TR - 142

ESTIMATION OF GROUND WATER RECHARGE DUE TO RAINFALL BY MODELLING OF SOIL MOISTURE MOVEMENT

आपो हि ष्ठा मयोभुवः

NATIONAL INSTITUTE OF HYDROLOGY
JAL VIGYAN BHAVAN
ROORKEE (UP)-247 667

1992-93

PREFACE

Quantification of the rate of natural ground water recharge is a basic pre-requisite for efficient ground water resource management. It is particularly important in regions with large demands for ground water supplies, where such resources are the key to economic development. However, the rate of aquifer recharge is one of the most difficult factors to measure in the evaluation of ground water resources. The main techniques used to estimate ground water recharge rates are the Darcian approach, the soil water balance approach and the ground water level fluctuation approach. Estimation of recharge, by whatever method, are normally subject to large uncertainties and errors. A reappraisal of the recharge process suggests a more realistic model.

This report entitled 'Estimation of Ground Water Recharge due to Rainfall by Modelling of Soil Moisture Movement' is a part of the research activities of 'Ground Water Assessment' division of the Institute. The purpose of this study is to estimate the ground water recharge due to rainfall by solving numerically the partial differential equation of downward moisture flow in unsaturated soils. The study has been carried out by Mr. Chandra Prakash Kumar, Scientist 'C' under the guidance of Dr. G. C. Mishra, Scientist 'F'.

(S. M. Seth)

Director

CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

The amount of water that may be extracted from an aquifer without causing depletion is primarily dependent upon the ground water recharge. Thus, a quantitative evaluation of spatial and temporal distribution of ground water recharge is a pre-requisite for operating ground water resources system in an optimal manner.

Rainfall is the principal means for replenishment of moisture in the soil water system and recharge to ground water. Moisture movement in the unsaturated zone is controlled by capillary pressure and hydraulic conductivity. The amount of moisture that will eventually reach the water table is defined as natural ground water recharge. The amount of this recharge depends upon the rate and duration of rainfall, the subsequent conditions at the upper boundary, the antecedent soil moisture conditions, the water table depth and the soil type.

The purpose of this study is to estimate the ground water recharge due to rainfall by studying one-dimensional vertical flow of water in the unsaturated zone. A model has been formulated for finite difference solution of the non-linear Richards equation applicable to transient, one-dimensional water flow through the unsaturated porous medium. Implicit scheme with implicit linearization (prediction-correction) has been used for discretization. The ground water recharge has been estimated using appropriate initial and boundary conditions for storm and interstorm periods.

1.0 INTRODUCTION

In many arid and semi-arid regions, surface water resources are limited and ground water is the major source for agricultural, industrial and domestic water supplies. Because of lowering of water tables and the consequently increased energy costs for pumping , it is recognized that ground water extraction should balance ground water recharge in areas with scarce fresh water supplies. This objective can be achieved either by restricting ground water use to the water volume which becomes available through the process of natural recharge or by recharging the aquifer artificially with surface water. Both options require knowledge of the ground water recharge process through the unsaturated zone from the land surface to the regional water table.

When water is supplied to the soil surface, whether by precipitation or irrigation, some of the arriving water penetrates the surface and is absorbed into the soil, while some may fail to penetrate but instead accrue at the surface or flow over it. The water which does penetrate is itself later partitioned between that amount which returns to the atmosphere by evapotranspiration and that which seeps downward, with some of the latter reemerging as stream flow while the remainder recharges the ground water reservoir.

When rain intensity exceeds soil infiltrability, in principle the infiltration process is similar to the case of shallow ponding. If rain intensity is less than the initial infiltrability value of the soil but greater than the final value, then at first the soil will absorb at less than its potential rate and the flow of water in the soil will occur under unsaturated conditions; however, if the rain is continued at the same intensity, and as the soil infiltrability decreases, the soil surface will eventually become saturated and henceforth the process will continue as in the case of ponding infiltration. Finally, if rain intensity is at all times lower than soil infiltrability (i.e., lower than the effective saturated hydraulic conductivity), the soil will continue to absorb the water as fast as it is applied without ever reaching saturation. After a long time, as the suction gradients become negligible, the wetted profile will attain a wetness for which the conductivity is equal to the water supply rate, and the lower this rate, the lower the degree of saturation of the infiltrating profile.

Recharge is the rate at which water is replenished in the aquifer. Surface water reaches the permanent water table via a number of different routes. Intergranular seepage augments the moisture content of the soil and satisfies any moisture deficit before recharge may occur. But water may also enter and flow through crack systems in the unsaturated zone, thus reaching the water table with little or no effect on general soil moisture conditions. Especially in hilly terrain, rainfall may run over the surface of the land and collect in ditches and stream channels which feed the aquifer.

Quantification of ground water recharge is a major problem in many water-resource investigations. It is a complex function of meteorological conditions, soil, vegetation, physiographic characteristics and properties of the geologic material within the paths of flow. Soil layering in the unsaturated zone plays an important role in facilitating or restricting downward water movement to the water table. Also, the depth to the water table is important in ground water recharge estimations. Of all the factors controlling ground water recharge, the antecedent soil moisture regime probably is the most important.

The conventional method of estimating recharge as precipitation minus evapotranspiration minus runoff, with allowance for changes in soil moisture storage, is very sensitive to measurement errors and to the time scale of analysis. The customary method of calculating ground water recharge by multiplying a constant specific yield value by the water table rise over a certain time interval may be erroneous, especially in shallow aquifers. The hydraulic approach, based on Darcy's equation, offers the most direct measurement of seepage rates and hence recharge. However, it is highly site specific and most laborious and expensive, requiring specialized field equipment and personnel.

In the present study, the ground water recharge due to rainfall events separated by interstorm periods has been estimated by studying one dimensional vertical flow of water in the unsaturated zone. The governing partial differential equation (Richards equation) has been numerically solved with appropriate initial and boundary conditions pertinent to interstorm and storm periods.

2.0 REVIEW

2.1 General

Estimating the rate of aquifer replenishment is probably the most difficult of all measures in the evaluation of ground water resources. Estimates are normally and almost inevitably subject to large errors. No single comprehensive estimation technique can yet be identified from the spectrum of those available, which does not give suspect results.

Recharge estimation can be based on a wide variety of models which are designed to represent the actual physical processes. Methods which are currently in use include the following :

(i) The soil water balance method (soil moisture budget);

(ii) The zero flux plane method;

(iii) The one-dimensional soil water flow model;

(iv) Inverse modelling for estimation of recharge (two-dimensional ground water flow model);

(v) The saturated volume fluctuation method (ground water balance); and

(vi) Isotope techniques and solute profile techniques.

The two-dimensional ground water flow model and the saturated volume fluctuation method are regarded as indirect methods, because ground water levels are used to determine the recharge.

2.2 Soil Water Balance Method

Water balance models were developed in the 1940s by Thornthwaite (1948) and revised by Thornthwaite and Mather (1955). The method is essentially a book-keeping procedure which estimates the balance between the inflow and outflow of water.

In a standard soil water balance calculation, the volume of water required to saturate the soil is expressed as an equivalent depth of water and is called the soil water deficit. The soil water balance can be represented by :

$$G_r = P - E_a + \Delta S - R_o \qquad \ldots(2.1)$$

where,

G_r = recharge;

P = precipitation;

E_a = actual evapotranspiration;

ΔS = change in soil water storage; and

R_o = run-off.

One condition that is enforced, is that if the soil water deficit is greater than a critical value (called the root constant), evapotranspiration will occur at a rate less than the potential rate. The magnitude of the root constant depends on the vegetation, the stage of plant growth and the nature of the soil. A range of techniques for estimating E_a, usually based on Penman-type equations, can be used.

The data requirement of the soil water balance method is large. When applying this method to estimate the recharge for a catchment area, the calculation should be repeated for areas with different precipitation, evapotranspiration, crop type and soil type. The soil water balance method is of limited practical value, because E_a is not directly measurable. Moreover, storage of moisture in the unsaturated zone and the rates of infiltration along the various possible routes to the aquifer form important and uncertain factors. Another aspect that deserves attention is the depth of the root zone which may vary in semi-arid regions between 1 and 30 metres. Results from this model are of very limited value without calibration and validation, because of the substantial uncertainty in input data (precipitation and potential evapotranspiration). The model parameters do not have a direct physical representation which can be measured in the field.

2.3 Zero Flux Plane Method

The zero flux plane method relies on the location of a plane of zero hydraulic gradient in the soil profile. Recharge over a time interval is obtained by summation of the changes in water contents below this plane. The position of the zero flux plane is usually determined by installation of tensiometers. Unfortunately, the method fails to work during periods of high infiltration, when the hydraulic gradient becomes positive downwards throughout the profile.

The flux q, defined as the volume of water per unit time passing through the unit area at any depth, is given by Darcy's law :

$$q = -K(\theta) \frac{\partial H}{\partial z} \qquad \ldots(2.2)$$

where,

$K(\theta)$ = unsaturated hydraulic conductivity;

H = total water potential = $h(\theta) - z$;

z = depth beneath the surface (positive);

h = matric potential (negative); and

θ = water content.

Thus, knowing the unsaturated hydraulic conductivity and the potential gradient, the flux may be determined. Water potentials may be measured, using tensiometers or the neutron scattering technique. The hydraulic conductivity estimation presents more of a problem. Firstly, K may vary by a factor of 10^3 or so over the normal water content range of a typical soil and, secondly, there are large variations of K from place to place, even in apparently homogeneous soils and over distances of a few metres at the same depth.

There is, however, an alternative to this approach which avoids the need to know values of K. From the one-dimensional vertical form of the water balance equation :

$$\frac{\partial \theta}{\partial t} = - \frac{\partial q}{\partial z} \qquad \text{...(2.3)}$$

by assuming negligible lateral soil moisture flow, one obtains by integration from depth z to depth z + dz :

$$q_z = q_{z+dz} + \int_{z}^{z+dz} \frac{\partial \theta}{\partial t} \, dz \qquad \text{...(2.4)}$$

where q_z is the vertical component of the Darcian water flux. At the zero flux plane depth, say z_0, the potential gradient is zero and the flux is also zero. If z_0 does not change with time, the accumulated flux, $F(z')$, between times t_1 and t_2 is

$$F(z') = \int_{t_1}^{t_2} q(z).dt = \int_{z_0}^{z'} \left|\theta(t_1) - \theta(t_2)\right|.dz \qquad \text{...(2.5)}$$

where,

$z' = z + dz$ and $z_0 = z$.

2.4 Soil Water Flow Model

For recharge to occur, water has to move through the unsaturated zone until it reaches the water table. Flow conditions within this zone are far more complex than the flow mechanisms in a saturated aquifer.

The equation of a moisture retention curve is a non-linear relation of the water content. In more physical terms, it is said to show a hysteresis effect. Since the moisture retention curve can only be determined experimentally, its true behaviour in practice is only known at a finite number of points. Two methods, to obtain values at non-experimental points, can be used. The first and most obvious method is to use interpolation, but this method can only be successful in those cases where the experimental points are closely spaced. The second approach is to fit an empirical equation to the experimental points. The equations mostly used today are the Brooks and Corey function (Brooks and Corey, 1964) and the Van Genuchten (1980) function. The Van Genuchten equation deserves special attention. In this equation, the moisture retention curve is expressed as :

$$s_r = [\ 1 + (\lambda h)^n\ ]^{-m} \qquad \ldots(2.6)$$

where λ, n and m are characteristic constants, which have to be determined for every soil type. Van Genuchten suggested that one should use the value $m = 1 - 1/n$. The Van Genuchten equation expresses the moisture retention curve not in terms of the water content, but rather in terms of the reduced water content, defined by the equation :

$$S_r = \frac{(\theta - \theta_r)}{(\theta_s - \theta_r)} \qquad \text{...(2.7)}$$

where,

θ_s = the saturated water content; and

θ_r = the residual water content.

