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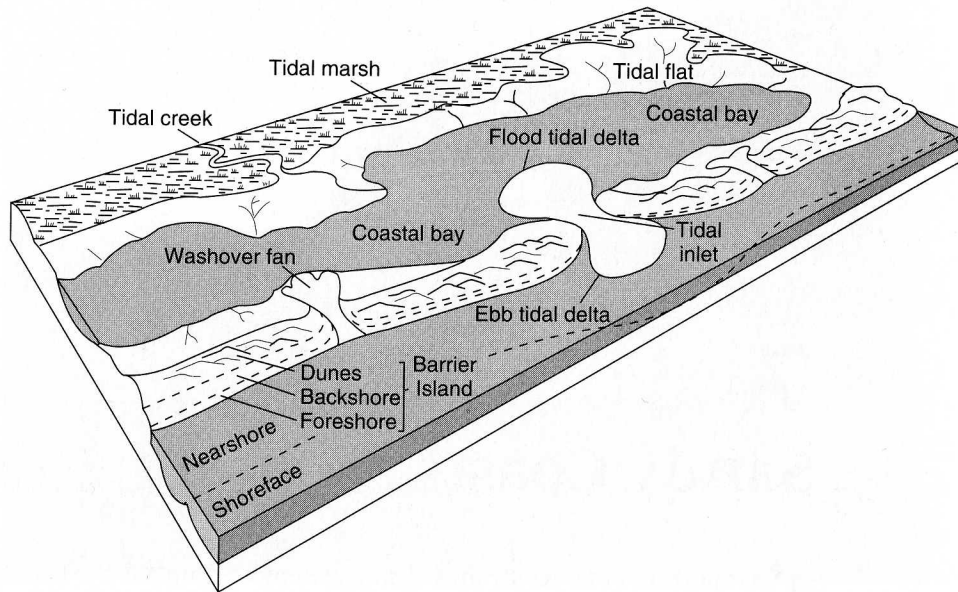
# 7

## Sandy Coasts

That portion of the coast most familiar to most people is the beach, closely followed by the dunes, which are typically landward of the beach. Although they also occur on the mainland, both of these coastal elements are also typical of barrier islands, quite complicated and dynamic parts of the coast. In addition to the beach and dune environments, these islands include washover fans that commonly become vegetated to form back-island flats (Fig. 7.1). Along the coast, barrier islands are typically interrupted by tidal inlets. The landward portion of the barrier island may also include widespread marshes and tidal flats similar in general character to those that border estuaries. In many barrier island systems, the back-barrier water body is parallel to the island and rather narrow. These water bodies are typically called lagoons, coastal bays that have little or no fresh-water input or tidal exchange with the open ocean.

Barrier islands only account for about 12 to 15 percent of the world's coast, but they tend to be the most expensive and most highly developed coastal type. In a cause and effect framework, barriers are the depositional portion of the wave-dominated coast as contrasted to the estuaries that are at least significantly influenced, if not dominated, by tides. The other important aspect of wave-dominated coasts is erosional, typically rocky and will be treated in Chapter 8.

Barrier islands come in a variety of sizes, shapes, and origins (Fig. 7.2). In general, they are elongate accumulations of sand that range from a kilometer or so to over a hundred kilometers in length. Some are barely above high tide, and others have dunes that rise tens of meters above the sea. Barrier islands occur throughout the world, from the north slope of Alaska to the tropics of South America and Australia. They are most prominent on stable, trailing edge and



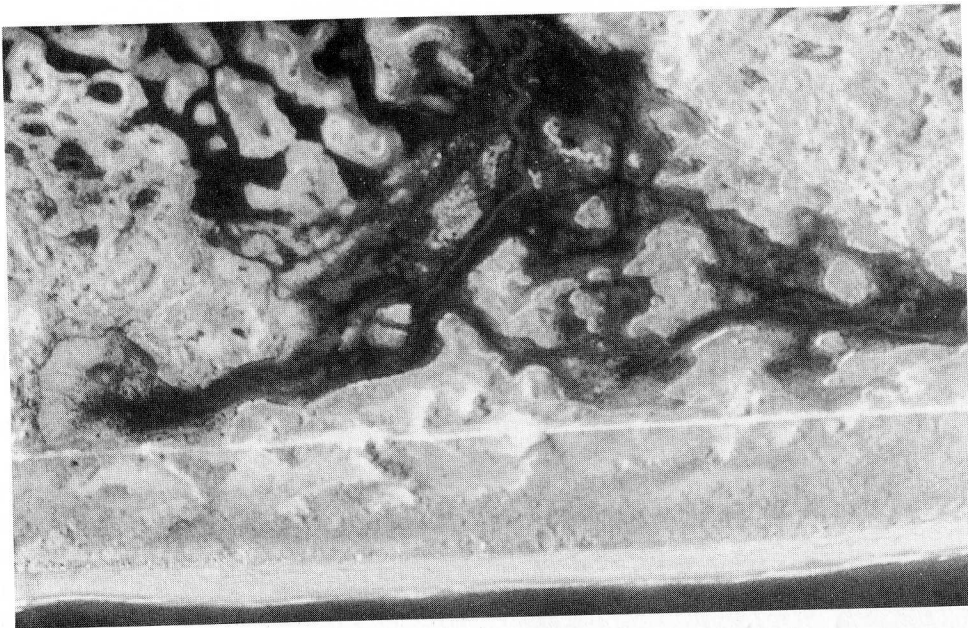
**FIGURE 7.1**

Schematic diagram of a barrier island system showing major environments.

marginal sea coasts but they are also present on some collision coasts such as on the Pacific coast of the United States, including Alaska.

Barrier islands may occupy virtually any geologic and tectonic setting if there is sediment available, processes to accumulate it, and a place where it can come to rest—a familiar set of requirements for coastal depositional environments. Given these constraints, tectonically active, leading edge coasts are not likely places for well-developed barriers. Even if sediment is available, and it is not always, the wave energy along these coasts tends to remove sediment. Additionally, there is little room for barriers along the generally steep nearshore of an irregular coast. Thus, the only common barriers along tectonically active coasts are short spits. These may, however, be quite common; along the Pacific coast of the United States there are nearly 100 such barrier spits at least a kilometer long.

Most well-developed barrier systems are on trailing edge coasts, where the inner shelf is gently sloping, wave energy is moderate to low, and sediment tends to abound. The Atlantic coast of the United States is essentially one continuous barrier island system from New Hampshire to Miami. Similar conditions may occur along the coasts of marginal seas for similar reasons.



(a)



(b)

**FIGURE 7.2**  
Oblique aerial photos of barrier islands on the a) Texas and b) South Carolina coast.

## FORMATION OF BARRIER ISLANDS

The location and configuration of barrier islands give some clues about their origin. They are generally separated from the mainland by open water and are present seaward of a wide variety of coastal environments and morphologies that include estuaries, lagoons, rocky headlands, and gently sloping coastal plains. Apparently the majority of the sediment from which they are made could not have come directly from the landward direction but must have been derived from offshore, alongshore, or some combination thereof.

Another important aspect to be considered in the origin of barrier islands is their age. The present, active barrier islands are all Holocene in age, in fact, they are quite young. Nearly all are less than 7000 years old and most were formed less than 4000 years ago. The dating part of their origin is rather straightforward; shells, wood, and other carbon-bearing material in the sediments provide radioactive carbon ( $^{14}\text{C}$ ), which can be counted thereby providing a date of origin.

