A Penman-Brutsaert Model for Wet Surface Evaporation

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An atmospheric stability Penman-Brutsaert model for wet surface evaporation is developed for short time prediction of the diurnal latent heat flux. Monin-Obukhov similarity theory is used in conjunction with a theoretical scalar roughness length for bluff rough surfaces in the formulation of the drying power of air. The robustness of the model is evaluated with lysimeter measurements of wet bare soil evaporation for a range of atmospheric conditions. Instances where radiation or local advection dominated were measured and modeled. Good agreement ($r^2 = 0.96$) was obtained overall between the Penman-Brutsaert derived evaporation rates and values measured by the lysimeters.

INTRODUCTION

Potential evaporation, loosely defined as evaporation from surfaces where water is not limiting, is an important reference state for understanding land surface processes in hydrology and climatology. *Granger* [1989], in a review of the concept of potential evaporation, formally defined "wet surface evaporation" as the evaporation governed by the available energy and atmospheric considerations. This is the basis of the classic *Penman* [1948] combination equation, which combines both energy and atmospheric vapor transport components to model potential evaporation. *Thom and Oliver* [1977] note that the so-called combination approach is attractive both for aesthetic and practical reasons, so that Penman type evaporation equations are widely used in hydrology today. In this study we use a modified Penman equation to calculate the diurnal evaporation from a wet bare soil field over 20-min periods. Since the objective is to describe the water vapor fluxes throughout the day, stability of the atmosphere should be accounted for in the formulation of the atmospheric drying power [*Stricker and Brutsaert*, 1978; *Mahrt and Ek*, 1984]. *Brutsaert* [1982] discusses how the atmospheric stability may be incorporated with the Monin-Obukhov [*Monin and Obukhov*, 1954] similarity theory in modeling the capacity of the atmosphere to transport vapor in conjunction with the surface energy budget. In addition, *Brutsaert* [1975, 1982] has obtained a theoretical formulation of the scalar roughness length for bluff-rough surfaces on the basis of a local Reynolds number. The Penman equation with the Brutsaert suggestions for the drying power of the air and the vapor roughness length is tested against lysimeter evaporation measurements.

Experiments were carried out for 5 days at Davis, California, over a bare soil field in late summer and early fall of 1990, where the soil surface was saturated by sprinkle irrigation. Atmospheric and energy measurements were collected, and evaporation was measured with two sensitive lysimeters over 20-min-periods throughout the 5 days. The predictive capacity of the Penman-Brutsaert model is studied for a range of radiant and atmospheric forcing at the surface. The importance of atmospheric stability corrections in the context of the Penman-Brutsaert model is also examined.

EXPERIMENT

The sensitive lysimeter research facility is located at the Campbell field at the University of California, Davis. The soil is a Yolo clay loam which was periodically raked to minimize surface crust and seal formation and remove any vegetation. The surface had an irregular pattern of furrows 1–4 cm deep with scattered small soil clods of roughly 2–4 cm diameter throughout the field. The furrow-clod surface is impermeable to air flow and can be classified as a hydrodynamically bluff-rough surface. A field average surface roughness (z_0) of 2 cm was selected [see *Brutsaert*, 1982; *Panofsky and Dutton*, 1984; *Stull*, 1988]. Applied water was furnished by an irrigation system, composed of six laterals with 14 sprinklers per lateral, wetting a soil surface area approximately 130 m by 130 m. The irrigated soil is located within a larger bare soil region some 500 m by 500 m. A 16-can network was used to monitor the net application rates and obtain a uniformity coefficient for each irrigation run. The uniformity coefficients ranged from 0.80 to 0.85 [see *Cuenca*, 1989]. Irrigation was carried out for 3 to 5 hours the evening prior to each experimental day to sufficiently saturate the upper soil surface and maintain potential evaporation conditions for at least the next 24 hours. For each of the days studied the color of the soil was noted to make certain the surface was wet throughout the day.

