TR- 119

PREDICTION OF EVAPORATION LOSSES FROM SHALLOW WATER TABLE USING A NUMERICAL MODEL

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PREFACE

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 waters are 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. Among the important considerations in this regard is the necessity to limit the rate of capillary rise to the surface.

This report entitled 'Prediction of Evaporation Losses from Shallow Water Table using a Numerical Model' 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 bare soils with a high water table. The study has been carried out by Mr. Chandra Prakash Kumar, Scientist 'C' under the guidance of Dr. G.C.Mishra, Scientist 'F'.

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ABSTRACT

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.

The purpose of this study is to estimate the steady state evaporation rates from bare soils under conditions of high water table. A finite difference numerical scheme based upon the one-dimensional Richards equation has been employed to attain the steady state moisture profiles and estimate the evaporation rates under conditions of high water table. The procedure takes into account the relevant atmospheric factors and the soil's capability to conduct water. Field data required include soil water retention curves, water table depth, and a record of air temperature and air humidity. Results obtained with the method demonstrate how the soil water evaporation rates depend on water table depth and suction prevailing at the soil surface.

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. Under annual field crops, the soil surface may remain largely bare throughout the periods of tillage, planting, germination, and early seedling growth, periods in which evaporation can deplete the moisture of the surface soil and thus hamper the growth of young plants during their most vulnerable stage. Rapid drying of a seedbed can thwart germination and thus doom an entire crop from the start. The problem can also be acute in young orchards, where the soil surface is often kept bare continuously for several years, and in dryland farming in arid zones, where the land is regularly fallowed for several months to collect and conserve rainwater from one season to the next.

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 the present study, steady state evaporation rates from bare soils under high water table conditions have been estimated 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 the climatic factors.

2.0 REVIEW

2.1 Physical Conditions

Three conditions are necessary if the evaporation process from a given body is to persist. First, there must be a continual supply of heat to meet the latent heat requirement (which is about 590 cal/gm of water evaporated at $15^{\circ}C$). This heat can come from the body itself, thus causing it to cool, or as is more commonly the case, it can come from the outside in the form of radiated or advected energy. Second, the vapour pressure in the atmosphere over the evaporating body must remain lower than the vapour pressure at the surface of that body (i.e., there must be a vapour pressure gradient between the body and the atmosphere), and the vapour must be transported away, by diffusion or convection, or both. These two conditions - namely, supply of energy and removal of vapour - are generally external to the evaporating body and are influenced by meteorological factors such as air temperature, humidity, wind velocity and radiation, which together determine the atmospheric evaporativity (the maximal flux at which the atmosphere can vapourize water from a free water surface).

The third condition is that there be a continual supply of water from or through the interior of the body to the site of evaporation. This condition depends upon the content and potential of water in the body as well as upon its conductive properties, which together determine the maximal rate at which the body can transmit water to the evaporation site. Accordingly, the actual evaporation rate is determined either by external evaporativity or by the soil's own ability to deliver water, whichever is the lesser (and hence the limiting factor).

2.2 Capillary Rise from a Water Table

The rise of water in the soil from a free water surface (i.e., a water table) has been termed capillary rise. This term derives from the capillary model, which regards the soil as analogous to a bundle of capillary tubes, predominantly wide in the case of a sandy soil and narrow in the case of a clay soil. Accordingly, the equation relating the equilibrium height of capillary rise h_c to the radii of the pores is

$$h_c = \frac{2 \gamma \cos \alpha}{r \rho_w g} \qquad \ldots(1)$$

where γ is the surface tension, r the capillary radius, ρ_w the water density, g the gravitational acceleration, and α the wetting angle, normally (though not always justifiably) taken as zero. This equation predicts that water will rise higher, albeit less rapidly, in a clay than in a sand. However, soil pores are not capillary tubes of uniform or constant radii, and hence the height of capillary rise will differ in different pores. Above the water table, matric suction will generally increase with height. Consequently, the number of water filled pores, and hence wetness, will decrease in each soil as a function of height. The rate of capillary rise, i.e., the flux , generally decreases with time as the soil is wetted to greater height and as equilibrium is approached.

The wetting of an initially uniformly dry soil by upward capillary flow, illustrated in figure 1, is a rare occurrence in the field. In its initial stages, this process is similar to infiltration, except that it takes place in the opposite

