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SMR/147-38

COLLEGE ON SOIL PHYSICS

15 April - 3 May 1985

COLLOQUIUM ON ENERGY FLUX AT THE SOIL ATMOSPHERE INTERFACE

6 - 10 May 1985

ESTIMATION OF ACTUAL EVAPORATION DURING A DRY PERIOD:
COMPARAISON BETWEEN MODEL AND MEASUREMENTS IN LOUVAIN-LA-NEUVE

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This work investigates some theorical and pratical aspects of modelling soil water flow and evaporation in and from an homoge - neous unsaturated finite soil core. Its main feature include a non linearized solution of the highly non linear surface boundary con - dition and, a specific way to prevent unstabilities development during the simulation run . The model is tested and validated by simulating the changes in soil wetness under natural conditions, and, by comparing the actual evaporation calculated to the potential one, - obtained using the Penman-Monteith equation.

To be presented at the International Centre for Theoretical Physics, Trieste (Italy), May 6-10 1985.

INTRODUCTION

Soils always remain bare through the periods of tillage, planting, germination and early seedling growth (Hillel, 1978). During these periods, soil evaporation can deplete completely the soil moisture and thus, seriously affects growth of youngs plants during their most vulnerable stage. On the other hand, injudicious management of both soil and atmospheric water (rain) generally cause, even on bare soil a great loss of water and, furthermore, give rise to phenomena like erosion and salinization which at long term would probably accelerate the soil deterioration process.

These two examples, rely on many others justify studies of soil evaporation process in nature that, are continuously made since a century by meteorologists, plant physiologists, agricultural engineers and hydrologists.

Measurements of soil evaporation are scattered and only few are available. The main reason is probably the lack of an accurate measurement technique (Reynolds and Walker, 1984).

Modelling is an alternate tool for evaporation assessment. However it is limited by the complexities and by the number of interracting and interdependent factors involved in the evaporation process ; thus only simplified approach are now used to estimate or to predict its magnitude and its rate.

Our purpose in this study are

i) to simulate soil water flow in a soil core with solution of an one-dimensional numerical soil water flow model;

ii) to evaluate the resulting actual soil evaporation;

iii) and finally to compare these latter model results to potential evaporation estimated in the same climatic conditions.

STRUCTURE OF THE MODEL

Two main systems ordinarily interact during the evaporation process. (i) the atmospheric system which provides energy required to meet the latent heat requirement of evaporation and (ii) the soil system which, continuously supplies water to the evaporation site.

The structure of the model consists thus of two sections, for separate treatments of energy flux in the atmospheric system and soil water flow in the soil system.

ATMOSPHERIC SECTION

Accounting for the energy balance at the soil surface, potential evaporation is calculated in this section with the Monteith (1965) version of the Penman equation (1946). This is written as follow :

$$L_{ETP} = \frac{s(R_n - G) + \rho C_p \delta e / r_a}{s + \gamma} \qquad (1)$$

where

R_n = net radiation ($W\ m^{-2}$)
ρ = air density ($Kg\ m^{-3}$)
C_p = specific heat of the air at constant pressure ($J\ Kg^{-1}\ °C^{-1}$)
δe = the vapor pressure saturation deficit in the air (P_a)
s = the slope of the saturation pressure (P_a/°C) curve at the air temperature
γ = the psychrometric constant (P_a/°C)
r_a = the aerodynamic resistance (s/m)
L = the latent heat of evaporation ($J\ Kg^{-1}$)

The aerodynamic resistance depends of the wind speed and soil roughness. It is calculated by the following equation (van Bavel and Hillel, 1976)

$$r_a = [\ln (Z/Z_0)]^2 \;/\; 0.16 \times U \tag{2}$$

where Z_0 is the roughness length (m), U(m/s) the wind speed, and Z the reference level at which U is measured (2m).

SOIL SECTION

The one-dimensional soil water flow model to be described consist of a set of non linear equations generally accepted for description of water flow in unsaturated homogeneous soils (Richards, 1931; Childs and Childs and Collis-George, 1950). The principal equation is derived from theory based on :

(i) total potential of water in soil,

$$\Phi = \Psi + Z + \Omega \tag{3}$$

(ii) Darcy law

$$\nu = -K \nabla_Z \Phi \tag{4}$$

(iii) continuity equation

$$\frac{\partial \theta}{\partial t} = -\nabla_Z \cdot \nu \tag{5}$$

where

Φ is the hydraulic potential

Ψ the matric potential, Z the gravitational potential and Ω the overburden potential.

