Estimates of Evapotranspiration with a One- and Two-Layer Model of Heat Transfer over Partial Canopy Cover

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ABSTRACT

One of the applications of remotely sensed surface temperature is to determine the latent heat flux \( LE \) or evapotranspiration \( ET \) from field to regional scales. A common approach has been to use surface–air temperature differences in a bulk resistance equation for estimating sensible heat flux, \( H \), and to subsequently solve for \( LE \) as a residual in the one-dimensional energy balance equation. This approach has been successfully applied over uniform terrain with nearly full, actively transpiring vegetative cover; however, serious discrepancies between estimated and measured \( ET \) have been observed when there is partial canopy cover.

In an attempt to improve the estimates of \( H \) and as a result compute more accurate values of \( ET \) over partial canopy cover, one- and two-layer resistance models are developed to account for some of the factors causing the poor agreement between computed and measured \( ET \).

The utility of these two approaches for estimating \( ET \) at the field scale is tested with remotely sensed and micrometeorological data collected in an arid environment from a furrowed cotton field with 20 percent cover and a dry soil surface. The estimates of \( LE \) are compared with values measured using eddy correlation and energy balance methods. It is found that the one-layer model generally performed better than the two-layer model under these conditions; but only when using a bluff-body correction to the resistance based on a conceptual model of heat and water vapor transfer at the surface taking place by molecular diffusion into Kolmogorov-scale eddies. The empirical adjustment to the surface resistance with the one-layer approach assumed to be applicable for a fairly wide range of conditions was found to be inappropriate. This result is attributed to the significant size of the furrows relative to the height of the vegetation.

Furthermore, a sensitivity analysis showed that the one-layer model with the empirical adjustment for the resistance was significantly affected by the changes in the surface roughness, whereas the physically based bluff-body correction was relatively insensitive to these variations. For the two-layer model, a large change in the input variable for computing soil evaporation had a relatively small impact on the computed fluxes while a significant change in the leaf area index appeared to amplify the deviations between measured and modeled \( LE \)-values.

1. Introduction

One of the applications of remote sensing data in the thermal infrared spectrum is to determine the skin temperature of the earth’s surface. This information has been incorporated in various models for computing the evapotranspiration \( (ET) \), which is usually the second largest quantity (precipitation being first) in the hydrologic cycle. A review of many of these approaches is given by Carlson (1986) and Jackson (1985).

This paper discusses a method whereby \( ET \) is solved as a residual in the one-dimensional surface energy balance equation, viz.,

\[
LE = Rn - G - H.
\]

In Eq. (1) \( LE \), \( Rn \), \( G \), and \( H \) are the latent heat flux, net radiation, soil heat flux, and sensible heat flux, respectively. Energy associated with photosynthesis or stored in the vegetation and in the canopy air space are neglected.

Estimates of \( Rn \) and \( G \) either come from micrometeorological measurements or from a combination of meteorological and remote sensing data (e.g., Jackson et al. 1987). In fact, recent developments in the latter approach may provide estimates of \( Rn \) and \( G \) with primarily remotely sensed information (Jackson et al. 1985; Clothier et al. 1986). Attempts, however, to provide as simplified parameterization for estimating sensible heat flux have not been as successful because of the relatively strong dependence of the transfer coefficient on various surface and meteorological conditions (e.g., Seguin and Itier 1983). Consequently, computing reliable values of \( H \) represents the most formidable obstacle in the residual method [i.e., Eq. (1)] for calculating satisfactory estimates of \( ET \) from hourly to daily time scales and over length scales from 100 m to 10 km.

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With a surface–air temperature difference, the transfer coefficient is normally represented by a resistance analogous to Ohm’s Law (Monteith 1973):

$$H = \rho C_p(T_s - T_a)/R_{ah}$$  \hspace{1cm} (2)

where $\rho$ is the air density (kg m$^{-3}$), $C_p$ the specific heat at constant pressure (J kg$^{-1}$ K$^{-1}$), $T_s$ the remotely sensed surface temperature (K), $T_a$ the air temperature (K) at some height z, and $R_{ah}$ the resistance to heat transfer (s m$^{-1}$). Equation (2) is a bulk transfer equation, and consequently it treats the vegetation-substrate surface as one layer. By using $T_s$ obtained from a thermal infrared sensor in (2) one implicitly assumes that the thermometric observation is a representative temperature for this idealized surface (see Fig. 1). Thus, it follows that the temperature difference ($T_s - T_a$) is assumed proportional, under the given meteorological conditions, to the rate of vertical turbulent transport of sensible heat. This single layer approach has been evaluated by comparing thermometric observations of $T_s$ with an “aerodynamic temperature,” $T_o$, determined by having measurements of $H$ and $T_a$, and estimates of $R_{ah}$ by invoking surface layer similarity. For full canopy cover Choudhury et al. (1986) and Huband and Monteith (1986a,b) found $T_s < T_o$ by 1 to 2 K on average under unstable conditions. Huband and Monteith (1986a) attributed most of the disagreement to the choice of emissivity and reflectivity of the vegetation. However, view angle of the sensor as well as canopy geometry and the canopy temperature profile will also significantly affect the value of $T_s$ (Huband and Monteith 1986a; Kimes 1983). Furthermore, more recent results indicate that the deviations between $T_s$ and $T_o$ are appreciably larger for partial canopy cover conditions (Stewart et al. 1988).