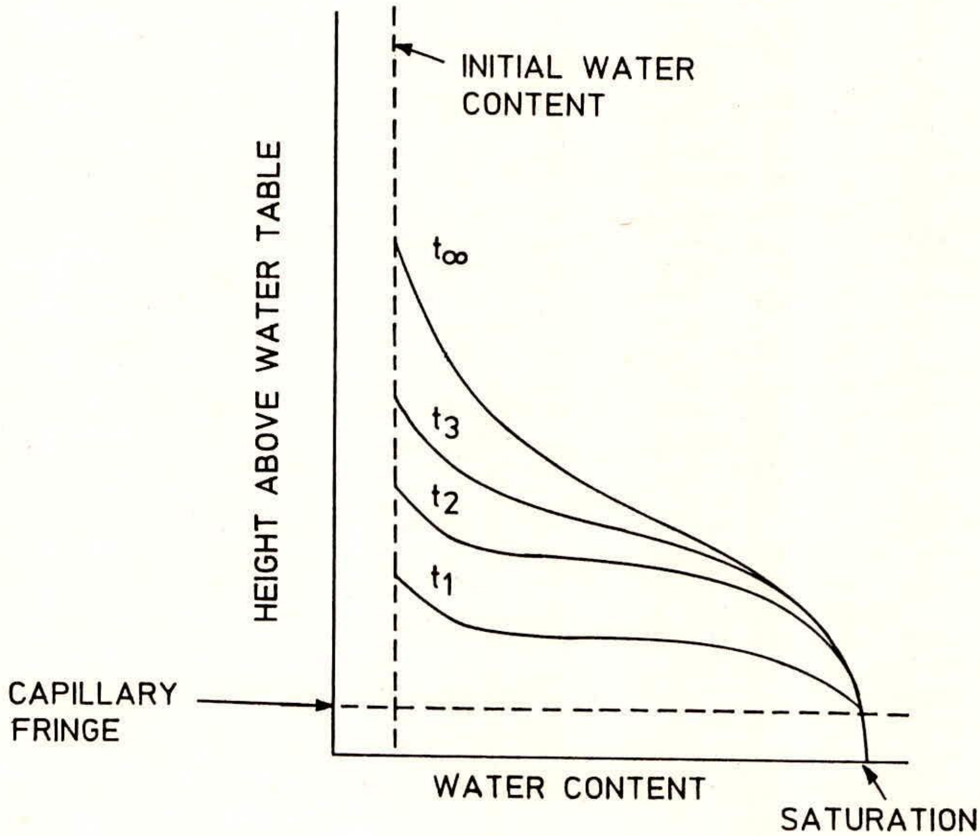

FIG. 1 - THE UPWARD INFILTRATION OF WATER FROM A WATER TABLE INTO A DRY SOIL

direction. At later stages of the process, the flux does not tend to a constant value, as in downward infiltration, but to zero. The reason is that the direction of the gravitational gradient is opposite to the direction of the matric suction gradient, and when the latter (which is large at first but decreases with time) approaches the magnitude of the former, the overall hydraulic gradient approaches zero.

Such an ideal state of static equilibrium between the gravitational head and the suction head is the exception rather than the rule under field conditions. In general, the condition of soil water is not static but dynamic - that is to say, constantly in a state of flux rather than at rest. Where a water table is present, soil water generally does not attain equilibrium even in the absence of vegetation, since the soil surface is subject to solar radiation and the evaporative demand of the ambient atmosphere. However, if soil and external conditions are constant, that is, if the soil is of stable structure, the water table is stationary, and atmospheric evaporativity also remains constant (at least approximately) - then, in time, a steady state flow situation can develop from water table to atmosphere via the soil. However, in the field the flow regime will at best be a quasi-steady state, since diurnal fluctuations and other perturbations will prevent attainment of truly stable flow conditions. Nevertheless, the representation of this process as a steady state flow is a useful approximation from the analytical point of view.

2.3 Steady Evaporation in the Presence of a Water Table

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

$$q = K(h) \left(\frac{dh}{dZ} - 1\right) \qquad \ldots(2)$$

or

$$q = D(\theta) \frac{d\theta}{dZ} - K(\theta) \qquad \ldots(3)$$

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) is

$$\frac{q}{K(h)} + 1 = \frac{dh}{dZ} \qquad \ldots(4)$$

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

$$Z = \int \frac{dh}{1 + \frac{q}{K(h)}} = \int \frac{K(h)}{K(h)+q}\, dh \qquad \ldots(5)$$

or

$$Z = \int \frac{D(\theta)}{K(\theta) + q}\, d\theta \qquad \ldots(6)$$

In order to perform the integration in equation (5), 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 (6) is to be integrated. An empirical equation for $K(h)$, given by Gardner (1958), is

$$K(h) = \frac{a}{h^n + b} \qquad \ldots(7)$$

where the parameters a, b, and n are constants which must be determined for each soil. Accordingly, equation (2) becomes

$$e = q = \frac{a}{h^n + b} \left(\frac{dh}{dZ} - 1\right) \qquad \ldots(8)$$

where e is the evaporation rate.

With equation (7), equation (5) 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 (7), Gardner (1958) obtained the function

$$q_{max} = \frac{Aa}{d^n} \quad \ldots(9)$$

where d is the depth of the water table below the soil surface, a and n are constants from equation (7), A is a constant which depends on n, and q_{max} 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 (9) 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:

$$e_{max} = \frac{1 + \dfrac{1.886}{(n^2 + 1)}}{d^n} \qquad \ldots(10)$$

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 (7) 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 (5) leads to

$$\int_0^L dZ + \int_L^{d+L} dZ = \int_{h_0}^{h_L} \frac{dh}{1+e/K_1(h)} + \int_{h_L}^{h_{L+d}} \frac{dh}{1+e/K_2(h)} \qquad \ldots(11)$$

where L and d are the thicknesses of the 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 indeed 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 field data required include soil moisture characteristic curves, water table depth, and 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 are still neglected.

3.0 PROBLEM DEFINITION

The direct evaporation of water from the soil surface often involves considerable losses of water and involves the hazard of salinizing the top layers of the soil. This danger 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 possibilities of reducing evaporation by studying the water flow patterns through the soil.

The objective of the present study is to determine the evaporation from shallow water tables through a homogeneous 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 profile is obtained for the given initial and boundary conditions and the steady state evaporation rate is 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 and suction at the soil surface (dictated by the atmospheric evaporativity) on the actual steady evaporation rate is therefore examined by varying depth of the water table, temperature and humidity of the air.

4.0 METHODOLOGY

4.1 General

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 an 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.

4.2 Soil Water 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 for 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

$$q = -K\left(\frac{\partial h}{\partial Z} - 1\right) \qquad (cm/h) \qquad \ldots(12)$$

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)

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial Z} \qquad (/h) \qquad \ldots(13)$$

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

Substitution of equation (12) into equation (13) yields the partial differential equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial Z}\left[K\left(\frac{\partial h}{\partial Z} - 1\right)\right] \qquad \ldots(14)$$

Equation (14) 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 only functions of the independent variables Z and t). 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

$$C = \frac{d\theta}{dh} \qquad (/cm) \qquad \ldots(15)$$

In equation (15) 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

$$\frac{\partial\theta}{\partial t} = \frac{d\theta}{dh} \cdot \frac{\partial h}{\partial t} \qquad \ldots(16)$$

and substituting equation (15) into equation (14) yields

$$C(h)\ \frac{\partial h}{\partial t} = \frac{\partial}{\partial Z}\ [\ K\ (h)\ (\ \frac{\partial h}{\partial Z} - 1)\ ] \qquad \ldots\ (17)$$

In equation (17) 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 (17) provides the basis for predicting soil water movement in layered soils of which each layer may have different physical properties.