K is the hydraulic conductivity, ν the water flux, θ the volumetric soil moisture content, t the time and ∇ the "del" operator, taking here only along the vertical since the flow is considered to be one-dimensional.

This Principal equation (The Richard's equation) is expressed as:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\theta) \frac{\partial \theta}{\partial z} \right) \tag{6}$$

where Z now designates the vertical coordinate (taken positive downward) and $D(\theta)$ the hydraulic diffusivity. Note that we disregard here the gravity flow, neglecting its effect during the evaporation process under our fixed conditions.

Equation (6) is then solved according to the set of flow conditions :

(i) the boundaries conditions

a) along the atmosphere interface (Philip, 1957; Reynolds and Walker, 1984)

$$\nu_s = -D(\theta_s) \left.\frac{\partial \theta}{\partial z}\right|_{z=0} = \nu_a = -\rho_a C_{p_a} \left(RH \cdot e_s(T_s) - e_a \right) / \gamma r_a \tag{7}$$

where ν_s is the evaporative flux accross the soil surface as defined by moisture flow.

ν_a, the evaporative flux accross the soil surface as defined by atmospheric water flow [M/s]. θ_s the soil moisture at the soil surface; RH the relative humidity of the air at the soil surface, $e_s(T_s)$, the saturated vapor pressure of the air at the soil surface temperature (P_a) and e_a the atmospheric vapor pressure (P_a). The relative humidity (RH) is calculated by :

$$RH = \exp \left(G \cdot \Psi(\theta_s) / R_w T_s \right) \tag{8}$$

where $\Psi(\theta_s)$ is the matric potential (M) at the volumetric surface moisture content (θ_s), G , the gravitational constant (9.81 m s^{-2}), R_w the gas constant for water vapor ($4.615 \cdot 10^2$ m^2 s^{-2} k^{-1}) and T_s the soil surface temperature (K).

The saturated vapor pressure at (T_s) is calculated using the following empirical equation suggested by Bögel (1981)

$$e_s(T_s) = 6.1121 \exp [(18.564-T_s/254.4)/(T_s+255.57)] \qquad (9)$$

b) boundary condition along the bottom of the soil column

$$v_b = -D(\theta_b)\frac{\partial\theta}{\partial z}\Big|_{z=L} = 0 \qquad (10)$$

where v_b is the evaporative flux at this boundary θ_b the volumetric soil moisture content and L the length of the column.

c) initial conditions

These conditions provide the initial soil moisture profile when running the model.

The critical part for the successful use of (6) together with (7) and (10) are the specification of flow functions; that are $D(\theta)$ and $\Psi(\theta)$ versus of the dependent variable .

For $D(\theta)$, used is made of an exponential type function of the form $D(\theta) = A \exp (B\theta)$ in which A and B are positive constants to be calibrated by curve fitting of the function $D(\theta)$ to the experimental data.

Since this calibration was not done in this study, these constants were chosen in the litterature according the soil type studied. For $\Psi(\theta)$, the selected function is of the form $Cx\,\theta_a^D$ where as above C and D are constants but negative and choosen in the litterature according to the soil type studied. On the other hand, the model neglects the effect of sequential drying and wetting that could occur in layered soils. This

assumption implies the uniqueness of the relationship between $\Psi(\theta)$ and $D(\theta)$ functions versus of θ.

SOLUTION TECHNIQUES

The set of non linear partial differential equations (eqs. 6, 7, 10) are solved numerically with a finite difference scheme

To perform these numerical solutions, a grid is applied over the entire (Z,T) plan (with z, the vertical coordonate, positive downward and t the time) and a discretisation is operated along each of these two Z and t axis. This is done at first in the computation procedure (Fig. 1).

Discretisations are governed by the numerical stability; i.e. the resolution in space and in time ...

