TR (AR) –   /2000-2001

SIMULATION OF SOIL MOISTURE MOVEMENT IN A HARD ROCK WATERSHED USING SWIM MODEL

NATIONAL INSTITUTE OF HYDROLOGY
JAL VIGYAN BHAWAN
ROORKEE - 247 667 (UTTARANCHAL)
INDIA

2000 - 2001

PREFACE

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.

This report entitled "*Simulation of Soil Moisture Movement in a Hard Rock Watershed using SWIM Model*" is a part of the research activities of 'Hard Rock Regional Centre' of the Institute. The purpose of this study is to simulate the soil moisture movement in a hard rock watershed through a numerical model and determine the ground water recharge from rainfall. The study has been carried out by Mr. C. P. Kumar, Scientist 'E1' and Dr. B. K. Purandara, Scientist 'B', Hard Rock Regional Centre, National Institute of Hydrology, Belgaum.

(**K. S. Ramasastri**)
Director

CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

A very large fraction of the water falling as rain on the land surfaces of the earth or applied irrigation water moves through unsaturated soil during the subsequent processes of infiltration, drainage, evaporation, and the absorption of soil-water by plant roots. The water movements in the unsaturated zone, together with the water holding capacity of this zone, are very important for the water demand of the vegetation, as well as for the recharge of the ground water storage. A fair description of the flow in the unsaturated zone is also crucial for predictions of the movement of pollutants into ground water aquifers.

A number of simulation models are available for investigating the soil water balance. SWIM (Soil Water Infiltration and Movement) is a physically based, isothermal, one dimensional model of water flow through the soil coupled with a simple crop water extraction model in which the growth of the canopy and of the root system is a predetermined input. SWIM is driven by rainfall and potential evaporation, and so appears to be more appropriate than few other similar models if the available meteorological data are limited.

The present study aims at modelling of soil moisture movement in Barchi watershed (Karnataka) using SWIM. Field and laboratory investigations were carried out to determine the saturated hydraulic conductivity at eight locations using Guelph Permeameter and soil moisture retention characteristics using the Pressure Plate Apparatus. The van Genuchten parameters of soil moisture retention function and hydraulic conductivity function were obtained through non-linear regression analysis. Daily rainfall and evaporation data of Barchi for the period 1996-97 to 1999-2000 were used for the simulations. Water balance components like runoff, evapotranspiration and drainage (groundwater recharge from rainfall) were determined through SWIM.

1.0 INTRODUCTION

Most of the processes involving soil-water interactions in the field, and particularly the flow of water in the rooting zone of most crop plants, occur while the soil is in an unsaturated condition. Unsaturated flow processes are in general complicated and difficult to describe quantitatively, since they often entail changes in the state and content of soil water during flow. Such changes involve complex relations among the variable soil wetness, suction, and conductivity, whose inter-relations may be further complicated by hysteresis. The formulation and solution of unsaturated flow problems very often require the use of indirect methods of analysis, based on approximations or numerical techniques. For this reason, the development of rigorous theoretical and experimental methods for treating these problems was rather late in coming. In recent decades, however, unsaturated flow has become one of the most important and active topics of research and this research has resulted in significant theoretical and practical advances.

Subsurface formations containing water may be divided vertically into several horizontal zones according to how large a portion of the pore space is occupied by water. Essentially, we have a zone of saturation in which all the pores are completely filled with water, and an overlaying zone of aeration in which the pores contain both gases (mainly air and water vapour) and water. The latter zone is called the unsaturated zone. Sometimes the term soil water is used for the water in this zone.

The vertical movement of soil moisture in the liquid phase between the surface and the water table can be subdivided into the following three categories according to predominant forces involved.

<u>Infiltration and exfiltration</u>

Alternate wetting and drying of soil surface during consecutive storm and interstorm periods will cause a penetration of the medium by an unsteady wave like diffusion of liquid soil moisture into the soil during wet surface (storm) periods under the complementary effects of capillarity and gravity and out of the soil during dry surface (interstorm) periods when capillarity opposes gravity. With increasing depth of penetration, diffusion reduces the soil moisture gradients and thus reduces the effect of capillarity until moisture movement becomes dominated by gravity. The depth at which surface induced capillary forces become negligible determines the penetration depth of the surface process and is used to define the thickness of the zone of soil moisture. The presence of transpiring vegetation adds another mechanism for moisture extraction distributed over a depth which is related to root structure.

Percolation

Liquid soil moisture moves out of the bottom of the zone of soil moisture and percolates downward under the domination of gravity forces until it encounters the increasing soil moisture gradients lying above the water table. At some depth upward capillary forces will be prominent defining the bottom of this intermediate zone.

Capillary rise

Between the water table and the intermediate zone there is a capillary fringe in which gravity and capillarity again jointly govern the liquid soil moisture movement.

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.

For analytical studies on soil moisture regime, critical review and accurate assessment of the different controlling factors is necessary. The controlling factors of soil moisture may be classified under two main groups viz. climatic factors and soil factors. Climatic factors include precipitation data containing rainfall intensity, storm duration, interstorm period, temperature of soil surface, relative humidity, radiation, evaporation, and evapotranspiration. The soil factors include soil matric potential and water content relationship, hydraulic conductivity and water content relationship of the soil, saturated hydraulic conductivity, and effective medium porosity. Besides these factors, the information about depth to water table is also required.

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 theory for transient isothermal flow of water into nonswelling unsaturated soil is well understood and has been developed to a large extent in terms of solutions of the non-linear Richards equation. In the field, the description of infiltration is highly complicated since the initial and boundary conditions are usually not constant while the soil characteristics may vary with time and space. In view of this, most efforts in recent past, have been concentrated on seeking numerical solutions.

The governing partial differential flow equation can be interpreted numerically by a finite difference, a finite element or a boundary element technique. Then a discretization scheme is applied for a system of nodal points that is superimposed on the soil depth-time region under consideration. Implementing the appropriate initial and boundary conditions then leads to a set of (linear) algebraic equations that can be solved by different methods. The operation by means of such a mathematical model is termed simulation, while the model is called simulation model.

The soil water movement may be modelled mathematically from bases provided by:

(a) the soil moisture characteristic,

(b) equations describing the volume flux of water and water vapour in response to potential gradients, and

(c) the law of continuity of matter and additionally, in the case of evaporation, the law of continuity of heat energy.

The objective of the present study is to simulate the movement of soil moisture in Barchi watershed (sub-basin of Kali river in North Kanara district of Karnataka) using the SWIM model. The study includes

- ♦ Measurement and determination of saturated hydraulic conductivity and soil moisture retention characteristics.
- ♦ Modelling of soil moisture movement using the SWIM model. Daily rainfall and evaporation data of Barchi for the period 1996-97 to 1999-2000 were used for the study.
- ♦ Determination of water balance components like runoff, evapotranspiration and drainage (recharge to groundwater from rainfall).

The SWIM (Soil Water Infiltration and Movement) is a software package developed by Division of Soils, CSIRO, Australia (Verburg et al., 1996) for simulating infiltration, evapotranspiration, and redistribution. The major features of the model include the ability to deal with:

- ➢ Layered and gradational soils such as occur in field soils where hydraulic properties vary with depth down the profile, either abruptly or gradually.
- ➢ Saturated/unsaturated conditions as can occur at layer interfaces, which result in locally perched water.
- ➢ Surface ponding as can occur under high rainfall intensities.
- ➢ Surface runoff, where 'excess' water can be removed from the system.
- ➢ Surface sealing, where the properties of the surface may vary directly as a function of rainfall energy, and hence as a function of time.
- ➢ Rainfall dynamics, so that real storm intensities (down to 1-minute resolution and below) can be simulated.
- ➢ Solute transport.
- ➢ Vapour flow, hysteresis, bypass flow, osmotic effects, and potential subsurface downslope flow.

- ‘Cultivations’ or ‘disturbances’ of the soil surface which enable the application of dry fertilizer (solute) and resetting of the surface conductance and surface roughness values at specified times.

SWIM has already been used world-wide in a variety of studies. It was originally written assuming that a preprocessor would be used to interface with the user. A preprocessor is not yet available; hence current input facilities (Version 2.1) are less than ideal.

2.0 STUDY AREA

The Barchi watershed upstream of Barchi is located in the leeward side of western ghat and is a sub-basin of Kali river. It lies in Haliyala taluk of Karwar (North Kanara) district in Karnataka. The location and drainage system of Barchi watershed is shown in Figure 1.

The Barchinala stream originates from Thavargatti in Belgaum district at an altitude of about 734 m, 20 km north of Dandeli and flows through North Kanara district of Karnataka State. The catchment is relatively short in width and river flows in a southerly direction and joins the main Barchi river near the gauging site. The geographical area covered by Barchi watershed is 21.126 km^2. The watershed lies between $74^o36'$ and $74^o39'$ East longitudes, and $15^o18'$ and $15^o24'$ North latitudes.

High land region consists of dissection of high hills and ridges forming part of the foot hills of western ghats. It consists of steep hills and valleys intercepted with thick forest. The slopes of the ghats are covered with dense deciduous forest. Forest cover occupies around 80% of the study area. The watershed is mainly covered with Bamboo, Teak and mixed plantations. The brownish and fine-grained soils are the principal types of soils found in the area. The following land uses were observed at the locations of field studies (Figure 1):

1. Bamboo plantation (near gauging site)
2. Teak plantation (ridge)
3. Mixed forest, disturbed fire.
4. Mixed forest with bamboo
5. Soil profile with high litter content
6. Agricultural land
7. Mixed forest with high litter content
8. Bamboo and mixed forest

The stream gauging site is located at an elevation of 480 m, where the nala crosses Dandeli-Thavargatti road, about 5 km from Dandeli. The stream is a 4^{th} order stream and joins main Barchi river downstream of the gauging site. A full fledged meteorological station, maintained by Water Resources Development Organisation (WRDO), Karnataka, is located near the gauging site.

