A THEORETICAL BASIS FOR THE PRIESTLEY-TAYLOR COEFFICIENT

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(Received in final form 24 July, 1996)

Abstract. The relationship between potential evaporation and areal evaporation is assessed using a closed-box model of the convective boundary layer (CBL). Potential evaporation is defined as the evaporation that would occur from a hypothetical saturated surface, with radiative properties similar to those of the whole area, and small enough that the excess moisture flux does not modify the characteristics of the CBL. It is shown that the equilibrium rate of potential evaporation is given by $E_{p0} = \alpha E_0$, where E_0 is the equilibrium evaporation (radiative term of the Penman formula), and α is a coefficient similar to the Priestley-Taylor coefficient. Its expression is $\alpha = 1+[1/(\epsilon+1)](\langle r_s \rangle / r_a)$, where $\langle r_s \rangle$ is the areal surface resistance, r_a is the local aerodynamic resistance, and ϵ is the dimensionless slope of the saturation specific humidity at the temperature of the air. Its calculated value is around 1 for any saturated surface surrounded by water, about 1.3 for saturated grass surrounded by well-watered grass and can be greater than 3 over saturated forest surrounded by forest. The formulation obtained provides a theoretical basis to the overall mean value of 1.26, empirically found by Priestley and Taylor for the coefficient α. Examining, at the light of this formulation, the complementary relationship between potential and actual evaporation (as proposed by Bouchet and Morton), it appears that the sum of these two magnitudes is not a constant at equilibrium, but depends on the value of the areal surface resistance.

Key words: evaporation, potential evaporation, Priestley–Taylor coefficient, boundary-layer model

1. Introduction

Slatyer and McIlroy (1961) were the first to introduce the concept of equilibrium evaporation (E_0), which they defined as the limit reached when unsaturated air is in contact with a wet surface over a long fetch. It is given by

$$\lambda E_0 = \frac{\epsilon}{\epsilon + 1}(R_n - G), \tag{1}$$

where $R_n - G$ is the available energy (with R_n the net radiation and G the soil heat flux), ϵ is the dimensionless slope of the saturation specific humidity at the temperature of the air, and λ is the latent heat of vaporization. McNaughton (1976) showed that air passing over a region with a constant surface resistance to evaporation achieved an equilibrium with the surface, where the gradient of saturation deficit vanishes, the Bowen ratio equals $1/\epsilon$, and the evaporation rate becomes E_0.

The definition of potential evaporation (E_p) has been the subject of some controversy (Granger, 1989). In this study, E_p corresponds to the evaporation rate of

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Boundary-Layer Meteorology **82:** 179–191, 1997.

saturated land sites or open water sites, when there is no significant surface or physiological control on the evaporation, i.e., when the surface or canopy resistance in the Penman-Monteith model (Equation (3)) is zero. The concept of potential evaporation differs from that of potential evapotranspiration, which refers to the maximum rate of evaporation from an area completely and uniformly covered by a vegetation with an adequate supply of water (Thornthwaite, 1948; Penman, 1956; Brutsaert, 1982). The precise definition of E_p adopted here is the one used by Morton (1969, 1983), i.e., the evaporation that would occur from a moist surface with an area small enough that the effects of the evaporation on the overpassing air would be negligible.

A coefficient α will be introduced, relating potential evaporation (as specified above) to equilibrium evaporation, and defined by

$$E_p = \alpha E_0. \tag{2}$$

The same coefficient has been proposed by Priestley and Taylor (1972), but these authors used a somewhat different definition of potential evaporation. For them, E_p represents the evaporation from a "horizontally uniform saturated surface (land and water)", sufficiently extended to obviate any significant advection of energy from outside, and they established experimentally that the best estimate of α was 1.26. Nevertheless, as a kind of generalization, the name of "Priestley-Taylor coefficient" will be conserved to the coefficient α specified by Equation (2), with E_p defined according to Morton. To support this generalization one can wonder wether, in trying to obtain the evaporation from a large saturated area, these two authors did not estimate a potential evaporation close to the one proposed by Morton, since (quotation) "... the only observations available as a basis [for the derivation of the value of α] are those from individual sites, subject in some cases to quite apparent small-scale non-uniformity and advection. ..." (Priestley and Taylor, 1972).

