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ESTIMATION OF THE SOIL HEAT FLUX FOR DETERMINATION OF EVAPOTRANSPIRATION IN ECOSYSTEMS

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ABSTRACT

This paper presents a method for the estimation of the soil heat flux that was used to determine the evapotranspiration in ecosystems. The great advantage of this approach is that there is the possibility to use this method for both daytime and nighttime as well as that the parameters are adaptively recalculated depending on three soil temperatures. It enables the estimation in situ of the soil thermal parameters and of the soil heat flux for various types of soil in various types of ecosystems without using special types of probes. This method allows a better estimation of evapotranspiration, which is of essential use in agriculture, forestry, botany, soil science, geography, ecology and geomorphology.

Keywords: evapotransipration, heat flux, soil

1. INTRODUCTION

Many hydrologic models and agricultural management applications require evapotranspiration estimates. Evapotranspiration (ET) is the term used to describe the combined process of water loss from the soil surface by evaporation and from the crops by transpiration [1, 3, 6, 9, 11]. The intensity of evapotranspiration is mainly determined using mathematical models rather than by direct measurement with lysimeters or the Eddy Covariance Technique [1, 11]. The reasons for this are high costs, difficulties and inaccuracies associated with the use of the direct measurement. There are several mathematical models available for estimating evapotranspiration from measured climatic data [1, 2, 3, 6, 9, 11]. Evaporation of water requires relatively large amounts of energy. The energy coming into the evaporation surface must equal the energy leaving the surface during the same time period. Therefore

$$Rn = \lambda \cdot ET + H + G + A_f + A \tag{1}$$

where Rn is the intensity of the net radiation [$W{\cdot}m^{-2}$] (i.e. the difference between incoming and outgoing radiation of both short and long wavelengths); $\lambda \cdot ET$ is the latent heat flux consumed during evapotranspiration [$W{\cdot}m^{-2}$]; H is the intensity of the sensible heat flux [$W{\cdot}m^{-2}$]; G is the intensity of the soil heat flux [$W{\cdot}m^{-2}$]; λ is the latent heat of vaporization [$J{\cdot}kg^{-1}$]; ET is the intensity of evapotranspiration [$kg{\cdot}m^{-2}{\cdot}s^{-1}$]; A_f is the intensity of the heat flux consumed during photosynthesis [$W{\cdot}m^{-2}$] and A_c is the intensity of the biomass thermal capacitance change [$W{\cdot}m^{-2}$]. Accordingly [9]

$$A_f \approx 2\%\, Rn \tag{2}$$

and

$$A_c < A_f\,, \tag{3}$$

therefore A_f and A_c are much less than the other factors in (1) and thus, they are negligible. This is in accordance with [1, 2, 3, 9, 11]

$$Rn \approx \lambda \cdot ET + H + G\,, \tag{4}$$

where only the vertical fluxes are considered and the horizontal fluxes are ignored. Model (4) requires methods to partition the intensity of the net radiation Rn into the latent heat flux $\lambda \cdot ET$, the intensity of the sensible heat flux H and the intensity of the soil heat flux G. The soil heat flux G is typically smaller than H or $\lambda \cdot ET$ and for daylight G is commonly approximated according to [1]

$$G = 0.4 \cdot e^{-0.5 \cdot LAI} \cdot Rn = \delta \cdot Rn, \qquad (5)$$

where *LAI* is the leaf area index and

$$\delta \approx 0.4 \cdot e^{-0.5 \cdot LAI}. \qquad (6)$$

In some papers authors consider G as a residual term of energy balance or assume it to be negligible on daily timescales [10]. Many empirical studies [1, 7, 10] shown that G is not constant and that G/Rn can range from 0.05 to 0.50 which depends, except *LAI,* on the time of day, the soil moisture and the thermal properties too. Sensitivity analyses show that if these changes are ignored then it can result in significant errors in modeled flux terms. Because Rn-G is a measure of the energy available for $\lambda \cdot ET$ and H, these differences cannot be ignored. Using a constant ratio for G/Rn will lead to overestimation of sensible and latent heating in the early part of the day and vice versa in the afternoon. For time periods of one hour or less it is necessary for the intensity of the soil heat G to be estimated in a more sophisticated way. Therefore the soil heat flux G was estimated using the two soil temperatures measured at the depth 0.01m and 0.02m. In contrast with [7], it was not assumed the exponential soil temperature profile.

