

CHAPTER 8

OCEAN IN THE EARTH SYSTEM

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Case-in-Point

By mid-summer, a large volume of water significantly depleted in dissolved oxygen develops in the Gulf of Mexico just offshore from the Louisiana coast. Waters become *hypoxic*; that is, dissolved oxygen drops to less than 2 milligrams per liter of water, causing considerable stress and perhaps death for bottom-dwelling organisms. In the summer of 1999, the hypoxic zone covered about 20,000 square kilometers--approximately the area of New Jersey and double its extent in 1985.

What causes these Gulf waters to become hypoxic? The answer is a combination of factors. Discharge of the Mississippi River into the Gulf reaches its annual peak in spring and summer. The warmer, fresh river water is less dense than the colder saltier seawater so the river water spreads over and floats on top

of the seawater. The Mississippi River also delivers copious amounts of nutrients to the Gulf waters that spur growth of algae populations. Decomposer organisms (i.e., bacteria) consume the organic remains of dead algae and deplete the seawater of much of its dissolved oxygen. Stratification of fresh water over saltwater blocks the transfer of oxygen into the bottom water and hypoxia develops. Seasonal changes in weather cause hypoxic conditions to disappear by late fall. Reduced discharge of fresh water plus mixing of surface waters by storm winds eliminate the stratification of water and the seawater is replenished in dissolved oxygen as it interfaces with the atmosphere.

Nutrients responsible for hypoxic conditions in Gulf waters are primarily compounds of nitrogen,

particularly nitrate, applied as fertilizer to agricultural lands in the Mississippi River drainage basin. The Mississippi drains about 41% of the contiguous U.S., from the Rockies eastward to the Appalachians, with about 58% of the basin in cropland. About 7 million metric tons of nitrogen in commercial fertilizer is applied annually in the basin and about 1.6 million metric tons wash into the Gulf of Mexico. Scientists at the U.S. Geological Survey found that fertilizer use during the prior two years explained about 68% of the variation in nitrate flux into the Gulf and stream flow (due to changes in precipitation and runoff) explained about 18% of the variation. Considerable amounts of nitrate are stored in the soil-groundwater system so that the flux of nitrate is relatively low in dry years and high in wet years.

Driving Question:

What are the general characteristics of the ocean system?

Viewed from space, the ocean is the most conspicuous component of the Earth system, covering 71% of the planet's surface. For this reason, Earth is sometimes referred to as the *water planet*. The ocean consists of salt water occupying huge topographic depressions, the ocean basins. The ocean is by far the largest reservoir in the global water cycle and a major reservoir in the other biogeochemical cycles operating on the planet.

The basic properties of seawater largely depend on the ocean's interactions with the other components of the Earth system, including the atmosphere and lithosphere. Energy and matter are continually exchanged between the ocean and its surroundings. Heat is transferred between the ocean surface and overlying atmosphere with the direction of heat transfer depending on the temperature gradient (i.e., always from warm to cold). The kinetic energy in wind drives waves and surface ocean currents. Chemicals in solution and sediments in suspension are delivered to the ocean by rivers while lava spewing through fractures on the ocean bottom reacts with and chemically alters seawater. Some windblown dust also ends up in the ocean. Celestial bodies external to the Earth system (i.e., the moon and sun) also exert an important influence on the ocean (e.g., tides, solar radiational heating).

This is the first of two chapters that summarize the physical properties of the ocean and its interactions in

the Earth system. In this chapter, we examine the chemical composition of seawater, the impact of solar radiation on the ocean including its temperature structure and productivity, and how seawater density differences cause ocean waters to circulate. We also explore the characteristics of the ocean basins and the causes of the astronomical tides.

Properties of Seawater

Ocean water contains a higher concentration of dissolved materials than water in any other reservoir of the global water cycle. As pointed out earlier in this book, the high salinity (a measure of the concentration of dissolved solids) of seawater precludes its use for most domestic, agricultural, and industrial purposes. Salinity also affects the density and other physical properties of seawater. This section summarizes the composition, pressure, and density of seawater.

COMPOSITION

Trace amounts of just about every element found in Earth's crust are dissolved in ocean water. A fortune awaits the inventor who develops an economically feasible way to extract specific trace elements from seawater. Consider gold, for example. As of this writing, gold is selling for about \$325 a troy ounce, about a penny a milligram. A typical one-kilogram sample of seawater contains about 0.000006 (6×10^{-6}) milligrams of gold, so that about 16 million kg (4 million gallons) of seawater would have to flow through a gold-extraction apparatus to obtain just \$1 worth of gold. But the cost of energy required just to pump this much water would greatly exceed \$1, which is why no ships are at sea attempting to separate gold from seawater.

In the 19th century, chemists began to examine the composition of seawater in some detail, and one of them, C.R. Dittmar, made a very important discovery while analyzing the composition of 77 water samples collected during the world-wide voyage of *H.M.S. Challenger* (from December 1872 to May 1876). Dittmar discovered that although the total amount of dissolved solids varied among water samples, the ratio of the concentrations of the major constituents of seawater was the same in all samples. The major constituents of seawater occur in the same relative concentrations

throughout the ocean system (Table 8.1), a characteristic of seawater described as the **principle of constant proportions**.

The major dissolved constituents of ocean water, such as chloride (Cl^-) and sodium (Na^+) ions, are *conservative constituents*; that is, they occur in constant proportions or change concentration very slowly. Seawater constituents that participate in biogeochemical or seasonal cycles have variable concentrations and are described as *non-conservative*. Recall from Chapter 1 that the ocean is a reservoir in all biogeochemical cycles operating on the planet. Examples of non-conservative constituents of seawater are oxygen, carbon dioxide, silica and calcium compounds, nitrates, phosphates, and aluminum.

The principle of constant proportions means that measurement of the concentration of a major conservative constituent of seawater is sufficient to determine the concentration of any other conservative constituent (or the sum of all other constituents). This principle made it possible for oceanographers to define and measure salinity. Prior to World War II, *salinity* was defined as the total amount of solid materials in grams dissolved in one kilogram of seawater. The standard method for measuring salinity was chemical titration to determine the concentration of the chloride ion (Cl^-), which was then entered into a simple formula to calculate salinity. The value of salinity was expressed in grams of dissolved material per kilogram of seawater (g/kg) or parts per thousand (ppt). The accuracy of this method was about ± 0.02 ppt.

After World War II, scientists found that

salinity could be determined from measurements of the electrical conductivity of seawater with 10 times greater accuracy than the old chemical titration method. In 1969 the international oceanographic community accepted a new definition of salinity based on conductivity measurements; that is, the salinity of a water sample is defined as the ratio of the conductivity of the sample to the conductivity of *standard seawater*. The Institute of Oceanographic Services in Wormly, England prepares and supplies samples of standard seawater to laboratories around the world for use in calibrating salinity-measuring instruments. As a ratio, salinity has no units and may be written as a pure number, but for convenience we express salinity in *practical salinity units (psu)*. The conductivity-based definition of salinity was designed to retain the validity of measurements made the old way; that is, one psu essentially equals 1.0 part per thousand. A seawater sample having a salinity of 34.82 psu contains about 34.82 grams of dissolved materials per kilogram.

Rivers are a major route whereby the global water cycle delivers dissolved materials to the sea, but river water and seawater differ markedly in salinity and the type of dissolved components. The average dissolved load of rivers is 119 parts per *million*, as compared to 35 parts per *thousand* for typical ocean water; that is, the ocean is about 300 times saltier than most rivers. Chloride and sodium ions, the most common constituents of seawater, are very limited in river water. And the carbonate ion, the number one dissolved constituent of river water, does not even appear in seawater. After entering the ocean, dissolved materials follow four possible pathways: remain in solution, precipitate out of solution, adsorb onto sinking sediments, or taken up by living organisms.

Processes that operate at the atmosphere/ocean interface add or remove water molecules from seawater and largely explain spatial variations in salinity. Most dissolved substances are left behind when seawater evaporates or freezes, increasing the surface salinity locally. On the other hand, precipitation, runoff from rivers, and melting ice add fresh water and decrease the local surface salinity. Mixing of two or more volumes of water having different salinities also alters salinity. On an average annual basis, the net effect of freezing of water and melting of ice on near sea-surface salinity is negligible because the increase in salinity produced by ice formation in winter is offset by the decrease in salinity produced by melting ice during summer. Where

TABLE 8.1
Dissolved Ions in Average Seawater

<i>Ion</i>	<i>Psu</i>	<i>Percent by weight</i>
Chloride (Cl^-)	18.980	55.04
Sodium (Na^+)	10.556	30.61
Sulfate (SO_4^{2-})	2.649	7.68
Magnesium (Mg^{2+})	1.272	3.69
Calcium (Ca^{2+})	0.400	1.16
Potassium (K^+)	0.380	1.10
Bicarbonate (HCO_3^-)	0.140	0.41
Bromide (Br^-)	0.065	0.19
Boric acid (H_3BO_3)	0.026	0.07
Strontium (Sr^{2+})	0.013	0.04
Fluoride (F^-)	0.001	0.001
TOTAL	34.482	99.99

large rivers enter the ocean, salinity is reduced as fresh water and seawater mix; salinity is also less where rainfall is heavy. Runoff is only important in coastal regions, so evaporation, precipitation, and mixing are the three principal processes governing surface salinity over most of the ocean.

