LABORATORY STUDIES ON SOIL MOISTURE FLOW AND RUNOFF GENERATION IN LATERITE SOIL

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THESIS

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1998

DECLARATION

I hereby declare that this thesis entitled **"Laboratory studies on soil moisture flow and runoff generation in laterite soil"** 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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SYMBOLS AND ABBREVIATIONS

Agric.	- agricultural
Am.	- America
ASAE	- American Society of Agricultural Engineers
Atm.	- atmospheric
Bull.	- bulletin
cm	- centimetre
cm/h	- centimetre per hour
Cons.	- conservation
CPCRI	- Central Plantation Crops Research Institute
ed.	- edition
Engg.	- engineering
et. al.	- and others
g	- gram
G	- gage
Geol.	- geological
Geophys.	- geophysical
GI	- galvanized iron
h	- hour
Hydrol.	- hydrology
J.	- journal
KCAET	- Kelappaji College of Agricultural Engineering and Technology
kg	- kilogram
l	- litre
Mag.	- magazine
Manage.	- management

Meteor.	- meteorological
mm	- millimetre
mm/h	- millimetre per hour
MS	- mild steel
Proc.	- proceedings
Res.	- research
Resour.	- resources
Sci.	- science
Soc.	- society
STEC	- State Committee on Science, Technology and Environment
Tr.	- transactions
USDA	- United States Department of Agriculture
viz.	- namely
%	- percent
&	- and

Introduction

INTRODUCTION

Over a great part of the earth, water limits plant growth either because there is too much or too little of it in the soil. The extreme cases are in humid and arid areas where if crops are to be grown at all they have to be provided with proper drainage or irrigation. Between these extremes lie the established agricultural land, many of which can be improved by having less water or more in the soil at certain times of a year. A great deal is done to control water in agriculture and if both land and water are to be put to the best use, a full understanding of soil-water relation is necessary. Much of the research on retention and movement of water in soil and the use of water by plants is done with this objective.

The water that enters into the soil either by rain or irrigation may pass through the surface soil layers to the watertable and is known as groundwater. The water which is not drained deep into the soil profile is either retained in the soil pores and channels or on the surface of the soil particles. In soil, water is held in the pore space and in a non-equilibrium state, the fluid is in motion in response to the gradient of potential. Under isothermal conditions, generally two types of flows are recognised - diffusion and bulk flow. The flow of water in liquid state under water potential gradient is the case of bulk flow.

Infiltration and redistribution of water within soil profiles are of significant importance to present day water conservation and groundwater contamination problems. The rates at which these processes occur depend on the water transmission characteristics of the soil profile. The water in the soil is in a highly dynamic state. It is essentially moving in the soil profile from one point to another in response to water-moving forces. Knowledge of the pattern of water movement within the soil profile is essential in the solution of problems involving irrigation, drainage, water conservation, groundwater recharge and pollution as well as infiltration and runoff control.

It is now well known that two basic hydrologic characteristics of a porous material must be defined experimentally before it is possible to carry out numerical analyses of water movement in the unsaturated phase for that material. These characteristics are the soil water suction - water content relation and the hydraulic conductivity - water content relation. Although analytic expressions based on Darcy's equation for unsaturated soil-water movement and retention have been available, relatively few investigators have described soil-water behaviour using measured values of both hydraulic conductivity and hydraulic gradients manifested within field soil profiles as a function of soil depth. Modelling of water movement in soils requires knowledge of hydraulic conductivity as a function of volumetric water content [$K(\theta)$] or soil water pressure head [$K(h)$], and the soil water retention curve (h).

The study was done on laterite soil. Laterite soils are by far the most important group occurring in Kerala and cover the largest area. The broad belt of land lying between the sea and eastern hilly region of the State, varying in width from 50 - 100 Km. is a lateritic belt. The soil is porous, well-drained and have poor capacity for retaining moisture. Almost every crop grown in the State is cultivated on laterite soils. They include paddy, coconut, tapioca, rubber, pepper, ginger, bananas, sugarcane, arecanut and cashewnut. The lateritic terrain of Kerala can be considered as the backbone of the State, as its economy depends upon this lateritic terrain which produces most of its cash crops. The State's agriculture is mainly confined to laterite soils.

The factors which affect the hydraulic conductivity of unsaturated soil are those related to the nature of the soil and the soil water content. The saturated and unsaturated flow processes play a conspicuous role in hydrology and thus in watershed management. It functions in generating runoff from rainfall, base flow and groundwater recharge. That is, the study helps in quantifying the moisture and transport in the soil during the rainy season.

In this study, a rainfall simulator was fabricated and various rainfall intensities were simulated. At different rainfall intensities, the flow of water was monitored in terms of moisture content, tension, surface runoff and outflow. Experiments were conducted at varying slopes also. A relationship between hydraulic conductivity and volumetric moisture content was established. The relation between runoff and soil moisture status was determined.

The main objective of this research work is to study the saturated and unsaturated flow of water in laterite soil under selected precipitation intensities. The specific objectives are as follows :

1. To design and develop a rainfall simulator and a soil trough.
2. To monitor the saturated and unsaturated flow of water in selected laterite soil.
3. To establish a relation between hydraulic conductivity, moisture content and hydraulic head.
4. To study the generation of surface runoff and the relationship between runoff and soil moisture status.

Review of Literature

REVIEW OF LITERATURE

The soil water suction - water content relation and the hydraulic conductivity-water content relation are the two basic hydrologic characteristics of a porous material to be defined experimentally before carrying out numerical analyses of water movement in the unsaturated phase. This chapter gives an idea of the previous works done in this field. The experimental setup for conducting laboratory studies were also reviewed.

2.1 Laterite soil

In Kerala, at Angadippuram, a ferraginous, vesicular, soft material occurring within the soil which hardens irreversibly on exposure and used as a building material was first recognised as "laterite" by Francis Buchanan(1807). He coined the term laterite, from "later", the Latin word for brick. Laterites are products of intense rock weathering. A number of theories were propounded to explain the genesis of laterite soils. D'Hoore(1954) grouped these theories into : (a)concentration of sesquioxides by removal of silica and bases ie, relative accumulation (b)concentration of sesquioxides by accumulation either across the profile or between profiles ie, absolute accumulation.

The earliest of the modern suggestions regarding the origin of laterite may be that of Russell(1889) wherein he emphasised the fact that in warm, moist temperate and tropical climates, the water percolating through the rocks have greater solvent action and the red colour of the soils was due to iron oxides presumed to be residual after considerable amount of alkaline earths have been removed, largely by chemical solution. Harrison and Reid(1910) contributed to the view that laterite is a weathering product of rocks in which the chemical decomposition of the silicates had resulted in the formation of secondary silicious compounds, secondary silica and alumina and oxides of iron in more or less hydrated forms. Campbell(1917), distinguished between the process of "alteration" and "weathering" in rocks and emphasised that laterite was essentially a precipitation and not a residual product. According to Alexander and Cady(1962) laterite is a highly weathered material rich in secondary oxides of iron, aluminium or both. It is nearly void of bases and primary silicates, but it may contain large amounts of quartz and kaolinite. The laterite soils are those in which laterization is the dominant soil forming process.

Typical laterite soils are characterized by a vesicular structure and the accumulation of hydrated oxides of iron and aluminium. Laterite soils may vary in depth from 1.8 to 3m and may have a thick layer of kaolin clay below. These soils do not manifest typical clay properties such as plasticity, cohesion, expansion and shrinkage to any great extent. They are porous and well-drained and have poor capacity for retaining moisture. The base exchange capacity is also low. In some regions of Kerala the soils have not developed into true laterites. There is an accumulation of iron and aluminium in these soils and they show many of the properties of laterites. They do not have the vesicular structure, however, which is peculiar to true laterites. These soils are called lateritic soils. The deep red colour of these soils is due to haematite or anhydrous ferric oxide. According to Buchanan(1807), laterite soil is one in which a laterite horizon is found and lateritic soil is one wherein exists an under-developed laterite horizon which under appropriate conditions will become a true laterite. Harrassowitz(1926) gave a chemical definition on the basis of their silica/alumina molar ratio; soils with this value less than 1.33 were laterite and those with value above it were lateritic soils.

From the distribution of the laterite soils it can be seen that these vast regions have a large proportion of favourable topography for agriculture and adequate temperature for the plant growth. There are only very few physical constraints for crop production. Physical constraints include susceptibility to erosion, low water holding capacity and drought stress. When the chemical constraints are eliminated by liming and the application of the necessary amounts of fertilizers, the productivity of these soils are among the highest in the world. Laterite soils are by far the most important group occurring in Kerala and cover the largest area. On laterites at a lower elevation, rice is grown and on those at higher elevation, plantation crops grow well under good soil management. Almost every crop grown in Kerala is cultivated on laterite soils. They include paddy, coconut, tapioca, rubber, pepper, ginger, bananas, sugarcane, arecanut and cashewnut. It is to be noted that the broad belt of land lying between the sea and the eastern hilly regions of the State, varying in width from 50-100 Km is a lateritic belt and the State's agriculture is mainly confined to laterite soils. The lateritic terrain can be considered as the backbone of the State, as its economy depends upon this lateritic terrain which produces most of its cash crops.

2.2 Soil moisture flow

Soil consists of four fractions : the mineral particles and nonliving matter which form the matrix, and the soil solution and air which occupy the pore spaces within the matrix. It provides the anchorage which enables the roots to maintain plants in an erect position and acts as a reservoir for water and salt. Much of the success of plants in any given habitat depends on the suitability of the soil as a medium for root growth and functioning.

When water, whose source is either rainfall or irrigation, is applied to soil, it enters the soil pores replacing the air contained in them. In soil, water is held in the pore space and in a non-equilibrium state, the fluid is in motion in response to the gradient of potential. Under isothermal conditions, generally two types of flows are recognised- diffusion and bulk flow. The flow of water in liquid state under water potential gradient is the case of bulk flow.

2.2.1 Movement of water into soils

The rate of infiltration into soil is an extremely important factor in soil moisture recharge by rain or irrigation. The path of downward movement of water in the soil following its application to the surface was described in detail by Bodman and Colman(1944) for a uniform profile and by Colman and Bodman(1945) for a non-uniform profile. They found that the wetted portion of a column of uniform soil into which water was entering at the top and moving downward appeared to comprise a stable gradient through which water was transmitted, ranging from a saturated zone at the top to a wetting zone at the lower end. Five zones in series were described as (1)a saturated zone ie, a zone presumed saturated which reached a maximum depth of 1.5cm (2)a transition zone, a region of rapid decrease of water content extending to a depth of about 5cm from the surface (3)the main transition zone, a region in which only small changes in water content occurred (4)the wetting zone, a region of fairly rapid change in water content and (5)the wetting front, a region of very steep gradient in water content which represents the visible limit of water penetration. The typical wetting pattern for a loamy sand is given in fig.1.

2.2.2 Movement of water within soils

The movement of water within soils controls not only the rate of infiltration but also the rate of supply to roots and the rate of underground flow to springs and streams.

2.2.2.1 Movement of liquid water

It has been customary to differentiate between saturated flow or saturated conductivity in saturated soils and unsaturated flow or capillary conductivity in

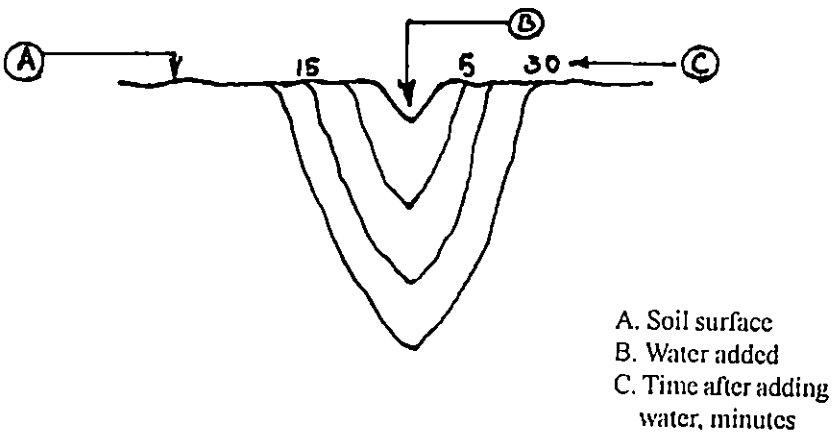


Fig. 1 **Typical wetting pattern for loamy sand**

soils which are unsaturated. However, the term hydraulic conductivity; formerly used for water flow in saturated soils, is now being used for both saturated and unsaturated flow. The chief difference is that in saturated soils, gravity controls the water potential gradient, while in drained soils it is controlled by the matric potential and water moves in films surrounding the soil particles rather than by gravity flow through the pores.

2.2.2.1.1 Saturated flow

If sufficient quantity of water is available, the entire pore space may be filled with water and the excess water would move downward by a physical process known as saturated flow. In saturated soils, water movement takes place throughout a soil-pore space that contains little, if any, air.

According to Darcy's law(1856) the velocity of flow of water through a column of soil is directly proportional to the difference in pressure head and inversely proportional to the length of the column,

$$v = k\Delta H/l$$

where,

v - the velocity, cm/s

ΔH - the difference in pressure head, cm

l - the length of the column, cm

k- the proportionality constant

Richards(1940), in a discussion of permeability units for soils, has suggested that a permeability unit that is based upon a flow equation adequate to cover a variety of cases should be used generally in soil investigations. The modified Darcy formula, as given below, is suggested as being rather convenient for such work.

$$v = k(\eta_s/\eta)g.i$$

This is the same as the previous equation with the addition of the gravity constant(g) and a relative viscosity factor.

Downward movement of water in soils must take place through different horizons. The porosity and hydraulic conductivity of the various layers may be greatly different. In a saturated soil, the percolation rate is determined by the rate of movement through the least pervious horizon. If an impervious layer exists in the subsoil, water movement of any consequence through such a horizon can take place only through cracks and fissures or old root channels and worm holes. It is difficult to granulate these deeper layers adequately. The movement of water in the larger pores may be influenced considerably by the resistance of entrapped soil air. If water that enters the soil leaves some pores in contact with the atmosphere, the soil-air pressure remains unchanged. However, if a rapid entrance of water into the soil entraps appreciable quantities of air the soil air pressure will increase and percolation will decrease. It can be said that the downward movement of water by gravitational forces in natural soils is related to (a)the amount and continuity of the non-capillary pores as determined by soil structure, texture, volume changes and biological channels (b) the hydration of pores, and (c)to the resistance of entrapped air.

2.2.2.1.1.1 Flow over a sloping plane

Considering the flow in a porous medium, the surface tension forces is small in comparison with gravity forces. The porous medium is saturated with water to a depth h(x,t), at the surface of which the pressure is taken constant. (Fig.2) Above the saturated region, the only significant water transport arises from infiltration, which is assumed to contribute a constant inflow 'v_0' to the saturated layer for a limited time.

This yields the following equation,

$$\partial H/\partial \tau + 2(\partial H/\partial X) = \partial/\partial X(H\partial H/\partial X)+\lambda$$

$$\text{ie, } X=x/L \text{ , } H=2h/(L\tan\theta) \text{ , } \tau=(kgSt)/(2\varepsilon vL)$$

Then $\lambda=4vCv_0/(kgS^2)$ is a dimensionless source term. The magnitude of λ is a measure of the rate of infiltration relative to the rate at which water flows downhill in the saturated gravity flow. (Henderson & Wooding, 1964)

The basic governing equation used for the problem of unsteady-state subsurface drainage of sloping lands has been the Boussinesq equation :

$$z = \partial^2 z/\partial x^2 + (\partial z/\partial x)^2 - a\partial z/\partial x = E/k(\partial z/\partial t)$$

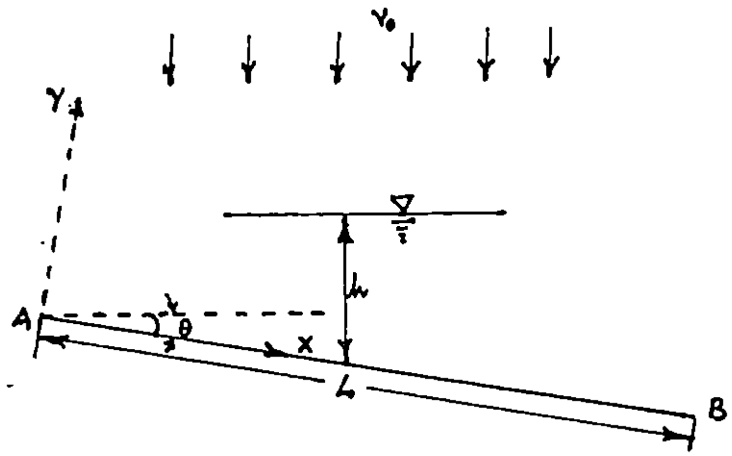


Fig. 2 Schematic representation of flow over a sloping plane

where

z - height of watertable above impermeable layer at horizontal distance x from origin at time t

E- specific yield of soil

K - hydraulic conductivity and

α- slope of impermeable layer

Chapman(1980) suggested a modification incorporating an extended Dupuit-Forchheimer assumption in the Boussinesq equation. This modification is given by the following equations.

$$q = -K\cos^2\alpha h(\partial H/\partial X)$$

where

q - discharge

K - hydraulic conductivity

$$\partial/\partial X(h\partial H/\partial X)\cos^2 a + P/K = (S/K)\,\partial H/\partial t$$

where

P - net rate of vertical accretion to the free surface

S - storage coefficient

2.2.2.1.2 Unsaturated flow

If the application of water is limited or if evaporation is taking place at the soil surface and plants are absorbing water from wetter zones, the soil pores will be only partially filled with water and the soil water potential decreases. Water moves from a region of high potential to depleted soil water zones. Such a type of water movement is termed as unsaturated flow. An unsaturated soil is one in which the larger pores are filled with air, and consequently, where movement is closely dependent upon a large number of air-water interfaces. The distribution and movement of water may be laterally, vertically upward, vertically downward or at any angle between the vertical and horizontal. Soil-moisture movements under these conditions are discussed from the view point of the older concepts of capillarity as well as from the more recent analogies to the flow of heat or electricity.

