Introduction to Marine Environmental Science and Oceanography

CHAPTER

1

Although a turning point would be difficult to pinpoint, sometime within the past several decades a new environmental awareness was developed with the understanding that conservation of living resources cannot be adequately achieved by focusing only on imperiled species or single components of ecosystems. **Environmental science** was born out of a need to consider all aspects of the physical and living world in solving complex environmental problems. **Environmental biology** is primarily an incorporation of biological processes that realizes the need to embrace environmental processes when studying living resources in the natural world and solving conservation problems. This is especially true in the marine environment where environmental systems processes such as currents, waves, chemical processes, and geologic activity are dynamic and often unpredictable and strong influences on the living components. Although the perspective of this text is biological, rarely can conservation be achieved without considering the non-living environment's effect on life processes.

Marine conservation is concerned with the preservation and management of the ocean environment and the organisms that live within it or are otherwise dependent upon it, that is, the **marine ecosystem** (**Figure 1-1**). The extremely complex, dynamic, and variable ecosystems of the oceans cannot be understood without first considering oceanography. *Oceanography*—the study of the physical, chemical, geologic, and biological processes of the oceans—is the foundation upon which a conservation ethic can be built. This chapter provides a general overview of the non-biological components of oceanography. This includes studies of the water column as well as land areas adjacent to the sea, the continental coastlines, and islands. The physical dynamics of water movements are discussed, including waves, tides and the tidal cycle, and currents occurring at the surface and throughout the water column. Finally, some general aspects of ocean chemistry are covered, introducing basic aspects of the chemistry that comprise the biological components of the sea.

1.1 Geological Oceanography

Geological oceanography or **marine geology** is the study of the sea floor and the lands bordering the sea. Geological oceanographers often study aspects of the formation of the sea floor and coastlines by major geologic processes and sediment formation and accumulation. These processes govern the current mineral makeup of the seafloor and coastline and the physical features of these regions. Marine geology is closely linked with physical and chemical oceanography because physical and chemical processes determine the geology of oceanic regions.

Seafloor Spreading

The sea floor is remarkably complex with a diversity of physical features, some similar to those found in terrestrial

Figure 1-1 A view of the Pacific Ocean and volcanic shoreline of Japan.

regions and others unique to the marine environment (**Figure 1-2**). This region, especially of the deeper ocean basins, was virtually unknown to humans until about the past 50 years. Most of the major physical features are formed through the processes of **seafloor spreading** and **plate tectonics** (**Figure 1-3**). This concept is based on observations that new seafloor is constantly, but slowly, being extruded from the ocean bottom at mid-ocean ridges. Extensive ridges can be seen running through the middle of the Atlantic Ocean bottom and along the east side of the Pacific Ocean up through the North American continent. From a human perspective, these processes are extremely slow. At an average rate of about 1 cm/yr, it takes almost a century for the seafloor to move 1 meter, and it has taken almost 200 million years for the seafloor formed at the mid-Atlantic ridge to move to the U.S. Atlantic coast. As the forces of seafloor spreading gradually push the seafloor away from the mid-ocean spreading centers, it eventually **subsides**, sinking down below the continental land masses or into deep ocean **trenches**. Some of the deepest of the trenches formed during this process are the western Pacific Ocean, reaching depths as great as 7,000 meters below the average depth of the seafloor. These forces affect the seafloor and the continental land masses worldwide and result in a movement of the large **plates** that interlock to form the ocean basin and support the continents. For example, the North American continent is mostly over one large oceanic plate. Seafloor spreading processes result in the production of some remarkable features, both under the sea and above the sea surface, including mountains, islands, and volcanoes. They also produce some of the most extreme events on the planet: earthquakes and tsunamis.

Seafloor Regions

The features of the sea floor are commonly categorized according to their association with land masses, physical characteristics, and depth below the ocean's surface. Although the divisions between adjacent regions are not always exclusive and discrete, the categories prove a convenient way to make generalizations concerning physical process, ecosystem categories, and conservation needs.

Continental Margins

The Earth's continents might be described as sitting on or riding over the oceanic plates, and the edges of the continents where they meet the sea are called **continental margins** (**Figure 1-4**). This is the location of most of the shallow ocean waters. Because the waters covering the continental margin are typically the most productive and accessible to humans, many of the conservations issues covered in later chapters will be focused in this region. Many of the living

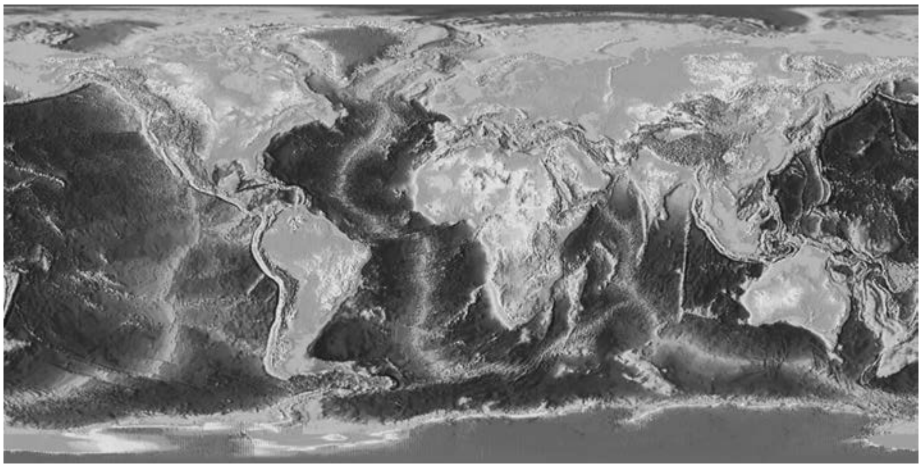


Figure 1-2 Relief map of the Earth. Shades of gray indicate depths and elevations; ocean regions are indicated from light to dark gray, shallow to deep. Note the mid-ocean ridges and other seafloor features. (See color plate 1-2.)

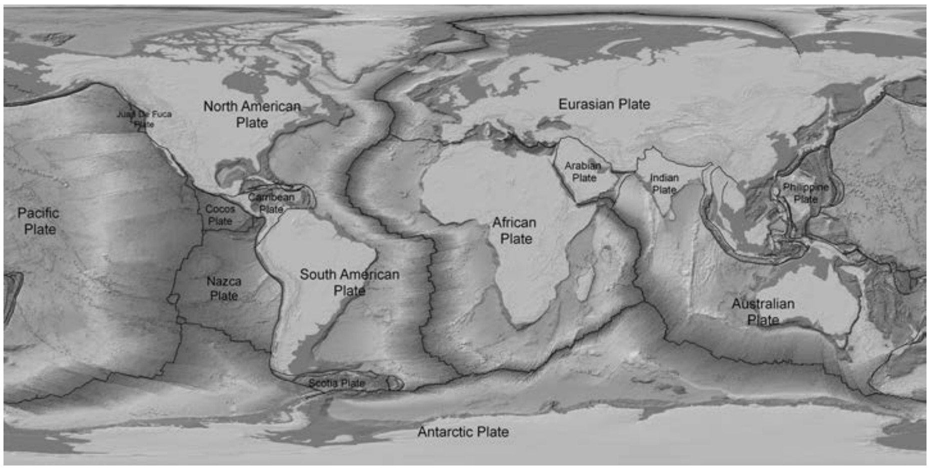


Figure 1-3 Map of the sea floor with evidence of seafloor spreading based on age of ocean sediments. Newer areas of the seafloor near the spreading centers are indicated by dark gray with jagged lines. (See color plate 1-3.)

resources are obtained from water over the continental margin and this is also where about ⅓ of the Earth's oil and gas reserves are located.

Continental margins can be classified according to the plate interactions associated with the margin. **Active margins** are common where the plates are interacting or colliding, such as along the Pacific coast of North America. In this region the intersection of plates continues onto the continent in California, and result in frequent earthquake activities along the U.S. west coast. There is another active margin in the west Pacific; the plate interactions there have resulted in the formation of the many volcanic islands, including the islands that make up Japan. The slope from the coast out to the deep ocean is typically steeper along active margins than along passive margins. Some of the most biologically productive areas of the oceans occur in the deep near-shore waters along active margins as a result of ocean processes, discussed below.

Passive margins (Figure 1-4) are typically in the middle of one of the ocean plates and, thus, are the source of little geologic activity. The U.S. Atlantic coast is along a passive margin. The slope of the sea floor here is very gradual and the margin is much broader. There is often a buildup of sediments well offshore along passive margins, especially where there is significant river input. The biological productivity can be elevated along passive margins where there are large nutrient inputs from continental sources.

The continental margin can be subdivided into three regions: the shelf, slope, and rise, progressing seaward from the coast. The **continental shelf** is generally the most biologically rich region of the ocean. Several factors contribute to this high productivity. These include the input

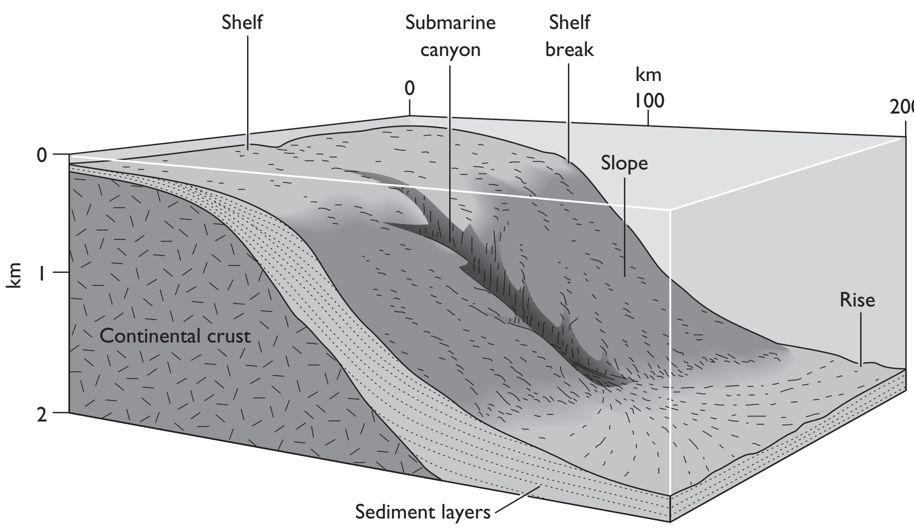


Figure 1-4 Representation of a continental margin, consisting of the continental shelf, slope, and rise. Submarine canyons cut through the continental margin in some regions. (See color plate 1-4.)

of nutrients from rivers flowing onto the shelf or rising up from deeper ocean waters, and the higher sunlight penetration into shallow depths resulting in high rates of photosynthesis. The continental shelf is less than 10% of the ocean bottom. High biological production, biodiversity, and accessibility make this region heavily exploited and impacted by human activities, however. The focus of marine conservation efforts thus is in this region. The downward slope along the shelf is only about 1 degree on average, resulting in only a gradual decrease in depth of less than 2 meter per kilometer moving away from the coast. If the shelf could easily be viewed, it would appear as flat as the flattest plains on the continents. Still, the average depth over the shelf is about 75 meters: deep from a human perspective but very shallow compared to the average depth of the ocean (about 13,000 feet or about 4,000 meters). The width of the continental shelf varies from being almost nonexistent along some active margins, to being hundreds of miles across along some passive margins. This can have an important influence on what types of organisms live above the shelf in different regions of the world. The end of the continental shelf is at the shelf break, where the slope begins to increase onto the continental slope.

The **continental slope** is the region of transition between the flat continental shelf and the deep-sea floor. This slope is somewhat steeper than the shelf, at about 4 degrees. Still, this equates to about 75 meters vertical change per 1 kilometer distance. The slope comprises about 6% of the sea floor. The slope region is richer in features than the shelf, and drop-offs, ridges, and canyons are evident. When first discovered by oceanographers, the submarine canyons were believed to be formed only in association with coastal river input; many are found near rivers and the canyons may cut into the continental shelf. However, many canyons are present in areas where there is no river input. These, it was determined, are formed by slumping of sediments off the continental slope. This slumping can result in an undersea flow of sediments, and water currents digging into the slope and forming a canyon. These features in the slope are important for fishes and other organisms since they provide shelter and hiding places in an otherwise mostly barren undersea terrain.

The sediments that slump off the continental slope may be transported toward the deep ocean basins. This forms a transitional region dividing the continental margin from the deep sea, called the **continental rise**. The rise is composed of loosely consolidated sediments and ends at the edge of the deep-sea bottom, making up about 5% of the sea floor. The slope is steeper, but similar features occur here as on the continental slope.

Deep Ocean Basins

The **deep ocean basins** (**Figure 1-5**), sometimes called the **abyss**, make up about 40% of the sea floor. Much of this area is extremely flat, flatter even than the continental shelf. This is because most geologic features in the deep ocean basin have been slowly covered with sediments, many of them of biological origin. Although the sediment accumulation is at a very slow rate, many areas of the deep ocean have remained fairly undisturbed for millions of years. This allows sediments accumulations (called **oozes**) up to almost 5 kilometers deep in some regions (though sediment accumulation off river deltas may be as thick as 20 kilometers). The origin of deep-sea sediments varies from region to region, largely as a reflection of ocean chemistry and depth. They are composed of a combination of

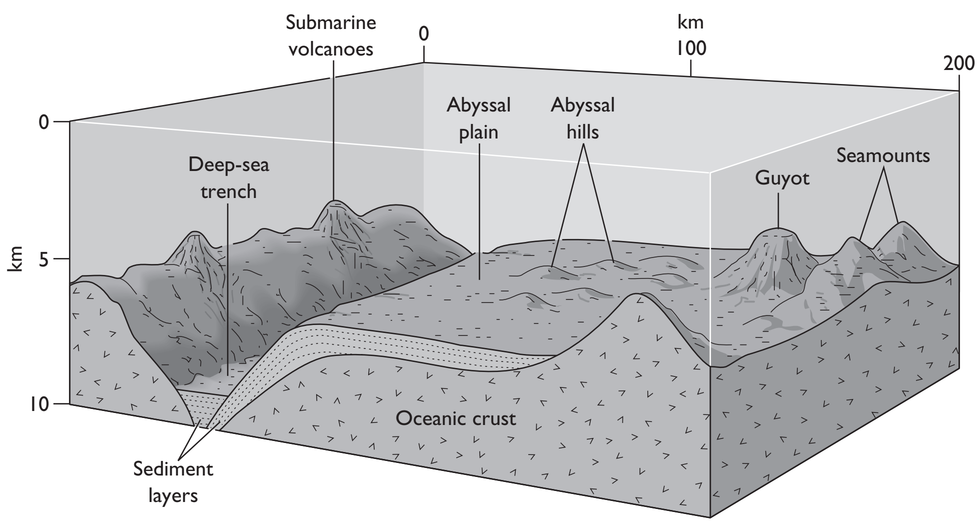


Figure 1-5 Seafloor features of the deep ocean basin. Deep-sea trenches cut into the abyssal plains. Abyssal hills, guyots, seamounts, and submarine volcanoes are scattered over the plains.

fine clays and the remains of organisms, primarily the shells of microscopic organisms that die and sink to the bottom. Abyssal hills break the relief along the ocean basins. These are extinct volcanoes partially exposed above the deep sea sediments. Of course they may have been much higher above the sea floor in past times, but are now mostly buried.

Oceanic Ridges

Oceanic ridges are the mountain ranges of the sea floor. These occur at the spreading centers, often in the center of deep ocean basins. The region associated with the ridges comprises about 30% of the seafloor. The geologic activity at these spreading centers results in mountain chains more extensive than any occurring on the continents, extending about 60,000 kilometers in total length. There is extensive volcanic activity associated with the ridges. Because of the extreme water depths, the ridges are still far below the sea surface, even though the mountains may rise up to 2 kilometers above the seafloor. Because of the relative inaccessibility to humans, little was known of these regions until late in the 20th century, and most are still unexplored. Recent explorations of the ocean ridges have resulted in the discovery of some of the most remarkable geologic features and unique ecosystems on Earth.

