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EVAPORATION FROM LAYERED SOILS IN THE PRESENCE OF A WATER TABLE

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PREFACE

A steady state flow problem of interest and importance is the upward movement of water from a water table and subsequent evaporation at the soil surface. This information is desirable when estimating water loss from soils by evaporation and estimating the amount of ground water available to plants due to the upward movement of water from a water table. Soils may also become saline due to the upward movement of saline ground water and its subsequent evaporation at the soil surface. To minimize the rate of salt accumulation and thus reduce the salinity hazard, attempts are usually made to lower the water table by pumping or by installation of drains. In order to determine what depth to water table should be maintained, the relation between depth to water table, soil properties and evaporation rate must be known.

This report entitled "Evaporation from Layered Soils in the Presence of a Water Table" is a part of the research activities of 'Ground Water Assessment' division of the Institute. The purpose of this study is to estimate the steady state evaporation from layered soils with a high water table. The study has been carried out by Mr. C. P. Kumar, Scientist 'C' under the guidance of Dr. G. C. Mishra, Scientist 'F'.

(S. M. SETH)

Director

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ABSTRACT

Evaporation of water from soil surface causes loss of water and is also responsible for salinizing the top layers of the soil. The danger of soil salinization becomes more acute in regions where a high ground water table exists. In order to minimize water losses as well as reduce the rate of soil salinization, one has to evaluate the effect of the depth of soil layers overlying the water table.

Evaporation from shallow water table through a homogeneous soil profile has been studied theoretically and experimentally by many workers. However, uniform soil profiles rarely occur in nature. It is more common to find the soils having well-defined layers differing from each other either in texture or in structure. Therefore, it becomes necessary to determine the effect of layered soils on evaporation from a shallow water table.

The purpose of this study is to estimate the steady state evaporation rates from layered soils in the presence of high water table under isothermal conditions. A finite difference numerical scheme based upon the one-dimensional Richards equation has been employed to estimate the evaporation rates from a two-layered soil profile overlying a shallow water table for appropriate initial and boundary conditions. The method takes into account the relevant atmospheric factors and soil moisture characteristics of the two layers. The effects of sequence and thickness of the soil layers and water table depth on the evaporation rates have been examined.

1.0 INTRODUCTION

Evaporation in the field can take place from plant canopies, from the soil surface, or from a free-water surface. Evaporation from plants, called transpiration, is the principal mechanism of soil-water transfer to the atmosphere when the soil surface is covered with vegetation. When the surface is at least partly bare, evaporation can take place from the soil as well as from plants. These two interdependent processes are commonly lumped together and treated as if they were a single process, called evapotranspiration. In the absence of vegetation, and when the soil surface is subject to radiation and wind effects, evaporation occurs directly and entirely from the soil. It is a process which, if uncontrolled, can involve very considerable losses of water in both irrigated and unirrigated agriculture.

Evaporation of soil water involves not only loss of water but also the danger of soil salinization. This danger is felt most in regions where irrigation water is scarce and possibly brackish and where annual rainfall is low, as well as in regions with a high ground water table. Where a ground water table occurs close to the surface, continual flow may take place from the saturated zone beneath through the unsaturated soil to the surface. If this flow is more or less steady, continued evaporation can occur without materially changing the soil moisture content (though cumulative salinization may take place at the surface). In the absence of shallow ground water, on the other hand, the loss of water at the surface and the resulting upward flow of water in the profile will necessarily be a transient state process causing the soil to dry. A proper formulation of an evaporation process should account for spatial and temporal variability, as well as for interactions with the above ground and below ground environment.

It is desirable to estimate the evaporation rates from bare land surfaces and to predict the variation of these rates with meteorological conditions or with man-imposed changes in the water table level. This estimate might be rather important in certain regions during the appraisal of ground water availability. For such purposes, it is often both permissible and useful to assume steady state of the hydraulic gradient driven upward flux of water and to neglect certain effects of soil temperature and of solute accumulations. The basic approaches required for the development of this method can be found in the literature. Convenient equations were suggested for describing hydraulic conductivity, the most relevant soil parameter, and from it methods were developed for evaluating soil-limited evaporation in cases of high water table. It was also shown how the effects of the soil factors on bare soil evaporation interact with the effects of the atmospheric parameters on bare soil evaporation. However, all the studies concerned themselves with homogeneous soils and mainly with cases involving liquid transfer.

In homogeneous profiles, the soil moisture characteristics are same everywhere. However, soil is frequently stratified near the surface, containing layers with markedly different water retention and water conducting properties. The mathematical description of water transport through unsaturated layered soil is very complex because of subtle effects that can occur at the interface between layers.

The layer of least permeability often has a dominant influence on the transport through the system. For example, even though steep hydraulic head gradients are often present, flow through a series of layers of unsaturated soil can be nearly zero under conditions where large and nearly empty pores with small hydraulic conductivities are encountered. Such a condition exists where a wetting front moving through homogeneous soil encounters a layer of coarse sand or gravel. The hydraulic head of the soil just above the wetting front may be of the order of -100 cm of water and that in the dry sand below the front may be as low as -10^3 or -10^4 cm. Thus the potential gradient at the interface will be large. Despite this, the flow nearly drops to zero as the front reaches the coarse sand layer because there is very little water in fine pores in the sand and the large pores can not fill at the low matric potentials present in the upper region. Thus, the cross section for liquid flow is very small. Before any appreciable flow can occur, the hydraulic head in the upper layers of the finer textured soil must rise to a value near zero, at which time some of the pores or channels in the sand will begin to fill and the hydraulic conductivity of the lower layer will rise.

Coarse materials such as straw or other organic matter or holes in the soil created by burrowing insects and animals restrict, rather than aid, flow as long as the hydraulic potentials surrounding them are too low for them to fill with water. For this reason, dry soil often persists through the wet season beneath straw turned under by ploughing. Only when large pores and channels connect with the surface where free water can get to them or are beneath the water table, such channels contribute appreciably to liquid flow.

Fine pores in hard pans and clay pans also seriously restrict flow. Such materials become wet rapidly for short distances when first contacted by water because of the high absorptive capacity of fine pores. However, as the distance through which water must move in fine pores increases, the rate of flow decreases, in tight clay it becomes extremely slow. Flow in such materials is often so slow that water table build up above them.

Water retention following wetting or redistribution of water is greatly affected by stratification. Clay pans and hard pans often create serious waterlogging because retention is so pronounced above such layers. Coarse layers act much the same as a check valve. Water tables can not be maintained above a coarse layer. However, since a coarse layer restricts flow at relatively high hydraulic potentials, retention of water above such layers often is appreciably more than it would be in the absence of such a layer.

The steady state flow equation describing upward movement of water from a water table and evaporation at the surface of a homogeneous soil has been solved for a number of different analytical expressions. However, it is of particular importance to determine how evaporation in the layered case may differ from the homogeneous case.

The actual evaporation rate is governed by the atmospheric conditions, thickness and transmitting properties of the soil layers and the water table depth. While the maximum possible (potential) rate of evaporation from a given soil depends only on atmospheric conditions, the actual flux across the soil surface is limited by the ability of the porous medium to transmit water from below.

In the present study, an attempt has been made to examine the effect of layered soil profiles on steady state evaporation rates from a shallow water table under isothermal conditions by using a finite difference numerical scheme for solution of the one-dimensional Richards equation. The evaporation rates are shown to be related to the water table depth and sequence and thickness of the soil layers.

2.0 REVIEW

The steady state upward flow of water from a water table through the soil profile to an evaporation zone at the soil surface was first studied by Moore (1939). Theoretical solutions of the flow equation for this process were given by several workers including Gardner (1958), Anat et al. (1965) and Ripple et al. (1972).

The equation describing steady upward flow is

$$q = K(h)\left(\frac{dh}{dz} - 1\right) \qquad \ldots(2.1)$$

or

$$q = D(\theta)\frac{d\theta}{dz} - K(\theta) \qquad \ldots(2.2)$$

where q is flux (equal to the evaporation rate under steady state conditions), h suction head (soil water pressure), K hydraulic conductivity of the soil, D soil water diffusivity, θ volumetric water content and z height above the water table. The equation shows that flow stops (q = 0) when dh/dz = 1. Another form of equation (2.1) is

$$\frac{q}{K(h)} + 1 = \frac{dh}{dz} \qquad \ldots(2.3)$$

Integration should give the relation between depth and suction or wetness :

$$z = \int \frac{dh}{1 + \frac{q}{K(h)}} = \int \frac{K(n)}{K(h)+q}\, dh \qquad \ldots(2.4)$$

or

$$z = \int \frac{D(\theta)}{K(\theta) + q} d\theta \qquad \ldots(2.5)$$

In order to perform the integration in equation (2.4), we must know the functional relation between K and h, i.e. K(h). Similarly, the functions $D(\theta)$ and $K(\theta)$ must be known if equation (2.5) is to be integrated. An empirical equation for K(h), given by Gardner (1958), is

$$K(h) = \frac{a}{h^n + b} \qquad ..(2.6)$$

where the parameters a, b and n are constants which must be determined for each soil. Accordingly, equation (2.1) becomes

$$e = q = \frac{a}{h^n + b} \left(\frac{dh}{dz} - 1\right) \qquad \ldots(2.7)$$

where e is the evaporation rate.

With equation (2.6), equation (2:4) can be used to obtain suction distributions with height for different fluxes, as well as fluxes for different surface suction values. The steady rate of capillary rise and evaporation therefore depend on the depth of the water table and on the suction at the soil surface This suction is dictated largely by the external conditions, since the greater the atmospheric evaporativity, the greater the suction at the soil surface upon which the atmosphere is acting. However, increasing the suction at the soil surface, even to the extent of making it infinite, can increase the flux through the soil only upto an asymptotic maximal rate which depends on the depth of the water table. Even the driest and most evaporative atmosphere can not steadily extract water from the surface any faster than the soil profile can transmit from the water table to that surface. The fact that the soil profile can limit the rate of evaporation, is a remarkable and useful feature of the unsaturated flow system. The maximal transmitting ability of the profile depends on the hydraulic conductivity of the soil in relation to the suction.

Disregarding the constant b of equation (2.6), Gardner (1958) obtained the function

$$E_{lim} = \frac{Aa}{d^n} \qquad \ldots(2.8)$$

where d is the depth of the water table below the soil surface, a and n are constants from equation (2.6), A is a constant which depends on n and E_{lim} is the limiting (maximal) rate at which the soil can transmit water from the water table to the evaporation zone at the surface.

The actual steady evaporation rate is determined either by the external evaporativity or by the water transmitting properties of the soil, depending on which of the two is lower, and therefore limiting. Where the water table is near the surface, the suction at the soil surface is low and the evaporation rate is determined by external conditions. However, as the water table becomes deeper and the suction at the soil surface increases, the evaporation rate approaches a limiting value regardless of how high external evaporativity may be.

Equation (2.8) suggests that the maximal evaporation rate decreases with water table depth more steeply in coarse-textured soils (in which n is greater) than in fine-textured soils. Nevertheless, a sandy loam soil can still evaporate water at an appreciable rate even when the water table is as deep as 180 cm.

The subsequent contributions of a number of workers have generally accorded with the above theory. Anat et al. (1965) developed a modified set of equations employing dimensionless variables. Their theory also leads to a maximal evaporation rate e_{max} varying inversely with water table depth d to the power of n:

$$e_{max} = \frac{1 + \frac{1.886}{(n^2 + 1)}}{d^n} \qquad \ldots(2.9)$$

Subsequently, Ripple et al. (1972) derived the following relation for soil-limited evaporation.

$$E_{\infty} \simeq K_s \left[\frac{|h_{1/2}|}{L} \cdot \frac{\pi}{n \sin \frac{\pi}{n}} \right]^n \qquad \ldots(2.10)$$

where,

E_{∞} = soil-limited rate of evaporation from the soil (cm/day);

K_s = hydraulic conductivity of water-saturated soil (cm/day);

$h_{1/2}$ = a constant soil coefficient representing h at $K = \frac{1}{2} K_s$ (cm of water);

L = total distance between the water table and soil surface (cm); and

n = a soil coefficient (which usually ranges from 2 for clay to 5 for sands) in K-h relationship of equation (2.6).

Equation (2.10) is similar to the formulas for E_{lim} given without derivation by Gardner (1958) for n = $\frac{3}{2}$, 2, 3, 4 and yields identical numerical coefficients.

The equations given above deal with the movement of soil water in the liquid phase only. Under isothermal conditions, if the water content is above the wilting point, any vapour pressure gradient present in the soil will be sufficiently small so that movement in the vapour phase can be neglected. In the case of evaporation from a soil, however, if the potential evaporation rate is appreciably greater than the rate at which water can be transmitted from the water table to the soil surface, the soil near the surface will dry out. A vapour pressure gradient will be set up near the soil surface, causing movement of water in the vapour phase and thus allowing the soil to dry below the surface. Under these conditions, the movement of water in the vapour phase must be taken into account.

The effect of a surface mulch upon steady-state evaporation can be treated in a simple manner. For this purpose, a mulch is defined as a medium which transports water in the vapour phase only. The steady-state rate of evaporation from a soil with a surface mulch should be inversely proportional to the thickness of the mulch (Gardner, 1958).

A theoretical analysis of steady evaporation from a two layered soil profile was carried out by Willis (1960), with the following assumptions :

(a) the steady flow through the layered profile is governed only by the transmission properties of the profile (external evaporativity taken to be infinite);

(b) matric suction is continuous at and through the interlayer boundary, though wetness and conductivity may be discontinuous (i.e., change abruptly);

(c) the same empirical K(h) function given by equation (2.6) holds for both layers but the values of parameters a, b and n are different; and

(d) each soil layer is internally homogeneous.

