ELSEVIER Agricultural and Forest Meteorology 72 (1995) 317–325

AGRICULTURAL AND FOREST METEOROLOGY

Difficulties of estimating evapotranspiration from the water balance equation

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Received 16 August 1993; revision accepted 17 March 1994

Abstract

The effect of soil spatial variability on the estimation of evapotranspiration, using the water balance equation, is evaluated using data from 25 experimental plots, distributed along a transect of 125 m, on a dark red tropical latosol. The variability of soil water storage, total hydraulic gradients, soil hydraulic conductivity and soil water flux densities, and their influence on the calculation of evapotranspiration, are discussed. The variability of these parameters confers a coefficient of variation of the order of 40% to evapotranspiration estimates, indicating that aerodynamic and empiric approaches are a better choice for evapotranspiration estimation of extensive field areas, in which spatial variability of soil hydraulic characteristics is relevant.

1. Introduction

Field water balances demand considerable instrumentation at both upper and lower boundaries of the soil–plant–atmosphere system under study. This fact causes a significant limitation on studies of their spatial variability, which require large numbers of replicates. As a consequence, only few studies discuss the difficulties imposed by the spatial variability of the system on the estabilishment of water balances. Pukkala et al. (1991) describe the spatial distribution of direct radiation below forest canopies, indicating the importance of horizontal variations in the use of

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SSDI 0168-1923(94)02168-J

prediction models. Chason et al. (1991) point to spatial and temporal variabilities of LAI in comparing direct and indirect methods for estimating forest canopy leaf area. At the lower soil boundary, Cisse and Vachaud (1988) discuss the variability of soil physical parameters and their implications on estimates of below root-zone soil water fluxes. Although several other reports discuss aspects of variabilities in space and time, there is a significant lack of contributions on water balance variability as a whole. Here this variability is analysed over a relatively small land strip (125 m long), on which spatial variabilities of incoming rainfall and radiation, and of air mass transfer are likely to be minimal, but in which soil and plant spatial variabilities play an important role.

2. Materials and methods

The experiment was carried out at the field station of ESALQ/USP, county of Piracicaba, SP, Brazil, located in the tropics, 23° S and 47° W, at an altitude of 580 m above sea level, 250 km inside the continent. The climate is Cwa according to Köppens's classification and has one typical rainy season (October–February) and one typical dry season (April–August). Average air temperature is +21°C and average rainfall 1250 mm year^{-1}. The soil is a dark red latosol, known as terra roxa estruturada, which according to the American classification is an Oxic Paleudalf. It has a pronounced textural B horizon at depths varying from 20 to 70 cm from soil surface. Average granulometric analysis showed 25% sand, 15% silt and 60% clay.

Measurements of water balance components were carried out during 1989 and 1990 with the soil having different covers: weeds, corn and bare soil. In the equation

$$P - ET \pm Q_L \pm R \pm \Delta S_L = 0 \tag{1}$$

ΔS_L represents the change in soil water storage in the layer 0–150 cm during a period Δt, measured twice a month with neutron probes calibrated at the same field site; P represents the rainfall during Δt; ET represents the evapotranspiration during Δt; Q_L represents the integral of soil water fluxes q_L at the lower soil boundary ($z = 150$ cm), considered negative downward, also during Δt; and R is the runoff, not measured directly, but estimated from balance.

