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Evapotranspiration models compared on a Sierra Nevada forest ecosystem

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Abstract

Evapotranspiration, a major component in terrestrial water balance and net primary productivity models, is difficult to measure and predict. This study compared five models of potential evapotranspiration (PET) applied to a ponderosa pine forest ecosystem at an AmeriFlux site in Northern California. The AmeriFlux sites are research forests across the United States, Canada, Brazil, and Costa Rica with instruments on towers that measure carbon, water, and energy fluxes into and out of the ecosystems. The evapotranspiration models ranged from simple temperature and solar radiation-driven equations to physically-based combination approaches and included reference surface and surface cover-dependent algorithms. For each evapotranspiration model, results were compared against mean daily latent heat from half-hourly measurements recorded on a tower above the forest canopy. All models calculate potential evapotranspiration (assuming well-watered soils at field capacity), rather than actual evapotranspiration (based on soil moisture limitations), and thus overpredicted values from the dry summer seasons of 1997 and 1998. A soil moisture function was integrated to estimate actual evapotranspiration, resulting in improved accuracy in model simulations. A modified Priestley–Taylor model performed well given its relative simplicity.

Keywords: Evapotranspiration; AmeriFlux; FLUXNET; Biosphere–atmosphere interactions; Shuttleworth–Wallace; Penman–Monteith; McNaughton–Black; Penman; Priestley–Taylor

1. Introduction

Over the entire land surface of the globe, rainfall averages around 750 mm year^{-1}, of which some two thirds is returned to the atmosphere as evapotranspiration, making evapotranspiration the largest single component of the terrestrial hydrological cycle (Baumgartner and Reichel, 1975). Carbon dioxide (CO_2)-induced greenhouse warming has accelerated the necessity to understand the hydrologic cycle and climate change (Houghton et al., 1990; IGP-BAHC, 1993; GCIP-GEWEX, 1993; Watson et al., 1995; Kaczmarek et al., 1996). Evapotranspiration and CO_2 uptake by vegetation are intrinsically coupled, leading to links and feedbacks between land surface and climate that have only begun to be explored (Hutjes et al., 1998; Savabi and Stockle, 2001). Forests can strongly influence the global hydrologic and carbon cycles and thus climate (Musselman and Fox, 1991). A limited number of canopy-scale eddy covariance studies have shown that drought stress plays a significant role in net ecosystem exchange (e.g. Baldocchi, 1997). The development of models describing climate-landscape relationships, such as water and carbon fluxes at ecosystem levels, is a necessary step in understanding and predicting the effects of changes in climate on landscape and on water resources (Kite, 1998).

Evapotranspiration model estimates and field measurements vary widely. Differences in the treatment of

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evapotranspiration are prominent among both climate and terrestrial ecosystem models (Shuttleworth, 1991; VEMAP, 1995). Evapotranspiration has always been difficult to measure, especially on an ecosystem or watershed spatial scale. Methods have been developed to measure evapotranspiration at the leaf level, the tree level, and the stand level. At the stand level, instruments mounted on a tower above the canopy are routinely used to measure humidity and wind velocities at high frequency, with water fluxes out of the forest canopy calculated by the eddy covariance method. Since the majority of moisture supplied by precipitation returns to the atmosphere as evapotranspiration, and because evapotranspiration is one of the most difficult processes to evaluate in hydrologic analysis, estimates are generally considered to be a significant source of error in streamflow simulation (Burnash, 1995). Nonetheless, effective characterization of the evapotranspiration process is critical for completing the water balance in terrestrial ecosystems, and accurately predicting the effects of global climate and land use change. A process-based understanding of evapotranspiration is needed to quantify likely changes in evapotranspiration due to climate and land surface change (Choudhury and DiGirolamo, 1998; Hutjes et al., 1998). Therefore, we find it necessary to evaluate various evapotranspiration methods employed in these modeling efforts.

The current modeling approach for estimating evapotranspiration is to calculate potential evapotranspiration (PET) using methods driven by meteorological data and/or vegetation characteristics, and to scale this estimate down to actual evapotranspiration (AET) based on limitations in available water (i.e. soil moisture) (Stannard, 1993; Federer et al., 1996; Vörösmarty et al., 1998). PET has been used to describe the evapotranspiration that would occur given an adequate water supply at all times (Linsley et al., 1958). However, the term PET is somewhat ambiguous, because the upper limit to evapotranspiration is dependent on vegetation type as well as soil water and climatic conditions (Burman and Pochop, 1994). The historical development of the PET concept has led to a variety of both PET definitions and methods (Shuttleworth, 1991; Federer et al., 1996). Following the nomenclature of Shuttleworth (1991), we examined two types of PET modeling approaches: reference-surface PET methods and surface-dependent PET methods. Reference-surface evapotranspiration is defined as evapotranspiration that would occur from a land surface specified as a ‘reference crop’ (usually a short, uniform, green plant cover such as alfalfa or grass) under designated weather conditions and soil at field capacity (also termed well-watered soil) (Shuttleworth, 1991; Federer et al., 1996). Reference-surface methods generally focus on an empirical relationship between temperature and PET, but neglect vegetation. Surface-dependent evapotranspiration is defined as evapotranspiration that would occur from a specified land surface, and the methods generally include a combination of vegetation and soil characteristics.

Several models for estimating evapotranspiration have been introduced in the literature, and our study includes many of them. Vörösmarty et al. (1998) compared nine models on all the watersheds of the continental United States using a water-balance model. Other studies compared PET models at the fetch scale for sparsely vegetated rangeland (Stannard, 1993), wildland vegetation in semiarid rangeland (Di, 1993), partial canopy/residue-covered fields (Farahani and Ahuja, 1996), maize with bare soil (Farahani and Bausch, 1995), and barley (Tourula and Heikinheimo, 1998). Federer et al. (1996) compared PET models at seven locations, but did not compare the PET estimates with measurements, because the purpose was to inform global modeling efforts on the relative variation between methods. McNaughton and Black (1973) performed energy balance measurements of evapotranspiration in a Douglas fir forest with soil at field capacity, resulting in a PET model used in our study. Few studies have analyzed evapotranspiration dynamics in forest ecosystems not only because of the general focus on agriculture, but also because of the difficulty of obtaining evapotranspiration measurements in forests.

