# 10

# Water Balance

## 10.1 | Chapter Summary

This chapter introduces the hydrologic cycle on land. The overall hydrologic cycle is reviewed and simplified into an expression where the change in soil moisture is the balance between precipitation input and losses from evapotranspiration and runoff. The various terms in this equation are discussed. Some precipitation is intercepted by the plant canopy and evaporates back to the atmosphere. Evaporation is the physical process by which liquid water changes to vapor. Transpiration is the evaporation of water held internal to plants as water moves from the soil through plants into the atmosphere along a continuum of decreasing water potential. The basic meteorological and biological processes controlling evapotranspiration are introduced, but Chapters 12 and 13 provide a more in-depth discussion of meteorological processes, and Chapters 15–17 review the biological control of transpiration by leaves and plant canopies. Runoff is discussed in separate sections on infiltration and overland flow. The section on overland flow reviews the effects of soil texture, soil water, and land cover on runoff. A simple model of the water balance illustrates geographic patterns of water availability.

## 10.2 | Cycling of Water on Land

The cycling of water over land depicted in Figure 10.1 involves much more detail than that shown in Figure 3.4. Solar energy drives the hydrologic cycle, evaporating water from soil, lakes, and rivers. When conditions are favorable, this water vapor condenses to form clouds and eventually *precipitation*, which replenishes soil water and renews the cycle. Precipitation occurs when air cools and water vapor condenses to form cloud droplets or ice crystals. Cooling is caused by air rising in altitude, such as when air is lifted over mountains. This type of precipitation is known as orographic precipitation. Precipitation also occurs through frontal convergence, when a warm air mass rises over a colder air mass. In summer, thunderstorms develop when strong solar radiation heats the ground, warming the surface air and causing the less dense air to rise. This type of rainfall is known as convective precipitation.

Not all of the precipitation reaches the ground. Leaves, twigs, and branches of plants intercept rain and snow. *Interception* is the process by which precipitation is temporarily stored on plant surfaces. This water quickly evaporates and never replenishes the soil. The water that is not intercepted falls to the ground as throughfall or stemflow. *Throughfall* is water reaching the ground directly through openings in the plant canopy or by dripping down from leaves, twigs, and branches. *Stemflow* is water that reaches the ground by flowing down plant stems and tree trunks.

In many regions, a significant portion of winter precipitation falls as *snow*. If temperatures are cold enough, this water accumulates and is stored for periods of hours, days, or months

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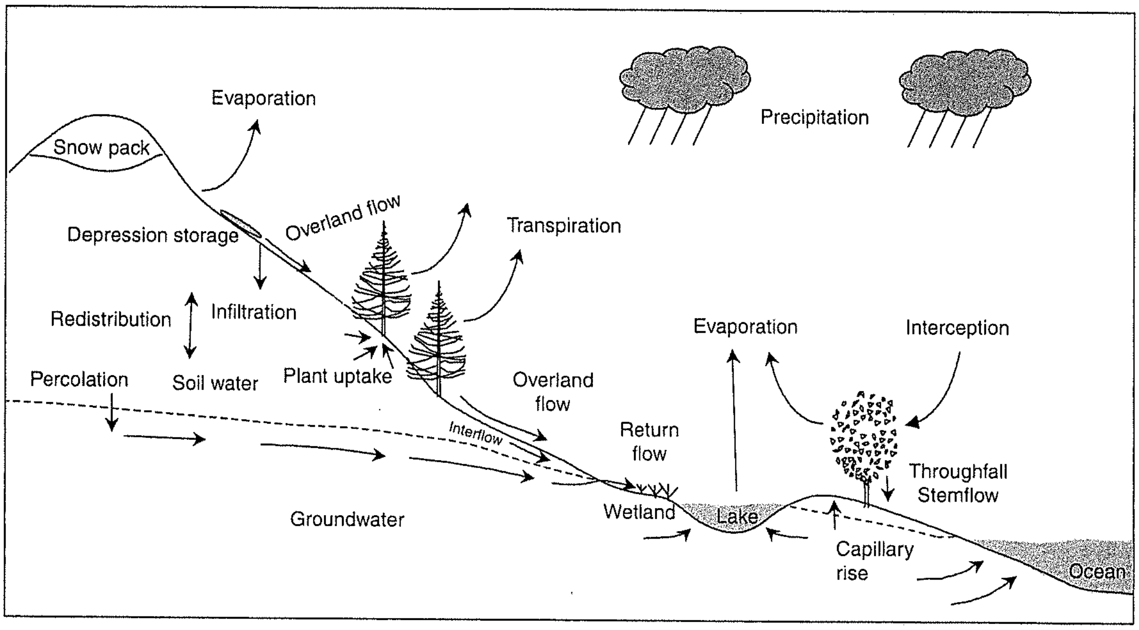


Fig. 10.1 The hydrologic cycle on land.

before melting. Winter storage of precipitation in snow and subsequent spring snowmelt is a large source of water in seasonally-cold climates.

Liquid water reaches the ground as rainfall where vegetation is absent, as throughfall and stemflow under vegetation, or from snowmelt. Some of this water infiltrates into the soil. *Infiltration* is the physical process by which water moves into the soil. When the infiltration capacity of soil is exceeded, water collects as puddles in small depressions (*depression storage*). When these are filled, water runs off over the ground surface as overland flow. *Overland flow* is runoff generated when the infiltration capacity of soil is exceeded by the rainfall intensity, resulting first in ponding of water on the soil surface and then flow across the surface. It moves downhill, first in small rills and gullies and then into creeks and streams that feed large rivers.

The water that infiltrates into the soil wets the soil and is stored as soil water. *Soil water* is water held in the unsaturated zone between the soil surface and the water table. Soil water returns to the atmosphere through evaporation from bare ground and transpiration from plants. *Evaporation* is the physical process by which water changes from a liquid to vapor in the air. *Transpiration* is evaporation of water held inside plants.

Within the soil, water is removed during evaporation and by plant roots (*plant uptake*) when plants replenish water lost during transpiration. Water also moves vertically and horizontally due to internal forces determined by how wet or dry the soil is and by gravity. In most cases, gravity is the greatest force, causing water to flow downwards. This movement is known as *redistribution*, or more commonly percolation. If the water movement is deep, the percolating water will *recharge* the groundwater. On very shallow soils underlain with impermeable material (e.g., bedrock), the infiltrating water may move downhill as subsurface interflow. *Interflow* is the lateral movement of water in upper soil layers. For most landscapes, interflow is not thought to be important relative to overland flow.

*Groundwater* is the subsurface region that is saturated with water. The top is defined by the water table, which separates the saturated and unsaturated zones, and the bottom is defined by an impermeable layer (e.g., bedrock). Water moves horizontally within these aquifers, typically at a rate of about one-half to one meter per day. This lateral water flow recharges rivers, lakes, wetlands, and oceans and provides the base flow to maintain riverflow in the absence of rainfall. When this flowing water reaches the surface, it is known as return flow. *Return flow* is the process by which groundwater re-emerges from the soil in a saturated area and flows downslope as overland flow. Groundwater is typically recharged by water percolating through the soil column. Water also flows upward, wetting the area of soil immediately above the water table (*capillary rise*). As a result, the upper boundary of groundwater (i.e., the water table) fluctuates seasonally as water enters and leaves the aquifer.

Insights to the hydrologic cycle can be gained by reducing the full cycle shown in Figure 10.1 to a more simple form, $\Delta S = P - E - R$. In this water budget, the change in water storage on land ($\Delta S$) is the difference between precipitation input ($P$), evapotranspiration loss to the atmosphere ($E$), and runoff to the oceans ($R$). The next sections examine the terms in this equation in more detail.

