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A STUDY OF FLOODED AND RAIN INFILTRATION
RELATIONS WITH SURFACE PONDING

by

Yehia Z. El-Shafei

A dissertation submitted in partial fulfillment
of the requirements for the degree

of

DOCTOR OF PHILOSOPHY

in

Irrigation Science

Approved:

Major Professor

Committee Member

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UTAH STATE UNIVERSITY
Logan, Utah

1970

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Yehia Z. El-Shafei

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NOTATIONS

Symbol	Definition
a	constant represents the slope of an infiltration curve on log-log paper, dimensionless
A	the final infiltration rate, L/T
\bar{A}	cross section-area, L^2
$A'(i)$	constant dependent on the initial moisture content, dimensionally inconsistent
\bar{a} and \bar{b}	constants dependent on the liquid, the soil, and the capillary pressure history, dimensionally inconsistent
b and B	parameters determined from infiltration data, dimensionally inconsistent
C	constant representing Y at unit time, dimensionally inconsistent
\underline{C}	constant, dimensionally inconsistent
C'	constant mainly dependent on sprinkler intensity, R, and initial moisture content, W_i , in the soil, dimensionally inconsistent
D	soil-water diffusivity, L^2/T
\bar{D}	mean value of D, L^2/T
dW'	the area of the portion of the cross-section of the pore space that would desaturate if S underwent a change, dS, at the capillary pressure, P_c , L^2
D_s	soil-water diffusivity at saturation, L^2/T
e	base of the natural logarithms
e'	the effective porosity or water filled porosity, which is the ratio of the volume of water to the bulk volume, dimensionless
E	porosity, the pore volume expressed as a decimal fraction of the medium bulk volume, dimensionless
E_a	percent of pores not filled with water, dimensionless

NOTATIONS (Continued)

Symbol	Definition
f	shape factor, dimensionless
F	a factor, dimensionless
g	the gravity constant, L/T^2
G	bulk specific gravity, dimensionless
\bar{G}	soil parameter which results primarily from gravity forces, L/T
h	depth from water surface to wetting front, L
h_o	depth of water on soil surface, L
h_T	head loss in transmission zone, L
H	specific moisture capacity, L^{-1}
I	intake rate (instantaneous infiltration rate), L/T
I_o	the initial infiltration rate, L/T
k	the pore shape factor having an average value of about 2.5, dimensionless
K	capillary conductivity, L/T
K'	the effective permeability or effective capillary conductivity of a porous medium having any volumetric moisture content θ , L^2
\underline{K}	a coefficient called permeability or "intrinsic permeability," it is a function of the geometry of the medium, L^2
K_c	constant depending on the capillary forces on the moving water-soil boundary, dimensionless
K_s	the saturated hydraulic conductivity, L/T
K_T	hydraulic conductivity of the transmission zone, L/T
L	the length or distance between two points in a porous medium in the direction of flow, L
\bar{L}	the actual length of the path taken by the fluid as it passes through a porous medium, L

NOTATIONS (Continued)

Symbol	Definition
m	constant represents the slope of C' , on log-log paper, when C' is drawn as a function of R , dimensionless
\bar{m} and \underline{m}	constants, dimensionally inconsistent
n	is $(a-1)$, which represents the slope of infiltration rate on log-log paper, dimensionless
n'	constant having value between 1.0 and 0.80 dependent on the physical properties of the soil or soil type, dimensionless
N	constant representing I at unit time, dimensionally inconsistent
P_c	the capillary pressure at a particular saturation, S , F/L^2
q	volume of flow per amount of time, L^3/T
\bar{q}	the flow per unit cross-sectional area of stationary material, L/T
Q	the volume of fluid passing in time, t , L^3
r	the average capillary radii, L
R	water application rate, rain (sprinkler) intensity, L/T
\bar{R}	hydraulic radius, L
\bar{R}^2	the mean squared value of the hydraulic radius, dimensionless
R_t	the radius of capillary tube, L
S	degree of saturation, which is the ratio of the volume of water to the volume of voids ($S = 1.0$ for saturated flow), dimensionless
S_e	a reduced relative saturation and is equal to $(S - S_o)/(1.0 - S_o)$, dimensionless
S_o	the critical saturation at which the capillary conductivity is considered to approach zero, dimensionless
S_p	soil factor called sorptivity, $L/T^{1/2}$
S_u	the ultimate saturation achieved, dimensionless

NOTATIONS (Continued)

Symbol	Definition
t	is time, T
t _s	time of irrigation until flooding occurs or until appearance of surface ponding, T
T	tortuosity, which is equal to the ratio $(L/L)^2$, dimensionless
\bar{u}	the average tangential flow velocity, L/T
V	flow velocity, L/T
V _c	the average incremental velocity in the flow channel whatever its size or shape, L/T
V _{macro}	the velocity of water flow within completely filled voids, L/T
W	the soil moisture content on dry-weight basis, dimensionless
W _o	the moisture content at which the capillary conductivity is considered to approach zero, dimensionless
W _i	the initial moisture content, dimensionless
x	the horizontal distance coordinate, L
Y	accumulated infiltration, L
Y _R	cumulative rain (sprinkler) infiltration, L
Z	depth from soil surface to wetting front (vertical coordinate), L
α	the contact angle of the interface between wetting and non-wetting fluid at the medium solids, it is often assumed to be zero
β' and γ'	are constants, dimensionless
σ	surface tension of wetting fluid, F/L
η	viscosity of fluid FT/L ²
θ	the moisture content on volume basis, dimensionless
θ _i	the initial moisture content, dimensionless
θ _m	the maximum moisture content which can be obtained under flooded infiltration, dimensionless

NOTATIONS (Continued)

Symbol	Definition
θ_s	the moisture content at saturation, dimensionless
ρ	density of fluid, FT^2/L^4
ϕ	hydraulic potential, L
ψ	matric or capillary potential (suction), L
ψ_T	capillary potential in the transmission zone, L
τ	a unique suction (tension) function

ABSTRACT

A Study of Flooded and Rain Infiltration
Relations with Surface Ponding

by

Yehia Z. El-Shafei, Doctor of Philosophy

Utah State University, 1970

Major Professor: Joel E. Fletcher
Department: Agricultural and Irrigation Engineering

The non-linear partial differential equation (combination of Darcy's law and continuity equation) has been used in this investigation to predict the flooded infiltration through soils possessing appreciable amount of clay and initially drier than its field capacity. One of the most important assumptions made in solving the differential equation is that the capillary conductivity-moisture content relationship is unique for each initial moisture content computation due to the different reaction of clay minerals with different initial moisture contents.

Mathematical equations were also derived to predict:

1. The rate of the wetting front advance, prior to the occurrence of surface ponding, taking into account the effect of initial soil moisture content and rate of water application.

2. The time at which surface ponding takes place under different rain (sprinkler) intensities by utilizing the intake rate curve obtained under flooded infiltration.

The derived equations enable us to estimate a definite period of time, during which a field can be sprinkled at a given application rate, beyond which if sprinkling continues runoff will take place, and to estimate the accumulative rain (sprinkler) uptake at the time of surface ponding.

The theory was tested and firmly supported by the results of a multipurpose laboratory experiment conducted on samples of a Nibley silty clay loam soil packed into columns to a density of 1.25 gm/cm^3 .

(188 pages)

CHAPTER I

INTRODUCTION

Origin and Nature of the Problem

The rate at which water can be applied to the soil without causing runoff is an important factor in irrigation and hydrology. This importance has made the study of rain or sprinkler and irrigation infiltration receive attention in the last few years. The term rain or sprinkler infiltration as used here refers to a downward entry of water into the soil at a rate sufficing to prevent the formation of an enduring water cover. Such a cover may be defined as a free water layer, which accumulated upon the soil surface and which once established, persists at least as long as water is supplied to the soil surface.

On the other hand, ponded water infiltration is considered here as characterized by the pressure of an enduring water cover and as brought about either by flooding, rainfall, or by sprinkling. It has been shown that the relative amount of surface runoff which occurs following precipitation or sprinkling is dominantly affected by the rate of infiltration and the total amount of infiltration. When the intensity of rainfall or sprinkling is greater than the intake rate of the soil, the soil resists penetration of water. A capillary film is immediately formed over the ground surface followed by surface ponding, then surface runoff.

If the intake rate curve (obtained under flooding condition) for the soil is used as a basis there would be a definite period of time, during

which a field could be sprinkled at a given application rate, beyond which if sprinkling continues runoff (or small water pits) will appear.

Hence, if the intensity of sprinkling is not coordinated with the soil's intake rate (and usually the former is higher than the latter), then the irrigation level may be less than is required to wet the soil to the full depth of the roots.

Soils must be irrigated when their moisture levels are in the range between air dry and field capacity. Therefore, it is important to investigate the effect of the initial moisture content and the water application rate on surface ponding and infiltration for a soil initially drier than its field capacity.

Objectives

The objectives of this study are:

1. To investigate the possibility of obtaining the intake rate curve (under flooding infiltration) theoretically from the physical properties of the soil such as hydraulic conductivity, capillary potential, and initial soil moisture content for a soil initially drier than its field capacity.
2. To develop a mathematical equation to predict the rate of the wetting front advance, prior to the occurrence of surface ponding, taking into account the effect of initial soil moisture content and rate of water application.
3. To investigate the possibility of using the intake rate curve for the soil to predict the time at which surface ponding takes place under different water application rates.

