Assessment of Evapotranspiration in Ecosystems

M. Hofreiter and P. Trnka

***Abstract*—**This paper deals with the online and offline assessment of evapotranspiration using mathematical models instead of direct measurement. The evapotranspiration was estimated using the Penman-Monteith Method and the Bowen Ratio Method on data obtained from 14 ground meteorological stations deployed around the landscape of South Bohemia in the Třeboň region. These data were recorded in 10-minute intervals and transferred via the GPRSS network to a server where they were accessed through Internet. Several times a year, these data were complemented by infrared images captured by aerial photography. The thermographic screening was done with an airship and by an aircraft for different altitudes. These infrared images helped to improve the evapotranspiration estimate in the vicinity of the meteorological stations. The data processing and modeling of the evapotranspiration of the selected ecosystem was done using the Matlab programming environment and its toolboxes. This paper also describes the method for the assessment of the soil heat flux which can be used for the online evapotranspiration estimation.

***Keywords*—**Climatic data, evapotranspiration, infrared image, mathematical model.

I. Introduction

Evapotranspiration (ET) is the term used to describe the combined process of water loss from the soil surface by evaporation and from crops by transpiration. More than half of the water that enters the soil returns to the atmosphere through evapotranspiration.

Evapotranspiration rate and amount are the basic information needed for hydrologic models and agricultural management applications. This data is also essential for water quality management and other environmental concerns.

II. Evapotranspiration Estimate

The intensity of evapotranspiration is mainly determined using mathematical models [2], [5], [18] rather than by direct measurement with lysimeters (weighing or compensational) [4],[10] and [19] or by using the Eddy Covariance Technique [9]. The main reasons for this is that there are costs, difficulties and inaccuracies associated with the use of the direct measurement. There are several mathematical models available to determine the evapotranspiration estimate. Most of these models were developed for estimating evapotranspiration from measured climatic data. In our case we used two methods for ET estimation: the Penman-Monteith Method [1],[11],[13] and the Bowen Ratio Method (BR Method) [3], [12]. Both of these methods are based on the fact that the evaporation of water requires relatively large amounts of energy. The energy entering the evaporation surface must equal the energy leaving the surface during the same time period. Therefore

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

where Rn is the intensity of the net radiation [W·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·m^{-2}]; H is the intensity of the sensible heat flux [W·m^{-2}]; G is the intensity of the soil heat flux [W·m^{-2}]; λ is the latent heat of vaporization [J·kg^{-1}]; ET is the intensity of evapotranspiration [kg·m^{-2}·s^{-1}]; A_f is the intensity of the heat flux consumed during photosynthesis [W·m^{-2}] and A_c is the intensity of the biomass thermal capacitance change [W·m^{-2}]. According to [12]

$$A_f \doteq 2\%\ Rn \tag{2}$$

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]

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

where only the vertical fluxes are considered and the horizontal fluxes are ignored.

Evapotranspiration is much more intensive during daylight hours. Therefore, the main consideration is limited to daylight conditions. It holds [16] that

$$Rn = Rs + Rl\,, \tag{5}$$

where R_s is the intensity of the net shortwave (solar) radiation and R_l is the intensity of the net longwave radiation between the Earth and the atmosphere. The boundary between the shortwave and longwave radiation has a wavelength of 3 μm. The intensity of the net shortwave radiation R_s is the difference between the solar radiation $Rs_{\downarrow}$ [W·m^{-2}] reaching the Earth's surface and the shortwave radiation reflected away from the Earth's surface as $Rs_{\uparrow}$ [W·m^{-2}] and thus

M. Hofreiter is with the Institute of Instrumentation and Control Technology, Faculty of Mechanical Engineering, Czech Technical University in Prague, Czech Republic (phone: +420 224352566; fax: +420 233335604; e-mail: Milan.Hofreiter@fs.cvut.cz).

