SEA|MESTER

Introduction to Oceanography

OCE 2002

Lecture Notes

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Version 2.3.1

Introduction to Oceanography; OCE 2001

This is a 3 credit class, which involves a total of 45 contact hours. These will include both lectures and field activities. The material of the class is divided into 5 sections: geological oceanography, chemical oceanography, physical oceanography, biological oceanography, and humans & the oceans. Besides quizzes and exams on material covered in class, you will also be evaluated on papers and presentations that will allow you to further your knowledge of one given topic related to oceanography, and conduct your own independent research project.

Grading will be as follows:

5 quizzes (one at the end of each section) @ 4% each	20%
1 individual literature review paper	10%
1 individual literature review presentation	10%
1 mid-term exam	20%
1 group research project paper	10%
1 group research project presentation	10%
1 final exam	20%

The midterm exam covers material from two sections, chemical oceanography (30%), geological oceanography (30%), the introduction lecture and labs (10%) as well as questions on literature review presentations (30%). The final exam is not cumulative. It covers material from the other three sections; physical oceanography (30%), biological oceanography/coasts (30%), humans and the oceans (30%), as well as the concluding class and questions on labs (10%).

The literature review paper is 4 pages long, and the literature review presentation is 8 minutes. The group research paper should as long as it needs to be to present your findings in a clear and concise way. The group research presentation will be 10 minutes. Full details of expectations for papers and presentation are presented in the appendices at the end of the lecture notes.

This course is mostly based on the textbook *Essentials of Oceanography* (11th edition) by Trujillo and Thurman. A few chapters also incorporate elements taken from *An Introduction to the World's Oceans* by Sverdrup, Duxbury and Duxbury, and from other sources. This booklet includes only quick explanations of processes and concepts covered in class. It is meant to help jog your memory and help you review for exams. If you need more thorough explanations, you should read the associated chapters in the textbooks and/or speak to your instructor.

1. Introduction to Oceanography (Trujillo, Introduction & Chapter 1)

1.1. Background

Oceanography is the study of the oceans. It is not a pure science, but rather incorporates many different sciences, such as chemistry, biology and physics, with a common goal of understanding the oceans.

Why study the oceans? They cover 71% of our planet (Figure 1.1), and play an important role in regulating global climate through their interaction with the atmosphere. Oceans have been present for about 4 billion years, and life originated in oceans. Moreover, the majority of the human population lives by the sea, and modern societies use biological and mineral resources from the sea. Understanding the oceans is critical for optimal and sustainable harvest of these resources.

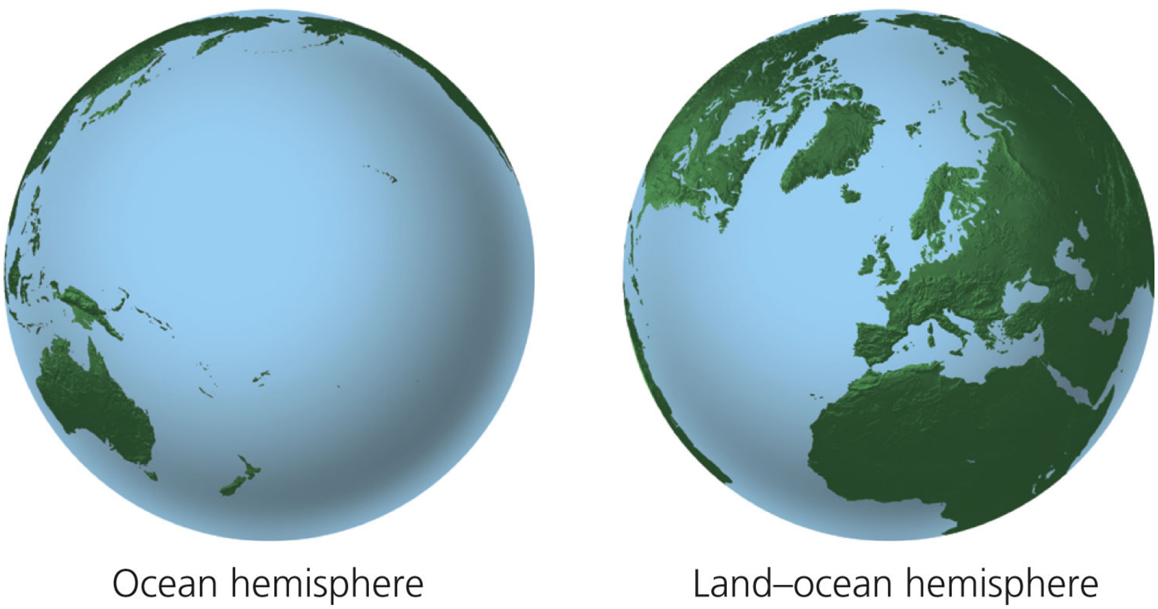


Figure 1.1. Oceans cover 71% of the planet. The Pacific Ocean is the largest and when viewed from that side the planet shows much more ocean than land.

Four major oceans have traditionally been recognized: Pacific, Atlantic, Indian and Arctic. Additionally, most oceanographers now recognize the Southern Ocean as its own entity (Figure 1.2). Their average depth of all oceans is ~3.5 km. The Pacific means peaceful or tranquil, but that is a misnomer as the Pacific Ocean has numerous earthquakes and volcanoes along its edge (the Ring of Fire, see Chapter 3). The Pacific is the oldest ocean, about 200 million years old, and the deepest, with an average depth of 4.2 km. It is the largest (13,000 km wide) and covers 1/3 of the earth's surface. The Atlantic Ocean is half as old as the Pacific, and much smaller (6,600 km wide). It is 3.6 km deep on average. The Indian Ocean is 7,000 km wide and has an average depth of 3.7 km. It is confined to the Southern hemisphere. The Arctic Ocean is mostly frozen and has an average depth of

1.1 km. The Southern Ocean is physically connected to the Pacific, Atlantic and Indian Oceans but this body of water, south of about 50 degrees south, is defined by the distinct circulation of the Antarctic convergence.

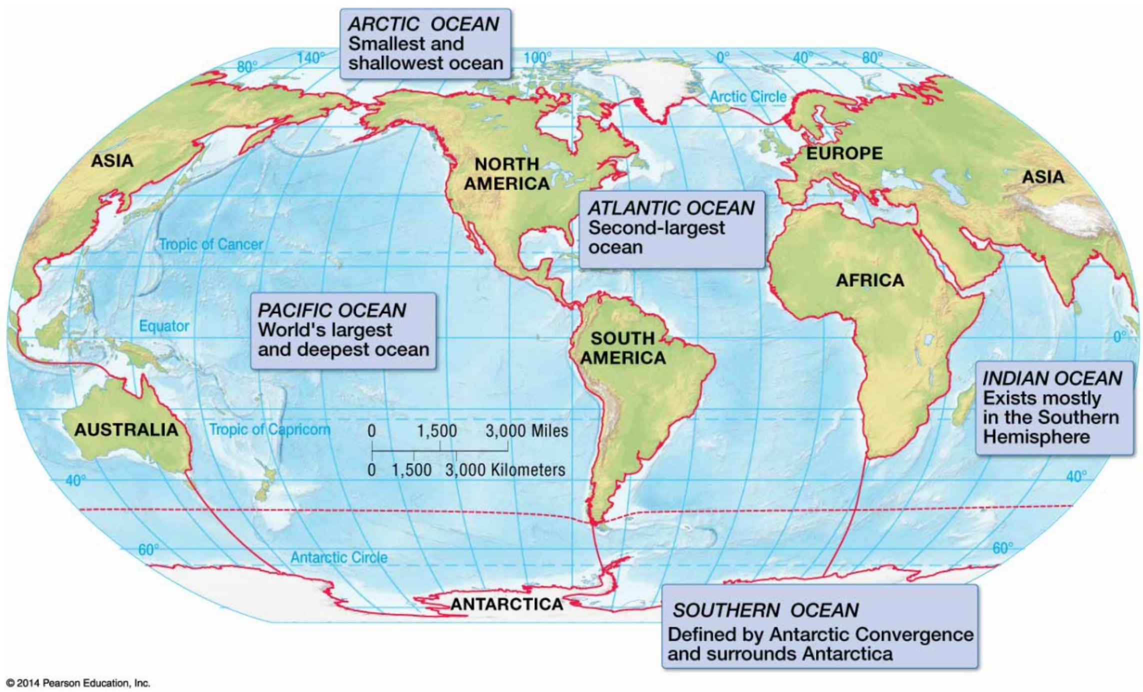


Figure 1.2. The world's oceans.

Seas are bodies of salt water that are smaller and shallower than oceans. They have a direct connection to an ocean and are partially enclosed by land, often as indentations into continents, or delineated by an island arc. There are many seas around the world, including the Caribbean, Mediterranean and Red Sea.

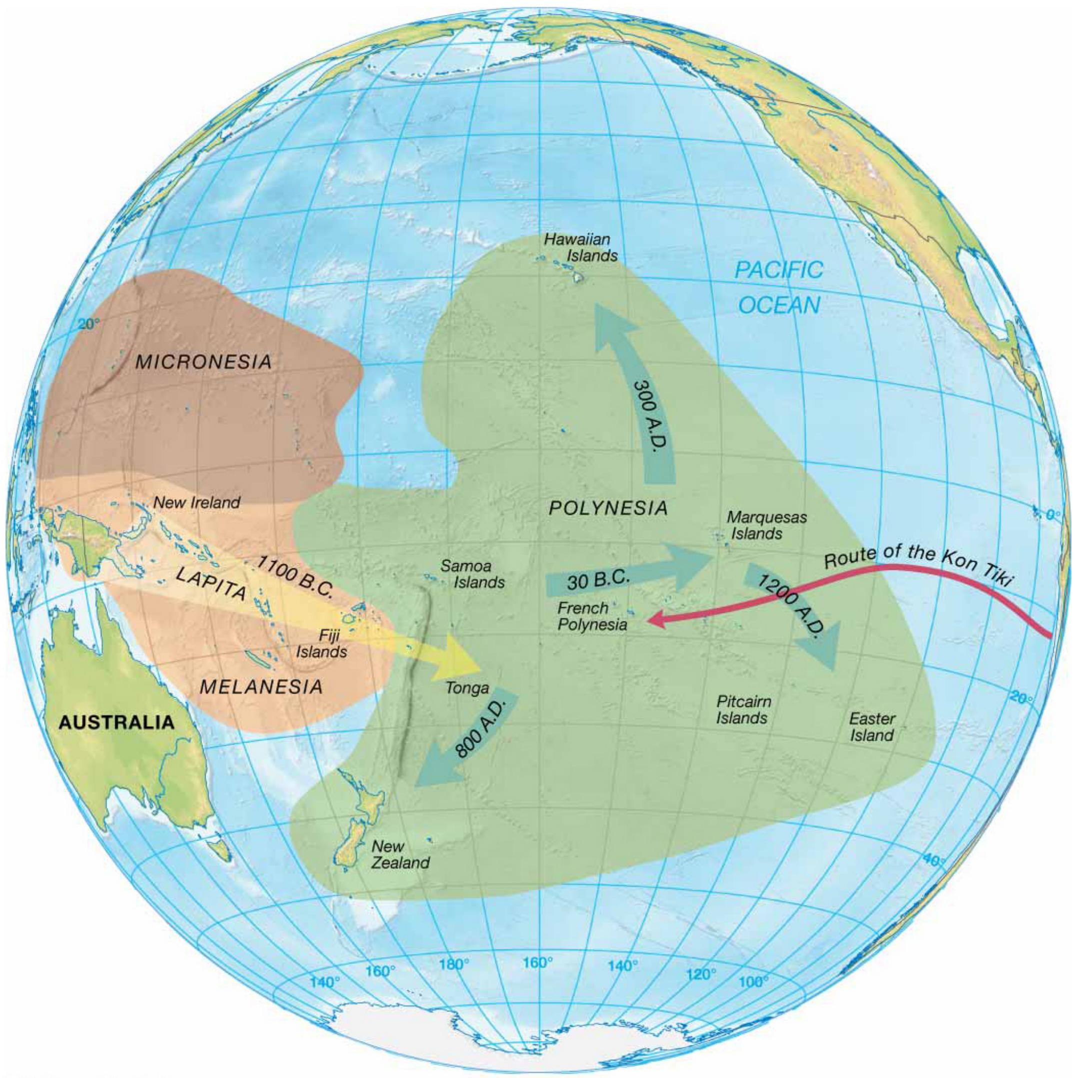


Figure 1.3. Movements of humans from Asia to the Pacific islands with approximate dates of arrival. The map also shows the route of the balsa raft Kon Tiki, from South America to Polynesia in 1947.

1.2. History of Oceanography

Early history

Oceanography is a relatively new science, compared to mathematics or physics. Modern oceanographic research really only started towards the end of the 19th century. However, humans have interacted with the ocean for thousands of years and in doing so, started to develop knowledge of the oceans to gather food and to travel. As early as 4000 BC, the first boats were built for trading in the South Pacific, and tools were developed for navigation. Asian people populated the Pacific islands by sailing and paddling in rudimentary canoes to Micronesia, Melanesia and Polynesia (Figure 1.3). While Thor Heyerdahl showed in 1947 that it was possible to sail in a simple raft from South America to Polynesia, there is clear genetic and anthropologic evidence that supports migration from Asia rather than South America. Around 450 BC, Herodotus made one of the first world maps. It was centered on the Mediterranean and showed the world known at that time (Figure 1.4). Around 325 BC, Pytheas found a method to determine latitude based on the angle of the horizon and the North Star. He also made a connection between the moon and the tides. Around this time, Greek scholars suggested that the earth was round and the Greek Eratosthenes made remarkably precise estimates of the earth's circumference. Ptolemy produced around 150 BC a world map similar to those produced before him, but which also included lines of longitude and latitude.

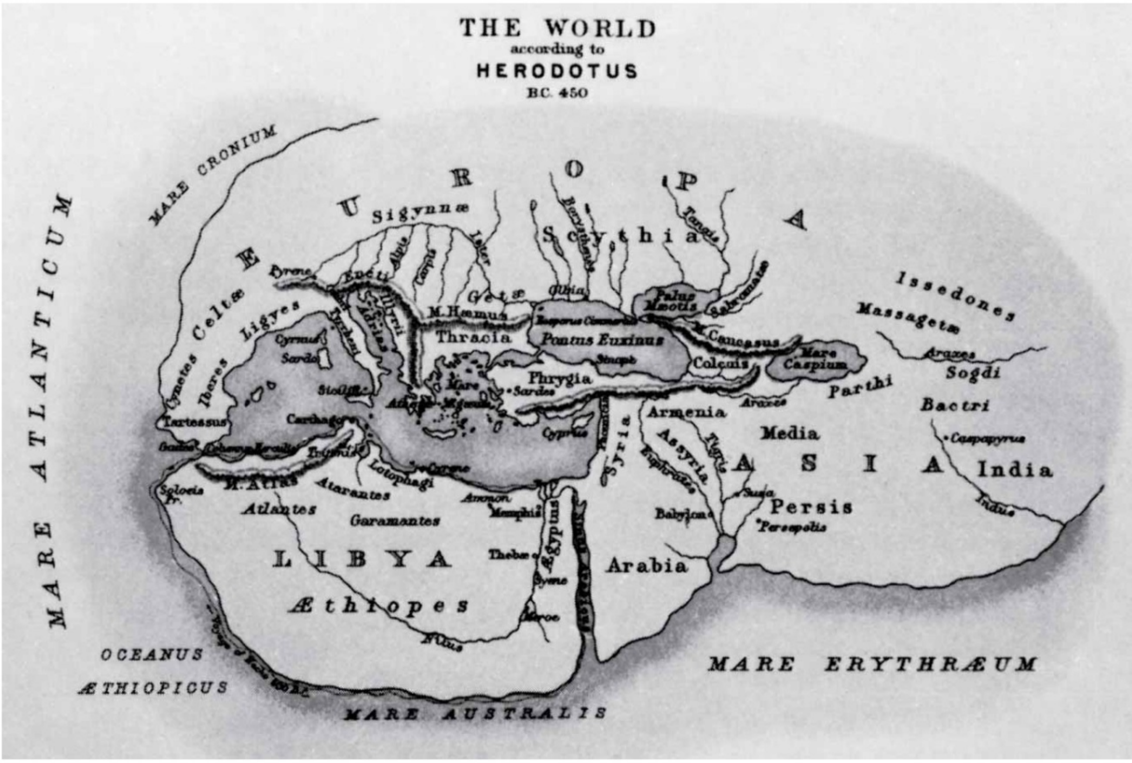


Figure 1.4. Herodotus' world map (450 BC)

The middle ages

Much of the early achievements and knowledge of the Greeks were lost after the destruction of the library of Alexandria in 415 AD and the fall of the Roman Empire in 476. The Library of Alexandria had served as one of the most important depository of scientific information at that time. Some of this knowledge was retained by the Arabs who sailed extensively for trading in the Mediterranean Sea and Indian Ocean. In much of Europe, scientific inquiry was suppressed as Christianity rose. However in Northern Europe, the Vikings of Scandinavia started building sturdy ships and exploring the Atlantic. Around 981 AD, the Viking Erik the Red sailed from Iceland to Greenland. His son Leif Eriksson, sailed west from Greenland and landed in North America, on what is now Newfoundland.

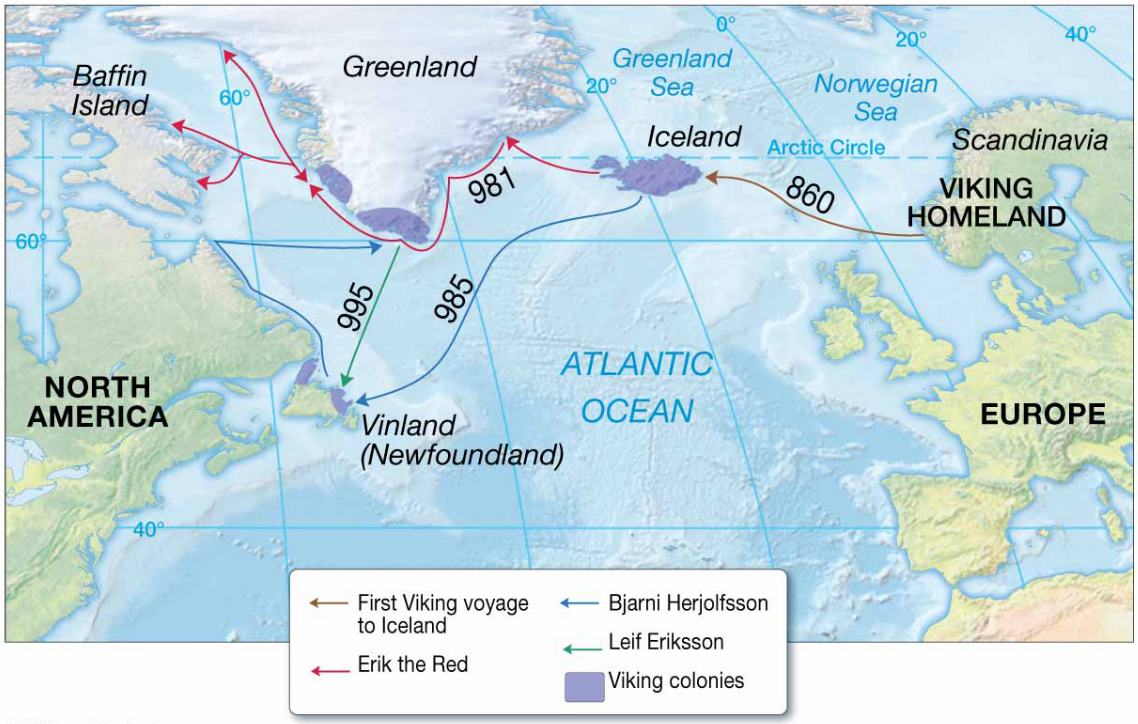


Figure 1.5. Viking voyages and colonies

European discoveries

In the mid-15th century, the search for new trade routes to the east led Europeans to start maritime explorations. Christopher Columbus was an Italian navigator charged by the Spanish royalty to find a new route to the East Indies across the Atlantic Ocean. In his first voyage in 1492 (Figure 1.6), Columbus made landfall in the then uncharted Caribbean, thinking he had arrived near India. Columbus made three more voyages to the Caribbean, which inspired other explorers. John Cabot is thought to be the first explorer to reach mainland North American in 1497, making landfall in the northeast of the continent. In 1519, Magellan organized the first successful circumnavigation of the world, which lasted until 1522 (Figure 1.6.). He started with 5 ships and 280 sailors, and of those only 1 ship with 18 men returned. Magellan himself was killed in the Philippines in 1521.

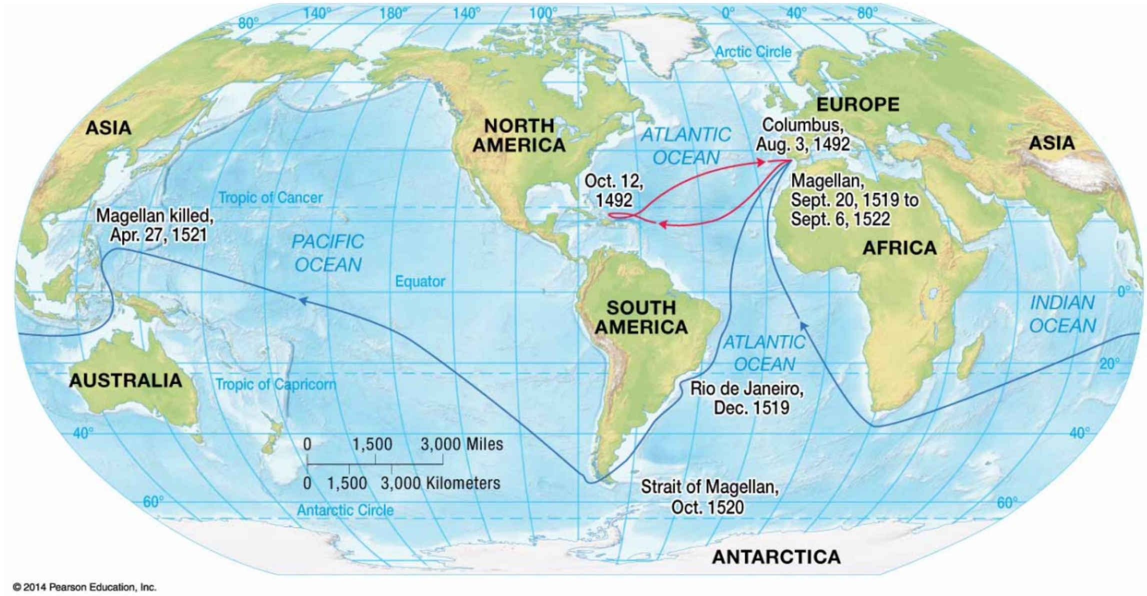


Figure 1.6. Voyages of Columbus and Magellan

Improving navigation and scientific knowledge of oceans

As colonies were established in the new world, interest in developing accurate charts was increased. James Cook sailed extensively around the Pacific Ocean from 1768-1779 and discovered and charted numerous islands. During his voyages, James Cook made depth soundings and recorded information on winds, currents and water temperature (Figure 1.7). In 1769, Ben Franklin charted the Gulf Stream for trading purposes, in particular to increase the speed and efficiency of mail ships between England and the North America (Figure 1.8). The voyage of the Beagle, which started in 1831 with Darwin on board, really marks the beginning of mankind's interest in the oceans for scientific purposes. During the Beagle's five year voyage around the world (Figure 1.9), Darwin collected and catalogued many species of plants and animals and developed theories not only on the origin of species but also on the formation of coral reefs.

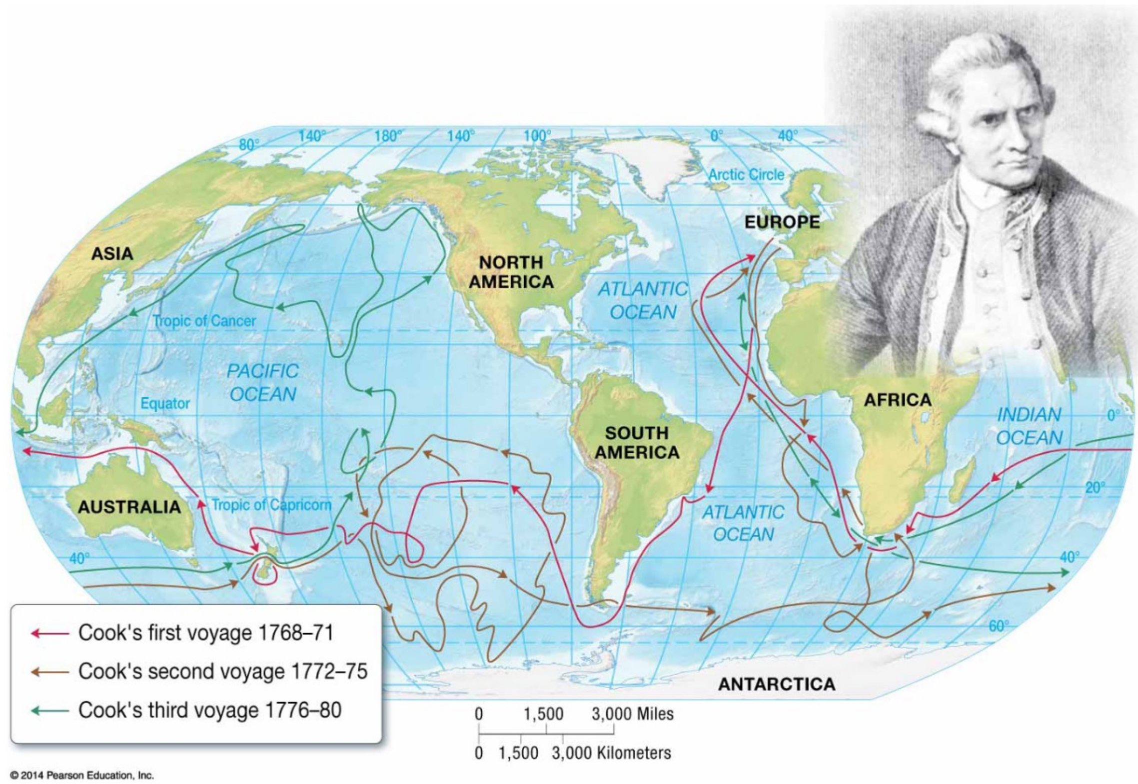


Figure 1.7. James Cook's voyages

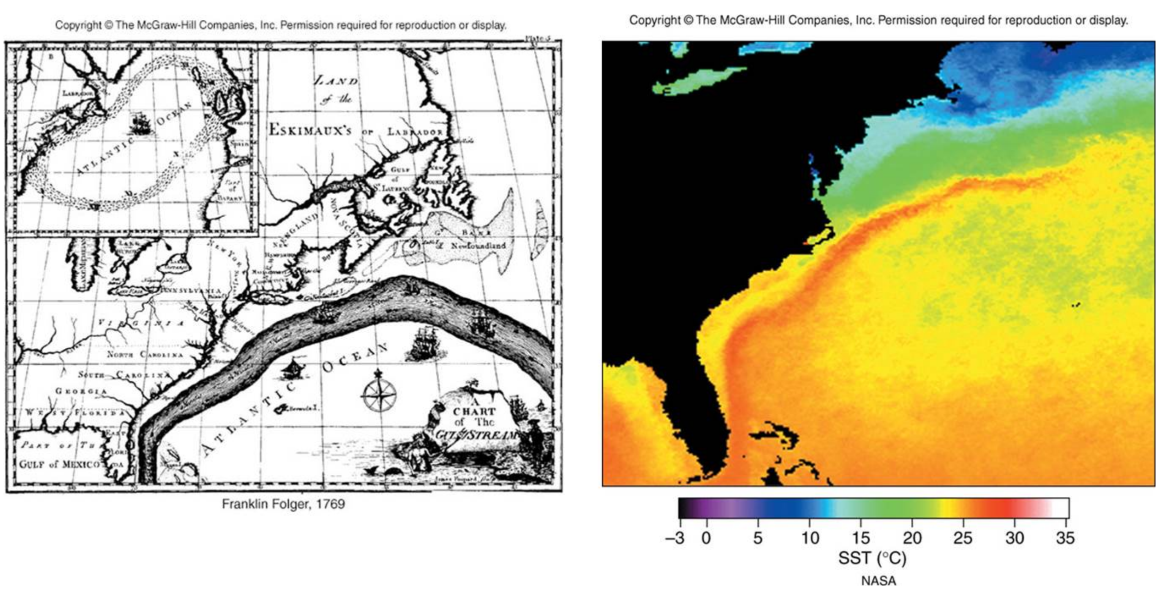


Figure 1.8. The Gulf Stream, as drawn by Benjamin Franklin in the 18th century (left) and captured by remote sensing of sea surface temperature in the 21st century (right).

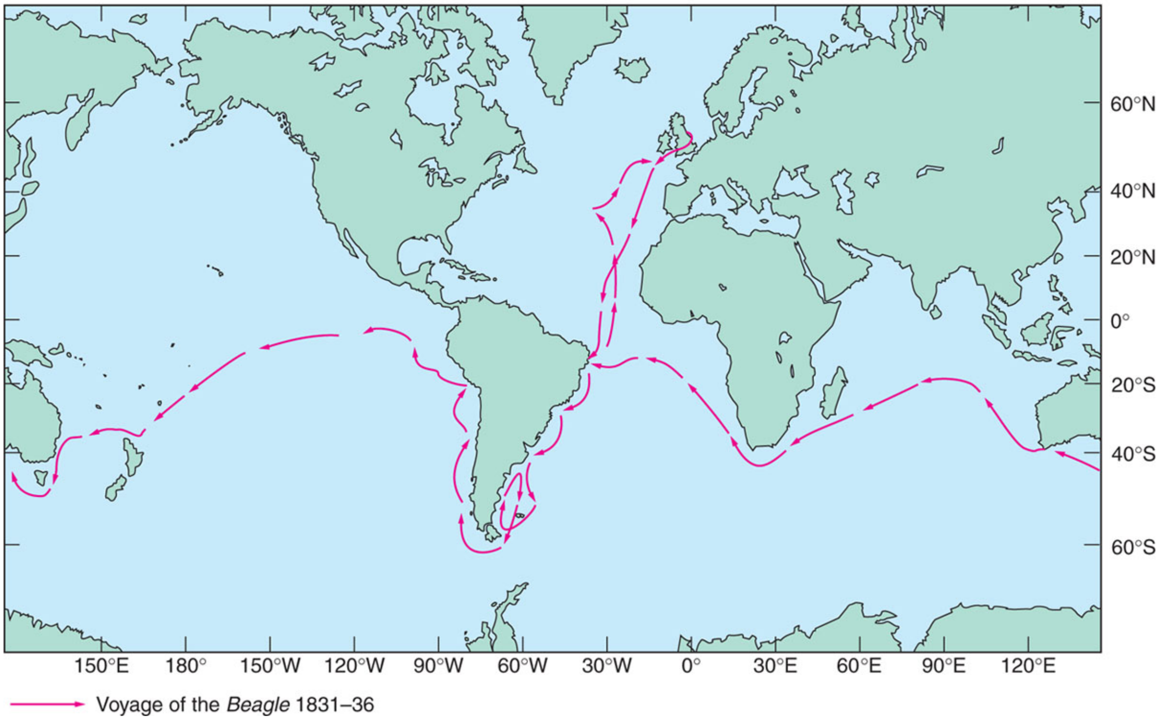


Figure 1.9. Route of Charles Darwin on the Beagle.

Around that time, interest in ocean science grew, but we lacked the technology to verify hypotheses, which lead to a lot of unverified speculation. For example, Edward Forbes an English naturalist who surveyed the shores of Europe, erroneously hypothesized in the mid-1800s that there would be no life below a depth of 550m. His hypothesis disregarded the fact that explorers Sir John Ross and his nephew Sir James Clark Ross had found organisms in mud samples taken at 1.8 km and 7 km deep in polar waters, as early as 1818.

Much of the interest in ocean science in the 17th and 18th century was linked to exploration or trading purposes. In the early 19th century, interest in ocean science grew but was still limited. In the latter part of the 19th century, the laying of telegraph cables increased the necessity of understanding the seafloor, its currents and organisms. Cables retrieved after extended periods underwater led to the discovery of many deep-sea organisms.

Modern oceanography

Modern oceanography started in 1872 with the Challenger expedition, organized by the British government. This expedition involved four years of circumnavigating the earth, covering 110,840 km (Figure 1.10). It conducted 492 deep-sea soundings and discovered the deepest part of the oceans, the Mariana trench in the Pacific (the Challenger made its deepest sounding at 8,185m; the Mariana trench is now known to be over 11 km deep). The Challenger sampled waters around the globe and led to the identification of 4,717 new species. The Challenger is still to date the most comprehensive oceanographic expedition ever undertaken. The success of the Challenger spurred other nations to undertake oceanographic research.

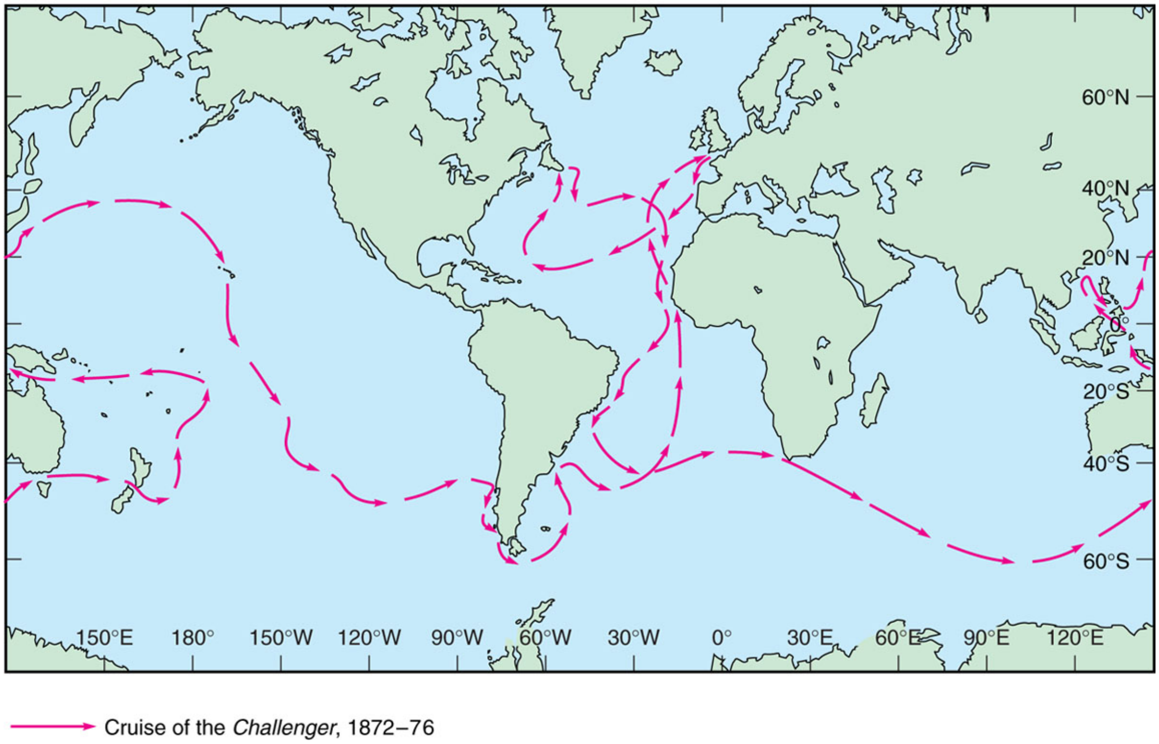


Figure 1.10. Route of the Challenger expedition, 1872-1876.

As research continued, there was a transition from a descriptive science to a quantitative science and efforts were made to test hypotheses. In 1893, Fridtjof Nansen, a Scandinavian scientist, put forward the hypothesis that there are no land masses under the Arctic ice. To test his hypothesis, he decided to freeze his ship, the Fram, into polar ice to see if it would drift. The Fram was locked in ice until August 1896. While frozen in the Arctic, Nansen took temperature and salinity measurements, and observed great phytoplankton blooms. He measured depth and found the Arctic to be a deep ocean. After his return from that expedition, Nansen continued research in oceanography and is now well-known for the sampling bottle named after him.

The 20th century saw important technological development that could be used in marine science, and the establishment of oceanographic institutes dedicated to research. Wars also contributed greatly to technological improvements and during World War II the military supported research on sound transmission and charting of the sea floor, in partnership with Scripps and Woods Hole Oceanography Institutes. Nowadays there are many more oceanographic institutes, and a lot of research is conducted in universities. Governments fund a lot of research concerning fisheries, and species that cross borders lead to international collaborations. Large-scale international collaborations have been developed in recent years, including the Deep-sea Drilling Project and the Ocean Drilling Project. Technology such as GPS, satellites, submersibles, hydrophones, CTD, sonar, digital photography and scuba is used in modern research.

1.3. The earth's structure

Density and density stratification

Density is a measure of mass per unit volume of a material; the concept is important for many sub-disciplines of oceanography. The early earth was a ball of hot liquid rock and the elements in it started to segregate according to their density, with the highest density materials in the center and lower density materials near the outside, in a process called density stratification

Earth's internal structure

Because of density stratification, the earth is composed of distinct layers which can be viewed based on their chemical composition or their physical composition (Figure 1.11). Based on chemical composition, the earth is divided into the core, the mantle and the crust.

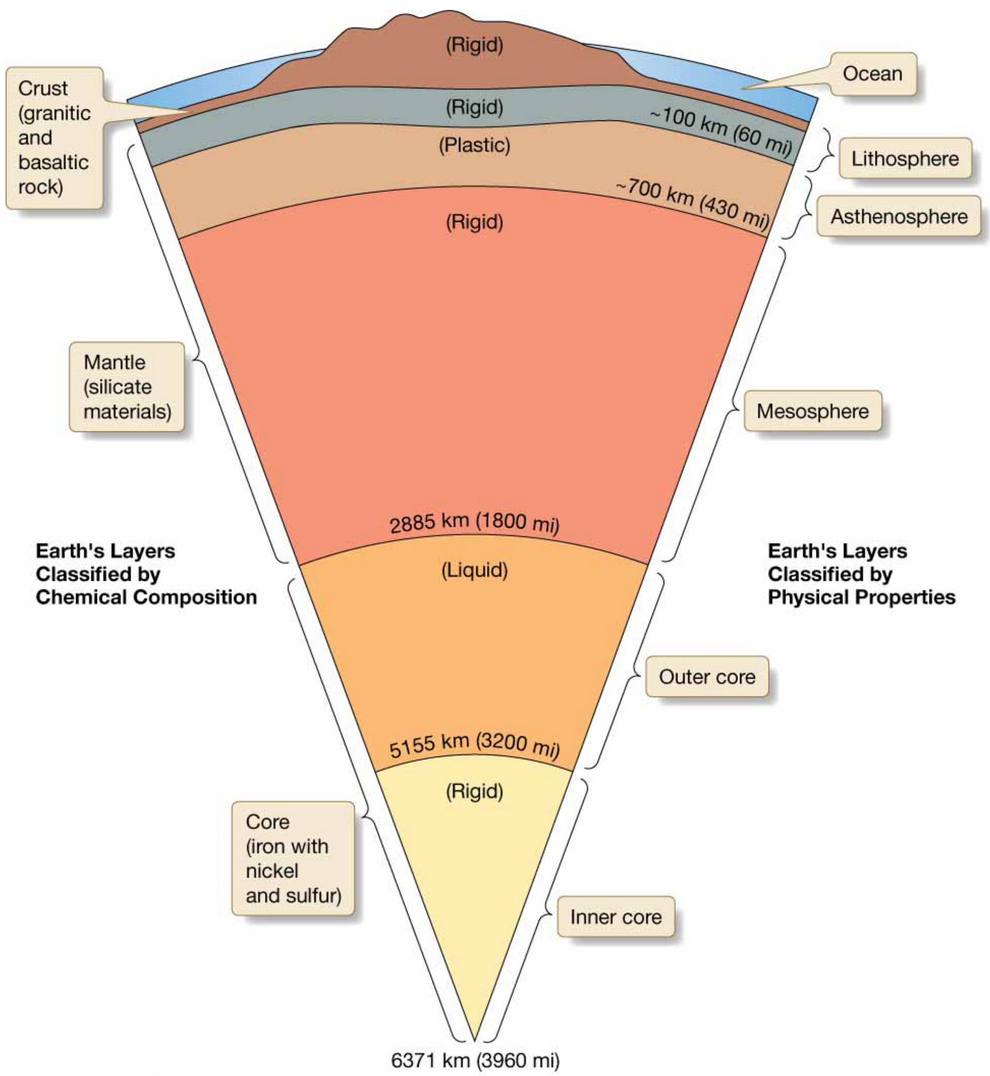


Figure 1.11. Physical and chemical stratification of the Earth.

The core is the innermost layer; it is composed primarily of iron, but also contains nickel, sulfur and oxygen. The mantle is composed of magnesium-iron silicates. It is 2,866 km thick and makes up 70% of the earth's volume. The outermost layer is the crust. There are two types of crusts, oceanic and continental crust. Oceanic crust is thinner (8 km thick on average), solid, and relatively dense (~3.0 g/cm^3). It is composed of basalt-type rock, which is high is iron, magnesium, and calcium. Continental crust is on average 35 km thick, solid and less dense (~2.67 g/cm^3). It is granite-type rock which has a high content of sodium, potassium, aluminum and silica.

When looking at the earth's stratification by physical properties, the breakdown of layers is slightly different (Figure 1.11). Although the crust is chemically different from the mantle underneath, it is actually fused with the outermost part of the mantle in a strong and rigid surface shell called the lithosphere, which extends to 100-150 km thick. Under the lithosphere is the asthenosphere, where the rock is partially molten and is weak and deformable. The asthenosphere ranges from the base of the lithosphere to approximately 350 km deep. Under the asthenosphere, the increased pressure creates a rock of greater strength and rigidity in the zone known as the mesosphere from the base of the asthenosphere to the mantle-core boundary.

The outer core is found between the mesosphere and the inner core. The outer core is partly molten because of the very high temperatures in this region. The innermost layer is called the inner core. It is solid and very dense (~13 g/cm^3) because it is under great pressure. The inner core is also very hot (4,000-5,500^{o}C) because of pressure and radioactive decay.

Isostatic adjustments

The asthenosphere offers buoyant support to each section of the lithosphere. Elevated continents, which are thicker and less dense than oceanic crust, extend higher because they "float" on the mantle in a similar way that icebergs float on water, with a good portion of their mass below the water line (Figure 1.12). Each section of the lithosphere sinks into the asthenosphere until its mass is balanced by its buoyant force in a process called isostasy. Added mass on the lithosphere, such as a mountain chain or glaciers, makes it sink deeper in the asthenosphere. If this mass is removed (i.e. glaciers melting), the lithosphere would slowly rise in response, until its new mass in is balance with its buoyant force. This process is known as isostatic crustal rebound (Figure 1.13).

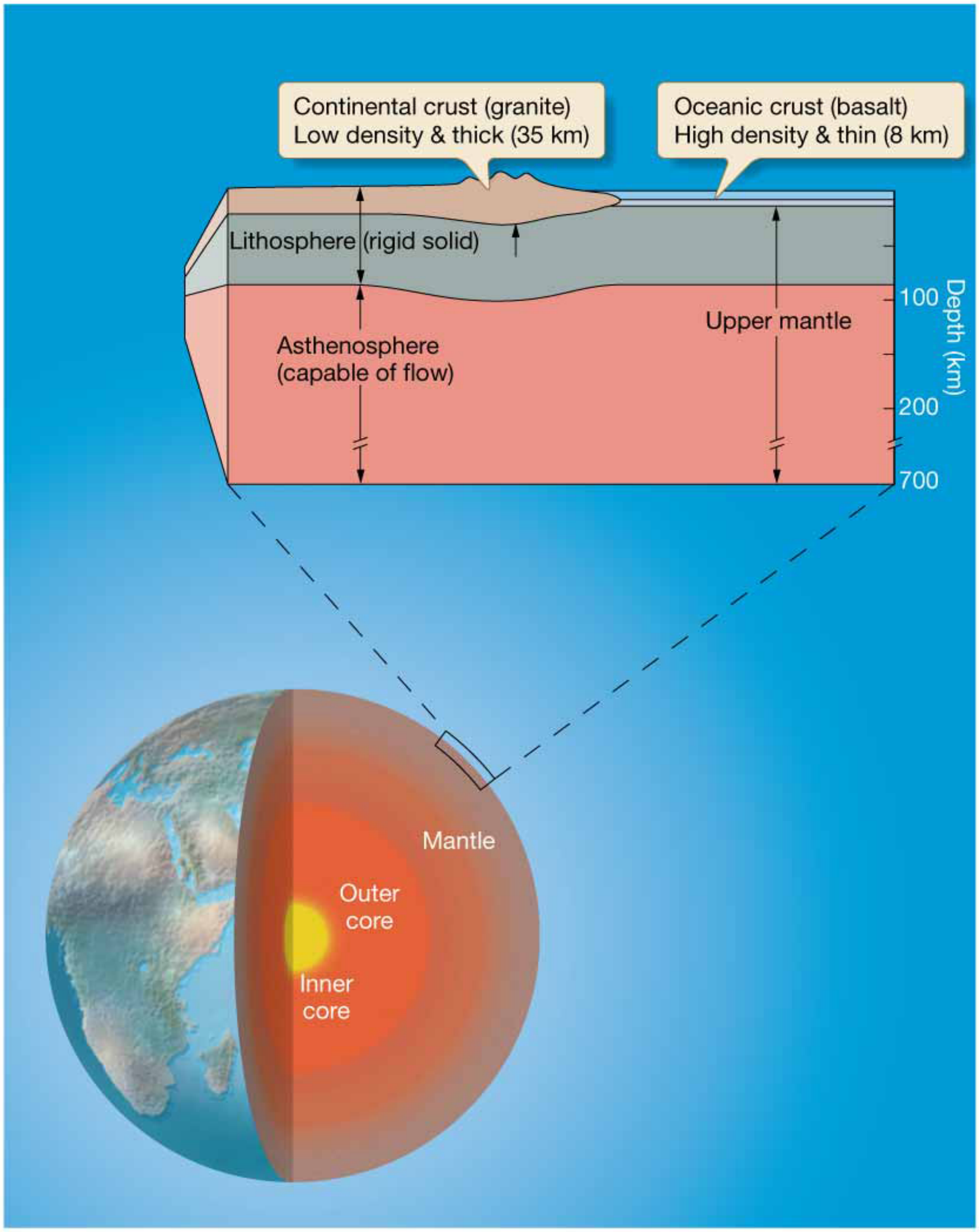


Figure 1.12. The internal structure of the earth with details of the lithosphere, showing continental and oceanic crusts, all buoyed by the asthenosphere.

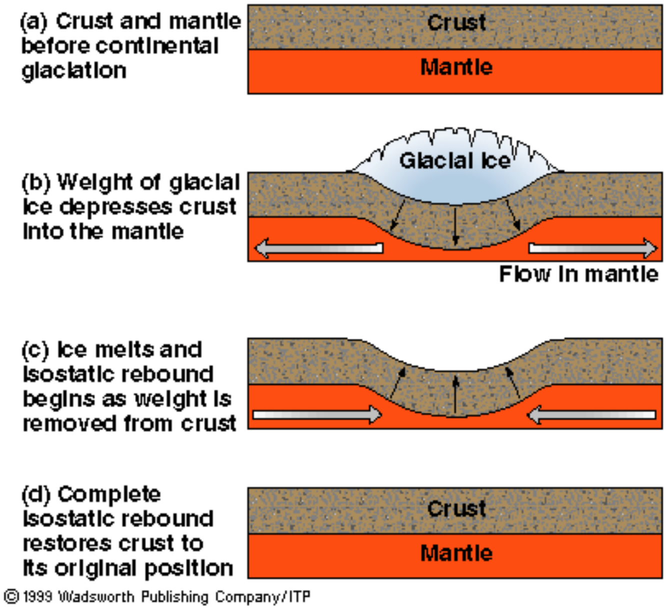


Figure 1.13. Isostatic crustal rebound in response to the formation and melting of glaciers.

1.4. Formation of the earth's atmosphere and oceans

Our atmosphere was expelled from the earth's mantle by volcanoes in a process called outgassing. The composition of the early atmosphere was similar to the gasses emitted from volcanoes and included a lot of water vapor. This water vapor condensed in the atmosphere, and formed rain which filled basins, becoming oceans (Figure 1.14). The minerals that make sea water salty came mainly from runoff from land, as rivers and acidic rainwater eroded rocks. The oceans have been present for about 4 billion years, and chemical and geological evidence suggests that ocean salinity has remained constant for the last ~1.5 billion years. If salinity remains constant despite the constant addition of salts from river runoff, then there must be a mechanism that simultaneously removes salts from the oceans. One of these mechanisms is biological organisms, which use minerals to make their shells and skeletons.

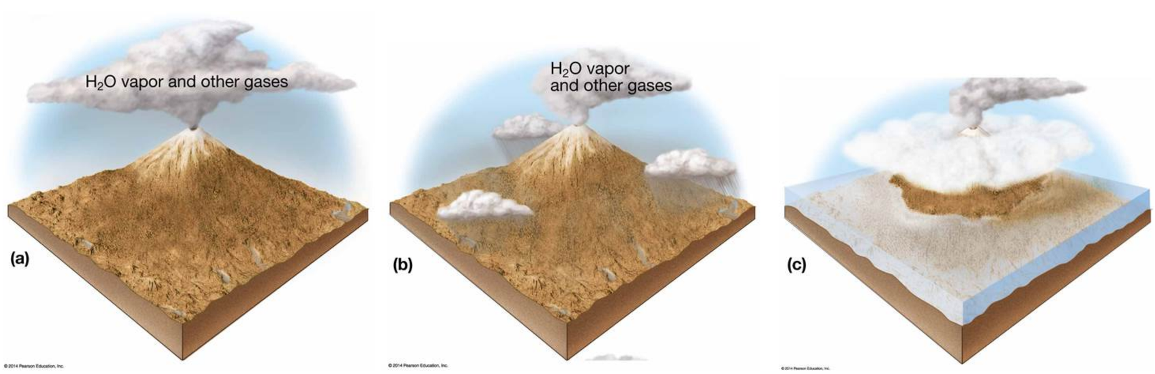


Figure 1.14. The formation of the earth's oceans, from volcanic outgassing to condensation in clouds and precipitation to the earth's surface.

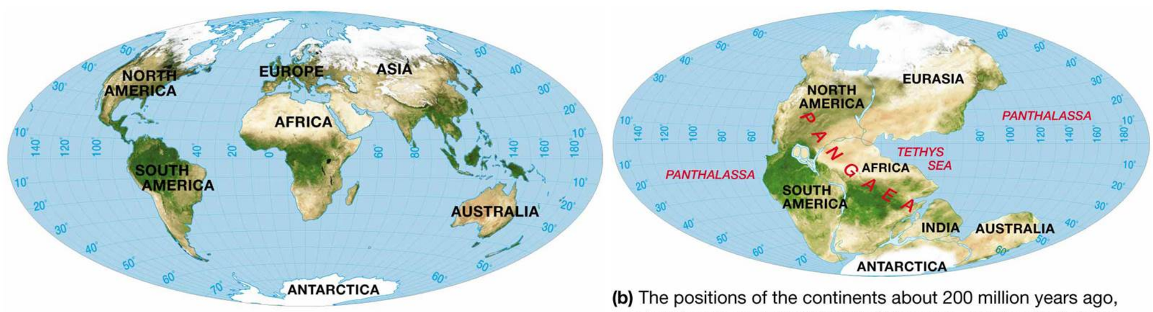
1.5. Review Questions

1. What percentage of the Earth's surface is covered with oceans?
2. Where is the Mariana Trench?
3. Which ocean is the deepest?
4. Which ocean is the oldest?
5. Who was the first person to circumnavigate the world?
6. Where did Polynesians come from?
7. Which part of North America did Leif Eriksson reach?
8. Who was the first explorer to reach mainland North America?
9. Who charted the Gulf Stream, and for what purpose?
10. What are two theories developed by Charles Darwin?
11. Which expedition is referred to as the most comprehensive oceanographic expedition ever, identifying over 4,000 new species?
12. Who froze his ship the *Fram* into the Arctic ice? And why did he do this?
13. What is density?
14. What are the 3 layers of the earth, based on chemical composition?
15. What are the 5 layers of the earth, based on physical properties?
16. What is the lithosphere?
17. Explain isostatic crustal rebound
18. Which type of crust is densest? Which is thickest? Which extends further in the asthenosphere?
19. Where did the ocean's water originally come from?
20. What processes add salt to the oceans? What processes remove salt from the oceans?
21. Which part of the core of the earth is solid?
22. What is the mineral composition of the inner core?
23. What is the mineral composition of the mantle?
24. What type of rocks generally form oceanic crust

2. Plate Tectonics & the Ocean Floor (Trujillo, chap. 2)

2.1. Continental Drift

In 1912, a German meteorologist called Alfred Wegener proposed that approximately 200 million years ago, all continents existed together in a single supercontinent that he called Pangaea. He suggested that Pangaea was then broken into two continents, Laurasia (today's North America and Eurasia) and Gondwanaland (the southern continents), by the rotation of the Earth and tidal forces.



(a) The positions of the continents today.

(b) The positions of the continents about 200 million years ago, showing the supercontinent of Pangaea and the single large ocean, Panthalassa.

Figure 2.1. Current position of continents and position of continents when they formed the supercontinent Pangaea, 200 to 300 million years before present.

Many lines of evidence supported Wegener's hypothesis:

1. The geographic fit of continents (Figure 2.2). This fit is even more striking if one looks at a map of the world with sea level lower than at present day.
2. Similar rock types in older mountain ranges and rock formations between continents, for example between North America and Europe (Figure 2.3).
3. Fossils older than 150 million years old are very similar between continents, whereas younger fossils are not. This suggests that species roamed freely between the continents until 150 million years ago, at which point populations on each continent no longer interbred and evolved into different species. For example, 250 million year old fossils of the reptile mesosaurus are found only in South America and Africa, with a limited distribution in both continents (Figure 2.4). This suggests that the continents were once joined.
4. The southern continents contain tillites, a type of glacial deposits, which were formed 250-300 million years ago. This suggests that the continents were at some point closer to the South Pole than they are today and were covered by ice sheets.
5. Some northern continents have tropical deposits of coal (e.g. in the US and China), suggesting they were closer to the equator in the past.

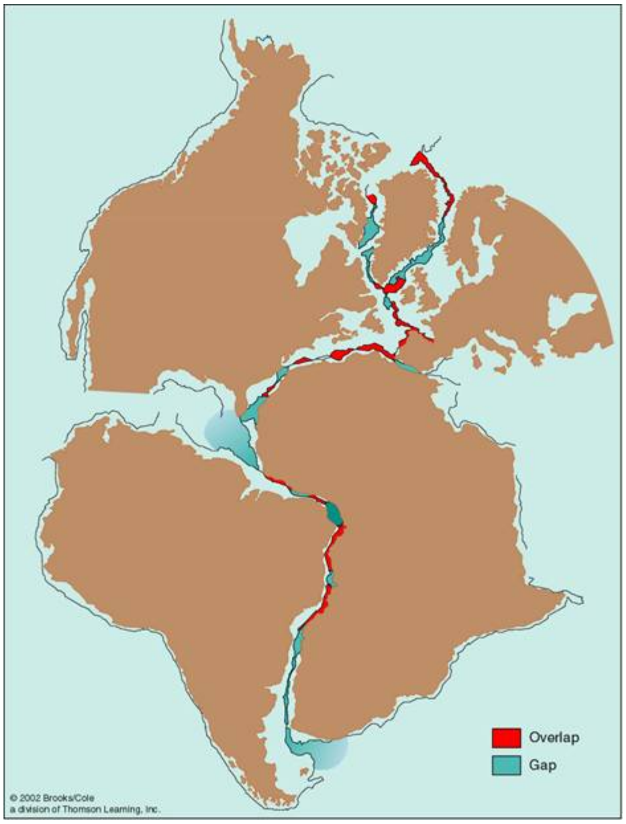


Figure 2.2. The continents fit together remarkably well, especially when using the edge of the ocean basin (2000m) as the boundary.

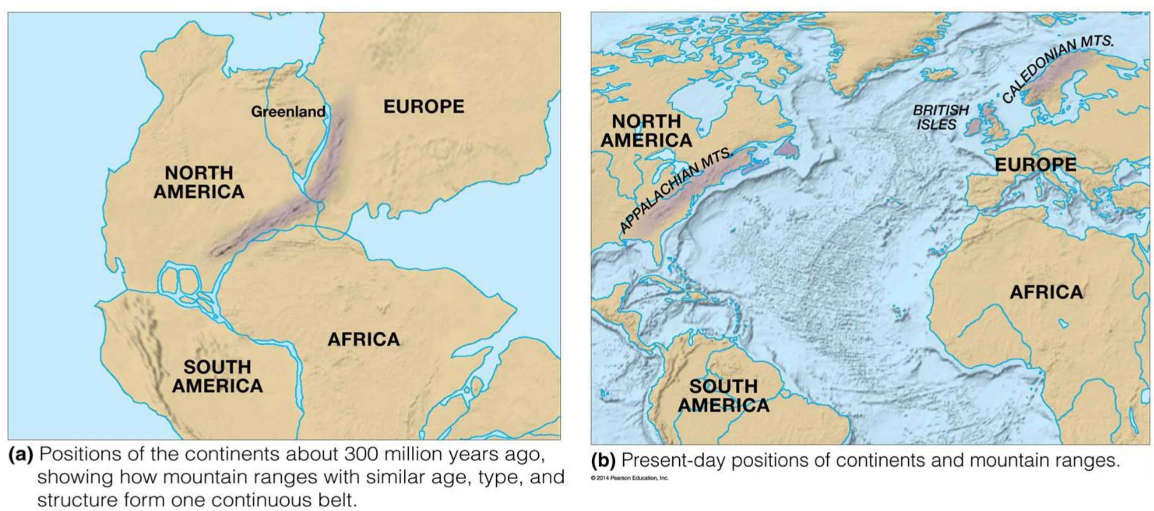


Figure 2.3. Matching mountain ranges between North American and Europe

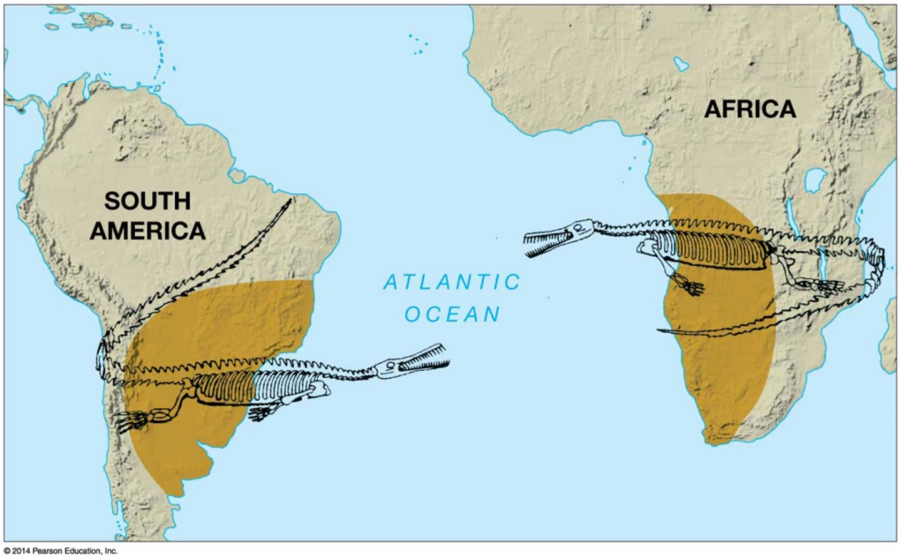


Figure 2.4. The limited range of Mesosaurus fossils in South America and Africa suggests that the continents were once joined.

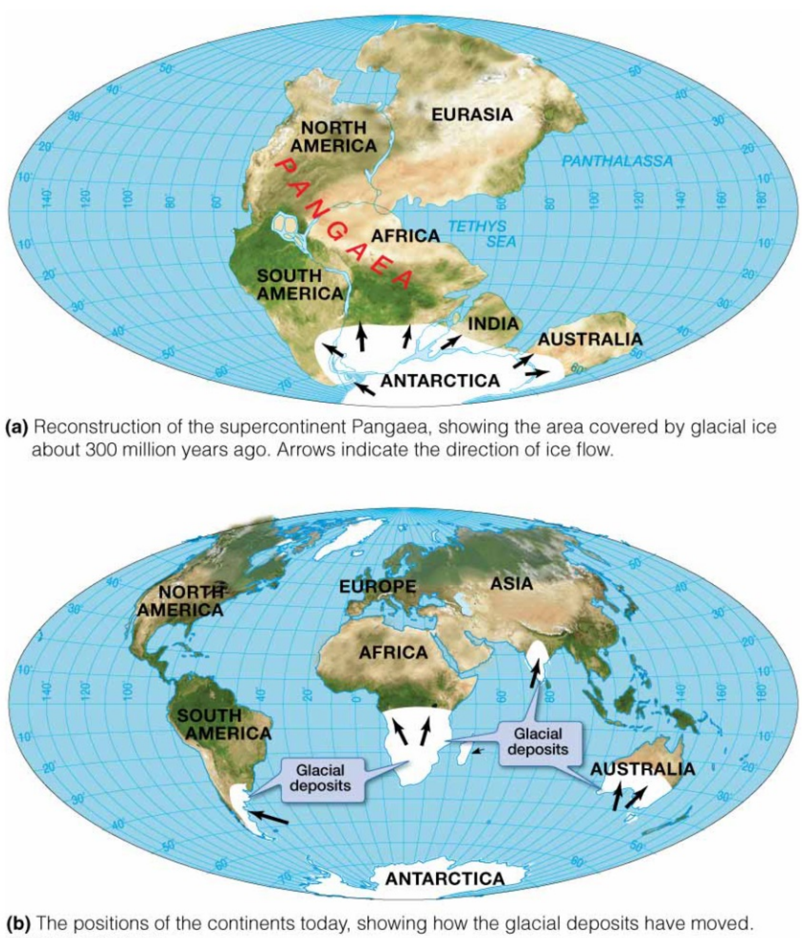


Figure 2.5. Glacial deposits found in all southern continents had to have been produced when those continents were much further south.

Though all this supported Wegener's theory, the mechanism he suggested (rotation of the Earth and tidal forces) did not provide a reasonable explanation for the movement of continents and for that reason the theory was not accepted by the scientific community at that time.

2.2. Seafloor Spreading

Large-scale efforts to survey the ocean during and after World War II showed interesting features on the sea floor. The first of these was a chain of mountain, approximately 65,000 km long, 2-3 km high, and 1,000-3,000 km wide, that ran through the middle of the ocean basins. This is called a mid-ocean ridge, and has an associated narrow volcanic zone near its axis. Another feature observed were deep ocean trenches, 6-11 km deep underwater canyons characteristics of the Pacific Ocean, and often associated with island chains (e.g. Japan, Indonesia, Philippines, Aleutians).

In the 1960s, Harry H. Hess proposed the theory of seafloor spreading to explain these features of the seafloor. Accordingly to this theory, material inside the earth is heated by the earth's natural radioactivity. Heated magma, as it becomes less dense, moves upward towards the lithosphere, setting up large convection cells within the molten asthenosphere (Figure 2.6). As this magma moves horizontally beneath the lithosphere, it cools down and sinks down towards the core. This theory suggests that the currents created by the convection cells carry along large sections of the overlying lithosphere. Where currents diverge, molten magma breaks through the lithosphere at the mid-ocean ridge, solidifying into new crust material. These are spreading centers. Since the diameter of the Earth has remained constant for about a billion years, if there is new crust being formed at spreading centers, old crust must be destroyed elsewhere. Seafloor spreading suggests that crust is destroyed at ocean trenches, or subduction zones, where older, cooler, denser crust sinks back into the earth's interior.

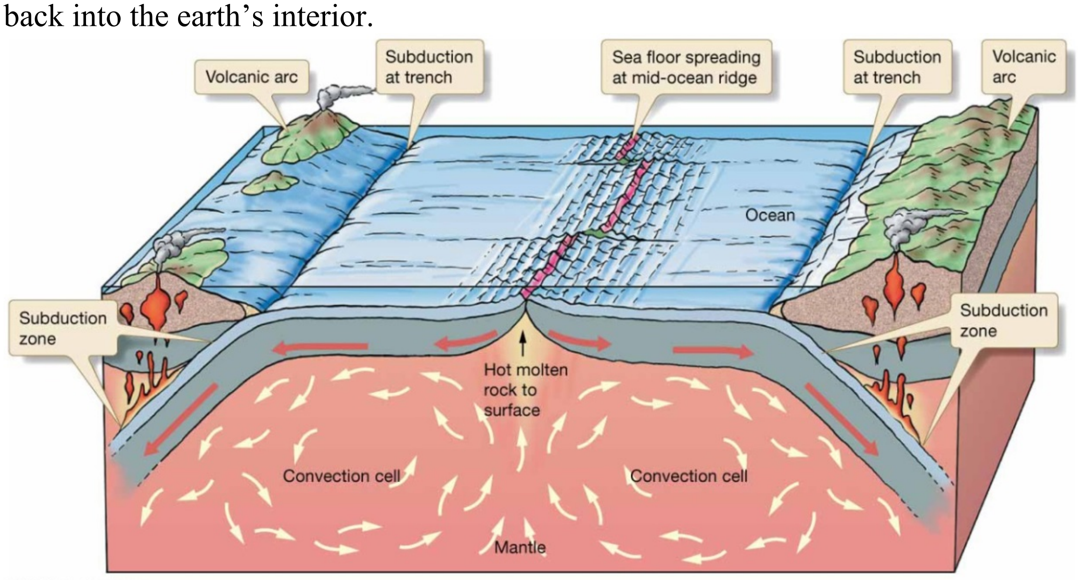
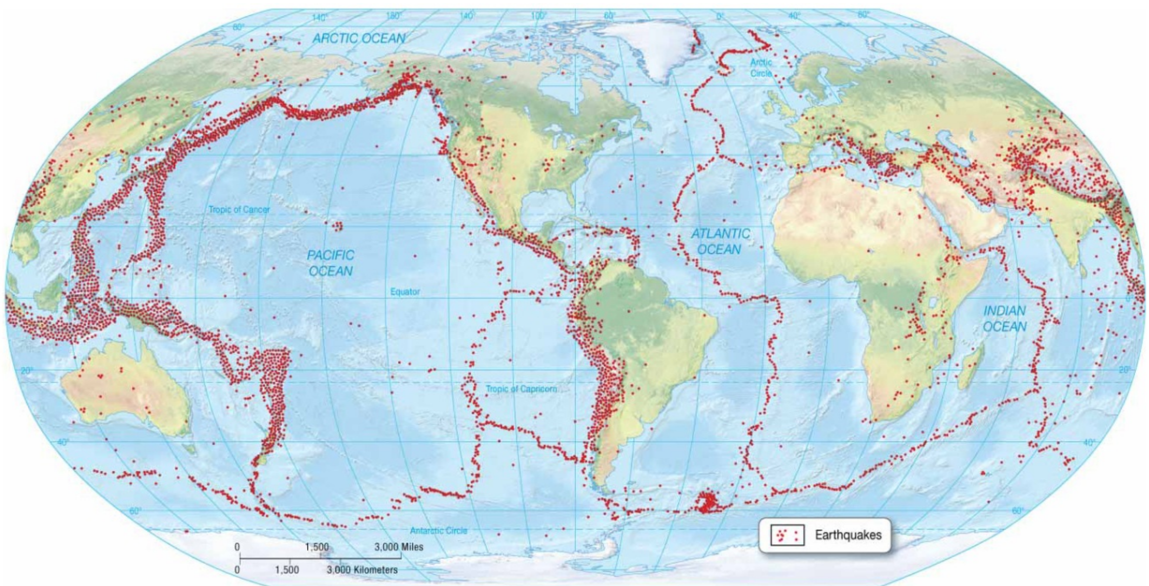


Figure 2.6. Spreading centers (mid-ocean ridges) and subduction zones (trenches) and island arcs associated with the theory of seafloor spreading.

There are many lines of evidence that support seafloor spreading. First, an examination of the distribution of earthquake epicenters shows that they are mainly distributed along ridges and trenches (Figure 2.7); shallow earthquakes (< 100km deep) occur primarily along ridges, while deep earthquakes occur along trenches. Moreover, the oldest oceanic crust is approximately 200 million years old whereas continental rocks are much older (up to 5 billion years old), supporting the idea that oceanic crust is constantly recycled back in the mantle. The age and thickness of sediments were also found to increase with increased distance from mid-ocean ridge systems (Figure 2.8). Finally, there is a mirror pattern of recorded magnetic polar reversals in oceanic crust on either side of ridges (Figure 2.9), supporting the theory that this crust was formed along the ridge and then spread apart.



(a) Distribution of earthquakes with magnitude equal to or greater than M_W = 5.0 for the period 1980-1990.

Figure 2.7. Global distribution of earthquakes.

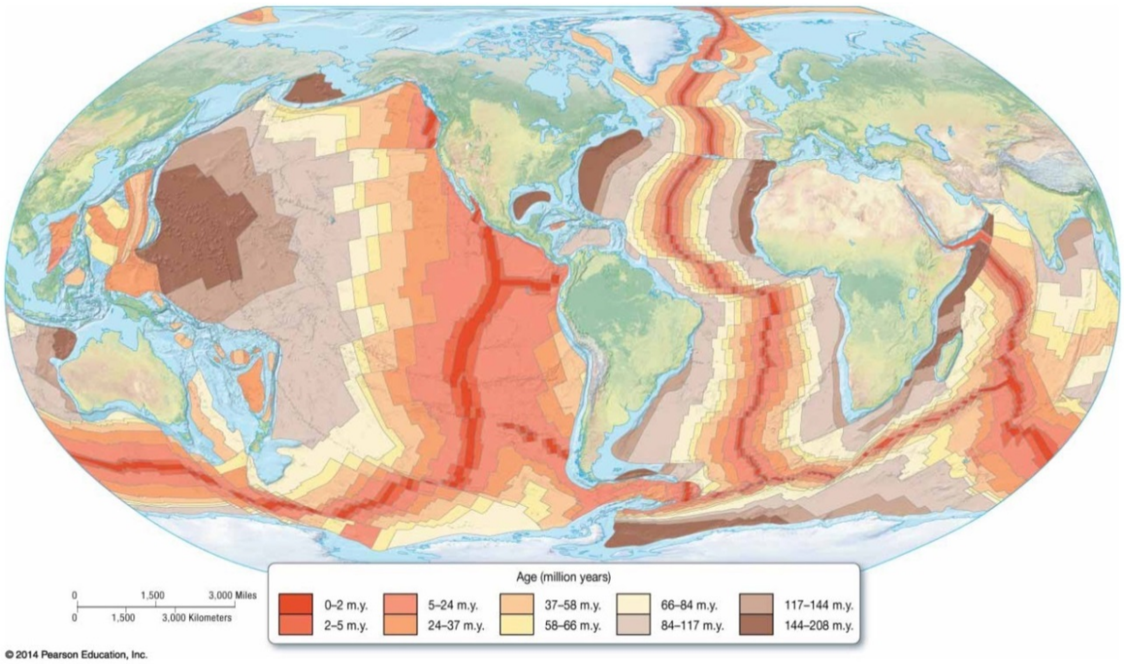


Figure 2.8. Age of the oceanic crust increases with distance from mid-ocean ridges

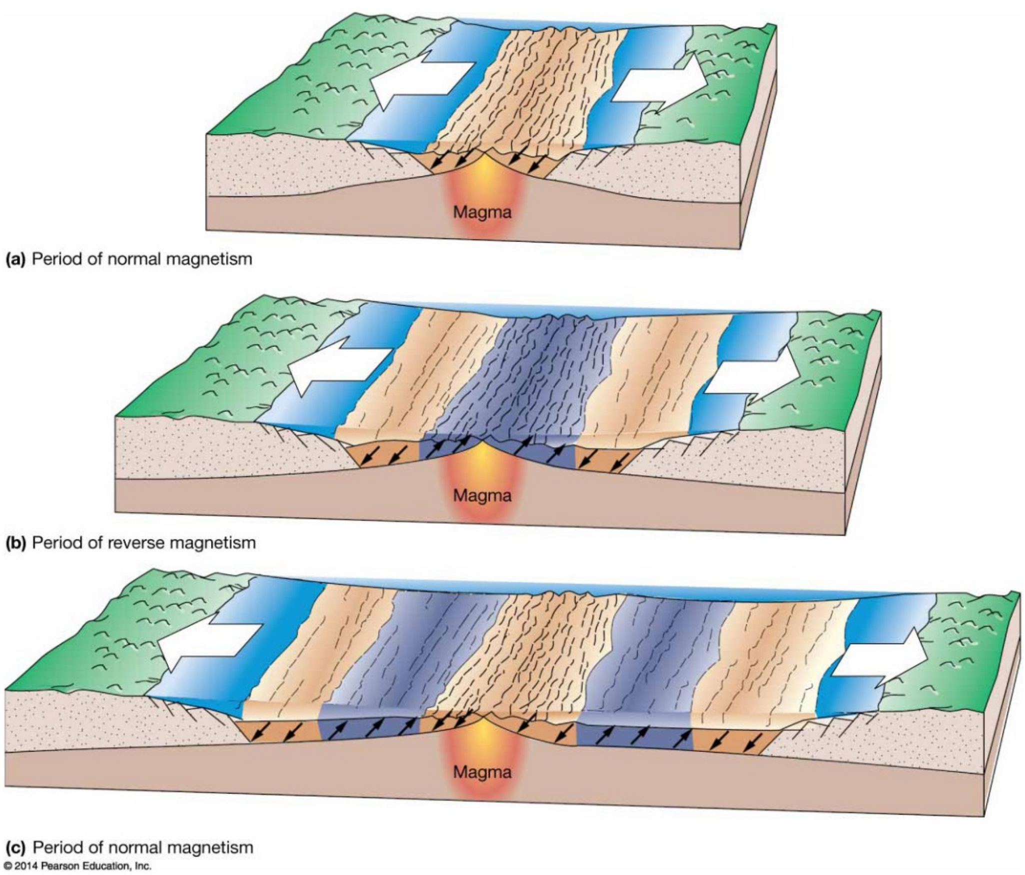
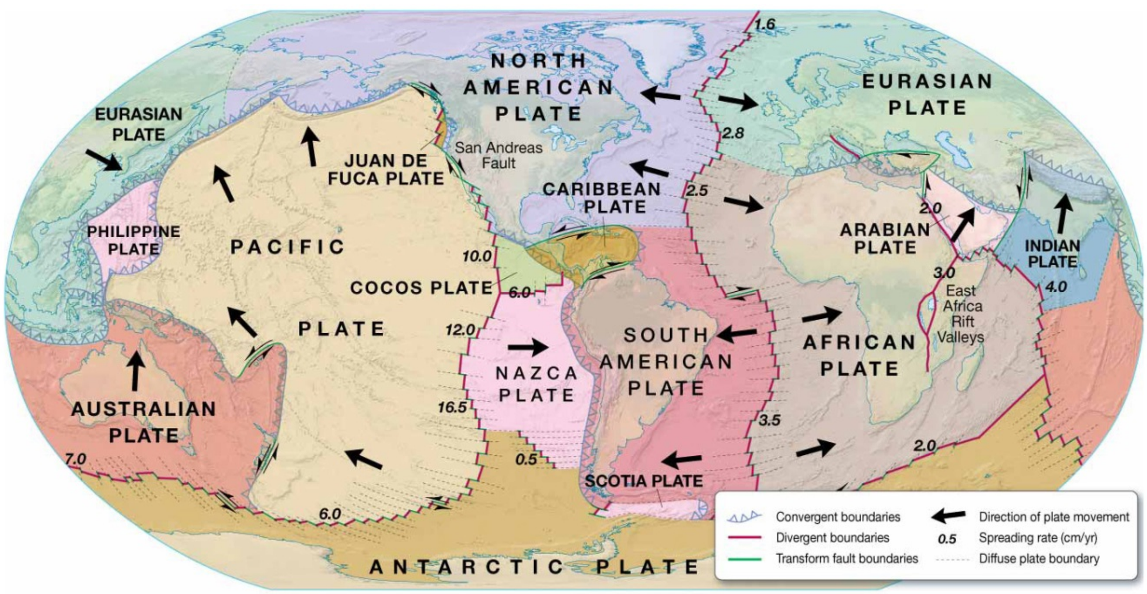


Figure 2.9. Patterns of magnetic reversals are recorded along mid-ocean ridge since basalt magnetizes according to the current magnetic field when it solidifies.

2.3. Plate Tectonics

In 1965, John Tuzo Wilson, a professor at the University of Toronto, combined the ideas of continental drift with those of seafloor spreading to produce the currently accepted theory of Plate Tectonics. Movements of the lithosphere produce both drifting continents and seafloor spreading.

We now know that the lithosphere comprises several lithospheric plates, each 80-100 km thick, capped by continental crust, oceanic crust, or both. Major plates include the Pacific, Eurasian, African, Australian, North American, South American and Antarctic plates, whereas minor plates include the Cocos, Caribbean, Nazca, Philippine, Arabian and Indian plates. These plates move mostly independently of each other, and may diverge, converge or move side by side. Earthquakes are concentrated along plate boundaries along trenches, ridges or faults (Figure 2.10; compare plate boundaries with location of earthquakes in figure 2.7).



(b) Plate boundaries define the major tectonic plates (shaded), with arrows indicating the direction of motion and numbers representing the rate of motion in centimeters per year.

Figure 2.10. Tectonic plates of the world.

Features of plate boundaries

Plate boundaries can be divergent (pulling apart), convergent (moving together) or transform (moving side by side) (Figure 2.11). Specific geological features are associated with each type of plate boundary. Plates move slowly, on average 2 to 10 cm/year, but they may move faster or slower depending on the plate and type of ridge system.

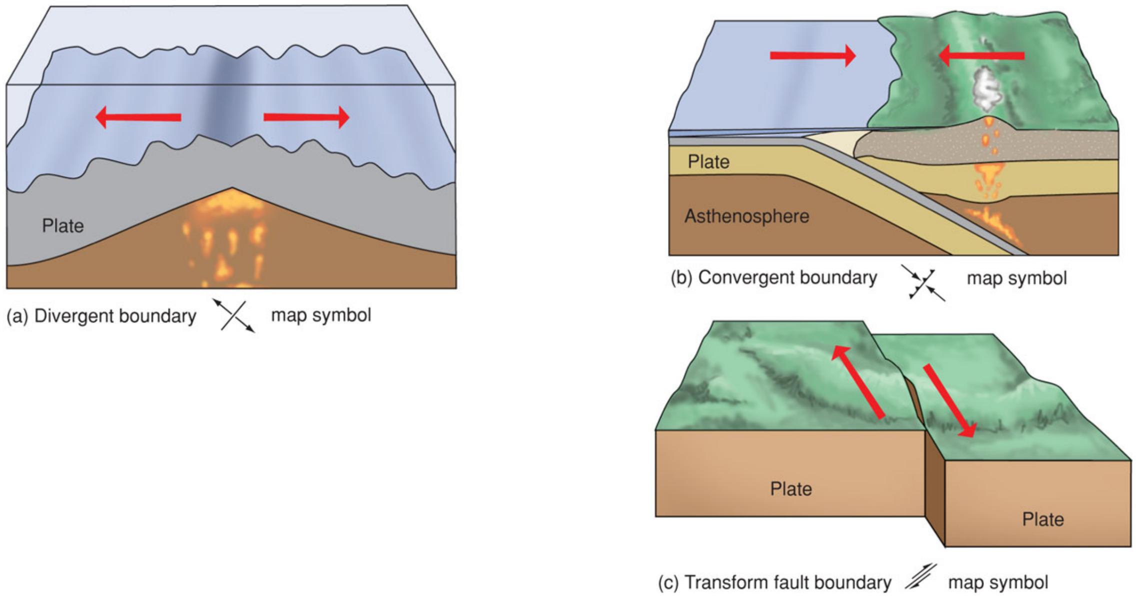


Figure 2.11. Three types of plate boundaries

Divergent boundary features

Divergent plate boundaries occur when plates move apart at the spreading centers. The lithosphere splits, separates and is forced apart as new crust material intrudes the crack. This mechanism over time can form new ocean basins (Figure 2.12). This occurs mainly at mid-ocean ridges, where seafloor spreading creates new lithosphere, but can occur on land as well, e.g. the East African Rift Valley (Figure 2.13). Rift valleys are a central linear depression commonly found in divergent plate boundaries. Mid-ocean ridges may be periodically offset by fracture zones where rocks move side by side relative to each other (green lines in figure 2.13; see also section on transform faults below).

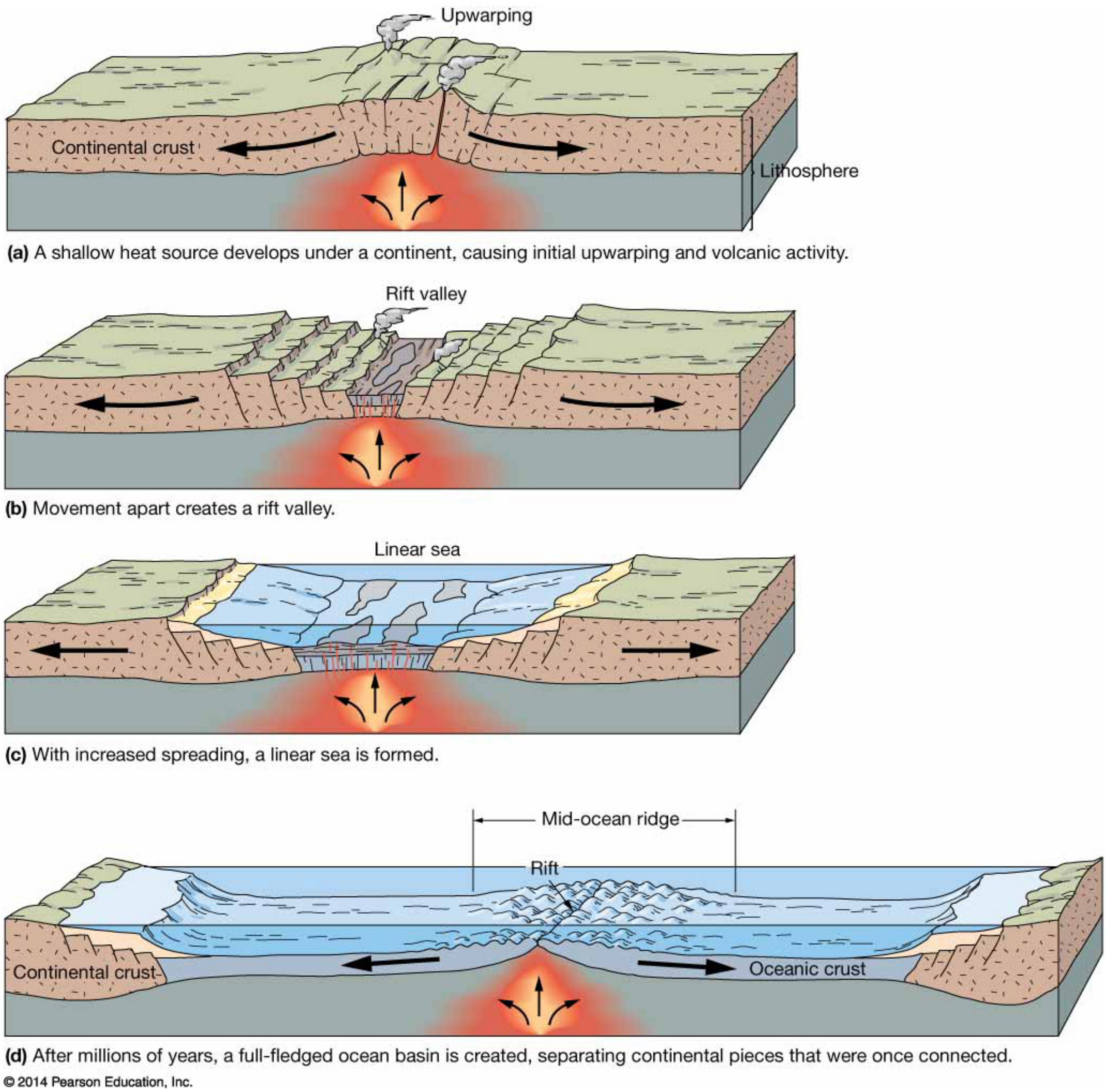
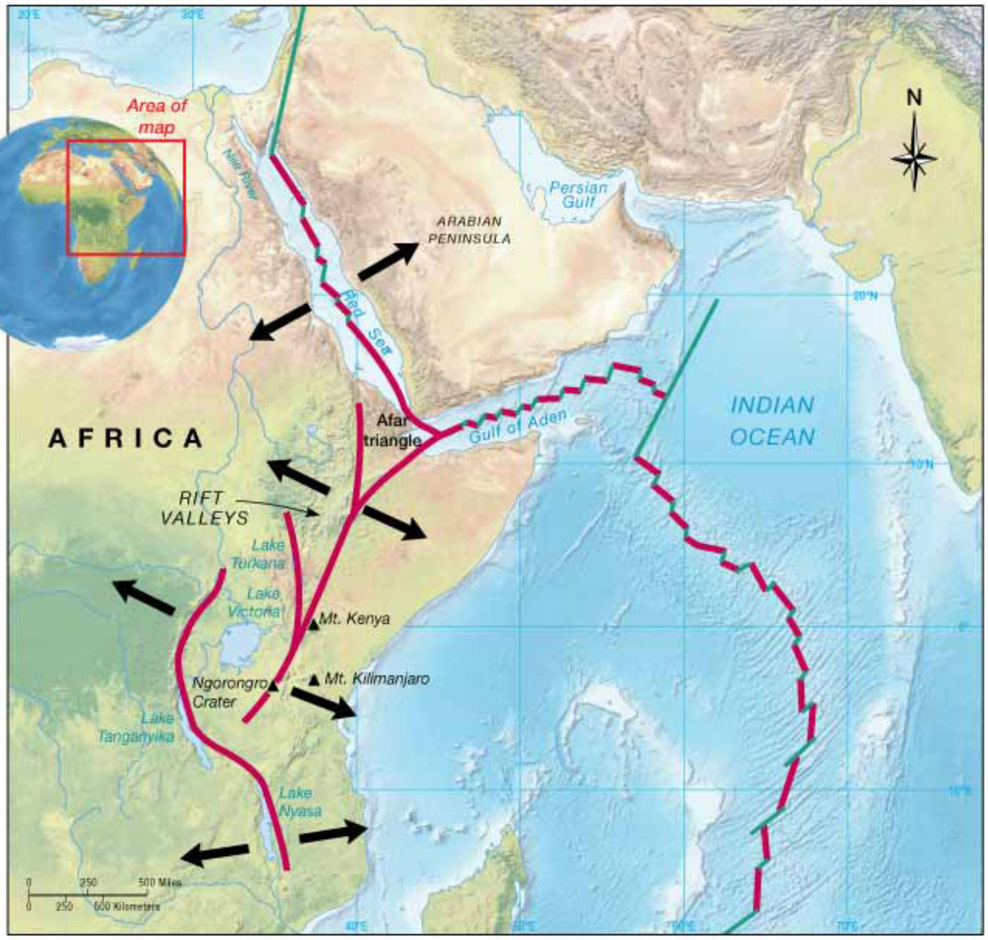
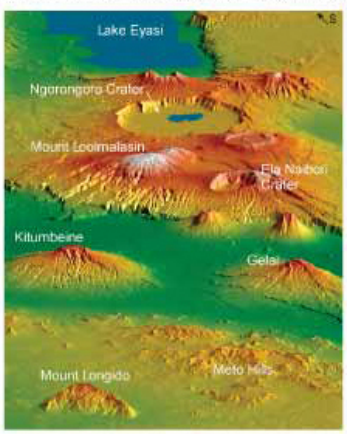


Figure 2.12. Formation of an ocean basin through divergent plates. Note that the rift valley is clearly seen when the plates start moving apart, and a rift can be identified on the axis of the mid-ocean ridge.



(a) Parts of east Africa are splitting apart (arrows), creating a series of linear downdropped rift valleys (red lines) along with prominent volcanoes (triangles). Similarly, the Red Sea and Gulf of Aden have split apart so far that they are now below sea level. The mid-ocean ridge in the Indian Ocean has experienced similar stages of development.



(b) Land elevation perspective view, looking southwest along part of the East Africa Rift in Tanzania, showing the downdropped Lake Eyasi and numerous volcanic peaks and craters of the Crater Highlands. Color indicates elevation, where green is lower elevation and brown/white is higher.

(c) Photo of a rift that formed in 2005 after seismic activity and a volcanic eruption of Mount Dabbahu in Ethiopia's Afar triangle, Africa; note people at left for scale.

Figure 2.13. The East Africa Rift Valley and associated features.

Convergent boundary features

Convergent plate boundaries occur where plates move toward each other, mainly along subduction zones. The result of convergent plates depends on the type of crust involved. Between two oceanic plates or an oceanic plate and a continental plate, the denser plate sinks below the less dense plate. In the case of an interaction between and oceanic plate and a continental plate, the oceanic plate is denser and therefore sinks beneath the continental plate. The friction of the plates as one is subducted creates a series of deep earthquake along the Wadati-Benioff zone, which marks the location of the subducted plate. As this plate sinks further down in the asthenosphere, temperature increases and melts the plate and its associated sediments. This molten material of lesser density mixes with mantle magma and rises to the surface, where it forms belt of volcanoes (e.g. Caribbean or Andes; Figure 2.14, 2.15). When two continental plates converge, their thickness and low density prevents them from being subducted, and instead they collide to form tall uplifted mountain ranges (e.g. Himalayas; Figure 2.16).

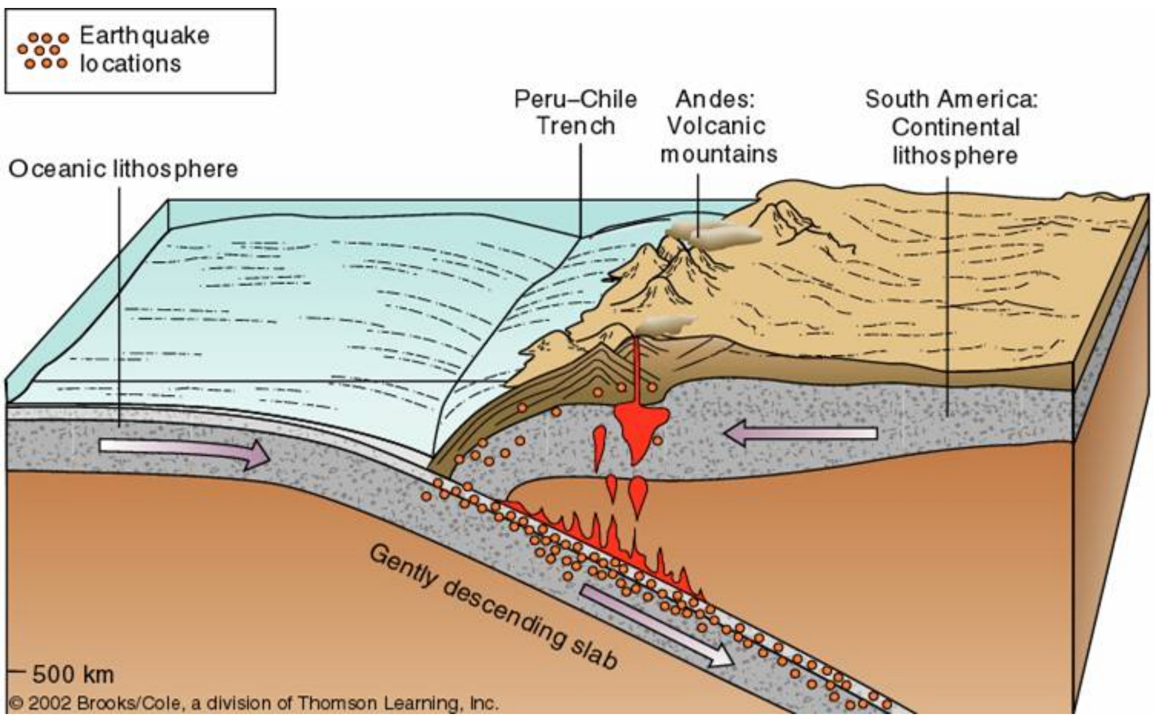


Figure 2.14. A convergent plate boundary at a subduction zone between a continental and an oceanic crust, resulting in a series of volcanic peaks (e.g. Andes).

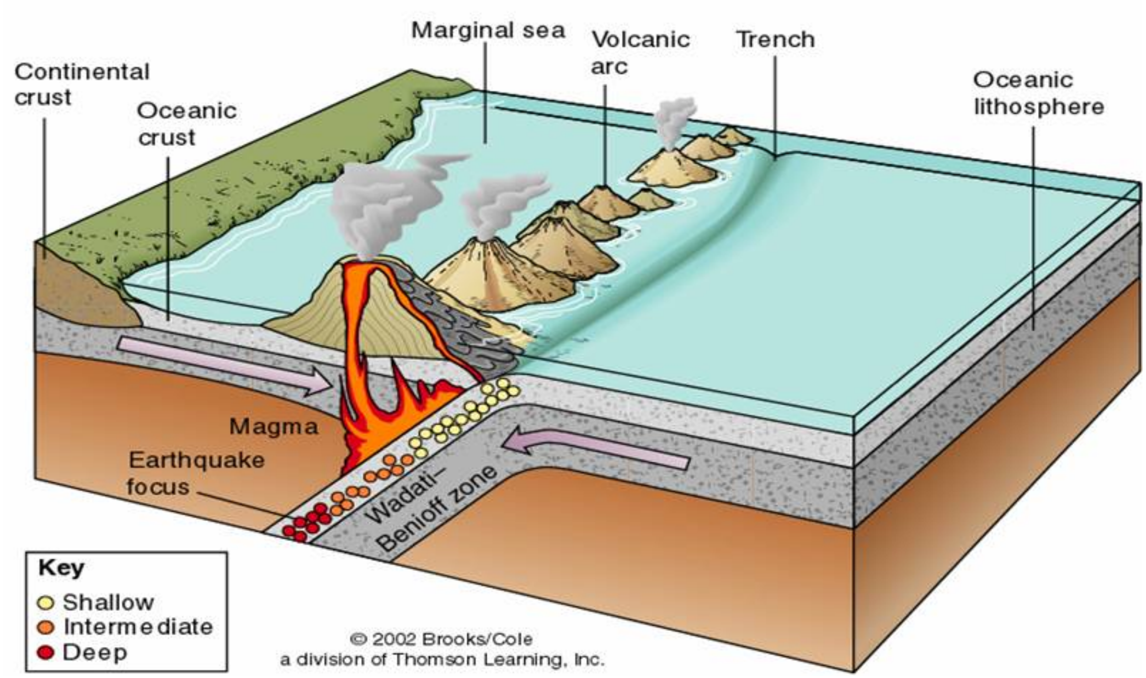


Figure 2.15. A convergent plate boundary at a subduction zone between two oceanic crusts, resulting in the formation of an island arc (e.g. Caribbean).

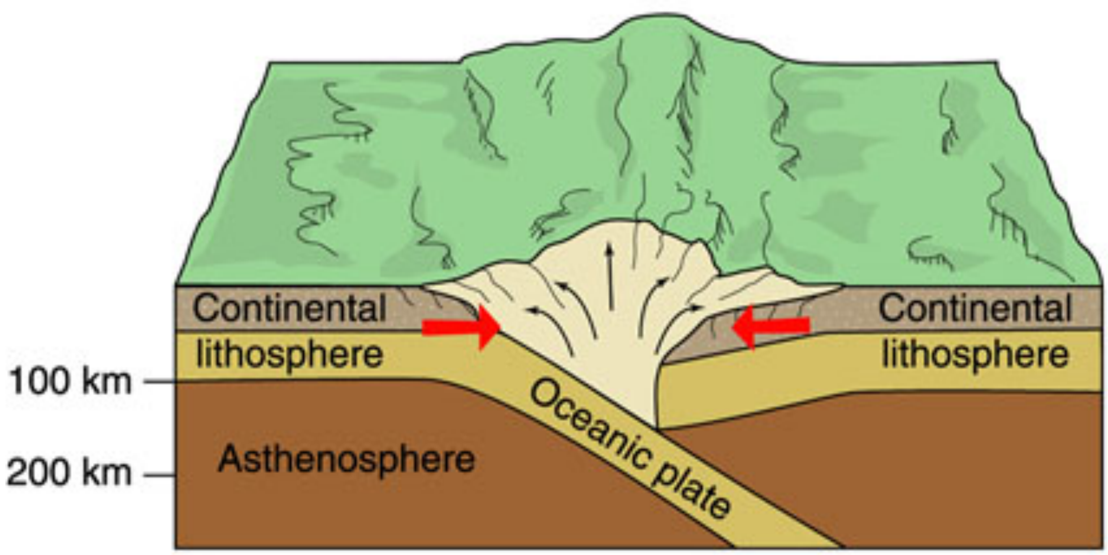


Figure 2.16. A convergent plate boundary between two continental plates, which resist subduction, forming a mountain range (e.g. the Himalayas).

Transform boundary features

Transform faults occur when two plates move side by side relative to each other. This type of boundary movement is associated with spreading centers, where regular splits in the mid-ocean-ridge called fracture zones occur perpendicular to the ridge. Transform faults are the active area of a fracture zone, where the adjacent plates move in opposite direction. This results in a high occurrence of earthquakes. Transform faults are found regularly along mid-ocean ridges (see green lines in Figure 2.13a), but can also occur on land as in the San Andreas Fault (Figure 2.17).

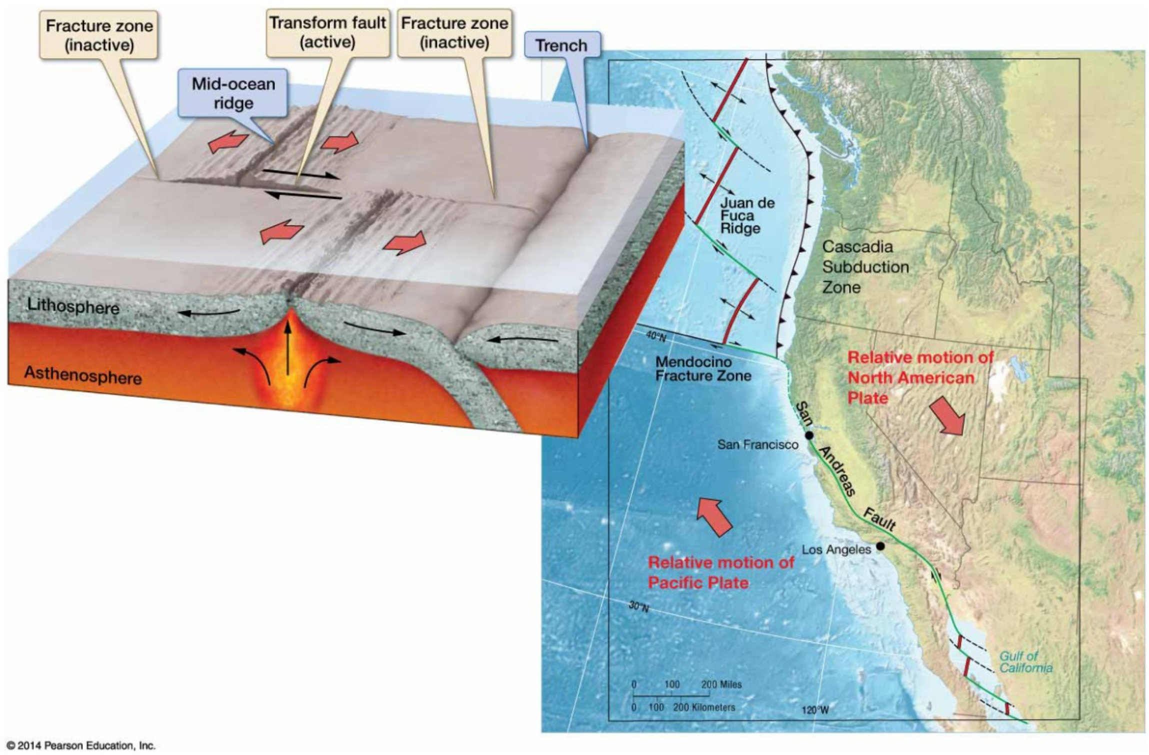


Figure 2.17. A transform fault, where two plates move sideways relative to each other, occurs along fracture zones of spreading centers where plates move in opposite direction. Transform faults are colored in green in the map on the right.

Though most volcanic activity around the world is associated with divergent and convergent plate boundaries, there are isolated areas of volcanic activity scattered throughout the oceans known as hot spots. Hot spots are found under continents and oceans, in the center of plates and along ridges. They periodically channel hot material to the surface, which may break through the lithosphere to form a volcano (Figure 2.18). Hot spots are relatively stationary, but as the overlying lithosphere moves, successive eruptions create chains of volcanoes such as the Hawaiian Islands and the Galápagos. The orientation of these island chains indicates the direction of the plate movement, with age of islands increasing with distance from the hot spot (Figure 2.19).

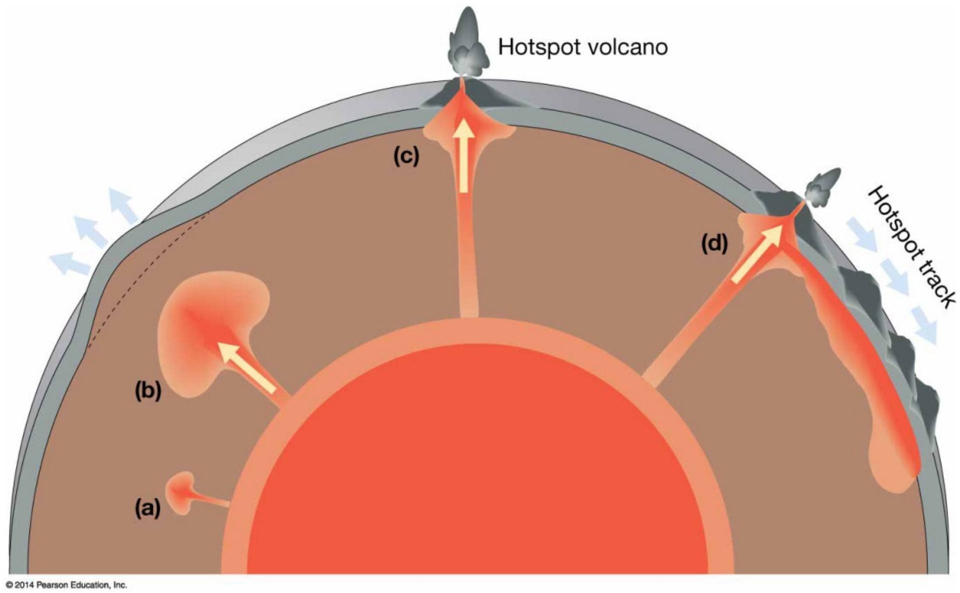


Figure 2.18. Formation of hotspots: a) hot buoyant material detaches from the deep mantle, b) rises to the surface and c) forms a hot spot volcano. d) as the lithosphere move on top of a stationary hot spot, a chain of volcanoes is created.

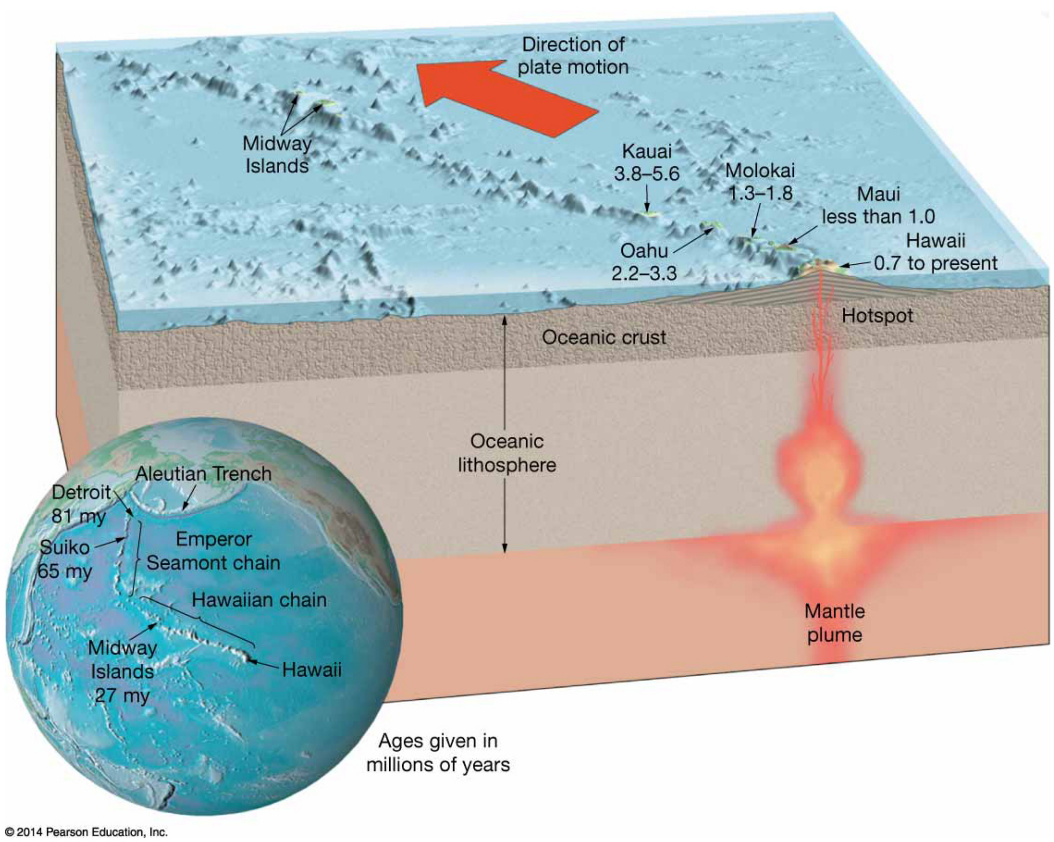


Figure 2.19. Formation of successive volcanic peaks by a hot spot.

Seamounts and tablemounts

As volcanoes move away from a hot spot or spreading center, they subside and erode, creating seamounts (volcano that does not reach the surface of the water) or guyots (seamount with a flat top from erosion when it was near sea level; Figure 2.20).

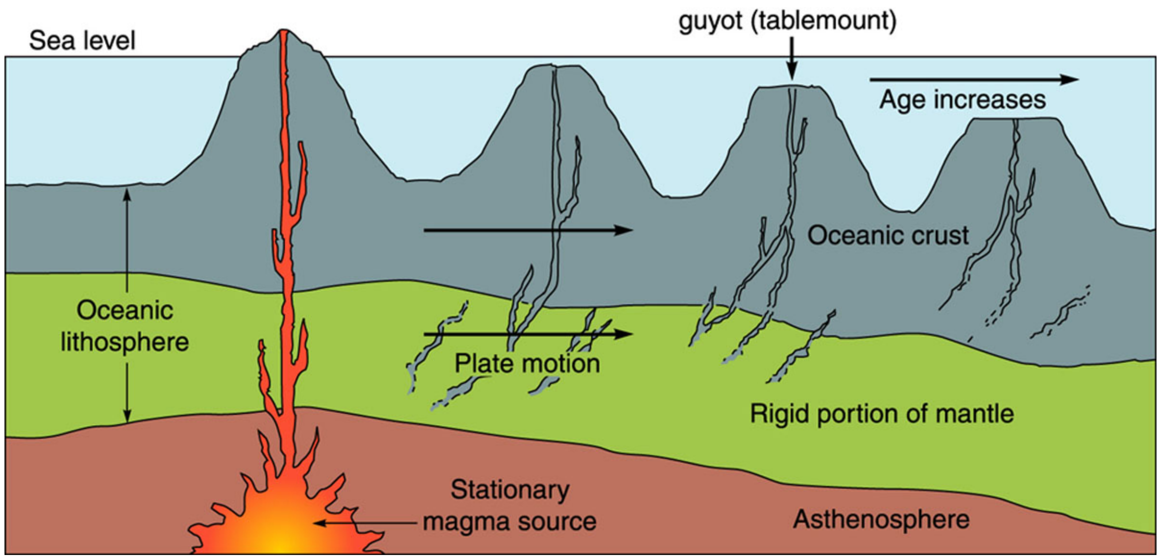


Figure 2.20. As volcanoes formed by hot spots move away from the heat source, they subside due to the increased weight on the lithosphere. Volcanoes that don't reach the surface of the water are called seamounts when they are conical, and table mounts or guyots when they are flat at the top. Seamounts and tablemounts can also be associated with spreading centers.

Coral Reef Development

Darwin observed a lot of coral reefs on his voyage around the globe on the Beagle. He was the first to form the hypothesis that atolls originated as fringing reefs around a subsiding volcano, which proved to be correct.

Coral reefs slowly grow vertically at a rate of 3-15 mm/yr. They grow only in shallow water, where sunlight is abundant, and therefore young reefs grow close to shore around and are called fringing reefs. As an island subsides (sinks) or the surrounding sea level rises, the reef grows vertically to remain in the shallow zone, and develops into a barrier reef, separated from shore by a lagoon. With time, as the island sinks below sea level, the barrier reef keeps growing vertically to form a ring of coral reef surrounding a lagoon; this is called an atoll (Figure 2.21; Figure 2.22).

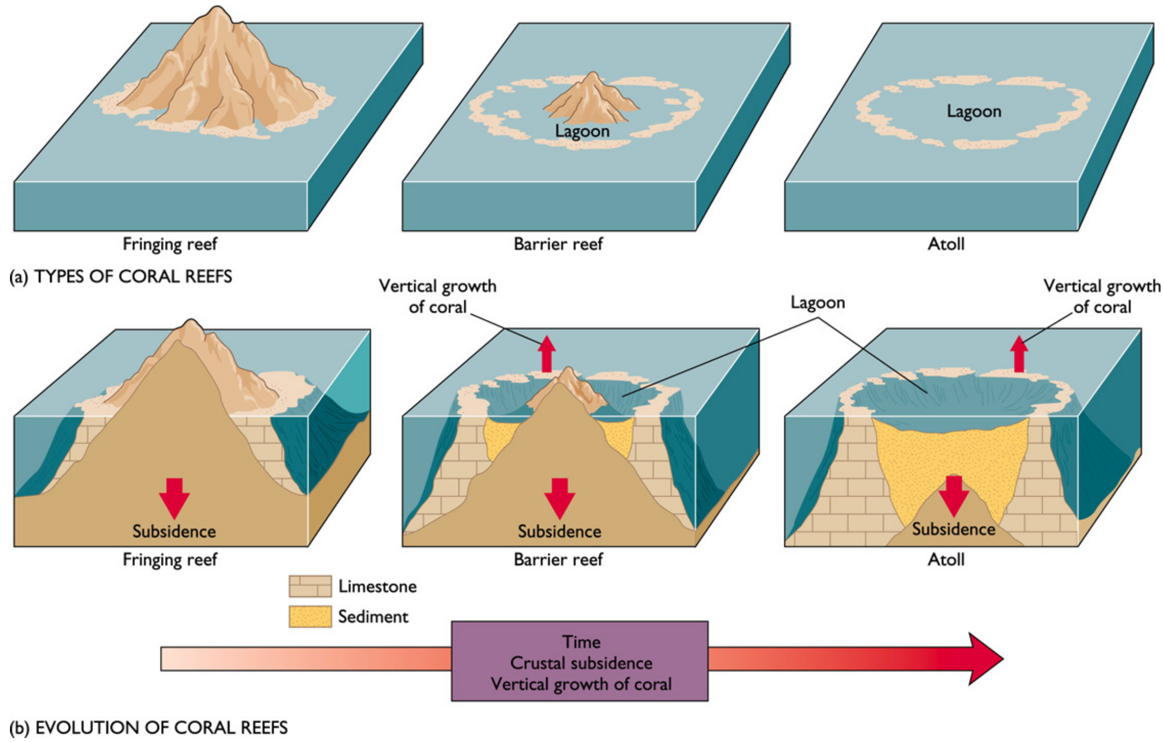


Figure 2.21. Formation of fringing reefs, barrier reefs and atolls.