4. To investigate the moisture content profiles during sprinkling at different application rates.

CHAPTER 2

REVIEW OF LITERATURE

Flow of Water Through Unsaturated Soil

Fourier (1822) presented his very complete mathematical theory of the transport of heat in conducting materials based on the law that the rate of conduction of heat is proportional to the temperature gradient. In 1827 Ohm enunciated his law to the effect that the rate of transport of electricity (i.e. the strength of an electric current) in a conductor of electricity is proportional to the difference of electric potential between its ends, i.e. that the electric current is proportional to the electric potential gradient. In 1822 Navier developed equations describing the flow of viscous fluids in terms of the distribution of hydraulic potential, equations which were later again derived by Stokes (1845) in a more general law. Where the boundaries of the moving fluid are simple, as in the case of flow through a tube of uniform radius, these equations permit one to derive the rate of flow in terms of the dimensions of the conductor and the potential difference between the ends, and indeed Poiseuille's (1842) experimentally derived equation of flow of fluid through a tube readily can be obtained from the earlier theoretical work. His equation is quoted here in the form,

$$\frac{Q}{t} = \frac{\Delta\phi}{L} \frac{\pi}{8\eta} R_t^4 g \dots \dots \dots (1)$$

wherein

- Q = the volume of fluid passing through the tube in time t, L³
 L = the length of the tube between the ends, L

- $\Delta\phi$ = the potential difference between the ends of the tube, L
 η = the viscosity of the fluid, FT/L²
 R_t = the radius of the tube, L
 g = gravity constant, L/T²
 ρ = the density of fluid, FT²/L⁴

Here again the rate of flow is proportional to the potential gradient. Gardner (1920) developed a theory about the flow of water in unsaturated soil analogous with heat flow. In his concept the so-called "capillary transmission constant" was assumed to be independent of the moisture content of the soil. In a later article Gardner (1936) followed the theory of Buckingham (1907) who stated that the capillary conductivity must be dependent on the moisture content. Richards (1931) developed a theory from which the validity was later proved by Childs and Collis-George (1950). According to this theory the one-dimensional flow of water in unsaturated soil obeys Darcy's law:

$$V = -K \frac{\partial\phi}{\partial Z} \dots \dots \dots (2)$$

in which

- V = the flow velocity, L/T
 K = the capillary conductivity, L/T
 Z = the depth of wetting front, L
 $\frac{\partial\phi}{\partial Z}$ = the potential gradient, dimensionless
 ϕ = the potential, L

The negative sign indicates that the direction of flow is opposite to that in which the potential increases. This is the most general statement of Darcy's law for isotropic porous materials, i.e. materials

which do not have any preferred directions of flow by reason of their structure, and in which, therefore, the flow is in the direction of the potential gradient.

Gardner (1956) divided the term ‘‘velocity of flow’’ into two different terms, the macroscopic velocity V_{macro} and the channel velocity V_c . He also introduced the F-function into Darcy’s law to account for the fact the voids are not completely filled with water.

$$V_{macro} = V_c \theta = - K_s F \frac{d\theta}{dz} \dots \dots \dots (3)$$

in which

- V_{macro} = the velocity of water flow within completely filled voids, L/T
- V_c = the average incremental velocity in the flow channel whatever its size or shape, L/T
- θ = the moisture content on volume basis, dimensionless
- K_s = the saturated hydraulic conductivity, L/T
- F = a factor, dimensionless

The potential ϕ is given by the equation

$$\phi = \psi + Z \dots \dots \dots (4)$$

in which

- ψ = the matrix or capillary potential (suction), L
- Z = the gravitational potential, L

Gardner (1956) introduced a method for determining the F-function by using the moisture content and the gradient of the moisture potential, both as a function of distance from water source and time.

The flow of a fluid in an unsaturated porous system must obey the law of conservation of matter which is expressed in the equation of continuity. This equation in turn expresses the fact that the differences between the rates of flow into and out of an element of the soil equals the rate of storage, thus

$$\frac{\partial V}{\partial Z} = - \frac{\partial \theta}{\partial t} \quad \dots \dots \dots (5)$$

Substituting Equation (2) into Equation (5) yields

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial Z} \left(K \frac{\partial \Phi}{\partial Z} \right) \quad \dots \dots \dots (6)$$

Using the fact that in unsaturated soil Ψ is negative and for vertical flow $\partial Z/\partial Z = 1$, Equation (6) can be written

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial Z} \left(K \frac{\partial \Psi}{\partial Z} \right) + \frac{\partial K}{\partial Z} \quad \dots \dots \dots (7)$$

where Z is taken as the vertical ordinate positive upward.

For horizontal flow $\partial Z/\partial Z = 0$ and Equation (6) changes into

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial X} \left(K \frac{\partial \Psi}{\partial X} \right) \quad \dots \dots \dots (7a)$$

in which

x = the horizontal distance coordinate

When the concept of the diffusion coefficient of water in soil is introduced, the problem becomes analogous to the thermal diffusion of heat. This introduction can be accomplished by splitting the gradient of the matric potential into two parts, assuming that the matric potential is a unique function of the moisture content θ , to yield

$$\frac{\partial \Psi}{\partial Z} = \frac{\partial \Psi}{\partial \theta} \cdot \frac{\partial \theta}{\partial Z} \quad \dots \dots \dots (8)$$

boundary conditions which are defined in the particular problem under consideration. The value of $\frac{\partial \Psi}{\partial \theta}$ may be determined from the moisture characteristics. K is either determined experimentally or calculated and the product D can then easily be obtained. The analysis of water flow in unsaturated soil, described by Equation (10), has been possible mathematically only through numerical methods. While both the experimental difficulties of determining the relationship between D or K and θ , and the mathematical difficulties of analysis are considerable, and complicated by hysteresis effects, the main features of water movement in unsaturated soil can be understood and predicted by the equation.

Olsen (1961) introduced four conditions to be satisfied for the applicability of Darcy's law:

The soil particles are approximately

- (a) uniform
- (b) larger than one micron
- (c) small enough so that the liquid flow is laminar
- (d) flow channels are uniform in size

Swartz (1968) introduced the τ function which he assumed is a unique suction (tension) function and has written Equation (2) in the form

$$V = K(\tau) \frac{\partial \tau}{\partial x} \dots \dots \dots (11)$$

$$V = - D(\theta) \frac{\partial \theta}{\partial x} \dots \dots \dots (11a)$$

wherein

$$D(\theta) = \left[- \frac{d\tau}{d\theta} K(\theta) \right] \dots \dots \dots (12)$$

in which

$$D = \text{soil-water diffusivity}$$

$$\frac{\partial \tau}{\partial x} = \text{the suction (tension) gradient}$$

Determination of Capillary Conductivity

Laboratory measurement of the capillary or unsaturated conductivity of soils at various values of capillary pressure or saturation is a routine but tedious procedure. Often, the time spent in obtaining precise measurements of capillary conductivity is not warranted for the application intended. But, as pointed out in the previous section, first of all the relation between the capillary conductivity and the moisture content must be known to solve the flow problems in unsaturated soil.

In order to calculate the diffusivity, the moisture characteristic must be available. Otherwise the conductivity may be derived from D-values and the moisture characteristic of the soil.

Several schemes have been proposed for the calculation of capillary conductivity for flow in partially saturated media. Kozeny in 1927 (Keller, 1967) presented an expression relating the conductivity of granular earth materials to such geometric properties as the pore space, porosity, specific surface, pore shape, and tortuosity of the pores. The ratio of the porosity to the specific surface was used to evaluate the mean hydraulic radius of the pores.

The expression presented by Kozeny provides a reasonable approximation to the conductivity of sands with fairly uniform pore size. It fails, however, for media having a wide range of pore sizes such as soils or rocks containing clay. The reason is that the conductivity is actually

proportional to the weighted mean of the squares of the hydraulic radii of pores and not to the square of the mean hydraulic radius as was assumed by Kozeny.

Richards (1931) used tensiometers maintained at measured suctions to supply water to the inflow face and extract it from the outflow face to measure the potential difference between these two separated points in the unsaturated column, thus reproducing the features of the constant head permeameter with unsaturated soil under suction. The capillary conductivity, then, can be calculated with the aid of Equation (2). This method was used in a somewhat modified form by Christensen (1944). The method is not, however, a good one, because in general the maintenance of a potential gradient imposes a suction (tension) gradient and, with it, a consequent gradient of moisture content and of capillary conductivity. Thus the potential gradient is not uniform for a constant rate of water movement, and the potential difference becomes greater across the layers of lower conductivity. This difficulty may be overcome by securing a uniform moisture content throughout the length of the sample column, or by endeavoring to measure the moisture content and potential gradient at a point. The latter course was pursued by Moore (1939), who plotted the suction distribution along the column by means of an array of tensiometers and the moisture distribution by sampling at the end of the experiment. In a similar experiment, Wyckoff and Botest (1936) used water in which gas was dissolved under pressure. As the pressure was reduced, the gas came out of solution in the pore space, thereby displacing liquid and causing a reduction of the degree of saturation. The moisture content was determined by measuring the electric conductivity of the unsaturated

material while the potential gradient was measured by manometers. It cannot be said with certainty that the distribution of water and air in the pore space in such an experiment are separately continuous and move independently. It may well be that the air is in the form of bubbles carried along by the water, and in fact the authors, in their discussion, envisage that this is the case. If so, the results of the experiment are not relevant to the movement of soil water in which the air in the pore space is in continuous equilibrium with the external atmosphere, although the curves of conductivity expressed as functions of the degree of saturation proved to be of the same kind as those resulting from more valid kinds of experiments.

Childs and Collis-George (1950) developed a method of relating capillary conductivity to pore-size distribution, the method taking into account the variation in pore size. Because their method requires matching theoretical and experimental curves at a single point, it is suitable only for calculating values of conductivity as functions of capillary pressure or saturation relative to the experimentally determined values. The method cannot be used to calculate absolute values of conductivity. Furthermore, the calculations necessary in applying their method are laborious.

