A Comparison of a Hierarchy of Models for Determining Energy Balance Components over Vegetation Canopies

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ABSTRACT

Several methods for estimating surface energy balance components over a vegetated surface are compared. These include Penman–Monteith, Deardorff, and multilayer canopy (CANWHT) models for evaporation. Measurements taken during the 1991 DOE-sponsored Boardman Area Regional Flux Experiment over a well-irrigated, closed wheat canopy are used in the comparison. The relative performance of each model is then evaluated. It is found that the Penman–Monteith approach using a simple parameterization for stomatal conductance performs best for evaporation flux. The Deardorff model is found to have the best relative performance for sensible heat, while the CANWHT model gives the best results for net radiation and soil heat flux. The Priestley–Taylor model for evaporation and a resistance-analog equation for sensible heat flux are also tested.

1. Introduction

The complexity of surface–atmosphere interactions has led to the development of a hierarchy of numerical models that can determine one or more of the four energy balance components—net radiation $R_n$, soil heat flux $G$, sensible heat flux $H$, and latent heat flux $LE$—and there is ongoing debate on which methods are the most desirable to use in large-scale climate models. In general, we would expect that the model that incorporated the most of the physical and biological processes that govern the surface fluxes—from soil characteristics to plant physiology to micrometeorology—would perform the best. However, increasing the physics in a numerical scheme costs computer time and requires detailed knowledge of several parameters. A condition can easily be reached where errors introduced in parameterizations can overwhelm the accuracy the increased rigor might provide. As computers become faster, the demands on computer time are diminishing. However, we are still challenged with trying to develop models that not only can parameterize surface influences accurately but are general enough to be used globally.

As an example of classifying current numerical schemes, Shuttleworth (1991) recently discussed three types of models of increasing complexity for predicting evaporation over vegetated surfaces: “energy balance,” “single-source,” and “simulation.” Energy balance models such as the Priestley–Taylor equation (Priestley and Taylor 1972; hereafter PT) are typically characterized by simple equations and rely only on standard meteorological inputs to drive them. Single-source or “big-leaf” models like the Penman–Monteith approach (e.g., Thom 1975; hereafter PM), or those of Deardorff (1978; hereafter DF) and Shuttleworth and Wallace (1985; hereafter SW), treat the canopy and soil as a single- or two-layer system, and use submodels for the canopy and/or soil resistance. Finally, simulation models such as SPAM (Soil–Plant–Atmosphere Model) (Stewart and Lemon 1969; Sinclair et al. 1976) and CANWHT (multilayer canopy) (Baldocchi 1992) give, among other factors, a detailed treatment of the physiological response of the plant species at many levels within a canopy. They are considered the most rigorous.

Studies comparing the performance of different models have been previously done, most notably, those involved in the determination of moisture flux. Sinclair et al. (1976) compared the SPAM model, a simplified SPAM model, and a modified PM approach for determining transpiration. They found that a big-leaf model such as the PM gave transpiration rates within 10% of SPAM in conditions where there was appreciable moisture flux. Stannard (1993) compared the PM, SW, and PT models for determining moisture flux in semi-arid conditions, and found that SW and PT approaches performed equally well, while the PM scheme demonstrated its inadequacy in conditions of sparse vegetation and limited soil moisture. Many other studies have used simpler approaches such as the PT or PM (e.g., Beijers and Holtslag 1991) and obtained reasonable results. However, it is still not clear whether the simpler approaches are universal enough or whether

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those that are more rigorous are practical enough to be useful for incorporation into large-scale models. Further studies are needed to determine which classification of models are presently sufficient to give satisfactory results on a global scale.

At the Atmospheric Turbulence and Diffusion Division of the National Oceanic and Atmospheric Administration’s (NOAA) Air Resources Laboratory in Oak Ridge, Tennessee, we have been using evaporation models that fall in each of Shuttleworth’s three classifications mentioned above. In this paper, we describe and compare several surface energy balance models of differing physical sophistication in order to evaluate their performance for determining one or more components of the energy budget. We utilize measured surface fluxes and other data from a well-irrigated wheat crop to test and compare the accuracy of the model’s predictions. In addition, the performance of a simple resistance-analog approach for determining sensible heat flux is briefly described. We discuss the applicability of these models under different conditions and attempt to answer the question if a less-rigorous approach can be used to obtain reasonable estimates of the surface energy fluxes in large-scale atmospheric modeling.

2. Descriptions of the models

To understand the limitations inherent in different approaches, a brief description of each model used in this study is presented in the order of increasing complexity.

a. The Priestley–Taylor equation

The Priestley–Taylor equation originates from Penman’s (1948) formula for evaporation over a saturated soil surface:

\[ LE = \frac{\epsilon F_A}{\epsilon + 1} + \frac{A_qD}{\epsilon + 1}. \]  

(1)

The latent heat flux LE is a function of the available energy \( F_A = R_n - G \), the specific humidity deficit \( D = q_s(T) - q \), the normalized derivative of the saturation specific humidity \( \epsilon = (\lambda/c_p)(\partial q_s/\partial T) \), and an exchange coefficient \( A_q \). In the usual notation \( \lambda \) and \( c_p \) represent the latent heat of vaporization and specific heat of air at constant pressure, respectively, while \( q_s(T) \) and \( q \) are the saturated specific humidity and specific humidity at height \( z \) and temperature \( T \) above the surface. Priestley and Taylor (1972) proposed that in conditions of negligible advection, ample water supply, positive (upward) heat flux, and positive (upward) moisture flux (i.e., no condensation), this equation could be simplified by allowing the latent heat flux to be directly proportional to the equilibrium evaporation rate \( \epsilon F_A(\epsilon + 1)^{-1} \):

\[ LE = \alpha_{PT} \frac{\epsilon F_A}{\epsilon + 1}. \]  

(2)

where \( \alpha_{PT} \) is the constant of proportionality. Thus, in this equation, the latent heat flux \( LE \) is to a large extent only a function of temperature through \( \epsilon \) (not entirely true since \( c_p \) is a function of humidity as well) and available energy \( F_A \). The value of 1.26 proposed by Priestley and Taylor for \( \alpha_{PT} \) was obtained by averaging the results of comparisons between the equilibrium evaporation and latent heat flux measurements over both a saturated land surface and open water. From Eq. (1) it is evident that an absence of condensation (positive \( D \)) leads to the condition \( LE > \epsilon F_A(\epsilon + 1)^{-1} \). Further, from the surface energy budget, an upward heat flux (positive \( H \)) requires that \( LE < F_A \). It follows in equation (2), that these two constraints dictate a range for \( \alpha_{PT} \): 1 < \( \alpha_{PT} < [(\epsilon + 1)\epsilon^{-1}] \).

