15 Shoreline Systems

Water in oceans and lakes is in constant motion. It moves by wind-generated waves, tides, tsunamis (seismic sea waves), and a variety of density currents. As it moves, it constantly modifies the shores of all the continents and islands of the world, reshaping coastlines with the ceaseless activity of waves and currents. Shoreline processes can change in intensity from day to day, and from season to season, but they never stop.

Shoreline systems are complex open systems where the principal source of energy is wind-generated waves. Ultimately, the wind's energy is derived from the Sun. Gravity is also an important source of energy in the system—its influence is felt in tides and near-shore currents. The materials in the system include the shore itself, sand (and other shoreline sediment), and seawater.

The world's present shorelines, however, are not the result of present-day processes alone. Nearly all coasts were profoundly affected by the rise in sea level caused by the melting of the Pleistocene glaciers, which began between 15,000 and 20,000 years ago. The rising sea flooded large parts of the low coastal areas, and shorelines moved inland over landscapes formed by continental processes. The shapes of many coastlines are largely the result

of processes other than marine and may owe their outlines to stream erosion or deposition, glaciation, volcanism, tectonism, or even the growth of plants and animals.

The photograph above shows the coast of California's Big Sur. As waves pound the shore, they erode cliffs and cause them to recede from the beach, leaving a wide, flat terrace or wave-cut platform in their wake. The platform is produced just below sea level where wave action is still vigorous. The result of this process is evident in the lower right of the photo. The flat terrace above the cliffs is an ancient wave-cut platform that was uplifted out of the sea by tectonic movements. Before the last uplift, the shoreline was near the base of the mountains in the background. Thus, wave erosion, followed by tectonic uplift and renewed wave erosion, created this landscape.

Why is this shoreline so different from those of the Atlantic coast of the United States, or the coasts of tropical islands such as Tahiti? In this chapter, we will consider these and other questions of coastal dynamics. Shorelines are especially important to our society because of the concentration of population on or near the coasts. In fact, over twenty percent of the world's population is within 100 km of the shoreline. To live in harmony with these rapidly changing environments, we must understand their histories and dynamics.

MAJOR CONCEPTS

1. Wind-generated waves provide most of the energy for shoreline processes.
2. Wave refraction concentrates energy on headlands and disperses it in bays.
3. Longshore drift, generated by waves advancing obliquely toward the shore, transports sediment parallel to the coast. It is one of the most important shoreline processes.
4. Erosion along a coast tends to develop sea cliffs by the undercutting action of waves and longshore currents. As a cliff recedes, a wave-cut platform develops, until equilibrium is established between wave energy and the shape of the coast.
5. Sediment transported by waves and longshore current is deposited in areas of low energy to form beaches, spits, and barrier islands.
6. Erosion and deposition along a coast tend to develop a straight or gently curving shoreline that is in equilibrium with the energy expended upon it.
7. Reefs grow in tropical climates and thrive only in shallow, clear marine waters. Fringing reefs around volcanic islands can evolve into atolls.
8. The worldwide rise in sea level, associated with the melting of the Pleistocene glaciers, drowned many coasts. Coasts are classified on the basis of either the process—subaerial or marine—that has been most significant in developing their configurations or their tectonic setting.
9. Tides are produced by the gravitational attraction of the Moon and locally exert a major influence on shorelines.
10. Tsunamis are waves generated by earthquakes, volcanic eruptions, and subaqueous landslides that disturb the seafloor.

WAVES

Shorelines are dynamic systems involving the energy of waves and currents. Wind-generated waves provide most of the energy for erosion, transportation, and deposition of sediment. Waves approaching a shore are bent, or refracted, so that energy is concentrated on headlands and dispersed in bays.

Most shoreline processes are directly or indirectly the result of wave action. An understanding of wave phenomena is therefore fundamental to the study of shoreline processes. All waves move some form of energy from one place to another. This is true of sound waves, radio waves, and water waves. The most important types of ocean waves are generated by wind.

What is the nature of the motion of water in wind-generated waves?

As wind moves over the open ocean, the turbulent air distorts the surface of the water. Gusts of wind depress the surface where they move downward; as they move upward, they cause a decrease in pressure, elevating the water's surface. These changes in atmospheric pressure produce an irregular, wavy surface in the ocean and transfer part of the wind's energy to the water. In a stormy area, waves are choppy and irregular, and wave systems of different sizes and orientations may be superposed on each other. As the waves move out from their place of origin, however, the shorter waves move more slowly and are left behind, and the wave patterns develop some measure of order.

Wave Motion in Water

Water waves are described in the same terms as those applied to other wave phenomena. These are illustrated in Figure 15.1. The **wavelength** is the horizontal distance between adjacent **wave crests.** The **wave height** is the vertical distance

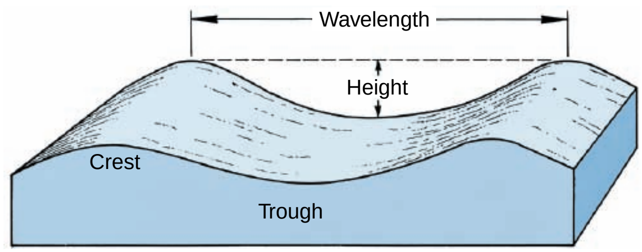


FIGURE 15.1 **The morphology of a wave** can be described in terms of its length (the distance from crest to crest), height (the vertical distance between crest and trough), and period (the time between the passage of two successive crests).

between wave crest and wave trough. The time between the passage of two successive crests is the **wave period.**

Wave motion can be observed easily by watching a floating object move upward (as the crest of a wave approaches) and then sink into the following trough. Viewed from the side, the object moves in a circular orbit with a diameter equal to the wave's height (Figure 15.2). Beneath the surface, this orbital motion dies out rapidly, becoming negligible at a depth equal to about one-half the wavelength. This level is known as the **wave base.** The motion of water in waves is therefore distinctly different from the motion in currents, in which water moves in a given direction and does not return to its original position.

A wave's energy depends on its length and height. The greater the wave's height, the greater the size of the orbit in which the water moves. The total energy of a wave can be represented by a column of water in orbital motion.

Breakers

Wave action produces little or no net forward motion of the water because the water moves in an orbital path as the wave advances. As a wave approaches shallow water, however, some important changes occur (Figure 15.3). First, the wavelength decreases because the wave base encounters the ocean bottom, and the resulting friction gradually slows the wave. Second, the wave height increases as the column of orbiting water encounters the seafloor. As the wave form becomes progressively higher and the velocity decreases, a critical point is reached at which the forward velocity of the orbit distorts the wave form. The wave crest then extends beyond the support range of the underlying column of water, and the wave collapses, or breaks. At this point, all of the water in the column moves forward, releasing its energy as a wall of moving, turbulent surf known as a **breaker.**

After a breaker collapses, the **swash** (a turbulent sheet of water) flows up the beach slope. The swash is a powerful surge that causes a landward movement of sand and gravel on the beach. After the force of the swash is dissipated against the slope of the beach, the water flows down the beach slope as **backwash,** although some seeps into the permeable sand and gravel.

In summary, waves are generated by the wind on the open ocean. The wave form moves out from the storm area, but the water itself moves in a circular orbit with little or no forward motion. As a wave approaches the shore, it breaks, and the energy of the forward-moving surf is expended on the shore, causing erosion, transportation, and deposition of sediment.

Wave Refraction

A key factor in shoreline processes is **wave refraction** because it influences the distribution of energy along the shore as well as the direction in which coastal water and sediment move. It occurs because the part of a wave in shallow water begins to drag the bottom and slows, whereas the segments of the same wave in deeper water move forward at normal velocity. As a result, the wave is bent, or

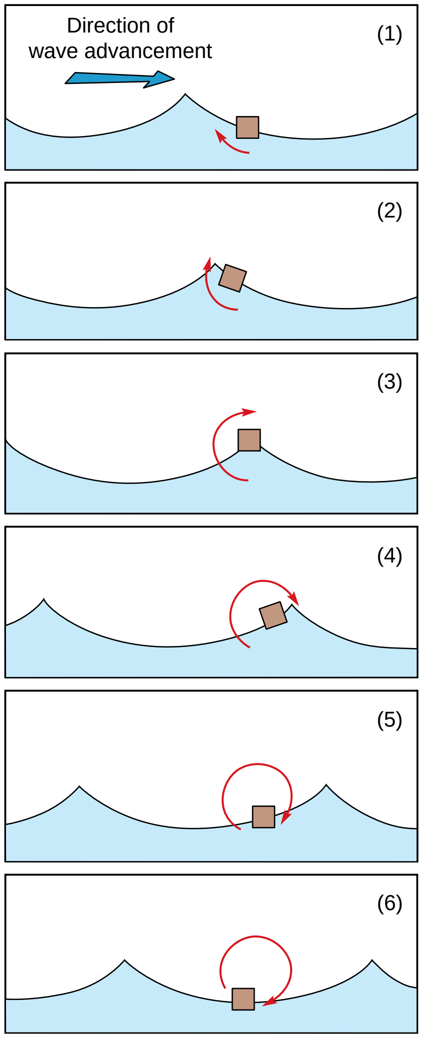


FIGURE 15.2 **The motion of a water particle as a wave advances** is shown by the movement of a floating object. As the wave advances (from left to right), the object is lifted up to the crest and then drops down to the trough (top). The wave form advances, but the water particles move in orbits, returning to their original position.

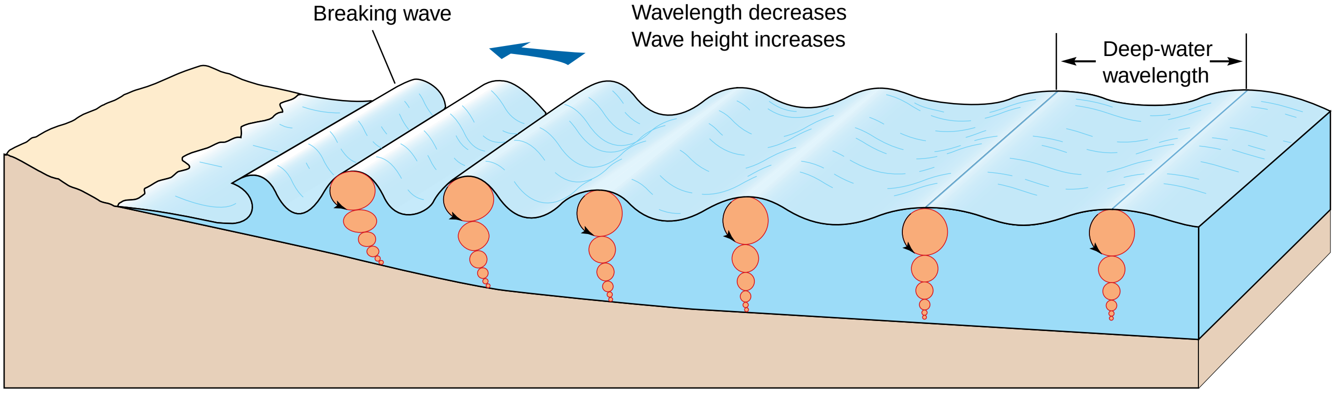


FIGURE 15.3 **A wave approaching the shore** undergoes several significant changes as the water in orbital motion encounters the seafloor. (1) The wavelength decreases because of frictional drag, and the waves become crowded together as they move closer to shore. Note also that the orbital motion of water in a wave decreases with depth and dies out at a depth equal to about half its wavelength. (2) The wave height increases as the column of water, moving in an orbit, stacks up on the shallow seafloor. (3) The wave becomes asymmetrical, because of increasing height and frictional drag on the seafloor, and ultimately breaks. The water then ceases to move in an orbit and rushes forward to the shore. The photograph shows the characteristic shape of a breaking wave. *(Courtesy of Leroy Grannis/Masterfile Corporation)*

refracted, so the crest line tends to become parallel to the shore. Wave refraction thus concentrates energy on headlands and disperses it in bays.

How does wave refraction influence erosion and deposition along coasts?

To appreciate the effect of wave refraction on the concentration and dispersion of energy, consider the energy in a single wave. In Figure 15.4, the unrefracted wave is divided into three equal parts (AB, BC, and CD), each having an equal amount of energy. As the wave moves toward the shore, segment BC, in front of the **headland,** first interacts with the shallow floor and is slowed down. Meanwhile, the rest of the wave (segments AB and CD) moves forward at normal velocity. This difference in velocity causes the crest line of the wave to bend as it advances shoreward. The wave energy between points B and C is concentrated on a relatively short segment (B′C′) of the headland, whereas the equal amounts of energy between A and B, and between C and D, are distributed over much greater distances (A′B′ and C′D′). Breaking waves are thus powerful erosional agents on the headlands but are relatively weak in bays, where they commonly deposit sediment to form beaches. Where major wave fronts are refracted around islands and headlands, the refraction patterns are obvious from the air (Figure 15.5).

