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The correction of soil heat flux measurements to derive an accurate surface energy balance by the Bowen ratio method

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Abstract

A method is presented for calculating conductive heat flux at the soil surface (G_0) from measured soil heat flux (G) some centimetres beneath the ground surface. The method does not require estimation of thermal properties and is valid for inhomogeneous soils with regard to their thermal properties. Data from the central sub-site of the Eastern Super Site of the HAPEX-Sahel experiment are used to illustrate the method. Finally, the influence of using corrected values of surface soil heat flux G_0, rather than measured values of G, in the energy budget with the Bowen ratio is evaluated. The corrections for G are small in the case of the highly diffusive soil of the Sahel. Errors in estimating latent heat flux with G instead of G_0 are negligible. However, calculations show that these errors could be much more important for other soils with lower soil thermal diffusivity.© 1997 Elsevier Science B.V.

1. Introduction

The Bowen ratio method is commonly used to estimate sensible and latent heat fluxes at the Earth's surface. The method is based on the energy budget equation, for which soil heat flux at the surface has to be known. Common methods for estimating soil heat flux at the surface, G_0, include the null-alignment (Kimball and Jackson, 1975), the harmonic (Horton and Wierenga, 1983) and the finite difference method (Balabanis, 1987; Sharratt et al., 1992), which are all based on soil temperature measurements. The combination method (Fuchs, 1986; Massman, 1992) combines temperature with heat flux measurements. Soil heat flux measurements cannot be made directly at the surface because of exposure of the sensor to radiation; nor can they be made very close to the soil surface, because of the modification induced by the sensor in moisture movement. When heat flux sensors are installed a few centimetres

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beneath the soil surface, the combination method provides a correction yielding estimates of the soil heat flux at the surface. A large number of works concerning sensible and latent heat fluxes at the soil surface computed by the Bowen ratio method do not make this correction. They assume that soil heat flux measured a few centimetres into the soil is a good approximation of the actual value at the surface. Although this might be realistic when temperature gradients at the soil surface are small, such as beneath a dense canopy, it will not be for the case of a sparse canopy.

Temperature-based methods for determining soil heat flux at the surface require knowledge of the volumetric heat capacity as well as of the thermal conductivity of the soil. For the combination method, knowledge of the volumetric heat capacity only is needed. Thermal properties of soils are difficult to measure or to estimate. However, in the case of the HAPEX-Sahel experiment, soil thermal diffusivity has been estimated for the whole period of intensive observations (IOP) at a subsite of the Eastern Central Super Site, by Passerat de Silans et al., (1996). Their work shows vertical inhomogeneity in soil thermal properties in the 0–0.25 m layer, and therefore common methods for surface soil heat flux estimation are not applicable. Inhomogeneity of the thermal properties is caused by variation in bulk dry soil density and soil moisture. In that study, no attempt was made to relate soil thermal diffusivity to soil moisture because water content profiles are unknown in the upper soil layer (0–0.025 m).

The purpose of this paper is to propose an analytical method for estimating the soil heat flux at the surface, G_0, when soil heat flux is measured at a depth z, and to examine the consequences of correcting the soil heat flux on the latent heat fluxes. The proposed method is applicable in inhomogeneous soil, i.e. when thermal properties vary with depth, and does not require the knowledge of thermal properties. Variation of the thermal properties in the upper soil layer is very difficult to estimate because of the link between thermal properties and soil water content. For the proposed method, only temperature measurements at the same depth z where the heat flux plate is placed and at the soil surface are necessary. Data collected during the HAPEX-Sahel IOP at a grass layer in the savanna site are used for the purpose.

