop an en echelon pattern, in which individual bars are oblique and attached to the shoreline (Fig. 35). Study of an attached oblique bar on the southern coast of Oregon (Hunter et al., 1979) showed that the sedimentary facies reflect the circulation cell. Water flows landward across the bar in the form of very asymmetric oscillatory flow, generating either a flat bed or lunate megaripples. In the longshore trough and rip channel, medium to small subaqueous dunes migrate in the direction of flow, respectively, producing longshore- and offshore-directed cross-bedding (Fig. 36).

Some coasts are characterized by irregular bar systems. Davidson-Arnott and Greenwood (1976) describe the facies that form in mostly medium- to-fine-grained sand along the shore of Kouchibouguac Bay, New Brunswick. Two sets of bars occur, broadly shore-parallel, but with much irregularity. Bar crests here are composed of a combination of flat bedding and cross-bedding, and the troughs are largely underlain by ripple-laminated sand. Rip channels that cut through the inner bar are underlain by seaward-facing ripple lamination and cross-bedding.

In all three examples, the bars shift landward and seaward as wave conditions change. The oblique bars on the Oregon coast

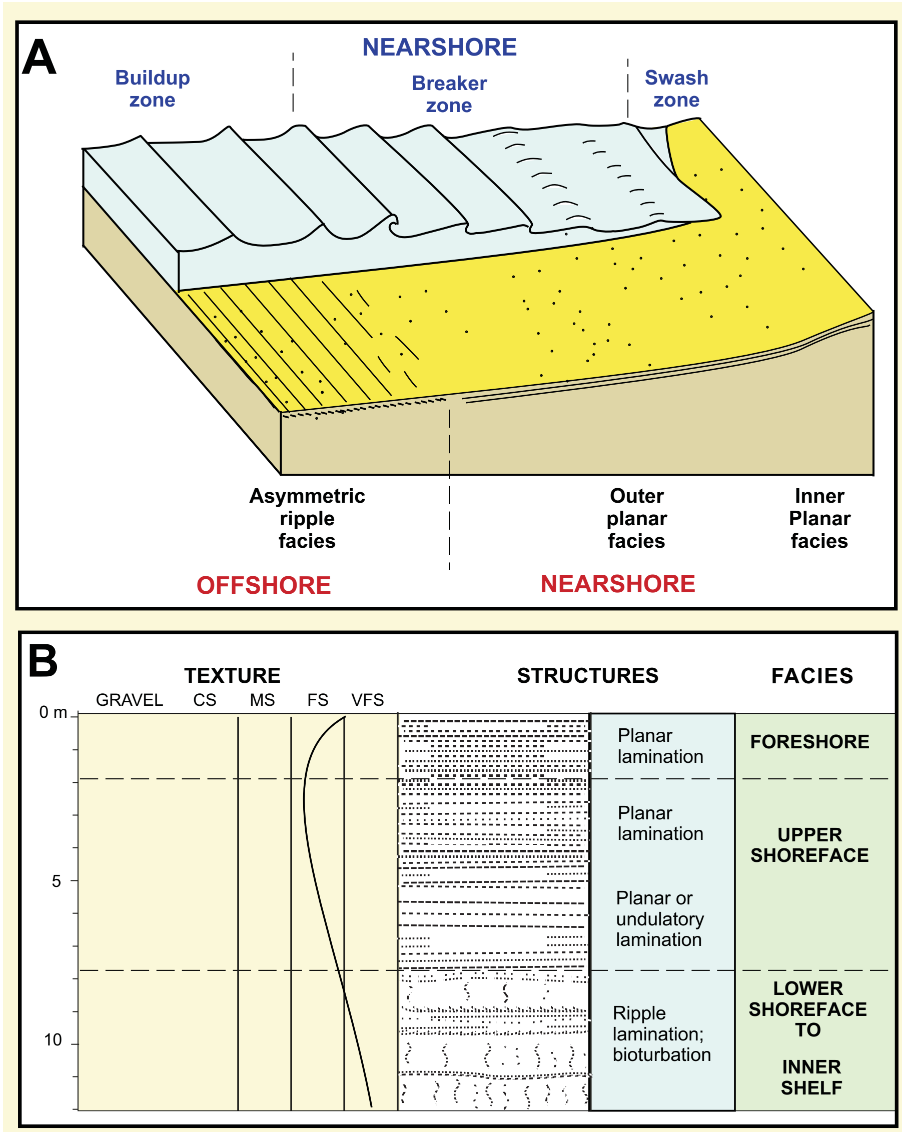

FIG. 32.—**A)** Sedimentary structural facies in the non-barred nearshore (upper shoreface) in fine sandy sediment on a high-energy coast under fair-weather conditions. Flatter beach–nearshore profile expands the surf zone relative to coarser shorelines. No medium- to large-scale bedforms. Surf and swash zones are underlain by planar parallel lamination. **B)** Vertical succession produced by progradation of such a system. Section lacks cross-bedding that typifies the upper shoreface of coarser shorelines.

also migrate laterally at rates of 100–200 m/ month and generate an envelope of bar–trough sedimentary facies (Fig. 37A). During progradation of this envelope, the currents in the trough landward from the bar erode previously deposited bar facies (Fig. 37B). Therefore the bar itself, although apparently the dominant feature in the system, has a low potential for preservation. Davidson-Arnott and Greenwood (1976) reach a similar conclusion for the bars on the New Brunswick coast. Most interpretations of sand bodies as "offshore bars" are probably wrong. The vertical succession produced by a prograding nearshore bar–trough system contains an erosional surface that separates rip-channel and longshore-trough facies from subjacent finer sand deposited on the seaward side of the bar (Fig. 38).

The net effect of bar–trough systems is to enhance the unidirectional flow of rip currents and longshore currents and to create internal erosional surfaces marking the seaward migration of the troughs during progradation. The unidirectional currents can generate bedforms in fine sand that would be shaped into a flat bed or ripples by oscillatory flow (Fig. 39A). As a result, a progradational succession produced by a barred nearshore system composed of fine sand can show abundant cross-bedding that otherwise would be absent (Fig. 39B).

FAIR-WEATHER OBSERVATIONS AND STORM-DOMINATED SYSTEMS

Most of our direct observations of nearshore processes come from studies conducted under conditions of fair weather, when data can be collected most easily. Yet it is likely that processes operating during storms dominate much of the

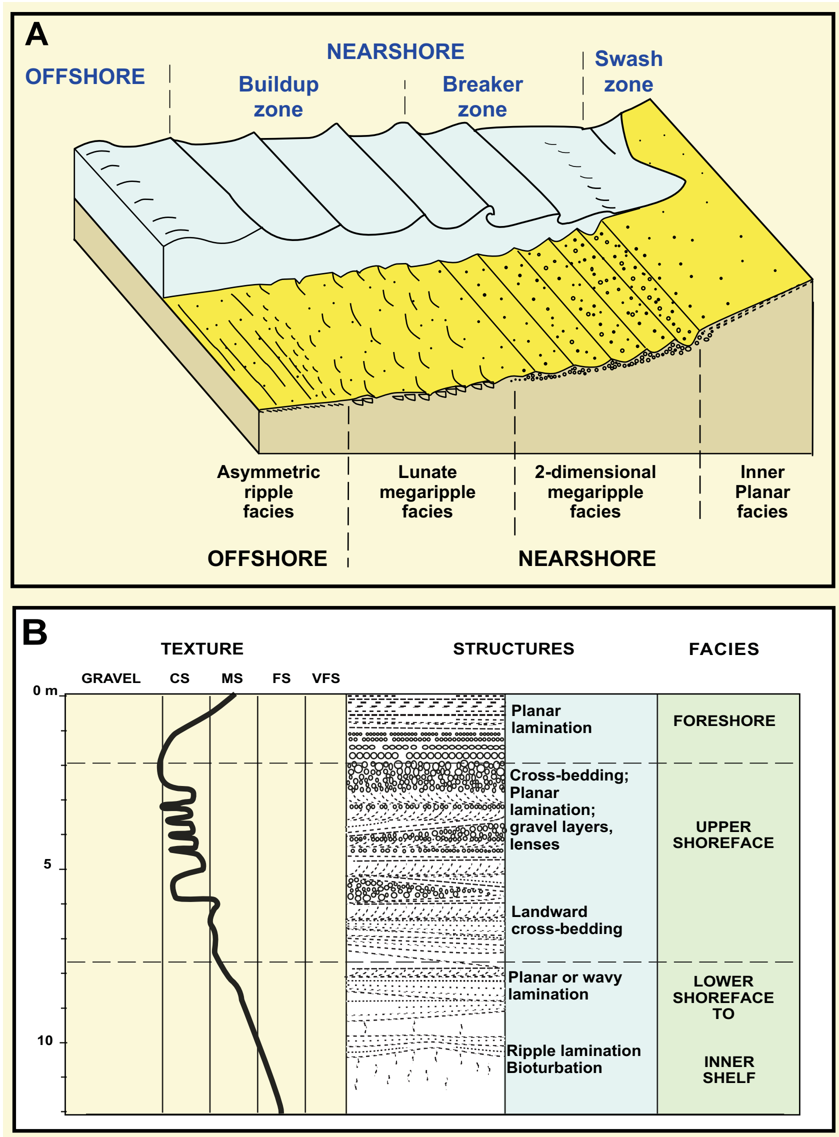

FIG. 33.—**A)** Sedimentary structural facies in the non-barred nearshore (upper shoreface) in coarse gravelly sand on a high-energy coast under fair-weather conditions. Large two-dimensional, straight-crested ripples occur in the gravel. These ripples tend to face landward near the beach and be symmetrical in deeper water. **B)** Stratigraphic succession produced by progradation of such a system.

nearshore stratigraphic record. Storms influence almost all coastlines, and, compared to fair-weather waves, are capable of eroding, transporting, and depositing vast quantities of sediment. Analysis of wave records along most coasts indicates a pattern where most of the time is occupied by fair-weather conditions, a small but significant component of time is occupied by typical large annual storms, and a tiny fraction of time is occupied by very infrequent major storms (Fig. 40). Each of these marks the sedimentary record in different ways, depending on water depth. On the inner shelf, extreme events are likely to produce the only physical structures in sediment otherwise dominated by bioturbation. The presence of sand in this environment, however, by itself probably attests to transport and deposition during storms of a wide range of sizes. On the shoreface, the effects of storm and fair-weather cross-shore transport become important (Fig. 41). The enhancement of rip currents and their extent into deeper water during storms (Fig. 42) provides a mechanism for transporting a wide range of grain sizes to or beyond the base of the shoreface. Shoaling waves following the storm drive much of this material back onto the upper shoreface, but some of the coarser grains (small pebbles, granules) are likely to be left behind, trapped in burrows or other depressions. The resulting bimodal sediment forms a distinctive lower-shoreface facies in coarse sediment (Figs. 43, 44). In the absence of these pebbles, it may be very difficult to distinguish between lower-shoreface and subjacent sandy-shelf facies. The upper shoreface is likely to be dominated by storm processes and consist of rip-current deposits, storm lags, and other storm-generated features (Fig. 43). Beach foreshores are eroded during storms, and aggrade in fair-weather

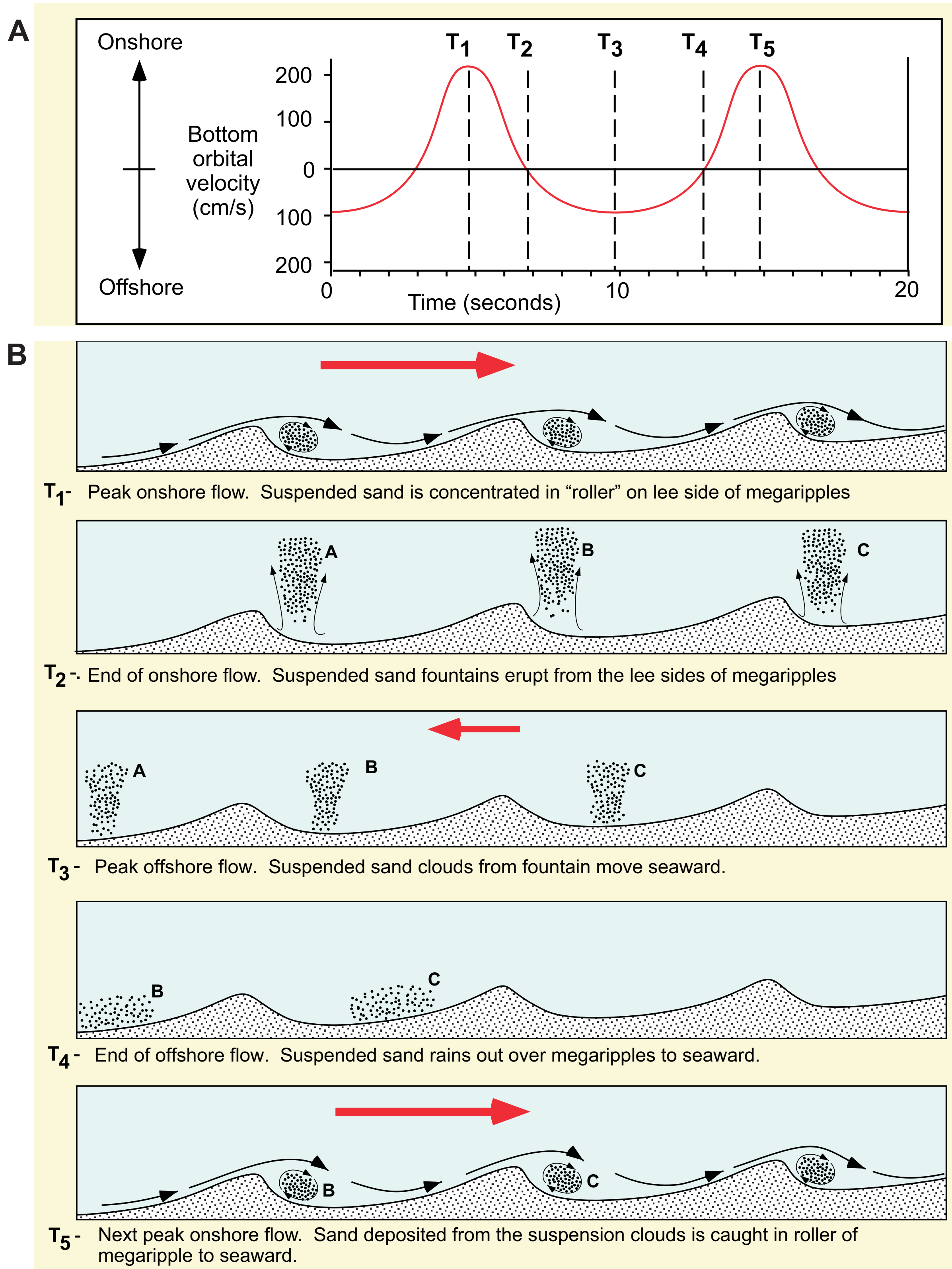

FIG. 34.—Differential transport of coarse and fine sand under the same set of waves. **A)** Velocity profile of currents generated at the sea bed by a passing waves. **B)** T1–T5 correspond to times shown in Part A. Coarse sand moves shoreward as bed load, whereas fine sand is thrown into suspension and drifts seaward, where it is trapped in rollers on the lee side of megaripples farther offshore. Transport of sand to seaward in a field of megaripples under strongly asymmetric orbital flow.

