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Uncertainty Analysis of Evapotranspiration Estimates in Ecosystems

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***Abstract* - Many hydrologic models and agricultural management applications require evapotranspiration estimates. The intensity of evapotranspiration is mainly determined by mathematical models rather than by direct measurement. In addition to its own estimate of evapotranspiration, it is necessary to determine the uncertainty of this estimate. This uncertainty is not usually mentioned. In this paper these formulas are derived for the uncertainty estimate of evapotranspiration under simplifying assumptions. These assumptions enabled one to derive an expression of evapotranspiration estimation uncertainty suitable for practical applications. The paper focuses on both the absolute and the relative uncertainty of evapotranspiration estimation. The derived formulas can be used for determining the uncertainty in evapotranspiration estimation, but as well as for the accuracy estimate which is necessary for the measuring of input variables. The derived relationship shows that the net radiation should be more accurately measured than the other energy fluxes that have an influence on evapotranspiration. It follows that the relative uncertainty of evapotranspiration is primarily influenced by the relative uncertainty of net radiation. The uncertainty in the measurement of net radiation was derived from data obtained by using a radiometer which was equipped with a pair of pyranometers and with a pair of pyrgeometers. Planck's Law was used for spectral analysis. The possible presence of systematic errors in the measuring of net radiation was evaluated for its potential impact on the errors of the evapotranspiration estimate. This paper is accompanied by measurement records and graphs documenting the achieved results.**

***Keywords:* Evapotranspiration, Uncertainty, Estimate, Ecosystem, Radiation, Radiometer**

I. Introduction

The status of each ecosystem in terms of biodiversity and stability is directly dependent on two factors. The first is energy balance, including incoming and outgoing energy flows; the other is the water balance (hydrological). The monitoring and examination of ecosystems allows us to describe the link between directly and indirectly measured values as well as landscape elements. Monitored ecosystems are examples of complex dynamic systems with distributed parameters which have a number of interactive variables [12].

Evapotranspiration (ET) is the term used to describe the combined process of water loss from the soil surface by evaporation and the 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. The principal factors affecting the rate of evapotranspiration are:

a) ***Weather Conditions:*** Solar radiation, air temperature, humidity, wind speed, etc.

b) ***Crop Factors:*** Crop height variations, crop roughness, reflection, ground cover, crop root system, transpiration resistance, etc.

c) ***Management and Environmental Conditions:*** Soil salinity, land fertility, soil water content, plant density, etc.

The intensity of evapotranspiration is mainly determined using mathematical models rather than by direct measurement with lysimeters (weighing or compensational) or the Eddy Covariance Technique. 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 (PM Method) [1, 5, 6, 8] and the Bowen Ratio Method (BR Method) [7, 9, 13]. Both of these methods are based on the fact that the 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

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

where R_n 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); $H•ET$ is the latent heat flux consumed during evapotranspiration [W•m^{-2}]; λ is the intensity of the sensible heat flux [W•m^{-2}]; G is the intensity of the soil heat flux [W•m^{-2}]; $\lambda$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 [6]

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

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]

$$Rn = \gamma \bullet ET + H + G \tag{4}$$

where only the vertical fluxes are considered and the horizontal fluxes are ignored. The intensity of these energy fluxes (Rn, $\lambda•ET$, H, G), during a 24 hour period on a cloudless day and with a well-watered transpiring surface are schematic sketched in Fig. 1 [4].

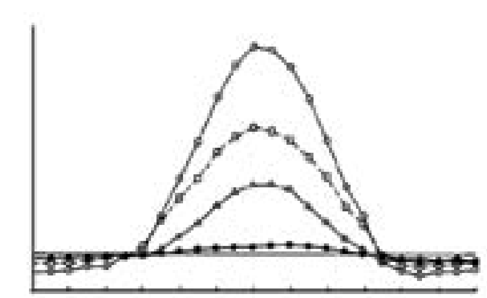

Fig. 1 The intensity of the energy fluxes $Rn, \lambda \cdot ET, H, G$

Evapotranspiration is much more intensive during daylight hours. Therefore the next consideration is restricted to daylight conditions. It holds [10] that