4.3 Initial and Boundary Conditions

To obtain a solution for the one dimensional vertical flow equation, equation (17) 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

$$h\ (Z,\ t = 0) = h_o \qquad \ldots(18)$$

As hysteresis is not considered in this study, this condition is equivalent to

$$\theta\ (Z,\ t = 0) = \theta_o \qquad \ldots(19)$$

One can then 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

$$\left.\begin{array}{l} h\ (Z=0,\ t) = h_u \\ h\ (Z=L,\ t) = h_l \end{array}\right\} \qquad \ldots(20)$$

These conditions are equivalent to

$$\left.\begin{array}{l} \theta\ (Z=0,t) = \theta_u \\ \theta\ (Z=L,t) = \theta_l \end{array}\right\} \qquad \ldots(21)$$

(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

$$q(t) = - K(h) \left(-\frac{\partial h}{\partial Z} - 1 \right) \qquad \text{...(22)}$$

(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.

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):

$$h(0,t) = -\frac{RT(t)}{Mg} \ln [f(t)] \qquad \text{...(23)}$$

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.

For the present study, initial condition has been defined by equation (19) as

$$\theta(Z, t = 0) = 0.10 \qquad \text{...(24)}$$

and the upper boundary condition has been obtained by equation (23). The phreatic surface acts as lower boundary of the system in case of a shallow ground water table. The lower boundary condition has therefore been set as (Dirichlet type, equation 21):

$$\theta\ (Z = L,\ t) = \theta_s \qquad \ldots(25)$$

where L is the depth of the ground water table and the subscript s denotes saturated condition.

4.4 Soil Moisture Characteristics

For the present study, functional relations, as reported by Haverkamp et al.(1977), for characterizing the hydraulic properties of a soil, 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 an infiltration experiment carried out in the laboratory. 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 of 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 $\theta = 0.10$ 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:

$$K = K_s \frac{A}{A + |h|^{\beta_1}} ; \qquad \ldots(26)$$

$$K_s = 34 \text{ cm/h},$$

$$A = 1.175 \times 10^6,$$

$$\beta_1 = 4.74.$$

and

$$\theta = \frac{\alpha(\theta_s - \theta_r)}{\alpha + |h|^{\beta_2}} + \theta_r ; \qquad \ldots(27)$$

$$\theta_s = 0.287,$$

$$\theta_r = 0.075,$$

$$\alpha = 1.611 \times 10^6,$$

$$\beta_2 = 3.96.$$

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

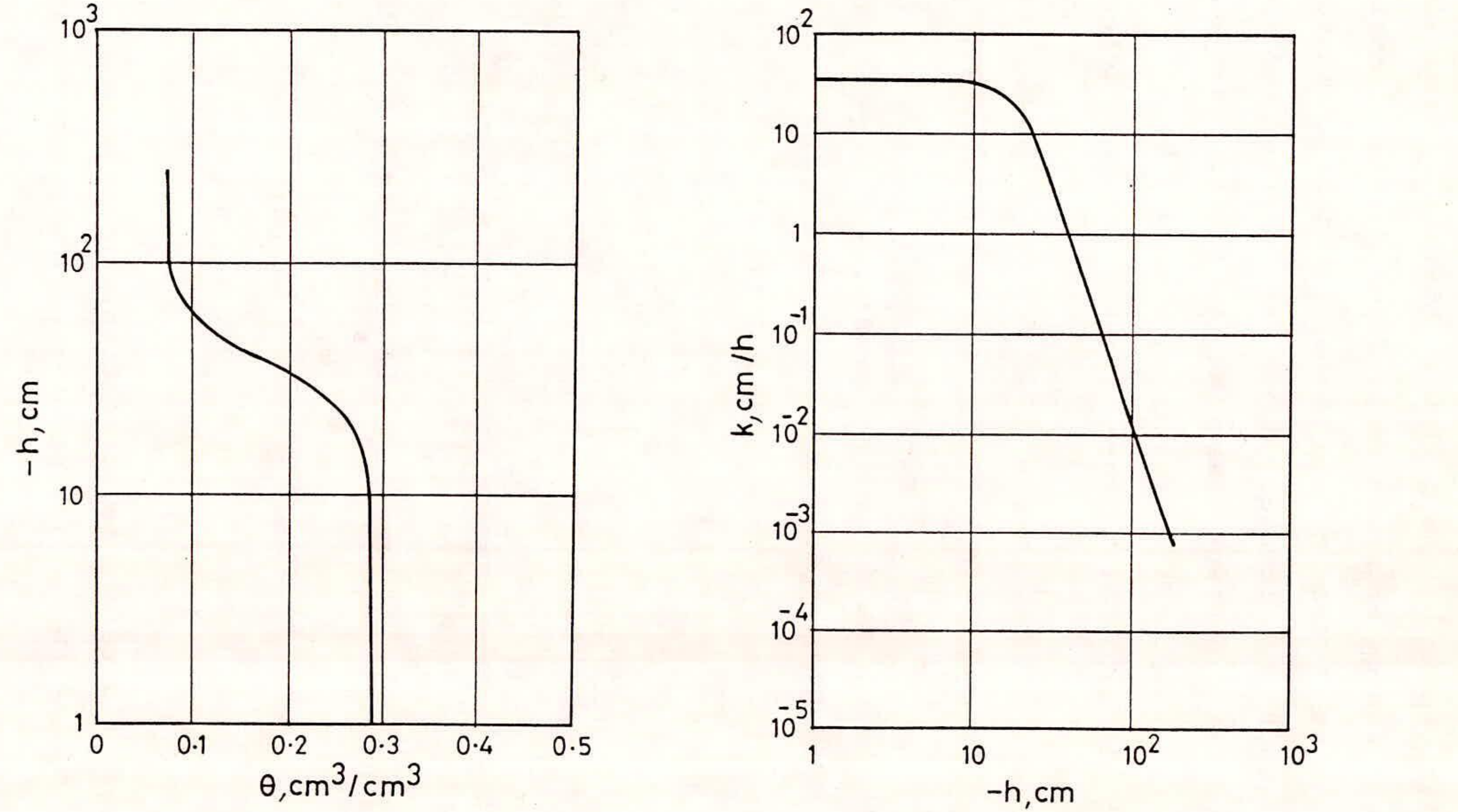

FIG.2- RELATIONSHIPS BETWEEN THE SOIL WATER PRESSURE h, THE WATER CONTENT θ AND THE HYDRAULIC CONDUCTIVITY K FOR THE SOIL USED IN THE STUDY

Figure 2 present the relationships between the soil water pressure h, the water content θ and the hydraulic conductivity K for the above soil used in this study.

