The Earth System

2

Climate depends not only on atmospheric processes, but also on physical, chemical, and biological processes involving other components of the Earth system. This chapter reviews the structure and behavior of those other components. We show how the cycling of water, carbon, and oxygen among the components of the Earth system has affected the evolution of the atmosphere. Drawing on this background, we summarize the history of climate over the lifetime of the Earth, with emphasis on causal mechanisms. The final section discusses why Earth is so much more habitable than its neighbors in the solar system. Chapter 10 revisits some of these same topics in the context of climate dynamics, with a quantitative discussion of feedbacks and climate sensitivity.

2.1 Components of the Earth System

This section introduces the cast of characters and briefly describes their roles and interrelations in the ongoing drama of climate. The atmosphere, which in some sense plays the starring role, has already been introduced in Chapter 1. The interplay between atmospheric radiation and convection regulates the temperature at the Earth's surface, setting the limits for snow and ice cover and for the various life zones in the biosphere. The stratospheric ozone layer protects the biosphere from the lethal effects of solar ultraviolet radiation. Atmospheric wind patterns regulate the patterns of oceanic upwelling that supplies nutrients to the marine biosphere, they determine the distribution of water that sustains the terrestrial (land) biosphere, and they transport trace gases, smoke, dust, insects, seeds, and spores over long distances. Rain, frost, and wind erode the Earth's crust, wearing down mountain ranges, reshaping the landscape, and replenishing the soils and the supply of metallic ions needed to sustain life.

Other components of the Earth system also play important roles in climate. The oceans are notable for their large "thermal inertia" and their central role in the cycling of carbon, which controls atmospheric carbon dioxide concentrations. Extensive snow and ice-covered surfaces render the Earth more reflective, and consequently cooler, than it would be in their absence. By evaporating large quantities of water through their leaves, land plants exert a strong moderating influence on tropical and extratropical summer climate. Living organisms on land and in the sea have been instrumental in liberating oxygen and sequestering of carbon in the Earth's crust, thereby reducing the atmospheric concentration carbon dioxide. On timescales of millions of years or longer, plate tectonics exerts an influence on climate through continental drift, mountain building, and volcanism. This section describes these processes and the media in which they occur.

2.1.1 The Oceans

The oceans cover 72% of the area of the Earth's surface and they reach an extreme depth of nearly 11 km. Their total volume is equivalent to that of a layer 2.6 km deep, covering the entire surface of the Earth. The mass of the oceans is ~250 times as large as that of the atmosphere.

a. Composition and vertical structure

The density of sea water is linearly dependent on the concentration of dissolved salt. On average, sea water in the open oceans contains ~35 g of dissolved salts per kg of fresh water, with values typically ranging from 34 to 36 g kg^{-1} (or parts per thousand by mass, abbreviated as o/oo). Due to the presence of these dissolved salts, sea water is ~2.4% denser than fresh water at the same temperature.

The density σ of sea water (expressed as the departure from 1 in g kg^{-1} or o/oo) typically ranges from 1.02 to 1.03. It is a rather complicated function of temperature T, salinity s, and pressure p; i.e., $\sigma = \sigma(T, s, p)$. The pressure dependence of density in liquids is much weaker than in gases and, for purposes of this qualitative discussion, will be ignored.[1] As in fresh water, $\partial\sigma/\partial T$ is temperature dependent, but the fact that sea water is saline makes the relationship somewhat different: in fresh water, density increases with increasing temperature between 0 and 4 °C, whereas in sea water, density decreases monotonically with increasing temperature.[2] In both fresh water and sea water, $\partial\sigma/\partial T$ is smaller near the freezing point than at higher temperatures. Hence, a salinity change of a prescribed magnitude δs is equivalent, in terms of its effect on density, to a larger temperature change δT in the polar oceans than in the tropical oceans, as illustrated in Fig. 2.1.

Over most of the world's oceans, the density of the water in the wind-stirred, *mixed layer* is smaller, by a few tenths of a percent, than the density of the water below it. Most of the density gradient tends to be concentrated within a layer called the *pycnocline*, which ranges in depth from a few tens of meters to a few hundred meters below the ocean surface. The density gradient within the pycnocline tends to inhibit vertical mixing in the ocean in much the same manner that the increase of temperature with height inhibits vertical mixing in atmospheric temperature inversions and in the stratosphere. In particular, the pycnocline strongly inhibits the exchange of heat and salt between the mixed layer, which is in direct contact with the atmosphere, and the deeper layers of the ocean. At lower latitudes, *pycnocline* is synonymous with the *thermocline* (i.e., the layer in which temperature increases with height), but in polar oceans, *haloclines* (layers with fresher water above and saltier water below) also play an important role in inhibiting vertical mixing. The strength and depth of the thermocline vary with latitude and season, as illustrated in the idealized profiles shown in Fig. 2.2.

Within the oceanic mixed layer, temperature and salinity (and hence density) vary in response to

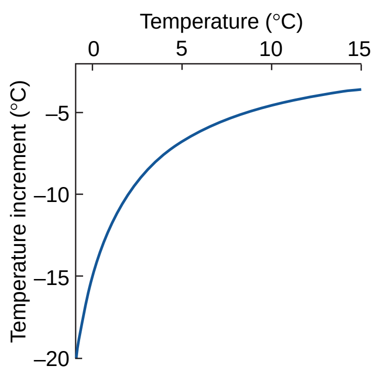

Fig. 2.1 The change in temperature of a water parcel required to raise the density of sea water at sea level as much as a salinity increase of 1 g kg^{-1}, plotted as a function of the temperature of the parcel. For example, for sea water at a temperature of 10 °C, a salinity increase of 1 g kg^{-1} would raise the density as much as a temperature decrease of ~5 °C, whereas for sea water at 0 °C the same salinity increase would be equivalent to a temperature change of ~17 °C. [Adapted from data in M. Winton, Ph.D. thesis, University of Washington, p. 124 (1993).]

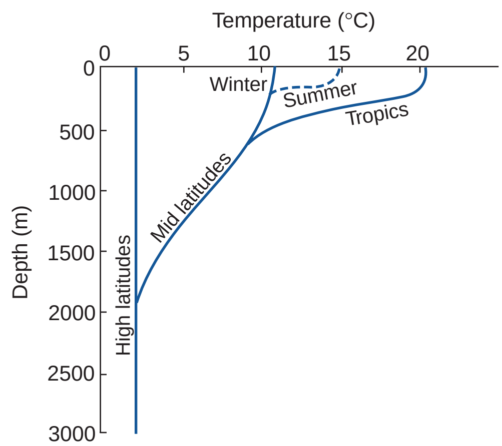

Fig. 2.2 Idealized profiles of the temperature plotted as a function of depth in different regions of the world's oceans. The layer in which the vertical temperature gradient is strongest corresponds to the thermocline. [From J. A. Knauss, *Introduction to Physical Oceanography*, 2nd Edition, p. 2, © 1997. Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ.]

[1] The small effect of pressure upon density is taken into account through the use of *potential density*, the density that a submerged water parcel would exhibit if it were brought up to sea level, conserving temperature and salinity. (See Exercise 3.54.)

[2] Ice floats on lakes because the density of fresh water decreases with temperature from 0 to 4 °C. In contrast, sea ice floats because water rejects salt as it freezes.

exchanges of heat and water with the atmosphere. Precipitation lowers the salinity by diluting the salts that are present in the oceanic mixed layer, and evaporation raises the salinity by removing fresh water and thereby concentrating the residual salts, as illustrated in the following example.

Exercise 2.1 A heavy tropical storm dumps 20 cm of rainfall in a region of the ocean in which the salinity is 35.00 g kg^{-1} and the mixed layer depth is 50 m. Assuming that the water is well mixed, by how much does the salinity decrease?

Solution: The volume of water in a column extending from the surface of the ocean to the bottom of the mixed layer is increased by a factor

$$\frac{0.2 \text{ m}}{50 \text{ m}} = 4 \times 10^{-3}$$

and (ignoring the small difference between the densities of salt water and fresh water) the mass of the water in the column increases by a corresponding amount. The mass of salt dissolved in the water remains unchanged. Hence, the salinity drops to

$$\frac{35.00 \text{ g of salt}}{1.004 \text{ kg of water}} = 34.86 \text{ g kg}^{-1}. \quad ■$$

Water parcels that are not in contact with the ocean surface tend to conserve temperature and salinity as they move over long distances. Hence, *water masses* (layers of water extending over large areas that exhibit nearly uniform temperature and salinity) can be tracked back to the regions of the mixed layer in which they were formed by exchanges of heat and mass with the atmosphere. Among the important water masses in the Atlantic Ocean, in order of increasing density, are:

- *Mediterranean outflow*, which is conspicuously warm and saline due to the excess of evaporation over precipitation in the Mediterranean Sea.
- *North Atlantic deep water (NADW)*, formed by the sinking of water along the ice edge in the Greenland, Iceland, and Norwegian Seas.
- *Antarctic bottom water (AABW)*, formed by sinking along the ice edge in the Weddell Sea.