P. Trnka. is with the Institute of Instrumentation and Control Technology, Faculty of Mechanical Engineering, Czech Technical University in Prague, Czech Republic (e-mail: Pavel.Trnka@fs.cvut.cz).

$$Rs = Rs_{\downarrow} - Rs_{\uparrow} . \quad (6)$$

Similarly, the intensity of the net longwave radiation Rl is the difference between the longwave radiation $Rl_{\downarrow}$ [W·m^{-2}] coming from the atmosphere to the Earth's surface and the longwave radiation $Rl_{\uparrow}$ [W·m-2] emitted by the Earth.

$$Rl = Rl_{\downarrow} - Rl_{\uparrow} . \quad (7)$$

From (4), it is obvious that the intensity of evapotranspiration ET can be determined by the relationship

$$ET = \frac{1}{\lambda} \cdot \left(Rn - G - H \right) \quad (8)$$

if the magnitudes Rn, G and H are known. The latent heat of vaporization λ [J·kg^{-1}] at the air temperature t [°C] can be expressed according to [1]

$$\lambda = 2501 \cdot 10^3 - 2361 \cdot t \quad (9)$$

The intensity of evapotranspiration ET was not directly determined by formula (8), but it is based on the Bowen ratio β [3], [14] defined by

$$\beta \triangleq \frac{H}{\lambda \cdot ET} . \quad (10)$$

In accordance with the Bowen method [3], the Bowen ratio β was calculated

$$\beta = \gamma \frac{t_h - t_l}{e_h - e_l} , \quad (11)$$

where t_h is the temperature at a height of 2m [K], t_l is the temperature at a height of 0.3 m [K], e_h is the water vapour pressure at a height of 2 m [kPa], e_l is the water vapour pressure at a height of 0.3 m [kPa], γ is the psychrometric constant [kg·m^{-1}·s^{-2}·K^{-1}].

The intensity of evapotranspiration ET was calculated using formula (12) resulting from (8) and (10).

$$ET = \frac{Rn - G}{\lambda \left(1 + \beta \right)} . \quad (12)$$

The variables Rn, G, λ and β were either directly measured or were calculated using observed data.

III. Climatic Data Measuring

The climatic data were measured by meteorological stations; see Fig. 1, which were deployed in the southern part of the Czech Republic. The following quantities were monitored and recorded at 10-minute intervals: Precipitation, soil humidity, air temperature and humidity at 30 cm and 2 m, incidental and reflected global solar radiation in the short-wave region (0.3 – 2.8 μm), incidental and emitted radiation in the IR region (4.5 - 45 μm), wind speed and direction at 2 m, ten values for soil temperature measured every 10 mm of soil depth and then at depths of 200 mm and at 250 mm.

The intensity of the net radiation Rn was determined using the relationships (5), (6) and (7); where $Rs_{\downarrow}$ and $Rs_{\uparrow}$ were measured by pyranometers and where $Rl_{\downarrow}$ and $Rl_{\uparrow}$ were measured by pyrgeometers, see Fig. 2. The atmospheric pressure p [Pa] was measured centrally for the whole area.

