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Estimating surface energy fluxes using a micro-meteorological model and satellite images

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Abstract

The increasing interest of meteorological, climatic, and hydrological scientific communities in the different components of the surface energy balance, and particularly in the evapotranspiration, has encouraged the development of different micro-meteorological models for estimating surface energy fluxes at a local scale. Recent advances in satellite remote sensing techniques may allow the monitoring of these surface energy fluxes over extensive areas. However, most of the current models require in situ calibration or empirically derived parameters, which limit their operational application at a large scale. The objective of this work is to present a micro-meteorological approach that can be operationally used together with satellite images to monitor surface energy fluxes at a regional scale. Firstly, we introduce the framework and details of the proposed micro-meteorological model, which is based on a two-source patch representation of the soil-canopy-atmosphere system. The feasibility of the model is explored at a local scale using data collected over two completely different ecosystems. On the one hand, data collected over a maize (corn) crop in Beltsville, Maryland, USA, during the 2004 summer growing season. On the other hand, data from an experimental campaign carried out in a boreal forest in Finland in 2002. Comparison of the results with ground measurements shows errors between 15 and 60 W m^{-2} for the retrieval of net radiation, soil heat flux, and sensible and latent heat fluxes in both sites.

Key words: STSEB model, surface energy fluxes, radiometric temperatures

1 Introduction

A very important issue in meteorology, climatology, and hydrology studies is the determination of evapotranspiration, but the direct measurement of this parameter is usually difficult. In recent decades a wide variety of methods have been proposed for estimating it, from the classic methods that allow us to calculate evapotranspiration on a local scale using in situ measurements, to models based on remote sensing techniques.

The direct relationship between surface thermodynamic temperature and energy balance has long been recognized by meteorologists and hydrologists (Idso et al., 1975; Kustas and Norman, 1999). This is the starting point for models based on remote sensing thermal infrared (TIR) data for determining evapotranspiration, since the surface thermodynamic temperature can be obtained from the temperature value directly sensed by the remote instrument, the brightness temperature. The atmosphere emits a downwelling sky radiation reaching the surface and being in part upwards reflected. The atmospheric layer between the surface and the remote instrument absorbs part of this reflected radiation and the radiation directly emitted by the surface. Thus, a correction of the brightness temperature values is required to account for these atmospheric effects. Besides, real surface emissivities (ratio between the radiation emitted by a body and that emitted by a blackbody at the same temperature) are always lower than the unity (blackbody assumption). Thus, an additional emissivity correction is necessary to

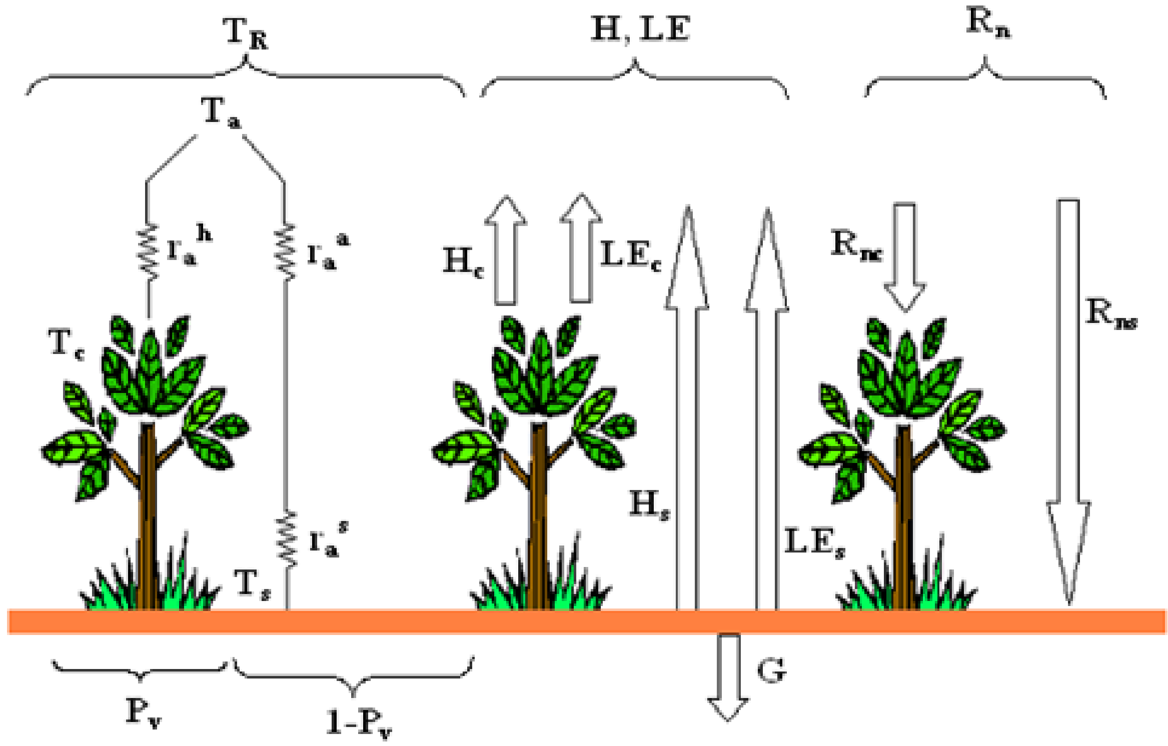

Figure 1. Scheme of resistances and flux partitioning between soil and canopy, corresponding to the STSEB approach. Symbols are defined in the text.

finally obtain the surface thermodynamic temperature from the sensed value of brightness temperature.

Most of these methods are based on principles of energy conservation. The governing equation is thus the Energy Balance Equation (EBE) of the land surface, which models a system formed by vegetation, surrounding soil, and atmosphere:

$$R_n = G + H + LE \tag{1}$$

where R_n is the net radiation flux (W m^{-2}) (balance between the incoming and outgoing long-wave and shortwave radiation), G is the soil heat flux (W m^{-2}) (heat flux through the soil layer underneath the surface), H is the sensible heat flux (W m^{-2}) (heat flux between two different- temperature levels), and LE is the latent heat flux in the atmosphere boundary layer (W m^{-2}) (heat flux between two different-vapour pressure levels). The last two terms, H and LE, are the most difficult to estimate, and, obtaining the sensible heat flux, evapotranspiration can be estimated as a residual from the EBE. Mention at this point that other minor terms such as advection, heat storage in the air and canopy, and photosynthesis have been removed from Equation 1, since it has been shown that their contribution to the energy balance is normally negligible and difficult to reliably estimate with standard micrometeorological measurements (Wilson et al., 2002).