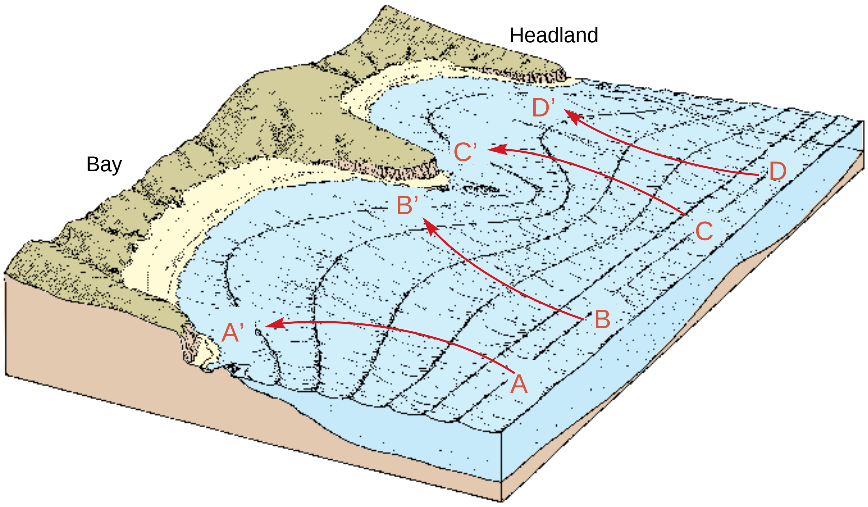


FIGURE 15.4 Wave refraction concentrates energy on headlands and disperses it across bays. Each segment of the unrefracted wave—AB, BC, and CD—is the same length and therefore has the same amount of energy as the other segments. As the wave approaches shore, segment BC encounters the seafloor sooner than AB or CD and moves more slowly. This difference in the velocities of the three segments causes the wave to bend, so that the energy contained in segment BC is concentrated on the headland (B′C′), while the energy contained in AB and CD is dispersed along the beach (A′B′ and C′D′).

LONGSHORE DRIFT

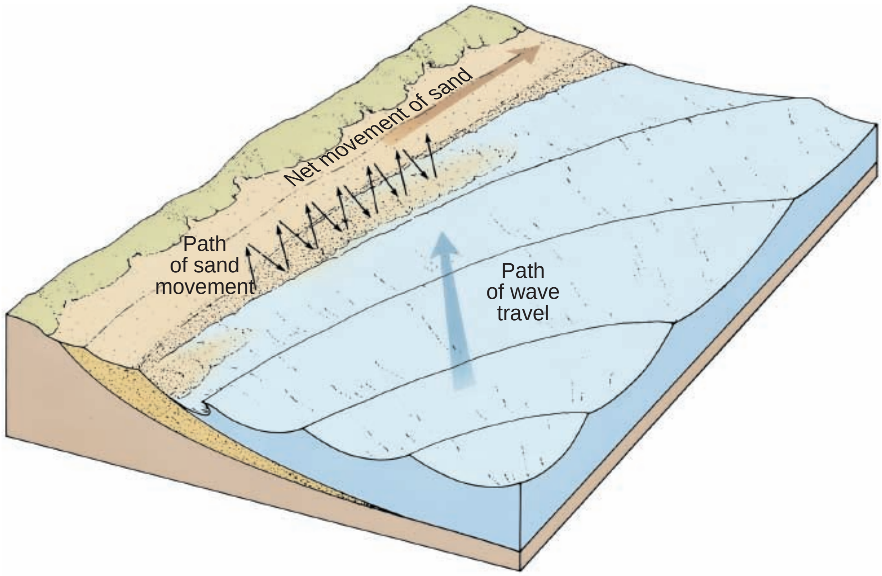
Longshore drift is generated as waves strike the shore at an angle. Water and sediment move obliquely up the beach face but return directly down the beach, perpendicular to the shoreline. This movement results in a net transport parallel to the shore. As a result, an enormous amount of sediment is constantly moving parallel to the shore.

Longshore drift is one of the most important shoreline processes. It is generated as waves advance obliquely to the shore (Figure 15.6). As a wave strikes the shore at an angle of less than 90°, water and sediment moved by the breaker are transported obliquely up the beach, in the direction of the wave's advance. When the wave's energy is spent, the water and sediment return with the backwash, directly down the beach, perpendicular to the shore. The next wave moves the material obliquely up the shore again, and the backwash returns it again directly down the beach slope. A single grain of sand is thus moved in an endless series of small steps, with a resulting net transport parallel to the shore. This process is known as **beach drift.** A similar process, known as a **longshore current,** develops in the breaker zone; thus, longshore movement occurs in two zones. One is along the upper limits of wave action and is related to the surge and backwash of the waves. The other is in the surf and breaker zone, where material is transported in suspension and by saltation. The two processes work together, and their combined action is known as **longshore drift.**

Longshore drift results in the movement of an enormous volume of sediment. A beach can be thought of as a river of sand, moving by the action of beach drift. If the wave direction is constant, longshore drift occurs in one direction only. If waves approach the shore at different angles during different seasons, longshore drift is periodically reversed. Longshore currents can pile significant volumes of water on the beach, which return seaward through the breaker zone as a narrow **rip current.** These currents can be strong enough to be dangerous to swimmers.

FIGURE 15.5 Wave refraction around headlands and islands is clearly shown in aerial photographs taken along the coast of Oregon. The energy concentrated on the headlands has reduced some of them to offshore islands. *(Courtesy of U.S. Geological Survey)*

(A) As a wave breaks on the shore, sediment lifted by the surf is moved diagonally up the beach slope. The backwash then carries the particles back down the beach at a right angle to the shoreline. This action is repeated by each successive wave and transports the sediment along the coast in a zigzag pattern. Particles also are moved underwater in the breaker surf zone by this action.



(B) Longshore drift can be seen in patterns of sediment in this aerial photograph. The sediment is moved parallel to the shore in a series of waves.

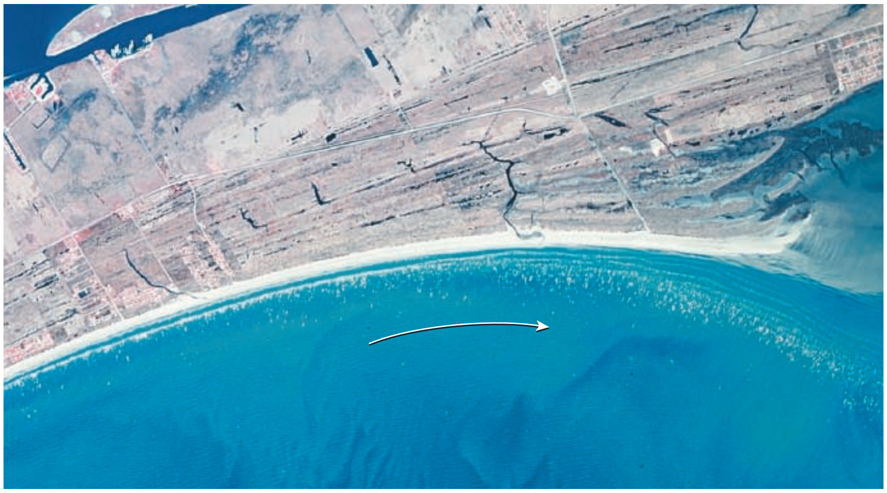


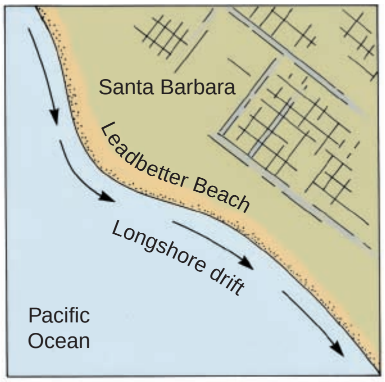
FIGURE 15.6 **Longshore drift** occurs where waves strike the beach at an oblique angle.

Longshore Drift at Santa Barbara, California

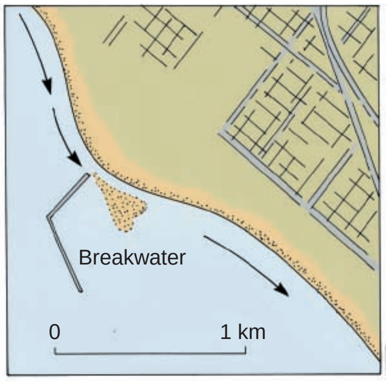
Longshore Drift

One of the best ways to appreciate the process of longshore drift is to consider how it has influenced human affairs. A good example occurred near Santa Barbara, along the southern coast of California, where data have been collected over a considerable period. Santa Barbara is a picturesque coastal town at the base of the Santa Ynez Mountains. It is an important educational, agricultural, and recreational area, and the people there wanted a harbor that could accommodate deep-water vessels. Studies by the U.S. Army Corps of Engineers suggested that the site was unfavorable because of the strong longshore currents, which carry large volumes of sand to the south (Figure 15.7A). Rivers draining the mountains of the coastal ranges supply new sediment to the coast at a rate of 600 m^3/day. Longshore drift continually moves the sand southward from beach to beach. The currents are so strong that boulders 0.6 m in diameter can be transported. Ultimately, the sand transported by longshore drift is delivered to the head of a submarine canyon and then moves down the canyon to the deep-sea floor. Despite reports advising against the project, a breakwater 460 m long was built and a deep-water harbor was constructed in 1925 at a cost of $750,000. This breakwater was not tied to the shore. Sand, moved by longshore drift, began to pour through the gap and fill the harbor, which was protected from wave refraction and longshore currents by the breakwater (Figure 15.7B). To stop the filling of the harbor, the town had to connect the breakwater to the shore. Sand then

Why is a beach commonly called "a river of sand"?



(A) The Santa Barbara coast had significant longshore drift before the breakwater was built.



(B) The initial breakwater prevented the generation of longshore currents in the protected area behind it, and therefore the harbor filled with sand.

(C) After the breakwater was connected to the shore, longshore currents moved sand around the breakwater and filled the mouth of the harbor. Sand is now dredged from the harbor and pumped down the coast.

(D) Photo of Santa Barbara Harbor *(Courtesy of National Oceanic and Atmospheric Administration)*

FIGURE 15.7 The effect of a breakwater on longshore drift in Santa Barbara, California, is documented by a series of maps of the coast from 1925 to 1938.

accumulated behind the breakwater, at its southern end. Soon a smooth, curving beach developed around the breakwater, and longshore drift carried sand around the breakwater and deposited it inside the harbor (Figure 15.7C). Two disastrous effects were produced: First, the harbor became so choked with sand that it could accommodate only vessels with very shallow draft; second, the beaches downcoast were deprived of their source of sand and began to erode. Within 12 years, more than $2 million worth of damage had been done to property down the coast from Santa Barbara, as the beach in some areas was cut back 75 m. The problem was solved by the installation of a dredge in the Santa Barbara harbor to pump out the sand and return it to the longshore drift system on the downcurrent side of the harbor. Most of the beaches have been partly replenished, but dredging is very expensive.

EROSION ALONG COASTS

Erosion along coasts results from the abrasive action of sand and gravel, moved by the waves and currents and, to a lesser extent, from solution and hydraulic action. The undercutting action of waves and currents typically produces sea cliffs. As a sea cliff recedes, a wave-cut platform develops. Minor erosional forms associated with the development of sea cliffs include sea caves, sea arches, and sea stacks.

Coastal regions are sculpted in many shapes and forms, such as rocky cliffs, low beaches, quiet bays, tidal flats, and marshes. The topography of a coast results from the same basic forces that shape other land surfaces: erosion, deposition, tectonic uplift, and subsidence.

Wave action is the major agent of erosion along coasts, and its power is awesome during storms. When a wave breaks against a sea cliff, the sheer impact of the water can exert a pressure exceeding 100 kg/m^2. Water is driven into every crack and crevice of the rocks, compressing the air within. The compressed air then acts as a wedge, widening the cracks and loosening the blocks.

Solution activity also takes place along the coast and is especially effective in eroding limestone. Even noncalcareous rocks can be weathered rapidly by solution activity because the chemical action of seawater is stronger than that of fresh water.

Sea Cliffs and Wave-Cut Platforms

The most effective process of erosion along coasts, however, is the abrasive action of sand and gravel moved by the waves. These tools of erosion operate like the bed load of a river. Instead of cutting a vertical channel, however, the sand and gravel moved by waves cut horizontally, forming wave-cut cliffs and wave-cut platforms.

Why are most shores undergoing vigorous erosion?

To understand the nature of wave erosion and the principal features of its forms, study the diagram and photographs in Figure 15.8. Where steeply sloping land descends beneath the water, waves act like a horizontal saw, cutting a notch into the bedrock at sea level. This undercutting produces an overhanging **sea cliff,** or **wave-cut cliff,** which ultimately collapses. The fallen debris is broken up and removed by wave action, and the process is repeated on the fresh surface of the new cliff face. As the sea cliff retreats, a **wave-cut platform** is produced at its base, the upper part of which commonly is visible near shore at low tide. Sediment derived from the erosion of the cliff, and transported by longshore drift, may be deposited in deeper water to form a **wave-built terrace.** Stream valleys that formerly reached the coast at sea level are shortened and left as **hanging valleys** when the cliff recedes.