2. Theoretical background

2.1. Development of the method

The transient heat conduction equation is

$$C\frac{\partial T}{\partial t}=-\frac{\partial}{\partial z}\left(-\lambda\frac{\partial T}{\partial z}\right) \tag{1}$$

where C is the volumetric heat capacity (J m^{-3} K^{-1}), and λ the thermal conductivity (W m^{-1} K^{-1}). Considering that both these parameters are depth dependent, Eq. (1) gives

$$C\frac{\partial T}{\partial t}=\lambda\frac{\partial^2 T}{\partial z^2}+\lambda'\frac{\partial T}{\partial z} \tag{2}$$

Using the following transformations (Nerpin and Chudnovskii (1984), given by Massman (1993)):

$$T=\left(\frac{C_0\lambda_0}{C\lambda}\right)^{1/4}\tau \text{ and } \xi=\alpha_0^{1/2}\int_0^z\left(\frac{1}{\alpha}\right)^{1/2}\mathrm{d}z \tag{3}$$

this latter being a Kirschoff type transformation, where α is soil thermal diffusivity, Eq. (2) becomes

$$\frac{\partial\tau}{\partial t}=\frac{\lambda_0}{C_0}\frac{\partial^2\tau}{\partial\xi^2}+\omega_{\mathrm{T}}(z)\tau \tag{4}$$

with

$$\omega_{\mathrm{T}}(z)=\frac{\lambda}{16C}\left[\left(\frac{C'}{C}+\frac{\lambda'}{\lambda}\right)^2-4\left(\frac{\lambda'^2}{\lambda^2}+\frac{\lambda'C'}{\lambda C}\right)-4\left(\frac{C''C-C'^2}{C^2}+\frac{\lambda''\lambda-\lambda'^2}{\lambda^2}\right)\right]$$

Here, single and double primes indicate respectively the first and second derivative with respect to depth. The product λC is the so-called thermal admittance. Subscript zero in Eq. (4) indicates properties for the soil surface level.

Assuming $\omega_{\mathrm{T}}(z)=0$, Eq. (4) is transformed in the transient heat conduction equation for constant thermal properties with depth, taking its values at the soil surface:

$$\frac{\partial\tau}{\partial t}=\frac{\lambda_0}{C_0}\frac{\partial^2\tau}{\partial\xi^2} \tag{5}$$

$\tau(\xi,t)$ can be found by any analytical solution of the heat conduction equation taking account of its initial and boundary conditions.

Introducing $G(z,t)=-\lambda\frac{\partial T}{\partial z}$ in Eq. (1) and differentiating with respect to z, it follows that

$$C\frac{\partial G}{\partial t}=\lambda\frac{\partial^2 G}{\partial z^2}-\lambda\frac{C'}{C}\frac{\partial G}{\partial z} \tag{6}$$

Using the above Kirshoff transformation and a new variable Γ, Eq. (6) gives

$$\frac{\partial\Gamma}{\partial t}=\frac{\lambda_0}{C_0}\frac{\partial^2\Gamma}{\partial\xi^2}+\omega_{\mathrm{G}}(z)\Gamma \tag{7}$$

where Γ is defined by

$$G=\left(\frac{C\lambda}{C_0\lambda_0}\right)^{1/4}\Gamma \tag{8}$$

and

$$\omega_{\mathrm{G}}(z)=\frac{\lambda}{16C}\left[\left(\frac{C'}{C}+\frac{\lambda'}{\lambda}\right)^2-4\left(\frac{C'2}{C^2}+\frac{\lambda'C'}{\lambda C}\right)+4\left(\frac{C''C-C'^2}{C^2}+\frac{\lambda''\lambda-\lambda'^2}{\lambda^2}\right)\right] \tag{9}$$

The inverted ratio of thermal admittance in relation to the expression of variable τ

(Eq. (3)) should be noted. Assuming $\omega_G(z)=0$, Eq. (7) is formally identical to Eq. (5). Therefore they both have the same analytical solution provided they have the same initial and boundary conditions.

The formalism to obtain $\tau(0,t)$ from Eq. (5) or $\Gamma(0,t)$ from Eq. (7), can be written

$$\tau[\xi(z),t] \xrightarrow{\mathcal{L}(\xi,t)} \tau(0,t)=T(0,t)$$

and

$$\Gamma[\xi(z),t] \xrightarrow{\mathcal{L}(\xi,t)} \Gamma(0,t)=G(0,t) \quad (10)$$

where $L(\xi,t)$ is an operator indicating the analytical integration of the differential-equation. As the transient heat conduction equation is linear, Eq. (10) becomes

$$T(z,t) \xrightarrow{\mathcal{F}_1(\xi,t)} T(0,t)$$

and

$$G(z,t) \xrightarrow{\mathcal{F}_2(\xi,t)} G(0,t) \quad (11)$$

with $\mathcal{F}_1(\xi,t)=(C\lambda/C_0\lambda_0)^{1/4}\mathcal{L}(\xi,t)$ and $\mathcal{F}_2(\xi,t)=(C_0\lambda_0/C\lambda)^{1/4}\mathcal{L}(\xi,t)$.

