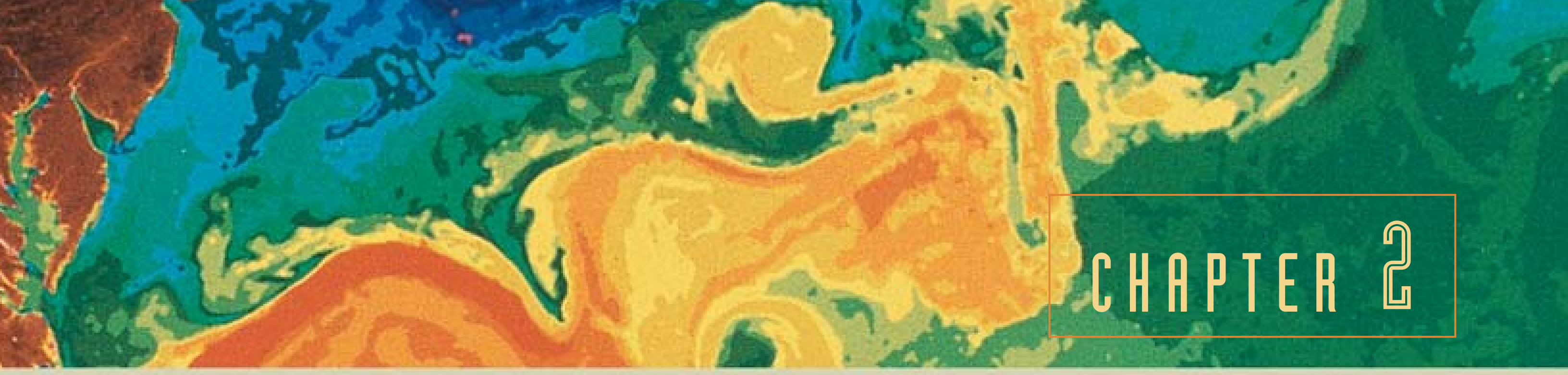


CHAPTER 2

Earth's Climate System Today

In this chapter we examine Earth's climate system, which basically functions as a heat engine driven by the Sun. Earth receives and absorbs more heat in the tropics than at the poles, and the climate system works to compensate for this imbalance by transferring energy from low to high latitudes. The combination of solar heating and movement of heat energy determines the basic distributions of temperature, precipitation, ice, and vegetation on Earth.

We first look at Earth from the viewpoint of outer space as a sphere warmed by the Sun and characterized by global average properties. Then we explore how solar radiation is absorbed by the ocean and land at lower latitudes, transformed into several kinds of heat energy, and transported to higher altitudes and latitudes by wind and ocean currents. Next we examine the kinds of ice that accumulate at the high latitudes and high altitudes where little of this redistributed heat reaches. The chapter ends by reviewing characteristics of life that are relevant to climate.

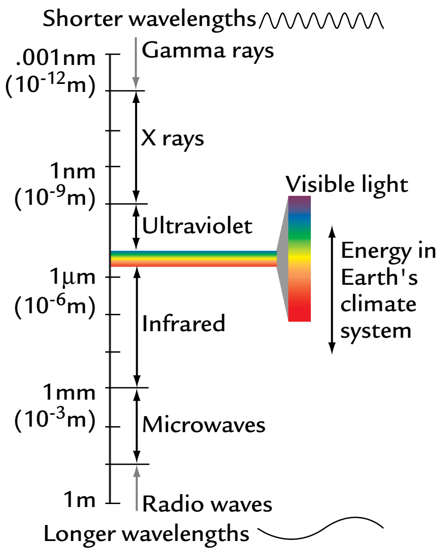


FIGURE 2-1 The electromagnetic spectrum Energy moves through space in a wide range of wave forms that vary by wavelength. Energy from the Sun that heats Earth arrives mainly in the visible part of the spectrum. Energy radiated back from Earth's surface moves in the longer-wavelength infrared part of the spectrum. (Modified from W. J. Kaufman III and N. F. Comins, *Discovering the Universe*, 4th ed., © 1996 by W. H. Freeman and Company.)

Heating Earth

Earth's climate system is driven primarily by heat energy arriving from the Sun. Energy travels through space in the form of waves called **electromagnetic radiation**. These waves span many orders of magnitude in size, or wavelength, and this entire range of wave sizes is known as the **electromagnetic spectrum** (Figure 2-1).

The energy that drives Earth's climate system occupies only a narrow part of this spectrum. Much of the incoming radiation energy from the Sun consists of visible light at wavelengths between 0.4 and 0.7 µm (1 µm, or micrometer = 1 millionth of a meter), sometimes referred to as **shortwave radiation**. Some ultraviolet radiation from the Sun also enters Earth's atmosphere, but radiation at still shorter wavelengths (X rays and gamma rays, measured in nanometers, or billionths of a meter) does not affect climate.

2-1 Incoming Solar Radiation

The most basic way to look at Earth's climate system is to consider its average properties as a sphere. In this way we reduce its three-dimensional complexities to a single global average value typical of the entire planet, as if we were space travelers looking at Earth from a great distance.

Radiation from the Sun arrives at the top of Earth's atmosphere with an average energy of 1368 watts per square meter (W/m^2). These watts are the same units of energy used to measure the brightness (or more accurately the power) of a household light bulb. If Earth were a flat, one-sided disk directly facing the Sun, and if it had no atmosphere, 1368 W/m^2 of solar radiation would fall evenly across its entire surface (Figure 2-2, top).

But Earth is a three-dimensional sphere, not a flat disk. A sphere has a surface area of $4\pi r^2$ (r being its radius) that is exactly four times larger than the surface area of a flat one-sided disk (πr^2). Because the same amount of incoming radiation must be distributed across this larger surface area, the average radiation received per unit of surface area on a sphere is only one-quarter as strong (1368/4 = 342 W/m^2). Another way of looking at this is to see that half of Earth's rotating surface is dark at any time because it faces away from the Sun at night, while the daytime side, warmed by the Sun, receives radiation at indirect angles except at the one latitude in the tropics where the Sun is directly overhead in any given season (Figure 2-2, bottom).

The 342 W/m^2 of solar energy arrives at the top of the atmosphere, mainly in the form of visible radiation. About 70% of this shortwave radiation passes through Earth's atmosphere and enters the climate system (Figure 2-3). The other 30% is sent directly back out into space after reflection (or scattering) by clouds,

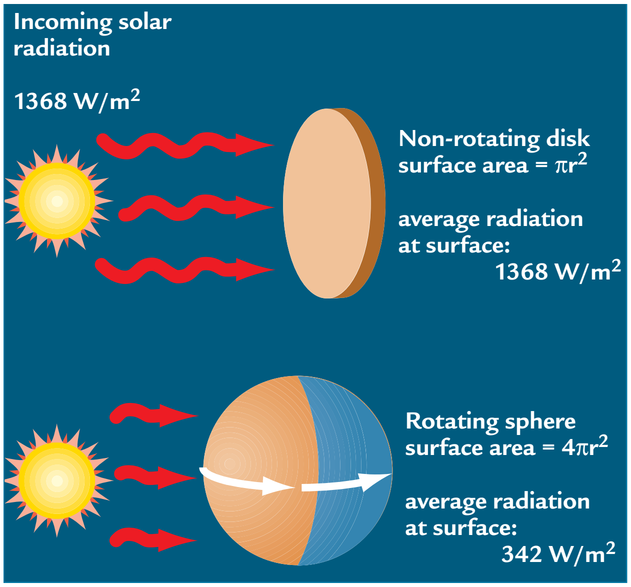


FIGURE 2-2 Average solar radiation on a disk and a sphere The surface of a flat nonrotating disk that faces the Sun (top) receives exactly four times as much solar radiation per unit of area as the surface of a rotating sphere, such as Earth (bottom).

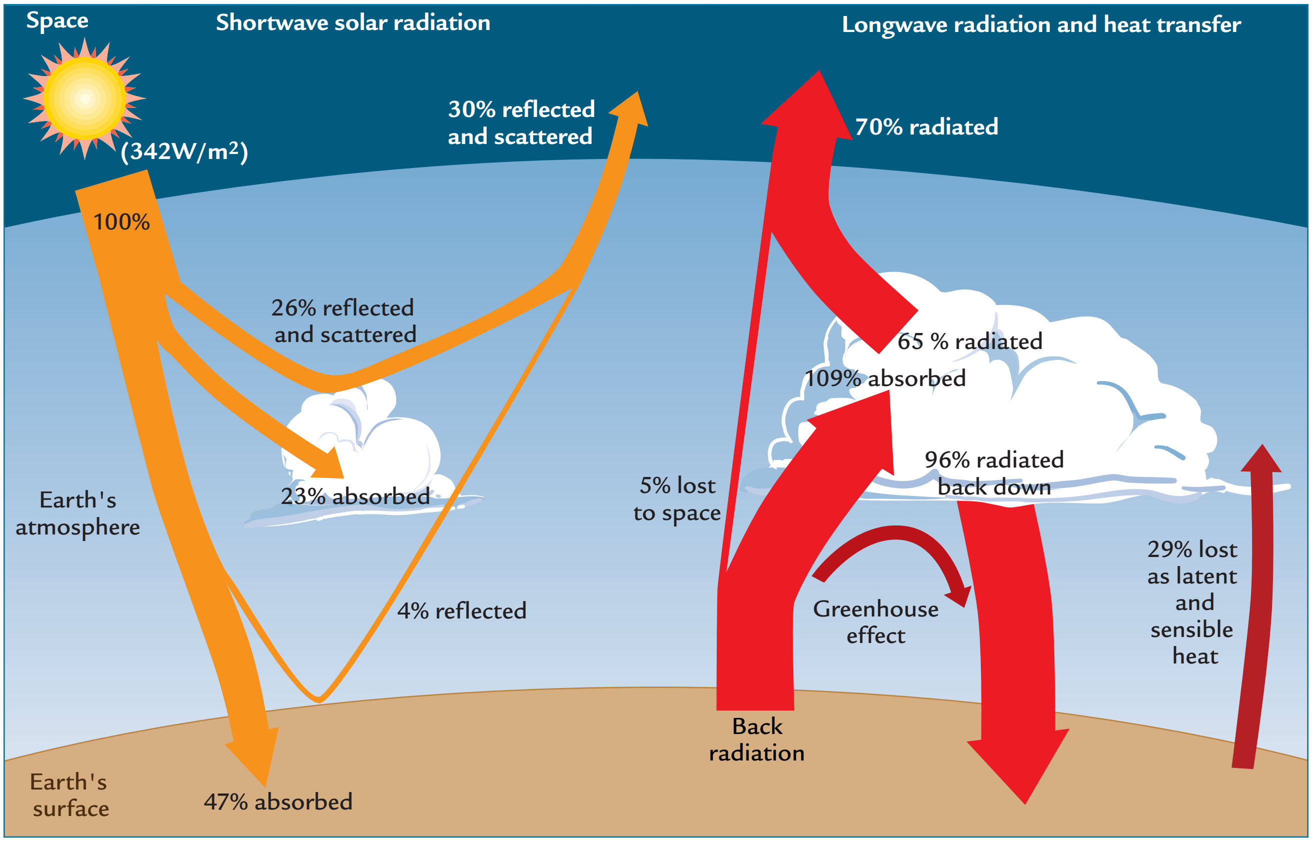


FIGURE 2-3 Earth's radiation budget Solar radiation arriving at the top of Earth's atmosphere averages 342 W/m^2, indicated here as 100% (upper left). About 30% of the incoming radiation is reflected and scattered back to space, and the other 240 W/m^2 (70%) enters the climate system. Some of this entering radiation warms Earth's surface and causes it to radiate heat upward (right). The greenhouse effect (lower right) retains 95% of the heat radiated back from Earth's heated surface and warms Earth by 31°C. (Adapted from T. E. Graedel and P. J. Crutzen, *Atmosphere, Climate, and Change,* Scientific American Library, © 1997 by Lucent Technologies, after S. H. Schneider and R. Londer, *Co-evolution of Climate and Life* [San Francisco: Sierra Club Books, 1984], and National Research Council, *Understanding Climate Change: A Program for Action* [Washington, D.C.: National Academy of Sciences, 1975].)

dust, and the more reflective regions at Earth's surface. As a result, the average amount of solar energy retained by Earth is 240 W/m^2 (0.7 × 342 W/m^2).

Of the 70% of solar radiation that is retained within the climate system, about two-thirds is absorbed at Earth's surface and about one-third by clouds and water vapor in the atmosphere (see Figure 2-3). This absorbed radiation heats Earth and its lower atmosphere and provides energy that drives the climate system.

2-2 Receipt and Storage of Solar Heat

Because Earth is continually receiving heat from the Sun but is also maintaining a constant (or very nearly constant) temperature through time, it must be losing an equal amount of heat (240 W/m^2) back to space. This heat loss, called **back radiation**, occurs at wavelengths lying in the infrared part of the electromagnetic spectrum (see Figure 2-1). Because it occurs at longer wavelengths (5–20 μm) than the incoming shortwave solar radiation, back radiation is also called **longwave radiation**.

Any object with a temperature above absolute zero (−273°C, or 0K) contains some amount of heat that is constantly being radiated away toward cooler regions. Radiated heat can come from such objects as red-hot burners on stovetops, but it is also emitted from objects not warm enough to glow, such as the asphalt pavement that emits shimmering ripples of heat on a summer day.

Although it may seem counterintuitive, even objects with temperatures *far* below the freezing point of water are radiators; that is, they emit some heat. The amount of heat radiated by an object increases with its temperature. The radiation emitted is proportional to T^4, where T is the absolute temperature of the object in Kelvins. Objects with temperatures of −272°C (1K) emit at least a tiny bit of heat energy and so can technically be considered radiators!

What is the average radiating temperature of our own planet? One way to try to answer this question is to average the countless measurements of Earth's surface temperature made over many decades to derive its mean surface temperature. If we do this, we find that the average surface temperature of Earth as a whole is +15°C (288K, or 59°F). This value sounds like a reasonable middle ground between the very large area of hot tropics, averaging 25°–30°C, and the much smaller polar regions, which average well below freezing (0°C). It seems reasonable that the *surface* of planet Earth radiates heat with this average temperature.

But there is a problem with this simple way of looking at radiator Earth. If we now take all available measurements from orbiting satellites and space stations high above Earth, they tell us our planet is sending heat out to space as if it had an average temperature of −16°C (257K or 3°F). This value is 31°C colder than the +15°C average temperature we are certain is correct for Earth's surface. If both sets of measurements are accurate, why are these values so different?

The reason for this discrepancy is the **greenhouse effect** (see Figure 2-3). Earth's atmosphere contains greenhouse gases that absorb 95% of the longwave back radiation emitted from the surface, thus making it impossible for most heat to escape into space. The trapped radiation is retained within the climate system and reradiated down to Earth's surface. This extra heat retained by the greenhouse effect makes Earth's surface temperature 31°C warmer than it would otherwise be.

In effect, measurements made by satellites and space stations in outer space cannot detect the radiation emitted directly from the warmer surface of the Earth because of the muffling effect of the blanket of greenhouse gases and clouds. Instead, most of the heat actually radiated back to space is emitted from an average elevation of 5 kilometers, equivalent to the tops of many clouds—still well within the lowest layer of Earth's atmosphere (Box 2-1). These cold (−16°C) cloud tops emit radiation at an average value of about 240 W/m^2, exactly the level needed to offset the amount of solar radiation retained within Earth's climate system and keep it in balance.

The two main gases in Earth's atmosphere are N_2 (nitrogen) at 78% of the total and O_2 (oxygen) at 21%, but neither is a greenhouse gas because neither traps outgoing radiation. In contrast, the three most important greenhouse gases form very small fractions of the atmosphere. Water vapor (H_2O_v) averages less than 1% of a dry atmosphere, but it can range to above 3% in the moist tropics. Carbon dioxide (CO_2) and methane (CH_4) occur in much smaller concentrations of 0.035% and 0.00018%, but they are also important greenhouse gases.

Clouds also contribute to the retention of heat within the climate system by trapping outgoing radiation from Earth's surface. This role in warming Earth's climate works exactly opposite to the impact of clouds in reflecting incoming solar radiation and cooling our climate. The relative strength of these two competing roles varies with region and season.

Many important characteristics of Earth's climate, such as the amount of incoming sunlight, vary with latitude. Incoming solar radiation is stronger at low latitudes, where sunlight is concentrated more nearly overhead, than at high latitudes, where the Sun's rays strike Earth at a more indirect angle and cover a wider area (Figure 2-4). As a result, larger amounts of solar radiation reach the same unit area of Earth's surface in the tropics than near the poles (Figure 2-5).

This unequal distribution of incoming solar radiation is aggravated by unequal absorption and reflection

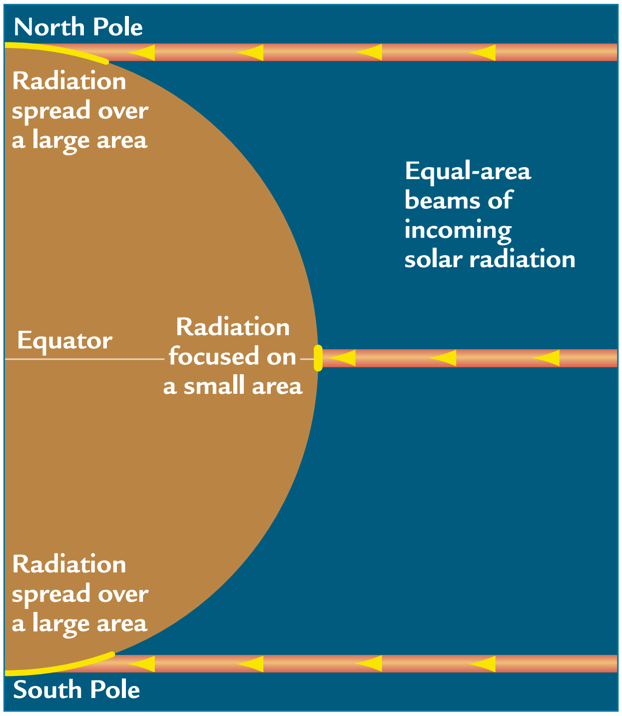


FIGURE 2-4 **Unequal radiation on a sphere** More solar radiation falls on a unit area of Earth's surface near the equator than at the poles because of the more direct angle of incoming radiation. (Adapted from L. J. Battan, *Fundamentals of Meteorology* [Englewood Cliffs, N.J.: Prentice-Hall, 1979].)

BOX 2-1 Looking Deeper into Climate Science

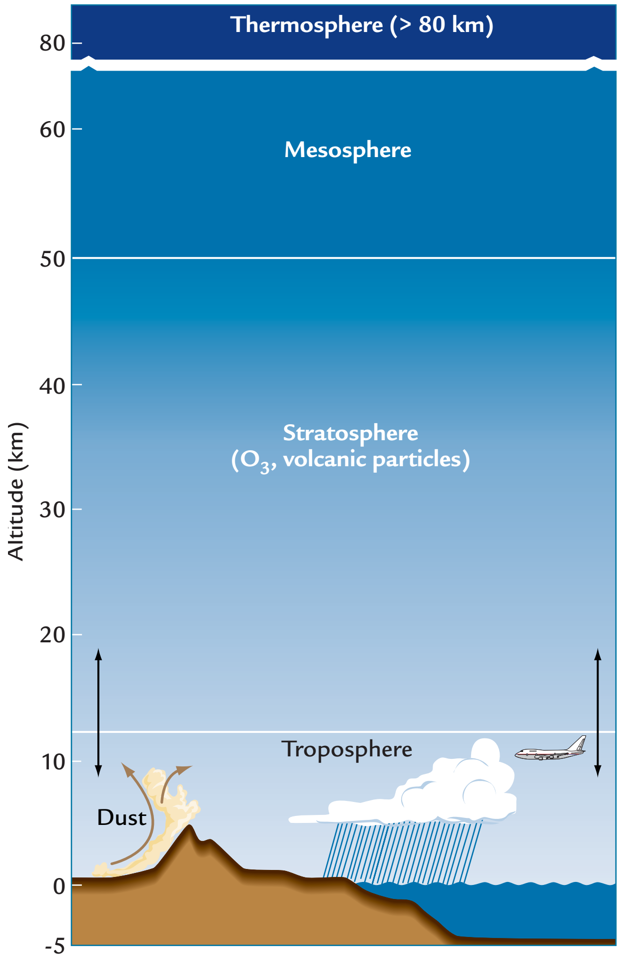
The Structure of Earth's Atmosphere

Earth's atmosphere is divided into four layers that extend far above its surface: the **troposphere,** which extends from the surface to between 8 and 18 kilometers; the **stratosphere,** extending up to 50 kilometers; and the overlying mesosphere (50–80 km) and thermosphere (above 80 km). Only the lowest two are critical in climatic changes.

The troposphere is both the layer within which we live and the layer within which most of Earth's weather happens. Storm systems that produce clouds and rainfall or snowfall are almost entirely confined to this layer. Dust or soot particles that are lifted by strong winds from Earth's surface into the lower parts of this layer are quickly removed by precipitation every few days or weeks. The troposphere is also the main layer within which we measure Earth's climate and its changes, particularly those at Earth's surface. As we will see later, about 80% of the gases that form Earth's atmosphere are contained within the troposphere.