The NADW and AABW, each marked by its own distinctive range of temperatures and salinities, are

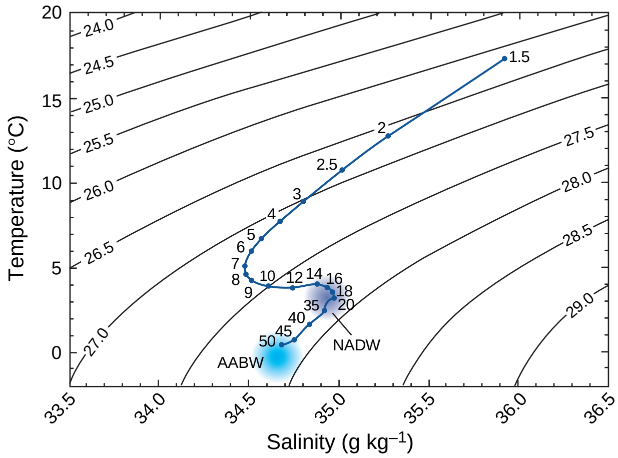

Fig. 2.3 Vertical sounding of water temperature and salinity in a vertical sounding in the subtropical Atlantic Ocean. Numbers along the sounding indicate depths in hundreds of meters. Potential (i.e., pressure-adjusted) density (in *o/oo*) is indicated by the contours. Characteristic temperature and salinity ranges for North Atlantic deep water (NADW) and Antarctic bottom water (AABW) are indicated by shading. [Reprinted from *Seawater: Its Composition, Properties and Behavior*, The Open University in association with Pergamon Press, p. 48 (1989), with permission from Elsevier.]

both clearly evident near the bottom of the tropical sounding shown in Fig. 2.3. The AABW is slightly colder and fresher than the NADW. When both temperature and salinity are taken into account, the AABW is slightly denser than the NADW, consistent with its placement at the bottom of the water column.

b. The ocean circulation

The ocean circulation is composed of a *wind-driven* component and a *thermohaline* component. The wind-driven circulation dominates the surface currents, but it is largely restricted to the topmost few hundred meters. The circulation deeper in the oceans is dominated by the slower thermohaline circulation.

By generating ocean waves, surface winds transfer horizontal momentum from the atmosphere into the ocean. The waves stir the uppermost layer of the ocean, mixing the momentum downward. The momentum, as reflected in the distribution of surface currents shown in Fig. 2.4, mirrors the pattern of surface winds shown in Figs. 1.18 and 1.19, with closed anticyclonic circulations (referred to as *gyres*) at subtropical latitudes and cyclonic gyres at subpolar latitudes. Another notable feature of the wind-driven

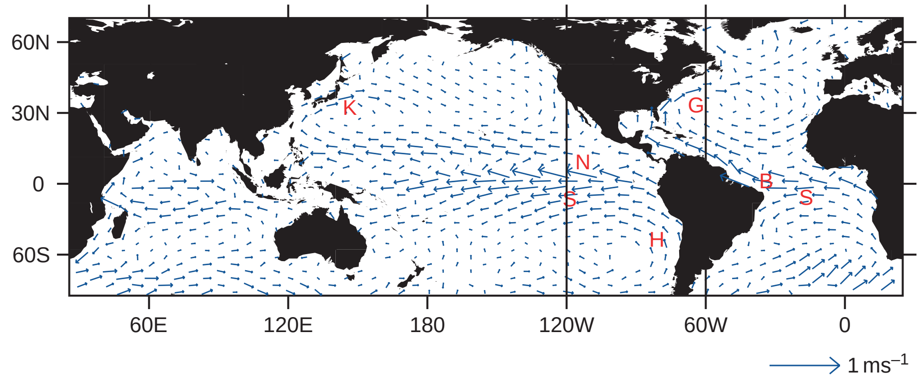

Fig. 2.4 Annual mean ocean surface currents based on the rate of drift of ships. The *Gulf Stream* (G) and the *Kuroshio Current* (K) are warm, *western boundary currents*. The *Humboldt Current* (H) is the most prominent of the cold, equatorward currents driven by the winds along the eastern flanks of the subtropical anticyclones. The westward *South Equatorial Current* (S) is driven by the easterlies along the equator and the weaker eastward *North Equatorial Countercurrent* (N) is a response to the winds in the vicinity of the ITCZ. [Data courtesy of Philip Richardson, WHOI; graphic courtesy of Todd P. Mitchell.]

circulation is the west-to-east *Antarctic circumpolar current* along 55 °S, the latitude of the Drake passage that separates Antarctica and South America. Velocities in these wind-driven currents are typically on the order of 10 cm s^{-1}, a few percent of the speeds of the surface winds that drive them, but in the narrow *western boundary currents* such as the *Gulf Stream* off the east coast of the United States (Figs. 2.4 and 2.5) velocities approach 1 m s^{-1}. The relatively warm water transported poleward by the western boundary currents contributes to moderating winter temperatures over high latitude coastal regions.

Over certain regions of the polar oceans, water in the mixed layer can become sufficiently dense, by virtue of its high salinity, to break through the pycnocline and sink all the way to the ocean floor to become what oceanographers refer to as *deep water* or *bottom water*. In some sense, these negatively buoyant plumes are analogous to the plumes of warm, moist air in low latitudes that succeed in breaking through the top of the atmospheric mixed layer and continue ascending until they encounter the tropopause. The presence of CFCs[3] in NADW and AABW indicates that these water

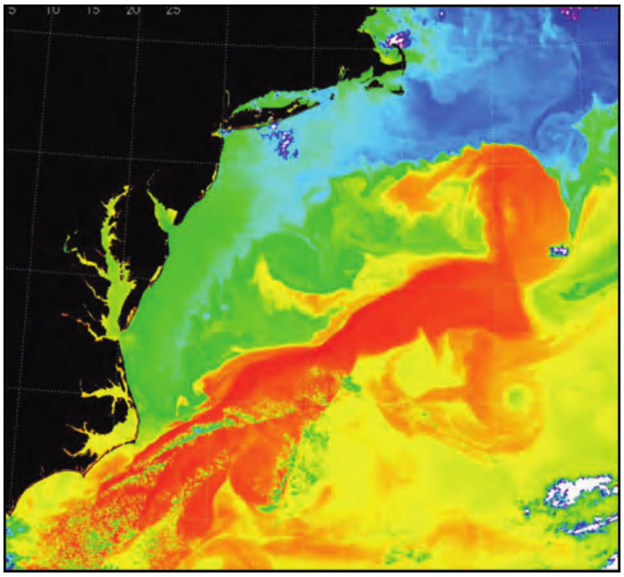

Fig. 2.5 Eddies along the landward edge of the Gulf Stream, as revealed by the pattern of sea surface temperature. Temperatures range from ~20 °C in the orange regions down to ~6 °C in the darkest blue regions. Note the sharpness of the boundary and the indications of turbulent mixing between the waters of the Gulf Stream and the colder Labrador Current to the north of it. [Based on NASA Terra/MODIS imagery. Courtesy of Otis Brown.]

[3] The term *chlorofluorocarbons (CFCs)* refers to a family of gaseous compounds that have no natural sources; first synthesized in 1928. Atmospheric concentrations of CFCs rose rapidly during the 1960s and 1970s as these gases began to be used for a widening range of purposes.

masses were in relatively recent contact with the atmosphere.

By virtue of their distinctive chemical and isotopic signatures, it is possible to track the flow of water masses and to infer how long ago water in various parts of the world's oceans was in contact with the atmosphere. Such chemical analyses indicate the existence of a slow overturning characterized by a spreading of deep water from the high latitude sinking regions, a resurfacing of the deep water, and a return flow of surface waters toward the sinking regions, as illustrated in Fig. 2.6. The timescale in which a parcel completes a circuit of this so-called *thermohaline circulation* is on the order of hundreds of years.

The resurfacing of deep water in the thermohaline circulation requires that it be *ventilated* (i.e., mixed with and ultimately replaced by less dense water that has recently been in contact with the ocean surface). Still at issue is just how this ventilation occurs in the presence of the pycnocline. One school of thought attributes the ventilation to mixing along sloping isopycnal (constant density) surfaces that cut through the pycnocline. Another school of thought attributes it to irreversible mixing produced by tidal motions propagating downward into the deep oceans along the continental shelves, and yet another to vertical mixing in restricted regions characterized by strong winds and steeply sloping isopycnal surfaces, the most important of which coincides with the *Antarctic circumpolar current*, which lies beneath the ring of strong westerly surface winds that encircles Antarctica.

Although most of the deep and bottom water masses are formed in the Atlantic sector, the thermohaline circulation involves the entire world's oceans, as illustrated in Fig. 2.7. Within the Atlantic sector itself, the thermohaline circulation is comprised of two different cells: one involving NADW and the other involving AABW, as illustrated in Fig. 2.8.

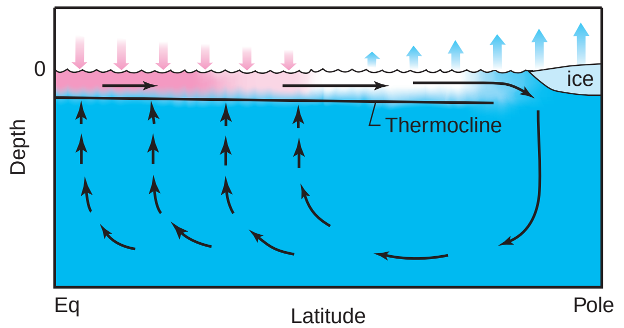

Fig. 2.6 Idealized schematic of the thermohaline circulation in an equatorially symmetric ocean. The domain extends from the sea floor to the ocean surface and from equator to pole. Pink shading indicates warmer water and blue shading indicates colder water. The shaded arrows represent the exchange of energy at the air–sea interface: pink downward arrows indicate a heating of the ocean mixed layer and blue upward arrows indicate a cooling. The role of salinity is not specifically represented in this schematic but it is the rejection of salt when water freezes along the ice edge that makes the water dense enough to enable it to sink to the bottom.

