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Use of the Priestley-Taylor evaporation equation for soil water limited conditions in a small forest clearcut

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ABSTRACT

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The Priestley-Taylor equation, a simplification of the Penman equation, was used to allow calculations of evapotranspiration under conditions where soil water supply limits evapotranspiration. The Priestley-Taylor coefficient, α, was calculated to incorporate an exponential decrease in evapotranspiration as soil water content decreases. The method is appropriate for use when detailed meteorological measurements are not available. The data required to determine the parameter for the α coefficient are net radiation, soil heat flux, average air temperature, and soil water content. These values can be obtained from measurements or models.

The dataset used in this report pertains to a partially vegetated clearcut forest site in southwest Oregon with soil depths ranging from 0.48 to 0.70 m and weathered bedrock below that. Evapotranspiration was estimated using the Bowen ratio method, and the calculated Priestley-Taylor coefficient was fitted to these estimates by nonlinear regression. The calculated Priestley-Taylor coefficient (α') was found to be approximately 0.9 when the soil was near field capacity (0.225 cm^3 cm^{-3}). It was not until soil water content was less than 0.14 cm^3 cm^{-3} that soil water supply limited evapotranspiration. The soil reached a final residual water content near 0.05 cm^3 cm^{-3} at the end of the growing season.

INTRODUCTION

A major limitation of many reforestation efforts is a lack of plant-available water during the growing season. In areas with xeric climates, this is often combined with high air and soil temperatures which act to increase plant stress. Therefore, information regarding both water supply and environmental demand is required to properly assess the harshness of sites for reforestation. The measurement and application of the surface energy budget is a useful analytical approach because components of radiation, soil heat flow and soil water environments are included. This approach does, however, require detailed, site-specific measurements for the initial calibration.

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Previous work

A number of simplifications of surface energy budget techniques used to estimate evapotranspiration have been used to decrease the quantity and the intensity of measurements required. The Penman equation (Penman, 1948) is commonly used in situations where detailed meteorological data are available. The assumptions and simplifications used by Penman (1948) to model the aerodynamic components of evapotranspiration make the Penman equation useful only for calculating potential evapotranspiration. Furthermore, the equation requires calibration for specific sites. A modified version of Penman's original equation, the Penman–Monteith equation (Monteith, 1964), allows calculation of actual evapotranspiration but requires detailed knowledge about the resistances to heat and water flow at the evaporating surface, making its use impractical without intensive meteorological and surface measurements. Priestley and Taylor (1972) suggested a modification of the Penman equation, which requires less extensive measurements:

$$\lambda E_a = \alpha \frac{s}{s+\gamma}(R_n - G) \tag{1}$$

where E_a is actual evapotranspiration, λ is the latent heat of vaporization, α is a model coefficient (which Priestley and Taylor allowed to vary for drying conditions), s is the slope of the saturation vapor density curve, γ is the psychrometric constant, R_n is net radiation, and G is soil heat flux. To relate this formulation to the Penman equation, the aerodynamic term is modeled as $(\alpha - 1)[s/(s+\lambda)](R_n - G)$. This simplification has proven to be successful because the radiation term generally dominates the aerodynamic term (Stewart, 1983).

The Priestley–Taylor coefficient, α, for daily calculations was determined to be 1.26 for freely evaporating surfaces (Priestley and Taylor, 1972; Stewart and Rouse, 1977) and E_a in eqn. (1) becomes potential evapotranspiration (E_p). In general, for wet surfaces with α set at 1.26, the combined term $\alpha[s/(s+\gamma)]$ is approximately 1 at air temperatures near 24°C and therefore evapotranspiration is approximately equal to available energy, $\lambda E_a = (R_n - G)$. For specific sites, the coefficient α depends on surface vegetation and microclimatic conditions and has a measured range from 0.72 for forest conditions to 1.57 for conditions of strong advection (Table 1). When $\alpha[s/(s+\gamma)]$ is less than 1, actual evapotranspiration is less than the potential energy-limited rate. This often occurs for dry canopy conditions and is controlled by surface resistance to evapotranspiration (De Bruin, 1983). When $\alpha[s/(s+\gamma)]$ is greater than 1, surface resistance is often low (for instance, a wet forest canopy), and α is a function of wind speed and aerodynamic resistance and $s/(s+\gamma)$ is a function of air temperature. Under these conditions, actual evapotranspiration may still be less than the potential energy-limited rate, be-