The formation of the barriers is not so straightforward. Three origins for barriers have been presented in the geologic literature, all in the mid- to late-nineteenth century. The first was proposed by a French geomorphologist, Ille de Beaumont, in 1852. He advocated that waves caused sediment to accumulate in an upward shoaling scheme that eventually led to a supratidal sand bar that continued to accumulate sediment, became vegetated, stabilized, and formed a barrier island. A subsequent theory by McGee in 1890 proposed that coastal dune and beach ridges were drowned by rising water level leaving a strip of sand separated from the mainland by a narrow and shallow water body, also forming a barrier island. The third mode of origin is quite different and applies only to some barriers. In a 1885 publication Gilbert advocated formation of a spit emanating from a headland with eventual breaching at one or more places to form a barrier island with tidal inlets. This mode of occurrence is well documented in some parts of the world but because it requires a headland, most of the barriers of the Gulf of Mexico or of the Atlantic coast could not have formed in this manner. It is most common on the Pacific coast and in New England.

The purpose of this discussion is not to argue about the preferred interpretation of barrier island formation, however, a few comments on the first two modes of origin are warranted. Numerous very young barriers have formed during historical time documented by sequential aerial photography (Fig. 7.3). Many have developed according to the scheme of upward shoaling by wave action, whereas the only ones that have been demonstrated to result from drowning of a coastal beach-dune ridge or mainland detachment are in the Mississippi Delta area where land subsidence is rapid. In other locations it can be argued that sea-level rise has been too slow for any drowning of these ridges to occur in visually recordable history. The key here is that a slow rise in sea level will encourage either destruction or landward migration of the beach-dune ridges, not their drowning.

This brings us back to the question of why we have barrier islands along large sections of the coast. Remembering the discussion of sea-level rise since the



**FIGURE 7.3**

North Bunces Key, an example of a barrier that was formed on the Florida coast in less than a decade.

melting of the glaciers, for example, the Holocene transgression, provides some important clues. Sea-level rise curves show that the rate of rise was very rapid for several thousand years. This rapid rise probably did not provide a stable shoreline for enough time to develop a barrier island under any of the aforementioned scenarios, although some researchers would disagree.

At about 6000 to 7000 years before present the rate of sea-level rise slowed dramatically. In much of North America sea level was as much as 10 meters below its present position. Regardless of the absolute position of the shoreline relative to now, the rate of sea-level rise slowed to a maximum of about 1.4 millimeters per year, similar to the present rate in much of the United States. The slow rate of sea-level rise permitted enough time for waves and wave-generated currents to mold sediments into coast-parallel barrier islands, and permitted them to persist; some through slow landward migration as sea level rose but others were built vertically or prograded seaward from where they formed.

## **BEACH AND NEARSHORE**

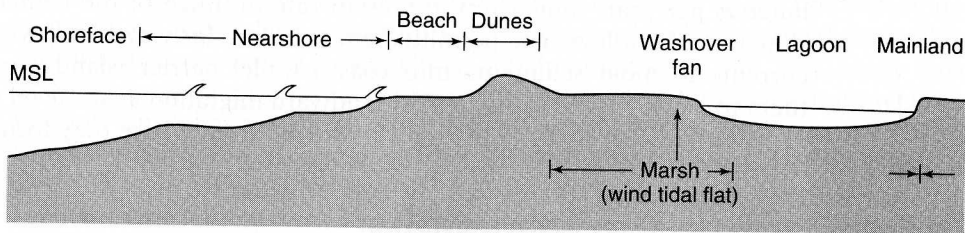
Most people visit or live along the coast to enjoy the beach. The constant change, interesting landscape, fascinating creatures, and the beautiful scenery always have captured peoples' attention. The beach acts as the seaward protection for

the coast, whether it be a barrier island or the mainland. Coastal residents commonly observe many aspects of beach dynamics, especially the results of storms and seasonal or long-term changes.

A proper understanding of beaches must include the adjacent shallow marine waters generally called the nearshore environment. It includes the region that extends from the low tide line, the seaward extent of the beach, out across the surf zone and includes the sandbars that are common along most coasts (Fig. 7.4). The width and depth to which the nearshore extends ranges widely depending upon its slope, wave climate, and abundance of sediment. In many places there are rather persistent sandbars that parallel the beach and over which waves break during storms (Fig. 7.5). Although some nearshore areas may be essentially smooth and gently sloping, most are a combination of sand bars and intervening troughs.

The beach itself extends from the low tide line landward across the unvegetated sediment to the beginning of vegetation or to the next geomorphic feature in the landward direction, which may be a dune, a bedrock cliff, or nowadays a seawall (Fig. 7.6). The beach is the most actively changing part of the coast; each wave causes shifts in sediment. It is composed of all kinds and sizes of sediment from mud to large boulders. The typical sandy beach displays two or three different profile configurations. These shapes depend upon the presence and configuration of the major elements that comprise the beach, the foreshore, the backshore, and the storm ridge. The foreshore includes the intertidal zone and extends to the landward break in slope. It is typically gently sloping toward the sea and may display a small, ephemeral bar and trough variously called a ridge and runnel or a swash bar (Fig. 7.7). The foreshore includes the swash zone in which waves rush up and back as they meet the shoreline.

The backshore or backbeach extends from the foreshore across the remainder of the beach and generally is not covered with water. Only during storm surge conditions is the backbeach subjected to wave activity; most of the time it is subjected to wind. This portion of the beach is generally nearly horizontal or



**FIGURE 7.4**

Profile diagram across the nearshore and barrier showing each of the major environments that are typically represented.

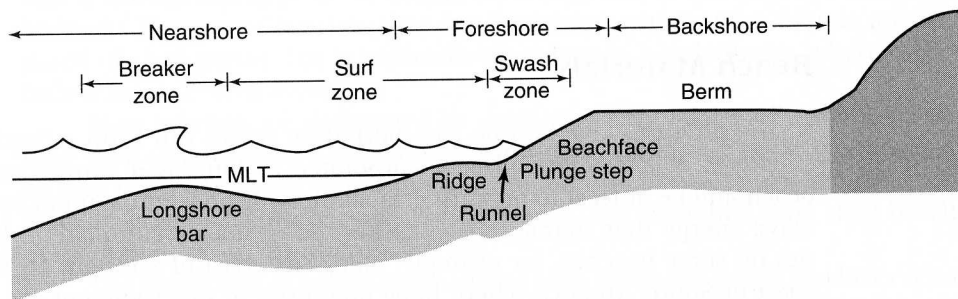
(From Davis, R. A., 1983, *Depositional Systems*, Prentice-Hall, Englewood Cliffs, New Jersey, p. 405)

**FIGURE 7.5**  
Oblique photo of waves  
breaking over longshore sand  
bars on Mustang Island, Texas.



slightly landward sloping. After major periods of erosion it may be greatly reduced or even absent. Gravel beaches of shell and/or rock fragments commonly include a storm ridge that is just landward of the foreshore. This feature may rise several meters above high tide and can totally replace the typical backbeach. Its composition depends upon the nature of the gravel material in the immediate area, and its size is typically directly related to the rigor of the storms that produce it.

The overall profile of the beach and adjacent nearshore may be steep or gently sloping. This depends upon a variety of factors such as sediment supply, wave climate, overall slope of the inner shelf, and perhaps tidal range. Generally,



**FIGURE 7.6**  
Beach profile showing major elements of a typical accretional beach.  
(From Davis, R. A., 1985, *Coastal Sedimentary Environments*, Springer-Verlag, New York, p. 386)



**FIGURE 7.7**

Photograph of ridge and runnel adjacent to a beach on the Sand Key, Florida coast.

the steep beach either has poorly developed nearshore sand bars or none at all. Little wave energy is dissipated across the nearshore, thus waves reach the shore with a fair amount of energy. These are the characteristics of reflective beaches. The other end of the spectrum is the gently sloping beach in which energy is gradually lost as the waves move toward the shore. These are dissipative beaches. The intermediate beaches tend to exhibit good development of the nearshore bars and formation of rip channels.

