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Introduction to the Global Environment: The Water and Energy Cycles and Atmospheric and Oceanic Circulation

Introduction

In this book we shall be concerned with the principal constituents of rocks, water, and life as they circulate through the land, the sea, and the air. In other words, our concern will be with the geochemistry of the Earth's surface and how it operates naturally and how it has been perturbed by human activities. The approach is global in scope, and because of their special importance in Earth surface geochemical cycles, water and air will receive major attention. Water moves from the atmosphere to the land surface as rain containing pollutants added to the air by humans. Humans also add other gases and solids to the atmosphere that result in changes in atmospheric composition and climate. The air flows from one place to another, distributing pollutants over wide areas. Water once on the ground reacts with minerals contained in rocks, via a process known as *chemical weathering*, with a change in chemical composition of the water and with soil formation. Plants circulate elements between the atmosphere and soils. Water eventually makes its way into rivers, which obtain dissolved pollutants as well as suspended material derived from human-accelerated erosion. The rivers then deliver their load to the oceans, where a variety of chemical and biological processes occur. Overall, it is these fluxes of water and air that ultimately act to maintain the overall chemical and physical conditions at the Earth's surface.

Throughout most of this book we will concentrate on the principal constituents transported by water and air. These are sodium, potassium, calcium, magnesium, silicon, carbon, nitrogen, sulfur, phosphorus, chlorine, and, of course, hydrogen and oxygen. We shall point out how the global cycles of some of these elements have been perturbed by humans, resulting in such things as greenhouse gases, acid rain, eutrophic lakes, and oceanic acidification. Although they are of geochemical and environmental interest, we shall not be concerned with minor and trace elements (e.g., lead and mercury) or exotic synthetic chemicals (e.g., pesticides), since our goal is not all-inclusive environmental coverage. (The interested reader is referred to books such as those by Laws 2000 and Mackenzie 2011 for detailed discussion of environmental problems.) Rather, the approach is that of geochemists trying to understand how global chemical cycles operate and how they affect the major constituents of rocks, water, air, and life.

Because of their importance to climate and chemical changes in the global environment,emphasis in this chapter will be placed on the atmosphere and the oceans, and how their circulation operates; a discussion of the water and energy cycles of the Earth and their role in meteorology and oceanography will be included. The chapter, then, helps to set the stage for discussions of such subjects as the atmospheric greenhouse effect (chapter 2), acid rain (chapter 3), and chemistry of the oceans (chapter 8).

The Global Water Cycle

Major Water Masses

Earth is the only planet in the solar system having an abundance of liquid water on its surface; about 70% of the Earth is covered by liquid water. Because of the particular combinations of temperature and pressure on the planet's surface, water can exist here in three states: as liquid water, as ice, and as water vapor. This is in sharp contrast to the surface of the planet Mars, for example, which is so cold and dry that water can exist there only as ice or as water vapor.

Water is by far the most abundant substance at the Earth's surface. There are 1444×10^6 km^3 of it in its three phases: liquid water, ice, and water vapor. As shown in table 1.1, most of the Earth's water (97%) is stored as seawater in the oceans. The remaining 3% is either on the continents or in the atmosphere. The amount of water in the atmosphere, in the form of water vapor, is very small in comparison with the other reservoirs, only around 0.001% of the total. However, it plays a very important role in the water cycle, as we shall see.

Of the fresh water stored on the continents, around two-thirds is in the form of ice in ice sheets, polar ice caps, and glaciers. Most of the rest of the continental water is present either as subsurface groundwater or in lakes and rivers. It is this small part of the Earth's total water (1%) that humans draw on for their water supplies. Here we shall focus on water near the Earth's surface and how it moves within and between the various reservoirs.

Fluxes between Reservoirs

Water does not remain in any one reservoir but is continually moving from one place to another in the hydrologic cycle. This is illustrated in figure 1.1. (For a more detailed discussion, see chapter 5 for runoff; also Penman 1970,

Table 1.1 Inventory of Water at the Earth's Surface

Reservoir	Volume $10^6 km^3$ (10^{18}kg)	Percent of Total
Oceans	1400	96.95
Mixed layer	50	
Thermocline	460	
Abyssal	890	
Ice shelves (floating)[a]	0.7	0.048
Ice caps and glaciers[b]	0.09	0.006
Ice sheets[a]	27.6	1.9
Greenland	2.9	
Antarctica	24.7	
Groundwater	15.3	1.06
Lakes	0.125	0.009
Rivers	0.0017	0.0001
Soil moisture	0.065	0.0045
Atmosphere total[b]	0.0155	0.001
Terrestrial	0.0045	
Oceanic	0.0110	
Biosphere	0.002	0.0001
Approximate Total	1444	

Source: NRC 1986; Berner and Berner 1987; Lemke et al. 2007.
[a] Lemke et al. 2007.
[b] As liquid volume equivalent of water vapor.

Baumgartner and Reichel 1975, NRC 1986, and Chahine 1992). Water is evaporated from the oceans and the land into the atmosphere, where it remains for only a short time, on the average about eleven days, before failing back to the surface as snow or rain. Part of the water falling onto the continents runs off in rivers and, in some places, accumulates temporarily in lakes. Some also passes underground only to emerge later in rivers, lakes, and the ocean. The remaining portion of the precipitation on the continents is returned directly to the atmosphere via evaporation. Over the oceans, evaporation exceeds precipitation, with the difference being made up by input via runoff from the continents. An idea of the sizes of these various fluxes of water (mass transported per unit time) is shown in figure 1.1.

In order to conserve total water, evaporation must balance precipitation for the Earth as a whole, since the total mass of water at the Earth's surface is believed to be constant over time. The average global precipitation rate, which is equal to the evaporation rate, is 0.506×10^6 km^3/yr. For any one portion of the Earth, by contrast, evaporation and precipitation generally do not balance. On the land, or continental part of the Earth, the precipitation rate (0.108×10^6 km^3/yr) exceeds the evaporation rate (0.071×10^6 km^3/yr), whereas over the oceans evaporation (0.435×10^6 km^3/yr) dominates over precipitation (0.398×10^6 km^3/yr). The difference in each case (0.037×10^6 km^3/yr) comprises water transported from the oceans to the continents as atmospheric water vapor, or that returned to the oceans as river runoff (see fig. 1.1). Although inaccurately

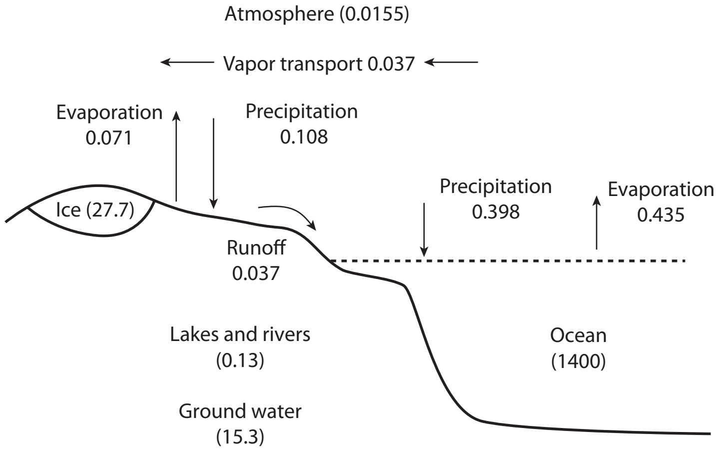

Figure 1.1 The hydrologic cycle. Numbers in parentheses represent inventories (in 10^6 km^3 = 10^{18} kg) for each reservoir. Fluxes are in 10^6 km^3/yr per year (10^{18} kg/yr).
Data from table 1.1, NRC 1986, and chapter 5.

known, there is a considerably smaller amount of direct groundwater discharge to the oceans (0.0022×10^6 km^3/yr)(Korzun et al. 1977).

The values given in figure 1.1 have built-in errors. Precipitation is difficult to measure, rainfall measurements over the oceans are few, and oceanic evaporation is necessarily estimated from models. Also, because the evaporation rates over many land areas of the Earth have not been measured, the values given in figure 1.1 are based, by necessity, on the difference between measured worldwide precipitation on land and river runoff values. (Recently, attempts have been made to obtain evaporation rates by other means, including atmospheric water vapor balance and heat balance; see the section Radiation and Energy Balance below.)

