CHAPTER 5

The Circulation of the Oceans

Key Questions

- Why do ocean currents form?
- How can the circulations of both the surface ocean and the deep ocean basins be driven by solar radiation and be closely linked, yet operate at very different time scales?
- What role does ocean circulation play in the global climate system?

Chapter Overview

We continue our discussion of Earth's circulatory subsystems by describing the processes that drive the circulation of the world's oceans. The movement and circulation of the oceans is tied very closely to the circulation of the atmosphere: Both are ultimately driven by the distribution of available solar energy, and their motions are linked by friction at the sea surface. In Chapter 4, we described an imbalance in the latitudinal distribution of energy that produces an equator-to-pole temperature gradient at the surface—the driving force for the pattern of Earth's surface wind. These wind patterns are responsible for the circulation of the ocean surface and the formation of the world's major ocean currents. As with the atmosphere, once the ocean starts to move, it comes under the influence of the Coriolis effect, which plays a significant role in the resulting circulation patterns. The oceans are vertically stratified, with denser water at the bottoms of the major ocean basins and less-dense water near the surface. The density is controlled by the temperature and by the salt content (*salinity*) of the water. The deep-ocean water is separated from the surface layer of the ocean by a transition zone with sharply defined density, temperature, and salinity gradients. This deep-ocean water moves as a response to small changes in density that occur over wide areas, and the movement is largely independent of the surface-ocean circulation. Together, however, both types of ocean circulation contribute to the redistribution of available energy in the Earth system, albeit over very different time scales. And both play a major role in the distribution of nutrient supplies in the oceans.

WINDS AND SURFACE CURRENTS

Chapter 4 showed that the circulation in the troposphere is caused by atmospheric pressure gradients that result from vertical or horizontal temperature differences. We saw that from a global perspective, these temperature variations are caused by latitudinal differences in solar heating. But ocean surfaces are also heated by incoming solar radiation. Do the oceans, therefore, circulate for the same reason as the atmosphere? The answer is no, because the solar heating of the ocean takes place at the upper surface of the fluid, whereas the solar heating of the atmosphere occurs largely at the lower surface of the fluid near Earth's surface. Solar heating results in warmer water at the surface of most of the world's oceans. But the Sun's rays warm only the top few hundred meters of the ocean; 90% of the radiation that penetrates the surface is absorbed in the first 100 m. The warmer water is less dense than the cooler water below, which is not affected by the surface heating. This situation is inherently stable, so there is very little vertical movement. It is similar to the situation in the stratosphere. Recall from Chapter 3 that the atmosphere at this level is stable because the maximum solar heating occurs high in the stratosphere, the site of peak absorption of ultraviolet radiation by ozone. Where temperature increases with height, there is no density imbalance, and convection cannot take place. The fluid—water or air—remains well stratified. The true situation in the ocean is actually more complicated than this, as we will see, because the density of seawater is also affected by its salt content. It remains true, however, that the ocean overturns very slowly.

At the same time, temperature changes in the ocean occur slowly. Remember from Chapter 4 that the oceans have a high heat capacity—it takes a considerable amount of heat to produce just small changes in temperature. Slight differences in incoming solar radiation from place to place thus have little impact on the surface temperature of the ocean, so lateral temperature and density differences are slight over large areas. Unlike the troposphere, therefore, the surface ocean does not circulate as a *direct* response to the surface heating. Instead, surface temperature plays a more indirect role: The surface temperature influences the atmospheric circulation, and the resulting pattern of global winds determines the circulation of the upper ocean.

The movement of the wind over the ocean causes friction at the surface. As a result of friction, the wind drags the ocean surface with it as it blows, thus setting up a pattern of surface-ocean *wind-drift* currents. The force of the wind acting on the surface is referred to as *wind stress*. The water movement is usually confined to the top 50 to 100 m of the ocean, although well-developed currents such as the Gulf Stream in the North Atlantic and the Kuroshio Current in the North Pacific may extend as much as 1–2 km below the surface. The Coriolis effect influences ocean currents just as it does winds, so the water is deflected to the right of the path of the wind in the Northern Hemisphere (and to the left of the wind's path in the Southern Hemisphere). Observations show that this deflection tends to be approximately 20–25° from the wind direction. Thus, as a first approximation of what the ocean circulation should look like, we can take the surface winds of Figure 4-11 and add a large ocean bounded by continents on the east and west. In doing so we obtain the surface-ocean circulation pattern shown in Figure 5-1. The trade winds produce westward-flowing currents in the tropics. When these currents reach the western continental boundary, they are deflected northward and southward. They then come under the influence of the westerlies, which cause the currents to flow eastward in the midlatitudes. When these currents reach the eastern landmass, some water is deflected to the pole and some toward the equator. The waters that flow toward the poles are replaced by equatorward flow along the western landmass. The waters that flow toward the equator come back

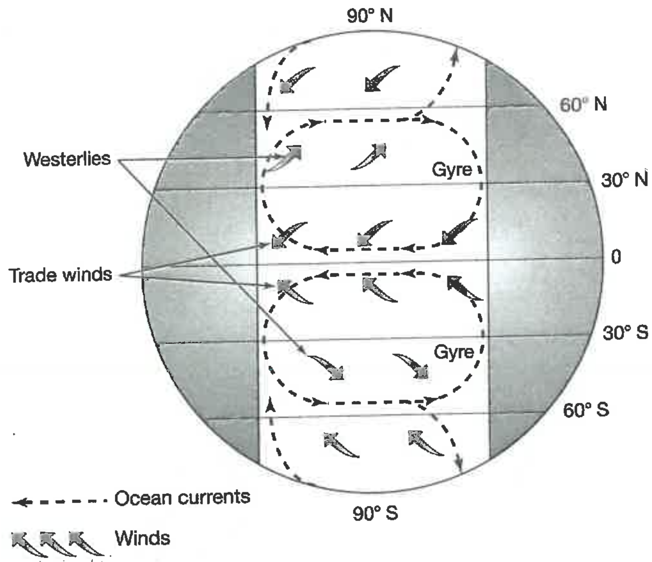


FIGURE 5-1 A simplified view of the surface-ocean circulation.

under the influence of the trade winds and are blown westward again. The currents complete a large, circular circulation pattern (called a **gyre**) in the subtropical oceans. The circulation of these gyres is clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere.

Compare the simplified model in Figure 5-1 with the observed distribution of ocean currents in Figure 5-2. The same general circulation features are apparent. Figure 5-2 shows counterclockwise gyres in the Southern Hemisphere and clockwise gyres in the Northern Hemisphere. Figure 5-2 identifies the world's major ocean currents and designates them as "warm" or "cool," labels that we will explain later in this chapter. The pattern in the real world is more complicated because the distribution of land and water is not as simple as it is in Figure 5-1. In the Southern Hemisphere, the westerlies result in an eastward-flowing current—the *West Wind Drift*—that extends around the globe because, in the real world, there is very little land in the middle and higher latitudes to deflect the water back toward the west.

Convergence

There are further differences between Figure 5-1 and the actual circulation of the ocean surface that are not apparent from the diagram. To begin with, although our predicted gyres are present, the explanation of their occurrence is a little more complicated than we previously suggested. To explain why they form, first we need to describe a few more processes that take place at the ocean surface. If we consider the circulation pattern shown in Figure 5-1, we might expect that water would pile up as it reached the coasts. In the Northern Hemisphere, therefore, we would expect to find water piling up in the northeast and southwest portions of

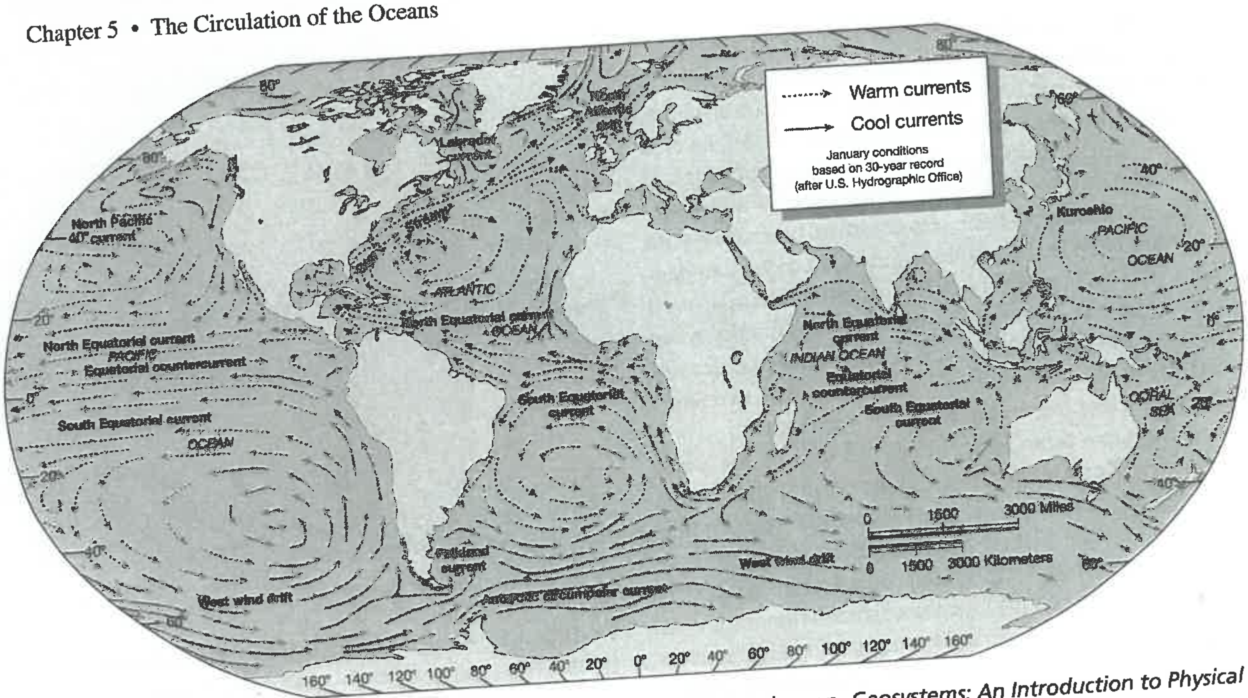


FIGURE 5-2 The major surface-ocean currents. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography,* 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

the gyre. This does not happen; rather, water piles up (or converges) in the *middle* of the gyre. This *convergence* results from the combined effects of the wind-driven surface-ocean currents, Earth's rotation, and, ultimately, friction.

The Norwegian explorer Fidtjof Nansen made a key observation that led to a better understanding of convergence during an expedition across the Arctic Ocean in the 1890s. His ship, the *Fram,* was frozen into the ice at the beginning of winter and drifted with the ice for over a year. It had long been thought that the surface-ocean currents were produced by the winds. Among many observations made during the expedition, however, Nansen noted that the ice (and the ship) did not drift *with* the wind but at 20–40° to the right of the surface wind path.

Walfrid Ekman, a Swedish physicist, first made the connection between wind-driven currents and Earth's rotation and derived a mathematical explanation of Nansen's observations. Due to friction between wind and the water surface, some of the kinetic energy of the air is transferred to the top layer of the water. As that layer moves, it drags along the water just below it, which in turn drags along the water just below that, and so on. The water appears to move as many thin, coupled layers, and kinetic energy is transferred down the water column. However, as the energy is transferred downward, friction causes some of the energy to be dissipated in the form of heat, so each level moves more slowly than the level above. At some depth below the surface, the effects of the wind-induced movement disappear. However, as each layer moves, it is again subject to the Coriolis effect. Once a layer starts to move, the water is deflected to the right of the path of the layer above (or the wind path, for the surface layer) in the Northern Hemisphere and to the left in the Southern Hemisphere. The deeper below the surface, the farther each layer is deflected to the right or left of the surface layer, producing a spiraling effect known as the **Ekman spiral** (Figure 5-3a).

Ekman's theory predicts, under strong and persistent winds in the open ocean, (1) that the surface current will flow at 45° to the surface-wind path, (2) that the flow will be reversed at approximately 100 m below the surface (that is, the current at 100 m will flow in a direction opposite to the surface current), and (3) that it will also be considerably reduced in speed. In practice, there are few observations of a well-developed Ekman spiral, but observations do show that the surface flow is to the right of the surface-wind path (although usually at an angle less than 45°). The observations also bear out a further prediction from the theory—that when the movements of all the individual layers of water in the spiral are added, the net direction of transport within the water column is at a right angle (90°) to the wind direction. This net movement of water is referred to as **Ekman transport.** In a clockwise gyre in the Northern Hemisphere, the effect of Ekman transport is to push water into the center of the gyre (Figure 5-3b). Note that the counterclockwise gyres in the Southern Hemisphere will produce exactly the same result, because the Coriolis effect deflects the water to the left.

Divergence

Just as there are parts of the ocean where convergence occurs, there are also parts of the ocean where divergence occurs. In the equatorial Atlantic of the Northern

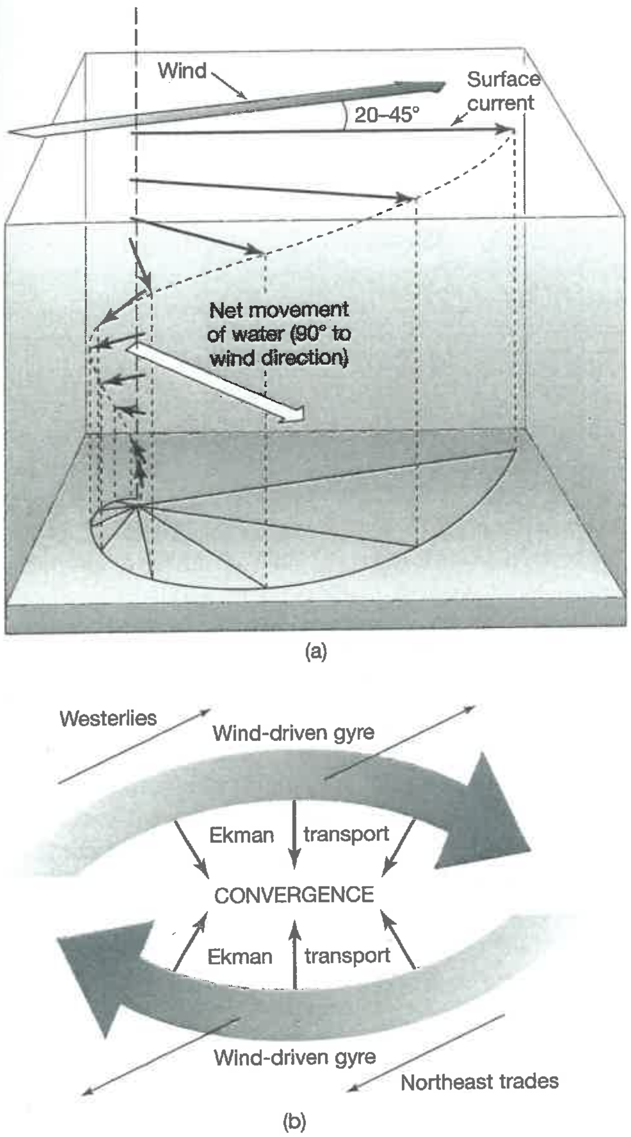


FIGURE 5-3 (a) The Ekman spiral. (b) Convergence in the center of a subtropical gyre due to Ekman transport.

