CHAPTER

5

The Hydrologic Cycle

5.1 WATER, ESSENTIAL TO CLIMATE AND LIFE

Water continually moves between the oceans, the atmosphere, the cryosphere, and the land. The movement of water among the reservoirs of ocean, atmosphere, and land is called the hydrologic cycle. The total amount of water on Earth remains effectively constant on time scales of thousands of years, but it changes state between its liquid, solid, and gaseous forms as it moves through the hydrologic system. The amount of water moved through the hydrologic cycle every year is equivalent to about a 1-m depth of liquid water spread uniformly over the surface of Earth. This amount of water annually enters the atmosphere through evaporation and returns to the surface as precipitation in rain or snow. To evaporate 1 m of water in a year requires an average energy input of 80 Wm^{-2}. The sun provides the energy necessary to evaporate water from the surface. Once within the atmosphere, water vapor can be transported horizontally for great distances and moved upward. This horizontal and vertical movement of water vapor is critical to the water balance of land areas, since about one-third of the precipitation that falls on the land areas of Earth is water that was evaporated from ocean areas and then transported to the land in the atmosphere (Fig. 5.1). The excess of precipitation over evaporation in land areas supports the return of water from the land to the ocean in rivers.

The atmosphere contains a relatively small amount of water (Tables 5.1 and 1.2). If all the water vapor in the atmosphere were condensed to liquid and spread evenly over the surface of Earth, it would be only about 2.5 cm deep. Since 100 cm of water is evaporated and condensed per annum, the atmospheric water is removed by precipitation about 40 times a year, or every 9 days. Because net evaporation is a small residual of a more rapid two-way exchange of water molecules across the air–water interface, the actual residence time of water molecules in the atmosphere is about 3 days. Since nearly a 3-km depth of water is present near the surface of Earth, most of which is in the oceans, and only 2.5 cm can reside in the

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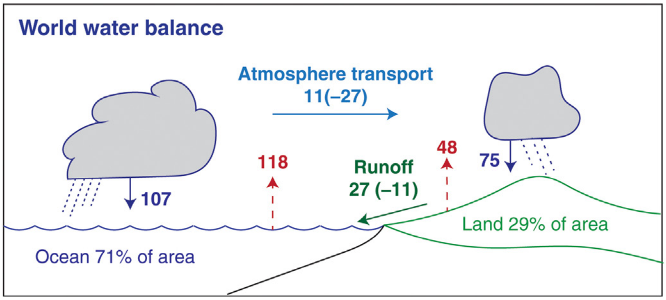

FIGURE 5.1 Schematic diagram showing the basic fluxes of water in the global hydrologic cycle. Units are centimeters per year spread over the area of the land or ocean. Since the areas of land and ocean are different, the land–ocean water exchanges by atmospheric transport and river runoff have different values depending on the reference area, as indicated by the parentheses. The smaller values are those referenced to the larger oceanic area.

TABLE 5.1 Water Volumes of Earth

Category	Volume (10^6 km^3)	Percent (%)
Oceans	1348.0	97.39
Polar ice caps, icebergs, glaciers	227.8	2.010
Ground water, soil moisture	8.062	0.580*
Lakes and rivers	0.225	0.020
Atmosphere	0.013	0.001
Total water amount	1384.0	100.0
Fresh water	36.00	2.60

Fresh water reservoirs as a percent of total fresh water	
Polar ice caps, icebergs, glacier	77.2
Ground water to 800 m depth	9.8*
Ground water from 800 m to 4000 m	12.3*
Soil moisture	0.17*
Lakes (fresh water)	0.35
Rivers	0.003
Hydrated earth minerals	0.001
Plants, animals, humans	0.003
Atmosphere	0.040
Sum	100.000

** Numbers uncertain.*
(From Baumgartner and Reichel, 1975.)

atmosphere, an average water molecule must wait a very long time in the ocean, in an ice sheet, or in an aquifer, between brief excursions into the atmosphere.

In earlier chapters we saw the important role of water in many aspects of the climate system. Water is crucial to life, and the existence of oceans on Earth has dramatically influenced the character and evolution of Earth's atmosphere. Chemical and biological processes that take place in the oceans continue to regulate atmospheric composition. In the current atmosphere, water vapor is the most important gaseous absorber of solar and terrestrial radiation and accounts for about half of the atmosphere's natural greenhouse effect. Clouds of liquid water and ice contribute about 30% of the atmosphere's natural opacity to thermal radiation and contribute about half of Earth's reflectivity for solar radiation. The evaporation of water accounts for about half of the cooling of the surface that balances the net radiative heating of Earth's surface. The energy required to evaporate 1 kg of water is about 593 times the energy required to warm that water by 1°. As the water vapor rises into the atmosphere, it eventually condenses and precipitates, but the energy released during the condensation of atmospheric water vapor helps to drive the circulation systems of the atmosphere. Water can alter the surface albedo of Earth through the deposition of snow and ice and by fostering the development of vegetative cover on land surfaces.

5.2 THE WATER BALANCE

To understand how local climates are maintained, it is instructive to consider the water budget for the surface. In order to model the climate, the surface water balance must be accurately represented. The surface water balance may be written as

$$g_w = P + D - E - \Delta f \tag{5.1}$$

where g_w is the storage of water at and below the surface, P is the precipitation by rain and snow, D is the surface condensation (dewfall or frost), E is the evapotranspiration, and Δf is the runoff.

Averaged over a long period of time, the storage term is small. In addition, dewfall is usually small, or can be incorporated into a generalized precipitation. The resulting hydrologic balance for a long-term average is

$$\Delta f = P - E \tag{5.2}$$

A complementary balance for the atmosphere must also be satisfied. Precipitation minus evaporation is the net flux of water from the atmosphere to the surface and occurs with opposite sign in the atmospheric water balance.

$$g_{wa} = -(P + D - E) - \Delta f_a \tag{5.3}$$

The terms have the same meaning as in (5.1), except that g_{wa} indicates storage of water in the atmosphere and Δf_a indicates horizontal export of water by atmospheric motions, primarily in the form of water vapor. Adding the budgets for the surface (5.1) and the atmosphere (5.3), we obtain a water balance for the surface–atmosphere system in which the exchange of water across the surface does not appear.

$$g_w + g_{wa} = -\Delta f - \Delta f_a \tag{5.4}$$

When averaged over a year, the storage terms on the left of (5.4) are generally small, and the horizontal transport of water out of a region by the atmosphere must be equal and opposite to the net horizontal transport at or below the surface. This means that water carried to continents by atmospheric transport must equal the runoff from rivers, and atmospheric export of water from the subtropics must be balanced by ocean currents.

The distributions with latitude of the terms in the annually averaged surface water balance are shown in Fig. 5.2. Precipitation peaks near the equator, with secondary maxima in middle latitudes of each hemisphere. The equatorial maximum is associated with heavy precipitation in the *intertropical convergence zone*, where the trade winds converge from either hemisphere. Moisture-laden air near the surface flows toward the equator from both hemispheres and converges near the equator, where it is

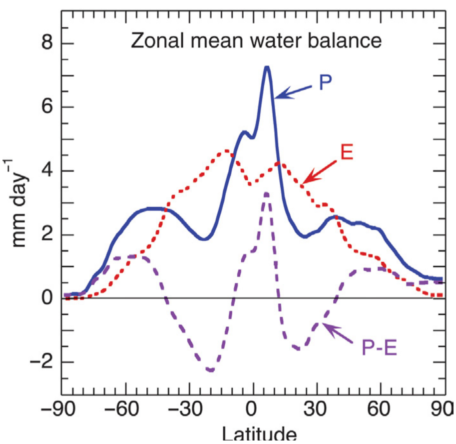

FIGURE 5.2 **Latitudinal distribution of the surface hydrologic balance, showing evaporation *E*, precipitation *P*, and *P* – *E* which equals runoff Δ*f*.** *Data from ERA Interim 1979–2009.*

released in thunderstorms, tropical cyclones, and other precipitation-producing weather systems (Fig. 5.3). The secondary maxima in mid-latitudes of each hemisphere are associated with the weather systems there. In middle latitudes, cyclonic disturbances with strong winds drive vertical motions that release water. Evaporation varies more smoothly