The evaporative fluxes were measured on a 20-min basis with both a circular weighing lysimeter and a circular floating lysimeter. The lysimeters, centrally located on the irrigated field, are 6 m in diameter and 1 m in depth. The floating lysimeter generally exhibits slightly more fluctuation in the measurement of evaporation than the weighing lysimeter which is less sensitive to surface wind shear. The weighing lysimeter measurements are also somewhat erratic due to short bursts and changes in the wind speed. To eliminate some of the noise a $1:2:1$ smoothing is applied to the raw flux measurements. The lysimeters are suitable for this study, as they have proven to be reliable for measuring evaporation to 0.03 mm of equivalent water depth [*Pruitt and Angus*, 1960; *Aboukhaled et al.*, 1982].

Atmospheric data were logged over the 20-min intervals which correspond to the evaporation measurements with the lysimeters. The atmospheric measurements include temperature (T_a) and relative humidity (RH) at 0.80 m and mean horizontal wind speed (V) at 2 m. The surface energy components of net radiation (R_n) and soil heat flux (G) were also monitored for these periods. The soil heat flux was

TABLE 1. Some Daily Weather Characteristics

Julian Day	Applied Water, mm	T_a, °C			Relative Humidity, %			V, m s^{-1}			Rs Inc, max	R_n, min
		max	min	avg	max	min	avg	max	min	avg		
257	21	28.4	12.0	19.6	88.6	28.1	62.3	3.73	0.40	2.36	800	600
271	18	34.5	12.2	22.5	93.9	12.6	57.8	4.52	0.41	2.24	680	510
279	17	25.5	12.2	18.3	90.0	12.9	38.2	10.04	0.81	5.81	700	460
286	16	29.0	10.0	18.5	57.9	16.5	41.5	4.90	0.49	41.5	648	445
297	09	29.1	7.7	17.2	78.0	34.2	58.5	2.51	0.4	1.01	586	372

Rs Inc, maximum incident shortwave radiation; max, maximum; min, minimum; avg, average.

measured with two soil heat flux plates installed just below the surface (0.5 cm), and the net radiation was measured at 2 m with a Q6 Fritchen type net radiometer. To maximize the upwind fetch conditions the instruments were located at a north central portion of the field, since the winds are predominately from the south and southeast. The instrument sensors are located sufficiently high above the soil surface to be situated within the local surface layer of the field. It is important that the atmospheric properties be measured above the roughness wake layer, since the Monin-Obukhov similarity theory used in the analysis is based upon steady state surface layer flux profile relationships [*Parlange and Brutsaert*, 1989; *Sugita and Brutsaert*, 1990]. The average lower limit of the surface layer may be roughly scaled with the roughness length as $35z_0$ [e.g., *Garratt*, 1978; W. Brutsaert and M. B. Parlange, The unstable surface layer above forest: Regional evaporation and heat flux, submitted to *Water Resources Research*, 1991]. (Hereafter referred to as W. Brutsaert and M. B. Parlange, 1991.) The sensors should be placed in the lower few tens of meters above the surface such that assumptions concerning the "constant stress layer" are not violated.

Experiments on potential evaporation were carried out on Julian days 257, 271, 279, 286, and 297 (1990). Table 1 summarizes general radiation and weather characteristics for each day of the experiment and the amount of water applied in the evening prior to each experimental run. The maximum incident shortwave radiation is tabulated.

PENMAN-BRUTSAERT POTENTIAL EVAPORATION THEORY

The general form of the *Penman* [1948] combination equation is

$$E_p = W(R_n - G) + (1 - W)E_A \tag{1}$$

where E_p is the latent heat flux, $W = \Delta/(\Delta + \gamma)$ is a dimensionless weighing function, Δ is the slope of the saturation vapor pressure-temperature curve, γ is the psychrometric constant, and E_A is an atmosphere drying power function which represents the capacity of the atmosphere to transport water vapor. The first and second terms on the right-hand side of (1) represent the energy and atmospheric contributions to evaporation, respectively. The E_A was originally presented as a bulk transfer function on the basis of empirical observations for daily or longer time steps and has been the subject of extensive experimentation [e.g., *Penman*, 1948, 1956; *Van Bavel*, 1966; *Jensen*, 1974; *Doorenbos and Pruitt*, 1977; *Allen*, 1986].