Besides the problems with estimating the appropriate surface temperature to be used in Eq. (2), several parameterizations leading to the expression for computing $R_{ah}$ [i.e. Eq. (3)] have been challenged. For example, Raupach (1979) and Hicks et al. (1979) demonstrated that the displacement heights for heat and water vapor differed from the value for momentum. Paw U and Meyers (1987), using a second-order closure model, showed displacement height for heat varied with stability. Denmead (1984) questioned the utility of using Eq. (2), since experimental evidence in a forest showed that the source/sink distributions of heat and water vapor vary diurnally as well as seasonally. Thus, neither the displacement height nor roughness length for heat and water vapor is uniquely defined. The existence of these anomalies is also supported by the fact that the resistance in (2) is inversely proportional to the eddy diffusivity for heat transfer integrated from a level inside the canopy to some height in the surface layer. Experimental data as well as higher order closure models show that the diffusivity approach is inappropriate inside the canopy (Denmead and Bradley 1987; Meyers and Paw U 1987). As a result the lower limit of integration producing $R_{ah}$ in (2) may not be a quantity that can be ascertained from gradient diffusion theory.

In light of the drawbacks for using Eq. (2), it would be difficult to support the use of a single layer model for partial canopy cover where the substrate as well as the vegetation significantly affect the distribution of sources/sinks of latent and sensible heat flux and the overall energy balance in (1). In addition, partial cover conditions where the vegetation is not randomly distributed (e.g., row crops) may seriously limit a one-dimensional treatment of the surface (Arkin and Pitter 1974; Graser et al. 1987). However, it can be ar-

![FIG. 1. Schematic diagram illustrating single layer model conceptualization for evaluating the surface energy balance. The symbols $T_{soil}$ and $T_c$ are the soil and canopy temperatures that make up the composite surface temperature obtained from the thermal infrared sensor (see text).](image-url)
gued that if model parameters are horizontally averaged over area scales in which persistent features exist in large numbers, a one-dimensional treatment may still yield satisfactory results (Shuttleworth and Wallace 1985). Moreover, under environmental conditions that promote strong vertical turbulent mixing (viz., convective conditions with moderate windspeeds) horizontal transport processes will be minimized; hence, a one-dimensional approach may still be applicable. There are two or more layer models separating vegetation and substrate (e.g., Choudhury and Monteith 1988; Sellers et al. 1986; Shuttleworth and Wallace 1985; van de Griend and van Boxel 1989); yet, these models also make use of the eddy diffusivity approximation inside the canopy. But model simulations and comparison with field measurements and higher order closure models suggest that the increased complexity in the parameterizations of the flux-gradient relationships inside the canopy results in the eddy diffusivity approach being a reasonable approximation (e.g., Sellers and Dorman 1987). This result comes from the fact that the sensitivity of model calculations of total turbulent fluxes to relatively large changes in the magnitude of the aerodynamic resistances inside the canopy air space appears to be rather small (e.g., Shuttleworth and Wallace 1985).

Still, it may be difficult to employ many of these models operationally because of the relatively large number of input parameters. Consequently, this paper investigates the applicability of a one- and two-layer energy balance model in which the number of input parameters is kept to a minimum so that both can be operational. This work follows others who have recognized the need for operational ET models (e.g., Gurney and Camillo 1984).

Estimates of LE using the single-layer approach and a two-layer model that partitions the energy between soil/substrate and the vegetation will be compared to eddy correlation/energy balance measurements. With the one-layer approach, a comparison is performed between a composite surface temperature \( T_s \), estimated with ground-based thermometric observations and an aerodynamic temperature, \( T_0 \), computed with profiles of wind speed, \( u \), and air temperature in the surface layer. An empirical expression for correcting the estimate of \( R_{ab} \) in (2), using \( T_s \) and \( u \) and calibrated over a different site (Kustas et al. 1989b), is tested and compared to an analytical expression for bluff-rough surfaces (e.g., furrowed field) from Brutsaert (1975).

