

An Advection-Aridity Evaporation Model

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Actual evaporation is calculated by means of an advection-aridity complementary model which requires, as input, the meteorological data used in classical combination models. The main advantage of the model is that it does not require site-specific calibration. The advection-aridity formulation presented here includes the effect of atmospheric stability, which can be important to describe the diurnal evaporation variation, and a theoretical model of the scalar roughness height. A parameterization for advection over the surface of interest is incorporated which allows the model to be used for a wide range of natural conditions. The model was found to give good agreement with evaporation measurements obtained with a large sensitive lysimeter over a bare soil field.

INTRODUCTION

Estimation of actual evaporation from standard meteorological input variables is an important problem in hydrology, water resources, and climatology. Some 30 years have passed since *Bouchet* [1963] suggested the hypothesis of symmetry between potential evaporation (E_p) and actual evaporation (E) when the transport of water vapor into the atmosphere is limited at the land surface. The symmetry between potential and actual evaporation has been the basis of complementary models which require standard one level meteorological variables as input: net radiation (R_n), soil heat flux (G), mean air temperature (T_a), mean horizontal wind speed (V), and mean air relative humidity (RH) [e.g., *Brutsaert and Stricker*, 1979; *Brutsaert*, 1982; *Morton*, 1983].

Brutsaert and Stricker [1979] proposed the advection-aridity model based on the *Bouchet* symmetric arguments. Their model has been found to work well for daily evaporation predictions [e.g., *Brutsaert and Stricker*, 1979; *Ali and Mawdsley*, 1987; *LeMeur and Lu*, 1990]. The main advantage of the advection-aridity model is that it does not require surface resistance, soil moisture content, or other land surface measures of aridity [*Brutsaert and Stricker*, 1979].

This paper studies the advection-aridity model for diurnal evaporation and uses a parameterization to describe the horizontal advection of dry air which does not rely on any calibration. The diurnal model presented here is based on the *Brutsaert-Stricker* model and is tested with evaporation measurements on a 20-min interval. The potential evaporation is computed using a *Penman* equation [*Katul and Parlange*, 1992], while the reference wet surface evaporation (E_{po}) is computed using the *Priestley and Taylor* [1972] model. The proposed model is compared with bare soil evaporation data collected at the *Davis* sensitive lysimeter facility for both wet and dry surface conditions and a range of atmospheric conditions.

ADVECTION-ARIDITY MODEL AND PROPOSED FORMULATION

Bouchet [1963] postulated that as an initially wet surface of a region dries, the decrease in actual evaporation corre-

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sponds to an equivalent increase in potential evaporation. *Bouchet* notes that initially when the surface is wet, the evaporation rates are equal ($E = E_p = E_{po}$). When the water for evaporation becomes limited at the land surface, for the same quantity of energy available for evaporation the actual evaporation drops below E_{po} by q_1 ,

$$E - E_{po} = -q_1 \quad (1)$$

so that q_1 becomes available which increases E_p . The decrease in E below E_{po} affects primarily the air temperature, the air humidity, and the stability of the atmosphere [*Bouchet*, 1963]. *Bouchet* then hypothesized that the change in the potential evaporation could be given by

$$E_p = q_1 + E_{po} \quad (2)$$

Addition of (1) and (2) yields the complementary relationship

$$E + E_p = 2E_{po} \quad (3)$$

based on the assumption that q_1 does not alter the available energy, and that no external energy suddenly enters the region. The computation of E requires appropriate expressions for E_p and E_{po} which should be given by standard measured meteorological variables if a practical model is to be obtained. *Brutsaert and Stricker* [1979] adopted the *Penman* combination equation for potential evaporation,

$$E_p = \frac{\Delta}{\Delta + \gamma} (R_n - G) + \frac{\gamma}{\Delta + \gamma} E_a \quad (4)$$

where Δ is the slope of the saturation vapor pressure temperature curve, γ is the psychrometric constant, and E_a is the drying power of the air.

