Coastal and Submarine Geomorphology 8

Introduction

Oceans cover 70 percent of Earth's surface, and more than half of Earth's human population lives within 40 kilometers of the marine shoreline. Coastal environments provide resource-rich estuaries, sheltered ports, and beaches for recreation. **Continental margins** are economically important because many oil and gas deposits form and are found in modern and ancient marine environments. Today, however, coastal areas are particularly vulnerable to change in a warming, stormier world in which sea levels are rising.

Much of Earth history is recorded in marine sediments. Unlike upland terrestrial environments that erode over time, depositional coastal and deep marine environments accumulate sediment and thus have the potential to hold a sedimentary record of terrestrial surface processes over time. Only a portion of this record gets preserved over geologic time because subduction recycles most oceanic crust and the overlying sedimentary cover.

Only recently have the principles of geomorphology been applied to understanding submarine landforms and processes. Seafloor topography, the most widespread type of topography on the planet, is hidden from direct observation by water. Thus, the study of seafloor topography, **bathymetry**, uses geophysical images, depth soundings, and seafloor sampling techniques. Geomorphologists interpret seafloor topography using the same general principles of regional context, conservation of mass and energy, and definition of boundary conditions that guide their interpretation of subaerial landforms.

M. Miller

Incoming wave trains shoaling, breaking, and causing sea cliffs to erode between Santa Cruz and Moss Beach, California.

IN THIS CHAPTER

Stream systems deliver sediment eroded from continental landscapes to coastal environments, the transitional zone between terrestrial and marine processes. A much smaller but potentially hazardous supply of sediment comes from wave erosion of the coastal margin. Unlike rivers, where water flows only one way (downhill), coastal and estuarine environments are subject to multidirectional flows, the result of rising and falling tides, storms, and wind-driven waves washing back and forth. The interaction of waves, currents, and tides results in coastal erosion, sediment transport along shorelines, and local deposition, giving rise to coastal landforms. Sediment deposition and wave action over glacial–interglacial cycles of rising and falling sea level have shaped the **continental shelves** across which sediment is delivered to the ultimate sink—the deep marine environment.

Sea-level changes greatly affect coastal sediment dynamics. **Estuaries,** partly enclosed coastal water bodies fed by one or more rivers or streams, may form and enlarge when sea level rises, or they may diminish as when, for example, barrier islands erode away. Human impacts are important in coastal zones around the world where land use, river modifications, and coastal engineering structures have substantially altered coastal sediment supply. For example, dams have reduced global sediment delivery to the oceans by roughly half.

This chapter explains the fundamental processes and major landforms of coastal and offshore environments. We consider a variety of coastal settings, their characteristic landforms, and the drivers of geomorphic change in these settings. We provide an overview of offshore geomorphology, using process-based understanding to explain marine geomorphology and the development of submarine landscapes.

Coastal Settings and Drivers

Coastal morphology is broadly determined by tectonic setting, although over time changing sea level is an important control on continental margin morphology and process. Salinity, river inputs of water and sediment, tidal range, and wave action all influence the dynamics of coastal systems and the resulting landforms.

Tectonic Setting

The large-scale geomorphology of a continental margin depends on whether it occurs in the middle or on the edge of a tectonic plate **[Figure 8.1]**. Convergent and transform margins are **active margins,** where the edge of the continent coincides with the boundary between two tectonic plates. Subduction zones (where denser oceanic lithosphere sinks beneath less dense continental lithosphere) typically have a coastal mountain range hosting volcanoes and affected by earthquakes, or a volcanic island arc (where subduction involves oceanic lithosphere). Both types of subduction zone typically have a narrow continental shelf that leads out to a continental slope that drops steeply to a deep submarine trench. Subduction zones surround much of the Pacific Ocean. These collisional (convergent) margins tend to have rocky coastlines that are dominated by erosional landforms **[Photograph 8.1a]**, though deltas and other coastal depositional landforms develop near significant sediment sources, such as large rivers.

In contrast, a **passive margin** (or **trailing-edge margin**) is one where the continental margin does not coincide with a plate margin. Consequently, the oceanic and continental crust move in the same direction and speed. Most trailing-edge margins have an exposed coastal plain, part of which was continental shelf during periods of higher sea level. Similarly, the modern continental shelf was the coastal plain during lower sea-level stands, such as during the glaciations that repeatedly punctuated the past 2.7 million years. The eastern continental margins of North America, South America, Australia, and the margins of Africa and the Gulf of Mexico coast are examples of trailing-edge margins. Trailing-edge margins typically have extensive depositional landforms that are built by sediment transport and storage along the coast **[Photograph 8.1b]**.

Sea-Level Change

The volume of water in Earth's interconnected oceans plays a major role in coastal dynamics and evolution. Over the past 35 million years, sea level has varied through glacial–interglacial cycles and coastal processes have acted across a wide zone that ranges from meters to tens of meters above today's sea level to the now-submerged edge of the continental shelf.

Shoreline position reflects changes in sea level. When sea level rises, the shoreline advances landward. Falling sea level causes the shoreline to retreat seaward. Depending on the slope of near-shore land, glacial–interglacial sea-level changes can shift coastlines up to tens or even hundreds of kilometers. **Emergent coastlines** are stretches of the coast that have been exposed by relative sea-level fall. **Submergent coastlines** are those that have been inundated by the sea due to a relative rise in sea level.

The cause of changes in the relative elevation of land and sea can be local or global. Local changes in relative sea level are most often caused by uplift or subsidence, driven either by tectonics or by isostatic response to loading or unloading. Sea-level rise and fall can be driven by changes in the volume and size of ice sheets on a millennial timescale and by basin sedimentation and continental erosion on longer timescales. Worldwide sea-level changes driven by changes in ocean volume are referred to as **eustatic.** Such changes are caused by fluctuations in global ice volume as well as **steric** changes (changes in volume) caused by fluctuations in ocean temperature or salinity.

Over the past few million years, the timescale most relevant to geomorphology, eustatic sea level has been closely tied to the amount of glacial ice on land and to the thermal expansion and contraction of ocean water that accompanies climate changes. Sea level falls when a glaciation begins, ice volume at the poles grows, and the average

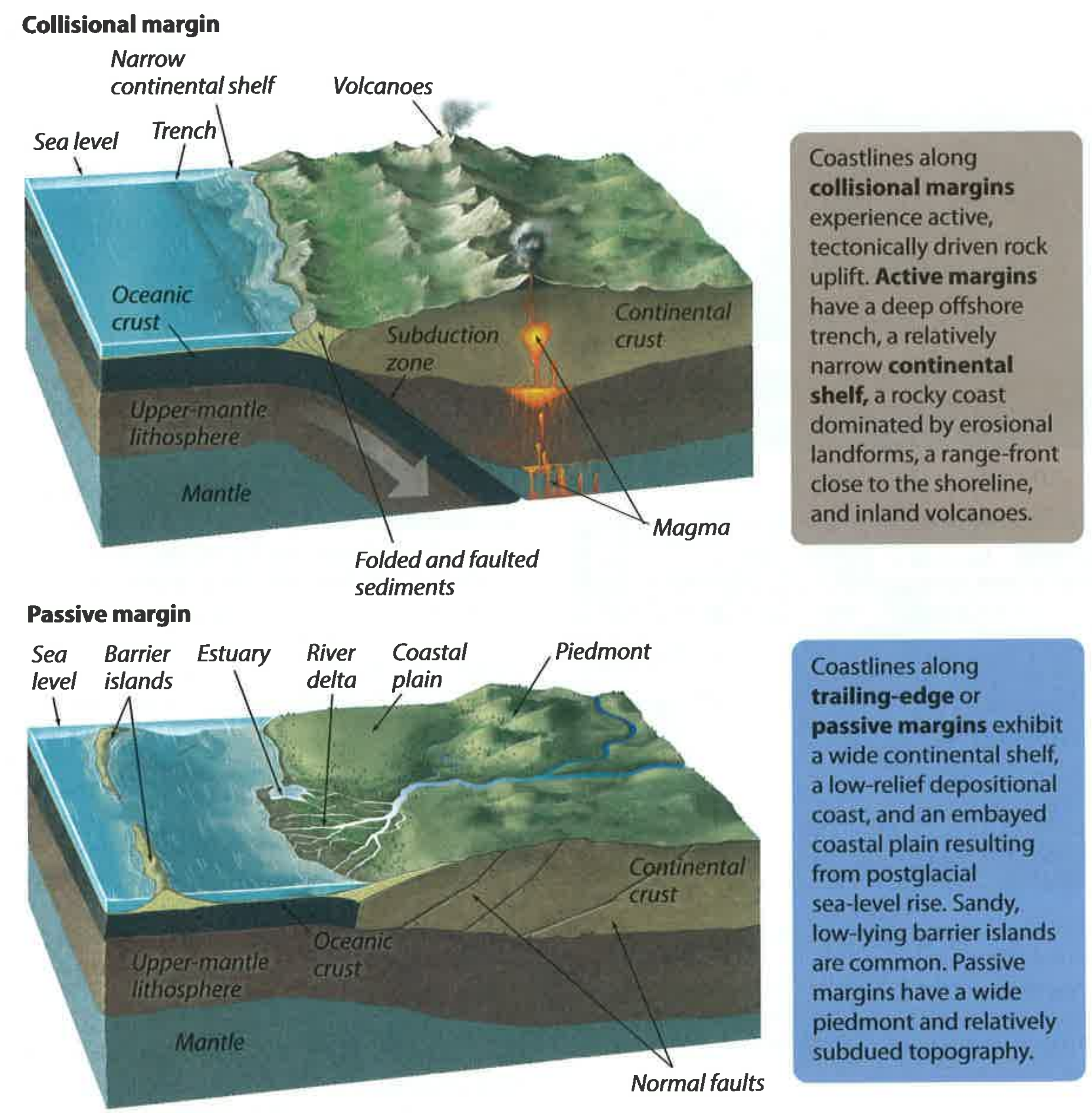
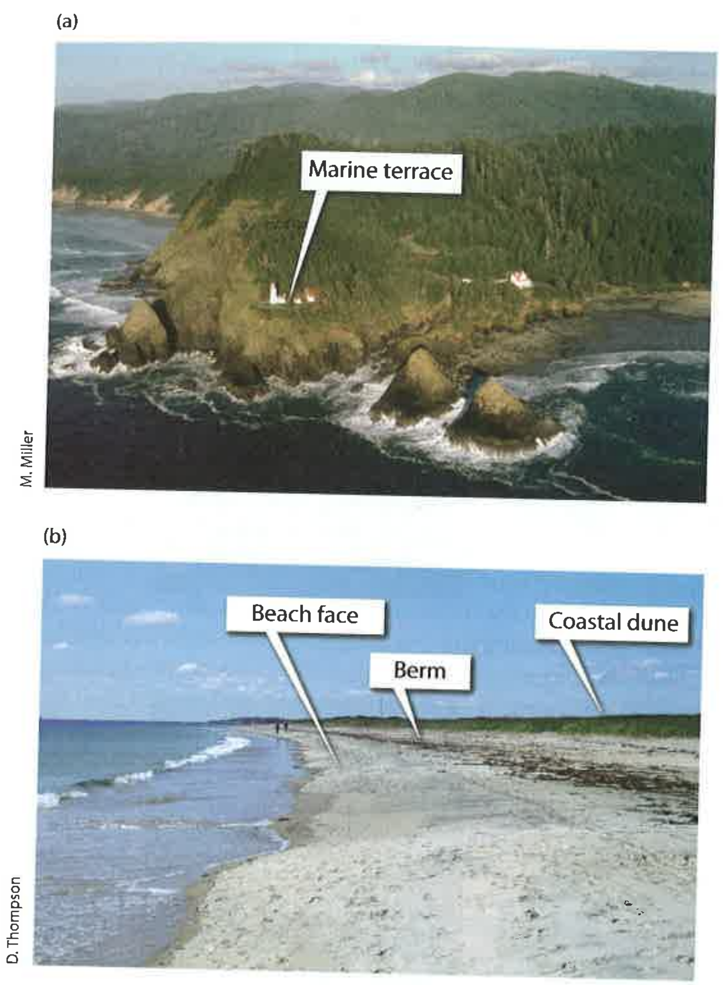


FIGURE 8.1 Collisional Versus Passive Margin Coastlines. The geomorphology of active or collisional margins is very different from that of passive or trailing-edge margins.

global ocean temperature cools, reducing ocean volume. The storage of ocean water as glacial ice and the cooling and contraction of the water left in the oceans leads to marine **regression;** coastlines migrate seaward and fluvial systems extend to reach the sea while cutting down into the coastal plain to adjust to falling base level. Conversely, when glaciers melt and transfer water back into the oceans at the end of a glaciation, sea level rises and causes marine **transgression,** during which shorelines move landward and seawater inundates formerly coastal areas. At the height of the last major glaciation, about 21,000 years ago, global average sea level stood about 130 m below its present level. It has been rising since then. During the most recent deglaciation, sea level rose about 1 cm/yr until about 7000 years ago, then the pace slowed to less than 2 mm/yr about 5000 years ago. Global average sea level then stabilized about 3000 years ago. Now, with climate warming and the ocean warming as well as glaciers melting, sea level is rising several millimeters per year.

The impact of postglacial sea-level rise on coastal geomorphology varies with time since deglaciation and with sediment supply. During the period of rapid sea-level rise from 18,000 to 7000 years ago, coastal valleys flooded and most sediment transported by rivers to the sea was trapped close to the coast. After sea-level rise slowed enough that coastline locations stabilized, those estuaries and fjords that receive little sediment remain unfilled, even though they efficiently trap the sediment supplied to them. In contrast, estuaries that received a large amount of sediment like that of the Columbia River in the Pacific Northwest, have filled to the point that some sediment is carried seaward of the estuary to feed along-coast transport and accumulate



PHOTOGRAPH 8.1 Coastline Morphologies. Coastlines exhibit a great variety of morphologies. (a) The rocky shoreline of a collisional margin at Heceta Head, Oregon. The lighthouse sits on an uplifted marine terrace remnant. (b) This sandy summer beach in Old Lyme, Connecticut, has a small beach face, a wide, flat berm, and a low coastal dune.

on the continental shelf. In areas with a very high sediment supply, like the Nile River delta before the construction of modern dams, the estuary filled to the point where sediment extended the shoreline seaward.

Salinity

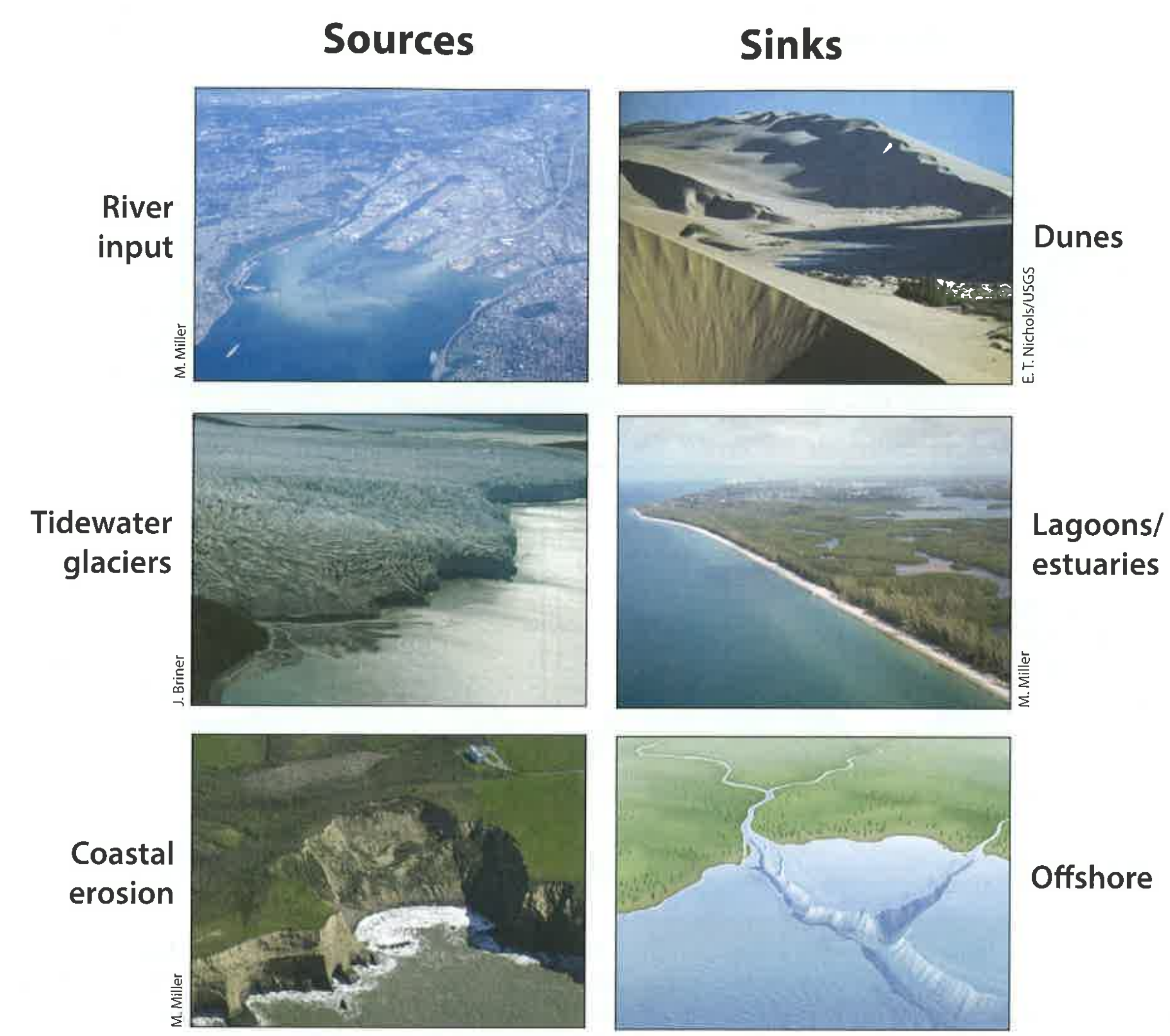
The salinity and temperature of seawater are fundamental drivers of coastal and marine processes. Ocean water has approximately 35 parts per thousand (ppt) of salt dissolved in it. For comparison, fresh water in lakes and rivers typically contains less than 0.5 ppt of dissolved mineral material. Brackish water in estuarine environments can have variable salinity but often is between 20 and 30 ppt. Lower salinity translates to lower density, so plumes of river water float on denser salt water where rivers enter coastal and marine environments. In some cases, however, high concentrations of suspended sediment at the mouths of rivers draining steep, rapidly eroding drainage basins increase the density of the water and create bottom-hugging density currents that cause erosion, displacement, and deposition of sediments in deep marine environments well below the depths affected by wave action. Gradients in density, which reflect differences in water-mass temperature and salinity, drive ocean circulation and thus some ocean-bottom sediment transport.

Substrate and Sediment Supply

Like terrestrial landscapes, coastal and marine landscapes are composed of either consolidated bedrock or unconsolidated materials. Many emergent coastlines and headlands are composed of rock outcrops with far greater erosional resistance than submergent coastlines that are mostly made of unconsolidated sediment. This variability in erosional resistance and sediment supply creates strikingly different coastal features and causes different dynamics on rocky, erosional coasts than on depositional shorelines.

Various mechanisms bring sediment to coastal and marine environments, and the particular mix of processes that deliver, store, and remove sediment sets the sediment budget for specific coastal and marine environments [Figure 8.2]. At any site, sediment comes from rivers, coastal erosion, sediment movement along the coast, in situ productivity of biogenic carbonate, aeolian inputs and, at high latitudes, glaciation. In most coastal settings, streams that discharge into coastal and estuarine environments deliver the majority of sediment, although there are exceptions. For example, on volcanic islands such as Hawaii, wave erosion of lava flows provides coastal sediment. Tropical beaches in areas where the uplands are eroding slowly may be made of biogenic coral sand, a product of coral eroded and broken up by wave action.

