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 $\alpha$ parameter can be set equal to the inverse of the McNaughton–Jarvis decoupling factor $\Omega$, that is, $\alpha = \Omega^{-1}$. This approach does not require local calibrations and it allows for a variable $\alpha$. 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 $\Omega$ values from 0.69 to 0.93. $\Omega$ decreased linearly as a function of the wind run.

The Perrier–Katerji–Goose–Itier model for $\alpha$ 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.

Keywords: Equilibrium evapotranspiration; Crop coefficient; Critical resistance; Penman–Monteith

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 ($\Omega$), defined by McNaughton and Jarvis (1983), based on the Penman–Monteith
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 $\Omega$ 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 Hisao, 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 $\Omega$. A 3D canopy simulation model indicated the existence of a spatial variation of $\Omega$ within the crown of an isolated walnut tree, with the sunlit leaves less coupled than the shaded ones (Daudet et al., 1999). Therefore, $\Omega$ 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, $E_0$ represents about 80% of $E_{\text{eq}}$, on average ($\Omega = 0.8$, McNaughton and Jarvis, 1983). Priestley and Taylor (1972) proposed to neglect the aerodynamic term and correct $E_{\text{eq}}$ by a dimensionless coefficient $\alpha$ (the Priestley–Taylor parameter) as $E_0 = \alpha E_{\text{eq}}$. Experimental results from several sites around the world, including vegetated surfaces and large water bodies (lake and oceans), gave $\alpha$ values in the range between 1.08 ± 0.01 and 1.34 ± 0.05, with average of 1.26 (Priestley and Taylor, 1972).

Empirically, $\alpha$ 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 $\alpha$; in general, $\alpha$ 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 $\Omega$ 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 $\Omega$ it will be tested here the goodness of the working hypothesis that the reference evapotranspiration can be computed as $E_0 = \Omega^{-1}E_{\text{eq}}$, or $\alpha = \Omega^{-1}$. It can be a mere coincidence but for a grass field $\alpha = 1.26 = 0.8^{-1}$ 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 $\alpha$ 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 ($E_{\text{rad}}$) and aerodynamic ($E_{\text{aero}}$) terms to the overall evapotranspiration ($E$, mm day$^{-1}$), McNaughton and Jarvis (1983) presented it in the following form:

$$E = E_{\text{rad}} + E_{\text{aero}} = \Omega E_{\text{eq}} + (1-\Omega)E_0$$

(1)

defining $\Omega$ as the dimensionless decoupling factor, computed by

$$\Omega = \left[1 + \frac{\gamma}{s + \frac{\rho_a}{\rho_d}}\right]^{-1}$$

(2)

being $\gamma$ (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;
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 $\lambda$ is latent heat of vaporization here taken as 2.45 MJ kg$^{-1}$.

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

$$E_m = \frac{p_c D_v}{2 s \gamma c}$$

where $p_c$ (kPa) is the volumetric specific heat of the air at constant pressure.

Pриестли и Тейлор (1972) предложили, что $E$ может быть рассчитано по формуле

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

где $\alpha$ компенсирует диском для $E_m$.

Выражение Пенман–Монтейт уравнение как безразмерное отношение доступной энергии ($R_n - G$), или

$$E = \frac{[(s/\gamma) + \left[(\rho c_p D_v/\gamma (R_n - G) + \gamma (r_c + r_a)/r_a)]}{(s/\gamma) + (r_c + r_a)/r_a}$$

Монтейт (1965) определил «неотходящую сопротивляемость» ($r_c$) как

$$r_c = \frac{\rho c_p D_v}{\gamma (R_n - G)}$$

или

$$E = \frac{\rho c_p D_v}{\gamma (R_n - G)}$$

и показал, что $r_c$ является критическим значением $r_c + r_a$, при котором

$$E = R_n - G$$

Подставив условие $\frac{\partial E}{\partial r} = 0$ в уравнение (1), изотермическое сопротивление принимает критическое значение

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

$$r^* = \frac{(\gamma + \lambda) C E_0}{\rho c_p D_v / \gamma (R_n - G)}$$

определяет условия, когда $E$ является независимым от $r_a$ (т.е., независимым от ветра) и $E = E_{eq}$.