### **Beach Materials**

Nearly anything that can be transported by waves can form a beach (Fig. 7.8). Whatever is available will be incorporated as long as it can be carried to the beach and be at least temporarily stable there. Mud is uncommon because of the wave energy that characterizes beaches, but it does accumulate in large quantities on some beaches, for example, along the coast of Surinam on the northern coast of South America where huge quantities of fine sediment are made available by the Amazon River.

Gravel particles of virtually all sizes and compositions are fairly common on beaches given the correct set of circumstances—a readily available source and

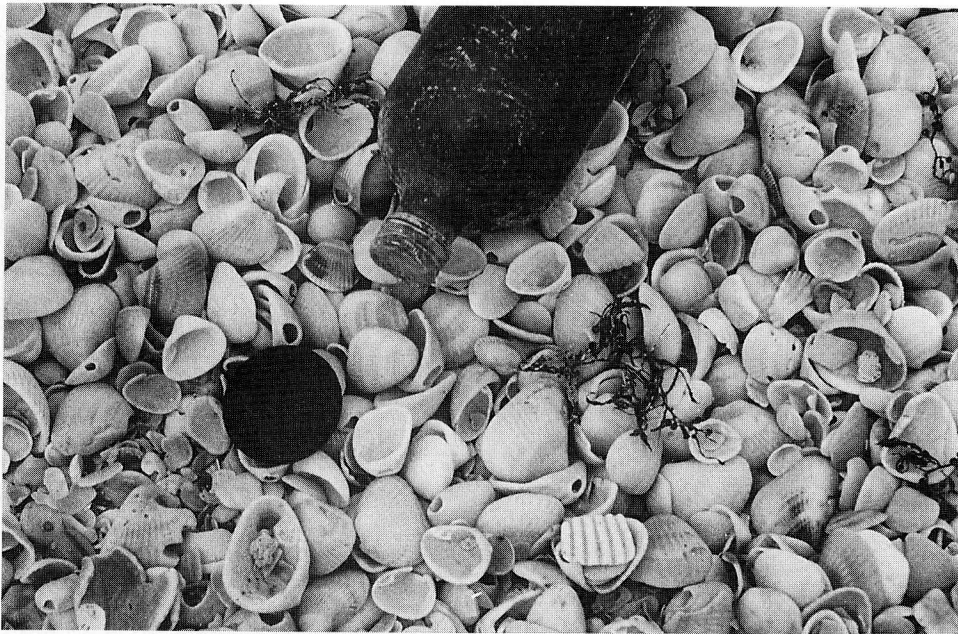
**FIGURE 7.8**

Example of beach material showing different grain sizes and compositions.

generally a high wave climate. They may range from small pocket beaches between headlands on an otherwise rocky coast to beaches that extend for many kilometers. Those in the northern latitudes are typically composed of rock fragments from glacial deposits or stream accumulations or are directly eroded from bedrock. The gravel beaches that are associated with barrier islands are composed of shell gravel that is especially abundant in low latitudes where there is little other sediment available.

Most beaches are dominated by sand; generally quartz in composition but they may contain nearly any mineral. Some are composed of only one mineral or rock fragment type if that is the only material available. Examples include the black or green beaches of some Pacific Islands and the white, fine shell beaches of the Bahamas and southern Australia.

The interaction of waves and currents with beach sediments produces certain textural characteristics. It is possible to generalize about these characteristics, especially for sandy beaches. As a consequence of wave activity, a given beach or a portion thereof displays a narrow range of particle size (Fig. 7.9). This is called *good*



(a)

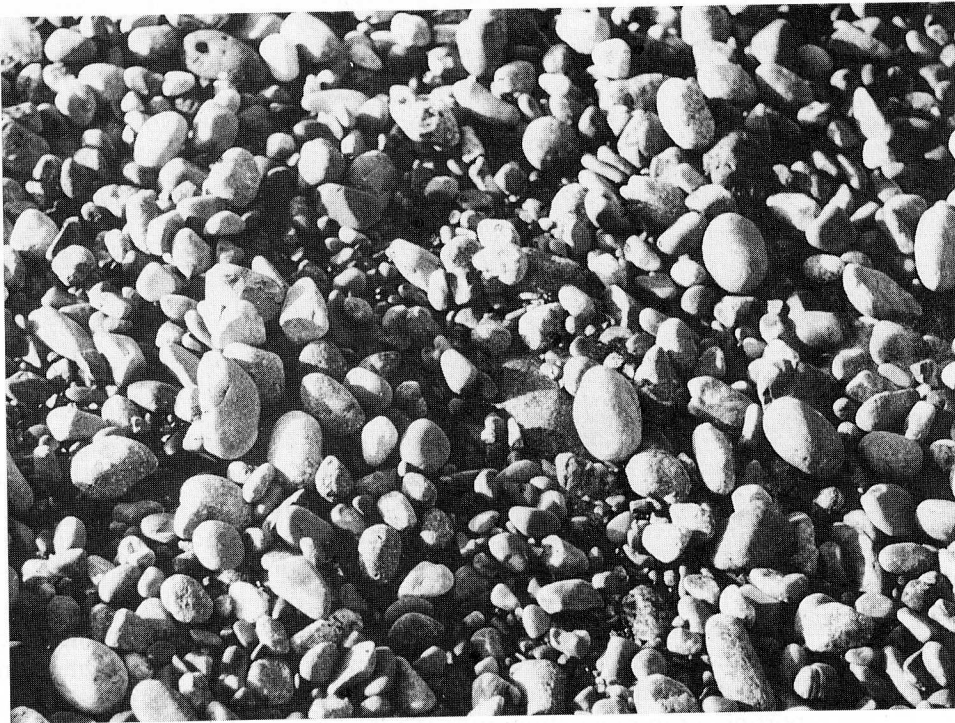
**FIGURE 7.9**

Examples of good sorting in beach sediment comprised of both a) biogenic and b) terrigenous grains.

*sorting*. That is, the waves and their related processes arrange the sediment particles in such a fashion that at any spot, one would expect to find sand grains that are nearly the same size. The same is true for gravel beaches; they are also well sorted, even though the absolute particle size is much larger than on sand beaches.

**Beach Dynamics**

The changes that take place on beaches range from the short-term ones, such as when individual waves strike the shore, to those that are long-term, taking at least decades. Likewise, the rate of change covers a very broad spectrum, to the extreme erosional conditions caused by severe extratropical storms and hurricanes. Regardless of the rate or scope of the changes that take place along beaches, there is some general level of predictability associated with each level of change. There are various cycles that beaches experience including those related to the tides and to the seasons of the year. The following discussion centers on



(b)

FIGURE 7.9 (Cont.)

the various levels of change, how they affect the beach, and what causes them. All of these cycles tend to be present on beaches regardless of where they occur—mainland, barrier island, rocky coast, or wave-dominated delta.

