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 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 movement of water among the reservoirs of ocean, atmosphere, and land is called the hydrologic cycle. 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. To evaporate 1 m of water in a year requires an average energy input of 80 W m^{-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 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

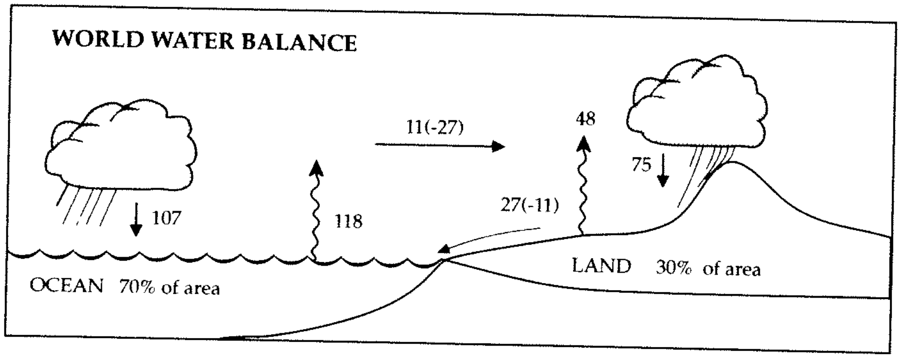

Fig. 5.1 Schematic diagram showing the basic fluxes of water in the global hydrologic cycle. Units are centimeters per year spread over the area of the land or ocean. Since the areas of land and ocean are different, the land–ocean water exchanges by atmospheric transport and river runoff have different 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
Groundwater, soil moisture	8.062	0.580[a]
Lakes and rivers	0.225	0.020
Atmosphere	0.013	0.001
Total water amount	1384.0	100.0
Freshwater	36.00	2.60
Freshwater reservoirs as a percent of total freshwater		
Polar ice caps, icebergs, glaciers		77.2
Groundwater to 800-m depth		9.8[a]
Groundwater 800–4000-m depth		12.3[a]
Soil moisture		0.17[a]
Lakes (freshwater)		0.35
Rivers		0.003
Hydrated earth minerals		0.001
Plants, animals, humans		0.003
Atmosphere		0.040
Sum		100.000

[From Baumgartner and Reichel (1975).]

[a] Numbers uncertain.

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 from Earth's surface accounts for about half of the cooling of the surface that balances the heating by absorption of solar radiation. 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

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

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. Also, 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.

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

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 Earth–atmosphere system in which the exchange of water across the surface does not appear.

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

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

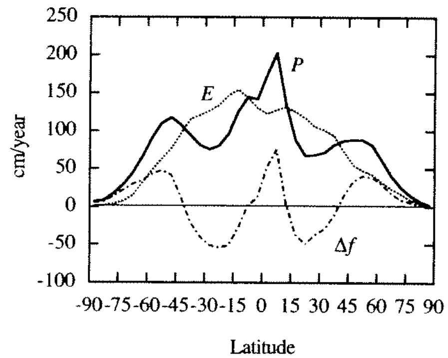

Fig. 5.2 Latitudinal distribution of the surface hydrologic balance, showing evaporation E, precipitation P, and runoff Δf. [Data from Baumgartner and Reichel (1975).]

and opposite to the net horizontal transport below the surface. This means that water carried to continents by atmospheric transport must equal the runoff from rivers.

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. Moisture-laden air near the surface flows toward the equator from both hemispheres and converges near the equator, where it is 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 of that region. In these latitudes cyclonic disturbances with strong winds drive vertical motions that release water. Evaporation varies more smoothly 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 degrees of latitude, and these regions export water vapor to be condensed in the latitudes where the precipitation maxima occur. The runoff 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. 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 water to sink to the bottom of the

