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"Difficulties of Estimating Evapotranspiration from the Water Balance Equation"

K. Reichardt
Universidade de Sao Paulo
Centro de Energia Nuclear na Agricultura (CENA)
Departamento Fisica del Solos
Campus de Piracicaba
Caixa Postal 96
13418-260 Piracicaba (SP)
Brazil

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DIFFICULTIES OF ESTIMATING EVAPOTRANSPIRATION FROM THE WATER BALANCE EQUATION[1]

M.M.Villagra[2]; O.O.S.Bacchi[3]; R.L.Tuon[4]; K.Reichardt[5]

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 great limitation on spatial variability studies, which require large number of replicates, and as a consequence, only few contributions discuss the difficulties imposed by the spatial variability of the system on the estabilishment of water balances. Pukkala et al. (1991) discuss the spatial distribution of direct radiation below forest canopies indicating the importance of horizontal variations in the use of prediction models. Chason et al. (1991) point to LAI's spatial and temporal variabilities 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 its implications on the estimatives of below root-zone soil water fluxes. Although several other reports discuss aspects of variabilities in space and time, there is a significant lack in contributions of 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 incomming rainfall and radiation, and of air mass transfer are surely minimal, but on 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 1,250 mm/year. 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 90 with the soil having different covers: weeds, corn and bare soil. In the equation:

$$P - ET \pm Q_L \pm R \pm \Delta S_L = 0 \qquad (1)$$

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[1]Contribution of Center for Nuclear Energy in Agriculture (CENA), University of São Paulo, C.P. 96, 13.418-970, Piracicaba, SP, Brasil.

[2]Department of Soil Physics, University of Gent, Coupure Links 653, 9000, Gent, Belgium

[3]Research Assitant at CENA, CNPq fellow.

[4]Research Technician at DFM/ESALQ, University of São Paulo.

[5]Professor of Physics and Meteorology at ESALQ, University of São Paulo, CNPq fellow.

i) ΔS_t 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; ii) P the rainfall during Δt; iii) ET the evapotranspiration during Δt; iv) Q_L the integral of soil water fluxes q_L at the lower soil boundary ($z = 150$ cm), considered negative downward, also during Δt; and v) R the runoff, not measured directly, but estimated from balance.

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

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

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

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

considering ψ = h (matric potential) + z (gravitational potential), all in cm 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

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

Due to diffculties of measuring Θ at soil surface with the neutron probe, Θ_0 was considered equal to Θ_{25}.

In periods where R = 0, ET was estimated from the balance equation (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 since the whole experiment covered only a strip of 5x125 m, and 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.cm^{-3} (1.69 10^{-4}) and 0.342 cm^3.cm^{-3} (8.1 10^{-5}), respectively.

In this study neutron probes showed again to be the most suitable methodology for the measurement of soil water contents, when many observations have to be performed in space and time. The consistency in time shown in Figure 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 great problems in the estimation of soil water storages S_{150}, had, however, a major effect on the calculations of hydraulic conductivities K.

Soil water matric potential (h) data showed a spatial coefficient of variation at each depth on the order of 8%, ranging from 5% to 84% for all the 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). Figure 2 is a sample for a wet period in which it can be seen that matric gradients oscilate from positive to negative, because the wetting front originated by a heavy rainfall did not reach all 25 observation points at the same time. 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 $\nabla\psi$.

