

the magnitudes of the fluxes involved, energy flux units of W m^{-2} can be interchanged with hydrologic units of millimeters per month (mm mo^{-1}) of liquid water evaporation.

When Q_n and either H or E can be determined independently, Equation (4.13) provides directly the remaining unknown flux. Usually, however, both H and E are unknown, and an indirect method must be used. From the methodological point of view, these indirect energy budget methods are analogous to the mean profile methods of Section 4.2.2. In both, essentially three equations are used which contain three unknowns E , u_* and H implicitly. In the profile methods these are the equations for \bar{q} , \bar{u} and $\bar{\theta}$. In the energy budget methods, (4.13) is used either with equations for \bar{q} and $\bar{\theta}$, or with equations for \bar{u} and $\bar{\theta}$ or \bar{q} , as will be shown next.

With Bowen ratio (EBBR)

When Q_n is known, the combination of the energy budget equation (4.13) with the Bowen ratio defined in Equation (4.9) produces

$$E = \frac{Q_{ne}}{1 + \text{Bo}} \quad (4.16)$$

Similarly, for the sensible heat flux one has

$$H_c = \frac{\text{Bo } Q_{ne}}{1 + \text{Bo}} \quad (4.17)$$

Bo can be determined as shown in (4.10), from profile data of temperature and specific humidity in the atmospheric surface layer. As discussed in Section 4.2.1, these data should be taken as averages over 15–30 min, approximately. Equation (4.16) shows that the energy budget with Bowen ratio (EBBR) method is most accurate when Bo is small. Both (4.16) and (4.17) produce a singularity when $\text{Bo} = -1$; but, as pointed out by Tanner (1960), over an active vegetation this is not a problem, as this situation usually occurs when H is low, around sunrise, sunset and occasionally at night. The situation does occur more often over cold water, and it may be necessary to use an alternative method when $-1 < \text{Bo} < -0.5$ to avoid the problem of a very small denominator in Equations (4.16) and (4.17). Tanner (1960) suggested the use of a bulk-transfer method for these special conditions. Another way consists of using mean values of Bo corrected by means of wind measurements, as outlined by Webb (1964); this method is especially useful when some terms in the available energy Q_n are only known for daily periods or longer.

The EBBR method has the advantage that no similarity functions for the atmospheric turbulence appear explicitly in the formulation. With Equation (4.10) no measurements of turbulence or of the mean wind speed are required, and the formulation, as written in (4.16) with (4.10), is independent of atmospheric stability. In addition, when Bo is small, the EBBR method may be less susceptible, albeit not immune, to imperfect fetch conditions, than mean profile methods, in which such effects are more directly apparent. The validity of the EBBR method depends critically on the similarity of the temperature and humidity profile; for the surface layer this requires the equality of the terms in the square brackets of Equations (2.51) and (2.52) (or (2.55) and (2.56)). The latest evidence

supports the view that $\Psi_v = \Psi_h$, even under stable conditions (see Dias and Brutsaert, 1996).

With profiles of wind and of a scalar (EBWSP)

If the profile data, either of the mean temperature or of the mean specific humidity, are lacking to apply the EBBR, the energy budget method can be applied instead with the mean wind speed profile. In fact, this procedure is potentially more powerful than the Bowen ratio method, because it yields not only E and H but also u_* .

As an illustration of this method, suppose that the specific humidity measurements are not available. It is then possible to use Equation (4.13) together with profile equations (2.50) and (2.51) (or (2.54) and (2.55)) in the surface layer, as a system of three equations with three implicit unknowns E , u_* and H . This system can be solved with measurements of Q_n , $\bar{\theta}_1 - \bar{\theta}_2$ (or $\theta_s - \bar{\theta}$) and $\bar{u}_2 - \bar{u}_1$ (or \bar{u} and z_0). The method can equally be applied with measurements higher up aloft in the mixed layer of the boundary layer. Thus in this case the system of three equations is (4.13) with (2.67) and (2.68). Similarly, if only humidity but no temperature measurements are available, to apply the method with surface layer data, the system of equations can consist of (4.13) with (2.50) and (2.52) (or (2.54) and (2.56)), or in the case of mixed layer data even with (2.67) and (2.71).

This EBWSP method and its simpler derivatives (see next section) are sometimes referred to as combination methods on the grounds, that both energy budget and hydrodynamic aspects of evaporation are considered. But this is misleading, since the Bowen ratio method is no less dependent on the validity of the hydrodynamics underlying (say) Equations (2.50)–(2.52) (or (2.54)–(2.56)), than the formulation of the mean wind speed profile.