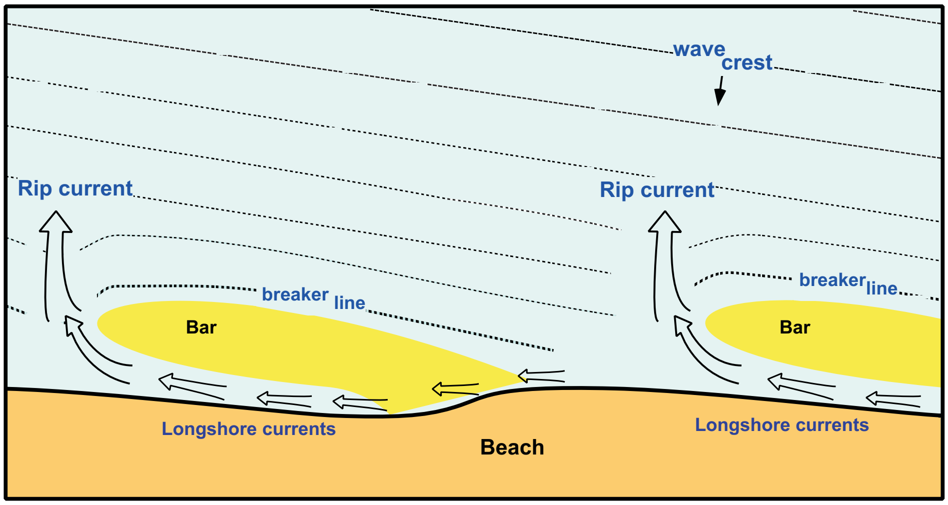

FIG. 35.—Nearshore circulation cells where wave incidence is oblique to coast. Longshore currents tend to flow in one direction only. Such cells promote the development of attached oblique bars (shown in yellow).

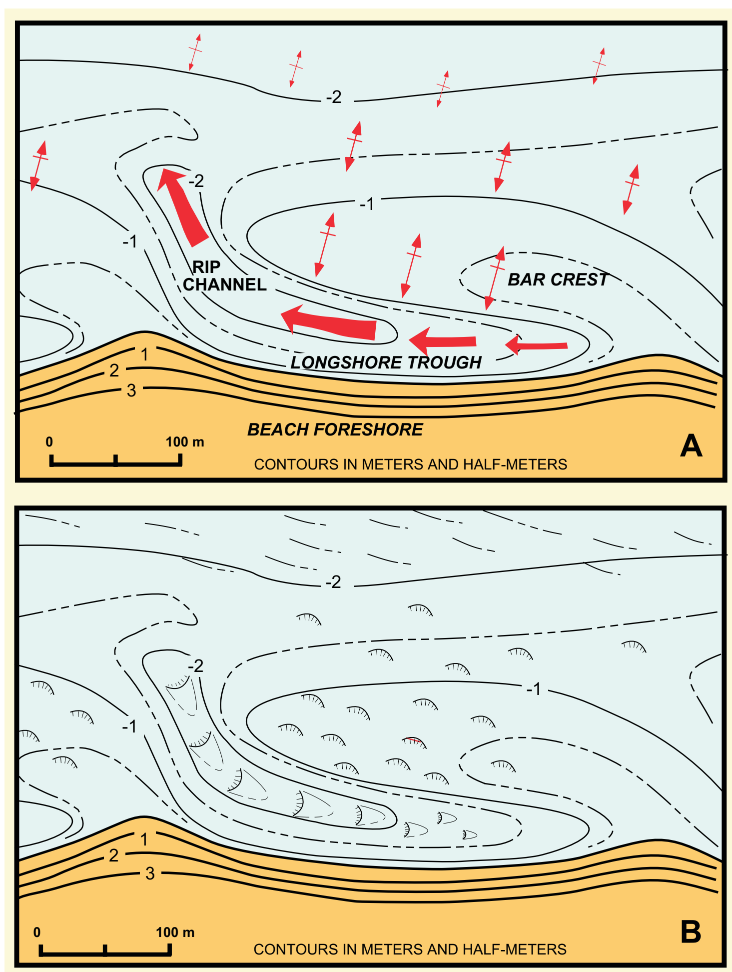

FIG. 36.—Bar and trough system on the southern coast of Oregon (Pistol River), U.S.A. **A)** Morphology and currents associated with the system. **B)** Sedimentary structures associated with this system. Dunes occupy the longshore trough and rip channel, whereas the bar crest is covered by lunate megaripples and/or a flat bed. Sediment in the longshore trough is coarser than that on the bar or in the rip channel.

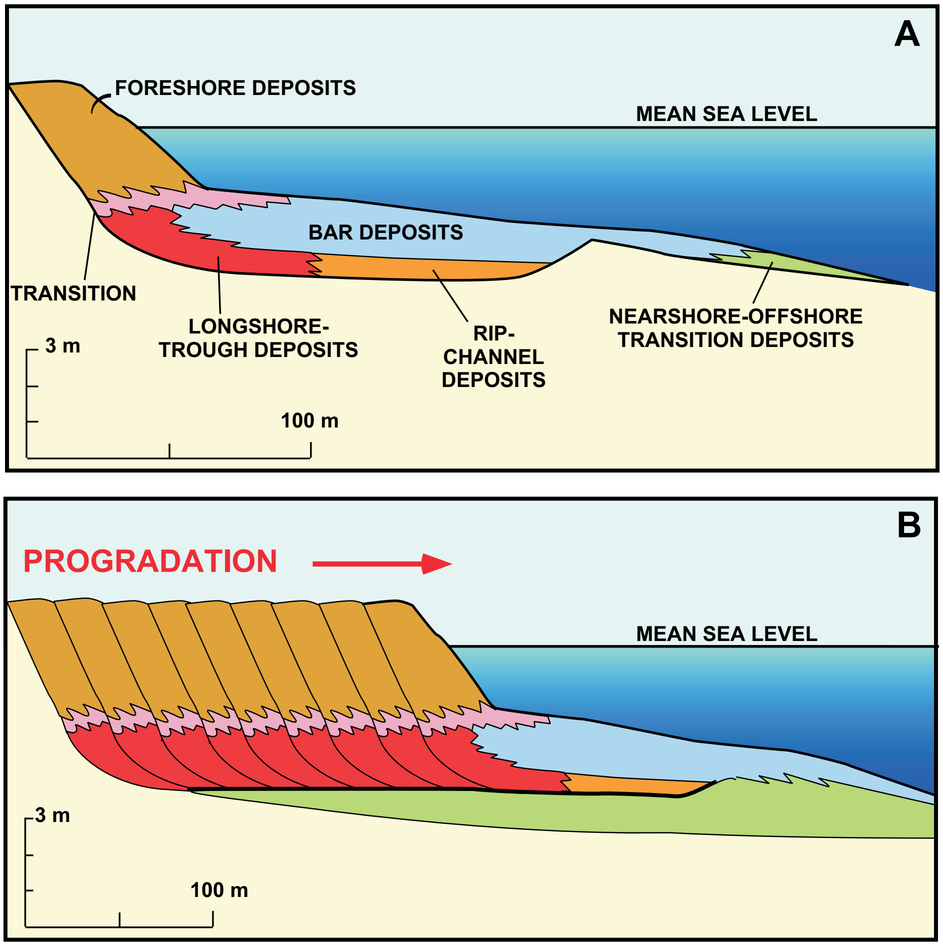

FIG. 37.—Depositional facies in a prograding bar–trough system. **A)** Envelope of sedimentary facies generated by laterally migrating shore-attached oblique bar system, as visualized in a cross section projected normal to the shoreline. **B)** Effects of progradation of a laterally migrating shore-attached oblique bar system, as visualized in a cross section normal to the shoreline. Figure shows the limited potential for preserving bar facies and the development of an internal erosional surface beneath the longshore trough and rip channel. Coarse sediment from these environments sits abruptly over finer sediment deposited at similar water depths outside the bar (nearshore–offshore transition). After Hunter et al. (1979).

intervals. Most beach foreshores thus consist primarily of fair-weather deposits.

The dominance of storm effects on the upper shoreface implies that features observed during fair-weather conditions may rarely be preserved. Examples include the lunate megaripples found in the high-energy nearshore of southern Oregon (Clifton et al., 1971). Decimeters-high dunes that migrate landward under the asymmetric flow of shoaling waves, these features produce medium-scale trough cross-bedding in the area just seaward of the breaker zone. Examination of many nearshore successions in the stratigraphic record shows that onshore trough cross-bedding is uncommon (Fig. 45). The lunate megaripples are fair-weather phenomena that are obliterated by the enhanced rip currents and longshore flow that accompanies storms.

Conversely, several structures that are common in the sedimentary record have never been seen during their formation. Gravel-filled gutter casts are common in pebbly nearshore sandstones. These shore-normal structures commonly have steep, or even undercut, sides indicating nearly simultaneous cutting and filling (Chiocci and Clifton, 1991). Their orientation and shape suggest that the gutter casts form when large waves drag gravel back and forth on a sandy bed during a storm. They have not been seen under fair-weather conditions.

Another structure attributed by most workers to storms is hummocky cross-stratification. This feature, which is common to nearly all shoreface and shelf fine-grained sandstone deposits, has been observed forming only once in the natural environment (Greenwood and Sherman, 1986). Many questions about the dynamics of its formation and its significance remain unanswered.

BIAS TOWARD LOW-ENERGY COASTS

Unsurprisingly, most of our knowledge about modern shoreface facies comes from the study of coasts with low wave energy. The few studies of high-energy systems, where the wave heights are routinely in the range of 1–2 m, have been focused on nearshore areas close to the shoreline. As a result, our understanding of modern open coasts is strongly biased toward low-energy systems. The result has been some errone-

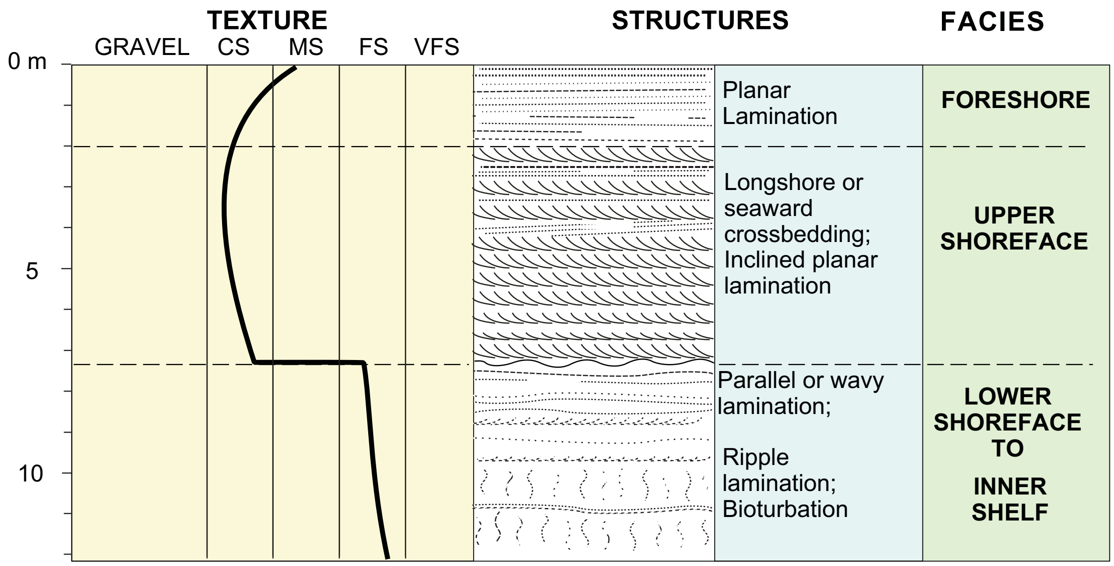

FIG. 38.—Vertical sequence produced by progradation of an attached, oblique bar system in a high-energy, coarse sandy setting.

ous generalizations that pervade the models of the wave-dominated coastal facies.

Wave Base

Perhaps the most broadly held misconception in the interpretation of shoreface systems is that of the role of fair-weather wave base relative to facies distributions. Reineck and Singh, in their justifiably influential book on depositional sedimentary environments (1973), state that the seaward limit of the shoreface corresponds to wave base, which in they identify as the "average maximum wave base". From their studies in the low-energy Gulf of Gaeta, they concluded that the boundary between the upper and lower shoreface corresponded to fair-weather wave base. In the low-energy setting of Long Island, New York, U.S.A., Shipp (1984) found that the maximum depth to which fair-weather waves moved sediment corresponded with the base of the shoreface. Such studies were incorporated into the models for a wave-dominated coast. Fair-weather wave base coincides with the base of the shoreface in the models provided by Walker and Plint (1992) and Reading (1996). The same relation is implicit in the text of Galloway and Hobday (1996), who note that the lower shoreface is influenced by both storm and fair-weather waves, whereas fair-weather waves (other than long-period swell) have little effect on the shelf. Consequently, a number of workers postulate that fair-weather wave base defines the shelf–shoreface boundary (e.g., MacEachearn and Pemberton, 1992; Maejima, 1993; Hettinger et al., 1994; Hart and Plint, 1995; Hampson and Storms, 2003).

