Chapter 4
Coastal Morphodynamics

4-1. Introduction

a. This chapter discusses the morphodynamics of four coastal environments: deltas, inlets, sandy shores, and cohesive shores. The divisions are somewhat arbitrary because, in many circumstances, the environments are found together in a limited area. This occurs, for example, within a major river delta like the Mississippi, where a researcher will encounter sandy beaches, bays where cohesive sediments accumulate, and inlets which channel water in and out of the bays.

b. Coastal features and environments are also not isolated in time. For example, as discussed in Chapter 3, estuaries, deltas, and beach ridge shores are elements of a landform continuum that extends over time. Which particular environment or shore type is found at any one time depends on sea level rise, sediment supply, wave and tide energy, underlying geology, climate, rainfall, runoff, and biological productivity.

c. Based on the fact that physical conditions along the coast are constantly changing, it can be argued that there is no such thing as an "equilibrium" state for any coastal form. This is true not only for shoreface profiles but also for deltas, which continue to shift over time in response to varying wave and meteorologic conditions. In addition, man continues to profoundly influence the coastal environment throughout the world, changing natural patterns of runoff and littoral sediment supply and constantly rebuilding and modifying engineering works. This is true even along undeveloped coastlines because of environmental damage such as deforestation, which causes drastic erosion and increased sediment load in rivers. The reader is urged to remember that coastal landforms are the result of the interactions of a myriad of physical processes, man-made influences, global tectonics, local underlying geology, and biology.

4-2. Introduction to Bed Forms

a. Introduction. When sediment is moved by flowing water, the individual grains are usually organized into morphological elements called *bed forms.* These occur in a baffling variety of shapes and scales. Some bed forms are stable only between certain values of flow strength. Often, small bed forms (ripples) are found superimposed on larger forms (dunes), suggesting that the flow field may vary dramatically over time. Bed forms may move in the same direction as the current flow, may move against the current (antidunes), or may not move at all except under specific circumstances. The study of bed form shape and size is of great value because it can assist in making quantitative estimates of the strength of currents in modern and ancient sediments (Harms 1969; Jopling 1966). Bed form orientations are indicators of flow pathways. This introduction to a complex subject is by necessity greatly condensed. For details on interpretation of surface structures and sediment laminae, readers are referred to textbooks on sedimentology such as Allen (1968, 1984, 1985); Komar (1976); Leeder (1982); Lewis (1984); Middleton (1965); Middleton and Southard (1984); and Reineck and Singh (1980).

b. Environments. In nature, bed forms are found in three environments of greatly differing characteristics:

- Rivers - unidirectional and channelized; large variety of grain sizes.
- Sandy coastal bays - semi-channelized, unsteady, reversing (tidal) flows.
- Continental shelves - deep, unchannelized; dominated by geostrophic flows, storms, tidal currents, wave-generated currents.

c. Classification. Because of the diverse natural settings and the differing disciplines of researchers who have studied sedimentology, the classification and nomenclature of bed forms have been confusing and contradictory. The following classification scheme, proposed by the Society for Sedimentary Geology (SEPM) Bed forms and Bedding Structures Research Group in 1987 (Ashley 1990) is suitable for all subaqueous bed forms:

d. Ripples. These are small bed forms with crest-to-crest spacing less than about 0.6 m and height less than about 0.03 m. It is generally agreed that ripples occur as assemblages of individuals similar in shape and scale. On the basis of crestline trace, Allen (1968) distinguished five basic patterns of ripples: straight, sinuous, catenary, linguoid, and lunate (Figure 4-1). The straight and sinuous forms may be symmetrical in cross section if subject to primarily oscillatory motion (waves) or may be asymmetrical if influenced by unidirectional flow (rivers or tidal currents). Ripples form a population distinct from larger-scale dunes, although the two forms share a similar geometry. The division between the two populations is caused by the interaction of ripple morphology and bed, and may be shear stress. At low shear stresses, ripples are formed. As shear stress increases above a certain threshold a

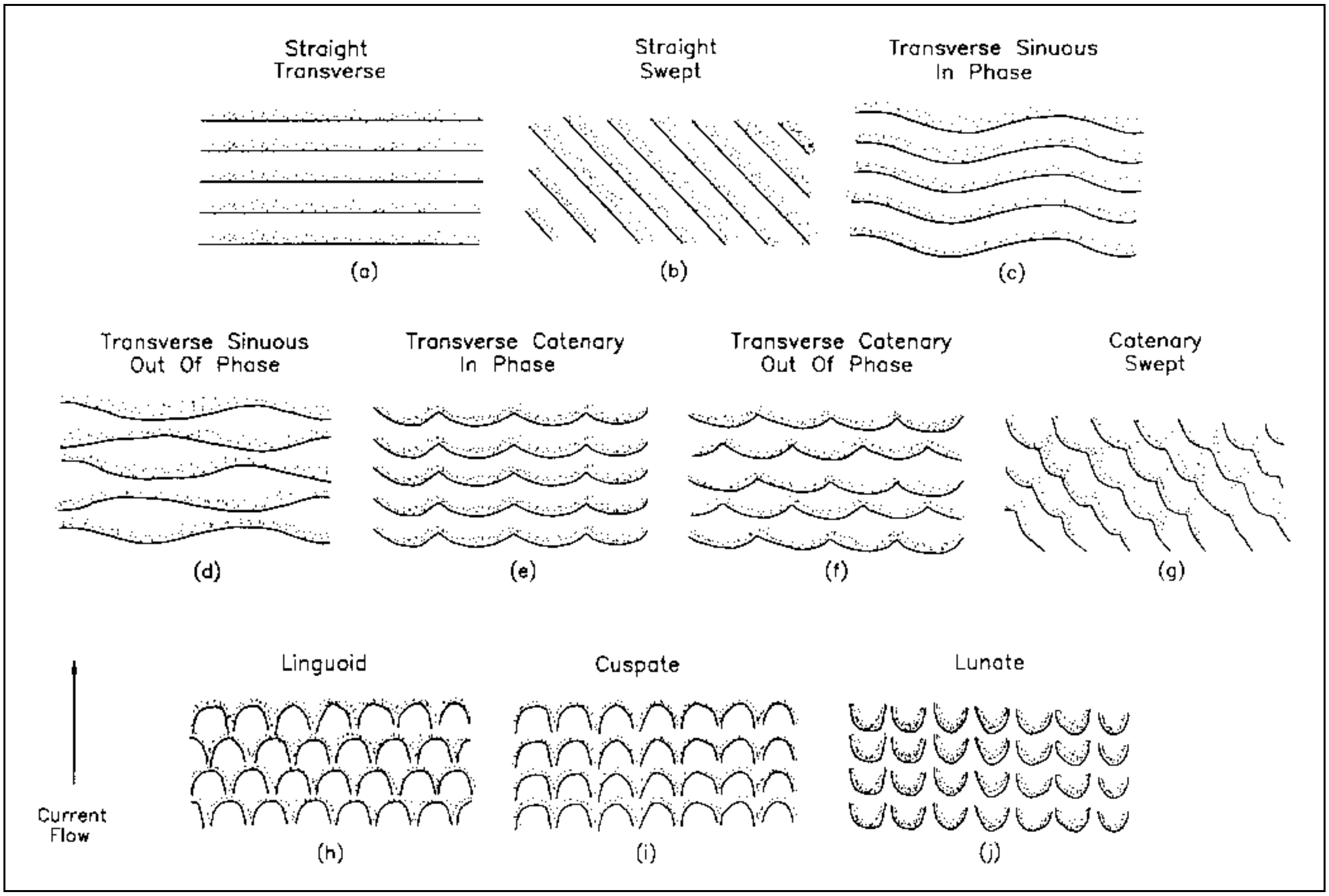

Figure 4-1. Sediment ripples. Water flow is from bottom to top, and lee sides and spurs are stippled (modified from Allen (1968))

"jump" in behavior occurs, resulting in the appearance of the larger dunes (Allen 1968).

e. Dunes. Dunes are flow-transverse bed forms with spacings from under 1 m to over 1,000 m that develop on a sediment bed under unidirectional currents. These large bed forms are ubiquitous in sandy environments where water depths are greater than about 1 m, sand size coarser than 0.15 mm (very fine sand), and current velocities greater than about 0.4 m/sec. In nature, these flow-transverse forms exist as a continuum of sizes without natural breaks or groupings (Ashley 1990). For this reason, "dune" replaces terms such as megaripple or sand wave, which were defined on the basis of arbitrary or perceived size distributions. For descriptive purposes, dunes can be subdivided as small (0.6 - 5 m wavelength), medium (5 - 10 m), large (10 - 100 m), and very large (> 100 m). In addition, the variation in pattern across the flow must be specified. If the flow pattern is relatively unchanged perpendicular to its overall direction and there are no eddies or vortices, the resulting bed form will be straight crested and can be termed two-dimensional (Figure 4-2a). If the flow structure varies significantly across the predominant direction and vortices capable of scouring the bed are present, a three-dimensional bed form is produced (Figure 4-2b).

f. Plane beds. A plane bed is a horizontal bed without elevations or depressions larger than the maximum size of the exposed sediment. The resistance to flow is small, resulting from grain roughness, which is a function of grain size. Plane beds occur under two hydraulic conditions:

- The transition zone between the region of no movement and the initiation of dunes (Figure 4-2).

- The transition zone between ripples and antidunes, at mean flow velocities between about 1 and 2 m/sec (Figure 4-2).

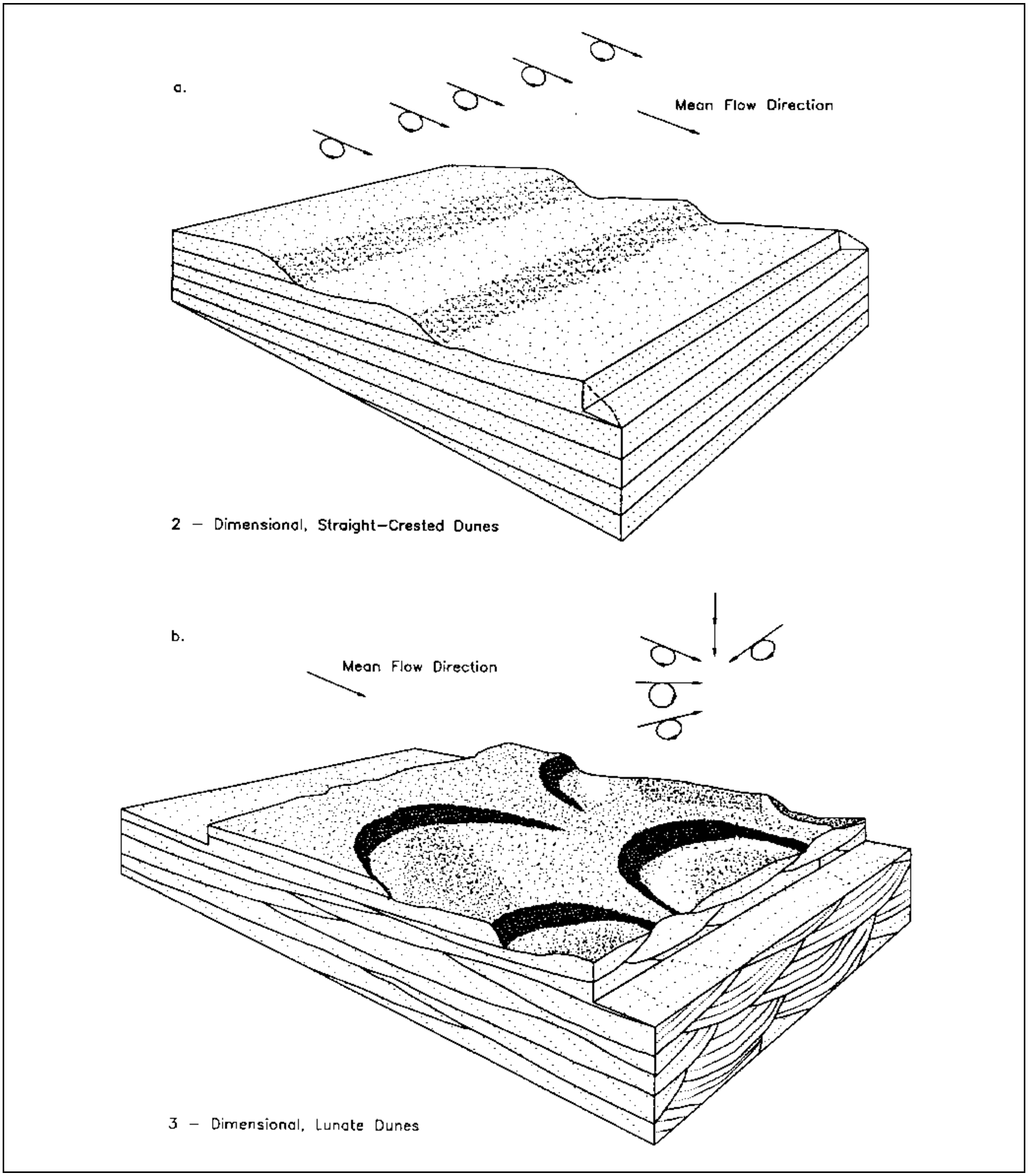

Figure 4-2. Two-dimensional and three-dimensional dunes. Vortices and flow patterns are shown by arrows above dunes. Adapted from Reineck and Singh (1980)

g. Antidunes. Antidunes are bed forms that are in phase with water surface gravity waves. Height and wavelength of these waves depend on the scale of the system and characteristics of the fluid and bed material (Reineck and Singh 1980). Trains of antidunes gradually build up from a plane bed as water velocity increases. As the antidunes increase in size, the water surface changes from planar to wave-like. The water waves may grow until they are unstable and break. As the sediment antidunes grow, they may migrate upstream or downstream, or may remain stationary (the name "antidune" is based on early observations of upstream migration).

h. Velocity - grain size relationships. Figure 4-3, from Ashley (1990) illustrates the zones where ripples, dunes, planar beds, and antidunes are found. The figure summarizes laboratory studies conducted by various researchers. These experiments appear to support the common belief that large flow-traverse bedforms (dunes) are a distinct entity separate from smaller current ripples. This plot is very similar to Figure 11.4 in Graf's (1984) hydraulics text, although Graf uses different axis units.

4-3. Deltaic Processes*

River deltas, which are found throughout the world, result from the interaction of fluvial and marine (or lacustrine) forces. According to Wright (1985), "deltas are defined more broadly as coastal accumulations, both subaqueous and subaerial, of river-derived sediments adjacent to, or in close proximity to, the source stream, including the deposits that have been secondarily molded by waves, currents, or tides." The processes that control delta development vary greatly in relative intensity around the world. As a result, delta-plain landforms span the spectrum of coastal features and include distributary channels, river-mouth bars, interdistributary bays, tidal flats, tidal ridges, beaches, beach ridges, dunes and dune fields, and swamps and marshes. Despite the pronounced variety of worldwide environments where deltas are found, all actively forming deltas have at least one common attribute: a river supplies clastic sediments to the coast and inner shelf more rapidly than marine processes can remove these materials. Whether a river is sufficiently large to transport enough sediment to overcome erosive marine processes depends upon the climate, geology, and nature of the drainage basin, and, most important, the overall size of the basin. The following paragraphs discuss delta classification, riverine flow, sediment deposition, and geomorphic structures associated with deltas.

* Material in this section adapted from Wright (1985).

a. General delta classification. Coleman and Wright (1975) identified six broad classes of deltas using an energy criteria. These models have been plotted on Figure 4-4 according to the relative importance of river, wave, and tide processes. However, Wright (1985) acknowledged that because each delta has unique and distinct features, no classification scheme can adequately encompass the wide variety of environments and structures found at deltas around the world.

b. Delta-forming processes.

(1) Force balance. Every delta is the result of a balance of forces that interact in the vicinity of the river mouth. A river carries sediment to the coast and deposits it beyond the mouth. Tidal currents and waves rework the newly deposited sediments, affecting the shape and form of the resulting structure. The long-term evolution of a delta plain becomes a function of the rate of riverine sediment input and the rate and pattern of sediment reworking, transport, and deposition by marine processes after the initial deposition. On a large scale, gross deltaic shape is also influenced by receiving basin geometry, regional tectonic stability, rates of subsidence caused by compaction of newly deposited sediment, and rate of sea level rise.

(2) River-dominant deltas.

(a) River-dominant deltas are found where rivers carry so much sediment to the coast that the deposition rate overwhelms the rate of reworking and removal due to local marine forces. In regions where wave energy is very low, even low-sediment-load rivers can form substantial deltas.

(b) When a river is completely dominant over marine forces, the delta shape develops as a pattern of prograding, branching distributary channels (resembling fingers branching from a hand). Interdistributary features include open bays and marshes. A generalized isopach map for this type of delta (Type I in Coleman and Wright's (1975) classification) is shown in Figure 4-5. A prime example is the Mississippi River, which not only transports an enormous amount of sediment, but also empties into the low wave-energy, low tide-range Gulf of Mexico. The Mississippi is discussed in detail later in this section.

(3) Wave-dominant deltas.

(a) At wave-dominant deltas, waves sort and redistribute sediments delivered to the coast by rivers and remold

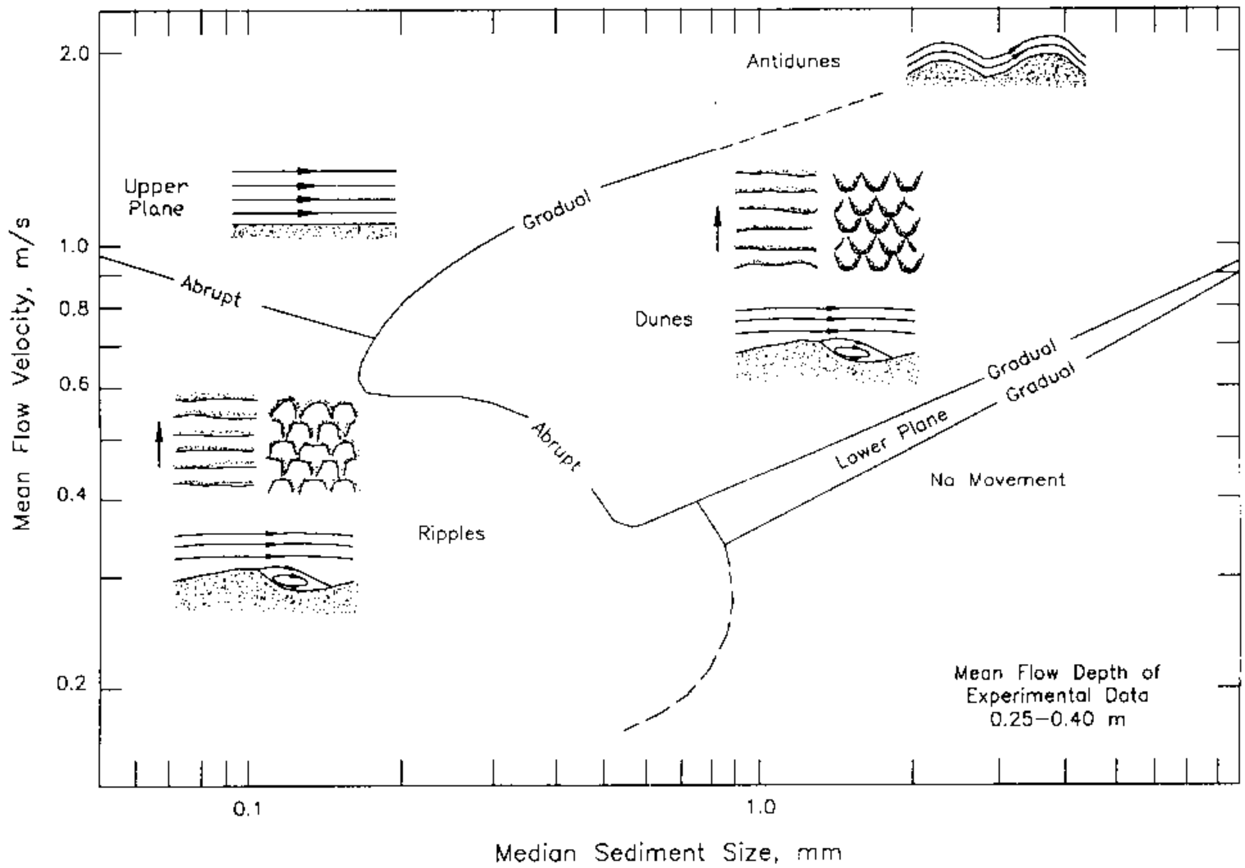

Figure 4-3. Plot of mean flow velocity against mean grain size, based on laboratory studies, showing stability phases of subaqueous bed forms (modified from Ashley (1990)). Original data from various sources, standardized to 10 °C water temperature (original data points not shown)

them into shoreline features such as beaches, barriers, and spits. The morphology of the resulting delta reflects the balance between sediment supply and the rate of wave reworking and redistribution. Wright and Coleman (1972; 1973) found that deltas in regions of the highest nearshore wave energy flux developed the straightest shoreline and best-developed interdistributary beaches and beach-ridge complexes.

(b) Of 16 deltas compared by Wright and Coleman (1972; 1973), the Mississippi was the most river-dominated while the Senegal in west Africa was the other extreme, the most wave-dominated. A model of the Senegal (Type VI in Figure 4-5) shows that abundant beach ridges are parallel to the prevailing shoreline trend and that the shore is relatively straight as a result of high wave energy and a strong unidirectional littoral drift.

(c) An intermediate delta form is represented by the delta of the Rio São Francisco del Norte in Brazil (Type V in Figure 4-5). Distributary-mouth-bar deposits are restricted to the immediate vicinity of the river mouth and are quickly remolded by waves. Persistent wave energy redistributes the riverine sediment to form extensive sand sheets. The exposed delta plain consists primarily of beach ridges and eolian dune fields.

(4) Tide-dominant deltas.

Three important processes characterize tide-dominated deltas:

(a) At the river mouths, mixing obliterates vertical density stratification, eliminating the effects of buoyancy.

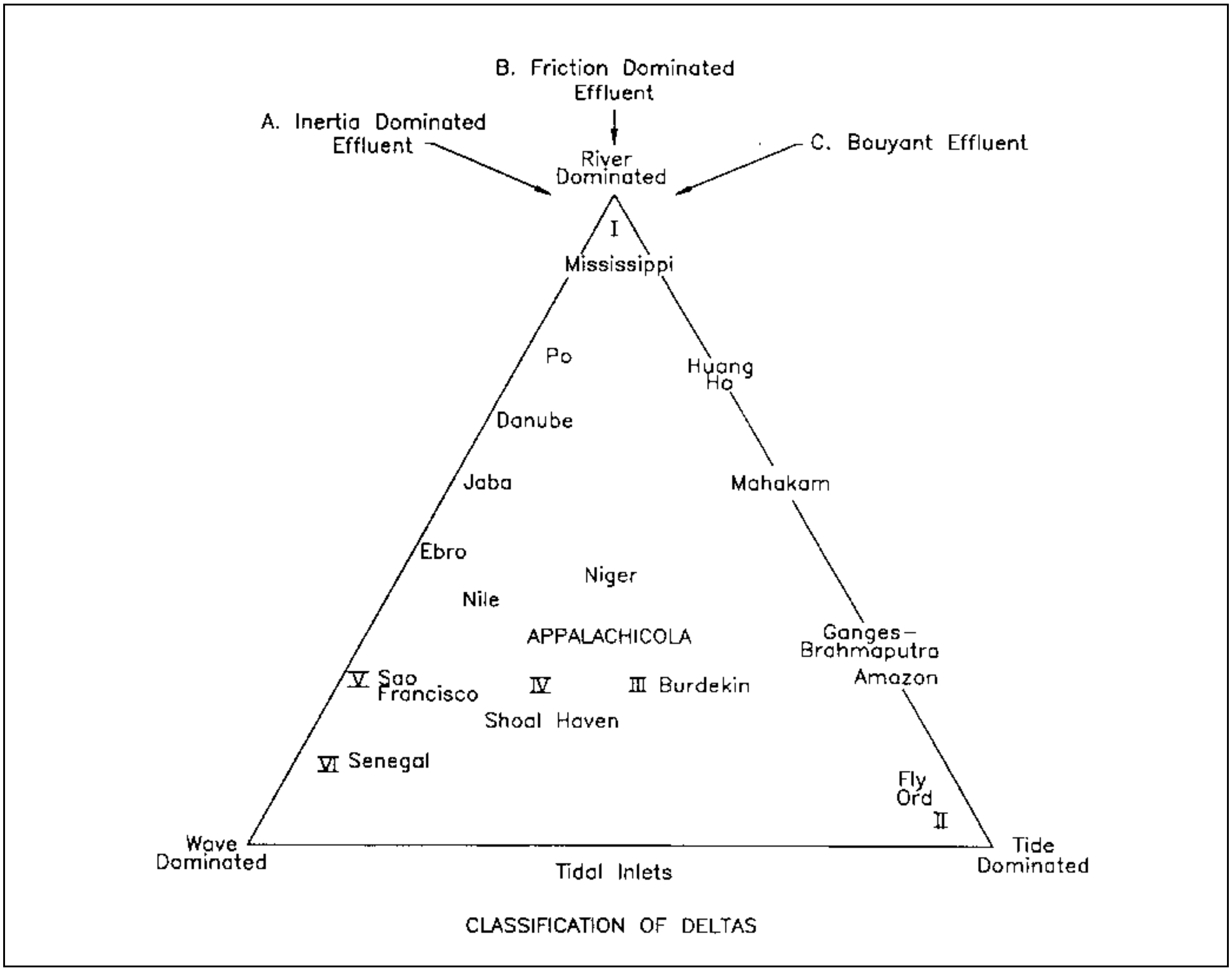

Figure 4-4. Comparison of deltaic dispositional models in terms of the relative importance of river, wave, and tide processes (from Wright (1985))

(b) For part of the year, tidal currents may be responsible for a greater fraction of the sediment-transporting energy than the river. As a result, sediment transport in and near the river mouth is bidirectional over a tidal cycle.

(c) The location of the land-sea interface and the zone of marine-riverine interactions is greatly extended both vertically and horizontally. Examples of deltas that are strongly influenced by tides include the Ord (Australia), Shatt-al-Arab (Iraq), Amazon (Brazil), Ganges-Brahmaputra (Bangladesh), and the Yangtze (China).

Characteristic features of river mouths in macrotidal environments are bell-shaped, sand-filled channels and linear tidal sand ridges. The crests of the ridges, which have relief of 10-20 m, may be exposed at low tide. The ridges replace the distributary-mouth bars found at other deltas and become the dominant sediment-accumulation form. As the delta progrades over time, the ridges grow until they are permanently exposed, forming large, straight tidal channels (Type II in Figure 4-5). An example of a macrotidal delta is the Ord of Western Australia.

(5) Intermediate forms.

(a) As stated above, the morphology of most deltas is a result of a combination of riverine, tidal, and wave forces. One example of an intermediate form is the Burdekin Delta of Australia (Type II in Figure 4-5).

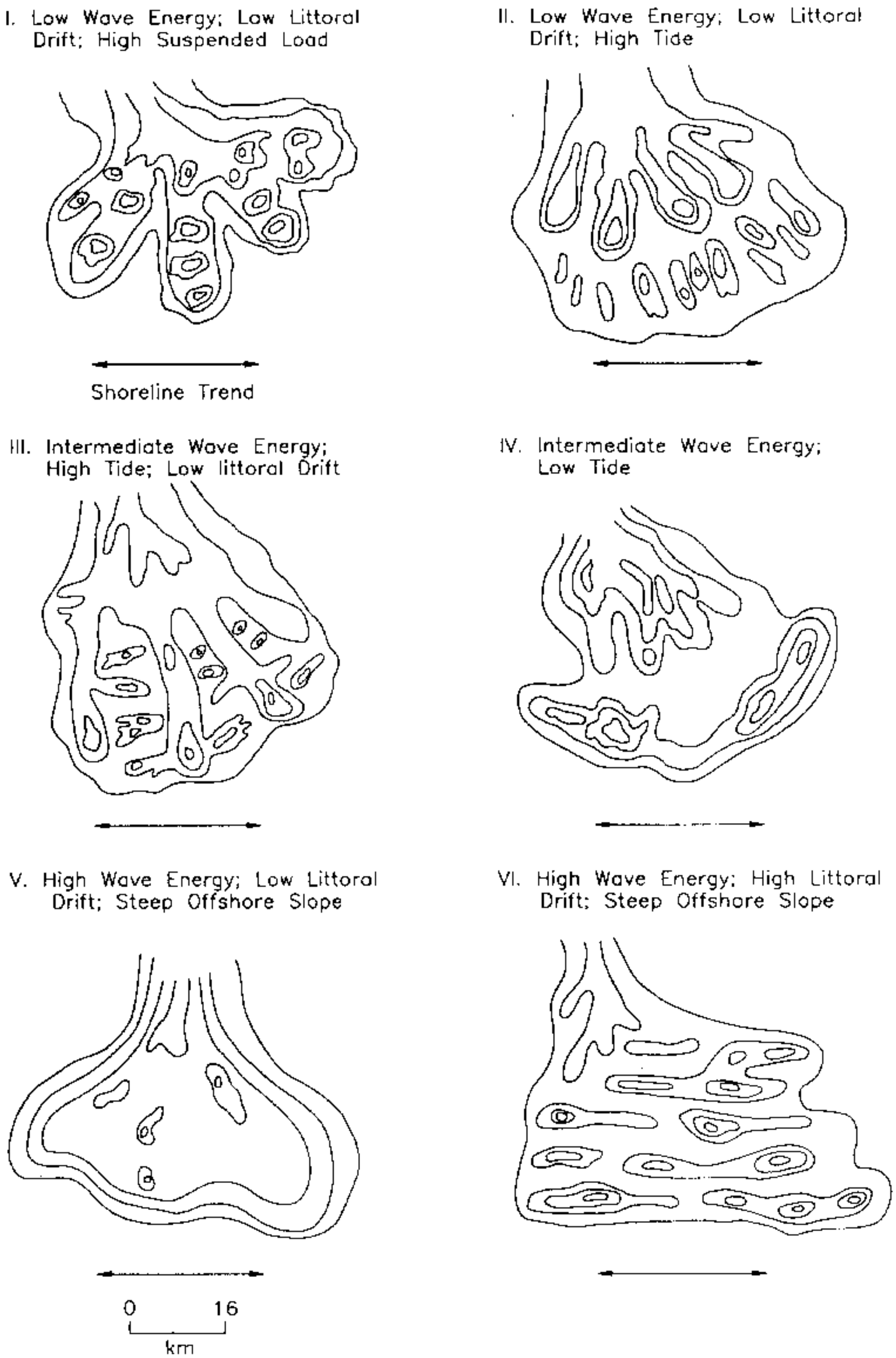

Figure 4-5. Isopach maps of six deltaic models (from Coleman and Wright (1973)). Locations of models with respect to energy factors are plotted in Figure 4-1

High waves redistribute sands parallel to the coastline trend and remold them into beach ridges and barriers. Within the river mouths, tidal currents produce sand-filled river channels and tidal creeks. This type of delta displays a broad range of characteristics, depending upon the relative strength of waves versus tides. In addition, features may vary seasonally if runoff and wave climate change. Other examples include the Irrawaddy (Burma), Mekong (Vietnam), and Red (Vietnam) Deltas (Wright 1985).

