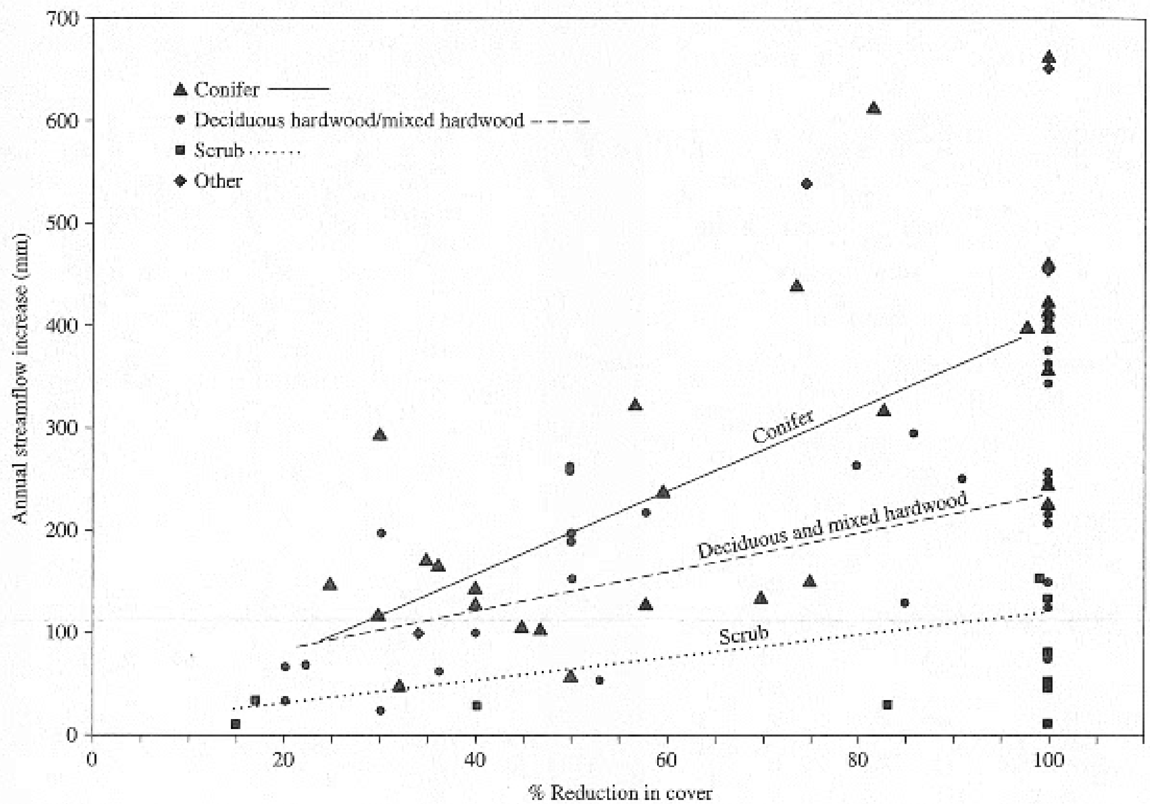



**FIGURE 7-18**
Annual streamflow increases due to reductions in vegetative cover as measured in watershed experiments. Reprinted from Bosch and Hewlett (1982) with permission of Elsevier Science.

### 7.6.5 Water-Quality Aspects

Even though intercepted water remains on plant canopies for only short times, the chemistry of throughfall is commonly significantly different from that of unintercepted precipitation (Figure 7-19). Much of the change is due to the dissolution of material that was deposited on the leaves from the air (dry deposition) and some—particularly organic carbon—represents leaching of the leaves themselves.

Lindberg and Garten (1988) found that more than 85% of the sulfate in throughfall in the southeastern United States is wash-off of dry deposition. They concluded that measurements of the sulfate content of throughfall are a good way to monitor the inputs of that constituent, which is of concern as an anthropogenic contributor to acidification of surface water.

## 7.7 POTENTIAL EVAPOTRANSPIRATION

### 7.7.1 Conceptual Definition

**Potential evapotranspiration** (PET) is the rate at which evapotranspiration would occur from a large area completely and uniformly covered with growing vegetation which has access to an unlimited supply of soil water, and without advection or heat-storage effects (Table 7-1). The concept was introduced as part of a scheme for climate classification by Thornthwaite (1948), who intended it to depend essentially on climate and to be largely independent of the surface characteristics.

However, we now know from Equation (7-56) that several characteristics of a vegetative surface have a strong influence on evapotranspiration rate even when there is no limit to the available water: (1) the albedo of the surface, which determines the net radiation (Table 7-5); (2) the maximum leaf conductance (Table 7-5); (3) the atmospheric conductance, which is largely determined by vegetation height (Table 7-5; Figure 7-11); and (4) the presence

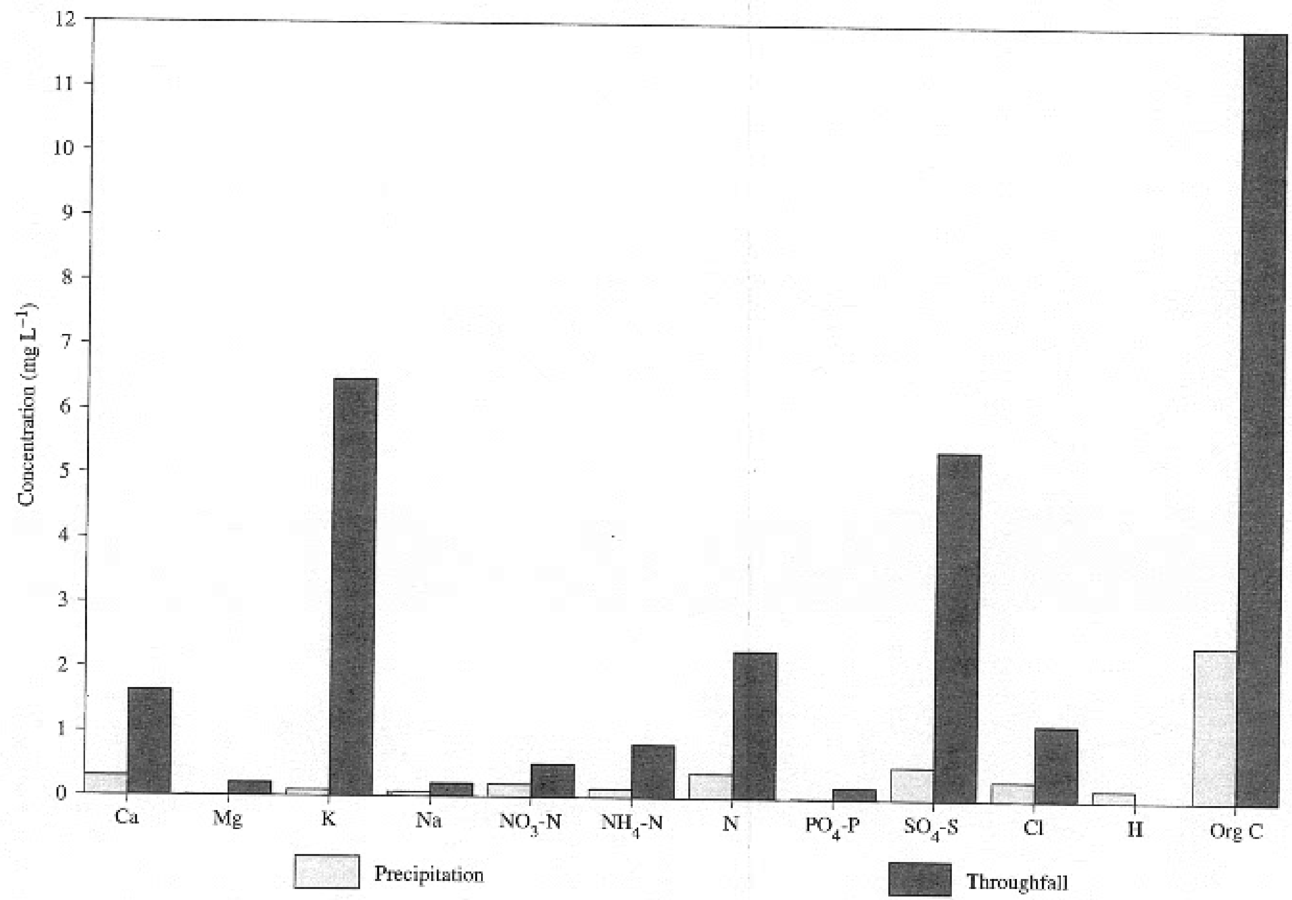



**FIGURE 7-19**
Comparison of chemical composition of incident precipitation and throughfall at Hubbard Brook Experimental Forest in the White Mountains of New Hampshire. Data from Likens et al. (1977).