Соединив уравнения (1)–(4) и (8), Перриер и др. (1980) показали, что

$$E = E_{eq}$$

с

$$C = \frac{1}{1 + (\gamma/s) (\gamma/\lambda)(r_a/r_c)}$$


3. Материалы и методы

Последовательность Приестли и Тейлор (1972), ежеходное значение испарения ($E_{eq}$) была получена с уравнением (3) с 24-часовым радиационным структурным уравнением, не учитывало теплового потока в почву, и с уравнением стратификации вверх от поверхности. Дневное значение испарения в почву, с уравнением (6) и (7) и последовательно, была использована в качестве функции $E_{eq}$, используя два схемы. Первый подход был использован Перриер и др. (1980) выражением (8)–(10) и далее и далее, и далее, и далее, и далее,


The second approach is an alternative hypothesis based on Monteith (1965) and Perrier (1975) arguments and it takes $a = G/\Omega$ (Alpha-Omega method), as defined in Eq. (2). For both schemes the parameterisation for a reference grass surface given by Allen et al. (1989, 1998) was adopted, i.e., $r_a/r_c = 0.34 U_2$, and $r_a = 208/U_2$ with the daily average wind at the 2 m ($U_2$, m s$^{-1}$) above the ground.
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_{\text{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_{\text{Lys}} were discarded based on the criteria $0.5 < E_{\text{Lys}}/R_{\text{es}} < 0.9$, 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 ($=0.92$ m$^2$), located about 80 m from the leading edge of the most prevailing wind direction, in a 35 m × 90 m field of Festuca pratensis 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 $E_{\text{Lys}}/R_{\text{es}}$ 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 ($\bar{E}$) and observed (O) ET. 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

$$\text{MAE} = \max(|O_i - \bar{E}_i|)^{1/N_{\text{days}}}$$

$$\text{MBE} = \frac{\sum_{i=1}^{N_{\text{days}}}(O_i - \bar{E}_i)}{N}$$

$$d = 1 - \frac{\sum_{i=1}^{N_{\text{days}}}(O_i - \bar{E}_i)^2}{\sum_{i=1}^{N_{\text{days}}}(O_i - \bar{O})^2}$$

$$\text{EF} = 1 - \frac{\sum_{i=1}^{N_{\text{days}}}(O_i - \bar{E}_i)^2}{\sum_{i=1}^{N_{\text{days}}}(O_i - \bar{O})^2}$$

$$t = \frac{\bar{O} - \bar{E}}{\left[\left(\frac{1}{N_{\text{days}}} \sum (O_i - \bar{O})^2 + \frac{1}{N_{\text{days}}} \sum (\bar{E}_i - \bar{O})^2\right)^{1/2}\right]}$$

where $\bar{O}$ and $\bar{E}$ are the means of the observed and estimated values; $s_{O}^2$ and $s_{\bar{E}}^2$ are the respective variances. The perfect model will have $b = 1$; $d = r^2 = 1$ 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 ($E_{\text{eq}}$) with lysimeter measurements ($E_{\text{Lys}}$) it is possible to determine an overall average Priestley–Taylor parameter $a$ (or $C$ for Perrier et al., 1980). For the tropical climate of Piracicaba, SP, Brazil, the comparisons resulted in $a = 1.20$ ($\pm 0.01$), $r^2 = 0.8218$ (Fig. 1A); for the semi-arid climate of Davis, CA, USA, it resulted in $a = 2.27$ ($\pm 0.05$), $r^2 = 0.9112$ (Fig. 1B). The original Priestley–Taylor proposal of a fixed average $a = 1.26$ worked well for Davis in the absence of heat advection ($E_{\text{ad}} = 1.0E_{\text{PT1.26}}$; $r^2 = 0.9088$). For Piracicaba there was a slight over-prediction of 5%, on average, as $E_{\text{Lys}} = 0.952E_{\text{PT1.26}}$; $r^2 = 0.8218$.