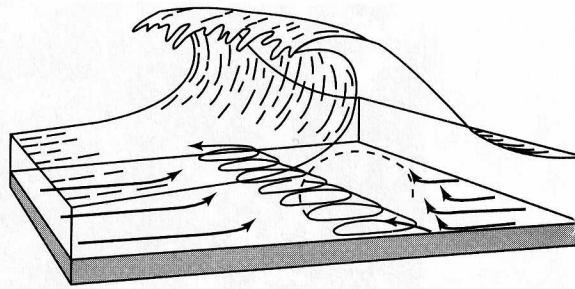
**Beach Processes** Both waves and tides and the currents that they generate may influence both the sediment and morphology of beaches. The breaking of waves produces obvious interaction with beach sediment (Fig. 7.10); each crash of the wave places sediment in temporary suspension, the amount directly related to the size of the wave. This suspended sediment is then moved by currents, primarily wave-generated longshore currents (see Chapter 4). The role of tides and tidal currents on beaches is more subtle. As the tides cause a rise and fall in the water level, the shoreline moves, causing the influence of waves to move accordingly. The greater the tidal range, the wider the band of change in incident wave energy. Tidal currents themselves do not appear to have a significant effect on beaches because they are so slow in this area. Only at the mouth of tidal inlets are these currents important.

Currents produced by waves are among the most important processes that

**FIGURE 7.10**

Diagram of the surf zone showing back and forth sediment movement under waves.

(From Ingle, J. C., 1966, *The Movement of Beach Sand*, Elsevier Publishing Co., Amsterdam p. 53)



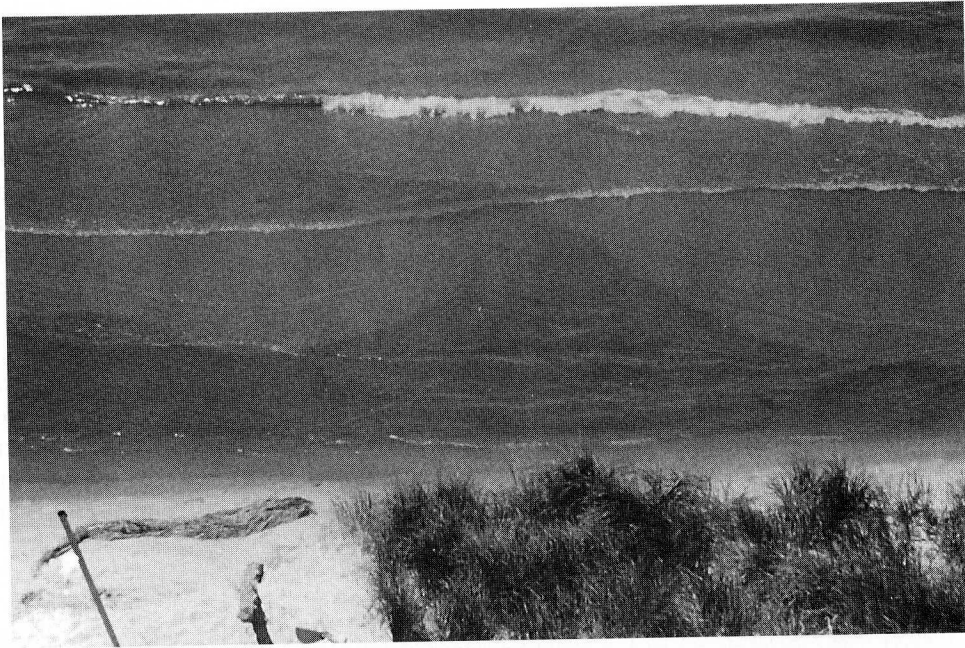
generate change in the beach. These include (1) combined flow currents, (2) longshore currents, (3) rip currents, and (4) the onshore–offshore currents produced in the surf zone and adjacent beach, and all are responsible or partially responsible for transport of sediment.

As waves propagate shoreward they are refracted, causing a vector of water movement along the shore that generates longshore currents. These currents can move in either direction along the beach, depending on the wave approach, and can range widely in their velocity, depending upon wave height and angle of approach; they may exceed 100 centimeters per second under storm conditions. The combination of waves stirring up sediment and the longshore current transporting it can move thousands of cubic meters of sediment in a day or two.

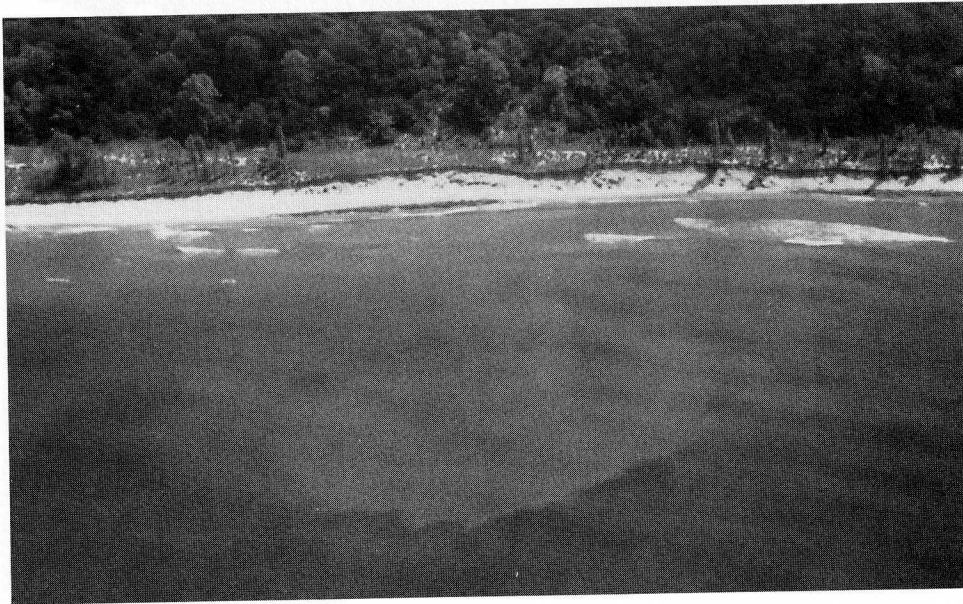
The shallow nearshore and commonly ephemeral sand bars that parallel the shoreline commonly have saddles or rip channels (Fig. 7.11). Rip currents that flow through these channels can also carry a modest amount of sediment (Fig. 7.12).

The other mechanism for sediment transport in this environment is on the swash zone where the uprush and backwash of the final breaking wave carries sediment across the foreshore. Depending upon wave conditions, slope of the foreshore, and sediment permeability, there may be a significant difference between the amount carried up the beach and that carried off the beach. One of the most important factors is permeability. With each wave a coarse and therefore permeable beach will experience a greater uprush of water than backwash. This is due to a portion of the water contained in the wave swash infiltrating into the sediment, limiting the return flow back in a seaward direction.

**Beach Cycles** The general conditions of beach processes and their interaction with beach sediment can be considered in two distinctly different but common scenarios. The most prevalent condition is one in which wave conditions are about equal to or below the mean energy conditions. At most beaches this includes a swell wave with a low wave height (generally less than 1 m) and a period of 8 to 12 seconds. Locally generated sea waves may be superimposed with a similar or smaller height and a period of 3 to 6 seconds. The sum of these condi-



**FIGURE 7.11**  
Oblique photo showing a saddle in a shallow nearshore sand bar; the location of rip currents.



**FIGURE 7.12**  
Large plume in the nearshore showing sediment transport in a rip current.

tions produces what is commonly called an *accretionary beach* because the dominant condition is one of deposition of sediment or stability; erosion is absent or quite limited (Fig. 7.13). The profile assumed by an accretional beach is one in which the backbeach is well-developed and wide with a relatively narrow foreshore. Nearshore sandbars display a relatively high amount of relief and are well-formed. This is a common configuration on most low-wave climate coasts such as the Gulf of Mexico and the Atlantic coasts of the United States.