or absence of intercepted water. Because of these surface effects, Penman (1956) redefined PET as "the amount of water transpired ... by a short green crop, completely shading the ground, of uniform height and never short of water," and the term **reference-crop evapotranspiration** is increasingly used as a synonym for PET.

Another concern about the definition of PET is that its magnitude is often calculated from meteorological data collected under conditions in which the actual evapotranspiration rate is less than the potential rate. However, if evapotranspiration had been occurring at the potential rate, the latent- and sensible-heat exchanges between the air and the surface, and hence the air temperature and humidity, would have been considerably different (Brutsaert 1982).

In spite of these considerable ambiguities, it has proven useful to retain the concept of PET as an index of the "drying power" of the climate or the ambient meteorological conditions, and we now examine some operational definitions that have been applied in climate classification and hydrologic modeling. Section 7.8.1 describes how estimates of actual evapotranspiration are derived from calculated values of potential evapotranspiration in hydrologic analysis.

### 7.7.2 Operational Definitions

In practice, PET is defined by the method used to calculate it, and many methods have been proffered. We limit our discussion to the methods most commonly applied in hydrologic studies. Following Jensen et al. (1990), these methods can be classified on the basis of their data requirements:

**Temperature-based:** Use only air temperature (often climatic averages) and sometimes day length (time from sunrise to sunset).

**Radiation-based:** Use net radiation and air temperature.

**Combination:** Based on the Penman combination equation; use net radiation, air temperature, wind speed, and relative humidity.

**Pan:** Use pan evaporation, sometimes with modifications depending on wind speed, temperature, and humidity.

Some of the methods do not require information about the nature of the surface and can be considered to give reference-crop evapotranspiration, others are surface-specific and require information about albedo, vegetation height, maximum stomatal conductance, leaf-area index, and other factors.

#### *Temperature-Based Methods*

Thornthwaite (1948) developed a complex empirical formula for calculating PET as a function of climatic average monthly temperature and day length. It turns out that Thornthwaite's temperature function has a form similar to the saturation vapor-pressure relation [Equation (7-4)], and some simplifications of his approach are based on this similarity. For example, Hamon (1963) estimated daily PET as

$$PET_H = 29.8 \cdot D \cdot \frac{e_a^*(T_a)}{T_a + 273.2}, \qquad \textbf{(7-63)}$$

where $PET_H$ is in mm day$^{-1}$, $D$ is day length in hr [calculated via Equations (E-5)], and $e_a^*(T_a)$ is the saturation vapor pressure at the mean daily temperature, $T_a$ (°C), in kPa. Equation (7-63) gives values close to those given by the original Thornthwaite formulation and has been used in several hydrologic models.

Malmstrom (1969) used an approach similar to Hamon's and claimed improved climate classification by estimating monthly climatic PET, $PET_M$, as

$$PET_M = 40.9 \cdot e_a^*, (T_a), \qquad \textbf{(7-64)}$$

where $PET_M$ is in mm month$^{-1}$, $e_a^*$ is in kPa, and the temperature is the climatic mean monthly air temperature in °C for months when average temperature exceeds 0 °C.

#### *Radiation-Based Methods*

Slatyer and McIlroy (1961) reasoned that air moving large distances over a homogeneous well-watered surface would become saturated, so that the mass-transfer term in the Penman Equation [Equation (7-55)] would disappear. They defined the evapotranspiration under these conditions as the **equilibrium potential evapotranspiration,** $PET_{eq}$. Subsequently, Priestley and Taylor (1972) compared $PET_{eq}$ with values determined by energy-balance methods over well-watered surfaces and found a close fit if $PET_{eq}$ was multiplied by a factor $\alpha_{PT}$ to give

$$PET_{PT} = \frac{\alpha_{PT} \cdot \Delta \cdot (K + L)}{\rho_w \cdot \lambda_v \cdot (\Delta + \gamma)}. \qquad \textbf{(7-65)}$$

A number of field studies of evapotranspiration in humid regions have found $\alpha_{PT} = 1.26$, and theoretical examination has shown that that value in fact represents equilibrium evapotranspiration over well-watered surfaces under a wide range of conditions (Eichinger et al. 1996). Thus $PET_{PT}$ is often referred to as the equilibrium potential evapotranspiration, and Equation (7-65) gives an estimate of PET that depends only on net radiation and air temperature. This relationship has proven useful in hydrologic analysis.

#### Combination Methods

If the required data are available, the Penman–Monteith Equation [Equation (7-56)], using a $C_{leaf}$ value calculated from Equation (7-52) with $f_{\theta}(\Delta\theta) = 1$, can be used to estimate PET for a specified vegetated surface. Shuttleworth (1994) defined the reference crop as grass with a height ($z_{veg}$) of 120 mm, an albedo ($a$) of 0.23, and a canopy conductance ($C_{can}$) of 14.5 mm s$^{-1}$.

#### Pan-Based Methods

The potential evapotranspiration for short vegetation is commonly very similar to free-water evaporation (Linsley et al. 1982; Brutsaert 1982). This may be because lower canopy conductance over the vegetation fortuitously compensates for the lower atmospheric conductance over the pan. In any case, annual values of pan evaporation (= free-water evaporation as shown in Figure 7-7) are essentially equal to annual PET, and pan evaporation corrected via Equations (7-41) and (7-42) can be used to estimate PET for shorter periods.

Potential evapotranspiration can also be directly measured by various forms of **atmometers** in which evaporation occurs from porous surfaces (Giambelluca et al. 1992) or flat plates (Fontaine and Todd 1993).

### 7.7.3 Comparison of PET Estimation Methods

Jensen et al. (1990) compared PET computed by 19 different approaches with measured reference-crop evapotranspiration in weighing lysimeters[11] at 11 locations covering a range of latitudes and elevations. The Penman–Monteith method gave the best overall results (Figure 7-20a). Equilibrium evapotranspiration [Equation (7-65) with $\alpha_{PT} = 1.26$] gave reasonable agreement up to rates of 4 mm day$^{-1}$ but considerable underestimation at higher rates (Figure 7-20b). Monthly Class-A pan evaporation correlated well with measured PET, but with considerable scatter presumably due to variability of heat exchange through the pan walls (Figure 7-20c).

For short vegetation, the Penman Equation [Equation (7-55)] gives nearly the same estimates as the Penman–Monteith Equation, and Van Bavel (1966) found a close correspondence for hourly and daily evapotranspiration computed by the Penman Equation and that measured for well-watered alfalfa growing in a lysimeter (Figure 7-21).