With these assumptions, equation (2.4) leads to

$$\int_0^L dz + \int_L^{d+L} dz = \int_{h_0}^{h_L} \frac{dh}{1+e/K_1(h)} + \int_{h_L}^{h_{L+d}} \frac{dh}{1+e/K_2(h)}$$

...(2.11)

where L and d are the thicknesses of bottom and top layers respectively. The integral in this equation relates water table depth (L+d) to the suction at the soil surface for any given evaporation rate. By assuming that the suction at the soil surface is infinite, one can calculate the limiting (maximal) evaporation rate for any given water table depth and profile layering sequence. Willis developed a graphical method for obtaining the necessary solution.

All of the above treatments apply to cases in which soil properties are the sole factor determining the evaporation rate. A more realistic approach should include cases in which meteorological conditions can also play a role. A more flexible treatment of steady state evaporation from multilayer profiles might also be based on numerical, rather than analytical or graphical, methods of solution. Such an approach was developed by Ripple et al. (1972). Their procedure makes it possible to estimate the steady state evaporation from bare soils (including layered ones) with a high water table. The required field data include soil moisture characteristic curves, water table depth, standard elevation records of air temperature, air humidity and wind velocity. The theory takes into account both the relevant atmospheric factors and the soil's capability to transmit water in liquid and vapour forms. The possible effects of thermal transfer (except in the vapour phase) and of salt accumulation were neglected.

3.0 PROBLEM DEFINITION

The objective of the present study is to determine the evaporation from shallow water tables through a two-layered soil profile under isothermal conditions on the basis of solutions of the water flow equation. The steady state upward water flow from a shallow water table through the soil toward its surface is described by the nonlinear Richards equation. A numerical model (finite difference scheme) is used for solving the partial differential equation describing one-dimensional water flow through the unsaturated porous medium. Steady state moisture profiles are obtained for the given initial and boundary conditions and the steady state evaporation rates are estimated by using Darcy's law.

The evaporation rate can be limited either by the external evaporative conditions or by the maximal rate at which the soil can transmit water to its surface. If the water table is near the soil surface, the external conditions will govern the evaporation rate, whereas if the water table becomes deeper, the evaporation rate approaches a limiting value which is determined by the soil profile capabilities of water transmission regardless of the external conditions. The effect of water table depth on the steady evaporation rate is therefore examined by varying depth of the water table. The effects of sequence of the two soil layers and their thicknesses on the evaporation rates are also determined.

4.0 METHODOLOGY

Most of the processes involving soil water flow in the field and in the rooting zone of most plant habitats, occur while the soil is in unsaturated condition. Unsaturated flow processes are in general complicated and difficult to describe quantitatively, since they often entail changes in the state and content of soil water during flow. Such changes involve complex relations among the variable water content, suction and conductivity, which may be affected by hysteresis. The formulation and solution of unsaturated flow problems very often require the use of indirect methods of analysis, based on approximations or numerical techniques.

4.1 General Equation of Unsaturated Flow

A proper physical description of water flow in the soil requires that three parameters be specified : flux, hydraulic gradient and conductivity. Knowledge of any two of these allows the calculation of the third, according to Darcy's law. This law states that the flux equals the product of conductivity by the hydraulic gradient. Darcy's law has been found to apply for unsaturated as well as saturated soils but the pressure gradient at unsaturation becomes a suction gradient and the hydraulic conductivity is no longer constant but a function of water content or suction. Since the conductivity depends on the number, sizes and shapes of the conducting pores, its value is greatest when the soil is saturated and decreases steeply when the soil water suction increases and the soil loses moisture. Darcy's law suffices to describe water flow under steady state conditions but must be combined with the continuity equation to describe unsteady (transient state) flow. According to Darcy's law, for one-dimensional vertical flow, the volumetric flux q $(cm^3/cm^2/h)$ can be written as

$$q = -K \frac{\partial}{\partial z} (h - z) \qquad (cm/h)$$

or

$$q = -K \left(\frac{\partial h}{\partial z} - 1\right) \qquad (cm/h) \qquad \ldots(4.1)$$

where K is the hydraulic conductivity (cm/h), h is the soil water pressure head (relative to the atmosphere) expressed in cm of water and z is the gravitational head (cm) considered positive in downward direction.

In order to get a complete mathematical description for unsaturated flow, we apply the continuity principle (law of conservation of matter)

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} \qquad (/h) \qquad \ldots(4.2)$$

where θ is soil moisture content expressed in cm^3/cm^3 and t is time in hours.

Substitution of equation (4.1) into equation (4.2) yields the partial differential equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[K \left(\frac{\partial h}{\partial z} - 1\right) \right] \qquad \ldots(4.3)$$

Equation (4.3) is a second order, parabolic type of partial differential equation (known as Richards equation) which is non-linear because of the dependency of K and h on θ (linearity means that the coefficients in a differential equation are functions of the independent variables z and t only). To avoid the problem of the two dependent variables θ and h, the derivative of θ with respect to h can be introduced, which is known as the specific water capacity C

$$C = \frac{d\theta}{dh} \quad ,(/cm) \qquad ...(4.4)$$

In equation (4.4), a normal instead of a partial derivative notation is used, because h is considered here as a single value function of θ (no hysteresis). Writing

$$\frac{\partial \theta}{\partial t} = \frac{d\theta}{dh} \cdot \frac{\partial h}{\partial t} \qquad ...(4.5)$$

and substituting equation (4.4) into equation (4.3) yields

$$C(h)\frac{\partial h}{\partial t} = \frac{\partial}{\partial z}\left[K(h) \left(\frac{\partial h}{\partial z} - 1\right) \right] \qquad ...(4.6)$$

In equation (4.6), the coefficients C and K are functions of the dependent variable h, but not functions of the derivatives $\partial h/\partial t$ and $\partial h/\partial z$. Written in this form, equation (4.6) provides the basis for predicting soil water movement in layered soils of which each layer may have different physical properties.

4.2 Initial and Boundary Conditions

To obtain a solution for the one-dimensional vertical flow, equation (4.6) must be supplemented by appropriate initial and boundary conditions.

As initial condition (at t=0), the pressure head is specified as a function of the depth z

$$h\ (z,\ t=0) = h_0 \qquad \text{...(4.7)}$$

If hysteresis is not considered, this condition is equivalent to

$$\theta\ (z,\ t=0) = \theta_0 \qquad \text{...(4.8)}$$

One can easily obtain the value of h (and vice versa) from the expression, $h = f(\theta)$.

To describe the boundary conditions, one can distinguish between three types :

(a) Dirichlet condition : specification of the dependent variable, the pressure head

$$\left.\begin{aligned} h\ (z=0,\ t) &= h_u \\ h\ (z=L,\ t) &= h_l \end{aligned}\right\} \qquad \text{...(4.9)}$$

These conditions are equivalent to

$$\left.\begin{aligned} \theta\ (z=0,\ t) &= \theta_u \\ \theta\ (z=L,\ t) &= \theta_l \end{aligned}\right\} \qquad \text{...(4.10)}$$

(b) Neumann condition : specification of the derivative of the pressure head. For the soil water problem, this condition means a specification of the flow through the boundaries

$$q(t) = -\ K(h)\ (\frac{\partial h}{\partial z} - 1) \qquad \text{...(4.11)}$$

(c) Mixed condition : a combination of the first two types. In particular, this can specify h at the lower boundary and q at the upper boundary.

For the present study, the initial and boundary conditions have been defined as follows.

I. Initial condition :

$$h(z,0) = h_0 \quad \text{for } z \geq 0, \quad t = 0 \qquad \ldots(4.12)$$

(approximate equilibrium moisture profile)

II. Upper boundary condition :

If the relative humidity (f) and the temperature of the air (T) as a function of time are known, and if it may be assumed that the pressure head at the soil surface is at equilibrium with the atmosphere, then h(0,t) can be derived from the thermodynamic relation (Edlefson and Anderson, 1943) .

$$h(0,t) = \frac{RT(t)}{Mg} \ln [f(t)] \qquad \ldots(4.13)$$

where R is the universal gas constant (8.314×10^7 erg/mole/K), T is the absolute temperature (K), g is acceleration due to gravity (980.665 cm/s^2), M is the molecular weight of water (18 gm/mole), f is the relative humidity of the air (fraction) and h is in bars. Knowing h(0,t), $\theta(0,t)$ can be derived from the soil water retention curve.

III. Lower boundary condition :

The phreatic surface acts as lower boundary of the system in case of evaporation from shallow water table.. The lower boundary condition has therefore been set as

$$h(z=L,\ t) = 0 \qquad \ldots(4.14)$$

where L is the depth of the ground water table.

4.3 Soil Moisture Characteristics

For the present study, functional relations, as reported by Haverkamp et al. (1977), for characterizing the hydraulic properties of two soils, sand and yolo light clay, were used. They compared six models, employing different ways of discretization of the non-linear infiltration equation in terms of execution time, accuracy and programming considerations. The models were tested by comparing water content profiles calculated at given times by each of the model with results obtained from infiltration experiments in a sand column. All models yielded excellent agreement with water content profiles measured at various times.

The infiltration experiments were done in the laboratory using a plexiglass column, 93.5 cm long and 6 cm inside diameter, uniformly packed with sand to a bulk density of 1.66 gm/cm^3. The column was equipped with tensiometers at depths of 7, 22, 37, 52, 67 and 82 cm below the soil surface. Each tensiometer had its own pressure transducer. The changes in water content at different depths were obtained by gamma ray attenuation using a source of Americium-241. A constant water pressure (θ = 0.10) was maintained at the lower end of the column, a constant flux (13.69 cm/h) was imposed at the soil surface (z = 0) and initial condition as θ = 0.10 throughout the depth. The hydraulic conductivity and water content relationship of the soil was obtained by analysis of the water content and water pressure profiles during transient flow. The soil water pressure and water content relationship was obtained at each tensiometer depth by correlating tensiometer readings and water content measurements during the experiments.

The following analytical expressions, obtained by a least square fit through all data points were chosen for characterizing the soil (sand) :

$$K = K_s \frac{A}{A + |h|^{\beta_1}} ; \qquad \ldots(4.15)$$

$K_s = 34$ cm/h,
$A = 1.175 \times 10^6$,
$\beta_1 = 4.74$.

and

$$\theta = \frac{\alpha (\theta_s - \theta_r)}{\alpha + |h|^{\beta_2}} + \theta_r ; \qquad \ldots(4.16)$$

$\theta_s = 0.287$,
$\theta_r = 0.075$,
$\alpha = 1.611 \times 10^6$,
$\beta_2 = 3.96$.

where subscript s refers to saturation, i.e. the value of θ for which $h = 0$, and the subscript r to residual water content.

The soil characteristics of yolo light clay are given in equations (4.17) and (4.18), using the same representation as in the previous case. The data points for $\theta(h)$ were taken from Philip (1969, pp. 221), values for $K(\theta)$ were presented by Philip (1957, pp. 353), points for $K(h)$ were determined from $\theta(h)$ and $K(\theta)$, as reported by Haverkamp et al. (1977).

$$K = K_s \frac{A}{A + |h|^{\beta_1}} ; \qquad \ldots(4.17)$$

$K_s = 4.428 \times 10^{-2}$ cm/h,
$A = 124.6$,
$\beta_1 = 1.77$.

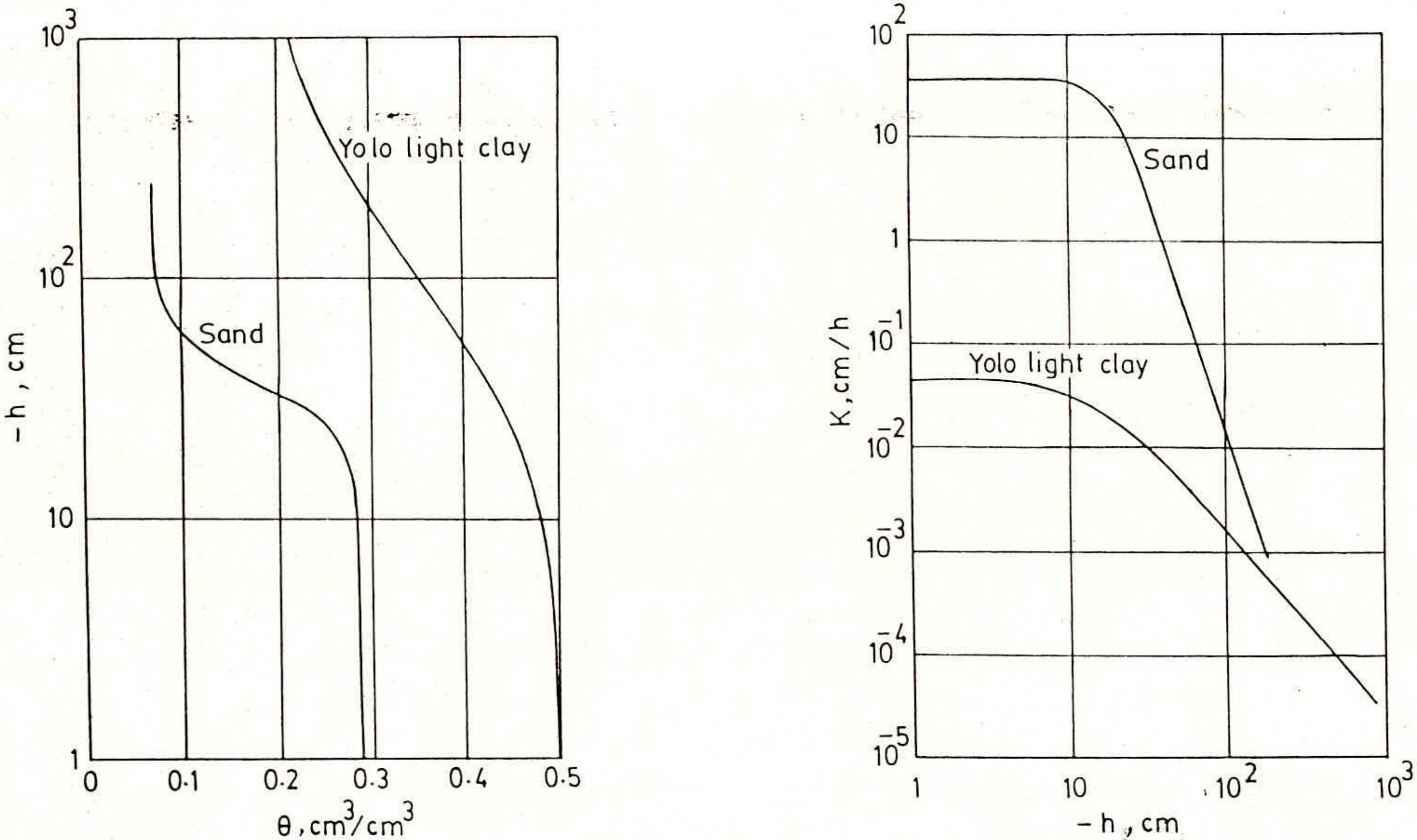

FIG.1 : RELATIONSHIPS BETWEEN THE SOIL WATER PRESSURE h, THE WATER CONTENT θ AND THE HYDRAULIC CONDUCTIVITY K FOR THE TWO SOILS USED IN THE STUDY

and
$$\theta = \frac{\alpha (\theta_s - \theta_r)}{\alpha + (\ln|h|)^{\beta_2}} + \theta_r ; \qquad \text{for } h < -1 \text{ cm} \qquad \ldots(4.18)$$

$\theta_s = 0.495$,
$\theta_r = 0.124$,
$\alpha = 739$,
$\beta_2 = 4$.

$$\theta = \theta_s \qquad \text{for } h \geq -1 \text{ cm}$$

The functional relationships represented by the above equations describe fairly well the data, as tabulated by Philip, except near h = -100 cm for the K(h) curve. Also the diffusivity values derived from equations (4.17) and (4.18) using $D(\theta) = K(\theta)/C(\theta)$ and $C(\theta) = d\theta/dh$ are somewhat different from those presented by Philip (1957, pp. 353), particularly near θ_s.