The $a$ 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 $E_{\text{Lys}}/R_{\text{es}}$ ratio than cooler days as indicated by relationship $E_{\text{Lys}}/R_{\text{es}} = 0.458 + 0.0108T_{\text{max}}$ ($r^2 = 0.4689$), with $3.9^\circ\text{C} < T_{\text{max}} < 37.2^\circ\text{C}$. 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 ($\alpha$), decoupling factor ($\Omega$).

<table>
<thead>
<tr>
<th>Method</th>
<th>$b$</th>
<th>$r^2$</th>
<th>MAE (mm day$^{-1}$)</th>
<th>MBE (mm day$^{-1}$)</th>
<th>RMSE (mm day$^{-1}$)</th>
<th>$d$</th>
<th>EF</th>
<th>$\bar{E}$ (mm day$^{-1}$)</th>
<th>$s_E$ (mm day$^{-1}$)</th>
<th>$t$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Piracicaba, SP, BR$^a$</td>
<td>$\alpha = 1.26$</td>
<td>0.952</td>
<td>0.8218</td>
<td>0.95</td>
<td>0.16</td>
<td>0.50</td>
<td>0.95</td>
<td>0.78</td>
<td>4.60</td>
<td>1.34</td>
</tr>
<tr>
<td></td>
<td>$\alpha = C$</td>
<td>1.345</td>
<td>0.7863</td>
<td>2.24</td>
<td>$-1.38$</td>
<td>0.54</td>
<td>0.73</td>
<td>$-0.21$</td>
<td>3.26</td>
<td>0.94</td>
</tr>
<tr>
<td></td>
<td>$\alpha = \Omega$</td>
<td>1.02</td>
<td>0.8636</td>
<td>1.24</td>
<td>$-0.12$</td>
<td>0.43</td>
<td>0.97</td>
<td>0.86</td>
<td>4.32</td>
<td>1.22</td>
</tr>
<tr>
<td>PM56</td>
<td>0.93</td>
<td>0.8634</td>
<td>1.42</td>
<td>0.33</td>
<td>0.46</td>
<td>0.94</td>
<td>0.78</td>
<td>4.77</td>
<td>1.19</td>
<td>2.22</td>
</tr>
<tr>
<td>Davis, CA, USA$^b$</td>
<td>$\alpha = 1.26$</td>
<td>1.00</td>
<td>0.9088</td>
<td>1.49</td>
<td>0.02</td>
<td>0.51</td>
<td>0.98</td>
<td>0.91</td>
<td>3.85</td>
<td>1.64</td>
</tr>
<tr>
<td></td>
<td>$\alpha = C$</td>
<td>0.949</td>
<td>0.8245</td>
<td>2.59</td>
<td>0.25</td>
<td>0.72</td>
<td>0.94</td>
<td>0.81</td>
<td>4.12</td>
<td>1.55</td>
</tr>
<tr>
<td></td>
<td>$\alpha = \Omega$</td>
<td>1.00</td>
<td>0.8723</td>
<td>2.62</td>
<td>$-0.04$</td>
<td>0.62</td>
<td>0.97</td>
<td>0.87</td>
<td>3.83</td>
<td>1.66</td>
</tr>
<tr>
<td>PM56</td>
<td>0.93</td>
<td>0.8952</td>
<td>2.53</td>
<td>0.28</td>
<td>0.60</td>
<td>0.97</td>
<td>0.86</td>
<td>4.15</td>
<td>1.76</td>
<td>2.33</td>
</tr>
</tbody>
</table>

$^a$ $O = 4.44$ mm day$^{-1}$, $s_E = 1.17$ mm day$^{-1}$, $n = 127$, $t_c = 2.62$.

$^b$ $O = 3.87$ mm day$^{-1}$, $s_E = 1.72$ mm day$^{-1}$, $n = 422$, $t_c = 2.60$.