2.2.2.1.2.1 Capillarity

The fundamental basis of the capillary tube hypothesis, with respect to the movement of capillary water in soils, is the well known height-of-capillary-rise equation.

$$h=2T/gDr$$

where

h - the height of the meniscus above the water level

T - the surface tension

D - the density of the liquid

g - the acceleration due to gravity

r - the radius of the capillary tube

Interpreting this equation in terms of the soil, we see that the height of rise is inversely proportional to the radius of the pores. The presence of many small pores in the soil would therefore suggest considerable capillary movement of water in soils.

Wollny(1885) studied various factors that affected the capillary movement of water in soils. He found that the rate of capillary rise in a loam soil increased with temperature, with the looseness of packing and with the original moisture content of the soil. He also observed that the rise was faster in a column of sand particles of mixed sizes than in columns having particles of uniform size. Harris and Turpin(1917) presented some interesting data on the capillary movement of water. They showed that the greatest rise and descent of water into a dry soil from a moist one always took place in the case of greatest initial moisture content of the source. Capillary movement downward was slightly faster than upward or laterally.

The capillary-tube hypothesis has emphasized distance of movement rather than rate. Experiments show, however, that the rate of capillary rise is very slow in these soils where the pore size mathematically suggest a high rise. In other words, these older concepts have indicated that there is a capacity factor and a conductivity factor in capillary phenomena; they seldom have been recognised as such in the analysis of experimental data.

2.2.2.1.2.2 Analogy of capillary movement with heat flow

The potential gradient is the change in potential per unit distance in the direction of the maximum rate of increase of potential. Since flow always takes place from a higher to a lower potential, the rate of flow of water through a pipe is expressed as :

$$\text{Rate of flow} = -k\,\mathrm{grad}\,\pi$$

where

k - the conductance of the pipe

$\mathrm{grad}\,\pi$ - the potential gradient

The flow of water in a soil may be expressed, according to the Darcy equation,

$$v = -k\,\mathrm{grad}\,\phi$$

where

$\mathrm{grad}\,\phi$ - the change in the total water- moving force per unit distance

k - the specific conductivity

here, $\mathrm{grad}\,\phi$ = capillary potential gradient + gravitational potential gradient

The capillary conductivity depends upon the kind of soil, its state of packing and the moisture content.

2.2.2.1.2.3 Hydraulic concept in moisture movement

Richards(1941) applied the principles of hydraulics to the movement of water in an unsaturated soil. The pressure measurements obtained with manometers attached to suitable porous media in contact with the soil can be interpreted in terms of hydraulic head and hydraulic gradient. The hydraulic gradient is defined by Richards as the loss in hydraulic head per unit distance along an average flow line. If soil tensiometers are placed in the soil at various depths, the direction of movement of soil moisture in an unsaturated soil can be determined by the hydraulic gradients between any two zones in the soil. If the soil surface is chosen as datum point, the depth of the porous cup subtracted from the hydraulic head gives the soil moisture tension in centimetres of water.

Richards(1928) has called attention to three common types of capillary movement of water : (a) the movement of precipitation or irrigation water downward through a comparatively dry soil, (b) movement of water upward from a saturated level and (c) the movement of water in the horizontal direction. The movement of capillary water downward takes place under the combined influences of the gravitational-potentialgradient and the capillary-potential gradient. If evaporation is prevented at the surface, downward movement will continue till the soil is drained or till equilibrium is attained with an impermeable layer or a saturated water table.

The nature of the soil profile, its texture, structure and pore size distribution determine to a large extent how water penetrates and is retained in the soil. The hydraulic conductivity is very sensitive to water content or suction. The factors which affect the hydraulic conductivity of unsaturated soil are those related to the nature of the soil and the soil water content.

In investigating the contribution of soil moisture to stream flow, Hewlett and Hibbert(1963) concluded from a laboratory slope drainage study that unsaturated flow alone was responsible for sustained base flow. These results were tested by constructing a laboratory slope drainage model. The results from the model, together with hydraulic conductivity determinations indicate the slope discharge to be controlled by saturated flow throughout drainage.(Anderson and Burt,1976).

Nieber and Walter(1981) presented a laboratory system for modelling hill slope soil moisture flow. Results from experiments with a 3.66m×0.58m×0.108m flume filled with sand were used to study the two-dimensional flow of soil moisture under the condition of rainfall infiltration. The experimental results were compared to results obtained by a numerical solution of the two-dimensional Richards' equation.

Ogawa *et al*(1992) investigated infiltration and discharge of rainwater and water movement in soil by a numerical analysis based on a one-dimensional infiltration model. Numerical analyses were performed for various precipitation intensities under various initial soil moisture conditions and with various soil properties. The speed of wetting front was found to approach that of the tracer as the precipitation intensity increased, as the initial soil moisture decreased and the grain size increased. Field studies were conducted on wetting front under drip irrigation at CPCRI, Kasaragod in laterite soil with gravelly-clay texture. It was found that vertical and horizontal movement of water was directly related to quantity of water applied.(Dhanapal *et al*,1995).

2.2.2.1.3 Steady flow

The flow of water satisfies the equation of continuity, which in general form may be expressed as

$$-[\partial (\rho v_x)/\rho x + \partial (\rho v_y)/ \partial y + \partial (\rho v_z)/ \partial z] = \partial \rho/\partial t$$

where

v_x, v_y, v_z - velocity in x,y and z directions

ρ - fluid density

t - time

For steady flow, there is no change in conditions with respect to time, and regarding water as an incompressible fluid makes 'ρ' a constant; therefore:

$$\partial v_x/\partial x + \partial v_y/\partial y + \partial v_z/\partial z = 0$$

2.3 Soil water properties

Although analytic expressions based on Darcy's equation for unsaturated soil water movement and retention has been available, relatively few investigators have described soil water behaviour using measured values of both hydraulic conductivity and hydraulic gradients manifested within field soil profile as a function of soil depth.

Davidson *et al*(1969) measured the hydraulic conductivity versus soil water content relation for different soil depths in the field for three soil profiles. The soils varied in physical properties from heterogeneous to homogeneous with depth and from loamy sand to silty clay in surface soil texture. Hydraulic conductivity values were calculated from drainage data taken during different time intervals. The soil water flux at various soil depths with and without evaporation at the soil surface was measured. The rate at which water drained from each of the profiles was predicted using Darcy's equation. The agreement between theoretical and measured results were explained in terms of soil heterogenity and depth.

The need for determining the hydraulic properties of soil profiles was pointed out by Hillel *et al*(1972). They used instantaneous profile method for determining soil hydraulic properties based on simultaneously monitoring the changing wetness and matric suction profiles during internal drainage. From the measured soil wetness and matric suction, the instantaneous values of the potential gradients and fluxes operating within the profile and the hydraulic conductivity values were obtained. The results were analysed to obtain the function of conductivity versus water content for each layer in the profile as well as for the profile as a composite whole.

A study was conducted by Libardi *et al*(1980) to examine the use of simplifying assumptions for the solution of Richards' equation in order to develop two simple methods for estimating hydraulic conductivity, as a function of soil water content, in the field. Values of hydraulic conductivity calculated by both the two (θ-method and flux method) proposed as well as by a third recently reported method(CGA-method) were compared with those calculated by integrating Richards' equation without simplifying assumptions. It was concluded that a greater number of observations made possible by simplifying assumptions, and hence less instrumentation and cost, is preferable to fewer observations with more exact methods that are not amenable to statistical analyses over large land areas.

A new and relatively simple equation for the soil-water content-pressure head curve, $\theta(h)$, was given by van Genuchten(1980). The particular form of the equation enables one to derive closed-form analytical expressions for the relative hydraulic conductivity, K_r, when substituted in the predictive conductivity models of N. T. Burdine or Y. Mualem. The resulting expression for $K_r(h)$ contain three independent parameters which may be obtained by fitting the proposed soil-water retention model to experimental data. It was found that a reasonable description of the soil-water retention curve at low water contents is important for an accurate prediction of the unsaturated hydraulic conductivity.

Shani *et al*(1987) proposed a method using 'drippers' to estimate the soil hydraulic properties based on assumed relationships of the hydraulic conductivity

(K) - matric head (h). The method is based on the observation that when water is applied at a constant rate to a point on the soil surface, a ponded zone is created that approaches a constant area in a short time. Thus, steady-state solutions of the two - dimensional flow equation can be applied. The method consists of the following steps:(1)Wet the soil from a dripper with several known discharge rates in a relatively dry soil. (2)Once the borders of the ponded zone are steady, saturated hydraulic conductivity(K_s) and the matric flux function(F)can be evaluated from a regression of flux versus the reciprocal of the ponded radius.

An alternative method for the determination of K(θ) was developed, in which a statistical model in a more general form with an additional unknown parameter is used in conjunction with the one-step outflow laboratory method. Previous results were reformulated on the basis of the Brooks and Corey model to obtain the cumulative outflow data by semi-analytical formulae.(Valiantzas & Sassalon,1991)

A method to estimate soil water diffusivity from experimental absorption data was explored by Warrick(1994). The purpose was to directly match measured sorption curves to scaled forms of the solutions and thus provide consistent hydraulic properties. Evaluation of such parameters is critical to prediction of water and solute transport within vadose zone. The results gave two independent relationships of the parameters. The soil water diffusivities were compared for the estimated parametric relationships.

Time domain reflectometry (TDR) probes installed vertically at the soil surface beneath a constant rate rainfall simulator was used to measure cumulative water storage and the soil's unsaturated hydraulic conductivity. The slope from linear regression of water storage on time before any applied water infiltrates to the bottom of the TDR probe gives an estimate of the local infiltration rate. Local infiltration rates measured by TDR in the field were plotted against the corresponding local steady state water contents to give an estimate of the soil's unsaturated hydraulic conductivity over a range in water content of 20% using only two applied rainfall rates. (Parkin *et al*, 1995).

Two important soil hydrologic properties viz. soil water retention and unsaturated hydraulic conductivity, were determined under simulated rainfall conditions. The h-θ relationship determined using rainfall simulator infiltrometer was used as input to van Genuchten's model for determining K-θ relationship. The results obtained showed similar trends as those obtained with infiltration profile method.(Mohanty & Singh,1996)

2.4 Richards' equation

It is widely recognized that the flow of water through an unsaturated soil is governed by spatial and temporal variations in the energy status of soil water. The search for methods through which to quantify this concept in a rigorous fashion has long been a principal activity in soil physics. The isothermal flow of water through an element of non-deformable, isotropic, homogeneous soil can be expressed mathematically with the Buckingham-Darcy law (Swartzendruber,1969):

$$J_{mw} = - K\nabla\phi_{\omega} \quad \text{--------------} \ (*)$$

where

J_{mw} - the water mass flux vector

K - the hydraulic conductivity

ϕ_w- the total potential of soil water

The gradient of ϕ_w is equal to the sum of the gradient of the chemical potential of soil water and that of the gravitational potential.

The derivation of a transport equation for soil water based on the Buckingham-Darcy flux law was accomplished first by L.A.Richards. This transport equation, known as the Richards' Equation (Swartzendruber,1969) may be found after combining equation(*) with the following expression:

$$-A\partial \psi_{m}/\partial t = \nabla J_{mw}$$

where

ψ_m - the soil water matric potential

A - $(\partial r_w/\partial \psi_m)_{T.P.Pa}$; the water capacity

ρ_w - the mass of water per unit volume of soil

T - absolute temperature

P - the applied pressure

P_a - the pressure of soil air

This equation, which can be generalised to water movement even in anisotropic deformable soils, is called the fundamental Richards' Equation.

2.4.1 The fundamental Richards' equation

The Richards' Equation in its original formulation(Richards,1931) is a partial differential equation that describes the time development and spatial variation of the soil water matric potential in a rigid, isotropic, homogeneous, unsaturated soil. This situation is expressed as follows:

$$dU_w = TdS_w + d\psi_m + gdz$$

where

U_w - a partial specific internal energy of water in the soil

S_w - entropy

ψ_m - soil water matric potential

g - gravitational acceleration

The differential balance law is given as:

$$\rho_w(TdS_w/dt + d\psi_m/dt) = \nabla(v_w.t^w) - (v_w \nabla t^w) + \rho Q^w - \nabla q_w + f_w$$

Further reduction and transformation of the equation results in:

$$\partial \psi_m/\partial t = -(\partial \psi_m/\partial \rho_w)_{T,P,Pa} \nabla . J_{mw}$$

where

$$J_{mw} = r_w v_w$$

This is the same as the fundamental Richards' Equation.

2.4.2 Solution of Richards' Equation

Conventional inspectional analysis was applied to the one-dimensional Richards' Equation to provide a unified classification scheme for three macroscopic scaling approaches which had been used to describe soil water flow phenomena under laboratory or field conditions. It was shown that the scaling parameters and similarity groups developed in an inspectional analysis of the Richards' Equation depend on the boundary and initial conditions imposed as well as on the special hypothesis invoked. (Sposito & Jury,1985)

Smith(1983) showed that using the Richards' equation in the Fokker-Planck non-linear diffusion form, unsaturated soil water flow may be treated as a diffusion-convection wave process. The mathematical development is as follows:

Description of unsaturated vertical flow commonly neglects the flow of air and combines the differential conservation of mass equation,

$$\partial \theta/\partial t + \partial \theta/\partial z = 0$$

with flux q, described by Darcy's law,

$$q = - K(\theta)[\partial \psi/\partial z -1],$$

to obtain Richards' equation:

$$\partial \theta/\partial t - \partial /\partial z(K\partial \psi/\partial z) + \partial K/\partial z = 0$$

where

θ - water content

z - depth, downward from the surface

$K(\theta)$ - hydraulic conductivity

t - time

ψ- soil water capillary potential

For analogy to a diffusion-convection equation, this equation is often transformed by defining 'diffusivity' D as

$$D(\theta) = K(\theta)\partial \psi/\partial \theta$$

to obtain

$$q = -D(\theta)\partial \theta/\partial z + K(\theta)$$

and

$$\partial \theta/\partial t - \partial /\partial z \, [D(\theta) \, \partial\theta/\partial z - K(\theta)] = 0$$

If $\partial\theta/\partial z$ is assumed a function of θ alone, the unsaturated flow equation may be solved by the method of characteristics, and when $\partial\theta/\partial z$ becomes sufficiently small, the Peclet number is assumed large enough to treat unsaturated flow kinematically. The method was compared to the complete solution to Richards' equation for a complex rain pattern and found to predict well the location of deeper moving fronts and also general θ patterns.

Sensitivity analysis is one of the tools available for analysing the effects of parameter uncertainity and soil heterogenity on the transport of moisture in unsaturated similar porous media. Kabala and Milly(1990) found that direct differentiation of the discretized Richards' Equation with respect to parameters defining spatial variability lead to linear systems of equations for elementary sensitivities that were readily solved in conjunction with the original equation. An empirical criterion was proposed for selection of time-step of simulation in the finite difference solution of explicitly linearized head form of non-linear Richards' Equation, under time-variant rainfall and irrigation intensity condition. The criterion was satisfactory in the exercises conducted, including a field evidence simulation exercise.(MohanRao *et al*, 1991)

Barry *et al*(1992) derived a new solution satisfying Richards' equation. The solution, which may be applied for infiltration or capillary rise, is valid for the condition of an arbitrary moisture tension imposed at the soil surface. The solution applies when the form of the soil moisture characteristic curve is a particular weighted integral of the gradient of the unsaturated hydraulic conductivity. Thus if the soil moisture characteristic curve is selected a priori, then this condition determines the hydraulic conductivity. Using the correspondence between Richards' equation and the convection-dispersion equation with a non-linear solute adsorption isotherm, a new exact solution for adsorptive solute transport is derived.

Lehua Pan and Wierenga, in 1995, presented a new approach to solve Richards' equation. It introduced a non-linear transformed pressure, P_t as the dependent variable. With the modified Picard method. The new approach was compared to, and contrasted with, two efficient existing methods: the ϕ-based transformation method and the h-based modified Picard method. The results showed that the new method offers excellent CPU efficiency and, unlike the h-based method, is numerically robust for all cases of variably saturated, heterogeneous media and pressure or flux type boundary conditions.