Figure 1 presents the relationships between the soil water pressure h, the water content θ and the hydraulic conductivity K for the two soils used in this study. These relations were used to estimate the evaporation rates from various combinations of layered soil profiles.

4.4 Finite Difference Approximation

Equation (4.6) is a non-linear partial differential equation (PDE) because the parameters K(h) and C(h) depend on the actual solution of h(z,t). The non-linearity of the equation causes problems in its solution. Analytical solutions are known for special cases only. The majority of practical field problems can only be solved by numerical methods. In this respect, one can use either explicit or implicit methods. Although an implicit approach is more complicated, it is preferable because of its better stability and convergence. Moreover, it permits relatively large time steps thus keeping computer costs low. For a given grid point at a given time, the values of the coefficients C(h) and K(h) can be expressed either from their values at the preceding time step (explicit linearization) or from a prediction at time (t+1/2 Δt) using a method described by Douglas and Jones, 1963 (implicit linearization).

Let us now solve equation (4.6) by a finite difference technique and appropriate initial and boundary conditions. We have

$$C \frac{\partial h}{\partial t} = \frac{\partial}{\partial z} \left[K \left(\frac{\partial h}{\partial z} - 1 \right) \right]$$

or

$$C \frac{\partial h}{\partial t} = \frac{\partial K}{\partial z} \left(\frac{\partial h}{\partial z} - 1 \right) + K \frac{\partial^2 h}{\partial z^2}$$

or

$$\frac{C}{K} \frac{\partial h}{\partial t} = \frac{\partial^2 h}{\partial z^2} + \frac{1}{K} \frac{\partial K}{\partial z} \left(\frac{\partial h}{\partial z} - 1 \right) \qquad \ldots (4.19)$$

Using implicit evaluation of the coefficients at time (t+1/2 Δt), that is values for K and C are obtained at time (t+1/2 Δt), then pressure distribution is evaluated at time (t+Δt). The partial differential equation is approximated by a finite difference equation replacing ∂t and ∂z by Δt and Δz, respectively.

Prediction (estimation of C_i^j and K_i^j)

From equation (4.19), by taking time step as Δt/2, we have

$$\frac{2C_i^j}{K_i^j} \cdot \frac{h_i^{j+1/2} - h_i^j}{\Delta t} = \frac{h_{i+1}^{j+1/2} - 2h_i^{j+1/2} + h_{i-1}^{j+1/2}}{(\Delta z)^2}$$

$$+ \frac{1}{K_i^j} \cdot \frac{K_{i+1}^j - K_{i-1}^j}{2\Delta z} \left[\frac{h_{i+1}^j - h_{i-1}^j}{2\Delta z} - 1\right]$$

where i refers to depth and j refers to time. Rearranging the terms, we get

$$- \frac{\Delta t}{(\Delta z)^2} h_{i-1}^{j+1/2} + \left[\frac{2C_i^j}{K_i^j} + \frac{2\Delta t}{(\Delta z)^2}\right] h_i^{j+1/2} - \frac{\Delta t}{(\Delta z)^2} h_{i+1}^{j+1/2}$$

$$= \frac{2C_i^j}{K_i^j} h_i^j + \frac{1}{2} \frac{K_{i+1}^j - K_{i-1}^j}{K_i^j} \frac{\Delta t}{\Delta z} \left[\frac{h_{i+1}^j - h_{i-1}^j}{2\Delta z} - 1\right] \quad \ldots(4.20)$$

Correction (estimation of h_i^j)

From equation (4.19), by taking time step as Δt, we have

$$\frac{C_i^{j+1/2}}{K_i^{j+1/2}} \cdot \frac{h_i^{j+1} - h_i^j}{\Delta t} = \frac{1}{2} \left[\frac{h_{i+1}^{j+1} - 2h_i^{j+1} + h_{i-1}^{j+1}}{(\Delta z)^2} + \frac{h_{i+1}^j - 2h_i^j + h_{i-1}^j}{(\Delta z)^2}\right]$$

$$+ \frac{1}{K_i^{j+1/2}} \cdot \frac{K_{i+1}^{j+1/2} - K_{i-1}^{j+1/2}}{2\Delta z} \left[\frac{h_{i+1}^{j+1/2} - h_{i-1}^{j+1/2}}{2\Delta z} - 1\right]$$

Rearranging the terms, we get

$$-\frac{1}{2}\frac{\Delta t}{(\Delta z)^2} h_{i-1}^{j+1} + \left[\frac{C_i^{j+1/2}}{K_i^{j+1/2}} + \frac{\Delta t}{(\Delta z)^2}\right] h_i^{j+1} - \frac{1}{2}\frac{\Delta t}{(\Delta z)^2} h_{i+1}^{j+1}$$

$$= \frac{C_i^{j+1/2}}{K_i^{j+1/2}} h_i^j + \frac{1}{2}\frac{\Delta t}{(\Delta z)^2}\left[h_{i+1}^j - 2 h_i^j + h_{i-1}^j\right]$$

$$+ \frac{1}{2}\frac{K_{i+1}^{j+1/2} - K_{i-1}^{j+1/2}}{K_i^{j+1/2}} \frac{\Delta t}{\Delta z}\left[\frac{h_{i+1}^{j+1/2} - h_{i-1}^{j+1/2}}{2\Delta z} - 1\right] \qquad \ldots(4.21)$$

When equation (4.20) or (4.21) is applied at all nodes, the result is a system of simultaneous linear algebraic equations with a tridiagonal coefficient matrix with zero elements outside the diagonals and unknown values of h. In solving this system of equations, a so-called direct method was used by applying a tridiagonal algorithm of the kind discussed by Remson et al. (1971).

4.5 Estimation of Evaporation Rates

Steady state evaporation rates from the layered profiles were estimated by applying equation (2.1) for two vertically adjacent nodal points after obtaining the equilibrium moisture profile for the given set of water table depth and thickness of the top soil layer.

$$q = K_{i+1/2}^j \left(\frac{h_{i+1}^j - h_i^j}{\Delta z} - 1\right) \qquad \ldots(4.22)$$

where, $$K^j_{i+1/2} = \sqrt{K^j_i \; K^j_{i+1}}$$

Geometric mean of K was taken following suggestions of Haverkamp and Vauclin (1979). Theoretically, we should get the same evaporation rates by considering any two vertical adjacent nodal points for the steady state condition. However, small variations may be observed due to values of K being not properly represented over the depth interval chosen. Therefore, arithmetic mean of computed evaporation rates at any time was taken by considering all nodes in succession for applying equation (4.22). Various sets of water table depths and thicknesses of the two soil layers were considered for the study.

The computer code, for discretization scheme used in the model and estimation of steady state evaporation rates from layered soils in the presence of a water table, as per the procedure described above, has been written in FORTRAN language and presented in Appendix-I and Appendix-II for the two soil layering sequences.

5.0 RESULTS

In a bare soil with a shallow water table, subject to atmospheric evaporation, steady flow can take place from the ground water source below to the evaporation sink above. When the water table is very near to the soil surface, and the soil transmits water readily, the actual evaporation rate will be limited by external evaporativity (i.e., the micrometeorological conditions). When the water table is relatively deep, the water-transmitting properties of the profile are likely to be limiting, and thus determine the evaporation rate. Capillary rise from a water table and evaporation at the soil surface entail the hazard of progressive salinization, even though this hazard is not always immediately apparent at the surface. To avoid this hazard, which is most severe in fine textured soils under irrigation, artificial ground water drainage may be necessary. The best way to conserve soil moisture against evaporation is to cause it to move as deeply as possible into the profile, by proper regulation of the irrigation regimen and by controlling the initial evaporation rate so as to allow maximal time for the post-irrigation redistribution of soil water.

The numerical model described in section 4.4 was tested by comparing water content profiles calculated at given times with results obtained from quasi-analytical solution of Philip subject to condition of a constant pressure at the soil surface ($\theta = 0.267$ cm^3/cm^3). The infiltration profiles at various times for infiltration in the sand (one of the two soils under consideration) obtained by quasi-analytical solution of Philip, were reported by Haverkamp et al. (1977). The model yielded good agreement with water content profiles at various times (Kumar and Mishra, 1991).

The present study was carried out for bare surface (i.e. no vegetation) and therefore transpiration by plants was not taken into account. The sub-surface profile was divided into uniform layers of thickness 4 cm each (depth interval, Δz) down to the water table position which was varied from 60 cm to 140 cm below the soil surface.

Since the steady state evaporation rates are estimated by considering the two vertically adjacent nodal points (equation 4.22), the size of the depth interval plays an important role. It was found that the numerical scheme is stable only when

$$\frac{\Delta t}{(\Delta z)^2} \leq 2.5 \qquad \ldots(5.1)$$

where Δt is the time step (seconds) and Δz is the depth interval (cm). Keeping in view the stability of the numerical scheme, the time step, Δt was taken as 40 seconds during the entire study period. Uniform evaporative conditions (temperature = 25°C, relative humidity = 0.75) were assumed for the study. The value of potential evaporation was obtained through Meyer's equation (for T = 25^{0} C, relative humidity = 0.75 and wind speed = 10 miles/hour) as 5.99 mm/day. Therefore, the maximum limit of evaporation from soil surface was imposed as the equivalent 0.025 cm/hour.

The following assumptions were made in carrying out the study :

i) The water table was considered as static at the lower boundary of unsaturated zone.

ii) Soil air was regarded as a continuous phase, essentially at atmospheric pressure.

iii) K(h) and θ were assumed to be single-valued, non-decreasing functions of h.

(iv) The matric suction was assumed to be continuous at and through the interlayer boundary.

(v) Each of the two soil layers was assumed to be internally homogeneous with respect to its hydraulic properties.

vi) Thermal and osmotic gradients were assumed to be negligible.

For the given external evaporative conditions, water table depth and the layered soil profiles, the equilibrium moisture profile was obtained by using the numerical scheme presented in section 4.4 and assuming the pressure head at the soil surface to be in equilibrium with the surrounding atmosphere ($h(0,t) = -396.14$ cm). The initial and boundary conditions were defined by the equations (4.12), (4.13) and (4.14) respectively. The rate of loss of water (Darcian flux q) served as a measure of the evaporation rate, once the steady state was attained. Ample time was allowed for steady state to be attained. The values of soil water pressures at different nodes during consecutive time steps gave assurance that steady state had been attained. At steady state, the rate of loss of water (the flux q), which is approximately the same at every depth, equals the evaporation rate.

The evaporation rates under each set of water table depth and soil layerings were evaluated by using computer programs (Appendix-I and Appendix-II). A sample of the input data to the model and corresponding output are given in Appendix-III and Appendix-IV respectively.

In order to examine the possible behaviour of layered profiles consisting of sand and yolo light clay, the dependence of steady evaporation rate upon top layer thickness was determined for various ground water depths. The computed results are given in tables 1 and 2. The variations of evaporation rates from layered profiles as a function of top layer depth and three water table depths are also presented in figures 2 and 3 for the two soil layering sequences. The evaporation rates for homogeneous soil profiles (i.e., top layer thickness = 0) are also indicated for comparison.

For the case of yolo light clay over sand (table 1 and figure 2), the top layer was found to have little effect on the evaporation rate when ground water is at a depth of 140 cm. However, for ground water depth of 120 cm, the evaporation rate increases markedly as the top layer thickness increases. For ground water depth of 100 cm, the evaporation occurs at potential rate (6 mm/day) for the top layer thickness greater than 20 cm.

For the case of sand overlying yolo light clay (table 2 and figure 3), the evaporation rate is seen to fall marginally upto top layer thickness of around 14 cm, for ground water depth of 100 cm. With increase in the top layer thickness beyond 14 cm, the evaporation rate increases gradually. However, for ground water depth of 80 cm, the evaporation rate marginally falls upto top layer thickness of around 10 cm, increases steeply thereafter and attains potential rate at top layer thickness of around 42 cm. For ground water depth of 60 cm, the effect of top layer thickness becomes more pronounced, the decrease in rate being upto top layer thickness of around 5 cm and attaining potential rate at top layer thickness of only 18 cm.