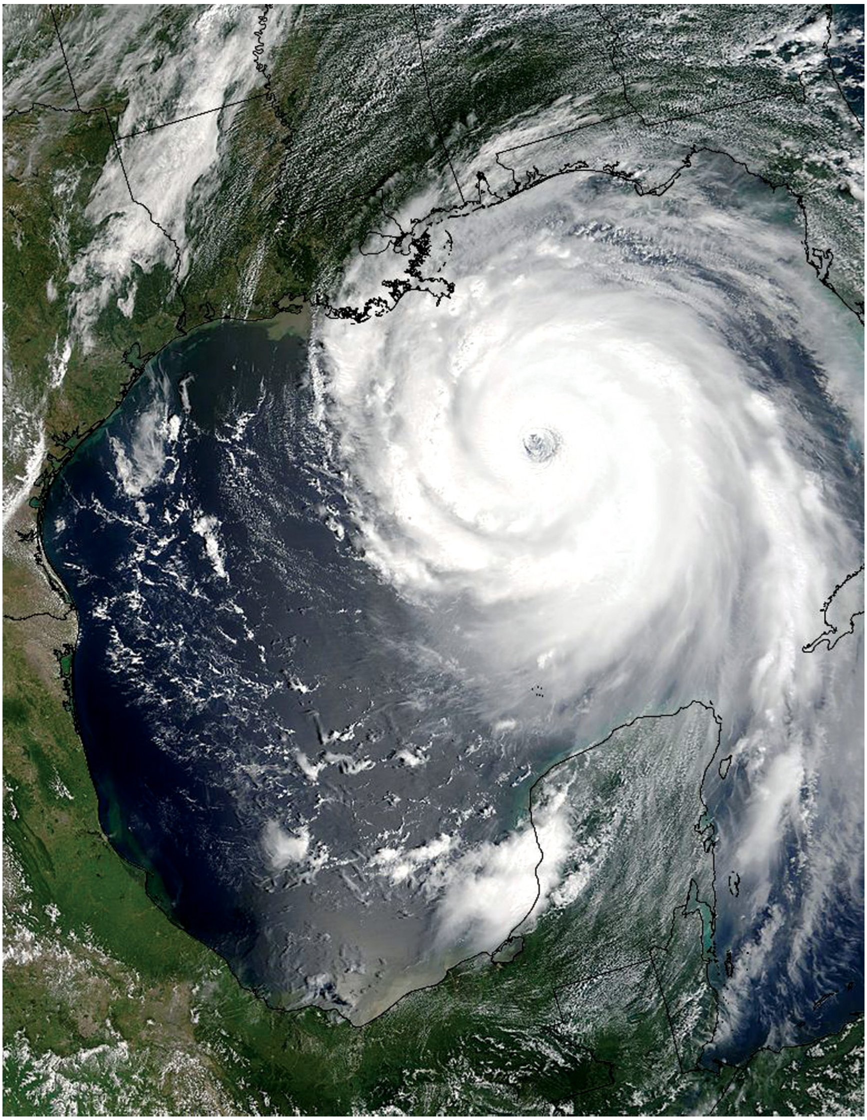

FIGURE 5.3 **Satellite image of hurricane Katrina approaching the Gulf Coast of the United States on August 29, 2005.** *NASA MODIS image.*

than precipitation, with a broad maximum in the tropics. Precipitation exceeds evaporation in the equatorial belt and again in middle to high latitudes. Evaporation exceeds precipitation in the belt from 15° to 40° of latitude, and these regions export water vapor to be condensed in the latitudes where the precipitation maxima occur. The runoff (also $P - E$) distribution shown in Fig. 5.2 implies transport of water vapor in the atmosphere from the subtropics to the equatorial and high latitude zones. A return flow in the oceans or rivers carries water back toward the subtropics.

The water balances of the continents and oceans are closely related to their climates and the processes that maintain climate (Table 5.2). The Atlantic and Indian oceans are net exporters of water vapor, whereas the Pacific and Arctic oceans receive more water in the form of precipitation than they give up to the atmosphere through evaporation. Comparison of the surface salinity of the Atlantic and Pacific oceans shows a much higher salinity in the north Atlantic than in the north Pacific (see Chapter 7). The surface hydrologic balance of the oceans plays an important role in determining their salinity and thereby the deep circulation of the oceans. The saline surface water of the Atlantic is a key factor for allowing surface

TABLE 5.2 Water Balance of the Continents and Oceans in mm/year

Region	E	P	Δf	$\Delta f/P$
LAND				
Europe	375	657	282	0.43
Asia	420	696	276	0.40
Africa	582	696	114	0.16
Australia	534	803	269	0.33
North America	403	645	242	0.37
South America	946	1564	618	0.39
Antarctica	28	169	141	0.83
All Land	480	746	266	0.36
OCEAN				
Arctic Ocean	53	97	44	0.45
Atlantic Ocean	1133	761	−372	−0.49
Indian Ocean	1294	1043	−251	−0.24
Pacific Ocean	1202	1292	90	0.07
All Ocean	1176	1066	−110	−0.10
Globe	973	973	0	

From Baumgartner and Reichel, 1975.

water to sink to the bottom of the ocean, since salinity is an important variable in determining seawater density, especially in high latitudes.

The *runoff ratio*, $\Delta f/P$, is a measure of the wetness of a continent. If it is large, then a significant fraction of the precipitation that falls on that continent flows into the ocean, rather than being evaporated over the land. The dry continents of Africa and Australia have relatively low runoff ratios. Typically, about 40% of the precipitation on a continent runs back to the global ocean in rivers. The evaporation from the surface of a continent typically makes up 60% of the precipitation that falls on that continent.

5.3 SURFACE WATER STORAGE AND RUNOFF

The storage term in (5.1) accounts for changes in the amount of water that is retained in the surface. Over land areas, this includes the water in the near surface soil and also water that flows deeper and becomes part of an underground water system. An additional important form of water storage is surface snow cover. Distinct seasons of precipitation and drying are a prominent feature of the climate in many regions. For such regions, storage of water in the soil and in snowpack is critical for determining the nature of the environment that develops during the dry season. In many mid-latitude regions, mountain snowpack is essential for spring and summer river flow, and at lower elevations spring snowmelt helps to replenish soil moisture and groundwater for the summer dry season. The combination of moist soil in springtime followed by summer warmth and sunshine makes many mid-latitude land areas agriculturally productive. Storage of precipitated water in snowpack depends only on the surface thermodynamics and physical structure. Storage of water that arrives at the surface as rainfall depends on the frequency and intensity of the precipitation and the characteristics of the soil, its vegetative cover, and the topography of the surface.

Climate interacts only with water that is on the surface or in the soil water zone. The soil water zone extends downward to the depth penetrated by the roots of the vegetation. Plants can draw water from this depth relatively quickly and release it to the atmosphere by *transpiration* through leaves. Because roots of plants can draw moisture from the soil more quickly than water is brought to the surface by nonbiological processes, vegetated surfaces normally release water more quickly to the atmosphere than bare soil does with the same water content. Depending on the conditions, one may need to consider a layer deeper than the root zone in order to predict surface moisture and evapotranspiration. Moisture stored deeper in the soil than the root zone must be brought upward by diffusion or capillary action. Transport through the soil in both liquid and vapor form is possible.

Water is suspended in the soil by adherence to soil particles in thin films. The amount of water that can be held in this manner is called the *field capacity* of the soil. The moisture balance of the soil layer and the average soil moisture content are critical to the local climate of land areas. The water in this zone is available for use by plants and can be transpired or evaporated. The soil layer and associated vegetation determine the fate of precipitated water, which may be quickly re-evaporated, absorbed by the soil, or run off in stream flow. The transfer of surface water to the soil is called infiltration. The fraction of precipitation that is retained by the soil is determined by soil and vegetation properties and by the rate and frequency of precipitation.

If the soil water content increases above the field capacity, then gravitational forces carry the water downward to the water table, where it becomes part of the groundwater. If the water encounters an impermeable obstacle such as bedrock, then it may flow laterally, seeking lower pressure. Gradual collection of water in subsurface reservoirs results in the formation of aquifers, from which freshwater can be extracted. In many areas, subsurface fresh water reservoirs that have taken many centuries to form, or that contain fossil water from previous wet epochs, are being pumped out for human use.

If the surface soil is saturated and the precipitation or snowmelt is more rapid than can be balanced by infiltration and evaporation, then surface ponding will occur. Once the surface depressions in the soil are filled with water, the surface water will begin to flow laterally toward streams and drainage systems. Water runoff from land areas in streams and rivers is important for navigation, fisheries, hydroelectric power generation, irrigation of dry land areas, and municipal water supplies.