The Barchi raingauge station is located at 15^{o}18’ N and 74^{o}37’ E. Average annual rainfall for the watershed is 1500 mm, majority of which occurs during the south-west monsoon period. Figures 2 to 5 present the variation of monthly rainfall and evaporation in Barchi watershed during the years 1996-97 to 1999-2000 respectively.

Depth to water table varies between 4 to 12 metres during pre- and post-monsoon periods. The yield of borewells in the study area is found to vary between 120 gallons per hour to 1170 gallons per hour.

3.0 METHODOLOGY

3.1 General

The present study involves modelling of soil moisture movement in Barchi watershed using the SWIM model. The following steps were undertaken for the study.

* **Field investigations** – measurement of saturated hydraulic conductivity at 8 locations using Guelph Permeameter and soil sampling.

* **Laboratory investigations –** Determination of saturated moisture content, and soil moisture retention characteristics using the Pressure Plate Apparatus.

* **Modelling** of soil moisture movement using the SWIM model. Daily rainfall and evaporation data of Barchi for the period 1996-97 to 1999-2000 were used for the study. Water balance components like runoff, evapotranspiration and drainage (recharge to groundwater from rainfall) were determined through SWIM.

Details of equipment and procedures adopted for field and laboratory investigations are presented below. Description of SWIM model is discussed in the next chapter.

3.2 Soil Moisture Characteristics

Quantitative measurements of soil physical properties are required for many purposes. In the area of land management, one may wish to know whether a particular management scheme will increase or decrease infiltration, runoff, erosion, leaching, salinization etc. We may need to predict material transport, such as the depth to a wetting front, position of a seepage face, time of arrival of a tracer plume, cumulative evaporation etc.

Any measurement of soil water in the field depends upon sampling at a given location, both in area and depth of soil profile, at a given time or times. These samples are then used to estimate the water condition of the entire area. Many methods are sufficiently accurate to measure the water condition in a given sample at a given time. Difficulty comes when one tries to apply these conditions to a large area or at a different time. In reality, the water condition measured is a transient one in a system that is continuously changing in three-dimensional space and time and the situation would likely be different at any other location at the same time, or at the same location at a different time.

In order to evaluate completely the condition of water in soil, one must know the energy of the water, the amount of water in the soil, and how these conditions change in space and time. This requires a complete understanding of water movement and flow in soils. Such complete evaluations of soil water conditions are not easily made, and are available only under controlled laboratory conditions.

There are two general reasons for measuring soil water. One is to determine the moisture content of a soil, that is, the amount of water contained in a unit mass or volume of soil. This information is necessary to calculate the water needed to restore the soil water in the root zone of the crop. The second reason is to determine the magnitude of the soil water potential, which is the negative of the work that must be done to remove a unit amount of the most loosely held water.

Plant response to water appears to be more closely related to the water potential than any other single factor, although the velocity of movement of water to the absorbing root is an important consideration. This movement rate is strongly related to the potential. Because of this relation, one desires to know the potential of the soil water whenever he is concerned about plant response. Knowledge of the soil water potential is also desired by irrigators since it indicates directly when water should be applied.

Prediction of infiltration is important in the design of irrigation areas and for the estimation of runoff in catchment management studies. Many predictive models exist and various methods have been employed in measuring infiltration behaviour. The proper evaluation of infiltration behaviour depends on knowledge of the hydrological soil properties.

Saturated hydraulic conductivity and unsaturated hydraulic conductivity are related to the degree of resistance from soil particles when water flows in pores. These resistances are affected by the forms, sizes, branchings, jointings, and tortuosities of pores as well as viscosity of water. In addition, unsaturated hydraulic conductivity is affected markedly by the volumetric water content of soil.

The relation between matric potential and volumetric water content in a soil is termed as the soil moisture characteristic curve because the curve is characteristic of each soil. The differences among soil moisture characteristic curves are attributed primarily to the differences in pore size distribution among soils. These curves are sensitive to the changes in bulk densities and disturbances of soil structures. In addition, the curves generally show hysteresis according to the wetting or drying of soils.

3.3 Soil Moisture Retention Curves

The graph giving the relation between soil moisture tension and soil moisture content is called moisture retention curve or soil moisture characteristic. If the tension is expressed as the logarithmic value of cm water, the graph is referred to as a pF-curve. Moisture retention curves are used:

- to determine an index of the available moisture in soil (the portion of water that can be readily absorbed by plant roots) and to classify soils accordingly, e.g. for irrigation purposes,
- to determine the drainable pore space (effective pore space, effective porosity, specific yield) for drainage design,
- to check changes in the structure of a soil, e.g. caused by tillage, mixing of soil layers etc.,
- to ascertain the relation between soil moisture tension and other physical properties of a soil (e.g. capillary conductivity, thermal conductivity, clay and organic matter content).

Clay soils show a slow and regular decrease in water content with increasing pF tension. Sandy soils may show only a slight decrease in moisture content in the lower pF range till the point where only a small rise in pF causes a considerable discharge of water due to a relatively large number of pores in a particular diameter range. The intersection point of the curve with the volumetric water content axis (tension: 1 cm water, pF = 0) gives the water content of the soil under nearly saturated conditions, which means that this point almost indicates the total pore space percentage (if no air entrapment has taken place). The zero moisture content is based on the oven-dry condition (105 ^{o}C), corresponding to a pF of approximately 7.

To construct the moisture retention curve of a soil sample, the moisture content of that sample must be measured. This is done by equilibrating the moist soil sample at a succession of known pF values and each time determining the amount of moisture that is retained. If the equilibrium moisture content (expressed preferably as volume percentage) is plotted against the corresponding tension (pF), the moisture retention curve (pF-curve) can be drawn. There is no single method of inducing the whole range of tensions from pF = - ∞ (total saturation) to pF = 7 (oven dry).

The ceramic plates equipment is suitable for determination of pF-curves in the pF range of 2.0-4.2 (0.1-15 bar of suction). Soil moisture is removed from the soil samples by raising air pressure in an extractor. A porous ceramic plate serves as a hydraulic link for water to move from the soil to the exterior of the extractor. The high-pressure air will not flow through the pores in the plate since the pores are filled with water. The smaller the pore size, the higher the pressure that can be exerted before air will pass through. During an experimental run, at any set pressure in the extractor, soil moisture will flow around each of the soil particles and out through the ceramic plate and outflow tube. Equilibrium is reached when water flow from the outflow tube ceases. At equilibrium, there is an exact relationship between the air pressure in the extractor and the soil suction (and hence the moisture content) in the samples. Accuracy of equilibrium values will be no more accurate than the regulation of air supply; therefore the pressure control panel has independent double regulators.

For each soil type, the characteristic pF-curve may be developed. These curves relate the soil suction to its moisture content. This relationship is important in studies of soil moisture movement and quantity and availability of soil moisture for plant growth.

3.3.1 Pressure Plate Apparatus

It consists of a ceramic pressure plate cell mounted in a pressure vessel, with the outflow tube running through the vessel wall to the atmosphere and soil sample held in place on the porous ceramic surface of the cell. Each ceramic pressure plate cell consists of a porous ceramic plate covered on one side by a thin neoprene diaphragm sealed to the edges of the ceramic plate. An internal screen between the plate and diaphragm provides a passage for flow of water. An outlet stem running through the plate connects this passage to an outflow tube fitting which connects to the atmosphere outside of the extractor.

To use the ceramic pressure plate cell, one or more soil samples are placed on the porous ceramic surface and held in place by retaining rings of appropriate height. The soil samples, together with the porous ceramic plate, are then saturated with water. This is usually done by allowing excess water to stand on the surface of the cell for several hours. When the saturation is complete, the cell can be mounted in the pressure vessel. Air pressure is used to effect extraction of moisture from the soil samples under controlled conditions.

As soon as air pressure inside the chamber is raised above the atmospheric pressure, higher pressure inside the chamber forces excess water through the microscopic pores in the ceramic plate and out through the outlet stem. The high pressure air, however, will not flow through the pores in the ceramic plate since the pores are filled with water and the surface tension of water, at the gas-liquid interface at each of the pores, supports the pressure similar to a flexible rubber diaphragm.

The maximum air pressure that any given wetted porous ceramic plate can stand before letting air pass through the pores, is determined by the diameter of pore. The smaller the pore sizes, the higher the pressure needed for air to pass through. The pressure value that finally breaks down the water meniscus, is called the “bubbling pressure” or the “air entry value” for the porous plate. Pressure plate cells must always be used at air pressure extraction values below the “bubbling pressure” or “air entry value” for the cell.

During an experimental run, for any set air pressure in the extractor, soil moisture will flow from around each of the soil particles and out through the ceramic plate until the effective curvature of water films throughout the soil are same as at the pores in the plate. When this occurs, an equilibrium is reached and the flow of moisture ceases. When air pressure in the extractor is increased, flow of soil moisture from the samples starts again and continues until a new equilibrium is reached. At equilibrium, there is an exact relationship between the air pressure in the extractor and the soil suction (and hence the moisture content) in the samples. For example, if air pressure in the extractor is maintained at 1/3 bar, the soil suction in the samples at equilibrium will be 1/3 bar. If air pressure is maintained at 1 bar, the soil suction at equilibrium will be 1 bar.

