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Source approach for estimating soil and vegetation energy fluxes in observations of directional radiometric surface temperature

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Abstract

A two-layer model of turbulent exchange that includes the view geometry associated with directional radiometric surface temperature is developed and evaluated by comparison of model predictions with field measurements. Required model inputs are directional brightness temperature and its angle of view, fractional vegetation cover or leaf area index, vegetation height and approximate leaf size, net radiation, and air temperature and wind speed. One advantage of the approach described in this paper is that directional brightness temperatures are considered so that the model should have wider applicability than single-layer models, and it opens the possibility of a simple solution if directional measurements are available from two substantially different view angles. Comparisons with several hundred measurements from two large-scale field experiments were performed. One study was conducted in a semiarid rangeland environment in Southern Arizona (Monsoon '90) while the other was conducted in a subhumid environment, namely the tall grass prairie in Eastern Kansas (FIFE). For the Monsoon '90 site, root-mean-square-differences (RMSD) between model predictions and measurements were between 35 and 60 W m^{-2} for soil, sensible and latent heat flux. With the FIFE site data RMSD values were between 50 and 60 W m^{-2}. The larger scatter with the FIFE data was mainly caused by the model having difficulty reproducing the fluxes for the observation period with dormant vegetation. Considerations of the expected variability associated with flux measurements over complex surfaces suggests that model-derived fluxes were in acceptable agreement with the observations. However refinements in formulations of soil heat flux probably would improve agreement between model predictions and measurements.

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1. Introduction

Reliable estimates of surface energy fluxes are essential for accurate assessment of surface–atmosphere interactions over a wide range of scales. Surface energy fluxes are important over a wide range of spatial and temporal scales: from determining crop water stress at the field scale to regional weather forecasts to simulating the effects of global climate change. The appropriate method of scaling the processes and fluxes has not been resolved, but many in the scientific community have regarded remote sensing as one of the few technologies that will provide the information required to deal with the scaling issue.

Surface sensible heat flux, H, can be estimated using remotely sensed (or radiometric) surface temperature. For a recent review of these approaches see Schmugge and Becker (1992) or Norman et al. (1995). Two major difficulties are associated with satellite estimates of surface temperature: accounting for the influence of atmospheric effects on measurements from a satellite-based sensor, and characterizing the influence of canopy architecture, fractional cover, sensor look angle, and sun zenith angle on the radiometric surface temperature. The problem of atmospheric corrections is discussed elsewhere (e.g., Becker and Li, 1990a; Perry and Moran, 1994; Sugita and Brutsaert, 1993). In this paper a model is presented to relate directional radiometric surface temperature to surface fluxes considering viewing geometry and canopy architecture. This issue is particularly troublesome in many semiarid environments where the surfaces are heterogeneous and contain partial canopy cover that can result in cooler vegetation and hot soil surfaces.

The soil–vegetation system can be approximated with a two-layer model, where the energy fluxes are partitioned between the soil and vegetation (e.g., Kustas, 1990; Shuttleworth and Gurney, 1990; Massman, 1992). In the past, these models have required parameters and observations that cannot be readily obtained from operational satellite and ground data, so their application has been limited. With satellite observations, especially over semi-arid areas, sensor resolution will always prohibit direct measurements of soil and vegetation temperatures. Some studies suggest these temperatures may be extracted from a scatterplot of surface temperature versus vegetation index (Carlson et al., 1994), while others (e.g., Lhomme and Monteny, 1993) have developed simplified two-layer models that can use a composite radiative temperature.

This paper will describe two-layer models for computing fluxes of latent and sensible heat using remote measurements of surface directional brightness temperature and some ancillary data. The relation among brightness, radiometric and aerodynamic temperatures is defined formally and used in the model to improve general applicability to other vegetation systems. Predicted fluxes are compared with measurements collected during the Monsoon '90 (Kustas et al., 1991) field experiment and the First ISLSCP Field Experiment (FIFE) (Sellers et al., 1988).

2. Model development

The approach used here accommodates the difference between radiometric temperature and aerodynamic temperature, and thus represents an advance over single-layer and two-layer models that ignore such effects. The difference between the thermodynamic temperatures of soil and vegetation components causes aerodynamic and radiometric temperatures to be different. Soil and vegetation temperatures contribute to the brightness surface temperature in proportion to the fraction of the radiometer view that is occupied by each component along with the component temperatures and emissivities. However, the soil and vegetation thermodynamic temperatures contribute to the aerodynamic temperature in proportion to the resistances to turbulent transport between the near-surface atmosphere and the soil and vegetation components. Hence, there is no obvious relationship between aerodynamic temperature and the directional brightness temperature, and they certainly can not be expected to be equal. Furthermore, if a scene contains a wide range of thermodynamic temperatures, the relationship between directional radiometric temperature and some average thermodynamic temperature depends on some knowledge of the distribution of temperatures or else uncertainties of a degree C or more can occur (Norman and Becker, 1995). In this example a single emissivity is used to represent the combined soil and vegetation. The ensemble directional radiometric temperature ($T_{RAD}(\theta)$) is determined by the fraction of the radiometer view that is occupied by soil versus vegetation expressed as

$$T_{RAD}(\theta) = \left[f(\theta)T_C^n + (1 - f(\theta))T_S^n\right]^{1/n}, \tag{1}$$

where T_C and T_S are vegetation canopy temperature and soil surface temperature. The factor n is the power of temperature, which when multiplied by a constant, approximates the integral of the black body function over the wavelength interval of the sensor. Usually $n = 4$ is reasonable for 8–14 and 10–12 μm wavelength bands (Becker and Li, 1990b). The fraction of the field of view of the infrared radiometer that is occupied by canopy is $f(\theta)$ and it can be related to view zenith angle (θ), canopy type and fraction of vegetative cover (f_C). Assuming a random canopy with a spherical leaf angle distribution,

$$f(\theta) = 1 - \exp\left(\frac{-0.5F}{\cos\theta}\right) \tag{2}$$

where F is the leaf area index (LAI), and

$$f_C = 1 - \exp(-0.5F). \tag{3}$$

The directional radiometric temperature is calculated from the brightness temperature ($T_B(\theta)$), which is directly measured by a radiometer. If the surface emissivity and sky conditions are known, then $T_B(\theta)$ is given by

$$T_B(\theta) = \left[\epsilon(\theta)(T_{RAD}(\theta))^n + (1 - \epsilon(\theta))T_{SKY}^n\right]^{1/n}. \tag{4}$$

In Eq. 4, $\epsilon(\theta)$ is the directional thermal emissivity and T_{SKY} is the hemispherical temperature of the sky (see Norman and Becker, 1995).

The relation between soil-plus-canopy sensible heat flux (H) and temperature is defined in terms of a quantity that is referred to as the aerodynamic temperature (T_{AERO}) so that

$$H = \rho C_P \frac{(T_{AERO} - T_A)}{R_A + R_{EX}}, \tag{5}$$

where ρC_P is the volumetric heat capacity of air, T_A is the air temperature at height z, and R_{EX} is an 'excess aerodynamic' resistance. R_A is the aerodynamic resistance calculated from the diabatically corrected log temperature profile equations (Brutsaert, 1982), expressed as

$$R_A = \frac{\left[\ln\left(\frac{z_U - d}{z_M}\right) - \Psi_M\right]\left[\ln\left(\frac{z_T - d}{z_M}\right) - \Psi_H\right]}{0.16U}, \tag{6}$$

where z_U and z_T are the height of wind speed measurement U and air temperature measurement T_A, d is displacement height ($d \approx 0.65 h_C$, where h_C is canopy height), z_M is roughness length for momentum ($z_M \approx h_C/8$), the 0.16 comes from the square of von Kármán's constant taken to be 0.4, and Ψ_M and Ψ_H are the diabatic correction factors for momentum and heat (Brutsaert, 1982). The 'excess aerodynamic' resistance is thought to arise from a higher resistance to heat transport than momentum transport. Although it probably is related to canopy characteristics (McNaughton and Van den Hurk, 1995), Garratt and Hicks (1973) postulated that it could be approximated by $[\ln(z_M/z_H)]/(0.4U^*)$, where $\ln(z_M/z_H) \approx 2$; U^* is the friction velocity, which is related to the momentum flux (τ) by $\tau = \rho(U^*)^2$. In this case, z_H is considered a roughness length for heat. This 'excess aerodynamic' resistance, which has been the subject of study in micrometeorology for more than 20 years, must be distinguished from the 'excess' resistance that has occasionally been used as an adjustment factor when directional radiometric temperature is substituted into Eq. 5 in place of the more appropriate aerodynamic temperature (For example Stewart et al., 1994). In this paper we develop a simple model for using measurements of directional radiometric surface temperature without resorting to such an empirical adjustment to Eq. 5. The approach used in this paper does not require re-definition of 'excess' resistance, which may be used in single-layer models that substitute radiometric temperature for aerodynamic temperature, and overcomes the problem of such an empirical resistance depending on radiometer view angle. For example, Vining and Blad (1992) showed that obtaining satisfactory results with a single-layer model required using directional radiometric temperatures from different view angles, depending on the magnitude of the wind speed.