The three parameters, namely, (i) the water content, (ii) the matric potential (fluid pressure), and (iii) the hydraulic conductivity, are interrelated. These relationships are very sensitive. For example, a change in the water content of a few percent, often corresponds to a change in the hydraulic conductivity of two or more orders of magnitude. The one-dimensional equation for vertical flow in the unsaturated zone can be expressed as (Richards, 1931) :

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} [K(h) \frac{\partial h}{\partial z}] - \frac{\partial}{\partial z} [K(h)] \qquad \text{...(2.8)}$$

where,

θ = volumetric water content;

K = hydraulic conductivity [= $K(\theta)$ or $K(h)$]; and

h = matric potential.

Both θ and K are functions of the unknown potential h. The solutions of equation (2.8) are more sensitive to $h(\theta)$ variations than $K(\theta)$ variations. No evidence in the literature exists that the $K(\theta)$ relationship exhibits a significant hysteresis therefore it is safe to assume that K is a unique function of θ.

Following Richards (1931), Darcy's law for unsaturated flow can be expressed as :

$$q = -K(\theta) \frac{\partial H}{\partial z} \qquad \text{...(2.9)}$$

where $H = h(\theta) - z$ and $K(\theta)$ are related to the relative permeability given by Van Genuchten (1980) :

$$k_r(S_r) = S_r^{1/2} \left[1 - (1 - S_r^{1/m})^m \right]^2 \qquad \ldots(2.10)$$

where,

$K(S_r) = K_s \cdot k_r(S_r)$; and

K_s = saturated hydraulic conductivity.

Equation (2.8) can be solved by either a finite difference or a finite element model.

2.5 Inverse Modelling Technique

The inverse modelling technique is a two-dimensional finite element (or finite difference) ground water model of the saturated zone. Current methods of calibrating ground water flow models are either indirect or direct. The indirect approach is essentially a trial and error procedure that seeks to improve an existing estimate approach of the parameters in an iterative manner, until the model response is sufficiently close to that of the real system. The direct approach is different in that it treats the model parameters as dependent variables in a formal inverse boundary value problem.

One of the main difficulties in dealing with the inverse problem stems from the inherent non-uniqueness of its solution. Many of the data entered into the inverse modelling technique represent imprecise measurements and processed information that give a distorted picture of the system's true state.

The calculation of recharge to an aquifer by the inverse modelling technique must be regarded with caution if the true

S-values (storage coefficient) of the aquifer are not known. If, however, the calibrated S-values can be regarded as being very close to the real values, this technique can be of much use in describing the behaviour of the aquifer to the recharge phenomena in general.

2.6 Saturated Volume Fluctuation (SVF) Method

Inputs and outputs for conventional hydrological models are generally water volumes per unit time, such as recharge, discharge and surface inflows and outflows. The fundamental idea common to a variety of situations, is that the hydrological balance equation or some other equation, empirically derived, is usually employed, for example :

$$I - O = \frac{\Delta W}{\Delta t} \qquad \ldots(2.11)$$

In the same manner, the geohydrological balance equation for a ground water reservoir is given as :

$$I - O + G_r - Q = \frac{\Delta W}{\Delta t} \qquad \ldots(2.12)$$

where,

$I = \frac{I_1 + I_2}{2}$ = mean lateral inflow (m^3/day) during time $t_2 - t_1 = \Delta t$;

$O = \frac{O_1 + O_2}{2}$ = mean lateral outflow (m^3/day) during Δt;

G_r = ground water recharge into the reservoir in m^3/day (also named percolation or deep infiltration);

Q = discharge out of (or into) the reservoir (bore holes, rivers, etc.) in m^3/day during Δt;

ΔW = change in ground water volume (m^3) = $S.\Delta V$;

S = specific yield (or effective porosity);

ΔV = change in saturated volume of aquifer material (= $V_2 - V_1$); and

Δt = $t_2 - t_1$ = time increment.

In equation (2.12), it is assumed that there is no vertical movement through the base of the water table aquifer. In the case where the roots of plants extract water from below the water table, the evapotranspiration must be added to the Q-term in equation (2.12). The accuracy of estimating this evapotranspiration term is not very reliable and for this reason, it is assumed that the G_r-term already includes the evapotranspiration, i.e.

effective G_r = actual G_r - evapotranspiration from the saturated zone.

The term I in equation (2.12) can be expanded with the aid of Darcy's law :

$$I = T_i . L_i . \frac{i_i^1 + i_i^2}{2} = A_1 . T_i \qquad \ldots(2.13)$$

where,

T_i = mean transmissivity at the inflow boundary (i.e. T_i = KD for a water table aquifer);

L_i = width of the inflow boundary;

i_i^1 = ground water gradient at inflow boundary at time t_1; and

i_i^2 = ground water gradient at inflow boundary

at time t_2.

The same reasoning can be followed at the outflow boundary to yield :

$$O = T_o . L_o . \frac{i_o^1 + i_o^2}{2} = A_2 . T_o \qquad \ldots(2.14)$$

Substitution into equation (2.12) yields :

$$G_r + A_1 . T_i - A_2 . T_o - A_3 . S = Q \qquad \ldots(2.15)$$

where,

$$A_3 = \frac{\Delta V}{\Delta t}$$

Equation (2.15) is the general ground water balance equation for an unconfined aquifer. The boundaries of an area usually studied, do not represent stream lines, i.e. they are not perpendicular to the equipotential lines. Hence, the lateral inflow and outflow of ground water crossing the area's boundaries must be accounted for in the balance equation. One of the factors influencing the change in water table is the effective porosity, S, of the zone in which the water table fluctuations occur. It has been recognized that S changes as the depth of the water table changes, especially for water tables less than 3 metres deep. Furthermore, it should be noted that if the water table drops, part of the water is retained by the soil particles; if it rises, air can be trapped in the interstices that are filling with water. Hence S for rising water is, in general, less than for a falling water table.

To apply equation (2.15) correctly, it is essential that both the area and the period for which the balance is assessed, be carefully chosen. By comparison of ground water levels of bore holes with similar water table fluctuation patterns, holes with the same pattern can be grouped together. It is also conceivable that the whole area be divided into sub-areas by the Thiessen method. Equation (2.15) can be applied for a number of specified assumptions.

(a) Where the inflow terms are balanced by the outflow terms, the change in ground water storage is zero (i.e. $\Delta V = 0$). This provides the necessary conditions to derive safe yield estimates and to predict recharge from precipitation :

$$G_r = A_2 . T_o - A_1 . T_i + Q \qquad \ldots(2.16)$$

When outflow occurs in the absence of inflow, a general recession model may be formulated. This permits an evaluation of the outflow quantities, the effects on ground water storage or the inflow that takes place following the recession.

(b) By incorporating the 'no recharge' recession (i.e. $G_r = 0$) for ΔV = maximum decrease during Δt, equation (2.15) reduces to :

$$A_1 . T_i - A_2 . T_o - A_3 . S = Q$$

from which S can be calculated as :

$$S = \frac{A_1 . T_i - A_2 . T_o - Q}{A_3} \qquad \ldots(2.17)$$

The S calculated with the above equation is a minimum, because G_r may not be zero as assumed.

(c) If the aquifer is bounded by impervious dykes or by ground water divides, A_1 and A_2 in equation (2.15) are zero. For this case,

$$G_r - A_3.S = Q \qquad ...(2.18)$$

from which the ground water recharge can be calculated, if S is known.

The following procedures for the application of equation (2.15) are recommended :

(1) Ground water levels in observation bore holes, which are well distributed over the whole of the aquifer within certain well-defined ground water boundaries (such as ground water divides or no-flow boundaries), are required on a regular, preferably monthly, basis.

(2) The region must be divided into a number of small triangles (constructing a mesh).

(3) Monthly water levels should be interpolated to every node of the mesh, after which the saturated volume, V_i at time t_i, is calculated. An arbitrary base level can be assumed, because only the difference in ΔV over a time Δt is needed.

(4) Repeat (3) for all months.

(5) Construct a graph of V_i against time (months). For times where $V_i = V_j$ (i.e. $\Delta V = 0$), equation (2.16) can be used to estimate the ground water recharge G_r, during the time $t_j - t_i$.

It is very important to realize that equation (2.15) is subject to a number of possible errors. The equation is a finite difference approach, with a solution accuracy which is dependent on the size of Δt. Interpolation errors always occur, but can be minimized, if the bore holes are well distributed over the domain of interest. The same bore holes must be used when interpolating ground water levels of different periods. It is equally important to always use the same interpolation technique (e.g. kriging) during the above calculations. If the aquifer is bounded by a flow or constant head boundary, the solution of equation (2.15) is dependent on the accuracy of the transmissivity values at these boundaries.

2.7 Isotope and Solute Profile Techniques

^{3}H, ^{2}H, ^{18}O and ^{14}C are commonly used in recharge studies, of which the first three most accurately simulate the movement of water, because they form a part of the water molecule. Many studies on recharge estimation using natural tritium, are listed in the literature. Although a proven tool for qualitative recharge estimation, environmental tritium has several disadvantages, e.g. (i) tritium is not conservative and is lost from the system by evapotranspiration; (ii) contamination during sampling and processing is a factor which is enhanced in remote areas and at low total moisture levels; (iii) analysis is highly specialized and costly; (iv) quantitative studies are difficult to achieve, since it is difficult to determine a tritium mass balance.

An environmental tracer suitable for determining the movement of water must be highly soluble, conservative and not substantially taken up by vegetation. The chloride ion satisfies most of these criteria and is therefore considered a suitable tracer, particularly in coastal areas where large quantities of aeolian chloride are precipitated.

If the assumption of chloride as a conservative ion is accepted, the ground water recharge is given by :

$$G_r = \frac{D}{C} \text{ (mm/year)} \qquad \ldots(2.19)$$

where,

D = wet and dry chloride deposition (mg/m^2/year); and

C = concentration in ground water.

The method is convenient, fast and cheap. The drawback of the technique is the uncertainty in the determination of the wet and dry deposition. The principal source of chloride in ground water, if there are no evaporite sources, is from the atmosphere. In this case, the recharge can be expressed as :

$$G_r = \text{rainfall} \times \frac{\text{Cl of rainfall}}{\text{Cl of ground water}} \qquad \ldots(2.20)$$

The chloride method must be treated with caution, as accession of chloride near the soil surface may violate the assumption of a steady state chloride flux density throughout the unsaturated zone, because of evapotranspiration. Furthermore, recharge under conditions of extremely high rainfall with a long recurrence period, is likely to influence the chloride concentration of ground water to a high degree, resulting in an overestimate of the mean annual recharge.

3.0 PROBLEM DEFINITION

There have been three modes of infiltration recognized due to rainfall : (1) nonponding infiltration, involving rain not intense enough to produce ponding, (2) preponding infiltration, due to rain that can produce ponding but that has not yet done so, and (3) rainpond infiltration, characterized by the presence of ponded water. Rainpond infiltration is usually preceded by preponding infiltration, the transition between the two being called incipient ponding. Thus, nonponding and preponding infiltration rates are dictated by rain intensity, and are therefore supply controlled (or flux controlled), whereas rainpond infiltration rate is determined by the pressure (or depth) of water above the soil surface as well as by the suction conditions and conductivity relations of the soil. Where the pressure at the surface is small, rainpond infiltration, like ponding infiltration in general, is profile controlled.

In the analysis of rainpond or ponding infiltration, the surface boundary condition generally assumed is that of a constant pressure at the surface, whereas in the analysis of nonponding and preponding infiltration, the water flux through the surface is considered to be equal either to the rainfall rate or to the soil's infiltrability, whichever is the lesser. In actual field conditions, rain intensity might increase and decrease alternately, at times exceeding the soil's saturated conductivity (and its infiltrability) and at other times dropping below it. However, since periods of decreasing rain intensity involve complicated hysteresis phenomena, the analysis of variable-intensity rainstorms is rather difficult.