2. ESTIMATING SOIL HEAT FLUX

At a depth z below the soil surface, the downward flux of heat in the soil is given by Fourier's law

$$G(z,\tau) = -\lambda_s \frac{\partial \vartheta(z,\tau)}{\partial \tau}, \qquad (7)$$

where λ_s is the soil thermal conductivity, $\vartheta(z,\tau)$ is the soil temperature at depth z at time τ. In the surface soil layer at a shallow depth z_1, the difference between the heat flux $G = G(0,\tau)$ entering the layer at time τ and at level $z = 0$ and leaving at $z = z_1$ is $G(0,\tau) - G(z_1,\tau)$, see Figure 1. The law of energy conversation holds for the surface soil layer after discretization

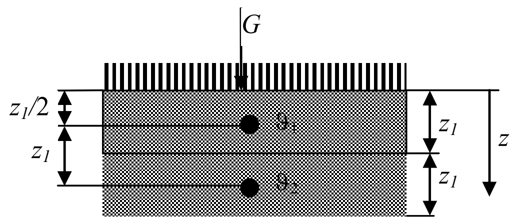

Figure 1 Surface soil layer

$$G(0,\tau) \Delta\tau - \lambda_s \frac{\left(\vartheta_1(\tau) - \vartheta_2(\tau)\right)}{z_1} \Delta\tau = \rho_s \cdot z_1 \cdot c_s \cdot \left(\vartheta_1(\tau + \Delta\tau) - \vartheta_1(\tau)\right), \qquad (8)$$

where ρ_s is the soil density, ($\vartheta_1(\tau) = \vartheta(z_1/2,\tau)$, $\vartheta_2(\tau) = \vartheta(z_1 + z_1/2,\tau)$), $\Delta\tau$ is the time interval, c_s is the soil specific heat. From (8) it follows

$$G = \rho_s \cdot z_1 \cdot c_s \cdot \left(\vartheta_1(\tau + \Delta\tau) - \vartheta_1(\tau)\right) + \lambda_s \frac{\left(\vartheta_1(\tau) - \vartheta_2(\tau)\right)}{z_1}, \qquad (9)$$

where the uniform soil thermal properties in the surface soil layer are assumed.

3. SOIL THERMAL PROPERTIES

A knowledge of the volume fractions of mineral components x_m, organic components x_o, water x_w, and air x_a, allows the determination of the volumetric heat capacity C_s, as follows

$$C_s = \rho_s \cdot c_s = \rho_m \cdot c_m \cdot x_m + \rho_o \cdot c_o \cdot x_o + \rho_w \cdot c_w \cdot x_w + \rho_a \cdot c_a \cdot x_a, \qquad (10)$$

where ρ is the density and c is the specific heat of the mineral, organic, water and air components of soil and which are distinguished by the subscripts *m, o, w* and *a* (from [3] $\rho_m = 2650\ \mathrm{kg \cdot m^{-3}}$, $\rho_o = 1300\ \mathrm{kg \cdot m^{-3}}$, $\rho_a = 1.2\ \mathrm{kg \cdot m^{-3}}$, $\rho_w = 1000\ \mathrm{kg \cdot m^{-3}}$, $c_m = 733\ \mathrm{J \cdot kg^{-1} \cdot K^{-1}}$, $c_o = 1296\ \mathrm{J \cdot kg^{-1} \cdot K^{-1}}$, $c_w = 4182\ \mathrm{J \cdot kg^{-1} \cdot K^{-1}}$, $c_a = 1010\ \mathrm{J \cdot kg^{-1} \cdot K^{-1}}$). It holds

$$x_m + x_o + x_w + x_a = 1 \tag{11}$$

and because $\rho_a \ll \rho_m$ and $\rho_a \ll \rho_w$, the fraction of air can be neglected in calculation (10).

The soil thermal conductivity λ_s can be a difficult parameter to estimate, since it depends not only on the volumetric water content, but also on mineral composition, porosity and dry density [8]. The thermal conductivity λ_s was calculated according to [4, 5, 7, 8] as a combination of dry λ_{dry} and saturated λ_{sat} thermal conductivities, weighted by the Kersten number

$$\lambda_s = K_e\left(\lambda_{sat} - \lambda_{dry}\right) + \lambda_{dry}, \tag{12}$$

where for unfrozen soils [5, 7, 8]

$$K_e = \begin{cases} 0.7\log S_w + 1.0 \text{ for } S_w > 0.05 \text{ and coarse soil} \\ \log S_w + 1.0 \text{ for } S_w > 0.1 \text{ and fine soil} \end{cases}, \tag{13}$$

$$\lambda_{dry} = \frac{0.135 \cdot \rho_{dry} + 64.7}{2700 - 0.947 \cdot \rho_{dry}}, \tag{14}$$

$$\lambda_{sat} = \lambda_{sol}^{1-x_{por}} \cdot \lambda_w^{x_{por}}, \tag{15}$$

S_w is the water saturation, x_{por} is the porosity, λ_{sol} is the thermal conductivity of solids, λ_w is the thermal conductivity of water, ρ_{dry} is the dry density.