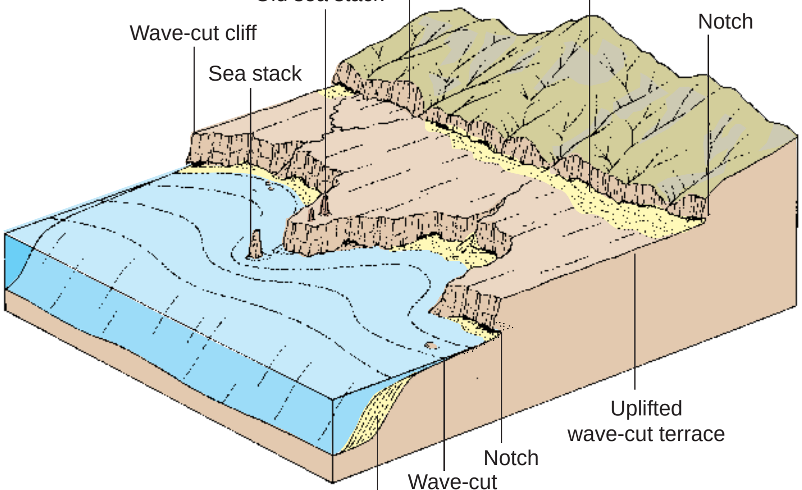
As the platform is enlarged, the waves break progressively farther from shore, losing much of their energy by friction as they travel across the shallow platform. Wave action on the cliff is consequently greatly reduced. Beaches can then develop at the base of the cliff, and the cliff face is gradually worn down, mainly by weathering and mass movement. Because wave-cut platforms effectively dissipate wave energy, the size to which they can grow is limited. However, some volcanic islands have been truncated completely by wave action and slope retreat so that only a flat-topped platform is left near low tide.

Sea Caves, Sea Arches, and Sea Stacks

The rate at which a sea cliff erodes depends on the durability of the rock and the degree to which the coast is exposed to direct wave attack. Zones of weakness (such as outcrops with joint systems, fault planes, and beds of shale between harder sandstones) are loci of accelerated erosion. If a joint extends across a headland, wave action can hollow out an alcove, which may later enlarge to a **sea cave.** Because the headland commonly is subjected to erosion from two sides, caves excavated along a zone of weakness can join to form a **sea arch** (Figure 15.9). Eventually, the arch collapses, and an isolated pinnacle known as a **sea stack** is left in front of the cliff. An excellent example of the development of sea arches and stacks is shown in Figure 15.10 in which the collapse of a sea arch is documented by photographs.

Summary of Coastal Erosion

Coastal erosion is a natural process that has altered the world's shorelines ever since the oceans were first formed at least 4 billion years ago. Every day, the surging action of waves, the movement of longshore currents, and the pounding of



(A) An uplifted wave-cut platform and a new sea cliff and platform in the process of forming. As erosion continues, the cliff recedes to form a wave-cut platform. Some sediment eroded from the shore is deposited in deeper water as a wave-built terrace.

(B) A wave-cut platform on the Washington coast. *(Photograph by D. Easterbrook)*

(C) Wave action operates like a horizontal saw cutting at the base of the cliff like this one in Oman.

FIGURE 15.8 **A wave-cut platform** is the fundamental landform produced by wave erosion.

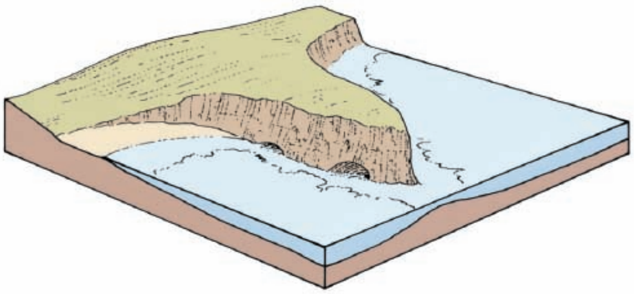
storms erode shorelines. In addition, sea level is constantly changing; with each rise or fall, a new coastline is formed and the process of reshaping the shore begins anew. Because of the recent rise in sea level, due to the melting of the glaciers, many shorelines of the world are several hundred meters higher than they were 30,000 years ago, Vigorous erosion by waves and currents will continue.

The rate at which wave action cuts away at the shore is extremely variable. It depends on the configuration of the coast, the size and strength of the waves, and the physical characteristics of the bedrock. In poorly consolidated material, such as glacial moraines, stream deposits, or sand dunes, the rate of cliff retreat may be as much as 30 m/yr, but rates of erosion along most coasts are much slower.

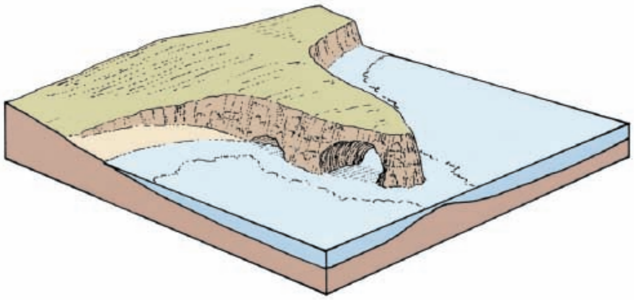
Can coastal erosion be stopped?

An interesting example of rates of coastal erosion over a longer time is documented by maps made by the ancient Romans when they conquered Britain. These maps show that in approximately 2000 years, parts of the British coast have been eroded back more than 5 km, and the sites of many villages and landmarks have

(A) Wave energy is concentrated on a headland as a result of wave refraction. Zones of weakness, such as joints, faults, and nonresistant beds, erode fastest, so sea caves develop in those areas.



(B) Sea caves enlarge to form a sea arch.



(C) Eventually, the arch collapses, leaving a sea stack. A new arch can develop from the remaining headland.

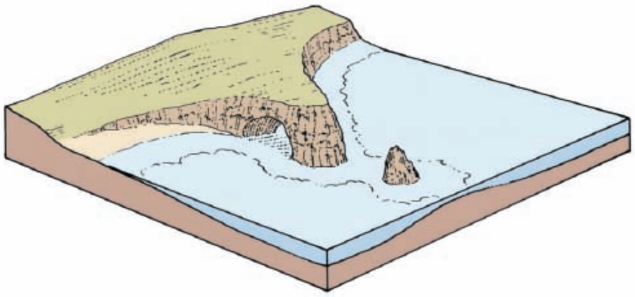


FIGURE 15.9 **The development of sea caves, sea arches, and sea stacks** is associated with differential erosion of a headland.

been swept away. Other examples of rapid wave erosion are found on new volcanic islands, such as Surtsey, near Iceland. The newly formed volcanic ash that makes up such islands can be completely planed off by wave action in a matter of only a few decades.

The reality of coastal erosion is made painfully clear by the passion of Americans to live and vacation on the seashore. Development projects unwittingly put more and more people and property on the shore, an area that by its very nature is dynamic and mobile. In a remarkably short period of time, waves can erode high cliffs like those that surround most of the island of Hawaii (Figure 15.11). About

(A) Sea arches and sea stack as they appeared in 1969.

(B) The same area in 1987, after collapse of arch.

FIGURE 15.10 **Collapse of a sea arch** along the coast of California. Between 1969 and 1987, wave erosion also eliminated a small sea stack.

86% of California's coast is receding at an average rate of 0.15 to 0.75 m/yr (Figure 15.12). Parts of Monterey Bay lose as much as 2 to 3 m/yr. Cape Shoalwater, Washington, about 100 km west of Olympia, has been eroding at a rate of more than 30 m/yr. Parts of Chambers County, Texas, have lost 3 m of coast in nine months. In parts of North Carolina, erosion in one year has cut into beachfront property up to 25 m.

To combat these losses, people build sea walls and breakwaters, but these are local and temporary solutions at best. A sea wall, or jetty (a long concrete or rock structure that juts out into water to restrain waves and currents), may protect threatened property near it, but it often hastens erosion in other areas. There may be no simple answer. Carefully written zoning laws that limit coastal development may be the best option. In our battle with nature, retreat might be the ultimate solution.

DEPOSITION ALONG COASTS

Sediment transported along the shore is deposited in areas of low wave energy and produces a variety of landforms, including beaches, spits, tombolos, and barrier islands.

A shoreline is a system that involves input of sediment from various sources, transportation of the sediment, and ultimate deposition. Much of the sediment is derived from the land and delivered to the sea by major rivers. The sediment is then transported by waves and longshore currents and is deposited in areas of low energy, where it builds a variety of landforms. Changes continue by both erosion and deposition until the coastline is smooth and straight, or gently curving.

Figure 15.13 shows some important elements in a coastal system. The primary sources of sediment, for beaches and assorted depositional features, are the rivers that drain the continents. Sediment from the rivers is transported along the shore by longshore drift and is deposited in areas of low energy. Erosion of headlands and sea cliffs is also a source of sediment. In tropical areas, the greatest source of sand commonly is shell debris, derived from wave erosion of near-shore coral reefs. Sediment can also leave the system by landward migration of coastal sand dunes and by transportation into deep areas of the ocean floor, where turbidites accumulate as submarine fans.

FIGURE 15.11 **High sea cliffs** of Hawaii indicate that the shoreline has receded tens of kilometers by wave action and slumping. Older islands in the chain have been completely planed off in a matter of only a few million years. *(Courtesy of Bob Abraham/Pacific Stock)*

FIGURE 15.12 **Erosion of sea cliffs** is a major process along shorelines. This photo shows erosion of a sea cliff along the coast of California. In the last two decades, dozens of homes and hundreds of other structures have been lost or damaged because of sea cliff recession, particularly during storms triggered by El Niño oscillations. *(Courtesy of CORBIS)*

Beaches

What is the source of sand on beaches?

A **beach** is a shore built of unconsolidated sediment. Sand is the most common material, but some beaches are composed of cobbles and boulders and others of fine silt and clay. The physical characteristics of a beach (such as slope, composition, and shape) depend largely on wave energy, but the supply and size of available sediment particles are also important. Beaches composed of fine-grained material generally are flatter than those composed of coarse sand and gravel.

Spits

In areas where a straight shoreline is indented by **bays** or **estuaries,** longshore drift can extend the beach from the mainland to form a **spit.** A spit can grow far out across the bay as material is deposited at its end (Figure 15.14). Eventually, it may extend completely across the front of the bay, forming a **baymouth bar.**

Tombolos

Beach deposits can also grow outward and connect the shore with an offshore island to form a **tombolo.** This feature commonly is produced by the island's effect on wave refraction and longshore drift (Figure 15.15). An island near a shore can cause wave refraction to such an extent that little or no wave energy strikes the shore behind it. Longshore drift, which moves sediment along the coast, is not generated in this wave shadow zone. Sediment carried by longshore currents is

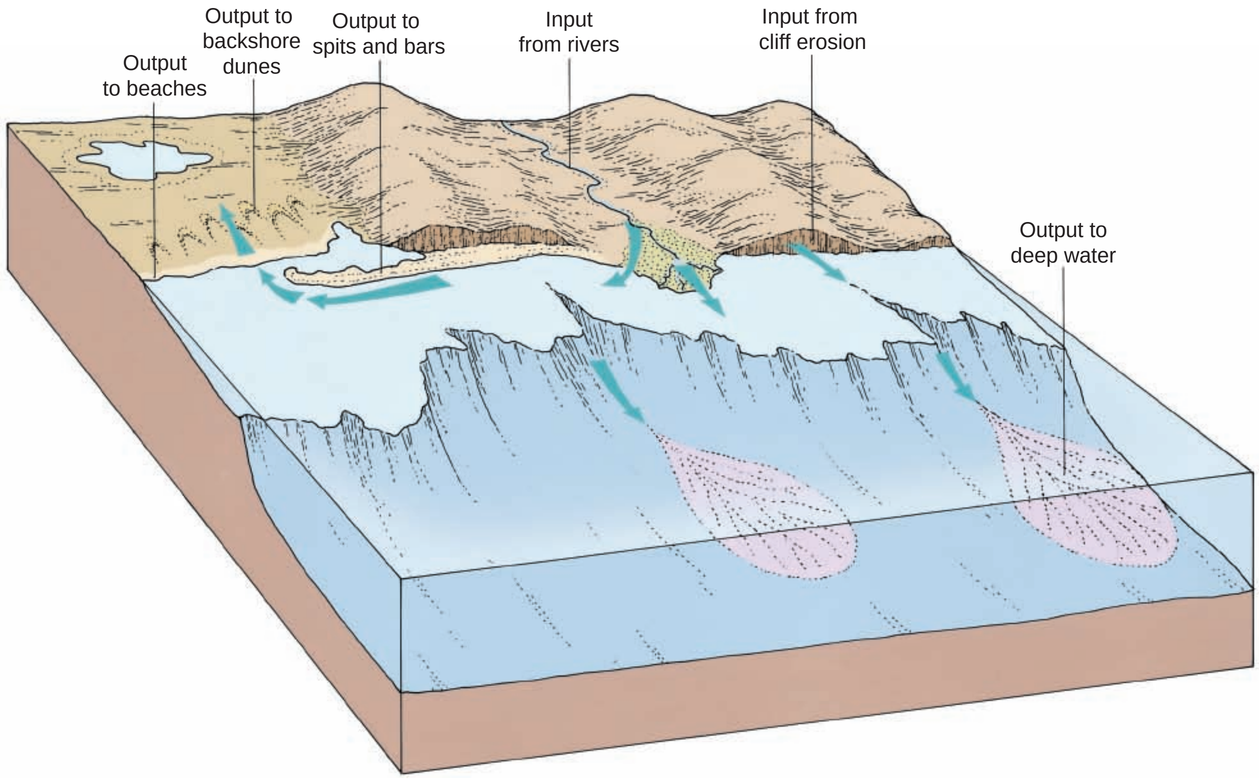


FIGURE 15.13 **A shoreline is a dynamic system** of moving sediment. Most of the sediment in a shoreline system is supplied by rivers bringing erosional debris from the continent and by the erosion of sea cliffs by wave action. This material is transported by longshore drift and can be deposited on growing beaches, spits, and bars. Some sediment, however, leaves the system either by transportation to deeper *water by turbidites or by the* landward migration of coastal sand dunes.

therefore deposited behind the island. The sediment deposit builds up and up and eventually forms a tombolo, a bar or beach connecting the shore to the island. Longshore currents then move uninterrupted along the shore and around the tombolo.