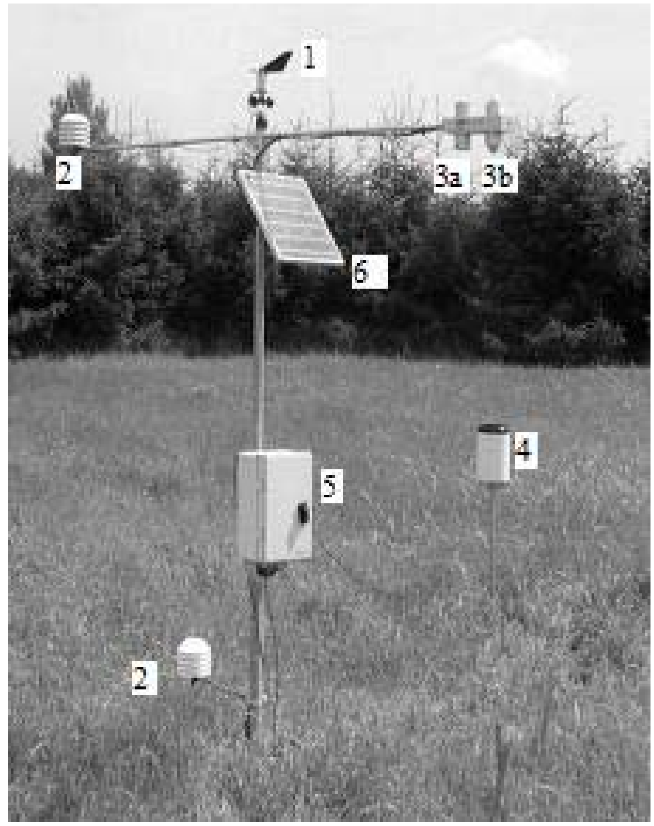

Fig.1 The meteorological station for measuring direction and speed wind (1), air temperature and humidity (2), incident and reflected solar radiation-pyranometer (3a), incident and emitted IR radiation-pyrgeometer (3b), precipitation (4). The other sensors are underground. The meteorological station is equipped with the M4016 control unit (5) and the solar panel (6).

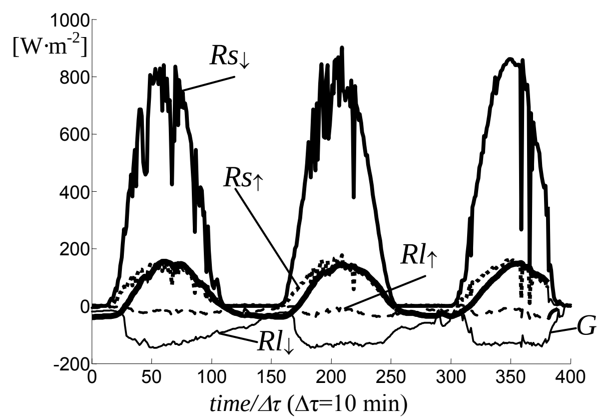

Fig. 2 The energy fluxes $Rs_{\downarrow}$, $Rs_{\uparrow}$, $Rl_{\downarrow}$ $Rl_{\uparrow}$,G

Direct measurement of the intensity of the soil heat flux G at the vegetation-soil boundary is extremely difficult because it depends on many factors, e.g. vegetation, vegetation period, climatic conditions, the thermal properties of the soil, the locality of the heat flux sensor. The soil heat flux G is typically smaller than H or $\lambda \cdot ET$ and for daylight G is commonly approximated according to [1]

$$G = 0.4 \cdot e^{-0.5 \cdot LAI} \cdot Rn = \delta \cdot Rn\,, \qquad (13)$$

where *LAI* is the leaf area index and

$$\delta \triangleq 0.4 \cdot e^{-0.5 \cdot LAI}\,. \qquad (14)$$

In some papers authors consider G as a residual term of energy balance or assume it to be negligible on daily timescales [17]. Many empirical studies e.g. [1], [9] and [17] 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 as well. Sensitivity analyses show that if these changes are ignored then significant errors in modeled flux terms can result. 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.

IV. Estimating Soil Heat Flux

The soil heat flux G was estimated using the two soil temperatures measured at the depth 0.01m and 0.02m. In contrast with [9], it was not assumed the exponential soil temperature profile.

