The Jensen-Haise (1963) formula, with adjusted units, reads

$$ET_p = (0.025T_a + 0.08)\frac{R_s}{28.6} \quad (5.7)$$

where

ET_p = potential evapotranspiration rate (mm/d)
R_s = incoming short-wave radiation (W/m^2)
T_a = average air temperature at 2 m (°C)

Equations 5.5 and 5.7 generally underestimate ET_p during spring, and overestimate it during summer, because T_a is given too much weight and R_s too little.

5.4.2 Air-Temperature and Day-Length Method

The formula of Blaney-Criddle (1950) was developed for the western part of the U.S.A. (i.e. for a climate of the Mediterranean type). It reads

$$ET_p = k\,p\,(0.457T_{am} + 8.13)(0.031T_{aa} + 0.24) \quad (5.8)$$

where

ET_p = monthly potential evapotranspiration (mm)
k = crop coefficient (–)
p = monthly percentage of annual daylight hours (–)
T_{am} = monthly average air temperature (°C)
T_{aa} = annual average air temperature (°C)

The last term, with T_{aa}, was added to adapt the equation to climates other than the Mediterranean type. The method yields good results for Mediterranean-type climates, but in tropical areas with high cloudiness the outcome is too high. The reason for this is that, besides air temperature, solar radiation plays an important role in evaporation. For more details, see Doorenbos and Pruitt (1977).

More commonly used nowadays are the more physically-oriented approaches (i.e. the Penman and Penman-Monteith equations), which give a much better explanation of the evaporation process.

5.5 Evaporation from Open Water: the Penman Method

The Penman method (1948), applied to open water, can be briefly described by the energy balance at the earth's surface, which equates all incoming and outgoing energy fluxes (Figure 5.4). It reads

$$R_n = H + \lambda E + G \quad (5.9)$$

where

R_n = energy flux density of net incoming radiation (W/m^2)
H = flux density of sensible heat into the air (W/m^2)
λE = flux density of latent heat into the air (W/m^2)
G = heat flux density into the water body (W/m^2)

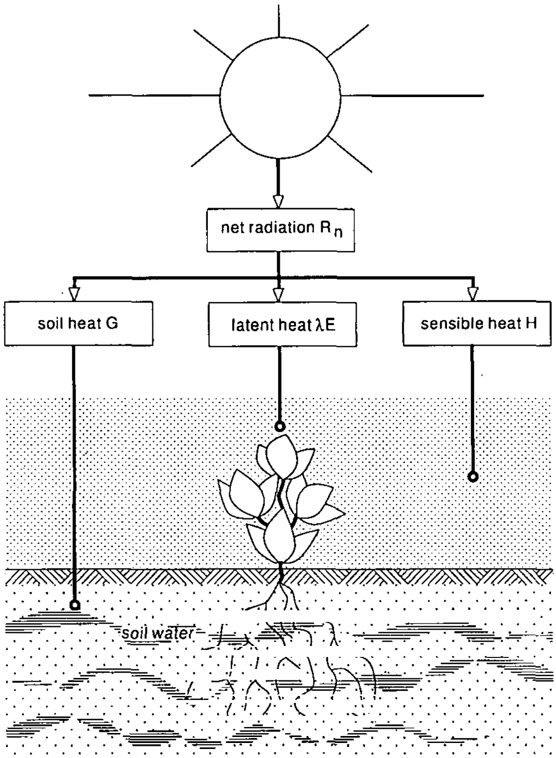

Figure 5.4 Illustration of the variables involved in the energy balance at the soil surface

The coefficient λ in λE is the latent heat of vaporization of water, and E is the vapour flux density in kg/m^2 s. Note that the evapo(transpi)ration in Equation 5.1 is expressed in mm water depth (e.g. over a period of one day). To convert the above λE in W/m^2 into an equivalent evapo(transpi)ration in units of mm/d, λE should be multiplied by a factor 0.0353. This factor equals the number of seconds in a day (86 400), divided by the value of λ (2.45×10^6 J/kg at 20°C), whereby a density of water of 1000 kg/m^3 is assumed.

Supposing that R_n and G can be measured, one can calculate E if the ratio H/λE (which is called the Bowen ratio) is known. This ratio can be derived from the transport equations of heat and water vapour in air.

The situation depicted in Figure 5.4 and described by Equation 5.9 shows that radiation energy, $R_n - G$, is transformed into sensible heat, H, and water vapour, λE, which are transported to the air in accordance with

$$H = c_1 \frac{(T_s - T_a)}{r_a} \tag{5.10}$$

$$\lambda E = c_2 \frac{(e_s - e_d)}{r_a} \tag{5.11}$$

where

c_1, c_2 = constants
T_s = temperature at the evaporating surface (°C)
T_a = air temperature at a certain height above the surface (°C)
e_s = saturated vapour pressure at the evaporating surface (kPa)
e_d = prevailing vapour pressure at the same height as T_a (kPa)
r_a = aerodynamic diffusion resistance, assumed to be the same for heat and water vapour (s/m)

When the concept of the similarity of transport of heat and water vapour is applied, the Bowen ratio yields

$$\frac{H}{\lambda E} = \frac{c_1 (T_s - T_a)}{c_2 (e_s - e_d)} \tag{5.12}$$

where

$c_1/c_2 = \gamma$ = psychrometric constant (kPa/°C)

The problem is that generally the surface temperature, T_s, is unknown. Penman therefore introduced the additional equation

$$e_s - e_a = \Delta (T_s - T_a) \tag{5.13}$$

where the proportionally constant Δ (kPa/°C) is the first derivative of the function $e_s(T)$, known as the saturated vapour pressure curve (Figure 5.5). Note that e_a in Equation 5.13 is the saturated vapour pressure at temperature T_a. Re-arranging gives

$$\Delta = \frac{e_s - e_a}{T_s - T_a} \approx \frac{de_a}{dT_a} \tag{5.14}$$

The slope Δ in Figure 5.5 can be determined at temperature T_a, provided that ($T_s - T_a$) is small.

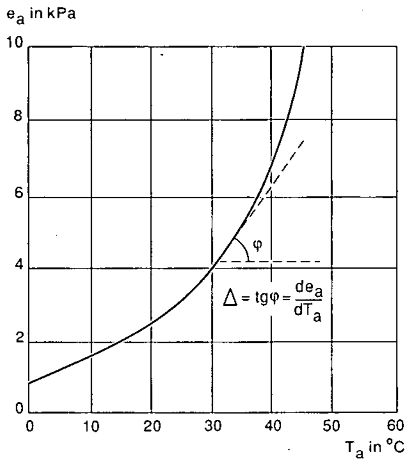

Figure 5.5 Saturated water vapour pressure e_a as a function of air temperature T_a.