The contribution of canopy and soil layers to fluxes of sensible heat depend on the temperature differences between each layer and the atmosphere, and on the coupling that is assumed between layers. For simplicity, we assume that the flux of heat from the soil surface is in parallel with the flux of heat from the leaves of the canopy (Fig. 1). This is a slightly different arrangement of resistances than has been assumed by others using two-layer models (e.g. Shuttleworth and Wallace, 1985; Sellers et al., 1986). However, for vegetated, semi-arid regions with low to moderate LAI and moderate wind speeds the resistance networks are indistinguishable because gradients of air temperature in the upper canopy are small. This is supported by Norman and Campbell (1983), who modeled fluxes in corn and observed that soil surface evaporation was coupled only weakly to canopy transpiration at moderate wind speeds. The parallel arrangement of resistance that we have assumed permits simpler solution to the equations and may be slightly more appropriate for more sparse, clumped vegetation common to semi-arid

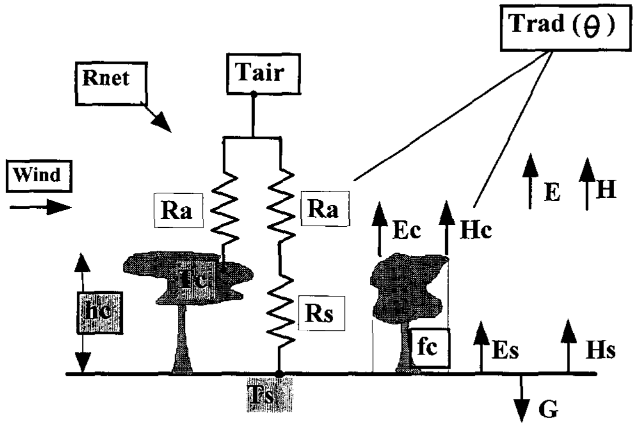

Fig. 1. Schematic diagram illustrating resistance network and key input/output parameters and variables for the 'parallel' model (see text for definition of symbols).

regions, where the canopy and soil tend to interact less with each other than they do for dense canopies. In fact, recently Chebhouni et al. (1993) used an area-weighted parallel sum of resistances for a three-component model of sensible heat flux and obtained good agreement with observed fluxes over sparse vegetation cover.

The approach that we use here with the parallel resistance network can be used with the more complex resistance network (See Fig. 11) of earlier models (Shuttleworth and Wallace, 1985; Sellers et al., 1986; McNaughton and Van den Hurk, 1995), but the equations are more complicated, even though no additional information is required. Appendix A contains a summary of equations for the more complicated resistance network, and it contains a comparison of sensible heat flux estimations from the parallel approach with the more complicated resistance networks using the same input data. Estimates of sensible heat flux from the two resistance network approaches agree well for the semi-arid sites studied. With the equations in Appendix A, the 'Lagrangian' revision of resistors suggested by McNaughton and Van den Hurk (1995) could be used; however these have not been used here because of present uncertainties in parameter values.

If the soil surface and vegetative canopy fluxes are taken to be in parallel with each other (See Fig. 1), then the sensible heat flux can be expressed as

$$H = H_C + H_S = \rho C_P \left[\frac{T_C - T_A}{R_A} + \frac{T_S - T_A}{R_A + R_S} \right], \tag{7}$$

where H_C and H_S are the sensible heat fluxes arising from the vegetative canopy and soil surface respectively, and R_S is the resistance to heat flow in the boundary layer immediately above the soil surface. The formulation for R_S is developed in Appendix B.

Unfortunately we do not have a remote measurement of T_C or T_{AERO}, but we have a measurement of the brightness temperature $T_B(\theta)$. From $T_B(\theta)$ we can calculate the radiometric temperature, $T_{RAD}(\theta)$ from Eq. 4. Since we do not have an observation of T_{AERO}, we can rewrite Eq. 5 as

$$H = \rho C_P \frac{T_{RAD}(\theta) - T_A}{R_*}, \tag{8}$$

where R_* is an effective radiometric-convective resistance given by

$$R_* = \frac{T_{RAD}(\theta) - T_A}{\dfrac{T_C - T_A}{R_A} + \dfrac{T_S - T_A}{R_A + R_S}}. \tag{9}$$

Given measurements of T_A, wind speed (U), the measurement heights (z_U and z_T), $T_{RAD}(\theta)$, canopy height (h_C), approximate leaf size, and fraction of vegetative cover (f_C) or LAI (F), the quantities that are needed to solve for $\boldsymbol{H}$ in Eq. 8 are T_C and T_S. If $T_B(\theta)$ is measured at two view angles (for example $\theta_1 = 0°$ and $\theta_2 = 50°$), then $T_{RAD}(\theta)$ can be estimated for the same angles from Eq. 4, and T_C and T_S can be obtained from the simultaneous solution of two equations and two unknowns using Eq. 1. This is similar to the method that Kimes (1983) used to separate canopy and soil-surface temperatures. The Along-Track-Scanning-Radiometer (ATSR) satellite is capable of making two nearly simultaneous measurements of brightness temperature from two different view angles so that this method may have practical application. Therefore, with the appropriate ground measurements and brightness temperature from two view angles, $\boldsymbol{H}$ can be estimated from Eq. 8 without resorting to empirically determined 'adjustment' factors for R_*, which are calibrated using the observations (e.g., Kustas et al., 1989).

Given an estimate of $\boldsymbol{H}$ from Eq. 8, evapotranspiration rate ($\boldsymbol{LE}$) for the canopy/soil system can be estimated from the canopy/soil energy balance equation,

$$\boldsymbol{LE} = \boldsymbol{R}_n - \boldsymbol{H} - \boldsymbol{G}, \tag{10}$$

if an estimate of net radiation ($\boldsymbol{R}_n$) from ground or satellite observations is available and the midday soil heat conduction flux ($\boldsymbol{G}$) can be approximated by

$$\boldsymbol{G} = 0.35\boldsymbol{R}_{n,SOIL}, \tag{11}$$

where $\boldsymbol{R}_{n,SOIL}$ is the net radiation at the soil surface and is estimated from Eq. 13. The constant of 0.35 is midway between its likely limits of 0.2 and 0.5 (Choudhury et al., 1987). If in the application of Eq. 10, $\boldsymbol{LE} \leq 0$, then $\boldsymbol{LE}$ is set to zero (no condensation under daytime convective conditions) and $\boldsymbol{G}$ is adjusted so that Eq. 10 is satisfied.

Frequently, brightness temperature is available at only a single viewing angle so that another equation for either T_C or T_S is required in addition to Eq. 1 to obtain $\boldsymbol{H}$ from Eqs. 8 and 9. A reasonable estimate of T_C can be obtained by partitioning the divergence of net radiation in the canopy ($\Delta \boldsymbol{R}_n$) into sensible and latent heat fluxes using a Priestly–Taylor approximation (Priestly and Taylor, 1972) for the green part of the canopy. The transpiration is given by

$$\boldsymbol{LE}_C = 1.3 f_g \frac{S}{S + \gamma} \Delta \boldsymbol{R}_n \tag{12}$$

where f_g is the fraction of the LAI that is green, S is the slope of the saturation vapor pressure versus temperature curve, γ is the psychrometer constant (0.066 kPa C^{-1}), and

$$\Delta \boldsymbol{R}_n = \boldsymbol{R}_n - \boldsymbol{R}_{n,SOIL} = \boldsymbol{R}_n - \boldsymbol{R}_n \exp(0.9 \ln(1 - f_C)). \quad (13)$$

The constant of 0.9 is obtained with the assumptions of random leaf positioning and spherical leaf angle distribution (cf. Eq. 3) and net radiation extinction coefficient of 0.45 to calculate the net radiation at the soil surface, $\boldsymbol{R}_{n,SOIL}$. This value of extinction is midway between its likely limits of 0.3 to 0.6 for most vegetation (Ross, 1981). The fraction of LAI that is green is required as an input and may be obtained from knowledge of the phenology of the vegetation. If no information is available on f_g, then it is assumed to be unity. As will become apparent later, Eq. 12 is only an initial calculation of $\boldsymbol{LE}_c$, and it can be overridden if the temperature difference between the canopy and the atmosphere is large.

A superficial examination of Eq. 12 may lead one to question the appropriateness of this equation. Eq. 12 is a better approximation to the transpiration than one might at first expect. The net radiation divergence is closely related to the intercepted photosynthetically active radiation (IPAR), because the dominant term in this divergence is the absorption of photosynthetically active radiation. Numerous investigators have documented the linear relationship between IPAR and vegetation dry matter accumulation in non-stress environments with the concept of light-use-efficiency (LUE) (Kiniry et al., 1989); even though some applications of this approach are not without controversy (Demetriades-Shah et al., 1992). The LUE concept also has been adapted to canopy net photosynthetic rate and Norman and Arkebauer (1991) discuss this. Thus canopy net photosynthetic rate is proportional to IPAR. Canopy net photosynthetic rate and transpiration rate are closely related through a conservative transpiration efficiency (Tanner and Sinclair, 1983), so one expects a relationship between net radiation divergence and transpiration rate. However, variation of atmospheric humidity, or vapor pressure deficit between the vegetation and the atmosphere, might be expected to cause transpiration to vary even though net radiation divergence is nearly constant. As atmospheric humidity decreases and vapor pressure deficit increases, transpiration increases if other factors (particularly stomatal conductance) remain constant. As humidity decreases stomatal conductance does not remain constant but decreases so that changes in conductance would tend to decrease transpiration. The effects on transpiration of increasing vapor pressure deficit and decreasing conductance tend to approximately cancel each other in the atmospheric humidity range from 25 to 75%, if ample soil moisture is available. This is the region of atmospheric humidity in which most transpiration occurs because midday humidities are normally in this range and above and below this range transpiration declines rapidly. This string of logic might be arguable, but careful analysis using the model of Ball et al. (1986), which relates assimilation and stomatal conductance, supports the hypothesis embodied in Eq. 12. Even though the constant of 1.3 in Eq. 12 may vary with vegetation type, the value of 1.3 appears reasonable for the study sites used to test the proposed models.