Brutsaert and Stricker were interested in modeling daily evaporation and used the *Penman* [1948] daily wind function in E_a . When short-term (<1 hour) estimates of the fluxes are desired, the effect of atmospheric stability can be important in the formulation of E_a [*Stricker and Brutsaert*, 1978; *Mahrt and Ek*, 1984; *Katul and Parlange*, 1992; *Parlange and Katul*, 1991]. *Brutsaert* [1982] suggested on the basis of *Monin and Obukhov* [1954] surface layer similarity theory that

$$E_a = ku_* \rho (q_a^* - q_a) \left[\ln \left(\frac{z_a - d_o}{z_{ov}} \right) - \Psi_v \left(\frac{z - d_o}{L} \right) \right]^{-1} \quad (5)$$

where d_o is the zero plane displacement height, z_a is the height of measurement above the surface, z_{ov} is the vapor roughness height, and q_a^* and q_a are the saturation specific humidity at air temperature T_a and the actual specific humidity of the air, respectively. The specific humidity, the vapor pressure (e), and the atmospheric pressure (P) can be approximately related through [$q = 0.622e/P$]. The *Monin and Obukhov* [1954] similarity stability function, ψ_v , depends on $y = (z - d_o)/L$; L is the Obukhov Length defined by

$$L = \frac{-u_*^3}{kg \left[\frac{H_v}{\rho c_p T_a} \right]} \quad (6)$$

where $H_v = (H + 0.61T_a c_p E)$ is the specific flux of virtual sensible heat and c_p is the specific heat of air at constant pressure. The *Businger-Dyer* [Dyer, 1974; Businger, 1988] stability correction functions for vapor are employed,

$$\Psi_v = 2 \ln \left[\frac{1 + x^2}{2} \right]; \quad y < 0 \quad (7)$$

$$\Psi_v = 5 \left(\frac{z_o}{L} - y \right); \quad 0 < y \leq 1 \quad (8)$$

$$\Psi_v = -5 \ln \left[\frac{z - d_o}{z_o} \right]; \quad 1 < y \quad (9)$$

where $x = (1 - 16y)^{1/4}$. Equation (7) is used for unstable atmospheric conditions and (8) and (9) for stable conditions. The friction velocity is also obtained from the Monin-Obukhov similarity equation for the mean horizontal wind speed V ,

$$V = \frac{u_*}{k} \left[\ln \left(\frac{z - d_o}{z_o} \right) - \Psi_m \left(\frac{z - d_o}{L} \right) \right] \quad (10)$$

where d_o is the zero plane displacement height and ψ_m is the momentum stability correction function. For stable atmospheric conditions the momentum stability correction function is assumed to be equal to the vapor stability correction function and for unstable atmospheric conditions it is given by

$$\Psi_m = \ln \left[\frac{(1 + x)^2(1 + x^2)}{(1 + x_o)^2(1 + x_o^2)} \right] - 2 \arctan(x) + 2 \arctan(x_o) \quad (11)$$

where $x_o = (1 - 16z_o/L)^{1/4}$.

For bluff rough surfaces the vapor roughness height can be determined with

$$\frac{z_{ov}}{z_o} = 7.4 \exp[-2.25(z_{o+})^{1/4}] \quad (12)$$

where $z_{o+} = (u_* z_o)/\nu$ is the roughness Reynolds number and ν is the kinematic viscosity [Brutsaert, 1975, 1982]. It is assumed that the displacement heights for vapor and momentum are approximately equal. This assumption does not introduce any significant error since z_a is large compared to d_o , so that $(z_a - d_o)$ is approximately equal to $(z_a - d_{ov})$. The value of E_a cannot be explicitly expressed in terms of

the measured variables (R_n , G , V , RH , T_a , and z_o); hence a numerical solution requiring an iterative procedure [Katul and Parlange, 1992] is used to solve the system for E_p , H_p , E_a , L , and u_* at each 20-min time step, where $H_p = R_n - G - E_p$. The system is initiated by assuming neutral conditions in the first pass and then corrected for atmospheric stability in successive passes. The system to solve for E_p converges rapidly with a closure specification of 0.1 W m^{-2} usually obtained in five passes.