From Equation 5.13, it follows that $T_s - T_a = (e_s - e_a)/\Delta$. Substitution into Equation 5.12 yields

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\frac{e_s - e_d}{e_s - e_a} \tag{5.15}$$

If $(e_s - e_a)$ is replaced by $(e_s - e_d - e_a + e_d)$, Equation 5.15 can be written as

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\left[1 - \frac{e_a - e_d}{e_s - e_d}\right] \tag{5.16}$$

Under isothermal conditions (i.e. if no heat is added to or removed from the system), we can assume that $T_s \approx T_a$. This implies that $e_s \approx e_a$. If we then introduce this assumption into Equation 5.11, the isothermal evaporation, λE_a, reads as

$$\lambda E_a = c_2 \frac{e_a - e_d}{r_a} \tag{5.17}$$

Dividing Equation 5.17 by Equation 5.11 yields

$$\frac{E_a}{E} = \frac{e_a - e_d}{e_s - e_d} \tag{5.18}$$

The ratio on the right also appeared in Equation 5.16, which can now be written as

$$\frac{H}{\lambda E} = \frac{\gamma}{\Delta}\left(1 - \frac{E_a}{E}\right) \tag{5.19}$$

From Equation 5.9, it follows that $H = R_n - \lambda E - G$. After some rearrangement, and writing E_o (subscript o denoting open water) for E, substitution into Equation 5.19 yields the formula of Penman (1948)

$$E_o = \frac{\Delta(R_n - G)/\lambda + \gamma E_a}{\Delta + \gamma} \tag{5.20}$$

where

- E_o = open water evaporation rate (kg/m^2 s)
- Δ = proportionality constant de_a/dT_a (kPa/°C)
- R_n = net radiation (W/m^2)
- λ = latent heat of vaporization (J/kg)
- γ = psychrometric constant (kPa/°C)
- E_a = isothermal evaporation rate (kg/m^2 s)

The term $\frac{\Delta}{\Delta + \gamma}(R_n - G)/\lambda$ is the evaporation equivalent of the net flux density of radiant energy to the surface, also called the radiation term. The term $\frac{\Delta}{\Delta + \gamma} E_a$ is the corresponding aerodynamic term. Equation 5.20 clearly shows the combination of the two processes in one formula.

For open water, the heat flux density into the water, G, is often ignored, especially for longer periods. Also note that the resulting E_o in kg/m^2 s should be multiplied by 86 400 to give the equivalent evaporation rate E_o in mm/d.

As was mentioned in Section 5.2, E_o has been used as a kind of reference evaporation for some time, but the practical value of estimating E_o with the original Penman formula (Equation 5.20) is generally limited to large water bodies such as lakes, and, possibly, flooded rice fields in the very early stages of cultivation.

The modification to the Penman method introduced by Doorenbos and Pruitt in FAO's Irrigation and Drainage Paper 24 (1977) started from the assumption that evapotranspiration from grass also largely occurs in response to climatic conditions. And short grass being the common surroundings for agrometeorological observations, they suggested that the evapotranspiration from 8 – 15 cm tall grass, not short of water, be used as a reference, instead of open water. The main changes in Penman's formula to compute this reference evapotranspiration, ET_g (g for grass), concerned the short-wave reflection coefficient (approximately 0.05 for water and 0.25 for grass), a more sensitive wind function in the aerodynamic term, and an adjustment factor to take into account local climatic conditions deviating from an assumed standard. The adjustment was mainly necessary for deviating combinations of radiation, relative humidity, and day/night wind ratios; relevant values can be obtained from a table in Doorenbos and Pruitt (1977).

If the heat flux, G, is set equal to zero for daily periods, the FAO Modified Penman equation can be written as

$$ET_g = c\left[\frac{\Delta}{\Delta + \gamma} R_n' + \frac{\gamma}{\Delta + \gamma} 2.7\, f(u)\,(e_a - e_d)\right] \tag{5.21}$$

where

ET_g = reference evapotranspiration rate (mm/d)
c = adjustment factor (–)
R_n' = equivalent net radiation (mm/d)
f(u) = wind function; $f(u) = 1 + 0.864\, u_2$
u_2 = wind speed (m/s)
$e_a - e_d$ = vapour pressure deficit (kPa)
Δ, γ = as defined earlier

Potential evapotranspiration from cropped surfaces was subsequently found from appropriate crop coefficients, for the determination of which Doorenbos and Pruitt (1977) also provided a procedure.

5.6 Evapotranspiration from Cropped Surfaces

5.6.1 Wet Crops with Full Soil Cover

In analogy with Section 5.5, which described evaporation from open water, evapotranspiration from a wet crop, ET_{wet}, can be described by an equation very similar to Equation 5.20. However, one has to take into account the differences between a crop surface and a water surface:

- The albedo (or reflection coefficient for solar radiation) is different for a crop surface (say, 0.23) and a water surface (0.05 – 0.07);

- A crop surface has a roughness (dependent on crop height and wind speed), and hence an aerodynamic resistance, r_a, that can differ considerably from that of a water surface.

Following the same reasoning as led to Equation 5.17, and replacing the coefficient c_2 by its proper expression, we can write E_a for a crop as

$$E_a = \frac{\varepsilon \rho_a}{p_a} \frac{(e_a - e_d)}{r_a} \tag{5.22}$$

where

ε = ratio of molecular masses of water vapour and dry air (–)
p_a = atmospheric pressure (kPa)
ρ_a = density of moist air (kg/m^3)

For a wet crop surface with an ample water supply, the Penman equation (5.20) can then be modified (Monteith 1965; Rijtema 1965) to read

$$ET_{wet} = \frac{\frac{\Delta(R_n - G)}{\lambda} + \gamma \frac{\varepsilon \rho_a}{p_a} \frac{(e_a - e_d)}{r_a}}{\Delta + \gamma} \tag{5.23}$$

Because the psychrometric constant $\gamma = c_p\, p_a/\lambda\varepsilon$, Equation 5.23 reduces to

$$ET_{wet} = \frac{\Delta(R_n - G) + c_p \rho_a \frac{(e_a - e_d)}{r_a}}{(\Delta + \gamma)\lambda} \tag{5.24}$$

where

ET_{wet} = wet-surface crop evapotranspiration rate (kg/m^2 s)
c_p = specific heat of dry air at constant pressure (J/kg K)

This ET_{wet} can easily be converted into equivalent mm/d by multiplying it by 86 400.