Hemisphere, for example, the northeast trades (which, remember, blow from the northeast toward the southwest) result in a westward-flowing surface current, the North Equatorial Current. The net Ekman transport is 90° to the right of the wind, which means that the bulk of water transport is directed almost due north. Conversely, the southeast trades (in the Southern Hemisphere) produce the westward-flowing South Equatorial Current, and the net Ekman transport is to the left of the wind flow, toward the south. Hence, divergence occurs near the equator. Both divergence and convergence can also be found along coastlines where the Ekman transport may push water toward or away from the coast, depending on the direction of wind movement and the surface current. Important areas of divergence occur off the southwest coast of North America and the west coast of North Africa due to the easterly winds and southward-moving currents in these regions. Similar areas of divergence are found off the west coasts of South America and southern Africa, where northward-moving currents have the same effect.

Upwelling and Downwelling

In areas of convergence, the surface water piles up, the sea surface rises, and the surface layer of water thickens (Figure 5-4a). In areas of divergence, the surface water moves away, the sea surface drops, and the surface layer thins. Where convergence occurs, the accumulation of water causes it to sink in a process known as **downwelling.** Conversely, where divergence occurs at the surface, water must rise from below to replace it (Figure 5-4b). Water at depth is cooler than water at the surface. The rising of cooler water to the surface to replace warm, divergent surface water is referred to as **upwelling.** As we will see later in this chapter, these deeper waters also tend to be rich in nutrients. Upwelling, therefore, brings these nutrient-rich waters to the surface.

Geostrophic Flow

Having explained Ekman transport and convergence, we now come to the real reason why the circulation in the subtropical oceans takes the form of very distinct oceanic gyres. Areas of convergence and divergence produce slight variations in sea-surface elevation across the ocean basins, so the sea surface actually slopes from one point to another. This difference in elevation is very slight—on the order of a few meters over 10^2 to 10^5 km (that is, slopes of 1 in 10^5 to 1 in 10^8). Yet these slight elevation gradients are sufficient to cause a downslope force on the water due to gravity. If we consider the subtropical ocean in the Northern Hemisphere, for example, we have already seen that the northeast trade winds produce a westward-flowing ocean current near the equator, whereas the prevailing westerly winds in the midlatitudes result in an eastward-flowing current. The circulation is completed by the deflection of water along the coastlines at the ocean margins. Ekman transport in the surface layers causes convergence and the pile-up of water in the middle of the ocean (Figures 5-3b and 5-5a).

The sea surface is only about 50 cm higher in the center of the gyre than at the edges, but gravity acting on this pile of water results in a force (referred to as the *pressure-gradient force*) that pushes outward, down the gradient, from the center. As the water flows, however, it is deflected by the Coriolis effect until that effect balances the pressure-gradient force acting down the slope. The result of the two forces acting in opposition is to cause a flow of water off to the side—to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (Figure 5-5b). Thus we end up with a circular flow of water around the gyre that is approximately parallel to the ocean slope (Figure 5-5c). Note the similarity to the description of the geostrophic wind in Chapter 4. In this case, the resulting current is

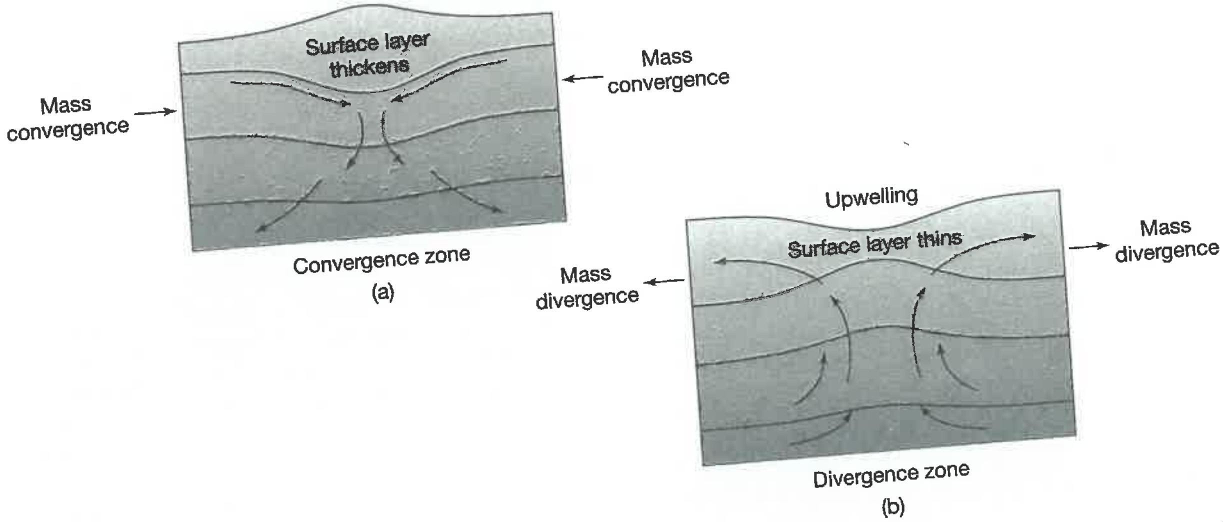


FIGURE 5-4 Schematic representation of zones of convergence and divergence. (a) Surface water accumulates in convergence zones, increasing the surface elevation (very exaggerated in the diagram) and thickening the surface layer. (b) The opposite happens in divergence zones—there is a decrease in surface elevation, and the surface layer thins.

called a **geostrophic current,** which flows around the gyre clockwise in the Northern Hemisphere (and counterclockwise in the Southern Hemisphere), in the same direction as the original wind-driven flow. In practice, the flow is a little less than 90° to the slope, so in fact the water tends to spiral inward as it moves around the gyre and the convergence in the gyre results in downwelling.

Boundary Currents

Ocean gyres are a prominent feature of the surface circulation. Figure 5-2, however, gives the impression that the flow around these gyres is not symmetric. Indeed, the flow in the western part of the gyre is confined to a narrow path with a fast-flowing current (a *western boundary current*),

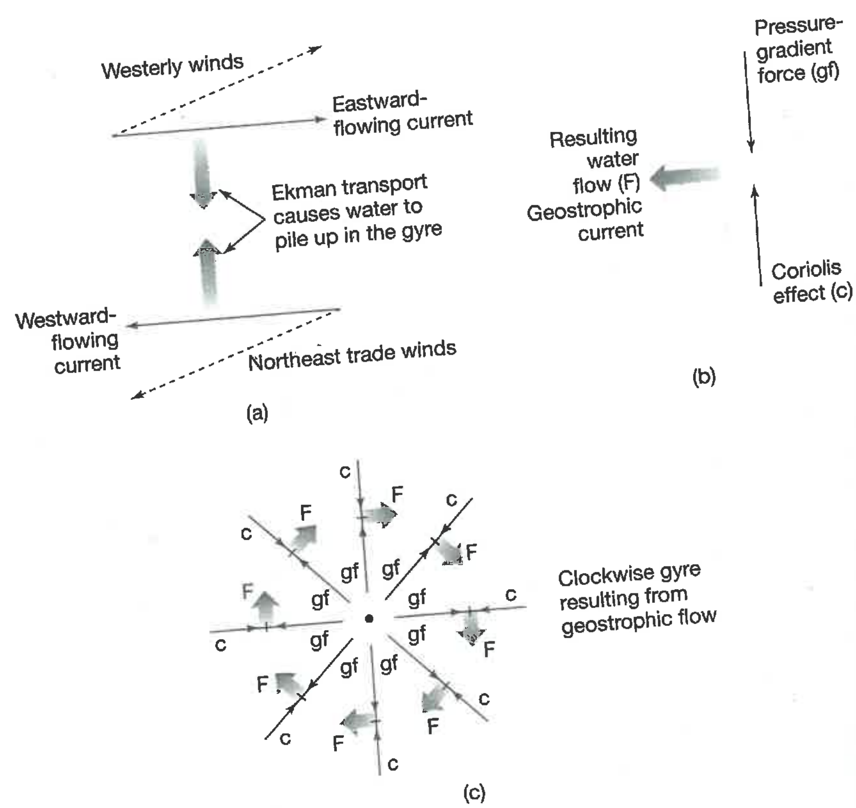


FIGURE 5-5 (a) The subtropical gyres are formed by geostrophic currents that occur when Ekman transport from the wind-driven currents causes water to pile up in the center of the gyre. (b) There is a force due to gravity, acting down the gradient of the surface slope, that is opposed by the Coriolis effect. The net effect is a flow of water at approximately 90° to the slope. (c) The result is a geostrophic current that flows approximately perpendicular to the slope of the sea surface around the gyre.

which in the east is more diffuse, spread over a much larger area and with much-reduced current speeds (an *eastern boundary current*). Eastern boundary currents also tend to be divergent; the Ekman transport is away from the continent, thinning the surface layer along the coastline. The thinner surface layer and the divergent flow promote upwelling in these regions.

The most-studied western boundary current is the Gulf Stream in the western North Atlantic. The Gulf Stream begins as a narrow (50–75 km wide), fast-flowing stream of warm water (20°C or higher) in the Florida Current, flowing northward between Bermuda and Cuba. The current can reach depths of more than a kilometer, with surface speeds between 3 and 10 km/hr. This current follows the coast northeastward to Cape Hatteras, North Carolina, where it continues across the North Atlantic as the Gulf Stream (see Figure 5-2). Moving northeastward across the Atlantic, the Gulf Stream decreases in speed and the flow broadens into the North Atlantic Drift, which eventually flows into the Arctic Basin north of Norway. On reaching Europe, the North Atlantic Drift splits north and south; the southward-flowing component becomes an eastern boundary current, the Canary Current. The Canary Current moves much more slowly than the Gulf Stream, is shallower (reaching depths of only about 500 m), and is much broader—up to 1000 km across. As the water flows back toward the tropics, it comes under the influence of the northeast trade winds that push it to the west in the North Equatorial Current to complete the gyre. This elliptical flow of water essentially isolates the area in the center of the gyre, the Sargasso Sea. The Sargasso Sea is named for the extensive cover of seaweed often found in this area. Low current speeds and light, variable winds made this region difficult to traverse back in the days of sailing ships. The ancient mariners were also afraid of being entangled by the huge mats of seaweed that covered the surface.

Our discussion of wind-driven currents illustrates how wind stress, the Coriolis effect, and the pressure-gradient force serve to produce convergence, geostrophic flow, and gyres in the subtropical oceans. However, our discussion still does not account for the asymmetric nature of the gyres and the very different modes of flow in the eastern and western boundary currents. This pattern is caused by dynamic forces that operate when fluids tend to move in a rotary motion. How this happens is explained in the Box "A Closer Look: Vorticity."

Ocean Circulation and Sea-Surface Temperatures

The large-scale surface-wind pattern produces gyres in the surface layer of the midlatitude oceans. In the Northern and Southern hemispheres, the Coriolis effect and Ekman transport cause a net movement of water into the center of the gyres. The higher surface elevation in the center of the gyres causes a geostrophic current to flow around the gyres in the same direction as the wind-driven flow, thus reinforcing the surface circulation. The shape of the gyre and the nature of the eastern and western boundary currents are then determined by the need to balance the forces that produce a tendency for water to rotate differently in different parts of the gyre.

The resulting circulation pattern has a significant impact on the redistribution of energy around the globe and on regional temperature (see Figure 5-2). The equatorial currents are warmed by the large input of solar radiation at low latitudes. When these currents are deflected poleward, they carry warmer water to middle and high latitudes. However, these currents lose heat as they travel poleward. When they are deflected northward and southward by the eastern landmass, the water moving poleward is warmer than the polar ocean, whereas the water moving equatorward is colder than the tropical ocean. At the same time, the surface water that originates in the polar oceans and moves equatorward is also colder than the midlatitude oceans. Ocean currents thereby aid in the latitudinal redistribution of energy: They move warmer water toward the poles and cooler water toward the equator.

Consider the warm North Atlantic Drift as an example. This northwest-flowing current brings fairly warm waters to northern Europe (see Figure 5-2). Due to the predominantly westerly flow of air in the midlatitude troposphere, most of northern Europe is warmed by the waters from the west and ultimately from the south. Contrast the seasonal variability and much milder conditions in southern Scandinavia with the more extreme conditions on the Labrador coast at the same latitude (see Figure 4-18). Southern Scandinavia benefits from the warmth of the North Atlantic Drift. The air passing over Labrador, however, comes from the cold interior parts of Canada, and a cold offshore current (the Labrador Current) brings sea ice down to this area in winter. The result is lower temperatures and a much greater seasonal temperature range in that area. Notice that we now also have the explanation for the cold offshore currents, such as the Benguela and Humboldt currents, that are responsible for the Namib and Atacama deserts on the west coasts of South Africa and South America, respectively.

We can see that the oceans play a significant role in determining the broad patterns of Earth's present-day climate. It is also very important in controlling much of the variability we expereience in climate from year to year. Recall from Chapter 4 that the atmosphere transports heat very rapidly, and any anomalies are quickly dissipated. The atmosphere has very little "memory" of change. In contrast, the oceans absorb and store large amounts of heat, and they release this heat very slowly. Consequently, the ocean's memory of any change is much longer than the atmosphere's. Transient anomalies that develop in sea-surface temperatures can be expected to have a lingering impact on climate for some time afterward. The oceans, therefore, are the most likely place to look for processes that might cause climate anomalies on the interannual or decadal time scales.