The term E_{po} in (4) is the reference evaporation rate which would occur if the surface was brought to saturation (wet) and the available energy supply ($R_n - G$) were held constant during that period. This term has been the subject of research and conflicting definitions [Seguin, 1975; Fortin and Seguin, 1975; Brutsaert and Stricker, 1979; Morton, 1983; Granger, 1989a, b; McNaughton and Spriggs, 1989] and is difficult to quantify from measurements over an unsaturated surface. In the advection-aridity model this is described with the *Priestley and Taylor* [1972] equation,

$$E_{po} = \alpha \frac{\Delta}{\Delta + \gamma} (R_n - G). \quad (13)$$

Priestley and Taylor [1972] concluded that α is a constant between 1 and $(\Delta + \gamma)/\Delta$, indicating that E_{po} from large saturated areas is bounded by the equilibrium evaporation and $(R_n - G)$, respectively. Therefore the formulation of the *Priestley and Taylor* [1972] model of wet surface evaporation always assumes a positive or zero sensible heat flux. The constant value of α was found experimentally to be 1.26 for conditions of minimal advection [Priestley and Taylor, 1972; Davies and Allen, 1973; Stewart and Rouse, 1976, 1977; Jury and Tanner, 1975; Parlange and Katul, 1991]. The energy difference ($R_n - G$) is identical to that used in the calculation of E_p . When $\Delta/(\Delta + \gamma)$ exceeds α^{-1} , which it does for temperatures exceeding approximately 30° , the Bowen ratio calculated using H_{po}/E_{po} is negative ($H_{po} = R_n - G - E_{po}$). This means that (13) may not always be suitable to define the reference E_{po} even for conditions of minimal advection. For wet to moderately dry surfaces this restriction is generally satisfied. For extremely dry surfaces this assumption can be severely violated, though under these conditions the evaporation is generally not large.

The main restriction on the Bouchet hypothesis is that it is valid only when there is no advection from upwind over the surface of interest. This restriction can be relaxed if a modification is made to the Bouchet complementary hypothesis and the formulation of the advection-aridity model to account for advection in the definition of the reference wet surface evaporation [Parlange and Katul, 1991]. In the event that E_p is larger than $(R_n - G)$ the increase in E_p is due not only to the energy q_1 becoming available but is also a result of horizontally advected drier air which increases E_a and hence E_p . Advection is already accounted for in the formulation of E_p since the calculation is closed using the surface energy balance, $R_n - G = E_p + H_p$. The impact of the advection must then be accounted for whenever $H_p < 0$ in the formulation of E_{po} . If the surface was wet and $H_p < 0$, then the *Priestley-Taylor* model obviously underpredicts E_{po} . *Parlange and Katul* [1991] suggested that the wet surface evaporation be increased by $|H_p|$ when advection is important, which linearly displaces the entire Bouchet symmetric relationship (see Figure 1). Operationally, whenever

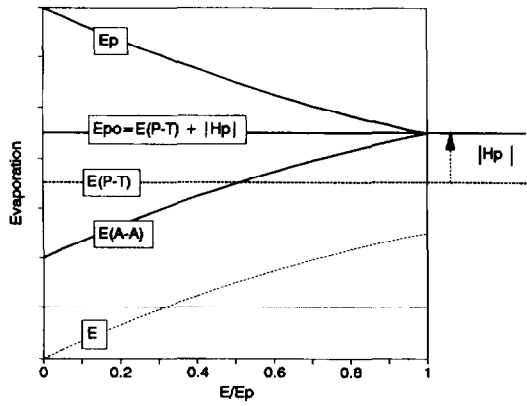


Fig. 1. Sketch of the adjustment of the Bouchet [1963] hypothesis for advection [after Parlange and Katul, 1991].

$H_p < 0$, the wet surface evaporation is increased, $E_{p0} = E_{pT} + |H_p|$. In general, the impact of horizontal advection is most pronounced when a strong dry front passes over the surface of interest and increases both the vapor pressure deficit and the friction velocity and consequently the drying power of the air E_a . This is of special interest in agricultural regions (e.g., California Central Valley) where dry air is often transported over recently irrigated surfaces. The importance of this correction will be discussed further on the basis of the experimental evidence.

EXPERIMENT

The sensitive lysimeter research facility is located at the Campbell Tract at the University of California, Davis. The evaporation rate was measured on a 20-min time step with a large sensitive weighing lysimeter (E_{wl}), 6 m in diameter and 1 m in depth. The lysimeter is reliable for measuring evaporation to 0.03 mm of equivalent water depth [Pruitt and Angus, 1960; Pruitt and Lorrence, 1985]. The advantage of using a lysimeter to test the model is that the evaporation measurements are independent of the surface energy budget components used in the model calculation.