The process of infiltration under rain is normally analysed based on the assumption of no hysteresis. The falling raindrops are taken to be so small and numerous that rain could be treated as a continuous body of 'thin' water reaching the soil surface at a specified rate. Soil air is regarded as a continuous phase, at atmospheric pressure. The soil is mostly assumed to be uniform and stable (i.e., no fabric changes such as swelling or surface crusting).

If a constant pressure head is maintained at the soil surface (as in rainpond infiltration), then the flux of water into this surface must be constantly decreasing with time. If a constant flux is maintained at the soil surface, then the pressure head at this surface must be constantly increasing with time. Infiltration of constant-intensity rain can result in ponding only if the relative rain intensity (i.e., the ratio of rain intensity to the saturated hydraulic conductivity of the soil) exceeds unity. During nonponding infiltration under a constant rain intensity q_r, the surface pressure head will tend to a limiting value h_{lim} such that $K(h_{lim}) = q_r$.

Under rainpond infiltration, the wetted profile consists of two parts: an upper, water-saturated part; and a lower, unsaturated part. The depth of the saturated zone continuously increases with time. Simultaneously, the steepness of the moisture gradient at the lower boundary of the saturated zone (i.e., at the wetting zone and the wetting front) is continuously decreasing. The higher the rain intensity is, the shallower is the saturated layer at incipient ponding and the steeper is the moisture gradient in the wetting zone.

A rainstorm of any considerable duration typically consists of spurts of high-intensity rain punctuated by periods of low-intensity rain. During such respite periods, surface soil moisture tends to decrease because of internal drainage, thus reestablishing a somewhat higher infiltrability. The next spurt of rainfall is therefore absorbed more readily at first, but soil infiltrability quickly falls back to, or even below, the value it had at the end of the last spurt of rain. A complete description would, of course, necessitate taking account of the hysteresis phenomenon in the alternately wetting-and-draining surface zone.

The objective of the present study is to estimate the amount and time distribution of ground water recharge due to a series of rainfall events with rain intensities approximately equal to soil infiltrability (i.e., constant pressure head maintained at the soil surface) and these rainfall events separated by interstorm periods. A numerical model (finite difference scheme) is used for solving the nonlinear partial differential equation (Richards equation) describing one-dimensional water flow through the unsaturated porous medium. It uses a one-dimensional (vertical) formulation of soil moisture movement in the following modes:

(a) into the soil through infiltration during rainstorms;
(b) out of the soil through evaporation of exfiltrated water between rainstorms;
(c) downward percolation to the water table continuously during the rainy season; and
(d) upward capillary rise from the water table.

The amount of ground water recharge due to rainfall is estimated based on Darcy's law and water balance of the unsaturated zone.

4.0 METHODOLOGY

4.1 General

The one-dimensional partial differential equation which describes the movement of moisture through unsaturated porous media subject to appropriate boundary and initial conditions has many field applications in the water environment. In hydrology, it describes the infiltration process that links the surface and sub-surface waters on land. In soil physics, it describes the capillary rise as well as drainage and evaporation of moisture in soils. In environmental pollution, it describes the longitudinal dispersion of pollutants in water courses. Therefore, the problem of seeking solutions to this equation has become a subject of concern for investigators from many different disciplines.

The unsaturated flow equation in its general form is highly non-linear. The parameters are often complex functions of the dependent variables. When the equation is used to describe the infiltration process, the problem is further complicated by the existence of two surface boundary conditions identified as the ponded infiltration condition and the rain infiltration condition. Under the latter condition, the problem formulation and the approach to the solution also depend upon the intensity of rainfall in relation to the surface saturated hydraulic conductivity. No analytical solution to the equation in its general form is available at the present time.

However, the linearized form of the equation is in mathematical form identical to the longitudinal dispersion equation with constant parameters. An analytical solution for the latter equation has been proposed by Ogata and Banks (1961), and can therefore be used for the linearized infiltration equation as well. A semi-analytical approach has also been proposed by Philip (1957). Both these solutions are for ponded infiltration condition only. Subsequently several researchers have proposed numerical solution procedures based upon the finite difference method for solving the ponded infiltration problem. For rain infiltration condition, Rubin and Steinhardt (1963, 1964) proposed a finite difference based numerical procedure for low rainfall intensities. Later, Rubin (1969) extended the method for analysing ponded rain infiltration. Similar finite difference based procedures have been proposed by Freeze (1969) and Whisler and Klute (1969). A finite element based procedure using complete discretization has been proposed by Bruch and Zyvoloski (1974) for vertical infiltration under ponded conditions. In most of these studies, the comparisons have been either with already published results or with data gathered from soil columns or horizontal field plots.

4.2 Constitutive Equations

Downward infiltration into an initially unsaturated soil generally occurs under the combined influence of suction and gravity gradients. As the water penetrates deeper and the wetted part of the profile lengthens, the average suction gradient decreases, since the overall difference in pressure head (between the saturated soil surface and the unwetted soil inside the profile) divides itself along an ever-increasing distance. This trend continues until eventually the suction gradient in the upper part of the profile becomes negligible, leaving the constant gravitational gradient in effect as the only remaining force moving water downward. Since the gravitational head gradient has the value of unity (the gravitational head decreasing at the rate of 1 cm with each centimeter of vertical depth below the surface), it follows that the flux tends to approach the hydraulic conductivity as a limiting value. In a uniform soil (without crust) under prolonged ponding, the water content of the wetted zone approaches saturation. However, in practice, because of air entrapment, the soil-water content may not attain total saturation but some maximal value lower than saturation which has been called 'satiation'. Total saturation is assured only when a soil sample is wetted under vacuum.

Darcy's equation for vertical flow is

$$q = -K \frac{\partial H}{\partial z} = -K \frac{\partial}{\partial z} (h - z) \qquad \ldots(4.1)$$

where q is the flux, H the total hydraulic head, h the soil water pressure head, z the vertical distance from the soil surface downward (i.e., the depth), and K the hydraulic conductivity. At the soil surface, $q = i$, the infiltration rate. In an unsaturated soil, h is negative. Combining this formulation of Darcy's equation (4.1) with the continuity equation $\partial\theta/\partial t = -\partial q/\partial z$ gives the general flow equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial H}{\partial z}\right) = \frac{\partial}{\partial z} \left(K \frac{\partial h}{\partial z}\right) - \frac{\partial K}{\partial z} \qquad \ldots(4.2)$$

If soil moisture content θ and pressure head h are uniquely related, then the left-hand side of equation (4.2) can be written

$$\frac{\partial \theta}{\partial t} = \frac{d\theta}{dh} \cdot \frac{\partial h}{\partial t}$$

which transforms equation (4.2) into

$$C \frac{\partial h}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial h}{\partial z}\right) - \frac{\partial K}{\partial z} \qquad \ldots(4.3)$$

where C (= $d\theta/dh$) is defined as the specific (or differential) water capacity (i.e., the change in water content in a unit volume of soil per unit change in matric potential).

Alternatively, we can transform the right-hand side of equation (4.2) once again using the chain rule to render

$$\frac{\partial h}{\partial z} = \frac{dh}{d\theta} \cdot \frac{\partial \theta}{\partial z} = \frac{1}{C} \cdot \frac{\partial \theta}{\partial z}$$

We thus obtain

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(\frac{K}{C} \cdot \frac{\partial \theta}{\partial z}\right) - \frac{\partial K}{\partial z}$$

or

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(D \frac{\partial \theta}{\partial z}\right) - \frac{\partial K}{\partial z} \qquad \ldots(4.4)$$

where D is the soil water diffusivity. Equations (4.2), (4.3) and (4.4) can all be considered as forms of the Richards equation.

Note that the above three equations contain two terms on their right-hand sides, the first term expressing the contribution of the suction (or wetness) gradient and the second term expressing the contribution of gravity. Whether the one or the other term predominates depends on the initial and boundary conditions and on the stage of the process considered. For instance, when infiltration takes place into an initially dry soil, the suction gradients at first can be much greater than the gravitational gradient and the initial infiltration rate into a horizontal column tends to approximate the infiltration rate into a vertical. On the other hand, when infiltration takes place into an initially wet soil, the suction gradients are small from the start and become negligible much sooner. The effects of ponding depth and initial wetness can be significant during early stages of infiltration, but decrease in time and eventually tend to vanish in a very deeply wetted profile.

4.3 Initial and Boundary Conditions

There are three different initial and boundary conditions that can be applied to equations (4.3) and (4.4) when describing infiltration. They are briefly defined in the following equations:

Condition 1

$$\theta(z,0) = \theta_i \quad \text{for } z \geq 0, \quad t = 0 \qquad \ldots(4.5\text{ a})$$

$$\theta(0,t) = \theta_0 \quad \text{for } z = 0, \quad t \geq 0 \qquad \ldots(4.5\text{ b})$$

where θ_i and θ_0 are the initial and surface moisture contents, respectively (usually $\theta_0 > \theta_i$). They may be constants or functions of z or t. The most common condition in infiltration is when there is a thin layer of water available at the surface. Then, the surface moisture content is the saturated value θ_s and is called the ponded infiltration condition. Then:

$$\theta(0,t) = \theta_s \quad \text{for } z = o, \quad t \geq 0 \qquad \ldots(4.5\text{ c})$$

Condition 2

$$\theta(z,0) = \theta_i \quad \text{for } z \geq 0, \quad t = 0 \qquad \ldots(4.6\text{ a})$$

$$\text{Flux} = -K\left(\frac{\partial h}{\partial z} - 1\right) = q_r \quad \text{for } z = 0, \quad t > 0 \qquad \ldots(4.6\text{ b})$$

where q_r is the rainfall intensity. The condition (4.6 b) can also be written as :

$$\frac{\partial \theta}{\partial z} = - \frac{q_r - K}{D} \qquad \text{...(4.6 c)}$$

This condition corresponds to rain infiltration and is applicable from the beginning of rainfall to the time of occurrence of incipient ponding. For low rainfall intensities $[q_r < K(\theta_s)]$ rain infiltration can continue without giving rise to ponding. As time passes, the surface moisture content approaches a limiting value θ_L.

Condition 3

$$h(z,0) = h_i \qquad \text{for } z \geq 0, \quad t = 0 \qquad \text{...(4.7 a)}$$

$$h(0,t) = h_f \geq 0 \qquad \text{for } z = 0, \quad t \geq t_p \qquad \text{...(4.7 b)}$$

$$\text{Flux} = -K \left(\frac{\partial h}{\partial z} - 1\right) = q_r \quad \text{for } z = 0, \quad 0 \leq t \leq t_p \qquad \text{...(4.7 c)}$$

where,

h_i = initial soil water pressure;

h_f = surface soil water pressure during ponding (hydrostatic); and

t_p = time of incipient ponding.

This condition corresponds to rain infiltration in which the rain intensity is greater than the surface saturated hydraulic conductivity. The physical meaning being that the rainfall intensity is exceeding the infiltration capacity of the soil, and therefore ponding of water at the surface is taking place. In equation (4.7 b), h_f can be taken as zero without loss of generality.