Above the troposphere lies the stratosphere, a much more stable layer almost completely separated from the turbulent storms and other processes so common in the troposphere. Only the largest storms penetrate the stratosphere, and only its lowermost layer. Large volcanic eruptions occasionally throw small particles up out of the troposphere and into the stratosphere. Because no rain or snow falls in most of this layer, it may take years for gravity to pull these particles back to Earth's surface. The stratosphere forms 19.9% of Earth's atmosphere; the troposphere and stratosphere together account for 99.9% of its mass.

The stratosphere is also important to Earth's climate because it contains small amounts of oxygen (O_2) and ozone (O_3), which block ultraviolet radiation arriving from the Sun. This shielding effect accounts for a small fraction of the 30% reduction in incoming heat energy from the Sun. It also greatly reduces the exposure of life-forms on Earth to the harmful effects of ultraviolet radiation, which can cause skin cancers and genetic mutations.



Structure of Earth's atmosphere Most day-to-day weather, as well as important changes in Earth's climate, occur in the lowermost layer of the atmosphere, called the troposphere, which varies in height from 8 to more than 18 km.

by Earth's surface at different latitudes. A smaller fraction of the incoming radiation is absorbed at higher latitudes than in the tropics mainly because (1) solar radiation arrives at a less direct angle (see Figure 2-4) and (2) snow and ice surfaces at high latitudes reflect more radiation (see Figure 2-5).

The percentage of incoming radiation that is reflected rather than absorbed by a particular surface is referred to as its **albedo** (Table 2-1). Snow and ice surfaces at high latitudes have albedos ranging from 60% to 90%, with larger values typical of freshly formed snow and ice, and somewhat lower values for snow or

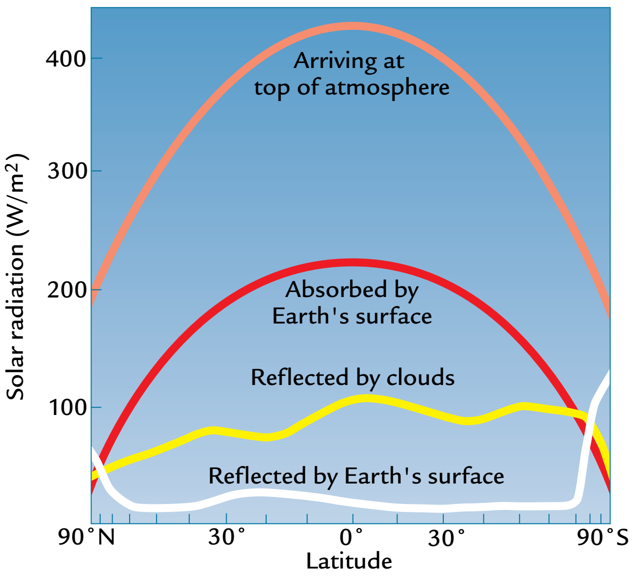


FIGURE 2-5 **Pole-to-equator heating imbalances** Incoming radiation is higher in the tropics than at the poles. Less reflective surfaces at low latitudes absorb a larger fraction of incoming radiation, while highly reflective snow and ice surfaces at the poles reflect more, aggravating the pole-to-equator imbalance in absorbed radiation. (Adapted from R. G. Barry and R. J. Chorley, *Atmosphere, Weather, and Climate*, 4th ed. [New York: Methuen, 1982].)

TABLE 2-1 Average Albedo Range of Earth's Surfaces

Surface	Albedo range (percent)
Fresh snow or ice	60–90%
Old, melting snow	40–70
Clouds	40–90
Desert sand	30–50
Soil	5–30
Tundra	15–35
Grasslands	18–25
Forest	5–20
Water	5–10

Adapted from W. D. Sellers, Physical Climatology *(Chicago: University of Chicago Press, 1965), and from R. G. Barry and R. J. Chorley,* Atmosphere, Weather, and Climate, *4th ed. (New York: Methuen, 1982).*

ice that contain dirt or are partly covered by pools of melted water. In contrast, snow-free land surfaces have much lower albedos (15–30%) and ice-free water reflects even less of the incoming radiation (below 5% when the Sun is overhead).

The albedo of any surface also varies with the angle at which incoming solar radiation arrives. For example, water reflects less than 5% of the radiation it receives from an overhead Sun, but a far higher fraction of the radiation from a Sun lying low in the sky (Figure 2-6). This same tendency holds true for the other surfaces. Because 70% of Earth's surface is low-albedo water, Earth's surface has an average albedo near 10%.

These factors combine to make Earth's surface more reflective near the poles than in the tropics. The Antarctic ice sheet and extensive areas of nearby sea ice in the southern hemisphere have very high albedos, as do the Arctic sea ice cover and the extensive winter

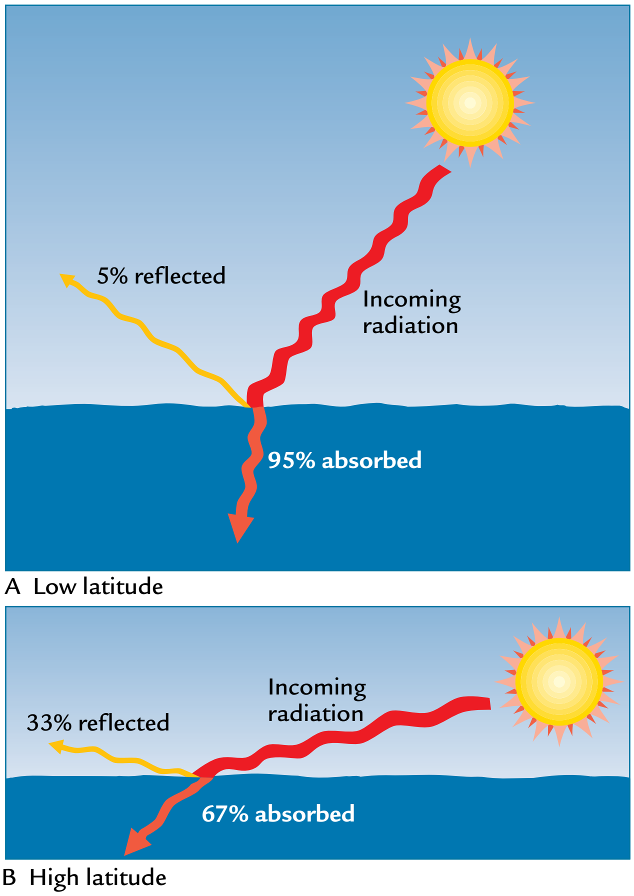


FIGURE 2-6 **Sun angle controls heat absorption** All of Earth's surfaces (here water) absorb more solar radiation from an overhead Sun (A) than from a Sun lying low in the sky (B). (Adapted from L. J. Battan, *Fundamentals of Meteorology* [Englewood Cliffs, N.J.: Prentice-Hall, 1979].)

snow cover in Eurasia and North America in the northern hemisphere. Clouds counteract some of this imbalance in surface albedo by reflecting somewhat more solar radiation at low latitudes than at high latitudes.

The net result is that both polar regions absorb even less of the already low amounts of incoming solar radiation, while the tropics absorb even more of the already-high radiation. The overall effect of these albedo differences is to increase the solar heating imbalance between the poles and the tropics (see Figure 2-5).

Radiation and albedo also vary seasonally. The 23.5° tilt of Earth's axis as it revolves around the Sun causes the northern and southern hemispheres to tilt alternately toward and away from the Sun, and this motion causes seasonal changes in solar radiation received in each hemisphere (Figure 2-7A). From our Earthbound perspective, we experience this orbital motion as a shift of the overhead Sun through the tropics from a latitude of 23.5°N on June 21 to 23.5°S on December 21. This change in the Sun's angle results in large seasonal changes in the amounts of solar radiation (W/m^2) received on Earth (Figure 2-7B).

Accompanying these seasonal changes in solar heating are large seasonal changes in the albedo of Earth's surface with latitude (Figure 2-8). In the south polar region, the ice sheet over Antarctica remains intact through the year, but an extensive ring of sea ice surrounding Antarctica expands and contracts every year across an area of 16 million square kilometers. In contrast, the Arctic Ocean has a multiyear cover of sea ice that fluctuates much less in extent through the year, and the main seasonal albedo change in the northern hemisphere comes from the winter expansion and summer retreat of snow cover on Asia, Europe, and North America. Both of these changes in albedo play important roles in long-term climate change (Box 2-2).

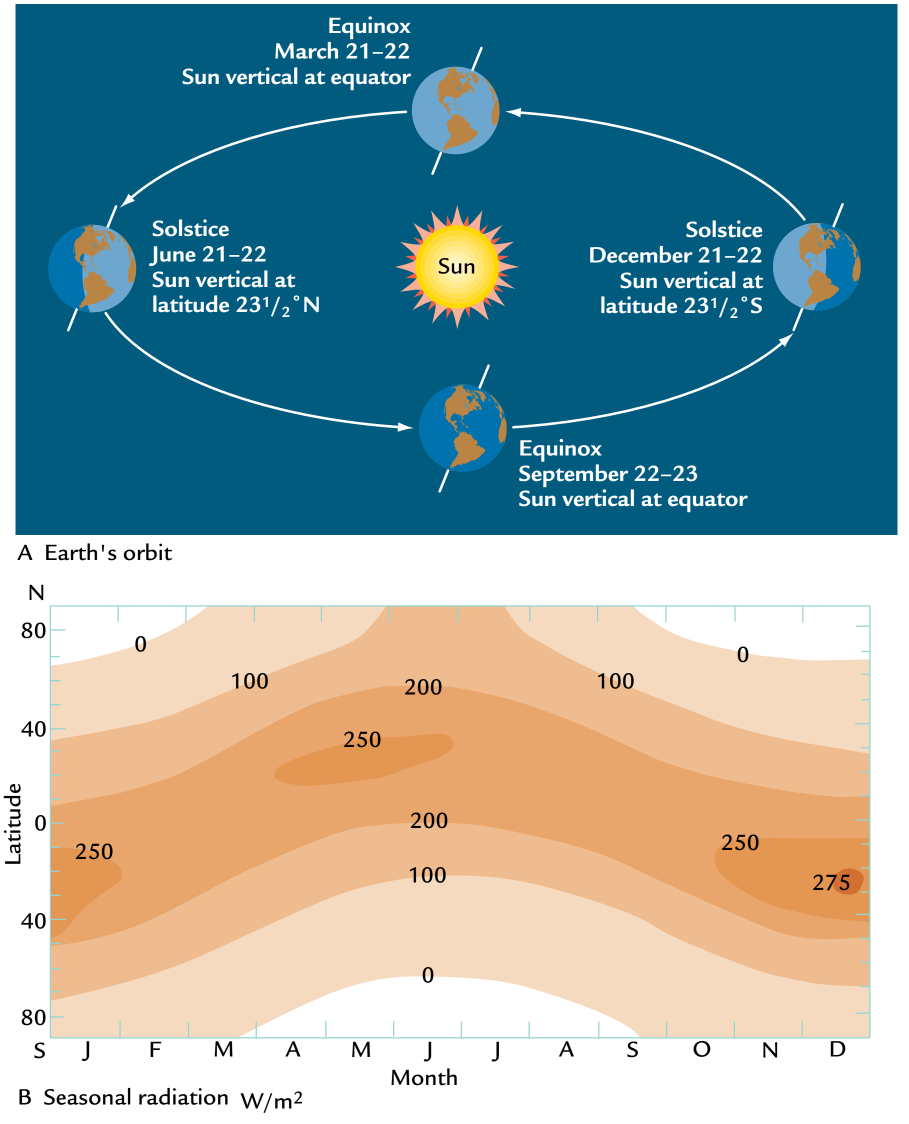


FIGURE 2-7 Earth's tilt and seasonal radiation (A) The tilt of Earth's axis in its annual orbit around the Sun causes the northern and southern hemispheres to lean directly toward and then away from the Sun at different times of the year. (B) This change in relative position causes seasonal shifts between the hemispheres in the amount of solar radiation received at Earth's surface. (A: adapted from F. K. Lutgens and E. J. Tarbuck, *The Atmosphere* [Englewood Cliffs, N.J.: Prentice-Hall, 1992; B: adapted from A. L. Berger, "Milankovitch Theory and Climate," *Reviews of Geophysics* 26 [1988]: 624–57.)

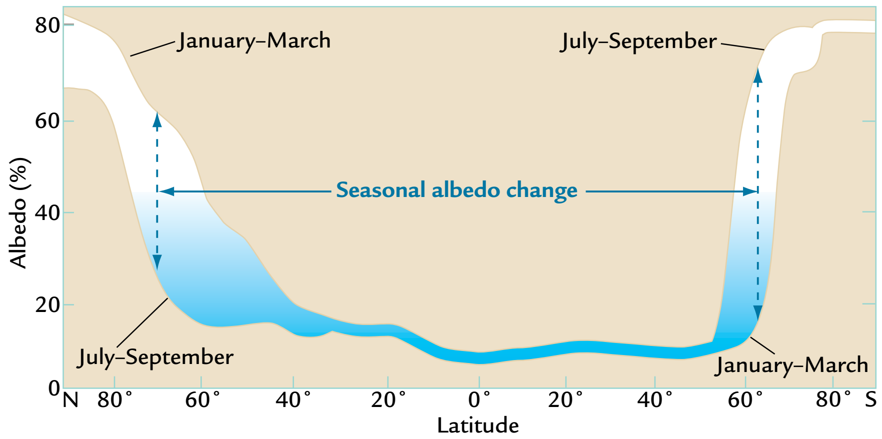
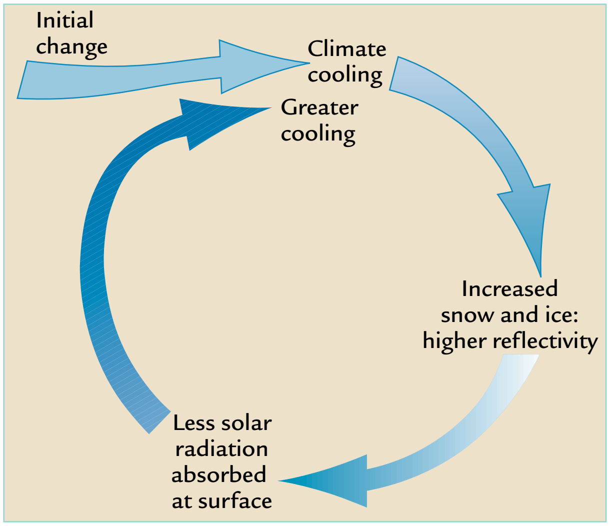


FIGURE 2-8 **Albedo changes with season** Average albedo increases in the northern hemisphere in winter (January–March) mainly because of increased snow cover over land and also because of more extensive sea ice. Albedo increases in the southern hemisphere's winter (July–September) because of more extensive sea ice. (Adapted from G. Kukla and D. Robinson, "Annual Cycle of Surface Albedo," *Monthly Weather Review* 108 [1980]: 56–68.)

BOX 2-2 CLIMATE INTERACTIONS AND FEEDBACKS

Albedo / Temperature

The large difference in albedo between highly reflective snow or ice and land or water surfaces that absorb radiation is important in climate change. Surface albedos can increase by 75% (from 15% to 90%) when snow-free land areas become covered with snow, and over oceans that become covered by sea ice. As a result, a surface that had previously absorbed most incoming radiation will now reflect it away, with significant implications for climate change.



Albedo–temperature feedback When climate cools, the increased extent of reflective snow and ice increases the albedo of Earth's surface in high-latitude regions, causing further cooling by positive feedback. The same feedback process amplifies climate warming.

Assume that climate abruptly cools, for any reason (perhaps a decrease in the output of the Sun). Part of the climate system's natural response to a cooling is an increase in the land area covered by snow and in the ocean area covered by sea ice. The expansion of these light, high-albedo surfaces will cause an increase in the percentage of incoming radiation reflected back out to space, and a decrease in the amount of heat absorbed at the surface.

The loss of absorbed heat in these regions will in turn cause the local climate to cool by an additional amount beyond the initial cooling. This is an example of the concept of positive feedback, introduced in Chapter 1. The positive feedback process also works in the opposite direction: an initial warming will reduce the cover of snow and ice, increase the amount of heat absorbed by exposed land or water, and further warm climate. Climate scientists estimate that any initial climate change will be amplified by about 40% by this positive feedback effect. An initial cooling of 1°C would be amplified to a total cooling of 1.4°C by this process.

Climate scientists call this positive feedback **albedo–temperature feedback.** In a larger sense, its net effect is to increase Earth's overall sensitivity to climate changes. The greater the area on Earth covered by snow and ice, the more sensitive the planet as a whole becomes to imposed changes in climate. Because most of the regional albedo contrast on Earth is localized at the equatorward limit of snow and sea ice, the albedo–temperature feedback most strongly affects climate at higher latitudes near these limits.

Clouds also affect the regional receipt of heat at Earth's surface. Areas such as subtropical deserts that are free of clouds receive far more solar radiation than do most parts of the oceans, especially subpolar oceans where frequent storms produce nearly continuous cloud cover. The cloud cover over tropical rain forests also reduces the receipt of radiation there. But areas without clouds also back-radiate more heat from Earth's surface than do areas with heavy cloud cover.

Water is the key to Earth's climate system (Box 2-3). Absorption and storage of solar heat are strongly affected by the presence of liquid water because of its high **heat capacity,** a measure of the ability of a material to absorb heat. Heat energy is measured in units of **calories** (one calorie is the amount of heat required to raise the temperature of 1 gram of water by 1°C). Heat capacity is the product of the density (in g/cm^3) of a heat-absorbing material and its **specific heat,** the number of calories absorbed as the temperature of 1 gram of this material increases by 1°C:

$$\underset{(\text{cal/cm}^3)}{\text{Heat capacity}} = \underset{(\text{g/cm}^3)}{\text{Density}} \times \underset{(\text{cal/g})}{\text{Specific heat}}$$

The specific heat of water is 1, higher by far than any of Earth's other surfaces. The ratios of the heat capacities of water:ice:air:land are 60:5:2:1. Much of the heat capacity of air is linked to the water vapor it contains. Likewise, much of the heat capacity of land is due to the small amount of water held in the soil.

The low-latitude oceans are Earth's main storage tanks of solar heat. Sunlight penetrates into and directly heats the upper tens of meters of the ocean, especially in the tropics, where the radiation arrives from a Sun high in the sky. Equally important, winds blowing across the ocean's surface stir the upper layers and rapidly mix solar heat as deep as 100 meters (Figure 2-9). In contrast, even though tropical and subtropical landmasses generally become very hot under the strong Sun, they are not capable of storing much heat because heat is conducted down into soil or rock at very slow rates (see

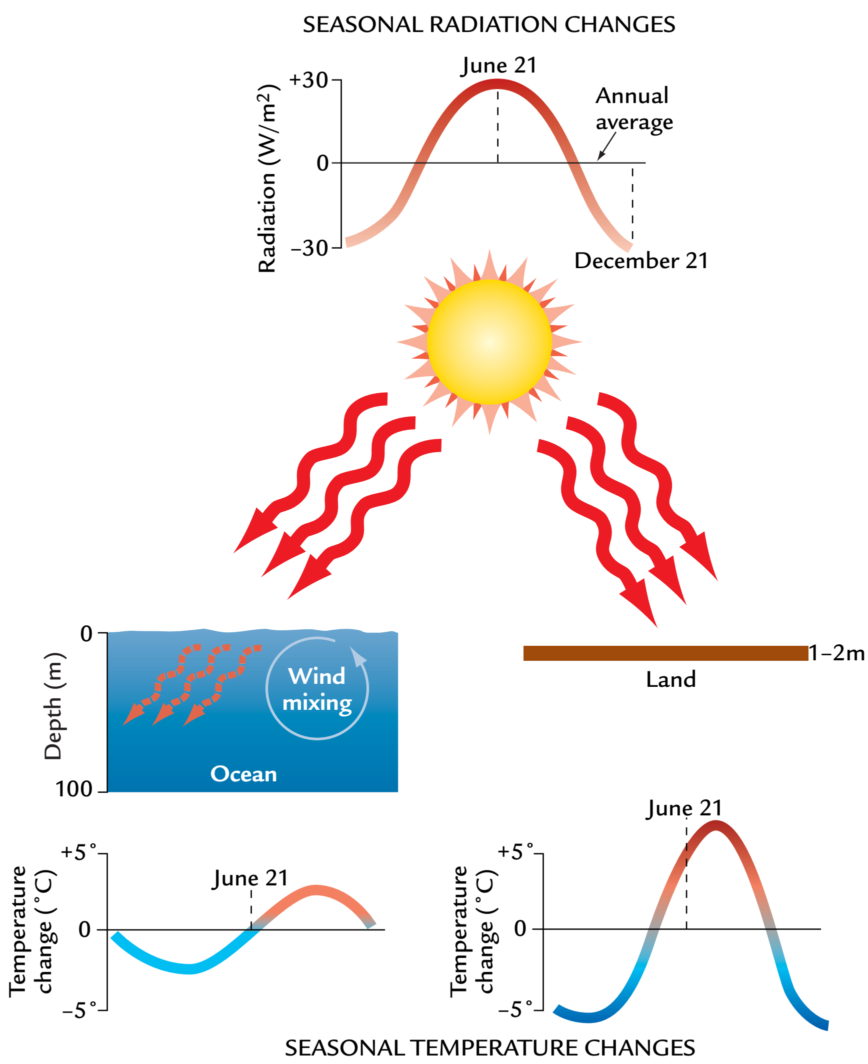


FIGURE 2-9 Difference in heating of land and oceans During the seasonal cycle of solar radiation (top), ocean surfaces heat and cool slowly and only by small amounts because temperature changes are mixed through a layer 100 m thick (lower left). In contrast, land surfaces heat and cool quickly and strongly because of their low capacity to conduct and store heat (lower right). (Adapted from J. E. Kutzbach and T. Webb III, "Late Quaternary Climatic and Vegetational Change in Eastern North America: Concepts, Models, and Data," in *Quaternary Landscapes,* ed. L. C. K. Shane and E. J. Cushing [Minneapolis: University of Minnesota Press, 1991].)