Assumption of constant volume of water in a given reservoir (water mass) enables the use of the concept of residence time. The residence time is defined as the volume of water in a reservoir divided by the rate of addition (or loss) of water to (from) it. It can be thought of as the average time a water molecule spends in a given reservoir. For the oceans, the volume of water present ($1{,}400 \times 10^6$ km^3; see fig. 1.1) divided by the rate of river runoff to the oceans (0.037×10^6 km^3/yr) gives a residence time of 38,000 years. This long residence time, which can also be thought of as a filling or replacement time, reflects the very large volume of water in the oceans. By contrast, the residence time of water in the atmosphere, relative to evaporation from both the oceans and the continents, is only about eleven days. Lakes, rivers, glaciers, and shallow groundwater have residence times lying between these two extremes, but because of extreme variability, no simple average residence time can be given for each of these reservoirs.

Geographic Variations in Precipitation and Evaporation

The values shown in figure 1.1 are only average values for precipitation and evaporation over the continents and oceans. From one region to another there

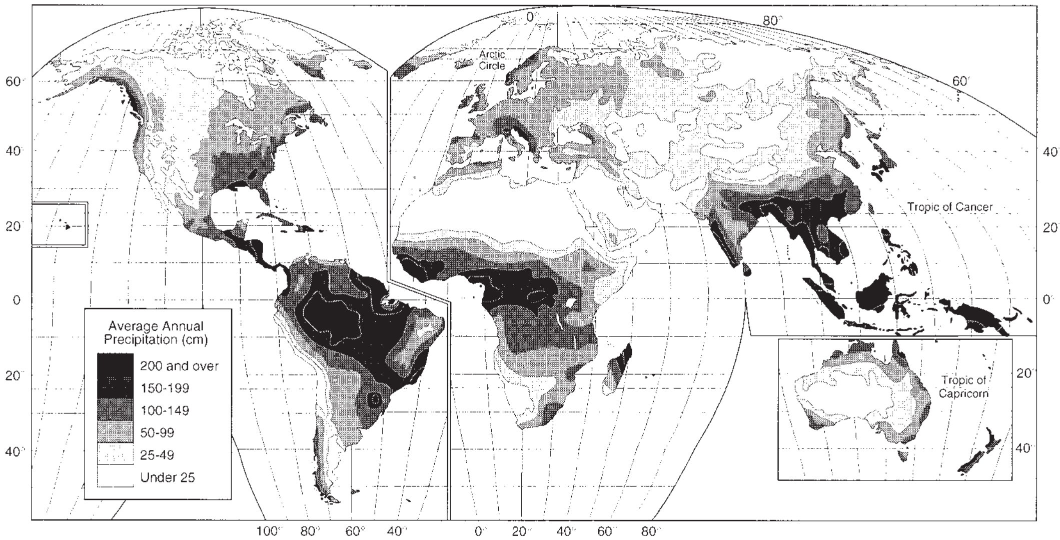

Figure 1.2 Global average annual precipitation.
After McKnight 1996.

is considerable variation, as can be seen in figure 1.2, which shows the mean annual precipitation for different areas of the continents.

In order to have rain or snow, there must be both sufficient water vapor in the atmosphere and rising air that can carry the water vapor up to a height where it is cold enough for condensation and precipitation to occur (see chapter 3). Net precipitation (precipitation minus evaporation), as shown in figure 1.3, is highest near the equator (10°N to 10°S) and at 35° to 60° north and south latitudes, where there is frequent storm activity with its accompanying air motion. Net precipitation is lowest in the subtropics (15° to 30° N and S), where the air is stable, and near the poles, which have both stable air and a very low moisture content due to low temperatures. (However, since there is also very low evaporation near the poles, precipitation can exceed evaporation in certain places, resulting in the formation of the ice caps of Greenland and Antarctica.) In continental areas of high rainfall, runoff is also high. Examples are the large rivers of equatorial regions (Amazon, Zaire, Orinoco) and those of middle latitudes such as the Mississippi and Hwanghe (Yellow).

Evaporation also varies considerably over the Earth's surface. For net evaporation to occur, there must be a heat source (i.e., radiation from the sun), a low moisture content in the air, and the presence of water available for evaporation. In arid regions the evaporation rate is high but is limited by the availability of water. Overall, as shown in figure 1.3, evaporation exceeds precipitation at subtropical latitudes (15° to 30° N and S). Over the continents this leads to the formation of large deserts at these latitudes (for example, the Sahara Desert in Africa and the Great Desert of Australia). The greatest evaporation rates on the Earth (more than 200 cm/yr), however, occur over the subtropical oceans, such as over the Gulf Stream, in winter where warm water, which is carried northward, encounters cooler, drier air and evaporates. High evaporation rates can

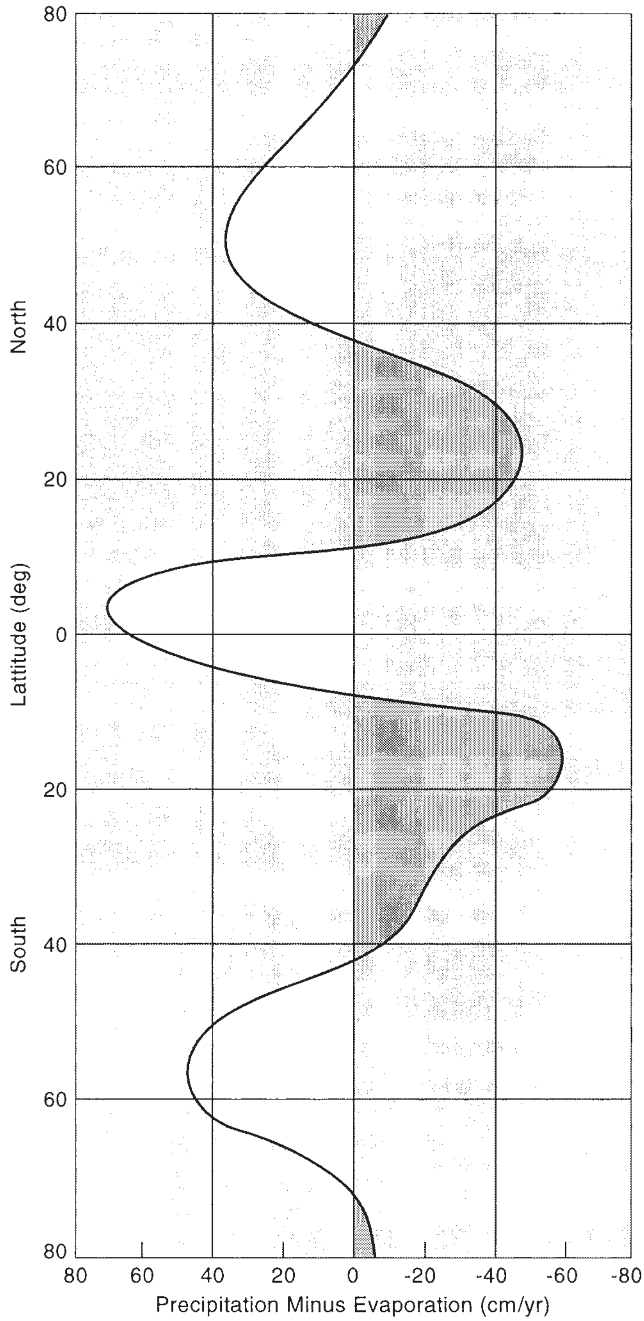

Figure 1.3 Net precipitation (precipitation minus evaporation) as a function of latitude. Positive values represent net precipitation while negative values represent net evaporation.

After Peixoto and Kettani 1973.

lead to locally higher ocean salinities due to the removal of almost pure water during evaporation, leaving dissolved sea salt behind. An outstanding example of this is the Mediterranean Sea, which has a salinity notably higher than that of the open ocean (e,g., see chapter 7).

The Energy Cycle

Introduction

In this section we shall take a look at the energy cycle of the Earth. This is what drives the water cycle and the circulation of the atmosphere and oceans. Atmospheric water vapor remains in the atmosphere for only about eleven days on the average. However, during this time it has travelled a mean distance of 1000 km (Peixoto and Kettani 1973), and this transport is controlled by the energy cycle of the Earth. The energy cycle, in turn, is greatly influenced by the presence of water vapor in the atmosphere. Thus, the Earth's energy and water cycles are intimately interconnected, and they exert strong influences on one another. (For more information on the Earth's energy cycle, see Ingersoll 1983; Ramanathan 1987; and Trenberth et al. 2009).

Radiation and Energy Balance

The primary energy sources for the Earth's surface are summarized in table 1.2. As can be seen, radiation from the sun is by far the most important source of energy (99.98%), and consequently it is the dominant influence on the circulation of the atmosphere and oceans. It is the only energy source that will be discussed here. The incoming solar radiation impinges on the top of the atmosphere, below which it is reflected, absorbed, and converted into other forms of energy. Apportionment of this energy into different forms is described in terms of the radiation or energy balance of the Earth (fig. 1.4).