b. The Penman–Monteith model

Monteith (1964) modified Penman’s equation to consider vegetation in various states of water stress. By combining aerodynamic and stomatal resistance equations for moisture:

\[ r_{av} = \frac{\rho c_p}{\gamma} \frac{q(0) - q(z)}{\lambda E} \]

(3)

\[ r_s = \frac{\rho c_p}{\gamma} \frac{q(T)(0) - q(0)}{\lambda E}, \]

(4)

where \( \gamma = c_p/\lambda \) is the psychrometric constant, and making the assumption that

\[ q(T)(0) = q_s(T(z)) + \epsilon(T(0) - T(z)), \]

(5)

Eq. (1) could take on a more universal form

\[ LE = \frac{\epsilon F_A + \rho \lambda D/r_{all}}{\epsilon + (r_{av} + r_s)/r_{all}}. \]

(6)

Here, the aerodynamic resistance to heat \( r_{all} \) arises from utilizing a resistance-analog equation for sensible heat similar to Eqs. (3) and (4) in the derivation.

Qualitatively, LE represents the latent heat flux from a source–sink height \( z \), the so-called big-leaf approach. Equation (5) demonstrates a key assumption, namely, that the saturation specific humidity at a leaf surface can be approximated by an extrapolation using \( \epsilon \), the derivative of the saturation specific humidity. Another assumption that further simplifies Eq. (6) is that of equating aerodynamic resistances for water vapor and heat to that for momentum: \( r_a = r_{av} = r_{all} = r_{gm} = U/u_s \), where \( U \) and \( u_s \) are the mean wind and friction velocity, respectively.

Approximating the two scalar resistances to that of momentum has long been recognized as dubious and one could expect significant errors if this assumption was made in a prognostic equation in which there was only a single resistance involved [e.g., Eq. (3)]. However, although estimations could be made for values of \( r_{all} \) through, for example, the specification of a sur-
face roughness length for heat (see Beljaars and Holtslag 1991) it can be demonstrated that the Penman–Monteith equation is relatively insensitive to errors in specifying an aerodynamic resistance. Figure 1 shows the percent change in LE due to a percent change in one of the resistances or scales. Twenty percent deviations in the aerodynamic resistance typically produce changes in LE of approximately 2%. Thus, for this exercise, the aerodynamic resistance for heat is assumed to be equal to that of momentum. This allows Eq. (6) to be reduced to

$$\text{LE} = \frac{\varepsilon F_A + \rho \lambda D / r_s}{\epsilon + 1 + r_s / r_a}. \quad (7)$$

In this study, the bulk stomatal resistance $r_s$ is parameterized in a form similar to the algorithm proposed by Stewart (1988), Shuttleworth (1989), and Raupach et al. (1992):

$$r_s = \frac{r_{s \text{ min}}}{f_R(F_A)} f_D(D). \quad (8)$$

Here $r_{s \text{ min}}$ represents the minimum attainable resistance of the canopy, while $f_R$ and $f_D$ are nondimensional conductances representing the (bulk) influence of the available energy and humidity on the canopy system. These conductances have the form

$$f_R(F_A) = 1 - \exp\left(-\frac{F_A}{F_{A0}}\right), \quad (9)$$

$$f_D(D) = \left(1 + \frac{D}{D_0}\right)^{-1}, \quad (10)$$

and represent the increased stomatal resistance that results from decreased available energy and increased specific humidity deficit. The scales $F_{A0}$ and $D_0$ as well as $r_{s \text{ min}}$ are indicative of the crop and are considered constant when working over short periods relative to its growing cycle. It should be noted that in general the primary factors affecting stomatal resistance also include the temperature and age of the leaf, the CO$_2$ concentration just above the leaf surface, and the leaf water potential. However, since the crop was well irrigated and since in this study it was desired to run the PM model using standard meteorological variables as much as possible, the affects of these other factors were assumed secondary.

c. Deardorff’s single-layer canopy model

Deardorff’s (1978) single-layer model incorporates more of the physical mechanisms that determine the different components of the energy balance above the canopy. The canopy is treated as a bulk layer with negligible heat capacity whose density is specified in terms of an area-averaged shielding factor $\sigma_f$ ($0 \leq \sigma_f \leq 1$). This factor is assigned a value 0 for no foliage and 1 for complete ground coverage. In the following brief description, the subscripts $a, f, g, h, \text{ and } o$ denote values, respectively, at a reference “anemometer-level” height, at the foliage surface, at the ground surface, at a height just above the top of the canopy, and at the surface of bare soil.

In the model, the fluxes of sensible and latent heat just above the canopy are determined as the sum of the individual contributions to the flux by the soil and the foliage:

$$H = H_s + H_f \quad (11)$$

$$\text{LE} = \text{LE}_s + \text{LE}_f, \quad (12)$$

where

$$H_s = \rho_a c_p c_{Hg} u_{af}(T_g - T_{af}) \quad (13)$$

$$H_f = 1.1 \text{ LAI } \rho_a c_p c_{Hg} u_{af}(T_f - T_{af}) \quad (14)$$

$$\text{LE}_s = \rho_a c_{Hg} u_{af}(q_g - q_{af}) \lambda \quad (15)$$

$$\text{LE}_f = \text{LAI } \rho_a c_{Hg} u_{af}(q_f(T_f) - q_{af}) r^* \lambda. \quad (16)$$

Here $\rho_a$ is air density, LAI is leaf area index, $c_f$ is a heat (or moisture) transfer coefficient for a foliage element, $c_{Hg}$ is a heat (or moisture) transfer coefficient for the ground, and $u_{af}$ is within-canopy mean wind. Here, $T$ and $q$ are temperature and specific humidity values, their subscripts indicating the origin. The factor 1.1 accounts for portions of the canopy architecture that exchange heat but do not transpire (Deardorff 1978).