Barrier Islands

Barrier islands are long, offshore islands of sediment, trending parallel to the shore (Figure 15.16). Almost invariably, they form long shorelines next to gently sloping coastal plains, and they typically are separated from the mainland by a lagoon. Most barrier islands are cut by one or more tidal inlets. Many barrier islands develop from the growth of spits across an irregular shoreline (Figure 15.14).

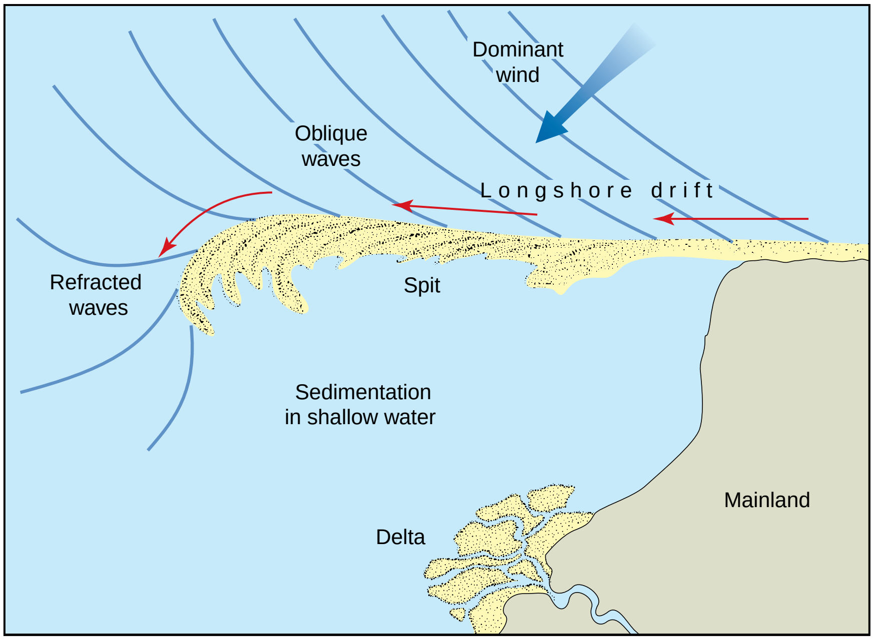
Transportation and deposition along many coasts can be measured using historical monuments, maps, and sequences of aerial photography. In northern France, a dike built at the shoreline in 1597 is now more than 3 km inland from the present shore, indicating an average rate of spit migration of about 1 km/100 yr.

Other rates of spit migration, based on dated maps, include the western end of Fire Island (off the southern coast of Long Island, New York) and the Rockaway spit (western Long Island). The rate of lateral migration for both Long Island spits is 65 m/yr, or 6.5 times the rate of migration along the French coast.

How does a spit evolve into a barrier island?

One of the best-documented examples of coastal modification is the barrier beach at Chatham, Massachusetts (Figure 15.17). Before 1987, the barrier beach known as Nauset, or North Beach, curved southward in a long, graceful arc and terminated as a spit south of the town of Chatham. On January 2, 1987, a storm and high tide cut a narrow slice (less than a half-meter deep and 5 m wide) through the barrier. By 1992, the break was more than 3 km wide and 8 m deep, and a new beach was connected to the mainland (Figure 15.17). The break in the protective barrier permitted extensive erosion on the mainland coast.

(A) Waves strike a shore obliquely and cause longshore sediment transport. The sediment moves in a series of sand waves around the end of a beach and into a bay.



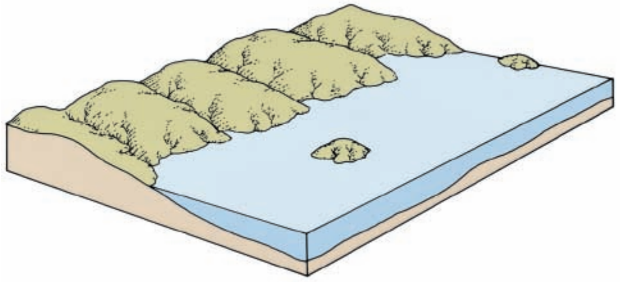
(B) A spit or barrier island can form by migration of a spit. Sediment moving along the shore is deposited as a spit in the deeper water near a bay. The spit grows parallel to the shore by longshore drift. *(Courtesy of National Oceanic and Atmospheric Administration)*

FIGURE 15.14 **Curved spits** develop as longshore drift moves sediment around the point of a barrier beach.

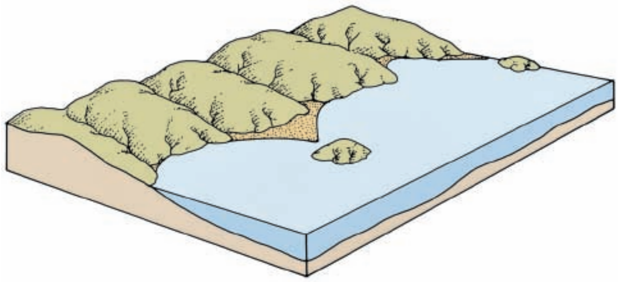
EVOLUTION OF SHORELINES

Processes of shoreline erosion and deposition tend to develop long, straight, or gently curving coastlines. Headlands are eroded, and bays and estuaries are filled with sediment. The configuration of the shoreline evolves until wave energy is distributed equally along the coast, and neither large-scale erosion nor deposition occurs.

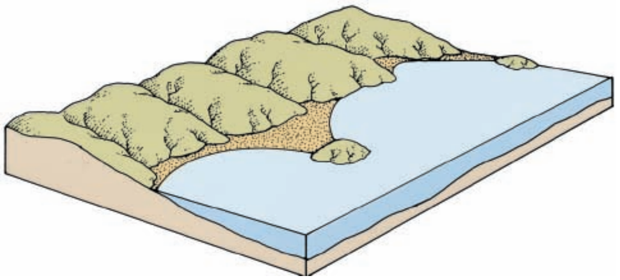
All of the coastlines throughout the world are constantly changing. In many areas, changes are rapid, and within only a few decades the local configuration of a shoreline can be significantly modified. Over a longer period, regional variations in coast configuration occur. This constant and rapid change in our coasts is due, in part, to the rise in sea level that accompanies the melting of glaciers. Other changes in the coast result from uplift or subsidence of the land or expansion or contraction of the sea. Thus, the shape of most coastlines is far from being at equilibrium with the wave energy expended upon them. The general trend is for headlands to be eroded and bays and estuaries to become filled with sediment. The change in the configuration of the shoreline is always in the direction so that energy is equally distributed along the shore, and neither large-scale erosion nor deposition occurs.



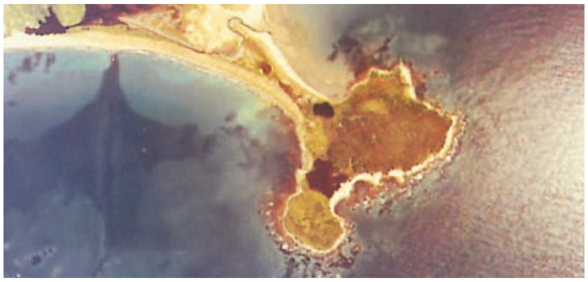
(A) An offshore island acts as a breakwater to incoming waves and creates a wave shadow along the coast behind it.



(B) Sediment moved by longshore drift is trapped in the shadow zone.



(C) The zone of sediment deposition eventually grows until it connects with the island. Longshore drift will then move sediment along the shore and around the tombolo.



(D) An aerial photograph of a tombolo. *(Courtesy of U.S. Geological Survey)*

FIGURE 15.15 **A tombolo** is a bar or beach that connects an island to the mainland. It forms because the island creates a wave shadow zone along the coast, in which longshore drift cannot occur.

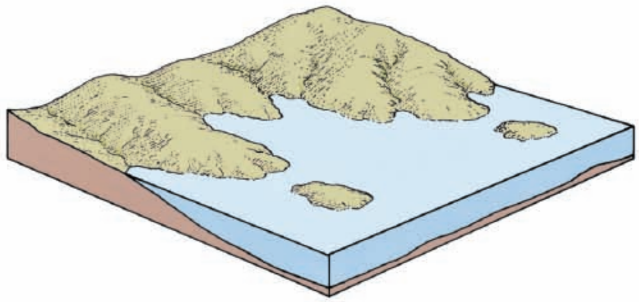
FIGURE 15.16 **A barrier island** along the Atlantic coast of the United States has a smooth seaward face, where wave action and longshore drift actively transport sediment. A tidal inlet may form a break in the island, and sediment transported through it is deposited as a tidal delta in the lagoon. *(Courtesy of NASA/Goddard Space Flight Center)*

(A) July 1984. Before the break, the barrier-spit acted as a shield protecting Chatham's shore and harbor.

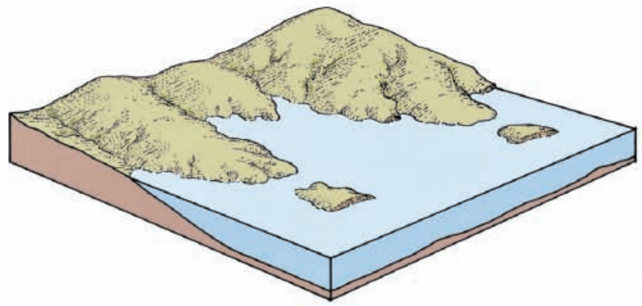
(B) March 1987. Two months after the break, the inlet enlarged and exposed the mainland to erosion.

(C) July 1992. Five years after the break, the inlet was 3 km wide, and a new spit connected the barrier to the mainland. Vigorous erosion continues along the coast of the mainland.

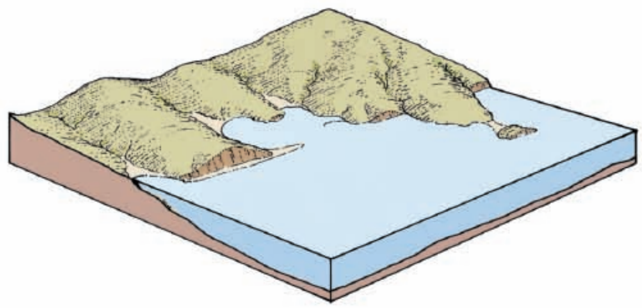
FIGURE 15.17 Changes of the shoreline near Chatham, Massachusetts. *(Courtesy of Kelsey-Kennard Photographers, Inc.)*



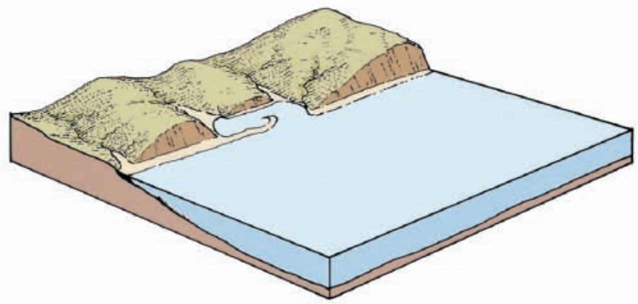
(A) A rise in sea level floods a landscape eroded by a river system and forms bays, headlands, and islands.



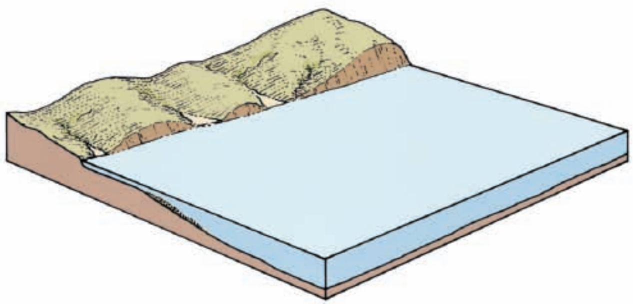
(B) Wave erosion cuts cliffs and, locally, sea stacks and arches on the islands and peninsulas.



(C) Wave-cut cliffs recede and grow higher, and headlands erode to form new sea cliffs. Sediment accumulates, forming beaches and spits.



(D) Islands are completely eroded, beaches and spits enlarge, and lagoons form in the bays.



(E) A straight shoreline is produced by the additional retreat of the cliffs and by sedimentation in bays and lagoons. The large wave-cut platform then limits further erosion by wave action.

FIGURE 15.18 The evolution of a shoreline of equilibrium from an embayed coastline involves changes due to both erosion and deposition. Eventually, a smooth coastline is produced, and the forces acting on it are essentially at equilibrium; thus, neither erosion nor deposition occurs on a large scale.

In such a condition, the energy of the waves and longshore drift is just sufficient to transport the sediment that is supplied. A shoreline with such a balance of forces is called a **shoreline of equilibrium.** As is the case with a stream profile of equilibrium, a delicate balance is maintained between the landforms and the geologic processes operating on them.

What is the configuration of a shoreline of equilibrium?