Although this interpretation may be valid for some successions, it is invalid as a generalization. First, wave base is so variously defined that it has lost much of its currency. Most geologists identify wave base as the greatest water depth in which passing waves disturb the bed, although some have used the term to separate the zones of "normal" wave erosion and wave deposition (Kowalewsky, 1982), which is postulated to occur at a depth of about 10 m (Schwartz, 1982). Plint (1988) seems to use this definition in his model of forced regression. Physical oceanographers have placed wave base at a depth where waves begin to "feel bottom", approximately equivalent to one-half of the wave length (Sverdrup et al., 1942), and many geologists have followed suit (Walker and Plint, 1992; Reading, 1996). This definition, however, does not consider the effect of wave height; large waves disturb the bottom at depth uninfluenced by smaller waves of the same period or wavelength. The concept that wave base equates with the water depth in which sediment first begins to move implies that wave base is partly dependent on sediment caliber; under the same set of waves, a fine bed might be above wave base whereas a coarser bed might not be. Moreover, wave theory predicts that a bed of fine sand is mobilized by fair-weather long-period swell at water depths much beyond the base of the shoreface (Fig. 46A). Shorter-period waves, such as those characteristic of a low-energy coast, move fine sand sediment in considerably shallow water (Fig. 46B). A coincidence of fair-weather wave base and the base of the shoreface, however, is likely to be just that: a coincidence depending on wave height and period and sediment grain size. It is noteworthy that for both long-period (10 s) and shorter-period (5 s) waves, a water depth equivalent to one-half of the deepwater wave length lies well seaward of the shoreface base on a prograding shoreline (Fig. 46A, B).

Finally, geologists can interpret wave base in the stratigraphic record only by inference, such as the balance between physical depositional structures and biogenic structures or the presence of mud layers in the section. In the first case, fine sand is commonly completely bioturbated even at water depths where everyday waves ripple the surface. The ripples would indicate deposition above fair-weather wave base, but the bioturbation could suggest deposition possibly below storm wave base. Layers of mud can accumulate in shallow water from the rapid settling of large volumes of silt and/or clay resuspended by storm waves or introduced by floods. Their presence is unrelated to fair-weather conditions. The extension of generalizations regarding wave base, drawn from studies of modern low-energy coasts, is largely unwarranted.

Wave Energy and Facies

Several workers have attempted to compare the facies of high-energy and low-energy coasts. Clifton (1976) contrasted the fair-

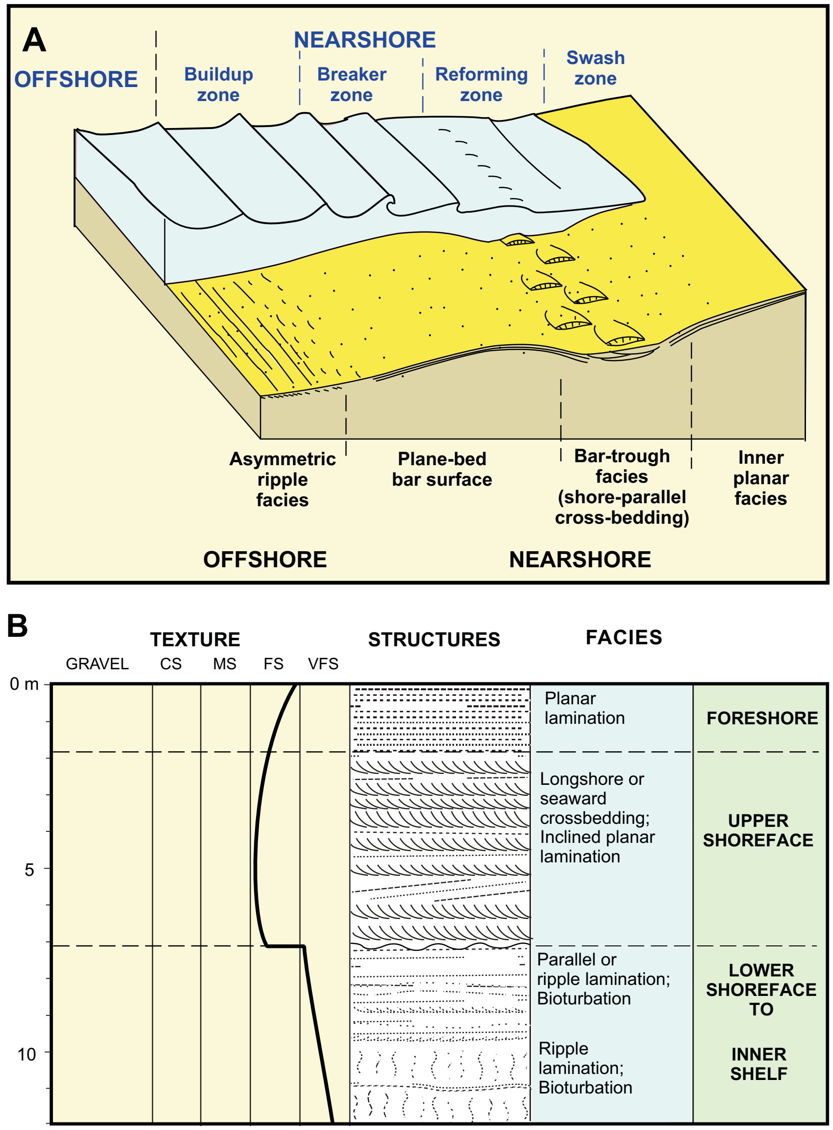

FIG. 39.—**A)** Sedimentary structural facies in a barred high- to moderate-energy fine-grained nearshore under fair-weather conditions. Longshore flow in the trough of the shore-parallel bar is strong enough to create small dunes that migrate alongshore in the trough. **B)** Vertical succession produced by progradation of the system shown in Part A. Fine sand shows much shore-parallel trough cross-bedding that would not be formed without unidirectional flow within a bar–tough system.

weather nearshore facies in sand of similar texture on the very high-energy coast of southern Oregon, the moderately low-energy coast of southeastern Spain, and the very low-wave-energy beach within Willapa Bay, Washington, U.S.A., (Fig. 47A). The facies distribution becomes increasingly compressed as facies requiring relatively high velocities progressively disappear as wave energy is decreased. The upper-flow-regime plane bed that characterizes the inner surf zone of southern Oregon is absent on the Spanish coast, and the lunate megaripples and cross-ripples observed in Oregon and Spain are absent in Willapa Bay. As the wave energy of the setting diminishes, the progradational successions (Fig. 47B) become thinner and increasingly impoverished in sedimentary structures requiring relatively strong currents.

Howard and Reineck (1981) contrast the facies succession from a high-energy coast in Southern California with that of a low-energy coast in Georgia, U.S.A. Their analysis is based on both box-core and vibracore analyses, which incorporates the effects of coastal storms. Their comparison shows a similar thinning of the facies progression and loss of higher-energy features in the low-energy setting (Fig. 48). The thickness of the shoreface section that would result from progradation of these coasts differs from about 9 m in California to 2 m in the example from Georgia. Cross-bedding, common in the shoreface deposits in California, is absent in the Georgia succession. Some textural differences may exist in these two examples. The California nearshore is typified by fine- to medium-grained sand (Howard and Reineck, 1981), whereas the Georgia nearshore seems to be composed of uniformly fine-grained sand (Howard et al., 1972). As noted below, textural differences may outweigh variations in wave energy in shaping coastal facies.

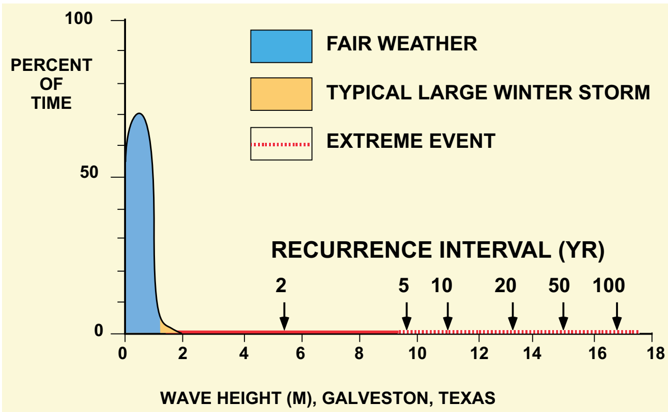

FIG. 40.—Energy (wave height) frequency for Galveston Island, Texas, U.S.A. During most of the time small waves prevail (fair-weather conditions). A small, but significant, amount of time is occupied by large-winter-storm conditions, and very large storm (hurricane) waves occur very infrequently. This pattern is typical for most coasts, although the wave heights involved may differ (for example, on the central California coast, waves 2 m high are fairly common, and typical large winter storm waves range from 4–5 m high).

Intuitively, it would seem that variations in energy regime would constitute a major influence on the character of shoreline facies. In reality, that influence can be difficult to resolve, largely because of textural complications. The nearshore profile off Padre Island (Fig. 49), where fine to very fine sand accumulates in a low-energy setting (fair-weather waves 0.2–0.5 m high tidal range > 1 m) differs markedly from that of the Oregon coast, (Fig. 31), where medium to coarse, pebbly sand accumulates in a high-energy setting (fair-weather waves 1–2 m high; tidal range 2–3 m). Much of the difference, however, may be due to the diverse textural character of these two environments. A comparison with a third shoreline, on the southeastern coast of Spain, helps to resolve the relative influences of texture and ambient energy. The environmental setting of this Spanish coast is very similar to that of the Texas Gulf coast (fair-weather wave heights in the range of 0.2–0.5 m, tidal range less than a meter), but texturally this tectonically active coast resembles that of Oregon. Profiles (Fig. 50) show development of a succession of structures similar to that seen on the Oregon coast. Even where bars composed of fine sand lie off the beaches, the intervening troughs are occupied by gravel shaped into large 2-D ripples like those shown in Figure 50B.

A comparison of outcrops representing each of these three environments (Figs. 51–54) likewise shows greater similarity between the Spanish and Oregon deposits. The primary difference between the two coarse-grained deposits is that the low-energy succession is significantly thinner that that formed under high-energy conditions, as a consequence of the deeper extent of wave influence on the high-energy California coast. Fine-grained, shallowing-up coastal succesions occur in the Eocene Jackson Group of West Texas, where presumably they accumulated under conditions similar to those on the present Texas Gulf Coast. Once exposed in now-covered uranium pits in West Texas, these successions are thinner than might be expected on a fine sandy coast with greater wave energy, and they also show more bioturbation than might occur on a high-energy coast. The successions differ strikingly from those formed in coarse sediment under similar oceanographic conditions off the coast of Spain. Although ambient wave energy is a factor in facies development, any interpretation of energy level must take into account the textural factor.

Low wave energy may be an important factor in shaping one particular type of shallow marine deposit. The shoreface is generally presumed to be an equilibrium profile for a given set of wave conditions. Where equilibrium is not achieved, owing to an inability of the waves to redistribute the introduced coarse sediment, and/or insufficient time to reshape the profile as new sediment accumulates, and/or a very steep offshore gradient, a Gilbert delta rather than a shoreface is likely to develop (Corner et al., 1990; Postma, 1990).

Gilbert deltas are characterized by steeply inclined foresets in tabular sets that can be tens of meters thick (Colella, 1988a, 1988b; Nemec, 1990). Sediment transport down the face of the delta occurs primarily by mass transport (Postma, 1984; Postma et al., 1988). Many, if not most, Gilbert deltas are conglomeratic, finer sediment being more easily shaped into a shoreface. But in areas of powerful waves, as along much of the U.S. West Coast, even the coarsest gravel can be reworked into an equilibrium profile and Gilbert deltas do not develop.

SHARP-BASED SHOREFACE DEPOSITS

The models for coastal deposits that formed during a forced regression derive almost entirely from the stratigraphic record. Although Pleistocene deposits on the outer part of modern continental shelves are cited as contemporary examples (Posamentier et al., 1992), little is known about the lithologic details of these deposits or the processes that attended their formation. The concept that forced regression can produce an extensive erosional base to shoreface deposits, as Plint (1988) postulated for the Cardium Formation, has been widely applied. Posamentier et al. (1992) note that, although coastal deposits in the Viking Formation in Joarcam Field, Alberta, Canada, seem produced by forced regression, shorefaces are sharp-based only in their most proximal position, rather than over the entire width of their occurrence, as in the Cardium. The difference between the two may

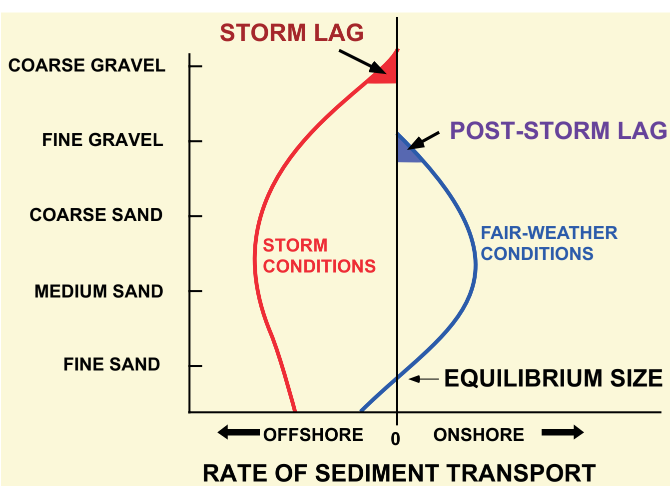

FIG. 41.—Idealized diagram showing rate and direction of transport on the shoreface as a function of grain size under storm and fair-weather conditions. Sediment carried seaward by storm rip currents is reworked in the aftermath of the storm. On the lower shoreface, fine sand in equilibrium with the fair-weather waves coexists with fine gravel deposited during the storm and left behind as a post-storm lag.

