Coast and Shore



Coastal areas join land and sea. The place where ocean meets land is usually called the **shore,** while **coast** refers to the larger zone affected by the processes occurring at this boundary. A sandy beach might form the shore in an area, but the coast (or coastal zone) includes the marshes, sand dunes, and cliffs just inland of the beach, as well as the sandbars and troughs immediately offshore. The world ocean is bounded by about 440,000 kilometers (273,000 miles) of shore.

Because of its proximity to both ocean and land, a coast is subject to natural events and processes common to both realms. A coast is an active place. Here is the battleground on which wind waves break and expend their energy. Tides sweep water on and off the rim of land, rivers drop most of their terrigenous sediments at the coasts, and ocean storms batter the continents. The *location* of a coast depends primarily on global tectonic activity and the volume of water in the ocean. The *shape* of a coast is a product of many processes: uplift and subsidence, the wearing down of land by erosion, and the redistribution of material by sediment transport and deposition.

Classifying Coasts



A good classification scheme allows us to group things into categories determined by common origins or the influence of common processes. The process of categorizing things often makes them easier to comprehend. Coasts have been hard to classify because, as we will see, many factors combine to shape them.

As we saw in Chapter 3, no area of geology has been left undisturbed by the revelations of plate tectonics. In the 1960s geologists began to classify coasts according to tectonic position. *Active* coasts, near the leading edge of moving continental plates, were found to be fundamentally different from the more *passive* coasts near trailing edges. The shapes, composition, and ages of coasts are better understood by taking plate movements into account. But the slow forces of plate movement are often obscured by the more rapid action of waves, the erosion of land, and the transport of sediments.

Another important factor that must be considered in the classification of coasts is long-term change in sea level. Five factors can cause sea level to change. Three of these factors are responsible for **eustatic change**—variations in sea level that can be measured all over the world ocean:

- The amount of water in the world ocean can vary. Sea level is lower during periods of global glaciation (ice ages) because there is less water in the ocean. It is higher during warm periods, when the glaciers are smaller. Periods of abundant volcanic outgassing or an increase in the number of icy comets striking Earth can also add water to the ocean, which raises sea level.
- The volume of the ocean's "container" can vary. As noted in Chapter 4, high rates of seafloor spreading are associated with the expansion in volume of the mid-ocean ridges. This displaces the ocean's water, which climbs higher on the edges of the continents. Sediments shed by the continents during periods of rapid erosion can also decrease the volume of ocean basins and raise sea level.
- The water itself may occupy more or less volume as its temperature varies. During times of global warming, seawater expands and occupies more volume, raising sea level.

Of course the continents rarely stay still as sea level rises and falls. Local changes are bound to occur, and two other factors produce variations in *local* sea level:

- Tectonic motions and isostatic adjustment can change the height and shape of the coast. Coasts can experience uplift as lithospheric plates converge, or can be weighted down by masses of ice during a period of widespread glaciation. The continents slowly rise when the ice melts.
- Wind and currents, seiches, storm surges, an El Niño or La Niña event, and other effects of water in motion can force water against the shore or draw it away.

Sea level has been at its current elevation (give or take 0.5 meter, or 1.5 feet) for only about 2,500 years. Over the past 2 million years, worldwide sea level has varied from about 6 meters (20 feet) above to about 125 meters (410 feet) below its present position. The recent low point occurred about 18,000 years ago at the height of the most recent glaciation (see Figure 4.14). Indeed, sea level has been at the modern "high" only rarely in the past 2 million years—the predominant state for Earth is a much lower eustatic sea level position. It is important to realize that coastlines have not yet come into equilibrium with modern sea level.

Changes in sea level produce major differences in the position and nature of coastlines, especially in areas where the edge of the continent slopes gradually or where the coast is rising or sinking. **Figure 12.1** shows an estimate of previous shore positions along the southeastern coast of the United States in the geologically recent past and a prediction for the distant future should the present warming trend cause more of the polar ice to melt.

What coastal classification scheme best combines these elements? One of the most useful was popularized by Francis Shepard and is used by many geological oceanographers: *primary* and *secondary coasts.* If a coast is essentially in almost the same condition as it was when sea level stabilized after the last ice age, Shepard called it a **primary coast.** Primary coasts are young coasts in which terrestrial influences (that is, processes that occur at the boundary between land and air) dominate. If the coast has been significantly changed by wave action and other marine processes since sea level stabilized, he termed it a **secondary coast.** Secondary coasts are usually older than primary coasts—they have been exposed to marine

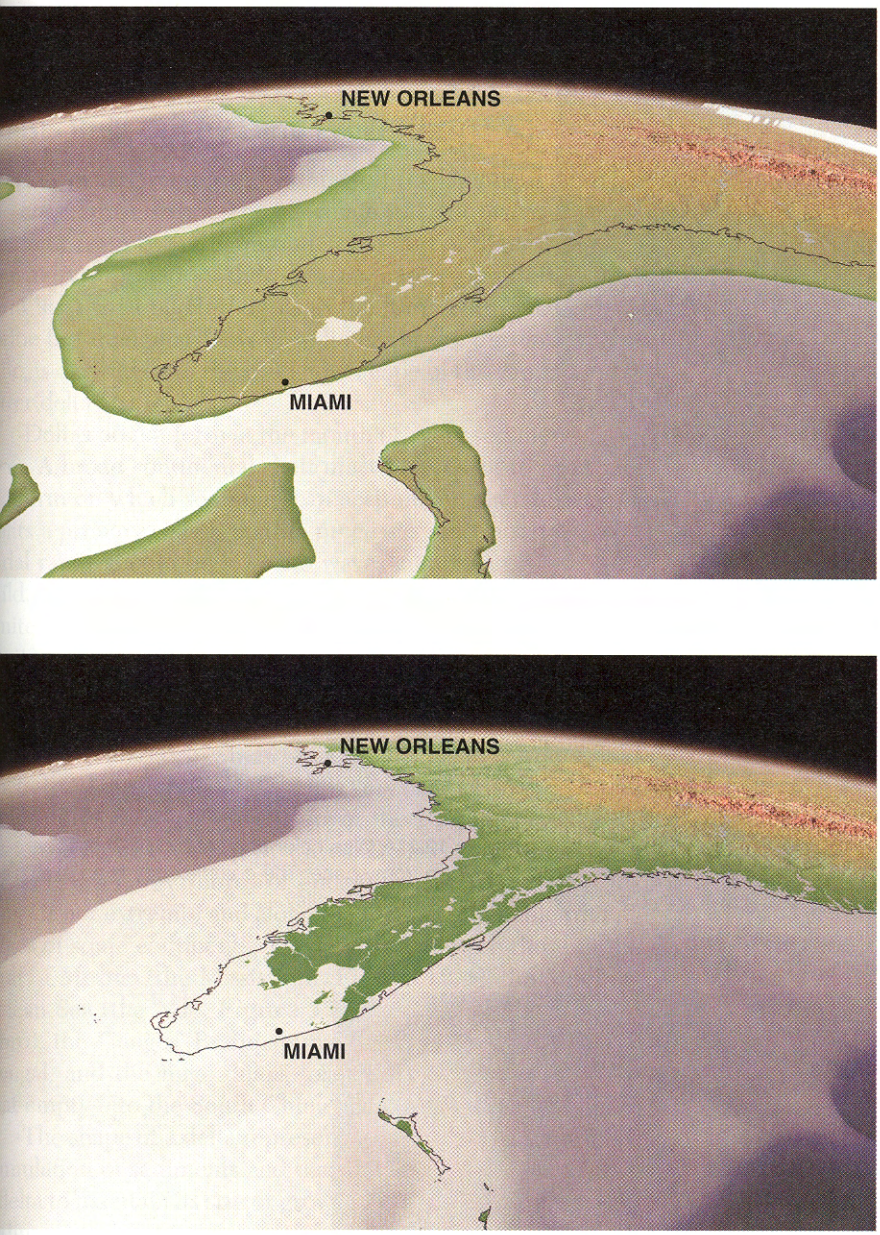


Figure 12.1 (**a**) The southeastern coast of the United States looked much different 18,000 years ago, during the last ice age. Because of lower sea level, the position of the gently sloping eastern coast has been as much as 200 kilometers (125 miles) seaward from the present shoreline, leaving much of the continental shelf exposed. (**b**) In the distant future, if the ocean were to expand and the polar ice caps were to melt because of global warming, sea level could rise perhaps 60 meters (200 feet), driving the coast inland as much as 250 kilometers (160 miles).

tion for a longer time. Secondary coasts retain little (if any) idence of the nonmarine processes that produced them. me coasts show both characteristics: A single, long shore- e can consist of both primary and secondary coasts.