Along Z axis, the soil profile length (L) is divided into a constant number of sublayers (N) (N arbitrarily choosen, here N=20) with constant thickness (X=L/N) from the soil surface (N=0) to the bottom of the profile (N=20)).

t, the time subinterval is variable, and depends as seen above on the stability of the process.

Next the equations (6,7, and 10) are then written in term of the Crank-Nicholson implicite scheme (Fig. 1) with ($i\Delta X, j\Delta t$) replaced by (i,j) for writting simplifications.

The second step in the computational procedure concerns with the calculation of $\theta_{i+1/2,\ j+1/2}$; that is volumetric water content at half space level and half time level. This is a critical part since $D(\theta)$ and $\Psi(\theta)$ are allocated at these points, leading one to write : $D(\theta_{i+1/2,\ j+1/2})$ and $\Psi(\theta_{i+1/2,\ j+1/2})$.

MODEL DISCRETISATION

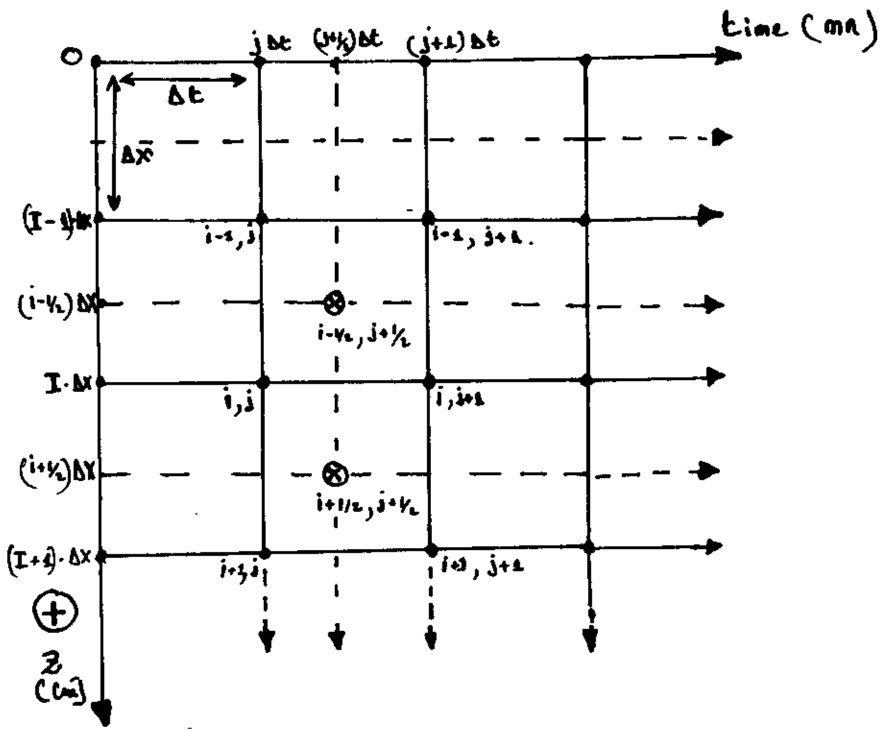


Figure I

θ (i.Δx,j.Δt) $\equiv$ θ (i,j) ; are the volumetric water conten at the time j.Δt and at the level i.Δx in the soil profile. The known values in the figure are:

θ (i-I,j) ; θ (i,j) ; and θ (i+I,j)

The unknown are:

θ (i-I,j+I) ; θ (i,j+I) ; θ (i+I,j+I) ;

θ (i-$\frac{1}{2}$,j+$\frac{1}{2}$) and θ (i+$\frac{1}{2}$,j+$\frac{1}{2}$)

Δx is the space subinterval, Δt the time subinterval and Z, the vertical coordonate,positive downward.

$\theta_{i\pm 1/2\,,\,j}$ are first calculated as an arithmetic average of values at nodes (i-1,j), (i,j) and (i+1,j) as follows :

$$\theta_{i+1/2\,,\,j} = (\theta_{i,j} + \theta_{i+1,j}) / 2. \tag{11a}$$

$$\theta_{i-1/2\,,\,j} = (\theta_{i,j} + \theta_{i-1,j}) / 2. \tag{11b}$$

Next, we make use of the backward Taylor series projection (Rosenberg, 1969) to estimate $\theta_{i\pm 1/2\,,\,j}$ at the half time level j + 1/2 .