The other condition is one of an erosive or storm beach (Fig. 7.14). Storms, although short in duration, are the dominant process along many coasts. During such conditions it is typically the increase in wind wave size that results in this dominance. These relatively steep waves cause considerable entrainment of sediment through combined flow motion and longshore sediment transport. Large quantities of sediment are in suspension, and the currents readily carry this sediment both offshore and alongshore. Additionally, the swash energy is high, and uprush and backwash are more extensive due to the large waves.

The result is removal of enough sediment from the beach to produce an erosional or storm profile. Such a profile is generally flat and featureless and has either a narrow backbeach or none at all, with most or all of the beach being in the foreshore portion. The post-storm beach typically contains a ridge and runnel



**FIGURE 7.13**

Photograph showing an example of an accretional beach.



**FIGURE 7.14**  
Photograph of an erosional beach profile.

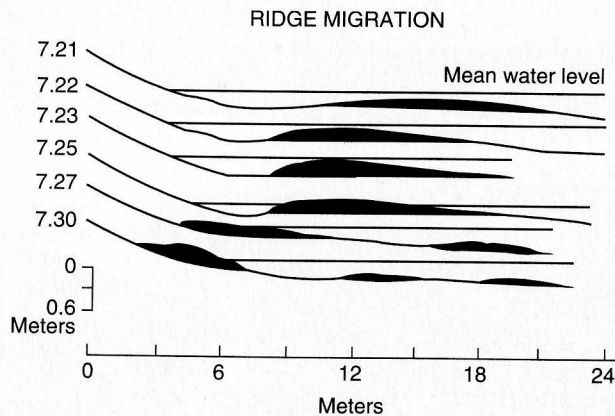


**FIGURE 7.15**  
Photograph of ridge and runnel along a recently eroded beach.

(Fig. 7.15) in the lower foreshore. The nearshore sandbars and related troughs tend to display less relief and move offshore up to a few tens of meters. The beach itself commonly is covered with a veneer of heavy minerals that accumulate as a lag deposit as the result of concentration during removal of the light fraction of the sand. This is really a placer deposit in much the same fashion as gold is concentrated in streams.

The storm beach is a temporary condition and in the absence of successive storms the recovery period begins as the storm starts to wane. This recovery process is initiated by the return to low-energy wave conditions and an absence of strong longshore currents. Along most coasts rip currents tend to develop best during low-wave energy conditions, but the amount of sediment they transport seaward is small compared to wave and longshore current transport. As swell and small wind waves persist there is significant landward transport of sediment. The nearshore sandbars return to their original position and configuration. More obvious is the shoreward migration of the ridge that results from washover during a flooding tide (Fig. 7.16). The original, somewhat symmetrical bar becomes asymmetric with the steep side landward. Over a period that ranges from about two weeks to as much as three months, the ridge repairs the beach, and an accretional profile that resembles the pre-storm situation is created. There may be some net loss, or even some net gain, depending upon how much sediment was lost or gained through longshore transport during the storm.

At some places where beaches are strongly influenced by waves that approach parallel to the coast, rip currents are strong during high-energy conditions, and longshore transport is minimal. This condition is common along the Pacific coast and also occurs in northern New England.



**FIGURE 7.16**

Diagram showing sequential migration of ridge and runnel as low energy conditions prevail after a storm.

(From Davis, R. A., et al, 1972, *Jour. Sed. Petrology*, 42, p. 416, Fig. 4)

The beach recovery process can be interrupted by storms that return the beach to an erosional condition. The somewhat closely spaced occurrence of these storms during the winter season has led to what is commonly called a *winter beach profile*, the same as a storm profile. If storms are frequent and severe, such as is common along the West Coast of the United States, the entire beach can be removed leaving a bedrock bench that is essentially devoid of sand. This condition may persist for a few months until the lower wave energy conditions of the spring and summer return sediment to the beach and eventually build an accretionary profile (Fig. 7.17). Good examples of this situation take place along the central Oregon coast and also in southern California.

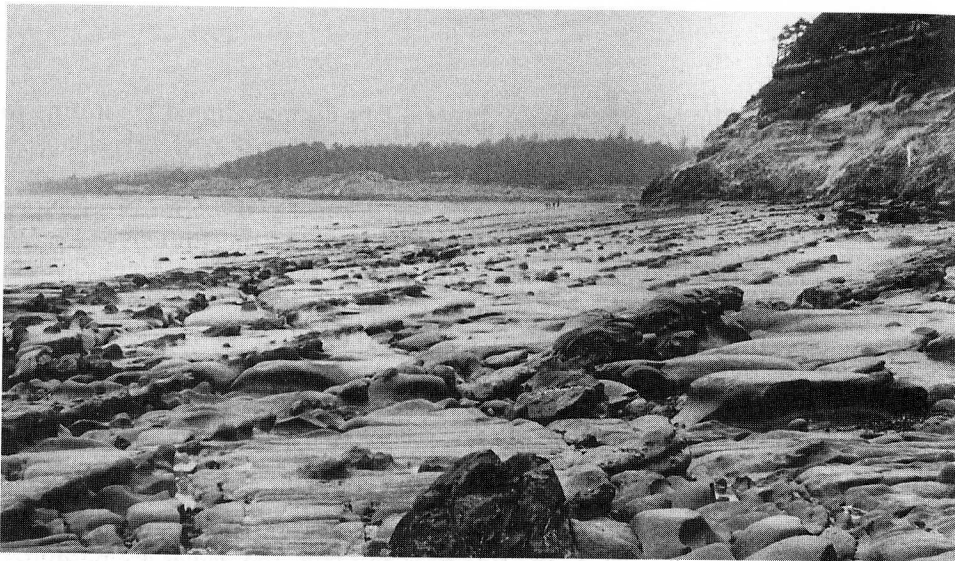
This sequence of beach conditions is quite seasonal and predictable, and produces the most obvious cyclic sequence in the dynamics of the beach environment. Other less pronounced cycles are associated with the diurnal tidal cycles and also with monthly lunar cycles. In each there is slight change in the foreshore beach. As the tide rises and falls, the waves are permitted to impact on different parts of the foreshore. Likewise, during the lunar cycle, the neap tidal range produces a narrower range of wave impact on the swash zone than does the spring tidal range.

Although beaches are quite dynamic by nature, and there is a wide variety of beach locations, sediments, and sizes, there is a predictable pattern to their behavior and to their morphology. Likewise the sediment we find on the beach is typically well-sorted but with a wide range in grain size. We can learn much more about beach dynamics from observing a single beach over an extended period of time than we can from visiting numerous beaches each a single time.