5.4 PRECIPITATION AND DEWFALL

Precipitation is produced when air parcels become supersaturated with water vapor, condensation and droplet formation occur, the droplets become large enough to fall, and the droplets or particles reach the surface without re-evaporation. Supersaturation is most often caused by cooling as air parcels expand adiabatically during ascent. Atmospheric motions associated with midlatitude frontal and synoptic weather systems can force ascent of air parcels. In the tropics and over continents during summer, ascent, condensation, and precipitation are often associated with convective instability, where parcels of air are forced upward by buoyancy in cumulonimbus clouds. In stratiform cloud systems, light but steady precipitation is generated through the radiative cooling of the tops of the clouds and steady overturning of moist air from beneath. Heavy and persistent precipitation may result when moist air is forced over mountain ranges by prevailing winds.

Global precipitation maps are produced by blending information from surface gauges and remote sensing from satellites. The geographic distribution of precipitation is shown in Fig. 5.4. The general features of the

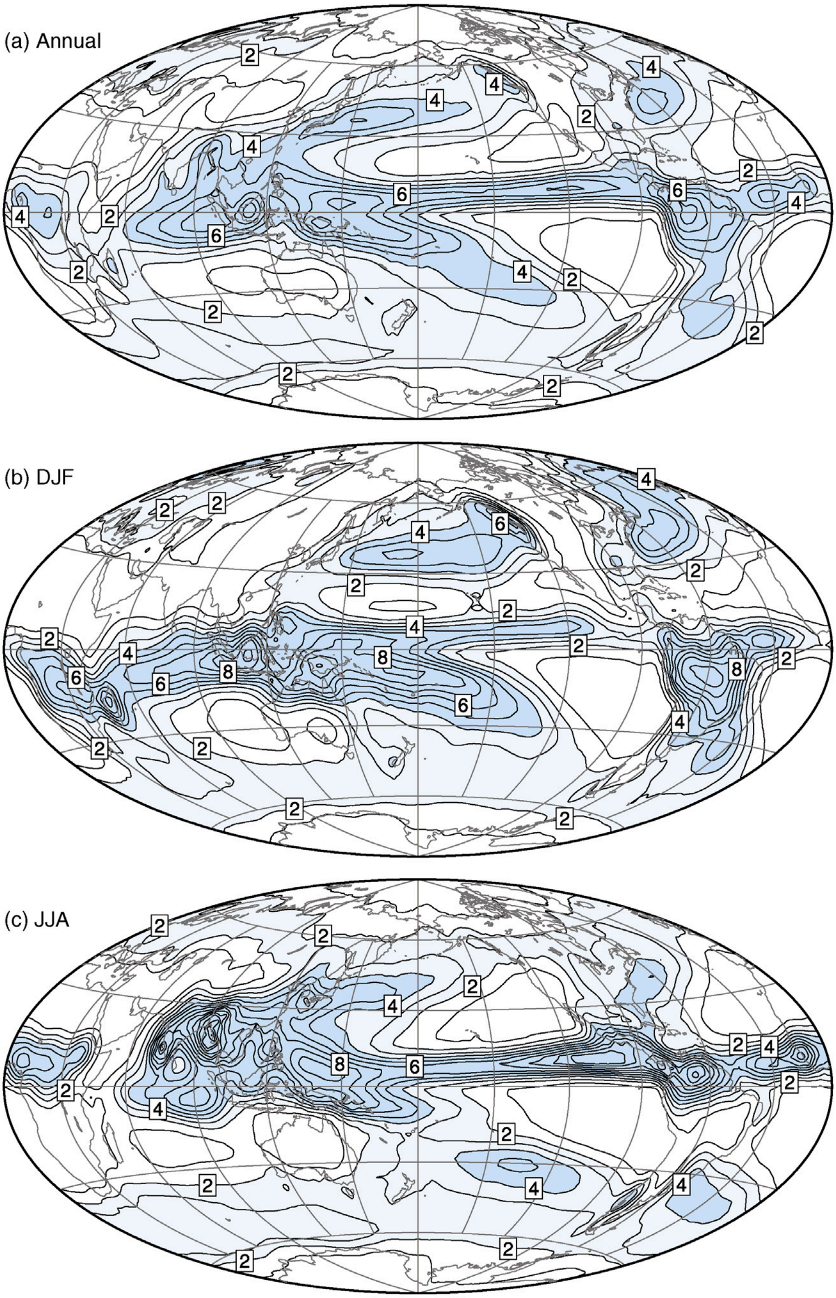

FIGURE 5.4 **Global maps of precipitation in millimeters per day for annual, DJF and JJA seasons.** Note that the Dateline is in the center of this plot. *Data from Global Precipitation Climatology Project.*

zonal-average precipitation in Fig. 5.2 are apparent, with the largest precipitation near the equator, where the average water content of the air is high and tropical convective systems are responsible for much of the rainfall. The tropical precipitation is concentrated near the Maritime Continent formed by the islands of the Western Pacific area and over the land areas of South America and Africa. In the eastern halves of the Pacific and Atlantic oceans, the precipitation forms a line north of the equator known as the Inter-Tropical Convergence Zone (ITCZ). Branching southeast from the ITCZ in the Pacific is the South Pacific Convergence Zone (SPCZ), which reaches from the Equator west of the Dateline toward mid-latitudes west of South America. In mid-latitudes of the Northern Hemisphere, the precipitation is concentrated near the western margins of the ocean, where storm tracks form, while in the Southern Hemisphere precipitation in mid-latitudes is strong at all longitudes. The tropical precipitation moves north and south following the Sun, particularly in the Asian sector, where heavy precipitation reaches nearly to 30°N in JJA. The forced ascent of moist surface air in mid-latitude weather systems and the westerly flow over obstacles such as coastal mountain ranges give rise to heavy precipitation, which can be seen in mid-latitudes west of the Rocky Mountains and the Andes. Precipitation declines toward polar regions. The entire hydrologic cycle is slowed down in polar regions because of the low temperatures and consequently low water-carrying capacity of the atmosphere.

When air comes into contact with a cold surface, usually on relatively clear nights, water vapor may condense directly onto the surface and form dew. Vapor flux from the soil may also be an important contributor to the accumulation of dew, especially at night when the underlying soil may be warmer than the surface. Dewfall is a significant contributor to the surface water balances in some arid climates, but is generally small and lumped together with the precipitation. Fog droplets that are too small to precipitate can be collected from the air by the leaves or needles of plants. In some climates such "combing" of liquid water from the air is an important mechanism whereby plants obtain moisture.

5.5 EVAPORATION AND TRANSPIRATION

Evapotranspiration is the removal of water from the surface to the air with an accompanying change in phase from the liquid to the vapor form. It is the sum of evaporation and transpiration. Evaporation refers to direct evaporation of water from the surface itself. *Transpiration* is the passage of water from plants to the atmosphere through leaf pores called *stomata*, which also serve as the point of entry for carbon dioxide required for photosynthesis. Water is absorbed from the soil and carried through the roots and stems of plants to the leaves, where it escapes as water vapor. Stomata normally close at night and open during the day, but they may also close at midday in response to high temperatures, temporary water deficit, or high carbon dioxide concentrations. The differences between evaporation and transpiration are important, but it is difficult to separate the effects of the two processes in practice, so they are generally added to form a single term in water budgets. Evapotranspiration may also include *sublimation*, which refers to the direct conversion of snow and ice to water vapor, without an intermediate liquid phase.

Evaporation from a wet surface is determined by the surface tension at the air–water interface and the rate of decrease of water-vapor concentration between the water surface and the adjacent air. The rate at which the water-vapor concentration changes with distance from a water surface depends on the molecular diffusivity and the ventilation of the air near the water surface by air motions. Normally, turbulent air motions are of primary importance for carrying water vapor away from a surface and dominate in determining gradients on scales larger than a few millimeters. The interaction of surface water waves with atmospheric turbulence can influence the rate of evaporation over the oceans. Over land, the structure of the surface and the vegetation covering it can have a substantial effect on the rate of evaporation. The collection of vegetable matter covering the land surface is called the *plant canopy*, which may be as thin as a layer of moss or as thick as a tall forest.

