Impact of soil vertical water movement on the energy balance of different land surfaces

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Abstract The soil heat flux determination method proposed by Gao (Boundary-Layer Meteorol 114:165–178, 2005) is discussed for (1) dry surfaces, (2) bare soil or sparse short-grass lands, and (3) dense-grass surfaces or forest. Our analysis shows that, when neglecting the contribution of soil vertical water movement to soil heat flux, the energy components measured independently will (1) still achieve balance over dry surfaces, and (2) be significantly in imbalance over bare soil or sparse short-grass lands. The mean of bare ground evaporation modeled by SiB2 is $1.58 \times 10^{-5} \text{ m}^3 \text{ s}^{-1} \text{ m}^{-2}$, and the mean of soil water flux obtained by the method of Gao is $1.22 \times 10^{-5} \text{ m}^3 \text{ s}^{-1} \text{ m}^{-2}$ for the Naqu site in the summer of 1998. Comparison of the bare ground evaporation with the mean of soil water flux shows a difference, the causes of which are investigated. Physically, the bare ground evaporation is equal to the sum of soil water flux and water content change in the soil surface layer. Because the bare ground evaporation is very limited for the dense-grass surfaces or forest, our analysis implies that the energy imbalance encountered over the dense-grass or forest is not caused by the fact that previous researchers neglected soil water movements in their energy budget analyses.

Keywords Land surface energy balance · Soil heat flux · Dry surface · Thermal convection · Soil water movement

Introduction

Micrometeorologists usually examine the land surface energy balance at a tower flux site, using the ratio of residual $Re (= Rn - H - LE - G_0)$ to available energy (i.e., the difference between the net radiation and soil-canopy surface heat flux, $Rn - G_0$) (e.g., Liu and Foken 2001), where Rn , H , LE , and G_0 are the net radiation, sensible heat, latent heat and soil heat fluxes, respectively. Table 1 shows some of the results given in previous literature. It is found that (1) the surface energy balance effectively achieves closure over dry surfaces; (2) the surface energy imbalance is significant for bare soil or short-grass land surfaces, and the values of $Re/(Rn - G_0)$ are higher than 30%; and (3) a surface energy imbalance is also encountered over vegetated soils, medium-grass lands or forest, and the values of $Re/(Rn - G_0)$ ranged from 10% to 20%.

Soil heat flux is an important component of the land surface energy balance (e.g., Kustas et al. 1999, 2000). Gao (2005) proposed a method to calculate soil heat. This method differs from traditional methods because it accounts for both soil thermal conduction and soil thermal convection. Using the data collected during the Global Energy and Water cycle Experiment (GEWEX), Asian Monsoon Experiment (GAME)–Tibet intensive observation period from mid-July to mid-September in 1998, Gao (2005) compared the energy balance flux components and energy budget closure calculated by the traditional methods with those by his methods, and found that the energy budget closure ε (defined as the ratio of the sum of sensible and latent heat fluxes to available energy) was 0.7 and 0.9 for the traditional and his method,

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Table 1 Comparison of the land surface energy imbalance $\Delta(=Re/(Rn-G_0)\times 100)$ or the residuum energy (Re), obtained from different land-cover classes

Experiment and/or reference	Δ or Re	Underlying surface
Dry surfaces		
Albertson et al. (1995)	0	Dry Owens Lake bed
Unland et al. (1996)	0	Semi-arid Sonoran Desert
Heusinkveld et al. 2004	0	Sandy desert
Bare soils or short-grass land surfaces		
TARTEX-90 (Foken et al. 1993)	33	Barley and bare soil
KUREX-91 (Panin et al. 1996)	33	Different surfaces
LINEX-97/1 (Foken 1998)	32	Short grass
LITFASS-98 (Beyrich et al. 2002)	37	Nearly bare soil
GAME/Tibet-98 (Gao 2005)	30	Sparse short grass
Vegetated soils, medium-grass lands or forest		
FIFE-89 (Kanemasu et al. 1992)	10	Steppe, grassland
Vancouver I-90 (Lee and Black 1993)	17	16-m forest
LINEX-96/2 (Foken et al. 1997)	20	Medium grass
FLUXNET (Wilson et al. 2002)	20	Tundra, conifer, agriculture, forest, grassland et al.
Kelliher et al. 1992	Re=50–150 Wm ⁻²	Broad-leaved forest
Oncley et al. 2002	Re=110 Wm ⁻² =16% of Rn	Forest

respectively. It is important to note that Gao's method was originally designed for land surfaces where the soil vertical water movements is significant.