The quantity $r^*$ is the fraction of the potential evaporation rate from the foliage defined as

$$r^* = 1 - \delta_s \left(\frac{r_s}{r_s + r_a}\right) \left[1 - \left(\frac{W_{\text{dew}}}{W_{\text{d max}}}\right)^{2/3}\right], \quad (17)$$

where $\delta_s$ is a delta function for whether or not condensation on the leaf surfaces is occurring (set to 0 for
condensation; 1 for no condensation), $W_{dew}$ is the mass of liquid water retained by foliage per unit horizontal ground area, and $W_{dew}^{max}$ is maximum value of $W_{dew}$ before runoff occurs. The aerodynamic resistance $r_a$ is the same used in the PM scheme and the bulk stomatal resistance $r_s$ is parameterized as

$$r_s = 0.5 \times 10^2 \left( \frac{S_{rain}}{S_{HI}} + 0.03 S_{max} + \Theta + \left( \frac{w_{wind}}{w_s} \right)^2 \right)^2,$$  

where $S_{HI}$ indicates the shortwave radiation flux downward above the canopy, $S_{max}$ the maximum daytime shortwave flux, and $\Theta$ the seasonal dependence of stomatal resistance. The parameters $w_{wind}$ and $w_s$ both dimensionless, are, respectively, a wilting point value of soil moisture in relation to its saturation amount and a root zone value of the volumetric soil water content. The coefficient $0.5 \times 10^2 \times m^{-1} \cdot s^{-1}$ in Eq. (18) is different from an originally suggested (but admittedly uncertain) value of $2.0 \times 10^2 \times m^{-1} \cdot s^{-1}$ (Deardoff 1978) and is based on field measurements made during the Boardman study (Baldocchi 1994).

In Eqs. (13)–(16), the ground surface transfer coefficient $c_H$, is a function of a transfer coefficient for bare soil $c_{H,0}$, a transfer coefficient for the top of the canopy $c_{H,th}$, and the shielding factor $\sigma_f$;

$$c_H = (1 - \sigma_f) c_{H,0} + \sigma_f c_{H,th}.$$  

The within-canopy mean wind $u_{a,f}$ is calculated from the above-canopy mean wind $U_c$, $c_{H,th}$, and $\sigma_f$. The mean within-canopy temperature $T_{a,f}$ is a function of $T_{g}$, $T_{f}$, and $T_{s}$; the mean within-canopy specific humidity $q_{a,f}$ is similarly determined from $q_{g}$, $q_{f}$, and $q_{d}$. Here $T_{f}$ is calculated from a foliage energy budget equation and $T_{g}$ from the force-restore method.

The net radiation at the top of the canopy is determined from $S_{HI}$ (assumed known), $R_{LH,I}$, $S_{HI}$, and $R_{LH,I}$. The longwave flux downward is calculated from Linke’s (1970) formula:

$$R_{LH,I} = \sigma T_c^4 \left[ 0.79 - 0.174 \exp(-0.095 e_a) \right],$$  

where $\sigma$ is the Stefan–Boltzmann constant and $e_a$ (mb) is the vapor pressure. The upward-directed shortwave and longwave components $S_{HI}$ and $R_{LH,I}$ are functions of the incoming $S_{HI}$ and $R_{LH,I}$ values at the top of the canopy, the foliage shielding factor $\sigma_f$, the foliage albedo, the ground surface albedo, and in the case of the upward longwave component: emissivities and temperatures of the ground surface and foliage.

The soil heat flux $G$ is determined from the energy budget equation at the ground surface:

$$-G = H_{g} + LE_{g} - (1 - \alpha_{g}) S_{HI} + R_{LH,I} - R_{LH,I} + r_s (1 - \sigma_f) S_{HI}.$$  

The multilayer canopy model (CANWHT)

The final model to be compared is a one-dimensional, multilayer photosynthesis–evaporation model (CANWHT) for simulating flux densities of moisture, sensible heat, and CO$_2$ as described in Baldocchi (1992). In this model, the integrated, canopy scale value of any function $f(L)$ describing canopy photosynthesis, transpiration, and stomatal conductance is evaluated as

$$f(L_A) = p_{sun} f(x_{sun}) + p_{shade} f(x_{shade})$$  

$$f(L_A) = \int_{0}^{x} p(x(L_A)) f(x(L_A)) dx$$  

$$f(L_A)_{canopy} = \int_{0}^{L_{out}} f(L_A) dL_A,$$

where $L_A$ is the leaf area, $x$ is the energy flux density normal to the leaf, and $p_{sun}$ and $p_{shade}$ are the probabilities of sunlit and shaded leaf areas. Physiological submodels for leaf photosynthesis, leaf and soil respiration, and stomatal conductance are coupled to a canopy micrometeorological model that describes leaf and soil energy exchange, turbulent diffusion, and radiative transfer within the canopy. A listing of the submodels is presented in Table 1.

A conservation budget for a passive scalar within a homogeneous canopy in steady conditions provides the foundation for computing local ambient concentrations and their scalar fluxes:

$$\frac{\partial F(C, z)}{\partial z} = S(C, z),$$

where $F(C, z)$ is the vertical turbulent flux and $S(C, z)$ is the diffusive source–sink strength. The diffusive source–sink strength of a scalar in a unit volume of leaves is proportional to the concentration gradient normal to individual leaves, the surface area $A_l$ of individual leaves, and the number $M$ of leaves in the volume (Finnigan 1985). The diffusive source–sink strength is expressed in the form of a resistance-analog relationship (Meyers and Paw U 1987):

$$S(C, z) = -\rho_{d}(z) \frac{[C(z) - C_{l}]}{r_b(z) + r_r(z)},$$

where $a(z)$ is the leaf area density, $[C(z) - C_{l}]$ is the concentration difference between air outside the laminar boundary layer of the leaves and the air within


Table 1. A summary of environmental inputs, submodels, and literature sources used to develop the multilayer canopy micrometrical model CANWHT (Baldocchi 1994) for computing CO₂ and moisture fluxes over wheat.

<table>
<thead>
<tr>
<th>Submodels</th>
<th>Attributes</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Environmental inputs</td>
<td>Time, photosynthetic photon flux density, air and soil temperature, humidity, wind speed</td>
<td>Farquhar et al. (1980)</td>
</tr>
<tr>
<td>Photosynthesis–respiration model</td>
<td>Biochemical/physiological</td>
<td>Collatz et al. (1991)</td>
</tr>
<tr>
<td>Stomatal conductance model</td>
<td>Dependent on photosynthesis, relative humidity, and CO₂ (Ball–Berry model)</td>
<td>Norman (1979)</td>
</tr>
<tr>
<td>Radiative transfer model</td>
<td>Random spatial and spherical leaf angle distributions</td>
<td>Thomson (1987)</td>
</tr>
<tr>
<td>Turbulent transfer model</td>
<td>Lagrangian random walk model</td>
<td>Thom (1975)</td>
</tr>
<tr>
<td>Surface energy balance model</td>
<td></td>
<td>Kreidemann and Anderson (1988)</td>
</tr>
<tr>
<td>Photosynthesis parameters</td>
<td></td>
<td>Rochette et al. (1991)</td>
</tr>
<tr>
<td>Soil–root respiration parameters</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The stomatal cavities, \( r_s \) is the boundary layer resistance to molecular diffusion, and \( r_l \) is the stomatal resistance.