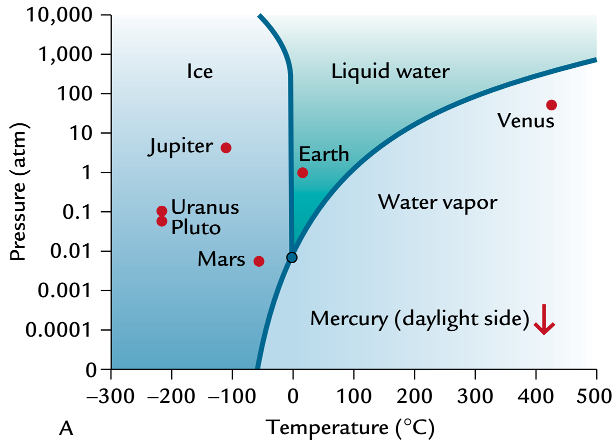
BOX 2-3 Climate Interactions and Feedbacks

Water in the Climate System

Water is a critical part of the miracle of life on Earth. Of all the planets in the solar system, only here on Earth can water exist in three forms: as a liquid (in oceans and lakes), as a frozen solid (ice), and as a gas (water vapor). The essence of our good fortune comes from the fact that water not only accounts for the largest fraction of our bodies and those of most other organisms, but also is the medium we drink for survival, the substance we swim in and skate or ski on for amusement, and part of the air we breathe from day to day without ill effect.

Water is also a vital component of the climate cycle. The oceans contain over 97% of Earth's water, with just 2% held in glacial ice and less than 1% in all the rest of Earth's reservoirs. The **hydrologic cycle,** or continual recycling of water among all these reservoirs, including the much smaller amounts held in the atmosphere, in plants, in lakes, in rivers, and in soil, is vital to the operation of the climate system, as we will see in this and future chapters.

Water and the climate system (A) Only on planet Earth can water exist in the form of a solid (ice), a gas (water vapor), or ordinary liquid water, depending on the prevailing temperature and pressure. (B). Water moves through the climate system in different forms as part of the global water cycle. (D. Merritts et al., *Environmental Geology,* © 1997 by W. H. Freeman and Company.)



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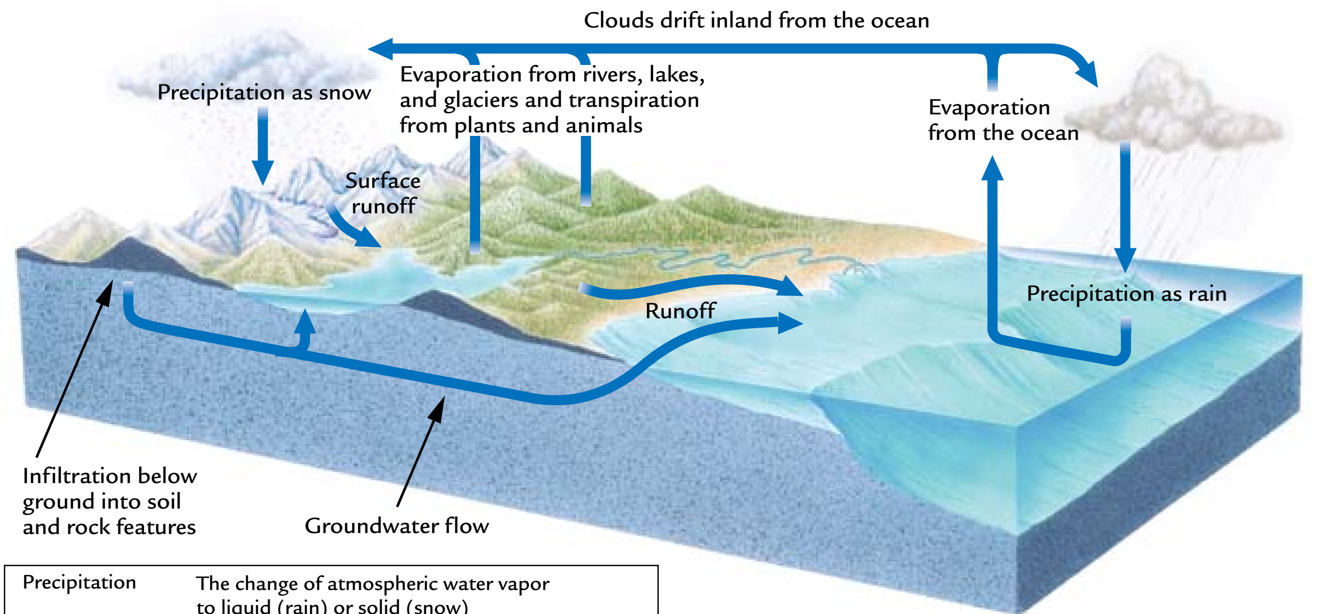


Precipitation	The change of atmospheric water vapor to liquid (rain) or solid (snow)
Evaporation	The exchange of water from a liquid to a vapor
Transpiration	The release of water into the atmosphere by plant and animal cells
Infiltration	The movement of liquid water downward from the land surface into and through the soil and rock
Runoff	The total amount of water flowing in a stream

B

Figure 2-9, bottom). As a result, the large amount of heat stored in the low-latitude ocean provides most of the fuel that runs Earth's climate system.

One consequence of the large difference in heat capacity between land and water is evident in Earth's response to seasonal changes in solar heating: a map showing only the seasonal *range* of temperatures at Earth's surface succeeds in outlining the shapes of most of its continents and oceans (Figure 2-10). The low heat capacity of landmasses allows them to respond strongly to seasonal changes in heating, while the high heat capacity and strong wind mixing of the ocean's upper layers limit its seasonal response (see Figure 2-9, bottom). The largest temperature changes occur over the largest, most continental landmass, Asia, with lesser changes on smaller continents such as Australia (see Figure 2-10). Small seasonal changes also characterize the tropics, which receive consistently strong solar radiation throughout the year and remain warm.

Land and ocean surfaces also differ markedly in their *rates* of response to seasonal heating changes (see Figure 2-9). Interior regions of large continents heat up and cool off quickly because of their low heat capacity, reaching maximum (or minimum) seasonal temperatures about one month after the seasonal radiation maximum (or minimum). In contrast, the upper ocean responds much more slowly because of its high heat capacity, reaching its extreme temperature responses only two or more months after the June 21 and December 21 solar radiation peaks in each hemisphere.

These differences in amplitude and timing of response between land surfaces and the upper ocean layers are referred to as differences in **thermal inertia**. The fast-responding land has a low thermal inertia; the slower-responding upper layers of the ocean have a high thermal inertia.

2-3 Heat Transformation

The heat energy received and stored in the climate system is exchanged among water, land, and air through several processes. As we saw earlier, some of the absorbed heat is lost from Earth's warm surface by longwave back radiation, but most back radiation is trapped by greenhouse gases and radiated back down toward Earth's surface (see Figure 2-3).

Two other important kinds of heat transfer occur within the climate system. One process involves the transfer of **sensible heat** by moving air. Sensible heat is the product of the temperature of the air and its specific heat. It is also the heat that a person directly senses as it is carried along in moving air masses. Surfaces heated

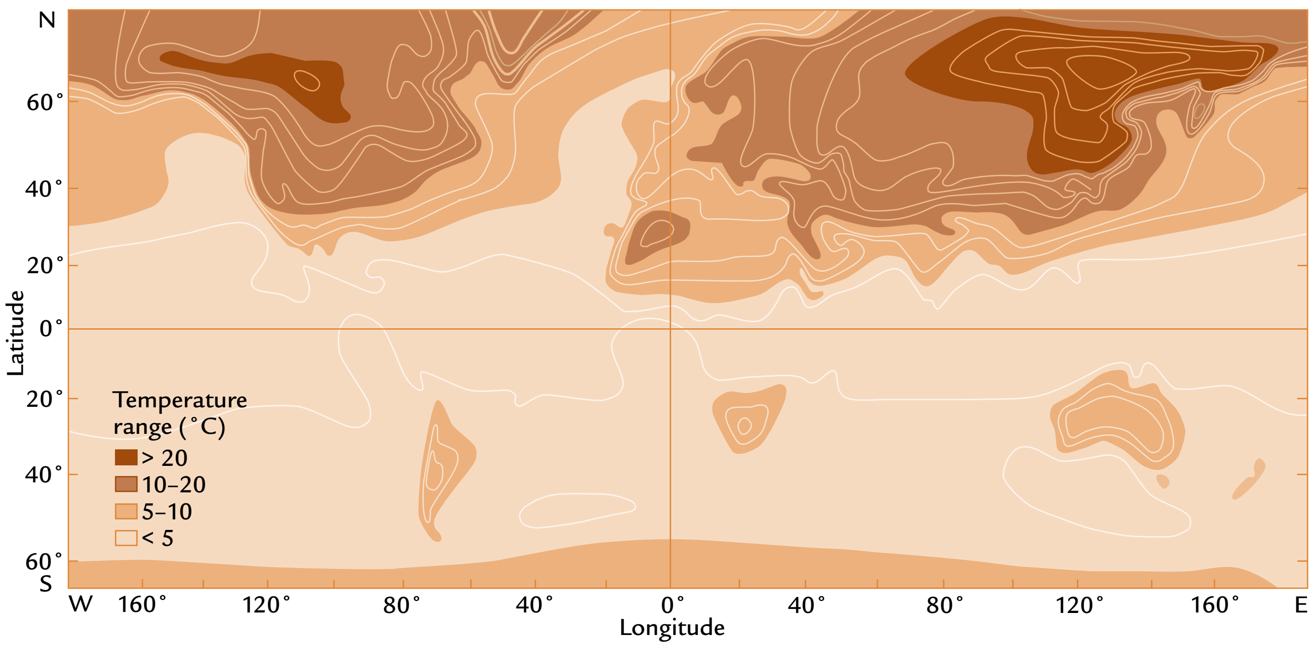


FIGURE 2-10 **Sensitivity of solar heating to land vs. ocean** The total change in mean daily surface temperature between summer and winter is greatest over large landmasses and much smaller over oceans and small continents. The locations of most continents and oceans can be detected on this map from their temperature responses alone. (Adapted from A. S. Monin, "Role of Oceans in Climate Models," in *Physical Basis of Climate and Climate Modeling,* Report no. 16, GARP Publication Series [Geneva: World Meteorological Organization, 1975].)

by the Sun warm the lowermost layer of the overlying atmosphere. The heated air expands in volume like any heated gas, becomes lighter (less dense), rises higher in the atmosphere, and carries sensible heat along with it in a large-scale process known as **convection**.

Convection of sensible heat by air is analogous to what occurs in a pot of water heated from below on a stovetop: the stovetop heating warms the lower layers of water and causes them to expand, rise, and transfer heat upward (Figure 2-11, top). The same process happens when air is heated and rises. Sensible heating at low latitudes is largest over land surfaces in summer, especially in dry regions, because the low water content and small heat conductance of land surfaces allow them to warm to much higher temperatures than the oceans and they transfer this heat to the overlying air (Figure 2-11, bottom). As air and water move horizontally across Earth's surface, they transport sensible heat.

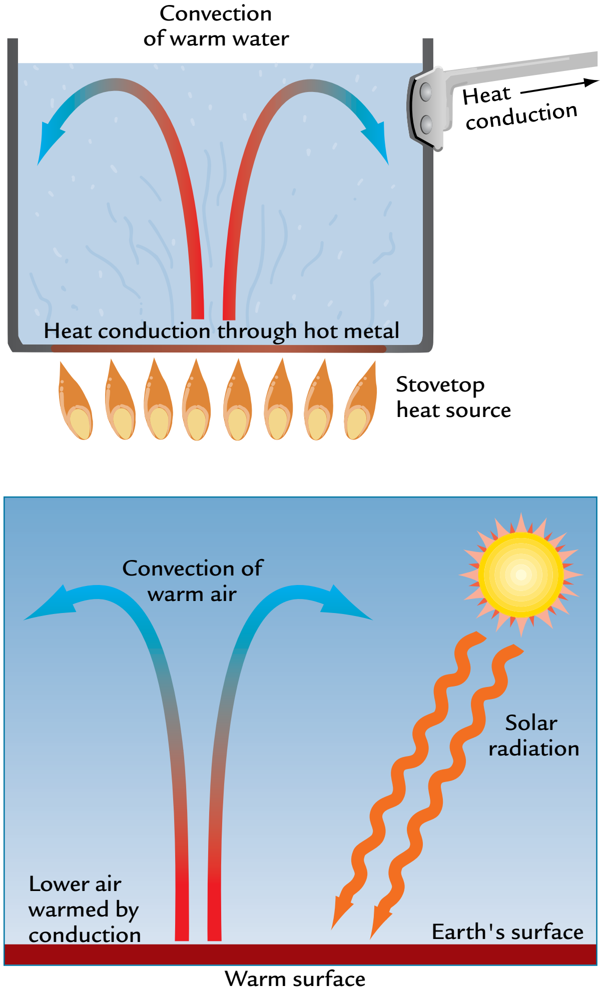


FIGURE 2-11 **Convection of heat** A kettle heated on a stovetop conducts heat to the lower layers of water, which then rise and convect heat upward (top). Similarly, a land surface warmed by the Sun transfers heat by conduction to the lower layers of the atmosphere, which then rise and convect heat upward (bottom).

The second form of heat transfer within the climate system involves the movement of **latent heat.** This more powerful process of heat transfer also depends on the convective movement of air, but in this case the heat carried by the air is temporarily hidden, latent in the water vapor. Transfer of this latent heat occurs in two steps: (1) initial evaporation of water and storage of heat in water vapor, and (2) later release of stored heat during condensation and precipitation, usually far from the site of initial evaporation.

To understand this process, we take a closer look at the behavior of water (Figure 2-12). We learned earlier that it takes 1 calorie of heat energy to raise the temperature of 1 gram of water by 1°C within the range of 0°–100°C. These stored calories are all available to be returned from the water to the air for each 1°C that the water subsequently cools.

But additional changes happen when water changes state, either to ice (a solid) or to water vapor (a gas). Large amounts of latent heat are stored during the warming process that transforms ice to water or water to water vapor, and this stored heat is available for later release during the cooling process that transforms water vapor back to water or water back to ice (see Figure 2-12).

When ice warms to a temperature of 0°C, any addition of heat (usually from the atmosphere) causes it to melt. Melting of solid ice to form liquid water requires a large input of energy (80 calories per gram of ice). During the melting process, the temperature of the ice holds at 0°C, because all the additional energy is being used for melting.

The opposite happens when water freezes: caloric energy (heat) is liberated. Water chilled to 0°C by loss of heat to the atmosphere begins to freeze if still more heat is extracted. The freezing process liberates the same 80 calories of heat energy per gram of water that had been stored in the water as the ice melted. Because the heat energy liberated during the freezing process was in effect hidden in the water, this released energy is called the **latent heat of melting**.

A similar process governs transitions between liquid water and water vapor (see Figure 2-12). Turning liquid water heated to 100°C into water vapor (gas) requires the input of 540 calories per gram of water, again with no change in temperature during the change of state. Condensation of water vapor back to liquid water in clouds or fog liberates the same amount of energy, called the **latent heat of vaporization**.

Despite what Figure 2-12 implies, water does not have to boil to be turned into vapor. Evaporation occurs across the entire range of temperatures at which liquid water exists, including all the temperatures of the surface ocean and the lower atmosphere. As with boiling,

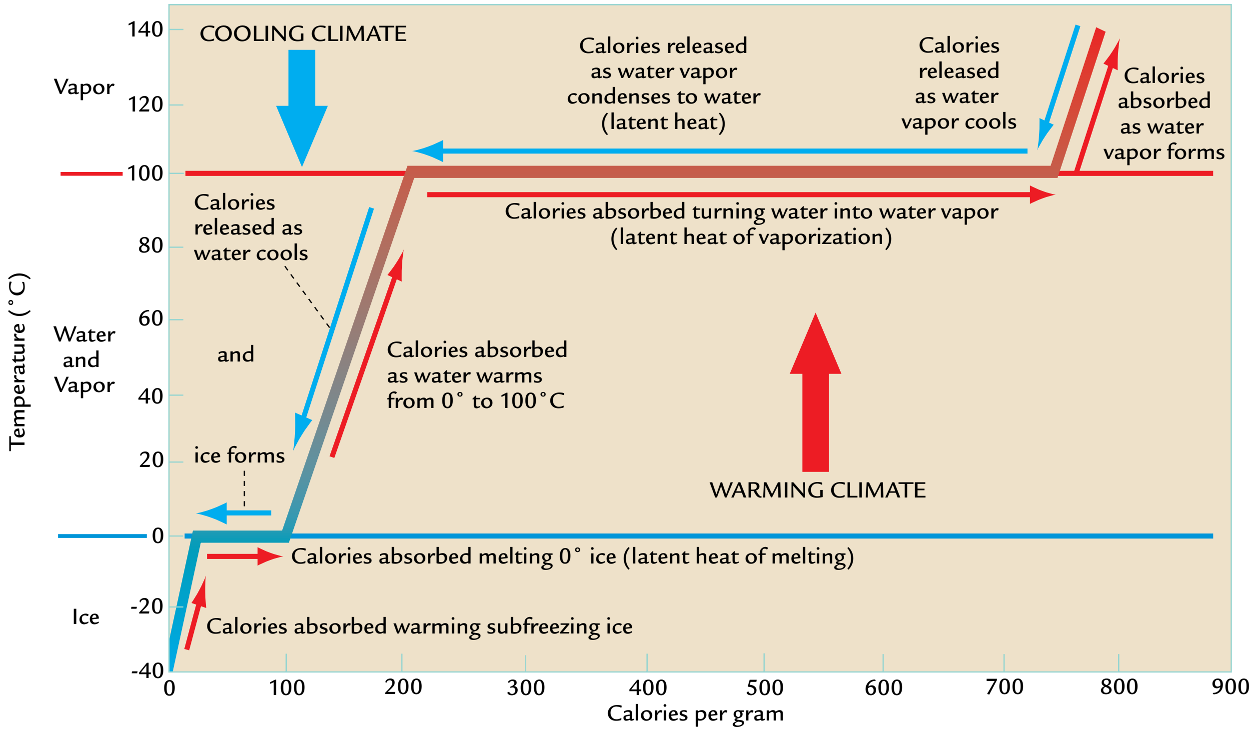


FIGURE 2-12 Heat transformations Heat calories are absorbed from the atmosphere when ice, water, or water vapor is warmed. This heat is released back to the atmosphere during cooling. Heat is also absorbed and stored during changes of state from ice to liquid water or from liquid water to water vapor and this stored latent heat is available for release when water vapor condenses to water or when water freezes to form ice. (Adapted from H. V. Thurman, *Introduction to Oceanography,* 6th ed. [New York: Macmillan, 1991].)

the change of state from liquid to vapor at lower temperatures requires the addition of a large amount of heat energy, and this energy is stored in the water vapor in latent form for later release.

The amount of water vapor that can be held in air is limited, much like the amount of sugar that can be dissolved in coffee. Attempts to add more water vapor to fully saturated air will cause condensation (the formation of dew or raindrop nuclei at the so-called **dew point** temperature). This limit of full saturation, measured in grams of water per cubic meter of air and called the **saturation vapor density**, roughly doubles for each 10°C increase in air temperature (Figure 2-13). Warm tropical air at 30°C can hold almost ten times as much water vapor as cold polar air masses near or below 0°C. As a result of this relationship, water vapor is an important positive feedback in the climate system (Box 2-4).

Evaporation of water from Earth's surface in warmer regions stores excess heat energy in the warm atmosphere. This energy stored in water vapor is carried along with the moving air, both vertically and horizontally. When condensation and precipitation occur, the stored latent energy is released as heat, far from the site of evaporation. The average parcel of water vapor stays in the air for 11 days and travels over 1000 kilometers, about the distance from Los Angeles to Denver.

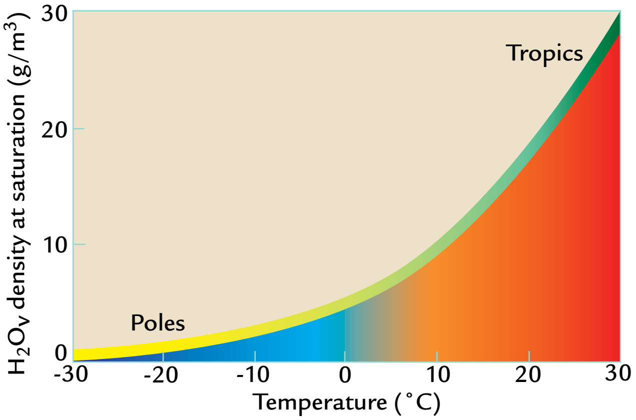


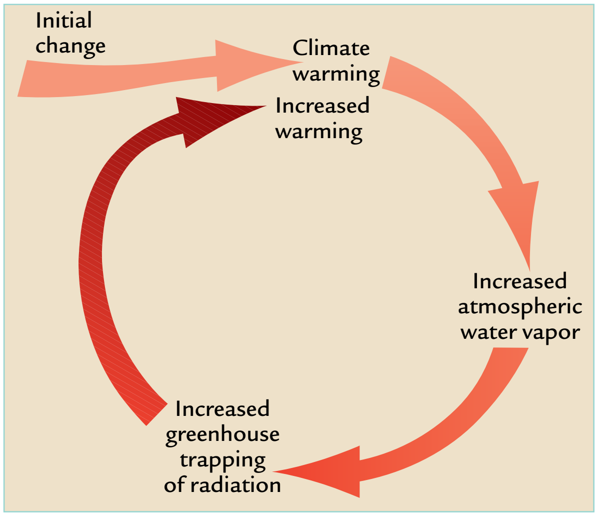
FIGURE 2-13 Water vapor content of air Warm air is capable of holding almost ten times as much water vapor (H_2O_v) as cold air. (Adapted from L. J. Battan, *Fundamentals of Meteorology* [Englewood Cliffs, N.J.: Prentice-Hall, 1979].)