Numerous field data (e.g. Davies and Allen, 1973) have proven that evaporation from unstressed agricultural crops and pastures is often in agreement with this value of 1.26 for α. However, the additional energy, implied by the factor 1.26, has not been clearly explained yet. It has been ascribed to the entrainment of relatively warm, dry air downwards through the top of the convective boundary layer (CBL), defined as the turbulent surface layer which develops during daytime, due to the input of sensible heat from the ground. But it is still not clear why the entrained energy should be a conservative fraction of the available energy at the ground (Monteith, 1981, 1985).

The aim of this paper is to study the impact of regional evaporation on local potential evaporation, and to infer a theoretical explanation of why α should be greater than 1.0. In our approach, the CBL is represented by a well-mixed slab of air capped by the free atmosphere. These slab models have been used by McNaughton (1976), Perrier (1980), de Bruin (1983), McNaughton and Spriggs (1986), Culf (1994) to investigate the feedback loop between surface and air characteristics.

2. Theoretical Development

2.1. The Closed Box Model

Essentially, the convective boundary layer comprises a relatively thin surface layer, where the gradients of temperature and humidity may be significant, and a well-mixed layer with a potential temperature θ_m and a specific humidity q_m. The first boundary condition at the ground is given by the energy balance equation $R_n - G = H + \lambda E$, where H is the upward sensible heat flux and λE is the upward latent heat flux. The latent heat flux at the ground is governed by the Penman–Monteith combination equation

$$\lambda E = \frac{\epsilon(R_n - G) + \rho\lambda D_m/r_a}{\epsilon + 1 + r_s/r_a}, \tag{3}$$

where $D_m = q^*(\theta_m) - q_m$ is the potential saturation deficit of the mixed layer, $\epsilon = (\lambda/c_p)dq^*(\theta_m)/dT$ is the dimensionless slope of the saturation specific humidity, ρ is the air density, c_p is the specific heat of air at constant pressure, r_a is the bulk aerodynamic resistance to heat and water vapour transfer through the surface layer, and r_s is the bulk surface resistance to water vapour transfer. Above the capping inversion of the mixed layer is the undisturbed atmosphere, whose properties are determined by synoptic scale processes. The inversion cap, whose height h grows during the daytime, is not impermeable. The incorporation of air from above the capping inversion (with a saturation deficit D_+) into the mixed-layer tends to raise the saturation deficit D_m, and consequently the evapotranspiration rate, because generally $D_+ > D_m$.

In the absence of entrainment (input of dry air from above the capping inversion), the conservation equations for sensible heat and water vapour can be combined to the energy balance equation and Equation (3) to yield the following differential equation (McNaughton and Spriggs, 1986; Raupach, 1991)

$$\frac{dD_m}{dt} + \frac{D_m}{\tau} = \frac{D_0}{\tau}, \tag{4}$$

with

$$\tau = h\left(r_a + \frac{r_s}{\epsilon + 1}\right), \tag{5}$$

and

$$D_0 = \frac{\epsilon(R_n - G)}{\epsilon + 1} \cdot \frac{r_s}{\rho\lambda}. \tag{6}$$

This case is the well-known closed-box model, in which the CBL has an impermeable lid at a fixed height h. The solution for steady forcing ($R_n - G$, r_s and r_a

kept constant) was given by Perrier (1980, 1982) and by McNaughton and Jarvis (1983). It reads

$$D_m(t) = D_0 + (D_i - D_0)\exp(-t/\tau), \tag{7}$$

where D_i and D_0 are respectively the initial and final value of D_m. The solution shows that D_0 is approached exponentially with a time constant τ (63% of the equilibrium achieved). The corresponding rate of areal evaporation E^a follows the same exponential approach, and the final value is the equilibrium evaporation E_0 given by Equation (1).

