# 2 Earth's Climate System

**Climate is the state factor that most strongly governs the global distribution of terrestrial biomes. This chapter provides a general background on the functioning of the climate system and its interactions with atmospheric chemistry, ocean, and land.**

## Introduction

**Climate exerts a key control over the functioning of Earth's ecosystems.** Temperature and water availability govern the rates of many biological and chemical reactions that in turn control critical ecosystem processes. These processes include the production of organic matter by plants, its decomposition by microbes, the weathering of rocks, and the development of soils. Understanding the causes of temporal and spatial variation in climate is therefore critical to understanding the global pattern of ecosystem processes.

The amount of incoming solar radiation, the chemical composition and dynamics of the atmosphere, and the surface properties of Earth determine climate and climate variability. The circulation of the atmosphere and ocean influences the transfer of heat and moisture around the planet and thus strongly influences climate patterns and their variability in space and time. This chapter describes the global energy budget and outlines the roles that the atmosphere, ocean, and land surface play in the redistribution of energy to produce observed patterns of climate and ecosystem distribution.

## A Focal Issue

**Human activities are modifying Earth's climate, thereby changing fundamental controls over ecosystem processes throughout the planet, often to the detriment of society**. Some climatic changes subtly alter the rates of ecosystem process, but other changes, such as the frequency of severe storms have direct devastating effects on society. Climate warming, for example, increases sea-surface temperature, which increases the energy transferred to tropical storms (Fig. 2.1). Although no individual storm can be attributed to climate change, the intensity of tropical storms may increase (IPCC 2007). Other expected effects of climate change include more frequent droughts in drylands such as sub-Saharan Africa, more frequent floods in wet climates and in low-lying coastal zones, warmer weather in cold climates, and more extensive wildfires in fire-prone forests. What determines the distribution of Earth's major climate zones? Why is climate changing, and why do regions differ in the climatic changes they experience? An understanding of the causes of temporal and spatial variation in the climate system facilitates predictions of the changes that are likely to occur in particular places.

## Earth's Energy Budget

**The sun is the source of the energy available to drive Earth's climate system.** The wavelength of energy produced by a body depends on its

F.S. Chapin, III et al., *Principles of Terrestrial Ecosystem Ecology*,
DOI 10.1007/978-1-4419-9504-9_2, © Springer Science+Business Media, LLC 2011

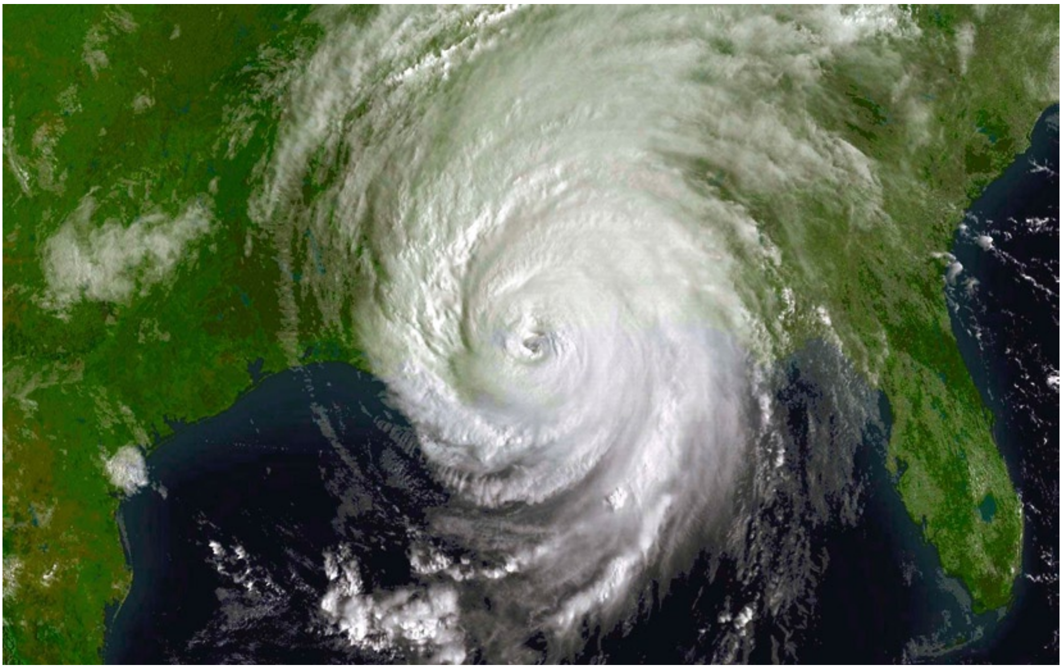

**Fig. 2.1** Satellite view of Hurricane Katrina over coastal Louisiana. This tropical storm flooded New Orleans in 2005, killing approximately 1,570 people and causing $40–50 billion of damage. Human-caused ecological changes in coastal Louisiana contributed to the impact of the hurricane. Climate warming is expected to increase the frequency of severe tropical storms like Hurricane Katrina. Image courtesy of NOAA (http://www.katrina.noaa.gov/satellite/satellite.html)

temperature. Because it is hot (6,000°C), the sun emits most energy as high-energy **shortwave radiation** with wavelengths of 0.2–4.0 µm (Fig. 2.2). These include ultraviolet (UV; 8% of the total), visible (39%), and near-infrared (53%) radiation. On average, about 30% of the incoming shortwave radiation is reflected back to space, due to **backscatter** (reflection) from clouds (16%); air molecules, dust, and haze (6%); and Earth's surface (7%; Fig. 2.3). Another 23% of the incoming shortwave radiation is absorbed by the atmosphere, especially by ozone in the upper atmosphere and by clouds and water vapor in the lower atmosphere. The remaining 47% reaches Earth's surface as direct or diffuse radiation and is absorbed there (Trenberth et al. 2009).

Earth also emits radiation, like all bodies, but, due to its lower surface temperature (about 15°C), Earth emits most energy as low-energy **longwave radiation** (Fig. 2.2). Although the atmosphere transmits about half of the incoming shortwave radiation to Earth's surface, **radiatively active gases** (water vapor, $CO_2$, $CH_4$, $N_2O$ and industrial products like chlorofluorocarbons [CFCs]) absorb 90% of the outgoing longwave radiation (Fig. 2.3). Of the approximately 10% of longwave radiation that escapes to space, most is in wavelengths where longwave absorption by the atmosphere is small (referred to as atmospheric windows; Fig. 2.2). The energy absorbed by radiatively active gases in the atmosphere is re-radiated in all directions (Fig. 2.3). The portion that is directed back toward the surface contributes to the warming of the planet, a phenomenon known as the **greenhouse effect**. Without these longwave-absorbing gases in the atmosphere, the average temperature at Earth's surface would be about 33°C lower than it is today, and Earth would probably not support life, except perhaps at hydrothermal vents in the deep ocean.

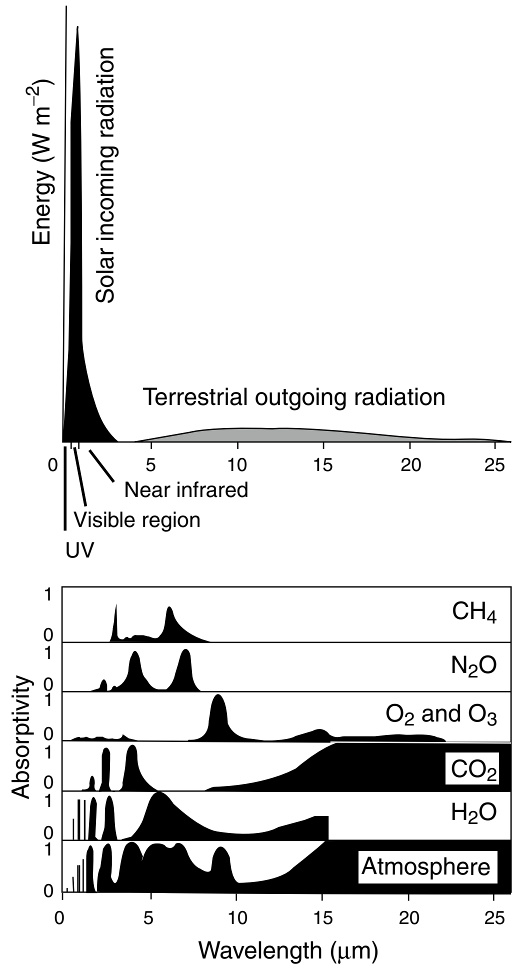


**Fig. 2.2** The spectral distribution of solar and terrestrial radiation and the absorption spectra of the major radiatively active gases and of the total atmosphere. These spectra show that the atmosphere absorbs a larger proportion of terrestrial radiation than solar radiation, explaining why the atmosphere is heated from below. Redrawn from Sturman and Tapper (1996) and Barry and Chorley (2003)

As a global long-term average, Earth is normally close to a state of radiative balance, meaning that it emits as much energy back to space (as longwave radiation) as it absorbs. However, human activities are changing the composition of the atmosphere enough to increase the heat retained by the planet, as described later. Assuming balance, the longwave radiation emitted to space must equal the sum of the solar radiation absorbed by both the surface and the atmosphere. The atmosphere is heated by longwave absorption by radiatively active gases and by the absorption of some incoming (shortwave) solar radiation; it is also heated from the surface by non-radiative fluxes of heat that are carried upward by atmospheric **turbulence** (mixing). These include **latent heat flux**, where heat that evaporates water at the surface is subsequently released to the atmosphere as air parcels rise and cool, and the water vapor condenses, forming clouds and precipitation. There is also an upward transfer of heat that is conducted from the warm surface to the air immediately above it and then moved upward by convection of the atmosphere as thermals (**sensible heat flux**). These heat sources collectively sustain the longwave emission to space, as well as a large flux of longwave radiation from the lower atmosphere back to Earth's surface. This back radiation to the surface represents the natural greenhouse effect described earlier.

Long-term records of atmospheric gases, obtained from atmospheric measurements since the 1950s and from air bubbles trapped in glacial ice, show large increases in the major radiatively active gases ($CO_2$, $CH_4$, $N_2O$, and CFCs) since the beginning of the industrial revolution 250 years ago (see Fig. 14.7). Human activities such as fossil fuel burning, industrial activities, animal husbandry, and fertilized and irrigated agriculture contribute to these increases (see Chap. 14). As concentrations of these gases rise, the atmosphere traps more of the longwave radiation emitted by Earth, enhancing the greenhouse effect and increasing Earth's surface temperature. A small imbalance thus exists in the radiative flows shown in Fig. 2.3, estimated to be about 0.26% of the incoming radiation. Most of this excess energy is absorbed in the ocean, causing water to expand and sea level to rise. The warming caused by radiative imbalance also contributes to widespread melting of glaciers and ice sheets (Greenland and Antarctica) and arctic sea ice.

The globally averaged annual energy budget outlined above gives a sense of the critical factors controlling the global climate system. Regional climates, however, reflect spatial variation in energy exchange and in lateral heat transport by the atmosphere and the ocean. Earth is heated more strongly at the equator than at the poles and

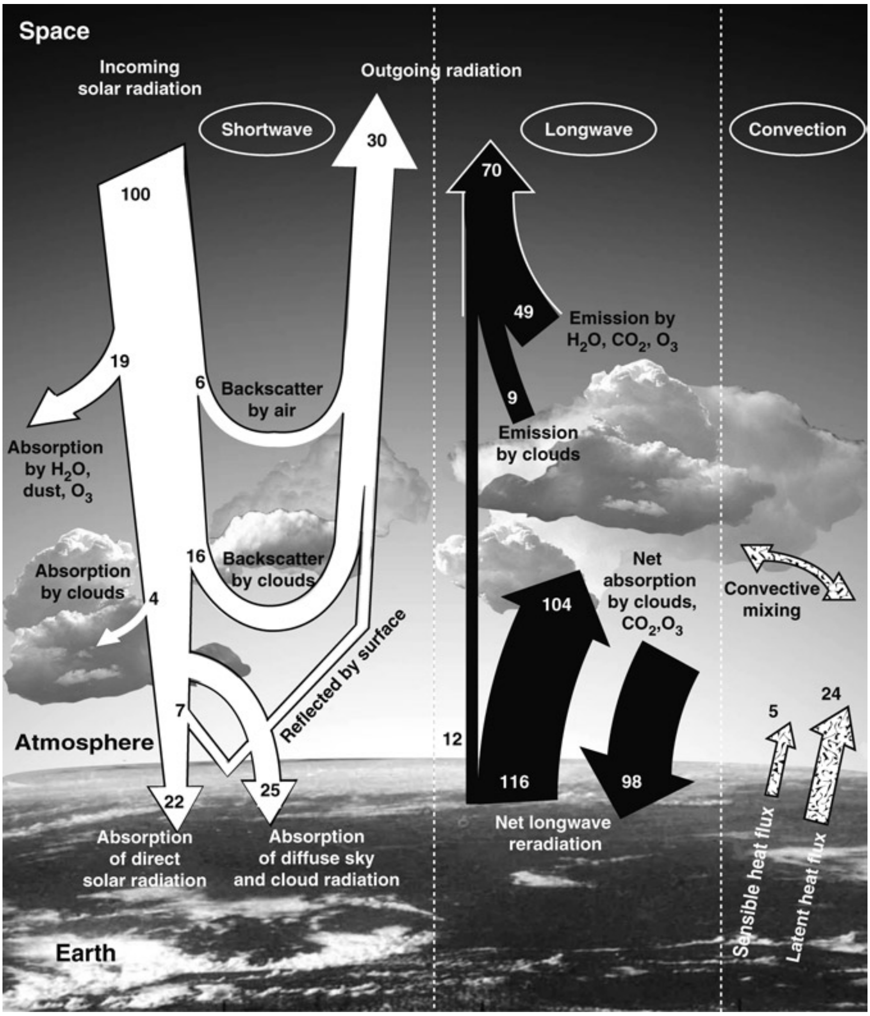


**Fig. 2.3** The average annual global energy balance during 2000–2004 for the Earth-atmosphere system. The numbers are percentages of the energy received from incoming solar radiation. At the top of the atmosphere, the incoming solar radiation (100 units or 341 W $m^{-2}$ [global average]) is balanced by reflected shortwave (30 units) and emitted longwave radiation (70 units). Within the atmosphere, the absorbed shortwave radiation (23 units) and absorbed longwave radiation (104 units) and latent + sensible heat flux (29 units) are balanced by longwave emission to space (58 units) and longwave emission to Earth's surface (98 units). At Earth's surface, the incoming shortwave (47 units) and incoming longwave radiation (98 units) are balanced by outgoing longwave radiation (116 units) and latent + sensible heat flux (29 units). Data are from Trenberth et al. (2009)

rotates on an axis that is tilted relative to the plane of its orbit around the sun. Its continents are spread unevenly over the surface, and its atmospheric and oceanic chemistry and physics are dynamic and spatially variable. A more thorough understanding of the atmosphere and ocean is therefore needed to understand the fate and processing of energy and its consequences for Earth's ecosystems.

## The Atmospheric System

### Atmospheric Composition and Chemistry

**The chemical composition of the atmosphere determines its role in Earth's energy budget.** The atmosphere is like a giant reaction flask,

**Table 2.1** Major chemical constituents of the atmosphere

| Compound | Formula | Concentration (%) |
|---|---|---|
| Nitrogen | $N_2$ | 78.082 |
| Oxygen | $O_2$ | 20.945 |
| Argon | Ar | 0.934 |
| Carbon dioxide | $CO_2$ | 0.039 |

Data from Schlesinger (1997) and IPCC (2007)

containing thousands of different chemical compounds in gas and particulate forms, undergoing slow and fast reactions, dissolutions, and precipitations. These reactions control the composition of the atmosphere and many of its physical processes, such as cloud formation and energy absorption. The associated heating and cooling, together with the uneven distribution of solar radiation, generate dynamical motions crucial for energy redistribution.

More than 99.9% by volume of Earth's dry atmosphere is composed of nitrogen, oxygen, and argon (Table 2.1). Carbon dioxide ($CO_2$), the next most abundant gas, accounts for only 0.039% of the atmosphere. These percentages are quite constant around the world and up to 80 km in height above the surface. That homogeneity reflects the fact that these gases have long **mean residence times** (MRT) in the atmosphere. MRT is calculated as the total mass divided by the flux into or out of the atmosphere over a given time period. Nitrogen has an MRT of 13 million years, $O_2$ 10,000 years, and $CO_2$ 5 years (see Chap. 14). Some of the most important radiatively active gases, such as $CO_2$, nitrous oxide ($N_2O$), methane ($CH_4$), and CFCs, react relatively slowly in the atmosphere and have residence times of years to decades. Other gases are much more reactive and have residence times of days to months. Highly reactive gases make up less than 0.001% of the dry volume of the atmosphere and are quite variable in time and space. These reactive gases influence ecological systems through their roles in nutrient delivery, smog, acid rain, and ozone depletion (Graedel and Crutzen 1995). Water vapor is also quite reactive and highly variable both seasonally and spatially.