In order to provide guidance to the modeling community on how to interpret flux measurements that do not appear to be consistent with the conservation principles (Kustas et al. 1999), the micrometeorological measurement community has been attempting to resolve these serious discrepancies in the surface energy balance closure (Twine et al. 2000). However, few previous articles have considered the phenomenon that the surface energy balance achieves closure well over dry hot surfaces while energy imbalance is encountered frequently for other types of land surface. The present research builds on the work of Gao (2005), and attempts to explain why energy balance achieves closure well over a dry surface. The second goal of this paper is to better understand the energy balance for different land surfaces by comparing soil water flux to bare ground evaporation rate using the data utilized in Gao (2005).

Materials and methods

Review of Gao et al. (2003) and Gao (2005)

Soil temperature

Gao et al. (2003) incorporated thermal conduction and convection equations to consider these two processes together as follows,

$$\frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2} + W \frac{\partial T}{\partial z}, \quad (1)$$

where k is the soil thermal diffusivity and W is the liquid water flux density. Given the boundary condition: $T|_{z=0} = \bar{T} + A \sin \omega t$, ($t \geq 0$), where $\bar{T}$ is a constant, A is the amplitude of the soil surface temperature, and ω is the angular velocity of the Earth's rotation, they derived the solution to Eq. (2) as follows,

$$T(z, t) = T_0 + A \times \exp \left[\left(-\frac{W}{2k} - \frac{\sqrt{2}}{4k} \sqrt{W^2 + \sqrt{W^4 + 16k^2\omega^2}} \right) z \right] \cdot \sin \left[\omega t - z \frac{\sqrt{2}\omega}{\sqrt{W^2 + \sqrt{W^4 + 16k^2\omega^2}}} \right] \quad (2)$$

Assuming A_1, A_2 are amplitudes, and Φ_1, Φ_2 are initial phases of soil temperatures at the depths of z_1 and z_2 in the soil, respectively, and assuming $z_1 > z_2$ (i.e., $A_1 < A_2$ and $\Phi_1 > \Phi_2$), Gao (2005) derived the followed equations.

$$W = \frac{\omega(z_1 - z_2)}{\Phi_1 - \Phi_2} \left[\frac{-(\Phi_1 - \Phi_2)^2 + \ln^2(A_1/A_2)}{(\Phi_1 - \Phi_2)^2 + \ln^2(A_1/A_2)} \right] \\ = \frac{\omega(z_1 - z_2)}{\Phi_1 - \Phi_2} \left[\frac{2 \ln^2(A_1/A_2)}{(\Phi_1 - \Phi_2)^2 + \ln^2(A_1/A_2)} - 1 \right], \quad (3)$$

and

$$k = - \frac{(z_1 - z_2)^2 \omega \ln(A_1/A_2)}{(\Phi_1 - \Phi_2) [(\Phi_1 - \Phi_2)^2 + \ln^2(A_1/A_2)]}. \quad (4)$$

Land surface energy budget closure

Surface soil heat flux (G_0) is usually estimated by adding the heat storage of the soil layer above the plate to the plate measurements, namely

$$G_0 = G_1 + C_g \Delta z \partial T / \partial t, \quad (5)$$

where G_0 is the surface soil heat flux, G_1 is the soil heat flux at depth (from the plate heat flow measurement in soil), C_g is the volumetric heat capacity of the soil, which can be easily derived from soil components (Van Wijk 1963), Δz is the thickness of a thin layer of the soil and T the mean soil temperature of the thin layer, ∂T is the change in mean soil temperature during the measurement period, and ∂t the change in time. $C_g \Delta z \partial T / \partial t$ represents the heat storage of the soil layer.

Applying Eq. (1) to Eq. (5) gives

$$G_0 = G_1 + C_g k \partial T / \partial z + C_w W \Delta T \quad (6)$$

where C_w is the water volumetric heat capacity. Gao (2005) pointed out that, physically, $C_g k \partial T / \partial z$ contributes to the soil heat flux and depends on thermal diffusivity, and $C_w W \Delta T$ contributes to the soil heat flux through thermal convection within the surface soil layer. Gao (2005) therefore suggested the surface energy balance equation to be

$$R_n = H + LE + G_1 + C_g k \partial T / \partial z + C_w W \Delta T \quad (7)$$

where R_n , H and λE are net radiation, sensible heat flux, and latent heat flux, respectively, all which can be obtained by direct measurements.