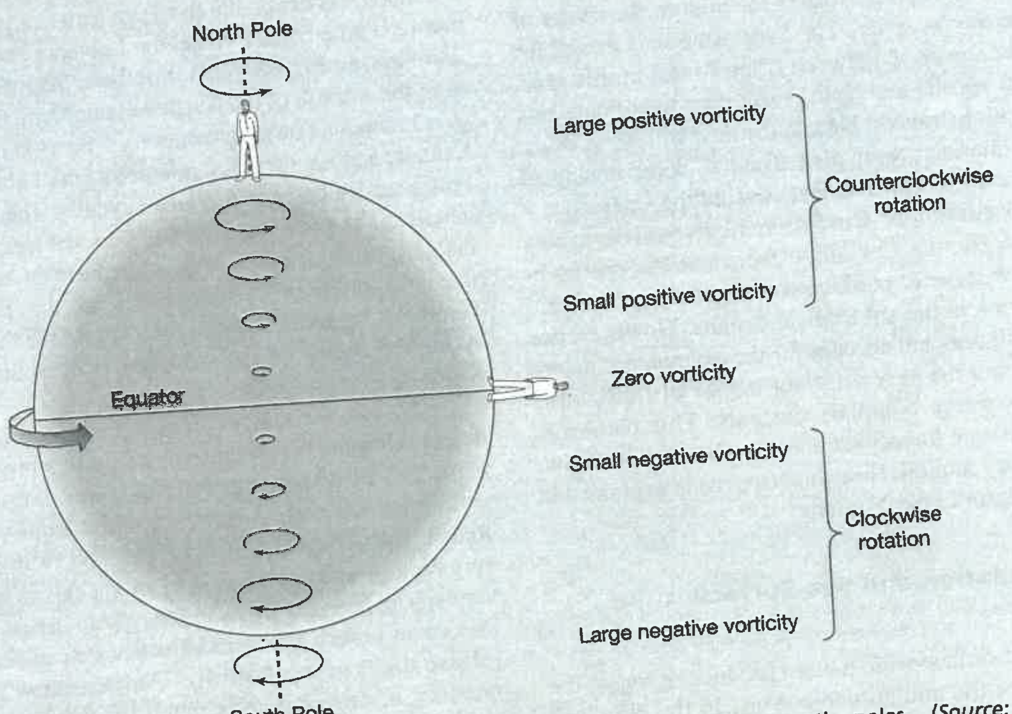
A CLOSER LOOK

Vorticity

The text explains why large-scale gyres form in the subtropical oceans. But to explain the asymmetric pattern of the gyres and the differences between eastern and western boundary currents, we need to introduce an additional concept—that of vorticity. **Vorticity** describes the tendency of a fluid to rotate. A tendency to rotate in a counterclockwise direction is referred to as positive vorticity, whereas a tendency for clockwise rotation is negative vorticity. We refer to vorticity as being the *tendency* to rotate (rather than the actual rotary motion of the fluid) because different forces could impose both a positive and a negative vorticity on the same mass of water at the same time, and the actual amount and direction of the rotation would then depend on the net effect of the different forces.

So what produces vorticity? Imagine yourself standing exactly on the North Pole. You would be spinning around a vertical axis at the rate of one rotation every 24 hours (actually, 23 hours and 56 minutes). In other words, you would be experiencing a counterclockwise rotation about your vertical axis. At the equator you would experience no angular rotation, because you would be standing at exactly 90° to Earth's axis of rotation (Box Figure 5-1). Anywhere between the equator and the pole you would experience some fraction of the pole's angular rotation. This angular rotation about a vertical axis at Earth's surface, brought about because of Earth's rotation, produces vorticity that is referred to as **planetary vorticity.** Mathematically, the planetary vorticity is identical to the Coriolis effect. Like the Coriolis effect, planetary vorticity acts in the opposite direction in the Southern Hemisphere.

A tendency for rotary motion can be created by a number of factors other than planetary rotation. Surface waters being driven by cyclonic or anticyclonic circulations in the atmosphere (i.e., low-pressure or high-pressure systems) will produce positive or negative vorticity. Similarly, *current shear,* in which the speed of the current changes across the current, will also produce vorticity. Representing the speed of the current by the length of the arrows in Box Figure 5-2a, for example, would produce negative (clockwise) vorticity; the faster-moving water tends to curl in toward the slower part of the current. The current shown in Box Figure 5-2b, conversely, would produce positive (counterclockwise) vorticity. Current shear can be quite dramatic where friction with a coastal boundary slows the edge of currents that flow parallel to the coastline. The vorticity produced by the pattern of surface winds and by current shear is referred to as **relative vorticity.** The **absolute vorticity** experienced by a body of water is then simply the sum of the planetary and relative vorticities.



BOX FIGURE 5-1 Planetary vorticity, increasing from zero at the equator to a maximum at the poles. (*Source:* Open University, *Ocean Circulation*, New York: Pegamon Press, 1989.)

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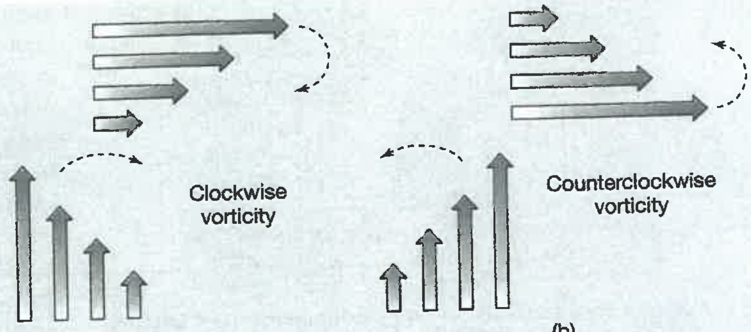


BOX FIGURE 5-2 Schematic diagram of current shear producing negative (clockwise) and positive (counterclockwise) relative vorticity. The lengths of the arrows represent relative speeds in the current.

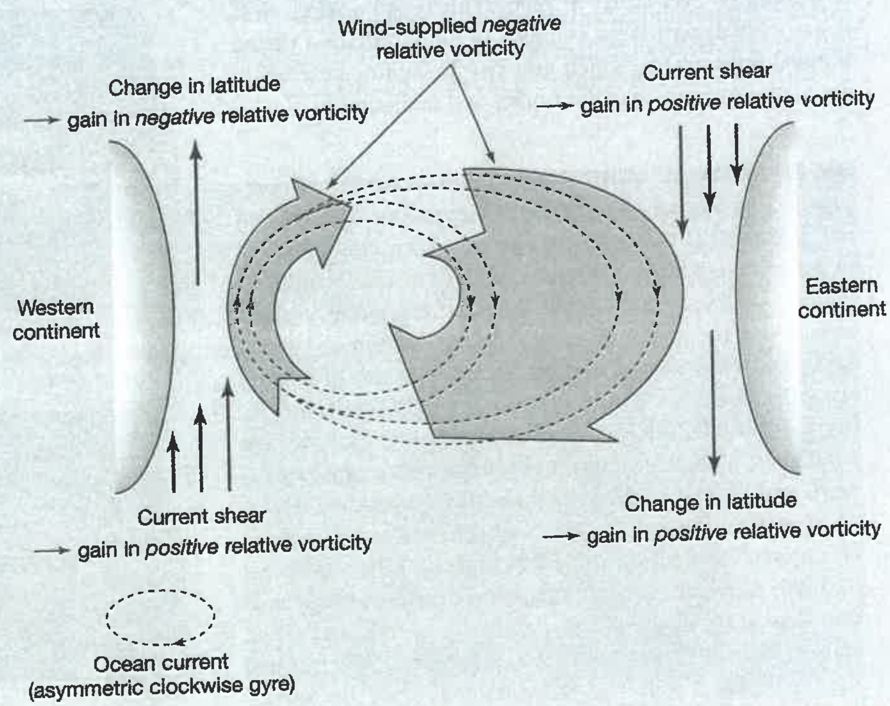
How does this help us explain the difference between western and eastern boundary currents? The answer lies in the fact that, like mass and energy, absolute vorticity is a conserved property. In Box Figure 5-3, we can see the various factors that contribute to vorticity around a gyre. To begin with, the anticyclonic surface-wind pattern produces *negative relative vorticity* all around the gyre. At the eastern boundary current, this wind-supplied negative relative vorticity is balanced:

- The negative relative vorticity is balanced in part by a small increase in positive relative vorticity due to friction (current shear) along the coast. This factor is not large; because of the width of the current, only a small portion interacts with the coastline.
- Positive relative vorticity also arises from the southward flow of the water. The water is moving into an area of lower positive planetary vorticity; for vorticity to be conserved, there has to be a decrease in negative relative vorticity (or, looked at another way, a gain in positive relative vorticity).

The situation is very different on the western boundary:

- Current shear again produces positive relative vorticity.
- The change in latitude, however, leads to a gain in *negative relative vorticity* (the water is moving into regions of large positive planetary vorticity and so must acquire negative relative vorticity). This gain in negative relative vorticity reinforces rather than offsets the negative relative vorticity imparted by the surface-wind pattern.

The only way balance can be achieved in the western boundary current is by increasing the friction along the boundary, which is achieved with a narrower, deeper, and faster-flowing current. Hence the asymmetric nature of the gyre.



BOX FIGURE 5-3 Contributions to the changes in relative vorticity around an asymmetric subtropical gyre. (*Source:* Open University, *Ocean Circulation*, New York: Pergamon Press, 1989.)

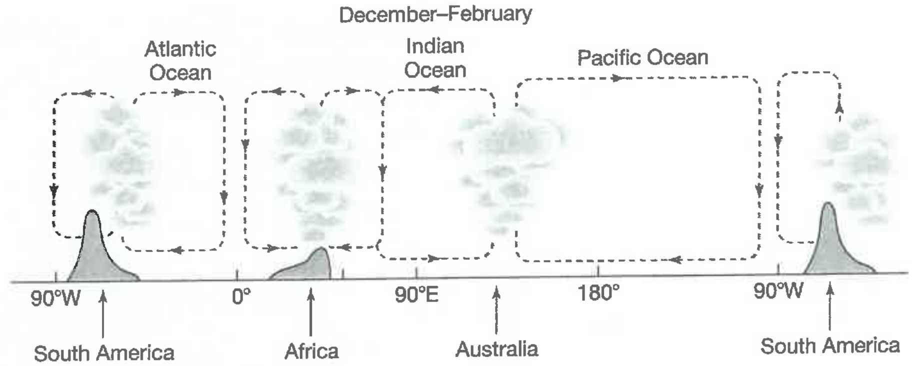


FIGURE 5-6 The east–west circulation in the equatorial troposphere. The shaded areas represent heavy precipitation.

So, where do we look in the oceans? One obvious place to start is in the tropics, where we find the large convective towers of the Hadley cells that drive the atmospheric circulation over a large portion of the planet. We will also look at the opposite extreme, and examine the role of the polar oceans in short-term climate variability in Chapter 6.

El Niño–Southern Oscillation (ENSO) Events

The name **El Niño** was originally given to a warm ocean current that appeared off the coasts of Peru and Ecuador. The current flowed for only a few weeks, and, because it usually occurred near Christmastime, local fishermen named it El Niño after the Christ child. The name has taken on a different meaning more recently; it is now used by researchers to describe a major shift in the oceanic circulation that occurs in this region every 2 to 10 years. This broad oceanic shift is associated with large changes in the circulation of the tropical atmosphere, which give rise to significant climate anomalies over much of the tropics and midlatitudes.

THE EQUATORIAL ATMOSPHERIC CIRCULATION Superimposed on the north–south Hadley circulation (described in Chapter 4) is a significant east–west circulation in the troposphere that is most prevalent over the equatorial Pacific. The western equatorial Pacific has the highest sea-surface temperatures on the globe. This region, which encompasses Australia and Indonesia, is a site of intense atmospheric convection. As is true of Hadley cells, the rising air diverges at high altitudes, but, in this case, we are concerned with a component of the flow that moves eastward and westward along the equator rather than northward and southward (Figure 5-6). The eastward-moving air crosses the Pacific, where it subsides off the west coast of South America. The circulation is completed by an easterly flow at the surface. This circulation is linked to other, smaller cells driven by convection over South America and Africa. Figure 5-6 shows the normal pattern of the equatorial east–west circulation. Figure 5-6 also indicates the normal pattern of precipitation, with heavy precipitation in the convective regions and drier conditions in the areas of subsidence. The circulation cells produce an oscillation in the sea-level pressure distribution between the western and the central/eastern portions of the tropical Pacific Ocean. When pressures are low in the west, they tend to be higher in the east, and vice versa. This oscillation in sea-level pressures is referred to as the **Southern Oscillation (SO)**.

THE OCEAN CIRCULATION The persistent easterly wind at the surface in the Pacific Ocean produces a westward-flowing ocean current, which results in the water piling up in the western part of the ocean. This causes very warm water to accumulate in the western Pacific. Figure 5-7 shows that water piles up in the west, which causes the ocean surface to slope downward from west to east. The

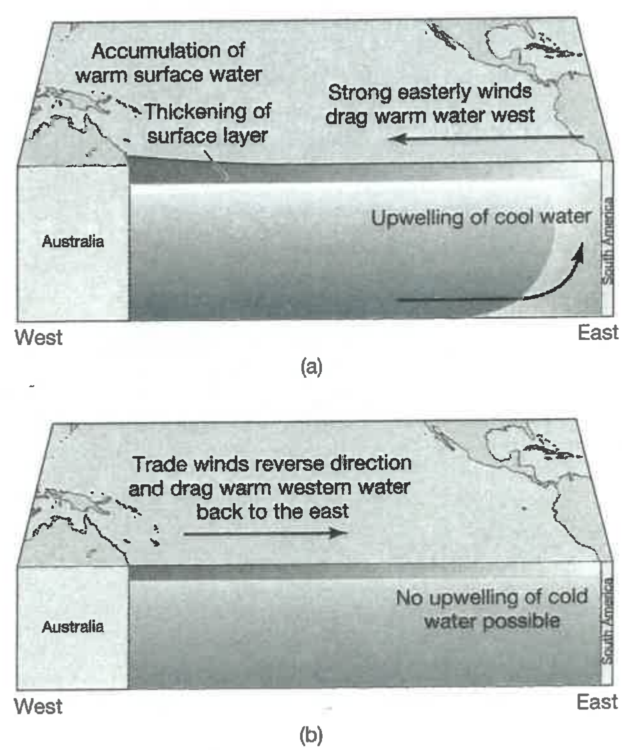


FIGURE 5-7 The ocean surface layer in the tropical Pacific Ocean.

slope is exaggerated in the diagram; the difference in surface elevation from west to east is only on the order of a couple of meters. This east-to-west movement of water thickens the warm surface layer in the west and thins it in the east. The thinner surface layer in the east allows the upwelling of colder, nutrient-rich water from below, which promotes high levels of biological productivity and large fish populations.