So, identifying $\mathcal{F}_1(\xi,t)$ from measurements of both $T(z,t)$ and $T(0,t)$, we can deduce $\mathcal{F}_2(\xi,t)$ by

$$\mathcal{F}_2(\xi,t)=(C_0\lambda_0/C\lambda)^{1/2}\mathcal{F}_1(\xi,t) \quad (12)$$

$G(0,t)$ can then be derived from measured $G(z,t)$, if the admittance ratio $(C_0\lambda_0/C\lambda)$ is known.

In all this theory, the assumption has been made that C and λ are independent of time, so $(C_0\lambda_0/C\lambda)^{1/2}$ (in Eq. (12)) is a 'bulk correction factor'. It may be deduced from information about a bulk variation in temperature during the time of integration; for instance, from the quotient of the mean daily surface temperature to the mean temperature at depth z, if the temperature wave evolution with time is steady periodic.

Massman (1993) compared the exact solution of Eq. (4) with the approximated solution of Nerpin and Chudnovskii (1984) (Eq. (5), in which $\omega_T=0$ is assumed). In his work, he assumed the steady periodicity of transformed temperature τ at various depths. His exact solution requires a two-layer soil: a layer of finite thickness in which thermal properties C and λ are assumed to vary monotonically with depth in an exponential way, overlying a semi-infinite layer in which thermal properties are constant. He showed that, in these conditions, ω_T can be disregarded, i.e. $\omega_T \ll \omega$ where ω is the fundamental frequency of the temperature wave, if concavity of the curve of thermal properties with depth is positive. Concavity of a curve is defined as the second derivative of its mathematical function. For other concavities, the assumption made in Eq. (5) may not be fulfilled. In the expressions of ω_T and ω_G, concavity of thermal properties is implicit in the last term. However, in both these expressions, this last term enters with an opposite sign. Therefore, following Massman (1993), if ω_T can be disregarded, the same cannot be said, all the more so, for ω_G, and vice versa.

2.2. Application of the method

Assuming that temperature and flux waves are steady periodic, transformed temperatures and heat fluxes at the soil surface may be described by a Fourier series

$$\tau(0,t)=T(0,t)=\overline{T}_0+\sum_{i=1}^{n}A_i\sin(i\omega t+\phi_i)$$

and

$$\Gamma(0,t)=G(0,t)=G_0+\sum_{i=1}^{n}B_i\sin(i\omega t+\delta_i) \quad (13)$$

where couple $[A_i,\phi_i]$ or $[B_i,\delta_i]$ means semi-amplitude and phase of the temperature or heat flux wave of harmonic i, respectively.

Analytical solution of the heat conduction equation for a semi-infinite medium with boundary conditions given in Eq. (13), is (Carslow and Jaeger, 1959)

$$\tau=(\xi,t)=T_0+\sum_{i=1}^{n}A_i\exp\left(\frac{-\xi}{D_{0i}}\right)\sin\left(i\omega t+\phi_i-\frac{\xi}{D_{0i}}\right) \quad (14)$$

where D_{0i} is the penetrating depth of the temperature wave for time equal to P/i (P is the period of the main harmonic: $P=24$ h). D_{0i} is a function of the thermal diffusivity at the soil surface (subscript zero). The expressions $\exp(\xi/D_{0i})$ and ξ/D_{0i} are the damping and phase shift difference, respectively. At any depth z, $T(z,t)$ will be obtained by multiplying $\tau(\xi,t)$ from Eq. (14) by $(C_0\lambda_0/C\lambda)^{1/4}$. Therefore, calling T_{mz} the average temperature at depth z, T_0 and T_{mz} will be related by

$$\left(\frac{C_0\lambda_0}{C_z\lambda_z}\right)^{1/2}=\left(\frac{T_{mz}}{T_0}\right)^2 \quad (15)$$

from Eq. (12), providing a simple way to determine the bulk correction factor. Here subscript z means that values of C and λ are at depth z.