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

I. Initial condition:

$\theta(z,0) = \theta_i$ for $z \geq 0$, $t = 0$ ...(4.5 a)

(Equilibrium moisture profile with surface moisture content = 0.10)

II. Upper boundary conditions:

(a) during rain infiltration -

$\theta(0,t) = (\theta_s - 0.001)$ for $z = 0$, $t \geq 0$ ...(4.5 c)

(b) during interstorm period -

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(o,t) can be derived from the thermodynamic relation (Edlefson and Anderson, 1943):

$$h(0,t) = \frac{RT(t)}{Mg} \ln [f(t)] \qquad ...(4.8)$$

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 ground water recharge due to rainfall. The lower boundary condition has therefore been set as

$\theta(z=L, t) = \theta_s - 0.001$ ...(4.9) 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 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:

$$K = K_s \frac{A}{A + |h|^{\beta_1}} ; \qquad \ldots(4.10)$$

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

and

$$\theta = \frac{\alpha\,(\theta_s - \theta_r)}{\alpha + |h|^{\beta_2}} + \theta_r ; \qquad \ldots(4.11)$$

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

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

Figure 1 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 (4.3) is a non-linear partial differential equation (PDE) because the parameters $K(h)$ and $C(h)$ depend on the

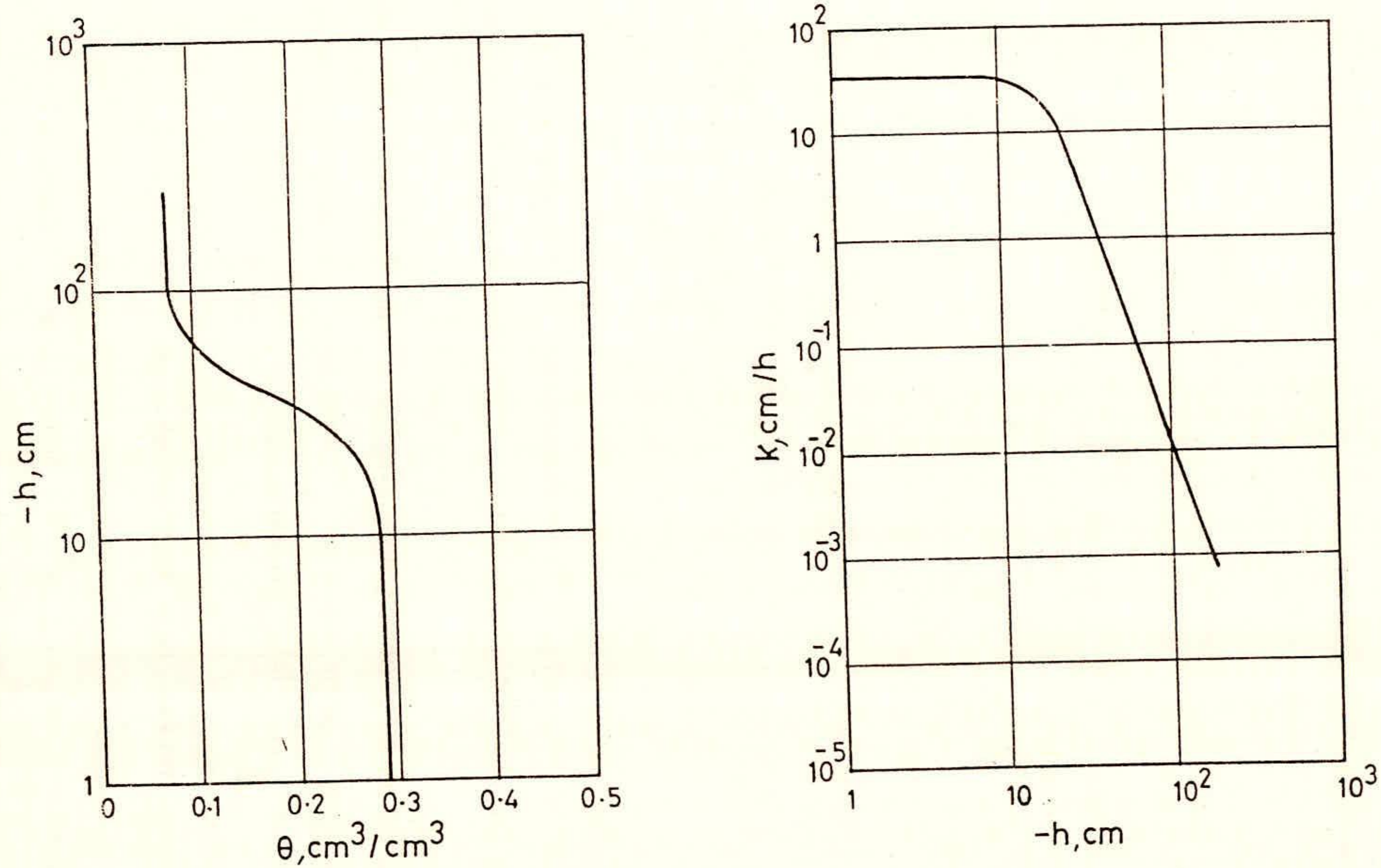

FIG. 1. RELATIONSHIPS BETWEEN THE SOIL WATER PRESSURE h, THE WATER CONTENT θ AND THE HYDRAULIC CONDUCTIVITY K FOR THE SOIL USED IN THE STUDY

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 (4.3) 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

$$\frac{C}{K} \frac{\partial h}{\partial t} = \frac{\partial^2 h}{\partial z^2} + \frac{1}{K} \frac{\partial K}{\partial z} \left(\frac{\partial h}{\partial z} - 1\right) \qquad \ldots (4.12)$$

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 (4.12), by taking time step as $\Delta t/2$, we have

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

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

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

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

$$= \frac{2C_i^j}{K_i^j} h_i^j + \frac{1}{2} \quad \frac{K_{i+1}^j - K_{i-1}^j}{K_i^j} \; \frac{\Delta t}{\Delta z} \left[\frac{h_{i+1}^j - h_{i-1}^j}{2\Delta z} - 1\right] \qquad \ldots(4.13)$$

Correction (estimation of h_i^j)

From equation (4.12), 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

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

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

$$+ \frac{1}{2} \frac{K_{i+1}^{j+1/2} - K_{i-1}^{j+1/2}}{K_i^{j+1/2}} \frac{\Delta t}{\Delta z} \left[\frac{h_{i+1}^{j+1/2} - h_{i-1}^{j+1/2}}{2\Delta z} - 1 \right] \qquad ...(4.14)$$

When equation (4.13) or (4.14)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).

4.6 Estimation of Ground Water Recharge

After obtaining the pressure (and soil moisture) distribution at each time step, the ground water recharge due to rainfall was estimated by the following two methods:

(i) Darcian flux method

The flux in the Darcian method is calculated as the product of the unsaturated hydraulic conductivity and the hydraulic gradient. According to Darcy's law, for one dimensional vertical flow, the volumetric flux q ($cm^3/cm^2/h$) can be written as

$$q = -K \frac{\partial}{\partial z} (h - z) \qquad (cm/h) \qquad ...(4.1)$$

or

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

The ground water recharge due to rainfall (RR) was estimated by applying the above equation for two vertically adjacent nodal points (at and above the water table) for each time step.

$$RR = -K^{j}_{i+1/2} \left(\frac{h^{j}_{i+1} - h^{j}_{i}}{\Delta z} - 1\right) \qquad ...(4.15)$$

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).

(ii) Water balance of the unsaturated zone

The soil water balance of the unsaturated zone can be represented as follows :

RECH = RAIN - EVAP - DELSM ...(4.16)

where,

RECH = ground water recharge ;

RAIN = rain infiltration ;

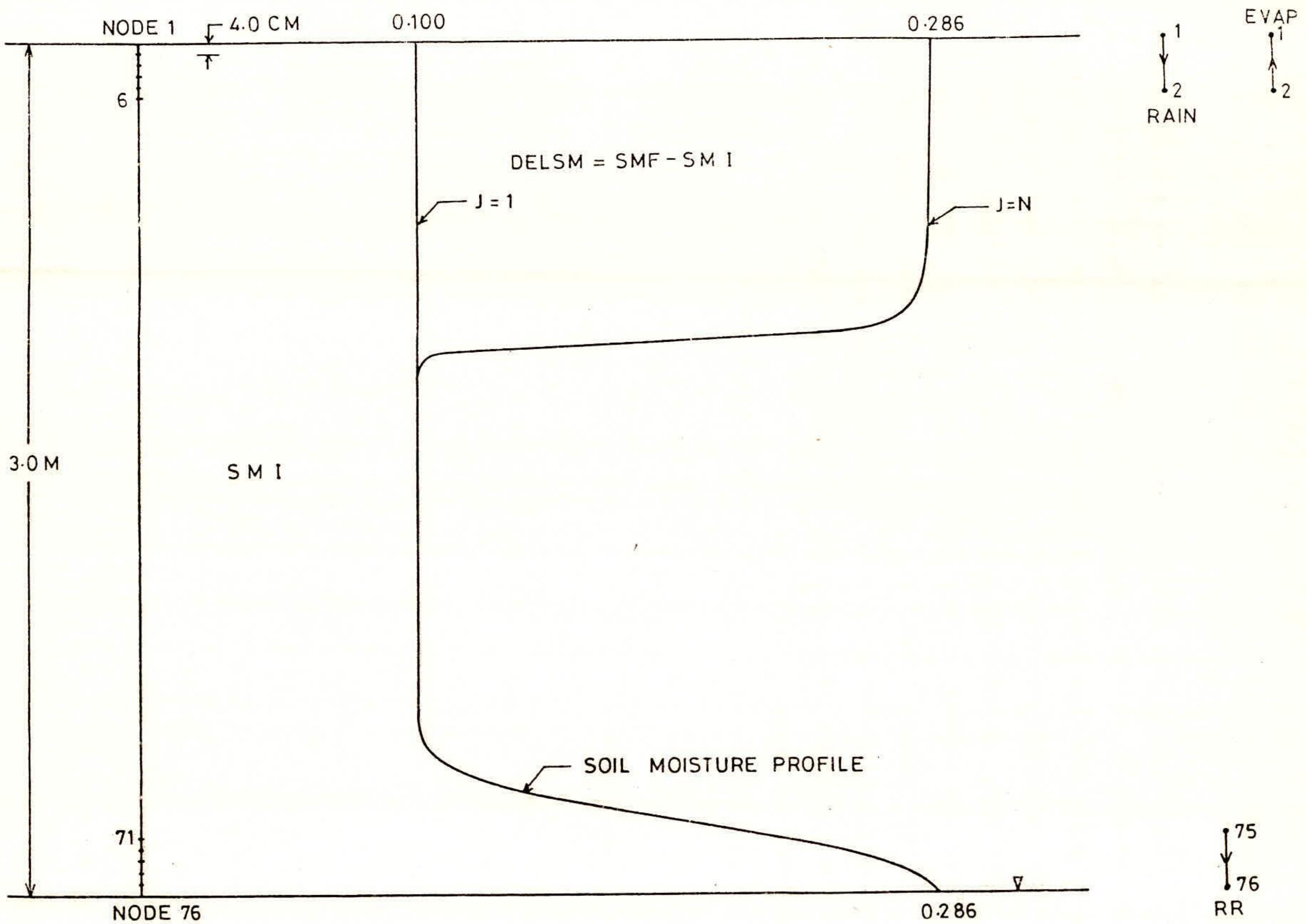

FIG. 2. SCHEMATIC REPRESENTATION OF COMPUTATIONAL SET-UP

EVAP = evaporation from the soil ; and

DELSM = change in soil moisture storage of the unsaturated zone.

Equation (4.16) provide a means of estimating ground water recharge due to rainfall during each time step. Rain infiltration and evaporation from the soil (assumed as zero during the storm period) were computed from the equation (4.1) for two vertically adjacent nodal points (at and below the ground surface). Figure 2 presents the schematic representation of computational set-up.

The computer code, for discretization scheme used in the model and estimation of ground water recharge due to rainfall as per the procedure described above, has been written in FORTRAN and presented in Appendix-I.

5.0 RESULTS

The numerical model described in section 4.5 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 (θ = 0.267 cm^3/cm^3). Haverkamp et al. (1977) has reported the infiltration profiles at various times for infiltration in the sand (under consideration) obtained by quasi-analytical solution of Philip. 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 75 layers of thickness 4 cm each (depth interval, Δz) down to the water table position assumed at a depth of 3 metres. Keeping in view the stability of the numerical scheme, the time step (Δt) was taken as 3 seconds during the entire study period. Three rainfall events of 3 hours duration each separated by interstorm periods of 3 hours duration were considered for the study (figure 3). Uniform evaporative conditions (temperature = 25°C , relative humidity = 0.75) were assumed during the interstorm periods. The upper boundary condition during the rain infiltration was defined as

$$\theta(0,t) = 0.286 \quad \text{for } z = 0, \quad t \geq 0$$

implying that a constant pressure head corresponding to θ = 0.286 (h = - 9.56 cm) was maintained at the soil surface during the rain infiltration. The lower boundary condition was defined as

$\theta(z=L, t) = 0.286$

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) The soil cover was assumed to be homogeneous and isotropic.