4.3.2 Evaporation from wet surfaces: simplified expressions

The EBWSP method with measurements at one level

When the surface is wet, the surface specific humidity may be assumed to be the saturation value at the surface temperature, i.e. $q_s = q^*(T_s)$. This allows an approximation, first introduced by Penman (1948) and given in Equation (4.20) below; the main advantage of this approximation is that it eliminates the need for measurements of \bar{q} , \bar{u} and $\bar{\theta}$ at two levels, as in the profile methods (Section 4.2.2) and standard energy budget methods (Section 4.3.1) and that measurements at one level suffice.

The equation derived by Penman (1948) was intended for an open water surface. Here a somewhat more general derivation is presented, which is applicable to any wet surface, but which retains the essential features. By virtue of Equation (4.6), the Bowen ratio (4.10) can also be written in terms of the vapor pressure; with lower measurements at the surface, where $e_s = e^*(T_s)$, the Bowen ratio is

$$\text{Bo} = \gamma \frac{(\bar{T}_s - \bar{T}_a)}{(\bar{e}_s - \bar{e}_a)} \quad (4.18)$$

where e_a and T_a are the vapor pressure and temperature in the air, respectively, at some reference level, and where

$$\gamma = \frac{c_p p}{0.622 L_e} \quad (4.19)$$

Table 4.1 Values of (γ/Δ) at 1000 mb (γ is defined by Equation (4.19) and Δ can be obtained from Table 2.4)

Air temperature T_a (°C)	(γ/Δ)
-20	5.864
-10	2.829
0	1.456
5	1.067
10	0.7934
15	0.5967
20	0.4549
25	0.3505
30	0.2731
35	0.2149
40	0.1707

is commonly referred to as the psychrometric constant; at 20 °C and $p = 1013.25$ hPa it is $\gamma = 0.67$ hPa K⁻¹. Note that the θ difference is replaced by that of T , since they are often practically the same in the surface layer. The crucial step in Penman's analysis is the assumption

$$\frac{e_s^* - e_a^*}{T_s - T_a} = \Delta \quad \text{Typ: Diff. Kreis } \approx \text{Dervée} \quad (4.20)$$

where $\Delta = (de^*/dT)$ is the slope of the saturation water vapor pressure curve $e^* = e^*(T)$ at the air temperature T_a (see Figure 2.1); $e_a^* = e^*(T_a)$ is the corresponding saturation vapor pressure and $e_s^* = e^*(T_s)$ is the vapor pressure at the temperature of the surface, as indicated by the subscript. Since e_s for a wet surface is the value at saturation, the Bowen ratio (4.18) is thus, approximately

$$\text{Bo} = \frac{\gamma}{\Delta} \left[1 - \frac{(e_a^* - \bar{e}_a)}{(\bar{e}_s - \bar{e}_a)} \right] \quad (4.21)$$

In this expression Δ depends only on temperature and γ depends on both temperature and pressure. Values of (γ/Δ) for different temperatures at $p = 1000$ hPa are presented in Table 4.1 and Figure 4.2; they were obtained by means of (4.19) and values of Δ and L_e listed in Table 2.4. Substitution of (4.21) into (4.16) produces

$$Q_{nc} = \left(1 + \frac{\gamma}{\Delta} \right) E - \frac{\gamma}{\Delta} \left(\frac{e_a^* - \bar{e}_a}{\bar{e}_s - \bar{e}_a} \right) E \quad (4.22)$$

In the second term on the right of Equation (4.22), a bulk-transfer equation can be used, such as (4.7), to replace the unknown $E/(\bar{e}_s - \bar{e}_a)$ by a wind function $f_e(\bar{u}_r)$. Thus (4.22) yields the desired result, the Penman (1948) equation in its usual form

$$E = \frac{\Delta}{\Delta + \gamma} Q_{nc} + \frac{\gamma}{\Delta + \gamma} E_A \quad (4.23)$$