We can regard this pattern of atmospheric and oceanic circulation as the norm, but in any year we see substantial differences. In some years this pattern intensifies: The sea-surface temperatures in the central and eastern Pacific are colder than normal, and convection over Indonesia is enhanced. These conditions are referred to as **La Niña** conditions—the pattern is similar to the "normal" conditions, but the circulation is enhanced. More drastic changes occur when the pattern breaks down in what is referred to as an **El Niño–Southern Oscillation (ENSO)** event. The Southern Oscillation may also be referred to as being in a "cold" phase during La Niña (or *anti-ENSO*) events and in a "warm" phase during ENSO events.

Before we describe what happens when an ENSO event occurs, let us look at this circulation again. The atmospheric convection in the western Pacific occurs as a result of the high sea-surface temperatures, but the high sea-surface temperatures are a result of the atmospheric circulation, which is driven by the convection. In other words, it is a classic chicken-and-egg situation. It is not a case of the ocean forcing the atmosphere or vice versa. Instead, it is a single integrated system with a positive feedback loop (Figure 5-8). If we perturb the system at any point, we should expect to see changes throughout all of its components.

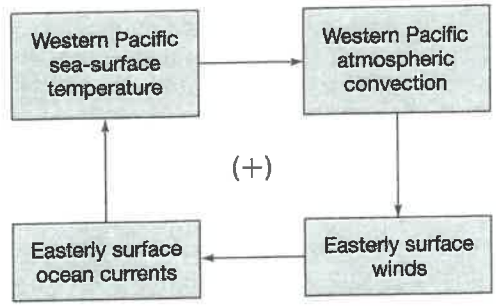


FIGURE 5-8 Schematic diagram of the interaction between the atmosphere and ocean in the tropical Pacific Ocean.

ENSO EVENTS Scientists are still debating what causes ENSO events to occur. For the purposes of this discussion we will simply break into the cycle in Figure 5-8 and ask what happens if, for some reason, there is a decline in the strength of the easterly winds.

If these winds weaken or reverse direction, which happens in some ENSO events, there is nothing to restrain the pile-up of warm water that has accumulated in the western Pacific. This water then comes sloshing back across the ocean in what is known as a *Kelvin wave.* It takes about 60 days for this wave to travel back across the Pacific. When it does, it has two major consequences. First, it shifts the pool of high sea-surface temperatures from the western to the central Pacific (Figure 5-9), which then completely changes the atmospheric circulation. Second, it shuts off the upwelling in the eastern Pacific, which has drastic consequences for biological productivity. The loss of the nutrient-rich water leads to a massive die-back of marine organisms and the bird life that feeds on them.

The changes in the atmospheric circulation are shown in Figure 5-10. The greatest area of convective activity during ENSO events lies over the central Pacific. The rising air

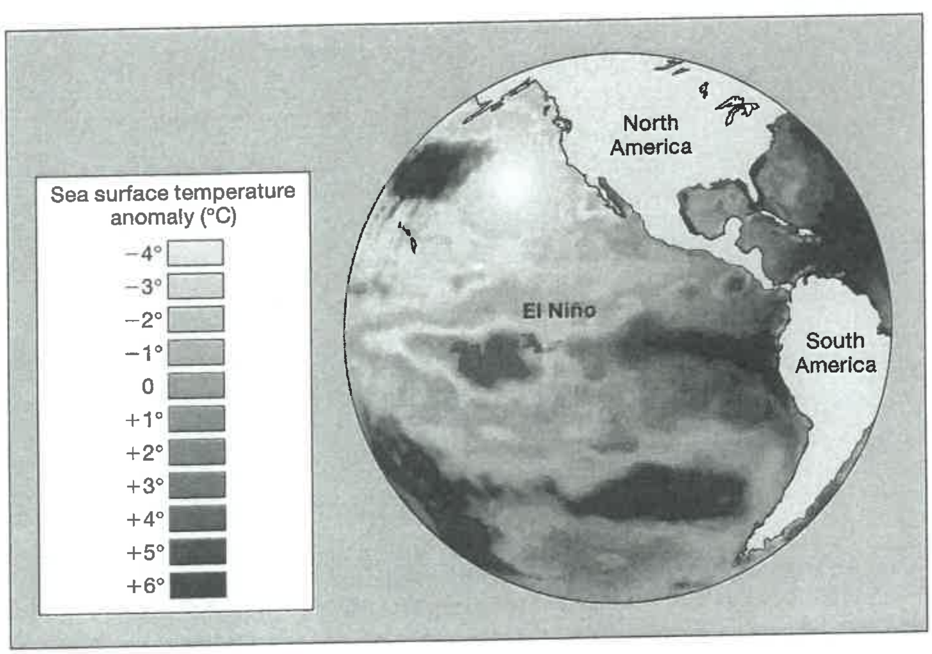


FIGURE 5-9 [See color section] Sea-surface temperature anomalies in the central and eastern Pacific during the 1997–1998 ENSO event.

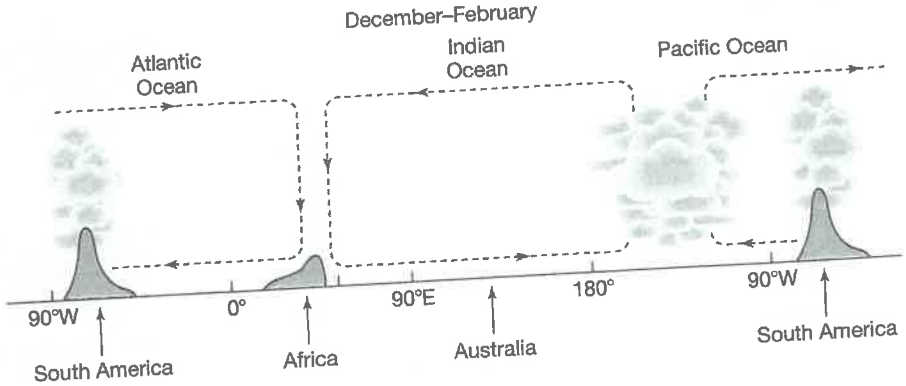


FIGURE 5-10 Atmospheric circulation during an ENSO event.

diverges to the east and west, meeting and subsiding over Africa, although there is also localized uplift on the western side of the Andes. In a non-ENSO year, there is low pressure (rising air) at the surface over Australia and Indonesia and high pressure (subsiding air) at the surface in the central and eastern Pacific (Figure 5-6). In an ENSO year, this pattern reverses: Pressure increases over Australia and decreases in the central Pacific. We can calculate the pressure difference between these two locations (Figure 5-11a) and plot this difference through time to produce the **Southern Oscillation Index (SOI)** (Figure 5-11b). This index is a measure of the pressure difference between the western and eastern parts of the tropical Pacific Ocean. Strong positive values indicate La Niña (non-ENSO) conditions; strong negative SOI values indicate ENSO conditions, which have occurred throughout the interval for which recorded data are available. We also find evidence from ice cores that ENSO events of various magnitudes

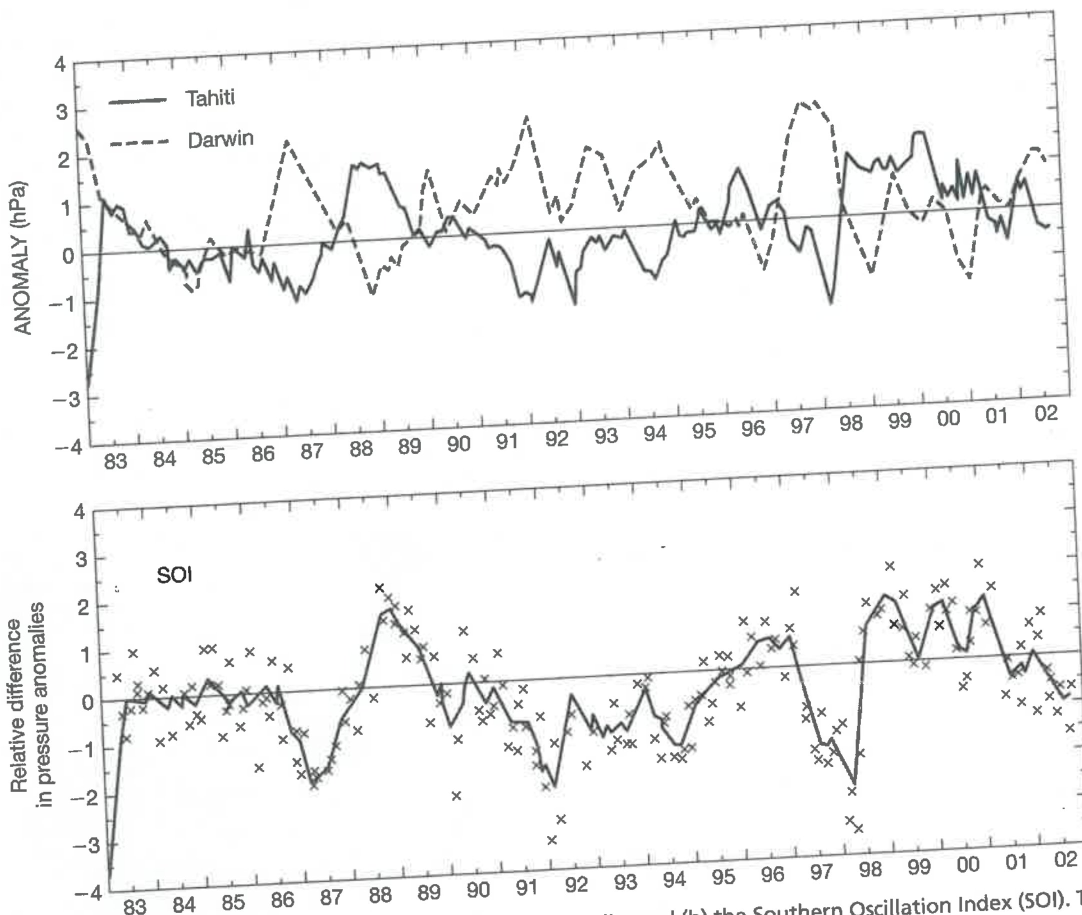


FIGURE 5-11 (a) The sea-level pressures at Tahiti and Darwin, Australia, and (b) the Southern Oscillation Index (SOI). The SOI is computed from the sea-level pressure differences. Negative values of the SOI indicate warm (El Niño) events. Note the strengths of the 1982–1983 and 1997–1998 events. (*Source:* NOAA Climate Prediction Center http://www.cpc.ncep.noaa.gov/data/indices/).

have been a feature of the climate system for at least the past 500 years.

CLIMATIC IMPACTS OF ENSO The most dramatic impacts of ENSO events are seen in their effects on rainfall patterns. In non-ENSO years, summertime convection and rainfall occur over Australia and Indonesia. There is also convective activity and rainfall over equatorial Africa and the Amazon Basin. In contrast, there is subsiding air (dry conditions) west of the Andes. However, in ENSO years the dominant convective region shifts toward the central Pacific, and convection and rainfall over Australia and Indonesia diminish. Figure 5-10 shows that there is also subsidence over Africa and some localized convection over the western Andes. The result is drought in central America, Brazil, Australia, Indonesia, and southeast Africa and anomalously high rainfall amounts in the central Pacific and on the western slopes of the Andes in Ecuador and Peru. These high rainfall amounts typically result in floods and landslides, with their accompanying high levels of soil erosion. ENSO events also appear to have some effect on the monsoon circulation over India, resulting in increased rain over southern India and reduced rainfall over northern India and the Himalayas.

The impact of ENSO events is not confined to the tropics (see the Box "A Closer Look: The 1982–1983 and 1997–1998 ENSO Events"). The changes in atmospheric circulation can also influence the midlatitudes. We saw in Chapter 4 that the location and strength of midlatitude weather systems are controlled in part by the subtropical highs and in part by the sea-surface temperature gradients. Because both of these factors change during ENSO events, it is not surprising that they should have some effect on midlatitude climate. The general pattern of temperature and rainfall anomalies associated with an ENSO event are shown in Figure 5-12 for the Northern Hemisphere winter. In practice, each ENSO event tends to be somewhat different from others in the record. For example, the strength and location of the sea-surface temperature anomalies are not

A CLOSER LOOK

The 1982–1983 and 1997–1998 ENSO Events

The 1982–1983 ENSO event was one of the most severe of the 20th century. It developed unexpectedly. The normal early warning sign of changing tropical sea-surface temperatures was missed because of the preceding eruption of a Mexican volcano (El Chichón). The volcanic aerosols reduced the outgoing infrared radiation, resulting in lower satellite-derived sea-surface temperature measurements. The resulting pattern of climate anomalies was also slightly different from a "normal" ENSO: They were much more intense than earlier ENSO events. The impact of this particular event was considerable, with major droughts and floods occurring throughout the tropics. It is estimated that climate-related catastrophes resulting from the 1982–1983 ENSO event left more than 1,000 people dead and caused almost $9 billion worth of damage:

- Ecuador and northern Peru experienced floods and landslides that left 600 dead and resulted in crop and property losses totaling approximately $400 million. Guayaquil, Ecuador, had 20 times its normal rainfall in May 1983.
- In Indonesia, there were crop failures and starvation.
- In Botswana, the ENSO-induced drought followed two previous years of drought and eventually led to the loss of thousands of livestock.
- Eastern Australia suffered the worst drought of the century. Animal feed supplies were so diminished that thousands of sheep, dying of starvation, had to be shot. The dry conditions resulted in huge dust storms, one of which deposited 11,000 tons of topsoil on the city of Melbourne. February 1983 also saw some of the worst bushfires in Australian history that caused extensive damage and killed 75 people.
- Tahiti and French Polynesia had last experienced a typhoon at the beginning of the 20th century. The warm-water pool that formed in the central Pacific during the 1982–1983 ENSO event generated several large storms, and the islands were hit by six typhoons in five months.
- In the United States, increased rain in the Midwest resulted in the flooding of the Mississippi River. An increase in storms on the West Coast resulted in severe flooding and landslides in California, where 10,000 houses were lost or damaged and farm losses totaled half a billion dollars. There was a record snowfall in the Rockies, which, when it melted, resulted in flooding in Salt Lake City and along the lower Colorado River.

The 1997–1998 event was equally severe. Anomalous weather patterns occurred in many parts of the world, including extensive impacts in the equatorial Pacific, North and South America, and East Africa:

- Flooding, mud slides, and disease killed more than 80 people in Peru, and flooding in Ecuador resulted in 90 deaths and the evacuation of 22,000 people.
- The central Pacific had eight tropical cyclones, but only two the year before.
- Drought conditions produced hunger in Indonesia and Papua New Guinea. The extreme drought in Indonesia resulted in forest fires that burned over 1 million acres of forest.
- In the United States, this event produced heavy rains, flooding, and mud slides along the California coast but a very mild winter in the northeastern region.