MRT provides a reasonable estimate of the lifetime of a gas in the atmosphere for those gases like $CH_4$ and $N_2O$ that undergo irreversible reactions to produce breakdown products. $CO_2$, however, is not "destroyed" when it is absorbed by the ocean or the biosphere, but continues to exchange with the atmosphere. If all fossil fuel emissions ceased instantly today, the excess fossil-fuel $CO_2$ in the atmosphere (about 35% higher than the "natural" background) would decline by 50% within 30 years, another 20% within a few centuries, but the remaining 30% excess $CO_2$ would remain in the atmosphere for thousands of years (IPCC 2007; Archer et al. 2009; see Chap. 14). This will create, from the perspective of a human lifetime, a permanently warmer world (Solomon et al. 2009). The magnitude of this climate warming will depend on the rates at which people reduce their emissions of fossil fuels and other trace gases.

Some atmospheric gases are critical for life. Photosynthetic organisms use $CO_2$ in the presence of light to produce organic matter that eventually becomes the basic food source for almost all animals and microbes (see Chaps. 5–7). Most organisms also require oxygen for metabolic respiration. Di-nitrogen ($N_2$) makes up 78% of the atmosphere. It is unavailable to most organisms, but nitrogen-fixing bacteria convert it to biologically available nitrogen that is ultimately used by all organisms to build proteins (see Chap. 9). Other gases, such as carbon monoxide (CO), nitric oxide (NO), nitrous oxide ($N_2O$), methane ($CH_4$), and volatile organic carbon compounds like terpenes and isoprene, are the products of plant and microbial activity. Some, like tropospheric ozone ($O_3$), are produced chemically in the atmosphere as products of chemical reactions involving both **biogenic** (biologically produced) and anthropogenic gases, and can, at high concentrations, damage plants, microbes, and people.

The atmosphere also contains **aerosols**, which are small solid or liquid particles suspended in air. Some aerosol particles arise from volcanic eruptions and from blowing dust and sea salt. Others are produced by reactions with gases from pollution sources and biomass burning. Some aerosols act as **cloud condensation nuclei** around which water vapor condenses to form cloud droplets. Aerosols, together with gases and clouds and characteristics of the surface, determine the reflectivity (**albedo**) of the planet and therefore exert major control over the energy budget and hence climate. The scattering (reflection) of incoming shortwave radiation by some aerosols reduces the radiation reaching Earth's surface and tends to cool the climate. For example, the sulfur dioxide injected into the atmosphere by the volcanic eruption of Mt. Pinatubo in the Philippines in 1991 and the subsequent creation of sulfate aerosols cooled Earth's climate for about a year.

Clouds have complex effects on Earth's radiation budget. All clouds have a high albedo, and hence reflect much more incoming shortwave radiation than does the darker Earth surface. Clouds, however, are composed of water droplets and ice crystals, which are very efficient absorbers of longwave radiation impinging on them from Earth's surface. The first process (reflecting shortwave radiation) has a cooling effect by reflecting incoming energy back to space. The second effect (absorbing longwave radiation) has a warming effect, by preventing energy from escaping to space. The balance of these two effects depends on many factors, including cloud type, temperature, thickness, and height. The reflection of shortwave radiation usually dominates the balance in high clouds, causing cooling, whereas the absorption and re-emission of longwave radiation generally dominates in low clouds, producing a warming effect. While clouds have a net cooling effect globally by reducing solar input, they have a net warming effect in the Arctic and Antarctic, where heat loss predominates.

## Atmospheric Structure

**Atmospheric pressure and density decline with height above Earth's surface.** The average vertical structure of the atmosphere defines four relatively distinct layers characterized by their temperature profiles. The atmosphere is highly compressible, and gravity keeps most of the mass of the atmosphere close to Earth's surface. Pressure, which is related to the mass of the overlying atmosphere, decreases logarithmically

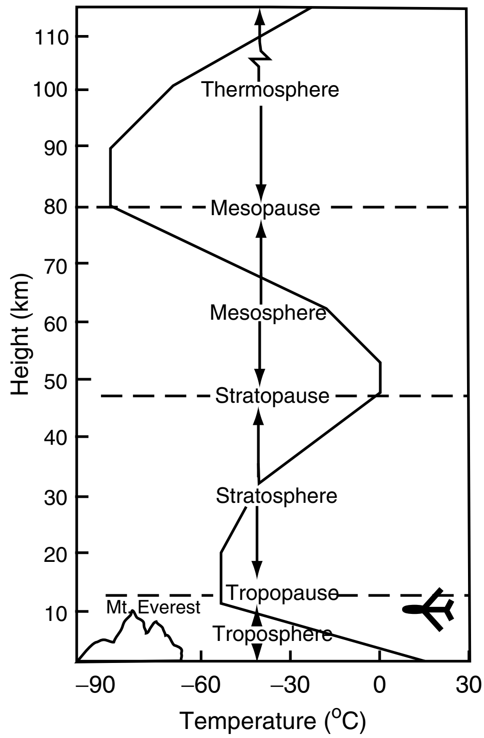


**Fig. 2.4** Average thermal structure of the atmosphere, showing the vertical gradients in temperature in Earth's major atmospheric layers. Redrawn from Schlesinger (1997)

with height, as does the density of air. As one moves above the surface toward lower pressure and density, the vertical pressure gradient also decreases. Furthermore, because warm air is less dense than cold air, pressure falls off with height more slowly for warm than for cold air.

The **troposphere** is the lowest atmospheric layer (Fig. 2.4). It contains 75% of the mass of the atmosphere and is heated primarily from the bottom by sensible and latent heat fluxes and by longwave radiation from Earth's surface. Because air heated at the surface cools as it rises and expands, temperature decreases with height in the troposphere.

Above the troposphere is the **stratosphere**, which, unlike the troposphere, is heated from the top, resulting in an increase in temperature with height (Fig. 2.4). Absorption of UV radiation by **ozone** ($O_3$) in the upper stratosphere warms the air.

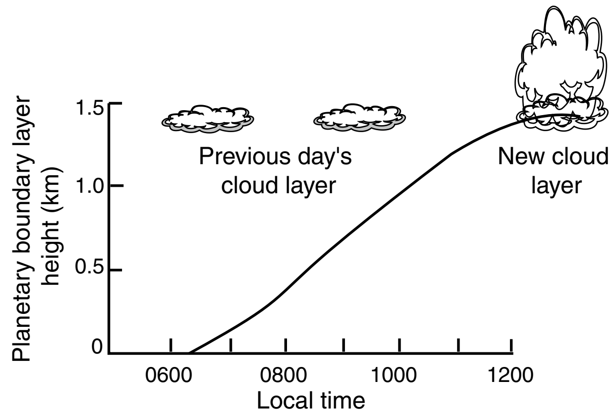


**Fig. 2.5** Growth in height of the planetary boundary layer (PBL) above the plant canopy between 6 a.m. and noon in the Amazon Basin on a day without thunderstorms. The increase in surface temperature drives evapotranspiration and convective mixing, which causes the boundary layer to increase in height until the rising air becomes cool enough that water vapor condenses to form clouds. Redrawn from Matson and Harriss (1988)

Ozone is most concentrated in the upper stratosphere due to a balance between the availability of shortwave UV necessary to split molecules of $O_2$ into atomic O and a high enough density of molecules to bring about the required collisions between atomic O and molecular $O_2$ to form $O_3$. The ozone layer protects the biota at Earth's surface from UV radiation. Biological systems are very sensitive to UV radiation because it damages DNA, which contains the information needed to drive cellular processes. The concentration of ozone in the stratosphere has been declining due to the production and emission of chlorofluorocarbon chemicals (CFCs) that destroy stratospheric ozone, particularly at the poles. This results in ozone "holes," regions where the transmission of UV radiation to Earth's surface is increased. Because the south polar region is colder and has more stratospheric clouds in which ozone-destroying reactions occur, the ozone hole over Antarctica is much larger than its arctic counterpart. Slow mixing between the troposphere and the stratosphere allows CFCs and other compounds to reach and accumulate in the ozone-rich stratosphere, where they have long residence times.

Above the stratosphere is the **mesosphere**, where temperature again decreases with height. The uppermost layer of the atmosphere, the **thermosphere**, begins at approximately 80 km and extends into space. The thermosphere has a very small fraction of the atmosphere's total mass, composed primarily of O and N atoms that can absorb energy of extremely short wavelengths, again causing an increase in heating with height (Fig. 2.4). The mesosphere and thermosphere have relatively little impact on the biosphere.

The troposphere is the atmospheric layer where most weather occurs, including thunderstorms, snowstorms, hurricanes, and high and low pressure systems. The troposphere is therefore the portion of the atmosphere that directly responds to and affects ecosystem processes. The **tropopause** is the boundary between the troposphere and the stratosphere. It occurs at a height of about 16 km in the tropics, where tropospheric temperatures are highest and hence where pressure falls off most slowly with height, and at about 9 km in polar regions, where tropospheric temperatures are lowest. The height of the tropopause varies seasonally, being lower in winter than in summer.

The **planetary boundary layer** (PBL) is the lower portion of the troposphere in which air is mixed by surface heating, which creates convective turbulence, and by mechanical turbulence as air moves across Earth's rough surface. The PBL increases in height during the day largely due to convective turbulence. The PBL mixes more rapidly with the free troposphere when the atmosphere is disturbed by storms. The boundary layer over the Amazon Basin, for example, generally grows in height until midday, when it is disrupted by convective activity (Fig. 2.5). The PBL becomes shallower at night when there is no solar energy to drive convective mixing. Air in the PBL is relatively isolated from the free troposphere and therefore functions like a chamber over Earth's surface. The changes in water vapor, $CO_2$, and other chemical constituents in the PBL therefore serve as an indicator of the biological and physiochemical processes occurring at the surface (Matson and Harriss 1988). The PBL in urban regions, for example, often has higher concentrations of pollutants than the cleaner, more stable air above. At night, gases emitted by the surface, such as $CO_2$ in natural ecosystems or pollutants in urban environments, often reach high concentrations because they are concentrated in a shallow boundary layer.

## Atmospheric Circulation

**The fundamental cause of atmospheric circulation is the uneven solar heating of Earth's surface.** The equator receives more incoming solar radiation than the poles because Earth is spherical. At the equator, the sun's rays are almost perpendicular to the surface at solar noon. At the lower sun angles characteristic of high latitudes, the sun's rays are spread over a larger surface area (Fig. 2.6), resulting in less radiation received per unit ground area. In addition, the sun's rays have a longer path through the atmosphere at high latitudes, so more of the incoming solar radiation is absorbed, reflected, or scattered before it reaches the surface. This unequal heating of Earth results in higher tropospheric temperatures in the tropics than at the poles, which in turn drives atmospheric circulation and transports atmospheric heat toward the poles. As a consequence of this, the input of shortwave solar radiation exceeds longwave radiation loss to space in the tropics, whereas longwave radiation loss exceeds solar input at temperate and high latitudes (Fig. 2.7).

Atmospheric circulation has both vertical and horizontal components (Fig. 2.8). Surface heating causes the surface air to expand and become less dense than surrounding air, so it rises. As air rises, the decrease in atmospheric pressure with height causes continued expansion, which decreases the average kinetic energy of air molecules, meaning that the rising air becomes cooler. Cooling causes condensation and precipitation because cool air has a lower capacity to hold water vapor than warm air. Condensation, in turn, releases latent heat, which can cause the rising air to remain warmer than surrounding air, so it continues to rise. The average **lapse rate** (the rate at which air temperature decreases with height) varies regionally depending on the strength of surface heating and the atmospheric moisture content but averages about 6.5°C $km^{-1}$.

Surface air rises most strongly at the equator because of the intense equatorial heating and the large amount of latent heat released as this moist tropical air rises, expands, cools, and releases heat by condensation of water vapor. This air often rises until it reaches the tropopause. The upward movement and expansion of equatorial air also creates a horizontal pressure gradient that causes the equatorial air aloft to flow horizontally from the equator toward the poles (Fig. 2.8). This poleward-moving air cools because of both emission of longwave radiation to space and mixing with cold air that moves toward the equator from the poles. In addition, the tropical air converges into a smaller volume as it moves poleward because the radius and surface area of Earth decrease from the equator toward the poles. Due to the cooling of the air and its convergence into a smaller volume, the density of air increases, creating a high pressure that causes upper air to subside and warm. Subtropical high-pressure zones typically have clear skies; the resulting high input of solar radiation drives abundant evaporation. This moist subtropical surface air moves back toward the equator to replace the rising equatorial air. Hadley proposed this model of atmospheric circulation in 1735, suggesting that there should be one large circulation cell in the northern hemisphere and another in the southern hemisphere, driven by atmospheric heating and uplift at the equator and subsidence at the poles. Based on observations, Ferrell proposed in 1865 the conceptual model that we still use today, although the actual dynamics are much more complex (Trenberth and Stepaniak 2003). This model describes atmospheric circulation as a series of three circulation cells in each hemisphere. (1) The **Hadley cell** is driven by expansion and uplift of equatorial

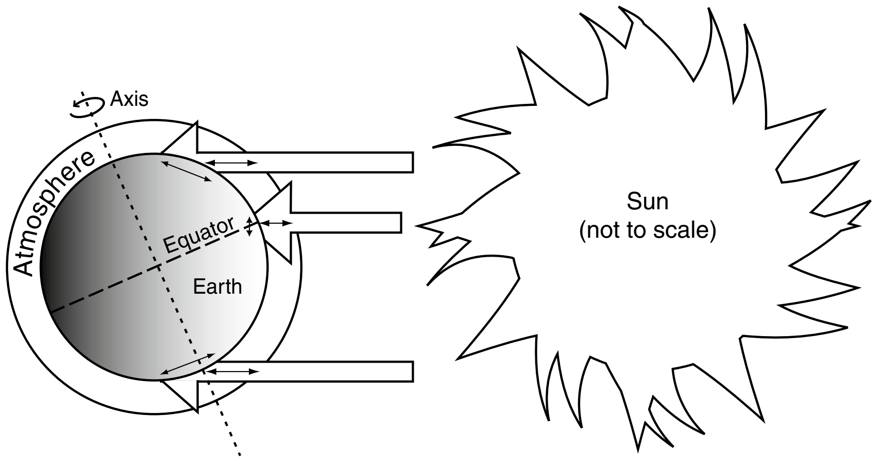


**Fig. 2.6** Atmospheric and angle effects on solar inputs to different latitudes. The *arrows* parallel to the sun's rays show the depth of the atmosphere that solar radiation must penetrate. The *arrows* parallel to Earth's surface show the surface area over which a given quantity of solar radiation is distributed. High-latitude ecosystems receive less radiation than those at the equator because radiation at high latitudes has a longer path length through the atmosphere and is spread over a larger ground area

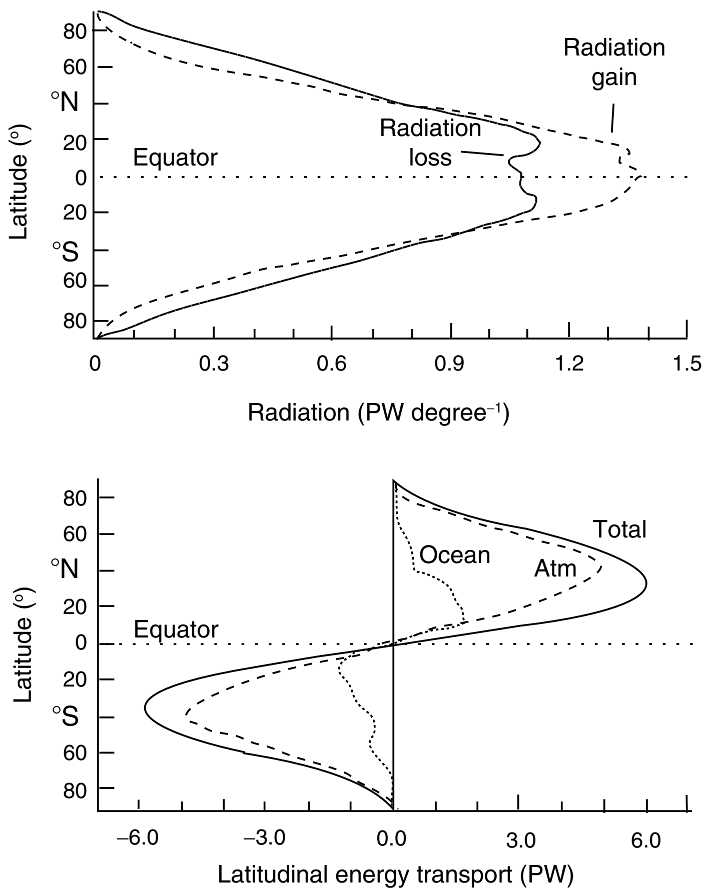


**Fig. 2.7** Latitudinal variation in heat input and loss to Earth (*top*; units PW [$10^{15}$ W] per degree of latitude) and in latitudinal heat transport by the ocean and the atmosphere (*bottom*; units PW). Redrawn from Fasullo and Trenberth (2008)