RIMARY COASTS



imary coasts are often rough and irregular. The ocean has t had time to modify the terrestrial features provided by anges in sea level, the scouring of glaciers, deposition of diment at the mouths of rivers, volcanic eruptions, or the ovement of earth along faults.

Land Erosion Coasts



When sea level was lower during the last glaciations, rivers cut across the land and eroded sediment to form coastal river valleys. When higher sea level returned, the valleys were flooded, or *drowned*, with seawater. Chesapeake Bay (**Figure 12.2**) and the Hudson River valley are examples of drowned river mouths. Coasts dominated by these sunken river valleys are sometimes called ria coasts, from the Spanish word for river.

Glaciers sometimes form in river valleys when rivers cut through the edges of continents at high latitudes. Deep, narrow bays known as **fjords** are often formed by tectonic forces and later modified by glaciers eroding valleys into deep, U-shaped troughs. Fjords (**Figure 12.3**) are found in British

Columbia, Greenland, Alaska, Norway, New Zealand, and other cold, mountainous places.

Coasts Built Out by Land Processes



In a few places, sediments washing off the land have built out the coasts extensively. The shoreline in such places is much different from its configuration at the end of the last ice age. The most important of these coastal features are **deltas.** This term, first used by Herodotus in the fifth century B.C. to describe the triangular mass of sediment at the terminus of the Nile, is derived from the triangular shape of the capital Greek letter delta (Δ).

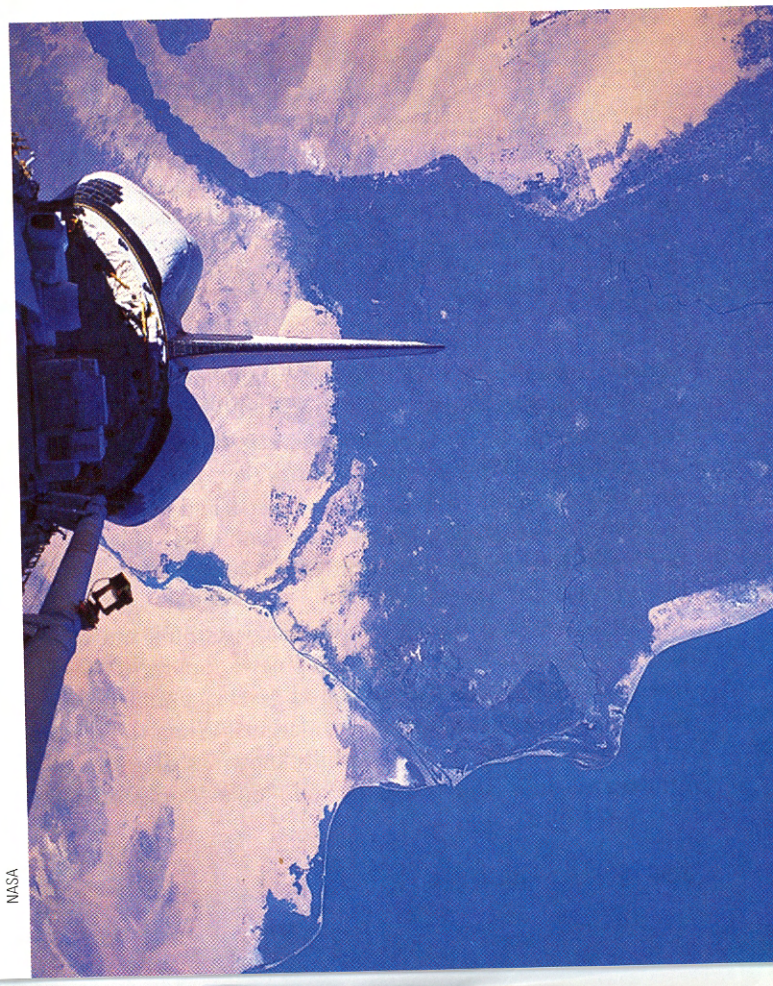
Deltas do not form at the mouth of every sediment-laden river. A broad continental shelf must be present to provide a platform on which sediment can accumulate. In addition (as befits a primary coast), marine processes must not dominate: Tidal range should be low, and waves and currents generally mild. There are no large deltas along the Atlantic coast of the United States because sediments arriving at the coast are deposited in the sunken river mouths or dispersed by tides and currents. Also, there are no large deltas along the western margins of North and South America because these coasts are converging margins, where an oceanic plate is being subducted and the continental shelf is very narrow; sediment that would form a delta is swept down the continental slope or dispersed along the coast by waves. Deltas are most common on the low-energy shores of enclosed seas (where the tidal range is not extreme) and along the tectonically stable trailing edges of some continents. The largest deltas are those of the Gulf of Mexico (the Mississippi, **Figure 12.4**), the Mediterranean Sea (the Nile, **Figure 12.5,** and the Rhône, Po, and Ebro), the Ganges–Brahmaputra river system in the Bay of Bengal, and the huge deltas formed by the rivers of China that empty into the South China Sea.

The shape of a delta represents a balance between the accumulation of sediments and their removal by the ocean. For a delta to maintain its size or grow, the river must carry enough sediment to keep marine processes in check. The combined effects of waves, tides, and river flow determine the shape of a delta. In 1975 William Galloway, a geologist at the University of Texas, classified deltas by the relative influence of those three factors. (See **Figure 12.6.** Note that any delta strongly affected by only one of the three factors is placed at an apex of the triangle. A delta with a balance of forces is placed near the middle of the triangle.) *River-dominated deltas* are fed by a strong flow of fresh water and continental sediments, and they form in protected marginal seas. They terminate in a well-developed set of *distributaries*—the split ends of the river—in a characteristic bird's-foot shape (as shown in the Mississippi, Figure 12.4). In *tide-dominated deltas*, freshwater discharge is overpowered by tidal currents that mold sediments into long



USGS

Figure 12.4 River deltas form at places where sediment-laden rivers enter enclosed or semienclosed seas, where wave energy is limited. The bird's-foot shape of the Mississippi Delta is seen clearly in this photograph. Submerged overlapping lobes of the Mississippi Delta extend completely across a wide continental shelf, and sediment from the river is now being deposited directly on the continental slope. Lobed and bird's-foot deltas form where deposition overwhelms the processes of coastal erosion and sediment transportation. The sediment-laden water looks brown or tan in this photograph taken from low orbit.



NASA

Figure 12.5 The Nile Delta, which has been smoothed into a more rounded shape by postdepositional coastal erosion and sediment transportation. (The tail of the space shuttle from which this photo was taken can be seen in the upper left of the photo.)



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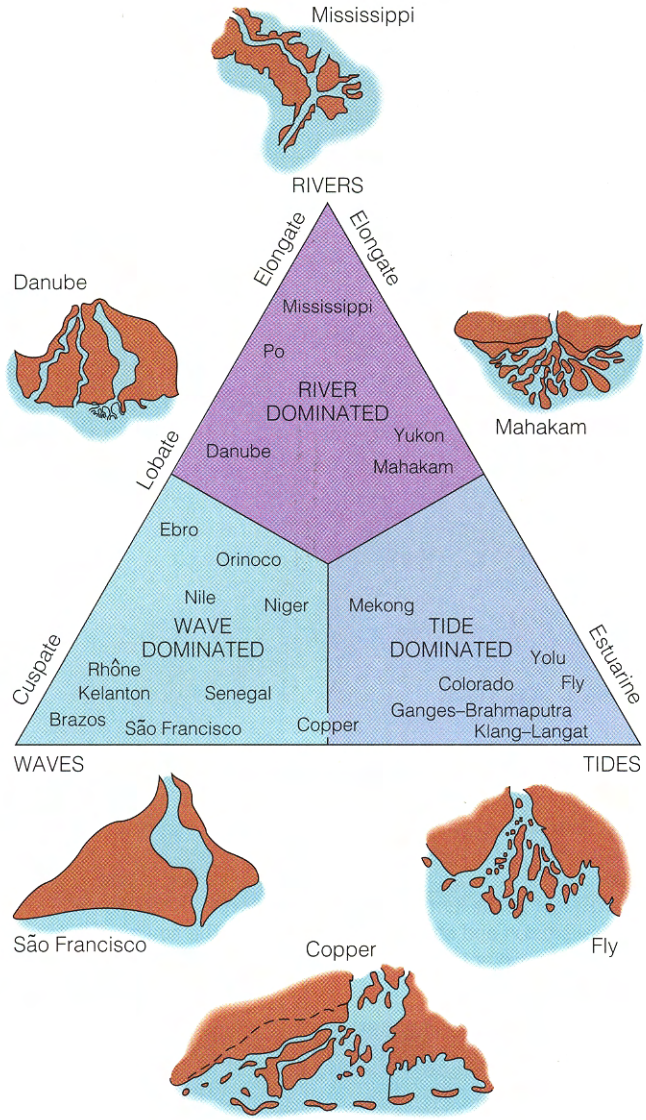