We can construct a simple conceptual model of a shoreline's evolution toward equilibrium and show the changes that would be expected to occur as erosion and deposition operate (Figure 15.18). Figure 15.18A shows an area originally shaped by stream erosion and subsequently partly drowned by rising sea level. River valleys are invaded by the sea to form irregular, branching bays, and some hilltops form peninsulas and islands. Next, as shown in Figure 15.18B, marine erosion begins to attack the shore. The islands and headlands are eroded into high wave-cut cliffs. As erosion proceeds (Figure 15.18C), the islands and headlands are worn back, and the sea cliffs increase in height. Minor features, such as sea caves, sea arches, and sea stacks, form by differential erosion in weak places in the bedrock. These are continually being formed and destroyed as the sea cliff recedes. A wave-cut platform develops, reducing wave energy, so a beach forms at the base of the cliff. In a more advanced stage of development (Figure 15.18D), the islands are eroded away and bays become sealed off, partly by the growth of spits, forming lagoons. The shoreline then becomes straight and simple. In the final stages of marine development (Figure 15.18E), the shoreline is cut back beyond the limits of the bay. Sediment moves along the coast by longshore drift, but the wave-cut platform is so wide that it effectively eliminates further erosion of the cliff by wave action. The shoreline of equilibrium is straight and essentially in equilibrium with the energy acting on it. Further modification of the cliffs results from weathering, mass movement, and stream erosion.

Naturally, the development of a shoreline is also affected by special conditions of structure and topography and by fluctuations of sea level or tectonics. The process of erosion of the headlands by wave action and the straightening of the shoreline by both erosion and deposition follow the general sequence of this idealized model; however, actual shorelines rarely proceed through all these stages because fluctuations of sea level upset the previously established balance.

The development of a shoreline is interrupted in many areas by tectonic uplift, which abruptly elevates sea cliffs and wave-cut platforms above the level of the waves. When this happens, wave erosion begins at a new, lower level, and the **elevated marine terraces,** stranded high above sea level, are attacked and eventually obliterated by weathering and stream erosion (Figure 15.19).

Storm Surges

Coastal changes are particularly great during intense storms such as hurricanes, typhoons, or "northeasters," as they are called in New England. These storm surges expend tremendous amounts of energy along coastal regions and produce considerable damage and rapid changes in coastal morphology. Intense storms, such as hurricanes and typhoons, are centered around strong low-pressure systems in the atmosphere that cause the sea surface to rise in a broad dome, while they depress the surface farther away. Such a buildup of water beneath the storm produces extensive flooding when it reaches shallow coastal areas. In addition, during the storm, the drag of the wind on the sea surface not only produces high waves but also creates currents that push the water in the direction of the wind. When the water reaches the coastal area, it impinges along the shore, resulting in abnormally high waves and tides. As a result, the rapid erosion, transportation, and deposition of sediment cause extensive damage to life and property.

The effect of storm surges on local coastal morphology and sediment transport is profound and extremely rapid (Figure 15.20). Large bodies of sediment may be washed over a barrier island to form a washover fan in the adjacent bay. New surge channels (tidal inlets) may be opened, while previously formed inlets may be sealed. Beach and dune sand may be moved inland, rearranging the area's surface features. Storm surges may be as high as 5 to 7 m.

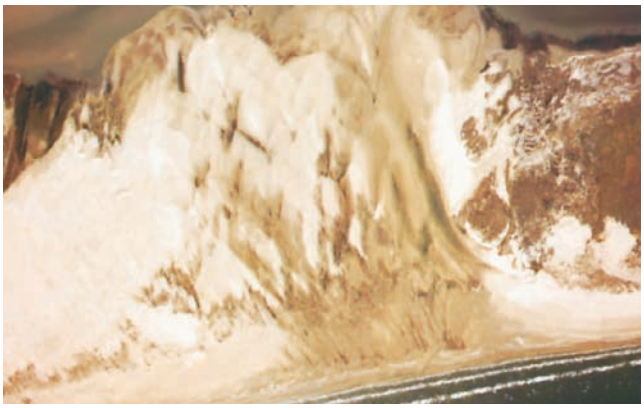
One way to appreciate the energy of storm surges, and their force as a geologic agent, is to consider the devastation they inflict on human life. On September 8, 1900, a storm surge hit the barrier island city of Galveston, Texas, killing 6000 people. Property damage from Hurricane Gilbert in 1988 exceeded $10 billion. One year later Hurricane Hugo hit South Carolina, destroying considerable amounts of property, but early warnings and mass evacuation averted great loss of life. In contrast, storm surges in the Bay of Bengal in 1876 took 100,000 lives, and another in 1970 killed an estimated 300,000 people.

FIGURE 15.19 **A series of elevated beach terraces** resulted from tectonic uplift along the southern coast of California and the offshore islands. This photograph of San Clemente Island was taken with the sun at a low angle, to emphasize the sequence of terraces. (*Photograph by J. Shelton)*

REEFS

Reefs form a unique type of coastal feature because they are biological in origin. Modern reefs are built by a complex community of corals, algae, sponges, and other marine invertebrates. Most reefs grow and thrive only in the warm, shallow waters of semitropical and tropical regions.

In many regions of the ocean, **coral** reefs grow and flourish to such an extent that they significantly modify, if not control, the configuration of a coastline. Reefs are especially important in the warm tropical waters of the South Pacific, where they are a major influence along the coasts of most islands (Figure 15.21). Reefs are constructed from invertebrate colonial animals that, instead of building separate isolated shells, build enormous "apartment houses" in which thousands of individuals live. When the animals die, the shell structure remains intact and subsequent generations build their apartments upon the abandoned homes of their predecessors; soon a reef develops into a "rocky" coast. The most impressive modern reef is Australia's Great Barrier Reef, which stretches a distance of 2000 km. Others are noted throughout the South Pacific as barrier reefs and atolls surrounding ancient volcanic islands. Indeed, coral reefs cover an area about the size of Italy.



(A) Washover fans on a barrier island formed in a hurricane. *(Courtesy of National Oceanic and Atmospheric Administration)*

(B) Beach erosion caused by Hurricane Hugo in South Carolina made this building collapse. *(Courtesy of David M. Bush/Orrin H. Pilkey, Jr.)*

FIGURE 15.20 **Storm surges** are produced by storms with strong winds like hurricanes. They can be extremely damaging to coastal buildings and cause rapid erosion.

Reef Ecology

The marine life that forms a reef can flourish only under strict conditions of temperature, salinity, and water depth. Most modern coral reefs occur in warm tropical waters between the limits of 30° S latitude and 30° N latitude (Figure 15.22). Colonial corals need sunlight, and they cannot live in water deeper than about 75 m. They do not grow up from abyssal depths on the sea floor. Instead, they grow most luxuriantly just a few meters below sea level. Dirty water inhibits rapid, healthy growth because it cuts off sunlight, and the suspended mud chokes the organisms that filter feed. Corals are therefore absent or stunted near the mouths of large muddy rivers. They can survive only if the salinity of the water ranges from 27 to 40 parts per thousand; thus, a reef can be killed if a flood of fresh water from the land reduces the salinity. Coral reefs are remarkably flat on top, the upper surface is usually are exposed at low tide but must be covered at high tide. Reefs can grow upward with rising sea level if the rate of rise is not excessive. They can also grow seaward over the flanks of reef debris. The fact that reefs form in such restricted environments makes them especially important as indicators of past climatic, geographic, and tectonic conditions.

How can an entire shoreline be formed by growth of organisms?

Types of Reefs

Fringing reefs, generally ranging from 0.5 to 1 km wide, are attached to such land masses as the shores of volcanic islands (Figure 15.23) or continents. The corals grow seaward, toward their food supply. Because coral and other reef-building lifeforms, need sunlight to grow, reefs are usually absent near deltas and mouths of rivers, where the waters are muddy. Heavy sedimentation and high runoff also make some tropical coasts of continents unattractive to fringing reefs.

Barrier reefs are separated from the mainland by a lagoon, which can be more than 20 km wide. As seen from the air, the barrier reefs of islands in the South Pacific are marked by a zone of white breakers. At intervals, narrow gaps occur, through which excess shore and tidal water can exit. The finest example of this type is the Great Barrier Reef, which stretches for 2000 km along the northern shore of Australia, from 30 to 160 km off the Queensland coast.

Platform reefs grow in isolated oval patches in warm, shallow water on the continental shelf. They were apparently more abundant during past geologic periods of warmer climates. Most modern platform reefs seem to be randomly distributed, although some appear to be oriented in belts. The latter feature suggests that they were formed on submarine topographic highs, such as drowned shorelines.

Atolls are roughly circular reefs that rise from deep water, enclosing a shallow lagoon in which there is no exposed central land mass. The outer margin of an atoll is naturally the site of most vigorous coral growth. It commonly forms an overhanging rim, from which pieces of coral rock break off, accumulating as submarine talus on the slopes below. A cross-sectional view of a typical atoll shows that the lagoon floor is shallow and is composed of calcareous sand and silt with rubble derived from erosion of the outer side (see the foreground of Figure 15.23).

Atolls are by far the most common type of coral reef. More than 330 are known, of which all but 10 lie within the Indo-Pacific tropical area. Drilling into the coral of atolls tends to confirm the theory that atolls form on submerged volcanic islands. In one instance, coral extends down as much as 1400 m below sea level, where it rests on a basalt platform carved on an ancient volcanic island. Because coral cannot grow at that depth, it presumably grew upward as the volcanic island sank. A reef this thick probably accumulated over 40 or 50 million years.

(A) This aerial view of Bora Bora shows an island in the intermediate stage in the evolution of an atoll. Note the outer margin of the reef, where the growth of organisms is most active. The shallow lagoon inside the reef, shown in light blue, is mostly calcareous sand formed by erosion of the reef. The remnant volcano in the center is highly dissected by stream erosion, indicating the elapse of a long period of time since the volcano was active. *(Courtesy of Jan Arthus-Bertrand and CORBIS)*

(C) An underwater view of the reef, showing the community of organisms involved in reef construction. *(Courtesy of Tahiti Tourist Board)*

(B) Reef mounds are visible through the shallow water in the lagoon. Note the boat for scale. *(Courtesy of Tahiti Tourist Board)*

FIGURE 15.21 **A barrier reef surrounds a volcanic island** in the Society Islands, French Polynesia.

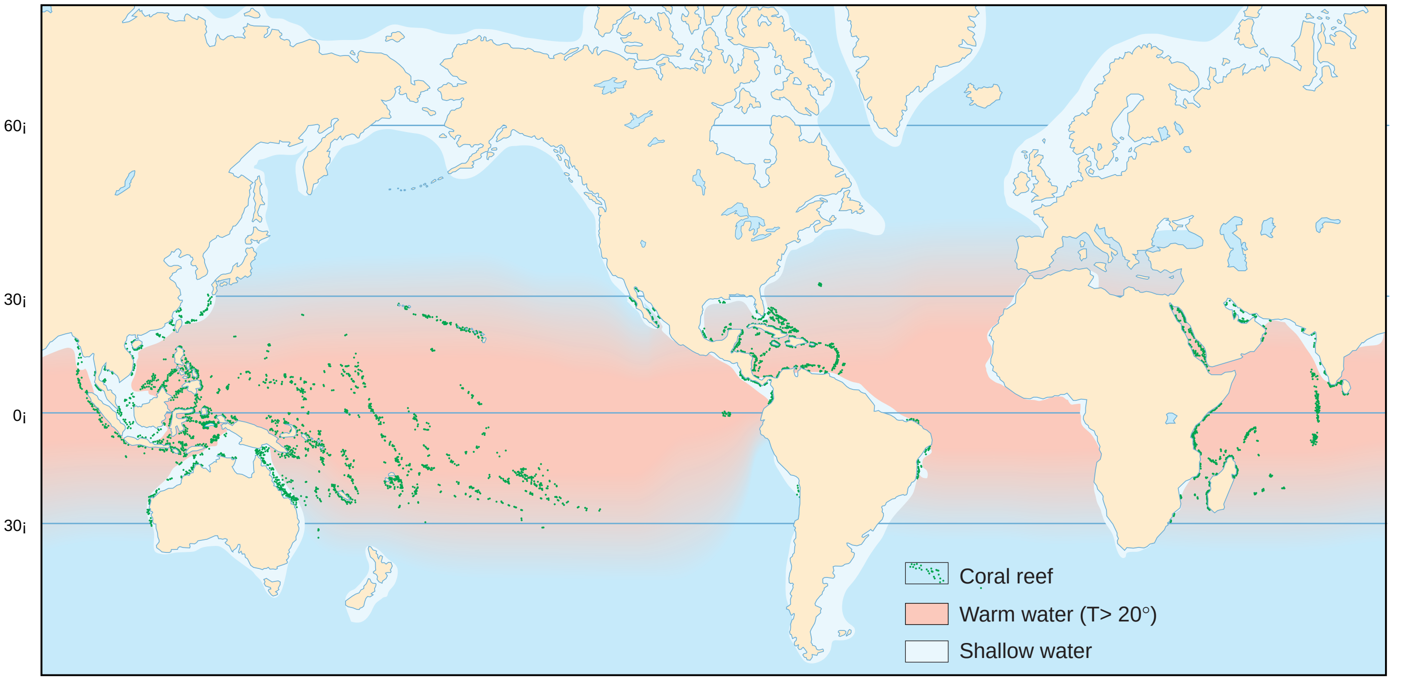


FIGURE 15.22 **The distribution of coral reefs** is restricted to low-latitude areas, where the average water temperature exceeds 20°C throughout the year. The water must also be shallow and clear. Reefs do not form where major rivers empty into the ocean, nor do they grow upward from the deep ocean floor of the abyssal plains. Reefs are widely developed throughout the Pacific and Indian oceans.