The primary objective of our study is to compare a subset of the PET models used by Vörösmarty et al. (1998), but at a smaller spatial scale with known evapotranspiration in a forest plantation calculated from micrometeorological measurements and eddy covariance methods at our study site. The models include surface-dependent methods developed by Shuttleworth and Wallace (1985), Monteith (1965), Priestley and Taylor (1972), and McNaughton and Black (1973), and a reference-surface method by Penman (1948); the models were chosen because they are commonly used in water balance models (e.g. Willmott et al., 1985; Arnell and Reynard, 1996), terrestrial ecosystem (net primary production) models (e.g. Melillo et al., 1993; Parton et al., 1993; Running and Hunt, 1993), hydroinformatics (Naoum and Tsanis, 2003), and in remote sensing (Li and Lyons, 2002). While scale and water-limitations at our site are important to this study, the critical factor driving our study is the use of ecosystem scale flux data. The flux measurements at the Blodgett Forest site, as part of AmeriFlux and the larger FLUXNET network of towers across the world, measure water, carbon, and energy fluxes as well as meteorological variables above the forest ecosystem canopy (e.g. Goldstein et al., 2000). Although the first modeling and analysis of forest evapotranspiration was done in the 1970s (e.g. Spittlehouse and Black, 1979), the flux data from the tower are novel and only recently have researchers used such data to drive larger-scale ecosystem models.

We assess the PET models at this stand-scale using data from a tower at the Blodgett Forest Research Station in California. We analyzed data acquired continuously over two climatically different growing seasons: 1997 was drier than the climatic mean, and 1998 was cooler and wetter than the climatic mean (influenced by El Niño). Evapotranspiration rates (measured evapotranspiration) were derived from eddy covariance measurements, and environmental variables were measured that are known to influence evapotranspiration rates, such as net radiation, wind speed, air temperature, relative humidity, and soil moisture.

2. Materials and methods

2.1. Site description

Data were gathered during the summers of 1997 and 1998 in the Sierra Nevada mountains at Blodgett Forest Research Station (38°53′42.9″N, 120°37′57.9″W, 1315 m), a research forest of the University of California, Berkeley (Goldstein et al., 2000). The forest was planted in 1990 and was dominated by ponderosa pine trees (*Pinus ponderosa* Doug. E. Laws), the most common conifer species in North America. The canopy also included individuals of Douglas fir (*Pseudotsuga menziesii*), white fir (*Abies concolor*), giant sequoia (*Sequoiadendron giganteum*), incense-cedar (*Calocedrus decurrens*) and California black oak (*Quercus kelloggii*). The major understory shrubs were manzanita (*Arctostaphylos* spp.) and *Ceonothus* spp. (Xu et al., 2001). In 1997, about 25% of the ground area was covered by shrubs, 30% by conifer trees, 2% by deciduous trees, 7% by forbs, 3% by grass and 3% by stumps. The forest area was in a stage of rapid growth, as exhibited by the 10% increase in leaf area index (LAI) between the 1997 (2.9–4.2) and 1998 (3.2–4.5) growing seasons. The site is characterized by a Mediterranean climate with an average annual precipitation of 163 cm (180 cm in 1997 and 117 cm in 1998), the majority of which falls between September and May, and almost no rain in the summer. The soil is a fine-loamy, mixed, mesic, ultic haploxeralf in the Cohasset series whose parent material was andesitic lahar (Goldstein et al., 2000).

2.2. Measurements

Infrastructure for the canopy scale flux measurements consisted of a 10 m measurement tower (Upright Inc.). From 1 June to 10 September, 1997 and from 1 May to 30 October, 1998, fluxes of CO_2, H_2O, and sensible heat were measured by the eddy covariance method. Environmental parameters such as wind speed and direction, air temperature and humidity, net and photosynthetically active radiation, soil temperature, soil moisture, soil heat flux, rain, and atmospheric pressure were also monitored. A system to measure the vertical profiles of CO_2 and H_2O was added in 1998. The data acquisition system was separated in two parts: (1) a fast response system which monitored data at high frequency (up to 10 Hz) used to calculate eddy covariance, with raw data stored in 30 min data sets; and (2) a slow response system which monitored environmental parameters and stored 30 min averaged data (Goldstein et al., 2000).

Wind velocity and temperature were measured at 10 Hz with a three-dimensional sonic anemometer (ATI Electronics Inc., Boulder, CO) mounted 5 m above the canopy to obtain an accurate reading of air above the canopy. The height will vary in other ecosystems depending on the vegetation characteristics, but the purpose is to measure representative wind characteristics. CO_2 and H_2O mixing ratios were measured with an infrared gas analyzer (IRGA, LICOR model 6262, Lincoln, NE). Fluxes of CO_2, H_2O, and sensible heat between the forest and the atmosphere were determined by the eddy covariance method (Goldstein et al., 2000). This method quantifies vertical fluxes of scalars between the forest and the atmosphere from the covariance between vertical wind velocity and scalar fluctuations averaged over 30-min periods (e.g. Shuttleworth et al., 1984; Baldocchi et al., 1988; Wofsy et al., 1993; Moncrieff et al., 1996). Environmental parameters were recorded on a CR10X datalogger (Campbell Scientific Inc., Logan, UT). Soil moisture probes were buried horizontally at 10 and 20 cm depth to record an upper soil moisture profile for the relatively flat and homogenous site. Brandes (1998) has also shown, through principal components analysis, that lateral spatial variability contributes only a small portion (<10%) of the total variance of a soil moisture data set, so our point measurements can be extrapolated laterally to the rest of the site. We used the soil moisture measurements at 20 cm because the plantation trees were young and the deepest roots had not extended past roughly a meter, though we acknowledge that moisture content can change throughout the soil profile. The 20 cm measurements represent our best available data at estimating moisture in the rooting zone, and the upper soil layers do not change drastically in composition, though the bedrock depth has not yet been measured. The soil moisture measurements at 10 cm were sensitive to variability in surface conditions. Total (all-sided) LAI was estimated using two techniques that resulted in similar estimates, (1) the LI-2000 (Li-Cor Inc., Lincoln, NE), which calculates LAI from diffuse sunlight measurements made with a 'fish eye' (148° conical field-of-view) optical sensor as fully described in Welles and Norman (1991), and (2) an allometric method that scaled up from leaf-level determination using the measured geometry of trees (Xu et al., 2001).