Marshall (1958) proposed a variation of their method which greatly facilitated the computation arbitrarily by choosing equal volume components of porosity. He calculated the capillary conductivity directly from the pore-size distribution based on the moisture characteristic curve. He divided the total saturated pore space into n_0 equal fractions by volume and calculated r_1, r_2, \dots, r_{n_0} , representing the mean radius

of pores (assumed capillaries) in each fraction where r_1 belongs with the class with the largest pores and r_{n_0} with the smallest. Summing the contribution from each combination of radii, he obtained:

$$K' = (e') \frac{n_0^{-2}}{8} \left[r_1^2 + 3 r_2^2 + \dots + (2n_0 - 1) r_{n_0}^2 \right] \quad (13)$$

in which

e' = the effective porosity or waterfilled porosity, which is the ratio of the volume of water to be bulk volume, dimensionless

n_0 = the number of equal fractions of pore space divided on a volume basis, dimensionless

K' = the effective permeability or effective capillary conductivity of a porous medium having any volumetric moisture content θ , L^2

Millington and Quirk (1961) proposed another variation of this method which, according to Jackson et al. (1965) predicted experimental curves more closely than did the procedures of Marshall or Childs and Collis-George.

Nielsen et al. (1960) also compared values of capillary conductivity determined by the methods of Marshall and of Childs and Collis-George with experimental results and found it necessary to accept an order of magnitude as the permissible error for agricultural soils.

Taking the case, wherein water at a constant rate is flowing down a sufficiently long column to a water table maintained at a constant level below the surface, that a sufficient time has elapsed for the rate of flow to be the same everywhere in the profile so that no further changes of θ

take place, a steady state of flow is obtained. According to Equation (4) and Equation (6), the equation of flow is given by:

$$\frac{\partial}{\partial Z} \left\{ K \left(\frac{\partial \Psi}{\partial Z} + 1 \right) \right\} = 0 \quad (14)$$

Then the first stage of integration gives

$$K \left(\frac{\partial \Psi}{\partial Z} + 1 \right) = K_0 \quad (15)$$

where K_0 is the constant of integration. Equation (15) simply gives an expression of Darcy's law where K_0 is the rate of flow down the column.

Taking Ψ a positive value, Equation (15) changes into

$$\frac{\partial \Psi}{\partial Z} = 1 - \frac{K_0}{K} \quad (16)$$

Integrating between the limits of the water table (where both Z and Ψ are zero) and Z (where the suction is Ψ) gives

$$Z = \int_0^{\Psi} \frac{d\Psi}{(1 - K_0/K)} \quad (17)$$

Childs and Collis-George (1950) now utilized the fact that in the case described above the moisture content and suction are uniform over an appreciable length. Zones of variable moisture content (suction) are located at the lower end of the column to the neighborhood of water table in a way which depends upon the pore size distribution and at the upper end to a zone in which is localized any intermittency of water supply. In the zone of uniform moisture content, the only pressure gradient is due to gravity. The moisture contents adjust themselves to provide the necessary conductivity to conduct the imposed flow with the gravitational potential gradient. For the computation of the conductivity one needs only to

measure the rate of flow and the cross sectional area of the column. For the calculation itself Equation (15) is used with $\frac{\partial \psi}{\partial z} = 0$. On the basis of the assumption that $K(\psi)$ and $\frac{d\psi}{d\theta}$ are constant over a small range of potential, one can derive the equation for horizontal flow:

$$\frac{\partial \psi}{\partial t} = \bar{D} \frac{\partial^2 \psi}{\partial x^2} \quad (18)$$

where \bar{D} is a mean value of D . This equation has the form of an ordinary diffusion equation and solutions are readily available for a wide variety of boundary and initial conditions (Carlshaw and Jaeger, 1947).

Equation (18) was used by Gardner (1956) for the measurement of capillary conductivity as a function of ψ but also the diffusion function may be obtained from it. The system used by Gardner consisted of a soil sample in a pressure membrane apparatus. The outflow of water as a function of time is measured and from this it is possible to evaluate the permeability. By repeating the measurements over a succession of small increments of potentials ($\psi_2 - \psi_1$) a series of conductivity-potential values is obtained. In a later article Gardner (1958) noted that the capillary conductivity, K , can be related approximately to the capillary potential ψ , by the empirical expression:

$$K = \bar{a} [\bar{\psi}^{\bar{m}} + \bar{b}]^{-1} \quad (19)$$

in which

\bar{a} , \bar{b} , and \bar{m} are constants and depend on the liquid, the soil, and the capillary pressure history, dimensionally inconsistent.

The ratio \bar{a}/\bar{b} gives the hydraulic (saturated) conductivity, i.e.

when $\psi = 0$ and, according to Gardner, for most types of soil, m varies from

about 2 (for fine-textured soils) to 4 or more (for coarse-textured soils). For sand it may exceed 10 or 15.

Bruce and Klute (1956) used a horizontal semi-infinite column with initial constant moisture content. Water was applied at the end $x=0$ by maintaining this end at saturation. Water now will flow into the column and assume a distribution conditioned by the nature of the diffusivity-moisture content function. The initial and boundary conditions are:

$$\begin{aligned} \theta &= \theta_i & x > 0 & & t = 0 \\ \theta &= \theta_s & x = 0 & & t > 0 \end{aligned} \quad \dots \dots \dots (20)$$

in which

- θ_i = the initial moisture, dimensionless
- θ_s = the moisture content at saturation, dimensionless

The fundamental differential equation for this case is given by Equation (10a). Assuming that the solution of the last mentioned equation is a function of the variable $\phi = xt^{-1/2}$, one can derive the solution for Equation (10a) subject to the boundary conditions Equation (20) (Bruce and Klute, 1956).

$$D(\theta_x) = - \frac{1}{2t} \left[\frac{\partial x}{\partial \theta} \right]_{\theta_x} \int_{\theta_i}^{\theta_x} x \partial \theta \quad \dots \dots \dots (21)$$

The integral in this equation is evaluated from a moisture content-distance curve for the flow system described above. The derivative $\left[\frac{\partial x}{\partial \theta} \right]_{\theta_x}$ is the slope of this curve (θ - x curve at a constant value of time, t) evaluated at θ_x and t is the time at which the distribution is measured.

Bruce and Klute mentioned that for making the diffusivity calculations, a smooth curve was drawn through the scattered points which is one of the most serious objections to the whole procedure. An interesting aspect of his distribution curves is that the slope, $\frac{dx}{d\theta}$, increases as the initial moisture content increases. Furthermore, the magnitude of the integral will be decreased as the initial moisture content increases. This means that for a given θ_x , the value of $\frac{dx}{d\theta}$ must increase in order to obtain the same value of $D(\theta_x)$ from a distribution curve with a higher initial moisture content. In view of the large possible errors in the calculated diffusivities, Klute concluded that it is not possible to say, for example, whether or not the three curves calculated for Bloom-field sand at different initial moisture contents are different.

King (1965) converted Equation (19) into a dimensionless equation for relative conductivity by dividing it by a comparable expression for hydraulic or saturated conductivity. He also presented another empirical formula for relative permeability based on capillary potential and depending upon the liquid, the soil, and the capillary potential history. Furthermore, he proposed an empirical expression which describes the moisture release curve in terms of saturation and capillary potential. By combining the equations expressing capillary conductivity in terms of capillary potential, i.e., Equation (19), with the expression for capillary potential in terms of saturation, he obtained empirical equations for the relative permeability in terms of saturation. King (1965) compared both his equation and Equation (19) against experimental steady state imbibition data, using oil with sands, a loam soil, and a clay soil in the capillary potential range of 0 to 100 centimeters. King made both his equation

$d(wp)$ = the corresponding change in wetted perimeter of the desaturated area at the soil-wetting phase interphase, L
 σ = surface tension of the wetting fluid, F/L

The expression which Brooks and Corey (1964) present for \overline{R}^2 when the contact angle between the water and solid is zero, is:

$$\overline{R}^2 = \frac{\sigma^2}{S} \cdot \int_{S_0}^S \frac{dS}{P_c^2} \dots \dots \dots (23)$$

in which

\overline{R}^2 = the mean squared value of the hydraulic radius, dimensionless

S = the saturation, which is the ratio of the volume of water to the volume of voids ($S = 1.0$ for saturated flow), dimensionless

S_0 = the critical saturation at which the capillary conductivity is considered to approach zero, dimensionless

They call the ratio $(\underline{L}/L)^2$ tortuosity, T , dimensionless where

L = the length or distance between two points in a porous medium in the direction of flow, L

\underline{L} = the actual length of the path taken by the fluid as it passes through a porous medium, L

The variation in tortuosity with saturation was studied both analytically and experimentally by Burdine (1953) and the following expression was evolved:

$$\frac{1}{T_r} = \frac{T_{1.0}}{T_s} = \left(\frac{S-S_0}{1-S_0} \right)^2 = S_e^2 \dots \dots \dots (24)$$

in which

- T = tortuosity, which is equal to the ratio $(L/L)^2$, the subscript s denotes the degree of saturation, dimensionless
- T_r = the relative tortuosity for unsaturated flow, \bar{T}_s , and saturated flow, T_{1.0}, dimensionless
- S_e = a reduced relative saturation and is equal to $(S-S_0)/(1.0-S_0)$, dimensionless

Brooks and Corey (1964) combine Equation (23) and Equation (24) with an equation similar to Carman's equation (1939) to obtain the following expression for capillary conductivity:

$$K' = K (S_e)^2 \frac{\int_0^S \frac{dS}{P_c^2}}{\int_0^{1.0} \frac{dS}{P_c^2}} \dots \dots \dots (25)$$

in which

- K' = the effective permeability or capillary permeability of an unsaturated porous medium having any volumetric moisture content, θ , L²
- K = a coefficient called permeability or "intrinsic permeability." It is a function of the geometry of the medium, L².

Keller (1967) used the Burdine and Kozeny-Carman capillary conductivity equations and extended them to an exponential form.

$$K = \left(\frac{G_1}{G''}\right)^{2/3} \cdot \left(\frac{2D_1^3}{2T_{11}}\right) \cdot (W_e)^2 \cdot \left(\frac{1}{3C''+1}\right) \cdot \left[X^{(3C''+1)}\right]_0^{\bar{X}} \dots (26)$$

soils from parameters that can be obtained from capillary pressure-desaturation data. The procedure is applicable to the drainage cycle and at moisture contents greater than field capacity

$$K' = \frac{E \sigma^2 \cos^2 \alpha}{kT} \int_0^S \frac{dS}{p_c^2} \dots \dots \dots (28)$$

in which

- σ = surface tension of wetting fluid, F/L
- E = the porosity, the pore volume expressed as a decimal fraction of the medium bulk volume, dimensionless
- α = the contact angle of the interface between wetting and non-wetting fluid at the medium solids, it is often assumed to be zero
- k = the pore shape factor having an average value of about 2.5, dimensionless

One of the objections to this procedure is the assumption that the pore shape factor, k , remains constant as the saturation varies.