The 1 bar ceramic plate cells are ideal for the routine determination of the 1/10 bar and 1/3 bar percentages in the cataloging of soils as well as all other soil moisture equilibrium studies in the 0-1 bar range of soil suction. The bubbling pressure of these cells is in excess of 1 bar. These cells also have the highest permeability amongst the pressure plate cells and hence time to reach the equilibrium will be the shortest possible. The 3 bar ceramic plate cells can also be used for determination of the 1/10 bar and 1/3 bar percentages as well as soil moisture equilibrium studies in the extended range of 0-3 bars of soil suction. Bubbling pressure of these cells is in excess of 3 bars. The 15 bar ceramic plate cells are not suitable for work in the 0-1 bar range of soil suction due to their small pore size. They can, however, be used effectively for soil moisture equilibrium studies in the 1-5 bar range of soil suction. Bubbling pressure of these cells is in excess of 15 bars. To use full range, these cells must be used in the 15 bar ceramic plate extractor.

The various pressure plate cells are not suitable for extracting solution from soils for chemical analysis. The immense surface area within the porous ceramic plate can cause disturbance and contamination of the soil solution. Where experiments for moisture equilibrium studies are being run, it is desirable to keep the sample heights small in order to reach equilibrium in reasonable time. The time required to reach equilibrium varies as the square of sample height. For example, a soil sample 2 cm high will require four times as long to reach equilibrium as a sample of 1 cm high. Whenever possible, soil sample heights should be limited to 1 cm.

Moisture retention studies can be made with prepared soil samples or undisturbed soil cores. Frequently, soil structure is quite an important determining factor in the value of 1/10 bar and 1/3 bar percentages and this aspect should be considered before electing to use undisturbed soil cores or prepared samples.

A source of regulated gas pressure is required for all extraction work. If the extractor is to be used extensively, compressed air from a compressor is the most satisfactory source of supply. Accuracy of equilibrium values will be no more accurate than the regulation of air supply. For working in the low soil suction range and particularly determination of the 1/10 bar and 1/3 bar percentages, it is essential to have excellent pressure regulation. If a laboratory compressed air supply line is available; the pressure control panel can be conveniently attached to the laboratory wall adjacent to the extractor and connected directly to the supply line.

The moisture retention curves can be developed for different soil types with this type of equipment. These "moisture characteristic" curves for each soil are extremely important in soils research and development of practical, effective irrigation practices.

3.4 Saturated Hydraulic Conductivity

The hydraulic conductivity is not an exclusive property of the soil alone, since it depends upon the attributes of the soil and the fluid together. The soil characteristics, which affect the hydraulic conductivity, are the total porosity, the distribution of pore sizes and the tortuosity – in short, the pore geometry of the soil. The fluid attributes, which affect the hydraulic conductivity, are fluid density and viscosity.

The simplest technique to measure the saturated hydraulic conductivity (K_s) is to take an 'undisturbed' cylindrical sample of the soil, saturate it, and let water flow through it in the laboratory. From the velocity and the hydraulic gradient observed on the sample, K_s can be calculated with Darcy's equation. Because truly undisturbed samples are difficult to obtain and the sample size is relatively small, laboratory methods have limited usefulness and direct measurement of K_s in the field is usually preferred.

3.4.1 Guelph Permeameter

The Guelph Permeameter (Figure 6) is a constant-head device that operates on the Mariotte siphon principle and provides a quick and simple method for simultaneously determining field saturated hydraulic conductivity, matrix flux potential and soil sorptivity in the field.

Theory

Some of the most important factors governing liquid transmission in unsaturated soils are field-saturated hydraulic conductivity K_{fs}, matric flux potential ϕ_m, and sorptivity S. Hydraulic conductivity is a measure of the ability of a soil to conduct water under a unit hydraulic potential gradient. K_{fs} or field-saturated hydraulic conductivity refers to the saturated hydraulic conductivity of soil containing entrapped air. K_{fs} is more appropriate than the truly saturated hydraulic conductivity for vadose (unsaturated) zone investigations because positive pressure heads do not persist in unsaturated conditions long enough for entrapped air to dissolve.

Matric flux potential, ϕ_m is a measure of the soil's ability to pull water by capillary force through a unit cross-sectional area in a unit time. Sorptivity, S is a measure of the ability of a soil to absorb a wetting liquid. In general, the greater the volume of a wetting liquid that can be absorbed, the more rapidly the liquid is absorbed. Since sorptivity is defined in part by matric flux potential, they are essentially two different ways of describing the same phenomenon. The Guelph Permeameter is used to determine K_{fs} and ϕ_m for a particular soil.

Mode of Operation

The Guelph Permeameter is an in-hole constant-head permeameter, employing the Mariotte principle. The method involves measuring the steady state rate of water recharge into unsaturated soil from a cylindrical well hole, in which a constant depth (head) of water is maintained.

Constant head level in the well hole is established and maintained by regulating the level of the bottom of the air tube, which is located in the centre of the permeameter. As the water level in the reservoir falls, a vacuum is created in the air space above the water. The vacuum can only be relieved when air, which enters at the top of the air tube, bubbles out of the air inlet tip and rises to the top of the reservoir. Whenever the water level in the well begins to drop below the air inlet tip, air bubbles emerge from the tip and rise into reservoir air space. The vacuum is then partially relieved and water from the reservoir replenishes water in the well. The size of opening and geometry of the air inlet tip is designed to control the size of air bubbles in order to prevent the well water level from fluctuating.

When the permeameter is operating, an equilibrium is established. The reduced pressure (vacuum) in the air above the water in the reservoir together with the pressure of the water column extending from the surface of well to the surface of water in the reservoir always equals the atmospheric pressure.

When a constant well height of water is established in a cored hole in the soil, a "bulb" of saturated soil with specific dimensions is rather quickly established. This "bulb" is very stable and its shape depends on the type of soil, the radius of the well and the head of water in the well. The shape of the "bulb" is numerically described by the C- factor (Reynolds et al., Groundwater Monitoring Review, 6:1:84-95, 1986) used in the calculations. Once the unique "bulb" shape is established, the outflow of water from the well reaches a steady state flow rate that can be measured. The rate of this constant outflow of water together with the diameter of the well and height of water in the well can be used to accurately determine the field saturated conductivity, matrix flux potential and sorptivity of the soil.

Governing Analytic Equations

The Richards' analysis of steady-state discharge from a cylindrical well in unsaturated soil, as measured by the Guelph Permeameter technique, accounts for all the forces that contribute to three dimensional flow of water into soils viz. the hydraulic push of water into soil, the gravitational pull of liquid out through the bottom of the well, and the capillary pull of water out of the well into the surrounding soil. The Richards' analysis is the basis for the calculations used to determine hydraulic conductivity and matric flux potential.

The following formulae are used to determine hydraulic conductivity, K_{fs} and matric flux potential, ϕ_m when following the standardized procedure.

When using both reservoirs:

$$K_{fs} = (0.0041)(X)(\bar{R}_2) - (0.0054)(X)(\bar{R}_1) \quad \ldots(3.1)$$

$$\phi_m = (0.0572)(X)(\bar{R}_1) - (0.0237)(X)(\bar{R}_2) \quad \ldots(3.2)$$

When using the inner reservoir:

$$K_{fs} = (0.0041)(Y)(\bar{R}_2) - (0.0054)(Y)(\bar{R}_1) \quad \ldots(3.3)$$

$$\phi_m = (0.0572)(Y)(\bar{R}_1) - (0.0237)(Y)(\bar{R}_2) \quad \ldots(3.4)$$

where,

X = Reservoir constant used when the reservoir combination is selected;

Y = Reservoir constant used when only the inner reservoir is selected;

$\bar{R}_1$ = Steady state rate of fall of water in the reservoir at first well height (always 5 cm in the standardized procedure); and

$\bar{R}_2$ = Steady state rate of fall of water in the reservoir at second well height (always 10 cm in the standardized procedure).

Sorptivity

When the volumetric water content of the soil can be measured or estimated with reasonable accuracy, soil sorptivity S can be calculated as follows:

$$S = \sqrt{\{2(\Delta\theta)\phi_m\}} \qquad \ldots (3.5)$$

where,

$\Delta\theta = \theta_{fs} - \theta_i$;

θ_i = initial volumetric water content; and

θ_{fs} = field-saturated volumetric water content.

Alpha Constant and the Conductivity – Pressure Head Relationship

Alpha is a constant that is dependent on the porous properties of soil. It is calculated as follows:

$$\alpha = K_{fs} / \phi_m \qquad \ldots (3.6)$$

The hydraulic conductivity and pressure head relationship, K(φ) describes the change in K with soil suction. Generally, as soil suction increases, hydraulic conductivity decreases exponentially. For any soil suction (as measured in cm of water), the hydraulic conductivity can be predicted by the following equation.

$$K = K_{fs} [e^{(\alpha)(\varphi)}] \qquad \ldots (3.7)$$

where,

φ = soil water suction (in cm of water); and

e = 2.71828 (base of natural logarithm).

The results of measurements with the Guelph Permeameter can indicate soil heterogeneity. When a negative K_{fs} or ϕ_m value is calculated, it is indicative of the presence of a hydrologic discontinuity, typically caused by soil stratification or the presence of rodent and/or root holes. This underlines the value of a profile description. When a negative value for K_{fs} or ϕ_m is obtained, it indicates that further measurements are needed to account for the degree and kind of soil heterogeneity.

Soils typically have three-dimensional heterogeneity. The Guelph Permeameter method yields essentially a “point” measurement. The size of land under investigation, degree of soil heterogeneity, soil type and kind of application will dictate the number of measurements needed to adequately characterize a given area and depth of soil. A soil profile description and soil survey report will greatly enhance the value and understanding of data obtained with the Guelph Permeameter. Because of the ease and simplicity of Guelph Permeameter and its depth profiling capability, it is a very useful method for understanding the three dimensional distribution of the water transmission properties of soils.