The meteorological data V , T_a , and RH were measured over a 20-min time step at 2, 0.80, and 0.80 m, respectively. The meteorological tower is located some 20 m northwest of the lysimeter. The research site, which can be uniformly irrigated (150 m \times 130 m), is situated within a larger bare soil region (500 m \times 500 m). Net radiation (R_n) was measured by

a Q6 Fritchen type net radiometer, and the soil heat flux by two plates of constant thermal conductivity placed just below the soil surface. In order to study the availability of water below the soil surface, five neutron probe access tubes were installed along an East-West transect at 30 m spacing (see M. B. Parlange et al., Physical basis for a time series model of moisture content using a simple hydrologic budget submitted to *Water Resources Research*, 1991). The volumetric moisture content (VMC) was measured in the top 30 cm by neutron probe scattering techniques using a Campbell Pacific Nuclear probe (CPN-model 503). The albedo (α_s) was monitored by two shortwave sensors measuring incident shortwave radiation (R_{si}) and reflected shortwave radiation (R_{sr}). In general, wet soil surfaces are darker than dry soil surfaces, and therefore the albedo of the soil surface can serve as an indication of the availability of water at the surface [e.g., Idso et al., 1975]. Experiments were carried out from September 14, 1990, until January 18, 1991. Table 1 presents a summary of the general energy, meteorological, and surface conditions for the days used in this study.

RESULTS AND DISCUSSION

The evaporation fluxes measured by the weighing lysimeter (E_{wl}) and calculated by the advection-aridity approach are compared for 9 days in Figure 2 for Julian days 257/90, 288/90, 297/90, 324/90, 338/90, 340/90, 343/90, 351/90, 017/91. The advection-aridity formulation $E_{(A-A)}$ without the adjustment of the reference wet surface evaporation by H_p , the advection aridity formulation with the H_p adjustment $E_{(adj)}$, H_p , and R_n are plotted in Figure 2. For each individual day, two linear regression models of the form $E_{wl} = AE_{adj} + B$ and $E_{wl} = AE_{adj}$ were determined, where A and B are the slope and intercept of the linear regression models, respectively. The coefficient of determination (r^2), and the standard error of estimate (SEE) for the regression model $E_{wl} = AE_{adj} + B$ are presented in Table 2. Brutsaert [1982] reports the range of daily albedo values for moist dark soils in plowed fields to be 0.05–0.13, while for dry soils in bare fields to be 0.15–0.25. The range of daily albedo values measured in this study fall within these two ranges. Values used to characterize the surface moisture condition were taken as $\alpha_s < 0.11$ for moist surfaces and $\alpha_s > 0.13$ for dry surfaces.

For moist surface conditions ($\alpha_s < 0.11$; Julian days 257-90; 297-90; 324-90; 351-90 Figure 2a, 2c, 2d, and 2h) the E_{adj} compared well with lysimeter measurements E_{wl} with an average standard error of estimate (33 $W m^{-2}$) which is

TABLE 1. Surface, Radiation, Energy, and Meteorological Conditions

Julian Day	Average VMC, % (0–30 cm)	Average Daily α_s	Radiation State, Maximum $W m^{-2}$		Energy State, Maximum $W m^{-2}$		Meteorological State		
			R_{si}	R_{sr}	R_n	G	T_a max, $^{\circ}C$	RH min, %	V_{max} , $m s^{-1}$
257-90	33.64	0.09	788	69.1	593	119	28.4	28.1	3.73
288-90	30.79	0.13	627	80.0	375	242	27.6	36.7	3.70
297-90	28.52	0.10	586	68.1	372	80	29.1	34.2	2.50
324-90	26.78	0.09	412	40.1	252	45	13.7	57.5	2.91
338-90	25.77	0.15	381	56.5	196	92	14.4	52.0	2.98
340-90	25.41	0.15	435	66.1	228	106	16.1	45.2	2.80
343-90	25.08	0.15	482	72.9	280	123	17.4	36.3	2.71
351-90	35.26	0.11	433	46.5	265	118	11.4	77.1	2.64
017-91	32.46	0.14	471	70.8	283	87	17.1	51.9	4.97

