The Priestley–Taylor parameter and the decoupling factor for estimating reference evapotranspiration

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Abstract

Using two extensive data sets of daily lysimetric measurements of reference evapotranspiration (E_0, mm day^{-1}) under climatic conditions ranging from humid tropical (Piracicaba, SP, Brazil) to semi-arid (Davis, CA, USA) it is here presented an evidence that the Priestley–Taylor α parameter can be set equal to the inverse of the McNaughton–Jarvis decoupling factor Ω, that is, $\alpha = \Omega^{-1}$. This approach does not require local calibrations and it allows for a variable α. Several statistical measures of the goodness of fit indicate that this simplification represents a good alternative to the theoretically more robust Penman–Monteith with the parameterization scheme of the FAO-56. Selected days with almost perfect match between FAO-56 estimates and observations, with E_0 varying from 0.60 to 7.0 mm day^{-1}, showed Ω values from 0.69 to 0.93. Ω decreased linearly as a function of the wind run.

The Perrier–Katerji–Goose–Itier model for α did not work well for the present data sets. The main problem here is the determination of a critical resistance with daily averages of vapor pressure deficit.
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1. Introduction

Evapotranspiration (ET) has been the subject of many reports for almost a century. At first the estimates were based on empirical relationships but Penman (1948) presented an equation incorporating all the weather variables with direct effect upon the evapotranspiration process. The Penman equation separates the contribution of the weather elements into two terms. One term isolates the effect of the available radiation, while the other term groups the elements that represents the surface–atmosphere interactions. The radiation term expresses the ability of

the surface in capturing the incoming radiation, and it sets the lower limit of ET if soil water supply is not limiting and also if it is not influenced by upwind or overhead conditions (McNaughton, 1976a,b). Under such conditions ET is defined as the equilibrium evapotranspiration (E_{eq}) by Slatyer and McIlroy (1961).

The surface–atmosphere interactions term depends on the aerodynamic characteristics of the surface and it expresses the conversion of sensible heat of the surrounding air into latent heat. As a rough surface interacts more strongly with the atmosphere it is able to extract more sensible heat from the passing air than a fairly smooth surface. The degree of atmosphere–vegetation interaction can be estimated by a decoupling coefficient (Ω), defined by McNaughton and Jarvis (1983), based on the Penman–Monteith

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equation. The decoupling coefficient sets the relative importance of the equilibrium term to the overall ET and it varies from 0 (a perfect coupling condition with the atmosphere providing all the needed energy for the ET) to 1 (a complete isolation being the radiation the only contributor to the ET process).

Likely values of Ω are inversely related to the leaf width and plant height (Jarvis, 1985; Jones, 1994; Jones and Tardieu, 1998). However, experimental results from agricultural crops (Baldocchi, 1994; Steduto and Hsiao, 1998), forests (Wullschleger et al., 2000; Martin et al., 2001), and windbreak trees (Smith and Jarvis, 1998) show large day-to-day and hour-to-hour variation in Ω. A 3D canopy simulation model indicated the existence of a spatial variation of Ω within the crown of an isolated walnut tree, with the sunlit leaves less coupled than the shaded ones (Daudet et al., 1999). Therefore, Ω can be dealt with in any time or space scale.

Based also on the Penman–Monteith equation, Allen et al. (1998) defined reference evapotranspiration (E_0) as the amount of water used by a hypothetical extensive reference surface of green, well-watered grass of uniform height of 0.12 m, actively growing and completely shading the ground, with a fixed surface resistance of 70 s m^{-1}, and an albedo of 0.23. Under such conditions, with advection having the least effect upon the evapotranspiration, E_{eq} represents about 80% of E_0, on average ($\Omega = 0.8$, McNaughton and Jarvis, 1983). Priestley and Taylor (1972) proposed to neglect the aerodynamic term and correct E_{eq} by a dimensionless coefficient α (the Priestley–Taylor parameter) as $E_0 = \alpha E_{\mathrm{eq}}$. Experimental results from several sites around the world, including vegetated surfaces and large water bodies (lake and oceans), gave α values in the range between 1.08 ± 0.01 and 1.34 ± 0.05, with average of 1.26 (Priestley and Taylor, 1972).

Empirically, α has been found to be related to: (i) the atmospheric vapor pressure deficit (Jury and Tanner, 1975); (ii) the soil moisture content (Davies and Allen, 1973; Williams et al., 1978; Barton, 1979); (iii) the soil moisture and the wind run (Mukammal and Neumann, 1977); (iv) the soil moisture and the incoming solar radiation (Flint and Childs, 1991); (v) the atmospheric stability condition (Viswanadham et al., 1991); and (vi) the turbulent sensible heat flux (Pereira and Villa Nova, 1992).

Theoretical atmospheric boundary layer models (McNaughton, 1976a,b; De Bruin, 1983; Monteith, 1995; Raupach, 2000) have been used to study the variation of α; in general, α decreases as the surface resistance to the evapotranspiration increases. Such models contribute towards the development of a theory of the evaporation process of large areas (McNaughton, 1976b), but an operational model for the everyday management activities is still necessary.