Once sediment is introduced to a coastal setting, it is generally transported parallel to the coast by the alongshore component of currents generated by waves approaching the shore at an oblique angle. In areas with limited sediment input from rivers and streams, coastal sediment is supplied mainly by cliff erosion that delivers sediment to coastal settings, or by carbonate-producing organisms like corals and shellfish that thrive in clear water away from fluvial inputs of clastic sediment. In modern, high-latitude environments, glaciers and icebergs that calve from tidewater glaciers deliver sediment directly to coastal and marine environments, and have done so in many settings during past periods of cooler climate. Coastal sediment sinks include depositional environments like **dunes** and **lagoons,** as well as deltas, marshes, and carbonate platforms. The dominant sediment source in the deep ocean basins is a steady "rain" of the bodies of single-celled marine plants and animals that sink from the near-surface **photic zone,** and the slow accumulation of wind-delivered dust. Coarse-grained, clastic sediment from the continents occasionally makes its way down the continental slope and into the **abyssal basins** of the deep sea, usually transported by density-driven flows.

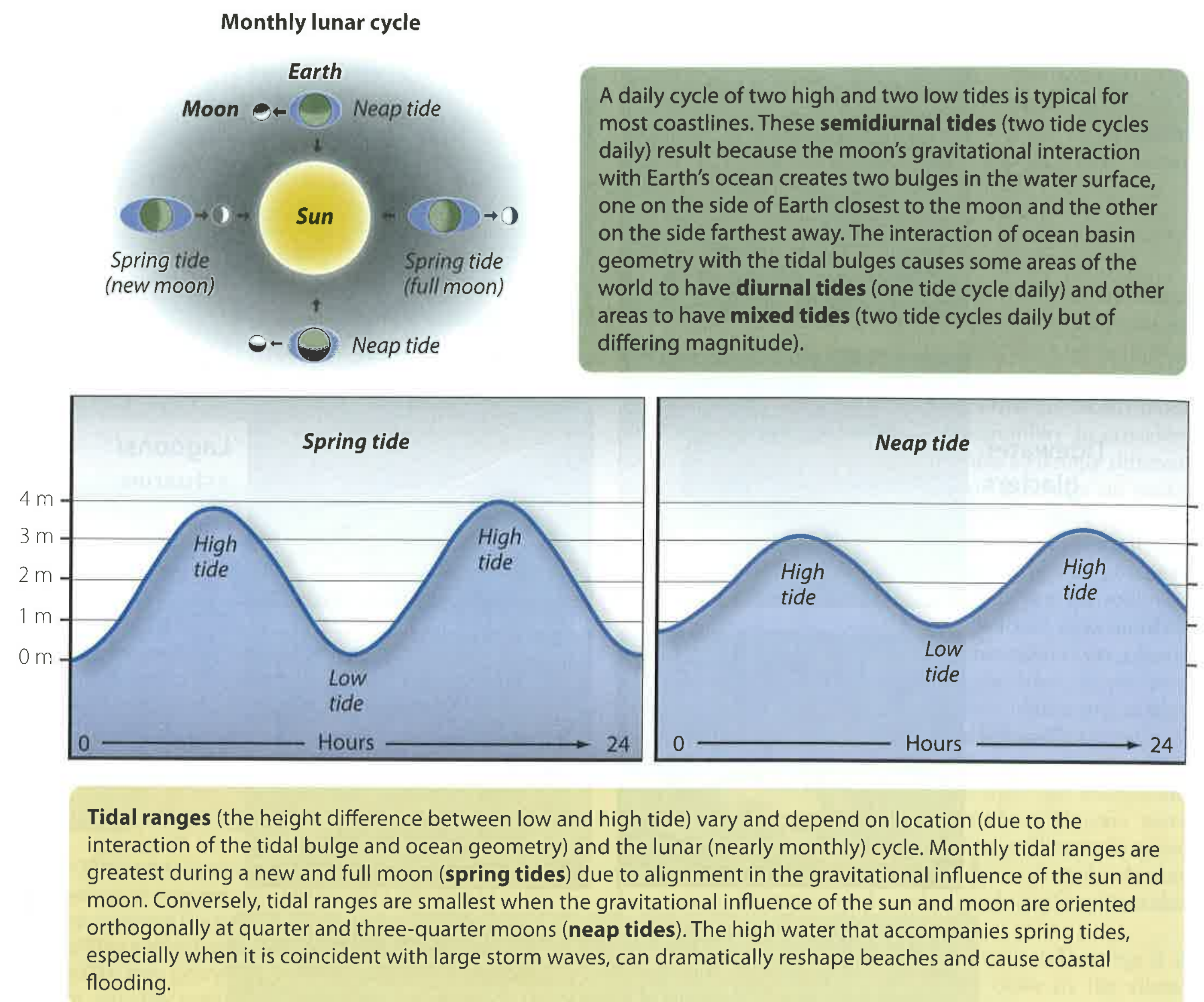


Sources of sediment to coastal environments include rivers, tidewater glaciers, coastal erosion, and aeolian inputs. Sediment sinks in coastal environments include deposits of material in coastal dune fields, beaches, lagoons, and estuaries, and off-shore transport that delivers sediment to the coastal shelf and deeper-water environments. Coastal environments can be either a net sediment source or a sink depending on the configuration of the coast and the time scale of analysis. The relative importance of each type of source and sink varies in different tectonic and climatic settings.

FIGURE 8.2 Coastal Sources and Sinks. Sediment moves onto and away from coasts through a variety of different pathways.

Tides

The daily rhythm of the **tides** produces changes in sea level that can be important geomorphically because they control the location at which wave energy is dissipated and sediment deposited. Tidal changes are driven by astronomical factors, specifically, the gravitational interaction of the Earth–moon system, which creates a bulge of ocean water between Earth and the moon and a second antipodal bulge on the other side of the planet [Figure 8.3]. The Earth rotates through these bulges, causing two lunar tides per day; in actuality, tidal cycles are about 12.5 hours, the time required for one Earth rotation relative to the moon. The sun also exerts a much weaker gravitational force on the ocean water, producing one tidal cycle per day of lesser amplitude. The geometry of the continents, the bathymetry of the ocean basins and continental margins, the latitude of the coastline, ocean currents, and weather systems all interact with the astronomical factors to determine the water level over time on any specific coastline.



A daily cycle of two high and two low tides is typical for most coastlines. These **semidiurnal tides** (two tide cycles daily) result because the moon's gravitational interaction with Earth's ocean creates two bulges in the water surface, one on the side of Earth closest to the moon and the other on the side farthest away. The interaction of ocean basin geometry with the tidal bulges causes some areas of the world to have **diurnal tides** (one tide cycle daily) and other areas to have **mixed tides** (two tide cycles daily but of differing magnitude).

Tidal ranges (the height difference between low and high tide) vary and depend on location (due to the interaction of the tidal bulge and ocean geometry) and the lunar (nearly monthly) cycle. Monthly tidal ranges are greatest during a new and full moon (**spring tides**) due to alignment in the gravitational influence of the sun and moon. Conversely, tidal ranges are smallest when the gravitational influence of the sun and moon are oriented orthogonally at quarter and three-quarter moons (**neap tides**). The high water that accompanies spring tides, especially when it is coincident with large storm waves, can dramatically reshape beaches and cause coastal flooding.

FIGURE 8.3 Tides and Tidal Range. Tides are an important driver of coastal geomorphic processes. Tidal range changes cyclically through time.

Tidal range, the difference between the elevation of high and low tides, is geomorphically important. In the open ocean, tidal ranges are low, usually less than a meter. Along coastlines, tidal ranges can be much greater; thus, the geomorphic effects of the tides as ocean levels rise and fall can be significant. Tidal range is controlled by ocean basin geometry and tends to be lower in basins with restricted connection to the global ocean, such as the Mediterranean Sea, and high where forced by local convergence of flow, such as in the narrow inlet of eastern Canada's Bay of Fundy that has an extreme tidal range of 15 m. **Macrotidal** coastlines experience tides greater than 4 m, **mesotidal** coasts experience intermediate tidal ranges between 2 and 4 m (as is common on the West Coast of the United States), and **microtidal** coasts like those around the Caribbean Sea experience fluctuations of less than 2 m.

Tidal range is driven in part by astronomical forcing (sun and moon position). Maximum tidal range occurs when the gravitational fields of the sun and moon reinforce each other to create a higher-than-average **spring tide.** When the moon and sun are oriented at right angles, a **neap tide** occurs and produces a low tidal range. Consequently, tidal ranges vary over a 28-day period as the orientation of the moon and sun change during the moon's orbital cycle. Spring tidal ranges are typically about 20 percent greater than average, and neap tidal ranges are usually about 20 percent less than average. High spring tides accompany the new and full moon each month, and low neap tides accompany the waxing and waning quarter moons.

Spring tides are important geomorphologically because they allow waves and wave energy to reach farther inland and erode shoreline features. Combinations of high spring tides and high storm surges from storm-generated waves can produce unusually destructive high water and large geomorphic effects. Neap tides are important in the coastal zone because they allow fluvial processes to dominate; for example, at neap tides, flow near deltas may become unidirectional, moving water and sediment offshore.

Tide-generated currents generally flow perpendicular to coastlines, in and out of bays and lagoons. Tidal flows have enough energy to prevent sediment accumulation, maintain tidal inlets, and force seawater into tidal channels in estuaries. The geometry of the seabed along coastlines acts to increase or decrease local tidal ranges, so the effect of tidal flows varies from place to place. Incoming and outgoing tides are often asymmetrical and uneven in strength, and they may roll around a bay rather than moving directly in and out. Rapidly advancing tidal fronts known as **tidal bores** push breaking waves into estuaries and rivers, forcing salt water inland. In the Amazon River, for example, the tide influences river flow as much as 480 km inland.

Waves

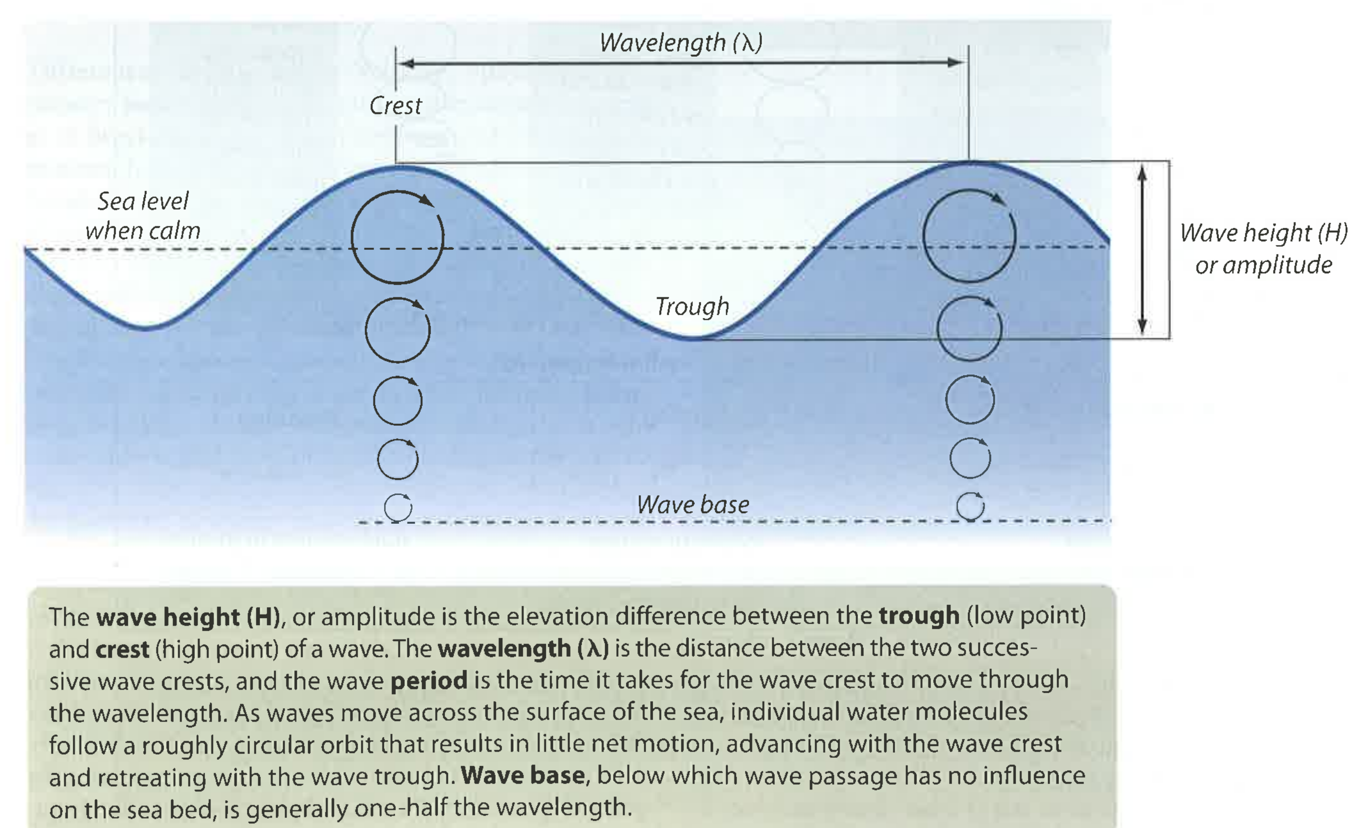
Waves are the dominant driving mechanism for coastal processes; they govern erosion and sediment transport in coastal environments. Blowing winds impart energy to the sea surface as the two fluids, air and water, drag on one another sufficiently that kinetic energy is transferred to the water from the air. Over time, that energy becomes focused into discrete water waves. The factors that affect the formation of waves are (1) the duration of the wind event, (2) the velocity of winds, and (3) the **fetch,** the distance that the wind blows across the water surface. Long-duration, high-velocity wind blowing across long distances produces large, energetic waves that travel out from the area where they were generated. Fetch is particularly important in determining wave characteristics. Waves from large storms often cause great changes in coastal environments, but the smaller waves that arrive day and night are the primary shapers of coastal landforms. Fair-weather waves and the longshore currents they generate gradually reshape coastlines after storms.

Waves are described by several basic characteristics [**Figure 8.4**]. Wavelength (λ) is the distance between successive wave crests. Wave period (T) is the time it takes for two wave crests to pass the same point. Wave velocity (V) is given by the ratio of wavelength to wave period:

$$V = \lambda / T \qquad \text{eq. 8.1}$$

Wave height (H), or amplitude, is the vertical distance between a wave crest and its low point or trough and thus equals the amplitude of the water surface rise and fall as a wave moves past a point on the sea surface.

Variable winds blowing across an area of the deep-ocean surface create waves with a wide range of wavelengths and different travel directions. In deep water, waves with longer periods and wavelengths travel more



The **wave height (H)**, or amplitude is the elevation difference between the **trough** (low point) and **crest** (high point) of a wave. The **wavelength (λ)** is the distance between the two successive wave crests, and the wave **period** is the time it takes for the wave crest to move through the wavelength. As waves move across the surface of the sea, individual water molecules follow a roughly circular orbit that results in little net motion, advancing with the wave crest and retreating with the wave trough. **Wave base**, below which wave passage has no influence on the sea bed, is generally one-half the wavelength.

FIGURE 8.4 Wave Definition. Waves are described in terms of wavelength and amplitude.

quickly than smaller waves, so waves sort out by size as they travel. This process, known as **wave dispersion**, creates a pattern of well-defined waves that travel outward from the area where they were generated. Wave velocity can be calculated since it is related to the wave period,

$$V = (gT)/2\pi \qquad \text{eq. 8.2}$$

where g is gravitational acceleration. The longest wavelengths create **swell**, regularly spaced waves with low, gently rounded crests that can travel thousands of kilometers across the open ocean with little loss of their original energy, until they approach land. As water depth shallows, pointed wave crests develop and then break to produce **surf**.

The water in a wave moves back and forth, but has little net motion in the deep ocean and does not move with the wave. The propagation of energy imparted to the waveform causes the rhythmic rise and fall of the sea surface in which the wave moves forward, but the water does not. In deep water, the water molecules near the sea surface oscillate and move in circular orbits as a wave passes their location [Figure 8.5]. The diameter of the orbit decreases with depth below the water surface and drops off to produce

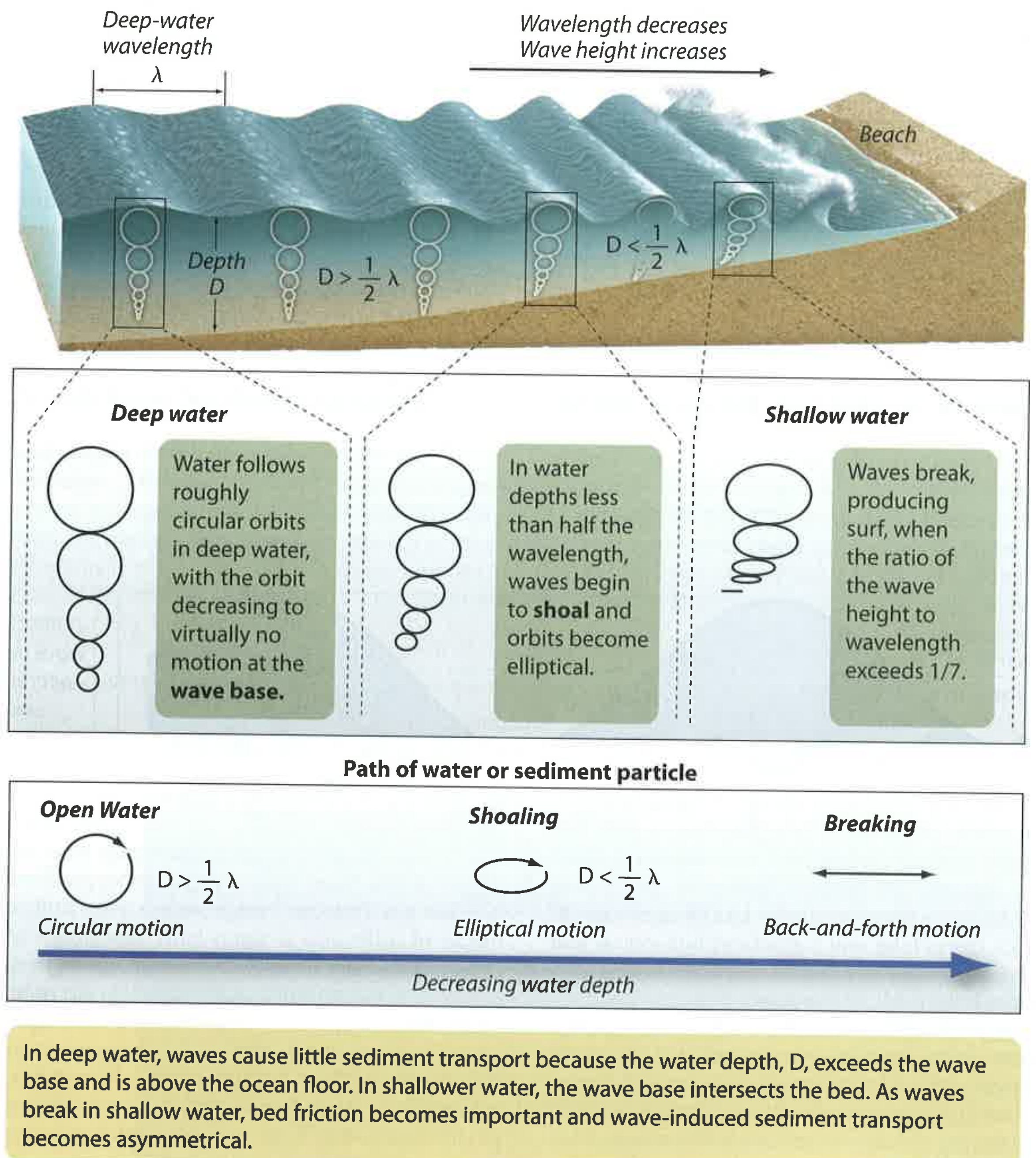


FIGURE 8.5 Sediment-Wave Asymmetry. Waves behave differently in shallow and deep water. In shallow water, waves break and move sediment shoreward.

no motion at a water depth greater than one-half the wavelength, a depth referred to as the **wave base.** Below this depth, water molecules are undisturbed as waves pass over the surface. This is why the calmest place to ride out a storm at sea is in a submarine and why waves have little geomorphic effect outside the coastal zone.