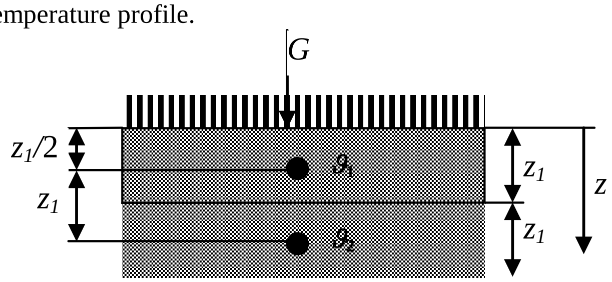

Fig. 3 Surface soil layer

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

$$G(z,\tau) = -\lambda_s \frac{\partial \vartheta(z,\tau)}{\partial \tau}\,, \qquad (15)$$

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 Fig 3. The law of energy conversation holds for the surface soil layer after discretization

$$\begin{aligned} G(0,\tau)\cdot\Delta\tau - \lambda_s \frac{(\vartheta_1(\tau)-\vartheta_2(\tau))}{z_1}\cdot\Delta\tau = \\ = \rho_s \cdot z_1 \cdot c_s \cdot (\vartheta_1(\tau+\Delta\tau)-\vartheta_1(\tau)) \end{aligned}\,, \qquad (16)$$

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 (16) it follows

$$G = \rho_s \cdot z_1 \cdot c_s \cdot (\vartheta_1(\tau+\Delta\tau)-\vartheta_1(\tau)) + \lambda_s \frac{(\vartheta_1(\tau)-\vartheta_2(\tau))}{z_1}\,, \quad (17)$$

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

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 soil thermal properties were estimated by the following method.

The volumetric heat capacity of soil C_s can be established

$$C_s = \rho_s \cdot c_s = \rho_m c_m x_m + \rho_o c_o x_o + \rho_w c_w x_w + \rho_a c_a x_a\,, \quad (18)$$

where the symbol x is the volume fraction of a component in soil, c is the specific heat, ρ is the density and the subscripts m, o, w and a represent the mineral, organic, water and air components. From [4] ρ_m=2650 kg·m^{-3}, ρ_o=1300 kg·m^{-3}, ρ_a=1.2 kg·m^{-3}, ρ_w=2650 kg·m^{-3}, c_m=733 J·kg^{-1}·K^{-1}, c_o=1296 J·kg^{-1}·K^{-1}, c_w=4182 J·kg^{-1}·K^{-1}, c_a=1010 J·kg^{-1}.K^{-1}. It holds

$$x_m + x_o + x_w + x_a = 1\,, \qquad (19)$$

where x_m, x_o, x_w and x_a follow from soil analysis.

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 [15]. The thermal conductivity λ_s was calculated according to [6], [8], [9] and [15] as a combination of dry λ_{dry} and saturated λ_{sat} thermal conductivities, weighted by the Kersten number K_e.

$$\lambda_s = K_e(\lambda_{sat} - \lambda_{dry}) + \lambda_{dry}\,, \qquad (20)$$

where for unfrozen soils

$$K_e = \begin{cases} 0.7\log S_w + 1.0 \text{ for } S_w > 0.05 \text{ and coarse soil} \\ \log S_w + 1.0 \text{ for } S_w > 0.1 \text{ and fine soil} \end{cases}\,, \quad (21)$$

and

$$\lambda_{dry} = \frac{0.135 \cdot \rho_{dry} + 64.7}{2700 - 0.947 \cdot \rho_{dry}}\,, \qquad (22)$$

$$\lambda_{sat} = \lambda_{sol}^{1-x_{por}} \cdot \lambda_w^{x_{por}}\,, \qquad (23)$$

$$x_{por} = x_a + x_w\,, \qquad (24)$$

$$S_w = \frac{x_w}{x_{por}}\,, \qquad (25)$$

$$\rho_{dry} = \rho_m \cdot x_m + \rho_o \cdot x_o + \rho_a \cdot x_{por}\,, \qquad (26)$$

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 and ρ_{dry} is the dry density.