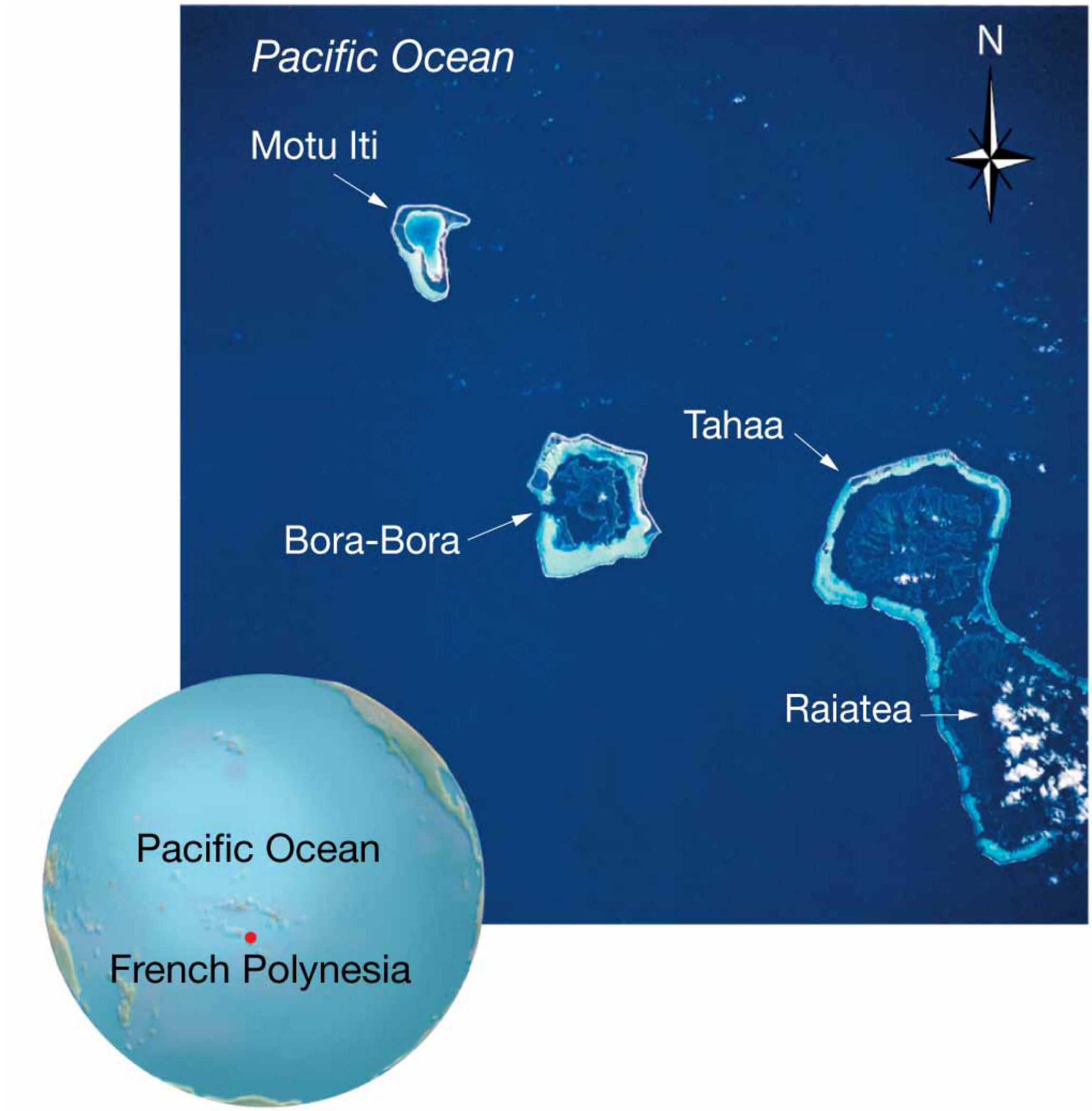


Figure 2.22. Islands in French Polynesia show various stages of atoll development. Bora Bora, Raiatea and Tahaa have a barrier reef around a central island while Motu Iti is a true atoll.

Ocean cycles

Plate tectonics teaches us that oceans are dynamic, and over the course of several hundred million years they can form, grow, and decline. John Tuzo Wilson classified oceans according to their stages of formation (Figure 2.23). An embryonic ocean occurs with uplifting and faulting in a land mass and the formation of a linear rift valley, as in East Africa. A juvenile ocean occurs when the fissuring between the two land masses is complete, as with the Red Sea. A mature ocean (e.g. the Atlantic Ocean) displays a broad ocean basin, with divergence still occurring but with little subduction. A declining ocean, such as the Pacific Ocean, occurs when the basin is getting smaller because of considerable subduction around the edges, which are accompanied by trenches and island arcs. The Mediterranean is a terminal ocean, which is small in size and declining, where the surrounding continents and island arcs collide and form young mountain ranges. A sutured ocean occurs when an ocean basin has been uplifted into a mountain range, such as the Himalayas, between Indian and Asia.

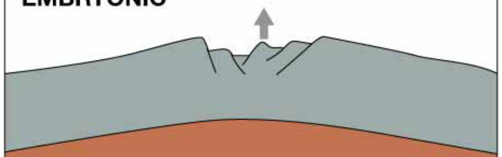
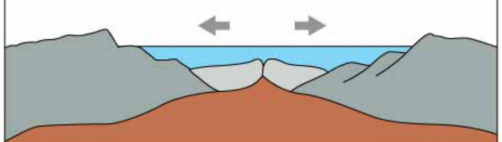
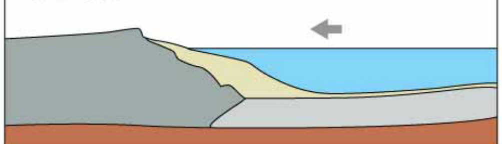
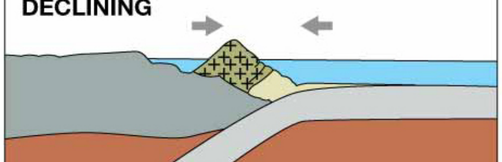
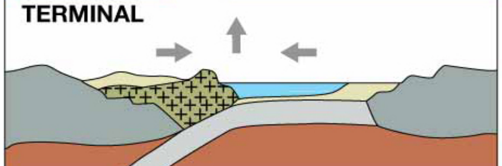
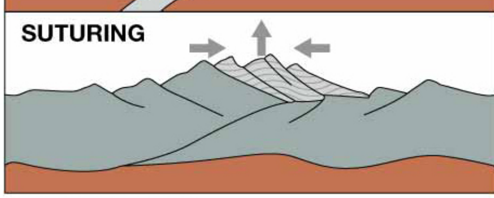
Stage, showing cross-sectional view	Motion	Physiography	Example
EMBRYONIC 	Uplift	Complex system of linear rift valleys on continent	East Africa rift valleys
JUVENILE 	Divergence (spreading)	Narrow seas with matching coasts	Red Sea
MATURE 	Divergence (spreading)	Ocean basin with continental margins	Atlantic and Arctic Oceans
DECLINING 	Convergence (subduction)	Island arcs and trenches around basin edge	Pacific Ocean
TERMINAL 	Convergence (collision) and uplift	Narrow, irregular seas with young mountains	Mediterranean Sea
SUTURING 	Convergence and uplift	Young to mature mountain belts	Himalaya Mountains

Figure 2.23. The Wilson cycle of ocean basin evolution

2.4. Review Questions

1. Who first proposed the theory of continental drift?
2. What is the name of the super continent of the continental drift theory?
3. When Pangea split in two, what were the names of the two new continents?
4. Name 3 lines of evidence supporting continental drift
5. What was the mechanism proposed by Wegener to explain continental drift?
6. What is the name of the underwater mountain range found at spreading centers in the middle of the oceans?
7. Where would you find a rift valley?
8. Who proposed the theory of Seafloor Spreading?
9. Give four pieces of evidence supporting seafloor spreading
10. Is lithosphere lost or formed at a subduction zone?
11. What is the name of the theory that unites continental drift and seafloor spreading
12. Name the 3 types of possible plate interactions
13. What features are typical of a divergent plate boundary?
14. What features are typical of a convergent plate boundary with 2 continental crusts?
15. What type of plate boundary is responsible for forming the Caribbean island arc?
16. What is the name of a pile of sediment scraped off by a plate overriding another plate at subduction zones?
17. Name a chain of islands formed by a hot spot
18. Where are most volcanoes found around the world?
19. What is the name of a seamount with a flattened top?
20. How is an coral atoll formed?
21. Give an example of an ocean in the mature stage?
22. Give an example of an ocean in the sutured stage?

3. Marine Provinces (Trujillo Chapter 3)

3.1 Determining ocean bathymetry

Bathymetry is the study of the depth and shape of the bottom of the ocean. Depth can be measured in various ways. The earliest depth readings were done using soundings: lowering a heavy weight on a line until it reached the bottom. In the early 1900s, the first echo sounders measured ocean depth by sending a sound signal to the bottom and measuring how long it takes for its echo to return to the surface. Modern echo sounders have tremendously increased the precision of the measurements, but these measurements are severely limited by ship time and resources. For this reason, satellite remote-sensing is increasingly used to infer bathymetry. Features on the ocean floor create sea level abnormalities above them, which can be measured accurately by satellites after correcting for waves, tides and other interferences (Figure 3.1). Because satellites can obtain much more data than ships, bathymetric charts derived from satellite data are more much detailed than those produced by acoustics alone (Figures 3.2 & 3.3).

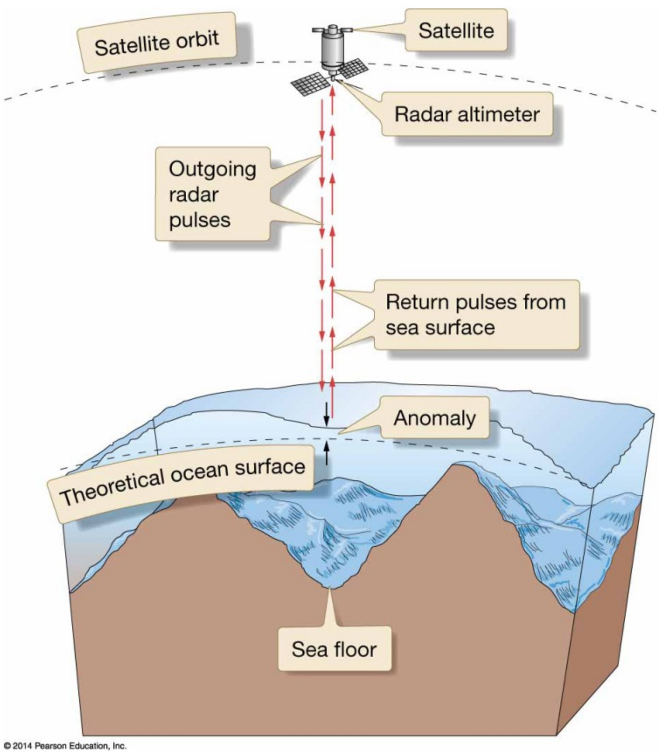


Figure 3.1. Satellite measurements abnormalities on the ocean's surface is used to infer submerged bathymetric features.

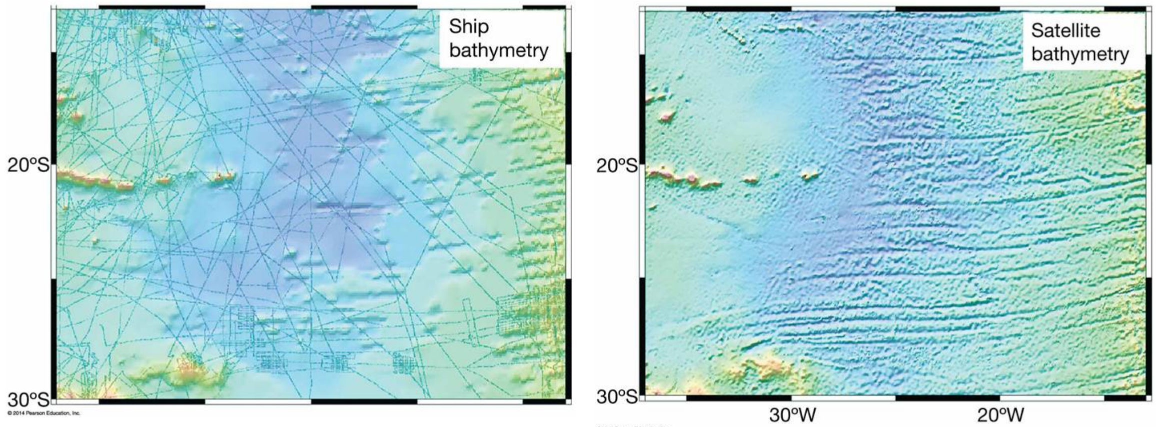


Figure 3.2. Bathymetric charts derived from satellite data (right) are much more detailed than those derived from ship data (left).

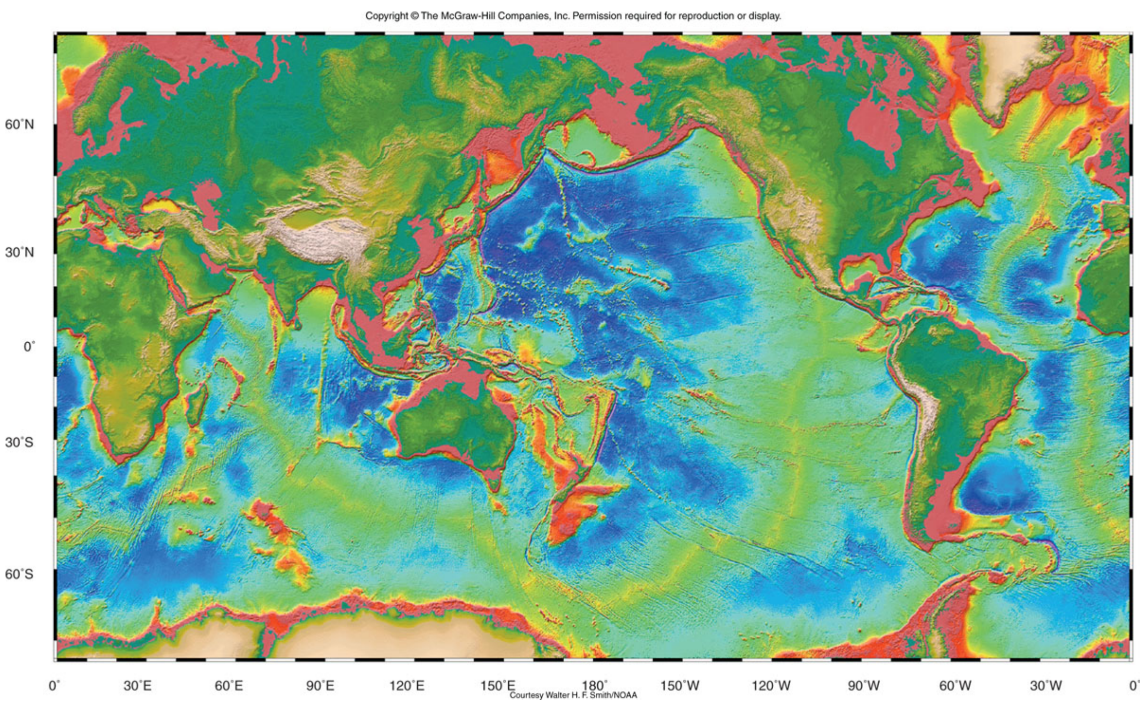


Figure 3.3. Global bathymetric chart derived from sea surface abnormalities, which reveals continental shelves and other shallow areas in pink, mid-ocean ridges in yellow-green and the deepest parts of the ocean in blue.

3.2. Features of the continental margins

The ocean floor can generally be divided in three regions: continental margins, the ocean basins, and mid-ocean ridges (Figure 3.4). The submerged edges of continents and the steep slopes that lead to the sea floor, both made of continental crust, are the continental margins. Continental margins may be passive or active. Passive margins are found where continents have moved apart (e.g. the Atlantic). Passive margins show little seismic or volcanic activity, and the transition from continental to oceanic crust occurs on the same plate. They are typically wide. Active continental margins, on the other hand, are associated with convergent plate boundaries and subduction of oceanic crust beneath continental crust (e.g. Pacific Ocean). Active continental margins are associated with earthquakes and volcanoes, and are typically narrow (Figure 3.5).

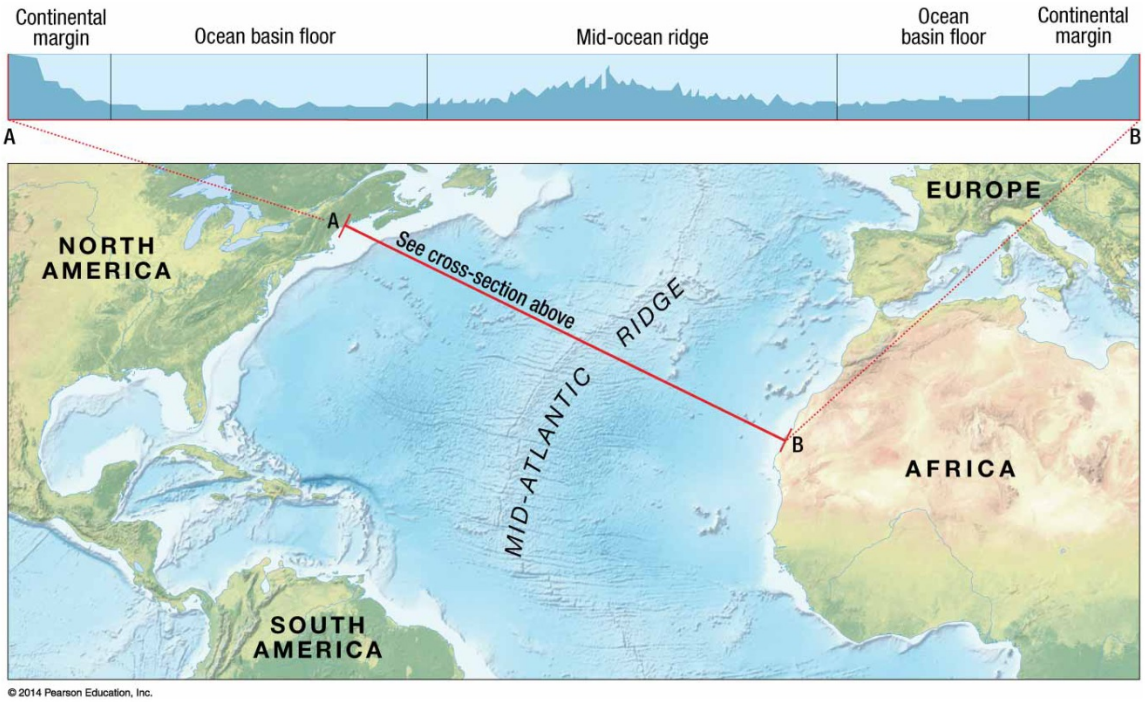


Figure 3.4. A cross-section of the North Atlantic reveals the three major ocean provinces.

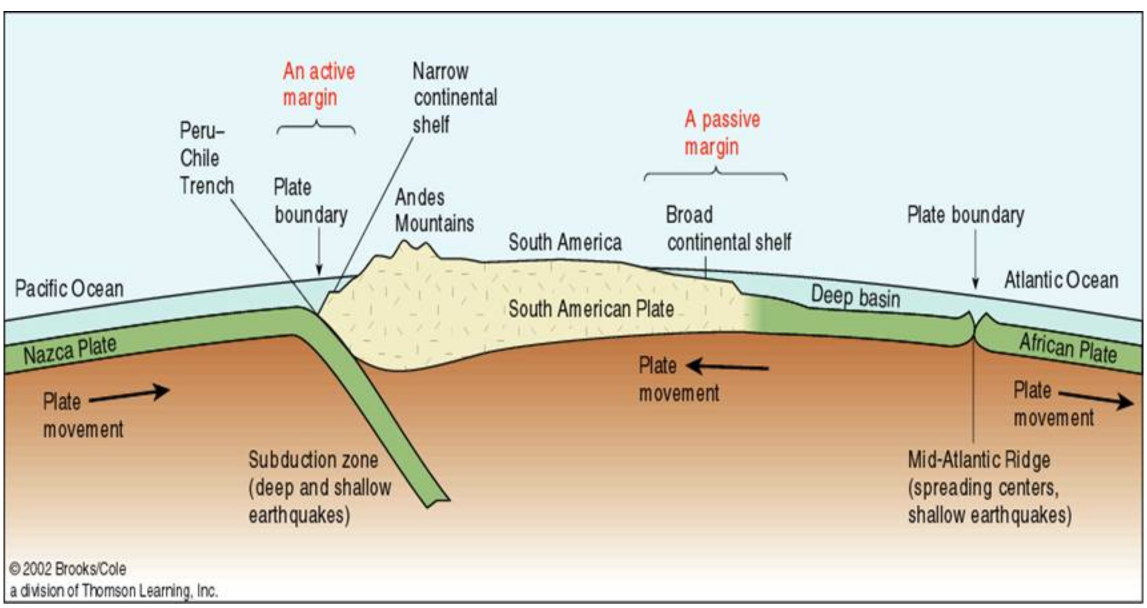


Figure 3.5. Passive and active continental margins, shown on each side of South America. The active margin (left/west) has a narrow shelf and an offshore trench where the oceanic plate is being subducted. The passive margin (right/east) faces the diverging plate boundary of the mid-ocean ridge; it is much wider.

Continental margins are made up of several sections (Figure 3.6). The continental shelf lies right at the edge of the continent and is nearly flat, with an average depth of 130m. The width of the continental shelf varies greatly, and is much greater in passive continental margins. Continental shelves have been alternately submerged and exposed through fluctuations in sea level during glacial ages, and when inundated, they may accumulate sediment derived from land and carried by rivers. The shelf break marks the abrupt change in slope from the nearly flat continental shelf to the continental slope. The angle of the slope varies greatly. Continental slopes have submarine canyons that were formed during periods of low sea level (Figure 3.7). These canyons are V-shaped with steep walls and transport sediments from the shelves to the deep sea floor. Turbidity currents are fast moving flows of sediments on the continental slope that may travel to speeds of 90 km/hr and carry enormous quantities of sediments. They are caused by earthquakes or overloading of sediments on the shelf. At the base of the continental slope, the accumulation of sediment creates a gentle slope. This portion of the continental margin is known as the continental rise, and is most prominent on passive continental margins. The continental rise marks the beginning of true deep ocean basins.

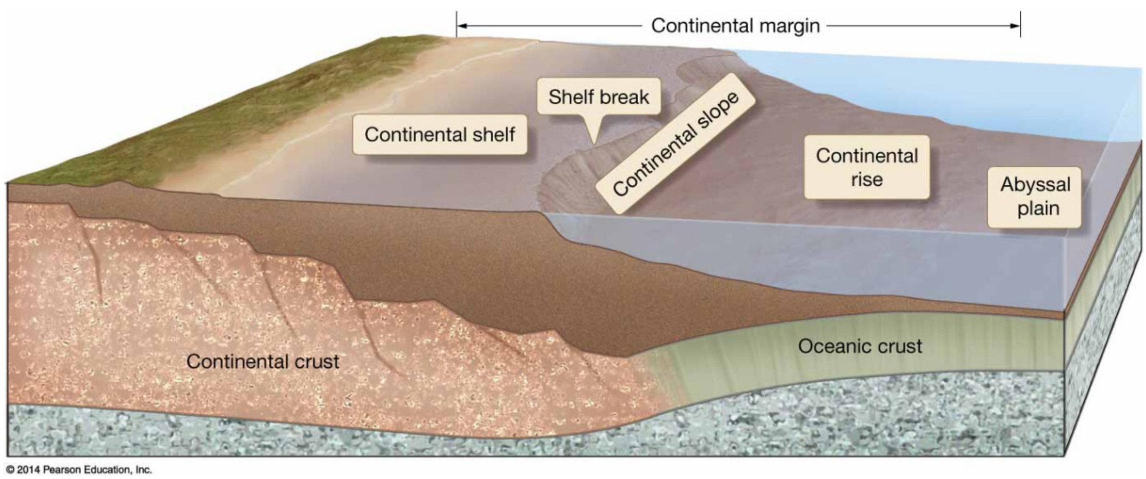


Figure 3.6. Continental margin, with continental shelf, slope and rise.

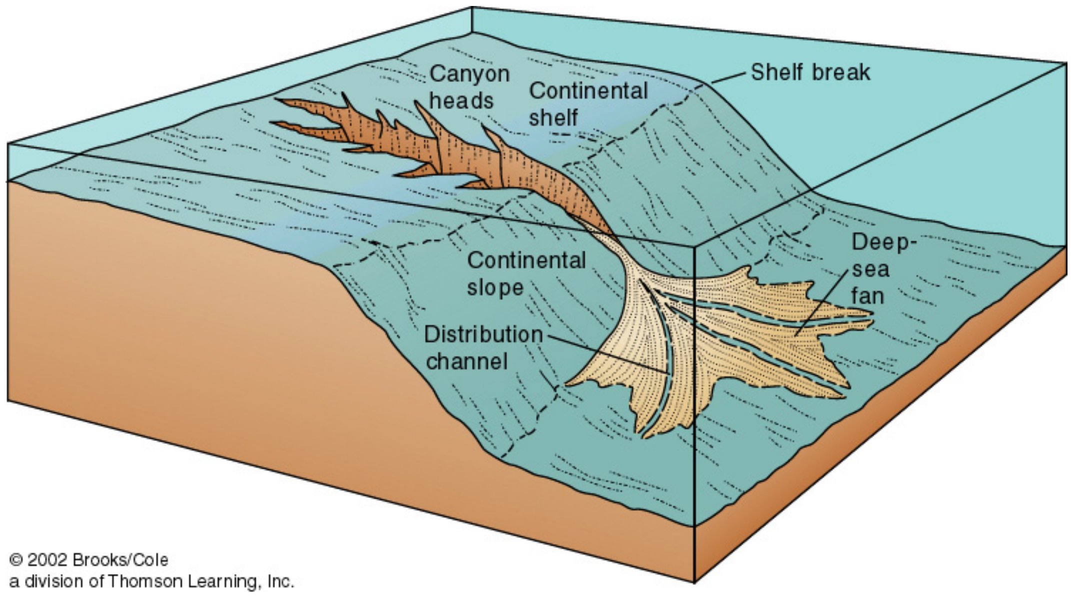


Figure 3.7. Submarine canyons are a common feature of continental shelves and slopes, and turbidity currents commonly flow down them bringing sediment at the base of the canyon.

3.3. Features of deep ocean basins

The deep sea floor covers a huge area of the oceans, and most of it consists of vast flat plains known as the abyssal plains (Figure 3.8). Sediments carried from continental shelves are eventually deposited on the deep sea floor, covering irregular topography and forming this flat abyssal plain. Abyssal hills and seamounts are scattered throughout the sea floor. Abyssal hills are short (less than 1,000m high) and are a very common feature of the deep oceans. Most are volcanic in origin. Seamounts are steep volcanoes that sometimes pierce the surface of the water and become islands. Submerged seamounts that have a flat top are known as guyots, and were flattened by wave erosion when they were at the surface (Figure 3.9). Deep sea trenches occur along convergent plate boundaries and typically have steep sides. The deepest part of the oceans, known as the Challenger Deep, occurs in the Mariana trench off Japan and is 11,020m deep. Ocean trenches are associated with volcanic arcs, on the side of the overriding plate, as material from the subducted plate melts and rises. These volcanoes can form island arcs such as the eastern Caribbean or volcanic mountain ranges such as the Andes. The Pacific Ocean is lined with such trenches, which create volcanoes and earthquakes around its perimeter, which has been dubbed the Pacific Ring of Fire (Figure 3.10).

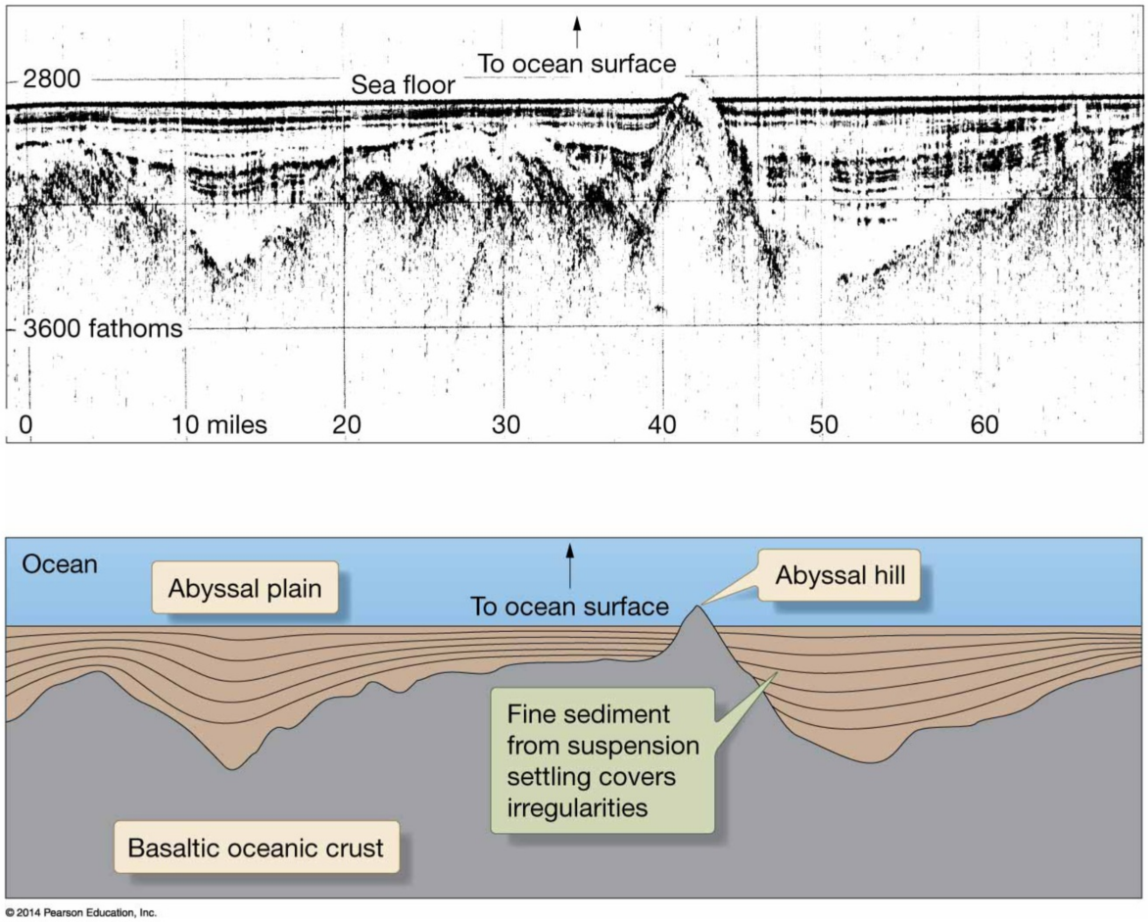


Figure 3.8. Seismic cross-section and drawing of an abyssal plain, where fine sediment cover most bathymetric features.

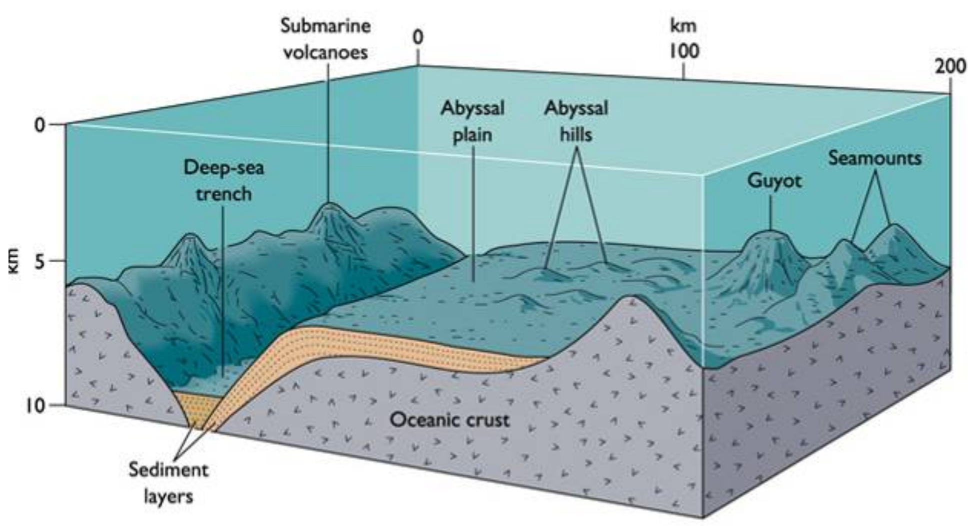
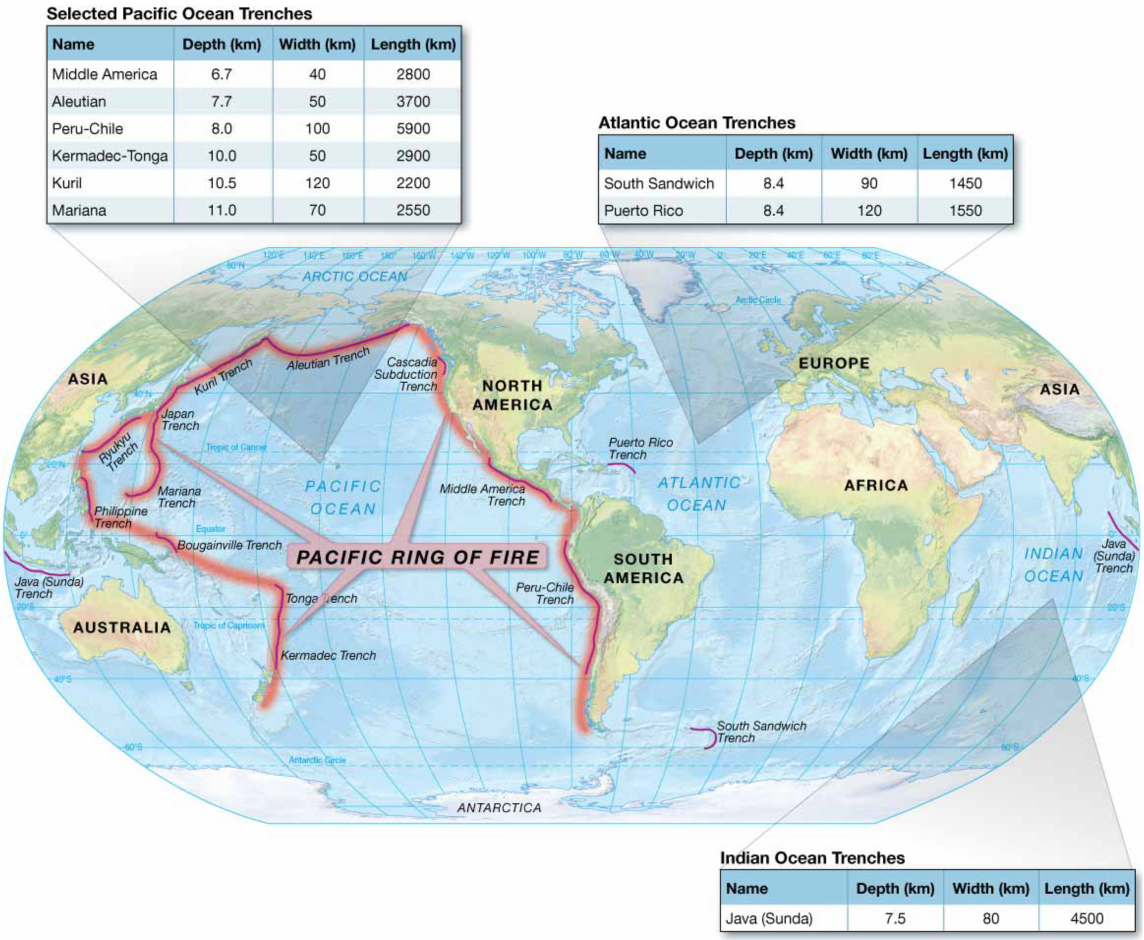


Figure 3.9. Features of the deep sea floor.



Selected Pacific Ocean Trenches

Name	Depth (km)	Width (km)	Length (km)
Middle America	6.7	40	2800
Aleutian	7.7	50	3700
Peru-Chile	8.0	100	5900
Kermadec-Tonga	10.0	50	2900
Kuril	10.5	120	2200
Mariana	11.0	70	2550

Atlantic Ocean Trenches

Name	Depth (km)	Width (km)	Length (km)
South Sandwich	8.4	90	1450
Puerto Rico	8.4	120	1550

Indian Ocean Trenches

Name	Depth (km)	Width (km)	Length (km)
Java (Sunda)	7.5	80	4500

Figure 3.10. Deep sea trenches are especially common around the perimeter of the Pacific Ocean, but are also found in the Indian and Atlantic Oceans.

3.4. Features of the Mid-Ocean Ridges

Mid-ocean ridges and rises are the longest continuous mountain chain on earth and are approximately 75,000 km long. Mid-ocean ridges are typically 2-3 km high and have a central rift valley along the axis of spreading (see section 2.3, and Figure 3.11). A prominent feature of the rift valley is hydrothermal vents (Figure 3.12). These unique features are created when water seeps down in cracks in the crust, gains heat and dissolved substances and is released by through the seafloor. Hydrothermal vents can reach temperatures of over 350°C and contain energy-rich inorganic compounds such as hydrogen sulfides which can be used as a source of energy by specialized communities than inhabit the vents. Volcanic seamounts can also be associated with mid-ocean ridges, as magma can escape through the oceanic crust by side chambers (Figure 3.13).

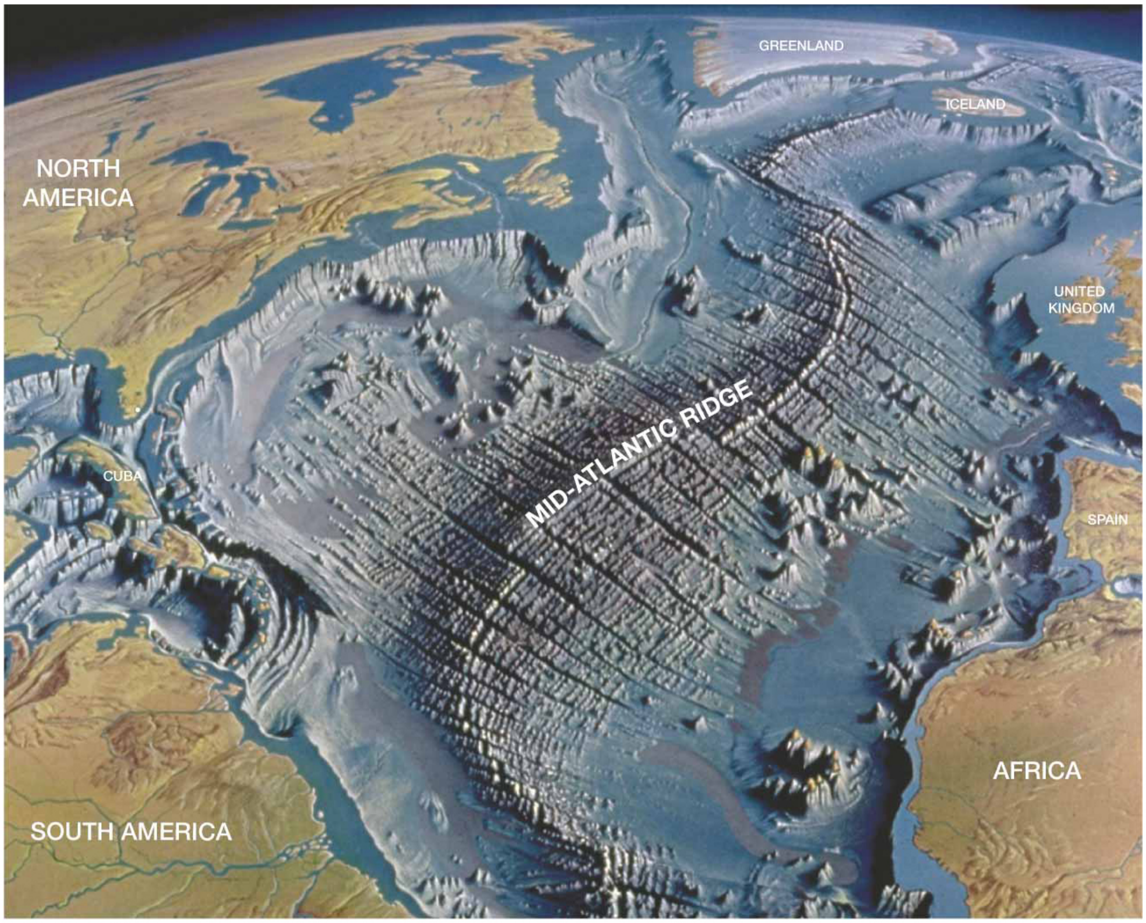


Figure 3.11. Bathymetric features of the North Atlantic, including the Mid-Atlantic Ridge clearly showing the central rift valley.

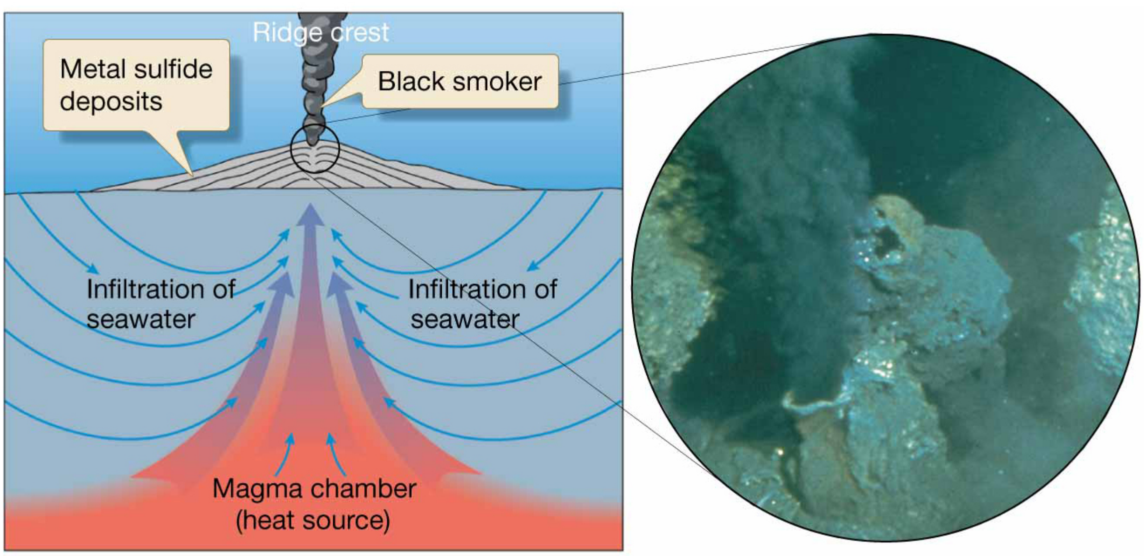
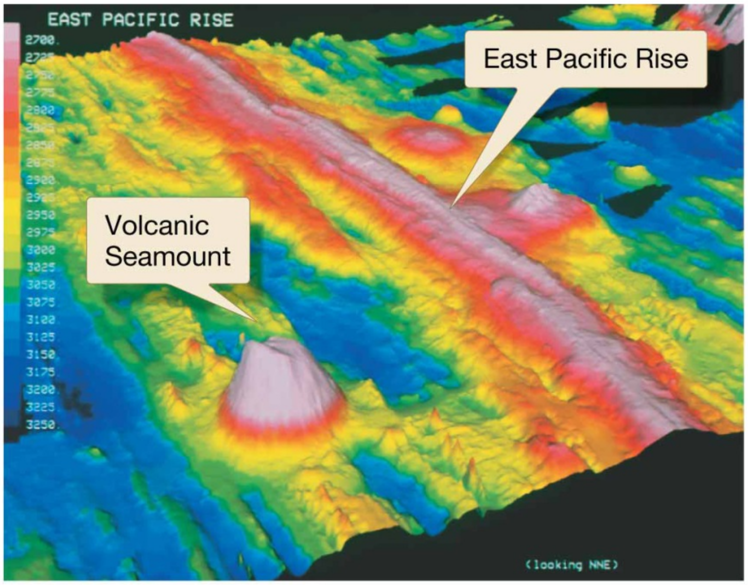


Figure 3.12. Structure and photo of a black smoker, a type of hydrothermal vent.



(a) False-color perspective view based on sonar mapping of a portion of the East Pacific Rise (*center*) showing volcanic seamount (*left*). The depth, in meters, is indicated by the color scale along the left margin; vertical exaggeration is six times.

Figure 3.13. A volcanic seamount associated with the spreading of the East Pacific Rise.

Mid-ocean ridges are cut by a number of fracture zones, parallel series of linear valleys perpendicular to the ridge (Figure 3.11). Transform faults are the region of the fracture zone where plates move in opposite direction (Figure 3.14). Earthquakes are frequent along transform faults.

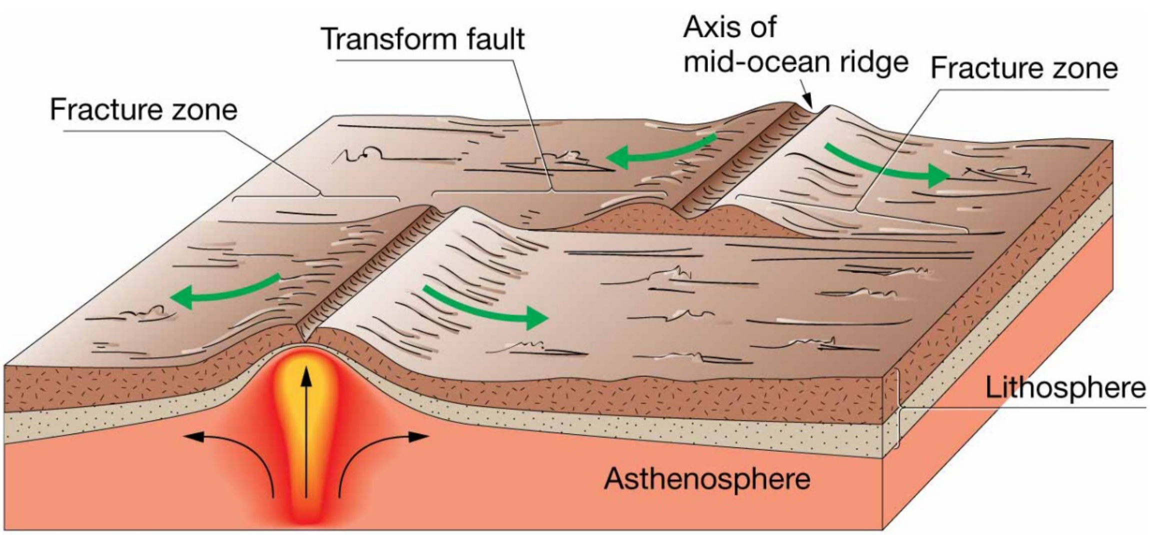


Figure 3.14. Fracture zones intersect mid-ocean ridges perpendicularly. In the transform fault area, the plates are moving in opposite direction.

The proportion of bathymetric and topographic features of the earth are shown in Figure 3.15. Note that abyssal plains (described as basins in Figure 3.15) comprise about 40% of the total area in the oceans.

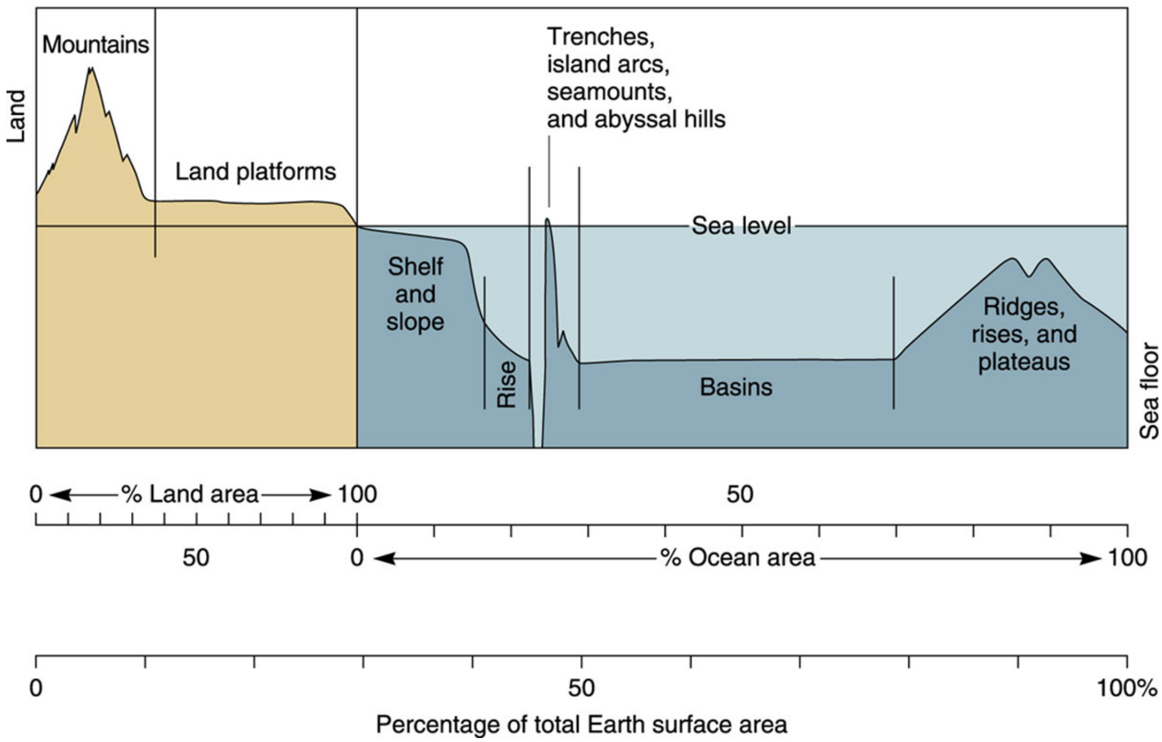


Figure 3.15. Topographic and bathymetric features of the earth, as a percent of total area of land or of oceans.

3.5. Review Questions

1. Which type of continental margin (passive or active) typically has a wide continental shelf?
2. What does bathymetry mean?
3. Which ocean has passive margins?
4. Which ocean has a lot of volcanoes and earthquakes along its margins?
5. What is the approximate average depth of a continental margin?
6. What is the name of the boundary between the continental shelf and the continental slope?
7. What are abyssal plains very flat?
8. Where are hydrothermal vents typically located?
9. Where are rift valleys located?
10. What is the difference between a transform fault and a fracture zone?
11. What size of sediments are typically found on the deep sea floor (fine or coarse)?
12. What is the continental rise?
13. What is a turbidity current?

4. Marine Sediments

4.1. Sediments and classification

Sediments are mineral and organic particles that have been removed by air, water and ice and accumulate on the sea floor. The margins of continents and the ocean floor constantly receive a supply of particles from various sources that can be classified by size and origin. They can originate from rocks that break down (lithogenous), biological organisms (biogenous), precipitation out of water (hydrogenous) and from outer space (cosmogenous) (Table 4.1). In this section we will examine each one of these types of sediment.

TABLE 4.1 CLASSIFICATION OF MARINE SEDIMENTS

Type	Composition		Sources		Main locations found
Lithogenous	Continental margin	Rock fragments Quartz sand Quartz silt Clay		Rivers; coastal erosion; landslides	Continental shelf
				Glaciers	Continental shelf in high latitudes
				Turbidity currents	Continental slope and rise; ocean basin margins
	Oceanic	Quartz silt Clay		Wind-blown dust; rivers	Abyssal plains and other regions of the deep-ocean basins
		Volcanic ash		Volcanic eruptions	
Biogenous	Calcium carbonate ($CaCO_3$)	Calcareous ooze (microscopic)	Warm surface waters	Coccolithophores (algae) Foraminifers (protozoans)	Low-latitude regions; sea floor above CCD; along mid-ocean ridges and the tops of volcanic peaks
		Shells and coral fragments (macroscopic)		Macroscopic shell-producing organisms	Continental shelf; beaches
				Coral reefs	Shallow low-latitude regions
	Silica ($SiO_2 \bullet nH_2O$)	Siliceous ooze	Cold surface waters	Diatoms (algae) Radiolarians (protozoans)	High-latitude regions; sea floor below CCD; upwelling areas where cold, deep water rises to the surface, especially that caused by surface current divergence near the equator
Hydrogenous	Manganese nodules (manganese, iron, copper, nickel, cobalt)		Precipitation of dissolved materials directly from seawater due to chemical reactions		Abyssal plain
	Phosphorite (phosphorous)				Continental shelf
	Oolites ($CaCO_3$)				Shallow shelf in low-latitude regions
	Metal sulfides (iron, nickel, copper, zinc, silver)				Hydrothermal vents at mid-ocean ridges
	Evaporites (gypsum, halite, other salts)				Shallow restricted basins where evaporation is high in low-latitude regions
Cosmogenous	Iron–nickel spherules Tektites (silica glass)		Space dust		In very small proportions mixed with all types of sediment and in all marine environments
	Iron–nickel meteorites		Meteors		Localized near meteor impact structures

Table 5.1 shows the names of sediment of various sizes. Sediment can be well sorted and homogeneous, or poorly sorted and heterogeneous. Typically, a rapid supply of sediment results in heterogeneous mixture, for example when a glacier melts and leaves sediment behind. Conversely, sediment deposited at the mouth of a continuously flowing river is more homogeneous. Water velocity in a given area is largely responsible for determining the size of sediment that is transported and deposited (Figure 4.1). Large particles require

a fast current to be transported, and are deposited quickly as current is reduced. Smaller particles (e.g. sand) are easily eroded and transported by slower currents, and take longer to be deposited. However, greater current velocities are required to erode clay than sand because it is more cohesive and stickier. Because water velocity determines the grain size that is deposited, the mean grain size of sediment in an area can serve as a rough measure of energy at the time of deposition.

Table 5.1 Particle Sizes and Settling Rates in Sediment

Type of Particle	Diameter	Settling Velocity	Time to Settle 4 km (2.5 mi)
Boulder	. 256 mm (10 in.)	—	—
Cobble	64–256 mm (. 2 1/2 in.)	—	—
Pebble	4–64 mm (1/6–2 1/2 in.)	—	—
Granule	2–4 mm (1/12–1/6 in.)	—	—
Sand	0.062–2 mm	2.5 cm/sec (1 in./sec)	1.8 days
Silt	0.004–0.062 mm	0.025 cm/sec (1/100 in./sec)	6 months
Clay	, 0.004 mm	0.00025 cm/sec	50 years[a]

[a]Though the theoretical settling time for individual clay particles is usually very long, under certain conditions clay particles in the ocean can interact chemically with seawater, clump together, and fall at a faster rate. Small biogenous particles are often compressed by organisms into fecal pellets that can fall more rapidly than would otherwise be possible. A fecal pellet is shown in Figure 5.11.

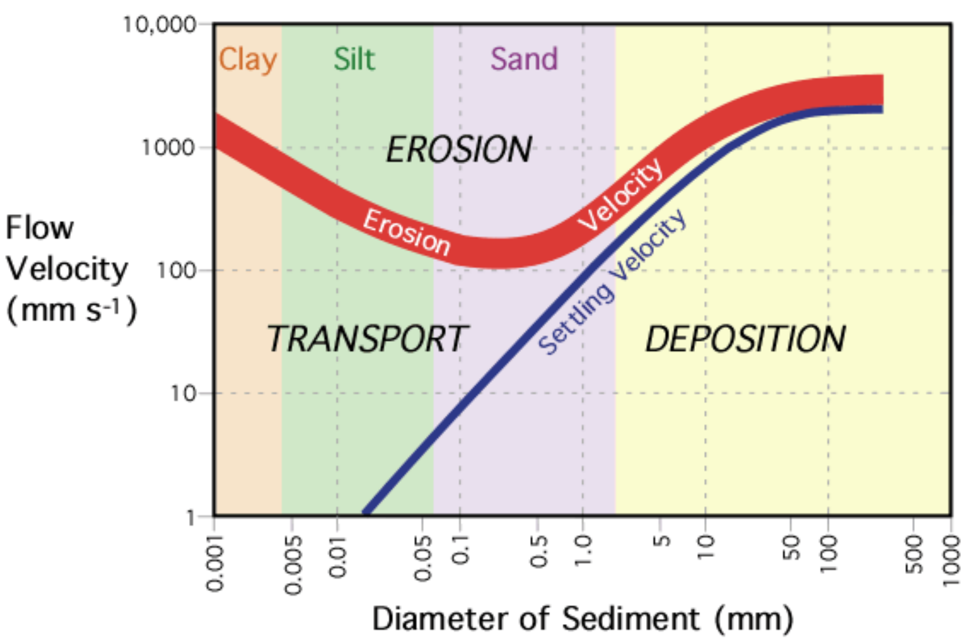


Figure 4.1. A Hjulstrom curve, showing velocity required to erode and deposit particles of various sizes.

4.2. Lithogenous sediment

Most abundant on continental margins are terrigenous or lithogenous sediments, sediments derived from land. This type of sediment is produced by the weathering and erosion of rocks on land (from wind, freezing of water, rivers and ice) and includes wind-blown dust and volcanic material. Though much of terrigenous sediment accumulates near continents, some of the very fine, clay-sized particles make it far offshore to the deep sea, transported by wind or current (Figure 4.2). Quartz is an important component of lithogenous sediments.

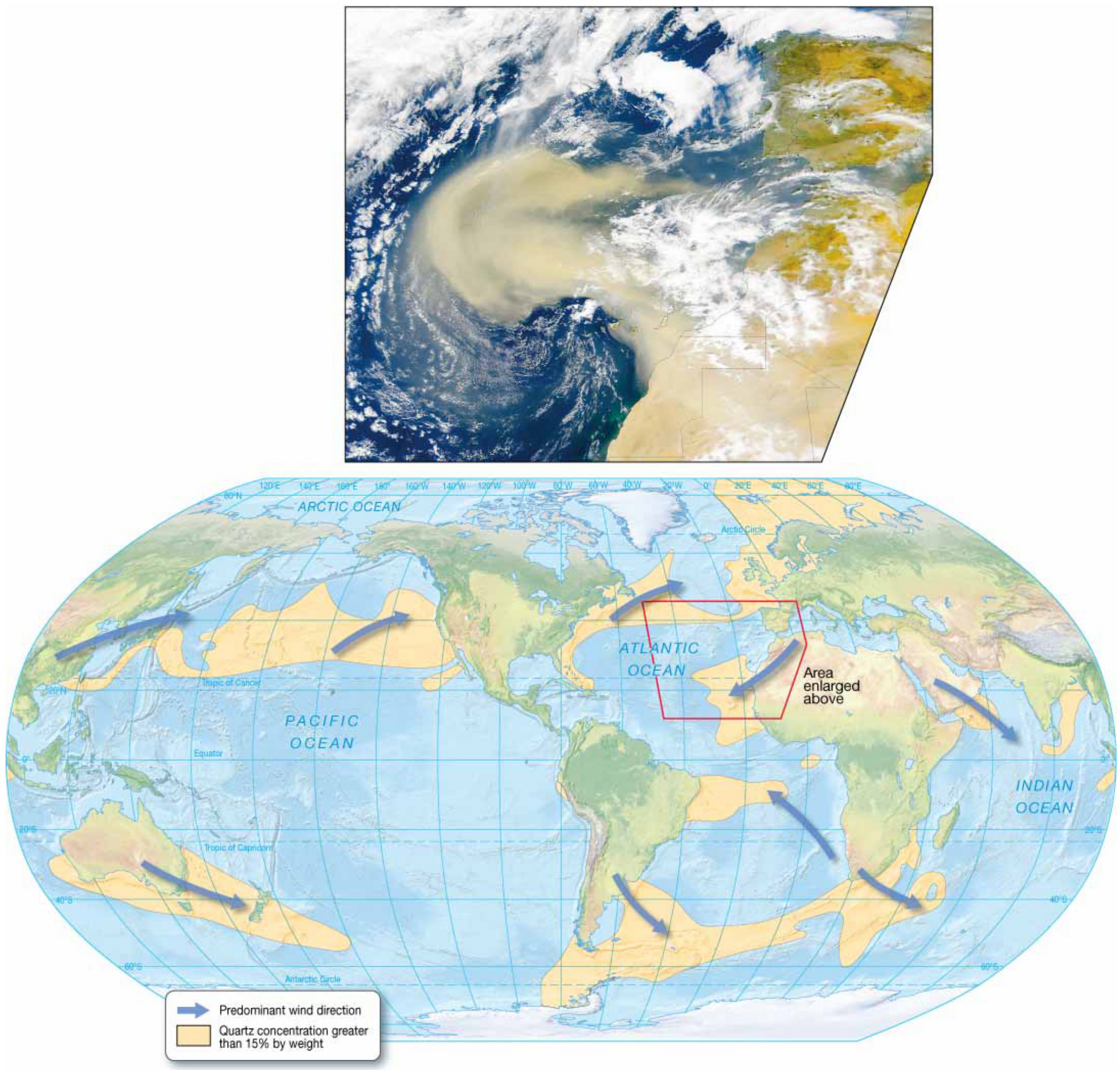
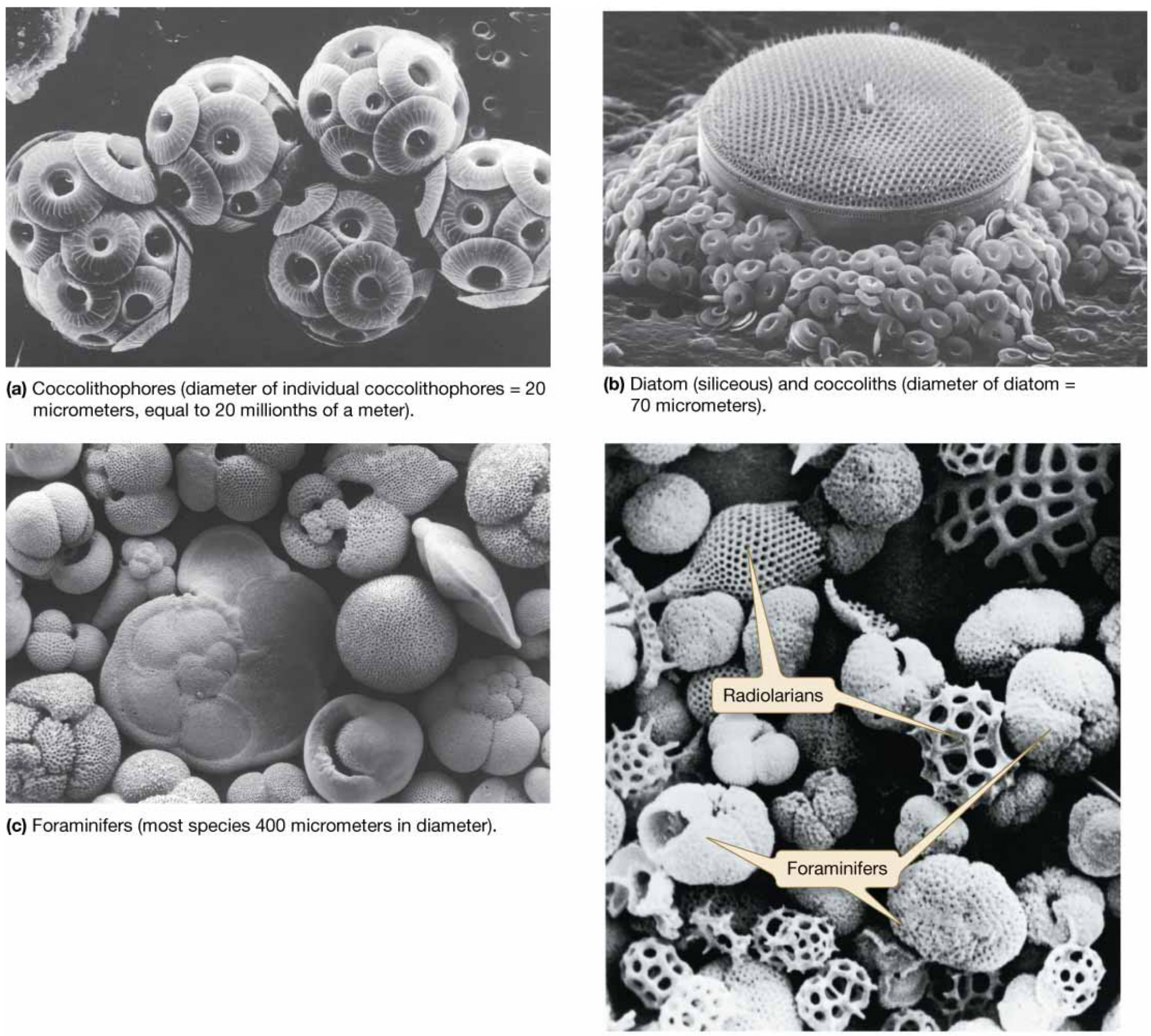


Figure 4.2. The highest concentration of lithogenous sediment in the deep sea follow dominant wind patterns. Wind can transport small sediment across large distances. Small sediments can also be carried a long way in current, since their settling rate is slow. Inset shows a dust storm blowing sediment from Africa into the Atlantic.

4.3 Biogenous sediments

Biogenous sediments are those derived from hard parts of organisms that do not degrade after the organism dies. They originate from single-celled organisms and from shell fragments of larger organisms. Deep-sea sediments that are composed by over 30% biogenic sediment are called oozes, and typically fall in two categories: calcareous ooze which comes primarily from foraminiferans, pteropods and coccolithophores (Figure 4.3), and siliceous ooze, which comes from diatoms and radiolarians (Figure 4.4). More details on these planktonic organisms are provided in chapter 15 of this booklet.



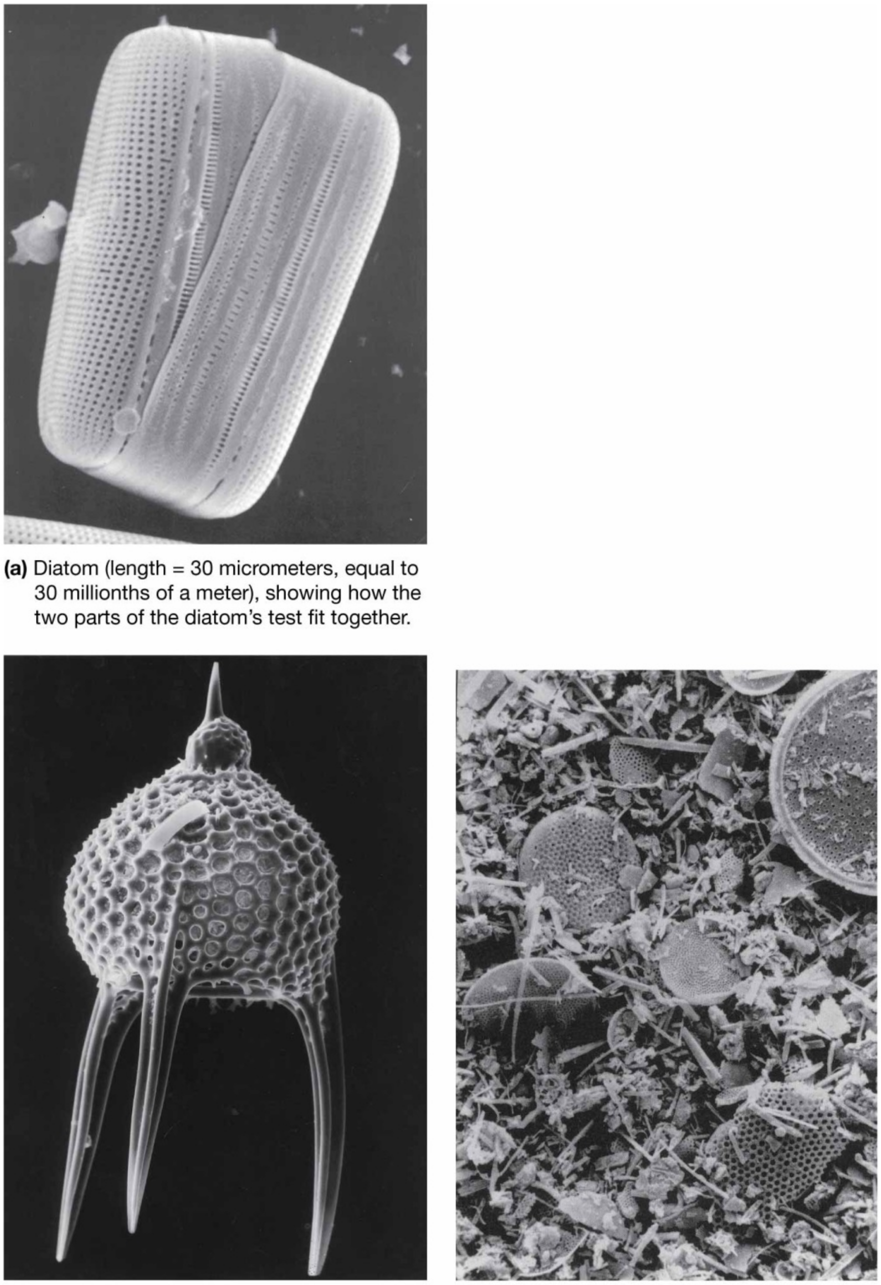
(a) Coccolithophores (diameter of individual coccolithophores = 20 micrometers, equal to 20 millionths of a meter).

(b) Diatom (siliceous) and coccoliths (diameter of diatom = 70 micrometers).

(c) Foraminifers (most species 400 micrometers in diameter).

(d) Calcareous ooze, which also includes some siliceous radiolarian tests (magnified 160 times).

Figure 4.3. Coccolithophores and foraminiferans are some of the most important organisms that contribute calcium carbonate shells to calcareous oozes.



(a) Diatom (length = 30 micrometers, equal to 30 millionths of a meter), showing how the two parts of the diatom's test fit together.

(b) Radiolarian (length = 100 micrometers).

(c) Siliceous ooze, showing mostly fragments of diatom tests (magnified 250 times).

Figure 4.4. Radiolarians and diatoms both produce silica shells that accumulate in siliceous oozes

Biogenic oozes form where a large biomass of shell-producing microorganisms occurs—in areas of high productivity. These organisms are most abundant in the surface layers of the ocean, and their tests (shells) sink to the bottom once they die. If the rate of accumulation is greater than the rate of dissolution, oozes form (Figure 4.5).

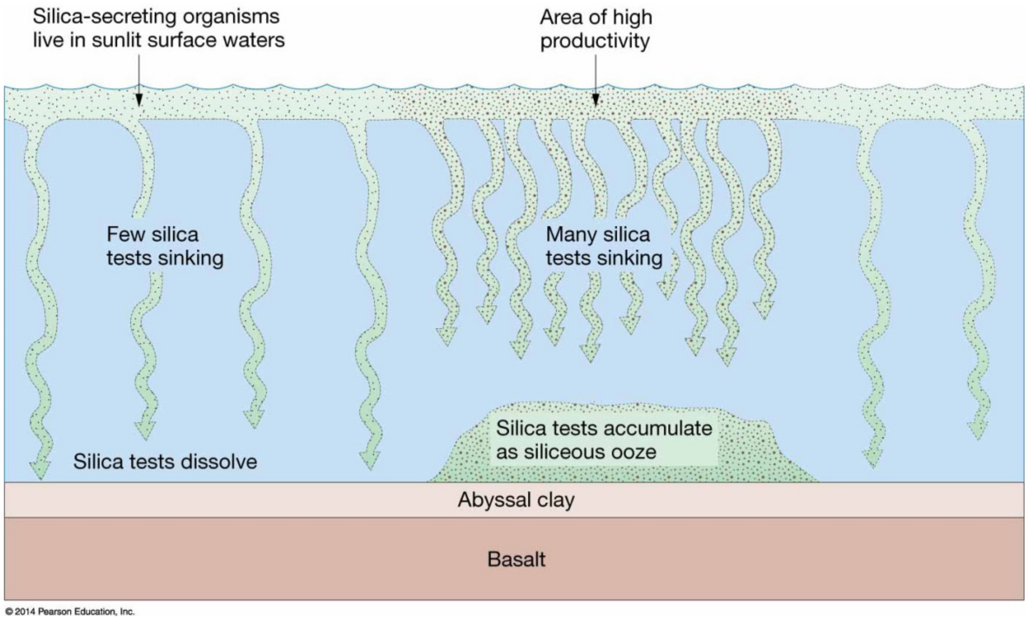


Figure 4.5. Accumulation of siliceous ooze below zone of high productivity, where siliceous material accumulates faster than it dissolves.

While siliceous oozes can be found at all depths, calcareous oozes are not found in the deepest waters, since calcium carbonate dissolves in the conditions found in the deep sea (Figure 4.6). The depth at which calcium carbonate dissolves is called the calcite compensation depth. It varies slightly depending on local conditions, but is generally around 4500m. Calcareous oozes are not found below this depth, except if they are rapidly covered by other sediments.

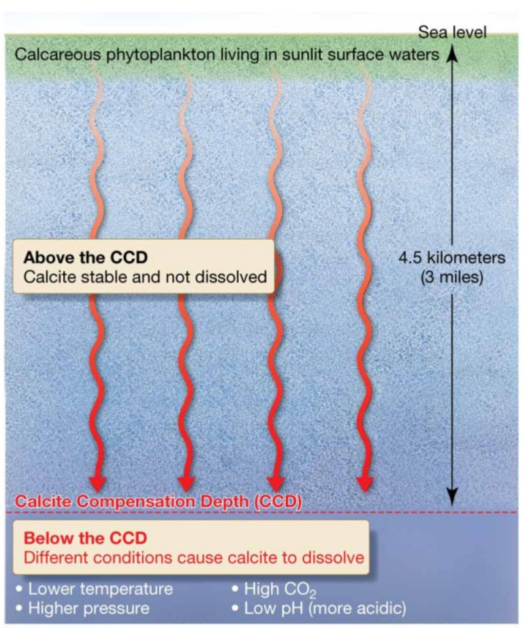


Figure 4.6. The calcite compensation depth is the point at which calcite dissolves in seawater.

Since mid-ocean ridges often rise aboove the calcite compensation depth, calcareous oozes are commonly found near them (Figure 4.7)

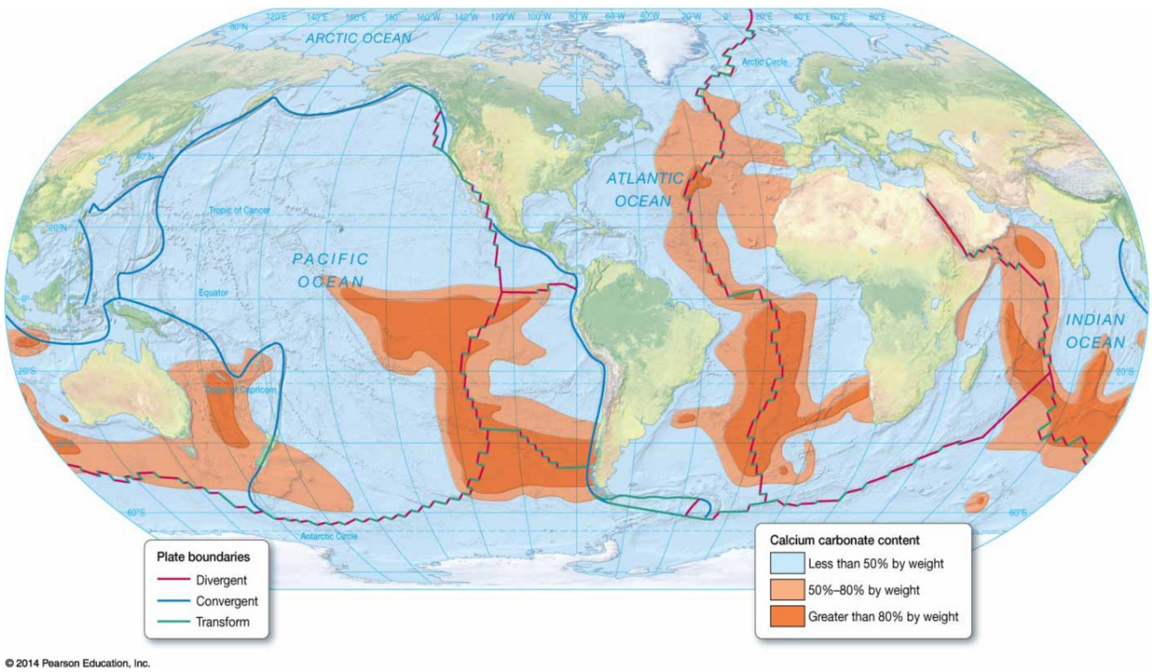


Figure 4.7. Since calcareous oozes are only found in shallower waters, the distribution of calcareous oozes follows mid-ocean ridges.

4.4 Hydrogenous & cosmogenous sediments

Hydrogenous sediments are produced in the water when minerals precipitate by chemical reactions. Most hydrogenous sediments are formed by precipitation of minerals on the sea floor, but some are formed in the water column or at hydrothermal vents. Hydrogenous sediments include salts, carbonates, phosphorites and manganese nodules.
Cosmogenous (or extraterrestrial) sediments originate from outer space, and are typically made of small, iron-rich, magnetic spheres.

4.5 Distribution of pelagic and neritic deposits

Sediment deposits found in the ocean are a mixture of particles from various origins, but one type of sediment often dominates in a given area and generalizations on sediment type can be made based on latitude and bathymetric features (Figure 4.8 & 4.9). For example, coarse-grained lithogenous sediments dominant near continents, while fine-grained lithogenous material (abyssal clay) and biogenic sediments are important in the middle of the oceans. Hydrogenous and cosmogenous sediment are much less abundant, except for isolated fields of manganese nodules in some regions of the abyssal plains (Figure 4.8).

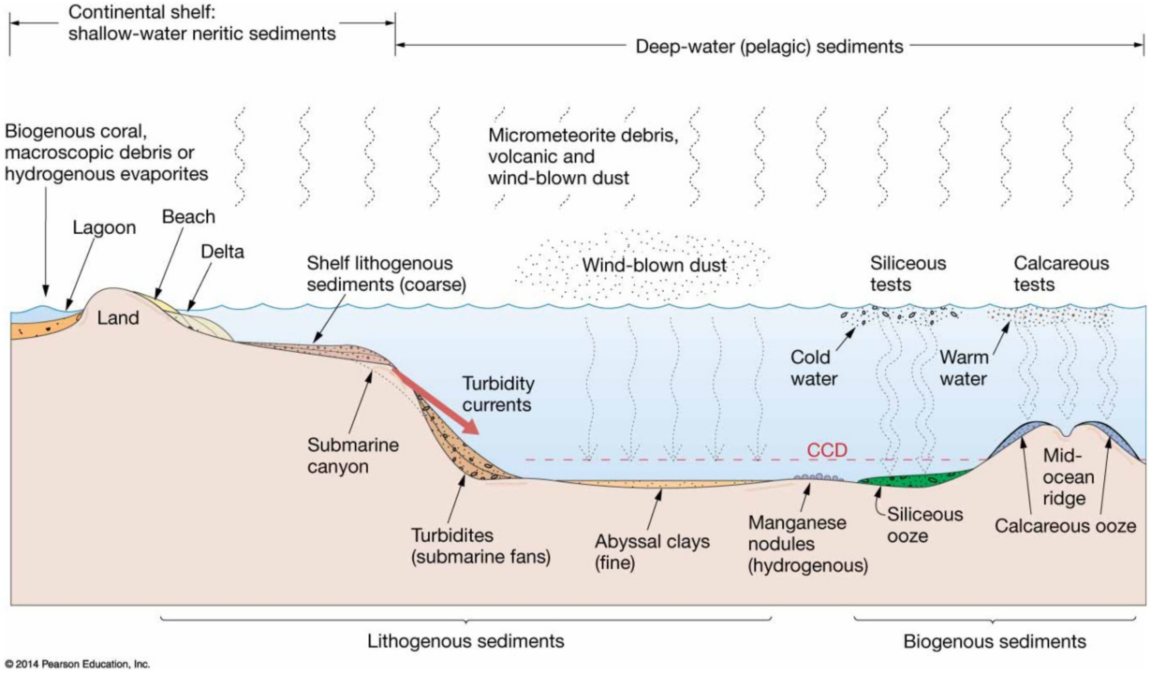


Figure 4.8. Typical distribution of sediments from a passive continental margin to a mid-ocean ridge

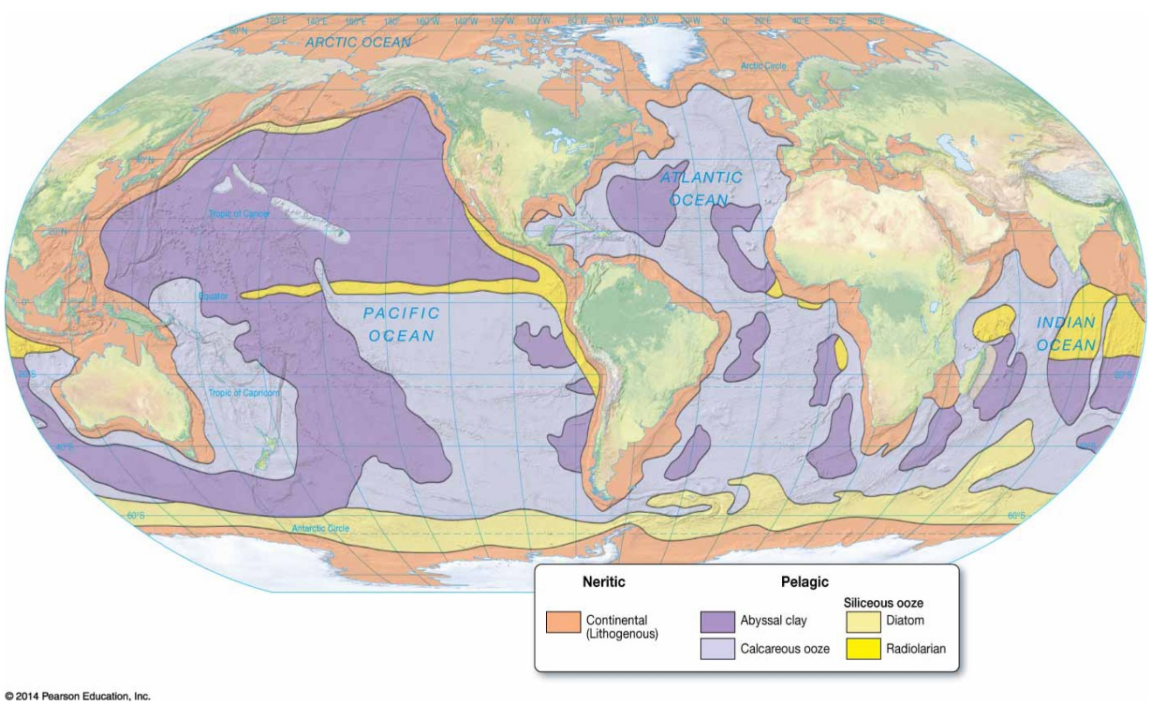


Figure 4.9. Distribution of marine sediments.

A quick glance at sinking rates of various particle sizes (table 5.1 above) shows that the smallest sediment particles can remain in suspension for months or years before reaching the seafloor. However it has become clear that particles tend to attract one another, producing larger particles that sink faster. Additionally, nearly all biogenic sediment particles are aggregated into fecal pellets as the organisms producing the tests are eaten by small animals. The resulting pellets, while still small, sink much faster and can reach the ocean floor within a few days. The type of ooze found in a given area therefore reflects the types of microorganisms that live above.

Marine sediments are thickest along passive continental shelves where large amounts of sediment is carried by rivers. Less sediment is found on active margins, where trenches and subduction zones are located close to shore. Thin sediment layers are found on mid-ocean ridges, with sediment thickness gradually increasing away from the ridge, as age of the ocean floor increases (Figure 4.10).

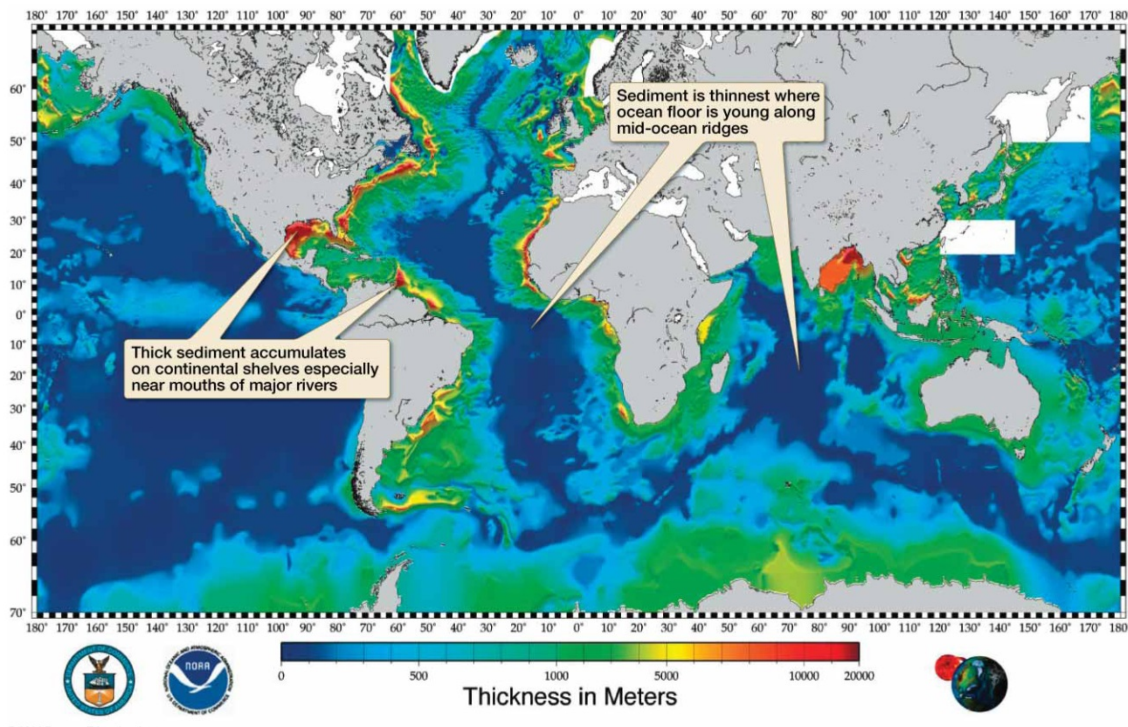


Figure 4.10. Thickness of marine sediment throughout the oceans.

4.6. Review Questions

1. What is the name for sediments that originated from rocks on land?
2. What are hydrogenous (or authigenic) sediments?
3. What is an ooze?
4. What are the two main types of oozes, and which organisms contribute to each?
5. Which kind of ooze is found deepest?
6. What kind of sediment is a manganese nodule?
7. Which areas of the seafloor have the thickest sediment accumulations?
8. How do clay-sized biogenic sediment reach the seafloor roughly in the same area where they were formed near the surface, if clay-sized sediment should take 50 years to settle in 4km?
9. Compare the velocities needed to deposit sand and silt
10. Where are lithogenous sediments most abundant?
11. Compare the velocities require to erode silt and clay
12. Why are calcareous oozes commonly found along mid-ocean ridges and not on abyssal plains?
13. What kind of sediment is African dust that blows to the Atlantic during storms?

5. Water (Trujillo, Chapter 5)

Water is a very common substance on earth; 71% of our planet is covered by water. Though it is very abundant, few people understand the special characteristics of water which make it so important to life on earth. This section examines various properties of water which are important to understand before one can understand the dynamics of the oceans and organisms living in it.

5.1 The Water Molecule

The water molecule is made up of two hydrogen atoms linked with one oxygen atom, and its chemical formula is H_2O (Figure 5.1). The angle between the two hydrogen atoms in liquid water is 105^o. The hydrogen atom possesses one electron in one ring, whereas the oxygen atom possesses 8 electron spread over rings; 2 electrons in the innermost level, and 6 in the outermost level. The hydrogen's ring would be full and most stable with 2 electrons, and the oxygen's outermost ring would be full and stable with 8 electrons. In the water molecule, the hydrogen atoms and the oxygen atom share electrons, thereby filling their outermost energy level to its full capacity, making it a very stable molecule. This type of bond is called a covalent bond, and it is very strong. Within this covalent bond, electrons are not shared equally, and spend more time towards the oxygen atom, making the water molecule a polar molecule: the oxygen side is negatively charged, and the hydrogen sides are positively charged. Water is a great solvent because of the polarity of its molecules.

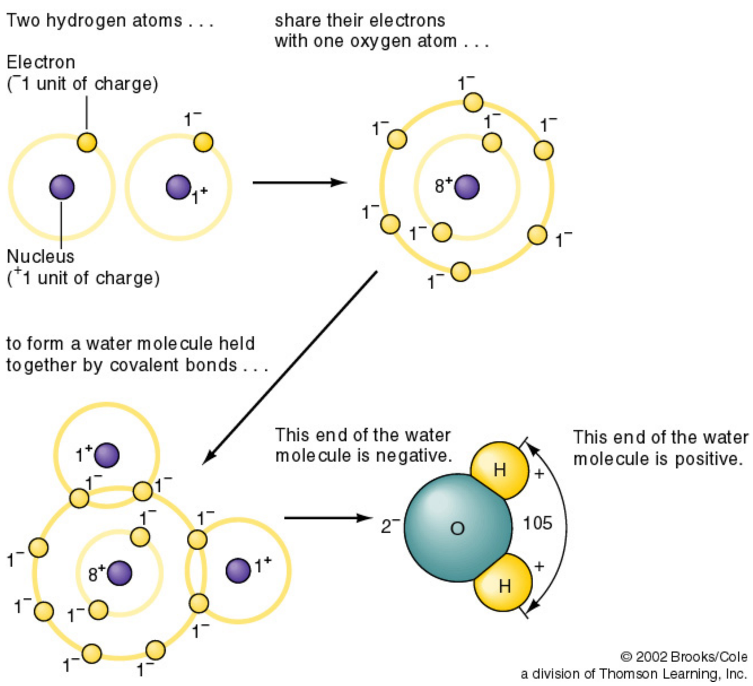


Figure 5.1. The water molecule

Covalent bonds are the strong bonds within the water molecule, linking the oxygen atom with the hydrogen atom. Water molecules also bond with one another through hydrogen bonds (Figure 5.2). Hydrogen bonds are formed between water molecules because of the polarity of the molecule. The negatively-charged oxygen from one molecule is attracted to the positively-charged hydrogen from another molecule. Each molecule can form hydrogen bonds with up to four other molecules. Hydrogen bonds are weak compared to covalent bonds, and are constantly forming and breaking as molecules move around in water. However, these bonds are relatively strong compared to similar bonds in other molecules. This relatively strong bond between water molecules is responsible for many characteristics of water, such as surface tension and high heat capacity.

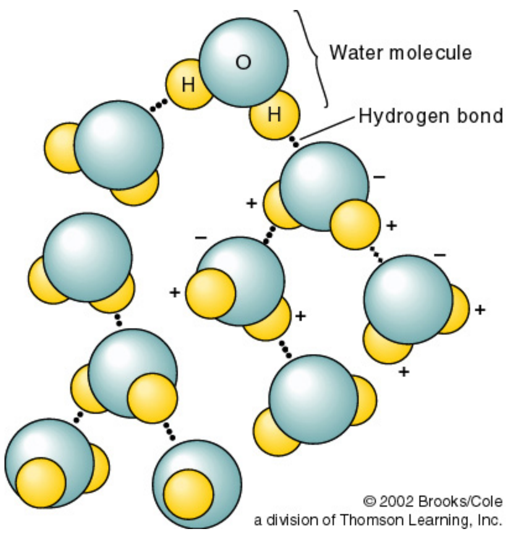


Figure 5.2. Hydrogen bonds between water molecules

The polarity of water molecule not only means that molecules are attracted to one another; they are also attracted to other polar compounds. In binding with these compounds, water can greatly reduce the attraction between ions of opposite charge. For example, the sodium and chloride ions in table salt are 80 times less attracted to one another when placed in water, as water molecules surround the ions in a process called hydration (Figure 5.3). Given enough time, water is able to dissolve nearly every substance in this fashion, and for this reason it is often called the university solvent.

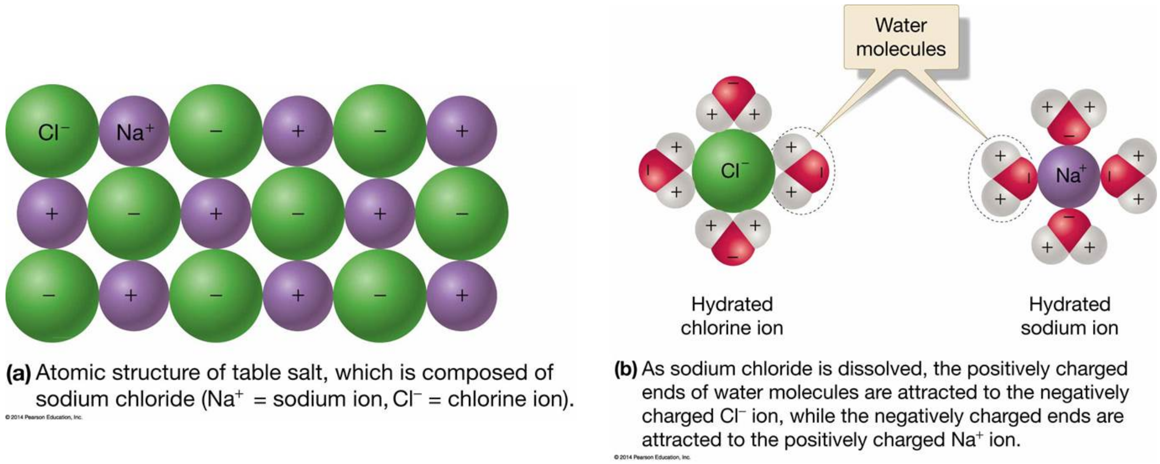


Figure 5.3. Water is a great solvent because of it polar nature. The negative side of the water molecule is attraction to positive ions (here, Na^+) while the positive side is attracted to negative ions (here, Cl^-).

5.2 Other properties of water

Thermal properties

As energy (heat) is added to water, the movement of molecules increases and molecules move further apart. It also takes heat to change the state of water from solid to liquid and gas (Figure 5.4). Conversely, changing states in the other direction (from gas to liquid to solid) releases energy to the environment. In the solid state, molecules remain firmly attached in a rigid structure. In the liquid state, water molecules have more kinetic energy (movement) and hydrogen bonds are broken and reformed at a much greater rate. In a gas state, molecules are much further apart with few hydrogen bonds at any given time.

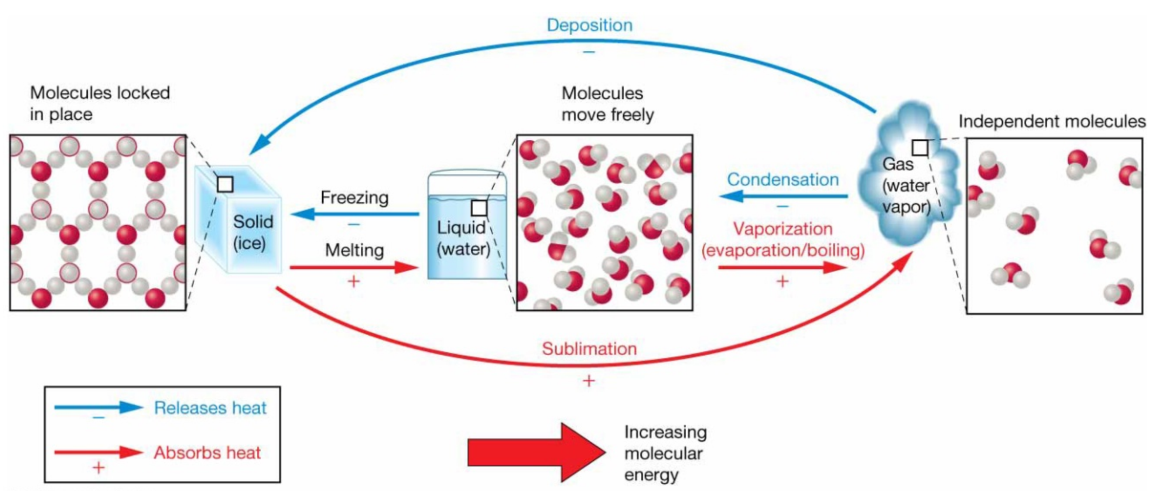


Figure 5.4. Water in each of the three states, showing the change in molecule arrangement and transfer of heat between water and the environment.

Within a state, the heat added to water directly increases the temperature of the water, e.g. it takes 1 calorie to raise 1 gram of water by 1ºC. This is called the heat capacity and it is high compared to other substances (Figure 5.5). The reason for this high heat capacity again resides in the geometry and polarity of water molecules, and the resulting hydrogen bonds which take more energy to break than the forces between molecules in other compounds (typically dominated by weaker interactions called van der Waals).

Pure freshwater melts (or freezes) at 0ºC and boils (or condenses) at 100ºC. These temperatures are very high; based on similar compounds one would expect water to melt at -90ºC and boil at -68ºC (Figure 5.6). The reason for this discrepancy lies in the geometry of the water and its polarity, which lead to strong hydrogen bonds that require a lot of energy to break.

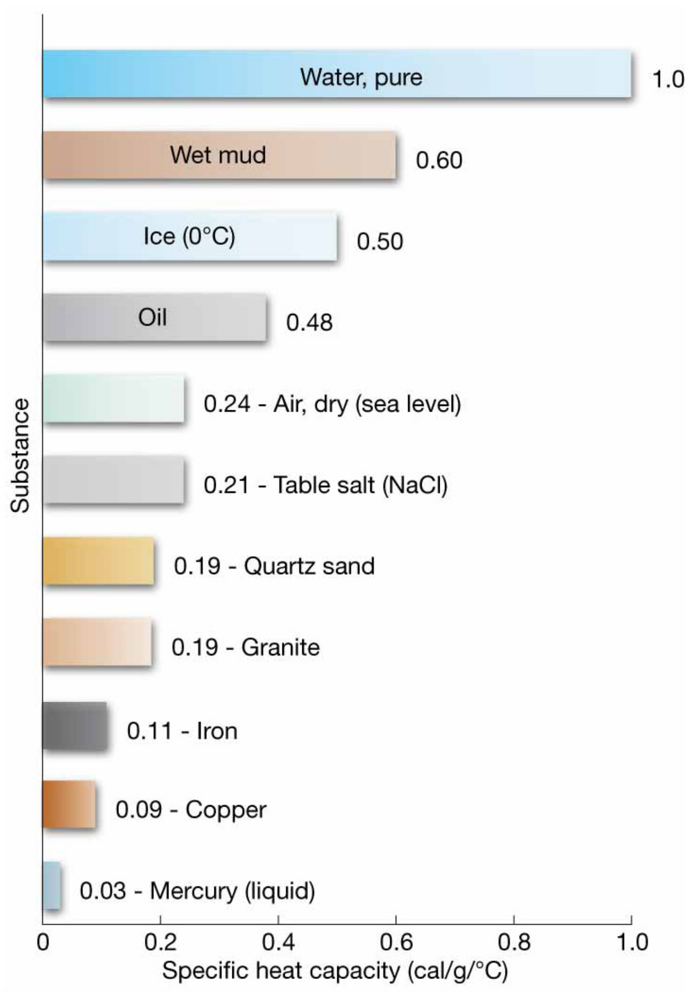


Figure 5.5. Specific heat capacity of water compared to other common substances. It takes more energy to increase the temperature of water than these other substances.

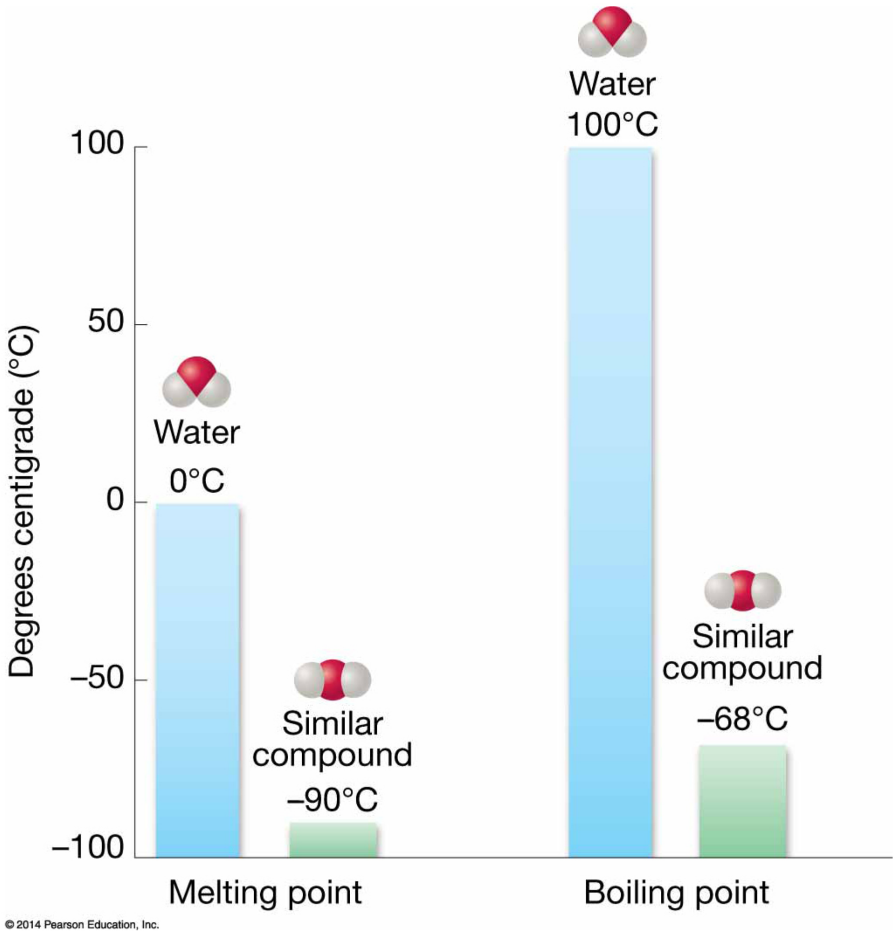


Figure 5.6. Comparison of melting and boiling point of water compared to other similar compounds.

As water changes state, a lot heat is absorbed or released from it. Going from solid to liquid and gaseous states takes a lot of energy to break hydrogen bonds and this energy does not increase the temperature of the water. Changing from ice to liquid water at 0ºC requires 80 calories that do not lead to a change in temperature. This is called the latent heat of melting. Similarly, changing from liquid water to gas (water vapor) requires 540 calories that do not change the temperature. This is called the latent heat of vaporization (Figure 5.7). The reason for a greater latent heat of vaporization compared to melting is that many more hydrogen bonds need to be broken to change water from liquid to gas (Figure 5.8).

While it is clear that water evaporates when it reaches 100ºC, we also know that water evaporates at the surface of the ocean at much lower temperature (e.g. 20ºC). In this case, the water molecules that evaporate obtain the energy from surrounding water molecules in the process of evaporation, and therefore this has a cooling effect on the remaining liquid water. One gram of water at 20ºC requires 585 calories to be evaporate, more energy than would be needed at 100ºC.

When water changes states from gas to liquid to solid, heat is instead released to the environment. This has important consequences for climate and air-sea interactions, for example in fueling tropical storms (more on this in chapter 6).

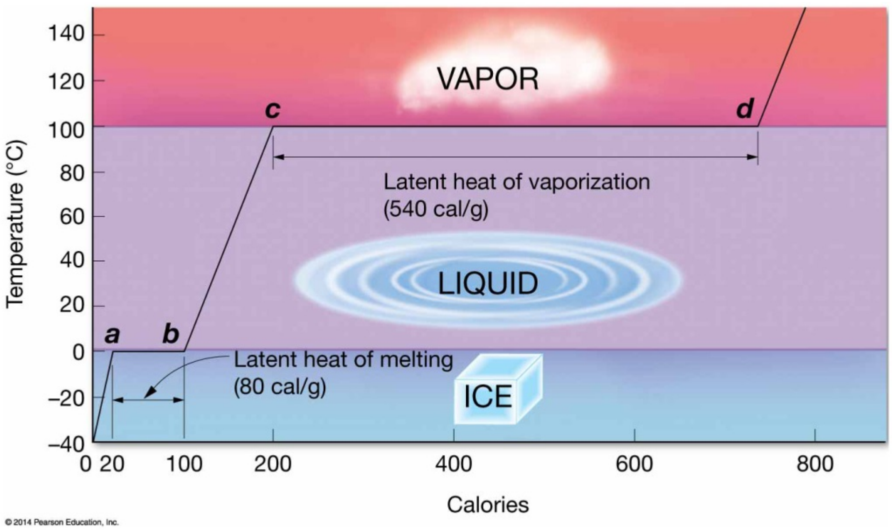


Figure 5.7. Latent heats and changes of state in water.

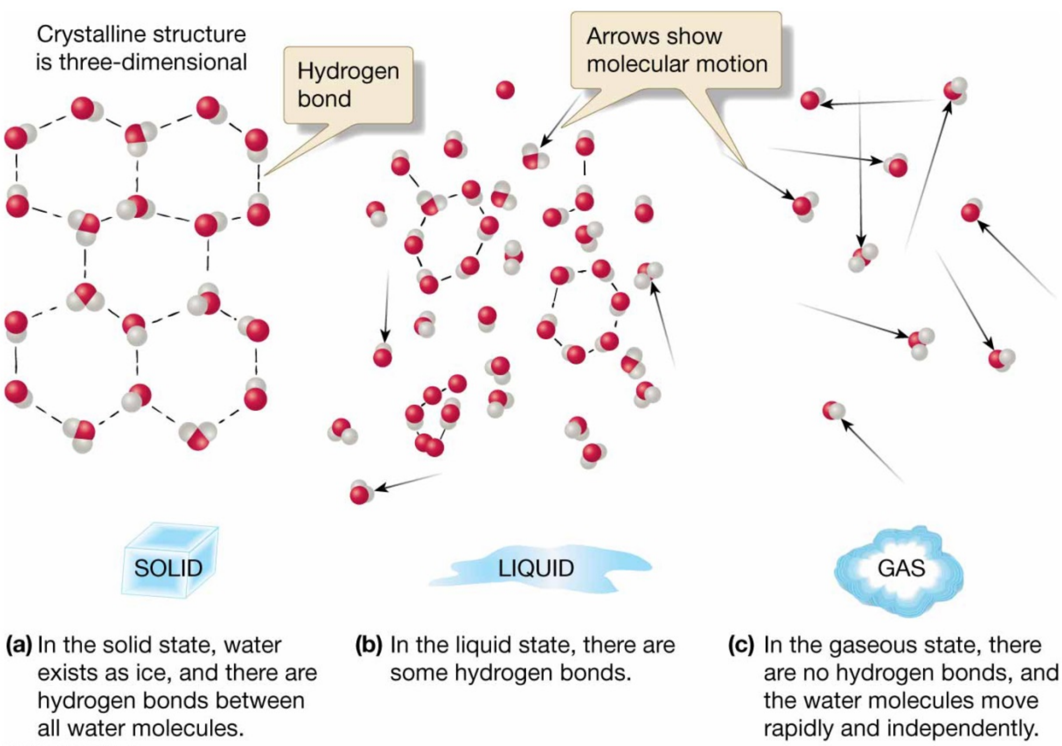


Figure 5.8. Water molecules in three states

Global thermostatic effects

Water's high heat capacity and latent heats both contribute to water's thermostatic effect: its ability to moderate temperature fluctuations. This is evident in two main ways. First, the evaporation/condensation cycle of water removes large amounts of heat from the tropics to redistribute it at higher latitude, thus moderating the Earth's climate (Figure 5.9). Second, the higher heat capacity of water than land helps reduce the temperature difference between day and night (Figure 5.10) and between summer and winter. Thus, areas on the coast have much milder climates than dry zones within continents.

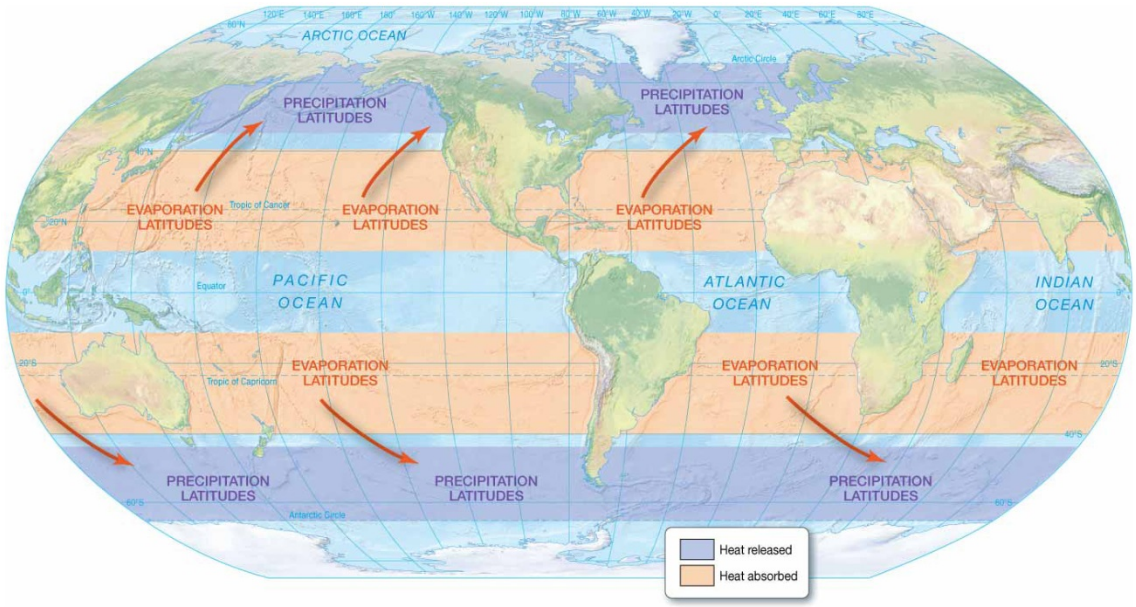


Figure 5.9. The cycle of evaporation and precipitation helps move surplus heat from the equator to higher latitudes. More detail on atmospheric circulation patterns are provided in chapter 9 of this booklet.

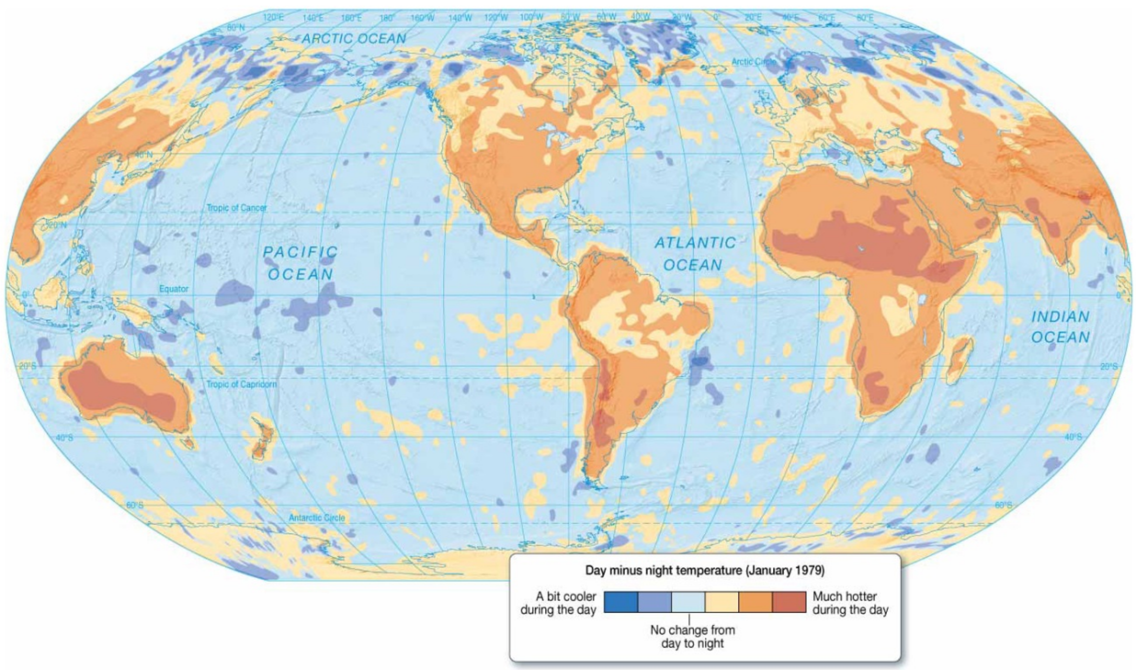


Figure 5.10. Temperature differences between day and night are much greater in dry areas than they are in the water and in coastal areas.

Surface tension
Surface tension refers to how difficult it is to break the surface of a liquid. Water has the highest surface tension of naturally occurring substances after mercury, because of the hydrogen bonds that bond molecules at the surface with those laterally and below. This high surface tension is important in the formation of wind-driven capillary waves as well as in the capillary effect. Increasing the temperature and salinity of water decreases surface tension.

Viscosity
Viscosity is the property of a fluid to resist flow. Water has a relatively low viscosity. Increased temperature decreases viscosity, but this is noticeable only to the smallest organisms, which must develop adaptations to prevent sinking in the less viscous, warm tropical waters. More on this in chapter 15.

Pressure
Water is much denser than air and for that reason pressure increases quickly with depth. For each 10 m (33 ft) of depth, the pressure increases by 1 atmosphere. Marine animals therefore have adaptations to deal with the pressure of the environment they live in. Water is nearly incompressible and therefore even the greatest pressures have only a very small effect on volume.