BOX 2-4 Climate Interactions and Feedbacks

Water Vapor

Water vapor, Earth's major greenhouse gas, varies in concentration from 0.2% in very dry air to over 3% in humid tropical air. This strong dependence on temperature produces an important positive feedback in the climate system called **water vapor feedback.**

Assume that climate warms for any reason. A warmer atmosphere can hold more water vapor, and the increased greenhouse gas traps more heat. This large greenhouse effect in turn further warms Earth, amplifying the initial warming through a positive feedback loop. The same positive feedback process works when the climate cools: initial cooling reduces the amount of water vapor held in the atmosphere and produces additional cooling because of the reduced greenhouse effect. It is estimated that direct positive feedback from water vapor can triple the size of an initial climate change. This estimate is based on the action of water vapor in a clear sky; it ignores the more complicated effects that occur when water vapor condenses and forms clouds.



Water vapor feedback When climate warms, the atmosphere is able to hold more water vapor (the major greenhouse gas in the atmosphere), and the increase in water vapor leads to further warming by means of a positive feedback. This feedback works in reverse during cooling.

Heat Transfer in Earth's Atmosphere

Tropical and subtropical latitudes below 35° have a net excess of incoming solar radiation over outgoing back radiation, while at latitudes higher than 35° the opposite is true (Figure 2-14). Most of this excess heat is stored within a thin upper layer of the tropical ocean, and it is this heating imbalance that drives the general circulation of the Earth's atmosphere and oceans. Roughly two-thirds of the net transport of heat from the Earth's equator toward its poles occurs in the lower atmosphere.

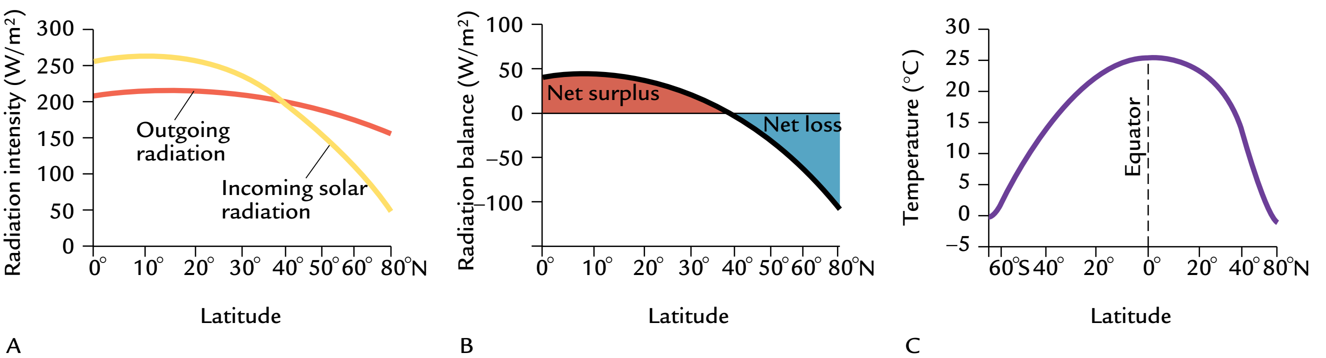


FIGURE 2-14 **Unequal heating of tropics and poles** Incoming solar radiation is strongest near the equator, while radiation emitted back to space is more evenly distributed between the tropics and poles (A). The resulting radiation surplus in the tropics and deficit at the poles (B) creates temperature imbalances (C) that drive the circulation of the atmosphere and oceans. (D. Merritts et al., *Environmental Geology,* © 1997 by W. H. Freeman and Company.)

2-4 Overcoming Stable Layering in the Atmosphere

Both sensible and latent heating are associated with upward motion or convection in the atmosphere. But how high does this convected air go, and why? Air is highly compressible, and most of the mass of the atmosphere is held close to Earth's surface by gravity, with 50% of the air molecules below 7 kilometers and 75% below 10 kilometers (Figure 2-15). Under the pull of gravity, each layer of the atmosphere presses down on the layers below, compressing them and increasing their density. As a result, atmospheric pressure—pressure exerted by the weight of the overlying column of air—increases toward the lower elevations.

Opposing this tendency of dense air to be held close to Earth's surface by gravity is a natural tendency of air to flow from the high pressures near Earth's surface to the lower pressures at higher elevations. In effect, the compressed air at lower elevations has nowhere to go and so it pushes back against the overlying layers. These two opposing forces, the downward pull of gravity and the resistance directed upward, tend to remain in a stable but delicate balance that limits the amount of vertical air motion.

Over limited areas, however, parcels of air will rise if they become less dense than the surrounding air. The major way this process occurs is by warming of Earth's surface and its lower atmosphere. Heating causes the lower layers of air to expand, become lower in density, and rise buoyantly in the atmosphere (like a balloon full of hot air).

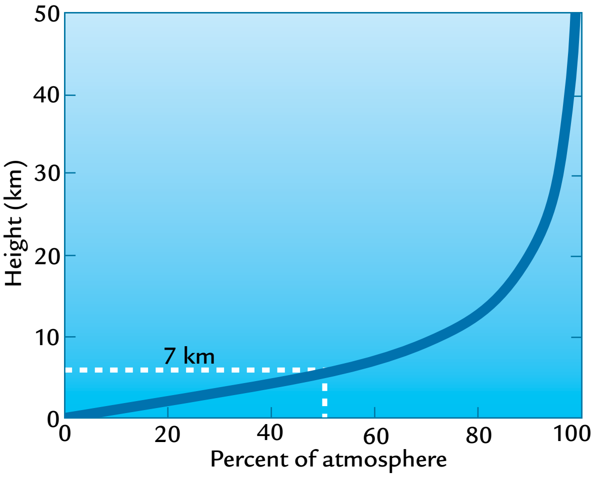


FIGURE 2-15 Distribution of air with elevation Each layer of air in Earth's atmosphere presses down on the underlying layers, increasing the pressure on the lower layers and at the surface. Most of the mass of Earth's atmosphere lies at lower elevations. (Modified from R. G. Barry and R. J. Chorley, *Atmosphere, Weather, and Climate,* 4th ed. [New York: Methuen, 1982].)

As these parcels of heated air rise to higher elevations, processes come into play that involve changes in temperature and in the density of the moving air due to changes in pressure. Climate scientists refer to these changes as **adiabatic** processes. In the case of rising parcels of air, the lower pressures encountered at higher elevations result in an additional expansion of the air beyond the amount originally caused by heating at Earth's surface. But expansion requires the expenditure, or loss, of heat energy, and no source of heat exists to replace the lost heat. As a result, the rising air parcels begin to cool and increase in density, especially in comparison with the surrounding air that is not rising. Eventually the parcel stops rising at the level where its density matches that of the surrounding air. This upward loss of heat is a dry process that is independent of the amount of water vapor in the air.

Latent heating is the second process that can destabilize the atmosphere, and it occurs as a wet process driven by water vapor, which weighs roughly a third less than the mixture of gases that form Earth's atmosphere. Evaporation adds water vapor to the atmosphere at low elevations and causes a net decrease in the density of the air, which can then rise to higher elevations. As before, the moist air rises, expands, and at some point cools to the temperature at which it becomes fully saturated with water vapor. Then condensation begins. Condensation releases latent heat, which partially opposes the cooling of the rising air parcel due to its expansion. The release of this heat in the air parcels also makes them more buoyant (less dense) and allows them to rise much higher in the troposphere (sometimes to 10 to 15 km). Eventually the cumulative loss of most of the water vapor in the parcels stops the release of latent heat that had kept them rising.

The rate at which Earth's atmosphere cools with elevation is called the **lapse rate**. This rate ranges from 5°C/km to as high as 9.8°C/km, but typically averages 6.5°C/km both at middle latitudes and for the planet as a whole. Localized lapse rates are highest in very dry regions where the only factor at work on rising air parcels is adiabatic expansion and cooling. For the areas of Earth's surface where rising air carries water vapor, the release of latent heat at higher elevations warms the air at upper levels and reduces the lapse rate to values near the lower end of this range.

2-5 Tropical–Subtropical Atmospheric Circulation

The large-scale circulation in Earth's atmosphere that transports heat from low to high latitudes is summarized in Figure 2-16. The red arrows on the left show

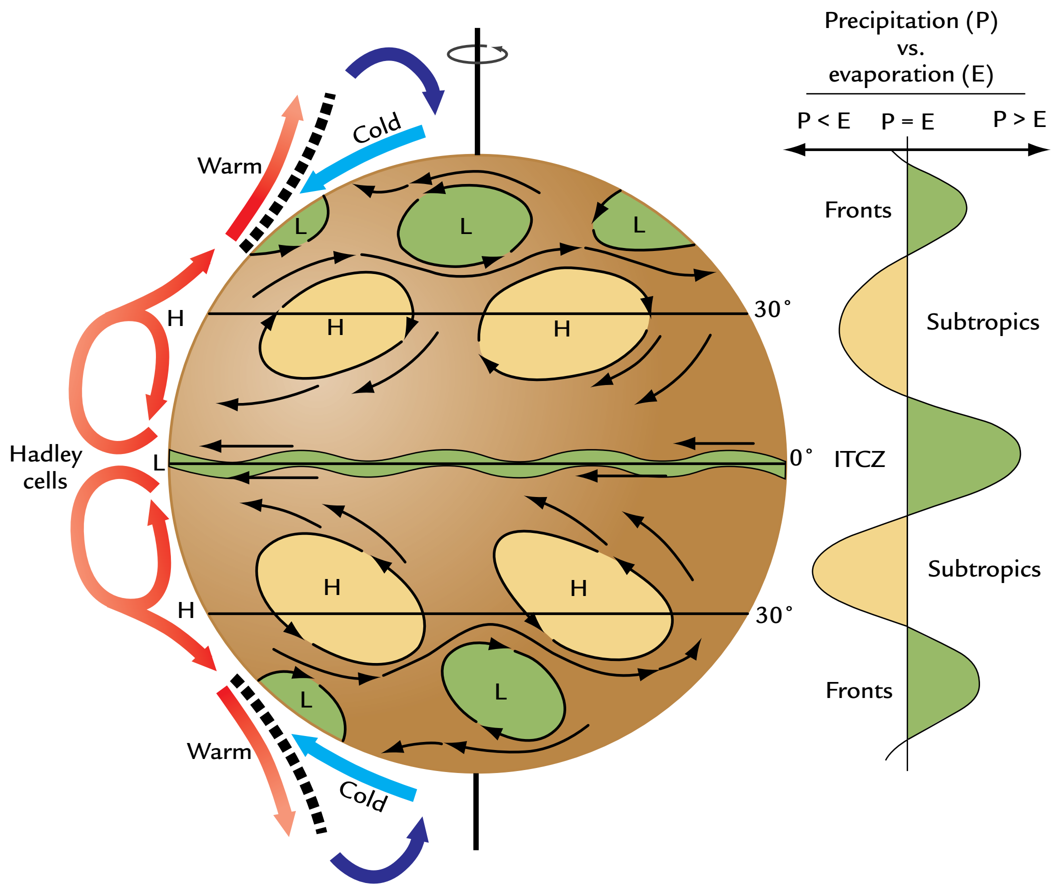


FIGURE 2-16 **General circulation of the atmosphere** (Left) Heated air rises in the tropics at the intertropical convergence zone (ITCZ) and sinks in the subtropics as part of the large-scale Hadley cell flow, which transports heat away from the equator. Additional poleward heat transfer occurs along moving weather systems (fronts) at middle and higher latitudes, with warm air rising and moving poleward and cold air sinking and moving equatorward. (Right) Rising air in the tropics causes a net excess of precipitation over evaporation, while dry air sinking in the subtropics produces more evaporation than precipitation. Higher latitudes tend to have small excesses of precipitation over evaporation. (Adapted from S. H. Schneider and R. Londer, *Co-evolution of Climate and Life* [San Francisco: Sierra Club Books, 1984]; E. Bryant, *Climate Process and Change* [Cambridge: Cambridge University Press, 1998]; and J. P. Peixoto and M. A. Kettani, "The Control of the Water Cycle," *Scientific American,* April 1973.)

heat being transported by warm air, with blue arrows indicating movement of cold air. The large-scale patterns of precipitation and evaporation resulting from these air motions are summarized on the right. This summary figure is important to the discussion that follows, so refer to it often as we go along.

Tropical heating drives a giant tropical circulation pattern called the **Hadley cell** (see Figure 2-16, left). Warm air rises in giant columns marked by towering, puffy (cumulonimbus) clouds created by evaporation of water vapor from tropical oceans and subsequent condensation at high altitudes. Condensation produces a narrow zone of high rainfall in the rising part of the Hadley cell near the equator. The rising motion in the tropical part of the Hadley cell represents an enormous transfer of heat through the atmosphere from low to high altitudes.

Air parcels that rise and lose water vapor in the tropics move toward the subtropics in both hemispheres, transporting sensible heat and other energy from lower to higher latitudes. In the subtropics, this air sinks toward the surface near 30° latitude. The sinking air is then warmed by the increasing pressure of the atmosphere at lower elevations (another adiabatic process), and it gradually becomes even drier and able to hold still more water vapor. This Hadley cell flow prevents condensation from occurring in much of the subtropics and makes these latitudes a zone of low average precipitation and high evaporation, in regions such as the Sahara Desert.

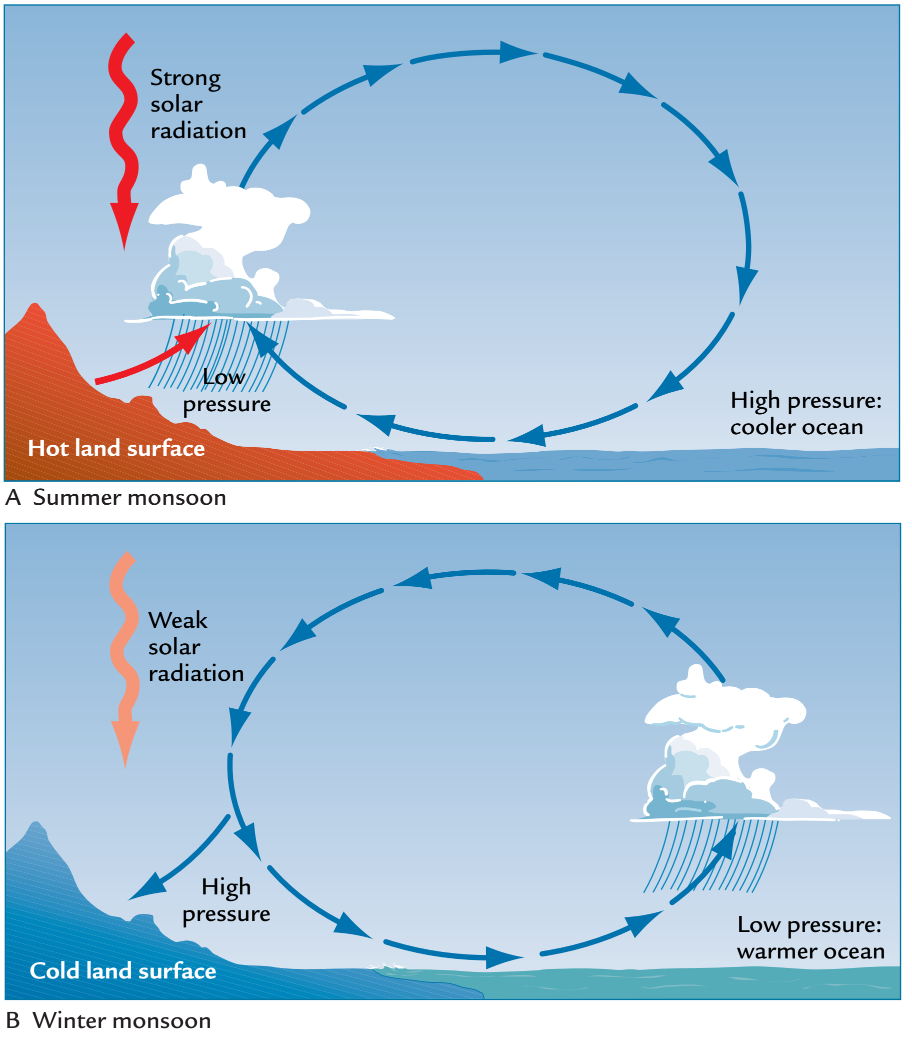
The Hadley cell circulation is completed at Earth's surface, where trade winds from both hemispheres blow from the subtropics toward the tropics and replace the rising air. As warm dry air carried by the trade winds passes over the tropical ocean, it continually extracts water vapor from the sea surface. The region near the equator where the northern and southern trade winds meet is called the **intertropical convergence zone (ITCZ)**. Water vapor carried by the trade winds contributes to the rising air motion and abundant rainfall along the ITCZ.

Viewed on a daily or weekly basis, the actual circulation at low latitudes consists of small-scale cloud systems that develop explosively in limited regions. But when the circulation is averaged over seasons or years, it takes the form shown by the schematic Hadley cell in Figure 2-16. This flow is an important part of the poleward transfer of heat, with a net upward movement and release of latent heat in the tropics followed by a net horizontal transfer of sensible heat to the subtropics at high elevations.

These large-scale movements of large masses of air alter the pressure (weight of air) at Earth's surface. Air movement upward and away from the tropics reduces the weight of the column of overlying air and produces low surface pressures near the ITCZ. Air moving into the subtropics and then down toward Earth's surface increases the pressure there.

Because solar heating is the basic driving force behind the Hadley cell circulation, the seasonal shifts of the Sun between hemispheres also affect the location of the ITCZ. It moves northward during the northern hemisphere's summer (June to September) and southward during the southern hemisphere's summer (December to March). The slow thermal response of the land and oceans causes the seasonal shifts of the

FIGURE 2-17 **Monsoonal circulations** (A) In summer, more rapid heating of land surfaces than of the ocean produces rising motion over the continents and draws moist air in from the ocean, producing precipitation over land. (B) In winter, more rapid cooling of the land surfaces than of the ocean produces sinking motion over the continents and sends cold dry air out over the warmer ocean, shifting most winter precipitation out to sea.



ITCZ to lag more than a month behind those of the Sun.

Important seasonal transfers of heat between the tropical ocean and land, called **monsoons**, arise from the fact that water responds more slowly than land to these seasonal changes in solar heating because of its larger heat capacity and high thermal inertia. The *summer monsoon* circulation is basically an in-and-up flow of moist air that produces precipitation. The strong, direct solar radiation in summer at low and middle latitudes heats Earth's surface (Figure 2-17A). Because soil contains relatively little water, land surfaces have low thermal inertia and heat up quickly. The ocean, with its much higher thermal inertia, absorbs the heat, mixes it through a layer up to 100 meters thick, and warms up far more slowly and to a smaller extent than the land.

These different responses to solar heating set in motion a large-scale land-sea circulation, the monsoon. Initially, dry air heated rapidly over the continental interior rises, and the upward movement of this mass of air produces a region of low pressure over the land. Air is then drawn in toward the low-pressure region from the cooler oceans. The moist air coming in from the oceans is slowly heated and joins in the prevailing upward motion.

As the moist air rises, it cools and its water vapor condenses. Condensation produces heavy precipitation and releases substantial amounts of latent heat, which fuels an even more powerful upward motion (see Figure 2-17A). This net in-and-up circulation in summer monsoons is a two-stage process: initially a dry process due to the rising of sensible heat, and later a wet process linked to ocean moisture and release of latent heat.

The strongest summer monsoon circulations on Earth today occur over India. Heating of the large high landmass of southern Asia focuses a strong wet summer monsoon against the Himalaya Mountains (see Chapter 7).

The *winter monsoon* circulation is the reverse of the summer monsoon. The basic flow is a down-and-out motion of cold, dry air from land to sea (Figure 2-17B). In winter the Sun's radiation is weaker, and land surfaces cool by back radiation. Because of differences in thermal inertia, land surfaces cool faster and more intensely than the oceans. Air cooled over the land sinks toward the surface and creates a region of high pressure where the extra mass of air piles up. Air flows outward from this cell toward the oceans at lower levels. Because the sinking air holds little moisture, the outflow from the continents to the oceans is cold and dry.

These near-surface monsoon circulations are part of a much larger circulation that links the subtropical continents and oceans (Figure 2-18). In summer the upward motion prevailing over land produces an excess of atmospheric mass (and high pressures) near 10 kilometers of altitude. This high pressure results in a high-level flow of air out toward the adjacent oceans, and this flow produces regions of high pressure in the lower atmosphere over the subtropical ocean in summer (Figure 2-18A).