Following Eq. 12, the sensible heat for the canopy ($\boldsymbol{H}_C$) is

$$\boldsymbol{H}_C = \rho C_P \frac{T_C - T_A}{R_A} = \Delta \boldsymbol{R}_n \left[1 - 1.3 f_g \frac{S}{S + \gamma} \right], \quad (14)$$

and an initial value of T_C can be obtained from Eq. 14. If these values of T_C and T_S are used in Eq. 8, the energy balance equation may not be satisfied. Therefore additional fluxes are calculated to assure that the energy balance equations for the canopy and soil are satisfied. The sensible heat flux from the soil ($\boldsymbol{H}_S$) is

$$H_S = \rho C_P \left(\frac{T_S - T_A}{R_S + R_A} \right) \tag{15}$$

and the soil evaporation ($\boldsymbol{LE}_S$) is

$$\boldsymbol{LE}_S = \boldsymbol{R}_{n,SOIL} - \boldsymbol{H}_S - \boldsymbol{G}. \tag{16}$$

If $\boldsymbol{LE}_S$ is positive, then a solution for soil and canopy energy fluxes is reached. If $\boldsymbol{LE}_S$ is negative, then the soil is likely to be dry so $\boldsymbol{LE}_S$ is set to zero and $\boldsymbol{H}_S$ is obtained from Eq. 16, a new T_S is calculated from Eq. 15, a new T_C is calculated from Eq. 1, a new $\boldsymbol{H}_C$ is given by

$$\boldsymbol{H}_C = \rho C_P \left(\frac{T_C - T_A}{R_A} \right), \tag{17}$$

and a new $\boldsymbol{LE}_C$ is calculated from

$$\boldsymbol{LE}_C = \Delta \boldsymbol{R}_n - \boldsymbol{H}_C. \tag{18}$$

Eq. 18 overrides the initial Priestly–Taylor approximation and allows transpiration to be less than potential. If $\boldsymbol{H}_C > \Delta \boldsymbol{R}_n$, which is impossible, then $\boldsymbol{LE}_C$ is set to zero so that $\boldsymbol{H}_C = \Delta \boldsymbol{R}_n$; T_C is calculated from Eq. 17 and T_S from Eq. 1, which yields $\boldsymbol{H}_S$ from Eq. 15. Then with $\boldsymbol{LE}_S$ already set to zero, $\boldsymbol{G}$ is computed from Eq. 16. This iteration among Eqs. 14–18 provides estimates of T_C and T_S that satisfy the soil-surface and canopy energy balances.

The two-layer model developed above is appropriate when directional brightness temperature is available from only a single view angle or from multiple view angles. Even if remote temperature observations are available from only a single angle, this two-layer model should have an advantage over single-layer models because it incorporates effects of view geometry on that single observation.

Use of typical fractions of canopy height for estimating d and z_M in Eq. 6 may not be appropriate for sparse, heterogeneous surfaces since density of the obstacles is also important (Brutsaert, 1982). Equations derived empirically by Shuttleworth (1991) from the modeling studies by Shaw and Pereira (1982) take into account the density of the vegetation for estimating z_M, and since these formulas use LAI no additional information is required. The accuracy of Eq. 3 in estimating fractional cover for heterogeneous surfaces having a wide range in vegetation types is not well known. For example, the present formula (Eq. 3) has f_C approaching one when LAI is on the order of 4 to 5. Others (e.g., Carlson et al., 1990) suggest $f_C \sim 1$ when LAI is on the order of 3 to 4. Modifications to these and other formulas in the present model will be not be attempted, but are planned for future studies. The emphasis here is to compare the original model formulations with flux observations.

3. Data description

The data to test the model come from two interdisciplinary experiments conducted in semiarid and subhumid areas. The Monsoon '90 experiment was conducted near Tombstone, AZ, in 1990 in the Walnut Gulch Experimental Watershed maintained by the USDA-ARS Southwest Watershed Research Center. For details of the experiment and measurements, see Kustas et al. (1991) and Kustas and Goodrich (1994). The First ISLSCP (International Satellite Landsurface Climatology Project) Field Experiment, FIFE, was conducted near Manhattan Kansas in 1987 and 1989 in and around the Konza Prairie. A thorough description of the experiment and measurements are given in Sellers et al. (1988) and Sellers et al. (1992a).

3.1. Monsoon '90 data

The remote sensing data from Monsoon '90 used in this study were collected at two Meteorological-Flux (METFLUX) stations during the pilot and main field campaign. The pilot field study was conducted for several days during the dry season in June when the vegetation is dormant. This gave baseline measurements of the fluxes and remote sensing data under these conditions. The main field study was conducted during the 'monsoon' season when the vegetation is most active (highest green biomass) and soil moisture is highly variable due to localized and frequent precipitation events. During the main field campaign, continuous measurements of canopy and soil temperatures were made with 2 Everest Interscience [1] radiometers (Model 1000) at each site, one pointed at the soil and the other at the vegetation. Instruments were positioned 1 to 2 m above the surface. This same measurement design was deployed by Nichols (1992) at semiarid sites in Nevada. The composite temperatures for the area surrounding the sites were estimated by using periodic observations with similar nadir viewing instruments mounted on yoke-type backpacks that collected the data over a prescribed set of transects. Details of the yoke measurements which were collected in both field campaigns are given by Moran et al. (1994a). Regression equations were obtained using the canopy and soil temperatures as independent variables and the yoke measurements as the dependent variable. This allowed the estimation of composite temperatures with the continuous infrared observations. METFLUX site number 1 was located in the Lucky Hills subwatershed, which is a shrub-dominated site. METFLUX site number 5 was located in the Kendall subwatershed and is primarily a grass ecosystem. Both sites are heterogeneous with canopy cover less than 50% and significant variability in vegetation height and architecture. Measurements of the canopy covers, height and LAI are summarized in Daughtry et al. (1991) and Weltz et al. (1994). The surface energy balance was computed using the variance technique described by Tillman (1972) for estimating $\boldsymbol{H}$ and measurements of net radiation ($\boldsymbol{R}_n$) and soil heat flux ($\boldsymbol{G}$) for computing latent heat flux as a residual. Comparisons with the one-dimensional eddy correlation technique by

[1] Company and trade names are given for the benefit of the reader and imply no endorsement by the USDA or University of Wisconsin.

Kustas et al. (1994a) showed differences were 20%, on average. A 20% disparity between different measurement techniques appears to be fairly common (e.g., Dugas et al., 1991; Nie et al., 1992) and therefore the fluxes estimated by the variance approach were considered reliable. Measurements of wind speed and air temperatures were made at a nominal height of 4 m. Twenty-minute data were recorded. Hourly-averaged data were used in the analysis.

3.2. FIFE data

The remote sensing data from FIFE used in this paper were collected by a helicopter-based remote sensing system (Walthall and Middleton, 1992). During the 1987 campaigns, data were acquired with this system at approximately 300 meters above ground level with a Barnes Modular Multiband Radiometer (MMR) instrument. The instrument had a 1° field of view, resulting in a nadir footprint approximately 5.2 m in diameter. The data used in this analysis were from the thermal band (band 8) of the MMR instrument which has a nominal bandpass of 10.4–12.3 μm. The data were acquired as the helicopter hovered over a number of different flux sites in the study area. As described by Walthall and Middleton (1992), coverage of the various sites was dependent on a number of factors and uncontrollable sky conditions. In this analysis, we have used data acquired in IFCs (Intensive Field Campaigns) 1, 3 and 4. The approximate magnitude of atmospheric effects on the surface temperatures estimated with the MMR data were evaluated by using near-simultaneous radiosonde profiles of atmospheric temperature and humidity (Brutsaert and Sugita, 1990) as input into the radiative transfer code Lowtran-7 (Kneizys et al., 1988). Analysis of data from four days (two days in IFC 1, and one each in IFCs 3 and 4) suggested that the magnitude of the atmospheric effects on the surface temperature estimated by the MMR was on the order of 0.2 to 0.8 degrees, depending on the atmospheric conditions for a particular day and the magnitude of the surface brightness temperature. For the brightness temperature values observed, the correction was usually between +0.3 and 0.4 degrees.

Measurements acquired at the different surface flux sites during the FIFE 1987 campaign are described in detail by Kanemasu et al. (1992) and Smith et al. (1992). The flux data and near-surface meterological data were reported at most stations at half-hourly intervals. The data used in these analyses were those reported for the half-hour intervals which most closely corresponded to the time of data acquisition by the helicopter-based system. The near-surface meterological data, including wind speed and air temperature were acquired at various heights (from about 1.5–2.5 m) at different stations. The surface energy balance was estimated using Bowen ratio and eddy correlation techniques. Thirteen sites were selected for testing the utility of the model for computing fluxes. The sites were numbers 2, 8, 10, 16, 20, 24, 26, 28, 30, 36, 40, 42, 44, and described in some detail by Smith et al. (1992). The variation in percent cover at these sites was mainly determined by whether the site was burned or unburned and grazed or ungrazed. Stewart and Verma (1992) give a detailed description of a burned-ungrazed site having fairly uniform cover and relatively high vegetation biomass (site 16) and a burned-heavily grazed site containing a significant amount of bare soil (site 26). A burned-grazed site and intermediate in cover and biomass relative to sites 16 and 26 is described by Fritschen and Qian (1992). By way of summary, Sites 16, 26, 28 and 30 had eddy correlation instruments and the remaining sites used Bowen ratio equipment. Sites 2 and 36 were unburned sites and remaining sites were burned in the Spring of the year that measurements were taken. Sites 2, 10, 16, 24, 26, 30, and 44 had slopes less than 3 degrees, site 42 had a slope greater than 8 degrees and the remaining sites had slopes between 3 and 8 degrees with various azimuths. The values of LAI came from Table 1b from Sellers et al. (1992b).

4. Results and discussion

The performance of the model was evaluated using difference measures suggested by Willmott (1982). These include the root mean squares difference (RMSD), which can be decomposed into systematic ($RMSD_S$) and unsystematic ($RMSD_U$) differences (Willmott, 1984; Willmott et al., 1985), and the mean absolute difference (MAD). The value of $RMSD_S$ is composed of an additive and proportional contribution and these relate to coefficients from the linear least squares fit between model and observations (Willmott, 1984). Willmott suggests that one should minimize $RMSD_S$ so that it is largely composed of $RMSD_U$ and hence indicate the best agreement with the observations one might expect with the model. See Table 1 for the formulas for computing the various difference statistics.