Brutsaert [1982] suggested on the basis of Monin-Obukhov similarity that the drying power of the air can be described as

$$E_A = ku_*\rho(q^* - q_a)\left(\ln\left[\frac{z - d_{0v}}{z_{0v}}\right] - \psi_v\left[\frac{z - d_{0v}}{L}\right]\right)^{-1} \tag{2}$$

where $k = 0.4$ is von Karman's constant, $u_* = (\tau_0/\rho)^{1/2}$ is the friction velocity, τ_0 is the surface shear stress, ρ is the density of the air, d_{0v} is the displacement height for water vapor, z is the height of measurement above the surface, z_{0v} is the vapor roughness height, and q_a and q^* are the specific humidity of the air and the saturation specific humidity at air temperature, respectively. The Monin-Obukhov similarity stability correction function ψ_v depends on $(z - d_{0v})/L$; L is the Obukhov length, defined by

$$L = \frac{-u_*^3}{kg[H_v/(\rho c_p T_a)]} \tag{3}$$

where $H_v = (H + 0.61\, T_a c_p E_p)$ is the specific flux of virtual sensible heat at the surface, c_p is the specific heat at constant pressure, T_a is the air temperature, and H is the specific flux of sensible heat. The atmospheric stability classification is based on L, where $L < 0$, $L > 0$, and $|L| \geq 100$ are the limits used to signify unstable, stable, and neutral conditions, respectively.

The friction velocity is obtained from the mean horizontal wind speed described in the context of Monin-Obukhov similarity:

$$V = \frac{u_*}{k}\left[\ln\left(\frac{z - d_0}{z_0}\right) - \psi_m\left(\frac{z - d_0}{L}\right)\right] \tag{4}$$

where d_0 is the momentum displacement height, and ψ_m is the momentum stability correction function. The stability correction functions have been obtained from numerous field experiments, and the Businger-Dyer formulation is now widely accepted [*Dyer*, 1974; *Businger*, 1988; *Högström*, 1988; *Sugita and Brutsaert*, 1990; W. Brutsaert and M. B. Parlange, 1991]. For unstable conditions these functions can be written as

$$\psi_m = \ln\left[\frac{(1 + x)^2(1 + x^2)}{(1 + x_0)^2(1 + x_0^2)}\right] - 2\arctan(x) + 2\arctan(x_0) \tag{5}$$

$$\psi_v = 2\ln\left(\frac{1 + x^2}{2}\right) \tag{6}$$

where $x = (1 - 16y)^{1/4}$, $y = (z - d_0)/L$, and $x_0 = (1 - 16z_0/L)^{1/4}$. For stable conditions, ψ_m and ψ_v are assumed to be equal where

$$\psi_v = \psi_m = 5(z_0/L - y) \qquad 0 < y \le 1 \tag{7}$$

$$\psi_v = \psi_m = -5 \ln\left(\frac{z - d_0}{z_0}\right) \qquad 1 < y \le 10 \tag{8}$$

and for very stable conditions the stability function used in this study was developed by *Hicks* [1976] from observations of the mean wind speed gradient:

$$\psi_v = \psi_m = \ln\left(\frac{z - d_0}{z_0}\right) + 0.8\left(\frac{z_0}{L} - y\right) \tag{9}$$