3.5 van Genuchten Parameters

The measurements of $\theta(h)$ from soil cores (obtained through pressure plate apparatus) can be fitted to the desired soil water retention model. Once the retention function is estimated, the hydraulic conductivity relation, K(h), can be evaluated if the saturated hydraulic conductivity, K_s, is known. In the present study, parameters of van Genuchten model were derived for soil moisture retention and hydraulic conductivity functions. For the van Genuchten model (1980), the water retention function is given by

$$S_e = (\theta - \theta_r)/(\theta_s - \theta_r) = [1 + (\alpha_v |h|)^n]^{-m} \qquad \text{for } h < 0$$
$$= 1 \qquad \text{for } h \geq 0 \qquad \ldots(3.8)$$

and the hydraulic conductivity function is described by

$$K = K_s S_e^{1/2} [1 - (1 - S_e^{1/m})^m]^2 \qquad \ldots(3.9)$$

where, α_v and n are van Genuchten model parameters, m = 1 – 1/n.

4.0 DESCRIPTION OF SWIM MODEL

4.1 Introduction

SWIM is an acronym that stands for Soil Water Infiltration and Movement. It is a software package developed within the CSIRO Division of Soils for simulating infiltration, evapotranspiration, and redistribution. The first version (SWIMv1) was published in 1990 (Ross, 1990b). Version 2 of the model (identified as SWIMv2.0), which combines water movement with transient solute transport and which accommodates a variety of soil property descriptions and more flexible boundary conditions, was completed in 1992. The latest version, SWIMv2.1, has been described here.

SWIMv2 is based on a numerical solution of the Richards' equation and the advection-dispersion equation. It can be used to simulate runoff, infiltration, redistribution, solute transport and redistribution of solutes, plant uptake and transpiration, soil evaporation, deep drainage and leaching. The physical system and the associated flows addressed by the model are shown schematically in Figure 7. Soil water and solute transport properties, initial conditions, and time dependent boundary conditions (e.g., precipitation, evaporative demand, solute input) need to be supplied by the user in order to run the model.

The model deals with a one-dimensional soil profile. For a vertical soil profile, this means that it may be vertically inhomogeneous, but must be horizontally uniform. This assumption has two consequences of importance in many common simulations. There is only one hydraulic conductivity function for each soil layer, so that any macropore, or bypass, flow can only be accounted for in a limited way. Secondly, the calculated solute concentrations apply to the whole soil layer, which means that there is no concentration gradient from the bulk soil to near the root surface. The presence of such a concentration gradient may in reality affect the soil osmotic potential and hence water and solute uptake. The overall purpose of the model is to address issues relating to the soil water and solute balance. As such, it is a research tool that can be integrated in laboratory and field studies concerned with soil water and solute transport.

4.2 Water Movement

4.2.1 Richards' Equation

One-dimensional flow of water through isothermal, rigid, unsaturated or saturated soil is governed by Darcy's law (Darcy, 1856; Buckingham, 1907)

$$q = -\mathrm{K}\frac{d\mathrm{H}}{dx} \qquad \ldots(4.1)$$

where,

q	=	water flux density
	=	volumetric water flow per unit cross-sectional area per unit time (cm^3 water/cm^2 soil/h)
K	=	hydraulic conductivity (cm^2 water/cm soil/h)
H	=	hydraulic head (cm water)
x	=	distance into the soil (cm soil)

Darcy's law states that water flows down a hydraulic gradient at a rate proportional to the gradient. The "constant" of proportionality, K, varies with conditions such as soil type and water content, but not with the gradient. Darcy's law has proven to be valid under most conditions of soil water flow provided the soil can be treated as a continuum, i.e., provided a suitable length scale for definition of variables such as q, K, and H can be established (Bear, 1979).

In flow situations where q, K, and H vary in time and space, so-called transient water flow, it is necessary to combine Darcy's equation with the continuity equation that conserves mass of water. For a fluid of constant density, this is expressed as conservation of volume.

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q}{\partial x} + S \qquad \ldots(4.2)$$

where,

θ	=	volumetric water content (cm^3/cm^3)
t	=	time (h)
S	=	source (or sink, if negative) strength (cm^3 water/cm^3 soil/h)

Combining equations (4.1) and (4.2) gives the Richards' equation (Richards, 1931).

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial x}\left[\mathrm{K}\frac{\partial \mathrm{H}}{\partial x}\right]+S \qquad \text{…(4.3)}$$

In rigid, unsaturated or saturated soil in which the gas pressure is always atmospheric (i.e. air can move freely) the hydraulic head, H, is the sum of the gravitational potential, z, and the matric potential, Ψ (which for convenience is extended to include positive values under saturated conditions). The gravitational potential, z, is equal to the elevation from some arbitrary reference level. The Richards' equation then becomes:

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial x}\mathrm{K}\left[\frac{\partial\psi}{\partial x}+\frac{dz}{dx}\right]+S \qquad \text{…(4.4)}$$

where, θ and Ψ are related by the water retention curve and K is related to θ by the hydraulic conductivity function. This so-called mixed θ and Ψ form of the Richards' equation can be conveniently solved numerically using the Newton-Raphson iterative method (Campbell, 1985). The numerical solution accurately conserves water during numerical solution, no matter how large the time step (Hornung and Messing, 1981; Milly, 1984; Celia et al., 1990; Ross, 1990a; Ross and Bristow, 1990).

Equation (4.4) is highly non-linear, especially in dry soils, where K and Ψ change over several orders of magnitude with changes in θ. SWIMv2.1, therefore, solves the Richards' equation by using a hyperbolic sine transform of Ψ (Ross, 1990a). For this purpose, equation (4.4) is written as

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial x}\mathrm{K}\left[\frac{d\psi}{dp}\frac{\partial p}{\partial x}+\frac{dz}{dx}\right]+S \qquad \text{…(4.5)}$$

with

$$-\frac{\psi-\psi_o}{\psi_1} = \sinh p \qquad \psi < \psi_o$$

$$-\frac{\psi-\psi_o}{\psi_1} = p \qquad \psi \geq \psi_o \qquad \ldots(4.6)$$

where, ψ_o (**psi0**) and ψ_1 (**psi1**) are shifting and scaling parameters, respectively. Appropriate choice of ψ_o allows the inverse hyperbolic sine transform to be applied over the dry range $\psi < \psi_o$ while using a linear transform for the wet range $\psi \geq \psi_o$. Ross (1990a) obtained good results with ψ_o = -50 cm and ψ_1 = -5 cm. Both $d\psi/dp$ and $d^2\psi/dp^2$ are continuous functions of p over the entire range of ψ (Ross, 1990a), which is desirable for the Newton-Raphson method. Use of this transform allows SWIMv2.1 to deal with unsaturated and saturated flow and dry soils with relatively large space steps in the numerical solution.

Equation (4.5) does not specify the direction of flow relative to the direction of gravity. Directions of flow other than vertical flow are, therefore, possible. In the input file to SWIMv2.1, (-dz/dx) is set equal to the gravity factor **gf,** which is equal to the cosine of the angle between the x-direction and gravity. A gravity factor of 1 gives vertical downward flow, a gravity factor of 0 gives horizontal flow (gravity ignored). Note, however, that flow is still strictly one-dimensional and that lateral flow is not taken into account.

4.2.2 Hydraulic Properties

In unsaturated systems, θ and Ψ are related by the water retention curve, and K is related to θ by the hydraulic conductivity function. Both these functions are strongly non-linear. For saturated conditions, these functions reduce to constants:

$$\theta = \theta_s$$

$$K = K_s$$

where,

θ = volumetric water content (cm^3/cm^3)

θ_s = saturated volumetric water content (cm^3/cm^3) (**ths**)

K = hydraulic conductivity (cm/h)

K_s = saturated hydraulic conductivity (cm/h) (**hks**)

Instead of relating θ and ψ directly, SWIM v2.1 uses a normalised parameter S, the effective saturation.

$$S = \frac{\theta - \theta_r}{\theta_s - \theta_r}$$

where,

S = effective saturation (cm^3/ cm^3)

θ_r = residual volumetric water content (cm^3/ cm^3) (**thr**)

The effective saturation, S, of a porous medium can be expressed as the cumulative distribution function of a capillary pore-size distribution (given as a function of the matric potential). Sometimes, several partly overlapping pore-size distributions can be distinguished. The water retention curve in SWIMv2.1 is described by considering it as the sum of overlapping pore-size distributions. The result is an overall water retention curve which can be expressed as a sum of simple functions. The parameters for these simple functions are input to the program HYPROPS, which processes them to an overall water retention curve. The output generated by HYPROPS is used to prepare the input file for SWIMv2.1. HYPROPS offers a choice of models for these simple functions (**terms**). The van Genuchten functions (vg), from van Genuchten (1980) are written as:

$$S(\Psi) = \left[1 + \left(\alpha|\Psi|\right)^n\right]^{-m} \qquad \alpha > 0,\ n > 1$$

$$K(\Psi) = K_s S^p \left[1 - \left(1 - S^{1/m}\right)^m\right]^2 \qquad m = 1 - 1/n,\ n > 0$$

where α, m and n are constants and p is the pore interaction index.

4.2.3 Initial and Boundary Conditions

Solution of equation (4.5) requires the specification of initial conditions, boundary conditions and the source/sink term S. The initial condition can be given as matric potential or water content at specified depths. If initial water contents are specified, then these are converted by SWIMv2.1 to matric potentials using the specified water retention curve. Boundary conditions for equation (4.5) can in general be defined in terms of matric potentials, potential gradients or fluxes, all defined as functions of time. The source/sink term S in equation (4.5) can be root water uptake, soil evaporation (surface node only), or bypassing of water to a certain node.

Top Boundary

One of the options in SWIMv2.1 is to define a constant potential top boundary condition (**itbc=1**), where the matric potential at the surface node is kept constant at the value given in the initial profile. Water needed to keep this value constant is “created” if necessary. The boundary acts as a source of water and the water is artificially “extracted” from or added to an imaginary runoff. This may lead to negative values being reported for runoff.