Monteith (1973) proposed an equation based on a convection analogue to Ohm's Law to obtain the sensible heat flux:

$$H = \rho C_p \frac{T_0 - T_a}{r_a} \tag{2}$$

where ρ is the air density (kg m^{-3}), C_p is the air specific heat at a constant pressure (J kg^{-1} K^{-1}), T_0 is the aerodynamic temperature (K) (temperature of the air in contact with the vegetation at the level at which energy exchanges atmosphere-soil-vegetation are produced), T_a is the air temperature (K) at a height z, and r_a is the aerodynamic resistance (s m^{-1}) which depends on the canopy height and the wind speed at height z (Monteith, 1973), and accounts for the resistance of the air to the heat transference between two points showing different temperature. However, the evaluation of this single source model has shown that there is an important limitation since an appreciable deviation appears between aerodynamic, T_0, and surface temperatures, T_s, (obtained from TIR radiometers), for partial canopy cover conditions.

Two-source (soil + vegetation) layer models, developed to accommodate partial canopy cover conditions, consider energy exchange between soil and canopy components, and hence interaction between soil and canopy elements (Shuttleworth and Wallace, 1985). Norman et al. (1995) introduced a remote sensing-based two-source layer modelling frame-

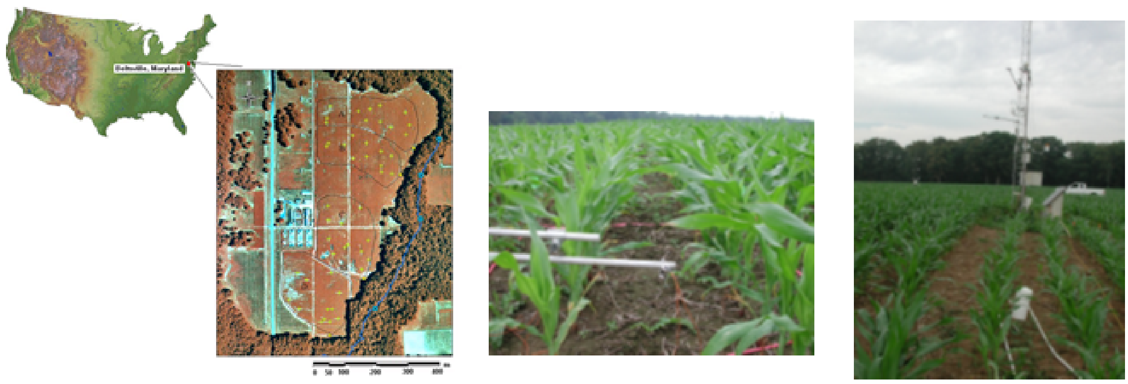

Figure 2. (a) Location and aerial picture of the OPE3 study site. (b) Experimental assembly of two Apogee IRTS-P3 infrared radiometers to measure T_c and T_s. (c) General view of the target and the micro-meteorological tower on which the instrumentation was mounted.

work for computing surface fluxes using directional brightness temperature observations. The Two-Source Energy Balance model (TSEB) was developed to require minimal inputs, similar to single-source models. Since typically only composite brightness temperature observations are available, an additional assumption is required for obtaining initial estimates of soil and vegetation canopy component temperatures and energy fluxes. For the TSEB scheme, the Priestley-Taylor (PT) equation applied to the vegetated canopy is used to obtain an initial solution. The PT equation allows us to estimate the component of the latent heat flux corresponding to the vegetation (transpiration). For that, additional information on vapour pressure, as well as the Priestley-Taylor parameter (α_{PT}), is required. The PT equation results are shown to be very sensitive to α_{PT}, which values depend on the particular surface and climatic conditions. The TSEB model has been widely applied, validated and modified to deal with unique landscapes over the past several years (Kustas and Norman, 1999; Li et al., 2005; French et al., 2003).

Alternatively, if the partitioning of composite land-surface temperature into soil and canopy temperatures is known a priori, e.g., through dual angle Thermal InfraRed (TIR) decomposition (e.g., François, 2002), the soil and canopy latent heat rates can be computed directly as a residual to the component energy budgets. In this case, the PT formulation is no longer required in the TSEB scheme. Encouraged by this fact, Sánchez et al. (2008) have recently proposed a Simplified Two-Source Energy Balance (STSEB) model, based on a patch representation of the energy exchange from soil and canopy, which permits estimation of surface fluxes under partial canopy cover conditions directly from component soil and canopy temperatures. A simple algorithm to predict the net radiation partitioning between soil and vegetation has been also developed as part of the STSEB model. This work focuses on the description and validation of the STSEB model at a local scale under variable surface conditions.

2 Description of the micro-meteorological model

The effective surface temperature in a soil-canopy-atmosphere system, T_R (K), can be obtained as a weighted composite of the soil temperature, T_s (K), and the canopy temperature, T_c (K):

$$T_R = \left[\frac{P_v(\theta)\varepsilon_c T_c^4 + (1 - P_v(\theta))\varepsilon_s T_s^4}{\varepsilon} \right]^{1/4} \tag{3}$$

where ε_c and ε_s are the canopy and soil emissivities, respectively, ε is the effective surface emissivity, and $P_v(\theta)$ is the fractional vegetation cover for the viewing angle θ. Note that Equation 3 is based on the Stefan- Boltzmann law.

The angular vegetation cover fraction $P_v(\theta)$ can be estimated from the measurements of Leaf Area Index (LAI) via:

$$P_v(\theta) = 1 - \exp\left(\frac{-0.5\Omega(\theta)\mathrm{LAI}}{\cos\theta} \right) \tag{4}$$

where $\Omega(\theta)$ is a clumping factor to characterize the heterogeneity of the surface (Anderson et al., 2005). This factor depends on the features of the vegetation structure, and it makes it possible to extend the typical equations for random canopies to heterogeneous cases.