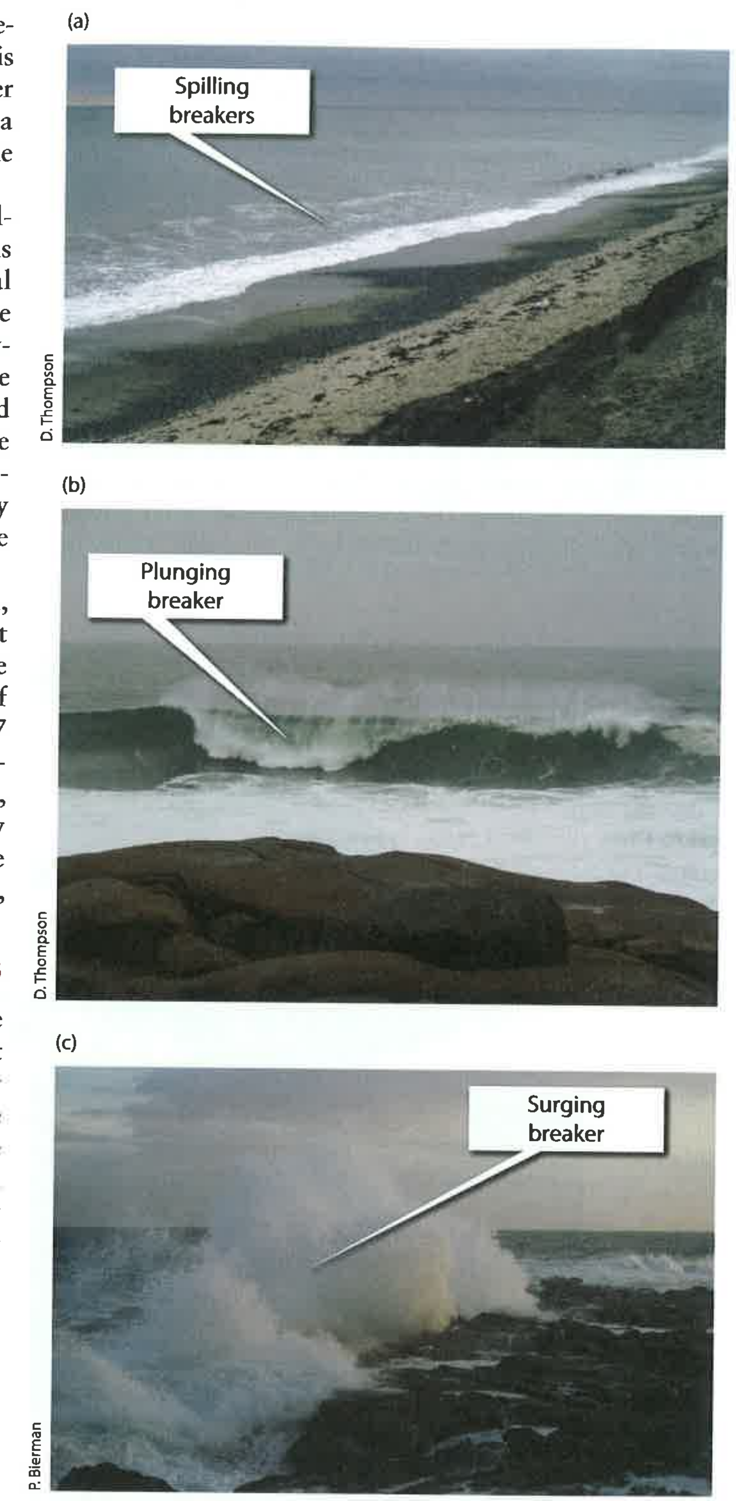
As a wave approaches the shore and enters water shallower than half its wavelength, the wave **shoals** as it begins to interact with the seabed. As a wave shoals, the orbital paths of water molecules in the wave flatten and become elliptical, with the water and sediment at the seabed moving back and forth. This interaction exerts drag on the water near the bed and the water exerts a force on the bed that can move sediment. Deeper shoreward flow under the wave crest moves more sediment than the shallower seaward flow under the wave trough, producing an asymmetry that causes net sediment transport toward shore. This is the process by which waves build up beaches.

When a wave enters shallow water and begins to shoal, wave speed decreases due to bottom friction, wave height increases, and wavelength decreases. This causes the wave crest to progressively steepen, and increases the ratio of wave height to wavelength (H/λ). When H/λ exceeds 1/7 (i.e., when wave height exceeds 0.14 times the wavelength), the wave crest loses support, becomes unstable, and breaks, producing surf that dissipates potential energy as kinetic energy in the near-shore environment. The velocity of these shallow-water or translational waves, where D is the depth of water, can be expressed as:

$$V = (gD)^{0.5} \qquad \text{eq. 8.3}$$

Differences in the water velocity, specifically the asymmetry beneath steep and gentle waves, create different types of breaking waves. Beach steepness, which is largely determined by beach sediment grain sizes, affects the type of breakers that develop on a particular beach; those breakers, in turn, affect the grain size of material left on the beach. Gently sloping, fine-sand beaches typically produce **spilling breakers,** through which one can casually wade out into the sea **[Photograph 8.2a]**. The crest of a spilling breaker becomes unstable first and gradually cascades down the advancing wave front as irregular foam. Spilling breakers characterize beaches along the southeastern Atlantic and Gulf of Mexico coasts of the United States.

In contrast, plunging and surging breakers that are more attractive to surfers are generally found along steeper beaches composed of coarse sand and gravel. **Plunging breakers** curl over the front of the advancing wave and fall onto the base of the wave over a short distance, producing a turbulent mass of water that churns up and suspends bottom sediment. Surfers call the tunnels of air under the curling crests of plunging breakers pipelines **[Photograph 8.2b]**. **Surging breakers** maintain unbroken wave crests as they run up the shore **[Photograph 8.2c]**. They typically develop where waves approach steeply sloping beaches and shoal very close to the shoreline.

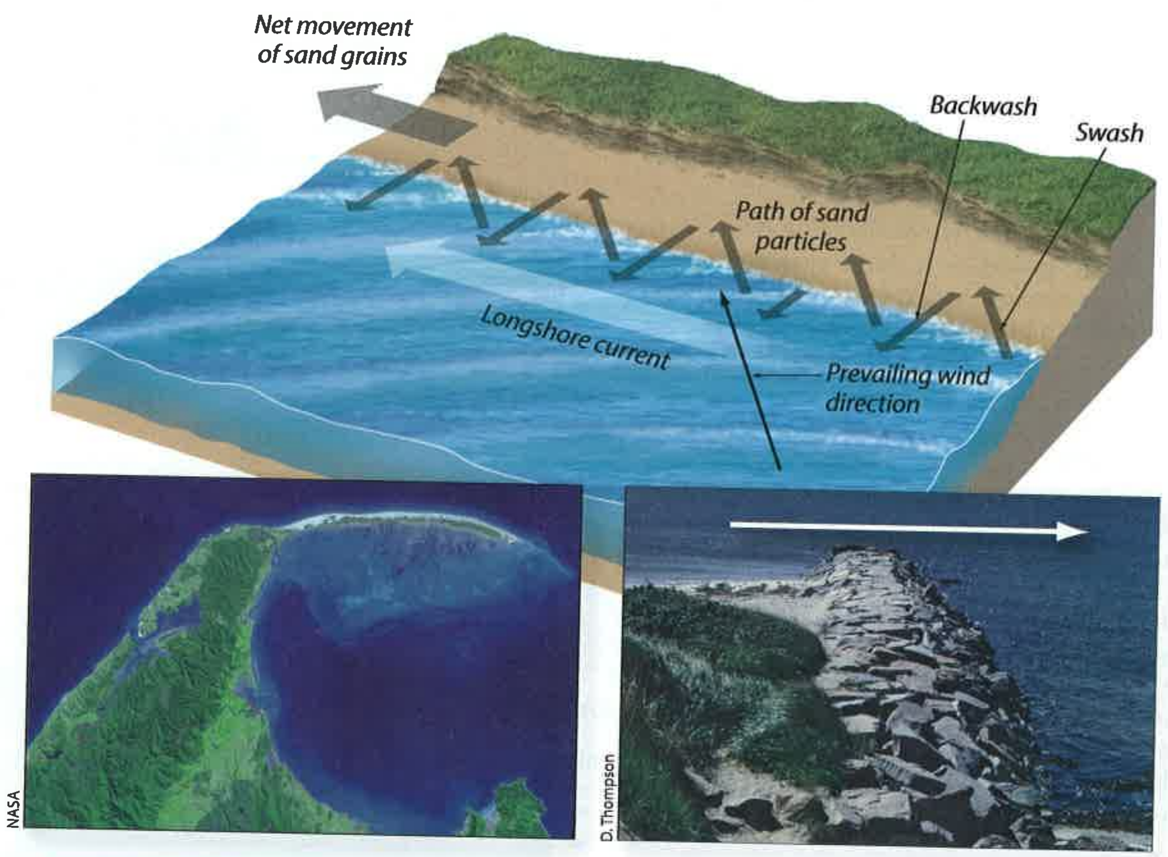


PHOTOGRAPH 8.2 Types of Breakers. (a) Spilling breakers move gently ashore at Charleston Beach, Rhode Island. (b) Plunging breaker, Cape Ann, Massachusetts. (c) Surging breakers wash over a marine platform at Cape St. Francis, South Africa.

The **surf zone** extends from the seaward limit of breakers to the landward extent of waves that run up the beach face. The **swash zone** is the area covered by **swash** that runs up the beach face. It is exposed after **backwash** flows back down the beach face. This oscillatory action sorts sediment grains by size, shape, and density and creates beaches that are composed of bedload material—the size of which reflects wave energy and sediment source characteristics. Alternation between vigorous swash and sluggish backwash often produces fine lamination in beach sands that is highlighted by contrasting hydraulically sorted layers of light quartz and dark, denser mafic minerals.

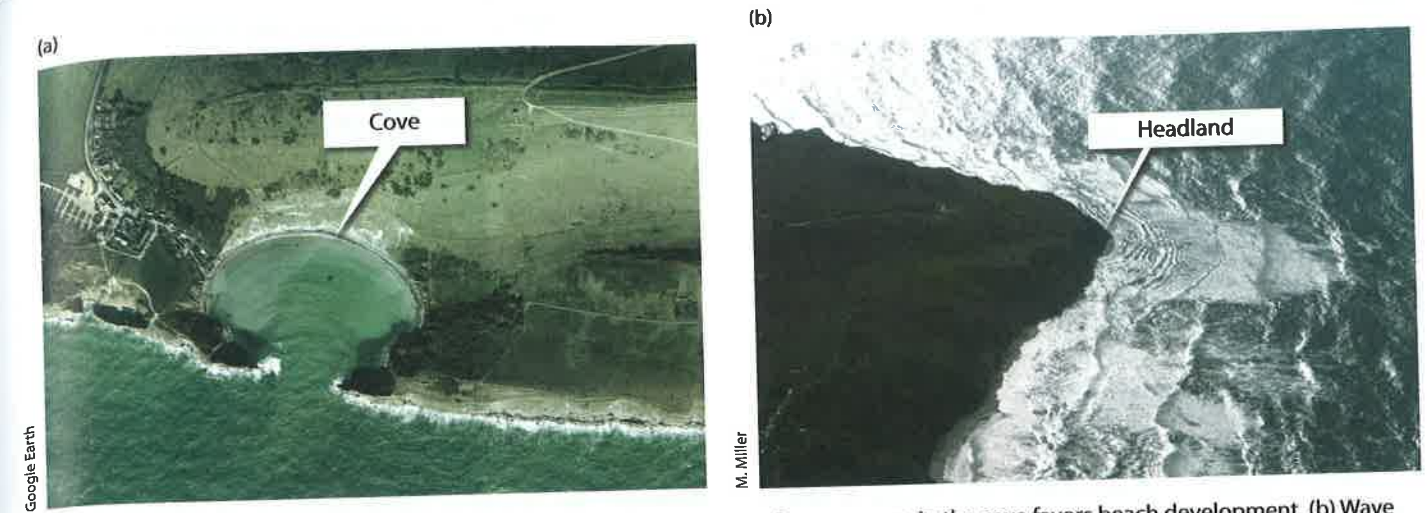
As waves shoal and break, they erode material from the seabed. Fine-grained silt and clay are transported seaward in suspension; waves move coarser sand (and gravel on high-energy beaches) back and forth on the seabed, producing net shoreward transport of the coarsest clasts. Swash moves up the beach in the direction the wave was traveling. In contrast, backwash is pulled back into the sea by gravity and moves directly down the slope of the foreshore, perpendicular to the coastline. Because wind-driven waves typically approach the shore at an angle, a zigzag pattern of sediment transport during swash and backwash produces **littoral** or **longshore currents**, called **longshore drift**, that move sediment along the shoreline [**Figure 8.6**]. The direction of longshore currents, and thus that of longshore drift, is governed by the prevailing direction at which waves strike a coastline. Longshore currents produce net transport along a coastline, so material delivered to coastal environments moves parallel to shore, nourishing beaches until it is transported offshore (in submarine canyons) to deeper marine environments below the wave base.

Where waves approach land at an angle to the shoreline, the shoreward ends of the waves reach wave base and begin to shoal before the seaward ends of the waves. This produces an effect known as **wave refraction** that bends the wave crests in response to the near-bottom portion of a wave traveling progressively slower upon moving into shallower water [**Photograph 8.3**]. Wave refraction along irregular coastlines concentrates erosive energy on headlands that project out into the sea and promotes deposition in protected embayments. Promontories become the focus of erosion, and protected coves become the focus of sedimentation. Thus, wave energy acts to straighten coastlines over time.



When waves approach the shoreline at an angle, sediment transport on the beach has a fundamental asymmetry that results in net **longshore transport** of sediment. Sediment is transported onto shore by **swash** in the direction the wave was traveling when it struck the coast. **Backwash** travels downslope back into the water under the influence of gravity. The net result is that wave action moves material along the coast. If the sediment moves out into open water, the result is the growth and extension of a **spit.** If the sediment is moving along a beach and encounters an obstruction, such as a groin that projects out from a beach, longshore transport will be disrupted and deposition of material will occur on the up-current side of the obstruction.

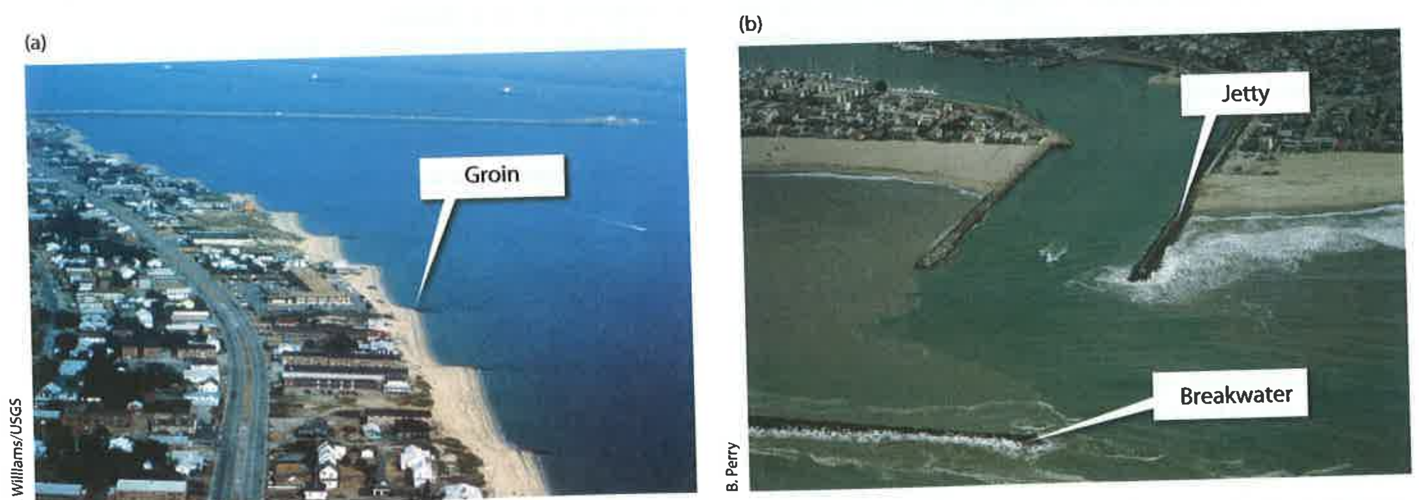
FIGURE 8.6 Longshore Drift. Longshore drift moves large amounts of sediment along beaches, due to a fundamental asymmetry of sediment transport by wind-driven waves and gravity.



PHOTOGRAPH 8.3 Wave Refraction. (a) Wave refraction into Lulworth Cove, on the southern coast of England. The dissipation of wave energy in the cove favors beach development. (b) Wave refraction around a headland in northern California favors erosion.

Engineered structures built in coastal environments to protect harbors and structures often have profound effects on longshore sediment transport by waves. **Groins** project from a shoreline into the sea [**Photograph 8.4a**]. Because they intercept sediment moving along the coast, sediment accumulates on the upcurrent sides of groins while longshore transport continues to move sediment away from their downcurrent sides. **Jetties** (barriers that project farther out into the ocean than groins) produce a similar effect, leading to a zone of sediment accretion on their upcurrent side and a zone of erosion on their sediment-starved, downcurrent side [**Photograph 8.4b**]. Breakwaters that armor a shoreline or deflect longshore currents also change local patterns of erosion and deposition. Progressive beach erosion because of sediment starvation that results from shutting off or intercepting longshore transport is a major problem in many coastal areas.

Very large, impulse-generated waves known as **tsunamis** wreak havoc on coastlines and can episodically but significantly change coastal morphology. Tsunamis are extremely long-wavelength (typically 100 to 200 km), low-amplitude (< several m) waves that are generated by sudden shifting of the seafloor due to impulses such as earthquakes, submarine landslides, or undersea volcanic eruptions. A tsunami can travel across an ocean and produce geomorphic effects on coastlines far from its source. With wavelengths that exceed the depth of the ocean, tsunamis behave as shallow water or translational waves. Even in water depths of kilometers, they can travel at remarkable velocities (>500 km/hr) and catastrophically inundate



PHOTOGRAPH 8.4 Coastal Engineering Structures. (a) Series of groins interrupt the natural transport of sand by longshore currents at Norfolk, Virginia. Can you determine which way the sand is moving? (b) Channel Islands Harbor breakwater and jetties, southern California.

coastal areas with little to no warning. Because tsunamis are often less than a meter high in open water, they pass unnoticed at sea. Arriving at a coastline, they **shoal** and can grow to produce waves more than 20 meters tall. Some tsunamis arrive as a large wave trough that produces what appears to be an anomalously low tide before the wave crest sweeps inland at devastating speeds (>10 m/s). Coastal configuration and the direction from which a tsunami strikes the coast affect the geometry of the waves and the pattern of coastal inundation at particular locations along a coastline. Like other types of waves, tsunamis interact with the coastal geometry as they approach land and produce wave-refraction and shoaling effects similar to those that result from normal wave action, only at a much larger scale.

Another source of anomalously high waves is **storm surge,** which is produced by strong storm winds that push water shoreward into the coast, in combination with a bulge in the sea surface under the area of lowest atmospheric pressure. Storm surges can raise local sea level 2 to 5 m during hurricanes and push water far inland in low-relief terrain. The additional temporary rise in local sea level because of storm surge amplifies the geomorphic impacts of coastal storms. For example, a storm surge of up to ~8 m during Hurricane Katrina was a key factor in the catastrophic levee breaks that flooded the city of New Orleans in 2005. Along the east coast of North America, storm surges cause major coastal erosion during long-lived nor'easters (very large extratropical storms) because the unusually high water continues over several tidal cycles.

Coastal Processes and Landforms

Coastal landforms reflect erosional and depositional processes as well as the grain size and volume of sediment available for transport in different environments. High-relief, rocky erosional shorelines typically develop on tectonically active continental margins, where erosion of bedrock by wave action creates most coastal landforms, and beaches are small and isolated. Low-relief and drowned, depositional shorelines, in contrast, typically develop on passive margins adjacent to extensive coastal plains composed of unconsolidated, clastic sediment derived from the continental interior. Coastal landforms in such settings typically consist of depositional features produced by longshore transport and deposition of sand by wave action. Beaches are the primary coastal landforms most people know and love, but there are many other distinctive landforms along sandy and rocky coastlines, including tidal flats, estuaries, and deltas **[Figure 8.7]**.