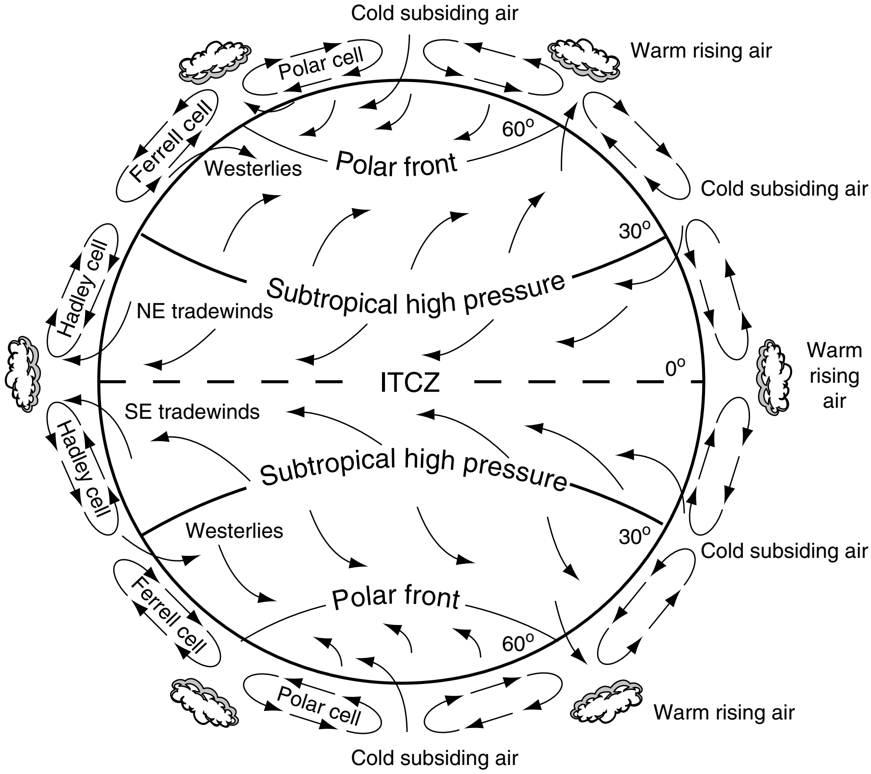


**Fig. 2.8** Earth's latitudinal atmospheric circulation is driven by rising air at the equator and subsiding air at the poles. These forces and the Coriolis effect produce three major cells of vertical atmospheric circulation (the Hadley, Ferrell, and polar cells). Air warms and rises at the equator due to intense heating. After reaching the tropopause, the equatorial air moves poleward to about 30°N and S latitudes, where it descends and either returns to the equator, forming the Hadley cell, or moves poleward. Cold dense air at the poles subsides and moves toward the equator until it encounters poleward-moving air at about 60°N and S. Here the air rises and moves either poleward to replace air that has subsided at the poles (the polar cell) or moves toward the equator to form the Ferrell cell. Also shown are the horizontal patterns of atmospheric circulation consisting of the prevailing surface winds (the easterly trade winds in the tropics and the westerlies in the temperate zones). The boundaries between these zones are either low-pressure zones of rising air (the ITCZ and the polar front) or high-pressure zones of subsiding air (the subtropical high-pressure belt and the poles)

air and subsidence of cool dense subtropical air. (2) The **polar cell** is driven by subsidence of cold converging air at the poles. (3) An intermediate **Ferrell cell** is driven indirectly by dynamical processes (Fig. 2.8). The Ferrell cell is actually the long-term average air movement of mid-latitude weather systems rather than a stable permanent atmospheric feature. The circular motion (eddies) of these mid-latitude weather systems produces a net poleward transport of heat. These three cells subdivide the atmosphere into several distinct circulations: tropical air masses between the equator and 30°N and S, temperate air masses between 30° and 60°N and S, and polar air masses between 60°N and S and the poles (Fig. 2.8).

**Earth's rotation causes winds to deflect to the right in the northern hemisphere and to the left in the southern hemisphere.** Earth and its atmosphere complete one rotation about Earth's axis every day. The direction of rotation is from west to east. Because the atmosphere in equatorial regions is further from Earth's axis of rotation than is the atmosphere at higher latitudes, it has a higher linear velocity than does polar air as it travels around Earth. As parcels of air move north or south, they tend to maintain their **angular**

**momentum** ($M_a$), just as a car tends to maintain its momentum, when you try to stop or turn on an icy road. This effect is summarized in the equation:

$$M_a = mvr \tag{2.1}$$

where $m$ is the mass, $v$ is the velocity, and $r$ is the radius of rotation. If the mass of a parcel of air remains constant, its velocity is inversely related to the radius of rotation (2.1). We know, for example, that a skater can increase her speed of rotation by pulling her arms close to her body, which reduces her effective radius. Air that moves from the equator toward the poles encounters a smaller radius of rotation around Earth's axis. Therefore, to conserve angular momentum, it moves more rapidly (i.e., moves from west to east *relative to Earth's surface*), as it moves poleward (Fig. 2.8). Conversely, air moving toward the equator encounters an increasing radius of rotation around Earth's axis and, to conserve angular momentum, moves more slowly (i.e., moves from east to west *relative to Earth's surface*). There is another effect at work. Air parcels moving eastward relative to the surface are subjected to a larger centrifugal force than parcels at rest with respect to the surface. While this extra centrifugal force acts outward from the axis of Earth's rotation, the fact that Earth's surface is curved means that a component of this centrifugal force is directed toward the equator. The opposite effect occurs if the air is moving east to west relative to the surface. Conservation of angular momentum and the centrifugal force represent the two components of the **Coriolis effect** that work together to deflect moving air parcels to the right in the northern hemisphere and to the left in the southern hemisphere. The Coriolis effect is a "pseudo force" that arises only because we view the motion of the atmosphere relative to Earth's rotating surface. The Coriolis effect explains why mid-latitude storms rotate clockwise (counterclockwise) in the northern (southern) hemisphere. The Coriolis effect also explains the rotation of the Hadley cells (Fig. 2.8).

The interaction of vertical and horizontal motions of the atmosphere creates Earth's **prevailing winds**, i.e., the most frequent wind directions. The direction of prevailing winds depends on whether air is moving toward or away from the equator. In the tropics, surface air in the Hadley cell moves from 30°N and S toward the equator, and the Coriolis effect causes these winds to blow from the east, forming easterly **tradewinds** (Fig. 2.8). The region where surface air from northern and southern hemispheres converges is called the **Intertropical Convergence Zone** (ITCZ). Here the rising air creates a zone with light winds and high humidity, known to early sailors as the **doldrums**. Subsiding air at 30°N and S latitudes also produces relatively light winds, known as the **horse latitudes**. The surface air that moves poleward from 30° to 60°N and S is deflected toward the east by the Coriolis effect, forming the prevailing **westerlies**, i.e., surface winds that blow from the west.

At the boundaries between the major cells of atmospheric circulation, relatively sharp gradients of temperature and pressure, together with the Coriolis effect, generate strong winds over a broad height range in the upper troposphere. These are the subtropical and polar **jet streams**. The Coriolis effect explains why these winds blow in a westerly direction, i.e., from west to east.

The locations of the ITCZ and of each circulation cell shifts seasonally because the zone of maximum solar radiation input varies from summer to winter due to Earth's 23.5° tilt with respect to the plane of its orbit around the sun. The seasonal changes in the location of these cells contribute to the seasonality of climate.

**The uneven distribution of land and the ocean on Earth's surface creates an uneven pattern of heating that modifies the general latitudinal trends in climate.** At 30°N and S, air descends more strongly over the cool ocean than over the relatively warm land because the air is cooler and more dense over the ocean than over the land. The greater subsidence over the ocean creates high-pressure zones over the Atlantic and Pacific (the Bermuda and Pacific highs, respectively) and over the Southern Ocean (Fig. 2.9).

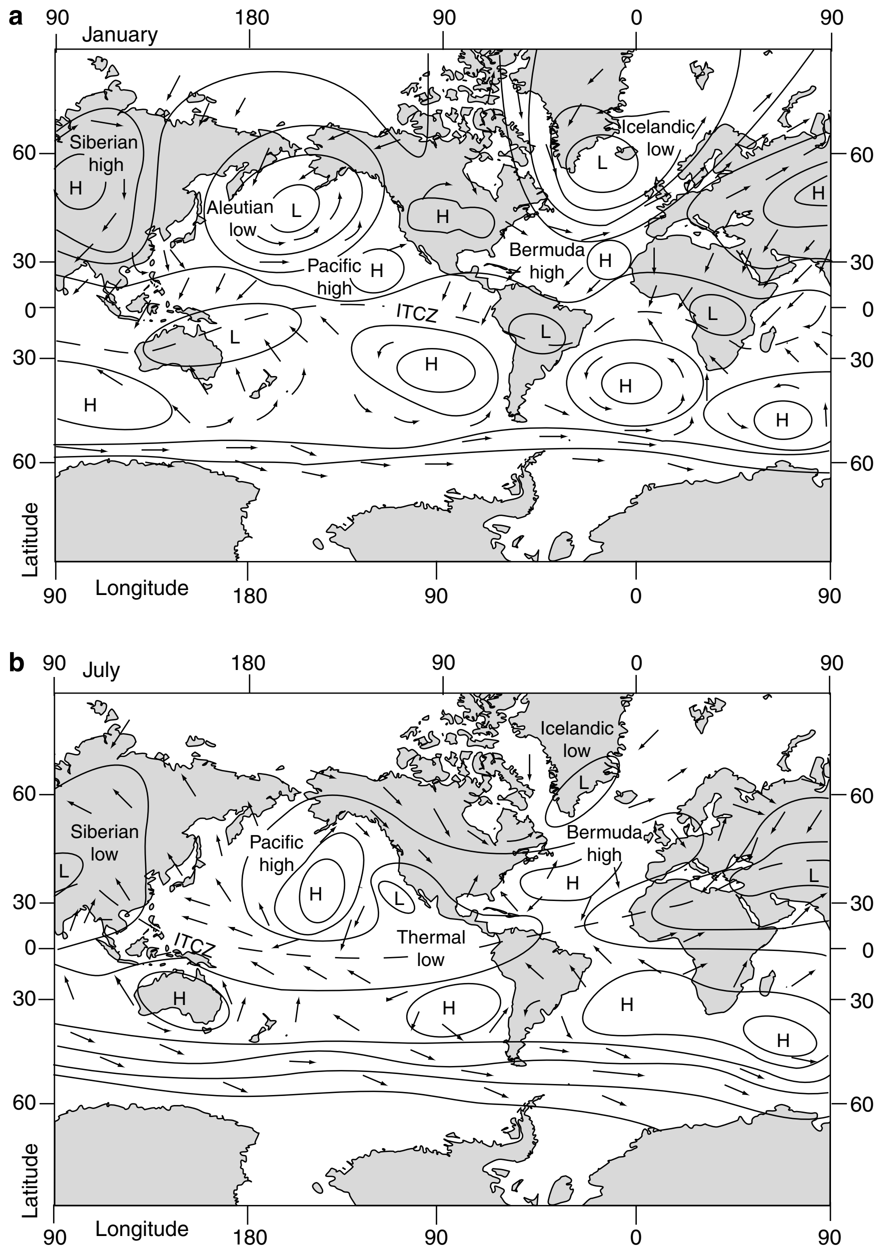


**Fig. 2.9** Average surface wind-flow patterns and the distribution of low (L)- and high (H)-pressure centers for January (*top*) and July (*bottom*). Redrawn from Ahrens (1998)

At 60°N, rising air generates semi-permanent low-pressure zones over Iceland and the Aleutian Islands (the Icelandic and Aleutian lows, respectively). These lows are actually time averages of mid-latitude storm tracks, rather than stable features of the circulation. In the southern hemisphere, there is little land at 60°S, leading to a broad trough of low pressure, rather than distinct centers. Air that subsides in high pressure centers spirals outward in a clockwise direction in the northern hemisphere and in a counter-clockwise direction in the southern hemisphere due to an interaction between friction, Coriolis forces, and the pressure gradient force produced by the subsiding air. Winds spiral inward toward low-pressure centers in a counter-clockwise direction in the northern hemisphere and in a clockwise direction in the southern hemisphere. Air in the low-pressure centers rises in balance with the subsiding air in high-pressure centers. The long-term average of these vertical and horizontal motions produces the vertical circulation described by the Ferrell cell (Fig. 2.8) and a horizontal pattern of high- and low-pressure centers commonly observed on weather charts (Fig. 2.9).

These deviations from the expected easterly or westerly direction of prevailing winds are organized on a planetary scale and are known as **planetary waves**. These waves are most pronounced in the northern hemisphere, where there is more land. They are influenced by the Coriolis effect, land–ocean heating contrasts, and the locations of large mountain ranges, such as the Rocky Mountains and Himalayas. These mountain barriers force the northern hemisphere westerlies vertically upward and to the north. Downwind of the mountains, air descends and moves to the south forming a trough, much like the standing waves in the rapids of a fast-moving river that are governed by the location of rocks in the riverbed. Temperatures are comparatively low in the troughs, due to the southward movement of polar air, and comparatively high in the ridges. The trough over eastern North America downwind of the Rocky Mountains (Fig. 2.9), for example, results in relatively cool temperatures and a more southerly location of the arctic tree line in eastern than in western North America. Although planetary waves have preferred locations, they are not static. Changes in their location or in the number of waves alter regional patterns of climate. These step changes in circulation pattern are referred to as shifts in **climate modes**.

Planetary waves and the distribution of major high- and low-pressure centers explain many details of horizontal motion in the atmosphere and therefore the patterns of ecosystem distribution. The locations of major high- and low-pressure centers, for example, explain the movement of mild moist air to the west coasts of continents at 50–60°N and S, where the temperate rainforests of the world occur (the northwestern U.S. and southwestern Chile, for example; Fig. 2.9). The subtropical high pressure centers at 30°N and S cause cool polar air to move toward the equator on the west coasts of continents, creating dry Mediterranean climates near 30°N and S. On the east coasts of continents, subtropical highs cause warm moist equatorial air to move northward at 30°N and S, creating a moist subtropical climate.

## The Ocean

### Ocean Structure

**Like the atmosphere, the ocean maintains rather stable layers with limited vertical mixing between them.** The sun heats the ocean from the top, whereas the atmosphere is heated from the bottom. Because warm water is less dense than cold water, the ocean maintains rather stable layers that do not easily mix. The uppermost warm layer of **surface water**, which interacts directly with the atmosphere, extends to depths of 75–200 m, depending on the depth of wind-driven mixing. Most primary production and decomposition occur in the surface waters (see Chaps. 5–7). Another major difference between atmospheric and oceanic circulation is that density of ocean waters is determined by both temperature and salinity, so, unlike warm air, warm water can sink, if it is salty enough.