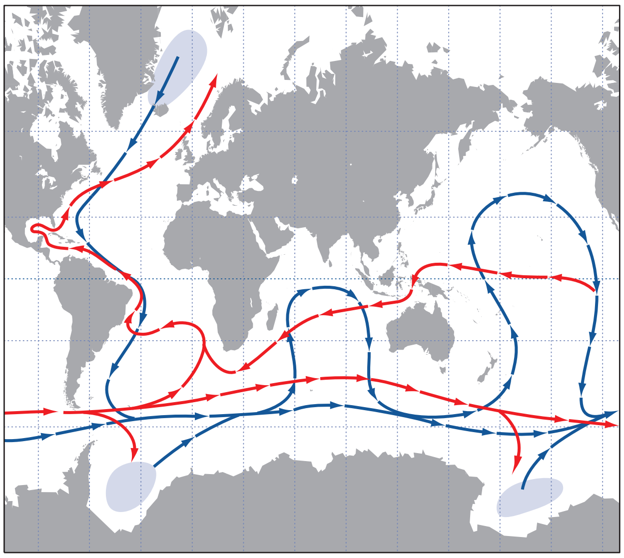

Fig. 2.7 Highly simplified schematic of the thermohaline circulation. Shading denotes regions of downwelling, blue arrows denote transport of bottom water, and red arrows denote the return flow of surface water. [Adapted from W. J. Schmitz, Jr., "On the interbasin-scale thermohaline circulation," *Rev. Geophys.*, **33**, p. 166, Copyright 1995 American Geophysical Union.]

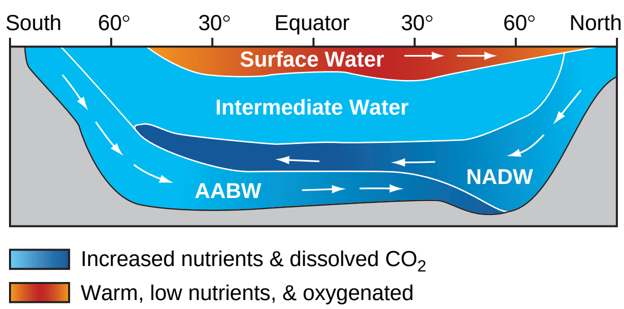

Fig. 2.8 Idealized cross section of the thermohaline circulation in the Atlantic Ocean. In this diagram, *Intermediate Water* comprises several different water masses formed at temperate latitudes. Note the consistency with Fig. 2.3. [Courtesy of Steve Hovan.]

c. The marine biosphere

Virtually all the sunlight that reaches the surface of the ocean is absorbed within the topmost hundred meters. Within this shallow *euphotic zone*,[4] life abounds wherever there are sufficient nutrients, such as phosphorous and iron, to sustain it. In regions of the ocean where the marine biosphere is active, the uppermost layers are enriched in dissolved oxygen (a product of photosynthesis) and depleted in nutrients and dissolved carbon, as illustrated in Fig. 2.9. *Phyto-* (i.e., plant) *plankton* are capable of consuming the nutrients in the euphotic zone within a matter of days. Hence, the maintenance of high *primary productivity* (i.e., photosynthesis) requires a continual supply of nutrients. The most productive regions of the oceans tend to be concentrated in regions of upwelling, where nutrient-rich sea water from below the euphotic zone is first exposed to sunlight.

Nutrients consumed within the euphotic zone by phytoplankton return to the deeper layers of the oceans when marine plants and animals that feed on them die, sink, and decompose. The continual exchange of nutrients between the euphotic zone and the deeper layers of the ocean plays an important role in the carbon cycle, as discussed in Section 2.3. The distribution of upwelling, in turn, is controlled by the pattern of surface winds discussed earlier. The distribution of *ocean color* (Fig. 2.10) shows evidence of high biological productivity and, by inference, upwelling

- beneath cyclonic circulations such as Aleutian and Icelandic lows,
- along the eastern shores of the oceans at subtropical latitudes,
- in a narrow strip along the equator in the equatorial Atlantic and Pacific Oceans.

In contrast, the ocean regions that lie beneath the subtropical anticyclones are biological deserts. The dynamical basis for these relationships is discussed in Section 7.2.5. Through their effect in mediating the geographical distribution of upwelling and the depth of the mixed layer, year-to-year changes in the atmospheric circulation, such as those that occur in association with El Niño, perturb the entire food chain that supports marine mammals, seabirds, and commercial fisheries.

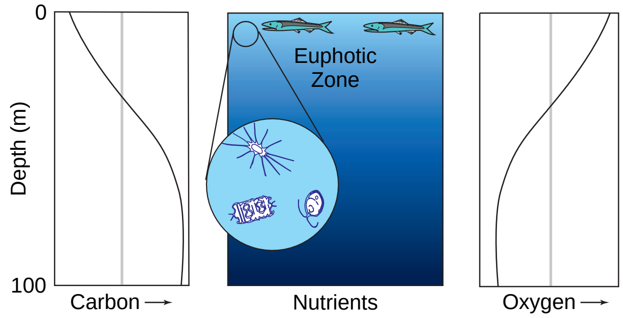

Fig. 2.9 Idealized vertical profiles of dissolved carbon (left) and oxygen (right) in biologically active regions of the oceans. The intensity of sunlight is indicated by the depth of the shading in the middle panel.

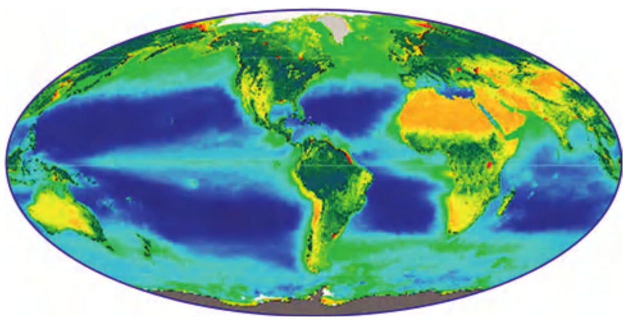

Fig. 2.10 Distribution of primary productivity in the marine and terrestrial biosphere, averaged over a 3-year period. Over the oceans the dark blue areas are indicative of very low productivity and the green and yellow areas are relatively more productive. Over land dark green is indicative of high productivity. [Imagery courtesy of SeaWiFS Project, NASA/GSFC and ORBIMAGE, Inc.]

d. Sea surface temperature

The global distribution of sea surface temperature is shaped by both radiative and dynamical factors relating to the pattern of seasonally varying, climatological-mean surface wind field over the oceans (Fig. 1.18). Radiative heating is the dominant factor. That incident solar radiation is so much stronger in the tropics than in the polar regions gives rise to a strong north–south temperature gradient, which dominates the annual-mean field shown in Fig. 2.11 (top).

The effects of the winds on the sea surface temperature pattern become more clearly apparent when the zonally averaged sea surface temperature at each latitude is removed from the total field, leaving just

[4] Greek: *eu*-good and *photic*-light.

the departures from the zonal-mean, shown in Fig. 2.11 (bottom). The coolness of the eastern oceans relative to the western oceans at subtropical latitudes derives from circulation around the subtropical anticyclones (Fig. 1.16). The equatorward flow of cool air around the eastern flanks of the anticyclones extracts a considerable quantity of heat from the ocean surface, as explained in Section 9.3.4, and drives cool, southward ocean currents (Fig. 2.4). In contrast, the warm, humid poleward flow around their western flanks extracts much less heat and drives warm western boundary currents such as the Gulf Stream. At higher latitudes the winds circulating around the subpolar cyclones have the opposite effect, cooling the western sides of the oceans and warming the eastern sides. The relative warmth of the eastern Atlantic at these higher latitudes is especially striking.

Wind-driven upwelling is responsible for the relative coolness of the equatorial eastern Pacific and Atlantic, where the southeasterly trade winds protrude northward across the equator (Fig. 1.18). Wind-driven upwelling along the coasts of Chile, California, and continents that occupy analogous positions with respect to the subtropical anticyclones, although not well resolved in Fig. 2.11, also contributes to the coolness of the subtropical eastern oceans, as do the highly reflective cloud layers that tend to develop at the top of the atmospheric boundary layer over these regions (Section 9.4.4).

The atmospheric circulation feels the influence of the underlying sea surface temperature pattern, particularly in the tropics. For example, from a comparison of Figs. 1.25 and 2.11 it is evident that the intertropical convergence zones in the Atlantic and Pacific sectors are located over bands of relatively warm sea surface temperature and that the dry zones lie over the *equatorial cold tongues* on the eastern sides of these ocean basins.