Hydrothermal Vents

One type of feature associated with the ocean ridge systems is a geologically active structure called a **hydrothermal vent** (**Figure 1-6**). At these vents, ocean water circulates below the seafloor to eventually spew out in undersea geyser-like formations. The minerals in these waters give them a distinct "smoky" appearance and they are usually named according to the appearance of the water venting from these features, for example, **black smokers** and **white smokers**. Hydrothermal vents support a unique ecosystem that is supported not by **photosynthesis** (the conversion of carbon dioxide into organic compounds using sunlight energy) but by **chemosynthesis** (the synthesis of organic compounds using inorganic chemical reactions as a source of energy). The minerals being released into the water eventually are deposited on the seafloor. These deposits are potentially a valuable recourse for human exploitation, which could lead to conservation issues, but for various reasons are not currently mined commercially. Conservation of these ecosystems will be discussed in Chapter 8.

Undersea Mountains

Away from the ocean ridges there are other, more isolated undersea mountains, called **seamounts**. These are formed by volcanic activity and most are inactive. Many were formed when they were located near a geologically active area like an oceanic **spreading center** at the mid-ocean ridges, or **hotspot**, a region of upwelling of hot mantle materials (one such hotspot formed the Hawaiian island chain). As the seafloor "conveyor belt" carries these mountains across ocean basins, they gradually get further away from the spreading center or hotspot where they were formed and are carried beneath the ocean's surface; thus, the "relative" sea level gradually changes around the seamount, resulting in the older seamounts being found well beneath the surface of deep ocean waters. Changes in global sea level can also affect whether these mountains are exposed as islands or hidden beneath the sea. With climate change and increasing sea level, low elevation volcanic islands could be flooded and become seamounts if covered by the ocean waters. Those living on such islands are gravely concerned about predicted sea-level rises (see Chapter 3).

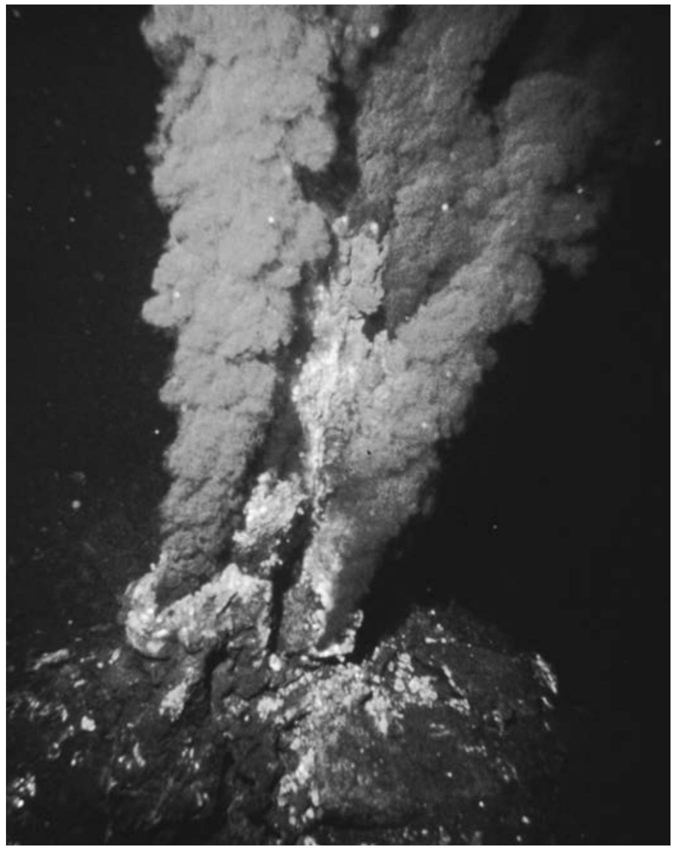


Figure 1-6 A black smoker at a mid-Atlantic ridge hydrothermal vent.

Seamounts vary in height, ranging to over 1 km above the sea floor. Some form a relatively sharp peak, but others are flat-topped, called **guyots** (pronounced gē′-ō). The flattened top of guyots is a result of erosion from weathering during the time period they were exposed as islands at the surface. These seamounts and guyots may provide habitats for rich ecosystems relatively near to the surface (but still up to hundreds of meters depth); in some cases they serve as an area for large fish to congregate for feeding or reproduction. Only in recent years have humans been able to access and exploit living resources from these undersea mountains, leading to concerns about potential overharvest. Conservation issues in the deep sea are discussed further in Chapter 8.

1.2 Oceanic and Coastal Regions

Sea Coast

The sea coast is the region where the land meets the sea. Although coastal regions are not covered by water at all times, they are so strongly influenced by the ocean that they must be considered in the conservation of marine environments. The access the sea coast gives to the resources of the sea results in high levels of human habitation. The resulting environmental modifications, pollution, and overexploitation of resources results in many conservation issues that will be discussed in late chapters. This section addresses how physical factors governed by oceanic and coastal processes determine the makeup of coastal areas.

Beaches

Beaches are areas where loose particles are deposited along the shore (**Figure 1-7**). Sands accumulate by erosion from terrestrial regions or from near-shore sources such as bottom sediments or coral reefs. Beach materials can range in size from boulders to sand. Typically, sandy beaches develop in areas where the slope of the beach is low; the flatter the beach, the finer the sands. For example, the low-gradient northern Gulf of Mexico coast is dominated by extensive white sandy beaches; the steeper and geologically more recent beaches along the eastern Pacific shoreline are more likely to be composed of gravel or boulders.

Figure 1-7 Beaches bordered by dunes at Gulf Shores, Alabama in the northern Gulf of Mexico; note the blowout from wave activity associated with 2004 Hurricane Ivan.

On sandy beaches, the energy from the pounding wave action causes continual movement of the sands and a constantly changing beach; in fact, beaches are sometimes referred to as "rivers of sand." Although there may not be an obvious rearrangement of the beach over a short time, if you return to a beach after several years or after a single large storm, the changes can be dramatic. This is one reason that people building homes too close to the ocean often regret it after a few years. Because of the stress from the constant water movement and the lack of nutrients in these areas, beaches are in some ways like deserts. Relatively few macroscopic organisms live their entire lives on or in the beach sands; however, many organisms do depend on beaches in various ways.

Unless there are rocks or cliffs adjacent to the beach, the wave and wind action will accumulate the sands in mounds forming **dunes** above the high tide zone. Sand dunes serve an important job of protecting the regions behind them from the action of the waves. Plants colonize the dunes and help to stabilize them. Behind the dunes, other types of vegetation that are tolerant of the wind and salt water will grow. Loss of this vegetation is a major concern for wildlife living in these areas, and should be a major concern for coastal residents because it protects them from the impact of the sea, especially during storms. Dune and beach ecology and conservation issues are discussed in further detail in Chapter 3.

Deltas

A **delta** is a land structure projecting out onto the continental shelf in a fanlike pattern where a river flows into the sea (**Figure 1-8**). The delta is formed by sediments deposited in a thick layer by the river. They only exist where sediment-laden rivers flow onto a broad continental shelf without much wave or tide action to wash away the sediments. Sediments are usually very fine because the larger, heavier particles settle out from the rivers before they reach the sea. One of the most prominent deltas is the Mississippi River Delta in the Gulf of Mexico. Along with the sediments, there are large amounts of nutrients deposited at deltas, resulting in a high biological productivity in these regions. Deltas are constantly changing, and under natural conditions will change shape or migrate back and forth along the coastline slowly over thousands of years. Human impacts on coastal rivers have dramatically altered these processes and led to coastal conservation issues (see Chapter 4).

Rocky Coasts

Rocky coasts are found where the coastline is steep and there is little sediment accumulation (**Figure 1-9**). Many of these rocky shorelines have little if any area that would be

Figure 1-8 Satellite image of the Mississippi River Delta in the northern Gulf of Mexico. (See color plate 1-8.)

considered a beach. Rocky coasts are typical in more geologically active regions along the active continental margins, often in areas with relatively recent volcanic activity. Although erosion and change occur gradually, these coasts are much more stable than beaches. The organisms that live in the **rocky intertidal** zone are specially adapted to withstand a constant pounding from wave action or frequent stranding with the change in the tide. The ecology and conservation concerns about rocky shorelines are covered in Chapter 3.

Estuaries

Estuaries are waters partially surrounded by land where fresh water and salt water meet as a river enters the sea.

Figure 1-9 A rocky coastline on the northern shore of Maui, Hawaii.

These may be associated with deltas, but estuaries are also formed by rivers that do not produce deltas. Because of the mixing of ocean and river waters, the salinity can range from fresh to full strength sea water. In isolated coves or tidal flats, evaporation may result in salinities substantially higher than those of ocean waters. These extreme and variable conditions mean that organisms living in estuaries must be adapted to withstand unique physiological stresses or have the mobility to avoid harsh conditions. Although physical stresses can limit organisms living in estuaries, they gain advantage from being in one of the most biologically productive environments on Earth due to the large input of nutrients and sunlight. Estuaries also provide protection from wave action and a diversity of habitats for shelter and protection. A large biomass and diversity of organisms, thus, are associated with estuaries, at least during some portion of their life.

The living resources of estuaries and access to the sea have attracted human settlements to estuaries for millennia. For example, the Nile Delta in Africa (**Figure 1-10**) and the lower Tigris and Euphrates Rivers have supported some of the oldest continuous civilizations on Earth. The Chesapeake Bay, where large U.S. cities have developed, is an estuary, and the waters flowing through the Mississippi River Delta form an estuary. The impact of activities associated with human settlements and the use of the biological

Figure 1-10 Satellite photo of the Nile River Delta and Estuary. (See color plate 1-10.)

production for human needs results in numerous conservation issues in estuaries. The ecology and conservation of estuaries and associated habitats will be discussed in Chapter 4.

Reefs

Reefs are made by living organisms attached or partially embedded into the bottom. Although reefs can form in temperate coastal regions or even in the deep sea (see Chapter 8), massive reefs are most prominent in clear, warm, shallow waters of the tropics as **coral reefs**. Although corals predominate, other organisms such as algae and sponges help to form the reef complex. The hard calcified skeletons of these organisms form massive structures in shallow waters. In fact, the Great Barrier Reef off Australia—considered the largest structure on Earth—is made by living organisms, comprised primarily of reef-forming corals.

Tropic coral reefs are mostly limited to shallow waters due to the dependence of the reef-forming organisms on sunlight for photosynthesis. Their structure, however, varies with the geologic history of the region in which they occur. These structures can be placed into three basic types based on their general appearance, age, and mechanism of formation (**Figure 1-11**). The first to document and propose an explanation of the variation in reef structures was Charles Darwin. He noticed in his global travels that while most reefs are associated with coastal areas, some consisted of rings of coral reef isolated in the open ocean. He developed a classification scheme for reefs into three types that are still used today. As scientists learned more about plate tectonics and island formation, Darwin's explanation of reef formation was refined to explain why these different reef formations exist. As a reef begins to form around a relatively new volcanic island, it grows in a fringe adjacent to the coastline. These are called **fringing reefs**. As oceanic plates sink as they move from the spreading center, the islands begin to sink as well. The coral organisms build the reef upward at a rate fast enough to keep the reef alive near the surface. The reef thus now encircles the island but is separated by a lagoon from the portion of the island remaining above the surface; these are **barrier reefs**. When the island eventually sinks below the water surface, the remaining circular reef encloses a shallow lagoon and is now referred to as an **atoll**. In some atolls, small islands form around the reefs. Because corals can only grow at a rate of about one centimeter per year, this evolution of reef formation occurs over an extremely long period of time; it may take tens of millions of years for an atoll to form. Some

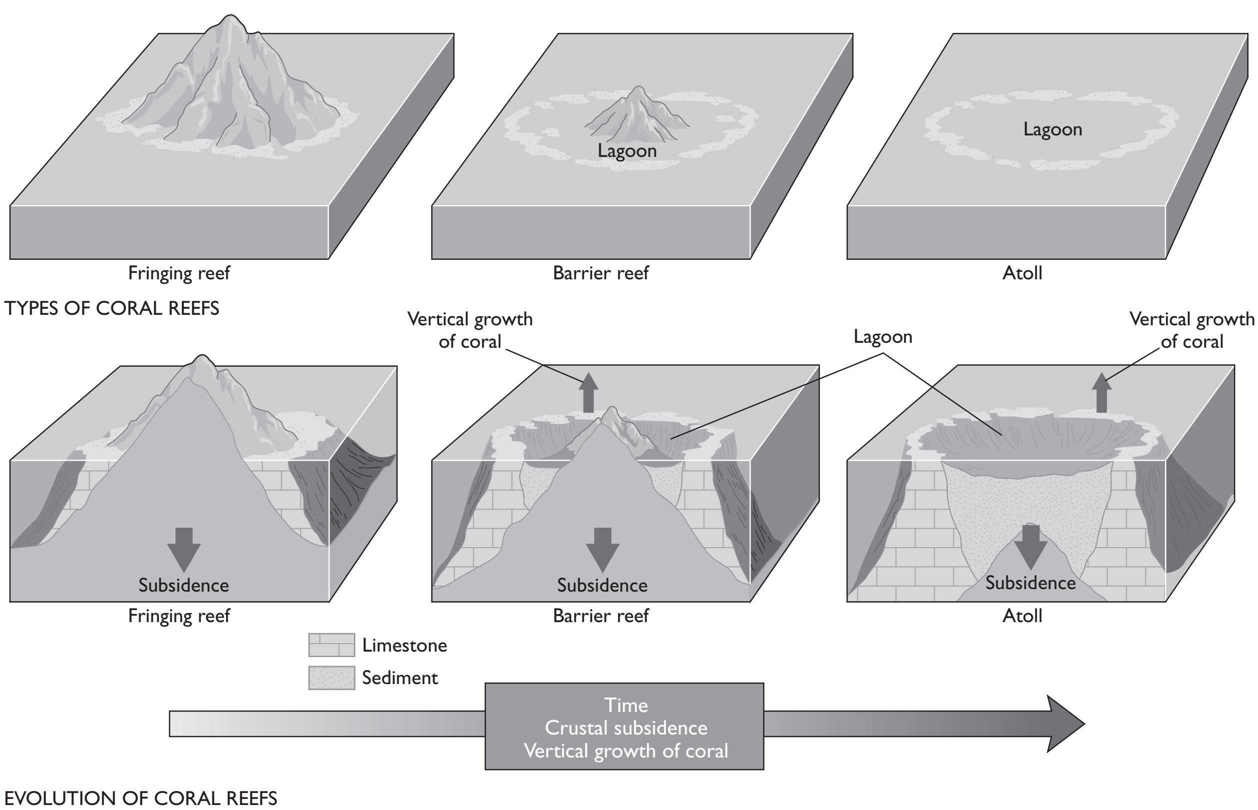


Figure 1-11 Representation of coral reef development. Fringing reefs form around new volcanic islands, developing into barrier reefs and atolls through geologic time in response to land sinking or sea level rising around the island.

scientists are now concerned that rapid global warming could result in the death of many coral reefs if coral growth cannot keep up with the rate of sea-level rise, leaving the corals in waters too deep and dark to survive.

Regardless of the class of reef formation, tropical coral reef habitats attract and support a similar assemblage of diverse organisms. In fact, coral reefs have higher species diversity than any other marine ecosystem. One reason for this high diversity is that organisms are more readily able to evolve specialized adaptations in the stable environment of the coral reef habitat. For example, physical variables such as sunlight, salinity, and temperature are relatively moderate and constant. Because many organisms on the reef have evolved in response to this stability, coral reefs can be very sensitive to disruptions. Human actions disrupt this stability in many ways. By blocking sunlight penetration with excess sediments, modifying temperature through global climate change, or removing major components of the ecosystem by excessive fish harvest, the coral reef ecosystem rapidly begins losing its biological diversity. Tropical coral reef ecology and conservation are discussed in Chapter 5.