The scalar roughness z_{0v} is estimated from an expression presented by *Brutsaert* [1975, 1982] for hydrodynamically bluff-rough surfaces:

$$z_{0v} = 7.4 z_0 \exp\left[-2.25(z_{0+}^{1/4})\right] \tag{10}$$

where $z_{0+} = (u_* z_0)/\nu$ is the roughness Reynolds number and ν is the kinematic viscosity. The sensible heat flux may be computed from the surface energy budget

$$H = R_n - G - E_p \tag{11}$$

which closes the system of equations to solve for E_p iteratively. The system from which E_p is computed includes (1), (2), (3), (4), and (11). The unknowns in this system of equations are E_p, E_A, H, L, and u_*. Because of the nonlinearity of the equations a closed form solution for E_p is not possible, hence an iterative scheme was employed with an error criterion of 0.1 W m^{-2} on E_p. The scheme is initiated by assuming neutral stability conditions and setting the stability correction functions equal to zero on the first pass. A first L value is obtained from the neutral approximation and then the full system with stability corrections is repeated until the E_p estimate converges. In general, this scheme converges with less than six passes to achieve closure of 0.1 W m^{-2}.

RESULTS AND DISCUSSION

The Penman-Brutsaert evaporation flux model is tested with measurements made at the Davis lysimeter facility. The model was employed on a 20-min time step corresponding to the 20-min lysimeter and micrometeorological measurements made in the field. The momentum and vapor displacement heights were assumed to be equal and set at zero. It is known that d_0 and d_{0v} are different, but for practical purposes they are approximately equal [*Brutsaert et al.*, 1989; *Kustas*, 1990]. It was noted that the system was insensitive to the choice of d_0 and d_{0v}.

The evaporation fluxes measured by the weighing lysimeter (E_{wl}) and the Penman-Brutsaert modeled flux estimates are plotted in Figures 1a, 1b, 1c, 1d, and 1e for Julian days 257, 271, 279, 286, and 297, respectively. The net radiation and the soil heat flux are also plotted in these figures as a reference energy state throughout the day. For each day, linear regression analysis between the latent heat flux calculated from the model and measured with the lysimeter is presented in Table 2. The coefficient of determination (r^2), the standard error of estimate (SEE), and the linear regression model, $E_p = AE_{wl} + B$, are listed in Table 2. The flux comparisons in Figure 1 and Table 2 indicate consistent agreement between the Penman-Brutsaert model and the

TABLE 2. Linear Regression Analysis for $E_{p,pn} = A\, E_{wl} + B$ and the t Statistic for the Hypothesis $A = 1$

Julian Day	r^2	Slope A	Constant B	t Statistic	SEE, W m^{-2}
257-E_p	0.95	0.98	-1.40	0.36	38.6
257-E_{pn}	0.94	0.93	6.72	1.01	39.1
271-E_p	0.98	0.95	-17.76	1.30	21.8
271-E_{pn}	0.97	0.92	-1.28	2.30	21.7
279-E_p	0.98	0.93	-4.88	1.94	24.1
279-E_{pn}	0.97	0.88	14.03	3.10	25.4
286-E_p	0.96	0.96	-14.05	0.76	30.9
286-E_{pn}	0.97	0.92	2.65	1.74	26.8
297-E_p	0.96	0.95	-15.01	0.82	24.4
297-E_{pn}	0.96	0.89	-6.39	2.03	21.9

SEE, standard error of estimate.

lysimeter measurements for each day. The intercepts for all the linear regression models were less than the standard error of estimate, and the slopes were not significantly different from unity at the 95% confidence level.

The comparisons provide evidence of the robustness of the Penman-Brutsaert model to provide accurate latent heat flux estimates when different combinations of atmospheric forcing play a role in evaporation at the land surface. Energetic considerations dominated the evaporation process on day 257 (see Figure 1a). The evaporation was less than or equal to the net radiation during the daytime hours. On day 271 the wind speed increased in the afternoon, and the mechanical production of turbulence was an important vapor removal mechanism. The model captured the advection in the afternoon so that the energy and aerodynamic effects were accurately described (see Figure 1b). The increase in wind speed in the afternoon due to the sea breeze influence is typical for the central California Valley. Julian day 279 could be considered a definitive test case for the study of local advection impacts. The evaporation rate was in excess of the net radiation for most of the day (see Figure 1c). Dry winds in excess of 10 m s^{-1} occurred, which resulted in high surface shear stress (see Figure 2) and high specific humidity deficit. The Obukhov length, plotted in Figure 2, indicates stable or neutral stability throughout the day. Since the latent heat flux exceeds the available energy, $(R_n - G)$, the sensible heat flux is negative. The flux comparison shown in Figure 1c demonstrates the ability of the model to describe the drying power of the air when strong dry winds pass over a wet surface. The model was also observed to work well on Julian days 286 and 297 (see Figures 1d and 1e). Day 297 was a relatively low flux day so that noise in the weighing lysimeter evaporation measurements are more apparent. These days were typically energy driven in the morning daytime hours and had advective effects in the afternoon when the sea breeze increased.