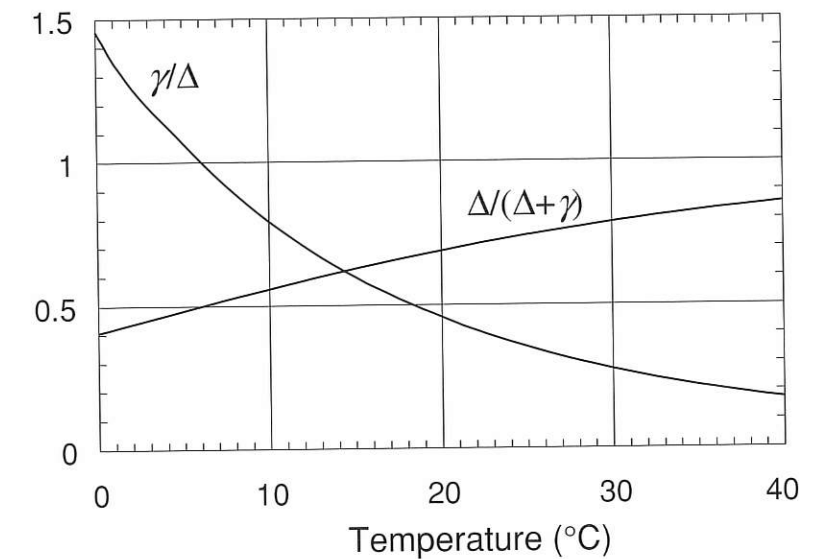


Fig. 4.2 Temperature dependence of (γ/Δ) and $\Delta/(\Delta + \gamma)$ at 1000 hPa; γ is defined by Equation (4.18) and $\Delta = de^*/dT$ is shown in Figure 2.1, and can be obtained from Equation (2.12) or from Table 2.4.

where E_A , a drying power of the air, is defined by

$$E_A = f_e(\bar{u}_r)(e_a^* - \bar{e}_a) \quad (4.24)$$

The ratio $\Delta/(\Delta + \gamma)$ is illustrated in Figure 4.2 for a pressure of 1000 hPa. Note that in Penman's (1948) original derivation it was assumed that $Q_{nc} = R_n/L_e$ and that all the other terms in Equation (4.14) are negligible. As mentioned, from the practical point of view, the main feature of this result is that it requires measurements of mean specific humidity, wind speed and temperature at one level only. This is a direct consequence of the approximation introduced in (4.20). For this reason, Penman's equation is useful when measurements at more than one level, needed for profile methods or standard energy budget methods, are unavailable or impractical.

Equation (4.23) has been widely used, but there is still no generally accepted way to formulate $f_e(\bar{u}_r)$, the wind function in E_A . Its definition in (4.24) suggests that any suitable mass transfer coefficient can be used for this purpose (see Section 4.2.2). Penman (1948) originally proposed an equation of the Stelling-type (4.5) as follows

$$f_e(\bar{u}_2) = 0.26 (1 + 0.54 \bar{u}_2) \quad (4.25)$$

where \bar{u}_2 is the mean wind speed at 2 m above the surface in m s⁻¹, and the constants require that E_A in Equation (4.24) is in mm d⁻¹ and the vapor pressure is in hPa. There are indications that Equation (4.25) yields reasonable results for natural terrain with small to moderate roughness (see Thom and Oliver, 1977); on the basis of experimental observations, it has also been suggested (Doorenbos and Pruitt, 1975) that for irrigated crops, the constant 0.54 should be replaced by 0.86. In calculations of long-term mean

values of E_A with equations like (4.25), to a first approximation the wind speed at 2 m can be estimated by assuming a power dependency on height, or

$$\bar{u}_2 = \bar{u}_r (2/z_r)^{1/7} \quad (4.26)$$

where z_r is the height (in m) at which the available wind data are measured.

A more fundamental approach to determine the wind function is based on turbulence similarity. Thus in terms of the bulk water vapor transfer coefficient as defined, for example, in (4.3), in which z_1 is the height of the measurement of \bar{u}_1 and z_2 that of \bar{e}_a , one obtains by virtue of (4.6), the wind function

$$f_e(\bar{u}_1) = 0.622 \rho p^{-1} C_e \bar{u}_1 \quad (4.27)$$

C_e can be determined by means of the similarity profile functions of Chapter 2. Under neutral conditions, on account of Equations (4.4), (4.6) and (4.7) this is (to a good approximation)

$$f_e(\bar{u}_1) = \frac{0.622 k^2 \bar{u}_1}{R_d T_a \ln[(z_2 - d_0)/z_{0v}] \ln[(z_1 - d_0)/z_0]} \quad (4.28)$$

where, again, z_1 is the level of the wind speed measurement and z_2 that of the water vapor pressure.