Note that evapotranspiration from a completely wet crop/soil surface is not restricted by crop or soil properties. ET_{wet} thus primarily depends on the governing atmospheric conditions.

5.6.2 Dry Crops with Full Soil Cover : the Penman-Monteith Approach

Following the discussion of De Bruin (1982) on Monteith's concept for a dry vegetated surface, we can treat the vegetation layer simply as if it were one big leaf. The actual transpiration process (liquid water changing into vapour) takes place in cavities below the stomata of this 'big leaf', and the air within these cavities will be saturated (pressure e_o) at leaf temperature, T_s (Figure 5.6). Water vapour escapes through the stomata to the outer 'leaf' surface, where a certain lower vapour pressure reigns. It is assumed that this vapour pressure at leaf temperature T_s equals the saturated vapour pressure e_a at air temperature T_a. During this diffusion, a 'big leaf' stomatal resistance, r_c, is

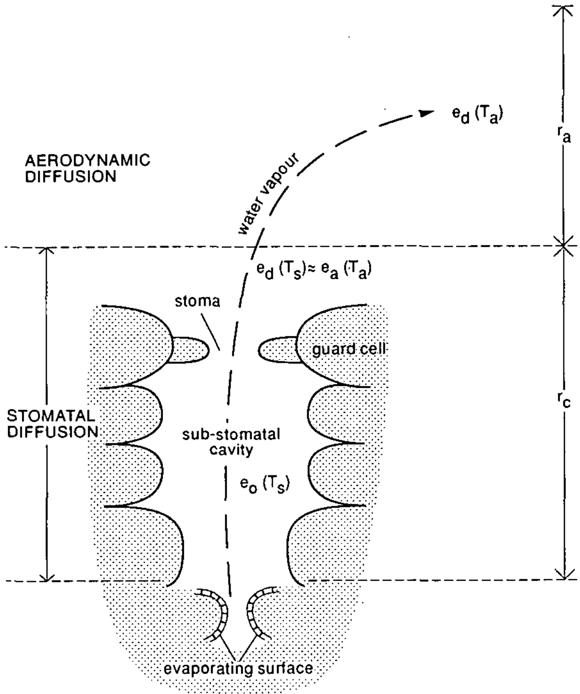

Figure 5.6 The path of water vapour through a leaf stoma, showing relevant vapour pressures, temperatures, and resistances

encountered. As the vapour subsequently moves from the leaf surface to the external air, where actual vapour pressure, e_d, is present, an aerodynamic resistance is encountered. When the vapour diffusion rate through the stomata equals the vapour transport rate into the external air, we can write

$$E_a = \frac{\varepsilon \rho_a}{p_a} \frac{e_o - e_a}{r_c} = \frac{\varepsilon \rho_a}{p_a} \frac{e_a - e_d}{r_a} = \frac{\varepsilon \rho_a}{p_a} \frac{e_o - e_d}{r_c + r_a} \tag{5.25}$$

where, in addition to the earlier defined ε, ρ_a, and p_a

E_a = isothermal evapotranspiration rate from the canopy (kg/m^2 s)
e_o = internal saturated vapour pressure at T_s (kPa)
e_a = saturated vapour pressure at the 'leaf' surface at T_a (kPa)
e_d = vapour pressure in the external air (kPa)
r_a = aerodynamic resistance (s/m)
r_c = canopy diffusion resistance (s/m)

From Equation 5.25, it follows that a canopy with r_c can be formally described with the same equation as ET_{wet}, if the vapour pressure difference ($e_a - e_d$) in Equation 5.24 is replaced by

$$e_a - e_d = \frac{e_o - e_d}{1 + \frac{r_c}{r_a}} \tag{5.26}$$

According to Monteith (1965), the same effect is obtained if γ is replaced by γ^*

$$\gamma^* = \gamma\left(1 + \frac{r_c}{r_a}\right) \tag{5.27}$$

The equation of Monteith for a dry vegetation then reads

$$ET = \frac{\dfrac{\Delta\,(R_n - G)}{\lambda} + \dfrac{c_p\rho_a}{\lambda}\dfrac{e_a - e_d}{r_a}}{\Delta + \gamma^*} \tag{5.28}$$

where

ET = evapotranspiration rate from a dry crop surface (kg/m^2 s)
γ^* = modified psychrometric constant (kPa/°C)

This Penman-Monteith equation is valid for a dry crop completely shading the ground.

Note that for a wet crop covered with a thin water layer, r_c becomes zero and the wet-crop formulation (Equation 5.24) is obtained again.

Equation 5.28 is, in principle, not able to describe the evapotranspiration from sparsely-cropped surfaces. With a sparsely-cropped surface, the evaporation from the soil might become dominant.

It appears that the canopy resistance, r_c, of a dry crop completely covering the ground has a non-zero minimum value if the water supply in the rootzone is optimal (i.e. under conditions of potential evapotranspiration). For arable crops, this minimum amounts to r_c = 30 s/m; that of a forest is about 150 s/m.

The canopy resistance is a complex function of incoming solar radiation, water vapour deficit, and soil moisture. The relationship between r_c and these environmental quantities varies from species to species and also depends on soil type. It is not possible to measure r_c directly. It is usually determined experimentally with the use of the Penman-Monteith equation, where ET is measured independently (e.g. by the soil water balance or micro-meteorological approach). The problem is that, with this approach, the aerodynamic resistance, r_a, has to be known. Owing to the crude description of the vegetation layer, this quantity is poorly defined. It is important, however, to know where to determine the surface temperature, T_s. Because, in a real vegetation, pronounced temperature gradients occur, it is very difficult to determine T_s precisely. In many studies, r_a is determined very crudely. This implies that some of the r_c values published in literature are biased because of errors made in r_a (De Bruin 1982).

Alternatively, one sometimes relates r_c to single-leaf resistances as measured with a porometer, and with the leaf area index, I_l, according to $r_c = r_{leaf}/0.5I_l$. If such measurements are not available, a rough indication of r_c can be obtained from taking r_{leaf} to be 100 s/m.