We recognize systematic errors associated with the eddy covariance method. Travel through the sampling tube and instrument response time from sensor separation between wind and scalar measurements cause damping of high frequency fluctuations by the closed-path IRGA and time lags between wind and scalar data (Rissman and Tetzlaff, 1994). Further, the sonic anemometer is unreliable in resolving fine-scale eddies in light winds (Goulden et al., 1996; Moncrieff et al., 1996). Generally, these types of errors result in the underestimation of flux (Leuning and King, 1992). The inability of the sonic anemometer to resolve the vertical wind occurs mainly at night as the fluctuations become dominated by small, high frequency eddies (Goulden et al. (1996) use $u^* < 0.17$ m s^{-1} as the threshold for reliable measurements). The inability of the sonic anemometer to resolve fine-scale eddies in light winds (e.g. during night) produced systematic errors in the sensible heat flux to correct the CO_2 and H_2O fluxes. Thus, although daytime turbulence was strong enough to produce reliable measurements, the calmer conditions during night rendered the nighttime flux measurements less reliable (Goldstein et al., 2000). For the purpose of this study, we corrected for outliers (greater than three standard deviations from the mean) and missing data points (via interpolation or backup sensors), and evaluated the evapotranspiration models using daytime (05:00–21:00 h) averages because nighttime measurements were unreliable. Sample size in 1997 was 87 daytime averages based on 3900 measurements, and sample size in 1998 was 149 daytime averages based on 8700 measurements.

2.3. Evapotranspiration models

Five PET models of increasing complexity were tested under two classes of land surface speciation (Shuttleworth, 1991; Federer et al., 1996). Reference-surface evapotranspiration is defined as evaporation that would result from a specific land surface, referred to as a 'reference crop' (Vörösmarty et al., 1998). Surface-dependent evapotranspiration is defined as the evaporation that would occur from any of a variety of designated land surfaces. A summary of the parameters and units used in each method is presented in Table 1.

The Priestley–Taylor model (Priestley and Taylor, 1972), the simplest of the five, is defined as:

$$\lambda E = \alpha \Delta A / (\Delta + \gamma)$$

where λE is total potential evapotranspiration (in flux units of W m^{-2}, for example), Δ is the derivative of saturated vapor pressure versus temperature, A is total available energy (net radiation minus soil heat flux), and γ is the psychrometric constant. Δ and γ are functions of air temperature (all models analyzed here are thus functions of air temperature). Priestley and Taylor (1972) determined an average value of 1.26 for α based on measurements of evapotranspiration from a variety of well-watered vegetated and water surfaces (i.e. PET). To estimate AET, α has been redefined to be a function of soil moisture (Flint and Childs, 1991).

The McNaughton–Black model (McNaughton and Black, 1973) is defined as follows:

$$\lambda E = c_p \rho D / \gamma r_{cs}$$

where c_p is specific heat at constant pressure, ρ is air density, D is vapor pressure deficit, and r_{cs} is bulk stomatal resistance of the canopy.

The Penman model (Penman, 1948) was the first effort to combine both energy and atmospheric vapor transport components to estimate PET and is defined as follows:

$$\lambda E = (\Delta A + 73.64 \rho \gamma (1 + 0.54 u) D) / (\Delta + \gamma)$$

where u is wind speed, and 73.64 is 2.6 times the latent heat of vaporization (units converted).

Table 1
Comparison of the increasing complexity of the models in terms of number of parameters required

Parameter	Symbol	Units	PT	MB	Penman	PM	SW
Rate of change of vapor pressure with temperature	Δ	kPa K^{-1}	✓		✓	✓	✓
Total available energy	A	W m^{-2}	✓		✓	✓	✓
Psychrometric constant	γ	kPa K^{-1}	✓	✓	✓	✓	✓
Air temperature	T_a	°C	✓	✓	✓	✓	✓
Specific heat at constant pressure	c_p	J kg^{-1} K^{-1}		✓		✓	✓
Air density	ρ	kg m^{-3}		✓	✓	✓	✓
Vapor pressure deficit	D	kPa		✓	✓	✓	✓
Bulk stomatal resistance of the canopy	r_{cs}	s m^{-1}		✓		✓	✓
Wind speed	u	m s^{-1}			✓	✓	✓
Aerodynamic resistance above the canopy	r_{aa}	s m^{-1}				✓	✓
Bulk boundary layer resistance of the vegetation	r_{ca}	s m^{-1}					✓
Aerodynamic resistance for substrate and canopy	r_{sa}	s m^{-1}					✓
Surface resistance of the substrate	r_{ss}	s m^{-1}					✓
Available soil energy	A_s	W m^{-2}					✓

PT, Priestley–Taylor; MB, McNaughton–Black; PM, Penman–Monteith; SW, Shuttleworth–Wallace.

The Penman–Monteith model (Monteith, 1965) expanded upon the Penman model:

$$\lambda E = (\Delta A + c_p \rho D / r_{aa}) / (\Delta + \gamma + \gamma (r_{cs}/r_{aa}))$$

where r_{aa} is the aerodynamic resistance above the canopy, and r_{cs} is stomatal resistance of the canopy.

For the Shuttleworth–Wallace model (Shuttleworth and Wallace, 1985), λE is separated into evaporation from the soil (λE_s) and transpiration from the canopy (λE_c), which are derived from the Penman–Monteith combination equations:

$$\lambda E_s = (\Delta A_s + \rho c_p D_0 / r_{sa}) / (\Delta + \gamma (1 + r_{ss}/r_{sa}))$$

$$\lambda E_c = (\Delta (A - A_s) + \rho c_p D_0 / r_{ca}) / (\Delta + \gamma (1 + r_{cs}/r_{ca}))$$

where A_s is available soil energy, and D_0 is vapor pressure deficit in the canopy; r_{sa} is the aerodynamic resistance between the substrate and canopy source height, r_{ca} is the boundary layer resistance of the vegetation, and r_{ss} is soil resistance. The aerodynamic resistance above the canopy (r_{aa}) and the aerodynamic resistance between the substrate and canopy source height (r_{sa}) are functions of leaf area index, eddy diffusivity decay constant, roughness length of the vegetation (function of vegetation height), zero plane displacement (function of vegetation height), a reference height above the canopy where meteorological measurements are available, wind speed, von Karman's constant, and roughness length of the substrate. D_0 is derived from the Ohm's law electrical analog for the vapor pressure and temperature difference between the canopy and the reference height above the canopy where fluxes out of the vegetation are measured. D_0 is a function of the measurable vapor pressure deficit at the reference height, D:

$$D_0 = D + (\Delta A - r_{aa} \lambda E_c (\Delta + \gamma)) / \rho c_p$$

and D can thus be substituted for D_0 into the combination equations. The total evaporation from the crop, λE, for the Shuttleworth–Wallace model is the sum of the Penman–Monteith combination equations with D substituted in for D_0:

$$\lambda E = C_c PM_c + C_s PM_s$$

where $\mathrm{PM_c}$ describes evaporation from the closed canopy, and $\mathrm{PM_s}$ describes evaporation from the bare substrate. The new Penman–Monteith equations have the form:

$$\mathrm{PM_c} = \frac{\left(\Delta A + (\rho c_p D - \Delta r_{ca} A_s)/(r_{aa} + r_{ca})\right)}{(\Delta + \gamma (1 + r_{cs}/(r_{aa} + r_{ca})))}$$

$$\mathrm{PM_s} = \frac{\left(\Delta A + (\rho c_p D - \Delta r_{sa} (A - A_s))/(r_{aa} + r_{sa})\right)}{(\Delta + \gamma (1 + r_{ss}/(r_{aa} + r_{sa})))}$$