In order to evaluate the influence of soil spatial variability on calculations using Eq. (1), S and Q_L were measured on a transect of 25 observation points, separated 5 m from each other. Each observation point, which covered an area of 25 m^2, consisted of one 2 m long neutron probe access tube and two mercury manometer tensiometers installed at depths of 135 and 165 cm, in order to estimate the hydraulic gradient at 150 cm. Soil water fluxes $\boldsymbol{q}_{150}$ were estimated through Darcy's equation

$$\boldsymbol{q}_{150} = K(\theta \text{ or } h)[\nabla\psi]_{150} \tag{2}$$

$K(\theta)$ and $K(h)$ relations, presented elsewhere (Reichardt et al., 1993) were determined, at the site, during an internal drainage test. $\nabla\psi$ was estimated by

$$[\nabla\psi]_{150} = \frac{\psi_{135} - \psi_{165}}{30} \tag{3}$$

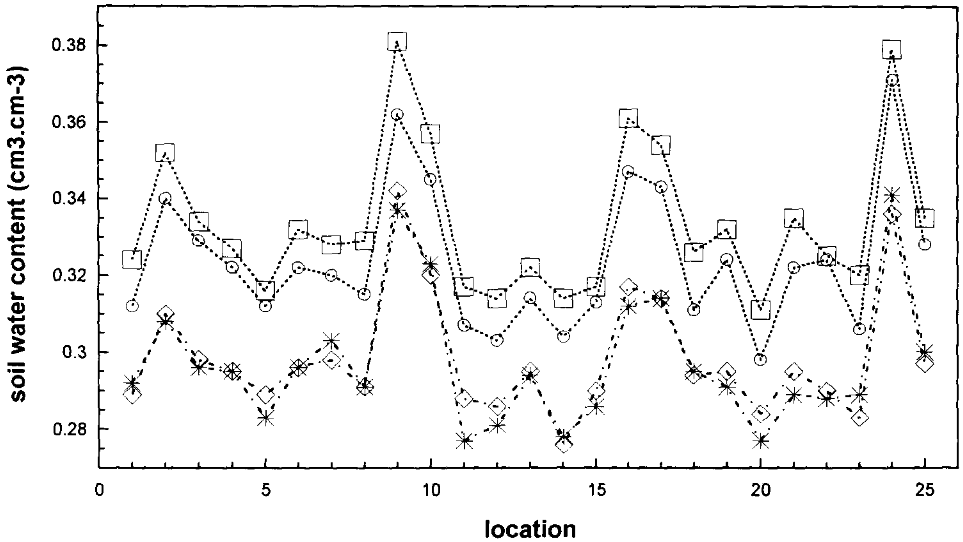

Fig. 1. Soil water content distributions along the transect for several dates. □, 2 March 1989; ○, 30 March 1989; ∗, 3 August 1989; △, 6 September 1989.

considering $\psi = h$ (matric water potential head) $+z$ (gravitational water potential head) in centimetres of water.

Soil water storage S_{150} was calculated from soil water content (θ, cm^3 cm^{-3}) data measured at depths of 25, 50, 75, 100, 125 and 150 cm, using Simpsons's rule

$$S_{150} = 1/3[\theta_0 + 4\theta_{25} + 2\theta_{50} + \ldots + 2\theta_{100} + 4\theta_{125} + \theta_{150}]\Delta z \tag{4}$$

Because of difficulties in measuring θ at soil surface with the neutron probe, and the fact that the measurement at 25 cm almost reaches soil surface, θ_0 was considered equal to θ_{25}. Negligible error was introduced by this procedure.

In periods where $R = 0$, ET was estimated from the balance Eq. (1) and in wet periods ET was taken as 0.75 Ev. 'A' (Class A pan evaporation), in order to estimate R. The rainfall P was taken as uniform because the whole experiment covered only a strip of 5×125 m^2, and was therefore measured by one single raingauge.

3. Results and discussion

*3.1. Raw data: soil water content (θ) and soil matric potential (*h*)*

Soil water content data, measured at least twice per month at the 25 plots over the 24 months of observations, showed spatial coefficients of variation over the time period and at each depth on the order of 3%, ranging from 1% to 9% for all the depths. For a given depth, e.g. 150 cm, typical means and variances for dry and wet periods were 0.267 cm^3 m^{-3} (1.69 10^{-4}) and 0.342 cm^3 cm^{-3} (8.1 10^{-5}), respectively.

