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

Alain Passerat de Silans[a,*], Bruno A. Monteny[b], Jean Paul Lhomme[b]

[a]Centro de Tecnologia—UFPb, 58050 João Pessoa, Brazil
[b]Centre ORSTOM de Montpellier, B.P. 5045, Montpellier, France

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

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

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

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

with

$$\omega_T(z)=\frac{\lambda}{16C}\left[\left(\frac{C'}{C}+\frac{\lambda'}{\lambda}\right)^2-4\left(\frac{\lambda'^2}{\lambda^2}+\frac{\lambda'C'}{\lambda C}\right)-4\left(\frac{C''C-C'^2}{C^2}+\frac{\lambda''\lambda-\lambda'^2}{\lambda^2}\right)\right]$$

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

(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

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

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

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

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

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

$G(0,t)$ 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

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

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

and

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

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)

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

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

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

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.

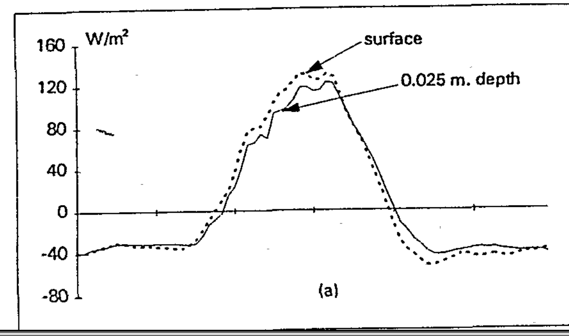
Fitting the measured heat flux and the measured temperature at depth z with a Fourier series with n harmonics, we obtain

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

and

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



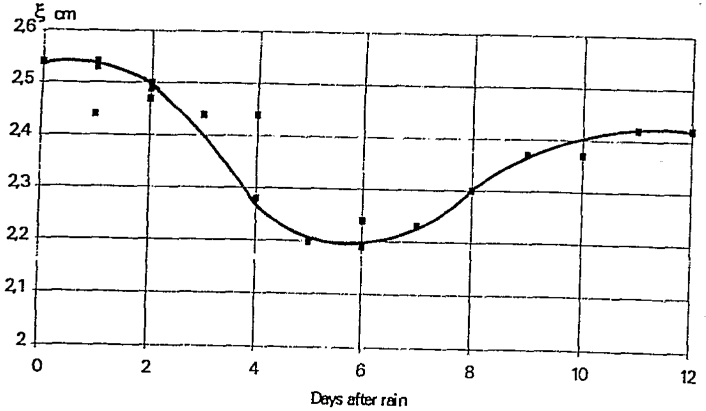


Fig. 2. Transformed depth ξ (cm) as a function of number of days after last rain.

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

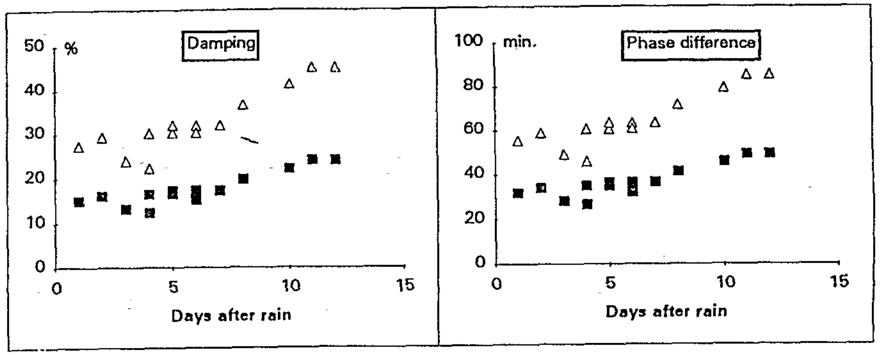


Fig. 4. Damping and phase shift differences for soil heat fluxes at the surface and at 0.025 m depth. ■, Calculated with soil thermal diffusivity profiles given by Passerat de Silans et al. (1996); Δ, calculated with soil thermal diffusivity equal to one-third of previous values.

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,

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

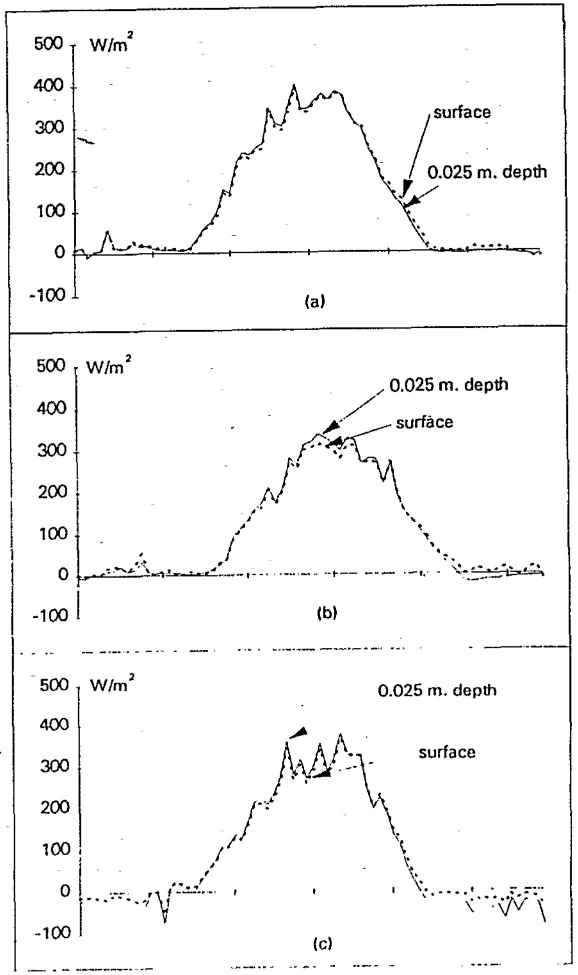


Fig. 5. Latent heat flux calculated with G measured at 0.025 m depth and with G_0 estimated at the soil surface:

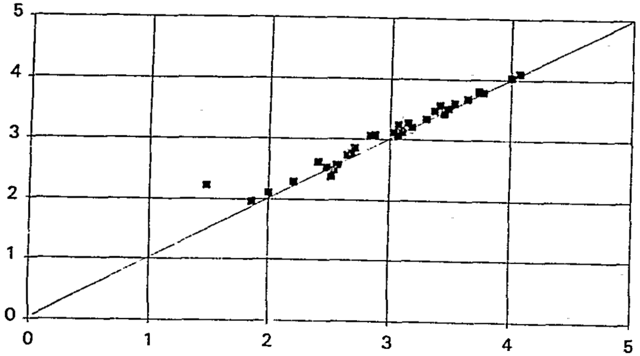


Fig. 6. Daily evapotranspiration (mm). On the abcissa, calculated with G measured at 0.025 m; on the ordinate, with G_0 estimated at soil surface. (The straight line is the 1:1 line).

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

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