Predictions of sensible heat from the model are compared to measurements in Fig. 2 for site 1 (shrub-dominated) and Fig. 3 for site 5 (grass-dominated). For a few observations collected during the dry season, $f_g = 0$ in Eq. 12 since the vegetation was

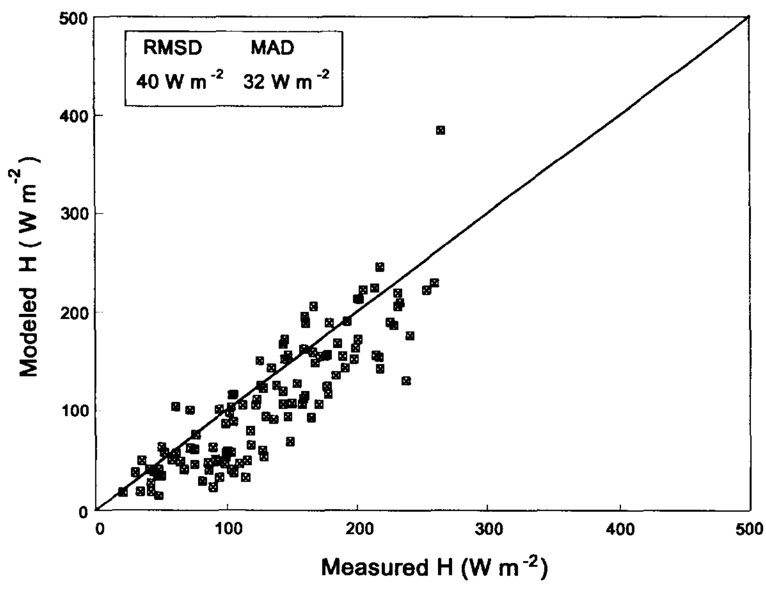

Fig. 2. Modeled versus measured sensible heat flux, *H*, for site 1 located in the shrub-dominated subwatershed. The line represents a 1:1 relationship.

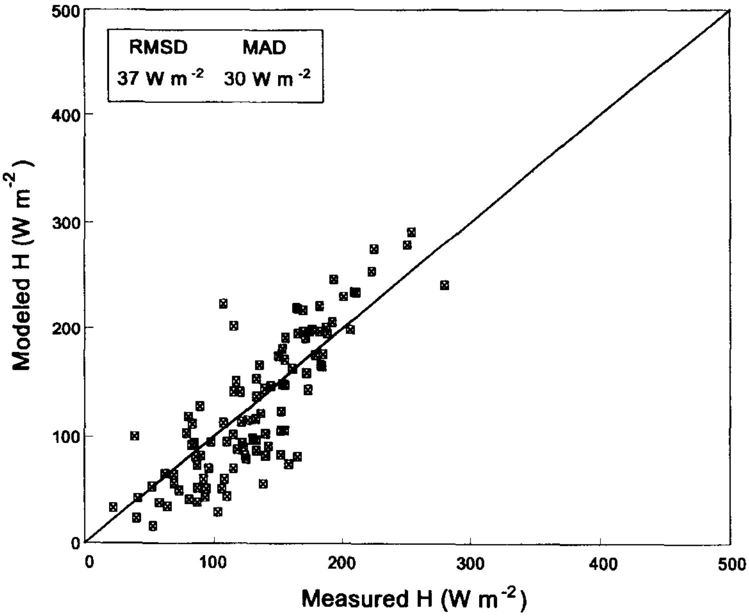

Fig. 3. Modeled versus measured sensible heat flux, $\boldsymbol{H}$, for site 5 located in the grass-dominated subwatershed. The line represents a 1:1 relationship.

dormant; all other cases $f_g = 1$. For site 1, predictions of $\boldsymbol{H}$ tend to be less than measurements, while for site 5 no such bias between modeled and measured $\boldsymbol{H}$ is evident. The plot of modeled $\boldsymbol{G}$ versus measured for site 1 (Fig. 4) and site 5 (Fig. 5) both show that the model overestimates the lower values of $\boldsymbol{G}$, say $\boldsymbol{G} < 100$ W m^{-2}. This has to do with the fact that $\boldsymbol{G}$ is not a constant fraction of $\boldsymbol{R}_n$ over the course of the day (Brutsaert, 1982) as assumed in formulating Eq. 11. The comparison of latent heat fluxes shows more scatter resulting in larger values of RMSD and MAD for site 1 (Fig. 6) and site 5 (Fig. 7).

Table 1 lists quantitative measures of model performance as suggested by Willmott (1982), except for the index of agreement since only one model is used in this comparison. For site 1, the mean bias of predicted minus observed in Table 1 ($\langle$P$\rangle$–$\langle$O$\rangle$) is fairly small for $\boldsymbol{G}$ but more significant for $\boldsymbol{H}$ (-21 W m^{-2}) and $\boldsymbol{LE}$ (15 W m^{-2}). However, the least squares regression results and the relative size of RMSD$_S$ to RMSD for $\boldsymbol{G}$ indicates a significant systematic difference. From the ratio RMSD$_S^2$/RMSD^2 one can obtain an idea of the proportion of RMSD that arises from systematic differences. For $\boldsymbol{G}$ this proportion is nearly 70%, while for $\boldsymbol{H}$ and $\boldsymbol{LE}$ this proportion is 30% and 11%, respectively. For site 5, the model-derived fluxes are generally in better agreement with the observations compared to site 1. The magnitude of the bias has reduced substantially, with $\boldsymbol{H}$ having a bias of 5 W m^{-2} and a bias of 1 W m^{-2} for $\boldsymbol{LE}$. Compared to site 1, $\boldsymbol{G}$ again has a relatively large RMSD$_S$, but $\boldsymbol{H}$ and $\boldsymbol{LE}$ have significantly smaller RMSD$_S$ values. The ratio RMSD$_S^2$/RMSD^2 is about 55% for $\boldsymbol{G}$, 5% for $\boldsymbol{H}$ and 1% for $\boldsymbol{LE}$.

Table 1
Quantitative measures of model performance with the Monsoon '90 data

Site	Flux	n [a]	$\langle O \rangle$ [b], $W m^{-2}$	$\langle P \rangle$ [c], $W m^{-2}$	S_o [d], $W m^{-2}$	S_p [e], $W m^{-2}$	a [f], $W m^{-2}$	b [g]	MAD [h], $W m^{-2}$	RMSD [i], $W m^{-2}$	$RMSD_S$ [j], $W m^{-2}$	$RMSD_U$ [k], $W m^{-2}$	r^2 [l]
1	*H*	117	133	112	60	66	−15	0.95	32	40	22	33	0.74
1	*G*	117	112	118	69	45	52	0.59	28	35	29	19	0.81
1	*LE*	117	162	177	67	92	−7	1.14	45	54	18	51	0.69
5	*H*	107	133	128	49	67	−22	1.12	30	37	8	36	0.70
5	*G*	107	99	108	57	41	50	0.57	24	35	26	24	0.64
5	*LE*	107	208	207	90	100	−4	1.01	31	41	3	41	0.83

[a] number of observations.
[b] mean of the observed variable.
[c] mean of the model predicted variable.
[d] standard deviation of the observed variable.
[e] standard deviation of the predicted variable.
[f] intercept of the least squares regression: $\overline{P}_i = a + b(O_i)$.
[g] slope of the least squares regression: $\overline{P}_i = a + b(O_i)$.
[h] Mean Absolute Difference: $\sum_{i=1}^{n} |P_i - O_i| / n$.
[i] Root Mean Square Difference: $\left[\sum_{i=1}^{n} (P_i - O_i)^2 / n\right]^{1/2}$.
[j] Systematic Root Mean Square Difference: $\left[\sum_{i=1}^{n} (\overline{P}_i - O_i)^2 / n\right]^{1/2}$.
[k] Unsystematic Root Mean Square Difference: $\left[\sum_{i=1}^{n} (P_i - \overline{P}_i)^2 / n\right]^{1/2}$.
[l] Coefficient of Determination.

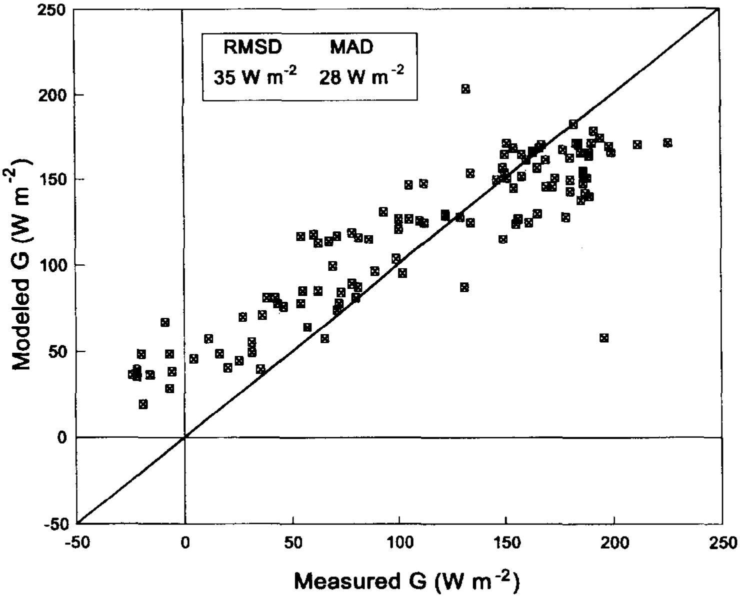

Fig. 4. Modeled versus measured soil heat flux, G, for site 1 located in the shrub-dominated subwatershed. The line represents a 1:1 relationship.