There are large differences in rates of coastal retreat and advance between sandy and rocky coasts due to differences in sediment supply, wave action, and resistance to erosion. Sandy coasts can advance if supplied with enough sediment. A falling sea level will also result in coastal advance (forming emergent coastlines), whereas a rising sea level pushes coastal processes and landforms inland (forming submergent coastlines). On eroding shorelines, coastal retreat may be very episodic; a coast retreating at an average rate of a meter per century may retreat 10 meters during a single large storm. Consequently, estimating long-term average rates of coastal retreat is best done using techniques that integrate over decades to millennia, such as air photo analysis and the dating of shoreline platforms.

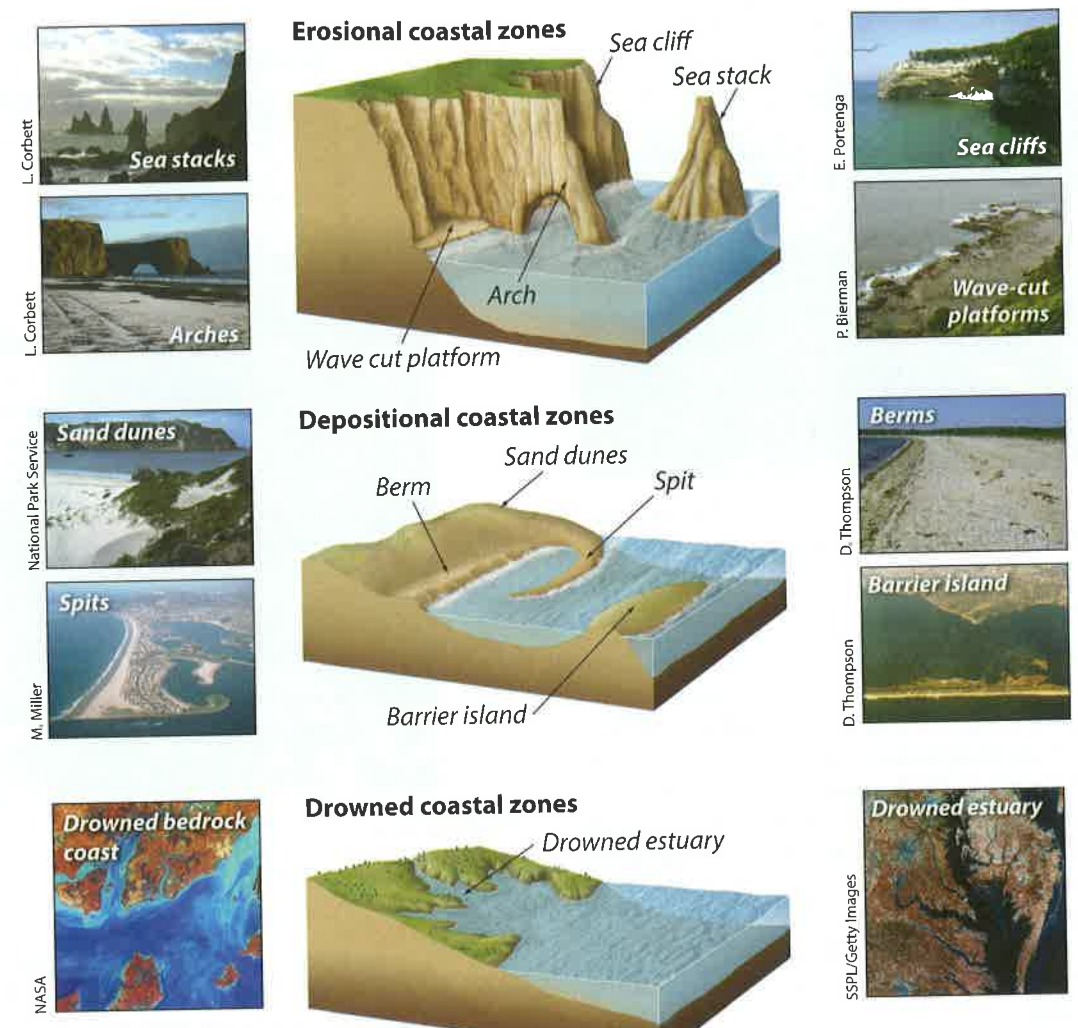
Rocky Coasts

Sea cliffs that are produced by landward retreat of bedrock slopes undercut by wave action are the most common and striking feature of rocky coasts **[Photograph 8.5]** as well as coasts where well-consolidated glacial deposits dominate the shoreline, such as those of eastern England and parts of Alaska. Active sea cliffs rise steeply from the shoreline at a sharp angle, the steepness of which is maintained by wave erosion at their bases. Inactive sea cliffs that have been raised above the zone of wave attack by tectonic uplift or isostatic rebound typically have a smoothly curving cliff base that has been shaped by subaerial weathering, erosion, and mass wasting. In places where sea-cliff bedrock is resistant, a **wave-cut notch** may develop at the base of the cliff **[Photograph 8.6]**. Relict wave-cut notches at the bases of inactive sea cliffs record former shoreline positions on tectonically active, uplifting coasts.

Landward sea-cliff retreat occurs where wave erosion, undercutting the cliff base, leads to mass failure of the cliff face. If wave action and longshore currents are sufficient to remove the debris, the process repeats and drives the cliff landward. The erosional resistance of the rock or sediment forming the sea cliff, the weathering processes acting to weaken the cliff face, and the energy and height of the waves striking the cliff base together determine the pace of cliff retreat. Sea-cliff erosion rates of several meters per year can occur in unconsolidated material, but rates as low as a millimeter per year are more typical in erosion-resistant rocks like granite.

Modern sea-cliff retreat is driven, in general, by an ongoing response to postglacial rise in global sea level. An additional 0.5 to 1.5 m of sea-level rise that is projected to result from melting of polar ice over the twenty-first century will cause a general acceleration of sea-cliff retreat, but the effect will vary locally depending on lithology, shoreline relief, sediment supply, beach slope, wave energy, and tidal range.

Cliff retreat on rocky coasts also produces a **wave-cut platform** that is beveled off just below the high-tide level (see Photograph 8.6). Wave-cut platforms typically slope seaward at a gentle angle of no more than several degrees and serve as breakwaters that dissipate wave energy and slow sea-cliff erosion. Wave-cut platforms are maintained by a number of processes, including erosion from wave action, the abrasive effect of suspended and bed sediment on the bedrock, and waves sweeping away the weathering products that result from mechanical and chemical disintegration of coastal bedrock that has been exposed



Rocky landforms are common along the erosional coastal zones that typify active, collisional continental margins. Landforms composed of unconsolidated sediment are common along depositional coastal zones typical of trailing-edge, or passive continental margins. Where sea level has risen, relict landscapes are drowned and may not yet have been altered beyond recognition by contemporary coastal processes. Examples of such drowned coastal zones include the coast of Maine and Chesapeake Bay in the United States and Marlborough Sound in New Zealand.

FIGURE 8.7 Erosional and Depositional Coastal Zones. Rocky shorelines and shorelines composed of unconsolidated materials have very different characteristic landforms. Rocky shorelines are dominated by bedrock and as a whole are resistant to erosion, changing slowly over time. Unconsolidated shorelines can change rapidly.

to repeated wetting and drying in the surf zone. On actively uplifting coasts, wave-cut platforms that become elevated above the surf zone provide long-term records of coastal uplift in the form of **marine terraces,** flat-lying features that indicate previous sea level stands.

Sea stacks, caves, and arches are prominent erosional features of irregular rocky coastlines. **Sea stacks** form where wave refraction concentrates wave attack around a headland and erodes through a narrow promontory to produce a small island or pillar that is isolated from the mainland at high tide **[Photograph 8.7]**. **Sea caves** form through preferential erosion of fractured or less resistant zones of rock exposed in sea cliffs. When a sea cave grows to extend completely through a rock promontory, it forms a **sea arch. Pocket beaches** are those restricted to small inlets and embayments nestled along rocky coasts between sea cliffs and such striking features as arches and sea stacks **[Photograph 8.8]**.

H. Stanchev

PHOTOGRAPH 8.5 Sea Cliff. This limestone sea cliff (70 m high) is on the coast of the north Bulgarian Black Sea, at Cape Kaliakra.

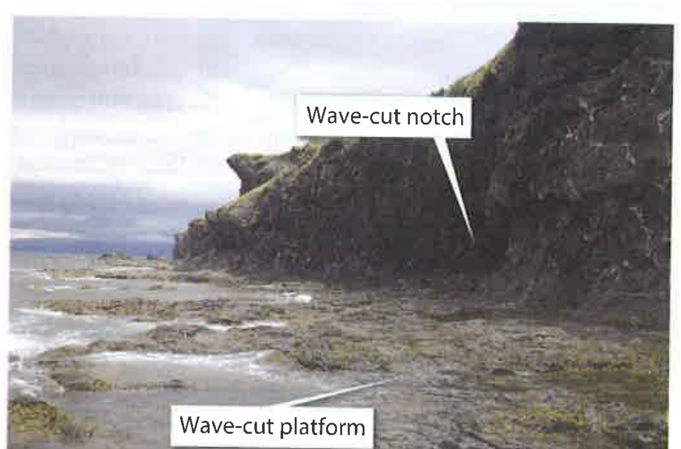
Beaches and Bars

Beaches are deposits of sand, gravel, or cobbles that form along shorelines directly affected by wave action. Local variations in the grain size and volume of sediment impart variability to beach morphology. Some beaches, such as those on Puget Sound in Washington State or along Cape Cod in Massachusetts, have adjacent stretches of sand, gravel, and cobble beach that result from local differences in the grain size and erosional resistance of glacial sediments exposed in active sea cliffs. Beaches reflect the interplay of sediment transport and sediment supply over a variety of timescales that range from individual storm events to millennia.

Beaches consist of several distinct, shore-parallel zones between mean lower low water (the average elevation of the lower of the two low-water heights of each tidal day) and coastal dunes, a sea cliff, or permanent vegetation [Figure 8.8]. The **offshore zone** extends seaward from the breakers, and a beach can be divided into the **inshore zone,** which consists of the breaker (surf) zone, extending from the base of the swash zone on the beach face to the beach trough and longshore bar; and the **foreshore,** the seaward-sloping beach face exposed at low tide in which **swash** rushes up and **backwash** runs down the beach face. The relatively flat **backshore** is often separated from the beach face by a distinct ridge, or **berm.** On a broad coast lacking a sea cliff, winds may deposit **dunes** inland of the berm (see Chapter 10).

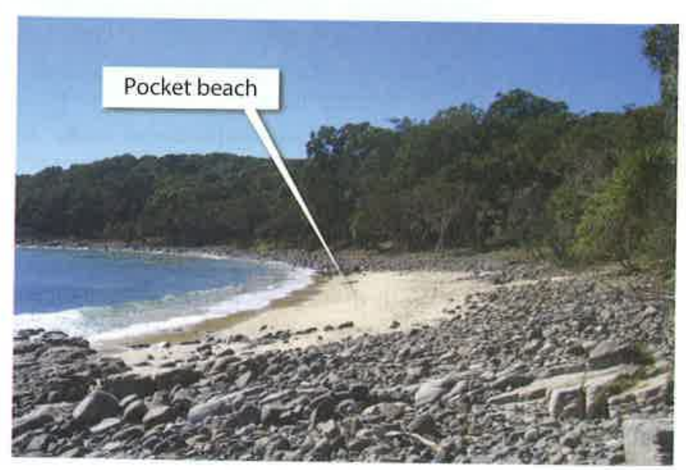
M. Miller

PHOTOGRAPH 8.7 Sea Stack. Sea stack at Myers Beach on the Oregon coast.



D. Thompson

PHOTOGRAPH 8.6 Wave-cut Notch and Platform. This wave-cut notch and wave-cut platform are in Gros Morne National Park, Newfoundland.



P. Bierman

PHOTOGRAPH 8.8 Pocket Beach. Sandy pocket beach between rocky outcrops in Noosa, northeastern Australia.

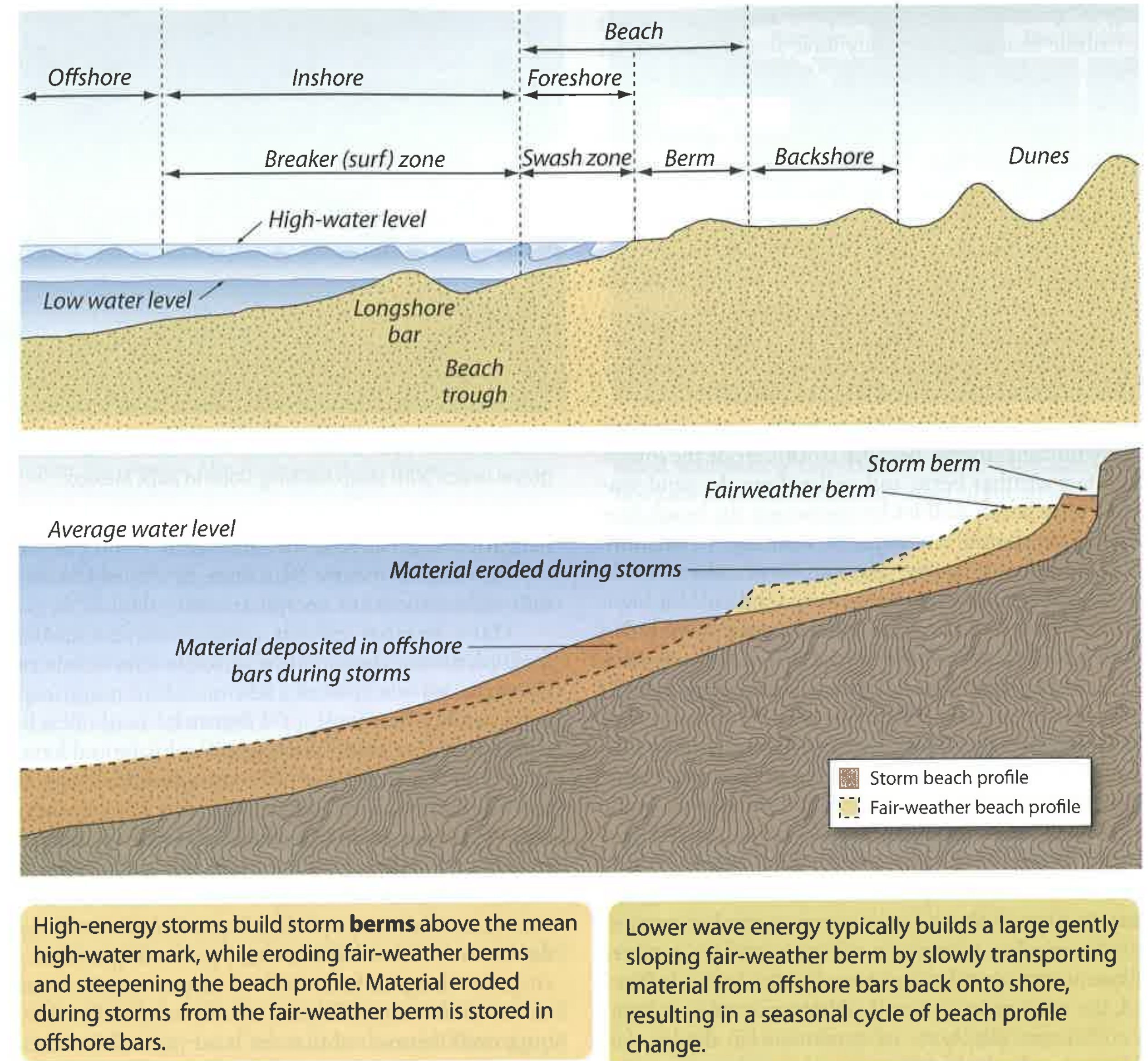


FIGURE 8.8 Beach in Cross Section (Storm Versus Fair-Weather Profiles). Beaches change their cross-sectional shape with the seasons as fair-weather waves move sediment shoreward onto berms and then higher-energy storm waves move sediment offshore into bars.

Beaches are dynamic. They are made up of readily transportable granular debris—sand, gravel, or cobbles—subject to movement by regular wave attack, longshore currents, and tidal flows. The fundamental asymmetry in sediment transport during swash and backwash in fair-weather conditions builds up beach berms over time. Waves rush onshore and come to a stop before flow reverses and the water runs back offshore. Because some seawater infiltrates into the permeable beach surface when swash flows up the **foreshore,** backwash has less transport capacity than swash. Less sediment flows back down the beach surface than moves up the beach. During rising and falling cycles of the tides, arriving waves build the beach up to create the berm. The elevation of the berm surface depends on how far swash moves up the beach face at high tide, a distance that scales with the energy (height) of incoming waves.

The slope of the beach face depends on the wave energy striking the coast as well as on the grain size and supply of sediment. Specifically, it is the interaction of swash and backwash processes that governs beach slope. The greater the difference between the ability of swash and backwash to transport sediment, the steeper the beach profile that develops. Backwash rapidly drains down into beaches composed of coarse pebbles or cobbles such that little of the material carried up the beach face is transported back down, building a steep beach face. In contrast, fine-grained beaches stay saturated in the short interval between waves due to their low permeability, limiting backwash infiltration into the beach face and resulting in greater seaward transport and flatter beach slopes. Cobble beaches commonly have slopes that exceed 20°, pebbly beaches have slopes of 10° to 15°, and sandy beaches have slopes of 3° or less.

Wave energy in part controls the flux of sediment on and off beaches and thus beach steepness. High-energy wave attack tends to increase the slope of a beach by eroding the beach face and moving sediment offshore. Thus, the beach slope changes seasonally in regions subject to erosion by large breaking waves accompanying seasonal storminess. Lower-energy waves that strike the coast as swell result in greater net transport of sediment onto the beach face. During calm periods between storms, low-energy waves move beach-forming material shoreward.

Waves build up a **fair-weather berm** during periods between significant storms. Because storms erode the lower-elevation, fair-weather berm and redistribute the sand seaward, they alter beach profiles by steepening the beach face and moving substantial amounts of sediment to offshore bars. Wave attack during very large storms can even erode a notch in the berm, creating temporary sand cliffs along a beach profile. Larger waves from higher-energy storms also may build a higher **storm berm** at the upper limit of wave action, just seaward of dunes. Between storms, beach faces gradually recover and flatten as fair-weather waves move sediment landward again and wave-cut smooth notches.

Storm patterns affect the degree to which beach morphology exhibits regular seasonal change. Years without major storms produce little seasonal change in beach profiles, and anomalous storm events dramatically alter beach profiles regardless of the seasonal pattern along a coastline. In the southeastern United States and eastern Australia, tropical depressions including hurricanes and typhoons are summer and fall events that periodically reform beach profiles. In New England, the most geomorphically effective storms are long-lasting nor'easters, which are most common in the late fall and early spring. In the northwestern United States, winters are stormy and summers are calm. At far northern latitudes, sea ice precludes near-shore winter waves, and there is little seasonal change in beach morphology.

Submerged, near-shore bars commonly form beneath breakers, and the interplay of wave action and sediment transport creates a feedback loop in which breaking waves build bars that, in turn, promote shoaling and thus more breakers. Consequently, bars commonly form at the seaward limit of the surf zone, where waves begin to break, and an associated trough grows between the bar and the beach. The complexity of the relationship between incoming waves and near-shore sediment transport is expressed in the variety of bar types that result from specific coastal currents and configurations. Multiple, shore-parallel bars may reflect the positions of breaking waves of different sizes, breaker positions during high and low tides, or development of breakers that form and break up again and again before reaching shore on low-gradient beaches. Deep bars that form beneath unusually large storm waves sometimes remain stable for long periods because they lie

M. Miller

PHOTOGRAPH 8.9 Beach Cusps. These beach cusps are on a gravel beach with steep backing slope in Baja, Mexico.

beneath the wave base for fair weather and even average storm waves. Offshore bars serve as natural breakwaters that reduce the wave energy affecting the beach.