In the two-layer model, resistances to heat transfer from the soil/substrate and vegetation will only be considered since \( LE \) will be solved using the residual approach [i.e., Eq. (1)]. Hence, estimates of the resistances to vapor transport from the soil and vegetation are not required. The partitioning of latent and sensible heat flux between the soil and vegetation as computed by the model will be investigated. Finally, a sensitivity analysis of both models will be performed. For the one-layer model, changes in the aerodynamic roughness parameters will be assessed. In the two-layer approach, adjustments in the contribution to \( LE \) from the soil and to the value of the leaf-area index will be investigated.

2. One- and two-layer model formulations

The resistance to heat transfer in the single-layer model has typically been determined by Monin-Obukhov surface layer similarity theory, which yields an expression of the form

\[
R_{ab} = \left[ \ln \left( \frac{z - d_0}{z_{0m}} \right) + \ln \left( \frac{z_{0m}}{z_{ah}} \right) - \psi_h \right] \times \left[ \ln \left( \frac{z - d_0}{z_{0m}} - \psi_m \right) \right] / k^2 u. \tag{3}
\]

The symbols \( \psi_h \) and \( \psi_m \) are stability correction functions of heat and momentum, \( d_0 \) and \( z_{0m} \), the displacement height and roughness length for momentum, \( z_{ah} \), the roughness height for heat, \( k \) \((\approx 0.4)\) Von Karman's constant, and \( u \) the wind speed at level \( z \). The distinction made in roughness lengths for heat and momentum is to account for the difference in transfer processes in close proximity to the obstacles (Thom 1972). This has led to empirical and theoretical treatment of the relationship between \( z_{ah} \) and \( z_{0m} \), historically expressed in a dimensionless form; viz., \( \ln (z_{0m}/z_{ah}) \) or \( kB^{-1} \) (Chamberlain 1968). For vegetative surfaces experimental data and physically based models suggest a constant \( kB^{-1} \sim 2 \); however, it has been shown analytically to be sensitive to foliage structure (Massman 1987). For bluff-rough elements (e.g., a furrowed or plowed field) Brutsaert (1982) reviews both theoretical and experimental evidence that suggests the magnitude of \( kB^{-1} \) can vary from order 2 to 10. A recent analysis with (2) over sparse canopy cover required that \( kB^{-1} \) be a function of the thermometric surface temperature observed from a nadir-viewing, thermal infrared sensor to obtain satisfactory results (Kustas et al. 1989a). Since the present analysis considers sparse cover, the equation for \( kB^{-1} \) developed by Kustas et al. (1989a) was adopted; i.e.,

\[
kB^{-1} = 0.17u(T_s - T_0). \tag{4a}
\]

However, because the field contained relatively large furrows compared to the height of the vegetation (see below) an analytical equation for bluff-rough obstacles was also employed (Brutsaert 1975):

\[
kB^{-1} = 2.46(z_{0s})^{1/4} - 2. \tag{4b}
\]

The symbol \( z_{0s} \) is a roughness Reynolds number \( (= u_* z_{0s}/\nu) \) where \( \nu \) is the kinematic viscosity.

In this paper a two-layer model is developed in which only the partitioning of net radiation and soil and sensible heat flux is explored since the utility of the residual approach [i.e, Eq. (1)] is to be tested. From Fig. 2 the energy balance of the vegetation is

\[
LE_c = Rn_c - H_c \tag{5}
\]
where \( Rn_c \) is the net radiation absorbed by the canopy and the subscript \( c \) pertains to the canopy. For the soil/substrate, the equation reads

\[
LE_s = Rn_s - H_s - G
\]

where subscript \( s \) pertains to the substrate. From Fig. 2, it follows that \( Rn = Rn_c + Rn_s \), \( H = H_c + H_s \), and \( LE = LE_c + LE_s \).

The mean canopy resistance to heat transfer \( r_{ac} \) was evaluated by first considering the leaf boundary layer resistance, \( r_b \). An equation from Jones (1983) was utilized:

\[
r_b = A(w/u)^{1/2}
\]

where \( A \) is a constant of the order 90 s\(^{1/2}\) m\(^{-1}\) (Goudriaan 1977); \( w \), a characteristic length scale taken as the average leaf width (\( \sim 0.05 \) m); and \( u \), the local wind speed at some level inside the canopy. In order to calculate a mean resistance per unit leaf area, \( r_b \), the wind profile inside the canopy had to be estimated. An exponentially decreasing \( u \) with depth was employed, which previous studies have shown to be a reasonable approximation (Brutsaert 1982):

\[
u = u_h \exp(-\beta(1-z/h)).
\]