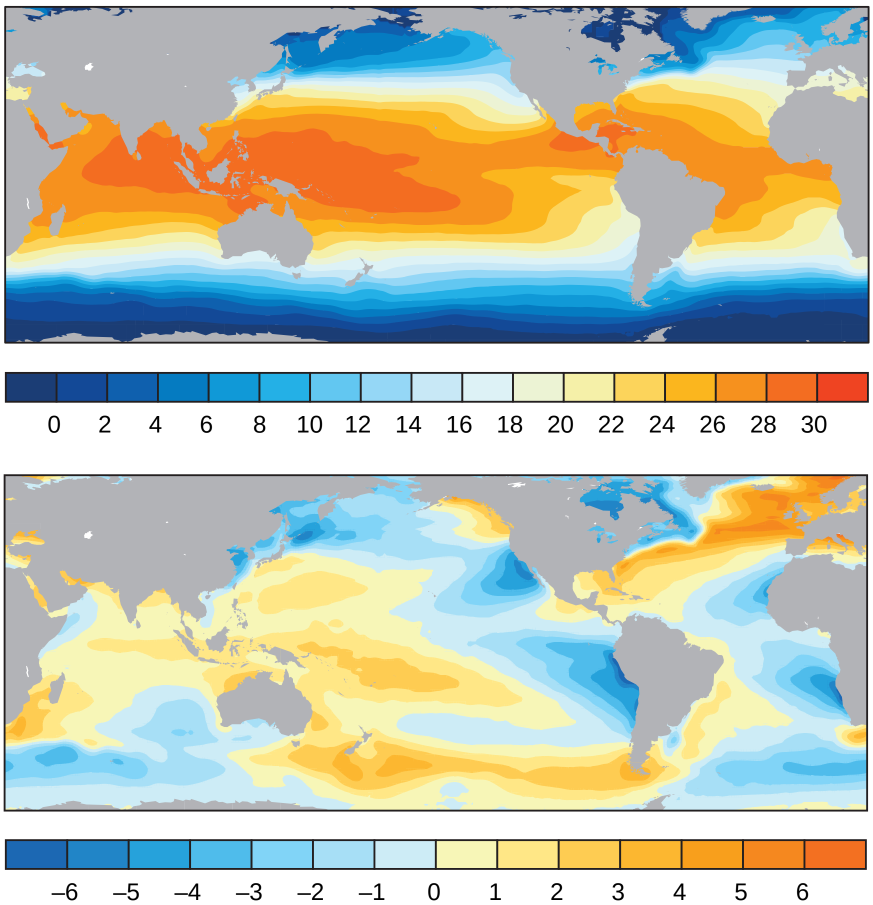

Fig. 2.11 Annual mean sea surface temperature. (Top) The total field. (Bottom) Departure of the local sea surface temperature at each location from the zonally average field. [Based on data from the U.K. Meteorological Office HadISST dataset. Courtesy of Todd P. Mitchell.]

2.1.2 The Cryosphere

The term *cryo-* (frozen) *sphere* refers to components of the Earth system comprised of water in its solid state, or in which frozen water is an essential component. The cryosphere contributes to the thermal inertia of the climate system; it contributes to the reflectivity or *albedo* of the Earth; by taking up and releasing fresh water in the polar regions, it influences oceanic thermohaline circulation; and it stores enough water to significantly the influence global sea level. The elements of the cryosphere are listed in Table 2.1 and all of them, with the exception of alpine glaciers, are represented in Fig. 2.12.

The *continental ice sheets*, dominated by Antarctica and Greenland, are the most massive elements of the cryosphere. The ice sheets are continually replenished by snowfall; they lose mass by sublimation, by the calving of icebergs, and, in summer, by runoff in streams and rivers along their periphery. The net *mass balance* (i.e., the balance between the mass sources and sinks) at any given time determines whether an ice sheet is growing or shrinking.

Over periods of tens of thousands of years and longer, annual layers of snow that fall in the relatively flat interior of the ice sheets are compressed by the accumulation of new snow on top of them. As the pressure increases, snow is transformed into ice. Due to the dome-like shape of the ice sheets and the plasticity of the ice itself, the compressed layers of ice gradually creep downhill toward the periphery of the ice sheet, causing the layer as a whole to spread out horizontally and (in accordance with the conservation of mass) to thin in the vertical dimension. Much of the flow toward the periphery tends to

Table 2.1 Surface area and mass of the various components of the cryosphere[a]

Cryospheric component	Area	Mass
Antarctic ice sheet	2.7	53
Greenland ice sheet	0.35	5
Alpine glaciers	0.1	0.2
Arctic sea ice (March)	3	0.04
Antarctic sea ice (September)	4	0.04
Seasonal snow cover	9	<0.01
Permafrost	5	1

[a] Surface area is expressed as percentage of the area of the surface of the Earth. Mass is expressed in units of 10^3 kg m^{-2} (numerically equivalent to meters of liquid water) averaged over the entire surface area of the Earth. For reference, the total surface area of the Earth and the area of the Earth covered by land are 5.12 and 1.45×10^{14} m^2, respectively. [Courtesy of S. G. Warren.]

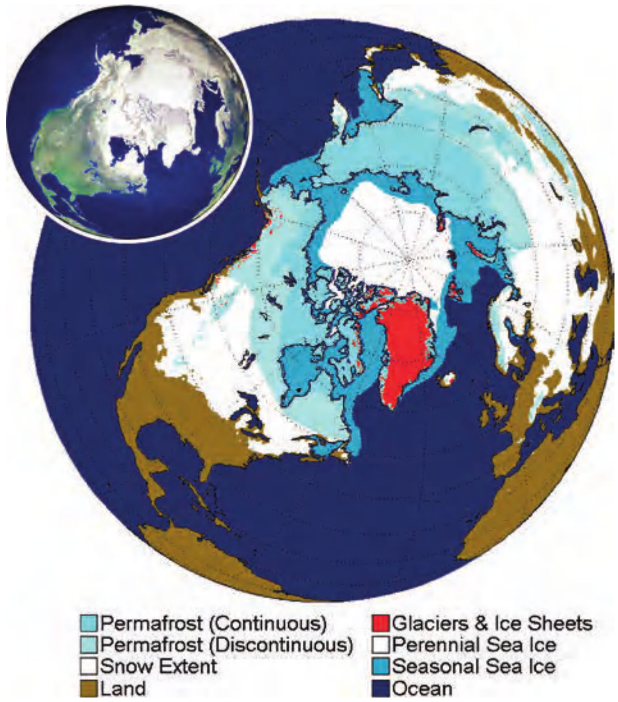

Fig. 2.12 Elements of the northern hemisphere cryosphere. The equatorward edge of the snow cover corresponds to ~50% coverage during the month of maximum snow extent. [Courtesy of Ignatius Rigor.] The inset at the upper left shows a NASA RADARSAT image highlighting these features.

be concentrated in relatively narrow, fast-moving ice streams tens of kilometers in width (Fig. 2.13).

Along the divides of the ice sheets the movement is very slow and the layering of the ice is relatively undisturbed. In *ice cores* extracted from these regions, the age of the ice increases monotonically with depth to ~100,000 years in the Greenland ice sheet and over 500,000 years in the Antarctic ice sheet. Analysis of air bubbles, dust, and chemical and biological tracers embedded within these ice cores is providing a wealth of information on the climate of the past few hundred thousand years, as discussed later in this chapter.

In many respects, *alpine* (i.e., mountain) *glaciers* behave like continental ice sheets, but they are much smaller in areal coverage and mass. Their fate is also determined by their mass balance. Parcels of ice within them flow continually downhill from an upper dome-like region where snow and ice accumulate toward their snouts where mass is lost continually due to melting. Because of their much smaller masses, glaciers respond much more quickly to climate change than continental ice sheets, and ice cycles through them much more rapidly. Some alpine glaciers also exhibit time-dependent behavior that is not climate

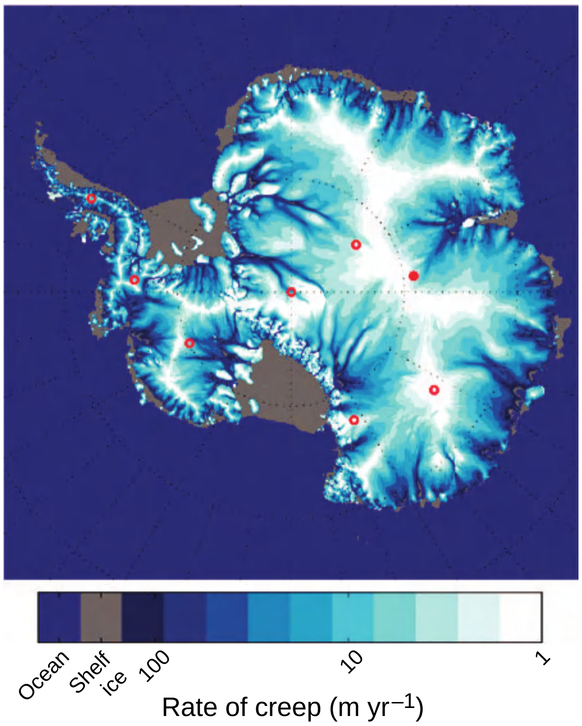

Fig. 2.13 Satellite image of the Antarctic ice sheet showing rate of creep of the ice (in m year^{-1}) on a logarithmic scale. Dots show the locations of ice core sites. Vostok, the site of the ice core shown in Fig. 2.31, is indicated by the solid red dot. [Adapted with permission from Bamber, J. L., D. G. Vaughan and I. Joughin, "Widespread Complex Flow in the Interior of the Antarctic Ice Sheet," *Science*, **287**, 1248–1250. Copyright 2000 AAAS. Courtesy of Ignatius Rigor.]

related: episodic surges of a few months' to a few years' duration interspersed with much longer periods of slow retreat.

Sea ice covers a larger area of the Earth's surface area than the continental ice sheets (Table 2.1) but, with typical thicknesses of only 1–3 m, is orders of magnitude less massive. The ice is not a continuous surface, but a fractal field comprised of ice floes (pieces) of various of shapes and sizes, as shown in Figs. 2.14 and 2.15. The individual floes are separated by patches of open water (called *leads*) that open and close as the ice pack moves, dragged by surface winds.