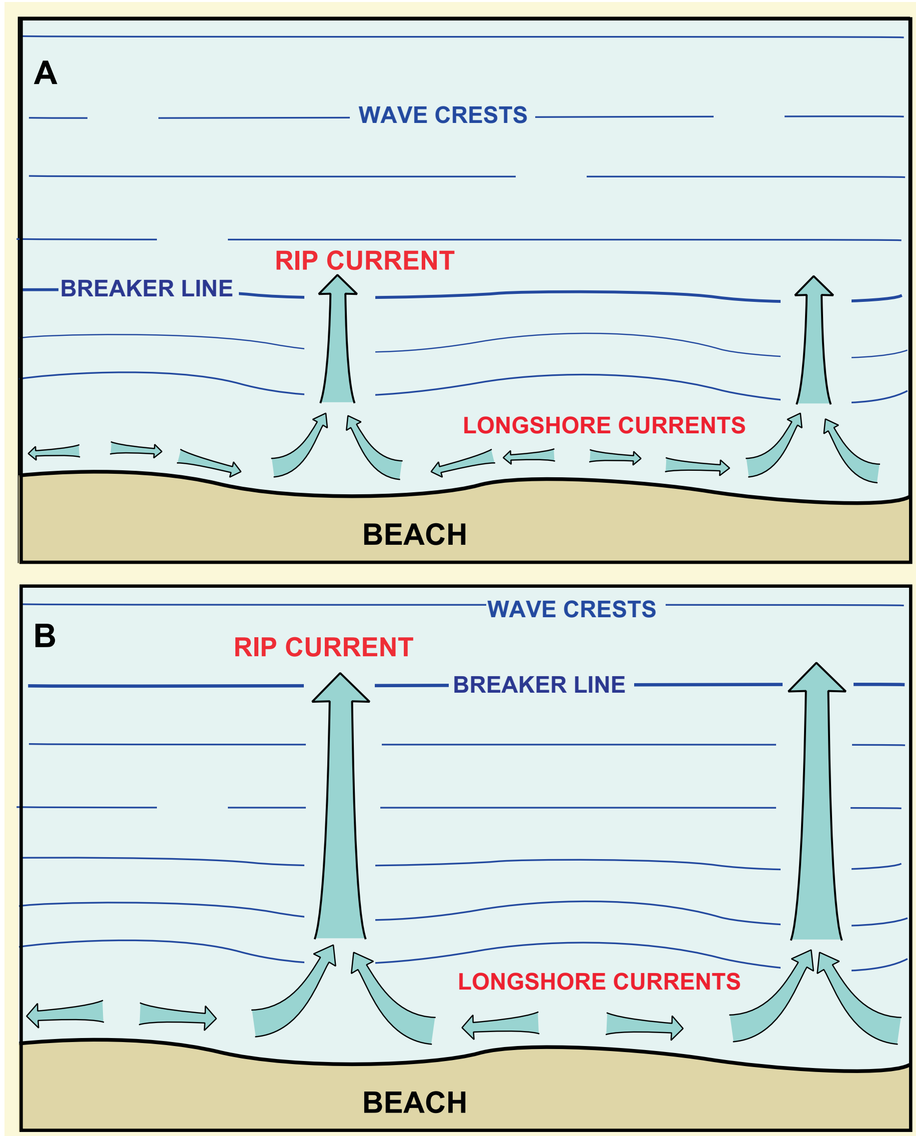

FIG. 42.—Nearshore circulation cells as a function of wave energy. **A)** Small waves break in shallower water than do large waves **B)**. Larger waves also generate a greater degree of setup/setdown and thus have more strongly developed rip and longshore currents. Rip currents under large waves extend farther offshore, probably to the base of the shoreface or slightly beyond.

reflect the finer grain size of the Viking sand or lower wave energy (Posamentier et al., 1992), or possibly a somewhat lower basinal gradient for the Cardium.

Although some sharp-based shoreface deposits unquestionably reflect a forced regression, as for example where the erosional surface is of regional extent (Plint, 1988; Hadley and Elliott, 1993), other explanations exist. Figure 6, for example, shows sharp-based shorefaces that result from incision during a stepped transgression (Walker and Plint, 1992; Bergman and Walker, 1988, 1999). As noted in a previous section, a prograding bar–tough system can create an erosional surface at the base of upper-shoreface deposits (Fig. 38). Examples of this feature occur repeatedly in middle Miocene shoreline deposits in the Caliente Range of California, U.S.A. (Clifton, 1981). Paleocurrent measurements support the conclusion that the sharp-based upper-shoreface deposits here result from progradation of a coast marked by oblique shore-attached bars, rather than from forced regression (Figs. 55, 56). In these deposits, the sharp contact separates coarser, cross-bedded sandstone from subjacent finer and more bioturbated sandstone. The common presence of small pebbles in the sandstone below the contact and occasional intertonguing of the facies suggests that the erosional surface does not represent a major break in facies succession.

Parasequences in the Blackhawk Formation in the Book Cliffs, Utah, U.S.A., also contain sharp-based nearshore sandstones that may be unrelated to forced regression. Two types of contacts occur (Fig. 57). One lies at the base of the upper shoreface deposits, as noted by Howard (1972), where clean, medium-grained cross-bedded sandstone sharply overlies finer bioturbated sandstone of the lower shoreface. As with the Caliente Range example, the break in succession is relatively minor and is probably attributable to a prograding bar–trough system. The

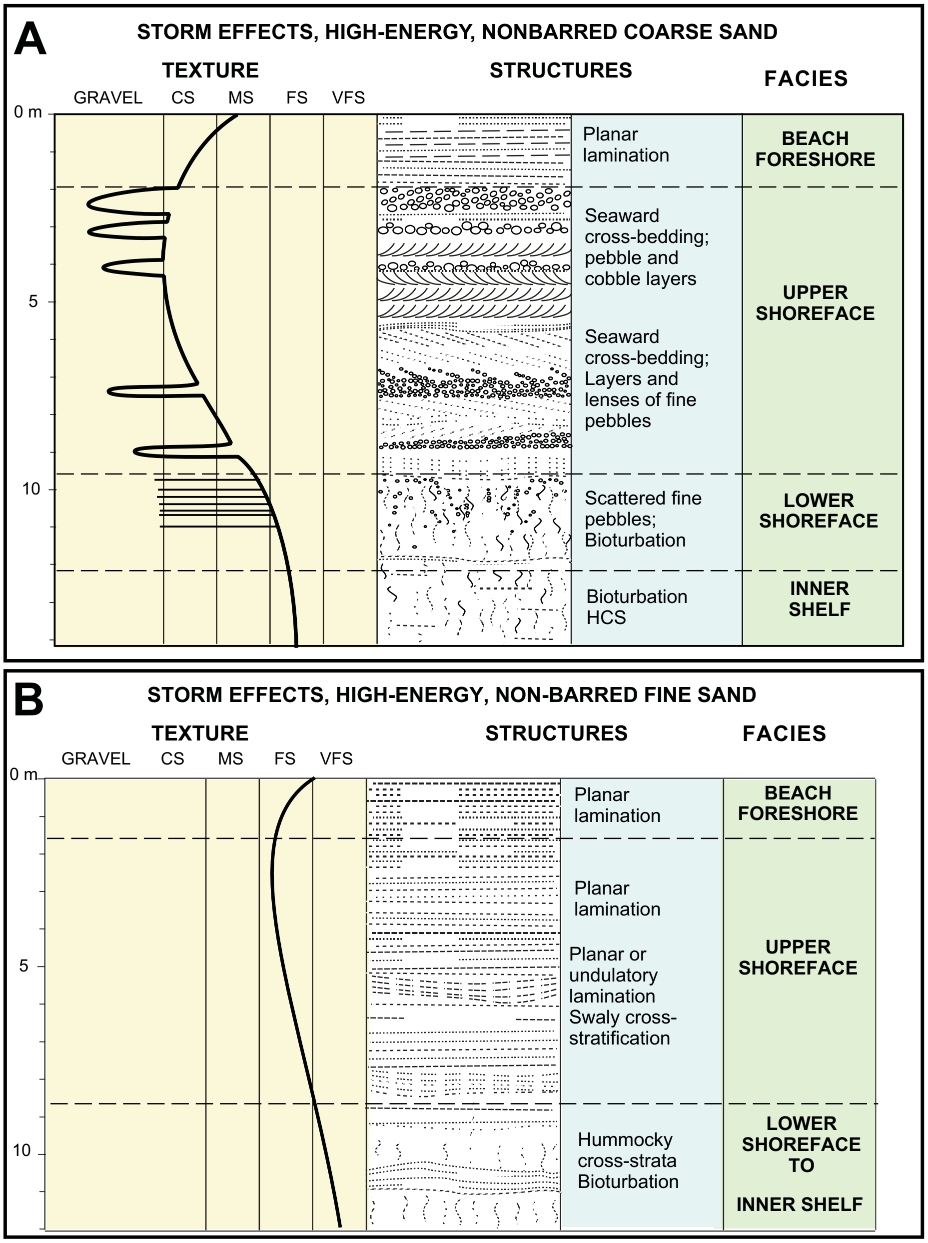

FIG. 43.—**A)** Vertical succession produced by progradation of a high-energy, nonbarred, coarse sandy nearshore, in which storm effects predominate. Scattered pebbles on lower shoreface interpreted as reworked post-storm lags. Scattered fine pebbles in the lower shoreface are interpreted as remnants of post-storm lags. **B)** Vertical sequence produced by progradation of a storm-dominated, nonbarred, high-energy, fine-sandy nearshore. Cross-bedding may result from increased nearshore circulation flow.

second contact lies at the base of amalgamated storm sands. In the section below, muddy intervals separate the storm sets. Although striking, the contact at the base of the amalgamated sandstone is probably comparable to that at the base of each of the subjacent storm sands. The absence of shale in the overlying section, because of either greater storm erosion in the shallower water or inability of mud to accumulate in the more energetic environment, defines the break. Such contacts cannot be traced laterally and are unlikely to represent forced regressions.

CONCLUSIONS

The basic facies model for open-coast clastic deposits is a simple upward-shallowing succession in a sand body bounded to seaward by shelf deposits and to landward by nonmarine facies. The sand body may be linear, as in a barrier, or sheet-like, as in a strand plain. The model exists with minor variations in the standard texts and has received broad application. It is flawed to a degree, in that it is based on a fairly limited set of modern

FIG. 44.—Photograph of transition from upper shoreface to lower shoreface (just below 10 cm scale) in the Plio-Pleistocene Merced Formation, San Francisco, California, U.S.A. Upper shoreface has cross-bedded sand and gravel. Lower shoreface is marked by a couple of meters of fine sand that bears scattered small pebbles in stringers and isolated clusters.

analogs. Many studies of modern coasts have been made in areas undergoing marine transgression, and conclusions drawn from these coasts are skewed in their view of sandbody geometry, shoreface profile, and sand–mud distribution. In addition, studies of modern coasts are biased toward fair-weather conditions

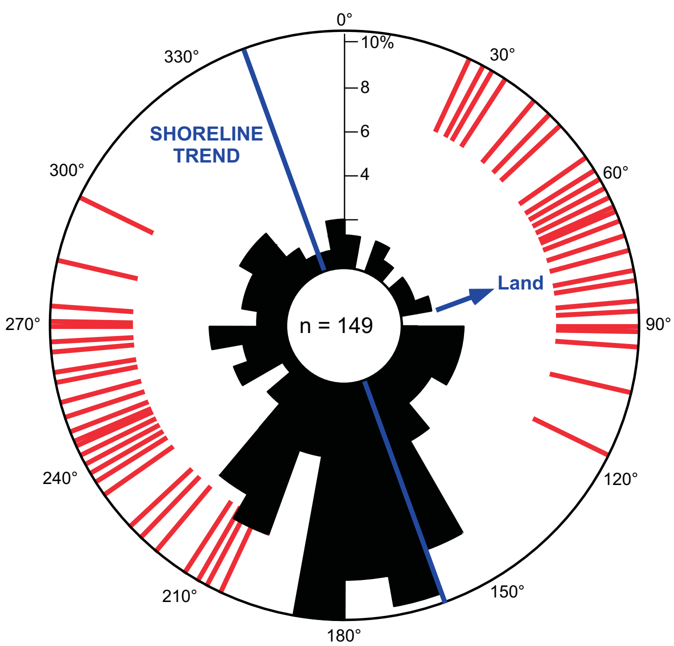

FIG. 45.— Summary of 149 cross-bedding measurements in nearshore facies of the Plio-Pleistocene Merced Formation. Red bars indicate orientation of gravel-filled gutter casts that are approximately normal to the shoreline. Most of the cross-bedding indicates south-flowing longshore currents. Very little cross-bedding is directed landward, indicating that lunate megaripples produced by fair-weather waves are rarely preserved. After Chiocci and Clifton (1991).

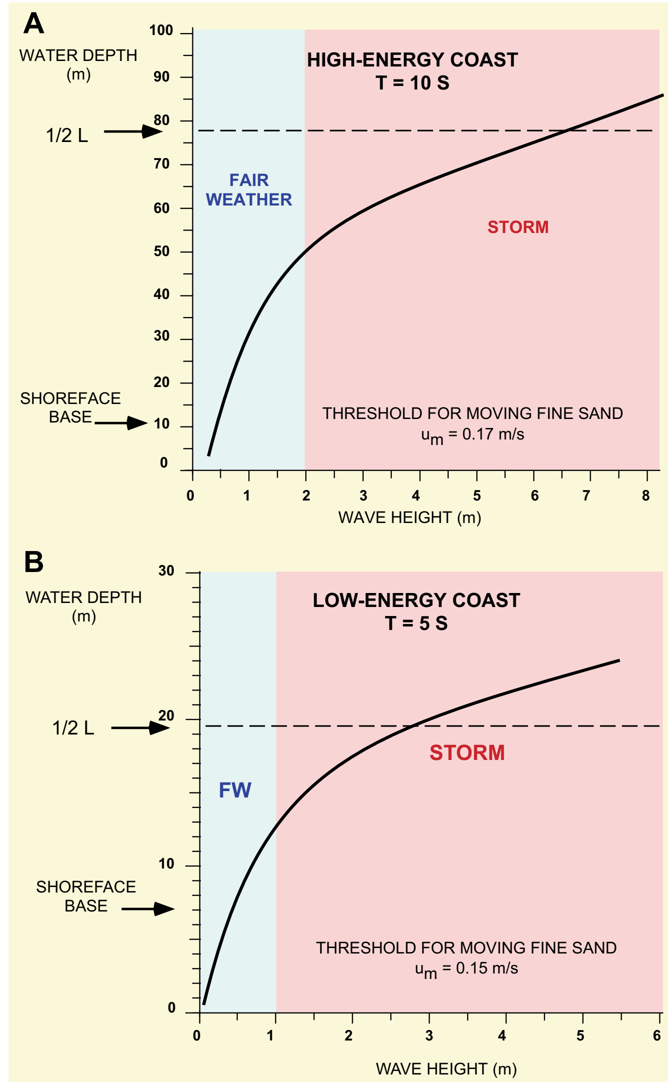

FIG. 46.—Combination of wave heights and water depths in which fine sand (D = 0.125 mm) will be moved by passing waves. **A)** 10-second waves. Combination of wave heights and water depths in which fine sand (D = 0.125 mm) will be moved by passing waves. **B)** 5-second waves. In both cases movement occurs in water depths well seaward of the base of the shoreface on a prograding coast. Water depths equal to one-half the deep water wave length are deeper still.

and settings of low wave energy. Most modern studies have been made on coasts in tectonically passive settings, where fine sand predominates on the beach and shoreface. The influence of texture, as an independent variable, has largely been overlooked.