The interdependence between sources and sinks \( S(z, t) \) and scalar concentrations \( C(z) \) is considered through use of a Lagrangian random-walk model based on the algorithm of Thomson (1987). A Markov sequence form of Thomson’s equation for Gaussian turbulence utilizes time steps equaling 5% of an integral timescale at canopy height: \( T_L = 0.3 \frac{h_c}{u_a} \). Concentration differences between an arbitrary level \( C_i \) and a reference level \( C_r \) (located above the plant canopy) are computed by summing the contributions of material diffusing to or from \( N \) layers within the canopy (denoted by the subscript \( j \); see Raupach 1988):

\[
C_i - C_r = \sum_{j=1}^{N} S_j(C_j) D_{ij} \Delta z_j
\]

where \( D_{ij} \) is a dispersion matrix and has the same units as a resistance \( (\text{s m}^{-1}) \). The dispersion matrix is calculated using the Lagrangian random-walk model. Details on how the random-walk model is implemented in this canopy model is reported by Baldocchi (1992, 1994).

The transfer of photons through the canopy is needed to evaluate photosynthesis, stomatal conductance, and leaf and soil energy balances. The radiative transfer model was derived from probabilistic theory (Myneni et al. 1989; Norman 1979) and assumes that foliage is randomly distributed in space, leaves have a spherical inclination angle distribution, and the sun is a point source (penumbral effects are insignificant) (Norman 1979). In this case the probability of beam penetration \( P_0 \) is calculated using a Poisson distribution:

\[
P_0 = \exp\left(-\frac{L_A G_f}{\sin \beta}\right)
\]

where \( \beta \) is the solar elevation angle and \( G_f \) is the foliage orientation function. The latter function represents the direction cosine between the sun and the mean leaf normal; for the ideal case \( G_f \) is constant and equal to 0.5 (Norman 1979). The radiative transfer model is also used to evaluate the probability of sunlit and shaded leaves, described in Eq. (24). For the random, spherical canopy, \( p_{sun} \) equals the Poisson probability function. The radiative transfer models have been tested thoroughly against measurements of radiation transfer through soybeans (Meyers and Paw U 1987).

Stomatal conductance is calculated according to the Ball–Berry submodule described, for example, in Collatz et al. (1991):

\[
g_s = m \frac{A_n \text{RH}}{M_{CO_2}} + g_0.
\]

Here \( g_s \) represents the stomatal conductance, \( A_n \) the net leaf photosynthesis, and \( RH \) and \( M_{CO_2} \) the relative humidity and CO₂ mole fraction of the air at the leaf surface. The constants \( m \) and \( g_0 \) are the slope and intercept of a linear regression from laboratory gas exchange studies relating the three other parameters to the right of the equal sign to the stomatal conductance. Thus, \( g_0 \) is essentially the cuticle resistance of the leaf.

Soil constitutes the lowest boundary of a canopy CO₂ and water exchange model. Soil/root respiration rates were computed using algorithms derived by Rochette et al. (1991). Flux densities of heat and water at the soil–air boundary were computed using a 10-layer numerical soil heat transfer model (Campbell 1985).

3. Measurements

The data used to perform the model comparison were collected by the NOAA Atmospheric Turbulence and Diffusion Division research group in the Boardman ARM Regional Flux Experiment (Doran et al. 1992) conducted near Boardman, Oregon, during 2–20 June 1991. Measurements were taken at 4 m over a well-irrigated wheat crop (Triticum Durum) with an average height of approximately 80 cm and leaf area index of 2.75. Plant rows were 15 cm apart so that the wheat crop presented a closed canopy. The data from 11 to 14 June (days 162–165) were chosen for the comparison, primarily because the dataset was uninterrupted.
with long periods of clear sky conditions. All data were reduced to 30-min averages.

Within-canopy measurements included photosynthetically active radiation (PAR) penetration, canopy infrared temperature (IRT), and stomatal resistances. The soil was sandy (mixed, mesic, xeric, Torrissaments Quincy series) with negligible organic matter, and a bulk density of 1.42 g cm$^{-3}$. The volumetric water content was between 0.170 and 0.200 cm$^3$ cm$^{-3}$. Measurements taken above the canopy included net radiation, PAR, specific humidity, air temperature, mean horizontal wind, and fluxes of momentum, sensible heat, and latent heat. The net radiation was measured by a net radiometer (Swissteco, model S-1, Melbourne, Australia), whereas soil heat flux was obtained as the average of three soil heat flux plates (Rebs model HFT-3, Seattle, Washington), buried 0.08 m below the soil surface. Sensible and latent heat fluxes were determined through the use of a three-axis sonic anemometer (Applied Technology, model SWS-211/3K, Boulder, Colorado), an open-path infrared absorption spectrometer (Auble and Meyers 1992), and the eddy correlation method.

A good test of the accuracy of the flux measurements is the degree to which one achieves energy balance closure among the four principal energy balance components. The mean energy balance residual between $R_n - G$ and $H + LE$ was $-20.7 \pm 1.36$ W m$^{-2}$, less than 5% of the mean available energy. Although a Student's paired t-test did show a significant departure from zero at the 0.05 probability level the energy balance closure observed can be considered equal to or better than many reported in the literature [see Baldocchi (1994) for further discussion]. A sample of the diurnal behavior of the four components in the energy balance on day 165 is shown in Fig. 2.

4. Model parameterizations

All of the models used in the comparison require assumptions. Primarily, values for constants and scaling factors must be specified. For example, in the Priestley–Taylor equation [Eq. (2)] a value of 1.26 was used for $c_{\text{PT}}$, as was originally suggested. Primary assumptions and parameterizations used in the remaining models are discussed below.

a. Penman–Monteith model

In the PM scheme, three parameters must be specified: $r_{\text{min}}$, $F_{\text{40}}$, and $D_0$. For a well-watered growing wheat crop, Raukach et al. (1992) suggested an $r_{\text{min}}$ equal to 20 s m$^{-1}$. This value is supported by measurements of stomatal resistances normalized by leaf area index taken during the Boardman study (Fig. 3). Raukach et al. also chose values of $F_{\text{40}} = 300$ W m$^{-2}$ and $D_0 = 0.015$ kg kg$^{-1}$ when working over growing wheat; these were adopted in this study as well.