Seasonal limits of the northern hemisphere pack ice are shown in Fig. 2.12. During winter, ice covers not only the Arctic, but also much of the Bering Sea and the Sea of Okhotsk, but during the brief polar summer the ice retreats dramatically and large leads are sometimes observed, even in the vicinity of the North Pole. Antarctic pack ice also advances and retreats with the seasons.

The annual-mean sea ice motion, shown in Fig. 2.16, is dominated by the clockwise *Beaufort Gyre* to the north of Alaska and the *transpolar drift stream* from Siberia toward Greenland and Spitzbergen.[5] Some ice floes remain in the Arctic for a decade or more, circulating around and around the Beaufort Gyre, whereas others spend just a year or two in the Arctic before they exit either through the Fram Strait between Greenland and Spitzbergen or through the Nares Strait into Baffin Bay along the west side of Greenland. Ice floes exiting the Arctic make a one-way trip into warmer waters, where they are joined by much thicker *icebergs* that break off the Greenland ice sheet.

New pack ice is formed during the cold season by the freezing of water in newly formed leads and in regions where offshore winds drag the pack ice away from the coastline, exposing open water. The new ice thickens rapidly at first and then more gradually as it begins to insulate the water beneath it from the subfreezing air above. Ice thicker than a meter is formed, not by a thickening of newly formed layer of

[5] The existence of a transpolar drift stream was hypothesized by Nansen[6] when he learned that debris from a shipwreck north of the Siberian coast had been recovered, years later, close to the southern tip of Greenland. Motivated by this idea, he resolved to sail a research ship as far east as possible off the coast of Siberia and allow it to be frozen into the pack ice in the expectation that it would be carried across the North Pole along the route suggested by Fig. 2.16. He supervised the design and construction of a research vessel, the *Fram* ("Forward"), with a hull strong enough to withstand the pressure of the ice. The remarkable voyage of the *Fram*, which began in summer of 1893 and lasted for 3 years, confirmed the existence of the transpolar drift stream and provided a wealth of scientific data.

[6] **Fridtjof Nansen** (1861–1930). Norwegian scientist, polar explorer, statesman, and humanitarian. Educated as a zoologist. Led the first traverse of the Greenland ice cap on skis in 1888. The drift of his research vessel the *Fram* across the Arctic (1893–1896) was hailed as a major achievement in polar research and exploration. Midway through this voyage, Nansen turned over command of the *Fram* to Harald Sverdrup and set out with a companion on what proved to be a 132-day trek across the pack ice with dog-drawn sledges and kayaks, reaching 86 °N before adverse conditions forced them to turn southward.

Sacrificed his subsequent aspirations for Antarctic exploration to serve the needs of his country and to pursue humanitarian concerns. Was instrumental in peacefully resolving a political dispute between Norway and Sweden in 1905–1906 and negotiating a relaxation of an American trade embargo that threatened Norwegian food security during World War I. Awarded the Nobel Peace Prize in 1922 in recognition of his extensive efforts on behalf of war refugees and famine victims.

Fig. 2.14 Ice floes and leads in Antarctic pack ice. The lead in the foreground is 4–5 m across. The floe behind it consists of multi-year ice that may have originated as an iceberg; it is unusually thick, extending from ~15 m below to ~1 m above sea level. Most of the portion of the floe that extends above sea level is snow. At the time this picture was taken, the pack ice in the vicinity was under lateral pressure, as evidenced by the fact that a *pressure ridge* had recently developed less than 100 m away. [Photograph courtesy of Miles McPhee.]

ice, but by mechanical processes involving collisions of ice floes. *Pressure ridges* up to 5 m in thickness are created when floes collide, and thickening occurs when part of one floe is pushed or *rafted* on top of another.

When sea water freezes, the ice that forms is composed entirely of fresh water. The concentrated salt water known as *brine* that is left behind mixes with the surrounding water, increasing its salinity. Brine rejection is instrumental in imparting enough negative buoyancy to parcels of water to enable them to break through the pycnocline and sink to the bottom. Hence, it is no accident that the sinking regions in the oceanic thermohaline circulation are in high latitudes, where sea water freezes.

Land snow cover occupies an even larger area of the northern hemisphere than sea ice and it varies

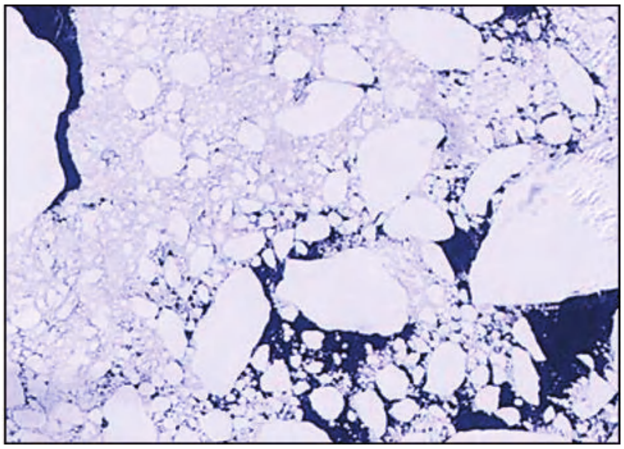

Fig. 2.15 Floes in pack ice streaming southward off the east coast of Greenland. The white area at the upper left is *landfast ice* that is attached to the coast, and the black channel adjacent to it is open water, where the mobile pack ice has become detached from the landfast ice. [NASA MODIS imagery.]

much more widely from week to week and month to month than does sea ice. With the warming of the land surface during spring, the snow virtually disappears, except in the higher mountain ranges.

Permafrost embedded in soils profoundly influences terrestrial ecology and human activities over large areas of Siberia, Alaska, and northern Canada. If the atmosphere and the underlying land surface

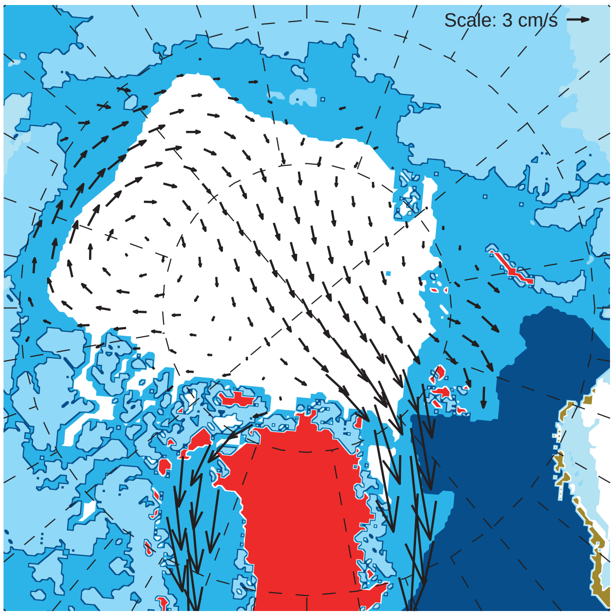

Fig. 2.16 Wintertime Arctic sea ice motion as inferred from the tracks of an array of buoys dropped on ice floes by aircraft. [Courtesy of Ignatius Rigor.]

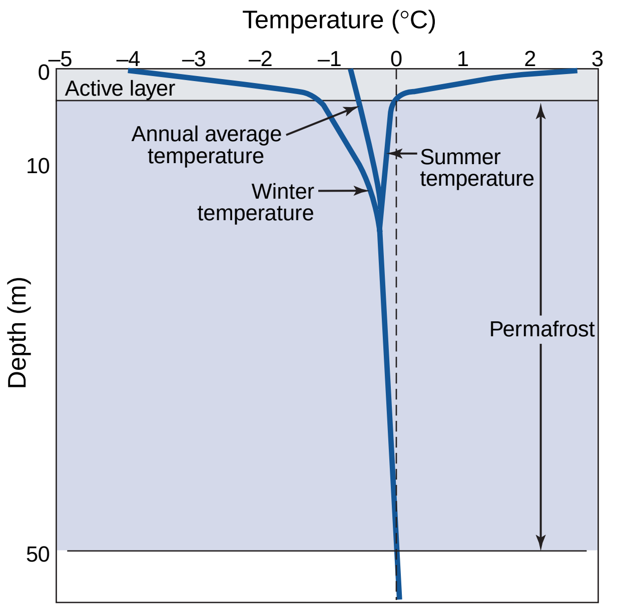

Fig. 2.17 Schematic vertical profile of summer and winter soil temperatures in a region of permafrost. The depth of the permafrost layer varies from as little as a few meters in zones of intermittent permafrost to as much as 1 km over the coldest regions of Siberia.

were in thermal equilibrium, the zones of continuous and intermittent permafrost in Fig. 2.12 would straddle the 0 °C isotherm in annual-mean surface air temperature. There is, in fact, a close correspondence between annual-mean surface air temperature and the limit of continuous permafrost, but the critical value of surface air temperature tends to be slightly above 0 °C due to the presence of snow cover, which insulates the land surface during the cold season, when it is losing heat.

Even in the zone of continuous permafrost, the topmost few meters of the soil thaw during summer in response to the downward diffusion of heat from the surface, as shown in Fig. 2.17. The upward diffusion of heat from the Earth's interior limits the vertical extent of the permafrost layer. Because the molecular diffusion of heat in soil is not an efficient heat transfer mechanism, hundreds of years are required for the permafrost layer to adjust to changes in the temperature of the overlying air.