Islands

Islands are non-continental land masses projecting above the surface of the water. The size of islands can vary considerably. In the strictest sense continents might be considered islands, and a boulder projecting above the sea surface also might be considered an island. Islands are typically considered as non-continental land masses of intermediate size, however, and often are classified according to the processes by which they are formed.

Barrier Islands

Barrier islands are relatively narrow bars of sand that form parallel to the shore along shallow gently-sloping coastlines. In order for barrier islands to form, there must be adequate sands and the forces of waves, tides, and currents must be strong enough to move and accumulate the sands. They will not form, however, where tides and currents are extreme. The proper combination of physical conditions is present for the formation of barrier islands along about 13% of the world's continental margins. Not only are barrier islands limited geographically, they also are more prevalent during some global climate conditions than others. Although some barrier island are simply accumulations of sand on submerged bars offshore, most are formed under special conditions during periods of rising sea levels. As sea levels rise along sandy shorelines, the waters break through the sands to form a lagoon behind the dunes. Eventually these dunes become islands separated from the mainland by open water. During periods of rising sea level, wave action causes the islands to migrate toward the mainland. Many current barrier islands actually began forming during the last major sea-level rise about 6,000 years ago.

One of the most extensive series of barrier islands is along the U.S. Atlantic and Gulf of Mexico coasts, where there are almost 300 separate barrier islands, with a combined length of over 4,000 km. These islands are continually changing as a result of current and wave action. Hurricanes or large storms will often overwash smaller islands, form new inlets, or even split them into separate islands. New islands can be formed from **spits** (sand barriers joined to the coast). For example, barrier islands along the Mississippi/Alabama Gulf of Mexico coast are regularly split or combined, especially in association with major hurricanes. In 2005, for example, Asbury Sallenger and colleagues reported substantial changes after a single event, Hurricane Katrina (**Figure 1-12**). Despite this instability, humans have connected many of the barrier islands to the mainland by bridges and developed the islands as cities or resort communities. These include Atlantic City, New Jersey; Miami Beach, Florida; Galveston, Texas; and Pensacola Beach, Florida. It can be a constant battle for residents of these islands to avoid loss of their homes or businesses to the actions of the sea, and no one can predict where the next hurricane might devastate developments on one of these islands. Fortunately, the U.S. barrier islands that have escaped development have been largely protected in the past few decades. Issues concerning conservation of barrier island habitats are discussed in Chapter 3.

Volcanic Islands

Most of the largest islands, as well as those found away from the continental margins, are formed by volcanic activity. As mentioned previously, this volcanic activity is usually associated with the convergence of oceanic plates with continental plates during seafloor spreading or the formation of hot spots in the middle of plates. Many volcanic islands are only a small portion of the volcano, with much more found under the water's surface. Volcanic islands formed at convergence zones are typically arranged in an arc adjacent to ocean trenches. These island chains are easily recognized on maps because of their shape (**Figure 1-13**). They include the Western Antilles in the Caribbean, the islands of Japan and the Mariana Islands in the western Pacific, and many others. These islands include some of the most active volcanoes on Earth; however, many are remnants of long-extinct volcanoes.

Volcanic islands that are formed in the middle of oceanic plates are also found in linear chains but are not near spreading centers or convergence zones. These are created by hotspots that push volcanic material to the surface. The classic example of this type of formation is the Hawaiian Island chain (Figure 1-13). These islands are formed in line

(a)

(b)

Figure 1-12 Satellite photos of barrier islands in the northern Gulf of Mexico **(a)** on October 15, 2004 and **(b)** taken on September 16, 2005 after Hurricane Katrina. Note that Dauphin Island, Alabama (second island from right) has been split in two and that Ship Islands (starting second from left) are substantially smaller. Dauphin Island also migrated landward, leaving some oceanfront homes in the sea. Petit Bois Island, 8 miles to the west, was a part of Dauphin Island 150 years ago. (See color plate 1-12.)

Figure 1-13 Satellite view of the Hawaiian Islands, formed by volcanic activity as the Pacific Ocean plate slides over the mid-ocean hot spot currently located beneath the island of Hawaii (lower right); the oldest islands (Kauai and Niihau) are in the upper left. (See color plate 1-13.)

because the hotspot creating them remains relatively stable while the oceanic plate passes over it. A series of islands thus is created over the millions of years it takes the plate to pass over the hotspot.

The Ocean Water Column

The waters of the oceans comprise vast, largely unexplored realms that continue to fascinate the human mind. Other than the thin layer at the surface, most of the waters of the oceans have been accessible to humans only in the past few decades. Even today much of the deep ocean is still relatively unexplored. The depths of the ocean are extreme from a human perspective, the most extreme being about 11,000 meters below the ocean surface, in the Mariana Trench of the western Pacific. The deepest in the Atlantic Ocean is the Puerto Rico Trench at 8,400 meters. Although these deep trenches compose only a relatively small area of the sea floor, the average depth of the oceans is still remarkable at about 4,000 meters. Most of the biological production in the ocean is limited to the upper 200 meters of the water column, where nutrients, oxygen, and light are most available. Biological production and oxygenation originating in surface waters, however, feed most of the biological activity in the deep sea, down to the deepest basins and trenches. These deep-sea regions have only recently been explored; development of technology that allows us to access these regions means that these are exciting times in ocean research. Still, the deep ocean remains as the last largely unexplored frontier on the planet. It remains to be seen whether increases in accessibility will lead primarily to conservation or to excess exploitation of deep ocean resources. These issues are discussed further in Chapter 8.

Ocean Zonation

For convenience and delineation of habitats, the ocean can be divided into separate zones according to the distance from shore and depth of the water (**Figure 1-14**). The divisions between zones are not precise lines, and there is typically no sharp transition from one zone to the next. The zones do give a useful reference, however, and some general characteristics of these zones are fairly consistent.

Moving horizontally from the coast, the **intertidal zone** (also called the **littoral zone**) is the region above the low tide line that is regularly covered and exposed by the movements of the tides. The **neritic zone** is the region over the continental shelf, from the intertidal zone offshore to the shelf break. This is the region of the highest biological productivity, primarily as a result of nutrient inputs from the continents and exposure to adequate sunlight for photosynthesis. High production and accessibility by humans results in high exploitation of resources from this region. The **oceanic zone** includes the remaining waters of the seas beyond the waters above the shelf break. In terms of biological production, this region is like a desert relative to the neritic zone. But because this ecosystem is larger than any other on Earth, there is a great diversity of organisms and large amounts of biomass. Virtually no region in the oceans is devoid of life.

The ocean waters also can be divided into vertical zones, classified according to their distance below the

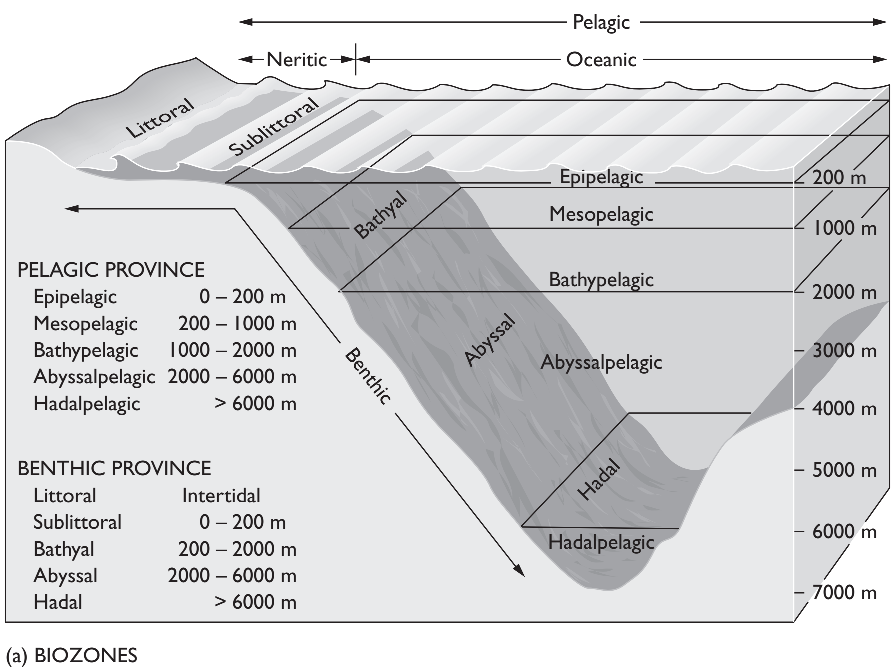


Figure 1-14 Ocean regions established to discriminate biological zones based on water depth.

ocean's surface. The sea floor is referred to as the **benthic region**, and the bottom habitat itself is called the **benthos**. The water column of the ocean is called the **pelagic zone**. The **epipelagic** generally corresponds to the **photic zone**, the region with adequate light penetration for photosynthesis. This is typically 100 to 200 m depth in the clear open ocean. The **mesopelagic zone** is the mid-region of the open ocean, from the photic zone down to about 1,000 meters depth. This is sometimes called the *twilight zone* because of the low levels of light. Many of the organisms in the mesopelagic depend on the sinking of food particles or nutrients from waters of the photic zone. Other organisms migrate to the surface at night to take advantage of the productive photic zone waters. The deep sea below 1,000 meters is divided into three zones. The **bathyl zone** (or bathypelagic when referring to the water column) is down to 4,000 m; the **abyssal** (or abyssopelagic) zone is below 4,000 meters to about 6,000 meters. The **hadal** (or hadopelagic) zone includes the deepest ocean trenches, down to 11,000 meters depth.

1.3 Water Movements: Wave, Tides, Currents

Coastal and seafaring civilizations have been studying waves, tides, and currents for practical purposes for thousands of years. The knowledge gained enhanced the abilities of humans to travel the oceans and exploit its resources. Scientific advancements over the past century have dramatically increased our understanding of these water movements so that information that used to require the accumulation of more than 100 years of knowledge now can be obtained almost instantaneously (for example, using satellite imagery). An understanding of the water movements not only allows us to navigate the ocean and exploit its resources but also gives us an understanding of how they influence marine organisms and ecosystems.

Waves are the most visible type of water movements. Waves differ from currents in that the water forming the wave does not move at the speed of the wave; it is the movement of energy that produces waves. This is why an object floating just offshore moves only slightly toward shore with each passing wave. The energy that drives the wave can come from various sources including the wind, movements of the Earth (e.g., earthquakes), gravitational forces, and atmospheric pressure. These energy sources are the basis for classifying waves.

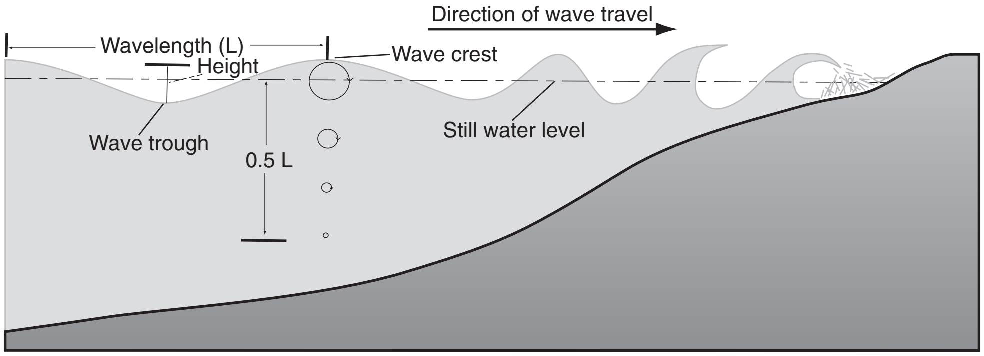
Wind Waves

Wind waves are disturbances in the ocean surface formed by the transfer of wind energy to the water (**Figure 1-15a**). Wind waves are typically formed in an undulating pattern with the waters rising to a **crest** (the highest point on the wave) and falling to a **trough** (the wave's lowest point). The size of the wave is defined by the **wave height**—the vertical distance from the crest to the trough—and the **wavelength**—the horizontal distance between adjacent crests. The **wave period** is the time it takes adjacent wave crests to pass a fixed point. Wind waves can vary tremendously in size, depending not only on the wind speed but also the wind **duration** (or how long it blows over the sea), and the **fetch** (the distance over which the wind blows).

Large waves may originate from distant winds; winds of greater duration and speed impart more energy to the water and make larger waves. Until the 1930s it was believed to be impossible for a wind wave to reach heights over 20 meters. This was challenged when a U.S. oceanographer documented waves higher than 30 meters during a long duration storm in the Pacific Ocean. Wave measurements during Hurricane Katrina in 2005, with winds up to 150 mph, showed that large waves are not as uncommon as once believed; waves over 18 meters were not uncommon and 25-meter waves were observed. Oceanographers have predicted that hurricane waves could reach heights as great as 40 meters. Although these waves are caused by storms or other unusual winds, extreme waves could occur miles from the actual storm. Sometimes single unusually large waves occur amongst average size waves in the open ocean. These **rogue waves** are surprisingly large for the sea conditions in which they occur. Though unusual, such waves over 25 meters appear to occur regularly in the open ocean. Their mechanism of formation is still debated, but they seem to often occur in deep water where there is a convergence of physical forces such as strong winds and fast currents, and may be a result of a consolidation of multiple smaller waves.

Regardless of what the maximum height of an ocean wave proves to be, you will not see wind waves approaching these heights near shore, especially in shallow coastal waters. This is because wind waves break as they move into shallow waters (Figure 1-15b). The motion of the wave actually extends below the surface to a distance of about 50% (½) of the wavelength; therefore, as the wave moves into shallower water it "feels" the bottom. As the lower portion of the wave begins to slow, the top of the wave continues forward making the wave steepen until it breaks. A wind wave breaks when its wave height is about 75% (¾) of the depth of the water. For example a 75-centimeter wave would break when it reaches waters of about one-meter depth; a three-meter wave would break in four-meter deep waters. Higher water levels come ashore and cause damage during hurricanes due to the combination of an increased water depth from the **storm surge** with the higher wind waves that form in these deeper waters.

(a)



(b)

Figure 1-15 **(a)** A wind wave breaking as it moves ashore. **(b)** The pattern of water motion in a wind wave approaching shore.

As waves break onshore their energy is transferred to the bottom. Because beaches are often composed of loose unconsolidated sands, they are easily moved by waves; this is why beaches are so unstable. Waves usually do not move onto shore directly parallel to the shoreline but approach at an angle, resulting in gradual movement of sediments alongshore down the beach in a process called **beach drift**. Although the movement of sand caused by one wave is not significant, the cumulative effect of continuous wave action can be substantial. Conservation issues resulting from human interactions with beaches and wave activity are discussed in Chapter 3.

Tsunami

The largest and most damaging waves in the ocean are formed not by the wind, but by geologic activity under the sea. These have often been called *tidal waves*, but because they are not a result of tidal activity this is a misnomer. Now the more acceptable term for these waves is **tsunami** (from a Japanese word meaning *harbor wave*). Tsunamis are the result of undersea earthquakes or volcanoes, or coastal landslides rapidly displacing the water. Tsunami waves can be extremely long, with wavelengths of over 160 km, and move extremely fast, at speeds over 640 km per hour. At this speed a tsunami could move from Alaska to Hawaii in about five hours. The height of these waves in the open ocean may only be a meter or two; thus, their steepness is so low that they would hardly be noticeable to a boat under which the tsunami passes. This is the reason that tsunami waves are so difficult to detect in the open ocean, requiring sensitive instruments to monitor changes in the water level. If the wave height is so low, then how can it cause so much damage when it reaches the shore? As the tsunami slows coming to shore, its long shallow waveform is translated into a narrow and much higher wave. Tsunamis can reach heights of over 30 meters as they hit the coastline. The waves appear to an observer as a fast flood of water rushing ashore, not as a breaking wave commonly imagined.