Figure 12.6 The Galloway classification of deltas. Each delta reflects the balance between the rate of sediment delivery at the river mouth and the mechanisms of its dispersal. This triangular diagram classifies river deltas according to the influence of the three major factors affecting their development: the river, waves, and tides. (Source: Used by permission of Dr. William Galloway, University of Texas, Austin and the Houston Geological Survey.)

islands parallel to the river flow and perpendicular to the trend of the coast. The largest tide-dominated delta has formed at the mouths of the Ganges–Brahmaputra river system on the Bay of Bengal. *Wave-dominated deltas* are generally smaller than either tide- or river-dominated deltas and have a smooth shoreline punctuated by beaches and sand dunes. Instead of a bird's-foot pattern of distributaries, a wave-dominated delta has one primary exit channel.

Deltas are not the only types of coasts built out by the land. The glaciers that covered the poleward parts of the continents during the ice age deposited great quantities of sediments and rocks near their outer margins. When the glaciers retreated, they left streamlined hills known as **drumlins** and **moraines**—hills and ridges of sediments—some of which still stand above sea level. Part of Long Island, New York, is a glacial moraine, as are the oval-shaped hills of Boston and sections of the Puget Sound coast at Seattle. Perhaps the most famous glacial moraine in the United States is the area around Cape Cod, Massachusetts. The cores of the islands of Martha's Vineyard and Nantucket represent the farthest advance of the glaciers around 18,000 years ago, and the spine of the cape itself is a remanent of material dropped by the glacier as it stabilized and then retreated northward when the climate warmed (see **Figure 12.7**).

Volcanic Coasts



As we saw in Chapter 4, most islands that rise from the deep ocean are of volcanic origin. If the volcanism has been recent, the coasts of a volcanic island will consist of lobed lava flows extending seaward, common features in the Hawaiian Islands. Concave shorelines (**Figure 12.8**) result when a small coastal volcano explodes or collapses and seawater fills the crater.

Coasts Shaped by Earth Movements



Coasts can coincide with places where Earth's crust is being warped or faulted. When the seabed on the seaward side of a coastal fault moves *downward*, a steep escarpment that continues to a greater depth than a wave-cut cliff can result. When the landward side of the fault moves *upward*, the part of the

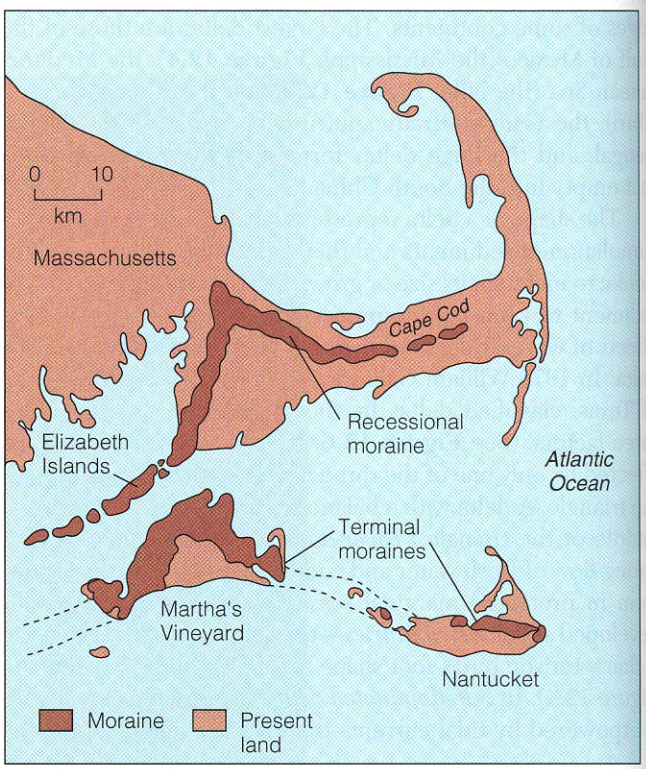


Figure 12.7 Most of the islands of Martha's Vineyard and Nantucket were formed from debris deposited or shaped by an advancing glacier. The spine of Cape Cod is built of material dropped as the glaciers stabilized and then retreated.



John S. Shelton

Figure 12.8 Concave shapes may also occur on a volcanic coast. The explosion and collapse of a volcano close to shore caused this indentation on the coast of Isla Encantada, Gulf of California, Mexico.



© George P. Lafaker, USGS

Figure 12.9 This former seafloor at Prince William Sound, Alaska, was raised 8 meters (26 feet) above sea level by tectonic uplift during the great earthquake of 27 March 1964. The exposed surface, which slopes gently from the base of the sea cliffs to the water, is about 400 meters (¼ mile) wide. The light-colored coating on the rocks consists mainly of the dried remains of small marine organisms. The photo was taken at about 0.0 tide, 30 May 1964.

coast previously submerged during high tides can be left high and dry. The 1964 Alaska earthquake (described in Chapter 3) caused parts of the Prince William Sound shore to rise more than 3.5 meters (12 feet). As can be seen in **Figure 12.9,** in some places a band of the old sea-cut bench nearly half a kilometer (¼ mile) wide was exposed, even at high tide.

Fault coasts sometimes occur along transform faults. The Pacific coast of North America provides two striking examples: the Gulf of California (located along the San Andreas Fault between Baja California and the mainland of Mexico) and the area around Point Reyes, north of San Francisco. Baja California was once a part of the North American continent, but movement at the boundary between the Pacific and North American Plates has torn the finger of land on the west side of the fault from the land mass to the east. The young, narrow, straight gulf formed as seawater intruded. As can be seen in **Figure 12.10,** a photo of Tomales Bay north of San Francisco, fault coasts can be startlingly straight.

Secondary Coasts



Secondary coasts are those coasts that have been significantly changed by wave action and other marine processes after sea level has stabilized. Land erosion and marine erosion both work to change a primary coast to a secondary coast. Secondary coasts are shaped and attacked from the land by stream erosion, the abrasion of wind-driven particles, the alternate freezing and thawing of water in rock cracks, the probing of plant roots, glacial activity, rainfall, dissolution by acids from soil, and slumping. (As any mountaineer will confirm, these factors are not limited to coasts.)

Attack from the sea is by waves and currents. The continual onslaught of waves does most of the work, with currents distributing the results of the waves' labor. Some indication of the violence of this activity may be inferred from **Figure 12.11.**

Large storm surf routinely generates pressures of 27,000 kilograms per square meter (25 tons per square yard). The crashing waves push air and water into tiny rock crevices. The repeated buildup and release of pressure within these crevices can weaken and fracture the rock. But it is not the hydraulic pressure of moving water alone that abrades the coasts. Tiny pieces of sand, bits of gravel, or stones hurled by waves toward the shore are even more effective at eroding the shore. **Dissolution,** the dissolving of minerals in the rocks by water, contributes to the erosion of easily soluble coastal rocks such as limestone. Even the digging and scraping activities of marine organisms have an effect.

The rate at which a shore erodes depends on the hardness and resistance of the rock, the violence of the wave shock to which it is exposed, and the local range of tides. Hard rock resists wear. Coasts made of granite or basalt may retreat an insignificant amount over a human lifetime; the granite coast of Maine erodes only a few centimeters per decade. Coasts of soft sandstone or other weak (or soluble) materials, however, may disappear at a rate of a few meters per year. In an extreme case, soft shore cliffs facing the North Sea in East Suffolk, England, eroded more than 11 meters (36 feet) during two hours of a severe storm!

Marine erosion is usually most rapid on **high-energy coasts,** areas frequently battered by large waves. High-energy coasts are most common adjacent to stormy ocean areas of great fetch and along the eastern edges of continents exposed to tropical storms. The coasts of Maine and British Columbia and the southern tips of South America and South Africa are typical high-energy coasts. **Low-energy coasts** are only infrequently attacked by large waves. Because of their generally protected location in the Gulf of Mexico, the U.S. Gulf states share a low-energy coast—at least between hurricanes!