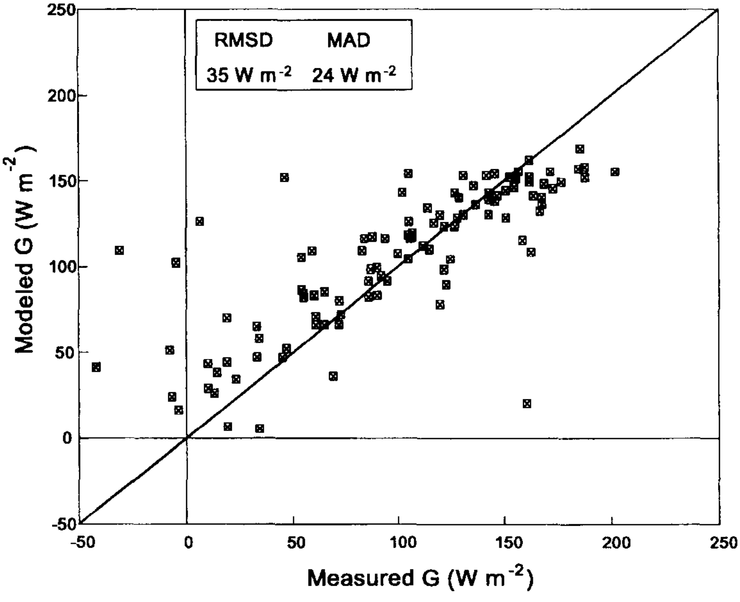

Fig. 5. Modeled versus measured soil heat flux, G, for site 5 located in the grass-dominated subwatershed. The line represents a 1:1 relationship.

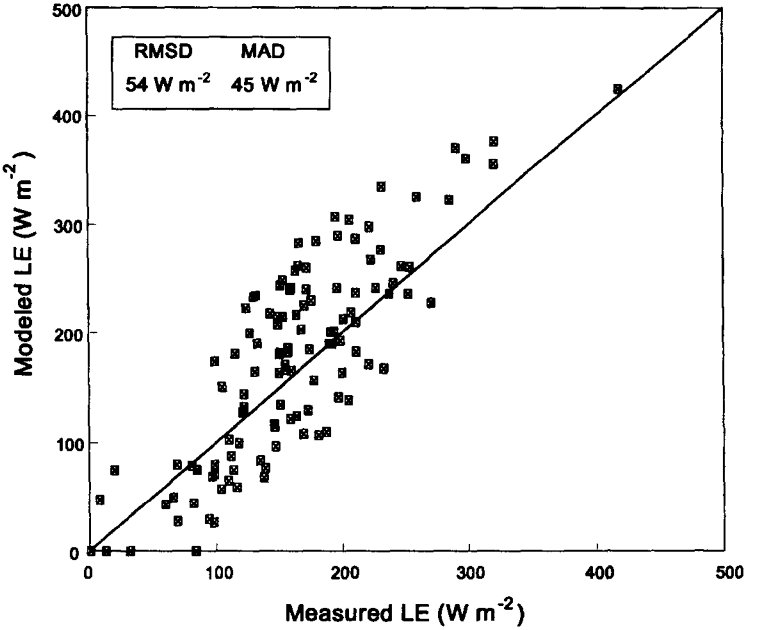

Fig. 6. Modeled versus measured latent heat flux, *LE*, for site 1 located in the shrub-dominated subwatershed. The line represents a 1:1 relationship.

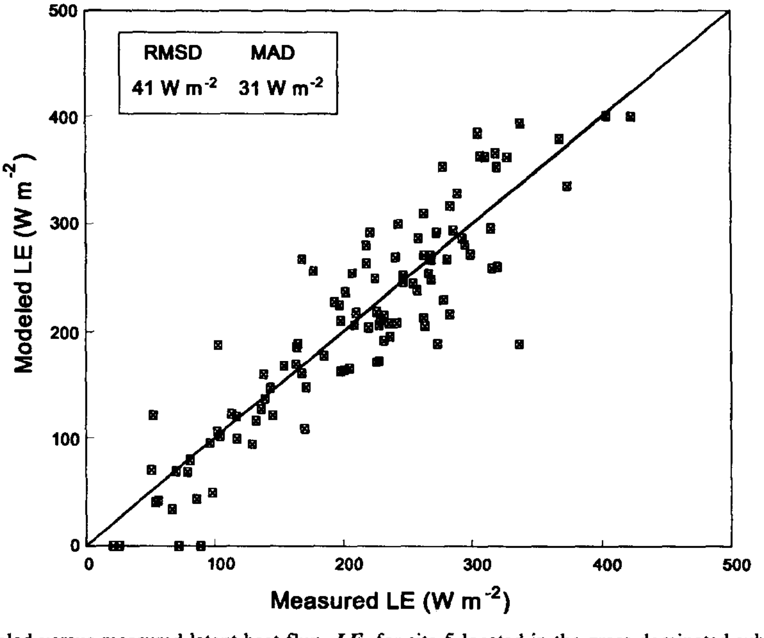

Fig. 7. Modeled versus measured latent heat flux, *LE*, for site 5 located in the grass-dominated subwatershed. The line represents a 1:1 relationship.

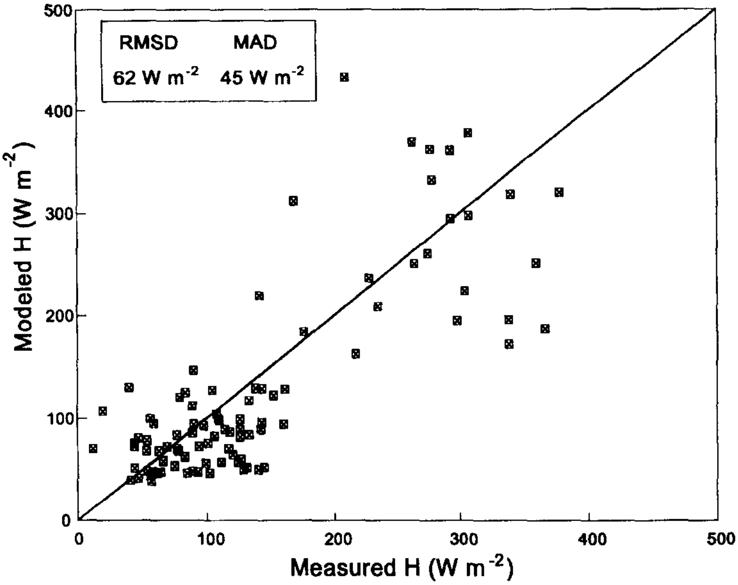

Fig. 8. Modeled versus measured sensible heat flux, H, for the FIFE sites. The line represents a 1:1 relationship.

For the FIFE data, model predictions of H versus measured is illustrated in Fig. 8. There are two distinct clouds of points, one for measured $H < 200$ W m^{-2} and for measured $H > 200$ W m^{-2}. The values of $H > 200$ W m^{-2}, where there is significantly larger scatter, are mainly from the observations collected during IFC 4. This IFC had essentially dormant vegetation. The total LAI was computed differently from Sellers et al. (1992b) for IFC 4 since the LAI of the dormant vegetation would be roughly 1/2 that of the live. Thus the ratio of dead to live biomass multiplying the green LAI values was reduced by 1/2 before it was added to the live LAI values to yield the total LAI. For IFC 4, the latent heat flux from the canopy computed with Eq. 12 was weighted by the fraction of green vegetation present, f_g, while it was assumed $f_g = 1$ for the other IFCs. For G, the RMSD is comparable to the results with H with the majority of model-derived values of G illustrated in Fig. 9 being significantly higher than measured, much more so than with the Monsoon '90 data (see Figs. 4 and 5). A change in the extinction coefficient to 0.6 in Eq. 13 and a reduction of the constant from 0.35 to 0.20 in Eq. 11 reduced the RMSD to 40 W m^{-2}. However, there was no significant reduction in RMSD for H or LE. For LE, the comparison in Fig. 10 shows less scatter than for H, yielding a lower RMSD. However, most of the LE values less than about 150 W m^{-2} are from IFC 4, suggesting that relatively large errors in estimating evaporation with the model can be expected under conditions with dormant vegetation.

Table 2 lists the quantitative measures of model performance for the FIFE data. The mean bias is significant for G, $\langle P \rangle - \langle O \rangle = 45$ W m^{-2}, but relatively smaller values for H and LE. For the FIFE data, the ratio $\mathrm{RMSD}_S^2/\mathrm{RMSD}^2$ for G of nearly 60%, for H of 8% and for LE at 18% are similar to the site 1 results from the Monsoon '90 data set.

Table 2
Quantitative measures [a] of model performance with the FIFE data

Flux	n	$\langle O \rangle$, $W m^{-2}$	$\langle P \rangle$, $W m^{-2}$	S_o, $W m^{-2}$	S_p, $W m^{-2}$	a, $W m^{-2}$	b	MAD, $W m^{-2}$	RMSD, $W m^{-2}$	$RMSD_S$, $W m^{-2}$	$RMSD_U$, $W m^{-2}$	r^2
H	97	139	130	91	97	14	0.84	45	62	17	60	0.62
G	97	59	104	42	45	71	0.56	49	62	48	39	0.27
LE	97	269	255	141	131	24	0.86	46	55	24	50	0.86

[a] See Table 1 for column definitions.