![Fig. 2. The diurnal behavior of the latent heat flux LE, sensible heat flux H, net radiation $R_n$, and soil heat flux G over the well-irrigated wheat crop on day 165 during BARFEX 1991.](image)

In testing the PM scheme, it is quickly apparent that the available energy stress factor $f_R$ in its present form [Eq. (9)] has an implausible influence on the overall canopy resistance when available energies become small. When $R_n - G = 5$ W m$^{-2}$, the stress factor causes the canopy resistance to have values at least 60 times the minimum. (In this context, it can be noted that the maximum resistance of the stomata is the cuticle resistance, which can be approximated by $r_{\text{max}} \sim 2000/LAI$. For an average LAI of 2.75, this gives $r_{\text{max}} \sim 725$ s m$^{-1}$.) To correct this misrepresentation, an assumption was made that below a certain available energy, the stress factor would have a constant influence on the overall canopy resistance. This defined an $f_{R \text{min}}$ so that once the available energy reached below a critical value, $f_R$ would be held constant and set equal to $f_{R \text{min}}$. In essence, this assumes that the decrease in available energy no longer influences the canopy resistance beyond a certain threshold and that nighttime evaporation rates are driven essentially by the specific humidity deficits, a finding supported by Baldocchi (1994). It was found that a value of $f_{R \text{min}} = 0.30$ corresponding to an available energy threshold of 107 W m$^{-2}$ gave good results. A summary of input values used in the PM scheme is given in Table 2.

b. Deardorff model

The principal assumptions made in running the Deardorff model are a foliage shielding factor ($\sigma_f$) of 0.95, a foliage albedo ($\alpha_f$) of 0.15, a foliage emissivity ($\varepsilon_f$) of 0.96, and a ground emissivity ($\varepsilon_g$) of 0.98. Based on values of 0.172 for the volumetric concentration of soil moisture at the ground surface (measured) and 0.2 for the averaged soil moisture over a 50-cm depth (assumed), the “root zone” value of the soil moisture $w$, was taken to be 0.197. The density $\rho_s$ and specific heat $c_p$ of the soil were taken to be $1.42 \times 10^3$ g cm$^{-3}$ and
librium latent heat fluxes is clearly evident during the day when the available energy is large and positive. However, at night when other assumptions made in deriving the equation (e.g., positive sensible heat flux) are significantly violated, it is not surprising that deviations from the relationship arise. For these cases, actual nighttime evaporation was often 100 W m\(^{-2}\) greater than those predicted by PT.

In fairness, one should evaluate the model in conditions for which the model was meant to apply. Figure 4b shows a reduced set of the same data in conditions of positive sensible heat flux, and positive moisture flux. The results are good; the best estimate for \(\alpha_{\text{PT}}\) is 1.29 ± 0.14 where the error represents one standard deviation. For the nearly continuous measurements over 19 consecutive days, these two criteria reduced the number of 30-min averages from 549 to 141 and involved only cases from 0700 to 1800 local time. Thus, although clearly the majority of moisture flux integrated over a day occurs during this time, appreciable flux occurring more than half of a typical day is not accurately predicted.

Figures 4c–e show the results of the PM, DF, and CW models in computing the moisture fluxes. For the rest of the analyses, only the 11–14 June data are used. All three models perform well during the day when the magnitudes of the evaporation fluxes are high. Linear fits through the data are shown for each case. The scatter in all three cases is comparable. The PM scheme provides the best overall result with a slope of 1.01 and virtually no offset. The fact that the PM scheme works well at night lends support to parameterizing the bulk stomatal resistance in the model so that it is driven solely by specific humidity deficits.

b. Sensible heat flux

A similar set of scatterplots is presented for the sensible heat flux. For illustrative purposes, a simple resistance-analog (Ohm’s law) equation (hereafter OL) is utilized initially where

\[
H = \rho c_p \frac{T(0) - T(z)}{r_{\text{at}}}. \tag{30}
\]

Two primary assumptions are often made in using this equation. First, researchers occasionally assume that the aerodynamic resistance for sensible heat is equivalent to that for momentum: \(r_{\text{at}} = r_{\text{am}} = \bar{U}/u^*\) (e.g.,

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(R_s)</td>
<td>measured</td>
<td>(r_{s,\text{min}})</td>
<td>20 s m(^{-1})</td>
</tr>
<tr>
<td>(G)</td>
<td>measured</td>
<td>(F_{\text{a0}})</td>
<td>300 W m(^{-2})</td>
</tr>
<tr>
<td>(D_m)</td>
<td>measured</td>
<td>(D_0)</td>
<td>0.015 kg kg(^{-1})</td>
</tr>
<tr>
<td>(r_s)</td>
<td>measured</td>
<td>(r_{\text{min}})</td>
<td>0.30</td>
</tr>
</tbody>
</table>
TABLE 3. Input parameters used in the Deardorff (DF) model.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Reference</th>
<th>Parameter</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha_f$</td>
<td>0.15</td>
<td>measured</td>
<td>$W_{\text{lift}}$</td>
<td>0.0677</td>
<td>Garratt (1992a)</td>
</tr>
<tr>
<td>$u_a$</td>
<td>measured</td>
<td></td>
<td>$W_{\text{sw}}$</td>
<td>negligible</td>
<td></td>
</tr>
<tr>
<td>$T_a$</td>
<td>measured</td>
<td></td>
<td>$\epsilon_f$, $\epsilon_s$</td>
<td>0.98, 0.96</td>
<td>Garratt (1992a)</td>
</tr>
<tr>
<td>$q_a$</td>
<td>measured</td>
<td></td>
<td>LAI</td>
<td>2.8</td>
<td>measured</td>
</tr>
<tr>
<td>$S_{\text{net}}$</td>
<td>measured</td>
<td></td>
<td>$c_i$</td>
<td>1260 J kg$^{-1}$ K$^{-1}$</td>
<td>Garratt (1992b)</td>
</tr>
<tr>
<td>$S_{\text{max}}$</td>
<td>measured</td>
<td></td>
<td>$\epsilon_{\text{sh}}$</td>
<td>0.00736</td>
<td>Garratt (1993)</td>
</tr>
<tr>
<td>$\rho_o$</td>
<td>1.18 kg m$^{-3}$</td>
<td>Garratt (1992b)</td>
<td>$\epsilon_{\text{no}}$</td>
<td>0.00276</td>
<td>Garratt (1993)</td>
</tr>
<tr>
<td>$P$</td>
<td>1.42 g cm$^{-3}$</td>
<td>measured</td>
<td>$\theta$</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>$P_o$</td>
<td>1025 kg m$^{-3}$</td>
<td>measured</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

this is done in deriving the Penman–Monteith equation). Second, the infrared canopy temperature measurements $T_{\text{IR}}$ are assumed to be the surface temperature of the vegetation elements $T(0)$. Figure 5a shows the results obtained when making these assumptions. The infrared canopy temperatures were measured directly in the field using an infrared radiometer (Everest Interscience, model 4000, Fullerton, California). The poor results in Fig. 5a can be attributed to both the inadequacy of the assumption equating the two aerodynamic resistances and the difficulty in obtaining true plant surface temperatures. Baldocchi (1994) showed that $1^\circ$C differences between $T_{\text{IR}}$ and $T(0)$ can cause 30–60 W m$^{-2}$ errors in calculating $H$.