As shown by the flux values near the top of figure 1.4, 341 Wm^{-2} of incoming, mainly short-wave (<4 μm), solar radiation at the top of the atmosphere is balanced by reflection of this radiation (102 Wm^{-2}) along with outgoing long-wave (>4 μm) energy flux from the Earth (239 Wm^{-2}). This must be true

Table 1.2 Primary Energy Sources for the Earth

Source	Energy Flux ($cal/cm^2/min$)	Percent of Total Energy Flux
Solar radiation	0.5[a]	99.98
Heat flow from interior of earth	0.9×10^{-4}	0.018
Tidal energy	0.9×10^{-5}	0.002

Source: Hubbert 1971; Flohn 1977.

[a] 0.5 $cal/cm^2/min$ is approximately equal to 341 $Watts/m^2$, the solar flux used in Fig.1.4, after Trenberth et al. 2009; (1 Watt = 0.2389 cal/sec).

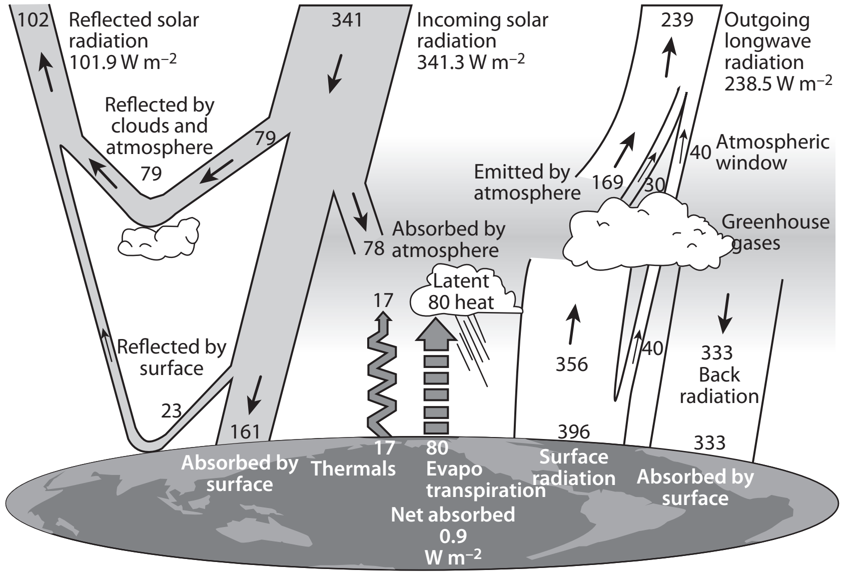

Figure 1.4 The mean annual radiation and heat balance of the atmosphere and Earth for the period March 2000 to May 2004. Units are in watts per square meter. Short-wave solar radiation is that with <4 μm wavelength and is shown by the arrows on the left portion of the figure; long-wave Earth radiation is >4 μm and is shown on the right side of the diagram.

After K. E. Trenberth et al. 2009; © Copyright 2009 American Meteorological Society.

to avoid rapid heating up or cooling of the earth and has been verified by satellite measurements (Ramanathan 1987). This does not mean, however, that very slight imbalances over extended periods cannot bring about cooling or heating. An example of this is the large climate changes that occurred at northern midlatitudes during the past 20,000 years, as evidenced by extensive glaciation and deglaciation that occurred during this time.

The 102 Wm^{-2} of reflected short-wave solar energy (fig. 1.4) represents about 30% of the incoming flux and is known as the Earth's *albedo*. It is a measure of how bright the Earth would appear if viewed from outer space. Note that 79 Wm^{-2} of this reflected radiation is by clouds and the atmosphere and does not reach the Earth's surface. At the surface just 23 Wm^{-2} is reflected back to space. Thus, changes in cloud cover are very important to radiative heating of the surface. The remaining 70% of the incoming solar radiation represents absorption and reradiation as long-wave (>4 μm) infrared radiation by the Earth's surface and various components of the atmosphere.

Because the sun is very hot (6000°C), most of the solar radiation is in shorter wavelengths (less than 4 μm), with the peak radiation in the visible wavelengths (0.4–0.7 μm). Much of this visible radiation, or light, penetrates the atmosphere and ultimately reaches the ground. This is important because life is dependent on the absorption of light by photosynthetic organisms. By contrast, almost all of the ultraviolet solar radiation (<0.4 μm) is absorbed in the upper atmosphere by ozone and 0_2, which protects life from its harmful

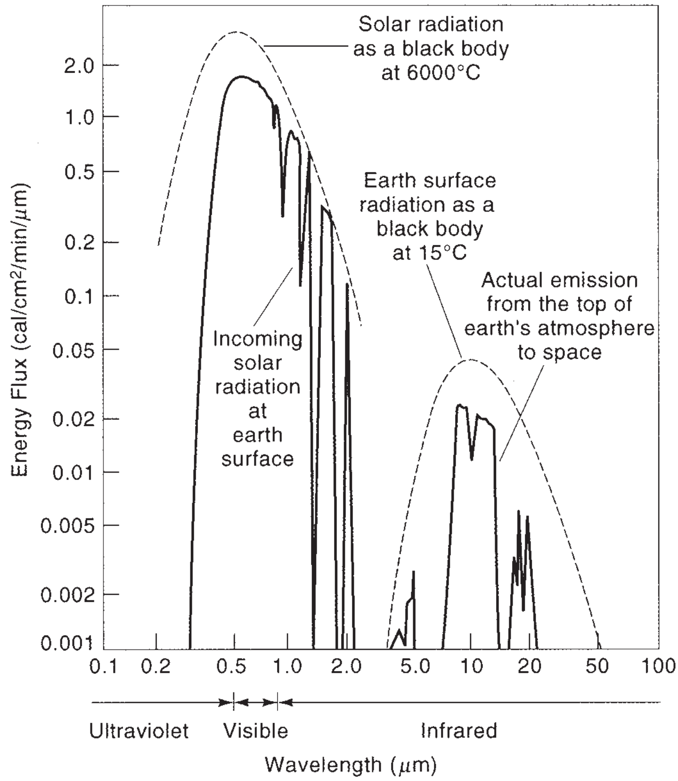

Figure 1.5 Incoming radiation from the sun at the earth's surface and outgoing radiation from the top of the earth's atmosphere as a function of wavelength. Also shown are the energy spectra expected for blackbodies at the same temperature as the sun and the Earth. (A blackbody emits the maximum radiation for its temperature.)
Adapted from Sellers 1965.

effects. (Because of the importance of UV absorption by ozone there is growing concern about anthropogenic ozone depletion—see chapter 2.) The sun's incoming long-wave, or infrared, radiation is also absorbed at certain wavelengths by atmospheric water vapor and CO_2 and by water droplets in clouds. Thus, the incoming solar radiation reaching the Earth's surface has gaps at those wavelengths where atmospheric absorption occurs (see fig. 1.5). Overall, due to the absorption of 78 Wm^{-2} by atmospheric ozone, water vapor, CO_2, and clouds combined with reflection of 102 Wm^{-2}, only 161 Wm^{-2} or 47% of the incoming solar radiation actually reaches the Earth's surface (fig. 1.4) to be absorbed there (Trenberth et al. 2009).

The surface of the Earth reradiates part of its absorbed solar energy back into space and the atmosphere, but since it is cooler than the sun (the average Earth surface temperature is 15°C [Ramanathan 1987]), the Earth radiates at longer wavelengths (>4 μm) with a maximum in the infrared at about 10 μm (see fig. 1.5). Atmospheric water vapor, carbon dioxide, and rarer gases such as methane are good absorbers of infrared energy in this range of wavelengths. Because of this, most of the infrared radiation originating from the Earth's surface is absorbed by these gases (mostly by water vapor), and very little of it escapes directly to space. As a result the spectrum of outgoing radiation leaving the Earth differs considerably from that expected for a black body at 15°C (see fig. 1.5).

Of the 396 Wm^{-2} of long-wave (infrared) radiation emitted by the Earth's surface, 333 Wm^{-2} are both absorbed by atmospheric water vapor and other gases and reradiated back to the ground. Here it is reabsorbed, keeping the Earth warm. Clouds also contribute to warming of the Earth by reducing long-wave emissions to space because at their bases they absorb radiation emitted by the warmer Earth surface and at their tops they emit to space at colder temperatures. However, as discussed above, clouds also reflect solar radiation, and because this process outweighs their warming effect, the net effect of clouds is global cooling (Meehl et al. 2007; see also chapter 2).