TABLE 1

Measured values of the Priestley–Taylor coefficient, α

α	Surface conditions	Reference
1.57	Strongly advective conditions	Jury and Tanner, 1975
1.29	Grass (soil at field capacity)	Mukammal and Neumann, 1977
1.27	Irrigated ryegrass	Davies and Allen, 1973
1.26	Saturated surface	Priestley and Taylor, 1972
1.26	Open-water surface	Priestley and Taylor, 1972
1.26	Wet meadow	Stewart and Rouse, 1977
1.18	Wet Douglas-fir forest	McNaughton and Black, 1973
1.12	Short grass	De Bruin and Holtslag, 1982
1.05	Douglas-fir forest	McNaughton and Black, 1973
1.04	Bare soil surface	Barton, 1979
0.84	Douglas-fir forest (unthinned)	Black, 1979
0.80	Douglas-fir forest (thinned)	Black, 1979
0.73	Douglas-fir forest (daytime)	Giles et al., 1984
0.72	Spruce forest (daytime)	Shuttleworth and Calder, 1979

cause actual evapotranspiration also depends on soil water availability, exchange surface properties, and the magnitude of net radiation (Priestley and Taylor, 1972; Jury and Tanner, 1975; Black, 1979; de Bruin, 1983). Calculations of actual evapotranspiration generally are based on modifications of a potential evapotranspiration equation to account for high surface resistance or low soil water supply. Methods involving calculation of surface resistance usually are based on the Penman–Monteith equation.

Modifications of the Priestley–Taylor equation to calculate actual evapotranspiration have used empirical relations based on soil water content. Often, α is redefined to be a function of soil water content (Davies and Allen, 1973; Mukammal and Neumann, 1977; Barton, 1979). Another approach is to define a soil water content below which evapotranspiration is limited and the Priestley–Taylor equation, with α fixed, is in error (Spittlehouse and Black, 1981) and no longer appropriate to use. This value of soil water content varies with soil type, vegetation and atmospheric conditions but covers a much smaller range when expressed as a percentage of total available soil water (Table 2), where available water (AWC) is the difference between root-zone water storage at field capacity and the amount remaining through the driest part of the year. For vegetated surfaces, from 50 to 80% of the available soil water can be extracted at the potential energy-limited rate. Bare-soil evapotranspiration is limited when approximately 40% of the available water is removed. This result is not unexpected (Tanner and Jury, 1976) and can be related to the depth distribution of water extraction in the soil profile.

TABLE 2

Percentage reduction in available water (AWC) before evapotranspiration is limited

AWC	Surface conditions	Reference
82	Douglas-fir forest (low demand)	Black and Spittlehouse, 1980
81	Lysimeter and bean crop	Priestley and Taylor, 1972
77	Lysimeter and field crop	Priestley and Taylor, 1972
75	Lysimeter and grass cover	Mukammal and Neumann, 1977
66	Douglas-fir forest (high demand)	Black and Spittlehouse, 1980
60	Douglas-fir forest	Black, 1979
55	Cropped surface	Davies and Allen, 1973
50	Forest clearcut	Flint and Childs, 1987a
50	Lysimeter and pasture crop	Priestley and Taylor, 1972
40	Bare soil surface	Estimate from Barton, 1979

OBJECTIVE AND APPROACH

The objective of this research was to calibrate the modified Priestley– Taylor equation for soil water limited conditions. This was done by substituting α' for the Priestley–Taylor coefficient α and making it a function of soil water content as suggested by Davies and Allen (1973) and Barton (1979). Although the original approach of Priestley and Taylor was to apply their formulation to large-scale environments, we applied the modified version to a small forest clearcut in Oregon.