The Deardorff model (Fig. 5b), on the other hand, determines the foliage surface temperature from a foliage surface energy budget equation and performs reasonably well although nighttime values were often slightly underestimated. The multilayer CW model (Fig. 5c) also gives reasonable results; however, the scatter is considerably greater. It is the DF model that appears to give the best results for this flux component.

c. Net radiation

Figures 6a and 6b show the results of the DF and CW models in determining net radiation. Demonstration of both models’ ability to calculate this quantity is important since both schemes use the net radiation to determine leaf temperature, in turn a critical factor in the calculation of sensible and latent heat flux. A linear regression through the DF model results show an error of approximately 10%. There is also significant scatter, however, as will become apparent in looking at the behavior of the model over a diurnal period, (Fig. 8c), a systematic under and overestimate of $R_n$ occurs during the morning and late afternoon, respectively. The source of the error is primarily in the calculation of the outgoing longwave component $R_{\text{LW}}$ and its feedback with leaf temperature. The specification of the shielding factor $\sigma_f$ and a leaf albedo $\alpha_f$ are both involved in determining these quantities and both can be greatly influenced, for example, by solar elevation angles (Deardorff 1978).

In contrast, the CW model performs extremely well in determining the net radiative flux and demonstrates the beneficial effect of budgeting the longwave and shortwave radiation through many levels in the canopy. Essentially, for this approach, negligible scatter is found for $R_n$ greater than zero and maximum errors of 20 W m$^{-2}$ are found when the fluxes are greatest.

d. Soil heat flux

Figures 7a and 7b present the results of DF and CW models in determining soil heat flux $G$. Both show significant nonlinear correlations as one moves from negative to positive values. Further, the DF model overestimates the flux for all conditions except at night when the flux becomes negative. The CW approach has problems achieving the higher flux magnitudes found during the day, but the scatter is significantly less than in the DF approach.

e. Diurnal cycle

It is useful and informative to compare the results over a diurnal period. Figures 8a–d show the behavior of the models for each of the four energy components on day 165, which can be considered consistent with the behavior over all four days (11–14 June). Again, the PM, DF, and CW models all do well in simulating latent heat flux, the PM model especially during the night. The PT equation was left out in this case (for clarity in viewing the plot); however, it has been already demonstrated that it works well during the day over this site. The DF model simulates the sensible heat fluxes well during the day while both the DF and CW models overestimate the fluxes 10%–40% at night. The multilayer approach certainly performs the best for daytime net radiation values and arguably is the most appropriate for soil heat flux although the DF model in this simulation seems to be better able to obtain the maximum positive fluxes.

6. Discussion

Table 4 summarizes the average errors and their standard deviations between the modeled and mea-
sured fluxes for each of the four energy components over the 11–14 June 1991 period. The means of the absolute values of the measured fluxes were rounded to the nearest 50 W m⁻² to define a scale $\xi_0$ for each of the energy balance components. This scale was used to normalize the differences between the measurements and the calculations. This defines the error, $(\xi_{\text{model}} - \xi_{\text{meas}})\xi_0^{-1}$, where $\xi$ is any of the four energy components. Although this method of normalizing is rather arbitrary it has the advantage of allowing the easy extraction of the mean unnormalized differences between modeled and measured values. The actual mean absolute values of $LE$, $H$, $R_a$, and $G$ are, respectively, 196, 65.8, 218, and 32.7 W m⁻².

In the case of latent heat flux, the PM model gives the best results with an average error of less than 1%. The CW model is on the average 4% down while the DF model follows with calculated values 13.5% too low. The PT equation gives the greatest error when used over the full measurement period (∼27%). How-

Fig. 4. A comparison of (a) the Priestley–Taylor (PT) equation for all cases, (b) the PT equation only for cases involving a positive heat flux, and the (c) Penman–Monteith (PM), (d) Deardorff (DF), and (e) CANWHT (CW) models in calculating 30-min averages of latent heat flux $LE$. The lines through the data in (a) and (b) show the results of the PT equation using a proportionality constant of 1.26.

ever, the error reduces to −2.9% when the limited set shown in Fig. 4b is used. As indicated by the standard deviations on the errors, all of the models show comparable scatter.

For sensible heat flux, the Deardorff model gives the best results with a mean error of +6%. The CW model is less accurate; calculated values are on average 17% less than the measurements. The scatter in the results is also considerably less in the DF model. The Ohm’s law approach is clearly the worst performer and should not be attempted unless there is confidence that accurate vegetation surface temperatures are available. If
Fig. 7. A comparison of the (a) DF and (b) CW models in determining soil heat flux $G$.

one assumes that a 1°C error causes a 45 W m$^{-2}$ error in computed heat fluxes, the results in Fig. 5a imply temperature errors ranging only from 0.5°C for positive measured heat fluxes to 1.5°C for the downward fluxes. Even then, errors in using the aerodynamic resistance for momentum (instead of that for sensible heat flux) can also be expected to cause significant errors in the computed heat fluxes. Because the transfer of momentum over vegetated surfaces is enhanced through bluffbody forces, the aerodynamic resistance to momentum

Fig. 8. A comparison of the models' results and the measurements on day 165 for (a) latent heat flux, (b) sensible heat flux, (c) net radiation, and (d) soil heat flux.
Table 4. Summary of normalized mean errors for the Priestley–Taylor (PT), resistance-analog (OL), Penman–Monteith (PM), Deardorff (DF), and CANWHT (CW) models. The average errors are defined according to $\xi = \frac{\xi_{\text{model}} - \xi_{\text{meas}}}{\bar{\xi}}$, where $\xi$ denotes a flux and subscripts indicate a modeled, measured, or scale value. Scale values were obtained through rounding the mean absolute values of the measured fluxes to the nearest 50 W m$^{-2}$. Standard deviations are also given for each error.