Waves can affect the coast only where they strike, so erosion is concentrated near average sea level. A shore with little tidal variation can erode quickly because the wave action is concentrated near one level for longer times. Low-energy coasts protected by offshore islands usually erode slowly, as do areas below the low-tide line. Some erosion does occur below the surface because of the orbital motion of water in waves, but even the largest waves have little erosive effect at depths greater than about 15 meters (50 feet) below average sea level. Cliffs above shore are subject to pounding either directly from waves or by rocks hurled by waves. In one case, windows of a Scottish lighthouse 100 meters (330 feet) above sea level were broken by wave-thrown rocks during an Atlantic winter storm.

Some Features of Secondary Coasts



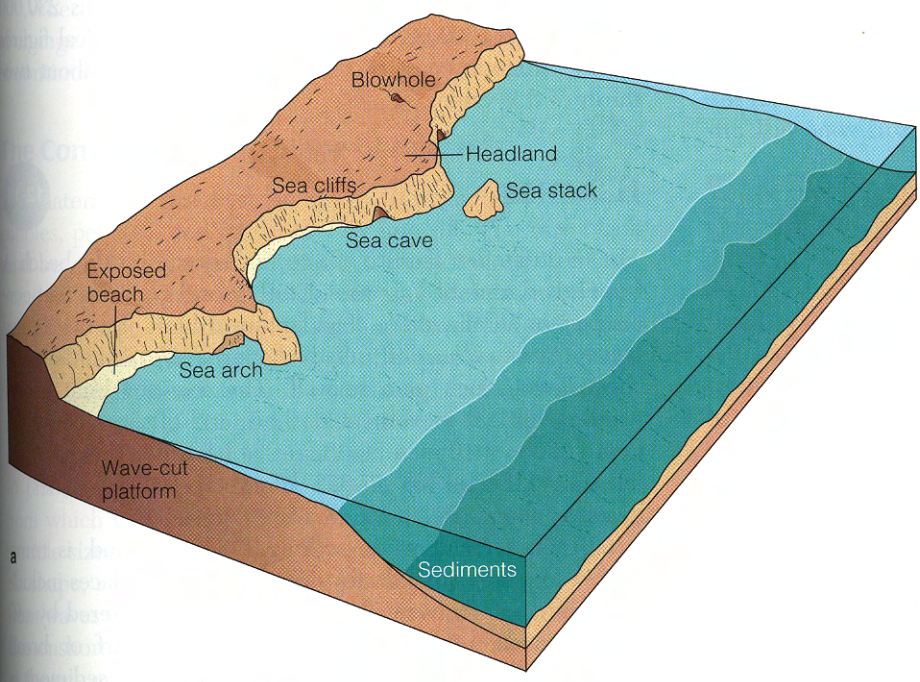
Erosive forces can produce a wave-cut shore showing some or all of the features illustrated in **Figure 12.12.** Note the complex, small-scale irregularities of this rocky erosional coastline. **Sea cliffs** slope abruptly from land into the ocean, their steepness usually resulting from the collapse of undercut notches. The position of the sea cliffs marks the shoreward limit of marine erosion on a coast. The parade of waves cuts **sea caves** into the cliffs at local zones of weakness in the rocks. Most sea caves are accessible only at low tide. A blowhole can form if erosion follows a zone of weakness upward to the top of the cliff. When the tide is at just the right height, spray can blast from the fissure as waves crash into the cliff. Offshore features of rocky coasts can include natural arches, rock stacks, and a smooth, nearly level **wave-cut platform** just offshore, which marks the submerged limit of rapid marine erosion. Much of the debris removed from cliffs during the formation of these structures is deposited in the quieter water farther offshore, but some can rest at the bottom of the cliffs as exposed beaches. Indeed, as we shall soon see, broad beaches are often features of secondary coasts.

Shore Straightening



The first effect of marine erosion on a newly exposed coast is to intensify the irregularity of the coastline. This happens because coastal rocks are usually not uniform in composition over long horizontal distances. Some hard rocks will resist erosion well, while softer rocks on the same coast may disappear almost overnight. (This explains the uneven character of stacks, arches, and sea cliffs.)

Eventually, however, coastal erosion tends to produce a smooth shoreline. Because of wave refraction (see Figure 10.20), wave energy is focused onto headlands and away from bays (**Figure 12.13**). Sediment eroded from the headlands tends to collect as beaches in the relatively calm bays. As erosion continues, the deposits may eventually protect the base



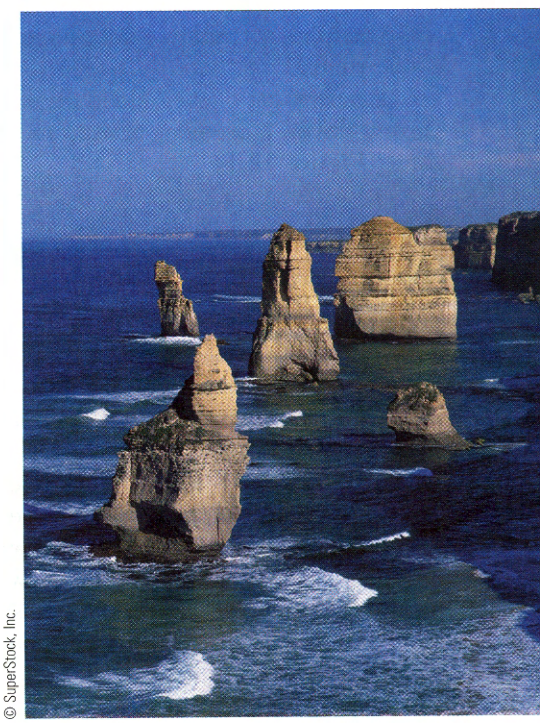


Figure 12.12 (**a**) Features of a secondary coast at low tide. (**b**) Sea stacks off the Australian coast.

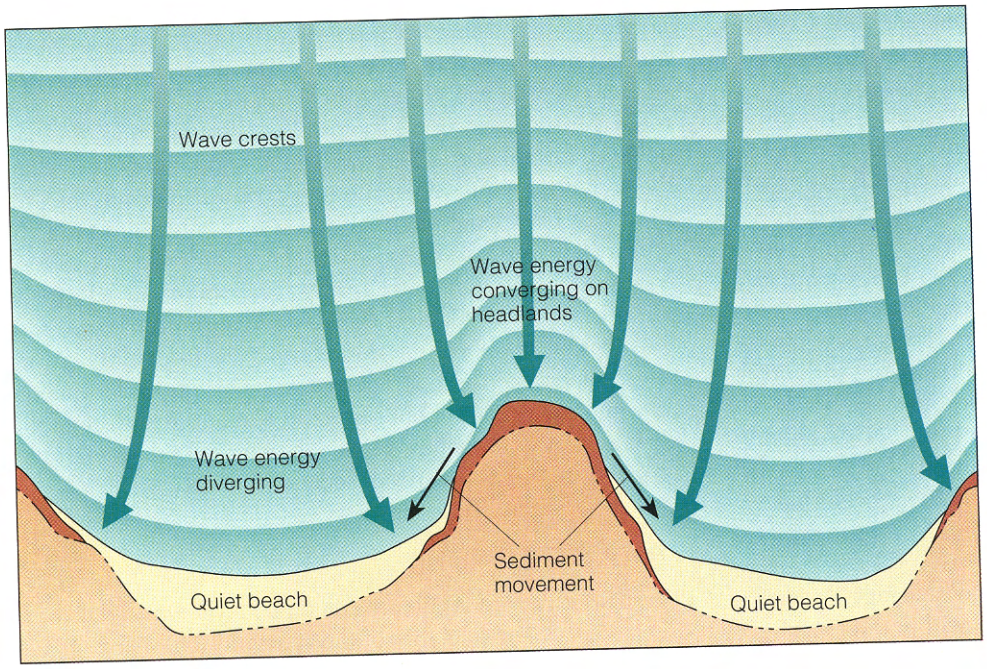


Figure 12.13 Wave energy converges on headlands and diverges in the adjoining bays. The accumulation of sediment derived from the headland in the tranquil bays eventually smooths the contours of the shore. See Figure 10.20 for a discussion of the process of wave refraction responsible for this pattern.

of the shore cliffs from the waves. Coastal irregularities are thus smoothed with the passage of time. As you might expect, straightening occurs most rapidly on high-energy coasts.