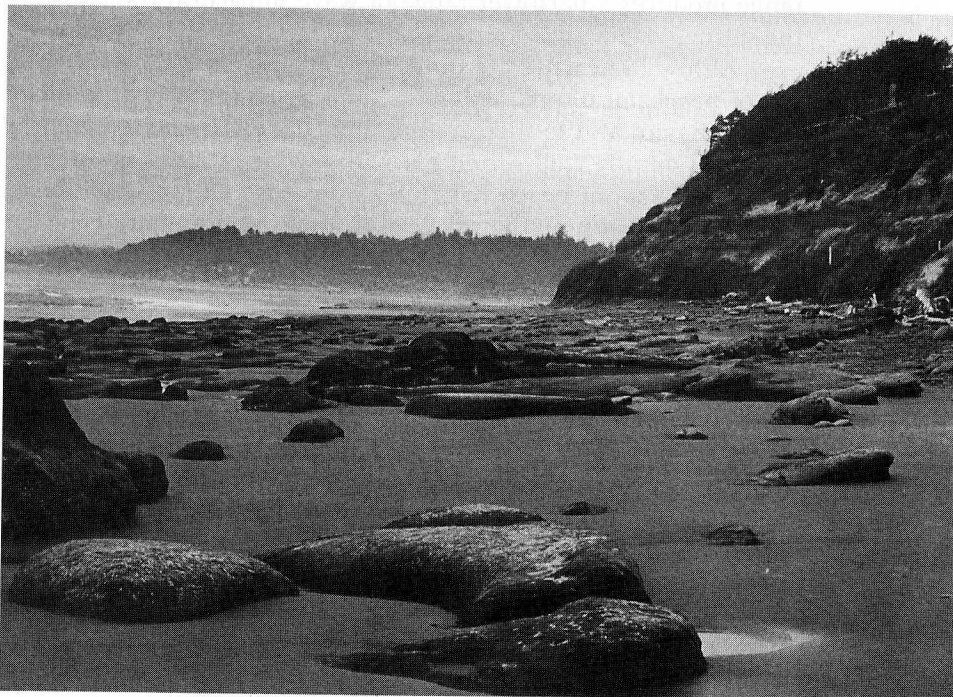
## DUNES

Sand dunes are an important part of many, but not all, coastal areas. These dunes are large piles of sand that accumulate from similar processes and in generally similar shapes and patterns as do the dunes on inland deserts. The fundamental prerequisites in both cases are a supply of sediment and the wind to move it. In the case of the coast, the wind is typically not a limiting factor although the sediment supply may be. Nevertheless, we have coasts in which dunes may exceed 100 meters in elevation (Fig. 7.18).

Coastal dunes are not restricted to barrier islands, although nearly all barriers have at least small dunes. Some coasts without barriers have tremendous dune fields. Particularly good examples are the southern coast of Oregon where the dunes extend a few kilometers inland from the coast and the southeastern part of Lake Michigan where dunes over 100 meters high have developed. Among the largest dunes in the world are those on the coast of Namibia in southwest Africa where barriers are absent.



(a)



(b)

**FIGURE 7.17**  
Beach on the Oregon coast a) after sediment is removed in the winter and b) when it returns during the spring.  
(Courtesy of William T. Fox)



**FIGURE 7.18**

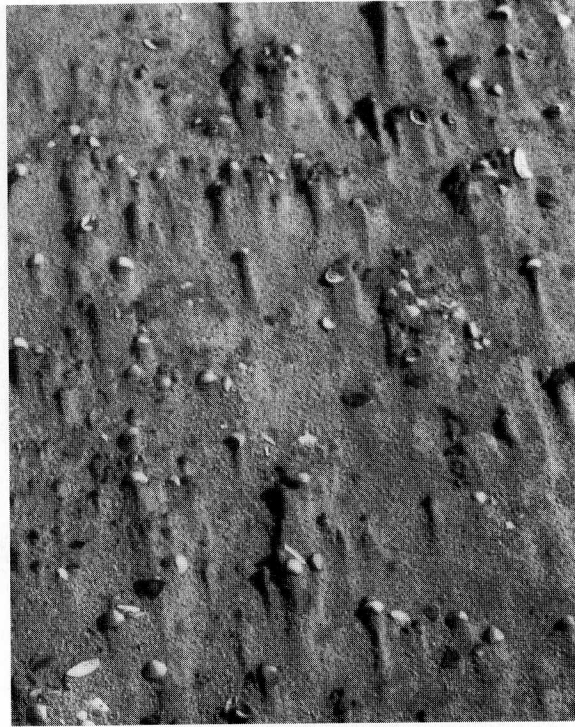
Photo showing examples of very large coastal dunes along the eastern coast of Australia.

### **Dune Formation and Distribution**

Any coast where sand accumulates has the potential to develop dunes. On most coasts prevailing winds or diurnal sea breezes provide the transport mechanism. Most areas have prevailing winds with some component that is onshore or shore-parallel, and in many areas the sea breeze may be a major factor, in fact, it is dominant along the southwest coast of Australia near Perth. In some places, such as along the New England coast, the dominant winds are the major factor in eolian sediment transport. Basically any dry part of the beach is subjected to eolian (wind) transport. The backbeach is especially susceptible to such processes because it is rarely wet. In fact, it is common for the backbeach to show various evidences of wind transport including ripples, sand shadows (Fig. 7.19), and heavy minerals or gravel lag concentrates. The sand shadows indicate a recent wind direction and may show scour around a shell or pebble. The gravel lag deposit is caused by wind blowing the fine sand from the beach and leaving the larger particles that cannot be transported. After a while the large particles become quite concentrated and actually form almost a pavement. Such a pavement inhibits further wind erosion and is called *desert armor* because of its importance in protection from wind erosion.

Much of the wind-blown beach sand tends to accumulate just landward of

**FIGURE 7.19**  
Photo showing sand shadows  
and shell lag on the backbeach.



the active backbeach. It is stopped from further transport by any type of obstruction that may be present, including bedrock cliffs, vegetation, existing dunes, or even human construction such as buildings or sea walls. Once the accumulation of eolian sediment begins, it continues unless conditions change, such as loss of sediment supply or the destruction of the stabilizing factor.

One of the best and most widespread aids in dune development is vegetation. Any type of plant serves as a focus for anchoring wind-blown sediment. Typically, the relatively inactive backbeach is covered with opportunistic plants such as the beach morning glory, beach *Spinifex*, and marram grass (Fig. 7.20). One of the most effective dune stabilizers on southern dunes, up to the latitude of Virginia, is sea oats (*Uniola*), while the American beach grass (*Ammophila*) extends from Virginia up to Nova Scotia in Canada. It is quite common to see small piles of sand around isolated plants on this part of the beach. After only months, these piles increase in size and the plants spread, thus increasing their effectiveness. In fact, even pieces of wood or any other sizable obstacle can act as a seed for dune development. Eventually small dunes or coppice mounds are present.

Such small incipient dunes are quite vulnerable; even a modest storm can destroy them, requiring the building process to begin again. This is the reason



**FIGURE 7.20**  
Opportunistic backbeach vegetation and small coppice mounds; the precursors to dunes.

that so much attention is paid to preserving vegetation on the backbeach and at the foot of the dunes. Absence of intense storms along with an abundant supply of sand and a regular mechanism for sand delivery eventually produces a dune. Dune size depends largely on the supply of appropriate sediment.

The typical configuration of dunes just landward of the beach is a linear arrangement that is essentially one dune wide. This dune ridge is called the *fore-dune* because of its location in the seaward or front of the barrier or the mainland as the case may be (Fig. 7.21). Many coasts contain numerous parallel dune ridges, each of which developed immediately landward of the active beach as a foredune ridge. The presence of several dune ridges indicates the portion of the coast that has had a history of growth or progradation toward the sea (Fig. 7.22). This is the desired condition for any coast because it indicates an overall lack of erosion. Dunes are about the best protection we have against severe storms and their related large surges. Some barrier islands contain a complicated assortment of dune ridge arrangements that show sets of ridges at acute angles to one another. This condition indicates periods of erosion that separated periods of dune accumulation and barrier progradation.