<table>
<thead>
<tr>
<th>Component</th>
<th>$\xi_0$</th>
<th>PT or OL</th>
<th>PM</th>
<th>DF</th>
<th>CW</th>
</tr>
</thead>
<tbody>
<tr>
<td>LE</td>
<td>200</td>
<td>-0.273 ± 0.213</td>
<td>0.002 ± 0.166</td>
<td>-0.135 ± 0.183</td>
<td>-0.041 ± 0.186</td>
</tr>
<tr>
<td>$H$</td>
<td>50</td>
<td>-0.673 ± 0.820</td>
<td>0.059 ± 0.544</td>
<td>0.122 ± 0.281</td>
<td>0.114 ± 0.136</td>
</tr>
<tr>
<td>$R_s$</td>
<td>200</td>
<td></td>
<td>0.336 ± 0.497</td>
<td>0.114 ± 0.410</td>
<td></td>
</tr>
<tr>
<td>$G$</td>
<td>50</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The transfer is less than that for heat. Using the logarithmic wind profile equation, the ratio of the two resistances can be expressed as

$$\frac{r_{wH}}{r_w} = \frac{\ln[(z - d)/z_{0H}]}{\ln[(z - d)/z_{0M}]}$$

where $d$ is the zero plane displacement and $z_{0H}$ and $z_{0M}$ are roughness lengths for heat and momentum, respectively. Also, using relations presented, for example, in Raupach et al. (1991), representative values of $d$ and $z_{0M}$ for an 0.80-m wheat crop are 0.60 and 0.065 m. Utilizing a further assumption on the ratio of the roughness lengths $\ln(z_{0M}/z_{0H}) \approx 2$ (Garratt 1992a), the ratio $r_{wH}/r_w = 1.51$. Thus, incorporating the more appropriate aerodynamic resistance in Eq. (30) would decrease the calculated heat fluxes by approximately 66%. The combination of errors associated with determining $T(0)$ in Eq. (30), and errors produced in assuming the eddy diffusivities of heat and momentum are equal, are easily sufficient to explain the poor performance of a resistance-analog equation for sensible heat flux.

A primary reason for adequately parameterizing surface heat fluxes is their direct influence on boundary layer development. During the early morning hours the nocturnal inversion near the ground is eliminated as incoming solar radiation leads to strong heating of the earth’s surface. This results in the rapid growth of the daytime boundary layer. Knowing the vertical extent of the daytime boundary layer is important to airspace exchange since it defines the volume of air involved. An oft-used approach to modeling boundary layer development is to consider only the thermodynamics, where one assumes the boundary layer is an isothermal “slab” of air with jump discontinuities in potential temperature and heat flux at the upper boundary. This generally involves a thermodynamic energy equation for the slab, an energy conservation equation at the capping inversion, and an equation for the downward heat flux at the top of the boundary layer due to entrainment (Tennekes 1973; Carson 1973). The rate of entrainment of free atmosphere air above the boundary layer is often not available and must be parameterized. A successful approach has been that of McNaughton and Spriggs (1986) where entrainment is parameterized by assuming that the product of the boundary layer height $h$ and the magnitude of the temperature jump at the top of the slab $\Delta \theta$ is constant. This leaves a prognostic equation for boundary layer growth:

$$\frac{dh}{dt} = \frac{H_{wH}}{\rho c_p \gamma_{\theta} h}$$

where $H_w$ is the virtual heat flux at the ground, and $\gamma_{\theta}$ is the lapse rate above the boundary layer. A value of 3 K km$^{-1}$ was assumed for $\gamma_{\theta}$ in this comparison. It is estimated that Eq. (32) can account for 80%–90% of convective boundary layer growth (Stull 1988; Culf 1992).

Figure 9 shows the growth of the boundary layer according to Eq. (32) using the measured heat fluxes, and the average errors on the measured values as described in Table 4; the model was initialized with a boundary layer height of 100 m. For the OL model it is apparent that average errors are of $-67\%$ to $-42\%$ in the calculations of the height of the boundary layer, while the CW and DF models involve errors of 9% and 3%, respectively.

The net radiation measurements involve only the DF and CW models. As is evident in Fig. 6b, the CW model gives exceptionally good results with a mean error of approximately 3%. The DF model gives values

![Fig. 9. The effect of heat flux errors on the McNaughton and Spriggs (MS) boundary layer growth model. Average errors from Table 4 were used to perturb the model.](image-url)
on average 12% higher than the measurements. For soil heat flux, the CW model again is the most accurate, although the calculated values are 11% too high. The DF model gave values that are on average 34% too high with greater scatter.

From these comparisons it is evident that a model that employs the most physical processes in determining components of the surface energy budget does not necessarily give the most accurate results. In the case of latent heat flux, Table 4 indicates that the Penman-Monteith equation with a stomatal resistance submodel can perform as well or better than the more rigorous DF and CW models. For sensible heat flux, the Dardorf model gives the best performance. However, for net radiation and soil heat flux the CANWHT model generates values closest to the measurements.

We now address the question that was posed in the beginning of this paper, namely, can the less-rigorous approaches be used to provide reasonably accurate estimates of the surface energy balance components or must one push the effort to actively employ the more rigorous simulation models in large-scale climate modeling? This study suggests that, for latent heat flux densities over closed canopies, a simple Penman-Monteith combination equation would be sufficient. The PT equation is easier to employ but has only limited use since it is designed to provide only daytime upwardly directed fluxes. The PM scheme, aside from its obvious sensitivity to the available energy and vapor pressure deficits, is most sensitive to the minimum resistance specified in the bulk stomatal resistance parameterization; 20% errors in this parameter can lead to 10% errors in calculated moisture flux. However, our knowledge of what the minimum resistances may be for different plant species in various stages of growth is improving, and this limitation appears less critical than that of attempting to accurately specify the numerous other parameters in the more rigorous approaches.