Monteith (1965) and Perrier (1975) had shown that Ω can be interpreted as the ratio between the evapotranspiration rate of a dry surface (E), given by the Penman–Monteith equation, and the rate of a wet surface (E_0), given by the original Penman equation, both in the same weather, or $\Omega = E/E_0$. With this definition of Ω it will be tested here the goodness of the working hypothesis that the reference evapotranspiration can be computed as $E_0 = \Omega^{-1} E_{\mathrm{eq}}$, or $\alpha = \Omega^{-1}$. It can be a mere coincidence but for a grass field $\alpha = 1.26 \approx 0.8^{-1}$ as indicated by McNaughton and Jarvis (1983). Experimental results will be presented indicating that $\alpha = \Omega^{-1}$ is a good approximation for computing E_0, on a daily time scale, under climatic conditions ranging from humid tropical to semi-arid. This approach avoids the use of a fixed α value, and it eliminates the need for local calibrations, and for measurements of atmospheric humidity.

2. Theory

Rearranging the Penman–Monteith equation (Monteith, 1965, 1981) to study the relative contribution of the radiative (E_{rad}) and aerodynamic (E_{aero}) terms to the overall evapotranspiration (E, mm day^{-1}), McNaughton and Jarvis (1983) presented it in the following form:

$$E = E_{\mathrm{rad}} + E_{\mathrm{aero}} = \Omega E_{\mathrm{eq}} + (1 - \Omega)E_{\mathrm{im}} \qquad (1)$$

defining Ω as the dimensionless decoupling factor, computed by

$$\Omega = \left[1 + \frac{\gamma}{s + \gamma}\frac{r_{\mathrm{c}}}{r_{\mathrm{a}}}\right]^{-1} \qquad (2)$$

being γ (kPa K^{-1}) the local psychrometric coefficient; s (kPa K^{-1}) the slope of the temperature–saturation vapor pressure relation at the mean temperature point;

r_c (s m^{-1}) the average canopy (surface) resistance; and r_a (s m^{-1}) the average aerodynamic resistance between the effective source of water vapor and a reference height above the surface (taken as 2 m for E_0). As noted by Jones (1994), Ω depends on the r_c/r_a ratio rather than on their absolute values.

The equilibrium evapotranspiration (E_{eq}, mm day^{-1}) is given by

$$E_{eq} = \frac{s}{s+\gamma}\frac{R_n - G}{\lambda} \tag{3}$$

where R_n (MJ m^{-2} day^{-1}) is the net radiation above the surface; and G (MJ m^{-2} day^{-1}) is the soil heat flux, presumed to be negligible on a daily time scale for reference evapotranspiration ($G = 0$, Allen et al., 1998); and λ is latent heat of vaporization here taken as 2.45 MJ kg^{-1}.

E_{im} (mm day^{-1}) is the part of the evapotranspiration *imposed* by the surrounding air (Jarvis and McNaughton, 1986). E_{im} is determined by the daily average vapor pressure deficit of the air (D_a, kPa), and by the canopy resistance, or

$$E_{im} = \frac{\rho c_P D_a}{\lambda \gamma r_c} \tag{4}$$

where ρc_P (kJ m^{-3} K^{-1}) is the volumetric specific heat of the air at constant pressure.

Priestley and Taylor (1972) suggested that E can be computed by

$$E = \alpha E_{eq} \tag{5}$$

with α compensating for the discard of E_{im}.

Expressing the Penman–Monteith equation as a dimensionless fraction of the available energy ($R_n - G$), or

$$\frac{E}{R_n - G} = \frac{(s/\gamma) + (\rho c_P D_a/\gamma(R_n - G)r_a)}{(s/\gamma) + ((r_c + r_a)/r_a)} \tag{6}$$

Monteith (1965) defined an "isothermal resistance" (r_i) as

$$r_i = \frac{\rho c_P D_a}{\gamma(R_n - G)} \tag{7}$$

indicating that r_i is a critical value of $r_c + r_a$ at which $E = R_n - G$.

Imposing the condition $\partial E/\partial r_a = 0$ in Eq. (1), the isothermal resistance takes the critical value

(Monteith, 1965; Jones, 1994; Rana et al., 1997, 1998):

$$r^* = \frac{(s+\gamma)\rho c_P D_a}{s\gamma(R_n - G)} \tag{8}$$

defining the condition when E is independent of r_a (i.e., independent of the wind speed) and $E = E_{eq}$.

Combining Eqs. (1)–(4) and (8), Perrier et al. (1980) showed that

$$E = CE_{eq} \tag{9}$$

with

$$C = \frac{1 + (\gamma/(s+\gamma))(r^*/r_a)}{1 + (\gamma/(s+\gamma))(r_c/r_a)} \tag{10}$$

Katerji and Perrier (1983) defined C as a crop coefficient, and it has been found to vary with the growth stage for wheat (Perrier et al., 1980), for rice and groundnut (Peterschmitt and Perrier, 1991). From Eqs. (2), (5) and (10) it can be seen that $\alpha = C = \Omega/\Omega^*$.