The method was implemented in the Matlab program environment and tested on the observed data measured at the meteorological station called "Vrt Domanin" near the town of Trebon in southern Bohemia. The following variables were measured for the determination of the soil heat flux (G): the volumetric soil moisture x_w and the soil temperatures ϑ_1 and ϑ_2 at the depth 0.01 m and 0.02 m. The soil thermal properties were obtained by the soil analysis. The course of the soil heat flux (G) calculated according formula (17) was compared with the data measured using Huxeflux Heat Flux Plate (HFP01) and with data calculated using formula (13). The intensity of the net radiation Rn was measured by the Net Radiometer (CNR1) manufactured by Kipp & Zonen. The data was recorded at 10 minute intervals. The results in Fig. 4 show the consistency between the values of the soil heat flux measured by the HFP01 and those calculated using formula (17). The fluctuation of the soil heat flux calculated using formula (13) is caused by clouds.

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.

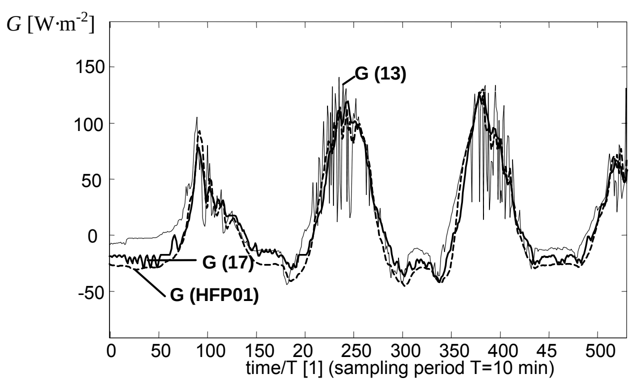

Fig. 4 Soil heat flux G calculated according formulas (13), (17) and measured using HFP01

V. DATA PROCESSING

Three times a day, the data collected from all the meteorological stations was transferred via the GPRS network to a server where it was stored in the Fidler –Magr database. The database was accessible via the Internet to all the participants in the project.

As the meteorological stations provide only point measurements, several times during the year, the countryside was also captured by aerial photography in the visible and infrared wavelengths, using both normal and infrared cameras mounted on a Cessna TU 206-F plane and on an airship as well [7]. The temperature map of the surface, knowledge of the surface and a sufficiently dense network of calibration stations make it thus possible to estimate (calculate) energy fluxes with sufficient precision even in locations off the calibration points. In comparison with satellite remote sensing, the records from the airship can capture detailed information about the selected location at the times for which we need to obtain the temperature map of the landscape. An important advantage is also minimal noise and ability to maintain fixed position of the airship. The main drawback of using airships resides in their limited applicability due to meteorological conditions.

