MONTHLY WATER BALANCE MODEL FOR LATERITIC HILL SLOPE - A CASE STUDY

BY

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THESIS

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TAVANUR - MALAPPURAM

1996

DECLARATION

I hereby declare that this thesis entitled "**Monthly water balance model for lateritic hill slope - A case study**" is a bonafide record of research work done by me during the course of research and that this thesis has not previously formed the basis for the award to me of any degree, diploma, associateship, fellowship or other similar title, of any other University or Society.

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Certified that this thesis entitled **"Monthly water balance model for lateritic hill slope - A case study"** is a record of research work done independently by **Smt. Beena Thomas** under my guidance and supervision and that it has not previously formed the basis for the award of any degree, diploma, fellowship or associateship to her.

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Dedicated to my Husband

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SYMBOLS AND ABBREVIATIONS

Accum.	-	Accumulated
AE	-	Actual Evaporation
Agric.	-	Agricultral
Amer.	-	American
ASAE	-	American Society of Agricultral Engineers
ASCE	-	American Society of Civil Engineers
Assoc.	-	Association
cm	-	centimetre(s)
cm/h	-	centimetre per hour
Co.	-	Company
Conserv.	-	Conservation
Contd.	-	continued
ET	-	evapotranspiration
ETc	-	crop evapotranspiration
et al.	-	and others
etc.	-	et cetra
FAO	-	Food and Agricultural Organisation
Fig.	-	Figure
Geol.	-	Geological
Geophys.	-	Geophysics
gm	-	gram(s)
h	-	hour(s)
ha	-	hectare(s)
Hydrol.	-	Hydrology

ICAR	-	Indian Council of Agricultural Research
i.e.	-	that is
ICRISAT	-	International Crop Research Institute for Semi Arid Tropics
IST	-	Indian Standard Time
inter.	-	international
J.	-	Journal
K.C.A.E.T	-	Kelappaji College of Agricultural Engineering and Technology
kg	-	kilogram(s)
km	-	kilometre(s)
km^2	-	square kilometre(s)
Ltd.	-	Limited
m	-	metre(s)
m^2	-	square metre(s)
Mgmt.	-	Management
Mha.m	-	Million hectare metre
min.	-	minute
mm	-	millimetre(s)
No.	-	Number
NPK	-	Nytrogen Phosphorus Potassium
PET	-	potential evapotranspiration
pp	-	pages
Proc.	-	Proceedings
pvt.	-	private
Res.	-	Research
Resour.	-	Resources

Sci.	-	Science
sec	-	second(s)
Ser.	-	service
Soc.	-	society
STEC	-	State Committee on Science, Technology and Environment
Surv.	-	survey
Trans.	-	Transactions
Univ.	-	University
USDA	-	United State Department of Agriculture
viz.	-	namely
Vs	-	versus
&	-	and
°	-	degree
%	-	per cent
/	-	per
'	-	minute
"	-	second

Introduction

INTRODUCTION

Water is a precious and most commonly used resource. The human civilization owes its existence to the benefits derived from the exploitation of water for power, irrigation, industries and domestic purposes. Surface water resources, being exploited from time to time, may become short of supply or may not be easily available at site. Groundwater is the largest source of fresh water on the planet excluding the polar ice caps and glaciers. The amount of groundwater within 800m from the ground surface is over 30 times the amount in all fresh water lakes and reservoirs, and about 3000 times the amount in stream channels, at any one time.

Groundwater is an important source of water supply throughout the world, and unlike any other mineral resource it gets its annual replenishment from the precipitation. At present nearly one fifth of all the water used in the world is obtained from groundwater resources. Its use in irrigation, industries and rural homes continues to increase. The present utilization of groundwater is roughly half of the available groundwater and the rest is available for further exploitation and utilization. The demand of water growing day by day, has stimulated the development of underground water supplies.

Kerala has a humid tropical climate with mean annual precipitation between 2000 and 4000mm. Laterites covers one third of the state's area, at heights of about 5 to 75 m above the sea level and is of thickness upto 40 m. In this loose strata a perched groundwater zone builds up in the rainy season. It is estimated that this groundwater reserve constitutes 55 per cent of Kerala's exploitable groundwater. About 50 per cent of this groundwater has traditionally been utilised for domestic purposes through open dug wells. But today the increased demands on our water resources requires intensive management and water conservation. This can be achieved only by increased storage and regulation of water during times of water surplus. The construction of dams alone cannot meet the total water requirement due to topographical, geological and environmental constraints. Groundwater management is thus of focal interest and a need is felt to identify groundwater potential areas and to monitor both quality and quantity aspects. Such groundwater information is basic to an increased development of these resources and their management which in turns increase the agricultural productivity.

Kerala, the small State in the south western corner of India, forms the `type locality' of laterite. The lateritic terrain of Kerala occupies the midland region of the State and this tract can be considered as the backbone of the State, as its economy depends upon this lateritic terrain which produces a variety of cash crops like coconut, cashew, pepper, banana etc.

In India, laterite soil occupy an area of 1,30,066 sq.km and is well developed on the summits of Deccan hills, Karnataka, Kerala, the Easternghats, West Maharashtra and central parts of Orissa and Assam. In the Indian peninsula laterite and related residual deposites of bauxite, iron, manganese and nickel ore have widespread distribution at varied altitudes. Two forms of laterites have been recognised in India, high level and low level, the latter frequently supposed to be of detrital origin. The high level form was found to cap the summits of hills and plateau on the high lands of central and western India whereas, the low level laterite was associated with large tracts in the neighbourhood of both coasts of the Deccan Peninsula.

The subsurface occurence of groundwater may be divided into zone of saturation and aeration. In the zone of saturation all interstices are filled with water under hydrostatic pressure. The zone of aeration consists of interstices occupied partially by water and partially by air. The saturated zone is bounded at the top by either a limiting surface of saturation or overlying impermeable strata, and extends down to underlying impermeable strata such as clay beds or bed rock. In the absence of overlying impermeable strata, the upper surface of the zone of saturation is the water table, or phreatic surface. This is defined as the surface of atmospheric pressure and would be revealed by the level at which water stands in a well penetrating the aquifer. Actually, saturation extends slightly above the water table owing to capillary attraction, however, water is held at less than atmospheric pressure.

Water occuring in the zone of saturation is commonly referred to as groundwater. In the zone of aeration suspended, or vadose water occurs. This general zone may be further divided into the soil water zone, intermediate zone and the capillary zone.

Practically, all groundwater originates as surface water. Principal source of natural recharge include precipitation, streamflow, lakes and reservoirs. Other contributions, known as artificial recharge, occur from excess irrigation, seepage from canals and water purposely applied to augment groundwater supplies. Water within the ground moves downward through the unsaturated zone under the action of gravity, where in the saturated zone it moves in a direction determined by the surrounding hydraulic situation.

Discharge of groundwater occurs when water emerges from underground. Most natural discharge occurs as flow into surface water bodies, such as streams, lakes and oceans; flow to the surface appears as a spring. Groundwater near the surface may return directly to the atmosphere by evaporation from within the soil and by transpiration from vegetation. Pumpage from wells constitutes the major artificial discharge of groundwater.

Groundwater studies are effected in order to ascertain the quantity of water available for development in a region and this can be done only after the identification of various physical features of the hydrologic systems involved, their hydraulic characteristics and their hydraulic interrelationships.

A balance exists between the quantity of water entering an area, change of storage water in the area, the evapotranspiration taking place in the area and the water leaving the area. This situation depicting the totality of all components, is known as hydraulic equilibrium, simply water balance. This water balance helps in analyzing the water budgeting in any area of interest.

The general equation for water balance is the quantitative statement of the balance between the total water gains and losses of a basin for a specific period of time. This balance considers all water, surface and subsurface entering and leaving or stored or depleated within the basin. Water entering the basin is equated to water leaving the basin plus or minus changes in the basin storage. Under specific conditions, precipitation is the source of water entering the basin and is the only water gain considered in the water balance study. Water leaving the basin includes streamflow, evapotranspiration and subsurface underflow. The storage belong to soil moisture, groundwater and surface water. There are many situations in which several items of the water balance equation can be eliminated because they do not measurably affect the balance between water gains and losses.

The flow of groundwater and water level fluctuations can be described and analysed mathematically using equations, provided that adequate hydrologic and geologic informations is given. A groundwater model can be designed as a non-unique, simplified version of a real groundwater system, describing the features essential to the purpose for which the model was developed, and including various assumptions and constriants pertinent to the system.

The present study aims at giving a deeper insight into the basic processes of saturated water flow in laterites and to quantify groundwater replenishment in relation to rainfall and geohydrology of the lateritic terrain.

The general objective of this research work is to analyse the water balance of a lateritic hillslope based on rainfall, evapotranspiration, water table fluctuations and soil properties of the area.

The specific objectives are:

1. Analysis of infiltration and surface runoff generation.

2. Study of recharge and saturated flow through lateritic hillslope.

3. Development of a model to predict seasonal water table variations.

Review of Literature

REVIEW OF LITERATURE

Groundwater can be developed at a small capital cost and the time taken for development is very small. The tapping of groundwater - location, spacing and yield, in a well field should be so phased that the annual recharge and discharge of the aquifer are almost balanced without causing an over draft in the area. The average annual groundwater recharge from rainfall and seepage from canals and irrigation systems is of the order of 67Mha.m of which 40 per cent i.e., 27Mha.m is extractable economically. The present utilization of groundwater is roughly half of this (13Mha.m) and about 14Mha.m is available for further exploitation and utilization.

The groundwater reservoirs gets water as a result of recharge from rainfall, rivers, streams, irrigation etc. and loses water due to regeneration in streams, movement towards other aquifers and man-made withdrawals. A study of groundwater balance is essential in order to evaluate the total groundwater resources of a basin.

Study of fluctuation of water table, infiltration characteristics, rainfall characteristics, groundwater recharge etc. are needed for the water balance study. In this chapter an attempt is made to give a brief review of literature relevant to the topic of study undertaken in the past.

2.1 Laterite soil

Laterite is an iron-rich product of long, intense tropical weathering. It forms in certain subsurface horizons where all but the most resistant materials are weathered away. Quartz, sand and kaolinite clay remain and iron accumulates as dark red mottles. Latosol and lateritic soils occur in tropical climates. They are the best known soils of the tropics, though extensive areas of red-yellow pedzolics and even some black soils also occur there. Some of the lateritic soils contain as much as 80 per cent oxide clays in their mineral matter. These soils generally have very deep profiles. The climate is warm and humid all or much of the year. The native vegetation is tropical forest. These soils generally are poor in NPK and organic matter. The pH ranges between 4.5 to 6.0.

In Kerala at Angadipuram, a ferruginous, vesicular, soft material occur within the soil, which hardens irreversibly on exposure and used as a building material, was first recognised as `laterite' by Francis Buchanan, a medical officer in the service of the East India Company. He (1807) suggested the name laterite, from `later', the Latin word for brick.

Harrassowitz (1930) presented a morphological definition for laterite, as one with a characteristic profile developing under tropical savannah and forming the following four levels or horizons in ascending order from subsurface to surface: (a) a fresh zone (b) a zone of primary alteration to kaolinite (c) a laterite bed (d) a surface zone with ferruginous incrustations and concretions.

Sathyanarayana and Thomas (1962) were the first to study the chemical properties of laterites occuring in the West coast of India. Their studies on two insitu laterites revealed that the profiles were acidic with pH range of 4-5 and extremely low in cation exchange capacity.

With respect to its colour, laterites vary considerably and the shades most frequently encountered are pink, ochre, red and brown, but some occurrences are mottled and streaked with violet, and others exhibit green marbling (Maignein, 1966). The apparent density will be higher at the surface and it decreases with increase in depth.

The term `plinthite' (USDA, 1975) replaces `laterite' in Buchanan's sense, and the already indurated materials are considered as ironstone or hard crusts not coming under the purview of soil.

2.2 Aquifer properties

Two properties of an aquifer related to its storage function are its porosity and specific yield. Porosity is the ratio of the volume of voids or pores in a soil mass to its total volume. The volume of water, expressed as a percentage of the total volume of the saturated aquifer, that can be drained by gravity is called the specific yield and the volume of water retained by molecular and surface tension forces against the force of gravity, expressed as a percentage of the total volume of the saturated aquifer, is called specific retension and corresponds to field capacity.

Porosity = Specific yield + Specific retension

Specific yield depends on grain size, shape and distribution of pores and compaction of the formation. The values of specific yield for alluvial aquifers are in the range of 10 to 20 per cent and for uniform sands about 30 per cent.

To evaluate the subsurface hydrologic consequences of any groundwater recharge technique, response of the subsurface flow system to induce recharge must be quantified. Quantification requires both tools and parameters which represent the aquifer's hydraulic properties. Knowledge of the hydraulic properties enables the hydrogeologist to calculate the subsurface storage.

Sophocleous (1986) found out the importance of capillary fringe and variable specific yield phenomena in groundwater recharge estimates. This study demonstrated the effect of a constant specific yield value on the behaviour of the water table rise. Calculation of specific yield value by any inverse procedure based on water level changes and on assumed or estimated values may be erroneous. Calculation of the recharge value based on observed water level fluctuations and the ultimate specific yield value usually are over estimated. He quantified the errors involved in such recharge calculations.

Neuman (1987) presented a comparative discussion of several methods for the determination of specific yield. The specific yield values are consistent with water balance considerations when all the components of the water budget are properly taken into account. The rate at which the groundwater level fluctuate in response to pumpage is controlled by the smaller specific yield that obtained from the time - drawdown analysis.

Nautiyal (1991) carried out systematic studies for the determination of specific yield and hydraulic conductivity and monitoring of water table fluctuations for the assessment of groundwater resources of shallow aquifers of Upper Ganga Basin, Uttar Pradesh. Specific yield was determined by three methods namely, column drainage method, method based on grain sizes of aquifer material and pump testing method. Pump testing method is the most common field method for the determination of specific yield and transmissivity. Pump testing method requires pumping at a constant discharge from the well and recording its effect (drawdown) at different time intervals in an observation well situated at a distance. In the column drainage method for the determination of specific yield, the volume of water drained at different time intervals from the aquifer sample was noted until the free drainage was negligible. The ratio of volume of water drained to the volume of saturated aquifer column in percentage gave the specific yield. The hydraulic conductivity was measured with a variable head permeameter. The hydraulic conductivity values show a larger variation and most of the values are in the range of 1 to 6m/day.

2.2.1 Infiltration

Infiltration is defined as the entry of water from the air side of the air soil interface into the soil profile. The rate of movement of water into the soil will depend on the magnitude of the forces and gradients and also on the factors determining the hydraulic conductivity of the soil. The aspects of infiltration which are being considered important in hydrology are cumulative infiltration and infiltration capacity. Cumulative infiltration is the total quantity of water that enters the soil in a given time and infiltration capacity is the maximum rate at which water can be absorbed by the soil in a given condition.

The physical properties and depth of the soil have probably the most important controls on subsurface flow production at a site. If the texture is coarse (with predominant sand and stones), vertical flow usually dominates; and when this soil is deep, subsurface flow response may be delayed. If the texture is fine, resistance to vertical flow results and lateral or shallow surface flow sometimes occurs quickly.

Horton (1933) reported that, of all hydrological variables, the infiltration capacity of the soil was the easiest to measure with accuracy, and that in conjunction with rainfall - intensity data, both surface runoff and total infiltration to ground water might be determined.

Sherman (1944) showed how rates of surface infiltration are inverse functions of the volume of capillary moisture in the soil column and that surface capillary intake decreases as the water penetrates deeper into the soil, although gravity flow in the large channels continues to provide water at depth for lateral capillary absorption.

Holtan (1961) proposed a variation of Horton's infiltration when supply of water at the surface is not the limiting factor, but the problems of estimating infiltration through wide ranges of surface supply remain. Horton's empirical infiltration equation gives poor results for short-term infiltration rates, which are precisely those most important in governing hillslope hydrology.

Poeson (1984) reported that the soil saturated on steep slopes will absorb, especially at the beginning of a rainfall event, more rain water compared to the soil saturated on a low slope. This is due to a spatially varying matric potential induced by a gravitational potential.

Varadan and Raghunath (1985) reported that infiltration rates for latesols of Kerala are 12-20 cm/h after 6 hours of study. They also reported that infiltration rates increase towards higher elevation and such variation occurred for latesols of Kerala even at an elevation difference of three metres.

2.2.2 Fluctuations of the water table

The water table represents the groundwater reservoir level and changes in its level represent changes in the groundwater storage. Maps showing the rise or fall of the water table in a specific time interval can be prepared from water level data of wells. Time dependent water level changes in wells can be depicted by hydrographs. A decline in the water table represents groundwater abstraction in excess of increment, while a rise represents groundwater increment in excess of abstraction. Most groundwater assessment studies involve correlation of watertable fluctuations as recorded in well with climatic elements such as rainfall, hydrologic influences such as fluctuations in surface water bodies and man made causes like application of irrigation water, artificial recharge, withdrawals from wells etc. All factors remaining the same, water table fluctuation is inversely proportional to specific yield.

In areas with well defined seasonal rainfall the water table rises and falls in annual cycles, the rise corresponding to the rainfall period and the low stage corresponding to the dry period. The water level rise does not commence immediately with the onset of the rainy season as the initial rain have to satisfy the soil moisture deficit which is at its maximum at the end of a dry spell. The magnitude of the water table fluctuation depends also on climatic factors, drainage, topography and geological conditions.

Marino (1974) studied the water table fluctuation in response to recharge. Solutions have been derived which describe the rise and fall of the water table in an extensive unconfined aquifer receiving uniform localised recharge and discharging into a surface reservoir in which the water level remains equal to that of the main flow before the incidence of recharge. The solutions are expressed in terms of the head averaged over the depth of saturation and are applicable when the rise of the water table is smaller than 50 per cent of the initial depth of saturation. When prediction of future water level is desired, the equation should be used in conjunction with the method of successive approximations.

Chapman *et al.* (1987) developed a simple model for predicting water table elevations at different locations on a hill slope, as a function of daily rainfall. The model predicts spatial variations in runoff, evaporation and recharge, without explicitly simulating the lateral transport processes of overland flow and groundwater flow. The model was calibrated against a nine month record of piezometric heads in three bores on a hill slope with spatially variable soil profiles, which was subject to land sliding.

Rai *et al.* (1988) derived an approximate solution of the non linear Boussinesq equation which describe the water table variations in a ditch-drainage system with a random initial condition and transient recharge. The numerical results reveal that the water table variation is significantly influenced by the random initial condition and the transient rate of recharge. The amplitude of variation is maximum at the groundwater divide.

Lal *et al.* (1991) compared the water table fluctuation predicted by different models. The model described in this work have been used to simulate the water table behaviour in response to subsurface drainage for climatological and soil conditions prvailing at Sampla in Haryana. A field experiment on subsurface drainage was conducted to control the water table and salinity in the water logged saline soils at Sampla. The experiment consists of three tile drain spacings of 25m, 50m and 75m. The average depth of tile line below ground surface was 1.75m. Two models for predicting water table namely, de Zeeuw - Hellinga and Van Schilfgaarde were selected to their field applicability by comparing the observed water table heights with the predicted water table for the period July to September 1985 for 75m drain spacing. Van Schilfgaarde model was found to be more satisfactory for its application in the field condition.