3. Material and methods

Following Priestley and Taylor (1972), daily equilibrium evapotranspiration (E_{eq}) was computed using Eq. (3) with the 24 h total of net radiation, neglecting the heat flux into the ground, and with the average air temperature. Daily reference evapotranspiration rates were computed as a function of E_{eq} using two schemes. The first approach was that used by Perrier et al. (1980) expressed by Eqs. (8)–(10) and hereafter referred to as the Perrier–Katerji–Goose–Itier (PKGI) method. In Eq. (8) it was used the daytime mean net radiation (R_{n+}, W m^{-2}), neglecting the soil heat flux, with the average vapor pressure deficit given by $D_a = e_s - e_a$, where $e_s = 0.5[e_s\{T_{max}\} + e_s\{T_{min}\}]$ and $e_a = [e_s\{T_{max}\}RH_{min} + e_s\{T_{min}\}RH_{max}]/200$, as described in Allen et al. (1998).

The second approach is an alternative hypothesis based on Monteith (1965) and Perrier (1975) arguments and it takes $\alpha = \Omega^{-1}$ (Alpha-Omega method), as defined in Eq. (2). For both schemes the parameterization for a reference grass surface given by Allen et al. (1989, 1998) was adopted, i.e., $r_c/r_a = 0.34 U_2$, and $r_a = 208/U_2$ with the daily average wind at the 2 m (U_2, m s^{-1}) above the ground.

Two data sets were used for testing the goodness of such schemes. One set was obtained from 1960 to 1963

at UC Davis, CA, USA (38°32′N; 121°46′W; 19 m a.m.s.l.), where a 6.1 m diameter weighing lysimeter located near the center of a 146 m × 355 m measured ET (E_{Lys}) from a field of perennial ryegrass mowed to keep an average plant height of 0.1 m, and irrigated weekly (Pruitt, 1964). Near the lysimeter, net radiation and wind run were measured at 2 m above the ground. Maximum and minimum air temperature and relative humidity were measured at 1 m above the surface. Days with strong advection as well as those with restricted E_{Lys} were discarded based on the criteria $0.5 < E_{Lys}/R_{n+} < 0.9$, remaining 422 days for analysis.

The other data set used came from measurements at Piracicaba, SP, Brazil (22°42′S; 47°30′W; 546 m a.m.s.l.) during 1996. ET was obtained also from a weighing lysimeter but with a much smaller surface area (≈0.92 m²), located about 80 m from the leading edge of the most prevailing wind direction, in a 35 m × 90 m field of *Paspalum notatum* L. grass. Soil moisture was monitored by tensiometers and kept near field capacity by underground irrigations. The grass was clipped to a height close to the 0.12 m of a reference surface defined by Allen et al. (1989, 1998). Net radiation was measured at 1 m above the ground. Maximum and minimum air temperature and relative humidity, and wind speed were obtained at 2 m height. The same E_{Lys}/R_{n+} ratio criteria used for the Davis data set was also applied to eliminate days with unreliable data, resulting in 127 days for analysis.

The measures of goodness of fit of the several approaches were based on linear regression between estimated (E) and observed (O) E_0. As the Y-intercept did not differ statistically from zero the regression was forced through the origin (i.e., $Y = bX$). Besides the traditional coefficient of determination (r^2), root mean square error (RMSE), maximum absolute error (MAE), mean bias error (MBE), the following statistics were also used: index of agreement (d; Willmott, 1981), model efficiency (EF; Zacharias et al., 1996), and the t-statistics for the mean with the level of significance chosen to be 0.5% (Stone, 1993; Jacovides and Kontonyiannis, 1995). Such indices are defined as

$$\text{MAE} = \max(|O_i - E_i|)_{i=1}^{N} \tag{11}$$

$$\text{MBE} = \frac{\sum_{i=1}^{N}(O_i - E_i)}{N} \tag{12}$$

$$d = 1 - \frac{\sum_{i=1}^{N}(O_i - E_i)^2}{\sum_{i=1}^{N}(|E_i - \bar{O}| + |O_i - \bar{O}|)^2} \tag{13}$$

$$\text{EF} = 1 - \frac{\sum_{i=1}^{N}(O_i - E_i)^2}{\sum_{i=1}^{N}(O_i - \bar{O})^2} \tag{14}$$

$$t = \frac{|\bar{O} - \bar{E}|}{[(s_O^2 + s_E^2)/N]^{0.5}} \tag{15}$$

where $\bar{O}$ and $\bar{E}$ are the means of the observed and estimated values; s_O^2 and s_E^2 are the respective variances. The perfect model will have $b = r^2 = d = \text{EF} = 1.0$ and $a = \text{MAE} = \text{MBE} = \text{RMSE} = t = 0$. The best model should tend to the above limits. EF is defined as an appropriate r^2 for the no-intercept fit and it can take negative values as well as values greater than 1 (Ksalveth, 1985; Montgomery and Peck, 1992, p. 49). A value of t below a critical threshold indicates a statistically reliable model.