The partitioning of the different fluxes into soil and canopy components was accomplished according to the scheme shown in Figure 1. In the patch approach, an analogy between Ohm's law and the heat transfer equation shows that the magnitude corresponding to H is the current density (intensity per unit area). Therefore, according to this configuration, the addition between the soil and canopy contributions (values per unit area of component) to the total sensible heat flux, H_s and H_c, respectively, are weighted by their respective partial areas as follows:

$$H = P_v H_c + (1 - P_v) H_s \tag{5}$$

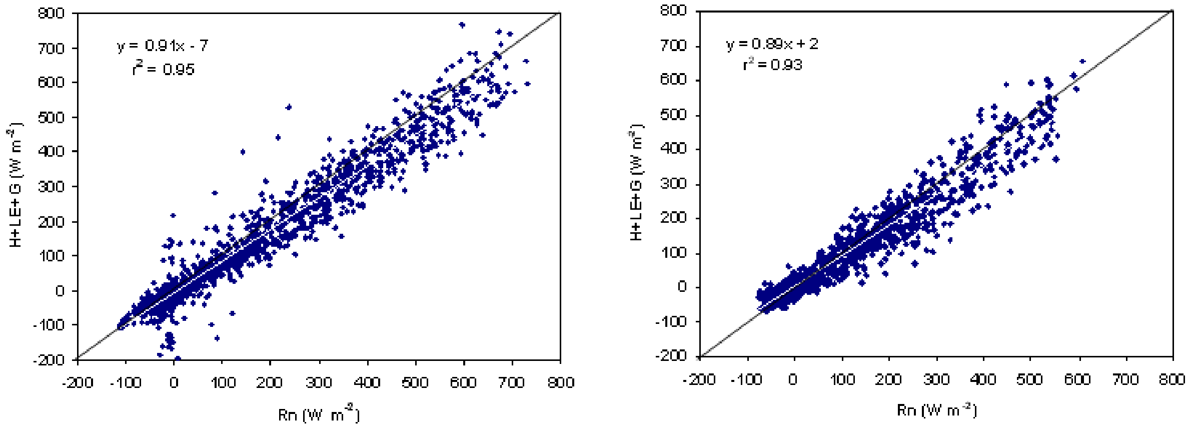

Figure 3. Linear regression between the two terms of the Energy Balance Equation: (a) (left) corn crop, (b) (right) boreal forest.

where P_v (without a view angle argument) refers to the cover fraction at nadir view (i.e. $\theta = 0°$). In Equation 5, H_s and H_c are expressed as:

$$H_c = \rho C_p \frac{T_c - T_a}{r_a^h} \tag{6}$$

$$H_s = \rho C_p \frac{T_s - T_a}{r_a^a + r_a^s} \tag{7}$$

where ρC_p is the volumetric heat capacity of air (J K^{-1}m^{-3}), T_a is the air temperature at a reference height (K), r_a^h is the aerodynamic resistance to heat transfer between the canopy and the reference height at which the atmospheric data are measured (s m^{-1}), r_a^a is the aerodynamic resistance to heat transfer between the point $z_{0M} + d$ (z_{0M}: canopy roughness length for momentum, d: displacement height) and the reference height (s m^{-1}), r_a^s is the aerodynamic resistance to heat flow in the boundary layer immediately above the soil surface (s m^{-1}). A summary of the expressions to estimate these resistances is shown in the Appendix.

To be consistent with the patch model configuration, a partitioning of the net radiation flux, R_n, between the soil and canopy is proposed as follows:

$$R_n = P_v R_{nc} + (1 - P_v) R_{ns} \tag{8}$$

where R_{nc} and R_{ns} are the contributions (values per unit area of component) of the canopy and soil, respectively, to the total net radiation flux. They are estimated by establishing a balance between the long-wave and the short- wave radiation separately for each component:

$$R_{nc} = (1 - \alpha_c) S + \varepsilon_c L_{sky} - \varepsilon_c \sigma T_c^4 \tag{9}$$

$$R_{ns} = (1 - \alpha_s) S + \varepsilon_s L_{sky} - \varepsilon_s \sigma T_s^4 \tag{10}$$

where S is the solar global radiation (W m^{-2}), α_s and α_c are soil and canopy albedos, respectively, σ is the Stefan-Boltzmann constant, and L_{sky} is the incident long-wave radiation (W m^{-2}).

A similar expression to Equation 5 is used to combine the soil and canopy contributions, LE_s and LE_c, respectively, to the total latent heat flux:

$$LE = P_v LE_c + (1 - P_v) LE_s \tag{11}$$

According to this framework, a complete and independent energy balance between the atmosphere and each component of the surface is established, from the assumption that all the fluxes act vertically. In this way, the component fluxes to the total latent heat flux can be written as:

$$LE_c = R_{nc} - H_c \tag{12}$$

$$LE_s = R_{ns} - H_s - \frac{G}{(1 - P_v)} \tag{13}$$

Finally, the direct relation observed traditionally between the heat transference in the soil and the available energy corresponding to this component makes it possible that G can be estimated as a fraction (C_G) of the soil contribution to the net radiation (Choudhury et al., 1987):

$$G = C_G (1 - P_v) R_{ns} \tag{14}$$

where C_G can vary in a range of 0.2 - 0.5 depending on the soil type and moisture. Recent studies have also expressed C_G as a function of time to accommodate temporal variation in this fraction.

In the STSEB scheme there is a real weighting of the soil and canopy elements and no direct coupling is allowed between soil and vegetation. Besides, the STSEB net radiation model does not consider attenuation of the downwelling sky and upwelling soil emission by an intervening canopy layer.