4.5 Finite Difference Approximation

Equation (17) 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 ($t+1/2\ \Delta t$) using a method described by Douglas and Jones, 1963 (implicit linearization).

Let us now solve equation (17) 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

$$\frac{C}{K} \frac{\partial h}{\partial t} = \frac{\partial^2 h}{\partial Z^2} + \frac{1}{K} \frac{\partial K}{\partial Z} \left(\frac{\partial h}{\partial Z} - 1 \right) \qquad \ldots (28)$$

Using implicit evaluation of the coefficients at time $(t+1/2\ \Delta t)$, that is values for K and C are obtained at time $(t+1/2\ \Delta t)$, then pressure distribution is evaluated at time $(t+\Delta t)$. 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 (28), by taking time step as $\Delta t/2$, 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

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

$$= \frac{2C_i^j}{K_i^j}\ h_i^j + \frac{1}{2}\ \frac{K_{i+1}^j - K_{i-1}^j}{K_i^j}\ \frac{\Delta t}{\Delta Z} \left[\frac{h_{i+1}^j - h_{i-1}^j}{2\Delta Z} - 1 \right] \quad \ldots(29)$$

Correction (estimation of h_i^j)

From equation (28), 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

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

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

$$+ \frac{1}{2} \frac{K_{i+1}^{j+1/2} - K_{i-1}^{j+1/2}}{K_i^{j+1/2}} \frac{\Delta t}{\Delta Z} \left[\frac{h_{i+1}^{j+1/2} - h_{i-1}^{j+1/2}}{2\Delta Z} - 1 \right] \qquad \text{...(30)}$$

When equation (29) or (30) 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).

Steady state evaporation rates were estimated by applying equation (2) for two vertically adjacent nodal points after obtaining the equilibrium moisture profile.

$$q = K^{j}_{i+1/2} \left(\frac{h^{j}_{i+1} - h^{j}_{i}}{\Delta Z} - 1 \right) \qquad \ldots(31)$$

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 adjacent nodal points. However, small variations may be observed due to values of K being not properly represented over the depth interval chosen. Therefore, arithmetic mean was taken by considering all nodes in sequence. Various sets of water table depths, air temperature and air humidity were considered for the study.

The computer code, for discretization scheme used in the model and estimation of steady state evaporation rates as per the procedure described above, has been written in FORTRAN IV and presented in appendix.

5.0 RESULTS

5.1 General

The actual evaporation rate is governed by the atmospheric conditions, transmitting properties of the soil 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.

For the given external evaporative conditions and the water table depth, the equilibrium moisture profile was obtained by using the numerical scheme presented in section 4.5 and assuming the pressure head at the soil surface to be in equilibrium with the surrounding atmosphere. The initial and boundary conditions were defined by the equations (24), (23), and (25) respectively. The rate of loss of water (Darcian flux q) served as a measure of the evaporation rate once the steady state was attained. The evaporative conditions were varied by varying the temperature and humidity of air. The evaporation rates under each set of external conditions and water table depth were evaluated.

Since the steady state evaporation rates are estimated by considering the two vertically adjacent nodal points (equation 31), the size of the depth interval plays an important role. It was found that the numerical scheme is stable only when

$$\frac{\Delta t}{(\Delta Z)^2} \leq 2.5 \qquad \ldots(32)$$

where Δt is the time step (seconds) and ΔZ is the depth interval (cm). At greater water table depths, matrix dimensions (number of nodes, number of time steps) may become quite large (insufficient virtual address space to complete the link). Also it takes more time to attain the equilibrium moisture profile in case of a deeper water table. Therefore, different sets of ΔZ and Δt had to be used for different water table depths, for condition of stability (equation 32).

5.2 Effect of Depth to Water Table

In order to study the relation between evaporation rate and depth to the water table, solutions were obtained by varying the water table depth and keeping the evaporative conditions same (temperature = $25^{o}C$, relative humidity = 0.75). Ample time was allowed for steady state to be attained at each depth. 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. Table 1 presents the data on depth to water table, depth interval, time step and the estimated evaporation rates. The results are also shown graphically in figure 3, where the steady state evaporation rates are plotted against the depth to the water table.

It can be expected that for a particular soil and meteorological condition, the evaporation rate remains essentially