The Accumulation of Beaches



If the submerged slope of the seafloor is steep, eroded sediments will soon drain to deeper waters. If the slope is not too steep, sediments will be transported along the coast by wave and current action. The movement of sediment (usually sand) along the coast, driven by wave action, is referred to as **longshore drift.** Longshore drift occurs in two ways: the wave-driven movement of sand along the exposed beach, and the current-driven movement of sand in the surf zone just offshore (**Figure 12.14**).

As we have seen, wind waves arriving from distant storms do not always approach the coast straight on. Most wind waves approach at an angle and then refract in shallow water to break almost parallel to the shore. Refraction is usually incomplete, however, and some angle remains when the waves break. If sediments have accumulated to form a beach, water from the breaking wave will rush up the beach at a slight angle but return to the ocean by running straight downhill under the influence of gravity. The sand grains disturbed by the wave will follow the water's path, moving up the beach at an angle but retreating down the beach straight down the slope. Net transport of the grains is longshore, parallel to the coast, away from the direction of the approaching waves. Net sand flow along the Pacific and Atlantic coasts of the United States is usually to the south, because the waves that drive the transport system usually approach from the north, where storms most commonly occur.

Sediments are also transported in the surf zone in a **longshore current.** The waves breaking at a slight angle distribute a portion of their energy away from their direction of approach. This energy propels a narrow current in which sediment already suspended by wave action can be transported downcoast. The speed of the longshore current sometimes approaches 4 kilometers (about 2 ½ miles) per hour.

Sand moving in the wash of waves along the beach and sediments propelled in the longshore current just offshore are often joined by much greater loads of sediments brought to the coast by local rivers. Net southward transport of all this material along the central California coast exceeds 230,000 cubic meters (300,000 cubic yards) per year. Typical figures for the Atlantic coast of the United States are about two-thirds of this value.

Beaches



The most familiar feature of a secondary coast is the beach. A **beach** is a zone of unconsolidated (loose) particles that covers part or all of a shore. The landward limit of a beach may be vegetation, a sea cliff, relatively permanent sand dunes, or construction such as a seawall. The seaward limit occurs where sediment movement onshore and offshore ceases—a depth of about 10 meters (33 feet) at low tide. The continental United States has 17,672 kilometers (10,983 miles) of beaches, about 30% of the total shoreline.

Beaches result when sediment, usually sand, is transported to places suitable for deposition. Such places include the calm spots between headlands, shores sheltered by offshore islands, and regions with moderate surf or broad stretches of high-energy coast. Sometimes the sediment is transported a very short distance—particles may simply fall from the cliff above and accumulate at the shoreline—

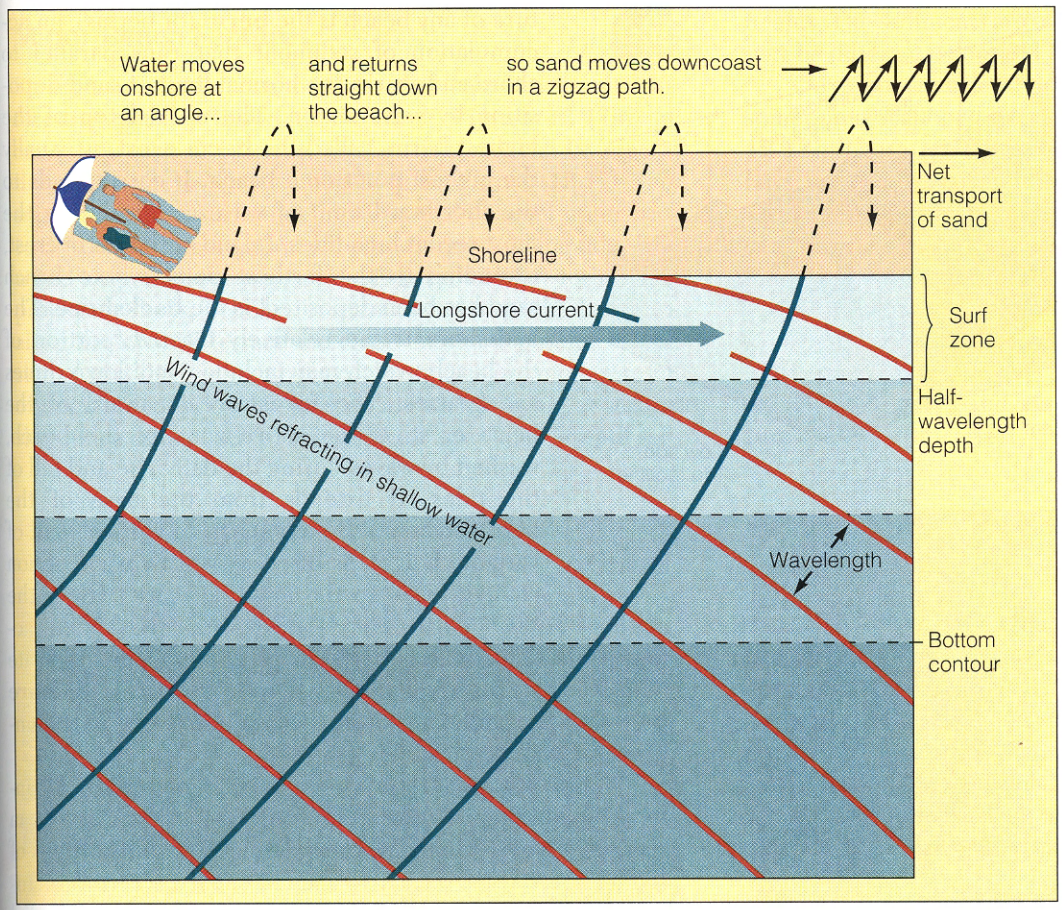


Figure 12.14 How wind-wave refraction and longshore current result in longshore drift of beach sand.

more often the sediment on a beach has been moved over long distances to its present location.

Wherever they are found, beaches are in a constant state of change. They may be thought of as rivers of sand—zones of continuous sediment transport.

The Composition and Slope of Beaches

12-14

The material of beaches can range from boulders through cobbles, pebbles, and gravel to very fine silt. The rare black sand beaches of Hawaii are made of finely fragmented lava. Some beaches consist of shells and shell debris or of fragments of coral. Unfortunately, some also include large quantities of human junk—glass or plastic beaches are not unknown. Cobble beaches can be very steep (occasionally with slopes in excess of 20°), but wide beaches of fine sand are sometimes nearly as flat as a parking lot.

In general, the flatter the beach, the finer the material from which it is made (**Table 12.1**). The relation between particle size and beach slope depends on wave energy, particle shape, and the porosity of the packed sediments. Water from waves washing onto a beach—the **swash**—carries particles onshore, increasing the beach's slope. If water returning to the ocean—the **backwash**—carries back the *same* amount of material as it delivered, the beach slope will be in equilibrium, so the beach will not become larger or steeper.

On fine-grain beaches, the ability of small, sharp-edged particles to interlock discourages water from percolating down into the beach itself, so water from waves runs quickly back down the beach, carrying surface particles toward the ocean. This process results in a very gradual slope. Broad, flat beaches also have a large area on which to dissipate wave energy, and

Table 12.1 The Relationship Between the Particle Size of Beach Material and the Average Slope of the Beach

Type of Beach Material	Size (mm)	Average Slope of Beach
Very fine sand	0.0625–0.125	1°
Fine sand	0.125–0.25	3°
Medium sand	0.25–0.50	5°
Coarse sand	0.50–1.0	7°
Very coarse sand	1–2	9°
Granules	2–4	11°
Pebbles	4–64	17°
Cobbles	64–256	24°

Source: Shepard, 1973.