Vörösmarty et al. (1998) compared the estimates of annual PET given by nine different methods in a global-scale water-balance model. The various methods were used as a basis for estimating actual evapotranspiration in a monthly water-balance model (as discussed in Section 7.8.1), and these estimates were compared with the difference between measured precipitation and measured streamflow. Considering the differences in conceptual basis and data requirements, the various methods gave surprisingly similar results overall. Interestingly, the Hamon method [Equation (7-63)], which is based only on temperature and day length, performed the best. Several variations of the Penman–Monteith method [Equation (7-56)] performed well, while the equilibrium method [Equation (7-65)] tended to overestimate in regions with higher ET rates and the Penman Equation [Equation (7-55)] overestimated for all locations.

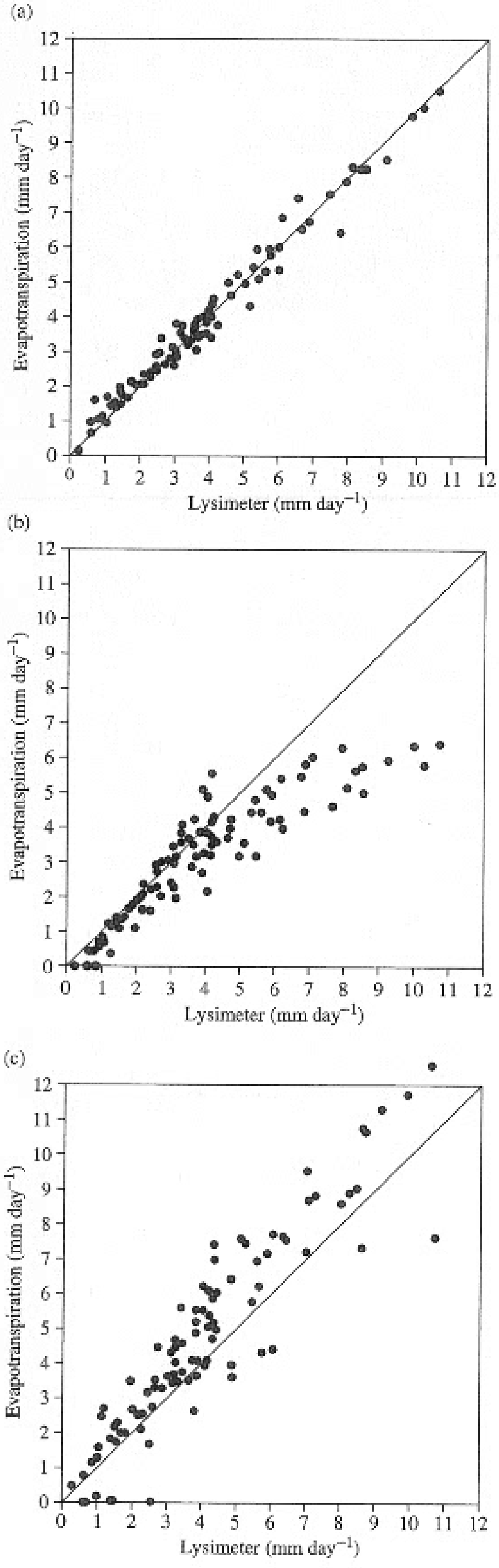
[11] A **weighing lysimeter** is an enclosed volume of soil for which the inflows and outflows of liquid water can be measured and changes in storage can be monitored by weighing. They are described further in Section 7.8.2.

## 7.8 ACTUAL EVAPOTRANSPIRATION

### 7.8.1 Potential-Evapotranspiration Approaches

#### Relation to Precipitation/Potential Evapotranspiration Ratio

In hot arid regions, potential evapotranspiration greatly exceeds precipitation so that average actual evapotranspiration is water-limited, and is essentially equal to average precipitation. In regions with abundant rainfall in all seasons, evapotranspiration is limited by the available energy, so that average evapotranspiration is essentially equal to average potential evapotranspiration. Pike (1964) thus reasoned that annual evapotranspiration, $ET$, is determined by




the ratio of average precipitation, $W$, to potential evapotranspiration, $PET$, and proposed the following relation for estimating annual evapotranspiration:

$$ET = \frac{W}{\left[1 + \left(\frac{W}{PET}\right)^2\right]^{1/2}}. \tag{7-66}$$

In spite of its purely empirical nature, Equation (7-66) gives reasonably good "first-cut" estimates of climatic average evapotranspiration (Figure 7-22).

### *Monthly Water-Balance Models*

Thornthwaite and Mather (1955) developed a water-balance model that estimates monthly actual evapotranspiration from monthly potential evapotranspiration. PET is calculated from monthly average temperature using the Hamon [Equation (7-63)], Malmstrom [Equation (7-64)], or other temperature-based approach. Monthly precipitation is input to a simple model of soil-moisture storage, which computes actual ET and updates soil moisture via a "bookkeeping" procedure. One version of this approach is described in detail in Box 7-3, and Table 7-9 and Figure 7-23 give an example of its application.

"Thornthwaite-type" monthly ET models can also be extended to estimate monthly ground-water recharge and runoff, and one can verify ET estimates by comparing estimated and measured runoff. In spite of their extremely simple structure, models of the Thornthwaite type generally estimate monthly runoff values reasonably well (Alley 1984; Calvo 1986), and this correspondence suggests that their estimates of actual evapotranspiration are also generally reasonable. Somewhat more elaborate versions of the basic monthly water-balance model described in Box 7-3 are used to simulate land-surface hydrology in many of the general circulation models used to forecast the impacts of climate change (e.g., Vörösmarty et al. 1998).

**FIGURE 7-20**
Comparison of average monthly potential evapotranspiration computed by (a) the Penman–Monteith equation [Equation (7-56)], (b) the Priestley–Taylor equilibrium method [Equation (7-65)], and (c) uncorrected Class-A pan evaporation, with values determined in lysimeters containing well-watered alfalfa at 11 locations. Overestimation by pan evaporation is probably due largely to heat exchange through the sides of the pan. From Jensen et al. (1990).

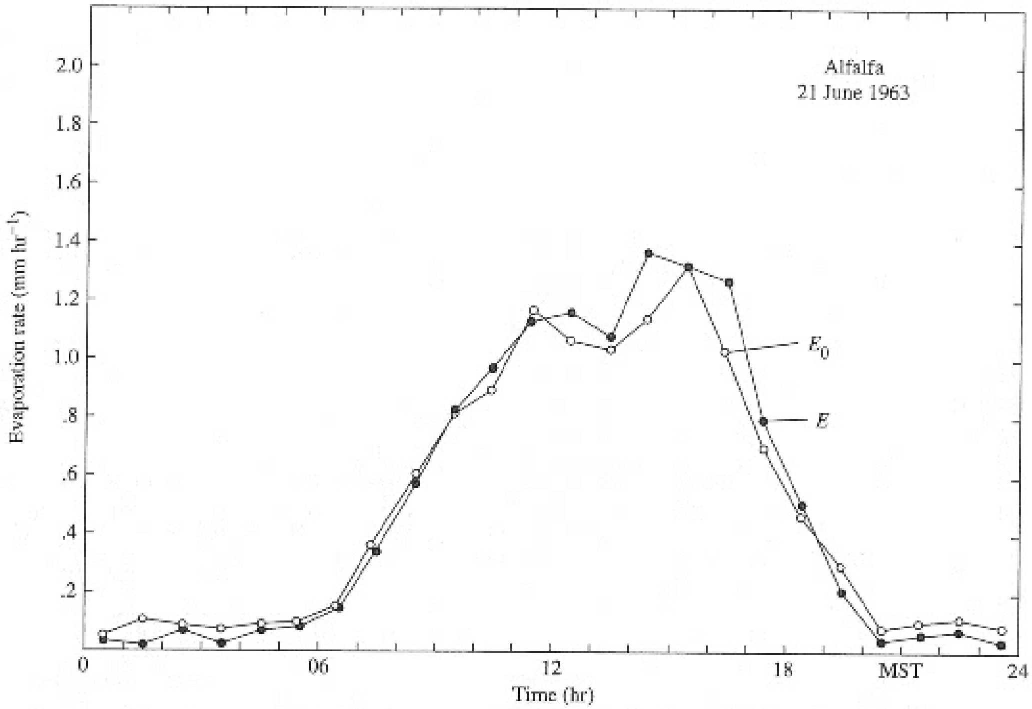



**FIGURE 7-21**
Comparison of observed hourly evapotranspiration for well-watered alfalfa (closed circles) and that calculated via the Penman equation (open circles). From Van Bavel (1966), used with permission of the American Geophysical Union.