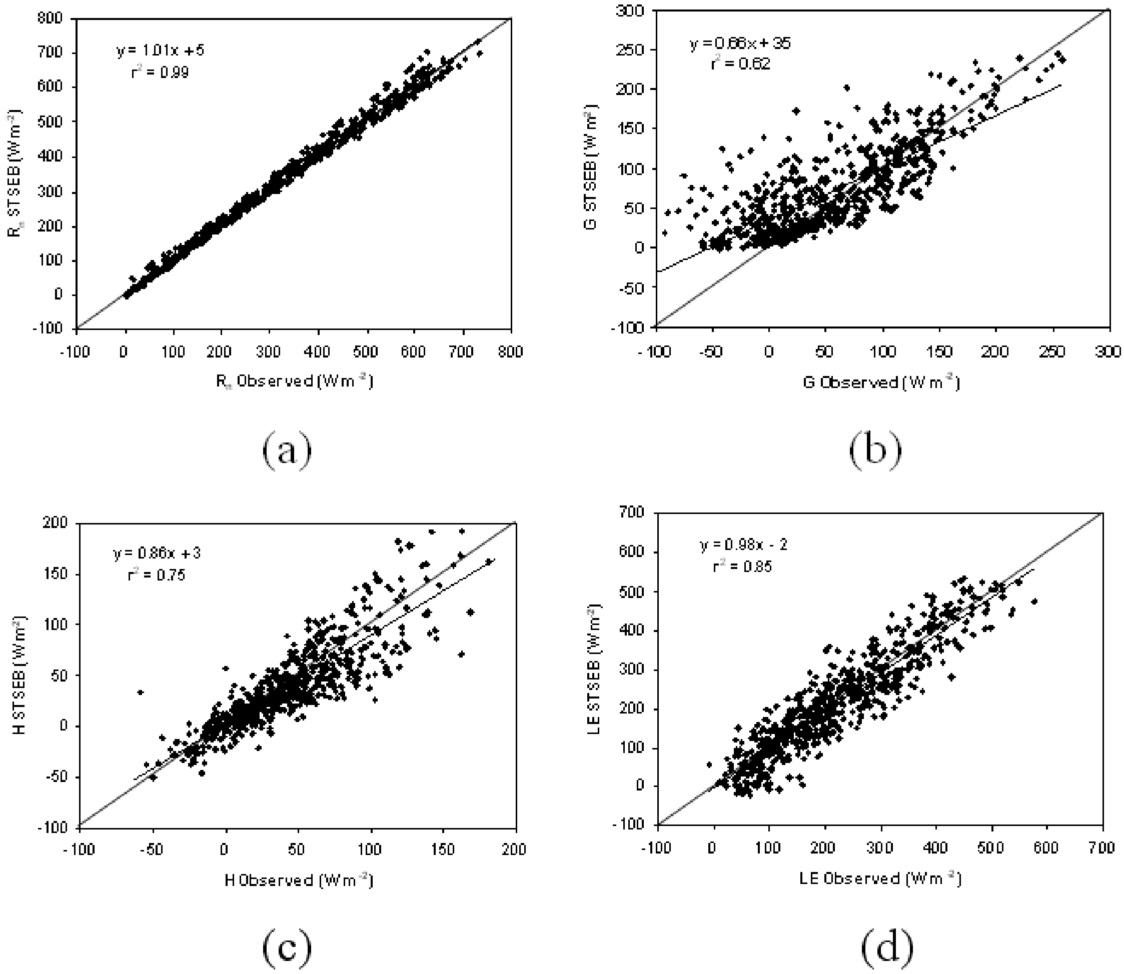

Figure 4. Linear regressions between the surface energy fluxes estimated by the STSEB model versus their corresponding ground measured values: (a) R_n, (b) G, (c) H (eddy-covariance measurements), (d) LE (RE technique applied).

3 Validation of the model

Firstly, the STSEB micro-meteorological model was validated at a local scale. For that, two completely different sites and ecosystems were selected, a corn crop field in USA and a boreal forest in Finland.

3.1 Validation in a corn crop field

3.1.1 Study site and measurements

Data were registered in a corn crop field associated with the Optimizing Production Inputs for Economic and Environmental Enhancement (OPE3) program, located at the USDA-ARS Beltsville Agricultural Research Center, Beltsville, Maryland (39°01'00"N, 76°52'00"W, 40 m above sea level) (Figure 2a). In this work we will focus on the experimental campaign carried out in the summer of 2004, encompassing all the stages in the corn growing season, from the beginning of June (plant emergence) to the end of July (cob formation).

Starting on June 9th, soil and canopy radiometric temperatures were measured simultaneously using Apogee IRTS- P3 infrared radiometers. This radiometer has a broad thermal band (7-14 μm) with an accuracy of 0.3°C, and 37° field of view. Soil temperature was measured with an Infrared Thermometer (IRT) mounted in the center of a row at an oblique angle (~45°) viewing parallel to the row crop. It was placed at a height appropriate to ensure the view of just the soil space between rows. Canopy temperature was sensed with a second IRT placed within the row and oriented horizontally, viewing the plants parallel to the row orientation (Figure 2b). The horizontal orientation ensured that this IRT was viewing only vegetation. Both temperature components were measured at two separated locations in the corn field, using two pairs of radiometers. Concurrently, the effective composite temperature of the corn+soil system was measured by a fifth IRT placed on a tower at 4.5 m height, viewing the surface at approximately a 45° viewing angle and an azimuth view perpendicular to the row direction. The

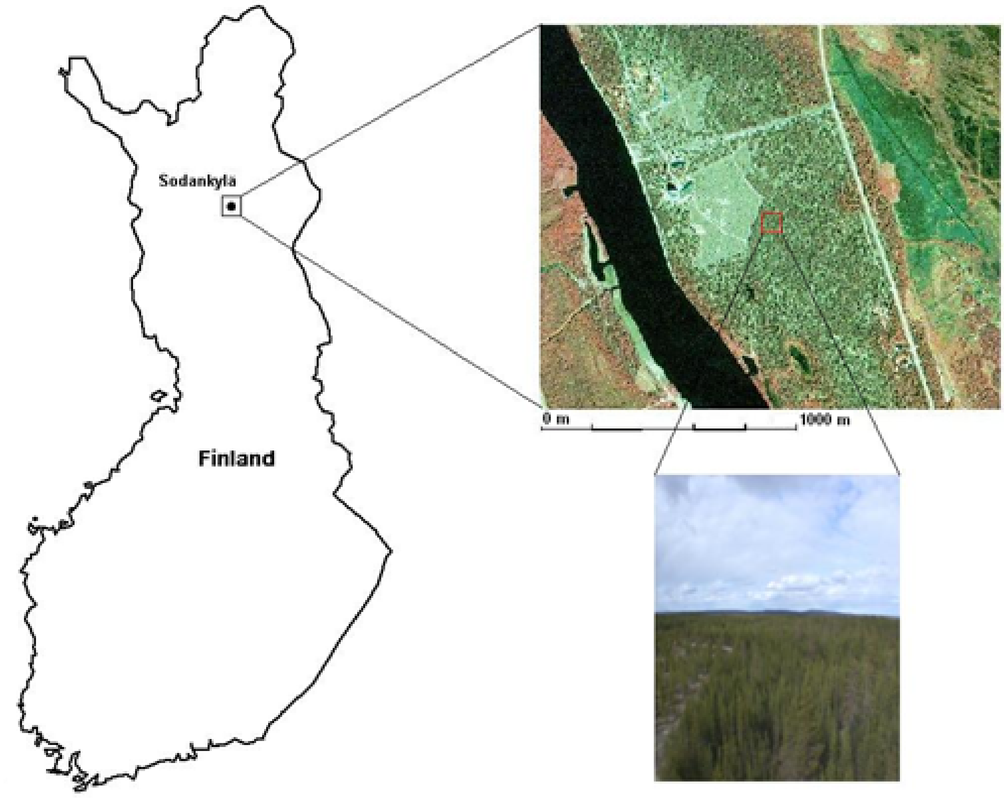

Figure 5. Location and aerial picture of the SIFLEX experimental site.

micro-meteorological and eddy covariance instrumentation were mounted on the same 10-m tower (Figure 2c).