In this study neutron probes proved again to be the most suitable methodology for

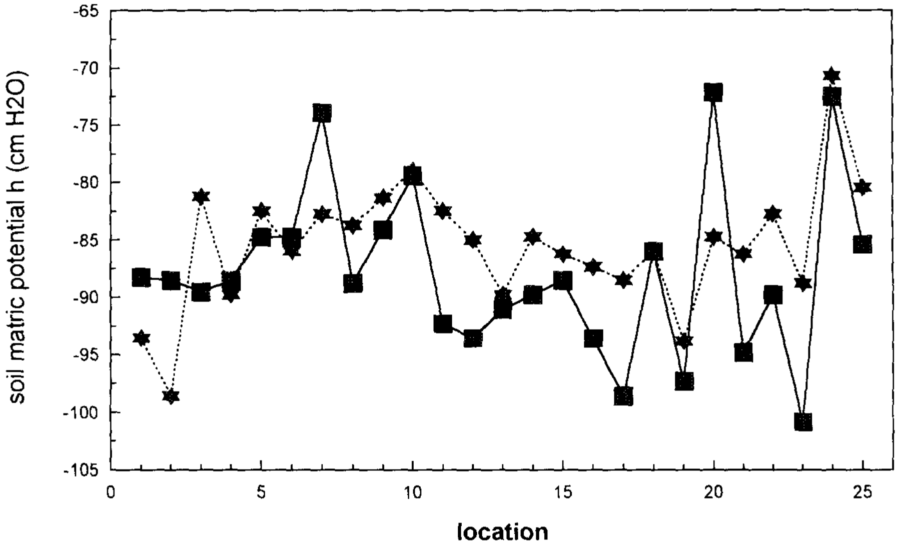

Fig. 2. Soil matric potential distribution along the transect for 12 February 1990. ■, 135 cm depth; ★, 165 cm depth.

the measurement of soil water contents, when many observations have to be performed in space and time. The consistency in time shown in Fig. 1 indicates that with neutron probes the 'same' soil volume is 'sampled' for soil water content, at each observation date.

As will be seen below, the observed variability of θ did not present significant problems in the estimation of soil water storages S_{150}, but it did have a major effect on the calculations of hydraulic conductivities K.

Soil water matric potential head (h) data showed over the same observation period spatial coefficients of variation at each depth on the order of 8%, ranging from 5% to 84% for both depths. For a given depth, e.g. 135 cm, typical means and variances during dry and wet periods were −426.7 cm H_2O (6988.9) and −72.6 cm H_2O (141.5). Fig. 2 is a sample for a wet period in which it can be seen that matric gradients oscilate from positive to negative, because at the observation time the wetting front originated by a heavy rainfall had not reached all 25 observation points. As discussed for soil water content data, the variability of h (and ψ) caused great difficulties in the estimation of hydraulic conductivities K and hydraulic gradients $\nabla\psi$.

3.2. Soil water storage (S_L) and its changes (ΔS_L)

Soil water storage, integrated according to Eq. (4), showed very low coefficients of variation, on the order of 3% ranging from 1% to 6%. Data indicate that the water storage capacity of this soil, for agricultural purposes, is low. Minimum and maximum values observed over the whole period were 432 mm and 565 mm, respectively, which represents a practical value of available water on the order of 133 mm in the 0–150 cm layer.