### *Use of Soil-Moisture Functions*

One of the most widely used methods for estimating actual evapotranspiration makes use of meteorologic data to estimate potential evapotranspiration by relations like those discussed earlier, and then computes actual evapotranspiration as

$$ET = F(\theta_{rel}) \cdot PET. \quad \textbf{(7-67)}$$

Here $\theta_{rel}$ is the **relative water content,** defined as

$$\theta_{rel} \equiv \frac{\theta - \theta_{pwp}}{\theta_{fc} - \theta_{pwp}}, \quad \textbf{(7-68)}$$

in which $\theta$ is the current water content, $\theta_{fc}$ is the field-capacity, and $\theta_{pwp}$ is the permanent wilting point of the root-zone soil (Section 6.4.1). The relation between $ET/PET$ and $\theta_{rel}$ usually has a form like that shown in Figure 7-24: $ET/PET$ increases quasi-linearly as $\theta_{rel}$ increases, and reaches 1 at some water content $\theta_{crit}$ (e.g., Davies and Allen 1973; Federer 1979, 1982; Spittlehouse and Black 1981). Typically $0.5 \cdot \theta_{fc} \leq \theta_{crit} \leq 0.8 \cdot \theta_{fc}$.

The Stewart (1988) model of canopy conductance [Equations (7-52) and (7-54)] includes a function $f_\theta(\Delta\theta)$ (Table 7-6) that can be used with the Penman–Monteith Equation [Equation (7-56)] to give

$$\frac{ET}{PET} = \frac{\Delta + \gamma \cdot \left\{1 + \dfrac{C_{at}}{C_{can}[f_\theta(\Delta\theta) = 1]}\right\}}{\Delta + \gamma \cdot \left\{1 + \dfrac{C_{at}}{C_{can}[f_\theta(\Delta\theta)]}\right\}}. \quad \textbf{(7-69)}$$

The form of this expression can be explored in Exercise 7-12.

A third approach to estimating $ET$ from $PET$ relates the value of $\alpha_{PT}$ in Equation (7-65) to some measure of soil-water content (e.g., Mukammal and Neumann 1977).

Methods of estimating $ET$ from $PET$ and soil moisture are well suited to use in "real-time" estimation, where $\theta$ is measured every few days, and in hydrologic models like the BROOK90 model, where $\theta$ is estimated by a "bookkeeping" algorithm

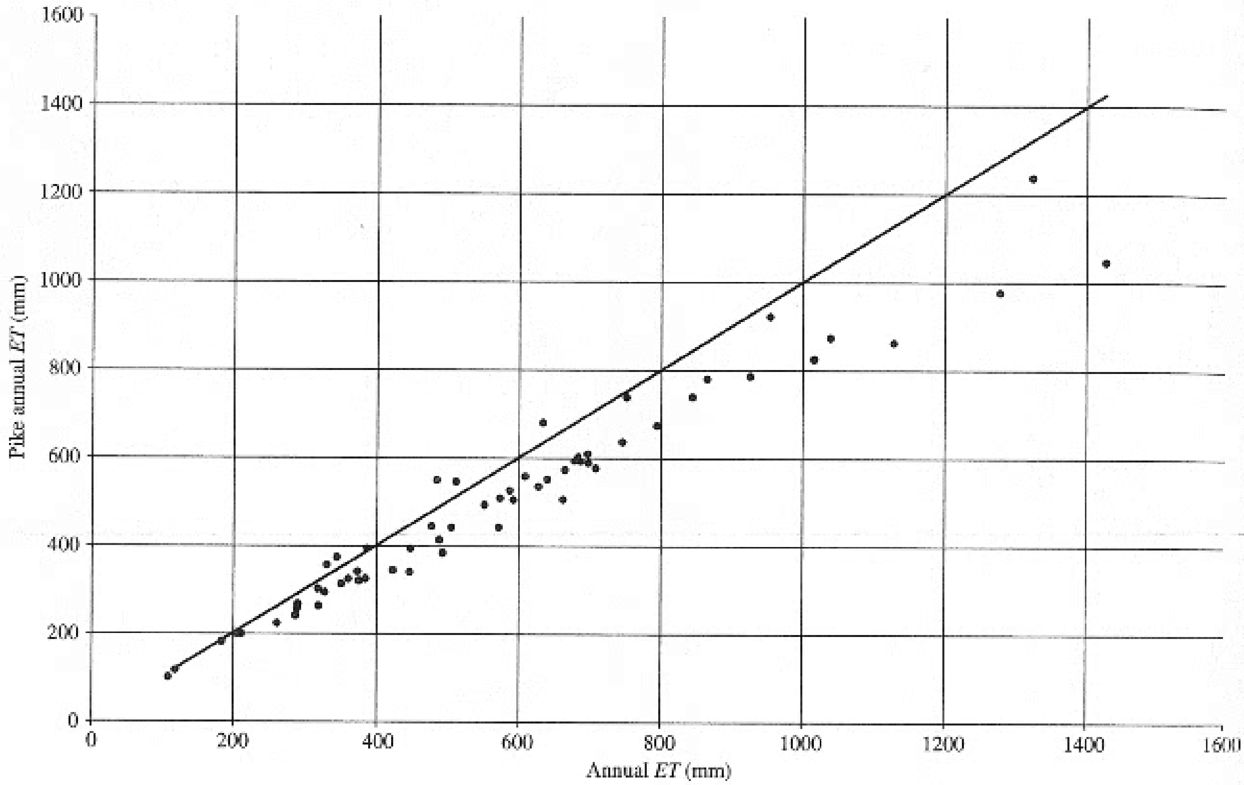



**FIGURE 7-22**
Comparison of annual evapotranspiration computed by the Pike Equation [Equation (7-66)] with that computed via a monthly water-balance model for selected North American stations.

along with equations for infiltration and deep drainage (Box 6-3). Box 7-4 describes the approach used by the BROOK90 model.

### *Complementary (Advection-Aridity) Approach*

Following Bouchet (1963), consider a uniform surface of 1 to 100 km$^2$ area evapotranspiring at the potential rate $ET = PET_0$ under a steady set of meteorological conditions. If these conditions remained constant, eventually the soil moisture would fall below field capacity and $ET$ would be less than $PET_0$. A flux of energy, $Q$, [E L$^{-2}$ T$^{-1}$] equivalent to the difference between $PET_0$ and $ET$ would then not be used for evapotranspiration and would become available to warm the atmosphere. Thus,

$$PET_0 - ET = \frac{Q}{\rho_w \cdot \lambda_v}. \quad \textbf{(7-70)}$$

The reduced evapotranspiration decreases the humidity, and the warming increases the air temperature. Under these circumstances, one would calculate a new potential evapotranspiration $PET$ that is larger than $PET_0$ by the amount $Q/(\rho_w \cdot \lambda_v)$:

$$PET - PET_0 = \frac{Q}{\rho_w \cdot \lambda_v}. \quad \textbf{(7-71)}$$

Combining Equations (7-70) and (7-71) yields the **complementary relationship** between $ET$ and $PET$:

$$ET = 2 \cdot PET_0 - PET. \quad \textbf{(7-72)}$$

(See Figure 7-25.)