(b) The fourth model of delta geometry is characterized by offshore bay-mouth barriers that shelter lagoons, bays, or estuaries into which low-energy deltas prograde (Type IV, Figure 4-5). Examples include the Appalachicola (Florida Panhandle), Sagavanirktok (Alaska), and Shoalhaven (southeastern Australia) Deltas (Wright 1985). In contrast to the river-dominant models, the major accumulation of prodelta mud occurs landward of the main sand body (the barrier), and at the same elevation, within the protected bay. Although suspended fines reach the open sea, wave action prevents mud accumulation as a distinct unit over the open shelf.

c. River mouth flow and sediment deposition.

(1) River mouth geometry and river mouth bars are influenced by, and in turn influence, effluent dynamics. This subject needs to be examined in detail because the principles are pertinent to both river mouths and tidal inlets. Diffusion of the river's effluent and the subsequent sediment dispersion depend on the relative strengths of three main factors:

- Inertia of the issuing water and associated turbulent diffusion.

- Friction between the effluent and the seabed immediately seaward of the mouth.

- Buoyancy resulting from density contrasts between river flow and ambient sea or lake water.

Based on these forces, three sub-classes of deltaic deposition have been identified for river-dominated deltas (Figure 4-4). Two of these are well illustrated by depositional features found on the Mississippi delta.

(2) Depositional model type A - inertia-dominated effluent.

(a) When outflow velocities are high, depths immediately seaward of the mouth tend to be large, density contrasts between the outflow and ambient water are low, and inertial forces dominate. As a result, the effluent spreads and diffuses as a turbulent jet (Figure 4-6a). As the jet expands, its momentum decreases, causing a reduction of its sediment carrying capacity. Sediments are deposited in a radial pattern, with the coarser bed load dropping just beyond the point where the effluent expansion is initiated. The result is basinward-dipping foreset beds.

(b) This ideal model is probably unstable under most natural conditions. As the river continues to discharge sediment into the receiving basin, shoaling eventually occurs in the region immediately beyond the mouth (Figure 4-6b). For this reason, under typical natural conditions, basin depths in the zone of the jet's diffusion are unlikely to be deeper than the outlet depth. Effluent expansion and diffusion become restricted horizontally as a plane jet. More important, friction becomes a major factor in causing rapid deceleration of the jet. Model 'A' eventually changes into friction-dominated Model 'B'.

(3) Depositional model type B - friction-dominated effluent.

(a) When homopycnal,[1] friction-dominated outflow issues over a shallow basin, a distinct pattern of bars and subaqueous levees is formed (Figure 4-7). Initially, the rapid expansion of the jet produces a broad, arcuate radial bar. As deposition continues, natural subaqueous levees form beneath the lateral boundaries of the expanding jet where the velocity decreases most rapidly. These levees constrict the jet from expanding further. As the central portion of the bar grows upward, channels form along the lines of greatest turbulence, which tend to follow the subaqueous levees. The result is the formation of a bifurcating channel that has a triangular middle-ground shoal separating the diverging channel arms. The flow tends to be concentrated into the divergent channels and to be tranquil over the middle ground under normal conditions.

(b) This type of bar pattern is most common where nonstratified outflow enters a shallow basin. Examples of this pattern (known as *crevasse splays* or *overbank splays*) are found at crevasses along the Mississippi River levees. These secondary channels run perpendicular to the main Mississippi channels and allow river water to debouch into the broad, shallow interdistributory bays. This

[1] River water and ambient water have the same density (for example, a stream entering a freshwater lake).

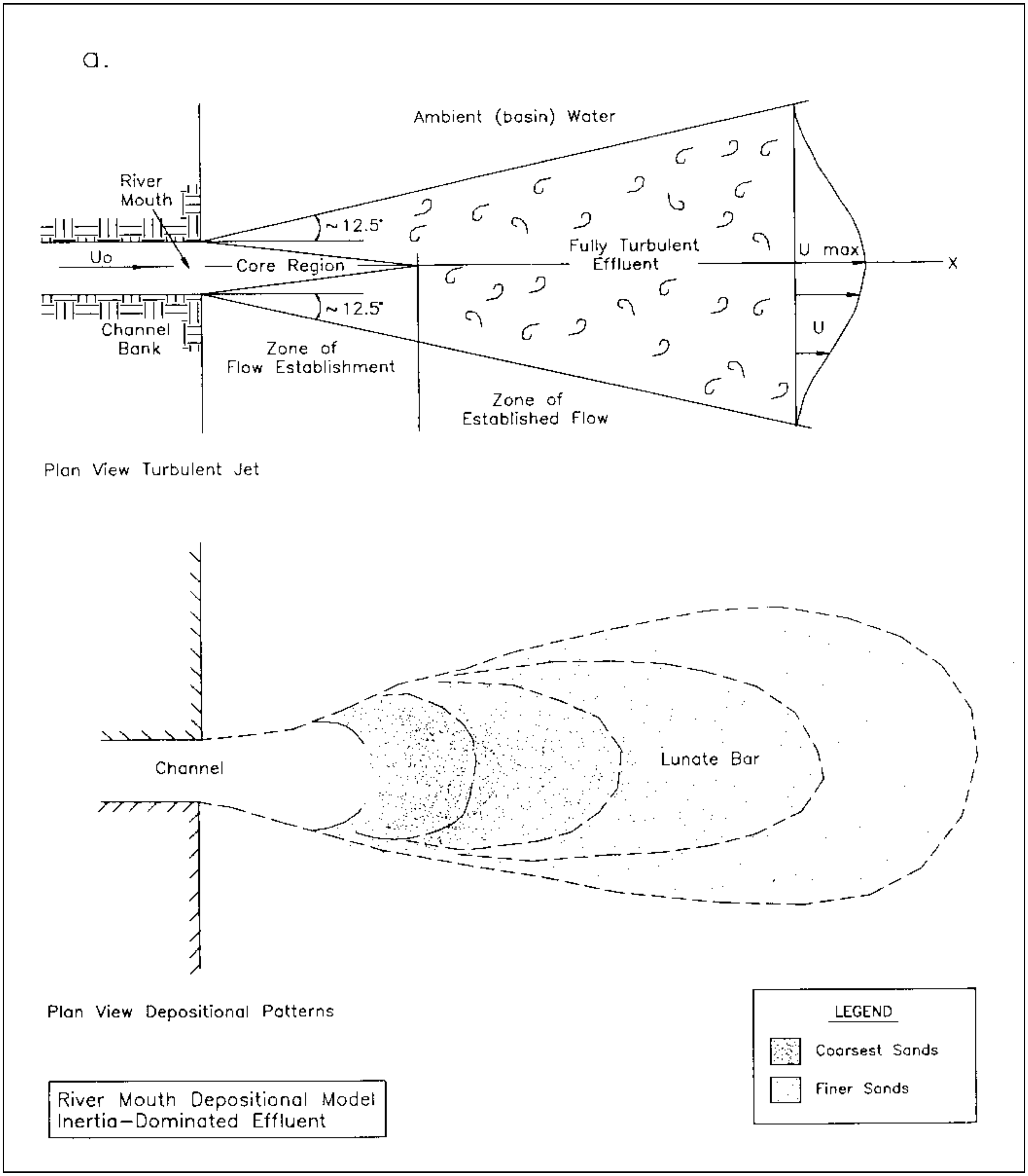

Figure 4-6. Plan view of depositional Model A, inertia-dominated effluent (adapted from Wright (1985)) (Continued)

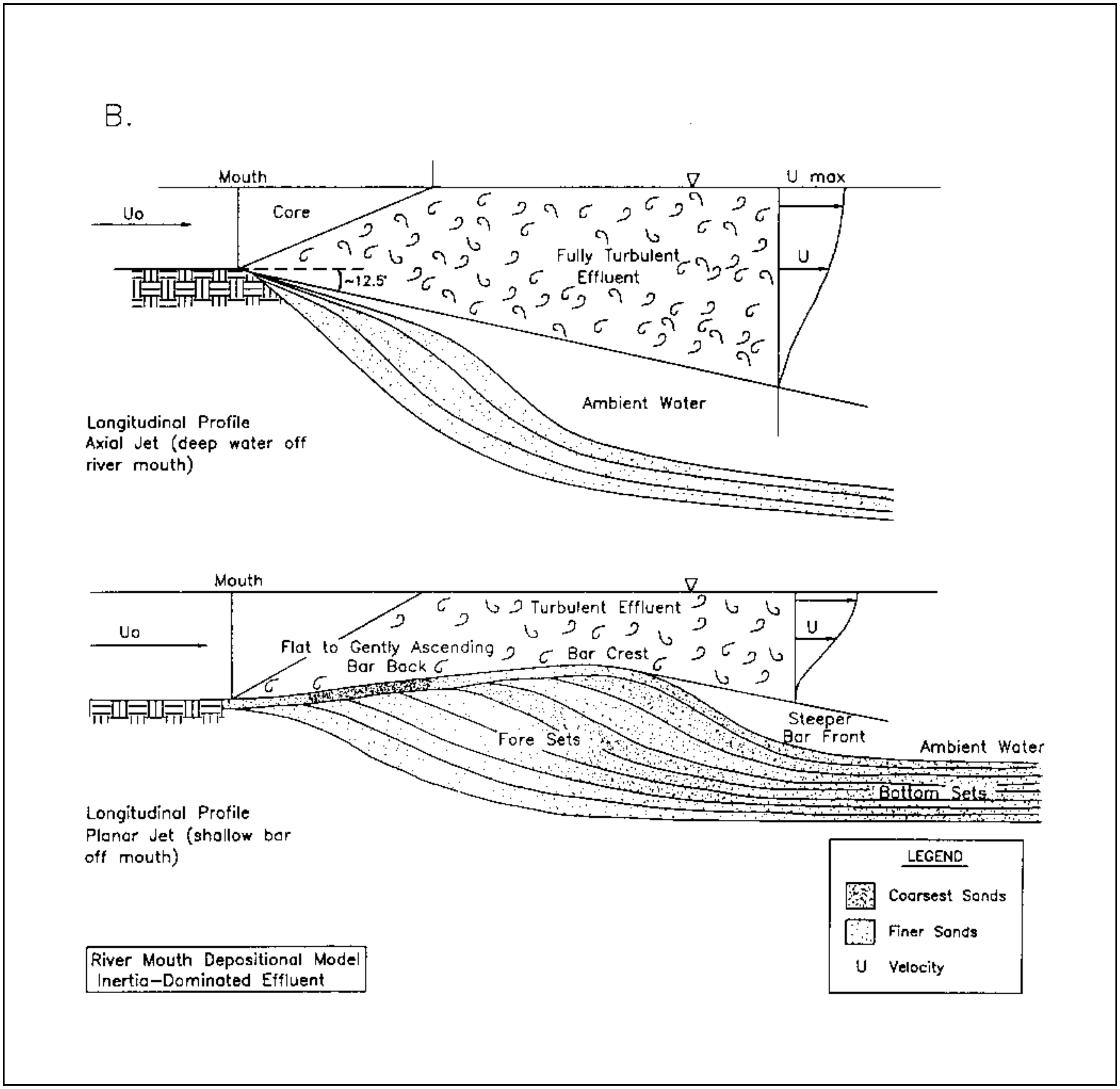

Figure 4-6. (Concluded)

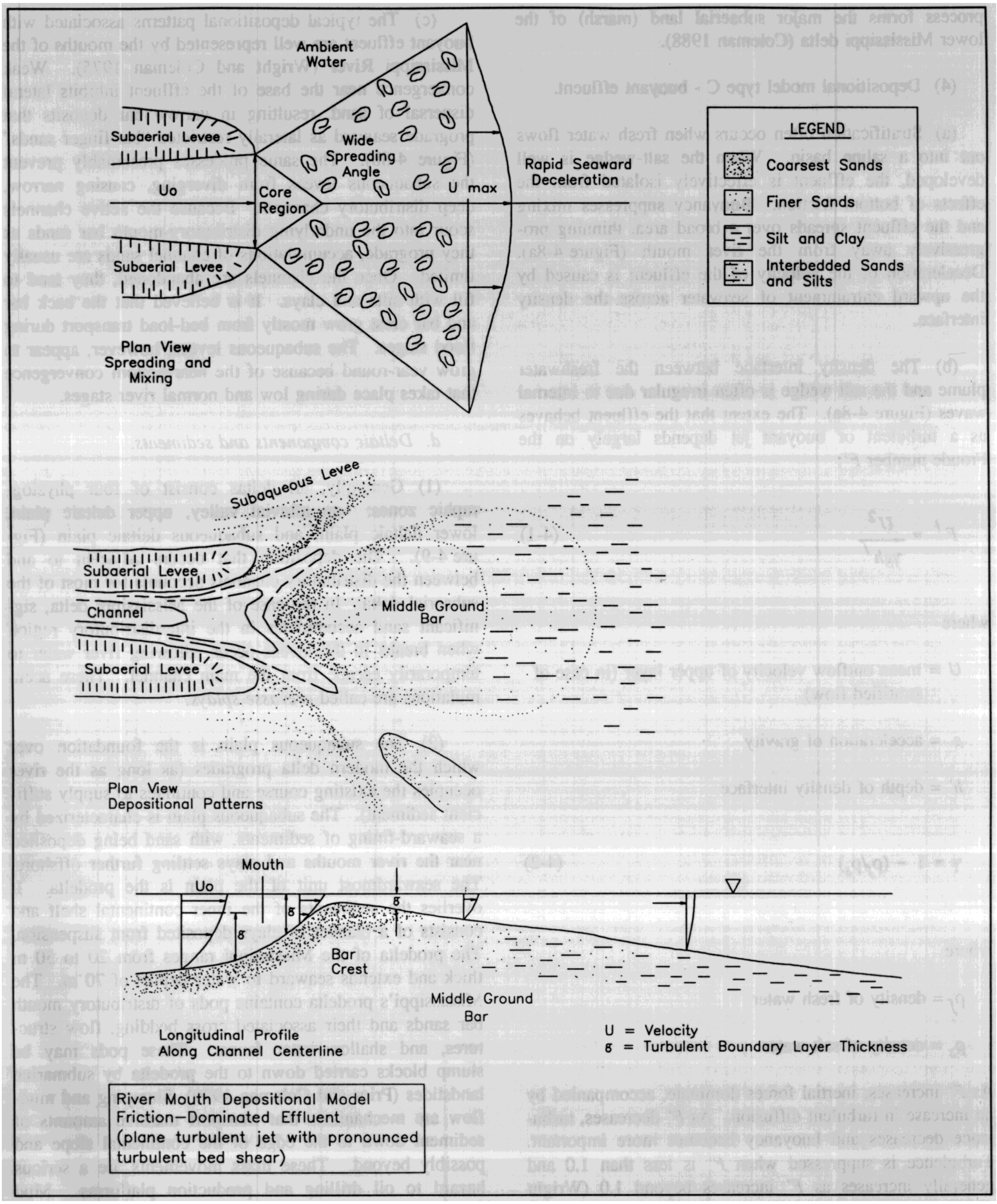

Figure 4-7. Depositional model type B, friction-dominated effluent (adapted from Wright (1985))

process forms the major subaerial land (marsh) of the lower Mississippi delta (Coleman 1988).

(4) Depositional model type C - buoyant effluent.

(a) Stratification often occurs when fresh water flows out into a saline basin. When the salt-wedge is well developed, the effluent is effectively isolated from the effects of bottom friction. Buoyancy suppresses mixing and the effluent spreads over a broad area, thinning progressively away from the river mouth (Figure 4-8a). Deceleration of the velocity of the effluent is caused by the upward entrainment of seawater across the density interface.

(b) The density interface between the freshwater plume and the salt wedge is often irregular due to internal waves (Figure 4-8a). The extent that the effluent behaves as a turbulent or buoyant jet depends largely on the Froude number *F'*:

$$F' = \frac{U^2}{\gamma g h'} \tag{4-1}$$

where

U = mean outflow velocity of upper layer (in case of stratified flow)

g = acceleration of gravity

h' = depth of density interface

$$\gamma = 1 - (\rho_f/\rho_s) \tag{4-2}$$

where

ρ_f = density of fresh water

ρ_s = density of salt water

As *F'* increases, inertial forces dominate, accompanied by an increase in turbulent diffusion. As *F'* decreases, turbulence decreases and buoyancy becomes more important. Turbulence is suppressed when *F'* is less than 1.0 and generally increases as *F'* increases beyond 1.0 (Wright 1985).

(c) The typical depositional patterns associated with buoyant effluent are well represented by the mouths of the Mississippi River (Wright and Coleman 1975). Weak convergence near the base of the effluent inhibits lateral dispersal of sand, resulting in narrow bar deposits that prograde seaward as laterally restricted "bar-finger sands" (Figure 4-8b). The same processes presumably prevent the subaqueous levees from diverging, causing narrow, deep distributory channels. Because the active channels scour into the underlying distributory-mouth bar sands as they prograde, accumulations of channel sands are usually limited. Once the channels are abandoned, they tend to fill with silts and clays. It is believed that the back bar and bar crest grow mostly from bed-load transport during flood stages. The subaqueous levees, however, appear to grow year-round because of the near-bottom convergence that takes place during low and normal river stages.

d. Deltaic components and sediments.

(1) Generally, all deltas consist of four physiographic zones: an alluvial valley, upper deltaic plain, lower deltaic plain, and subaqueous deltaic plain (Figure 4-9). The deposition that occurs adjacent to and between the distributory channels accounts for most of the subaerial delta. In the case of the Mississippi delta, significant sand accumulates in the interdistributory region when breaks in the levees occur, allowing river water to temporarily escape from the main channel. These accumulations are called *crevasse splays*.

(2) The subaqueous plain is the foundation over which the modern delta progrades (as long as the river occupies the existing course and continues to supply sufficient sediment). The subaqueous plain is characterized by a seaward-fining of sediments, with sand being deposited near the river mouths and clays settling further offshore. The seawardmost unit of the plain is the prodelta. It overlies the sediments of the inner continental shelf and consists of a blanket of clays deposited from suspension. The prodelta of the Mississippi ranges from 20 to 50 m thick and extends seaward to water depths of 70 m. The Mississippi's prodelta contains pods of distributory mouth bar sands and their associated cross bedding, flow structures, and shallow-water fauna. These pods may be slump blocks carried down to the prodelta by submarine landslides (Prior and Coleman 1979). Slumping and mudflow are mechanisms that transport massive amounts of sediment down to the edge of the continental slope and possibly beyond. These mass movements are a serious hazard to oil drilling and production platforms. Mud diapirs, growth faults, mud/gas vents, pressure ridges, and

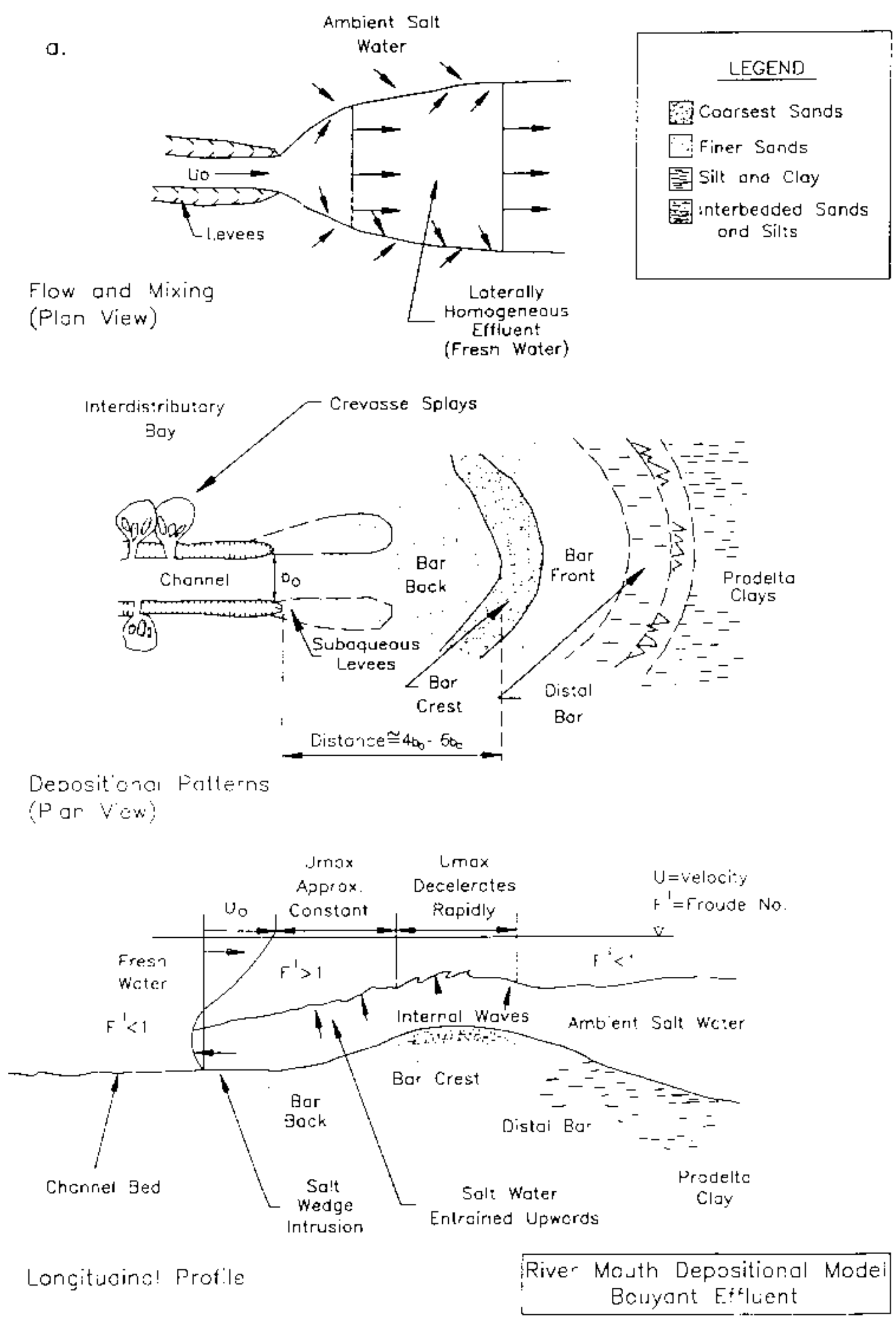

Figure 4-8. River mouth bar crest features, depositional model type C, buoyant effluent (adapted from Wright (1985)) (Continued)

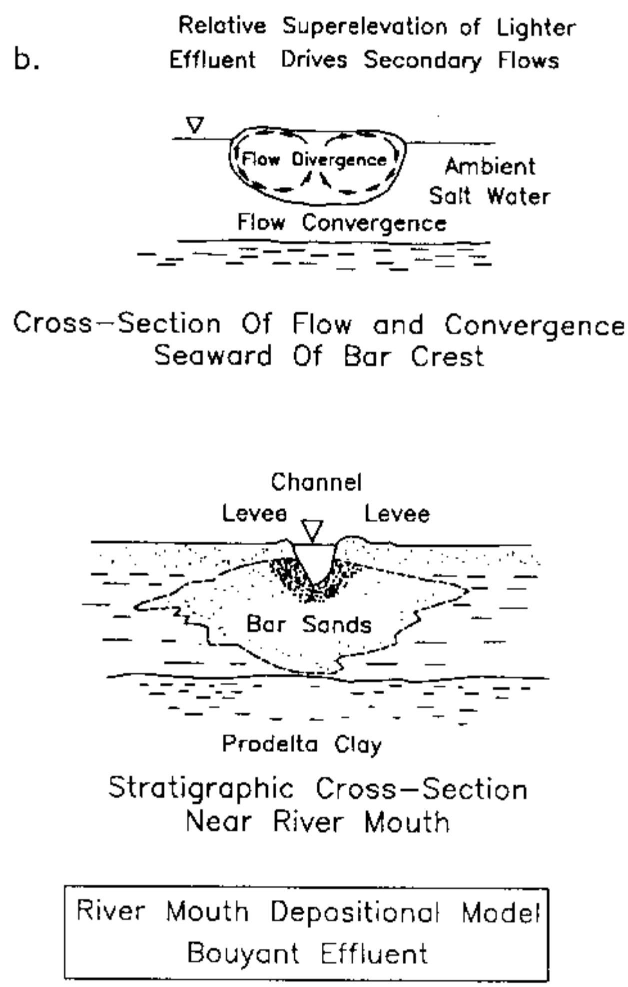

Figure 4-8. (Concluded)

mudflow gullies are other evidence of sediment instability on the Mississippi delta (Figure 4-10). Additional details of this interesting subject are covered in Coleman (1988), Coleman and Garrison (1977), Henkel (1970), and Prior and Coleman (1980).

(3) Above the delta front, there is a tremendous variability of sediment types. A combination of shallow marine processes, riverine influence, and brackish-water faunal activity causes the interdistributory bays to display an extreme range of lithologic and textural types. On deltas in high tide regions, the interdistributory bay deposits are replaced by tidal and intertidal flats. West of the Mississippi Delta is an extensive chenier plain. Cheniers are long sets of beach ridges, located on mudflats.

e. Mississippi Delta - Holocene history, dynamic changes.

(1) General. The Mississippi River, which drains a basin covering 41 percent of the continental United States (3,344,000 sq km), has built an enormous unconsolidated sediment accumulation in the Gulf of Mexico. The river has been active since at least late Jurassic times and has profoundly influenced deposition in the northern Gulf of Mexico. Many studies have been conducted on the Mississippi Delta, leading to much of our knowledge of deltaic sedimentation and structure. The ongoing research is a consequence of the river's critical importance to commerce and extensive petroleum exploration and

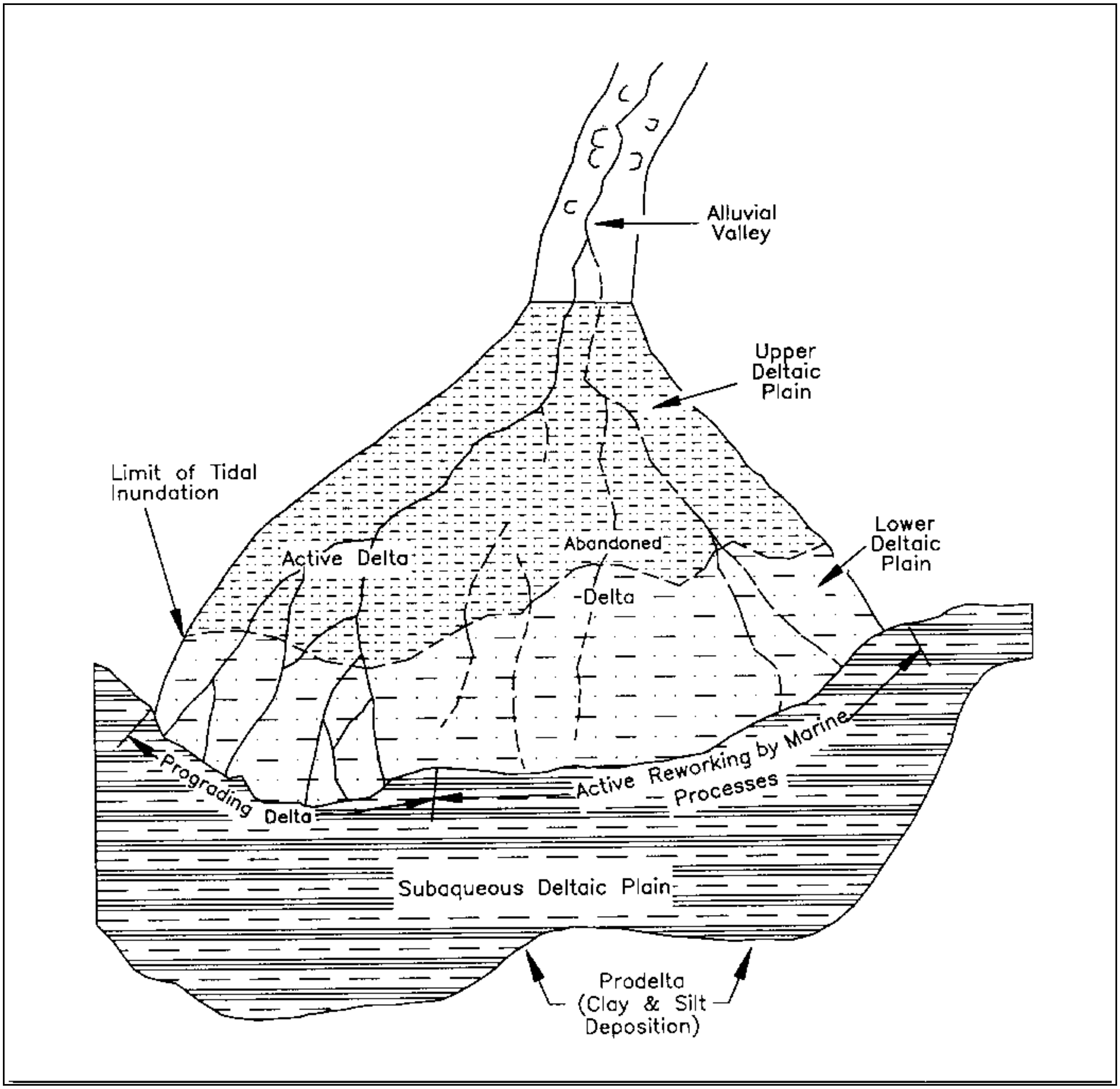

Figure 4-9. Basic physiographic units common to all deltas (from Wright (1985))

production in the northern Gulf of Mexico during the last 50 years.