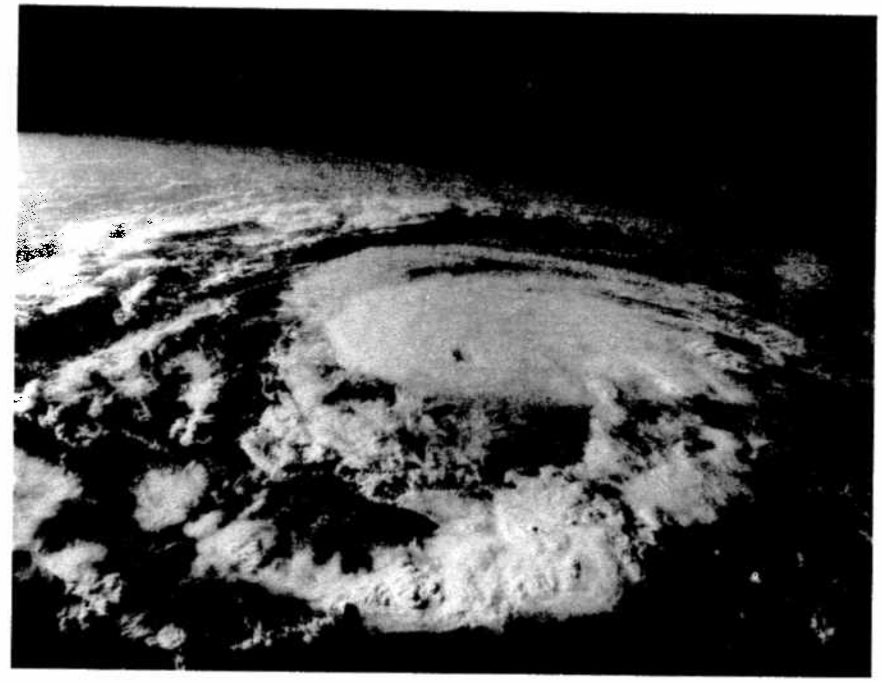

Fig. 5.3 Hurricane Bonnie located about 500 miles from Bermuda on 19 September 1992. (Photo credit: NASA.)

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).]

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 is typically 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 snowcover. 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 midlatitude 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 midlatitude 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 does bare soil 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. 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 ground-

water. 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.

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 reevaporated, 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 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, and the droplets or particles reach the surface without reevaporation. Supersaturation is normally caused by cooling of air parcels during ascent. Ascent of air parcels can be forced by atmospheric motions associated with midlatitude frontal and synoptic weather systems. 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.

The geographic distribution of precipitation is shown in Fig. 5.4. The general features of the 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. In the subtropics convection and precipitation are suppressed by the downward mean air motion that characterizes this region. In these latitudes precipitation is at a minimum, but high surface insolation and subsidence of dry air give rise to very strong evaporation. In midlatitudes precipitation increases again because of midlatitude synoptic storm systems. The forced ascent of moist surface air in midlatitude weather systems and the westerly flow over obstacles such as the Rocky Mountains give rise to heavy precipitation. In the polar regions precipitation decreases. The entire hydrologic cycle is slowed down in polar regions because of the low temperatures and consequently low water-carrying capacity of the atmosphere.

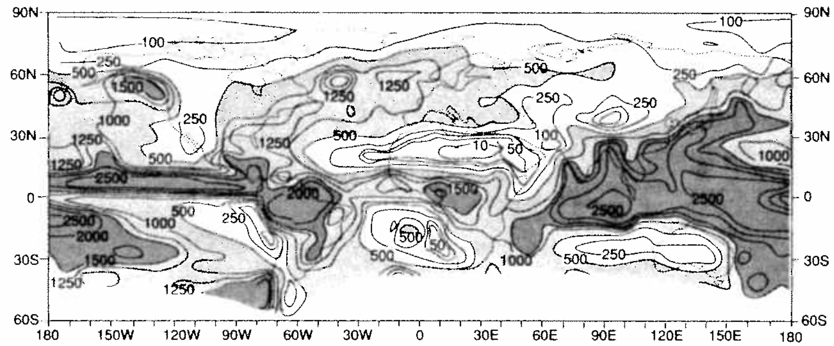

Fig. 5.4 Geographic distribution of annual mean precipitation in mm. [After Shea (1986). Reprinted with permission from the National Center of Atmospheric Research.]

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 or fall directly onto the soil.