Some authors have accounted for the entrainment effects. de Bruin (1983, 1989) made the downward fluxes of sensible heat and moisture proportional to the upward fluxes at the ground. McNaughton and Spriggs (1986) simulated the growth of the CBL by using an "encroachment model", based upon the assumption that the derivative of $h(t)$ with respect to time is proportional to H. Raupach (1991) made $h(t)$ proportional to the square root of time. The results obtained show that the predictions of surface fluxes are rather insensitive to the formulation used for entrainment (de Bruin, 1983; Raupach, 1991). McNaughton (1989) wrote: "It seems that useful models for regional evaporation can be developed without having to deal with difficult problems concerning the dynamics of entrainment". In the appendix a CBL model with entrainment is presented and used to predict the evaporation rates at equilibrium. It appears effectively that the impact of entrainment on evaporation is rather weak. Nevertheless, since entrainment always tends to raise the saturation deficit inside the CBL, the evaporation rates predicted by the closed-box model, and consequently the potential rate at equilibrium derived below and specified by Equation (10), are certainly underestimated with respect to the real world.

2.2. The Potential Evaporation at Equilibrium

Let us consider a small saturated surface, within the region, evaporating at the potential rate. Its radiative and aerodynamic characteristics are specified by primed variables. Those of the region surrounding this saturated surface are denoted by the areal averaging operator $\langle\ \rangle$. This case is a classic advection situation with a wet oasis surrounded by a drier region. The evaporation from the oasis can be obtained from Equation (3) by setting $r_s = 0$

$$\lambda E'_p = \frac{\epsilon(R_n - G)' + \rho\lambda D'_m/r'_a}{\epsilon + 1}, \tag{8}$$

r'_a is the aerodynamic resistance for heat and vapour of the growing internal boundary layer, and D'_m is the saturation deficit at the upper limit of this internal boundary layer. At a distance far enough from the leading edge, D'_m will not be very different from D_m (the saturation deficit within the well mixed layer, just above the surface layer of the whole area). And at equilibrium, when $D_m = D_0$, the rate of evaporation from this hypothetical small saturated surface can be approximated by Equation (8) in which D'_m is substituted by D_0, given by Equation (6). Then, the potential evaporation at equilibrium E'_{p0} reads

$$E'_{p0} = E_0 \left(\eta + \frac{1}{\epsilon + 1} \cdot \frac{\langle r_s \rangle}{r'_a} \right) \quad \text{with} \quad \eta = \frac{(R_n - G)'}{\langle R_n - G \rangle}, \tag{9}$$

where $\langle r_s \rangle$ is the areal surface resistance and $\langle R_n - G \rangle$ is the areal available energy (the small saturated surface excluded). The resistance r'_a has been examined with more details by Raupach (1991, p. 115), who defines it as the resistance to scalar transfer, from the saturated patch to the well-mixed layer, horizontally integrated over the whole patch.

For this relationship (9) to be valid, from an experimental viewpoint, the surface maintained at the potential rate must be small enough that the excess moisture flux and the different radiative and aerodynamic properties do not alter the characteristics of the CBL in equilibrium with the areal actual evaporation. But, at the same time, it must be large enough that the height of the internal boundary layer can reach the height of the areal surface layer (for D'_m to be equal to $D_m = D_0$). Such conditions can only be met for large areas. An order of magnitude can be obtained using the table given by Garratt (1994, p. 113), which shows the fetch x required to obtain an Internal Boundary Layer (IBL) of specified depth. The regional surface layer, which links the ground surface to the well-mixed layer, will be assumed to be about 50 m deep. According to this table, the fetch x required to reach this value of 50 m ranges from about 500 m to 1000 m depending on the downwind roughness length. Therefore, the minimum size of the surface maintained at the potential rate must lie between 0.5 and 1 km, and the aerodynamic resistance r'_a (like $(R_n - G)'$) have to be calculated at this distance of the leading edge. For this saturated surface not to alter the characteristics of the CBL, one may suppose that it must not represent more than 1% of the whole surface. This means that the minimum size of the region influencing the CBL ranges from 50 to 100 km.