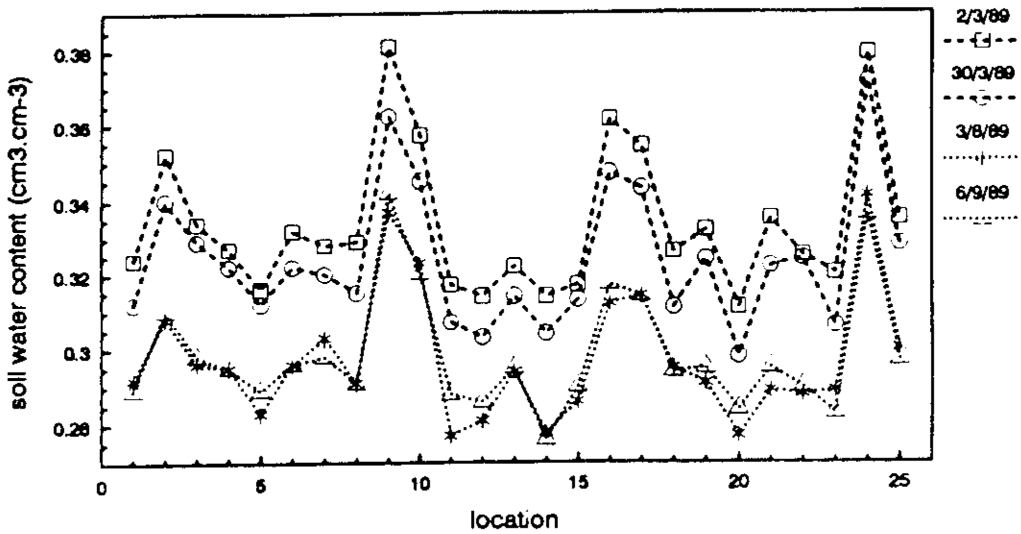

Figure 1. Soil water content distributions along the transect for several dates.

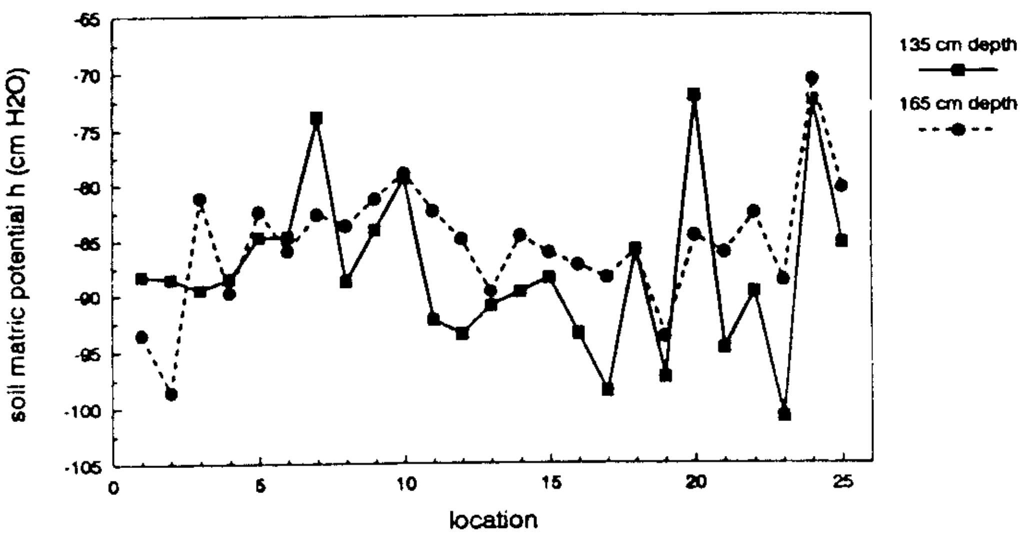

Figure 2. Soil matric potential distribution along the transect for (12/2/1990)

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

Soil water storage, integrated according to equation (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, which represents a practical value of available water on the order of 133 mm in the 150 cm layer.

Although S_L can be well estimated due to 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 high values of the coefficient of 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 equation (3), for the 25 locations had over the considered time period coefficients of variation on the order of 50%, ranging from 19% to 254%.

Figures 3 and 4 show ∇H spatial distribution for sample days in dry and wet periods. Negative values of ∇H indicate downward flow. As can be seen, even in dry periods, almost all plots showed downward flow. These figure 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.

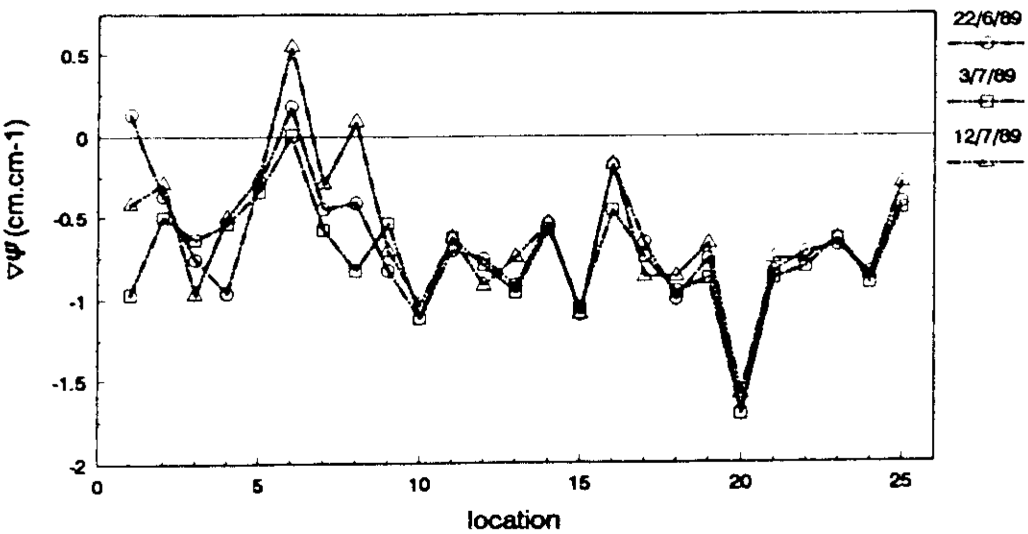

Figure 3. Hydraulic gradient distributions for sample days in a dry period.