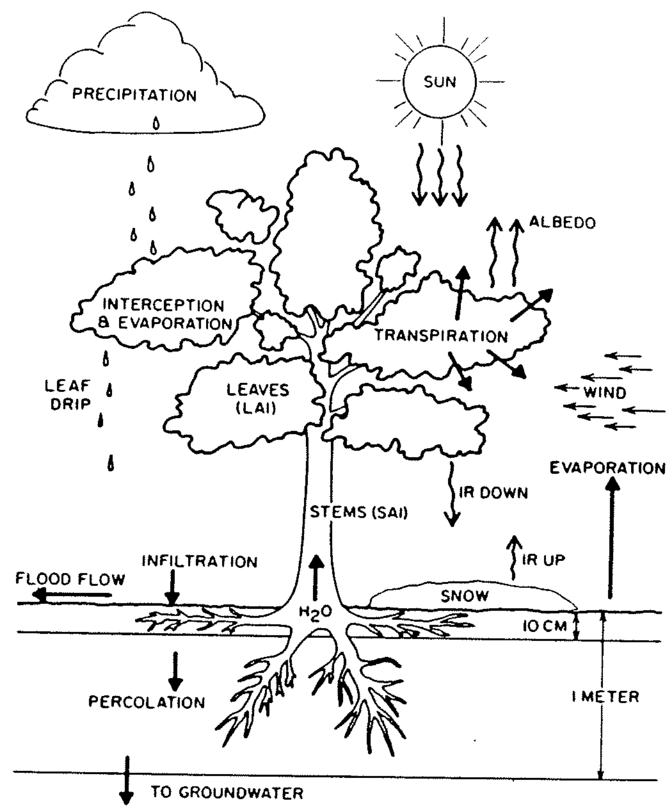

Fig. 5.5 Diagram showing the effects of the vegetation canopy on the water and energy fluxes. [From Dickinson (1984). © American Geophysical Union.]

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 top sides 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 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.[1] To obtain accurate results, the lysimeter must be large enough to contain the soil water zone and associated vegetation. The lysimeter 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. For example, a lysimeter containing a depth of growing grass and soil can be buried in a large grass field so that the grass in the lysimeter experiences exactly the same conditions as the adjacent grass outside the lysimeter. For environments where the vegetation is large and has a substantial root structure, such as a forest, the construction of a lysimeter for even a single tree is a considerable challenge.

Evapotranspiration can also be estimated by 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. It is also a challenge to obtain representative spatial and temporal means, particularly if the surface characteristics are spatially inhomogenous.

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

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

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 measure-

[1]Brutsaert (1982).

ment 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).

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

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 problematical 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

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

where

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

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

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

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.

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

The evaporating capacity of the air is defined by

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

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 evapotranspiration that would occur if the surface was wet, and is therefore the maximum possible evapotranspiration for the prevailing atmospheric conditions. It measures the effect of energy supply and air humidity on the evapotranspiration 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. Potential evaporation can be calculated using a variety of

theoretical and empirical techniques. One method would be to calculate it 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.

5.6 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 the air at the surface can always be assumed to be 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.

5.6.1 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.

$$\frac{\partial w_w}{\partial t} = \rho_w \frac{\partial h_w}{\partial t} = P_r - E + M_s - \Delta f \tag{5.14}$$

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 assumed to be 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 snowcover. The snowcover can be measured in terms of its water mass per unit area w_s, or an equivalent depth of water h_s.

$$\frac{\partial w_s}{\partial t} = \rho_w \frac{\partial h_s}{\partial t} = P_s - E_{\text{sub}} - M_s \tag{5.15}$$

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. Snowcover is removed by sublimation, E_{sub}, or melting. The snowcover 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 snowcover 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 potential evaporation.

$$\mathrm{PE} = \rho_a\, C_{\mathrm{DE}}\, U \left(q^*(T_s) - q_a\right) \tag{5.16}$$

If insufficient data to evaluate (5.16) are available, then another approximate formula can be used to estimate PE.

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

$$E = \beta_E \bullet \mathrm{PE} \tag{5.17}$$

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.

$$\beta_E = \begin{cases} 1.0, & h_w \geq h_v \\ \left(\dfrac{h_w}{h_v}\right), & 0 < h_w < h_v \end{cases} \tag{5.18}$$

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 (5.14). A thin layer near the surface can also be added to allow better treatment of short time scales associated with rainstorms or diurnal variations.

5.6.2 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.

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, and moisture. 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 reevaporation before it reaches the ground. Interception losses of this nature can range from 15 to 40% of precipitation for coniferous forests and 10–25% for deciduous forests in midlatitudes. Interception loss is greater if the precipitation rate is low or is intermittent, and less 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 deeper layer to which moisture is carried by gravity if the soil is saturated, and from which moisture can be drawn by capillary action.[2]

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, while 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 one meter, 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.