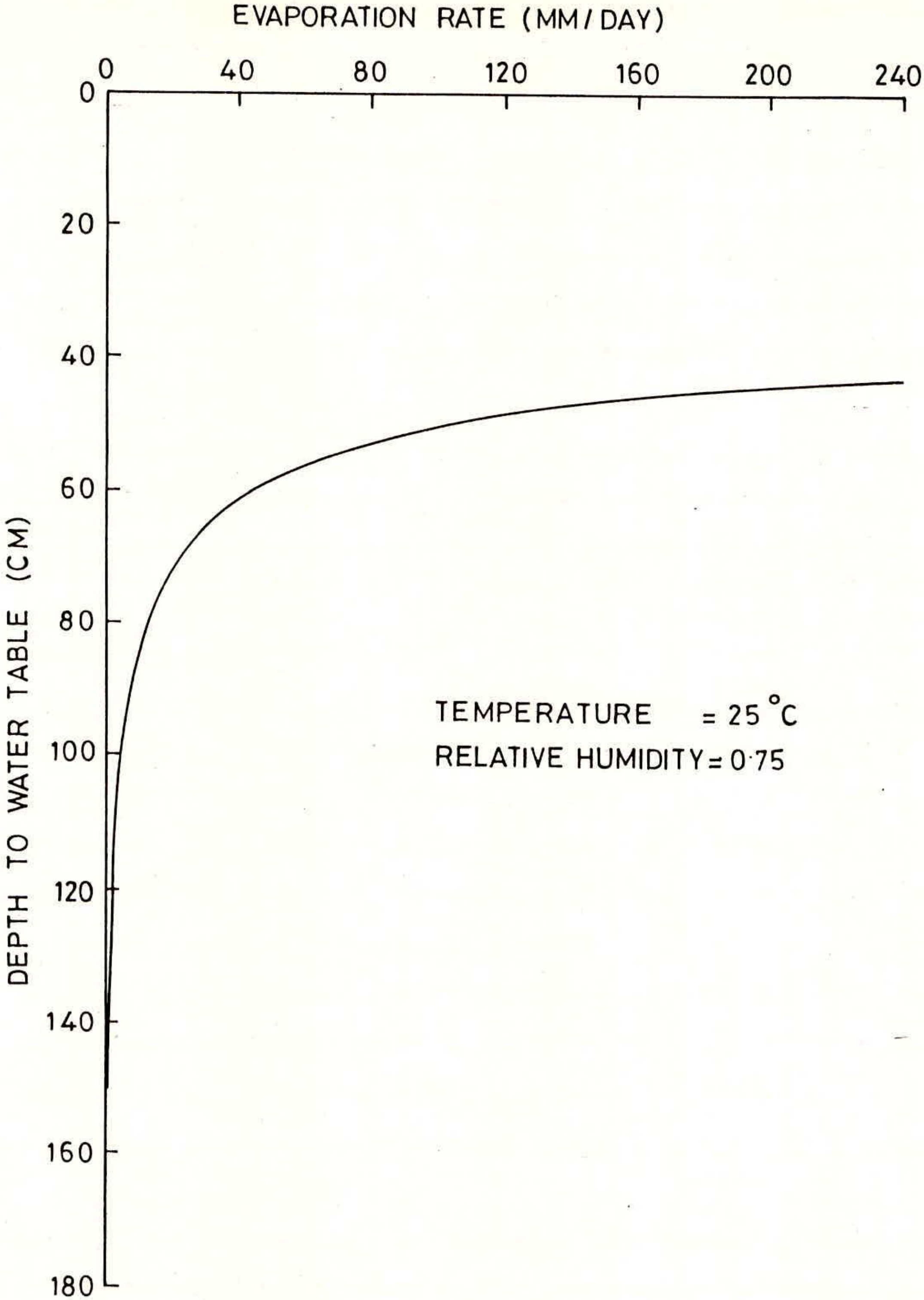

FIG. 3 - RELATION BETWEEN EVAPORATION RATE AND DEPTH TO WATER TABLE

constant and fixed by weather, if the water table depth does not exceed a certain value. With the water table at greater depths, the evaporative flux decreases markedly because the soil becomes the limiting factor. Figure 3 shows that under the conditions studied, evaporation was soil limited. This figure demonstrates that as the water table is lowered from 45 cm to 80 cm, the evaporation rate decreases markedly with depth. However, further lowering reduces the evaporation rate only slightly.

Table 1 - Steady State Evaporation Rates as a Function of Water Table Depth ($T = 25^{o}C$, $f = 0.75$, $h(0,t) = -396.14$)

S.No.	Depth to Water Table, L (cm)	Depth Interval, ΔZ (cm)	Time Step, Δt (sec)	$\frac{\Delta t}{(\Delta Z)^2}$	Number of Vertical Nodes	Evaporation Rate (mm/day)
1.	45	3	20	2.22	16	178.61
2.	60	4	40	2.50	16	44.36
3.	80	4	40	2.50	21	11.53
4.	100	4	40	2.50	26	4.03
5.	120	4	40	2.50	31	1.70
6.	150	5	60	2.40	31	0.56
7.	180	6	90	2.50	31	0.22
8.	210	6	90	2.50	36	0.10

Upward movement and evaporation of water is possible with the water table as deep as 150 cm and although the rate will be slow, accumulation of harmful amounts of soluble salts is possible if the ground water is sufficiently saline and if sufficient time is allowed. In such a case it may be feasible either lowering of the water table or periodically leaching out the salts by the application of excess water at the surface.

5.3 Effect of Suction Head at the Soil Surface

The variation of steady state evaporation rate with suction head at the soil surface was studied by varying the temperature and humidity values. All estimations were made at water table depth of 100 cm. Table 2 shows estimated values of steady state evaporation rates for different external evaporative conditions. It may be seen that relative humidity has more pronounced effect on evaporation rate than that of temperature. In figure 4, evaporation rates are plotted as a function of the suction at the soil surface, assuming a depth to water table of 100 cm. The evaporation rates do seem to approach definite maxima.

When the evaporation rate is low and is limited by external conditions, a large increase in the evaporation rate causes only a small increase in the suction at the soil surface. Evaporation under such conditions is virtually independent of the depth to water table and the capillary conductivity of the soil. The range of external conditions for which this is the case depends upon the depth to the water table. The shallower the water table the greater the range over which evaporation is controlled by external conditions.