4. Results and discussion

Table 1 summarizes the statistical measures for all approaches. Comparing equilibrium evapotranspiration (E_{eq}) with lysimetric measurements (E_{Lys}) it is possible to determine an overall average Priestley–Taylor parameter α (or C for Perrier et al., 1980). For the tropical climate of Piracicaba, SP, Brazil, the comparisons resulted in $\alpha = 1.20$ (±0.01), $r^2 = 0.8218$ (Fig. 1A); for the semi-arid climate of Davis, CA, USA, it resulted in $\alpha = 1.27$ (±0.01), $r^2 = 0.9112$ (Fig. 1B). The original Priestley–Taylor proposal of a fixed average $\alpha = 1.26$ worked well for Davis in the absence of heat advection ($E_{Lys} = 1.0E_{PT1.26}$, $r^2 = 0.9088$). For Piracicaba there was a slight overprediction of 5%, on average, as $E_{Lys} = 0.952E_{PT1.26}$, $r^2 = 0.8218$.

The α value here reported for Davis is slightly smaller than the 1.33–1.44 values found by Pruitt and Doorenbos (1977) using monthly averages and different screening criteria. In the present Davis data set it was found that, in general, there was a tendency for warmer days to have a higher E_{Lys}/R_{n+} ratio than cooler days as indicated by relationship $E_{Lys}/R_{n+} = 0.458 + 0.0108T_{max}$ ($r^2 = 0.4689$), with $3.9\,°C < T_{max} < 37.2\,°C$. For Piracicaba such

Table 1

Summary of the statistical indices used for analysis of the performance of the E_0 models: slope of the regression line (b), coefficient of determination (r^2), maximum absolute error, mean bias error, root mean square error, index of agreement (d), model efficiency (EF), standard deviation (s), t-statistic (t), mean of the observed values ($\bar{O}$), mean of the estimates ($\bar{E}$), Penman–Monteith FAO-56 (PM56), Priestley–Taylor parameter (α), decoupling factor (Ω)

Method	b	r^2	MAE ($\mathrm{mm\,day^{-1}}$)	MBE ($\mathrm{mm\,day^{-1}}$)	RMSE ($\mathrm{mm\,day^{-1}}$)	d	EF	$\bar{E}$ ($\mathrm{mm\,day^{-1}}$)	s_E ($\mathrm{mm\,day^{-1}}$)	t
Piracicaba, SP, BR[a]										
$\alpha = 1.26$	0.952	0.8218	1.91	0.16	0.50	0.95	0.78	4.60	1.34	1.01
$\alpha = C$	1.345	0.7883	2.24	−1.18	0.54	0.73	−0.21	3.26	0.94	8.86
$\alpha = \Omega^{-1}$	1.02	0.8628	1.24	−0.12	0.43	0.97	0.86	4.32	1.22	0.80
PM56	0.93	0.8634	1.42	0.33	0.46	0.94	0.78	4.77	1.19	2.22
Davis, CA, USA[b]										
$\alpha = 1.26$	1.00	0.9088	1.49	0.02	0.51	0.98	0.91	3.85	1.64	0.17
$\alpha = C$	0.949	0.8245	2.59	0.25	0.72	0.94	0.81	4.12	1.55	2.21
$\alpha = \Omega^{-1}$	1.00	0.8723	2.62	−0.04	0.62	0.97	0.87	3.83	1.66	0.34
PM56	0.93	0.8952	2.53	0.28	0.60	0.97	0.86	4.15	1.76	2.33

[a] $\bar{O} = 4.44\,\mathrm{mm\,day^{-1}}$, $s_0 = 1.17\,\mathrm{mm\,day^{-1}}$, $n = 127$, $t_c = 2.62$.
[b] $\bar{O} = 3.87\,\mathrm{mm\,day^{-1}}$, $s_0 = 1.72\,\mathrm{mm\,day^{-1}}$, $n = 422$, $t_c = 2.60$.

tendency was not detected perhaps due to a narrow maximum temperature range ($21.5\,°\mathrm{C} < T_{max} < 35.3\,°\mathrm{C}$).

In regard to the Perrier–Katerji–Goose–Itier method (Fig. 2), for Piracicaba it underestimated the measurements by 35%, on average, as $E_{Lys} = 1.346E_{PKGI}$, $r^2 = 0.7883$. The t-statistic above the critical value rejects the hypothesis that the PKGI gives reliable estimates for Piracicaba. For Davis, there was a slight overprediction of 5%, on average, or $E_{Lys} = 0.949E_{PKGI}$, $r^2 = 0.8245$. Days with large overes-