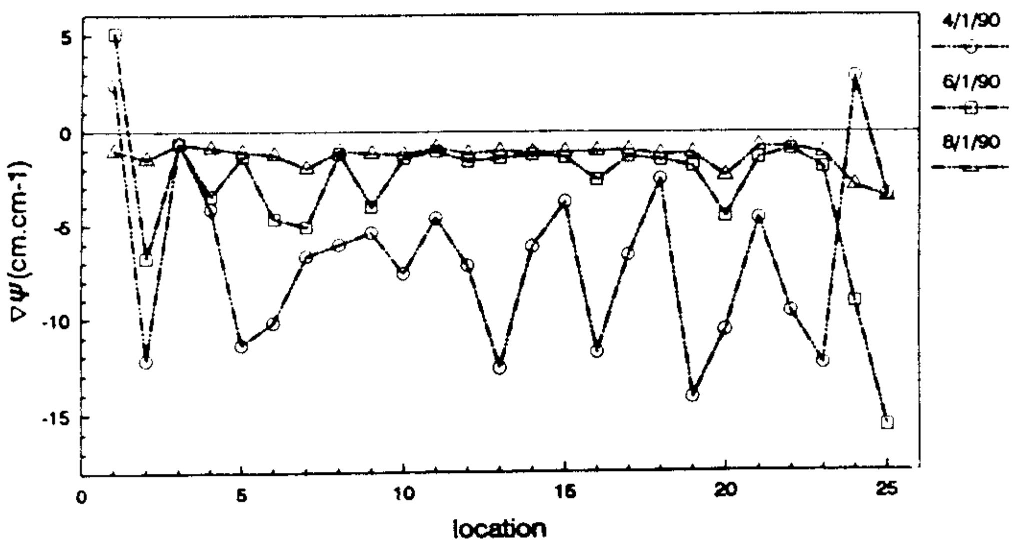

Figure 4. Hydraulic gradient distributions for sample days in a wet period.

3.4. Soil Hydraulic Conductivity

As stated in equation 2, soil hydraulic conductivity was related to Θ and h, during an internal drainage experiment, described elsewhere (Villagra, 1991). K(Θ) relations were of the form $K_o \exp \gamma(\Theta-\Theta_o)$ and K(h) relations equal to $\alpha \exp \beta h$. The variabilities of K_o and γ are characterized by coefficients of variation on the order of 30%, and those of α and β of 48 and 14%, respectively. As an example, for one day during a wet period, in which the water flow was significant (13/02/1990), values of K (mm/day) estimated for the 25 plots, at 150 cm depth, yielded average values of K, not significantly different, as follows: K(Θ) = 4.72 mm day^{-1} (CV = 72%) and K(h) = 3.42 mm.day^{-1} (CV = 27%). These C.V. values indicate that tensiometers are more suitable than neutron probes to estimate hydraulic conductivities in this field. However, in either case, the same difficullties discussed above for the hydraulic gradient, will not alow a precise description of the mean field behaviour with respect to hydraulic conductivities. Details of these difficulties are discussed in Reichardt et al (1993). Warrick & 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 variability.

3.5. Soil water flux densities

Figures 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 under field conditions.