Many of the limitations on the basic model could be obviated by considering the variations to be end members in a flexible or even multi-dimensional model based on parameters such as texture, sand supply, ambient wave energy, storm influence, coastal morphology, and nature of base-level change (Fig. 58). Using this approach, the basic model as presented in most texts becomes specifically a model for a storm-dominated, moderate-energy to low-energy setting with a moderate gradient in which fine sand was in somewhat limited supply and base level was static (Fig. 59). Coastal successions in Pleistocene deposits on the

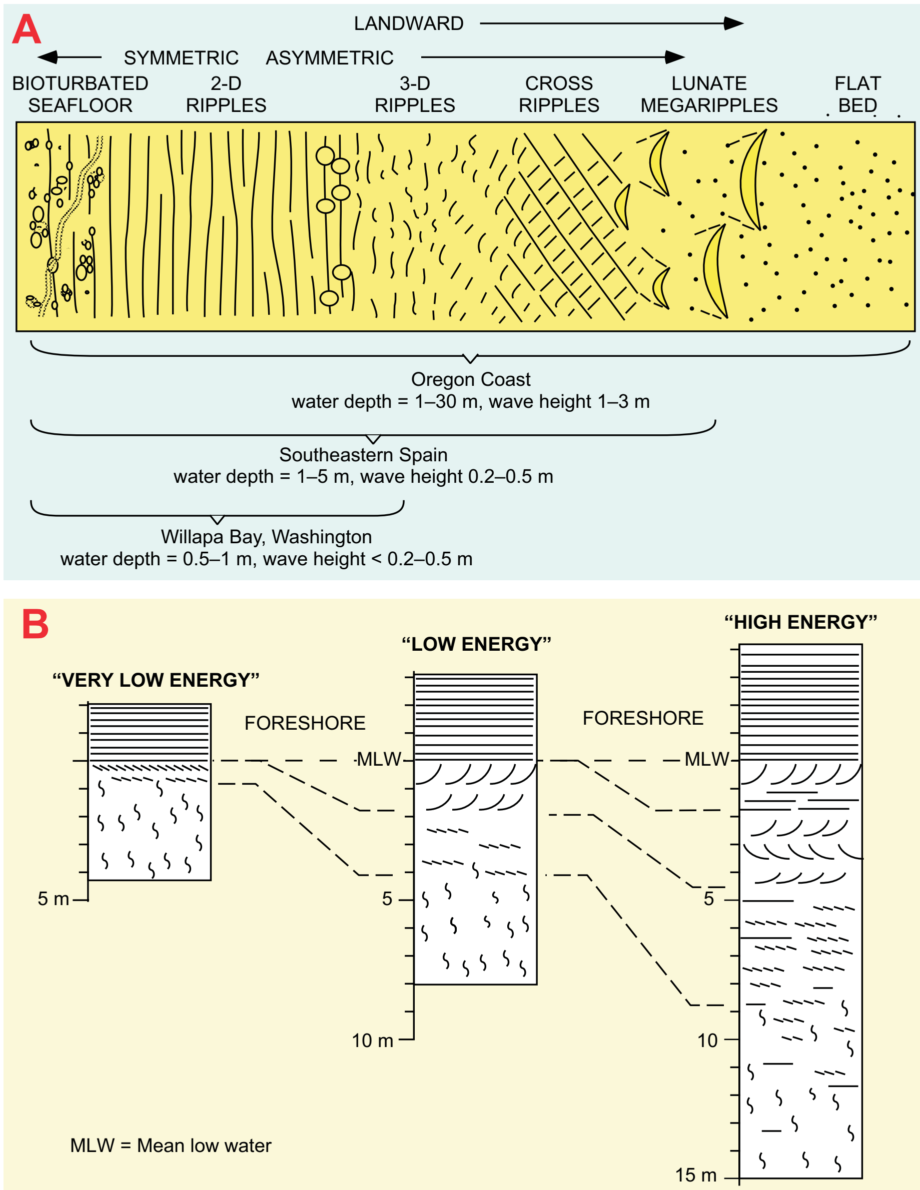

FIG. 47.— **A)** Distribution of sea-floor facies observed under fair-weather conditions on a high-energy coast (southern Oregon), U.S.A, a low-energy coast (southeastern Spain), and a very low-energy coast (Willapa Bay Washington, U.S.A.), and **B)** comparison of the hypothetical beach-to-offshore successions produced by progradation of facies shown in Part A.

California coast (Fig. 60) reflect accumulation in a storm-dominated, high-energy, barred setting with an abundant supply of sand, including coarse sand and gravel, under conditions of static base level. Pleistocene successions on the southeastern coast of Spain (Fig. 61) were deposited in a similar setting, but under conditions of much lower wave energy. Successions like those formed on the Texas Gulf Coast (Fig. 62) represent deposition in a similarly low-energy setting, but one dominated by fine sand, in which fair-weather processes predominate in the preserved deposit. Gilbert-delta deposits (Fig. 63) can be accommodated into the model as accumulating in a steep-gradient setting of low to very low energy. Although the supply of gravel may be substantial, not enough sand enters the system to develop an offshore profile in equilibrium with the waves.

Facies models have been proposed for open-coast sediment in settings other than progradation at constant or slowly rising relative sea level. In particular, models based on "forced regression" have provided an alternative for explaining isolated shallow marine sand bodies with sharp bases. The vertical succession produced thereby fits into the model as forming in a storm-dominated, moderate-energy to low-energy setting with a low gradient in which fine sand was in somewhat limited supply and falling (or fallen) base level (Fig. 64). Not all sharp-based shoreface successions, however, require sea-level change. Prograding bar–trough systems and the simple amalgamation of storm sands in a setting with limited sand supply can also produce erosionally based coastal sandstone deposits (Fig. 65).

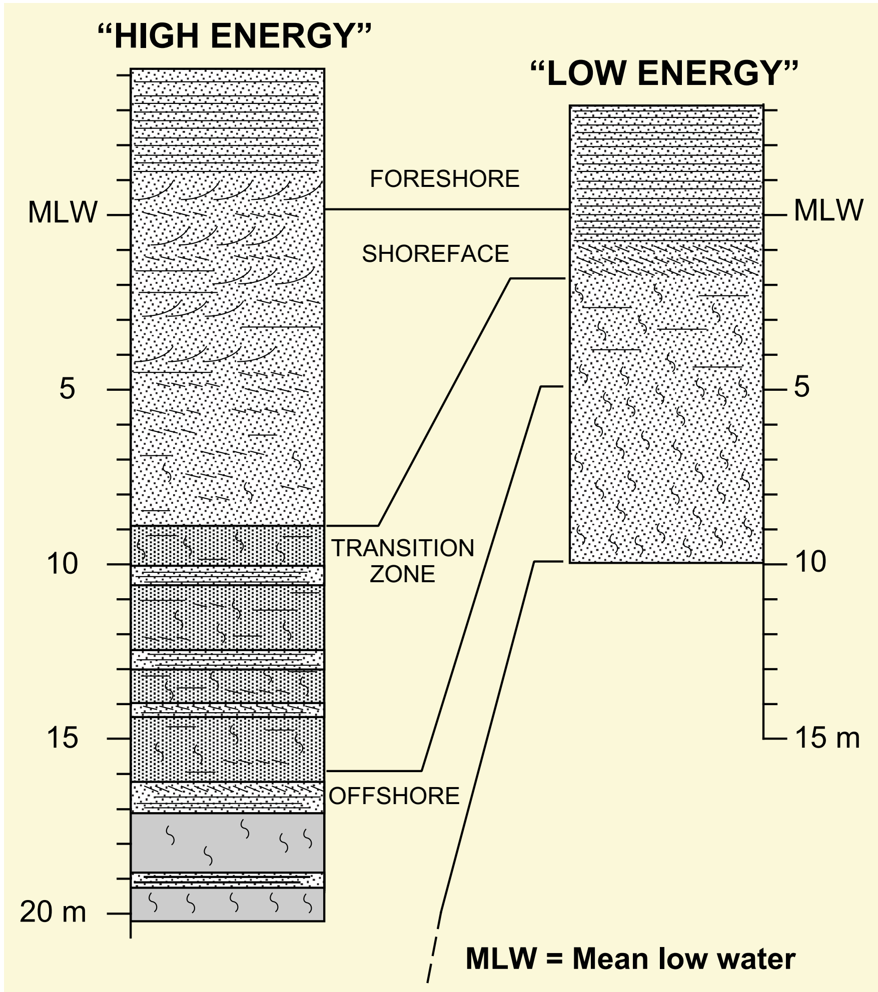

Fig. 48.—A comparison of high- and low-energy beach-to-offshore sequences, Ventura to Port Hueneme, California, and Sapelo Island, Georgia, U.S.A. After Howard and Reineck (1981).

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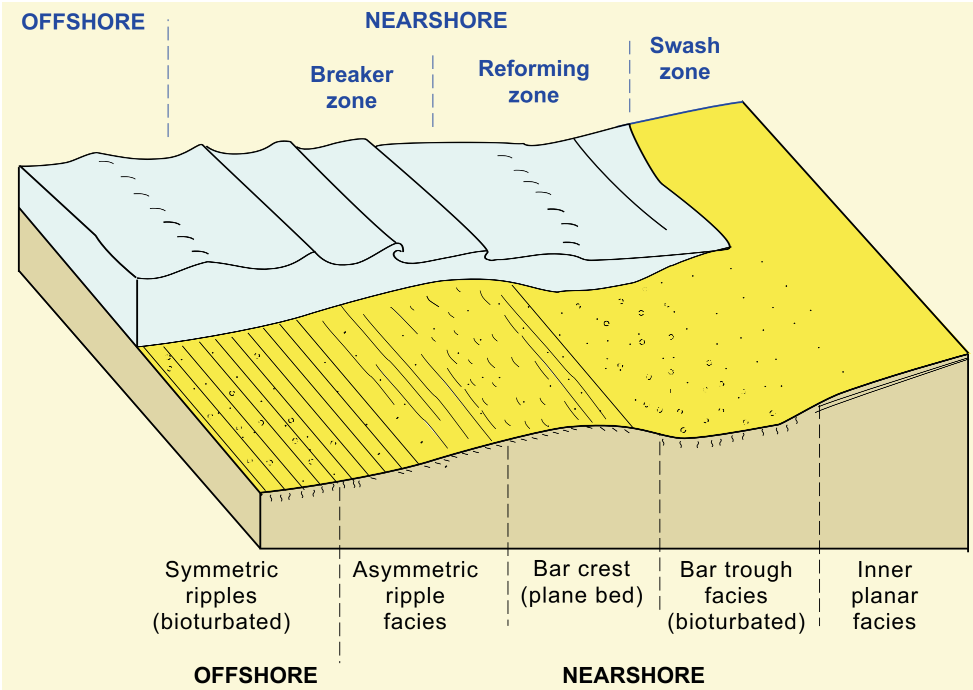

FIG. 49.—Sedimentary structural facies in a barred low-energy fine-grained nearshore under fair-weather conditions (example, Padre Island, Texas, U.S.A., after Hill and Hunter, 1976). Sediment in the longshore trough and seaward from the bar is intensely bioturbated.

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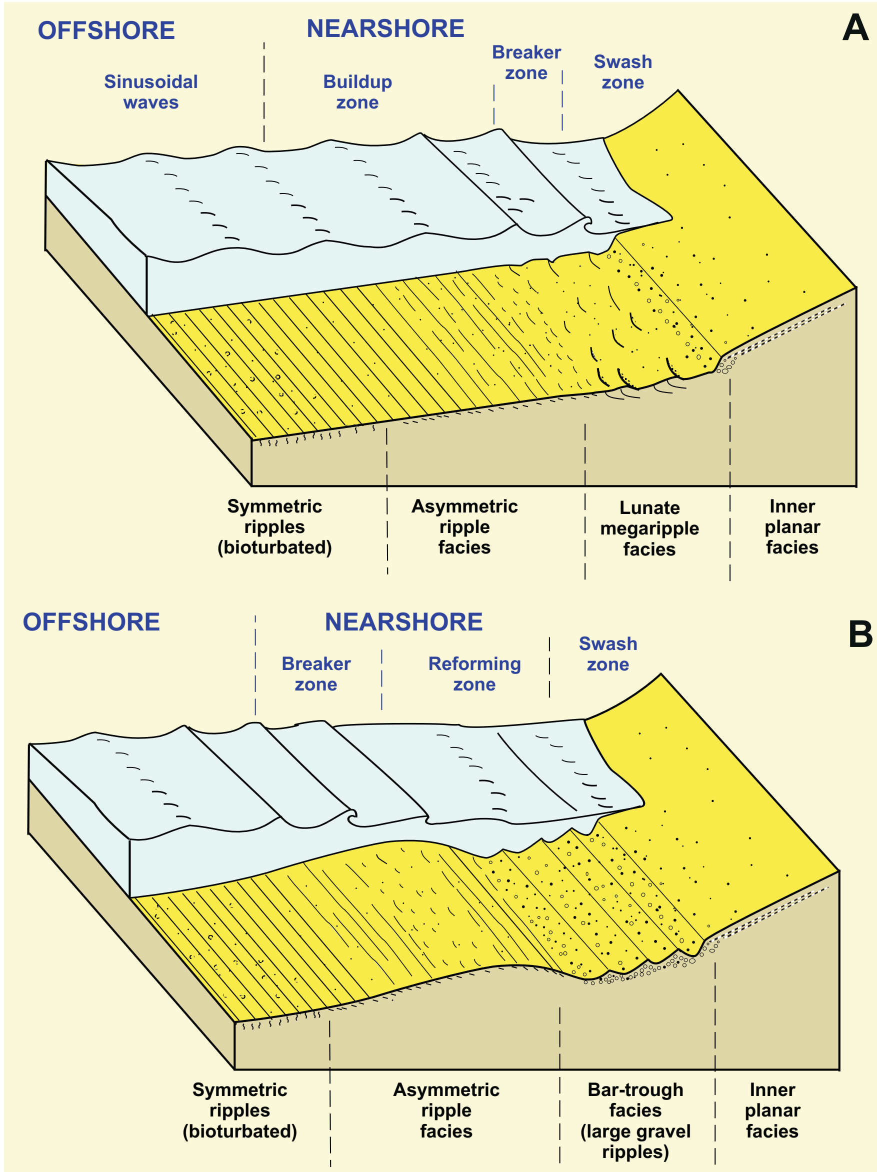

FIG. 50.—Coastal facies on an non-barred and barred nearshores on the southeastern coast of Spain. **A)** Upper-shoreface profile on a non-barred nearshore. Small (0.5 m) waves break directly on the edge of the foreshore. Lunate megaripples occur just seaward of the beach foreshore. **B)** Upper-shoreface profile on a barred nearshore, southeastern coast of Spain. Large 2-D megaripples occur in gravel at base of beach. Bar is composed of fine- to medium-grained sand.