Fitting the measured heat flux and the measured temperature at depth z with a Fourier series with n harmonics, we obtain

$$G(z,t)=G_{mz}+\sum_{i=1}^{n}B_{zi}\sin(i\omega t+\delta_{zi}) \quad (16)$$

and

$$T(z,t)=T_{mz}+\sum_{i=1}^{n}A_{iz}\sin(i\omega t+\phi_{zi}) \quad (17)$$

so, using Eq. (8), Eq. (13), Eq. (14), Eq. (16) and Eq. (17), the following expressions for B_i and δ_i can be derived:

$$B_i=\left(\frac{C_0\lambda_0}{C\lambda}\right)^{1/2}\frac{A_i}{A_{zi}}B_{zi} \quad (18)$$

$$\delta_i=\delta_{zi}+(\phi_i-\phi_{zi})$$

Therefore, $G(0,t)$ can be calculated when $G(z,t)$, $T(z,t)$ and $T(0,t)$ are known.

3. Experimental site and measurements

The HAPEX-Sahel project took place in Niger during the 1992 wet season, from mid-August to mid-October (Goutorbe et al., 1994). The region is generally covered by aeolian sand with a semi-arid vegetation. The field experiments were conducted at the Eastern Super Site of the HAPEX square at different sub-sites. One of them was a fallow savanna which consists of *Guiera* sp. bushes with an undergrowth of sparse grasses and herbs, on a flat sandy soil surface. Instrumentation was installed above the soil to evaluate the latent and sensible heat exchanges (Monteny, 1993; Monteny et al., 1994), and into the soil to measure soil heat flux and the vertical temperature profile. Soil heat flux has been measured at 0.025 m beneath the soil surface. It is the mean of four measurements located in a square metre with depth varying from 0.02 to 0.03 m. The depth of each sensor was measured after the experiment when the sensors were removed. Soil temperatures were measured at various depths (0.002, 0.02, 0.09, 0.14, 0.28, 0.51, 1.01 m) with thermocouple probes. Temperature at the surface was also routinely measured by an IR radiometric probe installed on the mast (9 m height) with the above-soil instrumentation. When the soil was bare, radiometric temperatures agreed well with the temperatures measured at 0.002 m, so the latter has been used as the temperature at the surface. Temperature measured at 0.02 m depth was lower than expected when compared with others and has been discarded. Therefore temperature at 0.025 m was generated by a cubic spline algorithm with the temperatures measured at 0.002, 0.09, 0.14, 0.28, 0.51 and 1.01 m as input.

All soil-based instrumentation was installed as close together as possible in a measurement area inside the fallow savanna. Signals from all instrumentation on the plot were sampled every 10 s with a Campbell Scientific (Logan, UT, USA) datalogger, and 20 min averages were computed for final storage.

4. Results and discussion

The method described above was applied from DOY (day of year) 230 to DOY 271. Eighteen harmonics were necessary to derive a correct fit to the diurnal experimental soil temperature and soil heat fluxes. However, small perturbations are not taken into account, which causes calculated values of G_0 to appear sometimes smoother than measured flux values at 0.025 m. In Fig. 1, measured soil heat flux G (0.025 m) is compared with soil heat flux calculated at the soil surface, G_0, for three typical days. The greatest differences in amplitudes and phase shifts are encountered 6 days after the last rain. The highest relative differences are obtained in the rising portion of the diurnal curve. During the afternoon, soil heat flux at the surface decreases faster than soil heat flux measured at 0.025 m depth.

The bulk correction factor is evaluated from Eq. (15). Its variation during the experiment is small, between 1.08 and 0.98. Lower values are obtained for the driest days.