Zomorodi (1991) derived a new method for the evaluation of the response of a water table to artificial recharge. The solution is formulated in the form of a simple numerical model. The model has several advantages over the traditional methods of mounting prediction. The effect of the unsaturated zone which modifies the recharge rate as compared with the infiltration rate is considered. Mounting is calculated for a variable recharge rate induced by a variable infiltration rate. Also, the effect of in-transit water in reducing the fillable pore space above a rising water table is considered.

The validity of the model results is illustrated using several sets of field data collected from the Ghazvin Plain, Iran. Sample calculations proved that the model predicts mounting more accurately than the traditional methods and therefore, more realistic recommendations for the design and operation of artificial schemes are possilble using this model.

2.2.3 Groundwater recharge

Groundwater recharge is that amount of surface water which reaches the permanent water table either by direct contact in the riparian zone or by downward percolation through the overlying zone of aeration. The methods for the estimation of recharge are generally based on the following parameters: intensity and duration of rainfall, evapotranspiration, soil moisture, runoff, infiltration capacities of soils, storage characteristics of aquifers, water level fluctuations and movement of groundwater. The various recharge components to be estimated and the methods employed are discussed below.

2.2.3.1 Soil moisture balance

Infiltration occuring at the land surface can be estimated by the soil moisture balance approach. The soil moisture balance for any time interval can be expressed as:

$$P = AE + I + R + \Delta Sm$$

where

P = rainfall

AE = actual evapotranspirtation

ΔSm = change in soil moisture storage

I = infiltration , and

R = surface runoff

Soil moisture budgeting, taking into account evapotranspirational abstraction from precipitation, provides a measure of moisture available for runoff and infiltration. This can be done by Thornthwaite's book keeping method of moisture balance. In this method, measurement of field capacity and wilting point are made to determine the available moisture down to the root zone. Monthly PET and rainfall are tabulated and compared. If rainfall P in a month is less than PET, then AE = P, the period being one of water deficit. If P is more than PET, then AE = PET, the balance of rainfall raising the moisture level of the soil to field capacity. After meeting the soil moisture deficit, the excess of rainfall over PET becomes the moisture surplus, also called water surplus. The moisture surplus results in surface runoff and recharge to the groundwater body. The runoff can be determined by gauging at the basin outlet, or estimated from the rainfall-runoff curves. The difference between the moisture surplus and runoff is the groundwater recharge.

The application of this method for the estimation of groundwater recharge requires information on runoff. However, in respect of arid and some semi-arid regions with no marked drainage courses, runoff can be ignored and the entire water surplus can be treated as groundwater recharge.

For long durations of rainfall, in excess of infiltration capacity, the hydrologic equation can be reduced to the following, as evapotranspirational loss is negligible:

$$P = R + Wp$$

where

P	=	Rainfall
R	=	Surface runoff , and
Wp	=	Recharge by infiltration from rainfall

For periods recording rainfall in excess of infiltration capacity, recharge can be estimated by superimposition of the rainfall-intensity curve over the infiltration-capacity curve. For this purpose, the rainfall should be recorded by self recording rain gauges. Rainfall in excess of infiltration capacity represents surface runoff while the rest contributes to recharge provided the moisture changes in the soil can be ignored.

The infiltration method can be used to estimate recharge from ephemeral streams, canals and flooded areas if the extent of wetted area and duration of wetting are known. However, recharge is limited to the available pressure head. The method is very approximate as soil moisture changes are not taken into account. Besides, rainfall of varying intensity results in a distortion of the capacity curve and recharge values.

2.2.3.2 Estimation from base flow

Base flow from a basin is an indirect measure of recharge as it represents the drainage of groundwater from aquifer storage after groundwater recharge has occured.

Mayboom (1961) suggested a method of determining groundwater recharge which involves analyses of a part of the runoff hydrograph representing groundwater recession, by applying Butler's equation

$$Q = \frac{K_1}{10^{t/k2}}$$

where

Q = discharge at any given time

K_1 = Q at t_0 , and

K_2 = Q at t when $Q = 0.1\ K_1$, or time increment corresponding to a log-cycle change in Q.

A major advantage of using base flow separation of stream hydrographs for recharge calculation is that the method does not require broad assumptions. A major disadvantage of this technique is that each worker uses his own arbitarary method of separating base flow from total stream discharge and hence comparisons cannot always be made between work of different authors. Moreover, stream gauges are seldom located at the positions desired.

2.2.3.3 Flow net analysis

A flow net has to be prepared using data from the wells. The data included piezometry of these locations, as well as all available transmissivity data from aquifer system. The annual average groundwater discharge can be calculated which will be approximately equivalent to annual recharge and the storage changes becomes insignificant with time. This method requires knowledge of the geometry of the system and aquifer properties, which may vary spatially. Caution is needed in translating water table rises into recharge, since factors such as air entrapment, changes in atmospheric pressure and hydrologic influence from surrounding areas may give rise to misleading conclusions.

2.2.3.4 Groundwater level fluctuation method

The groundwater table and its fluctuations are functions of groundwater recharge, water yielding properties, transmissivity and geometry of aquifer. The groundwater recharge, during the time period Δt, can be written as

$$GWR = h \times Sy \times A + Q \times \Delta t$$

where

GWR	=	groundwater recharge
h	=	change in groundwater level
A	=	area, and
Q	=	net groundwater flow

Here, the recessions derived from observations of decreasing groundwater levels when no recharge is supposed to occur is used. The shape of the recession curve depends on water yielding properties, transmissivity and geometry. If the groundwater table is shallow, the recessions may be influenced by evapotranspiration and frost penetration. To get the unaffected recession curve, caused only by groundwater flow, periods must be found when these processes are insignificant. The distance between the actual groundwater level and the groundwater level calculated from the recession curve, for every time interval, can then be summed up and multiplied by the specific yield to get the recharge.

The method is attractive since groundwater level observations often are available. The method could also give information of temporal and areal variations.

2.2.3.5 Nuclear methods

Nuclear techniques that have found application in groundwater estimations aims at tracing the movement of water in the saturated and unsaturated zones by one of the following methods.

1. Studying the variation in environmental isotopes in the water ($^{18}O/^{16}O$, $^{2}H/^{1}H$ etc.).

2. Tagging the water with artifcially produced radio isotopes such as ^{3}H.

3. Dating of groundwater.

The recharge is not obtained directly by these methods, but has to be inferred or estimated from the velocities of soil moisture or groundwater movement.

Three tracers considered promising for recharge studies are bromide, tritium and gamma tracers. One of the major drawbacks with environmental tracers is that their exact inputs are not accurately known. Furthermore, these inputs are not contolled in their amount, time of input and spatial distribution.

2.2.4 Estimation of groundwater recharge

Khan (1980) estimated the groundwater recharge in a basin from the water balance study. Rainfall, streamflow evapotranspiration, subsurface underflow, surface water soil moisture and groundwater storage are calculated on monthly basis to be used in the balance equation. The computation was performed by providing monthly hydrological input for the desired period. The average groundwater recharge to the basin was found as output by averaging the former values.

Leonard *et al.* (1981) evaluated the natural groundwater recharge of a tertiary aquifer system Barwon Downs Graben, Otway Basin, Victoria. Natural recharge of the aquifer system occurs by direct infiltration of precipitation in the Barongarook High intake area. A number of techniques, including flow net analysis and hydrograph separation, have been applied to quantify natural recharge. Results of the flow net analysis indicate that the natural recharge into the aquifer system in the Barwon Downs Graben is of the order of 4800 ML per year.

Sharma (1986) made a study on the measurement and prediction of natural groundwater recharge and gave a brief overview of recharge processes, merits and the limitations of estimation methods. Recharge rates are spatially variable on a relatively small scale and approximate methods are required to assess the response of aquifers at varying scales. For many systems, existance of preferred pathways of water flow has been identified.

Ghassemi *et al.* (1987) estimated the vertical recharge via. solutions of groundwater flow equation under approximate assumptions regarding the distributions for transmissivities and storativities through out the aquifer and its flow boundary conditions. This approach is illustrated with a case study wherein groundwater flow in the Parilla sand aquifer in the Mildura-Merbein area of the Sunraysin region of Victoria is simulated. Vertical recharge occurs mainly from rainfall and irrigation. From computations of the groundwater balance and salt flow from the aquifer to the river, vertical recharge from irrigation practices in the area is shown to contribute significantly to the flow of salt to the River Murray.

Nolan (1987) estimated the areal recharge in the Riverine plane of the South Eastern Australia. Net groundwater recharge within the Riverine plane shows significant spatial and temporal viability as a consequence of seasonal and long term variations in climatic conditions, land use practices including irrigation and hydrogeological heterogeneity. An empirical model based upon a conceptualisation of the dominant recharge mechanism has been developed to predict seasonal recharge rates for 60sq.km cells within the study area. When used in conjunction with a regional groundwater flow model, the calibrated coupled model was capable of simulating the seasonal fluctuations and long term trends of the regional groundwater table.

Simon (1987) developed a computer based process model designed to predict recharge to groundwater using daily data on rainfall and PET was described. A sensitivity study of the process model was presented. The model is shown to estimate recharge previously established for the Sherwood Borefield especially over longer time periods. The use of the process model to provide the monthly rec~~a~~harge data for the groundwater model was examined. The model was likely to be useful in estimating recharge variations when only long term estimates of recharge are available.

Farrington *et al.* (1988) did the water and chloride balances of Banksia Woodland on coastal deep sands of South Western Australia. Estimation of groundwater recharge using the water balance approach showed considerable variation between years. Recharge was highly correlated with the annual rise in groundwater table and the rainfall received during winter and spring. A long term estimate of groundwater recharge at the site, using the chloride balances was similar to the average value obtained using the water balance method when estimates of chloride concentration of soil water in the unsaturated zone and in the groundwater beneath the soil water were used.

Khan *et al.* (1989) developed an information system for implementing groundwater recharge models. Models were developed to estimate groundwater recharge by using weather data, soil properties and other land characteristics as inputs. These data, at best, are scattered and difficult to obtain even for single sites; since many are only available for selected locations, it usually impossible to obtain areal inputs required by even the simplest groundwater recharge model.

Nielsen (1989) developed a model for groundwater recharge in Southern Bali, Indonesia. A model for recharge was prepared, using all relevant available soil, landuse, hydrogeological and meteorological data for callibration. The calculated annual recharge in a year, which above average rainfall is to be 645mm for light soils, 538mm for medium soils and 376mm for heavy clay soils. Annual recharge in an average rainfall year was 605mm for light soils, 504mm for medium soils and 308mm for heavy soils. In a drought year recharge was 481mm in light soils, 401mm in medium soils and 267mm in heavy soils.

Thorpe (1989) determined a method in which tritium as a tool is used for assessing total and net groundwater recharge to the Gnangara groundwater mound, Perth, Western Australia. Total recharge rates were to be estimated by using the seasonal variation in tritium concentrations as indicators of rates of soil moisture movements through the unsaturated zone. However, at all sites tritium concentrations at shallow depths were significantly higher than those measured in the previous season's rainfall, consequently total recharge rates could not be estimated. Net recharge rates were determined by measurement of tritium-depth profiles in the saturated zone from piezometer clusters and a multilevel piezometer.

Reddy (1991) utilised the water balance model for the estimation of the groundwater recharge in a basin. The study was conducted at Dulapally basin which covers 34sq.km area in the granitic terrain near Hyderabad city. Recharge to the groundwater regime mostly takes place through vertical infiltration from the ground surface. In the water balance model, various components of water balance viz., surface runoff, actual soil evaporation, actual transpiration from vegetation, soil moisture status and ground water recharge have been computed, following a daily soil moisture accounting procedure. Groundwater recharge was estimated using the balance among various components.

Yong and Chun (1993) did a comparative study of calculation methods for recharge of rainfall seepage to groundwater in plain area. Groundwater in the plain area of China is recharged with rainfall and to a lesser extend with surface water and nearby groundwater. There are several methods for the calculation of recharge of rainfall seepage to groundwater. The calculated values using different methods vary due to the accuracy of parameters and calculation methods chosen. The comparative study is made using the water balance method and the groundwater regime analysis method applied to the plain area of China. The accuracy of groundwater regime analysis method is superior to that of the water balance method.

2.3 Water balance - Basic considerations

Quantification of groundwater and surface water resouces of any basin (or area) involves the application of the principle of conservation of mass, sometimes referred to as the continuity equation, to account for the quantitative changes occuring in the various components of the hydrologic cycle as applicable to the basin. The quantitative changes may be expressed as a water balance equation, in which the inflow, outflow and change in storage in a period are represented by individual components.

The equation of hydrologic equilibrium is

$$I = O + \Delta S$$

where

I = Inflow

O = Outflow, and

ΔS = Change in storage

Obviously, this is grounded in the premise that a balance exists between the amount of water which enters an area, change in the amount of water in storage in the area and the water which leaves the area. Clearly the items in the equation can include a number of factors of which the most significant may be listed as:

(a) Surficial, subsurface, precipitation, imported water

(b) Surficial, subsurface, consumptive usage, exported water

(c) Change in surface storage, change in groundwater storage, change in soil moisture

The factor, subsurface outflow can be eliminated in cases where outflow is measured at a point where impermeable strata direct all outflow to the surface. Storage can be neglected if the balance is determined for any time period during which no significant change in this is believed to occur. In view of this, it is possible to utilize a simplified version of the equation of hydrologic equilibrium in appropriate situations and this may be given as

$$P = O + ET$$

where

P = Precipitation

O = Outflow, and

ET = Evapotranspirative losses

In such cases, the following steps may be taken in order to obtain relevant information regarding these parameters.

2.3.1 Analysis of precipitation, outflow and ET

Firstly, the quantity of precipitation falling in a study area has to be determined.

Total outflow comprises of two components namely, water which moved over the surface of the land and water which after infiltration, moved through the zone of saturation so that

$$O = O_s - O_g$$

where

O_s = the overland runoff, and

O_g = the groundwater discharge

It is necessary to separate groundwater discharge from the total outflow. In doing this it is taken for granted that:

(a) During good weather periods, all the flow in many stream consists of water which has been discharged from the groundwater system.

(b) Groundwater discharge is proportional to groundwater level.

Groundwater level can be ascertained by making measurements in observation wells and the number of these must suffice to show the mean groundwater level for an area.

Evapotranspiration involves both the evaporation of moisture from exposed surface and the transpiration of moisture by plants. Usually, it is very difficult to separate these and so they are treated together as ET. It may refer to surficial ET or ET from the zone of saturation, the former including both the surface and soil zone.

Thus the relationship may be given as:

$$ET = ET_s + ET_g$$

where

ET_s = ET from surface and soil zone, and

ET_g = ET from zone of saturation

Losses due to ET may arise from a variety of factors such as the temperature, the duration of sunshine, relative humidity and wind velocity.

2.3.2 Water balance equations

Groundwater flows and surface water flows in a region are intimately connected. Hence water balance studies can be divided into two parts.

(i) water balance studies

(ii) groundwater balance studies

2.3.2.1 Water balance studies

Water balance of an area is defined by the hydrologic equation which states that in a specified period of time all water entering into a given area must be consumed, stored or go out as surface or subsurface flow. The following equation describes the water balance of a given region :

$$P + Wi = R + Et + Wo \pm \Delta Sg \pm \Delta Ss$$

where, all quantities are in volume for a specified period and

P = precipitation

Wi = surface and groundwater imported from area outside the region

R = stream outflow

Et = evapotranspiration

Wo = groundwater outflow

ΔSg = change in groundwater storage, and

ΔSs = change in soil moisture storage

Water balance for a given region should be worked out for a long period so that various items approach a steady state due to averaging out of climatic effects.

2.3.2.2 **Groundwater balance studies**

A groundwater inventory of an area quantifies the various means of recharge to or discharge from the groundwater reservoir as well as changes in storage therein. It may be stated as follows:

$$\Delta Sg = (Rp + Rn + Ra + Gi) - (Et + De + Da + Go + Ge)$$

where

ΔSg = change in ground water storage during the period in question

Rp = recharge due to precipitation

Rn = natural recharge from streams and lakes i.e., influent seepage

Ra = artificial recharge from canals, reservoirs, irrigation return flow, streading and injection wells

Gi = groundwater inflow from area outside the basin

Et	=	evapotranspiration from capillary fringe in shallow water table areas and from vegitation
De	=	natural discharge by seepage and stream flow i.e., effluent seepage loss
Da	=	artificial discharge due to pumping and consumptive use
Go	=	leakage from a bottom semi confined layer, and
Ge	=	groundwater outflow to areas outside the basin

General groundwater equation as given by Tolman (1937) is

$$R = E + S + I$$

where

R	=	rainfall
E	=	evaporation and transpiration loss
S	=	water discharged from the area, and
I	=	groundwater increment

Khan (1980) used the following equation for the estimation of groundwater recharge in a basin water balance study.

$$P = SF + AE + SW + SM + \Delta GW + GRC$$

where

P	=	precipitation over the basin
SF	=	stream flow from the basin
AE	=	actual evapotranspiration
SW	=	change in surface water
SM	=	change in soil moisture
ΔGW	=	change in groundwater storage, and
GRC	=	groundwater recharge

For a long term average annual water balance computation, the long term average of the annual changes in the storages are small and can be neglected. Thus the above equation becomes,

$$P = SP + AP + GRC.$$

Dhir (1980) studied the hydrological balance and the influence of utilisation of groundwater upon it. He related the role of groundwater in the hydrological cycle with independent variables representing the precipitation, PET, soil and plant characteristics and water table elevation. Uncertainity was introduced into the equation through the probability function of climatic variables and the probability distribution of water yields and the water balance elements. The mean values of runoff, ET and groundwater recharge have a long term average water balance which to the first order defines the annual water yield and water loss in terms of the annual precipitation, PET and physical parameters of the soil, vegetation, climate and water table.

Pinol *et al.* (1991) did the hydrological balance of two Mediteranean forested catchments located in Prades in Northeast Spain. Precipitation and discharge have been measured for several years in these catchments. Actual evapotranspiration has been calculated as the difference between annual precipitation and discharge. Results show that

(a) most of the precipitation is evaporated rather than loss by streamflow even in most humid years (b) there is a high inter-annual variability both in discharge and ET and (c) annual ET correlates significantly with annual precipitation, in contrast to the constancy of annual evaporation in catchments of wet and colder climates. Finally, a simple expression is proposed in order to calculate annual evaporation from the ratio of precipitation to PET. This expression uses a derived exponent which takes into account the characterstics of individual catchments.