## 10.3 Interception and Throughfall

Precipitation falls on foliage, branches, and bark, where it collects and readily evaporates back to the atmosphere. The amount collected depends on rainfall intensity, frequency, and duration. Most plants have a small capacity to store water. Early studies of interception suggested that trees, shrubs, and grasses can store about 1.3 mm of liquid water on their foliage (Zinke 1967). A slightly larger amount of water (3.8 mm) can be stored as snow in trees. This storage capacity varies with the leaf area index ($L$) of the canopy. Studies find that storage capacity (mm) ranges from $0.15L$ to $0.3L$, and may be as large as $0.5L$ in some forests (Carlyle-Moses and Gash 2011). The interception capacity and amount of water held in the canopy has a significant effect on climate because it readily evaporates (Davies-Barnard et al. 2014).

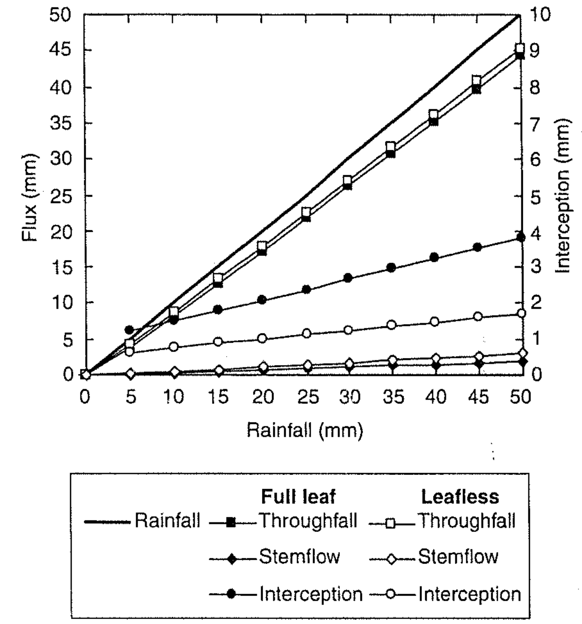


Fig. 10.2 Typical throughfall and stemflow (left axis) and interception (right axis) in relation to rainfall for a deciduous forest in full leaf and without leaves. Data from Zinke (1967). See also Helvey and Patric (1965).

During brief, moderate storms, this storage capacity may not be exceeded and most of the rainfall is intercepted. In long, intense storms, the storage capacity is quickly exceeded and water drips to the ground. Figure 10.2 shows typical relationships among rainfall amount and throughfall and stemflow. Stemflow is typically minor. Throughfall is the dominant component. For example, a deciduous forest in full leaf may allow 18 mm of rainfall to reach the ground during a 20 mm storm. Of this, 17 mm reaches the ground as throughfall and 1 mm is stemflow. The remaining water (2 mm) is intercepted. Intercepted water readily evaporates because it is held externally on foliage and wood. One study found that in pine forests the rate of evaporation of intercepted water was three times the rate of transpiration under the same radiation conditions (Stewart 1977).

Plant canopies generally intercept about 10–20 percent of annual precipitation (Carlyle-Moses and Gash 2011). Lowland tropical forests intercept, on average, 15 percent of annual precipitation (Krusche et al. 2011). Satellite-based estimates of interception loss range from 13 percent of annual rainfall for broadleaf evergreen forest to 19 percent for broadleaf deciduous forest and 22 percent for needleleaf forests (Miralles et al. 2010). One determinant of interception is vegetative cover; the denser the foliage the greater the amount of water stored. The seasonal emergence and senescence of leaves also matters. Deciduous forests intercept less rainfall when leafless than during the growing season when leaves are present (Figure 10.2). When in full leaf, grasses have the same storage capacity as trees, but because they have no woody material when dormant they intercept considerably less precipitation annually than trees. The type of leaf also determines the amount of interception. Broad-shaped leaves, such as found on deciduous trees, allow water droplets to run together, forming larger drops that readily drip off the leaf. Needle-shaped foliage, such as on coniferous trees, does not facilitate dripping because the separated small needles do not allow water to coalesce into large droplets.

## 10.4 | Evapotranspiration

Evaporation occurs when a moist surface is exposed to drier air. As air parcels move away from the surface, they carry with them moisture from the surface. Water evaporates from the surface, increasing the amount of water vapor in the surrounding air. When the air is saturated with water vapor, evaporation ceases. Transpiration is evaporation of water from plant leaves as it moves from the soil through plants and out through leaves to the air. Plants consume large amounts of water during growth. A field of corn covering 4000 m$^2$ can consume 10,000–15,000 liters of water (2.5–3.75 mm) in a day. A single well-watered tree can transpire 100–150 liters of water per day. Meteorological processes near the surface control evaporation and transpiration. Transpiration is also regulated by the physiology of plants. When plants cover a small portion of the surface, evaporation is the dominant flux. Transpiration becomes more important as plant cover increases. However, it is difficult to distinguish evaporation from transpiration, and the two terms are often combined into evapotranspiration.

Global evapotranspiration from Earth's land surface is about 550 mm of water per year (Jung et al. 2010). Other estimates range from 544 to 631 mm per year (Mueller et al. 2011). Transpiration accounts for 80–90 percent of global evapotranspiration (Jasechko et al. 2013). This water loss varies considerably among various biomes (Figure 10.3). Monthly evapotranspiration in tropical rainforest averages 2.5–3.5 mm per day and has little seasonal variation. Annual water loss is 1100 mm. Other biomes have comparable peak monthly rates, but strong seasonal variation yields much less annual water loss (200–500 mm). Seasonally cold biomes (temperate deciduous forest, boreal forest, and tundra) have a distinct annual cycle with high evapotranspiration during the warm summer months and low rates in the cold winter season. Temperate deciduous forest has peak monthly rates of 2–2.5 mm per day during the growing season. Boreal forest and tundra have lower maximum rates (1.5–2 mm per day) and a shorter growing season. Annual water loss declines from deciduous forest (380 mm) to boreal forest (230 mm) to tundra (170 mm). Ponderosa pine and grassland have distinct annual cycles related to precipitation. Peak rates in ponderosa pine (2.5 mm per day) and grassland (>2 mm per day) are comparable to other sites, but decline markedly in the dry season. Annual water loss is, however, similar to, or greater than, other sites (300 mm, grassland; 480 mm, ponderosa pine).

The rate of evapotranspiration depends on the availability of energy to evaporate water and the ability of water vapor to diffuse into the atmosphere. At 15°C, 2466 J are needed to change one gram of water to vapor (Table 3.3). Tropical climates, with a surplus of net radiation, have more energy to evaporate water than arctic climates. The capacity of air to remove water from the evaporating surface is also important. This is related to the humidity of the air. Dry air has a greater evaporative demand than humid air. It is also related to wind speed. As water evaporates