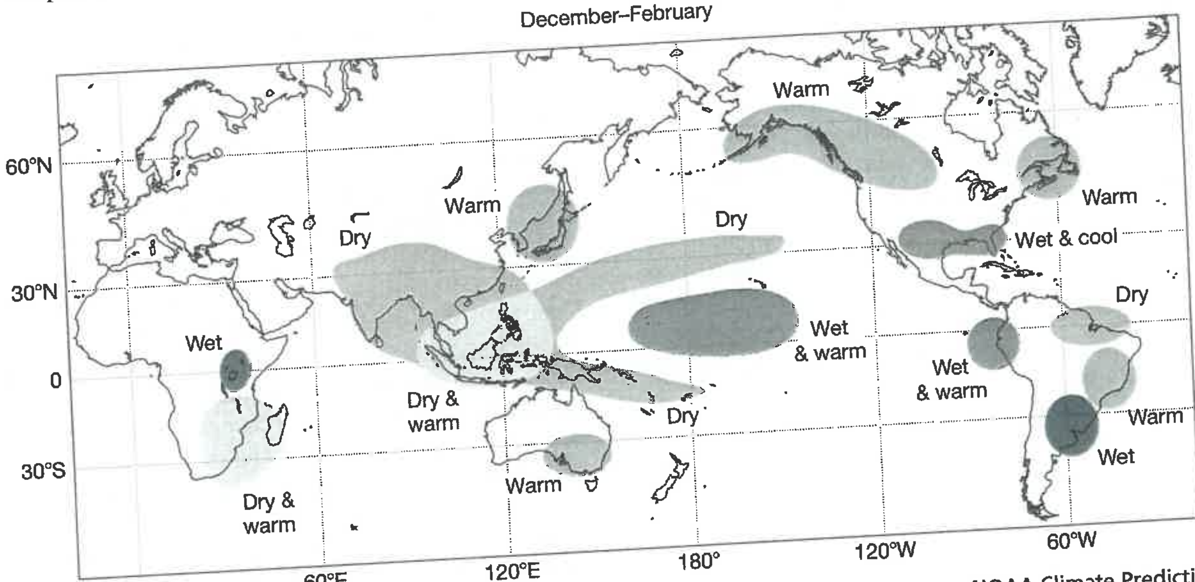


FIGURE 5-12 Rainfall and temperature anomaly patterns associated with an ENSO event. (*Source:* NOAA Climate Prediction Center http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/impacts/warm.gif)

always the same, and ENSO events may last one year or may extend over several years. As a result, the pattern of the climate anomaly in the tropics tends to be fairly consistent, although it may vary in magnitude, but the midlatitude effects are highly variable. The western United States, for example, was very dry in the 1976 event but very wet in 1982.

THE CIRCULATION OF THE DEEP OCEAN

Salinity

In contrast to the surface-ocean circulation, which is driven by atmospheric winds, the deep-ocean circulation is driven by differences in water density. These differences are caused by variations in temperature and in **salinity,** or the salt content of a water mass. Salinity is measured in terms of the proportion of dissolved salt to pure water. The salinity of the ocean is a measure of the quantity of different elements—sodium and chlorine (Na, Cl) being the most common—dissolved in a given mass of seawater (*g* salt/*kg* seawater). Oceanographers, until recently, expressed salinity in parts per thousand (‰), or *per mil.* (Now they express it without units.) A salinity of 1‰, or 1 out of a possible 100 parts salt, is equivalent to 10‰, or 10 out of a possible 1000 parts salt. The average salinity of the world's oceans is approximately 35‰, but there is some variability from ocean to ocean.

The primary constituents of sea salt are the ions chloride (Cl^-), sodium (Na^+), sulfate (SO_4^{2-}), magnesium (Mg^{2+}), calcium (Ca^{2+}), and potassium (K^+) (Table 5-1). Except for calcium, which shows some variability from place to place, these elements are found in nearly constant proportions around the globe. Many of the minor constituents—but not all—show a similar uniformity. Most of the variability occurs in those constituents that are utilized by marine organisms.

The salts contained in seawater are largely the result of the breakdown of crustal rocks, or *weathering.* Weathering occurs when rocks are altered by physical or chemical processes. When water flows over or through rocks, it removes soluble materials (ions). Rivers eventually carry the soluble ions to the ocean. In fact, it is estimated that rivers deliver between 2.5×10^{12} and 4×10^{12} kg

TABLE 5-1 Salt Content of the Earth's Oceans

Salt Ion	Grams per Kilogram (g/kg) of Ion in Seawater	Ion by Weight (%)
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
Flouride (F^-)	0.001	0.00
Total	34.482	99.99

Source: P. R. Pinet, *Oceanography*, St. Paul, MN: West, 1992.

(about 4 billion tons) of dissolved salts to the oceans each year. The oceans are much saltier than the river water because, when ocean water evaporates, the salt is left behind (increasing the salt concentration) in the ocean. Some of that evaporated freshwater falls on the land and eventually runs back to the ocean. It weathers rocks along the way and thus delivers a new supply of salts that accumulate over time to increase the saltiness of the ocean.

If such great volumes of salts reach the oceans each year, are the oceans still getting saltier with time? The answer is no, because many processes also remove salts from seawater. These processes include the following:

1. Evaporation of seawater from shallow seas. The remaining salts are concentrated and precipitate from solution as **evaporite deposits,** such as halite (table salt, NaCl) and gypsum ($CaSO_4 \cdot 2H_2O$).
2. Biological processes. For example, some marine microorganisms remove the elements calcium or silicon from seawater to form their shells, some of which are eventually deposited in ocean sediments.
3. Chemical reactions between seawater and newly formed volcanic rocks on the sea floor.
4. The formation of sea spray. As small droplets of seawater become airborne, salts, especially sodium and chlorine, are removed when the spray is deposited on land. These salts are eventually returned to the oceans via rivers.

Overall, salts are removed from seawater at a rate that essentially equals the rate of input, when averaged over geologic time scales (millions of years). In other words, the present salt content of the oceans does not represent the result of continuous accumulation but simply a balance between the rates of input and output of salts (see the Box "A Closer Look: The Salt Content of the Oceans and the Age of Earth").

Variations in salinity are caused by regional differences in evaporation, precipitation, sea ice formation and melt, and river runoff. Surface salinities increase where evaporation exceeds precipitation. We see this effect in such areas as the Mediterranean Sea, the Red Sea, and the Arabian Gulf. In contrast, the Gulf of Bothnia in the Baltic Sea, which also has little exchange of water with the open ocean but experiences much greater precipitation, has relatively low salinities. A similar effect is seen in the Chesapeake Bay on the Atlantic coast of the United States.

Thermohaline Circulation

Because deep-ocean circulation depends on temperature and salinity, this circulation is referred to as **thermohaline circulation** (*thermo* is Greek for "heat," and *haline* comes from the Greek *hals,* for "salt"). In discussing atmospheric circulation, we showed that large horizontal (in particular, latitudinal) changes in temperature and pressure lead to steep pressure gradients and a relatively rapid air circulation. In the deep oceans, horizontal changes in density are small, whereas vertical changes can be larger. But the densest water is at the bottom, so the structure is very stable. Consequently, the movement of water through the deep ocean is relatively slow. Although density-driven movements are much slower than the surface currents, they are no less important in shaping Earth's climate on time scales of hundreds to thousands of years.

THE VERTICAL STRUCTURE OF THE OCEANS The vertical structure of the deep oceans is determined by water density: The highest densities tend to occur in the deepest layers, while the lowest densities are typically found near the surface. Water density, in turn, is controlled by temperature and salinity: Usually, density increases as salinity increases or as temperature decreases; density decreases as

A CLOSER LOOK
The Salt Content of the Oceans and the Age of Earth

Following ideas first expressed by British astronomer Sir Edmund Halley in 1715, Irish scientist John Joly attempted to calculate the age of Earth on the basis of estimates of the salt content of the ocean and the rate of delivery of salts to the ocean. Two hundred years after Halley, Joly calculated Earth to be 80–89 million years old. However, we now know that Earth is approximately 4.6 billion years old. So where did Joly go wrong?

Joly assumed that the ocean had simply been accumulating all the salts delivered to it by rivers at a constant rate since Earth first formed. Joly neglected the various processes that remove salts from seawater (see the accompanying text). Repeating Joly's calculations but using current estimates of ocean volumes and salinities, we obtain the following:

- The total amount of salt in the oceans is approximately 5×10^{19} kg.
- The rate at which rivers deliver salt is 4×10^{12} kg/yr.
- Therefore, the "age" of Earth is $5 \times 10^{19}/4 \times 10^{12} = 13 \times 10^6$ yr.

Thirteen million years is somewhat less than Joly calculated with his knowledge of the world's river discharge, chemical composition, ocean volume, and salt content. The "age" that we have calculated is, in fact, the average length of time salt remains in the ocean. As we will see in Chapter 8, the length of time a substance remains in a given reservoir is called the *residence time*.

salinity decreases or as temperature increases. However, water is an unusual fluid in that the density of freshwater *increases* as temperature increases from 0 to 4°C. After that point, water acts more like other fluids in that its density *decreases* as temperatures continue to increase. The temperature at which maximum density occurs varies as salinity changes. Maximum density occurs at 4°C for freshwater (0‰), 2°C for a salinity of 10‰, and at the freezing point for a salinity of 24.6‰. As salinities continue to increase, the temperature of maximum density continues to decrease (it actually stays at the freezing point, which also decreases as salinity increases).

The lower-density zone, which occurs in the top 60–100 m of the ocean, is called the *surface zone,* a layer that interacts with the overlying atmosphere. This interaction takes place through evaporation, precipitation, exchanges of kinetic energy (the effect of winds and friction), radiative exchanges (the absorption of solar radiation and the emission of long-wavelength radiation), and the exchange of heat. This zone is well mixed by wind action, and so the surface zone is often referred to as the **mixed layer.**

The transition zone between the surface zone and the deep ocean is on the order of a kilometer in thickness and is characterized by a rapid increase in density with increasing water depth. The very sharp increase in density is called the **pycnocline;** the transition zone is referred to as the *pycnocline zone* (Figure 5-13). In some regions this density gradient is dominated by salinity changes, and salinity rises rapidly with increasing depth. In this case, the salinity gradient is specifically referred to as the **halocline.** In most other regions, temperature changes dominate the density gradient, and temperature drops rapidly with increasing depth. There the transition is called the **thermocline.** In either case, the steep density gradient makes this layer very stable. This stability limits vertical movements and insulates the deep ocean from seasonal changes in temperature and salinity. The deep ocean below the pycnocline (typically 1–5 km depth) contains about 80% of the volume of all oceans. Deep-ocean water is further stratified, with the highest densities at the sea floor. The water column within the deep ocean, therefore, is also stable, and little vertical movement takes place. The movement that does occur is subhorizontal, along sloping layers of equal density (*isopycnals*).

Figures 5-14 and 5-15 show temperature and salinity profiles, respectively, from the Atlantic and Pacific ocean basins. We see that the simple vertical structure outlined previously applies to most of the world's oceans, with the exception of those in high latitudes. The high-latitude seas are characterized by low temperatures and relatively low salinities at the surface, similar to the waters of the deep oceans. We will see next that there is a connection between these high-latitude surface waters and the deep oceans and that the formation of very dense surface water near the poles is, in fact, the primary driving force for the deep-ocean circulation.

BOTTOM-WATER FORMATION. Deep-ocean circulation begins with the production of dense (cold and/or salty) water at high latitudes. This very dense water can be produced by several processes. For example, cooling and increased salinity may result from a large difference between evaporation and precipitation or from the formation of sea ice. Forming along the sea ice margin in just a few regions in the polar oceans, **bottom water** constitutes the densest water produced in the oceans. Near the poles, the surface waters are cooled below the normal freezing point (–1.9°C in some areas) by contact with the cold overlying atmosphere. (The freezing point is lower than that of pure water because of the presence of salt.) When that water freezes, it forms a layer of sea ice several meters thick that floats on the surface of the polar oceans. When the ocean surface freezes, most of the sea salt is excluded, because the salt does not fit into the crystal structure of the ice. As a result,

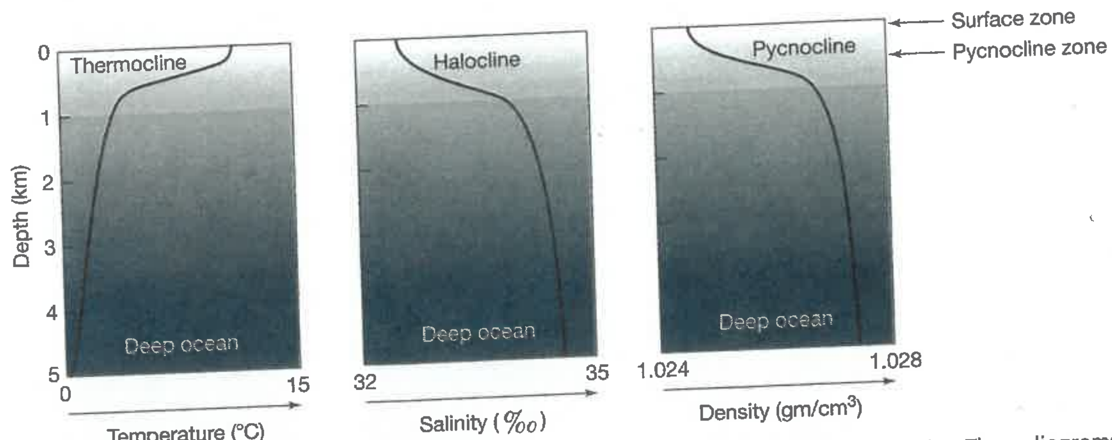
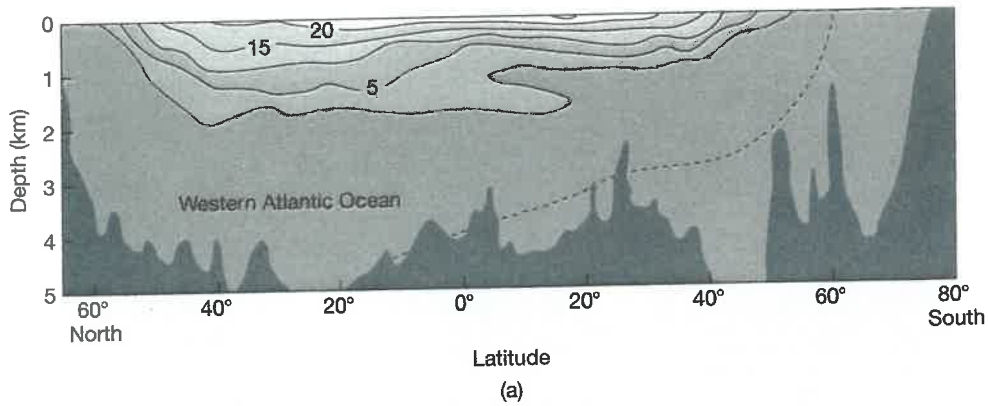


FIGURE 5-13 Generalized profiles of temperature, salinity, and density in the midlatitude ocean basins. These diagrams show the ocean to be divided into three layers: the surface zone, where there is little change in temperature, pressure, and density with depth; the pycnocline zone, where density increases rapidly (the pycnocline) and where there is an increase in salinity (the halocline) and a rapid decrease in temperature (the thermocline); and the deep ocean, where salinity generally increases slowly with depth, temperatures gradually decrease with depth, and there is little change in salinity.