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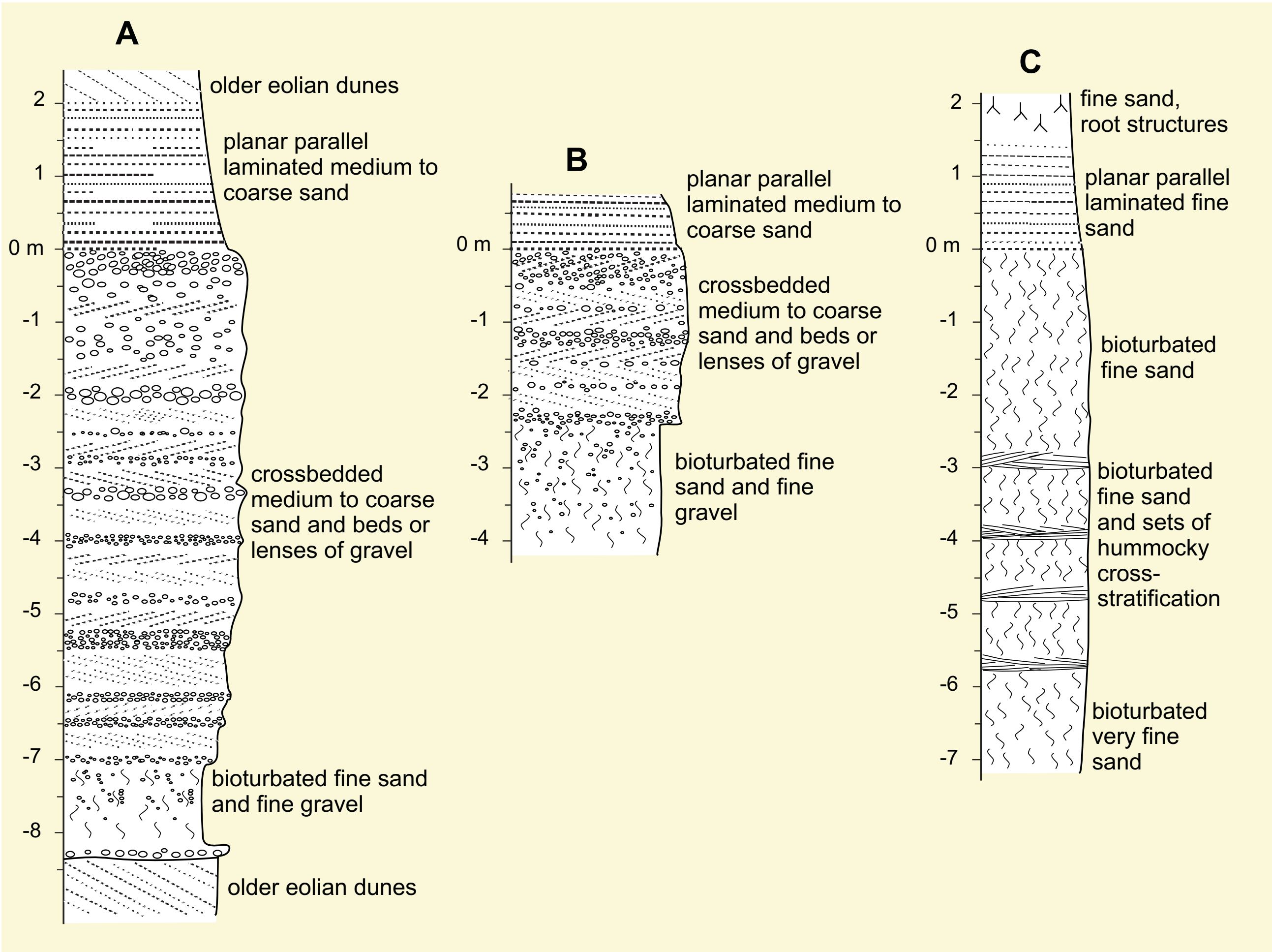

FIG. 51.—Comparison of stratal successions illustrating effects of ambient wave energy and grain size. **A)** Pleistocene terrace deposits, Monterey Bay, California, U.S.A., composed of medium to coarse, pebbly sand deposited in a high-energy setting (fair-weather waves 1–2 m high; tidal range 2–3 m). **B)** Pleistocene terrace deposits, Mediterranean coast, southeastern Spain, composed of coarse, pebbly sand and deposited in a low-energy setting (fair-weather waves 0.2–0.5 m high; tidal range > 1 m). **C)** Eocene Jackson Group, central Texas, U.S.A., composed of fine to very fine sand and deposited in a low-energy setting (fair-weather waves 0.2–0.5 m high; tidal range > 1 m). The Pleistocene deposit in southeastern Spain closely resembles the texturally similar Pleistocene deposit of Monterey Bay, despite the pronounced difference in energy regime. The succession in Spain differs markedly from that in the fine sandy deposits of the Jackson Group, which formed under a very similar energy regime. This comparison illustrates the need to factor in the effect of grain size when interpreting paleo–wave energy. The primary difference between the two coarse-grained deposits is that the low-energy succession is significantly thinner that that formed under high-energy conditions, as a consequence of the deeper wave base of the high-energy California deposit.

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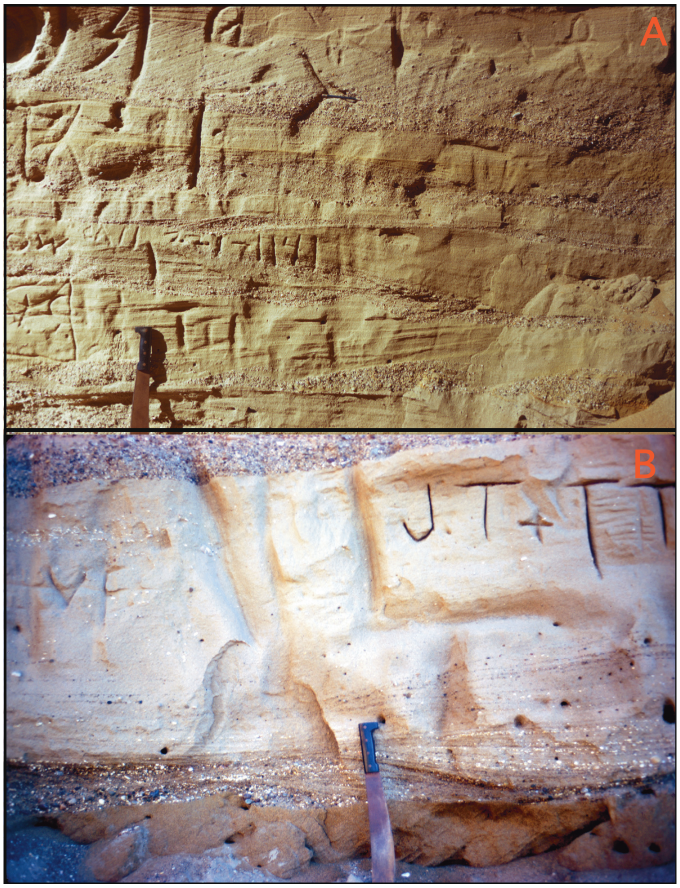

FIG. 52.—Pleistocene terrace deposits, Manresa Beach, California, U.S.A. **A)** Cross-bedded sand and gravel (upper shoreface facies) exposed several meters above the base of a shallowing-up section. **B)** Base of shallowing-up section. Note scattered pebbles in fine sand (lower-shoreface facies) above machete handle. Reddish sand with high-angle foresets at base of photo is older eolian dune sand.

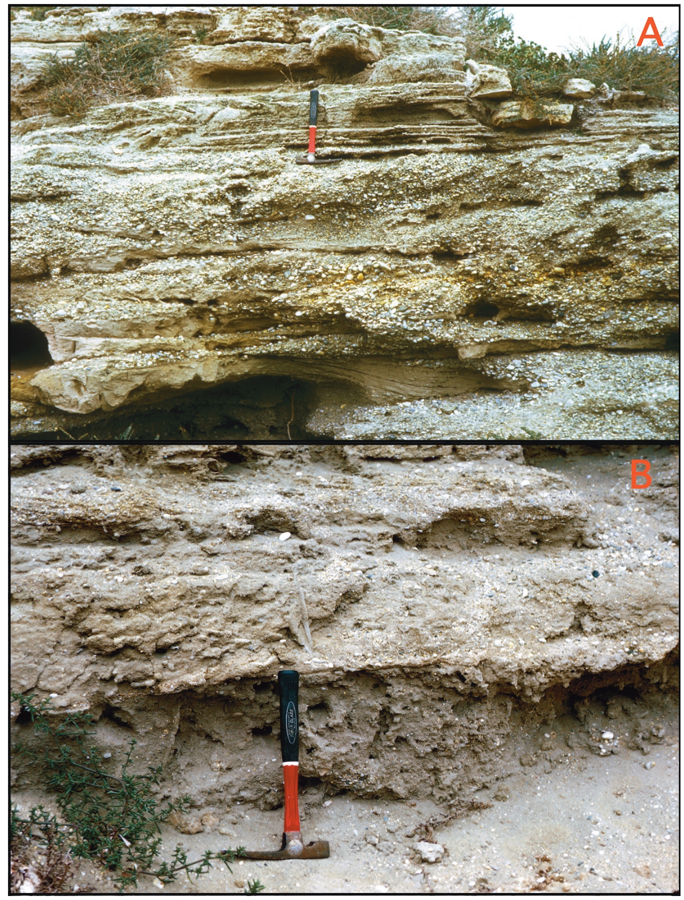

FIG. 53.—Pleistocene terrace deposits, southeastern Spain. **A)** Upper part of a section. Hammer head rests on contact between cross-bedded sand and gravel (inner shoreface) and upward-fining flat-bedded pebbly sandstone (beach foreshore). **B)** Lower part of section. Cross-bedded pebbly sand in upper part of photograph (upper shoreface) overlies burrowed sand with scattered small pebbles and pebble stringers (lower shoreface).

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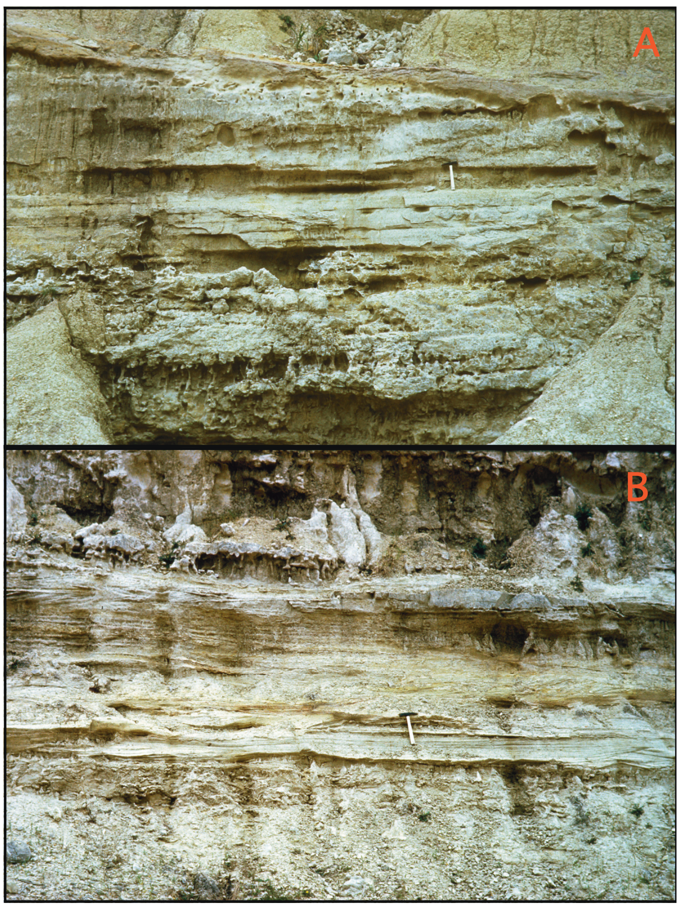

Fig. 54.—A) Exposure of the Jackson Group in a uranium pit, west Texas, U.S.A. **A)** Upper part of section. Hammer head rests on contact between flat-bedded sand (upper foreshore) and gently inclined large-scale foresets with root structures (backshore). Bioturbated sand in lower half of photo represents upper shoreface. **B)** Lower part of section. Hammer is on hummocky cross-stratified set in lower-shoreface or inner-shelf facies. Bioturbated interval in upper part of photo is upper shoreface.