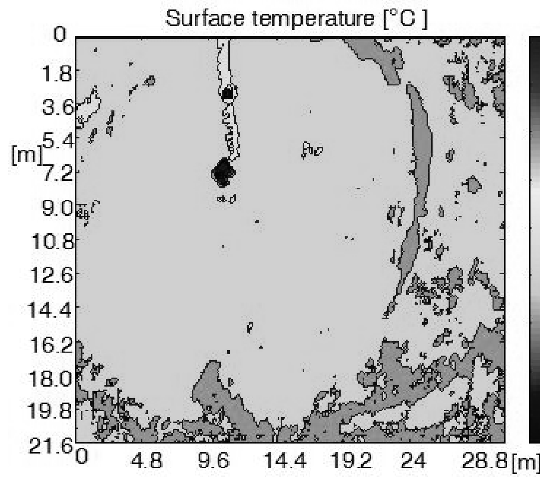

Fig. 5 Surface temperature

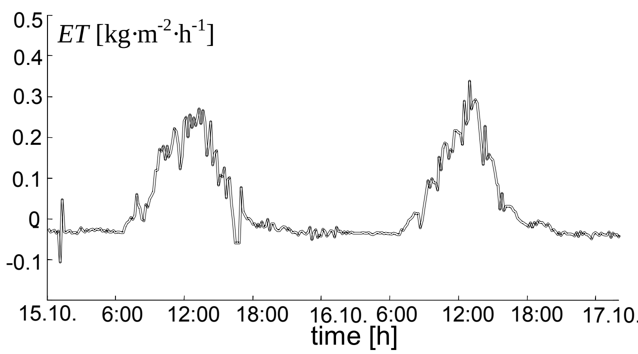

Fig. 6 The intensity of evapotranspiration ET

Matlab programming environment was used for modeling evapotranspiration and examination of hydrology and energy conditions in different types of biotopes. Except for direct processing and evaluation quantities (Fig. 2) observed through meteorological stations, the meteorological data and infrared images were linked. Two-dimensional and three-dimensional images of the energy fluxes and the intensity of evapotranspiration in selected location were obtained. For this purpose, the ThermaCam™ Researcher software from the FLIR company was used in order to process the results and determine the accurate surface temperatures (Fig. 5) from the infrared images. ThermaCam™ Researcher also allows the exporting of data from an infrared image into the Matlab environment. It was thus possible to obtain a visual idea about the distribution of energy fluxes and the evapotranspiration in the landscape.

The time course of the intensity of evapotranspiration during two days is depicted in Fig. 6. Fig. 7 illustrates the computed intensity of evapotranspiration based on data in the vicinity of the calibration point – the meteorological station and on the temperature field obtained from infrared images.

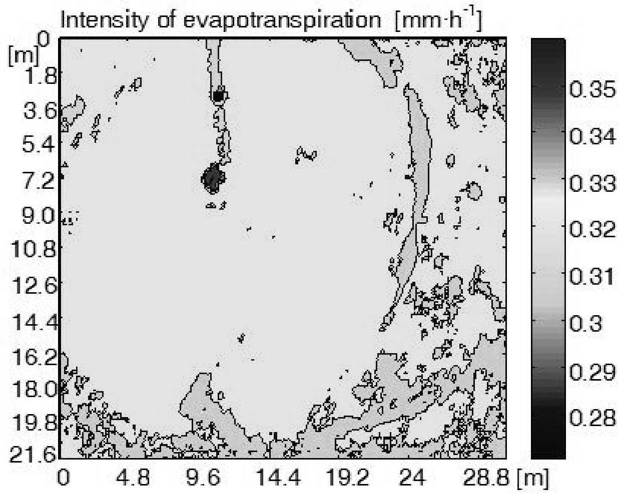

Fig. 7 Intensity of evapotranspiration

VI. Conclusion

The method presented in this paper makes it possible to quantify and visualise the main energy and evapotranspiration fluxes in the landscape based on a temperature map obtained through remote sensing technologies and the measured data obtained from meteorological stations. Due to the sampling period, it was thus possible to detect and evaluate the relatively rapid dynamic changes occurring in the ecosystem. In comparison with satellite remote sensing, records from an airship or a plane can capture detailed information about the selected location at the times needed to obtain the temperature map of the landscape.

An important advantage of using the airship is its minimal noise and its ability to maintain a fixed position. An added benefit is the opportunity to more objectively evaluate the intensity of energy fluxes and evapotranspiration. This is because the calculations are not based solely on spot metering, but also include the temperature profile of the location captured by infrared image. It was shown that in the Matlab environment it is possible to compute and display the main energy fluxes and the intensity of evapotranspiration when the measured data from the meteorological stations is supplemented with information in the form of these infrared images obtained via plane or airship.

Generally, this method has been developed in order to contribute to the assessment of the role of various ecosystems within the countryside and of the impact of human action upon the landscape. The results validate the idea that wetlands, sufficiently supplied with water, are important in the energy and water budget of drained agricultural landscapes.

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References

[1] R. G. Allen, L. S. Pereira, D. Raes and M. Smith, “Crop evapotranspiration – guidelines for computing crop water requirements”, FAO Irrigation and drainage paper 56, Food and Agriculture Organization of United Nations, Rome, 1998. Available: http://www.fao.org/docrep/X0490E/x0490e00.htm#Contents

[2] J. B. Aasher, A. Zilberman and B. B. Yosef: A novel system for remote sensing measurements of evapotranspiration and CWSI of agricultural crops, Proceedings of the 5th WSEAS Int. Conference Advances in Remote Sensing, Finite Differences and Information Security, Prague, 2012, pp. 142-150.