(2) Deposition time scales. The Mississippi Delta consists of overlapping deltaic lobes. Each lobe covers 30,000 sq km and has an average thickness of about 35 m. The lobes represent the major sites of the river's deposition. The process of switching from an existing lobe to a new outlet takes about 1500 years (Coleman 1988). Within a single lobe, deposition in the bays occurs from overbank flows, crevasse splays, and biological production. The bay fills, which cover areas of 250 sq km and have a thickness of only 15 m, accumulate in only about 150 years. Overbank splays, which cover areas of 2 sq km and are 3 m thick, occur during major floods when the natural levees are breached. The mouths of the Mississippi River have prograded seawards at remarkable rates. The distributory channels can form

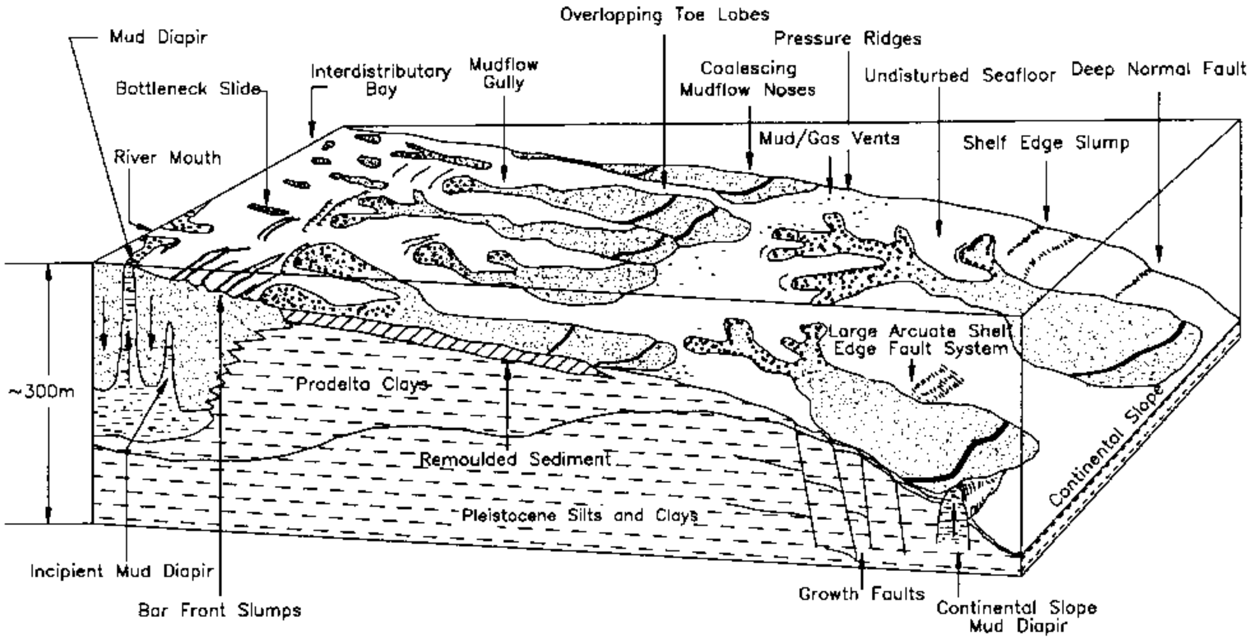

Figure 4-10. Structures and types of sediment instabilities on the Mississippi Delta (from Coleman (1988))

sand bodies that are 17 km long, 8 km wide, and over 80 m thick in only 200 years (Coleman 1988).

(3) Holocene history. During the last low sea level stand, 18,000 years ago, the Mississippi River entrenched its valley, numerous channels were scoured across the continental shelf, and deltas were formed near the shelf edge (Suter and Berryhill 1985). As sea level rose, the site of deposition moved upstream to the alluvial valley. By about 9,000 years before present, the river began to form its modern delta. In more recent times, the shifting deltas of the Mississippi have built a delta plain covering a total area of 28,500 sq km. The delta switching, which has occurred at high frequency, combined with a rapidly subsiding basin, has resulted in vertically stacked cyclic sequences. Because of rapid deposition and switching, in a short time the stacked cyclic deltaic sequences have attained thicknesses of thousands of meters and covered an area greater than 150,000 sq km (Coleman 1988). Figure 4-11 outlines six major lobes during the last 7,500 years.

(4) Modern delta. The modern delta, the Balize or Birdfoot, began to prograde about 800 to 1,000 years ago. Its rate of progradation has diminished recently and the river is presently seeking a new site of deposition. Within the last 100 years, a new distributory, the Atchafalaya, has begun to divert an increasing amount of the river's flow. Without river control structures, the new channel would by now have captured all of the Mississippi River's flow, leading to rapid erosion of the Balize Delta. (It is likely that there would be a commensurate deterioration of the economy of New Orleans if it lost its river.) Even with river control projects, the Atchafalaya is actively building a delta in Atchafalaya Bay (lobe 6 in Figure 4-11).

f. Sea level rise and deltas.

(1) Deltas experience rapid local relative sea level rise because of the natural compaction of deltaic sediments from dewatering and consolidation. Deltas are extremely vulnerable to storms because the subaerial surfaces are flat and only slightly above the local mean sea level. Only a slight rise in sea level can extend the zone subject to storm surges and waves further inland. As stated earlier, delta evolution is a balance between the accumulation of fluvially supplied sediment and the reworking, erosion, and transport of deltaic sediment by marine processes (Wright 1985). Even a river like the Mississippi, which has a high sediment load and drains into a low wave-energy basin, is prograding only in the vicinity of the present distributory channels, the area defined as the active delta (Figure 4-9).

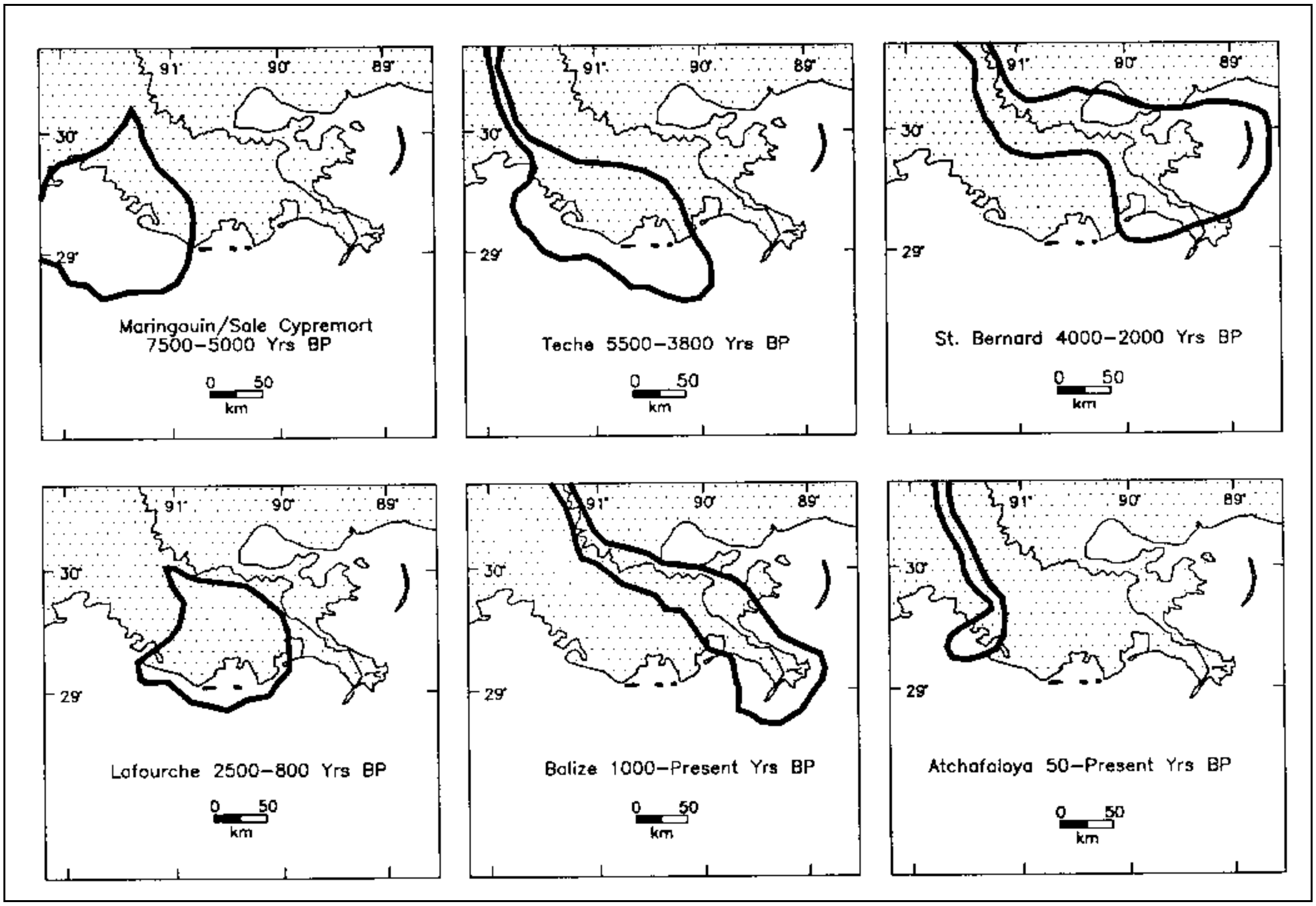

Figure 4-11. Shifting sites of deltaic sedimentation of the Mississippi River (from Coleman (1988))

(2) Deltas are highly fertile agriculturally because of the steady supply of nutrient-laden soil. As a result, some of the world's greatest population densities - over 200 inhabitants per sq km - are found on deltas (*The Times Atlas of the World* 1980):

- Nile Delta, Egypt.
- Chang Jiang (Yangtze), China.
- Mekong, Vietnam.
- Ganga (Ganges), Bangladesh.

These populations are very vulnerable to delta land loss caused by rising sea level and by changes in sediment supply due to natural movements of river channels or by upland man-made water control projects.

(3) Inhabitants of deltas are also in danger of short-term changes in sea level caused by storms. Tropical storms can be devastating: the Bay of Bengal cyclone of November 12, 1970, drowned over 200,000 persons in what is now Bangladesh (Carter 1988). Hopefully, public education, improving communications, better roads, and early warning systems will be able to prevent another disaster of this magnitude. Coastal management in western Europe, the United States, and Japan is oriented towards the orderly evacuation of populations in low-lying areas and has greatly reduced storm-related deaths. In contrast to the Bay of Bengal disaster, Hurricane Camille (August 17-20, 1969), caused only 236 deaths in Louisiana, Mississippi, Alabama, and Florida.

4-4. Inlet Processes and Dynamics

a. Introduction.

(1) Coastal inlets play an important role in nearshore processes around the world. *Inlets* are the openings in coastal barriers through which water, sediments, nutrients, planktonic organisms, and pollutants are exchanged

between the open sea and the protected embayments behind the barriers. Inlets are not restricted to barrier environments or to shores with tides; on the West Coast and in the Great Lakes, many river mouths are considered to be inlets, and in the Gulf of Mexico, the wide openings between the barriers, locally known as passes, are also inlets. Inlets can be cut through unconsolidated shoals or emergent barriers as well as through clay, rock, or organic reefs (Price 1968). There is no simple, restrictive definition of inlet - based on the geologic literature and on regional terminology, almost any opening in the coast, ranging from a few meters to several kilometers wide, can be called an inlet. Inlets are important economically to many coastal nations because harbors are often located in the back bays, requiring that the inlets be maintained for commercial navigation. At many inlets, the greatest maintenance cost is that incurred by repetitive dredging of the navigation channel. Because inlets are hydrodynamically very complex, predictions of shoaling and sedimentation have often been unsatisfactory. A better understanding of inlet sedimentation patterns and their relationship to tidal and other hydraulic processes can hopefully contribute to better management and engineering design.

(2) Tidal inlets are analogous to river mouths in that sediment transport and deposition patterns in both cases reflect the interaction of outflow inertia and associated turbulence, bottom friction, buoyancy caused by density stratification, and the energy regime of the receiving body of water (Wright and Sonu 1975). However, two major differences usually distinguish lagoonal inlets from river mouths, sometimes known as fluvial or riverine inlets (Oertel 1982).

a. Lagoonal tidal inlets experience diurnal or semidiurnal flow reversals.

b. Lagoonal inlets have two opposite-facing mouths, one seaward and the other lagoonward. The sedimentary structures which form at the two openings differ because of differing energy, water density, and geometric factors.

(3) This section reviews tidal flow in inlets and relates it to sedimentary structures found in the channels and near the mouths. Several conceptual models are reviewed and compared to processes that have been observed on the Atlantic and Gulf Coasts of the United States.

(4) The term *lagoon* refers to the coastal pond or embayment that is connected to the open sea by a tidal inlet. The *throat* of the inlet is the zone of smallest cross section, which, accordingly, has the highest flow velocities. The *gorge* is the deepest part of an inlet and may extend seaward and landward of the throat (Oertel 1988). *Shoal* and *delta* are often used interchangeably to describe the ebb-tidal sand body located at the seaward mouth of an inlet.

b. Technical literature. Pioneering research on the stability of inlets was performed by Francis Escoffier (1940, 1977). O'Brien (1931, 1976) derived general empirical relationships between tidal inlet dimension and tidal prism. Keulegan (1967) developed algorithms to relate tidal prism to inlet cross section. Bruun (1966) examined inlets and littoral drift, and Bruun and Gerritsen (1959, 1961) studied bypassing and the stability of inlets. Hubbard, Oertel, and Nummedal (1979) described the influence of waves and tidal currents on tidal inlets in the Carolinas and Georgia. Hundreds of other works are referenced in the USACE *General Investigation of Tidal Inlets* (GITI) reports (Barwis 1976), in special volumes like *Hydrodynamics and Sediment Dynamics of Tidal Inlets* (Aubrey and Weishar 1988), in textbooks on coastal environments (Carter 1988; Cronin 1975; Komar 1976), and in review papers (Boothroyd 1985; FitzGerald 1988). Older papers on engineering aspects of inlets are cited in Castañer (1971). There are also numerous foreign works on tidal inlets: Carter (1988) cites references from the British Isles; Sha (1990) from the Netherlands; Nummedal and Penland (1981) and FitzGerald, Penland, and Nummedal (1984) from the North Sea coast of Germany; and Hume and Herdendorf (1988, 1992) from New Zealand.

c. Classification of inlets and geographic distribution.

(1) Tidal inlets, which are found around the world in a broad range of sizes and shapes, encompass a variety of geomorphic features. Because of their diversity, it has been difficult to develop a suitable classification scheme. One approach has been to use an energy-based criteria, in which inlets are ranked according to the wave energy and tidal range of the environment in which they are located.

(2) Regional geological setting can be a limiting factor restricting barrier and, in turn, inlet development. High relief, leading-edge coastlines have little room for sediment to accumulate either above or below sea level. Sediment tends to collect in pockets between headlands, few lagoons are formed, and inlets are usually restricted to river mouths. An example is the Pacific coast of North America, which, in addition to being steep, is subject to high wave energy.

(3) Underlying geology may also control inlet location and stability. Price and Parker (1979) reported that certain areas along the Texas coast were always characterized by inlets, although the passes tended to migrate back and forth along a limited stretch of the coast. The positions of these permanent inlets were tectonically controlled, but the openings were maintained by tidal harmonics and hydraulics. If storm inlets across barriers were not located at the established stable pass areas, the inlets usually closed quickly. Some inlets in New England are anchored by bedrock outcrops.

d. Hydrodynamic processes in inlets.

(1) General patterns of inlet flow. The interaction of a jet that issues from an inlet or river mouth with the downstream water mass is a complex phenomenon. Three broad classes of flow have been identified (Wright 1985):

- *Hypopycnal* outflow, in which a wedge of less dense fresh water flows over the denser sea water beyond the mouth.
- *Hyperpycnal* outflow, where the issuing water is denser than and plunges beneath the basin water.
- *Homopycnal* outflow, in which the jet and the downstream water are of the same density or are vertically mixed.

(a) Hypopycnal flow. Horizontally stratified hypopycnal flow is usually associated with river mouths and estuaries (Carter 1988; Wright 1985). As an example, the freshwater plume from the Amazon is so enormous when it spreads over the sea surface, early explorers of the New World refilled their water casks while still out of sight of land (Morrison 1974).

(b) Hyperpycnal flow. This occurs when outflow from hypersaline lagoons or rivers with extreme sediment load concentrations is denser than the water into which it issues. The Huang Ho River of China is cited as an example, but little has been published in English about this uncommon situation (Wright 1985). It is unknown if hyperpycnal conditions occur at any tidal inlets around the United States.

(c) Homopycnal flow. At most tidal inlets, strong jets - steady unidirectional currents - are produced as the tide rises and falls along the open coast and the water level in the lagoon rises and falls accordingly. Joshi and Taylor (1983) describe three elements of a fully developed jet:

(1) The source area upstream where the water converges before entering the pass (inlet).

(2) The strong, confined flow within the throat (jet).

(3) A radially expanding, vortex-dominated lobe downstream of the opening of the inlet (Figure 4-12).

(d) Carter (1988) reports that most inlet jets are homopycnal, especially at narrow inlets that drain large lagoons having no other openings to the sea. Presumably, his statement refers to tidal lagoons that have only a limited freshwater inflow. Where there is a significant fluvial input, the water in the lagoon becomes brackish and a more complicated flow regime develops. As an example, at East Pass, Florida, on the northeast Gulf of Mexico, the flow within the inlet proper is dominated by either the ebb or flood tide, but stratification occurs in Choctawhatchee Bay at the flood-tide shoal and at the Gulf of Mexico exit over the ebb-tide shoal.

(2) Jets and converging source flow at inlet openings. At inlets with stable margins (especially ones with jetties), the stream of turbulent water that discharges through the orifice into a large unrestricted basin can be considered a free jet (Oertel 1988). Either axial or planar jets can form, depending on the density difference between the outflow and the water into which it is flowing.

(a) Axial jets. Homopycnal flow through an orifice forms an axial jet. In an ideal system without friction or waves, the near field (the zone of flow establishment) extends about 4D seaward of the inlet's mouth, where D equals the diameter of the orifice (Figure 4-13a). Beyond 4D, in the far field, the jet spreads and loses velocity. The current velocity in the near field is estimated to remain about the same as in the throat. Based on this model, Oertel (1988) suggests that well-established channels should form to a distance of about 4D from the inlet throat. In the far field, Unluata and Ozsoy (1977) calculated that there is an exponential growth in jet width and an exponential decay of center line velocity. Fort Pierce Inlet, on the Atlantic coast of Florida, is an example of a site where a distinct axial jet forms at ebb tide (Joshi and Taylor 1983).

(b) Planar jets. When the water emerging from an inlet is buoyant, a planar jet forms. This jet spreads more rapidly in the near field than the axial type, extending to a distance of about 4D, where D = width of the mouth (Oertel 1988).

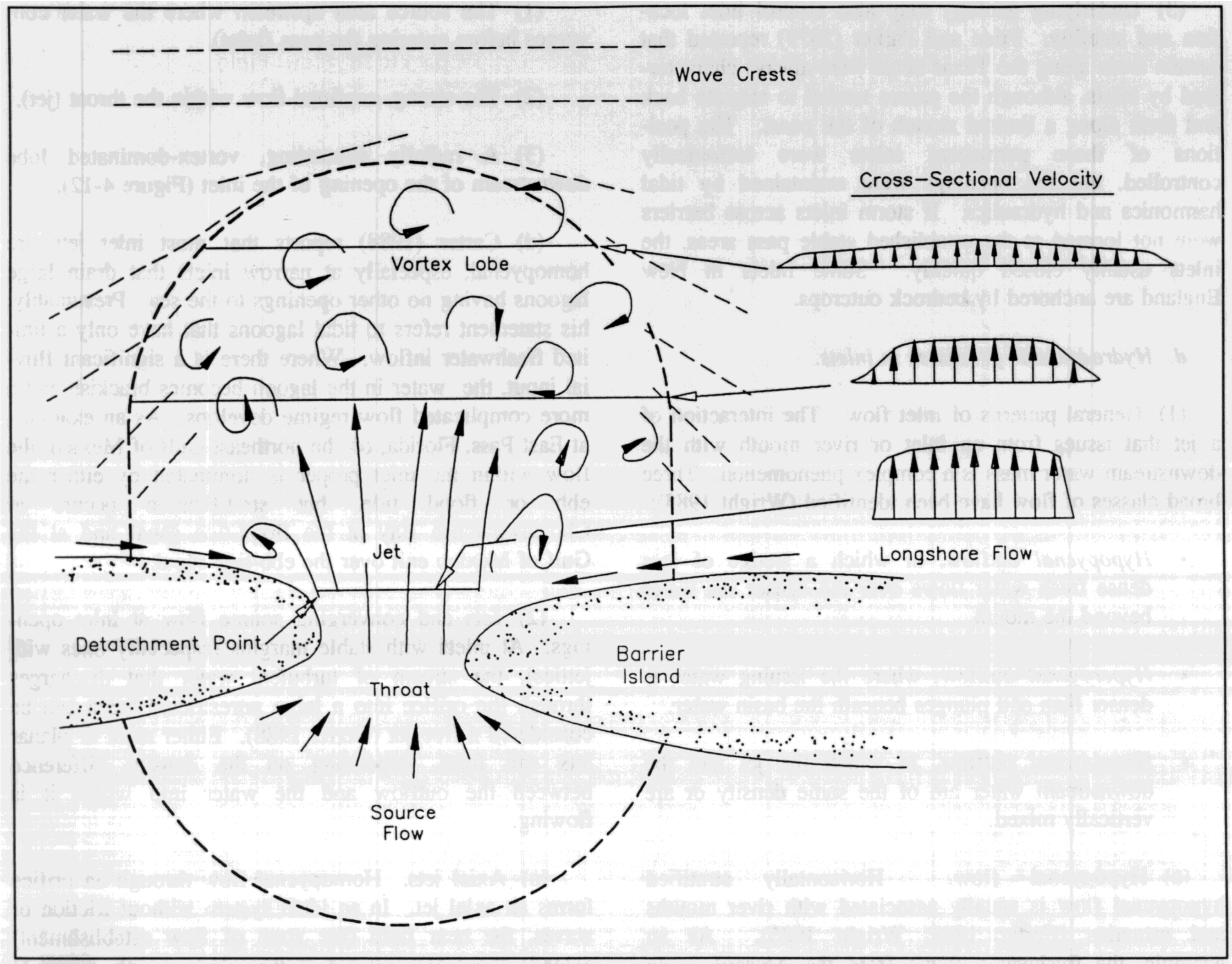

Figure 4-12. Three elements of flow through an idealized tidal inlet: source, jet, and expanding lobe (from Carter (1988))

(c) Planar jets at natural settings. In nature, the near and far fields of natural jets are affected by waves, littoral currents, friction, and bottom topography. Ismail and Wiegel (1983) have calculated that wave momentum flux is a major factor causing a jet's spreading rate to increase. The seafloor, especially if there is a shallow ebb-tide shoal, will squeeze the jet vertically and enhance spreading. Because of these factors, the planar jet model may be a more realistic description of the effluent at most tidal inlets. Aerial photographs from St. Mary's Entrance and Big Hickory and New Passes, Florida, clearly show jets spreading laterally immediately upon exiting the mouths (Joshi and Taylor 1983). At East Pass, Florida, dark, humate-stained water of the ebb tide expands beyond the jetties, forming an oval which covers the ebbtide shoal. Drogue studies in 1970 showed that the plume was buoyant and that below it, Gulf of Mexico water flowed in a westward direction (Sonu and Wright 1975).

(d) Flow at landward openings of inlets. Most of the technical literature has described jets that form at the seaward mouths of rivers or tidal inlets. On the landward side of inlets, a jet can only form if there is an open-water lagoon. In the back-bay areas of many barrier island systems, there are marshes and shoals, and flood flow is restricted to the deep channels (well-documented examples include North Inlet, South Carolina (Nummedal and Humphries 1978) and Sebastian Inlet, Florida (Stauble et al. 1988)). Both confined and jet-like flow may occur in lagoons in high tide-range coastlines. The

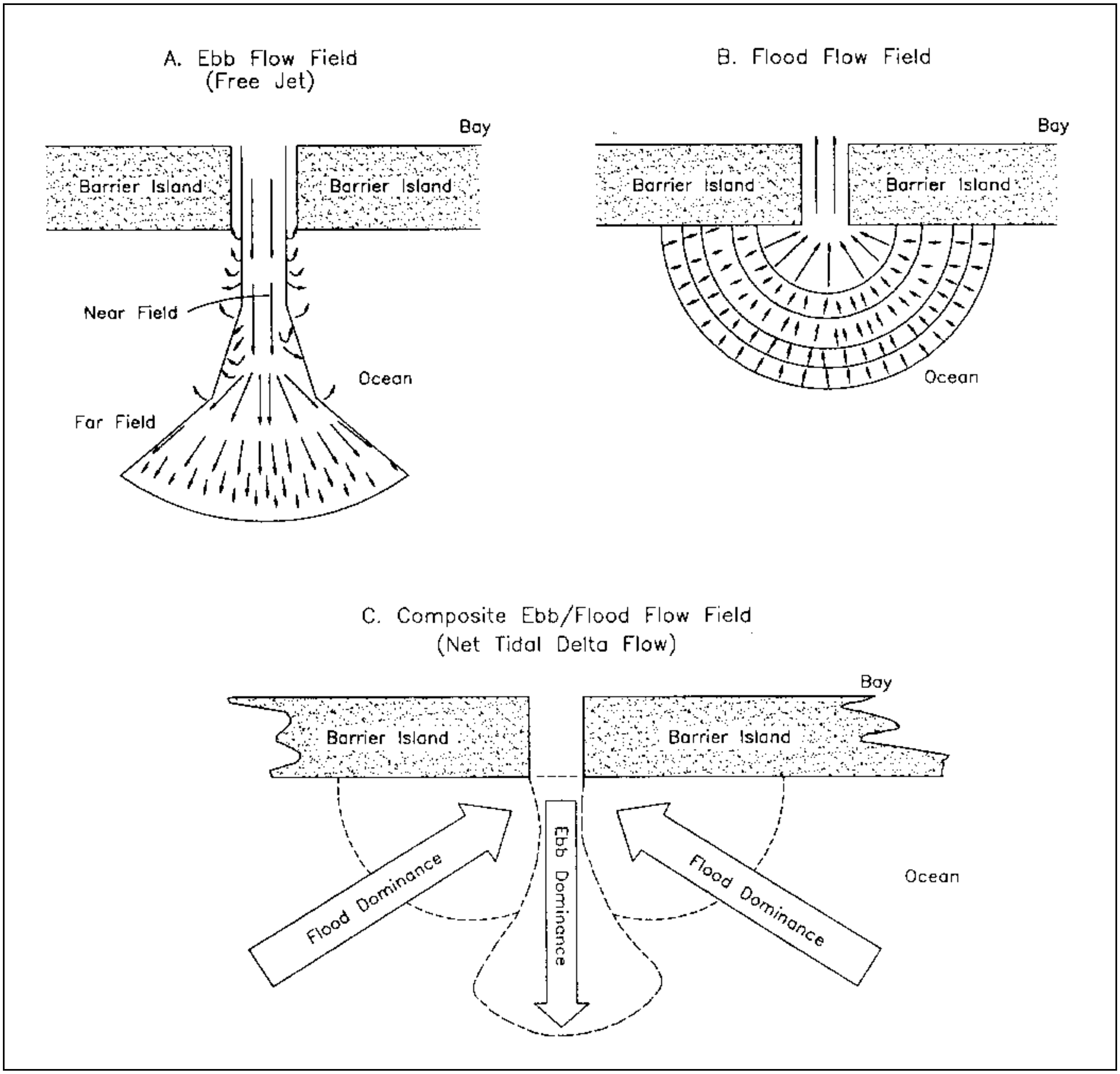

Figure 4-13. Sketch maps showing idealized flood and ebb flow fields (from Oertel (1988))

flood is initially restricted to established channels, but, as the water in the lagoon rises, the flow is able to spread beyond the confines of the channels and a plume develops. Nummedal and Penland (1981) documented this phenomenon in Norderneyer Seegat in Germany, where the tide range was 2.5 m.