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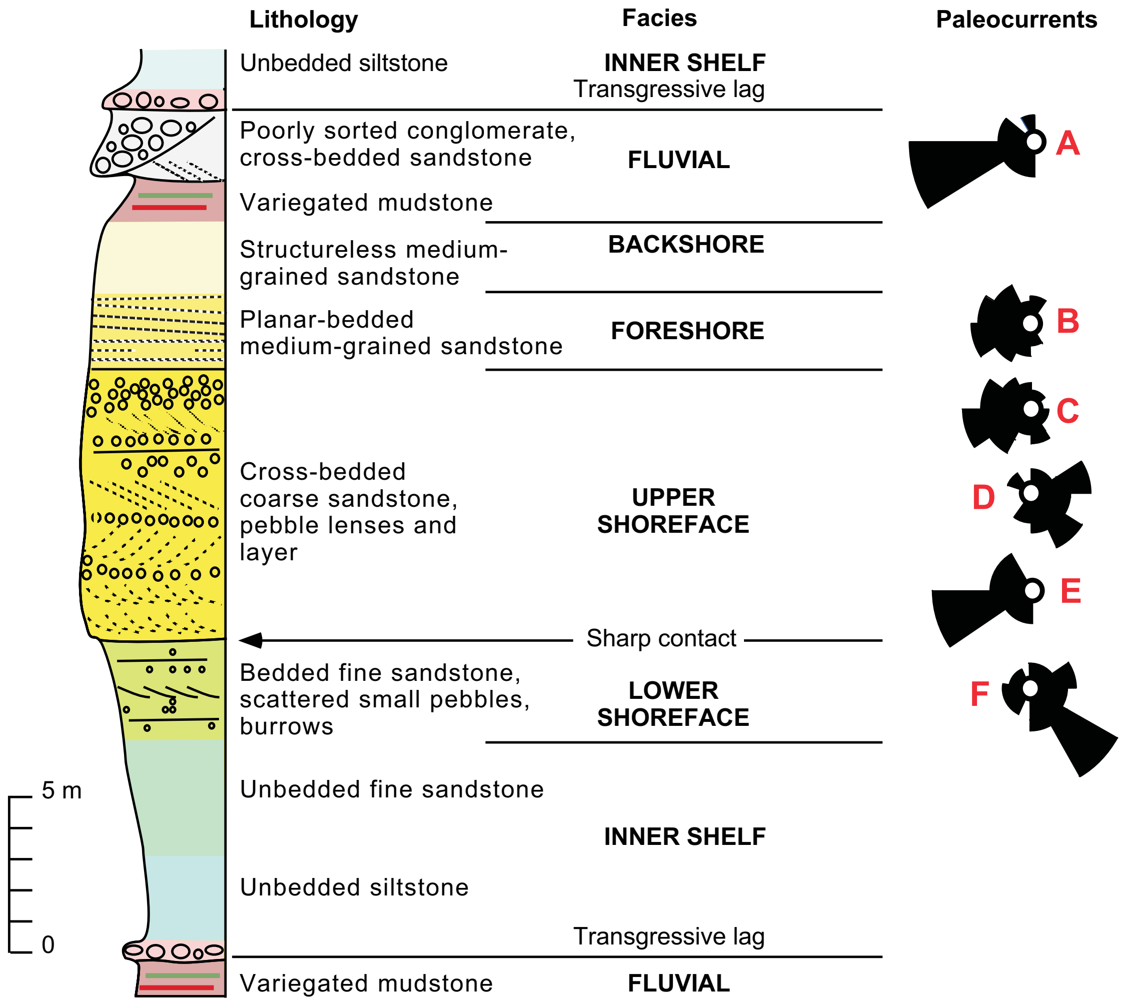

FIG. 55.—Generalized stratigraphic succession through middle Miocene shoreline deposits in the Caliente Range, California, U.S.A. (after Clifton, 1981). Note sharp contact at base of upper-shoreface facies. Paleocurrent roses: **A)** Fluvial sandstone; **B)** planar laminae of the foreshore; **C)** general summation of all (124) cross-bedding measurements within the upper shoreface facies; **D)** large-scale (> 1 m) cross-bedding in the upper-shoreface facies (inferred to represent bars); **E)** cross-bedding in sandstone beds immediately above the basal upper-shoreface contact (inferred to represent rip-channel deposits; **F)** foresets in isolated gravel ripples in the lower-foreshore facies (inferred to represent wave ripples and the direction of wave passage).

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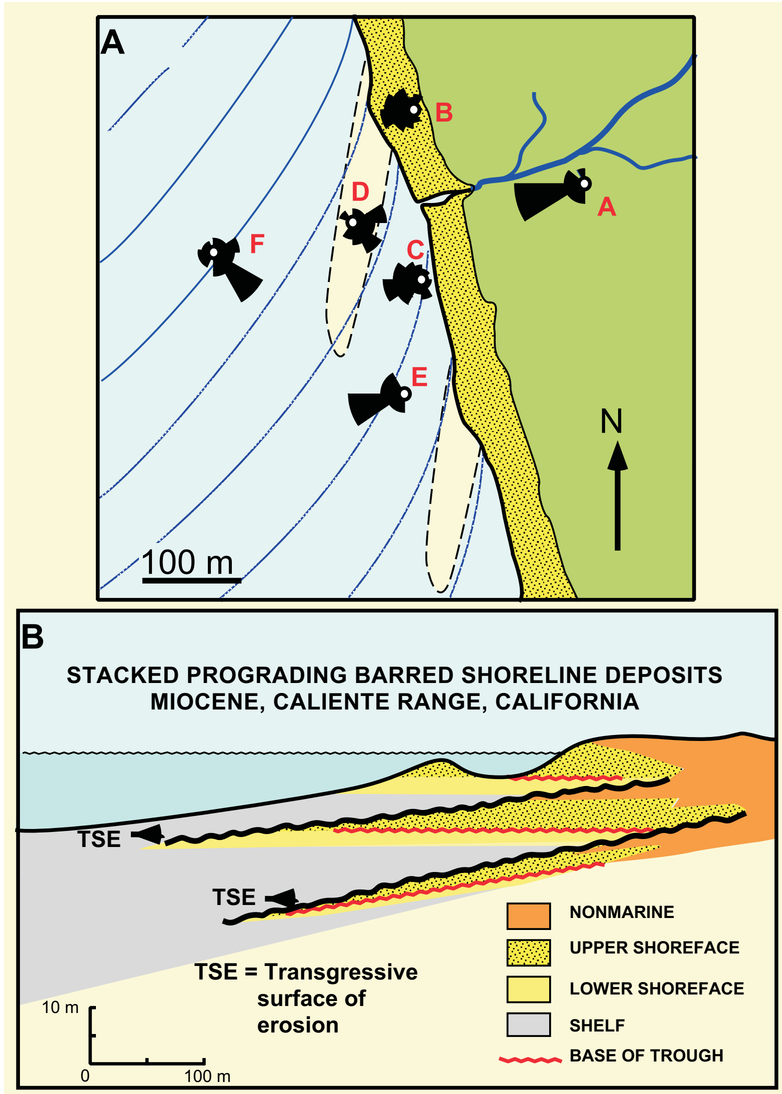

FIG. 56.—**A)** Inferred coastal bar–trough systems responsible for producing the middle Miocene shoreline succession in the Caliente Range, California, U.S.A. Shoreline trend was developed from independent evidence (after Clifton, 1981). **B)** Inferred stacking pattern of the shoreline deposits shown in Part A. Individual parasquences, separated by transgressive surfaces of erosion, contain internal erosional surfaces and sharp-based upper-shoreface deposits generated by prograding bar–trough systems.

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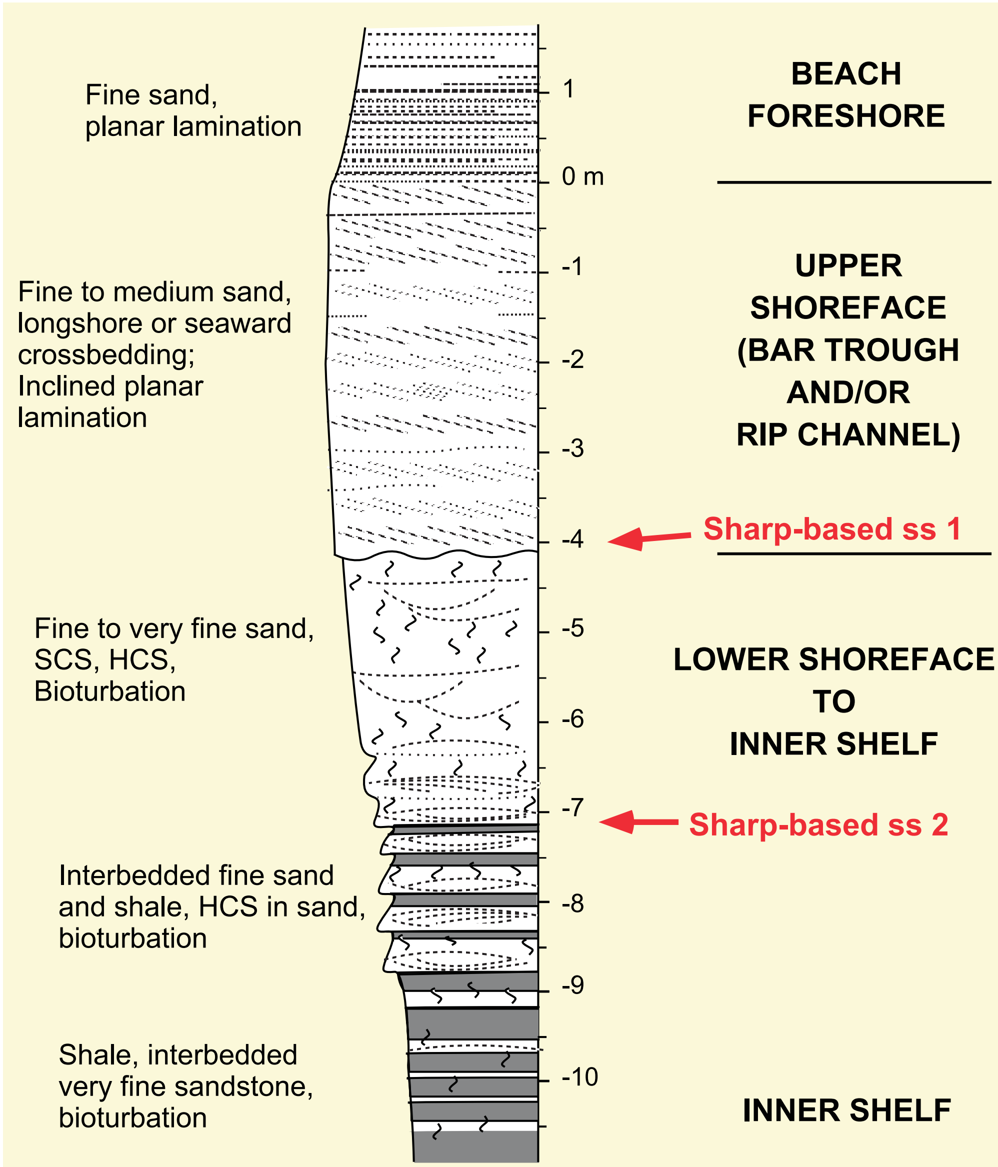

FIG. 57.—Generalized shoreface succession on a prograding fine-grained, non-pebbly, moderate- to low-energy coast (Cretaceous Blackhawk Fm., Book Cliffs, Utah, U.S.A.). Two sharp-based sandstones are shown: #1, produced by prograding bar–trough systems, and #2, produced by the amalgamation of storm sandstones.

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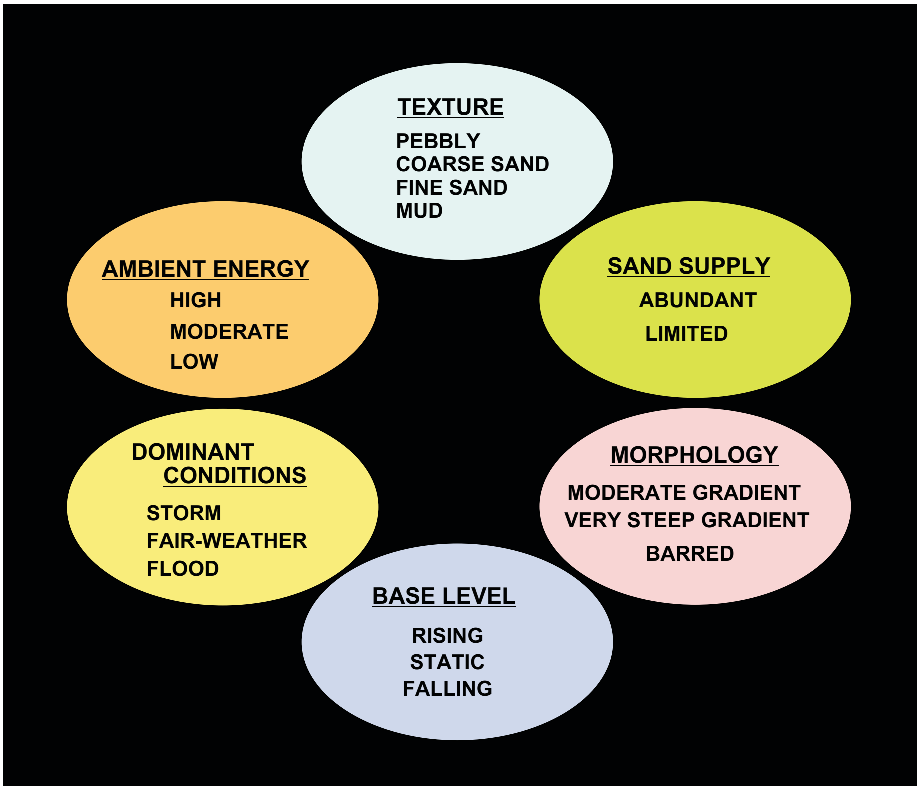

FIG. 58.—Parameters of open-coast settings that can influence the lithologies and stratigraphic succession of the deposit.

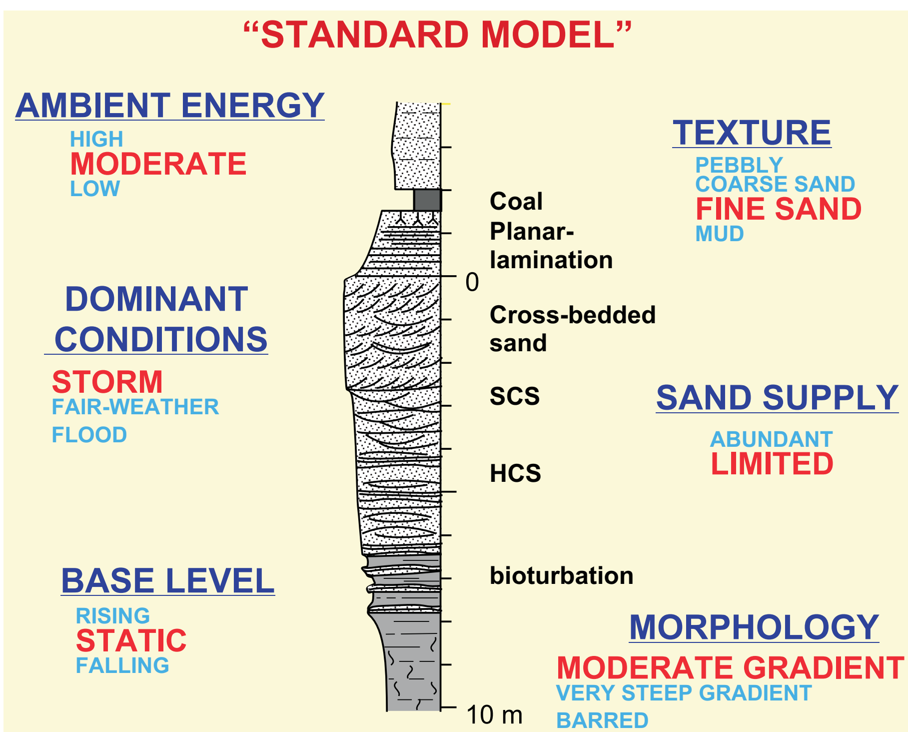

FIG. 59.—"Standard" model for a prograding shoreline (Walker and Plint, 1992). Specific to a storm-dominated, moderate- to low-energy setting with a moderate gradient in which fine sand was in somewhat limited supply and base level was static.