Theory of Infiltration

The rate of infiltration is the rate at which water crosses the soil surface and enters the profile.

The infiltration rate of a soil has been defined by Richards (1952) as the maximum rate at which a soil, in a given condition at a given time, can absorb rain. Quantitatively, infiltration rate is defined as the volume of water passing into the soil per unit area per unit of time. It has the dimensions of velocity. It is especially not to be confused

with the hydraulic conductivity of the soil, which is only one factor entering into the rate of infiltration.

The rate of water entry into the soil fluctuates widely between soil types and also wide differences can be found within a single soil type, depending upon the soil moisture level and management practices employed. A number of different formulas expressing the law of infiltration have been proposed, based on either the analysis of more or less simple models, on intuition, on frank empiricism or on genuinely ascertained physical properties of porous materials in general and of soil in particular. The aim of all such formulas is to express the rate of infiltration as a function of time elapsed from the inception of surface flooding, and, in particular, to account for the rapid decrease from initially very high values and, for uniform soils, the asymptotic approach to an ultimate constant value.

Flooded (ponded) infiltration

Kostiakov (1932) seems to have been the first to present the following equation for describing the infiltration of water into soils

$$Y = C t^a \quad (29)$$

and intake rate equation

$$I = \frac{dY}{dt} = aCt^{a-1} \quad (30)$$

in which

- Y = the accumulated infiltration, L
- t = time, T
- C = a constant representing Y at unit time, dimensionally inconsistent

Again, as time becomes large, infiltration rate approaches zero by this equation. In a later article, Lewis and Milne (1938) presented Equation (32) in the following form:

$$Y = At + B(1 - e^{-bt}) \quad (34)$$

Upon differentiation, one obtains

$$I = A + bBe^{-bt} \quad (35)$$

The difference between the above equation and Equation (33) is that the former gives a zero final infiltration rate, whereas the latter gives a finite one which is the known case. However, it is important to emphasize that this is merely another empirical equation which happens to fit the data.

Horton (1939, 1940) has given Equation (35) considerable attention. His form of it is:

$$I = A + (I_0 - A) e^{-bt} \quad (36)$$

wherein

A = the final infiltration rate, L/T

I_0 = the initial infiltration rate, L/T

Horton derived his equation empirically. He takes no account of the action of capillary and gravity forces. Philip (1957a) developed a numerical solution for Equation (10). It results in an equation of the form

$$Z = \lambda_1 t^{1/2} + \lambda_2 t + \lambda_3 t^{3/2} + \dots \quad (37)$$

satisfy the requirements of predicting the infiltration behavior at both extremes of lapsed time.

Horton (1940) defined infiltration capacity as the maximum rate at which rain can enter into the soil. He also stated that the rate of infiltration is the actual rate at which the rainfall can be absorbed into the soil. This could be any rate from zero to the capacity rate, but never exceeding the rainfall intensity. Richards (1952) did not, however, make any distinction between infiltration rate and infiltration capacity, but defined both as being the same as Horton's definition for infiltration capacity. Linsley and Franzini (1964) made a distinction between infiltration rate and capacity by defining infiltration rate as the actual rate at which rainfall enters into the soil.

Among the earliest of the proposed infiltration laws was that of Green and Ampt (1911). These authors prefaced their treatment with a discussion of the saturated conductivity of soil based on the capillary tube model, but their subsequent analysis is subject to this only to the extent that they suppose, consistently with such a model, that the advancing water front is a precisely defined surface at which the pressure, negative because of suction, is a constant characteristic of the soil. This front separates uniformly saturated soil behind it, of uniform hydraulic conductivity K_s , from uniformly unsaturated and as yet uninfluenced soil beyond it. Such a supposition is, in fact, an assumption that the moisture characteristic is a step-shaped curve indicating the reduction of moisture content sharply from saturation to a constant degree of unsaturation at an air entry value and would be satisfied by a granular

body in which the pore shapes and sizes were quite uniform. Their equation is in the form:

$$\frac{P}{E} t = Z - (h_o + K_c) \ln \left(1 + \frac{Z}{H + K_c} \right) \dots (40)$$

in which

- h_o = the depth of water on soil surface, L
 Z = the depth of wet soil from the soil surface to the wetting front, L
 K_c = a constant depending on the capillary forces on the moving water soil boundary, dimensionless
 P = a constant equal to $C''p$
 C'' = $\frac{\pi g \rho}{8 \eta}$
 p = $\frac{\Sigma r^4}{A}$
 r = the average capillary radii, L
 \bar{A} = cross section-area, L^2
 E = the porosity (pore space), dimensionless
 t = time, T
 η = the viscosity of the fluid, FT/L^2
 g = gravity constant, L/T^2
 ρ = the density of fluid, FT^2/L^4

Green and Ampt (1911) suggested that the measurement of E , P and K_c is of more importance than, and should replace, the determination of the sizes of the soil particles as in the usual "mechanical analysis" of soils.

It has to be admitted that none but the most artificial of granular media, such as an array of spheres of uniform size, presents a moisture profile which exhibits the feature of a well-defined plane of separation, at a well defined suction, between saturated material behind the water front and unaffected material ahead of it. Nevertheless Green and Ampt did in fact demonstrate the applicability of their formula to the case of infiltration into a column of disturbed soil material.

Swartzendruber and Huberty (1958) used an equation similar to Green and Ampt's equation, regarded as wholly empirically based, and compared it with observed rates of infiltration into natural soil profiles. Agreement was generally surprisingly good for total infiltration of up to three inches of water, the surprise being that natural profiles are often far from being uniform as is assumed when the equation is derived on physical grounds.

Foster (1948) observed that soil porosity was the most important factor affecting the soil intake rate. He stated that cultivation has a most significant effect on the initial rate of infiltration, but little effect on the final rates due to clogging of the pores by small soil products. Foster also reported that the intake rate was greater where there was cover than where the soil was bare. Fletcher (1949) proposed the use of the Poiseuille approximation to separate the parameters involved in the infiltration process. He proposed the following equation:

$$q = \pi r^3 \frac{(\rho r g h + 2\sigma \cos \alpha)}{8Z\eta} \dots \dots \dots (41)$$

in which for a single cylindrical capillary

- q = the volume of flow per amount of time, L^3/T
 h = depth from water surface to wetting front, L
 σ = surface tension of wetting fluid F/L
 α = the contact angle between liquid and solid
 Z = the depth from soil surface to wetting front, L

The usefulness of Poiseuille's approximation is to highlight some of the properties which affect infiltration.

Fletcher considered the infiltration as the total amount of water entering the soil from the surface from the time of its addition to the end of the first hour. He reported that the wettability α is possibly a function of a large number of variables. However, Equation (41) has been derived on the assumption that the soil pores are cylindrical and the radius of the pore have some degree of uniformity.

Recently, Fletcher (1969) modified Equation (41) to convert it to unit area of soil and to give the capillary a cross section of a three cusped hypocycloid of equivalent cross sectional area

$$q = \frac{E_a r_{50}}{161Z\eta} (r_{50} g\rho h + 19.513 \sigma \cos \alpha) \quad . \quad . \quad . \quad . \quad (42)$$

wherein

- r_{50} = the mode of particle size radius, L
 E_a = percent of pores not filled with water, dimensionless

Letey, Osborn, and Pelishek (1962) used the contact angle between the water and soil as an index to the wettability or hydro-philic properties of the soil. They found that water extracts of chaparral litter would increase the contact angle which, in turn, would decrease the soil intake rate as stated theoretically by Fletcher (1949).

Bertoni, Larson, and Shrader (1958) studied soil infiltration rate changes as influenced by seasons of the year. They found that the final infiltration rate tends to increase up to June or July and then decreases rather sharply during August and September. Philip (1958) and Al-Abdulla (1965) have shown the dependence of intake rate on initial soil moisture content. They stated that increasing the initial moisture content results in a decreasing infiltration rate. According to their work, antecedent moisture in the soil has a noticeable effect on soil intake rate.

Hansen (1955) developed an equation relating infiltration rate to hydraulic conductivity, capillary potential, and depth of wetting front.

$$Q_t = K_T \bar{A} \frac{\psi_T + Z}{Z} \dots \dots \dots (43)$$

in which

Q = volume of flow, L^3

\bar{A} = cross section-area, L^2

K_T = the hydraulic conductivity of the transmission zone, L/T

ψ_T = the capillary potential in the transmission zone extrapolated to the wetting front, L

Z = the length of wetting from the soil surface to the wetting front, L

Hansen developed Equation (43) on the assumption that the soil is homogeneous and its structure does not change after wetting. Under conditions where ψ_T does not appreciably increase with time, the potential $\left(\frac{\psi_T}{A} + 1\right)$ approaches one as time and Z increase. Hence Equation (41) can be written:

$$I = \frac{Q}{A} = K_T \dots \dots \dots (44)$$

which represents the constant value the infiltration generally approaches during downward movement of water through soils.

Hansen (1955) stated, in general, that the rate of entry in moist soils is less than in drier soils and the wetting front advances more rapidly when the soil is wet.