Net radiation was measured with a Kipp & Zonen CNR-1 net radiometer at 4.5 m above ground level (agl). This net radiometer measures separately the incoming and outgoing shortwave and long-wave radiation components. Six REBS soil heat flow transducers (HTF-1) were buried 6-cm below the soil surface. Soil temperatures were measured at 2 and 4-cm depth by two Type-T soil thermocouples to compute the storage component of the soil heat flux above the plates. A Campbell Scientific 3-D sonic anemometer and LiCor 7500 water vapour/carbon sensor positioned at 4-m above ground level was used to measure momentum, sensible heat, latent heat and carbon fluxes, as well as wind speed and direction. The eddy covariance instrumentation was oriented to the southwest, the predominant wind direction during the summer growing season. Unfavourable winds out of the north compromising the sonic measurements were flagged by the sensor system and discarded during post processing. Air temperature and vapour pressure were measured using a CSI HMP 45C sensor at 4-m above ground level. The sampling frequency was 10 Hz for the eddy covariance and 10 s for the energy balance and meteorological instrumentation. All data were stored as 30-minute averages on Campbell CR5000 and 23x data loggers.

Canopy geometry and LAI were sampled periodically during corn development. Also, canopy height was sampled weekly.

3.1.2 Processing and results

A lack of closure of the energy balance has been observed in numerous eddy covariance studies conducted over different landscapes. Possible causes of observed energy imbalances include instrumental effects, corrections applied when post- processing of the turbulence data, the length of the sampling interval, and the heterogeneity of the landscape (Wilson et al., 2002). Analysis of the energy balance over corn crops have shown closure ratios around 0.8, although 0.9 is attainable. In Figure 3a a linear regression between R_n and the sum $H + LE + G$ from Equation 1 for the OPE3 dataset yields a slope $\sim$0.9, which indicates approximately 10% of the estimated available energy is not accounted for, on average. Twine et al. (2000) suggested two methods to enforce the energy balance closure, namely by calculating the latent heat flux as a residual of the energy balance (RE method), and by conserving the measured Bowen ratio (BR method). In the RE method, the direct eddy-covariance measurements of H are assumed reliable and lack of closure is largely due to an under-measurement of LE. In the BR method, it is assumed both turbulent fluxes are under

Table 1. Statistical analysis of the STSEB model performance with the daytime OPE3-2004 dataset. RMSD (Root Mean Square Difference), MAD (Mean Absolute Difference), a (slope of the linear regression), b (interception of the linear regression), r^2 (regression coefficient).

Flux	Bias (W m^{-2})	RMSD (W m^{-2})	MAD (W m^{-2})	a	b (W m^{-2})	r^2
R_n	+8	18	13	1.01	5	0.99
G	+17	40	30	0.66	35	0.62
H_{EC}	-3	22	16	0.86	3	0.75
H_{BR}	-10	26	19	0.76	2	0.74
LE_{EC}	+30	60	50	1.04	22	0.82
LE_{RE}	-6	50	40	0.98	-2	0.85
LE_{BR}	0	50	40	1.00	0.6	0.85

measured with the amount distributed between H and LE based on the Bowen ratio (H/LE), which defines the fraction of available energy going into sensible versus latent heat. Since there is no consensus on how to resolve lack of energy balance closure with eddy covariance, comparison between model estimations and ground observations will be performed without closure and enforcing closure using both the RE and BR techniques.

In this work only daytime fluxes (when $R_n > 0$) have been considered. Daytime flux statistics are more descriptive of overall model utility since modelled nighttime flux estimates are constrained to be near zero. For estimating net radiation, Equations 8, 9 and 10 were applied using values of α_s = 0.12, α_c = 0.20, ε_s = 0.960, and ε_c = 0.985 (Campbell and Norman, 1998), characteristic of a corn canopy. The model reproduces measured net radiation with good accuracy, yielding a bias of +8 W m^{-2}, and RMSD = 18 W m^{-2} (Figure 4a). A constant value of C_G = 0.35 was used in Equation 14, corresponding to the midpoint between its likely limits (Choudhury et al., 1987). Soil heat flux results overestimate measurements by 17 W m^{-2} on average, with RMSD = 40 W m^{-2} (Figure 4b).

Table 1 lists statistics comparing turbulent fluxes estimates of H and LE with the eddy covariance fluxes in their original form (EC), and corrected for closure using the residual (RE) and Bowen ratio (BR) techniques. The RE closure technique, using H_{EC} and assigning all closure error to LE (LE_{RE}), yields the best agreement between STSEB and measured fluxes. Several studies with the TSEB model have also found optimal agreement using the RE method (e.g., Li et al., 2005).

Model comparisons with H_{EC} and LE_{RE} are shown in Figure 4c-d. After correcting the aerodynamic resistances for atmospheric stability, as described in the Appendix, H was estimated via Equations 5, 6, and 7. Comparisons between modelled and measured H show a negative bias of -3 W m^{-2}, and a RMSD of 22 W m^{-2} (Figure 4c). The slope (a) of a linear regression between STSEB H and H_{EC} is 0.86, indicating that the bias is multiplicative. The bias is further exacerbated when the Bowen ratio closure technique is applied, yielding a = 0.76.

For LE obtained by using Equations 11, 12, and 13, there is a tendency to overestimate the observed latent heat flux, LE_{EC} with a slope of 1.04 and a RMSD = 60 W m^{-2}. This overestimation may be due in part to an under-measurement problem with the eddy covariance system. The agreement of the LE results improves significantly when the RE closure technique is applied to the observations, decreasing the slope to 0.98 and the RMSD to 50 W m^{-2} (see Figure 4d and Table 1). If energy closure is enforced by the Bowen ratio technique (LE_{BR}), a similar slope is obtained (a = 1.00), but now with a null bias and a RMSD of 50 W m^{-2} (see Table 1).