When Penman's equation is applied to calculate mean values of E over periods of a day or longer, the use of wind functions like (4.25), (4.27) or (4.28) may be adequate. However, when hourly values are required, the effect of atmospheric stability, which varies through the day, may be important. It is possible to include the effect of the atmospheric stability in the wind function, by writing the drying power of the air (4.24) in a form similar to (2.56) (see also Brutsaert, 1982) as follows

$$E_A = k u_* \rho (q_a^* - \bar{q}_a) \left[\ln \left(\frac{z_a - d_0}{z_{0v}} \right) - \Psi_v \left(\frac{z_a - d_0}{L} \right) + \Psi_v \left(\frac{z_{0v}}{L} \right) \right]^{-1} \quad (4.29)$$

where \bar{q}_a and q_a^* are the specific humidity of the air and the saturation specific humidity at air temperature, respectively. The problem can be solved by the following iteration procedure. An initial value of E is calculated in the usual way by means of Equation (4.23) using a neutral E_A , say (4.24) with (4.28); it is also possible to use (4.29) with $\Psi_v = 0$, and u_* is calculated by means of (2.54) with $\Psi_m = 0$. The initial value of E is used to obtain H by means of (4.13). These initial values of E , u_* and H provide a first estimate of the Obukhov length L by means of (2.46). This value of L allows now the calculation of a second estimate of u_* by means of (2.54) and a second estimate of E_A by means of (4.29), which produces a second estimate of E by means of (4.23), and so on. An example of the application of this method has been presented by Katul and Parlange (1992).

Evaporation from wet surfaces in the absence of advection

The two-term structure of Equation (4.23) suggests an interpretation which may serve as an aid in understanding the effect of regional or large-scale advection. When the air has been in contact with a wet surface over a very long fetch, it could be argued that it may tend to become vapor saturated, so that E_A , shown in (4.24), should tend to zero.

Accordingly, Slatyer and McIlroy (1961) reasoned that the first term on the right of Equation (4.23) may be considered a lower limit for evaporation from moist surfaces. Thus

$$E_e = \frac{\Delta}{\Delta + \gamma} Q_{ne} \quad (4.30)$$

was referred to as *equilibrium evaporation*, and the second term of (4.23) may be interpreted a departure from that equilibrium. In the absence of cloud condensation or radiative divergence, this departure would stem from large-scale or regional advection effects, involving horizontal variation of surface or atmospheric conditions.

Subsequent investigations have shown, however, that over wet surfaces, true equilibrium conditions are encountered only rarely, if ever. The main reason for this is that the atmospheric boundary layer is never a perfectly homogeneous boundary layer, as would be the case in channel flow; rather, it is continually responding to unsteady large-scale weather patterns, involving condensation aloft and dry air entrainment, which tend to maintain a humidity deficit even over the ocean. Nevertheless, the idea underlying Equation (4.30) has led Priestley and Taylor (1972) to use equilibrium evaporation as the basis for an empirical relationship to describe evaporation from a wet surface under conditions of minimal advection, E_{pe} . With data obtained over ocean and moist land surfaces they concluded that it is roughly proportional to E_e , that is

$$E_{pe} = \alpha_e \frac{\Delta}{\Delta + \gamma} Q_{ne} \quad (4.31)$$

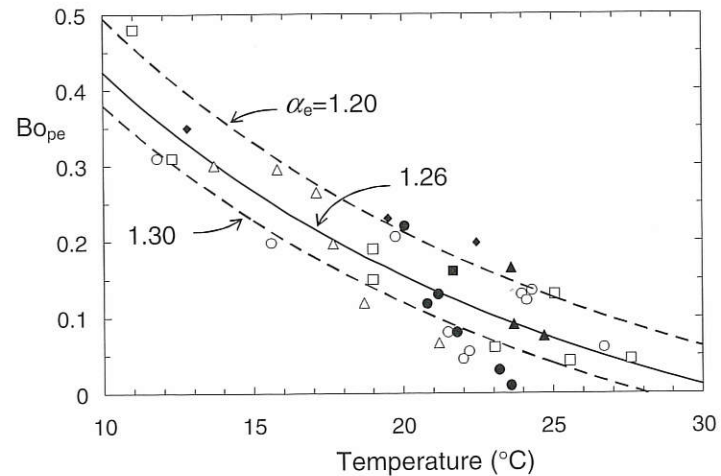
where α_e is a constant, which they found to be about 1.26. This value was later confirmed in many other studies (see Brutsaert, 1982) and α_e is now generally accepted to be of the order of 1.20–1.30, on average, for advection-free water surfaces and moist landsurfaces with short vegetation. Equation (4.31) is equivalent with a Bowen ratio

$$Bo_{pe} = \alpha_e^{-1} [(\gamma/\Delta) + 1] - 1 \quad (4.32)$$

which is illustrated in Figure 4.3 for different α_e values, together with some experimental data points.