Fok and Hansen (1966) assumed that throughout the transmission zone in a homogeneous, isotropic soil, the soil-moisture content and thus hydraulic conductivity remain constant. The wetting front is, in effect, a capillary fringe having a moisture content at the foremost part which approximates that of the original soil. By using these assumptions, they found that the physical properties of the soil, such as porosity, water holding capacity, antecedent moisture content, hydraulic conductivity, and capillary potential, could be related by utilizing a coaxial semi-logarithm plot to either the total or rate of infiltration. The following represents Fok and Hansen's equation:

$$\frac{K_T t}{ES\psi_T} = \frac{Z}{\psi_T} - \ln \left(1 + \frac{Z}{\psi_T} \right) \dots \dots \dots (45)$$

in which

ln = a natural log

S = the degree of saturation, dimensionless

E = the porosity of the soil, dimensionless

Keller (1967) developed a theory of water flow based on an analogy wherein the cross sectional flow area in the direction of flow is described by a single parabolic shaped pore-channel. He presented the following equation:

$$\bar{q} = 2 \int_0^{\bar{X}} \bar{u} Z dx \quad (46)$$

in which

\bar{q} = the flow per unit cross-sectional area of stationary material, L/T

\bar{u} = the average tangential flow velocity, L/T

X = the difference between the moisture content by weight, W, at which the flow takes place and the critical moisture content on a weight basis, W_0 , below which viscous flow is assumed to be negligible, dimensionless

Z = the vertical coordinate across the pore channel, L

Keller (1967) introduced the following equation to predict the basic infiltration rate, I_b :

$$I_b = \beta' \left(\frac{S_u \cdot E}{G} \right)^{\gamma'} \quad (47)$$

in which

E = the absolute porosity, dimensionless

S_u = the ultimate saturation achieved, dimensionless

G = the bulk density or apparent specific gravity, dimensionless

β' and γ' = constants

For the drainage cycle following complete saturation, $S_u = 1.0$, and for the imbibition cycle, $S_u \approx 0.85$.

Keller reported that Equation (45) may be used to predict I_b of a soil at any degree of compaction providing β' and γ' are known for the soil and a limiting degree of saturation, S_u , is assumed. Hanks (1965)

solved Equation (10), which is a combination of Darcy's law and continuity equation, numerically by the aid of a high speed digital computer to predict the infiltration rate. Hanks considers the influence of the many factors that affect infiltration from the point of view of the effect these factors have on the basic soil properties. Hanks stated five categories of information are needed to compute the infiltration:

1. The relationship between capillary conductivity, K , and water content, θ .
2. The relationship between capillary potential, ψ , and water content, θ .
3. The water content, θ , and the capillary potential, ψ , at the soil surface at the start of infiltration.
4. The depth and homogeneity of the soil.
5. The water content, θ , at the various depths at the beginning of infiltration (initial moisture conditions).

Hanks assumed the relationship between K and θ to be unique to predict the effect of initial moisture content on infiltration. The procedure showed the effect of dry and wet soil on infiltration, but failed to distinguish the effect of the other initial moisture content which lies in between.

Hanks and Bowers (1963) have shown that small changes in D values at high moisture contents effect a striking change in calculated infiltration rates, whereas relatively large changes in D (diffusivity) at low moisture contents have little effect on the calculated rates. Thus, it is essential that accurate measurements of D vs. θ or K vs. θ be obtained near moisture saturation if the numerical procedure is to yield accurate estimates of infiltration rate.

Youngs (1968) introduced an analysis of water uptake by the moment method. From Equation (6) the quantity of water, Q , held in the porous material per unit cross section of the column between Z_1 and Z_2 , is given by:

$$\frac{1}{A} \left(\frac{dQ}{dt} \right) = - \left(K \frac{\partial \Phi}{\partial Z} \right)_1 + \left(K \frac{\partial \Phi}{\partial Z} \right)_2 \quad \dots \quad (48)$$

and the moment M about Z_1 of the water held between Z_1 and Z_2 by:

$$\frac{1}{\rho g} \frac{dM}{dt} = - \int_{Z_2}^{Z_1} K \frac{\partial \Phi}{\partial Z} dZ + (Z_2 - Z_1) \left(K \frac{\partial \Phi}{\partial Z} \right)_2 \quad \dots \quad (49)$$

In Equations (48) and (49) the subscripts 1 and 2 refer to values at the positions Z_1 and Z_2 respectively.

While Equation (49) is quite general and has been suggested by Zaslavsky and Ravina (1965) to be used in experiments designed to measure hydraulic conductivity values of a porous material, it takes the simpler form:

$$\frac{1}{\rho g} \frac{dM}{dt} = - \int_{\Phi_1}^{\Phi_2} K d\Phi \quad \dots \quad (50)$$

Philip (1968) applied his theory of absorption on aggregated media. He regarded the aggregated medium as one made up of macroporosity, through which flow on the Darcy scale occurs, and microporosity, which is free to exchange water with the macroporosity. With the aid of the two models he used, he concluded, firstly, it is evident that for t sufficiently

small, the amount of absorption into the microporosity is negligibly small, and the phenomenon is indistinguishable from that of sorption into a classical medium with sorptivity S_p . Hence, for t small

$$Y_1 \approx S_p t^{1/2} \quad \text{and} \quad Y \approx S_p t^{1/2} \quad (51)$$

in which

Y_1 = cumulative net absorption into the macroporosity, L
 Y = the total cumulative absorption, L

Secondly, when t is sufficiently large, disequilibrium between the state of water in the macroporosity and in the microporosity is confined to a very small part of the total wetted region, and the aggregated medium can be expected to behave like a classical medium with a definite apparent sorptivity.

Philip (1957d, 1968) reported that the sorptivity of soils normally lies in the range 0.02 - 0.2 cm/sec^{1/2}. Under the assumption that the characteristic dimension of the microporosity is one-hundredth of that characteristic of the (macro) pores contributing to the sorptivity of such soils, one may estimate that the sorptivity of the microporosity lies in the range 0.002 - 0.02 cm/sec^{1/2}. Bodman and Colman (1943) and Colman and Bodman (1944) distinguished five zones in the soil during infiltration, which are as follows:

1. The saturated zone is a zone reaching a depth of about 1.5 centimeters;
2. The transition zone is a zone of about 5 centimeters in which a rapid decrease in moisture occurs;

3. The transmission zone is a zone in which moisture content is nearly constant;
4. The wetting zone is a zone of fairly rapid change in moisture content; and
5. The wetting front is a zone of very steep moisture gradients which shows a visible limit of moisture penetration into the soil.

The infiltration zones of Bodman and Colman are shown in Figure 1. The saturation and transition zone shown is usually presented in soils, but is not observed with porous materials of uniform particle size in experimental situations where there is little possibility of air being trapped under the downward advancing wetting front (Youngs, 1957; Childs, 1969; and Hanks, 1969).

Figure 2 shows the stages in the development of the moisture profile into an initially dry column of slate dust (Childs, 1969).

In a series of papers Philip (1954,1958) has shown how the features of the infiltration moisture profile illustrated in Figure 1 and Figure 2 (with the exception of the near-surface saturation and transition zone of Figure 1) can be understood. Furthermore if the dependence of soil-water suction Ψ and hydraulic conductivity K on water content θ is known, and is assumed unique, the progress of the moisture profile into a uniform incompressible soil can be quantitatively determined using Equation (10).

Nielsen et al. (1960) provided a comparison on two soils in the field which possessed fairly uniform properties to the depths of wetting investigated. He attributed the lack of agreement, as was found between measured and calculated profiles, to the assumptions of the theory not being satisfied.

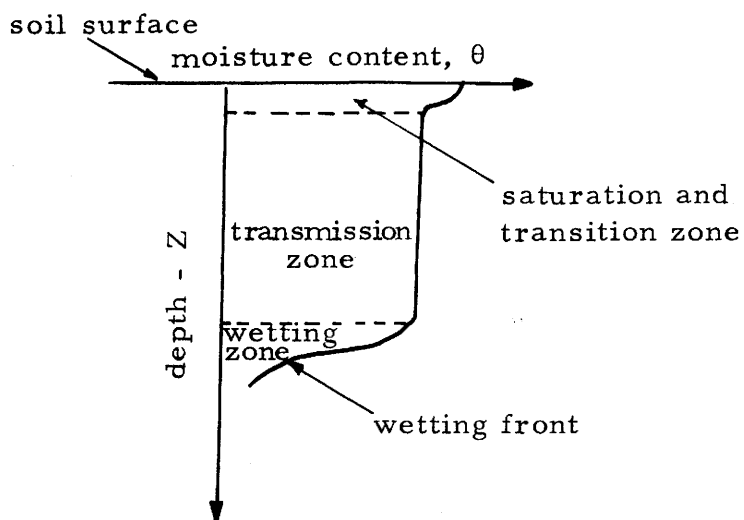


Figure 1. Zones of the moisture profile during infiltration as described by Bodman and Coleman (1943).

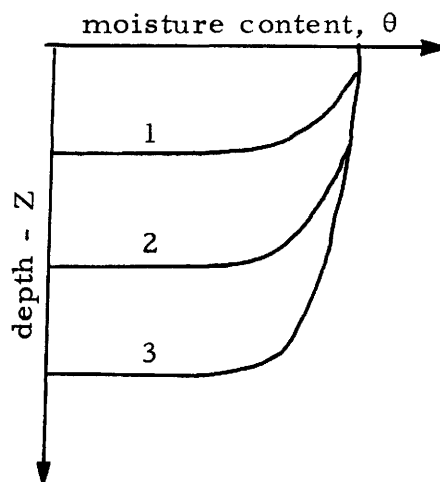


Figure 2. The development with time of moisture profile during vertical infiltration, Childs (1969).

Rain (sprinkler) infiltration

Linsley, Kohler, and Paulhus (1949) reported that rainfall intensity has little effect on the rate of infiltration when it exceeded the capacity rate. This disagrees with findings of Fletcher (1960). They further report that different slopes between 1 to 16 percent have little noticeable effect on infiltration rates.

Wisler and Brater (1959) reported that changes in the viscosity of the rainfall due to changes in its temperature influence the rate of infiltration because the flow in the interstitial space is nearly always laminar. This was also proposed in Fletcher's (1949) equation and in the field data for 72 measurements he presented. However, Butler (1957) reported that the net effect of temperature change is less than might be expected. This can be rationalized by noting that a rise in soil temperature is accompanied by increased microorganism activity, lowered surface tension, and swelling of the clay which, in turn, partly nullify the increase in infiltration capacity resulting from lowered viscosity.