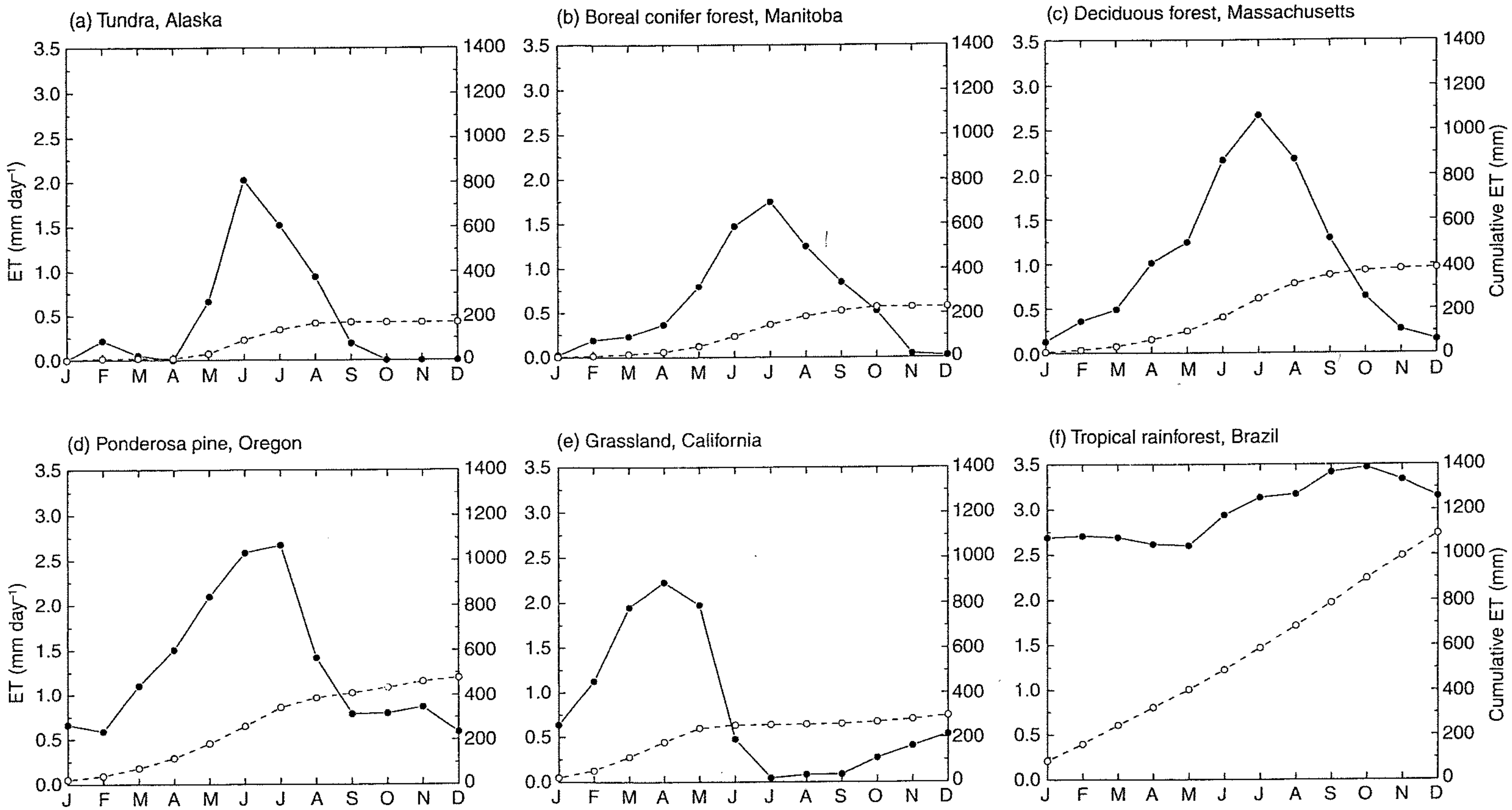


**Fig. 10.3** Monthly evapotranspiration (solid line, left axis, mm day$^{-1}$) and cumulative evapotranspiration (dashed line, right axis, mm) for (a) moist tundra, Alaska, (b) boreal conifer forest, Manitoba, (c) temperate deciduous forest, Massachusetts, (d) ponderosa pine, Oregon, (e) grassland, California, and (f) tropical rainforest, Brazil. See Figure 12.3 for monthly energy fluxes and site details.

from a surface, parcels of air near the surface become more humid. Under calm conditions, evapotranspiration decreases as the air becomes saturated with water vapor. With windy conditions, these parcels of air are carried away and replaced by less humid parcels.

The type of soil and its water content also regulate evapotranspiration. The rate of evaporation is determined by the rate at which water is supplied to the surface. An insufficient rate of soil water flow upward to the evaporating surface decreases the rate at which water can evaporate. The hydraulic properties of soil and the extent of drying determine the rate of replenishment (Chapter 9). A dry soil or a soil with low hydraulic conductivity provides less water for evapotranspiration than does a wet soil or one with high hydraulic conductivity.

The type of vegetation is also important. Leaves have microscopic pores called stomata that open to allow the plant to absorb $CO_2$ during photosynthesis. The plant cannot grow if stomata are not open, but when stomata are open, water inside the leaf diffuses out to the surrounding drier air during transpiration. If too much water is lost, the plant becomes desiccated and will die if its internal water is not replenished from water in the soil. Plants have evolved compromises between the need to open stomata to take up $CO_2$ and the need to close them to prevent water loss (Chapter 16).

Water lost from leaves during transpiration must be replenished from the soil. As transpiration increases during the day, water is first drawn from internal plant storage and then from soil near the roots. The movement of water from the soil through plants into the atmosphere occurs along a continuum of decreasing water potential. Water potential is a negative suction. The atmosphere exerts the most suction and has the lowest water potential. When wet, soil particles exert minimal suction on water and have a high water potential. Water flows from soil (on the order of −0.01 MPa when wet) into roots (−0.1 MPa) through the plant and out through foliage (−1 MPa) into the surrounding air (−100 MPa), moving from high to low water potential. At night, plant water uptake replenishes water depleted during the day. By morning, before transpiration begins, water in soil near the roots, water in plant storage, and water in foliage are again nearly equal in potential. The gradient in water potential re-establishes during the day as transpiration increases.

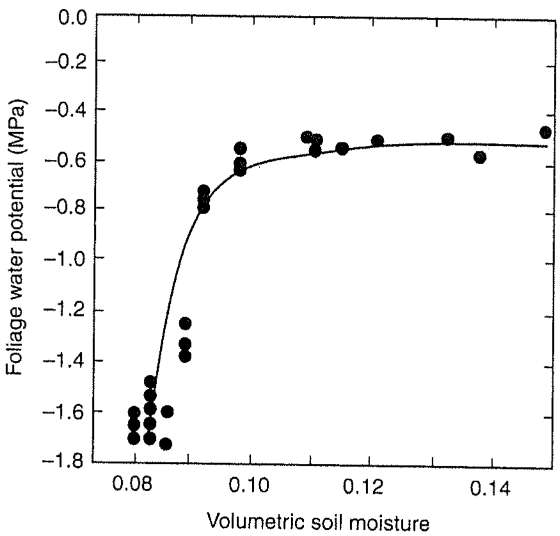


**Fig. 10.4** Pre-dawn foliage water potential in red pine trees in relation to volumetric soil water content for loamy sand. Volumetric water content is based on the depth of water in the upper 46 cm of soil. Adapted from Sucoff (1972).

As a plant extracts water from the soil, some critical water content is reached at which further decrease in soil water increases plant stress. In trees, water stress is seen in the pre-dawn water potential of foliage, which is a reliable indicator of soil moisture. Pre-dawn foliage water potential decreases as soil moisture drops below some threshold (Sucoff 1972; Hinckley and Ritchie 1973; Running et al. 1975). In the data shown in Figure 10.4, for example, pre-dawn foliage water potential of red pine trees growing on loamy sand is invariant of soil moisture at high water contents and decreases linearly with volumetric soil water less than about 0.1 $m^3\ m^{-3}$.