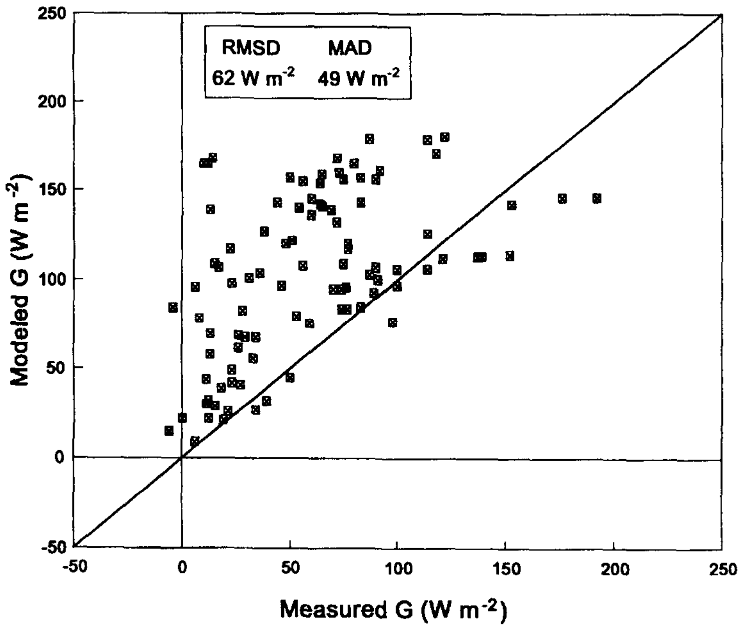

Fig. 9. Modeled versus measured soil heat flux, G, for the FIFE sites. The line represents a 1:1 relationship.

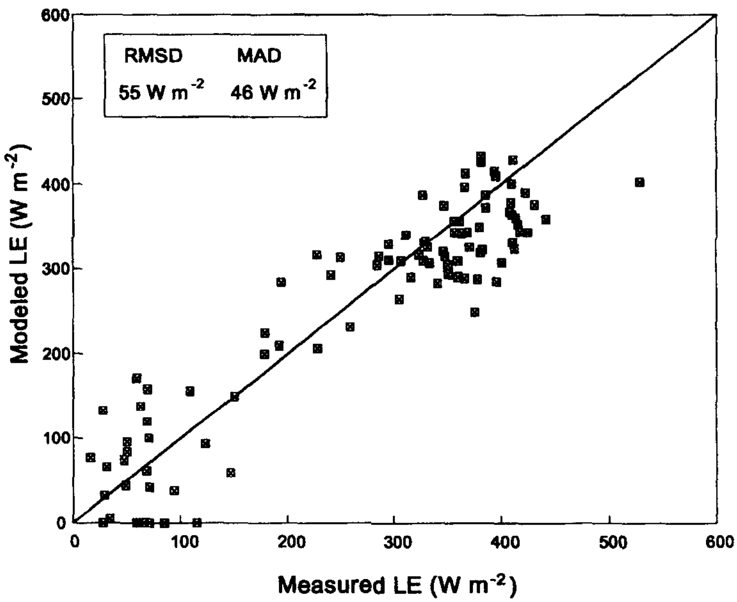

Fig. 10. Modeled versus measured latent heat flux, LE, for the FIFE sites. The line represents a 1:1 relationship.

The performance of the present two-layer model in comparison to other approaches cannot be directly assessed without running the other models with the same data set. This is beyond the scope of the present paper. However, some inferences can be made by comparing these results with other studies using radiometric surface temperature observations for computing the surface energy balance over the same experimental sites, and by contrasting these differences with typical variations obtained by using different flux measurement techniques.

For Monsoon '90, Moran et al. (1994b) computed H using a one-layer resistance model with T_{RAD} and by incorporating empirical 'excess resistance' parameters to estimate R. They employed remote sensing methods for estimating R_n and G. Values of MAD reported by Moran et al. (1994b) were between 30 and 40 W m^{-2} for H, about 20 W m^{-2} for G and around 40 W m^{-2} for LE. Kustas et al. (1994b), using a similar approach to Moran et al. (1994b), reported values of RMSD for H, G and LE of approximately 35 W m^{-2}, 30 W m^{-2} and 50 W m^{-2}, respectively and values of MAD of 25 W m^{-2} for H and G, and 35 W m^{-2} for LE. For FIFE, single-layer model results have been reported using regression statistics and correlation coefficient (r) or coefficient of determination (r^2). Sugita and Brutsaert (1990) using atmospheric boundary layer data in the surface layer with an empirically-derived z_H computed regional H values and compared the estimates with the average of six Bowen ratio stations located in the FIFE site. They found $r = 0.87$ or $r^2 = 0.76$ between predicted and observed H. Hall et al. (1992) used R_A with no adjustment to z_H and obtained poor agreement with measured H, namely a r^2 of 0.45.

The variation in flux estimation by various micrometeorological techniques used in Monsoon '90 was investigated by Kustas et al. (1994a), Stannard et al. (1994) and Moran et al. (1994b) and in FIFE by Nie et al. (1992) and Fritschen et al. (1992). Kustas et al. (1994a) found in comparing eddy correlation and variance approach for H that the RMSD and MAD were around 25 W m^{-2} and 20 W m^{-2} and for LE the RMSD and MAD were about 40 W m^{-2} and 30 W m^{-2}, respectively. Stannard et al. (1994) showed that standard errors in G (or RMSD$_U$ by definition) were around 30 W m^{-2}. Moran et al. (1994b) compared flux estimates at sites 1 and 5 and obtained similar results to Kustas et al. (1994a). For FIFE, Nie et al. (1992) in their intercomparison of flux measurements during FIFE found average differences in LE were about 20%. However Fritschen et al. (1992) found in comparing Bowen ratio and eddy correlation techniques during FIFE '89 that differences in the fluxes were typically larger than 20%.

Blad and Rosenberg (1974) performed an error analysis on the Bowen ratio method and predicted errors that ranged from 10 to 50% with 20% being a typical value. They also observed, as others have, that the Bowen ratio method systematically underestimated evapotranspiration measurements from a lysimeter by 20 to 30%, and noted an RMS error about the best-fit regression line (between lysimeter and Bowen ratio results) of about 20% of the mean with r^2 between 0.6 and 0.9.

Since there is no consistency in the statistics used among the past studies when comparing model to measurements and different measurement techniques, a qualitative evaluation of the present model results was deemed appropriate. For Monsoon '90, it appears that the agreement between two-layer model estimates of H and LE and the observations is not an improvement over the single-layer results since the values of

RMSD and MAD between the two techniques are similar. Nevertheless, the need for an 'excess resistance' term is not necessary with the present formulation, which makes it more robust and thus may require less calibration when applied to other landscapes. The RMSD and MAD values with the two-layer approach appear to be higher than that obtained by comparing different micrometeorological techniques used in Monsoon '90, so that improvement in the model may be possible. For FIFE, there are no direct comparisons with single-layer model results using the difference statistics listed in Tables 1 and 2. The resulting $r^2 = 0.62$ for $\boldsymbol{H}$ in Table 2 is higher than Hall et al. (1992) but lower than Sugita and Brutsaert (1990). However, Sugita and Brutsaert (1990) calibrated z_H for the FIFE site, which makes it considerably more difficult to apply their method to other surfaces without calibration. For the expected variation in the flux observations, Nie et al. (1992) find 20% for $\boldsymbol{LE}$. Given that the average measured $\boldsymbol{LE}$ for this data set is 270 $\mathrm{W\,m^{-2}}$, the average percentage difference computed by taking the ratio of RMSD with the average (RMSD/⟨O⟩) is about 20%. Hence, the model estimates of $\boldsymbol{LE}$ appear to be within the expected variability of the measured latent heat flux under most conditions. Furthermore, RMSD and MAD values from the two-layer approach are similar to results obtained in a comparison between Bowen ratio estimates and lysimeter measurements. Computing the same ratio for $\boldsymbol{H}$ yields an average percentage difference of nearly 45%. This may appear unacceptable, but the average Bowen ratio ($\boldsymbol{H/LE}$) for the intercomparison by Nie et al. (1992) and the flux observations used here was about 0.5. With $\boldsymbol{H}$ being half the magnitude of $\boldsymbol{LE}$, a variation of as high as 40% might be expected from the measurements, so that a value of RMSD/⟨O⟩ of 45% from the model may still be considered satisfactory.

5. Conclusions

Preliminary testing of a simple two-layer model for computing turbulent fluxes of sensible ($\boldsymbol{H}$) and latent ($\boldsymbol{LE}$) heat, and soil heat flux ($\boldsymbol{G}$) with radiometric surface temperature was evaluated with experimental data collected from semiarid and subhumid sites. Differences between the model and observations are within the expected accuracy of the observations, and are comparable to and in some cases an improvement over single-layer modeling results. The advantage of the present approach is that it does not require an empirical correction for the 'excess resistance' or calibration of z_H, which often is required by single-layer models when applied to heterogeneous surfaces in order to obtain acceptable results (Stewart et al., 1994). The incorporation of brightness temperature in the formulation is an improvement over current single-layer models, whether brightness temperature observations are available at a single or at multiple view angles. This accommodation of variable view angles is particularly important with the AVHRR (Advanced Very High Resolution Radiometer) satellite, which can have view angles from nadir to 55 degrees. Therefore, the model may not require calibration over a wider variety of surfaces than single-layer models, because it makes a distinction between the T_{RAD} and T_{AERO}. In addition, the relatively modest number of input parameters required by the model gives it operational capability.

A critical parameter in the model is the leaf area index (LAI) or percentage cover, f_C. Fortunately, both LAI and f_C have the potential of being estimated by remote sensing (Carlson et al., 1990; Baret and Guyot, 1991). However, a method is needed to account for senescent vegetation which doesn't contribute to canopy latent heat flux, $\boldsymbol{LE}_C$, but to the canopy sensible heat flux, $\boldsymbol{H}_C$. This may involve knowing the vegetation index–LAI relationship when the vegetation is in peak greenness, and then accounting for a reduction in total LAI in later stages where again, a remotely sensed vegetation index may assist in determining this value. A better method may be needed to estimate $\boldsymbol{G}$ if the model is applied to periods where $\boldsymbol{G}$ is not a constant fraction of $\boldsymbol{R}_n$.