Willis (1965) investigated the effect of the kinetic energy of rainfall. He reported that the rainfall infiltration rate of a bare soil was reduced by an increase in the kinetic energy of the rainfall. He further stated that the time at which the runoff began, and the rate of infiltration began to decrease, became less as the kinetic energy of the rainfall (which is a function of the velocity of impact of the raindrops and of the rainfall intensity) was increased.

Neal (1938) showed the effect of the degree of slope and rain intensity on runoff and soil erosion. He stated that the infiltration rate was increased with an increase in rain intensity. His conclusion stated that

Abramov stated that \underline{C} (constant) is greatly influenced by the state of moisture in the upper layer before irrigation. \underline{C} increases rapidly when there is greater moisture at the beginning. He further stated that the intake rate is significantly influenced by the state of moisture of the upper layer before irrigation. It is much greater when the moisture at the start is high. Figure 3 shows graphically the time, t_s , at which runoff occurs according to Abramov.

Neyestani (1968) supported Abramov and stated that the time at which rain induced ponding will occur is the same as the time at which

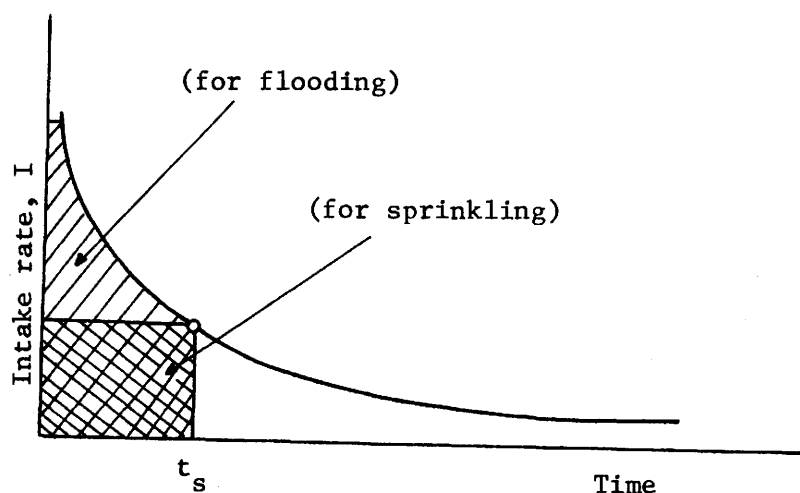


Figure 3. t_s -determination based on intake rate curve for both flooding and sprinkling.

the flood infiltration rate would have been equal to the rainfall rate.

In equation form:

$$I = aCt^{a-1} \dots \dots \dots (30)$$

or

$$I = N t^n \dots \dots \dots (54)$$

in which

N = a constant representing I at a unit time, dimensionally inconsistent

n = (a-1), which represents the slope of infiltration rate on log-log paper, dimensionless

Then the time, t_s , is:

$$t_s = (R/N)^{1/n} \dots \dots \dots (55)$$

Ram (1960) discussed the difference between ponded infiltration and rain infiltration. He stated that both practice and theory show that for an interval of time the depth of infiltration with ponded irrigation can be 1.4 to 5 times greater than with sprinkler irrigation. According to his discussion, the irrigation level with both gravitational and sprinkler irrigation methods approximately follows Kostaikov's (1932) equation.

Ram anticipated that soil intake rate, I, should be equal to rain intensity, R, when the rain intensity curve intersects the flood intake rate curve. At that time, therefore, the relation between soil intake rate and rain intensity can be written

$$I = R \dots \dots \dots (56)$$

with the principle of continuity, the equation for cumulative rain infiltration is

$$Y_R = R t \dots \dots \dots (57)$$

where

Y_R = the cumulative rain infiltration, L

Ram inserted the value of I for R from Equation (54) into Equation (57) and concluded that

$$Y_R = Rt = It = (Nt^n) t = Nt^{n+1} \dots \dots \dots (58)$$

Comparison of Equations (29) and (58) indicates that depth of rain infiltration is always less than flood infiltration. Ram, however, presents neither data nor experimental evidence to prove his theory. Rubin and Steinhardt (1963) showed that, in the case of rain infiltration, the non-linear differential Equation (10) is subject to the following initial and boundary conditions:

$$t = 0, Z = 0, \theta = \theta_i = \text{constant} \dots \dots \dots (59)$$

$$t > 0, Z > 0, \frac{\partial \theta}{\partial Z} = - \frac{R - K(\theta)}{D} \dots \dots \dots (60)$$

in which

- θ_i = the initial moisture content, dimensionless
- $K(\theta)$ = the capillary conductivity at moisture content θ , L/T
- R = sprinkler application rate or rain intensity, L/T

Equation (60) expresses the fact that, during rain infiltration, soil moisture flux at the surface is equal to rain intensity. Rubin and Steinhardt (1963) have proposed theoretically and demonstrated experimentally that if during rain infiltration

$$0 < R \leq K_s$$

in which

K_s = the capillary conductivity at saturation

and the media is homogeneous:

1. As time passes, the moisture gradients at all finite depths of any rain infiltration profile approach zero.
2. As time passes, the moisture contents at all finite depths of any rain infiltration profile will approach a constant value.
3. Rain infiltration may continue indefinitely into a deep profile without causing ponding.

Rubin and Steinhardt (1963) also proposed the following concerning rain infiltration. If during rain infiltration $R > K_s$ and the media is homogeneous:

1. As time passes the surface moisture content of the soil increases and it must reach the saturation moisture content, θ_s , after rain infiltration is allowed to proceed for a finite period of time.
2. Following this attainment of saturation, ponding ensues.
3. When ponding occurs, the accumulative water uptake cannot be greater than the flood-water uptake for the same period of time.

Proof of the first and second part of the later proposition by Rubin is of interest. Because of the monotonic properties of hydraulic conductivity, $K_{(\theta)}$, when $\theta_o < \theta_s$, $K_{(\theta_o)} < K_s$, in which $K_{(\theta_o)}$ is the value of K at the soil surface moisture content, θ_o . But the hypothesis of the proposition under consideration states that $K_s < R$. Hence, because of Equation (60), $\frac{\partial\theta}{\partial Z}$ of the soil surface is less than zero $[(\partial\theta/\partial Z) < 0]$ throughout the rain infiltration under consideration. However, it is obvious that if $(\partial\theta/\partial Z) < 0$, θ_o must increase because $\theta_o \leq \theta_s$. Hence, if the infiltration under consideration can continue indefinitely, limit θ_o exists. Therefore, in such a case, Philip's proof (1957b) is applicable. Thus it must be concluded that limit $(\partial\theta/\partial Z)_o = 0$. This result cannot be true, because at all times

$$(\partial\theta/\partial Z) \leq [K_s - R]/D_s = \text{constant} < 0,$$

in which

$$D_s = \text{soil diffusivity at saturation}$$

The above contradiction indicates that the process under consideration cannot continue indefinitely. Also, it is clear from such a result that θ_s as well as the lower moisture contents cannot constitute limit θ_o . But it is known that θ_o increases if $(\partial\theta/\partial Z)_o < 0$, i.e., if $\theta_o < \theta_s$. It follows that θ_o will become equal to θ_s after rain infiltration is allowed to proceed a finite period of time. This conclusion constitutes the first part of the last proposition. It was shown previously that as long as $(\partial\theta/\partial Z)_o < 0$ the surface moisture content must keep on increasing with time and simultaneously the surface moisture gradient must continue becoming less negative. It might be noted that in the cases considered

until now these two changes were off-setting each other's influence upon the surface moisture flux. This fact made it possible for the surface flux to remain constant and equal rain intensity. Since $(\partial\theta/\partial Z)_o$ must continue becoming less negative, such a change, when not accompanied by a surface moisture content increase, must result in ponding. This conclusion constitutes the second part of the last proposition.

Philip (1957d,1958) showed theoretically and proved experimentally that if during downward infiltration, the surface suction, ψ_o , exceeds the air-entry suction, ψ_A , and is kept constant, the infiltration velocity decreases and approaches the saturation hydraulic conductivity, K_s . Therefore, Rubin and Steinhardt (1964) concluded that under a constant infiltration velocity, $R > K_s$, the surface suction, ψ_o , decreases with time without limit. The proof of this contention begins with an expression of Darcy's law:

$$R = K_s \left(\frac{\partial\psi}{\partial Z} + 1 \right)_o \dots \dots \dots (61)$$

or

$$\left(\frac{\partial\psi}{\partial Z} \right)_o = \frac{R - K_s}{K_s} \dots \dots \dots (61a)$$

Rubin and Steinhardt (1964) showed that if a constant $R > K_s$ exists and the soil has a positive air-entry suction, ψ_A , the surface suction, ψ_o , must decrease to ψ_A within some finite time, t_A . At some later time which they called time of saturation, t_s , ψ_i must become equal to zero.

Maintenance of flux R at times exceeding t_s requires negative suction (i.e., positive hydrostatic pressures) at the soil surface. At time t_s commencement of ponding of rain water will occur since at $t > t_s$ the

Rubin and Steinhardt have shown experimentally that the ratio between the corresponding rain and floodwater uptake at incipient ponding (stage B) for Rehovot sand as used in their experiment was 0.55. An inspection of Rubin's work will show that the value of 0.55 is the slope of the flood infiltration curve or the a-value of Kostaiikov's equation.