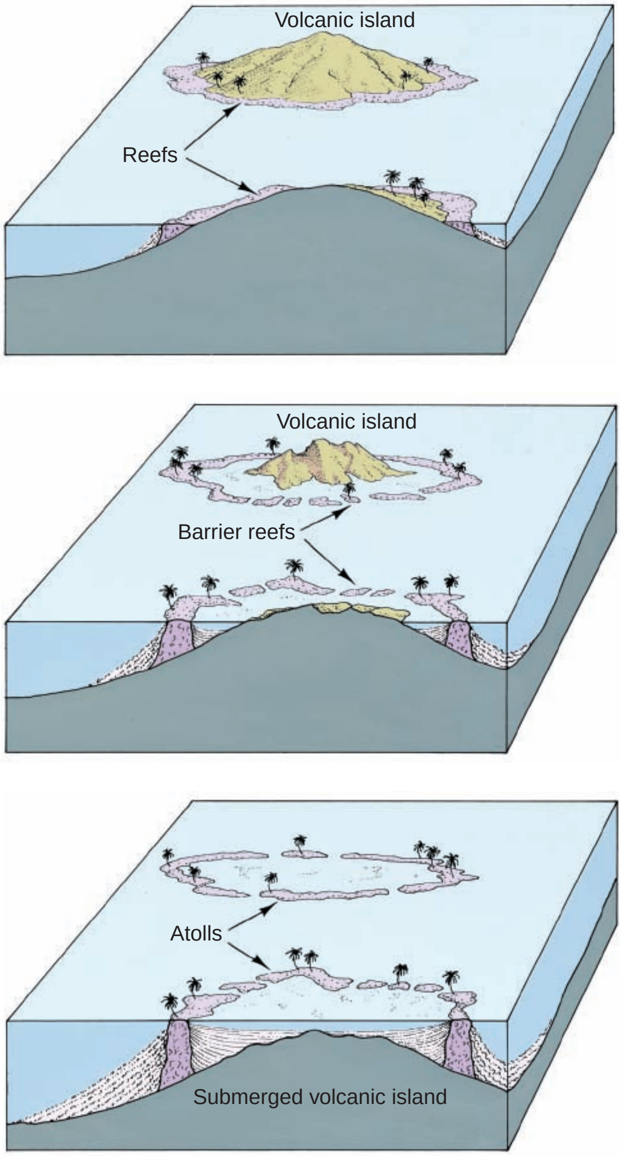
Origin of Atolls

In 1842 Charles Darwin first proposed a theory to explain the origin of atolls. As is indicated in Figure 15.23, the theory is based on the continued relative subsidence of a volcanic island. Darwin suggested that coral reefs are originally established as fringing reefs along the shores of new volcanic islands. As the island gradually subsides, the coral reef grows upward along its outer margins. The rate of upward growth essentially keeps pace with subsidence. With continued subsidence, the area of the island becomes smaller, and the reef becomes a barrier reef. Ultimately, the island is completely submerged, and the upward growth of the reef forms an atoll. Erosional debris from the reef fills the enclosed area of the atoll to form a shallow lagoon.

TYPES OF COASTS

On a global scale, coasts are classified on the basis of their tectonic setting. On a local scale, coasts are classified on the basis of the process most responsible for their configuration.

Coastal landscapes exist in an almost infinite variety of marvelous forms, ranging from the sandy barrier islands of the Atlantic, to the rocky shores of New England, to the swamps of southern Florida, to the rugged cliffs of California. Nearly all coasts are complex, both in the types of landforms and in their geologic history. All are dependent upon the landforms that preceded them, all are subject to the effects of changes in sea level that took place during the ice age, and all are influenced by the operation of present coastal processes. Despite these complexities, insight into the nature of shorelines can be gained by considering the processes responsible for their configuration. To do this effectively, we must consider coastal features on a regional scale and contrast them with features developed on a local scale.



(A) A fringe reef begins to grow along the coast of a newly formed volcanic island.

(B) As the island subsides, the reef grows upward and develops a barrier that separates the lagoon from open water.

(C) Further subsidence completely submerges the island, but if subsidence is not too rapid, the reef continues to grow upward to form an atoll.

FIGURE 15.23 **The evolution of an atoll** from a fringing reef was first recognized by Charles Darwin. The theory assumes that continued slow subsidence of the ocean floor allows the reef to continue growing upward.

Shoreline Classification Based on Plate Tectonics

On a regional basis there are fundamental reasons for similarities and differences along coasts, and it is not surprising that plate tectonics, the fundamental dynamic system of Earth, has a tremendous influence on the origin and evolution of coastlines. The broadest features of coasts, those extending for thousands of kilometers, are directly related to types of plate boundaries and may be classified as follows: (1) convergence coasts, (2) passive-margin coasts, and (3) marginal seacoasts.

A map showing the global distribution of the various types of coasts classified on the basis of plate tectonics is shown in Figure 15.24. This map highlights the relationship of various coastal types and the tectonic setting of the continents. You should refer to this map as you study the material that follows.

Convergence Coasts. Convergence coasts develop where one plate collides with another to form a subduction zone. They are all relatively straight and mountainous and are distinguished by rugged sea cliffs, raised marine terraces, and narrow continental shelves. Convergence coasts are regions of active seismicity, volcanism, and rapid tectonic uplift, all of which have profound influences on the characteristics of the coast. The high mountain ranges of convergence coasts have rivers that are short, steep, and straight because the drainage divide is at a high elevation near the coast. Consequently, the rivers do not transport enough sediment to build up large deltas; they deposit their sediment load into drowned river valleys or onto small open beaches. None of the world's 25 largest deltas occur on convergence coasts. Indeed, the only areas of significant sediment accumulation along convergence coasts lie in relatively small bays caused by drowning of stream valleys or depressions formed by faulting.

The steep gradients of the seafloor off convergence coasts descend to depths of hundreds of meters, and what sediment is transported by longshore currents is intercepted by submarine canyons and transported to the adjacent trenches, where they form submarine fans. The deep water close to shore also permits large waves to maintain their size because there is no shallow seafloor to interfere with wave motion and diminish their size. Thus, large waves strike the shore with high energy, resulting in rapid rates of erosion.

The west coasts of North and South America are excellent examples of continental convergence coasts, and the Aleutian Islands, Japan, and the Philippines are typical convergence coasts of island arcs.

How does tectonics influence the nature of shorelines?

Passive-Margin Coasts. Passive-margin coasts are initially formed by rifting and the movement of a continent away from a spreading ridge. The coastlines evolve through a series of stages and develop more diverse types. In the early stages of rifting, the separating land masses have high relief marked by steep cliffs. The drainage divide is close to the shore, so rivers and streams emptying into the sea are short, small, and carry little sediment. The coasts during this initial stage are very similar to convergence coasts. The cliffs formed by uplift and rifting are ultimately eroded and, as the continent moves away from the uparched oceanic ridge and as the lithosphere cools, the coast subsides. Passive-margin coasts then remain tectonically stable until they become involved with a converging plate or another rift. The tectonic stability of a passive continental margin is the fundamental distinguishing characteristic of this type of coast and is responsible for many local features. A cross section across a passive-margin coast shows a nearly flat, featureless surface extending from the continental

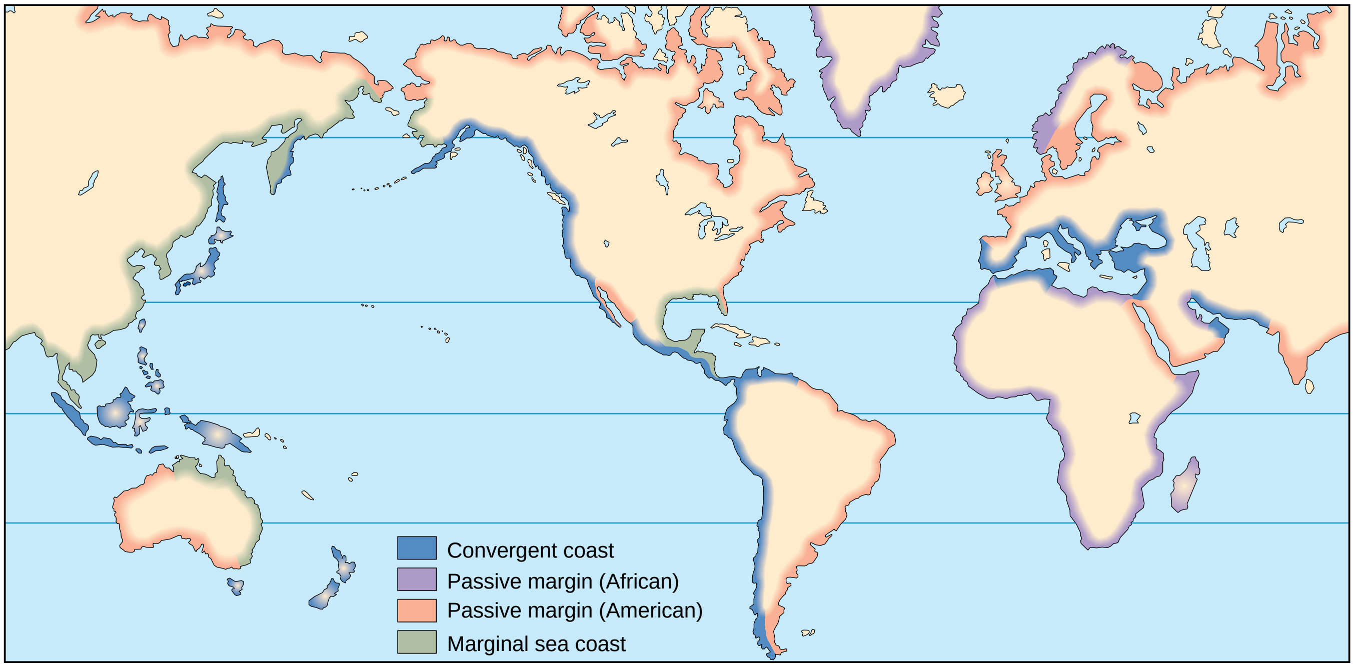


FIGURE 15.24 **The tectonic classification of coasts** is based on the tectonic setting of the continental margins. Passive margins characterize the eastern coasts of North and South America, Africa, and Australia. Mountainous coasts are typical of converging plate margins.

interior out to the edge of the continental shelf. The shoreline is located somewhere near the middle of this profile, and even a slight change in sea level causes a major shift in the position of the shore. The water offshore is shallow—rarely more than 50 m deep. This shallowness dampens the energy of approaching waves so that marine deposition is an important process in the development of local coastal landforms, such as beaches and barrier islands. Another distinguishing feature of passive-margin coasts is that they are the sites of the world's large deltas (see Figure 12.38 and 12.39). The drainage divide is at the crest of the mountain range, which is near the converging plate margin and thousands of kilometers away. Consequently, large collecting systems develop and funnel sediment to the passive margin.

The east coasts of North and South America are typical passive-margin coasts. Both have low relief and broad coastal plains and are bordered by wide continental shelves. The shores of the Red Sea and the Gulf of California are passive-margin coasts in the very early stages of development; they have high cliffs and narrow continental slopes. The coasts of Africa and Greenland represent a more advanced stage. Both the east and west coasts of these continents face ocean ridges, and both have relatively high relief but have developed narrow continental shelves. The coast of India is somewhat similar.

Marginal Sea Coasts. Some continental coasts are near converging plate boundaries but are removed from their influences by an offshore volcanic arc. Although they are near the plate margin and the subduction zone, they are far enough away to be unaffected by convergent tectonics. They thus behave more like passive-margin coasts. Major rivers commonly carry large quantities of sediment and build large deltas and other depositional features, such as beaches, bars, tidal flats, and marshes in the shallow seas, which are protected from vigorous wave action of the open ocean by the associated volcanic arc. For example, the South China Sea is protected from the open ocean by the Philippine island arc, and the Gulf of Mexico is protected by the island arc of the Caribbean.

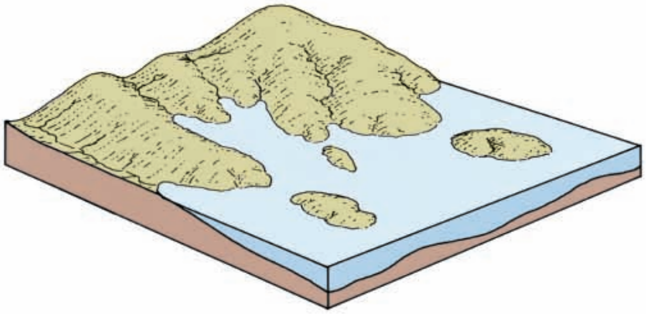
Shoreline Classification Based on Local Geologic Processes

It should be emphasized that the tectonic classification of coasts is intended to apply only on a regional scale—distances of thousands of kilometers. On a smaller scale, lengths of 100 km or so, the local coastal features are highly diverse. The configurations of the shorelines are controlled by geologic processes involving erosion and deposition. The recent rise of sea level associated with the melting of the last glaciers has had a profound effect on all shorelines. The rise of sea level developed new shorelines that are only a few thousands of years old. Many of these coasts are dramatically out of equilibrium with shoreline erosion. Climate is also a major controlling factor influencing local coastal features because it controls glacial systems, the location of major deltas, and the growth of reefs.