5.3 Water density

Density is the measure of mass per unit volume, and is commonly measured in grams per centimeter cubed (g/cm^3). Pure water has a density of $1 g/cm^3$ (at 4^oC) but density of water changes with its temperature, salinity and pressure. In liquid water, density increases as temperature drops from the boiling point, because water molecules lose kinetic energy and are increasingly close together. But this density increase is only true until water reaches 4^oC (Figure 5.11). At this temperature, the water molecules are the closest they will be, and start forming ice crystal, in which each molecule forms hydrogen bonds with four other molecules. To form these ice crystals, the angle between the hydrogen atoms within the molecule increases from 105^o to 109^o. This increase in the angle within the molecule decreases the density from 4^oC to 0^oC. Therefore, water is at its highest density at 4^oC, and ice floats on water.

Increased salinity and increased pressure also lead to higher water density. However, temperature has a greater influence on density than salinity and pressure. Adding salt to water also decreases its freezing point. For that reason, seawater typically freezes around -2^oC, rather than 0^oC for freshwater.

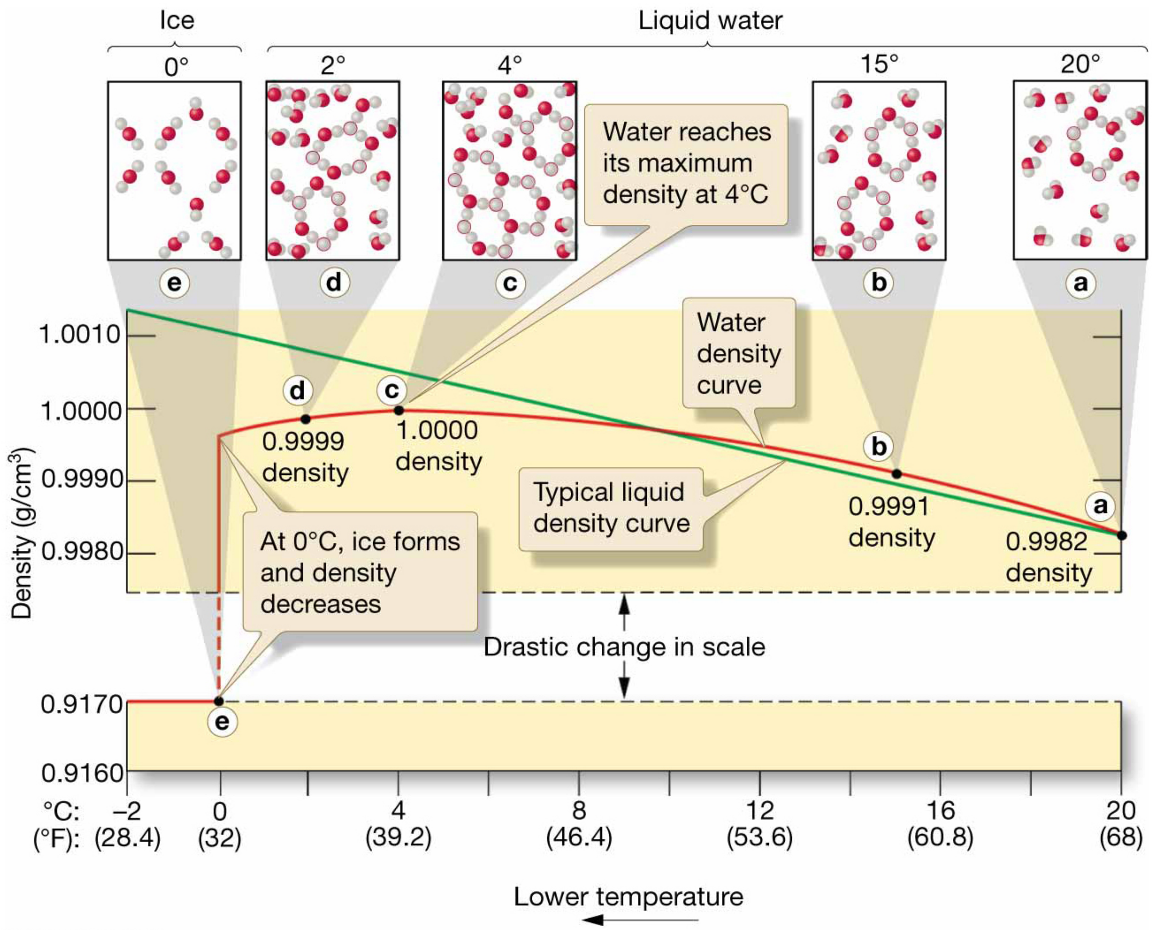


Figure 5.11. Density of pure water with temperature.

5.4. Review Questions

1. What type of bond links the oxygen to the hydrogen in the water molecule?

2. What is a hydrogen bond?

3. What characteristic of the water molecule makes hydrogen bonds relatively strong?

4. How many water molecules can one molecule form hydrogen bonds with?

5. What is the definition and value of the latent heat of vaporization?

6. How many calories are required to evaporate one gram of water at 75°C?

7. At what temperature is water densest?

8. What happens to the angle of covalent bonds within water molecules in ice?

9. What is heat capacity and why does water have high heat capacity compared to other substances?

10. What are the two main ways in which water reduces temperature fluctuations on Earth?

11. Give an example of the role of surface tension in the marine environment

12. What is the property of a fluid to resist flow?

6. Seawater (Trujillo Chapter 5)

6.1. Properties of Seawater

Salinity

Salinity refers to the amount of inorganic material dissolved in water. This excludes sediments held in suspension since those particles are not dissolved. The polar nature of the water molecule allows it to readily dissolve salts. As salts (e.g. NaCl) are added to water, they dissociate into ions (e.g. Na^+ and Cl^-), and bond with water molecules (Figure 5.3). Water can hold a certain quantity of salt in solution; this is called the saturation value. An increase in temperature increases the saturation value of salts.

On average the ocean has a salinity of 35 ‰ (or ppt, parts per thousand), which means that 1000 g of seawater is composed of 965g of water and 35 g of dissolved solids (Figure 6.1). The most abundant salts in water are referred to as major ions. Note that the six most abundant ions make up 99% of all salts in seawater (Figure 5.12). Minor constituents include more ions, as well as some gases and nutrients. Trace elements are present in concentrations lower than 1 ppm, and include aluminum, copper, cobalt, iron, mercury and silver, among others (Table 5.1). Though trace elements are present only in very small quantities, they may still play an important role biologically (e.g. iron). Nutrients are inorganic substances necessary for plant growth; nutrients and gases are discussed in chapter 7 of this textbook.

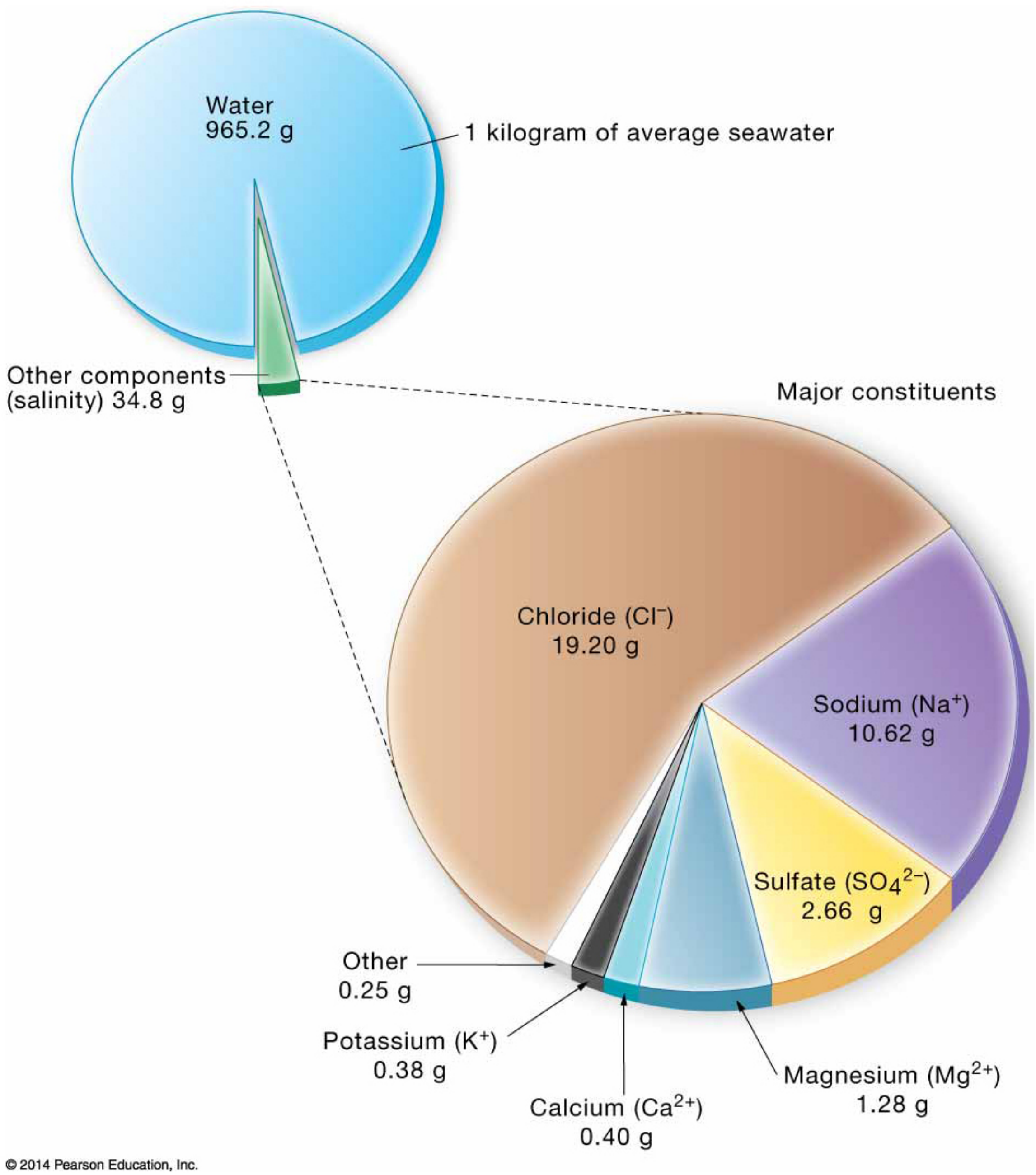


Figure 6.1. Major ions in seawater.

TABLE 5.1 SELECTED DISSOLVED MATERIALS IN 35‰ SEAWATER

1. Major constituents (in parts per thousand by weight, ‰)		
Constituent	**Concentration (‰)**	**Ratio of constituent/total salts (%)**
Chloride (Cl^-)	19.2	55.04
Sodium (Na^+)	10.6	30.61
Sulfate (SO_4^{2-})	2.7	7.68
Magnesium (Mg^{2+})	1.3	3.69
Calcium (Ca^{2+})	0.40	1.16
Potassium (K^+)	0.38	1.10
Total	**34.58‰**	**99.28%**

2. Minor constituents (in parts per million by weight, ppm[a])					
Gases		Nutrients		Others	
Constituent	**Concentration (ppm)**	**Constituent**	**Concentration (ppm)**	**Constituent**	**Concentration (ppm)**
Carbon dioxide (CO_2)	90	Silicon (Si)	3.0	Bromide (Br^-)	65.0
Nitrogen (N_2)	14	Nitrogen (N)	0.5	Carbon (C)	28.0
Oxygen (O_2)	6	Phosphorus (P)	0.07	Strontium (Sr)	8.0
		Iron (Fe)	0.002	Boron (B)	4.6

3. Trace constituents (in parts per billion by weight, ppb[b])					
Constituent	**Concentration (ppb)**	**Constituent**	**Concentration (ppb)**	**Constituent**	**Concentration (ppb)**
Lithium (Li)	185	Zinc (Zn)	10	Lead (Pb)	0.03
Rubidium (Rb)	120	Aluminum (Al)	2	Mercury (Hg)	0.03
Iodine (I)	60	Manganese (Mn)	2	Gold (Au)	0.005

[a]Note that 1000 ppm = 1‰.
[b]Note that 1000 ppb = 1 ppm.

Determining salinity

The principle of constant proportions states that the ratio of one major ion to another remains the same, regardless of variations in salinity. This applies to major conservative ions in open-ocean, not where rivers bring dissolved substances or reduce salinities. The ratios of minor non-conservative constituents (e.g. nutrients and gases) do not follow the principle of constant composition as they vary because they are connected to life cycles of organisms.

The salinity of water can be measured in a variety of ways. It can be measured with a salinometer, which measures the conductivity of water (the saltier the water, the more it conducts electricity). CTDs (Conductivity-Temperature Devices) are modern instruments that also measure salinity through conductivity. It is also possible to titrate the chlorine in the water, which is directly proportional to the total salinity because of the principle of constant proportions. A refractometer measures the bending of light as it passes from air to water; the saltier the water, the more dense it is, and the more it refracts light.

Pure water vs seawater

Seawater has slightly different properties than pure water: it freezes at a lower temperature, boils at a higher temperature, has higher density and higher pH (Table 5.2).

TABLE 5.2 COMPARISON OF SELECTED PROPERTIES OF PURE WATER AND SEAWATER

Property		Pure water	35‰ seawater
Color (light transmission)	Small quantities of water	Clear (high transparency)	Same as for pure water
	Large quantities of water	Blue-green because water molecules scatter blue and green wavelengths best	Same as for pure water
Odor		Odorless	Distinctly marine
Taste		Tasteless	Distinctly salty
pH		7.0 (neutral)	Surface waters, range = 8.0–8.3; average = 8.1 (slightly alkaline)
Density at 4°C (39°F)		1.000 g/cm^3	1.028 g/cm^3
Freezing point		0°C (32°F)	–1.9°C (28.6°F)
Boiling point		100°C (212°F)	100.6°C (213.1°F)

6.2. Processes affecting salinity

In the open ocean, salinity varies slightly, from about 33 to 38‰. Salinity variations are much more extreme in coastal areas, where freshwater input and evaporation can create brackish water (between seawater and freshwater) or hypersaline water (e.g. up to about 42‰ in the Red Sea). Salinity often varies seasonally based several factors that affect water input or removal (and to a much lesser degree, salt input or removal; Table 5.3).

TABLE 5.3 PROCESSES THAT AFFECT SEAWATER SALINITY

Process	How accomplished	Adds or removes	Effect on salt in seawater	Effect on H_2O in seawater	Salinity increase or decrease?	Source of freshwater from the sea?
Precipitation	Rain, sleet, hail, or snow falls directly on the ocean	Adds very fresh water	None	More H_2O	Decrease	N/A
Runoff	Streams carry water to the ocean	Adds mostly fresh water	Negligible addition of salt	More H_2O	Decrease	N/A
Icebergs melting	Glacial ice calves into the ocean and melts	Adds very fresh water	None	More H_2O	Decrease	Yes, icebergs from the Antarctic have been towed to South America
Sea ice melting	Sea ice melts in the ocean	Adds mostly fresh water and some salt	Adds a small amount of salt	More H_2O	Decrease	Yes, sea ice can be melted and is better than drinking seawater
Sea ice forming	Seawater freezes in cold ocean areas	Removes mostly freshwater	30% of salts in seawater are retained in ice	Less H_2O	Increase	Yes, through multiple freezings, called *freeze separation*
Evaporation	Seawater evaporates in hot climates	Removes very pure water	None (essentially all salts are left behind)	Less H_2O	Increase	Yes, through evaporation of seawater and condensation of water vapor, called *distillation*

Salts present in the oceans originally came from the crust and interior of the earth and were released through volcanism and hydrothermal vents. Physical and chemical weathering of rocks on land also adds salt to the oceans through river runoff. It is estimated that the oceans have been present for about 3.5 billion years, and salinity has remained stable for the last ~1.5 billion years. If salts are continuously added through the processes explained above, then they must also be removed by other processes for salinity to be stable (Figure 6.2). Removal of salts occurs in various ways. Sea spray leaves some salt on land. Shallow seas that evaporate over a long period of time leave salt deposits (evaporites). Biological organisms concentrate ions in their feces and shells, which may then be transferred to sediments. Salts in sediments may be returned to the interior of the earth as tectonic plates collide and one plate is subducted (pulled) under the other, along with some of the sediment overlying it. Finally, ions can adhere on the surface of small particles, e.g. clay, in a process called adsorption. Salts are then incorporated in sediments and do not return to seawater readily.

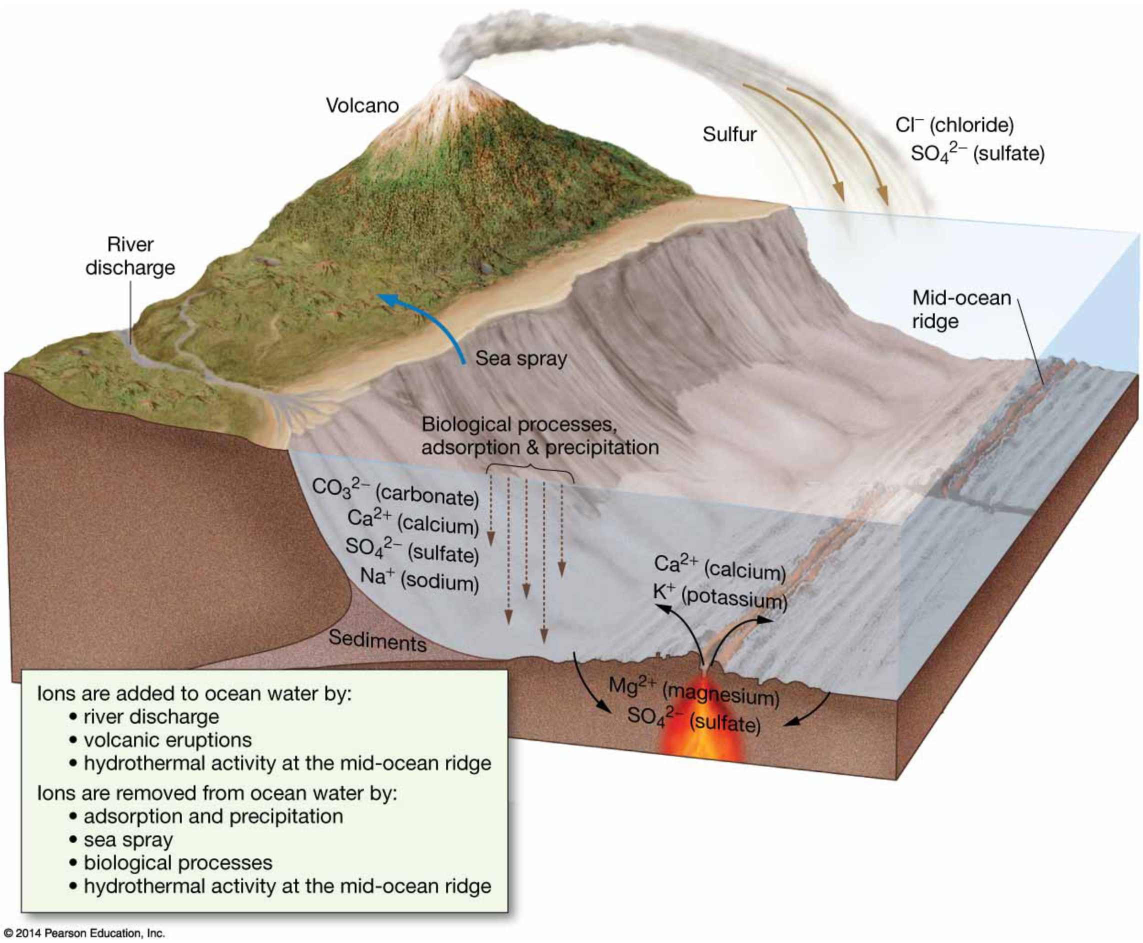


Figure 6.2. Cycling of dissolved materials in seawater

The residence time is the average time a substance remains in solution in the ocean. Conservative constituents have long residence times (e.g. millions of years), because they tend to be non-reactive with water and are not added or removed by biological processes.

Conservative constituents include major ions as well as some trace elements and some gases. Non-conservative constituents, on the other hand, are typically tied to biological, seasonal or geological cycles. They have a short residence time (< 1000 years) and include some trace elements, nutrients (Si, N, P) and some gases (e.g. O_2, CO_2).

6.3. Salinity variations with latitude and depth

While average ocean salinity is 35‰, there is considerable variation with latitude and depth. Salinity is highest around the Tropics of Cancer and Capricorn, where there is high evaporation and little precipitation. Near the equator, evaporation rates are also high, but high precipitation levels offset this loss of water. Salinity is lower at high latitudes because of low evaporation rates, high precipitation and runoff, and the melting of icebergs (Figure 6.3).

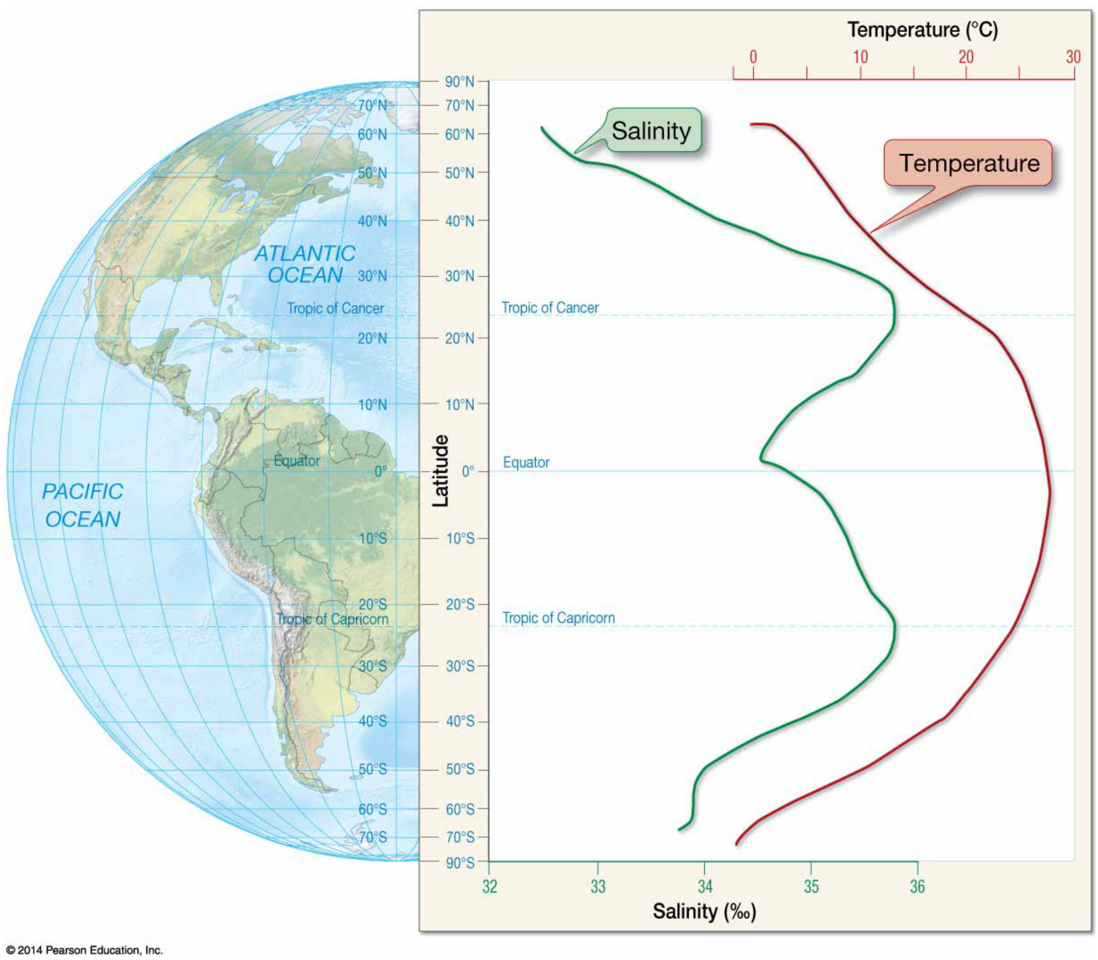


Figure 6.3. Variations in surface salinity with latitude.

Salinity is on average higher in the Atlantic than the Pacific, because this narrower ocean experiences higher levels of evaporation due to the land effects of the nearby continents

(Figure 6.4). Satellite-derived surface salinity maps show those patterns, as well as the lower salinity in coastal zones with high freshwater runoff (e.g. near the Amazon and Saint Lawrence rivers; notice also the low salinity of the Arctic ocean during the summer; Figure 6.4).

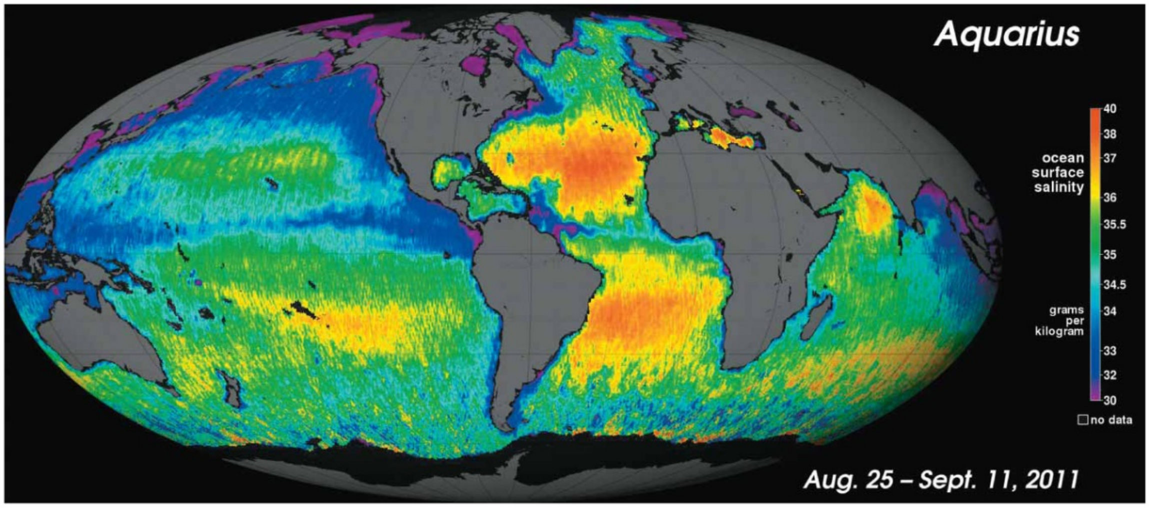


Figure 6.4. Surface salinity of the oceans derived from remote sensing.

Salinity also changes with depth, though the pattern depends on latitude. In the tropics, surface salinity is high and rapidly decreases from ~200-1000 m where it stabilizes around 35‰. At high latitudes the pattern is inversed: surface salinity is lower at the surface than at depth (Figure 6.5). Salinity at depth hardly changes since surface processes (evaporation, precipitation, melting of ice) are those that affect salinity. The region of sharp change in salinity with depth (either an increase or decrease, Figure 6.5) is called a halocline.

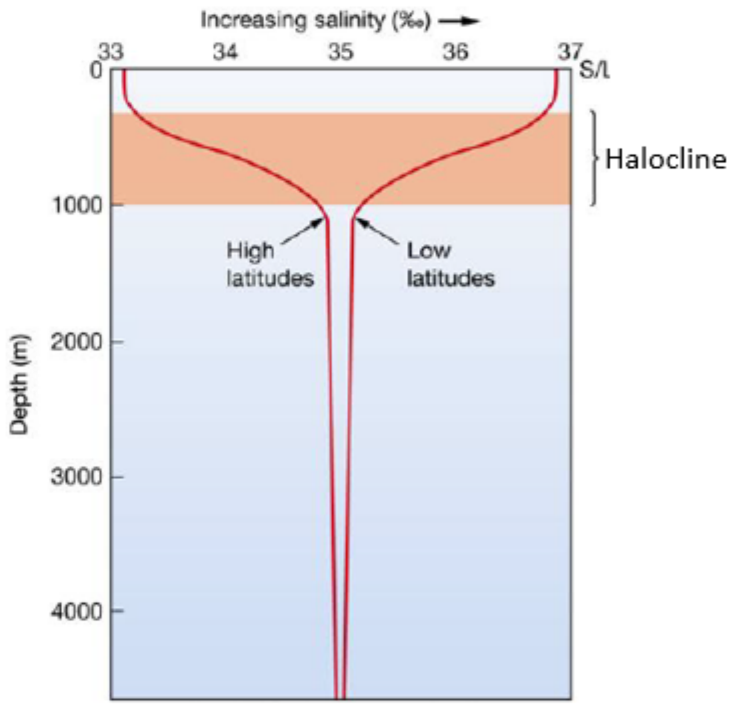


Figure 6.5. Typical patterns of salinity with depth at low and high latitudes.

6.5. Variation in seawater density with depth

The density of seawater is greater than that of freshwater (1.022-1.030 g/cm^3). Seawater density increases continuously with decreasing temperature until it reaches it freezing point (at -1.9^{o}C), where it rapidly increases in density (unlike freshwater, which has a highest at 4^{o}C). The density of a water mass determines whether it rises or sinks relative to other water masses. The ocean is layered according to density, with the densest water found at depth and less dense water found at the surface. Temperature, salinity and pressure all affect seawater density. Density increases with decreasing water, increasing salinity and increasing pressure. Of these pressure has the least effect and is only at play in the deep ocean, and can mostly be ignored. Temperature has the greatest effect on density because of the wide range of temperatures experienced in the ocean but salinity can also be important, especially in polar regions where temperature is stable year-round. The density of seawater resulting from its temperature and salinity is important to deep ocean circulation because high density water (e.g. cold, high-salinity water) sinks below less dense water.

In warm regions, where surface water is continuously heated by the sun, a layer of relatively well-mixed warm water extends to about 300m. Temperature then rapidly drops until ~1000m and then remains constant all the way to the ocean's bottom (Figure 6.6). The region of sharp change in temperature is called a thermocline. Because temperature has the greatest influence on density, the thermocline is accompanied by a pycnocline, a region of rapid change in density. This creates three main water masses: the mixed surface water above the permanent thermocline, which has the lowest density, the upper water which is the zone encompassing the thermocline (and corresponding pycnocline), and the deep water below, which has the highest density. At high latitudes, this thermocline is absent because the entire water column is cold (water is isothermal). Similarly, there is little change in density with depth (water is isopycnal). Therefore vertical water mixing occurs readily. Note that in temperate zones, a temporary thermocline and pycnocline is set up during the summer. More on this in chapter 16 of this booklet.

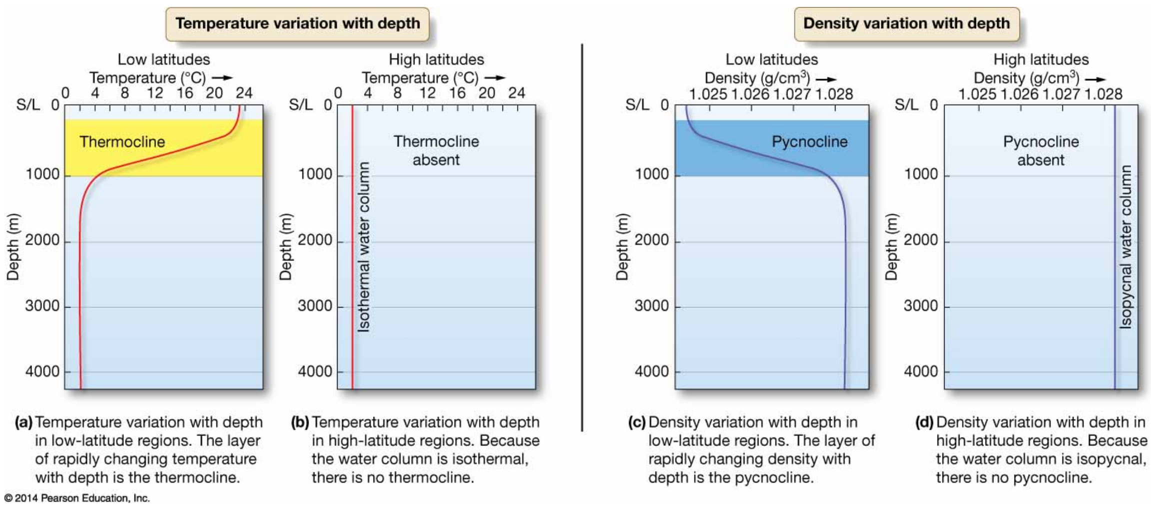


Figure 6.6. Temperature and density variations with depth, at low and high latitudes.

6.6. Review Questions

1. What is the average salinity of the oceans (include units)?

2. Where would you expect to find a higher average open ocean salinity; in the tropics (5^o N & S of equator) or in the desert belts (25^o N & S)?

3. How does salinity compared between the Atlantic and the Pacific?

4. What property of water makes it such a good solvent?

5. Name 3 different processes that add salts in the oceans?

6. Name 3 different processes that remove salt from the ocean?

7. What are the 6 major ions dissolved in seawater?

8. What is a trace element?

9. What is the residence time of a substance?

10. Is nitrate a conservative or a non-conservative constituent of seawater?

11. What is the principle of constant composition?

12. How does temperature affect the salt saturation value of seawater?

13. What is the name of the device used to determine salinity by measuring the bending of light as it passes from air to the water sample?

14. What effect does adding salt have on the boiling point and freezing point of water?

15. Is salt water denser or less dense than freshwater?

16. What do you call a rapid change in salinity with depth?

17. What do you call a rapid change in temperature with depth?

18. What do you call a rapid change in density with depth?

7. Gases and nutrients in seawater (Duxbury, Chapter 6)

7.1. Gases in Seawater

Several gases are dissolved in seawater. Atmospheric gases enter the surface of the ocean and distribute to depth by currents and mixing. Some (e.g. O_2 and CO_2) are also produced by biological processes.

The amount of gases that can be held in solution is known at the saturation value (or saturation concentration), and varies with temperature, salinity and pressure. A decrease in temperature, a decrease in salinity, and an increase in pressure all lead to a higher saturation value of gases.

Atmospheric gases enter the ocean at the surface through diffusion and wave action. They can be moved to depths with dense bodies of water that sink from the surface (e.g. cold surface water at the poles). Gases near the sea surface tend to be saturated with atmospheric values. At depth, they reflect biological processes (photosynthesis, respiration and decay) and geological processes (e.g. volcanoes; Table 7.1).

Atmospheric nitrogen cannot be used by most biological organisms, but can be fixed (i.e. made into ammonium, which is usable by other organisms) by some bacteria. Oxygen is produced by photosynthesis and used in respiration and decay. Carbon dioxide is produced by respiration and decay and used in photosynthesis.

Table 7.1. Abundance of gases in air and seawater (From Sverdrup *et al*, 2004)

Gas	Symbol	% in atmosphere	% in surface seawater	% in total oceans
Nitrogen	N_2	78.03	48	11
Oxygen	O_2	20.99	36	6
Carbon dioxide	CO_2	0.03	15	83
Argon, Helium, neon,	Ar, He, Ne	0.95	1	
Total		100.00	100	100

Concentration of Oxygen and Carbon Dioxide with Depth

Photosynthetic organisms are confined to the sea surface, where sunlight is available (to ~100-150m deep). In this region, oxygen is produced and carbon dioxide is consumed through photosynthesis in the following reaction:

$$CO_2 + H_2O \leftrightarrow C_6H_{12}O_6 + O_2$$

All organisms (animals and plants) also respire to meet their metabolic needs, using oxygen and glucose and producing carbon dioxide in the following reaction. Respiration occurs at all depths:

$$C_6H_{12}O_6 + O_2 \leftrightarrow CO_2 + H_2O$$

Plants have a lower rate of photosynthesis in lower light levels. Therefore, the rate of photosynthesis decreases with increasing depth, until light levels are so low that plants can only photosynthesize enough to meet their own metabolic needs. This depth, where the rate of photosynthesis is equal to the rate of respiration for plants and plant-like organisms, is known as the compensation depth (Figure 7.2). At this depth, the net oxygen production is zero. There can be no growth of photosynthetic organisms below it. The compensation depth varies with water clarity, and usually corresponds to approximately 1% of surface light levels, and can be as deep as 150m. Decomposition (decay) of dead organisms, like respiration, also occurs at all depths; it uses oxygen and releases carbon dioxide.

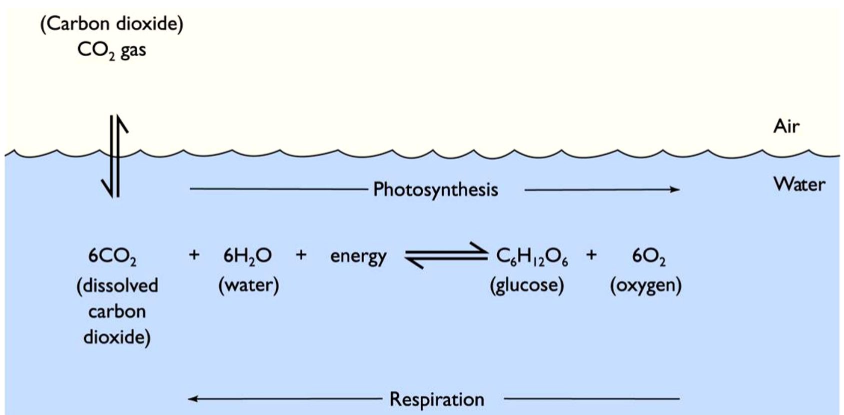


Figure 7.1. Photosynthesis and respiration reactions.

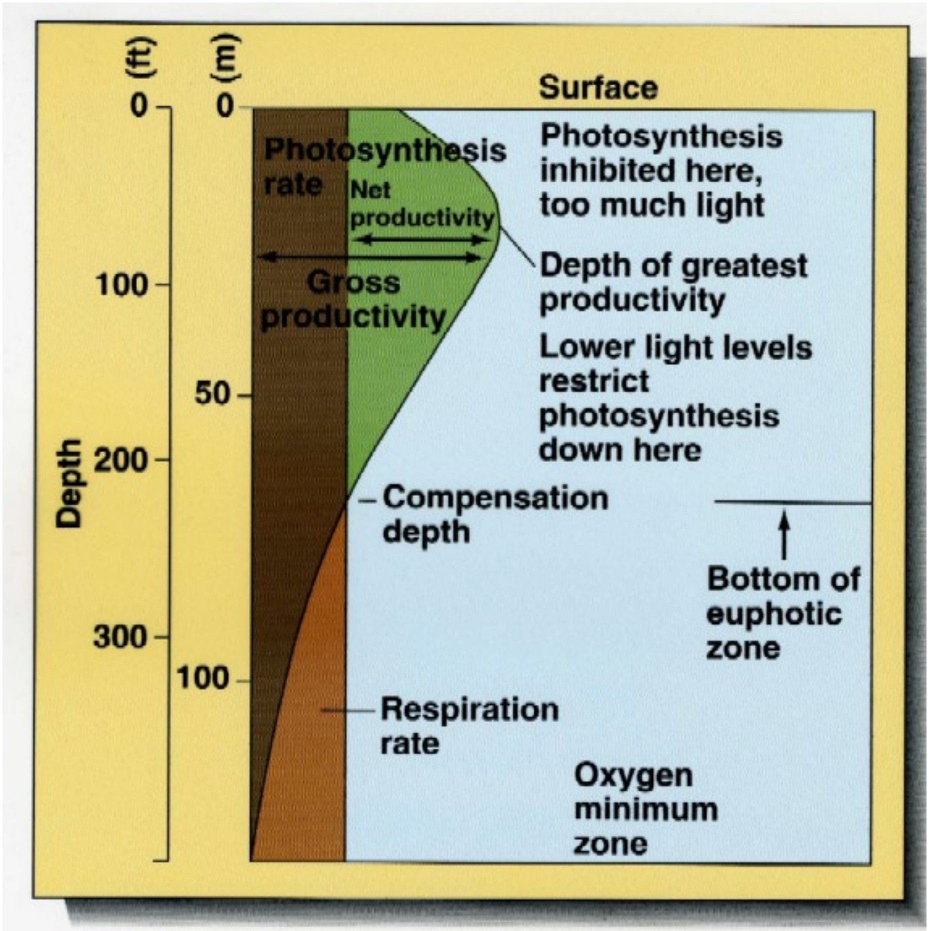


Figure 7.2. The compensation depth, where the rate of photosynthesis is equal to the rate of respiration.

The typical pattern of concentration of oxygen and carbon dioxide with depth can be seen in Figure 7.3. Oxygen concentration is highest near the surface, where sunlight is abundant and the rate of photosynthesis is high. It decreases slightly right at the surface because of a process called photoinhibition, where photosynthesis is reduced at very high light levels. Oxygen concentration decreases below the surface layer because it is used in respiration and decomposition. It reaches a minimum around 800m deep, which often coincides with a pycnocline (a rapid change in density with depth) and an accumulation of decaying organic material. Below the oxygen minimum, the density of animals is reduced and the rates of respiration and decomposition are reduced. Further, deep water typically originates from cold, dense, oxygen-rich surface water from polar regions. Carbon dioxide, on the other hand, is least abundant in surface waters, where it is used in photosynthesis, and gradually increases in concentration with depth as it is produced by respiration and decomposition.

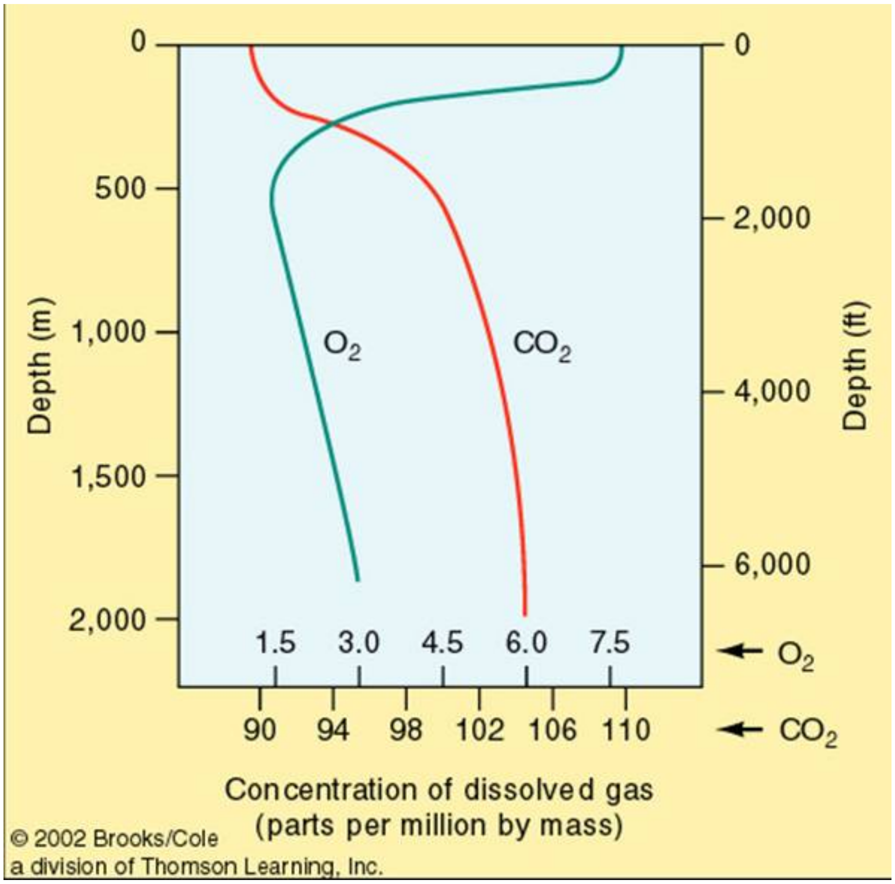


Figure 7.3. Vertical profile of oxygen and carbon dioxide in the Atlantic Ocean.

7.2. Seawater pH

The acidity or alkalinity of a substance is measured with the pH scale, a logarithmic scale from 1 to 14, which measures the concentration of hydrogen ions (H^+). A pH of 7 is neutral; lower pH is acidic (higher concentration of H^+) and higher pH is basic, or alkaline, with a low concentration of H^+ and high concentration of hydroxide ions OH^- (Figure 7.5). Pure water has a pH of 7; while there is constantly a small amount of free hydrogen ions and hydroxide ions due to the dissociation and reformation of water molecules ($H_2O \leftrightarrow H^+ + OH^-$), their concentration is equal and therefore the pH remains neutral. When other substances are dissolved in water, they can alter the balance of these ions and therefore change the pH of the solution. Since the pH scale is logarithmic, a solution with a pH of 6 has 10 times more free hydrogen ions (is 10 times more acidic) than one with a pH of 7. Therefore, seemingly small changes on the scale can mean large changes in water chemistry. This is important because biological processes must happen within a narrow range of pH. The average surface ocean pH is 8.1, and it varies between 8.0 and 8.3. Carbon dioxide readily combines with water to form carbonic acid (H_2CO_3), which can dissociate into bicarbonate (HCO_3^-) and a hydrogen ion (H^+):

$$CO_2 + H_2O \leftrightarrow H_2CO_3 \leftrightarrow HCO_3^- + H^+$$

This reaction clearly shows that increasing the amount of CO_2 in the oceans should increase ocean acidity. The change in pH is however buffered by another reaction involving calcium carbonate, which at lower pH dissociates in to Ca^{2+} and CO_3^{2-} (Figure 7.5). The carbonate ions (CO_3^{2-}) then react with free hydrogen ions to form bicarbonate (HCO_3^-). This process removes free H+ from the water, therefore buffers the acidifying effect of CO_2 and prevents large changes in pH. This is called the carbonate buffering system of the ocean. Recently however, large increases in CO_2 from man-made causes have overwhelmed this buffering system and oceans are becoming more acidic. More on this in chapter 20.

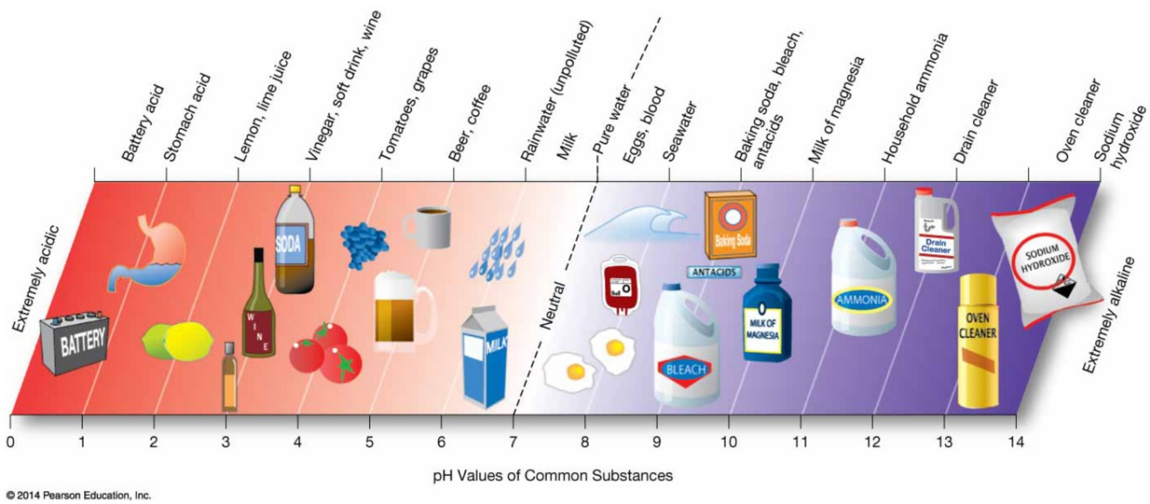
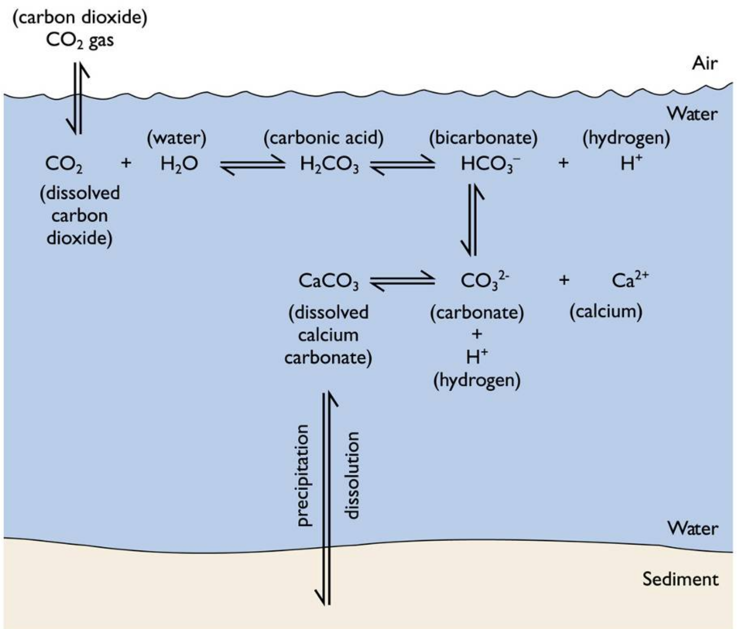


Figure 7.4. The pH scale, showing the pH of common substances. A pH 7 is neutral and a change of 1 pH unit translates to a 10-fold increase in free hydrogen ions.



(c) THE CO_2 SYSTEM

Seawater too basic: $H_2CO_3 \longrightarrow HCO_3^- + H^+$ pH drops

Seawater too acidic: $HCO_3^- + H^+ \longrightarrow H_2CO_3$ pH rises

(d) CARBONATE BUFFER

Figure 7.5. The carbonate buffering system, showing how calcium carbonate can dissociate to allow carbonate ions to take up free hydrogen ions, thereby reducing the change in pH associated with increased CO_2.

Because seawater pH is strongly tied to CO_2 concentration, it varies with depth in relation to CO_2 levels, with highest pH at the surface and a pH minimum around 800m, which coincides with high CO_2 levels caused by animal respiration at a depth where photosynthesis is not possible (Figure 6.6).

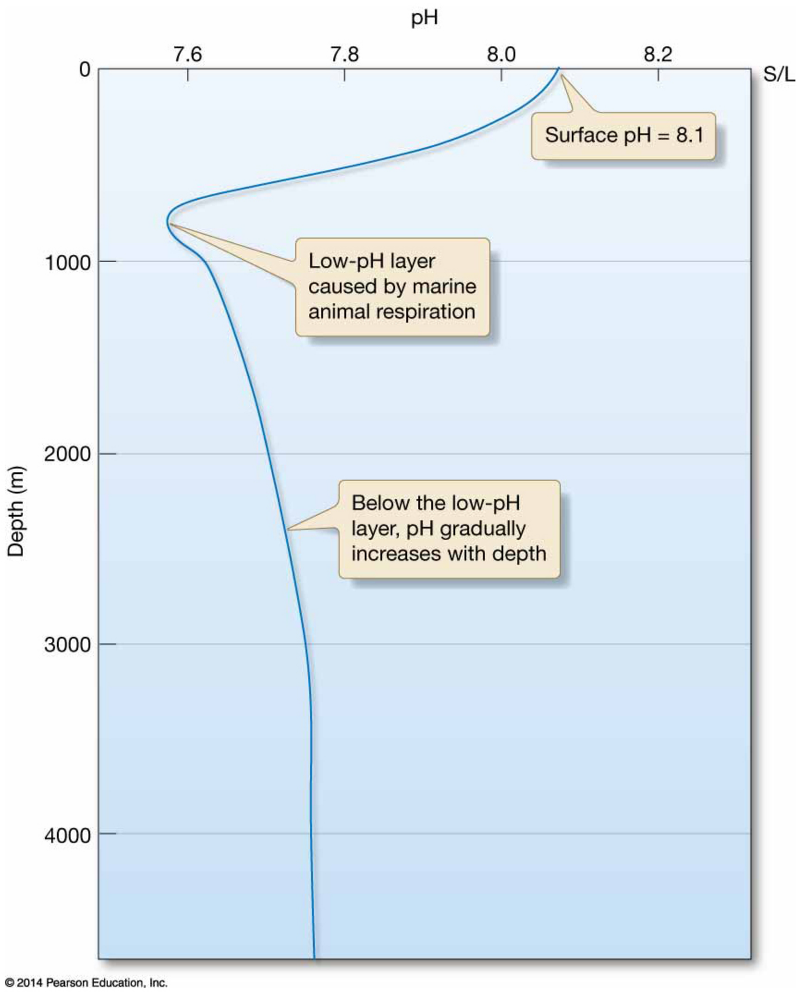


Figure 7.6. Changes in pH with depth in seawater.

7.3. Nutrients in seawater

Nutrients are inorganic molecules required for the growth of photosynthetic organisms. All plants require nitrogen and phosphorus, which for the most part must be in the form of nitrate (NO_3^-) and phosphate (PO_4^{3-}) for the plants to use them. Some types of phytoplankton, like diatoms, also need silicon in the form of silicate (SiO_4^-) to form their silica frustule (shell). As photosynthesis occurs and phytoplankton populations grow, these nutrients become depleted in the photic zone. There may be ample nutrients below the photic zone, but they must brought back up to the photic zone by vertical mixing for photosynthesis to continue. The layering and stability of the water column discussed in section 5.8 plays an important role in the availability of nutrients in surface waters where light is sufficient to sustain photosynthesis.

Permanent thermoclines occur in the tropics, where the surface of the ocean is warm year round and this warm water, being less dense, always remains near the surface. In areas with a permanent thermocline, nutrients are depleted in the photic zone and nutrient-rich deeper waters are not brought to the surface because the strong thermocline prevents vertical mixing, and therefore photosynthesis and productivity are low (Figure 7.7), resulting in clear waters.

Seasonal thermoclines, on the other hand, occur in temperate zones in the summer, when surface water is heated and becomes less dense. In the fall, as the temperature of the surface layer drops, this water becomes denser and sinks to the bottom (Figure 7.8). As surface water sinks to the bottom, nutrient-rich bottom water is brought to the surface in the photic zone, where it can be used by photosynthetic organisms. In temperate zones, phytoplankton biomass is at its highest in the spring, when nutrients are plentiful and there is enough sunlight to sustain photosynthesis; it decreases in the summer as the seasonal thermocline appears and nutrients become depleted; there is a second small phytoplankton bloom in the fall that corresponds to the start of the fall overturn and to the release of nutrients from zooplankton decay, until sunlight becomes limiting and production drops during the winter (Figure 7.7 & 7.8).

In polar regions, the water column is always well-mixed and nutrients are abundant, but light is only abundant enough for photosynthesis for a short period in the summer. Productivity in polar regions is limited by light, and there is a strong, yet short peak of primary production in the summer when sunlight is sufficient (Figure 7.7).

Nutrients may also be brought to the surface by upwelling: where deep water is forced to the surface. More on the factors controlling nutrient levels and primary productivity in chapter 16 of this booklet.

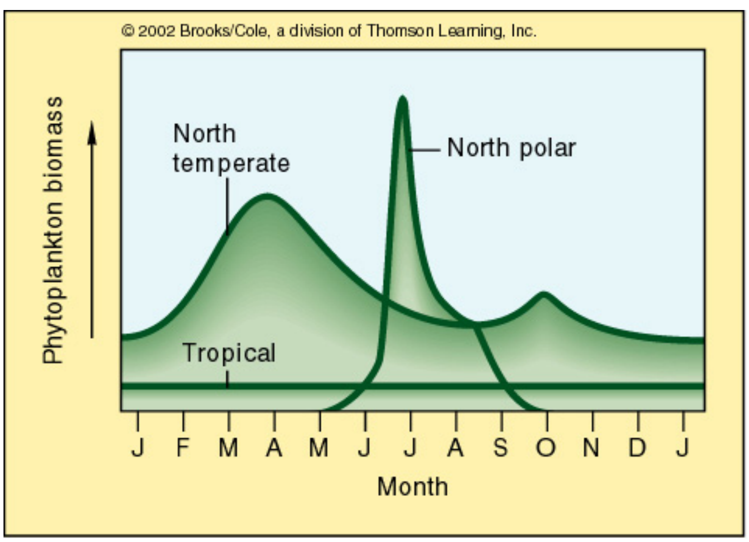


Figure 7.7. Annual variation in phytoplankton biomass (or primary production in tropical, temperate and polar regions.

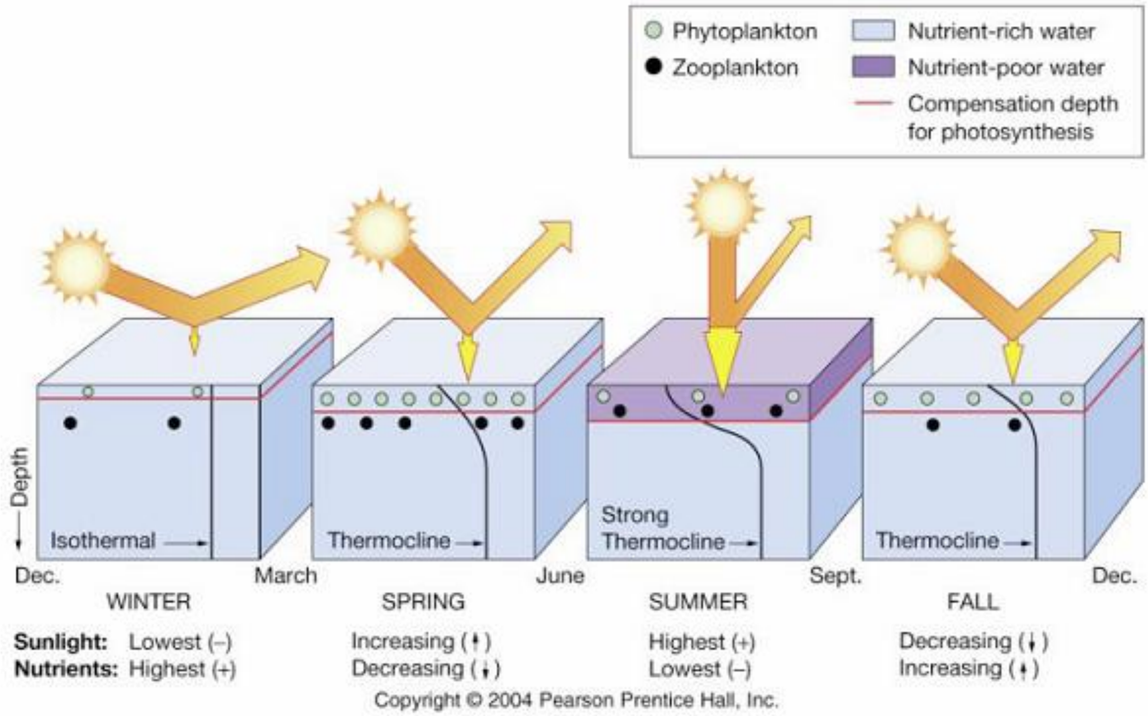


Figure 7.8. Seasonal variation in thermocline and mixing in temperate oceans. Nutrients are abundant when deep, nutrient-rich water is brought to the surface (from fall to spring). Primary production (phytoplankton growth) is high when nutrients are present and there is enough sunlight (spring and fall).

7.4. Review Questions

1. How does temperature affect the gas saturation value of seawater?

2. Describe the type body of water in which you would expect to find the greatest amount of gas held in solution

3. What is the compensation depth?

4. Where is oxygen most abundant in the water column?

5. Which biological process adds oxygen in the oceans?

6. Where is carbon dioxide most abundant in the water column?

7. Which two biological processes remove oxygen and add carbon dioxide?

8. What form must phosphorus be in to be used by photosynthetic organisms?

9. What is the average pH of ocean water?

10. Which compound plays an important role in the buffering capacity of seawater?

11. How does the addition of CO_2 affect ocean pH?

12. How would the addition of hydrogen ions affect ocean pH? How would it affect the buffering capacity reaction?

13. How many more free hydrogen ions are in a solution of pH 6 compared to a solution of pH 7?

14. At which latitude is primary production low year-round?

8. Transmission of Energy in the Water (Duxbury, Chap. 5)

8.1. Heat

Heat can be transmitted in three ways: conduction, convection and radiation (Figure 8.1). Conduction is the increase in molecular movement from fast-moving nearby molecules, e.g. when the handle of a metal spoon is heated after the other end of the spoon is placed in a hot liquid. Water is a poor conductor. Convection is driven by changes in density; warmer water is less dense and rises, carrying heat with it and spreading it to other areas. Radiation is the direct transmission of heat from its energy source, such as the heating of the earth from the sun. Heat can be radiated without a medium (e.g. in space).

Solar radiation hits the surface of the ocean, and warms the surface layer. This warm water tends to remain at the surface because it is less dense than the underlying colder water, and because water is a poor heat conductor. Cold warm from greater depths may return to the surface by seasonal mixing or upwelling, but otherwise stays under the thermocline.

Figure 8.1. Conduction, convection and radiation are all shown as this pot of water is heated over a fire.

8.2. Light

The sun radiates many forms of electromagnetic energy, which includes gamma rays, x-rays, ultraviolet, visible light, infrared, microwave and radio waves (Figure 8.2). However, of all the electromagnetic spectrum, only visible light is transmitted in water. About 60% of the light that reaches the sea surface is absorbed in the first meter; 80% in the first 10m, and 99% in the top 150m. No light penetrates below 1,000m. This decrease in light intensity with depth is called attenuation, and is affected by the amount of suspended particles and organisms. Light attenuation can be measured with a simple device called a Secchi disk. The visible light can be further divided into its various color components, from red (longest wavelength) to violet (shortest wavelength). When visible light is transmitted through seawater, the various colors are not all transmitted equally, and the longest wavelengths (red) are absorbed first. Blue wavelengths reach the deepest, which is why objects tend to look blue underwater (Figure 8.3).

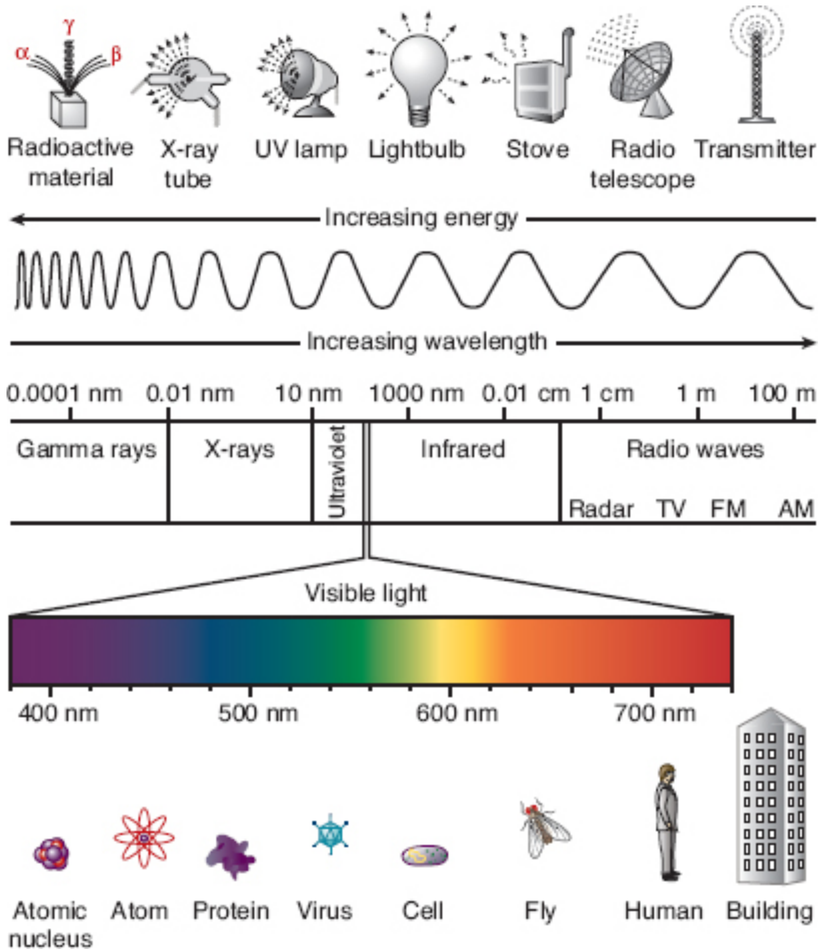


Figure 8.2. The electromagnetic spectrum, including visible light.

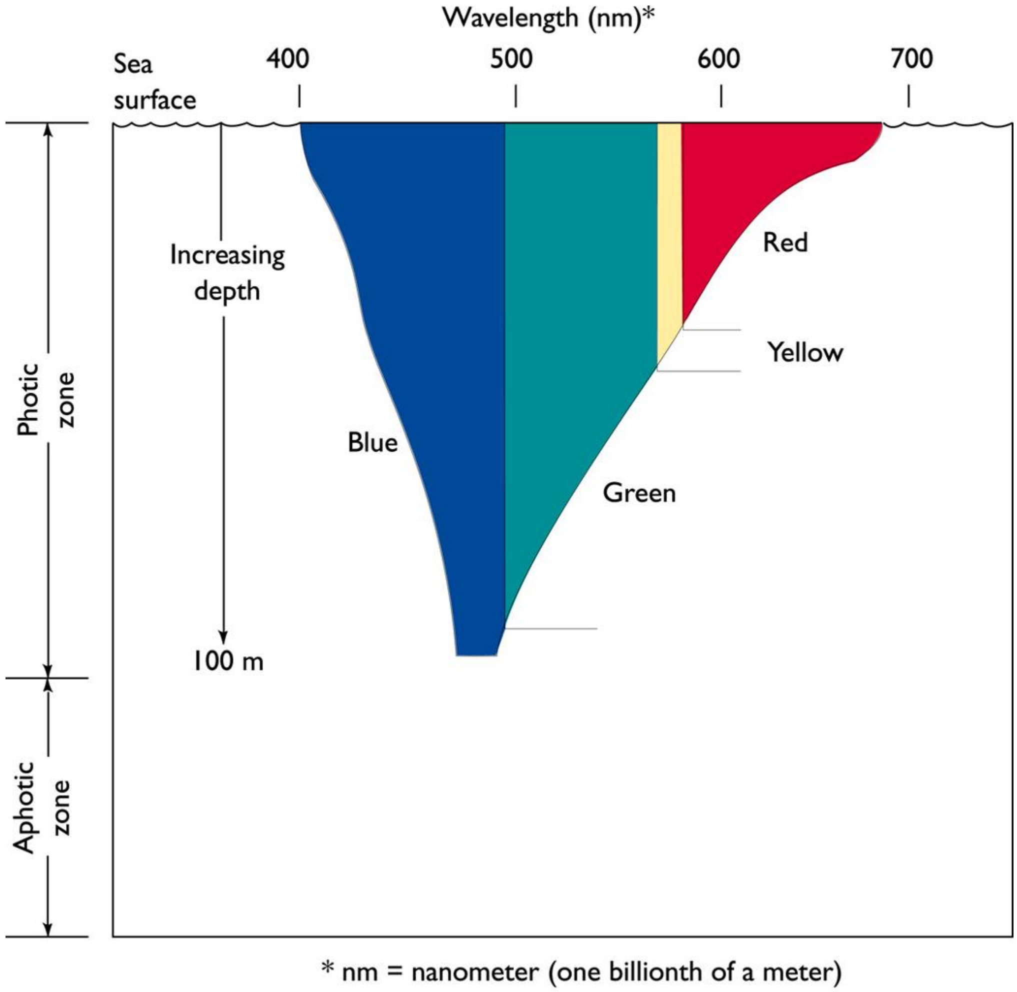


Figure 8.3. Light absorption in the open ocean.

In nearshore environments, the increased amount of particles reflecting light in the water increases the light attenuation and therefore decreases the depth of light penetration. Moreover, coastal waters often appear green rather than blue because abundant particles and organisms reflect those wavelengths (Figure 8.4).

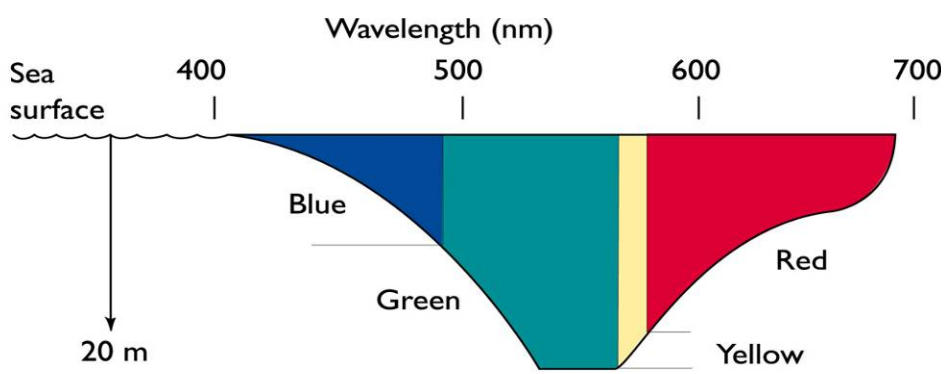


Figure 8.4. Light absorption in nearshore waters.

The speed of light is faster in air than in water and this causes light to bend as it enters the water (Figure 8.5). The degree of refraction (bending) is affected by temperature and salinity (the denser the water, the greater the refraction). This allows us to easily measure salinity in seawater by measuring the refraction of light with a device called a refractometer.

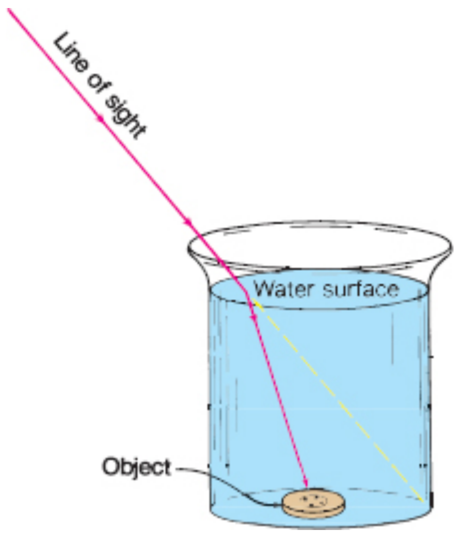


Figure 8.5. Objects that are not directly in the individual's line of sight can be seen in water because of the refraction of the light rays. The refraction is caused by the decreased speed of light in water.

8.3. Sound

Sound travels 4.5 times faster in water than in air. Many animals use sound to navigate, find their prey and communicate. Because the energy of high frequency sounds dissipates faster than that of low frequency sounds, the low frequencies travel further.

When sound hits an object, it is reflected back to the source and can be used to measure distance, direction and properties of the object. This is how marine mammals use sound to navigate and find prey through echolocation. This is also how SONAR and depth finders work aboard boats (Figure 8.6).

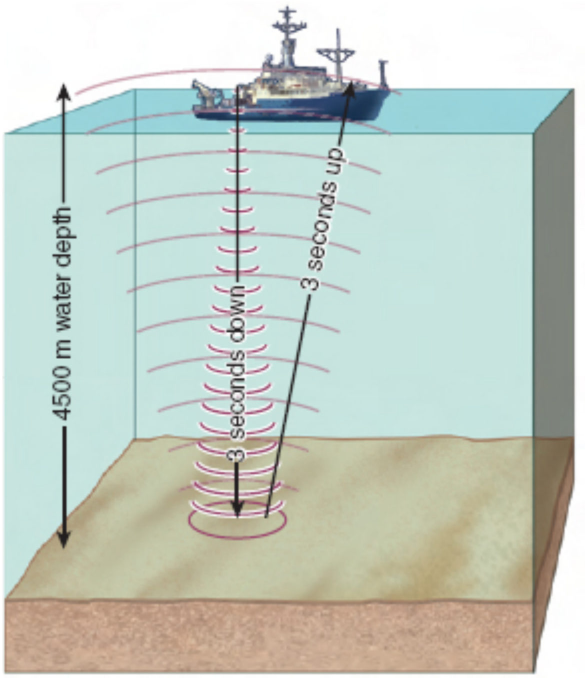


Figure 8.6. Traveling at an average speed of 1500 m per second, a sound pulse leaves the ship, travels downward, strikes the bottom, and returns. In 4500 m of water, the sound requires three seconds to reach the bottom and three seconds to return.

The speed of sound increases with increasing temperature, pressure and salinity (Figure 8.7). There is no constant trend of salinity with depth and therefore salinity is not as important as temperature and pressure when considering speed of sound with depth through the water column. Because of changes in temperature and pressure with depth, the speed of sound is not constant through the water column; it reaches a zone of minimum velocity around 1,000m (Figure 8.4). This zone of minimum sound velocity is known at the SOFAR channel (SOund Fixing and Ranging), and sound produced in this channel tends to be refracted back in the channel, and therefore travels great distances (Figure 8.8). Speed of sound increases above the SOFAR channel because of increased temperature, and increases below the SOFAR channel because of increased pressure.

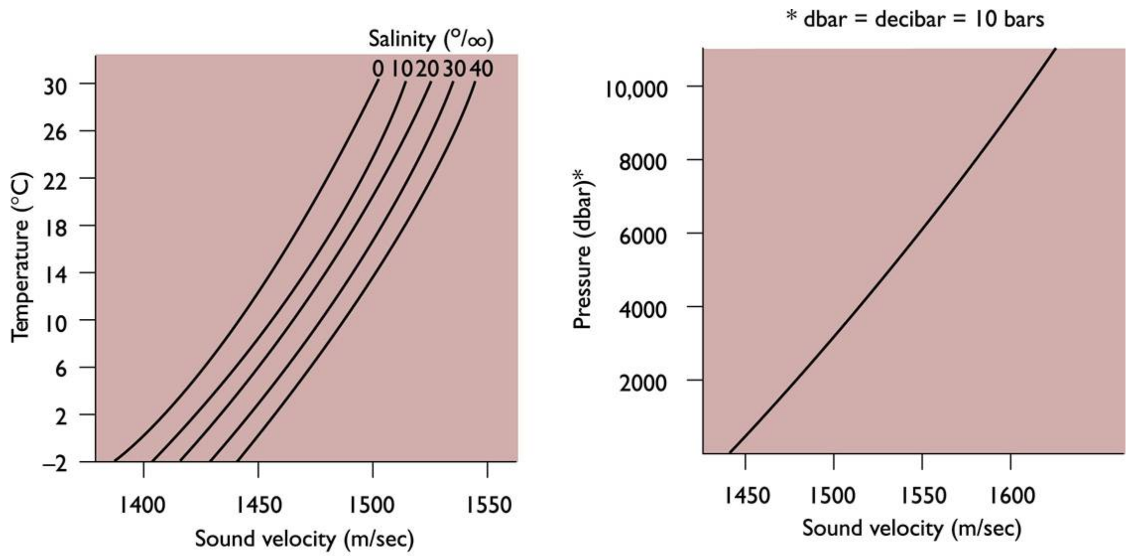


Figure 8.7. The speed of sound in water increases with temperature, salinity and pressure.

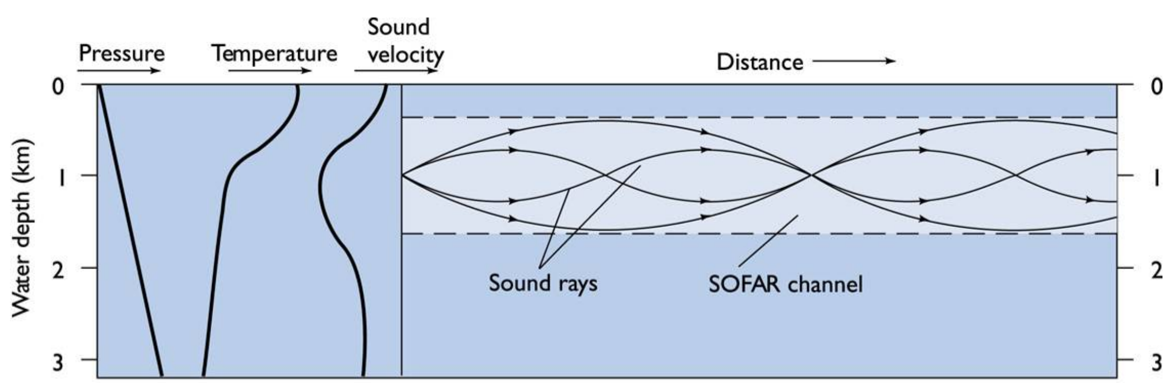


Figure 8.8. The SOFAR channel, around 1000m, where temperature and pressure create a zone of minimum sound velocity.

8.5. Review Questions

1. In which 3 ways can heat be transmitted?

2. How deep can light penetrate in the oceans?

3. Which wavelength of light is absorbed first?

4. How can light attenuation be measured?

5. What factors influence light attenuation?

6. What is refraction?

7. Which water properties influence its degree of refraction?

8. Which sound frequencies travel the farthest?

9. What property of sound allows for echolocation and SONAR?

10. How does the speed of sound change with temperature, pressure and salinity?

11. Where is the zone of minimum sound velocity?

12. How do sounds get transmitted greater distances when travelling in the SOFAR channel?

9. Air-Sea Interaction (Trujillo, chapter 6)

9.1. Variations in Solar Radiation

The earth experiences seasons because its axis of rotation is tilted in relation to the Earth's rotation around the sun. Therefore, as the earth rotates around the sun, the angle of the sun relative to the equator (the sun's declination) varies between 23.5°N and 23.5°S. The region between these latitudes receives much greater yearly solar radiation than those at higher latitudes, for four reasons (Figure 9.1). First, sunlight strikes low latitudes at a right (or high) angle, therefore the amount of radiation is concentrated over a smaller area than the same amount of sunlight at high latitudes. Second, sunlight travels through more atmosphere at high latitudes. Since the atmosphere absorbs of the incoming solar radiation, less radiation strikes the Earth at high latitudes. Third, more solar radiation is reflected at high latitude, where there is an increased amount of white material (snow and ice) which reflects it. This is called the albedo effect. Finally, the angle at which sunlight strikes the surface of the ocean determines how much of that light is absorbed, so less solar radiation is absorbed by oceans at high latitudes. For all these reasons, there is much more solar energy absorbed near the equator than at higher latitudes. Ocean and atmospheric circulation helps redistribute excess heat from the equator to the poles.

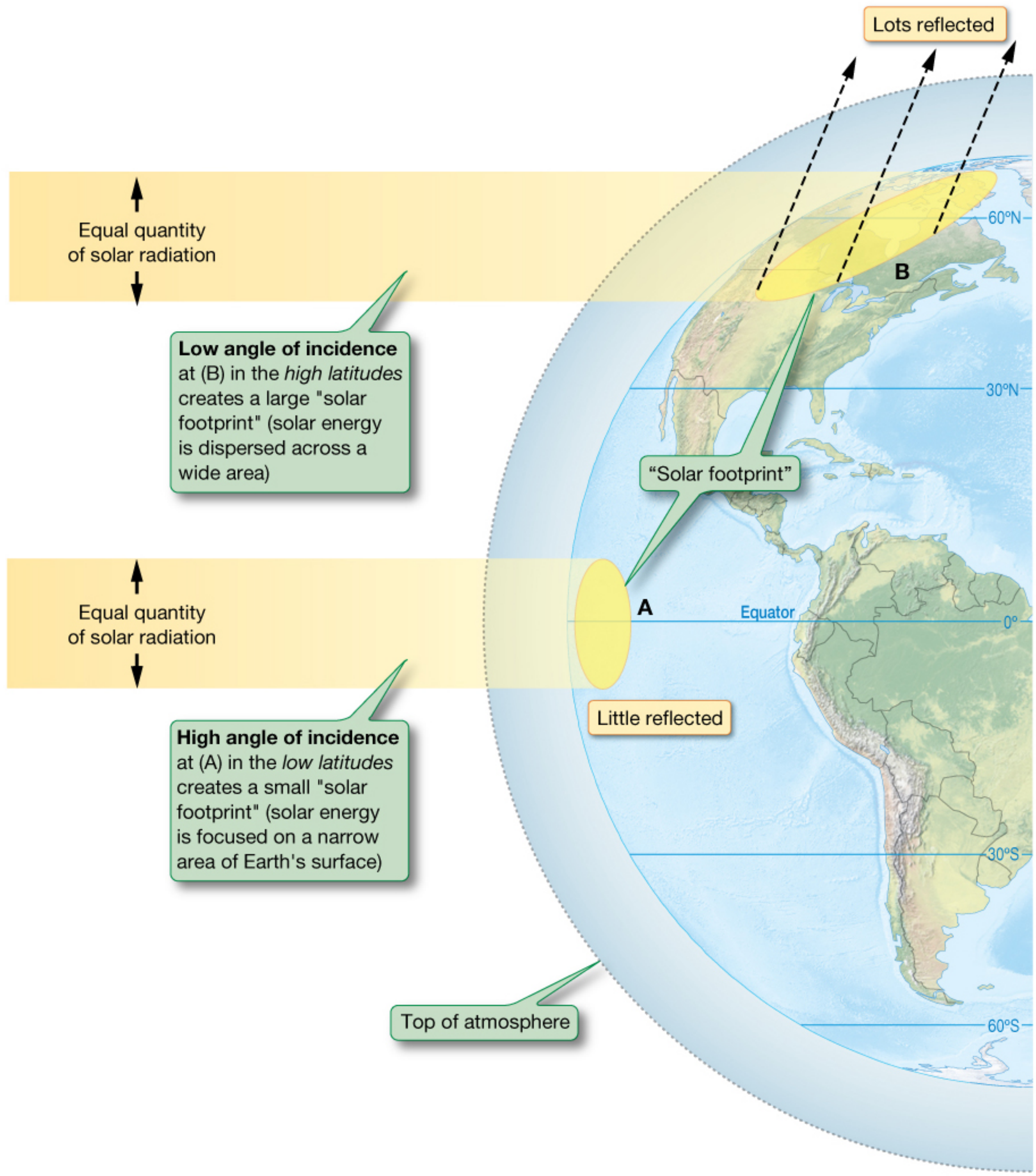


Figure 9.1. Solar radiation is more concentrated and passes through less atmosphere at low latitudes than high latitudes.

9.2. Physical Properties of the Atmosphere

The atmosphere is a mixture of gases that extends up about 90 km from the Earth's surface. Yet because of increased pressure near the surface of the earth, 99% of the gases in the atmosphere are below 30 km, and 90% of it is below 15km. The atmosphere is composed of nitrogen (N_2, 78.1%), oxygen (O_2, 20.9%), argon (0.9%) and carbon dioxide (CO_2, 0.039%) and some other inert gases (Figure 9.2).

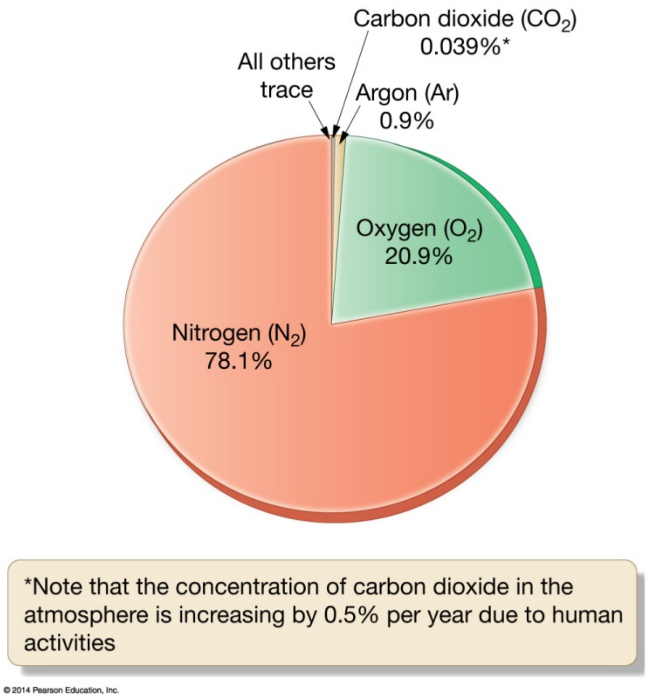


Figure 9.2. Composition of dry atmosphere.

Distinct layers can be identified in the atmosphere. The lowest layer is called the troposphere, and extends from the surface of the earth to an altitude of about 12 km. In the troposphere, temperature decreases with altitude because the troposphere is heated from the earth below. This decreasing in temperature occurs at a rate of approximately –10ºC for each 1,000m of elevation, to a minimum of about – 60ºC. In the stratosphere (from 12 to 50 km of altitude), temperature increases with altitude because of the ozone (O_3) layer located here, which absorbs UV from the sun. Temperature decreases again in the upper atmosphere (Figure 9.3).

Temperature has an important effect on the density of air. At higher temperature, air is less dense and it rises; cold air is denser and it sinks. These variations in density can create convection cells much the same way as would be set up in a room heated from below (Figure 9.4). Differential heating of the atmosphere at various latitudes sets the gases in motion, and creates winds. Air moves from areas of high pressure to areas of low pressure

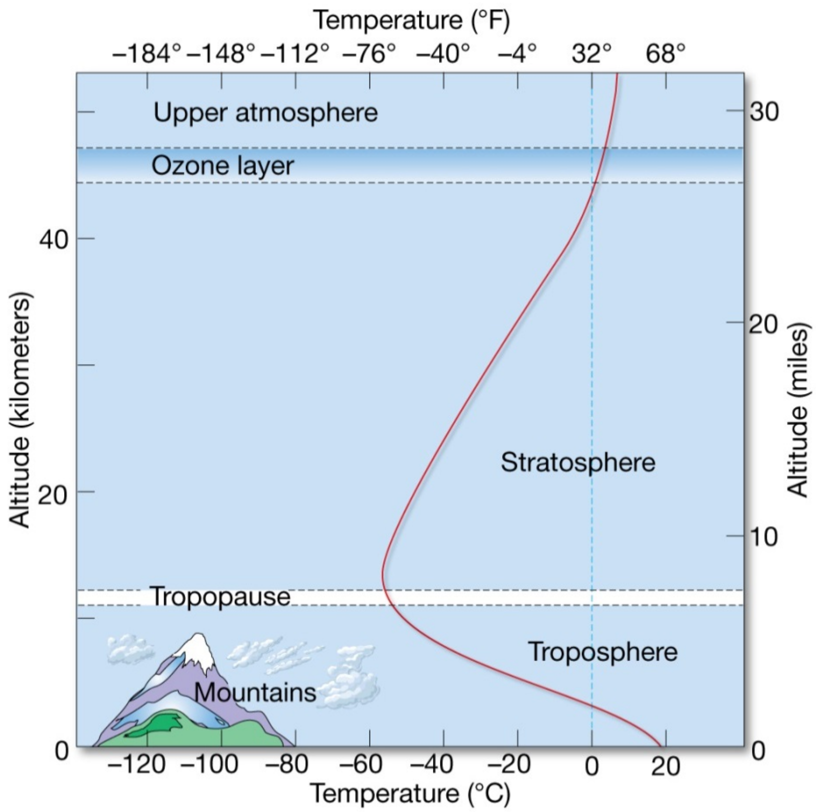


Figure 9.3. Temperature profile and the various layers of the atmosphere.

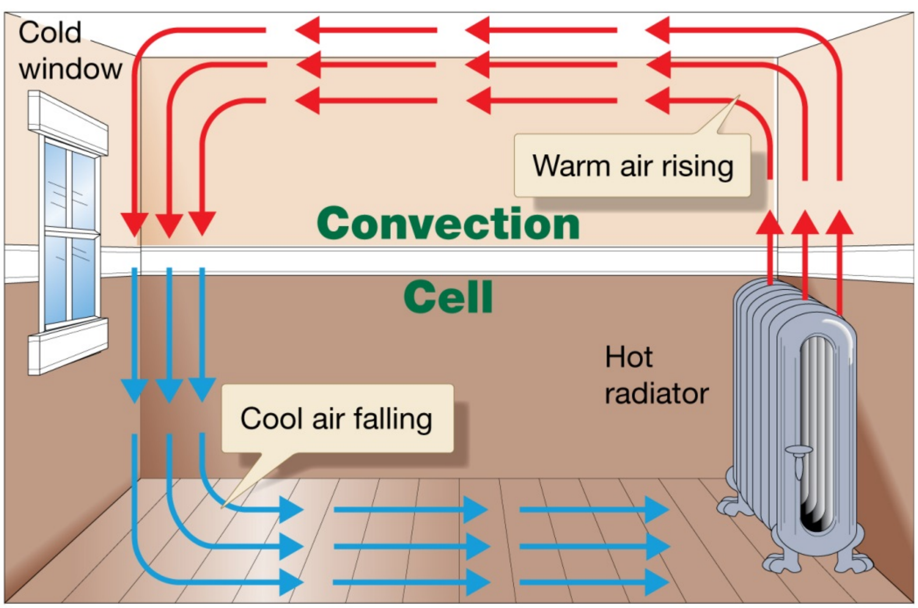


Figure 9.4. A convection cell if a room, created by the heating of the air below (by the radiator) and cooling of air higher up (by the window).

Non-Rotating Earth

Most of the atmosphere is located within the troposphere and winds in this layer are the most important when considering atmospheric circulation. Consider first a theoretical non-rotating earth with no land masses. As explained earlier, the equator receives more heat than higher latitudes, and warm, moist air (which is less dense) rises at the equator, creating an area of low pressure (Figure 9.5). As this moist air rises, its moisture condensates, creating precipitation at the equator. This air (which is now dry) then flows North or South towards higher latitudes, and sinks back down near the poles. The air then flows back towards the equator along the earth's surface, creating a large convection cell (Figure 9.6). The air flowing along the earth's surface is wind. In this model, the wind would blow steadily from the poles to the equator.

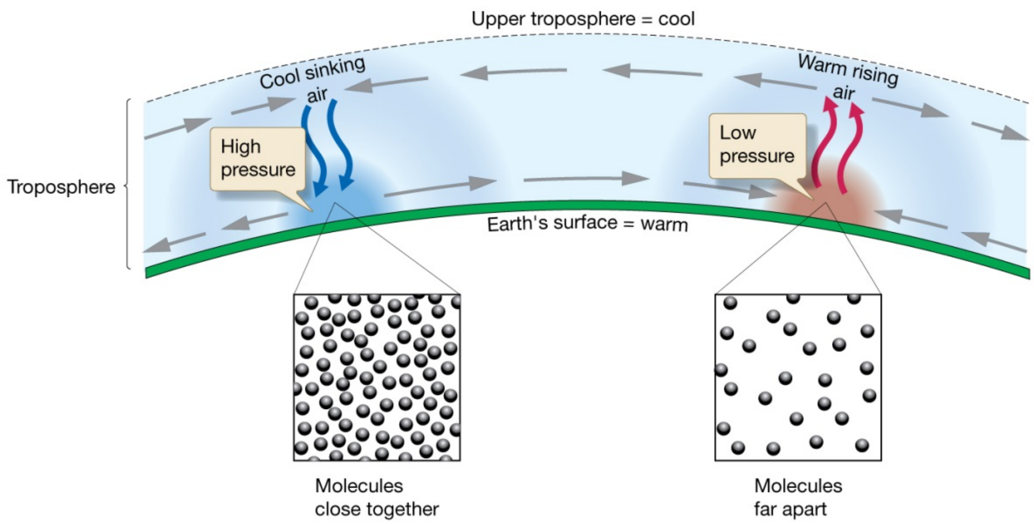


Figure 9.5. Zones of low atmospheric pressure are created in warm areas where the air above the earth is less dense and rises. High pressure zones are created where cool air sinks. Winds at the surface of the earth move from areas of high pressure to areas of low pressure.

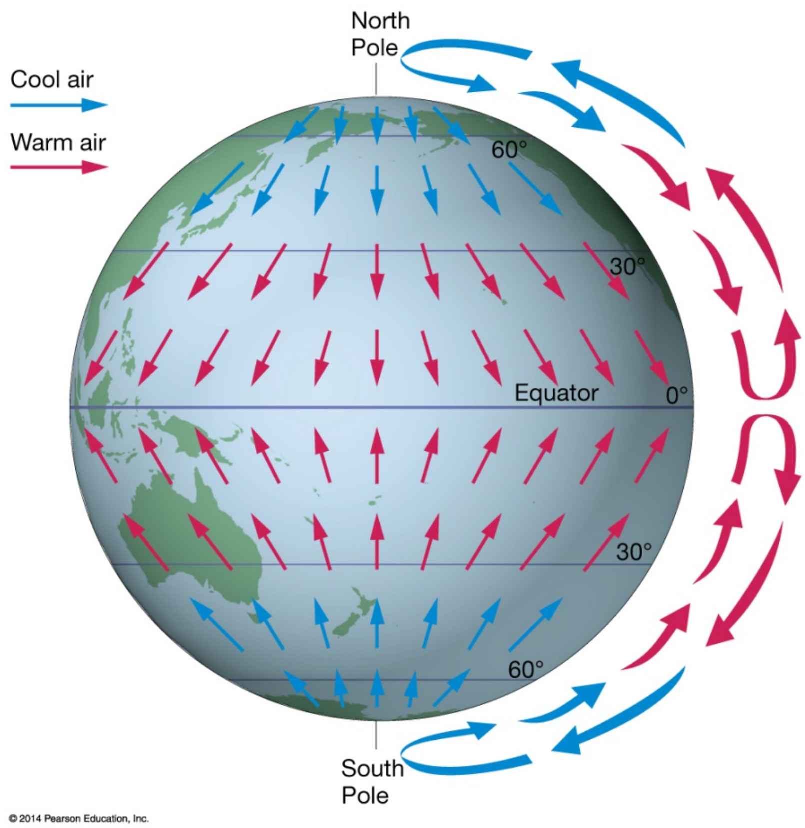


Figure 9.6. Large convection cell created by differential heating of the earth in a non-rotating earth with no land masses.

9.3. Influence of the Coriolis Effect

In reality, the earth turns on its axis in an easterly direction at a speed of 1,674 km/hr at the equator, and lesser speeds at higher latitudes (Figure 9.7). The rotation of the earth affects the movements of the atmosphere, the ocean, and any other object not directly attached to the earth. Consider an object that is directly on the equator moving East at a speed of 1,674 km/hr. If this object is set in motion directly North, it carries an easterly momentum as well as its northerly movement (Figure 9.7). As this object travels over land that has a slower easterly movement than itself, it appears to turn to the right. Similarly, an object set in motion from high latitude (slow rotational velocities) towards the south travels over parts of the earth that have a faster easterly movement, and again appears to veer to the right. In the same way, air or currents in movement in the atmosphere appears to turn to the right in the Northern hemisphere and to the left in the Southern hemisphere. This apparent deflection of objects in movement not directly attached to the earth is called the Coriolis effect. Rotational velocities of the earth decrease in a non-linear fashion from the equator to the poles; the rate of change becomes greater towards the poles. For this reason, the Coriolis effect is increasingly important at higher latitudes. It has no effect directly on the equator.

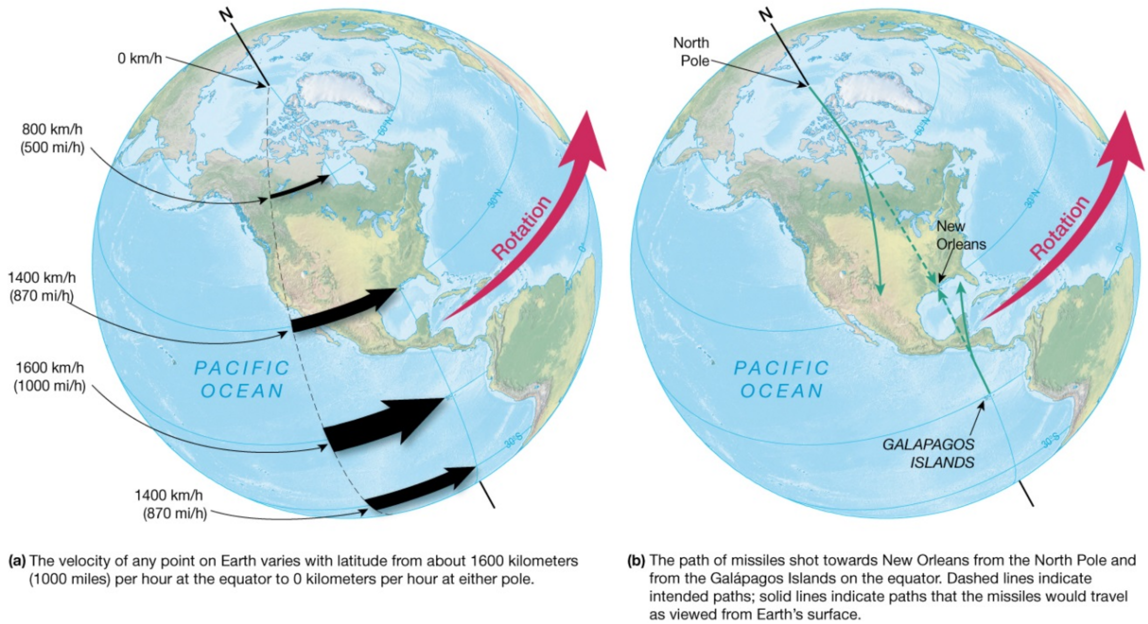


Figure 9.7. The earth spins on its axis in an easterly direction. Velocity is greatest at the equator and zero at the poles (left). The change in velocity with latitude leads to an apparent deflection to the right (Northern hemisphere), called the Coriolis effect (right).

9.4. Global Circulation Patterns

Taking into account the rotation of the earth and the Coriolis effect, the atmospheric circulation model is substantially modified. As in the previous model, warm, moist air rises at the equator and travels at high altitude towards the poles. But as this air moves to higher latitudes, it tends to turn to the right (Northern hemisphere) or left (Southern hemisphere). This shortens the large, hemisphere-wide convection cell, and instead creates three smaller convection cells in each hemisphere (Figure 9.8). The air rising at the equator sinks back down at 30^{o} N and S, and moves along the surface of the earth, either back towards the equator or towards higher latitude. At 60^{o} there is another area of low pressure where air rises, travels at altitude either towards the pole or towards 30^{o} N or S.

In this model, at 0 and 60^{o} North and South, warm, moist, low-density air rises, creating areas of low pressure, clouds and rain. At 30 and 90^{o} North and South, cool, dry, high-density air sinks, creating areas of high pressure, clear skies and low precipitation. The edges of convection cells where air movement is predominantly up or down results in areas of low and inconsistent winds at the surface, i.e. the doldrums (0^{o}) and the horse latitudes (30^{o}). The air moving across the surface of the earth is the resulting winds, which are also affected by Coriolis (and therefore veer to the right in the Northern hemisphere, and to the left in the Southern hemisphere). The resulting trade winds, westerlies and polar easterlies are shown in Figure 6.8.

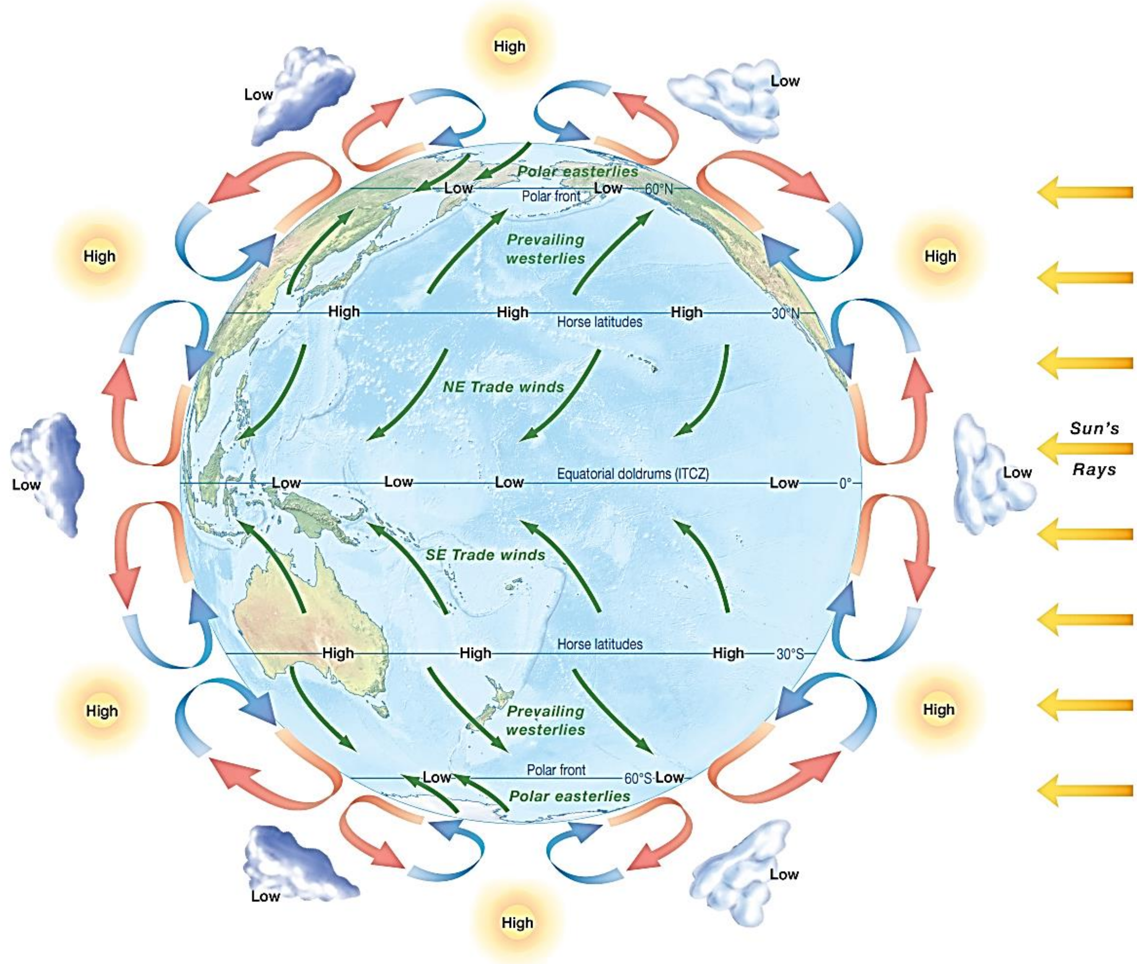


Figure 9.8. Atmospheric circulation resulting in a six-band surface wind system.

The model of atmospheric circulation presented in Figure 9.8 is simplified. It does not consider seasonal differences, how the presence of land masses affects atmospheric circulation. Land, because it has a lower heat capacity than water, absorbs and loses heat more rapidly, and large land masses therefore modify atmospheric circulation. Seventy percent of land masses are in the Northern hemisphere; since there are few land masses in the Southern hemisphere, it follows the model fairly closely. Land, with its low heat capacity, changes temperature rapidly. During the summer, land is warmer than the ocean, causing a low pressure area over land (hot air rises), so there is a continuous low pressure area between 0 and 60° North with no high pressure area at 30° North. However, the high pressure area still forms over the oceans. During the winter, land is cooler than the ocean, causing a high pressure area over land (cold air sinks), there is a continuous zone of high pressure between 30 and 90° North with no low at 60° N. However, the low pressure still forms over the oceans (Figure 9.9).

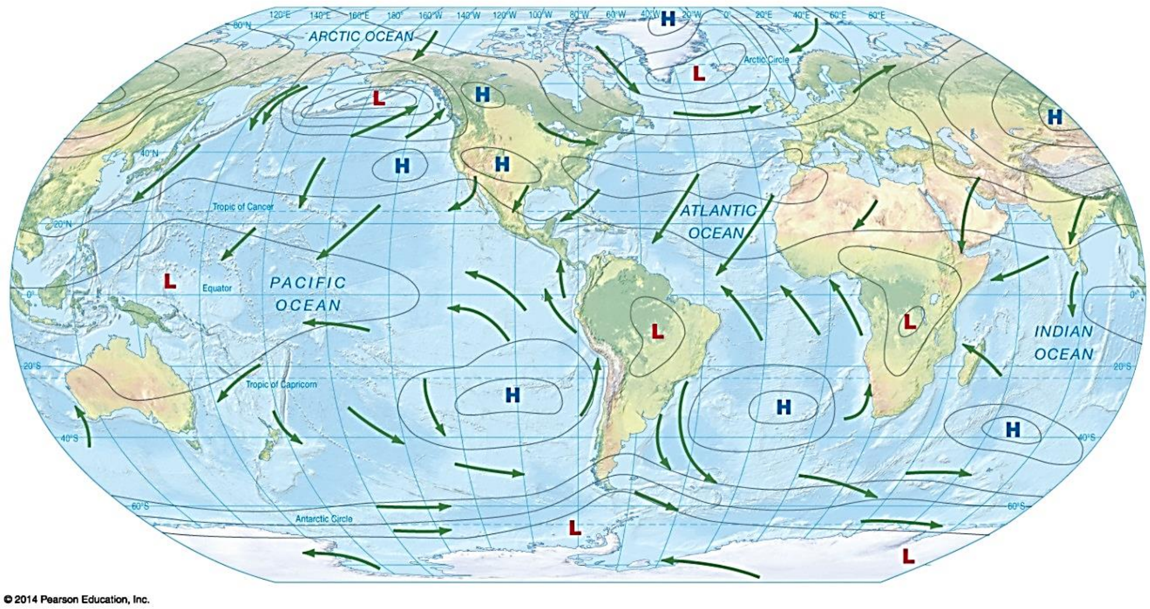


Figure 9.9. Typical atmospheric pressure in January shows the influence of land masses on global circulation patterns.

In the Indian Ocean, dominant wind direction reverses twice a year, from the South West monsoon in the summer to the North East monsoon in the winter. This wind pattern is created because of differences in temperature between the Indian Ocean and the Asian continent. In the summer, Asia warms up faster than the ocean, creating an area of low pressure over land. Winds fill in from high pressure over the ocean to low pressure over land. These summer winds come from the South West; this process is reversed in the winter (Figure 9.10).

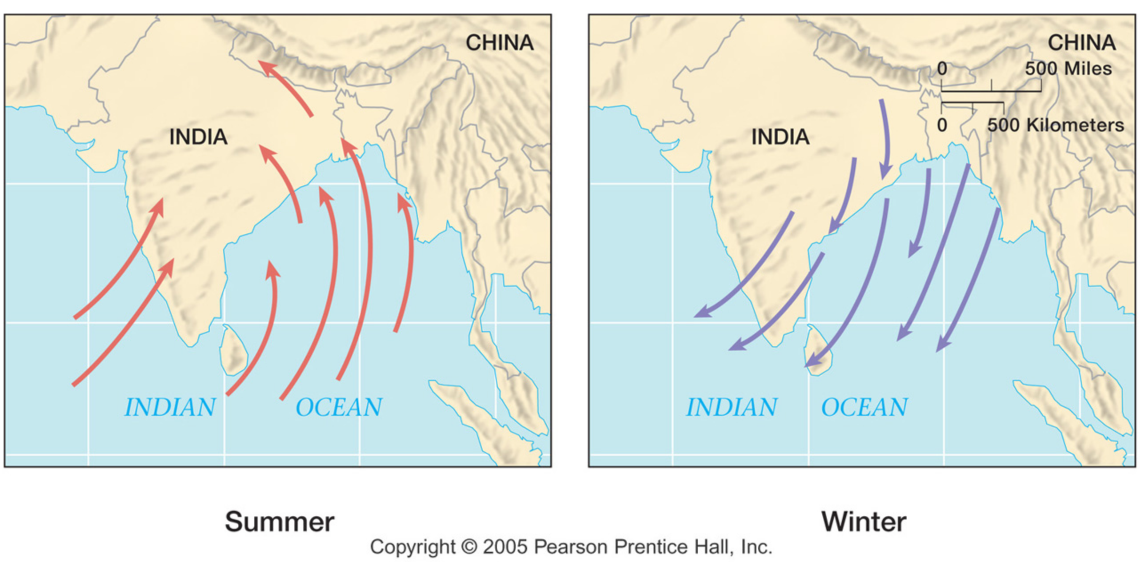


Figure 9.10. Dominant monsoon winds: South West during the summer (left) and North East during the winter (right).

9.5. Weather and Climate Patterns in the Oceans

The ocean has a large impact over the earth weather and climate, because of its size and of the thermal properties of water. Weather is the condition of the atmosphere at a given time and place; climate is the long-term average weather.

Winds

The movement of air from areas of high pressure to areas of low pressure along the surface of the earth is what we call wind. Except at very low latitudes, wind direction is modified as it moves because of the Coriolis effect. Therefore, in the northern hemisphere, wind tends to move in a clockwise direction around high pressure cells and in a counterclockwise (cyclonic) direction around low pressure cells (Figure 9.11).

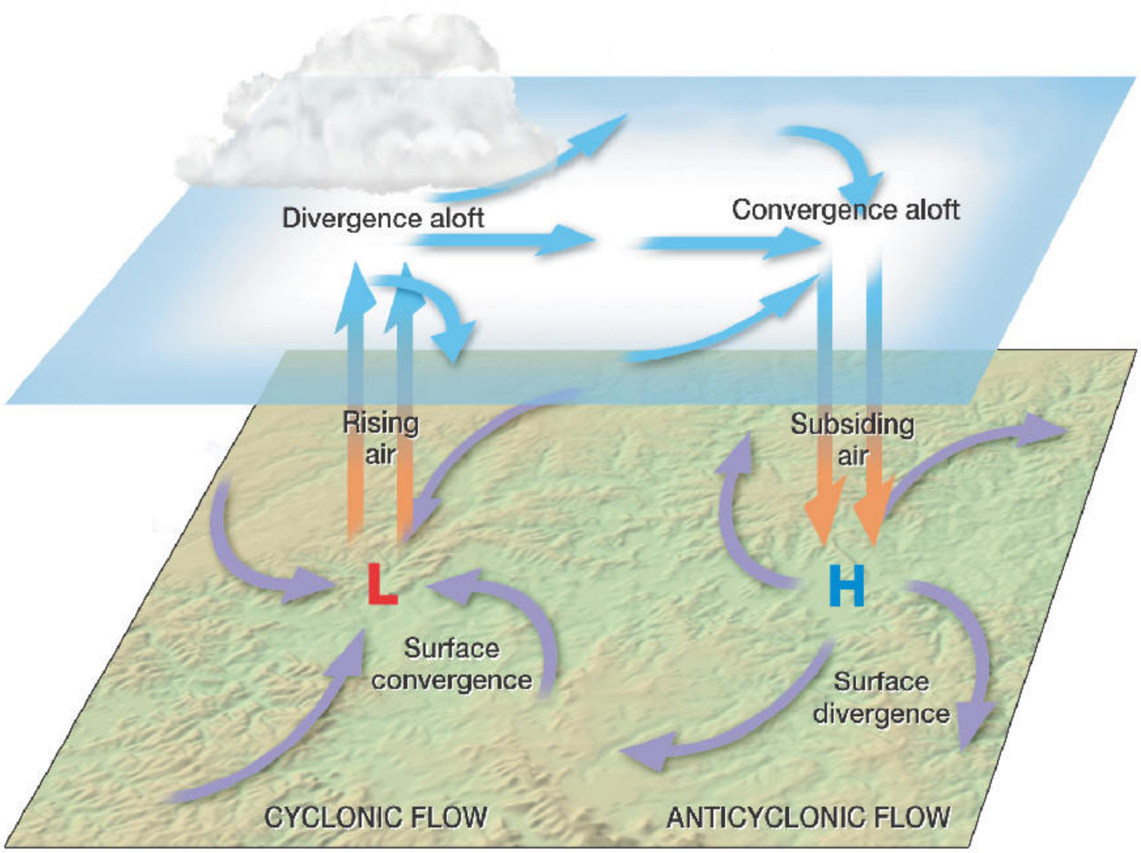


Figure 9.11. Air flow around high and low pressure zones in the northern hemisphere.

Land and sea breezes

The seasonally-shifting winds due seen as monsoons in the Indian Ocean can also happen at a smaller scale, in a daily cycle in some coastal areas. As land warms up faster than water during the day, overlying air rises and is replaced by air from over the ocean, creating an on-shore breeze (or sea breeze, or anabatic wind). At night, the air cools faster over land, sinks, and flows out towards the ocean, creating an off-shore breeze (or land breeze, or katabatic wind; Figure 9.12).

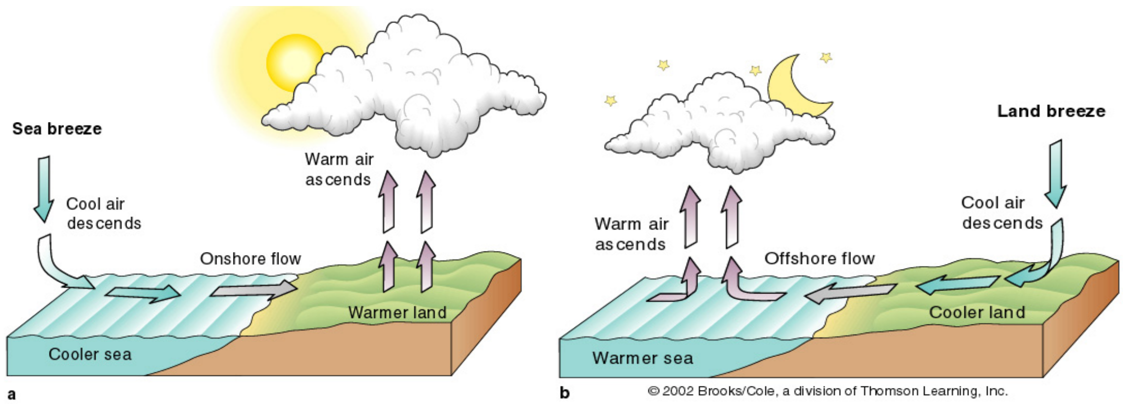


Figure 9.12. Sea breeze and land breeze, caused by the lower heat capacity of land compared to water.