Relatively sharp gradients in temperature (**thermocline**) and salinity (**halocline**) occur between warm surface waters of the ocean and cooler more saline waters at intermediate depths (200–1,000 m; Fig. 2.10). These two vertical gradients create a gradient in water density (**pycnocline**) that generates a relatively stable vertical stratification of low-density surface water above denser deep water. The deep layer therefore mixes with the surface waters very slowly over hundreds to thousands of years. These deeper layers nonetheless play critical roles in element cycling, productivity, and climate because they are long-term sinks for carbon and the sources of nutrients

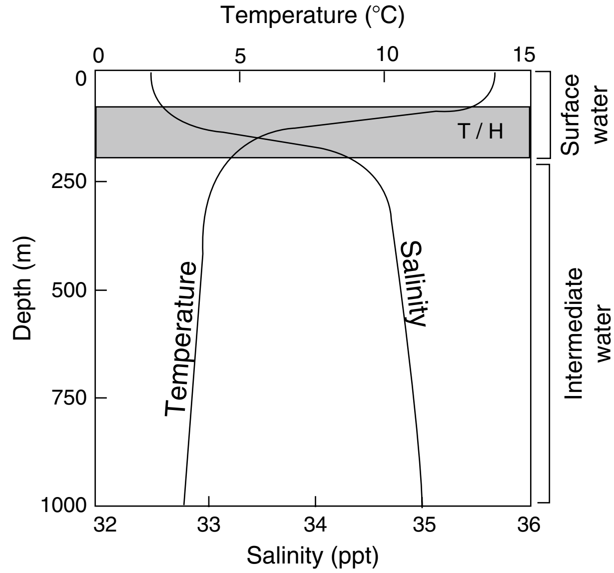


**Fig. 2.10** Typical vertical profiles of ocean temperature and salinity. The thermocline and halocline (T/H) are the zones where temperature and salinity, respectively, decline most strongly with depth. These transition zones usually coincide approximately

that drive ocean production (see Chaps. 5–9). **Upwelling** areas, where nutrient-rich deep waters move rapidly to the surface, support high levels of primary and secondary productivity (marine invertebrates and vertebrates) and are the locations of many of the world's major fisheries.

## Ocean Circulation

**Ocean circulation plays a critical role in Earth's climate system.** The ocean and atmosphere are about equally important in latitudinal heat transport in the tropics, but the atmosphere accounts for most latitudinal heat transport at mid- and high latitudes (Fig. 2.7). The surface currents of the ocean are driven by surface winds and therefore show global patterns (Fig. 2.11) that are generally similar to those of the prevailing surface winds (Fig. 2.9). The ocean currents are, however, deflected 20–40° relative to the wind direction by the Coriolis effect. This deflection and the edges of continents cause ocean currents to be more circular (termed **gyres**) than the winds that drive them. In equatorial regions, currents flow east to west, driven by the easterly trade winds, until they reach the continents, where they split and flow poleward along the western boundaries of the ocean, carrying warm tropical water to higher latitudes. On their way poleward, currents are deflected by the Coriolis effect. Once the water reaches the high latitudes, some returns in surface currents toward the tropics along the eastern edges of ocean basins (Fig. 2.11), and some continues poleward.

Deep-ocean waters show a circulation pattern quite different from the wind-driven surface circulation. In the polar regions, especially in the winter off southern Greenland and off Antarctica, cold air cools the surface waters, increasing their density. Formation of sea ice, which excludes salt from ice crystals (**brine rejection**), increases the salinity of surface waters, also increasing their density. The high density of these cold saline waters causes them to sink. This **downwelling** to form the North Atlantic Deep Water off of Greenland, and the Antarctic Bottom Water off of Antarctica drives the global **thermohaline circulation** in the mid and deep ocean that ultimately transfers water among the major ocean basins (Fig. 2.12). The descent of cold dense water at high latitudes is balanced by the upwelling of deep water on the eastern margins of ocean basins at lower latitudes, where along-shore surface currents are deflected offshore by the Coriolis effect and easterly trade winds. There is a net transfer of North Atlantic Deep Water to other ocean basins, particularly the eastern Pacific

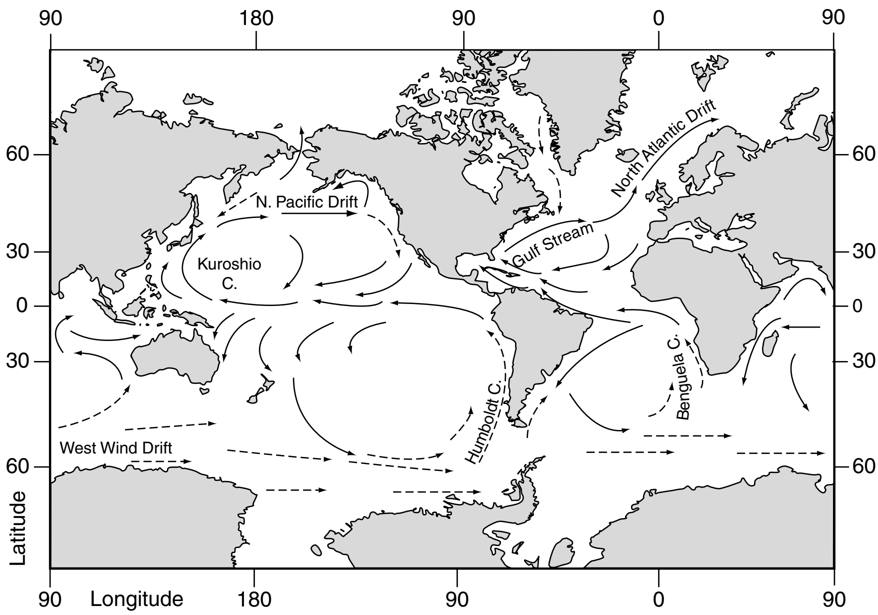


**Fig. 2.11** Major surface ocean currents. Warm currents (C) are shown by *solid arrows* and cold currents by *dashed arrows*. Redrawn from Ahrens (1998)

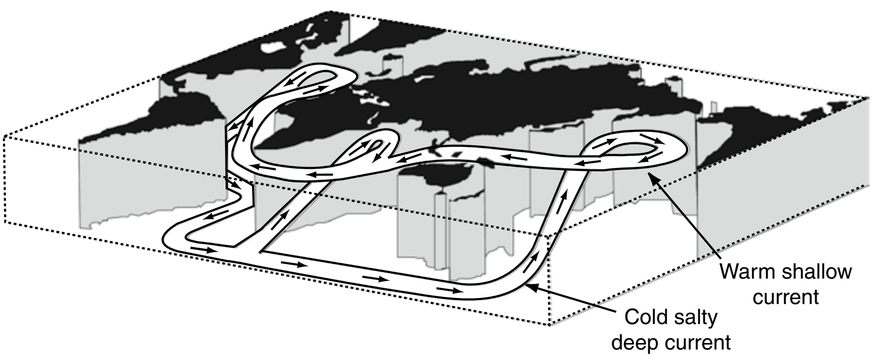


**Fig. 2.12** Circulation patterns of deep and surface waters among the major ocean basins

and Indian Oceans, where old phosphorus-rich waters emerge at the surface. Net poleward movement of warm surface waters balances the movement of cold deep water toward the equator. Changes in the strength of the thermohaline circulation can have significant effects on climate because of its control over latitudinal heat transport. In addition, the thermohaline circulation transfers carbon to depth, where it remains for centuries (see Chap. 14).

The ocean, with its high heat capacity, heats up and cools down much more slowly than does the land and therefore has a moderating influence on the climate of adjacent land. Wintertime temperatures in Great Britain and Western Europe, for example, are much milder than at similar latitudes on the east coast of North America due to the warm **North Atlantic drift** (the poleward extension of the Gulf Stream; Fig. 2.11). Conversely, cold upwelling currents or currents moving toward the equator from the poles cool adjacent landmasses in summer. The cold California current, for example, which runs north to south along the west coast of the U.S., keeps summer temperatures in Northern California lower than at similar latitudes along the east coast of the U.S. These temperature differences play critical roles in determining the distribution of different kinds of ecosystems across the globe.

## Landform Effects on Climate

**The spatial distribution of land, water, and mountains modify the general latitudinal trends in climate.** The greater heat capacity of the ocean has short-term regional as well as long-term global consequences. The ocean warms more slowly than land during the day and in summer and cools more slowly than land at night and in winter, influencing atmospheric circulation at local to continental scales. The seasonal reversal of winds (**monsoons**) in eastern Asia, for example, is driven largely by the temperature difference between the land and the adjacent seas. During the northern-hemisphere winter, the land is colder than the ocean, giving rise to cold dense continental air that flows southward from Siberian high-pressure centers across India to the ocean (Figs. 2.9 and 2.13). In summer, however, the land heats relative to the ocean, forcing the air to rise, in turn drawing in moist surface air from the ocean. Condensation of water vapor in the rising moist air produces large amounts of precipitation. Northward migration of the trade winds in summer enhances onshore flow of air, and the mountainous topography of northern India enhances vertical motion, increasing the proportion of water vapor that is converted to precipitation. Together, these seasonal changes in winds give rise to predictable seasonal patterns of temperature and precipitation that strongly influence the structure and functioning of ecosystems.

At scales of a few kilometers, the differential heating between land and ocean produces **land and sea breezes**. During the day, strong heating over land causes air to rise, drawing in cool air from the ocean (Fig. 2.13). The rising of air over the land increases the height at which a given pressure occurs, causing this upper air to move from land toward the ocean, if the large-scale prevailing winds are weak. The resulting increase in the mass of atmosphere over the ocean raises the surface pressure, which causes surface air to flow from the ocean toward the land. The resulting circulation cell is similar in principle to that which occurs in the Hadley cell (Fig. 2.8) or Asian monsoons (Fig. 2.13). At night, when the ocean is warmer than the land, air rises over the ocean, and the surface breeze blows from the land to the ocean, reversing the circulation cell. The net effect of sea breezes is to reduce temperature extremes and increase precipitation on land near the ocean or large lakes.

Mountain ranges affect local atmospheric circulation and climate through several types of **orographic effects**, which are effects due to the presence of mountains. As winds carry air up the windward sides of mountains, the air cools, and water vapor condenses and precipitates. Therefore, the windward side tends to be cold and wet. When the air moves down the leeward side of the mountain, it expands and warms, increasing its capacity to absorb and retain water. This creates a **rain shadow**, i.e., a zone of low precipitation downwind of the mountains. The rain shadow of the Rocky Mountains extends 1,500 km to the east, resulting in a strong west-to-east gradient in annual precipitation from eastern Colorado (300 mm) to Illinois (1,000 mm; see Fig. 13.3; Burke et al. 1989). Deserts or desert grasslands (steppes) are often found immediately downwind of the major mountain ranges of the world. Mountain systems can also influence climate by channeling winds through valleys. The Santa Ana winds of Southern California occur when high pressure over the interior deserts funnels warm dry winds through valleys toward the Pacific coast, creating dry windy conditions that promote intense wildfires.

Sloping terrain creates unique patterns of microclimate at scales ranging from anthills to mountain ranges. Slopes facing toward the equator (south-facing slopes in the northern hemisphere and north-facing slopes in the southern hemisphere) receive more radiation than opposing

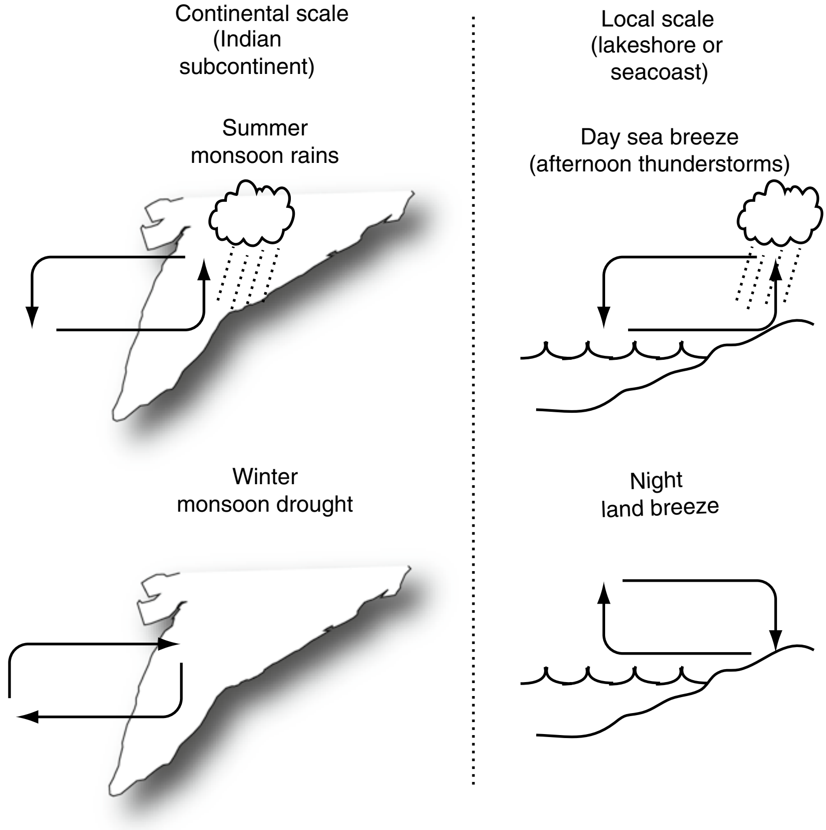


**Fig. 2.13** Effects of land–sea heating contrasts on winds and precipitation at continental and local scales. At the continental scale, the greater heating of land than sea during summer causes air to rise, drawing in cool moist ocean air over India that fuels precipitation. In winter, the ocean is warmer than the land, reversing these wind patterns. At the local scale, similar heating contrasts in coastal or lakeshore areas cause sea breezes and afternoon thunderstorms during the day and land breezes at night

slopes, creating warmer drier conditions. In cold or moist climates, the warmer microenvironment on equator-facing slopes provides conditions that enhance productivity, decomposition, and other ecosystem processes, whereas in dry climates, the greater drought on these slopes limits production. Microclimatic variation associated with slope and **aspect** (the compass direction that a slope faces) allows representatives of an ecosystem type to exist hundreds of kilometers beyond its major zone of distribution. These outlier populations are important sources of colonizing individuals during times of rapid climatic change and are therefore important in understanding species migration and the long-term dynamics of ecosystem changes (see Chap. 12).

Topography also influences climate through drainage of cold dense air. When air cools at night, it becomes denser and tends to flow downhill (**katabatic winds**) into valleys, where it accumulates. This can produce temperature **inversions** (cool air beneath warm air, a vertical temperature profile reversed from the typical pattern in the troposphere of decreasing temperature with increasing elevation; Fig. 2.4). Inversions occur primarily at night and in winter, when heating from the sun is insufficient to promote convective mixing. Clouds also tend to inhibit the formation of winter and nighttime inversions because they increase longwave emission to the surface. Increases in solar heating or windy conditions, such as might accompany the passage of frontal systems, break up inversions. Inversions are climatically important because they increase the seasonal and diurnal temperature extremes experienced by ecosystems in low-lying areas. In cool climates, inversions greatly reduce the length of the frost-free growing season.

## Vegetation Influences on Climate

**Vegetation influences climate through its effects on the surface energy budget.** Climate is quite sensitive to regional variations in vegetation and water content at Earth's surface. The albedo (the fraction of the incident shortwave radiation reflected from a surface) determines the quantity of solar energy absorbed by the surface, which is subsequently available for transfer to the atmosphere as longwave radiation and turbulent fluxes of sensible and latent heat. Water generally has a low albedo, so lakes and the ocean absorb considerable solar energy. At the opposite extreme, snow and ice have a high albedo and hence absorb little solar radiation, contributing to the cold conditions required for their persistence. Vegetation is intermediate in albedo, with values generally decreasing from grasslands, with their highly reflective standing dead leaves, to deciduous forests to dark conifer forests (see Chap. 4). Recent land-use changes have substantially altered regional albedo by increasing the area of exposed bare soil. The albedo of soil depends on soil type and wetness but is often higher than that of vegetation in dry climates. Consequently, overgrazing often increases albedo, reducing energy absorption and the transfer of energy to the atmosphere. This leads to cooling and subsidence, so moist ocean air is not drawn inland by sea breezes. This can reduce precipitation and the capacity of vegetation to recover from overgrazing (Foley et al. 2003a). The large magnitude of many land-surface feedbacks to climate suggests that land-surface change can be an important contributor to regional climatic change (Foley et al. 2003b).

Ecosystem structure influences the efficiency with which turbulent fluxes of sensible and latent heat are transferred to the atmosphere. Wind passing over tall uneven canopies creates mechanical turbulence that increases the efficiency of heat transfer from the surface to the atmosphere (see Chap. 4). Smooth surfaces, in contrast, tend to heat up because they transfer their heat only by convection and not by mechanical turbulence.

The effects of vegetation structure on the efficiency of water and energy exchange influence regional climate. About 25–40% of the precipitation in the Amazon basin comes from water that is recycled from land by evapotranspiration. (Costa and Foley 1999). Simulations by climate models suggest that, if the Amazon basin were completely converted from forest to pasture, this would lead to a permanently warmer drier climate over the Amazon basin (Foley et al. 2003b). The shallower roots of grasses would absorb less water than trees, leading to lower transpiration rates (Fig. 2.14). Pastures would therefore release more of the absorbed solar radiation as sensible heat, which directly warms the atmosphere. There are many uncertainties, however. Changes in cloudiness, for example, can have either a positive or a negative effect on radiative forcing, depending on cloud properties and height.