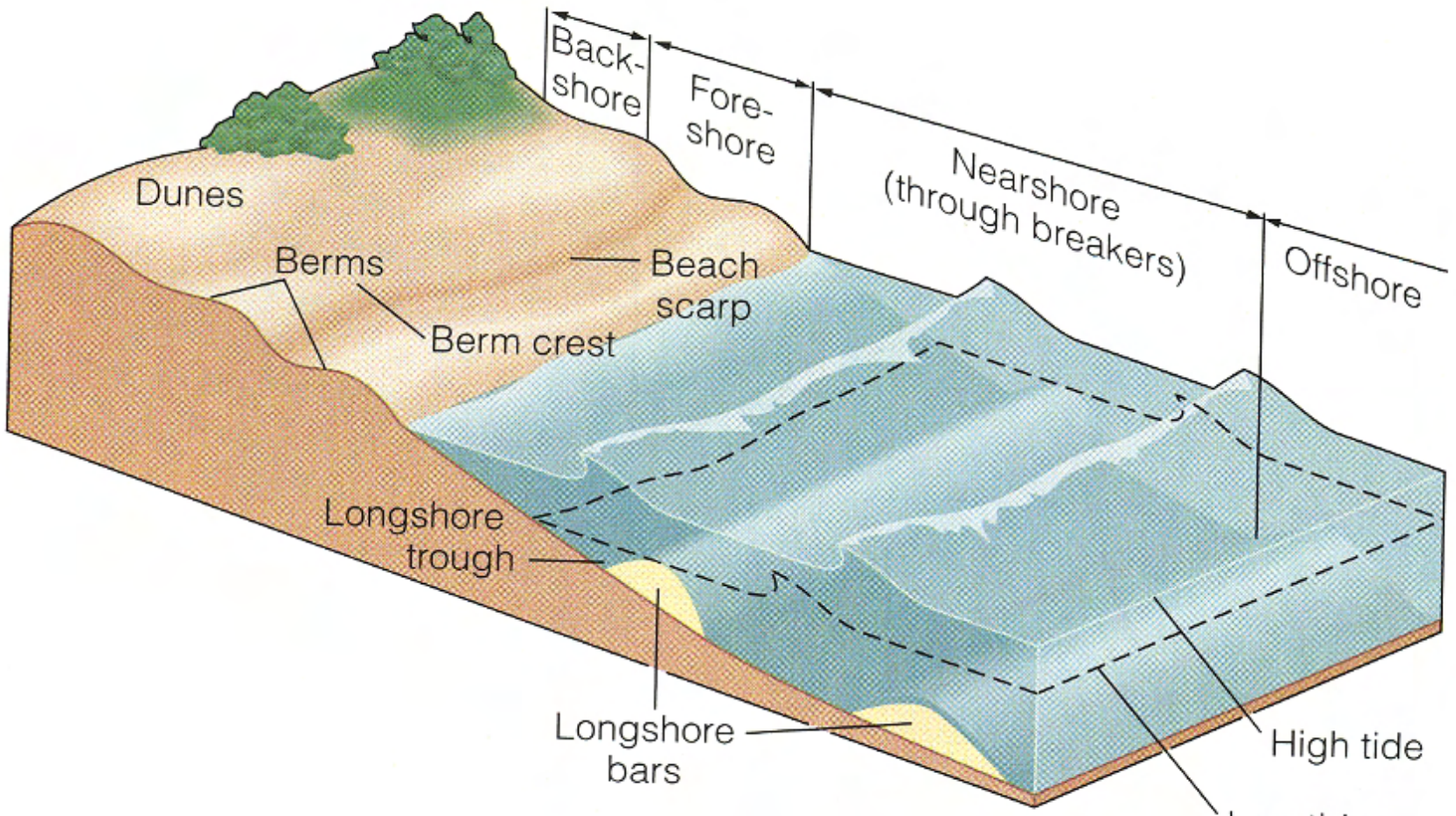
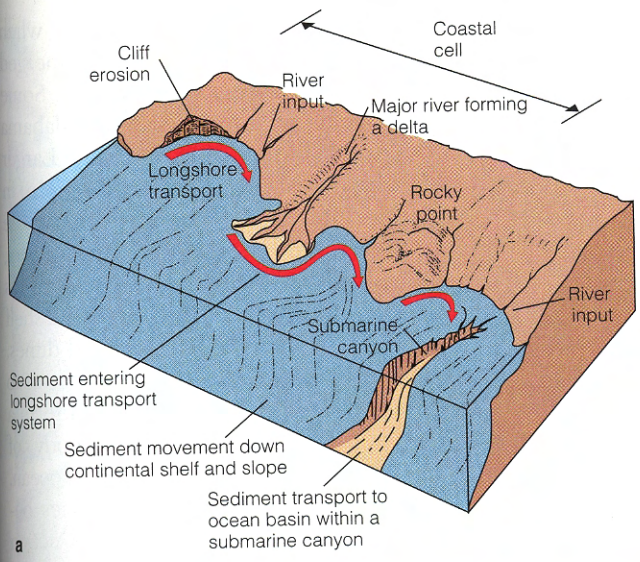


Figure 12.15 A typical beach profile.

they can provide a calm environment for the settling of fine sediment particles. In contrast, coarse particles (gravel, pebbles) do not fit together well and readily allow water to drain between them. Onrushing water disappears *into* a beach made of coarse particles, so little water is left to rush down the slope, thereby minimizing the transport of sediments back to the ocean. Thus, larger particles tend to build up at the back of the beach where they are thrown by large waves, increasing its steepness.

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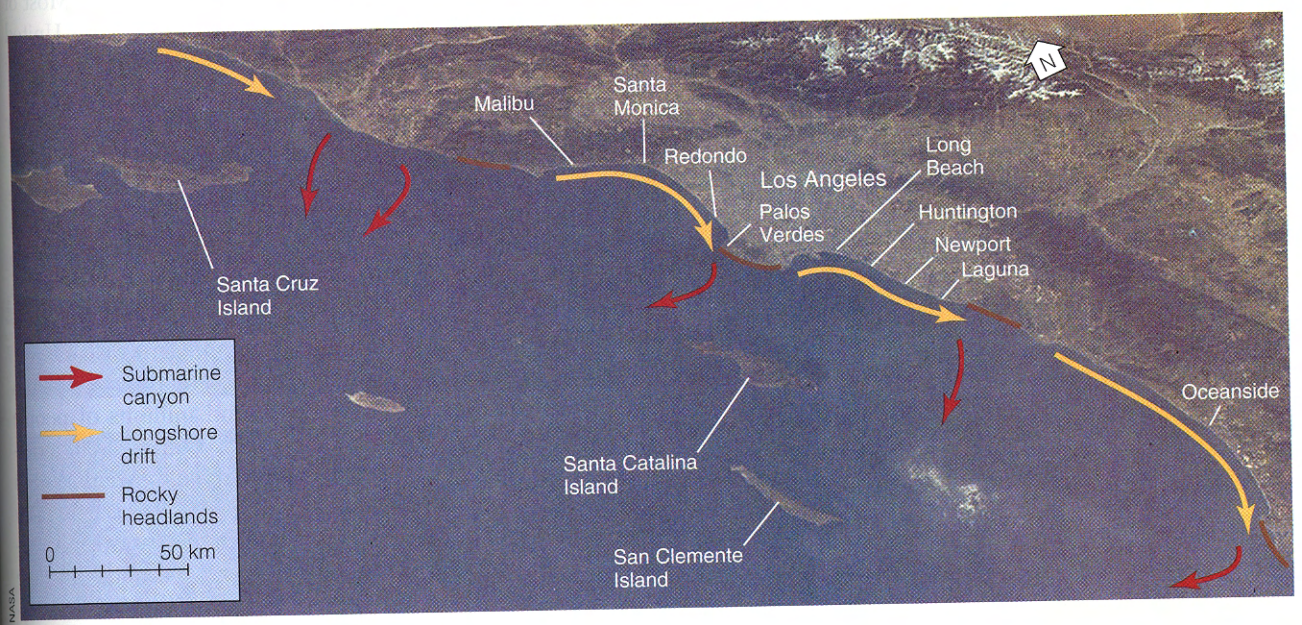


Figure 12.20 Coastal sediment transport cells. (**a**) The general features of coastal cells, in which sand is introduced by rivers, transported southward by the longshore drift, and trapped within the nearshore heads of submarine canyons. (**b**) A series of coastal cells in southern California. The yellow arrows show sand flowing in rivers toward the beaches; the red arrows show sand flowing into submarine canyons.

Large-Scale Features of Secondary Coasts

12-19

We have already noted the presence of small longshore bars off beaches (see Figure 12.15), but secondary coasts, especially the coasts along a subsiding irregular continental margin, also exhibit much larger features. Some of these features are illustrated in **Figure 12.21.**

Sand Spits and Bay Mouth Bars

12-20

Sand spits are among the most common features of secondary coasts. A sand spit forms where the longshore current slows as it clears a headland and approaches a quiet bay. The slower current in the mouth of the bay is unable to carry as much sediment, so sand and gravel are deposited in a line downcurrent of the headland (**Figure 12.22**). As can be seen in Figures 12.21 and 12.22, sand spits often have a curl at the tip. This is caused by the current-generating waves being refracted around the tip of the spit.