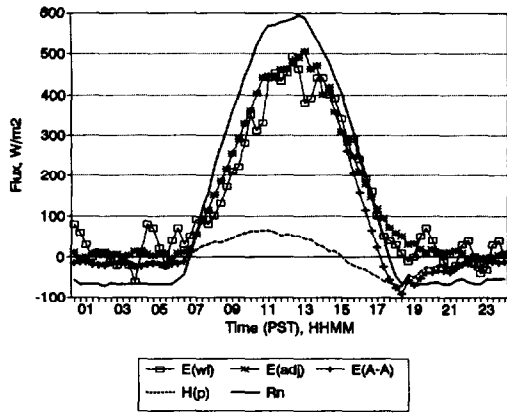


Fig. 2a. Comparison between the advection-aridity model (plus) without adjustment for advection, the adjusted advection-aridity model (asterisk), and the weighing lysimeter (rectangles). The potential sensible heat flux H_p (dotted line) and the net radiation R_n (solid line) are also plotted for Julian day 257, 1990.

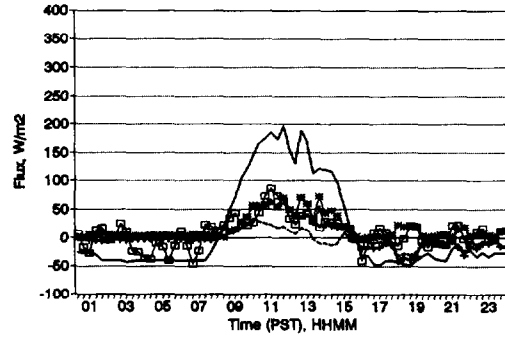


Fig. 2e. Julian day 338, 1990.

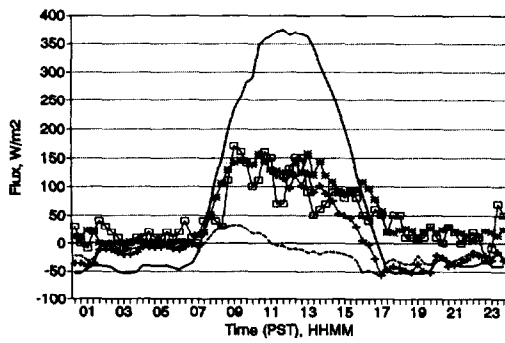


Fig. 2b. Julian day 288, 1990.

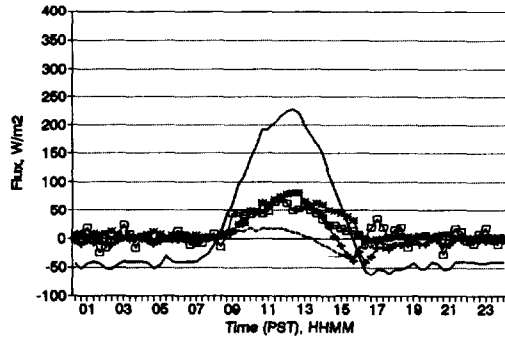


Fig. 2f. Julian day 340, 1990.

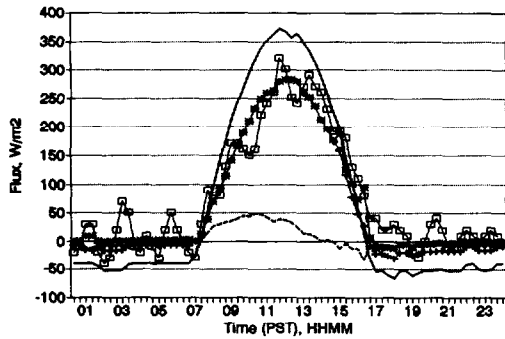


Fig. 2c. Julian day 297, 1990.

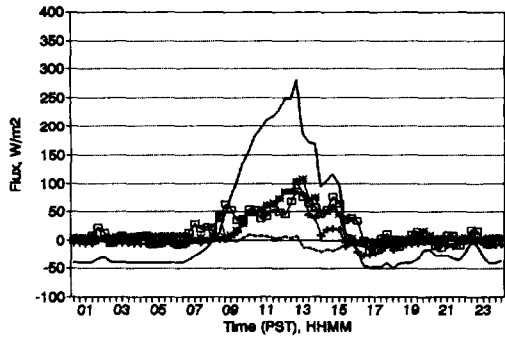


Fig. 2g. Julian day 343, 1990.