Under conditions of very low and very high vegetation cover, the scene is relatively homogeneous, and the STSEB model formulation should approximate a single source evaluation. However, under high cover, observations of T_s will be ill-constrained, whereas T_c is difficult to measure accurately under near-bare-soil conditions. To assess the performance of the STSEB under these potentially challenging limiting conditions, the experiment reported above was repeated considering only the period with 0.05$< P_v <$0.75. No significant improvement in the results for the STSEB validation was observed for this restricted time period, suggesting that the performance of the STSEB model does not degrade under low and high vegetation cover conditions.

3.2 Validation in a boreal forest

3.2.1 Study site and measurements

This experiment was carried out as part of the SIFLEX-2002 project. The measurement campaign was performed from April to June of 2002 at Sodankylä, in a northern boreal forest area of Finland. Sodankylä belongs to the Coordinated Enhanced Observing Period (CEOP) reference sites of the Global Energy and Water Cycle Experiment (GEWEX) of the World Climate Research Programme, and to the CO_2 flux station network of the CARBOEUROPE project (Figure 5).

Figure 6. (a) Soil covered by lichens, understory level below the trees. (b) Tower placed in the site. (c) Thermal sensor placed in the tower looking at the target.

The study area was placed at the Arctic Research Center, which belongs to the Finnish Meteorological Institute (FMI) at Sodankylä (67°21'42.7"N 26°38'16.2"E, 179 m above sea level), 100 km north of the Arctic Circle. This is a boreal forest area, with pines (Pinus sylvestris L.) of more than 10 m in height and 100 years old in average, and a tree density of about 2100 trunks per ha. The soil type is fluvial sandy podzol where lichens, cowberry and crowberry are common (see Figure 6a).

The meteorological variables were obtained from a 48 m height micro-meteorological tower placed in the site (Figure 6b). Air temperature and humidity were measured at different levels with a Vaisala HMP 45 sensor. Wind speed data was collected by a Vaisala WAA252 anemometer placed at a height of 23 m. Other weather parameters such as the fraction of cloud cover or the cloud height were estimated by human observation at the FMI-Arctic Research Center. Besides, atmospheric radiosoundings (Vaisala RS-80) of temperature, humidity and wind were launched twice a day at the study site.

LAI measurements were made twice at 12 points distributed around the study site. It was measured using a LICOR LAI-2000. A time series of soil and canopy albedo values was also registered during the campaign.

Thermal radiance measurements of pine trees were performed every 10 minutes by a multi-channel thermal infrared radiometer CIMEL ELECTRONIQUE CE 312 multi-channel. The CIMEL was placed on a tower observing the target from a height of 12 m, to assure a homogeneous view of the vegetation (Figure 6c). This radiometer has four spectral channels: one broadband, 8 - 14 μm (band 1), and three narrow channels, 11.5-12.5 μm, 10.5 -11.5 μm, and 8.2 - 9.2 μm (bands 2, 3, and 4, respectively). Temperature of the other component (vegetation covering the soil below the trees in this case) was inferred using Equation 3 and the outgoing long-wave radiation. Therefore, direct measurements of this variable can substitute the measurements taken by a thermal radiometer. This possibility increases the operativity of the STSEB model. Samples of pine branches and vegetation covering the soil below the trees were collected to measure their emissivity by means of the CE-312 radiometer using the Box Method (Rubio et al., 2003). Radiosounding data for several days with different atmospheric conditions were introduced into the MODTRAN (MODerate resolution atmospheric TRANsmission) 4 code to get estimates of the atmospheric parameters required to correct the brightness temperatures of atmospheric effects (transmittance, upwelling sky radiance and downwelling sky radiance). MODTRAN is a computer program designed to model atmospheric propagation of radiation in a wide spectral region.

Sensible and latent heat fluxes were measured with a SATI-3Sx sonic anemometer and a platinum thermal probe by using eddy-covariance methodology with an uncertainty in flux estimation of 15 - 20%. Measurements were taken at 23 m in height. Global and reflected short wave radiation were measured by a Kipp&Zonen CM11 sensor, and the incoming and outgoing long-wave components were registered by an Eppley Precision infrared radiometer. Net radiation was measured by a REBS Q-7 sensor. All these sensors were mounted at 46 m height. Soil heat flux was measured by a single HFT3 soil heat flux plate at a depth of 7 cm. Soil temperature was measured at 2, 5, 10, 20, 50, and 100 cm depth by using a set of thermocouples, and volumetric soil moisture at 5, 10, 20, 30, and 50 cm through Delta-T TDR-probes. The heat stored in the soil profile above the plate, computed from the temporal change in soil temperature and soil water content, was also considered. All data were recorded every 30 minutes.

Although the field campaign was conducted from April to June of 2002, this analysis is based on the data taken from