The most important cause of variation in the surface heat flux correction is due to the transformed depth variable ξ. The proposed method does not require the calculation of ξ as given in Eq. (3). However, we evaluate it from daily diffusivity profiles given

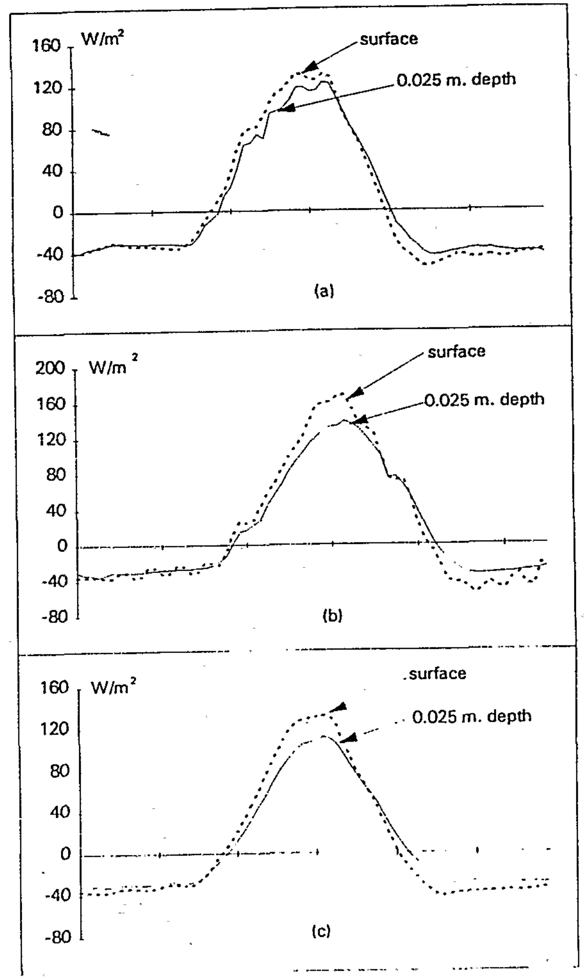

Fig. 1. Daily evolution of measured (0.025 m) and calculated (surface) soil heat fluxes: (a) 2 days after rain; (b) 6 days after rain; (c) 11 days after rain.

by Passerat de Silans et al. (1996). ξ value is strongly related to the drying dynamic of the upper soil layer and to dry bulk density variation with depth. This parameter does not reflect the daily pattern of the drying process, as it is evaluated as a bulk daily parameter owing to the method used for estimating the thermal diffusivity profile. We relate ξ to the number of days after the last rain event (only daily precipitation amounts greater than 1 mm are considered). The curve $\xi(t)$ is an inverted asymmetrical bell-shaped curve with horizontal edges (Fig. 2). Just after rain, ξ is greater than 0.025 m. This means that thermal diffusivity at the soil surface is slightly greater than

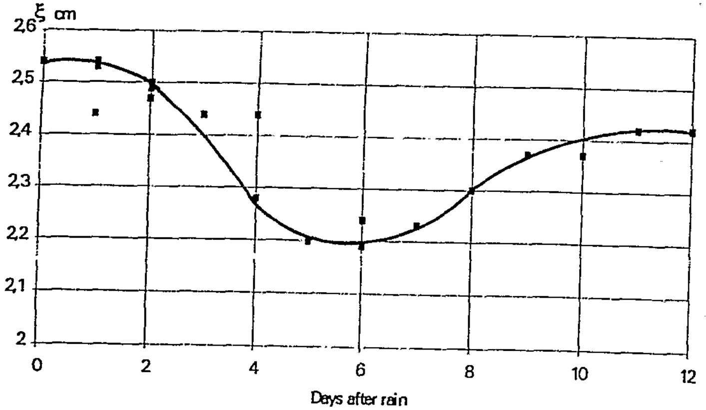

Fig. 2. Transformed depth ξ (cm) as a function of number of days after last rain.