The small saturated surface receives the same incoming shortwave and longwave radiations as the whole area. The main variables which can make the available energy of this saturated surface $(R_n - G)'$ different from the available energy of the whole area $\langle R_n - G \rangle$ are the albedo, the surface temperature and the soil heat flux G. Generally, their range of variation over different types of vegetation is relatively small, which means that η can never be very different from 1 (available energy is always higher over forests than over crops or grassland (Moore, 1976)). If we assume the available energy of the hypothetical saturated surface to be the same as that of the whole area (i.e. $\eta = 1$), Equation (9) simplifies into

$$E'_{p0} = \alpha E_0 \quad \text{with} \quad \alpha = 1 + \frac{1}{\epsilon + 1} \cdot \frac{\langle r_s \rangle}{r'_a}. \tag{10}$$

This equation gives, at equilibrium, a theoretical formulation to the Priestley-Taylor coefficient α, which appears to be proportional to the ratio $\langle r_s \rangle / r'_a$. In the appendix, using a CBL model developed by Raupach (1991), it is shown how entrainment (omitted in the closed-box approach) raises the value of α at equilibrium. For typical conditions of entrainment, the value of α appears to be increased by about 5%, which is fairly weak. Equation (10) describes also the feedback effect of areal actual evaporation on the value of local potential evaporation. It predicts that the drier a region ($\langle r_s \rangle$ high), the greater the local potential evaporation (E'_{p0}), which is conform to the common sense.

3. Discussion

3.1. Numerical Results

Values of the coefficient α and of the time constant τ of the closed-box model are shown in Table I for several combinations of surface types. Typical values of r_a and $\langle r_s \rangle$ have been chosen (the upperscript prime used for the characteristics of the small saturated surface will be omitted in the rest of the paper, as this will not lead to ambiguity). The height of the CBL has been set to 500 m and air temperature to 25 °C. When moist grass is surrounded by well-watered grass, the calculated value of α (1.26) is exactly equal to the overall mean value given by Priestley and Taylor (1972). This case corresponds approximately to the conditions envisaged by these authors. For saturated grass or forest surrounded by forest, the coefficient is much higher (2.04 and 3.62 respectively). These simulated results are roughly in agreement with the experimental data found in the literature. The rate of evaporation of intercepted water (i.e. the potential rate) is generally found higher on forest than on grassland. Stewart (1977) analysed measurements made by micrometeorological techniques of the rate of evaporation from the wet canopy of Thetford Forest in England (200 km^2 in size). He found that under identical radiation conditions, the rate of evaporation of intercepted precipitation was three times the transpiration rate that occurs when the canopy is dry. In contrast, earlier measurements for well-watered grass-covered surfaces (McMillan and Burgy, 1960; McIlroy and Angus, 1964) showed that there was no great difference between the transpiration rate and the rate of evaporation of intercepted water.

When a given saturated surface (water, grass or forest) is surrounded by open water, α is always equal to η, according to Equation (9), since $\langle r_s \rangle = 0$. In the case of very contrasting surfaces (such as grass-water or forest-water), η can be quite different from 1 because of different available energies over each surface. On the contrary, evaporation at the center of an extensive surface of water must be close to the equilibrium rate (E_0) since $\alpha = \eta = 1$. This theoretical result seems to contradict the experimental data obtained by de Bruin and Keijman (1979) over a large shallow lake in the Netherlands (about 460 km^2). During summer, they show that the parameter α exhibits a pronounced diurnal variation with a minimum value of about 1.15 early in the day and a maximum around 1.40 in the late afternoon.

Table I

Values at equilibrium of the coefficient α, defined by Equation (10), and of the time constant τ (63% of the equilibrium achieved), for a closed-box model of the CBL, with h = 500 m and θ = 25 °C (ϵ = 2.82). Two types of surface are considered: grass and forest. r_a is the aerodynamic resistance of the small saturated surface and $\langle r_s \rangle$ is the surface resistance of the surrounding environment, both expressed in s m^{-1}

Surface at potential rate	Surrounding environment	$\langle r_s \rangle$	α	τ (hr)
GRASS	Grass	50	1.26	8.8
(r_a = 50)	Forest	200	2.04	10.1
FOREST	Grass	50	1.65	8.8
(r_a = 20)	Forest	200	3.62	10.1

However, this relatively high value of α (with respect to the theoretical value of 1) can be ascribed to the entrainment effect (see appendix).