The attractive part of the model is the way surface boundary condition is solved, preserving its non linearity and preventing oscillations in θ_s to occur.

For this, the surface boundary condition (eq.7) is represented at any time by

$$\begin{aligned} V_s &= - D(\theta_{s,j}) \cdot \frac{\theta_{s-1,j} + \theta_{s+1,j}}{2 \cdot \Delta x} \\ &= - \rho_a c_{p_a} \cdot (RH \cdot e_s(T_s) - e_a) / \gamma r_a \end{aligned} \tag{12}$$

where $\theta_{s-1,j}$ and $\theta_{s+1,j}$ and $\theta_{s,j}$ respectively designate the volumetric soil moisture content at a ficticious node immediately above the surface, at a node immediately under the surface and, at the soil surface.

This equation is solved, together with eq. 6 written in the Crank-Nicholson terms, for $\theta_{s,j+1}$ by a regular falsi iterative method. The convergence criterion (CR) in this numerical integration is

$$CR = \left| \theta^{(k+1)}_{s,j+1} - \theta^{(k)}_{s,j+1} \right| \leq 10^{-4} \tag{13}$$

where (K) is the number of iterations.

This third part in the computational procedure, is done at each jΔt loop, allowing the solution for $\theta_{i,j+1}$ of the Crank-Nicholson transformed equation (eq. 6) to be obtained. The method used for this new integration is the Thomas' method, used for numerical solution of partial differential equation.

Actual evaporation is calculated in the fourth part of our (CP) (computational procedure) from the integration of the continuity equation :

$$(\partial V/\partial z)_t = -(\partial\theta/\partial t)_z \qquad (14)$$

which can also be written in its integration form as follow :

$$V(z) = -\int_{z=L}^{z} (\partial\theta/\partial t)\,.\,dz\,. \qquad (15)$$

providing the flux values at each level Z for each time integration and thus the flux value at the surface Z=0.

The integration is next achieved by a trapezoidal rule and actual evaporation at each time step given by $V(Z=0)$.

MODEL VARIABLES

Data used for the simulation run are :

- for the initial volumetric soil moisture profile, measurements made in the weighted lysimeter of the "Génie Rural Laboratory", UCL, Belgium. Its characteristics : 1.60 cm depth, 1 m^2 surface ; soil type : loam soil; vegetation covered : short homogeneous grass. - measurements were made with a neutron probe during August 1984 at time interval of one week; model run for one week, starting with August 1st measurements as initial data.

- for climatic data

(i) global radiation is provided by IRM-Uccle (Brussels) located at 30 km from the lysimeter experimental site. Furthermore, only daily data are available. Hourly values needed in the model are deduced by simulating its diurnal distribution with a sine function between sun rise and sun set. The day length, sunrise time and sunset time are calculated from astronomical formulae.

(ii) air temperature relative humidity of the air and wind speed are directly measured (or indirectly derived from made measurements) at the meteorological station of the ("Institut d'Astronomie et de Géophysique Georges Lemaître, UCL"), station located near the lysimeter experimental site. Heat flow in the soil is assumed proportional to global radiation, with proportionality constant equal to 0.1 (De Bruin and Holtslag, 1983). Surface temperature was assumed to be air temperature. The A,B,C and D parameters respectively used in $D(\theta)$ and $\Psi(\theta)$ functions are from Gardner (1960) and Gardner and Miller (1962). These are ($1.04\ 10^{-3}$ and 18.5) for A and B and (-1.2837 and -3.1365) for C and D respectively. These parameters are given for a sandy loam soil.

RESULTS

7-days simulation of the volumetric soil moisture profile and actual evaporation under natural fluctuating evaporativities are used to test the performance of the model. Some of the results are shown in fig. 2,3,4 and 5.

Figures 2 and 3 show the moisture profile at different time (indicated in the low left corner of the figure) since the start of simulation. The general expected feature is observed; that is the continuously humidity transfert from hight level to the evaporation site. However two striking phenomena are also observed :

(i) a greater variation of water content near the surface than at other depths. This is probably due to the evaporativity fluctuations.