The role of atmospheric water vapor and carbon dioxide in allowing the incoming short-wave solar radiation to pass through to the ground, while absorbing and reradiating to the Earth's surface most of the Earth's outgoing long-wave radiation, is referred to as the atmospheric *greenhouse effect* by comparison to the glass in a greenhouse. A greenhouse lets solar radiation in but keeps the greenhouse warm by preventing long-wave radiation from leaving because of absorption of the long-wave radiation by the glass. The greenhouse effect makes the Earth's surface much warmer (around 30°C warmer, according to IPCC 2007) than it would be otherwise. Lately, there has been much concern about the atmospheric buildup of carbon dioxide, released from fossil fuel burning, and other greenhouse gases in that they may absorb more than the normal amount of the Earth's outgoing radiation, resulting in an enhanced greenhouse effect and global warming (see chapter 2).

In order to maintain a constant Earth surface temperature, the amount of incoming solar radiation received at the surface (161 Wm^{-2}) should be balanced by the net loss of long-wave radiation from the surface to the atmosphere and space (396 – 333 = 63 Wm^{-2}) plus the fluxes of sensible heat via thermal updrafts (17 Wm^{-2}) and latent heat from evapotranspiration (80 Wm^{-2}). However, figure 1.4 shows that the sum of the three upward fluxes is less, by about 1 Wm^{-2}, than that of the incoming solar radiation. This represents the estimated warming of the Earth, mainly by human activities.

When liquid water is evaporated to form atmospheric water vapor, heat (energy) is absorbed. This is what is called *latent heat*, since upon subsequent condensation of the water vapor into rain and snow, the previously added energy is released as heat to the atmosphere. Since condensation can occur at great distances from the original site of evaporation, the transport of water vapor in the atmosphere also involves the transport of heat. Of the 161 Wm^{-2} of solar energy absorbed at the Earth's surface, 80 Wm^{-2} are used to evaporate water, giving rise to an equivalent latent heat flux from the land and oceans to the atmosphere (see fig. 1.4). It is this latent heat flux that drives the water cycle.

Knowledge of the latent heat flux to the atmosphere can be used to calculate the global rate of evaporation. Since it takes 2.46×10^9 joules to convert 1 m^3 of liquid water to water vapor at the average Earth surface temperature of 15°C, the rate of evaporation corresponding to the dissipation of 80 Wm^{-2} (fig. 1.4) as latent heat over the Earth's surface (510×10^6 km^2, as follows (after

Miller et al. 1983 and Budyko and Kondratiev 1964, updated to include accepted units and the energy data of Trenburth et al. 2009):

$$[(80 \text{ joule sec}^{-1} \text{ m}^{-2}) \times (510 \times 10^{12} \text{ m}^2) \times 3.15 \times 10^7 \text{ sec yr}^{-1}]/2.46 \times 10^9 \text{ joule m}^{-3} = 5.22 \times 10^{14} \text{ m}^3/\text{yr}$$

This is equivalent to 522,000 km^3 of water per year, which agrees fairly well with estimates of total annual evaporation from the Earth (506,000 km^3 from fig. 1.1) based on setting the total evaporation equal to measured total precipitation in the global water balance.

Variations in Solar Radiation: The Atmospheric and Oceanic Heat Engine

The amount of solar radiation absorbed by the Earth decreases with latitude from the equator to the poles. It is this variation in the Earth's heating that drives the circulation of the ocean and atmosphere and, thus, much of the hydrologic cycle.

Latitudinal variations in the input of solar energy are due to two factors. First, the Earth is a sphere and the angle at which the sun's rays hit its surface varies from 90° (or vertical) near the equator to 0° (or horizontal) near the poles. This is shown in figure 1.6. Less energy is received at the poles because the same amount of radiation is spread out over a much larger area at high latitudes (compare situation C with situations A and B in fig. 1.6) and because at high latitudes the sun's rays must travel through a much greater thickness of atmosphere where more absorption and reflection occur.

The second factor affecting latitudinal variations in heating is the duration of daylight. Because the polar axis of the Earth is tilted at an angle of 23.5° with respect to the ecliptic (the plane of the Earth's orbit about the sun), we have a progression of seasons where the angle of the sun's rays striking any given point varies over the year. This is shown in figure 1.7.

In the Northern Hemisphere winter no sunlight strikes the area around the North Pole, during a full day's rotation of the Earth, because it is in the Earth's shadow. Thus, little or no solar heating occurs in this area at this time. Conversely, at the South Pole there is continual daylight, but at a very low sun angle, during the Northern Hemisphere winter. As the seasons shift, the South Polar region eventually becomes plunged into twenty-four-hour darkness (during the Southern Hemisphere winter and Northern Hemisphere summer), just as the North Pole had been earlier. Low-latitude regions near the equator, by contrast, undergo little seasonal change in the duration of daylight, whereas intermediate latitudes are subjected to changes intermediate between those of the poles and the equator. Thus, because of seasonality, more annual radiation is received per unit area at lower, as compared to higher, latitudes.

The effects of variation in angle of the sun's rays with latitude and seasonal changes in the amounts of daylight result in strong variation with latitude in the total solar radiation received during a year, and this result applies equally to both the Northern and Southern Hemispheres. However, because

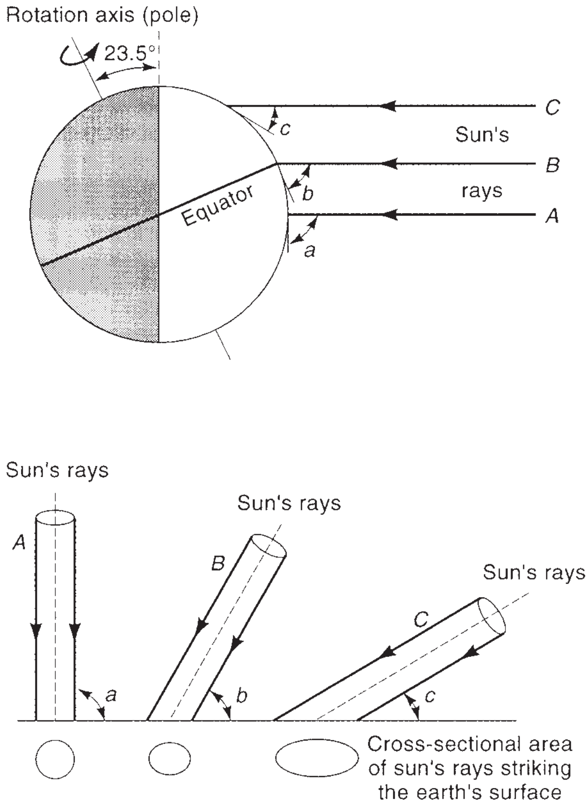

Figure 1.6 Schematic diagrams showing the variations of solar intensity (energy per unit area) with angle of incidence to the Earth's surface. Lower angles (higher latitudes) result in the same energy spread out over a larger area and, thus, in a lower intensity of radiation. Scene depicted is for Northern Hemisphere winter.

Adapted from Miller et al. 1983.

the Northern Hemisphere is dominated by land and the Southern Hemisphere by water, one might expect hemispheric differences in received radiation due to differences in the albedo of land vs water. Nevertheless, according to satellite measurements, the annual mean albedo (relection of solar radiation) of both hemispheres is nearly the same. This is because of the dominant influence of clouds (over surface effects) in determining the mean albedo of each hemisphere (Ramanathan 1987).