Figure 1 shows the evolution of the coefficient α as a function of the ratio $\langle r_s \rangle / r_a$, for different temperatures of the air. α increases linearly with $\langle r_s \rangle / r_a$ and the slope of the line is a decreasing function of air temperature (ϵ growing with temperature). Since the aerodynamic resistance r_a tends to decrease when wind velocity increases, α is an increasing function of wind speed, all other conditions being equal.

According to Equation (5), the time constant τ of the closed-box model is proportional to h, $\langle r_s \rangle$ and $\langle r_a \rangle$. The higher the CBL or the greater the areal resistances, the longer the time needed to achieve the equilibrium. The values of τ given in Table I are too great to obtain equilibration within a daily cycle. That means practically that the equilibrium potential evaporation E_{p0} (as E_0) can never be achieved over natural surfaces in the daytime span. They have to be considered as theoretical limits to which potential and actual evaporation tend.

3.2. The Complementary Relationship

Bouchet (1963) treated potential evaporation (E_p) and areal evaporation (E^a) as complementary quantities. He proposed an ingenious scheme, in which areal evaporation is considered to have a feedback effect on potential evaporation through the characteristics of the air. When an area dries out and areal evaporation decreases, the air becomes drier, which increases the potential rate of evaporation. The complementary relationship states that, when the external conditions do not change and in the absence of large-scale advection, the decrease in actual evaporation gen-

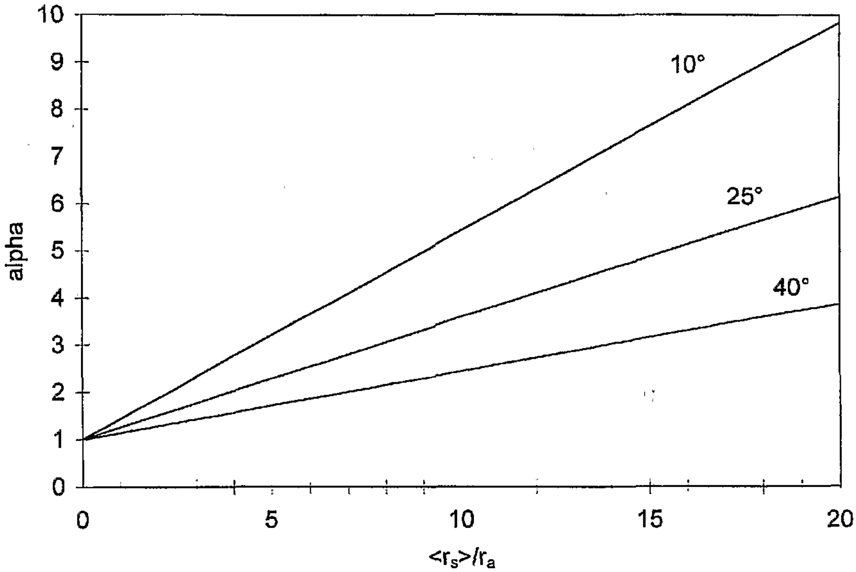

Figure 1. Coefficient α as a function of the ratio $\langle r_s \rangle / r_a$ (areal surface resistance over local aerodynamic resistance) for three different temperatures of the air.

erates an equal but opposite change in potential evaporation, implying a constant sum (Morton, 1969). This statement results in the following equation

$$E^a + E_p = 2E_p^a, \tag{11}$$

where E_p^a is the areal potential evaporation, obtained from a completely wet environment, when $E^a = E_p$. An important literature, reviewed in part by Morton (1983), has been produced on the legitimity of the derivation (Seguin, 1975; Fortin and Seguin, 1975; LeDrew, 1979; Perrier, 1982; Granger, 1989; Nash, 1989; McNaughton and Spriggs, 1989). The definition and value of E_p^a have also been widely discussed. On the basis of energy budget considerations, Bouchet had assumed that E_p^a was equal to half the absorbed solar radiation, cautioning that it would probably be less because of large scale advection. Later, Morton (1975), Brutsaert and Stricker (1979) suggested that E_p^a could be identified as the Priestley–Taylor equation ($E_p^a = \alpha E_0$, with $\alpha = 1.26$).