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

iv) The falling raindrops were assumed to be so small and numerous that rain may be treated as a continuous body of water reaching the soil surface at a certain rate.

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

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

The ground water recharge due to rainfall was estimated for a total duration of 30 hours by Darcian flux method and through water balance of the unsaturated zone. The input data to the model and output are given in Appendix-II and Appendix-III respectively.

Changes in the entire moisture content profile during rain infiltration are shown in figure 3. The wetted profile consists of two parts - an uppermost water-saturated part and a lower unsaturated, wetted part. The saturated layer of ever-increasing thickness propagates down through the profile. The time taken for recharge to occur was estimated as 1.42 hour and complete saturation was attained after 1.90 hour for the given rainfall, initial condition and soil characteristics.

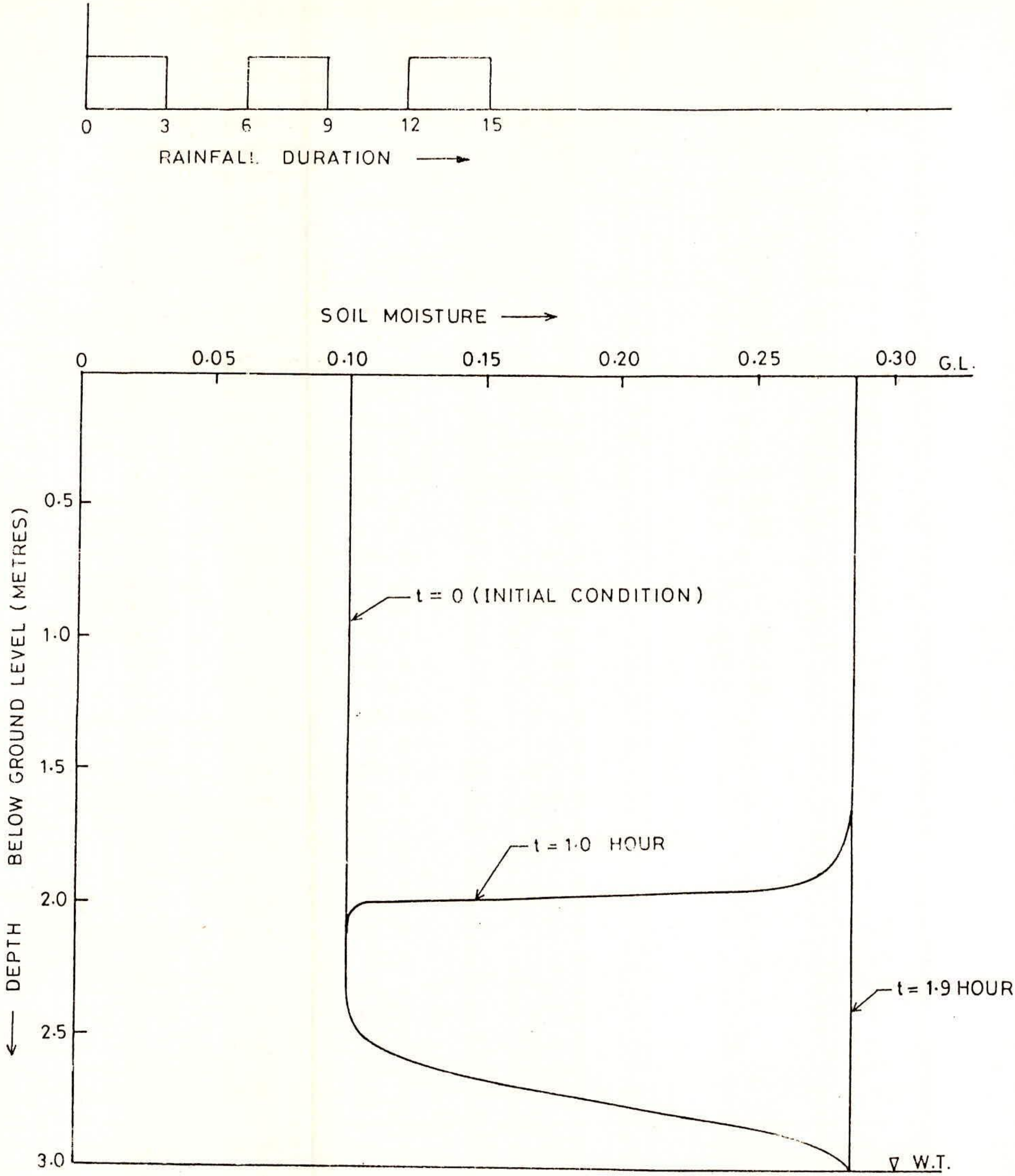

FIG.3. SOIL MOISTURE PROFILE AT DIFFERENT TIMES

Table 1 : Ground Water Recharge due to Rainfall

Hour	Rain Infiltration (cm)	Evaporation from the Soil (cm)	Change in Soil Moisture Storage (cm)	Ground Water Recharge (Water Balance) (cm)	Ground Water Recharge (Darcy's Law) (cm)
1	35.26	0	36.49	- 1.23	0
2	32.78	0	14.47	18.31	17.93
3	32.76	0	0	32.76	32.76
4	0	0.17	-18.56	18.39	16.93
5	0	0.05	- 6.54	6.49	6.15
6	0	0.03	- 4.30	4.27	3.99
7	35.15	0	29.36	5.79	4.73
8	32.76	0	0.04	32.72	32.72
9	32.76	0	0	32.76	32.76
10	0	0.17	-18.56	18.39	16.93
11	0	0.05	- 6.54	6.49	6.15
12	0	0.03	- 4.30	4.27	3.99
13	35.15	0	29.36	5.79	4.73
14	32.77	0	0.04	32.73	32.72
15	32.78	0	0	32.78	32.76
16	0	0.17	-18.56	18.39	16.93
17	0	0.05	- 6.54	6.49	6.15
18	0	0.03	- 4.30	4.27	3.99
19	0	0.03	- 3.24	3.21	2.96
20	0	0.02	- 2.58	2.56	2.34
21	0	0.02	- 2.14	2.12	1.91
22	0	0.01	- 1.82	1.81	1.61
23	0	0.01	- 1.57	1.56	1.37
24	0	0.01	- 1.37	1.36	1.19
25	0	0.01	- 1.21	1.20	1.04
26	0	0.01	- 1.08	1.07	0.92
27	0	0.01	- 0.96	0.95	0.82
28	0	0.01	- 0.87	0.86	0.73
29	0	0.01	- 0.78	0.77	0.66
30	0	0.01	- 0.71	0.70	0.59
Total	302.17	0.91	3.23	298.03	288.46

Table 1 presents the hourly values of rain infiltration, evaporation from the soil, change in soil moisture storage of the unsaturated zone and ground water recharge by the two methods for the study period. It can be observed that the ground water recharge due to rainfall estimated by Darcian flux method and water balance are in reasonable agreement with each other. The variation of cumulative ground water recharge with time is presented in figure 4.

It should be emphasized that the above results have not been subjected to empirical testing in the laboratory and in the field. Furthermore, the usefulness of the numerical model presented here is subject to several limitations as indicated below.

(a) A static water table has been considered at the base. This water table condition is not realistic from the point of view of continuity of flow between the saturated and unsaturated domains for various reasons. The existence of a static water table (pressure head equal to zero at a fixed location) does not take into account the fact that the water table will fluctuate in position, and that it will do so in response to the distribution of flow in both the unsaturated and saturated zones. A stronger objection can be raised in reference to flux calculations that show the flux across the water table to vary rapidly with time and in response only to the unsaturated flow conditions. In actual fact, the regional ground water flow pattern to which the water table is the upper boundary is only capable of accepting a given amount of recharge and thus offers a constraint on the possible flux of water across the water table. A basal boundary condition in which the pressure head equals zero at a fixed location is actually a statement of the gravity drainage problem, and the

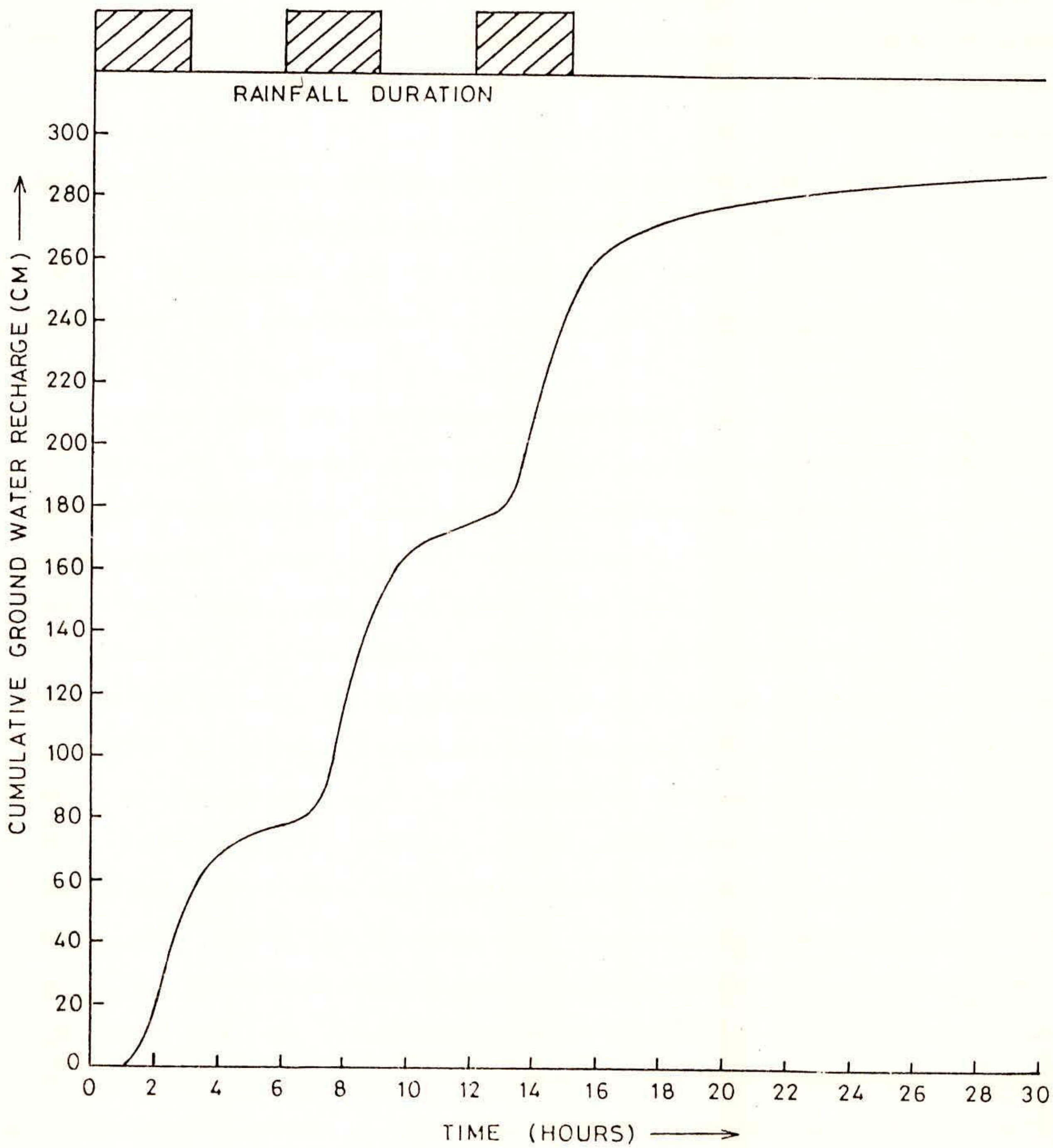

FIG. 4. VARIATION OF CUMULATIVE GROUND WATER RECHARGE WITH TIME

results should be interpreted in that light.