the average value over the 0.025 m layer. For very wet conditions (near saturation), such as just after rain, thermal diffusivity decreases when volumetric soil moisture increases (Fig. 3) and, at the soil surface, moisture is slightly smaller than at 0.025 m depth, as the drying process has just begun. When the upper soil layer is very dry (right edge of the inverted bell-shaped curve), a value of $\xi = 0.0242$ m is found, owing only to the bulk dry soil density variation. ξ reaches its minimum value ($\xi = 0.0219$ m) 6 days after the last rain. The minimum value of ξ corresponds to the highest gradient of thermal diffusivity in the 0–0.025 m soil layer when the moisture gradient is high. Six days after rain, the soil surface is dry for this Sahelian vegetation. Looking at the schematic curve of Fig. 3, which presents thermal diffusivity as a function of soil moisture, one can observe that, in relatively dry conditions such as may occur 6 days after rain, high gradients of thermal diffusivity will be obtained if soil moisture at the surface is at or near the critical value as defined by De Vries (1963) (this is the value for which liquid water is no longer a continuous medium) and soil water content at 0.025 m is near the

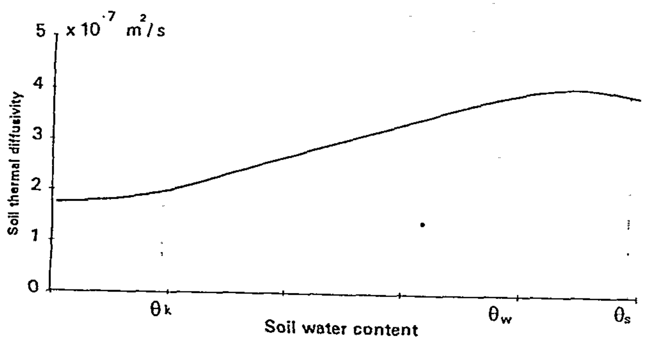

Fig. 3. Soil thermal diffusivity as a function of soil water content (example from Dantas-Antonino (1992)). θ_k, θ_w and θ_S are critical, wilting point and saturation soil water content, respectively.

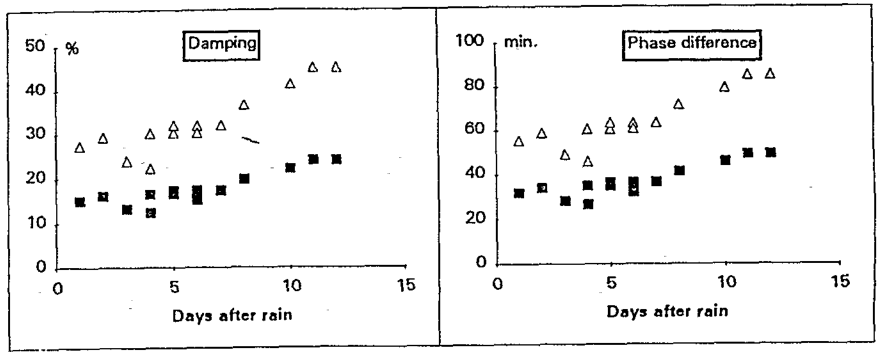

Fig. 4. Damping and phase shift differences for soil heat fluxes at the surface and at 0.025 m depth. ■, Calculated with soil thermal diffusivity profiles given by Passerat de Silans et al. (1996); Δ, calculated with soil thermal diffusivity equal to one-third of previous values.

wilting point. After this day, desorption at the surface is much slower than at 0.025 m depth, so the thermal diffusivity gradient decreases and ξ increases.

The soil in this experiment is a dense sandy soil, and thermal diffusivity is high. That is the reason why differences between estimated flux at the surface and measured flux 0.025 m beneath the soil surface are small. However, for some days between 06:00 and 09:00 h GMT, when temperature is rising and G is small, differences between G and G_0 reach 100%, because of the concomitant damping and difference of phase shift. From the apparent thermal diffusivity curves given by Passerat de Silans et al. (1996) and Fig. 2, we estimate the difference of phase shift and the damping corresponding to the main harmonic (see Eq. (14)). For comparison, we also calculate the values that would be obtained if thermal diffusivity was one-third of its actual value, corresponding to a less dense soil. The damping and difference of phase shift increase when dryness increases, and corrections would be higher for a less diffusive soil (Fig. 4). These values were calculated independently of the method used to correct the soil heat flux. Only the main harmonic is considered. The values encountered for the differences of phase shift and dampings are of the same order as the values that can be observed in Fig. 1.