(e) Source flow fields. During the flood at the seaward end of an unjettied inlet, the inflowing water uniformly converges in a semicircular pattern towards the inlet's throat (Figure 4-13b; Oertel 1988). Because the flow field is so broadly distributed, flood velocity is much lower than ebb jet velocity, particularly in the near field. It is unclear how the source flow field behaves at an inlet with seaward-projecting jetties. It seems likely that the streamlines wrap around the projecting jetties, but velocities along the outer side of the jetties are probably low. It may be difficult to verify this model at most sites because of the influence of waves, winds, currents, and local bathymetry.

(3) Influence of water mass stratification on inlet flow. When a lagoon contains brackish water, salt wedge dynamics can occur, where the incoming flood flows under less dense bay water. Mixing between the two waters occurs along a horizontal density interface. During ebb tide, a buoyant planar jet forms at the seaward opening of the inlet similar to the effluent from rivers.

(a) Wright, Sonu, and Kielhorn (1972) described how density stratification affected flow at the Gulf of Mexico and Choctawhatchee Bay openings of East Pass, Florida.

(b) During flood tide, drogues and dye showed that the incoming salty Gulf of Mexico water met the brackish bay water at a sharp density front and then dove underneath (Figure 4-14). The drogues indicated that the sea water intruded at least 100 m beyond the front into Choctawhatchee Bay. This was the reason that bed forms within the channels displayed a flood orientation over time.

(c) With the onset of ebb tide in East Pass, the seaward flow in the upper brackish layer increased in velocity and pushed the density front back towards the inlet. Initially, as the upper brackish layer flowed seaward, saline Gulf water underneath the interface continued to flow northwards into the bay. Within 2 hr after the onset of ebb flow, the current had reversed across the entire water column. As the brackish Choctawhatchee Bay water progressed southward through the inlet, it mixed to an increasing degree with the seawater underneath. By the time it reached the seaward mouth of the inlet, vertical mixing was nearly complete. As the ebb progressed, the wedge of brackish water continued to migrate seaward until it stopped near the edge of the flood-tide shoal bar crest, where it remained for the rest of the ebb cycle (Wright and Sonu 1975).

(4) Tidal flow and velocity asymmetry. *Tidal prism*, the amount of water that flows through an inlet, is determined by the tidal range, multiplied by the area of the bay which is supplied by the inlet. Prism may be one of the most important of the additional factors that determines the morphology of coastal inlets and their adjacent barrier islands (Davis and Hayes 1984). Along a reach of where tidal range is relatively constant, an inlet supplying a large bay will experience a much greater discharge than an inlet supplying a small bay. In addition, the inlet connecting the large bay to the sea will experience proportionately greater discharge during times when tide range is high (e.g. spring tides). However, it takes considerable time for a large bay to fill and empty as the tidal cycle progresses; therefore, the overall range of water levels in a large bay may be less than in a small bay.

(a) Effect of back bay salt marshes. Nummedal and Humphries (1978) describe how the bathymetry of a bay controls the degree of velocity asymmetry through an inlet gorge. The bays in the southeastern United States are typically filled with intertidal salt marshes, leaving only about 20 percent of the bay area as open water. The large variation in water surface area during the tide cycle tends to produce a strong ebb-dominant flow in these systems.

(b) Beginning of flood tide. As the open-water tide begins to rise at the beginning of the flood, water flows into the inlet and rapidly floods the limited-volume tidal channels in the back bay. The flow at this stage is reasonably efficient because the water level in the channels is able to rise almost as quickly as water outside the inlet (some delay is caused by friction).

(c) Near high tide. Once the water level in the bay rises enough to inundate the tidal channels, any additional water is free to spread laterally over a much greater expanse of marsh terrain. As a result, a lag develops because the flood tide cannot flow through the inlet quickly enough to fill the bay and keep pace with the rise in the open-water tide.

(d) Beginning of ebb tide. At high tide, the bay water level is below the open-coast level. As a result, although the open coast tide is beginning to drop, the bay is still rising. Eventually, the two water levels equalize, and the flow through the inlet turns to ebb.

(e) Near low tide. At the final stages of the ebb tide, the water in the bay has fallen below the marsh level and water is primarily confined to the back bay tidal channels. Because the channels contain only a limited volume, the water level drops almost as quickly as the open-coast level. (However, the process is not totally efficient because considerable water continues to drain out of the plants and saturated soil over time.)

(f) Low tide. At low tide, water levels within the bay and along the open coast are almost equal. Therefore, as soon as the tide begins to rise, the flow in the inlet turns to flood.

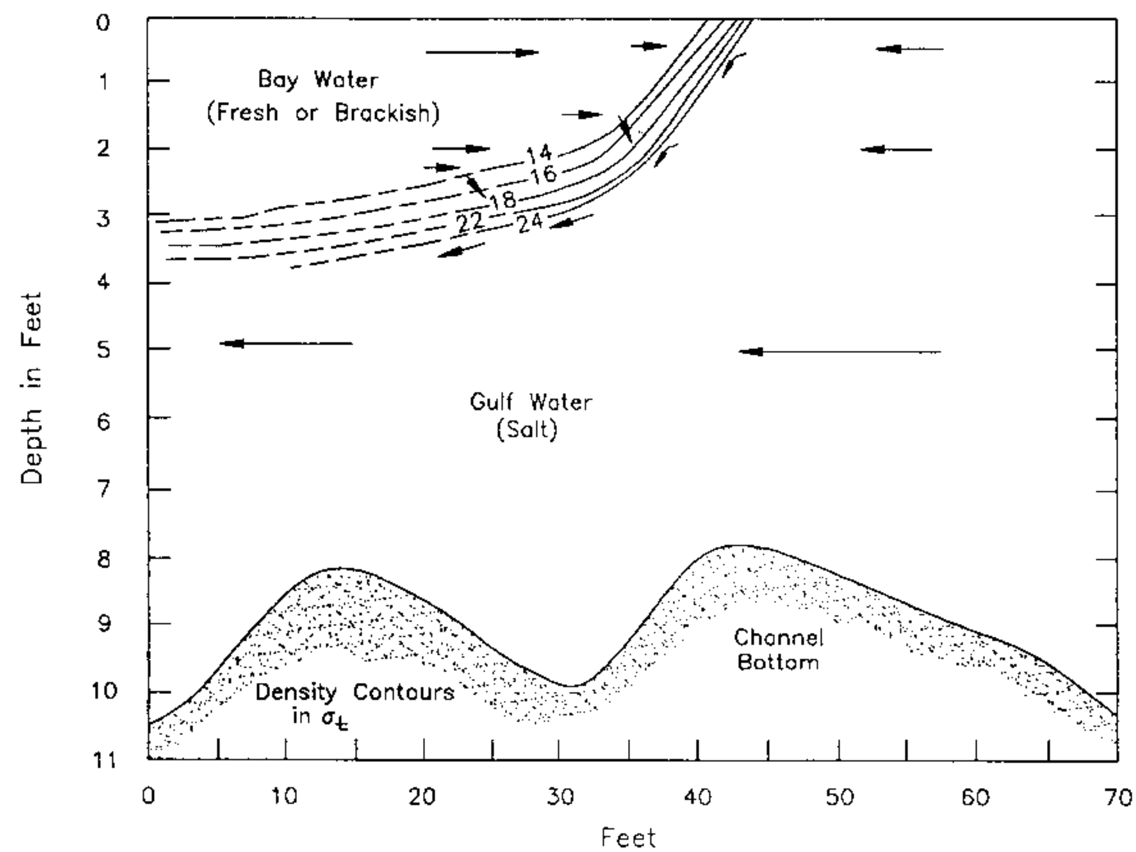

Figure 4-14. Stratified flow occurs during flood tide in Choctawhatchee Bay, Florida, as a wedge of sea water dives underneath the lower density bay water (after Wright, Sonu, and Kielhorn 1972). A similar phenomenon often occurs in estuaries

(g) Velocity asymmetry. The process described above results in a flood that is longer in duration than the ebb. As a result, average ebb velocity must be greater than flood. In addition, because of freshwater input, the total ebb volume may be greater that the flood, contributing to even higher velocities. Volumetric and velocity ebb dominance have been recorded at St. Marys Inlet and East Pass, Florida (Morang 1992).

(h) Net sediment movement. At Price Inlet, South Carolina (FitzGerald and Nummedal 1983) and North Inlet, South Carolina (Nummedal and Humphries 1978), because of peak ebb currents, the resulting seaward-directed sediment transport far exceeded the sediment moved landward during flood. However, ebb velocity dominance does not necessarily mean that net sediment movement is also seaward. At Sebastian Inlet, on Florida's east coast, Stauble et al. (1988) found that net sediment movement was landward although the tidal hydraulics displayed higher ebb currents. The authors concluded that sediment carried into the inlet with the flood tide was deposited on the large, and growing, flood shoal. During ebb tide, current velocities over the flood shoal were too low to remobilize as much sediment as had been deposited on the shoal by the flood tide. The threshold for sediment transport was not reached until the flow was in the relatively narrow throat. In this case, the shoal had become a sink for sediment carried into the inlet. Stauble et al. hypothesized that this pattern of net landward sediment movement, despite ebb hydraulic dominance, may occur at other inlets in microtidal shores that open into large lagoons.

d. Geomorphology of tidal inlets. Tidal inlets are characterized by large sand bodies that are deposited and shaped by tidal currents and waves. The *ebb-tide shoal* (or delta) is a sand mass that accumulates seaward of the mouth of the inlet. It is formed by ebb tidal currents and is modified by wave action. The *flood-tide shoal* is an accumulation of sand at the landward opening of an inlet that is mainly shaped by flood currents (Figure 4-15). Depending on the size and depth of the bay, an ebb shoal may extend into open water or may merge into a complex of meandering tributary channels, point bars, and muddy estuarine sediments.

(1) Ebb-tidal deltas (shoals).

(a) A simplified morphological model of a natural (unjettied) ebb-tidal delta is shown in Figure 4-15. The delta is formed from a combination of sand eroded from the gorge of the inlet and sand supplied by longshore currents. This model includes several components:

- A main *ebb channel*, scoured by the ebb jets.

- *Linear* bars that flank the main channel, the result of wave and tidal current interaction.

- A *terminal lobe*, located at the seaward (distal) end of the ebb channel. This is the zone where the ebb jet velocity drops, resulting in sediment deposition (the expanding lobe shown in Figure 4-11).

- *Swash platforms*, which are sand sheets located between the main ebb channel and the adjacent barrier islands.

- *Swash bars* that form and migrate across the swash platforms because of currents (the swash) generated by breaking waves.

- *Marginal flood channels*, which flank both updrift and downdrift barriers.

Inlets with jetties often display these components, although marginal flood channels are usually lacking.

(b) For the Georgia coast, Oertel (1988) described a simple model of ebb-delta shape and orientation which depended on the balance of currents (Figure 4-16). With modifications, these models could apply to most inlets. When longshore currents were approximately balanced and flood currents exceeded ebb, a squat, symmetrical delta developed (Figure 4-16a) (example: Panama City, FL). If the prevailing longshore currents exceeded the other components, the delta developed a distinct northerly or southerly orientation (Figures 4-16b and 4-16c). Note that some of the Georgia ebb deltas change their orientation seasonally, trending north for part of the year and south for the rest. Finally, when inlet currents exceeded the forces of longshore currents, the delta was narrower and extended further out to sea (Figure 4-16d) (example: Brunswick, GA).

(c) Based on studies of the German and Georgia bights, Nummedal and Fischer (1978) concluded that three factors were critical in determining the geometry of the inlet entrance and the associated sand shoals:

- Tide range.

- Nearshore wave energy.

- Bathymetry of the back-barrier bay.

For the German and Georgia bights the latter factor controls velocity asymmetry through the inlet gorge, resulting in greater seaward-directed sediment transport through the inlet than landward transport. This factor has aided the development of large ebb shoals along these coasts.

(d) The ebb-tidal deltas along mixed-energy coasts (e.g., East and West Friesian Islands of Germany, South Carolina, Georgia, Virginia, and Massachusetts) are huge reservoirs of sand. FitzGerald (1988) states that the amount of sand in these deltas is comparable in volume to that of the adjacent barrier islands. Therefore, on mixed-energy coasts, minor changes in volume of an ebb delta can drastically affect the supply of sand to the adjacent beaches. In comparison, on wave-dominated barrier coasts (e.g., Maryland, Outer Banks of North Carolina, northern New Jersey, Egypt's Nile Delta), ebb-tidal deltas are more rare and therefore represent a much smaller percentage of the overall coastal sand budget. As a result, volumetric changes in the ebb deltas have primarily local effects along the nearby beaches.

(e) Using data from tidal inlets throughout the United States, Walton and Adams (1976) showed that there is a direct correspondence between an inlet's tidal prism and the size of the ebb-tidal delta, with some variability caused by changes in wave energy. This research underscores how important it is that coastal managers thoroughly evaluate whether proposed structures might change the tidal prism, thereby changing the volume of the ebb-tide shoal and, in turn, affecting the sediment budget of nearby beaches.

(f) Ocean City, MD, is offered as an example of the effect of inlet formation on the adjacent coastline: the Ocean City Inlet was formed when Assateague Island was breached by the hurricane of 23 August 1933. The ebb-tide shoal has grown steadily since 1933 and now

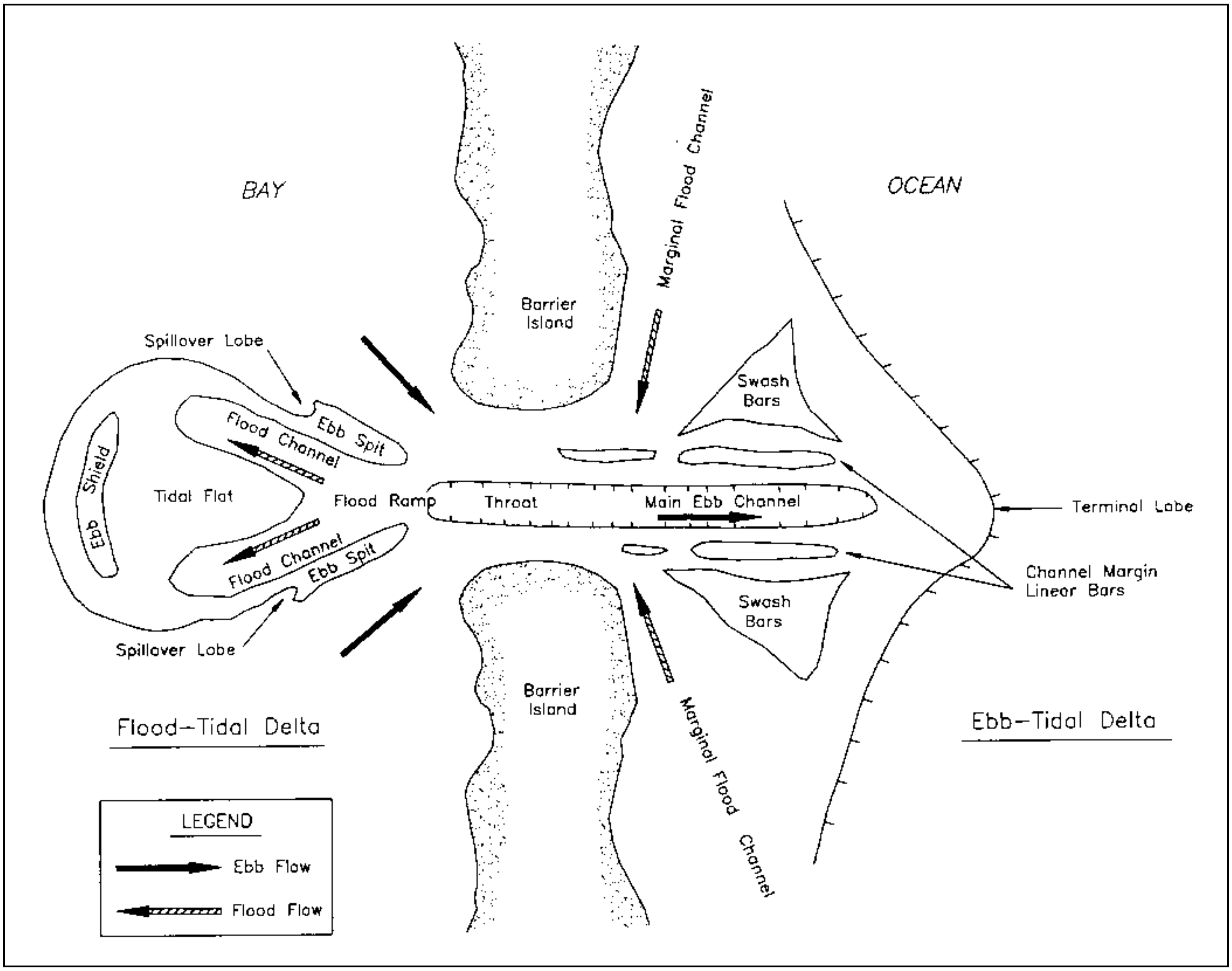

Figure 4-15. Geological model of a tidal inlet with well-developed flood and ebb deltas (from Boothroyd (1985) and other sources)

contains more than 6×10^6 m^3 of sand, located a mean distance of 1,200 m offshore. Since 1933, the growth of the ebb delta combined with trapping of sand updrift of the north jetty have starved the downdrift (southern) beaches, causing the shoreline along the northern few kilometers of Assateague Island to retreat at a rate of 11 m/year (data cited in FitzGerald (1988)).

(g) In contrast to Ocean City, the decrease in inlet tidal prisms along the East Friesian Islands has been beneficial to the barrier coast. Between 1650 and 1960, the area of the bays behind the island chain decreased by 80 percent, mostly due to historic reclamation of tidal flats and marshlands (FitzGerald, Penland, and Nummedal 1984). The reduction in area of the bays reduced tidal prisms, which led to smaller inlets, smaller ebb-tidal shoals, and longer barrier islands. Because of the reduced ebb discharge, less sediment was transported seaward. Waves moved ebb-tidal sands onshore, increasing the sediment supply to the barrier beaches.

(h) In many respects, ebb-tide deltas found at tidal inlets are similar to deltas formed at river mouths. The comparison is particularly applicable at rivers where the flow temporarily reverses during the flood stage of the tide. The main difference between the two settings is that river deltas grow over time, fed by fluvially supplied sediment. In contrast, at many tidal inlets, only limited sediment is supplied from the back bay, and the ebb deltas are largely composed of sand provided by longshore

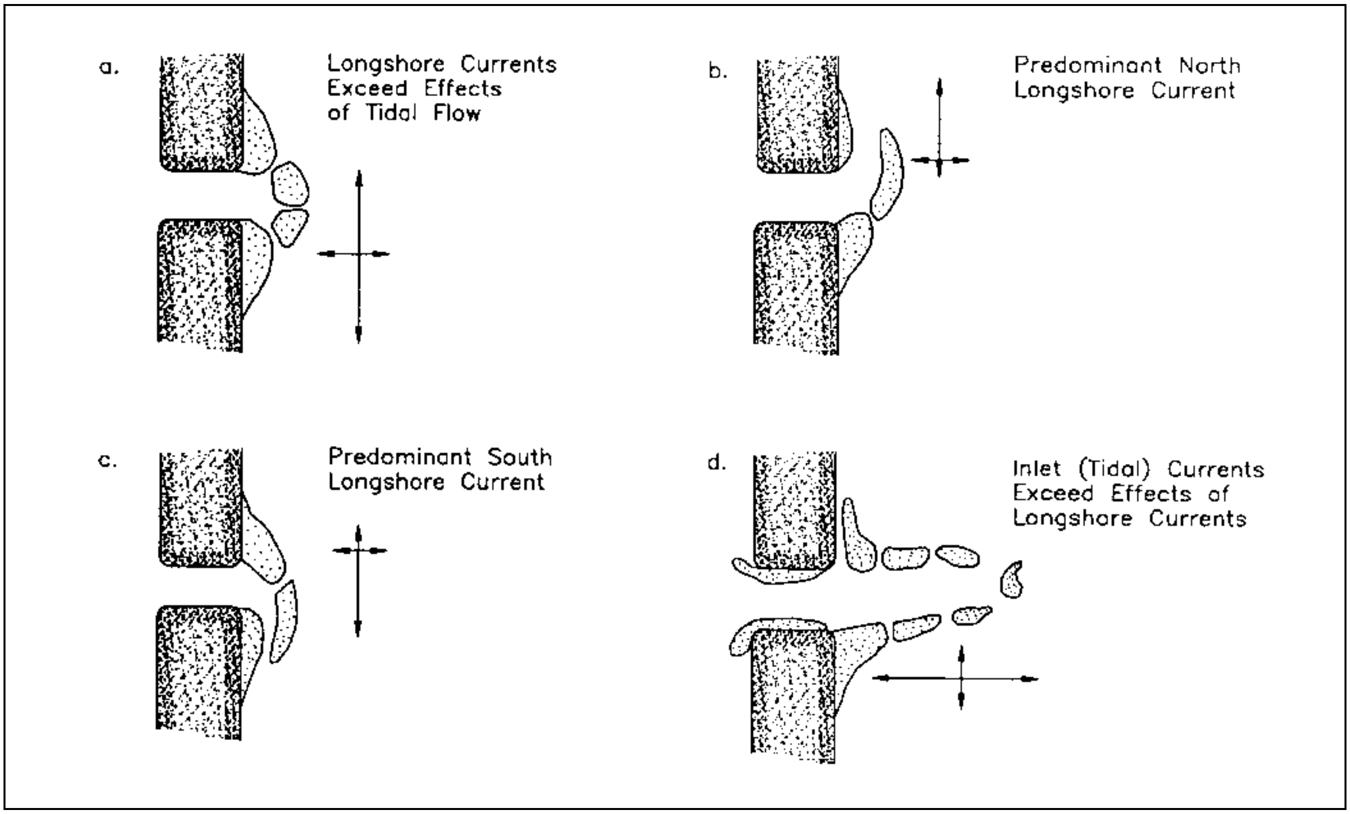

Figure 4-16. Four different shapes of tidal deltas, formed by the relative effects of longshore versus tidal currents (from Oertel (1988))

drift or reworked from the adjacent beaches. Under some circumstances, inlets and river mouths are in effect the same coastal form. During times of low river flow, the mouth assumes the characteristics of a tidal inlet with reversing tidal currents dominating sedimentation. During high river discharge, currents are unidirectional and fluvial sediment is deposited seaward of the mouth, where it can help feed the growth of a delta. Over time, a tidal inlet that connects a pond to the sea can be converted to a river mouth. This occurs when the back bay fills with fluvial sediment and organic matter. Eventually, rivers that formerly drained into the lagoon flow through channels to the inlet and discharge directly into the sea.

(2) Flood-tidal deltas (shoals).

(a) A model of a typical flood-tide shoal is shown in Figure 4-15. Flood shoals with many of these features have been described in meso- and micro-tidal environments around the world (Germany (Nummedal and Penland 1981), Florida's east coast (Stauble et al. 1988), Florida's Gulf of Mexico coast (Wright, Sonu, and Kielhorn 1972), and New England (Boothroyd 1985)). The major components are:

- The *flood ramp*, which is a seaward-dipping sand surface dominated by flood-tidal currents. Sediment movement occurs in the form of sand waves (dunes), which migrate up the ramp.

- *Flood channels*, subtidal continuations of the flood ramp.

- The *ebb shield*, the high, landward margin of the tidal delta that helps divert ebb-tide currents around the shoal.

- *Ebb spits*, high areas mainly formed by ebb currents with some interaction with flood currents.

- *Spillover lobes*, linguoid, bar-like features formed by ebb-tidal current flow over low areas of the ebb shield.

(b) Although this model was originally derived from studies in mesotidal, mixed-energy conditions, it appears to also be applicable to more wave-dominated, microtidal inlets (Boothroyd 1985). However, flood-tide shoals apparently are not formed in macrotidal shores.

(c) The high, central portion of a flood-tidal delta often extends some distance into an estuary or bay. This is the oldest portion of the delta and is usually vegetated by marsh plants. The marsh cap extends up to the elevation of the mean high water. The marsh expands aerially by growing out over the adjacent tidal flat. The highest, marsh-covered part of a flood shoal, or sometimes the entire shoal, is often identified on navigation charts as a "middle ground."

f. Sediment bypassing and inlet stability and migration.[1]

(1) Background. Inlets migrate along the coast - or remain fixed in one location - because of complex interactions between tidal prism, wave energy, and sediment supply. The littoral system is considered by some researchers to be the principal external sediment source that influences the stability of inlets (Oertel 1988). Not all of the sediment in littoral transport is trapped at the mouths of inlets; at many locations, a large proportion may be bypassed by a variety of mechanisms. Inlet sediment bypassing is defined as "the transport of sand from the updrift side of the tidal inlet to the downdrift shoreline" (FitzGerald 1988). Bruun and Gerritsen (1959) described three mechanisms by which sand moves past tidal inlets:

- Wave-induced transport along the outer edge of the ebb delta (the terminal lobe).
- The transport of sand in channels by tidal currents.
- The migration of tidal channels and sandbars.

They noted that at many inlets, bypassing occurred through a combination of these mechanisms. As an extension of this earlier work, FitzGerald, Hubbard, and Nummedal (1978) proposed three models to explain inlet sediment bypassing along mixed-energy coasts. The models are illustrated in Figure 4-17 and are discussed below.

(2) Inlet migration and spit breaching.

[1] Material in this section has been adapted from FitzGerald (1988).

(a) The first model describes the tendency of many inlets to migrate downdrift and then abruptly shift their course by breaching a barrier spit. The migration occurs because sediment supplied by the longshore current causes the updrift barrier to grow (spit accretion). The growth occurs in the form of low, curved beach ridges, which weld to the end of the spit, often forming a bulbous-tipped spit known as a "drumstick." The ridges are often separated by low, marshy swales. As the inlet becomes narrower, the opposite (downdrift) shore erodes because tidal currents attempt to maintain an opening.

(b) In environments where the back bay is largely filled with marshes or where the barrier is close to the mainland, migration of the inlet causes an elongation of the tidal channel. Over time, the tidal flow between bay and ocean becomes more and more inefficient. Under these conditions, if a storm breaches the updrift barrier, the newly opened channel is a more direct and efficient pathway for tidal exchange. This new, shorter channel is likely to remain open while the older, longer route gradually closes. The breaching is most likely to occur across an area where the barrier has eroded or where some of the inner-ridge swales have remained low. The end result of spit accretion and breaching is the transfer of large quantities of sediment from one side of the inlet to the other. An example of this process is Kiawah River Inlet, SC, whose migration between 1661 and 1978 was documented by FitzGerald, Hubbard, and Nummedal (1978). After a spit is breached and the old inlet closes, the former channel often becomes an elongated pond that parallels the coast.

(c) Several notes apply to the inlet migration model: First, not all inlets migrate. As discussed earlier, some inlets on microtidal shores are ephemeral, remaining open only a short time after a hurricane forces a breach through the barrier. If the normal tidal prism is small, these inlets are soon blocked by littoral drift. Short-lived inlets were documented along the Texas coast by Price and Parker (1979). The composition of the banks of the channel and the underlying geology are also critical factors. If an inlet abuts resistant sediments, migration is restricted (for example, Hillsboro Inlet, on the Atlantic coast of Florida, is anchored by rock reefs). The gorge of deep inlets may be cut into resistant sediment, which also will restrict migration.