Changes in albedo caused by vegetation change can create amplifying feedbacks. At high latitudes, for example, tree-covered landscapes absorb more solar radiation prior to snowmelt than does snow-covered tundra. Model simulations suggest that the northward movement of the tree line 6,000 years ago could have reduced the regional albedo and increased energy absorption enough to explain half of the climate warming that occurred at that time (Foley et al. 1994). The warmer regional climate would, in turn, favor tree reproduction and establishment at the tree line (Payette and Filion 1985), providing an amplifying (positive) feedback to regional warming (see Chap. 12). Predictions about the impact of future climate on vegetation should therefore also consider ecosystem feedbacks to climate (Field et al. 2007; Chapin et al. 2008).

Albedo, energy partitioning between latent and sensible heat fluxes, and surface structure also influence the amount of longwave radiation emitted to the atmosphere (Fig. 2.3). Longwave radiation depends on surface temperature, which tends to be high when the surface absorbs large

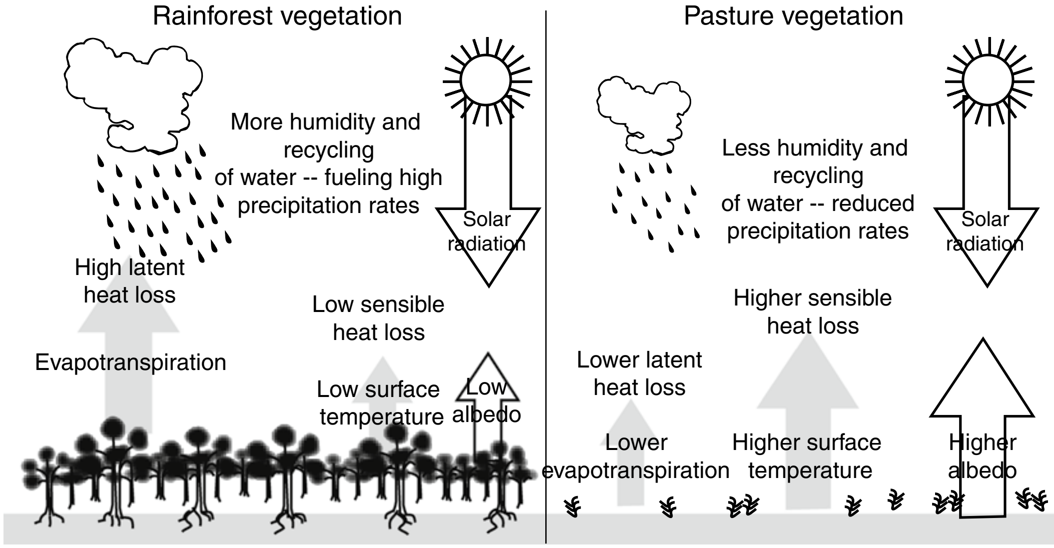


**Fig. 2.14** Climatic consequences of tropical deforestation and conversion to pasture. In forested conditions, the low albedo provides ample energy absorption to drive high transpiration rates that cool the surface and supply abundant moisture to the atmosphere to fuel high precipitation rates. In pasture conditions that develop after deforestation, low vegetation cover and shallow roots restrict transpiration and therefore the moisture available to support precipitation. This, together with high sensible heat flux leads to a warm, dry climate. Based on Foley et al. (2003b)

amounts of incoming radiation (low albedo), has little water to evaporate, or has a smooth surface that is inefficient in transferring turbulent fluxes of sensible and latent heat to the atmosphere (see Chap. 4). Deserts, for example, experience large net longwave energy losses because their dry smooth surfaces lead to high surface temperatures, and little moisture is available to support evaporation that would otherwise cool the soil.

## Temporal Variability in Climate

### Long-Term Changes

**Millennial-scale climatic change is driven primarily by changes in the distribution of solar input and changes in atmospheric composition.** Earth's climate is a dynamic system that has changed repeatedly, producing frequent, and sometimes abrupt, changes in climate, including dramatic glacial periods (Fig. 2.15) and sea-level changes. Volcanic eruptions and asteroid impacts alter climate on short time scales through changes in absorption or reflection of solar energy. Continental drift and mountain building and erosion have modified the patterns of atmospheric and ocean circulation on longer time scales. The primary force responsible for the evolution of Earth's climate, however, has been changes in the input of solar radiation, which has increased by about 30% over the past four billion years, as the sun matured (Schlesinger 1997). On millennial time scales, the distribution of solar input has varied primarily due to predictable variations in Earth's orbit.

Three types of variations in Earth's orbit influence the amount of solar radiation received at the surface at different times of the year and at different latitudes: **eccentricity** (the degree of ellipticity of Earth's orbit around the sun), **tilt** (the angle between Earth's axis of rotation relative to the plane of its orbit around the sun), and **precession** (a "wobbling" in Earth's axis of rotation with respect to the stars, determining the time of year when different locations on Earth are closest to the sun). The periodicities of these orbital parameters (eccentricity, tilt, and precession) are

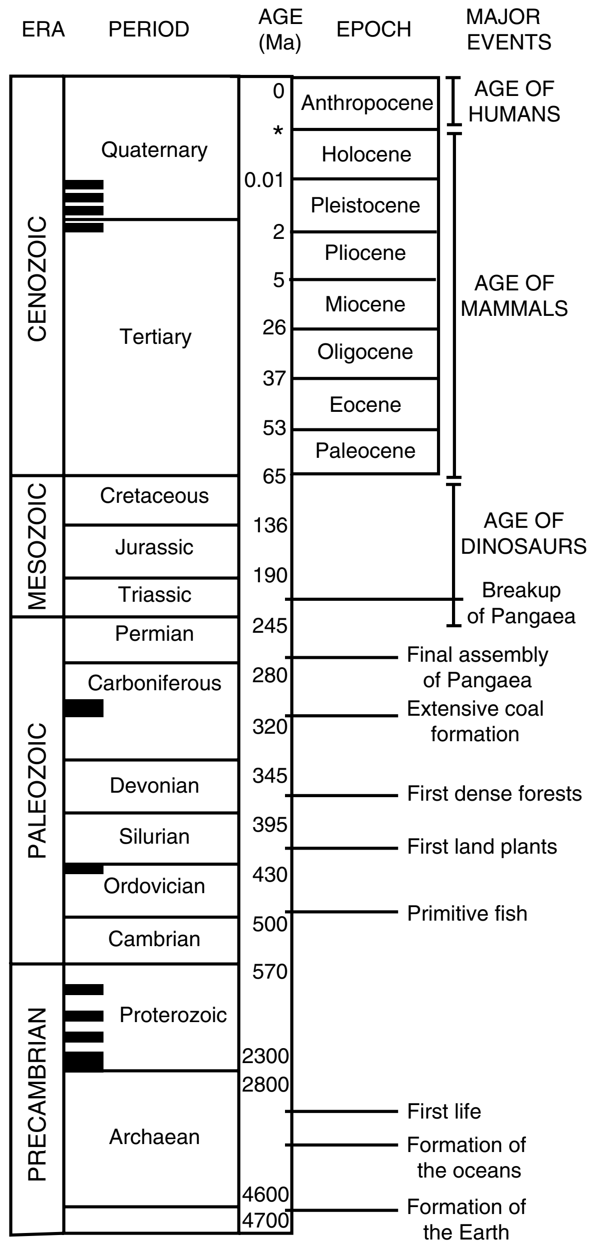


**Fig. 2.15** Geological time periods in Earth's history, showing major glacial events (*dark bars*) and ecological events that strongly influenced ecosystem processes. Note the changes in time scale in units of millions of years (Ma). The most recent geologic epoch (the Anthropocene) began about 1750 with the beginning of the industrial revolution and is characterized by human domination of the biosphere (Crutzen 2002). Modified from Sturman and Tapper (1996)

approximately 100,000, 41,000, and 23,000 years, respectively. Interactions among these cycles produce **Milankovitch cycles** of solar input that correlate with the glacial and interglacial cycles. Analysis of these cycles indicates that Earth would not naturally enter another ice age for at least 30,000 years, so natural cycles in solar input will not substantially offset human-driven warming of climate (IPCC 2007). Ice ages are triggered by minima in northern high-latitude summer radiation that enable winter snowfall to persist through the year and build northern-hemisphere ice sheets that reflect incoming radiation. These changes become globally amplified by feedbacks in Earth's climate system (such as changes in atmospheric $CO_2$ concentration) to cause large climatic changes throughout the planet.

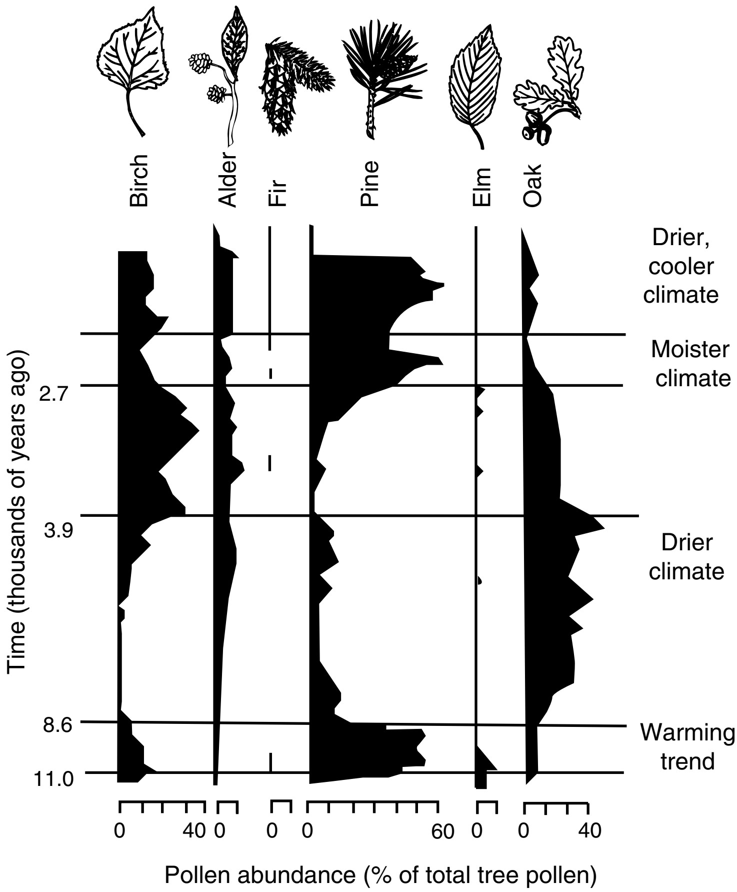


**Fig. 2.16** Pollen profile from a bog in northwestern Minnesota showing changes in the dominant tree species over the past 11,000 years. Redrawn from McAndrews (1966)

The chemistry of ice and trapped air bubbles provide a paleorecord of the climate when the ice formed. Ice cores drilled in Antarctica and Greenland indicate considerable climate variability over the past 650,000 years, in large part related to the Milankovitch cycles (see Fig. 14.6). Analysis of bubbles in these cores indicates that past warming events have been associated with increases in $CO_2$ and $CH_4$ concentrations, providing circumstantial evidence for a past role of radiatively active gases in climate change. The unique feature of the recent anthropogenic increases in these gases is that they are occurring during an *interglacial* period, when Earth's climate is already relatively warm. These cores indicate that the $CO_2$ concentration of the atmosphere is higher now than at any time in at least the last 650,000 years (IPCC 2007). Fine-scale analysis of ice cores from Greenland suggests that large changes from glacial to interglacial climate can occur in decades or less. Such rapid transitions in the climate system to a new state may be related to sudden changes in the strength of the thermohaline circulation that drives oceanic heat transport from the equator to the poles.

Past climates can also be reconstructed from other paleorecords. Tree-ring records, obtained from living and dead trees, provide information about climate during the last several thousand years. Variation in the width of tree rings records temperature and moisture, and chemical composition of wood reflects the characteristics of the atmosphere at the time the wood was formed. Pollen preserved in low-oxygen sediments of lakes provides a history of plant taxa and climate over the past tens of thousand years (Fig. 2.16).

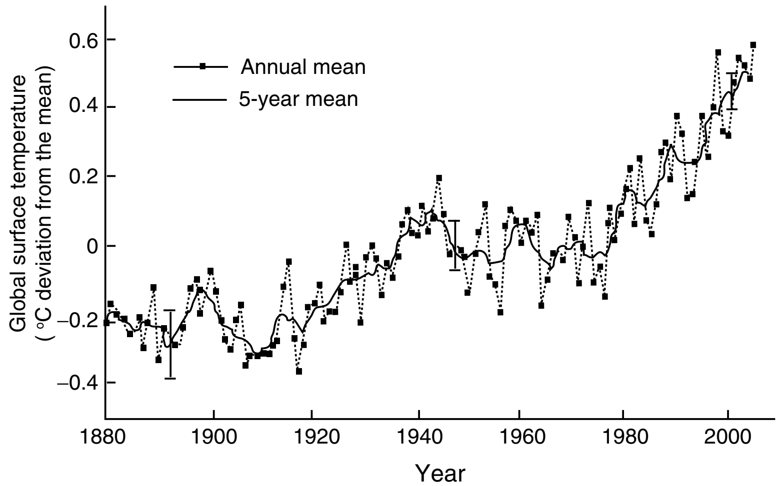


**Fig. 2.17** Time course of the average surface temperature of Earth from 1850 to 2005 (relative to the average temperature for this time period). Redrawn from IPCC (2007)

Pollen records from networks of sites can be used to construct maps of species distributions at various times in the past and provide a history of species migrations across continents after climatic changes (COHMAP 1988). Other proxy records provide measures of temperature (species composition of Chironomids), precipitation (lake level), pH, and geochemistry.

The combination of paleoclimate proxies indicates that climate is inherently variable over all time scales. Atmospheric, oceanic, and other environmental changes that are occurring now due to human activities must be viewed as overlays on the natural climate variability that stems from long-term changes in Earth's surface characteristics and orbital geometry.

## Anthropogenic Climate Change

**Earth's climate during the last half of the twentieth century was warmer than during any 50-year interval in the last 500 years and probably the last 1,300 years or longer** (Fig. 2.17; IPCC 2007; Serreze 2010). This warming is most pronounced near Earth's surface, where its ecological effects are greatest. A small amount of the recent warming reflects an increase in solar input, but most of the warming results from human activities that increase the concentrations of radiatively active gases in the atmosphere (Fig. 2.18). These gases trap more of the longwave radiation emitted by Earth's surface and warm the atmosphere, which retains more water vapor (another potent greenhouse gas) and further increases the trapping of longwave radiation. As a result, Earth is no longer in radiative equilibrium but is losing less energy to space than it is absorbing from the sun. Consequently, Earth's surface warmed about 0.7°C from 1880 to 2008 (Fig. 2.17) and is projected to warm an additional three to four times that amount by the end of the twenty-first century (Serreze 2010).

Climate models and recent observations indicate that warming will be most pronounced in the interiors of continents, far from the moderating effects of the ocean, and at high latitudes. The high-latitude warming reflects an amplifying feedback. As climate warms, the snow and sea ice melt earlier in the year, which replaces the reflective snow or ice cover with a low-albedo land or water surface. These darker surfaces absorb more radiation and transfer this energy to the atmosphere, which amplifies the rate of climate warming. Clouds, increases in water vapor, and increases in poleward energy transport also contribute to polar warming. Those changes in the climate system that occur over years to

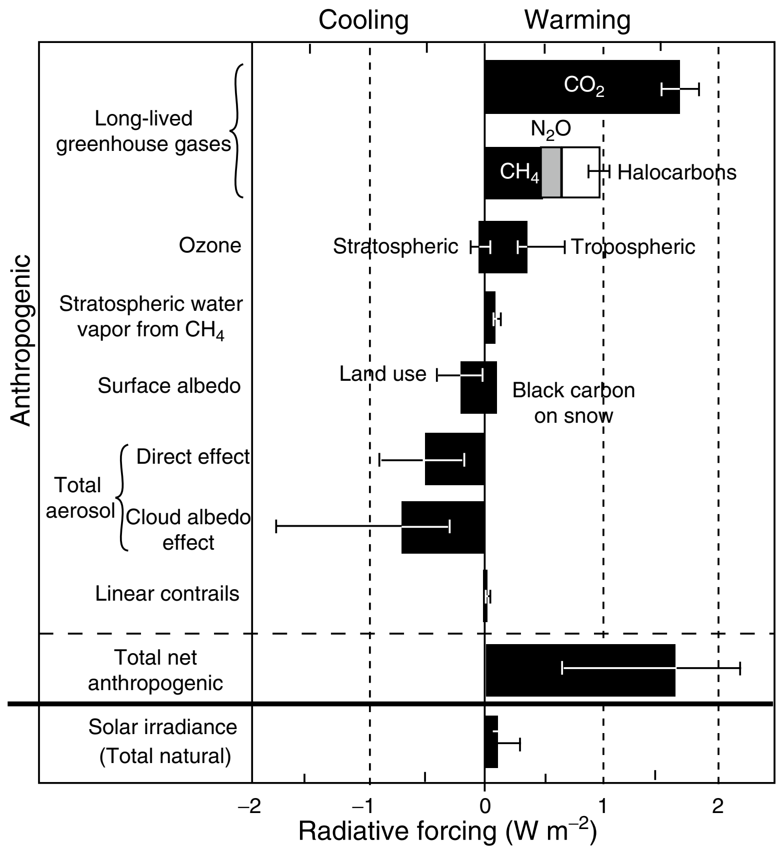


**Fig. 2.18** Global average radiative forcing of the climate system (i.e., external forces that modify the climate system) estimated for 2005. Some changes in the climate system lead to net warming; others lead to net cooling. The largest single cause of climate warming is the increased concentration of atmospheric $CO_2$, primarily as a result of burning fossil fuels. Redrawn from IPCC (2007)

decades are dominated by amplifying feedbacks, such as the ice–albedo feedback, causing anthropogenic warming to accelerate (Serreze 2010).