Rubin, Steinhardt and Reiniger (1964) showed that when rain intensity is smaller than the saturated soil's hydraulic conductivity, $R < K_s$, Philip's theory (1957a) for infinite-time flood water infiltration is applicable to the infinite-time rain advance of the wetting front during rain infiltration and can be represented as follows:

$$\left(\frac{dZ}{dt}\right)_R = \frac{R - K(\theta_i)}{\theta_L - \theta_i} \dots \dots \dots (66)$$

in which

$(dZ/dt)_R$ = the advance rate of the wetting front for rain infiltration,
L/T

θ_L = the limiting transmission-zone moisture content, dimensionless

They have shown that in the case where $R > K_s$, that

$$\theta_L = \theta_s \dots \dots \dots (67)$$

and when the initial soil moisture contents, θ_i , is very low, it can be assumed without significant error that

$$K(\theta_i) \approx 0 \dots \dots \dots (68)$$

Inserting values of Equations (67) and (68) into Equation (66), one obtains the following expression for the wetting front advance rate for

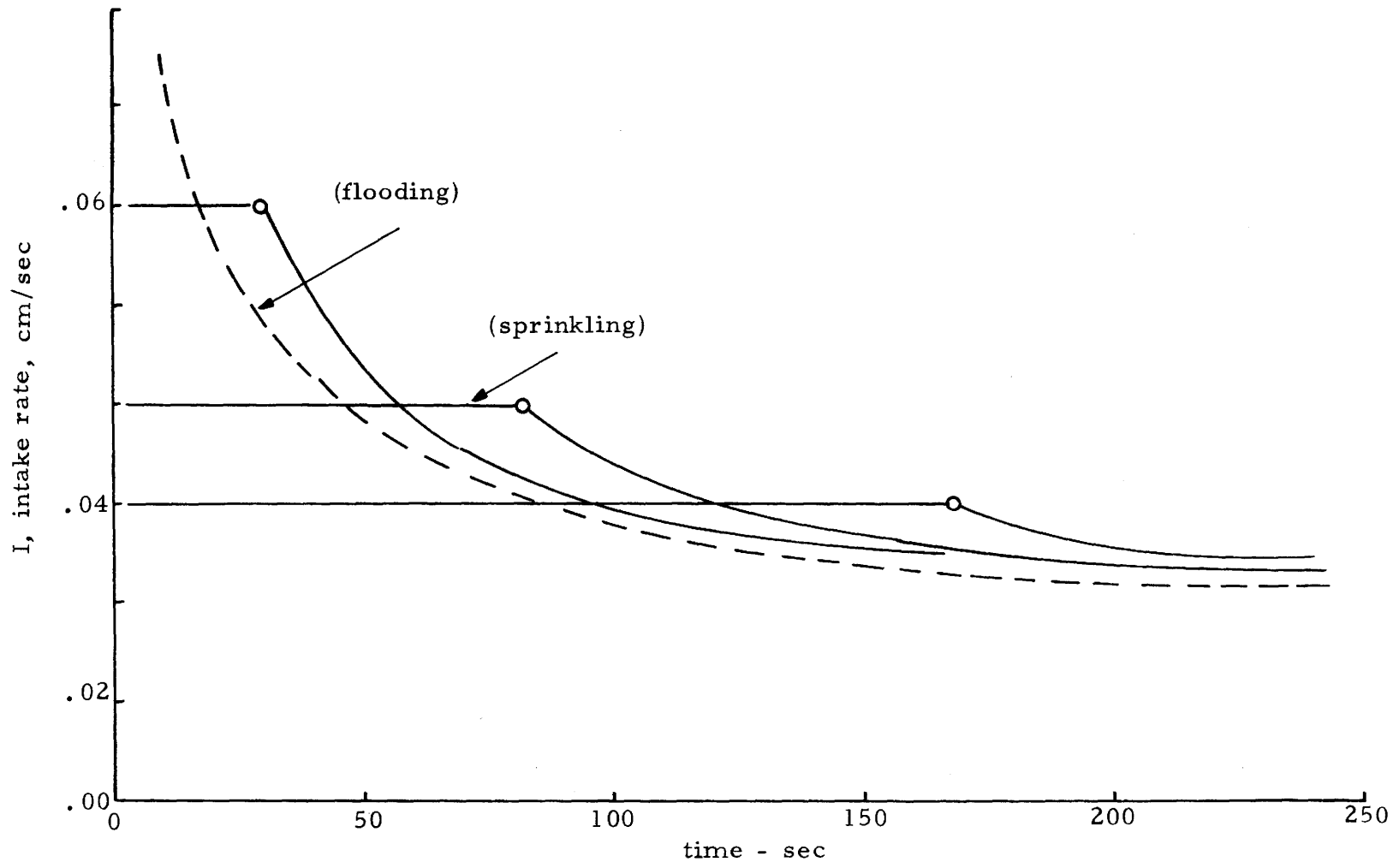


Figure 4. Relations between infiltration rates and infiltration duration during water uptake by initially air-dry Rehovot sand. The circles represent t_s .

Whisler and Klute (1967) extended the analysis of Rubin and Steinhardt (1963, 1964) to more complex situations. A nonuniform water content, nonuniform pressure head and hysteresis were included. They introduced a numerical analysis, carried on by computer, for the nonlinear partial differential flow Equation (10). Hanks (1969) developed a modified numerical method for Equation (10) considering the hysteresis effect in the suction vs. water content relationships, which allowed quantitative treatment of the effect of wetting rates on infiltration, redistribution and evaporation. The soil profile was assumed to be uniform with respect to its hydraulic properties.

CHAPTER 3

THEORY

The infiltration subject in the literature review has been divided into flooded (ponded) infiltration and rain (sprinkler) infiltration, which may at first thought appear to be rather disconnected subjects. For arable soils, however, they are brought together because one may need the intake rate curve (instantaneous flooded infiltration) of the soil to plan his sprinkler irrigation without causing runoff.

The theory was derived for water flow or infiltration into unsaturated soil possessing appreciable amounts of silt and clay and initially drier than its field capacity.

The following assumptions were made in developing the theory dealing with rain and/or flood infiltration into a vertical soil column:

1. The soil is regarded as a semi-infinite, homogeneous, isotropic body, the bulk density is uniform through the profile and remained constant during watering.
2. One-dimensional flow is assumed in the system.
3. The initial moisture content is assumed to be uniform throughout the profile.
4. Soil-air is regarded as a continuous phase and essentially at atmospheric pressure.
5. The water application rate is considered to be constant throughout the watering and great enough to eventually cause surface ponding or flooding.

6. The kinetic energy of the falling rain drops is so small that surface disturbance will be negligible.

Flooded (Ponded) Infiltration

The one-dimensional flow of water in unsaturated soil obeys Darcy's law:

$$V = -K \frac{\partial \phi}{\partial Z} \dots \dots \dots (2)$$

in which

- V = the flow velocity, L/T
 K = the capillary conductivity, L/T
 Z = the depth of wetting front, L
 ϕ = the potential, L

Assuming also, that flow obeys the law of conservation of matter which is expressed in the equation of continuity:

$$\frac{\partial V}{\partial Z} = - \frac{\partial \theta}{\partial t} \dots \dots \dots (5)$$

Substituting Equation (2) into Equation (5), one obtains the general form of the unsaturated flow equation in the vertical direction in a soil:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial Z} \left(K \frac{\partial \phi}{\partial Z} \right) \dots \dots \dots (6)$$

in which

- θ = the moisture content on volume basis, dimensionless
 t = time, T

The potential ϕ is given by:

$$\phi = \psi + Z$$

initial moisture contents should be connected separately instead of running one unique curve through all of them. Each curve itself is considered unique; and should be used in the numerical solution under the same initial circumstances of moisture content, at which it was obtained, to compute the ponded infiltration. In this study, the Crank-Nicholson implicit technique, adapted to include gravity, was used to solve Equation (7). This method was developed by Hanks and Bowers (1962).

The numerical solution for vertical infiltration can be written as

$$\frac{\theta_i^j - \theta_i^{j-1}}{\Delta t} = \frac{(\psi_{i-1}^{j-1} + \psi_{i-1}^j + \bar{g} - \psi_i^{j-1} - \psi_i^j) K_{i-\frac{1}{2}}^{j-\frac{1}{2}}}{2(\Delta Z)^2} - \frac{(\psi_i^{j-1} + \psi_i^j + \bar{g} - \psi_{i+1}^{j-1} - \psi_{i+1}^j) K_{i+\frac{1}{2}}^{j-\frac{1}{2}}}{2(\Delta Z)^2} \dots \dots \dots (71)$$

in which

\bar{g} = the gravitational term, being equal to ΔZ for the vertical infiltration

The subscripts "i" and "j" refer to distance and time, respectively.

An equation can be written for each depth interval involving unknowns of

$$\theta_i^{j-1} \quad \text{and} \quad \psi_i^{j-1}, \quad i = (1, 2, 3, \dots n-1)$$

The boundary condition will provide values of

$$\psi_o^j, \psi_o^j \text{ and } \theta_n^j$$

A series of n equations can be formed, but they will have more than n unknowns. These equations cannot be solved, hence additional information is required.