Flux type boundary conditions are set by appropriate choices of rainfall/irrigation and potential evaporation intensities. These conditions may, however, be modified if the soil hydraulics are limiting (even in the case of an infinite surface conductance (**itbc=0**)), or if there is a surface seal of limited conductance (conductance function, **itbc=2**). In both cases, not all water may infiltrate immediately. SWIMv2.1 gives three options to handle this deficit in infiltration.

- no ponding, all water runs off (**isbc=0**)
- ponding, no runoff (**isbc=1**)
- simple power-law runoff function (**isbc=2**)

If a fourth option (**isbc=3**) is chosen, surface runoff is treated as for isbc=2, but SWIMv2.1 will also calculate potential subsurface downslope flow.

Bottom boundary

SWIMv2.1 gives four options for the bottom boundary condition:

- given matric potential gradient of variable magnitude in time (**ibbc=0**)
- given potential, variable in time (**ibbc=1**)
- zero flux (**ibbc=2**)
- seepage, with threshold suction variable in time (**ibbc=3**)

If **ibbc=0**, then dψ/dx needs to be specified as a function of time (**ntb, tb(1,i), tb(2,i)**). When unit gradient of hydraulic head is chosen, dψ/dx=0 (**tb(2,i)=0**). If **ibbc=1**, then ψ needs to be specified as a function of time (**ntb, tb(1,i), tb(2,i)**). This option is useful for simulating measured values or a (fluctuating) water table (e.g. time record of positive values for ψ). Note, however, that this condition can cause the bottom boundary to act as a source of water (similar to the top boundary condition **itbc=1**). If **ibbc=2**, then q=0 (equation 4.1). This can be used to simulate an impermeable layer. If **ibbc=3**, then upward flow through the bottom boundary is not possible and drainage will only occur when ψ exceeds the specified limit (**ntb, tb(1,i), tb(2,i)**). Contrary to the situation with **ibbc=1**, the drained water is in this case lost to the system. Laboratory columns in which the base is held at a certain suction (e.g. by use of a suction plate, wick, or simply open to the air) are represented by this boundary condition.

4.3 Solute Transport

4.3.1 Advection-Dispersion Equation

Solute transport is governed by two processes: diffusion and advection with water. Differences in pore water velocities (both within an individual pore and between pores of different sizes) lead to an additional effect known as hydrodynamic dispersion. This process results in spreading of the solute, very much like diffusion does, so the two are often combined in the mathematical description of solute transport. In SWIMv2.1, solute movement is based on the following solute transport equation

$$\frac{\partial(\theta c)}{\partial t} + \frac{\partial(\rho s)}{\partial t} = \frac{\partial}{\partial x}\left[\theta D \frac{\partial c}{\partial x}\right] - \frac{\partial(qc)}{\partial x} + \phi \qquad \ldots(4.7)$$

where,

c	=	solute concentration in solution (μmol or μg solutes/cm^3 water)
s	=	adsorbed concentration (μmol/g soil or μg/g soil)
ρ	=	soil bulk density (g/cm^3) (**rhob**)
t	=	time (h)
x	=	depth (cm)

θ = water content (cm^3/cm^3)

q = water flux density (cm/h)

D = combined dispersion and diffusion coefficient (cm^2/h)

ϕ = source/sink term ($\mu mol/cm^3/h$ or $\mu g/cm^3/h$)

and

$$D = \tau D_0 + \varepsilon |v|^n$$
$$\tau = a(\theta - b)^{\xi} \qquad \text{…(4.8 a, b, c)}$$
$$s = kc^{\eta}$$

where,

D_0 = ionic or molecular diffusion coefficient in free water (cm^2/h) (**d0**)

τ = tortuosity factor [-]

ε = dispersivity of the medium [$(cm^2/h)/(cm/h)^n$] (**dis (i)**)

v = pore water velocity = q/θ (cm/h)

a = empirical constant [-] (**a**)

b = empirical constant [-] (**dthc**)

ξ = empirical constant [-] (**dthp**)

n = empirical constant [-] (**disp**)

k = coefficient of Freundlich isotherm
[(mol or g adsorbed solute/g soil)/(mol or g solute/cm^3 water)$^\eta$] (**exco**)

η = power of Freundlich isotherm [-] (**fip (i)**)

Currently, SWIMv2.1 can only account for one solute at a time. The choice of units for solute concentration, c, is flexible. Any units can be used as long as they are expressed in an amount/cm^3 soil. Units of s, ϕ, and k change accordingly, as well as the units of **slos.**

Values for the diffusion coefficient in water (**d0**) (equation 4.8a) can be found in several literature sources (e.g., Robinson and Stokes, 1965; Lehrman, 1979; Weast and Astle, 1980; Kemper, 1986; Sadeghi et al., 1988). The diffusion coefficient depends on the temperature, the concentration of the solute, and on the ions that the solute consists of. For example, if chloride moves as $CaCl_2$, it has a diffusion coefficient of approximately 0.045 cm^2/h, whereas if it moves as KCl, the appropriate value is about 0.071 cm^2/h (25°C, approx.

1.0 μmol/cm^3 (=0.001 M)). If there is a mixed electrolyte, the diffusion coefficient is determined by a contribution for each. Nye (1966) has outlined an approach to calculate the appropriate diffusion coefficient in that case (see e.g., Bond and Phillips, 1990a). If the cation and anion fronts move separately (e.g., Bond and Smiles, 1988; Bond and Phillips, 1990b), then the appropriate diffusion coefficient for the anion may be that in combination with the resident cation, rather than with the incoming cation. Note, however, that SWIMv2.1 currently does not handle separate fronts.

The equation for tortuosity (equation 4.8b) is general and allows flexible parameterisation. It is, for example, able to handle the commonly used relationship of Millington and Quirk (1961).

$$\tau = \theta^{7/3} / \theta_s^{\,2} \qquad \ldots(4.9)$$

For saturated systems, τ is often taken to be a constant equal to 0.67 (Rose, 1977), although Bond (1986) obtained a value of 0.442 in a saturated breakthrough experiment. While there is evidence that suggests that τ is a function of water content in unsaturated soils (e.g., Porter et al., 1960; Barraclough and Tinker, 1981), it has been assumed constant and equal to 1.0 in a number of successful descriptions of solute transport in unsteady, unsaturated flow experiments (Smiles et al., 1981; Bond et al., 1982; Bond et al., 1997). In practice, the difference in spreading resulting from values of τ between 1 and 0.5 is often small. Another frequently used equation for tortuosity was first proposed by Kemper and van Schaik (1966). Adapted to fit the definition of D and τ used by SWIMv2.1, it has the following functional form

$$\tau = \frac{a}{\theta}\exp(b\theta) \qquad \ldots(4.10)$$

While this equation cannot be converted to equation (4.8b), it can give an idea of the magnitude of τ. Comparing this equation with data collected on soils by Olsen et al. (1965) and Porter et al. (1960), Olsen and Kemper (1968) found that b was approximately 10 and $0.005 < a/\theta < 0.01$. They also point out, however, that this equation was applicable only in the range of moisture contents between 330 and 15000 cm suction. Indeed, at high water contents, values of τ above 1 may be obtained, and this is not consistent with the definition of τ.

Values for the dispersivity, ε, vary widely in the literature (see e.g., extensive review by Beven et al., 1993). Values obtained in field experiments are commonly an order of magnitude higher than those obtained in laboratory field columns (approx. 1 cm) (Rose et al., 1982). This is often a result of the inclusion of other effects in the dispersion term, such as heterogeneities in pore water velocities, preferential flow, and "immobile" water effects, because of using a steady-state analysis to fit the dispersion term to the field data. Spatial averaging of data and neglect of the contribution of diffusion can play a role as well. When the average flow velocity is low, the choice of the dispersivity is less critical because the second term on the right hand side of equation (4.8a) becomes small. The value of the exponent, n, is usually taken to be 1, although higher values have been found and 1.2 is sometimes used. Equation (4.8c) is the Freundlich isotherm for adsorption. If $\eta = 1$, then this reduces to a linear isotherm. While equations (4.8 a, b, c) have some physical basis; they are essentially empirical equations, so that the units in equations (4.8a) and (4.8c) vary with the powers n and η.

4.3.2 Solute Initial and Boundary Conditions

Solute can be added to the system in a variety of ways:

- <u>As part of initialisation</u>: Solute concentrations (in solution) are specified for each node (**csl(i)**). If there is adsorption ($k \neq 0$), then the initial amount of adsorbed solute is "created" using the specified adsorption isotherm.
- <u>In rainfall or irrigation</u>: Cumulative solute additions (in amounts rather than concentrations) are given in the input file (**nts** time-addition pairs). These solute additions are assumed to be mixed with the rainfall/irrigation. The units of the amounts need to be consistent with the unit of solute concentration used. For example if, over a certain time period, 0.32 cm of irrigation water is applied to the soil surface (0.32 cm^3 water/cm^2 surface area) with a solute concentration of 50 μmol solute/cm^3 water, then the amount of solute added (to be specified in input file) is 16 μmol solute/cm^2 surface area.

- As part of cultivation: In this case, the solute is added “dry” to the surface. It will enter the soil with infiltrating water at a concentration of **slsci** or disappear with surface runoff water at a concentration of **slscr** (**slsci** and **slscr** are specified in the input file).
- By production in the profile
- By artificial “creation” of solute when there is a constant potential top and/or bottom boundary condition for water flow (**itbc=1** or **ibbc=1**): The solute concentration at these boundaries is kept constant and in order to achieve this, solute may be “created”. At the top boundary, the solute concentration is held at the specified initial value (**csl(0)**); while for the bottom boundary, it is specified separately in the input file as **csl(n).**

4.4 Limitations of the Model

(i) Only one-dimensional flow is considered. Lateral equilibrium is, therefore, assumed. Net lateral surface runoff is treated as a sink term at the surface.