(ii) the tendency of soil surface moisture to increase during night time period after a drying during day-time period. The explanation of this feature is not easy while many authors have also observed it. (Hillel, 1978, Hansen, 1979). Following Hillel 1978, "the soil dries down in day time only to resorb the soil moisture content accumulated during night time pauses evaporativity"; fluctuations of these day-night soil surface wetness are of the order of 1% although Hillel (1978) found from measurements made in Israel (Gilat) a much higher values (3%).

Figure 4 shows the day to day 24h fluctuations of actual evaporation simulated by the model as well as the potential evaporation calculated using equation (1).

Figure 5 shows the daily up to 7 days corresponding to the 7 days simulations of actual potential and equilibrium evaporation. It is also represented here, the daily trends of potential evaporation calculated by an analog formula to (Eq.1) but with a daily total global radiation.

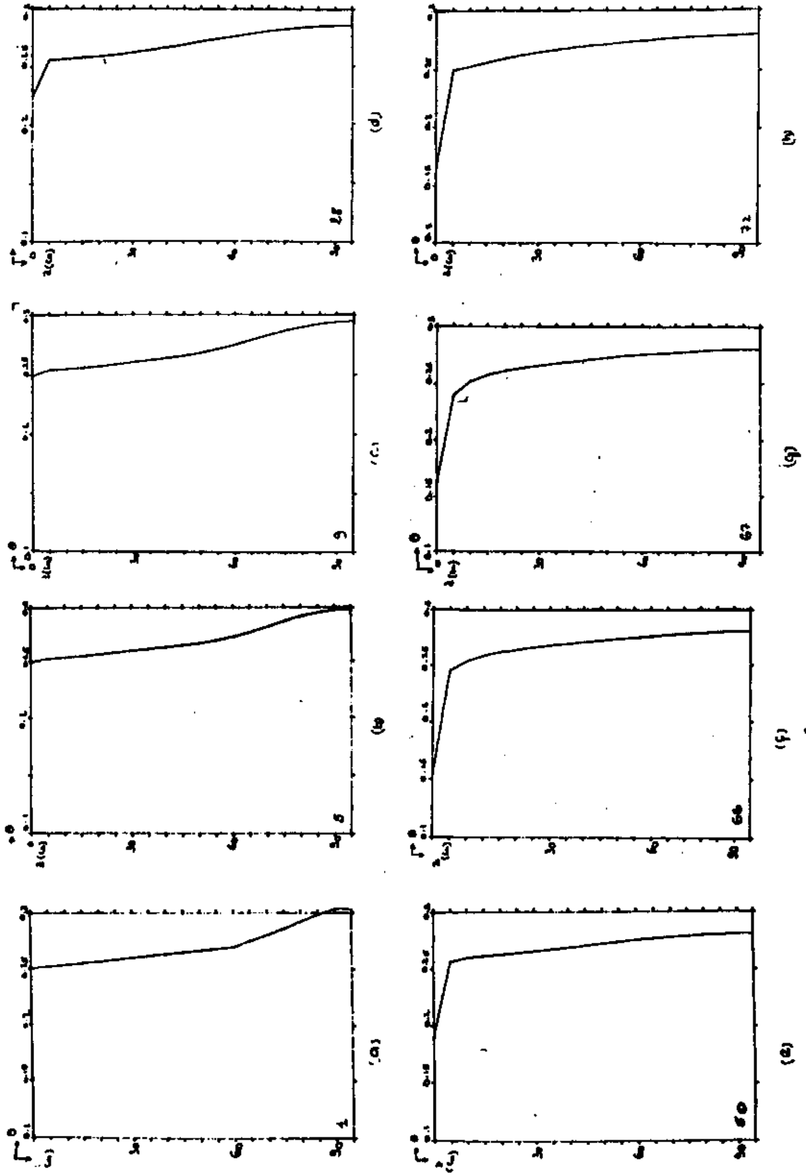


Figure 2

Soil moisture profile at different time since the starting of the simulation.