Many beaches exhibit crescent-shaped and cuspate forms with axes perpendicular to the coast that range in size from wavelengths of a few meters to major capes and embayments hundreds of kilometers long. **Beach cusps** commonly develop on the upper beach face and have wavelengths of less than 30 m [Photograph 8.9]. Cusps form in any size of beach sediment, from fine sand to large cobbles. Sediment in the seaward projections, or horns, of cusps is typically coarser than in the intervening embayments. Although their origin remains debated, there is a consensus that cusp spacing depends on wave height and direction and that higher waves generally produce greater cusp wavelengths. Some believe cusps result from standing waves that set up offshore; others believe they are a self-organized feature similar to river meanders. Cusps are typically destroyed by large storms and can reform within a day after storms. In some regions, like the Atlantic coast of the southeastern United States, the periodic spacing of very large shoreline crenulations, called **capes,** may reflect the spacing of rivers or variations in ocean currents.

Spits, Tidal Deltas, and Barrier Islands

Constructional coastal features are formed by sediment deposition in coastal environments. **Spits** are shore-parallel sediment deposits that are connected to the mainland at one end. Spits are found on many coastlines, including on large inland lakes. Longshore transport builds spits by delivering sediment to the end, or tip, of the spit where the sand is deposited, allowing the spit to build out into deeper water. A spit may grow across an existing embayment or extend from a headland to enclose a lagoon that fills with fine sediment once it is protected from wave action. Where tidal flux cuts through spits, it creates **inlets** through which tidal flow moves in and out of the embayment. Tidal currents often reshape the ends of spits into curved

M. Miller

PHOTOGRAPH 8.10 Spit. Dungeness Spit, Strait of Juan de Fuca, Washington; the Olympic Mountains are in the distance.

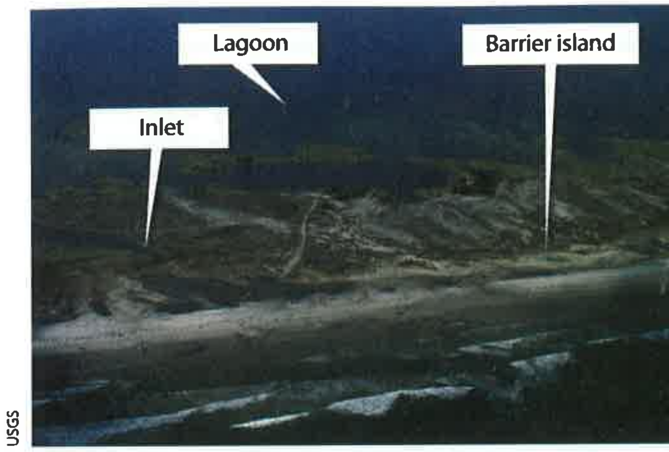
forms, or **hooks** [Photograph 8.10]. Storm surges may fill in existing tidal inlets with sediment and cut new passages through spits.

Flood-tide deltas form where sand carried by longshore currents reaches a tidal inlet and the incoming tide pushes it landward through the inlet. The flooding (rising) tide carries sand into the lagoon, where it rapidly settles out in the sheltered waters where the waves are weak. Sand transported out of a lagoon by an ebbing (falling) tide sometimes forms an **ebb-tide delta** on the seaward side of a barrier island [Photograph 8.11]. Ebb-tide deltas are generally small compared to flood-tide deltas because they are reworked by waves in the open marine environment and because the outgoing tide usually has a lower sediment-transport capacity than the incoming tide.

Barrier islands are similar to spits but are disconnected from the mainland by tidal inlets at both ends [Photograph 8.12]. They are typically shore-parallel islands located a short distance from the mainland across a lagoon or bay.

I. Buynevich

PHOTOGRAPH 8.11 Ebb-tide Delta. The Midway Inlet in South Carolina is a tidal inlet with an ebb-tide delta.



USGS

PHOTOGRAPH 8.12 Barrier Island. Ocracoke Island is a barrier island, southwest of Cape Hatteras. This photograph, taken after Hurricane Isabel (2003) passed within 30 km, shows sand washed over the island into the lagoon.

Barrier islands are typically several hundred meters to several kilometers across and from 10 to up to 100 km long. Windblown sand dunes on the landward side often form the spine of a barrier island, and tidal flats and marshes fringe the lagoon. The elevation of a barrier island depends on the available sand supply and the strength of dune-forming winds. Most barrier islands, like those that line the coast of South Carolina, are less than 10 m in height. Low barrier islands experience **overwash** by storm surges, which can carve new inlets and deliver sand to interior and back-barrier locations. Inlets migrate in the direction of longshore transport by accumulation of sediment on their upcurrent side and erosion of the downcurrent side of the inlet. Conditions on low-energy to moderate-energy coasts with limited tidal ranges favor development and maintenance of barrier islands.

Barrier islands form by three primary processes: spit elongation, bar submergence, and bar emergence. When a spit becomes long enough to slow transport of water between the ocean and lagoon, a new inlet forms as water levels rise over a spit during storms and splits off the end of the spit into an island. Bar submergence creates barrier islands when an old dune or other topographic coastal high point becomes isolated as sea level rises. Bar emergence occurs either during falling sea level or when a strong storm creates a large offshore bar that becomes exposed as a low island after the storm surge subsides and winds build dunes that raise the bar farther above sea level. Wave refraction sometimes focuses deposition and builds **tombolos,** sand or gravel bars that are inundated at high tide but connect an island with the mainland or another island at low tide.

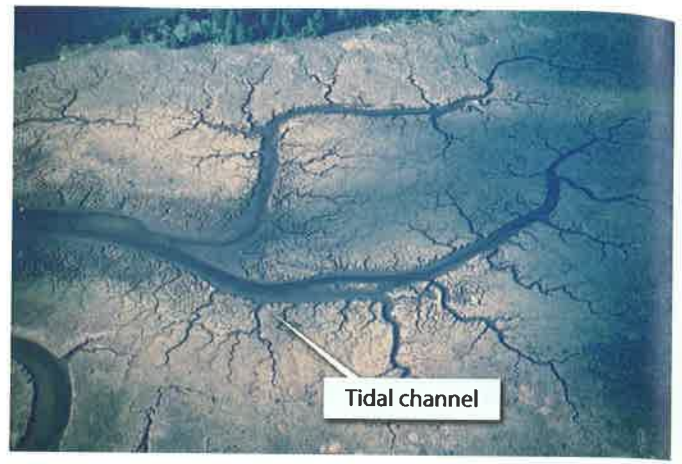
Coastal barriers migrate seaward or landward in response to changes in sea level, sediment supply, or coastal erosion. Landward migration of coastal barriers is common because of enhanced coastal erosion (due to wave action) driven by recent global sea-level rise. The main processes that move coastal barriers landward are wind transport to and through coastal dunes, washover during large storms, and tidal transport through inlets to lagoons and flood-tide deltas. Seaward growth of coastal barriers does occur close to large sediment sources. Sets of distinctive, prograding **beach ridges** typically accompany seaward movement of the shoreline in regions where rivers deliver abundant sediment to coastal environments or where uplift is occurring.

Offshore barrier islands are common on passive margins, with little tectonic uplift or subsidence—for example, the mid-Atlantic and Gulf of Mexico coasts of the United States. Barrier islands are much less common on active margins, but they can occur near substantial sediment inputs such as river mouths.

Lagoons, Tidal Flats, and Marshes

Lagoons are partially enclosed coastal water bodies on the landward side of spits, barrier islands, and reefs [**Photograph 8.13**]. The quiet waters of lagoons accumulate sediment that gradually fills in the lagoon embayment unless it is subsiding tectonically. Fill material includes sediment delivered from fluvial sources, marine sediment washed over spits and barrier islands, and biogenic material from high biologic productivity due to quiet water and high nutrient loads.

Tidal flats [**Photograph 8.14**] are depositional areas within the intertidal zone, the landscape between mean low-tide and mean high-tide levels. In coastal areas that are protected from direct wave attack, fine-grained mud (clay) transported as suspended load in the incoming tidal flux accumulates on a flat surface just below high-tide level. Coarser sand, transported as bedload, accumulates lower on tidal flats, closer to the mean low-tide level, or in intertidal creeks. If sea level is steady, then the rate of sediment supply controls the speed at which tidal flats aggrade or erode. If sea level falls, then tidal flats are more likely to transition from intertidal to terrestrial areas. If sea level rises, tidal flats require a substantial supply of sediment in order to aggrade rapidly enough to keep up with the rising water.



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PHOTOGRAPH 8.14 Tidal Flat. Aerial view of branching tidal channels on a tidal flat in Coos Bay, Oregon. Trees in the distance provide scale.

Tidal flats are best developed in macrotidal environments with large tidal ranges because the sediments that would otherwise form beaches, spits, and barriers are instead distributed throughout the coastal zone by waves attacking at a variety of different elevations. Waves die out as they cross broad tidal flats, and tidal currents come and go slowly at their outer, landward reaches, creating a low-energy zone where mud accumulates during slack water at high tides. Salt marshes form in the intertidal zone where salt-tolerant vegetation promotes sediment deposition by introducing roughness that dissipates wave energy and reduces tidal current velocity. Marshes are made up of 5 to 25 percent organic matter; thus, biologic productivity is important as a marsh sediment source, including below-ground productivity. In tropical regions, extensive mangrove vegetation protects coastal environments from erosion and promotes marsh growth by dissipating wave energy [**Photograph 8.15**].

Recent research suggests that many coastal marshes are recent additions to the landscape, the result of large volumes of sediment filling shallow areas of estuaries when upland deforestation increased sediment yields. Examples of sediment infilled tidal flats and emergent salt marshes that expanded greatly during the past several hundred years in estuaries include San Francisco Bay and the coast north of Boston, Massachusetts. Although afforestation and better land management practices have reduced sediment yields over the past century, the marshes remain because the vegetation changes water-flow dynamics, efficiently traps sediment, and produces organic matter that accumulates in the marsh. Even with sea level rising at >1 mm/yr, these relatively young marshes are so efficient at accreting sediment that they can keep from being submerged.

Blue Heron Realty

PHOTOGRAPH 8.13 Lagoon. The body of water in the right side of the image is a lagoon sheltered by a well-vegetated barrier island on the eastern shore of Virginia.

P. Bierman

PHOTOGRAPH 8.15 Mangrove Coast. Mangroves line a beach at the base of a steep coastal slope in Queensland, Australia, near Cape Tribulation.

Estuaries

Estuaries form where rivers discharge into the ocean in topographically confined settings that facilitate mixing of fresh and salt water and promote sedimentation [**Figure 8.9**].

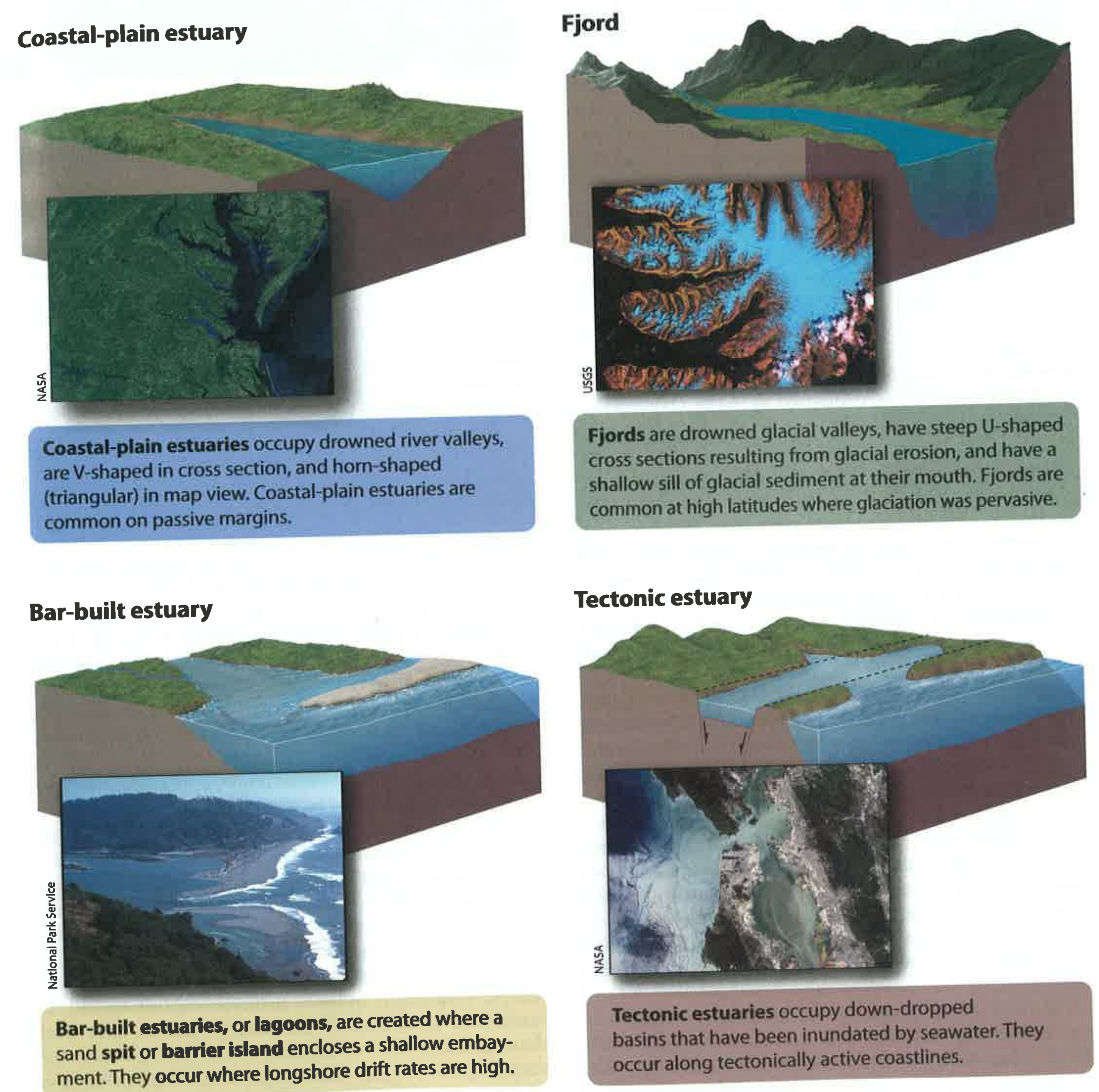


FIGURE 8.9 Estuary Types. Estuaries vary greatly in their plan form morphology, cross-sectional shape, and origin.

Sand traveling as fluvial bedload constitutes less than 10 percent of the riverine sediment that reaches estuaries, and most of that sand is deposited at sea level near the head (upriver end) of estuaries. The suspended load that a river delivers to an estuary, mainly silt and clay, is typically dispersed throughout the estuary and is either deposited or carried out into the ocean. Sediment that is not trapped in estuaries forms the riverine portion of the sediment supply for coastal beaches.

The dynamics of saltwater–freshwater mixing promote estuarine sedimentation. In estuaries, sediment-laden river water flows out over denser salt water and creates a zone of turbid surface water where mixing occurs. The boundary between the fresh water that is moving seaward at the surface and the underlying salt water is called the **halocline.** When silt and clay particles enter this mixing zone, they aggregate to form larger particles. In fresh water, the negative electrical charges on mineral surfaces tend to repel each other. But in brackish water, negatively charged clay mineral flakes and organic matter cluster around cations such as Na^+ and form loose clumps called flocs, a process called **flocculation.** As flocs sink and settle, they become trapped near the river mouth at the head of an estuary by the landward bottom flow of denser salt water. This process makes estuaries excellent sediment traps and leads to the formation of extensive tidal marshes and deltas in coastal environments.

A floc is typically composed of platy clay particles joined end to face in a "house of cards" structure that contains a lot of water. Consequently, estuarine sediments tend to consolidate and dewater as the weight of additional sediment squeezes water out of flocculated clays, leading to gradual subsidence over time. If uplift of the sediments and leaching of the cations occurs before consolidation (and the resultant strengthening) is complete, these estuarine sediments can liquefy quickly if shaken, for example, by an earthquake or an excavation. Such liquefaction caused substantial damage in the 1964 Alaska earthquake.

Different types of estuaries have different cross-sectional geometries and map patterns (see Figure 8.9). **Bar-built estuaries** are shallow features enclosed behind spits or barrier islands. **Tectonic estuaries** are fault-bounded coastal basins that have become flooded by seawater, such as San Francisco Bay. **Coastal-plain estuaries** are extensive embayments that formed as postglacial sea-level rise inundated valleys that were eroded by rivers when sea level was lower. Chesapeake Bay and Delaware Bay are coastal-plain estuaries. Because they were originally cut by rivers, coastal-plain estuaries are usually V-shaped in cross section and have a triangular map pattern that points upstream.

A **fjord** (from Norwegian for "passage") is an estuary in a drowned glacial valley where ice once eroded the valley floor **[Photograph 8.16]**. Fjords are generally deep and U-shaped in cross section and near their mouths have

P. Bierman

PHOTOGRAPH 8.16 Fjord. Fjord near Upernavik, central western Greenland, with steep walls and low-relief highlands. Cosmogenic exposure dating indicates that ice last occupied the fjord about 11,000 years ago.

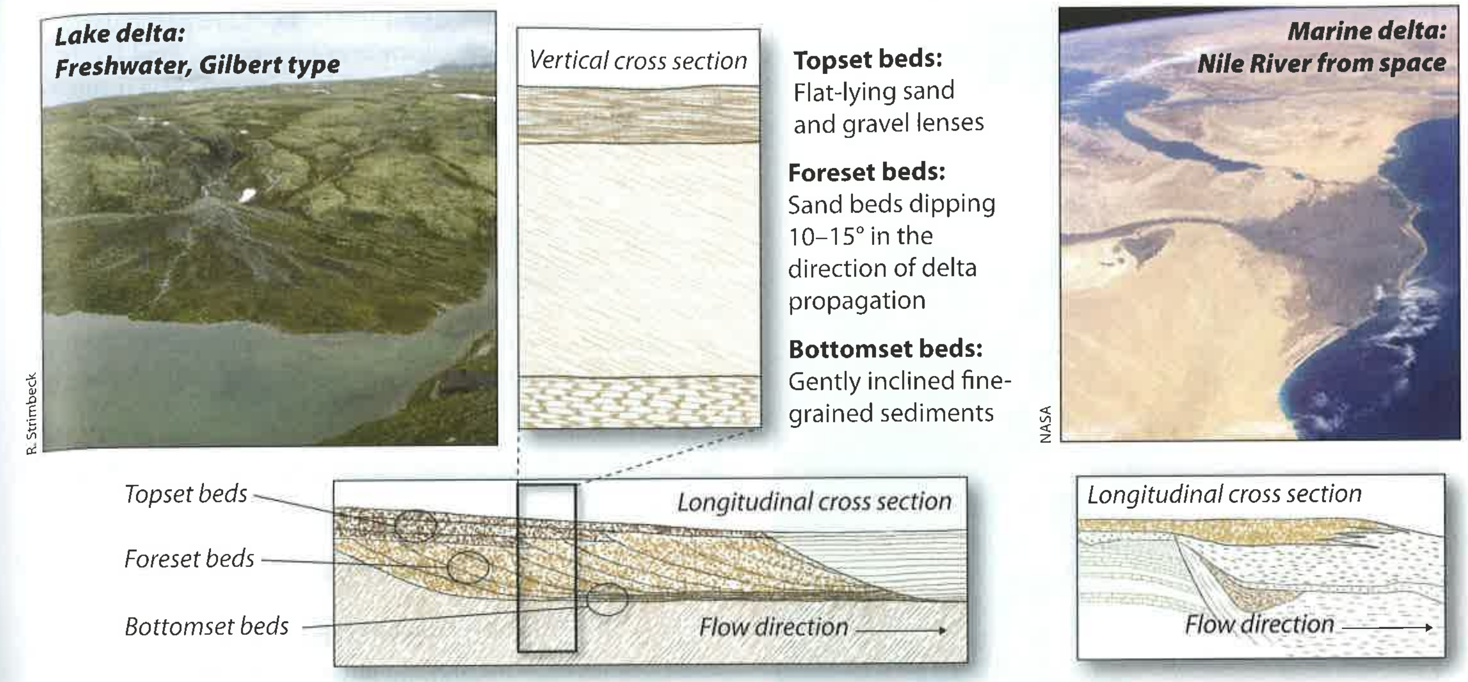
shallow sills that separate them from the marine environment. Most sills are submerged deposits of glacial sediment. Fjords occur along high-latitude coasts, and especially on the windward coasts of continents, where mountains catch snow and build glaciers from the ocean-derived moisture. They are thought to be the result of channelized ice flow excavating areas of weaker rock or preexisting lowland topography.