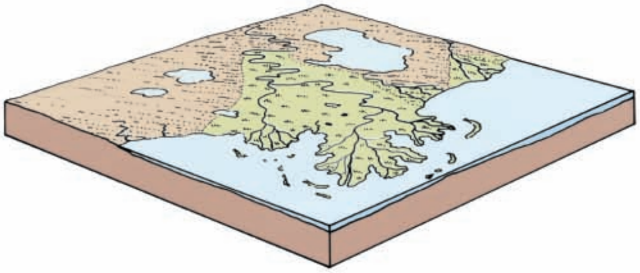
On this smaller scale, two principal types of coasts are recognized: (1) coasts shaped mainly by terrestrial processes of erosion and deposition, and (2) coasts shaped mainly by marine processes.

Coasts Formed by Subaerial Processes

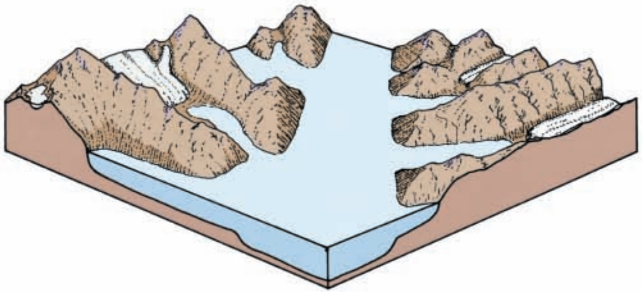
The configuration of many coasts is largely the result of subaerial geologic agents, such as streams, glaciers, volcanism, and earth movements. These processes produce highly irregular coastlines characterized by bays, estuaries, fjords, headlands, peninsulas, and offshore islands. The landforms can be either erosional or depositional, but they are only slightly modified by marine processes. Many of these coasts have experienced a relative rise in sea level. Some of the more common types are illustrated in Figure 15.25.



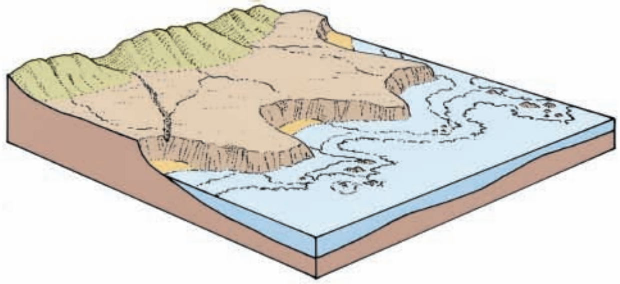
(A) Stream erosion produces an irregular, embayed coast with offshore islands.



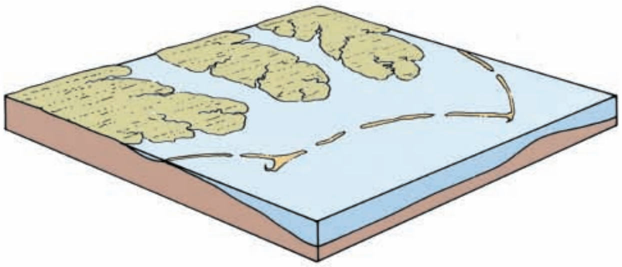
(B) Stream deposition produces deltaic coasts.



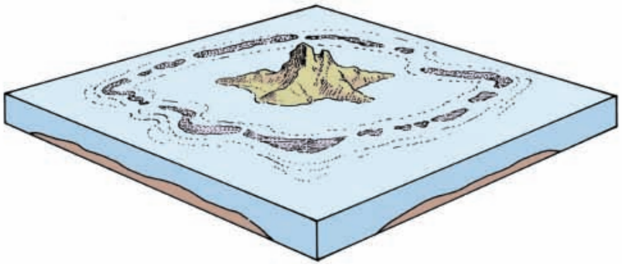
(C) Glacial erosion produces long, narrow, deep bays (drowned glacial valleys) called fjords.



(D) Marine erosion produces wave-cut cliffs.



(E) Marine deposition produces barrier islands and beaches.



(F) The growth of coral reefs produces barrier reefs and atolls.

FIGURE 15.25 **Classification of coasts** is based on the dominant geologic process in developing their configuration.

Stream Erosion Coasts. If an area eroded by running water is subsequently flooded by a rise in sea level, the landscape becomes partly drowned. Stream valleys become bays or estuaries, and hills become islands. The bays extend up the tributary valley system, forming a coastline with a dendritic pattern. Chesapeake Bay is a well-known example (Figure 15.25A).

Stream Deposition (Deltaic) Coasts. At the mouths of major rivers, fluvial deposition builds deltas out into the ocean. The deltas dominate the configuration of the coast. They can assume a variety of shapes and are locally modified by marine erosion and deposition (Figure 15.25B).

Glacial Erosion Coasts. Drowned glacial valleys, usually known as **fjords,** form some of the most rugged and scenic shorelines in the world. Fjords are characterized by long, troughlike bays that cut into mountainous coasts, extending inland as much as 100 km. In polar areas, glaciers still remain at the heads of many fjords. The walls of fjords are steep and straight. Hanging valleys with spectacular waterfalls are common (Figure 15.25C).

Glacial Deposition Coasts. Glacial deposition dominates some coastlines in the northern latitudes, where continental glaciers once extended beyond the present shoreline, over the continental shelf. The ice sheets left drumlins and moraines, to be drowned by the subsequent rise in sea level. Long Island, for example, is a partly submerged moraine. In Boston Harbor, partly submerged drumlins form elliptical islands.

Coasts Formed by Marine Processes

As marine erosion and deposition begin to act on a coastline, marine processes ultimately dominate and control the coastal configuration. These coasts are characterized by wave-cut cliffs, beaches, barrier islands, spits, and (in some cases) sediment deposited through the action of biological agents, such as marsh grass, mangroves, and coral reefs. Marine erosion and deposition smooth out and straighten shorelines and establish a balance between the energy of the waves and the configuration of the shore.

Wave Erosion Coasts. Wave erosion begins to modify the shoreline as soon as the landscape produced by other agents is submerged. Wave energy is concentrated on the headlands, and a wave-cut platform develops slightly below sea level. Ultimately, a straight cliff and a large wave-cut platform are created. The White Cliffs of Dover, England, are prime examples (Figure 15.25D).

Marine Deposition Coasts. Where abundant sediment is supplied by streams or ocean currents, marine deposits determine the characteristics of the coast. Barrier islands and beaches are the dominant features. The shoreline is modified as storm waves break over the barriers and transport sand inland. The barriers also increase in length and width as sand is added. The lagoons behind the barriers receive sediment and fresh water from streams; thus, they are often capable of supporting dense marsh vegetation. Gradually, a lagoon fills with stream sediment, with sand from the barrier bar (which enters through tidal deltas), and with plant debris from swamps. The barrier coasts of the southern Atlantic and Gulf Coast states are coasts of this type (Figure 15.25E).

Coasts Built by Organisms. Coral reefs develop a type of coast that is prominent in the islands of the southwestern Pacific. The reefs are built up to the surface by corals and algae, and they can ultimately evolve into an atoll. Another type of organic coast that is prevalent in the tropics is formed by intertwined root systems of mangrove trees, which grow in the water, particularly in shallow bays (Figure 15.25F).

TIDES

> Tides are produced by the gravitational attraction of the Moon and the centrifugal force of the Earth–Moon system. They affect coasts in two major ways: (1) by initiating a rise and fall of the water level and (2) by generating tidal currents.

Tides

On most shorelines throughout the world, the sea advances and retreats in a regular rhythm twice in approximately 24 hours. These changes are the **tides,** and their cause has intrigued people for thousands of years. In the Mediterranean, tides are almost imperceptible; in the Bay of Fundy, they are more than 20 m high. Tides raised by the Moon have gradually slowed Earth's spin. Nine hundred million years ago, a "day" on Earth was only 18 hours long. Ignorance of tides has had an impact on history. Caesar's war galleys were devastated on the British shore

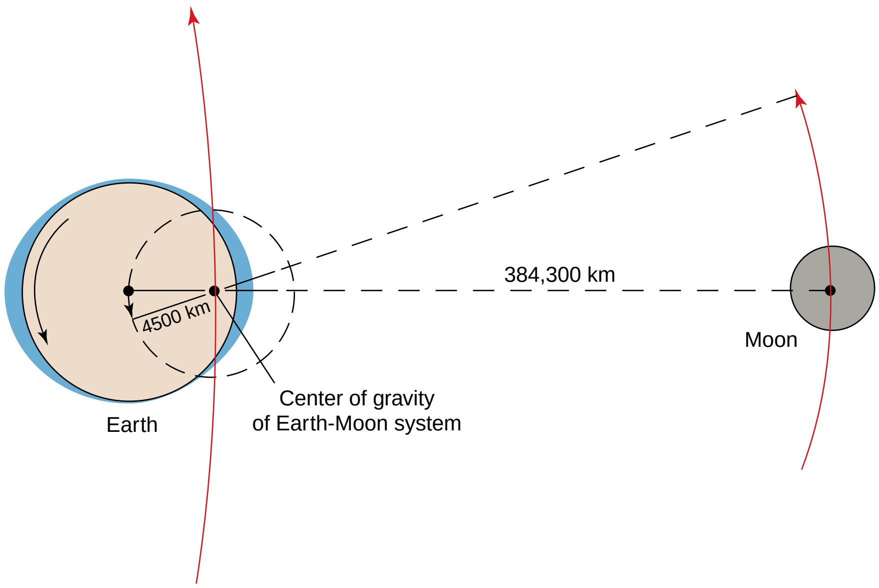


FIGURE 15.26 **Ocean tides** are caused by the gravitational attraction of the Moon and the centrifugal force of the Earth-Moon system. On the side of Earth facing the Moon, the gravitational attraction is greater, forming a tidal bulge in the ocean's water. On the other side of Earth, the centrifugal force is greater, causing another tidal bulge.

because he failed to pull them high enough out of the water to avoid the returning tide. King John of England (1167–1216) was caught in a high tide, lost his treasure and part of his army, and was so enraged he died a week later. The origin of tides was not known until Isaac Newton (1642–1727) showed how tides arise from the gravitational attraction of the Moon and Earth.

What are the major effects of the rise and fall of tides?

The diagram in Figure 15.26 illustrates, on a highly exaggerated scale, the principal forces that produce tides. The gravitational force exerted by the Moon tends to pull the oceans facing the Moon into a bulge. Another tidal bulge, on the side of Earth opposite the Moon, is caused by centrifugal force. Earth and the Moon rotate around a common center of mass, which lies approximately 4500 km from the center of Earth on a line directed toward the Moon. The eccentric motion of Earth, as it revolves around the center of mass of the Earth–Moon system, creates a large centrifugal force, which forms the second tidal bulge. Earth rotates beneath the bulges, so the tides rise and fall twice every day.

The major effect of the rise and fall of tides is the transportation of sediment along the coast and over the adjacent shallow seafloor. Extremely high tides are produced in shallow seas where the rising water is funneled into bays and estuaries. For example, in the Bay of Fundy, between New Brunswick and Nova Scotia, the tide range (the difference in height between high tide and low tide) is as much as 21 m (Figure 15.27). Where fine-grained sediment is plentiful and the tide range is great, the configuration of the coast is greatly influenced by tides and tidal currents.

TSUNAMIS

Movement of the ocean floor by earthquakes, volcanic eruptions, or submarine landslides may produce a wave called a tsunami, which has a long wavelength and travels across the open ocean at high speeds. As a tsunami approaches shore, its wavelength decreases and its wave height increases; therefore, a tsunami can be a formidable agent of destruction along shorelines.

Large waves, known as seismic sea waves or by the Japanese term tsunamis, originate from disturbances on the ocean floor. They are also commonly referred to as tidal waves, but they have no relationship with tides at all. **Tsunamis** can be caused

(A) Low tide near Halls Harbor, Nova Scotia.

(B) High tide near Halls Harbor, Nova Scotia.

FIGURE 15.27 **Tidal variations** are extreme in some restricted inlets such as the Bay of Fundy in Nova Scotia.

by volcanic eruptions, submarine landslides, or even meteorite impact, but most result from earthquakes that displace the ocean floor. It is not surprising then that most tsunamis occur in the Pacific Ocean, which is circled by active volcanoes and intense seismicity, both of which result from a series of subduction zones surrounding the Pacific. For example, in 1999 a magnitude 7.1 earthquake triggered in a subduction zone north of New Guinea created a tsunami that was 15 m high. When it struck the shore, it swept 2200 people to their deaths.