Future work is planned on testing the model over other vegetated surfaces and under different environmental conditions. Additionally, sensitivity analyses of the model will be conducted. These include investigating the potential variability in computed fluxes caused by modifications to the present model formulations in estimating surface roughness z_M, LAI, $\Delta\boldsymbol{R}_n$ and $\boldsymbol{G}$.

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Appendix A

An alternative vegetation-soil resistance network to that discussed in the main body of the paper is illustrated in Fig. 11. We will refer to this as the 'series' network, in contrast to the 'parallel' network discussed in the main body of the paper, even though resistors are in both series and parallel in both networks. It should be noted here that ALL temperatures in this appendix are absolute temperatures in units of K. The sensible heat fluxes from the vegetation ($\boldsymbol{H}_\mathrm{C}$), soil ($\boldsymbol{H}_\mathrm{S}$) and total soil-plus-vegetation ($\boldsymbol{H}$) are given by

$$\boldsymbol{H}_\mathrm{S} = \rho C_\mathrm{P} \frac{T_\mathrm{S} - T_{AC}}{R_\mathrm{S}} \tag{A.1}$$

$$\boldsymbol{H}_\mathrm{C} = \rho C_\mathrm{P} \frac{T_\mathrm{C} - T_{AC}}{R_\mathrm{X}} \tag{A.2}$$

$$\boldsymbol{H} = \rho C_\mathrm{P} \frac{T_{AC} - T_\mathrm{A}}{R_\mathrm{A}}. \tag{A.3}$$

Since $\boldsymbol{H}$ is the sum of $\boldsymbol{H}_\mathrm{S}$ and $\boldsymbol{H}_\mathrm{C}$, the following equation relates the air temperature in the canopy (T_AC) to resistance and other temperatures:

$$T_\mathrm{AC} = \frac{\dfrac{T_\mathrm{A}}{R_\mathrm{A}} + \dfrac{T_\mathrm{S}}{R_\mathrm{S}} + \dfrac{T_\mathrm{C}}{R_\mathrm{X}}}{\dfrac{1}{R_\mathrm{A}} + \dfrac{1}{R_\mathrm{S}} + \dfrac{1}{R_\mathrm{X}}}. \tag{A.4}$$

A general, closed form solution to this series resistance network is not obvious to the authors. Therefore, the approach here is to use a linearized form of Eq. 1 (in the main

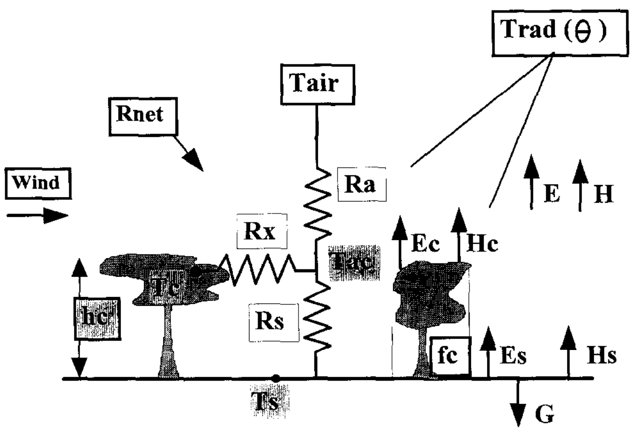


Fig. 11. Schematic diagram illustrating the resistance network and key input/output parameters and variables for the 'series' approach (See text for definition of symbols).

body of the paper) to obtain an approximate solution for $T_{C,Lin}$ and $T_{AC,Lin}$, and then derive small correction factors (ΔT_C and ΔT_{AC}) to accommodate the nonlinearity in Eq. 1 to calculate final estimates of T_C and T_S. The linear form of Eq. 1 is

$$T_{RAD}(\theta) \cong f(\theta)T_{C,Lin} + [1 - f(\theta)]T_{S,Lin}. \tag{A.5}$$

We want to solve Eq. A.4 for $T_{C,Lin}$, which is an initial approximation to T_C. $T_{S,Lin}$ can be eliminated from Eq. A.4 by solving Eq. A.5 for $T_{S,Lin}$ and substituting it into Eq. A.4. Further, T_{AC} can be obtained by equating the right hand side of Eq. 14 to Eq. A.2 resulting in

$$T_{AC,Lin} = T_{C,Lin} - \frac{\Delta R_n R_X}{\rho C_P}\left[1 - 1.3 f_g \frac{S}{S+\gamma}\right] \tag{A.6}$$

and this value substituted into Eq. A.4. The following equation represents the linear approximation to the vegetation temperature:

$$T_{C,Lin} = \frac{\dfrac{T_A}{R_A} + \dfrac{T_{RAD}(\theta)}{R_S[1 - f(\theta)]} + \dfrac{\Delta R_n R_X}{\rho C_P}\left[1 - 1.3 f_g \dfrac{S}{S+\gamma}\right]\left[\dfrac{1}{R_A} + \dfrac{1}{R_S} + \dfrac{1}{R_X}\right]}{\dfrac{1}{R_A} + \dfrac{1}{R_S} + \dfrac{f(\theta)}{R_S[1 - f(\theta)]}} \tag{A.7}$$

where all the quantities have the same definitions as given in the body of the paper except R_X. The total boundary layer resistance of the complete canopy of leaves (R_X) is approximated by

$$R_X = \frac{C'}{F}\left(\frac{s}{U_{d+z_M}}\right)^{1/2} \tag{A.8}$$

where C' is derived from weighting a coefficient in the equation for leaf boundary layer resistance over the height of the canopy (McNaughton and Van den Hurk, 1995), and even though the value of this coefficient is subject to some uncertainty, we use 90 $s^{1/2}\ m^{-1}$ (Grace, 1981). The leaf size is given by s, F is leaf area index, and U_{d+z_M} is given by

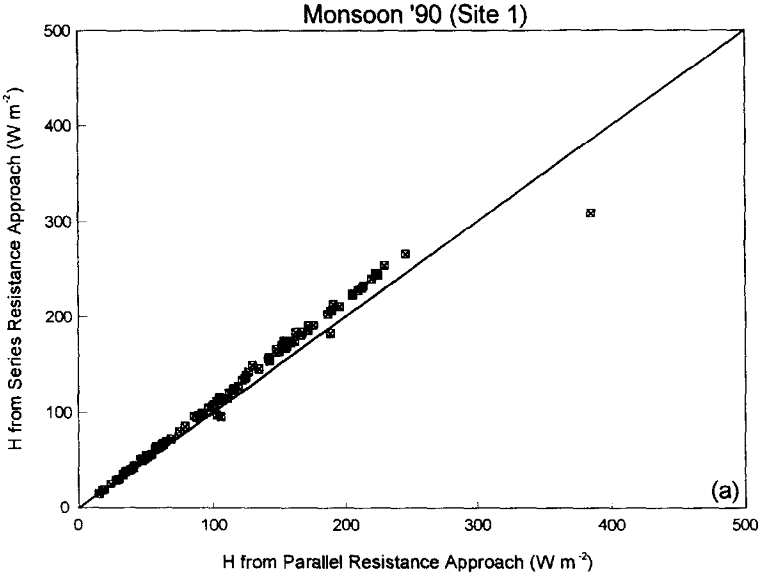
$$U_{d+z_M} = U_C \exp\left[-a\left(1 - \frac{d+z_M}{h_C}\right)\right]. \tag{A.9}$$

The quantities U_c, a, and h_C are defined in Appendix B. Eq. A.8 is related to R_{EX} in Eq. 5 of the main body of the paper (McNaughton and Van den Hurk, 1995). By making the following approximation,

$$(T_{C,Lin} + \Delta T_C)^n = T_{C,Lin}^n + nT_{C,Lin}^{n-1}\Delta T_C, \tag{A.10}$$

which is appropriate when $T_{\mathrm{C,Lin}} \gg \Delta T_{\mathrm{C}}$, a relation can be derived for ΔT_{C}:

$$\Delta T_{\mathrm{C}} = \frac{T_{\mathrm{RAD}}^{n}(\theta) - f(\theta)T_{\mathrm{C,Lin}}^{n} - [1 - f(\theta)]T_{\mathrm{D}}^{n}}{n[1 - f(\theta)]T_{\mathrm{D}}^{n-1}\left(1 + \dfrac{R_{\mathrm{S}}}{R_{\mathrm{A}}}\right) + nf(\theta)T_{\mathrm{C,Lin}}^{n-1}}, \tag{A.11}$$



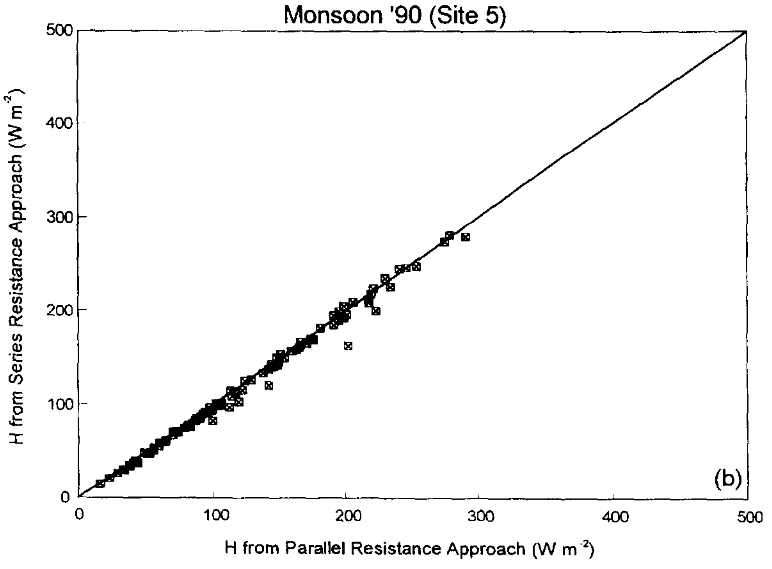


Fig. 12. Comparison of 'parallel' versus 'series' resistance approaches using input data from the Monsoon '90 experiment for Site 1 (a) and Site 5 (b) and from FIFE sites (c). The line represents a 1:1 relationship.