2.5 Rainfall simulators

Rainfall simulators have been used to accelerate research in soil moisture flow, soil erosion and runoff from agricultural lands. Meyer(1965) defined simulated rainfall as water applied in a form similar to natural rainfall. Simulated rainfall aids in creating a given rainstorm at a desired time and location. It enables investigators to obtain runoff data in a relatively short period of time.(Bubenzer & Meyer,1965)

In accordance to the report by Mutchler and Hermsmeier(1965),the rainfall simulators use one of the following drop forming methods:

1. Hanging yarns
2. Nozzles
3. Tubing tips

They reported the working of hanging yarn type simulators. For hanging yarn type simulator, a muslin cloth was laid loosely on a chicken wire screen so that depressions were formed in the cloth at each screen opening. A piece of yarn was attached to the cloth at each depression. Water applied as a spray to the cloth collected at the depressions and travelled down the hanging yarns to form drops. A low-cost, highly portable rainfall simulator infiltrometer was developed at IIT, Kharagpur for infiltration, runoff and erosion studies. Raindrops were produced at the end of wire loops inserted in capillary holes drilled through a 10 mm thick circular perspex sheet. The characteristics of the simulated rainfall were evaluated for intensities of 100 mm/h and 200 mm/h.(Bhardwaj & Singh,1992)

A simple and cheap rainfall simulator employing hypodermic needles as drop formers was developed by Choudhury *et al*(1978). Rainfall of varying intensity and drop size was produced by a combination of different needle sizes and water pressure. Floyd(1981) developed a rainfall simulator for use in small plot field experiments. The design was based on an oscillating boom housing a series of Vee-jet nozzles to which the water supply was periodically interrupted. The intensity of rainfall was 27 mm/h with a coefficient of variation of 11.3%. The drop size distribution approximated to that of natural rainfall of the same intensity but was deficient in drops of diameter greater than 3.5 mm. A portable rainfall simulator was developed for use in field infiltration experiments. The simulator, constructed of standard PVC pipe, takes a set up time of approximately 10 minutes. Eight sprinkler heads are attached to the top of the frame at an elevation of 1.83 m. One of the positive features of the rainfall is that the intensity of rainfall can be varied.(Bruce *et al*,1996)

2.5.1 Advantages and limitations of simulated rainfall

Meyer(1965) presented the advantages of simulated rainfall.

1. More rapid results can be obtained by applying selected simulated storms at selected treatment conditions.
2. Results from a few simulated storms at selected conditions often provide desirable informations.
3. Various measurements and observations which are difficult during natural rainstorms may be readily obtained during simulated storms.
4. Simulated rainfall is readily adaptable to highly controlled laboratory research.

The limitations of simulated rainfall as a research tool was indicated by Meeh(1965). The limitations may be of two types: modelling limitations and operating limitations.

Modelling limitations:

Soil and water research problems are usually associated with natural conditions of weather and soil. It is difficult to simulate factors like wind, light, temperature, humidity, vegetative influences etc.. Measurements of soil loss, water loss and infiltration are difficult to extrapolate to field conditions and natural rain.

Operating limitations:

The nature of most rainfall simulators limit the study to small plots. The need for an adequate supply of water in the vicinity of experimental plots limits the location of the work.

2.6 Rainfall characteristics

Rainfall can be characterized by the variations in raindrop sizes and the differences in impact velocities. To describe the physical characteristics of rainfall adequately, mathematical relationships were developed and tested with raindrop size distributions, fall velocity and their variations. The raindrop characteristics were found to be closely related to rainfall intensity. Rainfall parameters such as number of drops, momentum and energy were also defined to be a function of rainfall intensity.(Park *et al*,1983)

2.6.1 Raindrop deformation and fall velocity

Raindrop deformation

The deformation of raindrops in stagnant air depends upon their size. Small drops of less than 3 mm in diameter are nearly spherical at terminal velocity. Because of deformation, raindrops at terminal velocity may not exceed a certain threshold size. The threshold raindrop size has not been defined. Gunn and Kinzer(1949) reported that raindrops greater than 6 mm in diameter are unstable and easily broken into smaller sizes. However, raindrop size measurement data have contained dropsizes as large as 7 mm in diameter(Beard,1976).

Terminal velocity

The terminal velocity of a raindrop may be determined from the equation

$$C_D = (4/3)(\rho_w/\rho_a - 1)gd / v_T^2$$

where

C_D - the drag coefficient

ρ_w - the density of the raindrop

ρ_a - the air density

g - gravitational acceleration

d - diameter of the drop

v_T - the terminal velocity of a drop

For the Stoke's range, where the drag coefficient is inversely proportional to the Reynold's number(R_e<0.1), the terminal velocity is defined as

$$v_T = (\rho_w - \rho_a) g d^2 / 18\mu$$

Fall velocity

A raindrop released to air at initial velocity requires a certain fall distance before it reaches the terminal velocity. When the fall distance is insufficient, the impact velocity of a raindrop is less than terminal velocity.

To determine the distance that is required to obtain a given fall velocity, the following equation may be used.(Park *et al*,1983)

$$y = v_T^2/4g \ln[v_T^2 - v_0^2)/(v_T^2 - v_S^2)]$$

where

y - the travel distance, m

v_T - terminal velocity, m/s

v_0 - initial velocity, m/s

v_S - fall velocity, m/s

2.7 Measurement of rainfall characteristics

2.7.1 Droplet size

There are various methods for determining the droplet size viz. stain method, photographic method, momentum method, immersion method and flour pellet method.

The flour pellet method as described by Kohl(1974) consists of calibrating plain flour by dropping water drops of known diameter into trays containing about a 25 mm thick layer of sifted uncompacted flour. The flour pellets are oven-dried at 110°C. The dried pellets are weighed and a mass ratio determined. The flour trays are then exposed to natural or artificial rain and the drop sizes determined via the calibration curves. The flour pellet method does not require any special equipment to determine the drop size.

2.7.2 Uniformity of rainfall

Uniformity coefficient is a measure of the degree of uniformity of rainfall. The coefficient is computed from the field observations of the depth of water caught in open pans placed at regular intervals within the area. It is expressed by the equation developed by Christiansen(1942)

$$C_u = 100(1.0 - \Sigma X/mn)$$

where

m - average value of all observations, mm

n - total number of observation points

X - numerical deviation of individual observations from the average application rate, mm.

Materials and Methods

MATERIALS AND METHODS

The materials used and the methodology adopted for the study are discussed in this chapter.

3.1 Design and fabrication of a rainfall simulator

The design of the rainfall simulator is based on that of Bhardwaj *et al* (1992). Figure 3 shows the schematic diagram of the rainfall simulator. The portable rainfall simulator comprises of a drop forming mechanism mounted on a supporting frame. Plate1 shows the overall view of the experimental setup.

3.1.1 Drop forming mechanism

The drop forming mechanism consists of a tank with a perforated bottom. (Plate2). A 18G GI sheet of size 930 mm × 930 mm , with 1 mm holes drilled at a spacing of 13 mm, forms the base of the tank. The holes are provided with counter sink at the base. Copper wire of gage 20 is suspended through each hole by bending its upper end and making a loop at the lower end. The diameter of the loop is so selected such that the simulated water droplets have a size similar to that of natural rain drops. A 18G GI sheet was used for fabricating the sides of the tank with 250 mm height. The tank was provided with a water level indicator and an overflow pipe. (Plate 3)

Head of water in the tank is varied to get the desired intensity of simulated rainfall. The head is maintained during an experiment by means of a float valve.

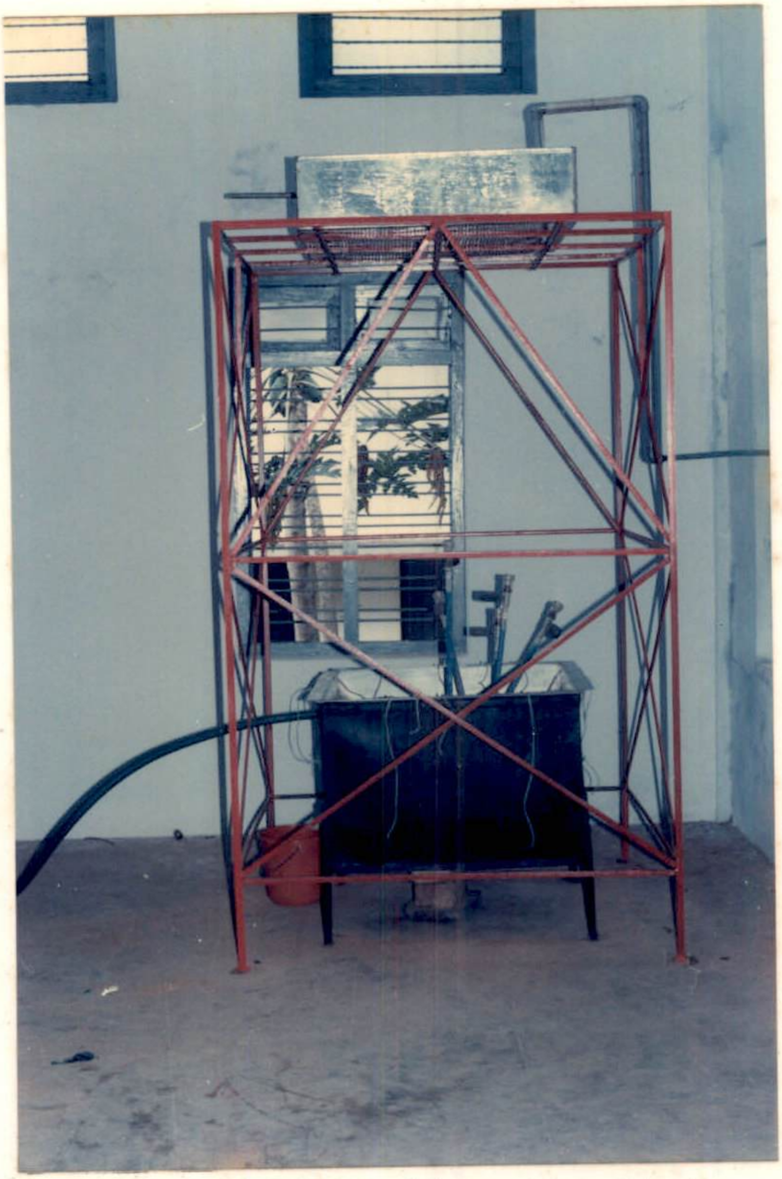
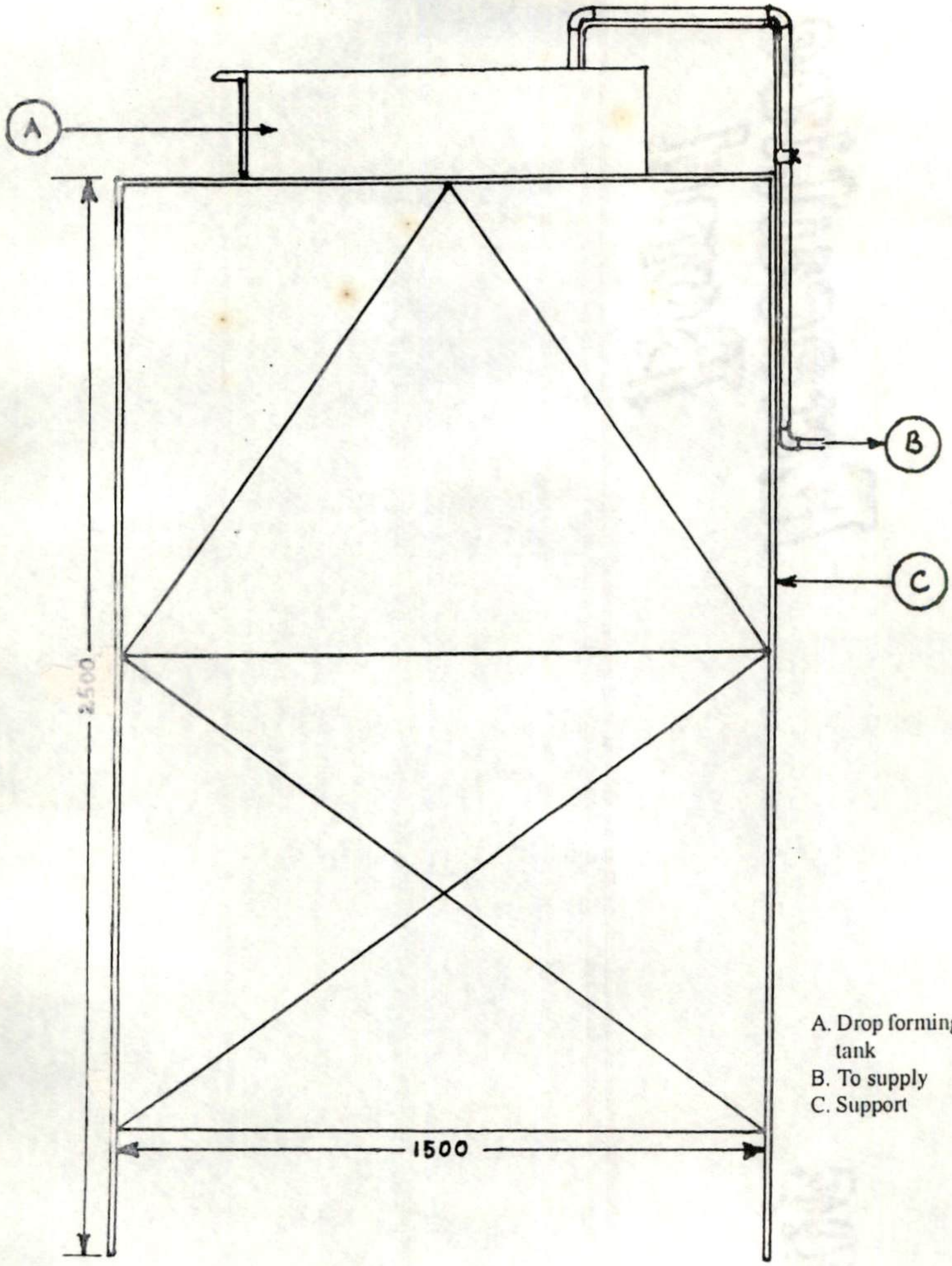


Plate 1.	Overall view of the experimental setup

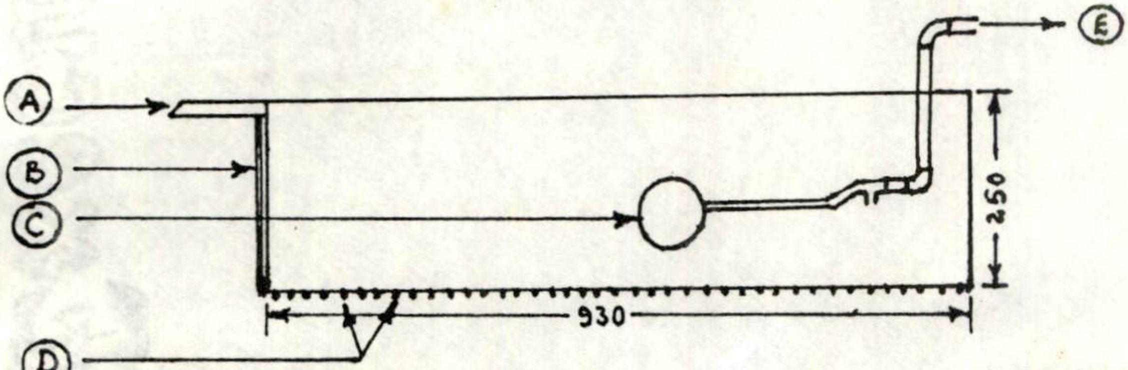


A. Drop forming tank
B. To supply
C. Support

All dimensions in mm

Fig. 3 Schematic diagram of rainfall simulator

Plate 2. Close view of the wire loops in the drop forming tank



A. Overflow pipe
B. Water level indicator
C. Float
D. Wire loops
E. To supply

All dimensions in mm

Fig. 4 Schematic diagram of drop forming mechanism

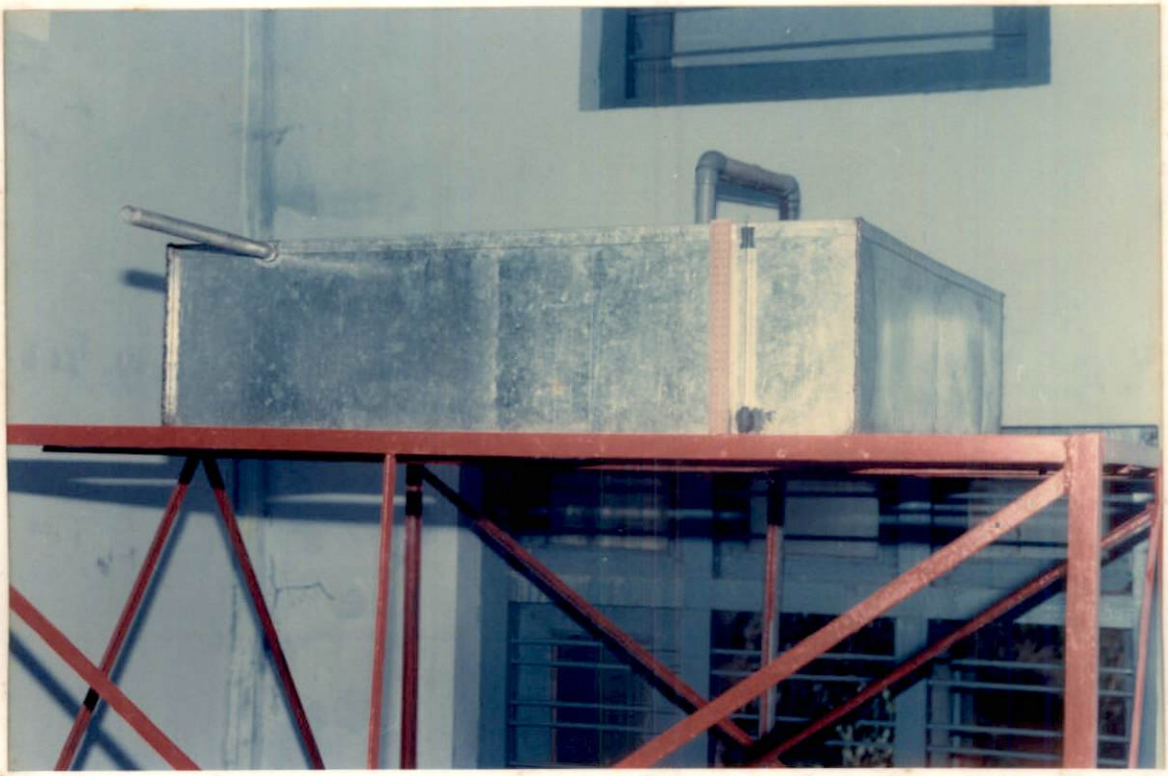


Plate 3. Accessories of the drop forming tank

The level of the float can be adjusted to maintain different heads of water and hence, to get different intensities. The head of water ranged from 5 to 22cm. A schematic diagram of the drop forming mechanism is shown in fig. 4.