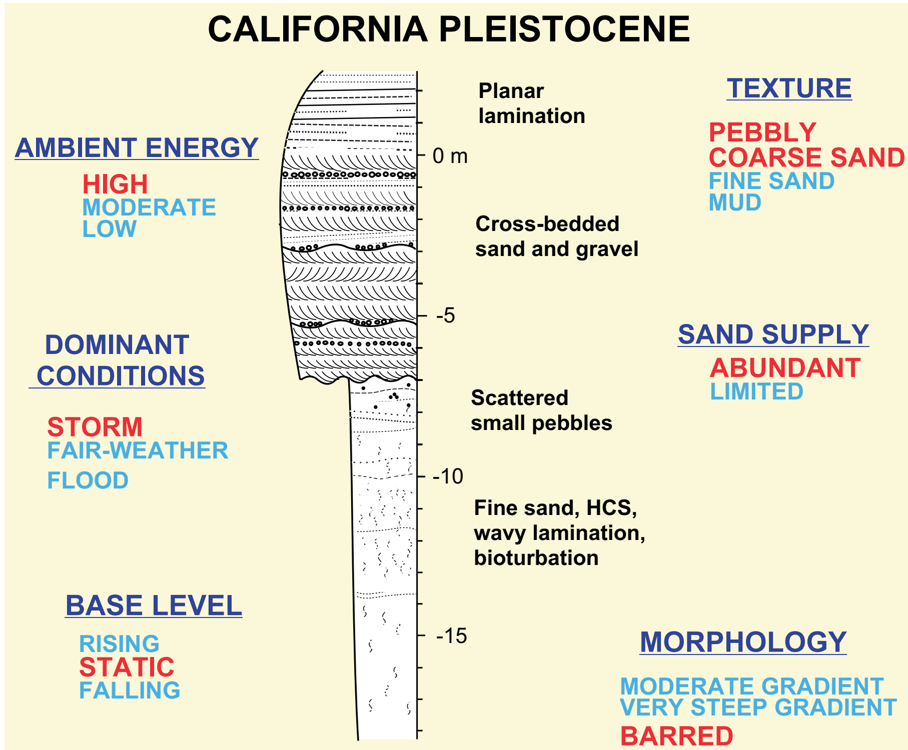

FIG. 60.—Prograding shoreline typical of that found in Pleistocene deposits found along the central California coast, U.S.A. Specific to a storm-dominated, high-energy, barred setting with an abundant supply of sand, including coarse sand and gravel, under conditions of static base level.

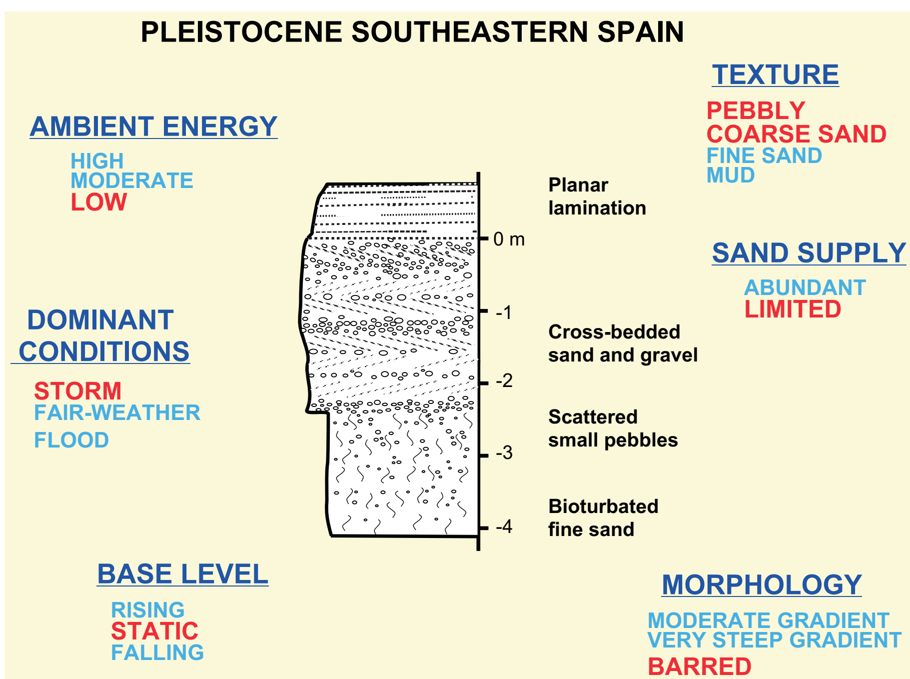

FIG. 61.—Prograding shoreline typical of that found in Pleistocene deposits found along the southeastern coast of Spain. Specific to a storm-dominated, low-energy, barred setting with a limited supply of sand, including coarse sand and gravel, under conditions of static base level.

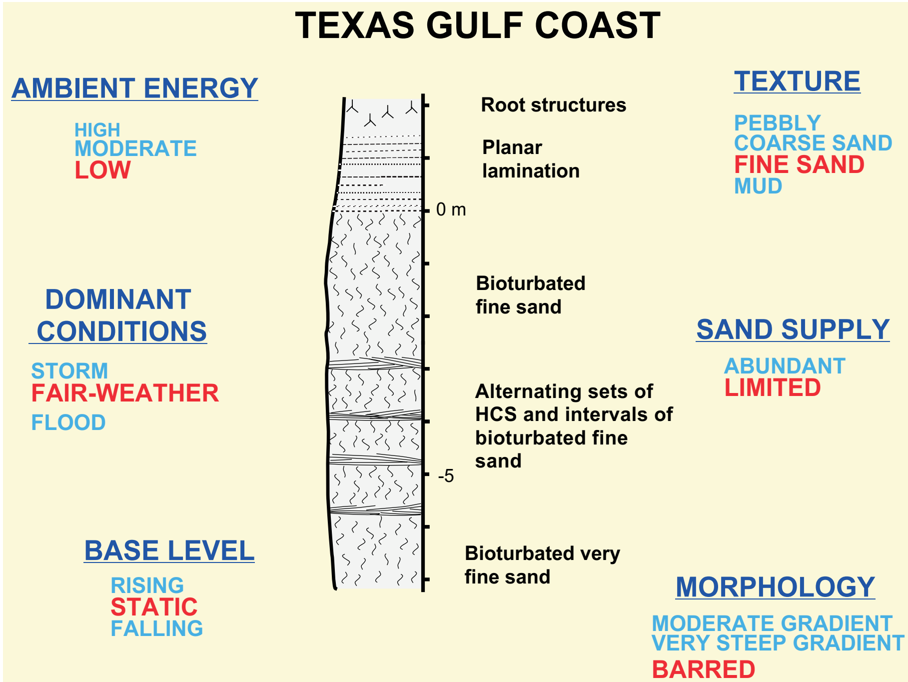

Figure 62. Prograding shoreline typical of that found on the Texas Gulf coast. Specific to a low-energy, barred setting with an limited supply of fine to very fine sand under conditions of static base level. Abundance of bioturbation indicates a dominance of fair-weather conditions, although storm deposits exist in the lower part of the succession.

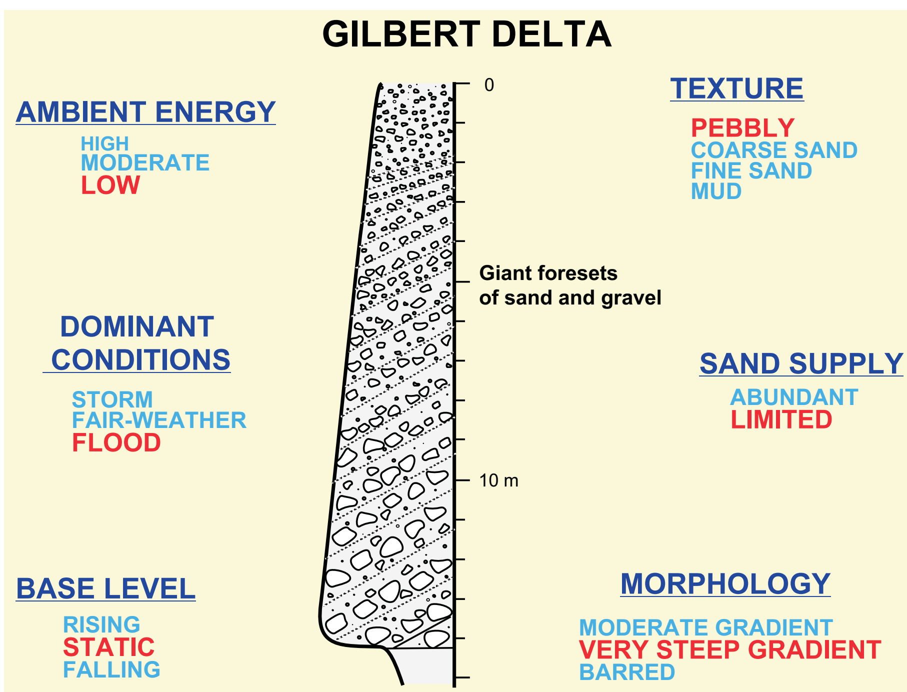

FIG. 63.—Prograding shoreline in a setting where coarse sediment is introduced into a setting of low to very low energy. Although the supply of gravel may be substantial, not enough sand enters the system to develop a offshore profile in equilibrium with the waves. Generally associated with steep gradients and, as shown here, under conditions of static base level.

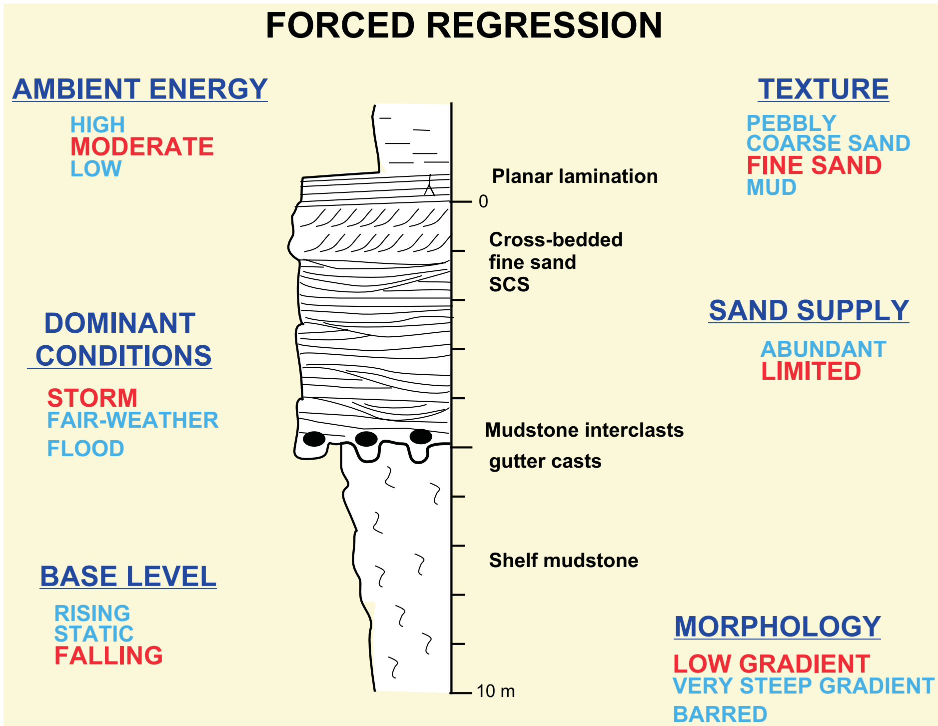

FIG. 64.—Forced-regression model for a prograding shoreline (Plint, 1988). Specific to a storm-dominated, moderate- to low-energy setting with a low gradient in which fine sand was in somewhat limited supply and base level was falling (or had fallen).

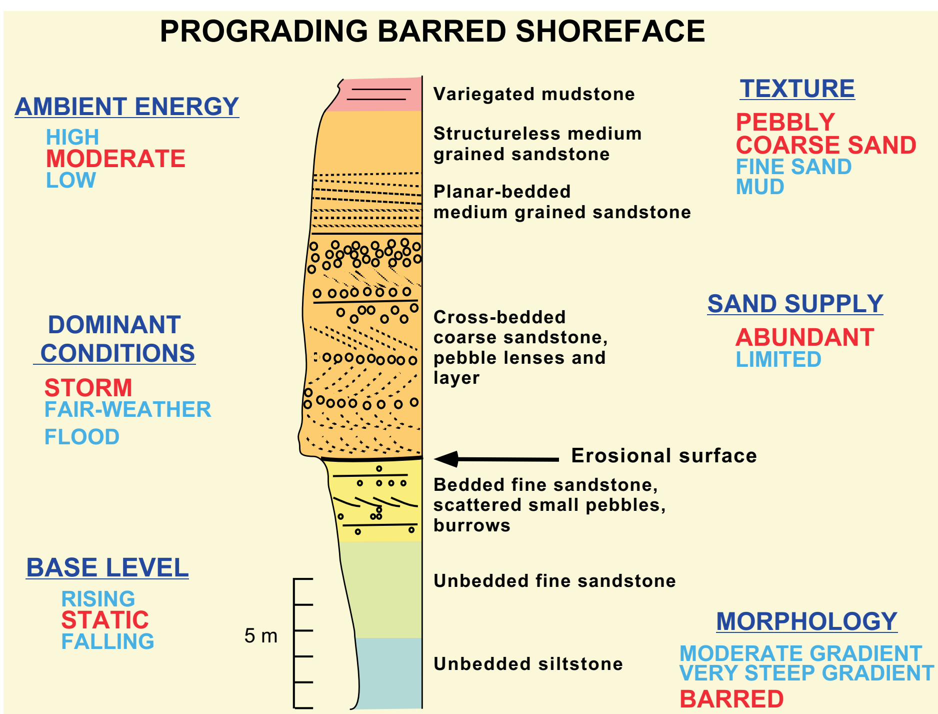

FIG. 65.—Prograding shoreline succession like that found in middle Miocene deposits in the Caliente Range, California, U.S.A. Specific to a storm-dominated, moderate-energy, barred setting with an abundant supply of sand, including coarse sand and gravel, under conditions of static sea level.