Langsholt (1992) has studied the water balance of a 600sq.m field site on a lateritic hill slope in Kerala in India, during two southwest monsoon seasons. Surface runoff was of minor importance while infiltration and evapotranspiration were the major components amounting to approximately 2/3 and 1/3 of the rainfall, respectively. Groundwater response was rapid, involving fluctuations of several meters. Groundwater recharge was found normally to takes place during the southwest monsoon season only. The field study demonstrated that seasonal shallow groundwater recharge representing a major portion of the rainfall may be observed in this lateritic terrain.

2.4 Water balance modelling

Hydrologists concerned with setting up various types of model simulating parts of the hydrological cycle or their mutual effects in the tropics. Many models have been developed which can be applied directly or with some modifications to tropical regions. Perhaps the later is more typical because many formulae need to be revised before being applied in tropical regions.

Brown *et al.* (1981) successfully applied one of the largest model for the hydrological balance of the upper Nile basin. The model was set up in order to evaluate alternative plans for water resources development and regulation of the regime in the basins of great African lakes. The model consists of three submodels, namely a catchment model, a lake model and a channel model.

Harikrishna (1982) developed a parametric water balance simulation model using hydrological data collected from small agricultural watersheds on vertisols at ICRISAT. From this, satisfactory prediction of runoff and other components of the water budget such as ET, soil moisture etc. can be obtained. The model parameters appears to be transferable between similar sites, which will be helpful in runoff prediction for new locations.

Blain *et al.* (1991) developed a vertically integrated two dimensional lateral flow model of soil moisture. Derivation of the governing equation is based on a physical interpretation of hillslope processes. The lateral subsurface flow model permits variability of precipitation and ET and allows arbitrary specification on soil moisture retention properties.

Maimone (1991) used a regional groundwater model for the water resource planning in Nassau country, New York. In addressing the fundamental question of the ability of the aquifer to yield sufficient quantities of water for public supply, two approaches have been used. The first is hydrograph analysis of water levels over a 45 year period. The second approach involves using a three dimensional finite element groundwater model. The model consists of 608 nodes and 1143 elements, and covers Queen Country, Nassau Country and the western part of Suffolk Country. The model has eight levels, and contains all the major geologic units. The model was calibrated by matching computer simulated groundwater levels and streamflows with field measured data. Water balances were developed for each aquifer under predevelopment and present conditions. This model can turn water level data into something that lends itself to hard conclusions about the state of the aquifer system, by providing a meaningful water balance of each aquifer for any given set of pumpage and rainfall conditions.

Vandewiele *et al.* (1991) studied more than 60 catchments from Northern Belgium ranging in size from 16 to 3160sq.km by means of a physically based stochastic water balance model. The parameter values derived from calibration of the model were regionally mapped for the study region. Associations between model parameters and basin lithological characteristics were established and tested. The results showed that the simple conceptual monthly water balance model with three parameters for actual ET, slow and fast runoff was capable either to generate monthly stream flow at ungauged sites or to extent river flow at gauged sites. This allows a fairly accurate estimation of monthly discharges at any location within the region.

Langsholt (1994) developed an analytical approximate model for an unsaturated flow in a spatially varied field, coupled with infiltration and ET at the lower boundary. The relation between hydraulic conductivity and soil moisture content on one hand and metric suction on the other, $K(\psi)$ and $\theta(\psi)$, are parameterised and the unsaturated flow equations depends on parameterisations of these. They are based on the notion of a moving, discontinous front. The field heterogeneity refers to saturated hydraulic conductivity only. Horizontal variability is considered, and the flow medium is approximated as a set of uncorrelated, vertically homogeneous columns. Expectations and variances obtained with this approach have been compared with observations of the field hydrological processes.

Materials and Methods

MATERIALS AND METHODS

A planned investigation of groundwater resources is essential to obtain full information about the occurrence of groundwater in a region. Water table which defines the upper boundary of groundwater is an important parameter in assessing the water potential in a region. Therefore, a model has been developed for locating the water table based on precipitation, evapotranspiration, specific yield etc. In this chapter, the methods adopted for the estimation of different parameters and the methodology used for the development and verification of the model are described.

3.1 Location

The selected study site is K.C.A.E.T. campus, Tavanur in Malappuram District, Kerala situated at 10°53'30" North latitude and 76° East longitude. Bharathapuzha river forms the northern boundary of the study area. The campus covers an area of 40.99ha., out of which, the total cropped area comprises 29.65ha.

The location map for the study area is given in fig.1.

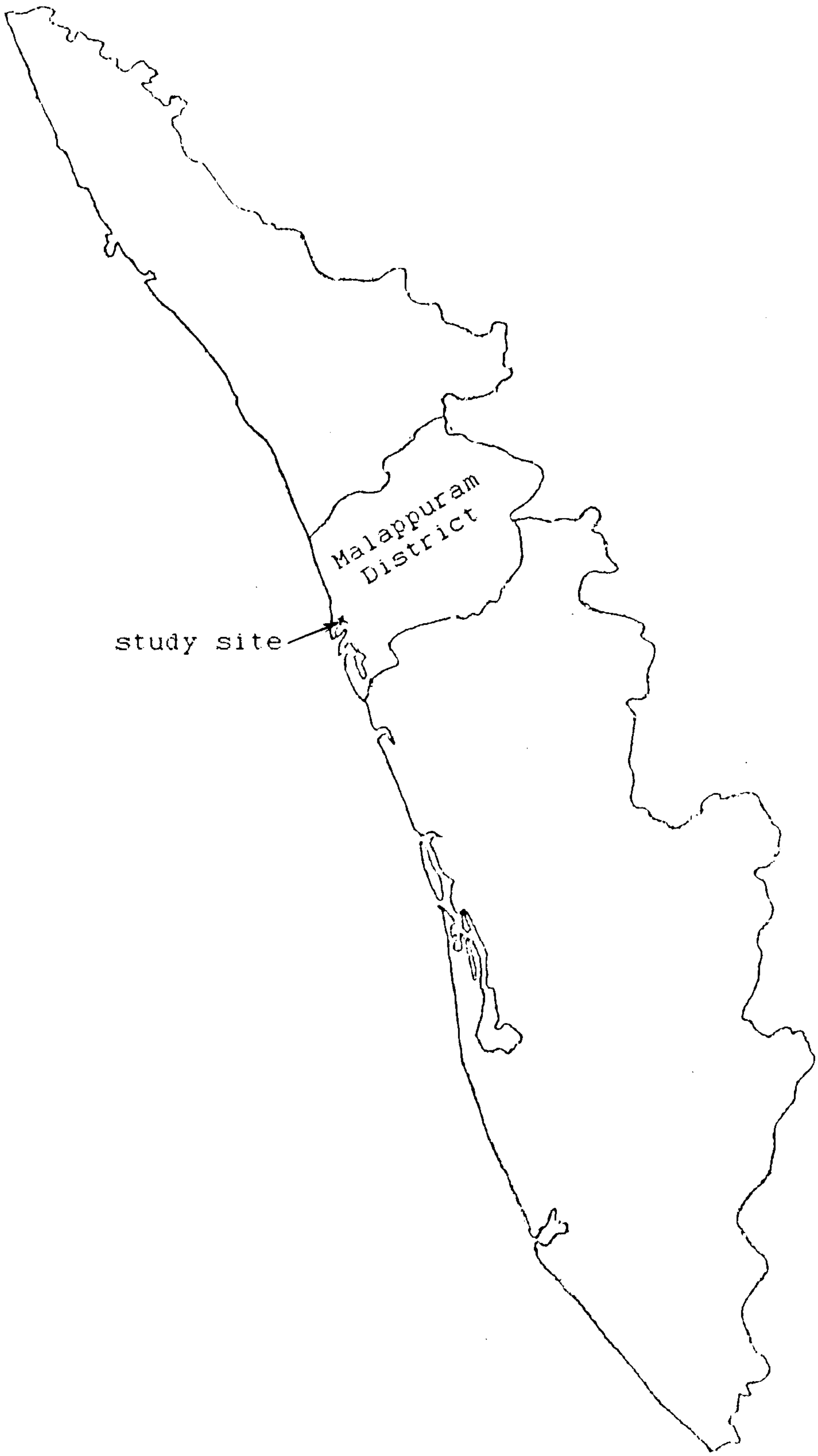

Fig.1 Location map of the study site

3.2 Geology

Geologically, the southern side of the study site is a hard rock laterite hill underlain by weathered rock and fresh rock of mostly gneisses and charnockites. The soil profile at the northern side is composed of sand, sandy clay, laterite, weathered rock and followed by hard crystalline rocks. The southern portion of the study site represents a ridge and the flow of groundwater is southwesterly towards river.

3.3 Climate

Kerala has a humid, tropical climate with temperature averaging between 20 and 32°C throughout the year. The mean annual precipitation varies between 2000 and 4000mm and is distributed over about 125 rainy days. Kerala is situated within the monsoon zone and is exposed to strong seasonal weather contrasts. One can differentiate between a `hot weather period' from March to May, a `southwest monsoon period' from June to September, a `northeast monsoon period' from October to December and a `winter season' in January and February. The southwest monsoon is the dominant rainy season.

Agroclimatically, the study area falls within the border line of Northern zone, central zone and kole zone of Kerala.

3.4 Topography

A topographical map of the study site is shown in Fig.2. The contours have been drawn at 2m interval with reduced level of an assumed bench mark arbitrarily chosen as 10m. The location of the selected wells are shown in the map.

3.5 Water Resources

The study site has a total of 10 dug wells and 5 ponds. Of these, 5 wells were selected for collecting the water table data. The water table contour along the northern side of the study site is assumed to be parallel to the river.

3.6 Methodology and estimation of parameters

3.6.1 Measurement of rainfall

The main source of water in the site is rainfall. The accurate measurement of rainfall is important and it was achieved with the use of non-recording and recording type raingauges. The non-recording type raingauge was used along with a recording type raingauge for cross checking. The total rainfall in a given period of time is measured by a non-recording type raingauge. The non-recording raingauge used is the Simon's raingauge, errected on a masonry

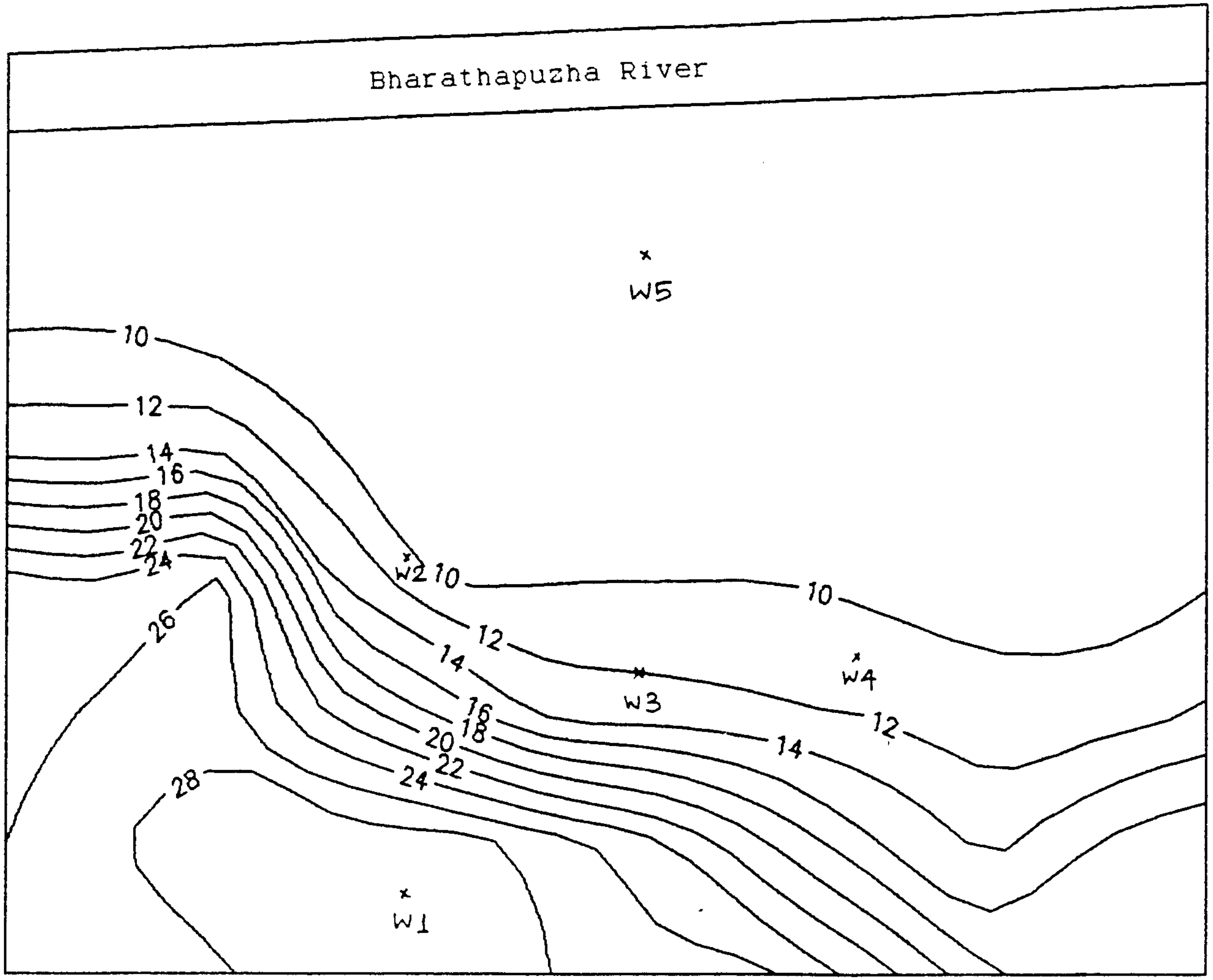

Fig.2 Topographical map of the study site

foundation with the rim 30.5cm above the ground. The measurement was done daily at 08.30 hours IST.

The natural siphon recording raingauge gives a continuous record of rainfall, its intensity and duration. The gauge is installed on a masonry foundation, with its rim 75cm above the ground, by the side of the ordinary raingauge (2m away from it) for checking and adjustment, if necessary. The rainfall data were collected everyday and the weekly values were calculated.

3.6.2 Estimation of evapotranspiration

The reference ET is calculated using a Computer package known as `CROPWAT' developed by FAO. In this package, Penman-Monteith method is used for the calculation. The input data required are the monthly maximum and minimum temperature, air humidity, sunshine hours and wind velocity. By inputing these data one by one, directly the ETo was obtained. The ETc is calculated as the product of ETo and the crop coefficient.

3.6.3 Measurement of infiltration

The infiltration characteristics of the study site were analysed using the double ring infiltrometer. The cylinders are made of 2mm rolled steel. The inner cylinder from which the infiltration measurements are taken is 30cm in diameter. The outer cylinder which is used to form the buffer pond to minimise the lateral spreading of water is 60cm in diameter. These are 25cm deep and is driven into the soil upto about 10cm for measurement. This is done by using a drop weight hammer after keeping a wooden plank on the top of the cylinder, to prevent damages to the edges of the cylinder.

The water level in the inner cylinder was read with a field type point gauge. Infiltration data is obtained from a cylinder infiltrometer by measuring the depth of the ponded water in the cylinder at frequent intervals in the beginning and this data provided the initial infiltration rate. The readings were taken until a constant value was obtained. These data correspond to the basic infiltration rate of the soil of the site. The buffer pond is filled with water immediately after filling the inner cylinder. Water level in the inner cylinder and the buffer pond are kept approximately the same.

Infiltration studies were conducted at different locations in the study site. At each location, the infiltration rate and accumulated infiltration were calculated. The average infiltration rate and average accumulated infiltration were also calculated. The infiltration equation for the average accumulated infiltration was formulated and the goodness of fit was evaluated.

3.6.4 Measurement of depth to water table

The water level fluctuation data were collected from 5 selected wells in the site. In one of the wells, the depth to water table was measured using a water level recorder and in the rest 4 wells, a weighted tape was used for the measurement. The water table data were measured from 26 May 1994 to 25 May 1995 and the weekly values were tabulated.

The recorder generally consists of a time element, a stage element and recording mechanism. The recording mechanism consists of a recording pen and the chart which is set on a metallic drum. The pen rests on the chart and gives a marking when actuated.

The rise of water in the well is proportionally transmitted to the drum with the help of a float, counter weight, float pulley and the float cable. The pen is moved at a uniform speed along the horizontal axis where as the drum moves in the vertical axis. The relative movements of the chart and the pen produce a curve called stage graph from which the change in water table is obtained.

3.7 Groundwater balance of the study area

The area is divided into grids in such a way that each selected well falls in seperate grids. The wells were numbered as W1, W2, W3, W4, and W5. W1 is located near the

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southern boundary which forms the ridge, W2, W3 & W4 are on the hill slope and W5 on the valley. The grid layout with selected well positions is shown in Fig.3. The groundwater balance in all these grids were considered for developing the model. The groundwater flow from the lowermost grids to the river is also considered. Generally, the groundwater balance in a grid is given by

$$T(i,j)_{t+1} - T(i,j)_t = R_t + V(i-1,j-1)_t - V(i,j)_t$$

$T(i,j)_{t+1} - T(i,j)_t$ = the change in groundwater storage during the period, t

R_t = recharge due to precipitation during the period, t

$V(i-1,j-1)_t$ = groundwater inflow from areas outside the grid, and

$V(i,j)_t$ = groundwater outflow to areas outside the grid

3.8 Model development

A model was developed for the prediction of water levels at selected points in the study site. The following steps were involved in the development of the model.