(d) Second, some inlets migrate updrift, against the direction of the predominate drift. Three mechanisms may account for updrift migration (Aubrey and Speer 1984):

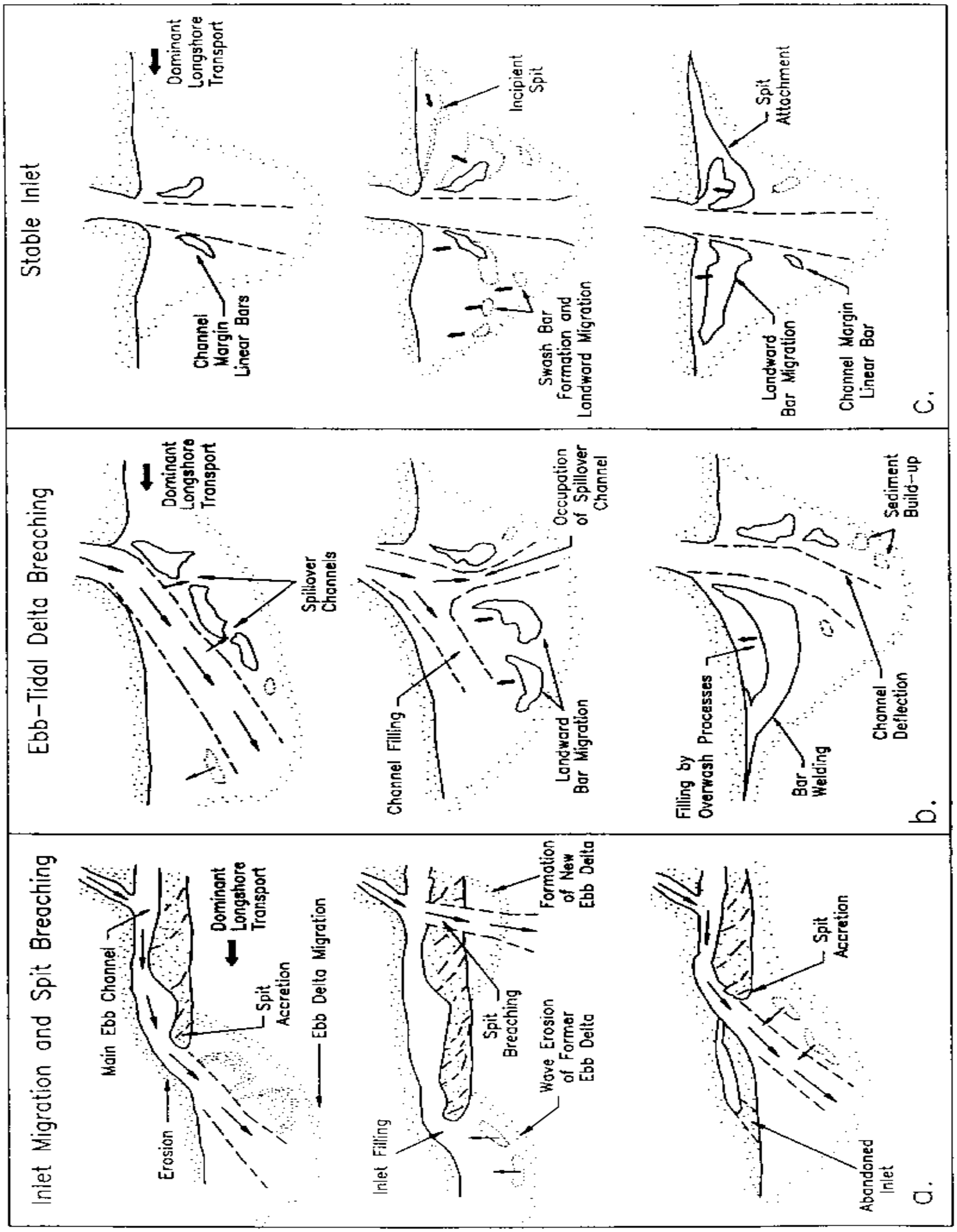

Figure 4-17. Three models of inlet behavior and sediment bypassing for mixed-energy coasts (adapted from FitzGerald (1988))

- Attachment of swash bars to the inlet's downdrift shoreline.

- Breaching of the spit updrift of an inlet.

- Cutbank erosion of an inlet's updrift shoreline caused by back-bay tidal channels that approach the inlet throat obliquely.

(3) Ebb-tidal delta breaching.

(a) At some inlets, the position of the throat is stable, but the main ebb channel migrates over the ebb delta (Figure 4-17b). This pattern is sometimes seen at inlets that are naturally anchored by rock or have been stabilized by jetties. Sediment supplied by longshore drift accumulates on the updrift side of the ebb-tidal delta, which results in a deflection of the main ebb channel. The ebb channel continues to deflect until, in some cases, it flows parallel to the downdrift shore. This usually causes serious beach erosion. In this orientation, the channel is hydraulically inefficient, and the flow is likely to divert to a more direct seaward route through a spillover channel. Diversion of the flow can occur gradually over a period of months or can occur abruptly during a major storm. Eventually, most of the tidal exchange flows through the new channel, and the abandoned old channel fills with sand.

(b) Ebb delta breaching results in the bypassing of large amounts of sand because swash bars, which had formerly been updrift of the channel, become downdrift after the inlet occupies one of the spillover channels. Under the influence of waves, the swash bars migrate landward. The bars fill the abandoned channel and eventually weld to the downdrift beach.

(4) Stable inlet processes.

(a) These inlets have a stable throat position and a main ebb channel that does not migrate (Figure 4-17c). Sand bypassing occurs by means of large bar complexes that form on the ebb delta, migrate landward, and weld to the downdrift shoreline (FitzGerald 1988). The bar complexes are composed of swash bars that stack and merge as they migrate onshore. Swash bars are wave-built accumulations of sand that form on the ebb delta from sand that has been transported seaward in the main ebb channel (Figure 4-15). The swash bars move landward because of the dominance of landward flow across the swash platform. The reason for landward dominance of flow is that waves shoal and break over the terminal lobe (or bar) that forms along the seaward edge of the ebb delta. The bore from the breaking waves augments flood tidal currents but retards ebb currents.

(b) The amount of bypassing that actually occurs around a stable inlet depends upon the geometry of the ebb-tidal shoal, wave approach angle, and wave refraction around the shoal. Three sediment pathways can be identified:

- Some (or possibly much) of the longshore drift accumulates on the updrift side of the shoal in the form of a bar that projects out from the shore (Figure 4-17c). As the incipient spit grows, it merges with growing bar complexes near the ebb channel. Flood currents move some of the sand from the complexes into the ebb channel. Then, during ebb tide, currents flush the sand out of the channel onto the delta (both the updrift and downdrift sides), where it is available to feed the growth of new swash bars.

- Depending on the angle of wave approach, longshore currents flow around the ebb shoal from the updrift to the downdrift side. Some of the drift is able to move past the ebb channel, where it either continues moving along the coast or accumulates on the downdrift side of the ebb shoal.

- Wave refraction around some ebb shoals causes a local reversal of longshore current direction along the downdrift shore. During this time, presumably, little sediment is able to escape the confines of the ebb-tidal shoal.

(5) Extension of bypassing models to other environments. The inlet migration models described above were originally based on moderate- to high-energy shores. However, research along the Florida Panhandle suggests that the models may be applicable to much lower energy environments than the original authors had anticipated. For example, between 1870 and 1990, the behavior of East Pass inlet, located in the low wave-energy, microtidal Florida Panhandle, followed all three models at various times (Figure 4-18; Morang 1992b, 1993). It would be valuable to conduct inlet studies around the world to further refine the models and evaluate their applicability to different shores.

g. Inlet response to jetty construction and other engineering activities.

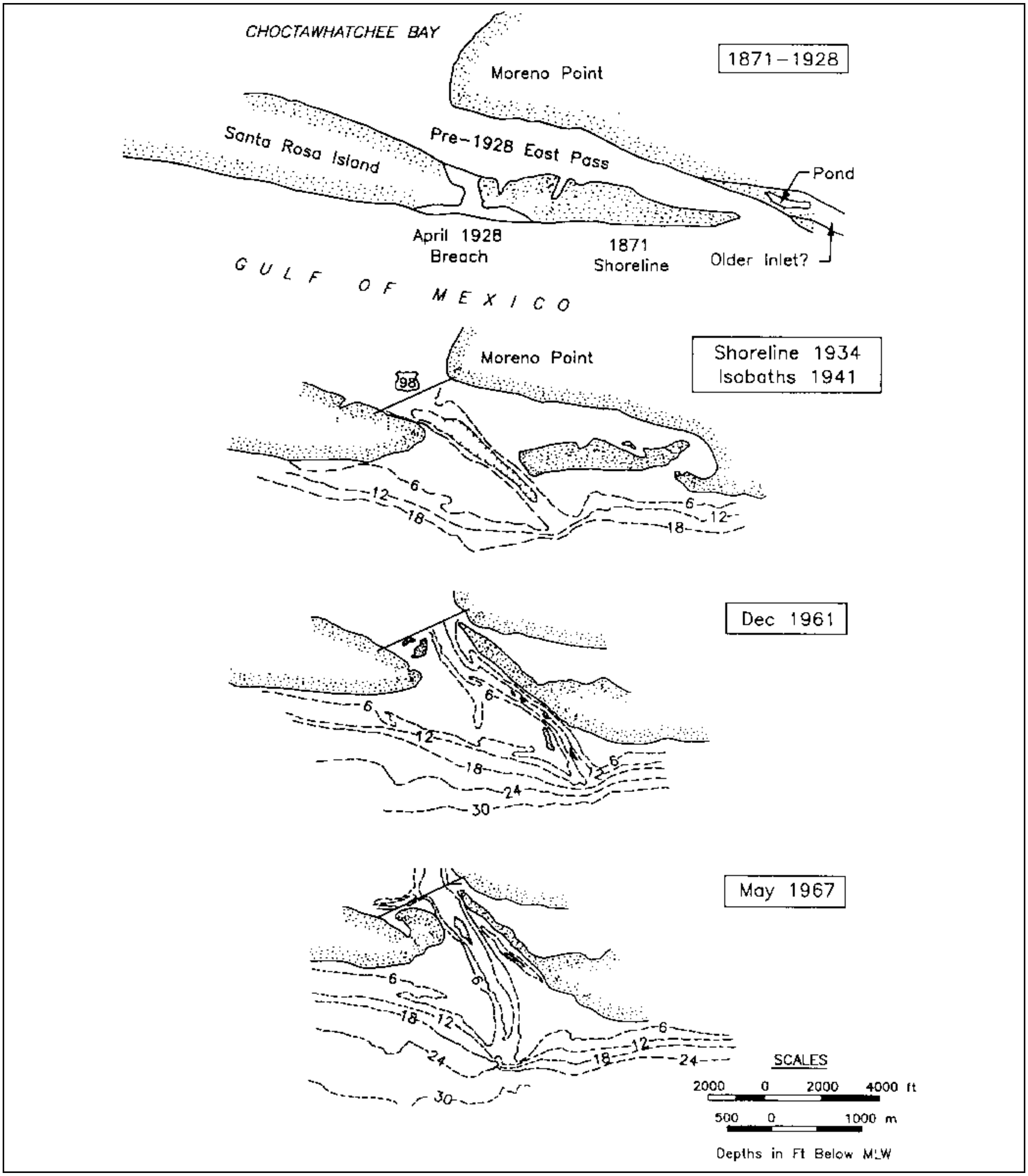

Figure 4-18. Spit breaching and inlet migration at East Pass, Florida (from Morang (1992b))

(1) Introduction. Typically, jetties are built at a site to stabilize a migrating inlet, to protect a navigation channel from waves, or to reduce the amount of dredging required to maintain a specified channel depth. However, jetties can profoundly affect bypassing and other processes at the mouths of inlets. Some of these effects can be predicted during the design phase of a project. Unfortunately, unanticipated geological conditions often arise, which lead to problems such as increased shoaling or changes in the tidal prism. Several classes of man-made activities affect inlets:

- Jetties stabilize inlets and prevent them from migrating.
- Jetties can block littoral drift.
- Walls or revetments can change the cross section of an inlet.
- Dredging can enlarge the cross section of a gorge.
- Dam construction and freshwater diversion reduce fluvial input.
- Weir sections (low portions of a jetty) allow sediment to pass into an inlet, where it can accumulate in a deposition basin and be bypassed.
- Landfilling and development in estuaries and bays can reduce tidal prism.

(2) Technical literatures. Many reports have documented the effects of jetties on littoral sediment transport. Early works are cited in Barwis (1976). Weirs and other structures are discussed in the *Shore Protection Manual* (1984). Dean (1988) discussed the response of modified Florida inlets, and many other case studies are reviewed in Aubrey and Weishar (1988). Examples of monitoring studies conducted to assess the effects of jetties include:

- Ocean City Inlet, Maryland (Bass et al., 1994).
- Little River Inlet, North and South Carolina (Chasten 1992, Chasten and Seabergh 1992).
- Murrells Inlet, South Carolina (Douglass 1987).
- St. Marys Entrance, Florida and Georgia (Kraus, Gorman, and Pope, 1994).
- East Pass, Florida (Morang 1992a).
- Port Mansfield Channel, Texas (Kieslich 1977).

(3) General inlet response.

(a) A model of the response of an ebb-tidal delta to jetty construction is shown in Figure 4-19. The first panel shows a natural inlet in a setting where the predominant drift direction is from right to left. The second panel shows the morphology after the jetties have been completed. At this time, sediment is accumulating on the updrift side of the channel because the updrift jetty (on the right) acts like a groin. As the new ebb delta grows, the abandoned tidal channel fills with sand, and swash bars on the former ebb delta migrate landward. With time, wave action erodes the former ebb delta, particularly if it is out of the sheltering lee of the jetties.

(b) The third panel shows the system after a new ebb delta has formed around the jetties. If the jetties are built across the old delta, then it essentially progrades seaward. If the jetties are built at a different site, then the abandoned ebb delta erodes and disappears while a new delta progrades out from the shore. At some projects, an abandoned ebb delta will disappear within a few years, even on low wave energy shores. The development of a new delta appears to take longer; while the initial growth is rapid, continued adjustment and growth occur for decades. The Charleston Harbor inlet has taken decades to respond to the jetties, which were constructed between 1879 and 1898 (Hansen and Knowles 1988).

(4) Interruption of sediment transport at engineered inlets.

(a) At most sites, the designers of a project must ensure that the structures do not block the littoral drift; otherwise, severe downdrift erosion can occur. Dean (1988) used the phrase “sand bridge” to describe the offshore bar (terminal lobe) across the mouth of most inlets. Net longshore sand transport occurs across the bridge. If the bar is not sufficiently broad and shallow, sediment is deposited until an effective sand bridge is reestablished. Unfortunately, this concept suggests that maintenance of a permanent channel deep enough for safe navigation is usually inconsistent with sediment transport around the entrance by natural processes. Sand bypassing using pumps or dredges can mitigate many of the negative effects of inlet jetties and navigation channels (EM 1110-2-1616).

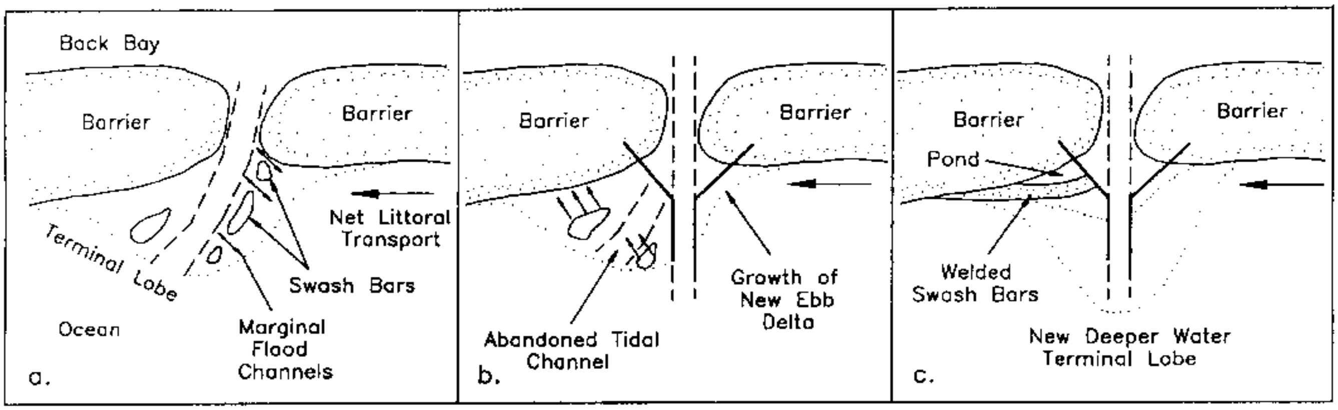

Figure 4-19. Model of the response of an ebb-tidal delta to jetty construction. The final result is development of a new ebb data seaward of the mouth of the jetties in deeper water than the original delta (adapted from Hansen and Knowles 1988)

(b) Dean (1988) also described the "sand sharing system" concept, which states that the sand bodies comprising an inlet, ebb-tidal shoal, and adjacent shorelines are interconnected and in equilibrium. In effect, an ebb shoal is in balance with the local shorelines, and any removal of sand from the shoal lowers the shoal's elevation, thereby causing a flow of sand to restore the local equilibrium. Some of this sand might be eroded from the nearby beaches. Dean (1988) proposed an axiom pertaining to a shoreline sand-sharing system:

> If sand is removed or blocked from a portion of the sand sharing system, the system will respond to restore equilibrium by transporting sand to the deficient area. The adverse erosional effect on the remainder of the system by this removal or blockage is certain, only the timing and degree of its manifestation are in doubt.

(c) Most engineering activities at inlets have some effect on the distribution of sediment. These effects are summarized in Table 4-1 and described in greater detail below.

(d) Storage against updrift jetty. A sand-tight jetty on the updrift side of an inlet will trap sand until the impoundment capacity is reached. If no mechanism has been incorporated into the project to bypass sediment, such as a weir section or a bypassing pumping station, the downdrift shoreline must erode at the same rate as the impoundment at the updrift jetty. This causes a redistribution of sediment, but not a net loss.

(e) Ebb-tidal shoal growth. When an existing inlet is modified by the addition of jetties, the ebb delta is often displaced further seaward to deeper water. The result is that the delta grows greatly in volume. This process may not always occur, depending on tidal prism and wave climate. For example, Hansen and Knowles (1988) concluded that the construction of jetties was eliminating the typical ebb-tidal delta morphology at Murrell's and Little River inlets in South Carolina. In contrast, at East Pass, Florida, the ebb delta has continued to grow seaward beyond the end of the jetties (Morang 1992a).

Table 4-1
Mechanisms Which Affect Sediment Budget of Shorelines Adjacent to Modified (Engineered) Tidal Inlets

Mechanism	Does Mechanism Cause a Net Deficit to Adjacent Shorelines?
1. Storage against updrift jetty	No
2. Ebb tidal shoal growth	Possibly
3. Flood tide shoal growth	Yes
4. Dredge disposal in deep water	Definitely
5. Leaky jetties	Can contribute sediment to nearby shorelines
6. Jetty "shadows"	No
7. Geometric control	No

Note: (From Dean (1988))

(f) Flood-tidal shoal growth. Flood-tide shoals can contain large amounts of sand transported from the adjacent shorelines. Under most circumstances, this sand is lost from the shoreface because there are few natural mechanisms which agitate a flood shoal to a great extent and carry the sand back out to sea. Major rainstorms can raise water elevations in back bays and greatly increase ebb flow, but even under these circumstances, much of the flood shoal is likely to remain. An exception may occur when an inlet is hardened, allowing the prism to increase. If jetties block incoming sand, the system may become sand starved and, over time, much of the flood shoal may be flushed out by the ebb flow.

(g) Dredge disposal in deep water. Until recently, much high-quality sand was dredged from navigation channels and disposed in deep water, where it was lost from the littoral zone. This was an unfortunate practice because beach sand is an extremely valuable mineral resource and is in short supply. Many states now require that all uncontaminated, beach-quality dredged sand be used for beach renourishment.

(h) Leaky jetties. Jetties with high permeability allow sand carried by longshore currents to pass into the channel. Dean (1988) states that this can result in increased erosion of both the updrift and downdrift beaches, whereas sand-tight jetties cause a redistribution, but not a net loss, of sand. However, if material that passes through leaky jetties is dredged and deposited on the adjacent beaches, the erosional impact is minimized. This is similar to the concept of a weir, which allows sand to pass into a deposition basin, where it can be dredged on a regular schedule.

(i) Jetty shadows. Sediment transported around an inlet (both modified and natural), may not reach the shore until some distance downdrift from the entrance. This results in a shadow zone where there may be a deficit of sediment.

(j) Geometric control. This refers to the refraction of waves around an ebb-tidal delta, resulting in local changes to the regional longshore drift pattern. A common result is that for some distance downdrift of a delta, the net drift is reversed and flows towards the delta, while further away from the delta, the drift moves in the opposite direction. The zone of divergence may experience erosion.

h. Summary. This section has discussed some of the many physical processes associated with water flow through tidal inlets. This complex topic has been the subject of a voluminous technical literature, of which it has been possible to cite only a few works. The following are among many interacting processes which affect sedimentation patterns in and near tidal inlets:

- Tidal range.

- Tidal prism - affects quantity of water flowing through the inlet.

- Wave energy - radiation stress drives longshore drift.

- Longshore drift - supplies sediment to vicinity of inlet.

- Fluvial input - affects stratification and sediment supply.

- Man-made intervention - dams upriver reduce sediment and fluvial input; jetties interrupt longshore drift.

- Meteorology - affects offshore water levels.

Recent research at tidal inlets around the world is enhancing our knowledge about these dynamic features of the coastline, but has also made it apparent that there is still much to learn with respect to engineering and management practices.

4-5. Morphodynamics and Shoreface Processes of Clastic Sediment Shores

a. Overview.

(1) Introduction. This section discusses morphodynamics - the interaction of physical processes and geomorphic response - of clastic sediment shores. The topic covers beach features larger than a meter (e.g., cusps and bars) on time scales of minutes to months. Details on grain-to-grain interactions, the initiation of sediment motion, and high frequency processes are not included. A principle guiding this section is that the overall shape of beaches and the morphology of the shoreface are largely a result of oscillatory (gravity) waves, although tide range, sediment supply, and overall geological setting impose limits. We introduce basic relationships and formulas, but the text is essentially descriptive. A brief introduction to waves has been presented in Chapter 2, Paragraph 2-5*b*; Chapter 5, Paragraph 5-5 gives details on the use of wave records.

(2) Literature. Beaches and sediment movement along the shore have been subjects of popular and scientific interest for over a century. A few of the many textbooks that cover these topics include Carter (1988), Davis (1985), Davis and Ethington (1976), Greenwood and Davis (1984), Komar (1976), and Zenkovich (1967). Small-amplitude (Airy) and higher-order wave mechanics are covered in EM 1110-2-1502; more detailed treatments are in Kinsman (1965), Horikawa (1988), and

Le Méhauté (1976). Interpreting and applying wave and water level data are covered in EM 1110-2-1414.

(3) Significance of clastic coasts. It is important to examine and understand how clastic shores respond to changes in wave climate, sediment supply, and engineering activities for economic and management reasons:

- Beaches are popular recreation areas.

- Beaches are critical buffer zones protecting wetlands and coastal plains from wave attack.

- Many people throughout the world live on or near beaches.

- Much engineering effort and expense are expended on planning and conducting beach renourishment.

- Sediment supply and, therefore, beach stability, is often adversely affected by the construction of navigation structures.

- Sand is a valuable mineral resource throughout most of the coastal United States.

(4) Geologic range of coastal environments. Around the world, the coasts vary greatly in steepness, sediment composition, and morphology. The most dynamic shores may well be those composed of unconsolidated clastic sediment because they change their form and state rapidly. Clastic coasts are part of a geologic continuum that extends from consolidated (rocky) to loose clastic to cohesive material (Figure 4-20). Waves are the primary mechanism that shape the morphology and move sediment, but geological setting imposes overall constraints by controlling sediment supply and underlying rock or sediment type. For example, waves have little effect on rocky cliffs; erosion does occur over years, but the response time is so long that rocky shores can be treated as being geologically controlled. At the other end of the continuum, cohesive shores respond very differently to wave action because of the electro-chemical nature of the sediment.

b. Tide range and overall beach morphology.

Most studies of beach morphology and processes have concentrated on microtidal (< 1 m) or low-mesotidal coasts (1-2 m). To date, many details concerning the processes that shape high-meso- and macrotidal beaches (tide range > 2 m) are still unknown. Based on a review of the literature, Short (1991) concluded that wave-dominated beaches where tide range is greater than about 2 m behave differently than their lower-tide counterparts. Short underscored that high-tide beaches are also molded by wave and sediment interactions. The difference is the increasing impact of tidal range on wave dynamics, shoreface morphodynamics, and shoreline mobility. Short developed a tentative grouping of various beach types (Figure 4-21). Discussion of the various shoreface morphologies follows: Section 4-5c describes coasts with tide range greater than about 2 m. Low tide-range shores, described by a model presented by Wright and Short (1984), are discussed in Section 4-5d.

c. High tidal range (> 2 m) beach morphodynamics.

(1) Review. Based on a review of earlier research on macrotidal beaches, Short (1991) summarized several points regarding their morphology:

- They are widespread globally, occurring in both sea and swell environments.

- Incident waves dominate the intertidal zone.

- Low-frequency (infragravity) standing waves may be present and may be responsible for multiple bars.

- The intertidal zone can be segregated into a coarser, steeper, wave-dominated high tide zone, an intermediate zone of finer sediment and decreasing gradient, and a low-gradient, low-tide zone. The highest zone is dominated by breaking waves, the lower two by shoaling waves.

- The cellular rip circulation and rhythmic topography that are so characteristic of micro-tidal beaches have not been reported for beaches with tide range greater than 3 m.

(2) Macrotidal beach groups. Using published studies and field data from Australia, Short (1991) divided macrotidal beaches into three groups based on gradient, topography, and relative sea-swell energy:

(a) GROUP 1 - High wave, planar, uniform slope. Beaches exposed to persistently high waves (H_b > 0.5 m) display a planar, flat, uniform surface (Figure 4-21). Shorefaces are steep, ranging from 1 to 3 deg, and have a flat surface without ripples, bed forms, or bars. The upper high tide beach is often relatively steep and cuspid and contains the coarsest sediment of the system. On

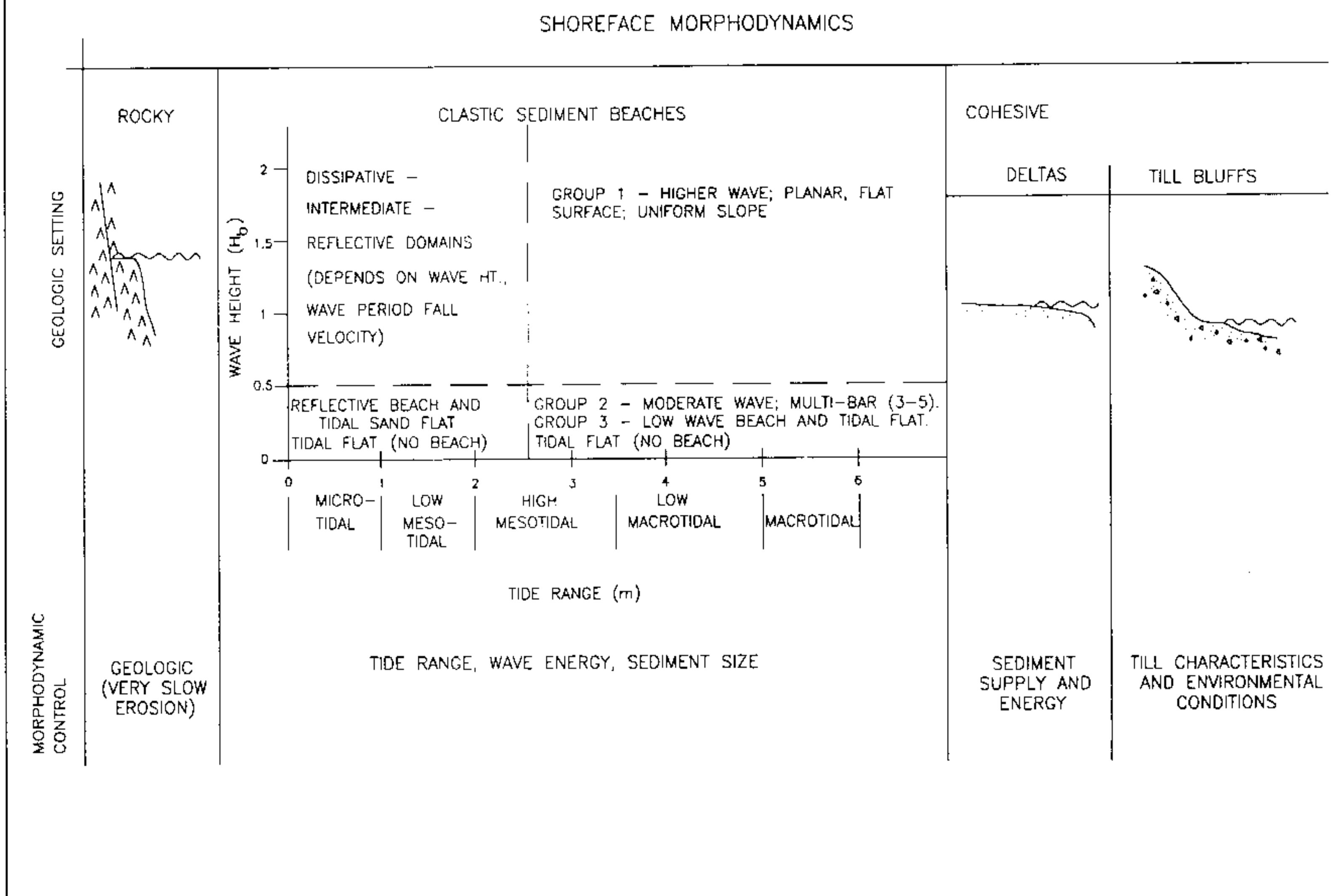

Figure 4-20. Summary of factors controlling morphodynamics along a range of coastal types. Clastic shore morphodynamics are detailed in Figure 4-21 and discussed in the text

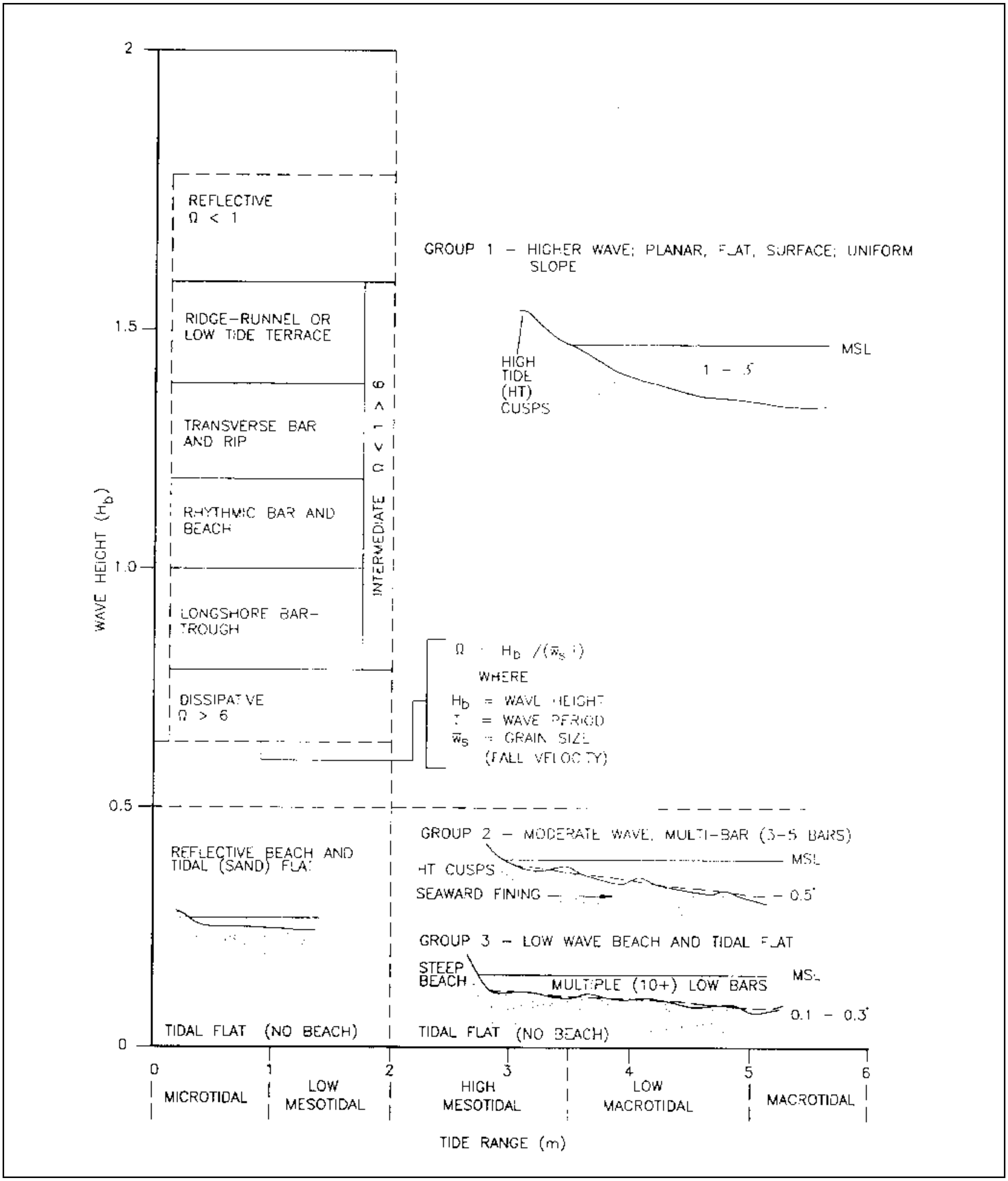

Figure 4-21. Micro- to macrotidal beach and tidal flat systems (adapted from Short (1991)). Dimensionless parameter Ω discussed in the text

both sand and gravel beaches, the high tide, upper foreshore zone is exposed to the highest waves. Plunging and surging breakers produce asymmetric swash flows, which maintain the coarse sediment and steep gradient. Further seaward, wave shoaling becomes a more important factor than wave breaking because waves are attenuated at low tide (due to shallower water and greater friction). Tidal currents also increase in dominance seaward. Wright (1981) found that tidal currents left no bed forms visible at low tide but were an important factor in longshore sediment transport.