Climate is the principal control of the average pattern of seawater salinity. On a global scale, the surface salinity of the ocean peaks near the center of each ocean basin, that is, in the subtropics (between 20 and 30 degrees N and between 15 and 30 degrees S) where annual evaporation rate exceeds precipitation (Figure 8.1). Semipermanent subtropical anticyclones dominate the climate of these areas and bring mostly cloud-free skies and little precipitation. Average surface salinity is generally lowest near the equator and in polar areas where annual precipitation is greater than the rate of evaporation.

Gases dissolved in seawater consist of oxygen, nitrogen, and carbon dioxide, all found in relatively large amounts, plus less abundant gases such as helium, neon,

and argon. Gases enter the sea from the atmosphere at the air-sea interface (and from volcanic activity on the ocean floor). The flux of gases entering seawater from the atmosphere depends on the amount of gas already in solution and the temperature. Recall from Chapter 2 that the saturation concentration of a gas in water decreases with rising temperature. When surface waters are saturated, a gas is expelled into the air at the same rate as the gas dissolves in water. When surface waters are unsaturated, gas dissolves in water at a greater rate than it is expelled into the atmosphere.

Oxygen and carbon dioxide are heavily involved in biological processes (i.e., photosynthesis and cellular respiration), so their concentrations vary markedly with depth and from one part of the ocean to another. Nitrogen (N_2), the principal atmospheric gas, is used minimally in biological processes and its concentration is relatively fixed throughout the water column. Two sources replenish dissolved oxygen in surface ocean waters: the atmosphere and marine plants. The atmosphere supplies additional oxygen directly to

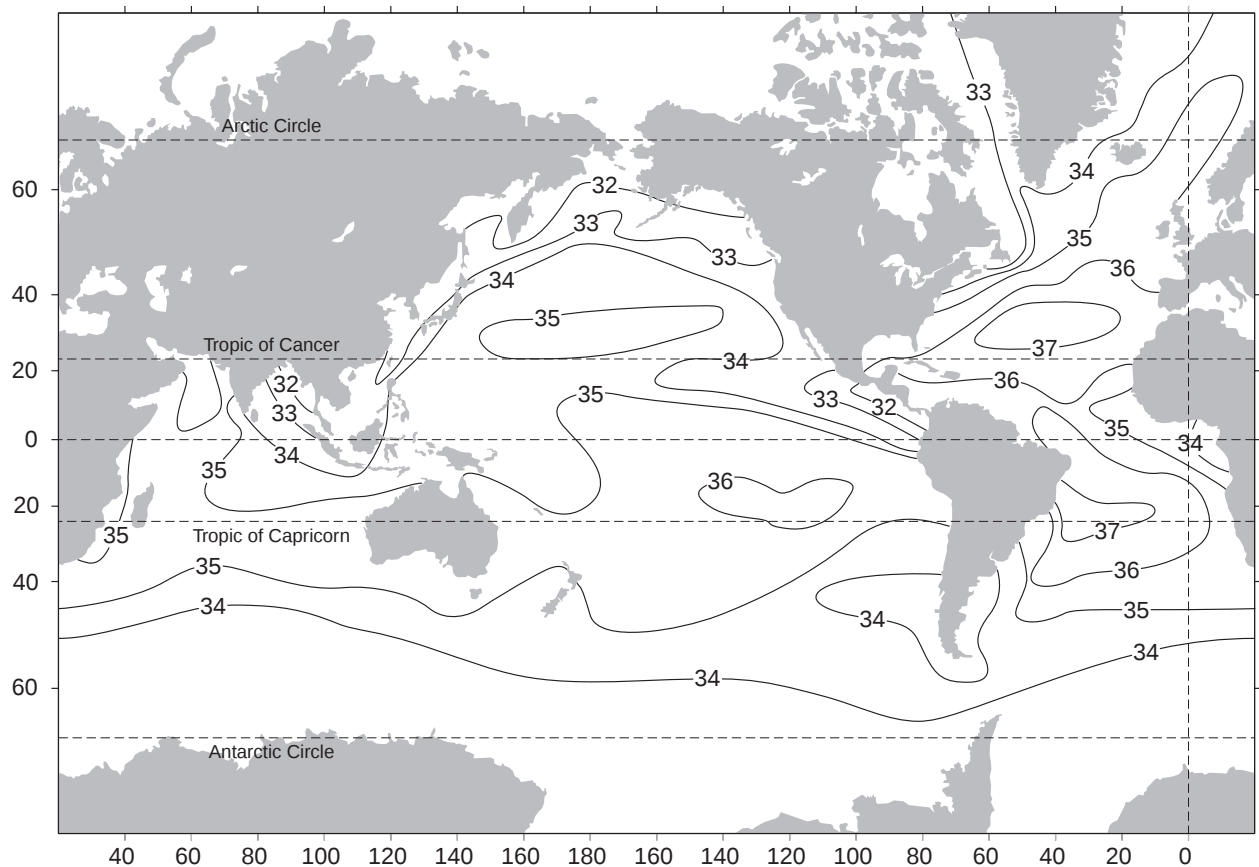


FIGURE 8.1
Global pattern of average sea-surface salinity (in psu) for August.

the ocean whenever the level of dissolved oxygen in surface waters dips below saturation. Marine plants (e.g., phytoplankton) produce oxygen (while utilizing carbon dioxide) in photosynthesis. As discussed later in this chapter, sunlight does not penetrate in quantities adequate for photosynthesis at depths greater than about 100 m (330 ft). *Downwelling*, the downward motion of surface water, transports oxygen-rich surface waters to great depths in the ocean. Although oxygenated water circulates within the ocean, new supplies of oxygen can only come from the surface layers. Once oxygen dissolves in water, it is available for use by animals in respiration, by bacteria in the decay process, and in chemical reactions involving oxidation. Animals use oxygen and generate carbon dioxide, whereas plants use carbon dioxide and generate oxygen so that oxygen and carbon dioxide are continually cycled between the marine plant and animal kingdoms.

PRESSURE

Ocean water exerts pressure. Recall from Chapter 4 that *pressure* is defined as a force per unit area. Atmospheric pressure can be thought of as the weight of a column of air acting over a unit area at the base of the column. By convention, *standard* atmospheric pressure is the average air pressure at sea level at latitude 45 degrees and an air temperature of 15 °C (59 °F); it equals 1.03 kg per square cm or 1013.25 mb (14.7 lbs per square in.). Water is much denser than air and a column of equivalent height produces much greater pressure.

A column of water having a height of 10.33 m (33.9 ft) exerts a pressure at its base equal to one standard atmosphere. Hence, the pressure at the bottom of the Marianas Deep, where the ocean depth is about 11,000 m (36,000 ft), is over 1000 times greater than standard atmospheric pressure. At that depth, the water pressure is 1097 kg per square cm (7.8 tons per square in.). For all practical purposes water is *incompressible*; that is, water density does not vary significantly with increasing pressure. To a good first approximation, the pressure at any point in a water column is directly related to depth and the relationship is linear; that is, doubling the depth doubles the pressure. At a depth of 10.33 m, the water pressure is 1.03 kg per square cm. At ten times this depth, 103.3 m (339 ft), the water pressure is ten times as great, 10.3 kg per square cm. The tremendous pressures encountered in deep ocean waters prevent

direct exploration except by special submersible vehicles designed to withstand potentially crushing stresses.

Oceanographers, like their counterparts in meteorology, commonly use the *bar* and its derivatives as a standard unit of pressure. Numerically, the water pressure expressed in *decibars* (0.1 bar) is the same as the water depth expressed in meters (with an error of less than 2%). The interchangeability of the two measures (i.e., water pressure in decibars and water depth in meters) greatly simplifies data analysis. For example, the water pressure at 100 m (330 ft) depth is about 100 decibars.

DENSITY

Relatively small but meteorologically significant changes in air pressure characterize the atmosphere. Consequently, one of the most important analytical tools used by meteorologists is analysis of the variation of air pressure with respect to both time and location. In the ocean, however, a small error in depth determination generates a large error in pressure measurement, so ocean scientists use water density rather than water pressure as their principal analysis parameter. Just as pressure fields in the atmosphere are plotted on weather maps using *isobars* (lines joining points of equal pressure), density fields in the ocean are plotted on charts using *isopycnals* (lines joining points of equal density).

Density is the mass contained in a unit volume, usually expressed as grams per cubic cm or kg per cubic m. The density of pure water at a temperature of 3.98 °C (39.16 °F) and standard atmospheric pressure is 1.0 gram per cubic cm, equivalent to 1000 kg per cubic m. As noted earlier, the density change accompanying a change in pressure is insignificant even for large pressure changes. But as discussed in Chapter 2, the variation in seawater density with changes in temperature and salinity is significant.

The density of fresh water is maximum at a temperature of 3.98 °C (39.16 °F). Any rise or fall of temperature from this value is accompanied by a decrease in water density. As salinity increases, the temperature of maximum density and the freezing temperature of water decline—but at different rates (Figure 2.12). At a salinity of 24.7 psu, the temperature of maximum density coincides with the freezing temperature. This is the conventional basis for distinguishing between seawater and brackish water. The salinity of *seawater* is greater than 24.7 psu and the

salinity of *brackish* water is less than 24.7 psu. Usually, brackish water is seawater that has been diluted by river water and occurs mainly in estuaries and coastal areas.

The density of seawater and brackish water increases with increasing salinity. Brackish water density either increases or decreases as the temperature changes depending on the initial temperature (as is the case with fresh water). The average salinity of the ocean is about 35 psu and almost always falls into the seawater class. Consequently, no matter what the temperature of seawater, a rise in temperature will always decrease its density and a fall in temperature will always increase its density.