3.1.2 Design of supporting frame

The supporting frame of the rainfall simulator is made of MS angle iron 25mm × 25mm × 3mm. Four angles, 2.5 m long, constitute the legs of the structure. The structure forms a square grid 1.5 m × 1.5 m at the upper end, also made of MS angle iron. The grid has further support with channel sections. This results in the division of the square grid into a number of smaller grids. It is on these grids that the drop-forming mechanism rests. The frame is further strengthened by trusses.

3.2 Design of the soil trough

The soil trough is a GI tank having an inner dimension of 900 mm × 600 mm (Plate 4). The tank has provision for collecting the surface and subsurface flow of water. There is a channel of width 5 mm at a depth of 120 mm from the top of the tank. Laterite soil is filled in the trough such that the channel collects the surface flow. The subsurface flow of water is collected through an outlet at the bottom. Two piezometers are installed, one each at the upper and lower ends of the soil trough. A frame of height 250mm supports the soil trough. It is constructed by welding 25mm × 25mm × 5mm MS angle iron pieces.

Plate 4. The soil trough

3.3 Installation

The experimental setup was installed in the Soil and Water Laboratory at KCAET. The soil trough was placed such that the simulated water drops fall in the trough. To vary the inclination of the trough, one side of the trough was raised. The water supply line was connected to the simulator tank through a float valve. Figure 5 shows the trough at a given inclination.

3.4 Testing of rainfall simulator

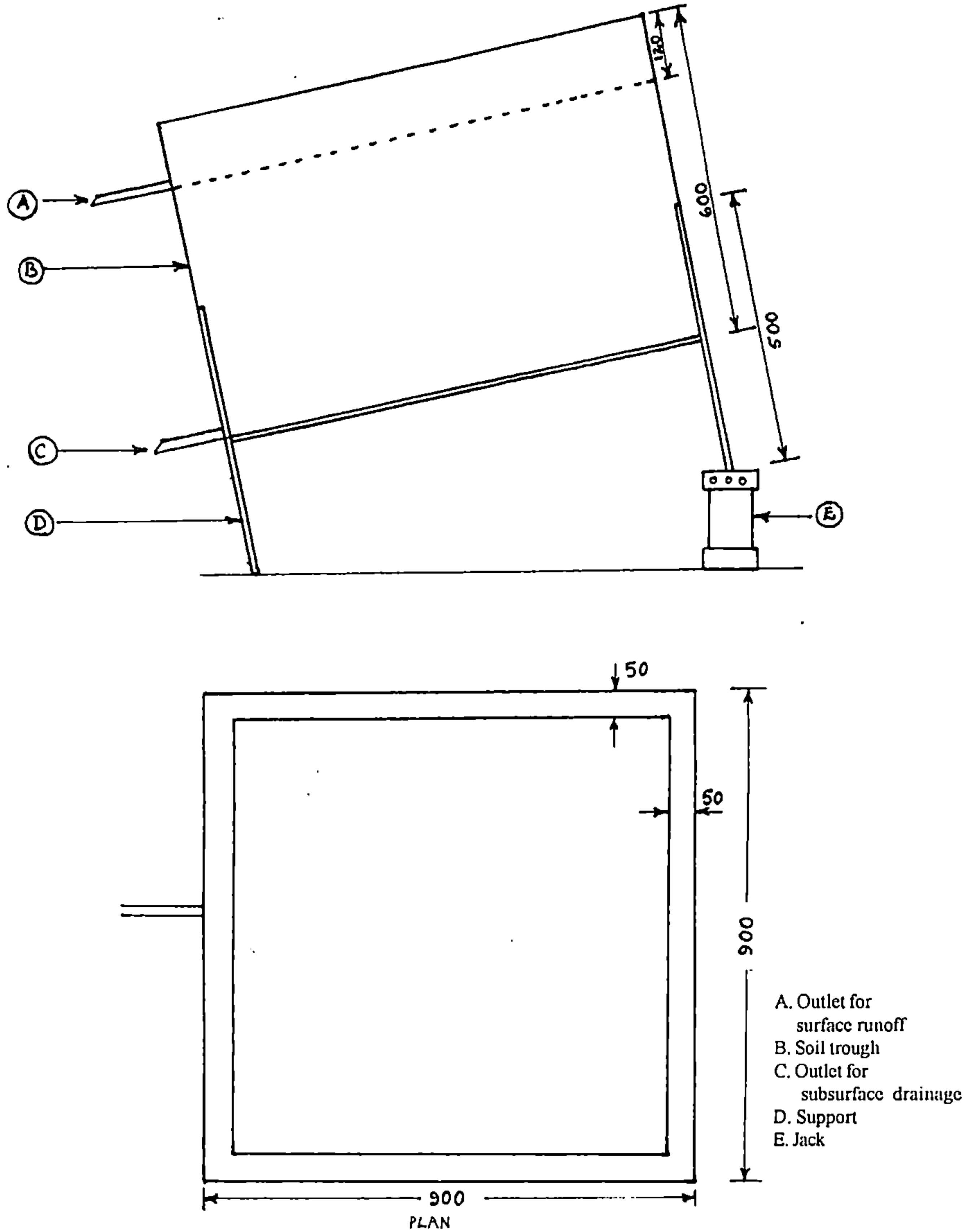
The rainfall simulator was tested for different intensities by changing the head of water in the tank. The head of water was changed by varying the position of the float valve.

3.4.1 Determination of simulated rainfall intensity

The rainfall simulator is operated at a particular head. Five catch cans of 10cm diameter are placed below the rainfall simulator. The simulator is run freely to attain uniformity. Four cans are placed at the corners and one at the centre. Then water is collected in the catch cans for a duration of 15 minutes. The volume of water collected in each can is measured and noted. This volume is then converted to equivalent depth. The process is repeated for different heads of water in the tank.

3.4.2 Determination of drop size

The drop size was determined using the flour-pellet method. This method consists of calibrating plain flour pellets formed by dropping water drops of known



All dimensions in mm

Fig. 5 Schematic diagram of inclined soil trough diameter into trays containing about a 25mm thick layer of sifted uncompacted flour. The flour pellets are dried, weighed and a mass ratio determined.

3.4.2.1 Calibration

A syringe and a set of hypodermic needles of sizes 16G, 18G, 24G and 26G were taken. A particular volume of water was filled in the syringe. The number of droplets produced by each needle for that volume was found. It was repeated five times for each needle. The diameter of droplets produced by each needle was found by using the following formula,

$$d = \sqrt[3]{6000\frac{V}{n}}$$

where

d - drop diameter, mm

V - volume of water in a syringe, ml.

n - number of droplets produced

A plate was filled with dry sifted flour. Droplets were allowed to fall drop by drop by pressing the syringe slowly. The pellets formed were oven dried for 24 hours at 110°C. The dried pellets were separated from the flour by sieving and the mass of the pellets produced by each droplet size was determined. The same process was repeated thrice.

The relationship between the droplet diameter (d) and droplet mass (M) was generated by regression analysis.

3.4.2.2 Determination of simulator raindrop size

Plates of 22cm diameter and 3.5cm depth were filled with dry sifted flour. The rainfall simulator was operated at a particular head. The plate was placed below the simulator so that the simulated droplets fell on the flour. The flour with the pellets was dried at 110°C for 24 hours. The pellets were sieved and weight of the known number of pellets formed was taken. Using the calibrated relation, the mean droplet diameter collected in the plate was determined. This procedure was repeated for different rainfall intensities.

3.4.3 Uniformity of rainfall

Five catch cans of 10 cm diameter were placed beneath the rainfall simulator. The cans were so arranged that one was at the centre and the others at the four corners. The soil trough remained covered with a polythene sheet throughout the test. After maintaining the head at the desired level, equivalent depth of water collected in each can during 15 minutes duration was determined. The coefficient of uniformity is calculated as

$$C_U = 100(1.0 - \Sigma X/mn)$$

where

C_U - Coefficient of uniformity, %

m - average depth of water, mm

n - number of catch cans

X - numerical deviation in depth of water of each can from the average depth, mm

The uniformity coefficient also was determined for different rainfall intensities.

3.5 Determination of basic soil properties

Soil was collected upto a depth of 50 cm from the campus at Tavanur. This soil comes under the Angadippuram series of laterite soil.

3.5.1 Moisture content

The initial moisture content of the soil was determined by oven drying method. The moisture content is given by

$$w = 100(M_2-M_3)/(M_3-M_1)$$

where

M_1 - mass of the container, g

M_2 - mass of container + wet soil, g

M_3 - mass of container + dry soil, g

3.5.2 Textural analysis

Textural analysis of the soil was done by determining the particle-size distribution. The analysis was performed in two stages: (1) sieve analysis and (2) sedimentation analysis.

3.5.2.1 Sieve analysis

A representative sample of the soil was dried in the oven at 104°C for 24 hours. From the dried soil, 500 g was taken for the analysis. The analysis consisted of coarse and fine analyses. A set of 100 mm, 63 mm, 20 mm, 10 mm and 4.75 mm sieves were used for coarse analysis. The weight of the materials retained on each sieve were noted. For fine analysis, 2 mm, 1 mm 500, 425, 300, 212, 150 and 75 micron IS sieves were used. First, the silt and clay particles were separated by washing the soil sample through a 75 micron sieve. The portion of soil retained on 75 micron sieve was dried and subjected to fine sieve analysis. The set of sieves were placed one above the other on a mechanical sieve shaker such that the 2 mm sieve containing the soil sample was on the top and the 75 micron sieve at the bottom, with a receiver below it. The sieve shaker was operated for about 10 minutes and the portion retained on each sieve weighed and noted.

The percentage of soil retained on each sieve is calculated on the basis of the total mass of the soil sample taken and from these results, percentage passing through each sieve is calculated.

3.5.2.2 Hydrometry

The hydrometer analysis is based on Stokes' law, according to which the velocity at which grains settle out of suspension, all other factors being equal, is dependent upon the shape, weight and size of the grain. The hydrometer and the sedimentation jar are calibrated before the start of the analysis.

After calibration, a relationship was established between effective depth (H_e) and the density readings (R_h) of the hydrometer. The necessary corrections to be made were also determined. Fifty grams of the soil was first treated with hydrogen peroxide solution to remove organic materials. Next, the soil was treated with 0.2N hydrochloric acid to remove calcium compounds, if any.

After washing the mixture with warm water till there was no acid reaction to litmus, the oven dried soil was weighed and 100 ml dispersing agent (Sodium hexametaphosphate) was added. The soil suspension was washed through a 75 micron IS sieve; the mass of those passing through the sieve was transferred to a 1000ml measuring cylinder, making up the volume accurately to 1000ml. The hydrometer was immersed in it and the readings were taken at different time intervals. The percentage finer(N) was determined and a particle size distribution curve was plotted.

3.5.3 Determination of bulk density

The core cutter method was used to determine the bulk density and dry density of the soil. A core cutter, consisting of a steel cutter, 10 cm in diameter and 13 cm high and a 2.5 cm high dolly was driven in the soil with the help of a rammer. The cutter, containing the soil, was dug out of the ground, and the mass of the soil in the cutter was found out. the bulk density was determined as follows:

Bulk density =(mass of soil in cutter)/(volume of cutter)

The moisture content of the excavated soil is found and then the dry density is computed as:

$$\rho_d = \rho/(1+w)$$

where

ρ_d- dry density, g/cm^3

ρ- bulk density, g/cm^3

w - moisture content (ratio)

3.5.4 Consistency limits

The liquid and plastic limits of the soil were determined.

3.5.4.1 Liquid limit

The liquid limit was determined with the help of the standard liquid limit apparatus designed by Cassagrande. About 120 g of the specimen passing through 425 micron sieve was mixed thoroughly with distilled water to form a uniform paste. A portion of the paste was placed in the cup of the Cassagrande apparatus and spread into position and a groove was cut in the soil pat using the Cassagrande BS tool. The number of blows required for the two parts of the soil sample to come into contact at the bottom of the groove was noted. The water content was determined by taking soil sample from near the closed groove and subjecting it to oven drying method. A graph was plotted between number of blows as abscissa on a logarithmic scale and the corresponding water content as ordinate. The water content corresponding to 25 blows was taken as the liquid limit.

3.5.4.2 Plastic limit

The soil specimen, passing through 425 micron sieve was mixed thoroughly with distilled water so that the soil mass could be easily moulded with fingers. A ball was formed of 10g of the soil mass and rolled between the fingers and a glass plate into a thread of uniform diameter. When the diameter was 3mm, the soil was remoulded again into a ball. The process of rolling and remoulding was repeated till the thread starts just crumbling at a diameter of 3mm. The water content of the crumbled threads was determined. The test was repeated twice with fresh samples. The plastic limit was taken as the average of the three water contents.

3.6 Monitoring saturated and unsaturated flow processes
3.6.1 Hydraulic head

The hydraulic head is the sum of gravitational and matric suction heads. The gravitational head is characterized by the distance from the datum to the point under consideration. Matric suction was measured using a tensiometer. Four tensiometers were installed at 10cm intervals in the soil plot. The tensiometer readings gave the matric suction which was then converted into matric suction head (cm of water).

3.6.2 Volumetric moisture content

The moisture contents at varying depths with an interval of 10 cm were determined gravimetrically. Gypsum blocks were also installed at four depths and at three different positions in the soil plot. Figure 6 shows the arrangement of the gypsum blocks. Volumetric moisture content is given by

$$m.c_{(vol.)} = m.c_{(grav.)} \times \text{bulk density}$$

3.6.3 Calculation of soil moisture flow and hydraulic conductivity

The soil moisture flux through each depth increment was calculated for particular time. Hydraulic conductivity at each depth and for different moisture

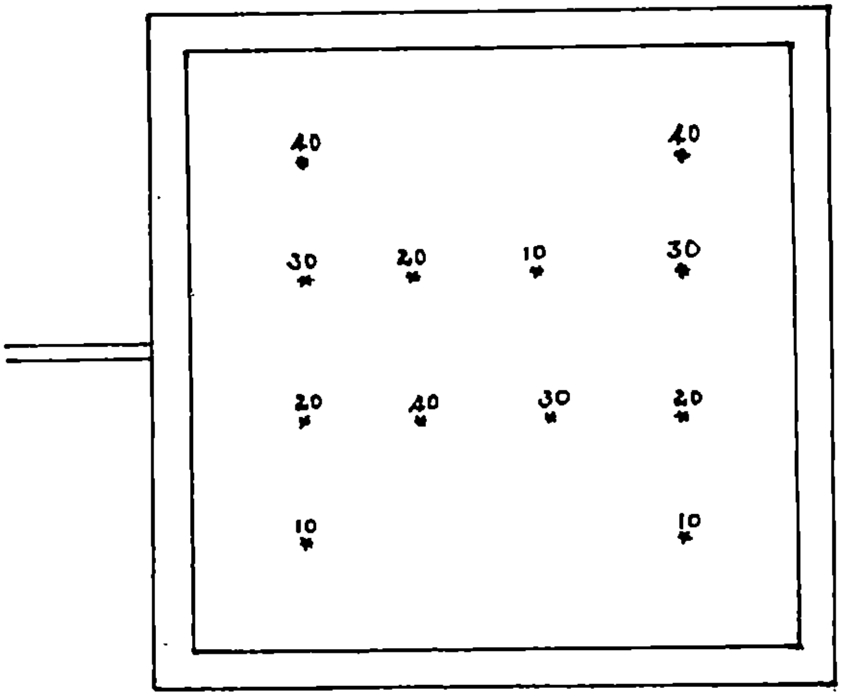


Fig. 6 Positions of gypsum blocks in soil trough

contents was calculated by dividing fluxes by the corresponding hydraulic gradient values.