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

where Rs is the intensity of the net shortwave (solar) radiation and Rl 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 3μ m. The fraction α (albedo) of the solar radiation $Rs_{\downarrow}$ [W-m^{-2}] reaching the Earth's surface is reflected as $Rs_{\uparrow}$ [W-m^{-2}] and thus

$$Rs_{\uparrow} = \alpha \cdot Rs_{\downarrow} \tag{6}$$

Therefore it holds that for the intensity of the net shortwave (solar) radiation Rs

$$Rs = Rs_{\downarrow} - Rs_{\uparrow} = Rs_{\downarrow} - \alpha \cdot Rs_{\downarrow} = Rs_{\downarrow}(1-\alpha) \tag{7}$$

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

$$Rl = Rl_{\downarrow} - Rl_{\uparrow} \tag{8}$$

From (4), it is obvious that for the intensity of evapotranspiration ET that

$$ET = \frac{1}{\lambda} \cdot (Rn - G - H) \tag{9}$$

The intensity of the soil heat flux G for daylight conditions can be approximated according to [1]

$$G = 0.4 \cdot e^{-0.5 \cdot LAI} \cdot Rn = \delta \cdot Rn \tag{10}$$

where LAI is the leaf area index and

$$\alpha = 0.4 \cdot e^{-0.5.LAI} \tag{11}$$

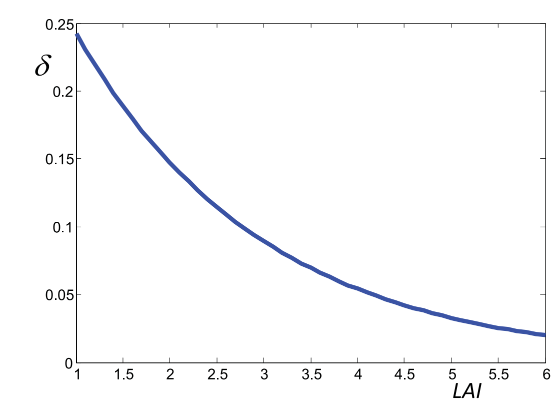

Fig. 2 Dependence δ on LAI

(δ 0.1 = for LAI - 2.8,which is typical for clipped grass, see also Fig. 2).

Let us consider Bowen ratio β defined by

$$\beta = \frac{H}{\lambda.ET} \tag{12}$$

then it follows from (9), (10) and (12)

$$H = \frac{Rn - G}{1+\beta^{-1}} = \frac{Rn - \delta \cdot Rn}{1+\beta^{-1}} = \frac{(1-\delta) \cdot Rn}{1+\beta^{-1}} \tag{13}$$

and

$$ET = \frac{Rn}{\lambda} \cdot \frac{(1-\delta)}{(1+\beta)} \tag{14}$$

II. Standard Uncertainty Of Evapotranspiration Measurement

If the quantity Y is not measured directly, but is determined from n quantities $X_1, X_2, \cdots, X_n$ through a functional relation f

$$Y = f(X_1, X_2, \cdots, X_n) \tag{15}$$

then the estimate y of the quantity Y is determined by the expression

$$y = f(x_1, x_2, \cdots, x_n) \tag{16}$$

where $x_1, x_2, \cdots, x_n$ are the input estimates for the n input quantities $X_1, X_2, \cdots, X_n$. The standard uncertainty $u(y)$ of the estimate y is the positive square root of the estimated variance $u^2(y)$ obtained from

$$u^2(y) = \sum_{i=1}^{n} A_i^2 \cdot u^2(x_i) + 2\sum_{i=2}^{n}\sum_{j<i}^{n-1} A_i \cdot A_j \cdot C(x_i, x_j), \tag{17}$$

where

$$A_i = \frac{\partial f(x_1, x_2, \cdots, x_n)}{\partial x_i} \tag{18}$$

and $C(x_i, x_j)$ is the estimated covariance associated with x_i and x_j [11]. The relative standard uncertainty of x_i is defined as

$$u_r(x_i) \triangleq \frac{u(x_i)}{|x_i|} \tag{19}$$

where $|x_i|$ is the absolute value of x_i and x_i is not equal to zero; $ur(y)$ is the relative standard uncertainty of y. The relative standard uncertainty of y is defined

$$u_r(y) = \frac{y}{|y|} \tag{20}$$

where $|y|$ is the absolute value of y and y is not equal to zero.