(b) GROUP 2 - Moderate wave, multi-bar. Multi-bar, macrotidal beaches are formed in fetch-limited environments with high tide range and abundant fine sand (King 1972). The common characteristic of these beaches is a relatively uniform 0.5- to 0.6-deg intertidal gradient and the occurrence of multiple bars (two to five sets) between msl and mlw (Short 1991). Bar amplitude is usually below 1 m and spacing ranges from 50 to 150 m, with spacing increasing offshore. Field observations indicate that the bars are formed by a wave mechanism, particularly during low wave, post-storm conditions. The bars appear to build up onsite rather than migrate into position. These multi-bar beaches probably cause dissipative conditions during most wave regimes, possibly resulting in the development of infragravity standing waves. This would account for the spacing of the bars; however, this hypothesis has not been tested with rigorous field measurements (Short 1991).

(c) GROUP 3 - Low wave beach and tidal flat. As wave energy decreases, macro-tidal beaches eventually grade into tide-dominated tidal flats. Between the two regimes, there is a transition stage that contains elements of both morphologies. These beach-tidal flat systems are usually characterized by a steep, coarse-grained reflective beach (no cusps usually present) which grades abruptly at some depth below msl into a fine-grained, very low gradient (0.1 deg), rippled tidal flat. The tidal flat may be uniform or may contain low, multiple bars. Beach-tidal flat shores are found in low-energy environments that are only infrequently exposed to wave attack, but the energy must be sufficient to produce the morphologic zonation.

(3) Spatial and temporal variations. Beaches on macro-tidal coasts vary morphologically as important environmental parameters change. Short (1991) cites one setting where the shoreface varies from high-energy, uniform steep beach (Group 1) to beach-tidal flat (Group 3) within 2 km. He suggests that the changes in morphology are due to variations in wave energy: as energy changes alongshore, important thresholds are crossed which result in different ratios of wave versus tide domination. In addition, there may be temporal variations throughout the lunar cycle. As tide range varies during the month, the transitions where one morphologic group merges into another may migrate cyclically along the coast. More field studies are needed to document this phenomenon.

(4) Summary. On tideless beaches, morphology is determined by waves and sediment character. On microtidal beaches, waves still dominate the morphodynamics, but tide exerts a greater influence. As tide range increases beyond 2-3 m, the shape of beaches becomes a function of waves coupled with tides. On the higher tide coasts, as water depth changes rapidly throughout the day, the shoreline and zone of wave breaking move horizontally across the foreshore and tidal currents move considerable sediment.

d. Morphodynamics of micro- and low-mesotidal coasts.

(1) Morphodynamic variability of microtidal beaches and surf zones. Based on field experiments in Australia, Wright and Short (1984) have presented a model of shoreface morphology as a function of wave parameters and sediment grain size. This model is a subset of Figure 4-21 that occupies the zone where tide range is between 0 and 2 m and H_b (breaker height) is greater than about 0.5 m.

(a) Wright and Short (1984) determined that the morphodynamic state of sandy beaches could be classified on the basis of assemblages of depositional forms and the signatures of associated hydrodynamic processes. They identified two end members of the morphodynamic continuum:

- Fully dissipative.

- Highly reflective. Between the extremes were four intermediate states, each of which possessed both reflective and dissipative elements (Figure 4-22).

(b) The most apparent differences between the beach states are morphological, but distinct process signatures, representing the relative velocities of different modes of fluid motion, accompany the characteristic morphology. As stated by Wright and Short (1984):

> Although wind-generated waves are the main source of the energy which drives beach changes, the complex processes, which operate in

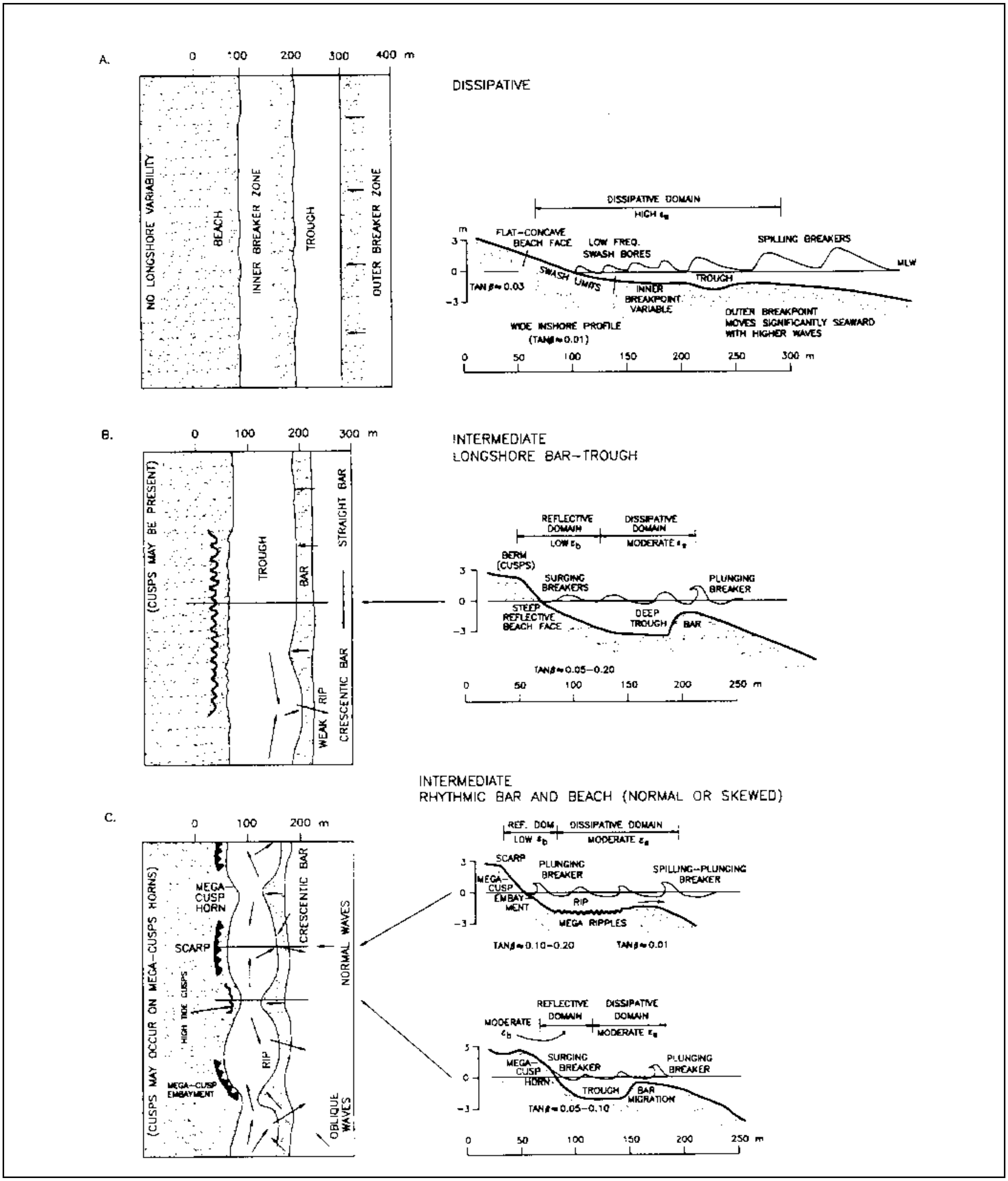

Figure 4-22. Plan and profile views of six major beach stages (adapted from Wright and Short (1984)). Surf-scaling parameter ε is discussed in the text; β represents beach gradient. Dimensions are based on Australian beaches, but morphologic configurations are applicable to other coastlines (Continued)

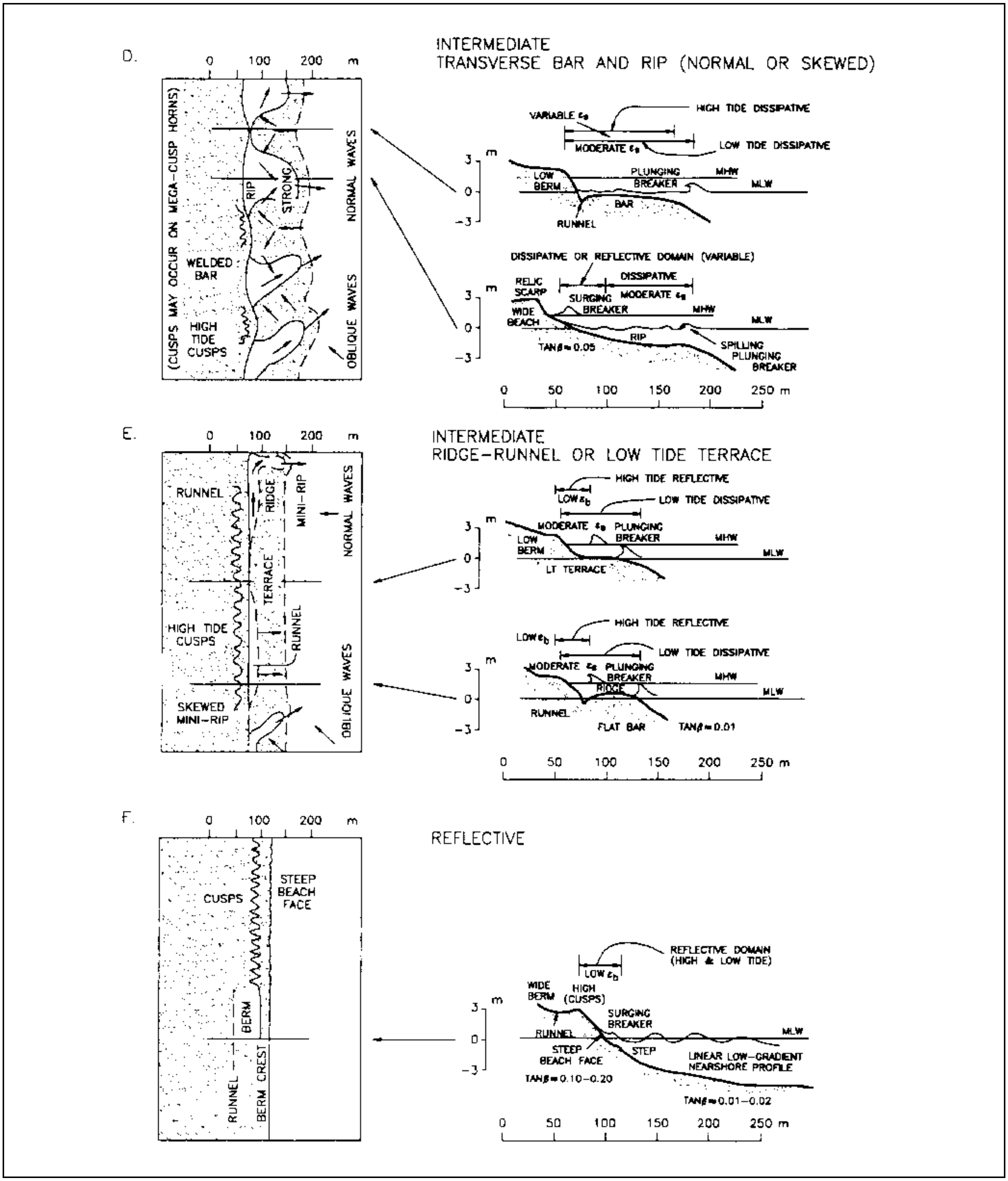

Figure 4-22. (Concluded)

natural surf zones and involve various combinations of dissipation and reflection, can lead to the transfer of incident wave energy to other modes of fluid motion, some of which may become dominant over the waves themselves.

Wright and Short grouped fluid motion into four categories (Table 4-2):

- Oscillatory flows.
- Oscillatory or quasi-oscillatory flows.
- Net circulations.
- Non-wave-generated currents.

(c) From repeated observations and surveys of beaches, Wright and Short (1984) concluded that beach state is clearly a function of breaker height and period and sediment size. Over time, a given beach tends to exhibit a *modal* or most frequent recurrent state, which depends on environmental conditions. Variations in shoreline position and profile are associated with temporal variations of beach state around the modal state. Wright and Short found that a dimensionless parameter Ω could be used to describe the modal state of the beach:

$$\Omega = \frac{H_b}{\overline{w_s} T} \tag{4-3}$$

where H_b is breaker height, $\overline{w_s}$ is sediment fall velocity, and T is wave period. A value of Ω about 1 defines the reflective/intermediate threshold; for intermediate beaches, $1 < \Omega < 6$; $\Omega \sim 6$ marks the threshold between intermediate and dissipative conditions (Figure 4-22).

Table 4-2
Modes of Fluid Motion Affecting Clastic Shorelines

Modes	Notes	Frequencies of flows	Examples
Oscillatory	Corresponds directly to incident waves	Frequency band of deep-water incident waves	Sediment-agitating oscillations
Oscillatory or quasi-oscillatory	Shore-normal oriented standing and edge waves	Wide range of frequencies	Trapped edge waves, "leaky" mode standing waves
Net circulations	Generated by wave energy dissipation	Minutes to days	Longshore currents, rip currents, rip feeder currents
Non-wave-generated currents	Generated by tides and wind shear	Minutes to hours (?)	Tidal currents

(Based on Wright and Short (1984))

(d) Beaches take time to adjust their state, and a change of Ω across a threshold boundary does not immediately result in a transformation from reflective to intermediate or from intermediate to dissipative. On the Pacific coasts of Australia and the United States, storms can cause a shift of beach state from reflective or intermediate to dissipative in a few days because the energy is high. The return to reflective conditions under low energy may require weeks or months or longer (the sequence of beach recovery is illustrated in stages a through f in Figure 4-22). In environments where the dominant variation in wave energy occurs on an annual cycle (e.g., high storm waves in winter and low swell in summer), the full range from a dissipative winter profile to a reflective summer profile may be expected.

(e) Wright and Short (1984) concluded that, in general, large temporal variations in Ω are accompanied by large changes in state. However, when the variations in Ω take place in the domains of $\Omega < 1$ or $\Omega > 6$, no corresponding changes in *state* result. Intermediate beaches, where Ω is between 1 and 6, are spatially and temporally the most dynamic. They can undergo rapid changes as wave height fluctuates, causing reversals in onshore/offshore and alongshore sediment transport.

(f) The parameter Ω depends critically upon $\overline{w_s}$, the sediment fall velocity. It is unclear how the relationships described above apply to shorefaces where the grain size varies widely or where there is a distinct bimodal distribution. For example, many Great Lakes beaches contain material ranging in size from silt and clay to cobble several centimeters in diameter. During storms, not only do wave height and period change, but fine-grain sediment is preferentially removed from the shoreface; therefore, the effective $\overline{w_s}$ may change greatly within a few hours.

Further research is needed to understand how Great Lakes beaches change modally and temporally.

(2) Highly dissipative stage (Figure 4-22a). The dissipative end of the continuum is analogous to the "storm" or "winter beach" profile described by Bascom (1964) for shores that vary seasonally. The characteristic feature of these beaches is that waves break by spilling and dissipating progressively as they cross a wide surf zone, finally becoming very small at the upper portion of the foreshore (Figure 4-23) (Wright and Short 1984). A dissipative surf zone is broad and shallow and may contain two or three sets of bars upon which breakers spill. Longshore beach variability is minimal.

(3) Highly reflective stage (Figure 4-22f). On a fully reflective beach, breakers impinge directly on the shore without breaking on offshore bars (Figures 4-24, 4-25). As breakers collapse, the wave uprush surges up a steep foreshore. At he bottom of the steep, usually linear beach is a pronounced step composed of coarser material. Seaward of the step, the slope of the bed decreases appreciably. Rhythmic beach cusps are often present in the swash zone. The fully reflective stage is analogous to the fully accreted "summer profile."

(4) Surf-scaling parameter. Morphodynamically, the two end members of the beach state model can be distinguished on the basis of the surf-scaling parameter (Guza and Inman 1975):

$$\varepsilon = \frac{a_b \omega^2}{g \tan^2 \beta} \tag{4-4}$$

where

a_b = breaker amplitude

ω = incident wave radian energy ($2\pi/T$ where T = period)

g = acceleration of gravity

β = the gradient of the beach and surf zone

Figure 4-23. Example of a dissipative beach: Southern California near San Diego

Figure 4-24. Example of a reflective sand beach: Newport Beach, CA, April, 1993

Strong reflection occurs when $\varepsilon \leq 2.0\text{-}2.5$; this situation defines the highly reflective extreme. When $\varepsilon > 2.5$, waves begin to plunge, dissipating energy. Finally, when $\varepsilon > 20$, spilling breakers occur, the surf zone widens, and turbulent dissipation of wave energy increases with increasing ε.

(5) Intermediate beach stages. These stages exhibit the most complex morphologies and process signatures.

(a) Longshore bar-trough state (Figure 4-22b). This beach form can develop from an antecedent dissipative profile during an accretionary period. Bar-trough relief is higher and the shoreface is much steeper than on the dissipative profile. Initial wave breaking occurs over the bar. However, in contrast to the dissipative beach, the broken waves do not continue to decay after passing over the steep inner face of the bar, but re-form in the deep trough. Low-steepness waves surge up the foreshore; steeper waves collapse or plunge at the base of the foreshore, followed by a violent surge up the subaerial beach (Wright and Short 1984). Runup is relatively high and cusps often occur in the swash zone.

(b) Rhymthic bar and beach (Figure 4-22c). Characteristics are similar to the longshore bar-trough state (described above). The distinguishing features of the rhymthic bar and beach state are the regular longshore undulations of the crescentic bar and of the subaerial beach (Figure 4-26). A weak rip current circulation is often present, with the rips flowing across the narrow portions of the bar. Wright and Short (1984) state that incident waves dominate circulation throughout the surf zone, but subharmonic and infragravity oscillations become important in some regions.

(c) Transverse-bar and rip state (Figure 4-22d). This morphology commonly develops in accretionary sequences when the horns of crescentic bars weld to the beach. This results in dissipative transverse bars (sometimes called “mega-cusps”) that alternate with reflective, deeper embayments. The dominant dynamic process of this beach state is extremely strong rip circulation, with the seaward-flowing rip currents concentrated in the embayments.

(d) Ridge and runnel/low tide terrace state (Figures 4-22e and 3-21). This beach state is characterized by a flat accumulation of sand at or just below the low tide level, backed by a steeper foreshore. The beach is typically dissipative at low tide and reflective at high tide.

Figure 4-25. Example of a reflective cobble beach: Aldeburgh, Suffolk (facing the North Sea), August 1983. Note the steep berm and the lack of sand-sized sediment

e. Processes responsible for shoreface sediment movement.

(1) Despite intense study for over a century, the subject of sand movement on the shoreface is still poorly understood. Sand is moved by a combination of processes including (Pilkey 1993; Wright et al. 1991):

- Wave orbital interactions with bottom sediments and with wave-induced longshore currents.
- Wind-induced longshore currents.
- Turbidity currents.
- Rip currents.
- Tidal currents.
- Storm surge ebb currents.
- Gravity-driven currents.
- Wind-induced upwelling and downwelling.
- Wave-induced upwelling and downwelling.
- Gravity-induced downslope transport.

Figure 4-26. Gravel cusps at St. Joseph, MI, November, 1993. This is an example of a rhymthic bar and beach on a freshwater coast without tides but subject to irregular seiching

Additional complications are imposed by constantly changing shoreface conditions:

- The relative contributions made by the different transport mechanisms vary over time.

- Because of differing regional geological configuration and energy climate, the frequencies of occurrence of the different mechanisms vary with location.

- Oscillatory flows normally occur at many frequencies and are superimposed on mean flows and other oscillatory flows of long period.

(2) Middle Atlantic Bight experiments of Wright et al. (1991).

(a) Wright et al. (1991) measured suspended sediment movement, wave heights, and mean current flows at Duck, NC, in 1985 and 1987 and at Sandbridge, VA, in 1988 using instrumented tripods. During their study, which included both fair weather and moderate energy conditions, onshore mean flows (interpreted to be related to tides), were dominant over incident waves in generating sediment fluxes. In contrast, during a storm, bottom conditions were strongly dominated by offshore-directed, wind-induced mean flows. Wright et al. attributed this offshore directed flow to a rise of 0.6 m in mean water level (during this particular storm) and a resultant strong seaward-directed downwelling flow.

(b) Wright et al. (1991) examined the mechanisms responsible for onshore and offshore sediment fluxes across the shoreface. They related two factors explicitly to incoming incident waves:

- Sediment diffusion arising from gradients in wave energy dissipation.

- Sediment advection caused by wave orbital asymmetries.

They found that four other processes may also play important roles in moving sediment:

- Interactions between groupy incident waves and forced long waves.

- Wind-induced upwelling and downwelling currents.

- Wave-current interactions.

- Turbidity currents.

Overall, Wright et al. found that incoming incident waves were of primary importance in bed agitation, while tide- and wind-induced currents were of primary importance in moving sediment. The incoming wave orbital energy was responsible for mobilizing the sand, but the unidirectional currents determined where the sand was going. Surprisingly, cross-shore sediment fluxes generated by mean flows were dominant or equal to sediment fluxes generated by incident waves in all cases and at all times.

(c) Based on the field measurements, Wright et al. (1991) concluded that "near-bottom mean flows play primary roles in transporting sand across isobaths on the upper shoreface" (p 49). It is possible that this dominance of mean flows is a feature which distinguished the Middle Atlantic Bight from other shorefaces. The oscillatory (wave) constituents may be proportionately much more important along coasts subject to persistent, high-energy swell, such as the U.S. west coast. Wright et al. also concluded that the directions, rates, and causes of cross-shore sediment flux varied temporally in ways that were only partly predictable with present theory.

f. Sea level change and the Bruun rule.

(1) General coastal response to changing sea level.[1] Many barrier islands around the United States have accreted vertically during the Holocene rise in global sea level, suggesting that in these areas the supply of sediment was sufficient to allow the beaches to keep pace with the rise of the sea. It is not clear how beaches respond to short-term variations in sea level. Examples of shorter processes include multi-year changes in Great Lakes water levels and multi-month sea level rises associated with the El Niño-Southern Oscillation in the Pacific.

(2) Storm response.

(a) Based on his pioneering research of southern California beaches in the 1940's, Shepard (1950) developed the classic model that there is an onshore-offshore exchange of sediment over winter-summer cycles. Studies since then have shown that this model applies mostly to beaches on swell-dominated coasts where the wave climate changes seasonally (particularly Pacific Ocean coasts) (Carter 1988). Many beaches do *not* show an obvious seasonal cycle. Instead, they erode during storms throughout the year and rebuild during subsequent fair weather periods.

(b) In some locations, such as the Gulf Coast, infrequent and irregular hurricanes may be the most important dynamic events affecting beaches. Following one of these storms, beach and dune rebuilding may take years (Figure 3-6 shows a portion of the Florida/Alabama shore that was damaged by Hurricane Frederick in 1979 and is slowly recovering). Recently, the popular belief that hurricanes are the most important morphodynamic events causing Gulf Coast beach erosion is being reevaluated with the benefit of new field data. Scientists have learned that, cumulatively, winter cold fronts produce significant annual barrier island retreat. Dingler, Reiss, and Plant (1993) monitored Louisiana's Isles Dernieres and found that Hurricane Gilbert (September 1988) produced substantial beach retreat initially, but it actually reduced the average erosion rate by modifying the slope of the shoreface from that produced by cold-front-generated storms. The different responses were related to the scale of the storms. Cold fronts, which individually were small storms, eroded the entire beach to the same degree. Most sand and mud was deposited offshore and only a small percentage of eroded sand was deposited on the backshore because the fronts usually did not raise the sea enough to cause overtopping. Hurricane Gilbert, in contrast, raised sea level substantially such that the primary erosion occurred on the upper beach, and much of the sand was deposited behind the island via overwash processes. Over a five-year period, the overall effect of this hurricane on the Isles Dernieres was to retard the retreat rate of the island by about 50 percent over that produced by cold fronts alone.

(3) Bruun Rule beach response model.

(a) One of the best-known shoreface response models was proposed by Bruun in 1962 (rederived in Bruun (1988)). Bruun's concept was that beaches adjust to the dominant wave conditions at the site. He reasoned that beaches had to respond in some manner because clearly they had adjusted and evolved historically as sea level had

[1] Chapter 2 reviewed sea level change and outlined some of the associated coastal effects and management issues. Table 2-6 outlined how shoreline advance or retreat at any particular location is a balance between sediment supply and the rate of sea level change. In this section, sea level change is meant in a general sense to be caused by a combination of factors, including eustatic (global) changes and local effects due to vertical movements of the coastal land.

changed. Beaches had not disappeared, they had moved. How was this translation accomplished? Earlier studies of summer/winter beach morphology provided clues that beaches responded even to seasonal changes in wave climate. The basic assumption behind Bruun's model is that with a rise in sea level, the equilibrium profile of the beach and the shallow offshore moves upward and landward. Bruun made several assumptions in his two-dimensional analysis:

- The upper beach erodes because of a landward translation of the profile.

- Sediment eroded from the upper beach is deposited immediately offshore; the eroded and deposited volumes are equal (i.e., longshore transport is not a factor).

- The rise in the seafloor offshore is equal to the rise in sea level. Thus, offshore the water depth stays constant.

(b) The Bruun Rule can be expressed as (Figure 4-27a):

$$R = \frac{L_*}{B + H_*} S \tag{4-5}$$

where

R = shoreline retreat

S = increase in sea level

L_* = cross-shore distance to the water depth H_*

B = berm height of the eroded area

Hands (1983) restated the Bruun Rule in simplified form:

$$x = \frac{zX}{Z} \tag{4-6}$$

where z is the change in water level. The ultimate retreat of the profile x can be calculated from the dimensions of the responding profile, X and Z, as shown in Figure 4-27b.