Plant canopies have important effects on the water and energy balances of the surface. Some of these effects are illustrated in Fig. 5.5. Precipitation that falls on a plant canopy can be intercepted by leaves and stems. Water that falls on the leaves can be evaporated from the leaves or drip to the surface. Interception of precipitation by leaves and evaporation from leaves can greatly decrease runoff if the rainfall rate is not too intense and the air is relatively dry. The leaf structure of a plant presents a much larger surface area on which evaporation can take place than the ground surface alone. The energy balance of the leaves is also of importance, since it determines how rapidly water can be evaporated. The structure and arrangement of leaves and branches affect the absorption of solar radiation, the emission of longwave radiation, and the ventilation of the surface by air motions. A parameter often used to characterize plant canopies is the *leaf area index* (LAI). It is defined as the ratio of the area of the tops of all the leaves in the canopy projected onto a flat surface to the area of the surface under the canopy. It is equal to the number of leaves that would be crossed by a vertical line passing through the canopy, on average.

5.5.1 Measurement of Evapotranspiration

Evapotranspiration can be estimated in a variety of ways. One of the most accurate methods is by weighing the moisture change in the soil

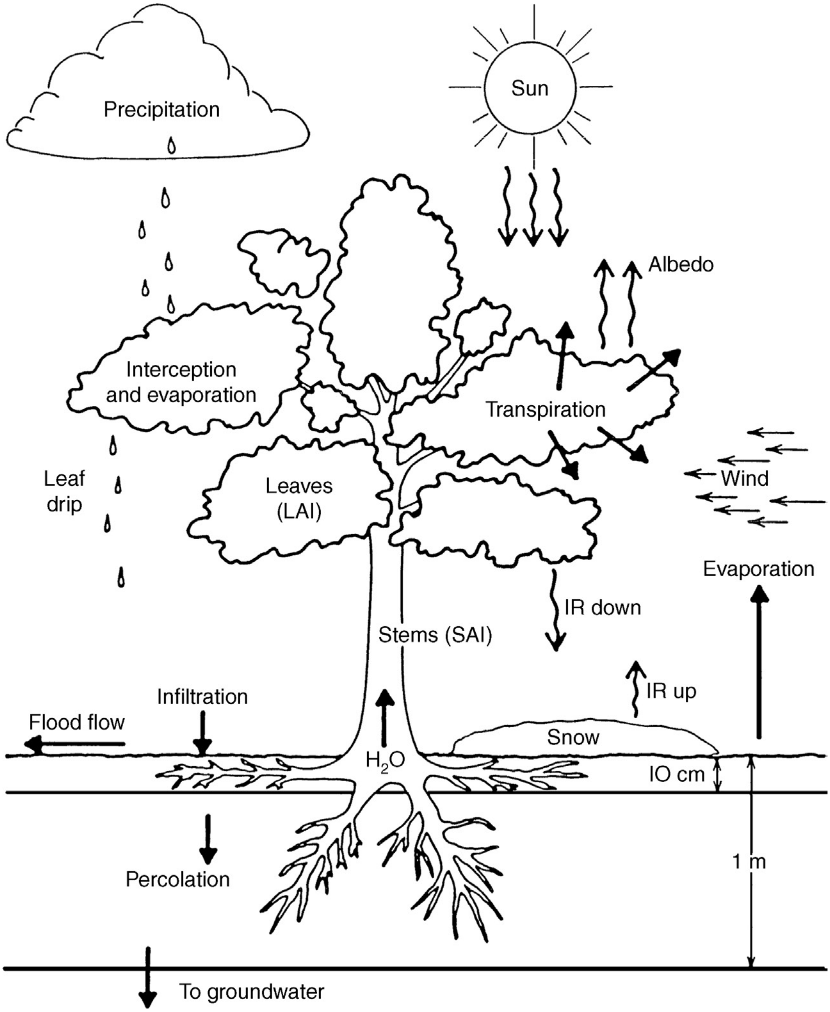

FIGURE 5.5 **Diagram showing the effects of the vegetation canopy on the water and energy fluxes.** *From Dickinson (1984). © American Geophysical Union.*

and its vegetative cover with a device called a *lysimeter*. A lysimeter is a container of soil set on a balance or provided with some other means of measuring water content. To obtain accurate results, the lysimeter must be large enough to contain the soil water zone and associated vegetation, and should be set in a larger environment where the surface conditions are similar to those under investigation, if results representative of the natural environment are desired. The lysimeter measures the weight of water in the soil. The net flux of moisture from the soil can be estimated from the change in weight.

Evapotranspiration can also be estimated by directly measuring the fluxes of moisture away from the surface by taking simultaneous measurements of vertical velocity and humidity. Because the moisture is carried upward by turbulent motions, the device used to measure wind and humidity fluctuations must respond on the time scale of seconds, but modern technology has made such measurements routine. It is also a challenge to obtain representative spatial and temporal means, particularly if the surface characteristics are spatially inhomogeneous.

An alternative to direct measurement of evapotranspiration is to infer it as a residual in the energy balance, if the other terms in the energy balance can be measured. Rearranging (4.1) to solve for the evaporation rate yields

$$E = \frac{1}{L}(R_s - SH - \Delta F_{eo} - G) \tag{5.5}$$

When the surface is moist, the net radiation and the evaporation are the largest terms in the surface energy balance (see Chapter 4), so that an accurate measurement of the net radiation and approximations to the other terms in the surface energy balance will provide a good estimate of evapotranspiration. Radiation can be measured very accurately and over long periods with relatively inexpensive instrumentation, and most weather stations have a device for measuring insolation. Sensible heat loss can be estimated from bulk aerodynamic formulas, if measurements of mean wind speed and temperature at two levels are available. Measurements of temperature profiles in the soil or water can be used to estimate energy storage below the surface.

5.5.2 Evaporation From a Wet Surface

Penman (1948) derived a method of calculating the evaporation from wet surfaces with minimal input data. The Bowen ratio is the ratio of sensible to latent surface energy flux. It may be estimated by using the bulk aerodynamic formulas (4.26) and (4.27).

$$B_o = \frac{SH}{LE} \cong \frac{c_p(T_s - T_a)}{L(q_s - q_a)} \tag{5.6}$$

Here we have assumed that the aerodynamic transfer coefficients for heat and moisture are equal. If the surface air is saturated, and the surface and reference-level air temperatures are not too different, we may make the following approximation:

$$\frac{(q_s^* - q_a^*)}{(T_s - T_a)} \cong \frac{dq^*}{dT} \tag{5.7}$$

where q^* is the saturation mixing ratio of water vapor. Using (5.7) in (5.6), and the assumption that the surface air is saturated, we obtain

$$B_o = B_e\left(1 - \frac{(q_a^* - q_a)}{(q_s^* - q_a)}\right) \tag{5.8}$$

where B_e is the equilibrium Bowen ratio defined in (4.33). It should be noted that the use of the Bowen ratio can be problematic in the presence of temperature inversions if the denominator in (5.7) is near zero.

The surface energy balance (5.5) may be rewritten as

$$E(1 + B_o) = E_{en} \tag{5.9}$$

where

$$E_{en} = \frac{1}{L}(R_s - \Delta F_{eo} - G) \tag{5.10}$$

E_{en} is the evaporation rate necessary to balance the energy supply to the surface by radiation, horizontal flux below the surface, and storage. Substituting (5.8) for the Bowen ratio in (5.9) yields

$$E(1 + B_e) = E_{en} + E\, B_e \frac{(q_a^* - q_a)}{(q_s^* - q_a)} \tag{5.11}$$

Using the bulk aerodynamic formulas, the evaporation may be eliminated from the second term on the right in (5.11) to yield an expression for the evaporation from a wet surface, which is often called Penman's equation.

$$E = \frac{1}{(1 + B_e)} E_{en} + \frac{B_e}{(1 + B_e)} E_{air} \tag{5.12}$$

The evaporating capacity of the air is defined by

$$E_{air} = \rho C_{DE} U (q_a^* - q_a) = \rho C_{DE} U q_a^* (1 - \mathrm{RH}) \tag{5.13}$$

and depends on the relative humidity of the air, RH, as well as the air temperature and wind speed.