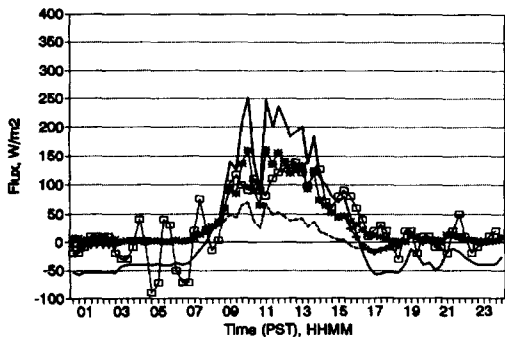


Fig. 2d. Julian day 324, 1990.

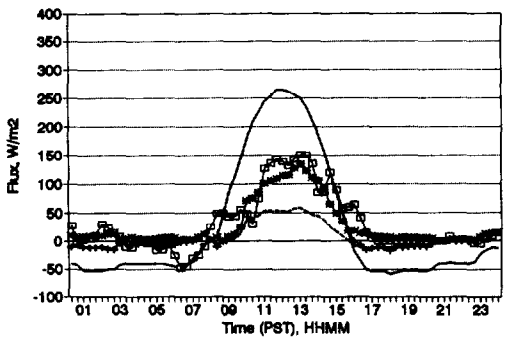


Fig. 2h. Julian day 351, 1990.

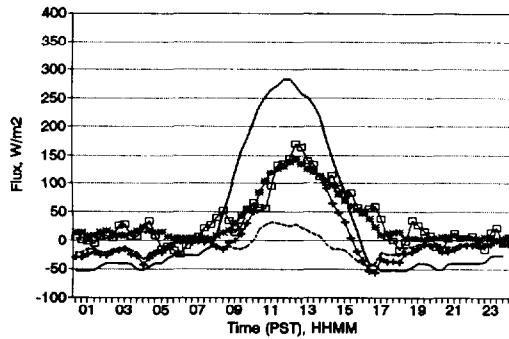


Fig. 2i. Julian day 017, 1991.

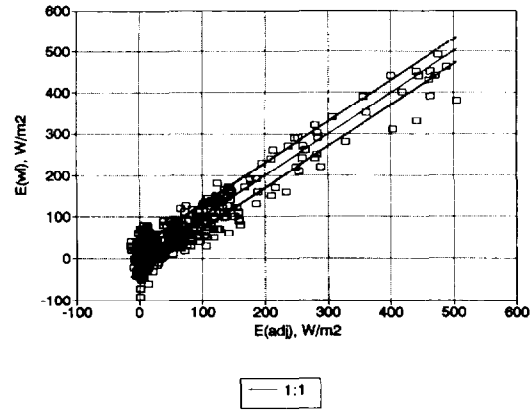


Fig. 3. Overall comparison between measured latent heat flux $E(wl)$ versus adjusted latent heat flux $E(adj)$ advection-aridity model predictions ($n = 648$ points). The 1:1 line and the $\pm 30 \text{ W m}^{-2}$ standard lysimeter error are plotted.

approximately equal to the estimated lysimeter measurement error (30 W m^{-2}). For day 257-90 (Figure 2a) the maximum latent heat flux was 500 W m^{-2} , and the adjusted advection-aridity model correlated well with lysimeter measurement ($r^2 = 0.94$) (see Table 2). The unadjusted advection-aridity clearly underpredicted the late afternoon fluxes when $H_p < 0$. On day 297 (Figure 2c) the sensible heat H_p was small ($< 50 \text{ W m}^{-2}$) throughout the day. The adjustment by H_p did not appreciably affect the flux calculations, though the unadjusted advection-aridity slightly underpredicted lysimeter measurements. On day 324 (Figure 2d) the atmosphere was humid and the sky was cloudy. The soil water content in the top 30 cm was 26.8%. The evaporation was low (maximum $< 150 \text{ W m}^{-2}$) and the impact of advection was minimal so that the unadjusted advection-aridity model performed well through out the day. The coefficient of determination ($r^2 = 0.67$) was rather low when compared to day 257 or 297 (see Table 2), nevertheless, the lower correlation can be attributed in part to the noise in the lysimeter measurement which affects the low flux end more than the high flux end. It should be noted that for this day the SEE was around 33 W m^{-2} (comparable to the measurement error = 30 W m^{-2}) while the maximum flux measured was about 140 W m^{-2} , signifying that the SEE was about 25% of the maximum flux. Between Julian days 345-349, several

precipitation events occurred in the Davis region. On day 351 (Figure 2h) the adjusted advection-aridity method correlated well with lysimeter measurements (see Table 2).