timations in Davis had minimum temperature below $5\,°\mathrm{C}$ and were all in the period October–May. The T_{min} threshold was selected based on visual inspection of the most discrepant points in Fig. 2B. Analysis of the 123 days with $T_{min} < 5\,°\mathrm{C}$ indicated gross overpredictions, or $E_{Lys} = 0.763E_{PKGI}$, $r^2 = 0.8243$. During such days the average E_{Lys} was equal to $2.31\,\mathrm{mm\,day^{-1}}$. Results for the remaining 298 days with $T_{min} > 5\,°\mathrm{C}$ displayed a much better fit with $E_{Lys} = 0.99E_{PKGI}$, $r^2 = 0.7963$, and the average E_{Lys} equal to $4.51\,\mathrm{mm\,day^{-1}}$.
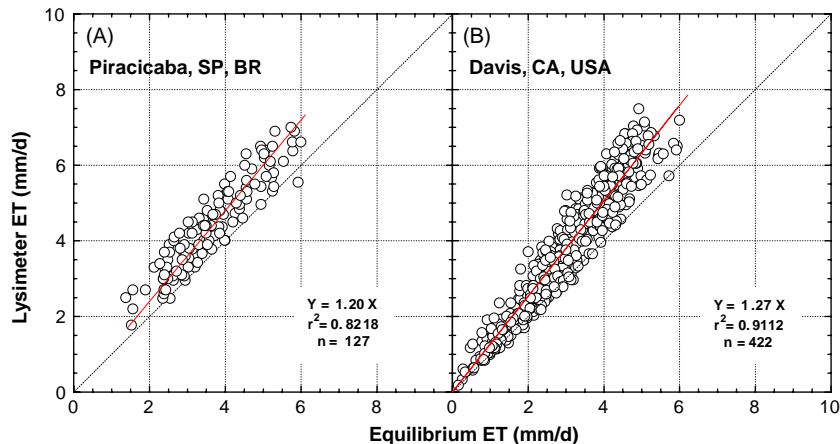

Fig. 1. Overall Priestley–Taylor α parameter for grass reference evapotranspiration: (A) Piracicaba, SP, Brazil; (B) Davis, CA, USA.
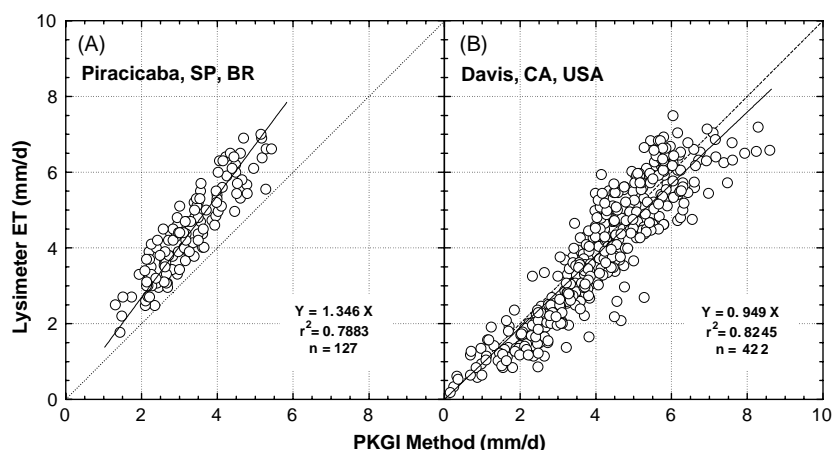
Fig. 2. Grass reference evapotranspiration predicted by the Perrier–Katerji–Goose–Itier method vs. lysimeter measurements: (A) Piracicaba, SP, Brazil; (B) Davis, CA, USA.

The above results indicate that the PKGI model for α did not work well for the present data sets. The main problem with this method is the determination of a critical resistance (r^* in Eq. (8)) with daily averages because it depends directly on the vapor pressure deficit (D_a), a non-linear function of the temperature. Such difficulty is expressed by the many methods available for estimating average D_a giving very distinctive results as shown in several reports (Doorenbos and Pruitt, 1977; Sadler and Evans, 1989; Allen et al., 1998). Rana et al. (1994) have found that the PKGI method works better on an hourly time scale rather

than on a daily basis. No attempt was made here to check their findings due to lack of the necessary hourly data set at this time. Their proposal of an empirical $C = 0.11(r^*/r_a) + 0.9$ for a daily time scale did not change statistically the results shown in Fig. 2 for both locations.

The Alpha-Omega (AO) approach ($\alpha = \Omega^{-1}$) represented a better alternative than the PKGI method as can be seen in Fig. 3. For both locations the points spread around the perfect fit line (1:1), with the following statistical relationships—(a) Piracicaba: $E_{Lys} = 1.02E_{AO}$, $r^2 = 0.8628$; (b) Davis: $E_{Lys} = 1.00E_{AO}$,
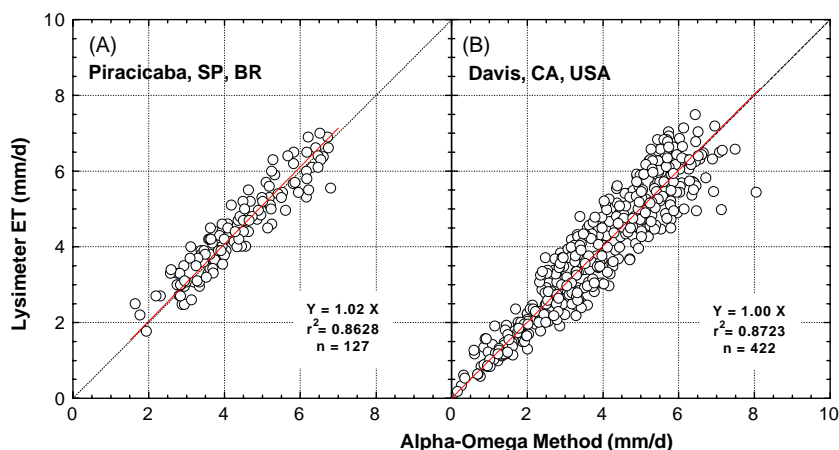