Temperature usually plays a greater role than salinity in producing variations in seawater density. (This is not generally true for brackish water, where salinity may dominate density.) More than 75% of the volume of ocean water is remarkably uniform in both temperature and salinity with temperatures ranging between 0 °C and 5 °C (32 °F to 41 °F) and a salinity of 34 to 35 psu. But considering 99% of ocean water volume, the full range of ocean water temperature is much greater than that of salinity. Temperature extremes are roughly –2 °C to 30 °C (28 °F to 86 °F) whereas salinity narrowly ranges between about 33 and 37 psu. The greater variability of ocean temperature (especially in surface waters) translates into a more important role for temperature than salinity in density variations.

Processes responsible for changes in temperature (e.g., solar heating) and salinity (e.g., precipitation, evaporation, river runoff) are confined essentially to the surface layer of the ocean, that is, where the ocean interfaces with the atmosphere. If relatively dense water sinks, it is no longer in contact with the atmosphere and its temperature and salinity (and therefore density) remains nearly constant. Temperature and salinity are *conservative* properties of volumes of water (called *water masses*) that allow them to be tracked from one place to another in the ocean. The density of seawater varies from about 1.02 gram per cubic cm for warm surface waters to around 1.05 gram per cubic cm for cold water at great depths. Although differing by at most only 3%, this density contrast is sufficient to drive large-scale thermohaline ocean circulation as we will see later in this chapter.

Density is very difficult to measure directly, especially at sea, but it may be determined readily from measurements of salinity, temperature, and pressure (or depth). An *in situ* **CTD** (conductivity-temperature-

depth) instrument monitors and records values of these three parameters at various locations within the ocean. These measurements are converted to densities, which form the basis for density field studies.

Ocean Temperature, Productivity, and Color

The interaction of incident solar radiation with ocean water governs ocean temperature, marine productivity, and the color of seawater. The ocean is a strong absorber of solar radiation. On a global annual average, the ocean surface reflects only about 8% of incident visible solar radiation; the other 92% is absorbed by water molecules and converted to heat. About 65% of visible light that enters seawater of the open ocean is absorbed within the upper 1 m (3.3 ft) and typically no more than 1% reaches a depth of 100 m (330 ft).

The temperature of seawater is largely determined by solar radiation absorbed in the surface layer of the ocean and heat transfer at the interface between the ocean and atmosphere (Chapter 10). In general, seawater temperature declines with increasing latitude and increasing depth. Most of the ocean is divided into three permanent layers based on the vertical profile of average water temperature: mixed layer, main thermocline, and deep layer (Figure 8.2).

At mid latitudes, winds (especially those associated with intense storms) circulate water from the ocean surface down to a maximum depth of about 450 m (1475 ft); this is the so-called **mixed layer**. Water temperature in the mixed layer averages about 15 °C (59

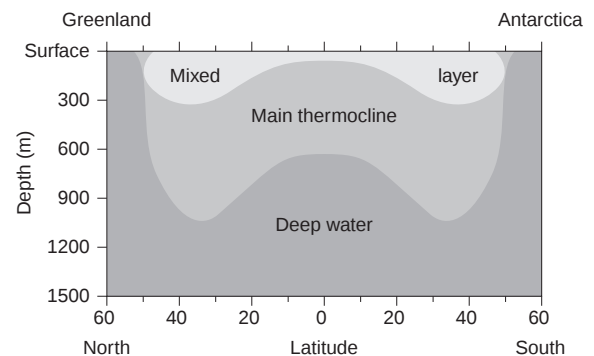


FIGURE 8.2
Subdivision of the ocean based on average vertical temperature profile from 60 degrees N to 60 degrees S.

°F). Below the mixed layer, no matter how violent a storm might be, the water is undisturbed. The **deep layer**, occurring below a depth of about 1200 m (3900 ft) in mid latitudes, is also at a nearly constant temperature, usually under 4 °C (39 °F). The transition zone between the warmer water of the mixed layer and the colder water of the deep layer is the **main thermocline** (*therm* for temperature and *cline* for rapid change with depth). At mid latitudes, the main thermocline is approximately 800 m (2600 ft) thick. In the tropics, the mixed layer is markedly thinner than at mid latitudes, as is the main thermocline. At high latitudes, poleward of about 50 degrees N and S, the deep layer extends to the surface; the mixed layer and thermocline are absent.

Variation in ocean temperature with depth for typical low, middle, and high latitude locations in winter is shown in Figure 8.3. Seasonal variations in temperature at high and low latitudes are almost negligible. At mid latitudes, however, the seasonal variation in surface water temperature is much more significant. In spring, in response to longer daylight and more intense solar radiation, surface waters warm by about 5 Celsius degrees (9 Fahrenheit degrees) to a depth of perhaps 15 or 20 m (50 or 65 ft), and a *seasonal thermocline* develops at that depth. As the water temperature continues to rise through the summer months, the surface warm layer thickens to about 50 m (165 ft), and the contrast in temperature between surface water and water at 50 or 60 m increases to as much as 10 Celsius degrees (18 Fahrenheit degrees). In the fall, in response to shorter daylight and weakening solar radiation, surface waters cool and become prone to wind-driven overturning. The seasonal thermocline lowers to a depth of perhaps 75 to 100 m (250 to 330 ft), with a

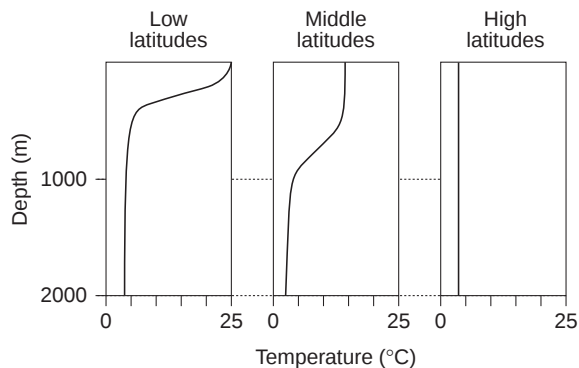


FIGURE 8.3
Average winter temperature profiles of the ocean at low, middle, and high latitudes.

temperature difference across the thermocline of about 5 Celsius degrees (9 Fahrenheit degrees).

Surface water at temperatures higher than 10 °C (50 °F) generally occurs only between 45 degrees N and 45 degrees S and extends downward to the permanent thermocline. In this broad latitude belt, intense solar radiation promotes high sea surface temperatures, considerable evaporation, and increased salinity. However, lower density caused by high temperatures outweighs higher density caused by increased salinity. Poleward of 45 degrees N and S, surface waters are denser, having been chilled by cold air and evaporation.

Based on the diminishing availability of light with depth, two layers are recognized in the ocean water column. The sunlit upper layer, known as the **photic zone**, is generally 100 m (333 ft) or less in thickness and receives adequate light for photosynthesis. Below is the **aphotic zone** where too little light penetrates for plants to survive. With few exceptions, marine life depends directly or indirectly on sunlight and organic productivity in the photic zone. Even the diverse community of animals that live at great depths in the ocean depend on organic substances produced within the photic zone that settle to the sea floor, the most notable exception being organisms that live near hydrothermal vents (described later).

The visible portion of sunlight is composed of all the colors and ocean water absorbs visible light selectively by wavelength. Selective absorption explains the blue color of the open ocean. Water absorbs the longer wavelengths of visible light (i.e., red and yellow) more efficiently than the shorter wavelengths (i.e., green and blue) so that green and blue penetrate to greater depths. More blue/green light is available to be scattered and reflected to our eyes. In shallow near-shore water, suspended solids such as particulate matter and tiny floating plants (phytoplankton) make the water turbid, so that light penetrates no deeper than about 20 m (65 ft) and sometimes much less. Suspended particles scatter yellow and green light preferentially, giving near-shore waters their characteristic color.

Thermohaline Circulation

Density contrasts that arise from differences in temperature and salinity generate subsurface movements of water masses, a phenomenon known as **thermohaline circulation**. A **water mass**, analogous to an air mass, is

a large, homogeneous volume of water featuring a characteristic range of temperature and salinity. Circulation resulting from density differences operates on a simple principle: a water mass of greater density (colder and/or more saline) will displace and flow beneath a water mass of lesser density. For example, relatively cold and dense waters generated at high latitudes drift equatorward and slide under the warmer, less dense surface waters formed at lower latitudes. Thermohaline circulation involves about 90% of all ocean water, but oceanographers know less about it than surface currents because studying deep-ocean circulation is expensive and time-consuming. Furthermore, the flow of water masses is very slow and difficult to measure. A thousand or more years may elapse before a water mass generated at high latitudes returns to the surface.

Density alone is not sufficient to identify water masses and their origins because many combinations of temperature and salinity can produce the same density value. As shown in the temperature-salinity (T-S) diagram in Figure 8.4, the relationship between a specific value of density and various combinations of temperature and salinity is non-linear; that is, lines of constant density (called *isopycnals*) are curved. This non-linear relationship has implications when water masses mix. That is, mixing of two water masses having the same density results in a new water mass of greater density, a process known as **caballing**. Consider, for example, water mass *A* and water mass *B* having different temperatures and salinities but the same density as plotted in Figure 8.5. When these two water masses mix

in equal proportions, the temperature (*T*) of the mixture (water mass *C*) is given by the simple arithmetic average of the temperatures of the two water masses. That is,

$$T_C = (T_A + T_B)/2$$

And the salinity (*S*) of water mass *C* is given by the simple arithmetic average of the salinities of the two water masses. That is,

$$S_C = (S_A + S_B)/2$$

But the temperature and salinity of water mass *C* (the mixture) corresponds to a density that is greater than the density of the original water masses *A* and *B*.