In general, the fluxes in the evening and early morning hours represent a small part of the total daily evaporation, though on days 271, 279, and 286 there was some evaporation measured and modeled when the horizontal wind speed increased in the evening. Strongly stable conditions at night are the most difficult conditions to describe. In a review, *Brutsaert* [1982] notes that little is known concerning the Monin-Obukhov flux-profile equations under strongly stable conditions and that the experimental evidence indicates that the stability correction functions are not equal [see *Monin*
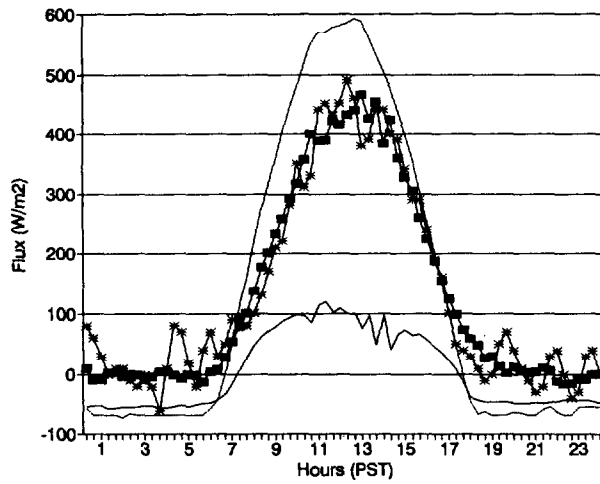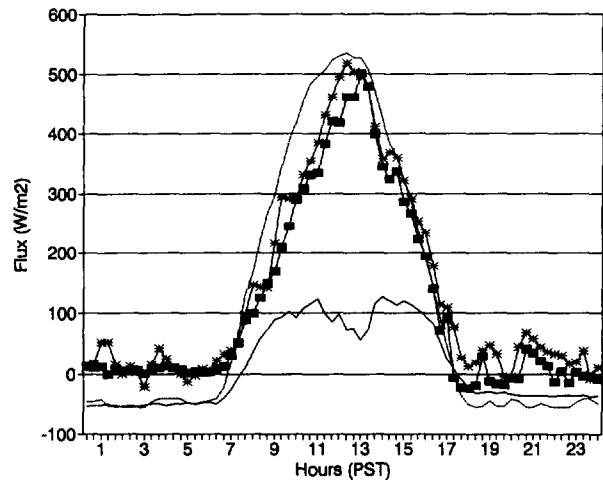
Fig. 1a.   Julian day 257, 1990.
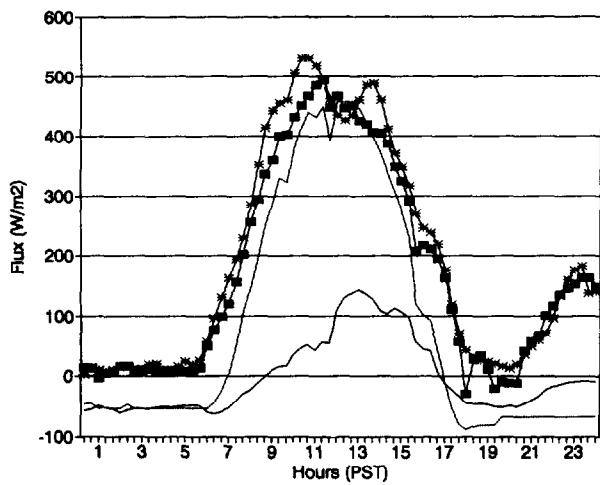
Fig. 1b.   Julian day 271, 1990.
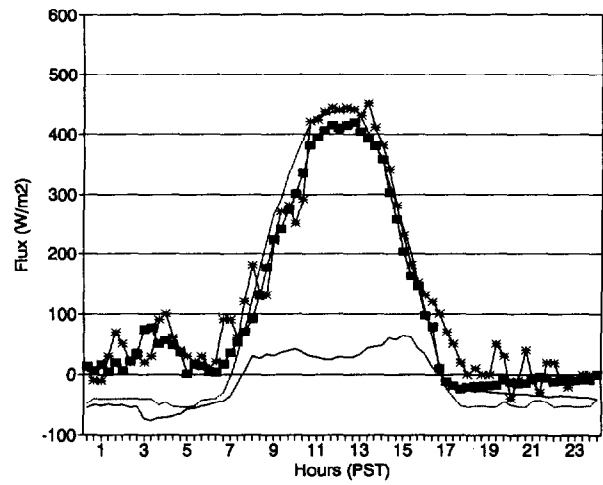
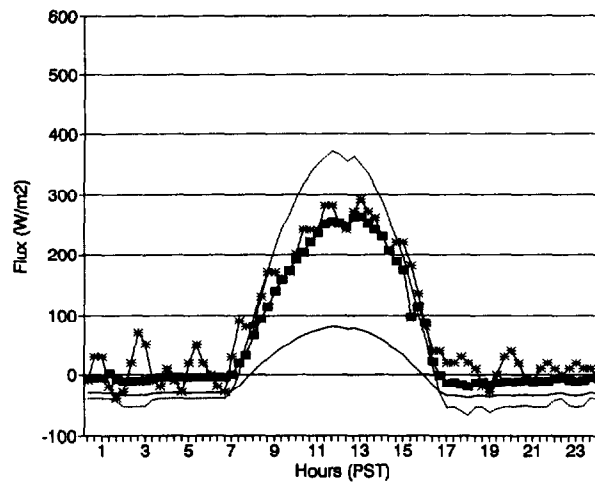Fig. 1c   Julian day 279, 1990.