In winter the slow thermal response of the oceans keeps them relatively warm, and they provide heat to the cold air flowing out from the land. The heat gained by the atmosphere results in rising motion and increased precipitation, and the upward movement of air produces strong low-pressure cells over the oceans at higher latitudes (Figure 2-18B).

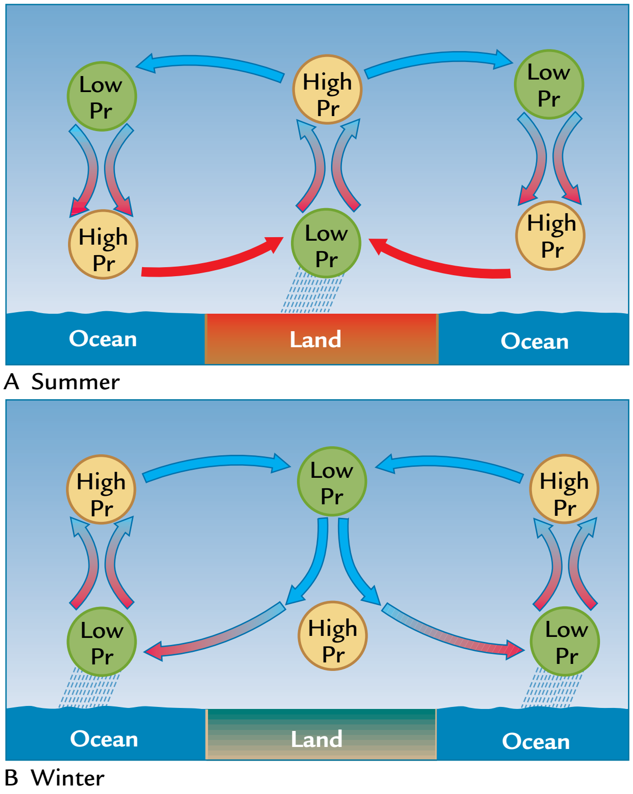


FIGURE 2-18 **Large-scale monsoon circulations** Air motion associated with monsoon circulations at larger scales is upward over the land and downward over the ocean in summer, but the exact reverse in winter. Precipitation is heaviest in regions of low pressure (Pr) and upward motion. (Adapted from J. E. Kutzbach and T. Webb III, "Late Quaternary Climatic and Vegetational Change in Eastern North America: Concepts, Models, and Data," in *Quaternary Landscapes,* ed. L. C. K. Shane and E. J. Cushing [Minneapolis: University of Minnesota Press, 1991].)

2-6 Atmospheric Circulation at Middle and High Latitudes

The giant Hadley cells are a simple and convenient summary of basic atmospheric circulation across that half of Earth's surface area lying between 35°S and

BOX 2-5 Looking Deeper into Climate Science

The Coriolis Effect

The direction of movement of fluids (air and water) is complicated by Earth's rotation. One way to visualize this effect is to contrast the actual motion of a person during a single day at either pole in comparison with that of a person on the equator. A person standing exactly at one of the poles spins around in place once each day, without ever moving from that point. In contrast, a person standing on the equator and facing east (the direction of Earth's rotation) zooms through space at 500 meters per second around the 40,000 km of Earth's circumference once each day, all the while facing east and not spinning at all. Earth's rotation accounts for this shift from a spinning motion at the poles to a one-way trip through space at the equator.

To appreciate the effect Earth's rotation has on moving objects, we can track the movement of a small airplane across the northern hemisphere. If the plane sets out from a point P on Earth's surface and flies in a straight line in relation to the stars, it will appear to an Earthbound observer located under the plane at point P_1 to be moving toward the northeast. Several hours later, with the plane still moving in exactly the same direction *in relation to the stars,* Earth's coordinate system (its reference directions of north, south, east, and west) will have rotated out from under the airplane, and the airplane will now appear from an Earthbound perspective at point P_2 to be moving to the southeast rather than the northeast. In effect, the rotation of Earth's coordinates out from under the plane makes the plane appear to have turned to the right, *although it has not really turned at all.*

Earth's rotation has the same effect on moving air and water. Air naturally tends to flow from regions of high to low pressure, but rotation causes its motion to appear to be deflected to the right (again, from an Earthbound perspective). This *apparent* deflection is called the **Coriolis effect,** after the French engineer who first described it.

In the northern hemisphere, air moving from a zone of high pressure to low pressure always appears to be deflected to the right. This deflection causes air moving outward from a high-pressure cell to acquire a net clockwise spin, and air moving in toward a low-pressure region acquires a counterclockwise spin for the same reason.

Both the direction of deflection and the spinning of air around the highs and lows are exactly reversed in the southern hemisphere because the direction of Earth's rotation viewed from the perspective of the South Pole is exactly opposite that of the view from the North Pole. At the equator, where Earth's rotation has no net spinning motion, the Coriolis effect drops to zero.

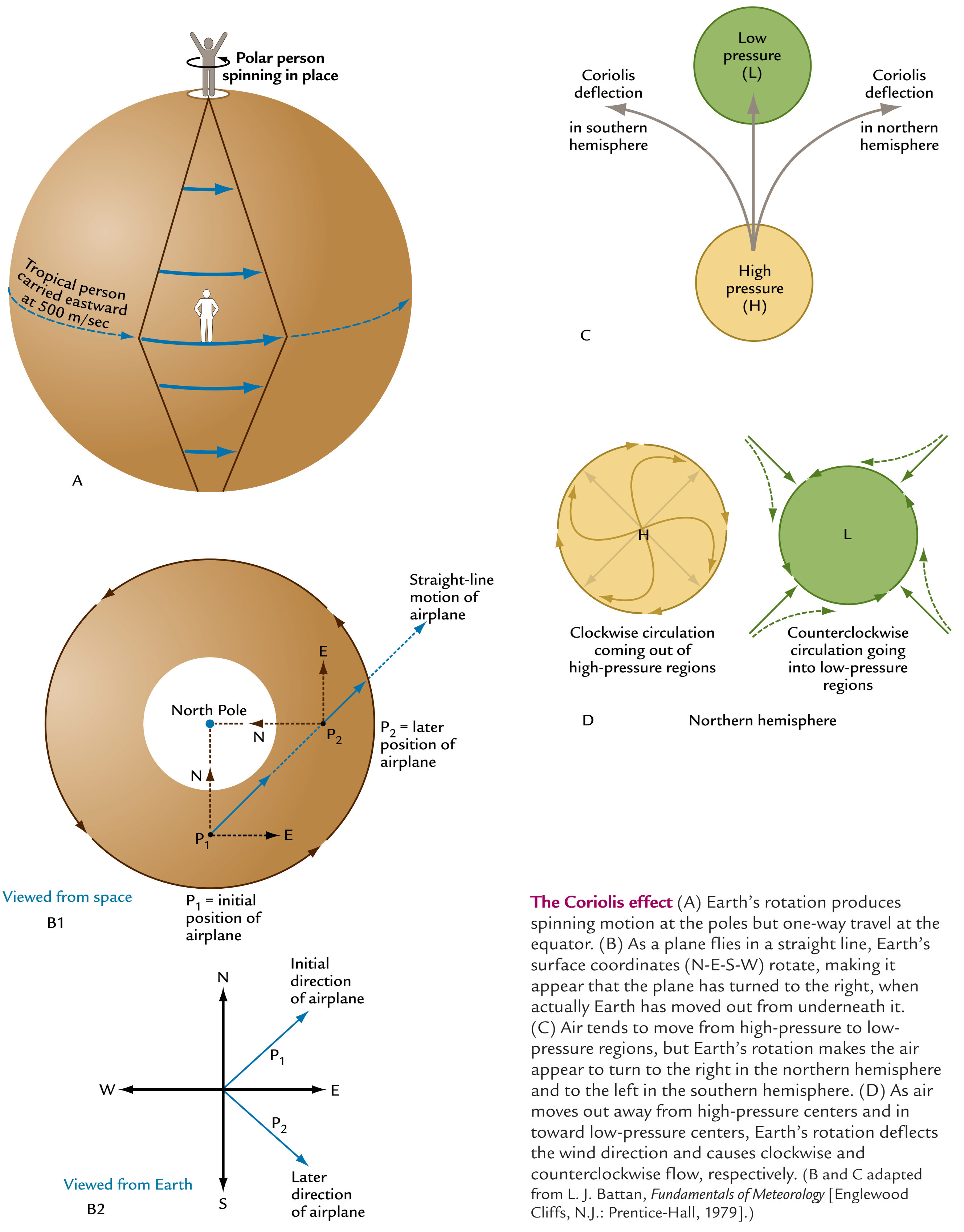
35°N latitude (see Figure 2-16). The circulation at latitudes above 35° is more difficult to summarize.

To understand the transfer of heat from middle to high latitudes, we begin with the regions of high surface pressure in the subtropics. The air that sinks to Earth's surface in the subtropical branch of the Hadley cell converges or piles up there, creating a nearly continuous band of high surface pressures in the subtropics (Figure 2-19). Superimposed on this basic pattern is a tendency for the monsoonal flow of air from land to sea in summer to produce oval-shaped cells of high pressure over the subtropical oceans (see Figure 2-17A).

Because air naturally flows away from regions of higher pressures toward areas where pressure is lower, the subtropical high-pressure cells send air moving outward near Earth's surface in all directions. Ultimately, however, the path taken by this air is not directly from regions of high to low pressure. Earth's rotation deflects the path of this air in a *relative* sense (Box 2-5), and this deflection produces a clockwise spin of air around subtropical highs in the northern hemisphere and a counterclockwise spin in the southern hemisphere.

Returning to the summary map of Earth's circulation in Figure 2-16, notice that the trade winds that flow out from the subtropical high-pressure zones toward the equator in the northern hemisphere are deflected to the west by Earth's rotation and given a net easterly trajectory: from northeast to southwest. These are the trade winds that move toward the tropics.

The same Coriolis deflection turns air flowing poleward from the subtropical highs toward the east, resulting in a net southwest-to-northeast ("westerly") flow. This surface flow of warm air out of the subtropics transports heat into high latitudes where the radiation balance is negative (see Figure 2-14B). Much of this heat transport in the atmosphere is ultimately tied to the large transfer of latent heat from the ocean to the atmosphere in the tropics.



The Coriolis effect (A) Earth's rotation produces spinning motion at the poles but one-way travel at the equator. (B) As a plane flies in a straight line, Earth's surface coordinates (N-E-S-W) rotate, making it appear that the plane has turned to the right, when actually Earth has moved out from underneath it. (C) Air tends to move from high-pressure to low-pressure regions, but Earth's rotation makes the air appear to turn to the right in the northern hemisphere and to the left in the southern hemisphere. (D) As air moves out away from high-pressure centers and in toward low-pressure centers, Earth's rotation deflects the wind direction and causes clockwise and counterclockwise flow, respectively. (B and C adapted from L. J. Battan, *Fundamentals of Meteorology* [Englewood Cliffs, N.J.: Prentice-Hall, 1979].)

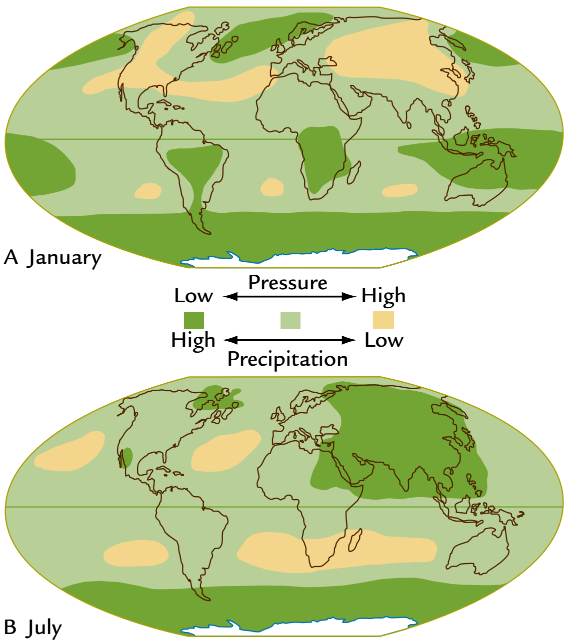


FIGURE 2-19 **Seasonal pressure patterns** A band of high surface pressure occurs in the subtropics in both hemispheres in both seasons. During summer, this zone is interrupted over land (especially Asia) by areas of low pressure produced by summer monsoon circulations. (Adapted from E. Bryant, *Climate Process and Change* [Cambridge: Cambridge University Press, 1998].)

Toward higher middle latitudes in both hemispheres, the circulation of the lower atmosphere is a complex zone of transition between the prevailing warm flow coming out of the subtropics and a much colder equatorward flow from higher latitudes (see Figure 2-16). The surface flow at these latitudes is dominated by an ever-changing procession of high- and low-pressure cells moving from west to east, separated by *frontal zones*, or regions near Earth's surface where large changes in temperature occur over short distances in association with fast-moving air. Both the poleward movement of warm air and the equatorward movement of cold air along these frontal systems have the effect of warming the higher latitudes in a net sense, adding a major contribution to heat redistribution on Earth.

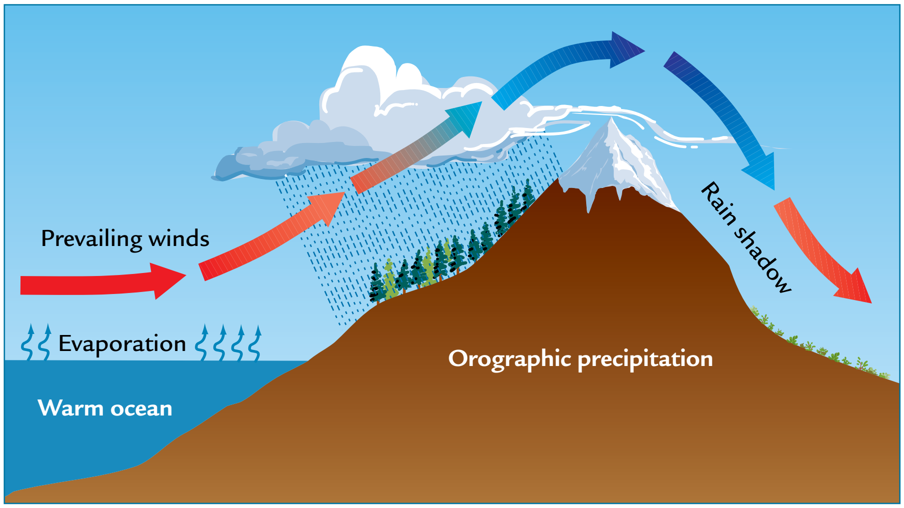
Precipitation generally exceeds evaporation in the temperate middle latitudes, for two reasons: the cooler air temperatures reduce the rates of evaporation, and precipitation associated with moving low-pressure cells is heavy.

Because the low-pressure cells move rapidly from west to east across the middle latitudes, they encounter and interact strongly with the topography of the land. Air flowing in from the ocean tends to carry large amounts of water vapor. As this air encounters mountain ranges that block its flow, it is forced to rise to higher elevations, and it cools. Water vapor condenses from the cooling air and produces heavy precipitation on the sides of mountains that face upwind toward warm oceans, such as the Olympic Mountains of Washington State. This is referred to as **orographic precipitation** (Figure 2-20).

Air that has been stripped of much of its water vapor then sinks on the downwind side of the mountains. As it moves to lower elevations, it is compressed and warmed, and it gains in capacity to store even more water vapor without condensation occurring. As a result, the lee or *rain shadow* sides of mountain ranges are areas of lower precipitation. This process also reinforces the natural tendency of mid-continental regions far from the oceanic sources of moisture, such as the Great Plains of the United States, to be dry.

At higher elevations in the mid-latitude atmosphere, winds flow more steadily from west to east. Narrow rib-

FIGURE 2-20 **Orographic precipitation** As moist air masses driven up against mountains rise and cool, water vapor condenses and produces precipitation. The air masses subsiding on the downwind side of the high topography warm, retain water vapor, and suppress precipitation. (Adapted from F. Press and R. Siever, *Understanding Earth,* 2nd ed., © 1998 by W. H. Freeman and Company.)



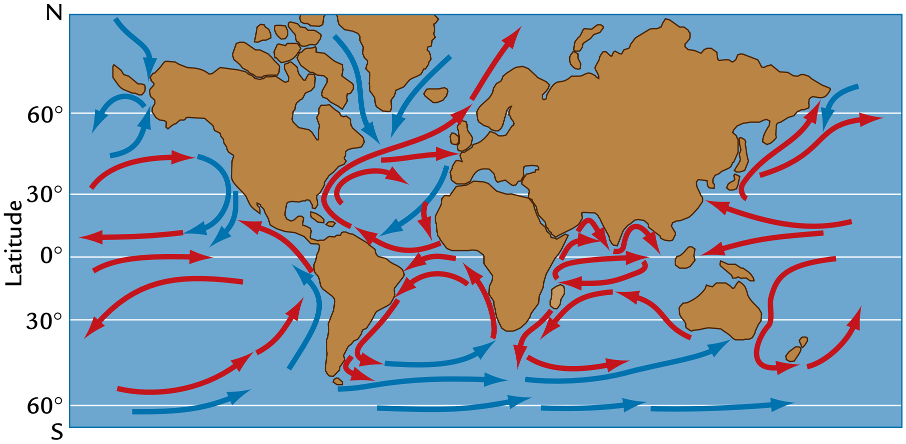


FIGURE 2-21 **Surface ocean circulation** The surface flow of the oceans is organized into strong wind-driven currents. These currents encircle large spinning gyres in the subtropical oceans. Currents moving out of the tropics carry heat poleward, while currents moving away from the poles carry cold water equatorward. (Modified from S. Stanley, *Earth Systems History*, © 1999 by W. H. Freeman and Company.)

bons of faster flow called **jet streams** occur at altitudes of 5 to 10 kilometers in two regions: a persistent but weaker jet near 30° latitude in the subtropics, near the sinking branch of the Hadley cell; and a more mobile jet that wanders between latitudes 30° and 60° above the moving high- and low-pressure cells. The jet at middle latitudes is especially strong in winter. Almost hidden by these prevailing west-to-east motions and the meandering paths of these jet streams is a net transport of heat and water vapor from low to high latitudes. Poleward meanders in the jet stream carry warm air to the north, and equatorward meanders carry cold air south.

Heat Transfer in Earth's Oceans

The uppermost layer of the ocean is heated by solar radiation. Like air, water expands as it warms and becomes less dense, but in this case the warmest layers are already at the top of the ocean, so they simply float on top of the colder, denser deep ocean. Winds mix the stored solar heat to maximum depths of 100 meters, a small fraction of the 4000-meter average depth of the oceans. Some of this warm water is transported from the tropics toward the poles, and this poleward flow carries about half as much heat as is transported by the atmosphere.

2-7 The Surface Ocean

Most of the surface circulation of the oceans is driven by winds, and one of the most prominent results is huge **gyres** of water at subtropical latitudes (Figure 2-21). These spinning gyres are mainly the result of an initial push (or drag) of the winds on the ocean surface, and of the Coriolis deflection of the moving water (see Box 2-5).

Blowing wind exerts a force on the upper layer of the ocean and sets it in motion in the same direction as the wind. The Coriolis effect turns this surface flow of water to the right in the northern hemisphere (and to the left in the southern hemisphere). The top layer of water in turn pushes (or drags) the underlying layers, which are deflected a little farther to the right than the surface layer and are also slowed by friction. This process continues down into the water column to a depth of about 100 meters, creating a downward spiral of water gradually deflected farther and farther to the right (Figure 2-22). The net transport of water in this

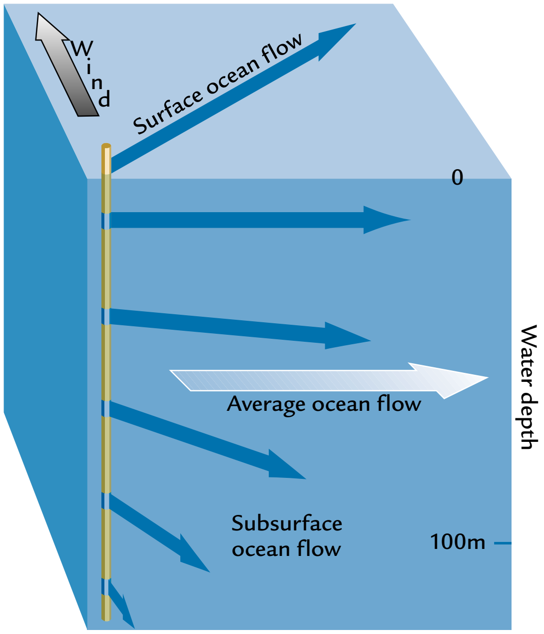


FIGURE 2-22 **Effect of surface winds on the ocean** In the northern hemisphere, low-level winds drive surface waters to the right of the direction in which the wind is moving. Subsurface water is turned progressively farther to the right, and the net transport of the upper layer of water is 90° to the right of the direction of the wind. (Modified from D. Merritts et al., *Environmental Geology*, © 1997 by W. H. Freeman and Company.)