The intensity of evapotranspiration ET depends on Rn, H, G. Let us assume that we know their estimates. Then

$$u^2(ET) = \frac{1}{\lambda^2}\left(u^2(Rn) + u^2(H) + u^2(G)\right) - \frac{2}{\lambda^2}C(Rn,H) - \frac{2}{\lambda^2}C(Rn,G) + \frac{2}{\lambda^2}C(G,H) \tag{21}$$

From (10) and (11) it follows

$$C(G,H) = \delta \cdot C(Rn,H) \tag{22}$$

$$C(Rn,G) = E\left[Rn^\circ \cdot \delta \cdot Rn^\circ\right] = \delta \cdot u^2(Rn) \tag{23}$$

where E is a symbol for the expected value and

$$Rn^\circ = Rn - E[Rn] \tag{24}$$

In the equation (21) it is necessary to replace the negative terms with zeroes in order that the uncertainty is not falsely reduced. Since it holds (22), then the value of $\frac{2}{\lambda^2}C(G,H)$ is $\frac{1}{\delta}$ times lower than the value of $\frac{2}{\lambda^2}C(Rn,H)$ and with respect to zeroising of the negative terms it is possible to disregard the positive value of $\frac{2}{\lambda^2}C(G,H)$ For the previous reasons equation (21) can be reduced to

$$u^2(ET) = \frac{1}{\lambda^2}\left(u^2(Rn) + u^2(H) + u^2(G)\right) \tag{25}$$

and the standard uncertainty of the intensity of the evapotranspiration

$$u(ET) = \frac{1}{\lambda}\sqrt{u^2(Rn) + u^2(H) + u^2(G)} \tag{26}$$

From (25) it is obvious that standard uncertainties $u(Rn)$, $u(H)$, $u(G)$ have the same influence on the standard uncertainty $u(ET)$.

The variance $u^2(ET)$ of the intensity of evapotranspiration ET is expressed by (14) and it is equal to

$$u^2(ET) = A_{Rn}^2 \cdot u^2(Rn) + A_\beta^2 \cdot u^2(\beta) + A_\delta^2 \cdot u^2(\delta) + 2 \cdot A_{Rn} \cdot A_\beta \cdot C(Rn,\beta) + \\ +2 \cdot A_{Rn} \cdot A_\delta \cdot C(Rn,\delta) + 2 \cdot A_\beta \cdot A_\delta \cdot C(\beta,\delta), \tag{27}$$

where according to (17) and (18)

$$A_{Rn} = \frac{\partial ET}{\partial Rn} = \frac{(1-\delta)}{\lambda \cdot (\beta+1)}, \tag{28}$$

$$A_\delta = \frac{\partial ET}{\partial \delta} = -\frac{Rn}{\lambda(\beta+1)}, \tag{29}$$

$$A_\beta = \frac{\partial ET}{\partial \beta} = -\frac{Rn(1-\delta)}{\lambda(\beta+1)^2}. \tag{30}$$

Provided that Rn, d, β are uncorrelated, it follows after arrangements with respect to (14) that

$$u^2(ET) = ET^2 \cdot \left(\frac{u^2(Rn)}{Rn^2} + \frac{1}{(1-\delta)^2} \cdot u^2(\delta) + \frac{1}{(\beta+1)^2} \cdot u^2(\beta)\right) \tag{31}$$

III. Relative Standard Uncertainty of Evapotranspiration Measurement

With respect to (19), (20) and (26), the relative standard uncertainty $u_r(ET)$ equals to

$$u_r(ET) = \frac{u(ET)}{|ET|} = \frac{1}{\lambda}\sqrt{\frac{Rn^2}{ET^2} \cdot u_r^2(Rn) + \frac{H^2}{ET^2} \cdot u_r^2(H) + \frac{G^2}{ET^2} \cdot u_r^2(G)} \tag{32}$$

hence

$$u_r(ET) = \sqrt{\left(\frac{Rn}{\lambda \cdot ET}\right)^2 \cdot u_r^2(Rn) + \left(\frac{H}{\lambda \cdot ET}\right)^2 \cdot u_r^2(H) + \left(\frac{G}{\lambda \cdot ET}\right)^2 \cdot u_r^2(G)} \tag{33}$$