Following the work of Massman (1993) on determining temperatures at the soil surface, the results presented in Fig. 1 will be quasi-exact if $\omega_T(z)$ and $\omega_G(z)$ are much less than ω, the fundamental frequency of the temperature wave or the flux wave.

Expressions for $\omega_T(z)$ and $\omega_G(z)$ can be respectively transformed in

$$\omega_T(z) = \frac{\lambda}{16C}\left(u'^2 - 4\frac{\lambda'}{\lambda} - 4u''\right) \tag{19}$$

and

$$\omega_G(z) = \frac{\lambda}{16C}\left(u'^2 - 4\frac{C'}{C} + 4u''\right) \tag{20}$$

Table 1
Values of C'/C and λ'/λ relative to u'

C'/C	$u'/4$	$u'/3$	$u'/2$	$2u'/3$
λ'/λ	$3u'/4$	$2u'/3$	$u'/2$	$u'/3$

in which $u = \ln(\lambda C)$ and a prime indicates the derivation with respect to depth. Using Eq. (15) and differentiating with respect to depth, we can calculate an expression for u':

$$u' = -4\frac{T_{mz}'}{T_{mz}} \tag{21}$$

Then values of $u'(z)$ and $u''(z)$ can be calculated from the profiles of mean daily soil temperature. We used a cubic spline algorithm to do this and calculate a family of values for $\omega_T(z)$ and $\omega_G(z)$ with the values of λ'/λ and C'/C indicated in Table 1 (remembering that the actual λ and C profile are unknown). Calculations are made for DOY 265; this is 6 days after the last rain, when differences in thermal diffusivity between the surface and 0.025 m depth are greatest.

Results show that the value of the term accounting for concavity (u'') is the more important, as outlined by Massman (1993). However, the higher value obtained at DOY 265 for $|\omega_T|$ or $|\omega_G|$ in the 0–0.2 m soil layer is $3.2 \times 10^{-10}\ s^{-1}$, whereas the fundamental frequency ω is $7.3 \times 10^{-5}\ s^{-1}$. Therefore the hypothesis of the theory presented in this paper, i.e. ω_T and ω_G equal to zero, is fulfilled.

Computation of latent heat flux by the energy balance Bowen ratio method is done using the expression

$$LE = \frac{R_n - G}{1+\beta} \tag{22}$$

where β is the measured Bowen ratio and R_n is the measured net radiation. We use Eq. (22) with both heat fluxes G (measured at 0.025 m) and G_0 (estimated at the soil surface). In Fig. 5, curves corresponding to both cases are drawn considering the same typical days as in Fig. 1. Differences during the day are small, occurring mainly before midday and at night whatever the day considered. However, as expected from Fig. 1, they are greater in (b), 6 days after rain, when the thermal diffusivity gradient in the 0–0.025 m soil layer is the most important. In Fig. 6, we compare the daily rate of evapotranspiration calculated with both soil heat fluxes (measured at 0.025 m and calculated at the soil surface). Daily values are unaffected by the soil heat flux corrections. This result is not surprising, as the daily mean of fluxes at the soil surface and at 0.025 m differ only by the bulk correction factor, which is always close to unity in this experiment.

No attempt has been made in this study to incorporate horizontal variability of soil thermal properties. However, differences up to 30% of the average value were observed between the maximum daily values of the four measured soil heat fluxes. Such differences are due to differences in the depth at which the sensor is located, to differences in exposure to the Sun at the surface because of the horizontal variability of vegetation cover, to differences between soil moisture at the four places and also to the horizontal variability of the thermal soil properties.