A water mass having acquired its temperature and salinity characteristics at the ocean surface then flows to a depth where its density is the same as the density of the surrounding water. When water evaporates, the remaining water becomes more saline, and its density increases. Likewise, water chilled in polar areas becomes denser. As noted earlier, once a water mass moves from its source region, its temperature and salinity remain remarkably constant, except at its margins where some mixing occurs. Most of the cold, dense water that occupies the ocean basins originated at polar latitudes where heat was conducted or radiated to the frigid atmosphere.

Oceanographers classify water masses by the depth at which they reach equilibrium with their surroundings and their geographical source region. In

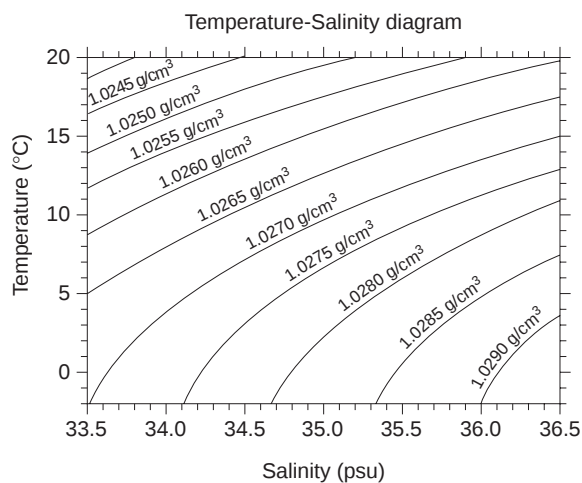


FIGURE 8.4
A temperature-salinity diagram.

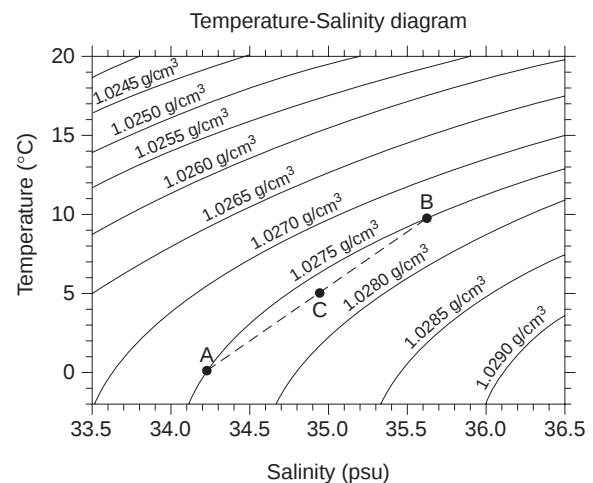
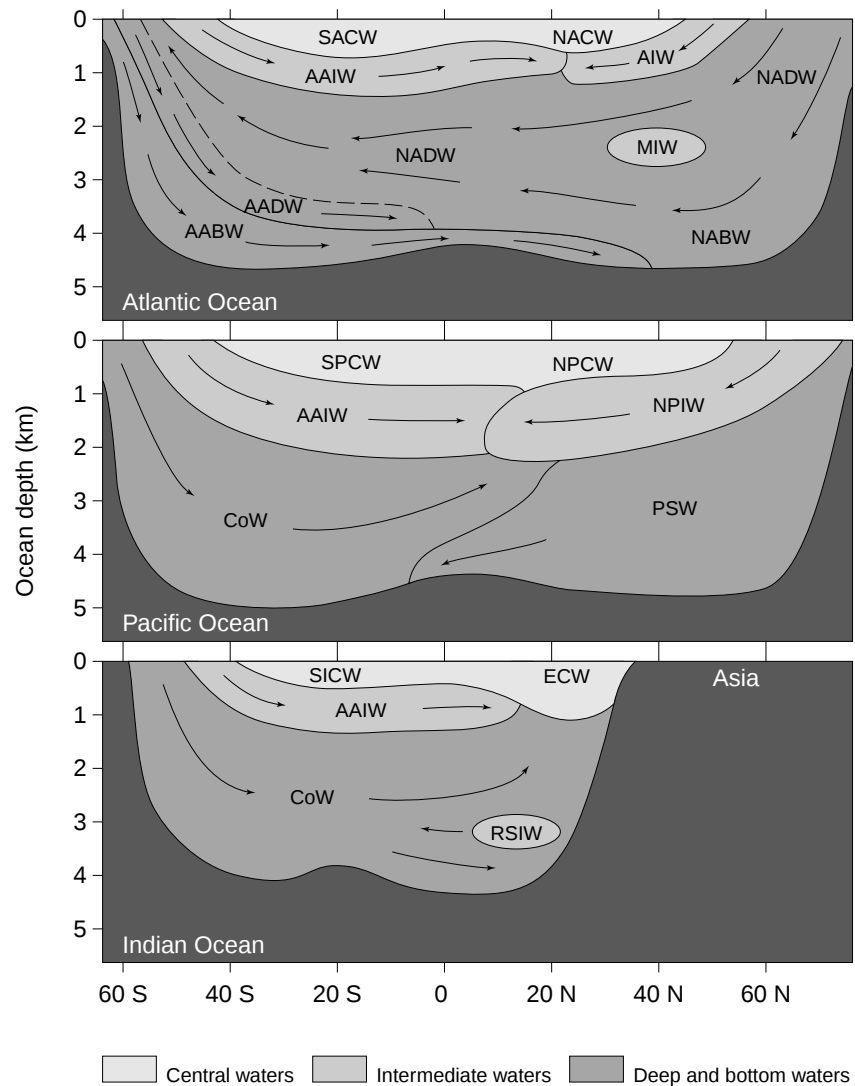


FIGURE 8.5
Mixing of water mass *A* with water mass *B* produces a denser water mass *C*.

FIGURE 8.6

Vertical distribution of water masses in the Atlantic, Pacific, and Indian Oceans.



order of increasing depth, water masses are designated as *central waters* (surface down to about 1.0 km), *intermediate waters* (1.0 to 2 km), and *deep and bottom waters* (greater than 2 km). Source regions are the Atlantic, Pacific, and Indian Oceans (Figure 8.6). As we will see in Chapter 10, heat transport by moving water masses—likened to conveyor belts—is an important control of global climate.

ATLANTIC WATER MASSES

In the immediate vicinity of the Antarctic continent, particularly the Weddell Sea, surface water temperatures drop to extremely low levels in winter. Chilled by exceptionally cold air masses and made more saline by ice formation, this water mass has the highest

temperature-salinity related density of any in the ocean. Having once gained these characteristics at the surface, *Antarctic Bottom Water* (AABW) sinks and flows northward along the ocean floor toward the equator and into the western North Atlantic basin. In fact, traces of this water mass have been detected as far north as 45 degrees N. *Antarctic Deep Water* (AADW) is slightly warmer and less saline than Antarctic Bottom Water, having formed at somewhat lower latitudes. AADW flows northward and plunges downward under the less-dense subpolar water of the South Atlantic at the so-called *Antarctic Convergence*. The *Antarctic Convergence* surrounds the Antarctic continent and is a zone where contrasting water masses come together. Beneath the surface, AADW is sandwiched between the denser Antarctic Bottom Water below and the less-dense

North Atlantic Deep Water above. *North Atlantic Deep Water (NADW)* originates near Greenland through winter cooling and evaporation and is the most abundant of any bottom water in the ocean.

Arctic Intermediate Water (AIW) forms in the subarctic, flows southward under the warm and highly saline water of the Sargasso Sea of the subtropical Atlantic, and converges with the northward flowing *Antarctic Intermediate Water (AAIW)*. Another intermediate water mass of some importance, but generated outside the Atlantic boundaries, is *Mediterranean Intermediate Water (MIW)*. It forms in the northern and eastern portions of the Mediterranean through intense heating and evaporation, which can raise the salinity to levels higher than 36.5 psu. MIW sinks to the bottom of the Mediterranean Basin and spills over the sill at the Strait of Gibraltar, spreading westward into the North Atlantic as a tongue of warm, salty water.

Atlantic waters are constantly being renewed at all depths at a very slow rate, but not as slowly as the Pacific renewal rate. Radioactive carbon measurements indicate that the *water age* (time since the water was last in direct contact with the atmosphere) of Antarctic Bottom Water is about 750 years in the Atlantic and approximately 1500 years in the Pacific.

PACIFIC WATER MASSES

As it flows around the Antarctic continent in the South Atlantic Ocean, Antarctic Bottom Water mixes with North Atlantic Deep Water producing what is referred to as *Common Water (CoW)*, the principal water type delivered to the bottom of the Pacific. CoW sinks and flows northward to near 10 degrees S where it converges with *Pacific Subarctic Water (PSW)*. The relatively shallow Bering Strait greatly reduces the influx of Arctic Bottom Water into the Pacific and very little Antarctic Intermediate Water is produced in the South Pacific. Consequently, the Pacific Ocean is only weakly layered below 2 km with remarkably uniform temperature and salinity; its deep water has relatively low density; and the circulation is particularly sluggish in response to minimal contrasts in density between water masses. Water mass convergence zones analogous to those in the Atlantic are discontinuous and misplaced, perhaps due to the width of the Pacific, about twice that of the Atlantic. *Antarctic Intermediate Water (AAIW)* and *North Pacific Intermediate Water (NPIW)*, the most

common water masses above 2 km in the Pacific apparently are formed by poorly understood mixing processes.