(b) The theory of rainfall infiltration presented here is not applicable when the assumption about soil air with approximately constant atmospheric pressure is not fulfilled.

(c) The theory under consideration can not be used whenever the effects of hysteresis in soil moisture properties are significant. Such effects may be created by the discreteness of raindrops. They might also be associated with decreases in rain intensity during flux-controlled rainfall uptake or with diminution in surface pressure heads during rainpond infiltration.

(d) The theory in question is also inapplicable to a soil in which infiltration-induced fabric transformations change the parametric moisture properties. If merely known time-dependencies of K(h) and h(θ) were involved, perhaps it would not be too difficult to extend the current numerical methods so as to take such a dependence into account, at least approximately. However, usually information on such a dependence is unavailable. Furthermore, fabric transformations under consideration usually decrease K(h). Such a decrease creates difficulties that can not be overcome, because it generates hysteresis effects.

(e) Difficulties in utilizing the theory in question are created also by the commonly met heterogeneity of soil cover. It is thought that an application of the methods developed in connection with flood water infiltration to the rainfall uptake case would not be difficult. Much more formidable is the areal treatment of infiltration into a soil with properties varying in the horizontal directions. In such a case one section of the area influences the infiltration into another by affecting the runoff.

(f) Finally, certain practical limitations on the utilization of the rainfall infiltration theory are due to the inadequacy of field methods for determining the pertinent soil moisture parametric functions. However, in certain cases of interest, the existing laboratory techniques may provide the required information.

In spite of the limitations outlined above, it is thought that under many conditions the theory presented here is applicable by incorporating the appropriate modifications in the initial and boundary conditions. However, to improve the reliability of ground water recharge estimates, we must monitor aquifer behaviour on a continuous or periodic basis to ensure that adequate data and hence representative averages of the spatially and temporally varying recharge process are obtained. The application of several independent or different ground water recharge estimation methods can complement one another and is likely to improve our knowledge of aquifer recharge, provided that an adequate hydrogeologic database and soil characteristics exist.

6.0 CONCLUSIONS

A numerical solution using an implicit finite-differencing technique is presented for a mathematical model of one-dimensional, vertical, unsteady, unsaturated flow above a water table. The solution is applicable to homogeneous, isotropic soils in which the functional relationships between hydraulic conductivity, moisture content, and soil moisture tension do not show hysteresis properties. The model has been applied for upper boundary condition of rain infiltration (equal to soil infiltrability) separated by interstorm periods and ground water recharge due to rainfall has been estimated. The model can furnish information useful in quantification of the rate of ground water recharge for soils with known moisture parameters and rains of a given intensity pattern by suitably modifying the initial and boundary conditions. However, the method is not utilizable when the soil exhibits significant air compression, parameter hysteresis, fabric transformations, or areal heterogeneity.

REFERENCES

1. Brooks, R. H. and A. T. Corey (1964), "Hydraulic Properties of Porous Media", Hydrol. Paper No. 3, Civil Engineering Department, Colorado State University, Fort Collins.
2. Bruch, J. C. and G. Zyvoloski (1974), "Solution of Equation for Vertical Unsaturated Flow of Soil Water", Soil Science, Volume 119, pp. 417-422.
3. Douglas, J. J. and B. F. Jones (1963), "On Predictor-Corrector Method for Non Linear Parabolic Differential Equations", J. Siam, Volume 11, pp. 195-204.
4. Edlefson, N. E. and A. B. C. Anderson (1943), "Thermodynamics of Soil Moisture", Hilgardia, Volume 15, pp. 31-298.
5. Freeze, R. A. (1969), "The Mechanism of Natural Ground Water Recharge and Discharge I. One Dimensional, Vertical, Unsteady, Unsaturated Flow above a Recharging or Discharging Ground Water Flow System", Water Resources Research, Volume 5, No. 1, pp. 153-171.
6. Haverkamp, R. and M. Vauclin (1979), "A Note on Estimating Finite Difference Interbloc Hydraulic Conductivity Values for Transient Unsaturated Flow Problems", Water Resources Research, Volume 15, No. 1, February 1979, pp. 181-187.
7. Haverkamp, R., M. Vauclin, J. Touma, P. J. Wierenga and G. Vachaud (1977), "A Comparison of Numerical Simulation Models for One-Dimensional Infiltration", Soil Sci. Soc. Am. J., Volume 41, pp. 285-294.
8. Hillel, Daniel (1980), "Applications of Soil Physics", Academic Press, New York, 385 pp.

9. Jayawardena, A. W. and J. J. Kaluarachchi (1986), "Infiltration into Decomposed Granite Soils: Numerical Modelling, Application and Some Laboratory Observations", Journal of Hydrology, Volume 84, pp. 231-260.

10. Kumar, C. P. and G. C. Mishra (1991), "Development of a Soil Moisture Prediction Model", Technical Report No. TR-104, National Institute of Hydrology, Roorkee, 1990-91, 52 pp.

11. Kumar, C. P. (1992), "Prediction of Evaporation Losses from Shallow Water Table using a Numerical Model", Technical Report No. TR-119, National Institute of Hydrology, Roorkee, 1991-92, 46 pp.

12. Ogata, A. and R. B. Banks (1961), "A Solution of the Differential Equation of Longitudinal Dispersion in Porous Media", U. S. Geological Survey, Prof. Paper No. 411-A, U. S. Govt. Printing Office, Washington, D. C.

13. Philip, J. R. (1957), "The Theory of Infiltration: 1. The Infiltration Equation and its Solution", Soil Science, Volume 83, pp. 345-357.

14. Remson, I., G. M. Hornberger and F. J. Molz (1971), "Numerical Methods in Subsurface Hydrology", Wiley-Interscience, New York, 389 pp.

15. Richards, L. A. (1931), "Capillary Conduction of Liquids through Porous Mediums", Physics, Vol. 1, pp. 318-333.

16. Rubin, J. (1969), "Numerical Analysis of Ponded Rainfall Infiltration", In: Water in the Unsaturated Zone, Proc. of the Wageningen Symposium, Wageningen, pp. 440-451.

17. Rubin, J. and R. Steinhardt (1963), "Soil Water Relations during Rain Infiltration: I. Theory", Soil Sci. Soc. Am., Proc., Volume 27, pp. 246-251.

18. Rubin, J. and R. Steinhardt (1964), "Soil Water Relations during Rain Infiltration: III. Water Uptake at Incipient Ponding", Soil Sci. Soc. Am., Proc., Volume 28, pp. 614-619.

19. Thornthwaite, C. W. (1948), "An Approach towards a Rational Classification of Climate", Geogr. Rev., Vol. 38, No. 1, pp. 55-94.

20. Thornthwaite, C. W. and J. W. Mather (1955), "The Water Balance", Publ. Climatol. Lab. Climatol. Drexel Inst. Technol., Vol. 8, No. 1, pp. 1-104.

21. Van Genuchten, M. Th. (1980), "A Closed-form Equation for predicting the Hydraulic Conductivity of Unsaturated Soils", Soil Sci. Soc. Am. J., Vol. 44, pp. 892-898.

22. Van Tonder, G. J. and J. Kirchner (1990), "Estimation of Natural Groundwater Recharge in the Karoo Aquifers of South Africa", Journal of Hydrology, Vol. 121, pp. 395-419.