5.7 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

[2]Bras (1990).

means of classifying climates.[3] 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. Evaporation and soil moisture are not routinely measured at most locations, however. Most climatological stations report surface air temperature and humidity, precipitation, and wind speed. With these variables potential evaporation can be estimated from semiempirical formulas like (5.12), (5.16), 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 (5.17). Figure 5.6 shows results of such a water balance analysis for a variety of locations.

The west coast of North America in middle latitudes experiences a wintertime maximum in precipitation associated with winter storms [Fig. 5.6(a–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 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 exceeds precipitation beginning in May, and by July the soil is sufficiently dry that the potential evaporation exceeds the actual evapotranspiration. This continues until October, when the precipitation again exceeds the evapotranspiration. In San Francisco the dry season begins in June 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 the potential evaporation. Acapulco, 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. (5.6(d)]. During the heavy precipitation months of June through October, the precipitation exceeds the potential evaporation. The remainder of the year is very dry. Because of the relatively constant temperature and humidity in the tropics, the potential evaporation varies little during the year.

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 [Fig. 5.6(e–h)]. As the summer season progresses, the soil dries out and the precipitation amount decreases. During middle and late summer the potential evaporation generally exceeds the actual evaporation, indicating relatively dry soil. At Sioux City, Iowa the precipitation peaks in June 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

[3]Thornthwaite (1948).

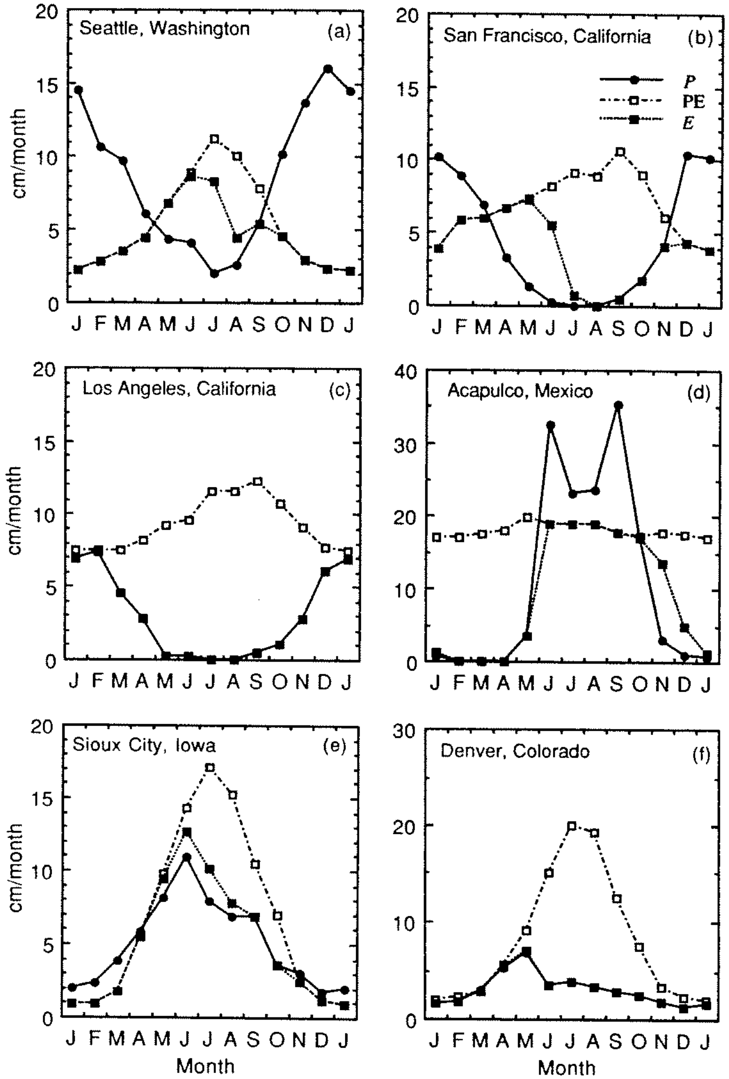

Fig. 5.6 Annual cycle of the water balance at various locations. Where precipitation does not appear, it is equal to evaporation (e.g., Los Angeles, Phoenix). Where potential evaporation does not appear, it is equal to evaporation (e.g., Churchill). [Data from Eagleman (1976).]

well suited for agriculture. Farther west and at the higher elevation of Denver, the precipitation is smaller and peaks earlier in the year. The air is dry and warm in the summer months and the potential evaporation greatly exceeds precipitation. At Phoenix, Arizona the precipitation is small at all times of year compared to the high