Fig. 1d.   Julian day 286, 1990.

Fig. 1e.   Julian day 297, 1990.

Fig. 1.   Comparison of lysimeter-measured (asterisk) and computed (solid rectangle) latent heat fluxes. Net radiation (dotted curve) and average soil heat flux (solid curve) are also shown.
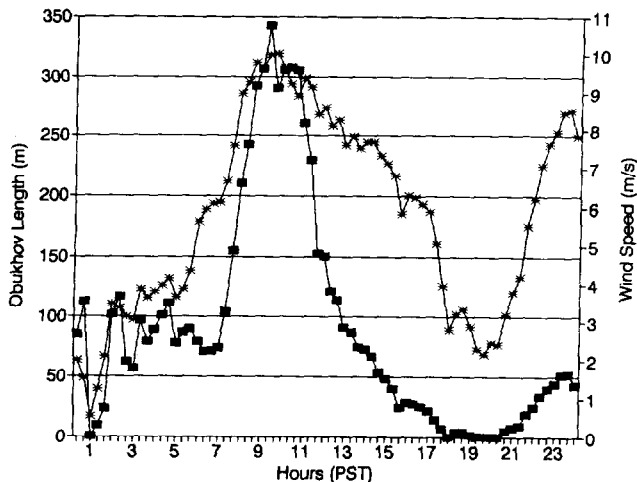
Fig. 2. Computed Obukhov length L (solid rectangle) and mean horizontal wind speed (asterisk) for Julian day 279, 1990.

and Yaglom, 1971; *Warhaft*, 1976; *Hicks*, 1976; *Kondo et al.*, 1978; *Lang et al.*, 1983]. Since the fluxes usually tend to be small, for practical flux estimation the stability correction formulation is usually not as important for strongly stable conditions. Our observations indicate that when the night water vapor fluxes contribute to the total daily losses, the suggestion of *Hicks* [1976] is adequate.