INDIAN WATER MASSES

Most of the Indian Ocean is in the Southern Hemisphere. Cold water sinking does not occur along its northern boundary so that its deep water mass exhibits less movement than the deep water of the North Atlantic Ocean. Bottom flow is well defined in southern regions and resembles the South Atlantic south of about 40 degrees S. Deep water in the Indian Ocean consists of *Common Water (CoW)*, a mixture of Antarctic Bottom Water and North Atlantic Deep Water. *Antarctic Intermediate Water (AAIW)* moves northward from the Antarctic Convergence and flows over the Common Water.

A tongue of *Red Sea Intermediate Water (RSIW)* moves over the submarine sill at the Strait of Bab el Mandeb and into the Indian Ocean at a depth of about 3 km. RSIW is distinguished by its exceptionally high salinity, typically greater than 40 psu, produced by extreme evaporation. RSIW provides the only significant modifying effect in the entire deep Indian Ocean. Equatorial shallow layers in the Indian Ocean are not well defined partially because of shifts in surface currents that accompany seasonal shifts in the monsoon circulation. Monsoon winds blow onshore during the high-sun season (summer) and offshore in the low-sun season (winter).

The Ocean Bottom

Ocean depth varies markedly from one location to another. Over large areas water depth is less than 200 m (650 ft); other areas are as deep as 11,000 m (36,000 ft); and the average ocean depth is 3796 m (12,454 ft). Here we examine the ocean bottom in vertical cross-sectional profile, including the continental margin and ocean basin. In places the ocean bottom is almost flat and featureless and in other places the ocean bottom is volcanically active with lava reacting chemically with seawater.

CONTINENTAL MARGIN

Imagine a small boat equipped with a device that measures the vertical distance from the water surface to the ocean bottom. The boat is docked at a beach on the East Coast of the United States. Your mission is to navigate the boat due east, continuously measuring and recording water depth, until the measured depth is a little more than 3000 m (9800 ft). When your mission is complete, you bring the record of measured depths back to the laboratory for analysis.

Back at the lab, you notice that our measurements do not show a uniform increase in water depth with distance from the beach. Three distinct zones are evident. The zone closest to the beach features a very gentle slope extending out to a depth of about 200 m (655 ft). Seaward from there to a depth of about 3000 m (9800 ft) the water depth increases much more rapidly with distance. Then a relatively narrow zone is transitional from the steep slope of the previous zone to the more-or-less flat ocean basin. The initial, gently sloping zone is the **continental shelf**; the second, steeply sloping zone is the **continental slope**; and the third transition region is the **continental rise**.

The continental shelf, slope, and rise together comprise the *continental margin* (Figure 8.7). This is not a misnomer because the bedrock of the continental margin is the same as the continental plates (Chapter 1). From a geological perspective, the continents do not end at the beach, but at the continental rise. At its outer edge, the continental margin merges with the deep-sea floor or

descends into an oceanic trench.

The continental shelf is nearly flat, sloping much less than 1 degree seaward. Although water depth generally increases with distance offshore, the rate of increase is quite small—averaging about 2 m per km (10 ft per mi). Shelf width generally ranges from a few tens of meters to 1000 km (620 mi). As a general rule, the shelf is narrowest where the continental margin is tectonically active (e.g., subduction zone) and is widest where the continental margin is passive (i.e., no plate boundary activity nearby). Hence, the North American shelf is wider along the *passive* East Coast (as much as several hundred kilometers) than the *tectonically active* west coast (a few kilometers). About 7.5% of the total area of the ocean overlies the continental shelf.

By international agreement, the *shelf break* (the demarcation between the continental shelf and the continental slope) is set at a water depth of 200 m (655 ft). Most of the world fisheries are confined to the continental shelf so that the convention on where the shelf ends has important political and economic ramifications. On 10 March 1983, the United States (later joined by 50 other nations) defined its jurisdiction over ocean resources (including fisheries, minerals, and fuels) to extend 371 km (200 nautical miles) offshore. This area of the ocean is about 1.7 times larger than the nation's total land area.

The inclination of the continental slope averages about 4 degrees but ranges between 1 and 25 degrees; water depth typically increases by about 50 m per km distance from the beach (265 ft per mi), a significantly

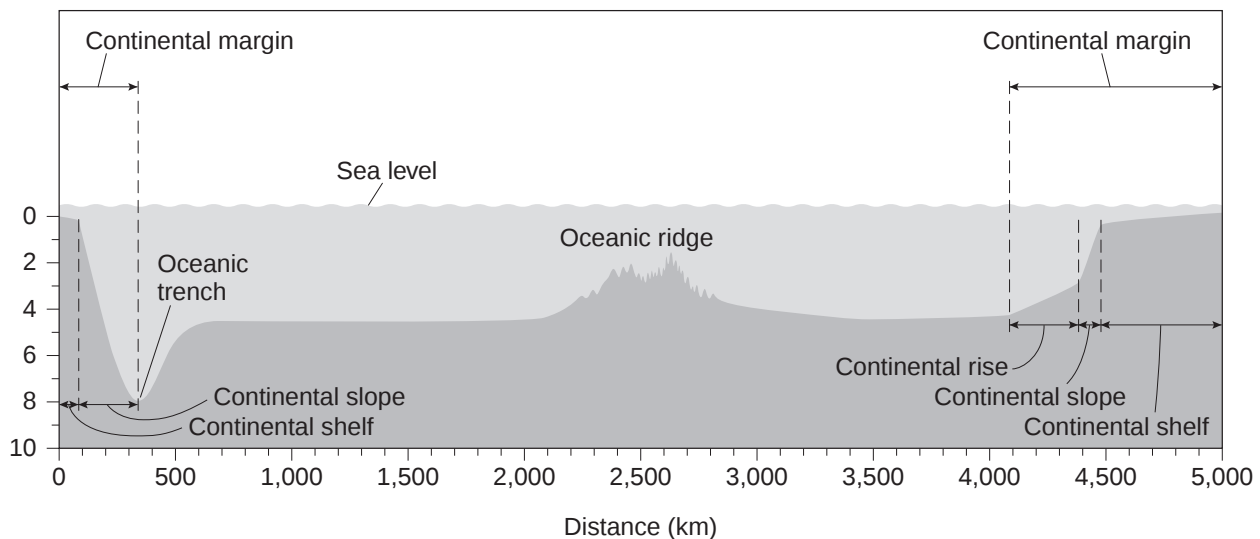


FIGURE 8.7
Cross-sectional profile of the ocean bottom with the vertical scale greatly exaggerated.

greater rate of increase than in the shelf zone. In many places such as around the margin of the Atlantic, the slope merges with the more gently sloping continental rise. In this passive continental margin, sediment spreads over the ocean floor forming vast, flat *abyssal plains* adjacent to the continental rise. In tectonically active continental margins, such as surrounding the Pacific Ocean basin, the slope descends directly into ocean trenches. Land-derived sediment flows into trenches and is not available to form either continental rise or abyssal plains. About 15% of the oceanic area is over the continental slope and rise.

Large numbers of deep, steep-sided *submarine canyons* slice into the continental slope and some run up onto the continental shelf; many of these canyons cut into solid rock. The most probable causative mechanism for most submarine canyons is erosion by turbidity currents. **Turbidity current** is a down-slope flow of water heavily laden with suspended sediment and denser than normal seawater. In many respects, a turbidity current resembles an underwater avalanche. Sediment delivered to the ocean by rivers accumulates on the ocean floor. Eventually the pile of sediments builds to such a size that it becomes unstable and moves downhill as a unit, scouring the ocean bottom in the process. Sediments accumulate at the base of the continental rise as a series of overlapping *submarine fans*. Turbidity currents have been clocked at speeds around 100 km (60 mi) per hour, so it is easy to imagine these flows gouging the ocean bottom and periodically clearing submarine canyons of sediments.

Many submarine canyons appear to be the natural extensions of existing rivers (e.g., the Hudson Canyon off the New York Bight), lending further support to the turbidity current explanation. For submarine canyons having no obvious association with existing rivers, sediment sources may be found upstream of regional longshore currents. As discussed in Chapter 9, longshore currents transport sand and other sediment along the shoreline. Also, some submarine canyons may be valleys cut by rivers during the Pleistocene Ice Age (1.7 million to 10,500 years ago). At times during the Pleistocene, sea level was as much as 130 m (425 ft) lower than today, so that much of the continental shelves was above sea level. Rivers flowed across these exposed shelves and eroded deep canyons that were subsequently flooded with the rise of sea level at the close of the Pleistocene (Chapter 7).