2.1.3 The Terrestrial Biosphere

Much of the impact of climate upon animals and humans is through its role in regulating the condition and geographical distribution of forests, grasslands, tundra, and deserts, elements of the *terrestrial* (land) *biosphere*. A simple conceptual framework for relating climate (as represented by annual-mean temperature and precipitation) and vegetation type is shown in Fig. 2.18. The boundary between tundra and forest corresponds closely to the limit of the permafrost zone, which, as noted earlier, is determined by annual-mean temperature. The other boundaries in Fig. 2.18 are determined largely by the water requirements of plants. Plants utilize water both as raw material in producing chlorophyll and to keep cool on hot summer days, as described later. Forests require more water than grasslands, and grasslands, in turn, require more water than desert vegetation. The water demands of any specified type of vegetation increase with temperature.

Biomes are geographical regions with climates that favor distinctive combinations of plant and animal species. For example, tundra is the dominant form of vegetation in regions in which the mean temperature of the warmest month is ≤10 °C, and sparse, desert vegetation prevails in regions in which potential evaporation (proportional to the quantity of solar radiation reaching the ground) exceeds precipitation. The global distribution of biomes is determined by the *insolation* (i.e., the incident solar radiation) at the top of the atmosphere and by the climatic variables:

- annual-mean temperature,
- the annual and diurnal temperature ranges,
- annual-mean precipitation, and
- the seasonal distributions of precipitation and cloudiness.

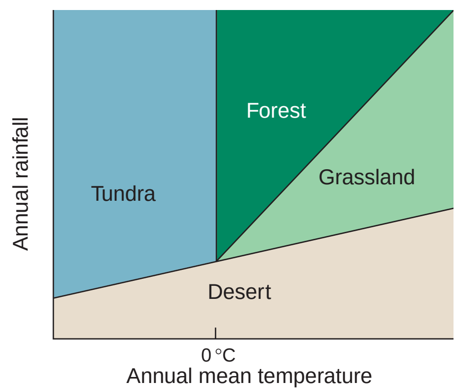

Fig. 2.18 A conceptual framework for understanding how the preferred types of land vegetation over various parts of the globe depend on annual-mean temperature and precipitation.

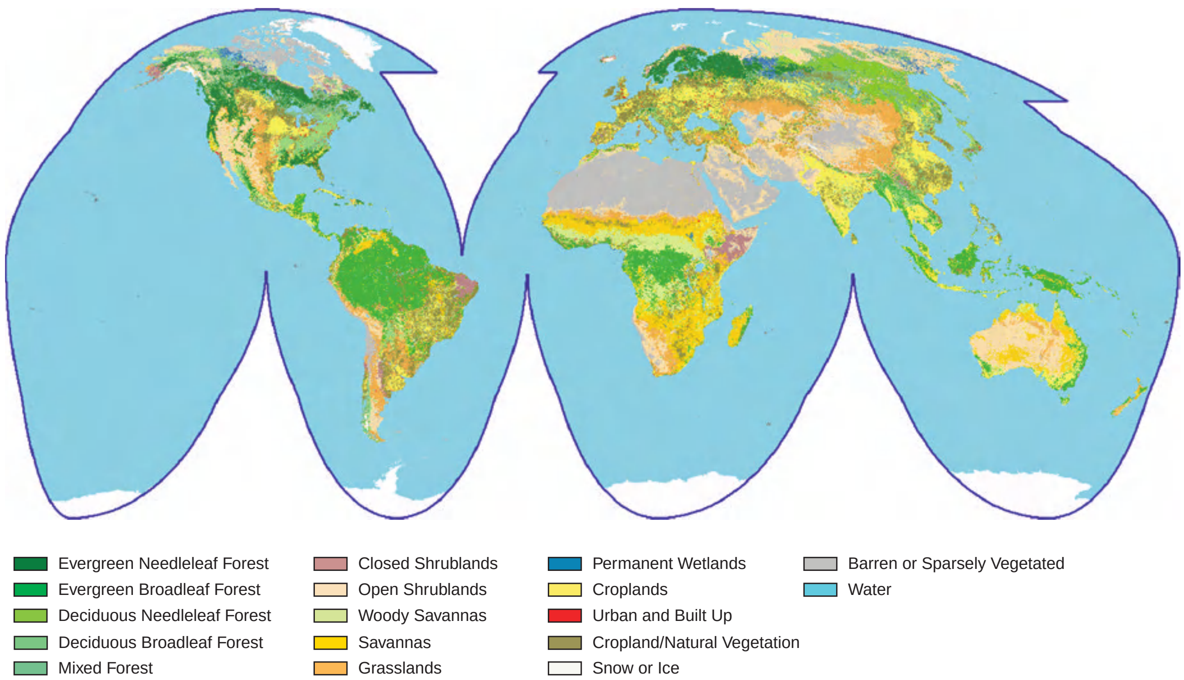

Fig. 2.19 Global land cover characterization, as inferred from NASA AVHRR NDVI satellite imagery and ground-based data relating to ecological regions, soils, vegetation, land use, and land cover. [From USGS Land Processes DAAC.]

Insolation and climate at a given location, in turn, are determined by latitude, altitude, and position with reference to the land–sea configuration and terrain. The combined influence of altitude upon temperature (Fig. 1.9), terrain upon precipitation (Fig. 1.25), and local terrain slope upon the incident solar radiation (Exercise 4.16) gives rise to a variegated distribution of biomes in mountainous regions.

Several different systems exist for assigning biomes, each of which consists of a comprehensive set of criteria that are applied to the climate statistics for each geographical location.[7] The "ground truth" for such classification schemes is the observed distribution of land cover, as inferred from ground-based measurements and high-resolution satellite imagery. An example is shown in Fig. 2.19.

The state of the terrestrial biosphere feeds back upon the climate through its effects on

- the hydrologic cycle: for example, during intervals of hot weather, plants control their temperatures by *evapo-transpiration* (i.e., by giving off water vapor through their leaves or needles). Energy derived from absorbed solar radiation that would otherwise contribute to heating the land surface is used instead to evaporate liquid water extracted from the soil by the roots of the plants. In this manner, the solar energy is transferred to the atmosphere without warming the land surface. Hence, on hot summer days, grass-covered surfaces tend to be cooler than paved surfaces and vegetated regions do not experience as high daily maximum temperatures as deserts and urban areas.
- the local albedo (the fraction of the incident solar radiation that is reflected, without being absorbed): for example, snow-covered tundra is more reflective, and therefore cooler during the daytime, than a snow-covered forest.
- the roughness of the land surface: wind speeds in the lowest few tens of meters above the ground tend to be higher over bare soil and tundra than over forested surfaces.

[7] These systems are elaborations of a scheme developed by Köppen[8] a century ago.

[8] **Wladimir Peter Köppen** (1846–1940) German meteorologist, climatologist, and amateur botanist. His Ph.D. thesis (1870) explored the effect of temperature on plant growth. His climate classification scheme, which introduced the concept of biomes, was published in 1900. For many years, Köppen's work was better known to physical geographers than to atmospheric scientists, but in recent years it is becoming more widely appreciated as a conceptual basis for describing and modeling the interactions between the atmosphere and the terrestrial biosphere.

2.1.4 The Earth's Crust and Mantle

The current configuration of continents, oceans, and mountain ranges is a consequence of plate tectonics and continental drift.[9] The Earth's crust and mantle also take part in chemical transformations that mediate the composition of the atmosphere on timescales of tens to hundreds of millions of years.

The Earth's crust is broken up into plates that float upon the denser and much thicker layer of porous but viscous material that makes up the Earth's mantle. Slow convection within the mantle moves the plates at speeds ranging up to a few centimeters per year (tens of kilometers per million years). Plates that lie above regions of upwelling in the mantle are spreading, whereas plates that lie above regions of downwelling in the mantle are being pushed together. Earthquakes tend to be concentrated along plate boundaries.

Oceanic plates are thinner, but slightly denser than continental plates so that when the two collide, the ocean plate is *subducted* (i.e., drawn under the continental plate) and incorporated into the Earth's mantle, as shown schematically in Fig. 2.20. Rocks in the subducted oceanic crust are subjected to increasingly higher temperatures and pressures as they descend, giving rise to physical and chemical transformations.

Collisions between plate boundaries are often associated with volcanic activity and with the uplift of mountain ranges. The highest of the Earth's mountain ranges, the Himalayas, was created by folding of the Earth's crust following the collision of the Indian and Asian plates, and it is still going on today. The Rockies, Cascades, and Sierra ranges in western North America have been created in a similar manner by the collision of the Pacific and North American plates. These features have all appeared within the past 100 million years.

Oceanic plates are continually being recycled. The Pacific plate is being subducted along much of the extent of its boundaries, while new oceanic crust is being formed along the mid-Atlantic ridge as magma

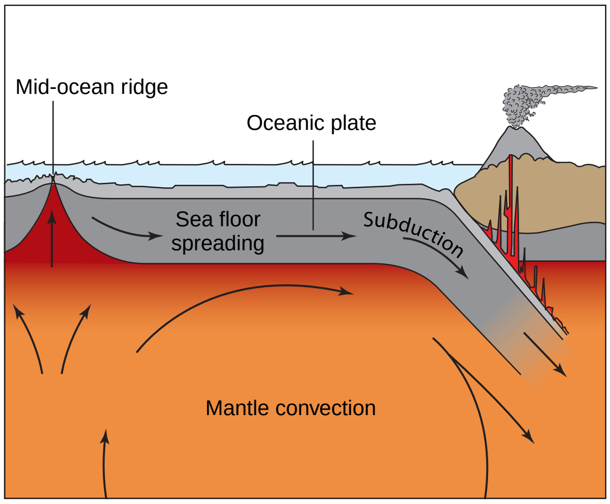

Fig. 2.20 Schematic showing subduction, sea floor spreading, and mountain building. [Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ. Edward J. Tarbuck, Frederick K. Lutgens and Dennis Tasa, Earth: An Introduction to Physical Geology, 8th Edition, © 2005, p. 426, Fig. 14.9.]

upwelling within the mantle rises to the surface, cools, and solidifies. As this newly formed crust diverges away from the mid-Atlantic ridge, the floor of the Atlantic Ocean is spreading, pushing other parts of the crust into the spaces formerly occupied by the subducted portions of the Pacific plate. As the Atlantic widens and the Pacific shrinks, the continents may be viewed as drifting away from the Atlantic sector on trajectories that will, in 100–200 million years, converge over what is now the mid-Pacific. A similar congregation of the continental plates is believed to have occurred about 200 million years ago, when they were clustered around the current position of Africa, forming a supercontinent called *Pangaea* (all Earth).