(i) Study of water table fluctuations

(ii) Formulation of the main structure of the model, and

(iii) Estimation of recharge percentage

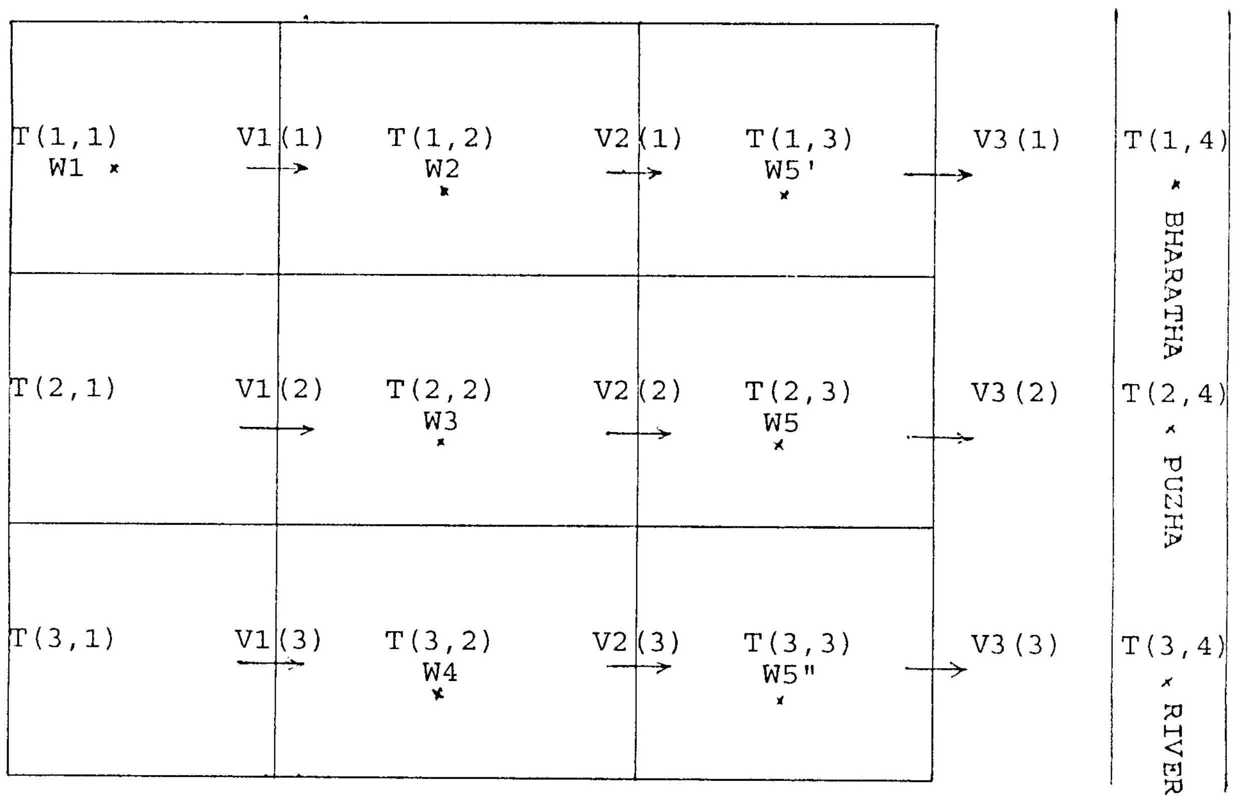


Fig.3 Grid showing the selected wells in the site

3.8.1 Study of water table fluctuations

The depletion in water table for a period of two months were studied. According to Darcy's law, the fluctuation is controlled by the elevation of the corresponding points, the distance between the points and the hydraulic conductivity.

The change in water table is also a function of outflow, inflow and storage assuming that the area of cross section of the column of aquifer releasing water remain constant and hydraulic conductivity at a particular depth remains constant. When we consider two wells at different elevations for a period without rainfall, the decrease in the water table at a point in the highest elevation is only due to the flow of water from this point to the next lower elevation. The change in storage at a point is the product of the change in water table in the well at that site and specific yield. The difference between the water table elevations in the two wells divided by the distance between the two wells gave the hydraulic gradient. At a particular depth when there is no rainfall and inflow, the change in storage divided by the hydraulic gradient is a constant of the following form.

$$S \times \Delta y \times Y \times X = K \times i \times Y \times b$$

$$\frac{S . \Delta y}{i} = \frac{K b}{X}$$

$$= \frac{T}{X}$$

$$= A$$

where

S	=	specific yield
K	=	hydraulic conductivity,m/week
T	=	transmissivity,m /week
b	=	saturated thickness of the aquifer contributing water to the well,m
Y	=	length of the grid,m
X	=	width of the grid,m
i	=	hydraulic gradient in the two wells
Δy	=	change in water table,m and
A	=	a constant

For a selected period, the weekly changes in storage and hydraulic gradient were calculated for different depths and the constants were determined. The constants for different locations were also calculated similarly, considering the inflow from grids lying above. These constants were used for the calculation of discharge from a grid during rainy season.

3.8.2 Model Structure

A Computer program was developed for the prediction of water table based on the input variables precipitation, evapotranspiration, initial levels in the wells, specific yield, elevation of the points under consideration and the distance between the points. The weekly precipitation, evapotranspiration and the above determined constants were fed through files and all the other data were directly fed through the key board.

From the measured rainfall depth and change in storage at a particular grid, the recharge and specific yield were estimated. The 50 per cent of the effective rainfall is estimated as the recharge. The storage depth can be obtained by dividing the recharge with the specific yield.

3.8.2.1 Calculation of groundwater flow

Two wells at different elevations and lying approximately along the same line were considered. Darcy's law was applied to determine the flow.

The flow was calculated as

$$v = \frac{K \times i \times Y \times b}{X \times Y}$$

$$= \frac{K\,b\,i}{X}$$

$$= \frac{T\,i}{X}$$

$$= A\,i$$

The groundwater flow is changed to flow depth by dividing it with the specific yield.

Therefore

$$v = \frac{A\,i}{S}$$

where

v = flow between two successive grids, m/week

i = hydraulic gradient between the two wells

A = the constant, and

S = specific yield

The flow for different sets of wells were calculated for the selected period.

3.8.2.2 Estimation of water table fluctuation

The initial depth to water table in all the selected wells were known. At each point there is a storage depth, a flow from higher elevation to the point and from the point to the lower elevation. The new water level at a particular point was obtained by subtracting the storage depth and inflow depth from the initial depth to the water table and adding the outflow depth to the same.

Mathematically,

$$S(i,j)_{t+1} = S(i,j)_t + V(i-1,j-1)_t - V(i,j)_t$$

$S(i,j)_t$ = groundwater storage at a particular point before a time interval, t

$V(i-1,j-1)_t$ = groundwater inflow during the time interval, t

$V(i,j)_t$ = groundwater outflow during the time interval, t and

$S(i,j)_{t+1}$ = groundwater storage after the time interval, t

The new level after a week was obtained, since the weekly changes were used. For the next week, these values were considered as the initial levels. Repeating the same procedure, the depths to water table for the selected points in the study area were estimated using the model for the study period.

3.8.3 Estimation of recharge percentage

The effective rainfall is obtained by subtracting ET from the rainfall, as groundwater is concerned. In order to find out the ground water recharge from the rainfall, the model has been simulated for different percentages of effective rainfall so that the error in the observed readings of the water table is minimised. Thus it was found that 50% of the effective rainfall can be taken as recharge.

3.9 Verification of the model

The depths to water table for the selected period were estimated using the model. The water table depth data for the same period were already observed. The observed and estimated data were compared for the verification of the model.

Results and Discussion

RESULTS AND DISCUSSION

This chapter highlights the preliminary test results, the developed model structure and the results of the model verification.

4.1 Precipitation

The cumulative annual rainfall during the study period (1994-95) was about 3100mm which is almost equal to the average annual rainfall and it is characterised by frequent occurrence of heavy rainfall and a few rainless days. This state prevailed from 31 May in 1994.

The weekly values of rainfall plotted is shown in Fig.4 and the data in tabulated form is given in Appendix-I. The maximum amount of weekly rainfall was 35.88cm during the last week of July 1994. The maximum intensity of rainfall was 50mm/h and it occured on 6 July 1994.

4.2 Evapotranspiration

The monthly evopotranspiration values are shown in Fig.5. The calculation of ET using `CROPWAT' and the weekly data are given in Appendix-II. ETc was found to be the maximum in March and the minimum in July.

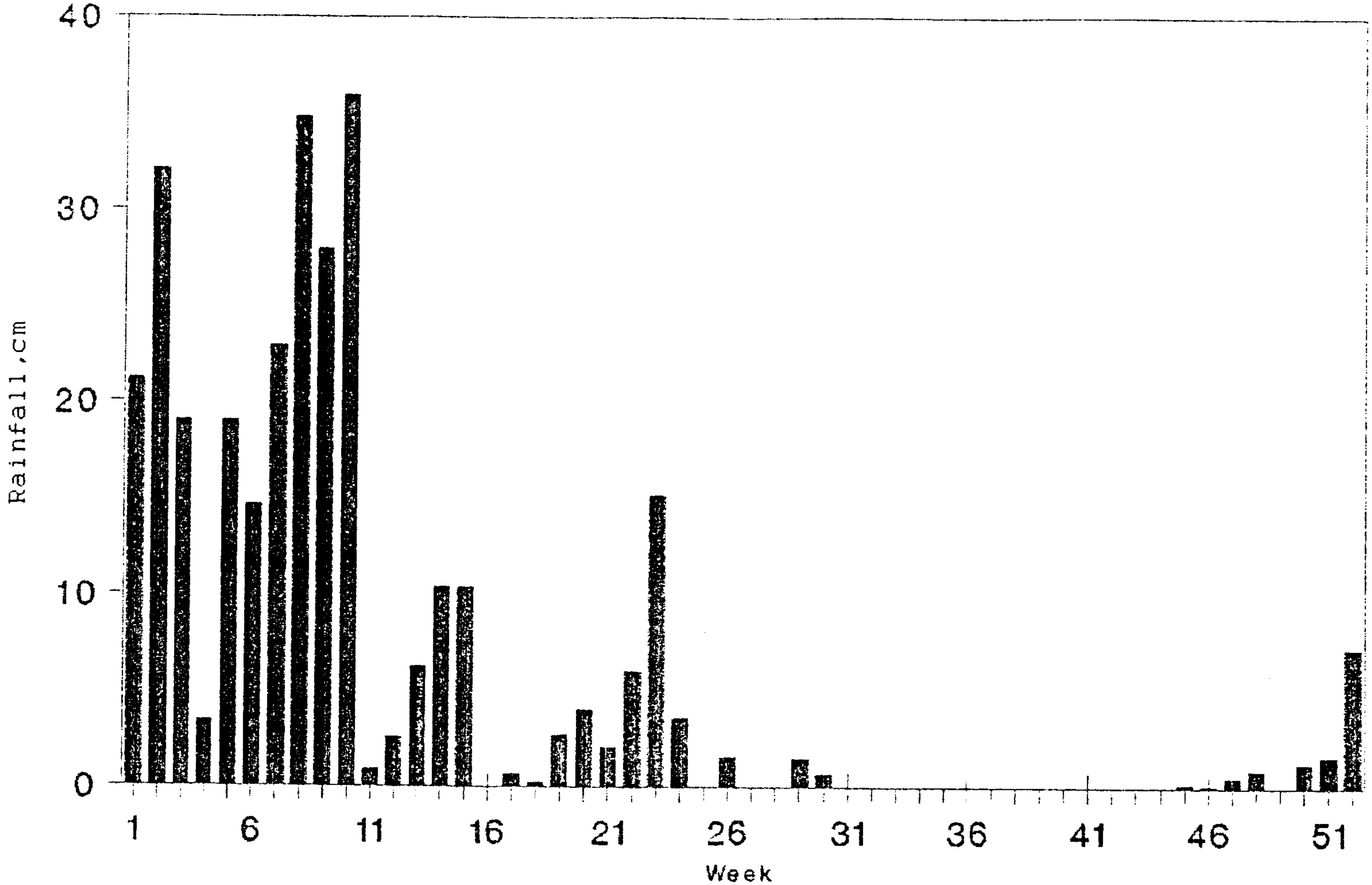


Fig. 4 Weekly values of rainfall from 26 May 1994 to 25 May 1995

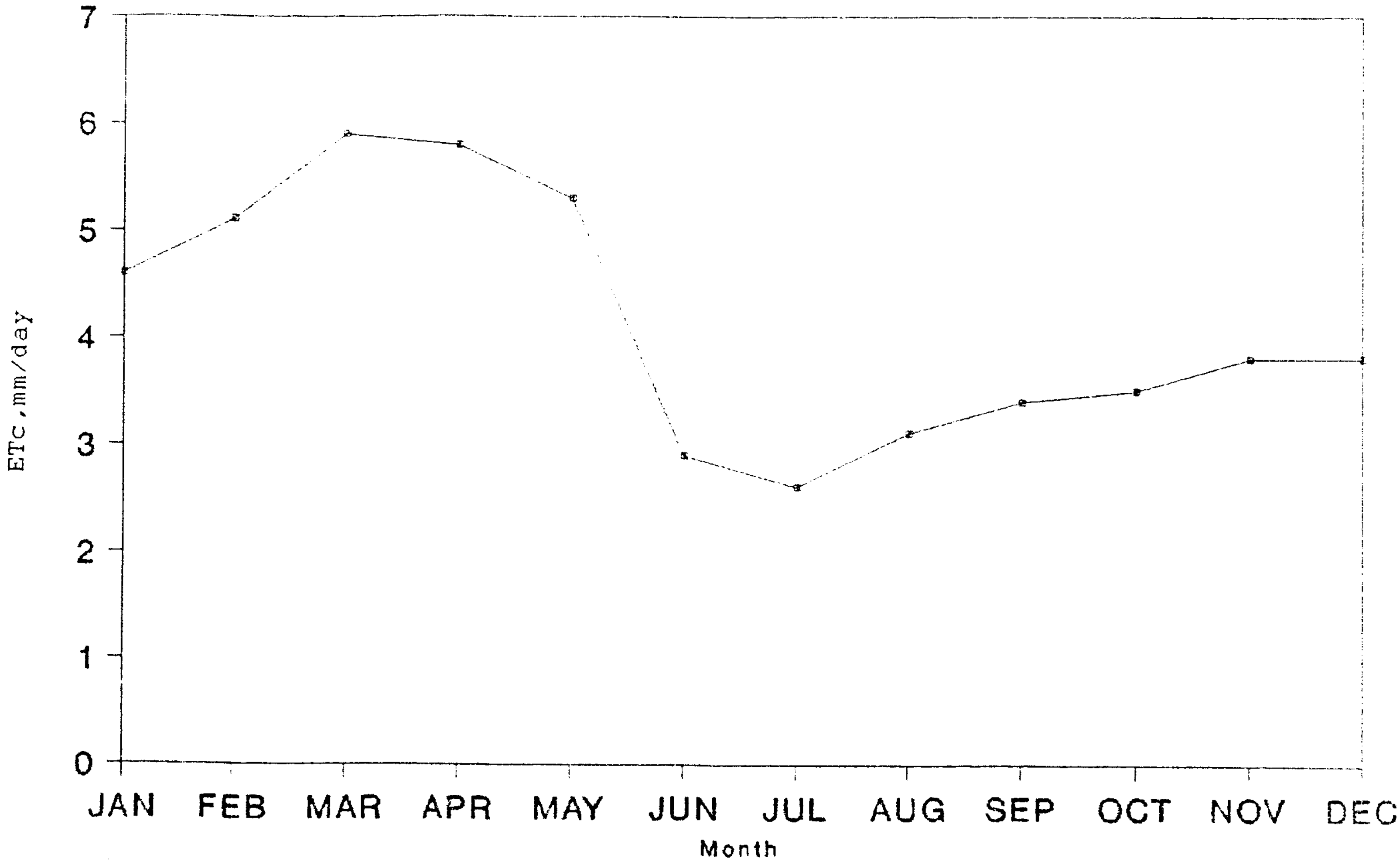


Fig. 6 Monthly evapotranspiration

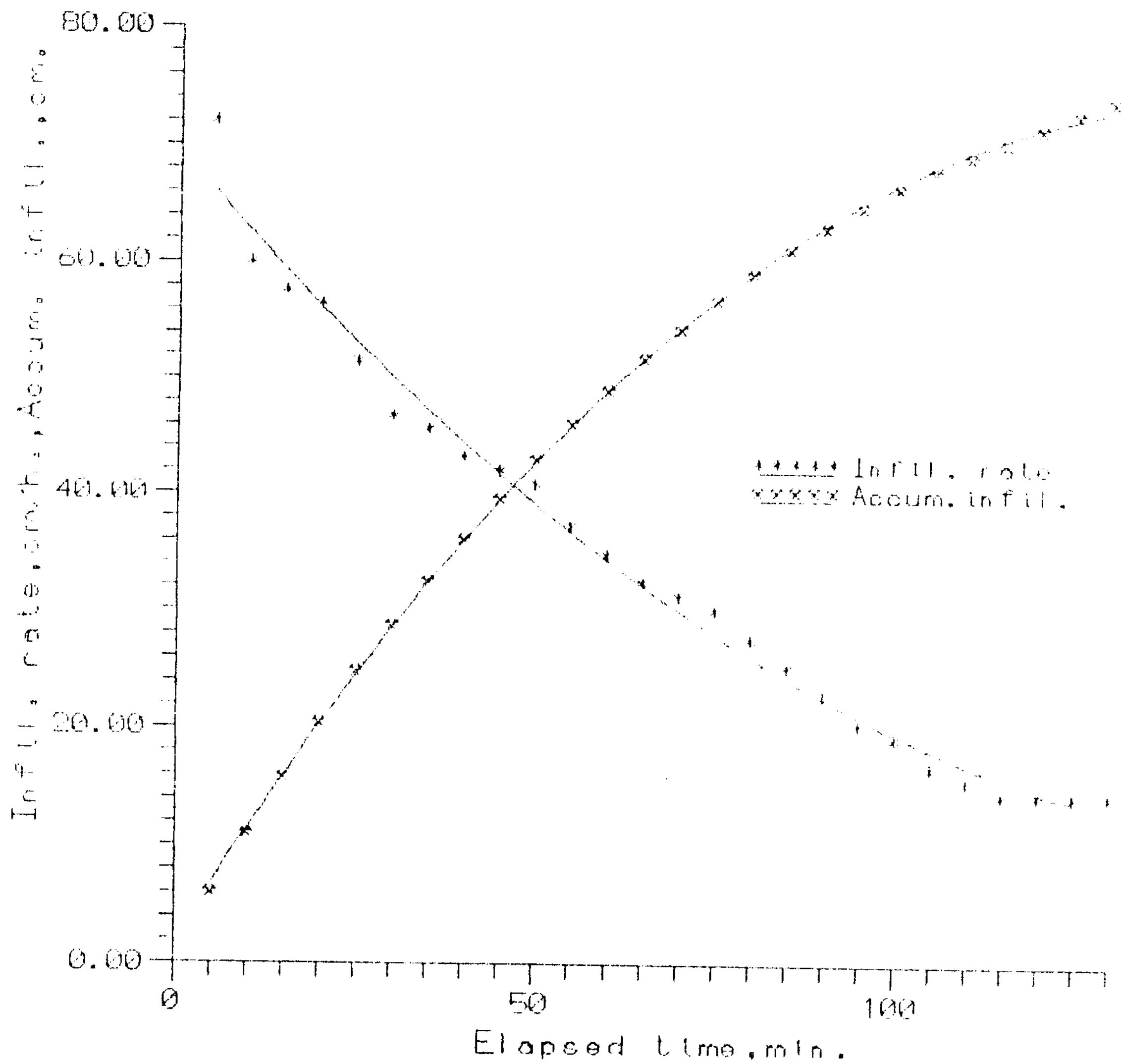


Fig.6 Infiltration characteristics at location I

4.3 Infiltration characteristics

The infiltration measurements were done at two representative locations in the study site. The infiltration rates and accumulated infiltrations at these locations are shown in Fig.6 and Fig.7 and the values are given in Appendix-III. The location I is on the lower portion of hillslope and the location II is on the upper portion. The infiltration rate at the lower portion was 14.4cm/h and at the upper hill was 16.8cm/h It was inferred from the figures that, the infiltration rate was more at the upper hill side than at the lower hill. Studies conducted by Varadan and Raghunath (1985) also concluded that the infiltration rate for latesols of Kerala are 12-20 cm/h and the infiltration rate increase towards higher elevation. The average infiltration rate and average accumulated infiltration at the study site are plotted against time and is shown in Fig.8 and the values are presented in Appendix-III. Comparing the infiltration rate and the maximum rainfall intensity during the study period, it is found that the infiltration rate is less than the rainfall intensity.