The symbol \( h \) is the canopy height (\( \sim 0.3 \) m), \( u_h \) the windspeed at \( h \), and \( \beta \) the extinction factor evaluated with the equations of Goudriaan (1977); see the Appendix for details. Values of \( u_h \) were calculated with the wind profile data and roughness parameters from Kustas et al. (1986b), and assuming near-neutral conditions due to roughness sublayer effects (Garratt 1980). With the leaf area index, LAI—assumed uniform over the canopy height—Eqs. (7) and (8) were
numerically integrated to obtain $\bar{r}_b$, the mean resistance per unit leaf area. Consequently, the resistance per unit ground area $r_{ac}$ was computed by the formula (Shuttleworth 1976):

$$ r_{ac} = \bar{r}_b / \text{LAI}. $$

(9)

Values of $Rn_t$ and $Rn_c$ were determined using Beer's law, assuming $Rn$ is proportional to the global shortwave radiation flux and neglecting longwave radiation in the canopy caused by vertical temperature gradients. For $Rn_t$, this yields an exponentially decreasing function

$$ Rn_t = Rn \exp(-\alpha \text{LAI}), $$

(10)

and hence,

$$ Rn_c = Rn(1 - \exp(-\alpha \text{LAI})). $$

(11)

Experimental observations for various crops (e.g., sorghum, cotton, soybean, and corn) compiled by Ritchie (1972) suggest the attenuation of $Rn$ given by (10) and (11) is a satisfactory parameterization (see also Ross 1975). This has more recently been supported by a surface temperature-based energy balance model that employs (10) in a net radiation-based empirical model for estimating $G$ under a growing wheat crop (Choudhury et al. 1987).

The attenuation coefficient, $\alpha$, is primarily a function of the leaf angle distribution and solar elevation (Monteith 1973). Adjustment to $\alpha$ for diffuse radiation was neglected. The value of $\alpha$ was estimated by taking a spherical leaf distribution exposed at a sun angle $\theta$, yielding $\alpha = 0.5 \csc \theta$.

From Fig. 2 and Eqs. (6)–(11) the estimate of soil evaporation and transpiration can be expressed by the energy balance for the two components:

$$ LE_s = Rn \exp(-\alpha \text{LAI}) - G - \rho C_p(T_{soil} - T_0)/r_{as}, $$

(12a)

$$ LE_c = Rn(1 - \exp(-\alpha \text{LAI})) - \rho C_p(T_c - T_0)/r_{ac}. $$

(12b)

The value, $T_0$, could not be evaluated independently with the above formulation. However, with some algebraic manipulation of the resistances in Fig. 2 and a few approximations, the following expression for $H_c$ was derived (Smith et al. 1988) and inserted into (12b):

$$ H_c = [\rho C_p(T_c - T_a) - H_s R_a]/(R_a + r_{ac}). $$

(13)

The resistance, $R_a$, replaces $R_{ab}$ in the single-layer model and is defined as a ball aerodynamic resistance. Hence it assumes a non-equivalence in the value of $T_a$ and $T_0$ as represented in Figs. 1 and 2, respectively. The magnitude of $R_a$ was evaluated with the formula from Mahrt and Ek (1984) for unstable conditions:

$$ R_a = (\ln((z - d_0 + z_{om})/z_{om})/k)^2(1 - 15 R_i)/(1 + C(-R_i)^{1/2})^{-1}/u $$

(14)

with

$$ R_i = (9.8/T_a)(T_a - T_s)(z - d_0)/u^2, $$

and

$$ C = 75k^2((z - d_0 + z_{om})/z_{om})^{1/2}/(\ln((z - d_0 + z_{om})/z_{om})). $$

An estimate of $LE_s$ came from the expression

$$ (Rn_t - G)/(1 + B_{os}) $$

(15)

where $B_{os}$ is the Bowen ratio for the substrate. This input variable was assigned a constant value of six from the simulation results of a drying soil by Choudhury and Monteith (1988) and a value of three for the sensitivity analysis (see the Appendix). Thus, the total latent heat flux of the substrate and canopy was estimated by combining (5), (6), (13), (14), and (15), yielding

$$ LE = Rn_c - \{\rho C_p(T_c - T_a) - [B_{os}(Rn_t - G)/(1 + B_{os})]R_a\}/(R_a + r_{ac}) $$

$$ + (Rn_t - G)/(1 + B_{os}) $$

(16)

where $Rn_c$ and $Rn_t$ are estimated by (10) and (11), respectively.

In summary, $LE$ was solved with the one-layer model using Eqs. (1)–(4a), $LE_{1L}$ (a), and (1)–(4b), $LE_{1L}$ (b). The two-layer model estimated $LE$ with Eq. (16), $LE_{2L}$.