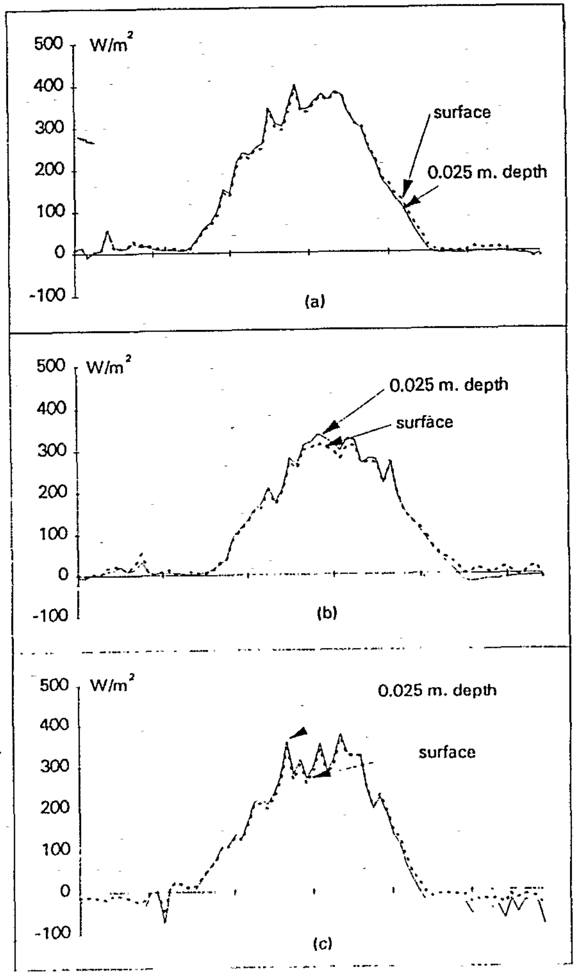

Fig. 5. Latent heat flux calculated with G measured at 0.025 m depth and with G_0 estimated at the soil surface: (a) 2 days after rain; (b) 6 days after rain; (c) 11 days after rain.

5. Conclusions

A method is presented to estimate soil heat flux at the soil surface when soil heat flux is measured some centimetres beneath the surface. The method requires measured temperatures at two levels: at or very close to the soil surface and at the depth at which the heat flux sensor is placed. The proposed method is valid when thermal properties vary with depth and does not require determination or knowledge of these thermal properties. It assumes that both terms ω_T and ω_G can be disregarded. The size

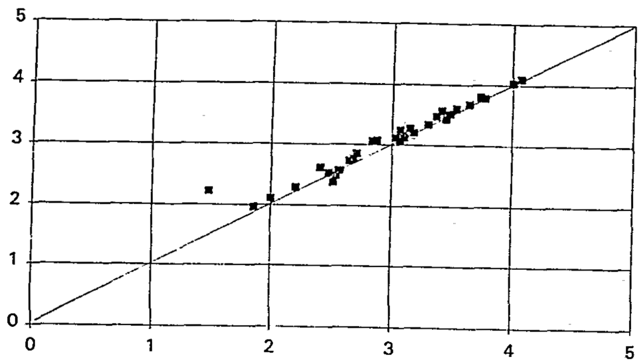

Fig. 6. Daily evapotranspiration (mm). On the abcissa, calculated with G measured at 0.025 m; on the ordinate, with G_0 estimated at soil surface. (The straight line is the 1:1 line).

of both these terms is essentially related to concavity of the curve of thermal properties with depth.

The proposed method has been applied to some data collected in the HAPEX-Sahel experiment. Both terms ω_T and ω_G are significantly smaller than the fundamental frequency ω of the temperature wave, so the above assumption is fulfilled. As the thermal diffusivity of the soil is high (varying between 0.6×10^{-7} and 2.2×10^{-7} m^2 s^{-1}), correction of the measured soil heat flux is small (at the top of the soil heat flux wave, the maximum correction observed is 25% of the measured value). The correction is more important when temperature rises than when it falls. Correction would be higher (probably by a factor of two) if thermal diffusivity was one-third of its actual value.

Corrections of soil heat flux are introduced for the determination of latent heat flux by energy balance coupled with Bowen ratio. The comparison between corrected and uncorrected latent heat fluxes shows small differences whatever the day considered. Probably greater differences would be observed for a soil with smaller thermal diffusivity.

In this paper, special attention has been devoted to the transformed depth ξ. The link between the ξ value and the moisture profile in the 0–0.025 m soil layer is explained and the daily evolution of the transformed depth reflects the daily evolution of the drying process. More work should be devoted to the horizontal variability of the soil heat fluxes, to improve the corrections for G_0 for application to the Bowen ratio method.

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