Brutsaert and Stricker (1979) reasoned that $PET_0$ is the PET under equilibrium conditions [Equation (7-65)] and $PET$ is the "actual" PET given by the Penman Equation using measured current values of the meteorological variables [Equation (7-33)]. Substituting those relationships into Equation (7-72) yields

$$ET = \frac{(2 \cdot \alpha_{PT} - 1) \cdot \Delta \cdot (K + L) - \gamma \cdot K_E \cdot \rho_w \cdot \lambda_v \cdot v_a \cdot e_a^* \cdot (1 - W_a)}{\rho_w \cdot \lambda_v \cdot (\Delta + \gamma)}, \quad \textbf{(7-73)}$$

## BOX 7-3

### Thornthwaite-Type Monthly Water-Balance Model

Referring to Table 2-3, Thornthwaite-type monthly water-balance models are lumped conceptual models that can be used to simulate steady-state seasonal (climatic average) or continuous values of watershed or regional water input, snowpack, soil moisture, and evapotranspiration. Input for such models consists of monthly values of precipitation, $P_m$, and temperature, $T_m$, representative of the region of interest. For steady-state applications, these values are monthly climatic averages, in which case $m = 1, 2, \ldots, 12$; for continuous simulations they are actual monthly averages, in which case $m = 1, 2, \ldots, 12 \cdot N$, where $N$ is the number of years of record. Such models typically have a single parameter, the soil-water storage capacity of the soil in the region, $SOIL_{max}$, which is defined as

$$SOIL_{max} = \theta_{fc} \cdot Z_{rz}, \qquad \textbf{(7B3-1)}$$

where $\theta_{fc}$ is the field capacity and $Z_{rz}$ the vertical extent of the root zone. Typically $SOIL_{max} = 100$ or $150$ mm. For continuous applications an initial value of soil moisture, $SOIL_0$, must also be specified.

All water quantities in the model represent depths (volumes per unit area) of liquid water; inputs and outputs are monthly totals and snowpack and soil storage are end-of-month values.

#### Snowpack, Snowmelt, and Water Input

Monthly precipitation is divided into rain, $RAIN_m$, and snow, $SNOW_m$, where

$$RAIN_m = F_m \cdot P_m \qquad \textbf{(7B3-2)}$$

and

$$SNOW_m = (1 - F_m) \cdot P_m, \qquad \textbf{(7B3-3)}$$

in which $F_m$ is the **melt factor.** Following Figure 4-14, $F_m$ is computed as follows:

$$\begin{aligned}
&\text{If } T_m \le 0^\circ\text{C:} \quad F_m = 0;\\
&\text{if } 0^\circ\text{C} < T_m < 6^\circ\text{C:} \quad F_m = 0.167 \cdot T_m;\\
&\text{if } T_m \ge 6^\circ\text{C:} \quad F_m = 1.
\end{aligned} \qquad \textbf{(7B3-4)}$$

The melt factor is also used in a temperature-index snowmelt model [Equation (5-57)] to determine the monthly snowmelt, $MELT_m$, as

$$MELT_m = F_m \cdot (PACK_{m-1} + SNOW_m), \qquad \textbf{(7B3-5)}$$

where $PACK_{m-1}$ is the snowpack water equivalent at the end of month $m - 1$. The snowpack at the end of month $m$ is then computed as

$$PACK_m = (1 - F_m)^2 \cdot P_m + (1 - F_m) \cdot PACK_{m-1}. \qquad \textbf{(7B3-6)}$$

By definition, the water input $W_m$, is

$$W_m = RAIN_m + MELT_m. \qquad \textbf{(7B3-7)}$$

#### Evapotranspiration and Soil Moisture

Following Alley (1984), if $W_m \ge PET_m$, ET takes place at the potential rate, i.e.,

$$ET_m = PET_m \qquad \textbf{(7B3-8)}$$

and soil moisture increases or, if already at $SOIL_{max}$, remains constant. Thus

$$SOIL_m = \min\{[(W_m - PET_m) + SOIL_{m-1}], SOIL_{max}\}, \qquad \textbf{(7B3-9)}$$

where min{...} indicates the smaller of the quantities in the braces. In the original formulation, Thornthwaite (1948) used an empirical function that depends on the climatic average monthly temperature to calculate $PET_m$. Most current applications of the method use a simpler temperature-based method [e.g., Equations (7-63) or (7-64)] or, if data are available, one of the other approaches for estimating $PET_m$.

If $W_m < PET_m$, $ET_m$ is the sum of water input and an increment removed from soil storage; that is,

$$ET_m = W_m + SOIL_{m-1} - SOIL_m, \qquad \textbf{(7B3-10)}$$

where the decrease in soil storage is modeled via the following conceptualization:

$$SOIL_{m-1} - SOIL_m = SOIL_{m-1} \cdot \left[1 - \exp\left(-\frac{PET_m - W_m}{SOIL_{max}}\right)\right]. \qquad \textbf{(7B3-11)}$$

#### Computation

If the model is used with climatic monthly averages, the computations in Equations (7B3-5), (-9), (-10), and (-11) are "wrapped around" from $m = 12$ to $m = 1$ so that

## BOX 7-3 (CONTINUED)

### Thornthwaite-Type Monthly Water-Balance Model

$m - 1 = 12$ when $m = 1$. Thus the computations are circular and must be iterated until all the monthly quantities converge to constant values. This iteration is automatically carried out in EXCEL spreadsheets when the "Iterations" box in the "Tools Option Calculation" menu is checked. Otherwise you will get an error message "Cannot resolve circular references."

**Overall Water Balance**

The model output is a table of monthly values that can be graphed to give a concise picture of the annual cycle of inputs, soil and snowpack storage, evaporation, and water available for ground-water recharge and streamflow at any location. Table 7-9 is a completed water-balance spreadsheet for Omaha, NE, at 41.3° N latitude; the annual values are the sums of the monthly values. Note that the annual precipitation and water input are equal, as must be the case. As shown in Figure 7-23, the snowpack begins to build up in December, reaches a peak in February, and melts in March and April. $ET = PET$ for the months when $W > PET$ (March–June). Soil-water storage is recharged by rain and snowmelt beginning in February and reaches its capacity in April. It declines in the period from July to October because $PET > P$ and some of the evaporative demand is satisfied by withdrawal of water from soil storage. The last line, $W - ET - \Delta SOIL$, is the average monthly "water surplus" (i.e., the water available for recharge and runoff).

This model has been programmed in a spreadsheet in file "ThornEx.xls" on the CD accompanying this text, and Exercise 7-13 suggests ways to experiment with the model.