As winds blow over elevated land, air near sea level is forced to rise, and the water vapor condenses as temperature drops, and forms clouds. This typically creates rain on the windward side of islands, which is called orographic rain. The leeward side of islands typically experiences less rainfall, as the precipitation on the windward side depleted the air mass of its moisture (Figure 9.13). Higher islands, which force moist air higher up, therefore experience more rainfall.

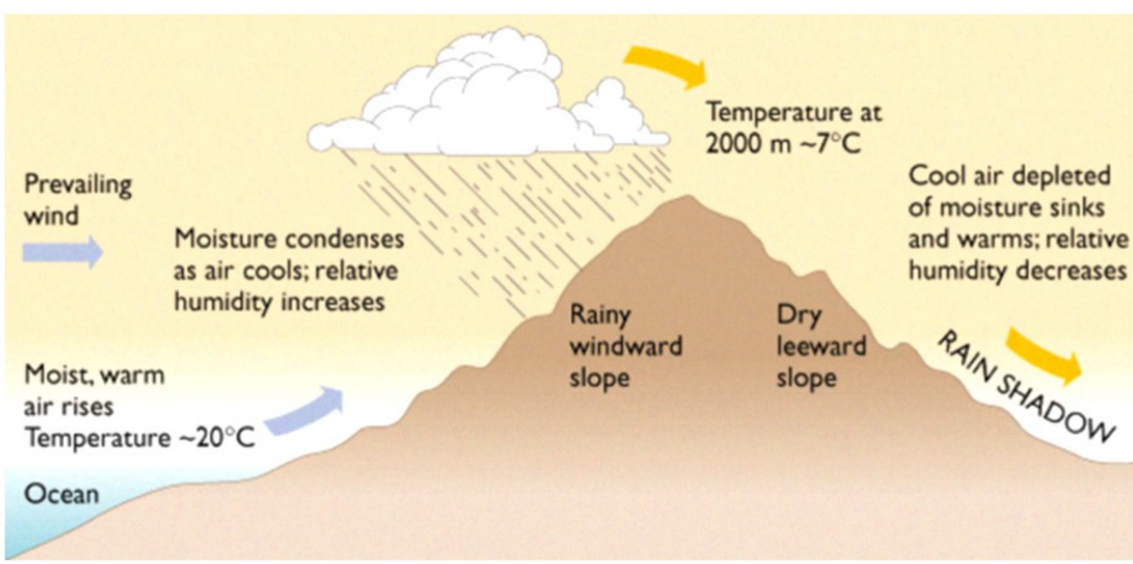


Figure 9.13. Orographic rain falls on the windward side of islands as moist air is forced towards higher altitude.

Hurricanes

Low pressure centers that form over water warmer than 27^{o}C, can form into tropical revolving storms when several conditions are present. These are called hurricanes, cyclones or typhoons depending on the area. Warm sea surface temperatures cause pressure to decrease further and increase evaporation, causing an intense, isolated, low pressure cell. If upper winds are weak and this low pressure cell forms at latitudes higher than 5^{o}, Coriolis affects winds moving towards the low pressure cell and set it in a circular motion. In the northern hemisphere, the winds moving from high pressure areas towards this low pressure center tend to veer to the right, setting the low pressure system in a counterclockwise motion (clockwise in the Southern hemisphere; Figure 9.11). As winds circle this low pressure system, they can grow in strength until the low pressure system becomes a tropical storm (winds 39 to 73 mph) and a hurricane (74 mph). As this storm grows, the winds evaporate more water (and heat), which fuels the storm. Tropical storms and hurricanes dissipate when they travel over cold water or land, and there is no more warm water. Once formed, a revolving storm system gets pushed by the dominant winds. Because there is no Coriolis force at the equator, tropical storms do not occur there (Figure 9.14). In the North Atlantic, hurricanes originally form off the Western coast of Africa and travel towards the Caribbean with the trade winds (Figure 9.15). Hurricane season here is from June to November, when sea surface temperatures are high enough to sustain the storms.

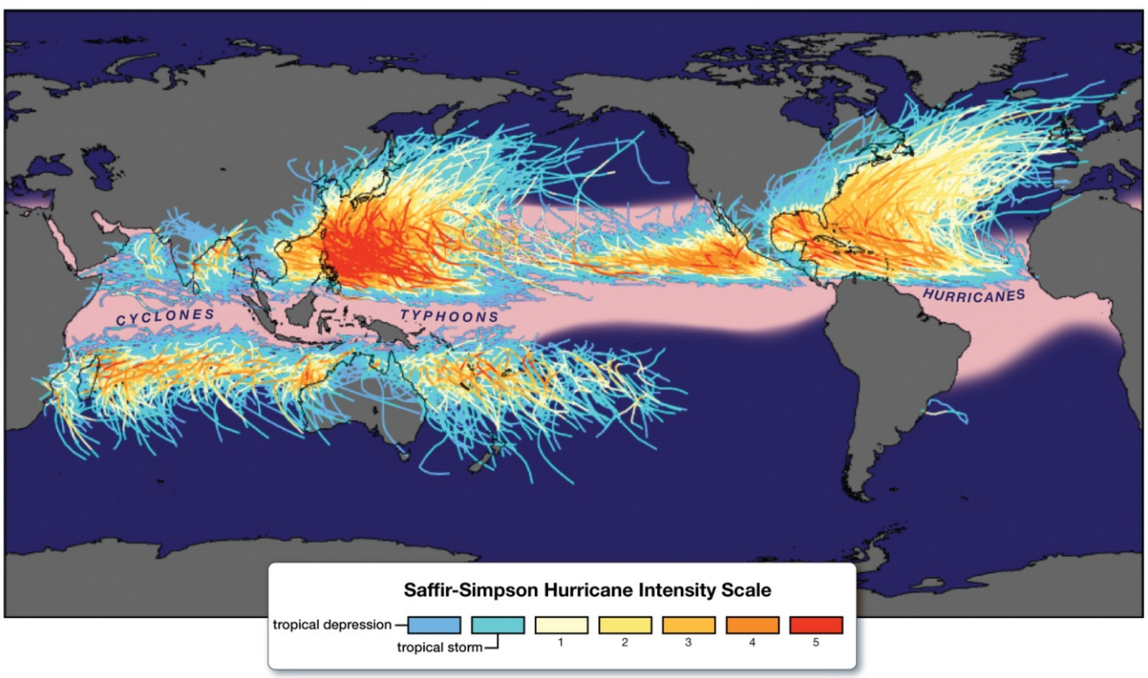


Figure 9.14. Historic paths of tropical storms

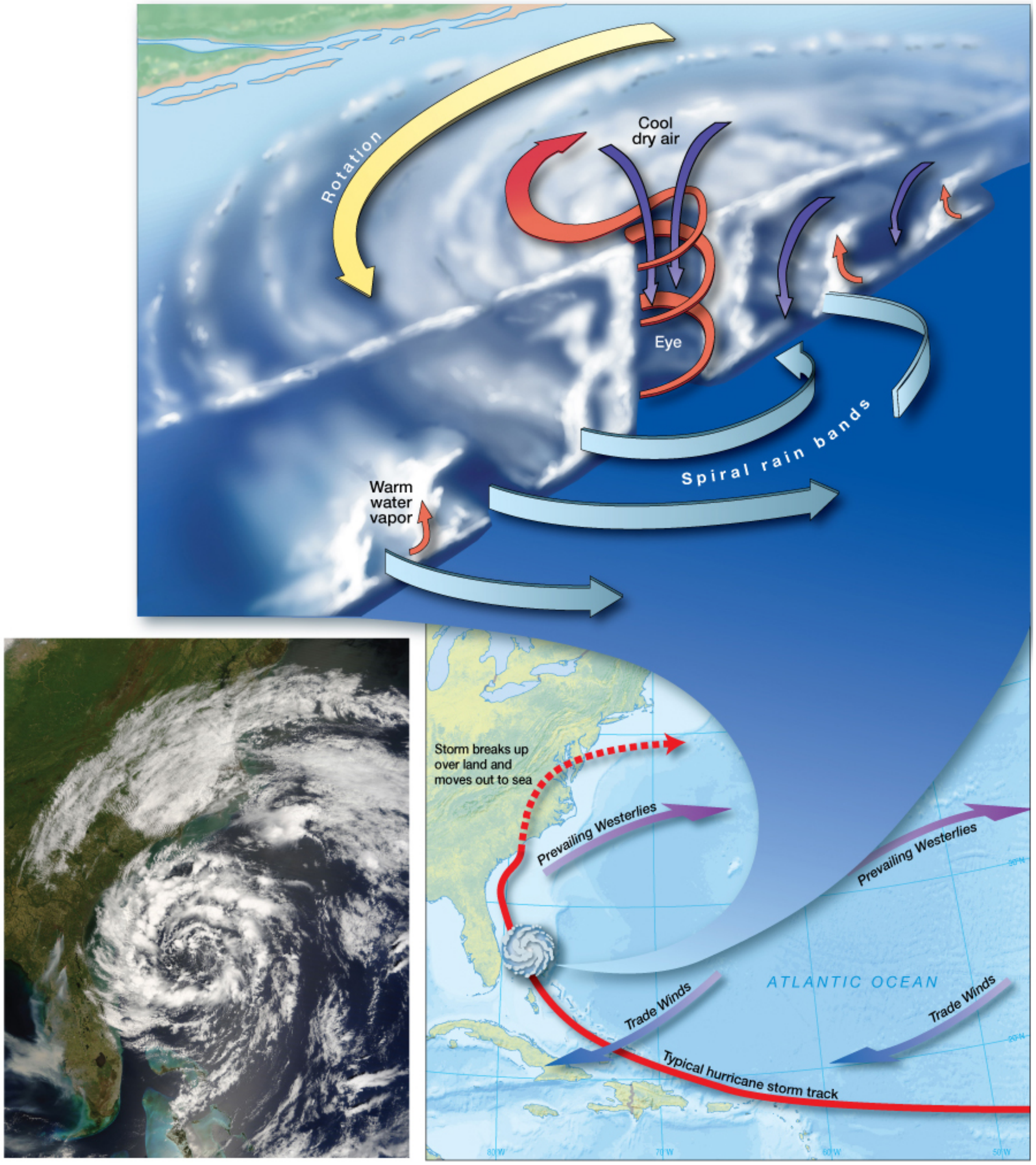


Figure 9.15. Structure and typical path of a hurricane.

Besides the clear danger to humans from extremely high sustained winds, hurricanes can also significantly damage coastal developments because of flooding caused by the storm surge. The storm surge is an area of elevated sea level created because of very low atmospheric pressure in intense storms. This dome of water can be 50-100 miles wide. Greatest damage occurs if the storm surge coincides with high tide (Figure 6.16).

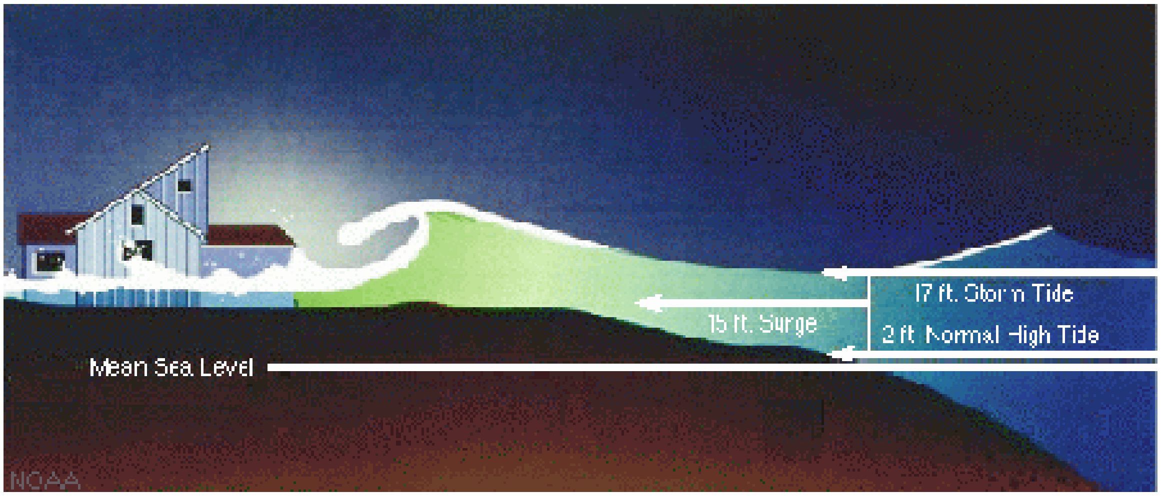


Figure 9.16. Storm surge combined with high tide leads to greatest coastal flooding.

9.6. Review Questions

1. Name and describe the major layers of the atmosphere

2. Which layer of the atmosphere is most important to consider for weather conditions and atmospheric circulation?

3. What gases is the atmosphere composed of? In what proportion?

4. What is the typical temperature change for each 1000m of altitude in the troposphere?

5. What gas present in the stratosphere is responsible for the increase in temperature in the stratosphere?

6. Where is solar radiation most intense in the globe? Name 2 reasons why.

7. The formation of convection cells in the atmosphere result in areas of low pressure at what latitudes? Areas of high pressure at what latitudes?

8. Why are the doldrums and the horse latitudes areas with little wind?

9. Name the dominant winds present between 0 and 30^o; 30 and 60^o; 60^o and the poles.

10. What is the name of the apparent deflection of objects not directly attached to the earth, which is caused by the rotation of the earth?

11. What is the direction of this deflection in the southern hemisphere?

12. Which hemisphere follows the theoretical model of convection cells more closely? Why?

13. Explain how land modifies winds on a daily basis.

14. What is anabatic wind? What is katabatic wind? When would you expect to see each?

15. Why do the dominant winds change twice a year in the Indian Ocean?

16. What is orographic rain?

17. Why do hurricanes dissipate when they go over land or cold water?

18. Which direction to hurricanes spin in the northern hemisphere and why?

10. Ocean Circulation (Trujillo, Chapter 7)

Major ocean currents are stable and predictable; they have been described as rivers without banks. There are two main types of ocean circulation: thermohaline, or density driven circulation, which affects circulation at depth, and wind-driven, which is most important in setting surface currents. This constant flow of ocean currents is caused by average weather conditions (e.g. trade winds, colder water at higher latitude). Because the density of water is 1,000 times greater than air, it has more momentum and once in motion it easily keeps flowing.

10.1. Surface currents

As winds blow over the oceans in the circulation patterns described in section 6.4, they set the surface of the water in motion. Currents generally follow wind patterns and are also affected by continents. However water, as it is not attached to the earth, is affected by Coriolis, and surface water is deflected at 45° from the direction of the wind (to the right in the Northern hemisphere, left in the Southern hemisphere). This surface layer then sets the underlying layer in motion through frictional drag, and this second layer is deflected even more. As depth increases, each subsequent layer of water is driven by the movement of the water above and moves at an increasing angle to the wind direction, while current speed decreases because of friction. This is known as the Ekman spiral (Figure 10.1). The Ekman spiral extends to a depth of approximately 150m; wind has no direct effect on deeper waters. The average direction of flow over the entire depth of the spiral (150m), or net transport of water, is approximately 90° from the direction of the wind (Figure 10.1).

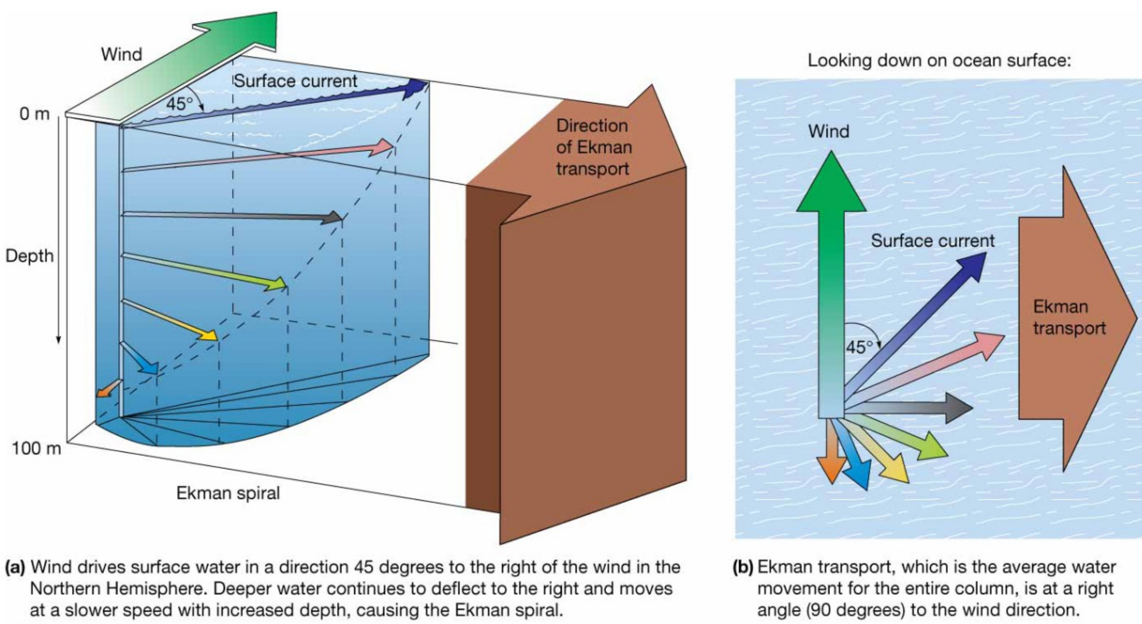


Figure 10.1. Flow of water in successive layers in the Ekman spiral. The overall transport of the top 150m is 90° to the wind direction. This is the Ekman transport.

The average atmospheric circulation patterns described in section 9.4 act to form gyres, large ocean basin circular motion current systems (Figure 10.2). As winds blow over the surface of the ocean, they set the surface water in motion at an angle of 45^o to the wind. For example in the North Atlantic, north east Trade Winds create the north equatorial current that flows west, and the south west Westerlies create the North Atlantic current that flows east. The Gulf Stream and Canary currents are created because of the continuity of flow between the equatorial and North Atlantic currents. Gyres move clockwise in the Northern hemisphere and counterclockwise in the Southern hemisphere.

Figure 10.2. Major surface currents in the North Atlantic Ocean are set in motion by dominant winds.

Once ocean gyres are in motion, the Coriolis deflection and Ekman transport drive water to the center of the gyres, creating a mound of water in the middle of the gyre (Figure 10.3). Gravity tends to push this water down and away from the center, and as this water is pushed away from the center it is again affected by Coriolis. This water movement contributes to the general flow around the gyre. When these forces (Ekman transport and gravity) are balanced, water flows smoothly around the gyre, and this is called geostrophic flow. Ideal geostrophic flow would continue perfectly around the gyre, but in reality, friction makes the current run somewhat downslope. The mound of water caused by geostrophic flow can be as high as 2m, and is a bit to the west of center because of the rotation of the Earth which pushes water to the west side of ocean basins.

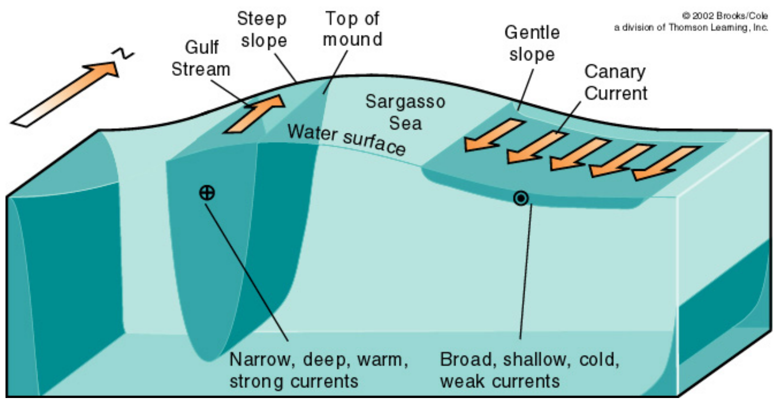


Figure 10.3. Formation of a mound in the middle of the North Atlantic, which leads to geostrophic flow.

Many major ocean currents are part of the subtropical and subpolar gyres formed by dominant winds (Figure 10.4). Currents flowing along the Western side of ocean basins (e.g. Gulf Stream) are typically stronger, deeper, and narrower due to the rotation of the Earth, which piles water along the eastern edge of land masses; the increase in Coriolis effect with latitude, and the changing strength and direction of East-West wind fields (tradewinds/westerlies) with latitude. Currents flowing along the eastern side of ocean basins are weaker, wider, and slower. This phenomenon is known as the western intensification of currents. Near the equator, the equatorial currents that are part of the North and South subtropical gyres flow west. The Ekman transport from these currents causes divergence of water right the equator, which forces both upwelling (discussed in the next section) and a narrow, easterly-flowing equatorial counter-current.

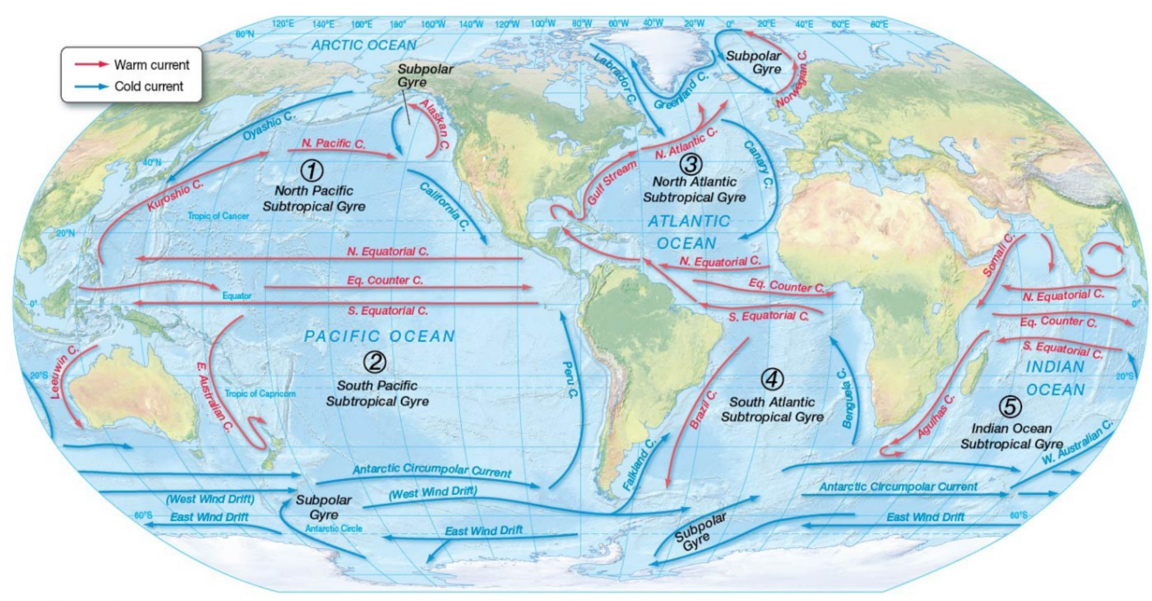


Figure 10.4. Major ocean currents of the world. Notice the clockwise flow in the Northern hemisphere and the counterclockwise flow in the Southern hemisphere.

10.2. Upwelling and Downwelling

Upwelling and downwelling refer to vertical movements of water. Upwelling refers to the movement of deep water towards the surface. This often occurs through divergence, when Ekman transport caused by winds cause surface waters to move away from each other (e.g. at the equator) or away from landmasses (e.g. on the West Coast of South America). This forces cold, nutrient-rich water from depth moves up to replace the surface water (Figure 7.5). The high levels of nutrients brought to the photic zone results in areas of high primary productivity. Downwelling occurs when wind-driven surface currents collide or are forced against landmasses (convergence) and surface water is force to sink down, bringing oxygen-rich water to depth. Upwelling and downwelling can also occur because of thermohaline circulation, e.g. downwelling occurs as cold, salty and dense Antarctic surface water sinks down (section 10.4), or because of geological features that force certain water movements.

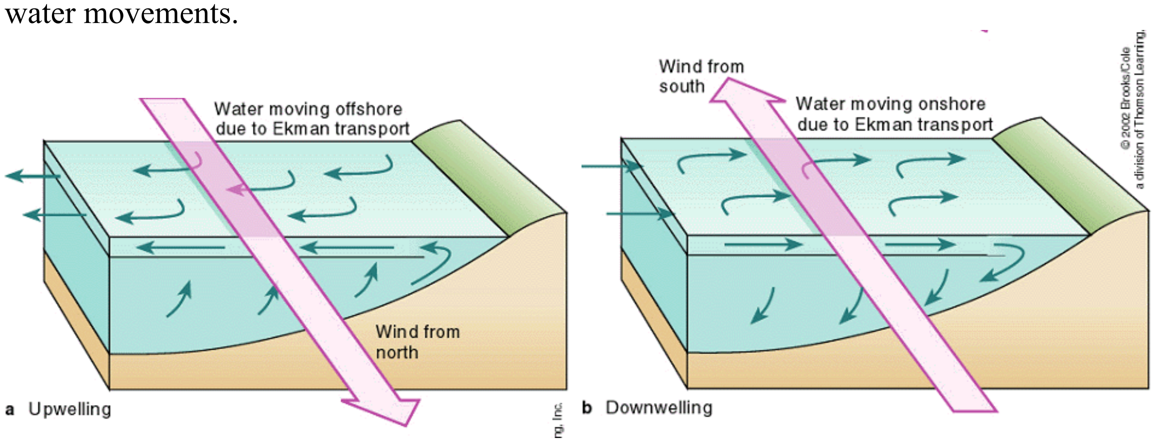


Figure 10.5. Wind-driven upwelling and downwelling in the Northern hemisphere.

10.3. Main surface circulation patterns

The main pattern of ocean circulation varies between ocean basins depending on the shape of the basin, its geographic location in relation to major wind belts, and seasonal changes in winds.

Antarctic Circulation

Antarctic surface circulation is dominated by the easterly-flowing Antarctic circumpolar current (Figure 10.6), set in motion by the strong westerly winds found from 40-60^{o} South (Figures 9.8 & 9.9). Between the Antarctic circumpolar current and the continent, the east wind drift is a surface current propelled by the polar easterlies, which flows from east to west. Between those two opposite surface currents there is a zone of divergence and upwelling, caused by the Ekman transport associated with each current. The edge of the Southern Ocean is marked by the Antarctic Convergence, a zone of downwelling north of the Antarctic Circumpolar Current where cold, dense Antarctic waters sink below warmer, less-dense sub-Antarctic waters.

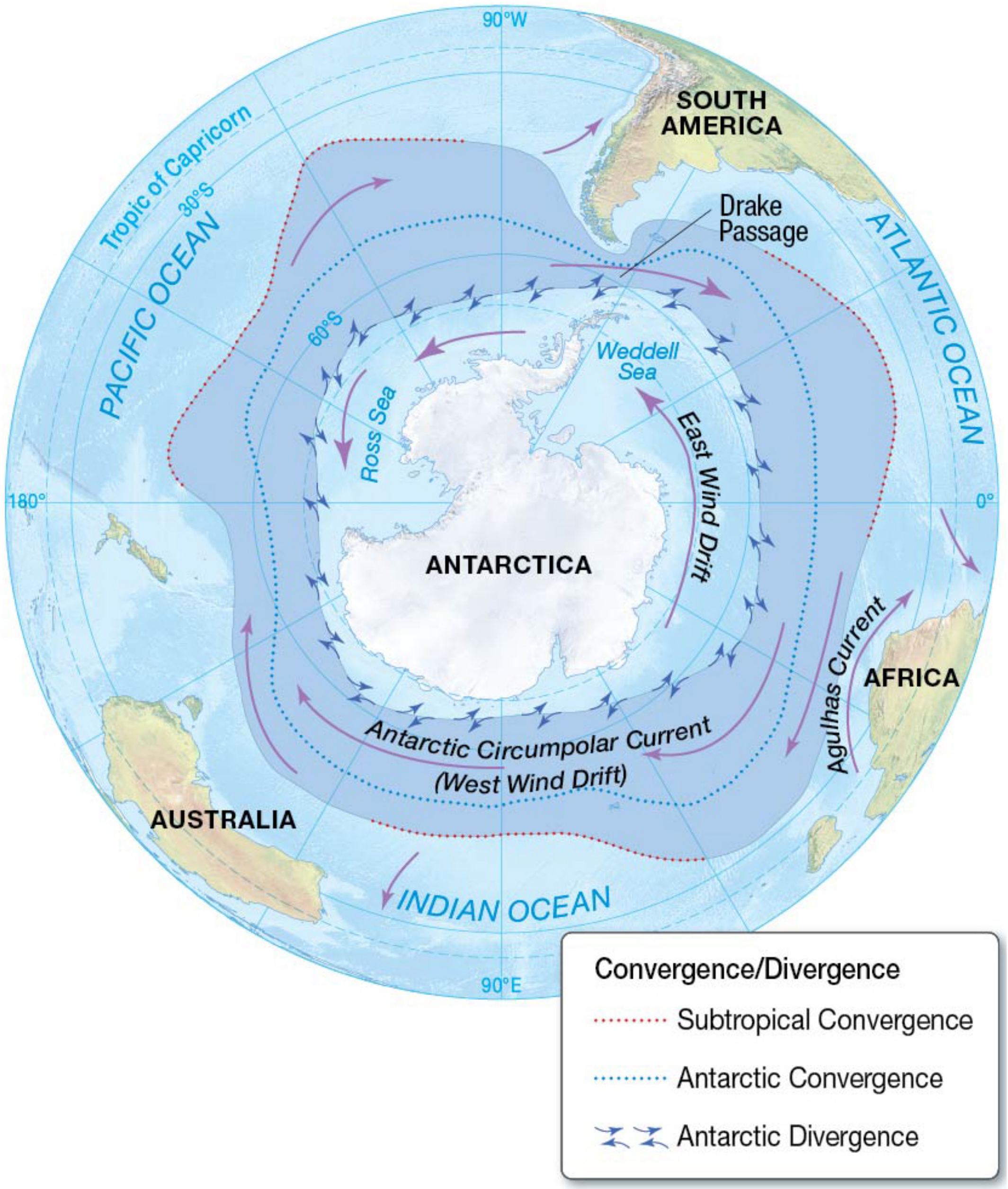


Figure 10.6. Antarctic surface circulation, dominated by the easterly flowing Antarctic circumpolar current.

Atlantic Ocean Circulation

The North Atlantic Subtropical gyre and the South Atlantic Subtropical gyre dominate the surface circulation patter in the Atlantic Ocean (Figure 10.7). Individual currents like the Gulf Stream and the Canary Current are parts of these gyres. The North gyre rotates clockwise while the South gyre rotates counterclockwise. An equatorial countercurrent flows easterly, in the opposite direction to the South and North equatorial currents on the east side of the basin. Interestingly, on the west side of the Atlantic, the South equatorial current meets the North equatorial current this forms the Antilles and Caribbean currents. The Caribbean current flows through the Caribbean Sea and into the Gulf of Mexico, forming the loop current before joining the Florida current and eventually the Gulf Stream (Figures 10.8 & 10.10). The Gulf Stream is an especially narrow and fast current because of western intensification (Figure 10.8)

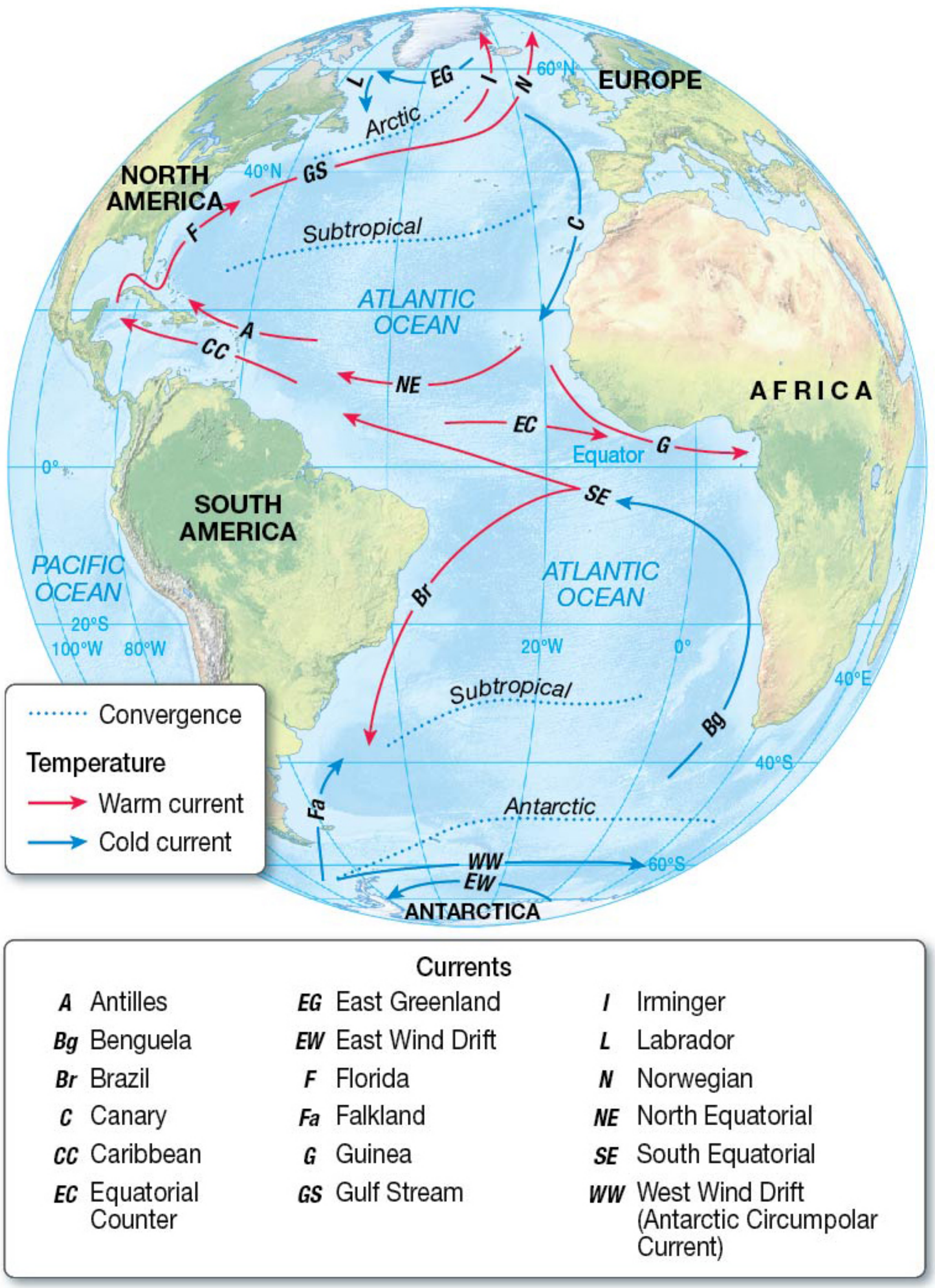


Figure 10.7. Dominant surface currents of the Atlantic Ocean.

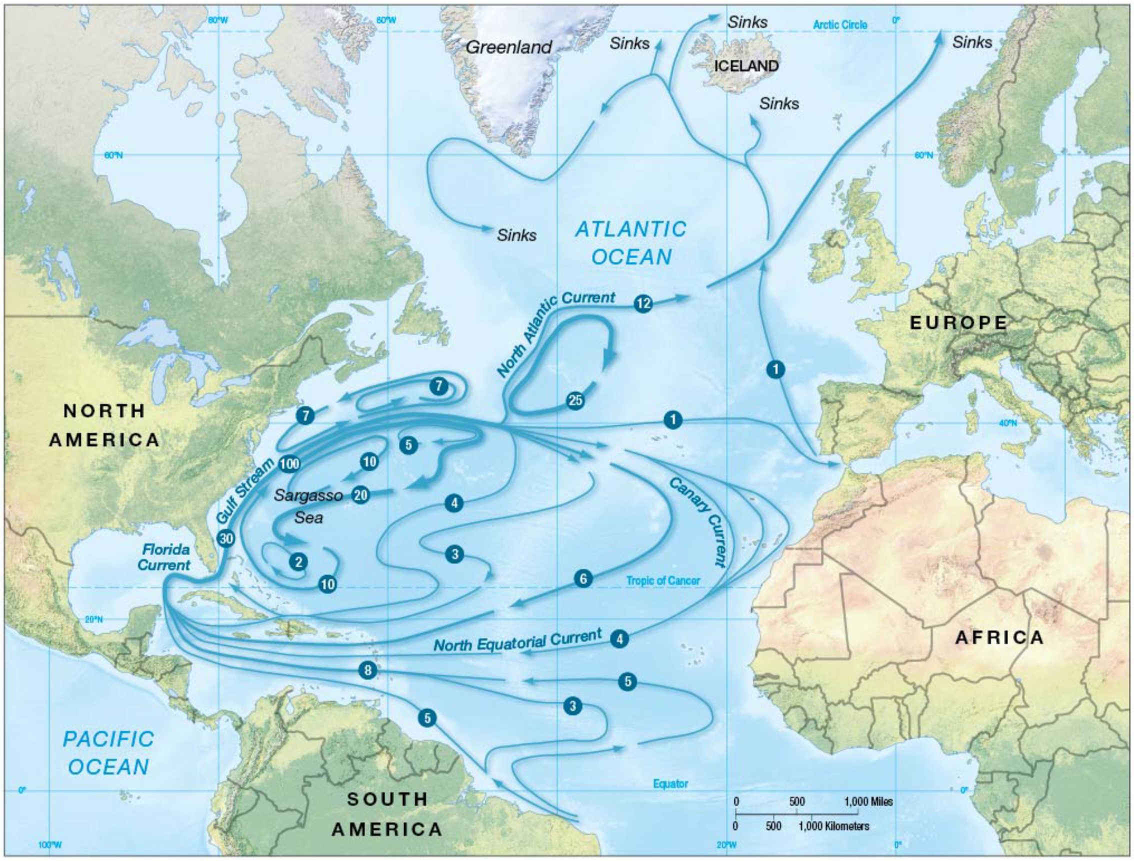
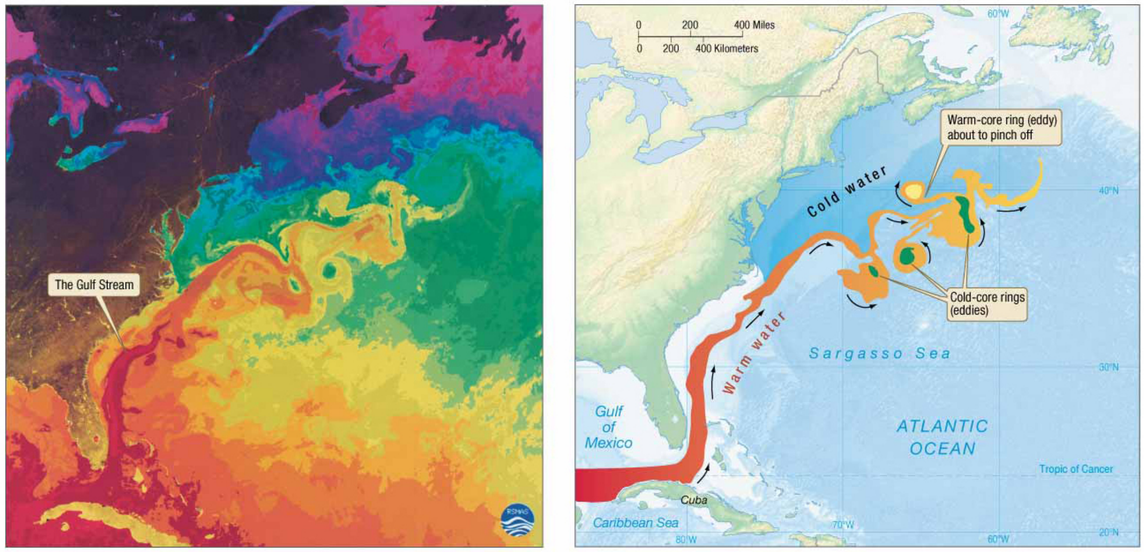


Figure 10.8. Circulation in the North Atlantic, showing average flow rates in Sverdrups (1 Sverdrup = 1 million cubic meters per second). Notice the difference in velocity and width between the Gulf Stream and the Canary Current.

The Sargasso Sea is found on the west side of the North Atlantic and is an area of slow moving water in the center of the North Atlantic Subtropical gyre, where floating algae called *Sargassum* accumulate.

As a narrow, fast-moving current moves through slower-moving water, the current oscillates and develops meanders along its boundary. Some of these meanders break off to form eddies, pockets of water moving with a circular motion. Clockwise rotating warm eddies or rings commonly form along the western edge of the Gulf Stream, creating isolated areas of low productivity in the cold waters north of the Gulf Stream (Figure 10.9) Similarly, cold water eddies moving in a counterclockwise direction pinch off and move into the warm Sargasso Sea as isolated areas of high productivity. Similar eddies are created by the loop current in the Gulf of Mexico (Figure 10.10). These eddies may play an important role in strengthening or weakening tropical storms that move over them.



(a) The Gulf Stream is shown flowing along the U.S. East Coast in this NOAA satellite false-color image of sea surface temperature (warm waters = red and orange).

(b) Matching map of the same area as part a.

Figure 10.9. Eddies pinching off the meandering Gulf Stream.

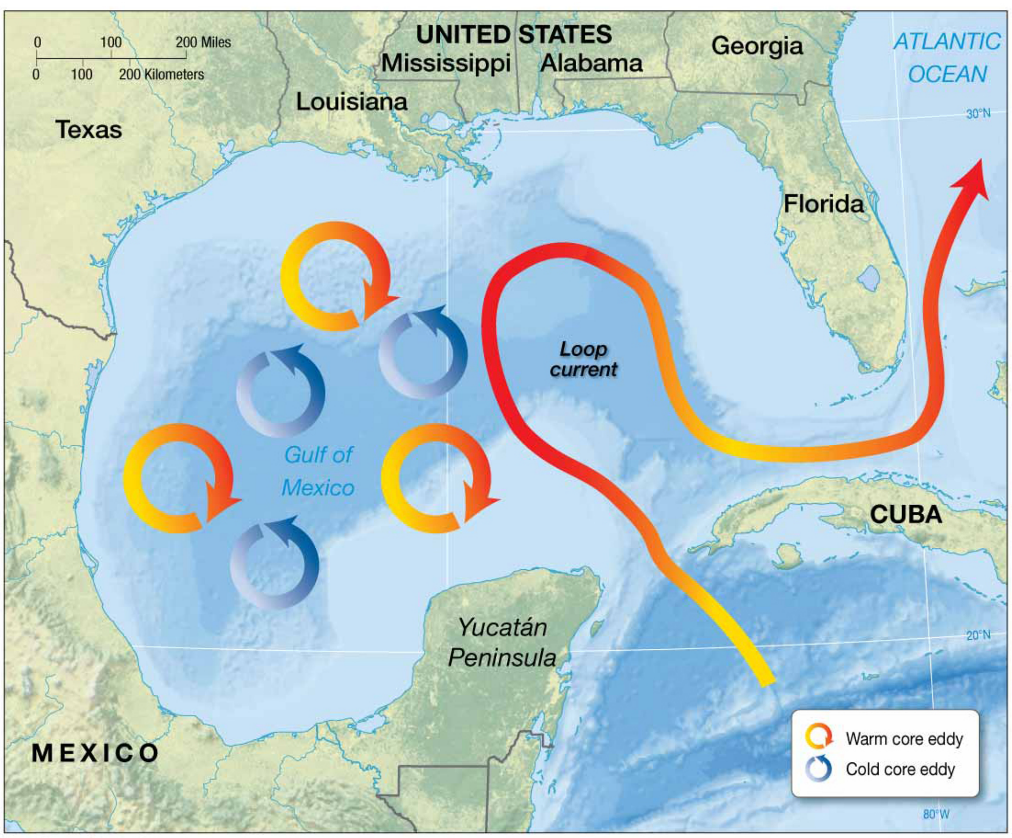


Figure 10.10. Eddies formed by the Loop Current in the Gulf of Mexico.

Indian Ocean Circulation

The monsoon winds of the Indian Ocean change seasonally (see section 6.4), which causes seasonal changes in surface current patterns (Figure 10.11). During the Northeast monsoon of the winter, the Northeast winds monsoon winds create the North Equatorial current to flow west and its extension the Somali current to flow south along the east coast of Africa. An equatorial countercurrent is established. When monsoon winds are reversed during the Southwest monsoon in the summer, the North Equatorial Current disappears and is replaced by the easterly-flowing Southwest Monsoon Current. The Somali Current also reverses. In the southern part of the Indian Ocean, a typical subtropical gyre is established.

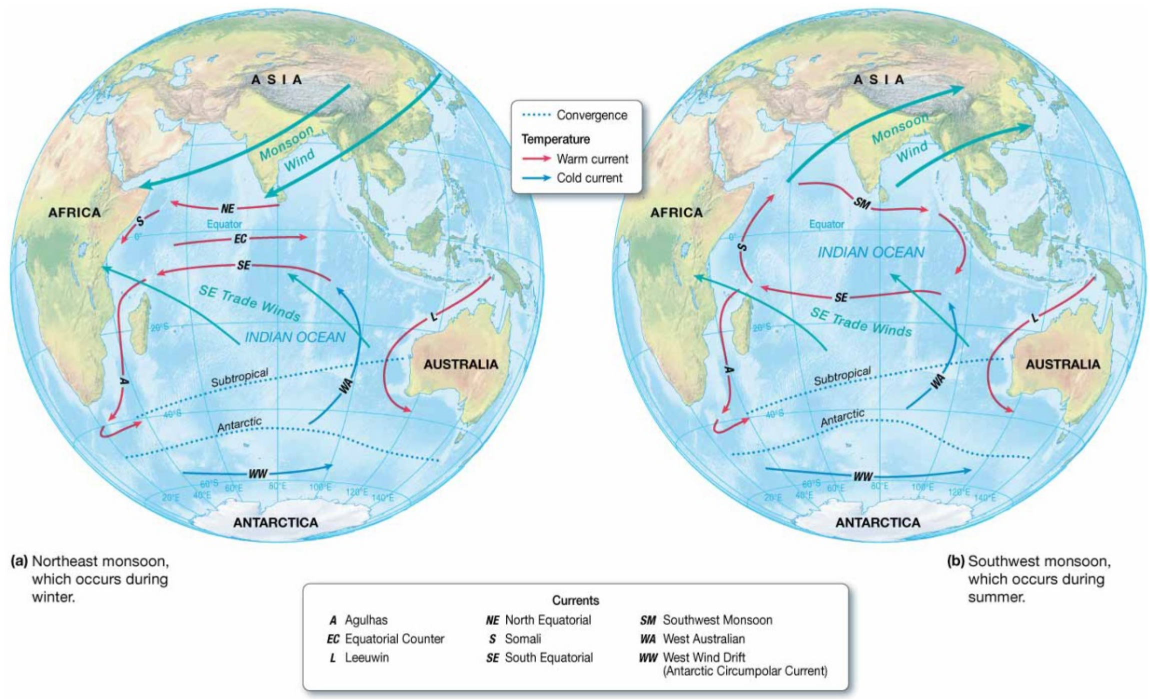


Figure 10.11. Surface circulation patterns in the Indian Ocean during the Northeast and the Southwest monsoons.

Pacific Ocean Circulation

Two large subtropical gyres dominate the circulation of the Pacific Ocean (Figure 10.12). Here, the Equatorial Countercurrent is well-developed and stronger than in the Atlantic. One of the most interesting features of the circulation patterns in the Pacific is the large-scale changes that occur every few years in associate with atmospheric and oceanographic disturbances, in particular in the South Pacific.

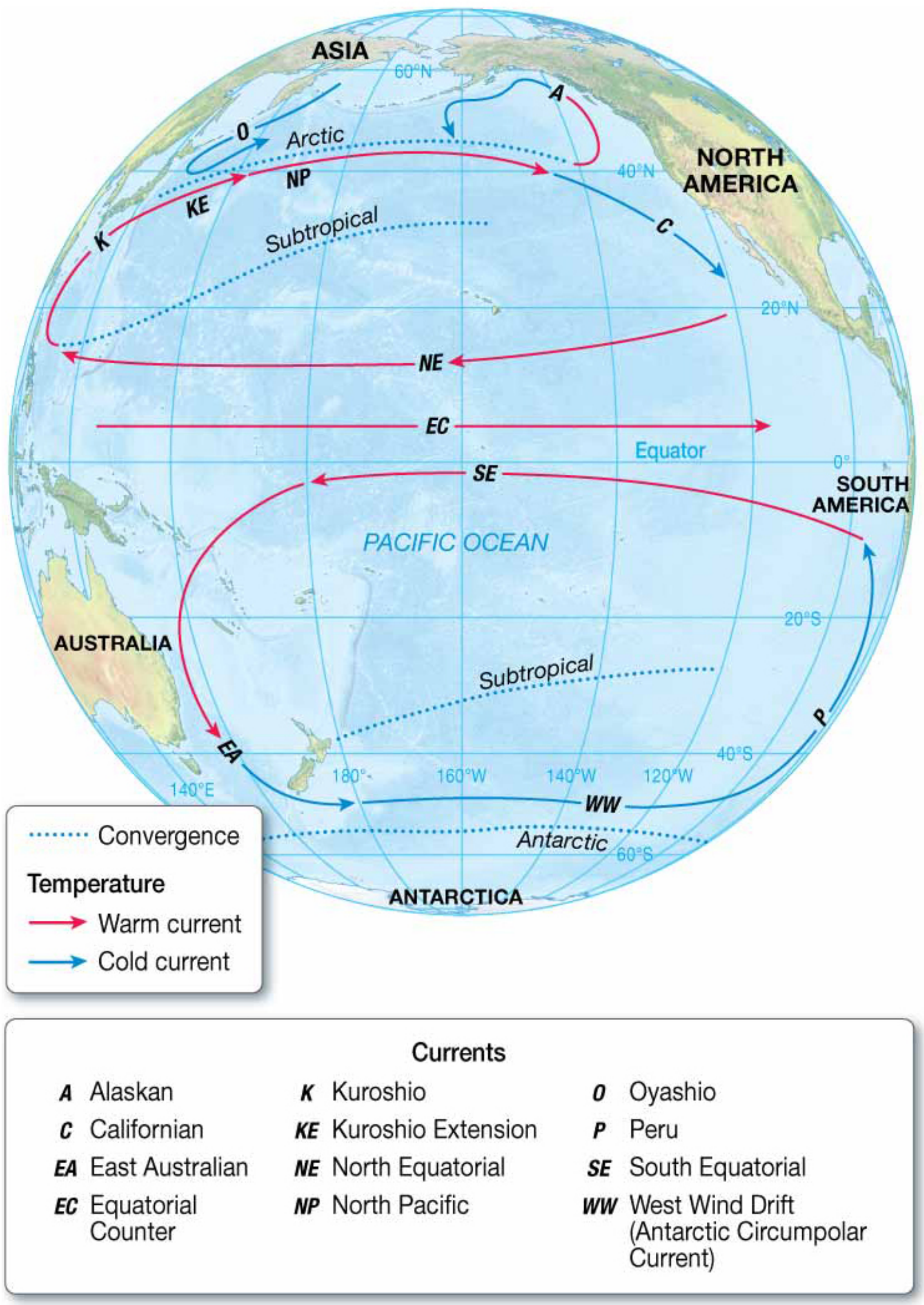


Figure 10.12. Surface currents of the Pacific Ocean.

The cold waters off the west coast of South America are one of the most productive fishing grounds in the world. The reason for this high productivity is the coastal upwelling that occurs in this region because of divergence of surface water and the continent caused by the dominant winds. This upwelling is strong during "normal" years (Figure 10.13a). Under "normal" conditions, an area of low pressure is established in the South West Pacific, and a zone of high pressure in the South East. This leads to the creation of a convection cell called the Walker Circulation Cell, which include strong southeast trade winds (Figure 10.13a).

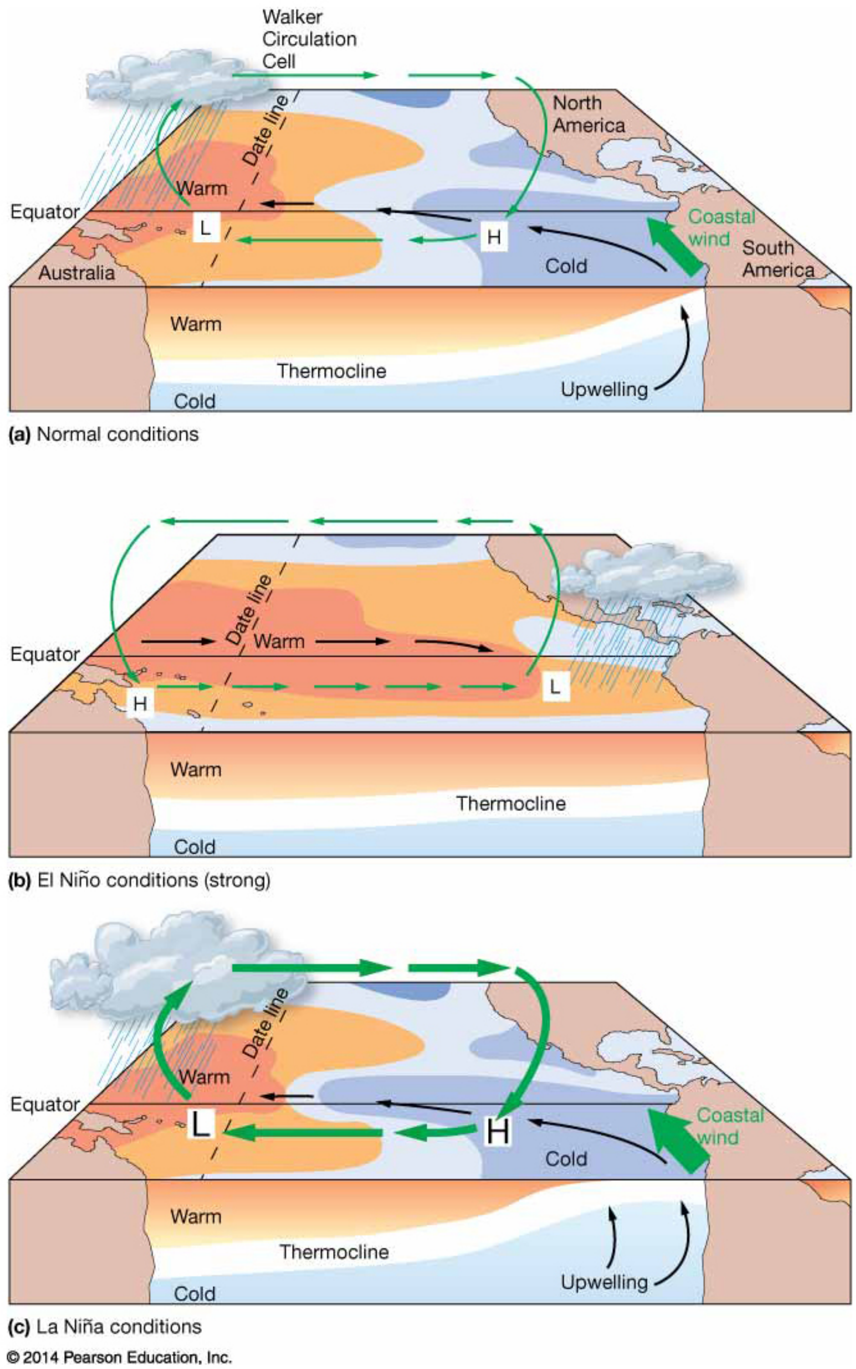


Figure 10.13. Atmospheric and oceanic circulation in the South Pacific during normal, El Niño and La Niña years.

This pattern of atmospheric and oceanic circulation periodically changes, in what is now called El Niño-Southern Oscillation (ENSO). In the ENSO warm phase (El Niño), the high pressure on the east side of the South Pacific weakens, which causes the trade winds to also weaken. This stops the upwelling off the coast of Peru and brings much warmer water than usual to this area (Figure 10.13b). The warm water is devastating to many coral reefs in the Pacific which cannot tolerate the elevated temperatures, and it dramatically reduces the productivity and fisheries yield off the coast of Peru. Further, because the Pacific is such a large ocean, large-scale circulation here have repercussions throughout the world (Figure 10.14). During the ENSO cool phase (La Niña), conditions are similar to "normal" yet intensified, with especially strong trade winds and upwelling (Figure 10.13c). The pattern of reversal between ENSO phases is irregular (Figure 10.15), but long-term data show that there is an increase in the frequency of warm phases caused by climate change.

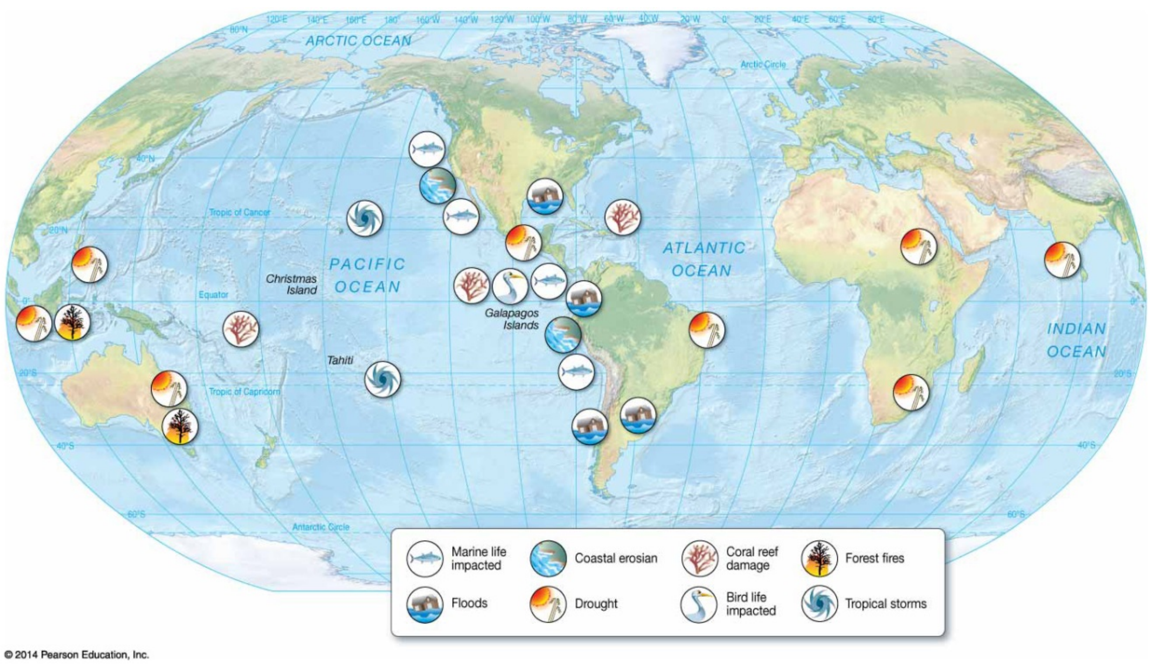


Figure 10.14. Some of the global impacts of ENSO warm phases.

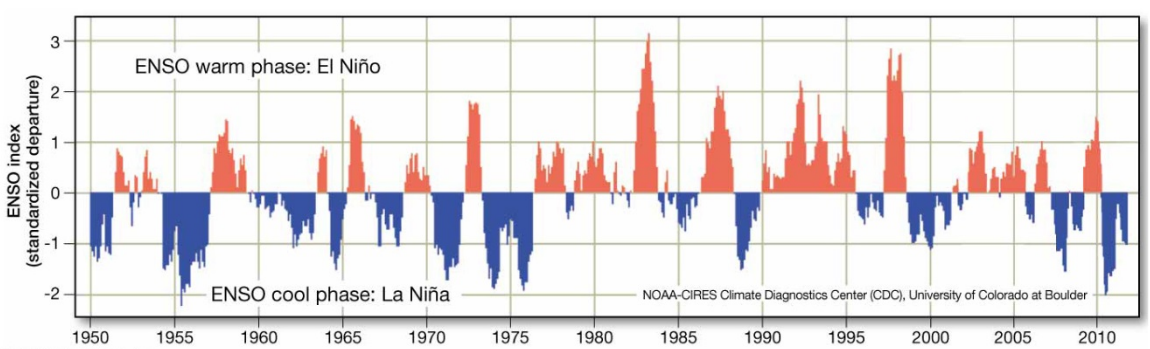


Figure 10.15. ENSO index from 1950 to 2013, showing an irregular patterns of changes between warm (positive) and cool (negative) phases.

10.4. Deep ocean currents

Deep currents that occur in the zone below the pycnocline affect about 90% of the ocean's water. Differences in density create these deep currents, which flow much slower than surface currents. Temperature and salinity are the most important factors affecting the density of surface water, and deep circulation is often referred to as thermohaline circulation. Salinity is affected by evaporation and precipitation, by the formation and melting of sea ice, and by the inflow of river water. Temperature is affected by global differences in solar radiation; it decreases with increasing latitude. Generally, the equator is warm with a relatively low salinity because of high precipitation (Figure 10.16). At 30º North and South, water is still warm and salinity increases due to areas of high pressure and low precipitation. Around 50-60º North and South, water is colder and salinity is lower, corresponding to areas of low pressure, high precipitation and large rivers. Polar waters are cold. They experience low precipitation but their salinity varies with season and the melting or formation of sea ice.

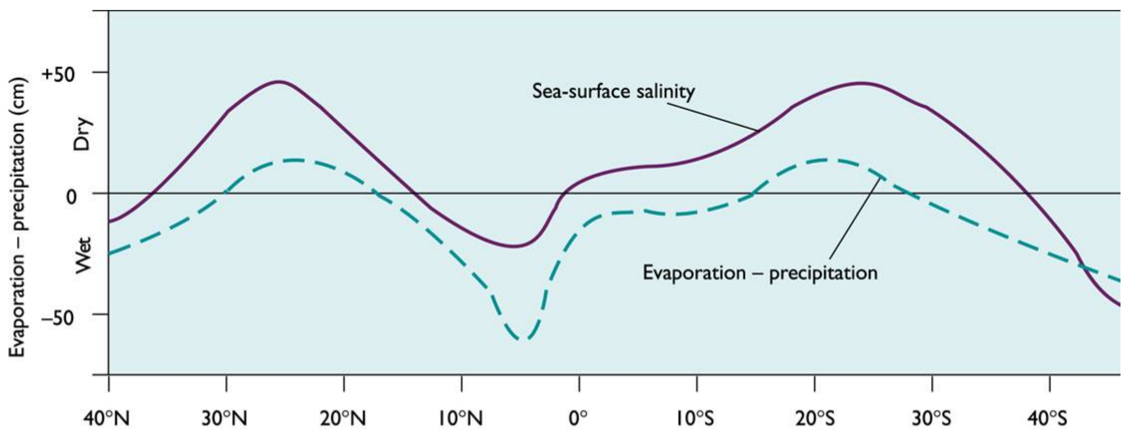


Figure 10.16. Latitudinal variations in salinity and "dryness".

Salinity also changes with distance from land and freshwater river input; the center of ocean basins tend to have a higher salinity than coastal areas. Minute changes in density (caused by changes in temperature and salinity) can cause large changes in vertical circulation; for this reason, oceanographers measure density to 5 decimal places. A body of water of a given density will sink until it reaches water of higher density, then it spreads horizontally. Water masses will not readily mix with water masses of a different density, and can retain their properties for extended periods of time. However, two bodies of water of the same density can readily mix. Because the density curve from various temperatures and salinities is not linear, two bodies of water of the same density but of different temperature and salinity can mix to produce a resulting body of water of increased density (Figure 10.17). This process is known as the caballing effect.

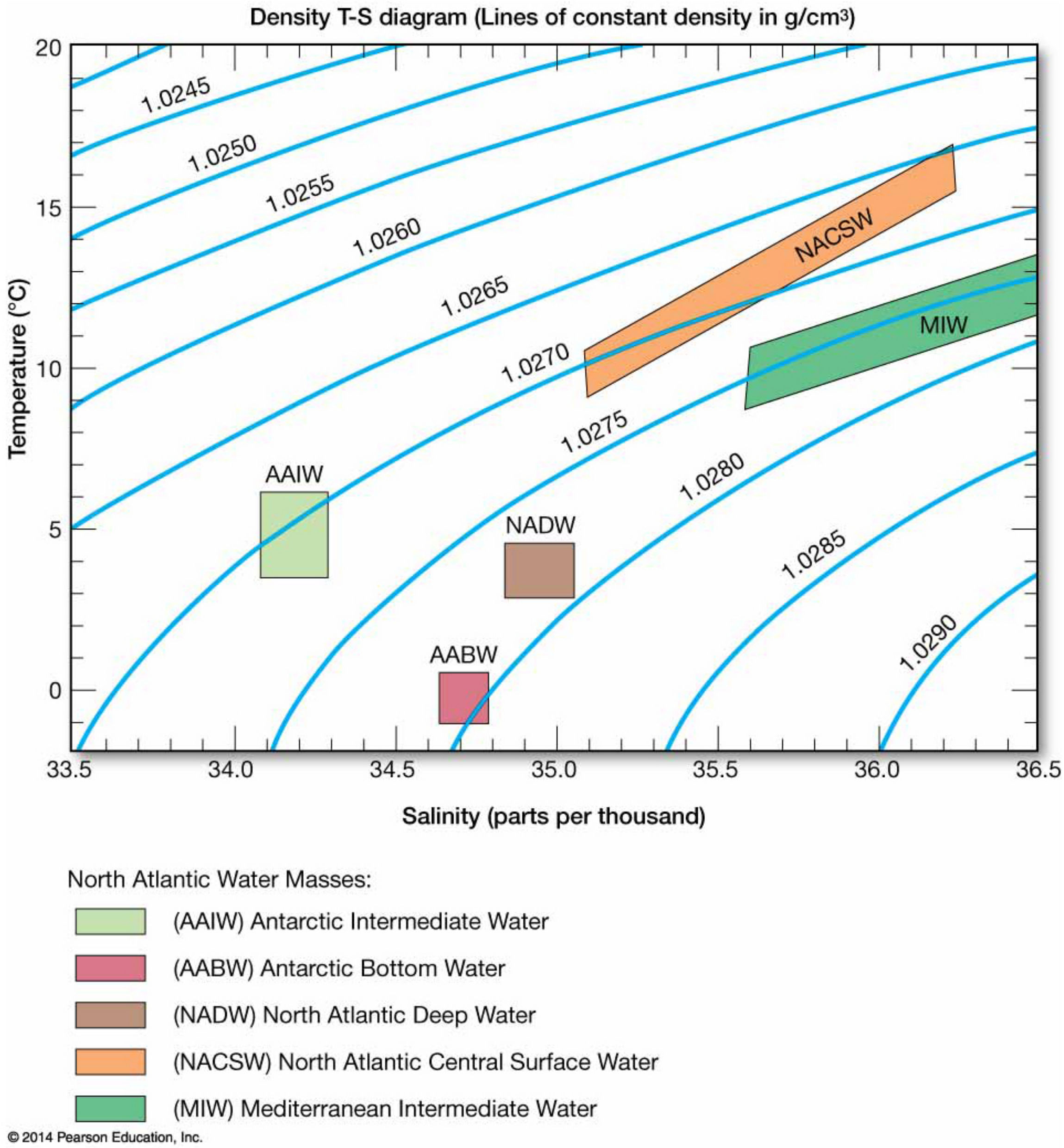


Figure 10.17. Density resulting from various temperature and salinity combinations.

Cold water from the poles is densest and sinks. This water then travels horizontally at depth throughout the ocean basins. The South Pole is slightly colder than the North Pole, and therefore surface water around Antarctica is denser than surface water in the Arctic. Consequently in the Atlantic, the densest water is produced around Antarctica and forms the Antarctic Bottom Water. The North Atlantic Deep Water, formed in the Arctic, sinks down and travels south at depth. It is the deepest body of water at high northern northern latitudes but it flows above the Antarctic Bottom Water once the two bodies of water meet (Figure 10.18). The Antarctic Convergence zone leads to the sinking of the Antarctic Intermediate Water, which is less dense than the North Atlantic Deep Water.

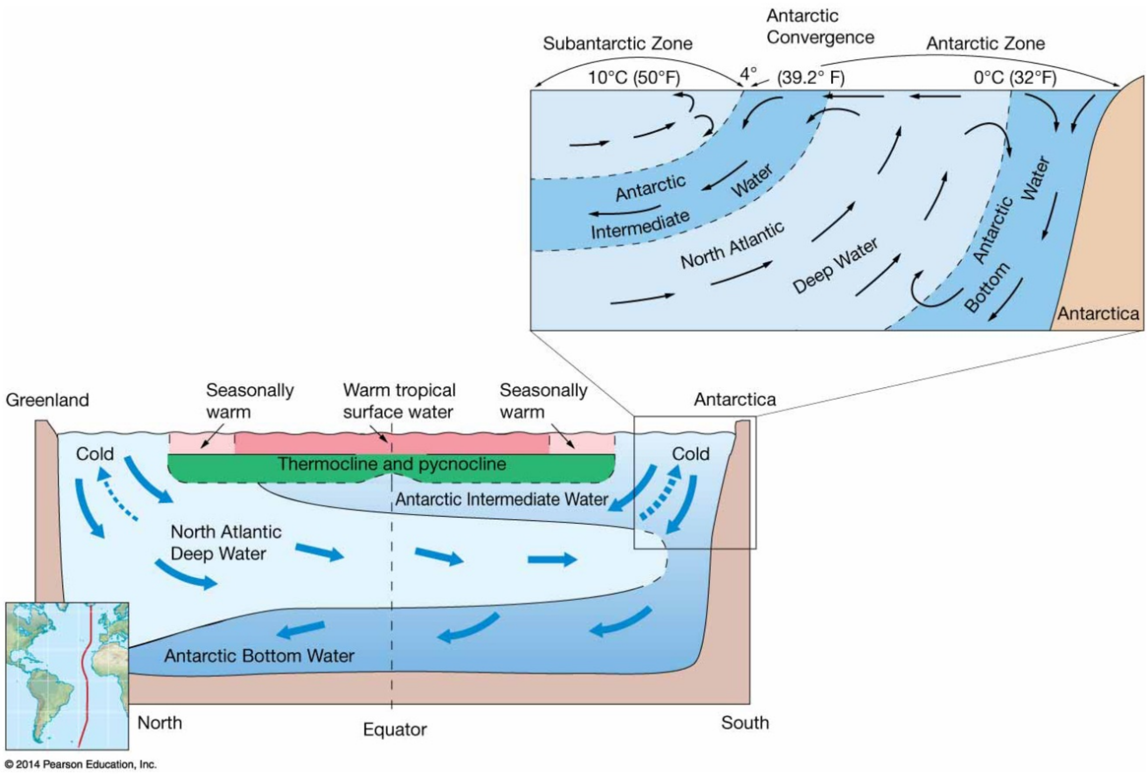


Figure 10.18. Cross-section of the Atlantic showing thermohaline flow.

Thermohaline circulation has the greatest impact on water movement at depth. Because of continuity of flow, water moving up or down in the water column also creates horizontal water movement (10.19).

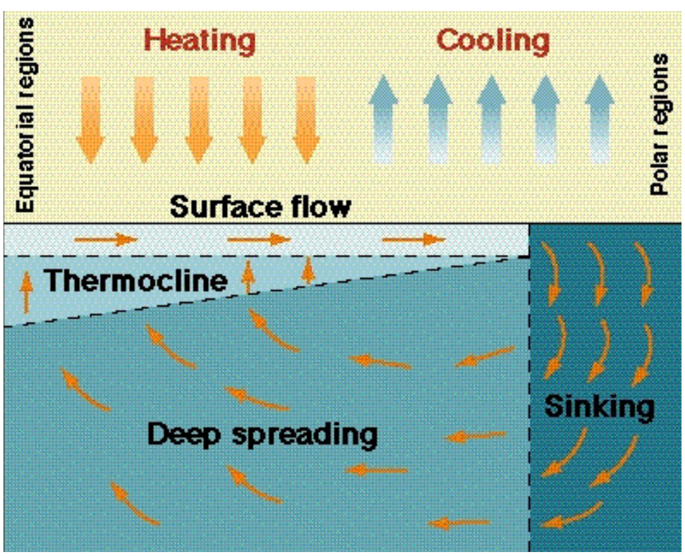


Figure 10.19. Generalized model of themohaline circulation, showing horizontal movement of surface water.

This movement of water created by thermohaline circulation is quite slow, and it creates a circulation pattern that exchanges water between the surface and depth and between ocean basins over a period of about 1000 years. This is called the global conveyor belt (Figure 10.20), and it has a great impact on global climate because of the potential to distribute heat throughout the oceans.

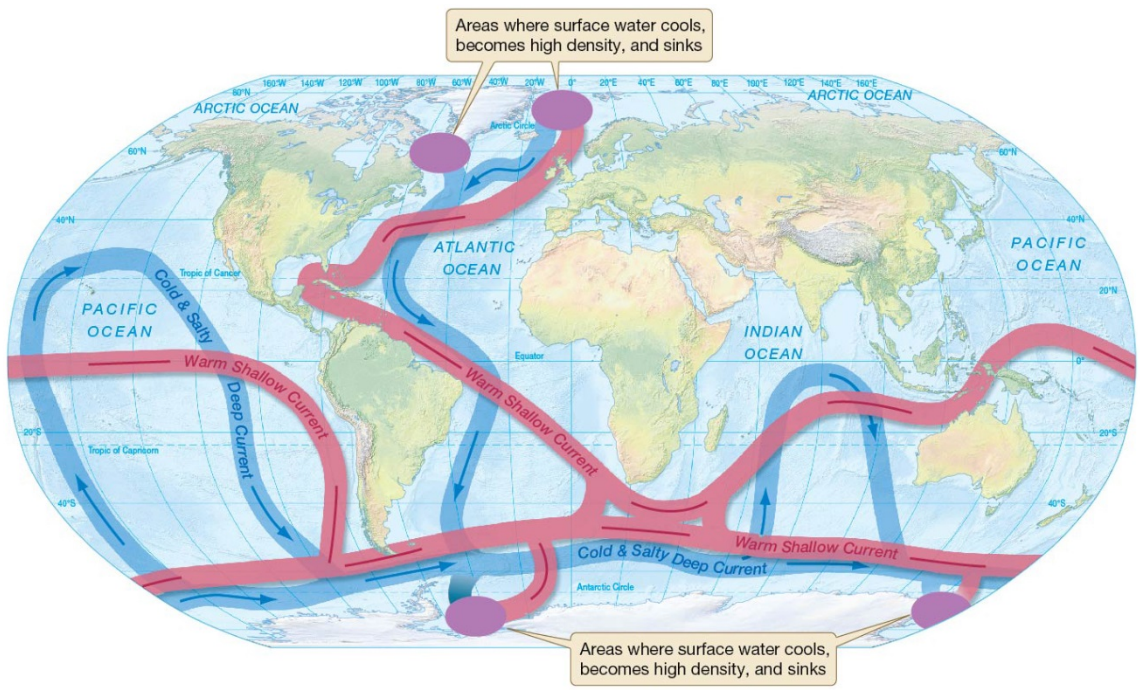


Figure 10.20. Global conveyor belt of thermohaline circulation.

10.5. Review Questions

1. Name the 2 major types of ocean circulation
2. Which type of ocean circulation is more important at depth?
3. True or False? Two bodies of the exact same density, when mixed, can create a resulting water of higher density.
4. The water at the very bottom of the ocean originated from what geographical area?
5. Why is the surface water at the equator colder than water a few degrees N and S?
6. What direction is surface current, relative to the wind that created it?
7. What is the average direction of the top 150 m of water, relative to the wind, and what term describes this?
8. What is a gyre?
9. What direction do gyres rotate in the northern hemisphere?
10. What two opposing forces create geostrophic flow around oceanic gyres?
11. On which side of an ocean basin would you find the strongest currents? Why?
12. What vertical water movement is created when currents collide or when surface current is forced against a landmass?
13. Explain why the fisheries on the western coast of Peru are so productive?
14. Explain why there are sometimes pockets of warm water and tropical marine life north of the Gulf Stream in cold waters of the Atlantic? In which direction do these pockets spin?
15. In El niño years, how to the areas of high and low atmospheric pressure differ from normal years in the South Pacific? How does the thermocline differ?
16. How does surface circulation pattern change with monsoons in the Indian Ocean?
17. In which direction do warm eddies from the Gulf Stream spin?
18. Where is the Sargasso Sea? Why does it form there?

11. Waves (Trujillo, Chapter 8)

11.1. Wave generation

Waves are moving energy that are created by disturbances and travel along the interface between two fluids. In this chapter, we will focus on surface waves which occur at the boundary between the ocean and atmosphere, but similar waves can be created between water masses of different density (Figure 11.1). Progressive waves are waves in which there is a direction to the transmission of the wave, away from the disturbing force. They can be created by a variety of disturbing forces, including wind, seismic activity, changes in atmospheric pressure, landslides, volcanic eruptions, or gravity (e.g. tides; Figure 11.2). Tides are discussed separately in the next chapter.

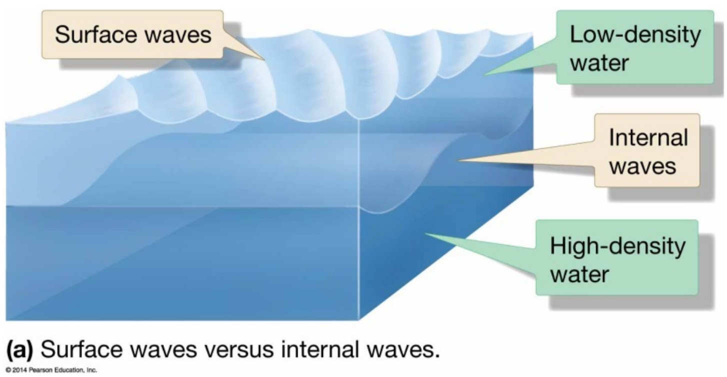


Figure 11.1. Surface waves and internal waves

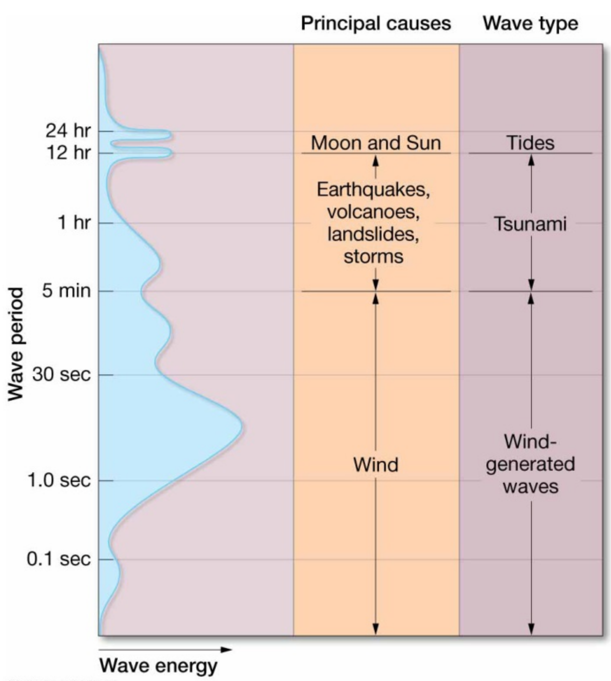


Figure 11.2. Types of waves and their causes. Most of the energy from waves in the ocean comes from wind-generated waves.

11.2. Wave characteristics

As a wave moves across the surface of the water, individual water particles are set in motion in a circular pattern called an orbit (Figure 11.3). The energy is transmitted to depth to other orbits, which decrease in diameter with increasing depth. The wave motion and orbits extend to a depth of about half the wavelength. The various parts of a wave are shown in Figure 11.3. The crest is the highest part of a wave; the trough is the lowest. The wavelength is the horizontal distance between 2 successive crests or troughs. The height is the vertical distance between a crest and a trough. The period is the time it takes for 2 troughs or 2 crests to pass a fixed point. The period is determined by the generating force and remains unchanged once a wave is created. The frequency is the number of wave crests or troughs that pass a fixed point each second. The celerity is the speed of the wave, equal to the wavelength divided by the period.

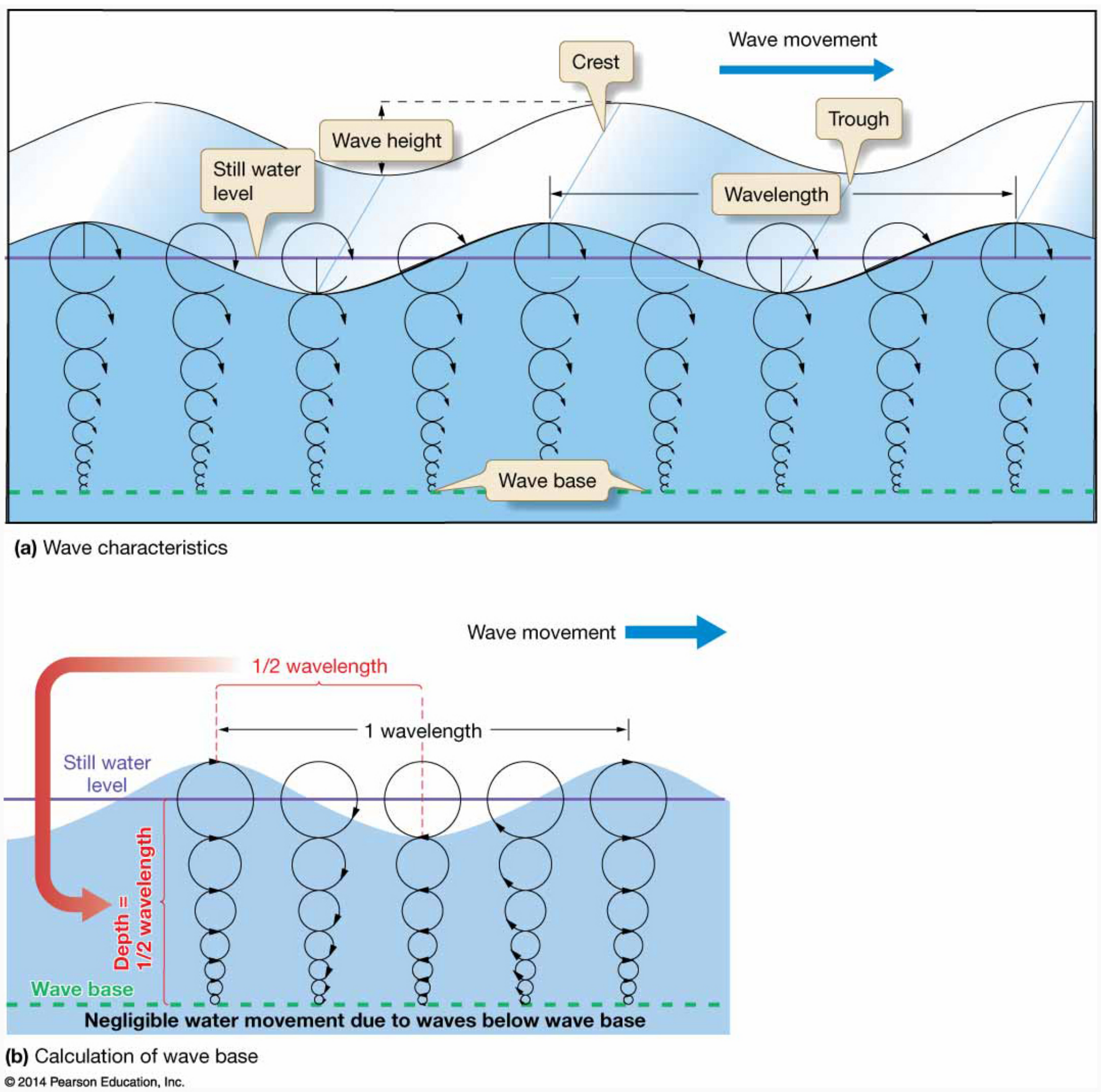


Figure 11.3. Orbits of water particles extending to a depth of ½ the wavelength.

The depth of a wave is defined by the ratio of the wavelength to the water depth (Figure 11.4). Deep-water waves occur in water that is deeper than one-half of its wavelength, where the orbits and water motion do not reach the sea floor. Under these conditions, the speed is dependent only on the wavelength, with longer wavelengths producing faster waves. Deep water waves are mostly wind-driven waves. Shallow-water waves occur where depth is less than 1/20 of the wavelength. The ocean floor interferes with the orbits and flattens them. The speed of shallow-water waves is influenced only by depth, as friction with the seafloor slows down the wave. Therefore shallow water waves have slower velocities at shallower depths. Shallow-water waves include wind-generated waves that have moved into shallow water, but also waves that have such long wavelengths that they are always a shallow-water wave, such as tsunamis and tides. Intermediate or transitional waves occur in water that is between ½ and 1/20 of the wavelength. The orbits start interacting with the sea floor and become elliptical because of friction.

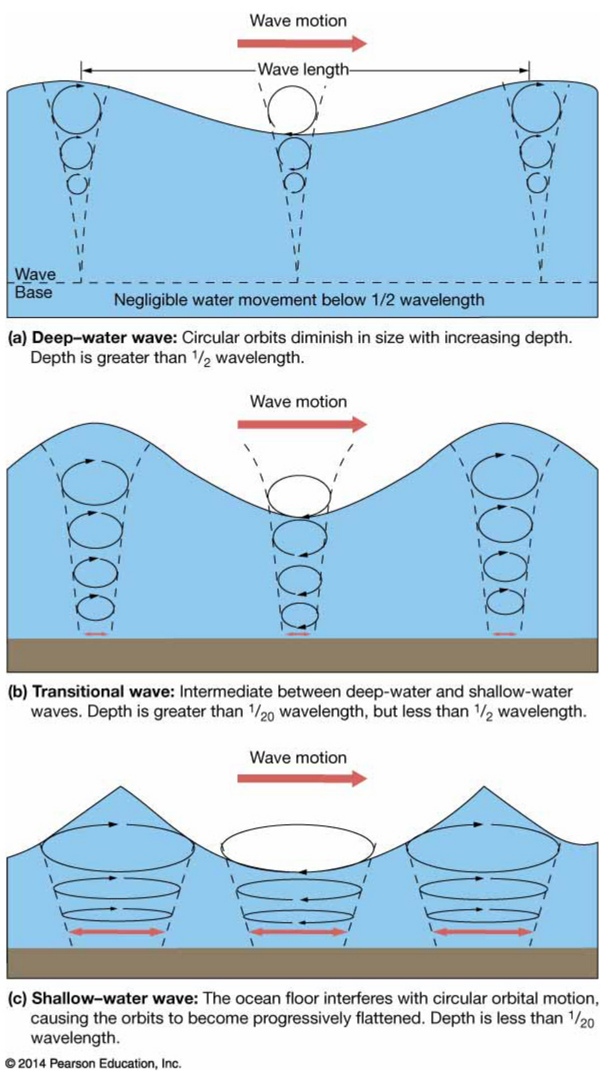


Figure 11.4. Deep, intermediate and shallow water waves, defined by the depth in relation to their wavelengths.

11.3. Development of wind-generated waves

Wave development

Wind-generated waves are formed by the transfer of wind energy to water. As wind blows across the surface of water, friction stretches the surface and creates ripples called capillary waves. If the wind stops blowing, capillary waves are restored by surface tension. If the wind keeps blowing, it grips the capillary waves and forms larger waves known as gravity waves, as they are restored by gravity. The height of a wind-generated wave is affected by three factors: the wind strength, the wind duration, and the fetch (the distance uninterrupted by land over which the wind has blown). Most waves observed at sea are progressive wind waves formed in local storm centers. Waves are known to be forced waves while they are under the influence of their generating force, and as free waves once they move away from the storm and are no longer under their generating force. Once away from the storm center, waves travel at speeds determined by their wavelengths. Waves with long wavelengths have a greater speed than those with short wavelengths, and will gradually move ahead of the shorter waves in a process called wave sorting (Figure 11.5). Those waves with long, regular wavelengths are called the swell, and they travel away from the storm center with little loss of energy.

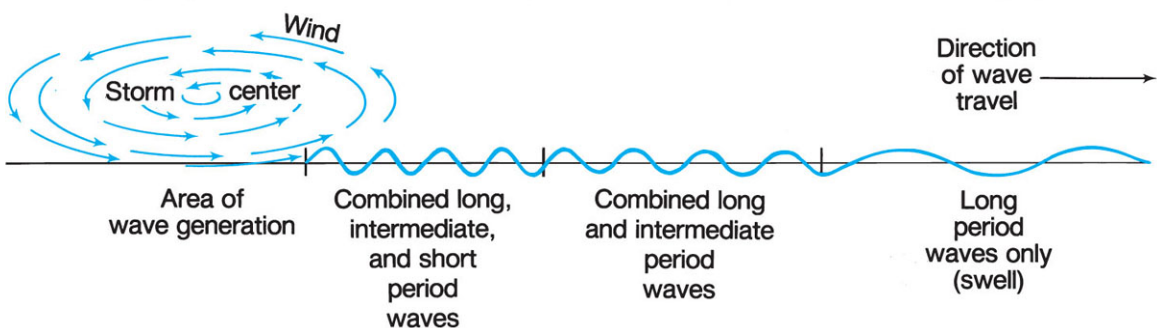


Figure 11.5. Formation of deep-water oceanic swell and wave sorting.

A group of waves leaving the area of the disturbing force becomes a wave train. Careful observation of waves as they travel away from the disturbing forces shows that the leading wave keeps disappearing. When a wave is lost from the leading edge, a new one is formed in the back of the group (Figure 11.6). Because of this phenomenon, wave trains travel along the surface of the ocean at half the velocity of individual waves in the group.

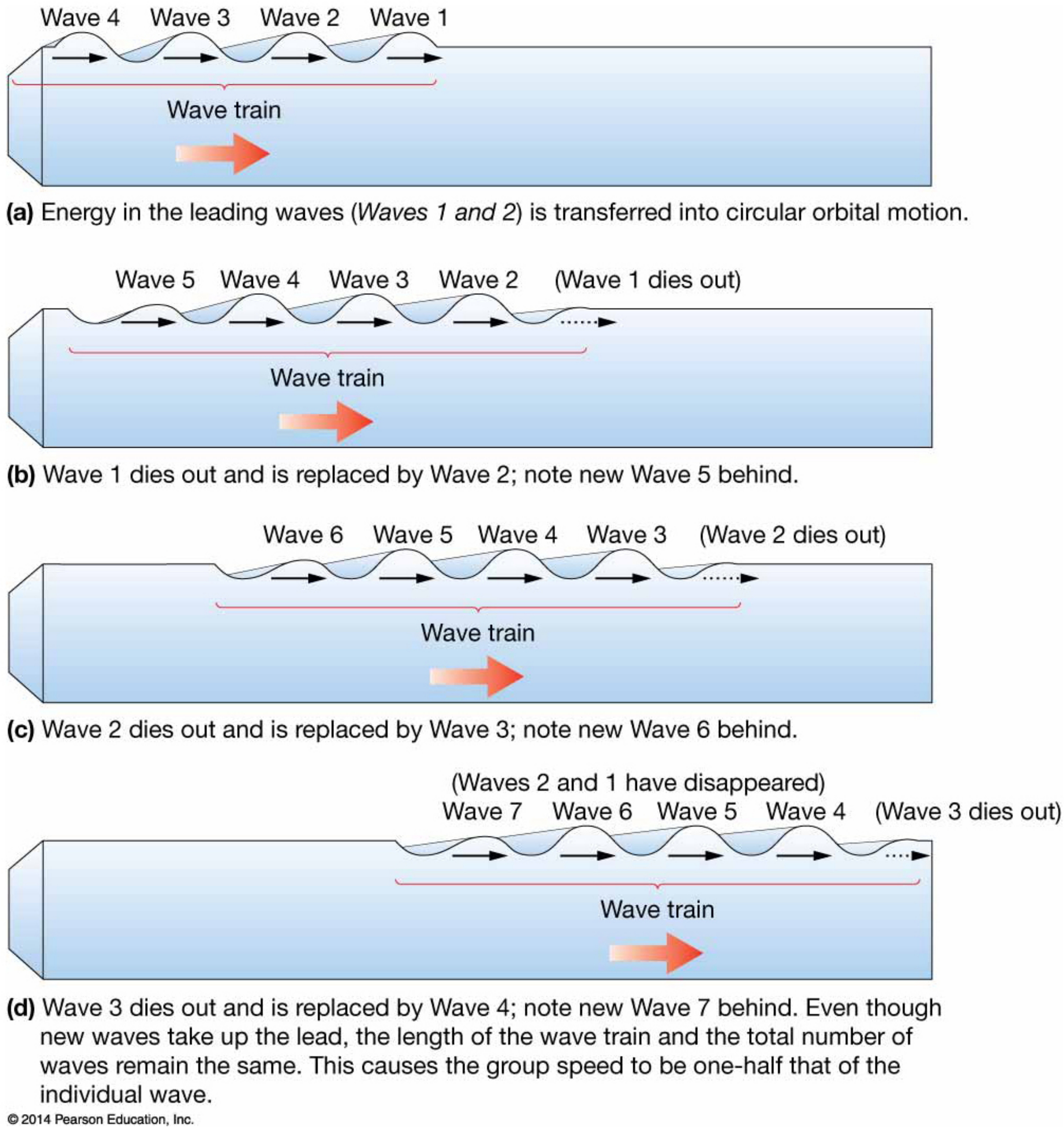


Figure 11.6. Movement of a wave train.

Interference patterns

When two waves meet and their crests and troughs match, their energy is added and the resulting wave is bigger than the original two (Figure 11.7); this is called constructive interference. Destructive wave interference, on the other hand, occurs when the crest of one wave meets the trough of another wave, and their energy is subtracted, resulting in a reduced wave.

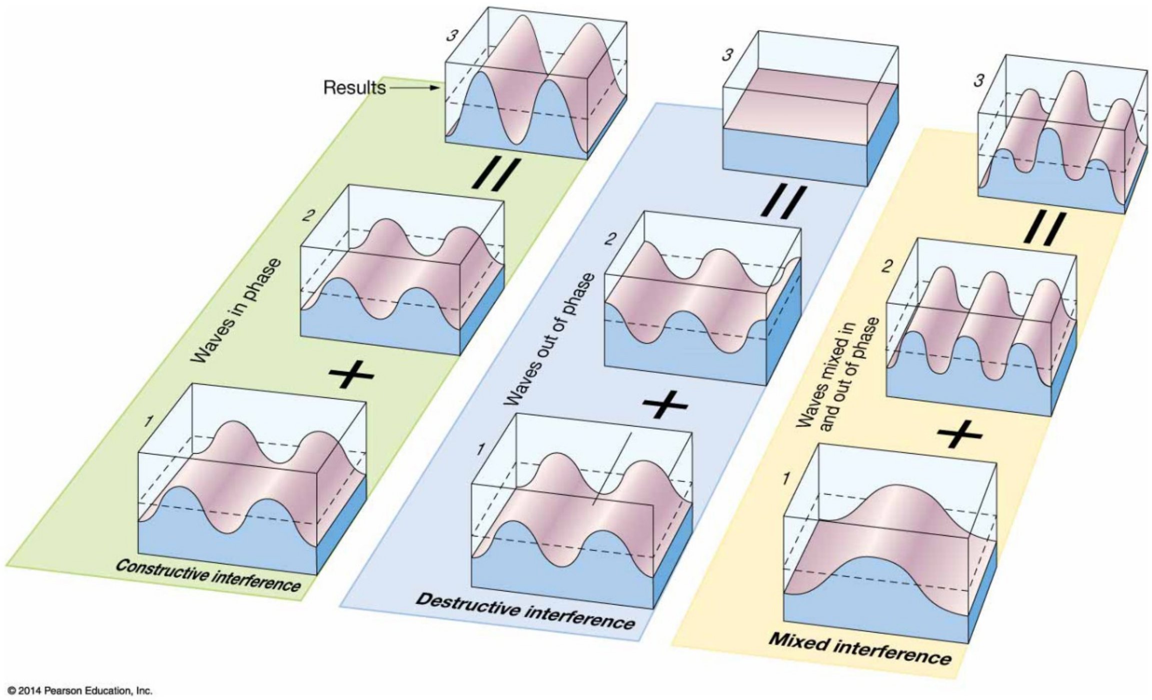


Figure 11.7. Various types of wave interference.

More complex patterns occur when waves of different wavelengths meet. Constructive interference occurs where their crests and troughs meet, and destructive interference where their crests and troughs are offset (Figure 11.8). This leads to commonly observed sequence of large waves coming in sets of 5 to 7, with a few minutes of smaller waves between sets.

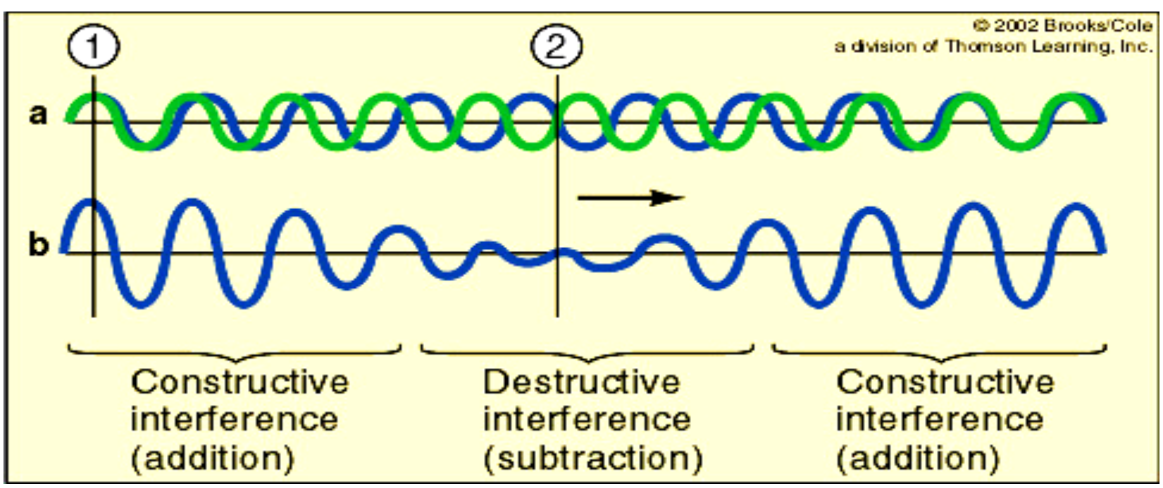


Figure 11.8. Wave interaction and the formation of wave sets.

Often, waves meet that come from various angles, which produces mixed interference patterns (Figure 11.9). For example, waves that meet at right angles produce a checkerboard pattern.

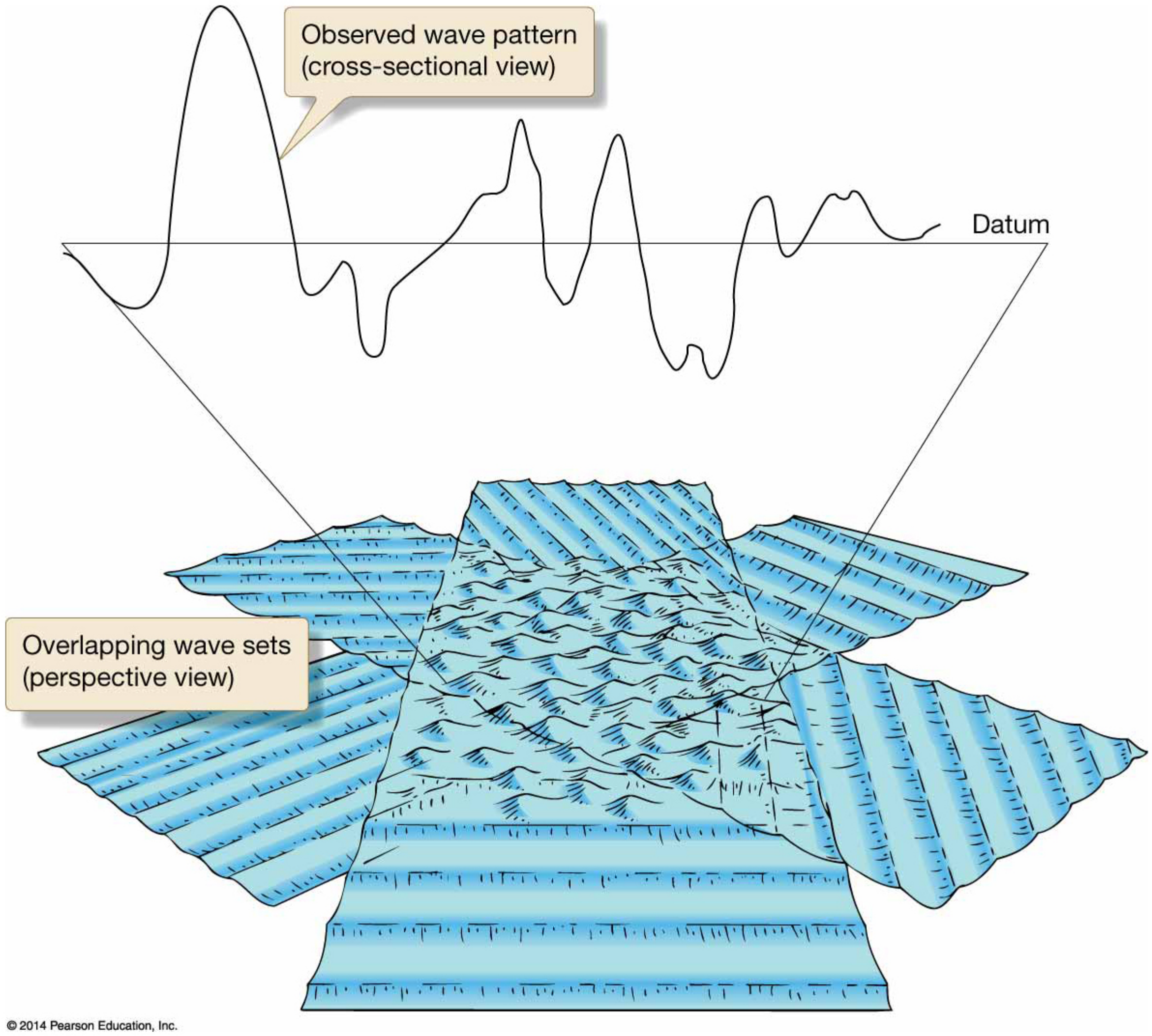


Figure 11.9. Mixed interference pattern resulting from waves meeting from various directions.

Rogue waves

Rogue waves are very large, isolated ocean waves that often occur when other ocean waves are not especially high. Their cause is usually an extreme case of constructive interference where multiple waves overlap exactly in phase to produce an extremely high one. Rogue waves may also be created when strong ocean currents amplify an opposing swell. This is seen along the southeast coast of Africa, where the Agulhas current (flowing south) runs against large waves formed near Antartica (Figure 11.10).

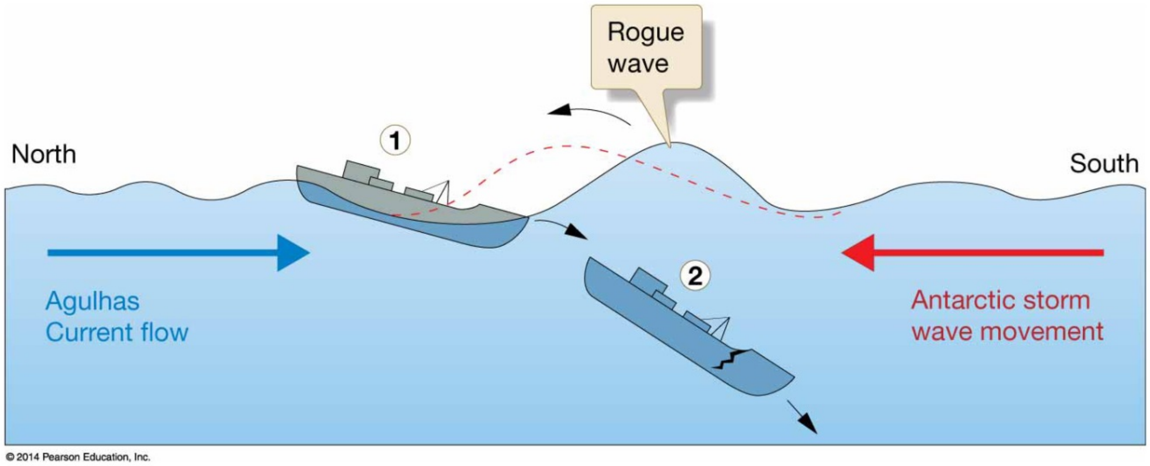


Figure 11.10. Rogue waves along the southeast coast of Africa, formed by the meeting of a strong current and large storm waves.

11.4. Wave changes in the surf zone

Waves that travel across the ocean as swell eventually release their energy in the surf zone, where waves break. As the deep-water waves of swells move into gradually shallower water that is less than ½ their wavelength, they become transitional waves and eventually shallow water waves. The interaction with the seafloor slows down the first wave that feels the bottom. Since the wave behind continues at the original speed, the wavelength decreases. As shallow-water waves slow down and the wavelength decreases, the height of the waves increases (Figure 11.11).

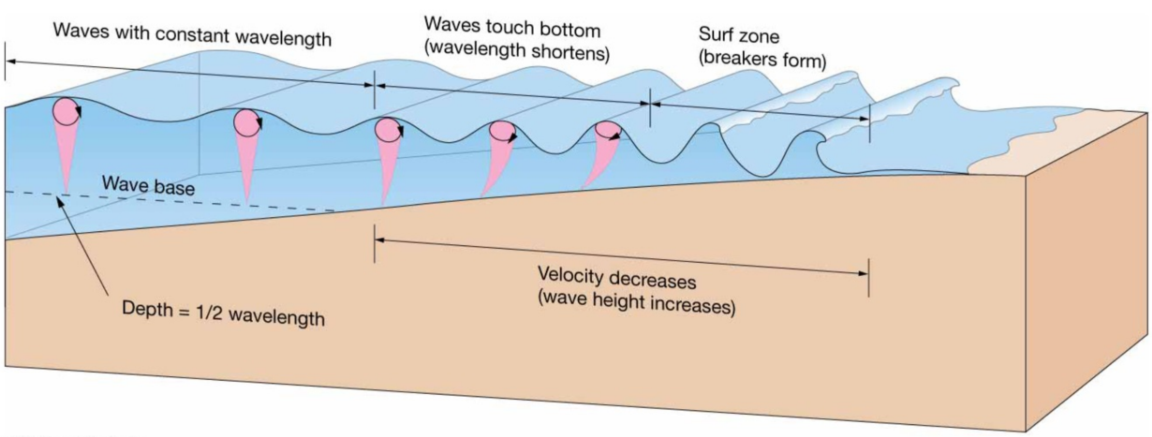


Figure 11.11. Profile of a shallow water wave. As the wave approaches shore, wave speed (celerity) and wave length decrease, while wave height increases.

Waves break when they reach a height to length ratio of ~1:7. The type of breaker depends on the slope of the shore and the speed at which the wave loses its energy (Figure 11.12). Spilling breakers occur on wider, flatter beaches and lose their energy gradually. Because of this slow release of energy, spilling breakers can provide surfer with a longer ride than other breakers. Plunging breakers occur on narrow, moderately sloped beaches where the crest of the wave curls over an air pocket. Plunging breakers lose their energy fast and provide an exciting ride for surfers. Surging breakers occur on very abrupt shores, where there is no opportunity for a proper break. They cannot be surfed on.

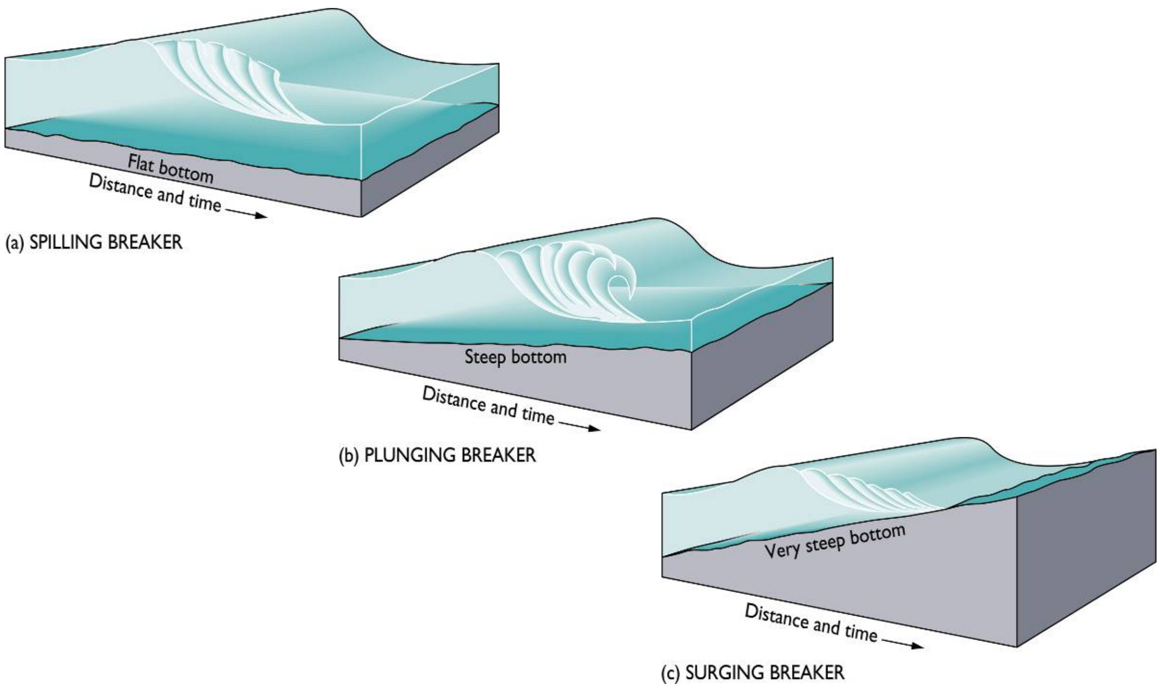
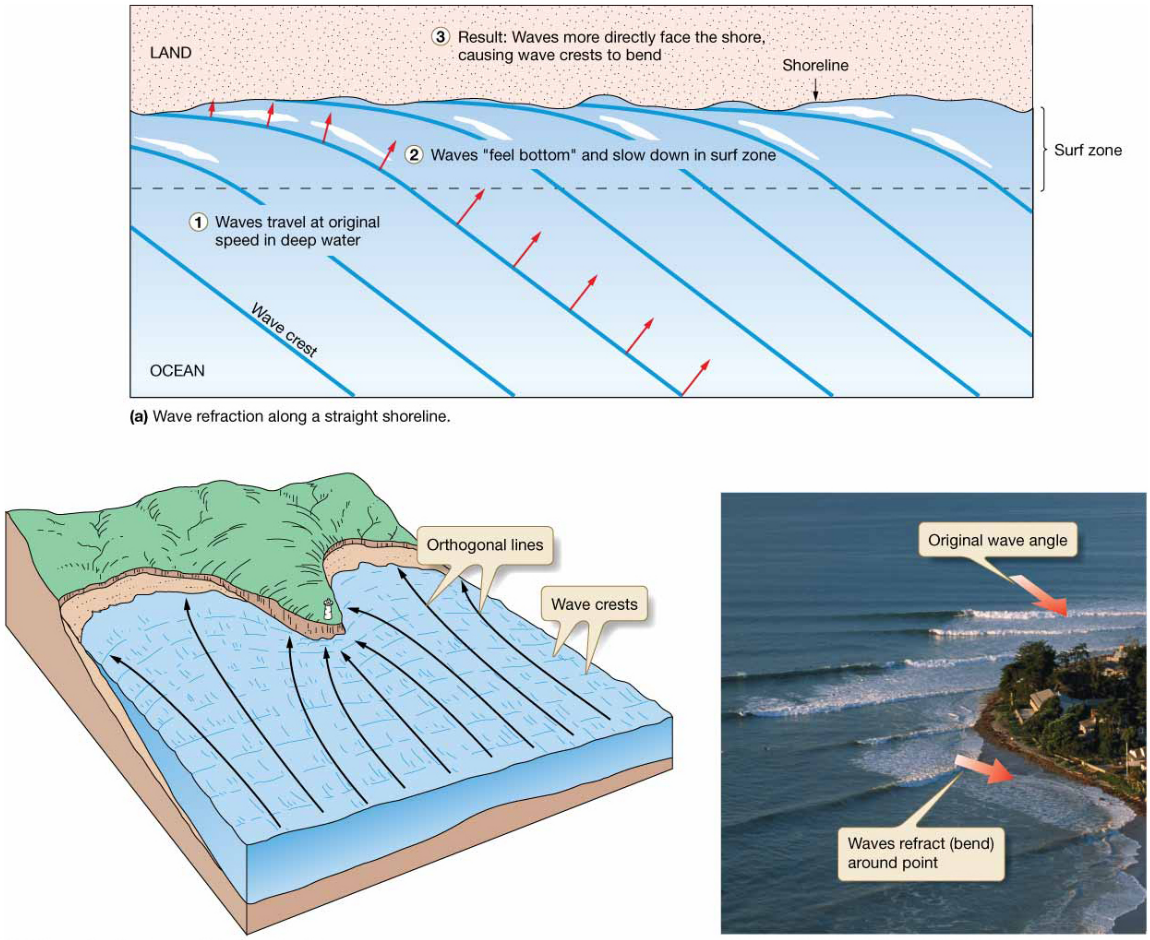


Figure 11.12. Different types of breakers as a wave enters shallow waters of different slopes.