The December 2004 tsunami in the Indian Ocean, caused by one of the largest earthquakes in decades, 240 km off the coast of Sumatra, created waves as high as 20 meters moving onto shore at speeds as high as 800 km per hour (**Figure 1-16**). This tsunami resulted in the deaths of over 280,000 people and damages totaling billions of dollars. Although damaging tsunamis are very infrequent in a single region, globally there is a tsunami that causes major damage and loss of lives on average about every seven years. For example, in the 50 years from 1925–1975 there were at least seven, including tsunamis in Russia, Alaska, Hawaii, Chili, and Newfoundland, Canada. There is nothing humans can do to eliminate tsunamis or predict their occurrence with any accuracy, but we can minimize their impacts by maintaining natural vegetation and structures along coastlines (see Chapter 4). The increased deployment of monitoring systems in the open ocean will increase our ability to warn coastal residents of a tsunami as it develops, but whether there is adequate time to respond depends on the distance from the wave's point of origin. The 2010 tsunami in Japan, resulting from an unprecedented earthquake approximately 70 kilometers off the coast, began moving ashore within about 30 minutes, giving little response time for avoidance; however, advanced earthquake warning systems likely saved the lives of thousands of people. Although there were few natural structures in this heavily developed region of Japan to buffer the wave, tsunami protection walls were in place; however, the unexpected height of the storm surge, over 30 meters in some locations, overtopped walls and inundated tsunami protection shelters.

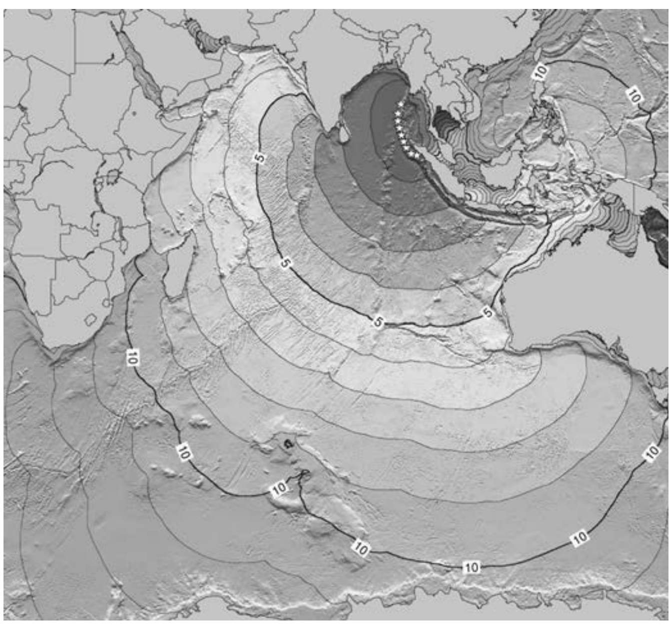


Figure 1-16 Movement of the tsunami wave produced by a magnitude 9.4 earthquake off Sumatra, Indonesia, on December 26, 2004, progressing in time from the epicenter (indicated by stars), from lighter to darker grays. (See color plate 1-16.)

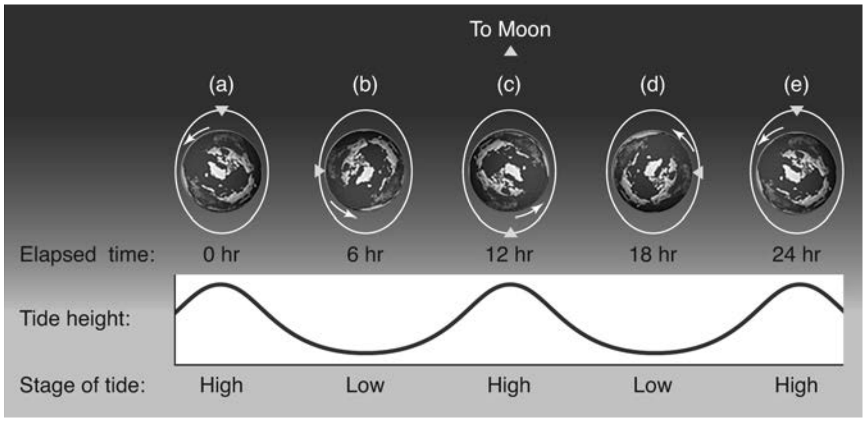
Tides

Technically speaking, tides are types of waves. They are not typically recognized as such because they rise and fall over a period of 12 or 24 hours and they have extreme wavelengths, as long as half the circumference of the Earth. Tides are periodic waves caused by the gravitational force of the Sun and Moon on the Earth, and the motion of the Earth (**Figure 1-17a**).

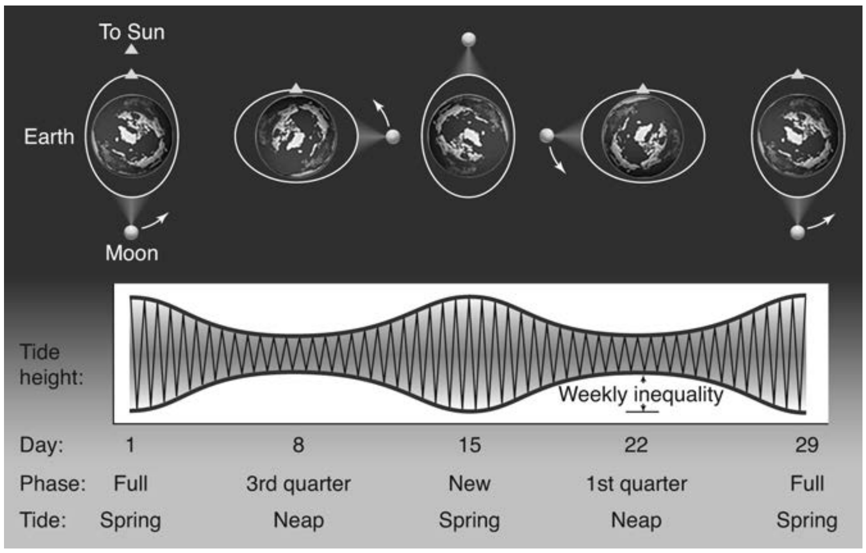
Tides vary over a diurnal (daily) cycle due to the Earth's rotation and a monthly cycle resulting from changes in the Earth's position relative to the Moon and Sun (Figure 1-17b). **Spring tides** are periods when the tide range is highest because the Sun, Moon, and Earth are in line (during new and full Moons) and the gravitational attractions of the Sun and Moon are combined. **Neap tides** occur when the Sun, Moon, and Earth form a right angle (first and third quarter Moons). Tides serve an important function in transporting nutrients, wastes, and organisms into and out of coastal inlets. Tidal currents can flow through estuaries as fast as 40 km per hour or more in some regions. Tidal currents differ from wind-driven currents in that they switch directions as the tides move in (**flood tide**) and move out (**ebb tide**) of an inlet or coastal area. In some coastal areas (i.e., the U.S. east coast) there are complex networks of tidal creeks miles inland along the coastline. The biological importance of these areas and conservation issues are discussed in Chapter 4.

Tides are observed as the daily rising and lowering of the sea level on a periodic cycle. Due to a number of factors, such as the shape of the ocean basin, the frequency, and height of tides vary around the globe (**Figure 1-18**). On some coastlines the tides are **diurnal**, with one high and one low tide per day. Regions with diurnal tides include the northern and western Gulf of Mexico, the north Pacific, and southwest Australia. On other coastlines there are two high and two low tides per day. These **semidiurnal** tides occur on the east coast of North America, most of the African and European coastlines, and much of the South American coastline. Along some coastlines the tides are mixed, having two cycles per day, but successive tides are of different heights, such as on the west coast of North America and in much of the western Pacific.

The **tidal range**, or difference in the level of high and low tides, varies substantially around the Earth. This is due to a complex interaction of various physical factors, including the shape and depth of the basin, coastline, or bay where the tide is observed. The typical tidal range is somewhere between 30 centimeters and 2 meters; however, the greatest tidal range is over 15 meters, in the Bay of Fundy on the



(a)



(b)

Figure 1-17 Tidal cycles. **(a)** Diurnal cycles in tides caused by the position on the Earth relative to the tidal bulge (arrow indicates a fixed position on the Earth). **(b)** Monthly change in tidal range due to the relative position of the Earth, Moon, and Sun; maximum (spring) tides occur when the Earth, Moon, and Sun are in line.

Atlantic coast of Canada. This variation in tides can have a large influence on organisms that use the tides for movement in and out of coastal areas. It also affects conservation issues through the mixing and transport of materials such as organic matter, nutrients, and pollutants, especially in areas where there is little river flow into the coastal region. These issues are discussed in Chapters 3 and 4.

Circulation Patterns

Surface Currents

Water and energy also can move through the ocean as **currents**. Whereas waves are primarily a movement of energy, currents are masses of moving water. **Surface currents**, driven mainly by surface winds, are layers of water flowing within the surface waters of the oceans; they can reach to depths as great as 400 meters. They serve an important ecological function by transporting drifting marine organisms, organic matter, and nutrients; they enhance the movements of migratory organisms.

Viewing a map of wind-driven ocean surface currents (**Figure 1-19**), a pattern of circular **gyres** is apparent as flows around the ocean basins of the northern and southern hemispheres. These gyres are primarily due to the **Coriolis Effect**, resulting from the forces imparted on the ocean due to the Earth's rotation. In the northern hemisphere, the Earth's rotation (toward the east) appears to deflect the flow to the right of the wind direction. The water continues to veer to the right around the basin resulting in a clockwise rotation. Not only is this obvious throughout the major ocean basins of the Atlantic and Pacific, but smaller gyres occur in smaller basins such as the Gulf of Mexico. (It is a myth that this rotation can have a significant effect on

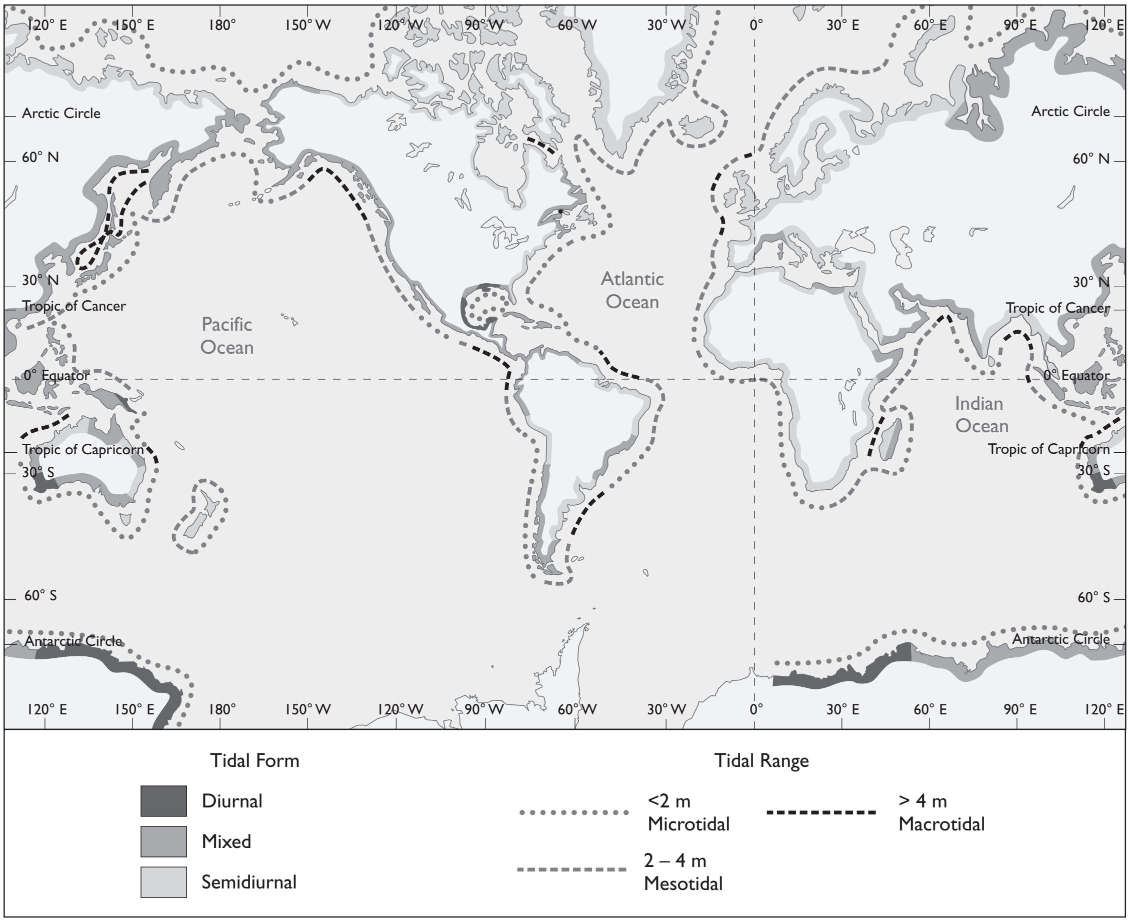


Figure 1-18 Global variation in tidal forms and ranges.

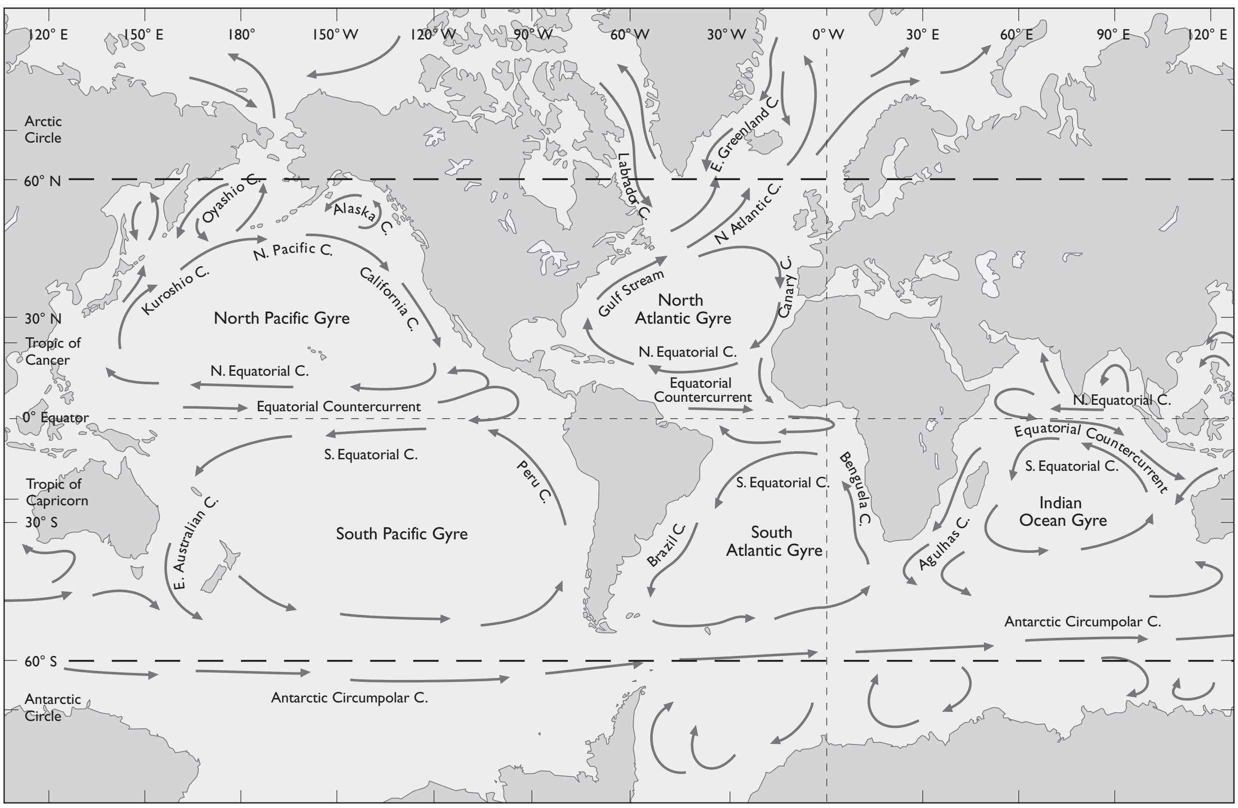


Figure 1-19 Global wind-driven surface currents indicating major circulation gyres.

water sitting in a small basin such as a tub or a toilet; other forces acting on the water greatly overwhelm the Coriolis force at this scale.) In the southern hemisphere the opposite occurs; currents are deflected to the left, resulting in a counterclockwise rotation. Wind patterns and deflection of currents by land masses also contribute to the formation of these gyres. Navigators of ocean vessels have taken advantage of these circulation patterns for centuries. It's not a coincidence that many early explorers followed similar paths around the seas; that's the direction the currents (and winds) carried them. Marine organisms take advantage of this circulation, riding the currents to spawning, feeding, and nursery grounds.