**TABLE 7-9**

Thornthwaite-type Monthly Water Balance (Box 7-3) for Omaha, NE, Computed by ThornEx.xls Spreadsheet Model on Text CD. Temperatures in °C, water-balance terms in mm. $SOIL_{max} = 100$ mm.

| | J | F | M | A | M | J | J | A | S | O | N | D | Year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| *P* | 21 | 24 | 37 | 65 | 88 | 115 | 86 | 101 | 67 | 44 | 32 | 20 | 700 |
| *T* | −5.4 | −3.1 | 2.7 | 10.9 | 17.2 | 22.8 | 25.8 | 24.6 | 19.4 | 13.2 | 3.8 | −2.1 | |
| *F* | 0 | 0 | 0.45 | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 0.63 | 0 | |
| *RAIN* | 0 | 0 | 17 | 65 | 88 | 115 | 86 | 101 | 67 | 44 | 20 | 0 | 603 |
| *SNOW* | 21 | 24 | 20 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 12 | 20 | 97 |
| *PACK* | 45 | 69 | 49 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 4 | 24 | |
| *MELT* | 0 | 0 | 40 | 49 | 0 | 0 | 0 | 0 | 0 | 0 | 7 | 0 | 97 |
| *W* | 0 | 0 | 57 | 114 | 88 | 115 | 86 | 101 | 67 | 44 | 28 | 0 | 700 |
| *PET* | 0 | 0 | 29 | 56 | 90 | 130 | 151 | 131 | 88 | 54 | 26 | 0 | 755 |
| *W-PET* | 0 | 0 | 27 | 61 | 8 | 1 | −50 | −26 | −25 | −18 | −5 | 0 | |
| *SOIL* | 29 | 29 | 55 | 100 | 100 | 100 | 60 | 47 | 36 | 30 | 29 | 29 | |
| *ΔSOIL* | 0 | 0 | 27 | 45 | 0 | 0 | −40 | −14 | −10 | −6 | −2 | 0 | |
| *ET* | 0 | 0 | 29 | 56 | 90 | 129 | 126 | 113 | 73 | 46 | 26 | 0 | 688 |
| *W-ET-ΔSOIL* | 0 | 0 | 0 | 12 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 12 |

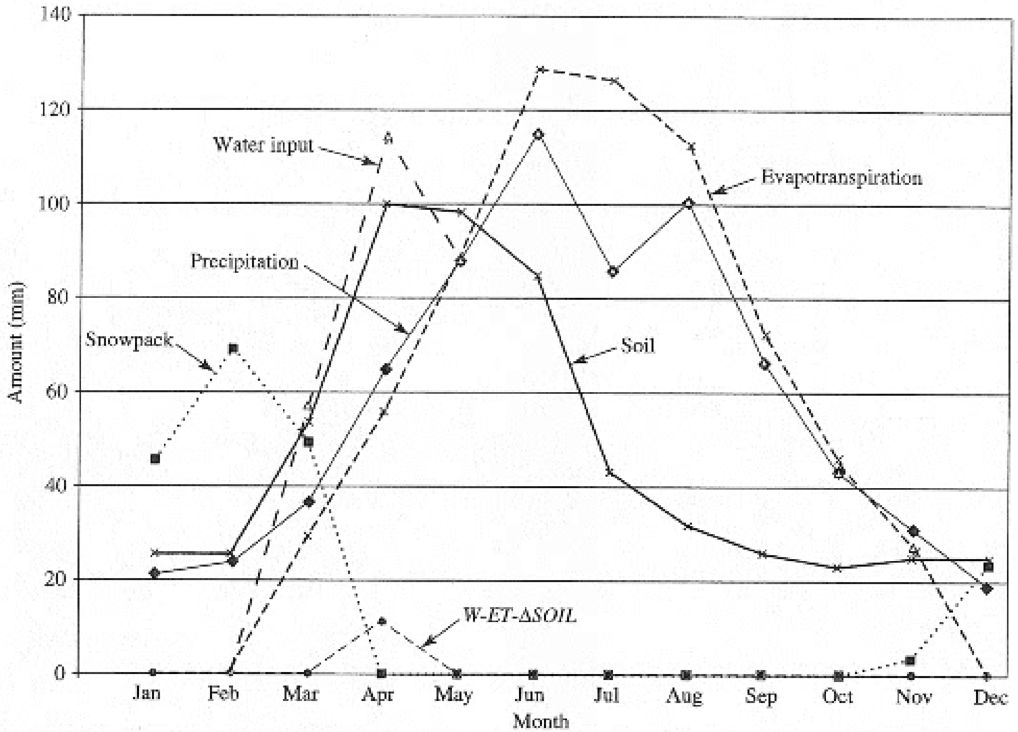
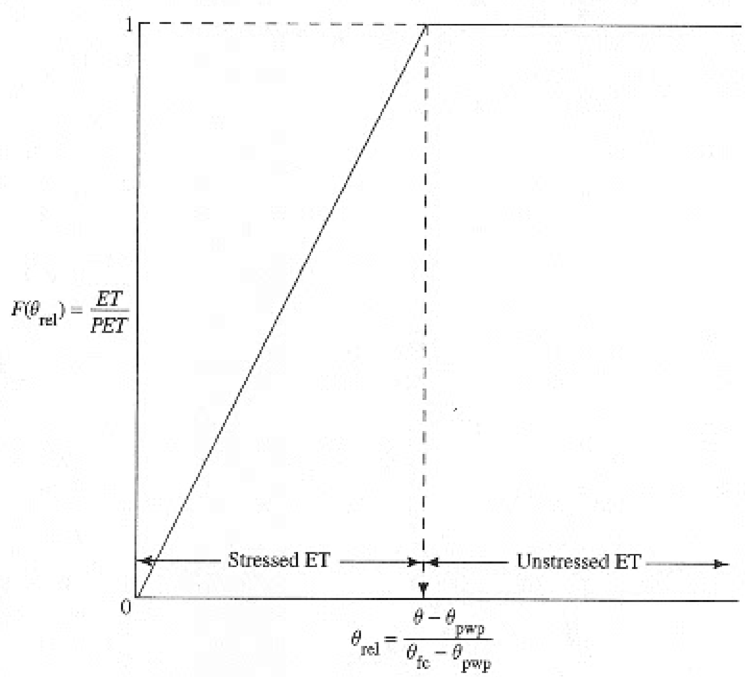



**FIGURE 7-23**
Annual cycle of water-balance components as computed by Thornthwaite-type monthly water-balance model for Omaha, NE (Table 7-9).

**FIGURE 7-24**
General form of relations between $ET/PET$ and soil-water content, $\theta$, used to estimate $ET$. Different studies have used different functions to express soil wetness. When the soil-water content variable is less than the critical value $\theta_{crit}$, $ET$ is less than $PET$ and plants are considered under water stress.



**BOX 7-4**

**Computation of Evapotranspiration in the BROOK90 Model**

BROOK90 uses the Shuttleworth and Wallace (1985) approach to modeling transpiration and soil-water evaporation. Briefly, this approach applies the Penman–Monteith Equation [Equation (7-56)] twice, once to compute potential transpiration for the vegetated fraction of the region (as in Example 7-6) and again for the ground-surface fraction. In computing potential soil evaporation, atmospheric conductance is modified by appropriate adjustments of the surface roughness and zero-plane displacement height, and the surface conductance decreases as surface soil-water content decreases. BROOK90 computes evaporation separately for the daylight and non-daylight hours using different values for air temperature and wind speed and, of course, solar radiation for the two periods. Total daily evapotranspiration is an appropriately weighted sum of daytime and nighttime transpiration and soil evaporation. Actual evapotranspiration and soil evaporation are then determined by multiplying potential values by factors that depend on internal plant resistance to water transport, the maximum potential plants can exert on soil water, and water contents in the various soil layers (Federer 1995).

which Brutsaert and Stricker (1979) called the **advection-aridity** interpretation of the complementary approach. Its main advantage is that it uses readily available meteorological data and does not require calibration to a specific site. It has been found to give estimates of daily ET that compare well with those using other approaches (Figure 7-26; see also Parlange and Katul 1992b).

### 7.8.2 Water-Balance Approaches

Actual evapotranspiration from a region over a time period $\Delta t$ can in principle be determined by measuring water inputs and outputs and changes in storage and solving the water-balance equation, just as for open-water evaporation. The application of this principle to various types of regions is discussed in the following sections; in all cases the precision of the determination is dictated by the precision with which all the other water-balance components can be measured.