Deltas

Deltas develop where rivers entering bodies of water supply more sediment than longshore drift and/or tidal currents remove. The result is a wedge of primarily terrestrial material built out into the marine (or lacustrine) environment. Deltas are best developed where rivers with high sediment loads enter sheltered coastal settings like Puget Sound, the Gulf of Mexico, or the Black Sea, and in calm areas behind barrier islands or reefs with small tidal ranges.

Deltaic deposits result when sediment conveyed by rivers enters standing water and river flow diverges because it is no longer constrained by channel banks. The Mississippi River, for example, approaches sea level at the landward side of its delta, but both built and natural levees contain the flow across the delta. Thus, the Mississippi River cannot deposit its sediment load until it finally enters the sea and flow diverges at the delta margin.

In many delta systems, deposition of the river's sediment load leads to development of **distributary channels** that bifurcate downstream. As channels diverge and thus get smaller downstream, sediment-transport capacity drops. This results in more sediment deposition and the formation of in-channel and channel-mouth bars. Such bars increase flow divergence, which reduces transport capacity and triggers further sedimentation. The locus of deltaic sedimentation shifts as coastal depressions become



Gilbert-type, freshwater deltas form where rivers deposit their load of sediment upon entering a body of water. Fine-grained sediment (silt and clay) define relatively flat-lying **bottomset beds** representing sedimentation of suspended material on the distal end of the delta. **Foreset beds** composed of gently dipping beds of sand and gravel represent deposition of bedload material off the submerged face of the delta front. Coarser, flat-lying **topset beds** composed of gravel, cobbles, and boulders represent fluvial deposition at water level. A delta advancing into a body of water will deposit a characteristic sequence of beds shown in the longitudinal cross section.

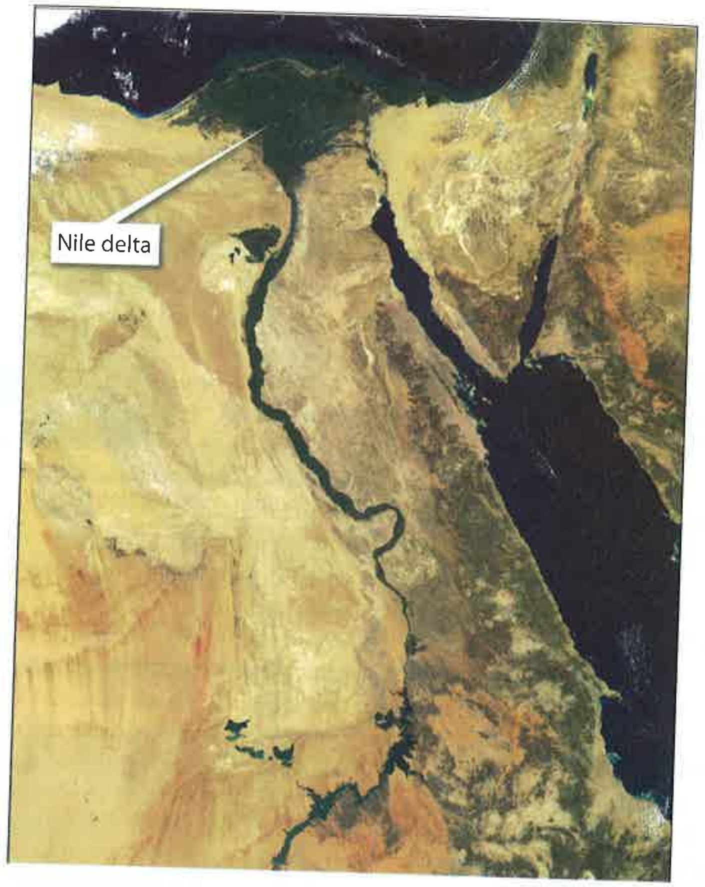
Marine deltas have different sedimentary architecture than freshwater deltas. Grain size in marine deltas depends on the caliber of material supplied by the drainage, wind strength, tides, and the size of the delta. There are thin, discontinuous lenses of sand and coarser material that represent paleochannels that once flowed over the delta surface.

FIGURE 8.10 Delta Anatomy. Deltas, formed as rivers deposit sediment into standing water, have a predictable internal architecture.

filled and streams switch to other, lower locations forming new depositional lobes. For example, the active portion of the Mississippi River delta has switched location many times over the past several thousand years.

The simplest deltas (known as **Gilbert-type deltas**) grow or **prograde** by deposition of thick sediment layers at river mouths, where fluvial sediments settle out and form distinctive topset, foreset, and bottomset deposits **[Figure 8.10]**. These Gilbert-type deltas are most common in smaller, freshwater lakes where the complexities introduced by tides, significant longshore drift, fast-moving currents, and the flocculation is minimal. **Topset beds** are made of sediment that is deposited near the surface of a delta at just above the average water level. **Foreset beds** are inclined lakeward (or seaward), and they form as sediment spills over the edge of the delta. Lateral accretion of foreset beds is the primary process by which a delta advances out into a body of standing water. **Bottomset beds** are horizontal or subhorizontal beds deposited in deep water. They are composed of fine-grained material and are the leading edge of the delta. In studies of now-vanished lakes, including those formed during glaciations and wet periods, the elevation of the topset-foreset contact is used to define the paleolake level.

More complex, marine deltas result when rivers empty into the ocean and deliver sediment at a rate greater than it can be transported away by coastal processes; they have different sedimentary architecture than lake deltas. The overall shape of a marine delta depends on the relative influences of fluvial sediment inputs and stream current strength, wave energy, and tidal energy. Some marine deltas, like Egypt's famous Nile River delta, form an upstream-pointing triangle that looks like the Greek letter delta (Δ) **[Photograph 8.17]**. Other deltas have very different morphologies.



NASA

PHOTOGRAPH 8.17 **Nile River Delta.** The Nile delta, bright green from vegetation, shows clearly in this remotely sensed image.

Marine deltas may be river-dominated, wave-dominated, or tide-dominated **[Figure 8.11]**. **River-dominated deltas** typically have a "bird's foot" morphology consisting of an elongated distributary mouth that protrudes at right angles from the coast (like the Mississippi River delta), where river delivery of sediment dominates delta form. River-dominated deltas typically form in areas with low wave energy, low tidal range, and high sediment supply that together allow fluvial processes to dominate. **Wave-dominated deltas** form on coasts where wave attack straightens the coastline, building extensive sandy spits and barrier islands. Wave-dominated deltas form in areas with high wave energy, unidirectional longshore transport, and steep offshore slopes. Longshore drift along wave-dominated deltas reworks the delta front into smooth, cuspate forms. **Tide-dominated deltas** have broad, seaward-flaring, fingerlike sand ridges and islands separating large distributary waterways (like the Ganges–Brahmaputra delta in Bangladesh or the Fly River delta in Papua New Guinea). They tend to form in areas with low wave energy, high tidal range, and little longshore transport. Strong tidal currents enlarge the seaward mouths of distributary channels and create large islands that are oriented orthogonally to the shoreline.

Deltas, marshes, estuaries, and lagoons are particularly sensitive to **subsidence**, which over time causes land levels to decrease unless more sediment is deposited. Much subsidence is the result of **compaction**, driven at least in part by the mass of sediment deposited later than and above the compacting layers. Marsh sediment is rich in organic matter, which both compresses under the weight of overlying sediments and decays over time, losing mass and volume. Parts of major deltas, such as the Mississippi River delta, are subsiding rapidly (6–8 mm/yr) because compaction is ongoing and no longer offset by sediment deposition. This sediment starvation results both from sediment bypassing the delta because the river channel is contained within flood-control levees and also because dams have reduced the sediment load of the Mississippi River by about 50 percent.

Coastal Rivers

Coastal rivers are subject to bidirectional flow and to tidally induced, water-level changes that influence channel morphology and dynamics. Deposition, particularly of suspended sediment, is enhanced in the zone of tidal influence in the lower reaches of coastal rivers. Tidal effects may extend many kilometers upstream in low-gradient river systems typical of passive margin settings. However, the tidal range changes over the year and through each month. The backwater effect of high tides and coincident storm surges can increase river flooding in coastal rivers, such as happened in South Carolina's coastal rivers during Hurricane Hugo in 1989, when the storm surge locally reached an extraordinary 6.1 m. Backwater effects of high tides may cause a rapid decrease in transport capacity near the upper limit of tidal influence. For example, historical maps and accounts indicate that enormous logjams occurred at the head of tidally influenced reaches on rivers around Puget Sound in the Pacific Northwest.

Marine Settings and Drivers

Seafloor processes and morphology typically reflect proximity to active geologic structures at divergent, convergent, and transform plate boundaries as well as proximity to continental and marine sources of sediment. In contrast to the continents, where topography commonly reflects ancient tectonic setting and erosional history, seafloor features are more likely to be the youthful products of recent and active tectonic processes and sedimentation. Portions of some continents are billions of years old, but new seafloor is continuously created at mid-ocean ridges and ancient oceanic lithosphere is consumed at subduction zones. The age of most of the world's seafloor can be measured in millions to tens of millions of years. In addition to the direct effect of plate tectonics

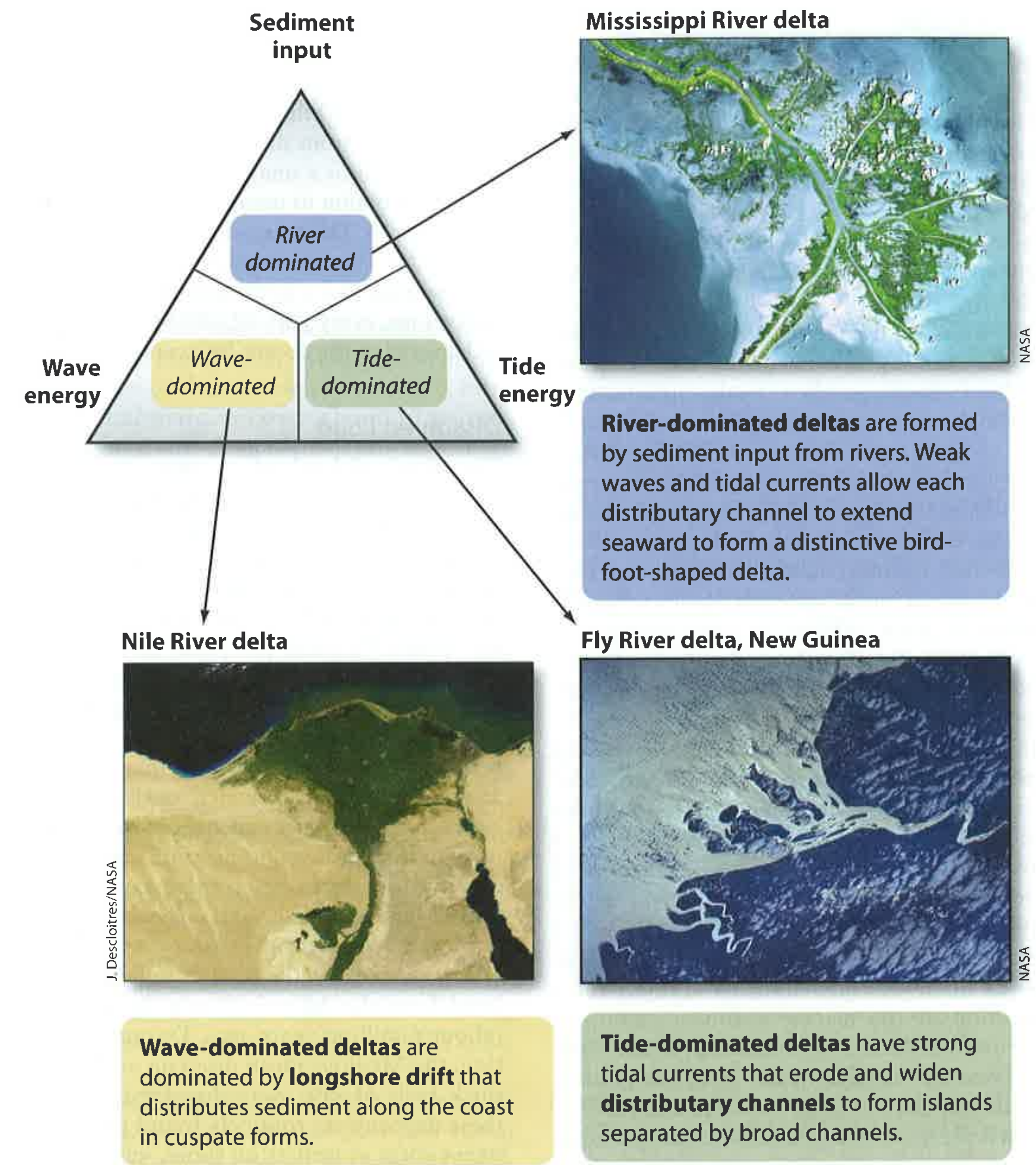


FIGURE 8.11 **Delta Process Triangle.** The shape of deltas depends strongly on the interaction between coastal and fluvial processes, specifically, mass fluxes as well as wave and tidal energy that affect sediment delivery and redistribution.

on the seafloor morphology, the movement of ocean currents, the delivery of sediment to ocean basins, and the precipitation of dissolved elements are geomorphically important processes.

Currents

Three primary types of currents influence marine topography and depositional environments in deep waters, areas below the wave base and thus protected from wave action. First, surface currents that are generated by friction between the atmosphere and sea surface under persistent wind belts govern the movement of water, heat, and sediment in the upper ~400 m of the ocean. Second, density and temperature gradients drive circulation of the deep ocean waters in what is known as the global conveyor belt (see Figure 1.5). Third, sediment-laden, gravity-driven density currents, called **turbidity currents**, are denser than average ocean water and thus flow down seafloor slopes in much the same manner as subaerial debris flows move down hillslopes. Due to their higher density and thus greater tractive force than clean-water flows, density currents can be highly erosive and can dramatically influence submarine topography.

Marine Sedimentation

Marine sediment consists of material derived from three different sources. **Lithogenic** or **terrigenous** sediments are produced by weathering and disintegration of rock on the continents and are delivered to marine environments by aeolian, fluvial, and glacial processes. **Authigenic** sediments are produced by inorganic precipitation of dissolved components out of supersaturated seawater. **Biogenic** sediments are generated by organic precipitation of dissolved components by organisms.

Terrigenous sediment dominates continental margins where rivers, the main agents of erosion and sediment transport on land, discharge their loads. Streamborne sediments that make it to estuarine and marine environments contain, on average, 90 percent silt and clay (mud) and 10 percent sand. Turbidity currents can carry terrestrial sediment hundreds of kilometers offshore and deposit this material onto the abyssal plains. Such deposits have been cored and used to understand the timing and thus recurrence intervals of such different geomorphically important events as great subduction zone earthquakes and the Lake Missoula glacial outburst floods that carved the Channeled Scablands of eastern Washington State (see Digging Deeper, Chapter 13).

Terrigenous particles are also introduced to marine environments by wind and glaciers. Windblown (aeolian) dust picked up from the continents settles out onto the sea surface far from land and can also be delivered to the oceans by mixing with rain water. Wind-deposited dust is a primary source of lithogenic sediment in the deepest parts of the ocean basins, far from continental margins, and can accumulate to form red clays that blanket the deep-sea floor. Glacial sediments account for a substantial portion of the marine sediment supply in many coastal environments at high latitudes, where flowing ice delivers material ranging in size from fine-grained glacial flour (silt and clay) to large boulders that raft out to sea frozen in icebergs. The cyclic occurrence of coarse, glacially derived sediment in marine cores allows one to infer the onset of glaciation as well as cyclicity in ice sheet behavior.

In tropical regions, where coral reefs are common, calcium carbonate sands derived from reef erosion can accumulate offshore. Such biogenic deposits are typically limited to areas with low terrestrial sediment input and relatively shallow water depths, as the carbonate minerals are not stable in deep, cold water and will dissolve.

Farther out to sea, sediment becomes less important in shaping marine landscapes. The morphology of the abyssal plain primarily reflects the effects of tectonism and volcanism during the creation of crust at the mid-ocean ridges. Over time, as the plate moves away from the ridge, these rocky, solid Earth landforms are draped by a thin blanket of authigenic sediment, marine sediment precipitated directly from seawater. Authigenic marine sediments dominate deep water deposition far from continents where biological productivity is low. In these locations, biologically produced calcium carbonate dissolves because it is soluble in deep, cold seawater. In addition, there are no large inputs of lithogenic sediment because the abyssal plains are far from the continents. Mixed with the authigenic sediment is a small but easily measurable extraterrestrial contribution to ocean sedimentation of interplanetary dust particles. On average, about 40,000 tons (0.04 megatons, or MT) of extraterrestrial dust fall to Earth every year. That sounds like a large amount until one considers that on average, every year, 12,600 MT of suspended sediment is delivered to the oceans by rivers draining the continents.

Dissolved Load

The dissolved load in seawater comes from rivers that carry the ionic products of chemical weathering on land, from volcanic eruptions at mid-ocean ridges, and from air fall and precipitation scavenging of the atmosphere. Some of the dissolved load remains in solution, producing and maintaining the saltiness of ocean water, which is on average about 35 parts per thousand (ppt) dissolved salts. Some of the dissolved load precipitates inorganically from seawater as authigenic chemical sediments. Other dissolved material is precipitated biologically when marine organisms produce mineral material like silica dioxide (SiO_2) and calcite ($CaCO_3$), which settle out as the biogenic oozes that cover vast expanses of the seafloor. Formation of evaporites from concentration and precipitation of the dissolved load can be an important sedimentation process in shallow marine and coastal settings. Tectonic and sea-level changes can isolate ocean basins, as happened when what is now the Mediterranean Sea was briefly isolated from the Atlantic Ocean during the late Miocene (about 6 million years ago). During this period of isolation, the Mediterranean dried up and evaporite deposits, thick beds of salt, were deposited in the basin. Today, these deposits are routinely found in Mediterranean sediment cores as well as on shore, where they are the basis for halite and gypsum extraction industries in Sicily. By the early Pliocene (about 5 million years ago), ocean water again flooded the Mediterranean.

Marine Landforms and Processes

Hidden beneath the waters of the world's oceans are a variety of submarine landscapes that rival in their variety and topography the terrestrial landscapes of the continents. A quick glance at a bathymetric map, on which contour lines represent depth below sea level, reveals an exciting and, until recently, little-known realm of sedimentation, mountain building, and volcanic activity. A variety of remote-sensing techniques—including side-scan sonar, depth profiling, ocean floor drilling, magnetic anomaly analysis, and deep-sea dives—has brought geomorphology to the ocean depths. Plate tectonics sets the location and character of submarine landscapes on the largest scale. Sedimentation and the erosive forces exerted by moving sediment and water continuously reshape the morphology of the ocean bottom.

Continental Margins

Continental margins have aprons of sediment draped over bedrock and extend out from the continents into the ocean basins. The crystalline rocks of continental interiors erode over time, and streams, wind, and ice transport the sediment to the edges of the continents and build them outward. Continental margins cover about 10 percent of Earth's surface and are composed of three primary subenvironments: the continental shelf, slope, and rise (see Figure 1.2).

The **continental shelf** is a smooth surface that slopes an average of less than one-tenth of a degree from the near-shore zone out to the shelf break at the top of the steeper continental slope. The seaward edge of modern continental shelves lies at an average water depth of 130 m, but continental shelves worldwide were subaerially exposed during major glaciations when more water was locked up in polar ice caps, reducing the volume of water in the oceans and lowering sea level. Continental shelves build seaward (prograde) over cycles of deposition and erosion as the shoreline migrates in response to changes in global sea level caused by changing global ice volume. Passive margins usually have wider continental shelves than active margins, the result of higher uplift rates on active margins. In tropical regions, the continental shelves can be covered with extensive submarine platforms built by coral reefs. Such reefs are common in the Caribbean and along the northern coast of Australia.