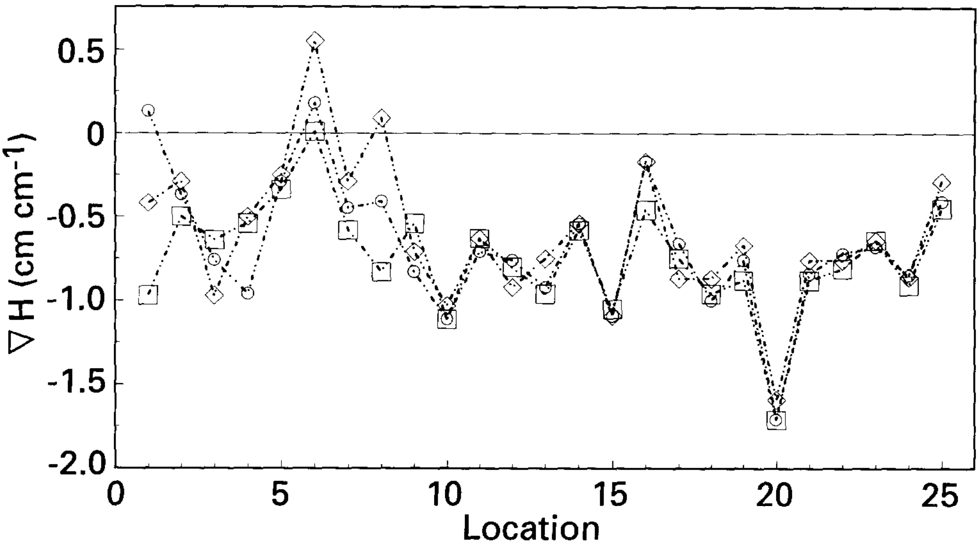

Fig. 3. Hydraulic gradient distributions for sample days in a dry period. ○, 22 June 1989; □, 3 July 1989; △, 12 July 1989.

Although S_L can be well estimated because of the low coefficient of variation, ΔS_L estimates might show very high coefficients of variation depending on the time interval Δt chosen to establish a water balance. In periods where initial and final values of S_L are similar, ΔS_L approaches zero and, logically, high values of the coefficient of

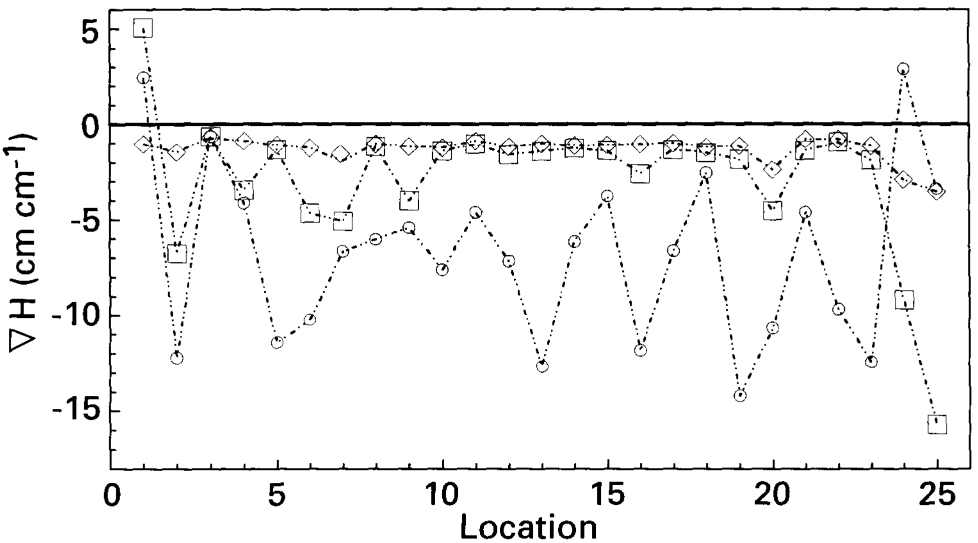

Fig. 4. Hydraulic gradient distributions for sample days in a wet period. ○, 4 January 1990; □, 6 January 1990; △, 8 January 1990.