Another limitation is the fact that the soil heat flux must be specified in the PM model; this quantity is not readily available when attempting larger-scale work. This argues perhaps for using the PM scheme in conjunction with another model that calculates the soil heat flux. The errors involved in calculating this component could be tolerable since its contribution to the available energy is usually much less than that of net radiation. Because of its relative simplicity, the DF model would be the likely candidate especially in light of its performance in calculating sensible heat fluxes compared to a simulation model such as CW. It should be mentioned, however, that studies like those of Wullschleger (1993), which provide parameters used in determining photosynthesis rates for many different plant species, allow simulation models to become more easily integrable now into the large-scale climate models. Also, inherent in the complexity of simulation models such as CW is the potential for improvement in many of the submodels used. To a large extent, however, the situation remains as discussed by Raupach and Finnigan (1988), that is, that multilayer models are useful and appropriate when studying canopy scales or less. For larger-scale work it is currently both appropriate and practical to apply the less rigorous schemes.

A final comment on this study is that it was performed over a closed canopy and the results presented here do not entirely apply when significant patches of bare soil are exposed. However, Dolman and Wallace (1991) compared a Lagrangian approach (Raupach 1989) with both single and dual-source big-leaf models over a crop with a low LAI and found that the more rigorous Lagrangian model did not give noticeable improvement over the dual-source approaches. The single-source PM model they used, predictably, did not work as well as the others. It appears that, for open canopies, the contributions of both soil and vegetation to sensible and latent heat fluxes must be treated explicitly.

Acknowledgments. The measurements used in this study were obtained through sponsorship of the Department of Energy as part of its Atmospheric Radiation Measurements (ARM) Program. The first author's research was supported in part by a Global Change Distinguished Postdoctoral Fellowship sponsored by the U.S. Department of Energy, Office of Health and Environmental Research, and administered by Oak Ridge Associated Universities. We are also grateful to D. R. Matt and T. P. Meyers for their comments on the manuscript.

APPENDIX

Notation

\( \alpha_g \) ground surface albedo
\( \alpha_{PT} \) constant of proportionality in the Priestley-Taylor equation
\( a \) leaf area density
\( A_I \) surface area of individual leaves
\( A_n \) net leaf photosynthesis
\( A_q \) exchange coefficient for water vapor
\( \beta \) solar elevation angle
\( c_f \) heat (or moisture) transfer coefficient for foliage elements
\( c_{Ho} \) heat (or moisture) transfer coefficient for bare soil
\( c_{He} \) heat (or moisture) transfer coefficient for soil under a canopy
\( c_{Hh} \) heat (or moisture) transfer coefficient for the top of the canopy
\( c_p \) specific heat of air at constant pressure
\( C \) scalar concentration
\( C_s \) scalar concentration within stomatal cavity
\( \delta_c \) delta function for condensation (0—no condensation, 1—condensation)
$d$ zero plane displacement
$\Delta$ specific humidity deficit at height $z$ above the
  canopy; $q_c(T) - q$
$D_0$ specific humidity scale
$D_{i,j}$ nondimensional matrix for calculating concentra-
  tion differences
$\epsilon$ nondimensional specific humidity derivative:
  $\left(\lambda/c_v\right)(\partial q_c/\partial T)$
$e_a$ vapor pressure of air at top of the canopy
$f_D$ nondimensional conductance due to specific
  humidity deficit
$f_R$ nondimensional conductance due to available
  energy
$F$ vertical turbulent flux
$F_A$ available energy: $R_n - G$
$F_{A0}$ available energy scale
$\gamma_\theta$ lapse rate immediately above the atmospheric
  boundary layer
$g_0$ cuticle conductance
$g_s$ stomatal conductance
$G$ soil heat flux
$G_f$ foliage orientation function
$h$ atmospheric boundary layer height
$h_c$ canopy height
$H$ sensible heat flux at height $z$ above the canopy
$H_f$ sensible heat flux at foliage surface
$H_s$ sensible heat flux at soil surface
$H_{es}$ virtual heat flux at the ground
$\lambda$ latent heat of vaporization
$L_a$ leaf area
$L_{AI}$ leaf area index
$LE$ latent heat flux at height $z$ above the canopy
$LE_f$ latent heat flux at foliage surface
$LE_s$ latent heat flux at soil surface
$m$ slope of stomatal conductance regression
$M$ number of leaves in a given volume
$M_{CO_2}$ mole fraction of CO$_2$ at leaf surface
$p_{shade}$ probability of shaded leaf area
$p_{sun}$ probability of sunlit leaf area
$P_0$ Poisson probability function for beam penetra-
  tion
$q$ specific humidity
$q_{af}$ mean within-canopy specific humidity
$q_e$ specific humidity at ground surface
$q_s$ saturation specific humidity
$r_a$ aerodynamic resistance
$r_{afH}$ aerodynamic resistance for sensible heat
$r_{am}$ aerodynamic resistance for momentum:
  $U/\mu_a$
$r_{av}$ aerodynamic resistance for moisture
$r_b$ leaf boundary layer resistance to molecular
  diffusion
$r^*$ fraction of potential evaporation rate from fo-
  liage
$RH$ relative humidity of air at leaf surface
$\rho_a$ air density
$r_s$ bulk stomatal resistance
$r_{s\text{ min}}$ minimum bulk stomatal resistance
$R_{L,e}$ longwave radiation downward at ground sur-
  face
$R_{L,eu}$ longwave radiation upward at ground surface
$R_{L,h}$ longwave radiation downward at top of the
  canopy
$R_{L,hu}$ longwave radiation upward at top of the can-
 opy
$R_n$ net radiation at height $z$ above the canopy
$\sigma$ Stephan–Boltzmann constant
$\sigma_f$ canopy shielding factor
$S$ diffusive source–sink strength
$S_{h,e}$ shortwave radiation flux downward at ground
  surface
$S_{h,u}$ shortwave radiation flux downward at top of the
  canopy
$S_{h,ue}$ shortwave radiation flux upward at top of the
  canopy
$S_{max}$ maximum daytime downward shortwave radia-
  tion flux
$T_a$ air temperature above the canopy
$T_{af}$ mean within-canopy air temperature
$T_f$ temperature of the foliage surface
$T_g$ temperature of the ground surface
$T_L$ integral timescale
$\theta$ seasonal dependence of stomatal resistance
  within-canopy mean wind
$u_{af}$ friction velocity
$U$ mean horizontal wind at height $z$ above the
  canopy
$w_s$ dimensionless root zone value of volumetric
  soil water content
$w_{wilt}$ dimensionless wilting point of soil water con-
  tent
$W_{dew}$ mass of liquid water retained by foliage per
  unit ground area
$W_{d,e}$ maximum $W_{dew}$ beyond which runoff occurs
$x_{shade}$ energy flux density normal to shaded leaf
$x_{sun}$ energy flux density normal to sunlit leaf
$z_{HF}$ roughness length for heat
$z_{OM}$ roughness length for momentum

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