Fig. 3. Grass reference evapotranspiration predicted by the Alpha-Omega method vs. lysimeter measurements: (A) Piracicaba, SP, Brazil; (B) Davis, CA, USA.
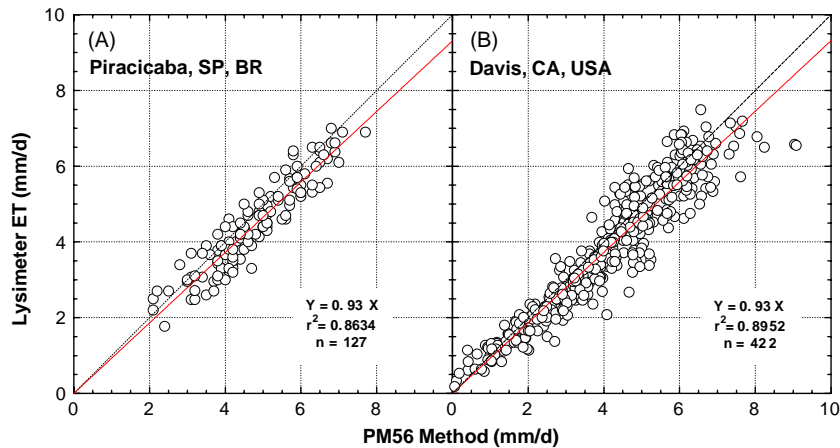
Fig. 4. Grass reference evapotranspiration predicted by the Penman–Monteith FAO-56 (PM56) vs. lysimeter measurements: (A) Piracicaba, SP, Brazil; (B) Davis, CA, USA.

$r^2 = 0.8723$. By all statistical standards (Table 1), the performance of the Alpha-Omega approach can be considered as excellent. This approach allows for a variable α parameter and it does not require local calibrations, eliminating the need for determining the critical resistance (r^*), and consequently the need for humidity measurements The FAO-56 parameterization $r_c/r_a = 0.34U_2$ worked well and no refinement was necessary at this time.

As the Penman–Monteith FAO-56 (PM56) is nowadays the only method recommended by FAO for estimating reference evapotranspiration (E_0), even when some input variables have to be estimated (Allen et al., 1998), the performance of this method is also here included for the sake of comparisons (Fig. 4). Similarly to the AO approach, the PM56 had almost identical statistical relationships for both locations, or (a) Piraci-

caba: $E_{Lys} = 0.93E_{PM56}$, $r^2 = 0.8634$; (b) Davis: $E_{Lys} = 0.93E_{PM56}$, $r^2 = 0.8952$. Such results indicate that the PM56 resulted in a overprediction of 7% on average.

Another way to evaluate the performance of the methods and approaches is to compare the corresponding accumulated E_0 (mm) against the accumulated E_{Lys}. The ratio estimated/measured close to 1 indicates good performance. Table 2 summarizes all results for both locations. As discussed before the Davis results were separated according to the daily minimum temperature (T_{min}). Days with $T_{min} < 5\,^{\circ}C$ had the worst estimates for all methods. Even though the PKGI approach uses the same input variables of the PM56 its performance varied from large underprediction for Piracicaba to large overprediction for the Davis cool days ($T_{min} < 5\,^{\circ}C$). Its performance is even inferior

Table 2

Comparisons of accumulated E_0 (mm) given by the Priestley–Taylor (three approaches) and Penman–Monteith (PM56) methods, and the lysimetric ET (E_{Lys}) for Piracicaba, SP, BR, and Davis, CA, USA

Method	Piracicaba		Davis					
	All	Ratio	$T_{min} > 5\,^{\circ}C$	Ratio	$T_{min} < 5\,^{\circ}C$	Ratio	All	Ratio
E_{lys}	564.1	–	1345.3	–	283.8	–	1629.1	–
$\alpha = 1.26$	584.7	1.04	1305.8	0.97	314.6	1.10	1620.4	0.99
$\alpha = C$	414.0	0.73	1356.6	1.01	376.7	1.32	1733.3	1.06
$\alpha = \Omega^{-1}$	548.8	0.97	1292.6	0.96	318.1	1.12	1610.7	0.99
PM56	605.6	1.07	1429.1	1.06	318.7	1.12	1747.8	1.07

The Davis results were separated according to the daily minimum temperature (T_{min}). Ratio: estimated/lysimeter.
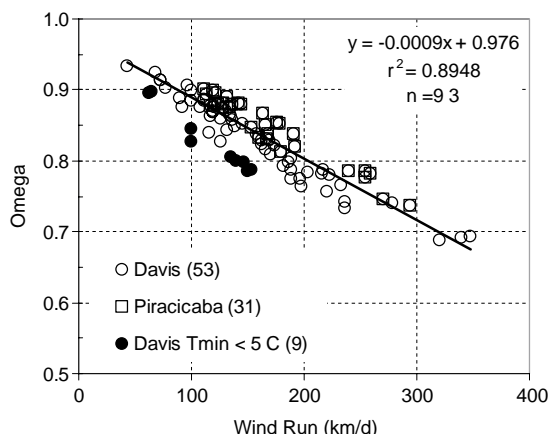
Fig. 5. The decoupling factor (Omega) as a function of the wind run.

to that obtained assuming the fixed $\alpha = 1.26$. Conversely, the less input demanding AO approach gave consistent good results for all conditions.