The aerodynamic resistance, r_a, can be represented as

$$r_a = \frac{\ln\left(\dfrac{z - d}{z_{om}}\right)\ln\left(\dfrac{z - d}{z_{ov}}\right)}{K^2 u_z} \tag{5.29}$$

where

z = height at which wind speed is measured (m)
d = displacement height (m)
z_{om} = roughness length for momentum (m)
z_{ov} = roughness length for water vapour (m)
K = von Kármán constant (–); equals 0.41
u_z = wind speed measured at height z (m/s)

One recognizes in Equation 5.29, the wind speed, u, increasing logarithmically with height, z. The canopy, however, shifts the horizontal asymptote upwards over a displacement height d, and u_z becomes zero at a height $d + z_o$ (Figure 5.7). Displacement d is dependent on crop height h and is often estimated as

$$d = 0.67\,h; \text{ with } z_{om} = 0.123\,h; \text{ and } z_{ov} = 0.1\,z_{om}$$

In practice, Equation 5.28 is often applied to calculate potential evapotranspiration ET_p, using the mentioned minimum value of r_c and the relevant value of r_a. It can also be used to demonstrate the effect of a sub-optimal water supply to a crop. The reduced turgor in the leaves will lead to a partial closing of the stomata, and thus to an increase in the canopy resistance, r_c. A higher r_c leads to a higher γ^*, and consequently to a lower ET than ET_p.

The superiority of the Penman-Monteith approach (Equation 5.28) over the FAO Modified Penman approach (Equation 5.21) is clearly shown in Figure 5.8. The Penman-Monteith estimates of monthly evapotranspiration of grass or alfalfa agreed better with lysimeter-measured values than FAO Modified Penman estimates.

Equation 5.28 is also used nowadays to calculate a reference evapotranspiration, ET_h. The reference crop is then the aforementioned (Section 5.2) hypothetical crop, with a canopy resistance r_c, and fully covering the ground. This crop is not short of water, so that the minimum r_c of 70 s/m applies. It has a crop height of 12 cm, so that the displacement height d and also the roughness lengths z_{om} and z_{ov} are fixed. For the standard measuring height z = 2 m and applying Equation 5.29 we find that

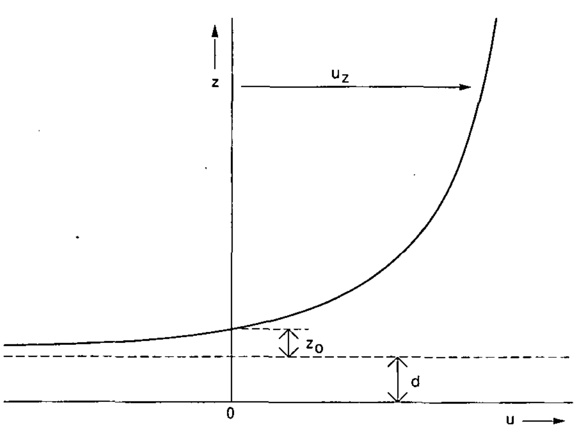

Figure 5.7 The aerodynamic wind profile, illustrating the displacement, d, and the roughness length, z_o

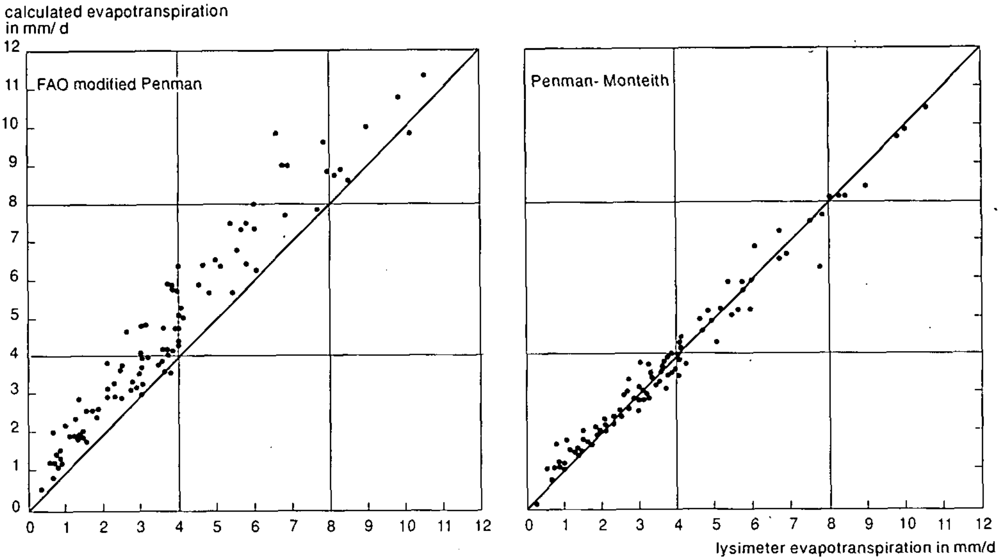

Figure 5.8 Comparison of monthly average lysimeter data for 11 locations with computed evapotranspiration rates for the FAO Modified Penman method and the Penman-Monteith approach (after Jensen et al. 1990)

$r_a = 208/u_2$. In that case, $\gamma^* = (1 + 0.337\ u_2)\gamma$. These values and values for other constants can be entered into Equation 5.28, which then produces, with the proper meteorological data, a value for the reference evapotranspiration, denoted by ET_h (see Section 5.7.2).

Potential evapotranspiration from other cropped surfaces could be calculated with minimum values of r_c and the appropriate crop height. As long as minimum r_c values are not available, one may use the above reference evapotranspiration, ET_h, and multiply it by the proper crop coefficient to arrive at the ET_p of that particular crop, as will be discussed further in Section 5.7.1.

5.6.3 Partial Soil Cover and Full Water Supply

If, under the governing meteorological conditions, enough water is available for evapotranspiration from the soil and the crop (and if the meteorological conditions are unaffected by the evapotranspiration process itself), we may consider evapotranspiration to be potential: ET_p. Hence, we can write

$$ET_p = E_p + T_p \tag{5.30}$$

where

E_p = potential soil evaporation
T_p = potential plant transpiration

As argued before, the Penman-Monteith approach (Equation 5.28) works only under the condition of a complete soil cover.