For the dry surface conditions ($\alpha_s > 0.11$; Julian days 288, 338, 340, 343, 017 Figure 2b, 2e, 2f, 2g, and 2i), the adjusted advection-aridity model again resulted in a substantial improvement over the unadjusted advection-aridity model. On Julian day 288 (Figure 2b), H_p was rather small in magnitude but negative all afternoon. The wind speed increased to 3.7 m s^{-1} at 1700 which resulted in an increase in $-H_p$. This is common in Davis when the adjacent unirrigated bare soil fields were subjected to strong radiation during the morning, and in the afternoon, strong winds transported the drier air over the research field. The H_p adjustment systematically improved the advection-aridity formulation to describe the impact of afternoon advection on the actual evaporation rate.

The average volumetric moisture content in the top 30-cm layer for Julian days 338, 340, and 343 (Figure 2e-2g) was about 25%, which limited the availability of water for evaporation. The maximum lysimeter-measured fluxes during these days were less than 70 W m^{-2} while the standard error of estimate of the model (SEE) was 18 W m^{-2} . This leads to the relatively small correlation between model estimation (E_{adj}) and lysimeter measurements (E_{wl}).

Several precipitation events occurred between January 3 and 10, 1991, creating a uniform moisture condition throughout the region. In addition, since most of the surrounding fields were bare ($\sim 10\text{--}15 \text{ km}$) the evaporation measured at the research site was representative of the Davis valley area. On Julian day 017, 1991, the average moisture content in the top 30 cm was 33%. Because of an increase in the afternoon wind speed (5 m s^{-1}) large-scale advection was important even though the minimum air relative humidity could be considered high (52%). The increase in wind speed resulted in an increase of the surface shear stress, which escalated the drying power of the air. The regression analysis indicated good correlation ($r^2 = 0.86$) between E_{wl} and $E_{(adj)}$ with a regression slope close to unity (1.04) and a standard error of estimate (SEE = 18.4 W m^{-2}) (see Figure 2i).

The 9 days combined together are indicative of the overall performance of the adjusted advection-aridity model (see Table

TABLE 2. Linear Regression Analysis for $E_{wl} = A E_{adj} + B$ and $E_{wl} = A E_{adj}$

Julian Day	r^2	SEE, W m^{-2}	A	B W m^{-2}
257-90	0.94	39.54	0.91	5.98
			0.93	0
288-90	0.71	26.63	0.75	7.98
			0.82	0
297-90	0.92	29.21	0.96	11.54
			1.01	0
324-90	0.67	32.83	0.94	-2.10
			0.92	0
338-90	0.32	19.44	0.64	-5.00
			0.51	0
340-90	0.56	14.56	0.65	0.83
			0.66	0
343-90	0.64	15.98	0.74	5.34
			0.83	0
351-90	0.82	22.51	1.16	-2.22
			1.14	0
017-91	0.86	18.40	1.05	-0.64
			1.04	0
Nine days Combined	0.90	26.10	0.94	0.95
			0.95	0

2 and Figure 3). The regression line ($E_{wet} = 0.94E_{adj} + 0.95$) and the coefficient of determination ($r^2 = 0.90$) indicate on average the adjusted advection-aridity model is robust for a range of surface moisture conditions and atmospheric forcing.

CONCLUSIONS

The simple complementary model presented, based on the Bouchet [1963] hypothesis and the advection-aridity evaporation formulation, is found to be reliable for evaporation estimation on a short time step (20 min). The adjustment of the apparent sensible heat flux (H_p) in the reference wet surface evaporation model [Parlange and Katul, 1991] improved the capability of the advection-aridity evaporation model to account for advection. The model computes the actual evaporation rate where the aridity of the land surface is deduced from the atmosphere so that no site specific calibration is necessary.

The model is compared with lysimeter evaporation measurements from a bare soil surface for a wide range of soil moisture and atmospheric stabilities. Overall the model predictions were well correlated with the lysimeter measurements ($r^2 = 0.90$) with a standard error of estimate of 26 W m^{-2} which is comparable to the error in the lysimeter measurement.

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