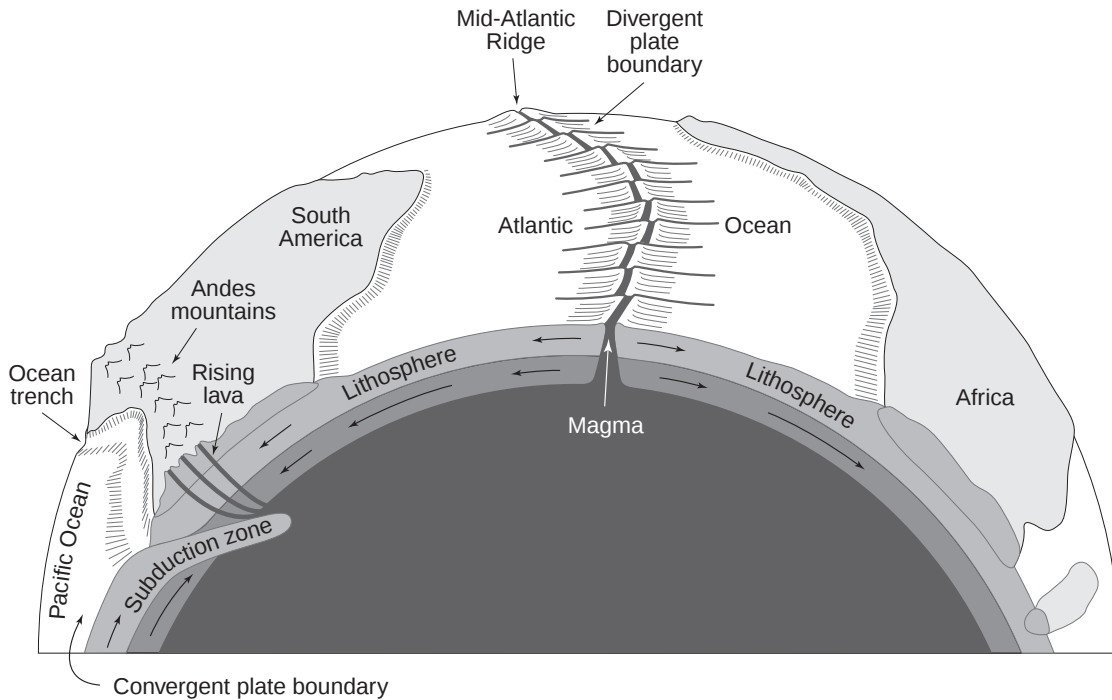
THE OCEAN BASIN

Fringed by continental margins, **ocean basins** encompass the remaining portion of the oceanic area. On average, the ocean is about 3800 m (12,500 ft) deep. Once thought to be largely flat and monotonous plains, the ocean basin actually has a varied topography featuring deep trenches, seamounts, and submarine mountain ranges. Indeed, undersea terrain is just as diverse as terrestrial terrain and exhibits even greater relief. Nevertheless, the major portion of the ocean bottom is comprised of plains and low hills, usually rising no more than about 100 m (330 ft) above the plains. Blanketed with sediments that tend to smooth out any irregularities in the bedrock below, abyssal plains and abyssal hills are typical of ocean bottom topography.

The most prominent features of the ocean basin, including trenches and ridge systems, are products of tectonic forces. Recall from Chapter 1 that Earth's lithosphere is divided into a dozen large plates some with landmasses (*continental plates*) and some under the sea (*oceanic plates*) that very slowly move over the face of the globe. Volcanic activity, earthquakes, and mountain building take place at boundaries between plates, many of which occur within or at the boundaries of the ocean basin. Depending on relative movement, plate boundaries are designated divergent, convergent, or transform.

Along a *divergent plate boundary*, adjacent plates move apart producing a rift (opening) along which magma wells up from below. Magma (or lava) cools and solidifies into new crust. The Mid-Atlantic Ridge, a submerged volcanic mountain range that divides the Atlantic Ocean into two roughly symmetrical basins, occurs along a boundary between oceanic plates that have been diverging for perhaps 200 million years (Figure 8.8). That divergence split apart the ancient supercontinent Pangaea, creating the Atlantic Ocean basin. At present, the Mid-Atlantic Ridge is spreading apart at about 2 cm per year and rifting is evident in the gaping fractures that dissect solidified lava flows in Iceland (Figure 8.9). Iceland is a volcanic island formed along the Mid-Atlantic Ridge. Other examples of divergent plate boundaries include the rifts occupied by the Red Sea and Gulf of Aden in North Africa.

The Mid-Atlantic Ridge is more than 2000 km (1240 mi) wide and rises about 2.5 km (1.5 mi) above the adjacent sea floor. It is part of a much longer ridge system that winds from the Arctic Ocean down the

**FIGURE 8.8**

Oceanic ridge systems (e.g., the Mid-Atlantic ridge) occur where oceanic plates diverge. Ocean trenches develop in subduction zones where plates converge. New crust forms at divergent plate boundaries and crust is destroyed in subduction zones.

middle of the Atlantic, curves around South Africa and into the Indian Ocean, and then into the Pacific Ocean. Its total length exceeds 65,000 km (40,000 mi), surpassing any mountain range on land.

Two plates move toward one another along a *convergent plate boundary*. A subduction zone forms where two oceanic plates converge or where an oceanic

plate converges with a continental plate (Figure 8.8). The denser oceanic plate slips under the other plate and descends into the mantle. With downward transport of mantle and crustal material, a trench forms on the ocean floor. Temperature and confining pressure increase with depth within the planet so that a plate is heated and compressed as it descends into the mantle. At depths approaching 100 km (60 mi), the plate begins to melt into magma. Whereas new crust forms at divergent plate boundaries, crust is destroyed in subduction zones at convergent plate boundaries. Magma migrates upward from a subduction zone toward Earth's surface and contributes to mountain building or volcanism.

During subduction, water that is chemically locked in hydrated minerals is released when they melt or are converted to higher density minerals. At divergent plate boundaries, water is released as steam from lava flows. In this way water cycles slowly out of the lithosphere, part of the global water cycle.

An **ocean trench** is relatively narrow, v-shaped and very deep. Depths greater than 8000 m (26,250 ft) are not uncommon and some trenches have deep holes called *deeps* that exceed 10,000 m (32,800 ft) in depth. Examples include the Japan Trench, southeast of the

**FIGURE 8.9**

This fracture in the volcanic bedrock in Iceland is an expression of the divergence of tectonic plates along the Mid Atlantic ridge. Iceland is a volcanic island formed along the ridge.

Japanese Islands in the Pacific, and the Puerto Rican Trench, east of the island of Puerto Rico in the Atlantic. The deepest place on the ocean floor is probably the Marianas Deep, in the Marianas Trench in the Pacific, with a maximum depth around 11,000 m (36,000 ft). A convergent plate boundary occurs just off the west coast of South America where the oceanic Nazca plate subducts under the continental South American plate. Plate convergence produced the Andes, a prominent mountain range that forms the backbone of South America, and the Peru-Chile Trench offshore. The Peru-Chile Trench is 5900 km (3660 mi) long, 100 km (62 mi) wide, and more than 8000 m (26,300 ft) deep.

Along a transform plate boundary, two plates slide laterally past one another. Although crust is neither created nor destroyed and these boundaries are free of volcanic activity, slippage can deform rock and trigger earthquakes. The San Andreas fault of California, site of frequent earthquakes, occurs along a transform plate boundary where the Pacific plate (carrying a piece of California) slides toward the northwest, past the North American plate.

Submarine volcanic activity is generated along divergent and convergent plate boundaries and over mantle hot spots (Chapter 1). A significant volcanic eruption, usually persisting over a long period of time, builds *seamounts*, underwater volcanic peaks rising more than 1000 m (3300 ft) above the ocean floor. Massive eruptions may cause the top of the volcano to break the water surface, producing an oceanic island. The Hawaiian Islands, Iceland, and most other oceanic islands are products of such submarine volcanic activity. With the passage of time and decline in volcanic activity, weathering and erosion dissect and greatly reduce the elevation of volcanic islands. Meanwhile, the subsurface void produced by the discharge of lava during island formation may start to collapse, causing the island to sink back into the sea. Such a seamount featuring a wave-eroded flat top is called a *guyot*.

Another example of an oceanic feature caused by volcanic activity followed by subsidence, is the coral atoll. Viewed from above, a *coral atoll* is a ring-shaped island surrounding a seawater lagoon (Figure 8.10). While serving as a naturalist on the British research ship H.M.S. *Beagle*, Charles Darwin first proposed that atolls were produced by coral growing around the edge of the island at the same rate that the island sank into the sea. Modern drilling of atolls finds coral remains continuously down to the underlying volcanic rock.

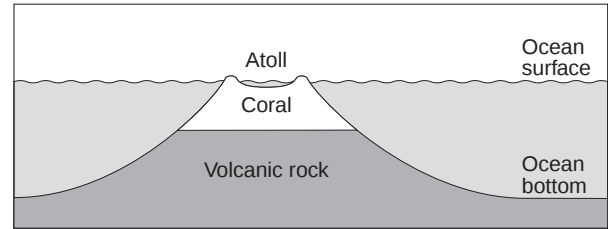


FIGURE 8.10
Vertical cross-section of a coral atoll.

Coral live in warm near-surface waters, so these drilling data indicate that the volcanic material was once at the surface, just as Darwin had speculated.

At divergent boundaries or hot spots, lava wells up from Earth's mantle and comes in contact with seawater. Lava cools and rapidly solidifies into new oceanic crust. Congealing lava contracts forming fractures through which seawater circulates and in the process, seawater is heated and dissolves minerals, altering its composition. This is another source of the dissolved constituents of seawater, in addition to the runoff from rivers. Submarine volcanic activity also releases gases into seawater. Sites where hot seawater exits fractures in the newly formed ocean crust are called **hydrothermal vents**. A few regions of extremely hot water have been known to exist for many years, but it was not until the Woods Hole submersible *Alvin* descended into the Galapagos Rift in 1979 that scientists observed hydrothermal vents directly. They observed chimneylike vents discharging plumes of hot, black water.

Communities of organisms, including bacteria, crabs, mussels, starfish, and tubeworms live near hydrothermal vents. In most marine communities, photosynthesizing organisms form the base of food chains. In the deep water where hydrothermal vents occur, no sunlight is available for photosynthesis, and the base of the food chain consists of bacteria that practice *chemosynthesis*. That is, they oxidize sulfur compounds from the hot vent waters and thereby provide nutrients for themselves and other members of marine food chains.

Astronomical Tides

Astronomical tides are the regular rise and fall of the ocean surface caused by the gravitational attraction of the moon and sun. Tides can be thought of as planetary-

scale waves propagating across the ocean (Chapter 9), with high tide being the crest of the wave and low tide being the trough of the wave. At sea the wave amplitude is well under a meter but the wave amplifies as it progresses into shallower coastal waters. Tides are measured at coastal locations as local diurnal changes in sea level. The difference in height between water levels at high and low tide is called the **tidal range** and the time between successive high tides is the **tidal period**.