The general equation describing the flow of water in a vertical soil profile is

$$\partial\theta/\partial t = \partial/\partial z\,[K(\theta)\,\partial H/\partial z]$$

where

θ - volumetric wetness (measurable by gypsum block)

t - time

z - the vertical depth

K - hydraulic conductivity, which is a function of soil wetness

H - the hydraulic head

Integrating, we get

$$\int_o^z \partial\theta/\partial t\, dz = (K\partial H/\partial z)_z$$

$$\text{or, } \partial\theta/\partial t\, Z = (K\partial H/\partial z)_z$$

Here, z is the soil depth to which the measurement applies.

$$(\partial W/\partial t)_z = K(\partial H/\partial z)_z = q$$

Here, W is the total water content of the profile to depth Z, ie,

$$W = \int_o^z \theta dz$$

$$\text{Finally, } K(\theta) = q / (\theta H/\theta z)$$

3.6.4 Soil moisture balance

The moisture balance in the soil plot is an itemized statement of gains, losses and changes of storage of moisture during a specified period of time. The soil moisture balance equation is

$$P - (R + D) = \Delta S$$

where

P - the applied rainfall

R - surface runoff from the soil plot

D - drainage (subsurface) of water

ΔS - change in storage of moisture

3.7 Runoff generation

The surface and subsurface flow of water for different rainfall intensities and at 10% and 20% slopes were measured. The subsurface flow was monitored periodically and the volume of runoff at each time was recorded.

Results and Discussion

RESULTS AND DISCUSSION

A rainfall simulator and soil trough were developed and installed in the Soil and Water Laboratory of KCAET. The rainfall simulator was calibrated for different intensities by varying the head of water in the tank. After calibration, various experiments were conducted to evaluate the hydraulic properties of laterite soil, the results of which are discussed in this chapter.

4.1 Testing of rainfall simulator

4.1.1 Intensity of rainfall

The rainfall simulator was tested for various intensities by changing the head of water in the drop forming tank. The head of water was varied in the range of 3 cm to 22 cm by altering the level of the float valve. At 3 cm depth of water the intensity of rainfall was 12.8 mm/h. It was found that intensity increased with the rise in head. A maximum intensity of 285.6mm/h was obtained for a head of 22 cm. The intensity corresponding to different values of head are given in Table 1. In accordance to the results, intensity is given by

$$I = 35.9489 - 7.13048\, H + 0.829516\, H^2$$

where

I - intensity of rainfall, mm/h

H - head of water, cm

Table 1. Variation of intensity of rainfall with the head of water in tank

Sl. No.	Head of water (cm)	Intensity of rainfall (mm/h)
1.	3.0	12.80
2.	5.0	20.00
3.	6.0	26.06
4.	9.0	40.00
5.	10.0	49.11
6.	12.0	77.90
7.	14.0	90.00
8.	16.0	124.50
9.	22.0	285.60

4.1.2 Simulated rain drop size

Drop size determination was done by the flour-pellet method. The method consisted of calibration and determining the drop size. After calibration, a relation between droplet diameter 'd' and mass of the droplet 'M' was established; using which the drop size was determined. Figure 7 shows the calibration curve. The relation is given by

$$d = 13.2763\ M^{0.254904}$$

where

d - diameter of drop, mm

M - mass of drop, g

Table 2 gives the variation in drop size with intensity. The droplet size ranged from 6.21 mm to 7.01 mm. It can be seen that the simulated raindrop size remained almost constant. The simulated raindrops resembled natural raindrops of size 5.4 mm having terminal velocity of 9.29 m/s.

4.1.3 Uniformity of rainfall

The uniformity coefficient of rainfall shows the uniformity of the simulated rainfall. The uniformity coefficient at different intensities are shown in Table 3. It is found that the uniformity of rainfall increased with intensity. Maximum uniformity was 85% for an intensity of 124.5 mm/h. This may be compared with the uniformity of 63% at 12.8 mm/h.

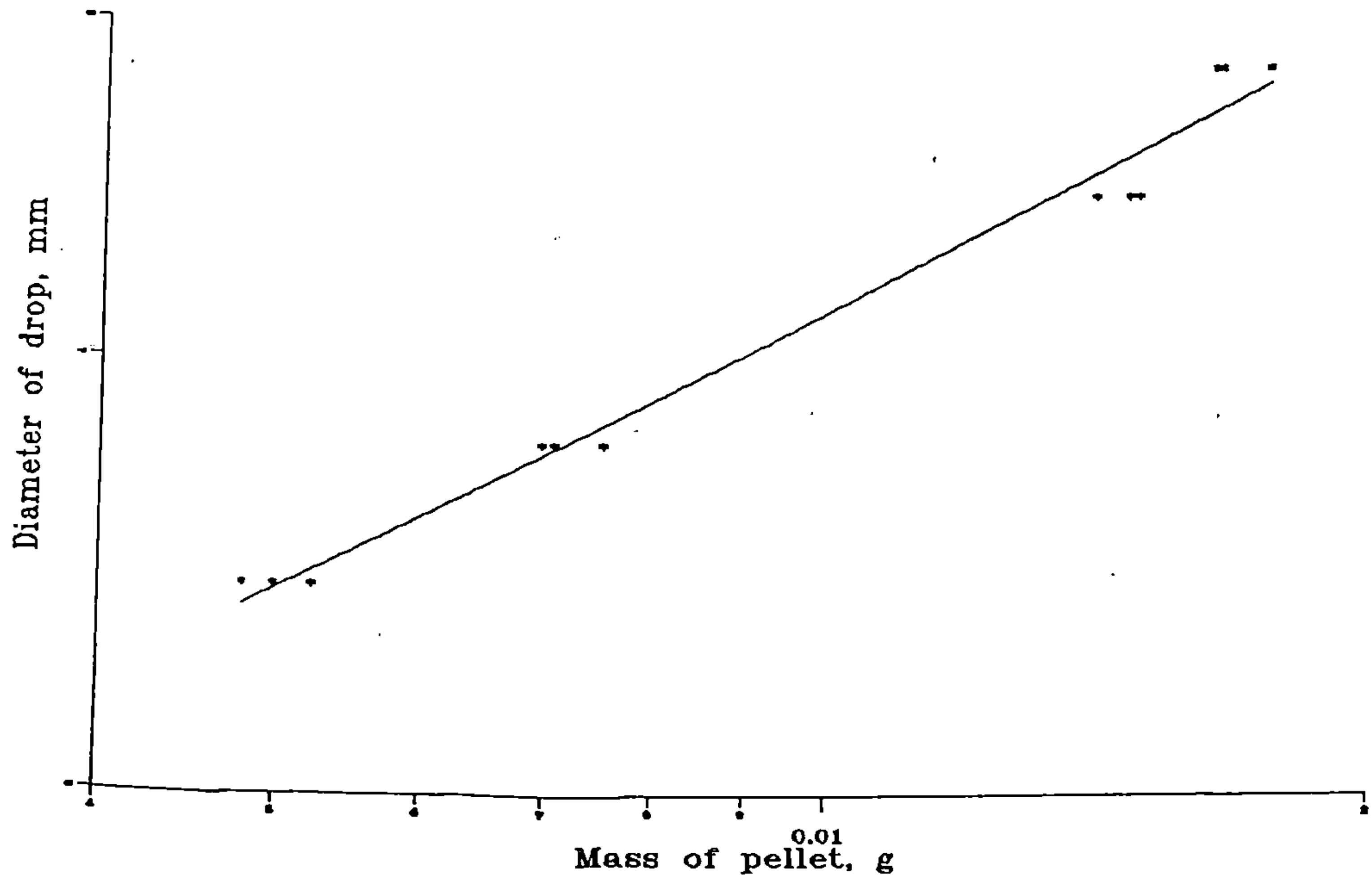


Fig.7 Calibration curve for dropsize determination.

Table 2. Variation in simulated raindrop size with intensity

Sl. No.	Intensity of rainfall (mm/h)	Diameter of raindrops (mm)
1.	20.00	6.62
2.	40.00	6.66
3.	49.11	6.70
4.	90.00	6.21
5.	124.50	7.01

Table 3. Uniformity of rainfall at different intensities

Sl. No.	Intensity of rainfall (mm/h)	Uniformity (%)
1.	12.80	63.00
2.	20.00	63.00
3.	36.06	70.00
4.	40.00	68.00
5.	49.11	70.00
6.	77.90	70.00
7.	90.00	84.50
8.	124.50	85.00

4.2 Soil properties

4.2.1 Textural analysis

The relative proportions of the different grain sizes which make up the soil mass was determined. Both mechanical (sieve) analysis and hydrometry were carried out. The particle size distribution curve is given in fig.8. The soil was found to be loamy sand with 88% sand, 4% silt and 8% clay. About 63% of the soil is gravel. From the orientation of the curve, it can be seen that the soil is coarse graded.

4.2.2 Bulk density

Bulk density was determined by using a core cutter. The soil was found to have a bulk density of 1.53 g/cm^3.

4.2.3 Consistency of the soil

Consistency limits denote the water content at which the soil mass passes from one state to the next. Experiments were conducted to evaluate the liquid and plastic limits of the soil. The soil had a liquid limit of 42.55% (Fig.9) and a plastic limit of 65.28%.

4.3 Flow processes

The readings of the piezometers installed in the soil trough showed that unsaturated flow takes place through the soil during the simulated rainfalls.

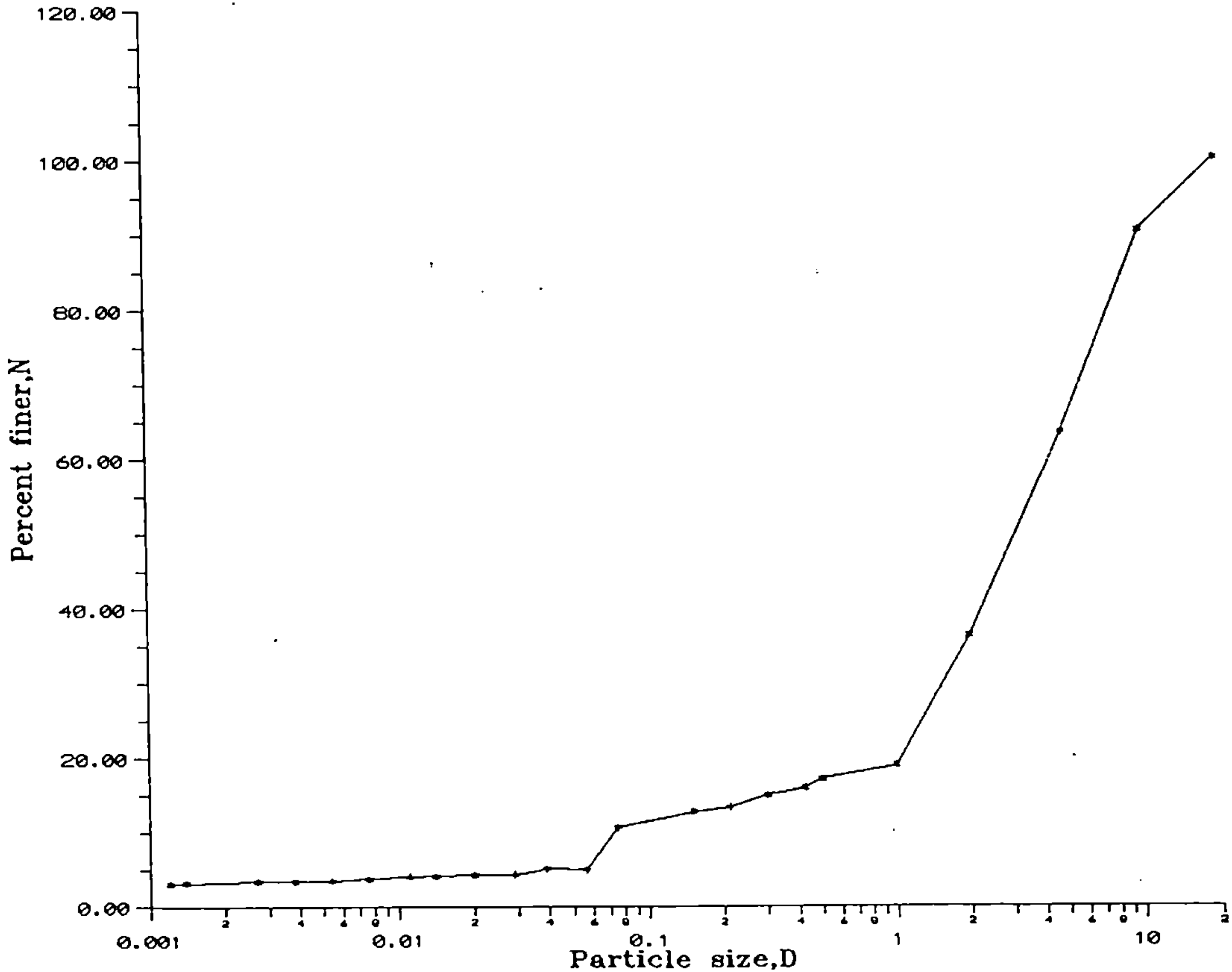


Fig.8 Particle size distribution curve.

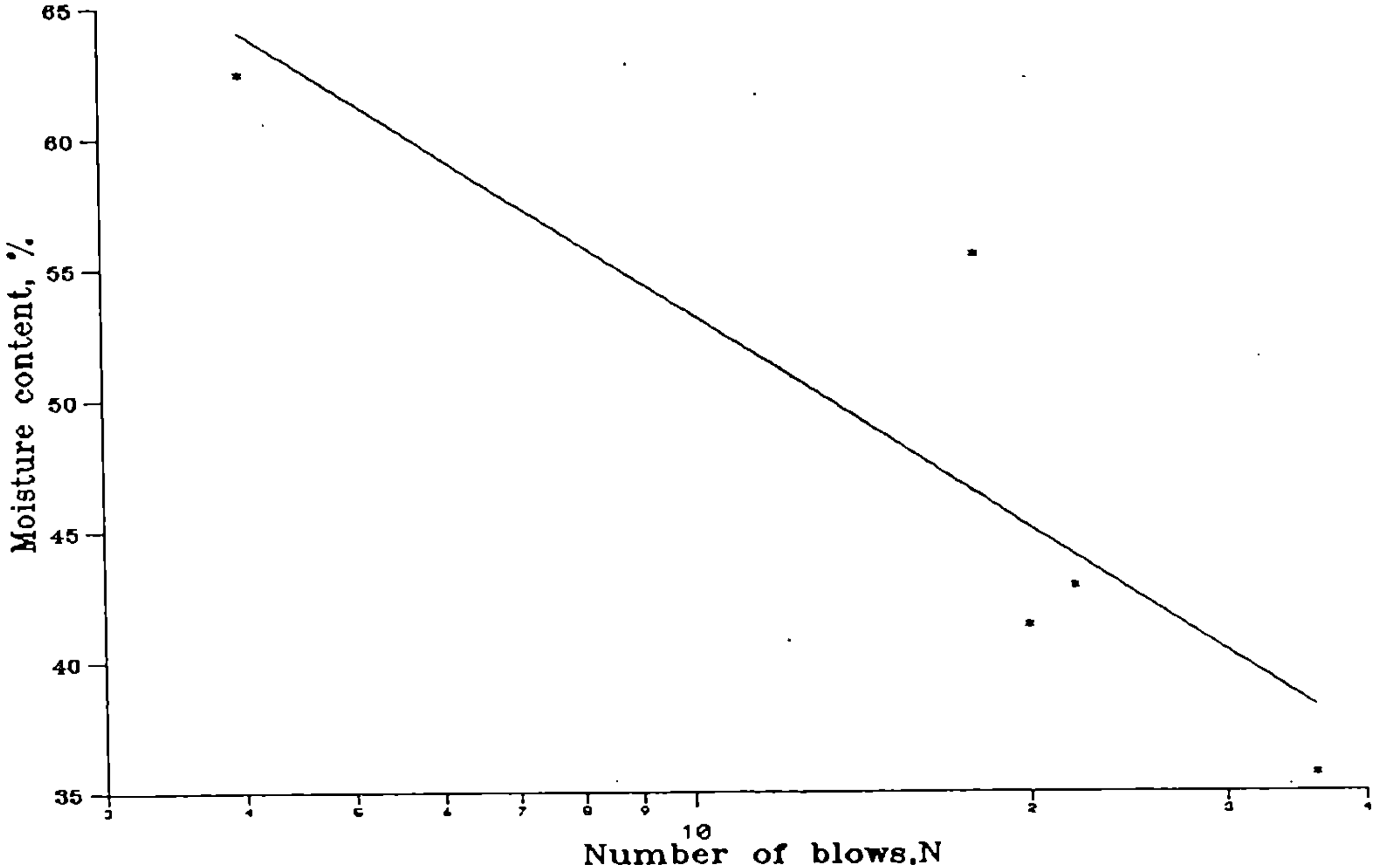


Fig.9 Flow curve for liquid limit determination.