3. The data

The data for this study were collected over a furrowed cotton field, 1500 m east–west × 300 m north–south, located in Maricopa Farms in central Arizona (33.08°N, 111.98°W) from 10 June 1987 day of year (DOY) 161 to 14 June 1987 DOY 165. A detailed discussion of the micrometeorological instrumentation and agronomic measurements made in the cotton field (field 28) is given by Kustas et al. (1989b,c). Briefly, the agronomic measurements gave a crop height of 0.3 m spaced 1 m apart and a furrow depth of about 0.2 m with rows running north–south. The percentage covered was estimated to be approximately 20 percent with a LAI ≈ 0.4. The last irrigation occurred on or about 23 May 1987 DOY 143. Hence, during the experimental period the soil surface was dry. Profiles of wind speed and temperature were determined at five levels above the surface (i.e., 1.2 m, 1.4 m, 1.8 m, 2.4 m, and 3.0 m above the furrow bottom) in the center of the field. Fluxes of latent and sensible heat were measured by the Bowen ratio/energy balance and eddy correlation method with instruments located within 150 m of each other and the profile mast. Differences in the half-hourly values of $H$ and $LE$ given by the two methods were investigated by Kunkel et al. (1988), un-
published manuscript). In general, the Bowen ratio/energy balance method tended to give higher values of \( LE \). However, Kustas et al. (1989b) found satisfactory agreement between eddy correlation measurements of \( H \) and estimates using the profile data in an eddy diffusivity approach. Consequently, it was felt that the eddy correlation measurements of \( H \) and \( LE \) computed by (1) gave the most reliable flux estimates. These values were used for comparison with model results.

A thorough discussion and analysis of the ground- and airborne-based thermometric observations (nadir-looking) is given in Kustas et al. (1990). One of the findings significant to this study was that the composite surface temperature given by the measurement scheme for the ground-based observations of the sunlit and shaded soil temperatures and the canopy temperature were comparable to values obtained for one of the days by an aircraft flying at an altitude of 150 m. Since only one day of airborne data is available, the larger quantity of the ground-based radiometric data, collected for field 28 between DOY 161–165, will be used.

Nineteen thermometric observations were made between DOY 161–165 that could be used in this analysis. A weighted average of the one-half hour-means of the terms in (1) and the meteorological data, \( u \) and \( T_o \), were determined using the mean time of the thermal measurements. Table 1 lists some of the pertinent micrometeorological and remote sensing data for this study, while Fig. 3 is a plot of the energy balance components versus time for all days. A reduced dataset, 10 out of the 19 observations, had optimal windspeed and fetch conditions. A comparison of \( LE \) estimated by the different formulations with the reduced dataset will be given.

### Table 1. Micrometeorological and remote sensing data with \( u \) and \( T_o \) given for the 3 m level (see text).

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<th>DOY</th>
<th>Time (decimal)</th>
<th>( R_n ) (W m(^{-2}))</th>
<th>( G ) (W m(^{-2}))</th>
<th>( H ) (W m(^{-2}))</th>
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* Observations with optimal fetch and wind speed conditions.

### 4. Analysis and results

As a means of illustrating the difficulties associated in using \( T_s \) with a one-dimensional surface layer formulation over partial canopy cover, Eq. (2) is inverted to solve for \( T_0 \) with values of \( H \) provided by the eddy correlation measurements, \( k_B^{-1} \) assumed a constant (\( \sim 2 \)), and all roughness parameters needed in (3) obtained from the results of Kustas et al. (1989b). As Fig. 4 illustrates, the values of \( T_0 \) were significantly less on average (\( \sim 10^\circ \)C) than the composite surface temperatures \( T_c \) measured with thermal infrared sensors. Inappropriate estimates of the roughness parameters (viz., \( z_{om} \) and \( d_0 \)) can significantly contribute to the differences between \( T_c \) and \( T_0 \). However, estimates of \( z_{om} \) and \( d_0 \) from Kustas et al. (1989b) are in fairly good agreement with the results from previous micrometeorological studies on cotton (Stanhill and Fuchs 1968; Stanhill 1976). This figure also shows that the deviation of \( T_0 \) from \( T_s \) grew as the magnitude of \( T_s \) increased. In fact, when the difference (\( T_s - T_0 \)) is plotted versus time (see Fig. 5) there is an apparent functional relationship with the amount of radiation received at the surface. This is more clearly evident when (\( T_s - T_0 \)) is regressed with the percent of sunlit soil computed from Kustas et al. (1990); a significant correlation of 0.86 is obtained. At first glance one may conclude that the magnitude of (\( T_s - T_0 \)) is mainly caused by sunlit soil temperatures strongly biasing the value of the composite temperature obtained by the sensor (Kustas et al. 1990). But there may be a more fundamental reason relating to the efficiency of heat relative to momentum transfer that is pronounced under these furrowed, sparse canopy-covered conditions (see below).
A major advantage of using a two-layer model is being able to estimate the contribution of the soil/substrate and the vegetation to the fluxes $H$ and $LE$ observed in the surface layer. This is illustrated in Fig. 6, which is a plot of the fluxes $H_a$, $LE_a$, $LE_v$, and $H_t$ versus time. The sign of $H_t$ shows that the vegetation not only converts radiant energy into latent heat, but also uses the energy in the surrounding air to transform water into water vapor. The magnitude of $H_t$ implied that nearly half of the latent heat released by the vegetation comes from the surrounding air, whose major source of sensible heat comes from the soil surface. Adveded energy well above the surface–atmosphere interface was not quantified; but, from the results of Kustas et al. (1989b) it probably can be neglected, especially for the wind directions with optimal fetch conditions.