Several classes of models are used to estimate evapotranspiration (Fisher et al. 2011). One method uses air temperature as a surrogate for the energy available to evaporate water. The Thornthwaite equation exemplifies this approach (Thornthwaite 1948). In this formulation, monthly potential evapotranspiration ($E_p$, mm) is:

$$E_p = 16\left(\frac{L}{12}\right)\left(\frac{N}{30}\right)\left(\frac{10T}{I}\right)^a \tag{10.1}$$

where $L$ is daylength (hours), $N$ is the number of days in a month, $T$ is mean monthly air temperature (°C), $a$ is defined as:

$$a = 6.75\times10^{-7}I^3 - 7.71\times10^{-5}I^2 + 1.79\times10^{-2}I + 0.49 \tag{10.2}$$

and $I$ is summed for months with $T > 0$°C as:

$$I = \Sigma(T/5)^{1.514} \tag{10.3}$$

Evapotranspiration from Thornthwaite's method is a potential evapotranspiration because it does not account for the reduction in evapotranspiration as a result of soil drying. However, simple relationships can be used to decrease potential evapotranspiration to the extent that soil water is limiting. One approach is to assume that evapotranspiration proceeds at its potential rate until the soil is depleted of water. However, this ignores the tighter binding of water to soil particles as the soil dries. An alternative is to assume a linear decrease in evapotranspiration as the soil becomes drier, scaled to give the potential rate when the soil is fully wet and zero when the soil is dry.

Another class of models relates evapotranspiration to available energy, given by net radiation. The Priestley–Taylor equation exemplifies this type of model (Priestley and Taylor 1972). This equation relates potential evapotranspiration ($E_p$, mm day$^{-1}$) to net radiation ($R_n$) as:

$$E_p = \alpha\frac{s}{s+\gamma}\frac{R_n}{\lambda} \tag{10.4}$$

Dividing by the latent heat of vaporization ($\lambda$, MJ kg$^{-1}$) converts $R_n$ from an energy flux (MJ m$^{-2}$ day$^{-1}$) to a mass flux (kg m$^{-2}$ day$^{-1}$), equivalent to a depth of water (mm day$^{-1}$) because the density of water is 1000 kg m$^{-3}$ (1 kg m$^{-2}$/1000 kg m$^{-3}$ = 0.001 m). In this equation, $s$ (kPa K$^{-1}$) is the change in saturation vapor pressure with respect to temperature (Table 3.3) and $\gamma$ is the psychrometric constant (a representative value is 0.0665 kPa K$^{-1}$, Chapter 12). The coefficient $\alpha$ equals 1.26 for a wet surface, but is lower for vegetation. For example, $\alpha$ equals 0.82 in tropical and temperate broadleaf forests and equals 0.65 and 0.55 in temperate and boreal conifer forests, respectively (Komatsu 2005; Baldocchi and Ryu 2011).

The Penman equation is a combination equation that includes both available energy and diffusion (Penman 1948). As given by Shuttleworth (1993, 2007), potential evapotranspiration ($E_p$, mm day$^{-1}$) over open water is:

$$E_p = \frac{s}{s+\gamma}\frac{R_n}{\lambda} + \frac{\gamma}{s+\gamma}\frac{6.43(1+0.536u)D}{\lambda} \tag{10.5}$$

This equation is similar to the Priestley–Taylor equation, but additionally includes wind speed ($u$, m s$^{-1}$) and vapor pressure deficit ($D$, kPa). Evapotranspiration is a weighted linear combination of available energy and vapor pressure deficit. Evapotranspiration increases as more energy is available and as the atmospheric demand (i.e., vapor pressure deficit) increases, all other factors being equal. The Penman–Monteith equation (Chapter 12) is an extension of the Penman equation and illustrates the thermodynamic, aerodynamic, and biological processes controlling evapotranspiration. It can be applied to calculate evapotranspiration for a reference crop (Shuttleworth 1993, 2007). The Penman–Monteith equation is also used with satellite remote sensing to estimate continental-scale evapotranspiration (Zhang et al. 2010; Mu et al. 2011; Ryu et al. 2011). Radiation, atmospheric humidity, and wind speed are critical determinants of evapotranspiration, and temperature-based estimates (such as Thornthwaite's method) give different (and poorer) estimates of the water balance than do radiation-based and combination-based methods (Fisher et al. 2009; Sheffield et al. 2012).

## 10.5 Runoff

The rate at which water infiltrates depends in part on the rate at which it is supplied to the soil surface. When the rainfall rate is less than the infiltration capacity, all the water infiltrates into the soil. Water delivered in excess of infiltration capacity initially accumulates as puddles in small depressions on the surface. Once this depression storage capacity is exceeded, the excess water flows downhill as overland flow, or surface runoff. The time when this occurs, known as time to ponding, depends on soil texture, antecedent soil water, and delivery rate. High infiltration capacity does not allow ponding on sand or sandy loam except under extremely high precipitation rates. In general, infiltration rate decreases, time to ponding decreases, and runoff increases from sand to loam to clay or with initially wetter soil.

The Green–Ampt equation (Chapter 9), and other such formulations, represents infiltration into idealized soil columns. In addition to micropores arising from the shape, arrangement, and aggregation of mineral and humus particles, soils have macropores formed by plant roots, earthworms, ants, and other burrowing organisms. These macropores can increase infiltration rates. Additionally, soil properties vary spatially, and other methods must be used to account for the effect of soil heterogeneity on infiltration (Chapter 11). Because of its importance to stormflow, several empirical formulas have been devised to determine runoff for application in landscape and urban planning. These equations illustrate the environmental controls of runoff.

One simple means to estimate runoff is the Rational method, which is commonly used in urban planning (Strom and Nathan 1993; Ferguson 1998). Runoff ($R$, m$^3$ s$^{-1}$) is:

$$R = 0.278cPA \tag{10.6}$$

where $P$ is rainfall intensity (mm per hour), $A$ is the drainage area (km$^2$), and $c$ is a coefficient that varies with land cover. The factor 0.278 converts units to m$^3$ s$^{-1}$. The equation states that peak runoff is equal to the fraction of the rainfall that runs off ($cP$) multiplied by the size of the drainage area ($A$). The runoff coefficient ranges from zero for a completely pervious surface to one for a completely impervious surface (Table 10.1). Urban landscapes generally generate more runoff than vegetated landscapes. Vegetated landscapes generate less runoff than bare ground.

**Table 10.1** Runoff coefficients for use with the Rational method

| Land cover | Range |
|---|---|
| *Natural landscapes* | |
| Forest, 0–5% slope | 0.10–0.40 |
| Forest, 5–10% slope | 0.25–0.50 |
| Forest, 10–30% slope | 0.30–0.60 |
| Grass, 0–5% slope | 0.10–0.40 |
| Grass, 5–10% slope | 0.16–0.55 |
| Grass, 10–30% slope | 0.22–0.60 |
| Bare ground, 0–5% slope | 0.30–0.60 |
| Bare ground, 5–10% slope | 0.40–0.70 |
| Bare ground, 10–30% slope | 0.52–0.82 |
| *Urban landscapes* | |
| Suburban residential | 0.25–0.40 |
| Row houses | 0.60–0.75 |
| Industrial | 0.50–0.90 |
| Central business district | 0.70–0.95 |

*Source*: From Strom and Nathan (1993, p. 106).

The United States Soil Conservation Service developed a method for estimating runoff based on soil type, land use, land cover, and antecedent soil moisture (SCS 1985, 1986). This method is also used in urban planning (Strom and Nathan 1993; Ferguson 1998). Runoff ($R$, mm) is:

$$R = \frac{(P - I_a)^2}{P - I_a + S_{max}} \tag{10.7}$$

where $P$ (mm) is rainfall over a 24-hour interval, $I_a$ (mm) is the initial loss of water to infiltration and in surface depressions before runoff begins (known as the initial abstraction), and $S_{max}$ is the potential maximum retention after runoff begins. This latter term is related to a curve number ($CN$) that depends on soil type, land use, land cover, and antecedent soil moisture. For fluxes in millimeters:

$$S_{max} = \left(\frac{1000}{CN} - 10\right)25.4 \tag{10.8}$$

The initial abstraction is $I_a = 0.2S_{max}$.