The coefficients C_c and C_s are resistance combination equations:

$$C_c = 1/(1 + R_c R_a / (R_s (R_c + R_a)))$$

$$C_s = 1/(1 + R_s R_a / (R_p (R_s + R_a)))$$

where

$$R_a = (\Delta + \gamma) r_{aa}$$

$$R_s = (\Delta + \gamma) r_{sa} + \gamma r_{ss}$$

$$R_c = (\Delta + \gamma) r_{ca} + \gamma r_{cs}$$

The evapotranspiration models described above calculate potential evapotranspiration rather than actual evapotranspiration. Potential evapotranspiration is defined as the evapotranspiration flux from the ecosystem under well-watered soil conditions (i.e. soil at or close to, field capacity). We derived actual evapotranspiration from potential evapotranspiration using a simple soil moisture function, $f(\phi)$ (Saxton et al., 1986):

$$\lambda E_{\mathrm{actual}} = f(\phi) * \lambda E$$

where $\lambda E_{\mathrm{actual}}$ is the actual evapotranspiration and the soil moisture function is a dimensionless variable estimated by a simple linear model:

$$f(\phi) = M/\text{Field capacity}$$

where M is soil volumetric moisture at 20 cm depth (at rooting zone). Field capacity (for which soil water potential is at -10 kPa) for the fine-loamy soil texture at our site was determined as 39% based on Saxton et al. (1986) and compared well to maximum soil moisture observed after rain events. Field capacity can be defined as the percentage of water remaining in a soil two or three days after its having been saturated and after free drainage has practically ceased. Brandes and Wilcox (2000) have shown that simple linear models of the evapotranspiration/soil moisture process are appropriate for hydrologic modeling. Soil moisture models have been developed with increasing complexity to better represent soil physics, such as the module in WATFLOOD that includes permanent wilting point, saturation, and a root fraction to simulate non-linear features of moisture extraction by vegetation from soil (Soulis et al., 2000); Munro et al. (1998) solve the Richards' equation and the temperature diffusion equation for multi-soil layers. However, we chose the simple, physically-based soil moisture function for three reasons: (1) the data requirements and modeling demands are greatly increased for more complex soil moisture functions; (2) these PET methods are often used in larger spatial scale models where detailed soil physics may not be possible to calculate accurately; and (3) our main goal is to assess the relative accuracy of the PET methods, rather than evaluate the merit of various approaches to modeling soil physics.

3. Results

3.1. Potential versus measured evapotranspiration

For all potential evapotranspiration models, simulated PET compared reasonably well with measured evapotranspiration at the beginning of the summer season (April–May). However, as the soil moisture decreased through the summer, all models tended to overpredict evapotranspiration because the PET models were designed for well-watered soil conditions rather than natural summertime Mediterranean drought conditions (Fig. 1). 1997 was drier than 1998, and greater evapotranspiration was observed in 1998. The summer of 1997 was also substantially windier than 1998—this fact influenced the wind-sensitive Penman method, which predicted unrealistically high amounts of evapotranspiration due to the fast winds. In fact, the Penman model predicts an increase of well over 100 W m^{-2} in evapotranspiration for every 0.5 m s^{-1} increase in wind speed, holding all other variables constant. The specific increase depends on the values of the fixed variables; VPD in particular effects that increase.

PET results from the Shuttleworth–Wallace, Penman–Monteith, and McNaughton–Black models had similar trends and magnitudes; McNaughton–Black tended to give the highest estimates followed by Penman–Monteith and Shuttleworth–Wallace, respectively. Penman–Monteith approximated Shuttleworth–Wallace in the dry season of 1997. The Priestley–Taylor model nearly approximated the measured evapotranspiration in both