The functional relationship between the average accumulated infiltration (y) and the elapsed time (t) was derived and the equation so obtained was

$$Y = 5.66\ t^{0.559} - 7.39$$

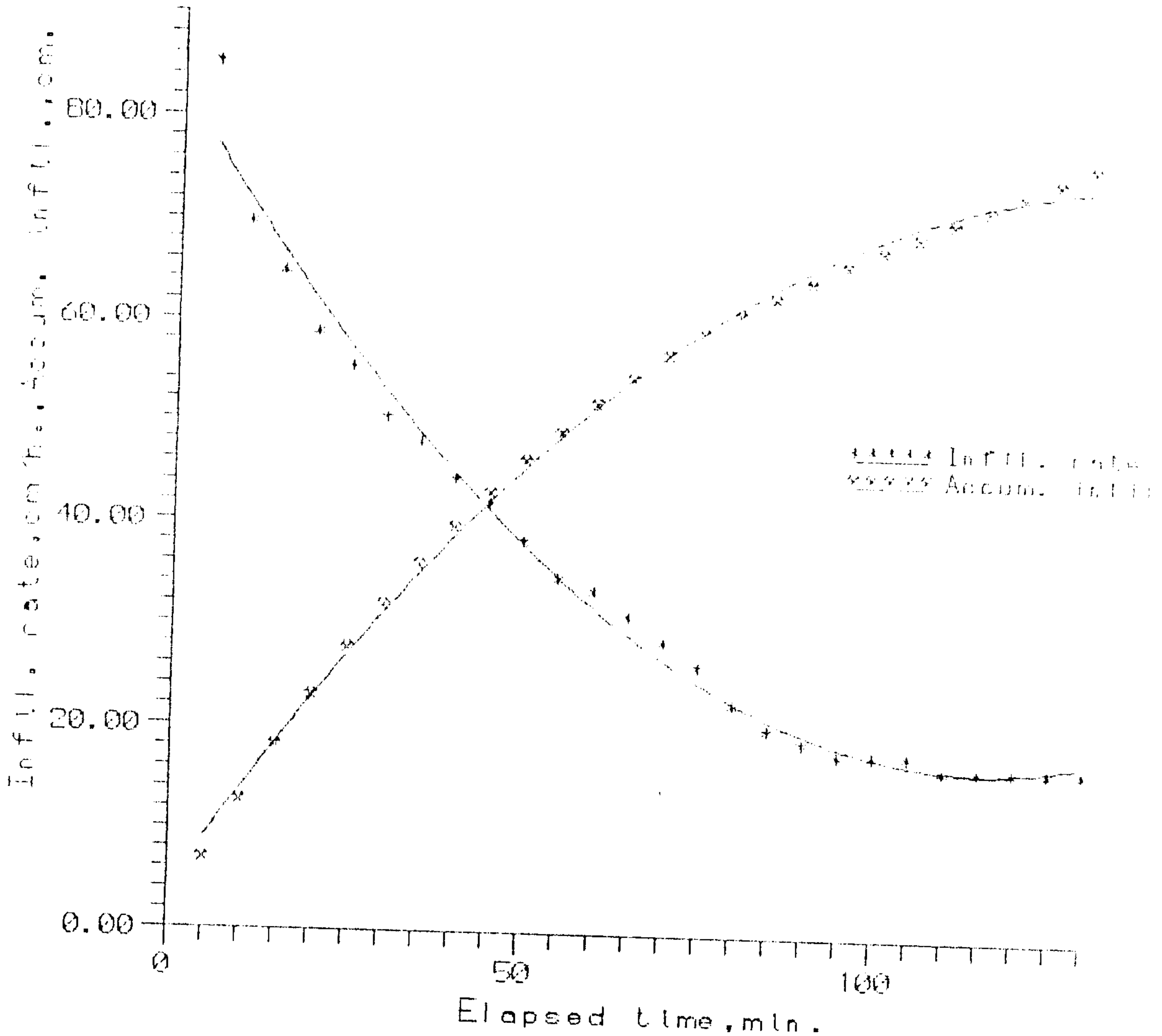


Fig.7 Infiltration characteristics
at location II

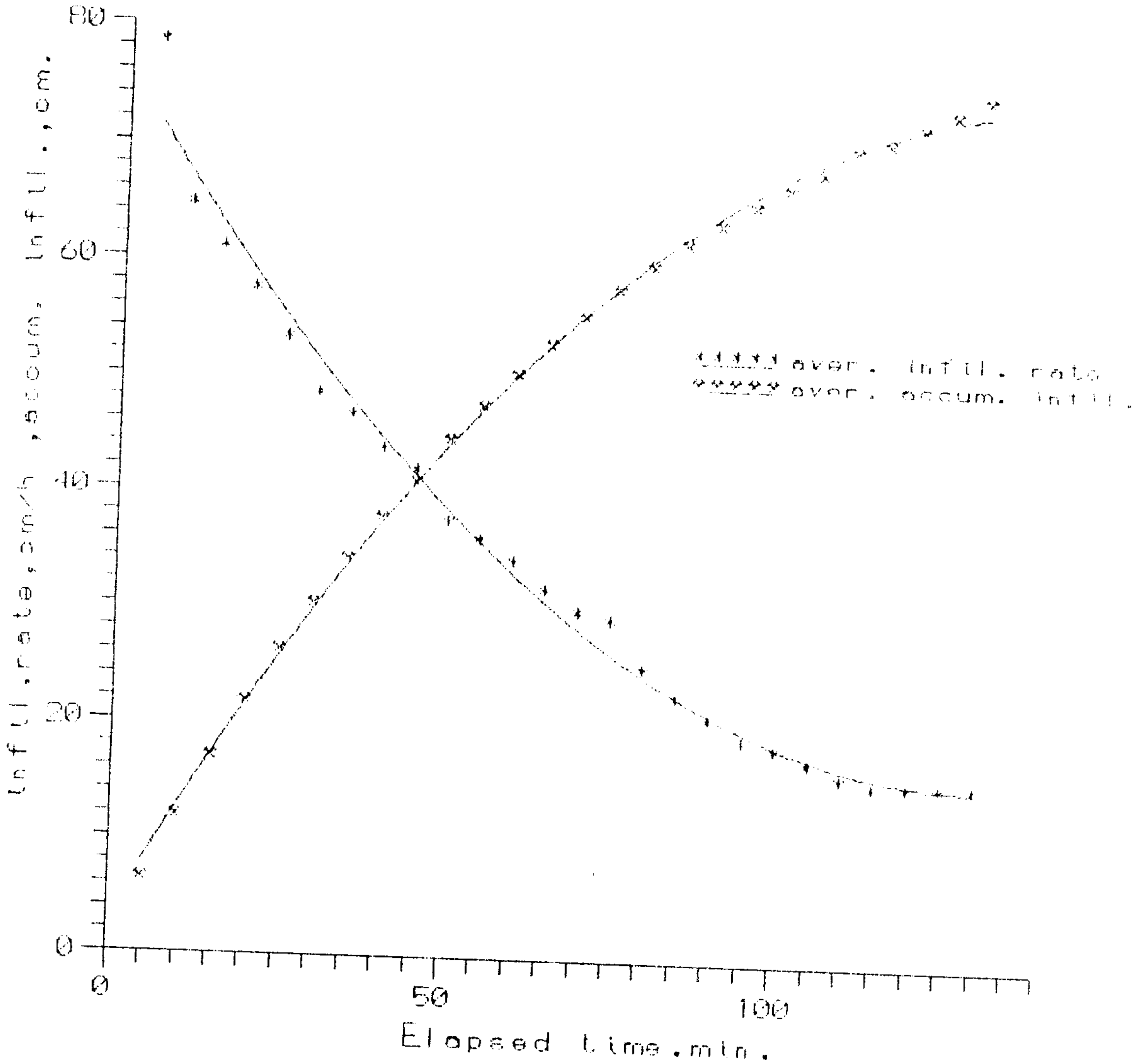


Fig.8 Average infiltration characteristics of the study site

The method of average used to determine the goodness of fit is presented in Appendix-IV. The goodness of fit was also evaluated.

4.4 Groundwater

The weekly observed depths to water table in all the selected wells in the site are given in Table 1. The observed weekly water table elevation of these wells are given in Fig.9 and the values are given in Appendix-V. It was observed that, the minimum depth to water table in wells W1, W4 and W5 were established in the last week of July and in W2 it was in third week of August and in W3 it occured in the first week of August. During the first week of the rainy season the lowest groundwater level for the season was established. Later in the season, intensive rainfall caused rapid responses. A thorough wetting of the soil was needed to bring about a moisture pulse moving from the ground surface to the saturated zone via, soil matrix which seems to account for the observed groundwater response in the present study. Due to the absence of rapid fluctuations in the water table during rainfall, the occurrence of preferred pathways and macropores in the soil matrix are neglected.

As the rain ceased, the profile dried out and the water table approached the pre-monsoon level. Data from recorded inter monsoon season showed no post-southwest monsoon groundwater response. The northeast monsoon was however, very weak during

Table 1. Weekly observed depth to water table in the selected wells in the study site

Week	Observed depth to WT in the wells, m W1	W2	W3	W4	W5
1	15.30	4.89	2.43	2.15	4.29
2	14.34	3.55	1.96	1.09	3.01
3	13.92	3.05	1.40	0.65	2.91
4	13.70	2.51	1.90	0.53	3.08
5	13.54	2.10	1.68	0.45	2.57
6	13.15	1.85	1.37	0.32	2.39
7	12.42	1.42	1.03	0.27	2.09
8	12.29	0.74	0.92	0.20	1.51
9	12.21	0.92	0.80	0.00	1.32
10	11.99	0.70	0.70	0.00	0.95
11	12.75	0.41	0.55	0.00	1.46
12	12.93	0.65	0.90	0.08	1.89
13	12.75	0.40	1.00	0.14	2.11
14	12.48	0.00	1.09	0.27	2.01
15	13.21	0.43	1.20	0.36	2.05
16	13.43	0.88	1.34	0.48	2.19
17	13.54	1.29	1.51	0.61	2.23
18	13.69	1.40	1.69	0.74	2.38
19	13.74	1.59	1.78	0.81	2.47
20	13.81	1.74	1.75	0.97	2.51

Table 1 (Contd.)

Week	Observed depth to WT in the wells, m W1	W2	W3	W4	W5
21	13.90	1.88	1.93	1.06	2.56
22	14.00	2.03	1.78	1.17	2.68
23	14.05	2.11	1.55	1.24	2.78
24	14.11	2.24	1.73	1.31	2.87
25	14.16	2.46	1.78	1.48	2.96
26	14.21	2.64	1.82	1.52	3.00
27	14.28	2.78	1.85	1.65	3.05
28	14.64	2.98	2.08	1.79	3.18
29	14.73	3.10	2.21	1.86	3.20
30	14.87	3.21	2.43	1.99	3.32
31	15.01	3.23	2.51	2.04	3.36
32	15.22	3.29	2.82	2.15	3.49
33	15.31	3.46	2.90	2.20	3.67
34	15.47	3.61	3.25	2.28	3.89
35	15.54	3.82	3.41	2.36	4.01
36	15.80	3.96	3.80	2.42	4.10
37	15.99	4.04	3.89	2.53	4.15
38	16.03	4.09	3.96	2.60	4.20
39	16.19	4.11	4.09	2.65	4.28
40	16.27	4.18	4.12	2.71	4.34
41	16.38	4.24	4.18	2.78	4.40

Table 1 (Contd.)

Week	Observed depth to WT in the wells, m				
	W1	W2	W3	W4	W5
42	16.51	4.31	4.22	2.84	4.47
43	16.60	4.42	4.31	2.89	4.58
44	16.72	4.63	4.40	2.95	4.71
45	16.80	4.71	4.54	3.03	4.79
46	16.86	4.80	4.63	3.14	4.86
47	16.91	4.91	4.79	3.25	4.95
48	17.03	4.99	4.85	3.34	5.04
49	17.21	5.09	5.00	3.43	5.09
50	17.30	5.13	5.08	3.54	5.11
51	17.51	5.20	5.11	3.60	5.16
52	17.43	5.12	5.03	3.54	5.08

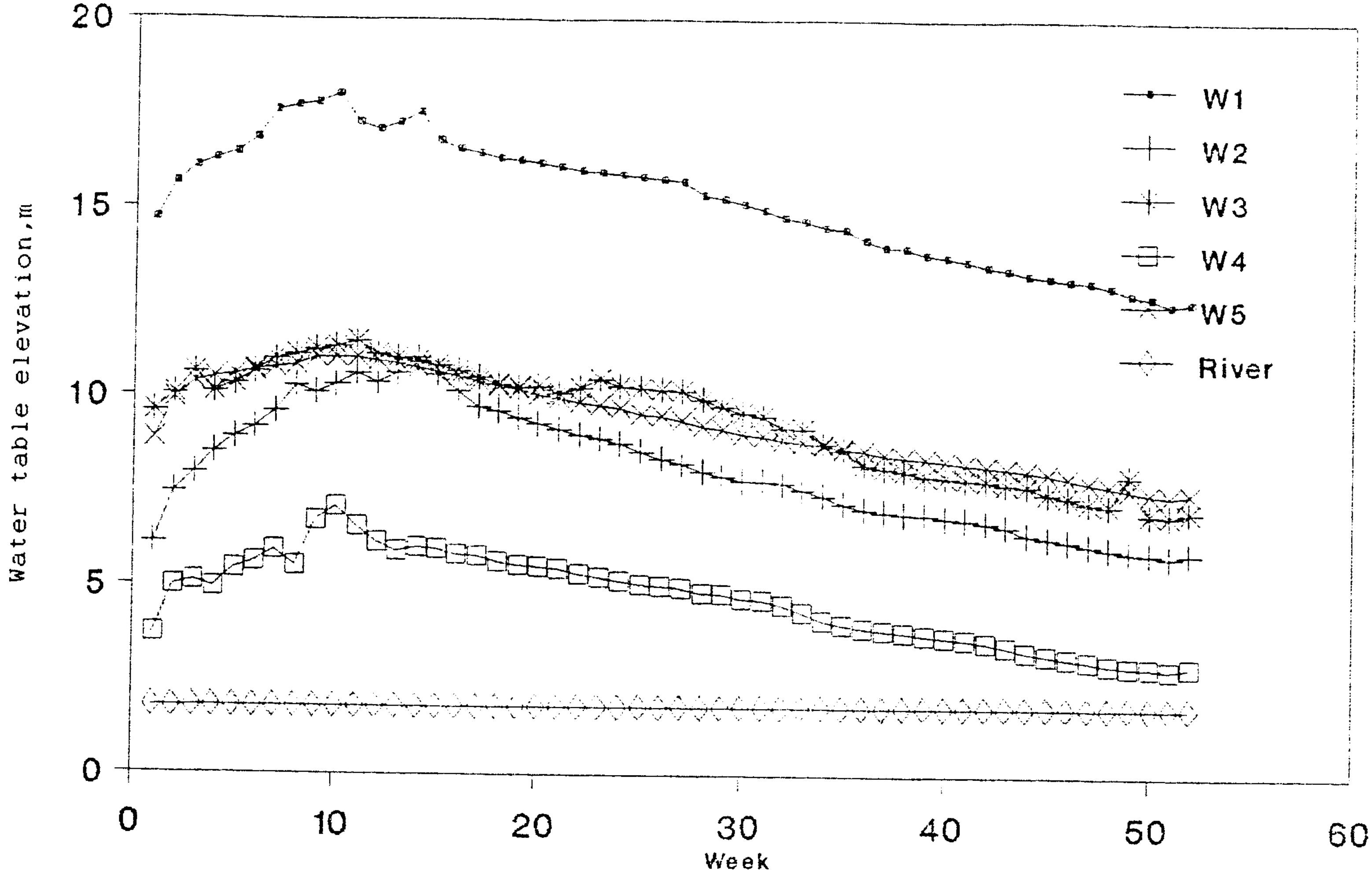


Fig.9 Weekly observed water levels in the selected locations

the study period. A thorough wetting of the profile, necessary to bring about changes in the groundwater level, required very intensive precipitation even though the profile was not as dry at the onset of the northeast monsoon as it was at the end of the dry winter and hot weather periods. Therefore, the groundwater recharge is a process concentrated chiefly in the southwest monsoon season and other occurrences are of a minor magnitude.

4.5 Study of recharge percentage

The effective rainfall is obtained by subtracting the ET from the rainfall. Weekly values of the effective rainfall is calculated and 50 per cent of it gives the recharge.

4.6 Model development

4.6.1 Evaluation of the existing flow pattern

A Computer program was developed for the evaluation of the flow pattern based on Darcy's equation. A constant, $S \times \Delta y / i = T/X = A$ has been calculated for different depths and grids and the flow chart for calculating the same is given in Fig.10. The program is listed in Appendix-VI. The depths for which the constants were estimated is given in Table 2 and the corresponding constants estimated using the program are given in Table 3. These constants were determined for a period without rainfall from 1 December 1994 to 16 February 1995 and

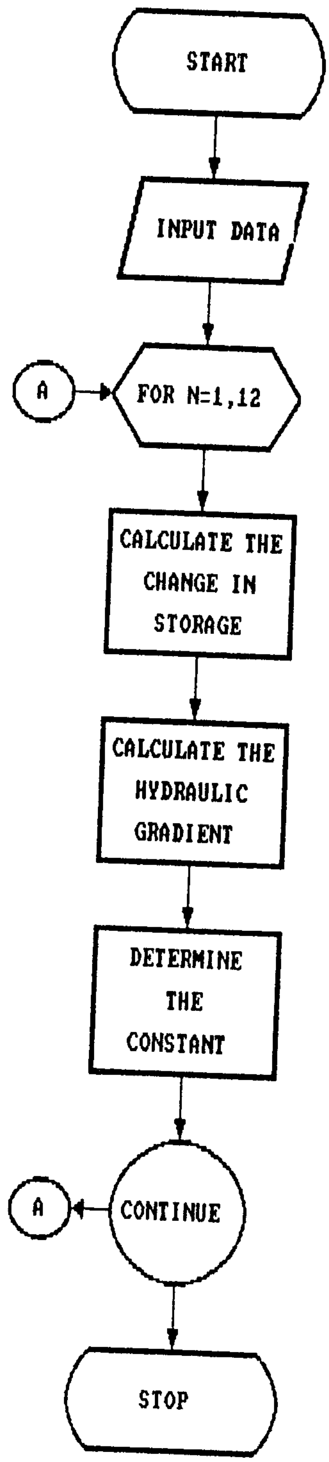


Fig.10 Flow chart of the program for determining the constant

Table 2. Depths to water table data for a period without rainfall

Weekly depth to water table, m				
W3	W5	W1	W2	W4
1.85	3.05	14.28	2.78	1.65
2.08	3.18	14.64	2.98	1.79
2.21	3.20	14.73	3.10	1.86
2.43	3.32	14.87	3.21	1.99
2.51	3.36	15.01	3.23	2.04
2.82	3.49	15.22	3.29	2.15
2.90	3.67	15.31	3.46	2.20
3.25	3.89	15.47	3.61	2.28
3.41	4.01	15.54	3.82	2.36
3.80	4.10	15.80	3.96	2.42
3.89	4.15	15.99	4.04	2.53
3.96	4.20	16.03	4.09	2.60

Table 3. $S.\Delta y/\hat{i}$ values for different depths and locations

W3	W1	W2	W4
2.92	1.85	3.36	2.17
1.68	0.47	2.06	1.09
2.91	0.73	1.95	2.04
1.68	0.73	0.35	0.79
4.22	1.12	1.05	1.74
1.13	0.49	2.92	0.79
4.84	0.86	2.57	1.22
2.28	0.38	3.52	1.18
5.60	1.38	2.41	0.88
1.38	1.02	1.40	1.60
1.08	0.22	0.88	1.03

these values were used for further calculation of the water level.