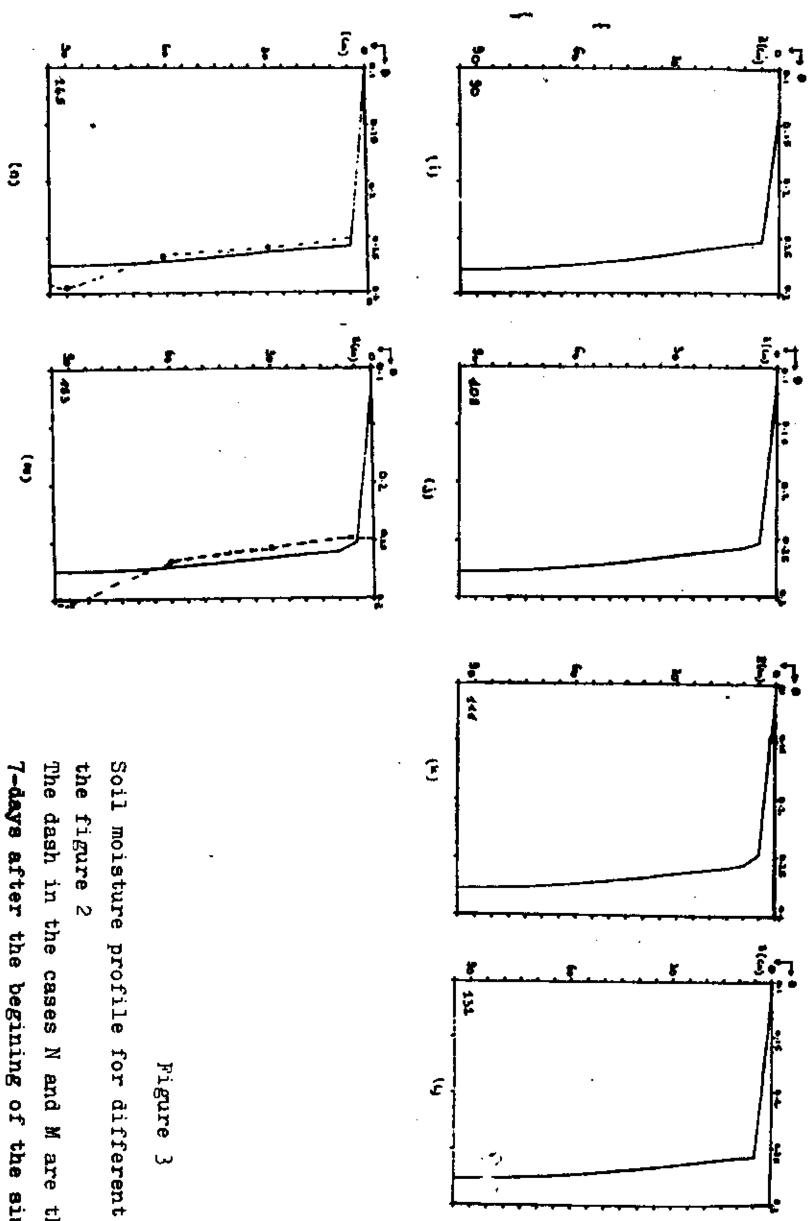
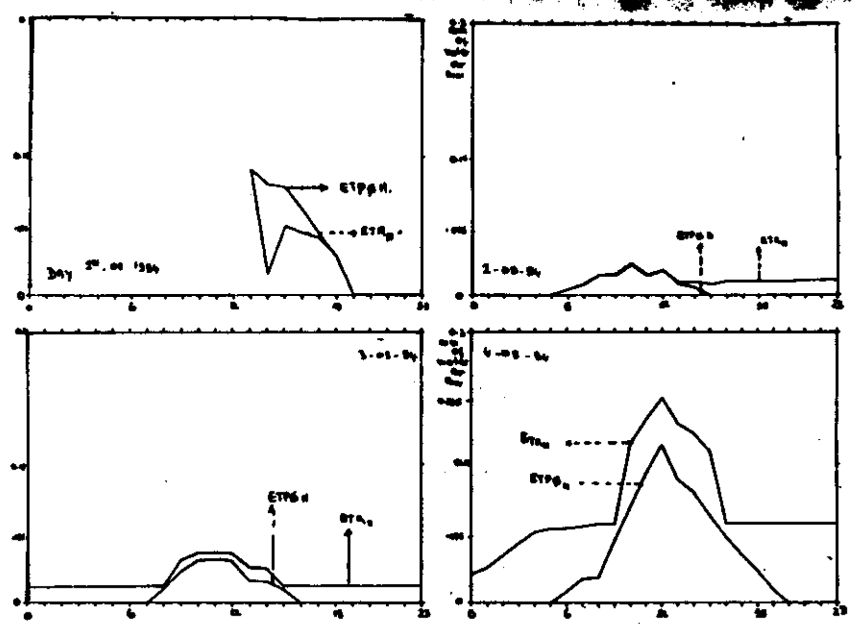