(c) Despite the continued interest in Bruun's concept, there has been only limited use of this method for predictive purposes. Hands (1983) listed several possible reasons for the reluctance to apply this approach:

- Skepticism as to the adequacy of an equilibrium model for explaining short-term dynamic changes.

- Difficulties in measuring sediment lost from the active zone (alongshore, offshore to deep water, and onshore via overwash).

- Problems in establishing a realistic closure depth below which water level changes have no measurable effect on the elevation or slope of the seafloor.

- The perplexity caused by a discontinuity in the profile at the closure depth which appeared in the original and in most subsequent diagrams illustrating the concept.

An additional, and unavoidable, limitation of this sediment budget approach is that it does not address the question of *when* the predicted shore response will occur (Hands 1983). It merely reveals the horizontal distance the shoreline must *ultimately* move to reestablish the equilibrium profile at its new elevation under the assumptions stated in Bruun's Rule.

(d) Hands (1983) demonstrated the geometric validity of the Bruun Rule in a series of figures which show the translation of the profile upward and landward (the figures are two-dimensional; volumes must be based on unit lengths of the shoreline):

- Figure 4-28a: The equilibrium profile at the initial water level.

- Figure 4-28b: The first translation moves the active profile up an amount z and reestablishes equilibrium depths below the now elevated water level. Hands defines the *active profile* as the zone between the closure depth and the upper point of profile adjustment. The volume of sediment required to maintain the equilibrium water depth is proportional to X (width of the active zone) times z (change in water level).

- Figure 4-28c: The required volume of sediment is provided by the second translation, which is a recession (horizontal movement) of the profile by an amount x. The amount of sediment is proportional to x times Z, where Z is the vertical extent of the active profile from the closure depth to the average elevation of the highest erosion on the backshore.

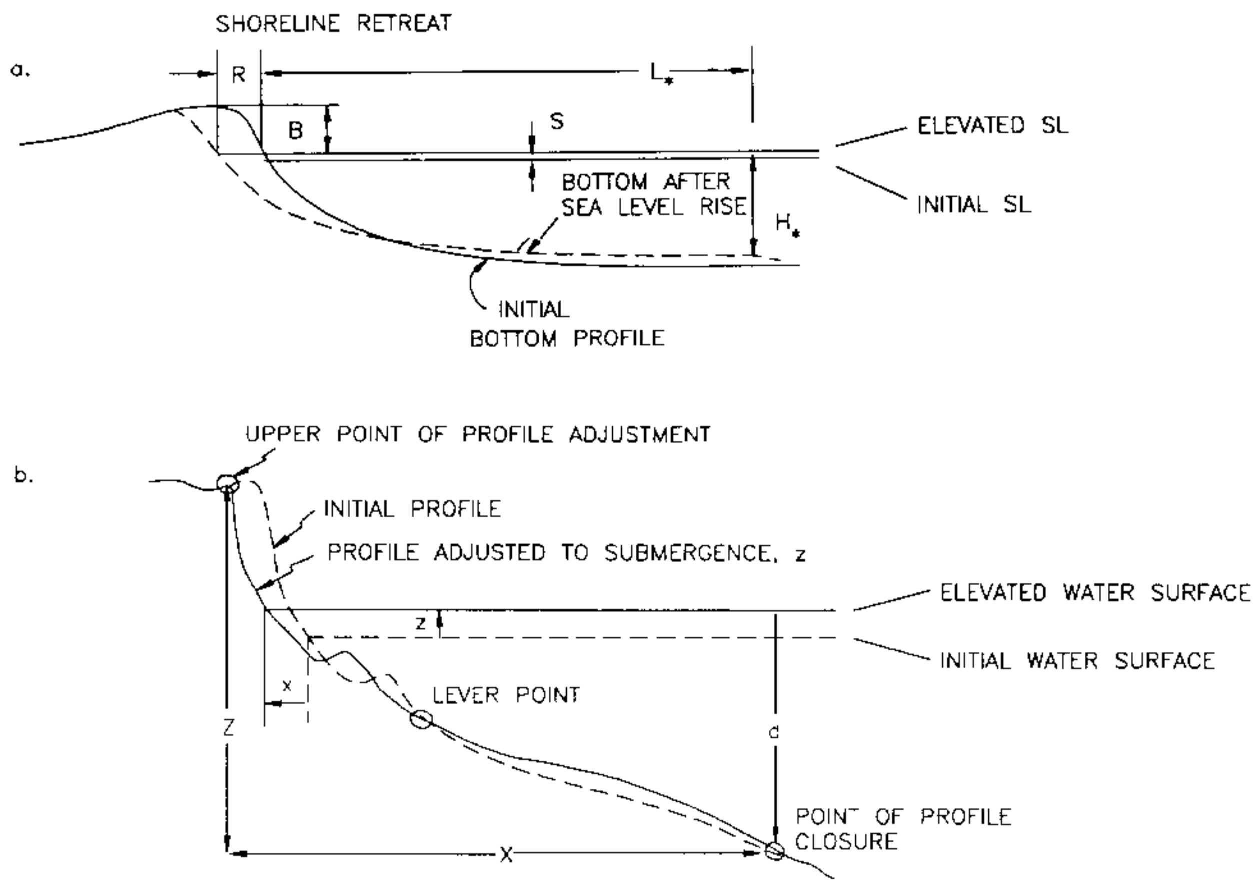

Figure 4-27. (a) Shoreline response to rising sea level (SL) depicted by the Bruun Rule. (b) Simplified nomenclature used by Hands (1983). The sandbar shows that the model is valid for complicated profile shapes

- Figure 4-28d: Equating the volume required by the vertical translation and the volume provided by the horizontal translation yields Equation 4-6. In reality, both translations occur simultaneously, causing the closure point to migrate upslope as the water level rises.

(e) One of the strengths of the Bruun concept is that the equations are valid regardless of the shape of the profile, for example, if bars are present (Figure 4-27b). It is important that an offshore distance and depth of closure be chosen that incorporate the entire zone where active sediment transport occurs. Thereby, sediment is conserved in spite of the complex processes of local erosion versus deposition as bars migrate (Komar et al. 1991). Another strength is that it is a simple relationship, a geometric conclusion based only on water level. Despite its simplicity and numerous assumptions, it works remarkably well in many settings. Even with its shortcomings, it can be used to predict how beaches can respond to changes in sea level.

(4) Use of models to predict shoreline recession. Although field studies have confirmed the assumptions made by Bruun and others concerning translations of the shoreface, there has been no convincing demonstration that the models can predict shoreline recession rates. Komar et al. (1991) cite several reasons for the inability to use the models as predictive tools:

- Existence of a considerable time lag of the beach response following a sustained water level rise (as shown by Hands (1983) for Lake Michigan).

- Uncertainty in the selection of the parameters used in the equations (in particular, closure depth).

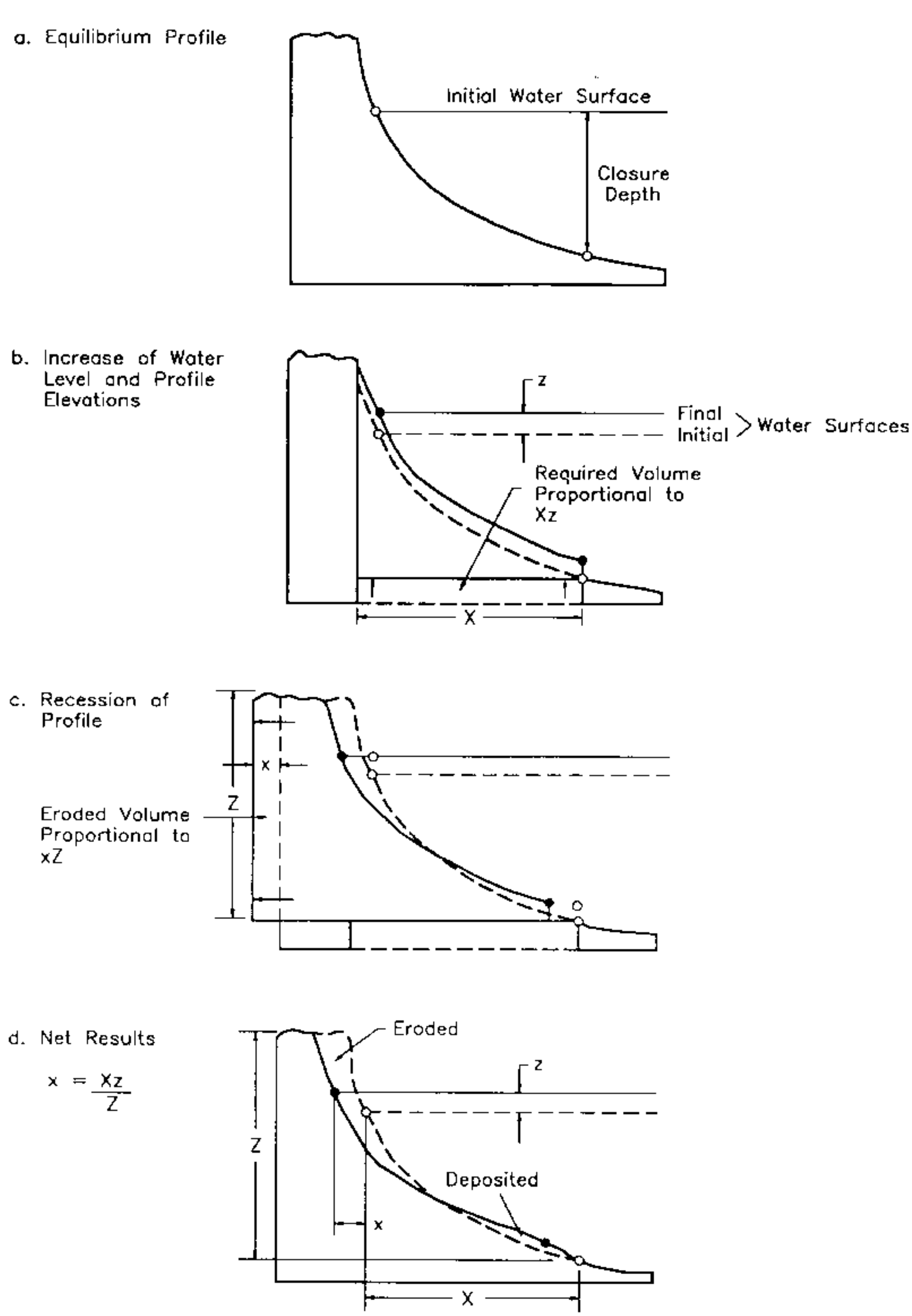

Figure 4-28. Profile adjustment in two stages, first vertical, then horizontal, demonstrating the basis for the Bruun Rule (Equation 4-6) (from Hands (1983)). Details are discussed in the text

- Local complexities of sediment budget considerations in the sand budget.

(5) Recommendations. We need more field and laboratory studies to better evaluate the response of beaches to rising (and falling) sea level. For example, it would be valuable to reoccupy the profile lines monitored by Hands (1976, 1979, 1980) in Lake Michigan in the 1970's to determine how the shores have responded to the high water of the mid-1980's and to the subsequent drop in the early 1990's. In addition, we need conceptual advances in the theoretical models. We also need to evaluate how sediment has moved onshore in some locations following sea level rise, because there is evidence that in some areas beach sand compositions reflect offshore rather than onshore sources (Komar et al. 1991).

g. Equilibrium profiles on sandy coasts.

(1) General characteristics and assumptions. The existence of an equilibrium shoreface profile (sometimes called equilibrium *beach* profile) is a basic assumption of many conceptual and numerical coastal models. Dean (1990) listed characteristic features of profiles:

- Profiles tend to be concave upwards.

- Fine sand is associated with mild slopes and coarse sand with steep slopes.

- The beach (above the surf zone) is approximately planar.

- Steep waves result in milder inshore slopes and a tendency for bar formation.

The main assumption underlying the concept of the shoreface equilibrium profile is that the seafloor is in equilibrium with *average* wave conditions. Presumably, the term *equilibrium* is meant to indicate a situation in which water level, waves, temperature, etc., are held constant for a sufficient time such that the beach profile arrives at a final, stable shape (Larson and Kraus 1989a). Larson (1991) described the profile as: "A beach of specific grain size, if exposed to constant forcing conditions, normally assumed to be short-period breaking waves, will develop a profile shape that displays no net change in time." This concept ignores the fact that, in addition to wave action, many other processes affect sediment transport. These simplifications, however, may represent the real strength of the concept because it has proven to be a useful way to characterize the shape of the shoreface in many locations around the world.

(2) Shape. Based on studies of beaches in many environments, Bruun (1954) and Dean (1976, 1977) have shown that many ocean beach profiles exhibit a concave shape such that the depth varies as the two-thirds power of distance offshore along the submerged portions:

$$h(x) = Ax^{2/3} \tag{4-7}$$

where

h = water depth at distance x from the shoreline

A = a scale parameter which depends mainly on sediment characteristics

This surprisingly simple expression asserts, in effect, that beach profile shape can be calculated from sediment characteristics (particle size or fall velocity) alone. Moore (1982) graphically related the parameter A, sometimes called the *profile shape parameter,* to the median grain size d_{50}. Hanson and Kraus (1989) approximated Moore's curve by a series of lines grouped as a function of the median nearshore grain size d_{50} (in mm):

$$\begin{aligned} A &= 0.41(d_{50})^{0.94} \quad , \quad d_{50} < 0.4 \\ A &= 0.23(d_{50})^{0.32} \quad , \quad 0.4 \le d_{50} < 10.0 \\ A &= 0.23(d_{50})^{0.28} \quad , \quad 10.0 \le d_{50} < 40.0 \\ A &= 0.46(d_{50})^{0.11} \quad , \quad 40.0 \le d_{50} \end{aligned} \tag{4-8}$$

Dean (1987) related the parameter A to the sediment fall velocity w. On a log-log plot, the relationship was almost linear and could be expressed as:

$$A = 0.067\,w^{0.44} \tag{4-9}$$

(3) Discussion of assumptions. Pilkey et al. (1993), in a detailed examination of the concept of the equilibrium shoreface profile, contended that several assumptions must hold true for the concept to be valid:

(a) *Assumption 1: All sediment movement is driven by incoming wave orbitals acting on a sandy shoreface.*

This assumption is incorrect because research by Wright et al. (1991) showed that sediment movement on the shoreface is an exceedingly complex phenomenon, driven by a wide range of wave, tidal, and gravity currents. Even in locations where the wave orbitals are responsible for mobilizing the sand, bottom currents frequently determine where the sand will go.

(b) *Assumption 2: Existence of closure depth and no net cross-shore (i.e., shore-normal) transport of sediment to and from the shoreface.*

Pilkey et al. (1993) state that this assumption is also invalid because considerable field evidence has shown that large volumes of sand may frequently move beyond the closure depth. Such movement can occur during both fair weather and storm periods, although offshore-directed storm flows are most likely the prime transport agent. Pilkey at al. cite studies in the Gulf of Mexico which measured offshore bottom currents of up to 200 cm/sec and sediment transport to the edge of the continental shelf. The amount of sediment moved offshore was large, but it was spread over such a large area that the change in sea bed elevation could not be detected by standard profiling methods[1]. Wright, Xu, and Madsen (1994) measured significant across-shelf benthic transport on the inner shelf of the Middle Atlantic Bight during the Halloween storm of 1991.

(c) *Assumption 3: There exists a sand-rich shoreface; the underlying and offshore geology must not play a part in determining the shape of the profile.*

Possibly the most important of the assumptions implicit in the equilibrium profile concept is that the entire profile is sand-rich, without excessive areas of hard bottom or mud within the active profile. Clearly these conditions do not apply in many parts of the world. Coasts that have limited sand supplies, such as much of the U.S. Atlantic margin, are significantly influenced by the geologic framework occurring underneath and in front of the shoreface. Many of the east coast barriers are perched on a platform of ancient sediment. Depending upon the physical state, this underlying platform can act as a subaqueous headland or hardground that dictates the shape of the shoreface profile and controls beach dynamics and the composition of the sediment.

[1] This latter statement underscores how important it is to develop improved methods to detect and measure sediment movement in deep water. With the present state of the science, the inability to measure changes in offshore sea bed elevation neither proves nor disproves the assumption of no significant sediment movement beyond the depth of closure.

Niederoda, Swift, and Hopkins (1985) believed that the seaward-thinning and fining veneer of modern shoreface sediments is ephemeral and is easily removed from the shoreface during major storms. During storms, Holocene and Pleistocene strata cropping out on the shoreface provide the immediate source of the bulk of barrier sands. Swift (1976) used the term *shoreface bypassing* to describe the process of older units supplying sediment to the shoreface of barrier islands.

Pilkey et al. (1993) contend that:

> ...a detailed survey of the world's shorefaces would show that the sand rich shoreface required by the equilibrium profile model is an exception rather than the rule. Instead, most shorefaces are underlain by older, consolidated or semi-consolidated units covered by only a relatively thin veneer of modern shoreface sands. These older units are a primary control on the shape of the shoreface profile. The profile shape is not determined by simple wave interaction with the relatively thin sand cover. Rather, the shape of the shoreface in these sediment poor areas is determined by a complex interaction between underlying geology, modern sand cover, and highly variable (and often highly diffracted and refracted) incoming wave climate. (p. 271)

(d) *Assumption 4: If a shoreface is, in fact, sand-rich, the smoothed profile described by the equilibrium profile equation (ignoring bars and troughs) must provide a useful approximation of the real shoreface shape.*

In addressing this assumption, Pilkey et al. (1993) cited studies conducted on the Gold Coast, in Queensland, Australia. The Gold Coast shoreface is sand-rich to well beyond a depth of 30 m. Without being directly influenced by underlying geology, the shoreface is highly dynamic. As a consequence, the Gold Coast shoreface shape cannot be described by one equilibrium profile; rather, it is best described by an ever-changing regime profile. Pilkey et al. concluded:

> The local shoreface profile shapes are entirely controlled by relative wave energy "thresholds"; for the sediment properties have not changed at all. Thus principal changes to the shoreface profiles of the Gold Coast are driven by wave power history with some modification by currents, and not by sediment size, or its parameter *A*, as defined within the equilibrium profile concept. (p. 272).

(4) General comments.

(a) The idea of a profile only adjusting to waves is fundamentally wrong as shown by Wright et al. (1991) and others. However, although the physical basis for the equilibrium profile concept is weak, critics of this approach have not proven that it always results in highly erroneous answers.

(b) Before the use of the equilibrium profile, coastal engineers had no way to predict beach change other than using crude approximations (e.g., sand loss of 1 cu yd/ft of beach retreat). The approximations were inadequate. Surveys from around the world have shown that shoreface profiles display a characteristic shape that differs with locality but is relatively stable for a particular place (i.e., Duck, NC). With many caveats (which are usually stated, then ignored), a profile can be reasonably represented by the equilibrium equation. The fit between the profile and the real seafloor on a daily, seasonal, and storm variation basis may not be perfect, but the differences may not matter in the long term.

(c) One critical problem for coastal engineers is to predict what a sequence of waves (storm) will do to a locality when little is known about the particular shape of the pre-storm beach. For this reason, numerical models like SBEACH (Larson and Kraus 1989), despite their reliance on the equilibrium profile concept, are still useful. The models allow a researcher to explore storm impact on a location using a general approximation of the beach. The method is very crude - however, the resulting numbers are of the right order of magnitude when compared with field data from many locations.

(d) Answers from the present models are not exact, and researchers still have much to learn about the weakness of the models and about physical processes responsible for the changes. Nevertheless, the models do work and they do provide numbers that are of the correct magnitudes when run by careful operators. Users of shoreface models must be aware of the limitations of the models and of special conditions that may exist at their project sites. In particular, profile-based numerical models are likely to be inadequate in locations where processes other than wave-orbital transport predominate.

h. Depth of closure.

(1) Background.

(a) *Depth of closure* is a concept that is often misinterpreted and misused. For engineering practice, depth of closure is commonly defined as the minimum water depth at which no measurable or significant change in bottom depth occurs (Stauble et al. 1993). The word *significant* in this definition is important because it leaves considerable room for interpretation. "Closure" has erroneously been interpreted to mean the depth at which no sediment moves on- or offshore, although numerous field studies have verified that much sediment moves in deep water (Wright et al. 1991). Another complication is introduced by the fact that it is impossible to define a single depth of closure for a project site because "closure" moves depending on waves and other hydrodynamic forces.

(b) For the Atlantic Coast of the United States, closure depth is often assumed to be about 9 m (30 ft) for use in engineering project design. However, at the Field Research Facility (FRF) in Duck, NC, Birkemeier (1985) calculated closure as deep as 6.3 m relative to mlw using CRAB surveys. Stauble et al. (1993) obtained 5.5 to 7.6 m at Ocean City, MD, from profile surveys. Obviously, it is invalid to assume that "closure" is a single fixed depth along the eastern United States.

(c) Closure depth is used in a number of applications such as the placement of mounds of dredged material, beach fill, placement of ocean outfalls, and the calculation of sediment budgets.

(2) Energy factors. As discussed above, the primary assumption behind the concept of the shoreface equilibrium profile is that sediment movement and the resultant changes in bottom elevation are a function of wave properties and sediment grain size. Therefore, the active portion of the shoreface varies in width throughout the year depending on wave conditions. In effect, "closure" is a time-dependent quantity that may be predicted based on wave climatology or may be interpreted statistically using profile surveys.

(3) Time considerations. The energy-dependent nature of the active portion of the shoreface also requires us to consider return period. The closure depth that accommodates the 100-year storm will be much deeper than one that merely needs to include the 10-year storm. Therefore, the choice of a closure depth must be made in light of a project's engineering requirements and design life. For example, if a berm is to be built in deep water where it will be immune from wave resuspension, what is the minimum depth at which it should be placed? This is an important question because of the high costs of transporting material and disposing of it at sea. It would be tempting to use a safe criteria such as the 100- or 500-year storm, but excessive costs may force the project

engineer to consider a shallower site that may be stable only for shorter return period events.

(4) Predictive methods.

(a) Hallermeier (1977, 1978, 1981a, 1981b, 1981c), using laboratory tests and limited field data, introduced equations to predict the limits of extreme wave-related sediment movement. He calculated two limits, d_ℓ and d_i, that included a buffer region on the shoreface called the shoal zone. Landward of d_ℓ, significant alongshore transport and intense onshore-offshore sediment transport occur (the littoral zone). Within the shoal zone, expected waves have neither a strong nor a negligible effect on the sandy bed during a typical annual cycle of wave action. Seaward of d_i, only insignificant onshore-offshore transport by waves occurs. The deeper limit was based on the median nearshore storm wave height (and the associated wave period). The boundary between the shoal zone and the littoral zone (d_ℓ) as defined represents the annual depth of closure. Hallermeier (1978) suggested an analytical approximation, using linear wave theory for shoaling waves, to predict an *annual* value of d_ℓ :

$$d_\ell = 2.28 H_e - 68.5 \left(\frac{H_e^2}{g T_e^2}\right) \tag{4-10}$$

where

d_ℓ = annual depth of closure below mean low water

H_e = the non-breaking significant wave height that is exceeded 12 hr per year (0.137% of the time)

T_e = the associated wave period

g = acceleration due to gravity

According to Equation 4-10, d_ℓ is primarily dependent on wave height with an adjustment for wave steepness. Hallermeier (1978) proposed using the 12-hr exceeded wave height, which allowed sufficient duration for "moderate adjustment towards profile equilibrium." Equation 4-10 is based on quartz sand with submerged density of γ' = 1.6 and median diameter between 0.16 and 0.42 mm, which typifies conditions in the nearshore for many beaches. If the grain size is larger than 0.42 mm, Equation 4-10 may not be appropriate. Because d_ℓ was derived from linear wave theory for shoaling waves, d_ℓ must be seaward of the influence of intense wave-induced nearshore circulation. However, because of various factors, Hallermeier (1978) "proposed that the calculated d_ℓ be used as a minimum estimate of profile close-out depth with respect to low(er) tide level." Because tidal or wind-induced currents may increase wave-induced near-bed flow velocities, Hallermeier suggested using mean low water (mlw) as a reference water level to obtain a conservative depth of closure. Note that Hallermeier's equations critically depend on the quality of wave data at a site. The reader is cautioned that Hallermeier's equations can be expressed in various forms depending on the assumptions made, the datums used as reference levels, and available wave data. The reader is referred to his original papers for clarification and for details of his assumptions. The equations may not be applicable at sites where currents are more important at moving sand than wave-induced flows.

(b) At the Lake Michigan sites that Hands (1983) surveyed, the closure depth was equal to about twice the height of the 5-year return period wave height (H_5):

$$Z \simeq 2 H_5 \tag{4-11}$$

In the absence of strong empirical evidence as to the correct closure depth, this relationship is recommended as a rule of thumb to estimate the 5-year profile response under Great Lakes conditions. The return period of the wave height should approximate the design life of interest. For example, the 20-year closure depth would be estimated by doubling the 20-year return period wave height ($Z \simeq 2H_{20}$).

(5) Empirical determination.

(a) When surveys covering several years are available for a project site, closure is best determined by plotting and analyzing the profiles. The closure depth computed in this manner reflects the influence of storms as well as of calmer conditions. Kraus and Harikai (1983) evaluated the depth of closure as the minimum depth where the standard deviation in depth change decreased markedly to a near-constant value. Using this procedure, they interpreted the landward region where the standard deviation increased to be the active profile where the seafloor was influenced by gravity waves and storm-driven water level changes. The offshore region of smaller and nearly constant standard deviation was primarily influenced by lower frequency sediment-transporting processes such as shelf and oceanic currents (Stauble et al. 1993). It must be noted that the smaller standard deviation values fall within the limit of measurement accuracy. This suggests that it is not possible to specify a closure depth unambiguously because of operational limits of present offshore profiling hardware and procedures.

(b) An example of how closure was determined empirically at Ocean City, MD, is shown in Figure 4-29 (from Stauble et al. (1993)). A clear reduction in standard deviation occurs at a depth of about 18 to 20 ft. Above the ~18-ft depth, the profile exhibits large variability, indicating active wave erosion, deposition, and littoral transport. Deeper (and seaward) of this zone, the lower and relatively constant deviation of about 3 to 4 inches is within the measurement error of the sled surveys. Nevertheless, despite the inability to precisely measure seafloor changes in this offshore region, it is apparent that less energetic erosion and sedimentation take place here than in water shallower than ~18 ft. This does not mean that there is no sediment transport in deep water, just that the sled surveys are unable to measure it. For the 5.6 km of shore surveyed at Ocean City, the depth of closure ranged between 18 and 25 ft. Scatter plots indicated that the average closure depth was 20 ft.

(c) Presumably, conducting surveys over a longer time span at Ocean City would reveal seafloor changes deeper than ~20 ft, depending on storms that passed the region. However, Stauble et al. (1993) noted that the "Halloween Storm" of October 29 to November 2, 1991, generated waves of peak period (T_p) 19.7 sec, extraordinarily long compared to normal conditions along the central Atlantic coast. Therefore, the profiles may already reflect the effects of an unusually severe storm.

(d) Figure 4-30 is an example of profiles from St. Joseph, MI, on the east shore of Lake Michigan. Along Line 14, dramatic bar movement occurs as far as 2,500 ft offshore to a depth of -25 ft with respect to International Great Lakes Datum (IGLD) 1985. This is where an abrupt decrease in standard deviation of lake floor elevation occurs and can be interpreted as closure depth. In September 1992, the mean water surface was 1.66 ft above IGLD 85. Therefore, closure was around 26-27 ft below *water* level.