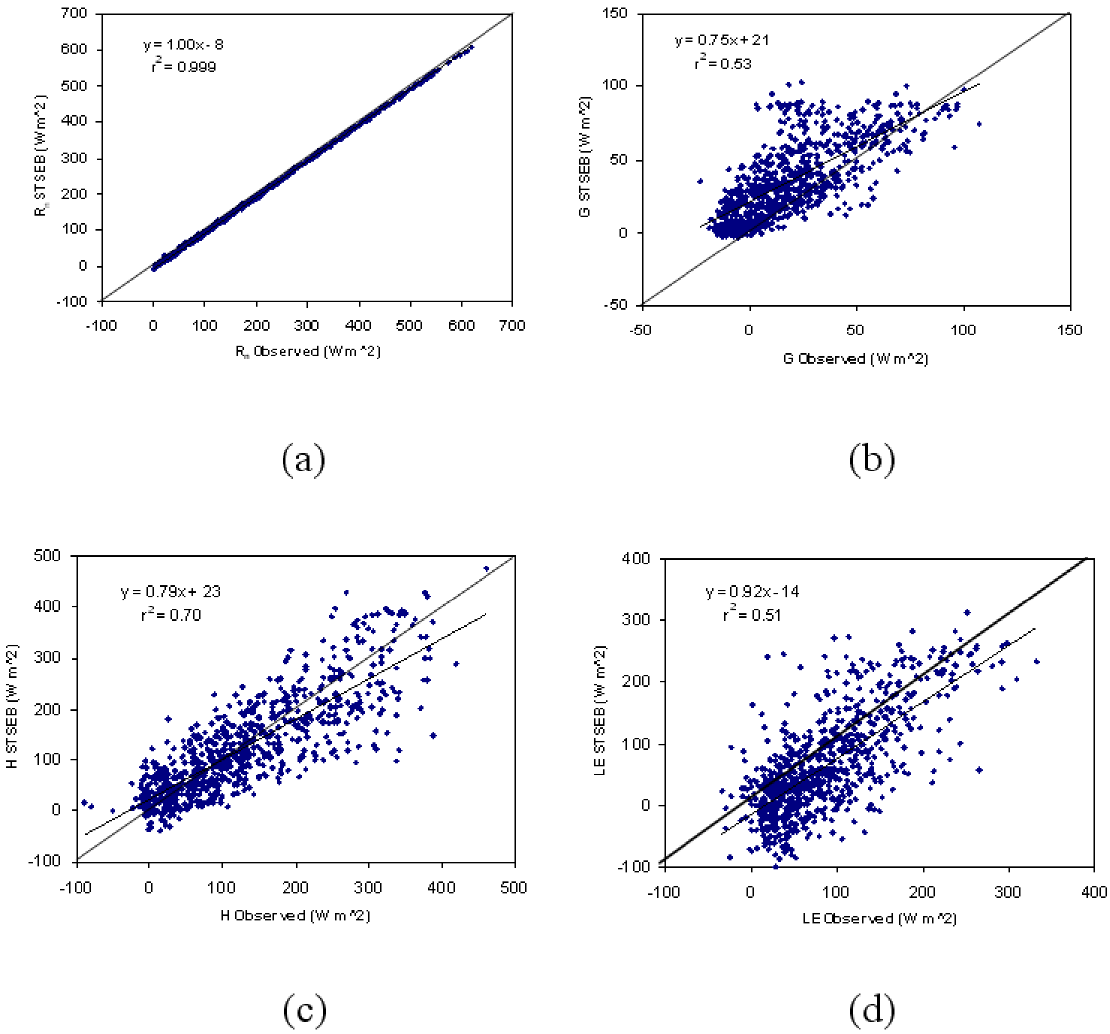

Figure 7. Linear regressions between the surface energy fluxes estimated by the STSEB model versus their corresponding ground measured values: (a) R_n, (b) G, (c), H (eddy-covariance measurements), (d) LE (RE technique applied).

the 5th of May in order to avoid additional complexities due to snow cover.

3.2.2 Processing and results

Modelling surface energy fluxes over these ecosystems is of great interest and importance since forests occupy an extension of about 30% of the terrestrial surface and, specifically, 11% is covered by boreal forests. The remote location and harsh climate of boreal forest kept them elusive to first studies on energy exchange and energy balance. Advances in infrastructures and technology allowed subsequent inclusion of these ecosystems in the scientific community. Testing a model under the conditions of SIFLEX is a challenge, not only because of the high latitude of the site, but also because evident limitations on the instrumentation assembly. In this configuration the surface layer is located at the level of the trees height, approximately 10 m, and according to the relations shown in the appendix, the displacement height and the roughness length result 6.7 and 1 m, respectively. Under these conditions the turbulent flux observations must be taken from a reference height, z, so that the distance to the surface layer is appropriate to characterize the energy exchanges with the atmospheric layer. In this work, as common in boreal forest sites (see FLUXNET website, http://www.fluxnet.ornl.gov/fluxnet/index.cfm), z was set to 23 m.

A slope ∼0.9 is observed in the energy balance closure for the case of the boreal forest, which again indicates that approximately 10% of the estimated available energy is not accounted for, on average (Figure 3b). This result is similar to the energy closure study conducted for the FLUXNET network by Wilson et al. (2002).

For estimating net radiation Equations 9 and 10 were applied using measured values of $\alpha_s = 0.13$, $\alpha_c = 0.08$, $\varepsilon_s = 0.953$, and $\varepsilon_c = 0.978$. Figure 7a shows the results of

Table 2. Statistical analysis of the STSEB model performance with the daytime SIFLEX-2002 dataset.

Flux	Bias ($W m^{-2}$)	RMSD ($W m^{-2}$)	MAD ($W m^{-2}$)	a	b ($W m^{-2}$)	r^2
R_n	-8	9	8	1.00	-8	0.999
G	+16	25	19	0.75	21	0.53
H_{EC}	-3	60	40	0.79	23	0.70
H_{BR}	-21	70	50	0.68	24	0.67
LE_{EC}	+6	70	50	1.30	-12	0.35
LE_{RE}	-21	60	50	0.92	-14	0.51
LE_{BR}	-3	60	50	1.39	-30	0.48

Equation 8 versus the measured values of R_n. A quantitative analysis of this regression is shown in Table 2. An underestimation of 8 W m^{-2}, and a RMSD of 9 W m^{-2} are observed in the total R_n for the campaign data. Regarding the soil heat flux, G, an overestimation of 16 W m^{-2}, and a RMSD of 25 W m^{-2} for the whole period were obtained (see Figure 7b). Equations 5 and 11 for estimating sensible and latent heat flux, respectively, were applied to each data of the dataset. Previously, the aerodynamic resistances in Equations 6 and 7 were obtained as described in the Appendix, and properly corrected for unstable or stable conditions. The quantitative analysis of this validation is summarized in Table 2. For the sensible heat flux, a mean underestimation of 3 W m^{-2}, a RMSD of 60 W m^{-2}, and a correlation coefficient of r = 0.84 were obtained (see also Figure 7c). For the latent heat flux, STSEB results show an underestimation of 21 W m^{-2}, and a RMSD = 60 W m^{-2} (Figure 7d) when the residual method is applied for the closure.

Comparing these results with those obtained in the corn crop, similar estimation errors are observed in LE, while those corresponding to H are higher in the boreal forest. At this point, it should be noted that the energetic exchange between the surface and the atmosphere is dominated by the evapotranspiration in the corn crop, and the sensible heat flux in the boreal forest conditions. Overall, the regression parameters are worse in the boreal forest, which may be due to both inaccuracies in the flux observations or uncertainties in the values of the soil temperature estimates.