Applying the results obtained at equilibrium from the closed-box approach leads to $E^a = E_p = E_0$, when the environment is completely wet ($\langle r_s \rangle = 0$), and when the environment is dry ($\langle r_s \rangle \neq 0$),

$$E^a + E_p = (1+\alpha)E_0 = \left(2 + \frac{1}{1+\epsilon} \cdot \frac{\langle r_s \rangle}{r_a}\right) E_0, \tag{12}$$

where r_a is the aerodynamic resistance of the small surface maintained at the potential rate. This result (valid only at equilibrium) suggests that there exists no real complementarity between areal evaporation and local potential evaporation, since their sum depends upon the ratio between the areal surface resistance and the local aerodynamic resistance. Any change in the regional surface resistance, and consequently in the areal evaporation, modify the value of the alleged constant ($2E_p^a$). This conclusion is in agreement with that of McNaughton and Spriggs (1989), whose calculations give no apparent support to the complementary relationship. They showed that large-scale advection (interpreted as entrainment of air from a capping inversion into the CBL) is strongly controlled by the surface energy balance. The conclusion of LeDrew (1979) is similar. He wrote: "Specifying the sum of actual and potential evapotranspiration as a constant is physically unrealistic". As a matter of fact, the relationship, at equilibrium, between areal actual evaporation and local potential evaporation is more soundly and more staightforwardly expressed by Equation (10).

4. Conclusion

By coupling a closed-box model of the convective boundary layer with a Penman-Monteith model of surface evaporation, it has been possible to assess the relationship, at equilibrium, between local potential evaporation (E_{p0}) and areal evaporation (E_0). A general expression of the form $E_{p0} = \alpha E_0$ has been inferred, where α is a coefficient function of the ratio of areal surface resistance over local aerodynamic resistance ($\langle r_s \rangle / r_a$), similar to the Priestley-Taylor coefficient. The formulation obtained provides a theoretical basis to the empirical value of 1.26 found by these authors for "the evaporation from a horizontally uniform saturated surface". It predicts that the drier a region ($\langle r_s \rangle$ high), the higher the value of α. It is no longer necessary to ascribe the additional energy (implied by $\alpha > 1$) to the entrainment effects at the top of the CBL. On the other hand, this formulation gives no support to the complementary relationship, which appears to be physically unrealistic at equilibrium.

As a concluding remark, it seems worthwhile pointing out that the analysis presented in this paper represents only a theoretical framework far from being applicable to all meteorological conditions. It lies on a relatively crude description of the convective boundary layer (the closed-box model), where entrainment and the positive feedback mechanism between saturation deficit and stomatal closure (which tends to reduce the effects of CBL characteristics on evaporation (Jacobs and de Bruin, 1992)) are ignored. Moreover, the formulation proposed for the coefficient α is only valid at equilibrium, reached when available energy and surface resistance are maintained constant during large time. As the time needed to achieve this equilibrium is generally much longer than the daytime span, the formulation obtained is purely theoretical and can not represent the real world. For a future study on the same topic, it would be interesting to investigate the behaviour of the coefficient α on a daytime span, using a CBL which accounts both for entrainment and for the diurnal variation of available energy, in order to compare the mean diurnal values obtained with those derived from this theoretical approach.

It is also important to stress that our definition of potential evaporation (similar to the one proposed by Morton (1983)) is not recognized by everybody. Consequently, the conclusions of this study are valid only in the framework imposed by this definition.