As climate warms, the air has a higher capacity to hold water vapor, so there is greater evaporation from the ocean and other moist surfaces. In areas where rising air leads to condensation, this leads to greater precipitation. Continental interiors are less likely to experience large precipitation increases but will be dried by increasing evaporation. Consequently, soil moisture and runoff to streams and rivers are likely to increase in coastal regions and mountains and to decrease in continental interiors. In other words, wet regions will likely become wetter and dry regions drier. Winter warming is likely to reduce the snowpack in mountains and therefore the spring runoff that fills reservoirs on which many cities depend for water supply. The complex controls and nonlinear feedbacks in the climate system make detailed climate projections problematic and are active areas of research (IPCC 2007).

## Interannual Climate Variability

**Much of the interannual variation in regional climate is associated with large-scale changes in the atmosphere–ocean system.** Superimposed on long-term climate variability are interannual variations that have been noted by farmers, fishermen, and naturalists for centuries. Some of this variability exhibits repeating geographic and temporal patterns. For example, **El Niño/ Southern Oscillation** or ENSO (Webster and Palmer 1997; Federov and Philander 2000) events are part of a large-scale, air–sea interaction that couples atmospheric pressure changes (the Southern Oscillation) with changes in ocean temperature (El Niño) over the equatorial Pacific Ocean. ENSO events have occurred, on average, every 3–7 years over the past century, with considerable irregularity (Trenberth and Haar 1996). No events occurred between 1943 and 1951, for example, and three major events occurred between 1988 and 1999.

In most years, the easterly trade winds push the warm surface waters of the Pacific westward, so the layer of warm surface waters is deeper in the western Pacific than in the east (Figs. 2.8 and 2.19). The resulting warm waters in the western Pacific are associated with a low-pressure center and promote convection and high rainfall in Indonesia. The offshore movement of surface waters in the eastern Pacific promotes upwelling of colder, deeper water off the coasts of Ecuador and Peru. These cold, nutrient-rich waters support a productive fishery (see Chap. 9) and promote subsidence of upper air, leading to the development of a high-pressure center and low precipitation. At times, however, the eastern-Pacific high-pressure center, Indonesian low-pressure center, and the easterly trades all weaken. The warm surface waters then move eastward, forming a deep layer of warm surface water in the eastern Pacific. This reduces or shuts down the upwelling of cold water, promoting atmospheric convection and rainfall in coastal Ecuador and Peru. The colder waters in the western Pacific, in contrast, inhibit convection, leading to droughts in Indonesia, Australia, and India. This pattern is commonly termed **El Niño**. Periods in which the "normal" pattern is particularly strong, with relatively cool surface waters in the eastern Pacific, are termed **La Niña**. The trigger for changes in this ocean–atmosphere system are uncertain, but may involve large-scale ocean waves, known as **Kelvin waves**, that travel back and forth across the tropical Pacific.

ENSO events have widespread climatic, ecosystem, and societal consequences. Strong El Niño phases cause dramatic reductions in anchovy fisheries in Peru with corresponding reproductive failure and mortality in sea birds and marine mammals. For the past four centuries, Peruvian potato farmers detected incipient El Niño conditions by looking at the brightness of stars in the summer, which corresponds to the high cirrus clouds that accompany El Niño (Orlove et al. 2000). This enabled them to adjust planting dates for their most critical crop. Similarly, annually variable harvest of shearwater chicks by New Zealand Maori provided early detection of El Niño events (Lyver et al. 1999). Extremes in precipitation linked to ENSO cycles are also evident in areas distant from the tropical Pacific. El Niño events bring hot, dry weather to the Amazon Basin, potentially affecting tree growth, soil carbon storage, and fire probability. Northward extension of warm tropical waters to the Northern Pacific brings rains to coastal California and high winter temperatures to Alaska. An important lesson from ENSO studies is that strong climatic events in one region have climatic consequences throughout the globe due to the dynamic interactions (termed **teleconnections**) associated with atmospheric circulation and ocean currents.

The Pacific North America (PNA) pattern is another large-scale pattern of climate variability. The positive mode of the PNA is characterized by above-average atmospheric pressure with warm, dry weather in western North America and below-average pressure and low temperatures in the east. Another large-scale climate pattern is the Pacific Decadal Oscillation (PDO), a multidecadal pattern of climate variability that appears to modulate ENSO events. More El Niño events tend to occur when the PDO is in its positive phase, as during the last 25 years of the twentieth century. The North Atlantic Oscillation (NAO) is still another large-scale circulation pattern. Positive phases of the NAO are associated with a strengthening of the pressure gradient between the Icelandic low- and the Bermuda high-pressure systems (Fig. 2.9). This increases heat transport

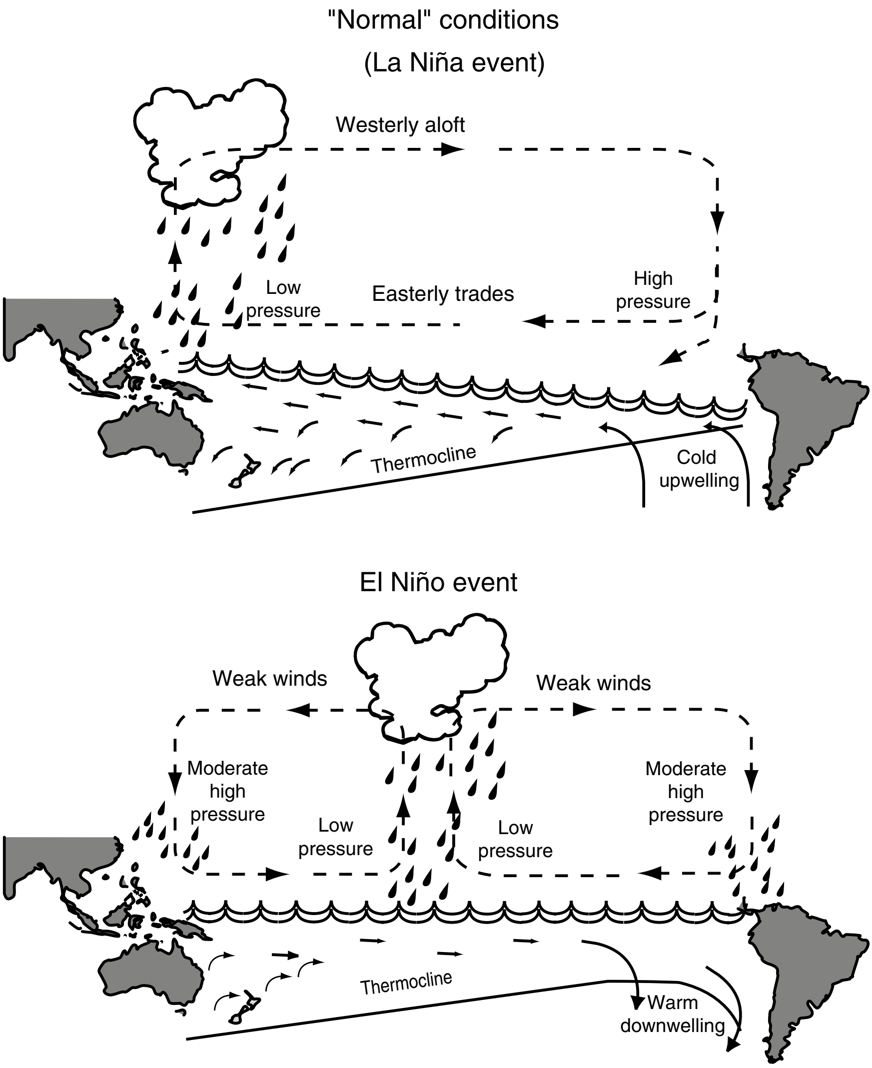


**Fig. 2.19** Circulation of the ocean and atmosphere in the tropical Pacific between South America and Indonesia during "normal (La Niña) years" and during El Niño years. In normal years, strong easterly trade winds push surface ocean waters to the west, producing deep, warm waters and high precipitation off the coast of Southeast Asia and cold, upwelling waters and low precipitation off the coast of South America. In El Niño years, however, weak easterly winds allow the surface waters to move from west to east across the Pacific Ocean, leading to cooler surface waters and less precipitation in Southeast Asia and warmer surface waters and more precipitation off South America. Redrawn from McElroy (2002)

to high latitudes by wind and ocean currents, leading to a warming of Scandinavia and western North America and a cooling of eastern Canada. Although the factors that initiate these large-scale climate features are poorly understood, the patterns themselves and their ecosystem consequences are becoming more predictable. Future climatic changes will likely be associated with changes in the strength and frequencies of certain phases of these large-scale climate patterns rather than simple linear trends in climate. Climate warming, for example, might increase the frequency of El Niño events and positive phases of the PDO and NAO.

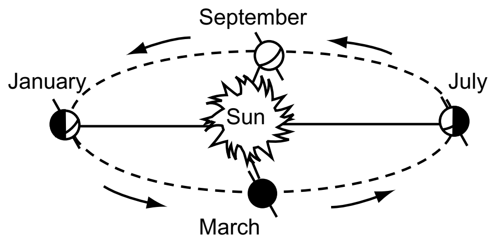


**Fig. 2.20** Earth's orbit around the sun, showing that the zone of greatest heating (the ITCZ) is south of the equator in January, north of the equator in July, and at the equator in March and September

## Seasonal and Daily Variation

**Seasonal and daily variations in solar input have profound but predictable effects on climate and ecosystems.** Perhaps the most obvious variations in the climate system are the patterns of seasonal and diurnal change. Earth rotates on its axis at 23.5° relative to its orbital plane about the sun. This tilt in Earth's axis results in strong seasonal variations in day length and solar **irradiance**, i.e., the quantity of solar energy received at Earth's surface per unit time. During the spring and autumn **equinoxes**, the sun is directly overhead at the equator, and the entire earth surface receives approximately 12 h of daylight (Fig. 2.20). At the northern-hemisphere summer **solstice**, the sun's rays strike Earth most directly in the northern hemisphere, and day length is maximized. At the northern-hemisphere winter solstice, the sun's rays strike Earth most obliquely in the northern hemisphere, and day length is minimized. The summer and winter solstices in the southern hemisphere are 6 months out of phase with those in the north. Variations in incident radiation become increasingly pronounced as latitude increases. Thus, tropical environments experience relatively small seasonal differences in solar irradiance and day length, whereas such differences are maximized in the Arctic and Antarctic. Above the Arctic and Antarctic circles, there are 24 h of daylight at the summer solstice, and the sun never rises at the winter solstice. The relative homogeneity of temperature and light throughout the year in the tropics contributes to their high productivity and diversity. At higher latitudes, the length of the warm season strongly influences the life forms and productivity of ecosystems.

Variations in light and temperature play an important role in determining the types of plants that grow in a given climate and the rates at which biological processes occur. Almost all biological processes are temperature dependent, with slower rates occurring at lower temperatures. Seasonal variations in day length (**photoperiod**) provide important cues that allow organisms to prepare for seasonal variations in climate.

**In aquatic ecosystems, seasonal changes in irradiance influence not only the temperature and light environment but also the fundamental structure of the ecosystem**. Both lakes and the ocean are heated from the top, with most solar radiation absorbed and converted to heat in the upper centimeters to meters of the water column. This surface heating tends to **stratify** lakes and the ocean, with warmer, less dense water at the surface (Fig. 2.21). This tendency for stratification is counter-balanced by turbulent mixing from wind, river inflow, and the cooling of surface waters that occurs at night and during periods of cold weather. Stratification is least pronounced in wind-exposed lakes or lakes with large river inputs (e.g., many reservoirs) where turbulence mixes water to substantial depth. In the open ocean, the turbulent mixed layer is often 100–200 m in depth. In shallow lakes, turbulence often mixes the entire water column.

Lake stratification is most stable in the temperate zone between about 25–40° N and S latitude (Kalff 2002). In colder climates, cold surface waters reduce the temperature (and therefore density) gradients from the surface to depth. In the tropics, deep waters are warm throughout the year, so there is only a weak temperature gradient (often about 1°C) from the surface to depth. Seasonal fluctuations in wind-driven evaporation and cloudiness account for much of the seasonal variation in surface-water temperatures of tropical lakes.

Stratification of nontropical lakes develops during summer, when the heating of surface waters is most intense. Weakly stratified lakes often mix water throughout the water column even

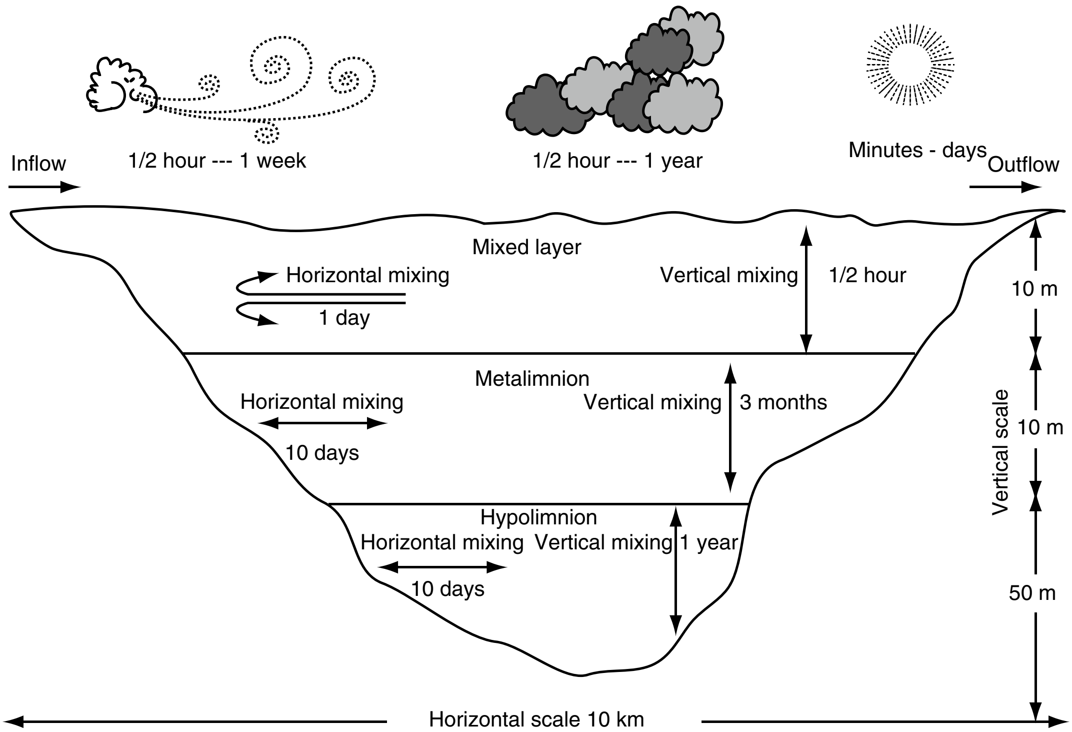


**Fig. 2.21** Estimates of horizontal and vertical mixing times in a medium-sized temperate lake. Redrawn from Kalff (2002)

during the summer. In these lakes, mixing may occur at night, if air temperatures are cooler than the surface waters, or during storms, when wind-driven mixing is more intense. In lakes that are more stably stratified (e.g., temperate lakes that are deep or protected from wind), two relatively discrete layers develop: an **epilimnion** at the surface that is heated by absorbed radiation and mixed by wind and a **hypolimnion** at depth that is colder, more dense, and unaffected by surface turbulence (Fig. 2.21). **Turnover** of these stably stratified temperate lakes occurs in the autumn, when air temperature declines below the temperature of the epilimnion, causing the epilimnion to cool. This surface cooling reduces the density gradient from the surface to depth so that wind-driven turbulence mixes waters more deeply in the lake. Even in wind-protected lakes, nighttime cooling makes surface waters cooler and denser, causing the water column to mix to depth.