Table 1 : Steady State Evaporation Rates from Layered Profiles of Yolo Light Clay over Sand

S.No.	Thickness of Top Layer (cm)	Evaporation Rates (mm/day)		
		L = 100 cm	L = 120 cm	L = 140 cm
1	0	4.030484	1.699340	0.814898
2	6	4.462968	1.875703	0.899408
3	10	4.840748	2.010576	0.958741
4	14	5.284893	2.161296	1.019669
5	18	5.809570	2.331051	1.086894
6	22	6.000000	2.522819	1.160336
7	26	6.000000	2.740535	1.240665
8	30	6.000000	2.985332	1.329966
9	34	6.000000	3.267940	1.429026
10	38	6.000000	3.590603	1.537989
11	42	6.000000	3.966448	1.660899
12	46	6.000000	4.400203	1.800100
13	50	6.000000	4.906572	1.955429

Table 2 : Steady State Evaporation Rates from Layered Profiles of Sand over Yolo Light Clay

S.No.	Thickness of Top Layer (cm)	Evaporation Rates (mm/day)		
		L = 60 cm	L = 80 cm	L = 100 cm
1	0	1.833106	1.088800	0.705278
2	6	1.685597	0.992043	0.647764
3	10	2.307044	0.954659	0.590985
4	14	4.937681	1.200225	0.582285
5	18	6.000000	1.720937	0.644286
6	22	6.000000	2.418169	0.766910
7	26	6.000000	3.202290	0.931421
8	30	6.000000	4.005234	1.119970
9	34	6.000000	4.789237	1.321853
10	38	6.000000	5.540245	1.526910
11	42	6.000000	6.000000	1.730477
12	46	6.000000	6.000000	1.930086
13	50	6.000000	6.000000	2.122212

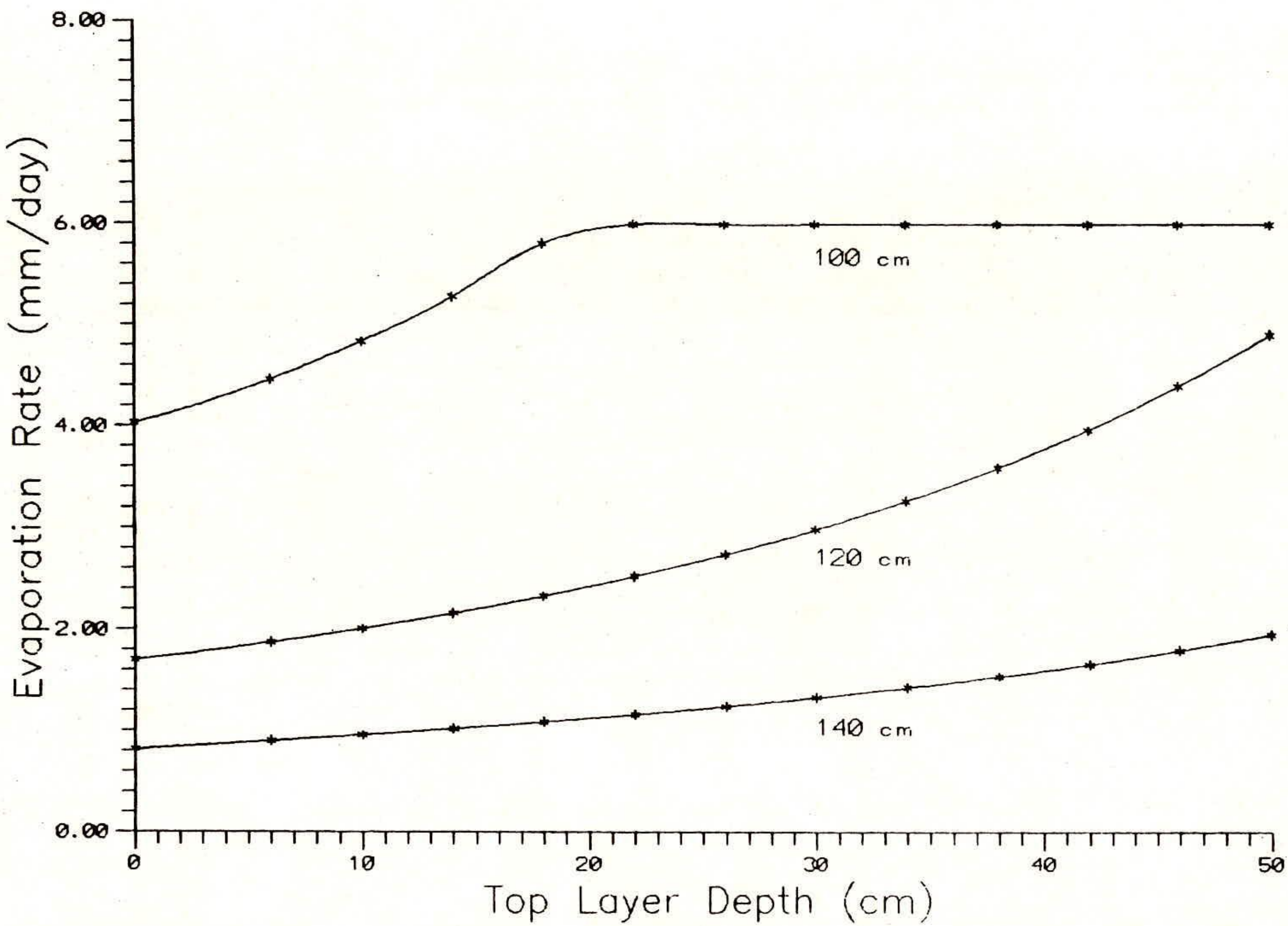

Figure 2 : Evaporation Rates from Layered Profiles of Yolo Light Clay over Sand as a Function of Top Layer Depth and Three Water Table Depths

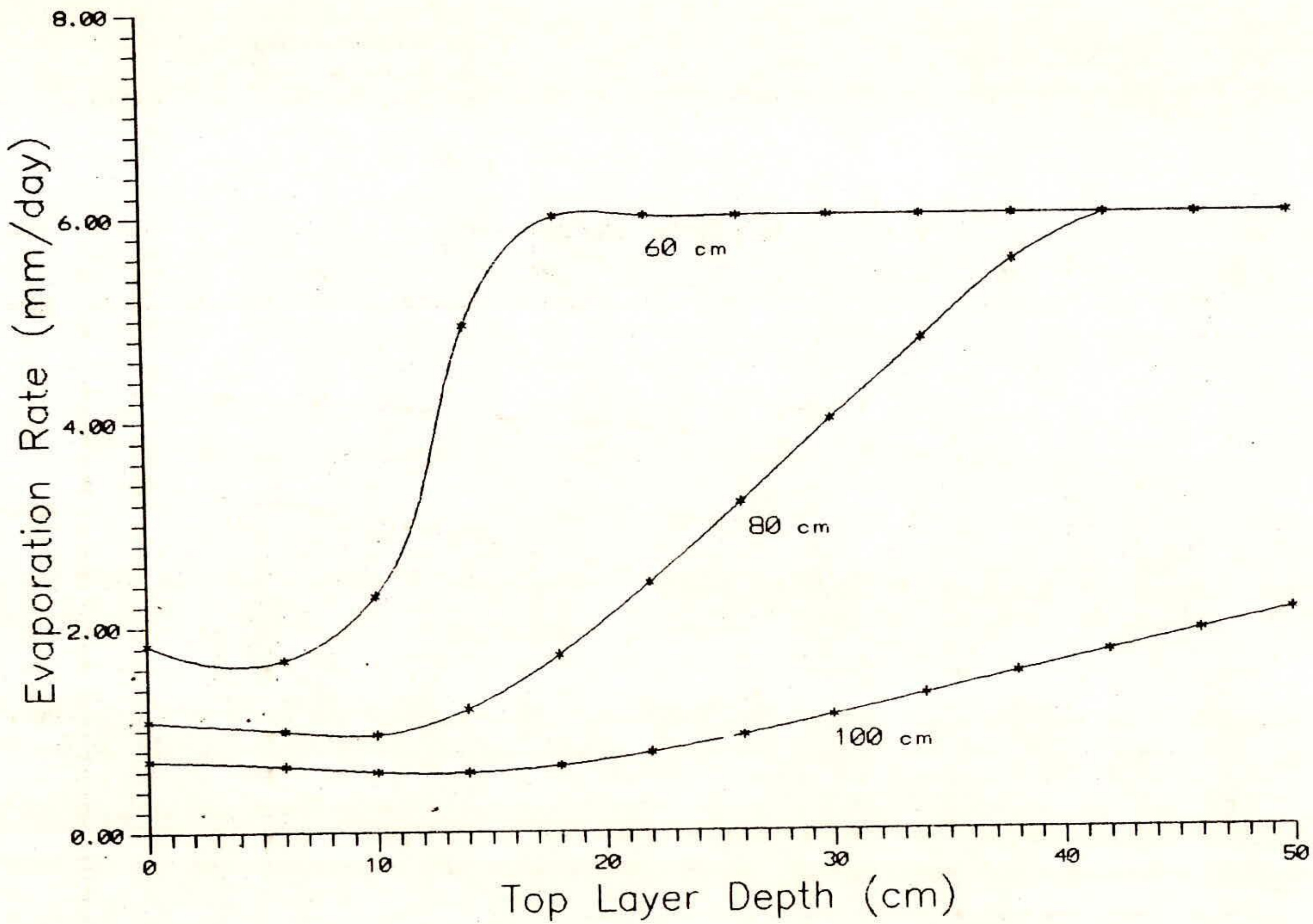

Figure 3 : Evaporation Rates from Layered Profiles of Sand over Yolo Light Clay as a Function of Top Layer Depth and Three Water Table Depths

The figures 2 and 3 also indicate that for deeper water tables, the evaporation rates were independent of the external evaporativity (6 mm/day). For shallower water tables, the evaporation rates followed the external "potential" rate as the top layer thickness was increased. It should be emphasized that the above results have not been subjected to empirical testing in the laboratory and field.

The significance of these results is that the presence of a fine-textured soil overlying a coarse-textured one, may increase the evaporative water loss for a given ground water depth. This effect will become more pronounced as the ground water becomes shallower. It is due to the higher unsaturated hydraulic conductivity of the finer material at the high suction values which prevail near the surface of the soil profile during evaporation. If the sequence of layers is a coarse-textured soil overlying a fine-textured one, the evaporative loss can be slightly reduced in case of thin top layers for corresponding water table depths. The effect of a tilled zone on the evaporation loss, as compared to a homogeneous soil profile, can thus be anticipated. The tilled top layer acts as if the soil has a coarse-textured layer on top, even though its texture is homogeneous. This implies that tillage operations can reduce evaporation and salinization to some extent in regions having shallow water tables.

The rate of accumulation of soluble salts due to upward movement of saline ground water can be obtained by multiplying the evaporation rate with salt concentration of the ground water. Even though the evaporation rate may be small, a significant quantity of salts may accumulate over a long period of time. If a crop is present, the above procedure may be used by taking bottom of the root zone as the upper boundary. The average suction at this point serves as the upper boundary condition. Then the amount of water and soluble salts moving up from a water table into the root zone can be calculated.

The above procedure, however, does not take into account the possibility that total hydraulic resistance or impedance to water flow of the entire soil profile may also include the existence of an inter-layer boundary impedance. The relation of the evaporation rate, e to the total soil profile impedance, ΣR can be given by

$$e = \frac{\Delta h}{\Sigma R} \qquad \ldots(5.2)$$

where Δh is the total hydraulic head difference between the water table and the soil surface. The total hydraulic impedance, ΣR is the sum of the impedances of the various layers. The impedance for any particular layer i is defined by the ratio $l_i/\bar{K}(h)_i$, where l_i is the layer thickness and $\bar{K}(h)_i$ its mean weighted hydraulic conductivity. The mean weighted conductivity, $\bar{K}(h)$ can be calculated by using the equation, $\bar{K}(h) = [\int K(h)dh]/[\int h(l)dl]$. Thus the total hydraulic impedance is given by

$$\Sigma R = R_B + R_T + R_{Int} \qquad \ldots(5.3)$$

where R_B and R_T are the impedances of the bottom and top layers respectively and R_{Int} is the possible impedance of the interlayer contact zone.

Under natural conditions, abrupt textural or structural discontinuity does not exist at the interlayer and it is more common to see a transition zone between the layers. However, any transition zone between the soil layers, howsoever gradual, is likely to exhibit differences in its hydraulic properties compared to those of either layer and thus acts as an additional soil layer having an impedance of its own.

Steady state conditions were assumed in this study, though the numerical model permits variable climatic parameters i.e. temperature and relative humidity. In nature, however, the systems considered are seldom in a steady state condition, principally because of the variations in meteorological conditions, soil salt content and water table depth. The changes in soil salt content and water table depth are relatively slow and therefore their short-period effects might be negligible. Their long-range influences, however, could be of considerable importance and should be taken into account, with different experimentally determined soil parameters and measured or predicted water table depths. Also under various conditions, the thermal transfer of water might significantly change the evaporation rate. In this study, the thermal transfer of liquidwater was entirely neglected.

6.0 CONCLUSIONS

The rise of water from a shallow water table can, in some cases, serve the useful purpose of supplying water to the root zone of crops. On the other hand, this process also entails the hazard of salinization, especially where the ground water is brackish and potential evaporativity is high. Excessive irrigation tends to raise the water table and thus aggravate the salinization problem. Lowering the water table by drainage can decisively reduce the rate of capillary rise and evaporation. Drainage is a costly operation, however, and it is therefore necessary, ahead of time, to determine the optimal depth to which the water table should be lowered. It requires the estimation of evaporation rates from bare soils with high water table conditions, specifically through multilayer profiles commonly found in nature.