23. Whisler, F. D. and A. Klute (1969), "Analysis of Infiltration into Stratified Soil Columns", In: Water in the Unsaturated Zone, Proc. of the Wageningen Symposium, Wageningen, pp. 451-470.

```
C       ESTIMATION OF GROUND WATER RECHARGE DUE TO RAINFALL
C       BY MODELLING OF SOIL MOISTURE MOVEMENT
C
C       IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION
C       (PREDICTION - CORRECTION)
C       (MODEL 4 OF HAVERKAMP ET AL., 1977)
C
        DIMENSION SUB(500),SUP(500),DIAG(500),B(500)
        DIMENSION H(500,2),CCC(500,2)
        DIMENSION THETA(500,2),HYDCON(500,2)
        DIMENSION HP(500,2),THETAP(500,2)
        OPEN(UNIT=1,FILE='HRECH.DAT',STATUS='OLD')
        OPEN(UNIT=2,FILE='HRECH.OUT',STATUS='NEW')
C
C       J REFERS TO TIME
C       I REFERS TO DEPTH
C       Z = DEPTH (CM), ORIENTED POSITIVELY DOWNWARD
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       THETA = VOLUMETRIC MOISTURE CONTENT (CUBIC CM / CUBIC CM)
C       H = SOIL WATER PRESSURE (RELATIVE TO THE ATMOSPHERE)
C           EXPRESSED IN CM OF WATER
C       THETAR = RESIDUAL MOISTURE CONTENT
C       THETAS = MOISTURE CONTENT AT SATURATION
C       THETAU = MOISTURE CONTENT AT THE SURFACE NODE
C                (UPPER BOUNDARY CONDITION)
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
C       STORM PERIODS = 0-LT1, LT2-LT3, LT4-LT5
C       INTERSTORM PERIODS = LT1-LT2, LT3-LT4
C
        READ(1,11)THETAR,THETAS,THETAU
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(I7,5X,I5)
      READ(1,61)LT1,LT2,LT3,LT4,LT5
 61   FORMAT(5I12)
      READ(1,62)T
 62   FORMAT(F5.2)
      READ(1,63)RH
 63   FORMAT(F5.2)
C
C     READING OF INITIAL CONDITIONS
C
      READ(1,17)(THETA(I,1),I=1,NNODE)
 17   FORMAT(5F12.6)
C
      WRITE(2,18)
 18   FORMAT(2X,'ESTIMATION OF GROUND WATER RECHARGE')
      WRITE(2,19)
 19   FORMAT(2X,'IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION')
      WRITE(2,20)
 20   FORMAT(2X,'(PREDICTION - CORRECTION)')
      WRITE(2,71)
 71   FORMAT(/2X,'TEMPERATURE IN CENTIGRADE')
      WRITE(2,72)T
 72   FORMAT(F7.2)
      WRITE(2,73)
 73   FORMAT(2X,'RELATIVE HUMIDITY OF THE AIR')
      WRITE(2,74)RH
 74   FORMAT(F7.3)
      WRITE(2,21)
 21   FORMAT(/2X,'THETAR',9X,'THETAS',9X,'THETAU')
      WRITE(2,31)THETAR,THETAS,THETAU
 31   FORMAT(2X,F5.3,10X,F5.3,10X,F5.3)
      WRITE(2,22)
 22   FORMAT(2X,'BETA1',10X,'BETA2')
      WRITE(2,32)BETA1,BETA2
 32   FORMAT(2X,F5.3,10X,F5.3)
      WRITE(2,23)
 23   FORMAT(2X,'CONA',11X,'ALPHA')
      WRITE(2,33)CONA,ALPHA
 33   FORMAT(2X,F11.3,4X,F11.3)
      WRITE(2,24)
 24   FORMAT(2X,'AKS')
      WRITE(2,34)AKS
 34   FORMAT(2X,F6.3)
      WRITE(2,25)
 25   FORMAT(2X,'DELT',11X,'DELZ')
      WRITE(2,35)DELT,DELZ
 35   FORMAT(2X,F10.8,5X,F6.3)
      WRITE(2,26)
 26   FORMAT(2X,'NTIME',10X,'NNODE')
```

```
      WRITE(2,36)NTIME,NNODE
36    FORMAT(I7,10X,I5)
      WRITE(2,75)
75    FORMAT(2X,'STORM AND INTERSTORM PERIODS')
      WRITE(2,76)LT1,LT2,LT3,LT4,LT5
76    FORMAT(5I12)
      WRITE(2,27)
27    FORMAT(/2X,'SOIL MOISTURE AT DIFFERENT NODES')
      WRITE(2,28)
28    FORMAT(/2X,'INITIAL CONDITIONS')
      WRITE(2,38)(THETA(I,1),I=1,NNODE)
38    FORMAT(5F12.6)
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 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
      RAIN=0.0
      EVAP=0.0
      RR=0.0
      VOL1=0.0
      DO 199 I = 2,NNODE-1
      VOL1=VOL1+THETA(I,1)*DELZ
199   CONTINUE
      SMI=THETA(1,1)*DELZ*0.5+VOL1+THETA(NNODE,1)*DELZ*0.5
      R=8.314E+7
      WM=18.0
      G=980.665
      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
C     GENERATION OF UPPER BOUNDARY CONDITION
C
      IF(J.LE.LT1)GO TO 300
      IF(J.GE.LT2.AND.J.LE.LT3)GO TO 300
      IF(J.GE.LT4.AND.J.LE.LT5)GO TO 300
      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
      THETA(1,1)=ALPHA*(THETAS-THETAR)/(ALPHA+
     1 ABS(H(1,1))**BETA2)+THETAR
      THETA(1,2)=THETA(1,1)
      THETAP(1,1)=THETA(1,1)
      THETAP(1,2)=THETA(1,1)
      GO TO 200
 300  THETA(1,1)=THETAU
      THETA(1,2)=THETAU
      THETAP(1,1)=THETAU
      THETAP(1,2)=THETAU
      H(1,1)=-(ALPHA*(THETAS-THETA(1,1))/(THETA(1,1)
     1 -THETAR))**(1./BETA2)
      H(1,2)=H(1,1)
      HP(1,1)=H(1,1)
      HP(1,2)=H(1,1)
 200  CONTINUE
C
      DO 500 I=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.)
 500  CONTINUE
C
      DO 600 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
      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)*((H(I+1,1)-
     2 H(I-1,1))/(2.*DELZ)-1.)
 600  CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      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+1,2)=B(I)
      DO 900 I=2,NNODE-1
      THETAP(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(HP(I,2))**
     1 BETA2)+THETAR
 900  CONTINUE
C
      DO 1000 I=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.)
 1000 CONTINUE
C
```

```
      DO 1100 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
      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)*((HP(I+1,2)-
     2 HP(I-1,2))/(2.*DELZ)-1.)+DELT/DELZ**2*.5*(H(I+1,1)-2.*
     3 H(I,1)+H(I-1,1))
1100  CONTINUE
C
      B(1)=B(1)-SUB(1)*H(1,2)
      B(NNODE-2)=B(NNODE-2)-SUP(NNODE-2)*H(NNODE,2)
      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+1,2)=B(I)
      DO 1400 I=2,NNODE-1
      THETA(I,2)=ALPHA*(THETAS-THETAR)/(ALPHA+ABS(H(I,2))**BETA2)+
     1 THETAR
1400  CONTINUE
C
      DO 1500 I = 1, NNODE
      HYDCON(I,2) = E4/(CONA+(ABS(H(I,2)))**BETA1)
1500  CONTINUE
C
      RR=RR-((HYDCON(NNODE-1,2)*HYDCON(NNODE,2))**0.5)*
     1 (((H(NNODE,2)-H(NNODE-1,2))/DELZ)-1.0)*DELT
      RINPUT=-((HYDCON(1,2)*HYDCON(2,2))**0.5)*
     1 (((H(2,2)-H(1,2))/DELZ)-1.0)*DELT
      IF(RINPUT.GT.0.0)RAIN=RAIN+RINPUT
      IF(RINPUT.LE.0.0)EVAP=EVAP+ABS(RINPUT)
      VOL2=0.0
      DO 99 I = 2,NNODE-1
      VOL2=VOL2+THETA(I,2)*DELZ
99    CONTINUE
      SMF=THETA(1,2)*DELZ*0.5+VOL2+THETA(NNODE,2)*DELZ*0.5
      DELSM=SMF-SMI
      RECH=RAIN-EVAP-DELSM
C
      IF (J.EQ.2)     GO TO 111
      IF (J.EQ.1701)  GO TO 111
      IF (J.EQ.2281)  GO TO 111
      IF (J.EQ.3601)  GO TO 111
      IF (J.EQ.5401)  GO TO 111
      IF (J.EQ.7201)  GO TO 111
      IF (J.EQ.9001)  GO TO 111
      IF (J.EQ.10801) GO TO 111
      IF (J.EQ.12601) GO TO 111
      IF (J.EQ.14401) GO TO 111
      IF (J.EQ.16201) GO TO 111
      IF (J.EQ.18001) GO TO 111
      IF (J.EQ.21601) GO TO 111
      IF (J.EQ.25201) GO TO 111
```

```
      IF (J.EQ.28801) GO TO 111
      IF (J.EQ.32401) GO TO 111
      IF (J.EQ.36001) GO TO 111
      GO TO 222
111   CONTINUE
      ITIME=J-1
      HOUR=ITIME*DELT
      WRITE(2,51)ITIME,HOUR
51    FORMAT(/2X,'TIME STEP =',I7,4X,'DURATION = ',F10.4,2X,'HOURS'/)
      WRITE(2,52)(THETA(I,2),I=1,NNODE)
52    FORMAT(5F12.6)
      WRITE(2,77)RAIN
77    FORMAT(/2X,'CUMULATIVE INFILTRATION             = ',F12.6,2X,'CM')
      WRITE(2,78)EVAP
78    FORMAT(2X,'CUMULATIVE EVAPORATION              = ',F12.6,2X,'CM')
      WRITE(2,79)DELSM
79    FORMAT(2X,'TOTAL INCREASE IN UZ SOIL MOISTURE  = ',F12.6,2X,'CM')
      WRITE(2,53)RECH
53    FORMAT(2X,'CUMULATIVE RECHARGE (WATER BALANCE) = ',F12.6,2X,'CM')
      WRITE(2,81)RR
81    FORMAT(2X,'CUMULATIVE RECHARGE (DARCY LAW)     = ',F12.6,2X,'CM')
      DIFF=RR-RECH
      WRITE(2,82)DIFF
82    FORMAT(2X,'DIFFERENCE BETWEEN TWO METHODS      = ',F12.6,2X,'CM')
222   CONTINUE
C
      DO 333 I = 2, NNODE-1
      THETA(I,1)=THETA(I,2)
      H(I,1)=H(I,2)
333   CONTINUE
C
400   CONTINUE
      STOP
      END
C
      SUBROUTINE TRID(M,SUP,SUB,DIAG,B)
      DIMENSION SUP(500),SUB(500),DIAG(500),B(500)
      N=M
      NN=N-1
      SUP(1)=SUP(1)/DIAG(1)
      B(1)=B(1)/DIAG(1)
      DO 51 I=2,N
      II=I-1
      DIAG(I)=DIAG(I)-SUP(II)*SUB(II)
      IF (I.EQ.N) GO TO 51
      SUP(I)=SUP(I)/DIAG(I)
51    B(I)=(B(I)-SUB(II)*B(II))/DIAG(I)
      DO 52 K=1,NN
      I=N-K
52    B(I)=B(I)-SUP(I)*B(I+1)
      RETURN
      END
```

0.075	0.287	0.286		
4.740	3.960			
1175000.000	1611000.000			
34.000				
0.00083333	4.000			
36001	76			
3601	7201	10801	14401	18001
25.00				
00.75				
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.106202	0.114577
0.125431	0.139324	0.156679	0.177478	0.200872
0.224933	0.246971	0.264515	0.276398	0.283064
0.286000				

ESTIMATION OF GROUND WATER RECHARGE
IMPLICIT SCHEME WITH IMPLICIT LINEARIZATION
(PREDICTION - CORRECTION)

TEMPERATURE IN CENTIGRADE
25.00
RELATIVE HUMIDITY OF THE AIR
0.750

THETAR THETAS THETAU
0.075 0.287 0.286
BETA1 BETA2
4.740 3.960
CONA ALPHA
1175000.000 1611000.000
AKS
34.000
DELT DELZ
0.00083333 4.000
NTIME NNODE
36001 76
STORM AND INTERSTORM PERIODS
3601 7201 10801 14401 18001

SOIL MOISTURE AT DIFFERENT NODES

INITIAL CONDITIONS

0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.106202	0.114577
0.125431	0.139324	0.156679	0.177478	0.200872
0.224933	0.246971	0.264515	0.276398	0.283064
0.286000				

TIME STEP = 1 DURATION = 0.0008 HOURS

0.286000	0.120462	0.100068	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100000	0.100000	0.100000
0.100000	0.100000	0.100024	0.106203	0.114577
0.125431	0.139324	0.156679	0.177478	0.200872
0.224933	0.246971	0.264515	0.276398	0.283065
0.286000				

CUMULATIVE INFILTRATION = 0.030399 CM
CUMULATIVE EVAPORATION = 0.000000 CM
TOTAL INCREASE IN UZ SOIL MOISTURE = 0.454220 CM
CUMULATIVE RECHARGE (WATER BALANCE) = 0.423821 CM
CUMULATIVE RECHARGE (DARCY LAW) = 0.000000 CM
DIFFERENCE BETWEEN TWO METHODS = 0.423821 CM

TIME STEP = 1700 DURATION = 1.4167 HOURS

0.286000	0.285999	0.285998	0.285997	0.285995
0.285994	0.285992	0.285991	0.285989	0.285987
0.285985	0.285983	0.285980	0.285978	0.285975
0.285972	0.285969	0.285966	0.285962	0.285958
0.285954	0.285950	0.285945	0.285940	0.285934
0.285928	0.285922	0.285915	0.285908	0.285900
0.285891	0.285882	0.285872	0.285861	0.285849
0.285837	0.285823	0.285808	0.285792	0.285774
0.285754	0.285733	0.285710	0.285684	0.285656
0.285625	0.285591	0.285552	0.285509	0.285461
0.285408	0.285347	0.285277	0.285197	0.285106
0.284999	0.284873	0.284723	0.284543	0.284322
0.284045	0.283689	0.283218	0.282568	0.281618
0.280116	0.277435	0.271644	0.255209	0.221607
0.228235	0.247662	0.264711	0.276456	0.283077
0.286000				

CUMULATIVE INFILTRATION = 48.923470 CM
CUMULATIVE EVAPORATION = 0.000000 CM
TOTAL INCREASE IN UZ SOIL MOISTURE = 49.821098 CM
CUMULATIVE RECHARGE (WATER BALANCE) = -0.897629 CM
CUMULATIVE RECHARGE (DARCY LAW) = 0.012946 CM
DIFFERENCE BETWEEN TWO METHODS = 0.910575 CM

TIME STEP = 2280 DURATION = 1.9000 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	64.760979	CM
CUMULATIVE EVAPORATION	=	0.000000	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962112	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	13.798866	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	14.655499	CM
DIFFERENCE BETWEEN TWO METHODS	=	0.856633	CM

TIME STEP = 3600 DURATION = 3.0000 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	100.794304	CM
CUMULATIVE EVAPORATION	=	0.000000	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962227	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	49.832077	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	50.693687	CM
DIFFERENCE BETWEEN TWO METHODS	=	0.861610	CM

TIME STEP = 5400 DURATION = 4.5000 HOURS

0.075018	0.081442	0.092557	0.104248	0.115486
0.125846	0.135194	0.143543	0.150974	0.157590
0.163498	0.168793	0.173559	0.177869	0.181785
0.185358	0.188633	0.191647	0.194431	0.197011
0.199410	0.201647	0.203741	0.205704	0.207550
0.209290	0.210933	0.212488	0.213963	0.215364
0.216697	0.217967	0.219179	0.220337	0.221446
0.222508	0.223527	0.224505	0.225446	0.226352
0.227224	0.228066	0.228877	0.229662	0.230420
0.231153	0.231863	0.232552	0.233219	0.233866
0.234494	0.235105	0.235698	0.236275	0.236836
0.237383	0.237917	0.238439	0.238950	0.239454
0.239956	0.240463	0.240990	0.241564	0.242232
0.243078	0.244248	0.245988	0.248670	0.252758
0.258591	0.265896	0.273465	0.279738	0.283867
0.286000				