The long-wave radiation leaving the Earth also varies with latitude but less strongly. The differences between the amount of solar radiation received and long-wave radiation emitted results in radiation imbalances over the surface of the Earth (fig. 1.8). From 35° north and south latitudes to the poles (mid- and polar latitudes) there is a net deficit of radiation (more leaves than enters), whereas from 35° to the equator (tropical and subtropical latitudes) there is a net surplus (more solar radiation enters than the Earth radiates back). To keep the poles from getting colder and the tropics from getting warmer, heat must be transported from lower to higher latitudes. This is accomplished by the

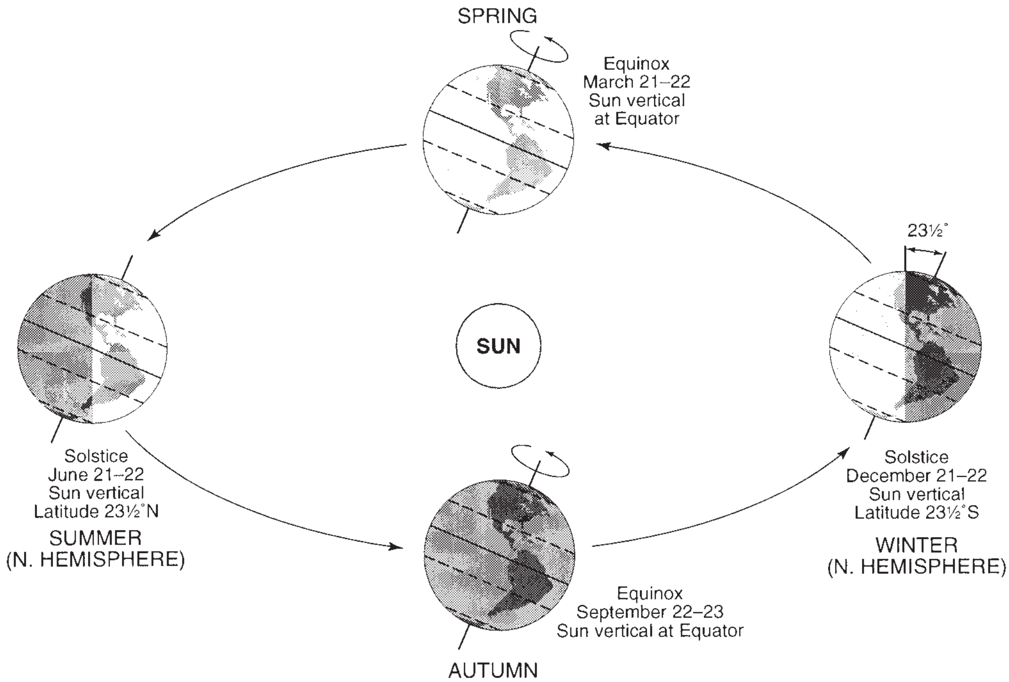

Figure 1.7 Revolution of the earth in its orbit around the sun, showing the changing seasons (also length of daylight). The seasons given are for the Northern Hemisphere; they are reversed in the Southern Hemisphere.

Modified from Lutgens and Tarbuck 1992, fig 2–3, p. 29.

circulation of the atmosphere and oceans. Thus, the atmosphere and oceans act like a "heat engine" driven by latitudinal variations in solar radiation. By contrast there is no transfer of heat across the equator because both hemispheres have similar zoned energy balances (Ramanathan 1987).

Heat is transported from the equator to the poles in three ways: (1) by ocean currents carrying warm water, (2) by atmospheric circulation (wind) carrying warm air, and (3) by atmospheric circulation carrying latent heat in the form of water vapor. The maximum meridional heat transport occurs around 30° latitude (Ramanathan 1987). In the northern hemisphere, the heat transport by the oceans and the atmosphere are roughly comparable in size but the exact fluxes are not well known (Chahine 1992).

Warm, wind-driven ocean currents transport heat poleward from the zone between 20°N and 20°S; they are well known, examples being the Gulf Stream in the North Atlantic and the Kuroshio Current in the North Pacific. The poleward transport of warm water by the oceans tends to warm the overlying atmosphere at higher latitudes, especially during the winter. This provides a moderating influence on climate and helps to explain the relatively mild winters experienced, for example, by western Europe (Gulf Stream) and by Japan (Kuroshio Current).

The atmosphere carries heat poleward as warm air and latent heat. A major source of warm air is the tropical region between 10°N and 10°S, which is the

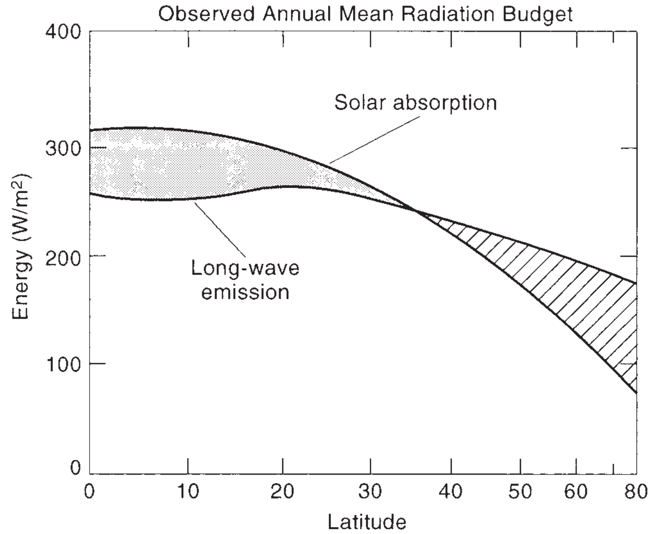

Figure 1.8 Annual zoned mean estimates for both hemispheres, which are nearly the same, of absorbed solar radiation and outgoing long-wave radiation emission obtained by satellites. Shaded regions denote net heating and dashed region denotes net cooling.

After Ramanathan 1987, fig. 3, p. 4076, based on data from Ellis and Vonder Haar 1976.

zone of maximum surplus of solar radiation. The tropical air is heated both by sensible heat and by the release of latent heat upon condensation of moisture. (The hot tropics are a zone of both high evaporation and high rainfall, with the latter predominating). Much additional latent heat in the form of water vapor is injected into the atmosphere in the subtropic zones (15°–30° N and S) where evaporation exceeds precipitation. From there, poleward transport of the latent heat takes place. The latent heat is subsequently released by condensation, which warms the atmosphere in the midlatitude zones of intense storm activity at 30°–50° north and south. Latent heat from condensation in clouds provides 30% of the heat energy that is carried by the atmospheric circulation (Chahine 1992).

A small part (about 0.7%) of the incoming solar radiation is converted into the energy of motion (kinetic energy) of ocean currents, winds, and waves. Although this is a small number, it represents an energy of major interest to the hydrologic cycle, that associated with the circulation of the atmosphere and oceans. This circulation will be discussed next.

Circulation of the Atmosphere

The atmosphere circulates as a consequence of the latitudinal heat imbalance discussed above. If the circulation were due solely to heating, hot air would rise at the equator and flow poleward at high levels. As the air was cooled in transit and piled up at the poles, it would tend to sink at the poles. To complete the

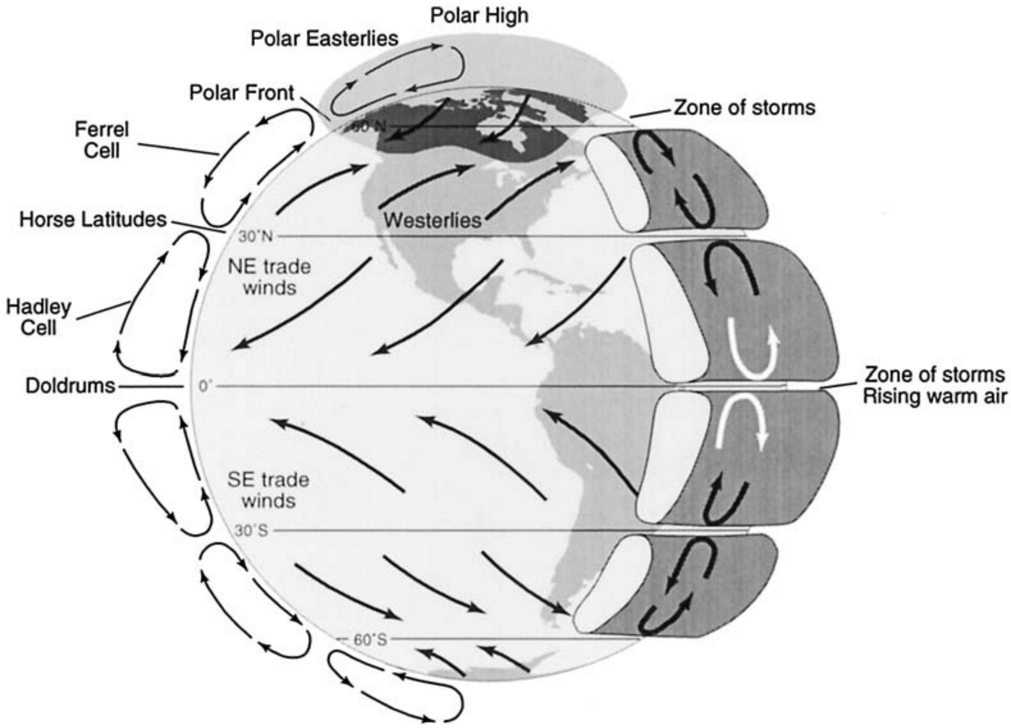

Figure 1.9 Schematic representation of the general circulation of the atmosphere. *Modified from Lutgens and Tarbuck 1992, fig. 8–3, p. 170.*

cycle, there would be a return flow near the Earth's surface of cool air toward the equator, where heating would produce two symmetrical closed circuits, or cells, one in the Northern and one in the Southern Hemisphere. Such a circulation (incorporating the Earth's rotation) was originally proposed in 1735 by George Hadley, but because of a number of factors, the actual circulation has turned out to be considerably more complicated.