If we want to estimate the potential evaporation of a soil under a crop cover, we can compute it from a simplified form of Equation 5.24 by neglecting the aerodynamic term and taking into account only that fraction of R_n which reaches the soil surface (Ritchie 1972)

$$E_p = \frac{\Delta}{(\Delta + \gamma)\lambda} R_n e^{-kI_l} \qquad (5.31)$$

where

- E_p = potential soil evaporation rate (kg/m^2 s)
- R_n = net radiation flux density reaching the soil (W/m^2)
- I_l = leaf area index (m^2 leaf area/m^2 soil area) (–)
- k = a proportionality factor, which may vary according to the geometrical properties of a crop (–)

Ritchie (1972) took $k = 0.39$ for crops like sorghum and cotton; Feddes et al. (1978) applied this value to crops like potatoes and grass. More recent views are based on considerations of the extinction coefficient for diffuse visible light, K_D, which varies with crop type from 0.4 to 1.1. A satisfactory relationship for k might be $k = 0.75 K_D$.

By subtracting E_p (Equation 5.31) from ET_p obtained through Equation 5.28, using minimum r_c values, we can then derive T_p from Equation 5.30 as $T_p = ET_p - E_p$. On soils with partial soil cover (e.g. row crops in their early growth stage), the condition of the soil – dry or wet – will considerably influence the partitioning of ET_p over E_p and T_p. Figure 5.9 gives an idea of the computed variation of T_p/ET_p as a function of the leaf area index, I_l, for a potato crop with optimum water supply to the roots for a dry and a wet soil, respectively, as computed by the simulation program SWATRE of Belmans et al. (1983).

If we assume that ET_p is the same for both dry and wet soil, it appears that for $I_l < 1$, with increasing drying of the soil and thus decreasing E_p, T_p will increase by a factor of approximately 1.5 to 2 per unit I_l. For $I_l > 2$-2.5, E_p is small and virtually independent of the moisture condition of the soil surface. This result agrees with the findings on red cabbage by Feddes (1971) that the soil must be covered for about 70 to 80% ($I_l = 2$) before E_p becomes constant. Similar results are reported for measurements on sorghum and cotton.

The above results show that the Penman-Monteith approach (Equation 5.28) can be considered reasonably valid for leaf area indices $I_l > 2$. Below this value, one can regard it as a better-than-nothing approximation.

Note: The partitioning of ET_p into T_p and E_p is important if one is interested in the effects of water use on crop growth and crop production. Crop growth is directly related to transpiration. (For more details, see Feddes 1985.)

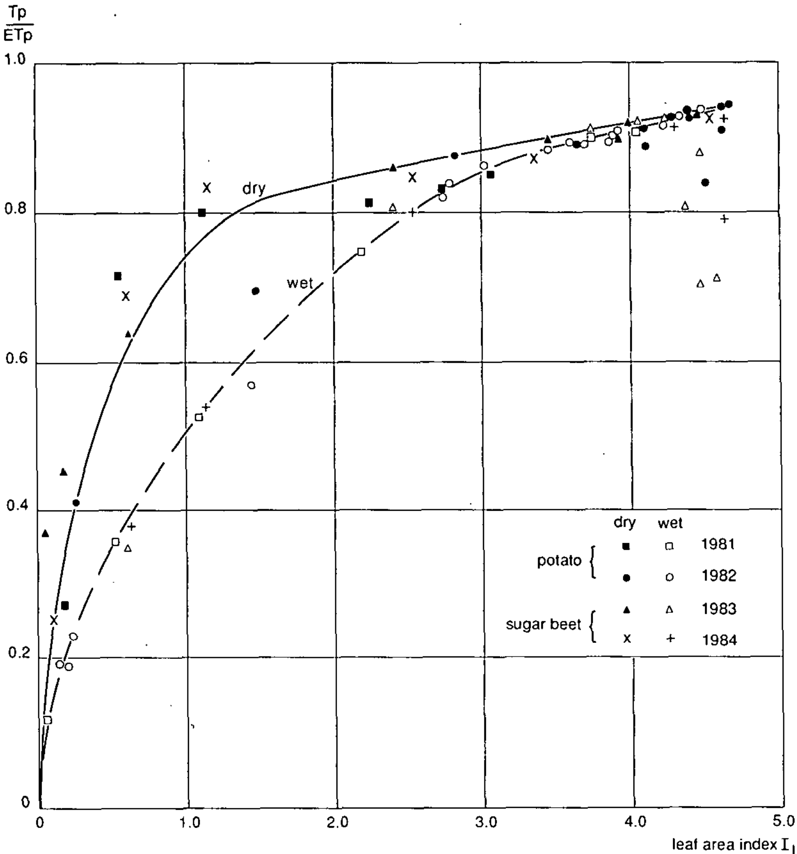

Figure 5.9 Potential transpiration, T_p, as a fraction of potential evapotranspiration, ET_p, in relation to the leaf area index, I_l, for a daily-wetted soil surface and for a dry soil surface

5.6.4 Limited Soil-Water Supply

Under limited soil-water availability, evapotranspiration will be reduced because the canopy resistance increases as a result of the partial closure of the stomata. Such a limitation in available soil water occurs naturally if soil water extracted from the rootzone by evapotranspiration is not replenished in time by rainfall, irrigation, or capillary rise. Another reason for a reduced water availability is a high soil-water salinity, whereby the osmotic potential of the soil solution prevents water from moving to the roots in a sufficient quantity.

Actual evapotranspiration, ET, can be determined from soil water balances by lysimetry, and with micro-meterological techniques, as were discussed in Section 5.3.

For large areas, remote sensing can provide an indirect measure of ET. Using reflection images to detect the type of crop, and thermal infra-red images from satellite or airplane observations for crop surface temperatures, one can transform these data into daily evapotranspiration rates using surface-energy-balance models (e.g. Thunnissen and Nieuwenhuis 1989; Visser et al. 1989). The underlying principle is that, for the same crop and growth stage, a below-potential evapotranspiration means a partial closure of the stomata (and increased r_c), a lower transpiration rate inside the sub-stomatal cavities, and hence a higher leaf/canopy temperature (Section 5.6.2).

Another way to estimate ET is by using a soil-water-balance model such as SWATRE (Feddes et al. 1978; Belmans et al. 1983), which describes the transient water flow in the heterogenous soil-root system that may or may not be influenced by groundwater.

An example of the output of such a model is presented in Figure 5.10. It shows the water-balance terms of the rootzone and the subsoil of a sandy soil that was covered with grass during the very dry year 1976 in The Netherlands. A relatively shallow watertable was present. Over 1976, the potential evapotranspiration, ET_p, was 502 mm, actual ET was 361 mm, which implies a strong reduction of potential evapotranspiration. Net infiltration, I, amounted to 197 mm. Water extraction from the rootzone in this rather light soil was 56 mm, which is only 16% of ET. The decrease in water storage in the subsoil amounted to 206 mm, of which 107 mm (30% of ET) had been delivered by capillary rise towards the rootzone, and 99 mm had been lost to the saturated zone by deep percolation.