These values of α_e indicate that over the ocean or other moist surfaces the second term of (4.23), that is the large-scale advection, accounts on average for about 20% to 23% of the evaporation rate. But this is only an average and large variations have been observed in different experimental settings. Still, it is remarkable that so many landsurfaces covered with fairly short vegetation, such as grass, which is not actually wet but with ample water available to the roots, yield about the same average values, ranging between 1.20 and 1.30, as open water surfaces. This may be the result of a fortuitous compensation of the specific humidity of non-wet leaf surfaces, which is lower than saturation, by a larger effective roughness, and thus transfer coefficient, of the vegetative surface. Still, in some studies drastically different values of α_e have been reported. This has been especially the case for very rough surfaces; for instance McNaughton and Black (1973) obtained $\alpha_e = 1.05$ for a young, 8 m high fir forest.

Fig. 4.3 Variation of Bowen ratio Bo_{pe} for moist surfaces as given by Equation (4.32); the solid curve represents an α_e value of 1.26 and the two dashed curves represent values of 1.20 and 1.30. The data points (for daily values) were collected by Davies and Allen (1973) from different sources.



Related empirical equations, but with R_s instead of Q_n as in (4.31), have been proposed by Makkink (1957), Jensen and Haise (1963), and Stephens and Stewart (1963). The short-wave radiation is often well correlated with the net radiation, which is the main component of Q_n over daily periods or longer. Such equations, which provide a good alternative to (4.31) when only short-wave radiation and temperature are available, have been used to determine irrigation requirements and as climatological indices of potential evaporation. However, the physical significance of such indices is not always clear, as is shown next.

✕ Potential evaporation

Because several of the simple energy budget-type methods for wet surfaces are often used as measures of potential evapotranspiration, a few comments are in order on this concept. The term potential evapotranspiration appears to have been introduced by Thornthwaite (1948) in the context of the classification of climate. It is now generally understood to refer to the maximal rate of evaporation from a large area covered completely and uniformly by an actively growing vegetation with adequate moisture at all times. The area is specified as large to avoid the possible effects of advection. Although the concept is widely used, it has also caused confusion, because it does not encompass all possible conditions and it involves several ambiguities. The concept requires closer specification if it is to serve as an unequivocal parameter.

Transpiration, even at the potential rate, involves such biological effects as stomatal impedance to the diffusion of water vapor, and the stage in the growth cycle of the vegetation. For this reason, the term potential evaporation is probably preferable. It can be defined to refer to the evaporation from any large uniform surface that is sufficiently moist or wet, so that the air in contact with it is fully saturated. Note that a wet or moist surface is not the same as one that has an adequate moisture supply for the roots of an actively growing vegetation; over short non-wet vegetation with adequate moisture the

evapotranspiration is often fairly similar to the evaporation from open water under the same conditions. As mentioned above, a possible explanation for this is that the stomatal impedance to water vapor diffusion may be compensated by the larger roughness values, resulting in larger transfer coefficients, of the vegetational surface.

Another point of ambiguity is that potential evaporation is often estimated by means of meteorological data observed under nonpotential conditions. Because the air interacts with the underlying surface, this is not the same rate as that which would be calculated or observed, if the surface had been moist or adequately supplied with water. Therefore, potential evaporation estimated on the basis of measurements carried out under nonpotential conditions should be called "apparent" to reflect this fact. Examples of apparent potential evaporation are the estimates made by means of an evaporation pan or by means of the Penman equation (4.23), on the basis of measurements in the actual, i.e. non-potential or arid, environment. Another example of apparent potential evaporation would be that obtained by means of Equation (4.3), in which q_s at the dry surface is assumed to be given by $q^*(T_s)$, i.e. the saturation specific humidity at the temperature of that surface. In what follows the "true" potential evaporation will be denoted by E_{po} , and the apparent potential evaporation by E_{pa} .

4.3.3 Operational methods for landsurfaces

Many operational procedures used in applied hydrology to predict evaporation involve some type of potential evaporation, used in conjunction with a procedure to derive the actual evaporation from it for the prevailing non-potential conditions.

Proportional fluxes with surface moisture "bucket"

Probably the oldest method, which follows work by Budyko (1955; 1974) and Thornthwaite and Mather (1955), is based on the following proportionality

$$E = \beta_e E_p \quad (4.33)$$

where E_p is a potential evaporation rate, and β_e a reduction factor reflecting the moisture availability. As mentioned above, potential evaporation is a somewhat ambiguous concept; not surprisingly, therefore, in practice Equation (4.33) has been applied with two different classes of E_p , such as E_{pa} , the apparent potential evaporation as defined in the previous section, and E_{pe} , the Priestley-Taylor equation given in (4.31).