(ii) Macropores and bypass flow are taken into account only in a limited way.

(iii) The soil matrix is assumed rigid, so that SWIMv2.1 is not strictly applicable to swelling soils.

(iv) Soil airflow is ignored.

(v) Vapour flow within the soil can be included as part of the conductivity term, but only in response to matric potential gradients.

(vi) Temperature effects on water movement are ignored.

(vii) Osmotic effects are ignored, except in water uptake and soil evaporation.

(viii) Wetting front instability or fingering (Glass et al., 1989; Hendrickx et al., 1993) is not taken into account.

5.0 ANALYSIS AND RESULTS

5.1 General

Knowledge of the physics of soil water movement is crucial to the solution of problems in watershed hydrology; for example, the prediction of runoff and infiltration following precipitation, the subsequent distribution of infiltrated water by drainage and evaporation, and the estimation of the contribution of various parts of a watershed to the ground water store. The present study deals with modelling of soil moisture movement in Barchi watershed (Karnataka) using the SWIM model. Daily rainfall and evaporation data of Barchi for the period 1996-97 to 1999-2000 were collected for the study. Water balance components like runoff, evapotranspiration and drainage (ground water recharge from rainfall) were computed through SWIM

5.2 Soil Moisture Characteristics

To model the retention and movement of water and chemicals in the unsaturated zone, it is necessary to know the relationships between soil water pressure, water content and hydraulic conductivity. It is often convenient to represent these functions by means of relatively simple parametric expressions. The problem of characterizing the soil hydraulic properties then reduces to estimating parameters of the appropriate constitutive model.

Based upon the available information, two distinct soil layers were identified (0-45 cm and 45-150 cm). Saturated hydraulic conductivity was measured at 8 locations in the study area by using Guelph Permeameter (locations are shown in Figure 1). The average saturated hydraulic conductivity values for the upper layer (0-45 cm) and lower layer (45-150 cm) were found to be 0.339 cm/hour and 0.648 cm/hour respectively.

Soil moisture retention characteristics were determined in the laboratory using the Pressure Plate Apparatus. Tables 1 and 2 present the values of saturated hydraulic conductivity (K_s), saturated moisture content (θ_s) and soil moisture retention data at 8 locations for upper (0-45 cm) and lower (45-150 cm) soil layers respectively. Soil moisture retention curves for the 8 field locations have been presented graphically in Figures 8 to 15 (upper soil layer) and Figures 16 to 23 (lower soil layer) respectively.

Table 1 : Soil Moisture Retention Data for Upper Soil Layer

Station	K_s (cm/ hour)	θ_s	Pressure (bar)							
			0.33	**1**	**3**	**5**	**7**	**10**	**12**	**15**
1	0.58	0.37	0.24	0.22	0.19	0.16	0.13	0.11	0.10	0.08
2	0.57	0.37	0.30	0.28	0.27	0.23	0.20	0.18	0.14	0.14
3	0.60	0.38	0.33	0.25	0.25	0.22	0.20	0.16	0.12	0.09
4	0.18	0.53	0.43	0.39	0.36	0.35	0.33	0.32	0.31	0.30
5	0.20	0.53	0.41	0.38	0.36	0.35	0.34	0.32	0.30	0.28
6	0.18	0.53	0.40	0.38	0.37	0.37	0.35	0.33	0.30	0.28
7	0.24	0.52	0.44	0.43	0.37	0.34	0.32	0.29	0.26	0.25
8	0.16	0.54	0.51	0.44	0.39	0.38	0.38	0.36	0.33	0.30

Table 2 : Soil Moisture Retention Data for Lower Soil Layer

Station	K_s (cm/ hour)	θ_s	Pressure (bar)						
			0.1	**0.33**	**1**	**3**	**5**	**10**	**15**
1	1.66	0.38	0.29	0.22	0.17	0.16	0.12	0.12	0.11
2	0.60	0.32	0.30	0.23	0.16	0.13	0.11	0.10	0.09
3	0.007	0.43	0.33	0.25	0.22	0.20	0.15	0.09	0.06
4	0.58	0.41	0.34	0.30	0.28	0.26	0.23	0.18	0.14
5	0.58	0.43	0.40	0.31	0.28	0.26	0.23	0.18	0.16
6	0.18	0.53	-	0.40	0.38	0.37	0.37	0.33	0.28
7	0.59	0.31	0.27	0.19	0.17	0.16	0.15	0.15	0.13
8	0.60	0.45	0.39	0.26	0.25	0.24	0.23	0.21	0.20

The experimental soil moisture retention data were fitted to the van Genuchten model (1980). Residual moisture content (θ_r) was assumed to be equivalent to moisture retained corresponding to 15 bar pressure. The parameters of soil moisture retention function and hydraulic conductivity function were obtained through non-linear regression analysis. Tables 3 and 4 present the van Genuchten parameters α and n (equations 3.8 and 3.9) for upper and lower soil layers in Barchi watershed. Average values of these parameters were also determined through non-linear regression analysis and used in modelling of soil moisture movement through SWIM.

Table 3 : van Genuchten Parameters for Upper Soil Layer

Station	K_s (cm/hour)	θ_r	θ_s	van Genuchten Parameters		Proportion of Variance Explained (%)
				α	n	
1	0.58	0.08	0.37	0.0073	1.434	80.78
2	0.57	0.14	0.37	0.0023	1.509	74.08
3	0.60	0.09	0.38	0.0021	1.465	79.07
4	0.18	0.30	0.53	0.0067	1.523	92.00
5	0.20	0.28	0.53	0.0129	1.373	80.66
6	0.18	0.28	0.53	0.0235	1.300	64.09
7	0.24	0.25	0.52	0.0020	1.580	84.07
8	0.16	0.30	0.54	0.0019	1.552	91.51
Average	0.339	0.215	0.471	0.0047	1.4385	24.43

Table 4 : van Genuchten Parameters for Lower Soil Layer

Station	K_s (cm/hour)	θ_r	θ_s	van Genuchten Parameters		Proportion of Variance Explained (%)
				α	n	
1	1.66	0.11	0.38	0.0148	1.563	97.04
2	0.60	0.09	0.32	0.0045	1.760	99.52
3	0.007	0.06	0.43	0.0154	1.358	87.12
4	0.58	0.14	0.41	0.0134	1.310	81.71
5	0.58	0.16	0.43	0.0070	1.444	91.68
6	0.18	0.28	0.53	0.0235	1.300	64.09
7	0.59	0.13	0.31	0.0120	1.596	95.35
8	0.60	0.20	0.45	0.0123	1.688	91.97
Average	0.648	0.121	0.394	0.0095	1.4212	58.31

Before running the SWIM model, it is necessary to convert measured or estimated hydraulic properties into a form that the SWIM program accepts. For this purpose, the program HYPROPS is provided. This program needs to be run first, as its output is used to prepare the SWIM input file. HYPROPS is not a hydraulic property fitting program. The user needs to provide the parameter values of the functions that are chosen to describe the water retention and conductivity data. HYPROPS performs any necessary summations and generates a hydraulic property table based on a piecewise cubic approximation. The average hydraulic property input sets for upper and lower soil layers, as obtained through HYPROPS, are given in Tables 5 and 6 respectively.

Table 5 : Hydraulic Property Input for Upper Soil Layer

S.No.	$Log_{10}\|\psi\|$	Water Content θ	Slope of θ vs. $Log_{10}\|\psi\|$	Log_{10} K	Slope of Log_{10} K vs. $Log_{10}\|\psi\|$
1.	-3.000000	0.471000	-0.000000	-0.473838	-0.004081
2.	-2.000000	0.471000	-0.000000	-0.480914	-0.011281
3.	-1.000000	0.470999	-0.000004	-0.500617	-0.031639
4.	-0.000000	0.470965	-0.000116	-0.557058	-0.092598
5.	1.000000	0.470047	-0.003129	-0.733332	-0.308302
6.	1.951918	0.452009	-0.053442	-1.358404	-1.224746
7.	2.600241	0.390345	-0.125789	-2.563588	-2.467220
8.	3.350125	0.305385	-0.088169	-4.684655	-3.023609
9.	4.969595	0.232830	-0.017980	-9.671609	-3.091979
10.	7.000000	0.217301	-0.002320	-15.950152	-3.092309

Table 6 : Hydraulic Property Input for Lower Soil Layer

S.No.	$Log_{10}\|\psi\|$	Water Content θ	Slope of θ vs. $Log_{10}\|\psi\|$	Log_{10} K	Slope of Log_{10} K vs. $Log_{10}\|\psi\|$
1.	-3.000000	0.394000	-0.000000	-0.195127	-0.006522
2.	-2.000000	0.394000	-0.000001	-0.206206	-0.017414
3.	-1.000000	0.393996	-0.000013	-0.236111	-0.047524
4.	-0.000000	0.393892	-0.000353	-0.320295	-0.138048
5.	1.050000	0.390728	-0.010427	-0.613347	-0.514558
6.	1.989175	0.346817	-0.103530	-1.632535	-1.887056
7.	2.814364	0.244994	-0.111824	-3.709778	-2.909980
8.	4.035274	0.159772	-0.037533	-7.394128	-3.051489
9.	5.932140	0.127168	-0.005979	-13.186908	-3.054320
10.	7.000000	0.123190	-0.002123	-16.448499	-3.054326

Note :

ψ = Suction
θ = Water Content
K = Hydraulic Conductivity

5.3 Model Conceptualization

Modelling of soil moisture movement in Barchi watershed has been done using SWIM. The model was simulated for 1461 days (1 May 1996 to 30 April 2000). One vegetation type (teak, covered in most parts of the watershed) was considered for the study. Exponential root growth with depth and linear interpolation with time was assumed. The following vegetation parameters were adopted for the simulations:

Root radius (rad)	=	0.5 cm
Root conductance (groot)	=	$4.0 * 10^{-7}$
Minimum xylem potential (psimin)	=	-15,000 cm
Root depth constant (xc)	=	150 cm
Maximum root length density (rldmax)	=	4 cm/cm^3

The profile is 150 cm deep with surface at 0 cm and bottom boundary condition applying at 150 cm. Vapour conductivity is not taken into account, nor is the effect of osmotic potential. There are two hydraulic property sets (for upper and lower soil layers) that are applied to 31 depth nodes of the 150 cm deep profile. Hysteresis is not taken into account.