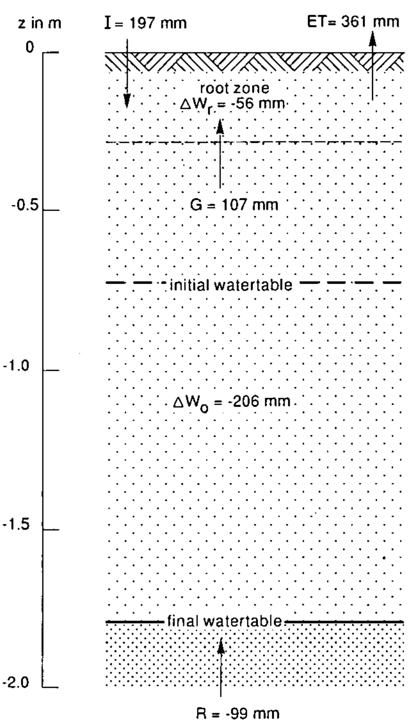

Figure 5.10 Schematic presentation of the water balance terms (mm) of the rootzone (0-0.3 m) and the subsoil (0.3-2.0 m) of a sandy soil over the growing season (1 April – 1 October) of the very dry year 1976 in The Netherlands. The watertable dropped from 0.7 m to 1.8 m during the growing season (after De Graaf and Feddes 1984)

The input data for SWATRE consist of:
- Data on the hydraulic conductivity and moisture retention curves of the major soil horizons;
- Rooting depths and watertables (if present);
- Calculated potential evapotranspiration;
- Precipitation and/or irrigation.

If such a water-balance model is coupled with a crop-growth and crop production model, the actual development of the crop over time can be generated. Hence, actual evapotranspiration can be determined, depending on the every-day history of the crop. Such a model can be helpful in irrigation scheduling, but it can also be used to analyze drainage situations.

5.7 Estimating Potential Evapotranspiration

5.7.1 Reference Evapotranspiration and Crop Coefficients

To estimate crop water requirements, one can relate ET_p from the crop under consideration to an estimated reference evapotranspiration, ET_{ref}, by means of a crop coefficient

$$ET_p = k_c \, ET_{ref} \tag{5.32}$$

where

ET_p = potential evapotranspiration rate (mm/d)
k_c = crop coefficient (–)
ET_{ref} = reference evapotranspiration rate (mm/d)

The reference evapotranspiration could, in principle, be any evaporation parameter, such as pan evaporation, the Blaney-Criddle ET (Equation 5.8 without the crop coefficient, k), the Penman open water evaporation, E_o (Equation 5.20), the FAO Modified Penman ET_g (Equation 5.21), or the Penman-Monteith ET_h (Equation 5.28).

For the calculation of ET_g and the corresponding crop coefficients, extensive procedures have been given by Doorenbos and Pruitt (1977). Smith (1990) concluded that the sound and practical methods of determining crop water requirements as introduced by Doorenbos and Pruitt (1977) are to a large extent still valid. And so, too, are their lists of crop factors for various crops at different growth stages, if used in combination with ET_g.

In the Penman-Monteith approach, we do not have sufficient data on minimum canopy resistance to apply Equation 5.28 generally, by inserting crop-specific minimum r_c values. Therefore, for the time being, a two-step approach may be followed, in which we represent the effects of climate on potential evapotranspiration by first calculating ET_h, and adding a crop coefficient to account for crop-specific influences on ET_p.

In the two-step approach, the crop coefficient, k_c, depends not only on the characteristic of the crop, its development stage, and the prevailing meteorological conditions, but also on the selected ET_{ref} method. Choosing the Penman-Monteith approach means that crop coefficients related to this method should be used.

Although it is recognized that alfalfa better resembles an average field crop, the new hypothetical reference crop closely resembles a short, dense grass cover, because most standard meteorological observations are made in grassed meteorological enclosures. In this way, the measured evapotranspiration of (reference) crops used in the various lysimeter and other evaporation studies (grass, alfalfa, Kikuyu grass) can be more meaningfully converted to the imaginary reference crop in the Penman-Monteith approach.

Standardization of certain parameters in the Penman-Monteith equation has led to the following definition (Smith 1990):

> 'The reference evapotranspiration, ET_h, is defined as the rate of evapotranspiration from an hypothetical crop with an assumed crop height (12 cm), and a fixed canopy resistance (70 s/m), and albedo (0.23), which would closely resemble evapotranspiration from an extensive surface of green grass cover of uniform height, actively growing, completely shading the ground, and not short of water.'

Procedures to calibrate measured potential evapotranspiration to the newly-adopted standard ET_h values in accordance with the above definition are then required.

To convert the Doorenbos and Pruitt (DP) crop factors, k_c^{DP}, to new crop factors, k_c^{PM}, and supposing that ET_p is the same in both cases, we can write

$$ET_p = k_c^{DP}\, ET_g = k_c^{PM}\, ET_h \tag{5.33}$$

from which

$$k_c^{PM} = \frac{ET_g}{ET_h}\, k_c^{DP} \tag{5.34}$$

The conversion factor ET_g/ET_h can easily be derived from long-term meteorological records (e.g. per 10-day period).

Note that crop factors are generally derived from fields with different local conditions and agricultural practices. These local effects may thus include size of fields, advection, irrigation and cultivation practices, climatological variations in time, distance, and altitude, and soil water availability. One should therefore always be careful in applying crop coefficients from experimental data.

As mentioned above, ET_{ref} is sometimes estimated with the pan evaporation method. Extensive use and testing of the evaporation from standardized evaporation pans such as the Class A pan have shown the great sensitivity of the daily evaporation of the water in the pan. It can be influenced by a range of environmental conditions such as wind, soil-heat flux, vegetative cover around the pan, painting and maintenance conditions, or the use of screens. Using the pan evaporation method to estimate reference evapotranspiration can only be recommended if the instrumentation and the site are properly calibrated and managed.