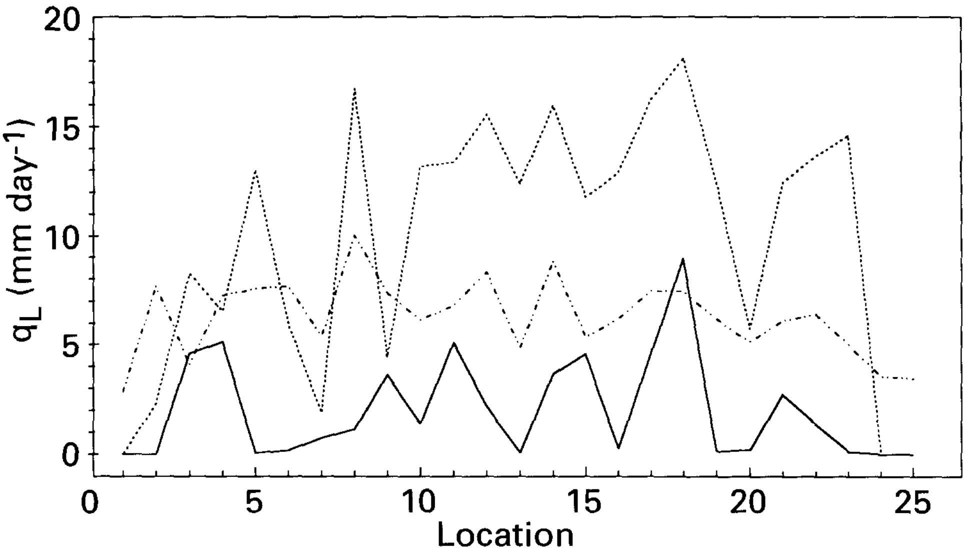

Fig. 5. Soil water flux density distributions at the lower soil boundary (150 cm) for sample days in a wet period. Solid line, 4 January 1990; dashed line, 6 January 1990; dotted dashed line, 8 January 1990.

variation are observed. In these cases, although the uncertainty of ΔS_L determination is high, their actual values are very small and do not significantly affect water balance calculations.

3.3. Total hydraulic gradients ($\nabla\psi$)

Total hydraulic gradients calculated according to Eq. (3), for the 25 locations had over the considered time period coefficients of variation on the order of 50%, ranging from 19% to 254%.

Figs. 3 and 4 show $\nabla\psi$ spatial distribution for sample days in dry and wet periods. Negative values of $\nabla\psi$ indicate downward flow. As can be seen, even in dry periods, almost all plots showed downward flow. These figures indicate that the estimation of soil water flux densities under field conditions is feasable from point to point; however, the use of the data in terms of a mean field behaviour, is still a problem to be solved.

3.4. Soil hydraulic conductivity

As stated in Eq. 2, soil hydraulic conductivity was related to θ and h, during an internal drainage experiment. $K(\theta)$ relations were of the form $K_0 \exp\gamma(\theta - \theta_0)$ and $K(h)$ relations $\alpha \exp\beta h$. For the θ relations, the variabilities of K_0 and γ are characterized by coefficients of variation on the order of 30%, for the h relations; α and β presented coefficients of 48 and 14%, respectively. As an example, for 1 day during a wet period, in which soil drainage was significant (13 February 1990), values of