Figure 10.5a illustrates runoff in relation to precipitation for a variety of curve numbers.

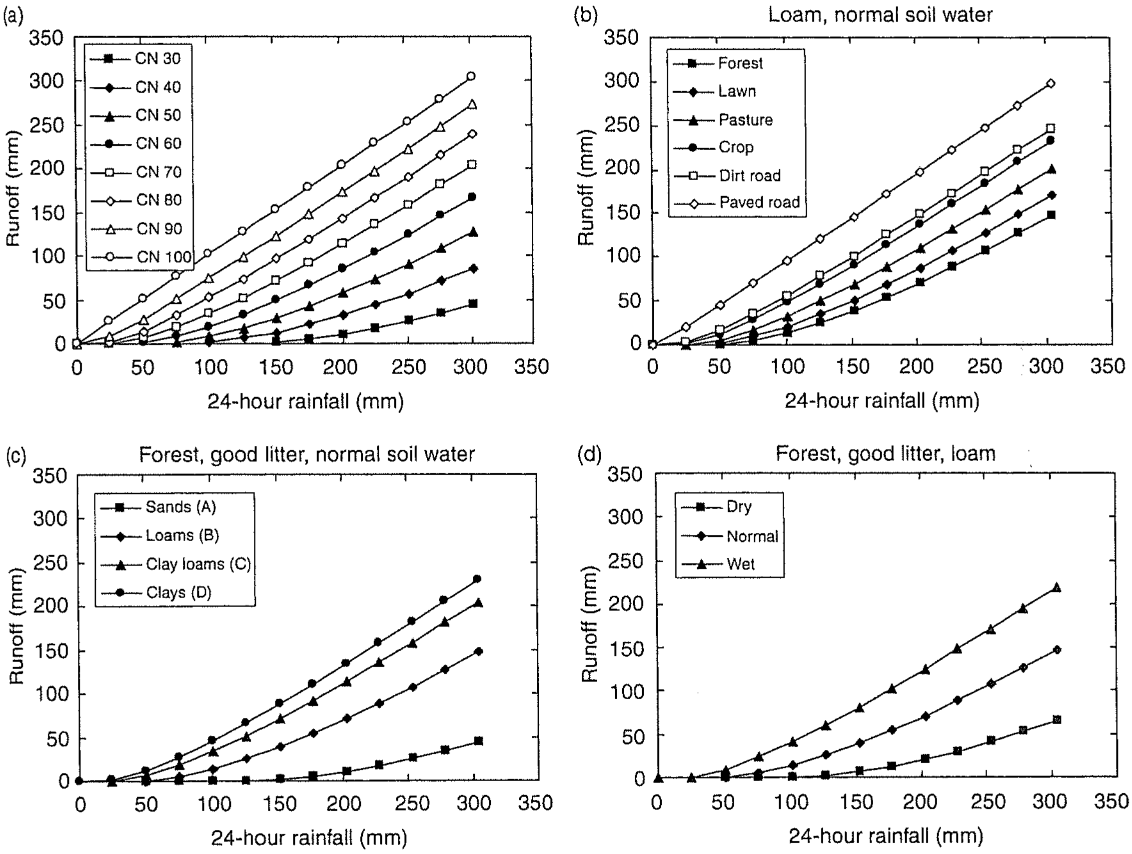


**Fig. 10.5** U.S. Soil Conservation Service runoff in relation to precipitation. Curve numbers are given in Table 10.2. (a) Effect of curve number from 30 to 100 on runoff. (b) Effect of land cover for a loam with normal soil water. (c) Effect of soil texture for a forest with good litter and normal soil water. (d) Effect of soil water for a forest with good litter and loam soil.

No surface runoff occurs when the duration of the storm is less than the time required to saturate the soil or if the intensity of rainfall is less than the soil's infiltration capacity. Runoff increases with curve number until $CN = 100$, when there is a one-to-one relationship between precipitation and runoff. The initial detention of precipitation prior to runoff decreases with curve number. For $CN = 30$, runoff does not begin until rainfall exceeds 150 mm. For $CN = 60$, runoff begins with rainfall in excess of 50 mm. Runoff begins almost immediately for $CN \geq 90$.

Table 10.2 shows curve numbers for several land cover types and soils with normal soil water, and Figure 10.5 shows resulting runoff in relation to storm rainfall for a variety of conditions. For a given soil type (e.g., loam), a forest with a good litter cover generates the least runoff. The litter cover retards surface water flow, giving the water additional time to enter the soil. In addition, large, extensive tree roots make the soil more porous, allowing more water to enter the soil. Crops and dirt roads generate high runoff. Paved roads generate the most runoff. Sands have high infiltration rates and low runoff potential. Clays have low infiltration rates and high runoff potential. Loams are intermediate soils, with moderate to low infiltration rates. Curve numbers must be adjusted for antecedent moisture conditions (SCS 1985, pp. 4.10–4.12, p. 10.7). Dry soils have lower curve numbers than wet soils so that dry soils have less runoff than wet soils.

Table 10.2 | Soil Conservation Service curve numbers (CN) in relation to soil texture and land cover for normal soil moisture conditions

| Land cover | Soil texture | | | |
|---|---|---|---|---|
| | Sands (A) | Loams (B) | Clay loams (C) | Clays (D) |
| Woods, good litter | 30 | 55 | 70 | 77 |
| Grass lawn, good condition | 39 | 61 | 74 | 80 |
| Pasture, good condition (moderately grazed) | 49 | 69 | 79 | 84 |
| Cropland, good condition | 67 | 78 | 85 | 89 |
| Dirt road | 72 | 82 | 87 | 89 |
| Paved road | 98 | 98 | 98 | 98 |

*Source*: From SCS (1985, pp. 9.1–9.11) and SCS (1986).

## 10.6 | Soil Water

In most locations, the water table, which defines the depth at which soil is saturated, is well below the ground surface. Wetlands, where the water is perched above the ground, are an exception. Between the ground surface and the water table is a region where soil is less than saturated. This is known as the unsaturated, or vadose, zone, and water in this region is known as soil water. Plant roots are typically restricted to the unsaturated zone, which supplies plants with necessary moisture.

The unsaturated zone is divided into three zones related to the distribution of water above the water table (Figure 10.6). The first 50–100 cm of soil, where plants typically have most of their roots, is known as the rooting zone. This zone is often saturated during rainfall when water infiltrates into the soil. However, the near-surface soil quickly dries as some of the water drains downward due to the force of gravity and some evaporates to the atmosphere. In addition, plant roots extract water to meet transpiration needs. Consequently, water contents in the root zone range from saturation during infiltration to wilting point in dry periods. Immediately below the root zone is the intermediate zone. This zone is recharged when water in excess of field capacity percolates down the soil column, eventually reaching a zone of saturation bounded by an impermeable layer (e.g., bedrock). Water contents can reach saturation during heavy storms, but mostly the water content is near field capacity. Below the intermediate zone is the groundwater, and immediately above this is a small zone called the capillary fringe, which is kept saturated by water rising from groundwater.

Figure 10.7 illustrates the drainage and wetting of the unsaturated zone over a 16-day period in a Canadian pine forest. On May 30, the overall water content was high, though the near-surface soil was dry. By June 13, the upper soil had dried as a result of evapotranspiration while the deeper soil had dried from drainage. Two days later, heavy rainfall wetted the upper soil. A distinct wetting front at a depth of 50 cm is apparent.