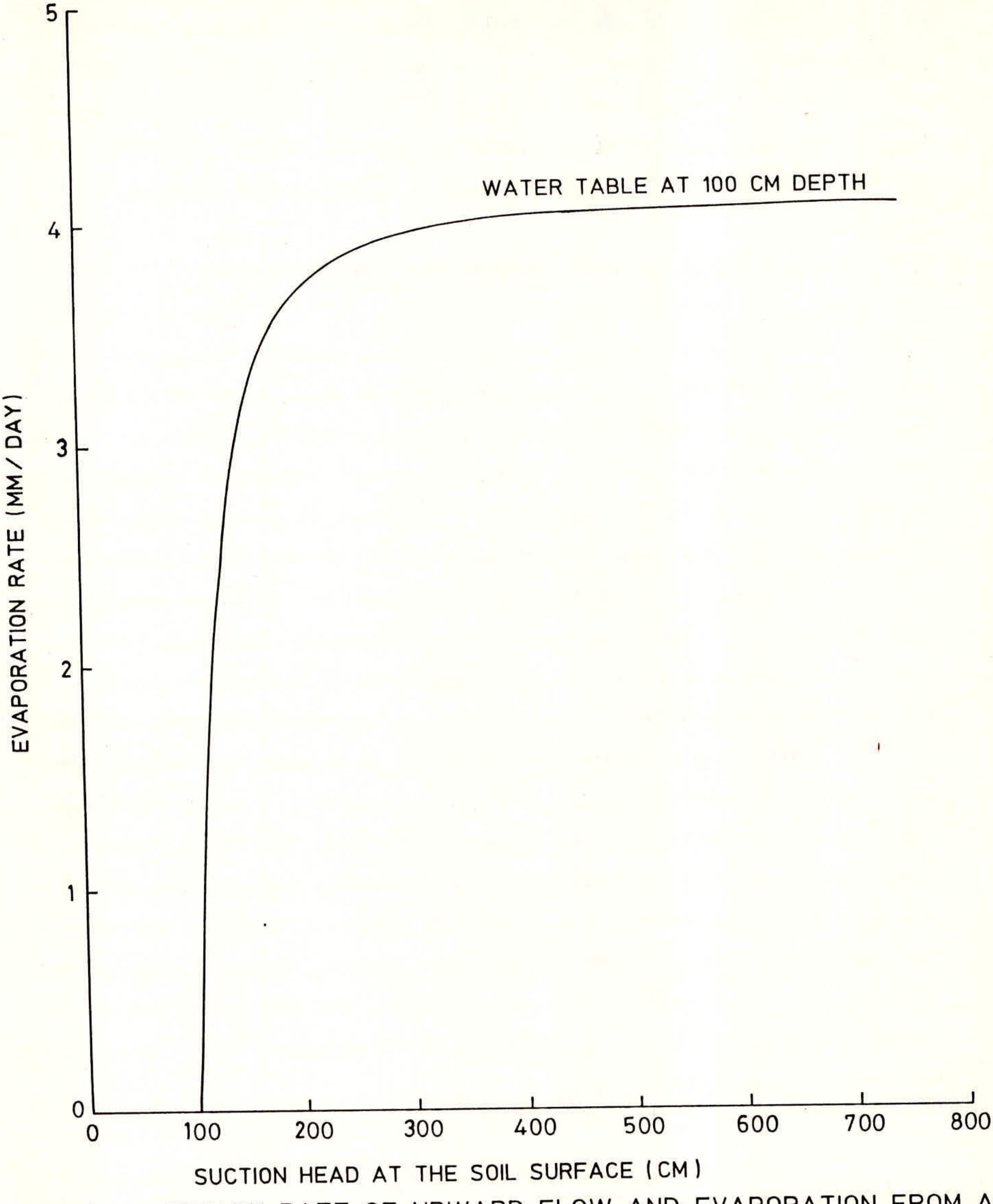

FIG. 4 -. STEADY RATE OF UPWARD FLOW AND EVAPORATION FROM A WATER TABLE AS FUNCTION OF THE SUCTION PREVAILING AT THE SOIL SURFACE

Table 2 - Steady State Evaporation Rates as a Function of Suction Head at the Soil Surface (L = 100 cm)

S.No.	Temperature, T (°C)	Relative Humidity, f	Suction Head at the Soil Surface, -h(0,t) (cm)	Evaporation Rate (mm/day)
1.	10	0.75	376.21	4.021
2.	15	0.75	382.85	4.024
3.	20	0.75	389.50	4.027
4.	25	0.75	396.14	4.031
5.	30	0.75	402.78	4.033
6.	35	0.75	409.43	4.037
7.	40	0.75	416.07	4.039
8.	25	0.60	703.41	4.084
9.	25	0.65	593.19	4.075
10.	25	0.70	491.14	4.060
11.	25	0.80	307.27	3.967
12.	25	0.85	223.79	3.800
13.	25	0.90	145.08	3.023
14.	25	0.91	129.87	2.537
15.	25	0.92	114.82	1.671
16.	25	0.929	101.41	0.213

5.4 Soil-Limited Evaporation

It is clear from physical considerations that an increase in the evaporative capacity of the atmosphere will produce an increased suction at the soil surface. This higher suction, in turn, must magnify the upward water flux through the soil. Such a flux can not increase without bound, rather a limiting soil-water flux and hence a soil-limited evaporation, E_∞, exists. For any particular soil system, the latter is approximately given by (Ripple et al.,1972)

$$E_\infty \approx K_s \left[\frac{|h_{1/2}|}{L} \cdot \frac{\pi}{n \sin \frac{\pi}{n}} \right]^n \qquad \ldots(32)$$

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 the form

$$K = \frac{a}{b + |h|^n} \qquad \ldots(33)$$

(a,b, and n are constants)

Equation (32) 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.

It may be noted from figure 4 that the curve approaches a limiting rate of evaporation with increasing water suction at the soil surface as expected. The rate of approach to the actual E_∞ mainly depends on the value of n characterizing the particular soil. It should be noted that most of the field soils commonly found show n values which lie between 2 and 5.

The value of E_∞ for the soil system under consideration is estimated for L = 100 cm, as below:

Comparing equation (26) and (33),

$$n = \beta_1 = 4.74$$

Also from equation (26),

$$\frac{K}{K_s} = \frac{A}{A + |h|^{\beta_1}}$$

or

$$\frac{1}{2} = \frac{A}{A + |h_{1/2}|^{\beta_1}}$$

or $$|h_{1/2}|^n = A = 1.175 \times 10^6$$

Also $$K_s = 34 \text{ cm/h} = 816 \text{ cm/day}$$

So, from equation (32), we get

$$E_\infty \approx 816 \times 1.175 \times 10^6 \left[\frac{1}{100} \times \frac{\pi}{4.74 \sin \frac{\pi}{4.74}} \right]^{4.74}$$

or $$E_\infty \approx 0.4516 \text{ cm/day}$$

or $$E_\infty \approx 4.516 \text{ mm/day}$$

This value of E_{∞}, as compared with figure 4, support the applicability of the equation (32) to the steady state evaporation problem.

5.5 Remarks

The dependence of the actual steady state evaporation rate on water table depth and weather (as demonstrated in figures 3 and 4), which can be computed with the aid of the approach presented in the preceding pages, might be most useful in hydrologic practice. The extent to which the above results can be applied quantitatively to the field depends upon the correspondence between capillary conductivity values and those existing in the field. The soil data employed might be less precise than desirable. In addition, it might be impossible to take into account adequately the variability of field soils.

Steady state conditions were assumed throughout this study. In nature, however, the systems considered are seldom in such a state, principally because of the variations in meteorological conditions, in soil salt content, and in 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 very 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

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 to 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.