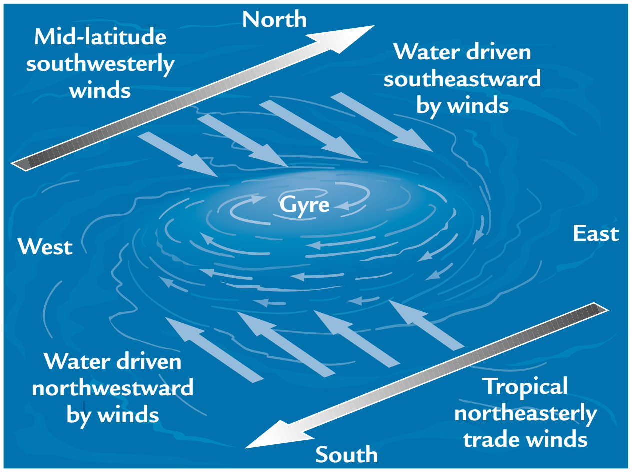


FIGURE 2-23 **Subtropical gyres** In the northern hemisphere, mid-latitude southwesterly winds and tropical northeasterly trade winds drive warm water toward the centers of subtropical gyres, forming a thick lens of warm water that circulates in a clockwise gyre.

entire 100-meter layer of ocean water is 90° to the right of the wind in the northern hemisphere (and to the left south of the equator).

In the North Atlantic Ocean, the prevailing low-altitude winds are the tropical trade winds and mid-latitude westerlies. Trade winds blowing toward the southwest push shallow waters toward the northwest, and mid-latitude westerlies blowing toward the northeast push surface water to the southeast (Figure 2-23). Together these winds drive the uppermost layer of water into the centers of the subtropical gyres and pile up a lens of warm water.

Sea level in the center of this lens sits 2 meters higher than the surrounding ocean. Water that flows away from this lens is turned to the right by the Coriolis deflection, and this creates a huge subtropical ocean gyre spinning in a clockwise direction (counterclockwise for gyres in the southern hemisphere). The edges of the continents also play a role in forming these gyres by acting as boundaries that contain the flow within individual ocean basins.

Subtropical gyres extend all the way to depths of 600 to 1000 meters. In the North Atlantic, most of the water moving into the deeper parts of the gyre comes from its northern margin, where the prevailing westerly winds are strong enough to push large volumes of water toward the south just underneath the lens of warm surface water.

Viewed over long intervals of time, most of the flow in subtropical gyres consists of water moving around in giant spirals. The volume of water circulating is enormous, about 100 times the transport of all Earth's rivers flowing into the oceans. Almost hidden in this recirculation is a much smaller amount moving *through* the gyres and carrying heat toward high latitudes.

The prevailing flow toward the equator in the deeper parts of the gyres must be balanced by a return flow toward the poles, and this flow is concentrated in narrow regions along the western gyre margins. In the North Atlantic, the poleward transport occurs in the **Gulf Stream** and its continuation, the **North Atlantic Drift** (see Figure 2-21). As the Gulf Stream emerges from the Gulf of Mexico, it forms a narrowly concentrated outflow of warm salty water headed north.

Another factor that affects poleward heat transport only in the North Atlantic Ocean is a giant vertical circulation cell linked to the deeper circulation of the ocean (Figure 2-24). A large volume of surface water sinks to depths below 2 kilometers in the higher latitudes of the North Atlantic, and this sinking water must be balanced by a compensating inflow of surface water from the south.

The effects of this circulation cell are felt even beyond the North Atlantic. The surface circulation typical of most oceans carries heat from the warm equator to the cold poles, as partial compensation for the imbalances set up by uneven solar heating and heat absorption at the surface. This normal equator-to-pole flow is

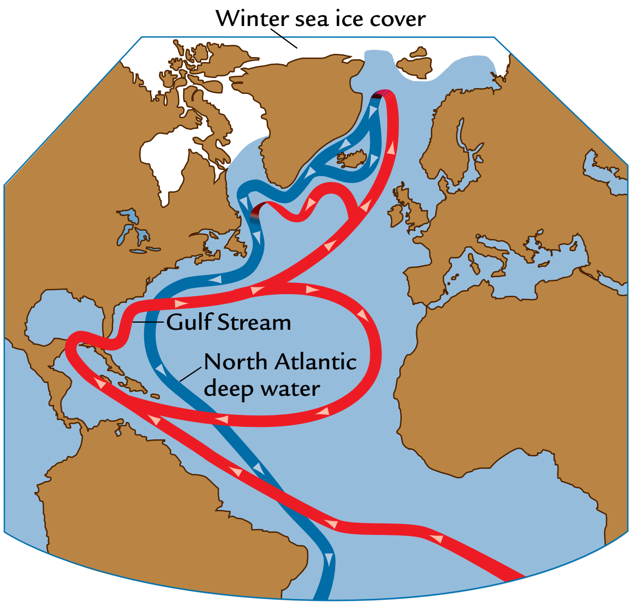


FIGURE 2-24 **Sinking of surface water** Warm salty water flowing northward in the North Atlantic Ocean chills and sinks north of Iceland and in the Labrador Sea, between North America and Greenland. This cold deep water flows south out of the Atlantic at depths of 2 to 4 km. (Modified from D. Merritts et al., *Environmental Geology,* © 1997 by W. H. Freeman and Company.)

reversed in the South Atlantic, where the net direction of heat transport is from south to north, *against* the planet's temperature gradient (see Figure 2-21). Temperate water flows into the South Atlantic from the middle latitudes of the Indian and Pacific oceans, moves northward across the equator, and sinks in the high-latitude North Atlantic.

The net northward transport of heat in the North Atlantic is often referred to as a "conveyor belt" (see Figure 2-24), but the overall flow has also been compared to an airport baggage carousel. Most of the luggage (heat) spins around and around the carousel (warm water recirculating in the subtropical gyre), while only a small amount enters the carousel (comes across the equator from the South Atlantic) or is removed (heads farther north in the Gulf Stream and its continuation).

This warm, northward-moving water in the Atlantic transfers a huge amount of heat to the atmosphere. At latitudes above 50°N, the large temperature contrast between the warm North Atlantic waters and the cold overlying air produces a loss of sensible heat from the ocean to the atmosphere that is comparable to the amount of heat delivered locally by incoming solar radiation.

The fundamental circulation and heat transport of the oceans are less well understood than those of the atmosphere, mainly because of the difficulty of maintaining long-term monitoring stations at sea. The closer oceanographers look at the surface flow, the more complicated it turns out to be, with smaller gyres of water recirculating within larger gyres. At even smaller scales, spinning cells of water 100 kilometers wide move erratically across the ocean and transport large amounts of heat. These are analogous to moving low-pressure and high-pressure cells in the atmosphere.

2-8 Deep-Ocean Circulation

The poleward flow of warm water that counters some of Earth's heat imbalance occurs above the **thermocline**, a zone of rapid temperature change between warm upper layers and cold water filling the deeper ocean basins. Actually, two thermoclines exist: (1) a deeper permanent portion that is maintained throughout the year, and (2) a shallower portion that changes as a result of seasonal heating by the Sun (Figure 2-25).

We've just seen how the warm poleward flow above the thermocline in the North Atlantic is balanced by sinking of cold water at high latitudes and movement of this cold deep water toward the equator. This overturning circulation is called the **thermohaline flow**. This term refers to the two main processes that control formation of deep water: temperature ("thermo-") and salinity ("-haline," from the same root as "halite," a synonym for rock salt).

Deep waters form and sink because they become more dense than the underlying water, as a result of any of several mechanisms. Seawater contains dissolved salt (on average near 35 parts per thousand, or ‰, by mass), and this salt content, or **salinity**, makes it 3.5% more dense than freshwater. The density of ocean water can be increased at lower latitudes when the atmosphere evaporates freshwater as water vapor, leaving the remaining water saltier (denser). The density of seawater can also can be increased at high latitudes by formation of sea ice during **salt rejection**, a process that stores freshwater in sea ice and leaves the salt behind.

Another way to increase the density of ocean water is by cooling it. Saltwater is slightly compressible, which means it loses volume and gains density when it is cooled by the atmosphere. Cooling can occur either because warm ocean water is carried poleward into cooler regions or because colder air masses move to lower latitudes.

Salinity and temperature often work together to raise the density of water. Initially, evaporation or formation of sea ice increases the salinity and density of

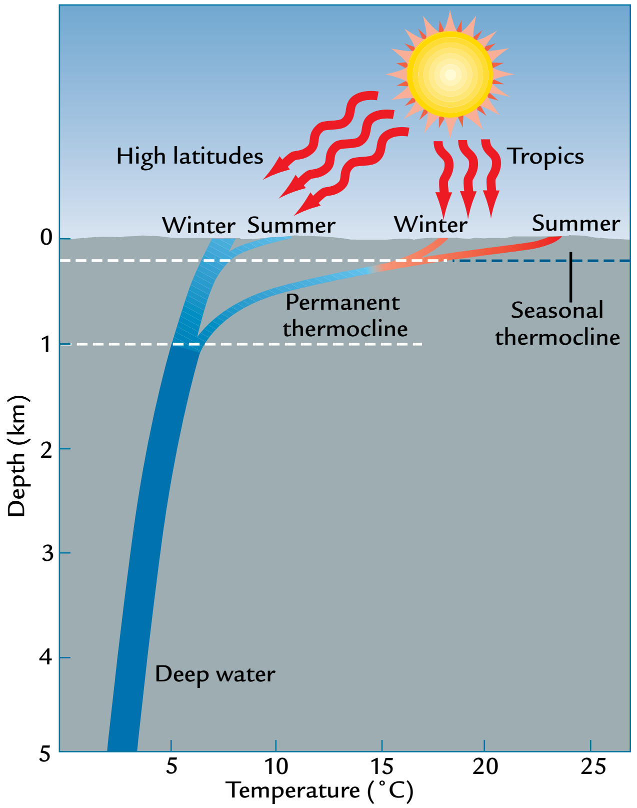


FIGURE 2-25 Thermoclines The permanent thermocline (100–1000 m) separates cold deep water from shallower layers affected by changes in Earth's surface temperature. Shallow seasonal thermoclines (0–100 m) vary in response to seasonal solar heating of the upper ocean layers.

surface waters as a kind of preconditioning process. Cold air then cools the water, further increases its density, and causes it to sink.

The large-scale thermohaline flow in the deep ocean is closely linked to temperature and its effect on density. The very dense waters filling the deepest ocean basins today form at higher, colder latitudes. The progressively less dense waters that fill successively shallower depths of the ocean form at less frigid latitudes farther from the poles.

Most of the deepest ocean of the world is filled by water delivered from just two regions, the high-latitude North Atlantic Ocean and the Southern Ocean, near Antarctica (Figure 2-26). No deep water forms in the high latitudes of the Pacific Ocean today because the salinity of the surface water is too low and the surface waters there are not sufficiently dense.

The surface waters that sink and form deep water in the North Atlantic initially acquire high salinity in the dry subtropics as a result of strong evaporation. Some of the water vapor taken from the tropical Atlantic Ocean is exported westward over the low mountains of Central America into the Pacific Ocean, leaving the Atlantic saltier than equivalent latitudes in the Pacific. Some of the salty water left in the Atlantic is carried northward by the Gulf Stream and North Atlantic Drift. Frigid air masses from the surrounding continents then extract sensible heat from the water in winter and further increase its density to the point where it sinks.

This water mass is called **North Atlantic deep water**. Sinking occurs in two regions in the North Atlantic, one north of Iceland, the other east of Labrador (see Figure 2-24). Together these two sources of deep water fill the Atlantic Ocean between depths of 2 and 4 kilometers. This flow moves southward with a total volume 15 times the combined flow of all the world's rivers. Eventually, much of this flow rises toward the sea surface in the Southern Ocean and joins the waters circling eastward around Antarctica.

An even colder and denser water mass forms in the Antarctic region and flows northward in the Atlantic below 4 kilometers (see Figure 2-26). This water mass, called **Antarctic bottom water**, fills the deep Pacific and Indian oceans. Some forms near the Antarctic coast when seawater is chilled by very cold air masses and as a result of salt rejection as sea ice forms. Some also forms by intense cooling in gaps in the extensive sea-ice cover well away from Antarctica.

Two smaller water masses are prominent at intermediate depths of the North Atlantic. **Antarctic intermediate water** forms far north of Antarctica at latitudes 45°–50°S. This water is warmer and less dense than North Atlantic deep water and flows northward above it at depths above 1.75 kilometers. **Mediterranean overflow water** forms in the subtropical Mediterranean Sea as a result of winter chilling of surface waters with a very high salt content caused by strong evaporation. Water vapor extracted by subtropical evaporation leaves Mediterranean waters much saltier than normal ocean water.

Deep ocean water is one of the slowest-responding parts of the climate system. On average, it takes more than 1000 years for a parcel of water that leaves the surface and sinks into the deep ocean to emerge back at the ocean's surface. The oldest and slowest-moving deep

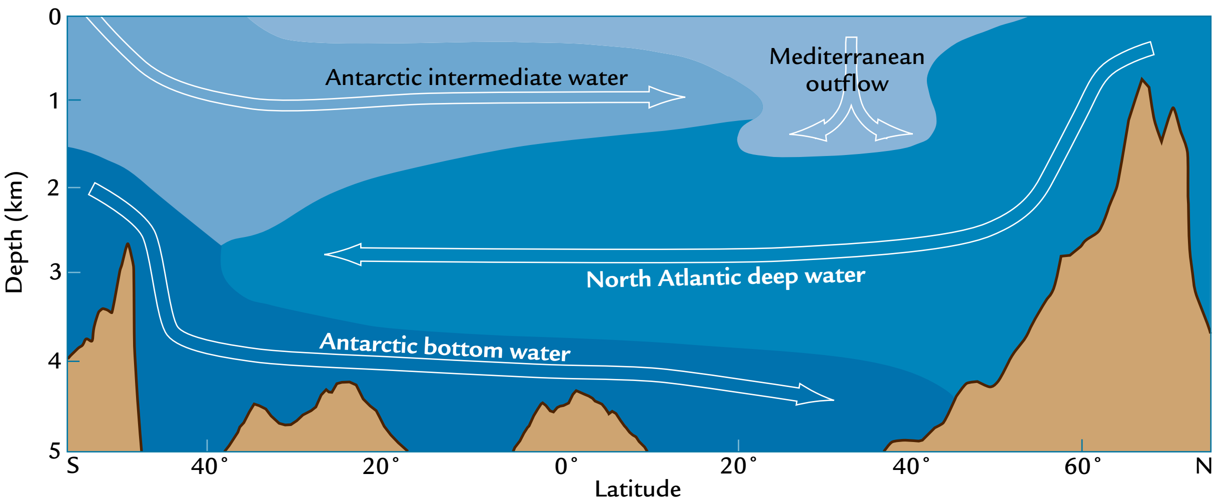


FIGURE 2-26 **Deep Atlantic circulation** Water filling the North Atlantic basin comes from sources in the high-latitude North Atlantic, the Southern Ocean near Antarctica, and (at shallower depths) the Mediterranean Sea. (Adapted from E. Berner and R. Berner, *Global Environment* [Englewood Cliffs, N.J.: Prentice-Hall, 1996].)

water is found in the Pacific Ocean, while the Atlantic has younger, faster-moving water. The long journey of water through the deep ocean keeps much of it out of touch with the climatic changes that affect the atmosphere and surface ocean.

With all this water sinking into the deep ocean, how does it get back to the surface? It turns out that climate scientists don't really know the answer to this question very well. In the Atlantic, some of the southward-flowing North Atlantic deep water rises to the surface near Antarctica, and very strong winds mix it into the upper layers of the Southern Ocean. But what happens to the large volume of Antarctic water moving north in the deep Pacific and Indian Oceans, where no deep water forms?

For decades a widely accepted explanation has been that deep water injected into the ocean in specific regions gradually mixes into the central ocean basins and slowly moves upward across the thermocline and into the warm surface waters. But this highly diffuse return flow has been difficult to detect because it is spread across such a large area. Recent measurements show that this upward diffusion of water is too slow to account for much of the return flow.

In addition, a process called **upwelling**, rapid upward movement of subsurface water from intermediate depths, occurs in two other kinds of ocean regions. Both upwelling processes are initiated by surface winds and aided by the Coriolis effect:

- When surface winds in the northern hemisphere blow parallel to coastlines along the path shown in Figure 2-27A, they push water away from the land. To replace surface water pushed offshore, water rises from below. The upwelling water is cooler than the nearby surface water that has remained at the surface and has been warmed by the Sun.

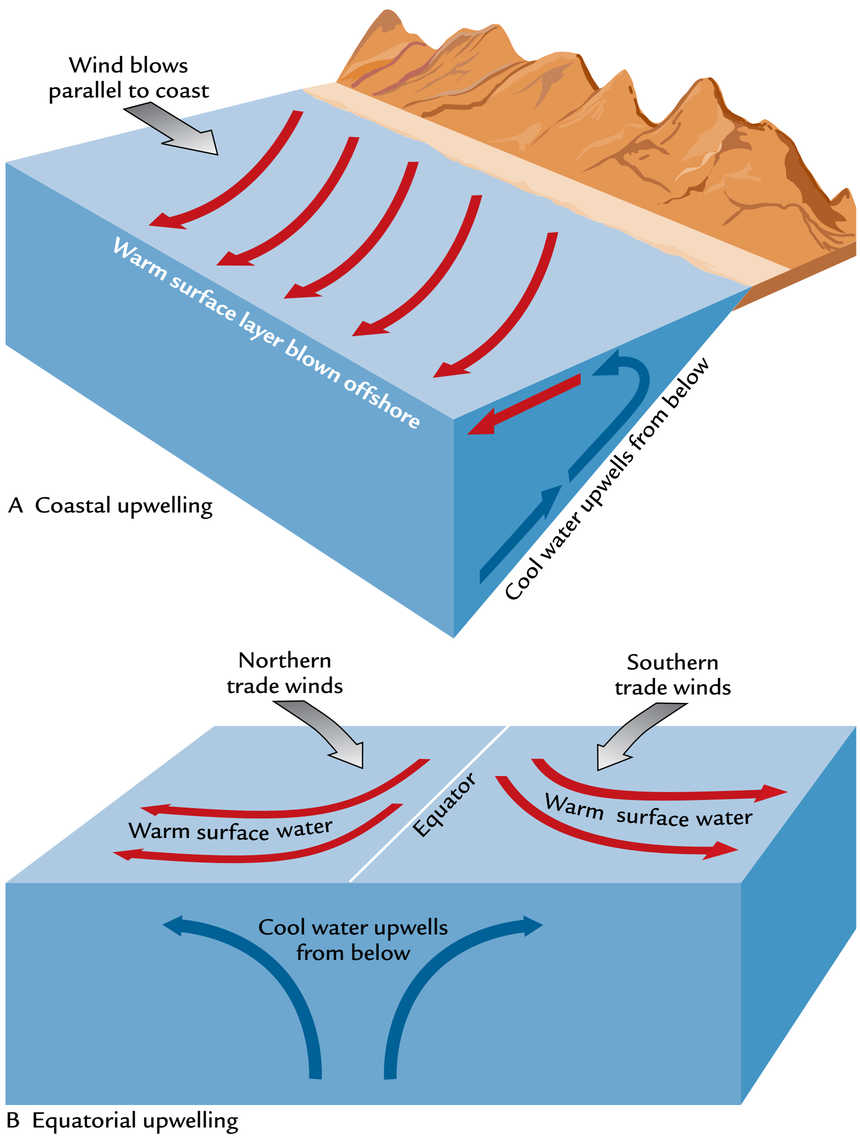


FIGURE 2-27 **Upwelling** Cool subsurface water rises along coastal margins (A), where winds drive warm water offshore, and near the equator (B), where winds drive surface waters away from the equator. (Modified from D. Merritts et al., *Environmental Geology,* © 1997 by W. H. Freeman and Company.)

- A second kind of upwelling occurs along the equator, especially in the eastern end of ocean basins (Figure 2-27B). Trade winds push surface waters away from the equator. Warm surface water is driven northward in the northern hemisphere and southward in the southern hemisphere by the opposing Coriolis deflections north and south of the equator. The movement of warm water away from the equator causes upwelling of cooler water from below.

Ice on Earth

Despite the redistribution of heat by air and water, temperatures cold enough to keep water frozen solid occur at high altitudes and latitudes. Ice is one of the most important components of the climate system, because its properties are so different from those of air, water, and land.

2-9 Sea Ice

Although freshwater freezes at 0°C, typical seawater resists freezing until it is cooled to −1.9°C. As sea ice forms, it rejects almost all the salt in the seawater. Because sea ice is less dense than seawater, it floats on top of the salty ocean.

When sea ice forms, it seals off the underlying ocean from interaction with the atmosphere. This change is vital to regional climates. Without an ice cover, high-latitude oceans transfer large amounts of heat to the atmosphere, especially in winter, when air temperatures are low (Figure 2-28A). This heat transfer keeps temperatures in the lower atmosphere close to those of the ocean surface (near 0°C).

But if an ice cover is present, this heat release stops, and the reflective ice surface absorbs little incoming solar radiation. Because of these changes, winter air temperatures can cool by 30°C or more in regions that develop a sea-ice cover (Figure 2-28B). In effect, an ice-covered ocean behaves like a snow-covered continent. This change forms a prominent part of the albedo–temperature feedback process examined in Box 2-2.

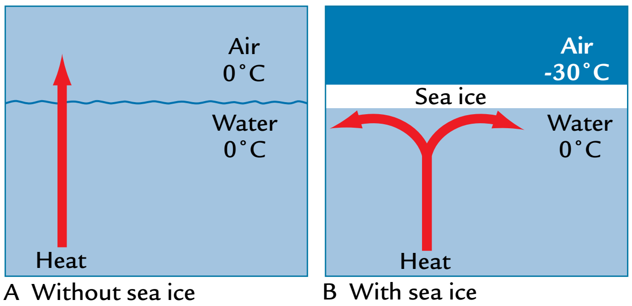


FIGURE 2-28 **Effect of sea ice on climate** Whereas heat can escape from an unfrozen ocean surface (A), a cover of sea ice (B) stops the release of heat from the ocean to the atmosphere in winter and causes air temperatures to chill by as much as 30°C.