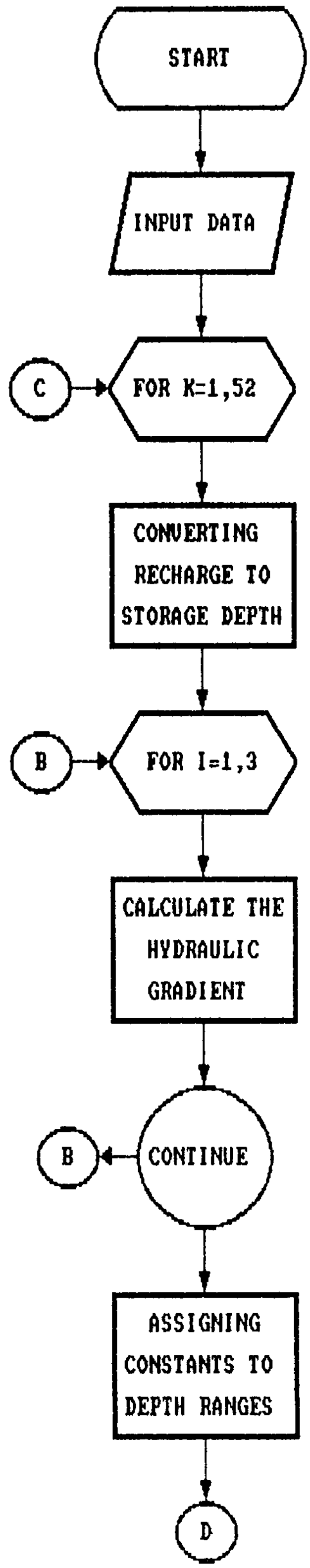
4.6.2 Main structure of the model

The flow chart presenting the main structure of the described model is given in Fig.11 and the program is listed in Appendix-VI. The depths to water table for the selected period were estimated using the model and it is given in Table 4. The constants used for different depth ranges and the values entering during the execution of the program are given in Appendix-VII.

4.6.3 Model verification

The model was verified by comparing the estimated values with the observed values for the same period. In Figures 12, 13, 14, 15 and 16 the observed and estimated depths to water table are plotted against the week for the selected wells. From the figures, it is inferred that, the estimated depths to water table is a reasonable estimate of the observed values for the same period.

The model is simulated for different percentages of effective rainfall. At 50 per cent it accounts as groundwater recharge and, the rest is assumed as surface runoff. Eventhough the infiltration rate is more, the amount of recharge is reduced because of the rock explosure in some part of the


START
INPUT DATA
C
FOR K=1,52
CONVERTING RECHARGE TO STORAGE DEPTH
B
FOR I=1,3
CALCULATE THE HYDRAULIC GRADIENT
B
CONTINUE
ASSIGNING CONSTANTS TO DEPTH RANGES
D

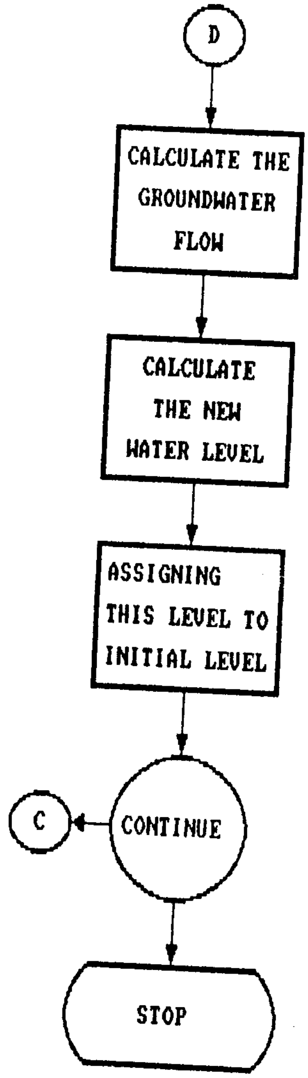


Fig.11 Flow chart presenting the main structure of the model

Table 4. Weekly estimated depth to water table in the selected wells in the study site

Week	Estimated depth to WT in the wells, m				
	W1	W2	W3	W4	W5
1	15.50	4.97	2.87	2.21	4.41
2	14.87	4.28	2.28	1.12	3.62
3	14.58	3.86	1.98	0.48	3.21
4	14.65	3.80	2.08	0.27	3.14
5	14.37	3.38	1.77	0.00	2.77
6	14.17	3.10	1.60	0.00	2.61
7	13.79	2.63	1.24	0.00	2.29
8	13.42	1.62	0.32	0.00	1.72
9	13.17	0.91	0.00	0.00	1.48
10	12.72	0.06	0.00	0.00	1.09
11	13.06	0.07	0.00	0.00	1.57
12	13.34	0.08	0.00	0.00	1.93
13	13.52	0.04	0.00	0.00	2.14
14	13.60	0.00	0.00	0.00	2.11
15	13.68	0.00	0.00	0.00	2.08
16	13.99	0.13	0.25	0.00	2.29
17	14.27	0.27	0.48	0.00	2.47
18	14.38	0.59	0.71	0.00	2.64
19	14.44	0.83	0.89	0.10	2.74
20	14.47	1.03	1.03	0.18	2.80

Table 4 (Contd.)

Week	W1	W2	W3	W4	W5
	Estimated depth to WT in the wells, m				
21	14.55	1.26	1.21	0.29	2.91
22	14.53	1.38	1.30	0.30	2.92
23	14.32	1.29	1.17	0.11	2.72
24	14.37	1.45	1.31	0.18	2.81
25	14.50	1.70	1.52	0.33	2.98
26	14.60	1.90	1.70	0.46	3.10
27	14.74	2.13	1.90	0.61	3.11
28	14.88	2.35	2.10	0.75	3.14
29	14.99	2.53	2.21	0.85	3.18
30	15.11	2.62	2.32	0.98	3.24
31	15.25	2.73	2.46	1.11	3.32
32	15.41	2.86	2.60	1.25	3.41
33	15.56	2.99	2.80	1.38	3.49
34	15.67	3.16	2.99	1.50	3.57
35	15.78	3.32	3.18	1.64	3.65
36	15.90	3.50	3.31	1.77	3.80
37	16.02	3.67	3.44	1.94	3.94
38	16.14	3.80	3.57	2.09	4.06
39	16.26	3.93	3.70	2.23	4.16
40	16.38	4.06	3.83	2.36	4.24
41	16.51	4.20	3.97	2.51	4.34

Table 4 (Contd.)

Week	Estimated depth to WT in the wells, m W1	W2	W3	W4	W5
42	16.64	4.34	4.11	2.64	4.44
43	16.78	4.48	4.23	2.78	4.55
44	16.91	4.62	4.35	2.92	4.65
45	17.03	4.72	4.46	3.06	4.74
46	17.16	4.83	4.58	3.17	4.83
47	17.28	4.92	4.68	3.27	4.91
48	17.39	5.00	4.78	3.36	4.98
49	17.51	5.10	4.89	3.47	5.05
50	17.61	5.17	4.97	3.55	5.10
51	17.69	5.23	5.04	3.62	5.13
52	17.65	5.16	4.98	3.56	5.04

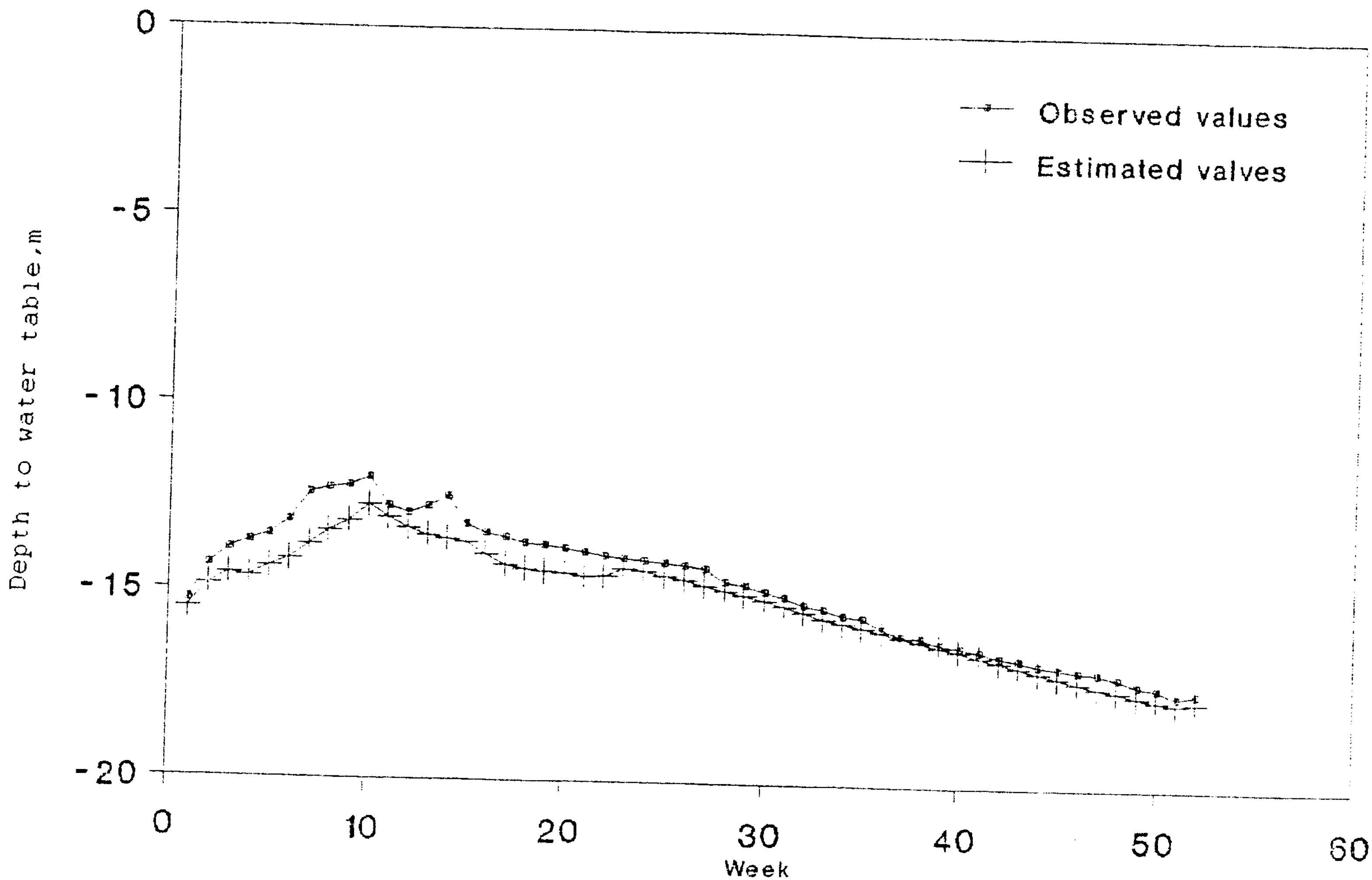


Fig.12 Observed and estimated weekly depths to water table in [illegible]