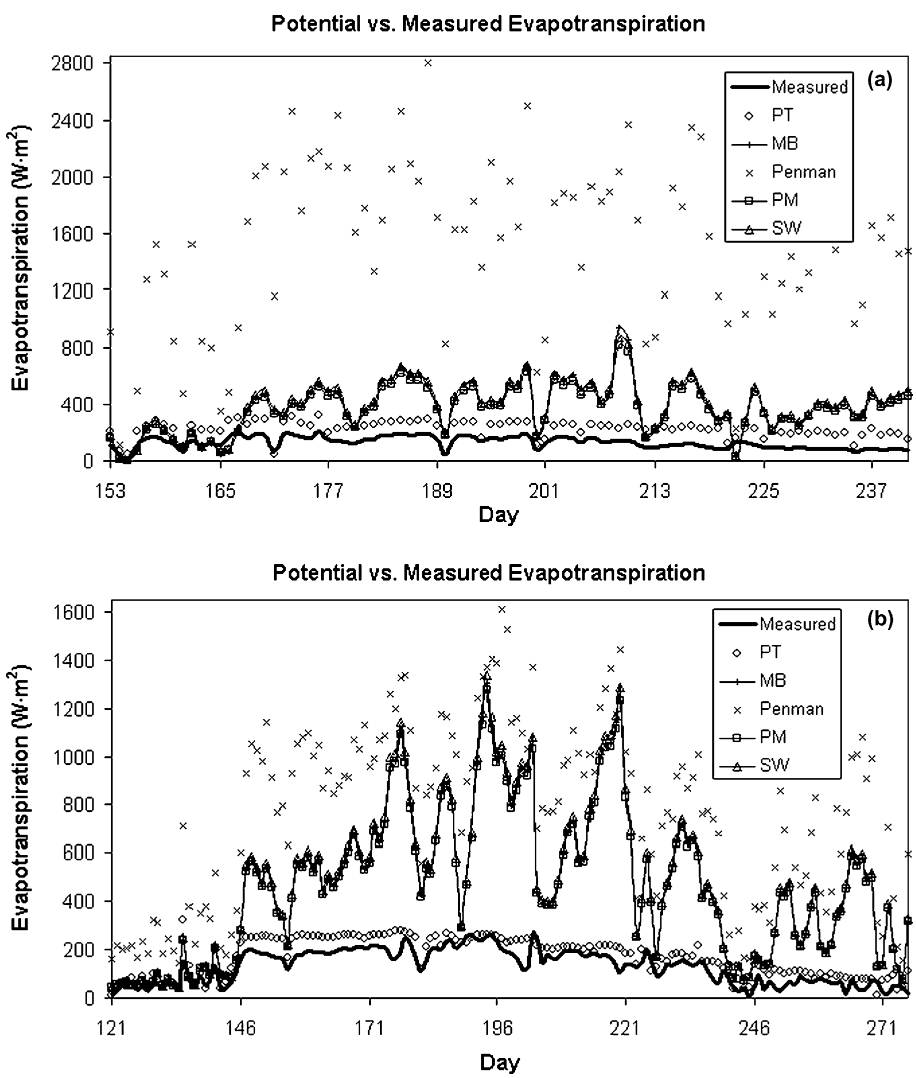


Fig. 1. Potential evapotranspiration without soil moisture function for (a) 1997 and (b) 1998. Shuttleworth–Wallace (SW), Penman–Monteith (PM), and McNaughton–Black (MB) all have similar trends and magnitudes, and Priestley–Taylor (PT) comes closest to measured evapotranspiration (Measured ET) in both years with $\alpha = 1.26$. The Penman method was highly sensitive to wind speed in 1997.

years, especially in the higher soil moisture year of 1998 when water was not as limiting.

3.2. Actual versus measured evapotranspiration

Actual evapotranspiration was derived from potential evapotranspiration for each model by applying the soil moisture function. The actual evapotranspiration estimations provided good approximations of measured evapotranspiration. With the soil moisture function, Shuttleworth–Wallace ($r^2 = 0.46$ in 1997; $r^2 = 0.69$ in 1998), Penman–Monteith ($r^2 = 0.43$ and 0.66), and McNaughton–Black ($r^2 = 0.37$ and 0.62) all performed well with similar trends and magnitudes. Penman–Monteith and McNaughton–Black approximated Shuttleworth–Wallace through both seasons, though McNaughton–Black began to more severely overpredict AET than the other models did late in both seasons. Priestley–Taylor ($r^2 = 0.73$ and 0.58) significantly underpredicted the measured evapotranspiration with the soil moisture function. Although the Penman simulations were improved, the Penman model still significantly overpredicted measured evapotranspiration. Flint and Childs (1991) state that the assumptions and simplifications used by the Penman model to model the aerodynamic components of evapotranspiration make the Penman model useful only for calculating potential evapotranspiration. The primary modified version of the Penman model, the Penman–Monteith model, allows for calculation of actual evapotranspiration given values for resistances. The soil moisture function performed better across the relatively wet season of 1998 than the dry season of 1997; the models tended to underpredict measured evapotranspiration in 1997 (Figs. 2 and 3). In addition, there was smaller scatter, but more sample points, in 1998.

3.3. Modified α for Priestley–Taylor model

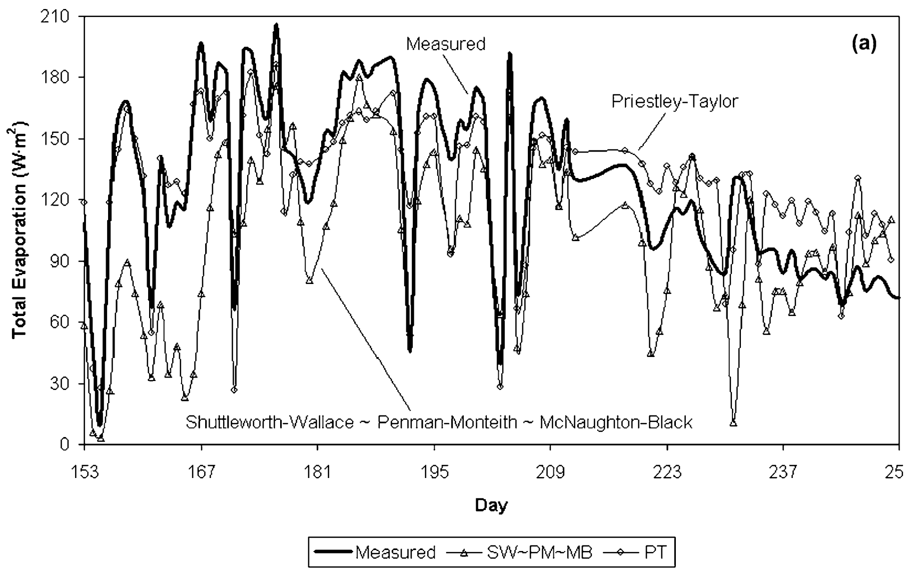
Whether or not use of a soil moisture function is appropriate for the Priestley–Taylor model has been answered, in part, by numerous studies on α. It has become increasingly common to redefine α based on soil moisture, rather than add on a soil moisture function to the original α value of 1.26. Priestley and Taylor originally proposed that α be reduced when soil water content falls below some critical soil moisture value where soil water supply limits evapotranspiration. Spittlehouse and Black (1981) state that the Priestley–Taylor model is in error and no longer appropriate to use if α is fixed. Based on the regression analysis prescribed by Flint and Childs (1991), our redefined α is:

$$\alpha = 0.84M + 0.72$$

where M is soil volumetric moisture of the top 20 cm (rooting zone). The average value for α across both years was 0.87, although that value was slightly lower for the drier 1997 and higher for the wetter 1998. It should be noted that although this equation follows the work of Flint and Childs (1991), the parameters for both this equation and for Flint and Childs' redefined α are not physically based and should be looked at critically in comparisons with other models. However, our newly calculated values for α approximate the actual measured values for α at similar sites, as tabled by Flint and Childs (1991); we append our value in Table 2. Determination of α based on either the regressed soil moisture function or measurements done at similar sites resulted in a greatly improved Priestley–Taylor model ($r^2 = 0.74$ and 0.85). With the new α value, Priestley–Taylor AET estimates were not significantly different from measured evapotranspiration across both years. Again, the Priestley–Taylor model performed well despite its relative simplicity.

4. Discussion

The Shuttleworth–Wallace, Penman–Monteith, and McNaughton–Black models resulted in similar simulations due to the common theoretical basis of their equations—the Penman model. McNaughton–Black, which excludes the radiation budget and any effect from the soil, is a simplification of Penman–Monteith, whereas Shuttleworth–Wallace adds a soil layer to the Penman–Monteith model. The simulations revealed that Penman–Monteith tended to give an intermediate result between these three models. Shuttleworth–Wallace is specifically designed for sparse crops where vegetation is not densely distributed and the soil surface may contribute significantly to total evapotranspiration, which is representative of the Blodgett site. Nonetheless, the substrate does not significantly contribute to total evapotranspiration because of low soil moisture, particularly in 1997. Thus, the Shuttleworth–Wallace model reduced back to the Penman–Monteith model and gave only slightly better results. In the relatively wet season of 1998, Shuttleworth–Wallace resulted in a more accurate simulation than in 1997 because the increased soil moisture lead to greater soil evaporation. Still, the soil evaporation was not a significant factor at this site and thus the McNaughton–Black model, which neglects the soil as an evaporation source, yielded similar results. Shuttleworth–Wallace has performed well in the literature as well (e.g. Di, 1993; Farahani and Bausch, 1995; Vörösmarty et al., 1998; Iritz et al., 1999), but the main drawback is the difficulty and extensiveness of the parameter estimation (Farahani and Ahuja, 1996). Because our study area is intensively measured as a research site, the parameters for Shuttleworth–Wallace were available for this analysis; other sites or large scale modeling efforts may not be so fortunate.