Plant roots extract water from the soil to replenish water lost through transpiration. In some instances, plant roots can also redistribute water within the soil profile. Such activity is known as hydraulic lift or more generally hydraulic redistribution. Through this process, drier upper soil layers are moistened by water from wetter deeper layers. Hydraulic redistribution is widespread among plant species and has been observed in grasses, shrubs, and trees in deserts, temperate forests, and tropical savannas (Richards and Caldwell 1987; Caldwell and Richards 1989; Dawson 1993a, 1996; Burgess et al. 1998; Caldwell et al. 1998; Jackson et al. 2000; Meinzer et al. 2004; Domec et al. 2010; Neumann and Cardon 2012). By keeping

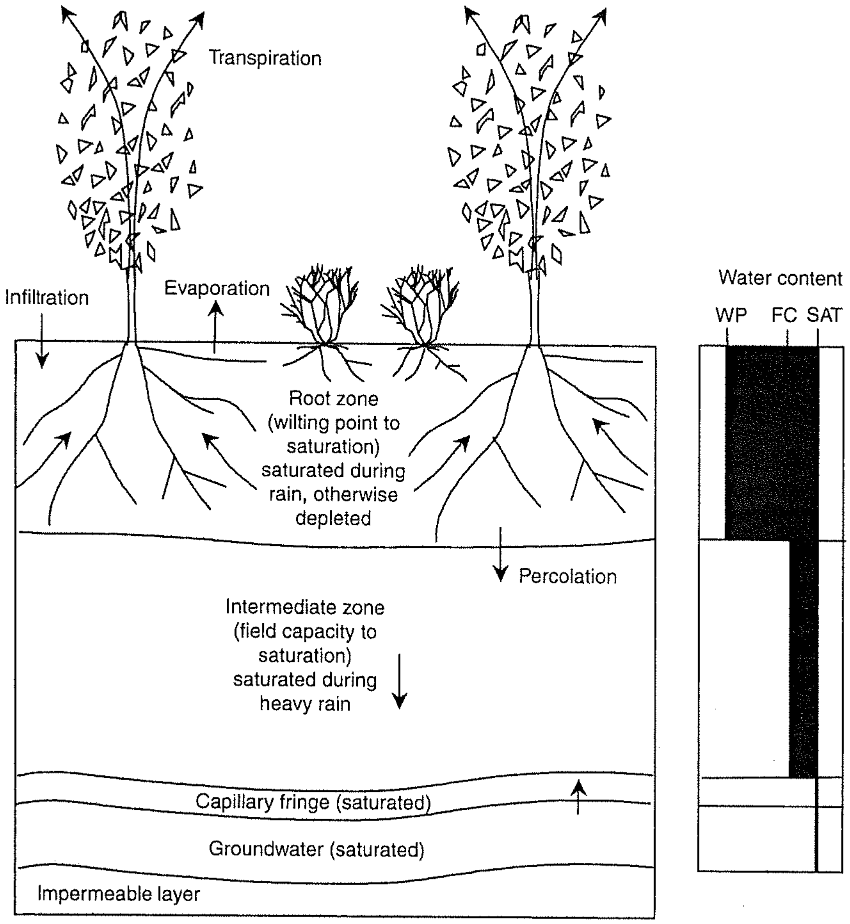


**Fig. 10.6** Typical soil water zones and movements.

upper soils moist, hydraulic redistribution can enhance water availability and sustain transpiration, with important effects on climate (Lee et al. 2005; Wang 2011).

## 10.7 | Water Balance Model

The complexities of the hydrologic cycle on land can be reduced to a simple form in which the change in soil water ($\Delta S$) is the balance between water input from precipitation ($P$), water loss from evapotranspiration ($E$), and water lost as runoff ($R$). Mathematically, $\Delta S = P - E - R$. Thornthwaite and Mather (1955, 1957) used this equation in a simple bucket model of monthly root zone soil water (Figure 10.8a). The soil is treated as a bucket with a maximum water-holding capacity. Precipitation fills the bucket, and evapotranspiration depletes the bucket. Any water in excess of the maximum water-holding capacity overflows the bucket and is lost as runoff. Potential evapotranspiration is calculated from Eq. (10.1) and reduced to actual evapotranspiration based on the ratio of soil water to maximum water-holding capacity.

Figure 10.9 illustrates this methodology applied over a 12-month period using the bucket model of Mintz and Walker (1993). From January to May and again from September to December, soil water does not limit evapotranspiration so that actual evapotranspiration equals potential evapotranspiration. In these months, evapotranspiration is less than precipitation and the excess precipitation runs off. In June, July, and August, soil water limits evapotranspiration to

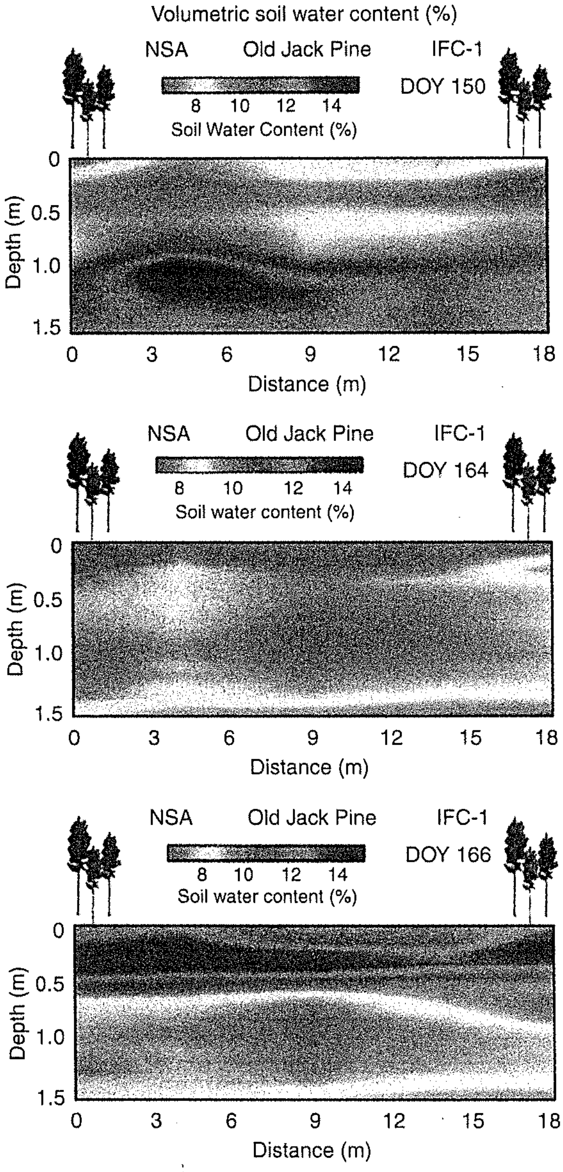


**Fig. 10.7** Soil water content (%) with depth across an 18 m transect in a Canadian jack pine forest in late spring 1994. Top: moist conditions on May 30. Middle: dry down on June 13. Bottom: wetting front on June 15. Reproduced from Cuenca et al. (1997). See color plate section.

less than the potential rate. There is no runoff in these months, because water loss from evapotranspiration balances water input from precipitation.

Figure 10.10 shows annual $P-E$ for the United States calculated by this method. The eastern half of the United States, except southern Florida, has a large surplus of water. The Pacific Northwest, where annual precipitation is high, and parts of the mountainous West, where cold temperatures reduce evapotranspiration, also have large water surplus. The least surplus water occurs in the hot, arid Southwest.