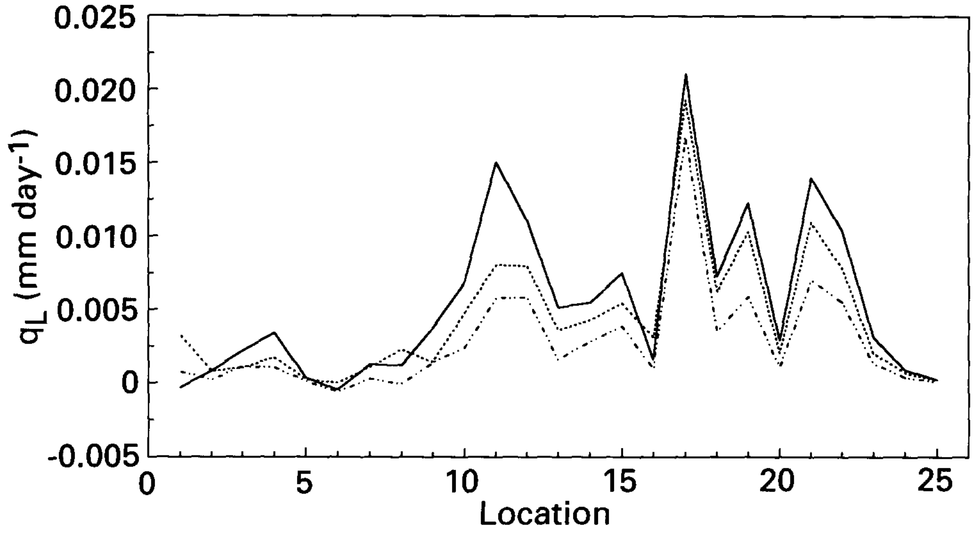

Fig. 6. Soil water flux density distributions at the lower soil boundary (150 cm) for sample days in a dry period. Solid line, 4 January 1989; dashed line, 3 July 1989; dotted dashed line, 12 July 1989.

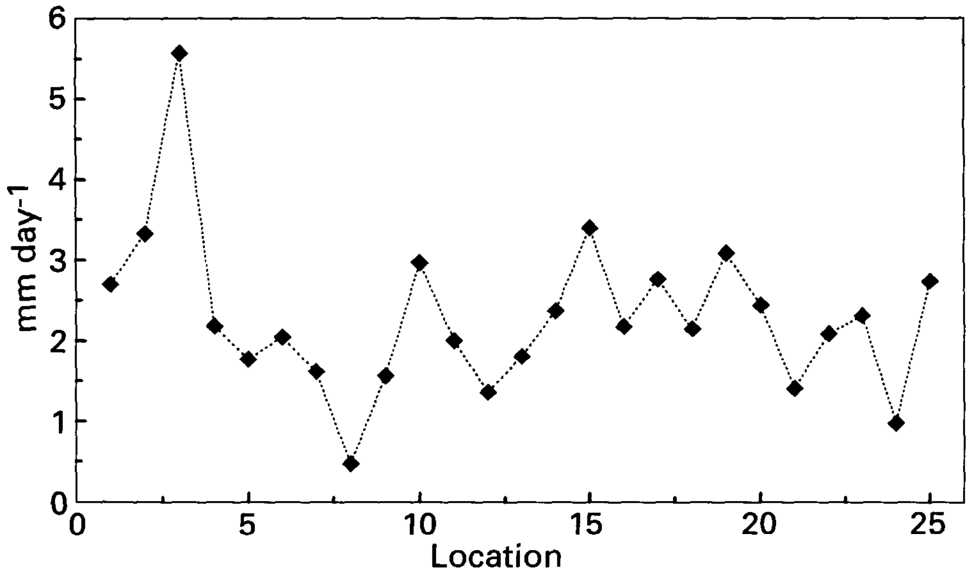

Fig. 7. Distribution of the ET_a, calculated from the balance equation, along the transect from 29 November 1990 to 14 December 1990 (corn).