Many ocean surfaces are only partially ice-covered. Gaps ("leads") produced in the ice by changing winds allow some heat exchange with the atmosphere and moderate the climate effects of a full sea-ice cover. Also, in summer, meltwater pools may form on the ice surface, and this water, along with a gradual darkening of the melting ice, may absorb more solar radiation.

The formation and melting of sea ice are driven mainly by seasonal changes in solar heating. In the Southern Ocean, most of the sea ice melts and forms again every year, over an area comparable in size to the entire Antarctic continent it surrounds. This annual ice cover averages 1 meter in thickness, except where strong winds cause the ice to buckle and pile up in ridges. In contrast, the landmasses surrounding the Arctic Ocean constrain the movement of sea ice and allow it to persist for 4 or 5 years. Older sea ice in the center of the Arctic may reach 4 meters in thickness, while annually formed ice around the margins is about 1 meter thick.

Recall that large inputs and extractions of heat calories from the atmosphere are required to form and melt sea ice, and the cycle of freezing and melting also involves exchanges of heat with the slow-responding ocean because of its high heat capacity. For these reasons, seasonal extremes in sea-ice cover lag well behind the seasonal extremes of heating by solar radiation. The maximum extent of sea ice is usually reached in the spring, the minimum extent in the autumn.

2-10 Glacial Ice

Glacier ice occurs mainly on land, in two forms. **Mountain glaciers** are found in mountain valleys at high elevations (Figure 2-29, top). Because glaciers can exist only where mean annual temperatures are below freezing, mountain glaciers near the equator are restricted to elevations above 5 kilometers (Figure 2-29, bottom). At higher and colder latitudes, mountain glaciers may reach down to sea level. Typical mountain glaciers are a few kilometers in length and tens to hundreds of meters in width and thickness. They typically flow down mountain valleys, constrained on both sides by rock walls.

Continental ice sheets are a much larger form of glacier ice, typically hundreds to thousands of kilometers in horizontal extent and 1 to 4 kilometers in thickness. The two existing ice sheets, which cover most of Antarctica and Greenland, represent roughly 3% of Earth's total surface area and 11% of its land surface.

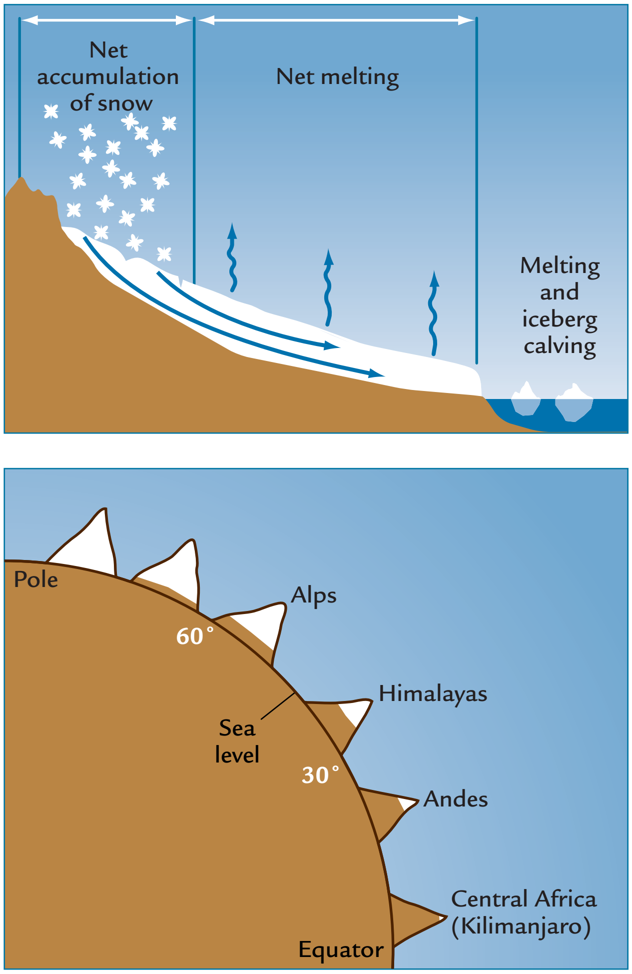


FIGURE 2-29 **Mountain glaciers** Mountain glaciers accumulate snow at colder, higher elevations. (Top) The snow turns to ice, flows to lower elevations where temperatures are warmer, and melts. (Bottom) Mountain glaciers can exist near sea level at high latitudes, but survive only at elevations above several kilometers in the warm tropics. (Modified from F. Press and R. Siever, *Understanding Earth*, 2nd ed., © 1998 by W. H. Freeman and Company.)

These ice sheets contain some 32 million cubic kilometers of ice, equivalent to about 70 meters of sea level change.

Continental ice sheets indeed have dimensions comparable to those of sheets—usually more than 1000 times as wide as they are thick—but their surfaces do have structure (Figure 2-30, top). The highest regions on the ice sheets are rounded **ice domes**, with the elevations sloping gently away in all directions. Domes may be connected by high broad ridges with gentle sags called **ice saddles**. On the sides of ice sheets, ice flows in fast-moving **ice streams** from which **ice lobes** protrude beyond the general ice margins.

These great masses of ice depress the underlying bedrock below the elevation it would have had if no ice were present (Figure 2-30, bottom). About 30% of the total ice thickness sits below the original (undepressed) bedrock level; the other 70% protrudes into the atmosphere as a broad smooth ice plateau.

Snow that falls on the higher parts of mountain glaciers and ice sheets gradually recrystallizes into ice. The ice then moves toward lower elevations under the force of gravity. Ice in mountain glaciers is affected by gravity because of the steep slopes of mountain valleys. Continental ice sheets are high plateaus, and gravity moves ice from higher to lower elevations.

Ice deforms and moves in the upper 50 meters of a glacier in a brittle way: fracturing and forming crevasses, emitting loud cracking noises in the process. Below about 50 meters, ice deforms more gradually, by slow plastic flow. Parcels of ice may take hundreds to thousands of years to travel through mountain glaciers. For the central domes of ice sheets, where the flow is directed deep into the center of the ice mass, the trip may take tens of thousands of years.

As ice flows, its layers are stretched and thinned. If the ice is very cold (below −30°C), it behaves in a stiff manner, and its sloping edges can be relatively steep. Impurities such as dust also tend to stiffen the ice. If the ice is somewhat warmer (close to the freezing point), it

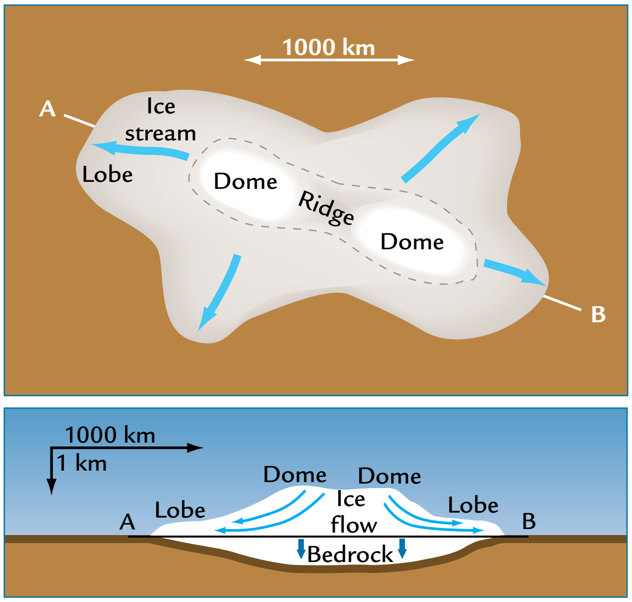


FIGURE 2-30 **Continental ice sheets** The central portions of large continent-sized ice sheets have high central domes connected by ridges. Ice streams on the flanks carry ice to lobes protruding from the ice margins (top). In cross section, snow accumulates on the high part of an ice sheet, turns to ice, and flows to the lower margins (bottom).

is more plastic—softer and easier to deform—and it will tend to relax into gentler slopes.

Ice also moves by mechanisms favorable to sliding on its basal layers. Large amounts of water can accumulate at the base of mountain glaciers, causing them to **surge** down valleys at rates far in excess of their normal movement. Lobes of continental ice sheets also move by sliding along their bases. In regions where ice streams occur, the ice may move several meters per day, or 100 to 1000 times faster than the rest of the ice mass.

Ice streams occur for two reasons. First, the pressure from the weight of the overlying ice may cause some of the ice at the bottom to melt and create a thin layer of water on which the ice can slide. Second, this water may percolate into and saturate soft unconsolidated sediments lying beneath the outer margins of the ice sheet, causing them to lose their mechanical strength or cohesiveness. These water-lubricated sediments provide a slippery *deformable bed* on which the overlying ice can easily slide.

Under certain conditions, **ice shelves** may form over shallow ocean embayments, and several shelves exist today on the margins of Antarctica. In these regions, gravity pulls ice out of the interior of the continent to the embayments, where it spreads out in shelves tens to hundreds of meters thick. Bedrock surrounding these embayments and at the shallow depths below it provides friction that keeps the ice from sliding away into the ocean. Immense **tabular icebergs** occasionally break off from these shelves and float away. An iceberg the size of the state of Connecticut broke off from Antarctica a few years ago.

The bottom of the western part of the Antarctic ice sheet lies below sea level, and this portion is called the West Antarctic **marine ice sheet**. Because marine ice sheets have bases lying below sea level, they are highly vulnerable to sea level changes and respond to changes in climate much more quickly than ice sheets that sit higher on the land.

Mountain glaciers and continental ice sheets ultimately exist for the same reason: the overall rate of snow falling across the entire ice mass equals or exceeds the overall rate at which ice is lost by melting and other means. Climate scientists analyze the conditions over present-day glaciers and ice sheets in terms of their **mass balance**, the average rate at which ice either grows or shrinks every year. The concept of mass balance can also be applied to different portions of glaciers: mass balances are positive at upper elevations, where **accumulation** of snow and ice dominates, but negative at lower elevations, where rapid **ablation** (loss of ice) occurs.

Ice accumulation occurs in regions where temperatures are cold enough both to cause precipitation to freeze and also to allow new-fallen snow to persist through the warm summer season. For mountain glaciers, subfreezing temperatures occur on the highest parts of mountains, where the air is coldest. For continent-sized ice sheets, which often exist at sea level, the cold required to sustain ice is found at high polar latitudes and on the high parts of the ice sheets.

Ablation of glacial ice by melting occurs when temperatures exceed the freezing point. Melting can occur because of absorption of solar radiation or by uptake of sensible or latent heat delivered by warm air masses (and by rain) moving across the ice. Ablation can also occur by **calving**, the shedding of icebergs to the ocean or to lakes. Calving differs from the other processes of ablation in that icebergs leave the main ice mass and move elsewhere to melt, often in an environment much warmer than that near the ice sheet.

The boundary between the high-elevation region of positive ice mass balance and the lower area of net loss of ice mass occurs at a mean annual temperature near −10°C for ice sheets but closer to 0°C for mountain glaciers. The mass balance at high elevations on the Greenland and Antarctic ice sheets is positive because of the absence of melting to offset the slow accumulation of snow (Figure 2-31). At elevations above 1 to 2 kilometers, air is so cold that it contains little water vapor. Although the precipitation that falls on these higher parts of the ice sheets is all snow, accumulation rates are low. This is especially the case for the frigid

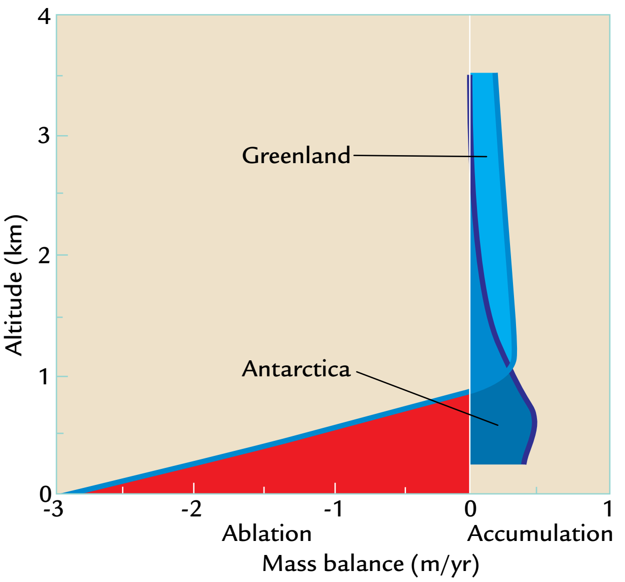


FIGURE 2-31 **Ice mass balance** Snow accumulates on the upper parts of ice sheets where melting does not occur. At lower elevations, accumulation is overwhelmed by net loss of ice by ablation due to melting (in Greenland) and to calving of icebergs (in Antarctica). The units shown convert snow to equivalent thicknesses of ice (in meters). (Adapted from J. Oerlemans, "The Role of Ice Sheets in the Pleistocene Climate," *Norsk Geologisk Tidsskrift* 71 [1991]: 155–161.)

Antarctic continent, centered on the South Pole and surrounded by an ice-covered ocean. The mass balance is more positive on the sides of these ice sheets, where air masses carry more moisture and cause more snowfall, yet ablation is not strong.

The mass balance on ice sheets is negative at lower elevations, usually because mean annual temperatures above 0°C accelerate the rate of melting. In Greenland, some of the low-elevation precipitation also falls as rain rather than snow, which further promotes ablation. In Antarctica, freezing conditions persist at sea level even in summer, and no melting occurs. The Antarctic ice sheet loses mass mainly by calving icebergs into the ocean.

Earth's Biosphere

To this point, we have examined only the physical side of the climate system, expressed mainly by variations in temperature, precipitation, winds, and pressure, but these physical parts of the climate system also interact with its organic parts (Earth's **biosphere**). Many of these interactions result from the movement of carbon (C) through the climate system and in turn affect the distribution of heat on Earth.

Carbon moves among and resides in several major reservoirs. The amount of carbon in each reservoir is typically quantified in gigatons (or 10^{15} grams) of carbon. Relatively small amounts of carbon reside in the atmosphere, the surface ocean, and vegetation; a slightly larger reservoir resides in soils, a much larger reservoir in the deep ocean, and a huge reservoir in rocks and sediments (Figure 2-32A).

Carbon takes different chemical forms in these different reservoirs. In the atmosphere, it is a gas (CO_2). Carbon in land vegetation is organic, as is most carbon in soils, while that in the ocean is mostly inorganic, occurring as dissolved ions (atoms carrying positive or

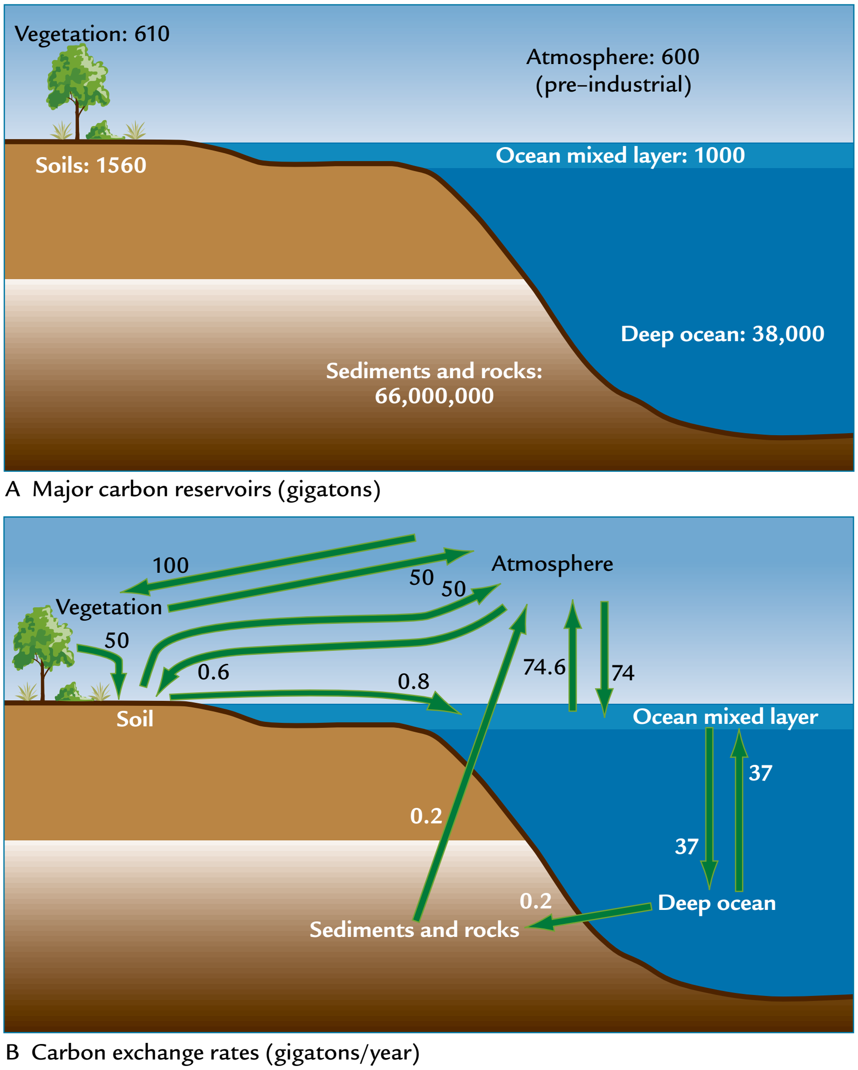


FIGURE 2-32 The carbon cycle The major carbon reservoirs on Earth vary widely in size (A) and exchange carbon at differing rates (B). Larger reservoirs (rocks, the deep ocean) exchange carbon much more slowly than smaller reservoirs (air, vegetation, the surface ocean). (Adapted from J. Horel and J. Geisler, *Global Environmental Change* [New York: John Wiley, 1997], and from National Research Council Board on Atmospheric Sciences and Climate, *Changing Climate,* Report of the Carbon Dioxide Assessment Committee [Washington, D.C.: National Academy Press, 1993].)

negative charges). Despite these differences in form, carbon is exchanged freely among all the reservoirs, changing back and forth between organic and inorganic forms as it moves.

Rates of carbon exchange among reservoirs vary widely (Figure 2-32B). In general, the sizes of the reservoirs are inversely related to their rates of carbon exchange. The small surface reservoirs (the atmosphere, surface ocean, and vegetation) exchange all their carbon with one another within just a few years. The much larger deep-ocean reservoir is partly isolated from the surface reservoirs by the thermocline and exchanges carbon with the surface ocean and the atmosphere over hundreds of years. The carbon buried in sediments and rocks interacts very slowly, moving in and out of the surface reservoirs only over hundreds of thousands of years or longer.

Plants grow on land if the conditions necessary for **photosynthesis** (the production of plant matter) are met: sunlight is needed to provide energy, and nutrients (mainly phosphorus and nitrogen) provide food for plant growth (Figure 2-33). With these conditions satisfied, plants draw CO_2 from the air and water from the soil to create new organic matter, while oxygen is liberated to the atmosphere:

$$6CO_2 + 6H_2O \underset{\text{Oxidation}}{\overset{\text{Photosynthesis}}{\rightleftarrows}} \underset{\text{Plants (organic C)}}{C_6H_{12}O_6} + 6O_2$$

During the time that plants grow by photosynthesis, and also during the time they are mature but no longer growing, plants take the water they need from the soil and give it back to the atmosphere (see Figure 2-33). This process, known as **transpiration** (and also as "respiration"), is a highly efficient way to return water vapor to the atmosphere, and it can occur at much faster rates than ordinary evaporation from vegetation-free ground.

After plants die (either by seasonal die-back or by reaching the end of their natural lifetime), oxygen is consumed in destroying their organic matter during **oxidation**. Oxidation can occur either through rapid burning (in fires) or by slow decomposition in the presence of oxygen, with the same ultimate result. Through either process, oxidation converts organic carbon back to inorganic form, as shown by the equation above.

2-11 Response of the Biosphere to the Physical Climate System

Trees, shrubs, and other plants accomplish most photosynthesis on land. CO_2 and sunlight are usually available over the continents, but the distributions of temperatures and rainfall critical to photosynthesis and plant life vary widely. To a large extent, rainfall (Figure 2-34A) determines the total amount of organic (live) matter present, called the **biomass**, and the predominant types of vegetation and associated organisms, called **biomes** (Figure 2-34B).