Stratification is important because it separates a well-lighted surface layer where photosynthesis exceeds respiration from a deeper, poorly illuminated hypolimnion where respiration exceeds photosynthesis. This spatial separation of these key ecosystem processes results in surface oxygenation and nutrient depletion and nutrient enrichment and oxygen depletion at depth. Seasonal and wind-driven mixing events are critical for resupplying nutrients to the epilimnion and oxygen to the hypolimnion. Lakes often experience a spring algal bloom when increases in light and temperature enable algae to take advantage of the nutrients that are resupplied to the epilimnion during autumn and winter. Eutrophication of lakes by nutrient inputs from fertilizers or sewage reduces water clarity, which concentrates the heating of water near the surface and reduces the depth of the epilimnion. Increased surface production also increases the rain of dead organic matter to depth, which depletes oxygen from the water column, making eutrophic lakes less suitable for fish despite their high algal productivity.

## Storms and Weather

**Storms, droughts, and other unpredictable weather events strongly influence ecosystems**. Because extreme events, by definition, occur infrequently, it is generally impossible to explain unambiguously the climatic cause of a particular event. The intensity of hurricanes and other tropical storms, for example, depends on sea-surface temperature, so it is not surprising that ocean warming is associated with an increase in hurricane intensity (IPCC 2007). Nonetheless, we cannot say that climate warming causes any particular event, such as Hurricane Katrina, which flooded New Orleans in 2005 (Fig. 2.1). Rather, intense hurricanes of that sort will probably occur more often, if climate continues to warm. Increased latitudinal heat transport associated with climate warming has also caused a strengthening and poleward shift in westerly winds, increasing the frequency of intense storms at high latitudes. These tropical and high-latitude storms are important agents of disturbance, so changes in their intensity are likely to alter the structure and long-term dynamics of ecosystems (see Chap. 12).

## Relationship of Climate to Ecosystem Distribution and Structure

**Climate is the major determinant of the global distribution of biomes.** The major types of ecosystems show predictable relationships with climatic variables such as temperature and moisture (Fig. 2.22; Holdridge 1947; Whittaker 1975; Bailey 1998). An understanding of the causes of geographic patterns of climate (Fig. 2.23), as presented in this chapter, therefore allows us to predict the distribution of Earth's major biomes (Fig. 2.24).

**Tropical wet forests** (rainforests) occur from 12°N to 3°S and correspond to the ITCZ. Day length and solar angle show little seasonal change within this zone, leading to consistently high temperatures (Figs. 2.22–2.25). High solar

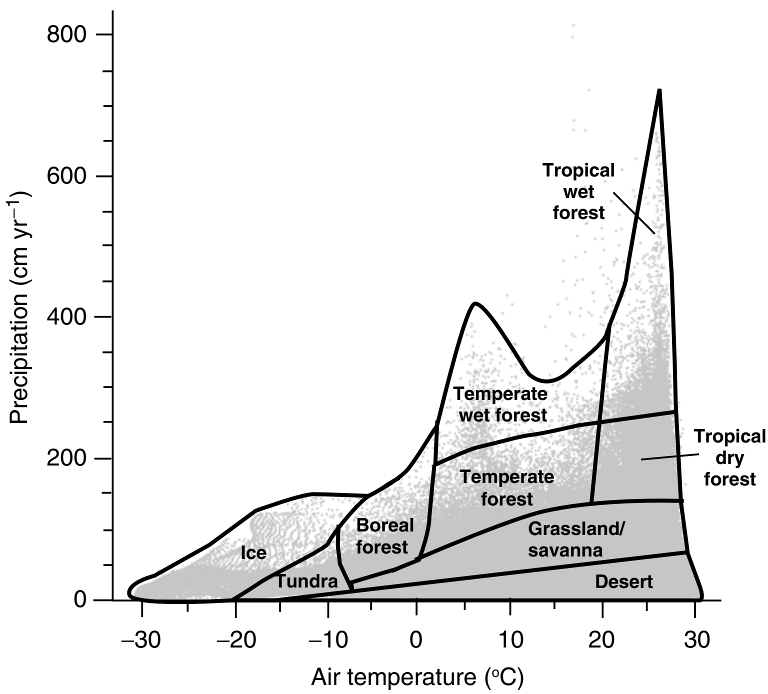


**Fig. 2.22** Distribution of major biomes in relation to average annual air temperature and total annual precipitation. *Gray dots* show the temperature–precipitation regime of all terrestrial locations (excluding Antarctica) at 18.5-km resolution (data from New et al. (2002)). Diagram kindly provided by Joseph Craine and Andrew Elmore

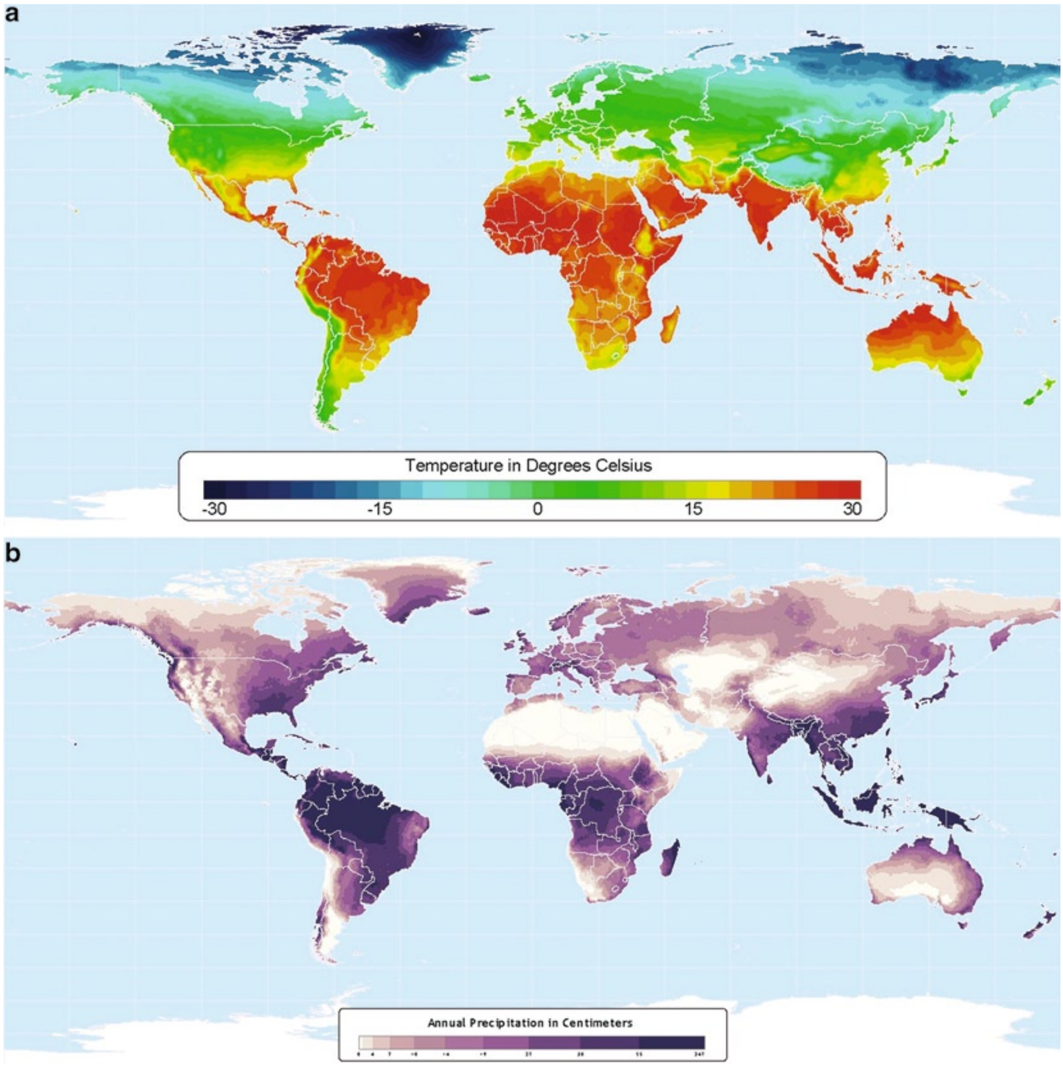


**Fig. 2.23** The global patterns of average annual temperature and total annual precipitation (New et al. 1999). Reproduced from the Atlas of the Biosphere (http://www.sage.wisc.edu/atlas/)

radiation and convergence of the easterly trade winds at the ITCZ promote strong convective uplift leading to high precipitation (175–400 cm annually). Periods of relatively low precipitation seldom last more than 1–2 months. **Tropical dry forests** (Fig. 2.26) occur north and south of tropical wet forests. Tropical dry forests have pronounced wet and dry seasons because of seasonal movement of ITCZ over (wet season) and away from these forests (dry season). **Tropical savannas** (Fig. 2.27) occur between the tropical dry forests and deserts. These savannas are warm and have low precipitation that is highly seasonal. **Subtropical deserts** (Fig. 2.28) at 25–30°N and S have a warm, dry climate because of the subsidence of air in the descending limb of the Hadley cell.

**Mid-latitude deserts, grasslands, and shrublands** (Fig. 2.29) occur in the interiors of continents, particularly in the rain shadow of mountain ranges. They have low unpredictable precipitation, low winter temperatures, and greater temperature extremes than tropical deserts. As precipitation increases, there is a gradual transition from desert

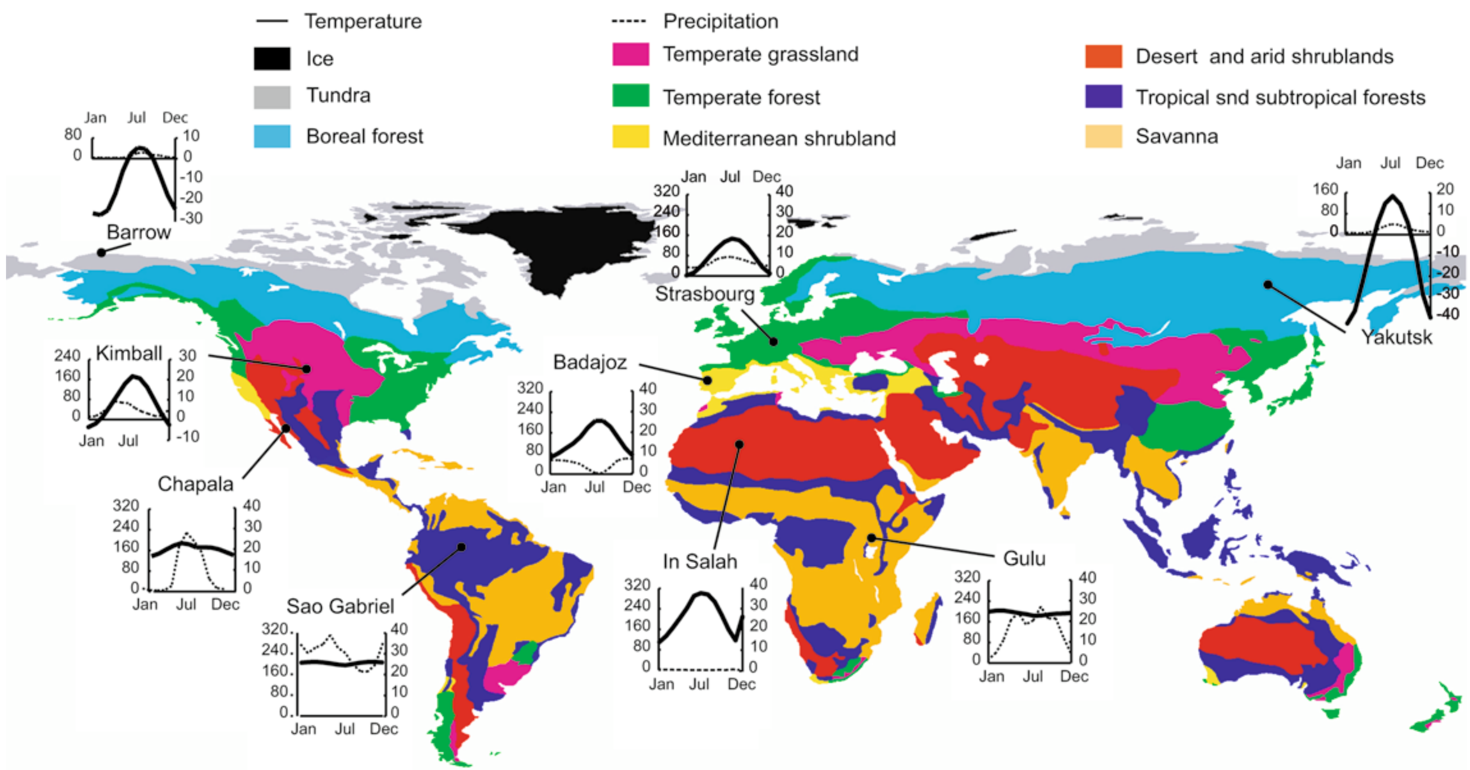


**Fig. 2.24** The global distribution of Earth's major biomes and the seasonal patterns of monthly average temperature and precipitation at one representative site for each biome. Climate data are monthly averages of the entire period of record for selected sites through the year 2000 (http://www.ncdc.noaa.gov/oa/climate/stationlocator.html). Map redrawn from Bailey (1998)

**Fig. 2.25** Tropical wet forest in Brazil. It is characterized by a diversity of life forms and species, including vines, epiphytes, and broadleafed evergreen trees. Photograph by Peter Vitousek

**Fig. 2.26** Subtropical dry forest in Chamela, western Mexico in the wet and dry seasons. The forest is dominated by drought-deciduous trees. Photograph by Peter Vitousek

**Fig. 2.27** Subtropical savanna in Kruger National Park, South Africa, showing a diversity of plants (grasses, shrubs, and trees) and grazing mammals. These fine-leaf savannas burn frequently, permitting both trees and grasses to coexist. Photograph courtesy of Alan K. Knapp

**Fig. 2.28** Sonoran desert landscape in the Superstition Mountains of Arizona, showing a diversity of drought-adapted life forms, with substantial bare ground between plants. Photograph courtesy of Jim Elser

**Fig. 2.29** Mid-latitude Kansas grassland (tallgrass prairie) in early summer with bison grazing. This landscape was burned early in the spring. Here, trees are restricted to the wetter portions of the landscape where they are also protected from fire. Photograph courtesy of Alan K. Knapp

**Fig. 2.30** Mediterranean shrubland in the Santa Monica Mountains of coastal California. It occurs on steep slopes with shallow soils and supporting drought-adapted deciduous and evergreen shrubs. Photograph courtesy of Stephen Davis

to grassland to shrubland. **Mediterranean shrublands** (Fig. 2.30) are situated on the west coasts of continents. In summer, subtropical oceanic high-pressure centers and cold upwelling coastal currents produce a warm dry climate. In winter, as wind and pressure systems move toward the equator, storms produced by polar fronts provide unpredictable precipitation. **Temperate forests** (Fig. 2.31) occur in mid-latitudes, where there is enough precipitation to support trees. The polar front, the boundary between the polar and subtropical air masses, migrates north and

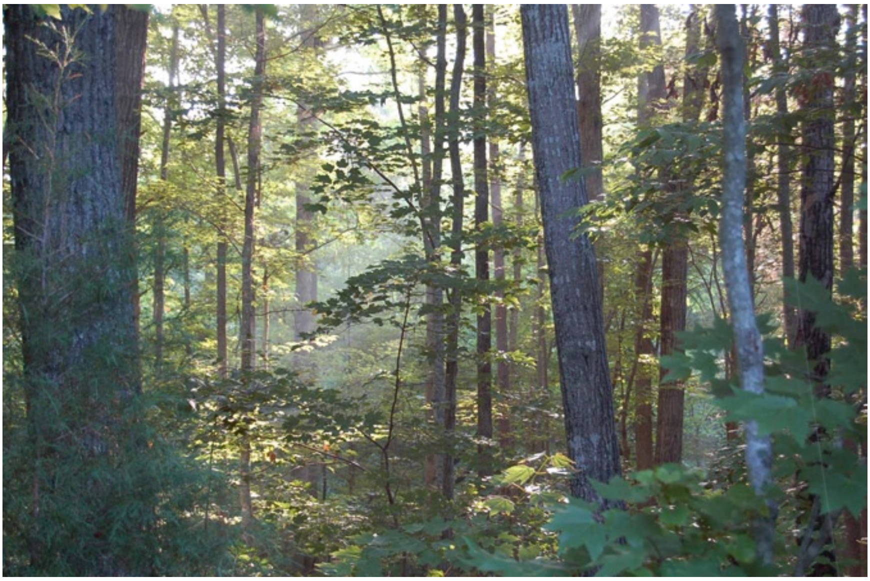

**Fig. 2.31** Temperate forest in the eastern U.S. (North Carolina), showing a complex multi-layered canopy with sunflecks common in all canopy layers. Photograph courtesy of Norm Christensen

**Fig. 2.32** Temperate wet forest in the Valley of the Giants in the Oregon Coast Range of the western U.S. The stand contains a range of tree ages up to five centuries. The understory has coarse woody debris and a flora of shrubs, ferns, herbs, mosses, and tree seedlings. Photograph courtesy of Mark E. Harmon

south of these forests from summer to winter, producing a strongly seasonal climate. **Temperate wet forests** (rainforests; Fig. 2.32) occur on the west coasts of continents at 40–65°N and S, where westerlies blowing across a relatively warm ocean provide an abundant moisture source, and migrating low-pressure centers associated with the polar front promote high precipitation. Winters are mild, and summers are cool.