4.3.1 Unsaturated flow

4.3.1.1 Hydraulic conductivity-volumetric moisture content relationship

The instantaneous profile method was carried out for different rainfall intensities to determine the hydraulic conductivity. The experiment was done for rainfall intensities of 20, 40, 90 and 285 mm/h. The results are furnished in tables 4-7. Soil moisture flux at different depths corresponding to different intensities is given in Appendix I. For an intensity of 20 mm/h, the volumetric soil moisture content at 10 cm depth, after an interval of one hour after wetting was 16.79%. The soil moisture content was found to deplete gradually with time. After 216 hours moisture content of the soil was reduced to 13.21%. At 20 cm depth, after one hour, the soil moisture content was 18.51%; whereas after 216 h it decreased to 13.21%. For the same intensity at lower depths, though the soil moisture content increased at first, it was found to decrease as time progressed. The increase in moisture content at the earlier stages can be attributed to the infiltration of water from the overlying layers.

When the intensity of rainfall was changed to 40 mm/h, the moisture content values at 10 cm depth of soil were 17.21% and 15.42% after 0.5 and 216 h respectively. The soil moisture content at the deeper layers were greater than that at 10 cm depth, but were depleted with the increase in time. Moisture content increased with increase in intensity. There was a corresponding increase in hydraulic conductivity also. The variation of soil moisture content with time at an

Table 4. Calculation of hydraulic conductivity at a rainfall intensity of 20 mm/h.

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂ H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	1	0.0390	0.088	0.443182	16.69
	2	0.0161	0.088	0.182955	16.79
	4	0.3990	0.088	4.534091	19.46
	23	0.0260	1.460	0.017808	16.18
	27	0.0030	2.150	0.001395	16.11
	29	0.0100	2.150	0.004651	15.97
	120	0.0009	2.150	0.000419	15.97
	144	0.00689	2.150	0.003205	15.42
	168	0.00900	2.150	0.004186	15.42
	192	0.00460	2.150	0.00214	14.68
	216	0.00918	2.150	0.00427	13.21
15 - 25	1	0.143	1.000	0.14300	18.51
	2	0.0161	1.000	0.0161	18.51
	4	0.876	1.000	0.876	18.36
	23	0.028	1.008	0.0277	18.65
	27	0.0045	0.032	0.1406	18.58
	29	0.0300	0.032	0.9375	18.19
	120	0.00111	1.006	0.0011	18.36
	144	0.00995	1.006	0.00989	17.62
	168	0.01605	1.006	0.01595	15.42
	192	0.00775	1.006	0.0077	14.68
	216	0.01528	1.006	0.01518	13.21
25 - 35	1	0.249	0.5495	0.453139	19.72
	2	0.0819	0.5495	0.149044	20.37
	4	0.898	0.5495	1.634212	18.84
	23	0.046	0.033	1.39393	21.11
	27	0.0366	0.033	1.10909	19.82
	29	0.045	0.033	1.36363	19.52
	120	0.00185	1.066	0.001735	20.19
	144	0.0115	1.066	0.010787	19.82
	168	0.01605	1.066	0.015056	17.99
	192	0.00925	1.066	0.008677	17.62
	216	0.01988	1.066	0.018649	16.52
35 - 50	1	0.3155	3.766	0.08377	19.29
	2	0.2014	3.766	0.05347	18.60
	4	1.5730	2.044	0.76956	18.27
	23	0.0810	2.044	0.03962	19.35
	27	0.1170	2.044	0.05724	18.58
	29	0.0840	2.044	0.04109	19.07
	120	0.01065	2.733	0.00389	16.52
	144	0.18130	2.733	0.06633	15.42
	168	0.03200	2.733	0.01170	17.99
	192	0.01155	2.733	0.00422	17.62
	216	0.02678	2.733	0.00979	16.52

Table 5. Calculation of hydraulic conductivity at a rainfall intensity of 40 mm/h

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	0.5	0.078000	0.776	0.100515	17.21
	4	0.011100	0.088	0.126136	16.95
	23	0.000480	1.465	0.000328	16.89
	27	0.001700	1.465	0.001160	16.93
	47	0.003000	2.154	0.001393	16.53
	50	0.014500	2.154	0.006732	16.83
	72	0.002086	2.843	0.000734	16.52
	96	0.002295	2.843	0.000807	16.15
	120	0.002295	2.843	0.000807	16.52
	144	0.006885	2.843	0.002422	15.42
	216	0.002295	2.843	0.000807	16.52
15 - 25	0.5	0.22200	1.001	0.22177	22.76
	4	0.04610	2.033	0.02267	23.99
	23	0.01930	0.325	0.05938	20.41
	27	0.01810	0.325	0.05569	21.06
	47	0.00770	1.066	0.00722	20.13
	50	0.03184	2.033	0.01566	16.83
	72	0.00334	0.999	0.00334	16.52
	96	0.009945	2.0025	0.00496	16.15
	120	0.005355	2.0025	0.00267	18.36
	144	0.009945	2.0025	0.00496	17.62
	216	0.005355	2.0025	0.00267	16.52
25 - 35	0.5	0.43010	2.231	0.19278	25.97
	4	0.05480	1.066	0.051407	25.94
	23	0.02070	1.066	0.019418	25.68
	27	0.03030	1.066	0.028424	25.19
	47	0.01145	1.066	0.012741	24.44
	50	0.03286	4.18	0.007861	24.48
	72	0.00445	4.18	0.001064	24.23
	96	0.013005	5.213	0.002494	23.50
	120	0.006885	5.213	0.001320	23.86
	144	0.017595	5.213	0.003375	22.03
	216	0.006885	5.213	0.001320	23.13
35 - 50	0.5	0.48060	4.123	0.11656	19.29
	4	0.34000	0.34	0.12445	18.84
	23	0.09350	0.0935	0.03422	16.89
	27	0.10320	0.1032	0.03777	18.58
	47	0.03060	0.0306	0.00894	18.69
	50	0.06958	0.06958	0.02027	18.36
	72	0.01840	0.0184	0.00536	18.72
	96	0.02907	0.02907	0.00847	17.62
	120	0.01836	0.01836	0.00535	18.36
	144	0.026775	0.02677	0.00780	17.62
	216	0.01377	0.00334	0.00334	17.62

Table 6. Calculation of hydraulic conductivity at a rainfall intensity of 90 mm/h.

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	$\partial H/\partial z$	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	0.5	0.2111	2.154	0.098004	17.22
	3	0.0542	1.465	0.036997	18.13
	23	0.006655	1.465	0.004543	17.24
	27	0.004016	1.465	0.002741	17.35
	47	0.000689	1.465	0.00047	17.25
	51	0.003443	1.465	0.00235	17.16
	72	0.00459	1.465	0.003133	16.52
	96	0.001052	1.465	0.000718	16.15
	144	0.002295	1.465	0.001567	15.42
15 - 25	0.5	0.3213	5.396	0.059544	20.37
	3	0.0550	4.3625	0.012607	20.15
	23	0.007114	4.3625	0.001630	20.05
	27	0.02639	4.3625	0.006049	20.65
	47	0.003289	4.3625	0.000753	20.13
	51	0.007268	4.3625	0.001665	20.28
	72	0.00590	2.0985	0.002811	19.82
	96	0.004112	2.0985	0.001959	20.56
	144	0.003285	2.0985	0.001565	19.82
25 - 35	0.5	0.9823	2.7475	0.35752	26.43
	3	0.07336	2.7475	0.02670	25.97
	23	0.008568	2.7475	0.00311	25.68
	27	0.0283	2.7475	0.01031	25.61
	47	0.004666	3.264	0.00142	25.88
	51	0.01071	3.264	0.00328	25.74
	72	0.007576	1.198	0.00632	25.33
	96	0.005642	1.198	0.00470	24.96
	144	0.00978	1.198	0.00816	24.23
35 - 50	0.5	3.249	4.455	0.72929	18.89
	3	0.4309	4.455	0.09672	18.86
	23	0.04269	4.455	0.00958	25.68
	27	0.0312	4.455	0.00700	25.61
	47	0.07948	4.11	0.01933	18.69
	51	0.0325	4.11	0.00790	18.72
	72	0.05478	4.11	0.01332	25.33
	96	0.0653	4.11	0.01588	17.62
	144	0.009785	4.11	0.00238	17.62

Table 7. Calculation of hydraulic conductivity at a rainfall intensity of 285.6 mm/h.

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	0.5	2.336	0.600	3.893333	23.21
	2	0.04743	0.600	0.07905	23.68
	4	0.02754	0.278	0.099065	23.31
	23	0.0511	0.278	0.183813	23.22
	47	0.00134	0.278	0.00482	23.01
	71	0.00803	0.278	0.028885	21.72
	96	0.00184	0.278	0.003319	22.03
15 - 25	0.5	3.1469	5.648	0.55717	15.77
	2	0.07497	3.065	0.02446	16.18
	4	0.0505	2.582	0.01955	15.72
	23	0.0523	2.582	0.02025	15.48
	47	0.01976	2.582	0.00765	15.10
	71	0.01766	2.582	0.00683	15.20
	96	0.01157	2.582	0.00448	14.68
25 - 35	0.5	3.1652	6.6815	0.47372	24.02
	2	0.2055	2.033	0.10208	25.97
	4	0.0520	2.033	0.02557	25.94
	23	0.0536	1.000	0.0536	25.68
	47	0.0249	1.000	0.0249	24.44
	71	0.01995	1.000	0.01995	23.89
	96	0.01587	1.000	0.01587	24.96
35 - 50	0.5	5.0899	7.208	0.70614	18.51
	2	0.7658	1.699	0.45073	18.51
	4	0.3756	1.699	0.22107	18.11
	23	0.1032	3.421	0.03016	18.2
	47	0.0521	3.421	0.01522	17.97
	71	0.0688	4.11	0.01673	18.46
	96	0.0548	4.799	0.01143	17.62

intensity of 90 mm/h can be seen in Table 6. It can be seen that the moisture content was comparatively greater than that for an intensity of 40 mm/h. Moisture content was then found to decrease with the advance in time. At an increased intensity of 285.6 mm/h also, the variation of soil moisture content with time was greater than that for lower rainfall intensities. The antecedent moisture content was low and hence the lower moisture percent at first.

Experiments were also carried out at varying slopes of 10% and 20% with rainfall intensities of 40 and 90 mm/h. At 10% slope, the change in moisture content with time was more at a higher intensity of 90 mm/h when compared to that at 40 mm/h. At a steeper slope of 20% too, the variation of soil moisture content with time was greater for higher rainfall intensity. Likewise, a steeper slope increased the rate of depletion with time. It can be inferred that slope of the plot has an emphatic effect on the amount of water depleted with time.

Hydraulic conductivity remained almost constant upto an intensity of 90 mm/h. As rainfall intensity increased to 285 mm/h, it was found that moisture content and hydraulic conductivity increased to 23.21% and 3.893 cm/h. This results in increase in the rate at which water infiltrates. Appendix II gives the hydraulic conductivity values for different rainfall intensities at 10% and 20% slopes. The hydraulic conductivity was found to remain constant at both slopes.

The functional dependence of soil hydraulic conductivity on volumetric moisture content was determined. A semi-logarithmic graph of hydraulic conductivity versus volumetric moisture content was plotted for different depths. As shown in fig.10, the functional dependence of soil hydraulic conductivity on moisture content obeys a straight line relationship on a semi-log plot. This means that the function is an exponential one. The functional dependence can be described empirically by a general equation:

$$K = ae^{b\theta}$$

where

K - hydraulic conductivity, mm/day

θ - volumetric moisture content (%)

a & b - constants of the soil

From the figure it can be seen that the K-θ curves for all depths follow almost the same pattern. Only the curve for 10 cm depth shows a slight variation. Table 8 shows the K-θ relationship at different depths.

4.3.1.2 Hydraulic head - volumetric moisture content relationship

The hydraulic head is the sum of gravitational and matric suction heads. The matric suction and volumetric moisture content were observed at different known depths in the soil. The observations were made at 10, 20, 30 and 40 cm depths. This procedure was done at different rainfall intensities of 20, 40, 90 and 285 mm/h and also for slopes of 10% and 20% and at rainfall intensities 40 and 90mm/h. Hydraulic head was plotted against volumetric moisture content. To plot the graph, readings at 20 cm depth was adopted. Figure 11 shows the variation of

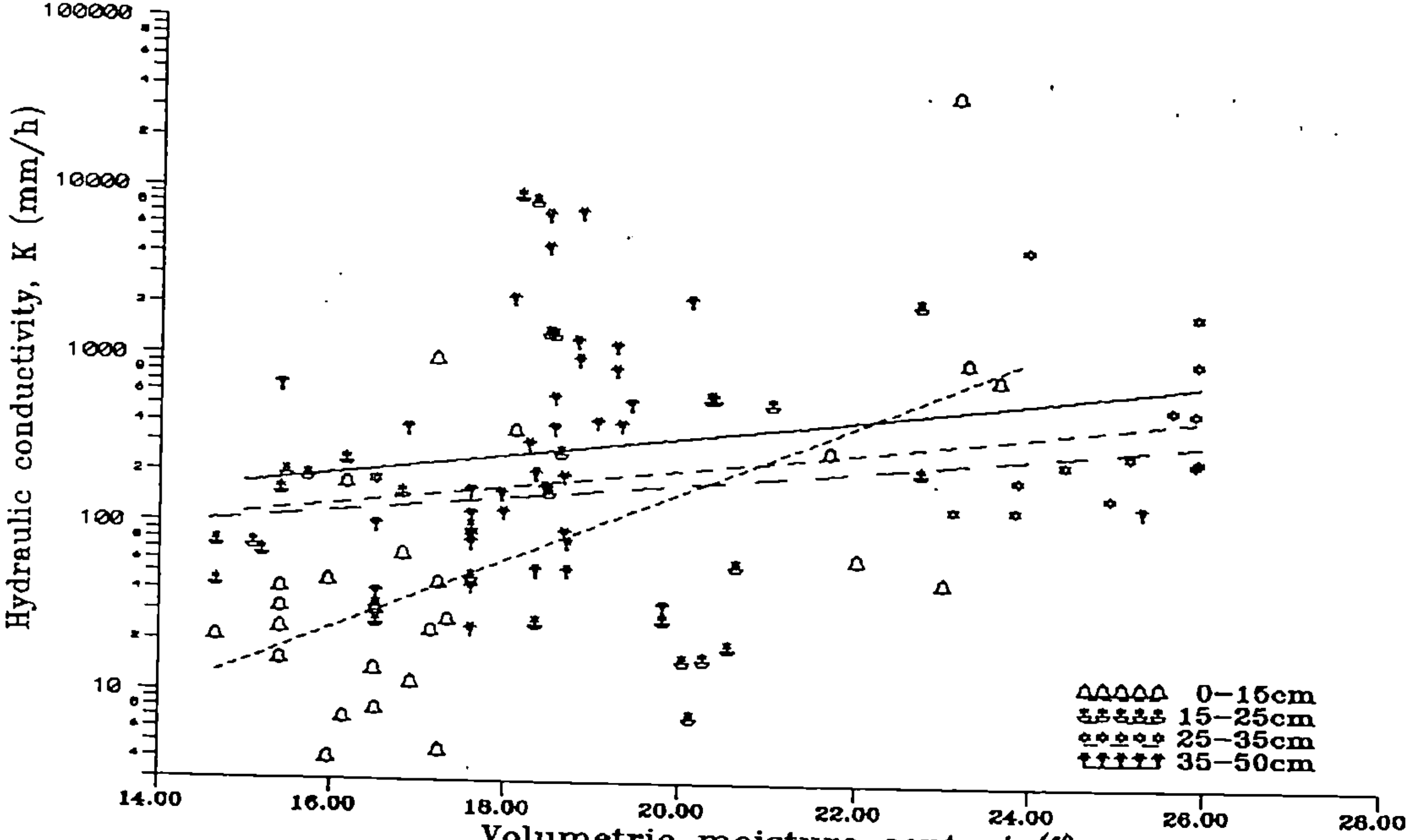


Fig.10 Variation of hydraulic conductivity with volumetric moisture content at different depths.