The currents flowing along the coastlines form one component of ocean circulation patterns. The **western boundary currents** (found on the west side of the ocean basins, along the east coast of the continents) are the fastest and deepest of the currents making up the ocean gyres. These currents tend to carry warm waters from equatorial regions toward the poles. Although this has been known for hundreds of years, only recently have we been able to get detailed surface temperature maps using satellite imagery (**Figure 1-20**). The warm waters flowing northward along the east coast of North America form the Gulf Stream, the largest of the western boundary currents. This current has an average width of 70 km, travels about 8 km per hour, to a depth of around 450 meters. It has important influences on fisheries along the U.S. Atlantic coast. For example, as it flows adjacent to the east coast of Florida it attracts oceanic fishery species into waters close to shore, and continues to influence marine populations as it moves north. The Gulf Stream also has a moderating effect on climate along the U.S. east coast due to its influence on air temperatures. If this current did not exist the climate along the eastern North American coastline would be much colder (see Box 1-1, Conservation Concern: Global Warming and the Ocean).

Eastern boundary currents flow along the east side of ocean basins (along the west coastlines of the continents). These are the opposite of the western boundary currents in

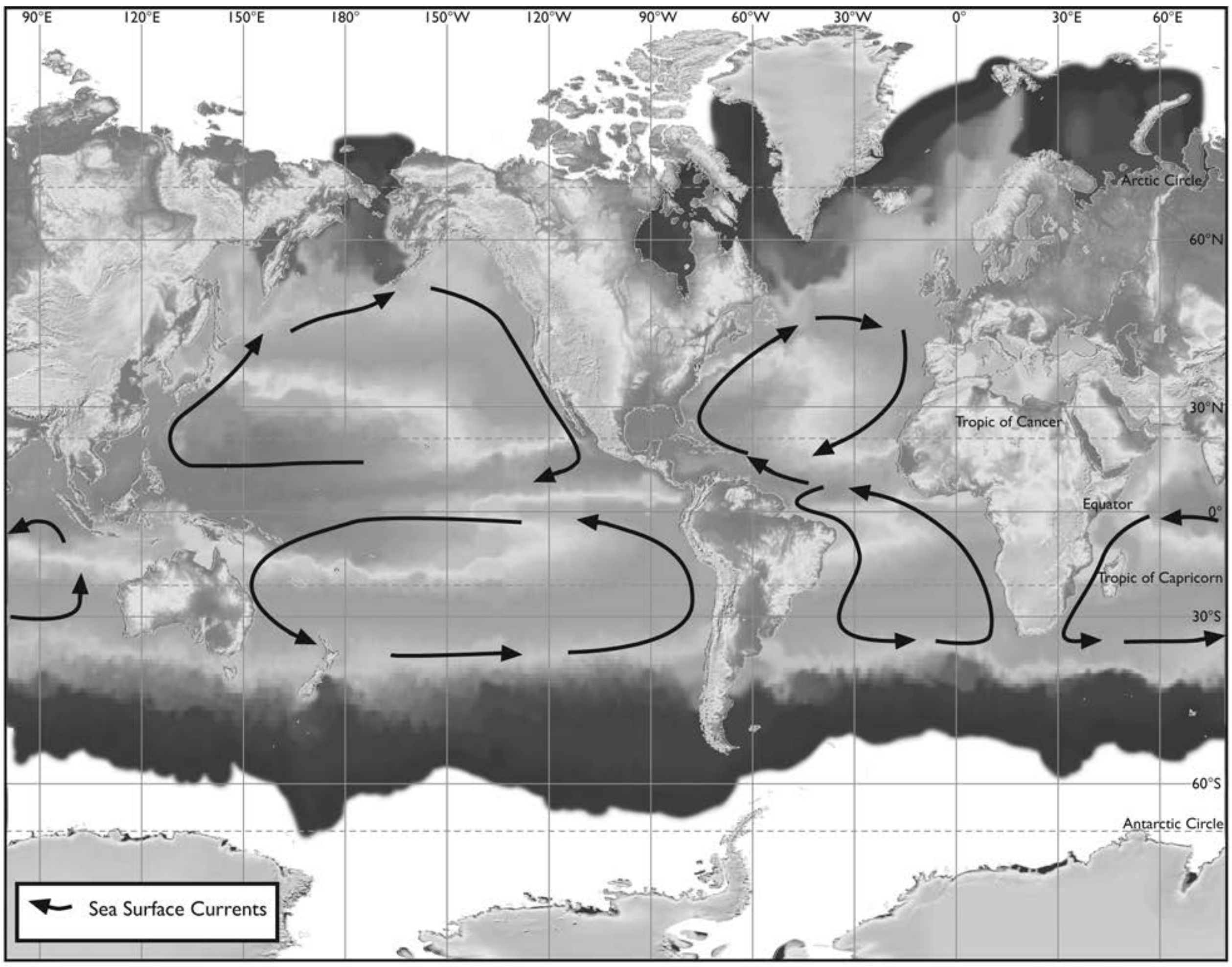


Figure 1-20 Global ocean surface temperatures during summer in the northern hemisphere. Grays indicate temperatures, with lighter gray being warmer and darkest grays being cooler. (See color plate 1-20.)

many ways. They tend to be smaller, weaker, broader, and shallower than western boundary currents. Flowing from the poles toward the equator, eastern boundary currents carry cold water southward in the northern hemisphere. For example, cold waters are transported well down the U.S. west coast. This is why, at comparable latitudes, waters along the U.S. east Pacific coast tend to be substantially colder than those along the Atlantic coast. The Peru Current in the South Atlantic is a southern hemisphere eastern boundary current with similar characteristics (but flowing equatorward from south to north).

Coastal Upwelling

Of course, the ocean is not a one dimensional plain. Along with the surface currents there are vertical components to ocean circulation. Coastal **upwelling** is one component of vertical circulation that influences conservation issues and is discussed in later chapters. Coastal upwelling can most easily be understood by looking at it in a stepwise fashion as it would develop (**Figure 1-21**). A prime example is the Peru Current, an eastern boundary current flowing along the Pacific coast of South America. As the surface currents flow northward along the shore, the waters are deflected offshore away from the coast due to surface winds and the Coriolis effect (to the left of the current direction in the southern hemisphere). As these waters are pushed away from shore, deeper waters are pulled up to replace them. This is the process of upwelling. These waters can be recognized by their temperature signature; the cold **thermocline** (a layer of relatively cold water) is extended to the surface along the coast. The waters upwelling from depth have high nutrient levels, resulting from the accumulation of organic matter that sinks from the productive surface waters toward the bottom. Because there is relatively little biological activity in deeper darker waters to use the nutrients, they accumulate until being drawn back to the surface. The nutrients upwelling to the surface support one of the most biologically productive regions and one of the most valuable fisheries in the ocean. A similar, but less pronounced pattern can be observed in other coastal upwelling zones, such as off portions of the coast of the western United States and western Africa. Later chapters will discuss the significance of these upwelling zones relative to ocean fisheries.

The pattern described above is the most typical circulation pattern along the western coast of South America. Sometimes this pattern is modified, however. For example, a change in surface winds can interfere with the surface currents, which disrupts the upwelling. In the tropical Pacific such a change occurs about every three to eight years, for reasons that are not completely understood. This change initiates events referred to as ***El Niño*** (Figure 1-21b). The effects of El Niño events have long been noticed off the coast of South America, because this is when upwelling practically stops, which reduces nutrient levels and leads to dramatic declines in fish populations (see Chapter 11). These events were named El Niño ("the Child") by Peruvian fishers because they are usually initiated around the Christmas season. Although their immediate impact is a change in weather and ocean circulation in the eastern Pacific, El Niño events have many effects on oceans and weather around the world. One of these, discussed in Chapter 5, is the increase in tropical water temperatures that impacts coral reef ecosystems.

Vertical Circulation Patterns

Ocean water circulates vertically not only along the coastlines but also in the open ocean (**Figure 1-22**). This vertical circulation, presented as The Great Ocean Conveyor by Wallace Broecker, is driven primarily by water density differences and is called **thermohaline** ("temperature-salinity") circulation. Saltier or colder water is denser and therefore tends to sink. Waters driven to the poles by surface

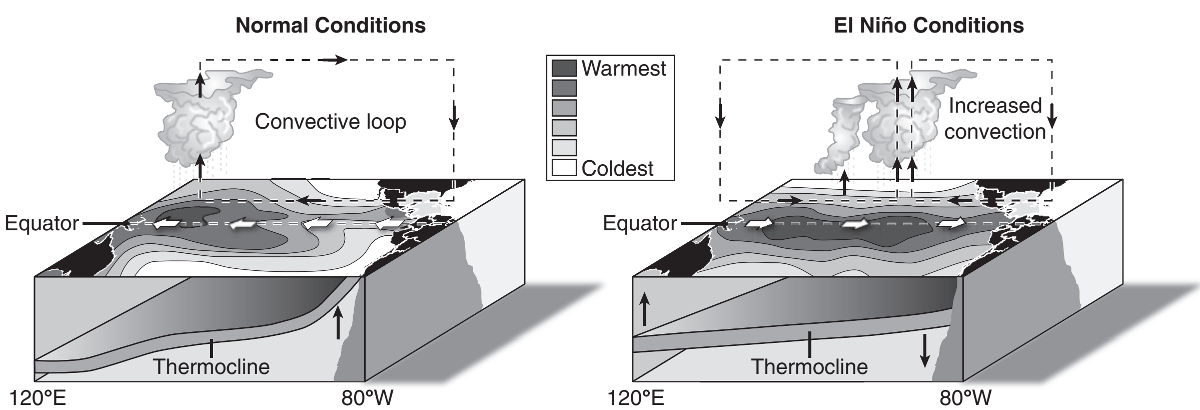


Figure 1-21 Coastal upwelling circulation off the Pacific coast of South America; grays indicate temperature, from darkest gray (warmest) to white (coldest). Left: During normal conditions the thermocline extends to the surface along the coast due to upwelling of deep waters. Right: During El Niño conditions; upwelling stops and the thermocline remains in deep waters. (See color plate 1-21.)

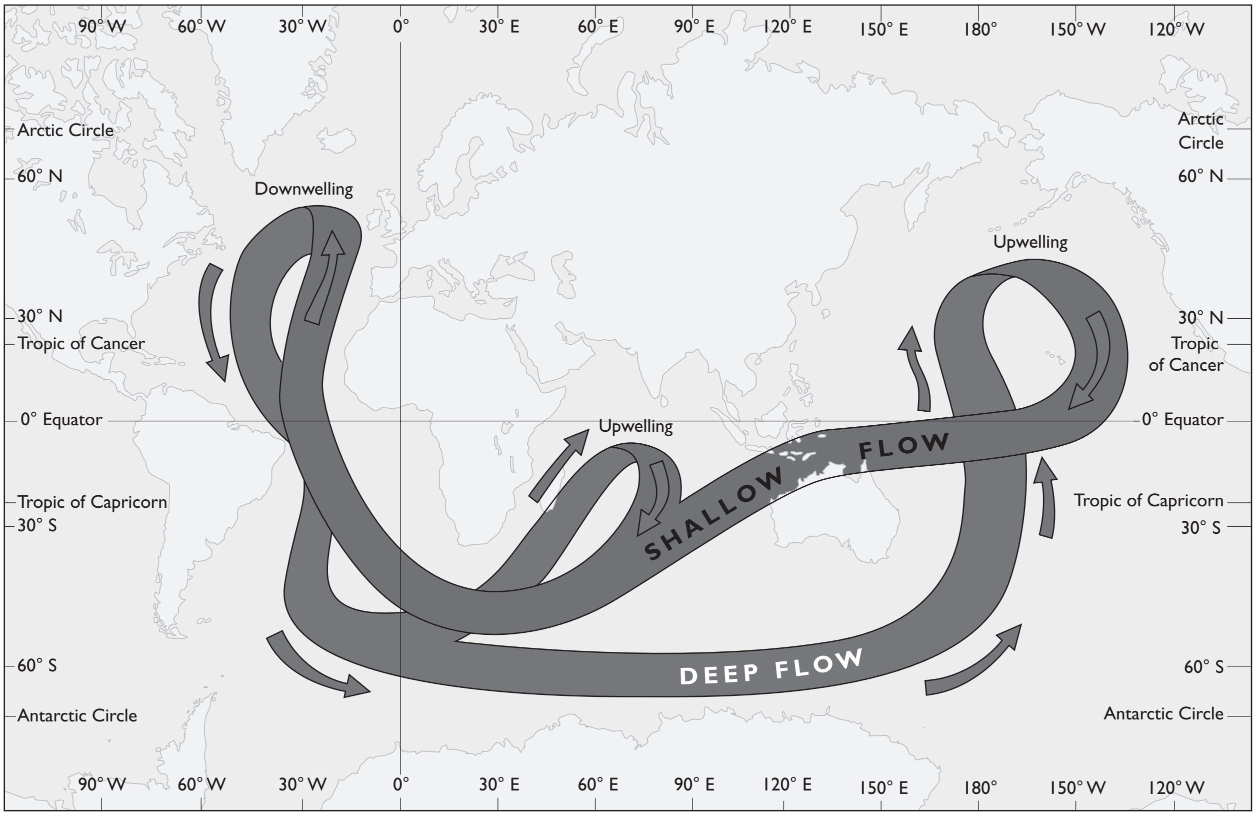


Figure 1-22 Global "conveyor belt" thermohaline circulation pattern between the surface and deep oceans.

circulation will tend to sink toward the bottom of the ocean as they get colder. These waters then slowly move across the deep sea floor toward the equator. Ecologically this is important because these waters carry dissolved oxygen throughout the deep open basins (see Chapter 8). In fact, without this circulation, the deep ocean could not support oxygen-dependent life, including all multicellular organisms. Eventually these bottom waters upwell to the surface at various locations. They are important in bringing nutrients to the surface in equatorial regions, making these tropical waters very productive. This deep-sea circulation is much slower than the surface circulation pattern discussed earlier. Most deep ocean bottom waters take 200 to 300 years or more on average to rise to the surface. Compare this to a bit more than a year it takes water in the North Atlantic surface current gyre to make a complete circuit. Although these water masses generally maintain some integrity, there is much mixing among water masses throughout the ocean. There is also seasonal and year-to-year variability in all oceanic circulation patterns. Many fear that changes in global climate could disrupt or change this circulation pattern and impact water exchange between the surface and the deep sea, with dramatic impact on deep-sea ecosystems (**Box 1-1. Conservation Concern: Global Warming and the Ocean**).

1.4 Sea-Level Changes

For the average person it is difficult to envision dramatic changes in sea levels because few have witnessed noticeable changes along ocean coastlines. It is well documented that in the past sea levels have been much higher and lower than they are at present, however (**Figure 1-23**). For example, during the last ice age, about 18,000 years ago, sea levels were over 120 meters below their current level. In fact, for most of the past 200,000 years the sea levels were at least 30 meters lower. And three million years ago sea levels were 35 meters higher than at present.