CUMULATIVE INFILTRATION	=	100.794304	CM
CUMULATIVE EVAPORATION	=	0.203161	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	28.710159	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	71.880981	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	71.105492	CM
DIFFERENCE BETWEEN TWO METHODS	=	-0.775490	CM

TIME STEP = 7200 DURATION = 6.0000 HOURS

0.286000	0.081804	0.083107	0.088646	0.094721
0.100939	0.107135	0.113187	0.119014	0.124565
0.129818	0.134763	0.139405	0.143755	0.147830
0.151647	0.155225	0.158581	0.161733	0.164698
0.167490	0.170124	0.172611	0.174965	0.177194
0.179310	0.181320	0.183233	0.185056	0.186795
0.188457	0.190047	0.191569	0.193028	0.194429
0.195775	0.197069	0.198315	0.199515	0.200672
0.201789	0.202868	0.203910	0.204919	0.205895
0.206840	0.207756	0.208646	0.209508	0.210346
0.211160	0.211952	0.212722	0.213472	0.214203
0.214916	0.215613	0.216297	0.216970	0.217638
0.218311	0.219006	0.219750	0.220596	0.221632
0.223012	0.224996	0.227991	0.232575	0.239375
0.248643	0.259578	0.270187	0.278432	0.283537
0.286000				

CUMULATIVE INFILTRATION	=	100.809769	CM
CUMULATIVE EVAPORATION	=	0.260641	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	21.996593	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	78.552536	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	77.766396	CM
DIFFERENCE BETWEEN TWO METHODS	=	-0.786140	CM

TIME STEP = 9000 DURATION = 7.5000 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	152.328125	CM
CUMULATIVE EVAPORATION	=	0.260641	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962212	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	101.105278	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	98.840401	CM
DIFFERENCE BETWEEN TWO METHODS	=	-2.264877	CM

TIME STEP = 10800 DURATION = 9.0000 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	201.464478	CM
CUMULATIVE EVAPORATION	=	0.260641	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962227	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	150.241623	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	147.976761	CM
DIFFERENCE BETWEEN TWO METHODS	=	-2.264862	CM

TIME STEP = 12600 DURATION = 10.5000 HOURS

0.075018	0.081442	0.092557	0.104248	0.115486
0.125846	0.135194	0.143543	0.150974	0.157590
0.163498	0.168793	0.173559	0.177869	0.181785
0.185358	0.188633	0.191647	0.194431	0.197011
0.199410	0.201647	0.203741	0.205704	0.207550
0.209290	0.210933	0.212488	0.213963	0.215364
0.216697	0.217967	0.219179	0.220337	0.221446
0.222508	0.223527	0.224505	0.225446	0.226352
0.227224	0.228066	0.228877	0.229662	0.230420
0.231153	0.231863	0.232552	0.233219	0.233866
0.234494	0.235105	0.235698	0.236275	0.236836
0.237383	0.237917	0.238439	0.238950	0.239454
0.239956	0.240463	0.240990	0.241564	0.242232
0.243078	0.244248	0.245988	0.248670	0.252758
0.258591	0.265896	0.273465	0.279738	0.283867
0.286000				

CUMULATIVE INFILTRATION	=	201.464478	CM
CUMULATIVE EVAPORATION	=	0.463803	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	28.710159	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	172.290512	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	168.388275	CM
DIFFERENCE BETWEEN TWO METHODS	=	-3.902237	CM

TIME STEP = 14400 DURATION = 12.0000 HOURS

0.286000	0.081804	0.083107	0.088646	0.094721
0.100939	0.107135	0.113187	0.119014	0.124565
0.129818	0.134763	0.139405	0.143755	0.147830
0.151647	0.155225	0.158581	0.161733	0.164698
0.167490	0.170124	0.172611	0.174965	0.177194
0.179310	0.181320	0.183233	0.185056	0.186795
0.188457	0.190047	0.191569	0.193028	0.194429
0.195775	0.197069	0.198315	0.199515	0.200672
0.201789	0.202868	0.203910	0.204919	0.205895
0.206840	0.207756	0.208646	0.209508	0.210346
0.211160	0.211952	0.212722	0.213472	0.214203
0.214916	0.215613	0.216297	0.216970	0.217638
0.218311	0.219006	0.219750	0.220596	0.221632
0.223012	0.224996	0.227991	0.232575	0.239375
0.248643	0.259578	0.270187	0.278432	0.283537
0.286000				

CUMULATIVE INFILTRATION	=	201.479935	CM
CUMULATIVE EVAPORATION	=	0.521282	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	21.996593	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	178.962051	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	175.049026	CM
DIFFERENCE BETWEEN TWO METHODS	=	-3.913025	CM

TIME STEP = 16200 DURATION = 13.4999 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	252.998154	CM
CUMULATIVE EVAPORATION	=	0.521282	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962212	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	201.514664	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	196.122986	CM
DIFFERENCE BETWEEN TWO METHODS	=	-5.391678	CM

TIME STEP = 18000 DURATION = 14.9999 HOURS

0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000	0.286000	0.286000	0.286000	0.286000
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.521282	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	50.962227	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	250.676819	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	245.259338	CM
DIFFERENCE BETWEEN TWO METHODS	=	-5.417480	CM

TIME STEP = 21600 DURATION = 17.9999 HOURS

0.075018	0.077903	0.082928	0.088646	0.094721
0.100939	0.107135	0.113187	0.119014	0.124565
0.129818	0.134763	0.139405	0.143755	0.147830
0.151647	0.155225	0.158581	0.161733	0.164698
0.167490	0.170124	0.172611	0.174965	0.177194
0.179310	0.181320	0.183233	0.185056	0.186795
0.188457	0.190047	0.191569	0.193028	0.194429
0.195775	0.197069	0.198315	0.199515	0.200672
0.201789	0.202868	0.203910	0.204919	0.205895
0.206840	0.207756	0.208646	0.209508	0.210346
0.211160	0.211952	0.212722	0.213472	0.214203
0.214916	0.215613	0.216297	0.216970	0.217638
0.218311	0.219006	0.219750	0.220596	0.221632
0.223012	0.224996	0.227991	0.232575	0.239375
0.248643	0.259578	0.270187	0.278432	0.283537
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.781947	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	21.558308	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	279.820038	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	272.331665	CM
DIFFERENCE BETWEEN TWO METHODS	=	-7.488373	CM

TIME STEP = 25200 DURATION = 20.9999 HOURS

0.075018	0.076180	0.078125	0.080400	0.082936
0.085694	0.088636	0.091724	0.094920	0.098187
0.101491	0.104803	0.108097	0.111353	0.114553
0.117685	0.120739	0.123708	0.126588	0.129377
0.132073	0.134677	0.137190	0.139614	0.141952
0.144206	0.146379	0.148474	0.150495	0.152444
0.154324	0.156140	0.157893	0.159586	0.161223
0.162806	0.164338	0.165820	0.167256	0.168647
0.169995	0.171302	0.172570	0.173801	0.174997
0.176158	0.177287	0.178385	0.179453	0.180492
0.181505	0.182491	0.183452	0.184392	0.185311
0.186210	0.187095	0.187971	0.188844	0.189729
0.190647	0.191636	0.192758	0.194118	0.195887
0.198344	0.201921	0.207233	0.215018	0.225849
0.239494	0.254319	0.267687	0.277503	0.283315
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.845007	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	13.595875	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	287.719421	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	279.541992	CM
DIFFERENCE BETWEEN TWO METHODS	=	-8.177429	CM

TIME STEP = 28800 DURATION = 23.9999 HOURS

0.075018	0.075666	0.076705	0.077920	0.079286
0.080793	0.082435	0.084201	0.086079	0.088058
0.090123	0.092261	0.094457	0.096697	0.098968
0.101258	0.103555	0.105851	0.108137	0.110403
0.112644	0.114856	0.117033	0.119172	0.121269
0.123325	0.125335	0.127301	0.129221	0.131095
0.132924	0.134707	0.136446	0.138141	0.139793
0.141402	0.142971	0.144499	0.145988	0.147439
0.148853	0.150232	0.151577	0.152888	0.154167
0.155415	0.156632	0.157821	0.158981	0.160115
0.161223	0.162306	0.163366	0.164405	0.165425
0.166431	0.167428	0.168425	0.169436	0.170483
0.171604	0.172858	0.174341	0.176210	0.178708
0.182211	0.187264	0.194580	0.204900	0.218585
0.234954	0.251899	0.266602	0.277116	0.283226
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.882073	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	8.840603	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	292.437622	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	283.713745	CM
DIFFERENCE BETWEEN TWO METHODS	=	-8.723877	CM

TIME STEP = 32400 DURATION = 26.9999 HOURS

0.075018	0.075436	0.076080	0.076830	0.077673
0.078607	0.079631	0.080743	0.081939	0.083217
0.084572	0.085998	0.087489	0.089038	0.090640
0.092286	0.093969	0.095683	0.097421	0.099177
0.100945	0.102718	0.104492	0.106264	0.108027
0.109780	0.111518	0.113239	0.114941	0.116621
0.118278	0.119911	0.121518	0.123098	0.124652
0.126178	0.127676	0.129147	0.130591	0.132006
0.133394	0.134754	0.136087	0.137394	0.138676
0.139932	0.141162	0.142369	0.143551	0.144711
0.145849	0.146966	0.148065	0.149148	0.150218
0.151280	0.152344	0.153422	0.154537	0.155719
0.157022	0.158528	0.160365	0.162735	0.165941
0.170430	0.176817	0.185844	0.198186	0.213982
0.232211	0.250494	0.265991	0.276902	0.283177
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.907419	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	5.589195	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	295.663696	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	286.494263	CM
DIFFERENCE BETWEEN TWO METHODS	=	-9.169434	CM

TIME STEP = 36000 DURATION = 29.9999 HOURS

0.075018	0.075312	0.075748	0.076253	0.076820
0.077447	0.078136	0.078886	0.079698	0.080572
0.081505	0.082496	0.083542	0.084642	0.085791
0.086987	0.088226	0.089504	0.090817	0.092160
0.093531	0.094923	0.096336	0.097763	0.099201
0.100647	0.102098	0.103550	0.105001	0.106449
0.107892	0.109327	0.110752	0.112166	0.113568
0.114957	0.116331	0.117689	0.119030	0.120354
0.121660	0.122948	0.124217	0.125469	0.126702
0.127916	0.129111	0.130288	0.131448	0.132590
0.133716	0.134827	0.135926	0.137016	0.138100
0.139187	0.140288	0.141423	0.142620	0.143920
0.145391	0.147137	0.149316	0.152168	0.156041
0.161433	0.168996	0.179469	0.193433	0.210829
0.230388	0.249585	0.265603	0.276768	0.283147
0.286000				

CUMULATIVE INFILTRATION	=	302.160309	CM
CUMULATIVE EVAPORATION	=	0.926125	CM
TOTAL INCREASE IN UZ SOIL MOISTURE	=	3.224998	CM
CUMULATIVE RECHARGE (WATER BALANCE)	=	298.009186	CM
CUMULATIVE RECHARGE (DARCY LAW)	=	288.475342	CM
DIFFERENCE BETWEEN TWO METHODS	=	-9.533844	CM

DIRECTOR	:	S. M. SETH
DIVISIONAL HEAD	:	G. C. MISHRA
SCIENTIST	:	CHANDRA PRAKASH KUMAR
DRAWING STAFF	:	NARENDRA KUMAR
		N. K. VARSHNEY
DOCUMENTATION STAFF	:	S. P. MODI