Table 8. Hydraulic conductivity - volumetric moisture content relationship

Depth of soil (cm)	k - θ relation
10	$k = \exp(0.469879\,\theta) \times 0.013098$
20	$k = \exp(0.10159\,\theta) \times 23.3915$
30	$k = \exp(0.126193\,\theta) \times 17.4555$
40	$k = \exp(0.13106\,\theta) \times 24.5655$

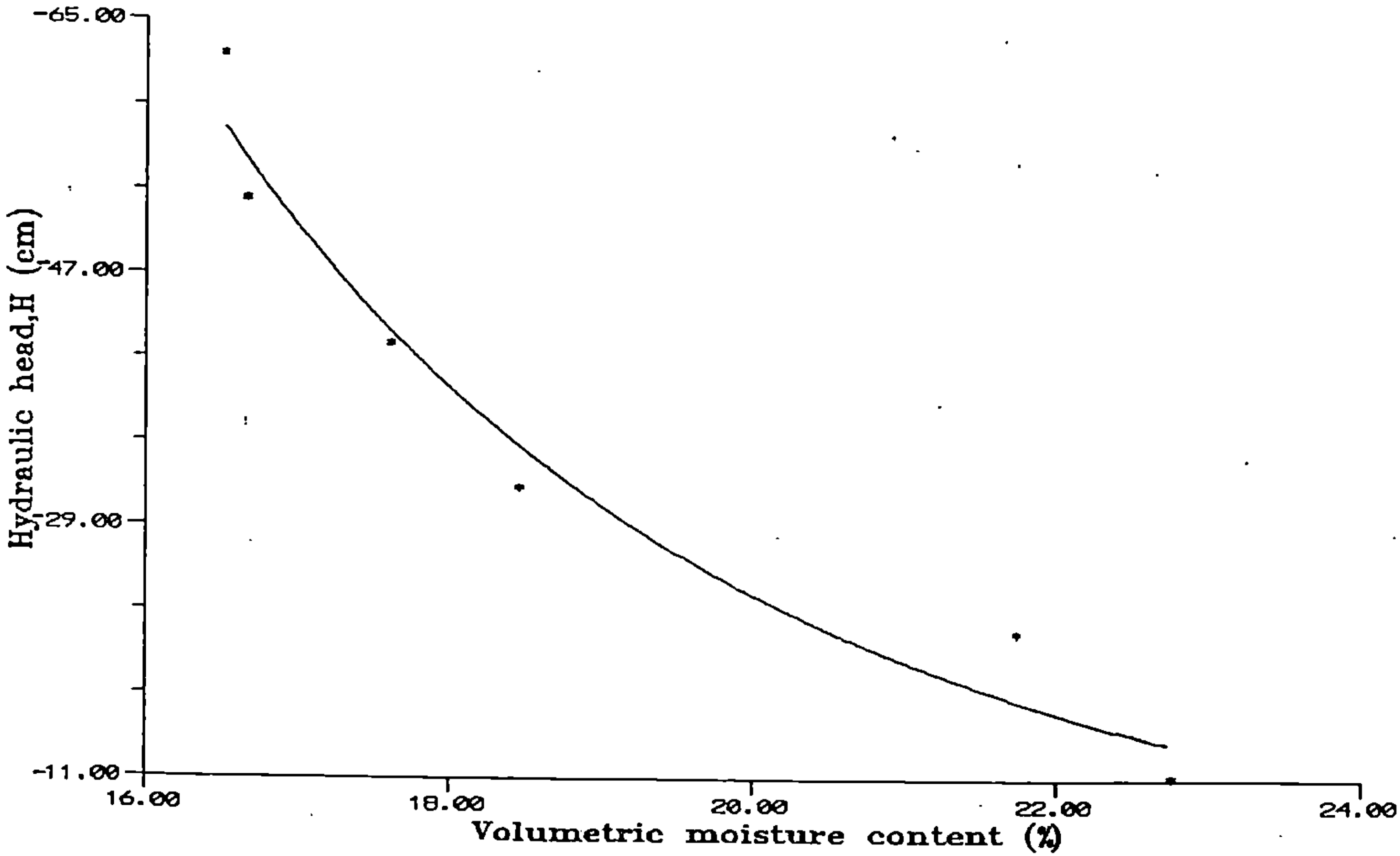


Fig.11 Variation of hydraulic head with volumetric moisture content.

hydraulic head with volumetric moisture content. Hydraulic head was found to increase with increase in moisture content. The hydraulic head at 16.52% moisture content was -62.7 cm. At a moisture content of 18.47%, the hydraulic head changed to -31.8 cm whereas at 22.76%, it further increased to - 11.2 cm.

In general, the θ-H relationship was found to be of the form,

$$H = a\theta^b$$

where

H - hydraulic head

θ - volumetric moisture content

a & b - constants

The θ-H relationship is given by :

$$H = 1.84531 \times 10^7 \times \theta^{-4.52089}$$

4.3.2 Soil moisture balance

It was seen that for the first three layers, the soil moisture content increased with depth before attaining field capacity. At 40 cm depth, however, the moisture content was less.

The moisture in the soil plot was accounted using the soil moisture balance equation. The resulting change in storage of moisture(ΔS) was compared to that observed during the experiment. A comparison of the change in moisture is given in Table 9. When the soil trough was horizontal, at an intensity of 20 mm/h, the change in soil moisture storage was found to be 6.199 litres as compared to the

Table 9. Comparison of change in moisture storage

Slope %	Rainfall intensity mm/h	Initial moisture storage %	Applied rainfall, P l	Surface runoff, R l	Sub-surface drainage, D l	Change in storage, ΔS l	ΔS (observed) l	Height of saturated zone, cm
Horizontal	20.0	12.9	8.1	-	1.901	6.199	7.9	-
	40.0	14.86	16.2	-	0.306	15.894	15.89	-
	90.0	11.85	36.45	-	1.51	34.94	35.23	0.3
	285.6	4.2	115.668	-	-	115.668	114.6	0.5
10	40	15.83	16.2	0.06	-	16.14	16.14	0.1
	90	11.0	36.45	0.045	3.73	32.675	32.40	0.5
20	40	15.46	16.2	0.032	1.116	15.052	15.42	0.5
	90	11.6	36.45	0.135	5.113	31.34	31.428	1.0

observed value of 7.9 litres. At an intensity of 40 mm/h, the calculated and observed ΔS values were almost the same ie, 15.894 l and 15.89 l respectively. For intensities of 90 mm/h and 285mm/h, these values were found to be 34.94 l & 35.23 l and 115.668 l & 114.6 l respectively. Storage increased with the increase in intensity. This means, an increase in moisture content and hence, higher hydraulic conductivity.

On adopting a slope of 10%, the observed and calculated values of change in moisture storage was equal (16.14 l) for an intensity of 40 mm/h. When the intensity was increased to 90 mm/h, the observed and calculated values were respectively 32.40 l and 32.675 l. At a slope of 20%; the computed and observed values of ΔS were 15.052 l and 15.42 l respectively for 40 mm/h, while those for 90 mm/h were 31.34l and 31.428 l respectively. This shows that soil moisture storage decreases by the change in slope. The change in slope also causes increased subsurface drainage.

4.4 Runoff generation

The surface and subsurface flow of water for different rainfall intensities and varying slope conditions were monitored.

When the soil plot was horizontal, without any slope, there was no surface runoff even at a high rainfall intensity of 285.6 mm/h. This was due to the high infiltration rate. Later, when the slope was changed to 10%, surface runoff volume of 0.06 litres was observed for a 0.5 h rainfall of intensity 40 mm/h. The surface runoff generation did not show any trend with the increase in slope and intensity.

4.5 Subsurface drainage

The variation in rainfall intensity did not cause any significant variation in drainage at no slope condition. At a maximum intensity of 285.6mm/h, there was no subsurface drainage. This was due to the low initial moisture content.

When the trough was inclined, the subsurface drainage also increased. For the same intensity of 90 mm/h, the subsurface drainage increased from 10.23% (3.73 l) to 14% (5.113 l) of the applied rainfall when the inclination was changed from 10 to 20%. There was no subsurface drainage at 10% slope for a rainfall intensity of 40mm/h; meanwhile at 20% slope 6.9% of the applied rainfall was obtained as subsurface drainage. This shows that inclination of the bedrock is a major factor in controlling the subsurface drainage of a region.

4.6 Variation in total outflow with time

The volume of total outflow from the trough was plotted against time, as in fig. 12. At 10% slope, for an intensity of 40 mm/h, only surface runoff of 0.06 l occurred. Subsurface runoff did not take place. As the experiment was conducted after a considerable interval, the soil was more dry and the infiltrated water was only sufficient to wet the lower layers. This was the reason that there was no subsurface outflow. For the same slope, but for an intensity of 90 mm/h, surface runoff was 0.05 l. After an interval of 1 h, there was a subsurface outflow of 0.71

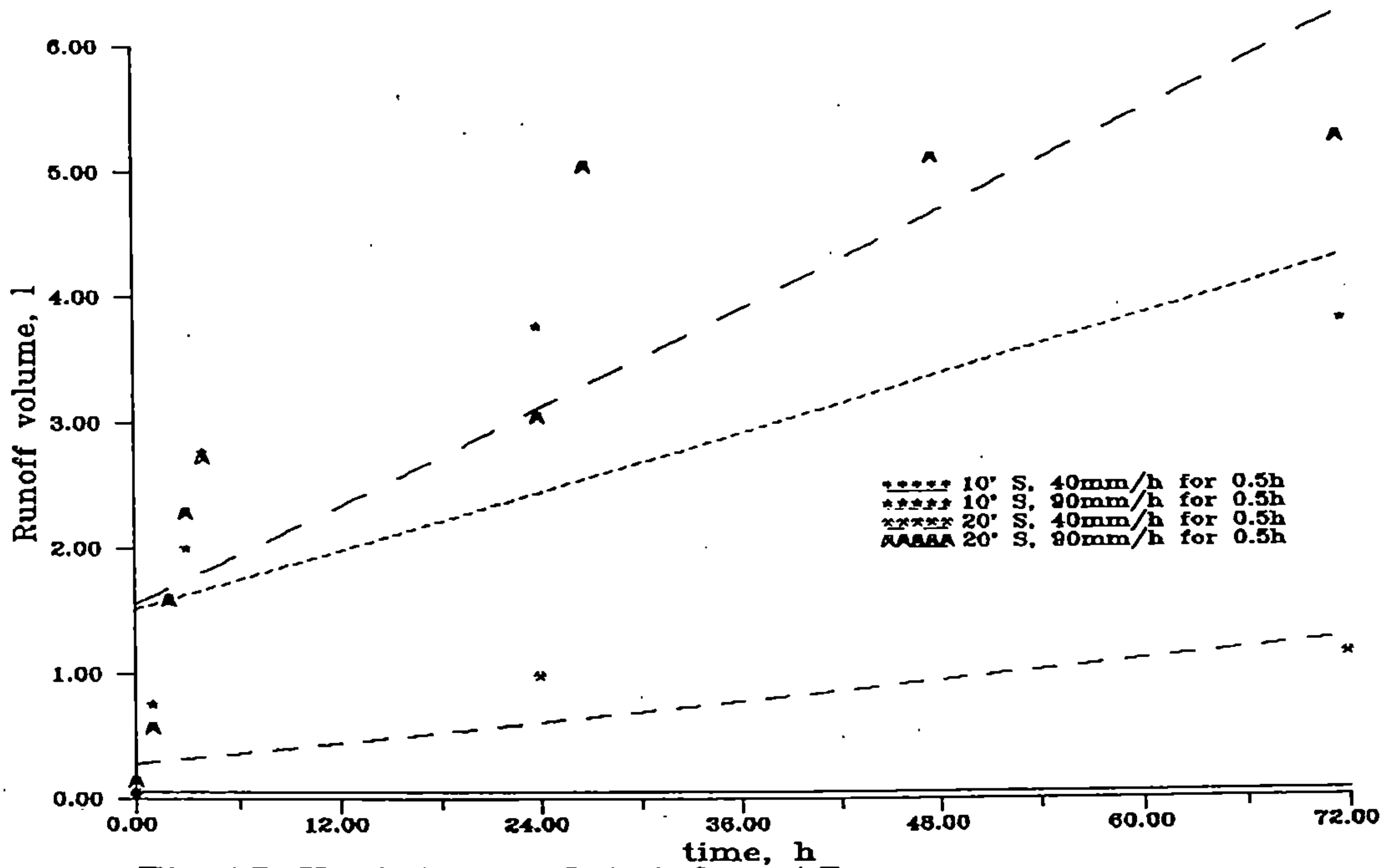


Fig.12 Variation of total outflow with time. (cumulative)

litres. A volume of 0.82 l was obtained at the end of the next interval of 1 h. The next two consecutive 1 h intervals yielded outflows of 0.42 and 0.77 l respectively. At the end of a 20 h interval, the outflow was 1 l. The next 48 h interval produced 0.03 l of outflow.

At 20% slope and a 0.5 h rainfall of 40 mm/h intensity, the surface runoff was 0.03 l. The subsurface outflow after 24 h was 0.96 l, while that after an interval of 48 h was 0.16 l. At the same slope and duration of rainfall, for a higher intensity, the surface runoff was 0.14 l. For the four consecutive 1 h intervals, the subsurface outflow observed was 0.43, 1.02, 0.69 and 0.44 l respectively. After a time interval of 20 h, the outflow was 0.33 l and that after the next 3 h was 2 l. A subsurface flow of 0.06 l was observed for the subsequent interval of 21 h. At the end of the next 24 h interval, the volume of outflow was 0.14 l. It can be concluded that the subsurface outflow is maximum at the end of 2 h after cessation of rainfall. The rate of drainage then decreases gradually.

Practical utility

In Kerala, the lateritic terrain is the backbone of the State's economy as a variety of cash crops are produced on this terrain. This study is useful in determining the transport of moisture and soil moisture storage in laterite soil. It can be inferred from this study that the inclination of the bed rock is the major governing factor in subsurface drainage.

Suggestions for future work

In this study, experiments were conducted at only two slope conditions of the soil trough. Further studies may be done with increased slopes. Also, studies may be carried out with longer duration of rainfall.

Summary

SUMMARY

Modelling of water movement in soils requires knowledge of the hydraulic conductivity as a function of volumetric moisture content [K(θ)] or soil water pressure head [K(h)] and the soil water retention curve. Laterite soils are by far the most important group occurring in Kerala and cover the largest area.

A study was conducted to understand the process of saturated and unsaturated flow of water in laterite soil under selected precipitation intensities. The objective was to establish a relation between hydraulic conductivity, moisture content and hydraulic head. Runoff generation and relation between runoff and soil moisture status were also considered.

A rainfall simulator and a soil trough were fabricated to conduct the soil hydraulic studies. The design of the rainfall simulator was based on that of Bhardwaj *et al* (1992). The portable rainfall simulator comprised of a drop forming mechanism mounted on a supporting frame. The drop forming mechanism consists of a tank with a perforated bottom. Wire loops of gage 20 copper are suspended through these perforations. A float valve ensures that constant head of water is maintained in the tank to get a desired intensity of rainfall. The position of the float can be adjusted to vary the head in the tank. The soil trough has provision for collecting surface and subsurface flow of water separately. The inclination of the

soil trough also can be varied as desired. The experimental setup was installed in the Soil and Water laboratory of KCAET.

The rainfall simulator was tested for different rainfall intensities by changing the head of water in the drop forming tank. A maximum intensity of 285.6 mm/h was attained at a head of 22cm. A 2^{nd} degree polynomial equation was established between intensity and head. The simulated raindrop size was determined by flour-pellet method. The experiments at different rainfall intensity showed that the simulated raindrop size remained almost constant at all intensities. The uniformity coefficient of rainfall was found to be a maximum ie, 85% for an intensity of 124.5 mm/h.

Properties of the soil were also determined. The particle size distribution curve when plotted showed that the soil was coarse graded. Both sieve analysis and hydrometry were conducted. The bulk density of the soil was determined by core cutter method. The soil had a bulk density of 1.53 g/cm^3. The liquid and plastic limits of the soil were determined by standard methods. The liquid limit was 42.55% and the plastic limit 65.28%.

The variation of soil moisture content with time for different rainfall intensities and slope conditions were studied. The moisture content increased with increase in rainfall intensity. The hydraulic conductivity was also found to increase at a high intensity of 285.6 mm/h. There was an increase in the rate of water movement too. A steeper slope was also found to increase the rate of drainage with time. But, the hydraulic conductivity remained almost constant for all slopes. The functional dependence of hydraulic conductivity (K) on volumetric moisture content (θ) was determined. It was observed that there is an exponential relation between K and θ.

The matric suction and volumetric moisture content were observed at different depths in the soil viz., 10, 20, 30 and 40 cm. Hydraulic head was found to increase with increase in moisture content. Hydraulic head- moisture content relation was given by

$$H = 1.84531 \times 10^7 \times \theta^{-4.52089}$$

Moisture balance in the soil trough accounted using the soil moisture balance equation was compared to that observed during the test. It was found that these were almost the same. Storage increased with the increase in intensity. It was not much affected by the change in slope.

A study on the runoff generation showed that there was no surface runoff even at a high intensity of 285.6mm/h when the soil trough was horizontal. Surface runoff was generated when the trough was inclined. Subsurface drainage also increased. It was concluded that the bed rock inclination is a major factor governing subsurface drainage. The variation in total outflow with time was analysed. The volume of subsurface runoff was found to be maximum at the end of two hours after rainfall. The rate of runoff then decreased gradually with time.