The reduction factor β_e is often assumed to be a function of soil water content. In the application of (4.33) with such expressions for apparent potential evaporation E_{pa} as (4.3) and (4.23), a common assumption has been

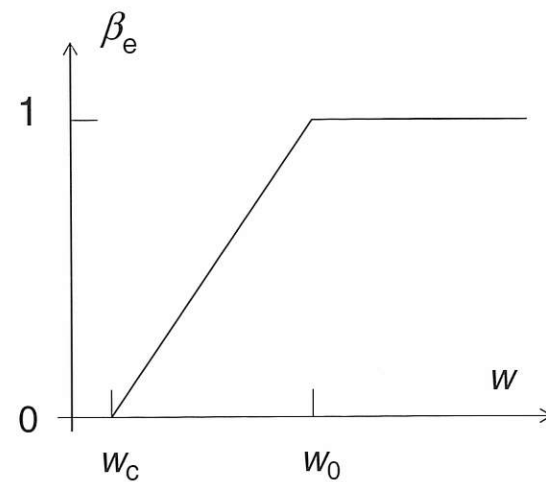
$$\beta_e = 1 \quad \text{for } w > w_0$$

$$\beta_e = (w - w_c)/(w_0 - w_c) \quad \text{for } w \leq w_0 \quad (4.34)$$

where w_0 is a critical soil water content above which E equals E_p , and w_c is a lower cut-off value below which E is zero. This is illustrated in Figure 4.4. The value of w can be determined on the basis of a soil water budget (see Thornthwaite and Mather,

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Fig. 4.4 Illustration of the commonly assumed relationship between the moisture availability factor β_e and the water content status w of the land surface complex. In practical applications this can be used with both apparent and actual potential evaporation.



1955; Budyko, 1974, p. 335; Manabe, 1969; Carson, 1982). The values of w_0 and w_c must be determined by calibration; for a surface soil layer with an assumed thickness of about 1 m, w_0 is generally taken to be of the order of 10–20 cm of water. The reduction factor β_e can also be related to some other surface moisture indices beside w , such as the accumulated actual evaporation minus precipitation (Priestley and Taylor, 1972), the local near-surface soil moisture content (see Davies and Allen, 1973; Crago and Brutsaert, 1992; Chen and Brutsaert, 1995), the soil moisture deficit (Grindley, 1970) and the antecedent precipitation index (Choudhury and Blanchard, 1983; Mawdsley and Ali, 1985; Owe *et al.*, 1989), again through calibration of the model with available data.

In some implementations of the same idea, the actual evaporation E is expressed in terms of the equilibrium evaporation E_e , by combining Equation (4.33) with (4.31), as follows

$$E = (\beta_e \alpha_e) E_e \quad (4.35)$$

in which E_e can be determined by means of (4.30). For instance, Figure 4.5 shows the results of Davies and Allen (1973) expressed as the product $(\beta_e \alpha_e)$ versus volumetric water content of the upper 5 cm of the soil. Although a nonlinear function is fitted to the data, it is similar to Equation (4.34). With data measured over prairie terrain it was found by Chen and Brutsaert (1995) that, with θ_{10} as the volumetric moisture content in the upper 10 cm of the soil profile, the relationship between E and E_e can be described by the following linear function

$$(\beta_e \alpha_e) = 1.26(\theta_{10} - 0.05)/0.22 \quad (4.36)$$

in the range $0.05 \leq \theta_{10} \leq 0.27$, and $(\beta_e \alpha_e) = 1.26$ for higher moisture contents; however, it was also observed that the relationship could be improved markedly by making $(\beta_e \alpha_e)$ not only dependent on soil moisture content but also on the density of the grassy vegetation cover, as expressed by the leaf area index La and the green vegetation fraction.

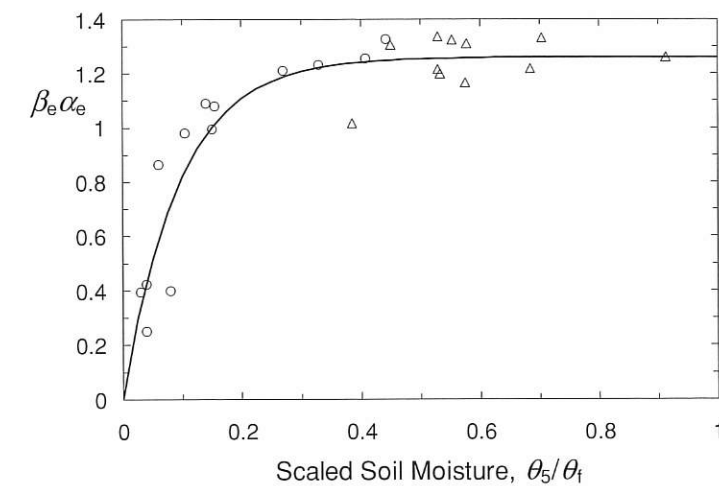


Fig. 4.5 Variation of $(\beta_e \alpha_e)$ with the water content in the upper 5 cm of the soil θ_5 , expressed relative to its value at field capacity θ_f for a sandy loam covered with perennial ryegrass in Ontario. The curve represents the function $\beta_e \alpha_e = 1.26[1 - \exp(-10.563\theta_5/\theta_f)]$. (After Davies and Allen, 1973.)