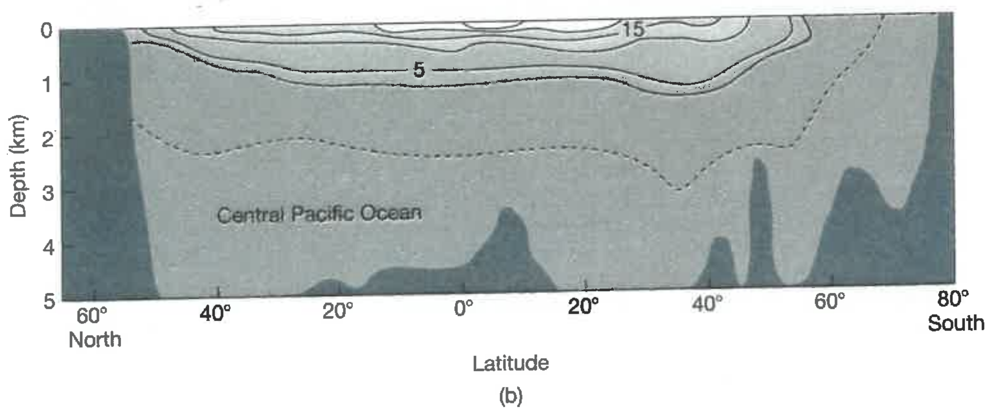
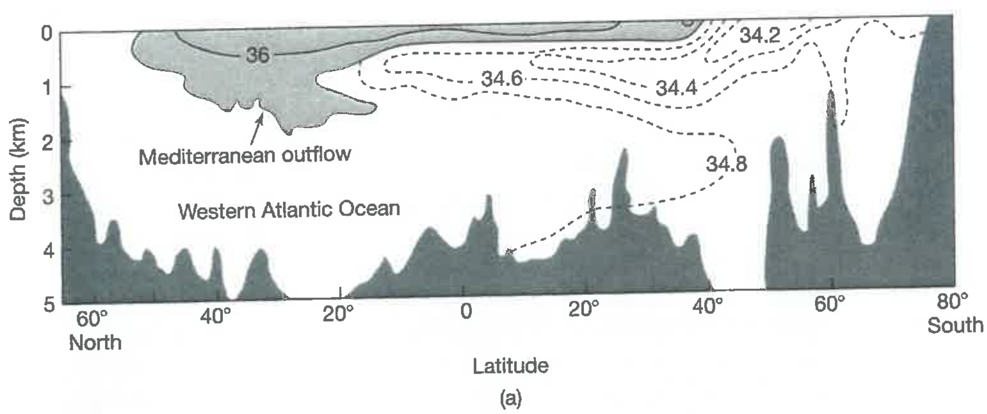


FIGURE 5-14 Vertical distribution of temperature in the (a) Atlantic and (b) Pacific oceans. The thermocline separates deep water from the surface layer in the tropics, but deep water extends to the surface at high latitudes. Temperature values are given in degrees Celsius.



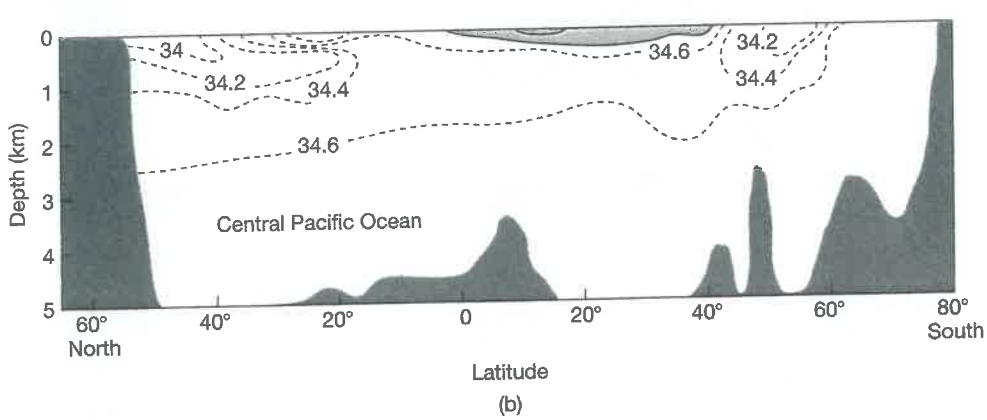


FIGURE 5-15 Vertical distribution of salinity in the (a) Atlantic and (b) Pacific oceans. The salinity profiles are more complicated than the temperature distributions in Figure 5-7. In the deep ocean, salinity tends to increase in the deeper waters. There is a salinity maximum at the surface in the tropics, however, due to evaporation. When water evaporates, the salts are left behind. Where evaporation exceeds precipitation, there is a net loss of water from the surface layer, and the remaining water has a higher salt concentration.

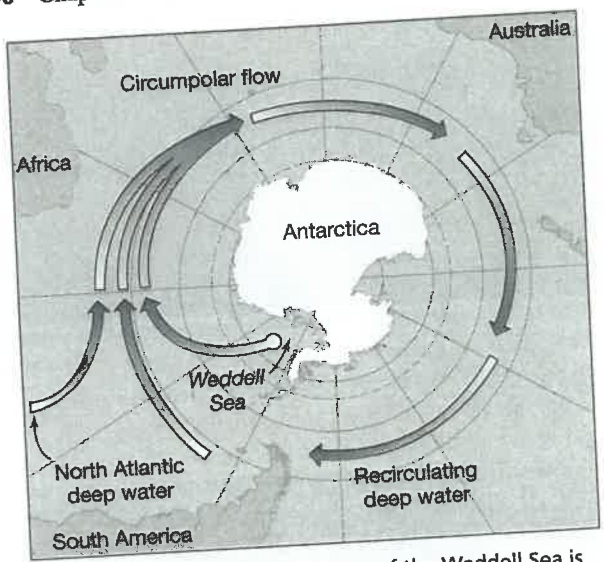


FIGURE 5-16 The Weddell Sea. Part of the Weddell Sea is occupied by an ice shelf (a mass of ice several hundred meters thick that flows from the West Antarctic ice cap). The remainder of the sea is covered by sea ice in winter. The ice forms near the coast and is pushed northward by persistent winds blowing off the ice cap. As the ice is pushed away from the coast, open water is exposed that freezes rapidly in the very cold temperatures. This ice, in turn, is pushed northward, allowing even more ice to form in a continuous process throughout the winter. This region is thus an ice-making factory. The result is the formation of very cold, highly saline water at the surface, which sinks to produce Antarctic Bottom Water. (*Source:* W. S. Broecker and T.-S. Peng, *Tracers in the Sea*, New York: Eldigio Press, 1982, Figure 7-17.)

the water just beneath the sea ice becomes saltier, and an underlayer of very cold, highly saline water forms. The combination of low temperatures and high salinity results in very dense water that sinks and flows down the slope of the basin and spreads toward the equator as the bottom layer of water in the deep-ocean basins.

One of the major sites of bottom-water formation is the Weddell Sea off Antarctica (Figure 5-16). The water formed there, called **Antarctic Bottom Water (AABW),** circles Antarctica and flows northward as the deepest layer in all three major ocean basins (Atlantic, Pacific, and Indian). Although few direct measurements of deep-ocean circulations have been made, reconstructions using observations of the distributions of temperature and salinity have identified AABW as far north as 45° N in the North Atlantic and 50° N (at the Aleutian Islands) in the North Pacific. Typical speeds for deep-ocean currents are only 0.03 to 0.06 km/hr, yet AABW has traveled more than 10,000 km from its formation site in the Weddell Sea, a trip that has taken some 250 years.

Similar masses of cold, dense water form in the Arctic Ocean—off the coast of Greenland—and flow south at depth into the western North Atlantic. These water masses are referred to as **North Atlantic Deep Water (NADW).** The processes of NADW formation are not entirely clear. Warm waters flowing northward in the southwest North Atlantic are evaporated and some portion of the water vapor is carried westward across Central America, where it falls as precipitation in the Pacific. This has the effect of slightly diluting (making less saline) the waters of the eastern Pacific, while producing higher salinities in the western Atlantic (the salt is left behind when the water evaporates). The Gulf Stream carries this more saline water northward where it cools, and the low temperatures and high salinities cause it to sink. At the same time, sea ice formation in the Greenland and Norwegian seas—like in the Weddell Sea—further increases the salinity and density of the surface waters in these regions. The NADW provides approximately half the input of deep water to the world's oceans, and the remainder comes from the Weddell Sea. The NADW that forms to the west of Greenland in the Labrador Sea sinks directly into the western Atlantic; the NADW that forms in the Norwegian Basin subsides and is dammed behind the *sills* (undersea ridges) that connect Greenland to Iceland and Iceland to the British Isles. This water periodically flows over the sills and cascades into the deep basins of the North Atlantic.

NADW flows southward through the Atlantic Ocean and joins the Antarctic Circumpolar Current, which flows around Antarctica. There the NADW and the AABW combine and circle the continent. They then proceed to branch off into the Indian and Pacific oceans (Figure 5-17). Some of the water completes the circle, reentering the Atlantic or continuing around for another circuit. The time scale over which this occurs is indicated in Figure 5-18, which shows the age of the water at various places in this flow. The map actually shows the change in the amount of radioactive carbon (^{14}C) present in the water masses, which represents the time since that body of water sank below the mixing layer and was no longer exchanging carbon dioxide with the atmosphere (see the two Boxes "Useful Concepts: Isotopes and Their Uses" and "A Closer Look: Carbon-14—A Radioactive Clock"). The decreasing proportions of ^{14}C indicate the pattern of flow. You can see the youngest water in the North Atlantic, getting progressively older as it flows south to the Southern Ocean. There is a further addition of young water off the Antarctic coast, and again the water gets progressively older as it flows around the Southern Ocean and up into the Indian Ocean or the northeastern Pacific Ocean. Some mixing of the deep water with surrounding water masses occurs, as does some biological addition of ^{14}C. Consequently, the ^{14}C age does not necessarily reflect the true age of the water masses. However, from these data it is possible to compute flow rates into the various basins, from which it is possible to determine the replacement (residence) time for deep water and the upwelling rates (how long it takes to get back to the surface). Taking residence times and upwelling rates into account, a parcel of Antarctic Bottom Water will reemerge at the surface in the Indian Ocean (on average) after 335 years, or in the Pacific Ocean after 595 years. The average residence time for the entire deep ocean is approximately 500 years.

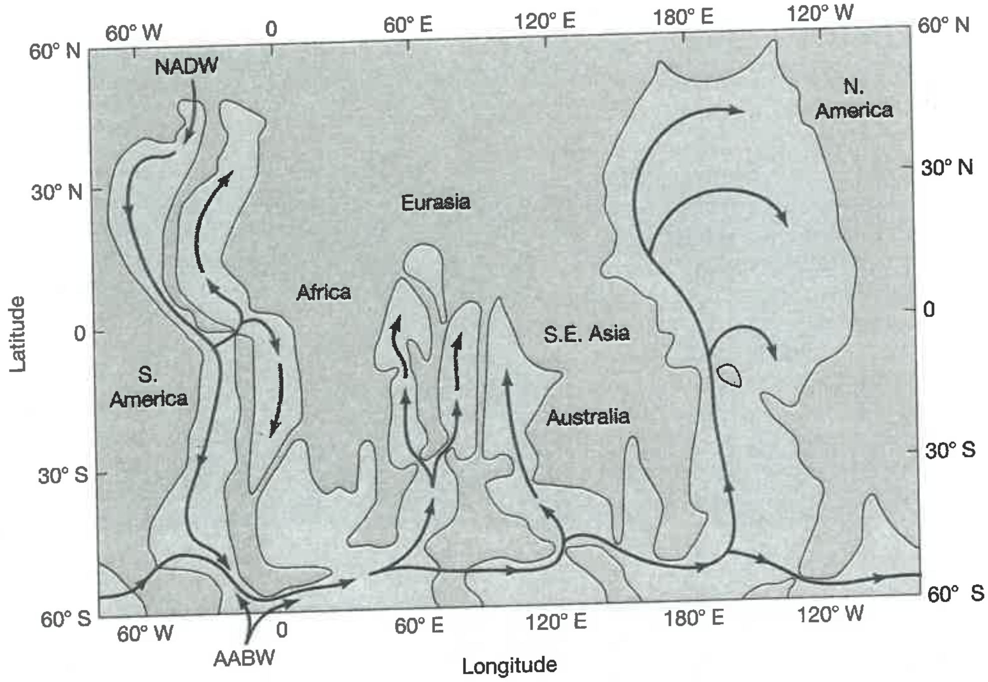


FIGURE 5-17 Flow pattern of the North Atlantic Deep Water and the Antarctic Bottom Water. This diagram represents the flow at a depth of 4000 m; the strange-looking continent/ocean configuration is what we would obtain if the oceans were drained to this depth. (*Source:* W. S. Broecker and T.-S. Peng, *Tracers in the Sea*, New York: Eldigio Press, 1982, Figure 1-12.)