A numerical model study has been carried out to examine the steady state evaporation from shallow water table through a soil. The evaporation rate is shown to be related to the depth to water table for a particular soil. The influence of external evaporativity is also considered.

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APPENDIX

```
C        SOIL MOISTURE PREDICTION MODEL
C
C
C        PREDICTION OF EVAPORATION LOSSES FROM SHALLOW WATER TABLE
C        USING A NUMERICAL MODEL
C
C        IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION (PREDICTION - CORRECTION)
C        (MODEL 4 OF HAVERKAMP ET AL., 1977)
C
         DIMENSION SUB(32),SUP(32),DIAG(32),B(32)
         DIMENSION H(32,5410),CCC(32,5410)
         DIMENSION THETA(32,5410),HYDCON(32,5410)
         DIMENSION HP(32,5410),THETAP(32,5410)
         OPEN(UNIT=1,FILE='EVAP4.DAT',STATUS='OLD')
         OPEN(UNIT=2,FILE='EVAP4.OUT',STATUS='NEW')
C
C        J REFERS TO TIME
C        I REFERS TO DEPTH
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 = RESIDUAL MOISTURE CONTENT
C        THETAS = MOISTURE CONTENT AT SATURATION
C        BETA1, CONA =  PARAMETERS IN THE HYDRAULIC CONDUCTIVITY
C                       AND SOIL WATER PRESSURE RELATIONSHIP
C        BETA2, ALPHA = PARAMETERS IN THE MOISTURE CONTENT AND
C                       SOIL WATER PRESSURE RELATIONSHIP
C        HYDCON = HYDRAULIC CONDUCTIVITY OF THE SOIL (CM/HOUR)
C        AKS = HYDRAULIC CONDUCTIVITY AT SATURATION (CM/HOUR)
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
         READ(1,11)THETAR,THETAS
11       FORMAT(3F12.3)
         READ(1,12)BETA1,BETA2
12       FORMAT(2F12.3)
         READ(1,13)CONA,ALPHA
13       FORMAT(2F12.3)
         READ(1,14)AKS
14       FORMAT(F12.3)
         READ(1,15)DELT,DELZ
15       FORMAT(F12.8,F12.3)
         READ(1,16)NTIME,NNODE
16       FORMAT(I4,6X,I4)
         READ(1,17)T
17       FORMAT(F5.2)
         READ(1,18)RH
18       FORMAT(F5.2)
C
```

```
C       READING OF INITIAL CONDITIONS
C
        READ(1,19)(THETA(I,1),I=1,NNODE)
 19     FORMAT(5F10.4)
C
        WRITE(2,20)
 20     FORMAT(2X,'* EVAPORATION LOSSES FROM SHALLOW WATER TABLE')
        WRITE(2,21)
 21     FORMAT(2X,'** ONE DIMENSIONAL RICHARDS EQUATION')
        WRITE(2,22)
 22     FORMAT(2X,'*** IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION')
        WRITE(2,23)
 23     FORMAT(/2X,'TEMPERATURE IN CENTIGRADE')
        WRITE(2,24)T
 24     FORMAT(F7.2)
        WRITE(2,25)
 25     FORMAT(2X,'RELATIVE HUMIDITY OF THE AIR')
        WRITE(2,26)RH
 26     FORMAT(F7.3)
        WRITE(2,27)
 27     FORMAT(2X,'THETAR',9X,'THETAS')
        WRITE(2,28)THETAR,THETAS
 28     FORMAT(2X,F5.3,10X,F5.3,10X,F5.3)
        WRITE(2,29)
 29     FORMAT(2X,'BETA1',10X,'BETA2')
        WRITE(2,30)BETA1,BETA2
 30     FORMAT(2X,F5.3,10X,F5.3)
        WRITE(2,31)
 31     FORMAT(2X,'CONA',11X,'ALPHA')
        WRITE(2,32)CONA,ALPHA
 32     FORMAT(2X,F11.3,4X,F11.3)
        WRITE(2,33)
 33     FORMAT(2X,'AKS')
        WRITE(2,34)AKS
 34     FORMAT(2X,F6.3)
        WRITE(2,35)
 35     FORMAT(2X,'DELT',11X,'DELZ')
        WRITE(2,36)DELT,DELZ
 36     FORMAT(2X,F10.8,5X,F5.3)
        WRITE(2,37)
 37     FORMAT(2X,'NTIME',10X,'NNODE')
        WRITE(2,38)NTIME,NNODE
 38     FORMAT(2X,I4,9X,I4)
        WRITE(2,39)
 39     FORMAT(/2X,'INITIAL CONDITIONS'/)
        WRITE(2,40)(THETA(I,1),I=1,NNODE)
 40     FORMAT(5F10.4)
        WRITE(2,41)
 41     FORMAT(/2X,'WATER CONTENT AND SOIL WATER PRESSURE
     1  AT DIFFERENT NODES')
C
        DO 100 I=1,NNODE
        H(I,1)=-(ALPHA*(THETAS-THETA(I,1))/(THETA(I,1)
     1  -THETAR))**(1./BETA2)
100     CONTINUE
C
```

```
C        GENERATION OF UPPER BOUNDARY CONDITION
C
         DO 200 J = 2, NTIME
         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,J)=HU
         THETA(1,J)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(1,J))**BETA2)+
     1   THETAR
         HP(1,J)=H(1,J)
         THETAP(1,J)=THETA(1,J)
 200     CONTINUE
C
C        GENERATION OF LOWER BOUNDARY CONDITION
C
         DO 300 J=1,NTIME
         THETA(NNODE,J)=THETA(NNODE,1)
         THETAP(NNODE,J)=THETA(NNODE,1)