METHODS

Field methods

Data for this study were obtained during a reforestation field experiment in southwest Oregon near Wolf Creek (42°43′N, 123°17′W, 715 m elevation) northeast of Grants Pass (see Flint and Childs (1987b) for complete details). The site is a steep, south-facing slope (190° azimuth, 30% slope) of approximately 16 ha. The soil is a moderately deep, loamy-skeletal, mesic mixed Typic Xerochrept. Soil water content at saturation averages 0.35 cm^3 cm^{-3}, residual water content is approximately 0.05 cm^3 cm^{-3}, and water content at field capacity averages 0.225 cm^3 cm^{-3}. Average soil depth is 0.66 m ranging between 0.48 and 0.70 m. Vegetation covered 81% of the site (30% shrubs from 0.3 to 0.6 m in height, 30% perennial forbs, 20% annual species, and 1% Douglas-fir seedlings). Climate measurements were made for the period 16 January–25 September 1983. Measurements of soil water content and soil temperature were made at ten locations and averaged for the site. Data were collected on ten dates between April and September 1983. Soil water content was measured using a two-probe gamma attenuation device (Model

2376, Troxler Laboratories, Research Triangle Park, NC) at 0.025-m depth intervals and corrected at the end of the season for dry soil density.

Soil temperatures were measured at five depths (0.02, 0.04, 0.08, 0.16, and 0.32 m) using five thermistors (YSI 44202, Yellow Springs Instruments, Yellow Springs, OH) in a plastic probe. Data were collected every 15 min and stored in a datalogger (Model CR-5, Campbell Scientific Inc., Logan, UT). Temperature data and soil heat capacities calculated from dry bulk density and water content were used to calculate soil heat flux using a calorimetric method (Fuchs, 1986).

Air temperatures were measured at 0.4 and 2.0 m using thermistors (YSI 44202) mounted in radiation shields. The thermistors were matched during laboratory calibration in a temperature controlled chamber. Dew point temperatures were measured at 0.4 and 2.0 m using LiCl dewcells (Holbo, 1981). The thermistors in the dewcells were also matched and calibrated in the laboratory using a humidity-controlled chamber (over salt solutions). Careful sensor matching was used to reduce the need for switching instruments for long-term field measurements. Net radiation was measured using a miniature all-wave net radiometer (C.W. Thornthwaite Associates, Camden, NJ). Rainfall data were collected with a tipping bucket rain gage (Texas Electronics, Arlington, TX) at 0.25 mm resolution. Sensor output was read every 10 s integrated for 30 min and stored (using a Model CR-21 datalogger, Campbell Scientific Inc., Logan, UT). Although soil water content and soil temperature were only collected periodically (ten dates for the season), the data necessary to calculate the Bowen ratio were continually measured over the season.

Modeling procedure

Actual evapotranspiration was calculated hourly using the Bowen ratio method

$$\lambda E_{\mathrm{a}} = \frac{(R_{\mathrm{n}} - G)}{1 + \beta} \qquad (2)$$

where β, the Bowen ratio, is the ratio of sensible to latent heat flux. β is calculated as

$$\beta = \frac{\rho C_{\mathrm{p}}(T_1 - T_2)}{\lambda(\rho_1 - \rho_2)} \qquad (3)$$

where ρC_{p} is the volumetric heat capacity of air, T_1 and T_2 are air temperatures at two heights (2.0 and 0.4 m), ρ_1 and ρ_2 are water vapor density at 2.0 and 0.4 m.