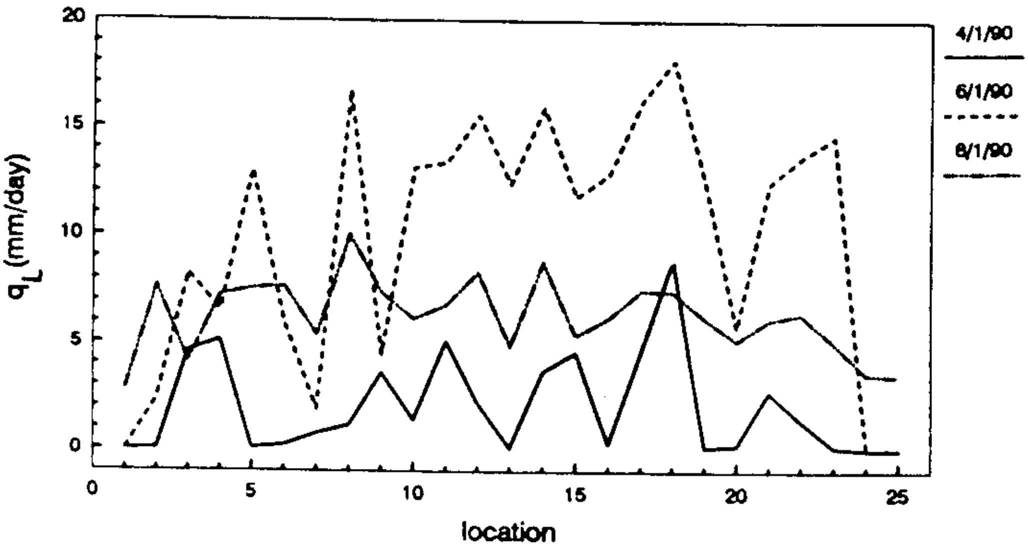

Figure 5. Soil water flux density distributions at the lower soil boundary(150cm) for sample days in a wet period.

For the sample day in the wet period (Figure 5), in which q_{150} fluxes are not negligible, maximum and minimum values were 10.5 and 2.0 mm.day^{-1}, the average 4.82 mm.day^{-1}, and the coefficient of variation 55%. Another complication is that, as observed in other situations (Rao et al. 1979 and Greminger et al. 1985), this data does not follow normal distribution.

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

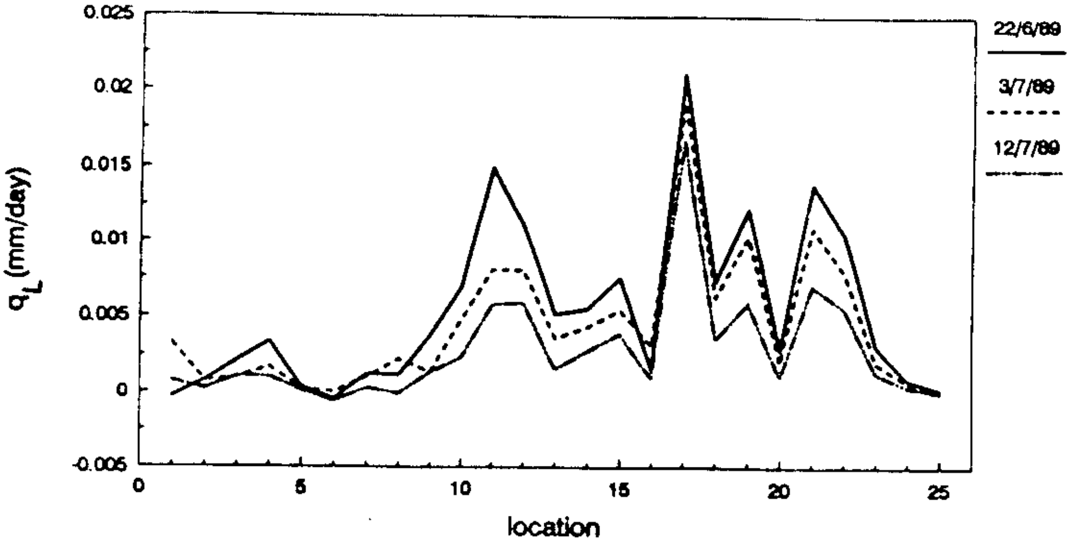

Figure 6. Soil water flux density distributions at the lower soil boundary(150cm) for sample days in a dry period.

3.6. Evapotranspiration

Figure 7 presents an example of ET variability, during a wet period, calculated from the water balance equation (1), applied plot by plot. For the period chosen, the runoff R component was zero. The average values for this period is 2.1 mm.day^{-1} with a C.V. of 42%. The variability is extremely high and indicates that spatial variability of soil properties and processes strongly limits the use of the water balance equation under field conditions.