How well the one- and two-layer models parameterize the energy exchange across the soil-vegetation–atmosphere interface was assessed by comparing esti-

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**Fig. 3.** Values of the energy balance components for all days (i.e., 19 observations) used in the study. The symbols represent $R_n$ (□), $G$ (+), $H$ (○), and $LE$ (△).

**Fig. 4.** Comparison of $T_s$, the composite surface temperature measured with an infrared thermometer versus the aerodynamic temperature $T_a$ estimated with Eqs. (2) and (3). The line represents perfect agreement.
Fig. 5. The difference between aerodynamic and thermometric temperature ($T_s - T_a$) versus local time of day.

Fig. 6. Plot of the partitioning of latent and sensible heat flux between the soil/substrate and canopy versus time of day. The symbols represent $H_s$ (□), $H_c$ (+), $LE_s$ (○), and $LE_c$ (△).

mates of $LE$ with the measured values. Figure 7 reveals for the one-layer model, $LE_{1L}$ (a) values are consistently and significantly smaller than measured $LE$ with appreciable scatter while $LE_{2L}$ (b) values are only slightly larger on average than those measured with much less scatter. The parameterizations used in the two-layer model produced appreciable scatter with $LE_{2L}$ values mostly underestimating measured $LE$.

Table 2 summarizes the results of the one- and two-layer models as the root mean square error (RMSE) between computed and observed values of latent heat flux for all the observations and for the reduced dataset. The results listed in Table 2 suggest that the one-layer model provides the most accurate estimates of $LE$ as long as $R_{sh}$ is appropriately adjusted for the surface conditions. For this particular case, it appears that with the nadir thermometric observations providing a composite surface temperature, the bluff-rough formulation
for $k B^{-1}$ [i.e. Eq. (4b)] is most applicable. This conclusion seems physically plausible given the low vegetative cover and the fact that the furrows were of relatively large size (i.e., $\sim 1/2$ the height of the vegetation) so that they significantly augmented momentum transfer relative to heat transfer resulting in a large $k B^{-1}$ required in (3).

However, an off-nadir-view angle of the thermal infrared sensor (say from aircraft and satellite altitudes), will result in significant changes in $T_s$ and consequently in the adjustment needed to $R_{ah}$. Moreover, the expression developed by Kustas et al. (1989a) was not applicable to this situation. Thus, the ability to adjust $R_{ah}$ with simple formulations like Eq. (4a) may not be feasible. On the other hand, two-layer models require an estimate of canopy temperature, which from a composite scene of soil and vegetation, may be difficult to obtain. Nonetheless, recent advances in multilayer models show some promise in this area (van de Griend and van Boxel 1989).

A sensitivity analysis of the one- and two-layer model estimates of $LE$ was performed by changing the values of several input parameters that were deemed either to have a significant affect on model calculations of $LE$ or whose value had a relatively high degree of uncertainty. Different values for the input variables applicable to the surface conditions of this study were estimated from the literature. The computed values of $LE$ were then evaluated by calculating the RMSE with the measured values.

For the one-layer model, the aerodynamic roughness parameters, $z_{om}$ and $d_0$, were estimated with the empirical equations of Hatfield (1989). His Table 1 equations developed for the cotton growth cycle, yields $z_{om} \sim 0.017$ m and $d_0 \sim 0.10$ m for a canopy height to planting width ratio of $\sim 0.3$. These estimates produce a RMSE with $LE_{1L}$ (a) values that are almost double the Table 2 results, while $LE_{1L}$ (b) values give essentially the same RMSE (see Table 3). Clearly, these results are caused by the fact that $k B^{-1}$ is insensitive to changes in roughness when estimated with (4a) while it is a function of $z_{om}$ in (4b). The use of significantly different roughness parameters had a small effect on the two-layer model estimates of $LE$, which is supported by other multilayer studies (e.g., van de Griend and van Boxel 1989).

In the two-layer model, a sensitivity analysis was performed by adjusting two of the input variables. One variable, $B_{om}$, was adjusted because its value is shown in the Appendix to have a high degree of uncertainty. The other parameter, LAI, was changed because it is likely to have the largest impact on the model calculations of $LE$.