A numerical model study has been carried out to estimate the steady state evaporation rates from shallow water table through a layered soil profile under isothermal conditions. An implicit finite-differencing technique is used for a mathematical model of one-dimensional, vertical, unsteady, unsaturated flow above a water table using the non-linear Richards equation.

The evaporation rates are shown to be related to the sequence and thickness of soil layers, their hydraulic properties and water table depths. It was found that soil layering reduces evaporation only marginally when a thin layer of coarse-textured soil overlies a fine-textured soil, as compared to a homogeneous profile. However, evaporation increases for thicker top layers of coarse-textured soil. The soil layering also increases the evaporation in case of fine-textured soil overlying a coarse-textured soil. The effect of layering was found to be more pronounced for shallow ground water depths. The possible existence of an interlayer contact zone impedance to water flow was not considered in the study.

The dependence of the actual steady state evaporation rate on water table depth and soil layerings, might be very useful in hydrologic practice. The extent, to which the above results can be applied quantitatively to the field, depends upon the correspondence between the input values of soil moisture characteristics for each layer and those existing in the field. The soil data employed might be less precise than desirable. In addition, it might not be possible to adequately take into account the variability of field soils.

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LAYERC.FOR

```
C       EVAPORATION FROM LAYERED SOILS
C       IN THE PRESENCE OF A WATER TABLE
C
C       LAYERED PROFILES OF YOLO LIGHT CLAY OVER SAND
C
C       IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION
C       (MODEL 4 OF HAVERKAMP ET AL., 1977)
C
        DIMENSION SUB(100),SUP(100),DIAG(100),B(100)
        DIMENSION H(100,2),CCC(100,2)
        DIMENSION THETA(100,2),HYDCON(100,2)
        DIMENSION HP(100,2),THETAP(100,2)
        OPEN(UNIT=1,FILE='LAYERC.DAT',STATUS='OLD')
        OPEN(UNIT=2,FILE='LAYERC.OUT',STATUS='NEW')
C
C       I REFERS TO DEPTH
C       J REFERS TO TIME
C       Z = DEPTH (CM), ORIENTED POSITIVELY DOWNWARD
C       THETA = VOLUMETRIC MOISTURE CONTENT (CUBIC CM / CUBIC CM)
C       H = SOIL WATER PRESSURE (RELATIVE TO THE ATMOSPHERE)
C           EXPRESSED IN CM OF WATER
C       R = UNIVERSAL GAS CONSTANT (ERGS/MOLE/K)
C       T = ABSOLUTE TEMPERATURE (K)
C           (READ IN CENTIGRADE AND CONVERTED IN K)
C       WM = MOLECULAR WEIGHT OF WATER (GM/MOLE)
C       G = ACCELERATION DUE TO GRAVITY (CM/SEC/SEC)
C       RH = RELATIVE HUMIDITY OF THE AIR (FRACTION)
C       THETAR, CTHETAR = RESIDUAL MOISTURE CONTENTS
C                         FOR THE TWO SOILS
C       THETAS, CTHETAS = MOISTURE CONTENTS AT SATURATION
C                         FOR THE TWO SOILS
C       BETA1, CONA, CBETA1, CCONA =  PARAMETERS IN THE HYDRAULIC
C                         CONDUCTIVITY AND SOIL WATER PRESSURE
C                         RELATIONSHIPS FOR THE TWO SOILS
C       BETA2, ALPHA, CBETA2, CALPHA = PARAMETERS IN THE MOISTURE
C                         CONTENT AND SOIL WATER PRESSURE RELATIONSHIPS
C                         FOR THE TWO SOILS
C       HYDCON = HYDRAULIC CONDUCTIVITY OF THE SOIL (CM/HOUR)
C       AKS, CAKS = HYDRAULIC CONDUCTIVITIES AT SATURATION (CM/HOUR)
C                   FOR THE TWO SOILS
C       DELT  = TIME STEP (HOURS)
C       DELZ  = DEPTH INTERVAL (CM)
C       NTIME = NUMBER OF TIME STEPS
C       NNODE = NUMBER OF NODES
C       CCC   = SPECIFIC WATER CAPACITY (/CM) DEFINED AS d(theta)/dh
C       DEPTH = WATER TABLE DEPTH
C       DT    = THICKNESS OF TOP LAYER
C
```

```
      READ(1,11)THETAR,THETAS,CTHETAR,CTHETAS
11    FORMAT(4F12.3)
      READ(1,12)BETA1,BETA2,CBETA1,CBETA2
12    FORMAT(4F12.3)
      READ(1,13)CONA,ALPHA,CCONA,CALPHA
13    FORMAT(4F12.3)
      READ(1,14)AKS,CAKS
14    FORMAT(2F12.5)
      READ(1,15)DELT,DELZ
15    FORMAT(F12.8,F12.3)
      READ(1,16)NTIME,NNODE
16    FORMAT(I7,5X,I5)
      READ(1,17)T
17    FORMAT(F5.2)
      READ(1,18)RH
18    FORMAT(F5.2)
      READ(1,19)DT
19    FORMAT(F5.2)
C
      TNODE = (DT+0.5*DELZ)/DELZ
      NODET = INT(TNODE)
C
C     READING OF INITIAL CONDITIONS
C
      READ(1,20)(H(I,1),I=1,NNODE)
20    FORMAT(5F12.6)
C
      IF(NODET.EQ.0) GO TO 102
      DO 101 I=1,NODET
      THETA(I,1)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1            +ALOG(ABS(H(I,1)))**CBETA2)+CTHETAR
      IF(H(I,1).GE.(-1.0))THETA(I,1)=CTHETAS
101   CONTINUE
102   DO 103 I=NODET+1,NNODE
      THETA(I,1)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,1))
     1            **BETA2)+THETAR
103   CONTINUE
C
      WRITE(2,21)
21    FORMAT(/2X,'EVAPORATION FROM LAYERED SOILS')
      WRITE(2,22)
22    FORMAT(2X,'IN THE PRESENCE OF A WATER TABLE')
      WRITE(2,23)
23    FORMAT(/2X,'YOLO LIGHT CLAY OVERLYING SAND')
      WRITE(2,24)
24    FORMAT(/2X,'ONE DIMENSIONAL RICHARDS EQUATION')
      WRITE(2,25)
25    FORMAT(2X,'IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION')
      DEPTH=(NNODE-1)*DELZ
      WRITE(2,26)
26    FORMAT(/2X,'WATER TABLE DEPTH')
      WRITE(2,27)DEPTH
27    FORMAT(2X,F7.3)
      WRITE(2,28)
28    FORMAT(/2X,'THICKNESS OF TOP LAYER')
```

```
      WRITE(2,29)DT
29    FORMAT(2X,F7.3)
      WRITE(2,30)
30    FORMAT(/2X,'TEMPERATURE IN CENTIGRADE')

      WRITE(2,31)T
31    FORMAT(F7.2)
      WRITE(2,32)
32    FORMAT(2X,'RELATIVE HUMIDITY OF THE AIR')
      WRITE(2,33)RH
33    FORMAT(F7.3)
      WRITE(2,34)
34    FORMAT(/2X,'THETAR',9X,'THETAS',9X,'CTHETAR',8X,'CTHETAS')
      WRITE(2,35)THETAR,THETAS,CTHETAR,CTHETAS
35    FORMAT(2X,F5.3,10X,F5.3,10X,F5.3,10X,F5.3)
      WRITE(2,36)
36    FORMAT(2X,'BETA1',10X,'BETA2',10X,'CBETA1',9X,'CBETA2')
      WRITE(2,37)BETA1,BETA2,CBETA1,CBETA2
37    FORMAT(2X,F5.3,10X,F5.3,10X,F5.3,10X,F5.3)
      WRITE(2,38)
38    FORMAT(2X,'CONA',11X,'ALPHA',10X,'CCONA',10X,'CALPHA')
      WRITE(2,39)CONA,ALPHA,CCONA,CALPHA
39    FORMAT(2X,F11.3,4X,F11.3,1X,F10.3,5X,F10.3)
      WRITE(2,40)
40    FORMAT(2X,'AKS',12X,'CAKS')
      WRITE(2,41)AKS,CAKS
41    FORMAT(2X,F8.5,7X,F8.5)
      WRITE(2,42)
42    FORMAT(2X,'DELT',11X,'DELZ')
      WRITE(2,43)DELT,DELZ
43    FORMAT(2X,F10.8,5X,F6.3)
      WRITE(2,44)
44    FORMAT(2X,'NTIME',10X,'NNODE')
      WRITE(2,45)NTIME,NNODE
45    FORMAT(I7,10X,I5)
      WRITE(2,46)
46    FORMAT(/2X,'SOIL MOISTURE AND SOIL WATER PRESSURE PROFILES')
      WRITE(2,47)
47    FORMAT(/2X,'INITIAL CONDITIONS'/)
      WRITE(2,48)(THETA(I,1),I=1,NNODE)
48    FORMAT(5F12.6)
      WRITE(2,*)
      WRITE(2,49)(H(I,1),I=1,NNODE)
49    FORMAT(5F12.6)
C
C     GENERATION OF UPPER BOUNDARY CONDITION
C
      R = 8.314E+7
      WM = 18.0
      G = 980.665
      TMP=T+273.15
      HU=R*TMP*ALOG(RH)/(WM*G)
      HU=HU/1019.80
      H(1,1)=HU
      H(1,2)=HU
```

```
      HP(1,1)=HU
      HP(1,2)=HU
      IF(NODET.EQ.0) GO TO 201
      THETA(1,1)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA+
     1           ALOG(ABS(H(1,1)))**CBETA2)+CTHETAR
      IF(H(1,1).GE.(-1.0))THETA(1,1)=CTHETAS
      GO TO 202
201   THETA(1,1)=ALPHA*(THETAS-THETAR)/(ALPHA+
     1           ABS(H(1,1))**BETA2)+THETAR
202   THETA(1,2)=THETA(1,1)
      THETAP(1,1)=THETA(1,1)
      THETAP(1,2)=THETA(1,1)
C
C     GENERATION OF LOWER BOUNDARY CONDITION
C
      THETA(NNODE,2)=THETA(NNODE,1)
      THETAP(NNODE,1)=THETA(NNODE,1)
      THETAP(NNODE,2)=THETA(NNODE,1)
      H(NNODE,2)=H(NNODE,1)
      HP(NNODE,1)=H(NNODE,1)
      HP(NNODE,2)=H(NNODE,1)
C
C     SIMULATION OF SOIL MOISTURE PROFILES
C
      E1=BETA1/BETA2
      E2=(THETAS-THETAR)
      E3=ALPHA**E1
      E4=CONA*AKS
      E5=1./BETA2*ALPHA**(1./BETA2)
      CE1=CBETA1/CBETA2
      CE2=(CTHETAS-CTHETAR)
      CE3=CALPHA**CE1
      CE4=CCONA*CAKS
      CE5=1./CBETA2*CALPHA**(1./CBETA2)
C
      DO 800 J=2,NTIME
      write(*,*)j
C
      IF(NODET.EQ.0) GO TO 302
      DO 301 I=1,NODET
      HYDCON(I,1)=CE4/(CCONA+(ABS(H(I,1)))**CBETA1)
      CCC(I,1)=EXP(-(CALPHA*(CTHETAS-THETA(I,1))/(THETA(I,1)
     1         -CTHETAR))**(1/CBETA2))*1./(CE5*CE2)*(CTHETAS
     2         -THETA(I,1))**(-1./CBETA2+1.)*(THETA(I,1)
     3         -CTHETAR)**(1./CBETA2+1.)
301   CONTINUE
302   DO 303 I=NODET+1, NNODE
      HYDCON(I,1)=E4/(CONA+(ABS(H(I,1)))**BETA1)
      CCC(I,1)=1./(E5*E2)*(THETAS-THETA(I,1))**(-1./BETA2+1.)*
     1         (THETA(I,1)-THETAR)**(1./BETA2+1.)
303   CONTINUE
C
      DO 903 I=2,NNODE-1
      DIAG(I-1)=2.*CCC(I,1)/HYDCON(I,1)+2.*DELT/DELZ**2
      SUB(I-1)=-DELT/DELZ**2
```

```
      SUP(I-1)=-DELT/DELZ**2
      IF(I.EQ.NODET)GO TO 901
      IF(I.EQ.NODET+1)GO TO 902
      B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ*.5
     1      *(HYDCON(I+1,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2      *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
      GO TO 903
901   B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1      *(HYDCON(I,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2      *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
      GO TO 903
902   B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1      *(HYDCON(I+1,1)-HYDCON(I,1))/HYDCON(I,1)
     2      *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
903   CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      DO 1000 I=1,NNODE-3
1000  SUB(I)=SUB(I+1)
      M=NNODE-2
      CALL TRID(M,SUP,SUB,DIAG,B)
      DO 1100 I=1,NNODE-2
1100  HP(I+1,2)=B(I)
C
      IF(NODET.LT.2) GO TO 403
      DO 401 I=2,NODET
      THETAP(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1           +ALOG(ABS(HP(I,2)))**CBETA2)+CTHETAR
      IF(HP(I,2).GE.(-1.0))THETAP(I,2)=CTHETAS
401   CONTINUE
      DO 402 I=NODET+1,NNODE-1
      THETAP(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(HP(I,2))
     1           **BETA2)+THETAR
402   CONTINUE
      GO TO 405
403   DO 404 I=2,NNODE-1
      THETAP(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(HP(I,2))
     1           **BETA2)+THETAR
404   CONTINUE
C
405   IF(NODET.EQ.0) GO TO 502
      DO 501 I=1,NODET
      HYDCON(I,1)=CE4/(CCONA+(ABS(HP(I,2)))**CBETA1)
      CCC(I,1)=EXP(-(CALPHA*(CTHETAS-THETAP(I,2))/(THETAP(I,2)
     1         -CTHETAR))**(1/CBETA2))*1./(CE5*CE2)*(CTHETAS
     2         -THETAP(I,2))**(-1./CBETA2+1.)*(THETAP(I,2)
     3         -CTHETAR)**(1./CBETA2+1.)
501   CONTINUE
502   DO 503 I=NODET+1,NNODE
      HYDCON(I,1)=E4/(CONA+(ABS(HP(I,2)))**BETA1)
      CCC(I,1)=1./(E5*E2)*(THETAS-THETAP(I,2))**(-1./BETA2+1.)*
     1         (THETAP(I,2)-THETAR)**(1./BETA2+1.)
503   CONTINUE
C
```

```
      DO 1203 I=2,NNODE-1
      DIAG(I-1)=CCC(I,1)/HYDCON(I,1)+DELT/DELZ**2
      SUB(I-1)=-DELT/DELZ**2*.5
      SUP(I-1)=-DELT/DELZ**2*.5
      IF(I.EQ.NODET)GO TO 1201
      IF(I.EQ.NODET+1)GO TO 1202
      B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ*.5
     1       *(HYDCON(I+1,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2       *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3       +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
      GO TO 1203
1201  B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1       *(HYDCON(I,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2       *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3       +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
      GO TO 1203
1202  B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1       *(HYDCON(I+1,1)-HYDCON(I,1))/HYDCON(I,1)
     2       *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3       +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
1203  CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      DO 1300 I=1,NNODE-3
1300  SUB(I)=SUB(I+1)
      M=NNODE-2
      CALL TRID(M,SUP,SUB,DIAG,B)
      DO 1400 I=1,NNODE-2
1400  H(I+1,2)=B(I)
C
      IF(NODET.LT.2) GO TO 603
      DO 601 I=2,NODET
      THETA(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1           +ALOG(ABS(H(I,2)))**CBETA2)+CTHETAR
      IF(H(I,2).GE.(-1.0))THETA(I,2)=CTHETAS
601   CONTINUE
      DO 602 I=NODET+1,NNODE-1
      THETA(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,2))
     1           **BETA2)+THETAR
602   CONTINUE
      GO TO 605
603   DO 604 I=2,NNODE-1
      THETA(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,2))
     1           **BETA2)+THETAR
604   CONTINUE
C
605   IF (J.EQ.2161)   GO TO 111
      IF (J.EQ.4321)   GO TO 111
      IF (J.EQ.6481)   GO TO 111
      IF (J.EQ.8641)   GO TO 111
      IF (J.EQ.10801)  GO TO 111
      IF (J.EQ.12961)  GO TO 111
      IF (J.EQ.15121)  GO TO 111
      IF (J.EQ.17281)  GO TO 111
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      IF (J.EQ.19441)  GO TO 111
      IF (J.EQ.21601)  GO TO 111
      IF (J.EQ.23761)  GO TO 111
      IF (J.EQ.25921)  GO TO 111
      IF (J.EQ.28081)  GO TO 111
      IF (J.EQ.30241)  GO TO 111
      IF (J.EQ.32401)  GO TO 111
      IF (J.EQ.34561)  GO TO 111
      IF (J.EQ.36721)  GO TO 111
      IF (J.EQ.38881)  GO TO 111
      IF (J.EQ.41041)  GO TO 111
      IF (J.EQ.43201)  GO TO 111
      IF (J.EQ.45361)  GO TO 111
      IF (J.EQ.47521)  GO TO 111
      IF (J.EQ.49681)  GO TO 111
      IF (J.EQ.51841)  GO TO 111
      IF (J.EQ.54001)  GO TO 111
      IF (J.EQ.56161)  GO TO 111
      IF (J.EQ.58321)  GO TO 111
      IF (J.EQ.60481)  GO TO 111
      IF (J.EQ.62641)  GO TO 111
      IF (J.EQ.64801)  GO TO 111
      IF (J.EQ.66961)  GO TO 111
      IF (J.EQ.69121)  GO TO 111
      IF (J.EQ.71281)  GO TO 111
      IF (J.EQ.73441)  GO TO 111
      IF (J.EQ.75601)  GO TO 111
      IF (J.EQ.77761)  GO TO 111
      IF (J.EQ.79921)  GO TO 111
      IF (J.EQ.82081)  GO TO 111
      IF (J.EQ.84241)  GO TO 111
      IF (J.EQ.86401)  GO TO 111
      IF (J.EQ.88561)  GO TO 111
      IF (J.EQ.90721)  GO TO 111
      IF (J.EQ.92881)  GO TO 111
      IF (J.EQ.95041)  GO TO 111
      IF (J.EQ.97201)  GO TO 111
      IF (J.EQ.99361)  GO TO 111
      IF (J.EQ.101521) GO TO 111
      IF (J.EQ.103681) GO TO 111
      IF (J.EQ.105841) GO TO 111
      IF (J.EQ.108001) GO TO 111
      IF (J.EQ.110161) GO TO 111
      IF (J.EQ.112321) GO TO 111
      GO TO 333
111   CONTINUE
C
C     ESTIMATION OF EVAPORATION RATES
C
      IF(NODET.EQ.0) GO TO 702
      DO 701 I=1,NODET
      HYDCON(I,2)=CE4/(CCONA+(ABS(H(I,2)))**CBETA1)
701   CONTINUE
702   DO 703 I=NODET+1,NNODE
      HYDCON(I,2)=E4/(CONA+(ABS(H(I,2)))**BETA1)
```

```
703     CONTINUE
C
        EVAP = 0.0
        DO 222 N = 2,NNODE
        EVAP=EVAP+((HYDCON(N,2)*HYDCON(N-1,2))**0.5)*
     1       (((H(N,2)-H(N-1,2))/DELZ)-1.0)*240.0
222     CONTINUE
        EVAP = EVAP/(NNODE-1)
        IF(EVAP.GT.6.0)EVAP=6.0
C
        ITIME=J-1
        HOUR=ITIME*DELT
        WRITE(2,51)ITIME,HOUR
51      FORMAT(/2X,'TIME STEP  =',I7,6X,'DURATION  = ',F10.4,1X,'HOURS'/)
        WRITE(2,52)(THETA(I,2),I=1,NNODE)
52      FORMAT(5F12.6)
        WRITE(2,*)
        WRITE(2,53)(H(I,2),I=1,NNODE)
53      FORMAT(5F12.6)
        WRITE(2,54)EVAP
54      FORMAT(/2X,'EVAPORATION LOSSES      =     ',F12.6,'    MM/DAY'//)
333     CONTINUE
C
        DO 1500 I = 2, NNODE-1
        THETA(I,1) = THETA(I,2)
        H(I,1) = H(I,2)
1500    CONTINUE
C
800     CONTINUE
        STOP
        END
C
        SUBROUTINE TRID(M,SUP,SUB,DIAG,B)
        DIMENSION SUP(100),SUB(100),DIAG(100),B(100)
        N=M
        NN=N-1
        SUP(1)=SUP(1)/DIAG(1)
        B(1)=B(1)/DIAG(1)
        DO 61 I=2,N
        II=I-1
        DIAG(I)=DIAG(I)-SUP(II)*SUB(II)
        IF (I.EQ.N) GO TO 61
        SUP(I)=SUP(I)/DIAG(I)
 61     B(I)=(B(I)-SUB(II)*B(II))/DIAG(I)
        DO 62 K=1,NN
        I=N-K
 62     B(I)=B(I)-SUP(I)*B(I+1)
        RETURN
        END
```