A **bay mouth bar** forms when a sand spit closes off a bay by attaching to a headland adjacent to the bay. The bay mouth bar protects the bay from waves and turbulence and encourages the accumulation of sediments there. An inlet—a passage to the ocean—may be cut through a bay mouth bar by tidal action, water flowing from a river emptying into the bay, or heavy rains. A bay mouth bar is shown in **Figure 12.23.**

Barrier Islands and Sea Islands

12-21

Secondary coasts can also develop narrow, exposed sandbars that are parallel to but separated from land. These are known as **barrier islands** (see **Figure 12.24**). About 13% of the world's coasts are fringed with barrier islands.

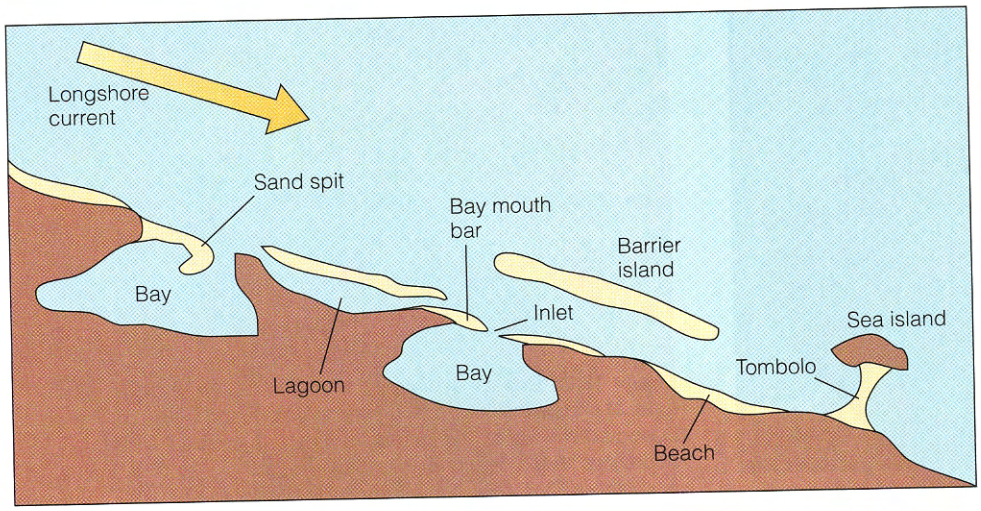


Figure 12.21 A composite diagram of the large-scale features of an imaginary secondary coast. Not all these features would be found in such close proximity on a real coast.

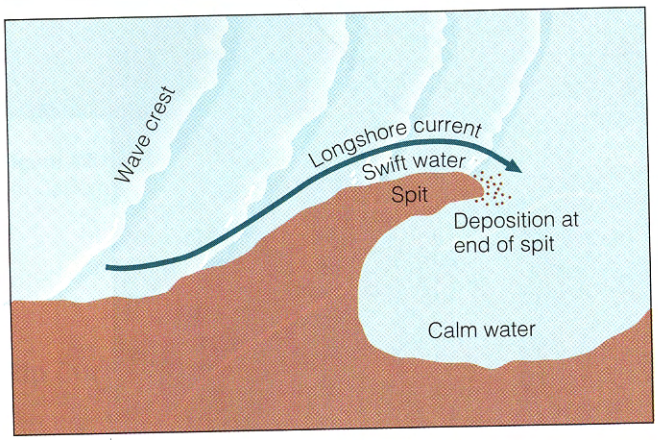


Figure 12.22 A sand spit forms where the longshore current slows as it clears a headland and approaches a quiet bay.



© Jack Mertz

Figure 12.23 A bay mouth bar. The inlet is now closed, but increased river flow (from inland rainfall) or large waves combined with very high and low tides could break through the bar.

Barrier islands can form when sediments accumulate on submerged rises paralleling the shoreline. Some islands off the Mississippi–Alabama coast developed in this way. Larger barrier islands are thought to form in a different way, however. Near the end of the last major rise in sea level, about 6,000 years ago, coastal plains near the edge of the continental shelf were fronted by lines of sand dunes. Rising sea level caused the ocean to break through the dunes and form a **lagoon**—a long, shallow body of seawater isolated from the ocean. The high lines of coastal dunes became islands. As sea level continued to rise, wave action caused the islands and lagoons to migrate landward. Most of the barrier islands off the southeastern coast of the United States originated in this way. They are still migrating slowly landward as sea level continues to rise.

There are 295 barrier islands along the Atlantic and Gulf coasts of the United States, with a combined length of 2,591 kilometers (1,610 miles). Every year severe storms generate waves intense enough to erode barrier island beaches. The largest of these storms can generate waves that overwash the low islands. Runoff from rivers swollen by rains, coupled with water driven by wind waves and storm surge, can rapidly flood a lagoon and cut new inlets through barrier islands.

Despite these dangers, about 70 of the barrier islands have been commercially developed, and millions of people live on them. The most famous barrier islands include Atlantic City, New Jersey; Ocean City, Maryland; Miami Beach and Palm Beach, Florida; and Galveston, Texas. Roughly once every hundred years a winter storm has catastrophic effects on populated areas of Atlantic barrier islands, and the southeastern Atlantic and Gulf coasts must contend with occasional large hurricanes.[1] The rise in sea level and continuing subsidence of these passive coasts (combined with changes caused by commercial development and the ongoing rise in sea level) will undoubtedly cost lives and destroy property. **Figure 12.25** suggests the extent of the threat.

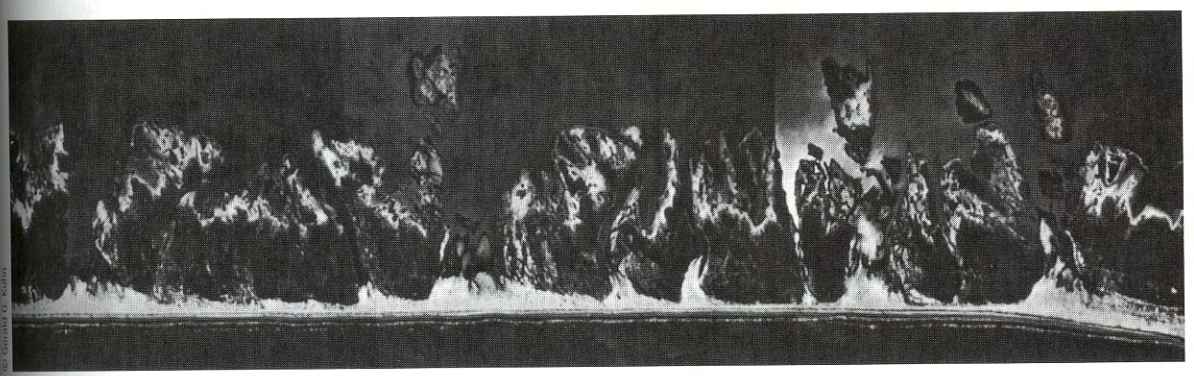
Unlike barrier islands, **sea islands** contain a firm central core that was part of the mainland when sea level was lower. The rising ocean separated these high points from land, and sedimentary processes

[1]Galveston, Texas, was the site of the worst natural disaster in U.S. history. On 8 September 1900 more than 6,000 people died when a hurricane-driven storm surge swept the barrier island.



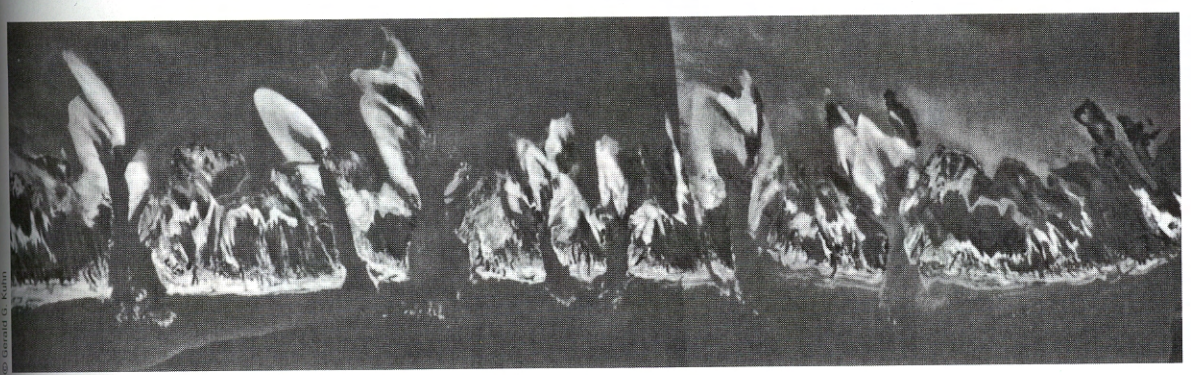
NASA

Figure 12.24 Padre Island and Laguna Madre, barrier islands off the Texas Gulf coast. (This photo, taken from space, is on a much larger scale than Figure 12.23.)