From the analysis in the Appendix it is conceivable that the value of $B_{om}$ is roughly one-half (i.e., $B_{om} \sim 3$) of what was used to produce the results in Table 2. With $B_{om} = 3$, the two-layer model computations of $LE$, $LE_{2L}$, are summarized in Table 3. The RMSE for both datasets is reduced by about 30 percent and 20 percent; yet, this is a rather marginal improvement when considering the relatively large change in the value of $B_{om}$ (i.e. 200 percent).

In the two-layer model formulation, an accurate estimate of the LAI is essential since it is directly pro-
portional to the bulk canopy resistance of heat transfer [cf. Eq. (9)], and indirectly influences the estimation of the canopy attenuation of \( R_n \). If this model is to be used operationally over large areas, LAI will need to be evaluated using remote sensing techniques (e.g., Choudhury 1987). However, for partial cover the soil background reflectance can significantly affect this estimate. For low zenith angles, the background reflectance of relatively bright soils, which is maximized by a dry soil surface, tends to reduce ratio-based vegetation indice-values used to evaluate LAI (Hue et al. 1985). Consequently, it is speculated for the dry soil conditions during this study that LAI estimated remotely would be around one-half the agronomic value. Albeit this is somewhat subjective, it will illustrate the importance of a reliable estimate of LAI. Table 3 lists the RMSE with the two-layer model using a LAI = 0.2. The resulting RMSE of the order 10^2 W m\(^{-2}\) substantiates the claim regarding the importance of having appropriate estimates of LAI\(^\dagger\). In fact, it is postulated that this sensitivity at low vegetative cover exists for essentially all multilayer energy budget models.

5. Conclusions

There was a significant improvement in estimating \( LE \) with the two-layer approach when compared to \( LE_{EL} \) (a). However, it fell short of being more accurate than the one-layer approach with the theoretical equation for \( k_B \); although for the reduced dataset the differences are rather small. Of course, as pointed out in the model formulation, only the resistances to heat and momentum transfer are considered. When Penman–Monteith equations for the water vapor transport from the vegetation and the soil are included, model sensitivity in estimating \( LE \) to large changes in \( r_{ve}, r_{va} \), and \( R_n \) is shown to be relatively small (Shuttleworth and Wallace 1985). Including water vapor requires estimating soil and stomatal resistances to water vapor transfer. This may be difficult to specify, especially under partial canopy conditions, without extensive ground-based measurements (e.g., Smith et al. 1988). In addition, from high-altitude aircraft or satellite altitudes, separation of vegetation and soil temperatures appears to be a rather formidable task. Finally, the sensitivity analysis revealed that the two-layer model flux calculations are strongly affected by LAI estimates, especially at low vegetative cover. It is speculated that since the bulk canopy resistance to water vapor transfer is usually computed as a ratio of leaf stomatal resistance to LAI [cf. Eq. (9)] that multilayer Penman–Monteith models will also be sensitive to errors in LAI estimates. This area of research should be actively pursued (e.g., van de Griend and van Boxel 1989) if two- and multilayer models can be made operational with satellite data. More field experiments with a larger number of observations over different surface types like HAPEX (Andre et al. 1988) and FIFE (Sellers et al. 1988) are necessary to evaluate the utility of the one-dimensional approach using the residual method and Penman–Monteith approach in one-, two- and multilayer models.

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APPENDIX

Estimation of Model Parameters \( B \) and \( B_{os} \)

This appendix contains the equations used for estimating the extinction coefficient or factor, \( \beta \), in Eq. 1 The value of LAI was also doubled in the model (i.e. LAI \( \sim 0.8 \)). This produced a RMSE of the order 70 W m\(^{-2}\) suggesting that underestimating LAI in model calculations results in significantly larger errors compared to overestimating LAI under these field conditions.

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Table 3. Values of the RMSE for one- and two-layer model estimates of LE using different values for several input variables (see text). Results are given with all 19 observations and the reduced dataset of 10.

<table>
<thead>
<tr>
<th>Model</th>
<th>Changes to input parameters</th>
<th>RMSE with 19 observations (W m(^{-2}))</th>
<th>RMSE with 10 observations (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>( LE_{EL} ) (a) ( z_{tan} = 0.17m ) ( d_0 = 0.10m )</td>
<td>161 151</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( LE_{EL} ) (b) ( z_{tan} = 0.17m ) ( d_0 = 0.10m )</td>
<td>27 27</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( LE_{FL} ) ( B_{so} = 3 )</td>
<td>34 26</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( LE_{FL} ) LAI = 0.2</td>
<td>130 117</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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Table 2. Values of the root mean square errors (RMSE) of one- and two-layer model estimates of \( LE \) with all 19 observations and the reduced dataset of 10.