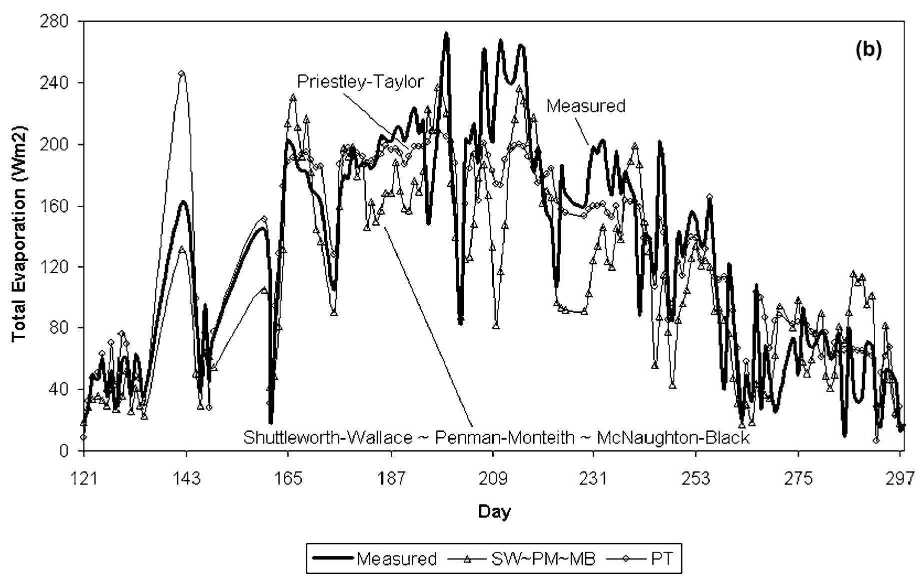


Fig. 2. Simulated versus measured (actual) evapotranspiration during (a) 1997 and (b) 1998. The soil moisture function is integrated in Shuttleworth–Wallace (SW), Penman–Monteith (PM), and McNaughton–Black (MB). Penman–Monteith and McNaughton–Black excluded due to similarity to Shuttleworth–Wallace; while PM approximates SW through both seasons, MB begins to diverge with overprediction late in both seasons as soil moisture decreases. The Priestley–Taylor (PT) graph is shown with $\alpha = 0.73$ in 1997 and 0.94 in 1998. The soil moisture function brought the simulations down to good approximations of measured evapotranspiration (Measured ET). Penman method excluded due to continued overprediction.

The upper bound to potential evapotranspiration should be the net radiation (under relatively stable conditions). According to energy balance models, incoming net radiation is partitioned into latent heat, sensible heat, and heat absorbed by the ground—therefore, latent heat, as a fraction of net radiation, should not exceed net radiation. The McNaughton–Black model is not a function of net radation and depends heavily on the accuracy of its other input parameters. An erroneously low stomatal resistance of the canopy (r_{cs}), for instance, can cause the McNaughton–Black model and other PET models that are function of r_{cs} to overpredict evapotranspiration.

For Shuttleworth–Wallace, Penman–Monteith, and McNaughton–Black, we used a constant r_{cs} throughout both seasons derived from the available minimum and maximum values measured at the site. The models are highly sensitive to r_{cs} and simulated evapotranspiration differed by as much as 26% at the minimum r_{cs} and 20% at the maximum r_{cs}. We evaluated and propagated error in r_{cs}, along with aerodynamic resistance above the canopy (r_{aa}), bulk boundary layer resistance of the vegetation (r_{ca}), aerodynamic resistance for the substrate and canopy (r_{sa}), and surface resistance of the substrate (r_{ss}) in the Shuttleworth–Wallace model via Gaussian error propagation (the final uncertainty for total