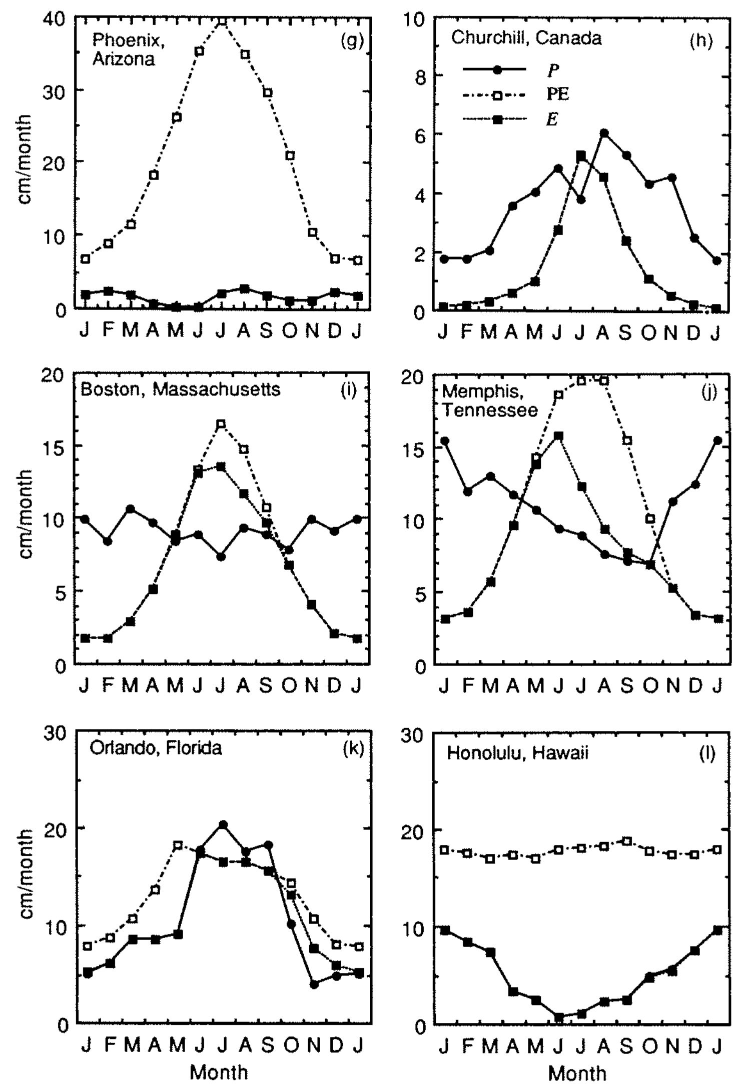

Fig. 5.6—*Continued*

drying capacity of the warm, dry air there. The potential evaporation greatly exceeds the precipitation at all seasons, indicating a desert climate.

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 the potential evaporation 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. Soils in high latitudes therefore typically have a high water content. At very high latitudes this water is mostly frozen. Churchill, Canada (~59°N) is an example of a cold continental climate where the evapotranspiration is always energy limited and exceeds the precipitation only during July [Fig. 5.6(h)].

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. The potential evapotranspiration 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 actual evapotranspiration from July to September [Fig. 5.6(i)]. In the southeastern United States, as characterized by Memphis, Tennessee, a modest wintertime peak in precipitation is observed. Farther south on the Florida peninsula, Orlando shows a summertime maximum in precipitation, 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. Honolulu is in the subtropics where potential evaporation exceeds precipitation at all seasons [Fig. 5.6(l)]. Precipitation peaks in winter and is balanced by evaporation in all seasons.

In general there are four basic types of water balance cycles, 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 midlatitudes 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.

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 $C_{DE} = 10^{-3}$, $U = 5$ 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 (a) $T_s = 0°C$, $q_s^* = 3.75$ g kg^{-1}, RH = 50%; (b) $T_s = 0°C$, $q_s^* = 3.75$ g kg^{-1}, RH = 100%; (c) $T_s = 30°C$, $q_s^* = 27$ g kg^{-1}, RH = 50%; and (d) $T_s = 30°C$, $q_s^* = 27$ g kg^{-1}, RH = 100%. Assume a fixed 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 midlatitude continental land areas? How do you think the August climate would change if the winter and spring snowfall were replaced by rainshowers?
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.