One difficulty in applying the formulation (4.33) with an apparent potential evaporation E_{pa} is that, as the surface dries out, the two quantities on the right-hand side of (4.33) move in opposite directions. Indeed, whenever β_e approaches zero, E_{pa} tends to become large; this may lead to an unstable product of a large with a small quantity, each with considerable noise. On the other hand, E_{pe} depends mainly on radiation and temperature, and not on the dryness of the air; hence application of (4.33) with E_{pe} is likely to be more robust and therefore preferable.

Surface resistance concept

A second procedure of reducing E_p to E is based on the realization that the release of water vapor from a vegetation is controlled by the stomata of the leaves. This is illustrated schematically in Figure 4.6. The underlying idea is that the air is assumed to be saturated with water vapor inside the stomatal cavities but not at the outside surface of the leaves, and the stomata provide an obstruction or a resistance to the diffusion of the water vapor from the inside to the outside of the leaves. This is often referred to a stomatal resistance r_{st} . Because evaporation also takes place from the soil surface, beside the leaves, the basic idea is usually extended to include this transport as well; thus the soil air at some depth below the surface can be assumed to be saturated and the soil pores can then be visualized as providing a resistance to the diffusion of water vapor to the soil surface. Hence more generally, the resistance approach is based on the concept of one or more resistance parameters in parallel and/or in series, which may account for the moisture stress in the vegetation and/or soil, and which relate saturation specific humidity q_s^* , at the temperature T_s of the evaporating surface, to the actual (non-saturated) specific humidity q_s at the evaporating surface.

Several such resistance parameters have been used for this purpose (see, for example, Monteith, 1973). The one given by Thom (1972) is instructive as an illustration and can be

Fig. 4.6 Sketch of a typical cross section through the underside of a leaf. Photosynthesis occurs in the mesophyll cells in the interior. The epidermis consists usually of a single layer of cells, covered by the protective cuticle; some of these cells are guard cells, which surround and control the size of the stomata by shrinking and swelling. The stomata are slit-like openings through which gas exchange and water vapor loss take place.

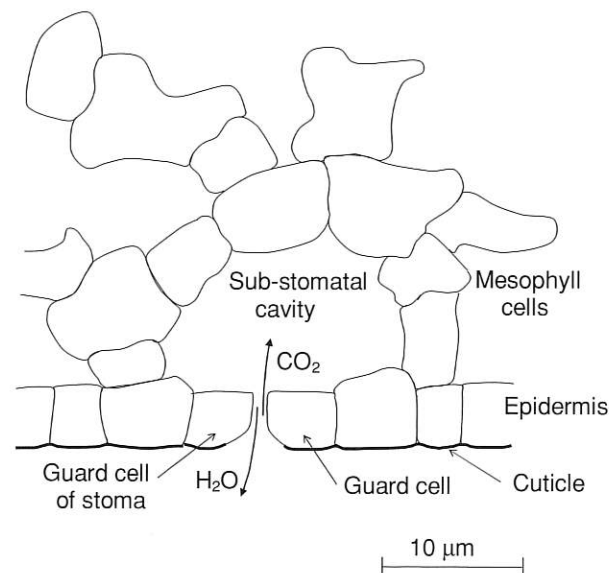
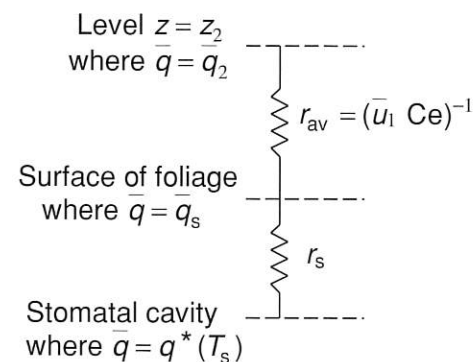


Fig. 4.7 Schematic diagram showing the resistance parameters, which can be used to describe transfer to and from a vegetational surface. The subscripts 1 and 2 refer to the levels of the wind speed and specific humidity measurements, respectively, in the atmospheric surface layer. In this context, Equation (4.3) is sometimes written in resistance notation as $E = \rho(\bar{q}_s - \bar{q}_2)/r_{av}$, which defines the aerodynamic resistance parameter for water vapor r_{av} in terms of the mass transfer coefficient C_e , as shown.