New derivation for dry surfaces

Soil temperature

We assume $W=0$ for dry surfaces, such as a dry hot desert, because W decreases with decreasing soil moisture (Gao 2005) and the soil moisture is extremely low there (Heusinkveld et al. 2004). In this way, Eq. (2) becomes

$$T(z, t) = T_0 + A \exp \left(-z \sqrt{\omega / 2k} \right) \sin \left(\omega t - z \sqrt{\omega / 2k} \right), \quad (8)$$

Equation (8) is widely used in textbooks (e.g., Stull 1988; Garratt 1992) for calculation of soil temperature. It means that the amplitude of the soil temperature exponentially decreases and the phase of the soil temperature delays linearly when the soil depth increases. Because Gao (2005) gave an explicit mathematical derivation and a physical explanation to Eq. (2), we believe that Eq. (2) reasonably describes the variation of soil temperature with soil depth. Comparing Eq. (8) with Eq. (2) means that a model derived from Eq. (8) will (1) generate a reasonable result of soil temperature for a dry hot sandy desert; and (2) overestimate

Fig. 1 **a** Inter-comparison of the measured ($H+\lambda E$) against available energy (G_1+G_2) using previous method, and **b** inter-comparison of the measured ($H+\lambda E$) against available energy $R_n-G_1-G_3-G_4$ using present method at the Naqu site in the Tibetan Plateau from DOY 195 to DOY 258, 1998. (After Gao 2005)

the amplitude of soil temperature, and delay the phase of soil temperature for other surfaces where $W \neq 0$, such as bare soil and short-grass lands.

When $W=0$, Eq. (3) means

$$\Phi_1 - \Phi_2 = -\ln(A_1/A_2) = \ln(A_2/A_1), \quad (9)$$

and Eq. (4) therefore becomes

$$k = \frac{(z_1 - z_2)^2 \omega}{2(\Phi_1 - \Phi_2)^2} = \frac{(z_1 - z_2)^2 \omega}{2 \ln^2(A_1/A_2)}. \quad (10)$$

Because $z_1 \neq z_2$, Eq. (10) means that $k > 0$ always. Eqs. (9–10) imply, for a dry hot desert ($W=0$), (1) that the phase delay equals to the logarithm of amplitude ratio of soil temperature collected at two depths; (2) that thermal conduction still occurs whereas thermal convection does not happen when $W=0$; and (3) that soil thermal diffusivity k can be determined using amplitudes or phases of soil

temperature collected at two depths of dry sandy soil. If surface temperature is used, i.e., $z_2=0$, $\Phi_2=0$, and $A_2=A$, Eq. (10) can be simplified to

$$k = \frac{z_1^2 \omega}{2\Phi_1^2} = \frac{z_1^2 \omega}{2 \ln^2(A_1/A)}, \quad (10')$$

which implies that the soil thermal diffusivity k can be determined by using one-level soil temperature when soil surface temperature is derived from the long-wave radiation omitted from the soil surface.

Surface energy balance budget

The same principle applies to Eq. (6). When $W=0$, Eq. (6) becomes

$$G_0 = G_1 + C_g k \partial T / \partial z, \quad (11)$$

and Eq. (7) becomes

$$R_n = H + LE + G_1 + C_g k \partial T / \partial z, \quad (12)$$

Fig. 2 Comparison of latent heat flux (λE) modeled using SiB2 against direct measurements. (After Gao et al. 2004)

which is also widely given in textbooks (e.g., Stull 1988) for calculation of soil heat flux. It means that the heat absorbed by the soil surface vertically transfers into the deep soil by thermal conduction only. So comparing Eq. (11) with Eq. (6) means that a model derived from Eq. (11) will (1) generate a reasonable result of soil heat flux, when it is applied to a dry hot sandy desert ($W=0$); and (2) underestimate soil heat flux, when it is applied to other surfaces where $W \neq 0$, such as bare soil and short-grass surfaces.

Results

Soil temperature for a dry hot desert

Because we have no data collected for dry soils, we have to use the illustration existing in previous literature for our analysis. Heusinkveld et al. (2004) analyzed the surface energy balance for a sandy desert. Because the average annual rainfall was 100 mm and there were sparsely vegetated linear dunes at their site, the soil moisture did not reach zero, with the result that the soil temperatures modeled by Heusinkveld et al. (2004) had somewhat larger amplitudes, and had somewhat delayed phases for all four depths (0.015 m, 0.034 m, 0.05 m and 0.1 m), as shown in Heusinkveld et al. (2004, Fig. 5).