Some of the material incorporated into the mantle when plates are subducted contains *volatile substances* (i.e., substances that can exist in a gaseous form, such as water in hydrated minerals). As the temperature of these materials rises, pressure builds

[9] The theory of continental drift was first proposed by Alfred Wegener[10] in 1912 on the basis of the similarity between the shapes of coastlines, rock formations, and fossils on the two sides of the Atlantic. Wegener's radical reinterpretation of the processes that shaped the Earth was largely rejected by the geological community and did not become widely accepted until the 1960s, with the advent of geomagnetic evidence of sea-floor spreading.

[10] **Alfred Wegener** (1880–1930). German meteorologist, professor at University of Graz. Began his career at the small University of Marburg. First to propose that ice particles play an important role in the growth of cloud droplets. Set endurance record for time a aloft in a hot air balloon (52 h) in 1906. Played a prominent role in the first expeditions to the interior of Greenland. Died on a relief mission on the Greenland icecap. The Alfred Wegener Institute in Bremerhaven is named in his honor. Son-in-law of Vladimir Köppen and co-authored a book with him.

up beneath the Earth's crust, leading to volcanic eruptions. As will be explained later in this chapter, gases expelled in volcanic eruptions are the source of the Earth's present atmosphere, and they are continually renewing it.

2.1.5 Roles of Various Components of the Earth System in Climate

Atmospheric processes play the lead role in determining such fundamental properties of climate as the disposition of incoming solar radiation, temperatures at the Earth's surface, the spatial distribution of water in the terrestrial biosphere, and the distribution of nutrients in the euphotic zone of the ocean. However, other components of the Earth system are also influential. Were it not for the large storage of heat in the ocean mixed layer and cryosphere during summer, and the extraction of that same heat during the following winter, seasonal variations in temperature over the middle and high latitude continents would be much larger than observed and, were it not for the existence of widespread vegetation, summertime daily maximum temperatures in excess of 40 °C would be commonplace over the continents. The oceanic thermohaline circulation warms the Arctic and coastal regions of Europe by several degrees, while wind-driven upwelling keeps the equatorial eastern Pacific cool enough to render the Galapagos Islands a suitable habitat for penguins!

Plate tectonics shaped the current configuration of continents and topography, which, in turn, shapes many of the distinctive regional features of today's climate. The associated recycling of minerals through the Earth's upper mantle is believed to have played a role in regulating the concentration of atmospheric carbon dioxide, which exerts a strong influence upon the Earth's surface temperature.

These are but a few examples of how climate depends not only on atmospheric processes, but on processes involving other components of the Earth system. As explained in Section 10.3, interactions between the atmosphere and other components of the Earth system give rise to feedbacks that can either amplify or dampen the climatic response to an imposed external forcing of the climate system, such as a change in the luminosity of the sun or human-induced changes in atmospheric composition.

The next three sections of this chapter describe the exchanges and cycling of water, carbon, and oxygen among the various components of the Earth system.

2.2 The Hydrologic Cycle

Life on Earth is critically dependent on the cycling of water back and forth among the various *reservoirs* in the Earth system listed in Table 2.2, which are collectively known as the *hydrosphere*. In discussing the exchanges between the smaller reservoirs, we make use of the concept of *residence time* of a substance within a specified reservoir, defined as the mass in the reservoir divided by the *efflux* (the rate at which the substance exits from the reservoir).[11] Residence time provides an indication of amount of time that a typical molecule spends in the reservoir between visits to other reservoirs. Long residence times are indicative of large reservoirs and/or slow rates of exchange with other reservoirs, and vice versa.

Based on current estimates, the largest reservoir of water in the Earth system is the mantle. The rate at which water is expelled from the mantle in volcanic emissions is estimated to be $\sim 2 \times 10^{-4}$ kg m^{-2} year^{-1} averaged over the Earth's surface, which is the basis for the 10^{11} year residence time in Table 2.2. At this

Table 2.2 Masses of the various reservoirs of water in the Earth system (in 10^3 kg m^{-2}) averaged over the surface of the Earth, and corresponding residence times

Reservoirs of water	Mass	Residence time
Atmosphere	0.01	Days
Fresh water (lakes and rivers)	0.6	Days to years
Fresh water (underground)	15	Up to hundreds of years
Alpine glaciers	0.2	Up to hundreds of years[a]
Greenland ice sheet	5	10,000 years[b]
Antarctic ice sheet	53	100,000 years
Oceans	2,700	
Crust and mantle	20,000	10^{11} years

[a] Estimated by dividing typical ice thicknesses of a large alpine glacier (~300 m) by the annual rate of ice accumulation (~1 m).
[b] Estimated by dividing typical ice thicknesses in the interior of the Greenland ice sheet (2000 m) by the annual rate of ice accumulation (~0.2 m).

[11] The concept of residence time is developed more fully in Chapter 5.1.

rate of exchange, only roughly 5% of the water estimated to reside in the mantle would be expelled over the $\sim 4.5 \times 10^9$-year lifetime of the Earth—not even enough to fill the oceans.

After the mantle and oceans, the next largest reservoir of water in the Earth system is the continental ice sheets, the volumes of which have varied widely on timescales of tens of thousands of years and longer, causing large variations in global sea level.

Exercise 2.2 Based on data provided in Table 2.1, estimate how much the sea level would rise if the entire Greenland ice sheet were to melt.

Solution: The mass of the Greenland ice sheet is equal to its mass per unit area averaged over the surface of the Earth (as listed in Table 2.1) times the area of the Earth or

$$(5 \times 10^3\,\mathrm{kg\,m^{-2}}) \times (5.10 \times 10^{14}\,\mathrm{m^2}) = 2.55 \times 10^{18}\,\mathrm{kg}$$

If the ice cap were to melt, this mass would be distributed uniformly over the ocean-covered area of the Earth's surface. Hence, if x is the sea level rise, we can write

$$\text{(Area of oceans)} \times \text{(density of water)}\, x = \text{mass of ice sheet}$$

$$((5.10 - 1.45) \times 10^{14}\,\mathrm{m^2}) \times (10^3\,\mathrm{kg\,m^{-2}})\, x = 2.55 \times 10^{18}\,\mathrm{kg}$$

Solving, we obtain $x = 7$ m.

Because the masses given in Table 2.2 are expressed in units numerically equivalent to the depth (in m) of a layer covering the entire surface of the Earth, we could have written simply

$$(5.10 - 1.45)\, x = 5.10 \times 5\,\mathrm{m}$$
$$x = 7\,\mathrm{m}$$

■

Of the reservoirs listed in Table 2.2, the atmosphere is by far the smallest and it is the one with the largest rates of exchange with the other components of the Earth system. The residence time of water in the atmosphere, estimated by dividing the mass of water residing in the atmosphere (~ 30 kg m^{-2}, equivalent to a layer of liquid water ~ 3 cm deep) by the mean rainfall rate averaged over the Earth's surface (roughly 1 m per year or 0.3 cm day^{-1}), is ~ 10 days. By virtue of the large exchange rate and the large latent heat of vaporization of water, the cycling of water vapor through the atmospheric branch of the hydrologic cycle is effective in transferring energy from the Earth's surface to the atmosphere.

Averaged over the globe, the rate of precipitation P equals the rate of evaporation E: any appreciable imbalance between these terms would result in a rapid accumulation or depletion of atmospheric water vapor, which is not observed. However, in analyzing the water balance for a limited region, the horizontal transport of water vapor by winds must also be considered. For example, within the region of the ITCZ, $P >> E$: the excess precipitation is derived from an influx of water vapor carried by the converging trade winds shown in Fig. 1.18. Conversely, in the region of the relatively dry, cloud-free subtropical anticyclones, $E >> P$: the excess water vapor is carried away, toward the ITCZ on the equatorward side and toward the midlatitude storm tracks on the poleward side, by the diverging low-level winds. For the continents as a whole, $P > E$: the excess precipitation returns to the sea in rivers. Local evapotranspiration E, as described in Section 2.1.3, accounts for an appreciable fraction of the moisture in summer rainfall P over the continents.

Under steady-state conditions, the mass balance for water vapor over in a column of area A, extending from the Earth's surface to the top of the atmosphere, can be written in the form

$$\overline{E} - \overline{P} = \overline{Tr} \tag{2.1}$$

where overbars denote averages over the area of the column and $\overline{Tr}$ denotes the horizontal transport (or flux) of water vapor out of the column by the winds, as discussed in the previous paragraph. Figure 2.21 shows the distributions of the export of water vapor (i.e., divergence of water vapor transport) by the winds over the low latitude oceans together with the observed distribution of $E - P$. Two aspects of Fig. 2.21 are worthy of note.