Global Sea Level Changes

Variations in global sea level are determined by measuring the mean sea level in relation to relatively stable continental coastlines. The major causes of variability in global sea level are changes in global climate. The most obvious factor is the conversion of ice water in glaciers and the polar ice caps to liquid water that flows into the oceans. When global temperatures are colder, more water is tied up in glaciers; therefore, sea levels are lower. When global temperatures are warmer, less ice is frozen; therefore, sea levels are higher. However, another important factor is **thermal expansion**. At temperatures above 4°C water expands as it is warmed.

Box 1-1 Conservation Concern: Global Warming and the Ocean

Possibly the most debated environmental issue in the world today is global climate change. Debates have centered on questions concerning: whether we are truly in a global warming trend, whether human activities are causing or influencing these changes, our capabilities to reverse recent trends, and what impact climate change will have on the Earth and its ecosystems, including the marine environment. An authoritative source of information on climate change is the Intergovernmental Panel on Climate Change (IPCC; http://www.ipcc.ch/), an international group of scientists established in 1988 through programs of the United Nations. One of the charges of this panel is to assess scientific documents from around the world and provide summaries of these documents. IPCC documents indicate that assessing and predicting global warming trends are complex processes. Because temperatures fluctuate seasonally, annually, and on longer term cycles, a few warmer years cannot be taken as proof of a long-term trend. On the other hand, a few cooler years cannot be taken as proof that global warming is *not* occurring. Decisions must be made concerning where to measure temperatures for documenting climate change, because they could be increasing in one place while decreasing in another. Consistent methods must be applied to enable comparisons of historic and recent temperatures; standard established criteria enable such comparisons over about the past century.

Comparisons of global average air temperatures near the Earth's surface show an increase of about 0.5 to 1 degrees Celsius during the 20th century (**Figure B1-1**). In the past 30 years, however, the temperature rise has averaged about 0.2°C per decade. Although this may appear trivial, it represents a large amount of energy globally and some regions have warmed faster than others. It is estimated that the Earth is within 1°C of the highest temperature in the past one million years. Although some of the popular news media may give the impression that there is wide disagreement among scientists, a 2008 scientific poll of over 3,000 Earth scientists showed 90% agreement that there has been a significant increase in mean global temperatures in the past 200-plus years.

The next critical question addressed by the IPCC is whether humans are influencing climate change. The conclusion of scientific studies is that most of the observed increase in temperature since the mid-1900s was caused by increased input of **greenhouse gases** (those that have an effect of trapping heat from solar radiation in the atmosphere, primarily water vapor, carbon dioxide, and methane), and that the primary source of the increase was carbon dioxide from fossil fuel burning and deforestation. Since the mid-1700s, carbon dioxide in the atmosphere has increased by over 30%, to levels not experienced on Earth for hundreds-of-thousands or possibly millions of years (**Figure B1-2**) (methane has increased by almost 150%). Carbon dioxide is estimated to contribute more than 60% of the warming attributed to greenhouse gases. Despite these data, scientists and political leaders debate as to the degree of influence these increases have on global warming. However, 97% of polled research climatologists—those scientists who would have the greatest expertise in climate change—agreed that humans have played a role in global climate change (fewer than 60% of the general public agreed).

The next question, whether we are capable of reversing recent global warming trends, is a social and political question as much as a scientific one. Decreasing deforestation and the burning of fossil fuels is physically possible, but politically difficult to initiate when it is viewed as causing economic hardships or as unnecessary. Recent agreements among climate

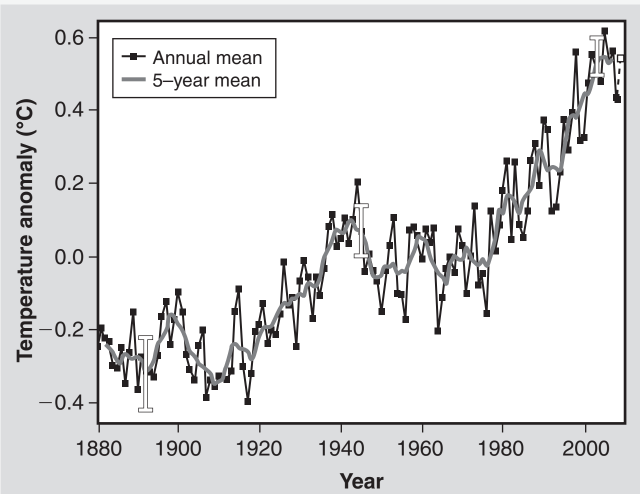


Figure B1-1 Global annual temperature anomaly indicating change in temperature from 1880 to 2009. The time period from 1951–1980 is used as the zero point for reference. The dotted line indicates the annual mean and the solid line is the five-year mean. White bars show the range of uncertainty. Data are compiled from surface air measurements at meteorological stations, ocean ships, and satellite measurements.

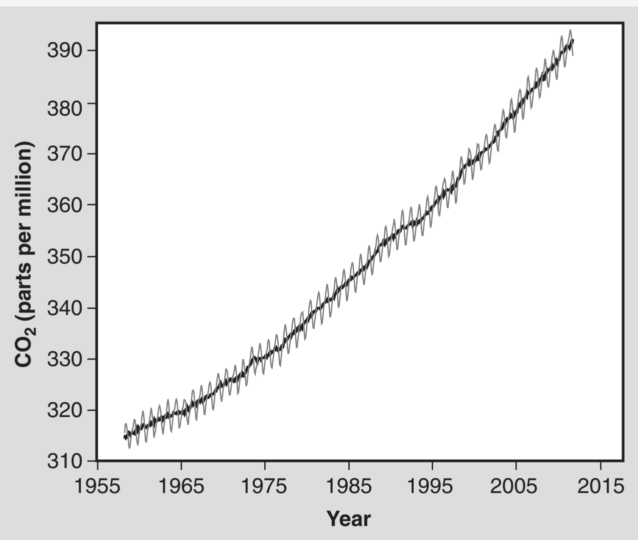


Figure B1-2 Plot of carbon dioxide record at Mauna Loa, Hawaii, by Scripps Institute of Oceanography (1958–1974) and the National Oceanic and Atmospheric Administration (1974–2011). The curves show monthly averages in parts per million (ppm). The solid line indicates annual averages.

scientists have led to increased actions around the world. The primary international agreement dealing with actions to reduce greenhouse gas emissions is the Kyoto Protocol, negotiated in 1977. Over 160 nations have ratified the treaty; the only industrialized nation that has not is the United States. Most nations, including the United States, however, have taken some measures toward moderating emissions. It is widely debated whether these actions will be enough to decrease global warming.

Other methods of reducing carbon dioxide levels in the atmosphere are being researched, including **carbon sequestration**, which is capturing carbon dioxide from burning of fossil fuels and storing it underground or elsewhere. Encouraging increased plant production is a more natural way of sequestering carbon dioxide from the atmosphere. This can be accomplished by planting trees or possibly by encouraging growth of phytoplankton in the ocean (see Box 1-2, Conservation Controversy: Fertilizing the Ocean). Some scientists argue that, even if we reduced greenhouse gas emissions considerably, global warming could still continue until the end of the 21st century. Another response to fears of global warming therefore is to begin adapting to predicted changes. These adaptations would include increased fortification of coastal cities (see Chapter 3) and abandoning islands and coastal areas at low sea levels. Depending on the amount of sea-level rise, it is debatable whether this should be viewed as a valid option, because much of the world's population resides in coastal areas.

The final question of what impact climate change will have on the Earth and its ecosystems may already be partially answered. Glacial retreat in mountainous regions around the globe has been linked to increases in atmospheric greenhouse gases. Examples include the Himalayas in Asia, the Alps in Europe, the Rocky Mountains and Cascade Range in North America, and the southern Andes in South America. Recent warming in the Arctic has resulted in shrinkage of ice cover. Since accurate measurements began in the late 1970s using satellite imagery, record lows in sea ice coverage were observed several times, all occurring in the 2000s. In 2007, the greatest summer decline in Arctic sea ice was recorded. The Northwest Passage, pursued in vain by early explorers for centuries, opened for the first time ever in human memory. Based on projections from recent ice melting, it is predicted that the Arctic Ocean could be free of summer sea ice by the mid to late 21st century. This is noteworthy because there is no scientific evidence of an ice-free Arctic for hundreds-of-thousands of years. Some scientists, however, question our ability to make such predictions with reliability.

With an increase in melting of ice, the percentage of liquid water on the planet must increase; thus, sea-level change is a predictable outcome of global warming. The rate and total amount of change is difficult to estimate precisely, however. Other changes in climate are not as predictable. Single weather events or anomalous weather years cannot necessarily be linked to global climate change. Climate predictions include changes in rainfall patterns causing changes in water distribution patterns, and increased frequency of extreme weather events. Predicting how these changes will affect land use patterns and the abundance and distribution of various organisms is now the focus of many scientific studies.

Other than ice melting and changes in sea level, there are numerous scenarios that might develop in oceans due to global warming and climate change. It is predicted that El Niño events will increase in frequency and intensity. One pattern that develops during El Niño conditions is an increase in water temperatures in shallow tropical waters. Recent extreme temperatures during El Niños have negatively impacted tropical reefs; it is still debated whether this is a direct result of global warming (see Chapter 5 for further discussion of these issues). Increased frequency of hurricanes and other tropical storms since the mid-1980s, especially in the North Atlantic, has been linked to increased water temperatures, but it is still debated whether this trend will continue with global warming.

Carbon dioxide (CO_2) plays an important role not only in global warming but also in ocean chemistry. Increases in atmospheric CO_2 result in higher CO_2 levels in the ocean. The reaction of CO_2 with water to form carbonic acid (H_2CO_3) results in ocean acidification. Ocean pH has decreased (become more acidic) by about 0.1 units since the industrial revolution and is predicted to continue its decline as the ocean absorbs more CO_2. Some climate models predict that the pH of ocean surface waters could decline by over 0.7 pH units due to anthropogenic input of CO_2 over the next few centuries; if so, this would probably be the lowest level experienced by ocean organisms in about the past 300 million years. An analysis by Timothy Wootton and colleagues has shown that calcareous species in regions around the globe are being stressed at current pH levels, and there is growing evidence that higher acidity is impacting coral reef ecosystems (see Chapter 5). Even if the input of CO_2 into the atmosphere were to be returned to normal levels, it could be a thousand years or more before the heat and CO_2 dissipate such that the ocean returns to normal.

One of the most dramatic predictions regarding climate change and the oceans involves changes in ocean circulation. According to one theory, melting of Arctic glaciers could disrupt vertical thermohaline circulation patterns in the North Atlantic as follows. Meltwater from glaciers and the polar ice cap is made up primarily of freshwater, which is lighter than the surrounding seawater (the salts in seawater make it about 3.5% denser than freshwater). The increased meltwater as a result of global warming therefore could lower the overall density of polar seawaters. Less dense polar waters would not sink as they typically do, and this could disrupt the global thermohaline vertical circulation pattern (see Figure 1-22). Because the sinking of North Atlantic waters is the major source of oxygen for the deep Atlantic Ocean, a disruption of this vertical current could result in death of most deep-sea organisms. In addition, upwelling of deep Atlantic waters is a major source of nutrients for North Atlantic plankton. A disruption of the circulation could reduce the plankton to less than half their current biomass, which would have a major impact on ocean food webs and fisheries. Less CO_2 would be taken up by ocean waters if thermohaline circulation is weakened; this would cause further increases in global warming.

Effects of the loss of the thermohaline circulation pattern would not be limited to the oceans. The disruption of near-surface flow of warm waters from the tropics into the North Atlantic would allow colder polar waters to flow equatorward and replace them. These warm currents play a major role not only in warming the north Atlantic but in moderating the climate and temperatures of eastern North America and western Europe; most of the heat that is transported to the atmosphere comes from ocean waters. This scenario therefore would result in major climate changes in these regions with some regions experiencing a paradox of much colder weather as an effect of global warming. Other predicted changes in climate include changes in rainfall pattern and storm frequency around the globe.

Since this hypothetical scenario was developed in the 1980s there has been much discussion of its likelihood, and some scientists have even suggested the pattern is already developing. There is strong historical evidence that such a disruption in thermohaline circulation has happened in the past. The last time was about 12,000 years ago, initiated by large inputs of meltwater at the end of the last glaciations period; temperatures in Scandinavia were lowered about 30°C. These changes probably happened rapidly, in a matter of years (but probably not as rapidly as presented in the 2004 film, *The Day After Tomorrow*, where within days the U.S. citizenry was rapidly being forced toward the Mexican border by an instant ice age). Whether we are in the midst of or near such a scenario is being debated among oceanographers. Mathematical models suggest that a shutdown of this circulation would require a global warming of at least 4°C above current temperatures. Although freshwater input into the North Atlantic is increasing, some models estimate that predicted melting rates will not put in enough freshwater to stop the circulation within this century. The IPCC predicted a possible slowing of 25% by the end of the 21st century, but not a complete shutdown. Some measurements suggest that the circulation may have already begun slowing substantially. This evidence has been questioned because large-scale automated systems have been in place to accurately monitor changes in North Atlantic thermohaline circulation only since the mid-1990s. A better understanding of this thermohaline circulation system that is so critical to ocean life and global climate is gradually being gained. We may soon have better, but not perfect, predictive capabilities. Even if/when we are able to establish the scenario under which large-scale changes in thermohaline circulation will develop, this will not answer the bigger question of how scientists convince the citizens of Earth of the need to make sacrifices to slow or stop the processes leading to global climate changes: will there be the commitment before it is too late? Once the process of ocean warming begins it is not easily reversed. Susan Solomon and colleagues presented observations and models indicating how persistent the effects of greenhouse gasses are on climate change and ocean warming. Atmospheric warming from carbon dioxide is nearly irreversible for over 1,000 years, even if emissions are stopped. The transfer of heat from the atmosphere to the ocean surface layer can take ten years or less, and centuries are necessary for transfer to the deep ocean. Once the impact of ocean warming has occurred, however, the dissipation of heat will take hundreds of years, even if atmospheric warming is reversed. This suggests that once critical levels of greenhouse gasses have accumulated, we could be dealing with the impacts for millennia regardless of what actions are taken.

The attempt to gain a better understanding of global warming, climate change, and ocean circulation is a good example of the process of science that is often lost to the general public in the hype that can be presented through the news media. Predictive models are generated; hypotheses are developed and tested, supported, or refuted; theories are generated; conclusions are drawn; paradigms become established. Although science may be able to come up with an ideal solution to conservation problems, unfortunately science cannot implement that solution. Conservation depends on human decision-making. What is the "right" answer to a scientist is not always going to be viewed as such to the government leader, politician, industry representative, non-governmental organization (NGO), or the general citizenry. The ongoing challenge thus is not just how to do the best science, but also how to get the science incorporated into conservation. Many scientists are now arguing for application of the **precautionary principle**, which is a responsibility to take cost-effective actions to limit global warming, even if there is a lack of full scientific certainty that those actions are necessary to limit climate change and the associated environmental damage. Even as world leaders become more willing to apply this philosophy, however, there remains the difficulty in defining and balancing what is "cost effective" and how much scientific uncertainty is too much (for example, how much hardship are people willing to accept to take a chance at solving unpredictable climate problems). To get world leaders to listen to these scientific concerns, a consensus must be reached as to what actions are most likely to limit global warming. These are a few of the dilemmas prevalent in marine conservation; many more are presented and discussed throughout this text.

Although the expansion is proportionately small relative to the volume of water, the cumulative effect over the entire ocean can be great. Most of the recent rise in sea level is due to thermal expansion, and at the predicted rate of glacial melting it will probably be the major factor through the 21st century. Thermal expansion is more predictable than glacial contributions to sea-level rise. Projecting anticipated glacial melt is a primary reason for the variation in predictions of future rates of sea level change.