#### *Land-Area Water Balance*

As discussed in Section 2.5.2, the most common method of estimating actual evapotranspiration from a land area is the application of the water-balance equation in the form

$$ET = W - Q - G_{out}, \tag{7-74}$$

where $W$ is precipitation, $Q$ is streamflow, and $G_{out}$ is ground-water outflow. The major problems in applying Equation (7-74) were discussed in Section 2.5.2, and include obtaining a reliable estimate of regional precipitation (see also Sections 4.2 and 4.3), obtaining reliable measurements of liquid outflows, especially where ground-water flow is significant (see also Section 8.5), and assuring that changes in storage over the period of measurement are negligible (or are measured and included in the equation).

In regions where most of the storage is in the form of soil water, the assumption of negligible change in storage typically leads to only small errors in estimating ET using data from time periods as short as a few years (Box 2-1; Hudson 1988). Such errors can be further minimized by selecting a water year that begins and ends during the season when soil moisture is near its maximum, as discussed in Section 2.5.2.

Apparently no studies have examined the validity of negligible storage change where other storage components, such as ground water and large lakes, are important. However, the levels of the Great Lakes and Great Salt Lake, U.S., show periods of several decades of steadily declining or rising levels (U.S. Geological Survey 1984). These trends suggest that significant errors are possible in estimating ET from Equation (7-74) for some large drainage basins, even when quantities are averaged over long periods.

#### *Lysimeter Measurement*

A **lysimeter** is an artificially enclosed volume of soil for which the inflows and outflows of liquid water can be measured and, commonly, changes in storage can be monitored by weighing. Lysimeters range from 1 m$^3$ or less to over 150 m$^3$ in size and are usu-

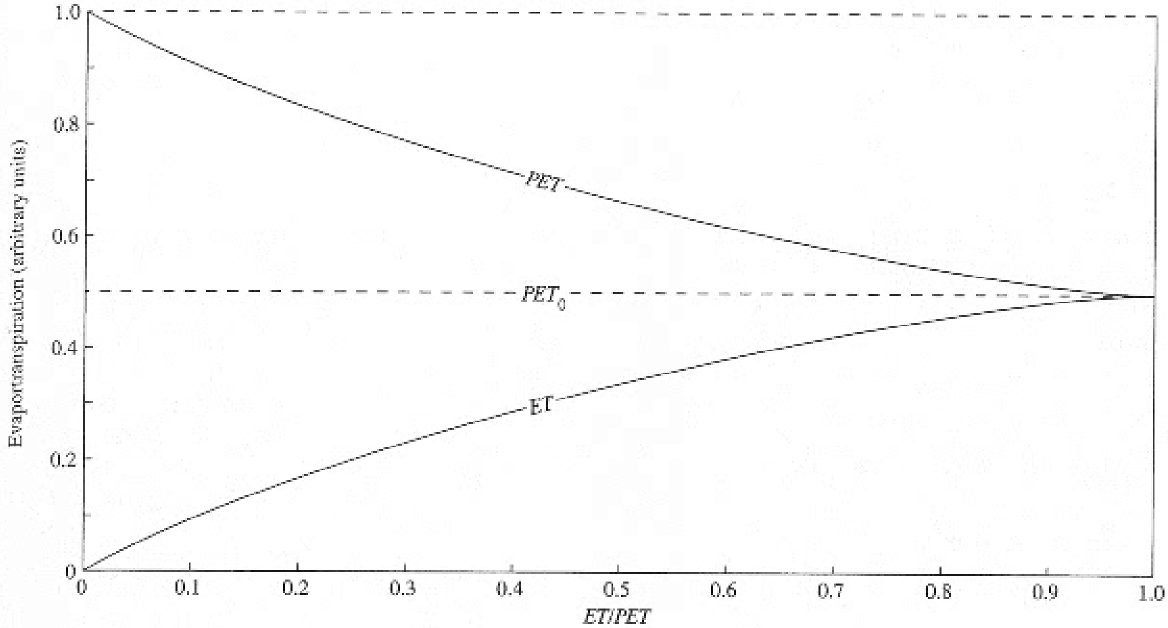



**FIGURE 7-25**
Bouchet's (1963) complementary relationship: $PET + ET = 2 \cdot PET_0$ [Equation (7-72)]. After Brutsaert (1982).

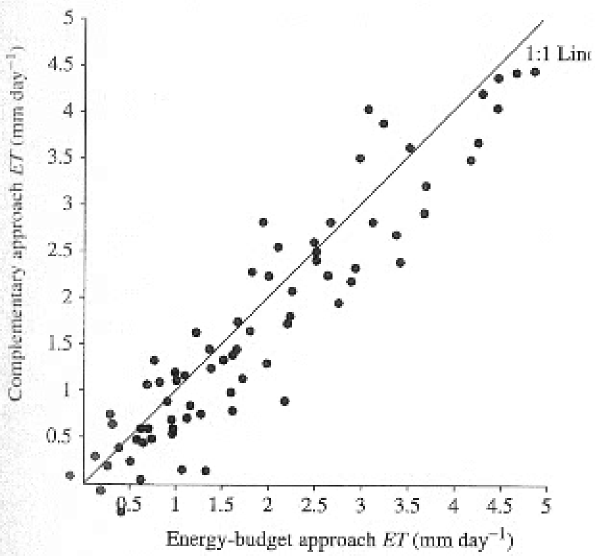



**FIGURE 7-26**
Comparison between estimates of daily *ET* obtained by the complementary (advection-aridity) approach [Equation (7-73)] and an energy-budget method. From Brutsaert (1982) used with permission of Kluwer Academic Publishers.

ally designed so that their soil and vegetation are as closely identical as possible to those of the surrounding area. Details of standard lysimeter construction can be found in Dunne and Leopold (1978), Brutsaert (1982), and Shaw (1988); Grimmond et al. (1992) describe a portable minilysimeter (< 0.2 m$^2$ area).

Carefully obtained lysimeter measurements are usually considered to give the best determinations of actual evapotranspiration during a time period, and are often taken as standards against which other methods are compared (e.g., Jensen et al. 1990). However, lysimeters must have provision for drainage, and the introduction of atmospheric pressure at depth can result in a water-content profile, and hence an evapotranspiration rate, different from those in the surrounding soil. Unfortunately, it is virtually impossible to use the technique for forest vegetation.

### *Soil-Moisture Balance*

One can estimate the total evapotranspiration in a rain-free time period $\Delta t$ by carefully monitoring soil-water content profiles $[\theta(z')]$ at spatially representative locations. As shown in Figure 7-27, the