         H(NNODE,J)=-(ALPHA*(THETAS-THETA(NNODE,J))/(THETA(NNODE,J)
     1   -THETAR))**(1./BETA2)
         HP(NNODE,J)=H(NNODE,J)
 300     CONTINUE
C
         E1=BETA1/BETA2
         E2=(THETAS-THETAR)
         E3=ALPHA**E1
         E4=CONA*AKS
         E5=1./BETA2*ALPHA**(1./BETA2)
C
         DO 400 J=2,NTIME
C
         DO 500 I=1,NNODE
C        IF(THETA(I,J-1).LT.(THETAR+0.005))THETA(I,J-1)=THETAR+0.005
C        TERM1=(THETA(I,J-1)-THETAR)/E2
C        HYDCON(I,J-1)=E4*TERM1**E1/(CONA*TERM1**E1+E3*((1.-TERM1))**E1)
         HYDCON(I,J-1)=E4/(CONA+(ABS(H(I,J-1)))**BETA1)
         CCC(I,J-1)=1./(E5*E2)*(THETAS-THETA(I,J-1) )**(-1./BETA2+1.)*
     1   ( THETA(I,J-1)-THETAR ) **(1./BETA2+1.)
 500     CONTINUE
C
         DO 600 I=2,NNODE-1
         DIAG(I-1)=2.*CCC(I,J-1)/HYDCON(I,J-1)+2.*DELT/DELZ**2
         SUB(I-1)=-DELT/DELZ**2
         SUP(I-1)=-DELT/DELZ**2
         B(I-1)=2.*CCC(I,J-1)/HYDCON(I,J-1)*H(I,J-1)+DELT/DELZ*.5
     1   *(HYDCON(I+1,J-1)-HYDCON(I-1,J-1))/HYDCON(I,J-1)*((H(I+1,J-1)-
     2   H(I-1,J-1))/(2.*DELZ)-1.)
 600     CONTINUE
C
         B(1)=B(1)-SUB(1)*H(1,J)
         B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,J)
         DO 700 I=1,NNODE-3
 700     SUB(I)=SUB(I+1)
         M=NNODE-2
         CALL TRID(M,SUP,SUB,DIAG,B)
         DO 800 I=1,NNODE-2
 800     HP(I+1,J)=B(I)
         DO 900 I=2,NNODE-1
         THETAP(I,J)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(HP(I,J))**
     1   BETA2)+THETAR
 900     CONTINUE
C
```

```
      DO 1000 I=1,NNODE
C     IF(THETAP(I,J).LT.(THETAR+0.005))THETAP(I,J)=THETAR+0.005
C     TERM1=(THETAP(I,J)-THETAR)/E2
C     HYDCON(I,J-1)=E4*TERM1**E1/(CONA*TERM1**E1+E3*((1.-TERM1))**E1)
      HYDCON(I,J-1)=E4/(CONA+(ABS(HP(I,J)))**BETA1)
      CCC(I,J-1)=1./(E5*E2)*(THETAS-THETAP(I,J) )**(-1./BETA2+1.)*
     1 ( THETAP(I,J)-THETAR ) **(1./BETA2+1.)
1000  CONTINUE
C
      DO 1100 I=2,NNODE-1
      DIAG(I-1)=CCC(I,J-1)/HYDCON(I,J-1)+DELT/DELZ**2
      SUB(I-1)=-DELT/DELZ**2*.5
      SUP(I-1)=-DELT/DELZ**2*.5
      B(I-1)=CCC(I,J-1)/HYDCON(I,J-1)*H(I,J-1)+DELT/DELZ*.5
     1 *(HYDCON(I+1,J-1)-HYDCON(I-1,J-1))/HYDCON(I,J-1)*((HP(I+1,J)-
     2 HP(I-1,J))/(2.*DELZ)-1.)+DELT/DELZ**2*.5*(H(I+1,J-1)-2.*
     3 H(I,J-1)+H(I-1,J-1))
1100  CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,J)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,J)
      DO 1200 I=1,NNODE-3
1200  SUB(I)=SUB(I+1)
      M=NNODE-2
      CALL TRID(M,SUP,SUB,DIAG,B)
      DO 1300 I=1,NNODE-2
1300  H(I+1,J)=B(I)
C
      DO 1400 I=2,NNODE-1
      THETA(I,J)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,J))**BETA2)+
     1 THETAR
1400  CONTINUE
C
      IF (J.EQ.90)   GO TO 111
      IF (J.EQ.180)  GO TO 111
      IF (J.EQ.270)  GO TO 111
      IF (J.EQ.360)  GO TO 111
      IF (J.EQ.450)  GO TO 111
      IF (J.EQ.540)  GO TO 111
      IF (J.EQ.1080) GO TO 111
      IF (J.EQ.1620) GO TO 111
      IF (J.EQ.2160) GO TO 111
      IF (J.EQ.2700) GO TO 111
      IF (J.EQ.3240) GO TO 111
      IF (J.EQ.3780) GO TO 111
      IF (J.EQ.4320) GO TO 111
      IF (J.EQ.4860) GO TO 111
      IF (J.EQ.5400) GO TO 111
      GO TO 400
 111  CONTINUE
C
      DO 666 I=1,NNODE
      HYDCON(I,J)=E4/(CONA+(ABS(H(I,J)))**BETA1)
 666  CONTINUE
C
      EVAP = 0.0
      DO 1500 N = 2,NNODE
      EVAP=EVAP+((HYDCON(N,J)*HYDCON(N-1,J))**0.5)*
     1 (((H(N,J)-H(N-1,J))/DELZ)-1.0)*240.0
1500  CONTINUE
      EVAP = EVAP/(NNODE-1)
C
```

```
      HOUR=J*DELT
      WRITE(2,51)J,HOUR
51    FORMAT(/2X,'TIME STEP = ',I4,4X,'DURATION  = ',F6.2,5X,'HOURS'/)
      WRITE(2,52)(THETA(I,J),I=1,NNODE)
52    FORMAT(5F10.4)
      WRITE(2,*)
      WRITE(2,333)(H(I,J),I=1,NNODE)
333   FORMAT(5F10.4)
      IF (J.EQ.5400) GO TO 444
      GO TO 555
444   WRITE(2,*)
      WRITE(2,777)(HYDCON(I,J),I=1,NNODE)
777   FORMAT (5F10.6)
555   CONTINUE
      WRITE(2,53)EVAP
53    FORMAT(/2X,'EVAPORATION LOSSES     =     ',F10.4,'     MM/DAY'/)
400   CONTINUE
      STOP
      END
C
      SUBROUTINE TRID(M,SUP,SUB,DIAG,B)
      DIMENSION SUP(32),SUB(32),DIAG(32),B(32)
      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
```

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