Average sea levels have been gradually increasing at around 2.5 millimeters per year. Although this is a relatively small change, the cumulative effect, especially if this rate increases, will have serious impacts in some coastal areas. Predictions of future sea-level rise vary greatly among scientists. Most predictions fall in a range from about 10 centimeters to 1 meter through the end of the 21st century. The reliability of these predictions will be important for conservation of coastal ecosystems. Although some debate continues, in recent years it has been well documented and generally agreed upon by experts that we are in a period of global warming. The most important debate from a conservation perspective is over how much humans are affecting

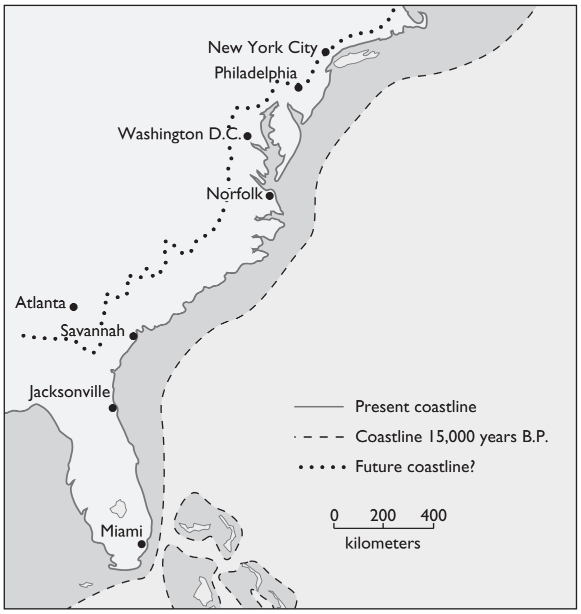


Figure 1-23 Position of the eastern U.S. coastline when sea levels were lower 15,000 years ago during the last glaciation, and predicted coastline if melting of the world's ice sheets continues at the projected rate.

global climate change and sea-level rise through air pollution and other activities, and what effects it will have on ocean currents and habitats (see Box 1-1, Conservation Concern: Global Warming and the Ocean). The effects of climate change on various marine and coastal ecosystems are discussed in later chapters.

Local Sea-Level Changes

Local sea-level changes are perceived changes in sea level relative to some point or structure on land. These can be caused by various factors. Because this is a relative change, either rising water levels or **subsidence** (sinking) of land can result in a local sea level change. One form of subsidence is the compaction of sediments in coastal areas. For example, acres of lands in coastal salt marshes of Louisiana are being lost each year as they literally sink into the sea. A major cause of this subsidence is the channelization of the Mississippi River system. The river normally has provided sediments to the marshes during periods of flooding to replenish the marshes. The channelized river transports these sediments out to sea and off the continental shelf. These issues and possible solutions are discussed in Chapter 4.

Another source of subsidence is on a much longer time scale of millions of years, associated with seafloor spreading. As seafloor gradually spreads and progresses toward the edge of the ocean plate, islands sink lower relative to ocean sea levels and eventually sink beneath the sea surface. For example, the flat-topped guyots are remnants of volcanoes that were exposed to weathering before sinking beneath the sea surface. This phenomenon can be observed in island chains, such as the Hawaiian Islands, where the newer islands rise higher above the sea surface, and the older islands to the west are lower or have disappeared below the surface.

The relative sea-level change at any point is a combination of all of these factors. Relative sea level is continually changing as a result of the ocean's volume, changes in coastal land formations, and processes of plate tectonic and sea floor spreading. From a human perspective these processes are typically slow; however, geologic events, including earthquakes or volcanic activity, or human modifications in the environment, including global climate change, could result in dramatic and even catastrophic short-term changes (see Box 1-1, Conservation Concern: Global Warming and the Ocean).

1.5 Ocean Chemistry

Seawater, Salts, and Trace Elements

The primary factor that distinguishes seawater from fresh water is the presence of relatively high concentrations of **inorganic solutes,** dissolved chemicals not formed by life processes. Within the water solution, these solutes exist as electrically charged atoms called **ions**. For example, sodium chloride (NaCl) disassociates into sodium ions with a positive charge (Na^+), and chloride ions with a negative charge (Cl^-). When these solutes are crystallized as sea water evaporates they form into **salts**. The primary salt dissolved in seawater is sodium chloride. Other ions are present at lower concentrations, including magnesium (Mg^{2+}), calcium (Ca^{2+}), potassium (K^+), sulfate (SO_4^{2-}), and bicarbonate (HCO_3^-). The ratio of these ions to each other in sea water is remarkably constant throughout the ocean. Many other elements are present at even lower concentrations and are considered minor or **trace elements**. The concentrations of these elements can vary dependent on local geological events or the input or incorporation by biological organisms. These include elements important to sea life such as nitrogen (N), phosphorus (P), and iron (Fe). They also include elements that are toxic at excess concentrations, such as lead (Pb) and mercury (Hg). The effects on organism and ecosystems of the increase in trace elements by human activity are discussed in Chapter 7. Some trace elements, such as manganese (Mn), accumulate in rock-like **nodules** around particles such as pieces of bone or algae on the deep sea floor over millions of years (**Figure 1-24**). The nodules are in concentrations great enough that commercial mining is being considered, with potential impacts to deep-seafloor ecosystems (see Chapter 8).

The measurement of the total concentration of the dissolved inorganic solids is called **salinity**. The average

Figure 1-24 A skate rests on a region of the deep-sea floor scattered with manganese nodules.

ocean salinity is about 3.5% (salinity is typically given in units of parts per thousand (ppt or ‰) rather than percent (%); for example, 3.5% is 35 ppt). The range of salinities in the open ocean is from about 33 to 37 ppt. In coastal areas where freshwater influence is high the salinities can be much lower. Divisions used to define habitats according to salinity are somewhat arbitrary, but useful, because salinity is an important factor governing the kinds of organisms that can survive in those habitats. Moving from the ocean through estuaries into rivers, there is a gradation in salinities from full strength (greater than 30 ppt) through **brackish** (about 3–30 ppt) to **freshwater** (< 3 ppt). In isolated intertidal pools where evaporation is high, salinities can become higher than those of natural seawater, and are called **hypersaline**.

Sodium chloride contributes the most to the ocean's salinity; the concentration of sodium and chloride ions in seawater is about 3.0% with the remaining salts made up of sulfate, magnesium, and other ions listed above. Minor or trace elements are present in concentrations measured in parts per million or parts per billion. Although some of the salts dissolved in seawater originate from the flow of freshwaters from the continents and coastal erosion, most of the sea salts are from other sources. For example, sodium ions originate from the weathering of rocks of the Earth's crust, but chloride ions originate in the Earth's mantle and are put into the ocean by hydrothermal vents and volcanic activity at the ocean ridges. Other ions originate from various combinations of these sources. Part of the ocean water is continually recycled into freshwater systems through evaporation, condensation, and rainfall; however, during the process of evaporation the salts are left behind. Much of the freshwater moving through this cycle originates in the oceans, because about 86% of global evaporation is from the oceans.

The global distribution of water is about 97% in the oceans and less than 1% as liquid freshwater (most of that being groundwater). The remaining 2% is in glaciers and polar ice; this is the storehouse of water that controls the large historic changes in absolute sea level, and the reason for concerns about melting of glaciers and the ice caps.

Dissolved Gases

Most gases in the atmosphere dissolve readily into ocean waters, and organisms in the oceans are highly dependent on these gases for biological processes. The oxygen required by almost all animals for respiration originates from near-surface waters either by dissolution from the atmosphere or from photosynthetic activities of plants (the oxygen bound up in water molecules cannot be obtained directly by animals for respiration). Deep-sea animals thus are dependent on vertical circulation of waters from the surface. Marine photosynthetic organisms are limited to water near the surface due to the necessity of adequate sunlight; however, they also need carbon dioxide from the air to support their metabolic processes.

The major gases in the ocean are the same as those in the atmosphere, although the proportions are somewhat different due to solubility differences. **Nitrogen**, about 78% of the volume of the atmosphere, makes up about 50% of the dissolved gases in the ocean. Nitrogen is needed by organisms to make proteins and other organic chemicals; however, most organisms cannot use the dissolved nitrogen directly. They depend on nitrogen that has been converted into organic forms. These processes are discussed below.

Oxygen makes up about 35% of the gas dissolved in the ocean; however, the concentration of oxygen in the ocean is extremely low compared to the atmosphere, which is about 21% oxygen. The average dissolved oxygen concentration in the ocean is 6 milligrams per liter of water (which can also be presented as 6 parts per million or ppm). This requires that most aquatic animals use gills to efficiently extract oxygen from the water. Even with efficient extraction and utilization of oxygen, most fish and large invertebrates cannot survive oxygen concentrations below about 2.5 ppm, a condition referred to as **hypoxia**. When waters reach near zero oxygen concentration they are called **anoxic**. Hypoxic conditions can develop when oxygen depletion occurs without adequate circulation. This can occur in coastal areas where there is excess nutrient input, as is discussed in Chapter 6.

Although the concentration of **carbon dioxide** naturally occurring in the atmosphere is very low (about 0.04%), due to its depletion by plants as a source of carbon, it is very soluble in seawater and thus comprises 15% of the dissolved gases in the ocean. Once CO_2 dissolves into the ocean it begins movement through a complex cycle (**Figure 1-25**). Carbon dioxide is continually extracted from water by plants and other photosynthetic organisms; however, it is still more concentrated in the ocean than in air. One reason for the high dissolution of CO_2 in seawater is its ability to combine with CO_2 to form carbonic acid (H_2CO_3; see Box 1-1, Conservation Concern: Global Warming and the Ocean). Because of these complex processes, CO_2 and pH levels vary widely among regions of the ocean, and within regions on a daily and seasonal basis. Factors affecting CO_2 and pH include photosynthesis and respiration, horizontal and vertical mixing, and exposure to the atmosphere. Anthropogenic increases in carbon dioxide in the atmosphere increase acidification of the oceans. Increased acidity could have a dramatic impact on ocean ecosystems, and may already be affecting some ecosystems (see Box 1-1, Conservation Concern: Global Warming and the Ocean and Chapter 7). Coral reefs may be especially sensitive because higher acidity can reduce the ability of organisms to build the reef (see Chapter 5).

Some of the dissolved carbon dioxide ends up in the shells and skeletons of marine animals and eventually into sediments. These processes begin with dissolved carbon dioxide forming carbonate ions that combine with calcium in sea water to form calcium carbonate ($CaCO_3$). This is used by marine organisms to build shells and skeletons. (The shells of many invertebrates and microorganisms are comprised primarily of calcium.) When these organisms die in the open ocean their shells gradually sink toward the bottom. On the sea floor above 4,500 meters depth calcium carbonate shells are deposited, mixing with other sediments to form thick deposits of **carbonate oozes**, recently discovered to house a rich assemblage of bacteria and other microorganisms (see Chapter 8). This calcium carbonate may eventually be incorporated into limestone rocks. Below 4,500 meters other materials, such as the remains of **silica** shells, take the place of the carbonates in forming deep sea **siliceous oozes**.

Chemicals of Life

Because living organisms are comprised of chemicals that are mostly derived directly from the ocean, understanding the biogeochemical cycle through which the elements move is critical to conservation. **Carbon** is considered the basic building block of marine (and other) organisms. The

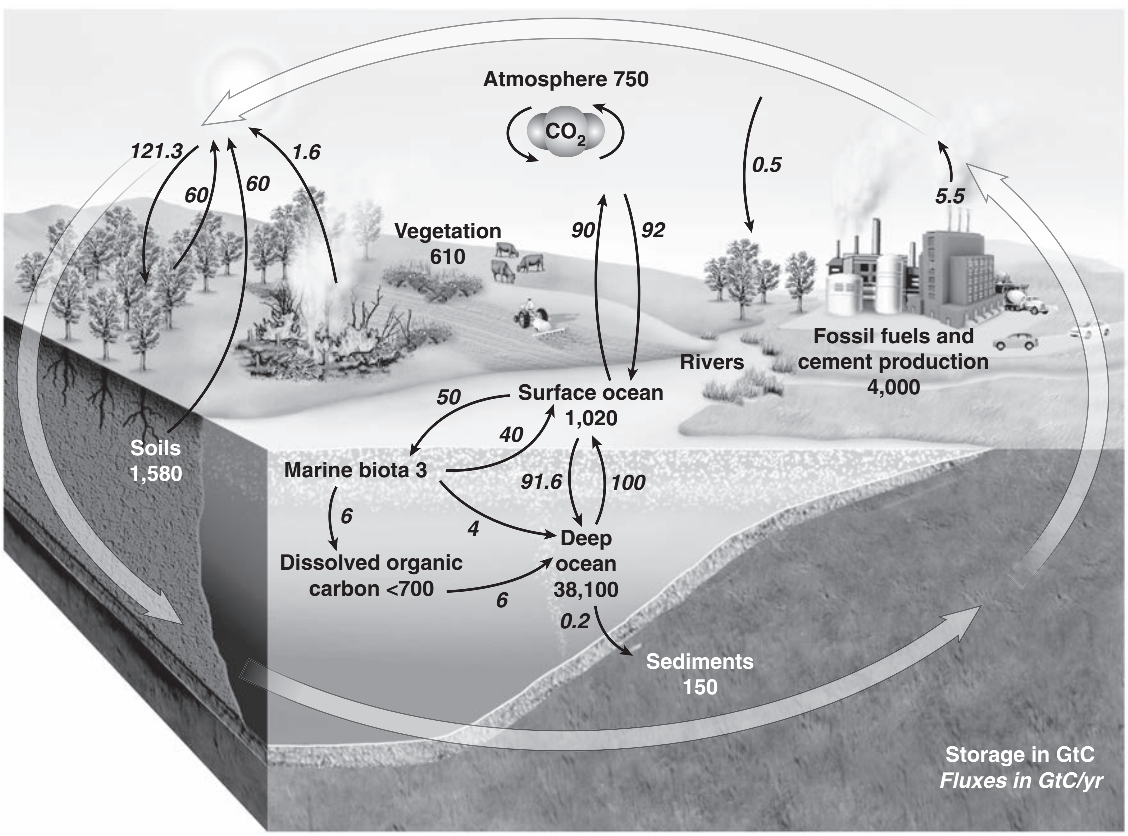


Figure 1-25 Global cycle of annual carbon fluxes and storage. (GtC = gigatons carbon.)

carbon dioxide from the atmosphere dissolves in water and is fixed into **organic molecules** using energy from sunlight during photosynthesis. This carbon is transferred, along with other chemicals, through the food web (see Figure 1-25 and Chapter 2). The carbon from photosynthetic organisms ends up either incorporated into the animal as new tissue (about 45%), expelled from the animal in carbon dioxide as a byproduct of respiration (45%), or excreted as waste into the water as **dissolved organic carbon** (**DOC**; 10%). DOC is used by bacteria and eventually ends up in other organisms. The organisms (or parts of organisms, such as shells) that escape being eaten, sink to the bottom and the carbon enters the geologic cycle as described above.

Many other elements combine to form organic chemicals in organisms; any compound that is used in the production of organic matter is considered a **nutrient**. Most nutrients are not typically considered as of concern in conservation issues because they are relatively abundant in the ocean compared to their need by organisms. The element that is most often considered as a **limiting nutrient** to ocean organisms is nitrogen (**Figure 1-26**). This is not due to a lack of nitrogen-based chemicals but because of limits in **available nitrogen**, that which is in a chemical form that organisms can utilize. Free nitrogen, dissolved in the water as a gas (N_2) in large concentrations, cannot be used by organisms directly. They depend on nitrogen that has been **fixed**, that is, converted from inorganic N_2 to organic forms (such as ammonia, NH_3), primarily by bacteria. Other major sources of available nitrogen are nitrates from rivers and precipitation. These sources are limited enough that ocean waters are frequently nitrogen-limited, especially in regions away from large river inputs. The other major nutrient that can be limiting under certain conditions is phosphorus. Ocean currents also can have a major influence on the distribution of nutrients, especially in upwelling regions where nutrients that have accumulated in deeper waters are brought to the surface. The distribution of nutrients in surface waters can be mapped indirectly by measures of surface chlorophyll concentration made possible by satellite imagery (**Figure 1-27**).