The general circulation of the atmosphere, showing the mean annual winds, is depicted in figure 1.9. Note that the winds do not simply blow along north-south, or meridional, lines. This is because they are deflected by the rotation of the Earth. The force that deflects moving objects—in this case, moving air masses to the right in the Northern Hemisphere and to the left in the Southern Hemisphere—is referred to as the *Coriolis force*. The general circulation differs from the simple Hadley circulation by being broken up into several latitudinal zones; however, there is still symmetry more or less between the Northern and Southern Hemispheres. For the sake of brevity, we will discuss only the Northern Hemisphere, but what is said applies equally to the Southern Hemisphere (with a leftward- instead of rightward-directed Coriolis force). (For details of circulation not discussed here, the interested reader is referred to books on meteorology and climatology such as Lutgens et al. 2009 and Barry and Chorley 1998.)

Hot moist air rises at the equator, and as it rises the moisture condenses and intense precipitation results, releasing latent heat. Here there are only very weak surface winds, giving rise to the equatorial doldrums. After rising, the air flows northward at high levels, cools, and eventually sinks around 30°N. The descending air is very dry (having lost most of its moisture in the tropics), and when it is warmed, its capacity to take up moisture is further increased. The resulting hot dry air causes intense evaporation at the Earth's surface, and this gives rise to the subtropical belt of deserts centered between 15° and 30°N. After reaching the surface, the air flows southward, picking up moisture as it flows over the ocean and being deflected to the right by the Coriolis force. This surface flow is known as the northeast trade winds. Upon reaching the equator the northeast trades converge with the southeast trades from the Southern Hemisphere, in the Intertropical Convergence Zone (ITC), and the air rises at the equator to complete the low-latitude cycle known as the Hadley cell (which behaves rather as Hadley expected the whole atmospheric circulation to behave).

At around 30°N, additional air descends and then flows north at the surface rather than south. This is the beginning of the Ferrel cell. The northward flowing air is deflected to the right, forming the prevailing westerlies, which flow from southwest to northeast in the Northern Hemisphere. The westerlies continue until they encounter a cold mass of air moving south from the North Pole at about 50°N. This zone where the air masses meet is known as the polar front, and it is a region of unstable air, storm activity, and abundant precipitation. The polar jet stream (a very fast air stream) occurs at this boundary. The warmer air from the south rises over the polar air and then turns south at high altitude to complete the Ferrel cell. Meanwhile the southward flowing polar air (polar easterlies) becomes warmed by condensation at the polar front and by contact with the southern air. As a result, it too rises and then flows northward at high altitude to the pole, where it sinks, thus completing the polar cell.

In the midlatitudes, the west-to-east flow (Northern Hemisphere westerlies) is subject to considerable turbulence because of the Earth's rotation. At higher levels of the atmosphere, the flow forms waves called *planetary waves*) that transport warm air from the surface to the top of the atmosphere (Ingersoll 1983). These waves are expressed, in the lower atmosphere, in a series of storms that travel west to east around the globe, transporting warm air poleward and cool air equatorward and releasing heat by precipitation.

Oceanic Circulation

Introduction

The oceans can be divided into two portions for the purpose of discussing circulation. The top 50–300 m, or surface layer (fig. 1.10), is stirred by the wind and is well mixed from top to bottom. Below the surface layer (also referred to as the *mixed* layer), the remaining deeper water is colder, less well mixed, and divided into a number of roughly horizontal layers of increasing density. The deep water is separated from the surface water by a region of steeply decreasing temperature gradient, known as the *thermocline*, about 1 km thick (fig. 1.10), across which there is limited communication between the surface and deep water. The deep part of the ocean below the thermocline is referred to as the *abyssal zone*. The volume of the surface (mixed) layer is small, comprising

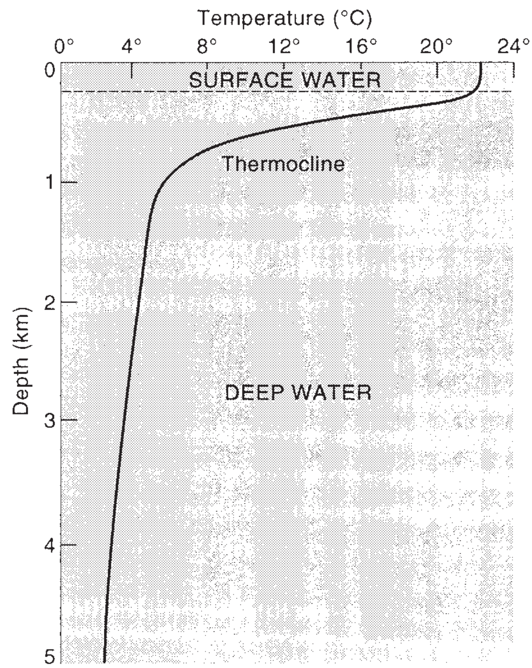

Figure 1.10 Generalized temperature-vs-depth profile for the oceans (at low to- mid-latitudes) showing vertical stratification into surface and deep water masses.

only 3.5% of the total ocean volume, while about one-third of the remaining volume is in the thermocline and two-thirds in the abyssal zone (see table 1.1).

In the surface ocean, lateral circulation is predominantly driven by the wind; in the deep ocean, circulation is driven by density variations due to differences in temperature and salinity, giving rise to the term *thermohaline circulation*. In this section the wind-driven and thermohaline circulations are discussed separately.

Wind-Driven (Shallow) Circulation

The circulation of the shallow ocean is driven by prevailing winds, which are in turn caused by uneven heating of the Earth's surface. The circulation pattern can be summarized as a number of current gyres that flow clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere due to the stresses imparted by the prevailing winds. These gyres extend poleward from about 10°N and S of the equator to about 45°N and S. This is shown in figure 1.11. Each gyre has a strong narrow poleward current on the western side with much weaker currents on the east. The most pronounced of these "westward intensification" currents are found in the Northern Hemisphere as the Gulf Stream in the Atlantic Ocean and the Kuroshio Current in the Pacific.

The circulation pattern shown in figure 1.11 is not what is expected if water were simply carried downwind. Transport of surface water, on the scale of

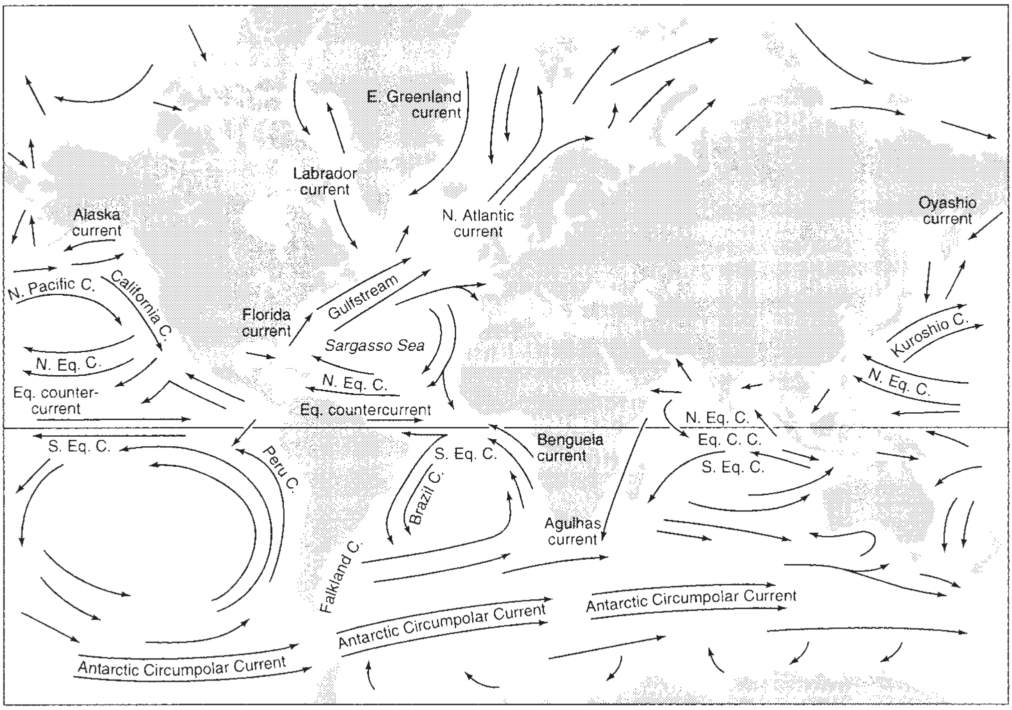

Figure 1.11 Surface currents of the oceans.
After Drake et al. 1978, fig. 6.1, p. 88.

the oceans, involves factors such as the Earth's rotation and friction against continents, as well as wind stresses on the surface. The interaction of wind and water is complicated, and only a brief summary will be given here. (For details, the interested reader should consult references on physical oceanography such as Pickard and Emery 1982, and Pedlosky 1990, or general oceanography texts such as Knauss 1997.)