LAYERS.FOR

```
C       EVAPORATION FROM LAYERED SOILS
C       IN THE PRESENCE OF A WATER TABLE
C
C       LAYERED PROFILES OF SAND OVER YOLO LIGHT CLAY
C
C       IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION
C       (MODEL 4 OF HAVERKAMP ET AL., 1977)
C
        DIMENSION SUB(100),SUP(100),DIAG(100),B(100)
        DIMENSION H(100,2),CCC(100,2)
        DIMENSION THETA(100,2),HYDCON(100,2)
        DIMENSION HP(100,2),THETAP(100,2)
        OPEN(UNIT=1,FILE='LAYERS.DAT',STATUS='OLD')
        OPEN(UNIT=2,FILE='LAYERS.OUT',STATUS='NEW')
C
C       I REFERS TO DEPTH
C       J REFERS TO TIME
C       Z = DEPTH (CM), ORIENTED POSITIVELY DOWNWARD
C       THETA = VOLUMETRIC MOISTURE CONTENT (CUBIC CM / CUBIC CM)
C       H = SOIL WATER PRESSURE (RELATIVE TO THE ATMOSPHERE)
C           EXPRESSED IN CM OF WATER
C       R = UNIVERSAL GAS CONSTANT (ERGS/MOLE/K)
C       T = ABSOLUTE TEMPERATURE (K)
C           (READ IN CENTIGRADE AND CONVERTED IN K)
C       WM = MOLECULAR WEIGHT OF WATER (GM/MOLE)
C       G = ACCELERATION DUE TO GRAVITY (CM/SEC/SEC)
C       RH = RELATIVE HUMIDITY OF THE AIR (FRACTION)
C       THETAR, CTHETAR = RESIDUAL MOISTURE CONTENTS
C                         FOR THE TWO SOILS
C       THETAS, CTHETAS = MOISTURE CONTENTS AT SATURATION
C                         FOR THE TWO SOILS
C       BETA1, CONA, CBETA1, CCONA =  PARAMETERS IN THE HYDRAULIC
C                         CONDUCTIVITY AND SOIL WATER PRESSURE
C                         RELATIONSHIPS FOR THE TWO SOILS
C       BETA2, ALPHA, CBETA2, CALPHA = PARAMETERS IN THE MOISTURE
C                         CONTENT AND SOIL WATER PRESSURE RELATIONSHIPS
C                         FOR THE TWO SOILS
C       HYDCON = HYDRAULIC CONDUCTIVITY OF THE SOIL (CM/HOUR)
C       AKS, CAKS = HYDRAULIC CONDUCTIVITIES AT SATURATION (CM/HOUR)
C                   FOR THE TWO SOILS
C       DELT  = TIME STEP (HOURS)
C       DELZ  = DEPTH INTERVAL (CM)
C       NTIME = NUMBER OF TIME STEPS
C       NNODE = NUMBER OF NODES
C       CCC   = SPECIFIC WATER CAPACITY (/CM) DEFINED AS d(theta)/dh
C       DEPTH = WATER TABLE DEPTH
C       DT    = THICKNESS OF TOP LAYER
C
```

```
      READ(1,11)THETAR,THETAS,CTHETAR,CTHETAS
11    FORMAT(4F12.3)
      READ(1,12)BETA1,BETA2,CBETA1,CBETA2
12    FORMAT(4F12.3)
      READ(1,13)CONA,ALPHA,CCONA,CALPHA
13    FORMAT(4F12.3)
      READ(1,14)AKS,CAKS
14    FORMAT(2F12.5)
      READ(1,15)DELT,DELZ
15    FORMAT(F12.8,F12.3)
      READ(1,16)NTIME,NNODE
16    FORMAT(I7,5X,I5)
      READ(1,17)T
17    FORMAT(F5.2)
      READ(1,18)RH
18    FORMAT(F5.2)
      READ(1,19)DT
19    FORMAT(F5.2)
C
      TNODE = (DT+0.5*DELZ)/DELZ
      NODET = INT(TNODE)
C
C     READING OF INITIAL CONDITIONS
C
      READ(1,20)(H(I,1),I=1,NNODE)
20    FORMAT(5F12.6)
C
      IF(NODET.EQ.0) GO TO 102
      DO 101 I=1,NODET
      THETA(I,1)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,1))
     1          **BETA2)+THETAR
101   CONTINUE
102   DO 103 I=NODET+1,NNODE-1
      THETA(I,1)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1          +ALOG(ABS(H(I,1)))**CBETA2)+CTHETAR
      IF(H(I,1).GE.(-1.0))THETA(I,1)=CTHETAS
103   CONTINUE
      THETA(NNODE,1)=CTHETAS
C
      WRITE(2,21)
21    FORMAT(/2X,'EVAPORATION FROM LAYERED SOILS')
      WRITE(2,22)
22    FORMAT(2X,'IN THE PRESENCE OF A WATER TABLE')
      WRITE(2,23)
23    FORMAT(/2X,'SAND OVERLYING YOLO LIGHT CLAY')
      WRITE(2,24)
24    FORMAT(/2X,'ONE DIMENSIONAL RICHARDS EQUATION')
      WRITE(2,25)
25    FORMAT(2X,'IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION')
      DEPTH=(NNODE-1)*DELZ
      WRITE(2,26)
26    FORMAT(/2X,'WATER TABLE DEPTH')
      WRITE(2,27)DEPTH
27    FORMAT(2X,F7.3)
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```
      WRITE(2,28)
28    FORMAT(/2X,'THICKNESS OF TOP LAYER')
      WRITE(2,29)DT
29    FORMAT(2X,F7.3)
      WRITE(2,30)
30    FORMAT(/2X,'TEMPERATURE IN CENTIGRADE')
      WRITE(2,31)T
31    FORMAT(F7.2)
      WRITE(2,32)
32    FORMAT(2X,'RELATIVE HUMIDITY OF THE AIR')
      WRITE(2,33)RH
33    FORMAT(F7.3)
      WRITE(2,34)
34    FORMAT(/2X,'THETAR',9X,'THETAS',9X,'CTHETAR',8X,'CTHETAS')
      WRITE(2,35)THETAR,THETAS,CTHETAR,CTHETAS
35    FORMAT(2X,F5.3,10X,F5.3,10X,F5.3,10X,F5.3)
      WRITE(2,36)
36    FORMAT(2X,'BETA1',10X,'BETA2',10X,'CBETA1',9X,'CBETA2')
      WRITE(2,37)BETA1,BETA2,CBETA1,CBETA2
37    FORMAT(2X,F5.3,10X,F5.3,10X,F5.3,10X,F5.3)
      WRITE(2,38)
38    FORMAT(2X,'CONA',11X,'ALPHA',10X,'CCONA',10X,'CALPHA')
      WRITE(2,39)CONA,ALPHA,CCONA,CALPHA
39    FORMAT(2X,F11.3,4X,F11.3,1X,F10.3,5X,F10.3)
      WRITE(2,40)
40    FORMAT(2X,'AKS',12X,'CAKS')
      WRITE(2,41)AKS,CAKS
41    FORMAT(2X,F8.5,7X,F8.5)
      WRITE(2,42)
42    FORMAT(2X,'DELT',11X,'DELZ')
      WRITE(2,43)DELT,DELZ
43    FORMAT(2X,F10.8,5X,F6.3)
      WRITE(2,44)
44    FORMAT(2X,'NTIME',10X,'NNODE')
      WRITE(2,45)NTIME,NNODE
45    FORMAT(I7,10X,I5)
      WRITE(2,46)
46    FORMAT(/2X,'SOIL MOISTURE AND SOIL WATER PRESSURE PROFILES')
      WRITE(2,47)
47    FORMAT(/2X,'INITIAL CONDITIONS'/)
      WRITE(2,48)(THETA(I,1),I=1,NNODE)
48    FORMAT(5F12.6)
      WRITE(2,*)
      WRITE(2,49)(H(I,1),I=1,NNODE)
49    FORMAT(5F12.6)
C
C     GENERATION OF UPPER BOUNDARY CONDITION
C
      R = 8.314E+7
      WM = 18.0
      G = 980.665
      TMP=T+273.15
      HU=R*TMP*ALOG(RH)/(WM*G)
      HU=HU/1019.80
      H(1,1)=HU
```

```
      H(1,2)=HU
      HP(1,1)=HU
      HP(1,2)=HU
      IF(NODET.EQ.0) GO TO 201
      THETA(1,1)=ALPHA*(THETAS-THETAR)/(ALPHA+
     1           ABS(H(1,1))**BETA2)+THETAR
      GO TO 202
201   THETA(1,1)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA+
     1           ALOG(ABS(H(1,1)))**CBETA2)+CTHETAR
      IF(H(1,1).GE.(-1.0))THETA(1,1)=CTHETAS
202   THETA(1,2)=THETA(1,1)
      THETAP(1,1)=THETA(1,1)
      THETAP(1,2)=THETA(1,1)
C
C     GENERATION OF LOWER BOUNDARY CONDITION
C
      THETA(NNODE,2)=THETA(NNODE,1)
      THETAP(NNODE,1)=THETA(NNODE,1)
      THETAP(NNODE,2)=THETA(NNODE,1)
      H(NNODE,2)=H(NNODE,1)
      HP(NNODE,1)=H(NNODE,1)
      HP(NNODE,2)=H(NNODE,1)
C
C     SIMULATION OF SOIL MOISTURE PROFILES
C
      E1=BETA1/BETA2
      E2=(THETAS-THETAR)
      E3=ALPHA**E1
      E4=CONA*AKS
      E5=1./BETA2*ALPHA**(1./BETA2)
      CE1=CBETA1/CBETA2
      CE2=(CTHETAS-CTHETAR)
      CE3=CALPHA**CE1
      CE4=CCONA*CAKS
      CE5=1./CBETA2*CALPHA**(1./CBETA2)
C
      DO 800 J=2,NTIME
      write(*,*)j
C
      IF(NODET.EQ.0) GO TO 302
      DO 301 I=1,NODET
      HYDCON(I,1)=E4/(CONA+(ABS(H(I,1)))**BETA1)
      CCC(I,1)=1./(E5*E2)*(THETAS-THETA(I,1))**(-1./BETA2+1.)*
     1         (THETA(I,1)-THETAR)**(1./BETA2+1.)
301   CONTINUE
302   DO 303 I=NODET+1, NNODE
      HYDCON(I,1)=CE4/(CCONA+(ABS(H(I,1)))**CBETA1)
      CCC(I,1)=EXP(-(CALPHA*(CTHETAS-THETA(I,1))/(THETA(I,1)
     1         -CTHETAR))**(1/CBETA2))*1./(CE5*CE2)*(CTHETAS
     2         -THETA(I,1))**(-1./CBETA2+1.)*(THETA(I,1)
     3         -CTHETAR)**(1./CBETA2+1.)
303   CONTINUE
C
```

```
      DO 903 I=2,NNODE-1
      DIAG(I-1)=2.*CCC(I,1)/HYDCON(I,1)+2.*DELT/DELZ**2
      SUB(I-1)=-DELT/DELZ**2
      SUP(I-1)=-DELT/DELZ**2
      IF(I.EQ.NODET)GO TO 901
      IF(I.EQ.NODET+1)GO TO 902
      B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ*.5
     1        *(HYDCON(I+1,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2        *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
      GO TO 903