4 Conclusions

In this work, the STSEB model has been introduced to estimate surface energy fluxes over sparse and heterogeneous canopies from radiometric surface temperatures and micrometeorological data. The strongest point of this model is that the Priestley-Taylor approach is not required in the equations scheme. Thus, problems linked to a non appropriate selection of the Priestley- Taylor constant are skipped. Also, the equation system presented to partition the net radiation between the soil and the vegetation is a simplified version of that used in the TSEB scheme. On the other hand, the STSEB model requires input measurements of both canopy and soil temperatures. The model has been tested at a local scale using two different ecosystems. On the one hand, under a full range of crop cover conditions using field data from a corn field at the USDA-ARS OPE3 experimental site in Beltsville, Maryland, USA, on the other hand, in a northern boreal forest area of Finland at the SIFLEX-2002 experimental site. Comparison of the results with ground- measured fluxes shows errors between 15 W m^{-2} and 60 W m^{-2} for R_n, G, H, and LE. These results are in reasonable agreement, and improve in some cases, with those obtained by the TSEB model shown in the recent literature.

In summary, these results demonstrate the utility of the STSEB model for a corn crop over a full range in cover conditions and a boreal forest when reliable measurements of soil and canopy temperatures are available. Further studies are needed to assess the utility of the STSEB model over semiarid conditions, for example, as well as the inclusion in the STSEB scheme of algorithms developed to use dual-angle satellite based radiometric temperature observations.

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Appendix A Expressions to estimate the aerodynamic resistances in the STSEB approach

The aerodynamic resistances used for this work are based on the general framework described in Norman et al. (1995), Kustas and Norman (1999) and Li et al. (2005), and adapted to modifications proposed by Brutsaert (1999). The aerodynamic resistance to heat transfer between the canopy and the reference height (z), r_a^h, is expressed as follows:

$$r_a^h = \frac{\left[Ln\left(\frac{z_u-d}{z_{0M}}\right) - \Psi_M\left(\frac{z_u-d}{L}\right) + \Psi_M\left(\frac{z_{0M}}{L}\right)\right] \cdot \left[Ln\left(\frac{z_r-d}{z_{0H}}\right) - \Psi_H\left(\frac{z_r-d}{L}\right) + \Psi_H\left(\frac{z_{0H}}{L}\right)\right]}{k^2 u} \quad \text{(A1)}$$

where, z_u and z_T are the measurement heights (m) for wind speed, u (m s^{-1}), and air temperature, respectively, d is displacement height (m), z_{0M} is the canopy roughness length for momentum (m), z_{0H} is the canopy roughness length for heat (m), and k is the Von Karman constant (≈0.41). The displacement height and the canopy roughness lengths are estimated by simplified expressions as functions of canopy height, h (m): $d = 2h/3$, $z_{0M} = h/10$, and z_{0H} is taken as a fraction of z_{0M} ($z_{0H} = z_{0M}/7$) to account for less efficient transport of heat versus momentum near the canopy elements. The stability functions for heat, Ψ_H, and for momentum, Ψ_M, are obtained from Brutsaert (1999).

A1 Unstable conditions

$$\Psi_M(y) = Ln(a+y) - 3by^{1/3} + \frac{ba^{1/3}}{2} Ln\left[\frac{(1+x^2)}{1-x+x^2}\right] + 3^{1/2} ba^{1/3} \tan^{-1}\left(\frac{2x-1}{3^{1/2}}\right) + \Psi_0 \quad \text{(A2)}$$

$$\Psi_H(y) = \frac{1-d}{n} Ln\left(\frac{c+y^n}{3^{1/2}}\right) \quad \text{(A3)}$$

in which $x = (y/a)^{1/3}$, and $y = -(z-d)/L$. The symbol Ψ_0 denotes a constant of integration, given by $\Psi_0 = (-Ln(a) + 3^{1/2} ba^{1/3}\pi/6)$. The parameters a, b, c, d, and n are assigned constant values of 0.33, 0.41, 0.33, 0.057, and 0.78, respectively (Brutsaert, 1999).

A2 Stable conditions

$$\Psi_M(y) = \Psi_H(y) = 5y \quad \text{(A4)}$$

L is the Monin-Obukhov length (m) and is expressed as:

$$L = \frac{-{u^*}^3 \rho}{kg\left[\left(\frac{H}{T_a C_p}\right) + 0.61E\right]} \quad \text{(A5)}$$

where u^* is the friction velocity, ρ is the air density (kg m^{-3}), g is the acceleration of gravity (m s^{-2}), C_p is the air specific heat at constant pressure (J kg^{-1} K^{-1}), H is the sensible heat flux, and E is the rate of surface evaporation (kg m^{-2} s^{-1}).

The aerodynamic resistance to heat transfer between point $z_{0M} + d$ and the reference height, r_a^a, is written as a simplified form of Equation A1. Since the transport of heat and momentum is equally efficient, in this case $z_{0M} = z_{0H}$ (see Kustas and Norman, 1999). Also, in this case $z_u = z_T$. Finally, r_a^a is given by the expression:

$$r_a^a = \frac{\left[Ln\left(\frac{z_u-d}{z_{0M}}\right) - \Psi_M\right]\left[Ln\left(\frac{z_u-d}{z_{0M}}\right) - \Psi_H\right]}{k^2 u} \quad \text{(A6)}$$

Finally, the aerodynamic resistance to heat flow in the boundary layer immediately above the soil surface, r_s^a, is estimated from an empirical expression developed by Sauer et al. (1995) from extensive studies of this soil- surface resistance in a wind tunnel and beneath a corn canopy. This expression was modified and improved later by Kustas and Norman (1999):

$$r_a^s = \frac{1}{0.0025(T_s - T_c)^{1/3} + 0.012u_s} \quad \text{(A7)}$$

where u_s is the wind speed at height above the soil surface where the effect of soil surface roughness on the free wind movement can be neglected, z' (m s^{-1}) (Sauer et al., 1995). This wind speed is determined assuming a logarithmic wind profile in the air space above the soil:

$$u_s = u\left[\frac{Ln\left(\frac{z'}{z'_0}\right)}{Ln\left(\frac{z_u}{z'_0}\right) - \Psi_M}\right] \quad \text{(A8)}$$

where z'_0 is the soil roughness length. In this expression the displacement height was zero while stability corrections were really not used because of the close proximity to the soil surface (Sauer et al., 1995). Unlike in two-layer schemes, the exponential wind profile in the canopy air space is not applied in the patch approach.

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