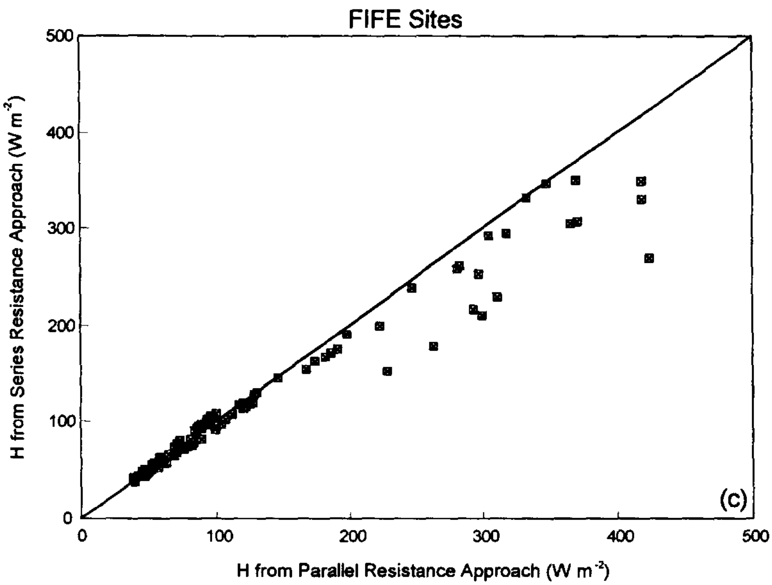


Fig. 12 (continued).

where

$$T_D = T_{C,Lin}\left(1 + \frac{R_S}{R_A}\right) - \frac{\Delta R_n R_X}{\rho C_P}\left[1 - 1.3 f_g \frac{S}{S+\gamma}\right]\left[1 + \frac{R_S}{R_X} + \frac{R_S}{R_A}\right] - T_A \frac{R_S}{R_A}. \tag{A.12}$$

Therefore, the final value of T_C is given by

$$T_C = T_{C,Lin} + \Delta T_C. \tag{A.13}$$

The soil temperature, T_S, is obtained from Eq. 1 in the main body of the paper so that all fluxes can be obtained from T_C, T_S, and T_A.

If the difference between radiometric temperature ($T_{RAD}(\theta)$) and air temperature (T_A) is sufficiently large, then this implies that the soil surface may be dry so that $E_S = 0$ and Eq. A.7 is not valid. A new equation must be derived subject to the constraint that $E_S = 0$ and removing the constraint of Eq. 12. Again the approach is to begin with a linear form of Eq. 1 and make a final adjustment for nonlinearity. If $E_S = 0$, then the soil sensible heat flux is given by

$$H_{S,E_S=0} = R_{n,SOIL} - G = 0.65 R_n \exp[0.9 \ln(1 - f_C)] \tag{A.14}$$

where G is given by Eq. 11. We want to solve Eq. A.4 for $T_{AC,Lin}$. The soil temperature can be obtained in terms of $T_{AC,Lin}$ from a combination of Eq. A.1Eq. A.14, and T_C can be obtained in terms of $T_{AC,Lin}$ by combining Eq. A.1Eq. A.5Eq. A.14 so that

$$T_{AC,Lin} = \frac{\frac{T_A}{R_A} + \frac{T_{RAD}(\theta)}{f(\theta)R_X} - \frac{[1-f(\theta)]}{f(\theta)R_X}\left(\frac{H_{S,E_S=0}R_S}{\rho C_P}\right) + \frac{H_{S,E_S=0}}{\rho C_P}}{\frac{1}{R_A} + \frac{1}{R_X} + \frac{[1-f(\theta)]}{f(\theta)R_X}}.$$

Using the approximation given by Eq. A.10 for T_{AC},

$$\Delta T_{AC} = \frac{T^n_{RAD}(\theta) - [1-f(\theta)]\left[\frac{H_{S,E_S=0}R_S}{\rho C_P} + T_{AC,Lin}\right]^n - f(\theta)T^n_E}{nf(\theta)T^{n-1}_E\left(1+\frac{R_X}{R_A}\right) + n[1-f(\theta)]\left[\frac{H_{S,E_S=0}R_S}{\rho C_P} + T_{AC,Lin}\right]^{n-1}} \tag{A.16}$$

where

$$T_E = T_{AC,Lin}\left[1 + \frac{R_X}{R_A}\right] - \frac{H_{S,E_S=0}R_X}{\rho C_P} - \frac{T_A R_X}{R_A} \tag{A.17}$$

so that

$$T_{AC} = T_{AC,Lin} + \Delta T_{AC}. \tag{A.18}$$

The soil temperature is calculated from

$$T_S = T_{AC} + \frac{H_{S,E_S=0}R_S}{\rho C_P} \tag{A.19}$$

and the vegetation temperature is obtained from Eq. 1. The vegetation sensible heat (H_C) is calculated from Eq. A.2 and transpiration (E_C) from Eq. 18 in the main body of the paper, subject to the constraint that $E_C \geq 0$. If $T_{RAD}(\theta) - T_A$ is very large, then transpiration and soil evaporation may both be zero.

Clearly the solution to the series network shown in Fig. 11 is more complicated than the parallel resistance model discussed in the main body of the paper and shown in Fig. 1. Therefore a comparison of the results from the two models is useful (Fig. 12). The results presented in Fig. 12 are outputs of the 'series' approach versus the 'parallel' approach for the same inputs as used in Figs. 2–10. For Site 1 of the Monsoon '90 data, the comparison indicates a slight bias between the 'series' and 'parallel' approaches yielding an RMSD = 14 W m^{-2}. Interestingly, the tendency for the 'series' model to predict larger values of $\boldsymbol{H}$ results in better agreement with the observations than achieved with the 'parallel' model (Fig. 2). For Site 5 the 'series' and 'parallel' models agree well with a RMSD = 8 W m^{-2}. For the FIFE data the scatter is larger with a RMSD = 32 W m^{-2}, and at $\boldsymbol{H}$ values greater than 200 W m^{-2} the 'series' approach results in smaller values of $\boldsymbol{H}$ than the 'parallel' approach. This result is supported by sensitivity tests where the 'series' network appears to yield more accurate results for dry canopies where evapotranspiration is small. Apparently when the vegetation-soil system is dry, the soil sensible heat exchange most affects the air temperature in the canopy (T_{AC}) and this is captured by the 'series' approach; however, the 'parallel' approach assumes that $\boldsymbol{H}_S$ and T_S do not affect $\boldsymbol{H}_C$ and T_C, and thus it predicts sensible heat fluxes that tend to be too large.

Appendix C

The resistance to transport of heat between the soil surface and a height representing the canopy, such as the canopy displacement height, has been most difficult to characterize. Recently, extensive studies of this soil-surface resistance have been done by Sauer et al. (1995) in a wind tunnel and beneath a corn canopy. Although soil-surface resistances depend on many factors, a reasonable, simplified equation is

$$R_S = \frac{1}{a' + b'U_S} \tag{B.1}$$

where $a' \simeq 0.004$ m s^{-1}, $b' \simeq 0.012$, R_S is in units of s m^{-1}, and U_S is the wind speed in m s^{-1} at a height above the soil surface where the effect of the soil surface roughness is minimal; typically 0.05 to 0.2 m. The coefficients in Eq. B.1 depend on turbulent length scale in the canopy, soil-surface roughness and turbulence intensity in the canopy and are discussed by Sauer et al. (1995). The wind speed just above the soil surface can be computed from the equations of Goudriaan (1977) as

$$U_S = U_C \exp(-a(1 - 0.05/h_C)), \tag{B.2}$$

where the factor a is given by Goudriaan (1977) as

$$a = 0.28 F^{2/3} h_C^{1/3} s^{-1/3}, \tag{B.3}$$

with the mean leaf size s given by four times the leaf area divided by the perimeter.

The term U_C in Eq. B.2 is the wind speed at the top of the canopy, given by

$$U_C = U\left[\frac{\ln\left(\dfrac{h_C - d}{z_M}\right)}{\ln\left(\dfrac{z_U - d}{z_M}\right) - \Psi_M}\right] \tag{B.4}$$

where U is the wind speed above the canopy at height z_U and the stability correction at the top of the canopy is assumed negligible due to roughness sublayer effects. Observations by Garratt (1980) and Cellier and Brunet (1992) show a significant reduction in the magnitude of the diabatic correction terms when applied near the roughness elements.

The use of the exponentially decaying wind profile in the canopy (Eq. A.2) is not always appropriate for wind speed predictions near the soil (Brutsaert, 1982). However, because an estimate of wind speed in the canopy space at a height sufficiently above the soil surface is used, and because it is assumed from Eq. 15 that the air flow above the canopy is influencing the transport of $\boldsymbol{H}_S$ from the soil surface, this approximation is considered reasonable.

The effect of the resistance at the surface of the soil, R_S, can be considerable because it may be several times larger than the aerodynamic resistance above the canopy (R_A), and hence have a considerable influence on the magnitude of R_* defined by Eq. 9. With partial vegetative cover, the soil can contribute much more to the ensemble radiometric temperature than to the aerodynamic temperature; this is apparent by comparing Eqs. 8 and 9 with Eq. 5. Therefore, the model-derived fluxes are significantly influenced by the magnitude of the soil-surface resistance. If vegetation cover is heterogeneous and especially if it is clumped so that relatively large areas are nearly bare soil, then the calculation of U_S from Eq. B.2 may have to be modified so that R_S is somewhat smaller.

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