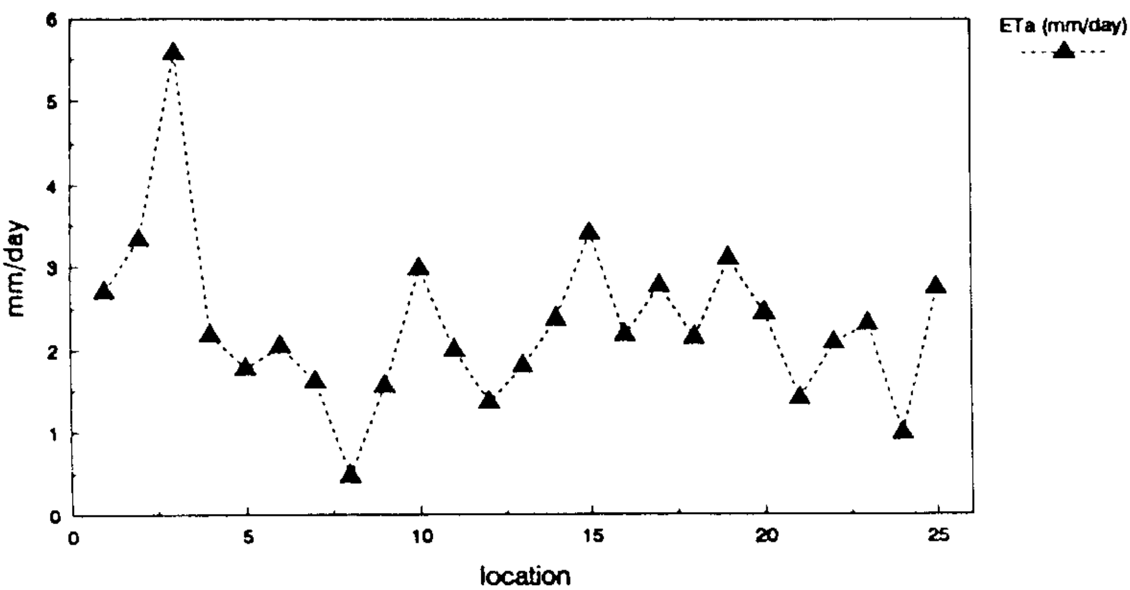

Figure 7. Distribution of the ET_a, calculated from the balance equation, along the transect from 29/11/1990 - 14/12/1990 (corn).

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 equation (1), plot by plot. The results, presented in figure (8) show again the great effect of soil variability on this water balance component.

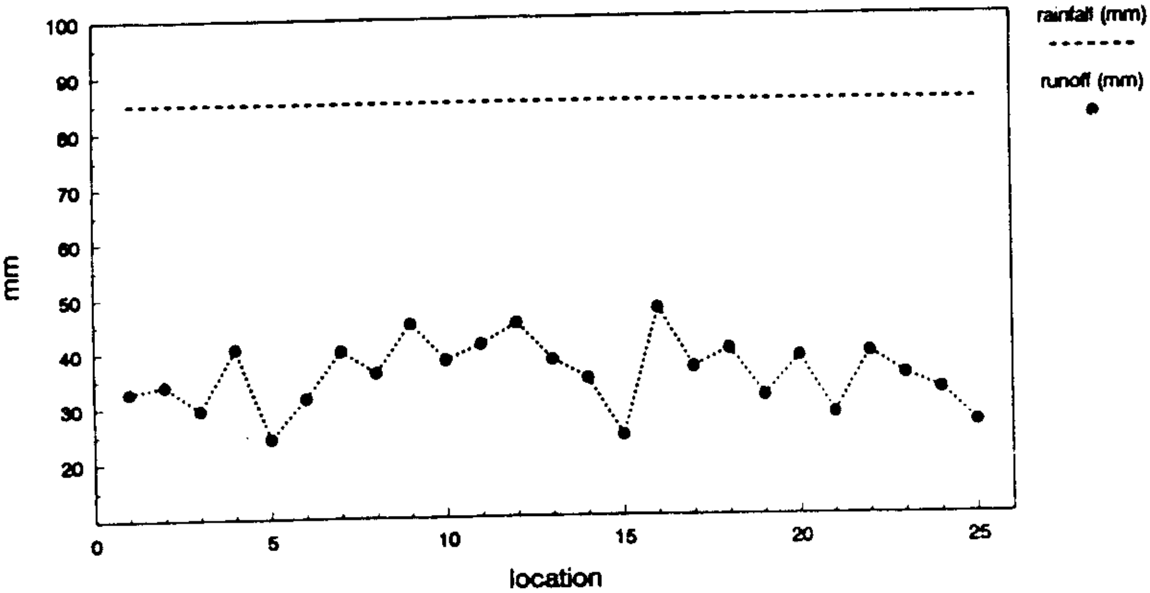

Figure 8. Distribution of the run-off, calculated from the balance equation, along the transect from 19/7/1989 - 3/8/1989 (bare soil).

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 due to the variability of soil parameters. This shows that aero-dynamic and empiric approaches to estimate evapotranspiration, although based mainly in atmospheric observations, are still the best choise for ET estimation.

5. REFERENCES

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