Offshore sandbanks and linear **sand ridges** occur on continental shelves where sand is abundant and ocean currents and waves are strong enough to transport sand-sized sediment. These features are generally 5 to 120 km long, 0.5 to 8 km wide, and can be up to 40 m tall, with heights that can equal 20 percent of the water depth. Although solitary sand ridges do exist, ridges typically occur in groups called sand-ridge fields. Individual ridges within ridge fields are typically spaced at distances of about 250 times the water depth. Sandbanks are sand ridges that develop in water shallower than 20 m. Most sandbanks are composed of fine to coarse sand, but gravelbanks do form in some areas with swift currents and abundant gravel. Sandbanks form in a variety of locations on continental margins, from the mouths of estuaries (like the Columbia River) to the edge of the continental shelf. Wave dissipation and refraction by offshore sandbanks helps protect beaches from erosion, and sandbanks also provide important nursery and feeding grounds for fish. Offshore sandbanks can present a significant hazard to shipping.

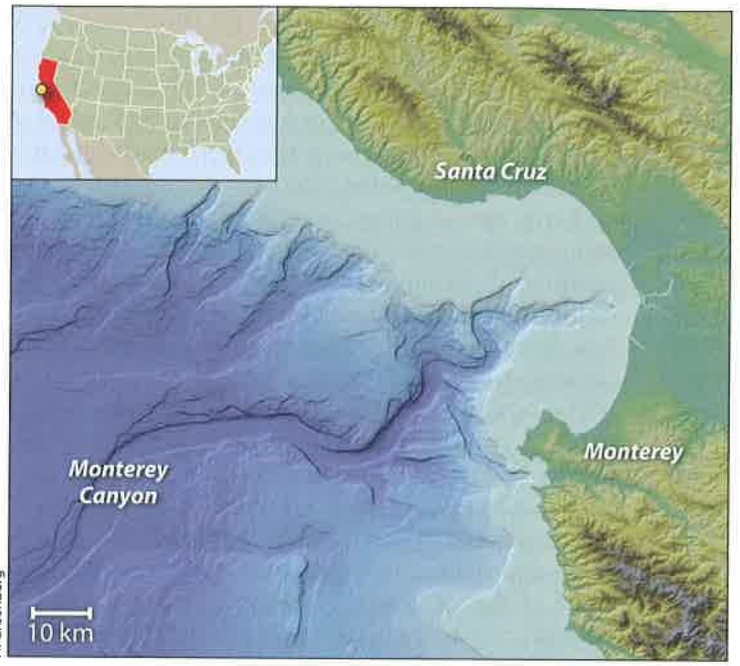
The **continental slope** drops off from the edge of the continental shelf at a depth of about 130 m and extends down to water depths of between 1.5 and 3.5 km. The continental slope is a rugged place. On average, it slopes more than 4 degrees and marks the buried edge of continental crust. Great **submarine canyons,** the largest canyons on Earth, extend down the continental slope [**Figure 8.12**]. Monterey Canyon, a modest submarine canyon on the continental margin south of Santa Cruz on the northern California coast, is about the size of Arizona's Grand Canyon. Even though submarine canyons sometimes exhibit sinuous meanders similar to those that develop in terrestrial rivers, we know that submarine canyons were not carved by rivers because they were submerged even during the lowest sea-level stands.

Submarine canyons are carved by turbidity currents, which transport clastic sediment down the continental slope and deliver it to the long-term geologic sediment sinks of the continental rise and abyssal basins. These currents are particularly erosive due to the high shear stress the dense flows exert on the continental slope. Turbidity currents sort sediments by grain size during transport and produce **turbidites,** distinctive deposits with upward-fining sequences that have coarse sand at the base and grade upward to fine silt and clay because the larger and heavier particles settle out before the lighter, smaller particles.

The **continental rise** is the ramp between the base of the continental slope and the deep abyssal seafloor at depths averaging around 4 km. The continental rise consists of oceanic crust buried beneath submarine fan sediment shed from land and transported across the continental shelf and slope. Accumulation at the head of the continental rise of sedimentary deposits carried by turbidites builds abyssal, or deep-sea, fans—the marine version of terrestrial fans extending down toward the deep seafloor.

During sea-level low stands, rivers and glaciers transport material out across the continental shelf to the shelf break 130 m below today's sea level. Delivery of a substantial volume of clastic sediment directly to the top of the continental slope creates large accumulations of unstable sediment. Large storms or earthquakes can trigger failure of these weak, generally poorly consolidated materials. Gigantic submarine landslides produce great slurries of sediment that move down the continental slope and may trigger tsunami waves.

During sea-level high stands, when base level is higher, continental-margin sediment is typically trapped in river-mouth estuaries and fjords because former fluvial and glacial valleys are inundated with seawater and streams drop their sediment loads farther inland. Deltas form either at the heads of estuaries or at the ocean in locations where fluvial sediment supply has been sufficient to fill estuaries and fjords. The sediment that does escape coastal entrapment is winnowed by waves, so sand is deposited on the near-shore continental shelf and mud is transported



Carved by submarine **turbidity currents,** bottom-hugging flows of sediment denser than the water through which they move, the largest, deepest canyons on Earth are submarine canyons that cross the continental shelf. Monterey Canyon, a modest submarine canyon off the coast of central California, is comparable in size to Arizona's Grand Canyon.

FIGURE 8.12 Submarine Canyons. Submarine canyons dissect the continental shelf and slope and deliver large amounts of terrestrial sediment to the deep ocean.

farther out to the middle shelf. On collisional margins with narrow continental shelves, sediment still makes it out to the continental slope even when sea level is relatively high, as is the case at present. Sedimentary deposits on the shelf grade outward from coarser modern beach and dune sands and gravels to finer-grained deposits that overlie coarse relict fluvial deposits that were laid down during periods of lower sea level. The morphology and sediments on continental margins reflect both the geometry of the continental shelf and the effects of repeated changes in sea level.

Abyssal Basins

The abyssal basins that constitute the deep seafloor between the continental rise and oceanic ridges are composed of linear hills, transform fault traces, normal fault scarps, and sediment-filled grabens in water depths of 3 to 6 km. The vast, low-relief abyssal basins cover more than half of Earth's surface and are divided into erosional bedrock uplands, **abyssal hills,** and depositional sedimentary lowlands that form the broad, smooth **abyssal plains.** Abyssal hills are elongate bedrock ridges that are parallel to the spreading segments of the associated mid-ocean ridge. They cover three-quarters of the floor of the Pacific Ocean and most of the Atlantic Ocean seafloor. They are a dominantly bedrock landscape and include relief up to 900 m that was generated by volcanic action during formation at a spreading center. Abyssal hills are typically blanketed by at least a thin layer of pelagic ooze. The abyssal basins are sedimentary environments that consist of smooth surfaces formed when sediment buried the basement rocks of oceanic crust. They generally characterize the seafloor near continental margins at the base of the continental rise.

Mid-Ocean Ridges

Mid-ocean ridges are underwater mountain ranges that circle the globe like the stitching on a baseball (see Figures 1.1 and 1.4). Also called **seafloor-spreading centers,** they form the network of fractures in Earth's lithosphere where plates diverge and spread apart above zones of convective upwelling in the underlying mantle. Mid-ocean ridges are about 1500 km wide and rise to 3 km above the surrounding ocean floor. Water depths are typically about 2 km along the crests of spreading ridges, but islands can form along ridges where volcanic output is high enough, as is the case in Iceland. Thermal expansion of the lithosphere associated with emplacement of hot basalt elevates the crust along spreading ridges, forming complex patterns of extensional grabens. Ridges typically have a major axial valley along the crest where volcanic eruptions create new oceanic crust. The lithosphere on opposite sides of a mid-ocean ridge cools and becomes more dense as the plates move away from each other; thus, the seafloor sinks to greater depths as it moves out from a spreading ridge.

Local mantle upwellings called **hot spots** produce volcanic eruptions that can build up to form island chains and basaltic plateaus in areas far from mid-ocean ridges. The 6000 km-long chain of the Hawaiian Islands and Emperor seamounts extends west from the "Big Island" of Hawaii to Midway Island, where the chain starts to bend northward to the Aleutian Trench off Alaska, recording changes in the direction of movement of the Pacific plate over a long-lived mantle hot spot.

Trenches

Marine trenches at convergent plate boundaries, where oceanic lithosphere is subducted into the mantle, are the deepest places in the ocean; they average about 8 km in depth. The Mariana Trench in the western Pacific Ocean, where terrestrial sediment loading is low, is the deepest point on Earth's surface at almost 11 km below sea level—several kilometers deeper than Mount Everest is high. In contrast to deep trenches with little sediment load, trenches along convergent margins with high sediment fluxes can be quite shallow because of sediment infilling. The Peru–Chile trench along the west coast of South America is shallow offshore of the northern Andes because steep slopes and intense tropical rainfall drive high rates of erosion that result in substantial sediment delivery. The trench is very deep off the hyperarid coast of Chile, where extremely dry conditions in the southern Andes deliver very little sediment to the marine environment. Farther south, offshore of the formerly glaciated regions of Chile, the trench is shallow due to the high delivery rate of sediment to the marine environment by glaciers.

Coral Reefs

Coral reefs are important features of tropical and subtropical continental margins where coral thrives in the warm waters. Reef-forming coral grows in clear, sediment-free water that is shallow enough to allow light to penetrate (i.e., less than about 60 m deep). Built up on fragments of coral broken by storm waves and the bodies of other marine organisms, coral reefs flourish where vigorous wave action supplies abundant nutrients and oxygen. Large freshwater inputs and turbid water with high sediment concentration inhibit coral growth, so reefs typically develop away from deltas and river mouths.

Charles Darwin first recognized the origin of coral reefs in the 1830s when he sailed through the South Pacific aboard H. M. S. *Beagle*. Based on the islands he visited on this epic voyage, Darwin proposed that there were three basic types of reefs. **Fringing reefs** grow near the shoreline or around islands as a narrow fringe usually a few hundred meters to a kilometer across [**Photograph 8.18**]. Breaks in fringing reefs often occur where streams deliver fresh water and sediment to coastal environments, impeding coral growth. **Barrier reefs** grow offshore and,

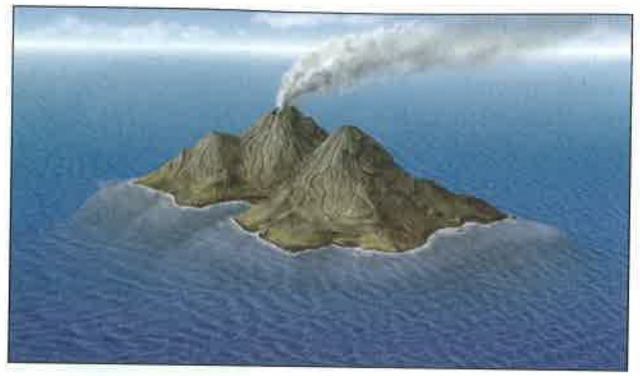
PHOTOGRAPH 8.18 Fringing Reef. Fringing reef surrounds a coral island and shelters the lagoon from wave energy on the southern Great Barrier Reef, Australia, near Heron Island.

like barrier islands, are separated from the mainland by lagoons. Barrier reefs, however, are generally broader than barrier islands and can be up to 15 km wide. The growth rate and pattern of a coral reef reflects the balance between coral growth and wave erosion, both of which are concentrated on the seaward side of the reef. The most famous example of a barrier reef, the Great Barrier Reef, extends for more than 2000 km along the eastern coast of Australia.

An **atoll,** Darwin's third variety of reef, is a circular reef that surrounds a lagoon. Atolls rise just a few meters above sea level and lack a central island. Darwin recognized that atolls form when a fringing reef around an island is able to keep pace with subsidence and builds up fast enough to maintain shallow water conditions even as the island sinks beneath sea level [**Figure 8.13**]. Borehole samples from some atolls have found more than 1000 m of coral reef rock above oceanic basement rocks. More recently, atolls were strategically important in World War II, when they were used as positioning sites for troops, as strategic air bases, and as nuclear test sites.

Applications

Coastal and marine geomorphology has immediate practical applications. Sand and gravel from coastal environments and from relict estuaries and deltas provide important sources of construction materials in many regions of the world. Most of the world's largest cities are near the ocean, and 75 percent of Americans live in coastal states. Many coastal cities, such as Venice, Italy, and New Orleans, Louisiana, rely upon groundwater pumped from the coastal marine and fluvial sediments beneath their buildings.



Atolls start as **fringing coral reefs** typically growing on the margins of volcanic islands in warm tropical waters.

When gradual growth of a coral reef can keep up with the ongoing **thermal subsidence** of the island (resulting from the cooling of the oceanic crust as it ages), the reef can remain close to sea level.

Eventually, continued subsidence of the volcanic island leads to development of an enclosed **lagoon** surrounded by the originally fringing reef.

FIGURE 8.13 Atoll Evolution. Atolls evolve if reef-building organisms can produce reef material at a rate equal to the rate of volcanic island erosion and subsidence.

These areas are vulnerable to settling and land subsidence, a problem that is compounded by levees and dams that restrict delivery of sediment to and deposition in coastal areas.

On average, North American coastlines are eroding at a rate of just under a meter a year, although rates vary greatly among different regions and depend on local conditions including wave heights, material properties, and land subsidence/uplift rates. The U.S. Army Corps of Engineers has classified about one-quarter of U.S. coastlines as seriously eroding. Severe coastal erosion has profound implications for risk to coastal properties and for geologic hazards during storms. Reduction in the supply of sediment to coastal environments, from both inland dam construction and coastal armoring, affects the sediment supply to beaches and barrier islands.

Historically, efforts to combat beach erosion have focused on engineered solutions. For instance, construction of groins to trap and retain sediment typically causes downcurrent erosion because of longshore sediment transport. Consequently, the construction of some coastal engineering structures leads to the need to build more structures along the shoreline. For example, construction of bulkheads by property owners along the shores of Puget Sound in Washington State and groin fields in New Jersey resulted in extensive beach engineering that greatly altered near-shore environments.

Understanding coastal sediment budgets that consider sediment carried in and out by longshore drift and other processes can help geomorphologists address such problems as harbor siltation and beach erosion. Longshore drift builds spits that block harbors and shipping channels and removes sand from prized beaches. The problems of delivering sand to eroding areas and removing it from aggrading areas have generated ongoing controversies over the role of shoreline stabilization versus planning to accommodate shoreline mobility. Some places have instituted expensive programs of beach nourishment, where sand is routinely pumped from offshore bars onto the beach in order to provide areas for recreation and to protect buildings and roads from erosion. In some of these areas, there is no expectation that the sand will stay in place; rather, the nourishment is regarded simply as beach maintenance.

Development on barrier islands can greatly affect their morphology through changes in sediment delivery and storage. Extensive development of a popular East Coast resort (Ocean City, Maryland) in the 1920s through 1970s motivated extensive efforts to stabilize the nearby barrier inlet by building a jetty at the southern end of the city. With the longshore current running north to south, the jetty stabilized the inlet but reduced the sediment supply to the barrier islands to the south, across the inlet. This caused extensive erosion of Assateague Island off southeastern Maryland, unintentionally shifting the shoreline there almost a kilometer inland.

Inundation from storm surges, tsunamis, and sea-level change are uniquely coastal issues. The ways in which waves interact and refract upon approaching land leads to profound differences in local tsunami and storm surge hazards along vulnerable coastlines. Lagoons and marshes dissipate wave energy and slow wave speed because they are rough; thus, they provide natural coastal defenses from hurricane-driven storm surges and tsunamis. The loss of such buffers can prove disastrous and was one factor contributing to Hurricane Katrina's devastation of New Orleans. The destruction of mangrove swamps that lessen wave energy and provide rich wetlands supporting high biodiversity is an ongoing problem in tropical regions around the world.

Coastal environments stand on the front lines of potential impacts from the anticipated sea-level rise projected to accompany global warming over the next century. Understanding how coastlines behaved during past sea-level rises can help predict future changes. The forecasted rise in sea level over the next century will have its first impacts on low-lying island nations and regions, and coastal geomorphologists expect large changes from even small amounts of sea-level rise in flat-lying coastal areas. Deltas and estuaries that will experience the greatest effects from sea-level rise also happen to be the environments where a majority of Earth's human population lives, where much of the world's seafood is nurtured, and where many of our cities are located.

Understanding the nature of the continental shelf as an environment exposed as land during glaciations and submerged during warmer interglacial periods has a major bearing on regional archaeology and biogeography. Because extensive portions of the continental shelves were land during glaciations, many archaeological sites are now submerged or are buried below sediments on the shelves. For example, ancient coastal environments in Siberia and Alaska and between New Guinea and Australia, now below sea level, are suspected of having been corridors for human migration. Similarly, during glacial sea-level low stands, rivers flowing across the modern continental shelves likely provided refugia for freshwater species, like salmon, during glacial advances that filled drainage basins with ice.

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DIGGING DEEPER What Is Happening to the World's Deltas?

Many of the world's major river deltas are slipping below sea level (Syvitski, 2008; Blum and Roberts, 2009; Syvitski et al., 2009). Why is this happening and why are scientists sounding the alarm in prominent scientific journals?

Deltas are vitally important to the billions of people who inhabit Earth. Their nutrient-rich alluvial soils and extremely flat topography provide wide expanses of excellent agricultural land on which to grow a variety of crops. The rivers that flow onto deltas provide a ready source of irrigation water, and the levees built to contain many of these rivers provide a sense of security from flooding that has encouraged people to settle on productive deltaic lowlands. Deltas are home to half a billion people, almost 10 percent of the world's population. Major cities, including Shanghai, New Orleans, and Bangkok are located on deltas. River deltas can be very large; the Amazon River delta covers almost half a million km^2.

Because deltas form where rivers empty into the ocean, and thus are fine-tuned to sea level, they are vulnerable to flooding and coastal erosion. Their flatness, with gradients as low as 0.00001 (1-mm rise per 100 m run), makes large areas vulnerable to flooding from overflowing rivers and rising seas; a 10-cm rise in sea level could move a delta shoreline landward by 10 km. Such low gradients make flooding on deltas deadly. Syvitski et al. (2009) report that in 2007–2008, delta floods killed more than 100,000 people and drove more than a million others from their homes.

The size and extent of deltas reflect a precarious three-way balancing act between the mass of sediment brought in by rivers, the volume of water in the world's oceans, and the sinking or subsidence of the delta. For example, deltas will shrink if ice sheets melt and/or the ocean warms, causing global ocean volume to increase and eustatic sea level to rise. Conversely, deltas expand if sediment yields increase and rivers supply more sediment to coastal environments. Deltas naturally subside because of the solid-Earth isostatic response to massive sediment loading and because delta sediments compact as they dewater and the organic matter within them decays and oxidizes.

Ocean volume is increasing and eustatic sea level is rising at an accelerating rate due to global warming **[Figure DD8.1]** (IPCC, 2007). After eustatic sea-level stabilized about 7000 years ago, long-term rates of eustatic sea-level rise have been <1 mm/yr. Since about 1940, the rate of global sea-level rise has increased to more than 2 mm/yr. Predictions of future rise are uncertain but suggest that the rate of sea-level rise will continue to increase as both the climate and the oceans warm, melting ice sheets and thermally expanding ocean waters.