Other water balance models follow the same principles, but represent more physical processes. For example, the monthly water balance model of McCabe and Wolock (2011a,b) additionally temporarily stores water as snow (Figure 10.8b). Precipitation is partitioned as rain or snow based on temperature. This snowfall accumulates and melts at a rate determined by temperature. Runoff is generated from infiltration-excess overland flow and additionally from surplus soil water. Monthly temperature and precipitation are climate inputs to the model. Seven site-specific inputs are: daylength (used in potential evapotranspiration); rain and snow threshold temperatures; the maximum snow melt rate; the fraction of rainfall that becomes direct runoff; the fraction of surplus water that becomes runoff; and soil water storage capacity.

## 10.8 | Isotope Hydrology

Isotopes are variants of a particular element that differ in their number of neutrons. Stable isotopes do not undergo radioactive decay and are commonly used in hydrological and biogeochemical analyses. Carbon, for example, has three naturally occurring isotopes: two stable isotopes ($^{12}C$ and $^{13}C$) and one radioactive isotope ($^{14}C$). Hydrogen has two stable isotopes, $^{1}H$ and $^{2}H$ (commonly called deuterium and abbreviated by the symbol D). Two stable isotopes of oxygen used in geochemical studies are $^{16}O$ and $^{18}O$. The lighter isotopes are much more common than the heavier isotopes; $^{12}C$, $^{1}H$, and $^{16}O$ have natural abundances of 98.89 percent, 99.98 percent, and 99.76 percent, respectively. The heavier isotopes ($^{13}C$, $^{2}H$, and $^{18}O$) are much less abundant.

The isotopic composition of a sample (given by the ratio of the heavy isotope to the light

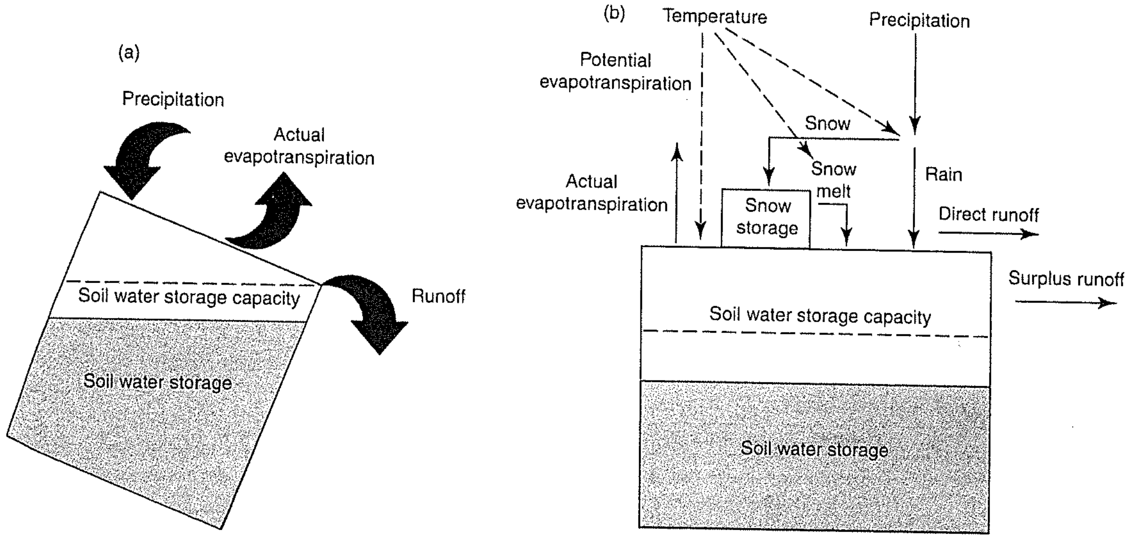


**Fig. 10.8** Schematic representation of (a) a bucket model with the water balance $\Delta S = P - E - R$ (Thornthwaite and Mather 1955, 1957) and (b) a more detailed water balance model (McCabe and Wolock 2011a).

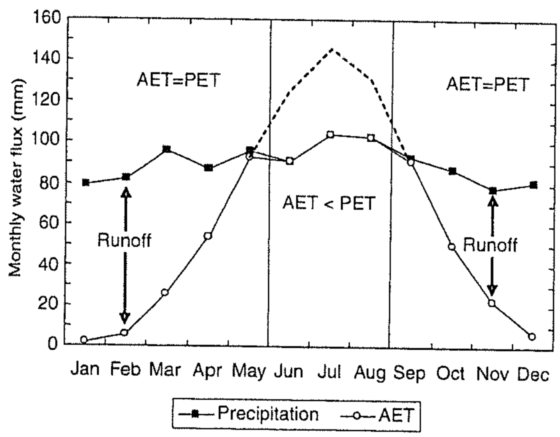


**Fig. 10.9** Monthly water balance. Runoff is the difference between precipitation and actual evapotranspiration (AET). The dashed line shows potential evapotranspiration (PET). In other months, AET = PET.

isotope) is measured relative to a standard. For $^{18}O$:

$$\delta^{18}O = \left[ \frac{\left(\frac{^{18}O}{^{16}O}\right)_{\text{sample}}}{\left(\frac{^{18}O}{^{16}O}\right)_{\text{standard}}} - 1 \right] 1000 \qquad (10.9)$$

and is reported as per mil (‰). The standard, by definition, has $\delta = 0‰$. A negative value indicates a ratio of heavy-to-light isotope that is less in the sample compared with the standard. The sample is said to be lighter or depleted relative to the standard. A positive value indicates a ratio of heavy-to-light isotope that is greater in the sample compared with the standard. The sample is said to be heavier or enriched. The $\delta^2H$ is based on the ratio $^2H/^1H$, and the $\delta^{13}C$ is based on $^{13}C/^{12}C$.

The isotopic composition of a product can differ from the source material. The process by which this occurs is termed isotope fractionation. In general, lighter isotopes are favored in evaporation and photosynthesis, leaving the source material heavier. This isotopic fractionation, or natural variation in the isotopic composition of substances, provides mechanistic understanding of geochemical cycles. For example, three isotopic types of water are $^1H_2{}^{16}O$ (>99%) and the heavier water $^1H^2H^{16}O$ (commonly abbreviated HDO) and $^1H_2{}^{18}O$. Isotopic fractionation occurs naturally through evaporation and condensation and imparts a discernible signature to the hydrologic cycle (Figure 10.11). Lighter isotopes evaporate more easily than heavier isotopes and

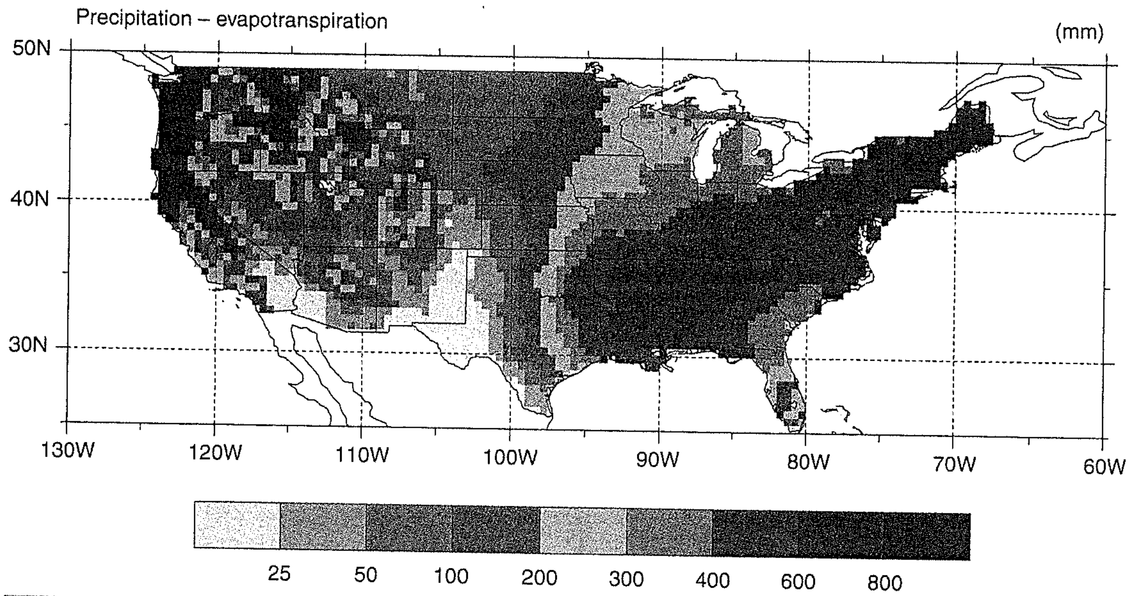


Fig. 10.10 Geographic distribution of the difference between annual precipitation and evapotranspiration in the United States. Evapotranspiration is based on the water balance model of Figure 10.9 using monthly temperature and precipitation climatologies (Legates and Willmott 1990a,b) and observed soil water-holding capacity (Rosenbloom and Kittel 1996). See color plate section.