Not all barrier island dunes are located adjacent to or are associated with the beach. Like those near the beach, their presence depends on an abundant



**FIGURE 7.21**  
Photo of foredune ridge along 90 Mile Beach on the Victoria coast of Australia.



**FIGURE 7.22**  
Oblique aerial photo of multiple dune ridges formed as the coast prograded seaward, Victoria, Australia.

sand supply and a means to move and accumulate it. In most instances one of two conditions leads to landward dune development on the coast—a tremendous sediment supply or an absence of stabilizing vegetation. Probably the best example of extreme abundance of sand is found along the coast of southern Oregon. Here the strong winds off the Pacific along with great amounts of sand have produced huge mobile dunes that extend 3 to 4 kilometers inland. These coastal dunes have inundated forests of mature trees as they migrate in a southerly direction perpendicular to the coast. Similar conditions exist elsewhere, for example, at Dauphin Island, Alabama, at Kitty Hawk, North Carolina, and Horseneck Beach, Massachusetts.

Generally subarid conditions in addition to abundant sand along the beaches have resulted in an extensive active dune complex in the Padre Island National Seashore, Texas, on the central part of the island. Extensive cattle grazing on the island during the late nineteenth and early twentieth century helped create these conditions. This portion of the Texas coast is one of considerable sediment accumulation and persistent onshore winds. As a result, the island is extremely wide, and the mainland is dominated by an extensive dune complex. The barrier itself contains extensive active dunes that range from being only a meter or so high on the landward side near Laguna Madre, to several meters high in the central island. The limited development of the dunes on the landward part of the island is due to their destruction from storm surge associated with hurricanes. These small dunes are on the wind tidal flats where storm surges of a meter or so are fairly common and can flood the dunes, destroying them by a combination of waves and currents. After the surge subsides, the wind and available sand must start again to construct the dunes.

## Dune Dynamics

The existence of dunes is testimony to how sand can be transported by wind on the coast. Attack by waves is an obvious factor in dune stability. Although vegetation is an effective stabilizer of these accumulations, there are conditions in which even vegetated dunes may become mobile (Fig. 7.23) or eroded.

The first and most obvious factor is the attack by waves. Even though dunes are out of the regular influence of waves, they are quite vulnerable even to only modest surge produced by storms. In areas of generally erosive beach conditions, dune retreat is a particular problem because there is no backbeach to protect them. Elevated water level with superimposed storm waves produces swash and in some cases, direct wave attack, at the base of the dune. The sand is easily washed away and carried both offshore and alongshore. Even though a dense dune grass cover is present, the sand is easily removed, commonly leaving a dense root system hanging over the scarp in the dune. Post-storm recovery may occur and return some, or even all, of the sand to the beach. Proper conditions can start the rebuilding process of the dune but it can take many years to



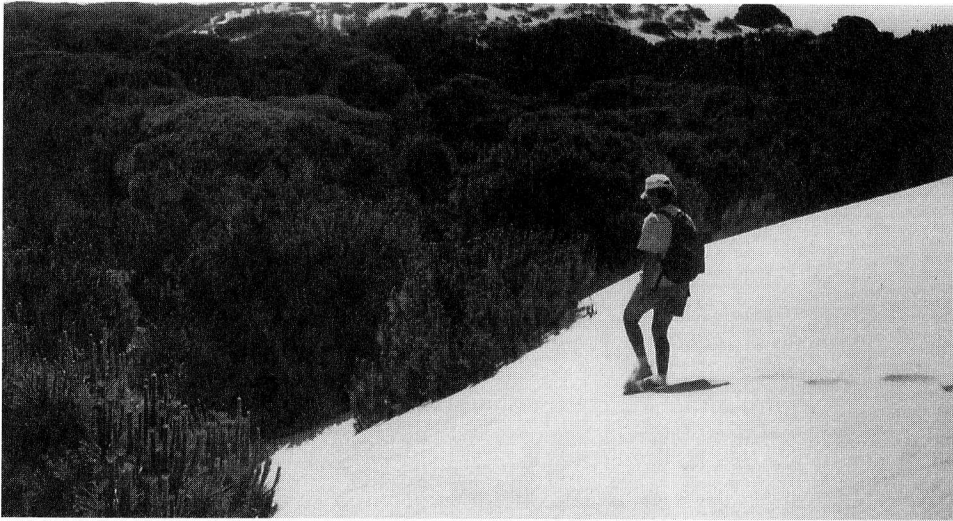
**FIGURE 7.23**

Large, mobile coastal dunes on the New South Wales coast of Australia.

restore the loss of just a single storm. It is generally rather easy to recognize dunes that have been eroded and then rebuilt by their profile change and perhaps even by their type of vegetation. By providing continual increase in the accessibility of the dunes to wave attack, rising sea level presents another scenario for dune erosion.

The other major aspect of dune dynamics is concerned with the migration of part or all of the dune through eolian processes. The same mechanism that forms the dune also can cause it to move, sometimes great distances. Generally dune mobility is associated with an absence of vegetation. Climatic conditions may reduce or eliminate vegetative cover, or overgrazing may remove much of the vegetation. Regardless of the reason, the result is the same—sediment begins to move.

The most common process for dune migration is called *blowover*. The on-shore wind component simply carries sand across the dune surface and permits it to move down the landward side by gravity. This creates a relatively steep slope called the *angle of repose*, generally about 30 degrees (Fig. 7.24). In other words, the sand is able to maintain a slope of this gradient as it migrates landward. This is true for all dunes regardless of their location or direction of migration. The sed-



**FIGURE 7.24**

Steep landward side of migrating coastal dunes that are transgressing over a forest on the coast of Spain.

iment may move as the result of individual grains rolling down the steep slope or as the result of grain flow. This is a type of sediment gravity process whereby oversteepening of the slope causes an instability that results in large numbers of grains moving down the slope in an avalanche fashion (Fig. 7.24). Anyone who has walked down a dune face has seen this phenomenon take place as the dune is disturbed. Migration of large dunes pays little attention to trees, buildings, or whatever is in its path as long as the dune is larger than the obstruction it will move over. Houses have been buried and then many years later exhumed as a result of migrating dunes.

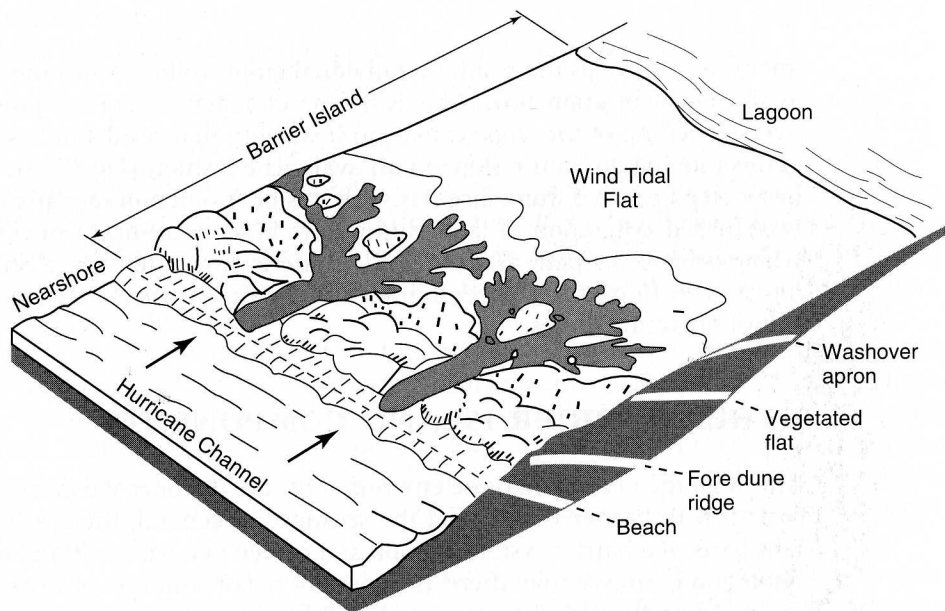
## **OTHER BARRIER ISLAND COMPONENTS**

Although the beach and dune environments are the ones we most commonly associate with barrier islands and the seashore in general, there are other important parts of a barrier system. A look at a profile of a barrier island or at an aerial photograph shows that there is generally a fair amount of area between the dunes and whatever the water body might be on the landward side of the barrier. Most of the landward portion of the barrier area is rather flat and only slightly above sea level. It typically continues with little change in profile across the back part of the island to a marsh and eventually to a tidal flat along the margin of the coastal bay that separates the island from the mainland. This flat, land-

ward portion of the barrier is readily divisible into at least three distinct environments based upon vegetation and elevation relative to mean sea level.