Figure 3

Soil moisture profile for different time as in the figure 2

The dash in the cases N and M are the measurements **7-days after the begining of the simulation and the initial soil moisture content respectively.**



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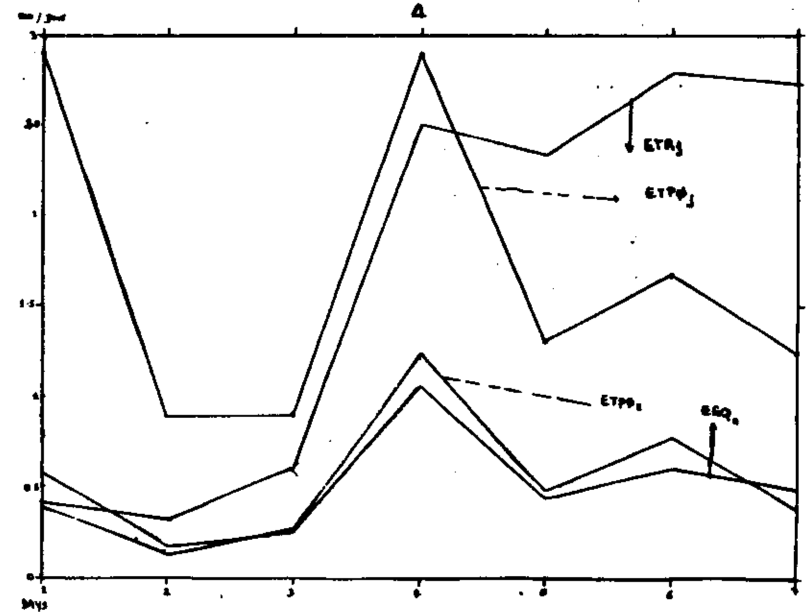

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Figure 4 and 5

daily and weekly courses of potential (ETP), actual (ETR), and equilibrum (EEQ) evaporation. The index h and j are used respectively for hourly and daily average.

It is observed in fig. 4 that actual evaporation follow the course of the potential one only during the day time period. During the night, potential evaporation is assumed to be zero. Soil evaporation reveals through this period a continuously water transfer to the atmosphere with a constant regime that, changes from day to day according to the deficit of saturation of the air and the wind speed. This night time evaporation can be used to explain why potential evaporation calculated with eq. 1 is always lesser that soil evaporation. Compared to the alternate daily total evaporation, the actual one follows its course until day 4, whenever more deviation begin. This discrepancy results from the net diminution in global radiation on day 5 which have drawn down the air evaporativity.

CONCLUSION

The present simulation model enables estimation of water flow and evaporation to be made. However, because the model does not take into account factors like soil water vapour transfer, gravity flow and hysteresis, inaccurate values may be obtained in soil layers where one or more of these factors have important effects. To illustrate this point a soil profile presented in figure 3M shows that the model under-estimated θ at greater depth. In the middle layers the estimates were much closer to measured values. Other possible sources of underestimation at greater depth could be incorrect choice of characteristic functions for this soil type and invalidation of the constraint of zero flux at this depth.

In conclusion, the model generates the most general features of water movement in soil and actual evaporation at different time scales (hour, day, week). Nonetheless, to be more accurate it needs suitable transfer functions($D(\theta)$, $\Psi(\theta)$) and an account of soil water vapour flow whose non linear numerical equations should be solved simultaneously with those used to describe the liquid flow; from which the model was derived (Philips, 1967). These are challenges in the future version of the model.

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