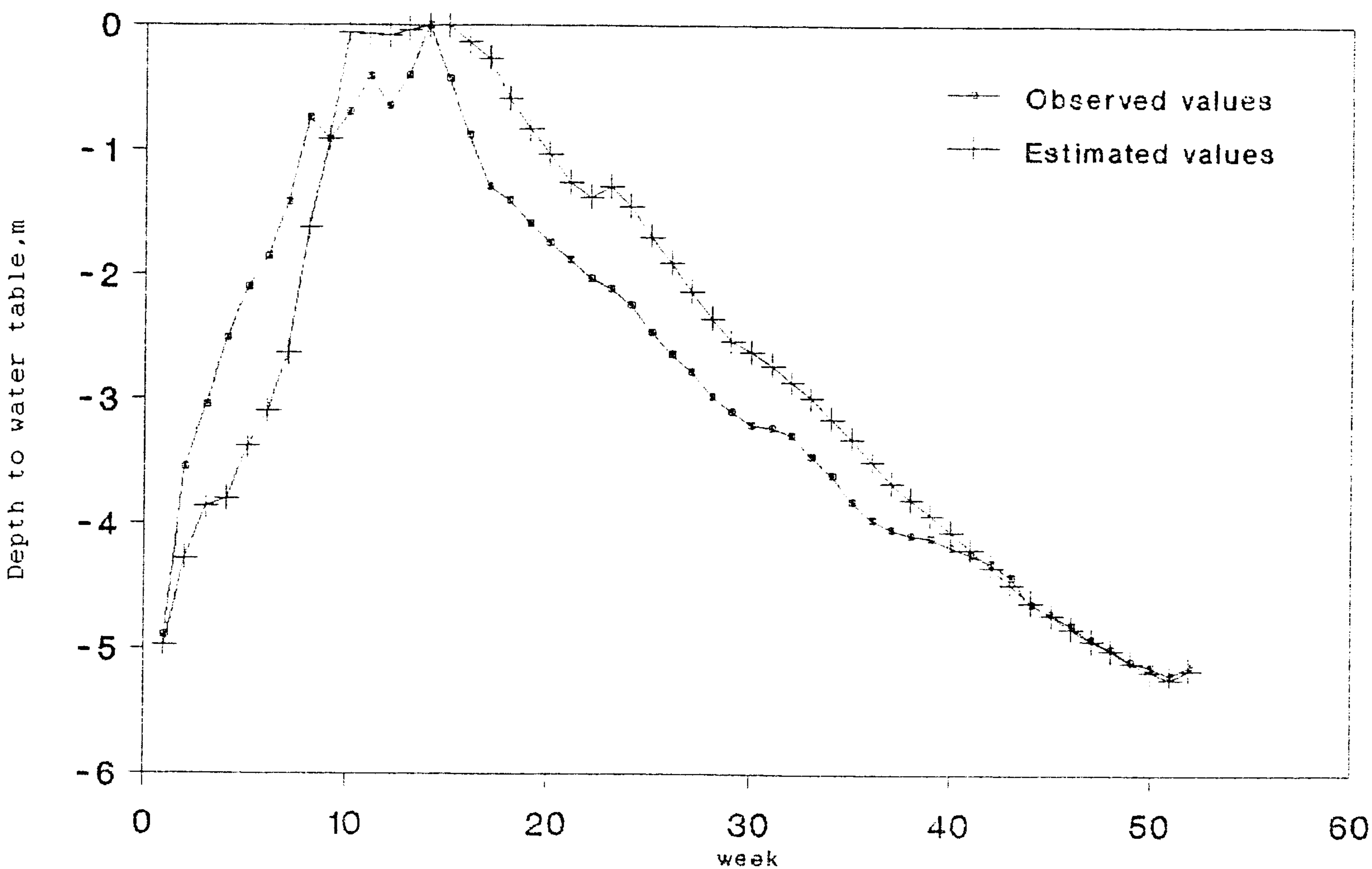


Fig.13 Observed and estimated weekly depths to water table in P 2

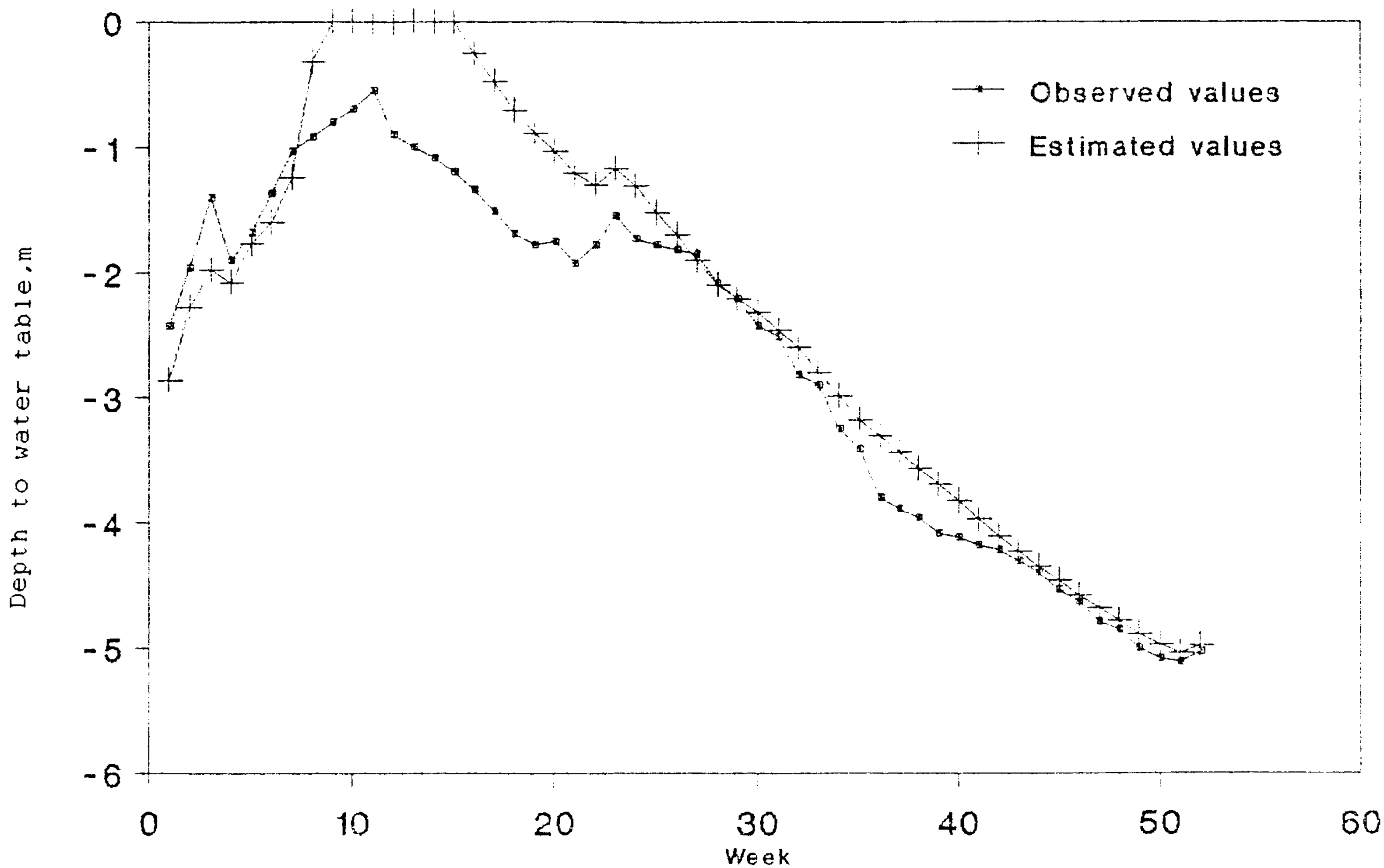


Fig.14 Observed and estimated weekly depths to water table in W3

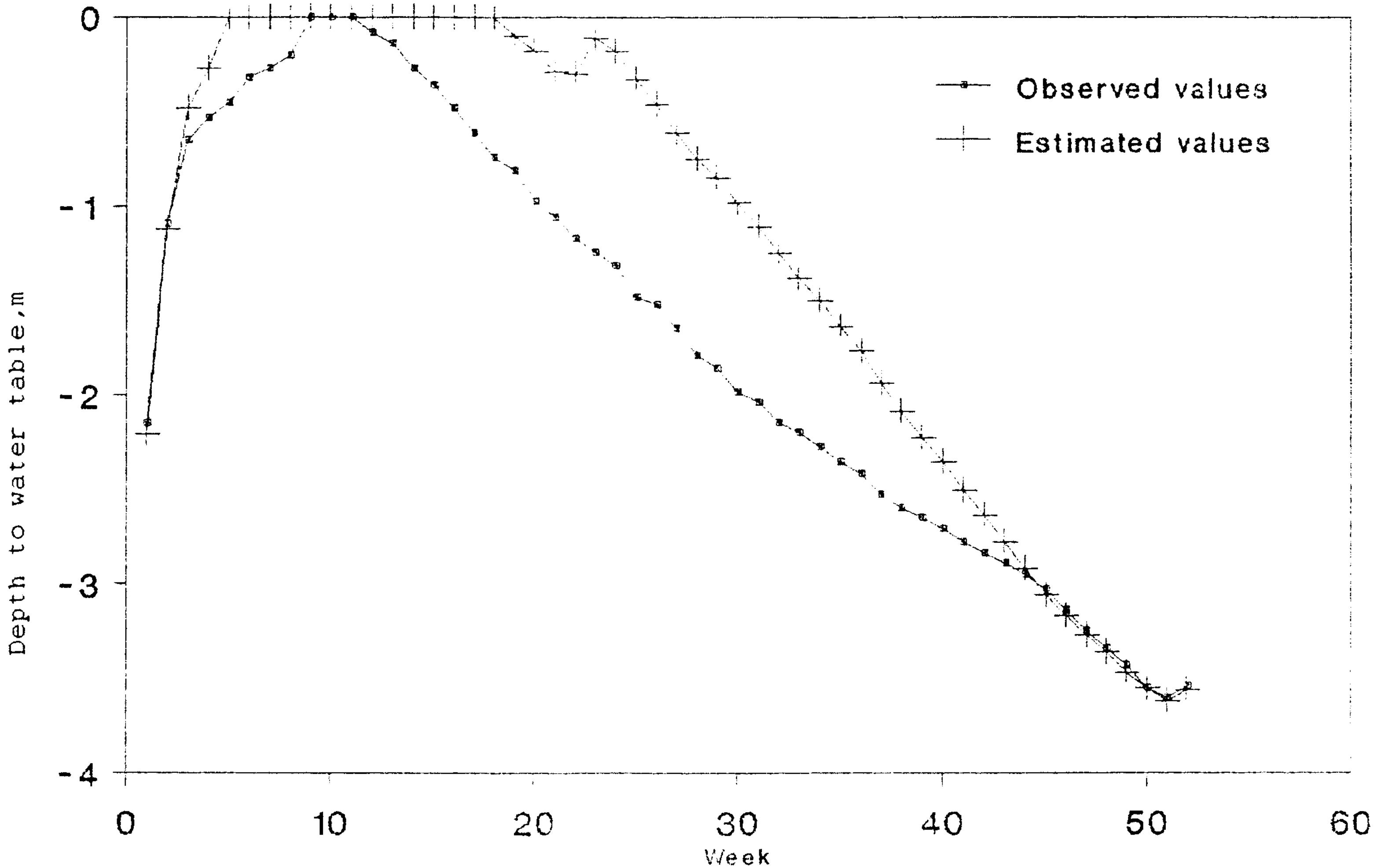


Fig.15 Observed and estimated weekly depths to water table in W4

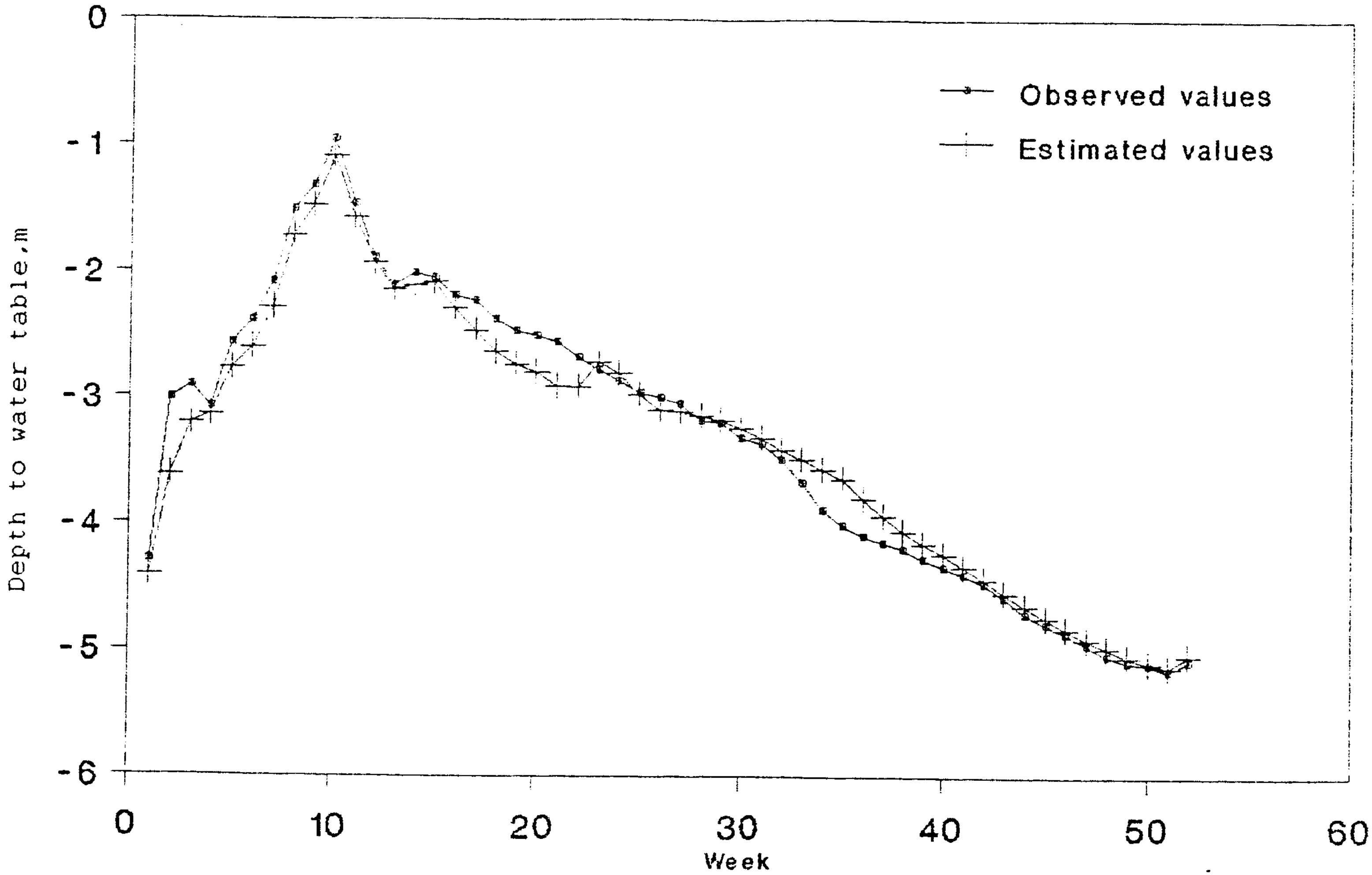


Fig.16 Observed and estimated weekly depths to water table in W5

study area. In the rocky area, the water is not infiltrated and it flows as surface runoff.

The same recharge percentage is assumed all over the study area. In flat land the recharge may be higher than that in the hill slope. This may cause a slight variation in the observed and estimated values. Another reason is that, in laterite soil, there are number of cavities in the soil profile. Through these pores, pipe flow may take place. But in the present study, pipe flow is not considered seperately.

Summary

SUMMARY AND CONCLUSION

Groundwater, compared to other water resources, is obtainable all the year around, and its use in conjunction with surface water is essential to maintain the water table within reasonable limits. Groundwater reservoir can also be used to store water in times of surplus availability.

For initiating a planned development of groundwater resources in a region, it is essential to obtain full information about the occurrence of groundwater in that region. Studies on groundwater movement in laterites are seldom. Therefore a study is conducted and a model has been developed for estimating the weekly changes in the water level in a humid lateritic region.

The infiltration characteristics of the study area is studied. It is found that the average infiltration rate at the upper hill was 16.8 cm/h and at the lower portion was 14.4 cm/h.

From the weekly values of rainfall and estimated evapotranspiration the effective rainfall was calculated. The surface runoff generation is also evaluated. It was found that 50 percent of the effective rainfall is recharged and the rest contributes to surface runoff.

The observed water table data from the wells in the site were analysed and found that the minimum depth to water table in wells W1, W4 and W5 were established in the last week of July and in W2 it was in third week of August and in W3 it occured in the first week of August. During the first week of the rainy season the lowest groundwater level for the season was established. Later in the season, intensive rainfall caused rapid responses. As the rain ceased, the profile dried out and the water table approached the pre-monsoon level. The northeast monsoon was very weak during the study period and the groundwater recharge was concentrated chiefly in the southwest monsoon season.

The model estimating water table at a purticular location and time based on the weekly precipitation, evapotranspiration, specific yield, initial levels, elevation of the points under consideration and the distance between the points is prepared.

The water table fluctuation data were taken and the existing flow pattern was evaluated for different location in the study site. The selected area was divided into grids and the saturated flow from upper grids to the lower grids were calculated.

Based on the groundwater balance of the study area, the model was developed and using this the depth to water table for a period from 26 May 1994 to 25 May 1995 were estimated. These values were compared with the observed values from the well in the site for the same period. It was found that the estimatd values gives a reasonable estimate of the depth to water table in the study area. Therefore this model can be used to predict the depth to water table in an area.

This model helps to predict the water table in the hill slope at any point. From this the amount of water that can be exploited from that location can be found out. The outflow from the hillslope continues throughout the year and by knowing the rate of discharge from the hill slope we can utilise this water efficiently.

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* Orginals not seen.

Appendices

Appendix - I

Weekly rainfall from 26 May 1994 to 25 May 1995(P.DAT)

Week	Rainfall,cm.	Week	Rainfall,cm.
1	21.18	27	0.00
2	32.10	28	0.00
3	19.00	29	1.50
4	3.39	30	0.70
5	18.99	31	0.00
6	14.63	32	0.00
7	22.88	33	0.00
8	34.80	34	0.00
9	27.93	35	0.00
10	35.88	36	0.00
11	0.90	37	0.00
12	2.58	38	0.00
13	6.22	39	0.00
14	10.36	40	0.00
15	10.34	41	0.00
16	0.00	42	0.00
17	0.68	43	0.00
18	0.26	44	0.00
19	2.73	45	0.20
20	4.04	46	0.10

Week	Rainfall, cm.	Week	Rainfall, cm.
21	2.08	47	0.50
22	6.00	48	0.90
23	15.12	49	0.00
24	3.60	50	1.26
25	0.00	51	1.60
26	1.58	52	7.20

Appendix - II

Calculation of evapotranspiration

Mon.	Temp. Max. °C	Temp. Min. °C	Relative humidity %	Wind velocity km./day	Sunshine hours	ETo mm./day	Kc	ETc mm./day
JAN	33.6	20.5	73	226	9.4	5.2	0.89	4.6
FEB	35.0	21.2	80	262	9.6	5.7	0.89	5.1
MAR	36.1	23.2	70	290	9.3	6.6	0.89	5.9
APR	35.7	24.9	70	283	8.7	6.5	0.95	5.8
MAY	33.7	24.4	75	307	7.0	5.6	0.95	5.3
JUN	30.1	23.7	94	223	3.3	3.1	0.95	2.9
JUL	28.8	22.6	95	221	2.8	2.8	0.94	2.6
AUG	29.1	22.7	92	199	4.3	3.3	0.93	3.1
SEP	30.6	22.9	91	214	5.8	3.8	0.89	3.4
OCT	31.4	22.7	90	199	6.4	3.9	0.89	3.5
NOV	32.2	21.9	85	187	7.6	4.1	0.94	3.8
DEC	32.7	20.9	85	190	8.5	4.2	0.91	3.8

Appendix - II(contd.)

Weekly ETc from 26 May 1994 to 25 May 1995(E.DAT)

Week	ETc, cm.	Week	ETc, cm.
1	3.71	27	2.66
2	2.03	28	2.66
3	2.03	29	2.66
4	2.03	30	2.66
5	2.03	31	2.66
6	1.82	32	3.22
7	1.82	33	3.22
8	1.82	34	3.22
9	1.82	35	3.22
10	1.82	36	3.57
11	2.17	37	3.57
12	2.17	38	3.57
13	2.17	39	3.57
14	2.17	40	3.57
15	2.38	41	4.13
16	2.38	42	4.13
17	2.38	43	4.13
18	2.38	44	4.13
19	2.45	45	4.06
20	2.45	46	4.06
21	2.45	47	4.06
22	2.45	48	4.06
23	2.45	49	3.71
24	2.66	50	3.71
25	2.66	51	3.71
26	2.66	52	3.71

Appendix - III

Infiltration characteristics of the study site

	Location I			Location II				
Elapsed time t min.	Depth of water infted. cm	Infil. rate cm/h	Accum. infil. cm	Depth of water infted. cm	Infil. rate cm/h	Accum. infil. cm	Aver. infil. rate cm/h	Aver. accum infil. cm
5	6.0	72.0	6.0	7.1	85.2	7.1	78.6	6.55
10	5.0	60.0	11.0	5.8	69.6	12.9	64.8	11.95
15	4.8	57.6	15.8	5.4	64.8	18.3	61.1	17.05
20	4.7	56.4	20.5	4.9	58.8	23.2	57.6	21.85
25	4.3	51.4	24.8	4.6	55.4	27.8	53.4	26.30
30	3.9	46.8	28.7	4.2	50.4	32.0	48.6	30.35
35	3.8	45.6	32.5	4.0	48.0	36.0	46.8	34.25
40	3.6	43.2	36.1	3.7	44.4	39.7	43.8	37.90
45	3.5	42.0	39.6	3.5	42.0	43.2	42.0	41.00
50	3.4	40.8	43.0	3.2	38.4	46.4	37.6	44.70
55	3.1	37.2	46.1	2.9	34.8	49.3	36.0	47.70
60	2.9	34.8	49.0	2.8	33.6	52.1	34.2	50.55
65	2.7	32.4	51.7	2.6	31.2	54.7	31.8	53.20
70	2.6	31.2	54.3	2.4	28.8	57.1	30.0	55.70
75	2.5	30.0	56.8	2.2	26.4	59.3	29.4	58.05
80	2.3	27.6	59.1	1.9	22.8	61.2	25.2	60.15
85	2.1	25.2	61.2	1.7	20.4	62.9	22.8	62.05
90	1.9	22.8	63.1	1.6	19.2	64.5	21.0	63.80

Appendix-III (Contd.)

	Location I			Location II				
Elapsed time t min.	Depth of water infted. cm	Infil. rate cm/h	Accum. infil. cm	Depth of water infted. cm	Infil. rate cm/h	Accum. infil. cm	Aver. infil. rate cm/h	Aver. accum infil cm
95	1.7	20.4	64.8	1.6	18.0	66.1	19.2	65.45
100	1.6	19.2	66.4	1.5	18.0	67.6	18.6	67.00
105	1.4	16.8	67.8	1.5	18.0	69.1	17.4	68.45
110	1.3	15.6	69.1	1.4	16.8	70.5	16.2	69.80
115	1.2	13.4	70.3	1.4	16.8	71.9	15.6	71.10
120	1.2	14.4	71.5	1.4	16.8	73.3	15.6	72.40
125	1.2	14.4	72.7	1.4	16.8	74.7	15.6	73.70
130	1.2	14.4	73.9	1.4	16.8	76.1	15.6	75.00

Appendix - IV

Method of averages to obtain infiltration parameters of the soil at the site.

For $t1 = 5$ min. $y1 = 6.55$ cm

For $t2 = 130$ min. $y2 = 75$ cm

$t3 = \sqrt{t1 \times t2} = \sqrt{5 \times 130} = 25.5$ min.

Corresponding value of y3 from graph = 26.5 cm

$$b = \frac{y1y2 - y3^2}{y1+y2-2y3} = -7.39$$

Evaluation of goodness of fit:

Sl. no.	Time t min.	observed accum. infil., y	y-b	log(y-b)	logt	y calculated	Deviation
1	5	6.55	13.94	1.144	0.699	6.542	0.112
2	10	11.95	19.34	1.286	1.000	12.900	-0.086
3	15	17.05	24.44	1.388	1.176	18.137	-0.064
4	20	21.85	29.24	1.476	1.301	22.602	-0.034
5	25	26.30	33.69	1.528	1.398	26.573	-0.010
6	30	30.35	37.74	1.577	1.477	30.280	0.002
7	35	34.25	41.64	1.619	1.544	33.630	0.018
8	40	37.90	45.29	1.656	1.602	36.869	0.027
9	45	41.00	48.39	1.685	1.653	39.816	0.029
10	50	44.70	52.09	1.717	1.699	42.729	0.044
11	55	47.70	55.09	1.741	1.740	45.455	0.047

Sl. no.	Time t min.	observed accum. infil., y	y-b	log(y-b)	logt	y calculated	Deviation
12	60	50.55	57.94	1.763	1.788	48.073	0.049
13	65	53.20	60.59	1.782	1.813	50.553	0.050
14	70	55.70	63.09	1.800	1.845	53.005	0.048
15	75	58.05	65.44	1.816	1.875	55.416	0.045
16	80	60.15	67.54	1.830	1.903	57.773	0.040
17	85	62.05	69.44	1.842	1.929	59.908	0.035
18	90	63.80	71.19	1.852	1.954	62.112	0.026
19	95	65.45	72.84	1.862	1.978	64.389	0.016
20	100	67.00	74.39	1.872	2.000	66.400	0.009
21	105	68.45	75.84	1.880	2.021	68.468	0.000
22	110	70.50	77.89	1.891	2.041	70.414	0.001
23	115	71.10	78.49	1.895	2.061	72.409	-0.018
24	120	72.40	79.79	1.902	2.079	74.268	-0.026
25	125	73.70	81.09	1.909	2.097	76.170	-0.034
26	130	75.00	82.39	1.906	2.114	78.117	-0.042

Appendix - V

Weekly observed water table elevation in the selected wells in the study site

Week	Observed water table elevation, m				
	W1	W2	W3	W4	W5
1	14.70	6.11	9.57	8.85	3.71
2	15.66	7.45	10.04	9.91	4.99
3	16.08	7.95	10.60	10.35	5.09
4	16.30	8.49	10.10	10.47	4.92
5	16.46	8.90	10.32	10.55	5.43
6	16.85	9.15	10.63	10.68	5.61
7	17.58	9.58	10.97	10.73	5.91
8	17.71	10.26	11.08	10.80	6.49
9	17.79	10.08	11.20	11.00	6.68
10	18.01	10.30	11.30	11.00	7.05
11	17.25	10.59	11.45	11.00	6.54
12	17.07	10.35	11.10	10.92	6.11
13	17.25	10.60	11.00	10.86	5.89
14	17.52	11.00	10.91	10.73	5.99
15	16.79	10.57	10.80	10.64	5.95
16	16.57	10.12	10.66	10.52	5.81
17	16.46	9.71	10.49	10.39	5.77
18	16.31	9.60	10.31	10.26	5.62
19	16.26	9.41	10.22	10.19	5.53

Appendix-V (Contd.)