Energy budget closure over land surfaces

Dry surface

We have pointed out that a model neglecting vertical soil water movement will generate a reasonable estimation of energy budget closure for a dry hot desert where W is extremely low, so it is not a surprise that energy balance closure was found to be very good at the site of Heusinkveld et al. (2004). Besides this, Albertson et al. (1995) examined the components of energy budget balance over the dry Owens Lake bed in Owens Valley, California, where the underlying surface is crusted sand with substantial amounts of evaporative salts. It is obvious that Albertson et al. (1995) did not account for the vertical soil water movements. However, their results (i.e., their Fig. 2a,b) show, not surprisingly, that energy budget was almost achieved closure at their site. Furthermore, Unland et al. (1996) examined surface flux measurements and model output at a semi-arid Sonoran Desert site, and their results show that the energy balance closure was found to be good, too, as shown in Unland et al. (1996, Fig. 5).

Bare soil or sparse short-grass lands

Our analysis above implies that if the vertical soil water movement is neglected in the calculation of soil heat flux (G_0),

Fig. 3 Scatter plot of latent heat flux (λE) modeled using SiB2 against direct measurements

a significant energy imbalance will be encountered for bare soils or short-grass lands where the vertical soil water movement is considerable. Table 1 gives values of $Re/(Rn-G_0) \cdot 100\%$ for the bare soil or short-grass land surfaces where the soil vertical water movement was not considered. It is obvious that the energy imbalance is significant and that values of $Re/(Rn-G_0) \cdot 100\%$ are larger than 30%. Note that the energy imbalance reaches 30% for a Tibetan short-grass prairie as shown in Table 1. Gao (2005) re-examined the soil heat flux for that site by considering soil vertical water movement, resulting in $Re/(Rn-G_0) \cdot 100\%$ being only 10%, as shown in Fig. 1.

Gao et al. (2004) simulated surface energy partitioning in the Tibetan prairie using the simple biosphere model 2 (SiB2), and the modeled values for the 2-month period were compared with micrometeorological field measurements made through the GAME-Tibet IOP 1998. The results show that, when underestimating net radiation by 11%, the model overestimated sensible, latent, and soil heat fluxed by 8, 3,

and 13% respectively. Because we are just concerned about soil water continuity, here we only give the modeled results of latent heat flux as shown in Fig. 2. We found that SiB2 significantly underestimated latent heat flux around 1200 hours on DOYs 202, 204, 207, 212, 222, 226, 245, and 246, although it gave reasonable estimates of latent heat flux for all other times. Figure 3 is the scatter plot of latent heat flux (λE) modeled using SiB2 against direct measurements. It shows that on average λE was overestimated by 3% with the correlation coefficient of 0.92.

The latent heat flux (λE) modeled by SiB2 consists of two parts: λE_c and λE_g , where λE_c is from the canopy and λE_g is from the bare soil surface. λ is the latent heat of vaporization, E_c is the evapotranspiration rate from canopy and E_g is the evaporation from the bare soil surface. The temporal variations of λE_c and λE_g are shown in Fig. 4. We found that the magnitude of λE_c is close to that of λE_g although, during the monsoon season, the mean canopy height was just 0.05 m and the leaf area index (hereinafter,

Fig. 4 Temporal variations of latent heat flux from canopy (λE_c) and latent heat flux from bare ground (λE_g) at the Naqu site in the Tibetan Plateau from DOY 195 to DOY 258, 1998

referred to as LAI) were assumed to be 0.5, respectively. Therefore, we deduced that λE_c would be more significant than λE_g for vegetated surfaces where LAI is much larger.

Because of soil water continuity in the vertical direction, in theory the daily average of the bare surface evaporation (E_g) should be equal to the sum of soil water flux (W) and water content change in the soil surface layer, although the phase change of water happens in the soil surface layer. W is calculated from the temperature measurements at two depths in soil (Gao 2005). Comparison of bare ground evaporation (E_g) against soil water flux (W) is shown in Fig. 5. It was found that soil water flux (W) is less and approaches the case of bare surface evaporation (E_g); on average, $E_g=1.58 \times 10^{-5} \text{ m}^3 \text{ s}^{-1} \text{ m}^{-2}$ and $W=1.22 \times 10^{-5} \text{ m}^3 \text{ s}^{-1} \text{ m}^{-2}$. The difference between E_g and W is caused by the following reasons:

- 1.) The difference was mainly caused by water content change in the soil surface layer. Because the whole observation period is within the monsoon season, rain events happened frequently (Gao 2005). The cold rainwater infiltrating into the soil created complexities in soil temperature distribution.
- 2.) Because it was impossible for us to directly measure soil evaporation from the bare ground (E_g), we have to use the modeled E_g for our comparison. Uncertainties of modeling bare surface evaporation (E_g) negatively affect the comparison results.