The advantage of (5.12) is that measurements of atmospheric variables at only one level are required. Over land surfaces the horizontal transport term is zero, and for time scales of a month or longer the storage term can also be ignored, so that only measurements of the net radiation, air temperature, specific humidity, and wind speed at one level are required to evaluate evaporation. The Penman equation (5.12) also shows the relative roles of air humidity and available radiation in driving evaporation over a wet surface. At high temperatures the equilibrium

Bowen ratio is small and evaporation is mostly dependent on available energy. As the equilibrium Bowen ratio becomes small, the evaporation rate approaches a value necessary to balance the energy input to the surface. This occurs at temperatures greater than about 25°C. At lower temperatures, and consequently higher equilibrium Bowen ratios, the evaporation rate is more dependent on the supply of unsaturated atmospheric air. At temperatures near or below freezing, the equilibrium Bowen ratio is large and the evaporation is dependent primarily on the drying capacity of the air.

5.5.3 Potential Evaporation

Evapotranspiration is constrained by the surface water supply, the energy available to provide the latent heat of vaporization, and the ability of the surface air to accommodate water vapor. The potential evaporation is defined as the rate of evaporation that would occur if the surface was wet, and is therefore the maximum possible evaporation for the prevailing atmospheric conditions. It measures the effect of energy supply and air humidity on the evaporation rate and avoids the more difficult issue of soil moisture availability and the physiological processes in plants that bring moisture from the soil to the atmosphere. If the potential evaporation exceeds the actual evapotranspiration, then a moisture deficit exists, and one may infer a dry surface. One method to calculate potential evaporation is from Penman's equation, which relates the evaporation from a wet surface to net radiative heating and mean air temperature, humidity, and wind speed at one level.

The potential evaporation can be used to understand how the hydrologic cycle at the surface might change with global mean temperature. The strongest variation in the potential evaporation is the saturation specific humidity, which increases at an exponential rate (1.11). We therefore expect that potential evaporation will increase in a warmer climate, meaning that water will be removed more efficiently from the surface in a warmed climate than in a cooler one. At the same time, if the atmospheric circulation does not change significantly, more moisture will be converged in regions of moisture convergence. We therefore expect that with warming will come greater contrast between areas in which precipitation exceeds evaporation and where evaporation exceeds precipitation. This is the "wet gets wetter, dry gets dryer" paradigm of global warming.

Since radiation dominates the energy supply for evaporation (5.10), we can write (5.12) as

$$PE = \frac{1}{(1 + B_e)} \left(\frac{R_s}{L} + B_e E_{air} \right) \tag{5.14}$$

where we have introduced the acronym *PE* for potential evaporation. Also, we see that the second term in the parentheses can be written,

$$B_e E_{air} = \frac{\rho C_D U(1-RH)}{\dfrac{L}{c_p}\dfrac{d\ln q^*}{dT}} = \frac{\rho C_D U(1-RH)}{\dfrac{L}{c_p}\dfrac{L}{R_v T^2}} \tag{5.15}$$

Here we have used,

$$\frac{d\ln q^*}{dT} = \frac{L}{R_v T^2} \approx 6.5\%\,\mathrm{K}^{-1} \tag{5.16}$$

Equation (5.16) states that the saturation specific humidity increases about 6.5% for each degree of warming for temperatures around 288 K. The exponential dependences of saturation vapor pressure in B_e and E_{air} cancel out in the second term of (5.14) so that neither term within the parentheses have the exponential dependence on temperature of saturation vapor pressure. If we assume, as is commonly done, that the relative humidity and the wind speed do not change very rapidly with warming, then the two terms inside the parentheses should vary only modestly with temperature, and the strongest temperature dependence resides in the $(1 + B_e)^{-1}$ in (5.14). This means that *PE* will increase with temperature, assuming that wind speed and relative humidity change slowly.

PE is, however, fairly sensitive to the relative humidity in the boundary layer. The relative humidity in the boundary layer is about 80%, so that if the relative humidity decreases to 79%, 1−RH would change by 5% from 20% to 21%. So small changes in relative humidity in the boundary layer can cause big changes in the gradient of moisture in the boundary layer, which would strongly affect *PE*. The relative humidity in the boundary layer is maintained by a complex set of processes, and it is not a simple calculation to predict how it would change in response to climate warming.

Since we have established that the terms inside the parentheses in (5.14) do not change as rapidly as saturation vapor pressure, the $(1 + B_e)^{-1}$ term multiplying the parenthesis determines the fractional sensitivity of *PE* to temperature changes, assuming constant relative humidity and wind speed. We can take the derivative with respect to temperature and show that,

$$\frac{\partial}{\partial T}(1+B_e)^{-1} = \left(\frac{B_e}{(1+B_e)^2}\right)\left(C - \frac{2}{T}\right) \tag{5.17}$$

where $C = (L/R_v)(1/T^2)$. B_e decreases rapidly with temperature and is one at about 278 K (Fig. 4.10), so that the first part of (5.17) is about 0.5 at about 278 K and is smaller for both colder and warmer temperatures. At very cold temperatures, there is so little water vapor in the atmosphere that its changes have little influence. At warm temperatures, $B_e << 1$ and all of the available energy at the surface is already being used to evaporate water. The second part $(C - 2/T)$ also decreases with increasing temperature, but the first part dominates the sensitivity of PE to temperature. The magnitude of the net effect is that the sensitivity of PE to temperature inferred from (5.17) decreases by about a factor of two between 0°C and 30°C, implying that PE is more sensitive to temperature in middle and high latitudes than in low latitudes. This is because at low temperatures the contribution to the total surface cooling from evaporation can increase at approximately the Clausius–Clapeyron rate, whereas at high temperatures virtually all the surface cooling is already being done by evaporation, which is constrained by the supply of energy to the surface (Scheff and Frierson, 2014). Since the supply of energy to the surface does not change rapidly with climate change, the potential evaporation does not change as rapidly in warmer latitudes as it does in colder latitudes where surface cooling can be shifted from sensible to latent cooling.

5.6 ANNUAL VARIATION OF THE TERRESTRIAL WATER BALANCE

The annual variation of the surface water balance at a location is intimately related to the local climate and its potential for human habitation and agriculture. The natural vegetation is adapted to the normal cycle of water surplus and water shortage that a region experiences. The annual variation of the water balance can be used as one means of classifying climates (Thornthwaite, 1948). The water balance depends on the annual variation of precipitation and evaporation, which together largely determine the soil moisture.

The accuracy with which the water balance can be determined depends on the amount and quality of the data available. Ideally, one would require good measurements of evaporation, precipitation, and soil moisture. However, evaporation and soil moisture are not routinely measured at most locations. Most climatological stations report surface air temperature and humidity, precipitation, and wind speed. With these variables, potential evaporation can be estimated from semi-empirical formulas such as (5.12), (5.20), or their simplified forms. If a soil moisture balance is calculated using the bucket model, then the actual evapotranspiration can be estimated from the potential evapotranspiration using (5.21). Figure 5.6 shows results of a simple water balance analysis for a variety of locations. These use daily ERA interim reanalysis data sets from 1979 to 2010, so that the surface radiation, precipitation, and evaporation are generated by the model, rather than by direct observation, but the model meteorology is very close to the observed weather, and the seasonal variations of