Initially, there is no water ponded on the surface. Runoff is governed by a simple power law function and a surface conductance function. No bypass flow was included. A matric potential gradient of 0, i.e. “unit gradient”, has been applied as bottom boundary condition throughout the simulation. Cumulative rainfall and evaporation records (daily) for the period 1996-97 to 1999-2000 were given in the input file for determination of water balance components (runoff, evapotranspiration and drainage). The SWIM input file for Barchi watershed is given at Annexure.

5.4 Simulation of Water Balance Components

Soil moisture movement in Barchi watershed was simulated for the period 1996-97 to 1999-2000. The water year (May to April) was considered for the study. The model computes runoff, infiltration, evapotranspiration and drainage for each time period. A reasonably good agreement was found between observed and simulated soil moisture profiles (Purandara and Kumar, 2000).

With the available input data and parameters, the model was found to underestimate the runoff values. Hourly rainfall values were not available for the watershed. Therefore, daily rainfall values were equally distributed to 4 hours for the periods exceeding 20 mm rainfall in a day. This made a better agreement between the observed and simulated runoff. However, the observed runoff values were suspected to be erroneous in view of inaccurate positioning of zero of gauge. The resulting water balance components for the simulation period have been presented in Table 7.

Table 7 : Water Balance Components for the Barchi Watershed

Year	Rainfall (mm)	Infiltration (mm)	Drainage (mm)	ET (mm)	Runoff (mm)	Runoff Coefficient (%)	Recharge Coefficient (%)
1996-1997	1345.85	1083.37	514.46	519.52	262.48	19.50	38.22
1997-1998	1765.25	1195.05	698.63	500.43	570.20	32.30	39.58
1998-1999	1241.30	1087.46	579.55	507.92	153.84	12.39	46.69
1999-2000	1886.80	1278.18	784.90	493.28	608.62	32.26	41.60
Total	6239.20	4644.06	2577.54	2021.15	1595.14	24.11	41.52

The yearly rainfall varied between 1241 mm to 1887 mm during the period under study. It can be observed from Table 7 that the drainage (recharge from rainfall) varies from 38% to 47% with the average value being 42%. The runoff coefficient was found to vary between 12% (low rainfall year) to 32% (high rainfall year) with the average value being 24%. Simulation of variable infiltration suggests that it has relatively little effect on evapotranspiration, but considerable effect on point drainage.

5.5 Concluding Remarks

A numerical model like SWIM can be used to predict water balance components for the unsaturated zone. However, the estimation of runoff is a complicated task and even simple models require several variables. Few of the runoff parameters in SWIM can only be estimated from published data. Also, the lower boundary condition in SWIM is faced with certain problems. Generally, one of three conditions may be chosen: gravity drainage, prescribed flux or prescribed potential (e.g. a water table where the potential is zero). The prescribed flux or potential conditions are easy to mimic mathematically, hence their solution is generally quite accurate. When gravity drainage is specified, there is some difficulty in mathematical formulation. Generally, this is modelled as a unit gradient and the flow is assumed to be equal to the hydraulic conductivity at the estimated potential. However, this implies that flow will always occur and under conditions of no or low rainfall, the modelled results yield unrealistically low potentials (hence high suctions). This is because the water table is in effect assumed to be at an infinite depth. As the model attempts to bring a profile to equilibrium, the potential continually drops indefinitely. It should be noted that when rainfall events are reasonably frequent, the problems associated with gravity drainage are often negligible.

6.0 CONCLUSION

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 main purpose of using dynamic simulation models is to assess the effects of water management measures such as irrigation, drainage, soil improvement and regional water supply plans, on the terms of the water balance of agricultural as well as nature conservation areas. Through the water balance terms one is generally able to evaluate effects of water management on e.g. crop yield and agricultural income. Transport of solutes is another aspect, which is directly related to the simulation of unsaturated water flow, i.e. the evaluation of pollution of the ground water reservoir, salinization, etc.

The output of a simulation model can include such variables as pressure head, moisture content and flux as a function of soil depth and time. However, most frequently one calculates the terms of the water balance, i.e. infiltration, actual evaporation, actual transpiration, change in soil water storage and the net flux through the region boundary.

Application of SWIM model is one of the simplest techniques, which is well suited for unsaturated zone. SWIM is a software package for simulating water infiltration and movement in soils. Water is added as precipitation and removed by runoff, drainage, evaporation from the soil surface and transpiration by vegetation. The simulator assumes that conditions can be treated as horizontally uniform, flow is described by the Richards equation and soil hydraulic properties can be described by simple functions. While this is adequate for many purposes, there are situations where it is not and the simulation results should never be applied uncritically.

In the present study, soil moisture movement in Barchi watershed (Karnataka) has been modelled using SWIM. Laboratory and field investigations were carried out to determine the soil moisture characterstics (hydraulic conductivity and soil moisture retention function). Water balance components like runoff, evapotranspiration and drainage were determined through SWIM for the period 1996-97 to 1999-2000. The ground water recharge was found to vary between 38% to 47% of rainfall while the runoff coefficient varied between 12% (low rainfall year) to 32% (high rainfall year) for the study period. Variable infiltration was observed to have relatively little effect on evapotranspiration, but considerable effect on drainage.

The SWIM model demonstrated the possibility of predicting water balance components of the unsaturated zone, but only with careful selection of input parameters. It would appear that when actual observed data is not available, it would be difficult to rely upon numerical models alone.

REFERENCES

1. Barraclough, P. B. and P. B. Tinker (1981), “The Determination of Ionic Diffusion Coefficients in Field Soils. I. Diffusion Coefficients in Sieved Soils in Relation to Water Content and Bulk Density”, Journal of Soil Science, Volume 32, pp. 225-236.

2. Bear, J. (1979), “Hydraulics of Groundwater”, McGraw-Hill, New York, U.S.A.

3. Beven, K. J., D. E. Henderson and A. D. Reeves (1993), “Dispersion Parameters for Undisturbed Partially Saturated Soil”, Journal of Hydrology, Volume 143, pp. 19-43.

4. Bond, W. J. (1986), “Velocity-dependent Hydrodynamic Dispersion during Unsteady, Unsaturated Soil Water Flow: Experiments”, Water Resources Research, Volume 22, pp. 1881-1889.

5. Bond, W. J., B. N. Gardiner and D. E. Smiles (1982), “Constant-flux Absorption of a Tritiated Calcium Chloride Solution by a Clay Soil with Anion Exclusion”, Soil Science Society of America Journal, Volume 46, pp. 1133-1137.

6. Bond, W. J. and D. E. Smiles (1988), “Predicting the Average Movement of Reactive Solutes in Soils”, Soil Use Management, Volume 4, pp. 115-120.

7. Bond, W. J. and I. R. Phillips (1990a), “Approximate Solutions for Cation Transport during Unsteady, Unsaturated Soil Water Flow”, Water Resources Research, Volume 26, pp. 2195-2205.

8. Bond, W. J. and I. R. Phillips (1990b), “Ion Transport during Unsteady Water Flow in an Unsaturated Clay Soil”, Soil Science Society of America Journal, Volume 54, pp. 636-645.

9. Bond, W. J., C. J. Smith and P. J. Ross (1997), “Field Validation of a Water and Solute Transport Model for the Unsaturated Zone”, In. Shallow Groundwater Systems: 2. Flow and Solute Transport Models (Ed. I. Simmers et al.), Heise verlag, Hannover, Germany.

10. Buckingham, E. (1907), “Studies on the Movement of Soil Moisture”, Bulletin 38, US Department of Agriculture Bureau of Soils, Washington DC, U.S.A.

11. Campbell, G. S. (1985), “Soil Physics with BASIC”, Elsevier, New York, U.S.A.

12. Celia, M. A., E. T. Bouloutas and R. L. Zarba (1990), “A General Mass- conservative Numerical Solution for the Unsaturated Flow Equation”, Water Resources Research, Volume 26, pp. 1483-1496.

13. Darcy, H. (1856), "Les Fontaines Publiques de la Ville de Dijon", Dalmont, Paris, France.

14. Glass, R. J., T. S. Steenhuis and J.-Y. Parlange (1989), "Mechanism for Finger Persistence in Homogenous, Unsaturated Porous Media: Theory and Verification", Soil Science, Volume 148, pp. 60-70.

15. Hendrickx, J. H. M., L. W. Dekker and O. H. Boersma (1993), "Unstable Fronts in Water-repellent Field Soils", Journal of Environmental Quality, Volume 22, pp. 109-118.

16. Hornung, U. and W. Messing (1981), "Simulation of Two-dimensional Saturated/ Unsaturated Flows with an Exact Water Balance", In. Flow and Transport in Porous Media (Ed. A. Verruijt and F. B. J. Barends), Balkema, Rotterdam, The Netherlands, pp. 91-96.

17. Kemper, W. D. (1986), "Solute Diffusivity", In. Methods of Soil Analysis (Ed. A. Klute), Part 1, 2nd Edition, Agronomy Monogram 9, ASA, Madison, Wisconsin, U.S.A., pp. 1007-1024.