**The boreal forest (taiga;** Fig. 2.33) occurs in continental interiors at 50–70°N. The winter climate is dominated by polar air masses and the

**Fig. 2.33** Boreal forest on the Tanana River of Interior Alaska. The landscape contains a spectrum of stand ages, ranging from early successional shrub stands on the point bar in the lower left and in the clearcut in the upper left to mature white spruce stands in the center of the photograph to muskegs on terraces in the distance that are thousands of years old. Photograph courtesy of Roger Ruess

summer climate by temperate air masses, producing cold winters and mild summers. The distance from oceanic moisture sources results in low precipitation. The subzero average annual temperature leads to **permafrost** (permanently frozen ground) that restricts drainage and creates poorly drained soils and peatlands in low-lying areas. **Arctic tundra** (Fig. 2.34) is a zone north of the polar front in both summer and winter, resulting in a climate that is too cold to support growth of trees. Short cool summers restrict biological activity and limit the range of life forms that can survive.

**Vegetation structure varies with climate both among and within biomes.** Predictable growth forms of plants dominate each biome type. Broadleaved evergreen trees, for example, dominate tropical wet forests, whereas areas that are periodically too cold or dry for growth of these trees are dominated by deciduous forests or, under more extreme conditions, by tundra or desert, respectively. Biomes are not discrete units with sharp boundaries but vary continuously in structure along climatic gradients. Along a moisture gradient in the tropics, for example, vegetation changes from tall evergreen trees in the wettest sites to a mix of evergreen and deciduous trees in areas with seasonal drought (Fig. 2.35; Ellenberg 1979). As the climate becomes still drier, the stature of the trees and shrubs declines because of less light competition and more competition for water (Fig. 2.36). Ultimately, this leads to a shrubless desert with herbaceous perennial herbs in dry habitats. With extreme drought, the dominant life form becomes annuals and bulbs (herbaceous perennials in which aboveground parts die during the dry season). A similar gradient of growth forms, leaf types, and life forms occurs along moisture gradients at other latitudes.

**The diversity of growth forms within some ecosystems can be nearly as great as the diversity of dominant growth forms across biomes.** In tropical wet forests, for example, continuous seasonal growth in a warm, moist climate produces large trees with dense canopies that intercept, and compete for, a large fraction of

**Fig. 2.34** Arctic tundra near Toolik Lake in northern foothills of the Brooks Range of Alaska. The landforms were shaped by Pleistocene glaciations, and the soils are kept wet and cold by a continuous layer of permafrost 30–50 cm beneath the surface. Photograph by Stuart Chapin

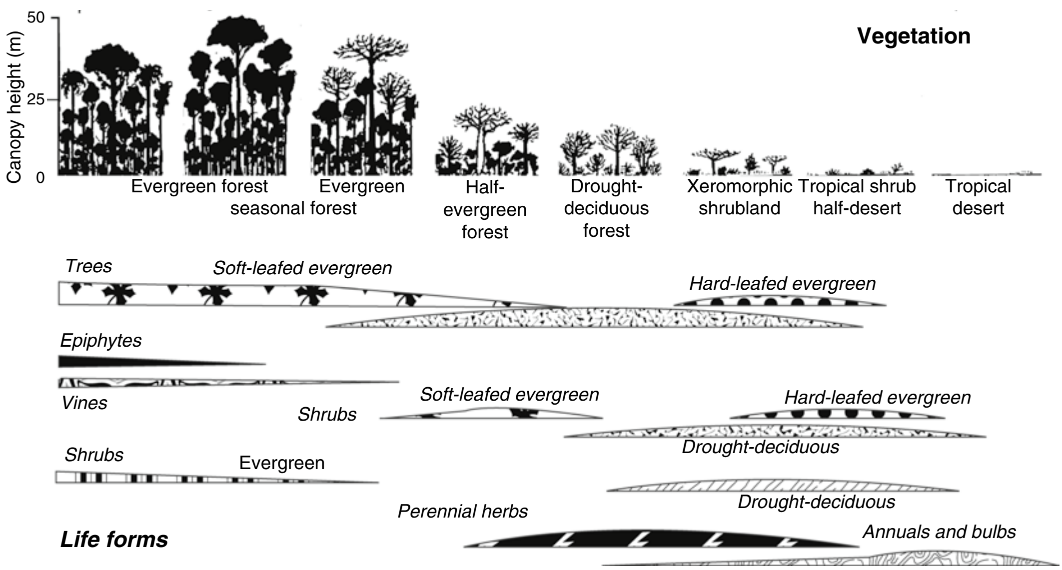


**Fig. 2.35** The change in life-form dominance along a tropical gradient where precipitation changes but temperature is relatively constant. Redrawn from Ellenberg (1979)

the incoming radiation. Light then becomes the main driver of diversity within the ecosystem. Plants that reach the canopy and have access to light compete well with tall trees. These growth forms include vines, which parasitize trees for support without investing carbon in strong stems. Epiphytes are also common in the canopies of tropical wet forests where they receive abundant light, but, because their roots are restricted to the canopy, their growth is often water-limited. Epiphytes have therefore evolved various specializations to trap water and nutrients. There is a wide range of sub-canopy trees, shrubs, and herbs that are adapted to grow slowly under the low-light conditions beneath the canopy (Fig. 2.35). Light is the most important general driver of structural diversity in the dense forests of wet tropical regions.

**Fig. 2.36** Patagonian steppe in cold, arid mountains of Argentina. Steppe is an example of a cold, dry ecosystem type intermediate between widespread "biomes." Photograph courtesy of Sandra Díaz

What determines structural diversity where moisture, rather than light, is limiting? Deserts, particularly warm deserts, have a great diversity of plant forms, including evergreen and deciduous small trees and shrubs, succulents, herbaceous perennials, and annuals. These growth forms do not show a well-defined vertical partitioning but show consistent horizontal patterns related to moisture availability. Trees and tall shrubs, for example, predominate adjacent to seasonal streams, evergreen shrubs in clay-rich soils that retain water, and succulents in the driest habitats. Competition for water results in diverse strategies for gaining, storing, and using the limited water supply. This leads to a wide range of rooting strategies and capacities to avoid or withstand drought.

**Species diversity declines from the tropics to high latitudes and in many cases from low to high elevation.** Species-rich tropical areas support more than 5,000 species of plants in a 10,000-km$^2$ area, whereas the high arctic has fewer than 200 species in the same area. Many animal groups show similar latitudinal patterns of diversity, in part because of their dependence on the underlying plant diversity. Climate, the evolutionary time available for species radiation, productivity, disturbance frequency, competitive interactions, land area available, and other factors have all been hypothesized to contribute to global patterns of diversity (Heywood and Watson 1995). Models that include only climate, acting as a filter on the plant functional types that can occur in a region, can reproduce the general global patterns of structural and species diversity (Fig. 2.37; Kleiden and Mooney 2000). The actual causes for geographic patterns of species diversity are undoubtedly more complex, but these models and other analyses suggest that human-induced changes in climate, land use, and invasions of exotic species may alter future patterns of diversity.

## Summary

The balance between incoming and outgoing radiation determines Earth's energy budget. The atmosphere transmits about half of the incoming shortwave solar radiation to Earth's surface but absorbs 90% of the outgoing longwave radiation emitted by Earth. This causes the atmosphere to be heated primarily from the bottom

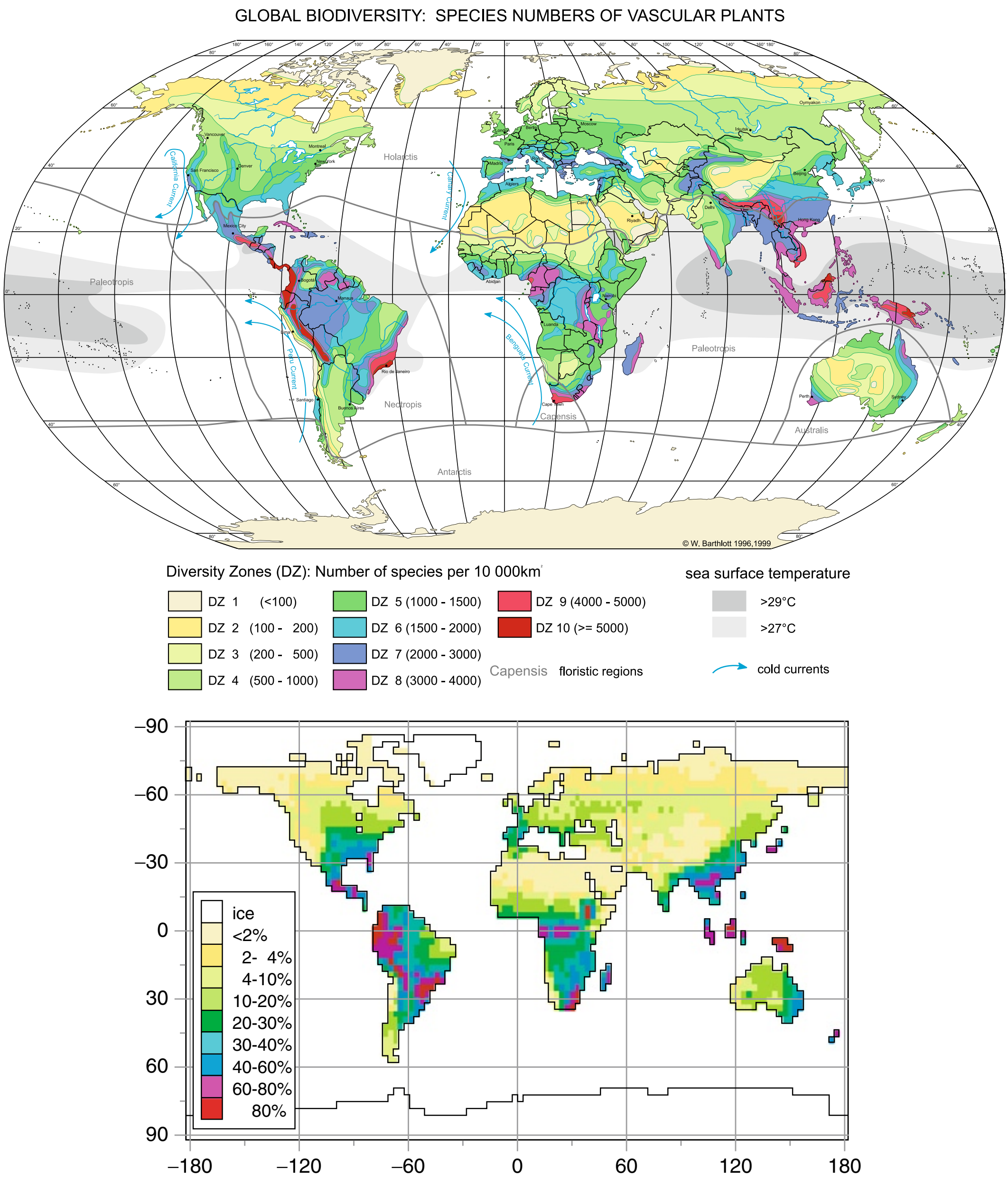


**Fig. 2.37** Global distribution of species richness based on observations (*top*; units number of species per 10,000 km$^2$) and on model simulations (*bottom*; units % of maximum diversity simulated) that use climate as a filter to reduce the number of allocation strategies. Reprinted from Kleiden and Mooney (2000)

and generates convective motion in the atmosphere. Large-scale patterns of atmospheric circulation occur because the tropics receive more energy from the sun than they emit to space, whereas the poles lose more energy to space than they receive from the sun. The resulting circulation cells transport heat from the equator to the poles to balance these inequalities. In the process, they create three relatively distinct air masses in each hemisphere, a tropical air mass (0–30°N and S), a temperate air mass (30–60°N and S), and a polar air mass (60–90°N and S). There are four major areas of high pressure (the two poles and 30°N and S), where air descends, and precipitation is low. The subtropical high-pressure belts are the zones of the world's major deserts. There are three major zones of low pressure (the equator and 60°N and S), where air rises, and precipitation is high. These areas support the tropical rainforests at the equator and the temperate rainforests of northwestern North America and southwestern South America. Ocean currents account for about 40% of the latitudinal heat transport from the equator to the poles. These currents are driven by surface winds and by the downwelling of cold saline waters at high latitudes, balanced by upwelling at lower latitudes.

Regional and local patterns of climate reflect heterogeneity in Earth's surface. Uneven heating between the land and the ocean modifies the general latitudinal patterns of climate by generating zones of prevailing high and low pressure. These pressure centers are associated with storm tracks that are guided by major mountain ranges in ways that strongly influence regional patterns of climate. The ocean and large lakes also moderate climate on adjacent lands because their high heat capacity causes them to heat or cool more slowly than land. These heating contrasts produce predictable seasonal winds (monsoons) and daily winds (land/sea breezes) that influence the adjacent land. Mountains also create heterogeneity in precipitation and in the quantity of solar radiation intercepted.

Vegetation influences climate through its effects on surface albedo, which determines the quantity of incoming radiation absorbed by the surface, and energy released to the atmosphere via longwave radiation and turbulent fluxes of latent and sensible heat. Sensible heat fluxes and longwave radiation directly heat the atmosphere, and latent heat transfers water vapor to the atmosphere, influencing local temperature and moisture sources for precipitation.

Climate is variable over all time scales. Long-term variations in climate are driven largely by changes in solar input and atmospheric composition. Superimposed on these long-term trends are predictable daily and seasonal patterns of climate, as well as repeating patterns such as those associated with El Niño/Southern Oscillation. These oscillations cause widespread changes in the geographic pattern of climate on time scales of years to decades. Future changes in climate may reflect changes in the frequencies of these large-scale climate modes.

## Review Questions

1. Describe the energy budget of Earth's surface and the atmosphere. What are the major pathways by which energy is absorbed by Earth's surface? By the atmosphere? What are the roles of clouds and radiatively active gases in determining the relative importance of these pathways?
2. Why is the troposphere warmest at the bottom but the stratosphere is warmest at the top? How does each of these atmospheric layers influence the environment of ecosystems?
3. Explain how unequal heating of Earth by the sun and the resulting atmospheric circulation produces the major latitudinal climate zones, such as those characterized by tropical forests, subtropical deserts, temperate forests, and arctic tundra.
4. How do the rotation of Earth (and the resulting Coriolis effect) and the separation of Earth's surface into the ocean and continents influence the global patterns of climate?
5. How does the chemical composition of Earth's atmosphere influence the climate of Earth?
6. What causes the global pattern in surface ocean currents? Why are the deep-water ocean currents different from those at the surface? What is the nature of the connection between deep- and surface-ocean currents?
7. How does ocean circulation influence climate at global, continental, and local scales?
8. How does topography affect climate at continental and local scales?
9. What are the major causes of long-term changes in climate? How would you expect future climate to differ from that of today in 100 years? 100,000 years? 1 billion years? Explain your answers.
10. Explain how the interannual variations in climate of Indonesia, Peru, and California are interconnected.

11. Explain the climatic basis for the global distribution of each major biome type. Use maps of global winds and ocean currents to explain these distributions.
12. Describe the climate of your birthplace. Using your understanding of the global climate system, explain why this location has its characteristic climate.

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