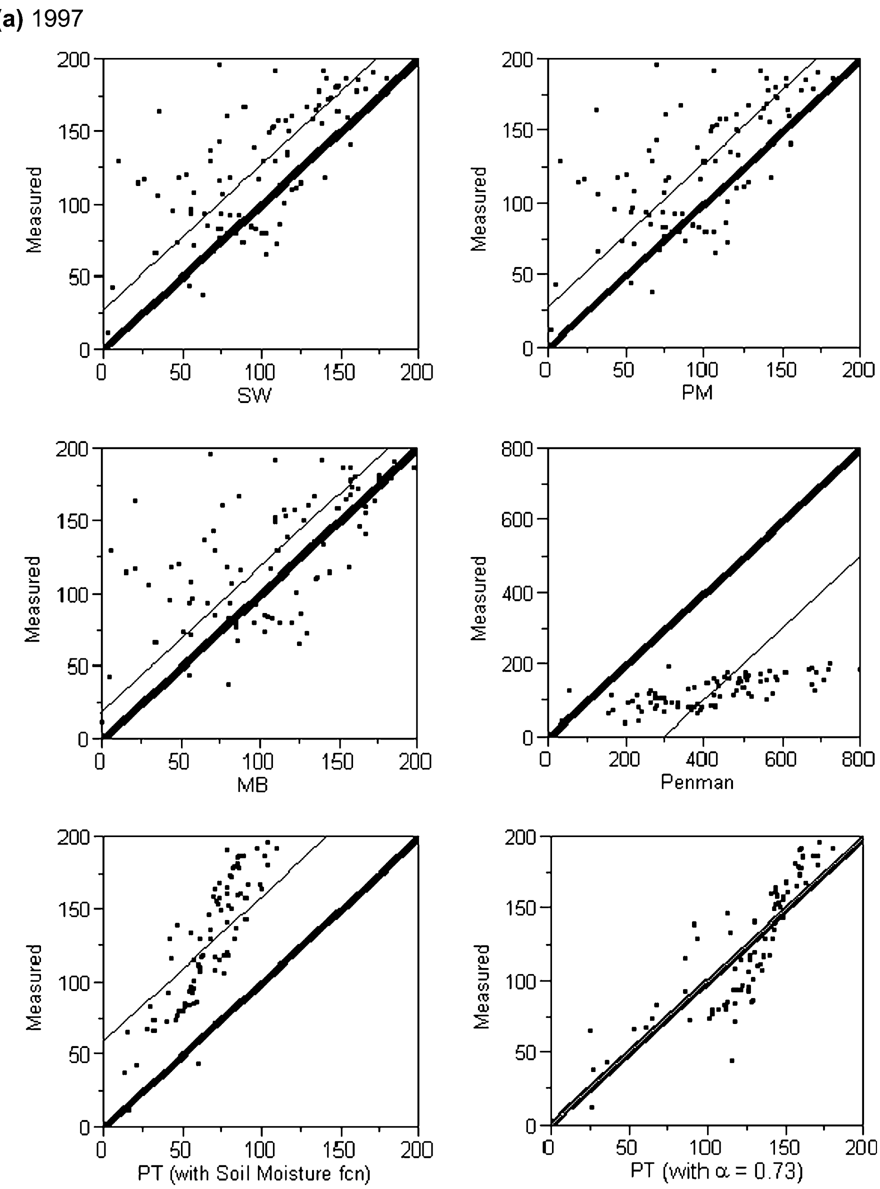


Fig. 3. Simulated versus measured evapotranspiration during (a) 1997 and (b) 1998. The soil moisture function is integrated in Shuttleworth–Wallace (SW), Penman–Monteith (PM), McNaughton–Black (MB), Penman, and Priestley–Taylor (PT with Soil Moisture function). The second Priestley–Taylor (PT) graph includes the modified α. The thick diagonal line is the 1:1 line, and the thin diagonal line is the actual difference between the means. Points below the 1:1 line represent over-prediction, and points above the 1:1 line represent under-prediction. Sample size in 1997 was 87 daytime averages based on 3900 measurements, and in 1998 was 149 daytime averages based on 8700 measurements.

evapotranspiration is equal to the square root sum of squares of the partial derivative of total uncertainty with respect to each resistance multiplied by the standard deviation of each resistance, respectively):

$$S_{\lambda E} \approx \Big[((\partial\lambda E/\partial r_{ca})(Sr_{ca}))^2 + ((\partial\lambda E/\partial r_{cs})(Sr_{cs}))^2 + ((\partial\lambda E/\partial r_{sa})(Sr_{sa}))^2 + ((\partial\lambda E/\partial r_{ss})(Sr_{ss}))^2 + ((\partial\lambda E/\partial r_{aa})(Sr_{aa}))^2\Big]^{0.5}$$

The model is nonlinear, and, for the purposes of the uncertainty analysis, we treated the variables as uncorrelated with one another. Gaussian error propagation is suitable here since it allows for examination of total uncertainty derived from the uncertainties in the parameters of the model. We found that uncertainty in r_{cs} contributes to 53% of the total uncertainty in the Shuttleworth–Wallace model. Based on this finding, much of the overprediction among the PET models may be attributed to uncertainty in r_{cs}. Again, a relatively more complex evapotranspiration model may not be more accurate than a simpler model because the complex model is more difficult to parameterize and results are vulnerable to large error resulting from propagating uncertainty in parameter values.

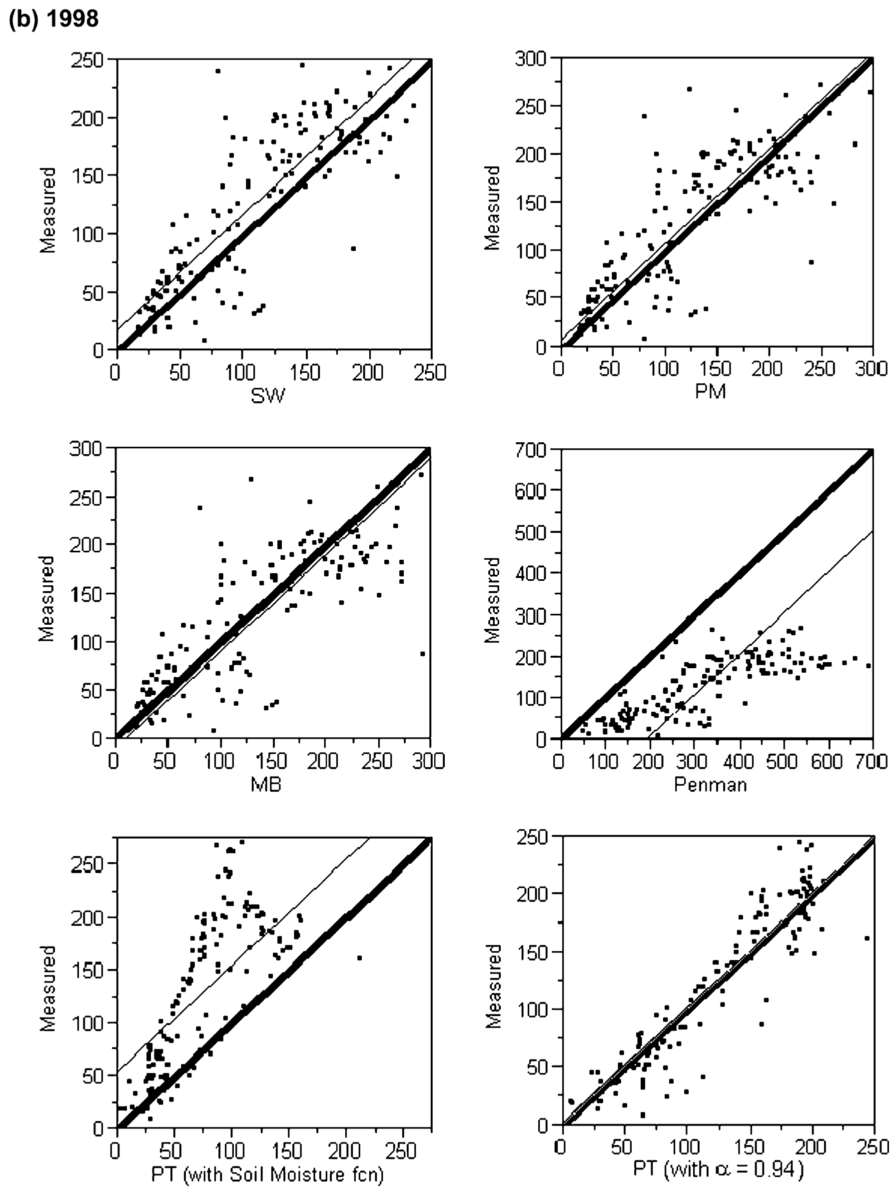


Fig. 3 (*continued*)

The success of Priestley–Taylor, given its relative simplicity, shows that this simple model may be preferable to the complex models by Shuttleworth–Wallace, Penman–Monteith, and McNaughton–Black for a partially-closed canopy under water-limitation. We add to the tabled values of the Priestley–Taylor coefficient, α, for different surface conditions. The data in Table 2 have the potential for playing a crucial role in the integration of the Priestley–Taylor method for estimating AET into larger scale ecosystem models. If most surface conditions have an associated α value under varying water stress conditions, or include α as a function of soil moisture, then it may be possible that ecosystem models could use those values with the method to accurately assess AET given relatively few input parameters.