- 3.) Equation (3) shows that the soil water flux (W) depends significantly on the phase difference and on the amplitude ratio of soil temperatures collected at two depths, so accurate determination of the soil temperature phase and the soil temperature amplitude therefore becomes essential. The error in determination of the soil temperature phase and the soil temperature amplitude also negatively affect the comparison results.

Grass-land surfaces or forest

Over the surfaces which were well vegetated, underlying surface energy absorption happens mainly at the vegetation canopy rather than at the soil surface. Table 1 gives values of $Re/(Rn-G_0) \cdot 100\%$ for vegetated soils, medium-grass lands and forest where the vertical soil water movement was not considered. It is obvious that values of $Re/(Rn-G_0) \cdot 100\%$ ranged from 10% to 20%.

There could be a number of reasons for the energy imbalances listed in the Table 1. Over vegetated surfaces, the canopy evapotranspiration flux contributes a significant part to the total latent heat flux, which means the soil water is withdrawn directly by the root systems. Wilson et al. (2002) gave a comprehensive evaluation of energy balance closure at FLUXNET sites. Their results show a general lack of closure at most sites, with a mean imbalance in the

Fig. 5 Comparison of evaporation from bare ground (E_g) against soil water flux (W) at the Naqu site in the Tibetan Plateau from DOY 195 to DOY 258, 1998

order of 20%, and they concluded that the sum of sensible heat flux and latent heat flux was underestimated and/or that available energy was overestimated. We certainly agree with them. Our above analysis implies that the magnitude of soil water flux (W) is less than and approaches that of bare soil surface evaporation (E_g). The bare soil surface evaporation (E_g) is insignificant over dense-grass surface or forest, so the soil water movement plays a minor role in surface energy balance there.

Discussion and conclusions

Tanaka et al. (2001, 2003) and Ma et al. (2003) investigated surface energy partitioning over the Tibetan Plateau by using the data collected during the GAME–Tibet IOP 1998. They pointed out that the surface energy imbalance was caused by measurement uncertainty of latent heat flux at their sites. Table 1 also supports their conclusion, because Table 1 shows that the energy imbalance is between 10% and 20% over the vegetated soil, medium-grass lands or forest where the latent heat flux is the main consumer of available energy. The present work mainly attempts to explain why the energy imbalance exceeds 30% of available energy for wet sparse short-grass land or wet bare soil, where latent heat flux is an important energy consumer of surface available energy and the magnitude of the evapotranspiration rate from canopy (E_c) is roughly close to that of, and the evaporation from, the bare soil surface (E_g).

Lee (1998) investigated the surface–air exchange over tall vegetation, and concluded that energy imbalance is caused by the energy advection. We think that the energy advection is important for tall vegetation such as forest, but it is not important for the wet sparse short-grass land or wet bare soil.

This paper explains why the energy balance achieves closure for a dry sandy desert while energy imbalance was encountered for other types of land surface, when traditional methods are used for determination of energy balance components. Furthermore, we classify the land surfaces into three groups: (1) dry and bare soil; (2) wet sparse short-grass land or wet bare soil; and (3) well vegetated soil, medium-grass lands or forest. Previous work indicates that energy balance is achieved for dry and bare soil surfaces; energy imbalance exceed 30% of available energy for wet sparse short-grass land or wet bare soil; and the energy imbalance is between 10 and 20%.

The traditional method of calculating soil heat does not account for the influence of soil vertical water movements on soil surface temperature and on soil heat flux. Although it can be used for dry and bare soil, where the vertical water movement is insignificant, it significantly underestimates the soil heat flux for wet sparse short-grass land or bare soil where the vertical water movement is significant.

The direct evaporation from the bare part of the surface is not significant over densely vegetated surfaces, so we conclude that the water movement plays a minor role in the surface energy balance. Work will continue to identify the source of the energy imbalance encountered over densely vegetated surfaces.

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