[3] I. S. Bowen, “The ratio of heat losses by conduction and by evaporation from any water surface”, *Physical Review*, Vol. 27, pp. 779-787, 1926.

[4] W. Brutsauert , *Evaporation into the Atmosphere*, Kluwer Academic Publishers, Dordrecht. 1991

[5] E. W. Harmsen, V. H. R. Builes, M. D. Dukes, X. Jia, J. E. Gonzalez and L. R. Perez Alegia.: “A ground-based method for calibrating remotely sensed surface temperature for use in estimating evapotranspiration”, WSEAS Transactions on Environment and Development, Issue 1, Volume 5, pp. 13-23, January 2009

[6] M. Hofreiter, “Estimating soil heat flux at surface soil layer”, Annals of DAAAM for 2011 & Proceedings, Vienna: DAAAM International, 2011, pp. 433-434.

[7] V. Jirka, M. Hofreiter, J. Pokorny and M. Novak, “Method for estimation of surface fluxes of solar energy in landscape based on remote sensing”, Proceedings of VI-International Conference on Environmental Hydrology, Cairo: American Society of Civil Engineers - Egypt Section, 13 p, 2009.

[8] O. Johansen , *Thermal conductivity of soils*, Ph.D. thesis, University of Trondheim, 1977, Available: http://www.dtic.mil/cgi-bin/GetTRDoc?AD=ADA044002&Location=U2&doc=GetTRDoc.pdf

[9] X. Liang, E. F. Wood and D. P. Lettenmaier, “Modeling Ground Heat Flux in Land Surface Parameterization Schemes”, *Journal of Geophysical Research*, vol. 104, no. D8, 1999, pp. 9851-9600.

[10] R. López-Urrea, F. M. S. Olalla, C. Fabeiro and A. Moratalla, “Comparison between various reference evapotranspiration equations for semiarid conditions”, Proceedings of the 4th IASME / WSEAS Int. Conference WHH’09, Cambridge, 2009, pp. 31-38.

[11] J. L. Monteith, *Principles of Environmental Physics*, Edward Arnold (Publishers) Limited, London, 1973.

[12] V. Novák, *Evaporation of water in nature and methods of its determining* (in Slovak), VEDA SAV, Bratislava, 1995.

[13] H. L. Penman, *“Natural evaporation from open water, bare soil and grass”*, Proc. Roy. Soc. London, A193, 1948, pp. 120 – 146

[14] P. J. Perez, F. Castellvi, M. Ibanez and J. I. Rosell, “Assessment of reliability of Bowen ratio method for partitioning fluxes”, *Agricultural and Forest Meteorology*, vol. 97, 1999, pp. 141-150.

[15] C. D. Peters-Lidard, E. Blackburn, X. Liang and E. F. Wood, “The effect of soil thermal conductivity parameterization on surface energy fluxes and temperatures”, *Journal of the Atmospheric Sciences*, vol. 55, pp. 1209-1224, 1998.

[16] R. Rösemann, *Solar Radiation Measurement*, Reichenbach/Fils, 2004.

[17] J. A. Santanello and M. A. Friedl, “Diurnal covariation in soil heat flux and net Radiation”, *J. of Applied Meteorology*, vol. 42, n. 6, pp. 851-862, 2003.,

[18] J. Teixeira, S. Shahidian and J. Rolim, “Selection of a simplified method to compute evapotranspiration for climatic input on an autonomous controller for landscape Irrigation”, 4th IASME/WSEAS Int. conference EEESD’08, Algave, 2008, pp. 380-385.

[19] A. D. Ward and S. W. Trimble, *Environmental Hydrology*, Lewis Publishers, 2004.