Appendix: Equilibrium Potential Evaporation in a CBL with Entrainment

When entainment is taken into account, Equation (4) must be replaced by the folowing more general equation (McNaughton and Spriggs, 1986; Raupach, 1991)

$$\frac{dD_m}{dt} + \frac{D_m}{\tau} = \frac{D_0}{\tau} + \left(\frac{D_+ - D_m}{h}\right)\frac{dh}{dt}, \tag{A1}$$

with τ and D_0 specified respectively by Equations (5) and (6). The last term of (A1) describes the effect of the growth of the CBL height (h) and the resulting incorporation of warmer and drier air. Raupach (1991) inferred an analytical solution to Equation (A1) in the particular case where $h(t)$ grows as square root of time $h(t) = (Kt)^{1/2}$ (K is a growth-rate parameter taken equal to 46 m^2 s^{-1}). Assuming a simple linear profile of saturation deficit above the CBL, in the undisturbed atmosphere, $D_+(z) = \gamma_D z$ (where γ_D is a positive parameter, with the dimension of m^{-1}, whose value can be adjusted as a function of the dryness of the air above the capping inversion), he obtained a first-order differential equation in D_m, with non- constant coefficients, the solution of which is relatively simple (with the initial condition $D_m(0) = 0$). At large time, the saturation deficit approaches the steady limiting value

$$D_{\text{eq}} = D_0 + \frac{K\gamma_D}{2}\left(\langle r_a\rangle + \frac{\langle r_s\rangle}{\epsilon + 1}\right), \tag{A2}$$

where $\langle r_a\rangle$ and $\langle r_s\rangle$ are the areal aerodynamic and surface resistance respectively. The corresponding steady evaporation is written as

$$E_{\text{eq}} = E_0 + \frac{K\rho\gamma_D}{2(\epsilon + 1)} = E_0(1 + \omega) \quad \text{with} \quad \omega = \frac{\rho\lambda K\gamma_D}{2\epsilon\langle R_n - G\rangle}, \tag{A3}$$

where E_0 is given by Equation (1) and ω represents the entrainment effect. When there is no entrainment ($K = 0$), $\omega = 0$, $E_{\text{eq}} = E_0$ and $D_{\text{eq}} = D_0$ (we retrieve the results obtained with the closed-box model). Raupach (1991) showed also that the time needed to obtain equilibrium with entrainment is substantially larger than in the case of a closed-box model.

Under these conditions, the potential evaporation of a small saturated surface, at equilibrium, is obtained from Equation (8), in which D'_m is substituted by D_{eq}

$$E'_{p,\text{eq}} = E_0 \left[\left(\eta + \frac{1}{\epsilon + 1} \frac{\langle r_s \rangle}{r'_a} \right) + \omega \left(\nu + \frac{1}{\epsilon + 1} \frac{\langle r_s \rangle}{r'_a} \right) \right], \tag{A4}$$

with $\nu = \langle r_a \rangle / r'_a$. Assuming $\eta = \nu = 1$ (i.e. $(R_n - G)' = \langle R_n - G \rangle$ and $r'_a = \langle r_a \rangle$), Equation (A4) simplifies into

$$E'_{p,\text{eq}} = \left(1 + \frac{1}{\epsilon + 1} \frac{\langle r_s \rangle}{r'_a} \right) E_{\text{eq}} = (1 + \omega) \left(1 + \frac{1}{\epsilon + 1} \frac{\langle r_s \rangle}{r'_a} \right) E_0 \tag{A5}$$

Therefore, in a CBL with entrainment (with the characteristics specified above), the Priestley-Taylor coefficient α_e can be approximated by $\alpha_e = (1 + \omega)\alpha$, α representing the coefficient obtained with the closed-box model. Considering a constant available energy $\langle R_n - G \rangle = 500$ W m^{-2}, and taking $\lambda = 2.4\ 10^6$ J kg^{-1}, $\rho = 1.20$ kg m^{-3} and $\epsilon = 2.2$, the corrective term ω can be written as $\omega = 4.9\ 10^4\ \gamma_D$. That means that the value of the Priestley-Taylor coefficient at equilibrium is directly proportional to the inversion strength (γ_D) in the undisturbed atmosphere just above the CBL. For a typical value of γ_D of 10^{-6}, the impact of the entrainment effect on the value of the Priestley-Taylor coefficient appears to be fairly weak, since it is around 5%.

Acknowledgements

We are pleased to acknowledge H.A.R de Bruin for his useful comments on the first version of this paper, and J.R. Garratt for having provided us with his internal and unpublished document on Potential Evaporation.

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