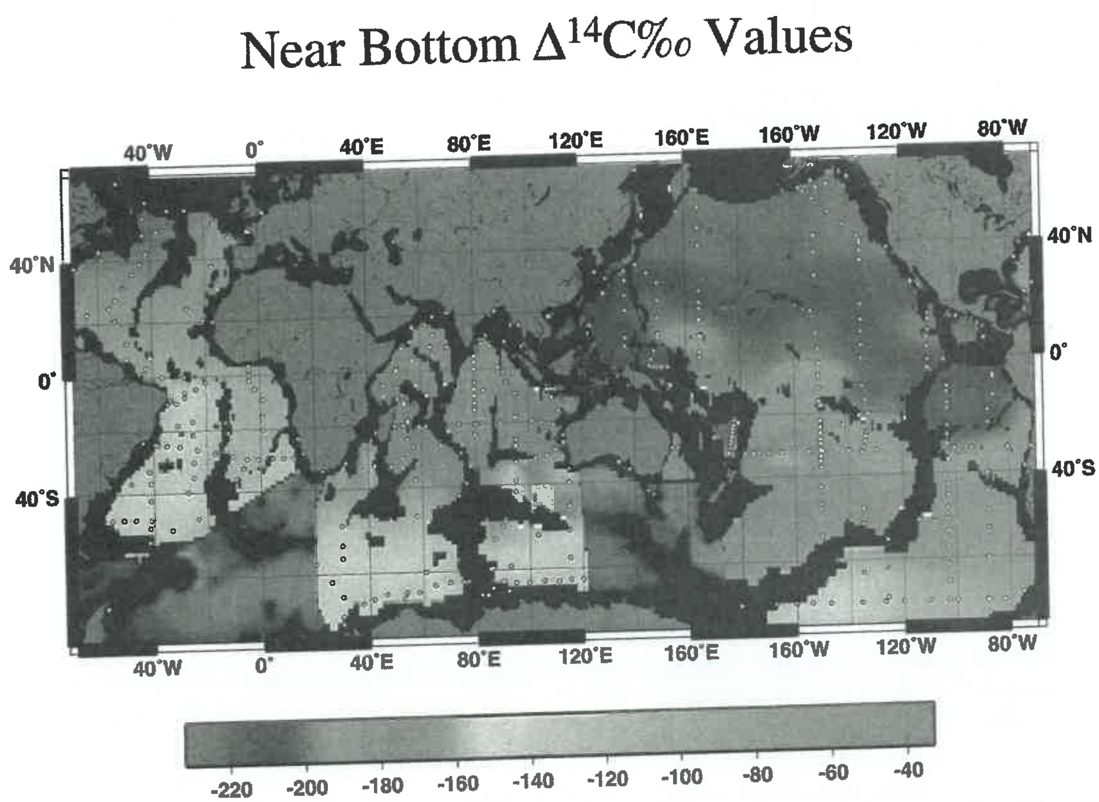


FIGURE 5-18 [See color section] ^{14}C difference values for the near bottom waters of the world's oceans. The values represent the change in the amount of radioactive carbon (^{14}C) present in the water body compared to present-day surface waters (see "A Closer Look: Carbon-14—A Radioactive Clock"). The smallest values represent waters where the ratio of radioactive is stable carbon are most similar to the present-day ocean values (i.e., the youngest water bodies). Regions with the largest difference values show the oldest water masses. The ^{14}C acts as a tracer that shows the path of water movement in the deep oceans. (*Source:* Diagram courtesy Robert M. Key, Princeton University.)

USEFUL CONCEPTS

Isotopes and Their Uses

Much of what we know about the present, and especially the past, Earth system comes from the use of isotopes. **Isotopes** are atoms of a given element that have different numbers of neutrons in their nuclei. Isotopes have the same **atomic number**—that is, the same number of protons in the nucleus—but a different **mass number,** which is the total number of protons plus neutrons in the nucleus. Carbon is a good example because it has several isotopes that are used for all sorts of different purposes in studying Earth and its biota. All carbon atoms have 6 protons. ^{12}C also has 6 neutrons, while ^{13}C has 7 neutrons and ^{14}C has 8 neutrons. The superscript preceding the element's symbol denotes the mass number.

Some isotopes of a given element are **stable isotopes,** which means that they do not spontaneously change into other isotopes or into atoms of another element. **Unstable isotopes** spontaneously change into other isotopes or elements by a process called **radioactive decay.** ^{12}C and ^{13}C are both stable, while ^{14}C is unstable and radioactive. Some 98.89% of all carbon is ^{12}C. ^{13}C constitutes most of the remaining carbon (~1.11%), while ^{14}C occurs only 0.0000000001% of the time. In other words, there are 10^{12} atoms of ^{12}C for every one atom of ^{14}C.

The stable and unstable carbon isotopes are both useful but for entirely different reasons. ^{13}C is used for studying the behavior of the carbon cycle over long time scales. Various microorganisms, especially photosynthetic ones, take up the different isotopes of carbon at different rates. Generally, the heavier isotope is taken up more slowly than the lighter one. We'll see in Chapter 11 that this leaves a useful record for understanding rates of organic carbon burial and variations in atmospheric oxygen in the geologic past. ^{14}C is also taken up more slowly by microorganisms, but that effect is dwarfed by the fact that it is also radioactive. Thus, the main use of ^{14}C is for **radiometric age dating.** The Box "A Closer Look: Carbon-14—A Radioactive Clock" describes how this technique is used to study deep-ocean circulation.

Linking the Thermohaline Circulation and the Wind Driven Surface Flow

We have now described how surface waters in the polar regions sink and spread at depth throughout the world's oceans. Ultimately these waters must return to the surface to complete the circulation (Figure 5-19). Water is only slightly compressible (we cannot cram more and more of it into the same space), so if water is sinking at high latitudes, it must be rising somewhere else. There must also be some flow of surface water into the high latitudes to replace the water that is subsiding and moving equatorward. This return flow is even more difficult to monitor than the flow of bottom water. It takes place very slowly through the pycnocline over the whole ocean and more rapidly through upwelling in currents along the eastern boundary of ocean basins and other regions of upwelling and deep mixing. Once the former deep water has reached the surface, the surface circulation that we discussed earlier returns the water to the polar regions. According to Wallace Broecker, a geochemist at Columbia University, this complete cycling of ocean water that is driven by thermohaline circulation can be likened to a giant conveyor belt.

The thermohaline conveyor belt is a significant feature of the Earth system in several respects: It plays a dominant role in the recycling of ocean nutrients, and it has a major impact on Earth's climate. Much of the life that exists in the oceans can be found in the near-surface layers, utilizing sunlight for photosynthesis—phytoplankton, for example—or living off the animals that feed on phytoplankton. These plants and animals use the nutrients in ocean water, so the surface layers become relatively depleted in nutrients. When these organisms die, they sink through the water column, decompose, and release the nutrients back into the water. The deeper ocean, therefore, is relatively rich in nutrients. The thermohaline circulation transports these nutrient-rich waters around the globe, returning the nutrients to the surface in areas of upwelling, primarily along the continental margins. Consequently, the concentrations of marine life are greatest in these upwelling regions. This is illustrated in Figure 5-20, a satellite image showing productivity in the oceans. The satellite measures the ocean color and the information is converted to concentrations of chlorophyll pigment in phytoplankton, so the image is showing where high concentrations of phytoplankton are found near the ocean surface. Note the high productivity in the North Atlantic and in upwelling coastal zones. In contrast, note the low concentrations in the middle of the primary ocean gyres.

Our description of the ocean circulation depicts a complex system of surface-wind-driven currents overlying a deep ocean with a relatively simple circulation driven by bottom-water formation and surface divergence. In reality, the deep oceans are much more complex. Clearly distinguishable water masses can be identified at different depths and in different geographic locations where variations in temperature and salinity impart different characteristics to the water bodies. For example, high evaporation rates and low rainfall (together with little river runoff) produce relatively warm, highly saline water in the Mediterranean Sea. This water flows out of the Mediterranean at depth through the Straits of Gibraltar and is clearly recognizable as a plume of warm, saline water spreading out into the mid-Atlantic at a depth of about

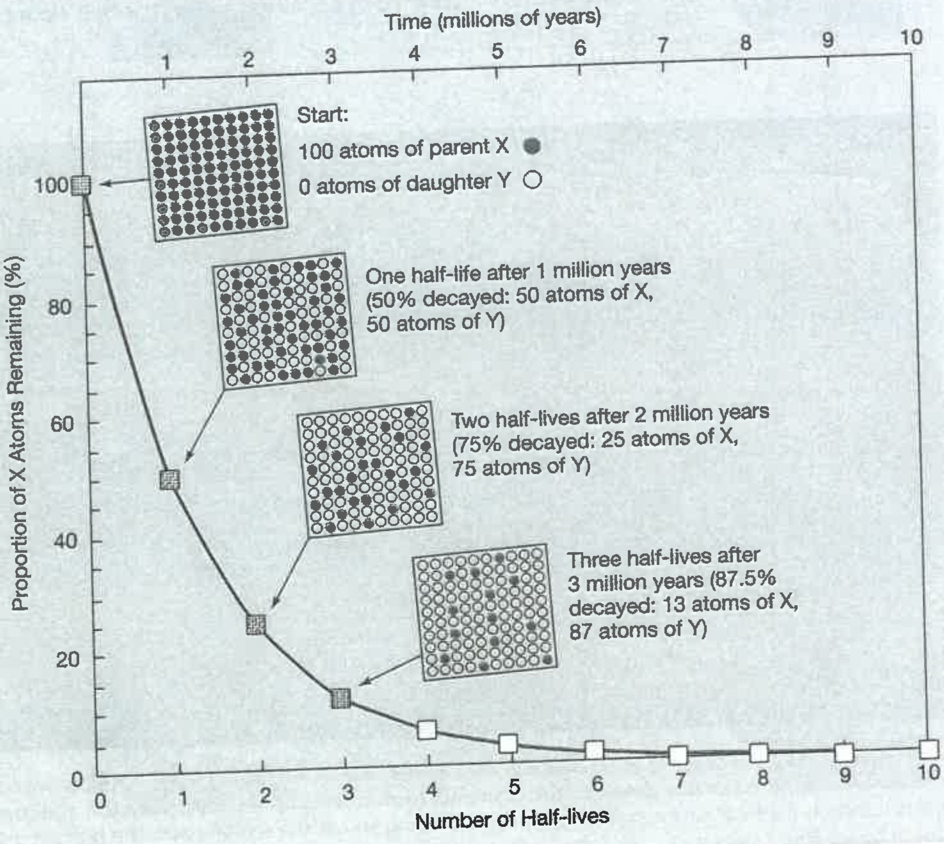
A CLOSER LOOK

Carbon-14—A Radioactive Clock

The radioactive isotope (or *radioisotope*) of carbon (^{14}C) is produced in the upper atmosphere through bombardment by cosmic rays from distant sources in the galaxy. This bombardment breaks apart atoms producing neutrons, which may then collide with other atoms. Nitrogen atoms (^{14}N) have 7 protons and 7 neutrons. When a nitrogen atom is struck with one of these "cosmic ray" neutrons, the neutron replaces one of the protons in the nucleus. The atom now has 6 protons and 8 neutrons—in other words, the ^{14}N becomes ^{14}C. The ^{14}C is unstable and so it decays back to nitrogen. Radioactive decay is exponential—half occurs in the first 5,730 years, half of the remainder in the next 5,730 years, and so on. This time period, in which half of the initial quantity of radioactive isotope decays, is referred to as the isotope's **half-life** (see Box Figure 5-4).

The ^{14}C is rapidly oxidized to $^{14}CO_2$ and is distributed through the atmosphere. Production of ^{14}C occurs at a relatively constant rate, so the proportion of ^{14}C to stable carbon in the atmosphere remains constant. Living organisms take up the unstable carbon and, although the ^{14}C immediately begins to decay, it is replenished by more ^{14}C from the atmosphere, maintaining an equilibrium that matches the proportions in the atmosphere. Once the organism dies, however, metabolic activity ceases so the ^{14}C continues to decay radioactively, but can't be replenished. Knowing the rate at which this decay occurs, it is possible, by looking at the ratios of ^{14}C to ^{12}C, to determine how long ago the organism lived. If you have a piece of wood, for example, that has half the amount of ^{14}C (in relation to ^{12}C) than we find in living trees, then we know that the piece of wood came from a tree that died roughly 5,700 years ago. This process, called **radiocarbon dating**, has been used to date materials back to 50,000 years ago and is used extensively in archeology and for reconstructing past climates.

^{14}C is also useful as a tracer of ocean circulation. Because the atmosphere exchanges CO_2 with the ocean surface, the surface waters of the ocean have nearly the same ratio of ^{14}C to ^{12}C as does the atmosphere. When surface waters sink, however, the ^{14}C that is present begins to decay, and it cannot be replenished. Consequently, the ratio of ^{14}C to ^{12}C in the deep waters gives a measure of how long it has been since the water was near the ocean surface: low $^{14}C/^{12}C$ ratios indicate "older" deep water. By measuring $^{14}C/^{12}C$ ratios in different localities, we can trace the time it takes for water to flow around the globe. The youngest water is found in the Weddell Sea near Antarctica and in the Norwegian/Barents Sea between Norway and Greenland. These are places where bottom water is being formed. The oldest deep water is found in the Northeast Pacific. By combining these and other data, the path and rate of the thermohaline circulation can be determined (Figure 5-18).