Week	Observed water table elevation, m				
	W1	W2	W3	W4	W5
20	16.19	9.26	10.25	10.03	5.49
21	16.10	9.12	10.07	9.94	5.44
22	16.00	8.97	10.22	9.83	5.32
23	15.95	8.89	10.45	9.76	5.22
24	15.89	8.76	10.27	9.69	5.13
25	15.84	8.54	10.22	9.52	5.04
26	15.79	8.36	10.18	9.48	5.00
27	15.72	8.22	10.15	9.35	4.95
28	15.36	8.02	9.92	9.21	4.82
29	15.27	7.90	9.75	9.14	4.80
30	15.13	7.79	9.57	9.01	4.68
31	14.99	7.77	9.49	8.96	4.64
32	14.78	7.71	9.18	8.85	4.51
33	14.69	7.54	9.10	8.80	4.33
34	14.53	7.39	8.75	8.72	4.11
35	14.46	7.18	8.59	8.64	3.99
36	14.20	7.02	8.20	8.58	3.90
37	14.01	6.96	8.11	8.47	3.85
38	13.97	6.91	8.04	8.40	3.80
39	13.81	6.89	7.91	8.35	3.72
40	13.73	6.82	7.88	8.29	3.66

Appendix-V (Contd.)

Week	Observed water table elevation, m W1	W2	W3	W4	W5
41	13.62	6.76	7.82	8.22	3.60
42	13.49	6.69	7.78	8.16	3.53
43	13.40	6.58	7.69	8.11	3.42
44	13.28	6.37	7.60	8.05	3.29
45	13.20	6.29	7.46	7.97	3.21
46	13.14	6.20	7.37	7.86	3.14
47	13.09	6.09	7.21	7.75	3.05
48	12.97	6.01	7.15	7.66	2.96
49	12.79	5.91	7.00	7.57	2.91
50	12.70	5.87	6.92	7.46	2.89
51	12.49	5.80	6.89	7.40	2.84
52	12.57	5.88	6.97	7.46	2.92

Appendix - VI

```
C     PROGRAM FOR THE DETERMINATION OF THE CONSTANT
      DIMENSION C(12,5),O1(12),O2(12),O3(12),O4(12)
C     S = SPECIFIC YIELD
C     L1 = ELEVATION OF W3
C     L2 = ELEVATION OF W5
C     L3 = ELEVATION OF W1
C     L4 = ELEVATION OF W2
C     L5 = ELEVATION OF W4
C     L6 = DISTANCE BET. W3 & W5
C     L7 = DISTANCE BET. W1 & W2
C     L8 = DISTANCE BET. W2 & W5'
C     L9 = DISTANCE BET. W4 & W5"
C     G = HYD. GRADIENT
C     O = CONSTANT
      OPEN(1,FILE='W.DAT')
      OPEN(2,FILE='O.DAT',STATUS='NEW')
      WRITE(*,*)'SPECIFIC YIELD ?'
      READ(*,*)S
      WRITE(*,*)'ELEVATIONS ?'
      READ(*,*)L1,L2,L3,L4,L5
      WRITE(*,*)'DISTANCE BET. THE WELLS ?'
      READ(*,*)L6,L7,L8,L9
      READ(1,*)((C(N,I),I=1,5),N=1,12)
      WRITE(2,20)
 20   FORMAT(2X,'O1(N)',7X,'O2(N)',7X,'O3(N)',7X,'O4(N)')
      DO 100 N=1,11
      U1=(C((N+1),1)-C(N,1))*S
      U2=(C((N+1),3)-C(N,3))*S
      U3=(C((N+1),4)-C(N,4))*S
      U4=(C((N+1),5)-C(N,5))*S
      G1=((L1-C(N,1))-(L2-C(N,2)))/L6
      G2=((L3-C(N,3))-(L4-C(N,4)))/L7
      G3=((L4-C(N,4))-(L2-C(N,2)))/L8
      G4=((L5-C(N,5))-(L2-C(N,2)))/L9
      O1(N)=U1/G1
      O2(N)=U2/G2
      O3(N)=U3/G3
      O4(N)=U4/G4
      WRITE(2,30)O1(N),O2(N),O3(N),O4(N)
 30   FORMAT(1X,F5.2,6X,F5.2,6X,F5.2,6X,F5.2)
100   CONTINUE
      STOP
      END
```

```
C     PROGRAM LISTING THE MAIN STRUCTURE OF THE MODEL
      DIMENSION L(3,4),T(3,4),M(3,3),R(52),G1(52),G2(52),G3(52)
      DIMENSION V1(52),V2(52),V3(52),O(4,9),P(52),E(52)
      DIMENSION A1(52),A2(52),A3(52),A4(52),A5(52),A6(52),A7(52)
      DIMENSION A8(52),A9(52),T1(52),T2(52),T3(52),T4(52)
      DIMENSION T5(52),T6(52),T7(52),T8(52),T9(52)
C     ***********************************************************
C     O,A =  CONSTANT
C     G   =  HYD. CONDUCTIVITY.
C     T   =  RESULTANT DEPTH TO WT,m
C     V   =  FLOW
C     W   =  STORAGE DEPTH,m
      OPEN(1,FILE='A.DAT')
      OPEN(2,FILE='P.DAT')
      OPEN(3,FILE='E.DAT')
      OPEN(4,FILE='RT.DAT',STATUS='NEW')
      OPEN(5,FILE='R.DAT',STATUS='NEW')
C     ***********************************************************
C     INPUT DATA
C     ***********************************************************
      WRITE(*,*)'SPECIFIC YIELD ?'
      READ(*,*)S
      WRITE(*,*)'INITIAL LEVELS ?'
      READ(*,*)((T(I,J),J=1,4),I=1,3)
      WRITE(*,*)'ELEVATIONS ?'
      READ(*,*)((L(I,J),J=1,4),I=1,3)
      WRITE(*,*)'DISTANCE BET. THE WELLS ?'
      READ(*,*)((M(I,J),J=1,3),I=1,3)
      READ(1,*)((O(I,J),J=1,9),I=1,4)
      READ(2,*)(P(J),J=1,52)
      READ(3,*)(E(J),J=1,52)
      WRITE(4,10)
10    FORMAT(13X,'PREDICTED DEPTHS')
      WRITE(4,15)
15    FORMAT(1X,'~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~')
      J=1
      DO 100 K=1,52
C     ***********************************************************
C     ASSIGNS CONSTANTS ACCORDING TO THE DEPTH
C     ***********************************************************
      IF (T(1,1) .LT. 14.00) THEN
      A1(K)=O(1,1)
      ELSE
      IF ((T(1,1) .LT. 15.00) .AND. (T(1,1) .GE. 14.00)) THEN
      A1(K)=O(2,1)
      ELSE
      IF ((T(1,1) .LT. 15.50) .AND. (T(1,1) .GE. 15.00)) THEN
      A1(K)=O(3,1)
      ELSE  IF (T(1,1) .GE. 15.50) THEN
      A1(K)=O(4,1)
      ENDIF
      ENDIF
      ENDIF
      ENDIF
      IF (T(1,2) .LT. 2.5) THEN
      A2(K)=O(1,2)
      ELSE
      IF ((T(1,2) .LT. 3.5) .AND. (T(1,2) .GE. 2.5)) THEN
      A2(K)=O(2,2)
      ELSE
      IF ((T(1,2) .LT. 4.5) .AND. (T(1,2) .GE. 3.5)) THEN
      A2(K)=O(3,2)
      ELSE
```

```
IF (T(1,2) .GE. 4.5) THEN
A2(K)=O(4,2)
ENDIF
ENDIF
ENDIF
ENDIF
IF (T(1,3) .LT. 2.5) THEN
A3(K)=O(1,3)
ELSE
IF ((T(1,3) .LT. 3.5) .AND. (T(1,3) .GE. 2.5)) THEN
A3(K)=O(2,3)
ELSE
IF ((T(1,3) .LT. 4.5) .AND. (T(1,3) .GE. 3.5)) THEN
A3(K)=O(3,3)
ELSE
IF (T(1,3) .GE. 4.5) THEN
A3(K)=O(4,3)
ENDIF
ENDIF
ENDIF
ENDIF
IF (T(2,1) .LT. 11.0) THEN
A4(K)=O(1,4)
ELSE
IF ((T(2,1) .LT. 12.0) .AND. (T(2,1) .GE. 11.0)) THEN
A4(K)=O(2,4)
ELSE
IF ((T(2,1) .LT. 12.5) .AND. (T(2,1) .GE. 12.0)) THEN
A4(K)=O(3,4)
ELSE
IF (T(2,1) .GE. 12.5) THEN
A4(K)=O(4,4)
ENDIF
ENDIF
ENDIF
ENDIF
IF (T(2,2) .LT. 2.0) THEN
A5(K)=O(1,5)
ELSE
IF ((T(2,2) .LT. 3.0) .AND. (T(2,2) .GE. 2.0)) THEN
A5(K)=O(2,5)
ELSE
IF ((T(2,2) .LT. 4.0) .AND. (T(2,2) .GE. 3.0)) THEN
A5(K)=O(3,5)
ELSE
IF (T(2,2) .GE. 4.0) THEN
A5(K)=O(4,5)
ENDIF
ENDIF
ENDIF
ENDIF
IF (T(2,3) .LT. 2.0) THEN
A6(K)=O(1,6)
ELSE
IF ((T(2,3) .LT. 3.0) .AND. (T(2,3) .GE. 2.0)) THEN
A6(K)=O(2,6)
ELSE
IF ((T(2,3) .LT. 4.0) .AND. (T(2,3) .GE. 3.0)) THEN
A6(K)=O(3,6)
ELSE
IF (T(2,3) .GE. 4.0) THEN
A6(K)=O(4,6)
ENDIF
```

```
      ENDIF
      ENDIF
      ENDIF
      IF (T(3,1) .LT. 3.0) THEN
      A7(K)=O(1,7)
      ELSE
      IF ((T(3,1) .LT. 4.0) .AND. (T(3,1) .GE. 3.0)) THEN
      A7(K)=O(2,7)
      ELSE
      IF ((T(3,1) .LT. 4.5) .AND. (T(3,1) .GE. 4.0)) THEN
      A7(K)=O(3,7)
      ELSE
      IF (T(3,1) .GE. 4.5) THEN
      A7(K)=O(4,7)
      ENDIF
      ENDIF
      ENDIF
      ENDIF
      IF (T(3,2) .LT. 2.0) THEN
      A8(K)=O(1,8)
      ELSE
      IF ((T(3,2) .LT. 2.5) .AND. (T(3,2) .GE. 2.0)) THEN
      A8(K)=O(2,8)
      ELSE
      IF ((T(3,2) .LT. 3.0) .AND. (T(3,2) .GE. 2.5)) THEN
      A8(K)=O(3,8)
      ELSE
      IF (T(3,2) .GE. 3.0) THEN
      A8(K)=O(4,8)
      ENDIF
      ENDIF
      ENDIF
      ENDIF
      IF (T(3,3) .LT. 2.0) THEN
      A9(K)=O(1,9)
      ELSE
      IF ((T(3,3) .LT. 2.5) .AND. (T(3,2) .GE. 2.0)) THEN
      A9(K)=O(2,9)
      ELSE
      IF ((T(3,3) .LT. 3.0) .AND. (T(3,2) .GE. 2.5)) THEN
      A9(K)=O(3,9)
      ELSE
      IF (T(3,3) .GE. 3.0) THEN
      A9(K)=O(4,9)
      ENDIF
      ENDIF
      ENDIF
      ENDIF
C     *****************************************************
C     CALCULATION OF STORAGE DEPTH
C     *****************************************************
      R(K)=0.5*(P(K)-E(K))/100
      WRITE(5,*)R(K)
      W1=R(K)/S
C     *****************************************************
C     CALCULATION OF HYDRAULIC GRADIENT
C     *****************************************************
      DO 200 I=1,3
      G1(I)=((L(I,J)-T(I,J))-(L(I,J+1)-T(I,J+1)))/M(I,J)
      G2(I)=((L(I,J+1)-T(I,J+1))-(L(I,J+2)-T(I,J+2)))/M(I,J+1)
      G3(I)=((L(I,J+2)-T(I,J+2))-(L(I,J+3)-T(I,J+3)))/M(I,J+2)
```

```
  200 CONTINUE
C     ************************************************************
C     CALCULATION OF GROUNDWATER FLOW
C     ************************************************************
      V1(1)=G1(1)*A1(K)/S
      V2(1)=G2(1)*A2(K)/S
      V3(1)=G3(1)*A3(K)/S
      V1(2)=G1(2)*A4(K)/S
      V2(2)=G2(2)*A5(K)/S
      V3(2)=G3(2)*A6(K)/S
      V1(3)=G1(3)*A7(K)/S
      V2(3)=G2(3)*A8(K)/S
      V3(3)=G3(3)*A9(K)/S
C     ************************************************************
C     CALCULATION OF THE RESULTANT DEPTHS
C     ************************************************************
      IF (T(1,1) .GE. (W1-V1(1))) THEN
      T1(K)=T(1,1)-W1+V1(1)
      ELSE
      T1(K)=0
      ENDIF
      IF (T(1,2) .GE. (W1+V1(1)-V2(1))) THEN
      T2(K)=T(1,2)-W1-V1(1)+V2(1)
      ELSE
      T2(K)=0
      ENDIF
      IF (T(1,3) .GE. (W1+V2(1)-V3(1))) THEN
      T3(K)=T(1,3)-W1-V2(1)+V3(1)
      ELSE
      T3(K)=0
      ENDIF
      IF (T(2,1) .GE. (W1-V1(2))) THEN
      T4(K)=T(2,1)-W1+V1(2)
      ELSE
      T4(K)=0
      ENDIF
      IF (T(2,2) .GE. (W1+V1(2)-V2(2))) THEN
      T5(K)=T(2,2)-W1-V1(2)+V2(2)
      ELSE
      T5(K)=0
      ENDIF
      IF (T(2,3) .GE. (W1+V2(2)-V3(2))) THEN
      T6(K)=T(2,3)-W1-V2(2)+V3(2)
      ELSE
      T6(K)=0
      ENDIF
      IF (T(3,1) .GE. (W1-V1(3))) THEN
      T7(K)=T(3,1)-W1+V1(3)
      ELSE
      T7(K)=0
      ENDIF
      IF (T(3,2) .GE. (W1+V1(3)-V2(3))) THEN
      T8(K)=T(3,2)-W1-V1(3)+V2(3)
      ELSE
      T8(K)=0
      ENDIF
      IF (T(3,3) .GE. (W1+V2(3)-V3(3))) THEN
      T9(K)=T(3,3)-W1-V2(3)+V3(3)
      ELSE
      T9(K)=0
      ENDIF
      WRITE(4,70)T1(K),T2(K),T3(K),T4(K),T5(K),T6(K),T7(K),T8(K),T9(K)
```

```
 70    FORMAT(1X,3F5.2,1X,3F5.2,1X,3F5.2)
C      **************************************************************
C      ASSIGN THE CALCULATED DEPTHS AS THE INITIAL LEVELS
C      **************************************************************
       T(1,1)=T1(K)
       T(1,2)=T2(K)
       T(1,3)=T3(K)
       T(2,1)=T4(K)
       T(2,2)=T5(K)
       T(2,3)=T6(K)
       T(3,1)=T7(K)
       T(3,2)=T8(K)
       T(3,3)=T9(K)
 100   CONTINUE
       STOP
       END

C      ~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~
```

Appendix - VII

The value of constant for different depth ranges (A.DAT)

1.85	3.36	3.36	1.85	2.92	2.92	1.85	2.17	2.17
0.47	1.95	1.95	0.47	2.28	2.28	0.47	1.74	1.74
0.49	1.40	1.40	0.49	1.38	1.38	0.49	1.60	1.60
0.22	0.88	0.88	0.22	1.08	1.08	0.22	1.03	1.03

Values entering during the execution of the program

Specific yield =0.22

Initial levels m	Elevation m	Distance between point, m
15.85	30	
5.38	11	175
4.86	8	250
3.20	5	100
12.85	27	115
3.20	12	300
4.86	8	100
3.20	5	80
1.85	16	300
3.20	11	100
4.86	8	
3.20	5	

MONTHLY WATER BALANCE MODEL FOR LATERITIC HILL SLOPE - A CASE STUDY

BY

BEENA THOMAS

ABSTRACT OF A THESIS

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ABSTRACT

The groundwater balance of K.C.A.E.T. campus, Tavanur, Kerala has been studied and a deterministic model has been developed for the prediction of depth to water table in this hill slope based on Darcy's equation. The input data required are the precipitation, evapotranspiration, specific yield, initial water table, elevation of the points under consideration and the distance between the points. The weekly precipitation and evapotranspiration were calculated and from this the recharge to groundwater was estimated. The groundwater recharge was taking place only during the southwest monsoon season.

The change in storage divided by the hydraulic gradient for a period without rainfall i.e., from 01 December 1994 to 16 February 1995 were determined. Using these values and other inputs in the model, the depth to water table for a period from 26 May 1994 to 25 May 1995 were estimated from the model. These values were compared with the observed values for the same period. It was found that the estimated values give a reasonable estimate of the depth to water table in the study area. This model can be used to predict the changes expecting in the water level in a particular region. The model has several areas of application in the fields of groundwater resource development and irrigation.

The infiltration characteristics of the selected area was evaluated and observed that the infiltration rate was increasing towards the higher elevation. The model was simulated for different percentages of effective rainfall and found that, at 50 per cent it accounts as groundwater recharge and the rest as surface runoff.