5.7.2 Computing the Reference Evapotranspiration

Accepting the definition of the reference crop as given in Section 5.7.1, we can find the reference evapotranspiration from the following combination formula, which is based on the Penman-Monteith approach (Verhoef and Feddes 1991)

$$ET_h = \frac{\Delta}{\Delta + \gamma^*} R_n' + \frac{\gamma}{\Delta + \gamma^*} E_a \qquad (5.35)$$

where

ET_h = reference crop evapotranspiration rate (mm/d)
Δ = slope of vapour pressure curve at T_a (kPa/°C)
γ = psychrometric constant (kPa/°C)
γ^* = modified psychrometric constant (kPa/°C)
R_n' = radiative evaporation equivalent (mm/d)
E_a = aerodynamic evaporation equivalent (mm/d)

This formula is generally applicable, but, to apply it in a certain situation, we have to know what meteorological data are available. As was indicated in Table 5.1, the Penman-Monteith approach requires data on air temperature, solar radiation, relative humidity, wind speed, aerodynamic resistance, and basic canopy resistance. For the computation method that will be presented in this section, we assume that we have the following information:

- General information:
  - The latitude of the station in degrees (positive for northern latitudes and negative for southern latitudes);
  - The altitude of the station above sea level;
  - The measuring height of wind speed and other data is 2 m above ground level;
  - The month of the year for which we want to compute the reference evapotranspiration;
- Crop-specific information:
  - The canopy resistance equals 70 s/m;
  - The crop height is 12 cm;
  - The reflection coefficient equals 0.23;
- Meteorological data:
  - Minimum and maximum temperatures (°C);
  - Solar radiation (W/m²);
  - Relative duration of bright sunshine (–);
  - Average relative humidity (%);
  - Wind speed (m/s).

To this situation, we apply the following computation procedure.

The weighting terms $\Delta/(\Delta + \gamma^*)$ and $\gamma/(\Delta + \gamma^*)$ in front of the radiation and aerodynamic evapotranspiration terms of Equation 5.35 contain γ, γ^*, and Δ. These variables are found as follows.

The psychrometric constant, γ

$$\gamma = 1615 \frac{p_a}{\lambda} \tag{5.36}$$

where

p_a = atmospheric pressure (kPa)
λ = latent heat of vaporization (J/kg); value 2.45 x 10^6
1615 = c_p/ε, or 1004.6 J/kg K divided by 0.622

The atmospheric pressure is related to altitude

$$p_a = 101.3 \left(\frac{T_a + 273.16 - 0.0065H}{T_a + 273.16} \right)^{5.256} \tag{5.37}$$

where

H = altitude above sea level (m)

The modified psychrometric constant, γ^*, can be found from Equation 5.27. We can insert the standard value of 70 s/m for the reference crop and use Equation 5.29 to find r_a. With the appropriate values, we find $r_a = 208/u_2$, so that

$$\gamma^* = (1 + 0.337\, u_2)\gamma \tag{5.38}$$

The slope of the vapour pressure curve, Δ

$$\Delta = \frac{4098\, e_a}{(T_a + 237.3)^2} \tag{5.39}$$

where

T_a = average air temperature (°C); $T_a = (T_{max} + T_{min})/2$
e_a = saturated vapour pressure (kPa), which follows from

$$e_a = 0.6108 \exp\left(\frac{17.27\, T_a}{T_a + 237.3} \right) \tag{5.40}$$

The radiative evaporation equivalent follows from

$$R'_n = 86400 \frac{R_n - G}{\lambda} \tag{5.41}$$

where

R_n = net radiation at the crop surface (W/m^2)
G = heat flux density to the soil (W/m^2); zero for periods of 10-30 days
λ = latent heat of vaporization (J/kg); value 2.45 × 10^6

Note that the number of seconds in a day (86 400) appears, and that the density of water (1000 kg/m^3) has been omitted on the right, because it is numerically cancelled out by the conversion from m to mm.

Net radiation is composed of two parts: net short-wave and net long-wave radiation: $R_n = R_{ns} - R_{nl}$. Net short-wave radiation can be described by

$$R_{ns} = (1 - \alpha)R_s \tag{5.42}$$

where

R_{ns} = net short-wave radiation (W/m^2)
α = albedo, or canopy reflection coefficient (-); value 0.23 for the standard reference crop
R_s = solar radiation (W/m^2)

The net long-wave radiation is represented by

$$R_{nl} = (0.9\frac{n}{N} + 0.1)(0.34 - 0.139\sqrt{e_d})\,\sigma\frac{(TK_{max}^4 + TK_{min}^4)}{2} \qquad (5.43)$$

where

R_{nl} = net long-wave radiation (W/m^2)
n = daily duration of bright sunshine (h)
N = day length (h)
e_d = actual vapour pressure (kPa)
TK_{max} = maximum absolute temperature (K)
TK_{min} = minimum absolute temperature (K)
σ = Stefan-Boltzmann constant ($W/m^2\,K^4$); equals 5.6745×10^{-8}

The actual vapour pressure, e_d, is found from

$$e_d = \frac{RH}{100}e_a \qquad (5.44)$$

where

RH = relative humidity percentage (–)

The aerodynamic evaporation equivalent is computed from

$$E_a = \frac{900}{(T_a + 275)}u_2(e_a - e_d) \qquad (5.45)$$

where

u_2 = wind speed measured at 2 m height (m/s)
e_a = saturated vapour pressure (kPa)
e_d = actual vapour pressure (kPa)

We arrive at Equation 5.45 by applying Equation 5.25, with $(e_a - e_d)$. The ratio of the molecular masses of water vapour and dry air equals 0.622. In addition, the density of moist air can be expressed as

$$\rho_a = \frac{p_a}{0.287\,(T_a + 275)} \qquad (5.46)$$

in which 0.287 equals R_a, the specific gas constant for dry air (0.287 kJ/kg K), and where the officially needed virtual temperature has been replaced by its approximate equivalent $(T_a + 275)$. Moreover, we can find r_a from Equation 5.29 by applying the standard measuring height of 2 m and the reference crop height of 0.12 m, which gives, as was indicated in Section 5.6.2, $r_a = 208/u_2$. Hence, calculating $0.622 \times 86400 / 0.287 \times 208$ produces the factor 900.

The vapour pressure deficit in the aerodynamic term is $e_a - e_d$.