BOX FIGURE 5-4 The graph of radioactive decay is exponential. In other words, half of the radioactive parent is left after one half-life. After a second half-life, a quarter of the parent is left, and so on. (*Source:* From J. P. Davidson, W. E. Reed, and P. M. Davis, *Exploring Earth: An Introduction to Physical Geology*, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

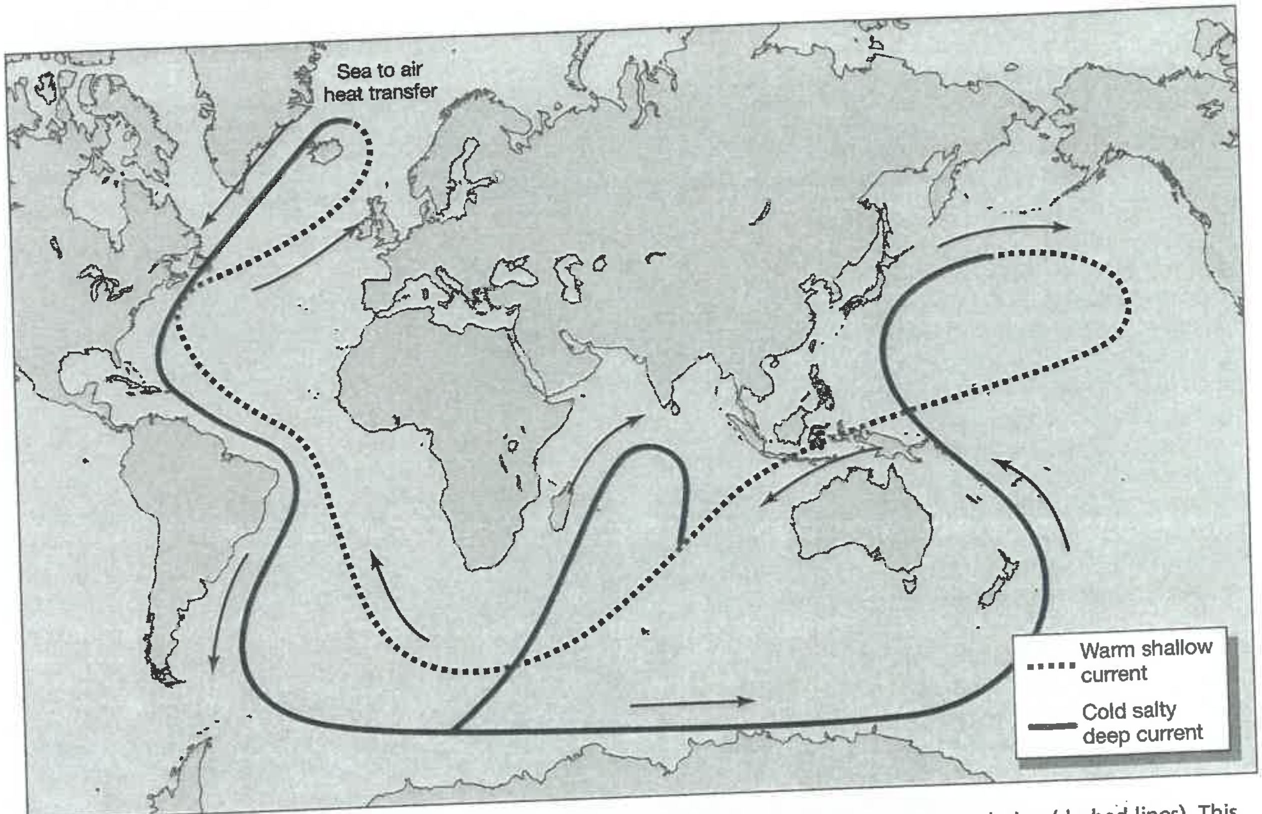


FIGURE 5-19 An idealized map of the deep-water flow (solid lines) and the returning surface circulation (dashed lines). This circulation has been described as a global conveyor belt. The deep water flows out of the North Atlantic, mixing with warmer water to the south. It is recooled by mixing with the cold surface water that subsides around Antarctica. Joining with the Antarctic Bottom Water, it flows around Antarctica in the Antarctic Circumpolar Current. Branches then flow back into the Atlantic as well as the Pacific and Indian oceans, where upwelling brings the cold waters to the surface. The water eventually returns via the surface currents to the North Atlantic to complete the circulation. (*Source:* From W. K. Hamblin and E. H. Christiansen, *Earth's Dynamic Systems,* 8/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

FIGURE 5-20 [See color section] The Coastal Zone Scanner carried on the Nimbus-7 satellite was one of the first instruments to record ocean color. The satellite detected the pigments from chlorophyll in phytoplankton and so measures the phytoplankton concentrations in the near-surface waters. The light shading shows the regions with the highest productivity. (*Source:* NASA/ Goddard Space Flight Center.)

1000 m (Figure 5-21). We need not concern ourselves with these added complexities, but it is worth noting that our knowledge of the deep-ocean circulation is limited and that much scientific investigation remains to be done in order for us to fully understand what is going on. This understanding is particularly important because the oceans play such a significant role in global climate.

Ocean Circulation and Climate

As we discussed earlier, the ocean circulation has a strong influence on global temperatures. The transport of warm surface water toward the poles, to replace the bottom water that forms near the sea ice margin, is one mechanism by which excess solar energy is transferred poleward. Figure 5-22

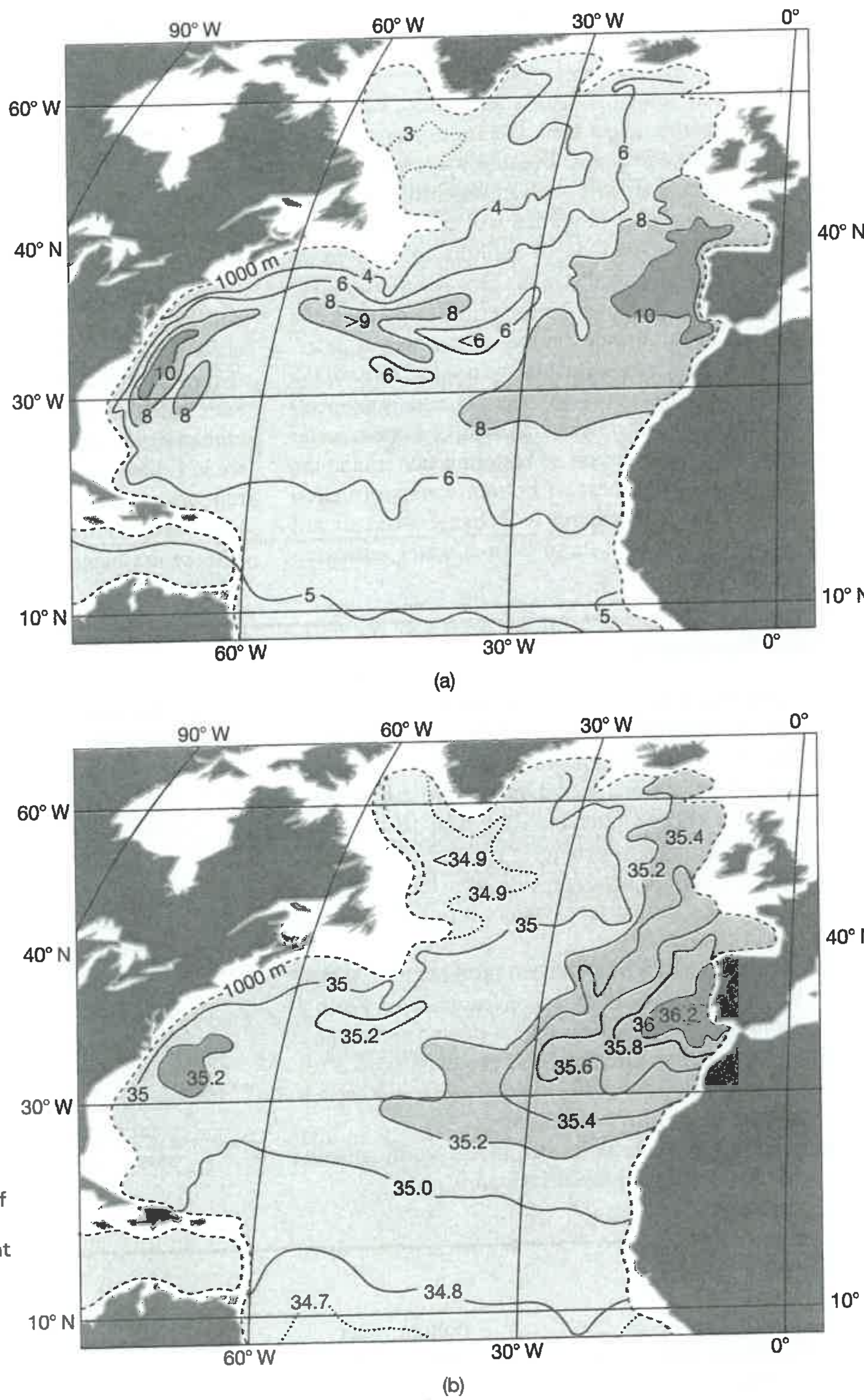


FIGURE 5-21 The distribution of (a) temperature (in degrees Celsius) and (b) salinity (per mil) at 1000-m depth in the North Atlantic, showing the spread of Mediterranean Sea water. (*Source:* Open University, *Ocean Circulation*, New York: Pergamon Press, 1989.)

shows the Northern Hemisphere poleward heat transport in the atmosphere and ocean. The total heat transport data are derived by calculating the heat transfer necessary to balance the radiation budget at each latitude (see Figure 4-2) and estimating how much of this transfer can be accomplished by the atmosphere. The ocean heat transport is then obtained by subtracting the atmospheric transport from the total. The estimates indicate that the ocean (1) provides almost as much poleward heat transport as does the atmosphere and (2) transports more heat than does the atmosphere at low latitudes, whereas the atmospheric transport dominates at middle to high latitudes.

At the same time, the oceans represent a vast reservoir of heat, absorbing heat from the atmosphere in some areas and releasing it in others. Because water heats up and cools down relatively slowly, pools of water that are cooler than normal or warmer than normal will cool or warm the atmosphere on time periods of months to seasons or years—the time needed for the pools of water to heat up or cool down. On much longer time periods, however, the average effect of the oceans on the atmosphere is determined by the overall temperature of the oceans. Most of the water in the oceans lies in the deep oceans, and its temperature is largely determined by the process of bottom-water formation and by the transport of bottom water around the ocean basins. If the process of bottom-water formation changes, the ocean temperatures will change—and so will climate. Estimates of the rate of bottom-water formation

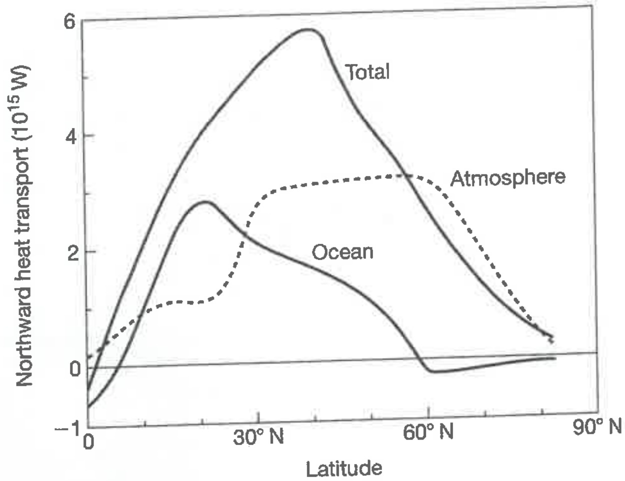


FIGURE 5-22 Poleward heat transport in the Northern Hemisphere. (*Source:* Open University, *Ocean Circulation*, New York: Pergamon Press, 1989.)

under the present climate, together with measurements of ocean volume, indicate that it would take about 1,000 years to recycle all of the deep water in the oceans. Hence, we can anticipate that the thermohaline circulation could moderate climate over time periods of about 1,000 years. However, we also have geologic evidence indicating that brief interruptions or changes in the thermohaline circulation can also have rapid and large impacts on regional climates, as we will see in Chapter 14.

Chapter Summary

1. As with the atmosphere, the driving force for the oceanic circulation is the global distribution of energy. Unlike the atmosphere, however, the oceanic circulation is driven indirectly by temperature differences: The surface-ocean circulation is, in fact, driven by the circulation of the atmosphere.
   a. Due to friction, wind blowing over the ocean surface drags the surface waters along, producing ocean currents.
   b. The pattern of surface-ocean currents is modified by the Coriolis effect, a consequence of Earth's rotation, and by the distribution of land and oceans.
2. The thermohaline circulation of the deep oceans results from temperature and salinity variations, which control the density of ocean waters.
   a. Cold, saline water is formed in the North Atlantic and in the Weddell Sea off Antarctica.
   b. The combination of low temperatures and high salinities produces very dense water that sinks to the ocean floor and flows as bottom water throughout the world's oceans.
3. Bottom water eventually rises to the surface in zones of upwelling and returns in surface currents back to the highlatitude source regions, completing a vast oceanic conveyor belt.
   a. In combination with the atmospheric circulation, the net effect of these oceanic circulations is to redistribute thermal energy from low latitudes, where Earth is hot, toward the poles, where it is cold.
   b. The thermohaline conveyor belt and the associated zones of upwelling and downwelling play a significant role in climate and in the distribution of nutrients in the oceans.

Key Terms

absolute vorticity
Antarctic Bottom Water (AABW)
atomic number
bottom water
downwelling
Ekman spiral
Ekman transport
El Niño
El Niño–Southern Oscillation (ENSO)

evaporite deposit
geostrophic current
gyre
half-life
halocline
isotope
La Niña
mass number
mixed layer
North Atlantic Deep Water (NADW)
planetary vorticity
pycnocline
radioactive decay
radiocarbon dating
radiometric age dating
relative vorticity
salinity
Southern Oscillation (SO)
Southern Oscillation Index (SOI)
stable isotope
thermocline
thermohaline circulation
unstable isotope
upwelling
vorticity

Review Questions

1. What effects does the surface-wind pattern have on the circulation of the oceans?
2. Why do ocean currents not move in exactly the same direction as the wind?
3. What is the Ekman spiral? Explain why Ekman transport occurs.
4. What is upwelling? Where does upwelling occur?
5. What is meant by a geostrophic current?
6. Explain the different characteristics of western and eastern boundary currents.
7. Explain what happens to the atmospheric and oceanic circulations in the tropical Pacific during an ENSO event.
8. Where does the salt in the oceans originate? Are the oceans getting saltier with time? If not, then why not?
9. Define the term *thermohaline circulation.* What are the processes that drive the circulation of the deep oceans?
10. Explain the differences among the pycnocline, the halocline, and the thermocline.
11. What is bottom water? Where and how does bottom water form?
12. What is meant by the term *thermohaline conveyor belt?*
13. Explain what effects the ocean has on modifying the global temperature distribution.

Critical-Thinking Problems

1. Explain what is meant by Ekman transport and what role it plays in producing oceanic gyres in the surface waters of the subtropical oceans.
2. Use a rough map sketch to help explain the role that the oceans play in determining the climates of southern South America and southern Africa, poleward of 20° S.
3. Water is a very unusual substance in that it reaches maximum density between the freezing point and 4°C, depending on salinity. As you cool the water surface to these temperatures it becomes denser and sinks (rather than immediately freezing). This means that you have to cool the whole water body (the lake or the surface, mixed, layer of the ocean) to this temperature before you can cool the surface layer enough to freeze, which is why some lakes can remain unfrozen even when the air temperature drops well below freezing. When you have cooled the surface layer to the freezing point, water is again unusual in that its solid form (ice) is actually less dense than the liquid, so ice floats. Consider how different the world would be if water behaved like most other substances and continued to increase in density down to the freezing point, and if ice were denser than liquid water. Speculate on what this might have meant for life on the planet.

Further Reading

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