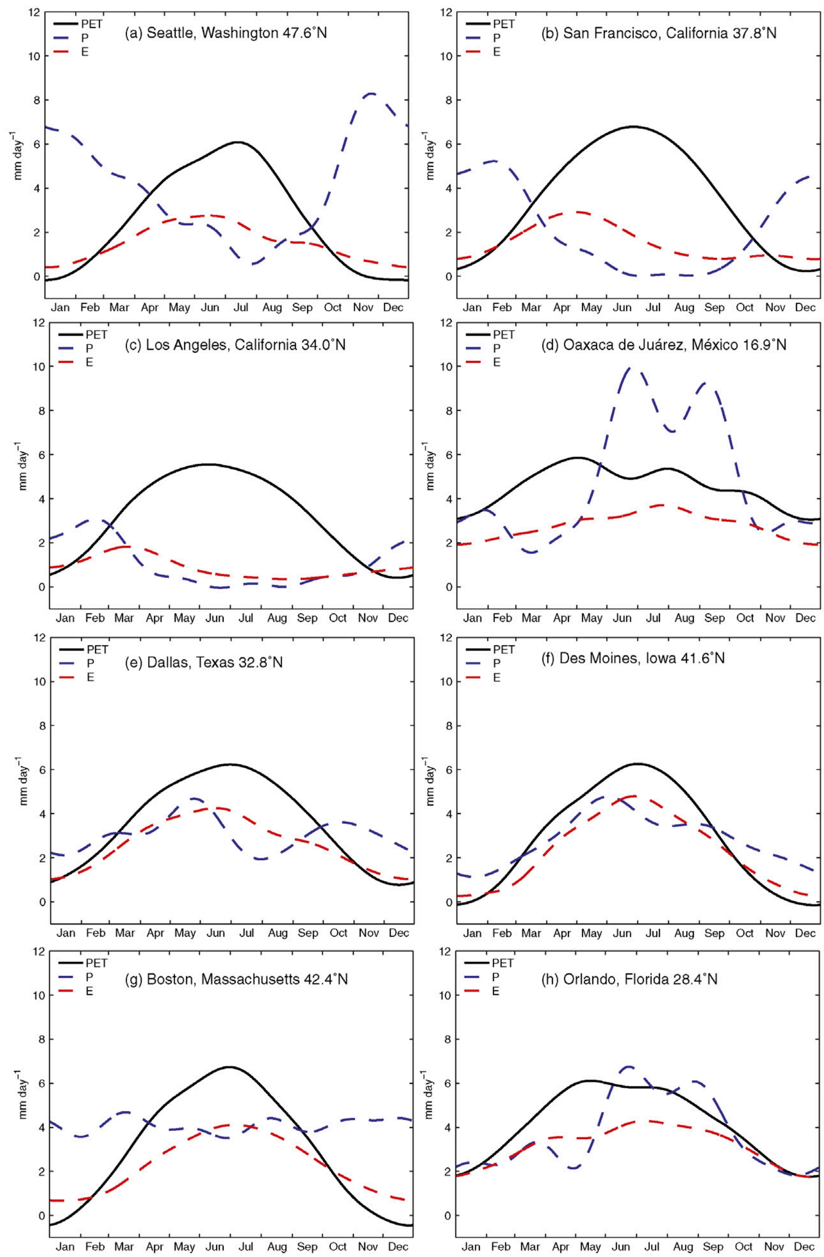

FIGURE 5.6 **Annual cycle of the water balance at twelve locations.** Data are from the ERA Interim Land Reanalysis data set for 1979–2010. Data are for the 0.75° box closest to the city indicated, and all data are from the model rather than direct observations. Potential evapotranspiration has been estimated simply as $R_s/\rho L$. The data have been smoothed.

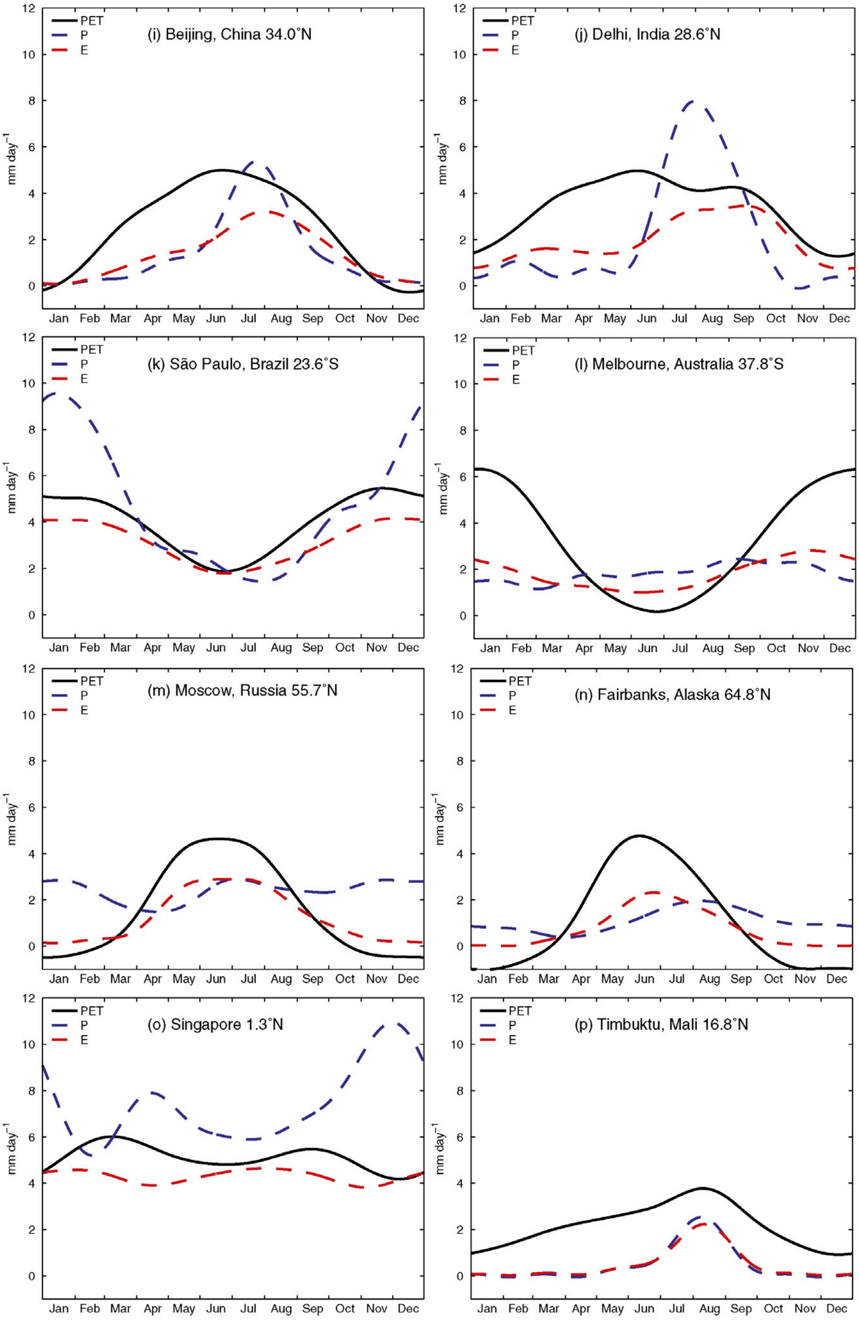

FIGURE 5.6 (*Continued*)

precipitation are very close to observations. Potential evapotranspiration (PET) is estimated from the net surface radiation alone as $\mathrm{PET} = R_s / \rho_w L$.

The west coast of North America in middle latitudes experiences a wintertime maximum in precipitation associated with winter storms (Fig. 5.6a–c). The wintertime peak of precipitation is smaller and occurs later in southern California than on the Pacific Coast of Canada. In Juneau, Alaska the monthly precipitation peaks in September or October, whereas in Los Angeles it does not peak until January or February. The summertime minimum of precipitation becomes deeper and of longer duration toward the south. Because the summer minimum of precipitation corresponds with the season of strongest insolation and warmest temperatures, the soil moisture can become depleted in the summer months. At Seattle, evapotranspiration (E) exceeds precipitation (P) beginning in May, and during summer the soil is sufficiently dry that PET greatly exceeds the actual evapotranspiration. This continues until October, when the precipitation again exceeds the evapotranspiration. In San Francisco, the dry season begins in April and extends until November. In Los Angeles, the soil is nearly always dry, and only during the months of December through February does the actual evapotranspiration approach PET. Oaxaca, Mexico is in the tropics (~17°N), and its precipitation maximum comes in the summer months, when tropical convection reaches northward from the equatorial region (Fig. 5.6d). During the heavy precipitation months of June through September, the precipitation exceeds PET. The remainder of the year is drier.

In much of the interior of North America the precipitation peaks in the spring or early summer when the soil is moist and temperatures are high enough to allow the air to carry large amounts of water vapor. As the summer season progresses, the soil dries out and the precipitation amount decreases. During middle and late summer PET generally exceeds the actual evaporation, indicating relatively dry soil. At Des Moines, Iowa the precipitation peaks in late May and continues into the middle and late summer. The combination of warmth, insolation, and precipitation during the summer in this part of the American Midwest makes it well suited for agriculture, especially corn. Dallas, Texas has a similar May maximum in precipitation, but the dip in summertime precipitation is greater, and the average difference between PET and precipitation is greater, indicating a drier climate.

In the northeastern United States, the monthly precipitation amount is almost independent of season. The frontal precipitation of winter is replaced in summer by more convective precipitation, such that the total precipitation remains almost constant. PET follows the insolation and temperature and peaks in the summer. During the summer months the evaporation exceeds the precipitation, causing the soil to dry out somewhat. In Boston, Massachusetts the potential evaporation exceeds the precipitation from

May to September (Fig. 5.6g). Farther south on the Florida Peninsula, Orlando shows a summertime maximum in precipitation, not unlike Oaxaca, with very heavy precipitation from June to September. This precipitation is mostly associated with thunderstorms, which are driven by solar heating of the land and begin during the hottest part of the day.