This calculation procedure may seem cumbersome at first, but scientific calculators and especially micro-computers can assist in the computations. Micro-computer programs that use the above equations to find the reference evapotranspiration are available. One example is the program REF-ET, which is a reference evapotranspiration calculator that calculates ET_{ref} according to eight selected methods (Allen 1991). These methods include Penman's open water evaporation, the FAO Modified Penman method, and also the Penman- Monteith approach. The program CROPWAT (Version 5.7) not only calculates the Penman-Monteith reference ET, but also allows a selection of crop coefficients to arrive at crop water requirements (Smith 1992). The program further helps in calculating the water requirements for irrigation schemes and in irrigation scheduling. For this program, a suitable database (CLIMWAT) with agro-meteorological data from many stations around the world is available. Verhoef and Feddes (1991) produced a micro-computer program in FORTRAN, which allows the rapid calculation of the reference crop evapotranspiration according to nine different methods, including the Penman-Monteith equation, and for a variety of available data.

The above mentioned computation methods contain a few empirical coefficients, which may be estimated differently by different authors. In the Penman-Monteith crop reference procedure presented here, however, we have used the recommended relationships and coefficients (Smith 1990), as were also used by Shuttleworth (1992). This procedure should reduce any still-existing confusion.

Calculation Examples

Table 5.2 shows the results of applying the above procedure to one year's monthly data from two meteorological stations in existing drainage areas: one in Mansoura, Egypt, and the other in Hyderabad, Pakistan, both from the database used by Verhoef and Feddes (1991). The relevant input data are listed as well as the calculated reference evapotranspiration.

A comparison of the ET_h-values for the two stations clearly shows the importance of wind speed, or, more generally, of the aerodynamic term. Radiation, sunshine duration, and temperatures do not differ greatly at the two stations, yet the ET_h for Hyderabad is up to twice that for Mansoura. This is mainly due to a large difference in wind speed, and, to a lesser extent, in relative humidity, which together determine the aerodynamic term.

It should be realized that the described procedure would be slightly different for other data availability. If solar radiation is not measured, R_s can be estimated from sunshine duration and radiation at the top of the atmosphere (extra-terrestrial radiation). Also, if relative humidity data are not available, the actual vapour pressure can be estimated from approximate relationships. Minimum and maximum temperatures may not be available, but only averages. Such different data conditions can be catered for (see e.g. Verhoef and Feddes 1991). We shall not mention all possible cases. The main computational structure for finding 10-day or monthly average ET_h-values has been adequately described above, and only one different condition (i.e. that of missing data on solar radiation) is discussed below.

Table 5.2 Computed reference evapotranspiration for two meteorological stations, following the described Penman-Monteith procedure

Month	T_{min} (°C)	T_{max} (°C)	R_s (W/m^2)	n/N (-)	RH (%)	u_2 (m/s)	ET_h (mm/d)
Mansoura, Egypt (Altitude 30 m)							
January	7.0	19.5	133	0.69	68	1.3	1.5
February	7.5	20.5	167	0.71	59	1.4	2.2
March	9.3	23.2	212	0.73	61	1.7	3.1
April	12.0	27.1	250	0.75	51	1.5	4.1
May	15.6	33.2	279	0.78	43	1.5	5.3
June	18.6	33.6	303	0.85	55	1.5	5.6
July	20.5	32.6	295	0.84	66	1.3	5.2
August	20.5	33.5	280	0.86	66	1.3	5.0
September	19.0	32.5	245	0.85	61	1.1	4.2
October	17.1	28.7	200	0.83	63	1.0	3.0
November	14.0	25.8	153	0.77	63	1.1	2.1
December	9.2	21.2	122	0.66	64	1.1	1.5
Hyderabad, Pakistan (Altitude 28 m)							
January	10.1	24.2	169	0.79	45	2.2	3.1
February	12.8	28.4	201	0.81	41	2.2	4.1
March	17.7	34.2	243	0.84	37	2.7	6.0
April	22.2	39.4	253	0.74	36	3.4	7.8
May	25.9	42.3	284	0.81	41	5.4	10.3
June	27.9	40.6	262	0.68	53	7.1	9.9
July	27.5	37.5	255	0.66	60	6.6	8.3
August	26.5	36.1	235	0.62	62	6.4	7.5
September	25.1	36.8	240	0.76	59	5.4	7.3
October	21.5	37.1	223	0.86	44	2.7	5.8
November	16.2	32.2	183	0.83	42	1.8	3.8
December	11.8	26.4	167	0.86	47	2.0	3.0

Missing Radiation Data

Many agrometerological stations that do not have a solarimeter to record the solar radiation do have a Campbell-Stokes sunshine recorder to record the duration of bright sunshine. In that case, R_s can be conveniently estimated from

$$R_s = \left(a + b\frac{n}{N}\right) R_A \tag{5.47}$$

where

R_s = solar radiation (W/m^2)
a = fraction of extraterrestrial radiation on overcast days (–)
a + b = fraction of extraterrestrial radiation on clear days (–)
R_A = extraterrestrial radiation, or Angot value (W/m^2)
n = duration of bright sunshine (h)
N = day length (h)

Although a distinction is sometimes made between (semi-)arid, humid tropical, and other climates, reasonable estimates of the Angstrom values, a and b, for average climatic conditions are a = 0.25 and b = 0.50. If locally established values are available, these should be used. The day length, N, and the extraterrestrial radiation, R_A, are astronomic values which can be approximated with the following equations. As extra input, they require the time of year and the station's latitude

$$R_A = 435\, d_r\, (\omega_s \sin\varphi \sin\delta + \cos\varphi \cos\delta \sin\omega_s) \qquad (5.48)$$

where

d_r = relative distance between the earth and the sun (-)
ω_s = sunset hour angle (rad)
δ = declination of the sun (rad)
φ = latitude (rad); northern latitude positive; southern negative

The relative distance, d_r, is found from

$$d_r = 1 + 0.033 \cos\frac{2\pi J}{365} \qquad (5.49)$$

where

J = Julian day, or day of the year (J = 1 for January 1); for monthly values, J can be found as the integer value of 30.42 × M – 15.23, where M is the number of the month (1-12)

The declination δ is calculated from

$$\delta = 0.4093 \sin\left(2\pi \frac{J + 284}{365}\right) \qquad (5.50)$$

The sunset-hour angle is found from

$$\omega_s = \arccos(-\tan\varphi \tan\delta) \qquad (5.51)$$

The maximum possible sunshine hours, or the day length, N, can be found from

$$N = \frac{24}{\pi}\omega_s \qquad (5.52)$$

For the Mansoura station (Table 5.2), which lies at 31.03° northern latitude, supposing that R_s is not available and that n = 7.1 hours, this amended procedure produces a January ET_h = 1.7 mm/d, not much different from the 1.5 mm/d mentioned in Table 5.2.

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