Refraction, reflection and diffraction of waves
Like other waves (e.g. light and sound), surface waves can be refracted, reflected and diffracted. Refraction occurs as waves move from deep to shallow water. If a wave comes in at an angle relative to shore, one end reaches shallow water first and begins to slow, while the rest of the wave remains at the same speed, resulting in bending of the wave (Figure 11.13). The shape of the refraction depends on the shape of the shoreline and sea floor. Reflection occurs when waves meet a straight, smooth, vertical barrier in water deep enough to prevent breaking. As the reflected waves pass through incoming waves they will produce interference (Figure 11.14). Reflected waves can also produce a standing wave (see below). Diffraction is the spread of wave energy sideways to the direction of wave travel. It occurs when waves pass around an obstacle (Figure 11.15).



(a) Wave refraction along a straight shoreline.

(b) Wave refraction along an irregular shoreline.

(c) Wave refraction at Rincon Point, California, looking west.

Figure 11.13. Refraction occurs as the area of the wave that enters shallow water first slows down, while the area in deeper water continues at the original speed, resulting in bending along the coast line.

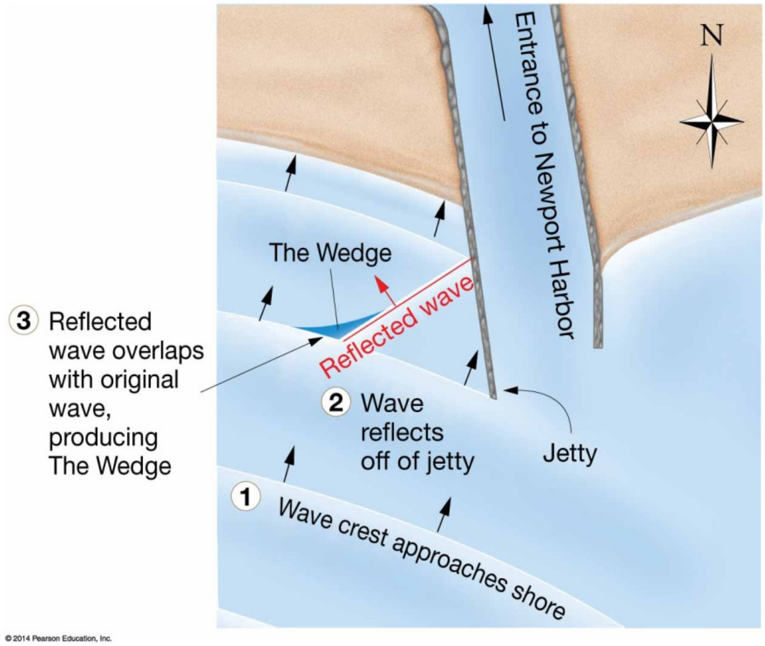


Figure 11.14. Wave reflection and interference in Newport Harbor, California

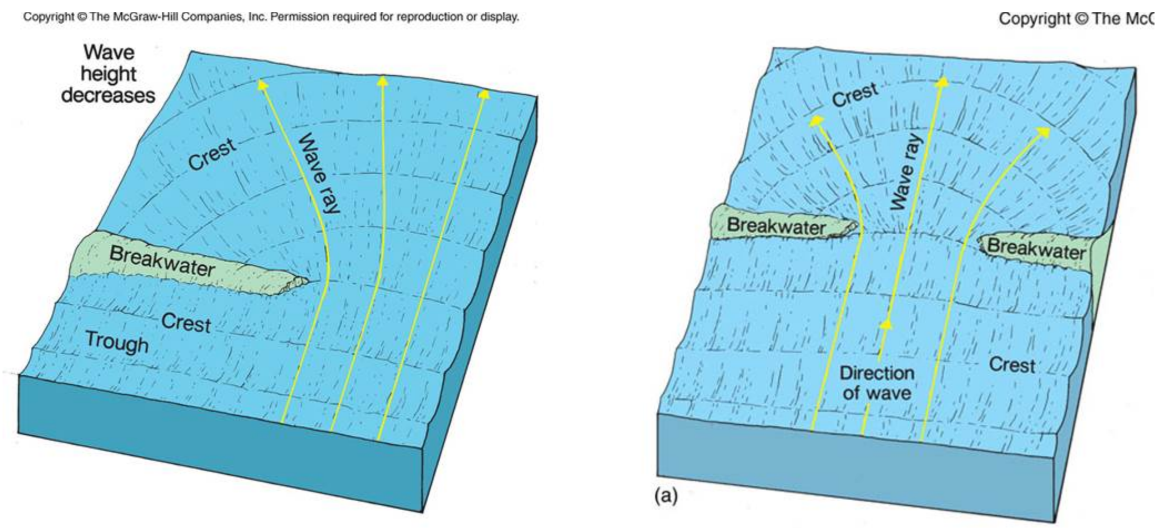


Figure 11.15. Diffraction occurs as the wave passes through an opening or around an object.

Standing waves

Standing waves oscillate vertically with no forward movement, which creates an alternation between a trough and a crest at a fixed position. The area of no or little vertical movement is called the node; the alternation of high and low water occurs at the antinode (Figure 11.16). Standing waves can be created by progressive waves reflected back on themselves, or by tectonic movement, storm surges, and tidal waves.

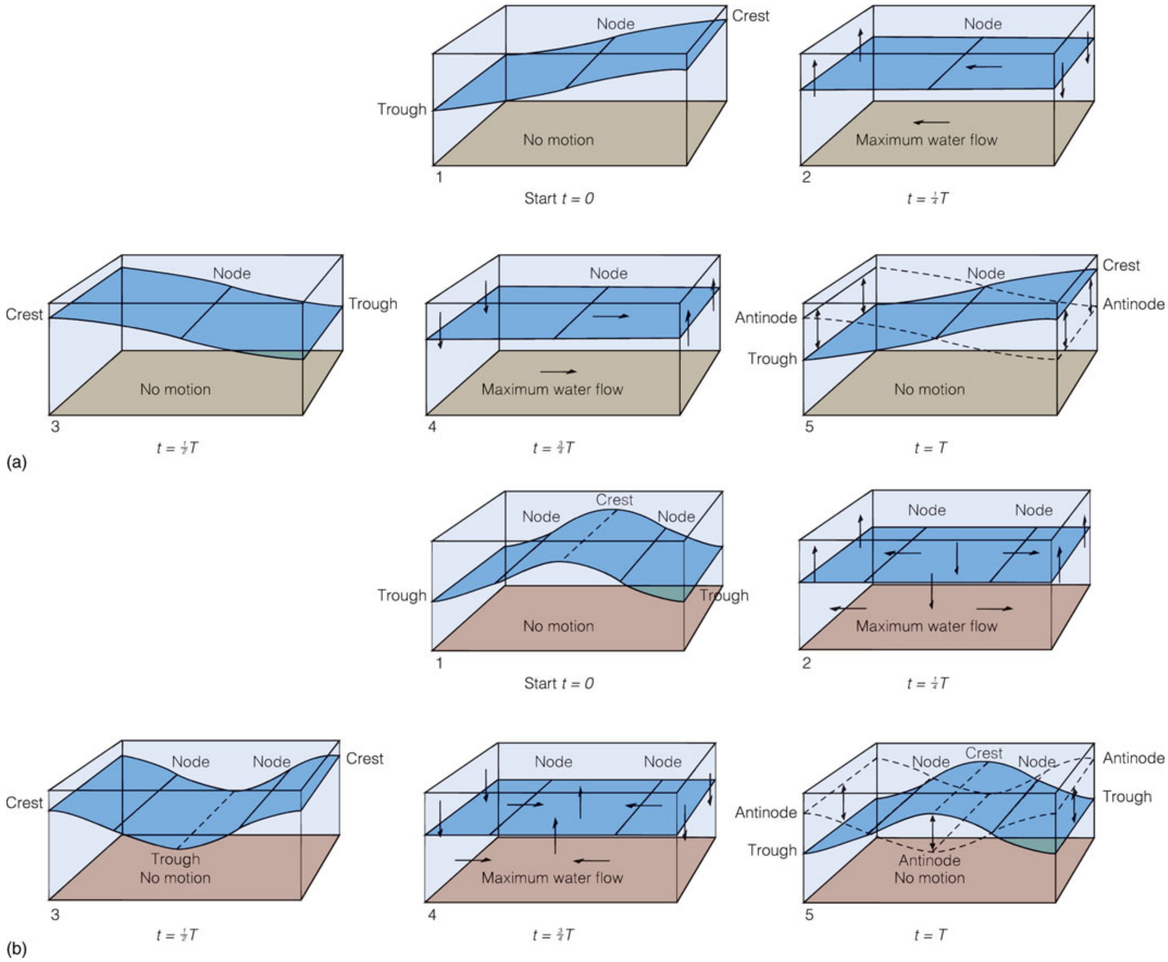


Figure 11.16. Sequence of water movement in a standing wave with 1 (a) or 2 (b) nodes, showing the area of highest vertical movement at the antinodes, and no vertical changes at the node.

11.5. Tsunamis

Tsunamis are sea waves cause by rapid displacement of water (Figure 11.17). They are typically triggered by earthquakes, landslides, volcanic eruptions, or iceberg calving. Tsunamis have extremely long wavelengths (100-200km), long periods (10-20 minutes) and are always shallow-water waves. As such, their speed is determined by ocean depth, and is typically about 200 m/s (720 km/hr). Tsunamis typically have a small wave height and are very difficult to detect in deep water, as the length is up to 600 times its height. However, when the wave reaches shore it slows down and greatly increases in height (Figure 11.18).

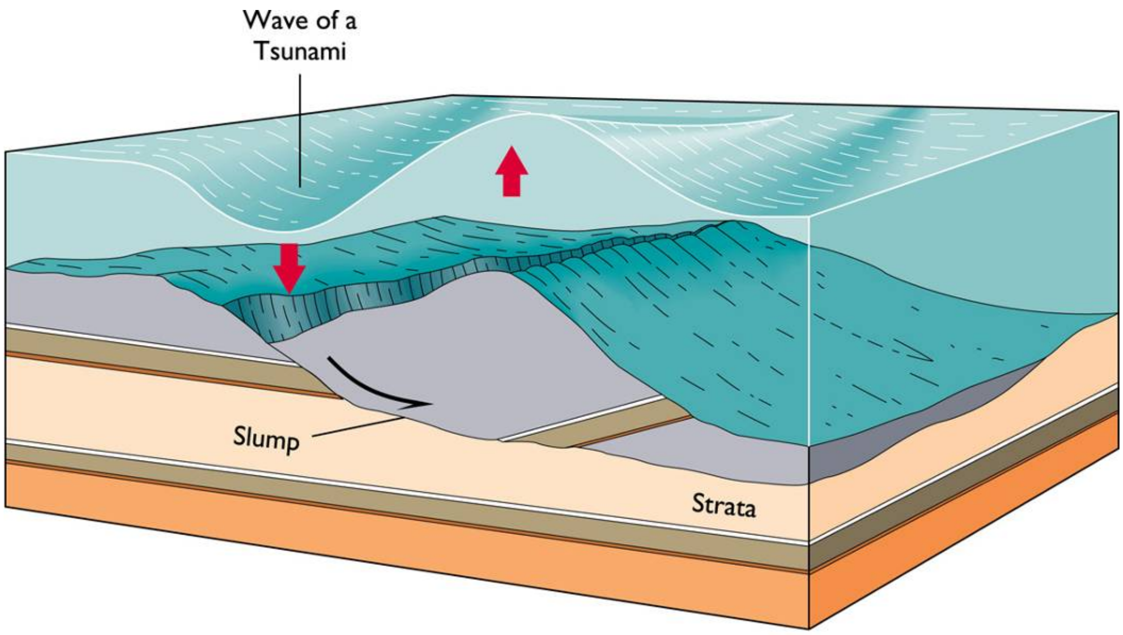


Figure 11.17. The generation of a tsunami by movement of the sea floor.

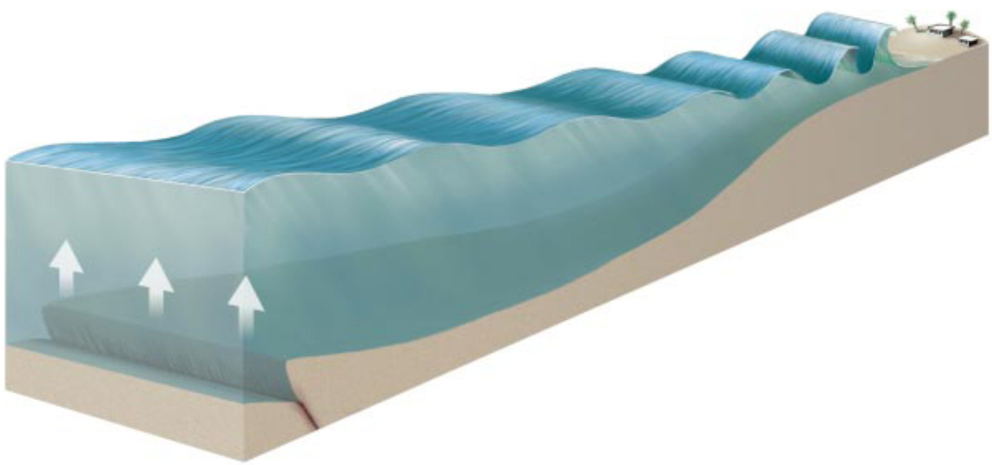


Figure 11.18. Decrease in wavelength and increase in wave height as a tsunami approaches shore.

11.6. Review Questions

1. What is the period of a wave?

2. How do you calculate the velocity of a wave?

3. What is a shallow water wave?

4. What is a deep water wave?

5. What is a free wave?

6. What is an example of a wave that is always a forced wave?

7. In a standing wave, what is the name of the point of least (or zero) vertical motion, and what is the name of the point with maximum vertical motion?

8. Which types of wave (short or long wavelengths) travel faster?

9. Which characteristic of a wave (velocity, wavelength or period) remains constant once the wave is formed?

10. What depth do orbits extend to?

11. What is the restoring force for capillary waves?

12. What three factors determine the height of a wind wave?

13. Which type of breaker is the best to surf on?

14. What kind of shore slope would create a surging breaker?

15. What causes refraction of waves as they approach shore?

16. In which condition is a wave reflected?

17. Why do waves come in sets of 5-7?

18. What is the name of the point of little vertical movement in a rotary standing wave?

19. True or false? The velocity of tsunamis is determined by ocean depth.

20. Are tsunamis shallow water or deep water waves?

12. Tides (Trujillo, Chapter 9, Duxbury chapter 10)

Tides are waves created by the gravitational pull of the moon and the sun. Their wavelength can equal half the circumference of the earth, and therefore they are always shallow-water waves and their speed is controlled by the depth of the water. The crest of the wave is high tide and the trough is low tide. Tides are forced waves as they are always under the influence of their generating force. Although the earth moves at a speed of 463 m/s at the equator, tide waves only move at a speed of about 200 m/s because of friction with the sea floor.

The tidal range is the vertical difference between high tide and low tide. Tidal datum is the reference level from which ocean depths and tide heights are measured. Typically on U.S. charts it is the mean lower low water. The current created as the tide comes in is called a flood tide, and as it comes out, an ebb tide. The time of minimal tidal current at high tide and at low tide is known as slack tide.

12.1. Causes of tides

Let's first consider a simplified earth with no land masses, and oceans of uniform depth. This is known as equilibrium tidal theory and is useful to understand the basic behavior of tide waves.

Tides are caused by three main factors: the gravitational pull of the moon, the gravitational pull of the sun, and the rotation of the earth. Though the moon has a much smaller mass than the sun, it is much closer and therefore has a greater effect on tides. The earth-moon system rotates about a common axis at the system's center of mass, which is about 4640 km from the earth's center (the barycenter; Figure 12.1).

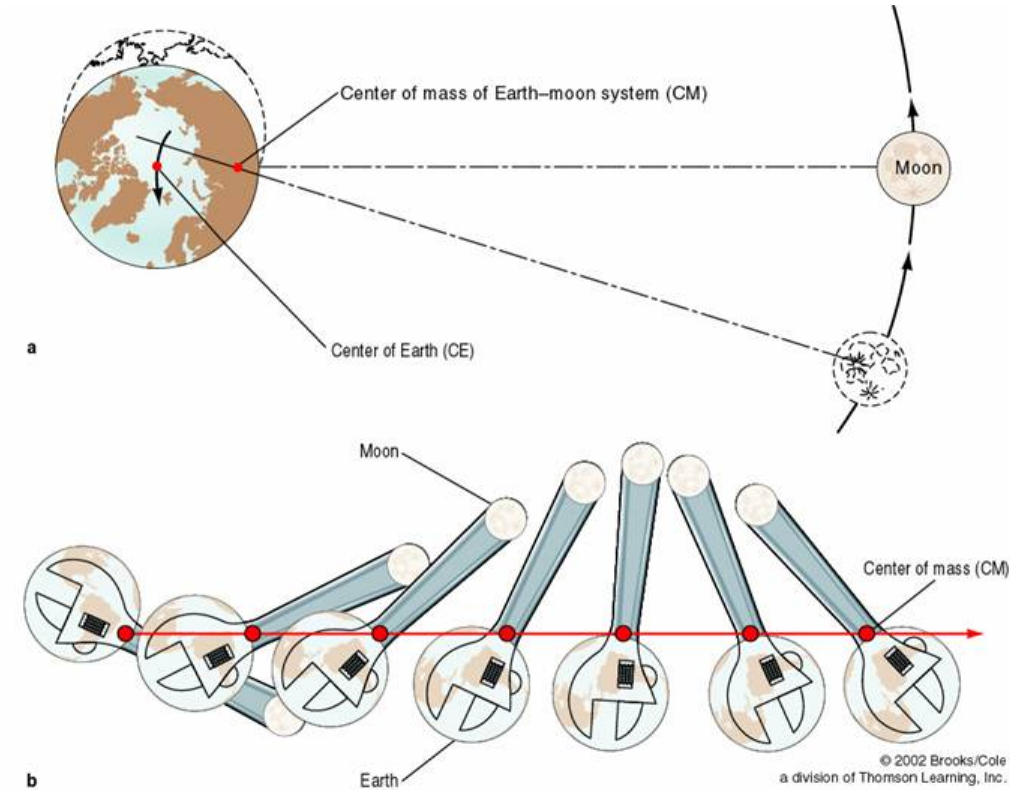


Figure 12.1. The earth and the moon rotate about their center of mass.

The gravitational force of the moon pulls water particles on the side of the earth directly under the moon to create a bulge. Additionally, as this system rotates, the centrifugal force of the earth-moon system pulls water particles on the side of the earth opposite the moon to create another bulge (Figure 12.2). These two bulges are the crests of the tidal wave, and the two depressions are the troughs.

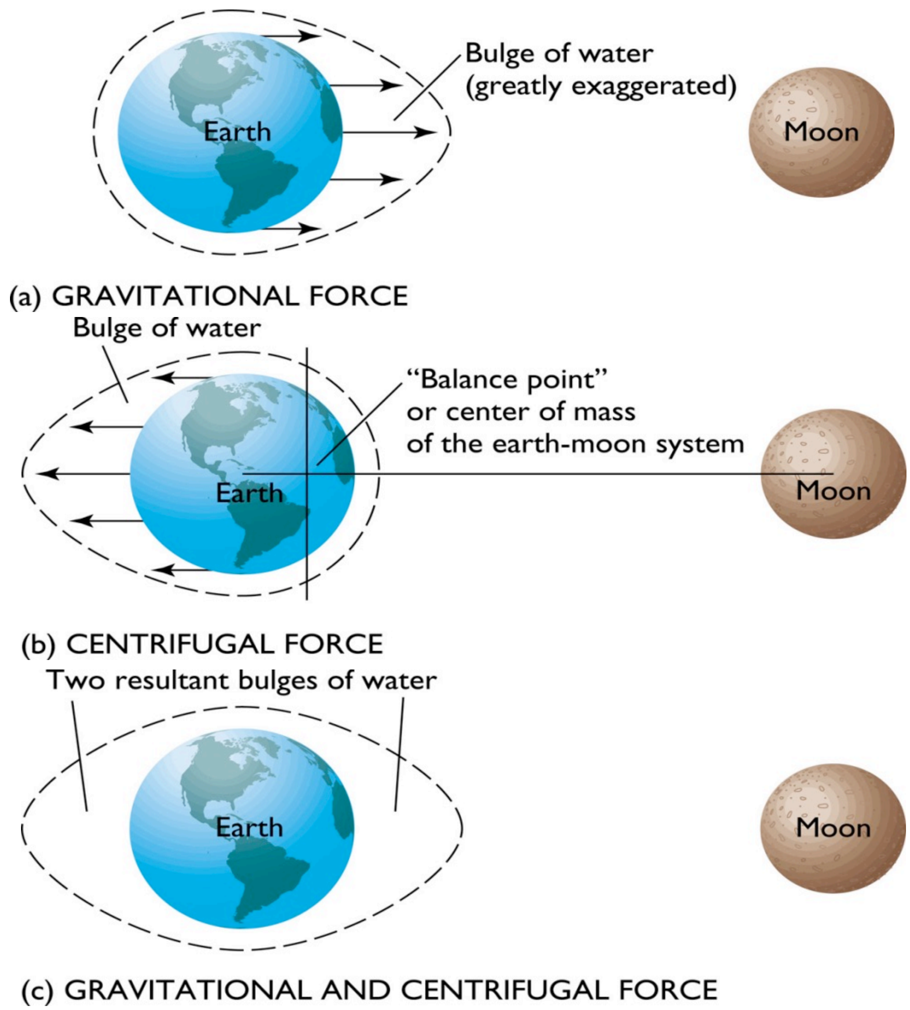


Figure 12.2. Two bulges of water are created by the gravitational pull of the moon and the centrifugal force of the rotation of the earth-moon system.

While these water bulges are created by the moon, the earth makes one full rotation in 24 hours. The bulges stay under the tide-producing body as the earth turns, and these bulges appear to move towards the West as the earth turns towards the East, creating high and low tides (Figure 12.3).

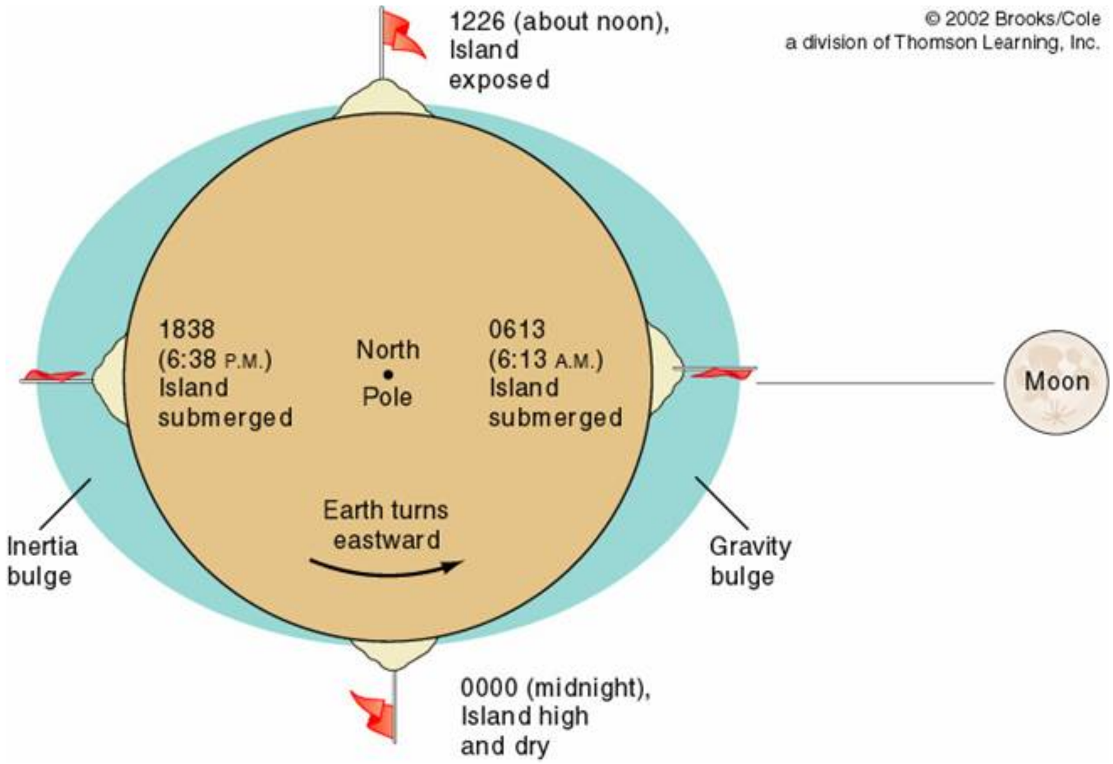


Figure 12.3. Island experiencing high and low tides that appear to move West as the earth rotates towards the East and the tidal bulges remain directly under and directly opposite to the moon.

A tidal day is 24 hours and 50 minutes, however, not 24h. Because the moon has a greater influence on tides than the sun, a tidal day is the time it takes for a point on the earth to be directly under the moon again. The earth completes a full rotation in 24 hours, but during that time the moon moves 12.2^o around the earth on its revolution cycle of 29.53 days. It takes the earth another 50 minutes of rotation for a point to be again directly under the moon (Figure 12.4). It is for this reason that the moon rises 50 minutes later each night, and that tides are 50 minutes later every day.

Figure 12.4. A tidal day of 24 hours 50 minutes, as the earth requires 50 minutes to complete the extra 12^{o} of rotation to be directly under the moon again.

Because the sun is much farther away (Figure 12.5), its tide-raising force is approximately 46% that of the moon. The sun tidal day is 24 hours (as opposed to 24 hours 50 minutes for the moon) and the sun tide is constantly moving West relative to the moon tide. The changes in the position of the moon relative to the earth and the sun creates monthly changes in tides.

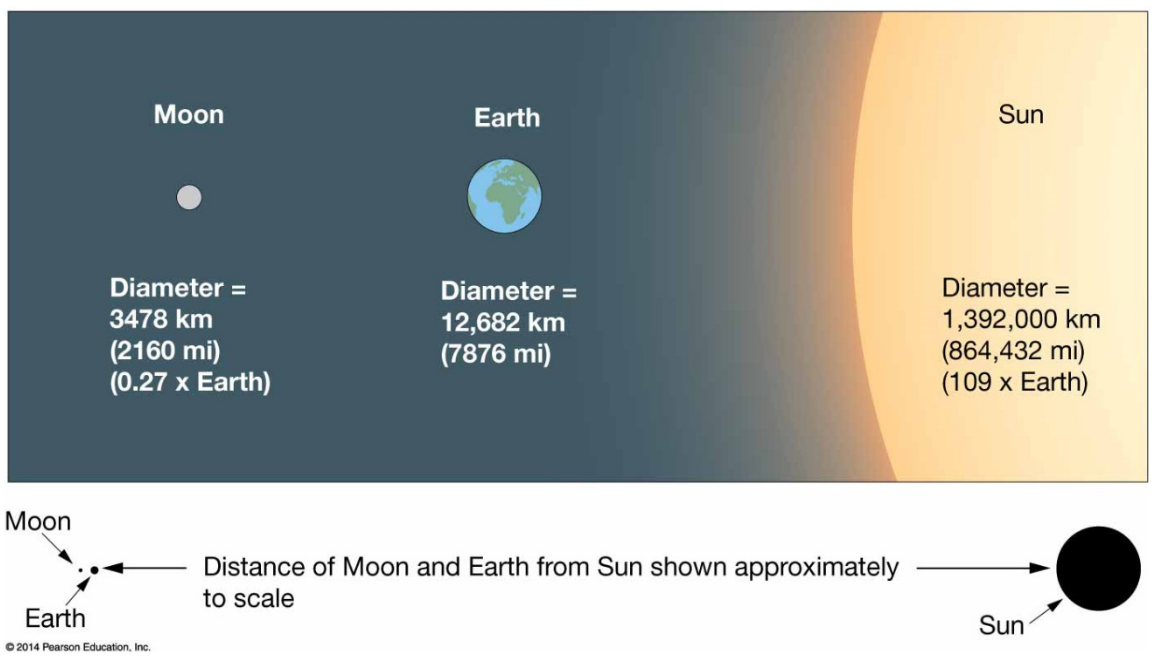


Figure 12.5. Relative sizes and distances of the Moon, Earth and Sun.

12.2. Monthly tidal cycle

When the moon, the earth and the sun are all in line, the bulges created by the moon and the sun coincide to create larger bulges (constructive interference) and a maximum tidal range. This is called a spring tide, and happens twice a month, during the new moon and again during the full moon. A neap tide, on the other hand, occurs when the moon, the earth and the sun are at right angle to each other, and the bulges created by the moon and the sun are out of phase. This leads to destructive interference, smaller bulges and minimum tidal range. Neap tides occur twice a month, during the first and third quarter (Figure 12.6).

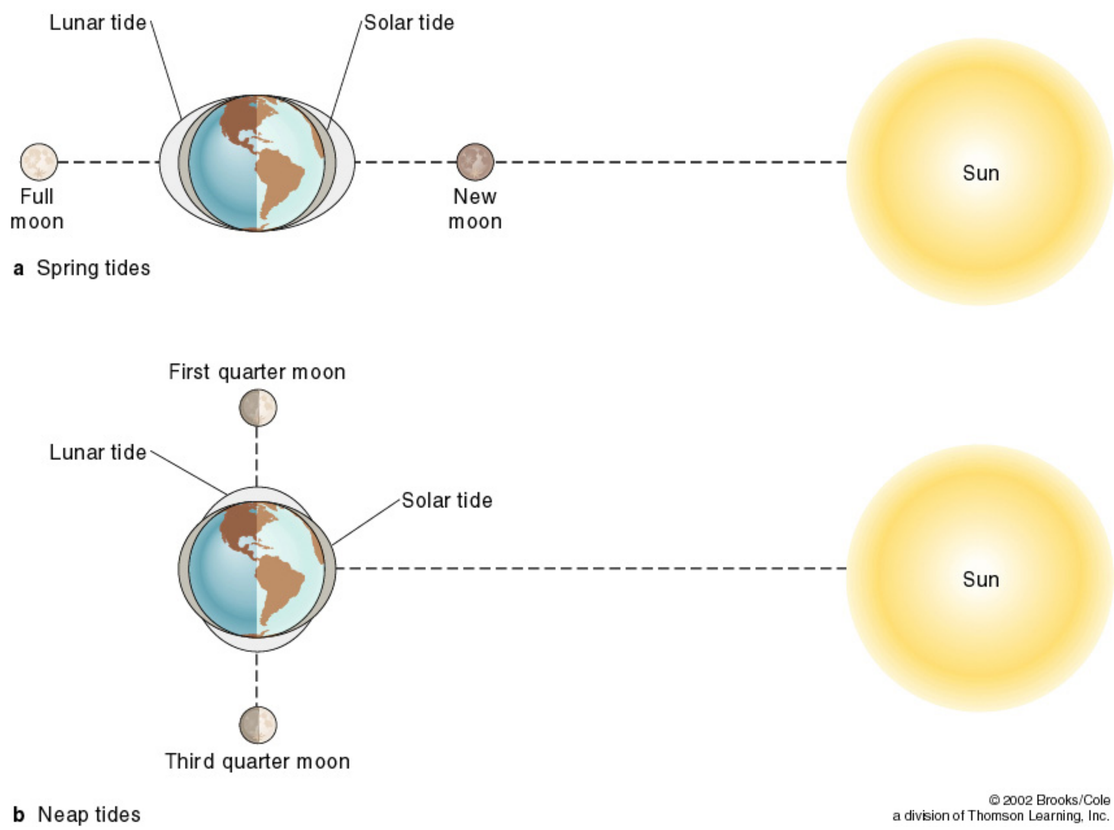


Figure 12.6. The relative position of the moon, earth and sun affects the tidal range.

Effect of declination

In the simple explanation of tides so far, it was assumed that the sun and the moon always remain directly above the equator. However, this is not the case. The axis of rotation of the earth is tilted 23.5^o relative to its plane of revolution around the sun (this is why we experience seasons). The plane of the moon's orbit is tilted about 5^o relative to the plane of the earth's revolution around the sun, so the angle of the moon relative to the equator, called declination, varies from 28.5^o north to 28.5^o south.

The tidal bulges will therefore rarely be aligned with the equator. Since the moon is the dominant force creating the tides, the bulges mainly follow the moon as it changes declination to a maximum of 28.5° North and South. In this case, one bulge is created in each hemisphere, creating different tidal patterns at different latitudes. In this model, high latitudes only experience one of the bulges every day, and therefore show a diurnal tide pattern. Equatorial regions experience each bulge, though not the highest section of it. Therefore they have semi-diurnal tides with a smaller range than high latitudes. Intermediate latitudes experience mixed semi-diurnal tides (Figure 12.7).

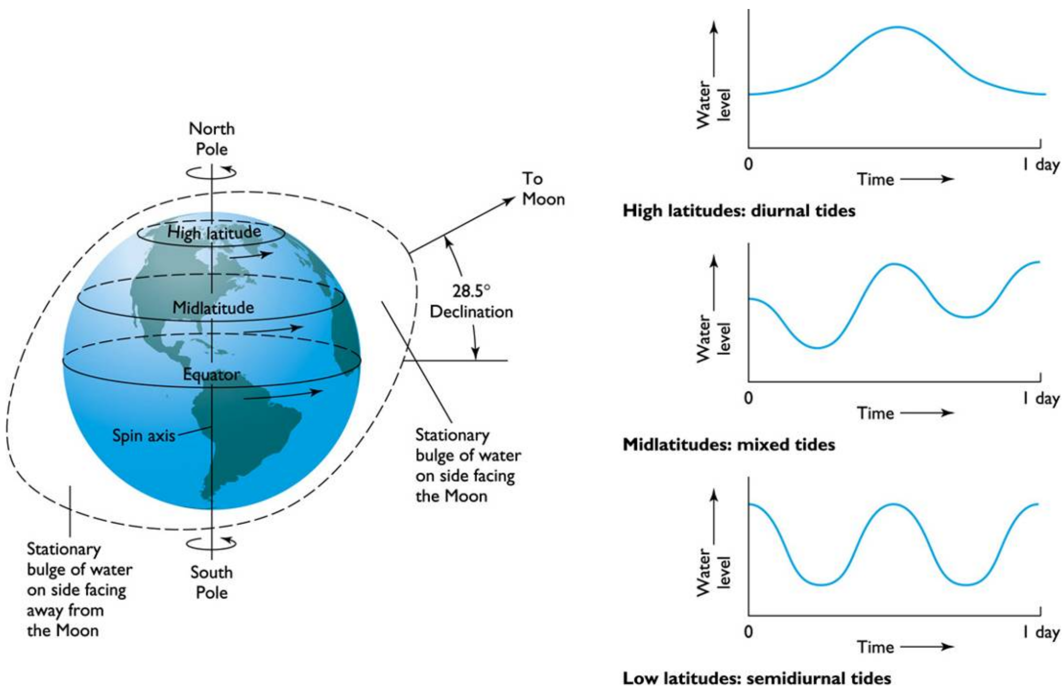


Figure 12.7. The declination of the moon creates one bulge in each hemisphere and creates different tidal patterns at different latitudes.

12.3. Actual tide patterns in the ocean and coastal zones

Because of their extremely long wavelengths (half the world's circumference), tide waves are always shallow water waves. Their speed is controlled by the depth of the ocean and the tidal bulges cannot keep up with the speed of the rotation of the earth. Moreover, continents prevent tide waves from travelling all the way around the world. Ocean tides instead break up into large circulation units called cells.

Tide waves can move either as progressive waves or standing waves. Progressive tide waves occur in large ocean basins or semi-enclosed basins where the tide wave moves across the sea-surface as a shallow-water wave, where orbits reach the sea floor and are elliptical. The crest of this wave lags behind the moon because friction with the seafloor slows down the speed. Progressive tide waves occur in the Western North Pacific and the South Atlantic (Figure 12.8). As the progressive tide wave moves across the ocean, the Coriolis effect deflects water movement to the right in the Northern hemisphere (and to the left in the Southern hemisphere). Progressive tide waves also occur in some narrow semi-enclosed basins (Figure 12.9).

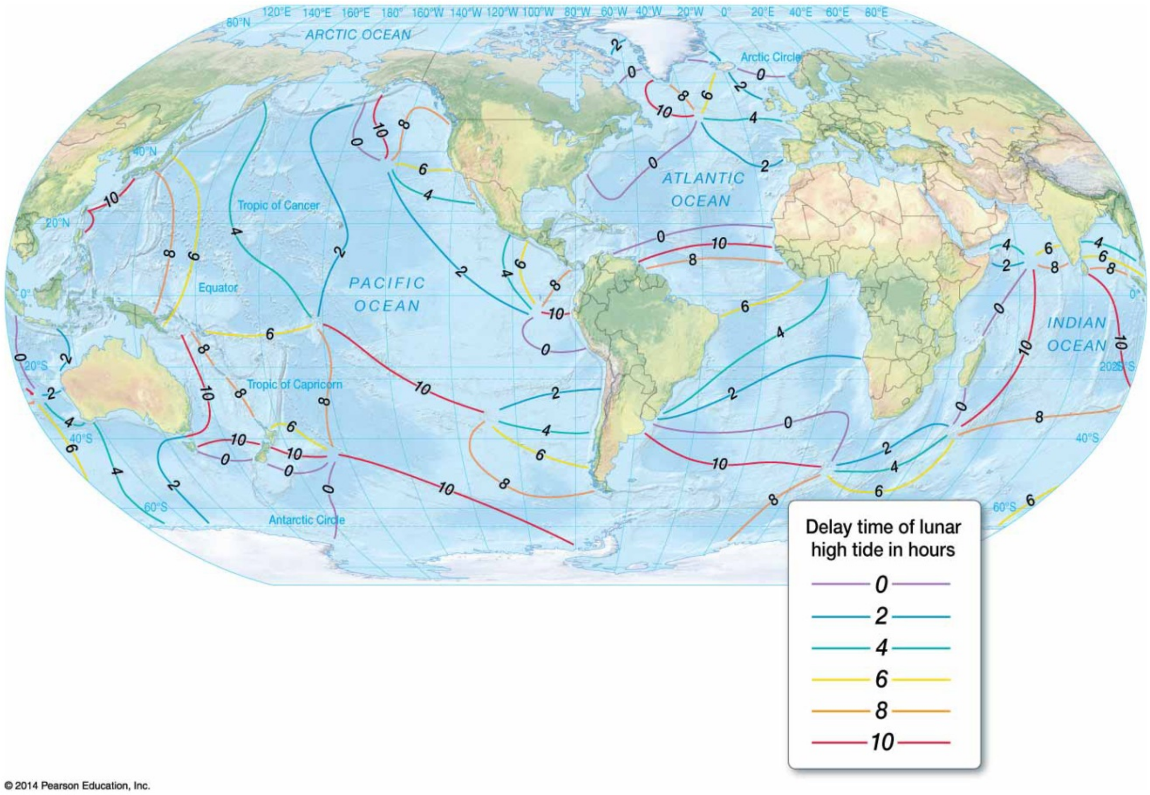


Figure 12.8. Cotidal lines across the oceans showing progressive tide waves in the Western North Pacific and the South Atlantic, and rotary standing waves in the North Atlantic. Cotidal lines connect all points experiencing the same phase of the tide (e.g. maximum or minimum) at a given time.

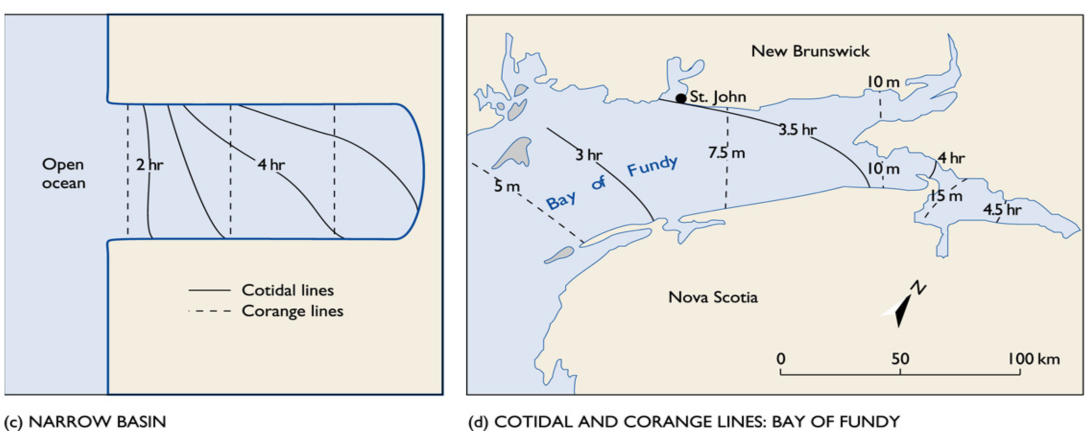


Figure 12.9. A progressive tide wave in a semi-enclosed basin such as the Bay of Fundy.

Standing tide waves occur in ocean basins where the wave is reflected by the edge of continents. In this case, the Coriolis Effect produces a rotary standing wave with an accompanying rotary tidal current. In the North Atlantic as the tide crest moves west it piles up on the coast of North America. As the tide comes out, the water is deflected to the right, towards the equator. Water accumulates at low latitudes and when it is forced back at higher latitudes by gravity, it is again affected by Coriolis and veers to the right, towards Europe (Figure 12.10). This process continues until a counter-clockwise rotary standing wave is created (clockwise in the Southern hemisphere).

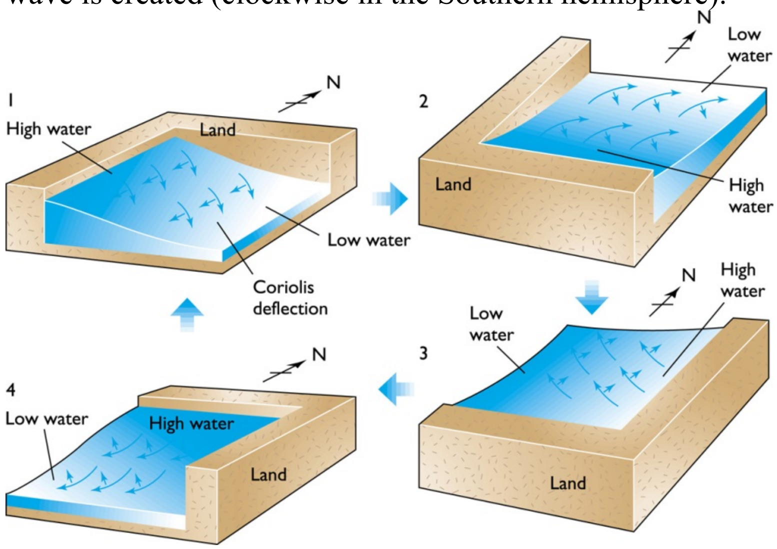


Figure 12.10. Reflection of the tide crest on the edge of continents and the Coriolis Effect result in a counter-clockwise rotary standing wave in the northern hemisphere.

In a rotary standing wave, the highest tidal range is found on the outside of the basins (antinodes) and no or little tidal range is found in the middle of the basin (node, also called amphidromic point; Figure 12.11). Rotary standing waves can be found in the open ocean as well as in broad semi-enclosed basins (Figure 12.12).

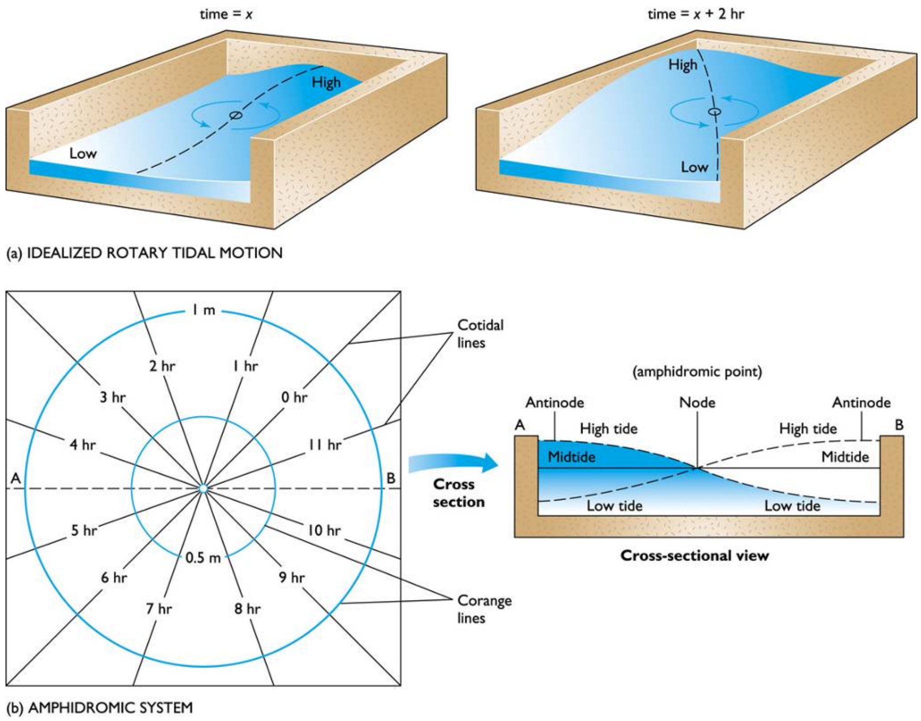


Figure 12.11. The formation of a rotary standing wave, showing the antinodes and amphidromic point.

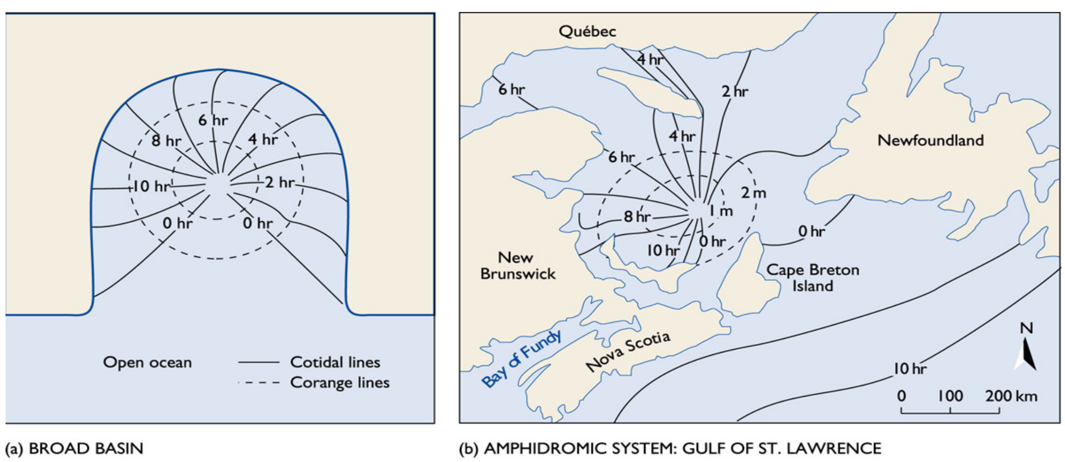


Figure 12.12. A rotary standing wave in a broad semi-enclosed basin such as the Gulf of St. Lawrence.

The shape of the coastline and semi-enclosed basins greatly affect the tide patterns observed in a particular location. Broad semi-enclosed basins tend to experience rotary standing waves. Long and narrow basins tend to experience progressive waves where the tide wave reverses direction between flood and ebb. Tidal range typically increases if a bay tapers landward and water is funneled towards narrow end, as in the case of the Bay of Fundy. Tidal resonance occurs if the period of the basin is similar to the tidal period, which increases tidal range even further, producing extreme tidal ranges of over 56 ft in the Bay of Fundy.

The orientation of semi-enclosed basins also affects the tides. East-West basins allow for a greater movement of the tide crest with the easterly rotation of the earth, and tend to experience higher tides. The Baltic Sea is connected to the Atlantic through an opening too small for the tide wave to move in. The Baltic itself has a north-south orientation and for that reason generates essentially no tides. The Mediterranean has an east-west orientation but has land masses in the middle that defeat the natural period of oscillation. Like the Baltic, the opening at Gibraltar is too narrow for the tide wave from the Atlantic to have much of an effect in the Mediterranean, and tidal range is extremely low (8cm in Malta).

12.4. Tidal Patterns

The various depths and shape of the ocean basins modify tidal patterns that are expected in a simple model (e.g. Figure 12.7). The type of tidal pattern experienced in a given location is a result of the gravitational forces acting on the tide (including the latitude and location relative to position of the moon), and the shape of the basin. There are three types of tidal patterns (Figure 12.13). Diurnal tides have one high tide and one low tide every day, with a period of 24 hours 50 minutes. A semidiurnal tide has two high tides and two low tides of approximately equal heights every day. A mixed semidiurnal tide has two high tides and two low tides of different heights every day. Semidiurnal tides and mixed semidiurnal tides both have a period of 12 hours 25 minutes.

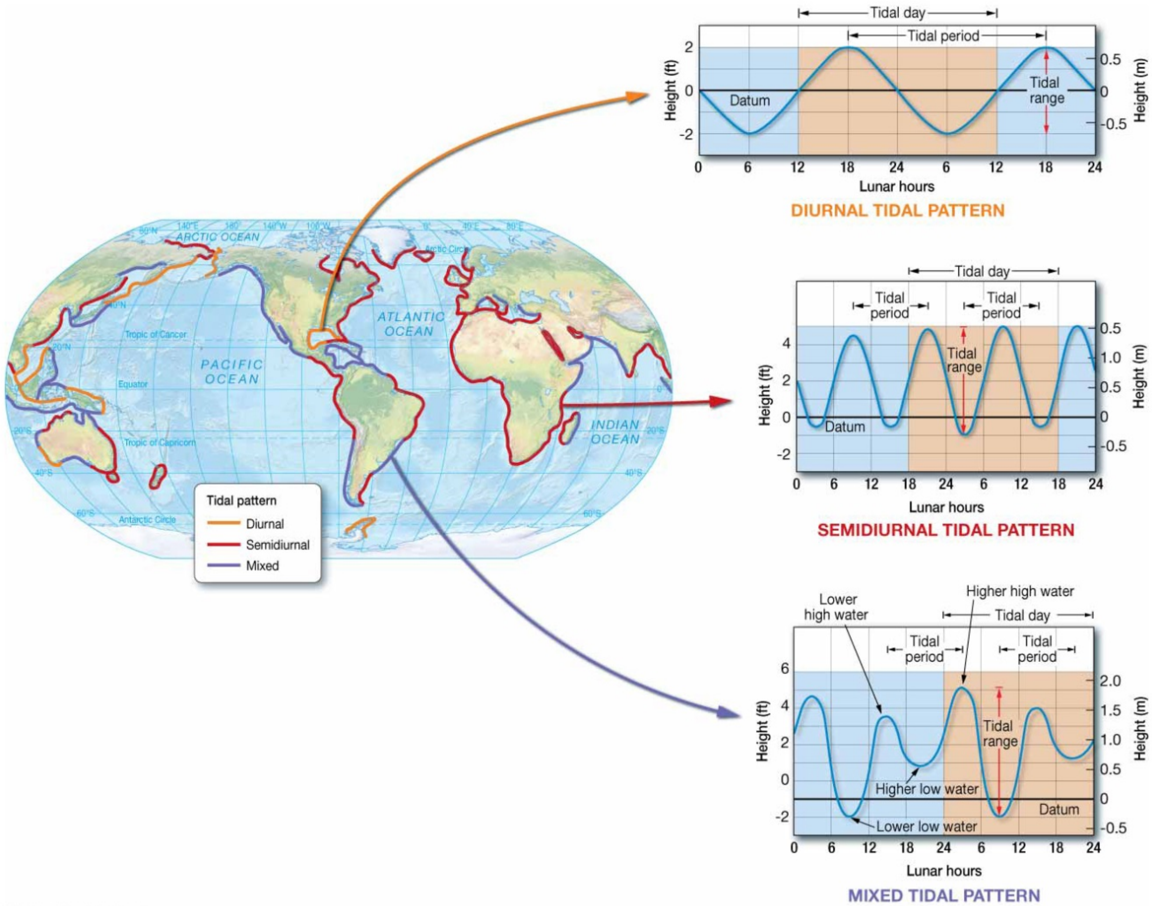


Figure 12.13. Tidal patterns around the world.

12.5. Review Questions

1. Are tides shallow water waves or deep water waves?
2. What is a mixed semi-diurnal tide?
3. What is the length of a tidal day?
4. What is the name of a rising or incoming tide?
5. What three factors cause tides?
6. What is the name of the tides of maximum tidal range, which occur during the new moon and full moon? Is this constructive or destructive interference?
7. Over how many years is mean low water level averaged to calculate tidal datum?
8. What is the tidal range?
9. Why two bulges are formed in the equilibrium tidal theory?
10. Does the sun or the moon have a greater effect in controlling the timing of the tides?
11. Where would you experience the greatest tidal range, of 40-50ft? Why?
12. What is the amphidromic point?
13. What is an antinode?
14. Which direction does a rotary standing wave rotate in the northern hemisphere?
15. Where would you see a rotary standing wave?
16. What force or effect causes a rotary standing wave in some ocean basins?

13. Coast: Beaches and Shoreline Processes (Trujillo, ch.10)

13.1. Coastal Regions

The shore is the zone between the lowest tide and the highest level affected by waves, which reaches some distance above the high tide mark. It is further subdivided into the foreshore and backshore. The backshore is above the high tide mark while the foreshore is the true intertidal: the zone between low tide and high tide. Seaward from the foreshore is the nearshore, from the low tide shoreline to the low tide breaker line. Beyond the breakers is the offshore zone, where water is deep enough that waves don't affect the bottom.

Beaches are depositional areas of the shore. They are composed of the sediment locally available from the erosion of local rocks, those brought by rivers, or from biogenous sediment (e.g. shells or shell fragments). Beaches are dynamic, and waves that crash on the beach are continuously moving material.

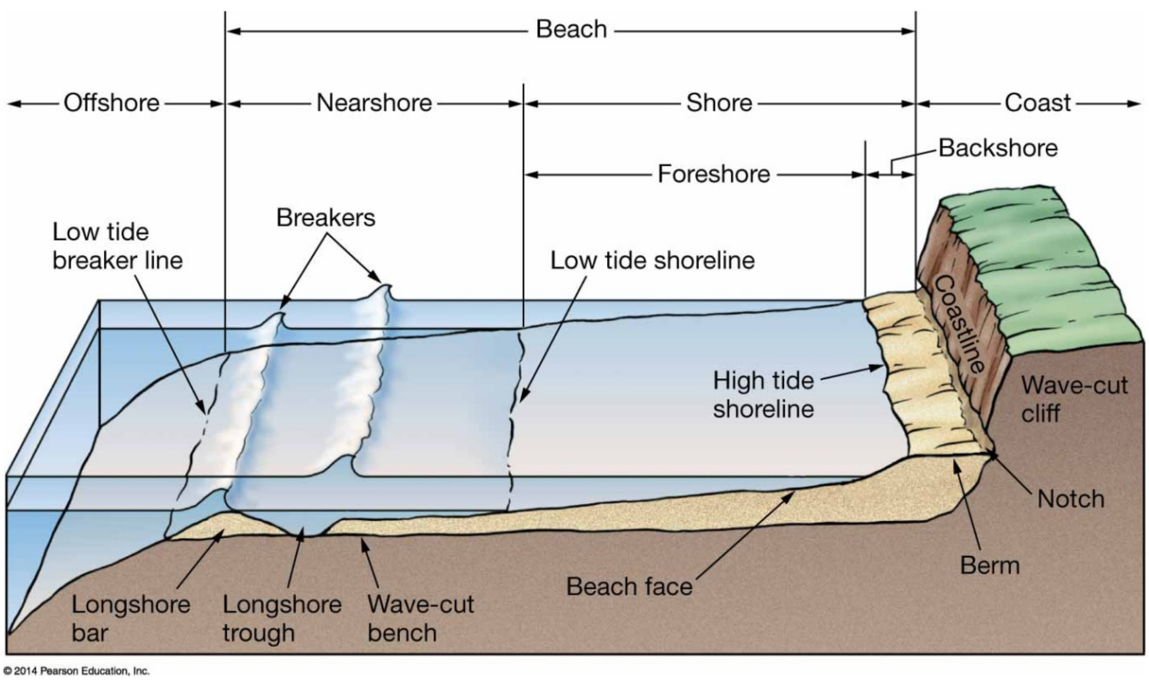


Figure 13.1. Features of a coastal region.

13.2. Movement of Sand on Beaches

There is a continuous movement of sediment on beaches. The net result of this reworking of sand depends on the direction and energy of waves hitting the beach. When waves break perpendicular to shore, sand also moves perpendicularly as a result of the waves breaking: sand moves up the beach in the swash, and drains away from the beach in the backwash.

When the wave energy is low, the swash dominates this transport because much of the water soaks into the beach, thereby reducing the transport associated with the backwash. The result is an increase in sand deposition. When wave energy is high, the beach is saturated with water from previous waves, therefore the backwash is important and sand is transported away from the beach. In many regions of the world, the amount of wave energy is light in the summer and heavy in the winter, resulting in seasonal changes to the beach (Figure 13.2).

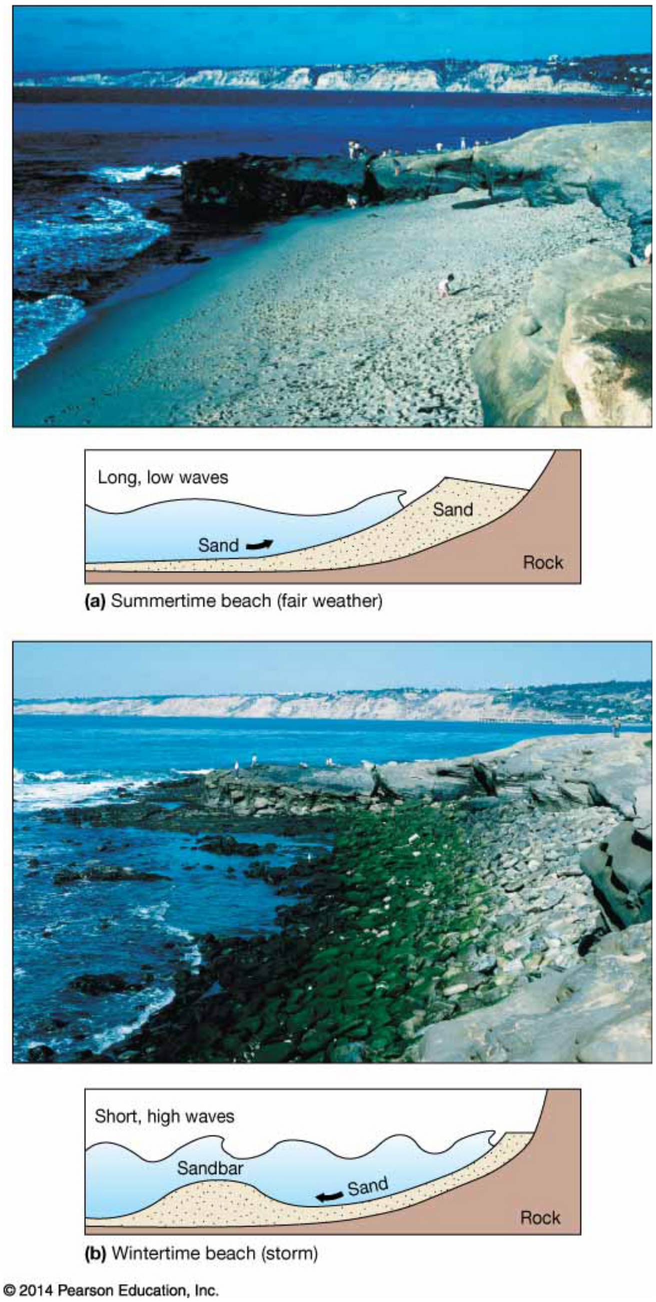
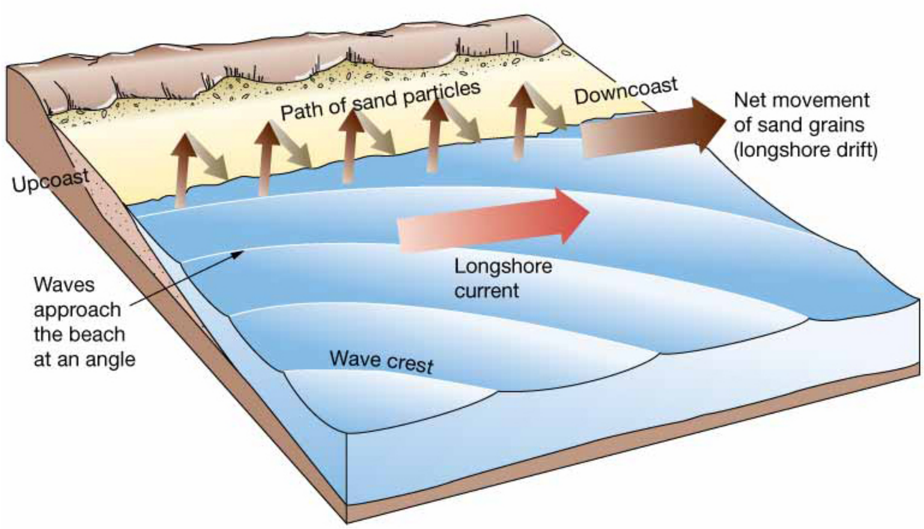


Figure 13.2. Seasonal changes to a beach in California is the result of higher wave energy in the winter than in the summer.

When waves break at an angle relative to shore, the swash moves up the beach at that angle, but the backwash is pulled down by gravity perpendicular to the shore. This creates a current of water parallel to shore called the longshore current. This current moves sand along the shore in the direction the waves are moving, called the longshore drift (Figure 13.3).

(a) Waves approaching the beach at a slight angle near Oceanside, California, producing a longshore current moving toward the right of the photo.



(b) A longshore current, caused by refracting waves, moves water in a zigzag fashion along the shoreline. This causes a net movement of sand grains (longshore drift) from upcoast to downcoast ends of a beach.

Figure 13.3. Waves breaking on shore at an angle create the longshore current, which runs parallel to shore.

13.3. Erosional and Depositional Shores

All shores experience some degree of erosion and deposition, but many can be categorized as being primarily erosional or primarily depositional. Energy is concentrated near headlands that jut out to sea because of wave refraction. Headlands therefore tend to erode and eventually the shoreline retreats (Figure 13.4).

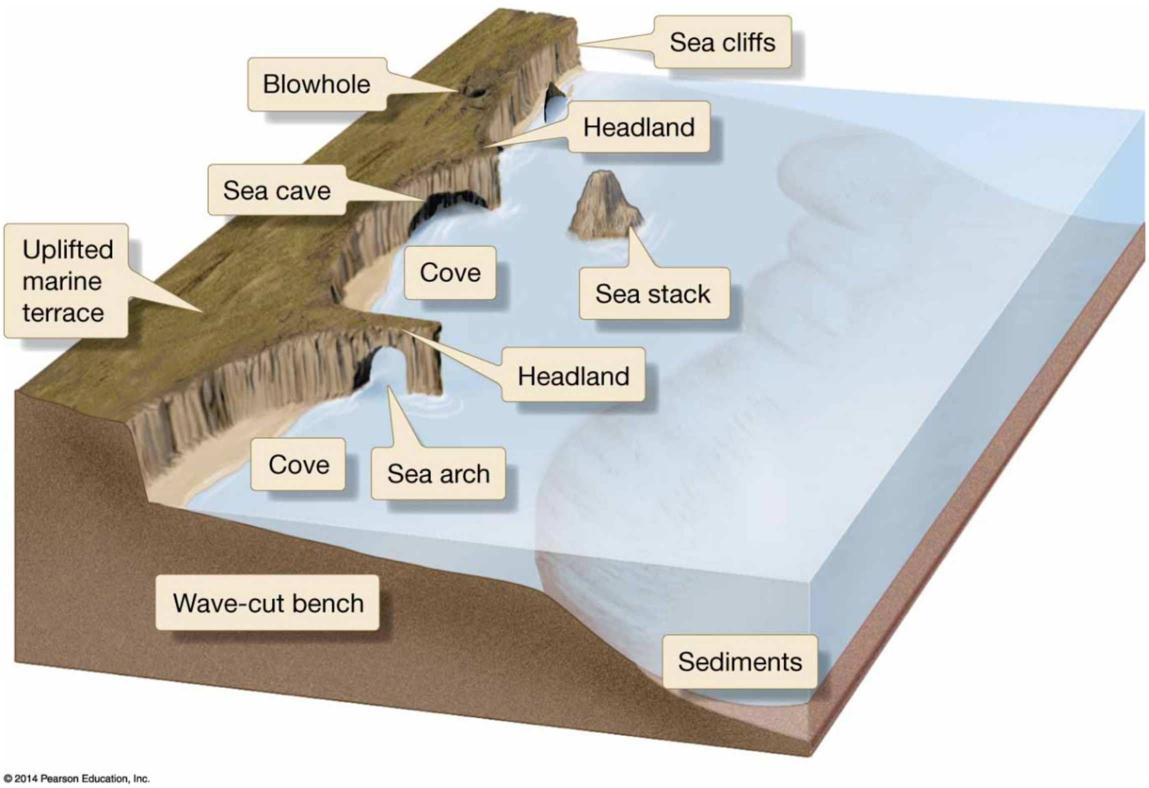


Figure 13.4. Features of erosional coasts.

Erosion of cliffs produces a large amount of sediment, and more sediment arrives at the coast through rivers from the erosion of inland rocks. Those sediments are carried along the coast by water movements and accumulate in low energy environments. Those depositional areas include many types of sand deposits which can be connected to or completely detached from shore (Figure 13.5). Spits are linear depositions of sand in the direction of the longshore drift that extend beyond a headland or entrance to a river. River outflow is often strong enough to keep the entrance open; when it's not, a spit can close off the bay turn in to a bay barrier (Figure 13.6). Tombolos are sand ridges that form in the wave shadow of an island and connect the island to the mainland—usually in the direction perpendicular to the average wave direction. Barrier islands are offshore sand deposits. Many appeared during the rise in sea level associated with the melting of glaciers at the end of the last ice age. Barrier islands are common off much of the US east coast and Gulf of Mexico (Figure 13.7).

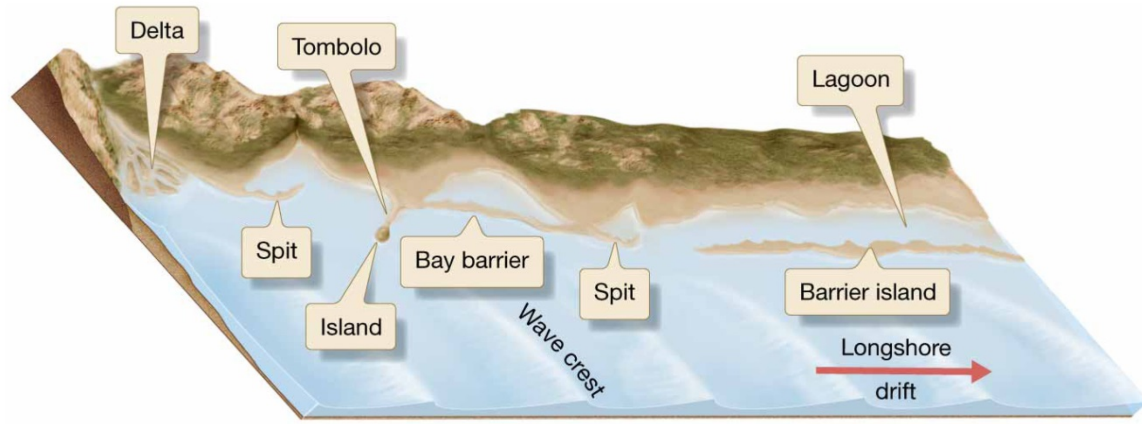
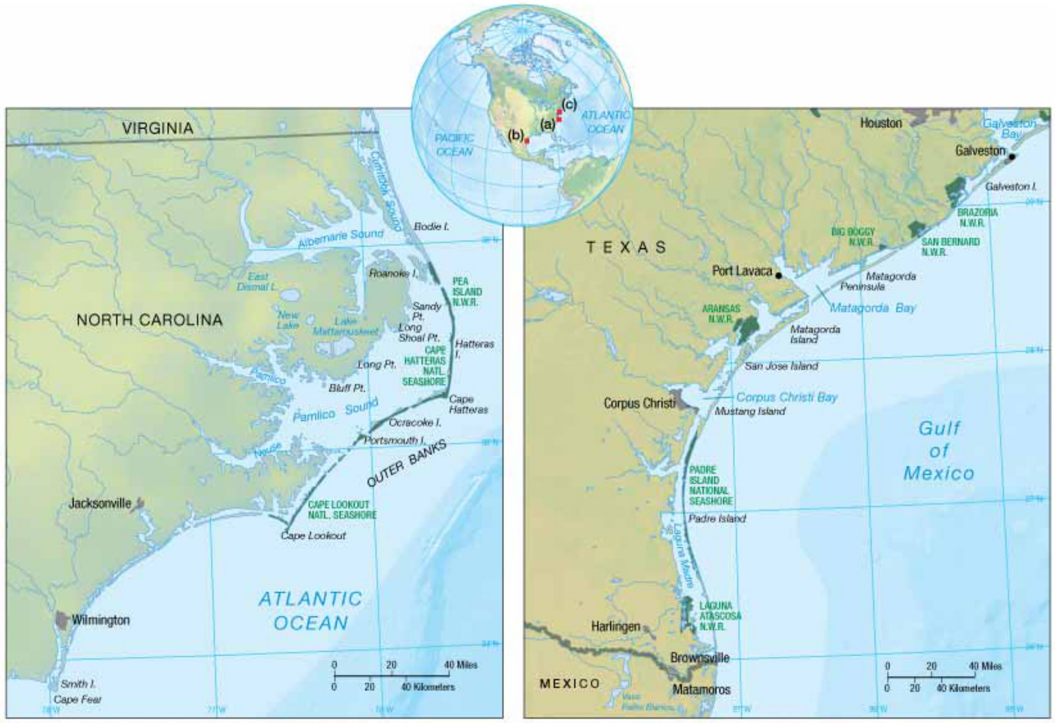


Figure 13.5. Features of depositional coasts.

(a) Barrier coast, spit, and bay barrier along the coast of Martha's Vineyard, Massachusetts.

(b) Tombolo at Goat Rock Beach, California.

Figure 13.6. Spit, bay barrier and tombolos are important features of depositional coasts.

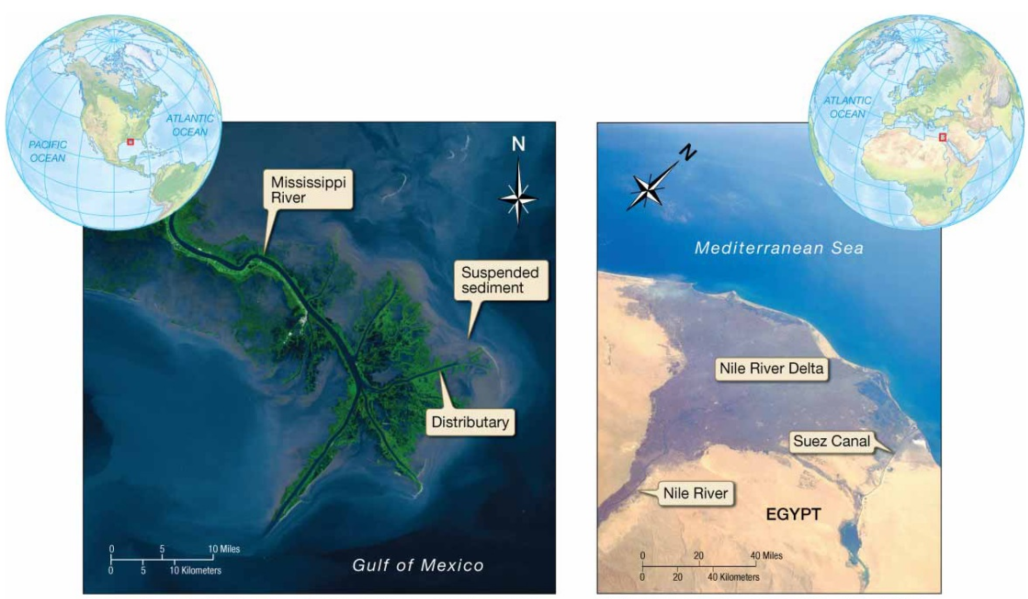


(a) Barrier islands along North Carolina's Outer Banks.

(b) Barrier islands along the south Texas coast.

Figure 13.7. Barrier islands are common along North Carolina and Texas.

Where rivers carry more sediment than is moved by longshore currents, deltas are formed at the river mouth (e.g. Mississippi and Nile deltas, Figure 13.8). Deltas are fertile, low-lying areas that often flood. The strength of the longshore current along with the amount of sediment transported by a river largely determines the shape of a delta.



(a) Satellite image of the branching "bird's foot" structure of the Mississippi River Delta, which flows into the Gulf of Mexico and shows suspended sediment in the water.

(b) Photograph from the Space Shuttle of Egypt's Nile River Delta, which has a smooth, curved shoreline as it extends into the Mediterranean Sea.

Figure 13.8. The Mississippi and Nile deltas.

Coastal areas can often be divided into distinct beach compartments, which consist of 3 components: 1) the rivers that supply the sediment, 2) the beach itself where sediment is transported to due to the longshore current and 3) submarine canyons which transport sand away from the coast. For example, the Southern California coast is comprised of 4 main beach compartments (Figure 13.9). When human activities modify sediment supply to the coast (e.g. through damming of rivers), the beaches that normally receive this sediment can become smaller as their existing sediment is still being swept away by the longshore current and submarine canyons. This process is called beach starvation.

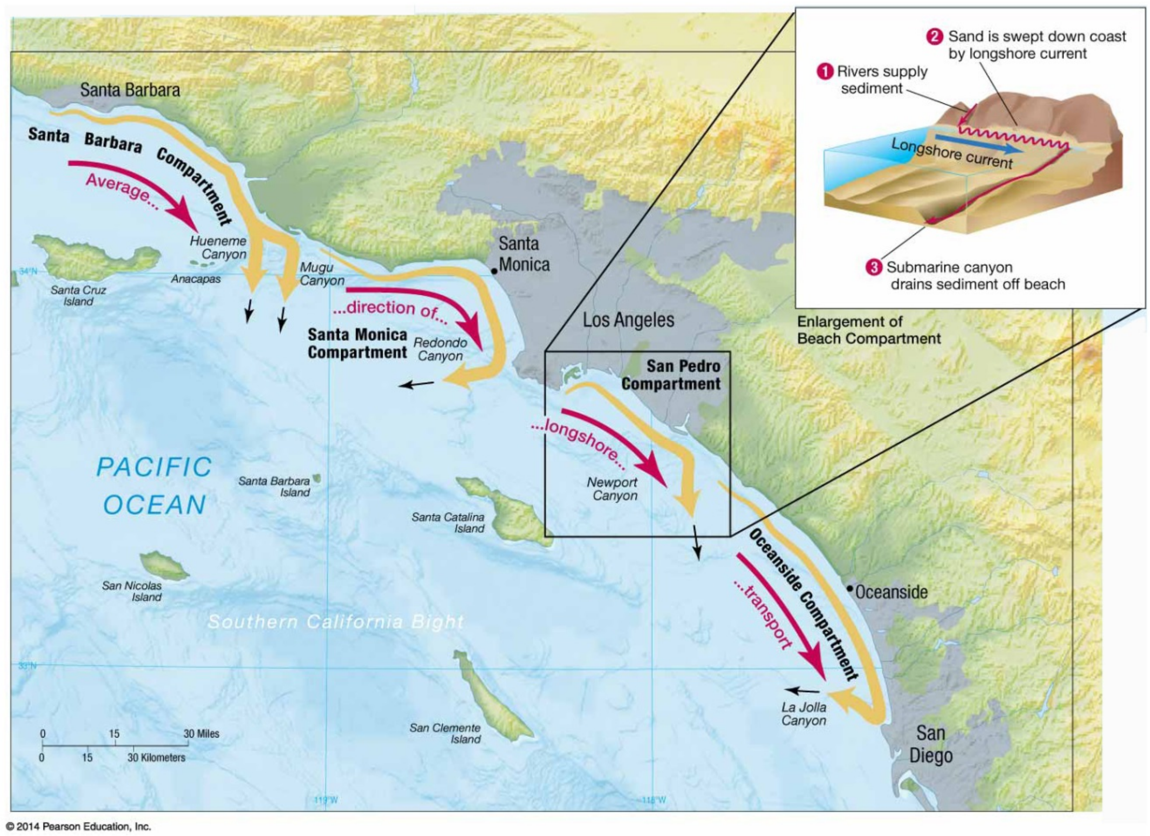


Figure 13.9. Beach compartments of Southern California.

13.4. Changes in Sea Level and Shorelines

Features on shorelines can also be caused by past changes in sea level. Sea level changes through time either through local movements of the earth's crust, or because of world-wide changes in sea level. Tectonic movements responsible for relative sea level changes include uplifting and subsidence of major continents and ocean basins, as well as localized folding, tilting and faulting of continental crust. For example, much of the west coast of North America is currently emerging from tectonic plate collisions. By contrast, the east coast of North America is mostly submerging, as the tectonic plate cools and accumulates more sediment with increased distance from the mid-Atlantic ridge. The Earth's crust also experiences isostatic adjustment, whereby the lithosphere sinks into the asthenosphere under increased weight (e.g. glaciers), and rises up when the weight is lifted (Figure 13.10).

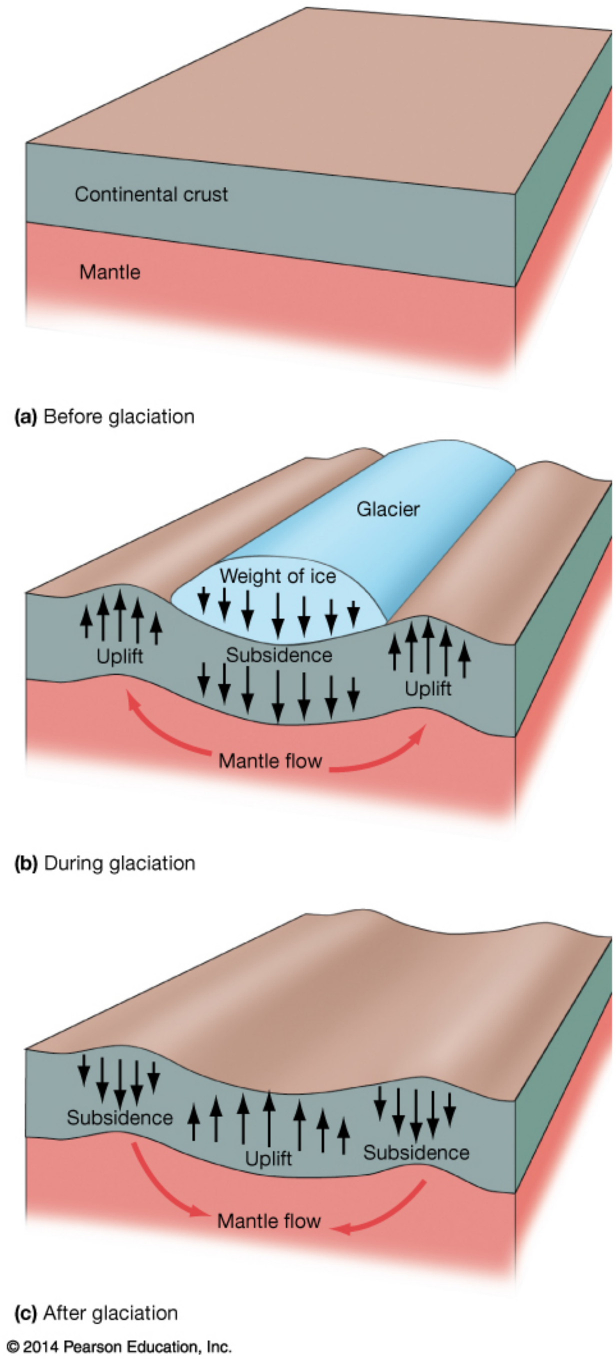


Figure 13.10. Isostatic adjustments caused by a glacier.

Tectonic movements and isostatic adjustments are responsible for localized changes in sea level. Worldwide (eustatic) changes is sea level also occur when there is a change in the amount of water in the oceans, or to the ocean's capacity. These changes can occur through the formation and destruction of large inland lakes, changes in seafloor spreading (which affect the size of mid-ocean ridges) or ice ages.

13.5. Stabilization of Coastlines

Many structures can be built to protect harbors and buildings built near shore (Figure 13.11). Groins are placed on eroding beaches to trap sand. Jetties are placed perpendicular to shore to protect from waves and prevent sediment deposition at the mouth of harbors. Breakwaters are built in front of harbors and shorelines, offshore and parallel to the beach, to redirect wave energy. Seawalls are built on shore to protect from storm waves. Because all these structures modify the longshore drift, they tend to increase erosion downstream and create sand accretion upstream. Beach renourishment consists of supplying beach sand from elsewhere to an eroding beach, but is pricey and is only a temporary solution.

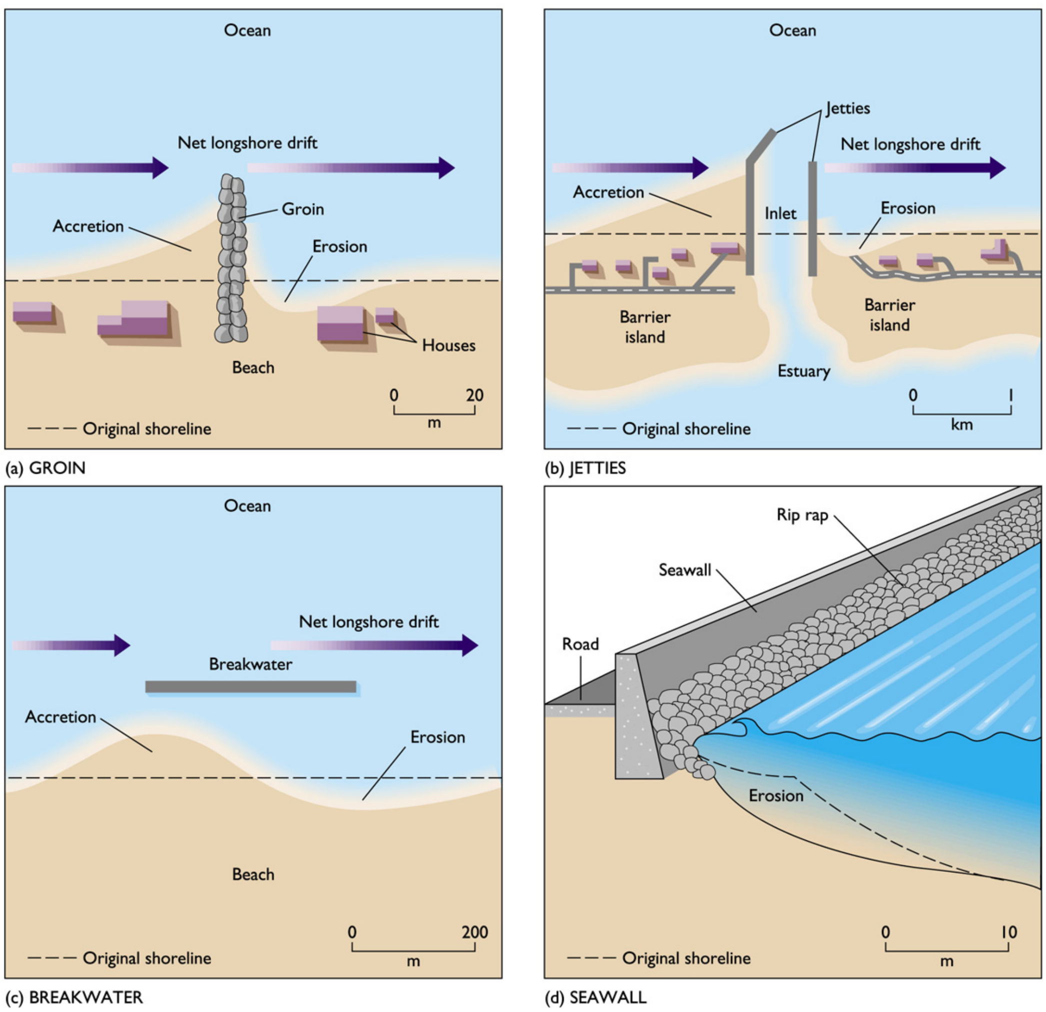


Figure 13.11. Four types of structures built to protect the coast.

13.6. Review questions

1. What is the difference between the terms shore and coast?
2. Where does beach sand come from?
3. How does wave action influence the swash and backwash, and in turn the beach profile?
4. What creates the longshore current?
5. Which kinds of features receive the most wave energy on an erosional coast?
6. What is the difference between a spit and a bay barrier?
7. Which angle do tomolos usually develop, relative to wave direction?
8. What two factors mostly determine the shape of a river delta?
9. What are the 3 components of a beach compartment>
10. Name 2 processes that can lead to localized changes in sea level.
11. Name 3 processes that can lead to eustatic sea level changes.
12. Describe 4 types of structures built to stabilize coastlines?

14. The Coastal Ocean (Trujillo, Chapter 11)

14.1. Laws governing ocean ownership

Historically, coastal zones were claimed by each individual country, typically 3 to 6 miles from the coastline. In the early 1900s, the US discovered oil and mineral resources on its continental shelf, and consequently proclaimed exclusive control over the continental shelf (over 200 nautical miles) in 1945. At the time there was no international agreement to prevent this. In 1958, 86 countries met for the first United Nations Conference on the Law of the Sea (UNCLOS). It took two more meetings and another 15 years to adopt Law of the Sea treaty 1973, which defined ocean boundaries and rights of using oceans. The treaty was ratified by the 60th nation in 1993 and became international law. The US still has not ratified UNCLOS because of concerns over restrictions of seabed mining.

The Law of the Sea treaty specifies how nations can protect their natural resources, settle maritime boundary disputes and extend their rights to resources adjacent to its territory. The four main elements are the following:

1. Coastal zone jurisdiction (Figure 14.1): The treaty establishes the territorial sea of a coastal nation as extending 12nm from its shore. This is considered part of the sovereign territory and countries set and enforce laws (e.g. customs, taxation, immigration and pollution), regulate use, and use any resource. Vessels have a right of innocent passage through the territorial sea. In the contiguous zone (12 nm beyond the territorial sea), countries can continue to enforce laws if the infringement started within the state's territory or territorial waters, or if this infringement is about to occur within the state's territory or territorial waters. The Exclusive Economic Zone extends 200 nm from the coastline, and the country has all rights to the resources in its EEZ. Where the continental shelf extends beyond 200nm, a country has exclusive rights to resources over to 350nm (or 100nm from the 2500m depth mark) but must share 7%.
2. Ship passage: Vessels have right of free passage in the high seas, and within territorial seas and straits used for navigation
3. Deep-ocean mineral resources: mineral resources in the high seas are under the jurisdiction of the UN International Seabed Authority (ISA). Private exploration of the seafloor may proceed under strict control of the ISA.
4. Arbitration of disputes: Should disputes arise on ownership and boundaries of the oceans, they will be settled by a UN Law of Sea tribunal.

In the US, individual states have jurisdiction from the shoreline to a distance of 3 to 9 nm in the territorial sea.

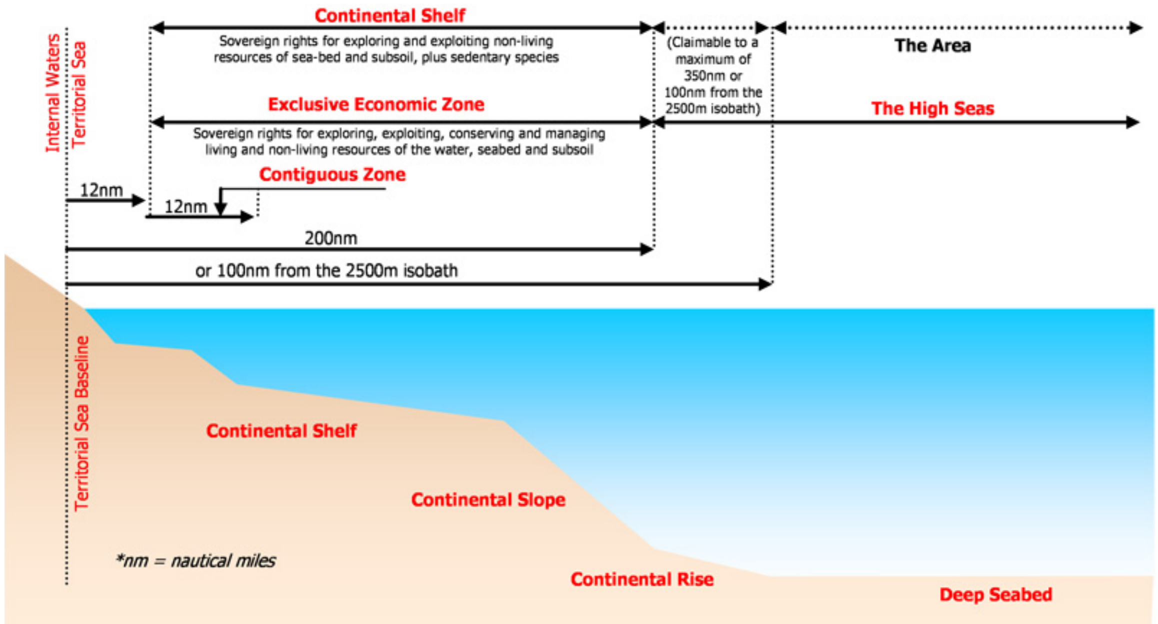


Figure 14.1. Coastal zones as defined by the United Nation's Law of the Sea treaty.

14.2. Characteristics of coastal waters

Coastal waters, above the continental shelf in the margin of continents, are shallower than the open ocean and have strong influences from land such as runoff and tidal mixing. Their physical and biological characteristics are therefore quite different than the open ocean.

Salinity

Freshwater runoff from continents generally lowers the salinity of coastal oceans compared to open oceans. Freshwater that comes from rivers is less dense than seawater. If river runoff occurs in a coastal area with limited vertical mixing, the freshwater layer remains at the surface and forms a halocline (Figure 14.2a). Runoff in a shallow area that has strong wind or tidal mixing simply reduces the salinity of the entire water column (Figure 14.2c). In some coastal regions, prevailing offshore winds can instead increasing evaporation and therefore increase surface salinity, resulting in a reverse halocline (Figure 14.2b).

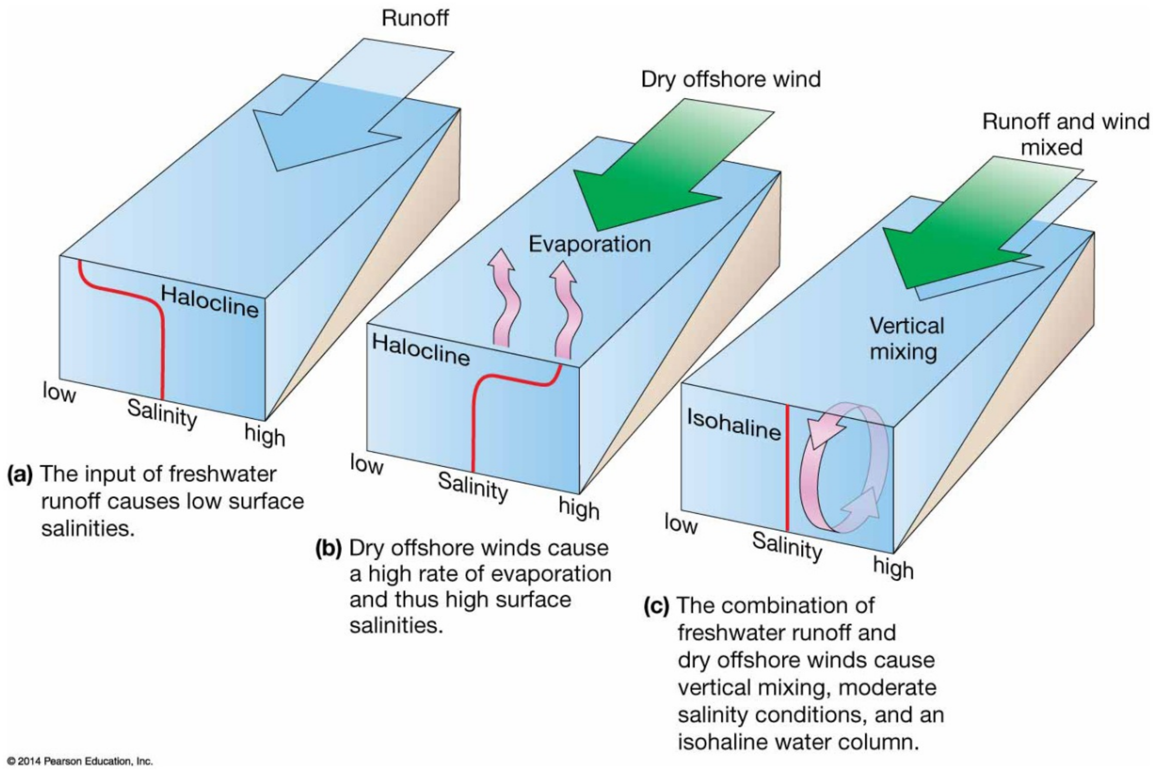


Figure 14.2. Variation in the salinity profile of the water column in the coastal ocean.

Temperature

The vertical temperature profile in any given location depends on how much surface waters are heated by the sun, and how much vertical mixing exists. High latitudes tend to have uniformly near-freezing temperature throughout the water column, with no thermocline at any point during the year (Figure 14.3b). Low latitudes receive a lot of solar heat, which often creates a permanent thermocline around 300-1000m in the open ocean (chapter 6). However, in shallow coastal areas where circulation with the open ocean is restricted, the entire water column may become uniformly warm (Figure 14.3a). In coastal zones in mid-latitudes, a thermocline is established in the summer when surface waters are heated by the sun. This thermocline breaks down in the fall and may even be reversed for short periods of time in the winter, when surface water might be colder than deeper water (Figure 14.3c).

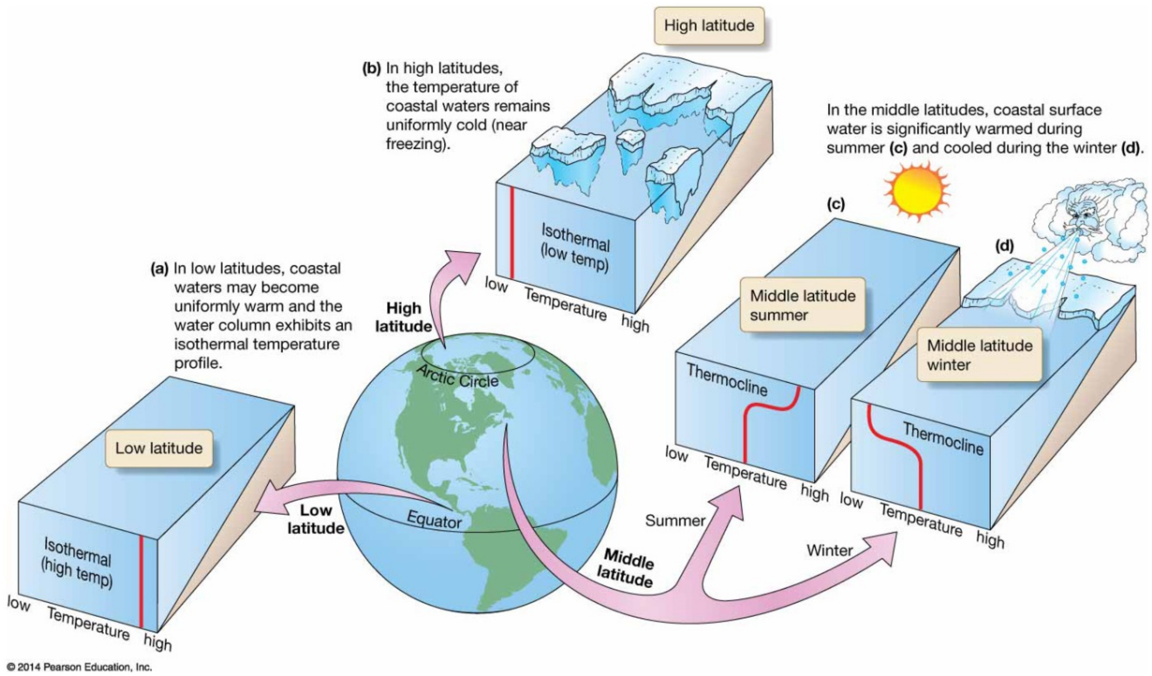


Figure 14.3. Temperature profile in coastal waters at various latitudes.

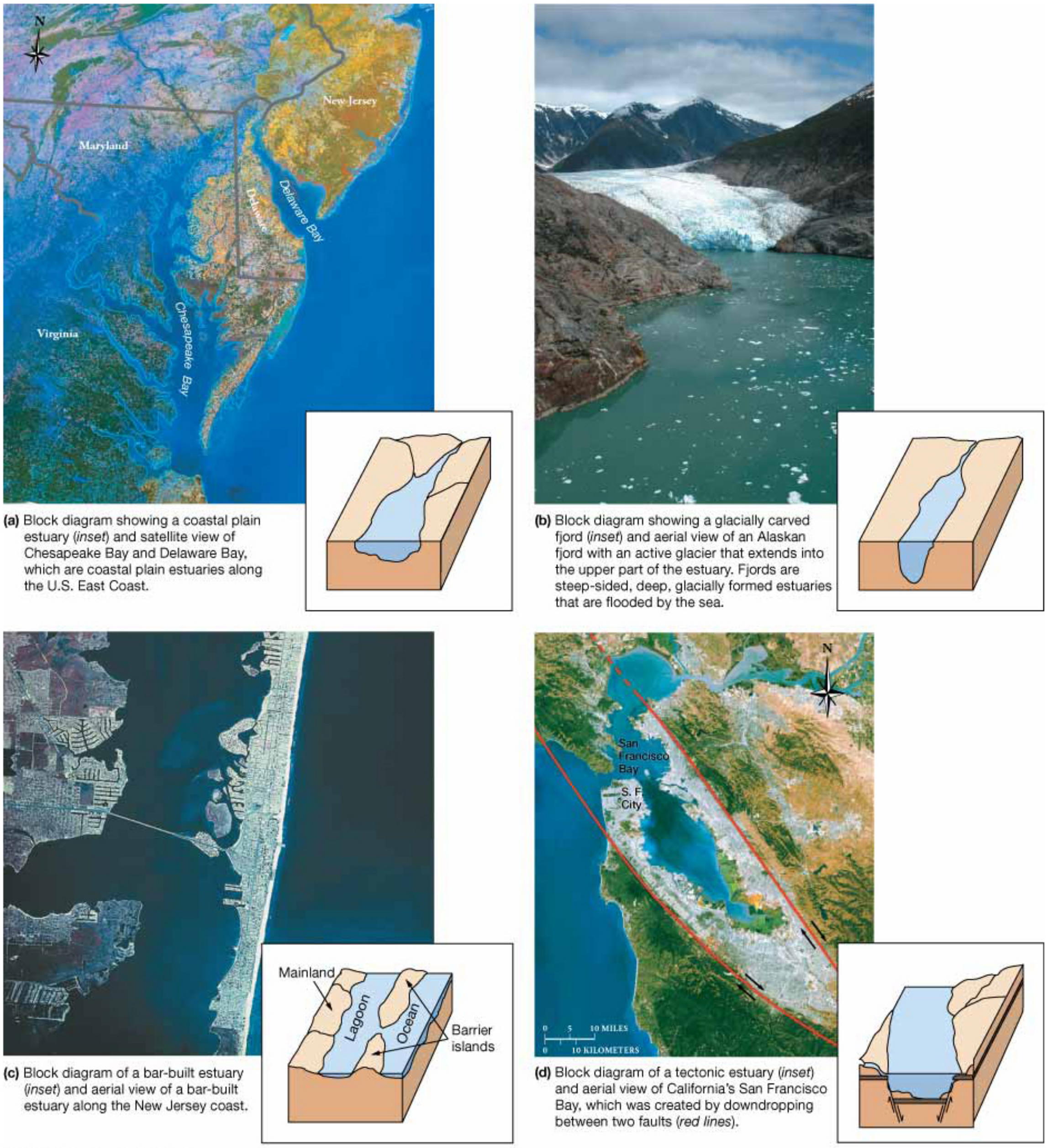
14.3. Types of coastal waters

The shape of continents and coastal zones creates several distinct types of coastal waters, including estuaries, lagoons and marginal seas.

Estuaries

Estuaries are partially enclosed coastal bodies of water where fresh and salt water mix. They are unique ecosystems that tend to be unstable because of the dynamic mixing of fresh and salt water. They are often highly productive environments because of high nutrient inputs from land. They however typically exhibit a low diversity because of the variability in salinity which few species can withstand.

Estuaries can be classified based on their geologic origin (Figure 14.4). Coastal plain estuaries (sometimes called drowned river valleys), formed when sea level rose and flooded existing river valleys. Fjords are U-shaped valleys with steep walls that were formed by glaciers and later flooded with sea level rise. Bar-built estuaries are formed by barrier islands deposited parallel to the coast. They are typically shallow. Finally, tectonic estuaries are created when tectonic movements drop a section of lithosphere which subsequently floods.



(a) Block diagram showing a coastal plain estuary (*inset*) and satellite view of Chesapeake Bay and Delaware Bay, which are coastal plain estuaries along the U.S. East Coast.

(b) Block diagram showing a glacially carved fjord (*inset*) and aerial view of an Alaskan fjord with an active glacier that extends into the upper part of the estuary. Fjords are steep-sided, deep, glacially formed estuaries that are flooded by the sea.

(c) Block diagram of a bar-built estuary (*inset*) and aerial view of a bar-built estuary along the New Jersey coast.

(d) Block diagram of a tectonic estuary (*inset*) and aerial view of California's San Francisco Bay, which was created by downdropping between two faults (*red lines*).

Figure 14.4. Types of estuaries based on geologic origin.

Estuaries can also be classified based on their salinity profile, which depends on the input of fresh water, the depth, and the slope of the bottom (Figure 14.5). The type of estuarine circulation and position of isohals (lines of equal salinity) may vary temporally with several factors including tide, season, rainfall, and melting of snow. Vertically mixed estuaries are shallow and have a strong outflow of water. Salinity increases from the head to the mouth of the estuary, and at any point is uniform from surface to bottom. In slightly stratified estuaries, the water column is slightly layered with saltier water moving towards the head at depth and a less salty, less dense layer of water moving in the opposite direction at the surface. This is called an estuarine circulation pattern. There is considerable mixing between layers and isohals are diagonal. In highly stratified estuaries, the estuarine circulation pattern is stronger and there is less mixing between layers. Deep water can reach full ocean salinity and the zone of mixing is narrow with a strong halocline. Salt wedges tend to develop near high volume rivers, which limits the intrusion of seawater at depth and results in a surface layer of freshwater reaching far from the head.

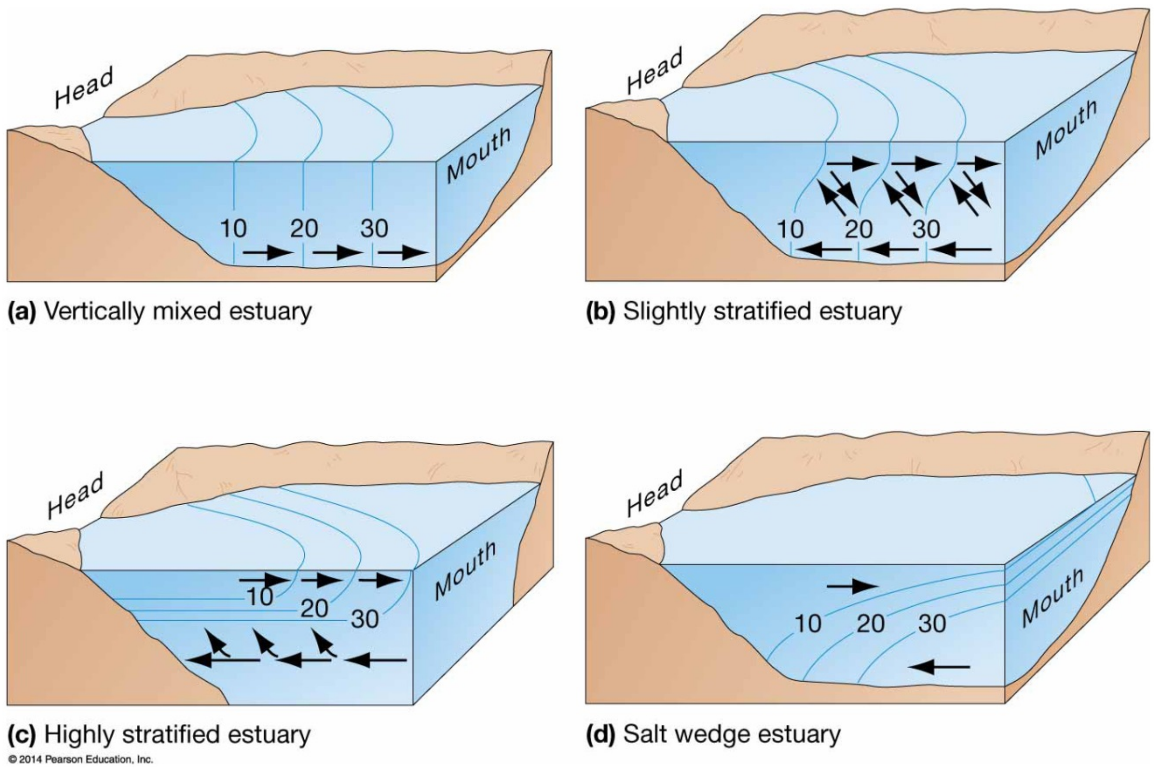


Figure 14.5. Four major types of estuarine circulation.

Lagoons

Lagoons are a special type of estuary that form behind barrier islands (Figure 14.6). They are shallow and protected, and have limited exchange with the ocean, which results unique physical characteristics. Three distinct zones can be identified in lagoons. The freshwater zone is near the head of the lagoon where freshwater enters; the transitional zone in the middle of the lagoon contains brackish water, and the saltwater zone near the mouth has close to full ocean salinity. Tidal variation is highest near the mouth. In some dry areas, lagoons can experience such a high level of evaporation that they become hypersaline. In this case, ocean water flowing into the lagoon is less dense than the high salinity water in the lagoon, and the layering and circulation pattern is inversed that usually found in estuaries (Figure 14.7). Such lagoons (or other estuaries) that lose more water to evaporation than gain it from runoff are called negative estuaries. Laguna Madre in Texas is an example of a negative estuary.

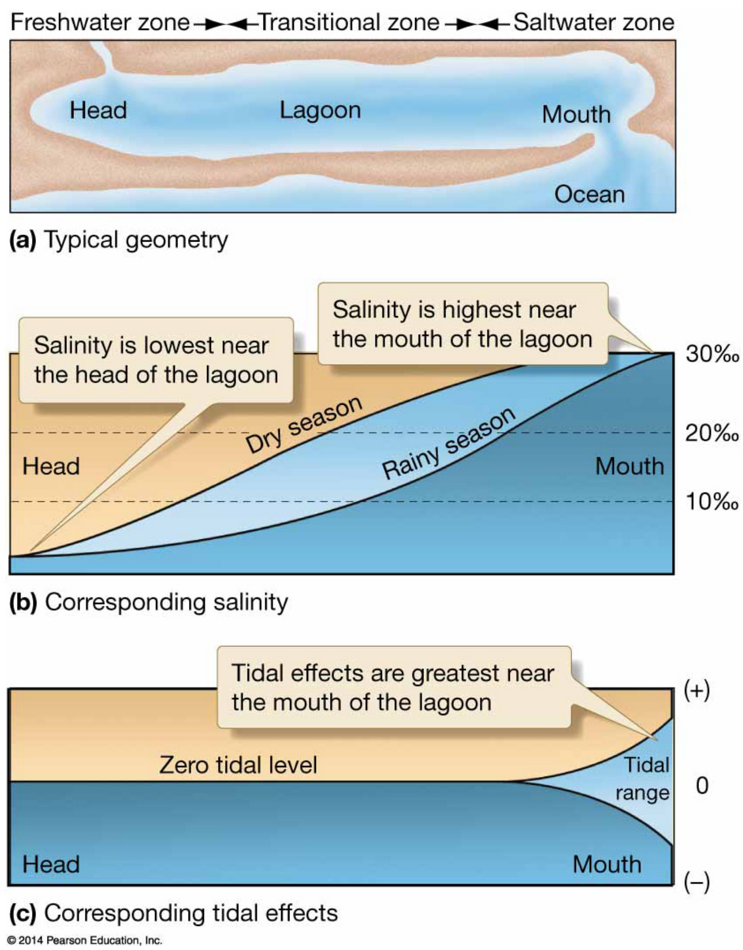


Figure 14.6. Typical characteristics of lagoons.

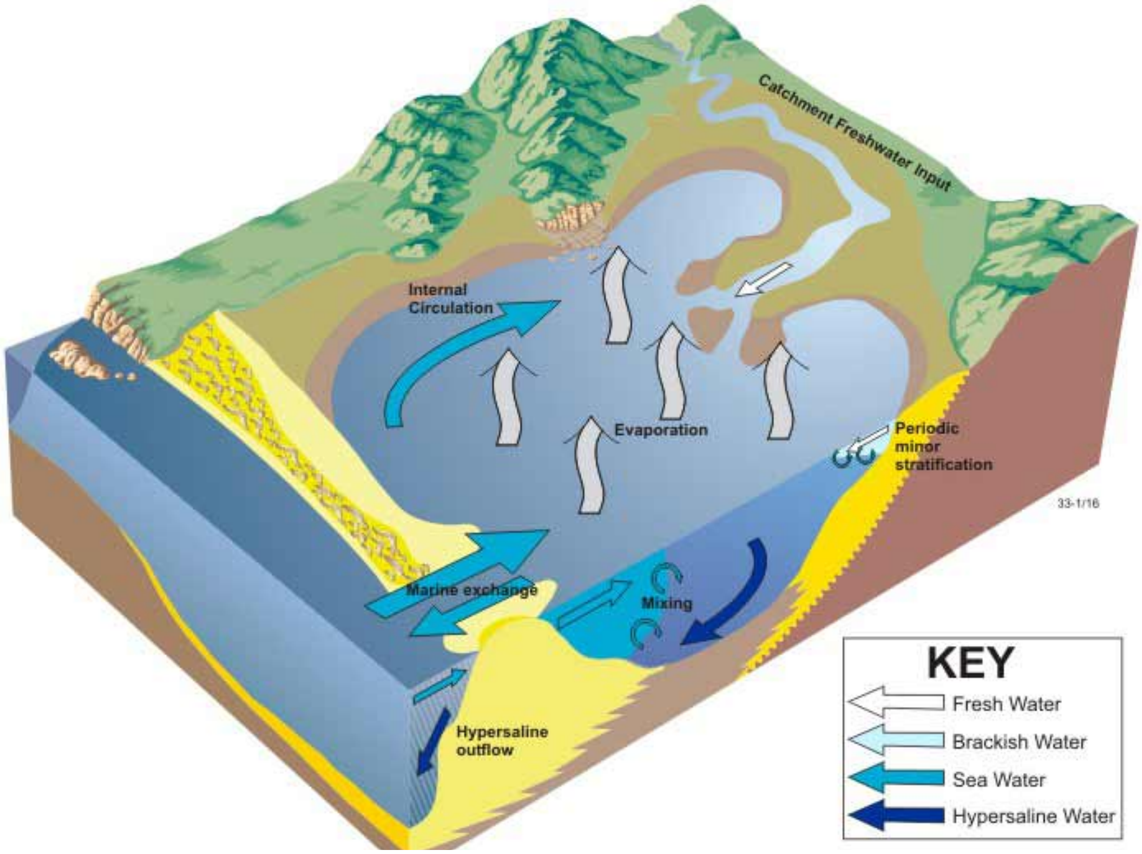


Figure 14.7. Circulation in a negative estuary, such as Laguna Madre.