Now, one can express with respect to (10),(12) and (14)

$$\frac{G}{\lambda \cdot ET} = \frac{\delta \cdot Rn}{\lambda \cdot ET} = \frac{\delta}{(1-\delta)} \cdot (\beta+1). \tag{34}$$

By means of (12) and (34) expression (33) takes after modifications the following form.

$$u_r(ET) = \sqrt{\left(\frac{1+\beta}{1-\delta}\right)^2 \cdot \left\{u_r^2(Rn) + \delta^2 \cdot u_r^2(G)\right\} + \beta^2 \cdot u_r^2(H)} \tag{35}$$

Result (35) quantitatively describes the dependence of the relative standard uncertainty of evapotranspiration measurement $u_r(ET)$ on the relative standard uncertainties $u_r(Rn)$, $u_r(H)$, $u_r(G)$. Formula (35) shows that the relative standard uncertainty $u_r(ET)$ mostly depends on $u_r(Rn)$, less on $u_r(H)$ and the least influence has $u_r(G)$. However one must be aware that observed ecosystems are examples of complex dynamical systems with distributed parameters. Therefore $u_r(Rn)$, $u_r(H)$, $u_r(G)$ must take into account all sources of variability (uncertainty components), such as instruments, different observers, samples, laboratories, variability of parameters.

Similarly it is possible to derive from (31) and (14) the following expression can for $u_r(ET)$

$$u_r(ET) = \sqrt{u_r^2(Rn) + \frac{\delta^2}{(1-\delta)^2}u_r^2(\delta) + \frac{\beta^2}{(\beta+1)^2}u_r^2(\beta)}. \tag{36}$$

Formulas (26), (35), (31), (36) can be used for uncertainty analyses and for corrections of methodology that is used for the evapotranspiration estimate.

Fig. 3 Meteorological station

IV. Net Radiation Measuring

From the previous sections it is obvious that the extra attention must be paid to the measuring of the net radiation Rn for the evapotranspiration estimate. This section focuses on the estimation of the net radiation Rn. The intensity of the net radiation Rn can be determined by means of (5), (6), (7) and (8), if they are measured the quantities $Rs_\downarrow$, $Rl_\downarrow$ and $Rl_\uparrow$. These quantities can be measured with net radiometers.

Fig. 4 Netradiometer CNRI

In the Czech Republic, a total of 14 meteorological stations were deployed in the selected ecosystem in the southern part of Bohemia. These meteorological stations (see Fig. 3) include recording and control unit M4016 from company Fiedler-Magr. Unit M4016 refers to telemetric stations with an encapsulated GSM / GPRS module, a programmable control machine, which uses various sensors for the reading of meteorological variables such as temperature, humidity, wind speed /direction, radiation, etc. The net radiation is measured by the Net Radiometer CNR 1 from the firm Kipp&Zonen [3]. It measures four radiation components separately because it is equipped with a pair of pyranometers CM3 and with a pair of pyrgeometers CG3 (see Fig. 4).

Let us assume that Rs and Rl are uncorrelated then

$$u^2(Rn)=u^2(Rs)+u^2(Rl) \quad (37)$$

where with regard to (7), (8) and (17),

$$u^2(Rs)=u^2(Rs_\downarrow)+u^2(Rs_\uparrow)-2C(Rs_\downarrow,Rs_\uparrow), \quad (38)$$

$$u^2(Rl)=u^2(Rl_\downarrow)+u^2(Rl_\uparrow)-2C(Rl_\downarrow,Rl_\uparrow). \quad (39)$$

As (6) holds then

$$C(Rs_\downarrow,Rs_\uparrow)=r(Rs_\downarrow,Rs_\uparrow)\cdot u(Rs_\downarrow)\cdot u(Rs_\uparrow)=u(Rs_\downarrow)\cdot u(Rs_\uparrow)=u^2(Rs_\uparrow), \quad (40)$$

because the correlation coefficient

$$r(Rs_\downarrow,Rs_\uparrow)=1 \quad (41)$$

and the measurements $Rs_\downarrow, Rs_\uparrow$ are realized with a pair of identical pyranometers,

where

$$u(Rs_\downarrow)=u(Rs_\uparrow). \quad (42)$$

In equation (38) it is necessary to replace the negative term with zero in order that the uncertainty $u(Rs)$ is not falsely reduced. After modification using (42)