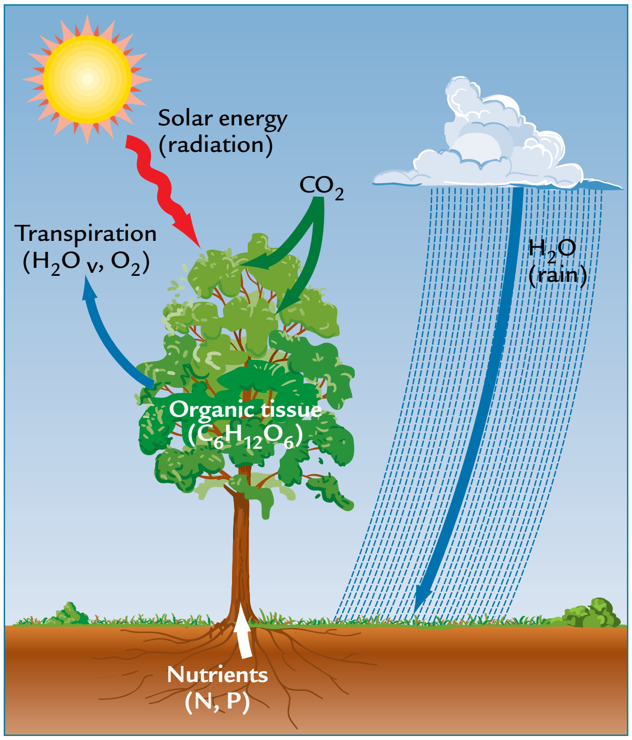


FIGURE 2-33 **Photosynthesis on land** Plant life on land uses sunlight, CO_2 and H_2O from the atmosphere, and soil nutrients in the process of photosynthesis. Plants also return water vapor (H_2O_v) and oxygen (O_2) to the atmosphere by transpiration. (Adapted from F. T. Mackenzie, *Our Changing Planet* [Englewood Cliffs, N.J.: Prentice-Hall, 1998].)

The tendency for rainfall to be abundant along the ITCZ produces tropical rain forest biomes with dense biomasses. Toward the dry subtropics, rain forests grade into **savanna** (scattered trees in a grassland setting) and then to the sparse scrub vegetation typical of deserts. Total biomass decreases toward the subtropics along with precipitation.

Large-biomass **hardwood forest** (maple, oak, hickory, and other leaf-bearing trees) occurs in the wetter portions of the middle latitudes (eastern North America, Europe, Asia) and on the upwind side of mountain ranges facing the ocean, while low-biomass biomes such as grasslands and desert scrub are found in drier interior regions in the rain shadow of mountain ranges (see Figure 2-20). **Conifer forest** (spruce and other trees with needles) dominates toward the higher

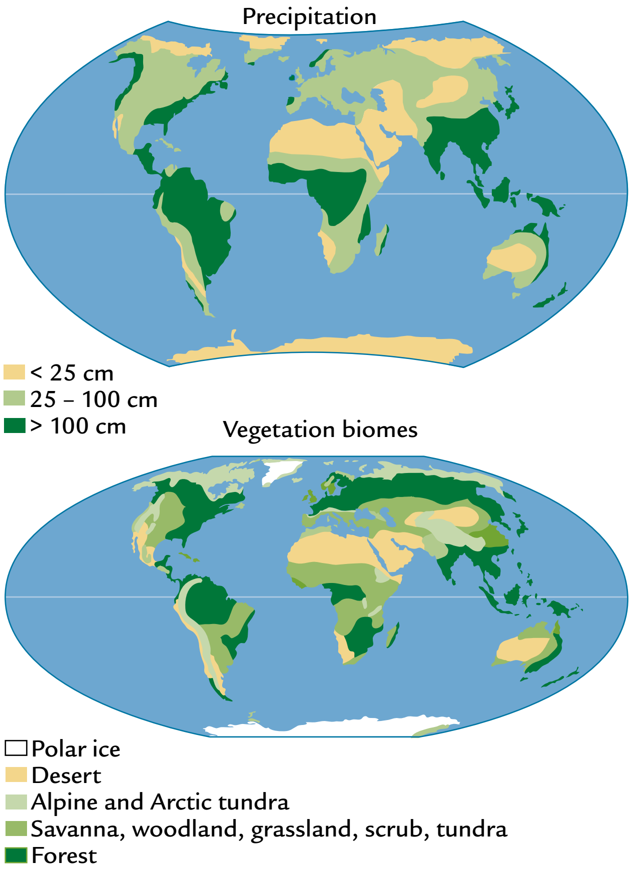


FIGURE 2-34 **Precipitation and vegetation** Global precipitation (top) is highest in the tropics and along mountain slopes that receive moisture-bearing winds from the ocean, and lowest in subtropical deserts and over polar ice. Vegetation biomes (bottom) largely reflect the patterns of precipitation, with high-biomass forests in regions of high precipitation and low evaporation. (Top: adapted from L. J. Battan, *Fundamentals of Meteorology* [Englewood Cliffs, N.J.: Prentice-Hall, 1979]; bottom: adapted from E. Bryant, *Climate Process and Change* [Cambridge: Cambridge University Press, 1998].)

latitudes of the northern hemisphere, but the fringes of the Arctic Ocean are surrounded by a wide band of scrubby **tundra** vegetation with low biomass above ground and large amounts of carbon stored below ground. Ice sheets are free of life forms except cold-tolerant bacteria.

Life in the oceans depends on a different combination of the same factors as on land. Obviously, water is abundantly available in the oceans, and CO_2 is plentiful in surface waters that exchange CO_2 with the atmosphere. In addition, light from the Sun is widely available in the upper layers of the ocean into which it penetrates (Figure 2-35).

With all these conditions favorable to photosynthesis, why isn't the surface ocean an enormous photosynthesis machine? The answer is simple: a lack of the nutrients nitrogen (N) and phosphorus (P). Nutrient food sources are scarce in most parts of the surface ocean.

A floating form of microscopic plant life called **phytoplankton** lives in the surface layers of the ocean and uses sunlight for photosynthesis. These minute organisms extract nutrients and incorporate them in the soft organic tissues of their bodies. Phytoplankton have short life spans (days to weeks), and when they die, they sink to deeper waters, leaving the surface layer depleted of nutrients. Thus the rates of photosynthesis in these sunlit surface waters are limited.

Initially near the surface, and mainly later at depths well below the surface, the slow decay and oxidation of the soft tissues of these sinking organisms releases nitrogen and phosphorus back into ocean water. Because most of these nutrients are released into and stay in the deeper ocean, their scarcity in surface waters limits the amount of life that can exist across most ocean areas.

In the few parts of the surface ocean where upwelling occurs, nutrients are more plentiful, and they result in greater **productivity**, or rates of photosynthesis by

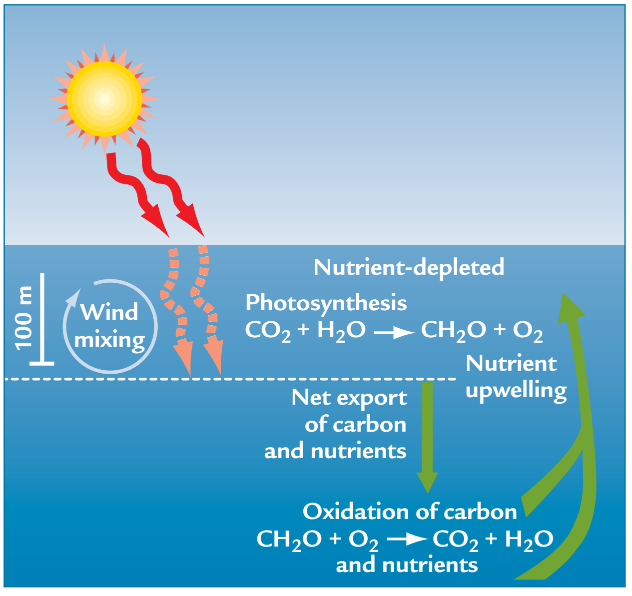


FIGURE 2-35 **Photosynthesis in the ocean** Sunlight penetrating the surface ocean causes photosynthesis by microscopic plants. As they die, their nutrient-bearing organic tissue descends to the seafloor. Oxidation of this tissue at depth returns nutrients and inorganic carbon to the surface ocean in regions of upwelling.

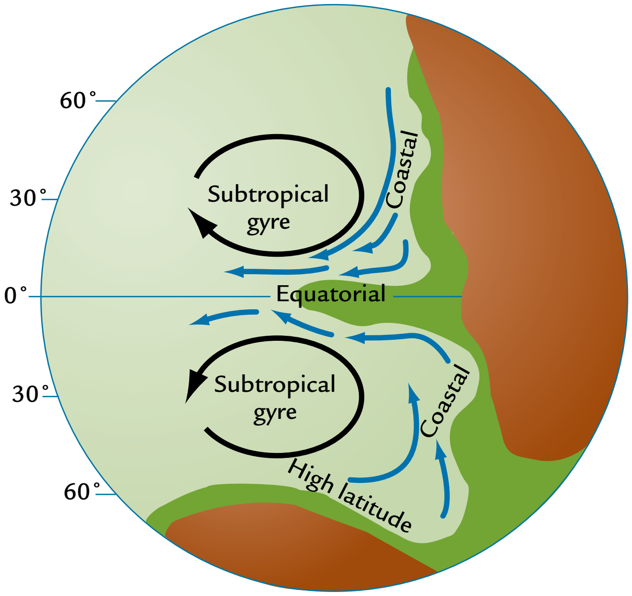


FIGURE 2-36 Ocean productivity The greatest amount of photosynthesis in the surface ocean occurs along shallow continental margins and in coastal, equatorial, and high-latitude regions where nutrients upwell from below.

phytoplankton (Figure 2-36). Wind-driven upwelling along some coastal margins returns nutrients to the surface from below and supplements nutrients delivered to continental shelves by rivers and resuspended during storms. As a result, surface waters near continental margins tend to be relatively productive. Upwelling in the eastern equatorial Pacific and Atlantic also increases rates of photosynthesis and productivity in those regions.

The Southern Ocean around Antarctica is another productive region (see Figure 2-36). Deep water from the Atlantic flows toward the surface in this area, bringing nutrients up from below. Strong winds mix these nutrients into the surface layers, producing the rich biomass of the Southern Ocean. Because sunlight is not plentiful, and because the long season of sea-ice cover limits the amount of time in which photosynthesis can occur, nutrients in the surface waters of the Southern Ocean are never depleted, even when productivity is highest.

2-12 Effects of the Biosphere on the Climate System

Life affects climate in many ways. One way is by providing positive feedback to physical processes that affect climate (Box 2-6). A second way is through changes in the amount of greenhouse gases in the atmosphere, especially carbon dioxide (CO_2) and methane (CH_4).

As we will see in later parts of this book, *all* the exchanges of carbon shown in Figure 2-32 have affected atmospheric CO_2 and climate, but at different time scales. The slow movement of carbon into and out of rock reservoirs affects atmospheric CO_2 and climate on the tectonic (million-year or longer) time scale, explored in Part II. Somewhat faster exchanges between the surface and deep ocean reservoirs affect climate on the orbital (ten-thousand-year) time scale, examined in Parts III and IV. Rapid exchanges among the surface reservoirs (vegetation, the ocean, and the atmosphere) affect climate over the shorter time scales reviewed in Parts IV and V.

Atmospheric CO_2 trends measured over the last four decades show two superimposed effects (Figure 2-37A). Each year a small drop in CO_2 values occurs in April–May and a comparable rise the following September–October. This oscillation reflects cycling of vegetation in the northern hemisphere: CO_2 is taken from the air by plant photosynthesis every spring and released by oxidation every autumn. The signal follows the tempo of the northern rather than the southern hemisphere because most of Earth's land (and land vegetation) lies north of the equator.

The second trend evident in the CO_2 curve is its gradual overall increase (see Figure 2-37A). This increase results mainly from burning of fossil-fuel carbon and secondarily from **deforestation** (clearing of vegetation from the land), which releases carbon to the atmosphere through burning and oxidation. The rapid increase in consumption of fossil fuels by humans over the last two centuries has tapped into huge reservoirs of coal, oil, and gas in rocks that naturally release their carbon at slow rates and has greatly accelerated these rates (Part V).

Methane (CH_4) is a second important atmospheric greenhouse gas, although far less plentiful than CO_2. It has many sources, including swampy lowland bogs, rice paddies, the stomachs and bowels of cows digesting vegetation, and termites. Common to all these CH_4 sources is the decay of organic matter in an oxygen-free environment. At the end of the twentieth century, methane concentrations in the atmosphere had risen by well over a factor of 2 above their natural (preindustrial) level to above 1700 parts per billion (Figure 2-37B). This recent increase in methane is the result of human activities (Part V).

As we noted earlier, both CO_2 and CH_4 trap part of Earth's back radiation, keep the heat in the atmosphere, and make Earth warmer than it would otherwise be. This warming in turn activates the positive feedback effect of water vapor (H_2O_v), the most important greenhouse gas. The combined effects of these three greenhouse gases in the recent past and near future are the focus of the last part of this book.

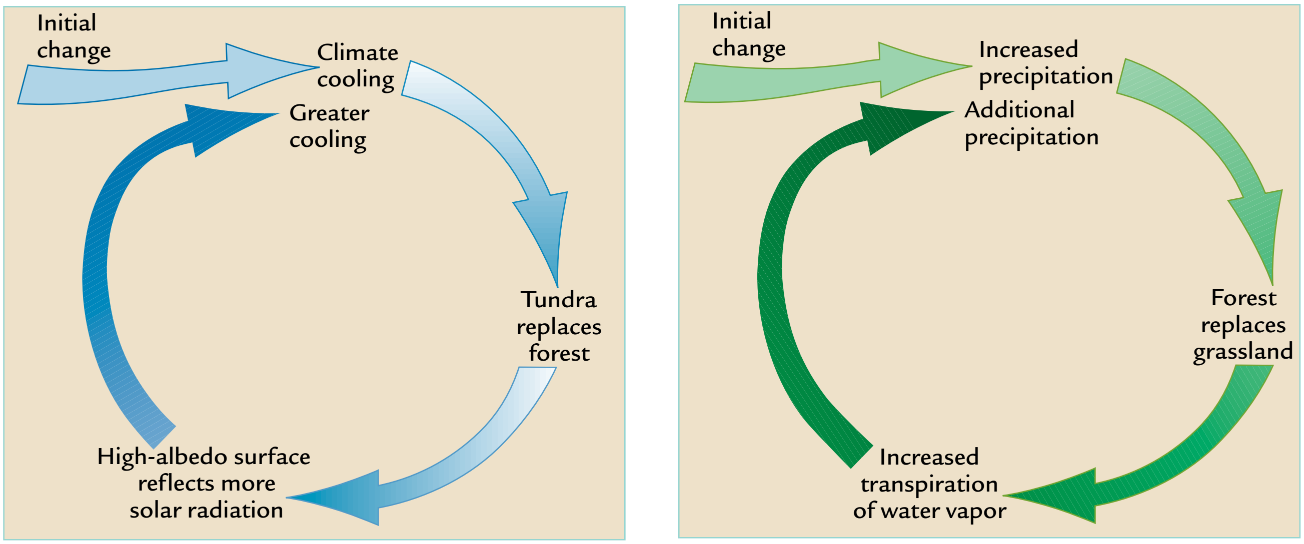
BOX 2-6 Climate Interactions and Feedbacks

Vegetation–Climate Feedbacks

The type of vegetation covering a land region can affect its average albedo. The two major types of vegetation in the Arctic, spruce forest and circumarctic tundra, interact in different ways with freshly fallen snow and produce surfaces with very different albedos. Snow that falls on tundra covers what little scrub vegetation exists and creates a high-albedo surface that reflects most incoming solar radiation. Snow that falls on spruce forests is blown from the trees and falls to the ground, allowing the dark-green surface of the exposed treetops to absorb most incoming solar radiation.

When climate cools, these contrasts in albedo produce an important positive feedback. Tundra gradually advances southward and replaces spruce forest, expanding Earth's high-albedo surface area. With more solar radiation reflected from this surface, the reduction in absorbed heat leads to further cooling. This process is called **vegetation–albedo feedback**. This same positive feedback works during times of climate warming: as forest replaces tundra, more solar heat is absorbed, and the climate warms even more.

A second type of vegetation feedback depends on the way vegetation recycles water. Land vegetation draws water needed for photosynthesis from the ground. Some of the water is handed off to the atmosphere as water vapor during times when the plants are actively extracting CO_2 from the air. Trees transpire much larger volumes of water vapor than grass or desert scrub, and this contrast is responsible for the second kind of positive feedback. When climate becomes wetter, forests gradually replace grasslands in some regions. The trees transpire more water vapor back to the atmosphere, thus increasing the amount available for rainfall. This positive feedback (called **vegetation–precipitation feedback**) works in the reverse sense when climate dries.



A Vegetation-albedo feedback

B Vegetation-precipitation feedback

Vegetation–climate feedbacks (A) When high-latitude climate cools, replacement of spruce forest by tundra raises the reflectivity (albedo) of the land in winter and causes additional cooling as a positive feedback. (B) When climate becomes wetter, replacement of grasslands by trees increases the release of water vapor back to the atmosphere and causes increases in local rainfall as a positive feedback.

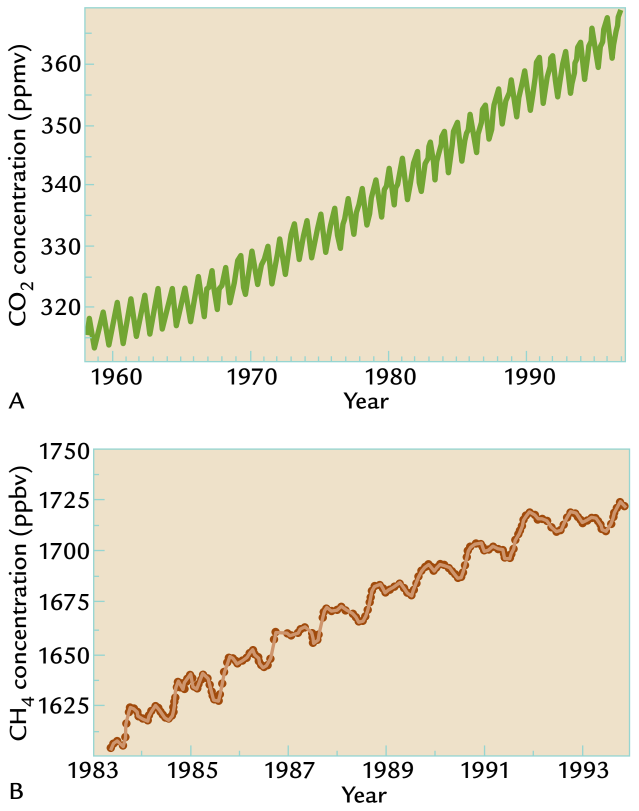


FIGURE 2-37 **Recent increases in carbon dioxide and methane** Instrument measurements record rapid rises of the greenhouse gases CO_2 (A) and CH_4 (B) in recent years. The gases are measured in parts per million by volume (ppmv) and parts per billion by volume (ppbv). (A: adapted from H. H. Friedli et al., "Ice Core Record of the $^{13}C/^{12}C$ Ratio of Atmospheric CO_2 in the Past Two Centuries," *Nature* 324 [1986]: 237–38; B: after M. A. K. Khalil and R. A. Rasmussen, "Atmospheric Methane: Trends over the Last 10,000 Years," *Atmospheric Environment* 21 [1987]: 2445–52.)

Key Terms

electromagnetic radiation (p. 19)
electromagnetic spectrum (p. 19)
shortwave radiation (p. 19)
back radiation (p. 20)
longwave radiation (p. 20)
greenhouse effect (p. 21)
troposphere (p. 22)
stratosphere (p. 22)
albedo (p. 22)
albedo–temperature feedback (p. 25)
heat capacity (p. 26)
calories (p. 26)
specific heat (p. 26)
hydrologic cycle (p. 27)
thermal inertia (p. 28)
sensible heat (p. 28)
convection (p. 29)
latent heat (p. 29)
latent heat of melting (p. 29)
latent heat of vaporization (p. 29)
dew point (p. 30)
saturation vapor density (p. 30)
water vapor feedback (p. 31)
adiabatic (p. 32)
lapse rate (p. 32)
Hadley cell (p. 33)
intertropical convergence zone (ITCZ) (p. 34)
monsoon (p. 35)
Coriolis effect (p. 36)
orographic precipitation (p. 38)
jet streams (p. 39)
gyres (p. 39)
Gulf Stream (p. 40)
North Atlantic Drift (p. 40)
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salinity (p. 41)
salt rejection (p. 41)
North Atlantic deep water (p. 42)
Antarctic bottom water (p. 42)
Antarctic intermediate water (p. 42)
Mediterranean overflow water (p. 42)
upwelling (p. 43)
mountain glaciers (p. 44)
continental ice sheets (p. 44)
ice domes (p. 45)
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ice streams (p. 45)
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productivity (p. 49)
deforestation (p. 50)
vegetation–albedo feedback (p. 51)
vegetation–precipitation feedback (p. 51)

Review Questions

1. How does solar radiation arriving on Earth differ from the back radiation emitted by Earth?
2. What kind of radiation is trapped by greenhouse gases? What is the effect on Earth's climate?
3. What different and opposing roles do clouds play in the climate system?
4. How does reflection of solar radiation from Earth's surface add to the effects of uneven solar heating in creating a pole-to-equator heat imbalance?
5. What processes cause air to rise from Earth's surface?

6. What causes the monsoon circulation to reverse from summer to winter?
7. Describe the main pathway by which heat in the atmosphere is transported toward the poles.
8. Why does rain fall on the sides of mountains in the path of winds from nearby oceans?
9. How do low-level winds create spinning gyres in the subtropical oceans?
10. Why does deep water form today at higher latitudes?
11. What effect does the formation of sea ice have on the overlying atmosphere?
12. What parts of ice sheets gain and lose mass? Why?
13. How closely does land vegetation (and total biomass) follow global precipitation trends?
14. What regions of the ocean are most productive? Why?
15. Describe two positive feedback processes discussed in this chapter.

Additional Resources

Barry, R. G., and R. J. Chorley. 1998. *Atmosphere, Weather, and Climate*. New York: Routledge.

Thurman, H. V. 1991. *Introductory Oceanography*. 6th ed. New York: Macmillan.