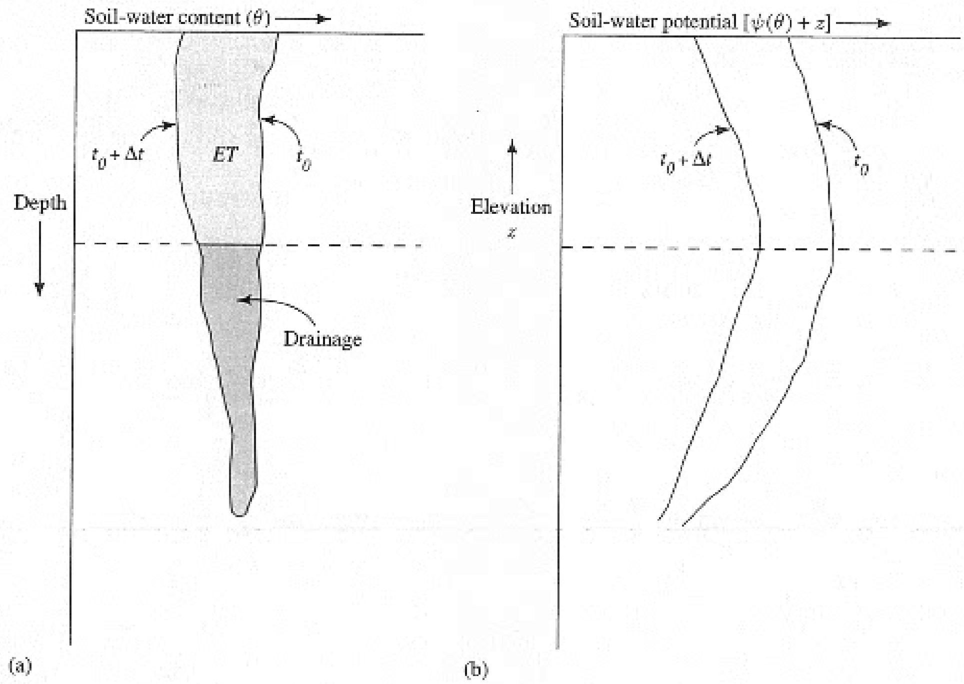



**FIGURE 7-27**
Conceptual basis for estimating evapotranspiration from the soil-water balance. (a) Change in soil-water content with depth during time period $\Delta t$. (b) Profiles of soil-water potential defining the zero-flux plane (dashed line) that divides water lost to evapotranspiration (ET) from that lost to drainage. After Shuttleworth (1992).

total soil-water loss is the difference in water content through the soil profile between times $t_0$ and $t_0 + \Delta t$. The portion of this loss due to evapotranspiration is determined by identifying the "zero-flux plane," which is the boundary between upward-directed water movement due to evapotranspiration and downward-directed movement due to drainage. The average location of the zero-flux plane is found by plotting profiles of the vertical soil-water potential $\psi(\theta) + z$, which is determined from the $\theta(z')$ values and the moisture-characteristic curve for the soil (Section 6.3.3).

This method essentially creates a "lysimeter without walls" that does not distort the soil water-content profile and can be especially useful in forests. However, obtaining representative values of $\theta(z)$ is not easy, and the method will not give good results if the water table is near the surface, if there is horizontal water movement, or if soil properties are highly variable. Rouse and Wilson (1972) found that the minimum length of $\Delta t$ for reliable results is 4 days, and is considerably longer under many conditions.

### *Atmospheric Water Balance*

Evaporation can also be estimated by applying the water-balance equation to a volume of the lower atmosphere. For a control volume of height $z_a$ and perimeter $X$ above an area $A$ and a time interval $\Delta t$, this equation becomes

$$ET = W - \frac{1}{A \cdot \rho_w} \cdot \int_0^{z_a} \int_X (\overline{\rho}_v \cdot \overline{v}_n) \cdot dx \cdot dz - M_2 + M_1, \quad \text{(7-75)}$$

where $ET$ and $W$ are the net evapotranspiration into and precipitation out of the volume (per unit area), $(\overline{\rho}_v \cdot \overline{v}_n)$ is the time-averaged product of the outward-directed wind velocity normal to the perimeter and the absolute humidity, and $M_1$ and $M_2$ are the total water content per unit area of the control volume at the beginning and end of $\Delta t$ respectively.

As summarized by Brutsaert (1982), this method has been applied in several studies using both routine and specially collected atmospheric data. Typically, $7 < z_a < 8$ km. The spatial and temporal coarseness of network upper-air observations limit its routine application to areas of $2.5 \times 10^5$ km$^2$ or more to provide estimates of monthly evaporation. Munley and Hipps (1991) showed the importance of vertical resolution in obtaining accurate estimates.

Several recent research efforts have applied the atmospheric water balance to estimate evapotranspiration from large areas of land, and it appears that the method will play an increasing role in expanding understanding of global-scale hydrology (Shuttleworth 1988; Brutsaert 1988). Kuznetsova (1990) summarized several applications of the approach at the large-river-basin, subcontinental, and continental scales.

### 7.8.3 Turbulent-Transfer/Energy-Balance Methods

#### Penman–Monteith Approach

The Penman-Monteith Equation [Equation (7-56)], with the vegetative canopy treated as a "big leaf" [Equations (7-52)–(7-54)], is commonly used to estimate land-area evapotranspiration. This approach can be refined by treating the vegetated and unvegetated portions of a given area separately, using Equations (7-43)–(7-45) for the bare soil areas. A detailed methodology combining canopy and bare-soil evapotranspiration was developed by Shuttleworth and Wallace (1985); this is the basis for modeling evapotranspiration in the BROOK90 model (Box 7-4).

#### Bowen-Ratio Approach

Direct application of the mass-transfer equation [Equation (7-17)] to estimating actual evapotranspiration from a land surface is generally infeasible because of the absence of surface-temperature data and, most of the time, the absence of a surface that is at saturation.

However, in principle evapotranspiration can be evaluated by applying the mass-transfer equation in the form that makes use of measurements of wind speed and humidity at two levels in the air near the surface [Equation (D-41)]. Dividing both sides of Equation (D-41) by $\lambda_v$ and $\rho_w$ gives

$$ET = \frac{0.622 \cdot \rho_a \cdot (v_2 - v_1) \cdot (e_1 - e_2)}{P \cdot \rho_w \cdot 6.25 \cdot \left[\ln\left(\frac{z_2 - z_d}{z_1 - z_d}\right)\right]^2}, \qquad (7\text{-}76)$$

where the subscripts 1 and 2 refer to measurements at the lower and upper levels, respectively.

However, rather than apply Equation (7-76) directly, we can eliminate the need for wind-speed data and for estimates of the roughness height by making use of the Bowen-ratio approach [as in the development of Equation (7-24)] and an energy-balance relation. This is done using Equation (D-48), which gives the sensible-heat transfer rate, $H$, as

$$H = \frac{c_a \cdot \rho_a \cdot (v_2 - v_1) \cdot (T_1 - T_2)}{6.25 \cdot \left[\ln\left(\frac{z_2 - z_d}{z_1 - z_d}\right)\right]^2}. \qquad (7\text{-}77)$$

From Equations (7-76) and (7-77), the Bowen ratio [Equation (7-11)] is

$$B = \frac{H}{\rho_w \cdot \lambda_v \cdot ET} = \frac{c_a \cdot P \cdot (T_1 - T_2)}{0.622 \cdot \lambda_v \cdot (e_1 - e_2)}. \qquad (7\text{-}78)$$

The energy balance, assuming that the change in heat storage and other terms are negligible, is

$$K + L - \rho_w \cdot \lambda_v \cdot ET - H = 0. \qquad (7\text{-}79)$$

Then making use of the Bowen ratio and solving for $ET$ yields

$$ET = \frac{K + L}{\rho_w \cdot \lambda_v \cdot (1 + B)}, \qquad (7\text{-}80)$$

where $K + L$ is the net radiation and $B$ is calculated from Equation (7-78).

Because of the need for measurements at two levels, Equation (7-80) is useful only in an elaborately instrumented research setting. Furthermore, the approach may not give good results for forest evapotranspiration because the diffusivities of momentum and water vapor may differ significantly over rough surfaces.

#### Eddy-Correlation Approach

The principles of the **eddy-correlation approach** are developed in Section D.6.9, leading to Equation (D-58), namely,

$$ET = \frac{\rho_a}{\rho_w} \cdot \overline{u_a' \cdot q'}\,, \qquad (7\text{-}81)$$

where $u_a'$ and $q'$ are concurrent instantaneous variations in the *vertical* component of wind speed and