The good performance of the AO approach is an indication that Ω is not a constant even for large fields of short grasses. Selecting days with almost perfect fit of the E_{PM56} with the E_{Lys} it was possible to analyze the range of Ω values under very distinctive weather conditions. During the 93 selected days (62 for Davis and 31 for Piracicaba) the ET varied from 0.60 to 7.0 mm day^{-1} covering the whole range of the measured values. For both locations Ω varied from 0.69 to 0.93 with average of 0.83, a value close to the 0.8 reported by McNaughton and Jarvis (1983). There was a negative linear relationship between Ω and the wind run (i.e., $\Omega = 0.976 - 0.0009 W_R$, $r^2 = 0.8948$, $n = 93$) with high values occurring during calm days and low values being characteristic of very wind days, as can be seen in Fig. 5. This statistical relationship is a simplification of Eq. (2) and it can be used in the AO approach for estimating E_0 with equivalent efficiency.

Acknowledgements

References

Allen, R.G., Jensen, M.E., Wright, J.L., Burman, R.D., 1989. Operational estimates of reference evapotranspiration. Agron. J. 81, 650–662.

Allen, R.G., Pereira, L.S., Raes, D., Smith, M., 1998. Crop evapotranspiration: guidelines for computing crop requirements. Irrigation and Drainage Paper 56. FAO, Rome, Italy, 300 pp.

Baldocchi, D., 1994. A comparative study of mass and energy exchange over a closed C3 (wheat) and an open C4 (corn) canopy. I. The partitioning of available energy into latent and sensible heat exchange. Agric. For. Meteorol. 67, 191–220.

Barton, I.J., 1979. A parameterization of the evaporation from nonsaturated surfaces. J. Appl. Meteorol. 18, 43–47.

Daudet, F.A., Le Roux, X., Sinoquet, H., Adam, F., 1999. Wind speed and leaf boundary layer conductance variation within tree crown: consequences on leaf-to-atmosphere coupling and tree functions. Agric. For. Meteorol. 97, 171–185.

Davies, J.A., Allen, C.D., 1973. Equilibrium, potential and actual evaporation from cropped surfaces in Southern Ontario. J. Appl. Meteorol. 12, 649–657.

De Bruin, H.A.R., 1983. A model for the Priestley–Taylor parameter α. J. Clim. Appl. Meteorol. 22, 572–578.

Doorenbos, J., Pruitt, W.O., 1977. Guidelines for predicting crop water requirements. FAO Irrigation and Drainage Paper 22, 2nd ed. FAO, Rome, Italy, 179 pp.

Flint, A.L., Childs, S.W., 1991. Use of the Priestley–Taylor evaporation equation for soil water limited conditions in a small forest clearcut. Agric. For. Meteorol. 56, 247–260.

Jacovides, C.P., Kontonyiannis, H., 1995. Statistical procedures for the evaluation of evapotranspiration computing models. Agric. Water Manage. 27, 365–371.

Jarvis, P.G., 1985. Coupling of transpiration to the atmosphere in horticultural crops: the omega factor. Acta Hortic. 171, 187–205.

Jarvis, P.G., McNaughton, K.G., 1986. Stomatal control of transpiration: scaling up from leaf to region. Adv. Ecol. Res. 15, 1–49.

Jones, H.G., 1994. Plants and Microclimate: A Quantitative Approach to Environmental Plant Physiology, 2nd ed. Cambridge University Press, 428 pp.

Jones, H.G., Tardieu, F., 1998. Modelling water relations of horticultural crops: a review. Sci. Hortic. 74, 21–46.

Jury, W.A., Tanner, C.B., 1975. Advection modification of the Priestley–Taylor evapotranspiration formula. Agron. J. 67, 840–842.

Katerji, N., Perrier, A., 1983. Modélisation de l'evapotranspiration réelle d'une parcelle de luzerne: role d'un coefficient culturalle. Agronomie 3 (6), 513–521.

Ksalveth, T.O., 1985. Cautionary note about R^2. Am. Stat. 39 (4), 279–285.

Martin, T.A., Brown, K.J., Kucera, J., Meinzer, F.C., Sprugel, D.G., Hinckley, T.M., 2001. Control of transpiration in a 220-year-old *Abies amabilis* forest. For. Ecol. Manage. 152, 211–224.

McNaughton, K.G., 1976a. Evaporation and advection. I. Evaporation from extensive homogeneous surfaces. Quart. J. Roy. Meteorol. Soc. 102, 181–191.