The overall comparison of E_p versus E_{wl} is presented in Figure 3. The overall regression analyses (n = 360), presented in Table 3, indicate that the standard error of estimate was of the order of the lysimeter accuracy and that the slope of the regression model was also not different from unity at the 95% confidence level.

The importance of the stability correction functions for the latent heat flux calculation is investigated by setting Ψ_m and Ψ_v to zero. This simplification could be justified if Ψ_m and Ψ_v are small compared to ln $[(z - d_0)/z_{0,0v}]$. By assuming neutral conditions and comparing the estimated latent heat flux E_{pn} with the lysimeter measurements the impact of stability on the flux estimation is tested. Under this assumption a linear regression analysis ($E_{pn} = AE_{wl} + B$) is presented in Table 2 for each day, and the overall model with
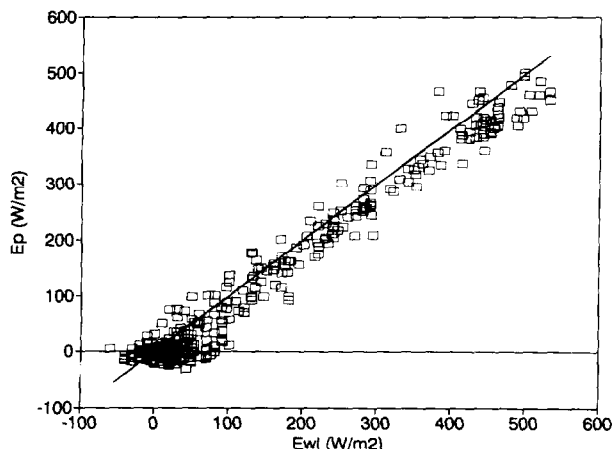

Fig. 3. Latent heat flux comparison between the weighing lysimeter measurements (E_{wl}) and the model predictions (E_p) for all days. The solid curve is 1:1 (number of points is 360).

TABLE 3. Linear Regression Analysis for All Lysimeter Observations (n = 360)

Linear Regression Model	Coefficient of Determination r^2	SEE, W m^{-2}
E_p = 0.95 E_{wl} − 11.02	0.96	28.3
E_{pn} = 0.91 E_{wl} + 1.86	0.95	28.5
E_{fl} = 0.86 E_{wl} + 8.15	0.91	42.37

E_{fl}, floating lysimeter flux measurement.

assumed neutral conditions is presented in Table 3. The standard error of estimate and the regression intercept for each day is similar to that obtained when the stability correction functions are included. Except for days 257 and 286 the slope of the regression model is found to be statistically (95%) different from unity. It is suggested then that the inclusion of the stability correction functions in the Penman-Brutsaert model is preferable to the simplified neutral formulation.

CONCLUSIONS

The Penman-Brutsaert model formulated with Monin-Obukhov similarity in the framework of an energy budget is accurate for potential flux calculation over short time intervals (<1 hour). The inclusion of atmospheric stability can be important in describing diurnal changes in both the wind speed and radiant energy patterns. Experiments were carried out on days when radiation forcing dominated the evaporation process as well as on days when the surface shear stress enhanced the vapor removal capacity of the atmosphere such that the latent heat exceeded the net radiation. Typically, both radiation and momentum fluxes were found to be important, with radiation dominating in the morning and surface shear stress enhancing the afternoon rate of evaporation. The advective effect in the afternoon sometimes caused the latent heat flux to exceed ($R_n - G$), resulting in conditions of stable stability. The Businger-Dyer stability correction functions were used for unstable and mildly stable conditions, and for strong stability the *Hicks* [1976] suggestion was applied.

The inputs used in the model can be routinely measured on a short time step in the field, although the soil heat flux and net radiation are susceptible to measurement error. In addition, the model is universal in that there is no local calibration based on site specific factors other than the surface roughness which is tabulated or may be measured in the field. The reliability of the model is strengthened since the model predictions are compared to flux measurements (lysimeter) which are independent of the model input.

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