The origin of the prominent gyres can best be understood by reference to the North Atlantic. Here the prevailing winds are westerlies (from the west) at 40°–50°N and trade winds (from the east) at 15°–30°N. Because of the Coriolis force (see previous section), water in the top layer does not simply move downwind, but instead is moved to the right of the wind direction (Ekman flow). This brings about a convergence or piling up of water from both the north and south into the central portion of the North Atlantic in the Subtropical Convergence. (This area is also known as the Sargasso Sea.) The piled-up water then sinks, and just below the surface, begins to return to the north and south. As it does, it is turned to the right by the Coriolis force, which results in a strong east-flowing current on the north and a strong west-flowing current on the south, just below the surface. These so-called geostrophic currents flow until they encounter the European continent at the northeast portion of the gyre and the North American continent along the southwest portion. Here they must turn. Since friction is strong along the continents and reduces the effect of the Coriolis force, the currents will flow from high pressure (where water is accumulating due to the current) to low pressure (where it is being removed), resulting in a north-to-south-flowing current on the European side and a south-to-north-flowing current (Gulf Stream) on the North American side. In this way a clockwise-flowing gyre results.

This type of explanation for the North Atlantic circulation can be applied to the other oceans, except that in the Southern Hemisphere the Coriolis force is to the left of the direction of motion, and as a result, the gyres are counterclockwise (see fig. 1.11). In the northern hemisphere the current on the west side of the Atlantic becomes intensified because of the superimposition of an additional pressure gradient (decreasing pressure from south to north due to the variation in Coriolis force) that reinforces the western current and opposes the eastern current.

Another prominent surface current is the Antarctic Circumpolar Current, which flows from west to east entirely around Antarctica and is driven by strong westerly winds. This is the primary current connecting the three ocean basins.

Coastal Upwelling

A special case of the wind-driven circulation is coastal upwelling, which has an important effect on biological productivity in the ocean (see chapter 8). Major upwelling occurs along the western boundaries of the continents, where the surface currents flowing toward the equator are broad and relatively weak. Winds blowing toward the equator along the coasts bring about a transport of surface water offshore. This is because the net transport of water (Ekman drift) is to the right of the wind in the Northern Hemisphere and to the left of the wind in the Southern Hemisphere. Transport of surface water away from shore leaves a nearshore deficit that is replaced by deeper water flowing up from below (see fig. 1.12). This deeper water is rich in nutrients, which results in high planktonic productivity and teeming life, including abundant fish. Some classic examples of upwelling areas are those off Peru and Chile, the bulge of West Africa, Namibia in southwest Africa, and the California coast.

Upwelling also results when surface waters are blown or transported away from an open-ocean area, bringing about the phenomenon of divergence. In such cases, subsurface waters move in to replace the missing surface water. Some noncoastal upwelling areas include the eastern equatorial Pacific, the sea around Antarctica, and the oceans at high northern latitudes (Kennett 1982).

Thermohaline (Deep) Circulation

Below the top few hundred meters, the oceans are not directly affected by the winds. Here circulation is brought about by density differences arising from differences in temperature and salinity. (The following discussion of deep ocean circulation is from Pickard and Emery 1982, Warren 1981, and Gordon 1986; see also Drake et al. 1978 and Knauss 1997.) In seawater, density increases continuously with decrease of temperature, and no density maximum, as is found in fresh water at 4°C (see chapter 6), is encountered. Density also increases with increasing salt content, or salinity. Overall the deep ocean is

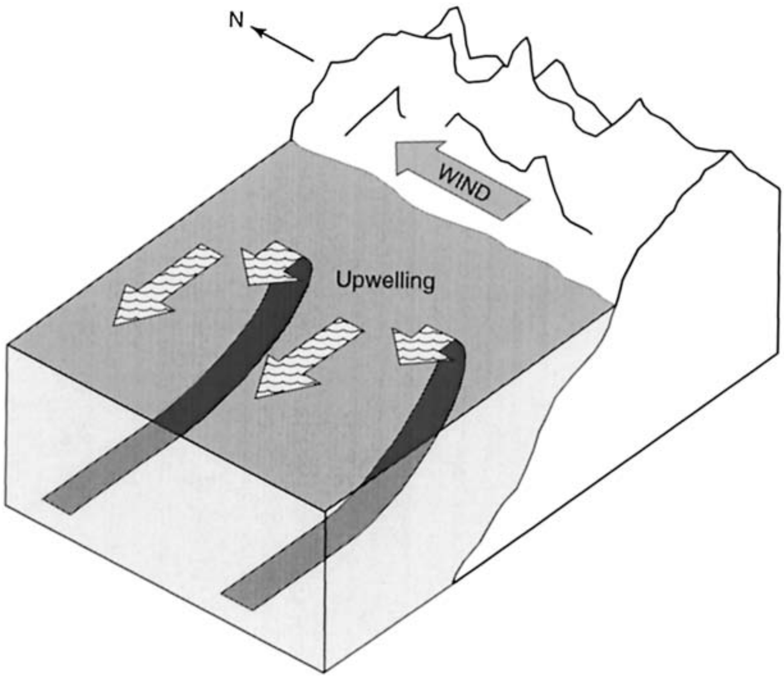

Figure 1.12 Upwelling, or the result of Ekman drift, in response to a north-blowing wind in the Southern Hemisphere.

After Turekian1976.

vertically stratified into various water masses, with the densest water at the bottom and the lightest at the top, and the density stratification is due primarily to the decrease of temperature with depth. The stratification severely inhibits vertical motion; in other words, it is much easier to move water along surfaces of constant density (isopycnals) than it is to move it across them. Thus, the deep-water circulation can be viewed as primarily horizontal.

The density stratification and density differences between water masses of the deep sea owe their origin to surface processes. Here changes in density are brought about by heating and cooling, evaporation, addition of fresh water, and freezing out of sea ice. Surface water migrating to high latitudes becomes more dense because of evaporation (which causes both cooling and increased salinity) and because of loss of sensible heat and sea-ice formation. (When sea ice forms, dissolved salts are excluded from the ice and the remaining water becomes more saline and, thus, more dense.) In certain locations in the far North and far South Atlantic, this surface density increase becomes so great that the cold, salty surface water occasionally, during winter, becomes denser than the underlying water and sinks downward to replace it. In this way deep ocean water originates and is replenished from above. Once it reaches great depths, the water tends to conserve its temperature and salinity as it flows laterally throughout the oceans, bringing about the deep water circulation.

An example of stratification and deep circulation for the Atlantic Ocean is shown in figure 1.13. Here each water mass is identified by its characteristic

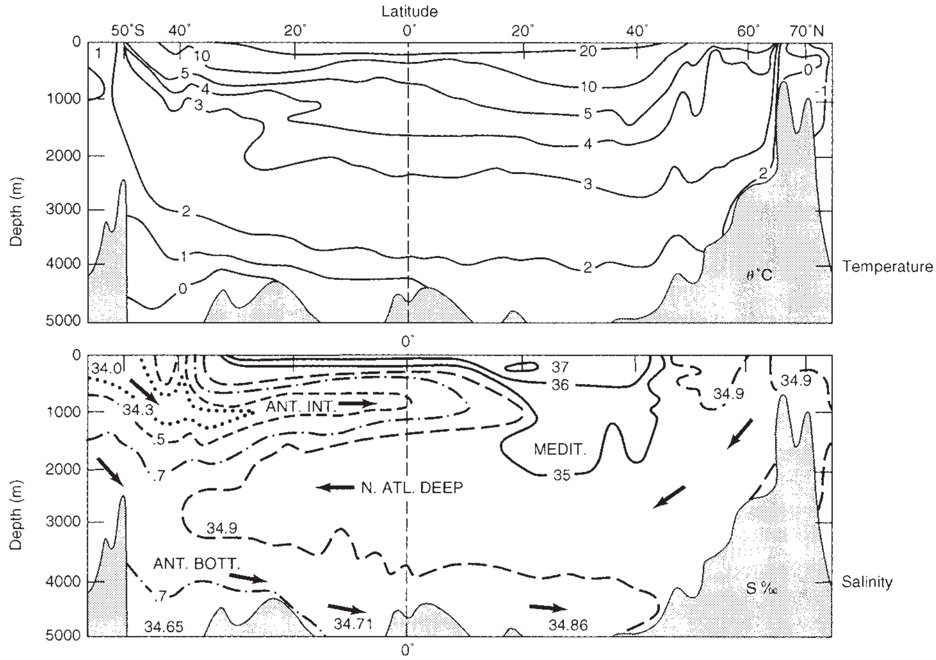

Figure 1.13 South-north vertical section of water properties of the Atlantic Ocean along the western trough as delineated by lines of constant temperature and salinity. N. Atl. Deep = North Atlantic Deep Water; Ant. Bott. = Antarctic Bottom Water; Ant. Int. = Antarctic Intermediate Water; Medit. = Mediterranean Water.