In general, four basic types of water balance cycles can be identified for the USA, although many locations show a combination of several types. The west coast of North America in middle latitudes has a winter precipitation maximum and summer dry period. The interior of the continent experiences a spring or summer rainfall maximum, followed by a drying period of varying intensity in the late summer. On the east coast of North America in mid-latitudes, the precipitation amount is almost independent of season, but the potential evaporation peaks in the summer, producing some reduction in soil moisture. In tropical or subtropical latitudes, the precipitation is very small in winter, but thunderstorms yield large amounts of rain during the warmest part of the summer.

Beijing, China, and Delhi, India have a monsoonal climate, with very dry conditions in winter and spring, then heavy summer rains starting in June (Fig. 5.6i,j). São Paulo, Brazil has a summertime precipitation maximum that exceeds the potential evapotranspiration and during the rest of the year the precipitation about equals the PET, implying a wet climate. Melbourne, Australia has nearly constant precipitation through the year, which is less than PET, except in winter.

In high latitudes, the low saturation vapor pressure associated with the relatively cold temperature constrains the rates of evaporation and precipitation. At many locations, the precipitation is greater than PET during most of the year. Water evaporated at warmer latitudes is transported into high latitudes by atmospheric motions and is precipitated when large-scale motions drive saturated air upward. The energy available at the surface is insufficient to allow the evaporation of this water. The growing season is short, so that the vegetation is not especially effective in bringing water from the soil to the atmosphere. Therefore, soils in high latitudes typically have a high water content. At very high latitudes, this water is mostly frozen. In high latitudes, it is common for the precipitation to exceed PET except during the summer, as in Moscow, Russia and Fairbanks, Alaska (Fig. 5.6m,n).

Singapore is near the equator in the Maritime Continent region, where the precipitation exceeds PET for most of the year and the annual variation of PET is small (Fig. 5.6o). Timbuktu, Mali is in the Sahel region of Africa, which gets precipitation only in the summer for a brief period, but even then the precipitation does not match the PET (Fig. 5.6p). PET is relatively low for such a tropical location because of the high albedo of the desert, and because the surface gets hot and loses heat through longwave emission during the daytime.

5.7 MODELING THE LAND SURFACE WATER BALANCE

The water balance of the surface is intimately coupled with the surface energy balance. Over water surfaces, the joint energy–water balance problem is simplified because one can assume that the air at the surface is saturated, and the storage and retrieval of water from below the surface is not an issue. Over land surfaces, the heat and water balances are very sensitive to the amount of water below the surface and the rate at which it can be brought to the surface and evaporated or transpired through plants. Transpiration through plants is dependent not simply on the soil moisture and atmospheric conditions, but also on the physiological state of the plant cover. Modeling and understanding of the exchanges of energy, water, and carbon dioxide at the surface are critically important for understanding climate variability and change. Vegetation on the land (land cover) is an important mediator of these exchanges, is important in its own right for agricultural and other uses, and will evolve to adapt to changing climate (land cover change).

5.7.1 The Bucket Model of Land Hydrology

The simplest model for the soil water budget is the bucket model. The soil is assumed to have a fixed capacity to store water that is available for evapotranspiration. The rate of change of the mass of water in the soil per unit area w_w, is determined by the rainfall rate P_r, the evapotranspiration rate E, the melting of snow M_s, and the runoff rate Δf.

$$\frac{\partial w_w}{\partial t} = \rho_w \frac{\partial h_w}{\partial t} = P_r - E + M_s - \Delta f \qquad (5.18)$$

The amount of available water in the soil can be expressed as an equivalent depth h_w, using a standard water density ρ_w. In the bucket model, the soil is assumed to have a fixed capacity to store moisture, corresponding to an equivalent water depth, h_c, which would typically be about 15 cm. If the soil moisture equals the capacity of the soil, then the soil is saturated. If the sum of rainfall plus snowmelt exceeds evaporation when the soil is saturated, then runoff at a rate just sufficient to keep the soil saturated is predicted.

To complete the soil moisture balance model for regions with snowfall, a separate budget for snowcover must be retained. If precipitation occurs when the surface temperature is below freezing P_s, it can be assumed to result in surface snow cover. The snow cover can be measured in terms of its water mass per unit area w_s, or an equivalent depth of water h_s.

$$\frac{\partial w_s}{\partial t} = \rho_w \frac{\partial h_s}{\partial t} = P_s - E_{sub} - M_s \qquad (5.19)$$

The maximum carrying capacity of the surface for snow or ice is determined by the lateral flow of ice sheets and does not become a factor until the ice is several hundred meters thick. Snow cover is removed by sublimation, E_{sub}, or melting. The snow cover lies on top of the soil and does not enter into the soil moisture balance unless it melts. Melting occurs when the surface temperature rises to the freezing point of water. The latent heat of fusion must be supplied to the surface energy balance when melting occurs. Melting continues at the rate necessary to keep the surface temperature from rising above 0°C until the temperature falls below freezing or the snow cover is completely removed.

The rate of evaporation depends on the soil moisture. The soil moisture can be used to relate the actual evaporation to the potential evaporation – the evaporation that would occur if the surface were wet. If measurements of air humidity, air temperature, wind speed, and surface temperature are available, the bulk aerodynamic formula can be used to calculate PET.

$$\mathrm{PET} = \rho_a C_{DE} U (q^*(T_s) - q_a) \tag{5.20}$$

If insufficient data to evaluate (5.20) are available, then another approximate formula can be used to estimate PET.

The actual evapotranspiration may be related to the potential evaporation and the soil moisture content.

$$E = \beta_E \,\mathrm{PET} \tag{5.21}$$

Healthy vegetation may transpire at the rate of potential evaporation, even when the soil is not saturated. When the soil moisture falls below a certain level h_v, the vegetation will no longer transpire at the potential rate. For soil moisture availability less than h_v, it is simplest to assume that β_E varies linearly between zero and one.

$$\beta_E = \begin{cases} 1.0, & h_w \geq h_v \\ \left(\dfrac{h_w}{h_v}\right), & 0 < h_w < h_v \end{cases} \tag{5.22}$$

The simple bucket model can easily be elaborated by adding a deep layer that exchanges water with the upper layer at a slow rate depending on the relative saturation of the two layers. This allows the soil water zone to be replenished with moisture from below without the occurrence of precipitation. In this case, an additional budget equation for the deep layer is required, and a term describing the exchange with the deep layer must be added to the soil-moisture equation (5.19). A thin layer near the surface can also be added to allow better treatment of short time scales associated with rainstorms or diurnal variations.

5.7.2 More Elaborate Models of Land Surface Processes

To improve significantly on the bucket model of land-surface hydrology, much more complex models must be introduced that describe the interactions of the atmosphere with vegetation and soil. Such models must fully couple the momentum, heat, and moisture budgets near the surface and describe each with compatible levels of sophistication and detail. The processes that must be considered include those illustrated in Fig. 5.5. Plants play a central role in the momentum, energy, and moisture transfers at the surface, and must be included in a model. Plants have effects on boundary processes through their physical properties and biological processes, and they have the ability to move water through their leaves at a rapid rate to facilitate photosynthesis when water is available. However, in times of water stress, plants can reduce their transpiration rate by closing their stomata. More advanced models of land–atmosphere interaction also seek to predict the exchange of carbon between the land surface and the atmosphere and to predict land cover change as the vegetation and soil co-evolve with a changing climate.

The rate at which plants transpire water depends on the availability of photosynthetically active radiation, temperature, air humidity, the availability of water within the plant, and the physiological state of the plant. The rate of water movement through plants is limited by the vapor phase in the leaves, rather than by the uptake of liquid water in the roots. The vapor pressure gradients that drive transpiration are strongly related to leaf temperature. For this and other reasons, leaf temperature is important to model, which requires calculation of the energy transfer through the plant canopy and a heat budget for the leaf structure of the plant. The transfer of solar radiation through the plant canopy is important, since it determines the distribution of heat input throughout the plant canopy and at the soil surface. Much of the insolation will be absorbed by the vegetative cover, rather than by the underlying soil. Because plants and other natural surfaces have very different albedos for visible and near-infrared radiation, these two frequency bands of solar radiation must be treated separately in accurate calculations. The obstacle to free airflow presented by the physical structure of the plant canopy affects the ventilation within the canopy, which is important to the turbulent fluxes of momentum, heat, moisture and carbon dioxide. The similarity hypotheses and resulting velocity profiles that lead to the bulk aerodynamic formulas are not valid within the plant canopy. Mean wind speeds and turbulent kinetic energy are much smaller within the canopy than just above it. The air properties within the plant canopy and at the soil surface can be very different from those near the top of the canopy, which is in direct contact with the atmosphere.