Marginal Seas

Marginal seas are large, semi-isolated bodies of water such as the Mediterranean and the Caribbean. Waters of marginal seas are relatively shallow and have limited exchange with the open ocean which leads to unique temperature and salinity characteristics. In the Mediterranean, as in Laguna Madre, evaporation exceeds precipitation leading to the formation of high-salinity, high-density water. This water is denser than Atlantic water at the Straits of Gibraltar, and the result is a similar circulation pattern as what is found in negative estuaries; seawater coming from open ocean flows on top of denser, saltier water leaving the Mediterranean (Figure 14.8).

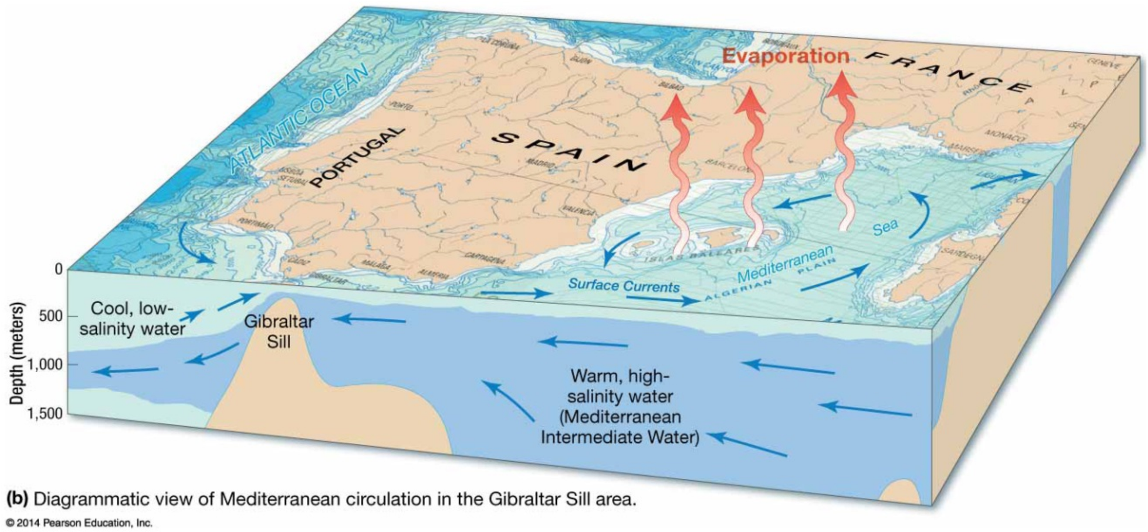


Figure 14.8. Circulation in the Mediterranean, opposite that found in estuaries.

14.4. Coastal wetlands

Wetlands are ecosystems that are typically saturated with water. They can border freshwater or saltwater. Coastal wetlands occur around the margins of coastal waters such as estuaries and lagoons. The two most important types of coastal wetlands are salt marshes and mangroves. Salt marshes occur in temperate zones while mangroves are tropical. Both occur in low energy environments, have oxygen-depleted muds and accumulations of organic matter, and are dominated by terrestrial plants that have adaptations for intermittent submersion in saltwater. Wetlands are important nursery grounds for diverse fish and invertebrates, protect shorelines from erosion during storms, and filter pollutants and excess nutrients. Unfortunately, wetlands around the world are being lost at an alarming rate due to coastal development, agriculture and aquaculture.

14.5. Marine pollution

Marine pollution refers to harmful substances released in the marine or estuarine environment, which may impact marine life, human health and human activities in the sea. It is often difficult to assess the degree of harm potentially caused by a substance. Environmental bioassays are a commonly used technique, in which experiments are run to determine the concentration of a particular substance that causes 50% mortality in a group of organisms. Yet environmental bioassays have several limitations, including the difficulties in extrapolating results to long-term effects, and to groups of organisms other than the one(s) used in the bioassay.

14.6. Main types of marine pollution

Pollution in the marine environment includes a variety of substances that are discussed in this section. Inputs of marine pollution mostly come from land-based activities in the form of runoff, discharges and emissions (Figure 14.9).

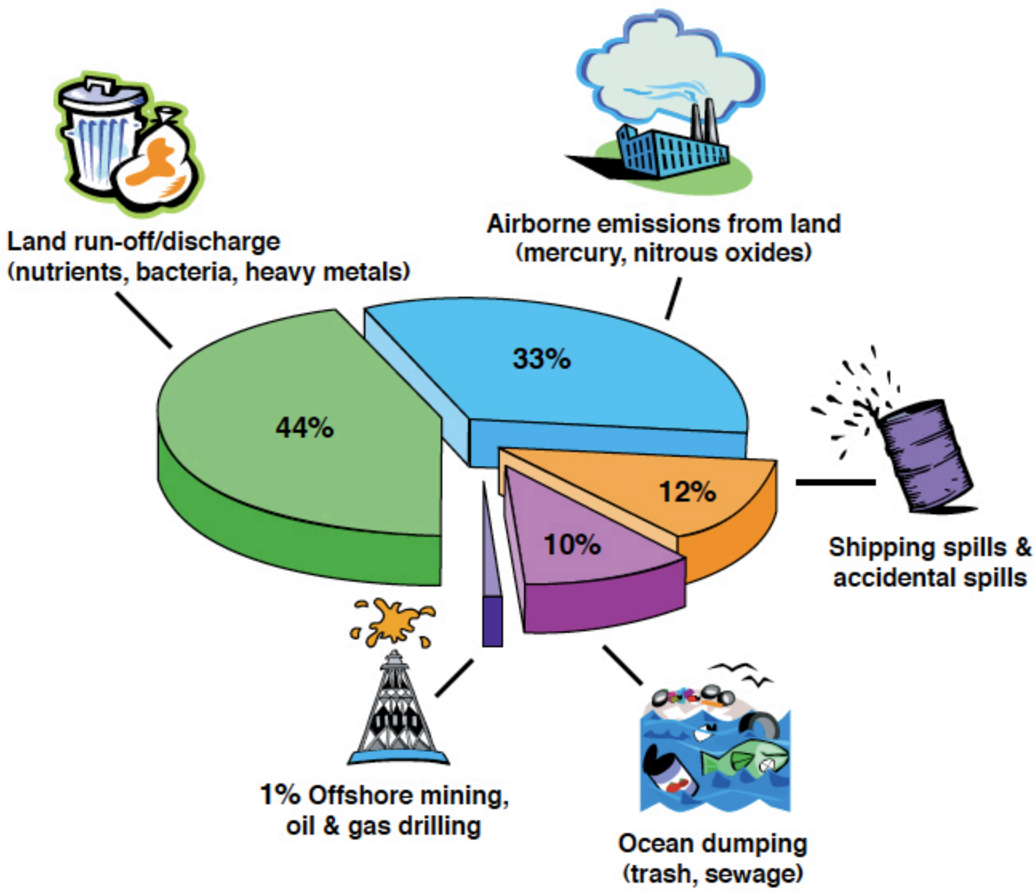


Figure 14.9. Sources of marine pollution.

Petroleum
Major oil spills contribute a large amount of petroleum to the marine environment, yet it should be noted that substantial oil pollution comes from many sources river runoff of oil washed off streets and parking lots (Figure 14.10).   Oil is a mixture of hydrocarbons and other substances. The most toxic compounds in crude oil include polycyclic aromatic hydrocarbons (e.g. napthlenes, benzene, toluene and xylenes), which can lead to disease in plants and animals in very small doses. They can also act as carcinogens when inhaled or ingested. Oil spills have been linked to mortality in fish eggs and many other small large marine organisms. Refined oil is much more toxic than crude oil.

Oil is less dense than water and when it is released in the ocean it floats at the surface. Volatile compounds can evaporate and leave behind a denser substance that can sink. Some oil can also mix with water. Photo-oxidation and bacteria break down the oil into water-soluble compounds. Floating oil can be skimmed and collected, though this collected oil still poses a disposal problem somewhere else. Several types of bacteria and fungi can break down a large proportion of the hydrocarbons in crude oil, and are therefore used to help clean up spills in a process called bioremediation. Other methods of cleaning up oil spills (e.g. adding chemical dispersants) are often more damaging to the environment than they are beneficial.

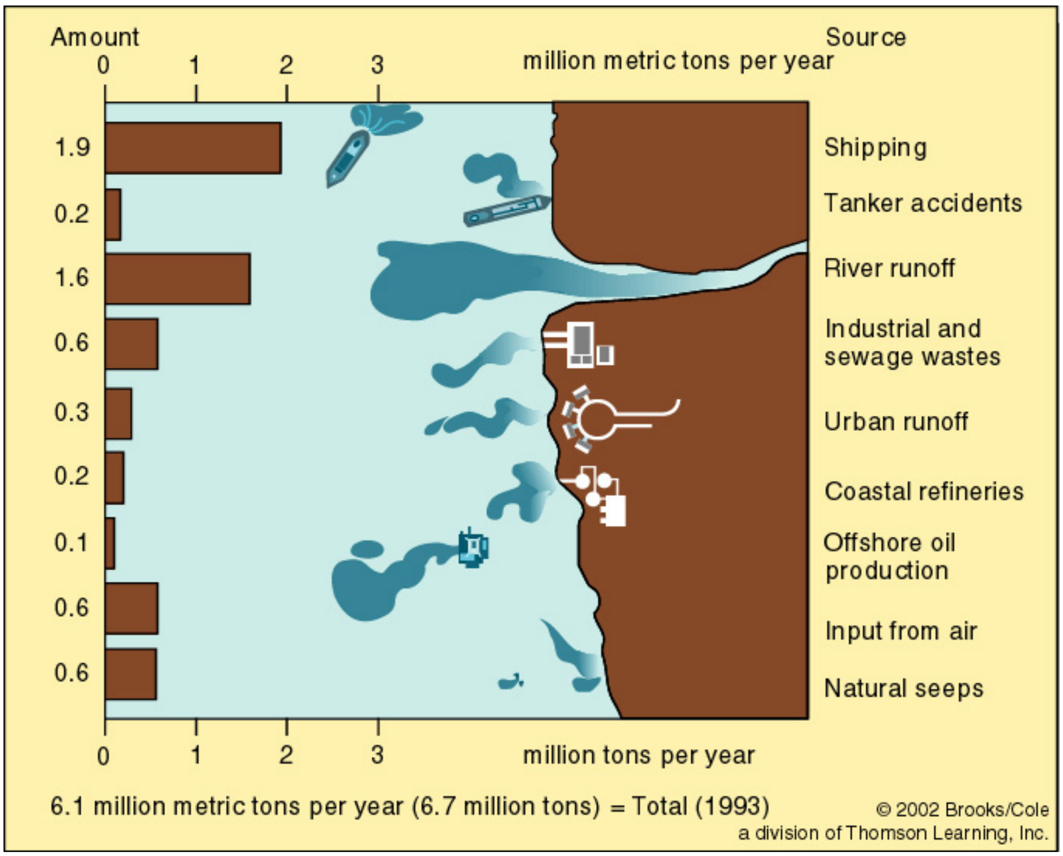


Figure 14.10. Main sources of oil pollution.

Because oil spill clean-up is difficult, the best strategy remains to prevent spills in the first place. In the US, the Oil Pollution Act of 1990 mandates that all oil tankers traveling in US waters be constructed with double hulls (Figure 14.11) by 2015. This aims to prevent a spill even in the case of ship grounding. The Oil Pollution Act also defined fiscal responsibility for clean-up and created the national Oil Spill Liability Trust Fund, which is available to provide up to one billion dollars per spill incident—to insure a rapid response.

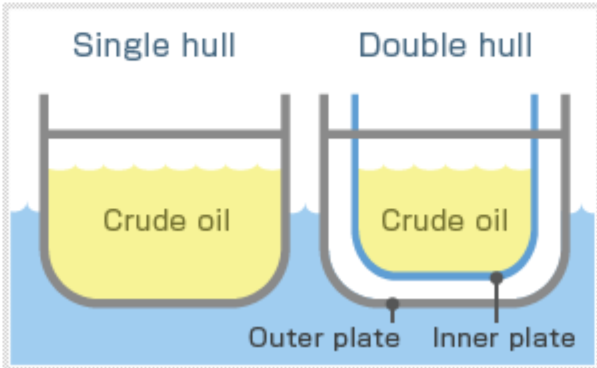


Figure 14.11. Cross-section of oil tankers with single hull and double hull construction.

Sewage sludge

The US Clean Water Act (1972), specifies that all sewage must have at a minimum a secondary treatment before it is discharged in the environment. The primary treatment removes half the suspended organic material and solids; the secondary treatment degrades the remaining solids. What remains is called sludge. Sludge is a semisolid mixture of human waste, oil, zinc, copper, lead, silver, mercury, pesticides and other chemicals. Sludge has commonly been released in the marine environment, some distance from the coastal zone (either transported by barge or pipes). Most municipalities now treat sludge to be released on land, yet in many areas including the Florida Keys, sludge still reaches coastal ecosystem with damaging consequences. For example, nutrients found in sludge can lead to coastal eutrophication (see chapter 16), and some coral diseases have been associated with exposure to human waste.

DDT and PCBs

The pesticide DDT and the chemicals PCBs are persistent chemicals that were released in the environment through human activities. Even though they have been banned in most countries, they are still found in many marine organisms around the globe. DDT and PCBs, along with some other chemicals, have been called persistent organic pollutants. DDT was used as a pesticide in agriculture during the 1950s. It is toxic not only to insect pests but also to many other organisms, including marine birds. Birds contaminated with DDT lay eggs with thin shells which have low survival. PCBs were used in industrial processes and have been shown to cause cancer and affect reproduction.

Mercury

Mercury is used in several industrial processes. When it enters the environment, it is converted to methylmercury, which is highly toxic to many organisms including humans. Mercury is known to affect the human nervous system and cause blindness, tremors, brain damage, birth defects, paralysis and even death. Mercury, along with many other toxins, can be concentrated in the tissue of organisms in a process called bioaccumulation. As these organisms are consumed by animals higher and higher up the food chain, the toxins magnify with each trophic level in a process called biomagnification (Figure 14.12). Top predators therefore have high amounts of toxins such as mercury. This is why US Food and Drug Administration advises that pregnant women, women of child-bearing age, nursing mothers and young children should avoid eating predatory fish like tuna and swordfish which tend to contain high levels of methylmercury. Methylmercury is not only found in areas close to sources of mercury; high methylmercury levels have also been found in marine mammals in the Arctic Ocean. This is of high concern for indigenous Inuit populations that rely on seals and other marine mammals as an important food source.

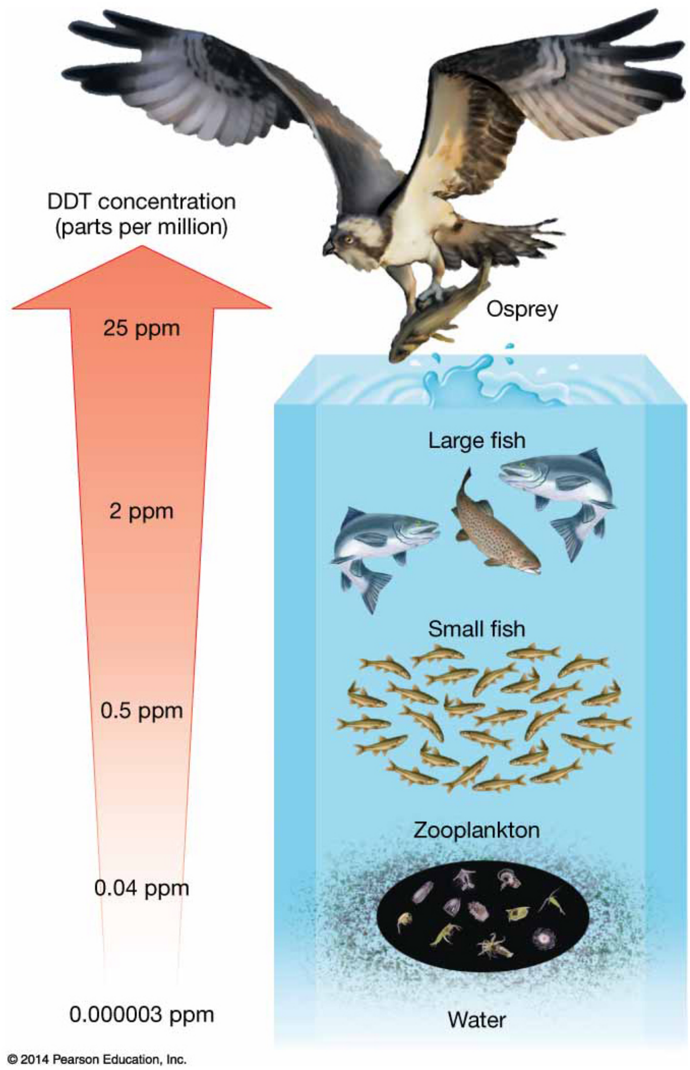


Figure 14.12. Biomagnification occurs when toxins become increasingly concentrated in organisms further up in the food chain.

Nonpoint source pollution and trash

Nonpoint sources of pollution refers to any kind of pollution that enters the ocean from multiple sources rather than one, single source. It includes much of the pollution that enters the oceans as runoff, since the actual source of it can be difficult to pinpoint. This runoff includes a lot of plastic and other trash that wash down in storm drains and rivers, along with pesticides, fertilizers, and small amounts of oil released from cars. Trash can also enter the oceans as a result of intentional dumping. Ocean dumping is regulated by the International Convention for the Prevention of Pollution from Ships (MARPOL) which was put in effect in 1983. MARPOL addresses the release of waste generated during normal vessel operations, regulates oil discharge from ships, sets limits on how far offshore sewage may be dumped, and addresses ship's trash (Figure 14.13).

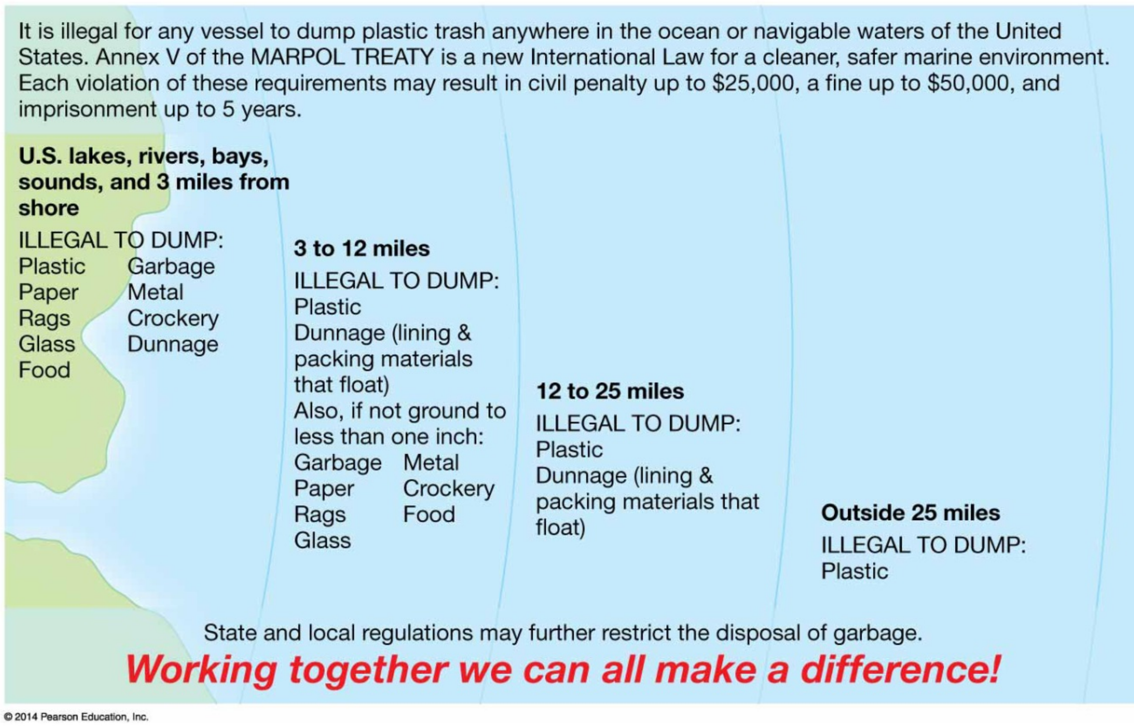


Figure 14.13. Zones defined by the International Convention on the Prevention of Pollution from Ships (MARPOL).

Much of the trash that arrives in the oceans sinks or biodegrades. One exception is plastic, which makes up 80% of the marine debris that comes from land, and does not readily biodegrade. As a result, plastic can remain in the ocean indefinitely. Plastic can impact marine life through entanglement and ingestion. Entanglement can become life-threatening in various ways, including through strangling. Marine birds have been documented to ingest so much plastic that it fills their stomachs, and the birds later die from starvation. Turtles have also died after eating plastic bags. Also important is the problem of toxicity associated with plastic trash, since plastic has a high affinity for toxic compounds that are not water-soluble. While much of the plastic trash in the ocean is clearly visible, a large proportion is made up of very small pre-production plastic pellets called nurdles. Nurdles are transported on vessels and commonly spill in the marine environment. Many beaches around the world have a high concentration of plastic nurdles along with mineral sediment.

Plastic is increasingly abundant in the oceans, and concentrates where currents converge. Large floating patches of plastic debris have recently been documents in the middle of subtropical ocean gyres. In these patches, plastic debris far outnumbers marine life. The most well-known of these patches occurs in the eastern North Pacific, in an area roughly twice the size of Texas (Figure 14.14). Plastic that remains at the surface of the ocean for extended periods of time eventually photodegrades in the sunlight, and breaks down in increasingly smaller components. This facilitates ingestion by more organisms. In response to the problem of plastic pollution, MARPOL bans all plastic dumping at sea (Figure 14.13).

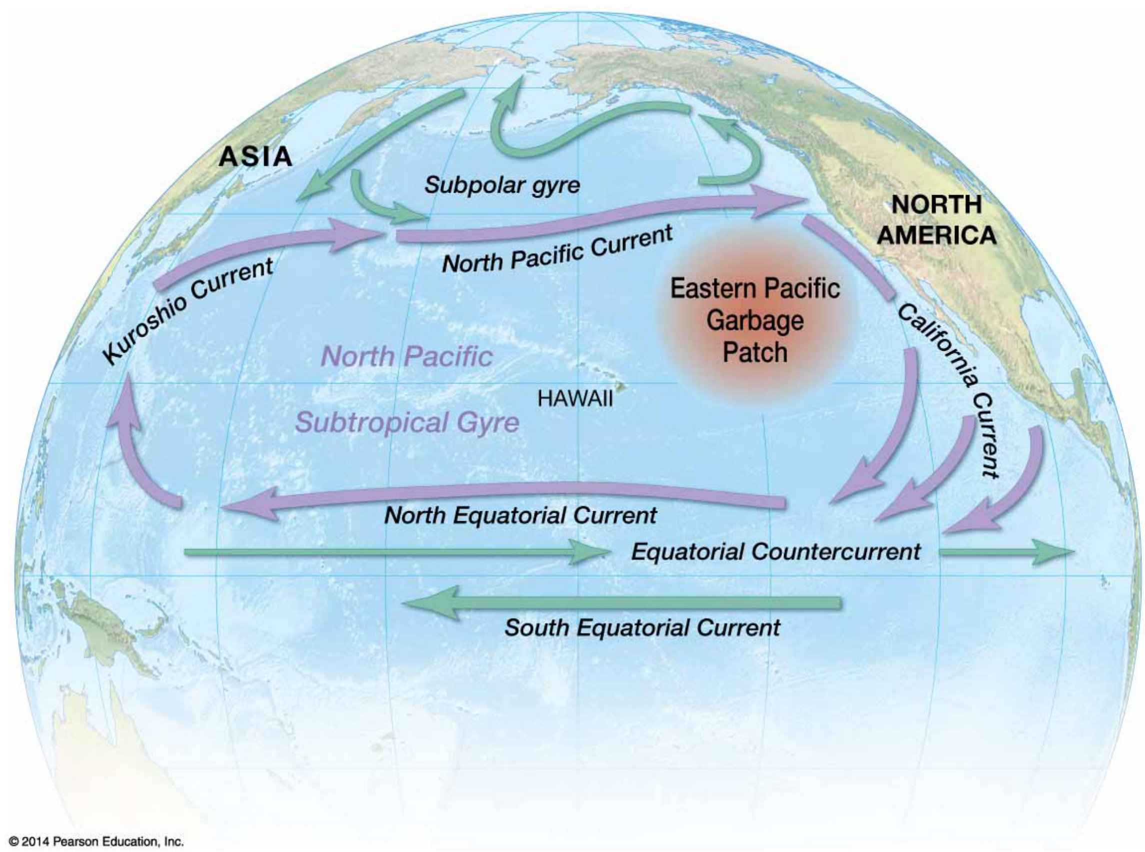


Figure 14.14. The Eastern Pacific Garbage Patch

Biological pollution: non-native species

Non-native species, or introduced species, are species that have been brought to a new area by human activity, either intentionally or accidentally. Organisms are transported to new environments through aquaculture, for recreational fishing, released form aquariums, on ships' hulls, and most importantly as larval stages in ships' ballast water. Though only a small proportion of introduced species are successful, those that become established can have a devastating effect on the local ecosystem as they compete with native organisms, and often have no predators to control their population. Those non-native species that cause negative consequences are called invasive species. There are many examples of marine invasive species that have cause vast ecological changes in their new location. For example, the algae *Caulerpa taxifolia*, native to Asia, was introduced in the Mediterranean (after being accidentally release from the Monaco Aquarium) where it has spread at an alarming rate and significantly changed benthic communities. It has also been introduced to California and Australia. The Indo-Pacific lionfish was recently introduced to the Caribbean, most likely released from aquariums in Florida. Large populations are now established throughout the Caribbean, where lionfish have strong impacts on prey populations and are not controlled by native predators.

14.7. Review Questions

1. Which United Nations conferences led to the development of international laws on maritime boundaries and rights to resources?
2. What is the name of the treaty that was created at the end of these conferences?
3. Where would you expect to have strong haloclines in the coastal ocean?
4. What is an estuary?
5. Why are estuaries often very productive?
6. Why do estuaries typically have low species diversity?
7. What are four ways in which estuaries are formed?
8. What are four types of salinity profiles in estuaries?
9. Describe typical estuarine circulation
10. Describe the circulation pattern in a negative estuary and in the Mediterranean
11. Where do salt marshes typically occur?
12. Is crude oil or refined oil more toxic to marine ecosystems?
13. What does the Oil Pollution Act mandate?
14. What does the Clean Water Act mandate?
15. Why are top predators at greater risk of mercury poisoning than animals low on the food chain?
16. What does MARPOL dictate?
17. Where does plastic debris accumulate in the oceans?

15. Marine Life and the Marine Environment (Trujillo, Ch 12)

Biological organisms can be classified by taxonomy, which reflects evolutionary links, or they can be classified by habitat or functional similarities. This section explores the major groups of organisms in the ocean and their adaptations to various habitats.

15.1. Taxonomic classification

Taxonomic classification aims to group organisms based on evolutionary links. An in-depth review of taxonomic classification is beyond the scope of this course; we will instead simply review the major taxonomic groups of living organisms. Living organisms are classified into three domains: the Bacteria, the Archaea and the Eukarya (Figure 15.1). The Bacteria and Archaea both include small, simple cells called prokaryotes. The Eukarya includes organisms with larger and more complex cells that have sub-cellular structures such as mitochondria and chloroplasts. Domains are further divided into kingdoms and the domain Eukarya has 3 clearly-defined kingdoms: Plantae, Animalia and Fungi.

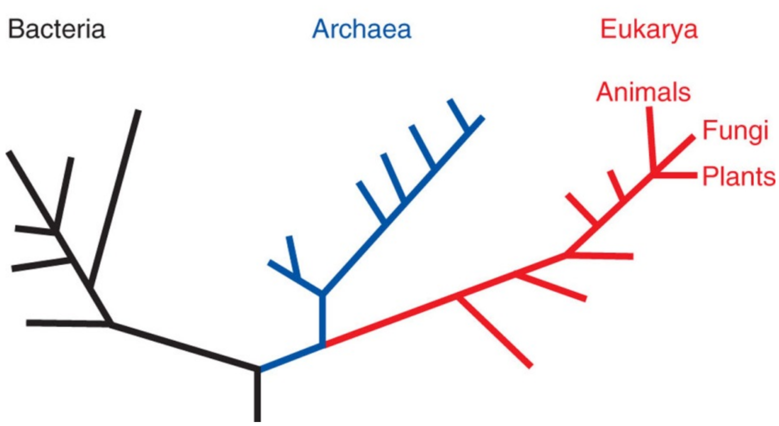


Figure 15.1. The three domains of life, and the 3 well-defined kingdoms of Eukarya.

Taxonomy further divides organisms into smaller and smaller groups of organisms, from phylum, class, order, family, genus and species (Figure 15.2). The fundamental unit in this classification is the species, a group of individuals that can mate to produce viable offspring. Every species is known in the scientific literature by a two-word name composed of the genus and species name. This is called binomial nomenclature, and ensures that species are clearly identified, even across languages. In binomial nomenclature, the genus name is capitalized, the species name is not capitalized, and both the genus and species are italicized. For example, the common dolphin is called *Delphinus delphis* in the scientific literature.

Taxonomic group	Gray wolf found in	Number of species
Domain	Eukarya	~4–10 million
Supergroup	Opisthokonta	>1 million
Kingdom	Animalia	>1 million
Phylum	Chordata	~50,000
Class	Mammalia	~5,000
Order	Carnivora	~270
Family	Canidae	34
Genus	*Canis*	7
Species	*lupus*	1

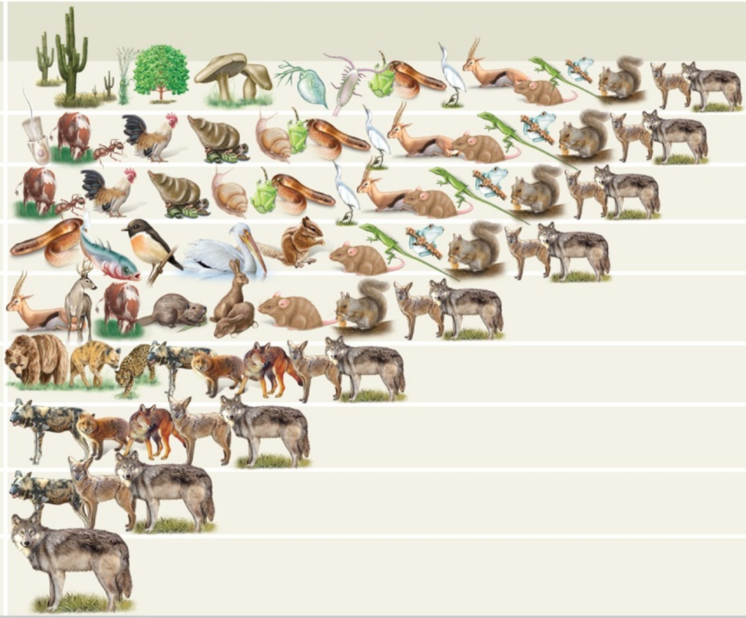


Figure 15.2. Hierarchical taxonomic groupings, using the example of the grey wolf, *Canis lupus*.

15.2. Other classification of marine organisms

Marine organisms are also grouped based on where they live and how well they move. Organisms that live in the water column are called plankton or nekton, while organisms that live on the bottom are the benthos.

Plankton

Planktonic organisms are free-floating organisms that cannot propel themselves against a current, therefore their movement is largely determined by water currents. Planktonic organisms span a large range of size from picoplankton that includes viruses (virioplankton) and bacteria (bacterioplankton) to macroplankton such as jellies. While jellies (like some other planktonic organisms) have some ability to swim, their swimming abilities are limited and their movements are largely determined by the current, so they are considered planktonic.

Phytoplankton are plant and plant-like members of the plankton community, capable of photosynthesis. They are microscopic, single-celled organisms (except *Sargassum*), and are limited to surface waters (down to a maximum of about 200m) by the availability of light. Because they have no or limited swimming ability, they possess adaptations that make them sink slowly and allow them to remain near the surface. They are small organisms, and therefore have a high surface-area-to-volume ratio which slows down sinking. Some also possess projections such as horns, spines or wings, which increase this ratio. They may store low-density fats and oils or have internal gas-filled vesicles, or form chains or aggregates. Distribution of phytoplankton is very heterogeneous, and varies with levels of light and nutrients, and with mixing and seasons. This is discussed in detail in

Chapter 16. Phytoplankton are most abundant in temperate zones and in areas of upwelling.

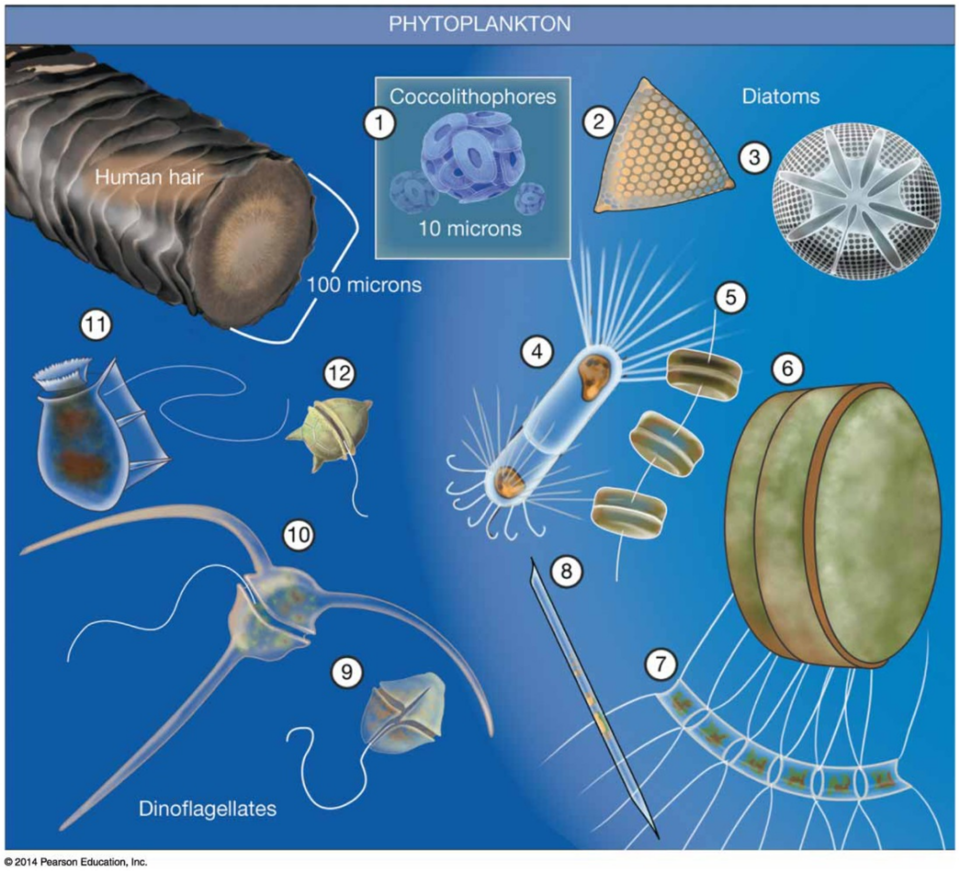


Figure 15.3. Common types of marine phytoplankton. 1) Coccolithophores; 2-8) Diatoms; 9-12) Photosynthetic dinoflagellates.

Zooplankton are the animals of the plankton community, organisms that must consume other organisms. The zooplankton is an extremely diverse group, and includes holoplankton (organisms that spend their entire life as plankton) and meroplankton (larval forms of benthic and nektonic adults that are temporary members of the plankton community). Many species of zooplankton exhibit a daily vertical migration, where they rise towards the surface at night to feed (where food is more concentrated) and descend to depth (200-600m) during the day. Advantages of spending the daytime in deeper waters include a decreased risk of predation from visual predators due to lower light levels, a decreased metabolic rate due to lower water temperature, and slower sinking rates due to increased density and viscosity. The zooplankton are an extremely diverse group that represent most animal phyla.

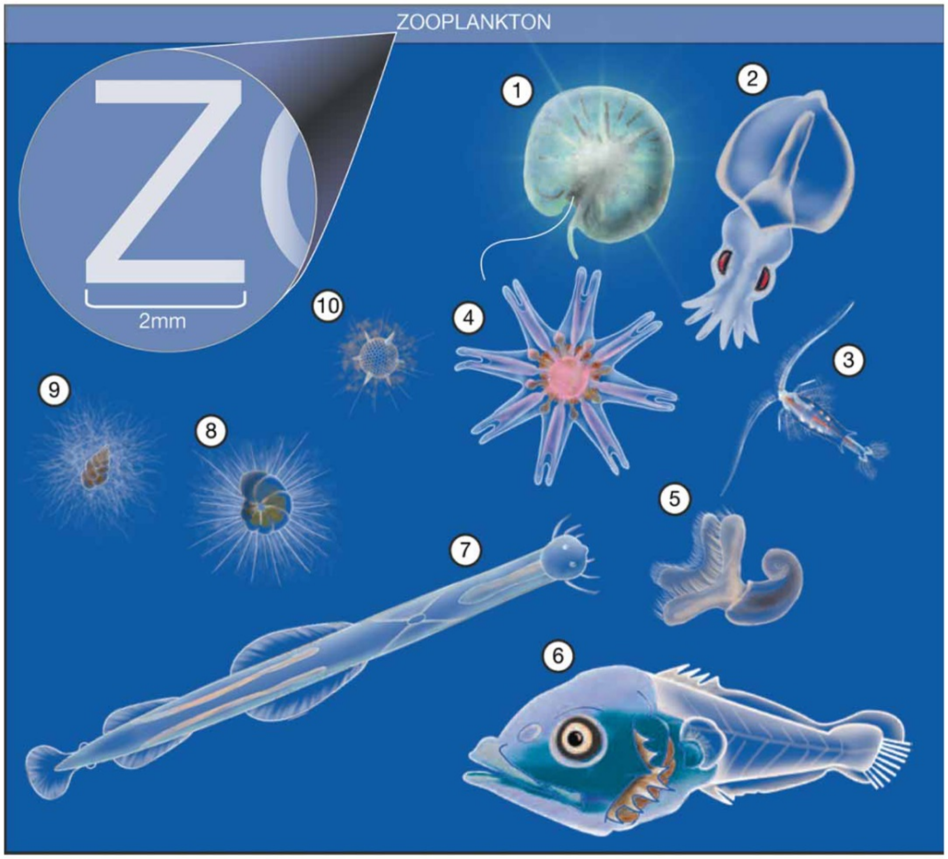


Figure 15.4. Common types of zooplankton: 1) *Noctilucta*, a predatory dinoflagellate that exhibits bioluminescence; 2) squid larva 3) Copepod; 4) Jelly larva; 5) Snail larva; 6) Fish larva; 7) Arrowworm; 8-9) Foraminiferas; 10) Radiolarians.

Nekton

Nekton are the pelagic organisms that are powerful enough swimmers to move at will through the water column (Figure 15.5). They are mostly vertebrates (fish, sharks, rays, marine mammals, reptiles, and birds) but also include invertebrates such as squid. The distribution of nekton is influenced by light, temperature, density and currents. Nekton are most abundant near the surface where food is most abundant, in the epipelagic zone which extends to approximately 200m and corresponds to the photic zone. Most nekton in this zone are large, carnivorous predators in high trophic levels. Some, like tuna, are migratory and follow food sources. Most exhibit a pelagic camouflage called **countershading**, being darker on the top and lighter on the bottom (e.g. shark and tuna, figure 15.5).

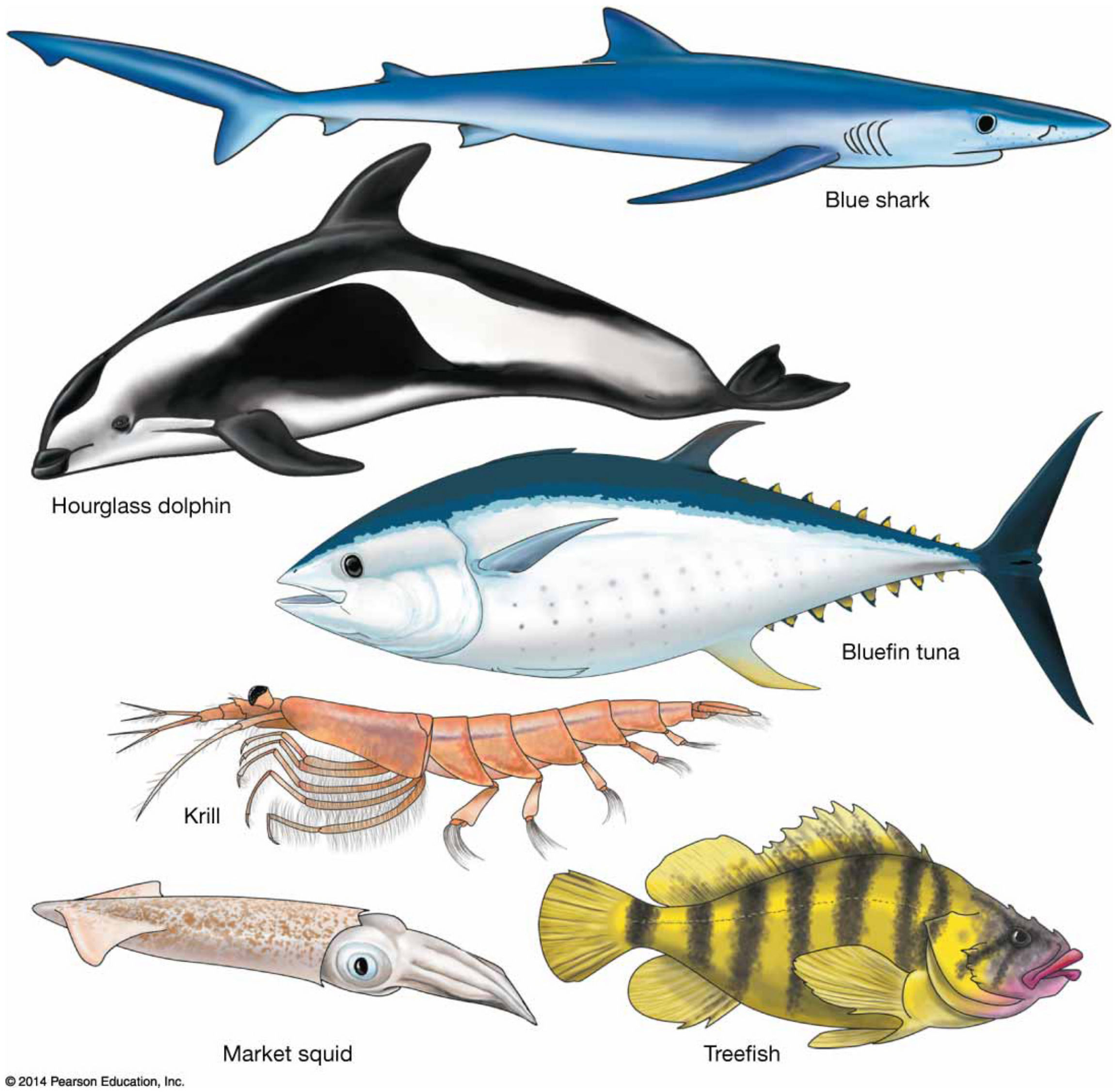


Figure 15.5. Nektonic organisms, with shark and tuna showing countershading, a type of camouflage common in the epipelagic zone.

Benthos
Benthic organisms are those that live on the ocean floor (Figure 15.6). Benthic plants and seaweeds can grow in shallow coastal areas where light is abundant, whereas benthic animals grow in all areas and depths of the oceans.

Benthic multicellular primary producers are restricted to the photic zone, and therefore do not grow deeper than 200m (though most plants are found much shallower than this). They are attached to the bottom, but float up to obtain sunlight. These large plants and seaweeds provide food and shelter from many animals in shallow areas.

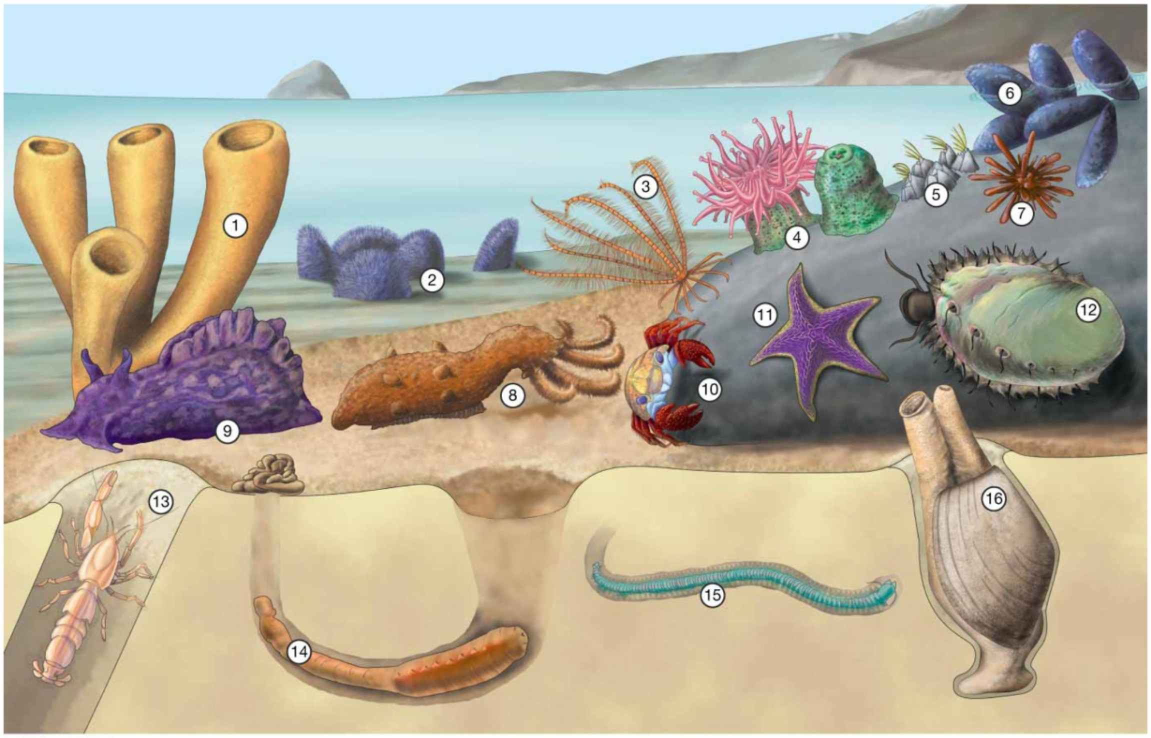


Figure 15.6. Benthic animals: 1) sponge; 2) sand dollars; 3) crinoid; 4) sea anemones; 5) barnacles; 6) mussels; 7) sea urchin; 8) sea cucumber; 9) sea hare; 10) shore crab; 11) sea star; 12) abalone; 13) ghost crab; 14) lug worm; 15) annelid worm; 16) clam.

Benthic animals are adapted to exist at all depths and are found associated with all substrates. Benthic habitats are varied and animals have evolved adaptations for all these habitats, therefore the number of benthic species is quite high, approximately 50 times that of pelagic species. Benthic organisms can be categorized as epifauna, organisms that live on the substrate, or infauna, organisms that live in or burrow through the sediment. Some benthic organisms are motile and can move around, others are sessile and are permanently fixed.

The type of animal community that develops in a given area depends on environmental conditions. Communities that develop on high energy rocky shores are very different than those that exist on low energy sediment areas. Similarly, animal communities in shallow water, amongst seaweeds and sea grasses are quite different from those that exist in the deep sea and depend on food that falls from above. See Trujillo, Chapter 15, for a description of some specialized communities that exist in specific benthic environments.

15.3. Marine diversity

It is estimated the marine species represent about 14% of all known species on Earth (Figure 15.7). This is surprising considering the vast size of the oceans, and is likely explained by the fact that the marine environment is more uniform than terrestrial environments. A higher diversity of environment leads to higher speciation. Within the marine environment, the vast majority of species are found in the benthic zone. This again can be explained by a much greater variation in environment on the benthos than in the water column.

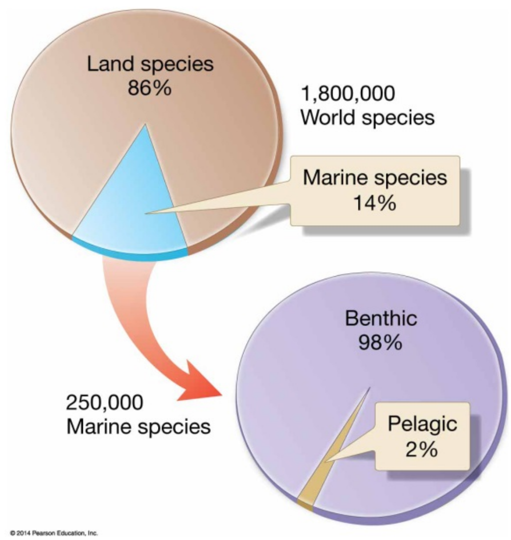


Figure 15.7. Distribution of species between various environments.

15.4. Adaptations of marine organisms

Organisms in the marine environment are adapted to conditions quite different from those on land. Different species might have different adaptations for the same problem and various parts of the oceans have different conditions. Therefore there is a wide variety of forms and systems in marine organisms.

Physical support and buoyancy
All organisms need to support their own body. Marine organisms experience much greater buoyancy from the water than do organisms on land, and therefore the need for structural support is greatly reduced. Marine organisms therefore have reduced support (e.g. seaweed compared to trees) and can reach a very large size (e.g. blue whale). Most marine organisms also need to stay within a narrow range of depths. To this end, many planktonic organisms have evolved adaptations to increase drag and reduce sinking, such as a low density body and long extensions of their shells.

Water viscosity
Viscosity is the resistance of a liquid to flow. Viscosity of water increases with increased salinity and decreased temperature, such that cold, salty water presents increased resistance to flow. While these changes in viscosity have little effect on large nektonic organisms, they do affect smaller organisms because frictional resistance increases with surface area to volume (SA:V) ratio. Organisms with a small SA:V ratio have high frictional resistance (Figure 15.8). This frictional drag is used by many phytoplankton and zooplankton to stay in upper part of the water column, where sunlight is abundant. In cold water (high

viscosity), plankton sink slowly and stay near the surface more easily. In tropical water, sinking rates are more rapid, and some organisms have evolved morphological adaptations to reduce their sinking rates such as more elaborate, feathery appendages (Figure 15.9), needle-like extensions and rings (Figure 15.10). Other organisms have small droplets of oil which decreases density and increases buoyancy. Despite adaptations to stay afloat, most planktonic organisms are denser than water and tend to sink slowly. This slow sinking is not a serious problem, however, because wind-induced turbulence causes enough mixing to keep most planktonic organisms in surface water, where they can photosynthesize or feed on abundant organisms.

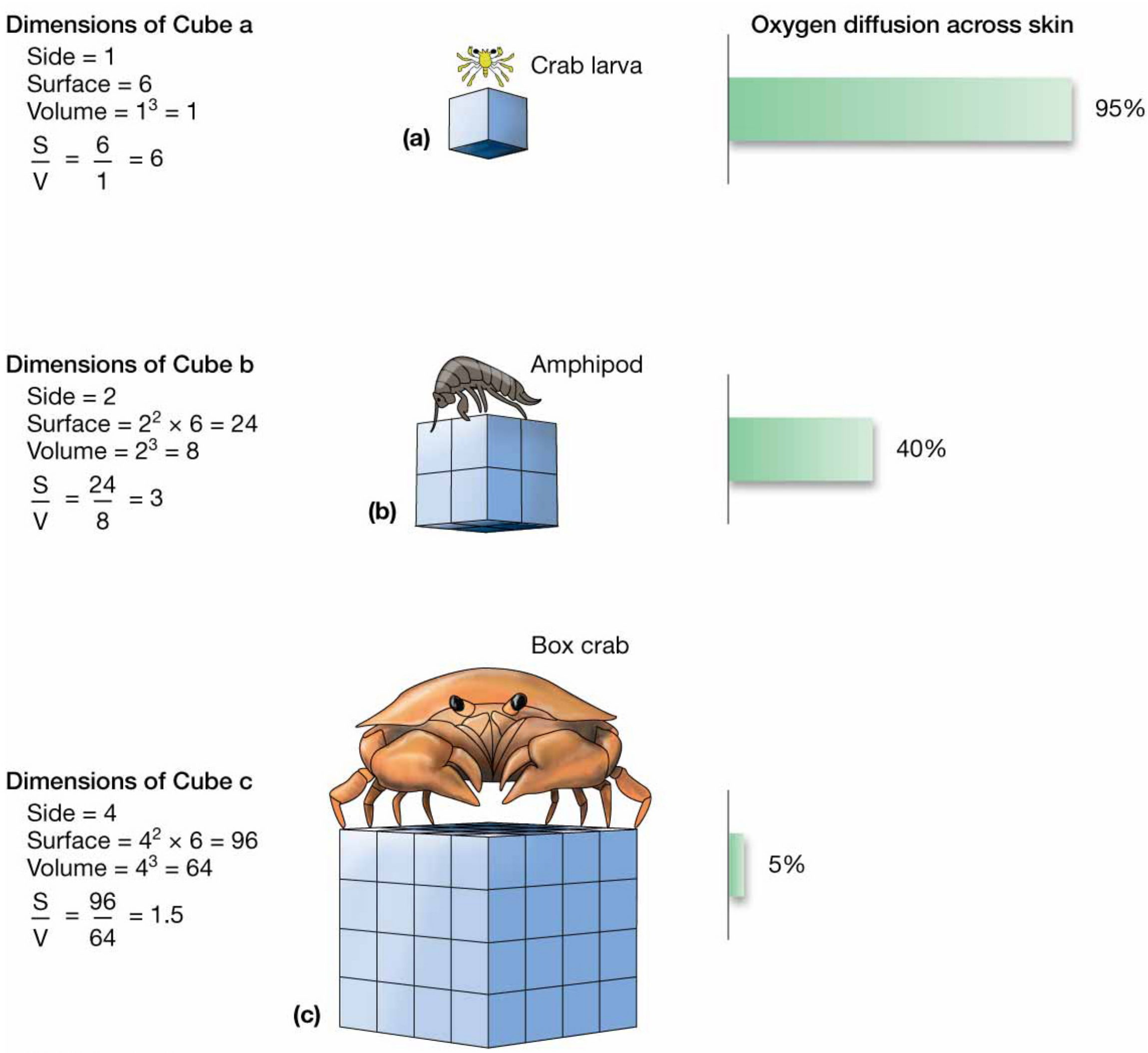


Figure 15.8. The surface area to volume ratio of organisms decreases with organism size. This has important implications for movement, sinking rate, gas exchange and nutrient update.

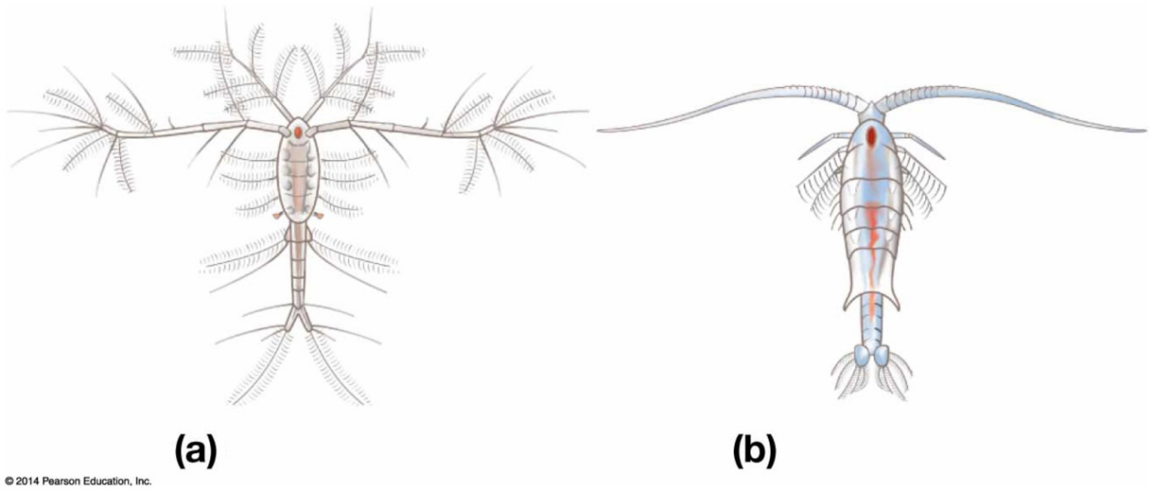


Figure 15.9. Variation in appendages between related species of copepods that live in water water (a) and cold water (b). Warm water species experience faster sinking rates due to lower water viscosity; ornate appendages help increase drag and slow down sinking (length of copepods: 2 mm)

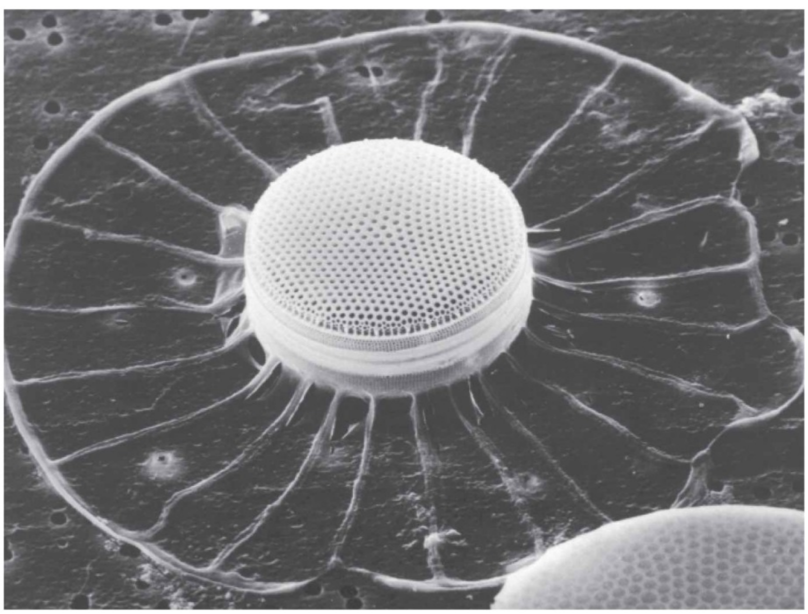


Figure 15.10. Warm-water diatom, showing a large marginal ring which increases surface area and drag (diameter = 60 microns).

High SA:V ratio is also an advantage for small phytoplankton as it allows for greater diffusion of gases and nutrients across the body wall (Figure 15.9). In this way, small phytoplankton are much more successful than larger phytoplankton in low-nutrient tropical waters. Conversely, high-nutrient areas (e.g. upwelling zones, polar seas) can support relatively large phytoplankton at the base of the marine food chain.

For large animals that swim freely in the ocean, water viscosity is an impediment to speed (by comparison, air offers much less drag to animals moving on land). For this reason, most fast-swimming marine animals are streamlined, with a shape that reduces resistance to water (Figure 15.11).

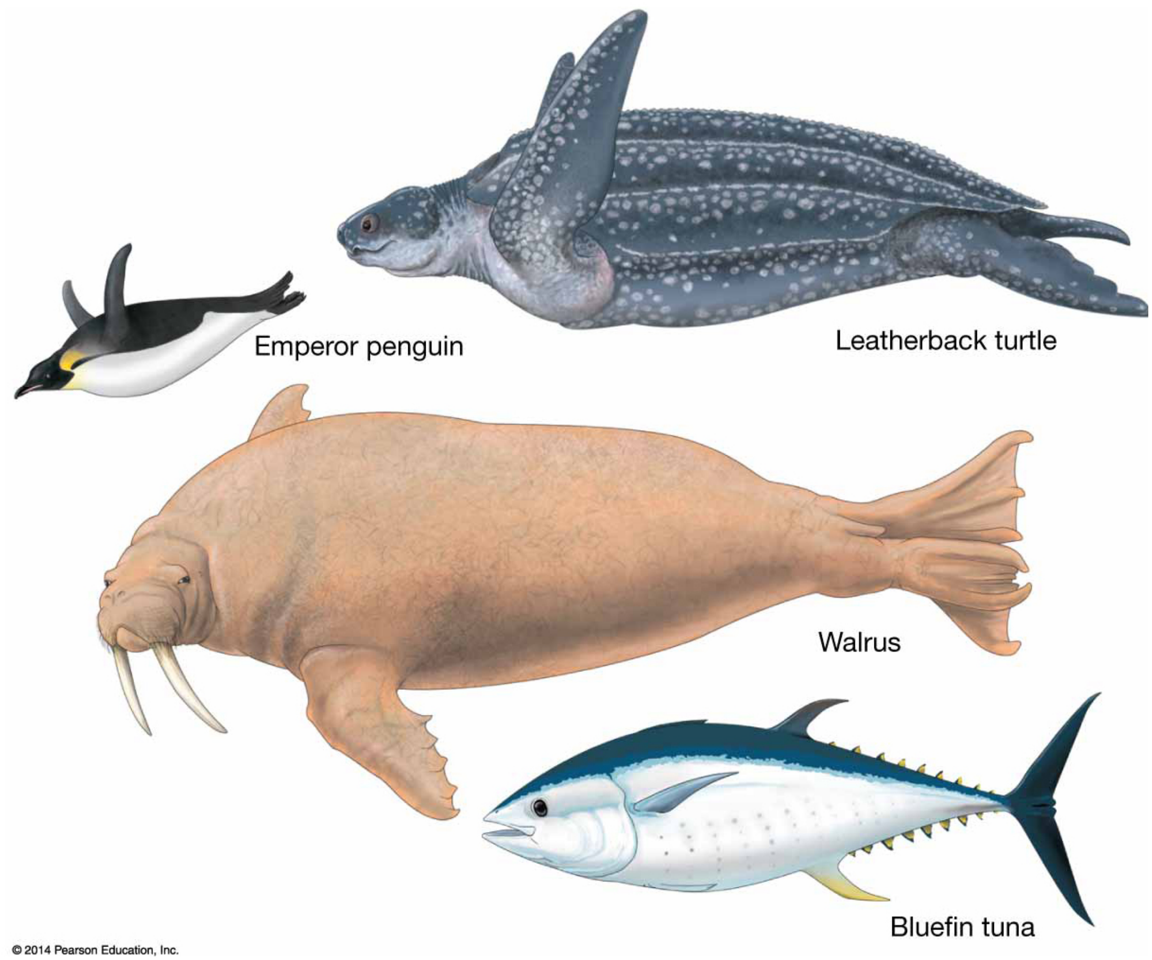


Figure 15.11. Streamlining in various groups of nektonic organisms.

Temperature
Organisms in the ocean experience a much narrower range of temperatures than those on land (Figure 15.12), for four reasons: 1) water has a high heat capacity, 2) the warming of the ocean is reduced substantially by evaporation, 3) solar energy that arrives at the surface of the ocean is distributed over a large body of water (several tens of meters or more), rather than a very thin layer and 4) water has good mixing mechanisms such as currents, waves and tides which redistribute heat. The small daily and seasonal variations in temperature are confined to the surface and their importance decreases with depth. Below 1500, water temperatures are around 3^{o}C year-round.

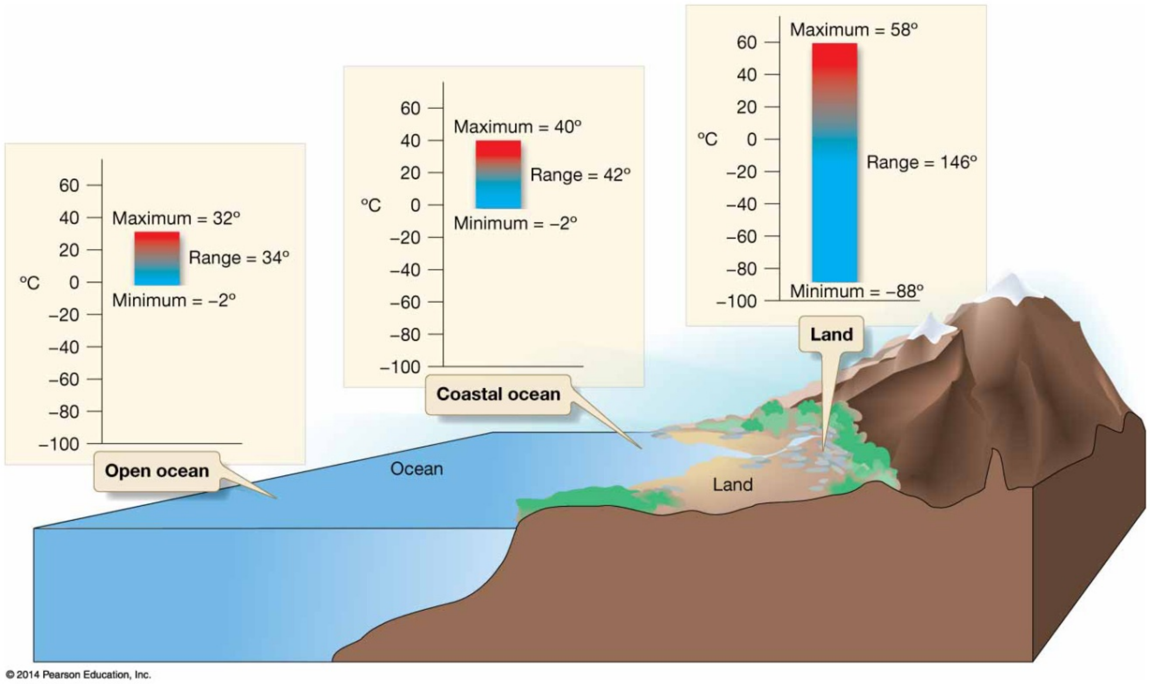


Figure 15.12. Differences in temperature range between land, the coastal ocean and the open ocean.

While temperature variations are less extreme in the ocean than on land, temperature still has important effects on marine organisms. Because of the lower density and viscosity of warm water, planktonic organisms in the tropics tend to have higher SA:V ratio, either by being small in size or by having extensions on their bodies. Because warm temperature increases metabolic rate, tropical species tend to grow faster and have shorter generation times than their cold-water relatives. Consistent with terrestrial patterns of diversity, tropical waters tend to have a higher diversity of planktonic species than cold waters. On the other hand, there is a higher biomass of plankton at high-latitudes because of higher availability of nutrients. Some organisms can survive over a wide range of temperatures (eurythermal), while others have very specific temperature requirements (stenothermal).

Salinity

Organisms also vary widely in their tolerance to fluctuations in salinity. Those that inhabit estuaries, where salinity is dynamic, typically can tolerate a wide range in salinity and are called euryhaline. On the other hand, those in regions that have very stable salinity, like the open ocean, typically have a much narrower tolerance and are called stenohaline. In euryhaline organisms, two main strategies can are found (Figure 15.13). Some organisms deal with changing salinity by having adaptations to maintain metabolic function through some range of internal salinities and are called osmoconformers. Others, called osmoregulators, are adapted to maintain a constant internal salinity across a range of environmental salinities. Osmoregulators include fishes, birds, mammals and several invertebrates (e.g. crustaceans).

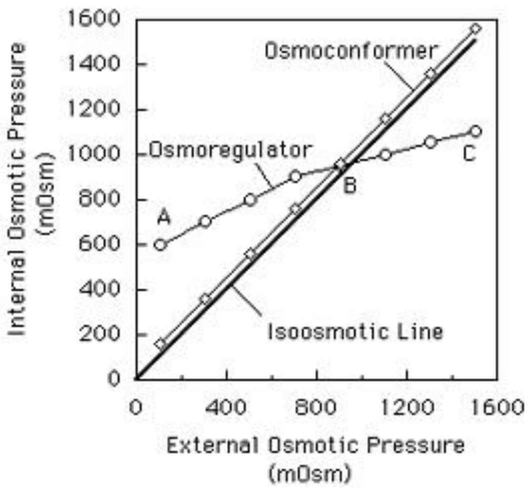


Figure 15.13. Relationship between internal and external salinity (presented here as osmotic pressure) for osmoconformers and osmoregulators.

When waters of unequal salinities are found on either side of a semi-permeable membrane (such as skin or a cell membrane), water moves from the side of low salinity (hypotonic) to the side of high salinity (hypertonic) in a process called osmosis (Figure 15.14). This affects organisms in both freshwater and saltwater, which have different adaptations to live in water of different salinity. This is seen clearly in bony fishes, which have freshwater and saltwater representatives (Figure 15.15). Freshwater fishes are hyperosmotic to their environment, and constantly gain water through osmosis. For this reason, they don't drink, and they produce a large amount of dilute urine. To compensate for the loss of salt, their gills have special cells that reuptake the very dilute salts found in freshwater. Marine fishes, on the other hand, are hypoosmotic to their environment and constantly lose water through osmosis. They therefore must drink large amounts of seawater and produce small amounts of very concentrated urine. Their gills also excrete excess salts that they intake through drinking water.

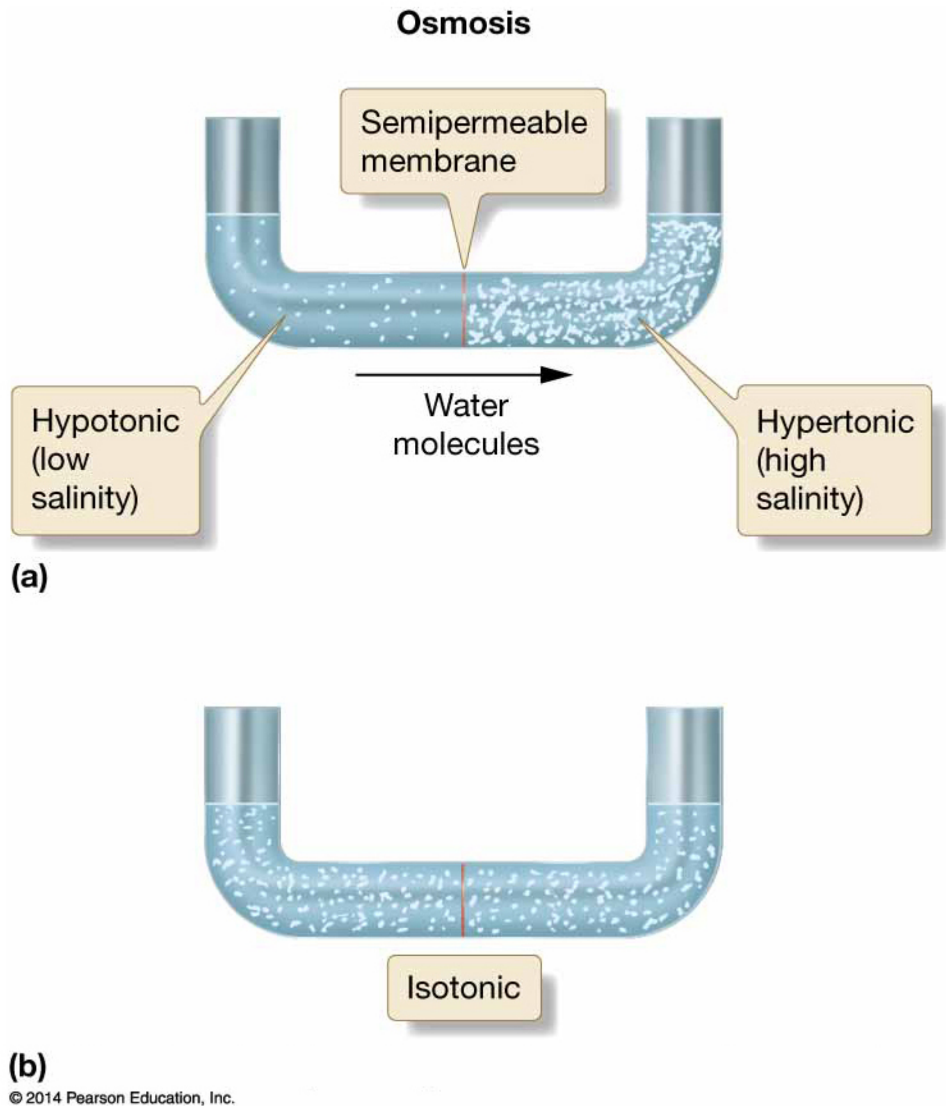


Figure 15.14. The process of osmosis.

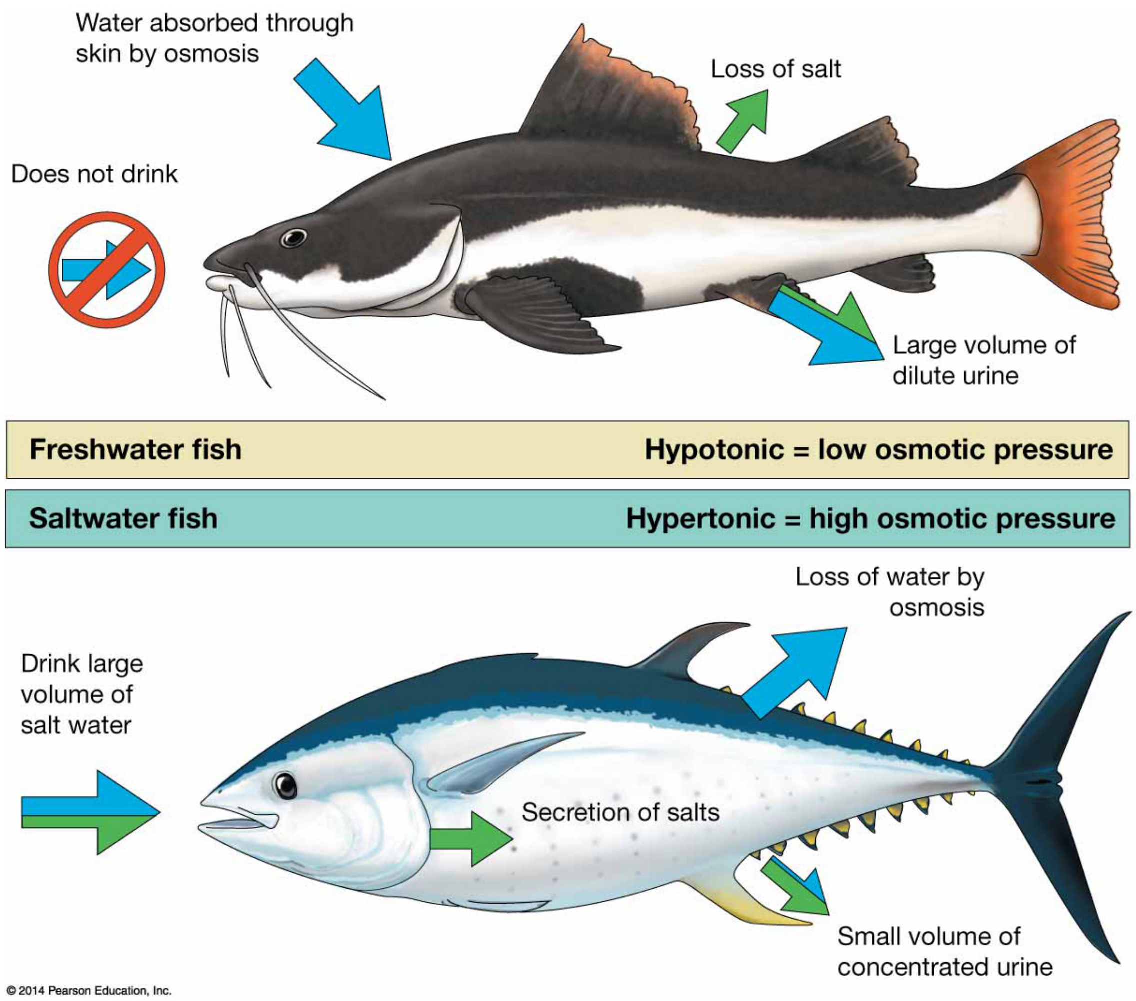


Figure 15.15. Osmotic adaptations of freshwater and marine bony fishes.

Dissolved gases
The amount of gases that are dissolved in seawater is important to the organisms that live in it, in particular for carbon dioxide (which phytoplankton need for photosynthesis) and oxygen (which all organisms need for their metabolism). Cold water holds more gases in solution than warm water, and this allows for high productivity in polar waters in the summer when sunlight and nutrients are available. Further, the cold, oxygen-rich water from the poles sinks to the bottom of the ocean and supplies deep-sea animals with dissolved oxygen. Except for marine birds and mammals, all marine animals must obtain dissolved oxygen from seawater. In very simple animals (e.g. zooplankton or sea jellies), this is achieved through simple diffusion. Most larger animals, on the other hand, require gills to efficiency extract oxygen from the water. Gills increase the surface area of capillaries over which gas exchange occurs (Figure 15.16).

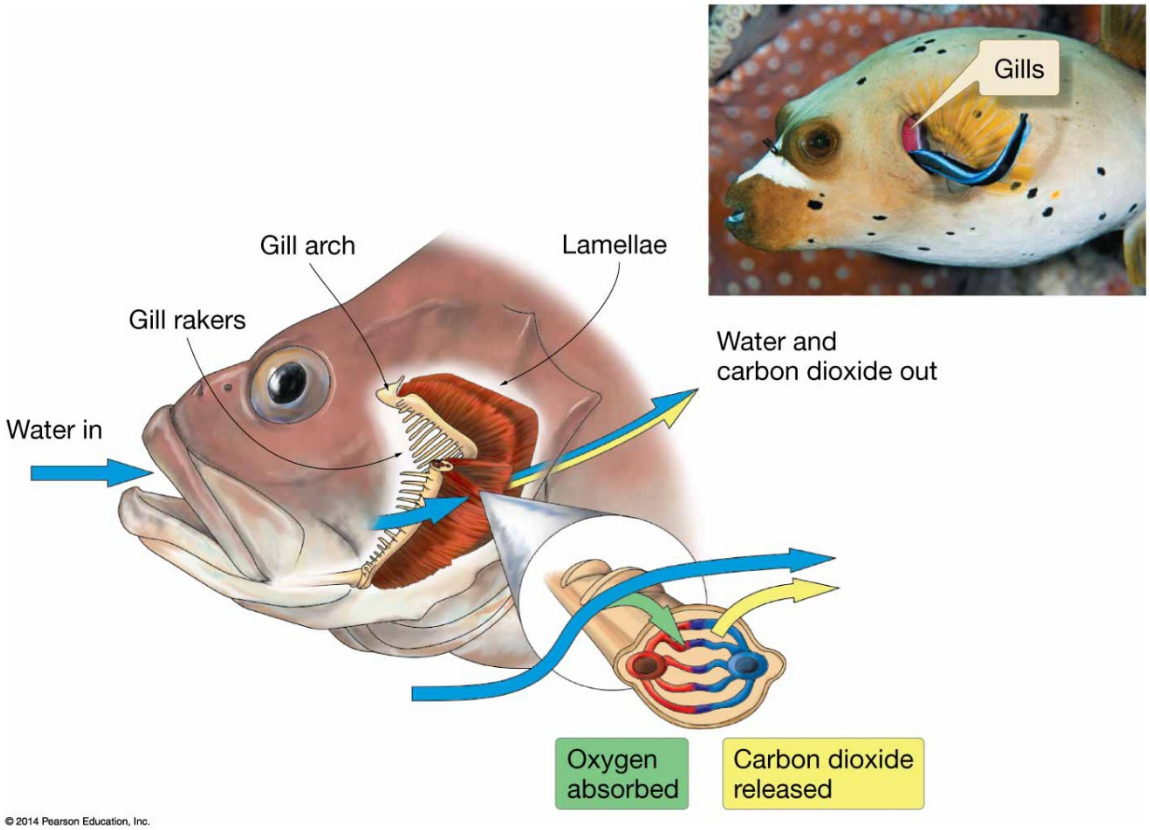


Figure 15.16. Fish gills function to extract oxygen from the water and release carbon dioxide.

Water's transparency

The depth of light penetration in the ocean depends on many factors including the amount of sediment & plankton in the water, latitude, time of day, and season. In the clearest ocean near the equator, maximum light penetration is about 1000m. In this clear, 3-dimensional environment, animals have evolved several strategies to avoid being seen by their predators or prey. Many organisms are nearly transparent or have silvery sides which reflect light. Others display countershading camouflage, with a dark dorsal side and a light ventral side (e.g. penguin and tuna in figure 15.11). With countershading, animals blend against a dark ocean when viewed from above, and against lit surface waters when viewed from below.

Many small animals avoid lit surface waters altogether during the day to avoid falling prey to visual predators and return to the epipelagic zone (where food is more abundant) only at night. This group of organisms forms the deep scattering layer, which is found around 1000m during day time and near the surface at night. The deep scattering layer (DSL) is often referred to as a "false bottom" because the density of organism can be so great that it presents a similar signal on sonars to the ocean floor. Many organisms make up the DSL, including copepods, krill and lantern fish (Figure 15.17).

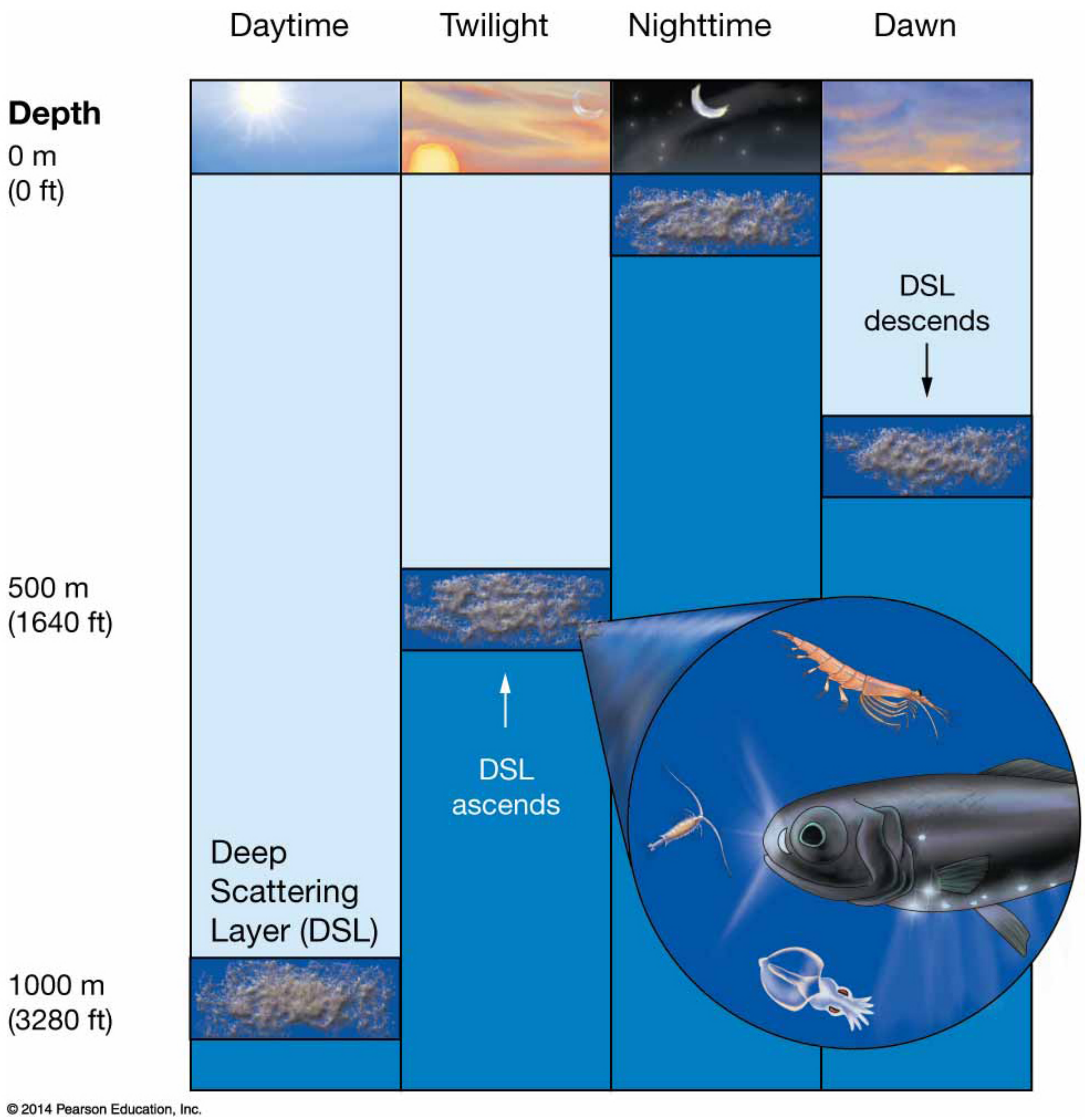


Figure 15.17. Daily movements and organisms of the deep scattering layer.

In shallow coastal waters, where animals have structures in which or against which to hide, different coloration patterns help camouflage. Some species blend exceptionally well with their background, while others use disruptive coloration, with bold patterns and colors which can help camouflage against colorful backgrounds such as a coral reef. Bright colors can also be important for mate recognition and to advertise readiness to mate, or to advertise defenses such as poison or spines.

Pressure

Water pressure increases by 1 atmosphere for every 10 meters of seawater depth. In the deepest parts of the ocean, pressure is several hundred times that of the surface. Water is nearly incompressible, so organisms that lack air-filled compartments (such as lungs) are much less affected by increased pressure than those who do. Deep sea fishes have equal pressure inside their bodies than there is outside, and are well-adapted to life at depth. Still, most organisms can only survive within a certain depth range. Many marine fishes have swim bladders that allow them to regulate their position in the water column (Figure 15.18). For this reason, deep-sea fishes that are brought to the surface rapidly with fishing gear often suffer barotrauma, with their swim bladders expanding more rapidly than they can compensate for (Figure 15.19).

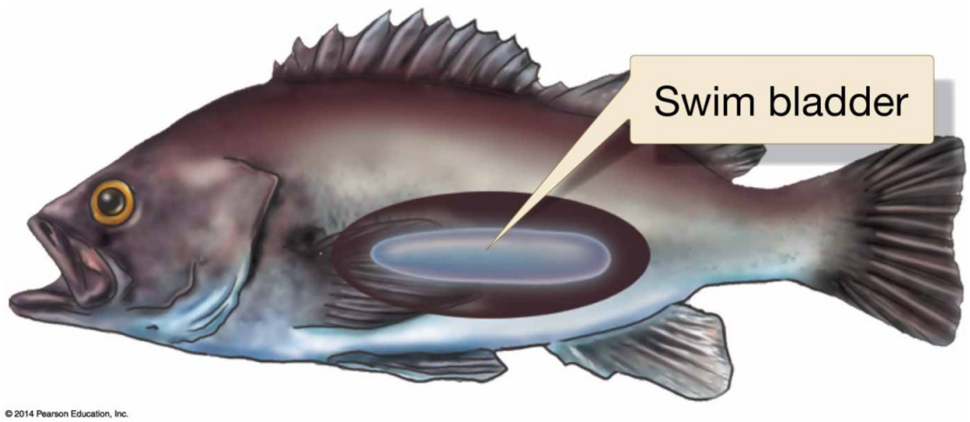


Figure 15.18. Swim bladder in a marine fish.

Figure 15.19. Bocaccio rockfish showing injuries from barotrauma, with the swim bladder push out of the mouth. Photo credit: NOAA/SWFSC.

15.5. Main divisions of the marine environment

The ocean is far from being a homogeneous environment, and it can be subdivided in many sections (Figure 15.20). There are two main ocean habitats: the pelagic zone, in the water column, and the benthic zone, on the ocean floor. The benthic zone can be divided further based on depth, from the intertidal zone (between high tide and low tide) to the hadal zone in deep trenches. Similarly, the pelagic zone can be divided horizontally between the neritic zone (close to shore, above the continental shelf) and the oceanic zone (offshore, away from the continental shelf), or vertically between the epipelagic, mesopelagic, bathypelagic and abyssopelagic.

The most important factor in determining the distribution of life in the pelagic zone is the availability of light. Therefore the pelagic zone is often divided vertically based on light levels. The photic (or euphotic) zone has enough light sustain photosynthesis (to ~100m deep) is where the vast majority of pelagic life exists. Most nekton in this zone are large, carnivorous predators in high trophic levels. The dysphotic zone has low levels of light, so countershading and other camouflage patterns are still useful. It extends from the bottom of the euphotic zone to the disappearance of all sunlight, between 600-1000m. Animals in the dysphotic mesopelagic zone are typically small and have low metabolism due to the low temperature and scarce food. They typically have large eyes adapted to the low light levels. Proteins and biological membranes are adapted to the increased pressure. Mesopelagic nekton are often scavengers and eat the scarce food they come across. Some have photophores, light-producing organs used in species identification, reproduction, camouflage, and attraction of prey. The aphotic zone (the dark zone, below 1000m) has not sunlight at all and in it, many animals are completely blind. Food is severely limited here; the deeper it is, the higher the chance of food falling from above having been eaten as it sinks. Animals have interesting adaptations to increase their predation and mating success.

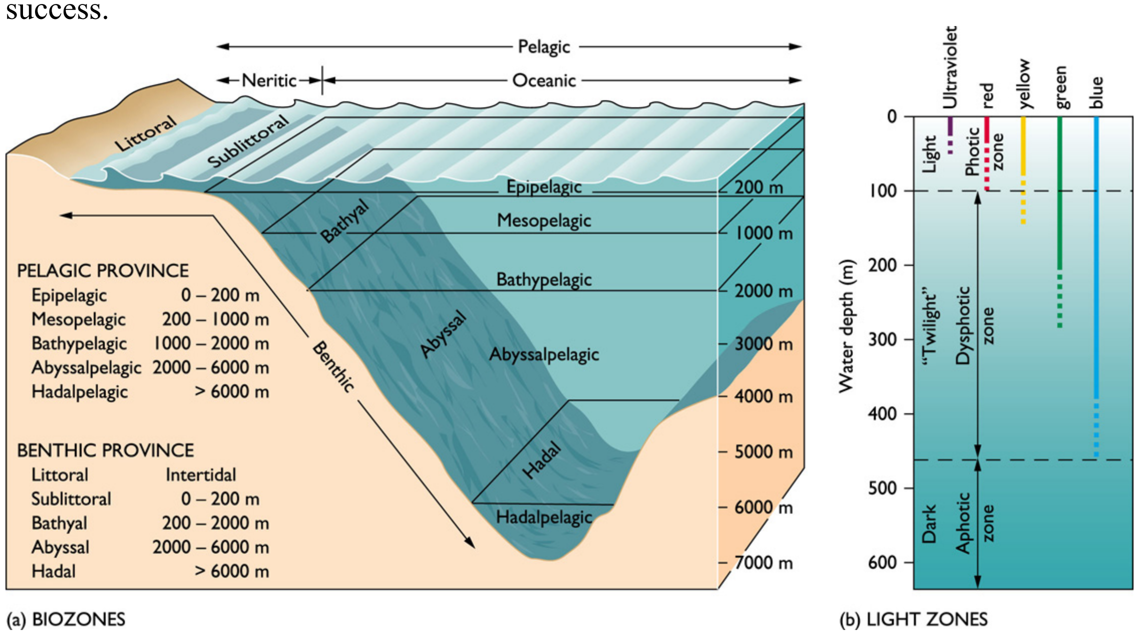


Figure 15.20. Ocean habitats

The upper part of the epipelagic zone is the only place in the ocean that has high enough light levels to support photosynthesis. This leads to high levels of oxygen and low carbon dioxide. Below the epipelagic zone oxygen levels drop rapidly because of the lack of photosynthesis, but continued decomposition and respiration. Oxygen levels reach a minimum in the mesopelagic zone around 700-1000m (called the oxygen minimum zone). Below this depth, oxygen levels increase, in large part because this deep water originated as cold, oxygen-rich water at the poles. Nutrient levels are typically low in the epipelagic zone because available nutrients are continuously used up by photosynthetic organisms. Nutrient concentration reaches a maximum near the oxygen minimum zone, because decomposition breaks down organic molecules into their inorganic components (the nutrients).

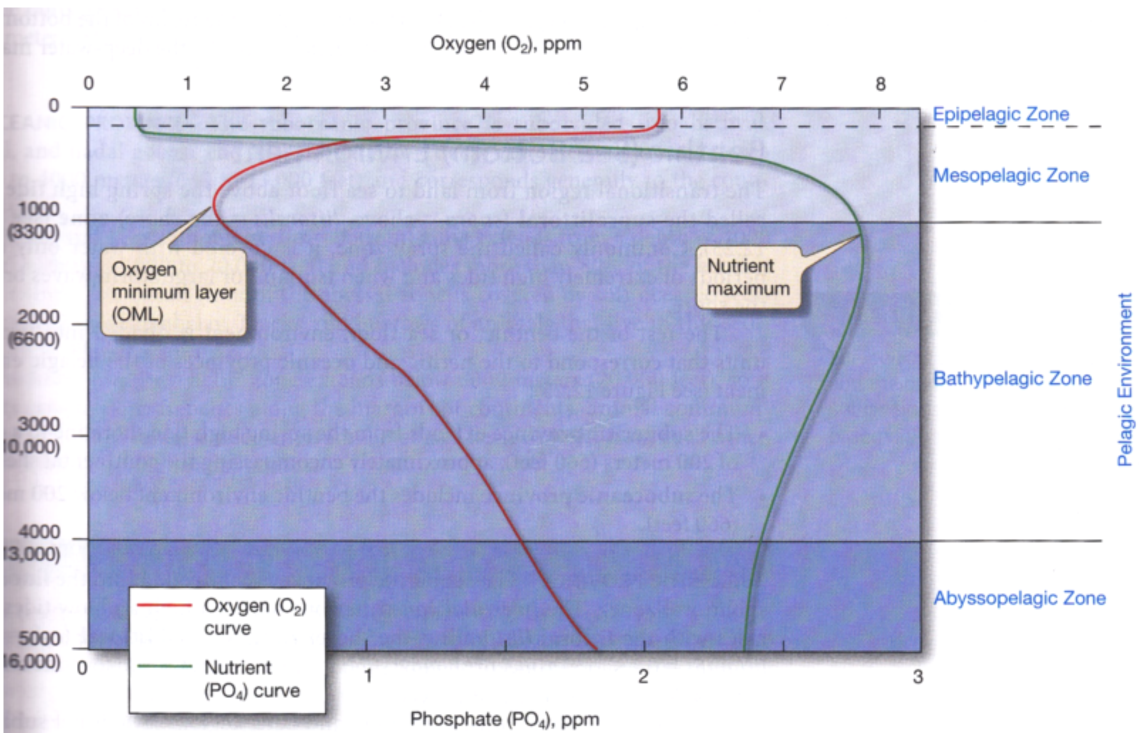


Figure 15.21. Dissolved oxygen and nutrient concentration with depth (vertical axis on the left, in meters (feet in brackets)).

15.6. Review Questions

1. What are the three domains of life?
2. What are the three well-defined kingdoms of Eukarya?
3. What is the basis of groupings in taxonomic classification?
4. What are the two words that are included in the scientific name of an organism?
5. What is phytoplankton?
6. What kind of camouflage is useful for nekton of the pelagic zone?
7. What is infauna?
8. Why is diversity in the marine environment lower than in the terrestrial environment?
9. What are two elements of an organism's biology that are influenced by its SA:V ratio?
10. Do plankton sink faster in the tropics or in the poles?
11. What adaptations do planktonic organisms have to lower their sinking rate?
12. What is the term used to describe an organism that can live in a wide range of salinity?
13. Explain how marine fishes maintain their osmotic balance
14. How does the depth of the deep scattering layer change with time of day?
15. What are the advantages to the zooplankton vertical migration?
16. Which zone has some light but not enough to sustain photosynthesis?
17. Where is the oxygen minimum zone? Why?

16. Biological Productivity & Energy Transfer (Trujillo, Ch. 13)

16.1. Primary Productivity

Primary productivity, or primary production, is the rate of formation of energy-rich organic matter from inorganic materials by autotrophs (photosynthetic or chemosynthetic organisms). Because life on earth is based on carbon (C), primary productivity is usually expressed in $gC/m^2/yr$ (weight/area/year). The rate of primary production varies widely across the oceans, from 25 to 1,250 $gC/m^2/yr$. Primary productivity is typically highest in estuaries and upwelling zones (where nutrients are highest), and lowest in open ocean (low nutrients). The biomass is the total mass of living organisms in a given volume of water. One can be interested in estimating the total biomass (all species combined) or the biomass of a specific species or group of organisms. The standing crop refers to the total biomass of photosynthetic organisms in a given volume of water at a given time.

Phytoplankton, though extremely small, are very abundant and have a high biomass, and therefore are responsible for 90-98% of the global marine organic carbon production. Seaweeds and other macroalgae, because they are mostly benthic, are limited to shallow near-shore areas. They have a smaller biomass and are responsible for about 2-10% of the marine organic production. Chemosynthetic organisms (mostly at hydrothermal vents) are responsible for less than 1%.

Most autotrophs use the sun's energy to convert inorganic nutrients into organic molecules through photosynthesis (Figure 16.1). The total amount of organic material produced in a given area over a given period of time is referred to as gross primary production. However, plants also respire (like all organisms), and use some of this organic material for their own metabolic needs (Figure 16.1). Net primary production is the amount of organic material available to higher trophic levels after respiration losses.

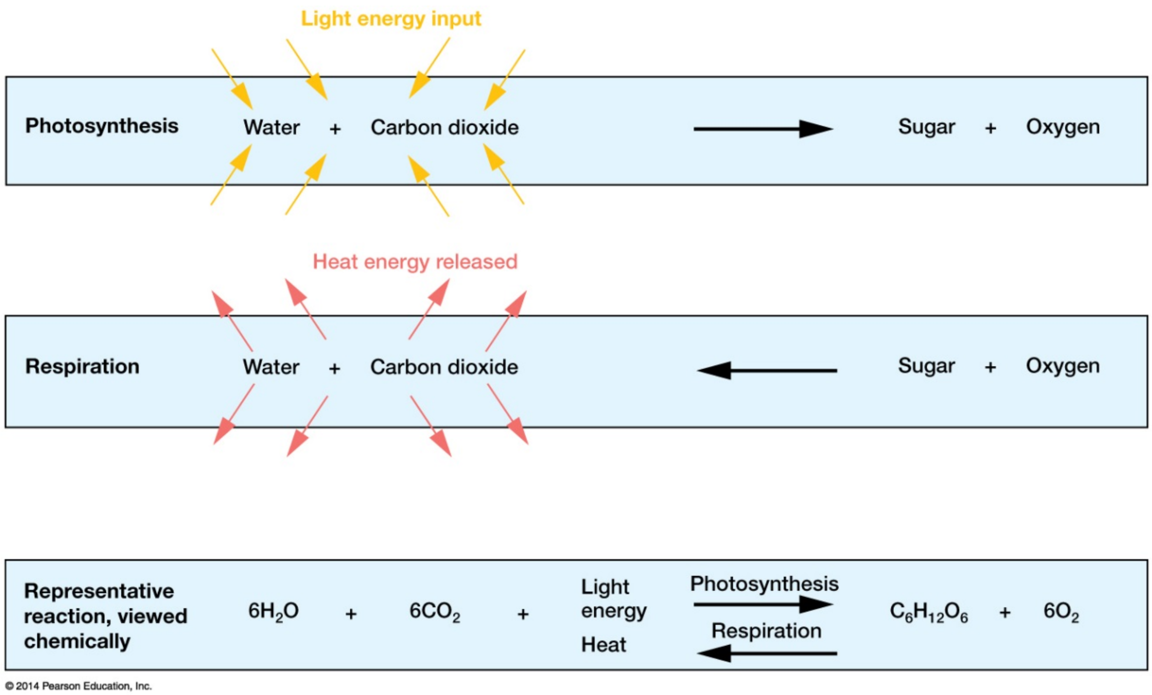


Figure 16.1. Photosynthesis and respiration reactions.

Because all photosynthetic organisms require sunlight, they must remain in the photic zone, which may reach as deep as 200m. The amount of light available for photosynthesis and the depth of the photic zone depend on many factors, including the amount of light absorption by the atmosphere, dust and clouds (e.g. less light reaches the surface of the ocean on a cloudy day); the angle between the sun and the sea surface, which varies with latitude and season; the transparency of the water (turbidity); the reflection of light at the surface, which increases with sea state (agitated seas reflect more light than flat seas). Because the amount of light decreases with depth, deeper regions of the ocean are light-limited and photosynthesis is low. The very surface of the ocean on the other hand, is often photo-inhibited; the light levels are so high that they reduce photosynthetic rate. The rate of photosynthesis is highest at intermediate light levels, and is typically at its highest between 5 and 20 m deep, though this is highly dependent on location and time. The depth at which photosynthetic organisms receive only enough light to produce the amount of oxygen they need for their own metabolic needs (i.e. rate of photosynthesis equals rate of respiration) is known as the compensation depth (Figure 16.2). The compensation depth usually occurs at about 1% of the light levels that reach the surface, and the net production at this depth is zero.

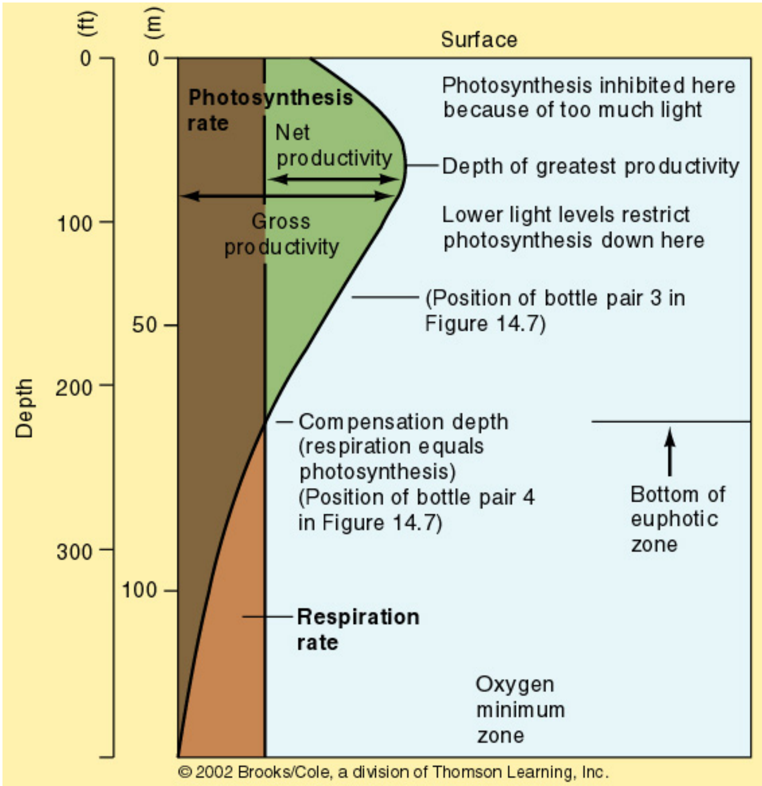


Figure 16.2. Compensation depth in relation to photosynthesis and respiration rates.

The compensation depth varies depending on factors that affect water clarity. For example, as the amount of phytoplankton increases during a bloom, the compensation depth becomes shallower as the bloom reduces the depth to which light can penetrate.

Photosynthetic organisms can adapt to changing light levels in four main ways. Various accessory pigments allow for light absorption over a wide range of wavelengths; chlorophyll, abundant in green algae, absorbs red wavelengths, which are quickly absorbed with depth, therefore green algae photosynthesize well near the surface. Red algae, on the other hand, have pigments that allow them to photosynthesize in lower light levels, and therefore grow deeper. Phytoplankton also have varying shapes of chloroplasts (the photosynthetic organelles within the cell): those in high light levels tend to have circular shaped chloroplasts, whereas those at depth in lower light levels may have oblong chloroplasts that increase the surface area-to-volume ratio to catch as much light as possible. Phytoplankton at depth can also increase the number of their chloroplasts and move chloroplast closer to the outside of the cell, to maximize photosynthesis in low light levels.

Inorganic nutrients are also necessary for the growth of phytoplankton. All photosynthetic organisms require nitrogen and phosphorus. Most organisms can only take up nitrogen in the form of nitrate (NO_3^{2-}), nitrite (NO_2^-), and ammonium (NH_4^+). Few organisms are able to use atmospheric nitrogen (N_2), though cyanobacteria (blue-green algae) are amongst the organisms that can. Similarly, phosphorus must be present in the form of phosphate (PO_4^{3-}). Besides nitrogen and phosphorus, some phytoplankton also require silicon in the form of silicate (SiO_2), for example, diatoms require silica to build their frustule (shell). Some trace elements (e.g. iron, manganese, cobalt, zinc copper) are also required, though in minute quantities, for the growth of phytoplankton.

Nutrients tend to be present in limited quantities in the photic zone because they are continually used up by primary producers. Below the photic zone, nutrients are abundant. Nutrients can be returned to surface waters (in the photic zone, where they can be used in photosynthesis), in four main ways. Upwelling is a process where deeper water is brought to the surface, because of the movement of surface currents or because of a lower density, and brings nutrients from deeper waters (Figure 16.3). Turbulence and mixing also bring deeper waters and nutrients back to the surface. Seasonal mixing occurs in temperate latitudes, where during the fall and winter, the lowered surface temperature creates dense surface water, which then sinks to depth and is replaced by nutrient-rich deep water that rises to the surface. Bacterial decomposition of phytoplankton, zooplankton and other organisms in the photic zone release nutrients in that zone. Similarly, the excretion of waste material by organisms near the surface releases nutrients in the photic zone.

Photosynthetic organisms must have sunlight and nutrients in order to convert the sun's energy into organic molecules. When both sunlight and nutrients are present, there is a rapid expansion of the phytoplankton population, called a phytoplankton bloom.

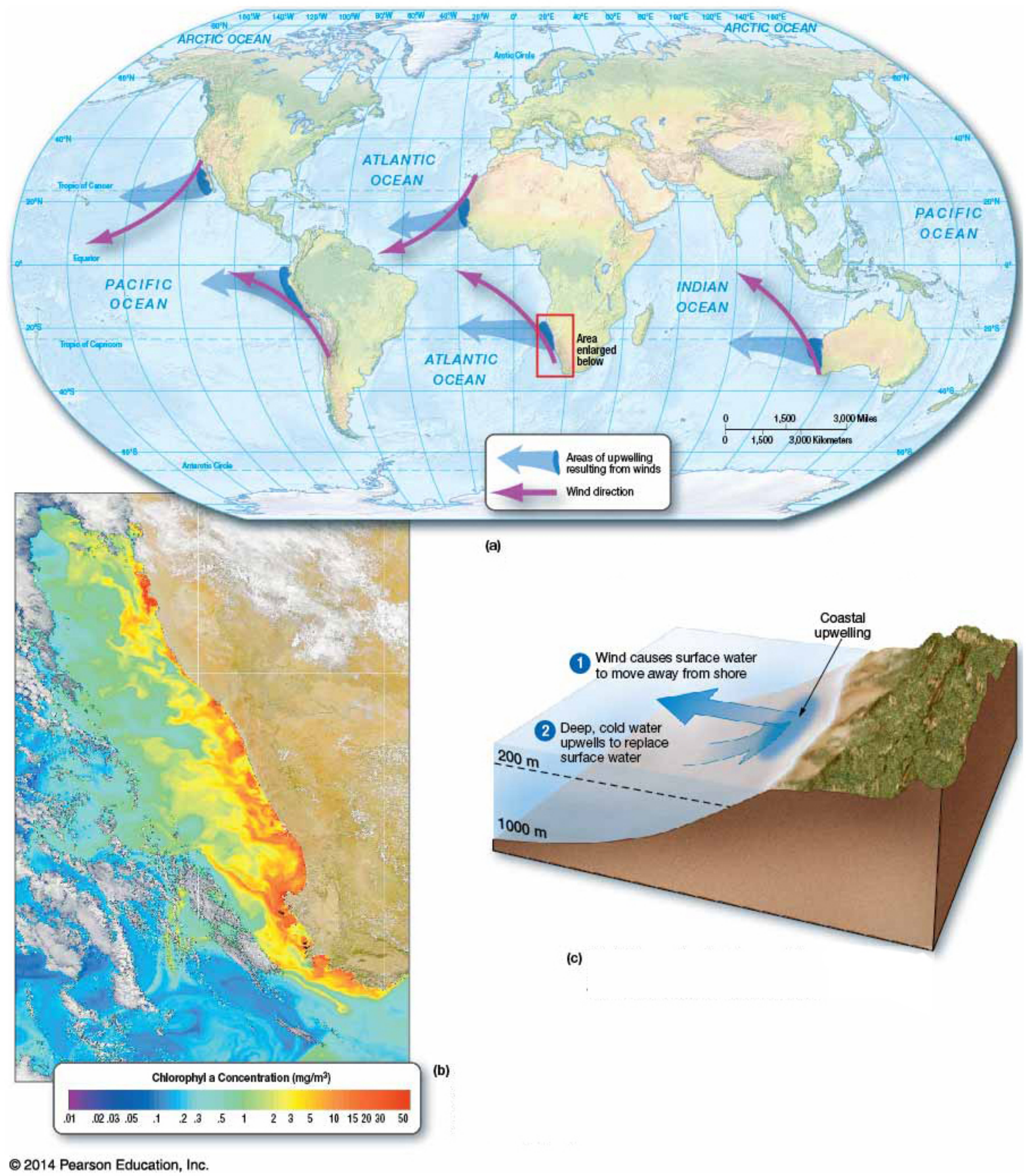


Figure 16.3. Wind-driven upwelling, which brings nutrient-rich deep water near the surface, enhances primary productivity.

16.2. Photosynthetic marine organisms

The vast majority of photosynthesis is in the ocean is carried out by microscopic organisms. However in shallow waters, algae and higher plants also contribute to primary production.

Seed-bearing plants
Seed-bearing plants have evolved on land from aquatic ancestors, and have roots, stems and leaves. They use their roots to anchor in sediments. They possess structures to conduct water and nutrients (xylem and phloem). Only a few species then evolved new adaptations for life at sea. All are confined to shallow coastal areas. Coastal seed-bearing plants include seagrasses, salt marsh grasses and mangroves (Figure 16.4).

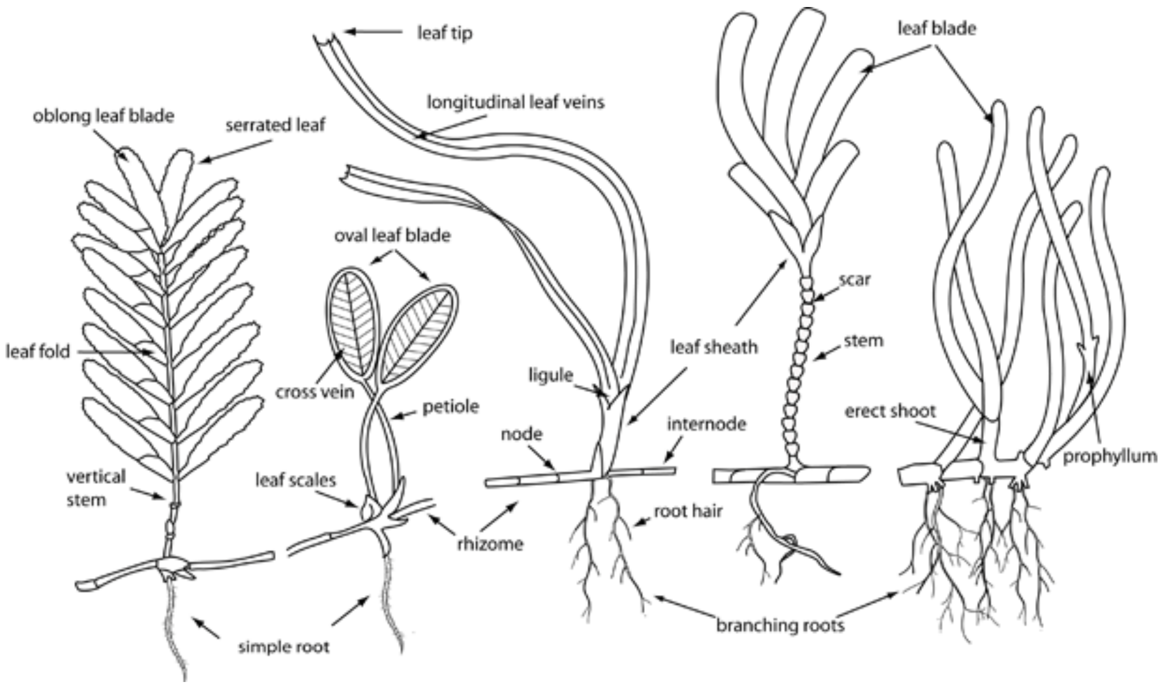


Figure 16.4. Typical growths in seagrasses.