$$u(Rs)=\sqrt{2}\cdot u(Rs_\uparrow). \quad (43)$$

Similarly it is possible to derive

$$u(Rl)=\sqrt{2}\cdot u(Rl_\uparrow), \quad (44)$$

because the measurements $Rl_\downarrow$ and $Rl_\uparrow$ are realized with a pair of identical pyrgeometers where

$$u(Rl_\downarrow)=u(Rl_\uparrow). \quad (45)$$

Formulas (37), (43) and (45) enable to express $u(Rn)$ in the

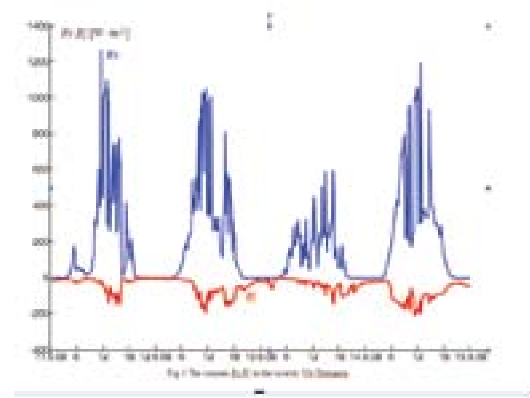

Fig.5 The courses Rs,Rl in the locality Vrt Domanin

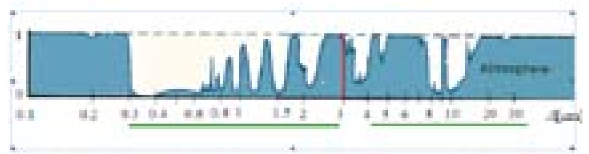

Fig.6 The spectral absorption of the atmosphere

form

$$u(Rn)=\sqrt{2\left[u^2(Rs_\uparrow)+u^2(Rl_\uparrow)\right]}. \quad (46)$$

The spectral range of pyranometer CM 3 is 305-2800 nm and the spectral range of pyrgeometer CG3 is 4.5-42 µm. Fig. 5 shows courses of the measured intensity of the net shortwave (solar) radiation Rs and the intensity of the net longwave radiation Rl in the locality Vrt Domanin (GPS 48°57'49.55"N, 14°44'41.132"E) near the city Trebon in the Czech Republic. The negative Rl means that mostly $Rl_\uparrow$ was greater then $Rl_\downarrow$ during this period.

Now our attention will be focused only on the systematic error in a measurement of the intensity of the net radiation due to the limited spectral range of the Net Radiometer CNR1. Problems related to a calibration, dust, bird droppings, moisture condensation inside the domes, a lack of green vegetation beneath the sensor etc. are not solved here. Fig. 6 shows the spectral absorption of the atmosphere, from [14]. The green lines in Fig. 6 illustrate the spectral range of the Net Radiometer CNR 1. It is obvious that the radiometer covers nearly all important wavelengths where the absorption of the atmosphere is less than 1. But the radiometer CNR 1 does not measure the radiation with wavelengths from 2.8 to 4.5 µm. In this range the absorption of the atmosphere is significantly less than the absorption for the wavelengths from 3.5 to 4 µm. The spectral radiance P [W-m^{-3}] of a black body at temperature T [K] per unit area and for wavelength $\Lambda[m]$ is described by Plank's Low

The spectral radiance of a black body at temperature 6000 K (roughly the surface temperature of the Sun) corresponds

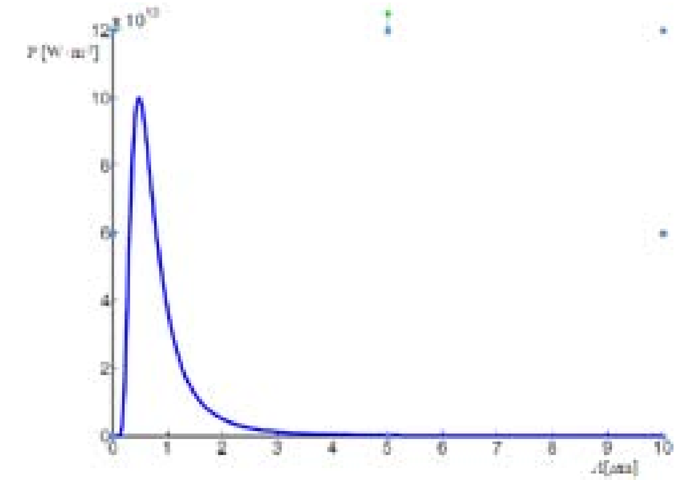

Fig. 7 The spectral radiance from a black body at temperature 6000 K [W-m^{-3}]

to the solar spectrum at the border of the atmosphere, see in Fig. 7.