The Priestley–Taylor equation (eqn. (1)) was used to solve for α' by substituting λE_{a} calculated from eqn. (2) into eqn. (1) and solving for α'

$$\alpha' = \frac{\lambda E_a}{[(s)/(s+\gamma)](R_n - G)} \tag{4}$$

Although the coefficient α' could be related to any process that limits evapotranspiration (e.g. soil hydraulic resistance, aerodynamic resistance, stomatal resistance), we chose to relate α' to soil moisture status. Our approach is similar to other published methods. Davies and Allen (1973) used soil water content divided by soil water content at field capacity (θ/θ_{fc}), Flint and Childs (1987a) used soil water content divided by soil water content at saturation (θ/θ_s), and Barton (1979) simply used gravimetric water content without any scaling. Our formulation is

$$\alpha' = A[1 - \exp(B\Theta)] \tag{5}$$

where A and B are regression coefficients, and Θ is the relative water saturation calculated as

$$\Theta = \frac{(\theta - \theta_r)}{(\theta_s - \theta_r)} \tag{6}$$

where θ is soil water content, θ_r is residual soil water content and θ_s is soil water content at saturation. Spittlehouse and Black (1981) use a similar ratio to determine the soil water content below which the Priestley–Taylor equation is no longer valid. In their work, θ_s is replaced by θ_{fc} (water content at field capacity) in eqn. (6)

The regression of Θ on α' (eqn. (5)) is performed using values for α' calculated from a substitution of eqn. (2) into eqn. (4)

$$\alpha' = \frac{(R_n - G)/(1+\beta)}{[(s)/(s+\gamma)](R_n - G)} = \frac{s+\gamma}{s(1+\beta)} \tag{7}$$

In earlier work, Flint and Childs (1987a) used only data collected on the ten dates over the growing season when soil heat flux could be calculated directly. This paper extends the previous work by adding a more complete dataset and a more appropriate formulation of α' dependence on water content (use of eqn. (6) rather than θ/θ_s). Inspection of eqn. (7) shows that the calculation of α' is a function of air temperature and water vapor density and is independent of R_n and G. In order to provide a continuous dataset of soil water content and soil heat flux to determine the empirical parameters for the relationship between α' and Θ (eqn. (5)), certain assumptions and models were used. Our Bowen ratio weather station operated continuously over the season, but the detailed dataset required to calculate soil heat flux and water content was collected on 10 days over the season. Those 10 days of complete data were used to develop a linear relationship between net radiation and soil heat flux.

Daily estimates of soil water content were made using a mass balance water budget approach. The components of this method are: (1) assume the soil profile is at field capacity on 16 January 1983 (our first day of data collection from the Bowen ratio weather station); (2) add daily rainfall to the soil water, evenly distributed over 0.5 m soil depth; (3) assume any water held above field capacity drained below the 0.5 m depth after 24 h; (4) subtract 20% of the evapotranspiration calculated using the Bowen ratio with our linear model of soil heat flux (eqn. (2)). (The 20% factor accounts for an estimated 20% evapotranspiration that comes from vegetation with rooting depths greater that 0.5 m — our measurement limit for cross-hole gamma — and is discussed later.)

RESULTS AND DISCUSSION

Results of this study include a relationship between net radiation and soil heat flux, measured and modeled soil water budgets, and a relationship between α' and Θ. Figure 1 shows the soil heat flux–net radiation relationship. This relationship shows more detail and higher soil heat flux values than are commonly encountered. Daytime estimates and measurements of soil heat flux in agricultural settings are generally between 2 and 10% of net radiation