Tides are important because of their impact on local navigation, moorings, coastal structures, legal boundaries, fisheries, and recreation. Today, considerable scientific research focuses on the global nature of tides, including their influence on other physical processes in the ocean such as circulation, mixing, and wave generation. Furthermore, oceanic and atmospheric scientists are interested in how storm-driven waves and storm surges combine with tides to affect the potential for coastal flooding and the rate of coastal erosion (Chapter 9). This section focuses on some fundamentals of ocean tides and their prediction.

ASTRONOMICAL CONTROLS

The gravitational attraction between Earth and the moon and sun depends on the masses of these objects and the distance between them. The greater the mass, the greater is the attraction whereas the greater the distance, the smaller is the attraction. Although its mass is considerably smaller than that of the sun, the moon has a greater influence on the tides because it is so much closer to Earth than is the sun.

Earth and moon revolve around a common center of mass once each lunar month. This turning motion, combined with the moon's gravitational attraction, causes an egg-shaped bulging of Earth's ocean surface. One ocean bulge faces toward the moon and the other is on the opposite side of the Earth, facing away from the moon (Figure 8.11). A similar interaction between Earth and sun produces ocean bulges on Earth that line up towards and away from the sun. On a theoretical Earth with no friction, ocean bulges would always align with the celestial body responsible for their formation. A place on the coast that is moved by Earth's rotation through the bulges would experience rising and falling sea level (tides). On a hypothetical Earth completely covered by water, each day a low-latitude location would experience two high tides (when under

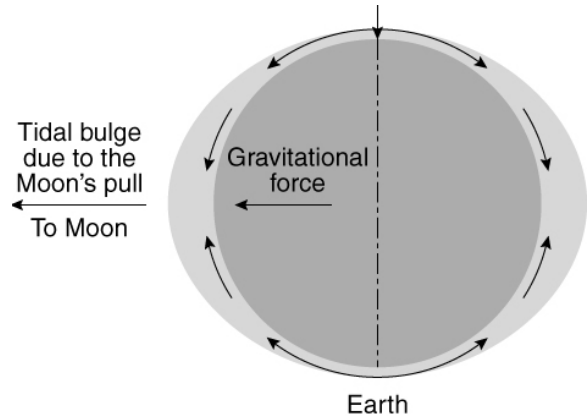


FIGURE 8.11

Two tidal bulges produced by the gravitational attraction of the moon.

bulges) and two low tides (when halfway between bulges).

The moon's monthly revolution about the Earth is in the same direction as Earth's daily rotation. A point on the Earth must make a full rotation and more (with about 24 hours and 50 minutes elapsing) to catch up with the advancing moon. The moon-based day is also called the *tidal day*. Ocean tidal bulges produced by the moon remain in the same alignment relative to the moon, but change their latitudinal positions on Earth from day to day as they follow the moon during its monthly revolution about the Earth. The plane of the moon's orbit is inclined to Earth's equatorial plane so that during one lunar month, the moon's latitudinal position moves from directly over the equator northward to within 5 degrees of the Tropic of Cancer (23.5 degrees N), back to the equator, on southward to within 5 degrees of the Tropic of Capricorn (23.5 degrees S), and then back to the equator where another cycle begins.

The sun produces ocean tidal effects that are about 45% of those related to the moon, due to the much greater distance between sun and Earth. Sun-related ocean tidal bulges are produced the same way as those resulting from Earth-moon interactions. The gravitational attraction between Earth and sun plus Earth's annual revolution around the mutual center of mass of sun and Earth produce a second set of similar but smaller tidal bulges aligned with the sun. The mid-points of sun-related tidal bulges follow the sun just as moon-related bulges track the moon. Their latitudinal positions go through a yearly cycle.

TYPES OF TIDES

When the moon is directly over the equator, its associated tidal bulges are centered on the equator. In theory, all coastal locations would rotate through the two tidal bulges and experience two equal high tides and two equal low tides per tidal day. This is known as a *semidiurnal tide*. Different tidal patterns exist when the moon, and the midpoints of its associated tidal bulges, are either north or south of the equator. While semidiurnal tides are observed at the equator at all times, most locations north or south of the equator experience two unequal high tides and two unequal low tides per tidal day. This is known as a *mixed tide* and the difference in heights between successive high (or low) tides is called the *diurnal inequality*. When the moon is above the Tropic of Cancer or Capricorn, the diurnal inequality is maximum and the tides are called *tropic tides*. When the moon is above or nearly above the equator, the diurnal inequality is minimum and the tides are known as *equatorial tides*. When the moon and its associated tidal bulges are either north or south of the equator, most points at high latitudes would in theory experience one high tide and one low tide per tidal day. This is known as a *diurnal tide*.

The separate sets of ocean bulges related to the moon and sun act at times together and at times in opposition. Twice monthly the positions of the sun, moon, and Earth are along a straight line (Figure 8.12). At these times of new and full moon phases, as viewed from Earth, the lunar- and solar-related ocean bulges also line up to produce tides having the greatest monthly tidal range. These are called *spring tides*. Twice monthly, at the first and third quarter phases of the moon, the sun pulls on the Earth along a line that is at right angles to the pull of the moon (Figure 8.13). At this time, tides have the least monthly range. These are called *neap tides*. Furthermore, the moon orbits Earth in an ellipse (rather than a circle) so that the moon is closest to Earth (stronger tide-generating force) at *perigee* and farthest

FIGURE 8.12

Configuration of Earth, moon, and sun responsible for spring tides.

FIGURE 8.13

Configuration of Earth, moon, and sun responsible for neap tides.

from Earth (weaker tide-generating force) at *apogee*. The moon completes one perigee-apogee-perigee cycle in 25.5 days.

Like ocean waves, tides are modified by many non-astronomical factors including friction, water depth, bottom topography, coastline configuration, and winds. Shallow basins with just the right length may have a natural period of oscillation that matches the period of the tide-generating force. This so-called *resonance* explains the extraordinary tidal range (about 15 m or 50 ft) observed in the Bay of Fundy, Canada. Non-astronomical factors help explain why many locations on the U.S. Atlantic coast have predominantly semidiurnal tides whereas many places on the Gulf Coast have predominantly diurnal tides, and many localities on the Pacific Coast have mostly mixed tides (Figure 8.14).

TIDE PREDICTION

Periods of motions of the Earth, sun, and moon in space (i.e., orbits, rotation) are fixed and known precisely. The predictability of the tide-generating celestial bodies means that astronomical tides can also be predicted with great accuracy.

Treated as a wave, the local tide can be resolved mathematically into various components, called *partial tides*. Partial tides are forecasted individually and added together to predict the height and timing of the future local tide. Although as few as four partial tides can account for 70% of the total tidal range, some 60 components are commonly used. More than 100 components must be considered to predict the tides along a complex irregular coast such as that of Alaska. In order to predict local tides, tide gauging stations must gather data for a minimum of 18.6 years to experience most of the astronomical configurations of the Earth-

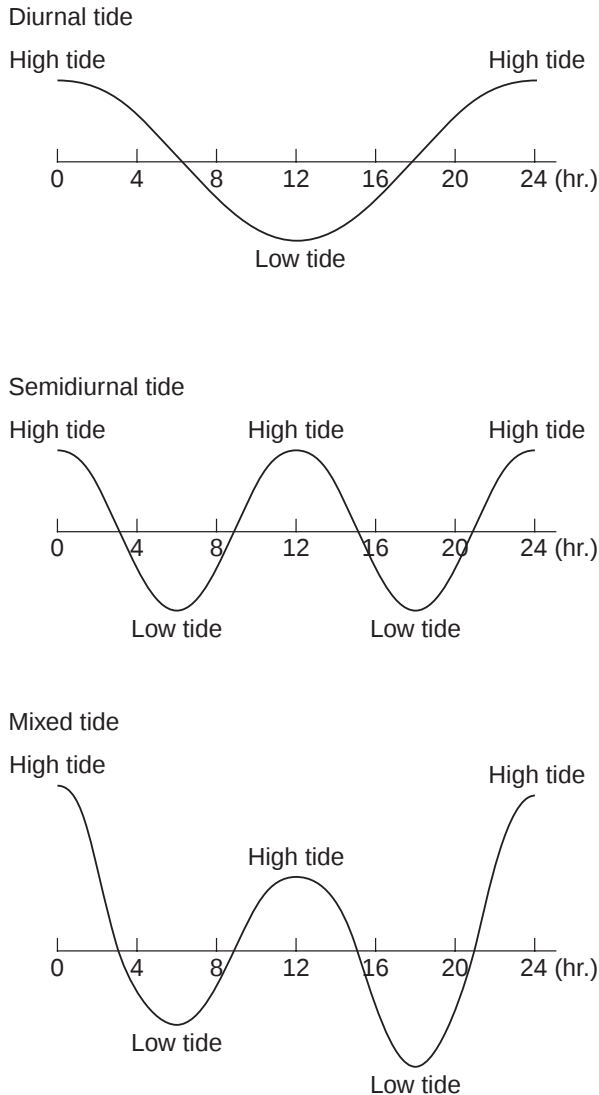


FIGURE 8.14
Types of astronomical tides observed in coastal locations.

moon-sun system generating the tides. However, only six months of tide gauging data provide an adequate basis for very reasonable tide predictions. Local winds or atmospheric pressure variations may explain differences between the actual and predicted tide.