McNaughton, K.G., 1976b. Evaporation and advection. II. Evaporation downwind of a boundary separating regions having different surface resistance and available energies. Quart. J. Roy. Meteorol. Soc. 102, 193–202.

McNaughton, K.G., Jarvis, P.G., 1983. Predicting effects of vegetation changes on transpiration and evaporation. In: Kozlowski, T.T. (Ed.), Water Deficits and Plant Growth, vol. VII. Academic Press, pp. 1–47.

Monteith, J.L., 1965. Evaporation and environment. Symp. Soc. Exptl. Biol. 19, 205–234.

Monteith, J.L., 1981. Evaporation and surface temperature. Quart. J. Roy. Meteorol. Soc. 107, 1–27.

Monteith, J.L., 1995. Accommodation between transpiring vegetation and the convective boundary layer. J. Hydrol. 166, 251–263.

Montgomery, D.C., Peck, E.A., 1992. Introduction to Linear Regression Analysis. Wiley, New York, 527 pp.

Mukammal, E.I., Neumann, H.H., 1977. Application of the Priestley–Taylor evaporation model to assess the influence of the soil moisture on the evaporation from a large weighing lysimeter and class A pan. Bound.-Lay. Meteorol. 12, 243–256.

Penman, H.L., 1948. Natural evaporation from open water, bare soil and grass. Proc. Roy. Soc. London A193, 120–146.

Pereira, A.R., Villa Nova, N.A., 1992. Analysis of the Priestley–Taylor parameter. Agric. For. Meteorol. 61, 1–9.

Perrier, A., 1975. Étude physique de l'évapotranspiration dans le conditions naturelles. II. expressions et paramètres donnant l'évapotranspiration réele d'une surface "mince". Ann. Agron. 26, 105–123.

Perrier, A., Katerji, N., Goose, G., Itier, B., 1980. Etude "in situ" de l'évapotranspiration reele d'une culture de blé. Agric. Meteorol. 21, 295–311.

Peterschmitt, J.M., Perrier, A., 1991. Evapotranspiration and canopy temperature of rice and groundnut in southeast coastal India. Crop coefficient approach and relationship between evapotranspiration and canopy temperature. Agric. For. Meteorol. 56, 273–298.

Priestley, C.H.B., Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Mon. Weather Rev. 100 (2), 81–92.

Pruitt, W.O., 1964. Cyclic relations between evapotranspiration and radiation. Trans. ASAE 7 (3) 271–275 and 280.

Pruitt, W.O., Doorenbos, J., 1977. Empirical calibration, a requisite for evapotranspiration formulae based on daily or longer mean climatic data? In: Proceedings of the International Round Table Conference on "Evapotranspiration", International Commission on Irrigation and Drainage, Budapest, 20 pp.

Rana, G., Katerji, N., Mastrorilli, M., El Moujabber, M., 1994. Evapotranspiration and canopy resistance of grass in a Mediterranean region. Theoret. Appl. Climatol. 50, 61–71.

Rana, G., Katerji, N., Mastrorilli, M., 1997. Environmental and soil-plant parameters for modelling actual crop evapotranspiration under water stress conditions. Ecol. Model. 101, 363–371.

Rana, G., Katerji, N., Mastrorilli, M., 1998. Canopy resistance modelling for crops in contrasting water conditions. Phys. Chem. Earth 23, 433–438.

Raupach, M.R., 2000. Equilibrium evaporation and the convective boundary layer. Bound.-Lay. Meteorol. 96, 107–141.

Sadler, E.J., Evans, D.E., 1989. Vapour pressure deficit calculations and their effect on the combination equation. Agric. For. Meteorol. 49, 55–80.

Slatyer, R.O., McIlroy, I.C., 1961. Practical Micrometeorology. CSIRO, Melbourne, 310 pp.

Smith, D.M., Jarvis, P.G., 1998. Physiological and environmental control of transpiration by trees in windbreaks. For. Ecol. Manage. 105, 159–173.

Steduto, P., Hsiao, T.C., 1998. Maize canopies under two soil water regimes. III. Variation in coupling with the atmosphere and the role of leaf water index. Agric. For. Meteorol. 89, 201–213.

Stone, R.J., 1993. Improved statistical procedure for the evaluation of solar radiation estimation models. Solar Energy 51, 289–291.

Viswanadham, Y., Silva Filho, V.P., André, R.G.B., 1991. The Priestley–Taylor parameter α for the Amazonian forest. For. Ecol. Manage. 38, 211–225.

Williams, R.J., Broersma, K., van Ryswyk, A.L., 1978. Equilibrium and actual evapotranspiration from a very dry vegetated surface. J. Appl. Meteorol. 17, 1827–1832.

Willmott, C.J., 1981. On the validation of models. Phys. Geogr. 2, 184–194.

Wullschleger, S.D., Wilson, K.B., Hanson, P.J., 2000. Environmental control of whole-plant transpiration, canopy conductance and estimates of the decoupling coefficient for large red maple trees. Agric. For. Meteorol. 104, 157–168.

Zacharias, S., Heatwole, C.D., Coakley, C.W., 1996. Robust quantitative techniques for validating pesticide transport models. Trans. ASAE 39, 47–54.