901   B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1        *(HYDCON(I,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2        *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
      GO TO 903
902   B(I-1)=2.*CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1        *(HYDCON(I+1,1)-HYDCON(I,1))/HYDCON(I,1)
     2        *((H(I+1,1)-H(I-1,1))/(2.*DELZ)-1.)
903   CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      DO 1000 I=1,NNODE-3
1000  SUB(I)=SUB(I+1)
      M=NNODE-2
      CALL TRID(M,SUP,SUB,DIAG,B)
      DO 1100 I=1,NNODE-2
1100  HP(I+1,2)=B(I)
C
      IF(NODET.LT.2) GO TO 403
      DO 401 I=2,NODET
      THETAP(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(HP(I,2))
     1              **BETA2)+THETAR
401   CONTINUE
      DO 402 I=NODET+1,NNODE-1
      THETAP(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1              +ALOG(ABS(HP(I,2)))**CBETA2)+CTHETAR
      IF(HP(I,2).GE.(-1.0))THETAP(I,2)=CTHETAS
402   CONTINUE
      GO TO 405
403   DO 404 I=2,NNODE-1
      THETAP(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1              +ALOG(ABS(HP(I,2)))**CBETA2)+CTHETAR
      IF(HP(I,2).GE.(-1.0))THETAP(I,2)=CTHETAS
404   CONTINUE
C
405   IF(NODET.EQ.0) GO TO 502
      DO 501 I=1,NODET
      HYDCON(I,1)=E4/(CONA+(ABS(HP(I,2)))**BETA1)
      CCC(I,1)=1./(E5*E2)*(THETAS-THETAP(I,2))**(-1./BETA2+1.)*
     1          (THETAP(I,2)-THETAR)**(1./BETA2+1.)
501   CONTINUE
502   DO 503 I=NODET+1,NNODE
      HYDCON(I,1)=CE4/(CCONA+(ABS(HP(I,2)))**CBETA1)
```

```
      CCC(I,1)=EXP(-(CALPHA*(CTHETAS-THETAP(I,2))/(THETAP(I,2)
     1          -CTHETAR))**(1/CBETA2))*1./(CE5*CE2)*(CTHETAS
     2          -THETAP(I,2))**(-1./CBETA2+1.)*(THETAP(I,2)
     3          -CTHETAR)**(1./CBETA2+1.)
503   CONTINUE
C
      DO 1203 I=2,NNODE-1
      DIAG(I-1)=CCC(I,1)/HYDCON(I,1)+DELT/DELZ**2
      SUB(I-1)=-DELT/DELZ**2*.5
      SUP(I-1)=-DELT/DELZ**2*.5
      IF(I.EQ.NODET)GO TO 1201
      IF(I.EQ.NODET+1)GO TO 1202
      B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ*.5
     1        *(HYDCON(I+1,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2        *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3        +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
      GO TO 1203
1201  B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1        *(HYDCON(I,1)-HYDCON(I-1,1))/HYDCON(I,1)
     2        *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3        +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
      GO TO 1203
1202  B(I-1)=CCC(I,1)/HYDCON(I,1)*H(I,1)+DELT/DELZ
     1        *(HYDCON(I+1,1)-HYDCON(I,1))/HYDCON(I,1)
     2        *((HP(I+1,2)-HP(I-1,2))/(2.*DELZ)-1.)
     3        +DELT/DELZ**2*.5*(H(I+1,1)-2.*H(I,1)+H(I-1,1))
1203  CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      DO 1300 I=1,NNODE-3
1300  SUB(I)=SUB(I+1)
      M=NNODE-2
      CALL TRID(M,SUP,SUB,DIAG,B)
      DO 1400 I=1,NNODE-2
1400  H(I+1,2)=B(I)
C
      IF(NODET.LT.2) GO TO 603
      DO 601 I=2,NODET
      THETA(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,2))
     1           **BETA2)+THETAR
601   CONTINUE
      DO 602 I=NODET+1,NNODE-1
      THETA(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1           +ALOG(ABS(H(I,2)))**CBETA2)+CTHETAR
      IF(H(I,2).GE.(-1.0))THETA(I,2)=CTHETAS
602   CONTINUE
      GO TO 605
603   DO 604 I=2,NNODE-1
      THETA(I,2)=CALPHA*(CTHETAS-CTHETAR)/(CALPHA
     1           +ALOG(ABS(H(I,2)))**CBETA2)+CTHETAR
      IF(H(I,2).GE.(-1.0))THETA(I,2)=CTHETAS
604   CONTINUE
C
```

```
605     IF (J.EQ.2161)    GO TO 111
        IF (J.EQ.4321)    GO TO 111
        IF (J.EQ.6481)    GO TO 111
        IF (J.EQ.8641)    GO TO 111
        IF (J.EQ.10801)   GO TO 111
        IF (J.EQ.12961)   GO TO 111
        IF (J.EQ.15121)   GO TO 111
        IF (J.EQ.17281)   GO TO 111
        IF (J.EQ.19441)   GO TO 111
        IF (J.EQ.21601)   GO TO 111
        IF (J.EQ.23761)   GO TO 111
        IF (J.EQ.25921)   GO TO 111
        IF (J.EQ.28081)   GO TO 111
        IF (J.EQ.30241)   GO TO 111
        IF (J.EQ.32401)   GO TO 111
        IF (J.EQ.34561)   GO TO 111
        IF (J.EQ.36721)   GO TO 111
        IF (J.EQ.38881)   GO TO 111
        IF (J.EQ.41041)   GO TO 111
        IF (J.EQ.43201)   GO TO 111
        IF (J.EQ.45361)   GO TO 111
        IF (J.EQ.47521)   GO TO 111
        IF (J.EQ.49681)   GO TO 111
        IF (J.EQ.51841)   GO TO 111
        IF (J.EQ.54001)   GO TO 111
        IF (J.EQ.56161)   GO TO 111
        IF (J.EQ.58321)   GO TO 111
        IF (J.EQ.60481)   GO TO 111
        IF (J.EQ.62641)   GO TO 111
        IF (J.EQ.64801)   GO TO 111
        IF (J.EQ.66961)   GO TO 111
        IF (J.EQ.69121)   GO TO 111
        IF (J.EQ.71281)   GO TO 111
        IF (J.EQ.73441)   GO TO 111
        IF (J.EQ.75601)   GO TO 111
        IF (J.EQ.77761)   GO TO 111
        IF (J.EQ.79921)   GO TO 111
        IF (J.EQ.82081)   GO TO 111
        IF (J.EQ.84241)   GO TO 111
        IF (J.EQ.86401)   GO TO 111
        IF (J.EQ.88561)   GO TO 111
        IF (J.EQ.90721)   GO TO 111
        IF (J.EQ.92881)   GO TO 111
        IF (J.EQ.95041)   GO TO 111
        IF (J.EQ.97201)   GO TO 111
        IF (J.EQ.99361)   GO TO 111
        IF (J.EQ.101521)  GO TO 111
        IF (J.EQ.103681)  GO TO 111
        IF (J.EQ.105841)  GO TO 111
        IF (J.EQ.108001)  GO TO 111
        IF (J.EQ.110161)  GO TO 111
        IF (J.EQ.112321)  GO TO 111
        IF (J.EQ.114481)  GO TO 111
        IF (J.EQ.116641)  GO TO 111
        IF (J.EQ.118801)  GO TO 111
```

```
      IF (J.EQ.120961) GO TO 111
      IF (J.EQ.123121) GO TO 111
      IF (J.EQ.125281) GO TO 111
      IF (J.EQ.127441) GO TO 111
      IF (J.EQ.129601) GO TO 111
      IF (J.EQ.131761) GO TO 111
      IF (J.EQ.133921) GO TO 111
      IF (J.EQ.136081) GO TO 111
      IF (J.EQ.138241) GO TO 111
      IF (J.EQ.140401) GO TO 111
      IF (J.EQ.142561) GO TO 111
      IF (J.EQ.144721) GO TO 111
      IF (J.EQ.146881) GO TO 111
      IF (J.EQ.149041) GO TO 111
      IF (J.EQ.151201) GO TO 111
      GO TO 333
111   CONTINUE
C
C     ESTIMATION OF EVAPORATION RATES
C
      IF(NODET.EQ.0) GO TO 702
      DO 701 I=1,NODET
      HYDCON(I,2)=E4/(CONA+(ABS(H(I,2)))**BETA1)
701   CONTINUE
702   DO 703 I=NODET+1,NNODE
      HYDCON(I,2)=CE4/(CCONA+(ABS(H(I,2)))**CBETA1)
703   CONTINUE
C
      EVAP = 0.0
      DO 222 N = 2,NNODE
      EVAP=EVAP+((HYDCON(N,2)*HYDCON(N-1,2))**0.5)*
     1     (((H(N,2)-H(N-1,2))/DELZ)-1.0)*240.0
222   CONTINUE
      EVAP = EVAP/(NNODE-1)
      IF(EVAP.GT.6.0)EVAP=6.0
C
      ITIME=J-1
      HOUR=ITIME*DELT
      WRITE(2,51)ITIME,HOUR
51    FORMAT(/2X,'TIME STEP  =',I7,6X,'DURATION  = ',F10.4,1X,'HOURS'/)
      WRITE(2,52)(THETA(I,2),I=1,NNODE)
52    FORMAT(5F12.6)
      WRITE(2,*)
      WRITE(2,53)(H(I,2),I=1,NNODE)
53    FORMAT(5F12.6)
      WRITE(2,54)EVAP
54    FORMAT(/2X,'EVAPORATION LOSSES      =     ',F12.6,'    MM/DAY'//)
333   CONTINUE
C
      DO 1500 I = 2, NNODE-1
      THETA(I,1) = THETA(I,2)
      H(I,1) = H(I,2)
1500  CONTINUE
C
800   CONTINUE
```

```
      STOP
      END
C
      SUBROUTINE TRID(M,SUP,SUB,DIAG,B)
      DIMENSION SUP(100),SUB(100),DIAG(100),B(100)
      N=M
      NN=N-1
      SUP(1)=SUP(1)/DIAG(1)
      B(1)=B(1)/DIAG(1)
      DO 61 I=2,N
      II=I-1
      DIAG(I)=DIAG(I)-SUP(II)*SUB(II)
      IF (I.EQ.N) GO TO 61
      SUP(I)=SUP(I)/DIAG(I)
61    B(I)=(B(I)-SUB(II)*B(II))/DIAG(I)
      DO 62 K=1,NN
      I=N-K
62    B(I)=B(I)-SUP(I)*B(I+1)
      RETURN
      END
```