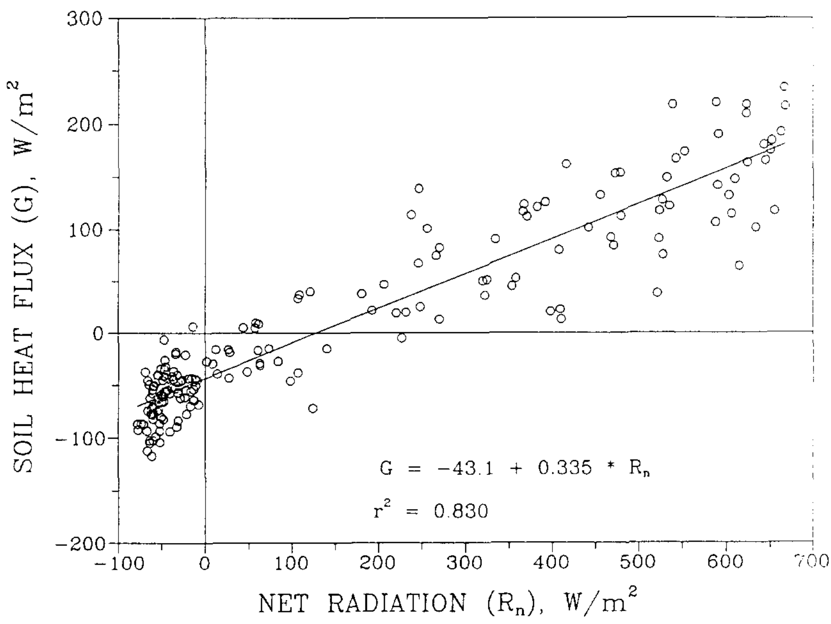

Fig. 1. Net radiation (R_n) vs. soil heat flex (G) for hourly values on 10 days during the growing season. Measured values (O) and model results from linear regression (–).

(Hanks and Ashcroft, 1980). Daytime soil heat flux from this rocky forest site averaged 23% of net radiation (Childs and Flint, 1989).

Daily evaporation was computed by summing 30-min Bowen ratio data (eqn. (2)). This method incorporated the net radiation–soil heat flux relationship of Fig. 1. Example datasets for spring and midsummer conditions are shown in Fig. 2. Solar radiation, net radiation, and soil heat flux patterns were shaped primarily by date, slope and aspect of the site (Figs. 2(a) and 2(d)). Evaporation rate (Figs. 2(b) and 2(e)) was calculated using the measured R_n, G, and the Bowen ratio (Figs. 2(c) and 2(f)). Springtime evapotranspiration (Fig. 2(b)) was almost twice as high as the midsummer rate (Fig. 2(e)). The modified Priestley–Taylor coefficient (α', eqns. (4) and (7)) was calculated every 30 min and was more variable than the Bowen ratio (β, eqns. (3) and (7)), particularly at those times when the energy budget terms were changing in sign and the estimate of λE_a was calculated erroneously using the Bowen ratio (eqn. (7) is undefined when $\beta = -1$) (Figs. 2(b) and 2(e)). In order to smooth the evapotranspiration calculations made using the Priestley–Taylor method, an average value of α' was calculated using daytime values of α' when $\beta > 0$. The λE_a in Fig. 2(b) and 2(c), using the Priestley–Taylor equation, had a fixed α' value of 0.87 for 20 May 1983 and 0.48 for 12 August 1983. For our analysis the value of α' was assumed to be dependent on Θ and therefore would not change significantly over a 24 h period. We feel our best estimate of α' is when β is holding steady over the midday period. The magnitude of error associated with applying the average of α' to early and late periods of the day is small because the values $(R_n - G)$ and, therefore, λE_a are small. The smoothing effect of the daytime average of α' can be seen in Figs. 2(c) and 2(f). The smoothed values were used in all subsequent calculations.

Soil water contents were estimated as discussed previously. The results of the mass balance water budget are shown in Fig. 3. Calculated values show good agreement with measured values for the top 0.5 m. The figure also shows the model fit obtained when the Priestley–Taylor equation is used to calculate λE_a rather than the measured Bowen ratio values. Any divergence between the two modeled curves represents a difference between modeled and measured α'.

The water removed by evapotranspiration and measured by our Bowen ratio station is most likely from some uneven distribution over the whole soil profile. As we only had site-specific data from the top 0.50 m (soil water determined using cross-hole gamma), we needed to determine or estimate the percentage of the total evapotranspiration that came from this region and the percentage that came from deeper in the soil. Site-specific vegetation data indicates that about 20% of the site was occupied by species rooting to depths of more than 0.50 m (Flint and Childs, 1987b). To estimate the correction factor for the evapotranspiration coming from depths of more than 0.50 m,