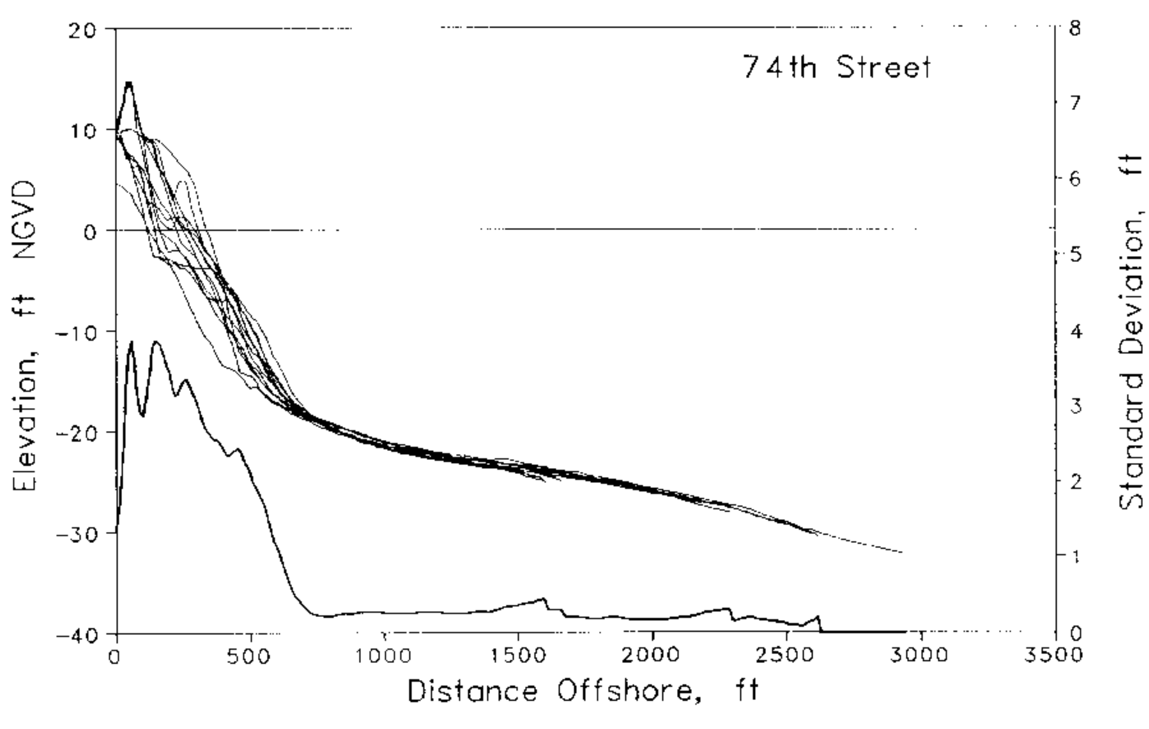

Figure 4-29. Profile surveys and standard deviation of seafloor elevation at 74th Street, Ocean City, MD (from Stauble et al. (1993)). Surveys conducted from 1988 to 1992. Large changes above the datum were caused by beach fill placement and storm erosion. Figure discussed in the text

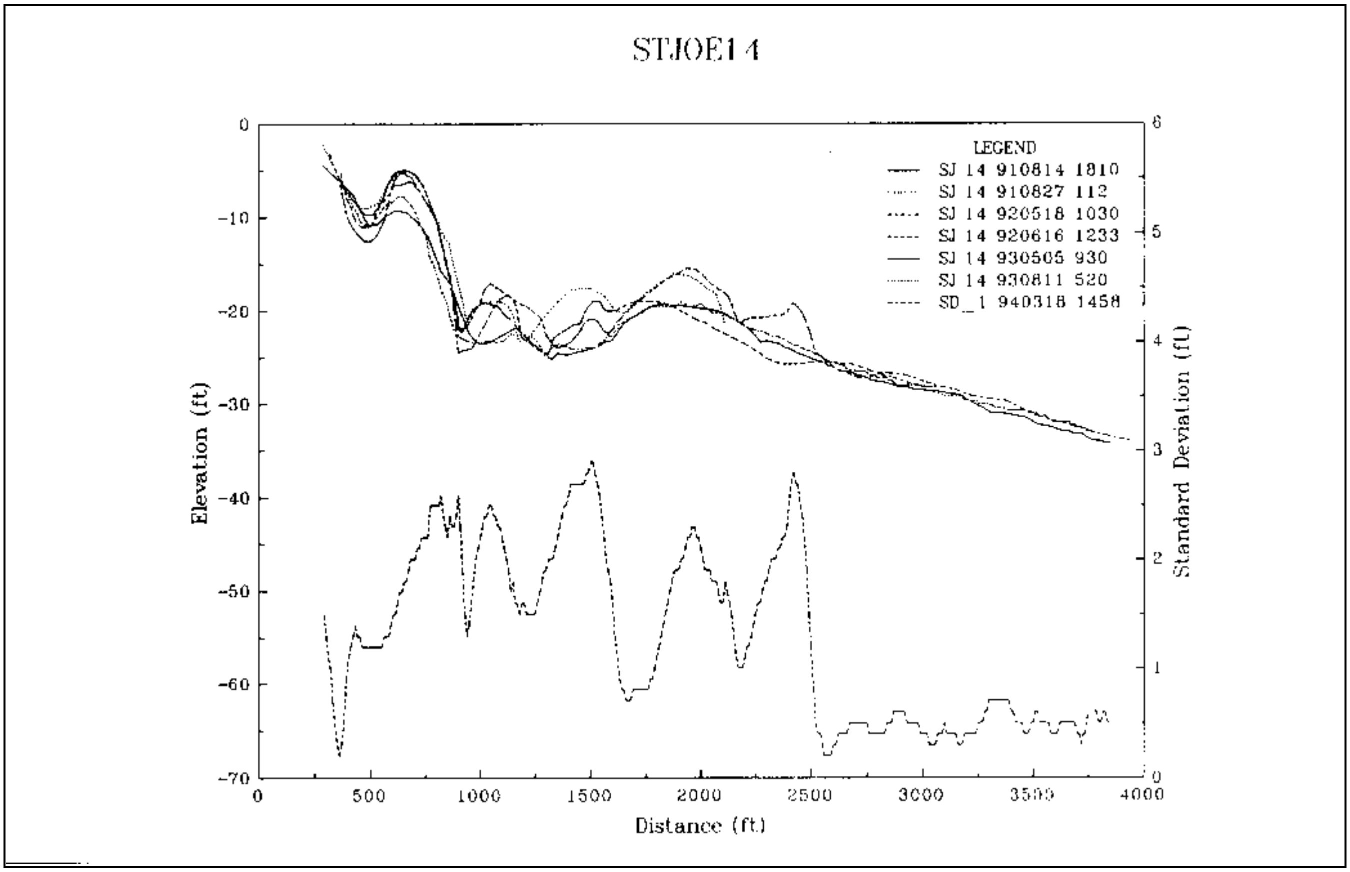

Figure 4-30. Profile surveys and standard deviation of lake floor elevation at St. Joseph, MI, on the east shore of Lake Michigan. Profiles are referenced to International Great Lakes Datum (IGLD) 1985. Surveys conducted between 1991 and 1994 (previously unpublished CERC data). Figure discussed in the text

(e) In the Great Lakes, water levels fluctuate over multi-year cycles. This raises some fundamental difficulties in calculating closure based on profile surveys. Presumably, during a period of high lake level, the zone of active sand movement would be higher on the shoreface than during a time of low lake level (this assumes similar wave conditions). Therefore, the depth where superimposed profiles converge should reflect the *deepest* limit of active shoreface sand movement. This would be a conservative value, but *only with respect to the hydrologic conditions that occurred during the survey program*. Presumably, if lake level dropped further at a later date, sediment movement might occur deeper on the shoreface. This suggests that closure on the lakes should be chosen to reflect the *lowest* likely water level that is expected to occur during the life of a project. (Note that this consideration does not arise on ocean coasts because year-to-year changes in relative sea level are minor, well within the error bounds of sled surveys. Sea level does change throughout the year because of thermal expansion, fresh-water runoff, and other factors as discussed in Chapter 2, but the multi-year mean is essentially stable.) In summary, determining closure depth in the Great Lakes is problematic because of changing water levels, and more research is needed to develop procedures that accomodate these non-periodic lake level fluctuations.

i. Longshore sediment movement.

The reader is referred to *Coastal Sediment Transport* (EM 1110-2-1502) for a detailed treatment of longshore transport.

j. Summary.

(1) A model of shoreface morphodynamics for micro- and low-mesotidal sandy coasts has been developed by Wright and Short (1984). The six stages of the model (Figure 4-22), illustrate the response of sandy beaches to various wave conditions.

(2) Sediment movement on the shoreface is a very complicated phenomenon. It is a result of numerous hydrodynamic processes, including: (1) wave orbital interactions with bottom sediments and with wave-induced longshore currents; (2) wind-induced longshore currents; (3) rip currents; (4) tidal currents; (5) storm surge ebb currents; (6) gravity-driven currents; (7) wind-induced upwelling and downwelling; (8) wave-induced upwelling and downwelling; and (9) gravity-induced downslope transport.

(3) The Bruun Rule (Equation 4-5 or 4-6) is a model of shoreface response to rising sea level. Despite the model's simplicity, it helps explain how barriers have accommodated rising sea level by translating upward and landward. A limitation is that the model does not address *when* the predicted shore response will occur (Hands 1983). It merely reveals the horizontal distance the shoreline must *ultimately* move to reestablish the equilibrium profile at its new elevation under the stated assumptions.

(4) The concept of the equilibrium shoreface profile applies to sandy coasts primarily shaped by wave action. It can be expressed by a simple equation (Equation 4-7) which depends only on sediment characteristics. Although the physical basis for the equilibrium profile concept is weak, it is a powerful tool because models based on the concept produce resulting numbers that are of the right order of magnitude when compared with field data from many locations.

(5) *Closure* is a concept that is often misinterpreted and misused. For engineering practice, depth of closure is commonly defined as the minimum water depth at which no measurable or significant change in bottom depth occurs (Stauble et al. 1993). Closure can be computed by two methods: (1) analytical approximations such as those developed by Hallermeier (1978) which are based on wave statistics at a project site (Equation 4-10); or (2) empirical methods based on profile data. When profiles are superimposed, a minimum value for closure can be interpreted as the depth where the standard deviation in depth change decreases markedly to a near-constant value. Both methods have weaknesses. Hallermeier's analytical equations depend on the quality of wave data. Empirical determinations depend on the availability of several years of profile data at a site. Determining closure in the Great Lakes is problematic because lake levels fluctuate due to changing hydrographic conditions.

4-6. Cohesive Shore Processes and Dynamics

a. Introduction.

(1) Cohesive sediments are typically homogenous mixtures of fine sand, silt, clay, and organic matter that have undergone consolidation during burial. These mixtures derive their strength from the cohesive (electrochemical attractive) properties of clay minerals, most commonly kaolinite, illite, chlorite, and montmorillonite. Clay particles exhibit a layered structure forming flaky, plate-like crystals that carry negative charges around their edges causing cations to be absorbed onto the particle surface. The presence of free cations is critical to the bonding of clay platelets. As clay particles become smaller, perimeters of the crystals become proportionally greater, which acts to increase the charge of each particle (Owen 1977). Owen (1977) describes a process in which some clays have the ability to absorb ions from solution into the layered structure of the clay, which allows the clay crystal to adjust its size and surface charge. In general, the higher the proportion of clay minerals, the more cohesive the sediment, although the type of clay mineral present, particle size, and the quantity and type of cations present in solution are also important factors.

(2) The presence of organic material may also be responsible for the cohesion of fine-grained sediments. Various organic substances are electrically charged and capable of acting as nuclei to attract clay minerals, forming particles having a clay-organic-clay structure (Owen 1977). Mucous secretions from various organisms can also bond fine particles together, forming cohesive sediments. These organic cohesive processes are quite common in low energy estuarine environments where fine-grained sediment sources are abundant and biological productivity is high.

(3) Detailed information on clay mineralogy and behavior is found in geotechnical engineering texts (Bowles 1979, 1986; Spangler and Hardy 1982).

(4) Hard, desiccated (dry), and well-compacted cohesive sediments are generally more erosion-resistant than cohesionless sediments exposed to the same physical conditions. Glacial till in some areas, such as the shores of the Great Lakes, is as consolidated and dense as sedimentary rock. Compacted and desiccated clay which is exposed on the seafloor in some formerly glaciated coasts (for example, off New England and Tierra del Fuego,

Argentina), is rock-hard and very difficult to penetrate with drilling equipment.

(5) In contrast, recent clayey sediments in river deltas or estuaries have a high water content and are readily resuspended by waves. As long as the receiving basins remain protected and there is a steady supply of new sediment, the soft clays accumulate and slowly compact (over thousands of years). Major storms like hurricanes can produce dramatic changes to marshy shores, especially if protective barrier islands are breached or overtopped by storm surges. A marshy coastline may also be severely eroded by normal (non-storm) waves if a river has changed its route to a different distributory channel, cutting off the sediment supply to this portion of the coast. The migration of the Mississippi River mouths is one of the factors contributing to coastal erosion in southern Louisiana (discussed in more detail in Chapter 4, Section 2).

(6) Coastal dynamic processes of cohesive shores are not as well understood and have not been as thoroughly studied as the dynamics of sandy shores. Because cohesive materials are very fine-grained, they are usually not found in recent deposits in exposed, high-energy coastlines. However, outcrops of ancient clay sediments may be present and may be surprisingly resistant to wave action. In protected environments where clays do accumulate, the shores develop distinctive morphological features in comparison with unconsolidated shorelines. Nairn (1992) defines a high-energy cohesive shore as being composed largely of a cohesive sediment substratum that plays a dominant role in the change of shoreline shape through the process of erosion. On the other hand, estuaries and tidal rivers are governed by quite different conditions: cohesive sediments are eroded, transported, and deposited on the seafloor primarily by tidal or fluvial currents (Owen 1977). This type of environment is also characterized by extremely high concentrations of suspended material in the nearshore water.

(7) The processes described here consider two categories of cohesive environments. The first deals with high-energy, erosional shorelines consisting of relict cohesive material being acted upon by contemporary processes. Materials from these environments are characterized by erosion-resistant, consolidated cohesive sediments that form distinctive geomorphic features along open shorelines. In contrast, the second category deals with low-energy, depositional environments of soft, unconsolidated muds, silts, and clays, characteristic of estuaries, deltas, and marshes.

b. High-energy cohesive coasts.

(1) High-energy cohesive coasts are those that do not permit abundant accumulation of fine-grained material due to sustained wave attack. Cohesive sediments in these environments are products of ancient geologic events that deposited and compacted the material into its present state. Coastal processes have exposed the material, leaving it vulnerable to the contemporary, high-energy wave conditions. The result is usually irreversible erosion across the entire active profile from the backshore bluff face to distances well offshore. These conditions are frequently found on open ocean shorelines in California and Massachusetts and are very common in the Great Lakes.

(2) Exposed cohesive coastlines have the ability to resist erosion due to the compressive, tensile, and consolidated properties exhibited by the sediment. Because these shores are primarily erosional rather than depositional, they exhibit distinctive morphological features in comparison with cohesionless shores. These distinct characteristics include steep vertical bluffs that constitute a marked discontinuity in slope between the upland and the shore (Mossa, Meisberger, and Morang 1992).

(3) The presence of a cohesive material underlying an unconsolidated sandy beach controls how the shoreface erodes. If the cohesive material is eroded by the high energy processes typical along open ocean and Great Lakes shorelines, the cohesive properties are lost. The fine-grained material does not have the ability to reconstitute itself, resulting in irreversible erosion. Most beach sand that results is quickly swept away during storms, preventing the formation of protective beaches. Where sand can accumulate, it has an important interactive role in cohesive shore processes. Sunamura (1976) states that sand introduced to the system acts as an abrasive agent on cohesive material, thereby increasing erosion rates. Nairn (1992) and Kamphius (1987, 1990) have shown that downcutting of the nearshore cohesive substratum by abrasion is the controlling factor in the recession of adjacent bluffs in the Great Lakes. The downcutting and deepening of the nearshore profile allows higher waves to attack the foreshore, resulting in accelerated bluff recession, as illustrated in Figure 4-31. However, as sand thickness increases over the cohesive surface, a threshold is reached where the sand protects the underlying material. At this stage, downcutting no longer occurs and shore recession is arrested.

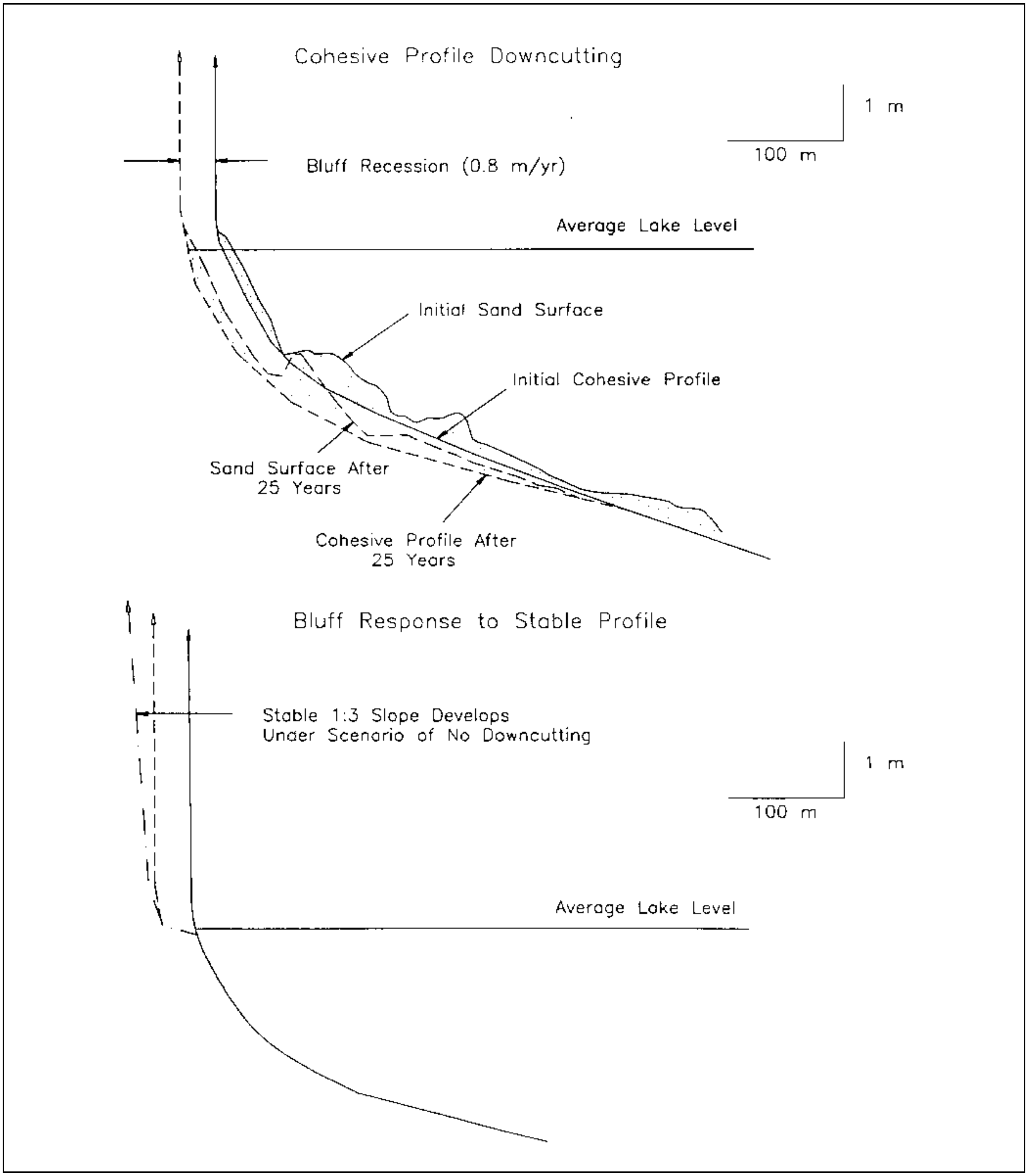

Figure 4-31. Illustration showing the relationship between downcutting of cohesive material in the nearshore and bluff recession (from Nairn (1992))

(4) Slopes and recession rates of the bluff faces depend on energy conditions as well as the geotechnical vproperties of the bluffs (grain size and degree of consolidation). Coastal processes, primarily waves, erode and undercut the base of the bluffs. This causes the upper portions to slump, resulting in a wide range of slope angles. In time, the bluffs may be fronted by a gently sloping beach or intertidal platform where debris may accumulate (Figure 3-22 and 4-32). If waves and currents remove the erosional debris faster than the rate of supply, then the bluff will rapidly retreat, resulting in a steep slope face. When the supply of eroded material exceeds the removal rate, debris accumulates at the base of the bluff, allowing for a lower angle slope face. Coasts shaped by these processes exhibit irregular shorelines. The formation of headlands and bays may be related to differential erosion rates of the various cohesive materials that are present. Once formed, irregular topography may have pronounced influence on waves, tides, sediment transport, and further shoreline evolution.

(5) Shorelines of the Great Lakes illustrate the processes described above. Cohesive shores on the Great Lakes are typically composed of hard glacial till deposits, remnants from the glacial processes that formed the lakes. Characteristic of Great Lakes cohesive shorelines is the existence of a backshore bluff (Figure 4-33). The bluff can be as low as a half meter, in the form of a wave cut terrace, or may be as high as 60 m or more (Nairn 1992). Where recession of the bluff has occurred, the face is steep and lacks vegetation. In some instances, there may be sandy beaches just seaward of the base of the bluff and there may be offshore sandbars. Other characteristics include the presence of exposed cohesive outcrops in the nearshore. Where sand cover is thin, intermittent, or nonexistent, downcutting of the nearshore lake bed occurs, leaving the base of the bluffs vulnerable to wave attack, allowing accelerated shoreline retreat.

(6) Much of Alaska's Bering Sea, Beaufort Sea, and Chukchi Sea coasts have low bluffs of permanently frozen glacial till. The water content of the till varies, and the bluffs thaw at varying rates on exposure to air during the summer. Storm surges cause dramatic bluff failures as ice in the toe turns to liquid and shear failures allow still-frozen blocks of bluff to fall. At times, these shores are protected by shore-fast ice that rides up at or near the summer water time, creating "ramparts" that may be several meters high. Some mechanical scour occurs, but often the net effect is armoring because the ramparts last beyond the time when the offshore ice is gone.

c. Estuaries and low-energy, open-shore coasts.

(1) Estuaries are semi-enclosed, protected, bodies of water where ocean tides and fresh water are exchanged. They function as sinks for enormous volumes of sediment. Estuarine sediments are derived from various sources including rivers, the continental shelf, local erosion, and biological activity, and sedimentation is controlled by tides, river flow, waves, and meteorology. The lower-energy conditions of estuaries, as opposed to those found on open coasts, allow for the deposition of fine-grained silts, muds, clays, and biogenic materials. Estuarine sediments are typically soft and tend to be deposited on smooth surfaces that limit turbulence of the moving water. When allowed to accumulate, these materials consolidate and undergo various chemical and organic changes, eventually forming cohesive sediments.

(2) The shores of estuaries and certain open-water coasts in low-energy environments (e.g., coastal Louisiana, Surinam, Bangladesh, and Indonesia) are characterized as having smooth, low-sloping profiles with turbid water occurring along the shore and extending well offshore (Suhayda 1984). These areas usually exhibit low and vegetated backshores and mud flats which are exposed at low tide. These conditions are also found in Chesapeake and Delaware Bays.

(3) Nichols and Biggs (1985) describe the movement of estuarine sediments as consisting of four processes:

- Erosion of bed material.
- Transportation.
- Deposition on the bed.
- Consolidation of deposited sediment.

These processes are strongly dependent on estuarine flow dynamics and sediment particle properties. The properties most important for cohesive sediments are interparticle bonding and chemical behavior because these parameters make cohesive sediment respond quite differently to hydrodynamic forces than to noncohesive sediments. Due to the cohesive bonding, consolidated materials (clays and silts) require higher forces to mobilize, making them more resistant to erosion. However, once the cohesive sediment is eroded, the fine-grained clays and silts can be transported at much lower velocity than is required for the initiation of erosion.

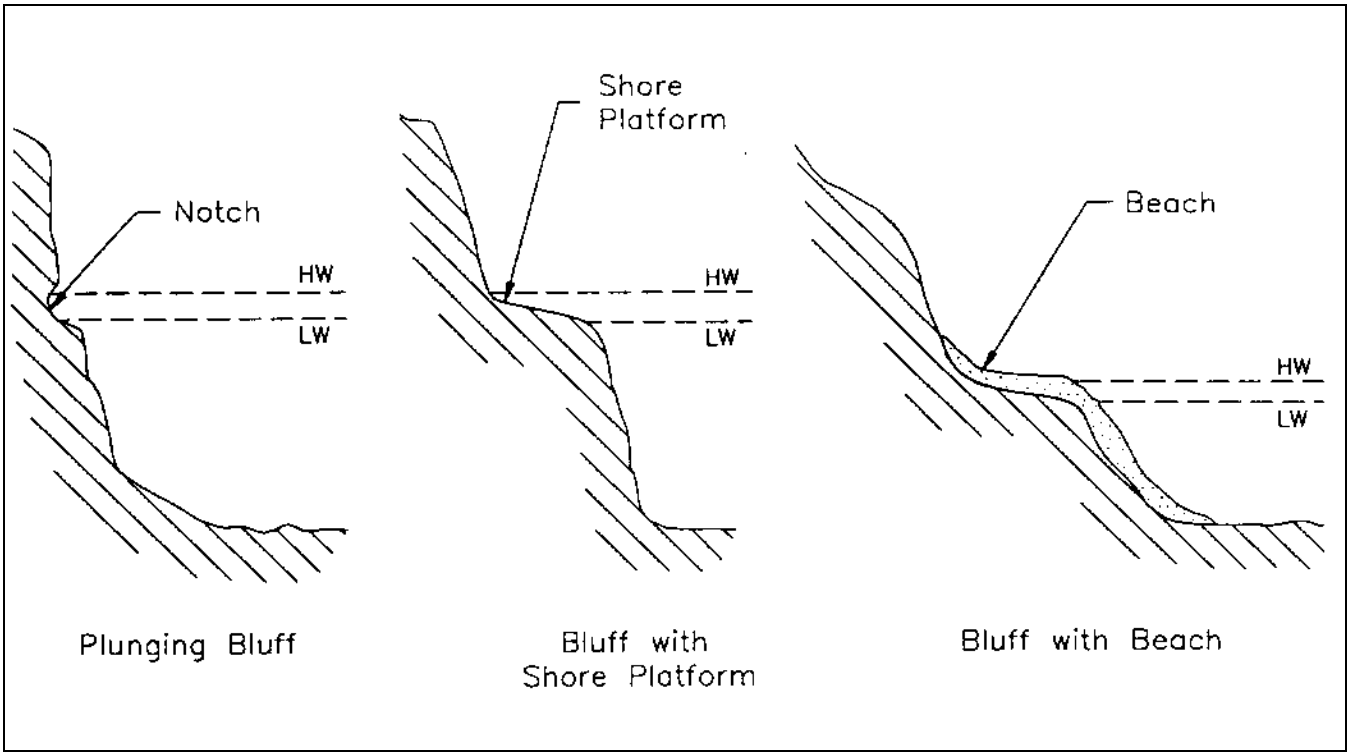

Figure 4-32. Variety of bluff morphology along cohesive shorelines (from Mossa, Meisburger, and Morang (1992))

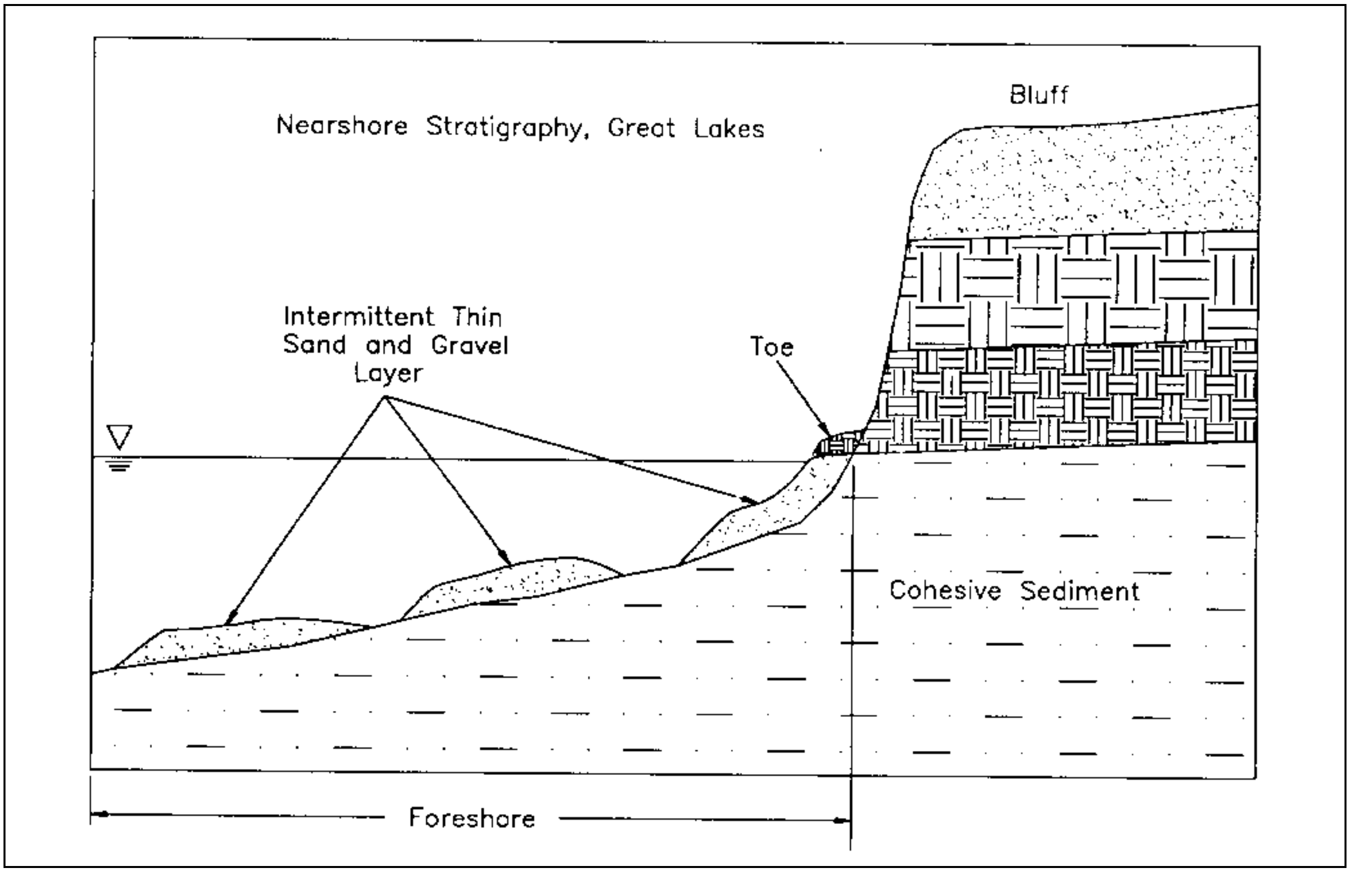

Figure 4-33. Characteristics of Great Lakes cohesive shorelines (great vertical exaggeration)