$$P(\Lambda,T)=\frac{3.73\cdot 10^{-16}}{\Lambda^5\cdot\left(e^{\frac{1.438\cdot 10^{-2}}{\Lambda\cdot T}}-1\right)} \quad (47)$$

If it is defined

$$\phi(\Lambda_1,\Lambda_2,T)\triangleq\frac{\int_{\Lambda_1}^{\Lambda_2}P(\Lambda,T)\,d\Lambda}{\int_0^\infty P(\Lambda,T)\,d\Lambda}=\frac{\int_{\Lambda_1}^{\Lambda_2}P(\Lambda,T)\,d\Lambda}{\sigma\cdot T^4} \quad (48)$$

where Boltzmann's constant $\sigma=5.6697\cdot 10^{-8}\ \mathrm{W\cdot m^{-2}\cdot K^{-4}}$ then the ratio Φ (Λ_1, Λ_2, T) expresses the ratio between the amount of energy emitted in the wavelength range from Λ_1

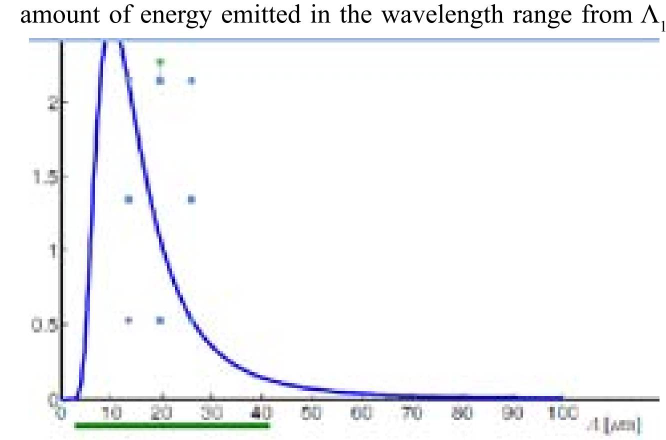

Fig.8 The spectral radiance from a black body at temperature 15°C

to Λ_2 by a black body at temperature T to the total amount energy emitted by this body. For the solved case: Λ_1=3.5µm, Λ_2=4µm and T=6000 K

$$\phi(\Lambda_1,\Lambda_2,T)=0.0039. \quad (49)$$

This ratio is very small and in addition to that the significant part of the radiation between wavelengths 3 to 4 µm is absorbed by the atmosphere. Therefore it is possible disregard this wavelength range.

A bit worse situation is in the monitoring of outgoing flows of longwave energy from the Earth. Fig. 8 shows for example the spectral radiance from a black body at temperature 15°C. The green line in Fig. 8 illustrates the spectral range of the Net Radiometer CNR 1. For the spectral range Λ_1=42µm, $\Lambda_2\to\infty$ and temperature T=288 K=15°C it holds

$$\phi(\Lambda_1,\Lambda_2,T)=0.0537. \quad (50)$$

This relative systematic error for the same spectral range, a black body and the temperature range from 1°C to 40°C is depicted in Fig. 9. The value (Λ_1,Λ_2,T) is the same for a black body with emissivity equals to one and a grey body with emissivity ε because

$$\phi(\Lambda_1,\Lambda_2,T)=\frac{\varepsilon\int_{\Lambda_1}^{\Lambda_2}P(\Lambda,T)\,d\Lambda}{\varepsilon\int_0^\infty P(\Lambda,T)\,d\Lambda}=\frac{\int_{\Lambda_1}^{\Lambda_2}P(\Lambda,T)\,d\Lambda}{\int_0^\infty P(\Lambda,T)\,d\Lambda}=\frac{\int_{\Lambda_1}^{\Lambda_2}P(\Lambda,T)\,d\Lambda}{\sigma\cdot T^4} \quad (51)$$