In addition to drawing moisture from the soil, plants can also intercept a substantial amount of precipitated water and store it on leaves and stems. The effectiveness of a plant in intercepting rainfall is dependent on the leaf structure, leaf orientation, the leaf area per unit of surface area, and on the frequency and intensity of precipitation. Tall vegetation with a large leaf area index can greatly decrease the supply of moisture to the soil by intercepting precipitation and facilitating its re-evaporation before it reaches the ground. Interception losses of this nature can range from 15% to 40% of precipitation for coniferous forests and from 10% to 25% for deciduous forests in mid-latitudes. Interception loss is greater if the precipitation rate is low or is intermittent, and lesser if the precipitation rate is high or continuous. To model interception and storage on leaves, budgets must be calculated for the amount of water stored on the surfaces of leaves. The removal of this surface water by evaporation depends on the supply of energy and unsaturated air at each level within the plant canopy.

The soil is the main reservoir from which evapotranspiration is drawn. Three layers within the soil may be distinguished by their interaction with the atmosphere. A thin layer very near the surface determines the interaction of the atmosphere with the bare ground surface on the time scale of individual precipitation episodes. If this thin layer becomes saturated during a rain shower, then runoff may occur. If this layer becomes dry, then surface evaporation is very small and transpiration by plants becomes the only mechanism whereby water can be efficiently removed from the soil. Below the surface layer is a deeper layer in which the roots of the vegetation reside and draw moisture from the soil. Below the root zone is a still deeper layer to which moisture is carried by gravity if the soil is saturated, and from which moisture can be drawn by capillary action.

The ability of soil to hold moisture may be measured by its field capacity, which is defined as the maximum volume fraction of water that the soil can retain against gravity. Typical values for loam are about 30%, but field capacities range from 10% for sand to 55% for peat. If the volumetric water content of the soil falls below a certain level, plants are unable to draw moisture from the soil and will remain wilted at all times of day. This permanent wilting threshold is typically one-third to one-half of the field capacity. The soil moisture available to plants is the difference between the volumetric soil water fraction and the wilting threshold. The maximum available soil moisture is the difference between the field capacity and the wilting threshold, and is 15–20% for typical soils.

The hydraulic conductivities of most soils decrease rapidly as the soil dries out, so that conduction of water becomes very slow if the water content is much below field capacity. If the surface layer of the soil becomes very dry, then infiltration of subsequent rainfall may be inhibited. Similarly, the soil in the root zone may become very dry, whereas the soil moisture several centimeters below the deepest roots remains near field capacity. The amount of water that is available to the vegetation is thus approximately equal to the available volumetric soil water fraction times the rooting depth. If the roots extend down about 1 m, and the available field capacity is about 15%, then the total amount of water available to plants when the soil is saturated is equivalent to about 15-cm depth of water. This may be greater or less depending on the type of soil and vegetation. Plants that live in sandy soil tend to develop deep roots, which offsets the low volumetric fraction of available water in sand.

To generalize the bulk aerodynamic formulas (4.26 and 4.27) to systems with plant canopies, they can be written in the following way.

$$SH = c_p \rho \frac{(T_s - T_a)}{r_h} \quad LE = L \rho \frac{(q_s - q_a)}{r_v} \tag{5.23}$$

In (5.23), r_h and r_v represent resistances to the transport of sensible and latent heat from the surface to the air above the plant canopy in units of s m^{-1}. Separate resistances can be specified for the plant canopy, the surrounding ground, and the air boundary above the plant canopy as in Fig. 5.7. Similar expressions could be written for carbon dioxide or other trace gases. One objective of land-surface modeling is to calculate values for these resistances that would cause a climate model to behave like nature.

Fig. 5.7 shows a schematic of the bucket model described previously along with a very generalized schematic of a more complex land-surface model including a plant canopy. The fluxes of sensible and latent heat that arise from the canopy model are a combination of what might escape directly from the soil and what is transmitted from the canopy. The great innovation of the more advanced land-surface models is to incorporate the biological effects of the plant canopy in controlling the exchanges of moisture and heat. The canopy has its own temperature and other variable

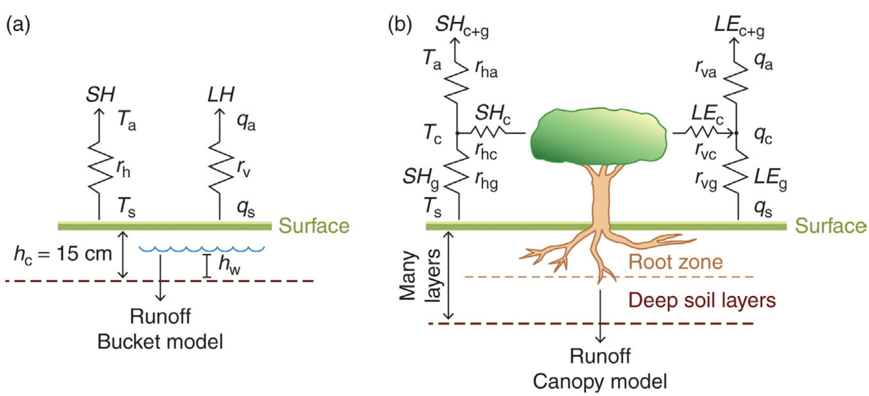

FIGURE 5.7 **Schematic diagrams of the bucket model and the canopy model of the land surface.** Subscripts g, c and a indicate values for the ground, canopy and air above the canopy. *Adapted from Sellers et al., 1997 and Pitman, 2003.*

properties that control the resistance to the transport of moisture and heat from the surface to the atmosphere. Variable stomatal-resistance models take into account the ability of plants to open or close their leaf pores depending on the condition of the plant and its environment. Further advances are the capability to model the response carbon uptake and release to climate conditions and CO_2 concentration. Even more advanced land-surface models can simulate change in the plant canopy and soil in response to changing climate conditions, such as the conversion of tundra to forests or the conversions of forest to grassland.

EXERCISES

1. The approximate volume of water retained in soil moisture and groundwater is given in Table 5.1. Use the data in Fig. 5.1 to calculate the time it would take for precipitation over land to deliver an amount of water equal to the soil water and groundwater. How long would it take to replace the groundwater and soil moisture if only 10% of the runoff could be redirected to replenishing the groundwater?
2. Use the bulk aerodynamic formula (4.32) to calculate the evaporation rate from the ocean, assuming that $C_{DE} = 10^{-3}$, $U = 5\ \text{m s}^{-1}$, and that the reference-level air temperature is always 2°C less than the sea surface temperature. Calculate the evaporation rate for the following:
   a. $T_s = 0°\text{C}, q_s^* = 3.75\,\text{g kg}^{-1}, \text{RH} = 50\%$
   b. $T_s = 0°\text{C}, q_s^* = 3.75\,\text{g kg}^{-1}, \text{RH} = 100\%$
   c. $T_s = 30°\text{C}, q_s^* = 27\,\text{g kg}^{-1}, \text{RH} = 50\%$
   d. $T_s = 30°\text{C}, q_s^* = 27\,\text{g kg}^{-1}, \text{RH} = 100\%$
   Assume a fixed air density of 1.2 kg m^{-3}. How would you evaluate the importance of relative humidity versus the importance of surface temperature for determining the evaporation rate?
3. Calculate the Bowen ratio using the bulk aerodynamic formulas for surface temperatures of 0, 15, and 30°C, if the relative humidity of the air at the reference level is 70% and the air–sea temperature difference is 2°C. Assume that the transfer coefficients for heat and moisture are equal.
4. Use the results of problem 3 to explain why high-latitude land areas often have high surface-moisture content.
5. Why is local winter and spring snow accumulation important for the summer soil moisture of mid-latitude continental land areas? How do you think the August climate would change if the winter and spring snowfall were replaced by rain showers?
6. What are some of the shortcomings of the bucket model of land hydrology? How are these limitations addressed by more sophisticated models for land-surface processes?
7. Derive (5.12) using the method outlined in the text.