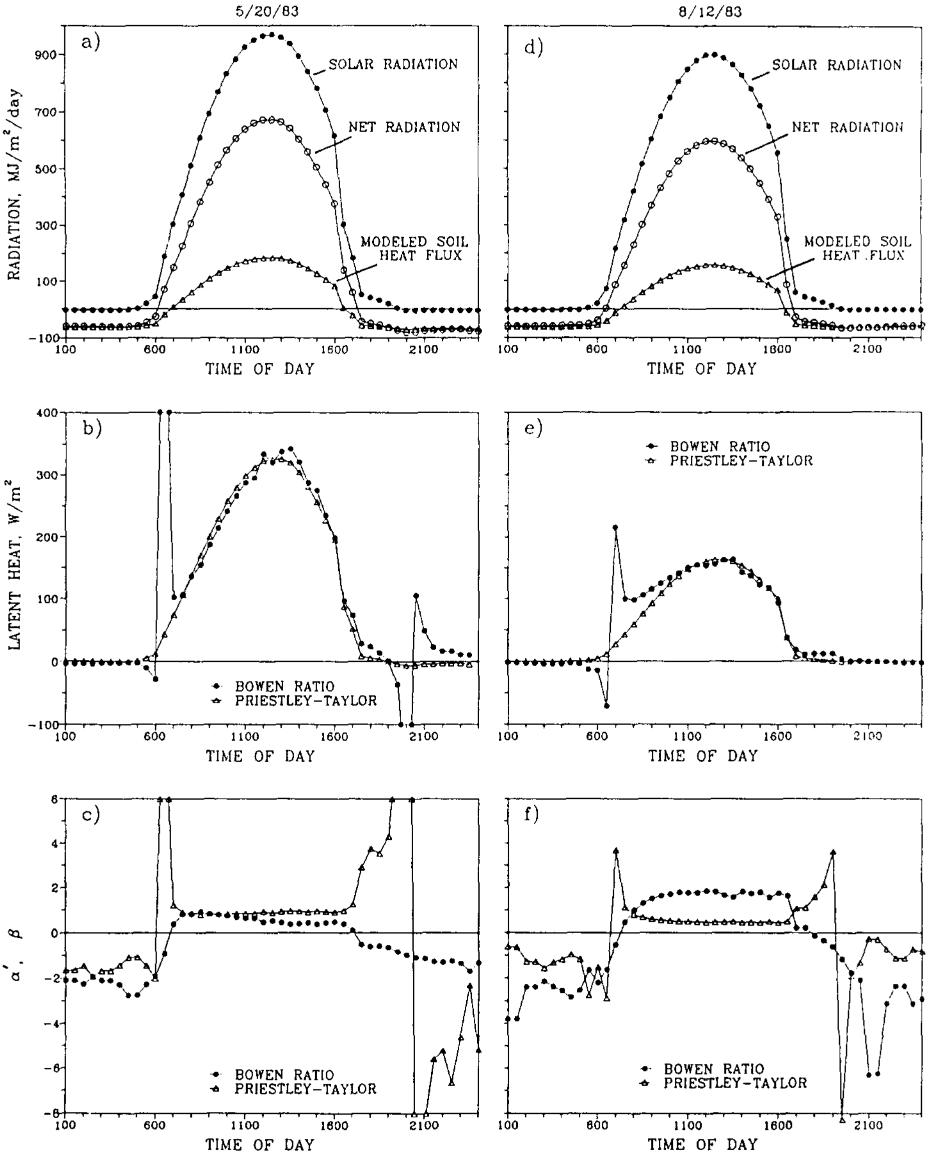

Fig. 2. Diurnal measured and calculated energy balance components for two dates: (a) and (d), total solar radiation, net all-wave radiation (R_n) and measured soil heat flux (G); (b) and e), measured (Bowen ratio) and calculated (Priestley-Taylor) evapotranspiration; (c) and f), the Bowen ratio (β) and the modified Priestley-Taylor coefficient (α').

we reduced the measured evapotranspiration value calculated from the Bowen ratio from its initial value in 5° increments until we obtained best fit (using least squares) between the Bowen ratio estimate of water content and the