18. Kemper, W. D. and J. C. van Schaik (1966), "Diffusion of Salts in Clay-Water Systems", Soil Science Society of America Proceedings, Volume 30, pp. 534-540.

19. Lehrman, A. (1979), "Geochemical Processes", Wiley-Interscience, New York, U.S.A.

20. Millington, R. J. and J. P. Quirk (1961), "Permeability of Porous Solids", Trans. Faraday Society, Volume 57, pp. 1200-1207.

21. Milly, P. C. D. (1984), "A Mass-conservative Procedure for Time-stepping in Models of Unsaturated Flow", Proceedings, Fifth International Conference on Finite Elements in Water Resources, Springer-Verlag, New York, U.S.A., pp. 103-112.

22. Nye, P. H. (1966), "The Measurement and Mechanism of Ion Diffusion in Soils. I. The Relation between Self-diffusion and Bulk Diffusion", Journal of Soil Science, Volume 17, pp. 16-23.

23. Olsen, S. R., W. D. Kemper and J. C. van Schaik. (1965), "Self-diffusion Coefficients of Phosphorous in Soils measured by Transient and Steady-state Methods", Soil Science Society of America Proceedings, Volume 29, pp. 154-158.

24. Olsen, S. R. and W. D. Kemper (1968), "Movement of Nutrients to Plant Roots", Advances in Agronomy, Volume 20, pp. 91-151.

25. Porter, L. K., W. D. Kemper, R. D. Jackson and B. A. Stewart (1960), "Chloride Diffusion in Soils as influenced by Moisture Content", Soil Science Society of America Proceedings, Volume 24, pp. 460-463.

26. Purandara, B. K. and C. P. Kumar (2000), "Simulation of Soil Moisture Movement in a Forested Watershed", ICIWRM-2000, Proceedings of International Conference on Integrated Water Resources Management for Sustainable Development, 19-21 December, 2000, New Delhi, India, pp. 833-839.

27. Ratnoji, Shilpa S. (2001), “Modelling of Soil Moisture Movement and Solute Transport in an Agricultural Field using SWIM”, M.Tech. Dissertation, Environmental Engineering, 2000-2001, K.L.E. Society's College of Engineering and Technology, Belgaum (Karnataka), India.

28. Richards, L. A. (1931), “Capillary Conduction of Liquids through Porous Media”, Physics, Volume 1, pp. 318-333.

29. Robinson, R. A. and R. H. Stokes (1965), “Electrolyte Solutions”, 2nd Edition, Butterworths, London, U.K.

30. Rose, D. A. (1977), “Hydrodynamic Dispersion in Porous Materials”, Soil Science, Volume 123, pp. 277-283.

31. Rose, C. W., F. W. Chichester, J. R. Williams and J. T. Ritchie (1982), “Application of an Approximate Analytic Method of computing Solute Profiles with Dispersion in Soils”, Journal of Environmental Quality, Volume 11, pp. 151-155.

32. Ross, P. J. (1990 a), “Efficient Numerical Methods for Infiltration using Richards’ Equation”, Water Resources Research, Volume 26, pp. 279-290.

33. Ross, P. J. (1990 b), “SWIM - A Simulation Model for Soil Water Infiltration and Movement”, Reference Manual to SWIMv1, CSIRO Division of Soils, Australia.

34. Ross, P. J. and K. L. Bristow (1990), “Simulating Water Movement in Layered and Gradational Soils using the Kirchoff Transform”, Soil Science Society of America Journal, Volume 54, pp. 1519-1524.

35. Sadeghi, A. M., D. E. Kissel and M. L. Cabrera (1988), “Temperature Effects on Urea Diffusion Coefficients and Urea Movement in Soil”, Soil Science Society of America Journal, Volume 52, pp. 46-49.

36. Smiles, D. E., K. M. Perroux, S. J. Zegelin and P. A. C. Raats (1981), “Hydrodynamic Dispersion during Constant Rate Absorption of Water by Soil”, Soil Science Society of America Journal, Volume 45, pp. 453-458.

37. van Genuchten, M. Th. (1980), “A Closed-form Equation for Predicting the Hydraulic Conductivity of Unsaturated Soils”, Soil Sci. Soc. Am. J., Volume 44, pp. 892-898.

38**.** Verburg, Kirsten, Peter J. Ross and Keith L. Bristow (1996), "SWIMv2.1 User Manual", CSIRO Division of Soils, Australia, Divisional Report No. 130.

39. Weast, R. C. and M. J. Astle (Editors) (1980), "CRC Handbook of Chemistry and Physics", 60th Edition, CRC Press, Boca Raton, Florida, U.S.A.

SWIM V2.1 INPUT FILE FOR BARCHI WATERHSED

```
* Simulation of Soil Moisture Movement in a Hard Rock Watershed
* (Barchi) using SWIM Model
*
* comments start with * in column 1
* ------------------------------------------------------------
* first line must be :
input file for swim v2.1@
* isol
1
* t0,tfin,pint,dw
0  35064  96  .1
* tcycle,ncult
0 0
* ----- i=1,ncult -----
* tcult(i),iprnt(i),idist(i),slapp(i)
* 0 0 1 0
* ---------------------
* dtmin,dtmax,ersoil,ernode,errex,dppl,dpnl,swt
0 24 .000001 .000001 .01 2 1 0
* slcerr,slswt
.000001 0
* psi0,psi1,hair,rad,groot
-50 -5 0.6 0.5 4.0d-7
* nveg
1
* ----- i=1,nveg -----
```

```
* psimin(i),xc(i),rldmax(i),fevmax(i),vcycle(i),iroot(i),igrow(i)
* f1,t1,f2,t2,f3,t3,f4,t4
-15000 150 4 1 0 0 1
2
0     1
35064 1
* --------------------
* ivap
0
* nprop,slmin,slmax,hyscon
20  -3.000000   7.000000 0
* ----- i=1,nprop -----
* sl(i),wc(i),wcd(i),hkl(i),hkld(i)
-3.000000   0.471000  -0.000000  -0.473838  -0.004081
-2.000000   0.471000  -0.000000  -0.480914  -0.011281
-1.000000   0.470999  -0.000004  -0.500617  -0.031639
-0.000000   0.470965  -0.000116  -0.557058  -0.092598
1.000000   0.470047  -0.003129  -0.733332  -0.308302
1.951918   0.452009  -0.053442  -1.358404  -1.224746
2.600241   0.390345  -0.125789  -2.563588  -2.467220
3.350125   0.305385  -0.088169  -4.684655  -3.023609
4.969595   0.232830  -0.017980  -9.671609  -3.091979
7.000000   0.217301  -0.002320 -15.950152  -3.092309
-3.000000   0.394000  -0.000000  -0.195127  -0.006522
-2.000000   0.394000  -0.000001  -0.206206  -0.017414
-1.000000   0.393996  -0.000013  -0.236111  -0.047524
-0.000000   0.393892  -0.000353  -0.320295  -0.138048
1.050000   0.390728  -0.010427  -0.613347  -0.514558
1.989175   0.346817  -0.103530  -1.632535  -1.887056
2.814364   0.244994  -0.111824  -3.709778  -2.909980
4.035274   0.159772  -0.037533  -7.394128  -3.051489
```

```
5.932140   0.127168  -0.005979 -13.186908  -3.054320
7.000000   0.123190  -0.002123 -16.448499  -3.054326
* ---------------------
* slxc,slpmax,slpc1,slpc2,scycle,idepth,itime
10 .0001 -1000 -15000 0 0 0
* f1,t1,f2,t2,f3,t3,f4,t4
.001 24 .009 48 .009 48 .001 72
* slupf,slos,slsci,slscr,d0,a,dthc,dthp,disp
1 0 0 0 0 0 0 1 1
* nprop
1
* ----- i=1,nprop -----
* rhob,exco,fip(i),dis(i),alpha(i),beta
2.65 3 1 2 -.01 -.005
* ---------------------
initial water content
* np
31
* ----- i=1,np -----
* x(i),psi(i),index(i,1),csl(i),indxsl(i)
0  0.215  1 0 1
5  0.215  0 0 1
10  0.215  0 0 1
15  0.215  0 0 1
20  0.215  0 0 1
25  0.215  0 0 1
30  0.215  0 0 1
35  0.215  0 0 1
40  0.215  0 0 1
45  0.215  0 0 1
50  0.121  0 0 1
```

```
55  0.121  0 0 1
60  0.121  0 0 1
65  0.121  0 0 1
70  0.121  0 0 1
75  0.121  0 0 1
80  0.121  0 0 1
85  0.121  0 0 1
90  0.121  0 0 1
95  0.121  0 0 1
100  0.121  0 0 1
105  0.121  0 0 1
110  0.121  0 0 1
115  0.121  0 0 1
120  0.121  0 0 1
125  0.121  0 0 1
130  0.121  0 0 1
135  0.121  0 0 1
140  0.121  0 0 1
145  0.121  0 0 1
150  0.121 11 0 1
* ------------------
* h,cslsur
0 0
* gf,isbc,itbc,ibp,ibbc
1 2 2 0 0
* hm1,hm0,hrc,roff0,roff1,slope (for isbc=3)
2 1 5 2 2
* g1,g0,grc
4 .02 2.5
* xbp,gbp,sbp
* ntb
```

```
2
0     0
35064 0
* csl(n) (for ibbc=1)
* 0
* nts
0
* ntr
-1
* effpar
0
* ----- i=1,ntr -----
* tv(1,i),tv(2,i)
'Rainsd.prn'
* nte
-1
* ----- i=1,nte -----
* tv(1,i),tv(2,i)
'Evaps.prn'
* -------------------
```

DIRECTOR	:	K. S. RAMASASTRI
COORDINATOR	:	B. SONI
STUDY GROUP	:	C. P. KUMAR
		B. K. PURANDARA
SCIENTIFIC STAFF	:	P. R. RAO