A tsunami differs from wind-produced ocean waves in that energy is transferred to the water by displacement of the seafloor during vertical faulting or by disturbances from volcanic eruptions or submarine landslides. When the seafloor is displaced rapidly, the entire body of water above it is affected. Whatever happens on the seafloor is reflected on the water surface above. Thus the entire body of water, 5000 to 6000 m deep, participates in the wave motion. Consequently, where part of the ocean floor is uplifted or subsides, a bulge and its adjacent depression are produced on the ocean surface (Figure 15.28). The alternating swell and collapse may

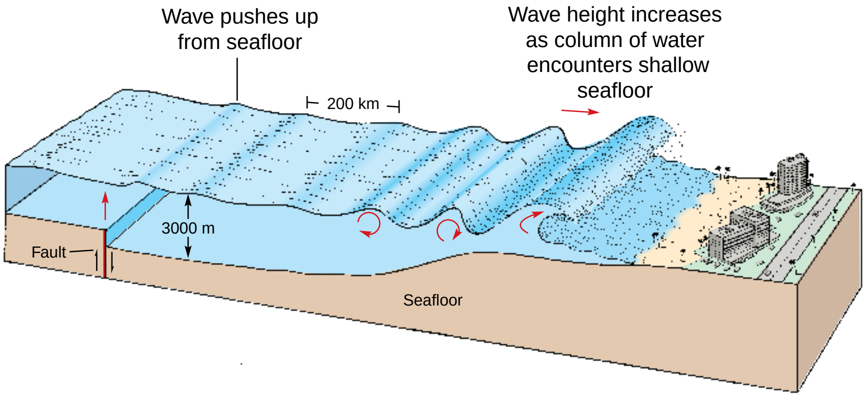


FIGURE 15.28 **Tsunamis** are produced by disruption of the seafloor (by earthquakes, volcanic eruptions, or landslides), which causes a large column of water to move. Initially, the wave is not very high, but as the waves approach the shore, the column piles up, dramatically increasing the wave's height. The wave may be as much as 30 m high when it strikes the shore.

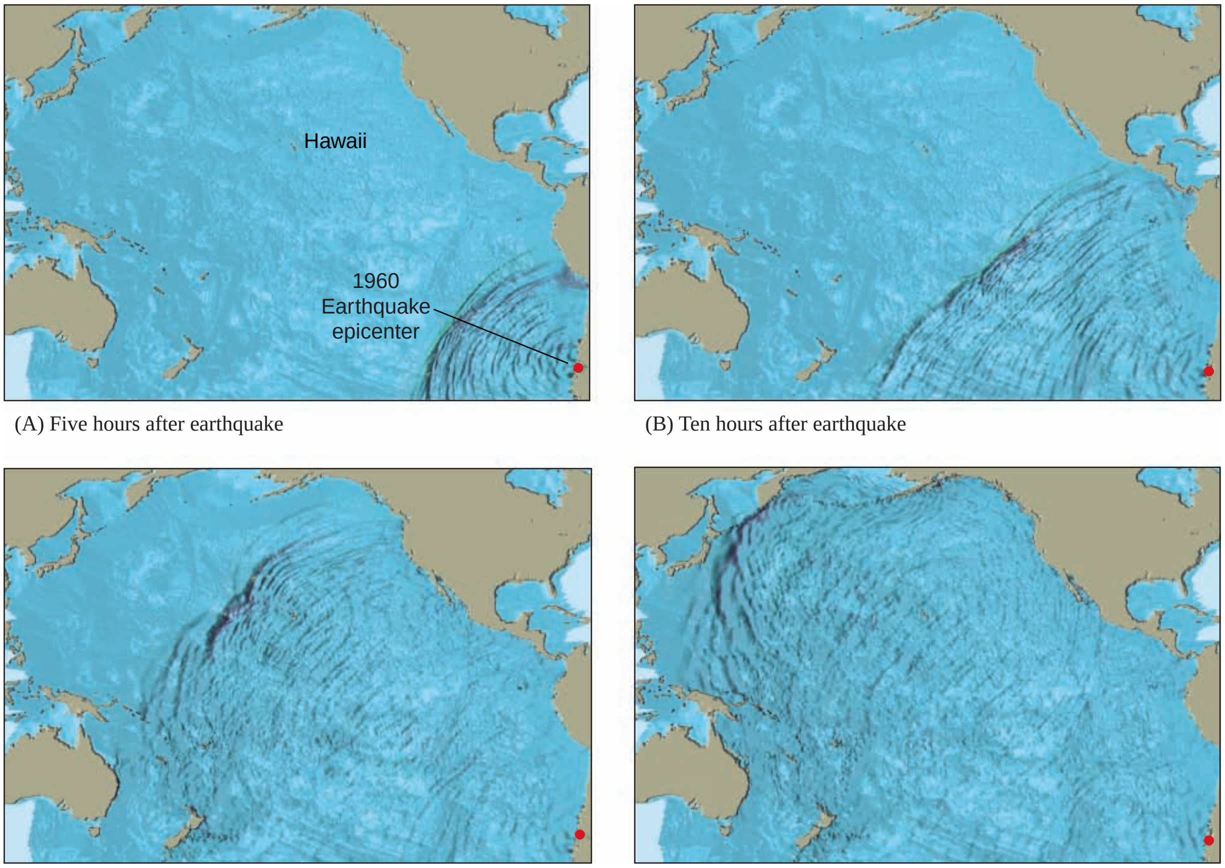
cover up to 10,000 km^2 and spread out across the ocean like ripples in a pond. In the open ocean, a tsunami is not a huge wall of water as many people might think; it is usually less than 1 to 2 m high, with a wavelength of up to 1000 km. Thus, the slope of the wave surface is very gentle (1 cm or so per kilometer). Such a wave in the open ocean is essentially invisible, because it is masked or hidden by the normal surface waves. Indeed, a passing tsunami would not even disturb a game of shuffleboard on a cruise ship.

Tsunamis

To understand tsunamis, scientists have made measurements using sensitive pressure meters lying on the seafloor and constructed computer models. In 4000 to 5000 m of water, these instruments are capable of detecting changes in sea level of less than a millimeter. These studies show that a tsunami is not a single wave, an image held by many, but is more like a series of concentric waves—similar to that produced by a pebble thrown into a pond (Figure 15.29). The wave front travels at tremendous speed ranging from 500 to 800 km/hr, roughly the speed of a jetliner. It can thus travel across the entire ocean in a few hours. Only as a tsunami approaches the shore does it reveal its tremendous energy (Figure 15.28). The energy distributed in the thick column of water becomes concentrated in a progressively shorter column, resulting in a rapid increase in wave height at the surface. Waves that are fewer than 60 cm high in the deep ocean can build rapidly to heights exceeding 15 m in many cases and well over 30 m in rare instances. They exert an enormous force against the shore and can inflict serious damage and great loss of life. For every meter along the coast, a tsunami can deliver more than 100,000 tons of water with a destructive power that is difficult to imagine. The tsunami that struck Japan in July 1993 was one of the largest in historical times, with wave surges 30 m above sea level.

Like all other waves, a tsunami consists of a crest and a trough. Commonly, the first sign that a tsunami is approaching is not an immense wall of water but the sudden withdrawal of the sea. Shorelines recede and harbors are emptied because the trough reaches the coast first. This seaward pull of the water from shore may extend out a great distance (over tens of kilometers), often with tragic results. When an earthquake and tsunami struck Lisbon in 1755, the withdrawal of the sea exposed the bottom of the city's harbor. This bizarre sight drew curious crowds, who drowned when the crest of the tsunami rushed in a few minutes later. Many died the same way when a tsunami hit Hawaii in 1946.

A number of tsunamis have been well documented by seismic stations and coastal observers. For example, the tsunami that hit Hawaii on April 1, 1946, originated in the Aleutian Trench off the island of Unimak. The waves moving across the open ocean were imperceptible to ships in their path because the wave height was only 30 cm. Moving at an average speed of 760 km/hr, they reached the



(A) Five hours after earthquake
(B) Ten hours after earthquake
(C) Seventeen hours after earthquake
(D) Twenty-two hours after earthquake

FIGURE 15.29 **Computer model of a 1960 tsunami** caused by a large earthquake along the subduction zone beneath Chile. The tsunami created a system of waves that rapidly spread across the entire Pacific Ocean. The wave front reached Japan about one day later. If you look carefully at the wave as it nears Japan, you will also see that tsunamis are reflected and refracted just like any other wave. *(Courtesy of P. L. F. Liu, S. N. Seo, S. B. Yoon, C. Devine/Cornell Theory Center)*

Hawaiian Islands, 3200 km away, in fewer than five hours. Because the wavelength was 150 km, the wave crests arrived about 12 minutes apart. As the waves approached the island, their height increased at least 17 m and thus produced an extremely destructive surf, which swept inland and demolished houses, trees, and almost everything else in their path.

Not all tsunamis are produced by violent earthquakes. For example, in 1896 in Japan, a mild earthquake barely felt on shore was followed by a large tsunami that drowned 22,000 people. Similarly, in Nicaragua in September 1992, no one felt the offshore quake that caused a destructive tsunami that swept coastal homes out to sea and killed 170 people. The reason for this seemingly anomalous condition is that some earthquakes release their energy very slowly, over a minute or more, rather than in a brief snap. This may happen if the boundary between the moving blocks of rock is lubricated. The seismic energy from such a quake moves Earth's surface in long undulations that humans do not feel.

Why is a tsunami so small that it is imperceptible in the open ocean, but may be more than 30 m high when it reaches the shore?

GeoLogic Tidal Inlet, Eastern Canada

(Courtesy of Department of Energy and Mines and Resources, Canada)

The shore is dynamic being constantly changed and reshaped by waves, currents and tides. Two major processes are active in this area, (1) longshore drift and (2) tides.

Observations

1. A barrier bar extends across the lower part of the photo, broken by a tidal inlet.
2. A lagoon (upper part of photo) separates the mainland from the barrier bar.
3. The open ocean is in the lower part of the area.

Interpretations

1. The curved spit indicates longshore drift from left to right.
2. A large tidal delta is built by incoming tides entering the quiet water of the lagoon through an inlet across the barrier bar.
3. A small tidal delta is built in the open ocean by outgoing tides.
4. Subaqueous sediment is moved by longshore currents forming subaqueous spits which restrict the tidal channel.
5. The source of sediment for the tidal delta is sediment transported by longshore currents.
6. The tidal delta has a series of distributary channels much like a river delta.
7. The tidal delta in the lagoon is much larger than that in the open ocean because wave-action in the open ocean redistributes the delta sediment and inhibits delta growth.

KEY TERMS

atoll (p. 440)
backwash (p. 423)
barrier island (p. 433)
barrier reef (p. 440)
bay (p. 432)
baymouth bar (p. 432)
beach (p. 432)
beach drift (p. 425)
breaker (p. 423)
coral (p. 439)
elevated marine terrace (p. 438)
estuary (p. 432)
fjord (p. 446)
fringing reef (p. 440)
hanging valley (p. 428)
headland (p. 424)
longshore current (p. 425)
longshore drift (p. 425)
platform reef (p. 440)
rip current (p. 425)
sea arch (p. 428)
sea cave (p. 428)
sea cliff (p. 428)
sea stack (p. 428)
shoreline of equilibrium (p. 437)
spit (p. 432)
swash (p. 423)
tide (p. 447)
tombolo (p. 432)
tsunami (p. 448)
wave base (p. 423)
wave-built terrace (p. 428)
wave crest (p. 422)
wave-cut cliff (p. 428)
wave-cut platform (p. 428)
wave height (p. 422)
wavelength (p. 422)
wave period (p. 423)
wave refraction (p. 423)

REVIEW QUESTIONS

1. Describe the motion of water in a wind-generated wave.
2. Explain how wave refraction alters the form of a coastline.
3. Explain the origin of longshore drift.
4. Describe the stages in the evolution of a sea cliff and wave-cut platform.
5. Name the major depositional landforms along a coast, and explain the origin of each.
6. What effect would the construction of dams on major rivers have on beaches along the coast?
7. How are elevated marine terraces formed?
8. What conditions are necessary for the formation of a coral reef?
9. Why are coral reefs so poorly developed along the shoreline of equatorial Brazil?
10. Explain the origin of atolls.
11. Describe six common types of shoreline.
12. What are the differences between a shoreline formed along a convergent continental margin and one on a passive continental margin?
13. Explain how ocean tides are generated.
14. Explain the origin of a tsunami.
15. What is the difference in the source of energy for a normal wave, a tide, and a tsunami?

ADDITIONAL READINGS

Bird, E. 2000. *Coastal Geomorphology: An Introduction*. New York: Wiley.
Bird, C. F., and M. L. Schwartz. 1985. *The World's Coastlines.* New York: Van Nostrand Reinhold.
Cartwright, D. E. 1999. *Tides: A Scientific History.* Oxford: Oxford University Press.
Davis, R. A. 1994. *The Evolving Coast.* New York: Freeman.
Dolan, R., and H. L. Lins. 1987. Beaches and barrier islands. *Scientific American* **250**(7):146.
Folger, T. 1994. Waves of destruction. *Discover* **15**(5):66.
Gonzalez, R. T. 1999. Tsunami! *Scientific American* **262**(5):57.
Hardisty, J. 1990. *Beaches, Form and Process.* New York: HarperCollins.
Snead, R. E. 1982. *Coastal Landforms and Surface Features: A Photographic Atlas and Glossary.* Stroudsburg, Penn.: Hutchinson Ross.
Trenhaile, A. S. 1997. *Coastal Dynamics and Landforms*. Oxford: Clarendon.

MULTIMEDIA TOOLS

Earth's Dynamic Systems Website
The Companion Website at **www.prenhall.com/hamblin** provides you with an on-line study guide and additional resources for each chapter, including:

- On-line Quizzes (Chapter Review, Visualizing Geology, Quick Review, Vocabulary Flash Cards) with instant feedback
- Quantitative Problems
- Critical Thinking Exercises
- Web Resources

Earth's Dynamic Systems CD
Examine the CD that came with your text. It is designed to help you visualize and thus understand the concepts in this chapter. It includes:

- Animations of tsunamis in the Pacific Ocean
- Simulations that show why tides and longshore drift develop
- Slide shows with examples of shoreline evolution
- A direct link to the Companion Website