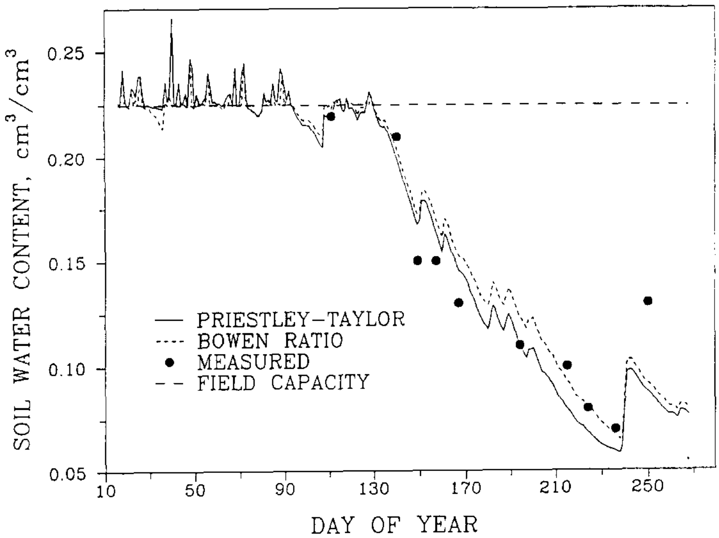

Fig. 3. Seasonal soil water contents using the modified Priestley–Taylor equation, the calculated values using the mass-balance water budget approach with the Bowen ratio, the measured data, and field capacity.

measured value from the cross-hole gamma. We found that reducing the Bowen ratio estimate of evapotranspiration by 20% would provide a best fit to our measured data (ten profiles averaged over the site, for each of 10 days measured over the season) (Fig. 3). The distribution of water use over the season is much more complex and the distribution changes dramatically over the season. Although we could incorporate other model values that provide information on the distribution (Childs et al., 1987) to provide a better fit, we believe that this kind of detail is unwarranted. This technique provides a best fit dataset that supplies enough information to determine the relationship between α' and Θ. Some of the difference between the Bowen ratio estimate of soil water content and the cross-hole gamma estimate could be due to bias or errors in the Bowen ratio measurements. As the correction for evapotranspiration coming from depths greater than 0.50 m was approximately 20%, the error in the Bowen ratio should be 20% or less. As a significant amount of vegetation (approximately 20%) had roots penetrating to depths of more than 0.50 m and contributed to some of the evapotranspiration, it appears that errors in the Bowen ratio were quite a bit less than 20%. This is fairly good considering the long-term nature of the experiment.

The values for α' were plotted against relative soil water saturation (Fig. 4). There are three different data points presented in the figure: (1) the original data points from Flint and Childs (1987a) which are calculated using only the 10 days where G and soil water content were measured; (2) values for days when incoming shortwave radiation was less than 12 MJ m^{-2} day^{-1};

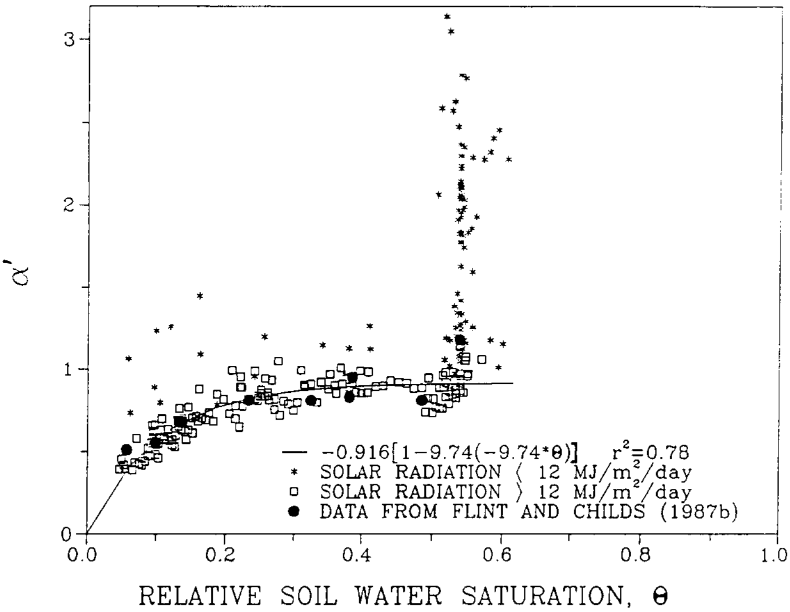

Fig. 4. Modified Priestley-Taylor coefficient α' vs. relative soil water saturation (Θ, 0–0.5 m depth). Points from measured data from Flint and Childs (1987b), solar radiation greater than 12 MJ m^{-2} day^{-1}, and solar radiation less than 12 MJ m^{-2} day^{-1}.