While many ocean areas are considered nutrient-limited, others have unnatural and undesirable excesses. Marine ecosystems have evolved and adapted to the typical levels of nutrients available in the water, and additional nutrients may support one component of the ecosystem to the detriment of others. This has led to numerous conservation issues, especially in coastal regions. For example, nutrients may support bacterial or algae growth in excess; and excess bacterial activity can deplete waters of oxygen, creating **hypoxic** (low oxygen) zones (see Chapter 6). Excess nutrients in coral reef ecosystems produce blooms of algae that grow over and outcompete the corals (see Chapter 5). The source of excess nutrients in coastal and nearshore ecosystems is typically fertilizers, sewage waste, or excesses of other organic materials flowing into the ocean from rivers. Trace elements in the ocean are typically not limiting because they are used by organisms in such small amounts. Iron concentrations, however, can be so low, due to its insolubility and tendency to adhere to falling particles, that it can be a limiting nutrient to biological production in open ocean waters. Proposals to enhance plankton production in the open ocean as a mechanism for removing carbon dioxide from the atmosphere have been met with controversy (**Box 1-2. Conservation Controversy: Fertilizing the Ocean**).

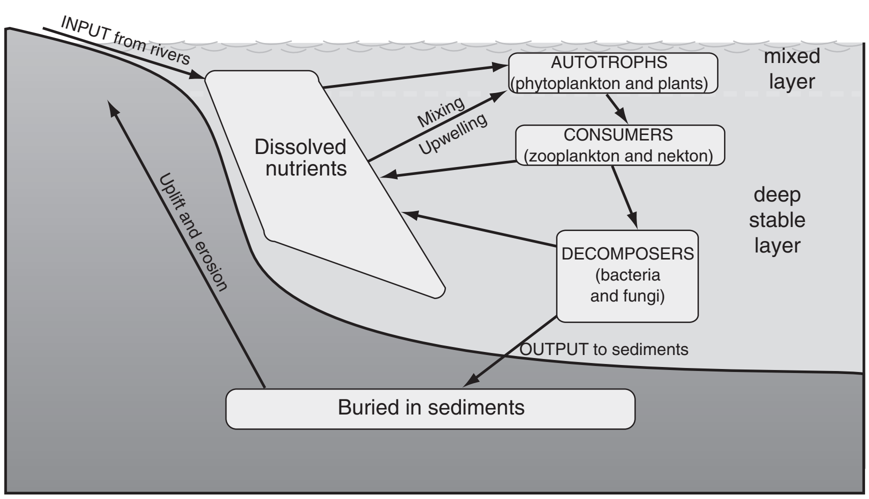


Figure 1-26 Cycle of the major ocean nutrients nitrogen and phosphorus.

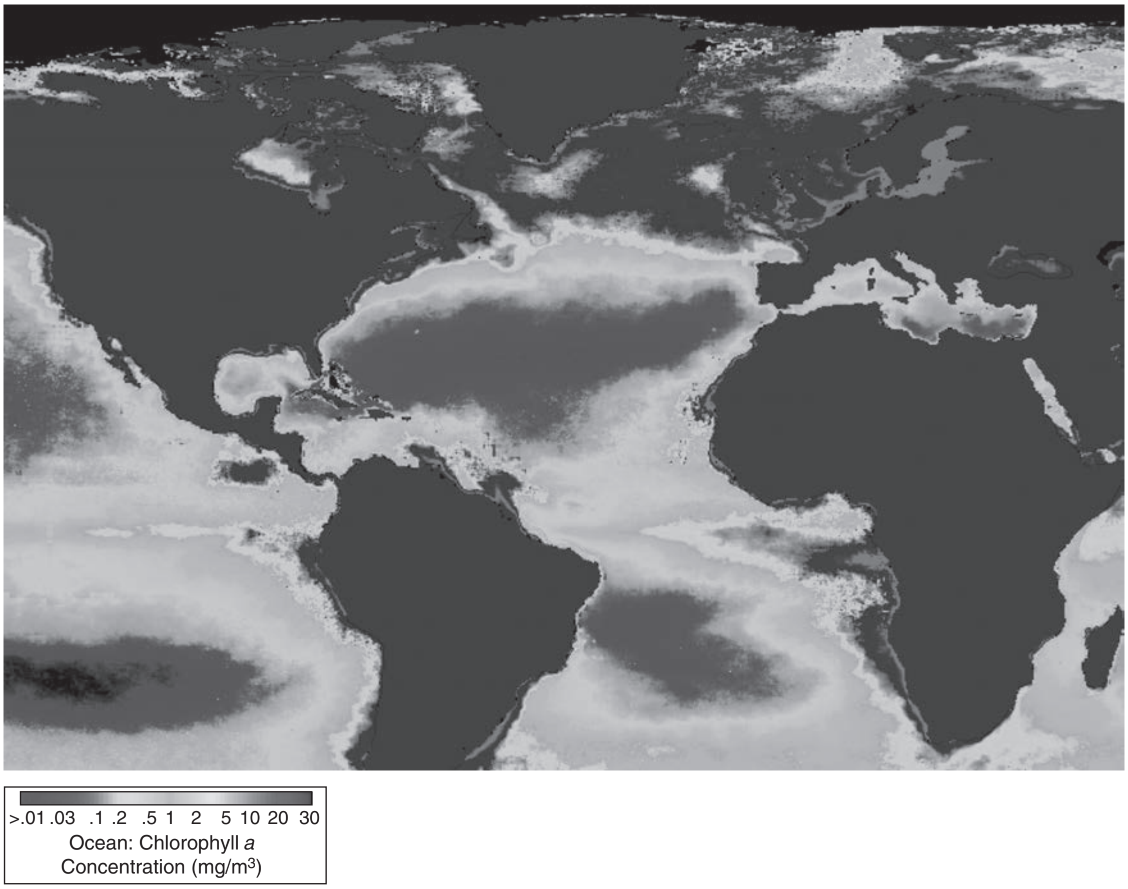


Figure 1-27 Image taken by the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) satellite. Gray ocean colors represent chlorophyll concentrations, which are indicative of primary production. Note high chlorophyll levels in the North Atlantic, in upwelling zones along the west coasts of South America and Africa and along the equator, and at the mouths of major rivers such as the Mississippi and Amazon. (See color plate 1-27.)

Box 1-2 Conservation Controversy: Fertilizing the Ocean

Iron is a **micronutrient** necessary for the survival and growth of phytoplankton in the open ocean. It is required in concentrations much lower than for macronutrients such as nitrogen and phosphorous (the ratio of iron to nitrogen in ocean phytoplankton is over 100,000 to 1), however. Despite the limited need for iron by phytoplankton, there are areas of the open ocean where it is the major limiting nutrient. This means that production will not increase without additional iron, even if excessive amounts of the other required nutrients are available. Although speculated since the early 1900s, this condition was not documented by scientists until the 1980s, primarily by oceanographer John Martin. This soon led to proposals to fertilize the ocean with iron in order to enhance biological production and reduce global warming. The reduction in global warming would be achieved through the uptake of atmospheric carbon dioxide by the phytoplankton. A large overall reduction in CO_2 is not achieved, however, if the plankton are eaten and incorporated into ocean food webs, because some of the CO_2 eventually will be released by organisms up the food chain during respiration. In order to achieve a net reduction in atmospheric carbon dioxide, the plankton or their remains would have to die and sink to the bottom. This would result in sequestering of the carbon in deep ocean waters or sediments (in the same manner that sequestering of carbon occurs naturally when carbonate tests of plankton sink into the deep sea).

Ocean experiments began in the 1990s to examine the effect of the iron "fertilizing" on biological production (**Figure B1-3**). Since then experiments have been carried out of various sizes in different areas of the ocean. The results have varied. Some experiments showed a substantial production of phytoplankton but little sequestering of carbon because most of the plankton was eaten and little carbon sank to the sea floor. In other experiments, a relatively large amount of carbon was sequestered and transported to deep waters; however, it is uncertain if this represented a permanent loss of the carbon to the deep sea. None of these experiments were carried out on a scale and in a time frame that would be necessary to see a significant change in atmospheric carbon dioxide. In a natural experiment, George Wolff and colleagues found that areas with elevated concentrations of iron from natural leaching

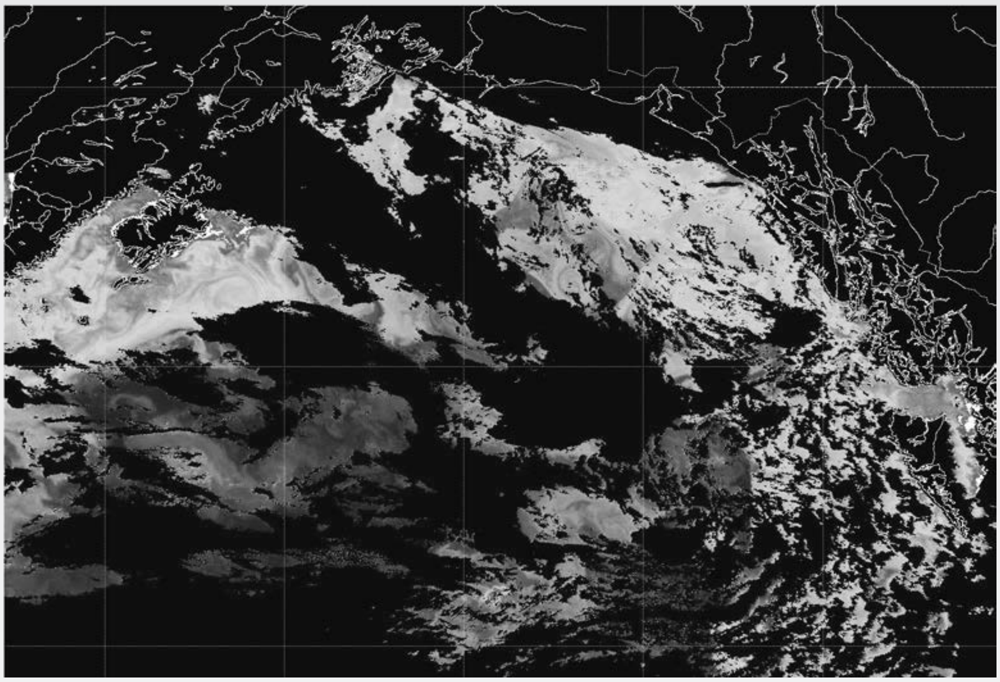


Figure B1-3 SeaWiFS image of the northeastern Pacific Ocean during a 2002 iron fertilization experiment. Chlorophyll concentration increases from dark to light gray in the open ocean. The fertilized bloom is the small, light area in the bottom center. (See color plate B1-3.)

from volcanic islands supported a larger biomass and density of deep-sea life and a species assemblage with lower evenness than unfertilized areas, and they concluded that large scale fertilization would likely affect deep-sea ecosystems. Supporters of proposals to use iron fertilization of the ocean as a means to reduce carbon dioxide in the atmosphere argue that there is enough evidence to begin large-scale programs. Some industries would like to see this used as a way to mitigate their release of carbon dioxide into the atmosphere.

There is still widespread opposition among conservation biologists to proposals for large-scale application of iron to the oceans, including Ken Buesseler and colleagues, who argued that a stronger scientific foundation of the risks and benefits of such actions are needed. Some doubt the effectiveness of iron fertilization in reducing atmospheric carbon dioxide. Even if it is shown to be effective, the unpredictability of impacts on marine ecosystems may never make it worth the risk. Ocean ecosystems are complex and changing those ecosystems intentionally is considered irresponsible. We do not know what indirect effects might occur. For example, encouraging blooms of certain algae could result in an increase in species that are detrimental to animals in the current ecosystem, such as toxic organisms or jellyfish, inedible to many fishes and marine mammals. Excess nutrients present the risk of initiating harmful algae blooms or creating hypoxic "dead zones" (see Chapter 4). Environmentalists argue that those pushing for ocean fertilization as a method to reduce carbon dioxide are looking for an excuse to allow carbon pollution to continue, rather than making the tougher decisions to radically reduce carbon emissions. This debate presents a somewhat unique type of conservation dilemma. Rather than suggesting we reduce actions that create climate change (e.g., burning fossil fuels), it is proposed that we increase actions that reduce the effects of climate change (e.g., ocean fertilizing). Following a conservation ethic, with a primary focus of maintaining the health of the natural world, one would argue in support of maintaining the natural balance and focusing on the reduction of carbon emissions. We have learned the hard way too many times that the best of human intentions, when they involve manipulating the environment, often return to haunt us. Other examples will be presented throughout the text.

STUDY GUIDE

Topics for Review

1. Distinguish and define the four major sub-disciplines of oceanography. Discuss how they overlap.
2. Describe the concepts of seafloor spreading and plate tectonics.
3. What is the difference in the subsidence at ocean trenches and subsidence in coastal marshes?
4. Compare the slope of the continental shelf, slope, rise, and deep-sea basin.
5. Compare characteristics and formation processes between active and passive margins.
6. What mineral resource from deep-sea regions is most likely to be exploited by humans?
7. How does the formation of island arcs differ from that of the Hawaiian island chains?
8. Why are guyots an important habitat in the deep sea?
9. What features differentiate beaches, deltas, and rocky coasts as habitats for coastal organisms?
10. What features make estuaries attractive to marine organisms as well as humans?
11. What are the major physical differences in the three basic reef types?
12. Why are barrier islands more vulnerable to erosion than volcanic islands?
13. What factors are responsible for higher levels of biological production in the neritic zone than the oceanic ocean?
14. What factors are responsible for higher levels of biological production in the epipelagic than the mesopelagic?
15. What distinguishes waves from currents relative to movements of energy and water?
16. Explain why wind waves of relatively large size are commonly seen coming ashore on calm, windless days.
17. Describe why the highest surfing waves are more commonly found along active coastlines (e.g., California) than passive coastlines (e.g., the eastern U.S.) even if wind conditions are similar.
18. How does a relatively shallow tsunami wave have enough energy to move ashore with such a great destructive force?
19. Describe the factors responsible for the variability in tidal range on a daily and monthly cycle at a single point along the coastline.
20. How are tides important to coastal organisms?
21. How does the Gulf Stream affect climate along the U.S. east coast?
23. How does coastal upwelling enhance fish populations off Peru?
24. How does thermohaline circulation enhance animal populations in the deep sea?
25. How does thermal expansion influence global sea levels?
26. What are the sources of minerals that make the ocean salty?
27. What chemical effect does excess carbon dioxide have on the oceans?
28. The concentration of dissolve nitrogen gas in the ocean is much greater than that of carbon dioxide. Explain why nitrogen is often a limiting nutrient to marine organisms, but carbon is not.
29. How do biological organisms contribute to sequestering of carbon in deep-sea sediments?

Conservation Exercises

Develop science-based arguments in support of each of the following statements:

1. Coral reef ecosystems are more sensitive than most marine ecosystems to environmental changes.
2. Barrier island beaches are inappropriate places for building homes.
3. Although tsunamis are unpredictable and unavoidable, preparations can reduce their impact on coastal settlements.
4. If the frequency of El Niño events increases it could have a substantial impact on coastal fish harvest.
5. We are currently experiencing a global warming trend, and humans are largely responsible.
6. Without a reduction in carbon dioxide emissions global warming will continue.
7. Because of the potential effects of the disruption of thermohaline circulation we must make sacrifices to stop global climate change.
8. Fertilization of the ocean with iron is not the best answer to resolving climate change problems.
9. Sea-level rise will continue with global warming even without large-scale melting of glaciers and sea ice.

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