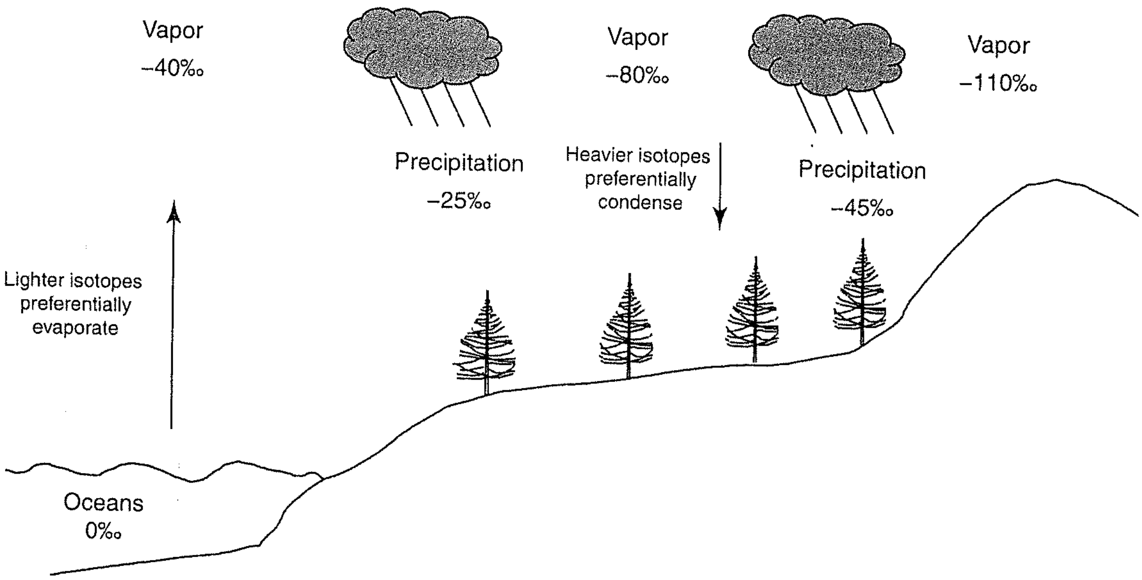


Fig. 10.11 Generalized representation of isotopes in the hydrologic cycle. Shown are representative values for $\delta^2H$ with evaporation from tropical oceans and subsequent precipitation over land. Adapted from Dawson (1993b).

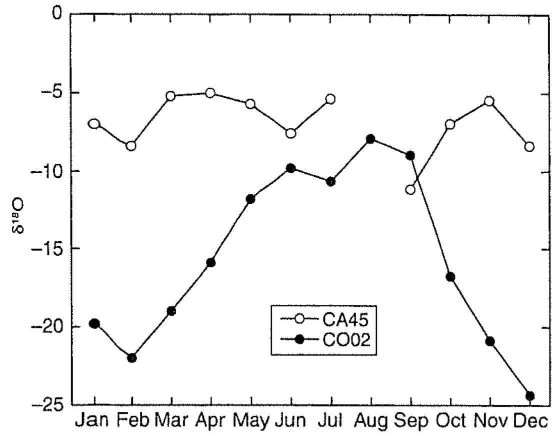


Fig. 10.12 Average monthly $\delta^{18}O$ of precipitation for a low elevation site near the California coast (CA45) and an inland site in the Colorado Rocky Mountains (CO02). Data from Vachon et al. (2010).

occur preferentially in the vapor phase. Heavier isotopes preferentially condense and occur preferentially in the liquid phase. Thus, water vapor that evaporates from tropical oceans is depleted in $^{18}O$ or $^{2}H$ (low $\delta^{18}O$ and $\delta^{2}H$). Heavier isotopes condense and precipitate preferentially to lighter isotopes, so that rainfall is enriched in $^{18}O$ or $^{2}H$ (higher $\delta^{18}O$ and $\delta^{2}H$) compared with water vapor. The air mass itself becomes further depleted in $^{18}O$ and $^{2}H$ as it moves inland. Temperature affects the isotopic composition of precipitation. With increasing temperature, precipitation is enriched in heavier isotopes ($^{18}O$ and $^{2}H$). Thus, precipitation is depleted of heavy isotopes in polar regions (low $\delta^{18}O$ and $\delta^{2}H$). The effect of temperature and source region on the isotopic signature of precipitation is seen in comparison of $\delta^{18}O$ for coastal California and inland Colorado sites (Figure 10.12). The coastal site has higher $\delta^{18}O$. The inland site is more depleted in $^{18}O$ (lower $\delta^{18}O$), especially in winter.

The isotopic composition of water can identify the source of atmospheric water (Henderson-Sellers et al. 2004; Noone et al. 2013). Isotopes can additionally discriminate among processes within a plant canopy, especially partitioning evapotranspiration into soil evaporation and transpiration (Yakir and Sternberg 2000; Yepez et al. 2003; Williams et al. 2004; Dawson and Simonin 2011; Jasechko et al. 2013). This is because evaporation enriches soil water in $^{18}O$ and $^{2}H$, but transpiration does not produce similar isotopic fractionation.

## 10.9 | Review Questions

**1.** Daily precipitation at a particular site is 40 mm; interception is 3 mm; transpiration is 4 mm; soil evaporation is 1 mm; and runoff is 6 mm. How much water is lost in evapotranspiration? How much water infiltrates into the soil? What is the change in soil water?

**2.** The data in Figure 10.2 show that a deciduous forest in full leaf intercepts 2 mm of rainfall in a 20 mm storm; 3 mm in a 35 mm storm; and 3.8 mm in a 50 mm storm. Why does the percentage of rainfall intercepted decrease with greater rainfall?

**3.** Calculate monthly potential evapotranspiration ($E_p$, mm) using Thornthwaite's method for a site with the following monthly temperature ($T$, °C), precipitation ($P$, mm), and daylength ($L$, hours):

| | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $T$ | −10.6 | −9.4 | −3.9 | 5.0 | 12.8 | 18.3 | 20.6 | 19.4 | 15.0 | 8.3 | 0.6 | −7.2 |
| $P$ | 94 | 81 | 94 | 61 | 79 | 89 | 97 | 86 | 89 | 84 | 86 | 94 |
| $L$ | 9.6 | 10.6 | 11.8 | 13.2 | 14.3 | 15.0 | 14.8 | 13.8 | 12.6 | 11.3 | 10.0 | 9.4 |
| Days | 31 | 28 | 31 | 30 | 31 | 30 | 31 | 31 | 30 | 31 | 30 | 31 |