The soil moisture function plays a key role in deriving actual evapotranspiration from potential evapotranspiration. PET assumes that soil water availability is not the limiting factor for AET (soil is at field capacity) and will thus overpredict evapotranspiration under drier soil conditions. In both years, the PET models performed well at the beginning of the summer season when soil moisture was still high from spring rainfall and residual moisture from snowmelt. But, as soil moisture declined throughout the summer, PET subsequently overpredicted measured evapotranspiration. It is crucial that our study included summertime drought conditions in a Mediterranean environment: it is under these, and similar, conditions that the assumptions of a given PET model can lead to inaccurate results. Analysis of PET models under environmental conditions that they were

Table 2
Measured values of the Priestley–Taylor coefficient, α, as tabled by Flint and Childs (1991) and appended with their and our values

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
0.90	Mixed reforestation (water limited)	Flint and Childs, 1991
0.87	Ponderosa pine (water limited, daytime)	This study
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

not initially designed for may be inappropriate. Many ecological models use PET functions on continental and global scales, and are subject to the same overprediction of actual evapotranspiration because of such assumptions (e.g. Running and Coughlan, 1988; Raich et al., 1991; Potter et al., 1993; Sellers et al., 1996; Thornton et al., 1997). Additionally, hydrologic model users at the local catchment scale must be able to handle soil moisture information, or the lack thereof (Wooldridge et al., 2003).

Thus, the relationship between potential and actual evapotranspiration must be addressed in these and future models. Furthermore, these large-scale models would benefit from a simple, but accurate, evapotranspiration component that would be relatively easy to parameterize. Within widely used general circulation models (GCMs) and numerical weather prediction models (NWPs), hydrological components such as evapotranspiration, precipitation and runoff have yet to be estimated with great accuracy. Models of the atmosphere, such as GCMs and NWPs, can be used to predict the impact of changes in the composition of the atmosphere using gas properties and the fundamental equations of energy transfer, mass conservation and atmospheric motion into changes in wind, temperature and moisture content. Despite the rigor of the atmospheric descriptions in such models, the atmosphere/land surface interaction is usually modeled quite crudely (Kite, 1998). Most GCMs (e.g. Dickinson and Kennedy, 1991; Raich et al., 1991; Sellers et al., 1996) use only a few hydrological parameters lumped over large grid squares, and contain no run-on/runoff transfer between grid squares. Such models generally overestimate precipitation and, as the vertical water balances in each grid square are independent of surrounding squares, they generally underestimate evapotranspiration, leading to an overestimate of water available for runoff (Kite, 1998). Information on land surface evapotranspiration is very important in the understanding of climate change. For example, the reduction in evapotranspiration (and the change in surface energy balance) associated with the removal of vegetation in the Sahel has been shown, via GCMs, to produce a reduction in rainfall (e.g. Charney, 1975; Cunnington and Rowntree, 1986). Therefore, there is a great need for process-based evapotranspiration models than can characterize different vegetation to allow us to better understand and predict any links between land-use change and climate change (Wallace, 1995).

We suggest that the Priestley–Taylor method may be most applicable to models run at large spatial scales because it is easier to parameterize than the widely used Penman–Monteith method, although further research is needed to confirm this suggestion. Furthermore, the models that we analyze here are atmosphere driven evapotranspiration models, which are appropriate for global or regional models since soil moisture information may not be as reliably extrapolated from point measurements as are atmospheric data. The difficulty in extrapolation of point source soil moisture data versus atmospheric data is due to heterogeneity of the soil medium relative to atmospheric mixing and well-understood physical relationships that govern the change in meteorological parameters (e.g. temperature, pressure, relative humidity). Unreliable soil moisture information will introduce a propagation of error in a soil water driven atmosphere exchange model. Nonetheless, as new data sources become increasingly available, especially with remote sensing, these models can and should be modified to allow for integration of the new data. As remotely sensed soil information develops, so too should the models that this information feeds into (Njoku et al., 2003).

Factors not taken into account that may affect the relationship between simulated and actual evapotranspiration include vegetative quality and other environmental variables. For example, ozone deposition, grazing of insects on leaves, the influence of animals such as cows on the environment, and disease are not taken into account when modeling evapotranspiration. Aside from systematic errors associated with the eddy covariance method, possible bias in the data and models include assumed values for three Shuttleworth–Wallace variables—surface resistance of the substrate, roughness length of bare substrate, and extinction coefficient of the crop for net radiation. We halved, doubled, and multiplied each parameter by a factor of 10 to test for sensitivity; nonetheless, Shuttleworth–Wallace is not highly sensitive to these parameters. Simulated evapotranspiration differed by less than 5% given the changes in these parameters. A major environmental phenomenon influencing the data was the occurrence of an El Niño event before the summer of 1998 that caused the vegetation to grow significantly in 1998; the heterogeneity across seasons allowed for ideal comparisons of the same site under different environmental conditions to see how robust the evapotranspiration models were.

5. Conclusions

Shuttleworth–Wallace, Penman–Monteith, McNaughton–Black, Priestley–Taylor, and Penman models for estimating evapotranspiration were compared using data from AmeriFlux tower measurements at a ponderosa pine ecosystem. Vörösmarty et al. (1998), in comparing these models on a global scale, found that the Shuttleworth–Wallace method performed best. In our study, Shuttleworth–Wallace, Penman–Monteith, and McNaughton–Black all yielded similar results, although Shuttleworth–Wallace performed slightly better than Penman–Monteith and McNaughton–Black; this similarity was because these models are derived from the Penman model, and because of the insignificant effect of the substrate on evapotranspiration at our site. Priestley–Taylor, with an appropriately defined α value, performed remarkably well, especially given its relative simplicity. The Penman model was very sensitive to wind speed in our study. When applying PET models, one must be aware of soil moisture conditions so that potential and actual evapotranspiration are differentiated. Integration of data from all the FLUXNET sites across the globe will be critical in determining the best possible evapotranspiration model to use at global scales for predicting changes in land-surface exchange due to climate change.

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