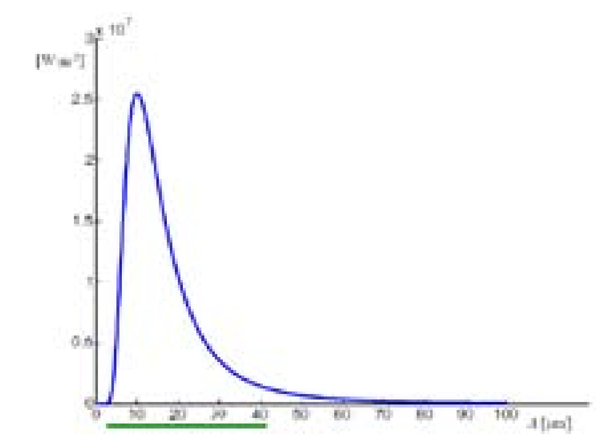

Fig. 9 The relative systematic error [%]

The intensity of radiation emitted over a wavelength range from a grey body with emissivity ε at temperature T can be obtained by means of (51)

$$\varepsilon\int_{\Lambda_1}^{\Lambda_2} P(\Lambda,T)\,d\Lambda = \varepsilon\cdot\phi(\Lambda_1,\Lambda_2,T)\cdot\sigma\cdot T^4 . \qquad (52)$$

This intensity of radiation for the spectral range Λ_1=42µm, $\Lambda_2\to\infty$ and temperature T=288 K is then

$$\varepsilon\int_{\Lambda_1}^{\Lambda_2} P(\Lambda,T)\,d\Lambda = \varepsilon\cdot 20.946\ \mathrm{W\cdot m^{-2}} \qquad (53)$$

The common emissivity range of an evaporating surface [6, 2] is

$$0.96 \le \varepsilon \le 0.98 \qquad (54)$$

Therefore, the intensity of radiation from a grey body within the above mentioned wavelength range and temperature for the average emissivity ε=0.97 is according to (53)

$$\varepsilon\int_{\Lambda_1}^{\Lambda_2} P(\Lambda,T)\,d\Lambda = 0.97\cdot 20.946 = 20.318\ \mathrm{W\cdot m^{-2}} \qquad (55)$$

The latent heat of vaporization λ [J·kg^{-1}] at air temperature t=15°C equals according to [1]

$$\lambda = 2501\cdot 10^3 - 2361\cdot t = 2465585\ \mathrm{J\cdot kg^{-1}} \qquad (56)$$

The intensity of the evaporation equivalent to the intensity of radiation 20.318 W·m^{-2} is

$$\frac{20.318}{\lambda} = 8.2406\cdot 10^{-6}\ \mathrm{kg\cdot s^{-1}\cdot m^{-2}} . \qquad (57)$$

This intensity of evaporation is equivalent to the intensity of evaporation 0.71 mm·day^{-1} (density of water ρ=1000 kg·m^{-3}). The intensity of evaporation [mm·day^{-1}], which is equivalent to the long wave radiation emitted from the Earth and not captured by the Net Radiometer CNR 1, is plotted for different temperatures and emissivity in Fig. 10.

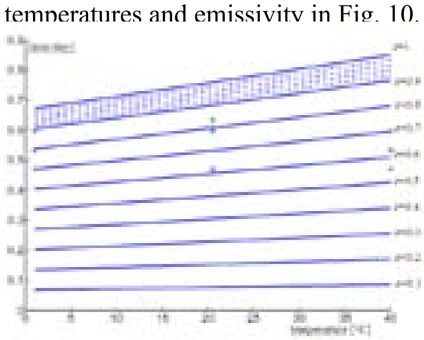

Fig. 10 Intensity of Evaporation Equivalent to the Earth Radiation Not Captured by Net Radiometer CNR 1

V. Conclusion

The purpose of this paper is the improvement of evapotranspiration monitoring. The derived formulas can be used for determining the uncertainty in evapotranspiration estimation, but as well as for the selection of sensors and methods used for evapotranspiration monitoring. These derived relationships show that the measurements of net radiation fluxes require greater attention than the other evapotranspiration influencing energy fluxes. Possible systematic errors in the measuring of net radiation by the Net Radiometer CNR 1 were evaluated for its potential impact on the evapotranspiration estimate. The revealed systematic errors will help to estimate the intensity of net radiation and the intensity of evapotranspiration in ecosystems more accurately.

Acknowledgements

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