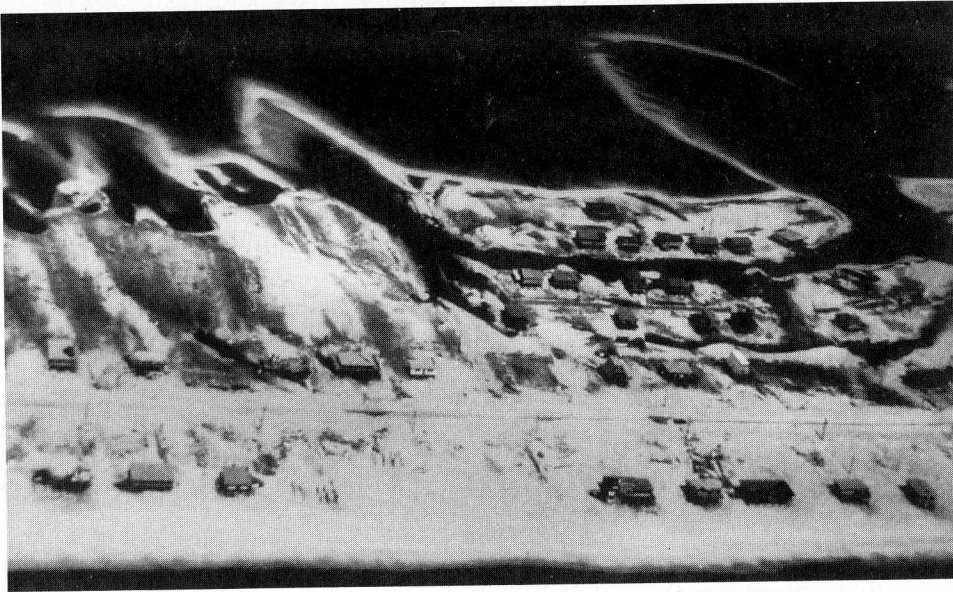
### Washover Fans

Virtually all of the sediment on which these environments have developed, however, was originally deposited by one mechanism and accumulated in one type of sediment body—washover fans (Fig. 7.25). Washover fans are thin, individually fan-shaped wedges of sand that are deposited during intense storm conditions when all or part of the beach–dune system is overtopped by waves and storm surge. On some barriers the individual fans coalesce to form a washover apron. The sediment accumulations that represent a single storm range up to about a meter in thickness but are more commonly few tens of centimeters. On narrow islands or under conditions of very high energy, the fans may extend completely across the island and into the adjacent coastal bay (Fig. 7.26). Fan sediments are primarily sand with some shell and are generally well-stratified in near horizontal layers.



**FIGURE 7.25**

Diagram showing washover fan and its sub-environments on the back-barrier area. (From Scott, A. J. et al, 1969, Effects of Hurricane Beulah, 1967, on Texas lagoons and barriers, *UNAM-UNESCO Symposium on Coastal Lagoons*, Mexico City, pp. 221–236)



**FIGURE 7.26**

Aerial photo showing washover fan extending into the back-barrier bay, Dauphin Island, Alabama.

(Courtesy of Dag Nummedal)

The sediment that accumulates in the washovers comes from the surf zone, beach, or even dunes and is carried by the combination of waves and unidirectional currents that overtop low barriers and travel relatively low paths. Many barriers have foredune systems that are too high to permit washover to take place. It is this condition that demonstrates dune protection for the more landward portion of the barrier.

The vegetation in the landward part of the barrier tends to be fairly resilient. Complete or partial burial by washover fans typically has only a modest effect, and the upland or marsh grasses that live here grow up through the fan sediment in several weeks to a few months (Fig. 7.27). The very thick fans may completely destroy the underlying plant community but the opportunistic nature of barrier island vegetation colonizes their surfaces in generally less than a year, depending upon climate. Even after the washover fan is completely vegetated, its general shape is detectable from aerial photographs, especially if there has been incursion into the bay on the landward side (Fig. 7.28).

It is the postdepositional phase of the washover fans that is fairly complicated. Once the fan has been deposited, it soon begins to be transformed. Partially buried plants penetrate upward through the fan, and the downward penetration of roots from surface plants causes disruption of the typically well-developed stratification of fan sediments. The part of the fan that extends into



**FIGURE 7.27**

Photo showing partial burial of vegetation by washover fan. Typically the vegetation is resilient and will grow up through the fan eventually causing it to lose its definition.



**FIGURE 7.28**

Relict washover fan that extends into the bay on the Delaware coast but is vegetated.  
(Courtesy of J. C. Kraft)

the intertidal zone or below is also rapidly occupied by a broad spectrum of benthic organisms that include marsh grasses and burrowing organisms. Bioturbation also destroys the stratification signature of the washover fan and contributes a small amount of fine sediment through the accumulation of fecal pellets. Additionally, the tidal and wave processes that act on the distal end of the fan cause sediment to be reworked. All of these processes may transform the package of sediment that was deposited as a washover fan into environments such as tidal flats, marshes, wind tidal flats, and back-island areas that have little or no resemblance to the original depositional character of a washover fan.

## Tidal Inlets and Tidal Deltas

Barrier islands do not continue without interruption. They are cut at various locations by tidal inlets that are the most important link between the open marine environment and the environments landward of barrier islands. After beaches, the tidal inlets are probably the most dynamic part of the barrier island system. Along many coasts where they occur, inlets may range widely in size, stability, and water flux. They owe their origin to a variety of circumstances with storms and human activity among the most important.

**Elements of Tidal Inlets** There are basically three major parts to a tidal inlet: (1) the inlet throat, the channel between the adjacent barrier islands and through which the tidal flux passes, (2) the ebb tidal delta, and (3) the flood tidal delta (Fig. 7.29). Tidal deltas are accumulations of sand that vary greatly in size and shape and that are located at the landward and seaward end of the inlet channel. They have many similarities to a fluvial delta in that they are caused by a sudden loss of the ability of currents to carry sediment at the end of a channel, resulting in accumulation of sediment. Because of tidal flow, the short inlet channel has a mouth at each end.

The main channel of the inlet is essentially as long as the width of the barrier island through which it cuts. It is deepest at the narrowest part, usually near the middle of its length, and the cross section commonly shows some asymmetry that typically has the deep side in the direction of the net longshore transport. The amount of asymmetry is generally proportional to the stability of the inlet; the more it migrates the more likely the channel is to be asymmetrical. Some quite stable inlets may have nearly symmetrical channels. The size of inlets ranges from only tens of meters to a few kilometers in width and up to depths of more than 30 meters.

Much of our terminology and understanding of the morphodynamics of tidal inlets and especially tidal deltas, comes from the work of Miles O. Hayes and his students who have provided most of the pioneering work on the dynamics of inlet systems in the late 1960s and early 1970s.

The two types of tidal deltas are best considered separately because they are