<table>
<thead>
<tr>
<th>Model</th>
<th>RMSE with 19 observations (W m(^{-2}))</th>
<th>RMSE with 10 observations (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>( LE_{EL} ) (a)</td>
<td>85 74</td>
<td></td>
</tr>
<tr>
<td>( LE_{EL} ) (b)</td>
<td>24 24</td>
<td></td>
</tr>
<tr>
<td>( LE_{FL} )</td>
<td>48 33</td>
<td></td>
</tr>
</tbody>
</table>

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\( \dagger \) The value of LAI was also doubled in the model (i.e. LAI \( \sim 0.8 \)). This produced a RMSE of the order 70 W m\(^{-2}\) suggesting that underestimating LAI in model calculations results in significantly larger errors compared to overestimating LAI under these field conditions.
(8). In addition, details regarding how the values of the Bowen ratio for soil evaporation, $B_{0}$, [cf. Eq. (15)] are provided.

a. Calculation of the extinction coefficient

The extinction coefficient was estimated from the formulation presented by Goudriaan (1977). The expression for $\beta$ is

$$\beta = \left( \frac{c_{d} LD_{A} h}{2l_{m} i_{w}} \right)^{0.5}$$

(A1)

where $c_{d}$ is the drag coefficient of the canopy, $l_{m}$ the mean mixing length inside the canopy, $LD_{A}$ the leaf area index, $h$ the mean canopy height, and $i_{w}$ the relative turbulence intensity factor. The value of $l_{m}$ is taken to be equal to the mean distance between leaves, which is estimated by the following formula:

$$l_{m} = 2 \left( \frac{3w^{2}}{4nLD_{A}} \right)^{1/3}$$

(A2)

where $LD_{A} = LAI/h$ is the leaf density and $w$ is the leaf width. Goudriaan (1977) found in his analysis of the aerodynamic characteristics of grasses, maize, and coniferous forests that it was adequate to use $c_{d} \sim 0.2$ and $i_{w} \sim 0.5$. Hence, these values were also used in Eq. (A1).

An adjustment for $\beta$ due to nonneutral conditions was not performed because it was not considered significant relative to the effect of the furrows on the wind profile. Their influence was quantified in the calculation of $u_{*}$ [cf. Eq. (8)] and by using $h = 0.5$ m, which included the furrow height, in Eq. (A1). The resulting value of $\beta$ was approximately 0.5.

b. Determination of $B_{0}$

The agronomic data indicated that at the start of the experiment nearly three weeks had elapsed since the last irrigation. Therefore, during the field campaign the soil was in stage 2 drying or the soil-limiting evaporation stage (Philip 1957). The simulation results of Choudhury and Monteith (1988) suggest a $B_{0}$ value of order 6 under these conditions. The variability in this value was assessed by using the results of several studies that show that the soil-limiting daily evaporation rate is related to the square root of time (Gardner 1959; Ritchie 1972; Black et al. 1969):

$$E_{s} = \frac{1}{2} D t^{1/2}.$$  

(A3)

The coefficient $D$ (mm d$^{-0.5}$) is a function of soil hydraulic properties and environmental conditions (see below), $E_{s}$ the daily evaporation rate (mm d$^{-1}$), and $t$ the time in days after reaching stage 2 drying conditions, which is also affected by the same factors as $D$. A review of experimental studies by Ritchie (1972) suggests that $D$ has a relatively narrow range from about 3 to 5 mm d$^{-0.5}$ going from a clay loam to a sand. However, Jackson et al. (1976) found that $D$ not only depends on soil properties but also on the time of year. They determined a range for $D$ from 4 mm d$^{-0.5}$ in the winter months to 8 mm d$^{-0.5}$ in the summertime (see their Fig. 2). Consequently, two values were chosen, namely $D = 4$ and $8$ mm d$^{-0.5}$.

Furthermore, the transition from energy-limiting (stage 1) to soil-limiting (stage 2) depended on the season. Since the last irrigation occurred about three weeks before the summer experiment, $t$ was approximated to be of order 15 days (see Jackson et al. 1976, Fig. 3). For $D = 4$ mm d$^{-0.5}$, Eq. (A3) yields $E_{s} \sim 0.5$ mm d$^{-1}$ while for $D = 8$ mm d$^{-0.5}$, Eq. (A3) produces $E_{s} \sim 1$ mm d$^{-1}$. With half-hourly values of $R_{n}$ and $G$ summed over the daytime periods, Eq. (15) required $B_{0}$ to be of order 6 and 3 for $E_{s} \sim 0.5$ mm d$^{-1}$ and $E_{s} \sim 1$ mm d$^{-1}$, respectively.

REFERENCES