Macroscopic algae
Also known as seaweeds, these are the largest group of benthic multicellular primary producers. They have no roots, and instead attach to hard substrates with their holdfast. They also do not have stems or leaves, and instead have a stipe and blade (Figure 16.5). Photosynthesis is most important in the blade. They have no system to conduct water or nutrients; molecules are transported by diffusion. Some seaweeds have pneumatocysts, gas-filled vesicles that provide flotation to keep the blades near the surface. There are three main groups of seaweeds, separated by their main pigments. The Chlorophyta, or green algae, possess large amounts of chlorophyll and are the shallowest group, for example *Halimeda*. The Phaeophyta, or brown algae, have accessory pigments that give them an olive-green or brown color. This group includes kelps and *Sargasssum*. The Rhodophyta, or red algae, have red accessory pigments that allow them to photosynthesize in low light levels, and therefore they are usually the deepest of all three groups. These include coralline algae and the seaweed used for sushi (nori).

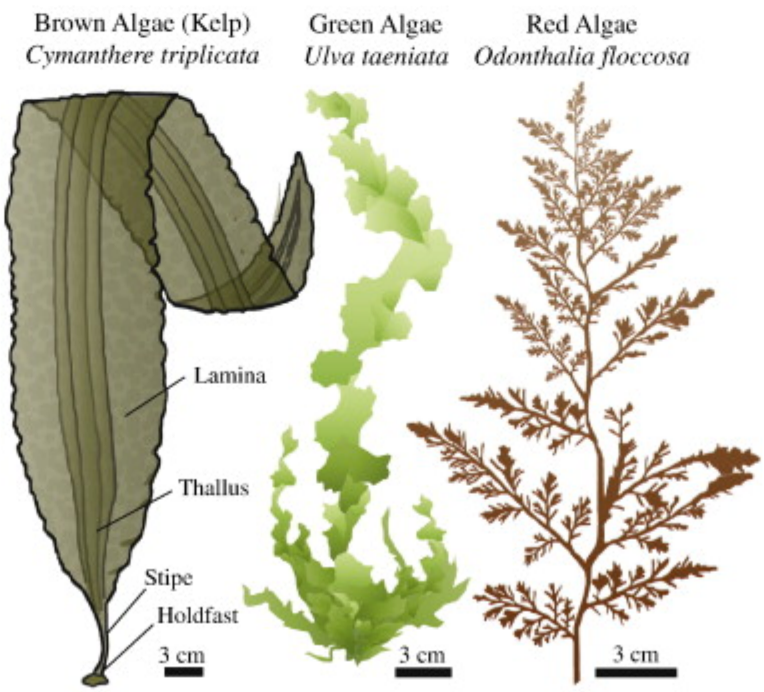
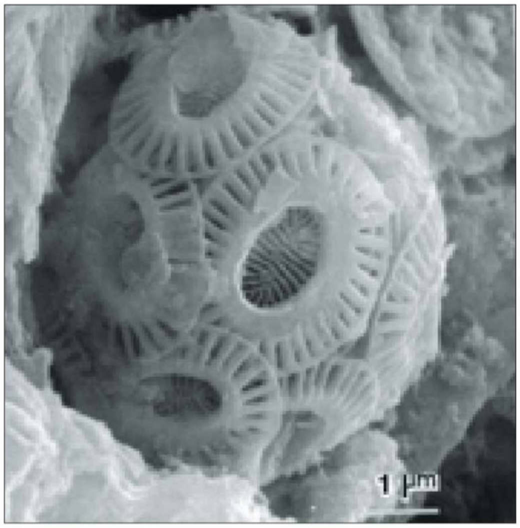


Figure 16.5. A few species of macroalgae.

Microscopic algae
The three main groups of phytoplankton are the golden algae, dinoflagellates, and blue-green algae (cyanobacteria).

The Chrysophyta are also called golden algae because they possess yellow-brown pigments that mask the chlorophyll. They include diatoms and coccolithophores. Diatoms (Figure 16.6) have a frustule (shell) impregnated with silicate, and therefore require silica as a nutrient. They are a very important part of temperate phytoplankton blooms, and are the major component of diatomaceous earth, siliceous sediments used as industrial filters and abrasives. Coccolithophores, on the other hand, are covered in carbonate plates called coccoliths and after they die contribute to calcareous deposits.

Figure 16.6. Diatoms



(b)

Figure 16.7. Coccolithophore

Dinoflagellates can be both autotrophic or heterotrophic. Autotrophic dinoflagellates are considered microalgae. They have red and green pigments, and can grow at greater depth than diatoms. Dinoflagellates possess two flagella that give them some swimming ability and some control over their vertical movements. Their test is made of cellulose, which breaks down after death; therefore dinoflagellates do not accumulate in sediment the way diatoms and coccolithophores do. Several species of dinoflagellates are bioluminescent, and the symbiotic algae in corals, zooxanthellae, are a type of dinoflagellates. Dinoflagellates are the main group responsible for red tides, the phenomenon by which microalgae become so abundant in the water that they discolor it, often (but not necessarily) red. Red tides are more accurately called harmful algal blooms and can be detrimental to humans and marine ecosystems. In many red tides, there is no direct harmful effect to marine animals or humans, but the scale of the bloom nonetheless has important ecological consequences in that once the blooms starts to die, the massive amount of decomposition can strip the water of oxygen and cause large die-offs of animals that could not move away from the anoxic waters. In other cases, red tide dinoflagellates produce toxins that have direct harmful effects on many of the organisms that consume them, or even simply come in contact with them. Dinoflagellates are responsible for paralytic shellfish poisoning, diarrhetic shellfish poisoning, neurotoxic shellfish poisoning and amnesic shellfish poisoning. These illnesses in humans tend to occur after consuming contaminated bivalves which have filtered the water and concentrated the toxins. Dinoflagellates also cause ciguatera poisoning, which occurs after ingestion of certain contaminated tropical reef fishes.

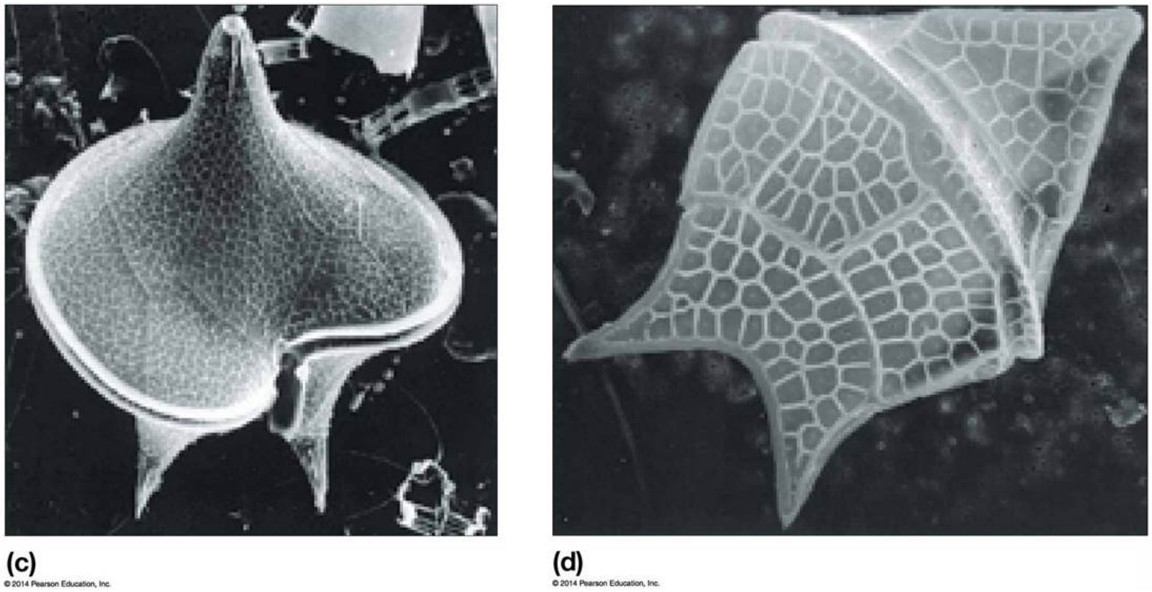


Figure 16.8. Dinoflagellates

Cyanobacteria, or blue-green algae, are the only photosynthetic prokaryotes. *Prochlorococcus,* a type of cyanobacteria, is the most abundant marine phytoplankton in the world. Cyanobacteria are especially important because they can use atmospheric nitrogen, N_2 as a source of nitrogen and therefore not limited by nitrate levels as most other photosynthesizers are. This process is called nitrogen fixation, and it allows cyanobacteria to thrive in areas that otherwise would have low primary production due to limiting nutrients.

Eutrophication and dead zones

Ocean eutrophication is the process of increasing nutrients in coastal waters dues to human activity. This occurs primarily through the runoff of fertilizer, sewage and animal waste. Increased nutrients enhances primary production, but can lead to large blooms which rapidly die and cause anoxic conditions because decomposition uses oxygen (Figireu 16.9). Large areas of hypoxic or anoxic waters now regularly develop near the mouth of major rivers of the world such as the Mississippi River, which drains a large portion of the US farm land. These low-oxygen zones are called dead zones, because they cause the death of animals that are trapped in them.

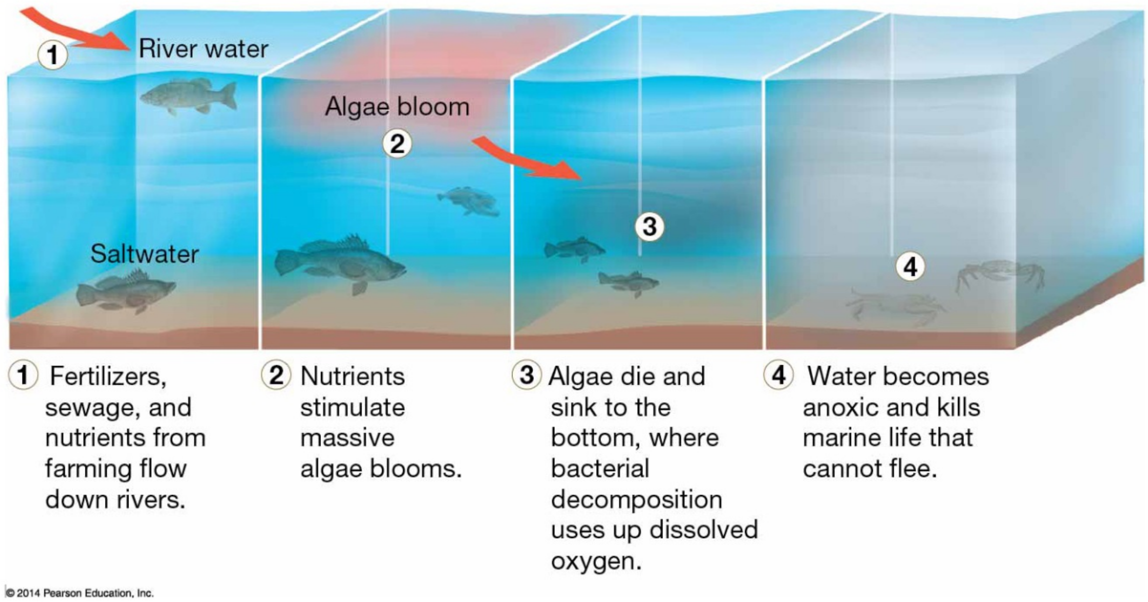


Figure 16.9. Formation of anoxic waters of dead zones.

16.3. Regional variation in primary productivity

Much of light levels and nutrient availability (and therefore primary production) varies with latitude. Tropical waters typically have high light levels but low nutrients; there is a permanent thermocline and nutrients are not regenerated rapidly in the photic zone (Figure 16.10), except in upwelling areas. Tropical regions therefore tend to have a low but constant primary productivity throughout the year.

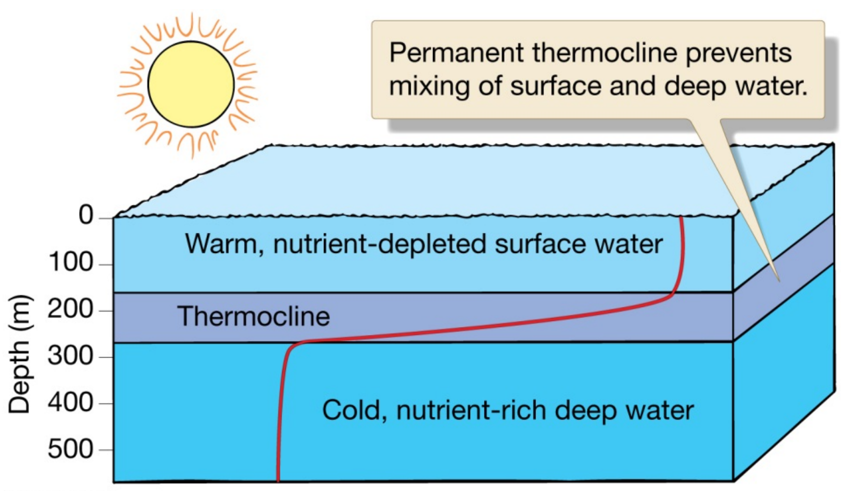


Figure 16.10. Productivity in tropical oceans.

There are 3 main exceptions to the low-productivity standard of tropical waters: zones equatorial upwelling, coastal upwelling, and coral reef ecosystems. Equatorial upwelling occurs where trade winds drive surface transport in diverging directions at the equator, forcing deep, cold, and nutrient-rich water back to the surface and thereby enhancing primary productivity. Coastal upwelling occurs where winds drive surface water away from a land mass, thereby forcing deep, nutrient-rich water to the surface. This typically occurs on the west side of continents. Coral reef ecosystems exist in low-nutrient waters and have few sources of new nutrients the way upwelling zones do. But here, there is a tight recycling of nutrients in large part because of the symbiosis between coral animals and their symbiotic algae. The algae use the coral's metabolic wastes as a source of nutrients and therefore can sustain much higher rates of primary production than would microalgae in the water column away from the reef.

Polar regions are characterized by low light levels for much of the year, but the water column is well-mixed and nutrients are abundant (Figure 16.11). Phytoplankton blooms occur for a short period in the summer, when light levels are sufficient to sustain photosynthesis. In the Southern Ocean, productivity is even higher than in the Arctic, because upwelling near Antarctica continually resupplies new nutrients and there are no rivers to create a pycnocline and water stratification at any point of the year. The main group of phytoplankton in polar seas are relatively large diatoms, which can grow well in the high nutrient levels. As soon as the phytoplankton bloom starts, zooplankton start feeding on phytoplankton and their populations rapidly increase.

Temperate regions are characterized by the highest annual primary productivity. Primary productivity varies with season according to temperature (which determines the stratification of the water column), light and nutrient availability (Figure 16.12). In the winter the standing crop is low because light is limited, even though the water column is well-mixed, and nutrients are abundant in the photic zone. The spring is characterized by a large bloom of phytoplankton as light levels increase while nutrients are still abundant. The phytoplankton declines in the summer, as surface temperature increases, which creates a thermocline that prevent nutrients from returning to the surface after they have been used up by the spring bloom. Grazers (herbivorous zooplankton) also contribute to the decline of phytoplankton. There is a second, smaller phytoplankton bloom in the fall, when surface temperatures become low enough to allow mixing and the return of nutrients to the photic zone while light is still abundant enough to sustain photosynthesis. Nutrients released in the photic zone by the decomposition of organisms (chiefly the summer zooplankton bloom) also contribute to the nutrients that trigger the fall phytoplankton bloom.

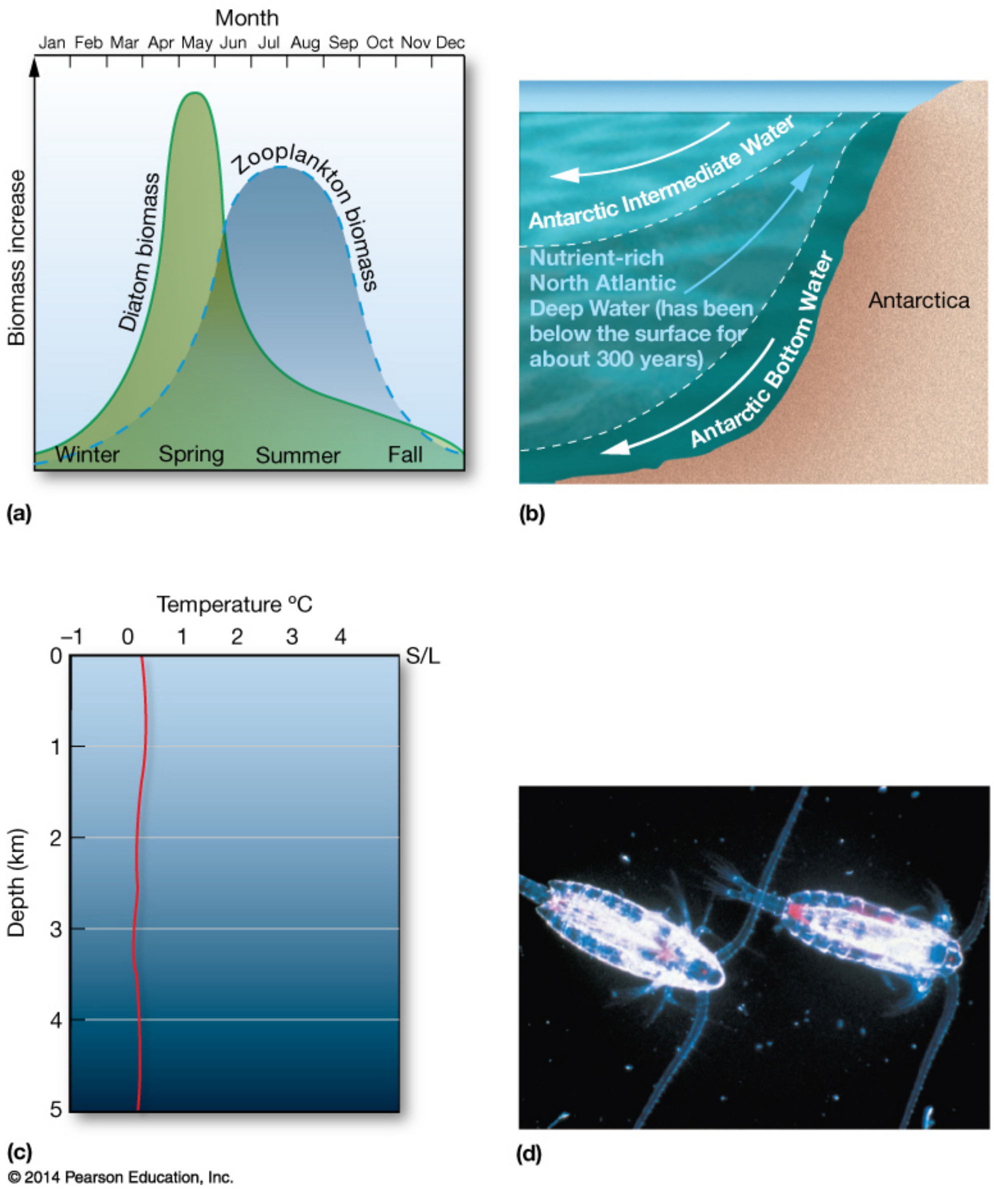


Figure 16.11. Productivity in polar oceans.

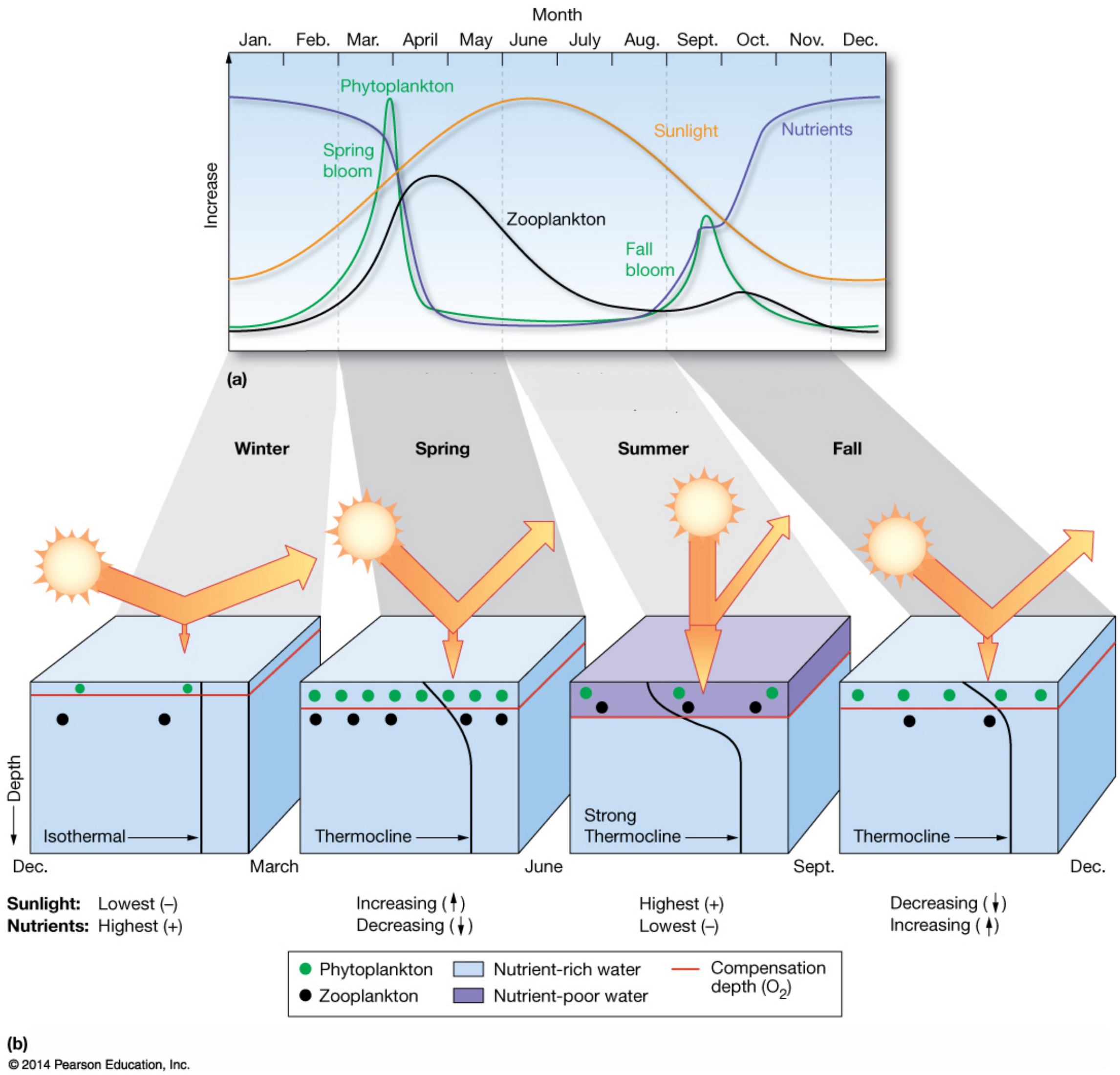


Figure 16.12. Productivity in temperate oceans.

The summer productivity of polar oceans is unmatched at other latitudes, but temperate oceans have the highest overall primary productivity, when averaged over a year (Figure 16.13). Tropical oceans typically have consistently low primary production throughout the year.

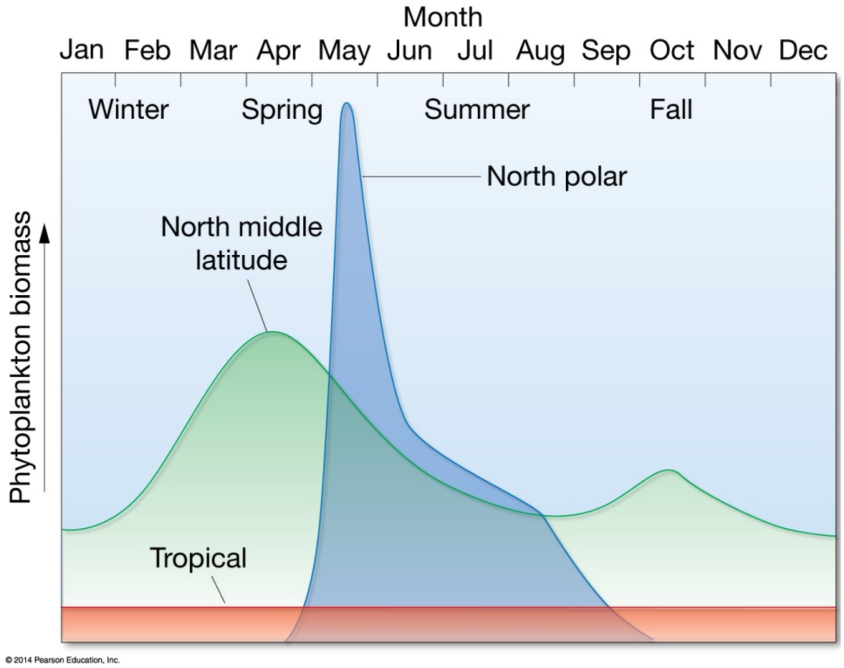


Figure 16.13. Annual patterns of primary productivity at three latitudes.

16.4. Energy and nutrient flow in marine ecosystems

In this chapter we have so far focused on primary production, but in examining the dynamics of marine ecosystems, one must consider all the organisms living in it, and how energy and materials flow through it. Biological communities include all the organisms that live together in a given area at a given time. Ecosystems include biological communities and all the non-living components of the environment.

Energy flow in marine ecosystems

Photosynthetic ecosystems are based on energy from the sun, which flows in one direction (Figure 16.14). Photosynthetic organisms convert this energy into chemical energy in the form of energy-rich organic compounds. Some of this energy is used by photosynthetic organisms for their own metabolic needs; some accumulates as biomass. Animals obtain energy by eating organic molecules and breaking them down. As organic molecules are eaten by animals, more and more energy is lost to the environment in the form of heat. New energy must constantly be supplied to an ecosystem.

Those organisms that can create organic molecules from inorganic components (either through photosynthesis or chemosynthesis) are called producers, or autotrophic organisms. Heterotrophic organisms, on the other hand, depend on organic molecules made by other organisms. These can be consumers, or decomposers. Consumers eat other organisms and can be categorized as herbivores (feed only on plant or algae), carnivores (feed only on animals), omnivores (feed on both plants and animals) or bacteriovores (feed on bacteria). Decomposers break down dead and decaying organic material without ingesting it. In this process, decomposers release the inorganic components of organic molecules, which are again available to producers at nutrients.

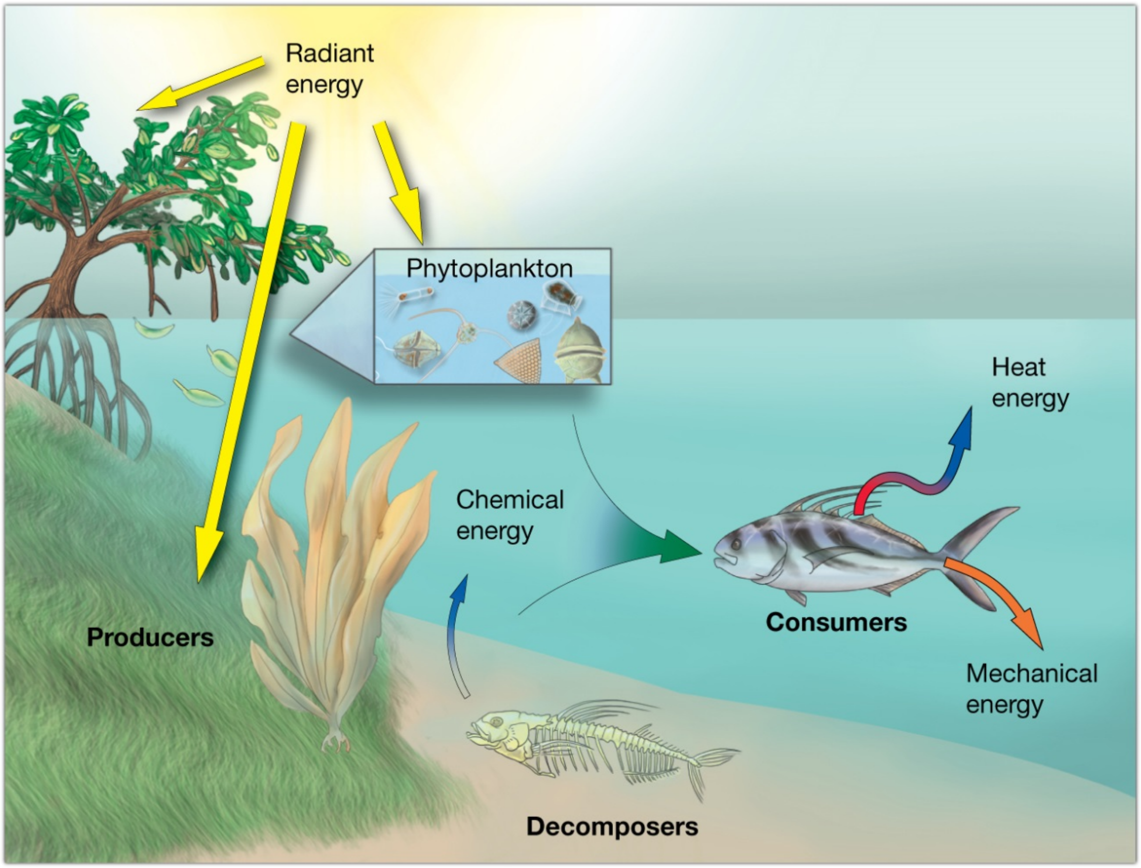


Figure 16.14. Energy flow through a marine ecosystem.

Nutrient cycles in marine ecosystems

Nutrients, as opposed to energy, are constantly recycled in ecosystems (Figure 16.15). For example, nitrogen can move from its form of atmospheric nitrogen N_2 to nitrates in the process of nitrogen fixation. Nitrates can be used by plants and algae in photosynthesis to be incorporated into amino acids (the building blocks of proteins). These amino acids are then incorporated into animal biomass as algae are eaten. When organisms break down, nitrates and other nutrients are released again the water column and can be used by other producers. Since photosynthetic rate is high near the surface, nutrients are used up quickly and tend to have low concentration. High nutrient levels found in deep water can be brought back to the surface through upwelling (Figure 16.15).

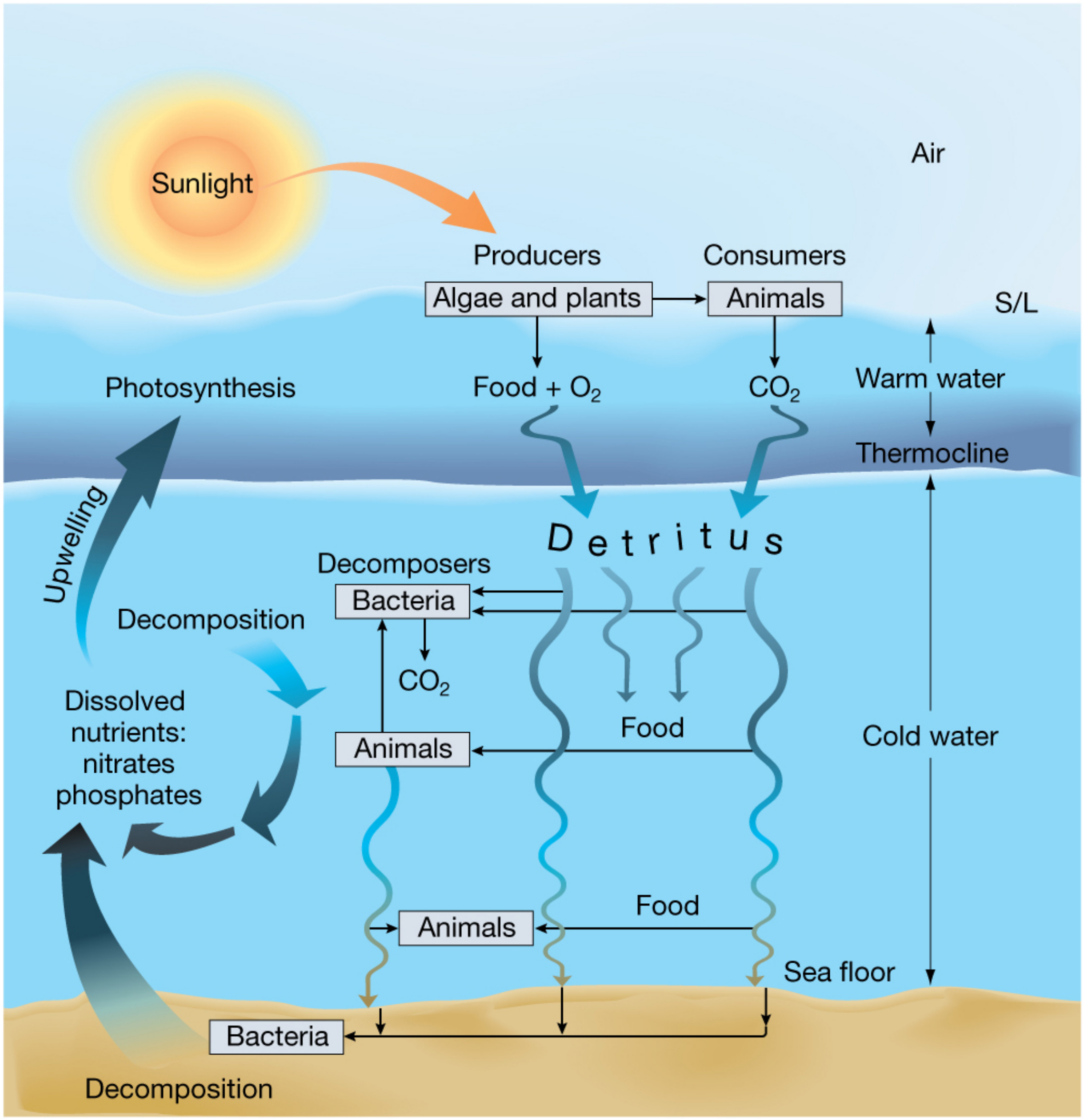


Figure 16.15. Biogeochemical cycling of matter.

Ocean feeding relationships
Marine animals obtain their food in three main ways. Filter feeders (or suspension feeders) filter plankton from seawater. Deposit feeders feed on organic matter that occurs in and on sediment. Carnivorous feeders directly capture and eat other animals. In shallow coastal waters with attached macroalgae, some animals are also specialized herbivores (or grazers).

The trophic level is the position of an organism within the trophic dynamics:

- Autotrophs form the first trophic level.
- Herbivores are the second trophic level.

- Carnivores occupy the third and higher trophic levels.
- Decomposers form the terminal level.

A food chain is the simple succession of organisms within an ecosystem based upon trophic dynamics. A food web is a more realistic representation of trophic interactions, where an organism may feed at several different trophic levels (Figure 16.16). The transfer of energy with each successive trophic level is highly inefficient and only about 10% of the energy contained in a trophic level is transferred to the next higher level (Figure 16.17). Some of the energy ingested is not assimilated and is lost in feces; of the energy assimilated, some is lost as heat in the process of respiration. Of the energy that contributes to increased biomass (increased size of the organism and increased population size through reproduction), some dies without being consumed by the next trophic level. Overall, only about 2% of the sun's energy is incorporated into producer biomass, and 10% of the energy contained in one trophic level is passed to the next level (Figure 16.18).

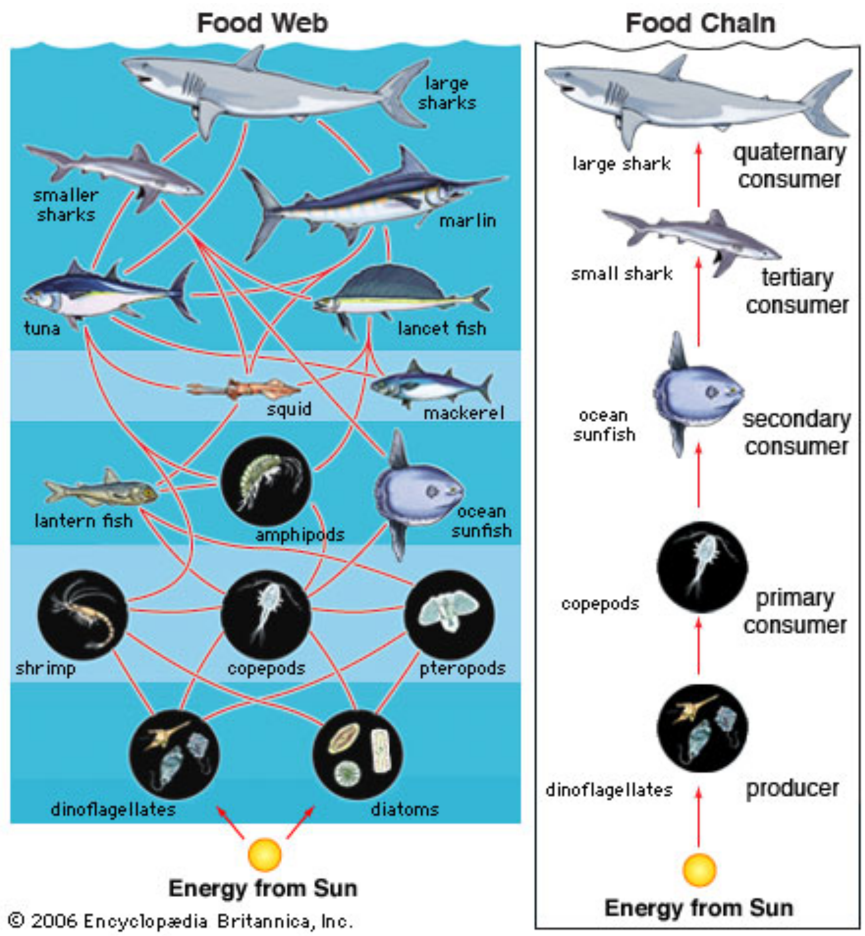


Figure 16.16. A simple marine food chain and more realistic marine food web.

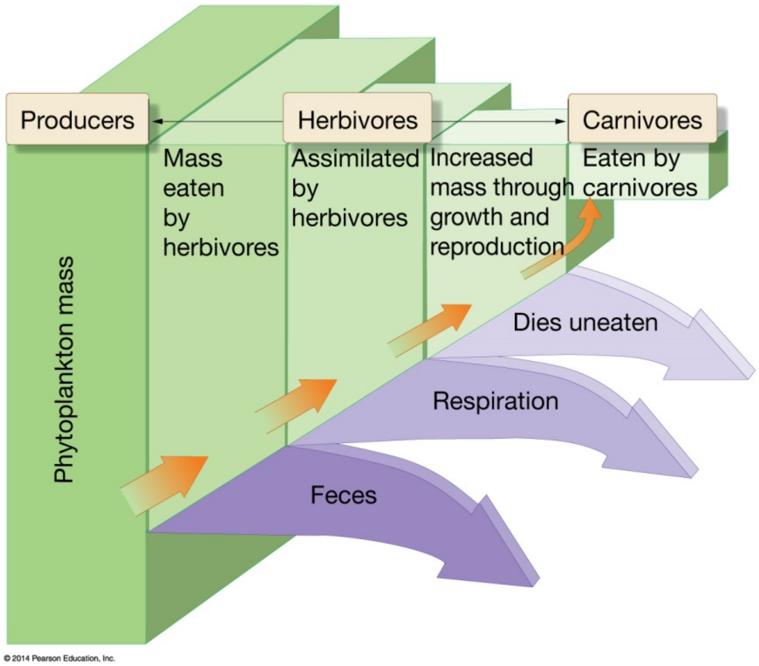


Figure 16.17. Energy losses with passage through a trophic level.

The loss of energy between trophic levels limits the number of trophic levels that can be sustained. With each trophic level, the size of individual organisms increases, but the overall biomass necessarily decreases. This is commonly represented in the form of an energy or biomass pyramid, showing the decrease in biomass and energy with each trophic level (Figure 16.19).

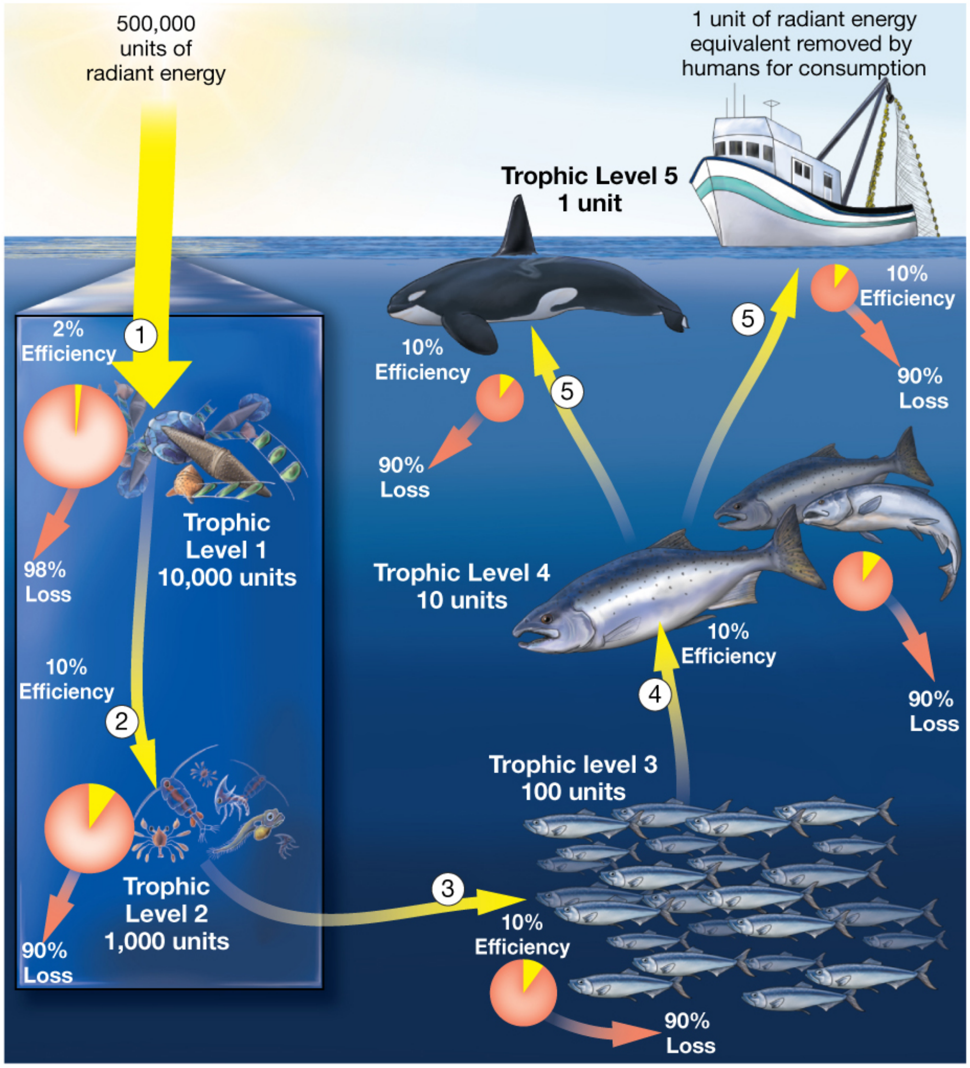


Figure 16.18. Ecosystem energy flow and efficiency.

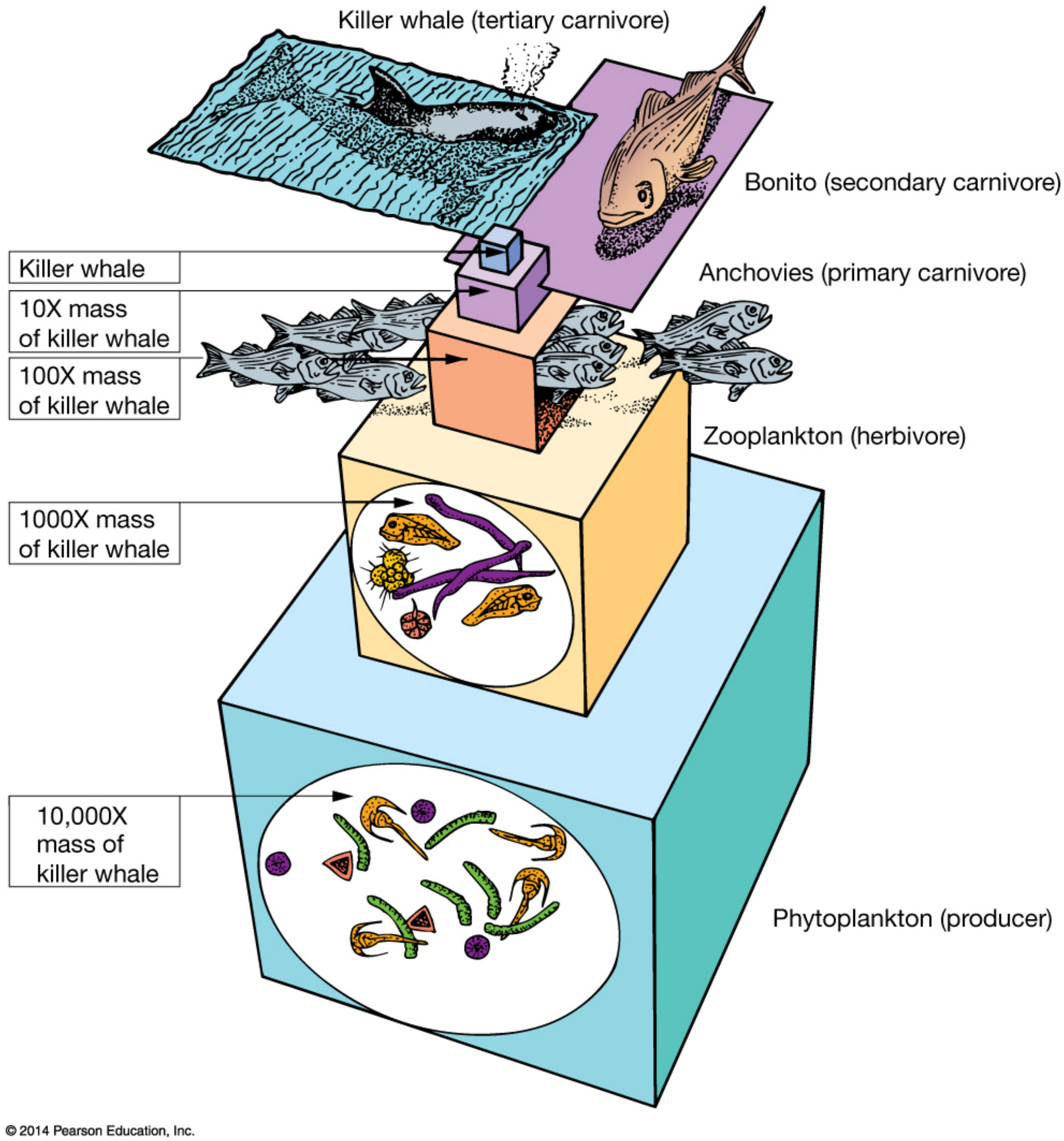


Figure 16.19. An biomass pyramid, showing the biomass and energy levels at each trophic level.

16.5. Review Questions

1. What is primary production? What unit is it typically measured in?
2. Which group of organisms contribute the most to global primary production?
3. Which deep-sea ecosystem does not rely on energy from the sun?
4. Which areas of the ocean exhibit the highest levels of primary production?
5. Why is photosynthetic rate a little lower at the surface compared to a few meters below?
6. What is the compensation depth?
7. What factors affect the compensation depth?
8. What are 4 ways in which photosynthetic organisms can adjust to varying light levels?
9. Which group of photosynthetic organism can fix nitrogen?
10. What is eutrophication?
11. Which side of ocean basins typically experience upwelling?
12. Why are nutrients usually limited in surface waters?
13. Name 2 types of seed-bearing plants found in the marine environment
14. Which group of seaweed typically can photosynthesize in the lowest light levels?
15. What are the 3 main groups of microscopic algae?
16. Are dinoflagellates autotrophic or heterotrophic?
17. How can eutrophication lead to dead zones?
18. When do zooplankton blooms occur in temperate zones?
19. What happens to the energy not transmitted to the next tropic level?

17. Marine Fisheries (Trujillo, Chap. 13 and other sources)

17.1. Fisheries and Fishing Techniques

Humans have always used the sea as a source of food. Although the sea provides a relatively small proportion of the overall food supply, this proportion has been increasing in recent years, and fisheries provide a significant portion of protein consumed in many regions of the world. Fisheries are the commercial extraction of fish or other organism from their natural environment for human purposes.

Fish have traditionally been extremely abundant in zones of high biological productivity. When the first explorers came to North America, cod was described as so abundant that it would virtually block ships. For this reason, as late as the 1880s, it was thought that fisheries could not be exhausted. As fishing fleets increased in size and technology allowed fishermen to catch more, the world catch steadily increased in the early part of 20th century, however the catch increased only from 60 million metric tons in 1970 to 86 million metric tons in 1989, despite a significantly increased fishing effort, indicating that stocks are depleted. World catch has now leveled off at approximately 90 million metric tons despite further increases in fishing efforts.

A variety of organisms are harvested from the sea. Fish provide by far the greatest tonnage, with small pelagic fish like herring, sardines and anchovies providing the largest catch, followed by demersal (closely associated with the bottom) fish like cod, hake and haddock. Invertebrates are also harvested, with mollusks (e.g. bivalves, squids), crustaceans (e.g. lobster) and echinoderms (e.g. urchins and sea cucumbers) among the most important. Seaweeds are harvested for direct consumption or for their compounds. Whales, though protected from exploitation by most countries, are still harvested to various levels by a few countries such as Japan, Norway, and St. Vincent and the Grenadines.

Most fisheries are located in zones of high biological productivity that are found near the coast on continental shelves (e.g. on the Grand Banks and Georges Banks), especially in upwelling areas, where high levels of nutrient input can sustain high primary production and large populations at higher trophic levels. Upwelling areas represent about 0.1% of the ocean surface area, yet account for about 58% of the world fishing catch. Some species, like tuna, swordfish and whales, are harvested in the pelagic zone offshore.

Different fishing techniques are employed to harvest organisms that live in different habitats. Trawling of nets on the bottom (Figure 17.1) is used extensively to harvest demersal fish, but tends reduce relief and destroy habitat. Gill nets are stretched in the water column and catch fish as they swim through the openings and get caught in it (Figure 17.2). Gill nets have been banned from many areas because of the high mortality they cause, of both targeted and non-targeted species (by-catch). Long-lining is used extensively to catch large fishes such as tuna and swordfish and involves setting out several miles of line (17.3 & 17.4). A purse seine net is a large net set around a school of fish and closed when a line passed through its bottom is drawn tight (Figure 17.5). Harpooning is used for large fish that demand a high price such as Bluefin tuna, as well as marine mammals. Traps are used to catch crustaceans and many coral reef fish. Many coral reef fish, especially in the South Pacific, are still harvested by cyanide poisoning and dynamite, with obvious impacts on human health and reef health.

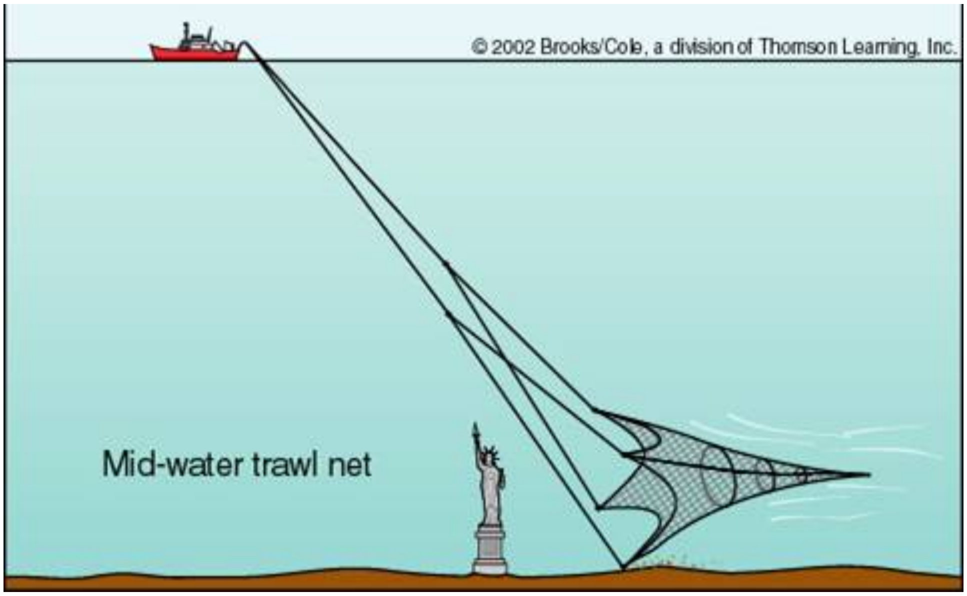


Figure 17.1. A mid-water trawl net dragged along the bottom, shown here with the statue of liberty for scale.

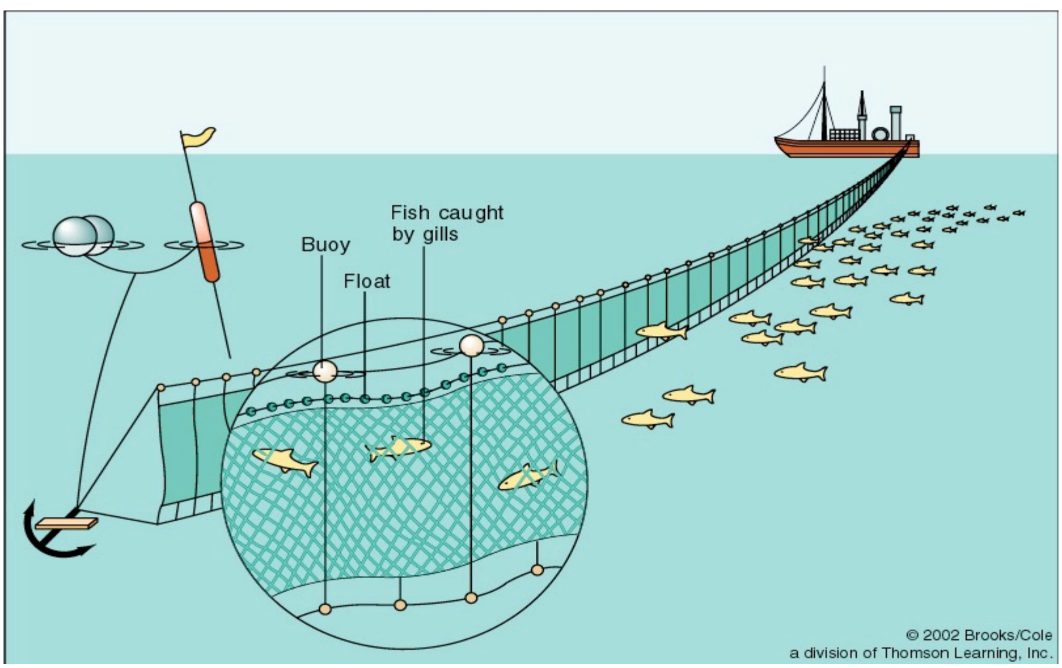


Figure 17.2. Gill nets, or drift nets, are large suspended nets that catch fish and other organisms that swim through it.

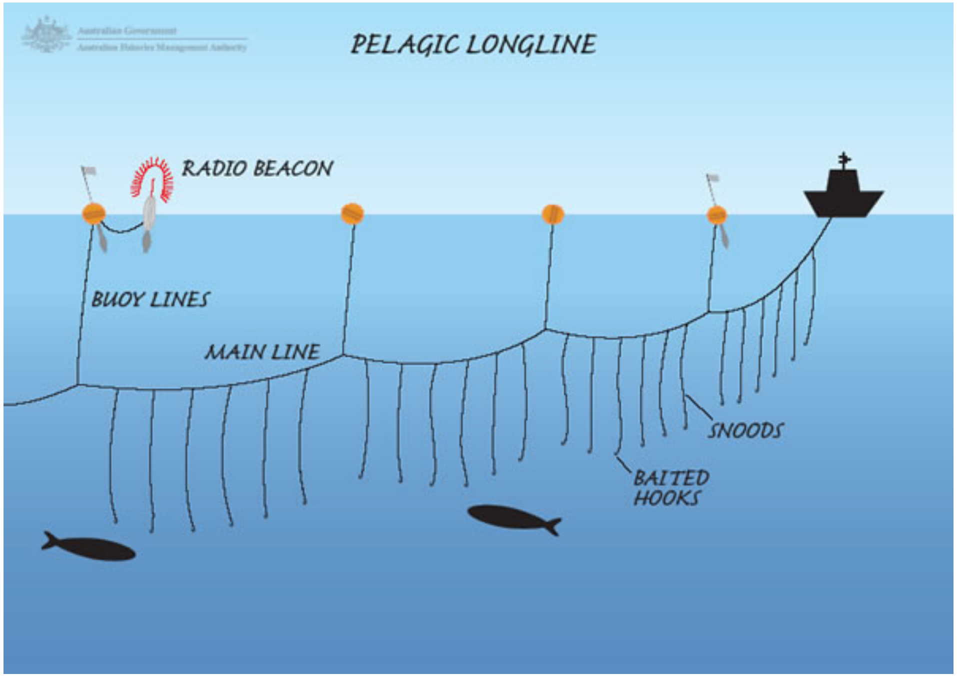


Figure 17.3. Pelagic longlines catch fishes in the water column.

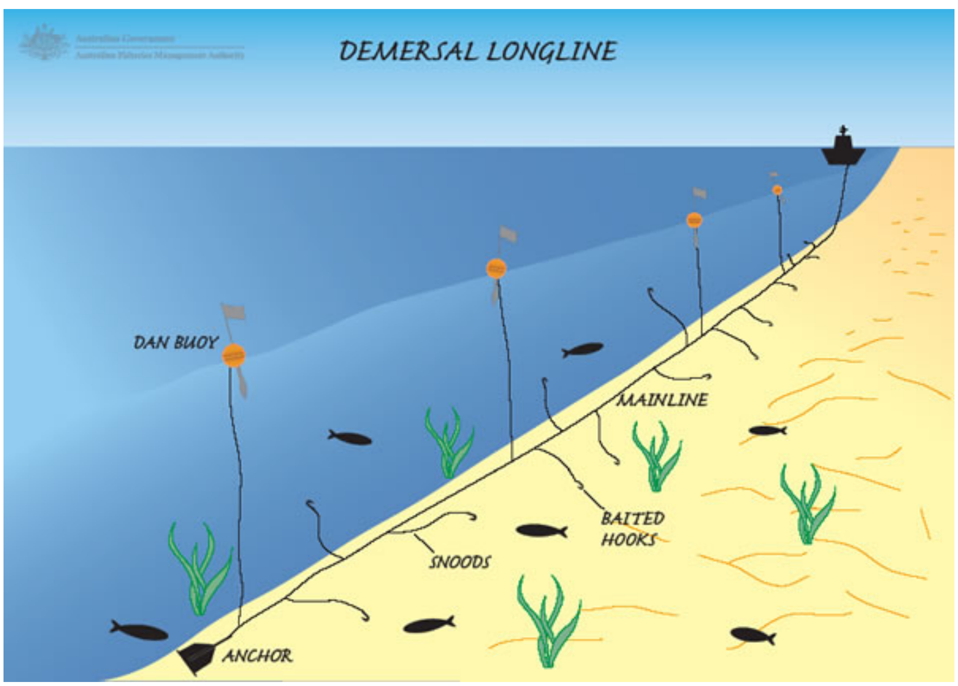


Figure 17.4. Demersal longlines catch fish in close associated with the bottom.

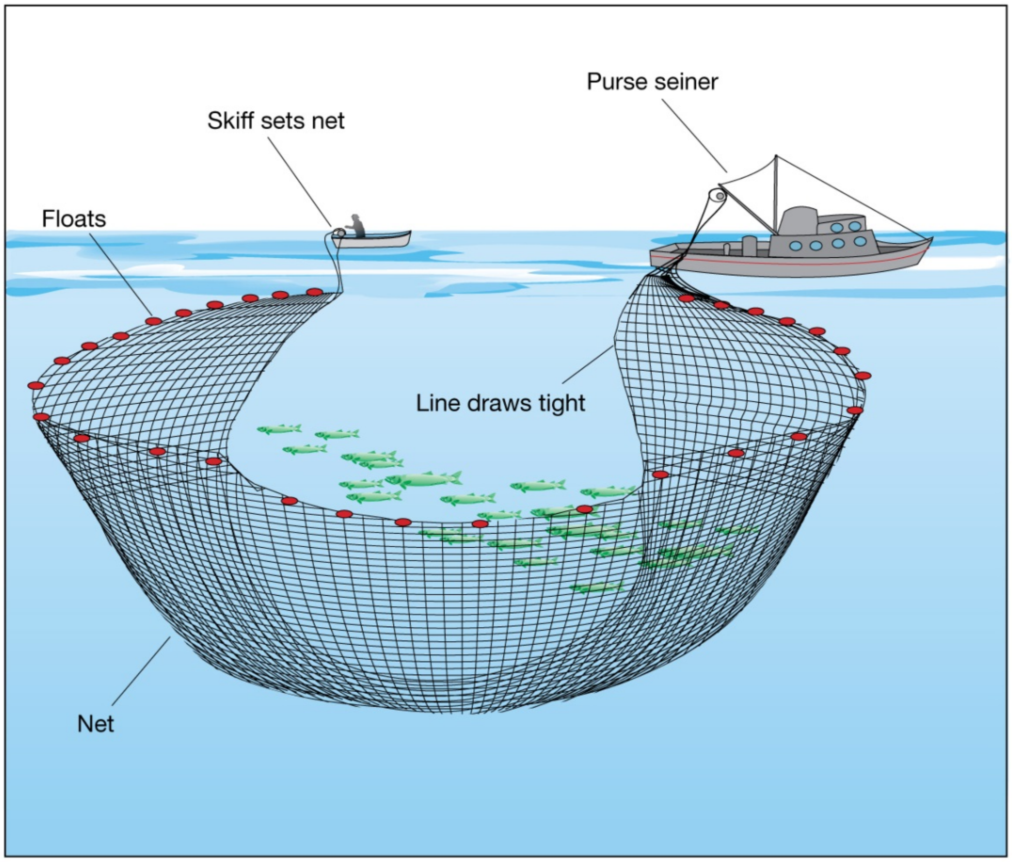


Figure 17.5. Purse seine net drawn around a school of tuna.

17.2. Overfishing

Fish (or other harvested organisms) must be allowed to reproduce if a fishery is to be maintained. If a population is harvested at a rate that exceeds its reproductive capacity, its size will diminish and fishing cannot be sustained. This is called overfishing; at present, 70-80% of the world's marine stocks are overfished and therefore are harvested in a way that is not sustainable. Also important to consider is that a fish population driven to low levels by overfishing is less resilient to stochastic mortality, ecosystem changes, and may have a lower reproductive success.

Many factors have contributed to the severe overfishing that has occurred in the last several decades. First, the world population has been steadily increasing, and there has been a shift in human food towards consuming more fish, both of which have led to a higher demand for fish and seafood. This higher demand could not be met with traditional means of fishing of hook and line, small nets and traps. Instead, the increased technology of the last century has allowed much more efficient fishing methods, which allow for the extraction of increasing biomass from the sea, meeting this higher demand. However, this increased fishing efficiency has also resulted in a rapid depletion of many fish stocks. Modern fishing vessels are more powerful than they were in the past, and can fish further offshore. Freezing holds also allow vessels to stay out at sea for weeks at a time, catching large amounts of fish offshore. Most fishing boats have electronics such as GPS as well as “fish finders” that use reflected sound to locate schools of fish. Nets are now much more effective than they were in the past with new technology like monofilament. In the North Pacific, 1500 fishing vessels set out 37 000 km of gill nets every day; at least 20% of their catch is thrown away because it consists of organisms that are badly damaged or cannot be sold. Some species like the Bluefin tuna command such a high price (up to $260/kg) that it justifies great expenses to catch them, including spotter planes that fly over vast areas of blue ocean to locate schools and then transmit the position to the fishing fleet. All this new technology has allowed fishermen to exploit most fish stocks at rates much higher than has ever been possible in the past, and beyond their reproductive capacity. Moreover, fishermen can now easily find a market for all this excess fish because of modern means of shipping (e.g. air transport) opens up the international market. Whereas fishermen would have traditionally stopped fishing once they had flooded the local market, modern fishermen can ship across the world in a matter of hours and therefore there is virtually always a market for their catch. However, shipping internationally increases costs, and fishermen must fish even more to make more profit, contributing even more to overfishing.

An overexploited population shows several signs, including a decreased size of the population (as estimated by scientific surveys), a decreased catch per unit effort, and a decreased average size of fish, as fish are harvested before they can grow big.

The case of the North Atlantic Cod

Problems of overfishing have been felt very acutely in the Western North Atlantic, where fishes used to be extremely abundant and in some places have been depleted possibly beyond recovery. The North Atlantic Cod of Newfoundland is one of the most striking cases of overfishing. Cod was so abundant when the first Europeans came to North America that they could lower a bucket over the side of their boat and catch fish in this way. Around 1600, low-scale commercial harvesting of cod started with the use of hand line and long lines. Cod fishing remained small-scale with these traditional methods for several hundred years. However in 1895 trawl nets started being used, and around 1950 huge factory ships were fishing on the Grand Banks off Newfoundland. Many of these ships came from overseas, fished outside the Canadian territorial waters (12 nm) and by processing fish right away on board, could conserve fish until their return overseas. Between 1966 and 1976, fishing capacity increased by 500%, but the associated catch only increased by 15%, a net decrease in catch per unit effort. Under such high fishing pressure, the Newfoundland cod plummeted in the latter part of the 20th century. Until 1950, catch was 250 000 tons/years. Catch was at an all-time high in 1968 at 810 000 tons, but quickly dropped to less than 300 000 tons by 1975 and then to 139 000 in 1978. Seeing catch dwindle, the Canadian government followed the US and in 1976 declared their Exclusive Economic Zone to extend 200 nm, to prevent foreign vessels exploit these resources. By the mid-1980s, the catch was back to about the same levels at 1950, but many more vessels (lower catch per unit effort), indicating the population was overexploited. Throughout the early 1980s, government scientists recorded many clear signs of overfishing, but their recommendations for reduced quotas were repeatedly ignored by the government. In 1990, an independent review of the state of the Northern Cod stock concluded that the population, biomass, spawning population and spawning biomass were all in decline. By 1992, the stocks were so low that the Canadian Department of Fisheries and Oceans (DFO) declared an unprecedented complete ban on cod fishing in Newfoundland. It is estimated that the total population was 400 000 tons in 1990, and plummeted to 1700 tons by 1994. Though the fishery has been closed since 1992, the population has not yet increased to harvestable levels, and may never come back to those levels. The case of the North Atlantic Cod, as well as many other overexploited species, demonstrates the importance of better managing our fish stocks, or using alternative methods of producing fish biomass (e.g. aquaculture) in a sustainable way.

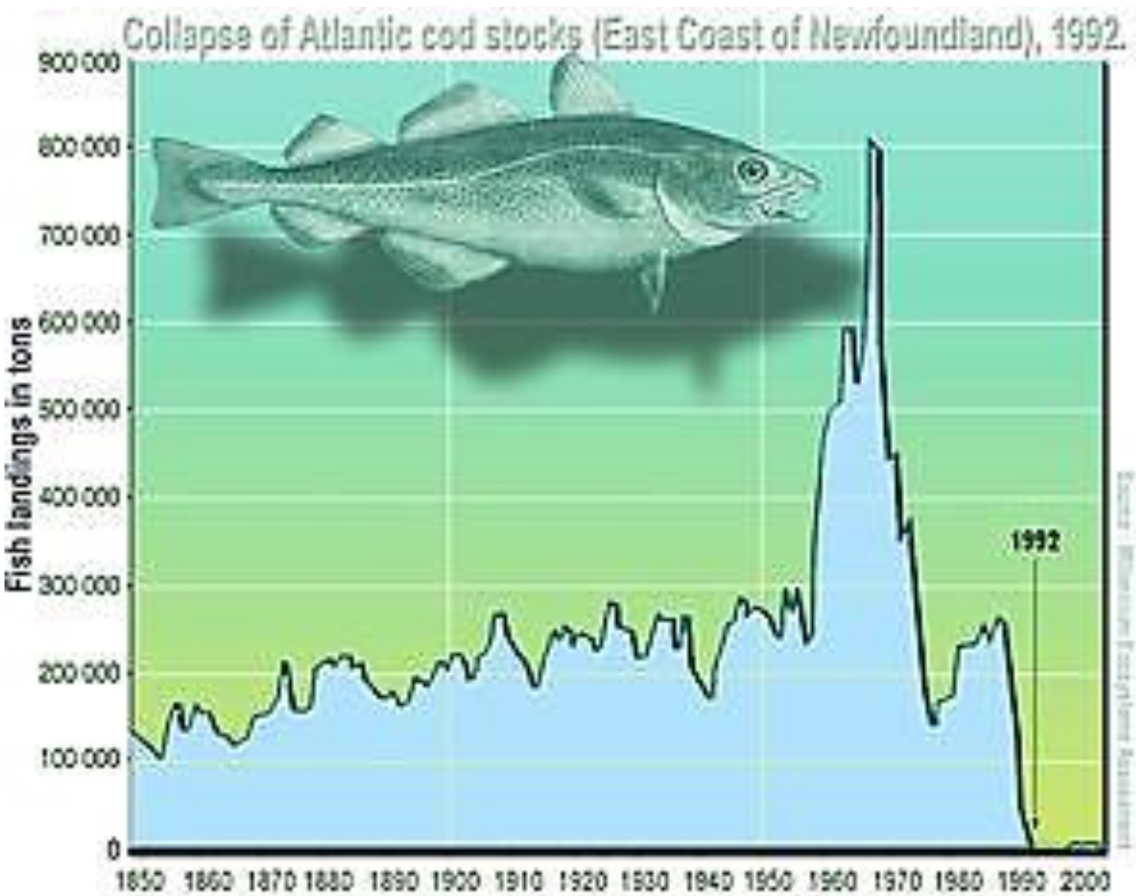


Figure 17.6. Cod landings from 1850 to 2000.

Coral reef fisheries
Most productive fisheries of the world are located in temperate zones or upwelling areas, where there is a good supply of nutrients to the system, which can sustain high levels of primary productivity and higher trophic levels. Coral reef systems, on the other hand, are fairly closed systems where there is a very tight recycling of nutrients. Harvesting coral reef fish takes away biomass that is then not decomposed and whose nutrients are not returned to the system. Moreover, coral reef fishes that are harvested are typically slow-growing, long-lived, and produce few offspring. For that reason, they can sustain much

lower rates of exploitation than fast-growing species in nutrient-rich environments. Many coral reef fish, such as groupers, are overexploited.

Overfishing of large predators
The most profitable fishing targets tend to be large predatory fishes such as cod, tunas and sharks, and these species have been the focus of intense fisheries. Recent studies have shown that about 90% of large predatory fishes have been removed from ocean ecosystems. This has led to important ecosystem changes. For example, the severe decline of the North Atlantic cod has released predation on benthic invertebrates such as shrimp, crabs and lobsters, and on planktivorous fishes. The populations of these groups have greatly increased in the last few decades. The increase in planktivorous fishes has led to a decrease in zooplankton biomass, which has allowed for an increase in phytoplankton abundance. This is an example of a trophic cascade, where the removal of a predator has alternating effects on lower trophic levels. This tends to happen in areas with simple food webs, or where entire trophic levels have been suppressed, as in the case for many fisheries (e.g. in the North Atlantic, other species of predatory demersal fishes have also been overexploited alongside cod). Similarly, the large declines in predatory sharks on the east coast of North America has led to a sharp increase in the abundance of cownose rays, which have decimated populations of bay scallops.

Fishing down the food web
Recent research has documented a systematic decline in the average size and trophic level of fishes targeted by fisheries. As big predatory fishes become overfished, fisheries move to the next lower trophic level below. This has been found across many fisheries in all areas of the oceans and has been called "fishing down the food web". It is often difficult to document such changes, unfortunately, because the concept of what is "normal" tends to shift over time with each generation. While an 85 year-old fisherman might think that current catches are small, his 25 year-old great grandson might think, based on his experience, that this year's catches is one of the biggest that's been seen. Such shifts with generations have been called shifting baselines (Figure 17.7). This is an important concept in conservation, as it is important to try to understand what a pristine ecosystem would look like in order to make better management decisions. The concept of shifting baselines was coined by Daniel Pauly, a fisheries biologist at the University of British Columbia. He summarized the problem in this way: "We transform the world, but we don't remember it. We adjust our baseline to the new level, and we don't recall what was there". To better understand changes that have occurred over decades or centuries of human exploitation, researchers have to resort to non-conventional sources of data such as logbooks, archeological evidence, and notes from newspapers and other writings. In this way, scientists have reconstructed a large decline in the average size of fish caught on fishing charters in Florida Keys (Figure 17.8), a clear example of fishing down the food web.

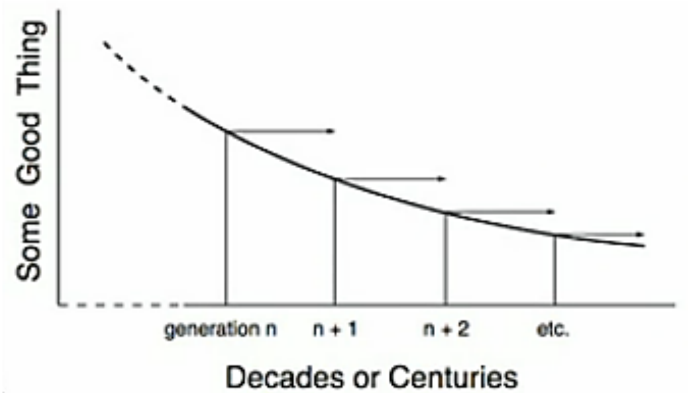


Figure 17.7. The concept of shifting baselines, showing that each generation tends to evaluate what's normal based on their own experience. In this example, people in generation n+2 see a small decline in fish stocks (or some other "good thing) over their lifetime, but they would likely be much more alarmed if they realized the severe declines that have occurred since their grandparents were the same age.

Figure 17.8. Fishing down the food web in the Florida Keys, reconstructed from historical photos.

17.3. Bycatch

Bycatch, or incidental catch, is the capture of organisms that are not targeted by a fishery. Because fishermen do not have permits to keep those organisms, they must return them to the ocean, but most have died by the time they are thrown back overboard. On average, about a quarter of the world's marine fisheries catch is made up of bycatch. Bycatch includes many marine fishes, as well as birds, turtles, sharks and dolphins. Because of this, a lot of research has been done to find harvest methods that reduce bycatch (e.g. Turtle Exclusion Devices or Bycatch Reduction Devices, the designs of which vary with fishing gear type). Some fishing methods have higher incidence of bycatch than others. Purse seine nets (Figure 17.5), which are commonly used to harvest tuna, often catch dolphins that swim with the school. Driftnets (or gillnets; Figure 17.2) indiscriminately catch most organisms that fit in the mesh size and have a high rate of bycatch, and have been banned by international law. Longlines, while they target species such as tuna and swordfish, also catch a lot of seabirds and turtles, which often have drowned by the time they are recovered (Figure 17.9)

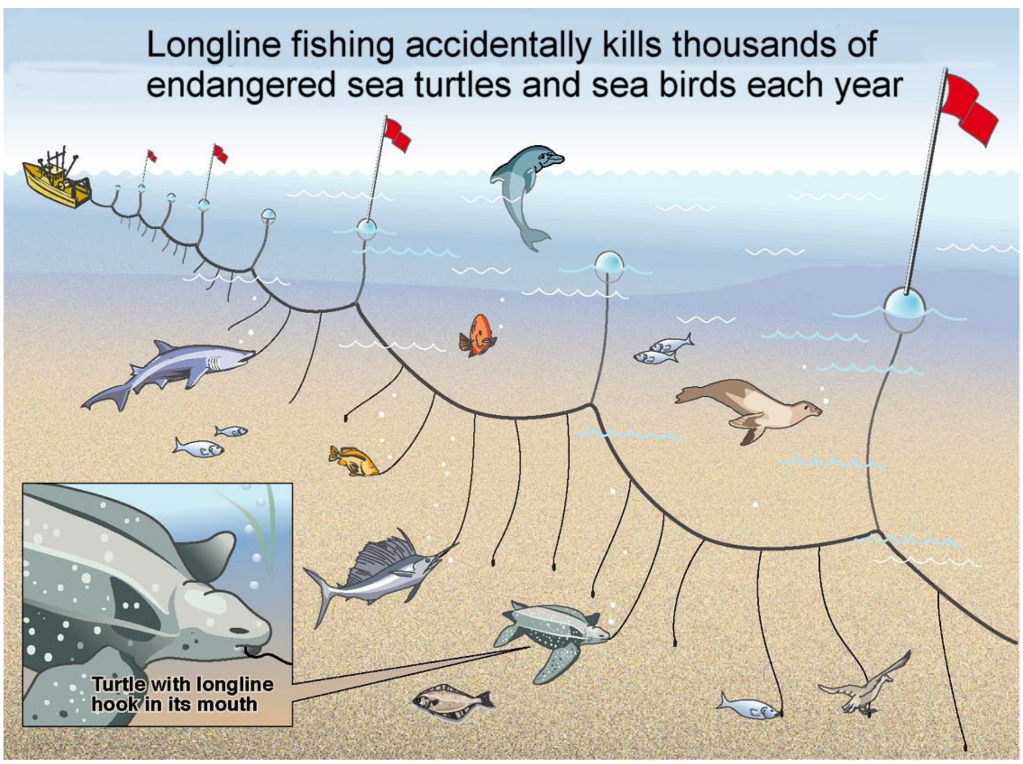


Figure 17.9. Bycatch on longlines.

17.4. Fisheries management

Though all fishermen understand that fish must be allowed to reproduce in order to sustain their population, it is typically up to scientists to study fish populations and determine sustainable levels of harvesting, and to local government to set quotas and enforce regulations.

Fisheries biologists must understand many aspects of a given species' physiology and ecology in order to establish sustainable harvest levels. One must know what environment the species lives in, including its habitat and food source. It is important to know the growth rate (we can determine the age of fish from rings on their vertebrae, scales, or otoliths) and the size at sexual maturity, to establish minimum landing sizes which allow individuals to reproduce at least once. It is also important to know the spawning season, the fecundity (number of eggs per spawn), survival rates, and life span of the species.

Estimates of population size must be made periodically. The easiest data to obtain is catch data from fishermen, on the assumption that catch is proportional to the population size. However, such data can be biased by misreporting of actual catch, and by the fact that bycatch is not reported even though most organisms caught as bycatch are dead and removed from the population. Scientists can also estimate stocks by tag and release programs, which also provides data on growth rate and movement, as long as the tagging process does not induce increased mortality or change in behavior. Stocks can be estimated by measuring the levels of eggs and larvae in the water, with the assumption that they directly correlate with the adult population. Finally, acoustics can be used to survey large areas and estimate biomass.

Once basic data on population dynamics of a given species has been gathered, biologists can calculate the Maximum Sustainable Yield (MSY), the theoretical quantity of fish that can be harvested each year without significantly interfering with the regeneration of fish stocks (Figure 17.10.). Fishing above the MSY leads to decreased catches as the population size decreases. Traditional fisheries management has aimed to regulate fishing effort right around the MSY, but this leads to two major problems. First, random fluctuations in stock recruitment leads to changes in the actual MSY, so fisheries that harvest at the MSY rates calculated based on years of overabundance might be overexploiting the stock a few years later. Second, maximum economic gain is usually achieved at fishing rates below the MSY, because costs of fishing tend to increase linearly while fishing revenues tend to follow a curve (Figure 17.11). Setting quotas lower than the MSY should therefore be more profitable; this level of fishing has been called the Optimal Yield.

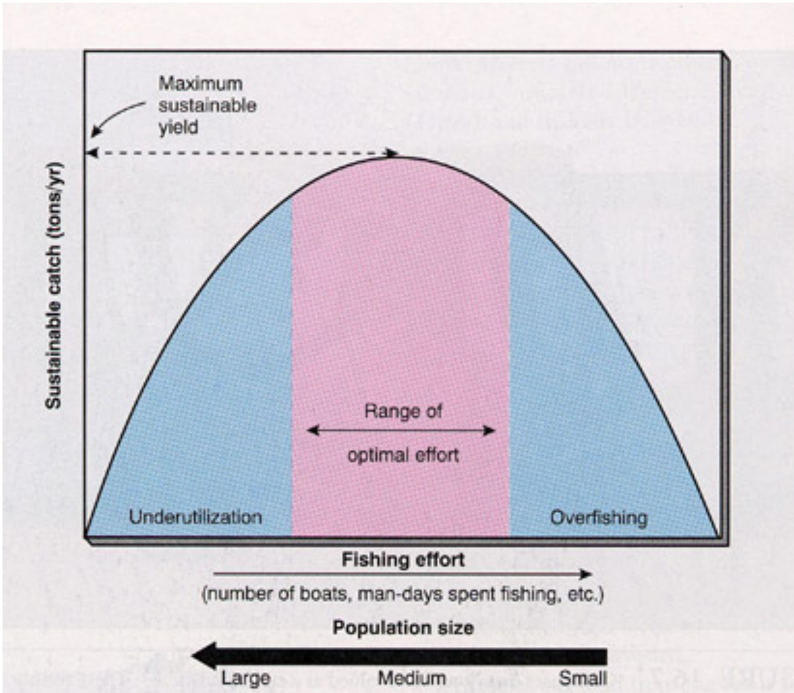


Figure 17.10. Maximum sustainable yield and range of optimal effort.

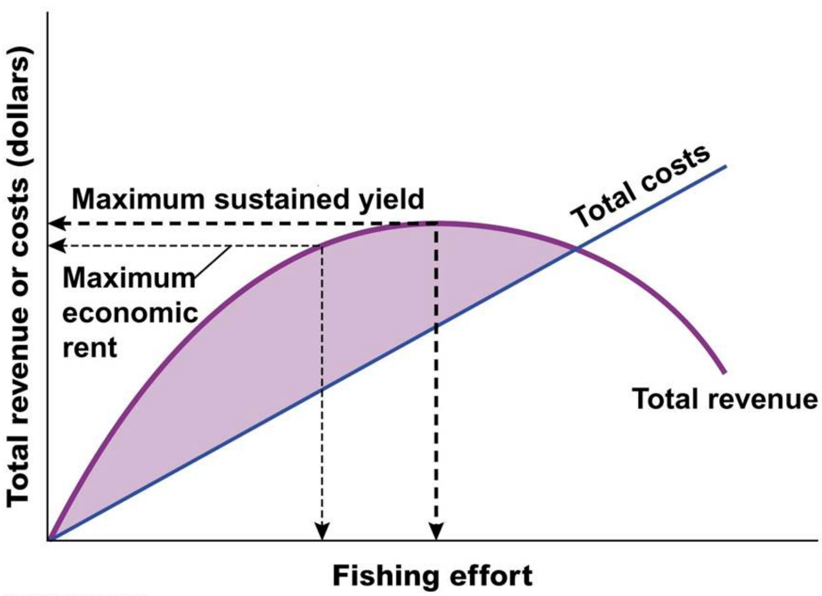


Figure 17.11. Fishing costs and revenues at various levels of fishing effort. Maximum profit is achieved at the Optimal Yield, which is below the Maximum Sustainable Yield.

Once the Maximum Sustainable Yield and/or the Optimal Yield has been calculated, the government can set limits such as quotas, total allowable catch (for individuals per season) and minimum landing size. Governments can also decide to close the fishery during the reproductive season, or establish regulations on fishing gear to minimize by-catch, or can establish refuges (e.g. marine protected areas). The specific regulations depend on biological factors of the species, as well as social and political factors. While traditional fisheries management has focused on single-species regulations, in recent decades more efforts have been made towards ecosystem-based fisheries management which aim to be a more comprehensive approach to managing fisheries by looking at entire ecosystems. Another big change is the recent increase in the establishment of marine protected areas in which harvesting is prohibited. Fish stocks in MPAs can grow and eventually fish migrate out of MPAs (a process called the spillover effect) where they can be caught. Scientists estimate that about 30% of the world's oceans should be protected in such a way to maintain sustainable fisheries. A much lower proportion of the world's oceans are currently protected from harvesting, but many large MPAs have been created in the past decade.

17.5. Seafood choices

Consumer demand can drive market forces to overharvest species, or to change practices and build more sustainable fisheries. Several groups now independently analyze the status of marine fish stocks to guide consumer choices toward sustainable options for fish and other seafood. A dominant scheme is the Seafood Watch from the Monterey Bay Aquarium. Its seafood classification can easily be found on wallet cards (Figure 17.12) or through a free smartphone app. The Marine Stewardship Council gives its seal of approval to fisheries that it deems sustainable so that consumers can choose to buy only seafood that has the MSC logo (Figure 17.13). While no such classification scheme is perfect, they present a smart start to consumer education that can help change demand and markets.

Best Choices	Good Alternatives	Avoid
Arctic Char (farmed)	Basa/Pangasius/Swai (farmed)	Caviar, Sturgeon* (imported wild)
Barramundi (US farmed)	Caviar, Sturgeon (US farmed)	Chilean Seabass/Toothfish*
Catfish (US farmed)	Clams (wild)	Cobia (imported farmed)
Clams (farmed)	Cod: Atlantic (imported)	Cod: Atlantic (Canada & US)
Cod: Pacific (US non-trawled)	Cod: Pacific (US trawled)	Crab: King (imported)
Crab: Dungeness, Stone	Crab: Blue*, King (US), Snow	Flounders, Halibut, Soles (US Atlanitc except summer flounder)
Halibut: Pacific (US)	Flounders, Soles (Pacific)	Groupers (US Atlantic)*
Lobster: California Spiny (US)	Flounder: Summer (US Atlantic)	Lobster: Spiny (Brazil)
Mussels (farmed)	Grouper: Black, Red (US Gulf of Mexico)	Mahi mahi (imported longline)
Oysters (farmed)	Herring: Atlantic	Marlin: Blue, Striped (Pacific)*
Saberfish/Black Cod (Alaska & Canada)	Lobster: American/Maine	Monkfish
Salmon (Alaska wild)	Mahi mahi (US)	Orange Roughy*
Sardines: Pacific (US)	Oysters (wild)	Salmon (farmed, including Atlantic)*
Scallops (farmed)	Pollock: Alaska (US)	Sharks* & Skates
Shrimp: Pink (OR)	Saberfish/Black Cod (CA, OR, WA)	Shrimp (imported)
Striped Bass (farmed & wild*)	Salmon (CA, OR, WA*, wild)	Snapper: Red (US Gulf of Mexico)
Tilapia (US farmed)	Scallops (wild)	Swordfish (imported)*
Trout: Rainbow (US farmed)	Shrimp (US, Canada)	Tilapia (Asia farmed)
Tuna: Albacore (Canada & US Pacific, troll/pole)	Squid	Tuna: Albacore*, Bigeye*, Skipjack, Tongol, Yellowfin* (except troll/pole)
Tuna: Skipjack, Yellowfin (US troll/pole)	Swai, Basa (farmer)	Tuna: Bluefin*
	Swordfish (US)*	Tuna: Canned (except troll/pole)
	Tilapia (Central & South America farmed)	
	Tuna: Bigeye, Tongoi, Yellowtail (US farmed)	

Support Ocean-friendly Seafood

Best Choices are abundant, well-managed and caught or farmed in environmental friendly ways.

Good Alternatives are an option, but there are concerns with how they're caught or farmed–or with the health of their habitat due to other human impacts.

Avoid for now as these items are caught or farmed in ways that harm other marine life or the environment.

Key
CA = California OR = Oregeon
WA = Washington
*Limit consumption due to concerns about mercury or other contaminants. Visit www.edf.org/seafoodhealth
Contaminant information provided by: ENVIRONMENTAL DEFENSE FUND

Seafood may appear in more than one column

Figure 17.12. Seafood watch classification of seafood choices.

Figure 17.13. The logo of the Marine Stewardship Council, which indicates seafood that is harvested sustainably.

17.6. Review Questions

1. On a global scale, what areas are the most productive fisheries?
2. Name 2 types of fish harvested in the pelagic zone
3. Which method of fishing is most commonly used for catching demersal (bottom dwelling fish), and destroys significant benthic habitats?
4. What is overfishing, and why do fish populations have a minimum size?
5. Name 4 causes of overfishing
6. Name 3 signs of overfishing
7. What is the term used to designate the theoretical maximum quantity of fish that can be harvested each year without significantly interfering with the regeneration of fish stocks?
8. At what fishing level is maximum profitability obtained?
9. Name 5 ways to regulate fisheries towards being sustainable
10. Name 5 methods of stock assessment?
11. What is by-catch?
12. Ultimately, why are coral reef fisheries less sustainable than upwelling fisheries?
13. What is a trophic cascade?
14. What is the concept of shifting baselines?
15. What is meant by "fishing down the food web"?

18. Aquaculture

It is clear that the world's fisheries are in crisis, and global catch has leveled off despite increased fishing effort (Figure 18.1). Increasingly, many are making the case that farming fish and other marine organisms offers a solution to meeting the demand for seafood of the world's growing population. Aquaculture now accounts for roughly one-third of the world's total supply of fish protein (Figure 18.1) and undoubtedly its contribution to seafood supplies will increase in the future.

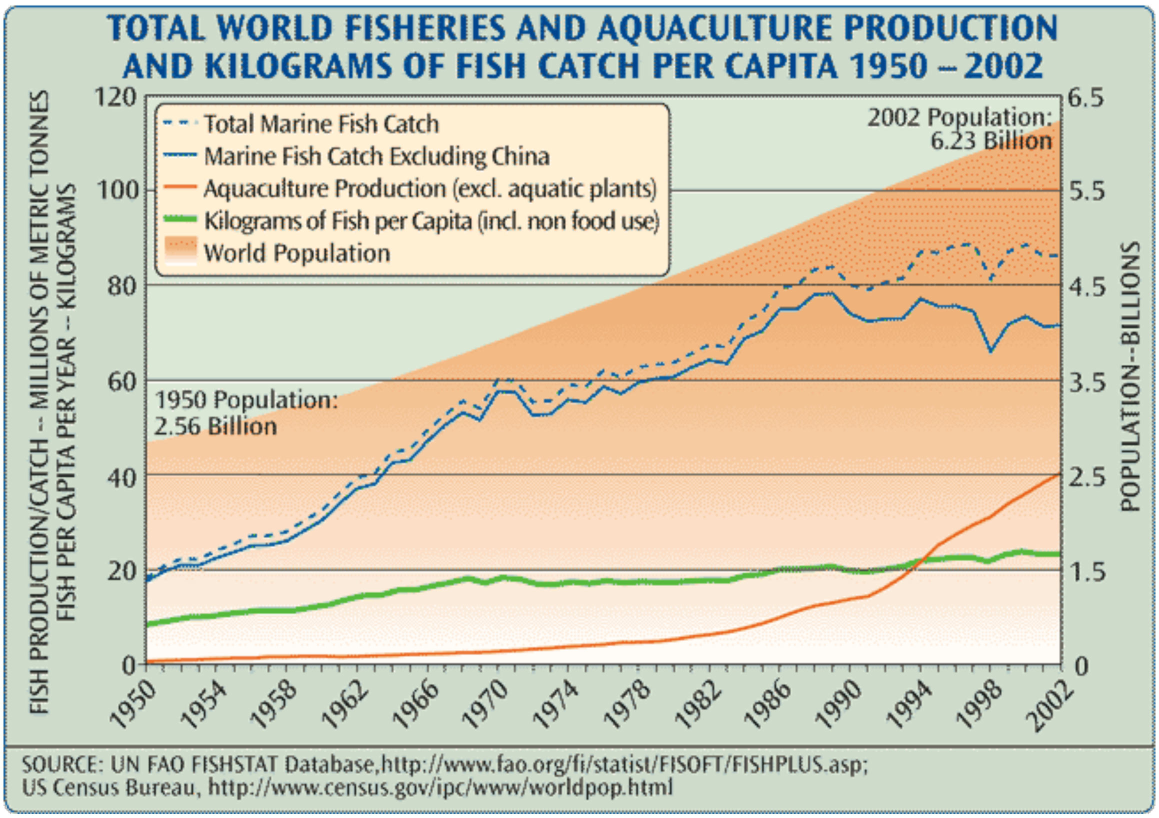


Figure 18.1.World marine fish catch, aquaculture production and human population.

While aquaculture has the potential to become a sustainable practice that can supplement capture fisheries, unsustainable aquaculture development could exacerbate the problems and create new ones, damaging our important and already-stressed coastal areas.

This section explores current practices aquaculture, environmental issues associated with these practices, and sustainable alternatives. It is mostly based on a report by the non-profit organization SeaWeb, as well as other sources.

18.1. Aquaculture history and current world trends

Aquaculture is the farming of aquatic organisms including fish, mollusks, crustaceans and aquatic plants. It is the fastest growing sector of the world food economy, increasing by more than 10% per year. While the world community has only recently viewed aquaculture as a potential solution to the dilemma of depleted oceans, it is by no means a new practice. It is not clear when aquaculture really began, but ancient Chinese manuscripts from the 5th century B.C. indicate the Chinese practiced fish culture. Egyptian hieroglyphics also indicate the Egyptians may have attempted fish farming as early at 2000 BC. Romans also developed aquaculture practices for cultivating oysters.

All of the early forms of aquaculture differed greatly from much of the aquaculture practiced today, with the major difference being that ancient aquaculture simply involved harvesting immature organisms and transferring them to an artificially created environment that is favorable to their growth. Fish farming in its modern form (i.e. involving successful reproduction in captivity) was first introduced in 1733 in Germany. Initially this "fish farming" was limited to freshwater fish. In the 20th century new techniques were developed to successfully breed saltwater species.

Today, the vast majority of aquaculture takes place in Asia. Most farmed fish and shellfish are grown in traditional small-scale systems that benefit local communities and minimize the environmental impact. Utilizing simple culture technologies and minimal inputs, these systems have been used for centuries. The net contribution of these traditional aquaculture systems can be great as they offer many benefits, including food security in developing nations.

However, an emerging trend is toward the increased development of farming high-value carnivorous fish species using intense aquaculture systems that often are environmentally damaging. Farming fish on an industrial scale, especially of carnivorous fish, is rapidly expanding. Largely controlled by multinational corporations, industrialized farming of carnivorous fish such as salmon requires the intensive use of resources and exports problems to the surrounding environment, often resulting in large environmental impacts and social conflicts.

18.2. Aquaculture Techniques

So far, freshwater aquaculture has been more successful than marine aquaculture. Profitable aquaculture species must be luxury items that demand a high price or species that grow fast and are easy to grow. Ideal aquaculture organisms should be hardy and resistant to disease and parasites, and grow well in crowded conditions. They should have a high degree of growth per unit intake, and can be brought to maturity and reproduce in captivity. Successful marine aquacultures include salmon, shrimps, oysters and mussels.

In order to successfully raise organisms in aquaculture, one must know a range of parameters on that species. We must know the ideal water chemistry, i.e. temperature and salinity for optimal growth; the optimal stocking density, i.e. the number of organisms per area that can be grown successfully; the habitat and behavior; the hardiness and growth rate; diet; the reproductive and life cycles.

Specific aquaculture techniques depend on the species raised, but there are generally three main steps in raising organisms. First, adult animals must spawn. This can be done naturally or artificially in a laboratory, or if mature animals cannot reproduce in captivity, larvae must be caught in the wild. Second, larvae are reared in a laboratory, under optimal conditions and without predators. This greatly increases the survival of larvae compared to the wild. Finally, the organisms must be raised to market size. Animals may be raised in laboratories, or in ponds or cages. Some of these steps may be skipped depending on the species and the intent of the culture. For example, shrimp larvae may be raised and then released in the natural environment to supplement natural populations, skipping the third step of raising them to market size.

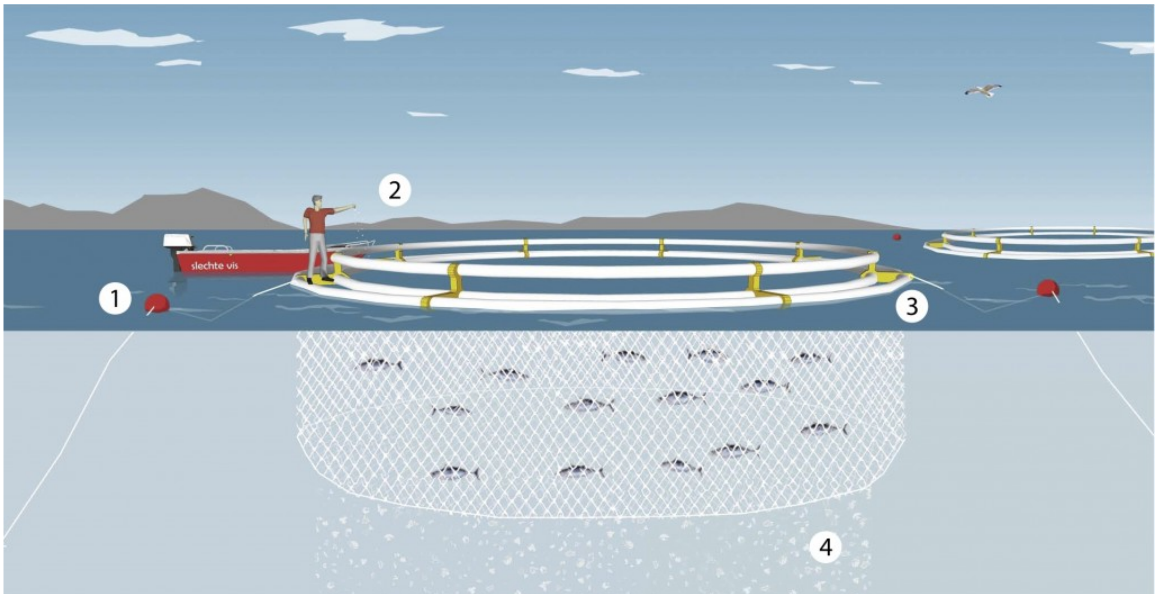
Intensive aquaculture is becoming increasingly abundant. It involves high stocking densities, external input of food, and mechanical aeration. There are three main methods of intensive aquaculture:

Open-sea cage aquaculture

Open-sea cage aquaculture refers to the rearing of aquatic species, within enclosures in natural waterways (Figure 18.2). Open systems are being implemented in a wide range of environments including freshwater rivers, brackish estuaries and coastal marine regions. Floating mesh cages are anchored to the seafloor and vary in size depending on the scale of operation and the species cultured.

Juvenile stock is sourced either from hatcheries or wild populations, and grown out in pens until a marketable size has been reached. Finfish grown in open systems are primarily carnivorous species which are fed on a diet of fishmeal (pellets comprising small schooling fish species).

There are numerous concerns associated with the expansion of open sea-cage aquaculture. One of the primary objections relates to the requirement of fishmeal to feed carnivorous species. In some cases the conversion ratio may be in the order of more than 5kg of fishmeal to produce just 1kg of marketable fish. Other significant issues include increased disease and parasite transmission due to high fish densities, the risk of escape and interbreeding with wild populations, and reduced water quality resulting from the accumulation of fecal waste.



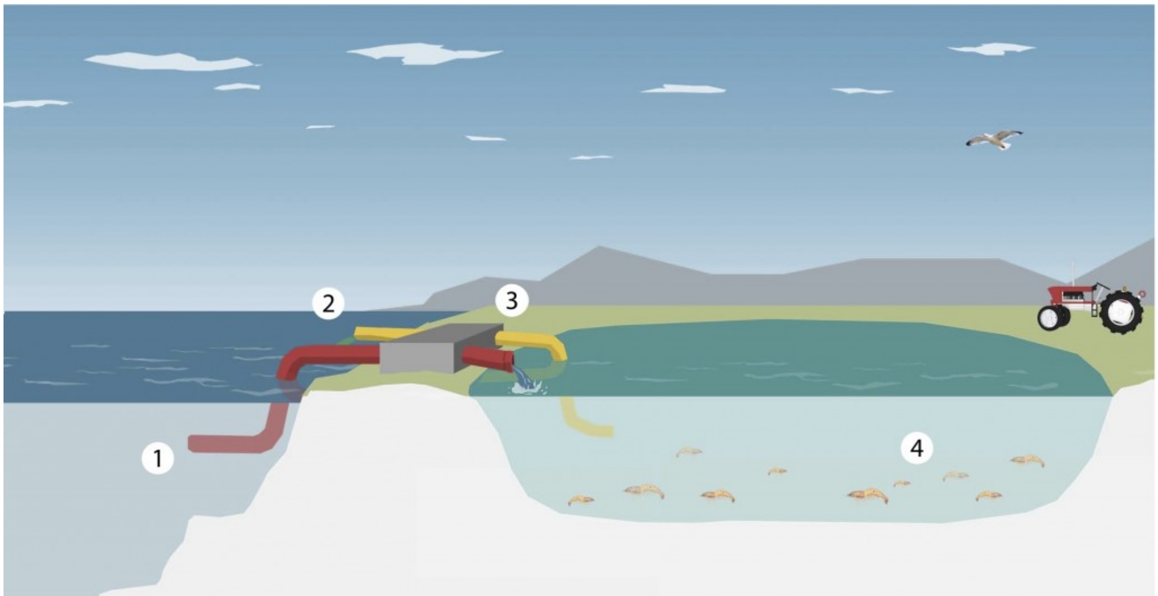
1. The cage is moored to the ocean floor 2. Fishemeal based feeds are added to the cages 3. Buoyant tubes keep the cages afloat 4. Fish faeces and waste fall through the cages

Figure 18.2. Open sea cage aquaculture

Semi-closed aquaculture or coastal ponds

Semi-closed aquaculture refers to the land-based production of a species, in which water is exchanged between the farm and a natural waterway (Figure 18.3). Waste water is released from the ponds into the local waterway, whilst the farm is replenished with fresh water pumped back into the system. Prawn farming is the predominant form of semi-closed aquaculture.

Semi-closed aquaculture operations can have large-scale effects on coastal ecosystems. As ponds require continual water exchange, they are often located adjacent to waterways, where coastal wetlands and mangroves are reclaimed for development. The result can be a vast loss of habitat which is critical for the juvenile stage of many species. Constant outflow of water may also reduce surrounding water quality if not treated adequately. Prawns are supplemented with fishmeal (pellets comprising small schooling fish species) at conversion ratios generally between 1-3kg of feed to 1kg of prawns, placing continued demand on wild fish stocks.



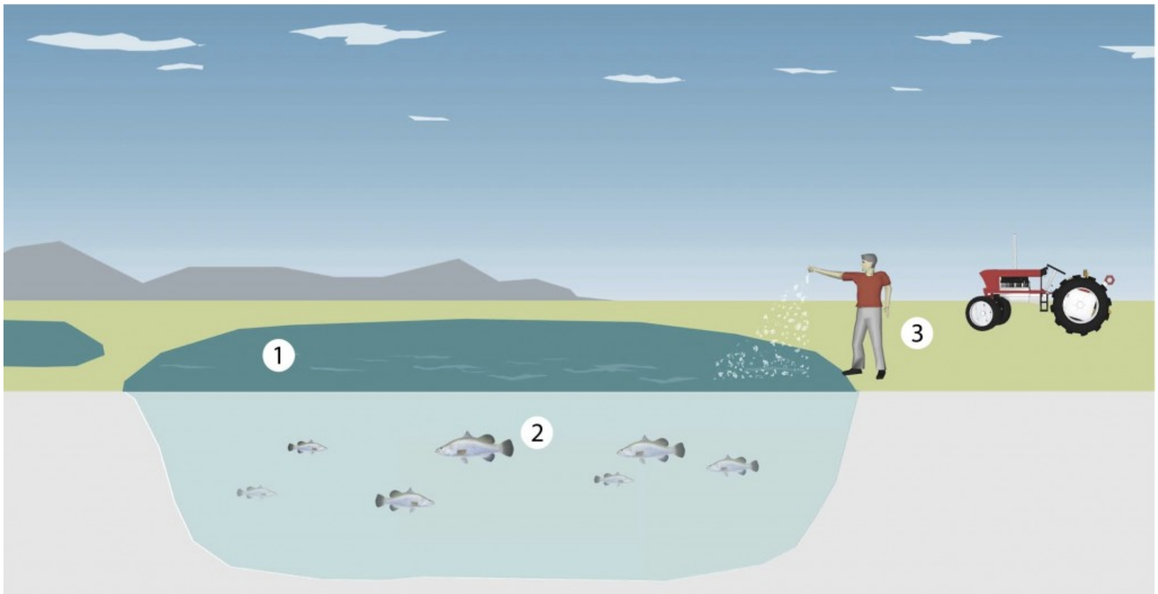
1. Inlet for ocean water 2. Outlet for waste water 3. Pump 4. Prawns are often cultivated using this method.

Figure 18.3. Semi-closed aquaculture

Inland pond or closed system aquaculture

Closed system aquaculture refers to the land-based rearing of aquatic species in tanks and ponds. Recirculation technology is implemented which cycles water through filtration processes and returns it back into the aquaculture system. This process aids in maintaining water quality while ensuring minimal exchange with natural waterways.

Closed aquaculture systems are primarily used for freshwater species but some marine fish and mollusks are also produced in closed systems. Closed system aquaculture is considered one of the more environmentally benign methods of rearing aquatic species. Fishmeal (pellets comprising small schooling fish species) may be added to feed carnivorous aquaculture species, and is a concern as it places continued demand on wild fish stocks. However, there is negligible interference with waterways as a result of tight control over waste water and the prevention of fish escape.



1. The pond or tank exists in a closed system 2. Species including barramundi are grown using this system 3. Feed is added

Figure 18.4. Closed system aquaculture.

18.3. Environmental costs

Like other forms of intensive food production, industrial-scale aquaculture generates significant environmental costs (summarized in Figure 18.5). Fish feed is often made from other fish species caught in the wild, often in unsustainable fisheries. In such cases, aquaculture can add to the problem of overfishing instead of alleviating it. This is particularly true of carnivorous species that require a large amount of food. The uneaten food, as well as fecal matter, can accumulate in high concentration near the aquaculture farm unless there is extensive water flow. High stocking densities lead to a high incidence of disease, requiring the use of antibiotics which can lead to the development of antibiotic-resistant strains of bacteria. Some of these diseases and parasites can spread to nearby wild populations and affect their survival and reproductive success. Cultured organisms often escape nets and pens, which can lead to problems of invasive species, or at least the dilution of the wild gene pool that is well-adapted to local conditions.

The expansion of some forms of aquaculture, particularly industrial-scale farming of salmon in netpens and shrimp in coastal ponds, has proven to be destructive to the natural environment and populations of aquatic animals. In British Columbia, a recent increase in the extent of coastal salmon aquaculture has been linked to increases in parasites such as sea lice (Figure 18.6) and diseases such as Infectious Salmon Anemia which have spread to wild populations. These parasites and disease are thought to be an important cause of the recent declines in wild salmon populations.

Unless significant changes are made, the increasing production of high-value carnivorous fish is unsustainable.

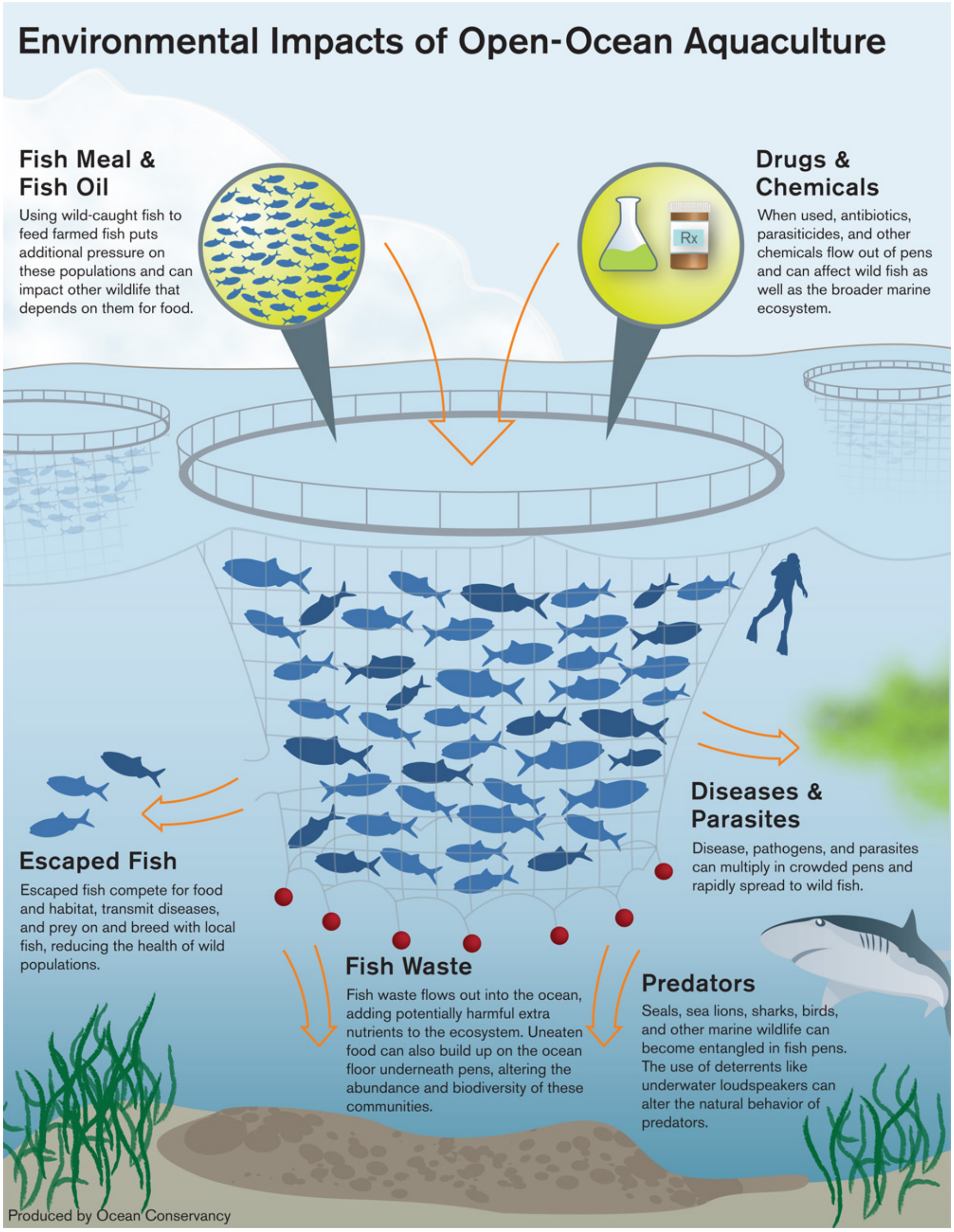


Figure 18.5. Problems associated with open-sea cage aquaculture.

Figure 18.6. Sea lice are increasingly abundant on wild salmon in British Columbia since the growth in salmon aquaculture in the region. They have been shown to negatively impact survival of wild stocks.

18.4. Moving towards sustainable aquaculture

There are a number of alternative ways forward in the development of aquaculture, which can offer more sustainable solutions. In some cases these methods have been around for centuries, but they have rarely been adopted in the modern aquaculture industry, and in other cases they are innovative practices that can be explored by aquaculture proponents. Alternatives include organic aquaculture, polyculture, open aquaculture systems, and closed and low discharge systems. These alternative practices have been successfully implemented in different areas of the world; however, they must be examined for their application on a wider scale. While each of the potentially sustainable practices mentioned in the following discussion does have some environmental impacts, they can be greatly minimized if the systems are managed well.

Organic aquaculture
Organic fish producers must comply with all of the same regulations that other organic certified producers do. Some substances or practices are prohibited from organic operations. For example, the addition of antibiotics to the fish feed is tightly regulated and the inclusion of genetically modified organisms is strictly forbidden in organic production. Rather than rely on the use of chemicals and drugs to improve the production of their fish, farmers instead optimize the living conditions, through lower stocking densities and cleaner, healthier water.

Polyculture
Polyculture or integrated aquaculture is a method of raising diverse organisms within the same farming system, where the wastes from one organism are used as inputs to another, resulting in the optimal use of resources and less pollution overall (Figure 18.7). Polyculture systems are not a new concept; on the contrary, they have been used for centuries. For over one thousand years fish farmers in China have produced four of the most widely cultivated fish species together in the same pond: silver carp (a phytoplankton filter feeder), grass carp (a herbivorous plant feeder), common carp (an omnivorous feeder), and bighead carp (a zooplankton filter feeder). Similar systems are currently being tested in the marine environment. It should also be noted that although polyculture systems based on netpens may prove beneficial for waste reduction, they fail to eliminate other problems associated with netpen aquaculture, specifically escape of fish, disease transfer, and discharge of chemicals.