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

In regard to the Perrier–Katerji–Goose–Itier method (Fig. 2), for Piracicaba it underestimated the measurements by 35%, on average, as $E_{\text{Lys}} = 1.346E_{\text{PKGI}}$, $r^2 = 0.7883$. 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 $E_{\text{Lys}} = 0.949E_{\text{PKGI}}$, $r^2 = 0.8243$. Days with large overestimations in Davis had minimum temperature below 5°C and were all in the period October–May. The $T_{\text{min}}$ threshold was selected based on visual inspection of the most discrepant points in Fig. 2B. Analysis of the 123 days with $T_{\text{min}} < 5$°C indicated gross overpredictions, or $E_{\text{Lys}} = 0.763E_{\text{PKGI}}$, $r^2 = 0.8243$. During such days the average $E_{\text{Lys}}$ was equal to 2.31 mm day$^{-1}$. Results for the remaining 298 days with $T_{\text{min}} > 5$°C displayed a much better fit with $E_{\text{Lys}} = 0.994E_{\text{PKGI}}$, $r^2 = 0.7963$, and the average $E_{\text{Lys}}$ equal to 4.51 mm day$^{-1}$.

Fig. 1. Overall Priestley–Taylor $\alpha$ parameter for grass reference evapotranspiration: (A) Piracicaba, SP, Brazil; (B) Davis, CA, USA.
The above results indicate that the PKGI model for \( \alpha \) did not work well for the present data sets. The main problem with this method is the determination of a critical resistance \( (r^* \text{ in Eq. (8)}) \) 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: \( E_{Lys} = 1.00E_{AO} \).
$r^2 = 0.8723$. By all statistical standards (Table 1), the performance of the Alpha-Omega approach can be considered as excellent. This approach allows for a variable $\alpha$ 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 now-a-days 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) Piracicaba: $E_{Lys} = 0.93E_{PM56}$, $r^2 = 0.8634$; (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

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

<table>
<thead>
<tr>
<th>Method</th>
<th>Piracicaba</th>
<th>Davis</th>
</tr>
</thead>
<tbody>
<tr>
<td>All $E_0$</td>
<td>564.1</td>
<td>1345.3</td>
</tr>
<tr>
<td>Ratio $T_{min} &gt; 5, ^\circ C$</td>
<td>283.8</td>
<td>-</td>
</tr>
<tr>
<td>Ratio $T_{min} &lt; 5, ^\circ C$</td>
<td>314.6</td>
<td>1.10</td>
</tr>
<tr>
<td>Ratio All</td>
<td>1629.1</td>
<td>0.99</td>
</tr>
<tr>
<td>$\alpha = 1.26$</td>
<td>564.7</td>
<td>1.04</td>
</tr>
<tr>
<td>$a = 0.73$</td>
<td>1305.8</td>
<td>0.97</td>
</tr>
<tr>
<td>$a = 1.0$</td>
<td>767.3</td>
<td>1.01</td>
</tr>
<tr>
<td>$a = 1.32$</td>
<td>318.4</td>
<td>1.10</td>
</tr>
<tr>
<td>$a = 1.42$</td>
<td>318.7</td>
<td>1.12</td>
</tr>
<tr>
<td>$a = 1.67$</td>
<td>1747.8</td>
<td>1.07</td>
</tr>
<tr>
<td>PM56</td>
<td>605.6</td>
<td>1.07</td>
</tr>
<tr>
<td>$T_{min} = 5, ^\circ C$</td>
<td>1429.1</td>
<td>1.06</td>
</tr>
<tr>
<td>$T_{min} &lt; 5, ^\circ C$</td>
<td>318.1</td>
<td>1.12</td>
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<tr>
<td>$T_{min} &lt; 5, ^\circ C$</td>
<td>1610.7</td>
<td>0.99</td>
</tr>
</tbody>
</table>

The Davis results were separated according to the daily minimum temperature ($T_{min}$). Ratio: estimated/lysimeter.
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 $\Omega$ 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 $\Omega$ 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 $\Omega$ 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 $\Omega$ and the wind run (i.e., $\Omega = 0.976 - 0.0009W_R$, $r^2 = 0.8948$, $n = 93$) with high values occurring during calm days and low values being characteristic of very windy 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.

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References