Conclusions

Density is one of the most important variables used to describe the state of the ocean system. Variations in seawater density are primarily determined by temperature and, to a lesser extent, salinity. Processes

that take place at the ocean/atmosphere interface, in turn, ultimately govern seawater temperature and salinity.

Density contrasts of water masses spur thermohaline circulation that involves about 90% of the ocean's volume. Denser water masses slowly sink under and push up less dense water masses. Surface water is also set in motion in the form of currents and waves by the action of wind on water. This interaction between the ocean and atmosphere is a major focus of the next chapter.

Basic Understandings

- The ocean consists of saltwater that occupies huge topographic depressions known as ocean basins. The ocean is by far the largest reservoir in the global water cycle and a major reservoir in the other biogeochemical cycles operating on the planet.
- The basic properties of ocean water (e.g., salinity, temperature, density) largely depend on the interactions of the ocean with the other components of the Earth system, including the atmosphere and lithosphere. Energy and material are continually exchanged between the ocean and its surroundings.
- The principle of constant proportions states that the major conservative constituents of seawater occur in the same relative concentrations throughout the ocean system. Constituents of seawater that participate in biogeochemical or seasonal cycles have variable concentrations and are described as *non-conservative*.
- The salinity of seawater is measured as the ratio of the electrical conductivity of the water sample to the conductivity of a standard seawater sample. As a ratio, salinity has no units but is commonly expressed as practical salinity units (psu). One psu essentially equals 1 part per thousand (ppt) of dissolved solids.
- Processes operating at the atmosphere/ocean interface that add or remove water molecules from seawater largely control salinity. Salinity increases through evaporation and ice formation; salinity decreases through precipitation, runoff from rivers, and melting ice.
- The flux of gases into seawater is related to the amount of gas already in solution and the temperature, the single most important control of the saturation level. The saturation of a gas in water

decreases with rising temperature. The dissolved oxygen supply of surface ocean waters is replenished directly by the atmosphere and as a byproduct of photosynthesis by marine plants.

- A column of water having a height of 10.33 m (33.9 ft) exerts a pressure equal to one standard atmosphere. To a good first approximation, the pressure at any point in a water column is directly related to depth and the relationship is linear; that is, doubling the depth of water doubles the pressure. Water pressure expressed in *decibars* is numerically the same as water depth expressed in *meters*.
- For many oceanic processes, seawater is considered incompressible; that is, water density does not vary significantly with depth or pressure. Water density does vary significantly with temperature and salinity. As salinity increases, the temperature of maximum density and the freezing temperature of water decline but at different rates. The salinity of seawater is greater than 24.7 psu and the salinity of brackish water is less than 24.7 psu. At a salinity of 24.7, the temperature of maximum density and the freezing temperature are equivalent. Seawater density increases with decreasing temperature and/or increasing salinity. Temperature is usually much more important than salinity in explaining observed changes in seawater density.
- Processes responsible for changes in temperature and salinity primarily affect the surface layer of the ocean, that is, where the ocean interfaces with the atmosphere.
- Ocean temperature is largely governed by solar radiation absorbed in the surface layer of the ocean and heat transfer at the interface between the ocean and atmosphere. In general, seawater temperature declines with increasing latitude and increasing depth.
- The interaction of solar radiation with ocean water governs the color of seawater, marine productivity, and ocean temperature. On a global annual average, the ocean absorbs about 92% of incident solar radiation with almost two-thirds of visible light absorbed within the upper one meter. Selective absorption of visible light by wavelength explains the color of ocean water. Based on the diminishing availability of light with depth, two layers are recognized in the ocean water column: the upper photic zone (100 m or less in thickness) in which sunlight is adequate for photosynthesis, and aphotic

zone (below the photic zone) where light intensity is insufficient for plants to survive.

- Based on the vertical profile of average water temperature, most of the ocean is stratified into the mixed layer, main thermocline, and deep layer.
- Density contrasts that arise from differences in temperature and salinity generate subsurface movement of water masses, a phenomenon known as thermohaline circulation. A water mass is a large, homogeneous volume of water featuring a characteristic range of temperature and salinity.
- The continental margin, transitional between land and sea, consists of the continental shelf, continental slope, and continental rise. Together, these zones underlie about 22.5% of the ocean's total area.
- Oceanic ridge systems (e.g., mid-Atlantic ridge) occur along divergent plate boundaries and deep oceanic trenches develop in subduction zones along convergent plate boundaries.
- Submarine volcanic activity is generated along divergent and convergent plate boundaries and over mantle hot spots. Seamounts, guyots, and islands are among the products of submarine volcanism.
- Astronomical tides are the regular rise and fall of the ocean surface caused by the gravitational attraction of the moon and sun. Although its mass is considerably smaller than that of the sun, the moon has a greater influence on the tides because it is much closer to Earth.
- Tides are classified as semidiurnal (two equal high and low tides per tidal day), mixed (two unequal high and low tides), and diurnal (one high tide and one low tide per tidal day).
- Treated as a wave, the tide at a particular location can be resolved into component partial tides. Partial tides are predicted individually and added to forecast the local tide.

ESSAY: Probing the Ocean Floor

Many different instruments and techniques are available to examine the configuration and composition of the ocean bottom. Some are direct and simple, such as the devices that take physical samples of the ocean bottom, whereas others, such as acoustic and gravimetric instruments, are more complicated.

Sampling the ocean bottom is straightforward. If a sample of bottom material is desired, a grab sampler is usually employed. This mechanical device operates in much the same manner as a human hand when it grabs a handful of dirt. If, on the other hand, scientists require a cross-section of the ocean bottom showing sediment layering, a weighted coring tube is lowered to within a few meters of the bottom and then allowed to fall with the weight of the system driving it into the bottom sediments. The sediment core, retrieved when the tube is brought to the surface, is a time record of local sedimentation, and it is even possible to date the various layers. Where the ocean bottom is bare rock, deep-sea drilling devices are available that obtain rock samples much like sediment cores.

A major effort currently underway in deep-sea drilling is the *Ocean Drilling Program (ODP)*, an international partnership of scientists and research institutions organized to explore the evolution and structure of Earth through study of deep ocean cores. ODP is sponsored by the National Science Foundation and agencies in 20 other participating countries under the management of *JOIDES (Joint Oceanographic Institutions for Deep Earth Sampling)*. The drill ship JOIDES (pronounced joy-deez) Resolution, is the centerpiece of the Ocean Drilling Program. With this ship, ODP scientists can obtain sediment and rock cores in water depths as great as 8200 m (27,000 ft). Built in 1978 in Halifax, Nova Scotia, the ship was originally a conventional oil-drilling vessel. She was refitted in 1984 and is now equipped with some of the world's finest shipboard laboratories. The ship's complement for each cruise is a mix of 30 scientists from around the world, 20 engineers and technicians, and a crew (including drilling personnel) of 52.

Each year the JOIDES Resolution embarks on six scientific expeditions, each approximately two months in length. The research ship has drilled in all of the world's oceans, including north of the Arctic and south of the Antarctic circles. Since January 1985, ODP has recovered cores that end-to-end total more than 160 km (99 mi) in length. After a core is brought to the surface, it is marked with its original location on the seafloor, coded to distinguish top from bottom, measured, and cut into smaller sections for study and storage. Each section of the core is sliced lengthwise. One half is used for nondestructive analysis before being stored. Scientists seeking to reconstruct a chapter in Earth's history sample the other half. No aspect of the core is overlooked. Paleontologists examine microfossils in the cores to determine the age of the material. Other scientists measure physical properties such as density, strength, and ability to conduct heat. Paleomagnetists use state-of-the-art equipment to read the record of Earth's magnetic field changes, information that helps determine when and where specific rocks were formed. At the end of an expedition, cores are stored at one of four repositories. ODP scientists are able to access these repositories much as the general public uses a library.

In conjunction with the Ocean Drilling Program, other methodologies are available for study of the composition, thickness, and nature of the oceanic crust. One of these techniques uses a sound source, such as an underwater explosion, coupled with a number of strategically positioned acoustic receiving devices. Knowing that sound travels at different speeds through different materials, and different acoustic components of the initial sound travel at different speeds through the same material, much can be learned about the thickness, extent, and composition of crustal layers. This information is then combined with gravimetric data, usually obtained using a gravimeter that measures the acceleration of gravity. From knowledge of the local gravitational acceleration, information about local crustal mass, and thus composition, may be inferred.

Probably the oldest type of measurement involving the ocean bottom is determination of depth. Classically this was done with a sounding line, a weighted rope marked off in fathoms (where 1.0 fathom = 6 ft). The line was thrown over the side, allowing the weight to reach the bottom. Today, an acoustic sounder (sometimes called a fathometer) is commonly used, where a sound pulse is directed toward the bottom and travel time to the bottom and back of an acoustic pulse is recorded. Knowing the speed of sound in seawater, the depth can be calculated.

Depth data are needed for chart preparation, but acquiring these data by acoustic sounder can be time consuming because ship speeds only average about 30 km per hour (15 knots). An alternative is a system called laser bathymetry. A downward facing laser gun is mounted in the floor of an aircraft so that light pulses can be aimed at the ocean bottom while the aircraft is in flight. Travel time for a light pulse to reach the bottom and return is then measured by computer. An aircraft can travel nearly ten times faster than a ship, so that close to 1000 square km (386 square mi) can be surveyed in a single day. The only limitation on this system is how far the laser light can penetrate into a given water column, limiting the technique to relatively clean and shallow waters. For depths beyond the shelf break, ocean scientists depend on acoustic sounding.