1. Apart from the sign reversal, the distribution of $E - P$ in the lower panel resembles the rainfall distribution in Fig. 1.25. That $-P$ and $E - P$ exhibit similar distributions indicates that the horizontal gradients of P must be much stronger than those in E. It follows that the strong observed gradients in climatological-mean rainfall are due to wind patterns rather than to gradients in local evaporation.

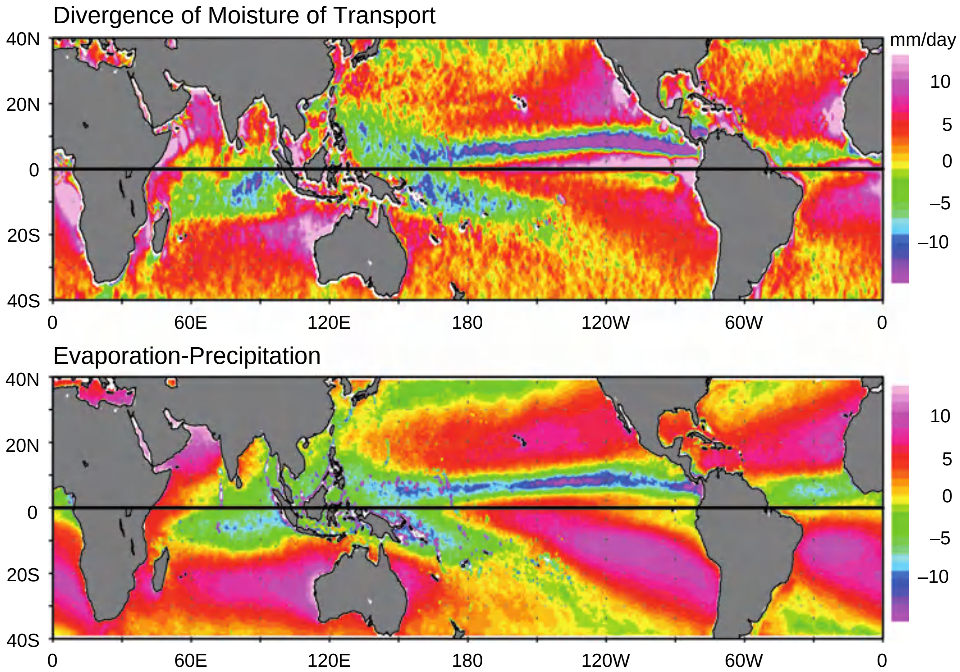

Fig. 2.21 Terms in the annual mean mass balance of atmospheric water vapor in units of mm day^{-1} of liquid water. (Top) The local rate of change of vertically integrated water vapor due to horizontal transport by the winds. (Bottom) Difference between local evaporation and local precipitation. If the estimates were perfect, the maps would be identical. [Based on data from NASA's QuikSCAT and Tropical Rain Measuring Mission (TRMM). Courtesy of W. Timothy Liu and Xiaosu Xie.]

2. In accordance with (2.1), the geographical distributions of $E - P$ and Tr in Fig. 2.21 are similar. The agreement is noteworthy because the measurements used in constructing these two maps are entirely different. The distribution of Tr is constructed from data on winds and atmospheric water vapor concentrations, without reference to evaporation and precipitation.

Also of interest is the time-dependent hydrologic mass balance over land for a layer extending from the land surface downward to the base of the deepest aquifers. In this case

$$\frac{d\overline{St}}{dt} = \overline{P} - \overline{E} - \overline{T} \tag{2.2}$$

where $\overline{St}$ is the area averaged storage of water within some prescribed region and the transport term involves the inflow or outflow of water in rivers and subsurface aquifers. For the special case of a land-locked basin from which there is no inflow or outflow of surface water, the transport term vanishes, and

$$\frac{d\overline{St}}{dt} = \overline{P} - \overline{E} \tag{2.3}$$

Hence the storage of water within the basin, which is reflected in the level of the lake into which the rivers within the basin drain, increases and decreases in response to time variations in $\overline{P} - \overline{E}$.

Figure 2.22 shows how the level of the Great Salt Lake in the reat Basin of the western United States has varied in response to variations in precipitation. From the time of its historic low[12] in 1963 to the time of its high in 1987, the level of the Great Salt Lake rose by 6.65 m, the area of the lake increased by a factor of 3.5, and the volume increased by a factor of 4. The average precipitation during this 14-year interval was heavier than the long-term average, but there were large, year-to-year ups and downs. It is notable that the lake level rose smoothly and

[12] The historical time series dates back to 1847.

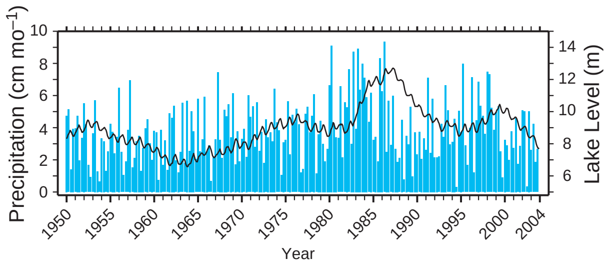

Fig. 2.22 The black curve shows variations in the depth of the Great Salt Lake based on a reference level of 4170 feet above sea level (in m). Depth scale (in m) at right. Blue bars indicate seasonal-mean precipitation at nearby Logan, Utah (in cm month^{-1}). [Lake level data from the U.S. Geological Survey. Courtesy of John D. Horel and Todd P. Mitchell.]

monotonically despite the large year-to-year variations in the precipitation time series. Exercises 2.11–2.13 at the end of this chapter are designed to provide some insight into this behavior.

2.3 The Carbon Cycle

Most of the exchanges between reservoirs in the hydrologic cycle considered in the previous section involve phase changes and transports of a single chemical species, H_2O. In contrast, the cycling of carbon involves chemical transformations. The carbon cycle is of interest from the point of view of climate because it regulates the concentrations of two of the atmosphere's two most important greenhouse gases: carbon dioxide (CO_2) and methane (CH_4).

The important carbon reservoirs in the Earth system are listed in Table 2.3 together with their masses and the residence times, in the same units as in Table 2.2. The atmospheric CO_2 reservoir is intermediate in size between the active biospheric reservoir (green plants, plankton, and the entire food web) and the gigantic reservoirs in the Earth's crust. The exchange rates into and out of the small reservoirs are many orders of magnitude faster than those that involve the large reservoirs. The carbon reservoirs in the Earth's crust have residence times many orders of magnitude longer than the atmospheric reservoirs, reflecting not only their larger sizes, but also the much slower rates at which they exchange carbon with the other components of the Earth system. Figure 2.23 provides an overview of the cycling of carbon between the various carbon reservoirs.

Table 2.3 Major carbon reservoirs in the Earth system and their present capacities in units of kg m^{-2} averaged over the Earth's surface and their residence times[a]

Reservoir	Capacity	Residence time
Atmospheric CO_2	1.6	10 years
Atmospheric CH_4	0.02	9 years
Green part of the biosphere	0.2	Days to seasons
Tree trunks and roots	1.2	Up to centuries
Soils and sediments	3	Decades to millennia
Fossil fuels	10	–
Organic C in sedimentary rocks	20,000	2×10^8 years
Ocean: dissolved CO_2	1.5	12 years
Ocean CO_3^{2-}	2.5	6,500 years
Ocean HCO^-	70	200,000 years
Inorganic C in sedimentary rocks	80,000	10^8 years

[a] Capacities based on data in Fig. 8.3 (p. 150) of Kump, Lee R.; Kasting, James F.; Crane, Robert G., The Earth System, 2nd Edition, © 2004. Adapted by permission of Pearson Education, Inc., Upper Saddle River, NJ.

Exercise 2.3 Carbon inventories are often expressed in terms of gigatons of carbon (Gt C), where the prefix *giga* indicates 10^9 and *t* indicates a metric ton or 10^3 kg. (Gt is equivalent to Pg in cgs units, where the prefix *peta* denotes 10^{15}.) What is the conversion factor between these units and the units used in Table 2.3?

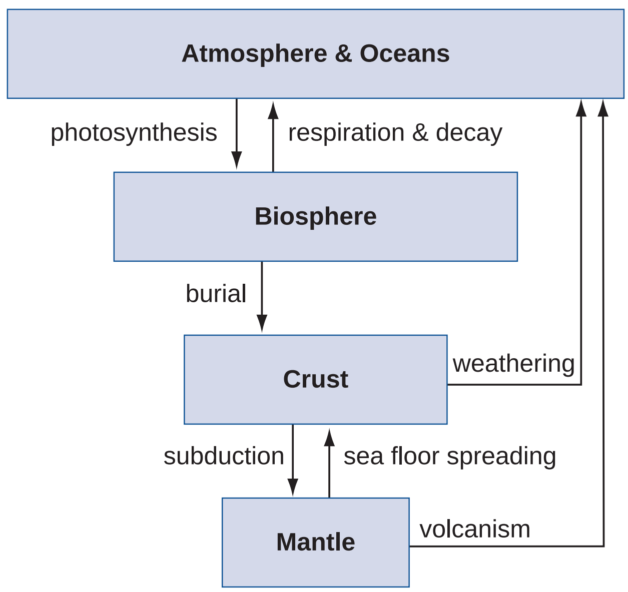

Fig. 2.23 Processes responsible for the cycling of carbon between the various reservoirs in the Earth system.