© Gerald G. Kuhn

a



© Gerald G. Kuhn

b



© W. Demetrakas, O.C. Camara

c

Figure 12.25 Barrier island modification, actual and potential. (**a**) The beach extending along the Matagorda Peninsula (Texas) barrier in September 1960. (**b**) The same area six days after the passage of Hurricane Carla in September 1961. The beach and island have been breached, and washover deltas are clearly seen. (**c**) Ocean City, Maryland, a developed barrier island. Host to 8 million visitors a year, this city (and others similarly situated) has no effective protection against flooding and damage from severe storms.

surrounded them with beaches. Hilton Head, South Carolina, and Cumberland Island, Georgia, are sea islands. If the island is close to shore, a bridge of sediments called a **tombolo** may accumulate to connect the island to the mainland. Tombolos can also connect offshore rocky outcrops or volcanoes to the mainland. A sea island and a tombolo are shown in Figure 12.21.

Coasts Formed by Biological Activity



Coasts can be extensively modified by the activities of animals and plants. As we will see, some kinds of marine algae and plants can build coasts, but the most dramatic biological modifications occur in the tropics, where coral polyps form reefs around volcanic islands or along the margin of a continent.

Coral Reefs



The greatest of all reefs is the Australian Great Barrier Reef (**Figure 12.26**), which begins in the Torres Strait separating New Guinea and Australia and runs down the northeastern coast of Australia for 2,500 kilometers (1,560 miles). The reef is not a single object, but a composite of more than 3,000 individual coral reefs covering 350,000 square kilometers (135,000 square miles)—collectively the largest structure made by living organisms on Earth.

Coral animals (which you will learn more about in Chapter 15) are related to the familiar sea anemones found along U.S. coasts. Reef-building coral polyps secrete a cup-shaped calcium carbonate skeleton, which remains behind after the animal dies; the accumulation of skeletons gradually forms the reef. These corals grow best in brightly lighted water about 5 to 10 meters (16 to 33 feet) deep, and in ideal conditions they grow at a rate of about 1 centimeter (½ inch) per year.

Coral Reef Types



In 1842 Charles Darwin classified tropical reef structures into three types: fringing reefs, barrier reefs, and atolls. We still use his classification today.

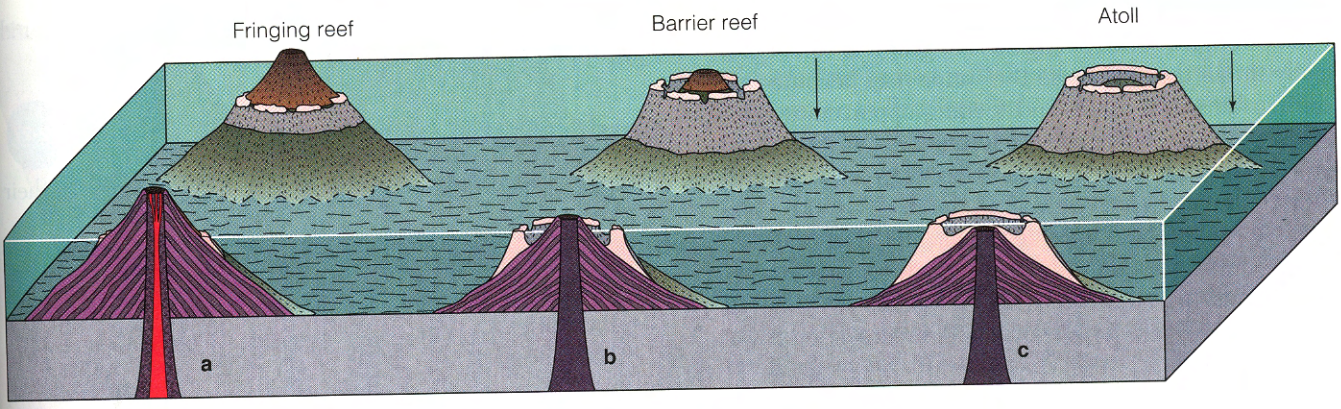
FRINGING REEFS As their name implies, **fringing reefs** cling to the margin of land. As can be seen in **Figure 12.27a**, a fringing reef connects to shore near the water surface. Fringing reefs form in areas of low rainfall runoff primarily on the leeward (downwind) side of tropical islands. The greatest concentration of living material will be at the reef's seaward edge, where plankton and clear water of normal salinity are dependably available. Most new islands anywhere in the tropics have fringing reefs as their first reef form. Permanent fringing reefs are common in the Hawaiian Islands and in similar areas near the boundaries of the tropics.

BARRIER REEFS **Barrier reefs** are separated from land by a lagoon (**Figure 12.27b**). They tend to occur at lower latitudes than fringing reefs and can form around islands or



Great Barrier Reef Marine Park Authority, Queensland, Australia

Figure 12.26 A small section of the Great Barrier Reef, Queensland, Australia. This coast has been extensively modified by biological activity.





d

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Figure 12.27 The development of an atoll. (**a**) A fringing reef forms around an island in the tropics. (**b**) The island sinks as the oceanic plate on which it rides moves away from a spreading center. In this case, the island does not sink at a rate faster than coral organisms can build upward. (**c**) The island eventually disappears beneath the surface, but the coral remains at the surface as an atoll. (**d**) The typical ring shape of an atoll.

lines parallel to continental shores. The outer edge—the barrier—is raised because the seaward part of the reef is supplied with more food and is able to grow more rapidly than the shore side. The lagoon may be from a few meters to 60 meters (200 feet) deep, and it may separate the barrier from shore by only tens of meters or as much as 300 kilometers (190 miles). Coral grows slowly within the lagoon because fewer nutrients are available and sediments and fresh water run off from shore. As you would expect, conditions and species within the lagoon are much different from those of the wave-swept barrier. The calm lagoon is often littered with eroded coral debris moved from the barrier by storms.

The Great Barrier Reef of Australia is the largest biological construction on the planet. It isn't a single reef but a conglomeration of thousands of interlinked segments. The segments present a steep outer wall to the prevailing currents and trade winds. At a growth rate of 1 centimeter (½ inch) per year, the structure is obviously of great age and astonishing volume. The huge reef is younger and thinner at its southern end; the slow northward movement of the Indian–Australian lithospheric plate in which Australia is embedded is thought responsible for this. The variety of organisms within the Australian Barrier Reef staggers the imagination.

ATOLLS An **atoll** (**Figure 12.27 c and d**) is a ring-shaped island of coral reefs and coral debris enclosing, or almost enclosing, a shallow lagoon from which no land protrudes. Coral debris may be driven onto the reef by waves and wind to form an emergent arc on which coconut palms and other land plants take root. These plants stabilize the sand and lead to colonization by birds and other species. Here is the tropical island of the travel posters.

Though an atoll's central lagoon connects to the deep water outside through a series of channels or grooves, coral does not usually thrive there because the water may become too fresh during rains or too hot, and feeding opportunities for the coral in the enclosed lagoon are limited.

Some atolls are isolated, but most occur in loose groups in shallow continental-shelf areas or in the deep open ocean. More than 300 atolls exist—most in the Pacific. They range in size from a few kilometers in diameter to Kwajalein in the Marshall Islands, whose slender 280-kilometer (176-mile) ring of coral encloses a lagoon of 2,850 square kilometers (1,100 square miles).

How do atolls form? Scientists began speculating on the cause of their ring shape soon after the first scientific voyages published their reports. Charles Darwin imagined a volcanic island growing from the sea, accumulating a skirt of coral around its shore, and then slowly subsiding at a rate equal to the growth rate of the coral. The central volcanic island would eventually sink from view, but the coral could grow continuously atop skeletons of past generations to maintain a living presence near the surface. Figure 12.27a–c shows the progression. Note that the island begins with a fringing reef, passes through a barrier-reef stage as it sinks, and eventually becomes an atoll as the peak disappears beneath the ocean surface. Should the island subside faster than about 1 centimeter (½ inch) per year (the growth rate of coral), all traces of both the island and the reefs would disappear. (The submerged island might become a guyot.)

This theory seemed reasonable, but Darwin couldn't explain what would cause volcanic islands to subside because he didn't know about plate tectonics. Now we know that volcanoes can form near spreading centers, ride outward and downward from their birthplaces, cease to be active as they leave their source of mantle heat, and sink as they are carried into deeper water—just slowly enough to permit coral growth to continue as they go.