Adapted from Pickard and Emery 1982, based on data from Bainbridge 1976.

temperature and salinity. The deep water is dominated by two cold water masses, the North Atlantic Deep Water and the Antarctic Bottom Water. The North Atlantic Deep Water (NADW) originates in the Norwegian Sea off Greenland from the cooling of Gulf Stream surface water by evaporation and by surface cooling to form sea ice, leaving denser salty water behind. This water sinks and flows at depth southward, where it is joined by water sinking in the Labrador Sea off Canada. Ultimately this water crosses the equator. In the Antarctic an even denser water, the Antarctic Bottom Water, is formed in the Weddell Sea by the cooling and freezing-out of sea ice in the winter. This sinks to the bottom and flows north. In the Antarctic region further north, at the Antarctic Convergence, is formed the Antarctic Intermediate Water, which also sinks, but not to the bottom, because it cannot displace the underlying denser North Atlantic Deep or Antarctic Bottom waters. This water thus occupies intermediate depths, as implied by its name.

Another intermediate-type water is formed in the Mediterranean Sea. Here intense evaporation causes the water to be sufficiently saline and dense that, upon passing out into the Atlantic through the Straits of Gibraltar, it sinks and fills intermediate depths. It does not sink to the bottom because, although it is more saline than North Atlantic Deep Water, it is also warmer and the temperature difference counteracts the salinity effect making it less dense

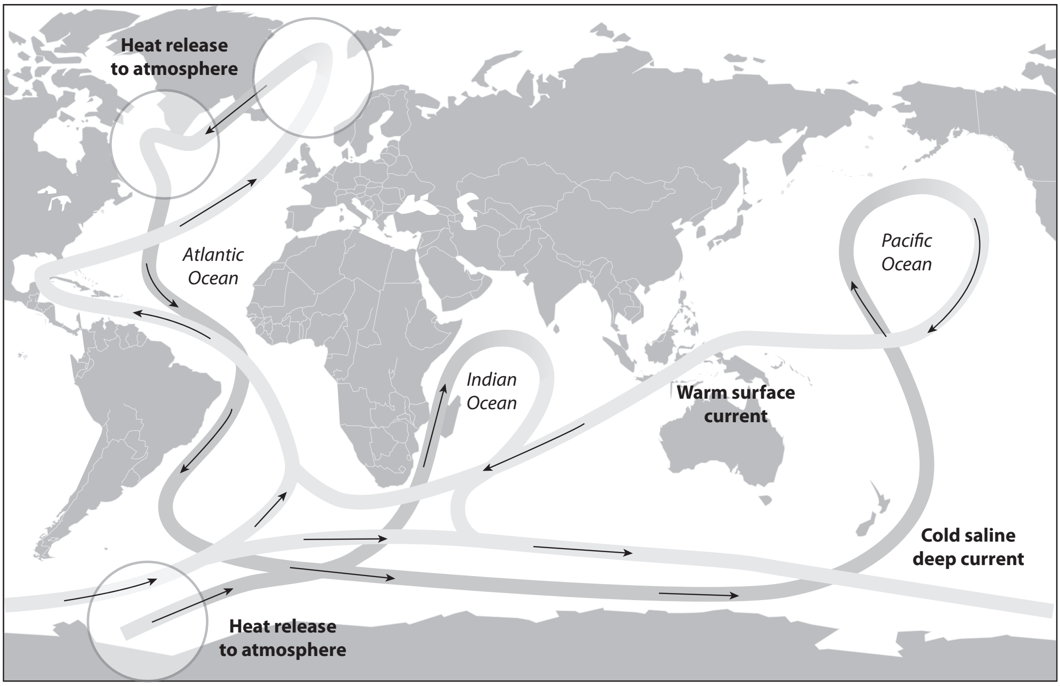

Figure 1.14 Global cycle of thermohaline circulation. Deep water circulation (darker lines) originates in the North Atlantic by sinking of North Atlantic Deep Water (NADW). NADW starts with water from the Norwegian Sea that is joined by water that sinks in the Labrador Bay. Both areas are denoted as sources of heat to the atmosphere as a result of cooling of the downwelling water. This water then flows south at depth as intensified currents along the western side of the ocean basin to the South Atlantic, where it is joined by Antarctic Bottom Water that sinks in the Weddell Sea (another "heat source" due to cooling). The deep water then flows into the deep Indian Ocean and Pacific Ocean. There is upwelling to the surface in all oceans. In addition, a warm water return circulation in the thermocline layer (shown by the lighter lines) serves as a way to return water to the North Atlantic. There is also return flow of surface water to the Atlantic through the Drake Passage south of South America.

After Intergovernmental Panel on Climate Change (IPCC) 2001.

than NADW. For the same reason, surface water at lower latitudes remains at the surface, even though it is more saline (see fig. 1.13, bottom); that is, the density-lowering effect of higher temperature overpowers the density-raising effect of higher salinity. Ultimately the Mediterranean water moves west and south across the Atlantic and joins the NADW in the western boundary current (Gordon 1986).

The lateral deep-water circulation over the entire ocean, which was originally modeled by Stommel (1958), is shown in figure 1.14. (Deepwater circulation is not well documented and maps, such as that shown in figure 1.14, are based largely on theoretical models). North Atlantic Deep Water flows away southward from its source as an intense bottom current on the western side of the North Atlantic. This meets a strong northward flowing current of Antarctic

Bottom Water in the South Atlantic, and as a result they merge and flow east through the Antarctic into the deep Indian Ocean and ultimately into the deep Pacific Ocean (Gordon 1986, Warren 1981). Thus, the deep water originates only in two areas, and both are in the Atlantic Ocean.

During the thermohaline circulation, deep waters remain out of contact with the atmosphere for long periods of time, which can result in appreciable change in their chemical composition (see chapter 8). The residence time of deep water, or average time it spends out of contact with the atmosphere, is about 200–500 years for the Atlantic and 1000–2000 years for the Pacific.

As the deep water traverses the ocean floor, there is very slow, diffuse upwelling (at the rate of about 1 m/yr) of deep water into surface layers. This slow upwelling is supplemented by the more intense but localized coastal and open-ocean upwelling discussed in the previous section. In addition, there are two direct routes proposed to return water to feed North Atlantic Deep Water formation. One is the flow of cold surface water through the Drake Passage (south of South America) into the South Atlantic. The other is a proposed return flow of warm water toward the North Atlantic in the ocean's thermocline layer. (Both routes are shown by lighter lines in fig. 1.14)

There is considerable interest in the thermohaline circulation of the North Atlantic and rate of formation of NADW because of its climatic implications. Possible changes in NADW formation have been cited as a cause of rapid climate change over the last 18,000 years (e.g., Street-Perrott and Perrott 1990), and increased global temperatures from greenhouse warming (see chapter 2) may affect the rate of formation of NADW in the future (e.g., Gates et al. 1992 and Broecker 1987). Manabe et al. (1991) find that enhanced greenhouse warming may reduce the rate of NADW formation, thus reducing the surface water flow from the south and delaying the future greenhouse response in the northern North Atlantic. NADW formation is sensitive to changes in surface water temperature and salinity (the latter due to the addition of fresh water from ice melting [Street-Perrott and Perrott 1990]) and changes in ocean currents (Shaffer and Bendtsen 1994). Lozier et al. (2010) have recently emphasized that the thermohaline circulation changes in response to natural variations in such things as winds and oceanic eddy circulation.