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Appendices

APPENDIX I

A. Calculation of soil moisture flux (20 mm/h, no slope)

Time, t (h)	Depth, z (cm)	$\partial\theta/\partial t$	$\partial z(\partial\theta/\partial t)$ (cm/h)	$q = \Sigma\partial z(\partial\theta/\partial t)$ (cm/h)
1	0 - 15 15 - 25 25 -35 35 - 50	0.002601 0.0104 0.0106 0.00443	0.039 0.104 0.106 0.0665	0.039 0.143 0.249 0.3155
2	0 - 15 15 - 25 25 -35 35 - 50	0.001071 0.000000 0.00658 0.007967	0.0161 0.000 0.0658 0.1195	0.0161 0.0161 0.0819 0.2014
4	0 - 15 15 - 25 25 -35 35 - 50	0.0266 0.0477 0.0022 0.045	0.399 0.477 0.022 0.675	0.399. 0.876 0.898 1.573
23	0 - 15 15 - 25 25 -35 35 - 50	0.00173 0.0002 0.0018 0.00233	0.026 0.002 0.018 0.035	0.026 0.028 0.046 0.081
27	0 - 15 15 - 25 25 -35 35 - 50	0.0002 0.00015 0.000321 0.00536	0.003 0.0015 0.0321 0.0804	0.003 0.0045 0.0366 0.117
29	0 - 15 15 - 25 25 -35 35 - 50	0.000667 0.002 0.0015 0.0026	0.01 0.02 0.015 0.039	0.01 0.03 0.045 0.084
120	0 - 15 15 - 25 25 -35 35 - 50	0.00006 0.000021 0.000074 0.000586	0.0009 0.00021 0.00074 0.0088	0.0009 0.00111 0.00185 0.01065
144	0 - 15 15 - 25 25 -35 35 - 50	0.000459 0.000306 0.000155 0.01132	0.00689 0.00306 0.00155 0.1698	0.00689 0.00995 0.0115 0.1813
168	0 - 15 15 - 25 25 -35 35 - 50	0.0006 0.000705 0.000 0.001063	0.009 0.00705 0.000 0.01595	0.009 0.01605 0.01605 0.032
192	0 - 15 15 - 25 25 -35 35 - 50	0.000307 0.000315 0.00015 0.000153	0.0046 0.00315 0.0015 0.0023	0.0046 0.00775 0.00925 0.001155
216	0 - 15 15 - 25 25 -35 35 - 50	0.000612 0.00061 0.00046 0.00046	0.00918 0.0061 0.0046 0.0069	0.00918 0.01528 0.01988 0.02678

B. Calculation of soil moisture flux
(40 mm/h, no slope)

Time, t (h)	Depth, z (cm)	$\partial\theta/\partial t$	$\partial z(\partial\theta/\partial t)$ (cm/h)	$q = \Sigma\partial z(\partial\theta/\partial t)$ (cm/h)
0.5	0 - 15	0.0052	0.078	0.078
	15 - 25	0.0144	0.144	0.222
	25 -35	0.04079	0.4079	0.4301
	35 - 50	0.00337	0.0505	0.4806
4	0 - 15	0.00074	0.0111	0.0111
	15 - 25	0.0035	0.035	0.0461
	25 -35	0.00087	0.0087	0.0548
	35 - 50	0.0190	0.2852	0.3400
23	0 - 15	0.000032	0.00048	0.00048
	15 - 25	0.001882	0.01882	0.0193
	25 -35	0.00014	0.0014	0.0207
	35 - 50	0.00485	0.0728	0.0935
27	0 - 15	0.000113	0.0017	0.0017
	15 - 25	0.00164	0.0164	0.0181
	25 -35	0.00122	0.0122	0.0303
	35 - 50	0.00486	0.0729	0.1032
47	0 - 15	0.0002	0.0002	0.003
	15 - 25	0.00047	0.00047	0.0077
	25 -35	0.000375	0.000375	0.01145
	35 - 50	0.00128	0.01915	0.0306
50	0 - 15	0.00096	0.0145	0.0145
	15 - 25	0.001734	0.01734	0.03184
	25 -35	0.000102	0.00102	0.03286
	35 - 50	0.00245	0.03672	0.06958
72	0 - 15	0.000139	0.002086	0.002086
	15 - 25	0.0001254	0.001254	0.00334
	25 -35	0.000111	0.00111	0.00445
	35 - 50	0.00093	0.01395	0.0784
96	0 - 15	0.0001966	0.002295	0.002295
	15 - 25	0.000765	0.00765	0.009945
	25 -35	0.000306	0.00306	0.013005
	35 - 50	0.001071	0.016065	0.02907
120	0 - 15	0.000153	0.002295	0.002295
	15 - 25	0.000306	0.00306	0.005355
	25 -35	0.000153	0.00153	0.006885
	35 - 50	0.000765	0.011475	0.01836
144	0 - 15	0.000459	0.006885	0.006885
	15 - 25	0.000306	0.00306	0.009945
	25 -35	0.000765	0.00765	0.017595
	35 - 50	0.000612	0.00918	0.026775
216	0 - 15	0.000153	0.002295	0.002795
	15 - 25	0.000306	0.00306	0.005355
	25 -35	0.000153	0.00153	0.006885
	35 - 50	0.000459	0.006885	0.01377

C. Calculation of soil moisture flux
(90 mm/h, no slope)

Time, t (h)	Depth, z (cm)	$\partial\theta/\partial t$	$\partial z(\partial\theta/\partial t)$ (cm/h)	$q = \Sigma\partial z(\partial\theta/\partial t)$ (cm/h)
0.5	0 - 15	0.01407	0.2111	0.2111
	15 - 25	0.01102	0.1102	0.3213
	25 -35	0.0661	0.6610	0.9823
	35 - 50	0.1511	2.2667	3.2490
3	0 - 15	0.003613	0.0542	0.0542
	15 - 25	0.00008	0.0008	0.055
	25 -35	0.001836	0.01836	0.07336
	35 - 50	0.023836	0.35754	0.4309
23	0 - 15	0.0004436	0.006655	0.006655
	15 - 25	0.0000459	0.000459	0.007114
	25 -35	0.0001451	0.001451	0.008568
	35 - 50	0.0022748	0.034122	0.04269
27	0 - 15	0.0002677	0.004016	0.004016
	15 - 25	0.0022374	0.022374	0.02639
	25 -35	0.000191	0.00191	0.0283
	35 - 50	0.0001933	0.0029	0.0312
47	0 - 15	0.0000458	0.000688	0.000688
	15 - 25	0.0002601	0.002601	0.003289
	25 -35	0.0001377	0.001377	0.004666
	35 - 50	0.0049876	0.074814	0.07948
51	0 - 15	0.0002294	0.003442	0.003442
	15 - 25	0.0003825	0.003825	0.007267
	25 -35	0.0003443	0.003443	0.01071
	35 - 50	0.0014526	0.02179	0.0325
72	0 - 15	0.000306	0.00459	0.00459
	15 - 25	0.000131	0.00131	0.0059
	25 -35	0.0001376	0.001676	0.007576
	35 - 50	0.003146	0.0472	0.05478
96	0 - 15	0.000070	0.001052	0.001052
	15 - 25	0.000306	0.00306	0.004112
	25 -35	0.0002582	0.002582	0.005642
	35 - 50	0.0039772	0.059658	0.0653
144	0 - 15	0.000153	0.002295	0.002295
	15 - 25	0.000099	0.00099	0.003285
	25 -35	0.0006495	0.006495	0.00978
	35 - 50	0.0000003	0.000005	0.009785

D. Calculation of soil moisture flux
(285.6 mm/h, no slope)

Time, t (h)	Depth, z (cm)	$\partial\theta/\partial t$	$\partial z(\partial\theta/\partial t)$ (cm/h)	$q = \Sigma\partial z(\partial\theta/\partial t)$ (cm/h)
0.5	0 - 15	0.15573	2.336	2.336
	15 - 25	0.08109	0.8109	3.1469
	25 -35	0.00183	0.0183	3.1652
	35 - 50	0.12831	1.9247	5.0899
2	0 - 15	0.003162	0.04743	0.04743
	15 - 25	0.002754	0.02754	0.07497
	25 -35	0.013053	0.13053	0.2055
	35 - 50	0.03735	0.5603	0.7658
4	0 - 15	0.001836	0.02754	0.02754
	15 - 25	0.002296	0.02296	0.0505
	25 -35	0.00015	0.0015	0.0520
	35 - 50	0.02157	0.3236	0.3756
23	0 - 15	0.0034066	0.0511	0.0511
	15 - 25	0.00012	0.0012	0.0523
	25 -35	0.00013	0.0013	0.0536
	35 - 50	0.0033066	0.0496	0.1032
47	0 - 15	0.0000893	0.00134	0.00134
	15 - 25	0.001842	0.01842	0.01976
	25 -35	0.000514	0.00514	0.0249
	35 - 50	0.0018133	0.0272	0.0521
71	0 - 15	0.0005353	0.00803	0.00803
	15 - 25	0.000963	0.00963	0.01766
	25 -35	0.000229	0..229	0.01995
	35 - 50	0.0032566	0.04885	0.0688
96	0 - 15	0.0001226	0.00184	0.00184
	15 - 25	0.000973	0.00973	0.01157
	25 -35	0.00043	0.0043	0.01587
	35 - 50	0.002601	0.039015	0.054885

APPENDIX II

A. Calculation of hydraulic conductivity (10% slope & rainfall intensity 40 mm/h)

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	$\partial H/\partial z$	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	0.5	0.26163	1.989	0.131538	23.21
	2	0.002295	1.989	0.001153	23.19
	23	0.048195	1.989	0.024230	22.87
	27	0.039015	1.989	0.019615	22.72
	47	0.01836	0.7688	0.02388	22.62
	51	0.057375	0.7688	0.07462	22.29
	72	0.089505	0.7688	0.11642	22.03
15 - 25	0.5	0.27999	5.1395	0.05447	24.02
	2	0.022185	5.1395	0.00431	24.93
	23	0.071145	5.1395	0.01384	24.35
	27	0.065025	5.1395	0.01265	24.28
	47	0.06732	2.0308	0.03314	24.18
	51	0.101745	2.0308	0.05010	23.86
	72	0.139995	2.0308	0.06893	23.73
25 - 35	0.5	1.07712	2.096	0.51389	16.20
	2	0.103275	2.096	0.04927	16.18
	23	0.237915	2.096	0.11351	15.78
	27	0.072675	2.096	0.03467	15.28
	47	0.09486	2.632	0.03603	15.10
	51	0.115515	2.632	0.04388	14.82
	72	0.181305	2.632	0.06887	14.82
35 - 50	0.5	1.224	4.105	0.29817	19.29
	2	0.204255	4.105	0.04975	19.27
	23	0.35955	4.439	0.08099	19.00
	27	0.081855	4.439	0.01843	19.00
	47	0.18666	4.111	0.04540	18.72
	51	0.120105	4.111	0.02921	18.72
	72	0.29835	4.111	0.07257	18.69

B. Calculation of hydraulic conductivity
(10% slope & rainfall intensity 90 mm/h)

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c., θ (%)
0 - 15	0.5	0.199665	1.989	0.100384	23.68
	1	0.02295	1.989	0.011538	23.68
	2	0.002295	1.989	0.001153	23.36
	3	0.002295	1.989	0.001153	23.19
	4	0.002295	1.989	0.001153	23.17
	24	0.00918	1.989	0.004615	23.13
	120	0.16524	1.297	0.127401	22.03
15 - 25	0.5	0.254745	2.0308	0.12544	25.97
	1	0.02295	2.0308	0.01130	25.97
	2	0.04284	2.0308	0.02109	25.95
	3	0.005355	1.0025	0.00534	26.67
	4	0.005355	1.0025	0.00534	23.99
	24	0.03978	1.0025	0.03968	23.86
	120	0.0918	0.9928	0.09246	22.03
25 - 35	0.5	0.282285	7.1877	0.03927	15.16
	1	0.04131	7.1877	0.00574	14.97
	2	0.07803	7.1877	0.01085	14.96
	3	0.009945	7.6930	0.00129	14.68
	4	0.013005	7.6930	0.00169	14.62
	24	0.10098	8.2190	0.01228	14.32
	120	0.23868	7.1907	0.03319	14.32
35 - 50	0.5	0.29376	0.3203	0.91714	19.46
	1	0.052785	0.3203	0.16479	19.27
	2	0.07803	0.3203	0.24361	19.27
	3	0.01683	0.3370	0.04994	19.27
	4	0.045135	0.3370	0.13393	19.27
	24	0.2295	0.3203	0.71650	19.27
	120	0.29376	0.3203	0.91714	18.65

C. Calculation of hydraulic conductivity (20% slope & rainfall intensity 40 mm/h)

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c, θ (%)
0 - 15	0.5	0.24786	1.297	0.191102	25.97
	2	0.07344	1.989	0.036923	23.53
	3	0.002295	1.989	0.001153	23.5
	4	0.002295	1.989	0.001153	23.17
	24	0.00918	1.989	0.004615	23.01
	120	0.16524	1.297	0.127401	20.19
15 - 25	0.5	0.27693	6.6835	0.04143	24.37
	2	0.07497	7.7215	0.00970	25.33
	3	0.002295	7.1955	0.00031	25.33
	4	0.005355	7.1955	0.00074	24.72
	24	0.02142	6.1675	0.00347	24.23
	120	0.20196	5.1295	0.03937	20.19
25 - 35	0.5	0.44523	1.589	0.280195	22.49
	2	0.09497	2.642	0.028376	21.84
	3	0.005355	2.116	0.002530	21.84
	4	0.005355	2.116	0.002530	21.42
	24	0.03978	2.106	0.018888	20.93
	120	0.22338	2.633	0.084838	18.36
35 - 50	0.5	0.493425	4.118	0.11982	21.49
	2	0.077265	4.118	0.01876	24.93
	3	0.05814	4.118	0.01411	25.52
	4	0.00765	4.118	0.00185	25.33
	24	0.11322	4.797	0.02360	24.60
	120	0.27846	5.851	0.04759	18.36

D. Calculation of hydraulic conductivity (20% slope & rainfall intensity of 90 mm/h)

Depth z (cm)	Time t (h)	Soil moisture flux, q (cm/h)	∂H/∂z	Hydraulic conductivity, K (cm/h)	Volumetric m.c, θ (%)
0 - 15	0.5	0.176715	1.297	0.136249	23.68
	2	0.025245	1.297	0.019464	23.68
	3	0.04131	0.956	0.043211	23.36
	27	0.006885	0.956	0.007201	19.82
	48	0.006885	0.0769	0.08953	19.82
	72	0.11016	0.0769	1.43250	19.09
15 - 25	0.5	0.181305	6.157	0.02944	25.97
	2	0.042075	6.157	0.00683	25.55
	3	0.074970	5.646	0.013278	25.33
	27	0.072675	6.674	0.01088	21.06
	48	0.066555	5.355	0.01242	20.56
	72	0.183600	5.355	0.03428	19.82
25 - 35	0.5	0.392445	3.625	0.108260	24.32
	2	0.084615	3.625	0.023424	25.55
	3	0.07803	3.625	0.021525	25.52
	27	0.104805	2.094	0.050050	20.65
	48	0.089505	2.094	0.042743	19.82
	72	0.22032	2.094	0.105214	19.82
35 - 50	0.5	0.406215	4.122	0.09854	18.89
	2	0.103275	4.122	0.02505	19.27
	3	0.172125	4.122	0.04175	19.27
	27	0.228735	3.417	0.06694	19.00
	48	0.11016	3.417	0.03223	18.36
	72	0.22032	4.118	0.05350	18.72

LABORATORY STUDIES ON SOIL MOISTURE FLOW AND RUNOFF GENERATION IN LATERITE SOIL

By
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ABSTRACT OF A THESIS

Submitted in partial fulfilment of the requirement for the degree

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ABSTRACT

If land and water are to be put to the best use, a full understanding of soil-water relation is necessary. Much of the research on retention and movement of water in soil and the use of water by plants is done with this objective. Modelling of water movement in soils requires knowledge of hydraulic conductivity as a function of volumetric water content or soil water pressure head, and the soil water retention curve.

Laterite soils are by far the most important group occurring in Kerala and cover the largest area. A study was conducted to analyse the saturated and unsaturated flow of water in laterite soil under selected precipitation intensities. The objective was to establish a relation between hydraulic conductivity, moisture content and hydraulic head. Runoff generation and relation between runoff and soil moisture status were also considered.

A rainfall simulator and a soil trough were fabricated. The design of the rainfall simulator was based on that of Bhardwaj *et al* (1992). The rainfall simulator comprised of a drop forming mechanism mounted on a supporting frame. A float valve maintains a constant head of water in the drop forming tank to get a desired rainfall intensity. The soil trough had provision for collecting surface and subsurface outflow of water. Provision was also made to incline the soil trough when a sloping plot was required. The experimental set up was installed in the Soil and Water laboratory of KCAET, Tavanur.

The rainfall simulator was tested for different rainfall intensity by changing the head of water in the tank. Rainfall intensity increased as the head of water in the tank was increased. At a head of 22 cm, a maximum rainfall intensity of 285.6 mm/h was obtained. Drop size determination by flour - pellet method showed that the simulated raindrop size remained almost constant. The uniformity coefficient also increased with increase in intensity. The basic soil properties were also determined. The particle size distribution curve showed that the soil was coarse graded. The bulk density of the soil was 1.53 g/cm^3.

The piezometer readings in the soil trough showed that unsaturated flow takes place through the soil under simulated rainfall. Experiments were done for different rainfall intensities of 20, 40, 90, and 285.6 mm/h and also at 10% and 20% slopes with 40 and 90 mm/h intensity. The variation in moisture content with time was found to increase with intensity of rainfall. Steeper slopes also increased the rate of depletion of moisture with time. Instantaneous profile method was adopted to determine hydraulic conductivity. An exponential relation was observed between hydraulic conductivity and volumetric soil moisture content.

Matric suction and volumetric moisture content were observed at different depths - 10, 20, 30 and 40 cm. The hydraulic conductivity - moisture content relationships were established. Studies on runoff generation showed that inclination of the bed rock is a major factor controlling subsurface outflow. Variation in intensity of rainfall did not cause considerable variation in outflow. Surface runoff also was affected only by the slope of the soil trough.