4. APLICATION

This method was used for determination of evapotranspiration in ecosystems in the south Bohemia. For this purpose 12 meteorological stations were deployed there. The following variables were measured for the determination of the soil heat flux G: the volumetric soil moisture x_w, the soil temperatures ϑ_1 and ϑ_2 at the depth 0.01 m and 0.02 m, the incoming shortwave radiation $Rs\downarrow$, the outgoing shortwave radiation $Rs\uparrow$, the incoming longwave radiation $Rl\downarrow$ from the atmosphere, the outgoing terrestrial radiation $Rl\uparrow$. For verifying received results the soil heat flux using Huxeflux Heat Flux Plate HFP01 was used. Data was recorded at 10 minute intervals. The soil thermal properties were obtained by the soil analysis.

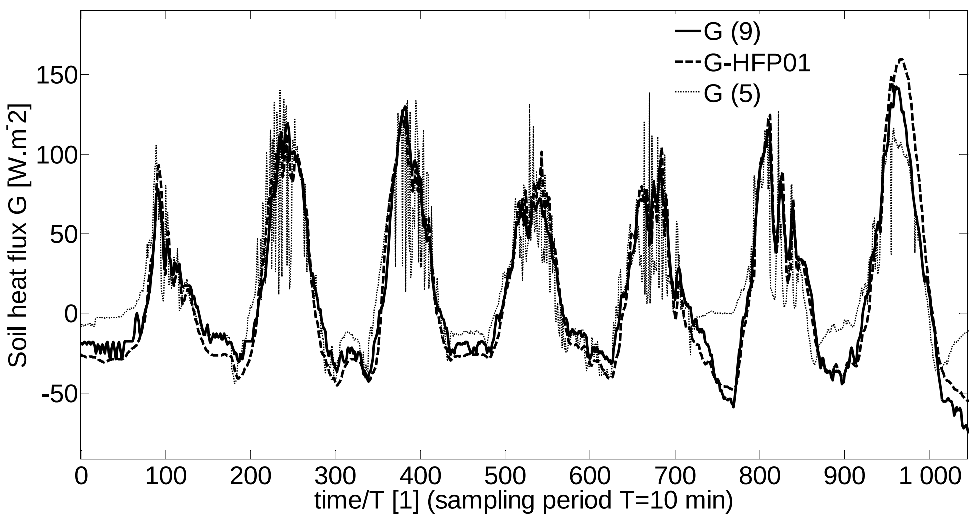

Figure 2. Soil heat flux G calculated according formulas (5), (9) and measured using HFP01

The soil heat fluxes calculated by formulas (5) and (9) with the verifying measurement using Huxeflux HFP01 are depicted in Figure 2. The environmental data was collected in June 2010 at the meteorological station called "Vrt Domanin" near the town of Třeboň in the southern part of the Czech Republic. In formula (5) the parameter δ was calculated according to formula (6), where LAI is the leaf area index. The parameter δ was equal to 0.18. The intensity of the net radiation *Rn* was obtained from the relationship

$$Rn = Rs \downarrow - Rs \uparrow + Rl \downarrow - Rl \uparrow . \qquad (16)$$

The results show the consistency between the values of the soil heat flux measured by HFP01 and those calculated according to formula (9). The fluctuation of the soil heat flux calculated by (5) is caused by clauds.

5. CONCLUSION

This paper introduces a simple method for the estimation of soil heat flux. The method requires measurement of only two soil temperatures near the surface and the volumetric water content. The other parameters can be found using one-off soil analysis. The method enables one to estimate the soil heat flux continuously during day and night conditions. This approach does not assume a monotonous temperature profile in soil which is usually assumed. The influence of clouds to the estimate is partly filtered out by the soil capacity. The method requires accurate temperature measuring. This approach is more accurate than the method based on model (5) and does not require a knowledge of the leaf area index (*LAI*) or the use of special probes. The more accurate estimate of the soil heat flux with regard to (4) results in a more accurate estimate of the intensity of evapotranspiration *ET*.

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