(3) values for days when incoming shortwave radiation was more than 12 MJ m^{-2} day^{-1}. Low solar radiation days have been separated as suggested by Black (1979); it may be inappropriate to use the modified Priestley–Taylor approach on such days because even soils with low water content can supply enough water to meet potential evapotranspiration. (The high values of α' may also be attributed to errors in the Bowen ratio if gradients (fluxes) were small, as small bias errors are amplified with small gradients. However, these values were not used in the determination of the relation between α' and Θ because of the small radiation values.)

The relation of α' to Θ is given in Fig. 4 for a profile depth of 0.50 m. Flint and Childs (1987a) evaluated the influence of using water content averaged over several soil thicknesses and found distinct differences. The calculation of α' can be determined for any depth where measured values are known. The depth of 0.50 m was chosen as it covered the average range of soil thickness at our site.

A simplified formulation of α' would be to set an upper limit of $\alpha' = 0.92$, where $\lambda E_p = \lambda E_a$. α' could be reduced when soil water content falls below some critical value of Θ where soil water supply limits evapotranspiration. (This was originally proposed by Priestley and Taylor when α was set at 1.26.) By estimating total available water capacity as the difference between field capacity ($\Theta \approx 0.60$) and driest seasonal water content ($\Theta \approx 0$), it can be seen that more than 50% of this total available water is used ($\Theta \approx 0.35$, Fig. 4), before soil water becomes limiting. This value is in general agreement with the data in Table 2. The coefficients A and B in eqn. (5) are 0.916 and -9.74 respectively, with an r^2 value of 0.78.

SUMMARY AND OBSERVATIONS

The Priestley–Taylor equation can be used to calculate actual evapotranspiration by incorporating α' as a variable parameter dependent on Θ. The relation of α' to Θ should only be used for high environmental demand days (solar radiation in excess of 12 MJ m^{-2} day^{-1}). The method also requires measurements or estimates of net radiation, soil heat flux, air temperature and atmospheric vapor density.

An argument can be made to separate the effects of soil moisture availability and environmental demand rather than to combine them in a complex α' (Spittlehouse and Black, 1984). Such a separation results in a more detailed model for evapotranspiration, which requires a more detailed description of the soil and the site. Such models (e.g. Spittlehouse and Black, 1981; Childs et al., 1987) are quite effective but not always practical for users without the resources for site-specific measurements. In such situations, the modified Priestley–Taylor equation may be advantageous, because little soils information is needed to calculate evapotranspiration.

The technique is robust in that the calculation of α' is independent of R_n and G, which are subject to measurement errors, and can be determined by periodic measurements of β. (Although the measurement of β is also subject to errors, the elimination of R_n and G is helpful in minimizing errors in α' by relating it only to β, as was done by Priestley and Taylor.) The measurements of β can come from a Bowen ratio station, eddy correlation or other flux-variance techniques that measure latent and sensible heat flux. When α' is determined from periodic measurements, then seasonal calculation can be made from simpler techniques if R_n, G, air temperature and vapor density deficits are known. If R_n is measured for the Bowen ratio or variance technique then models of R_n and G can be used to supplement G if it is not routinely measured. The application of this technique does point out the dependence of evapotranspiration modeling on R_n, perhaps more than any other measurement. Therefore, great care should be taken and a thorough understanding applied to using this or any other technique which relies on site-specific data. This technique provides total evapotranspiration. Partitioning between plants and soil and within soil layers is dependent on the application.

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