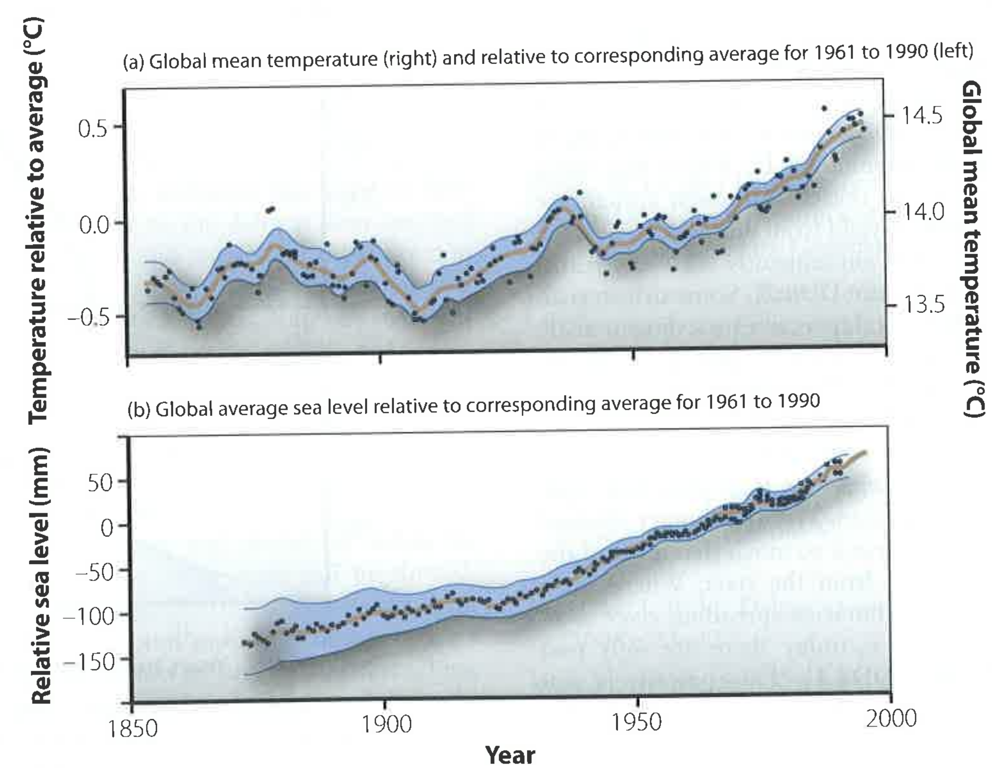

FIGURE DD8.1 Global warming began in earnest after 1900 (top), about the same time that the global average rate of sea-level rise increased (bottom). Global, eustatic sea level has risen about 200 mm in the past 150 years. [From IPCC (2007).]

DIGGING DEEPER What Is Happening to the World's Deltas? *(continued)*

For individual deltas, what matters is the sum of eustatic sea-level change and subsidence. Natural deltaic subsidence rates of a few millimeters per year have been accelerated by human activities. Examples include centimeters per year of subsidence on the Po River delta, from which methane was pumped, and the Chao Phraya delta on which Bangkok is located, where groundwater was withdrawn (Syvitski et al., 2009). Building levees along rivers contributes to subsidence by preventing sediment delivery to large parts of the delta surface. In some places, like the mouth of the Mississippi River, sediment that once spread over the delta surface is now delivered by levee-constrained channels right to the continental shelf. The result of this sediment starvation is that some delta surfaces inhabited by people have subsided enough that they lie below sea level and are kept safe and dry only by levees. Alternately, sediment accumulation in the channel can cause the river bed elevation to rise above the level of the surrounding floodplain.

Syvitski et al. (2009) investigated the critical balance of supply, subsidence, and ocean volume for 33 of the world's deltas. They used SRTM data (see Chapter 2) to determine delta surface elevation, historical maps to determine the number and location of distributary channels in the past, and satellite images to establish the extent of recent flooding. They supplemented these observations with existing data about river sediment loads and eustatic sea-level rise. Their conclusions are sobering. It appears that reduced sediment delivery is the biggest culprit in delta shrinkage. Most deltas are now sinking much faster than sea level is rising because so much sediment is trapped upstream by dams or impounded by levees that delta aggradation has slowed or ceased. Although agriculture initially increased sediment yield from drainage basins, the construction of large dams subsequently starved streams and deltas of sediment [**Figure DD8.2**]. Some deltas, such as that of the Yellow River, today receive no sediment at all.

The Nile delta, home to ~50 million people, is a prime example of a delta in trouble. The Aswan high dam captures about 98 percent of the sediment load that the Nile River carries north from central Africa (Syvitski, 2008). Irrigation channels on the delta now trap what little sediment makes it through the dam. The distributary channel network on the delta has shrunk so much that most of the delta is now disconnected from the river. Where there were once dozens of distributaries spreading river sediment over the delta surface, today there are only two (Syvitski, 2008) [**Figure DD8.3**]. This pattern is not unique to the Nile delta. Syvitski et al. (2009) report that the distributary network shrank through human actions in nearly half of the 33 deltas they studied.

Predicting the future of deltas requires an understanding of the mass balance of sediment in the system as well as predicting how fast the land is going down and the sea is going up. Using estimates of sea-level rise, historical rates of subsidence, and long-term average rates of sediment storage, Blum and Roberts (2009) conclude that by 2100 more than 10,000 km^2 of the Mississippi River delta will be inundated by seawater. The culprit: lowered sediment loading. More than half the sediment that used to flow down the Mississippi River to the delta is now trapped by the numerous dams in the watershed. Syvitski and Milliman (2007) report that the Mississippi River watershed contains more than 50,000 dams. Rates of sediment transfer from the Mississippi River to its delta are so low that sedimentation cannot keep up with even modest rates of eustatic sea-level rise and subsidence. Relative sea-level rise near the delta is anticipated to be on the order of a meter over the next century, flooding much of the low-lying wetlands and coastal marshes.

Can the trend of sinking and shrinking deltas be reversed? Probably not, but the process can be slowed. Kim et al. (2009) created a model of the Mississippi delta and used what we know about sediment loads, sea-level rise, and sediment compaction to suggest that breaching Mississippi River levees below New Orleans could build large amounts of new land [**Figure DD8.4**]. They conclude that if about half of the river flow were spilled through levees onto the subsiding delta, enough flow would remain for navigation to continue in the main channel while 700 to 1200 km^2 of new land would be created over the next century. This would offset about half the expected land loss. Of course, doing so would require abandoning the century-old policy of containing the Mississippi River in levees, at least for the reach below New Orleans.

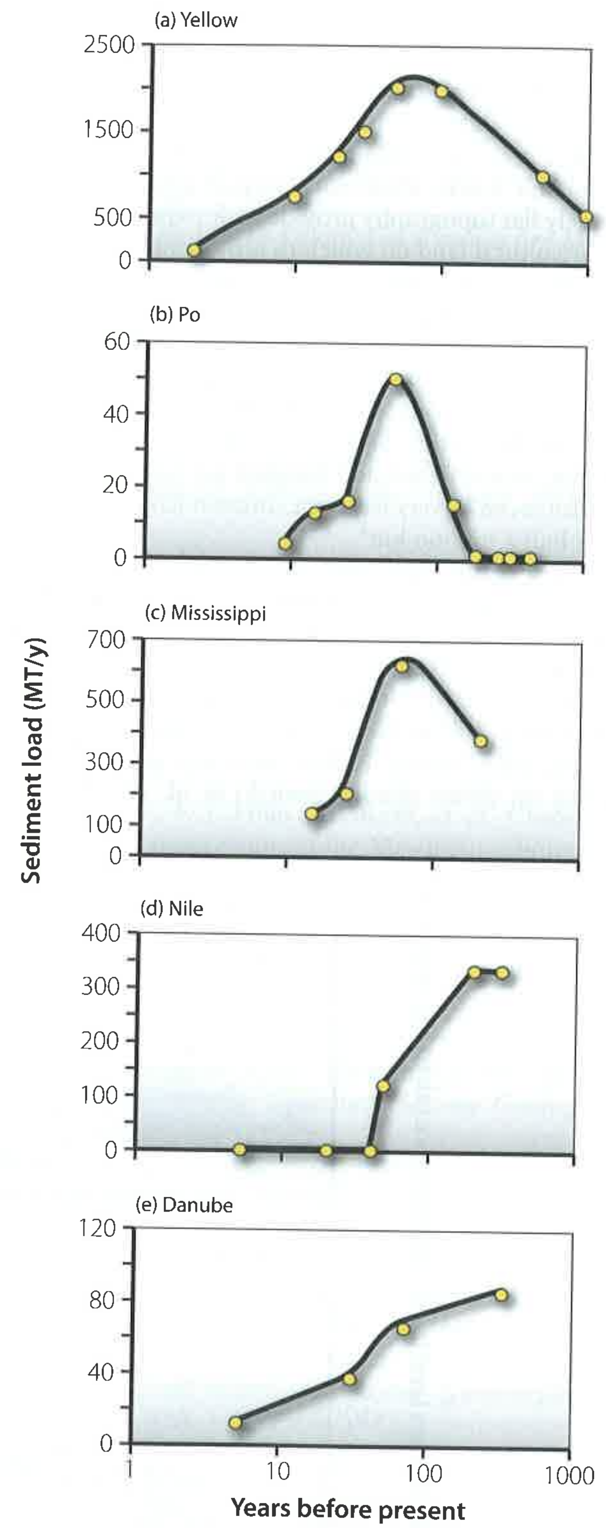

FIGURE DD8.2 The sediment loads (expressed as MT, megatons, millions of tons) of many rivers have changed over recent centuries. Many rivers experienced a peak in sediment loading during periods of intensive agriculture and then a drop in loads as dams were built and sediment moving downstream was trapped. Note the logarithmic scale for time. [From Syvitski (2008).]

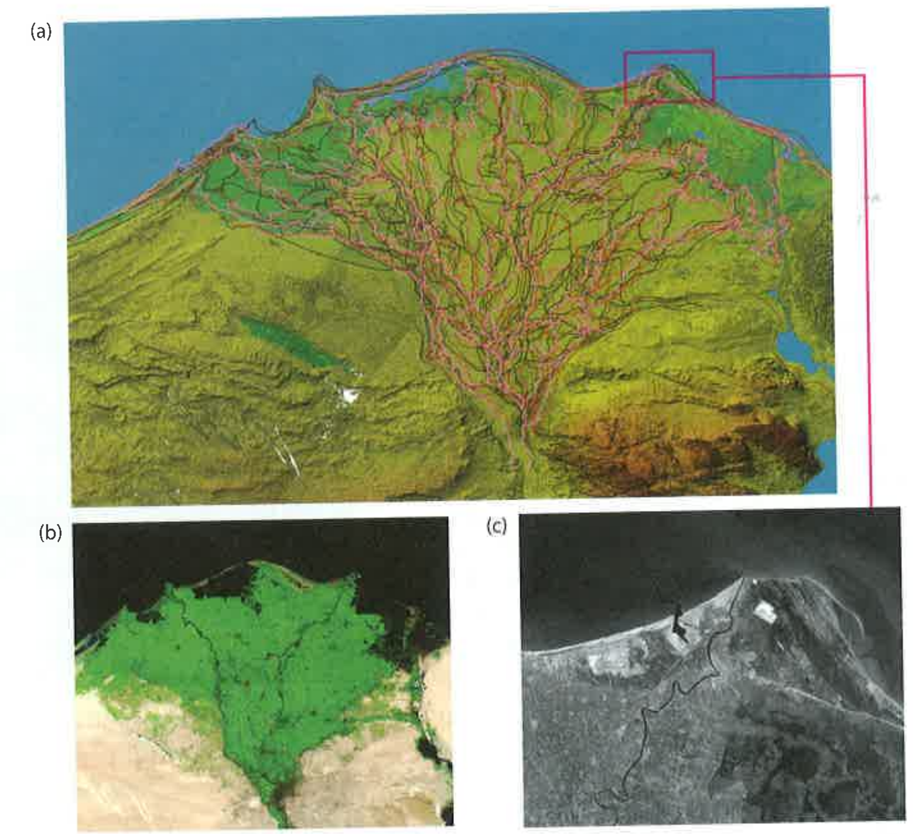

FIGURE DD8.3 The number and pattern of distributary channels on the Nile River delta has changed since the Aswan high dam was built. (a) All the distributary channels indicated on maps made of the delta in 1813, 1831, and 1897 superimposed on SRTM topographic data. (b) A remotely sensed image showing the current situation—only two main distributary channels are still carrying Nile River waters to the ocean. Look closely at part (c) to note the coastal erosion moving sand offshore to the northwest. Compare the older shorelines for this area with the inset in part (a) to see how much loss of coastal land has occurred. [From Syvitski, (2008).]

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DIGGING DEEPER What Is Happening to the World's Deltas? *(continued)*

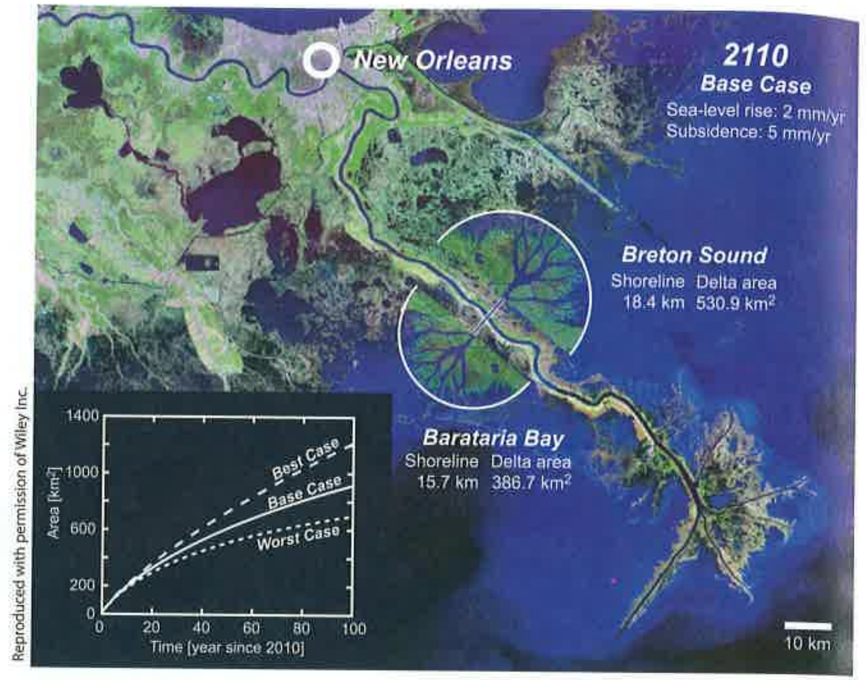
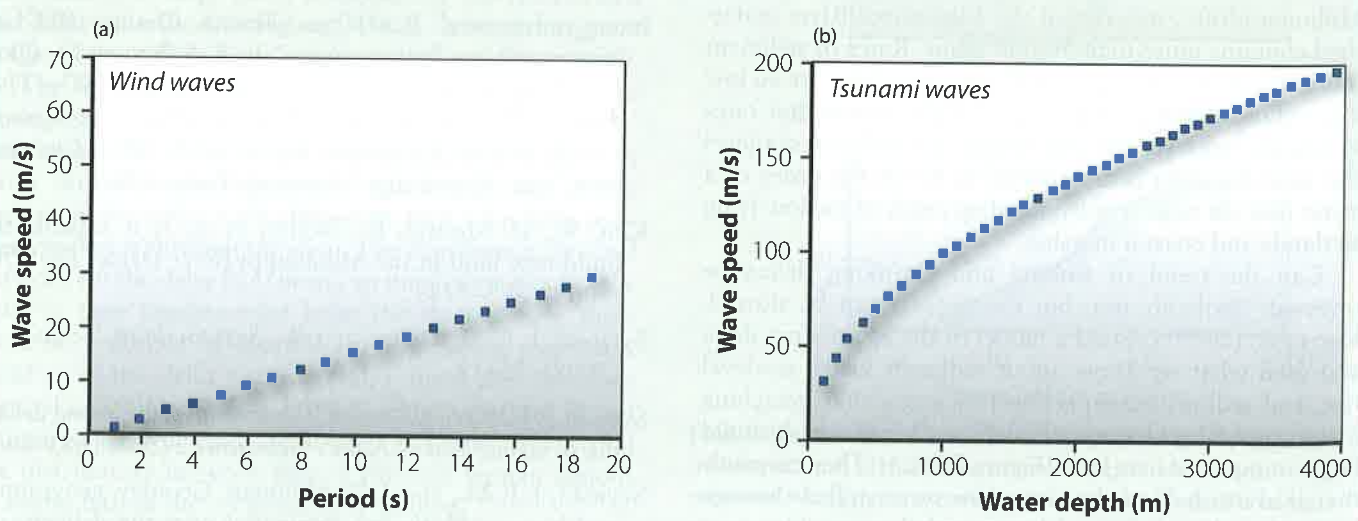


FIGURE DD8.4 This synthetic view imagines what the Mississippi River delta might look like in 2110 if the levees downstream of New Orleans had been breached in 2010. Two new delta lobes form at the breaches. The inset graph shows the rate at which new land will be built once the levees are breached, considering a best-case scenario in which sea level holds steady and subsidence is 1 mm/yr, a base case where sea level rises 2 mm/yr and subsidence is 5 mm/yr, and a worst case where sea level rises at 4 mm/yr while the land subsides at 10 mm/yr. [From Kim et al., (2009).]

WORKED PROBLEM

Question: Consider deep-water waves, created out at sea by a strong storm. Do waves with shorter or longer periods travel more quickly? How does the speed of these deep-water wind waves compare to the speed of a tsunami wave moving over the open ocean? Illustrate your answer graphically by calculating and plotting wave speeds for deep-water wind waves with periods between 1 and 20 s and for tsunami waves with a period of 10 min traveling over the ocean toward shore where water depths vary between 100 and 4000 m.

Answer: For the deep-water wind waves created by a storm, use eq. 8.2 to calculate wave speed. Your graph, as in the example below, should show that wave speed increases linearly as wave period increases. For the tsunami, which behaves as a shallow-water wave, use eq. 8.3. You should find that tsunami waves move much more rapidly than wind-driven waves and that tsunami waves slow when they approach land and the water depth shallows. In general, tsunami waves move much more quickly than wind waves.



Graphs indicating how (a) the speed of wind waves is linearly and positively related to the wave period, and (b) how tsunami wave speed increases nonlinearly with water depth.

KNOWLEDGE ASSESSMENT Chapter 8

1. How do geomorphologists study Earth's surface under the ocean waters?
2. Contrast shoreline geomorphology at active and passive continental margins.
3. Compare emergent and submergent coastlines and explain the processes that lead to these two different types of coasts.
4. How much salt does ocean water contain and how does this affect its mixing with fresh water from rivers?
5. List and explain the two major controls on local sea level and shoreline position.
6. Identify and explain which part of the tidal cycle is most likely to cause geomorphic change along the coast.
7. Describe the history of sea level globally over the past 21,000 years.
8. Describe the mechanisms that supply sediment to the coastal zone.
9. Explain how tides are geomorphically important.
10. Predict how fetch and wind speed will affect the ability of waves to do geomorphic work.
11. Explain wave base and how it is geomorphically important.
12. Sketch the difference between plunging, spilling, and surging breakers.
13. Explain wave refraction and how it affects the evolution of coastlines.
14. Define longshore drift and explain why it is important for coastal geomorphology.
15. Explain why shoaling occurs and why it is geomorphically important.
16. Predict the geomorphic effects of installing a jetty or a groin in the coastal zone.
17. Define storm surge and explain how it catalyzes coastal geomorphic change.
18. Explain how tsunami waves differ from wind waves.
19. What can a geomorphologist learn from the elevation of wave-cut notches and wave-cut platforms?
20. Sketch a sea stack and a sea arch and explain how they form.
21. Define a beach berm and explain how it is created and changes with the seasons.
22. Explain how spits and barrier islands are genetically related.
23. Predict where a flood-tide delta will form and explain how and why its size differs from an ebb-tide delta.
24. Predict how lagoons evolve over time, specifying the relevant processes.
25. Explain why clay and salt water are important in the behavior of estuaries.
26. List the four major types of estuaries and explain how each forms.
27. Sketch the internal architecture of a delta deposited into a freshwater lake by a stream or river.
28. Explain, from a process perspective, the reason that river-dominated, wave-dominated, and tide-dominated deltas have different shapes.
29. Define lithogenic, authigenic, and biogenic sediments and predict where in the marine realm you would expect to find each kind of sediment.
30. Sketch the large-scale geomorphology of an active and a passive continental margin.
31. Describe how continental margin sedimentation differs between glacial and interglacial periods.
32. Describe a submarine canyon and explain how it forms.
33. Explain, with sketches, Darwin's theory of atoll formation.
34. What type of sediment settles out and covers the abyssal plains of the deep ocean?
35. Explain beach nourishment, how it works, and why it is done.