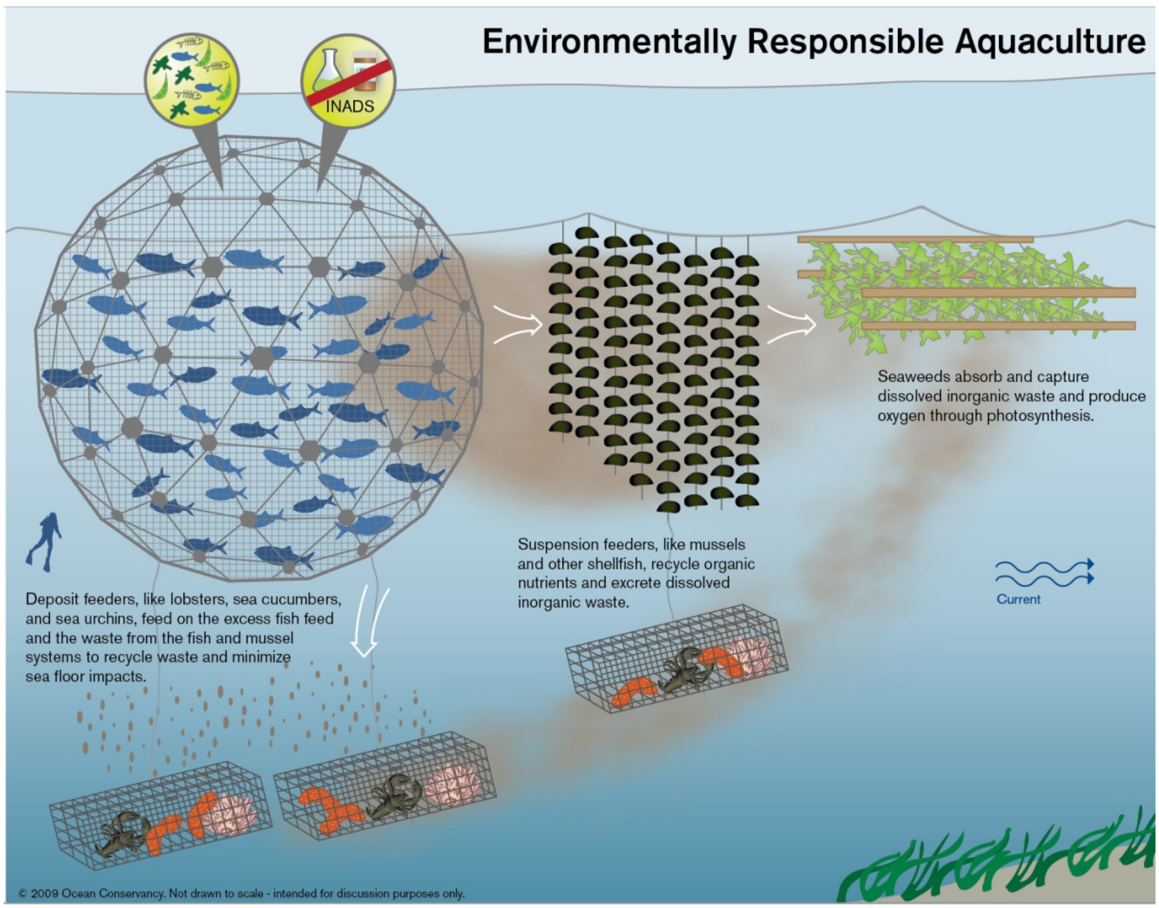
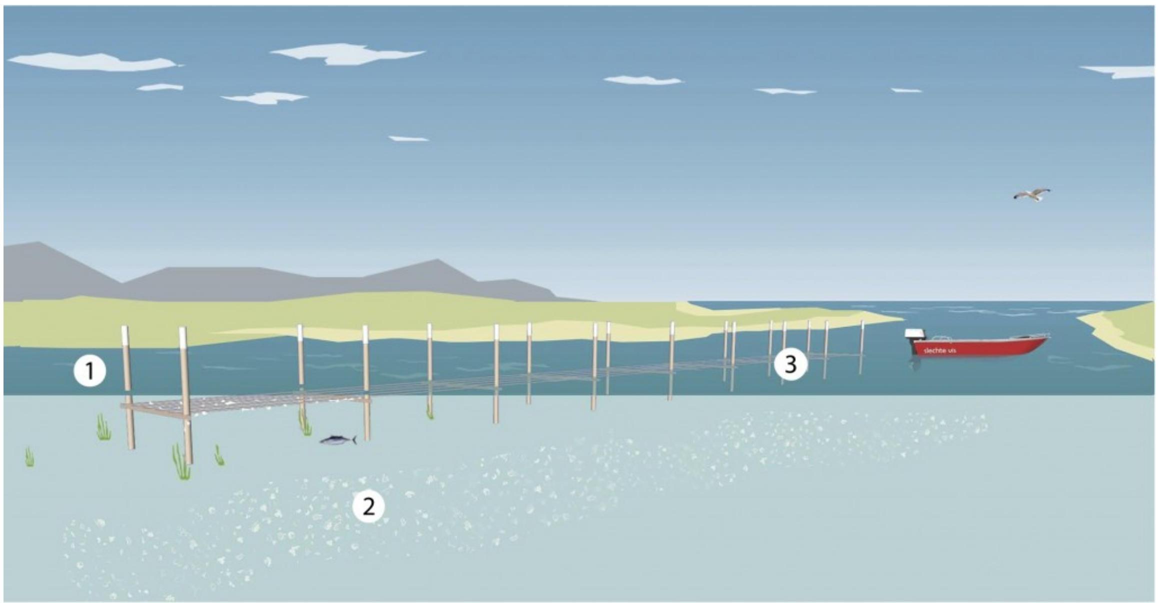


Figure 18.7. An aquaculture system designed to be sustainable, which includes principles of organic aquaculture (e.g. no antibiotics or fertilizers) and polyculture. From the Ocean Conservancy.

Open aquaculture systems
The culture of numerous bivalve species is carried out in systems open to natural waterways. Larval stages may be collected from the wild or produced in hatcheries. These are then placed into the water column by methods including attachment to sticks or

ropes, or containment in cages. The main species cultured with these methods are mussels and oysters. As these species are filter-feeders, they are capable of extracting nutritional requirements from the water column, with no fishmeal being added.

The farming of mussels, oysters and other filter feeders in this system is typically considered a sustainable method of aquaculture. With proper siting and planning, they have minimal impact upon coastal ecosystems and communities. In some cases the presence of bivalve farms can actually improve the water quality of the existing environment.



1. Mussels or oysters are grown on racks or in cages 2. Nutrients are taken from the water 3. Passive systems occur in estuaries as well as open ocean

Figure 18.8. An example of open aquaculture, often used for filter feeders like oysters or mussels.

Closed systems
Closed recirculating systems have a low impact upon the environment because of their closed nature – wastes and uneaten feed are not simply released into the ambient environment in the manner that they are with netpens and exotic species and diseases are not introduced into the environment. In recirculating systems, wastes are filtered out of the culture system and disposed of in a responsible manner. Recirculating systems can be built just about anywhere, including in urban settings where they can use existing structures and be placed close to markets, thereby reducing transportation costs. Recirculating systems can be used to grow a wide variety of fish species year-round in controlled environments. Species commonly grown in recirculating systems include hybrid striped bass and tilapia. Additionally, much research has been dedicated to developing recirculating systems for marine species of fish and this technology holds much promise.

Recirculating systems, however, can be costly to operate, as they are highly dependent on electricity or other power sources. Pumps must be used in order to maintain the constant flow of water and often water must be heated or cooled to the desired temperature. Backup systems must be in place in case of a power failure. A less expensive and more environmentally friendly option would be to take advantage of alternative energy and heating sources. Solar, wind and geothermal power are being considered as is heated water obtained from the waste products of manufacturing, electricity production, and composting. For example, tilapia farms can use the cooling water from power plants as a low cost warm water source. The warm water, which is necessary for growing tilapia, would otherwise be wasted.

18.5 Review questions

1. What is aquaculture?

2. How does modern aquaculture differ from that practiced several hundred years ago?

3. What parameters must be known to successfully raise a species in aquaculture?

4. What are the three main methods of intensive aquaculture?

5. Explain 4 different environmental issues associated with aquaculture

6. How can aquaculture be done more sustainably?

19. Climate Change: Assessment, Causes and Control (Trujillo, Chapter 16)

Global warming and climate change is a hot topic in the news these days. While climate change is often presented by mainstream media as a matter of vigorous scientific debate, in fact over 97% of climate scientists agree about the degree of climate change and the role of human activities in it. The current rate of temperature increase has not been seen for at least 1000 years. The rate of temperature and pH changes that are projected for the next 100 years are much greater than most populations' rate of evolution, therefore climate change is expected to lead to dramatic changes in the world's ecosystems. In this chapter, we explore the causes and mechanisms of climate change. The following chapter examines the impact on the oceans.

19.1. The Earth's climate system

The term climate refers to average weather conditions (temperature, rainfall, wind) that occur in an area over a long period of time. To understand the earth's climate, we need to study the atmosphere as well as the rocks, water, ice and organisms that interact with it (Figure 19.1). Changes in climate involve large-scale and often complex processes, including positive and negative feedback loops, which either reinforce (positive loops) or balance (negative loops) initial changes (Figure 19.2). For example, warming of the earth causes more water to evaporate, leading to a higher concentration of water vapor in the atmosphere. Water vapor acts as a greenhouse gas and exacerbates the warming which leads to even more evaporation and water vapor—a positive feedback loop. On the other hand, temperature increases lead to the formation of more clouds, and increased cloud cover can decrease the solar radiation absorbed by the earth. Such clouds therefore counteract or diminish the initial change in temperature, in a negative feedback loop. (Note, however, that the effect of clouds on climate is complex. High-level clouds trap heat and contribute to the greenhouse effect, therefore act as a positive feedback loop rather than the negative feedback loop described above. This small example shows how difficult it is to predict future climate change). Understanding climate change requires understanding of these and many other feedback loops, and to understand which have the greater effect on climate. For example, research shows that the negative feedback loop described in figure 19.2 is stronger than the positive one shown in the same figure. However, there are many other positive feedback loops such as melting of polar ice which increases absorption of solar radiation near the poles because of a lesser amount of highly reflective white ice, leading to even more melting of ice. Overall, positive feedback loops have a greater impact than negative ones, with the result that the earth is continuing to warm.

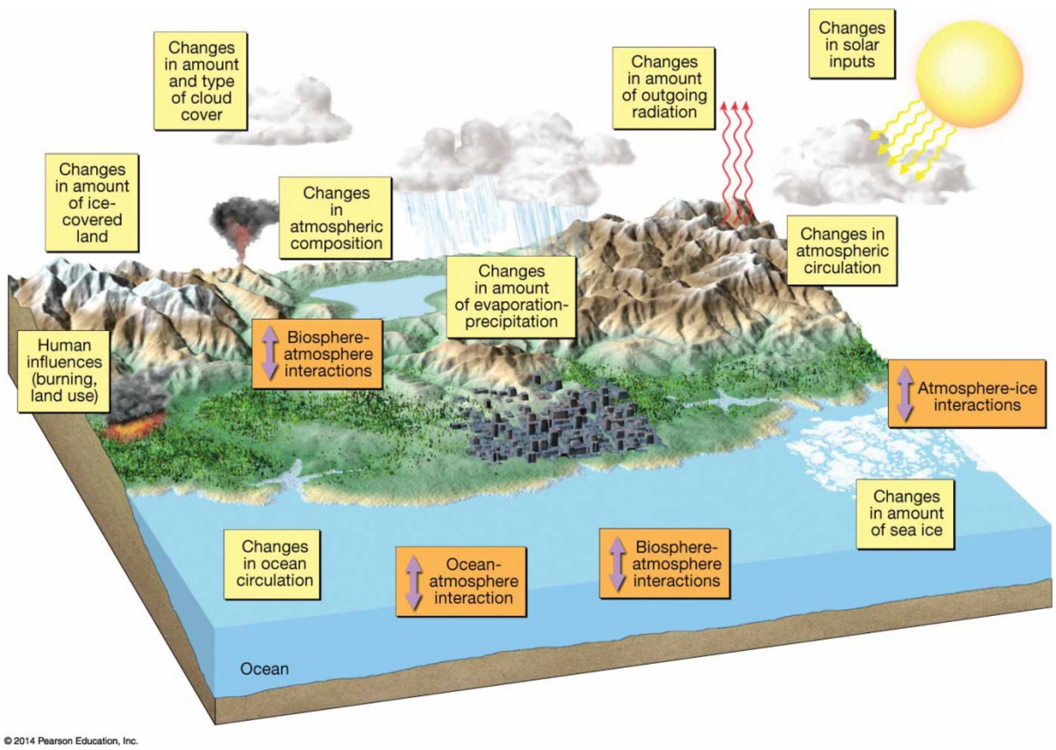


Figure 19.1. Major components of the earth's climate system.

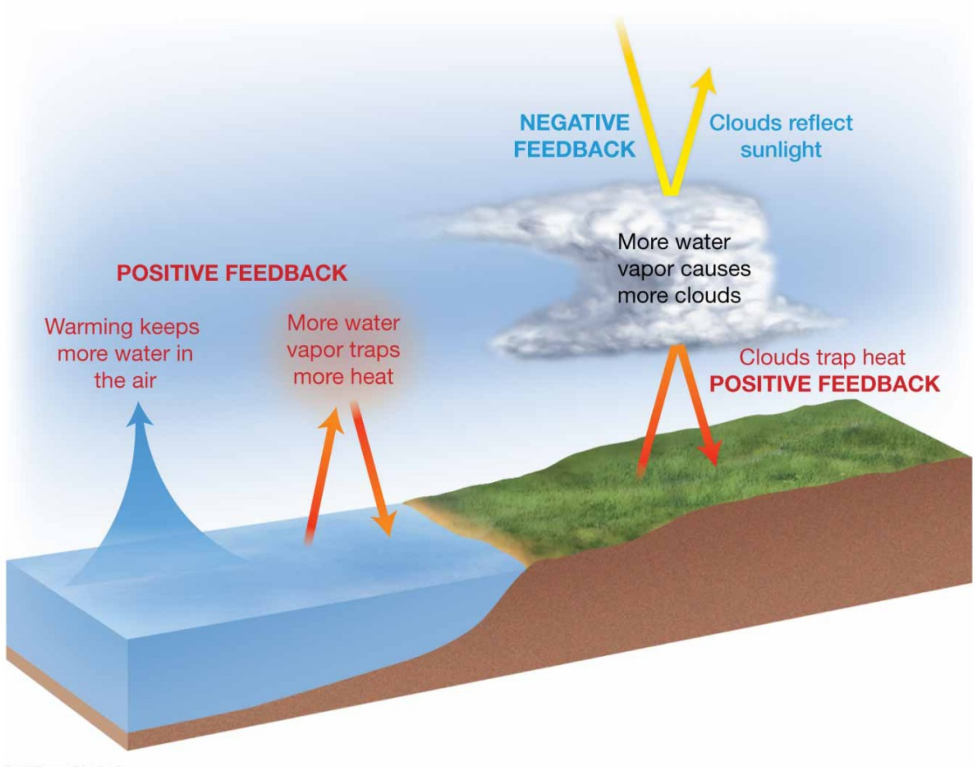


Figure 19.2. Examples of positive (left) and negative (right) climate feedback loops.

19.2. The Earth's recent climate change: natural or anthropogenic?

Proxy climate data and paleoclimatology
Scientists have long known that the earth's climate has fluctuated naturally over the history of the Earth. The important question is to determine how current rates of change compare to past changes, and how much of current climate change is due to human activities. Since records of temperature and atmospheric composition taken directly with instruments go back only about two centuries (at the most), scientists must reconstruct and estimate past climate conditions based on proxy data. Tree growth rings, seafloor sediment, coral skeleton composition, bubbles trapped in ice can all help reconstruct past climate. Paleoclimatology is the study of past climates using such proxy data.

Natural causes of climate change
The earth's climate has been dynamic over geological time, and many natural factors have been linked to past climate change including variations in solar energy, volcanic eruptions, variations in the earth's orbit and tectonic movements.

The energy output of the sun varies over time, in large part based on patterns of magnetic storms and the ejection of solar particles. Solar activity has been recorded to increase sharply every 11 years or so in the past ~130 years, each time followed by a decrease in irradiance (Figure 19.3). Overall the changes in solar irradiance do not explain the recorded increase in temperature (Figure 19.3).

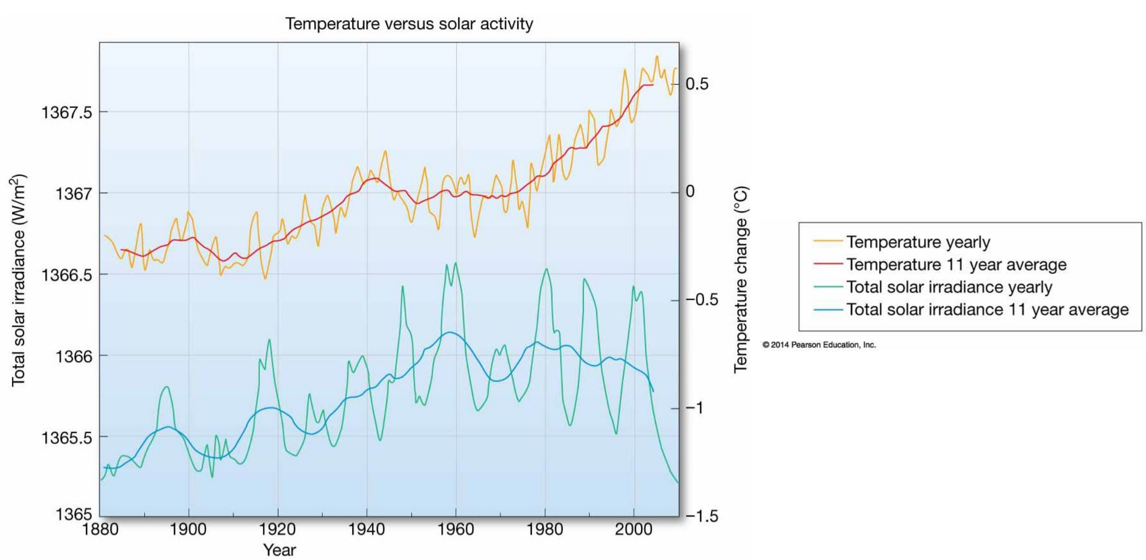


Figure 19.3. Earth temperature and solar irradiance from 1830 to 2010.

Several changes in the earth's pattern of orbit around the sun can also contribute to climate change. These include changes the shape of the earth's orbit, changes in its axis in relation to the plane of the orbit, and a wobbling of the earth's axis over time (Figure 19.4). These changes have been implicated in the long-term climate fluctuations that lead to glacial and interglacial periods, including the most recent ice age. However, changes due to the earth's orbit occur over very long cycles (~100,000 years, 41,000 years and 26,000 years, respectively) and are unlikely to explain the rapid changes in climate measure in the past ~150 years.

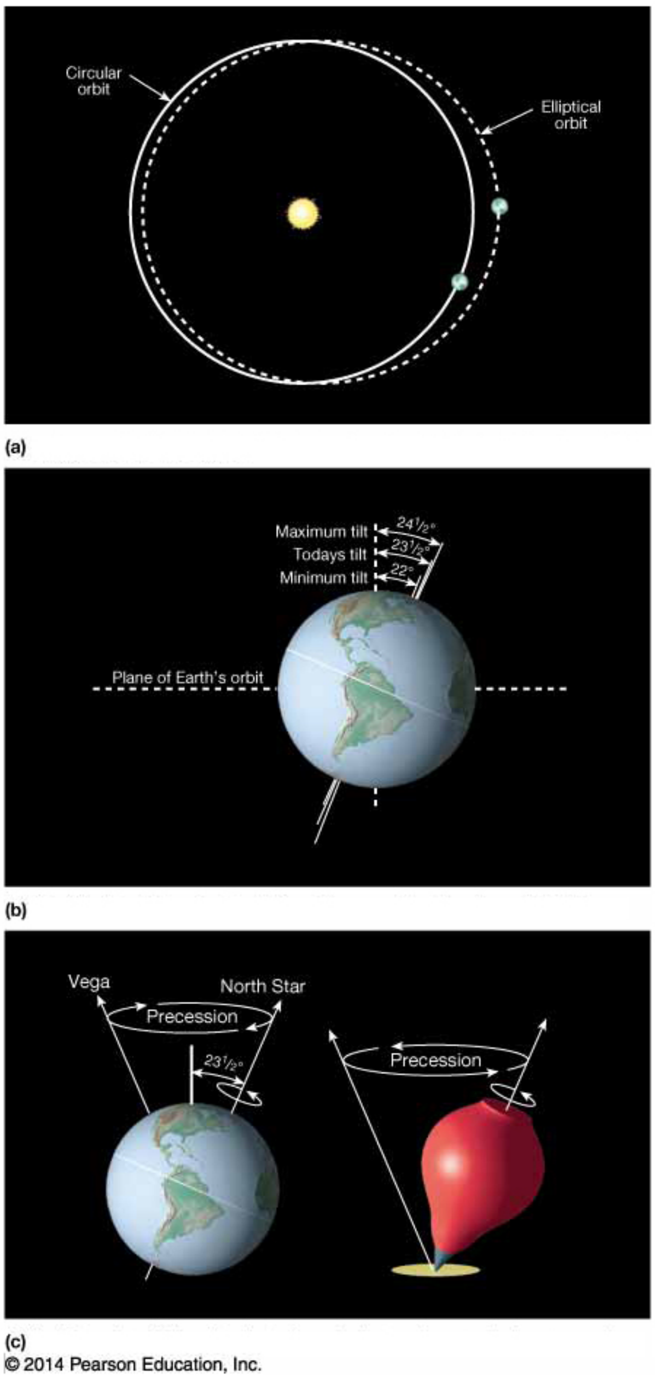


Figure 19.4. Three types of long-term variations in the earth's orbit.

Movement of tectonic plates, which are important over geological time, can reshape ocean basins and thereby change major ocean currents. This in turn can impact the atmosphere and global climate. However, as with variations in the Earth's orbit, tectonic changes happen very slowly and cannot account for the current rate of change.

Volcanic eruptions can emit a large quantity of greenhouse gases in the atmosphere, in particular carbon dioxide. However the emissions from volcanoes are very small compared to human emissions (Figure 19.5).

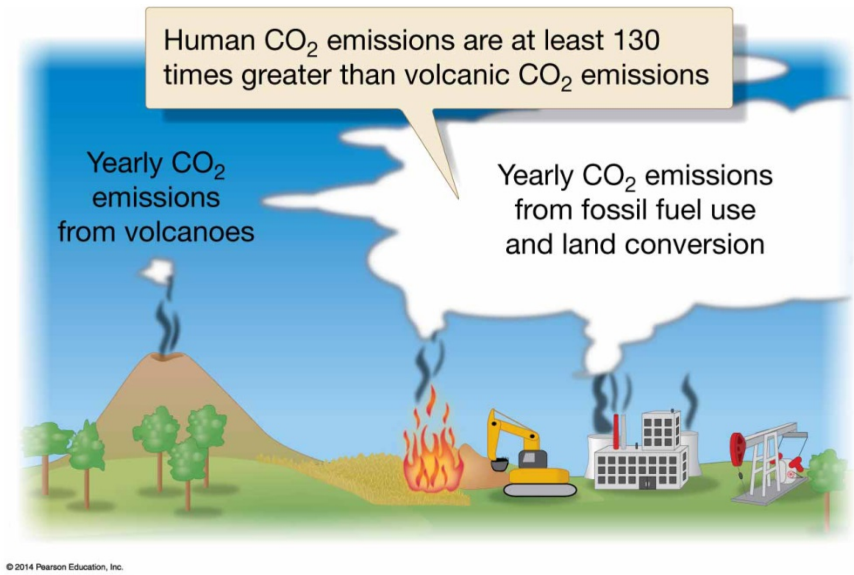


Figure 19.5. Carbon dioxide emissions from volcanoes compared to those from human activities.

While the Earth's climate has clearly changed over time due to natural factors, none of these natural occurrences can account for the current rate of change in temperature and atmospheric composition. There is widespread consensus among climate scientists that the climate change recorded in the past few decades is most directly linked to human activities, in particular carbon dioxide emissions. Scientific organizations such as the US National Academy of Science, the American Meteorological Society, the American Geophysical Union, the American Association for the Advancement of Science all publicly support the scientific conclusion that current climate change is caused by humans. Further, the Intergovermental Panel of Climate Change (IPCC), sponsored by the United Nations and the World Meteorological Organization, has issued a series of reports since 1990 that link climate change to human activity with increasing certainty. Drafts of latest report, released in 2013, states "There is consistent evidence from observations of a net energy uptake of the earth system due to an imbalance in the energy budget. **It is virtually certain that this is caused by human activities, primarily by the increase in CO_2 concentrations**. There is very high confidence that natural forcing contributes only a small fraction to this imbalance." This consensus has been broadly accepted in the scientific community for about 20 years, yet the US public is still largely reluctant to accept the idea (Figure 19.6).

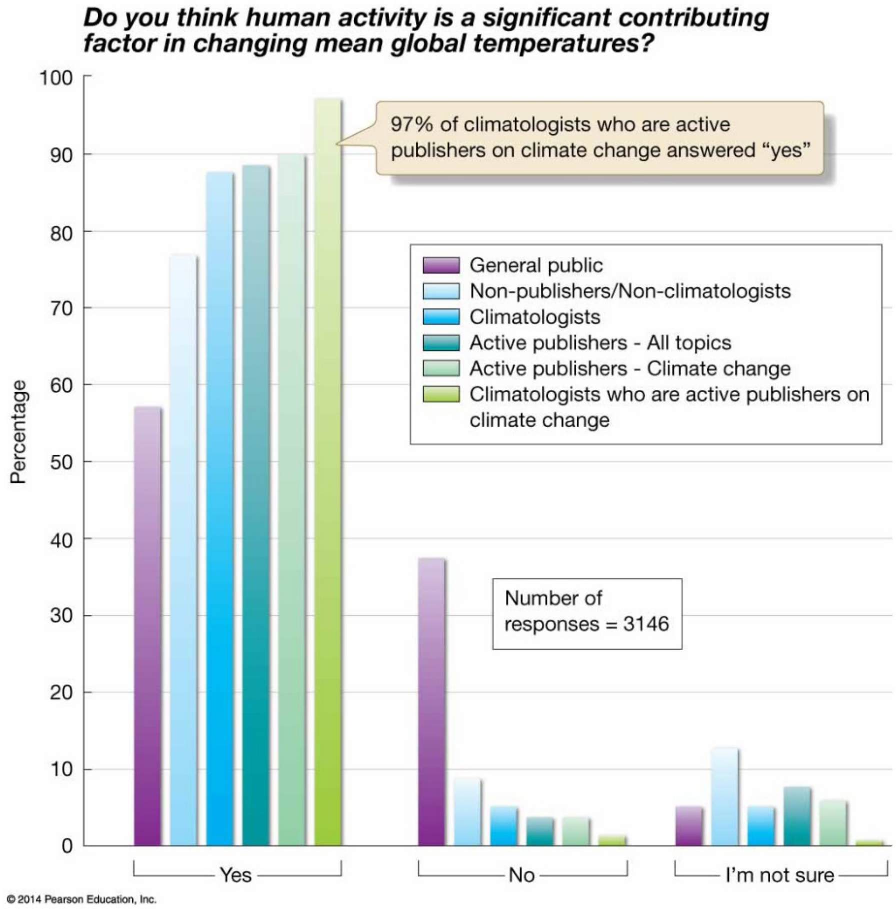


Figure 19.6. Consensus on humans being the main cause of climate change, in various groups of scientists and in the general public, in the USA.

19.3. Causes of the atmosphere's greenhouse effect

It is increasingly clear that human emissions of certain gases are responsible for global warming. These gases contribute to the greenhouse effect, which traps more sun energy in the atmosphere much like a greenhouse does (Figure 19.7). While the greenhouse effect occurs naturally (in fact the Earth would not be inhabitable to humans without it), human emissions enhance the effect, causing a warming of the Earth.

The Earth's heat budget

For the Earth's climate to be stable, it must have a balanced heat budget—meaning that the energy coming in roughly equals the energy coming out (Figure 19.8). Of the incoming solar energy, about 30% is reflected back in space by atmospheric backscatter, clouds or the Earth's surface; about 23% is absorbed in the atmosphere, and ~47% is absorbed at the surface of the Earth (land or oceans). Much of the sun's radiation occurs within the visible light spectrum, which cannot be absorbed by the atmosphere but is absorbed by rocks and water. These materials then re-radiate the energy back toward space, but at a different wavelength within the infrared spectrum (Figure 19.9). Radiation within this range of wavelengths is readily absorbed by greenhouse gases, and therefore heat up the atmosphere.

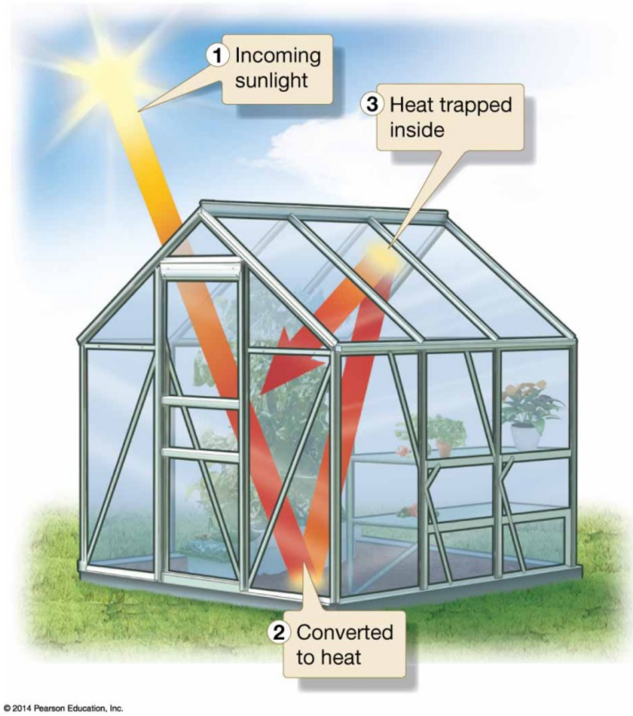


Figure 19.7. Greenhouses trap solar energy as heat.

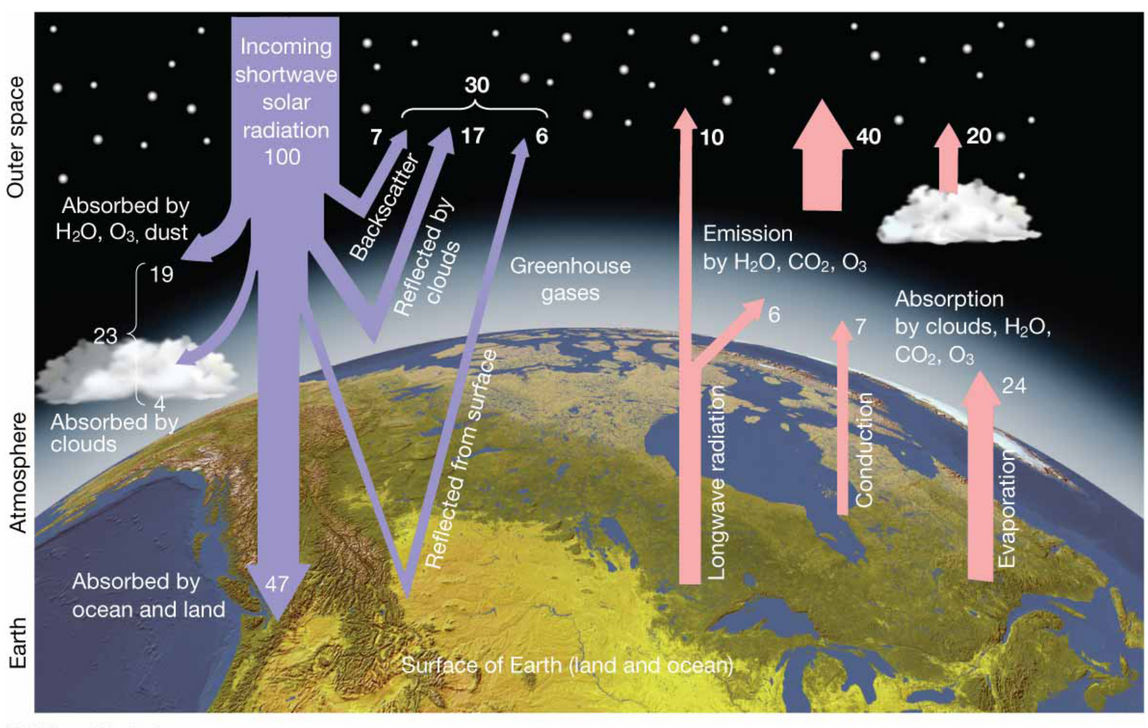


Figure 19.8. The Earth's energy budget.

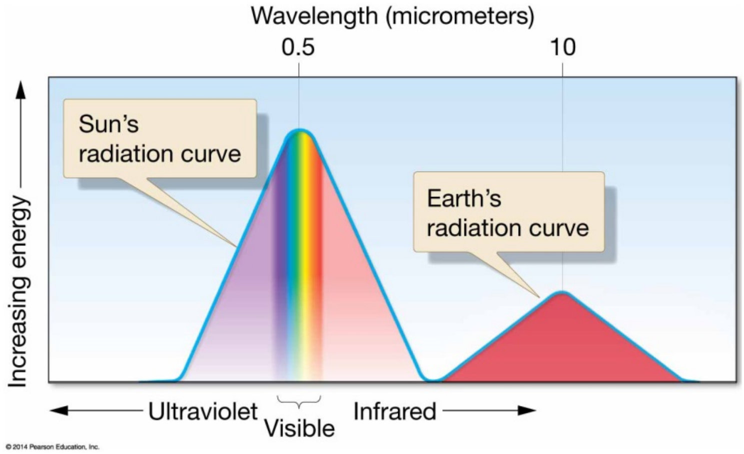


Figure 19.9. Energy radiated by the Sun and the Earth, in relation to wavelength along the electromagnetic spectrum.

Greenhouse gases

The greenhouse effect is caused by several atmospheric gases. Water vapor is single the most important gas contributing to the greenhouse effect. It enters the atmosphere through natural processes such as evaporation, though there is some evidence that human-induced warming has also contributed to an increase in water vapor in the atmosphere. Several other gases are clearly increasing in the atmosphere as a result of human activities (Table 16.1). Chief among them is carbon dioxide, which has increased dramatically in the atmosphere since industrialization, with an accelerating rate of increase in recent decades (Figure 19.10). Carbon dioxide enters the atmosphere through the burning of fossil fuels. Methane is the second most important anthropogenic greenhouse gas. It is produced by decomposing trash, cattle raising, and some agriculture. Additionally, some hydroelectric project, can contribute significant inputs of methane through the decaying material that accumulates in areas that are intermittently flooded and dry.

TABLE 16.1 HUMAN-CAUSED GREENHOUSE GASES AND THEIR CONTRIBUTION TO INCREASING THE GREENHOUSE EFFECT

Atmospheric gas	Human-caused sources of gas	Pre-industrial (circa 1750) concentration (ppbv[a])	Present concentration (ppbv[a])	Current rate of increase or decrease (% per year)	Relative contribution to increasing the greenhouse effect (%)	Infrared radiation absorption per molecule (number of times greater than CO_2)
Carbon dioxide (CO_2)	Combustion of fossil fuels	280,000	395,000	+0.5	60	1
Methane (CH_4)	Leakage, domestic cattle, rice agriculture	700	1775	+1.0	15	25
Nitrous oxide (N_2O)	Combustion of fossil fuels, industrial processes	270	315	+0.2	5	200
Tropospheric ozone (O_3)	Byproduct of combustion	0	10–80	+0.5	8	2000
Chlorofluorocarbon (CFC-11)	Refrigerants, industrial uses	0	0.26	−1.0	4	12,000
Chlorofluorocarbon (CFC-12)	Refrigerants, industrial uses	0	0.54	0.0	8	15,000
Total					100	

[a]ppbv = parts per billion by volume (not by weight).

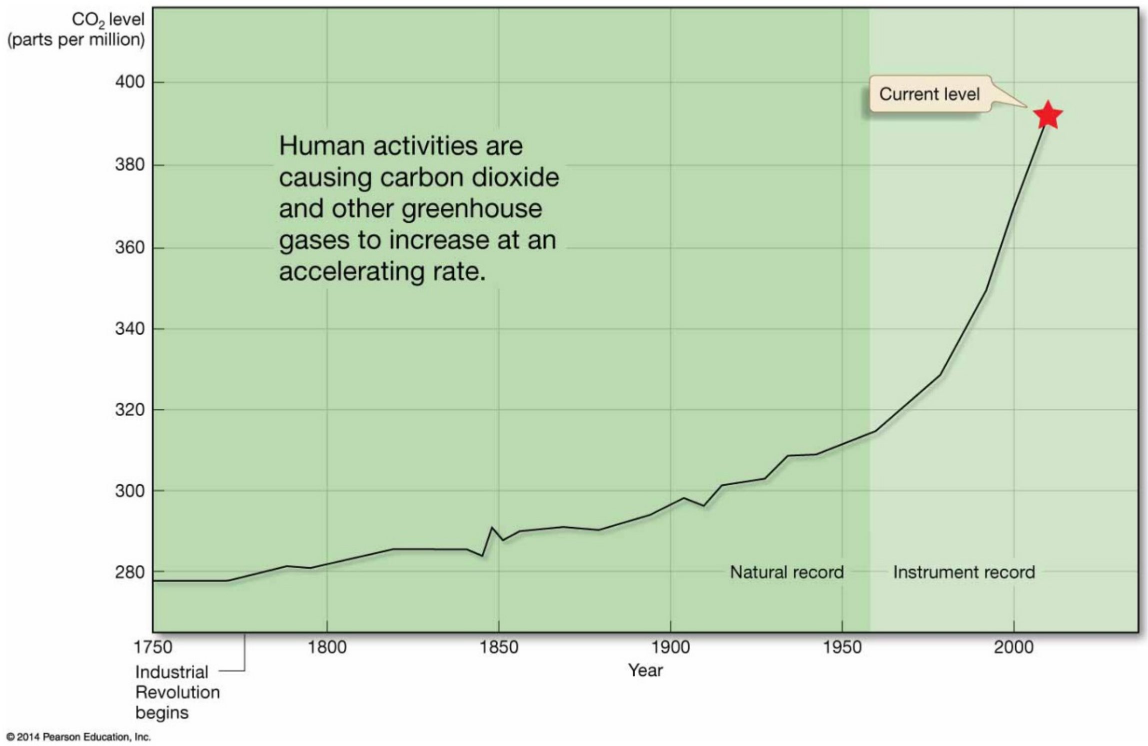


Figure 19.10. Atmospheric carbon dioxide concentration since the industrial revolution.

Other anthropogenic greenhouse gases such as nitrous oxide, ozone, and CFCs are present in much lower concentrations than carbon dioxide and methane but they are still important to consider because each molecule of those gases has a much stronger ability to absorb heat than carbon dioxide or methane.

Changes in atmospheric composition (and levels of greenhouse gases) can be reconstructed by analyzing small bubbles that get trapped in polar ice. Once the bubbles are isolated from the atmosphere, their composition remains unchanged. Using ice cores, scientists have been able to analyze changes in carbon dioxide and methane dating back ~800,000 years. The results show that while there are cyclical variations in concentration of both gases, current concentrations are higher than they have been for that entire time period (Figure 19.11). As greenhouse gases have increased in the past ~200 years, so has temperature. Current temperature increases have not been seen for at least a 1000 years (Figure 19.12). IPCC climate scientists have made several projections for continued increase in the Earth's temperature for the next 100 years, based on various scenarios of carbon dioxide emissions (Figure 19.13). Note that even with no further increase in carbon dioxide emissions, temperatures are expected to increase by about 1^{o}C over that time period.

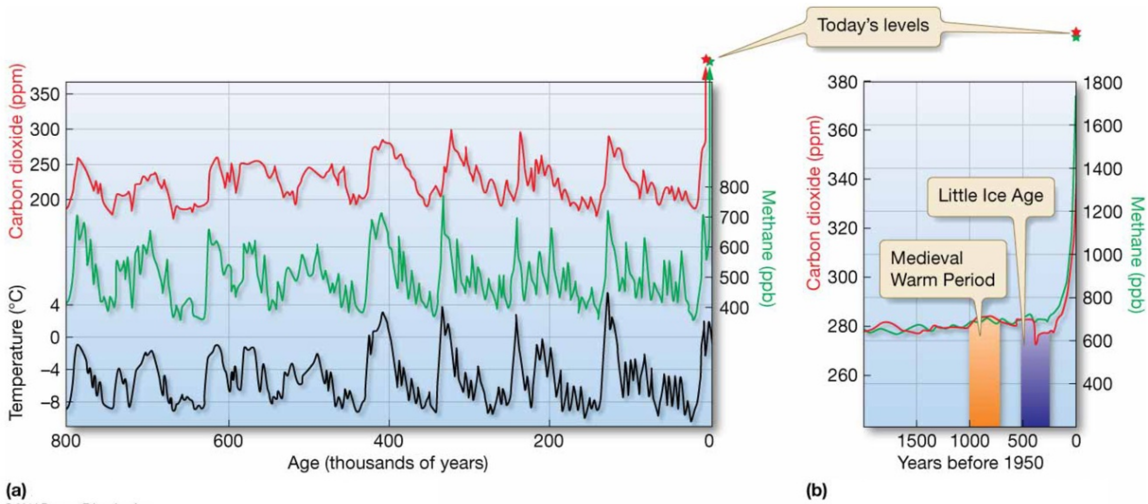


Figure 19.11. Atmospheric composition for the last 800,000 based on ice core data.

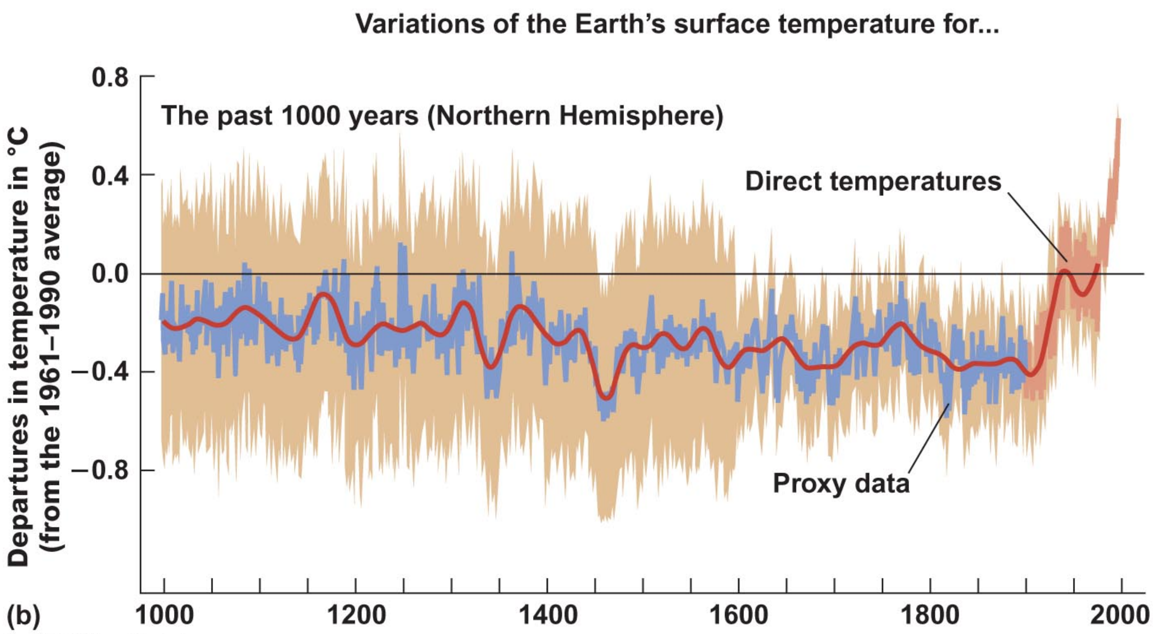


Figure 19.12. Variations in the Earth's temperature over the past 1000 years.

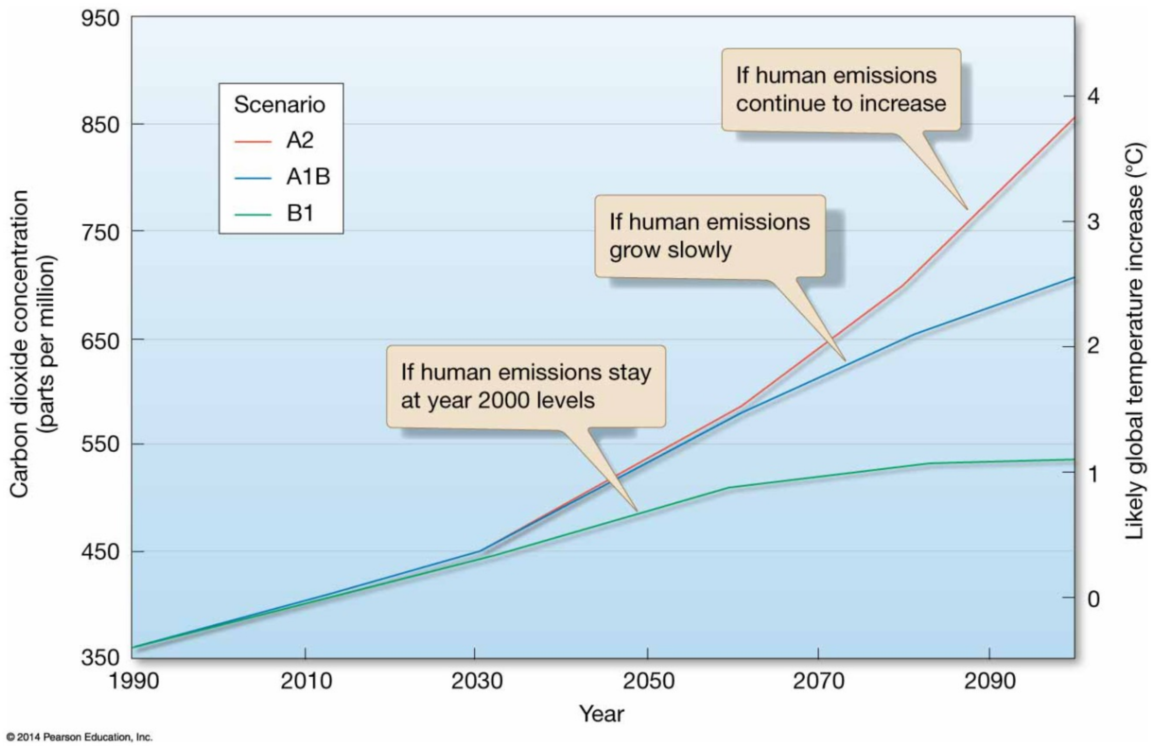


Figure 19.13. Three scenarios of carbon dioxide and temperature increases, from the IPCC.

Documented changes due to global warming
Increased temperatures lead to melting of glaciers and polar ice caps, shifts in species distributions toward the poles, changes in timing of seasonal events, and direct impacts related to increased temperatures such as coral bleaching. All of these changes have been documented already. They will be discussed in more detail in chapter 20.

19.4. Limiting and reducing greenhouse gases

The ocean's role in reducing atmospheric carbon dioxide
Gases are exchanged at the interface between atmosphere and oceans and for that reason, excess carbon dioxide in the atmosphere is in part absorbed by oceans. There, carbon can be incorporated into the biomass of living organisms through photosynthesis as well as into carbonate shells of certain organisms (e.g. coccolithophores, mollusks, echinoderms). Much of this carbon-based material eventually sinks below the photic zone and ends up buried on the ocean floor as fossil fuels and biogenous calcium carbonate. This is called the ocean's biological pump (Figure 16.31). However, it should be noted that the biological pump does not remove carbon dioxide indefinitely from the atmosphere. The recent increase in carbon dioxide is greater than the ability of the biological pump to move it to deep waters, and oceans are now becoming more acidic as more carbon dioxide dissolves in water (section 20.4). Increased temperature and acidity in turn reduce the efficiency of the biological pump, an example of a positive feedback loop.

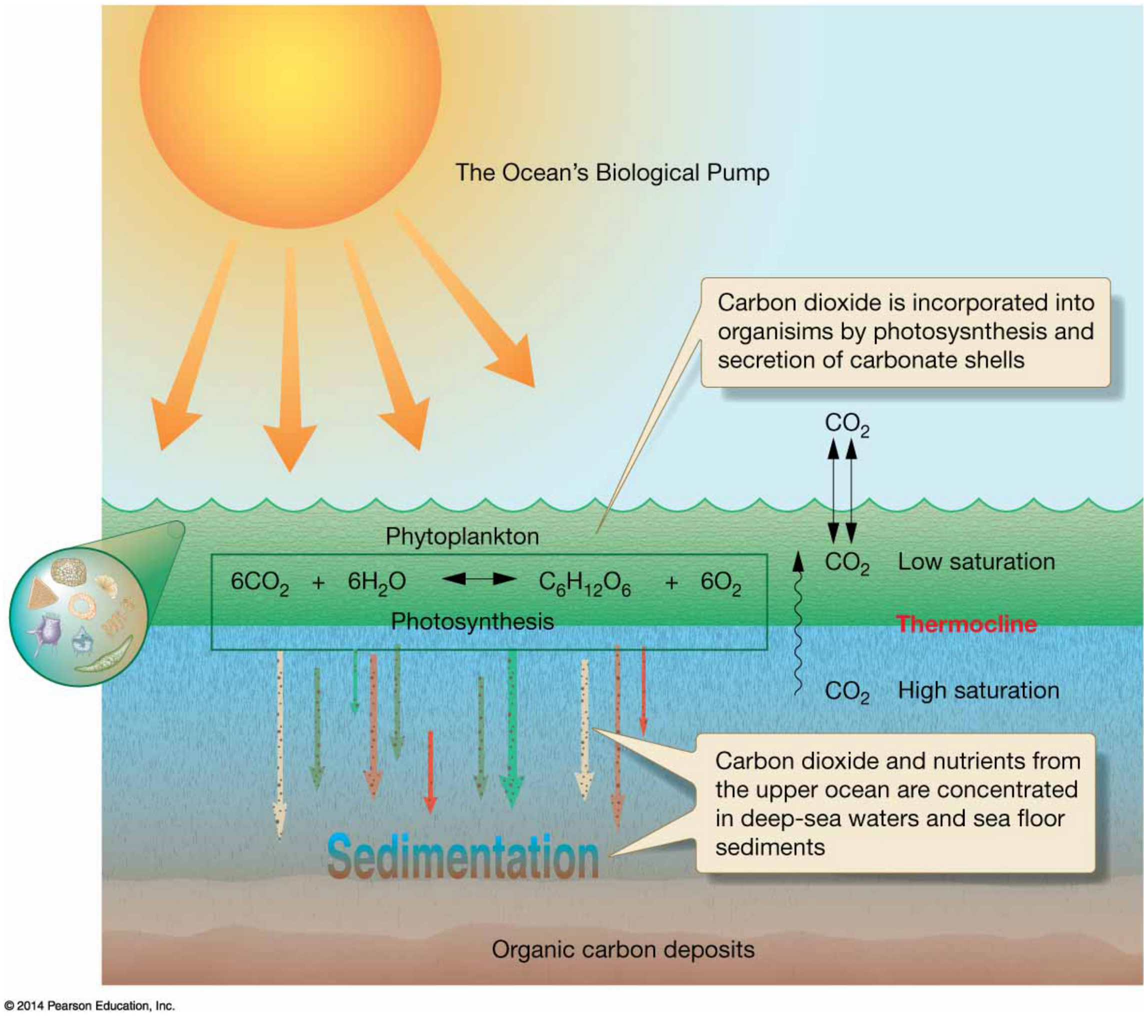


Figure 19.14. The ocean's biological pump.

The ocean's role in slowing down the Earth's temperature increase
Because water can absorb large amounts of heat with small changes in temperature, the oceans buffer global increases in temperature. Without oceans, warming would be much more rapid.

Possibilities for reducing greenhouse gases
There have been a number of global geo-engineering proposals that aim to reduce the level of carbon dioxide currently in the atmosphere, from installing shades in orbit to reduce solar radiation reaching the earth to spraying sulfate aerosols in the atmosphere to induce cooling. We'll examine two particular ideas that have been around for some time and have had some field testing: iron fertilization and oceanic carbon sequestration.

The idea of fertilization the oceans with iron came in the mid1980s, when oceanographer John Martin found that many parts of the ocean that had abundant sunlight and major nutrients (nitrates, phosphates, silicates) had low levels of productivity because they had extremely low concentration of iron. Martin suggested that productivity could be increased by fertilizing those areas with iron, which would boost primary productivity and help reduce atmospheric carbon dioxide concentration and global warming (Figure 19.15). This idea became widely known as the iron hypothesis. The iron hypothesis was first tested in 1993 near the Galápagos Islands and similar experiments have been done in many others parts of the oceans. These experiments consistently show that iron does increase primary production and decrease carbon dioxide levels. However, all open-ocean experiments have been done at a relatively small scale and it is still unclear what full-ecosystem consequences should be expected when adding large amounts of iron to the oceans. There are two main fears related to this idea. First, that large phytoplankton blooms will lead to oxygen depletion when they die (e.g. see the discussion on eutrophication and dead zones in section 16.2). Second, that large-scale iron input could lead to changes in plankton ecology, including changes in plankton species composition, and other effects further up the food web.

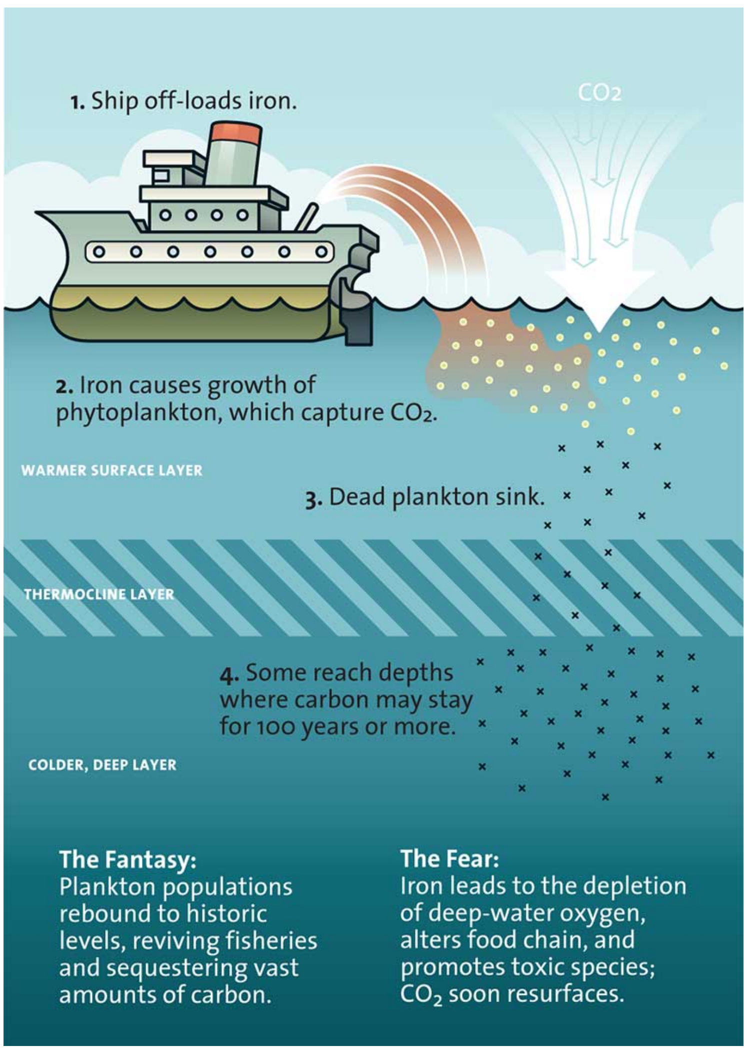


Figure 19.15. A diagram showing the idea of ocean iron fertilization, with the potential pros (“the fantasy”) and cons (“the fear”). From *The New Yorker* magazine.

There have been many experiments that have taken carbon dioxide from the atmosphere and pumped it in the deep ocean or underground reservoirs. This is called carbon sequestration. In the short-term, carbon sequestration would undoubtedly lead to a reduction in atmospheric carbon dioxide. However it is still unclear if carbon dioxide would remain locked in the ocean for long periods of time, since oceanic current patterns bring deep water to the surface in many regions. Additionally, there are concerns about the impacts of added carbon dioxide to the ocean's chemistry.

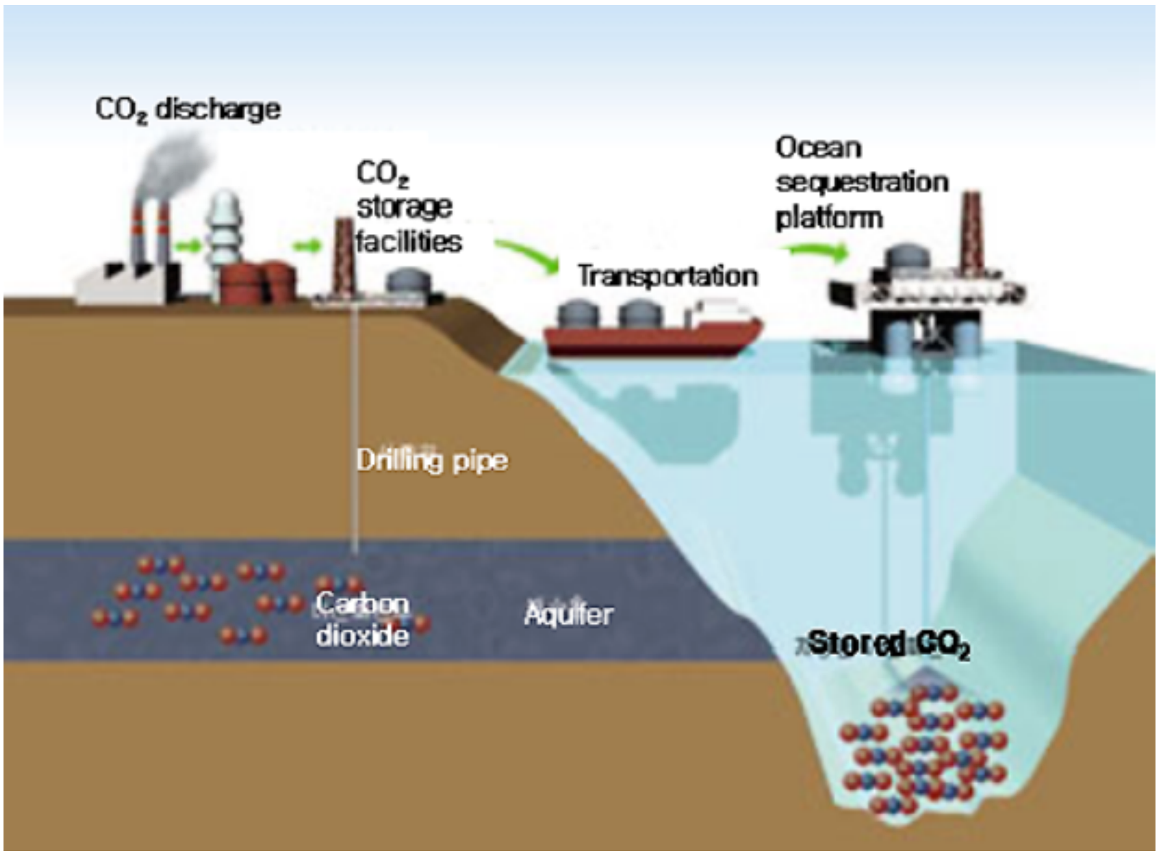


Figure 19.16. Carbon sequestration in underground reservoirs and in the deep sea.

The Kyoto Protocol for limiting greenhouse emissions

Given that there is no perfect way in which to reduce carbon dioxide concentration in the atmosphere, most efforts should be aimed at reducing new emissions. There have been a number of international meetings to attempt to come to agreements on targets for reductions of greenhouse gas emissions. The Kyoto protocol was developed in 1997 and set emission reduction targets for each country. However, most countries lag behind their targets; the USA and China (the two largest contributors to carbon emissions) never ratified the protocol, and Canada withdrew from it in 2011. In 2012, the United Nations Climate Change Conference in Doha, Qatar, led to an agreement to extend the Kyoto Protocol until 2020 (it was set to expire in 2012) and to develop a successor to the protocol by 2015, to be implemented by 2020. Several countries made specific statements at this conference. One of the most poignant ones came from Philippine envoy Naderev Sano who called for urgent action to halt climate change, emphasizing the Philippines' recent experience with a climate change-enhanced deadly typhoons and noting that countries may face more extreme weather disturbances frequently if climate change is left unchecked. In his words: "As we vacillate and procrastinate here, we are suffering. There is massive and widespread devastation back home. Heartbreaking tragedies like this are not unique to the Philippines."

19.5. Review Questions

1. What is the difference between climate and weather?
2. What is a positive feedback loop? What is an example of one in the context of climate?
3. How can scientists reconstruct changes in temperatures and atmospheric composition from thousands of years ago?
4. Explain 3 natural causes of climate change
5. Can these natural causes explain current climate change patterns? Why or why not?
6. What is the mechanism of the greenhouse effect?
7. What gases contribute to the greenhouse effect?
8. Which anthropogenic greenhouse gas contributes the most to global warming?
9. How can the biological pump of the oceans reduce atmospheric carbon dioxide?
10. What is the iron hypothesis?
11. What are 2 potential concerns with ocean carbon sequestration?
12. What does the Kyoto protocol dictate?

20. Climate Change: Consequences to the Ocean Environment (Trujillo, Chapter 16)

Global climate change is having a strong and increasing impact on ocean ecosystems. Some of these impacts are related to warming of seawater and associated processes such melting ice, rising sea level, increased storms and changes in ocean circulation. Another group of impacts is related to changes in ocean chemistry, in particular a decrease in ocean pH. Increased atmospheric carbon dioxide concentration is involved in both cases.

20.1. Increasing ocean temperature

Ocean temperatures have increased globally by about 0.6°C since 1970. However, the warming is far from uniform with the greatest temperature rise in the Arctic Ocean (Figure 20.1). The Southern ocean and some tropical regions have also seen higher than average temperature increases. Rising temperatures can have important impacts on many physical processes in the ocean (e.g. hurricane activity, melting of ice, changes in circulation patterns, see next few sections), and can also directly affect marine organisms. One of the most directly effect of warming temperatures is a poleward shift in the range of organisms. This has been documented for many marine species (Figure 20.2). Additionally, some species are found in increasingly deep waters; in the North Sea, the range of several commercial fish species was found to deepen by a rate of about 5m/decade.

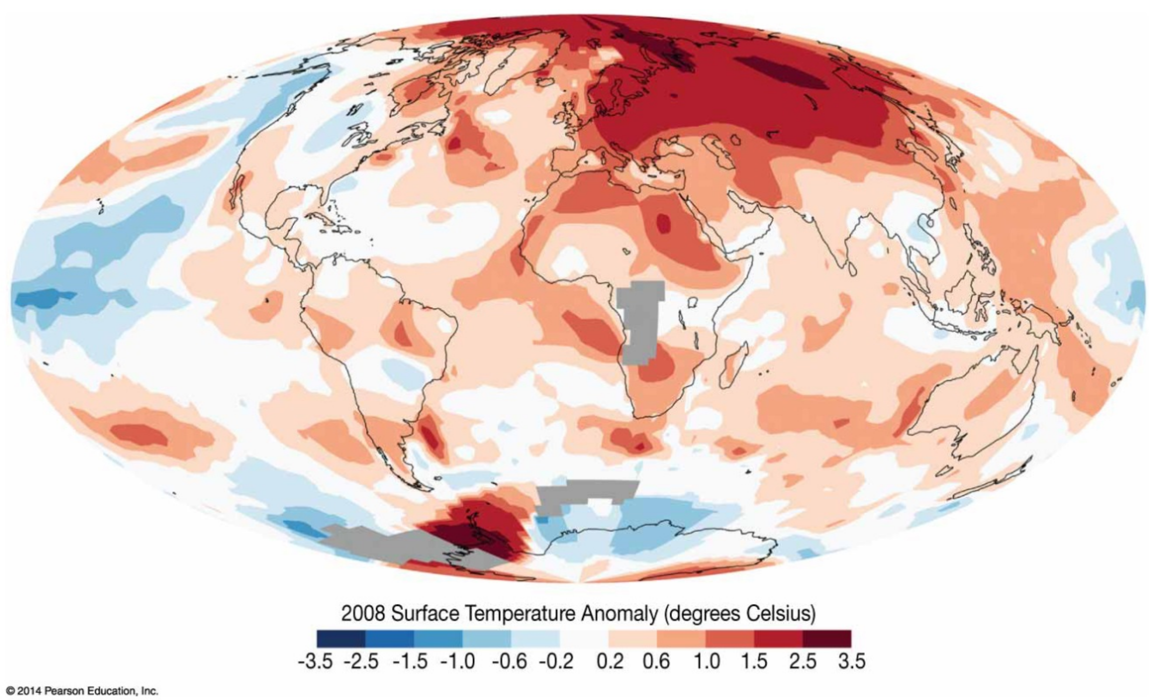


Figure 20.1. Sea surface temperature anomalies for 2008 when compared to 1950-1980 average temperatures.

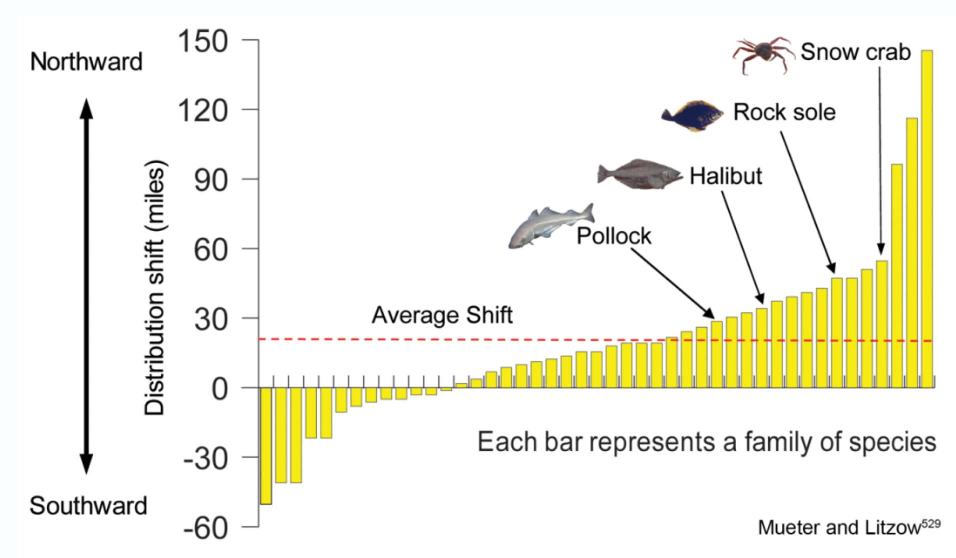


Figure 20.2. Many marine species are shifting their range poleward as water temperatures increase. On average, Alaskan marine species have migrated 19 miles to the north from 1982 to 2006.

Coral reefs are one of the marine ecosystems most directly affected by warming temperatures. Several species of coral are now found at higher latitudes than previously, another example of range shift. But while coral species might migrate poleward through larval dispersal, an individual colony is sessile and does not move during it lifetime. Therefore corals cannot simply move when temperature rise above their threshold of tolerance. When temperatures are too high, corals loose the symbiotic algae that live in their tissues in a process called coral bleaching (Figure 20.3). Corals can recover from short-term bleaching events, but die if the warm episode is sustained over a long enough period. Since 1998, several bleaching episodes worldwide have contributed to widespread coral mortality (Figure 20.4). Coral bleaching is especially frequent in El Niño years.

Figure 20.3. Coral bleaching is triggered by stresses such as increased temperature. Corals lose their symbiotic algae and the color they bring, and turn bright white as the calcium carbonate skeleton is visible through translucent tissue.

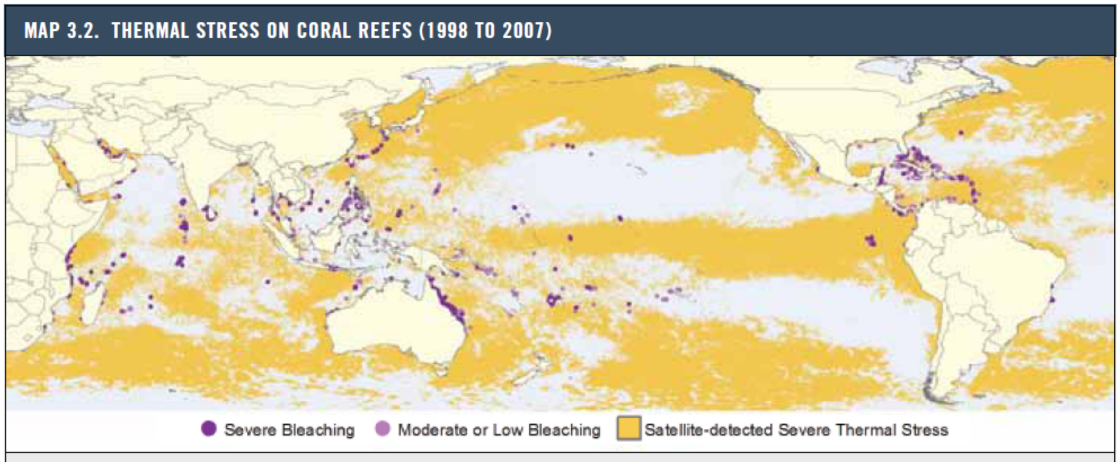


Figure 20.4. Zones of coral bleaching in relation to thermal stress, 1998-2007.

20.2. Increasing hurricane activity

Because warm water fuels tropical storms, it is reasonable to expect that global warming should lead to an increase in the number and intensity of tropical storms. Thus far, there is limited evidence of an increase in the number of tropical storms, but scientists are finding that storms have indeed become more intense, with more hurricanes of categories 4 and 5 in the past few years (Figure 20.5). Hurricanes are expected to continue becoming stronger and more frequent (Figure 20.6).

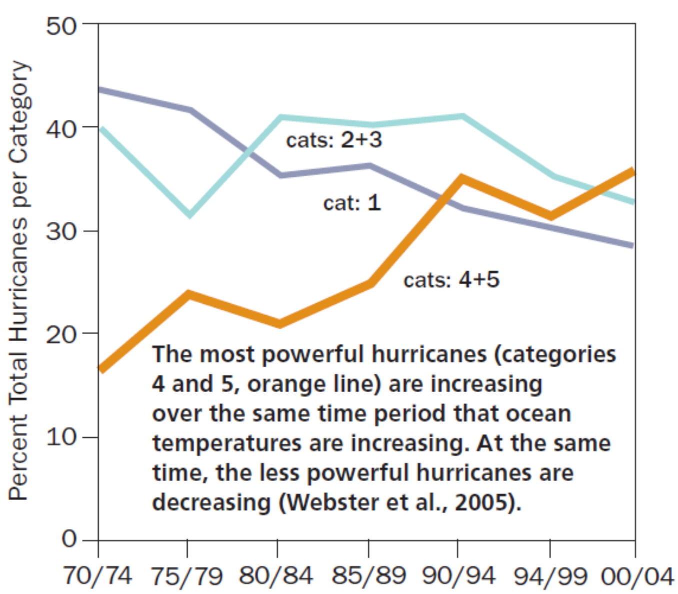


Figure 20.5. Warming of the oceans is leading to an increase in the proportion of category 4 & 5 hurricanes.

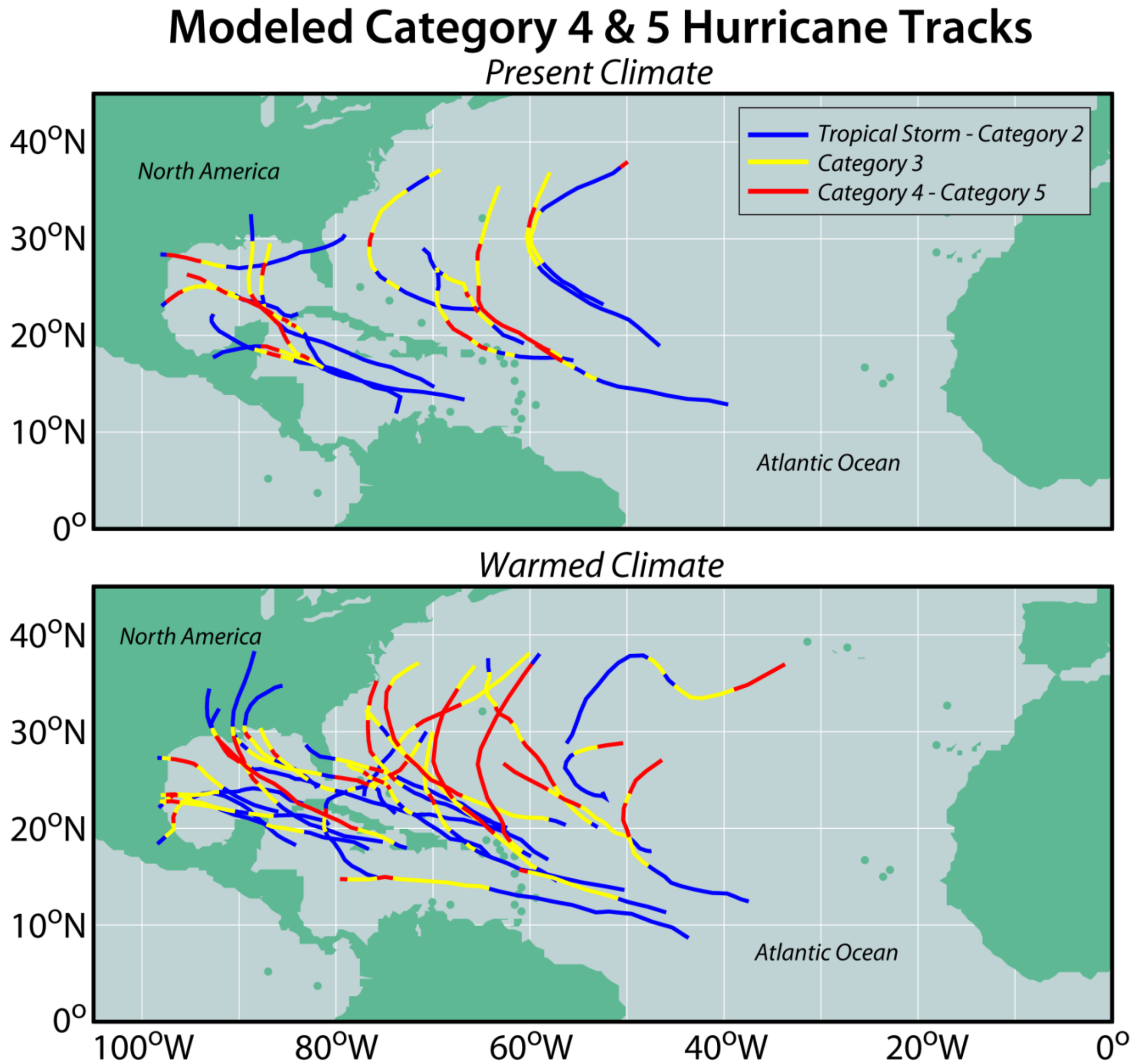


Figure 20.6. Tracks of simulated Atlantic Category 4 and 5 hurricanes for the present climate and for a warmer climate condition projected for the late 21st century. The hurricanes were simulated using higher resolution atmospheric models, with large-scale conditions taken from an ensemble of 18 global climate models. Bender et al. 2010 Science 327 pp. 454-458.

20.3. Changes in deep water circulation

The North Atlantic Ocean is an important source of deep water in the global conveyor belt of thermohaline circulation (section 10.4), as cold, salty water sinks and flows south (Figure 20.7). Global thermohaline circulation patterns could be altered as surface water in this region warms and becomes less salty (due to melting of ice). Global thermohaline circulation is important in the redistribution of heat on Earth, and experts agree that a reduction in the downwelling of North Atlantic water could contribute to further global climate change.

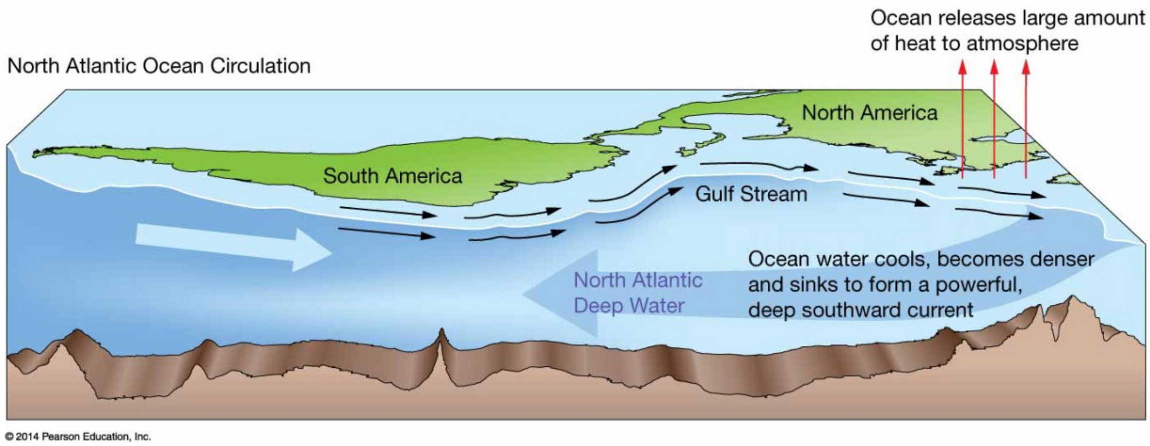


Figure 20.7. North-South section of the Atlantic Ocean showing major currents.

20.4. Melting of polar ice

Polar regions are predicted to experience large changes due to global warming, but the processes are largely different at the South Pole (a continent under a thick ice cap) the North Pole (an ocean with extensive but thinner drift ice, surrounded by land masses). The Arctic Ocean is experiencing the most dramatic changes so far, and is predicted to continue being an area of important climate change impacts. Sea ice cover in the Arctic has declined by half in the last 30 years, and based on proxy data it appears that the current levels have not been seen for at least 1450 years.

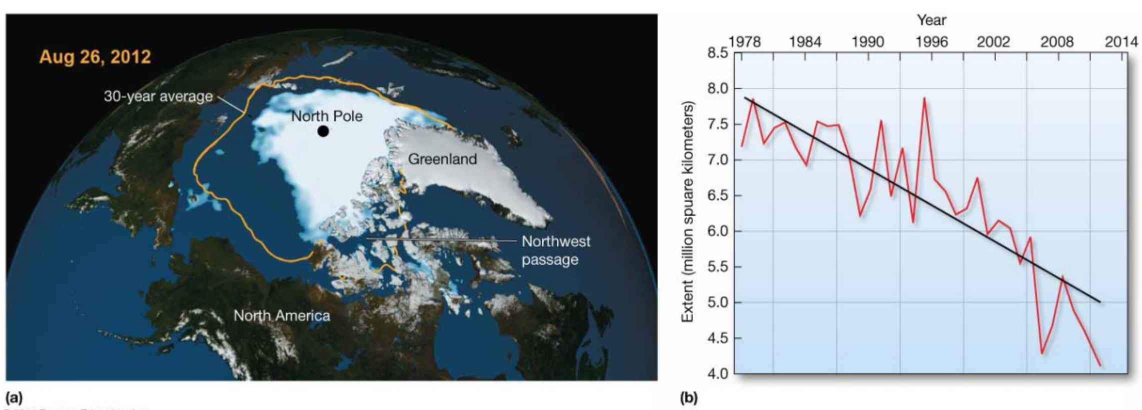


Figure 20.8. Decline of ice cover in the Arctic Ocean.

The decline of ice has very important implications for the ecology of the Arctic Ocean. The loss of ice will have strong impact on the distribution & success of animals that depend on ice. Most notably, polar bears hunt for seals, belugas and other animals on Arctic ice. As the extent of Arctic ice declines, their hunting grounds shrink. Polar bears are also found in increasing numbers on land in the Canadian Arctic, where they find less food and are often a nuisance for human communities. Seals need sea ice to give birth to their pups and to mate, so a reduction in sea ice can also mean decreased reproductive success for these animals. Finally, primary production in the Arctic is strongly tied to sea ice. Here, much of the sea ice is thin and many species of diatoms flourish on the underside of this ice. These are very important primary producers in the Arctic. Loss of ice cover decreases primary productivity which leads to less food available for higher trophic levels.

At the South Pole, warming is greatest around the Western part of Antarctica in the region that includes the Antarctic peninsula. Warming has caused a decrease in the thickness and extend of Antarctic ice sheets, and an increase in the amount of icebergs that calves form the ice sheet. Changes in sea ice extent and Southern Ocean ecology is shifts in the abundance of dominant species. For example, Adelie penguins seem to benefit from the openings in sea ice near their nesting grounds, while Emperor penguins are negatively affected by climate change and are expected to decline by up to 80% by 2100.

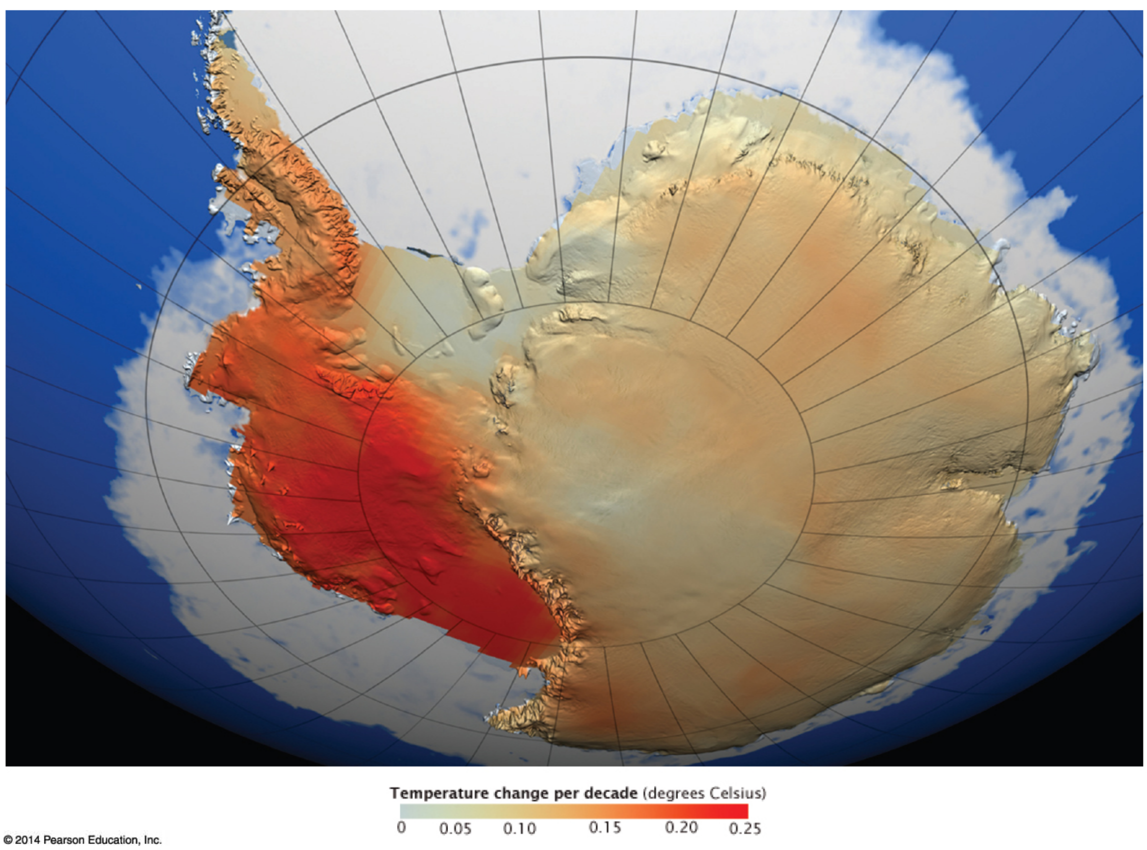


Figure 20.9. Warming trends in Antarctica.

20.5. Rising sea level

Sea level has been rising globally for the last century, due to the melting of polar ice and the thermal expansion of water as it warms (Figure 20.10). In some locations, sea level has increased by as much as 40cm in the past 150 years (Figure 20.11) and it is expected to rise by 0.6-1.6m by the year 2100. Even small changes in sea level can have large impacts on coastal processes, especially in low-lying coastal regions such as Florida. Increased sea level is expected to lead to increased flooding, drowning of beaches, accelerated coastal erosion, loss of coastal wetlands and more extensive damage from storms. Future rises in sea level are difficult to predict as they depend largely on potential sudden changes in the world's major ice sheets. The collapse of the West Antarctic ice sheet, for example, would raise global sea level by about 3.2m.

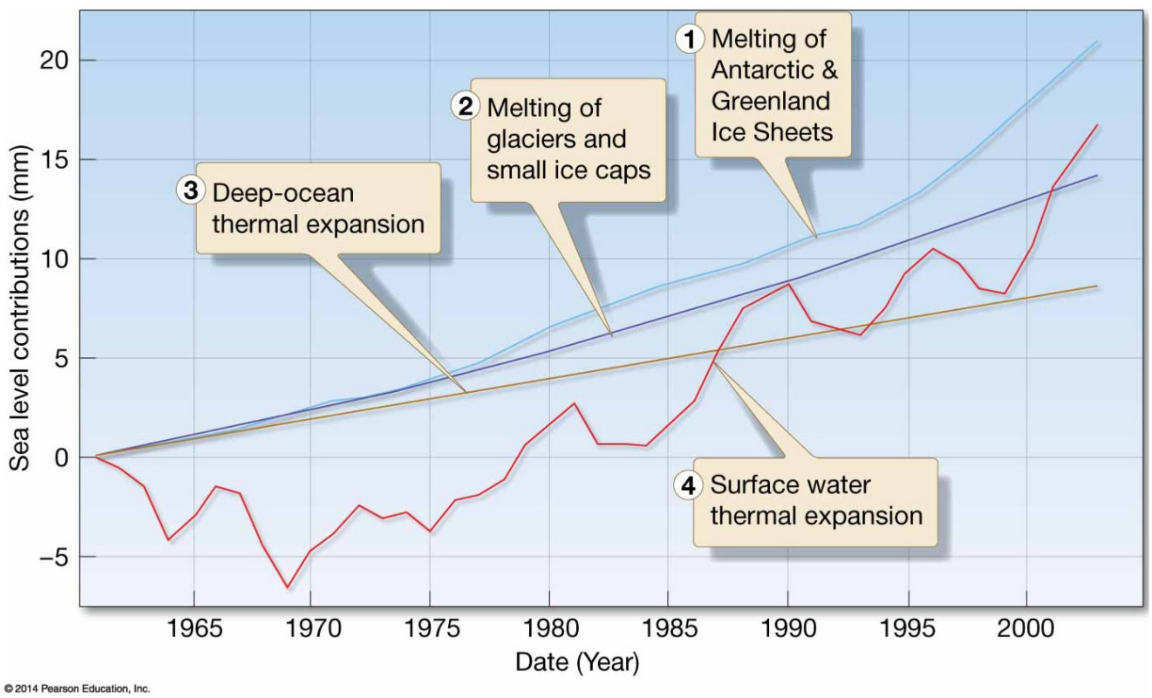


Figure 20.10. Individual contribution to global sea level rise of the four main contributing factors.

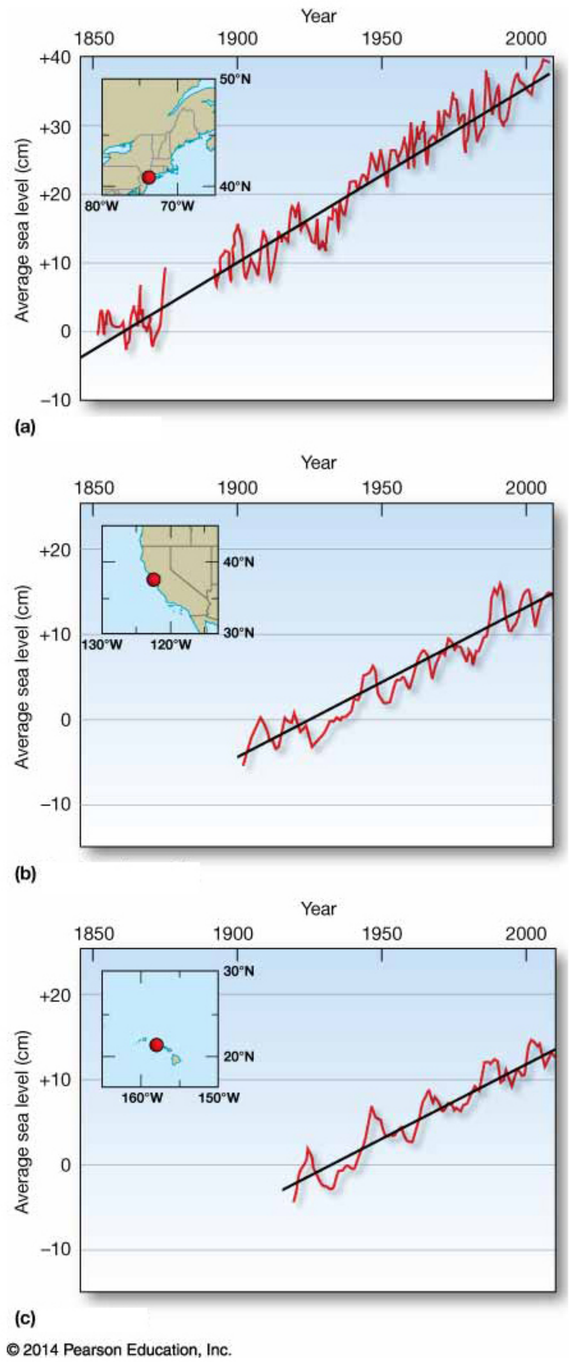


Figure 20.11. Sea level changes at three locations.

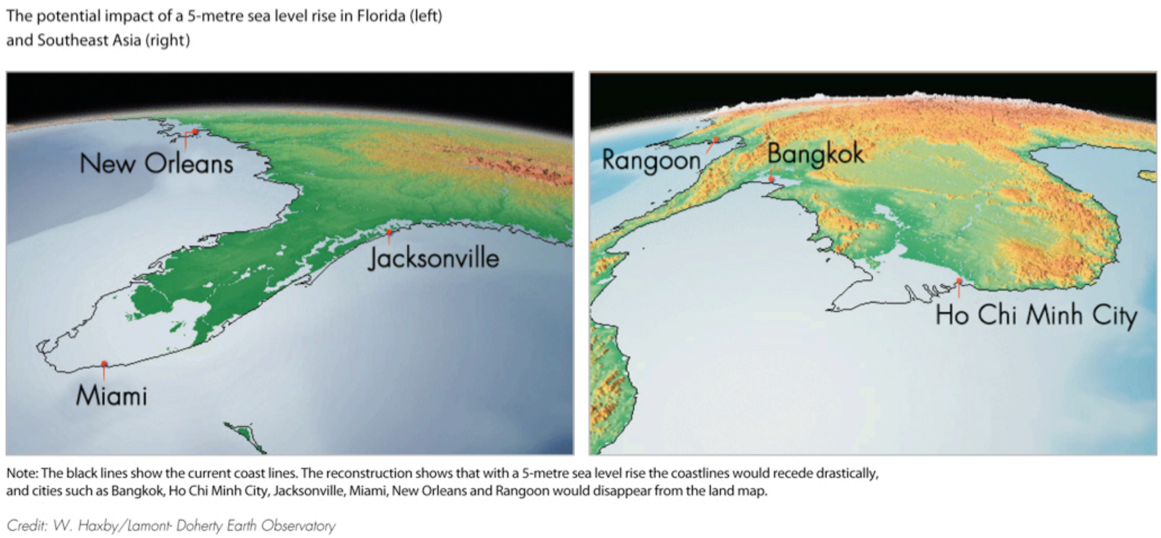


Figure 20.12. Changes in coastlines associated with a 5m rise in sea level.

20.6. Ocean acidification

The effects of climate change discussed in previous sections of this chapter are all linked to the change in temperature caused by increased greenhouse gases (in particular carbon dioxide). Another major impact of increased carbon dioxide in the atmosphere is a change to ocean chemistry. About 1/3 of the CO_2 created by anthropogenic activities ends up in the oceans. While this has the beneficial impact of reducing the greenhouse effect and slowing down global warming but also the consequence of reducing ocean pH (Figure 20.13). The large increases in CO_2 seen in recent decades have overwhelmed the buffering capacity of the ocean. As CO_2 concentration in the oceans increases, it reacts water to form carbonic acid which readily releases free H^+ ions (Figure 20.14). This has led to a decrease in pH of about 0.1 units since industrial revolution in a process called ocean acidification. Because the pH scale is logarithmic, this translates to a ~30% increase in free H^+ ions. A decrease in pH has the potential to impact many physiological processes for a variety of marine species. For example, acidification has been shown to impair hearing and olfaction in reef fish larvae, which in turn can decrease their ability to find a suitable habitat to settle. Ocean acidification impacts most dramatically the groups of organisms that make shells out of calcium carbonate ($CaCO_3$), since lower pH makes it harder to form or maintain these shells (the increased H^+ ions react with dissolved carbonate, an important component of the $CaCO^3$ shells; Figure 20.14). This includes organisms such as corals, echinoderms, coccolithophores and mollusks. Polar waters, because they hold a higher concentration of dissolved gases (including CO_2), have seen some of the largest declines in pH (Figure 20.15). Southern Ocean organisms exposed in lab experiments to projected changes in pH by 2100 show severe responses; krill eggs fail to hatch and the shell of pteropods (planktonic snails) have severe erosion within 45 days (Figure 20.16). As oceans continue to warm and absorb CO_2, the rate of CO_2 uptake will lower, because seawater will near its CO_2 saturation value, and the saturation value decreases with increasing temperature. This means that more of the CO_2 released in the atmosphere will be left there to contribute even more to global warming.

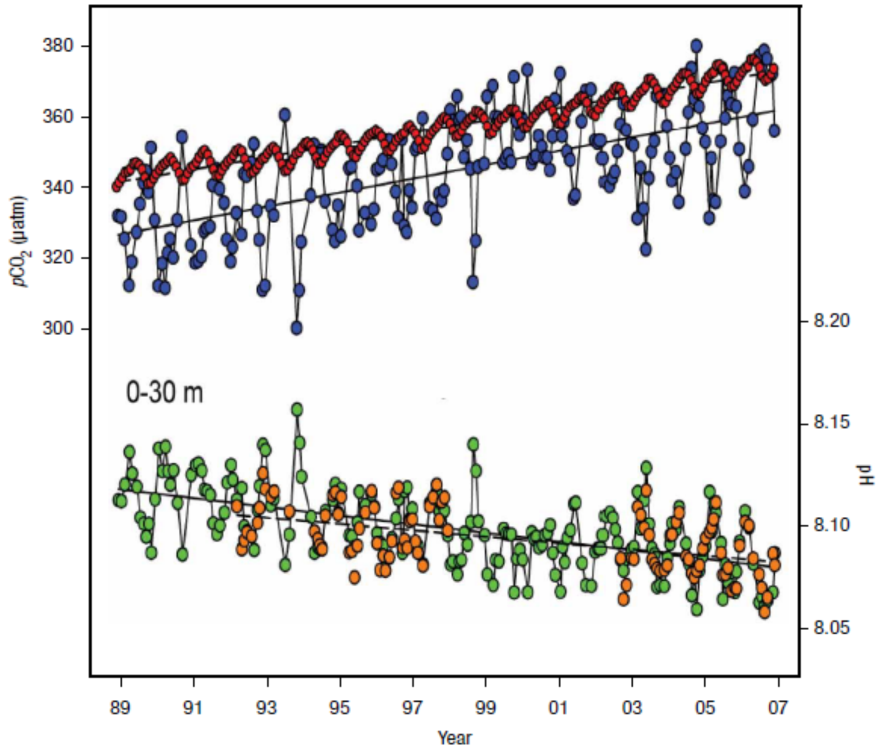


Figure 20.13. As atmospheric CO_2 concentration in the atmosphere (read) and ocean surface increase, ocean pH (green and yellow) decreases. Data from Hawaii.

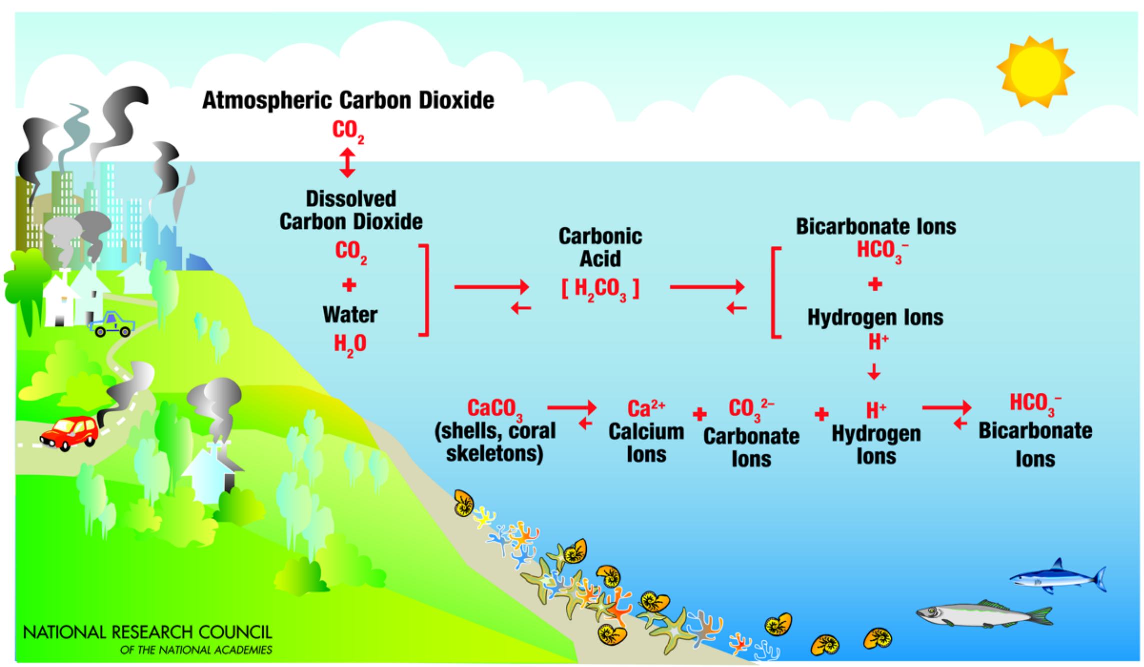


Figure 20.14. Increased dissolved CO_2 leads to ocean acidification: a higher concentration of free H^+ ions. This in turns inhibits the formation of $CaCO_3$ shells.

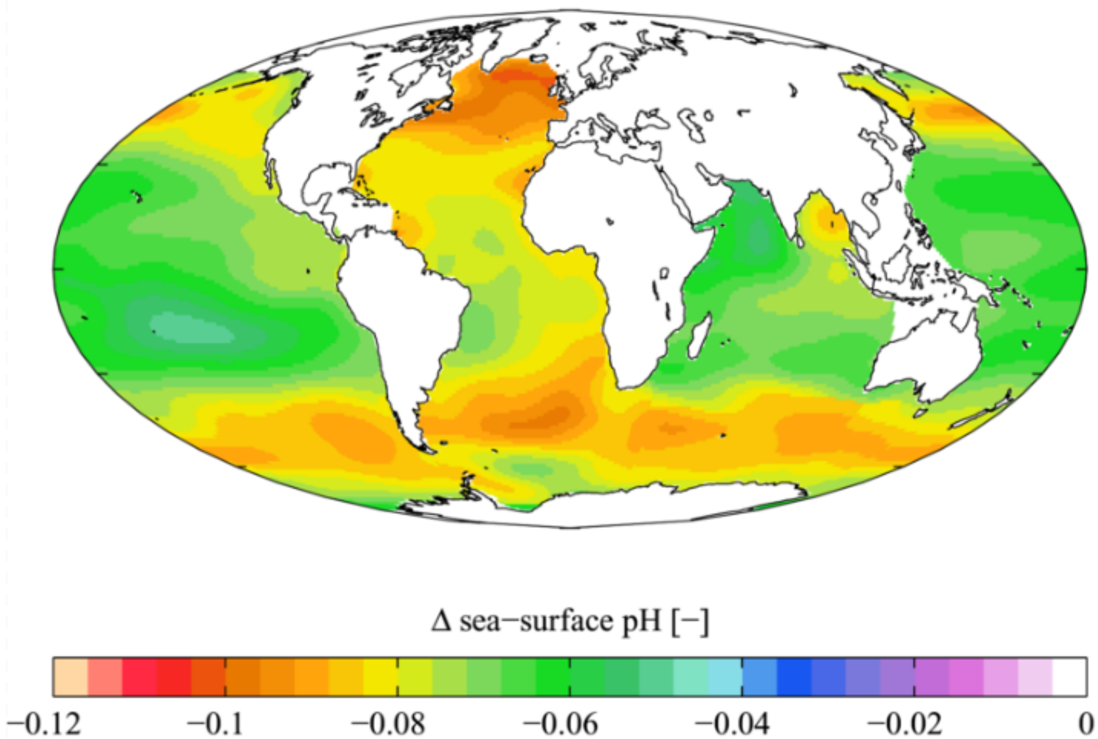


Figure 20.15. Changes in ocean surface pH from 1700 to 1990.

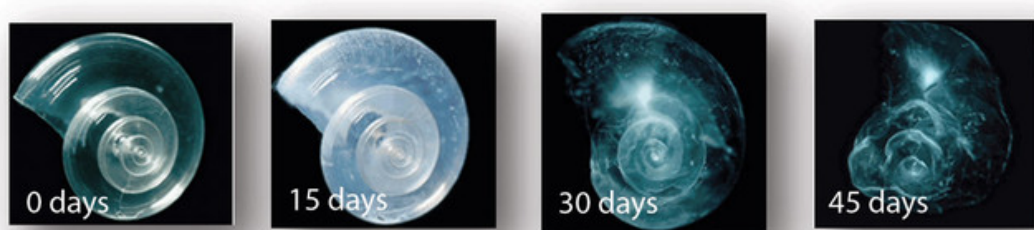


Figure 20.16. Changes in pteropod shells exposed to seawater with the pH projected for the year 2100.

20.7. Other predicted and observed changes

Several other changes can be expected in the oceans with continued warming. First, oceans are expected to have lower dissolved oxygen levels, as warm waters hold less gas (lower saturation value). At the same time, the metabolic rate of many marine organisms will increase along with the temperature rise, leading to more shortages of oxygen and more dead zones. Wind speeds and height of some waves may change. Global ocean productivity may diminish as a larger area of the ocean is stratified with limited nutrients in surface waters. Changes in range towards the poles could modify interactions between species in marine communities, as different species migrate at different rates (Figure

20.17). In summary, changes in the oceans caused by global climate changes are important and far-reaching, altering physical and biological processes throughout ocean basins. Many more potential consequences are likely to occur as our climate continues to change.

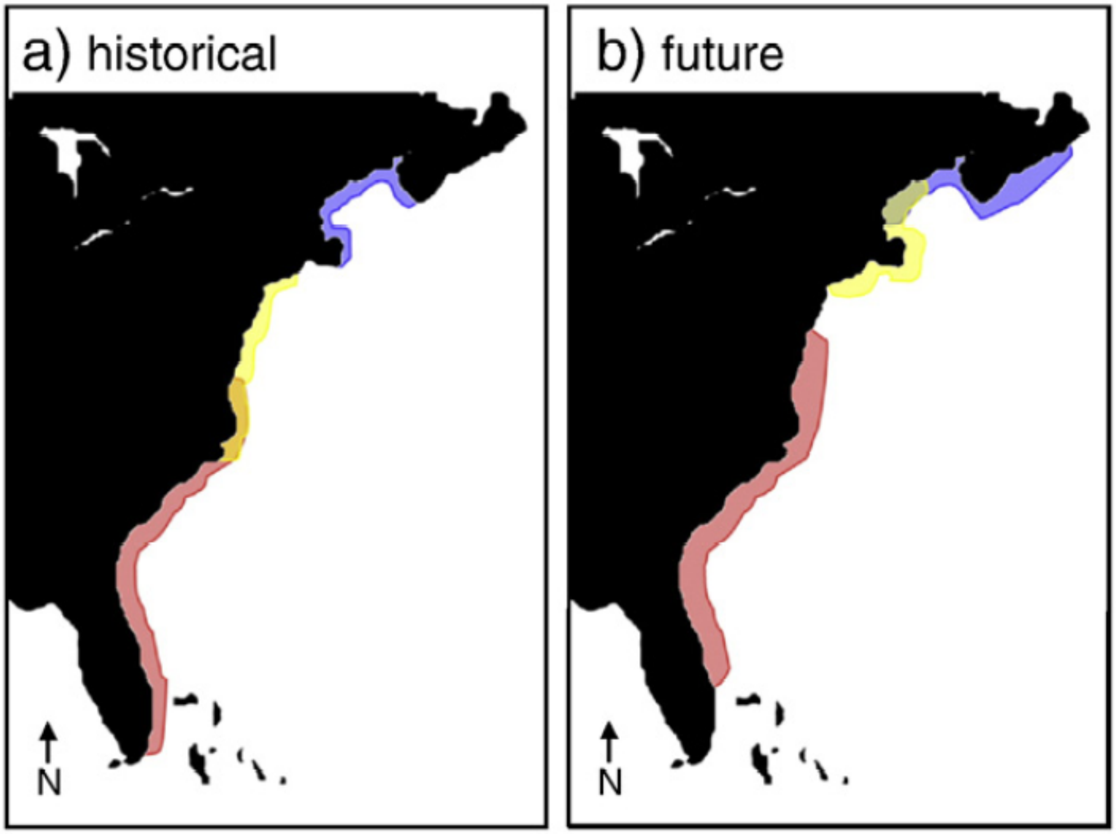


Figure 20.17. Hypothetical changes in interactions between species with poleward migration caused by climate change. In historical ranges, the red species and yellow species overlap around the mid-Atlantic US coast. As all species move northward, if the yellow species migrates faster than others it could eventually overlap with the blue species forming new interspecific interactions (e.g. predation or competition).

20.8. Review Questions

1. In which region of the world have temperatures risen most dramatically in the past few decades?
2. What is coral bleaching? When does it occur?
3. What changes might be expected in the frequency and strength of tropical storms in the next century?
4. How does warming impact deep ocean circulation?
5. What are 3 ecological impacts of the decline of Arctic ice cover?
6. What are the 2 mechanisms by which global warming increases sea level?
7. Why are oceans becoming more acidic?
8. Which regions of the oceans are seeing the greatest changes in pH?
9. Which organisms are especially vulnerable to acidification?
10. How can warming change interactions between species in biological communities?

Appendix 1. Guidelines for Papers & Presentations

1. Literature review assignment

a) Literature review paper

This paper is designed to examine your ability to identify key parameters and issues rising from a literature packet for a given topic. You are encouraged to use additional sources of information, such as the internet, and textbooks, but most of your paper should be based on the scientific papers provided by your instructor.

The paper should be 4 pages long (excluding references and figures), typed in Times New Roman, 12 font, 1.5 spaced, fully justified with 1.2 inch margins.

You will be evaluated on three main components: 1) the overall organization and structure of your paper, 2) the degree of understanding and clarity of explanations and 3) style and grammar. Specifically:

1) Make sure that the paragraph structure is logical (one main idea per paragraph, and each paragraph should start with a topic sentence), and that you have a good introduction and a good conclusion. You should aim to synthesize information from different references into your own logical argument. AVOID simply summarize one reference in one paragraph, and then moving to a different one.
2) Make sure you understand the topic well, and choose what angle you want to present. This is a short literature review paper and it's OK to choose to focus on one particular aspect (e.g. Pacific gyres and Pacific plastic garbage patch, rather than plastic pollution in general). But even if you choose to focus on one topic, make sure that you provide enough general background information for a non-specialist reader to follow. Make sure that you understand the references that you cite, and that you cite anything that is not common knowledge.
3) Proofread your paper to fix any typos and spelling errors. Try to write clear and concise sentences.

b) Presentation

You are also required to present your findings in an 8 minute. You will be evaluated on 1) the general organization of your talk, including a quality introduction and conclusion, 2) how well you understand the topic both in the presentation and in questions that follow, 3) your delivery including body language, speech, and use of visual aids and 4) your timing (should be between 7 and 9 minutes).

2. Group research projects

a) Research Paper

Research papers should be divided into 6 sections: Abstract, Introduction, Methods, Results, Discussion/Conclusion and References. Each of these sections should be labeled. The paper should be typed in Times New Roman, 12 font, 1.5 spaced, fully justified with 1.2 inch margins. There is no limit on how long the paper should be.

Abstract
Short (max. 250 words) summary of your research and its findings.

Introduction
Provide enough background information to explain your project.
State the relevance, i.e. why this research is important.
Clearly state the problem/purpose/hypothesis.
Justify your hypothesis, i.e. why you think that your hypothesis is true.
State the objectives of the project and what you intend to accomplish.
In order to do any of the above, you must refer to external sources such as textbooks, articles and the internet (take caution).

Methods
Make sure you mention all the equipment you used in your research.
Materials should be mentioned in sentence form describing how they were used, i.e. a refractometer was used to measure the salinity of the water.
Describe all the procedures well enough that someone could replicate your experiment.
Describe how the data was analyzed.
Include the time, date and location in this section.
You may find it necessary to refer to external sources and articles when writing your protocol.

Results
Choose tables, graphs and/or statistics that best present your results (eg. means and ranges)
Give a written description of the general trends of your results.
Do not interpret your results
In the written description refer to tables and graphs, i.e. As temperature increases the concentration of dissolved oxygen decreases (Figure 2).
Give actual values to illustrate your point, i.e. The average dissolved oxygen concentration was 6.7 ppm.
Label your graphs and tables. Figure 1 The Relationship between Dissolved Oxygen and Salinity.
If you present your data in a graph you don't need to include a table as well unless it helps you illustrate a point.
Include all your relevant data.

Discussion/Conclusion
Interpret your data and draw valid conclusions.
State problems with your data and how they affect your results.
Suggest ideas for further research.
Refer to your actual data, i.e. The dissolved oxygen concentration was higher in cold water (7.0 ppm) than in warm water (6.5 ppm).
You may find it worthwhile to refer to external sources and articles when writing your discussion.
Science <u>does not</u> prove anything. When interpreting your data, do not say things like" this proves that cold water holds more oxygen than warm water." Instead say "the data suggest that cold water holds more oxygen than warm water". Only if you conduct a test of significance (statistical test) can you say "there is a significant difference between the amount of dissolved oxygen cold water can hold and the amount of dissolved oxygen warm water can hold."

References
See below

Research papers will also be graded on several other factors: creativity, writing style and presentation of ideas, use of complete sentences and grammar, depth of understanding, amount of effort and organization.

b) Presentation

You are also required to present your findings to the group in a 10 minute talk. You will be evaluated on 1) the general organization of your talk, including a quality introduction and conclusion, 2) how well you understand the topic both in the presentation and in questions that follow, 3) your delivery including body language, speech, and use of visual aids and 4) your timing (should be between 9 and 11 minutes).

3. Citing References

When referring to external sources and articles, use both in-text citations and a literature cited section. If there is an in-text citation, it should also appear in the literature cited section and vice versa. External sources should only be included if you used their ideas in your paper. If you read an external source but did not include those ideas in your paper, do not refer to them in in-text citations or in the literature cited section.

In your essays, every time you refer to someone else's work you need to cite the author. For example: "In the 1990's total fisheries leveled off at 90 million metric tons (Halloway, 1999)." You will need a new citation every time you refer to a different author even if that occurs within the same sentence, i.e.: "Dissolved oxygen concentration increases as temperature decreases (Pinet, 1999) and salinity decreases (Duxbury and Duxbury, 1996)." **Each new point** should have at least one in-text citation unless common knowledge. You should reference the same author multiple times if you write several sentences that come from the same source.

In-Text Citations (References)

Most in-text citations will have one of the three following formats:

For one author: (Pinet, 1999)
For two authors: (Duxbury and Duxbury, 1996)
For more than two authors: (Duxbury et. al. 2000)

For example: Dissolved oxygen concentration increases as temperature decreases (Pinet, 1999; Duxbury et al. 2000) and salinity decreases (Duxbury and Duxbury, 1996). When more than one paper is being referenced they should be in order of date (eg Pinet, 1999; Duxbury et al. 2000)

Reference Section

This section should be alphabetized by author first and then by date.

For journal articles: Authors. Year. Name of Article. *Name of Journal* Volume # (Edition #): pages.

With one author: Gonzalez, F. I. 1999. Tsunami! *Scientific American* 280 (5) 56-65.
With two authors: Giese, G. and D. Chapman. 1993. Coastal Seiches. *Oceanus* 36 (1) 38-46.
With more than two authors: Schlee, J.S., D. W. Folger, W.P. Wilson, K.D. Klitgord and J.A. Grow. 1979. The Continental Margins of the Western North Atlantic. *Oceanus* 22 (3) 40-7.

For textbooks: Authors. Year. *Title of Book*. City, State: Name of Press.