defined by

$$r_s = \rho (q_s^* - q_s) / E \quad (4.37)$$

in which r_s is the surface resistance, and q_s is the actual (not saturated) mean specific humidity at the evaporating surface; the basic concept is illustrated in Figure 4.7. For practical use there have generally been two types of evaporation equation based on the resistance concept. In the derivation of the first type, q_s , which is unknown, is eliminated between Equation (4.37) and the standard mass transfer equation (4.3) to yield the expression

$$E = \frac{C_e \bar{u}_1}{(1 + r_s C_e \bar{u}_1)} \rho (q_s^* - \bar{q}_2) \quad (4.38)$$

In the derivation of the second type, Equation (4.38) is used (instead of (4.3)) to obtain an expression in a way analogous to (4.23), namely

$$E = \frac{\Delta Q_{ne} + \gamma C_e \bar{u}_1 \rho (q_2^* - \bar{q}_2)}{[\Delta + \gamma(1 + r_s C_e \bar{u}_1)]} \quad (4.39)$$

Equation (4.39) is in the form of the Penman–Monteith equation (see, however, Monteith, 1973; 1981; Thom, 1975)

Numerous experiments have been conducted to determine resistance values for different types of vegetation. This has been mostly done in the context of expressions related to Equation (4.39). A few examples are beans (Black *et al.*, 1970), sugar beets (Brown and Rosenberg, 1977), tropical rainforest (Dolman *et al.*, 1991), eucalyptus forest (Dunin and Greenwood, 1986), pine forest (Gash and Stewart, 1975; Lindroth, 1985), maize (Mascart *et al.*, 1991), barley (Monteith *et al.*, 1965), sorghum (Szeicz *et al.*, 1973), and fir forest (Tan and Black, 1976). In addition, many attempts have been made to relate resistance parameters with such factors as Bowen ratio, soil moisture suction in the root zone, soil moisture deficit, humidity deficit in the air, solar radiation, temperature, leaf area index and others (see VanBavel, 1967; Szeicz and Long, 1969; Federer, 1977; Garratt, 1978; Lindroth, 1985; Stewart, 1988; Gash *et al.* 1989). The relationships developed so far are mainly statistical, so that they are vegetation and site dependent. Therefore, the resistance formulation is probably not yet sufficiently general to be practical for predictive purposes, but it has been useful as a diagnostic index in certain simulation studies (for example, to calculate missing data).

As a note of caution, in previous studies the resistance formulation has not always been used with consistent definitions for C_e (or r_{av}) and r_s (Thom, 1972; Brutsaert, 1982, p. 111). For instance, the drag coefficient C_d (or the related so-called aerodynamic conductance) is often used instead of C_e , as required in the rigorous derivation of Equation (4.23) with (4.24) and (4.27). This drag coefficient is defined in (2.37). Because it is not likely that above vegetation $z_0 = z_{0v}$, nor that $\Psi_m = \Psi_v$ (or Ψ_h), C_d is rarely equal to C_e . As a result of this inappropriate use of C_d (instead of C_e), it is not clear how the turbulence aspects of the transport, normally embodied in C_e , can be partitioned or separated from the strictly vegetational and/or soil moisture aspects of the transport supposedly embodied in r_s . This has undoubtedly contributed to the difficulty in deriving general relationships for both C_e and r_s on the basis of (4.39).

Although the resistance formulation with r_s may appear conceptually quite different from Equation (4.33) with the reduction factor β_c , both approaches are, in fact, practically the same. Indeed, (4.38) is equivalent with (4.33) (in which (4.3) is used to represent E_p for a wet surface) and a reduction factor

$$\beta_c = (1 + r_s C_e \bar{u}_1)^{-1} \quad (4.40)$$

Similarly, (4.39) is the same as (4.33) with (4.23) and a reduction factor

$$\beta_c = [1 + r_s C_e \bar{u}_1 \gamma / (\Delta + \gamma)]^{-1} \quad (4.41)$$

and as (4.33) with (4.31) and a reduction factor

$$\beta_c = \alpha_c^{-1} [1 + \gamma C_e \bar{u}_1 \rho (q_2^* - \bar{q}_2) / \Delta Q_{ne}] [1 + r_s C_e \bar{u}_1 \gamma / (\Delta + \gamma)]^{-1} \quad (4.42)$$

In practical applications of Equations (4.38) and (4.39) a knowledge of the parameters C_e and r_s is essential. The physical nature of C_e is well understood and based on sound turbulence theory. But the conceptual significance of the resistance concept remains problematic, in spite of the many studies devoted to it.