LAYERS.DAT

```
0.075       0.287          0.124      0.495
4.740       3.960          1.770      4.000
1175000.000 1611000.000  124.600    739.000
34.00000    0.04428
0.01111111  4.000
32401       21
25.00
00.75
34.00
 -80.000000  -76.000000  -72.000000  -68.000000  -64.000000
 -60.000000  -56.000000  -52.000000  -48.000000  -44.000000
 -40.000000  -36.000000  -32.000000  -28.000000  -24.000000
 -20.000000  -16.000000  -12.000000   -8.000000   -4.000000
  -0.000000
```

LAYERS.OUT

EVAPORATION FROM LAYERED SOILS
IN THE PRESENCE OF A WATER TABLE

SAND OVERLYING YOLO LIGHT CLAY

ONE DIMENSIONAL RICHARDS EQUATION
IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION

WATER TABLE DEPTH
80.000

THICKNESS OF TOP LAYER
34.000

TEMPERATURE IN CENTIGRADE
25.00
RELATIVE HUMIDITY OF THE AIR
.750

THETAR	THETAS	CTHETAR	CTHETAS
.075	.287	.124	.495
BETA1	BETA2	CBETA1	CBETA2
4.740	3.960	1.770	4.000
CONA	ALPHA	CCONA	CALPHA
1175000.000	1611000.000	124.600	739.000
AKS	CAKS		
34.00000	.04428		
DELT	DELZ		
.01111111	4.000		
NTIME	NNODE		
32401	21		

SOIL MOISTURE AND SOIL WATER PRESSURE PROFILES

INITIAL CONDITIONS

.084491	.086512	.089078	.092362	.096593
.102077	.109212	.118487	.130462	.414413
.420664	.427316	.434402	.441955	.449999
.458540	.467530	.476798	.485845	.493155
.495000				

-80.000000	-76.000000	-72.000000	-68.000000	-64.000000
-60.000000	-56.000000	-52.000000	-48.000000	-44.000000
-40.000000	-36.000000	-32.000000	-28.000000	-24.000000
-20.000000	-16.000000	-12.000000	-8.000000	-4.000000
.000000				

TIME STEP = 90 DURATION = 1.0000 HOURS

.075018	.077448	.081679	.086704	.092514
.099304	.107439	.117440	.129943	.414400
.420664	.427316	.434402	.441955	.449999
.458540	.467530	.476798	.485845	.493155
.495000				

-396.140700	-113.614200	-87.727610	-75.664870	-67.836300
-61.891670	-56.899960	-52.402660	-48.154470	-44.009010
-40.000360	-36.000000	-32.000000	-28.000000	-24.000000
-20.000000	-16.000000	-12.000000	-8.000000	-4.000000
.000000				

EVAPORATION LOSSES = 4.456627 MM/DAY

TIME STEP = 2160 DURATION = 24.0000 HOURS

.075018	.076901	.080162	.084208	.089204
.095449	.103333	.113336	.125998	.413092
.419797	.426766	.434064	.441752	.449879
.458470	.467489	.476775	.485833	.493152
.495000				

-396.140700	-121.195400	-93.799650	-80.642970	-71.826940
-64.983470	-59.214910	-54.091780	-49.377400	-44.877640
-40.540370	-36.320990	-32.184840	-28.104160	-24.058020
-20.032050	-16.017810	-12.009990	-8.005422	-4.002445
.000000				

EVAPORATION LOSSES = 5.094581 MM/DAY

TIME STEP = 4320 DURATION = 48.0000 HOURS

.075018	.076854	.080029	.083969	.088834
.094914	.102592	.112340	.124697	.412477
.419207	.426236	.433613	.441386	.449594
.458255	.467335	.476673	.485777	.493136
.495000				

-396.140700	-121.973300	-94.432570	-81.203700	-72.342020
-65.466160	-59.673200	-54.531120	-49.801910	-45.290020
-40.910630	-36.632130	-32.432740	-28.292580	-24.195880
-20.129850	-16.084810	-12.053690	-8.031530	-4.014567
.000000				

EVAPORATION LOSSES = 5.000982 MM/DAY

TIME STEP = 6480 DURATION = 72.0000 HOURS

.075018	.076823	.079944	.083815	.088594
.094567	.102110	.111691	.123845	.412067
.418804	.425857	.433271	.441089	.449347
.458060	.467189	.476573	.485721	.493119
.495000				

-396.140700	-122.489000	-94.852460	-81.575970	-72.684360
-65.787480	-59.978860	-54.824730	-50.086120	-45.566640
-41.164970	-36.855930	-32.621540	-28.445650	-24.315190
-20.219240	-16.148750	-12.096790	-8.057665	-4.026789
.000000				

EVAPORATION LOSSES = 4.940605 MM/DAY

TIME STEP = 8640 DURATION = 96.0000 HOURS

.075018	.076802	.079884	.083706	.088425
.094323	.101771	.111232	.123243	.411774
.418512	.425580	.433019	.440868	.449161
.457910	.467076	.476496	.485677	.493107
.495000				

-396.140700	-122.858100	-95.153080	-81.842580	-72.929640
-66.017780	-60.197960	-55.035210	-50.289890	-45.765010
-41.349380	-37.019540	-32.760890	-28.560120	-24.405570
-20.287630	-16.198080	-12.130170	-8.078013	-4.036304
.000000				

EVAPORATION LOSSES = 4.898285 MM/DAY

TIME STEP = 10800 DURATION = 120.0000 HOURS

.075018	.076786	.079840	.083627	.088303
.094145	.101524	.110899	.122805	.411558
.418299	.425378	.432834	.440706	.449025
.457801	.466993	.476439	.485645	.493097
.495000				

-396.140700	-123.129700	-95.374380	-82.038850	-73.110160
-66.187190	-60.359150	-55.190080	-50.439850	-45.911060
-41.484620	-37.139650	-32.863090	-28.643840	-24.471570
-20.337370	-16.234080	-12.154610	-8.092898	-4.043263
.000000				

EVAPORATION LOSSES = 4.867475 MM/DAY

TIME STEP = 12960 DURATION = 144.0000 HOURS

.075018	.076775	.079808	.083570	.088213
.094016	.101344	.110657	.122485	.411401
.418143	.425230	.432700	.440589	.448926
.457722	.466934	.476398	.485622	.493091
.495000				

-396.140700	-123.329100	-95.536930	-82.183060	-73.242860
-66.311830	-60.477710	-55.303980	-50.550130	-46.018410
-41.583930	-37.227390	-32.937770	-28.704790	-24.519630
-20.373780	-16.260250	-12.172320	-8.103634	-4.048285
.000000				

EVAPORATION LOSSES = 4.845006 MM/DAY

TIME STEP = 15120 DURATION = 168.0000 HOURS

.075018	.076767	.079785	.083528	.088148
.093922	.101213	.110480	.122252	.411285
.418028	.425121	.432600	.440502	.448853
.457663	.466889	.476367	.485605	.493086
.495000				

-396.140700	-123.475400	-95.656170	-82.288800	-73.340160
-66.403190	-60.564660	-55.387540	-50.631060	-46.097200
-41.656990	-37.292130	-32.992840	-28.750030	-24.555290
-20.400690	-16.279680	-12.185450	-8.111624	-4.052022
.000000				

EVAPORATION LOSSES = 4.828610 MM/DAY

TIME STEP = 17280 DURATION = 192.0000 HOURS

.075018	.076761	.079768	.083498	.088101
.093853	.101118	.110351	.122083	.411200
.417944	.425040	.432527	.440437	.448799
.457619	.466856	.476345	.485592	.493082
.495000				

-396.140700	-123.582900	-95.743800	-82.366550	-73.411640
-66.470250	-60.628410	-55.448740	-50.690270	-46.154790
-41.710670	-37.340240	-33.033880	-28.783470	-24.581620
-20.420680	-16.294060	-12.195140	-8.117532	-4.054773
.000000				

EVAPORATION LOSSES = 4.816775 MM/DAY

TIME STEP = 19440 DURATION = 216.0000 HOURS

.075018	.076756	.079756	.083475	.088065
.093801	.101046	.110253	.121954	.411136
.417881	.424981	.432474	.440391	.448760
.457588	.466832	.476328	.485583	.493079
.495000				

-396.140700	-123.663600	-95.809590	-82.424940	-73.465420
-66.520820	-60.676570	-55.495100	-50.735210	-46.198620
-41.750860	-37.375230	-33.063350	-28.807540	-24.600640
-20.435060	-16.304500	-12.202230	-8.121819	-4.056775
.000000				

EVAPORATION LOSSES = 4.807635 MM/DAY

TIME STEP = 21600 DURATION = 240.0000 HOURS

.075018	.076753	.079747	.083458	.088040
.093765	.100995	.110185	.121864	.411091
.417835	.424937	.432433	.440355	.448730
.457564	.466814	.476316	.485576	.493077
.495000				

-396.140700	-123.721200	-95.856570	-82.466580	-73.503690
-66.556690	-60.710690	-55.527810	-50.766830	-46.229370
-41.779980	-37.401550	-33.085810	-28.825950	-24.614840
-20.445880	-16.312330	-12.207520	-8.125056	-4.058288
.000000				

EVAPORATION LOSSES = 4.801403 MM/DAY

TIME STEP = 23760 DURATION = 264.0000 HOURS

.075018	.076751	.079740	.083447	.088022
.093739	.100959	.110136	.121800	.411059
.417804	.424909	.432408	.440333	.448711
.457549	.466803	.476308	.485572	.493076
.495000				

-396.140700	-123.761800	-95.889740	-82.495990	-73.530760
-66.582150	-60.734940	-55.551140	-50.789450	-46.251380
-41.799910	-37.418380	-33.099910	-28.837520	-24.624110
-20.452790	-16.317140	-12.210740	-8.126981	-4.059187
.000000				

EVAPORATION LOSSES = 4.796802 MM/DAY

TIME STEP = 25920 DURATION = 288.0000 HOURS

.075018	.076749	.079735	.083438	.088008
.093718	.100931	.110098	.121749	.411033
.417779	.424884	.432385	.440313	.448695
.457536	.466793	.476301	.485568	.493075
.495000				

-396.140700	-123.793900	-95.915850	-82.519180	-73.552150
-66.602260	-60.754030	-55.569480	-50.807250	-46.268760
-41.815480	-37.433000	-33.112730	-28.848000	-24.632130
-20.458840	-16.321560	-12.213730	-8.128796	-4.060034
.000000				

EVAPORATION LOSSES = 4.793239 MM/DAY

TIME STEP = 28080 DURATION = 312.0000 HOURS

.075018	.076747	.079731	.083431	.087997
.093703	.100909	.110068	.121711	.411014
.417760	.424867	.432369	.440300	.448684
.457528	.466788	.476298	.485565	.493074
.495000				

-396.140700	-123.818000	-95.935570	-82.536710	-73.568300
-66.617440	-60.768560	-55.583510	-50.820820	-46.281960
-41.828110	-37.443500	-33.121350	-28.854800	-24.637290
-20.462520	-16.324030	-12.215330	-8.129799	-4.060500
.000000				

EVAPORATION LOSSES = 4.790444 MM/DAY

TIME STEP = 30240 DURATION = 336.0000 HOURS

.075018	.076747	.079730	.083428	.087992
.093696	.100899	.110055	.121693	.411005
.417752	.424860	.432364	.440295	.448680
.457525	.466785	.476296	.485565	.493074
.495000				

-396.140700	-123.829700	-95.945110	-82.545180	-73.575970
-66.624620	-60.775350	-55.589960	-50.827010	-46.287950
-41.833180	-37.447560	-33.124570	-28.857330	-24.639210
-20.463910	-16.324980	-12.215930	-8.130152	-4.060676
.000000				

EVAPORATION LOSSES = 4.789237 MM/DAY

TIME STEP = 32400 DURATION = 360.0000 HOURS

.075018	.076747	.079730	.083428	.087992
.093696	.100899	.110055	.121693	.411005
.417752	.424860	.432364	.440295	.448680
.457525	.466785	.476296	.485565	.493074
.495000				

-396.140700	-123.829700	-95.945110	-82.545180	-73.575970
-66.624620	-60.775350	-55.589960	-50.827010	-46.287950
-41.833180	-37.447560	-33.124570	-28.857330	-24.639210
-20.463910	-16.324980	-12.215930	-8.130152	-4.060676
.000000				

EVAPORATION LOSSES = 4.789237 MM/DAY

DIRECTOR	:	S. M. SETH
DIVISIONAL HEAD	:	G. C. MISHRA
SCIENTIST	:	C. P. KUMAR
DRAWING STAFF	:	NARENDRA KUMAR