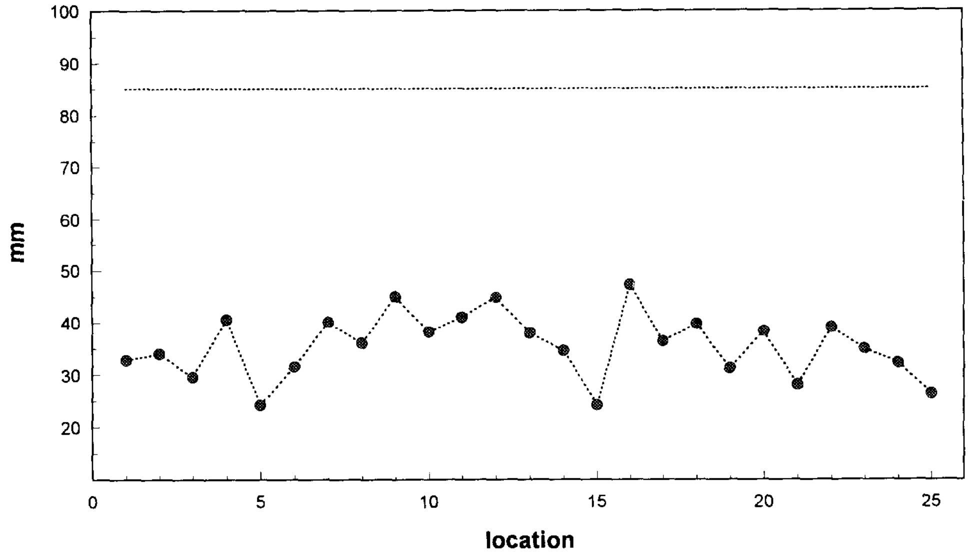

Fig. 8. Distribution of the run-off, calculated from the balance equation, along the transect from 19 July 1989 to 3 August 1989 (bare soil). Dashed line, rainfall (mm); ●, runoff (mm).

K(mm day^{-1}) estimated for the 25 plots, at the 150 cm depth, yielded average values of $K(\theta)$ and $K(h)$, not significantly different, as follows: $K(\theta) = 4.72$ mm day^{-1} (CV = 72%) and $K(h) = 3.42$ mm day^{-1} (CV = 27%). These CV values indicate that tensiometers give better estimates of hydraulic conductivities than neutron probes. However, in either case, the same difficulties discussed above for the hydraulic gradient will not allow a precise description of the mean field behaviour with respect to hydraulic conductivities. Details of these difficulties are discussed by Reichardt et al. (1993). Warrick and Nielsen (1980) classify soil parameters according to their variability in low, medium and high variation. In the high variation class they show hydraulic conductivity data with coefficients of variation ranging from 170 to 400%, much higher than in our case (CV on the order of 50%), which indicates that the soil here studied is not an exception with respect to spatial variability.

3.5. Soil water flux densities

Figs. 5 and 6 show examples of soil water flux densities at the depth of 150 cm, for dry and wet periods. Coefficients of variation are on the order of 60%. This shows that it is practically impossible to obtain representative estimates of soil water flux densities for this field.

For the sample day in the wet period (Fig. 5), in which q_{150} fluxes represent a major component of the balance, several plots presented maximum fluxes between 15 and 20 mm day^{-1}, whereas others presented negligible values. For this day the average flux was 4.2 mm day^{-1}, with a coefficient of variation of 55%. Another complication is that, as observed in other situations (Rao et al., 1979; Greminger et al., 1985), these data do not follow normal distribution.

A similar variability of q_L was observed for dry periods (Fig. 6) but because the values are insignificant, there is no implication in water balance calculations.

3.6. Evapotranspiration

Fig. 7 presents an example of ET variability, during a wet period. For the chosen period, the runoff R component was zero. The average of ET values for this period is 2.1 mm day^{-1} with a CV of 42%. The variability is extremely high and indicates that spatial variability of soil properties and processes strongly limit the use of the water balance equation under field conditions.

3.7. Run-off

During a dry period, after an exceptional rainfall of high intensity when run-off was observed, ET was assumed constant and equal to 0.75 of class A evaporation pan (34.3 mm) and run-off was estimated from the application of the water balance Eq. (1), plot by plot. The results, presented in Fig. 8 show again the great effect of soil variability on this water balance component.

4. Conclusions

The use of the water balance equation to estimate evapotranspiration yields average values with coefficients of variation on the order of 42%, mainly as a result of the variability of soil parameters. This shows that aerodynamic and empiric approaches to estimate evapotranspiration, although based mainly in atmospheric observations, are a better choice for ET estimation in soils with expressive spatial variability of soil hydraulic characteristics.

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