The derivation of Equation (7) assumes a unique relationship between capillary potential, ψ , and moisture content, θ . Based on this assumption,

$$\frac{\theta_i^j - \theta_i^{j-1}}{\Delta t} \approx \frac{\psi_i^j - \psi_i^{j-1}}{\Delta t} \cdot H_i^{j-\frac{1}{2}} \dots \dots \dots (72)$$

where the specific moisture capacity, H, is defined as

$$H_i^{j-\frac{1}{2}} = \left(\frac{\partial \theta}{\partial \psi} \right)_i^{j-\frac{1}{2}}$$

Substitution of Equation (72) into Equation (71) yields the final working equation:

$$\begin{aligned} \frac{\psi_i^j - \psi_i^{j-1}}{\Delta t} &= \frac{(\psi_{i-1}^{j-1} + \psi_{i-1}^j + 2\bar{g} - \psi_i^{j-1} - \psi_i^j) K_{i-\frac{1}{2}}^{j-\frac{1}{2}}}{2(\Delta Z)^2 H_i^{j-\frac{1}{2}}} \\ &- \frac{(\psi_i^{j-1} + \psi_i^j + 2\bar{g} - \psi_{i+1}^{j-1} - \psi_{i+1}^j) K_{i+\frac{1}{2}}^{j-\frac{1}{2}}}{2(\Delta Z)^2 H_i^{j-\frac{1}{2}}} \\ &\dots \dots \dots (73) \end{aligned}$$

The cumulative infiltration, Y , obtained from the numerical methods can be described by the empirical equation:

$$Y = C t^a \quad (29)$$

and the infiltration rate by:

$$I = N t^n \quad (54)$$

in which

- C = a constant representing Y at unit time, dimensionally inconsistent
- a = a constant representing the slope of the cumulative infiltration curve on log-log paper, dimensionless
- N = (aC) , constant representing I at a unit time, dimensionally inconsistent
- n = $(a-1)$, which represents the slope of infiltration rate on log-log paper, dimensionless

Rain (Sprinkler) Infiltration

Assuming steady state flow in the vertical direction in a soil,

V is a constant

One can write the following equation:

$$V = \left(\frac{dZ}{dt} \right)_R (\theta_T - \theta_i) \quad (75)$$

in which

$\left(\frac{dZ}{dt} \right)_R$ = the advance rate of the wetting front for rain infiltration, L/T

θ_i = the initial moisture content, on a volume basis, dimensionless

θ_T = the moisture content in the transmission zone, dimensionless

Darcy's law, Equation (2) can be written:

$$v = K_T \frac{\partial \phi}{\partial Z} \dots \dots \dots (76)$$

in which

K_T = hydraulic conductivity of the transmission zone, L/T

or

$$v = K_T \frac{h_T + Z}{Z} \dots \dots \dots (77)$$

in which

h_T = equal to $d\psi$, head loss in transmission zone, L

Substituting Equation (77) into Equation (75) one obtains:

$$\frac{dZ}{dt}_R = \frac{K_T}{(\theta_T - \theta_i)} \left(\frac{h_T + Z}{Z} \right) \dots \dots \dots (78)$$

By integrating the equation with initial condition $Z = 0$, $t = 0$, and assuming the capillary potential, ψ , is constant along the transmission zone ($h_T = 0$), and letting $K_T = R$:

$$\left(\frac{dZ}{dt} \right)_R = \frac{R}{\theta_T - \theta_i} \dots \dots \dots (79)$$

$$\int_0^Z dZ = \int_0^t \frac{R}{(\theta_T - \theta_i)} dt$$

$$Z = \frac{Rt}{\theta_T - \theta_i} \dots \dots \dots (80)$$

Equation (80), so far, has been derived on the assumption of unit hydraulic gradient between any two points in the transmission zone which

is followed by a sharply defined wetting front, with slope zero, assuming the depth of wetting zone to be zero.

In other words, Equation (80) has been derived by using piston flow characteristics. But, in reality, especially with a soil having appreciable amounts of clay, that is not quite the case. Usually, the moisture content of the transmission zone decreases slightly with depth and there exists a zone, with decreasing moisture content, between the transmission zone and wetting front, in agreement with the observations (Bodman and Colman, 1943). Also, $(\theta_T - \theta_i)$ is not quite independent of Z and t , since the moisture content of the transmission zone slightly decreases with depth and increases with time.

If one takes all the above considerations into account, Equation (80) can be written:

$$Z = \left(\frac{Rt}{\theta_T - \theta_i} \right)^{n'} \dots \dots \dots (81)$$

and letting the value

$$\left(\frac{R}{\theta_T - \theta_i} \right)^{n'} = C' = \text{constant} \dots \dots \dots (82)$$

Since, θ_T actually takes different values depending on the initial moisture content, θ_i , and the rate of application, R , one can consider C' as:

- a constant mainly dependent on sprinkler intensity, R , and initial moisture content in the soil, dimensionally inconsistent

Equation (81) can be written:

$$Z = C't^{n'} \dots \dots \dots (83)$$

in which

n' = a constant having values between 1.0 and 0.80 dependent on the physical properties of the soil, or soil type

Plotting Z as a function of t on log-log paper, one expects to obtain a series of parallel lines for different values of R . On the other hand, plotting C' at a unit time as a function of R , a series of parallel lines for different values of θ_i are expected. Mathematically, this may be expressed as:

$$C' = A'(\theta_i) R^m \dots \dots \dots (84)$$

in which

m = the slope of parallel lines as a result of plotting C' as a function of R , dimensionless

$A'(\theta_i)$ = a constant dependent on the initial moisture content, dimensionally inconsistent

The constant $A'(\theta_i)$ can be obtained from the following relationship:

$$\frac{A'(\theta_i)_1}{A'(\theta_i)_2} = \frac{(\theta_i)_1}{(\theta_i)_2} \dots \dots \dots (85)$$

Equation (83) can be written in a general form:

$$Z = A'(\theta_i) R^m t^{n'} \dots \dots \dots (86)$$

and the advance rate of the wetting front can be obtained by:

$$\left(\frac{dZ}{dt}\right)_R = A'(\theta_i) n' R^m t^{n'-1} \dots \dots \dots (86a)$$

$$t_s = (R/C)^{\frac{1}{a-1}}$$

$$t_s = (R/C)^{1/n} \dots \dots \dots (91)$$

in which

C = a constant representing cumulative flooded infiltration,
Y, at unit time, dimensionally inconsistent

n = a constant representing the slope of flooded (ponded)
infiltration rate on log-log paper, dimensionless

Figures 5 and 6 show graphically the time, t_s , at which surface ponding occurs according to the above theory.

From Equation (91), t_s depends on water application rate, R, and since $(1/n)$ is a negative exponent, one may conclude that the time at which ponding will occur, will decrease as the water application rate increases.

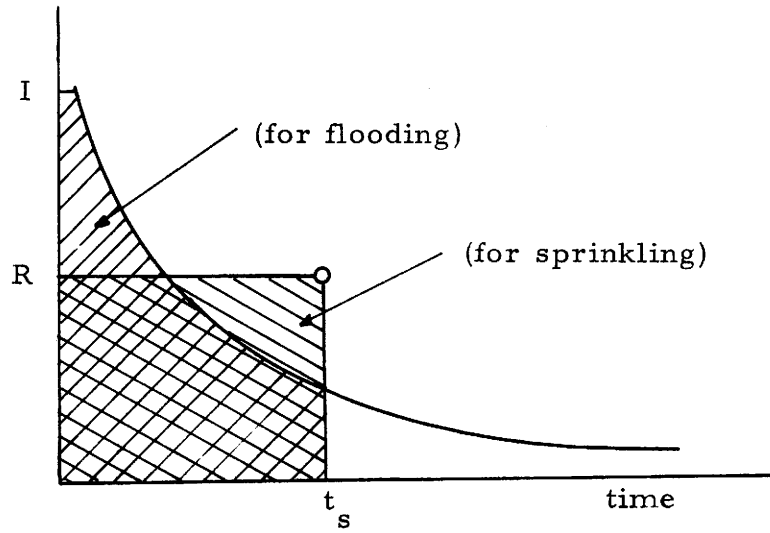


Figure 5. t_s - determination based on intake rate curve for both flooding and sprinkling.

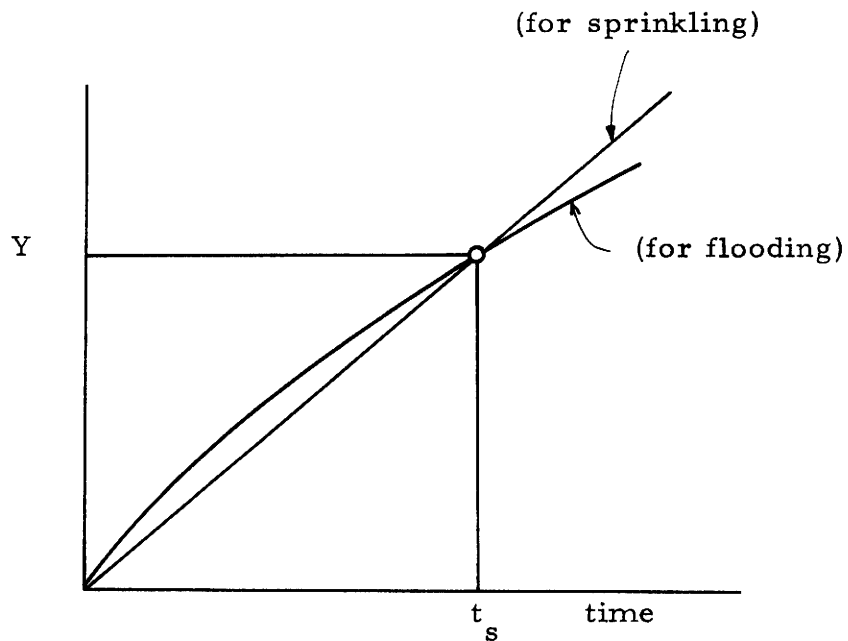


Figure 6. t_s - determination based on cumulative intake curve for both flooding and sprinkling.

CHAPTER 4

METHOD OF PROCEDURE

The experiment was conducted in the laboratory on soil columns contained in 5-cm diameter by 30 to 90-cm long lucite cylinders. The soil, a Nibley silty clay loam, was carefully sifted, passed through a 2-mm screen, mixed, and stored to assure samples for various experimental runs and replications which were as uniform as possible both chemically and mechanically.

The mechanical analysis of the soil along with other pertinent information is given in Table 1.

Table 1. Chemical and physical composition of the Nibley silty clay loam soil.

Characteristic	Diameter	Percent
Clay	< 2 micron	38.0
Silt	2-50 micron	58.9
Sand		
a.	0.05 - 0.1 mm	2.3
b.	0.1 - 0.25 mm	0.8
c.	0.25 - 2.0 mm	-
Organic matter	-	2.32
Soluble salts	-	0.15
CaCO ₃ Equiv.	-	1.7

Soil Containers

The soil container, shown in Figure 7, used in this experiment was composed of transparent plastic rings 5 centimeters inside diameter and

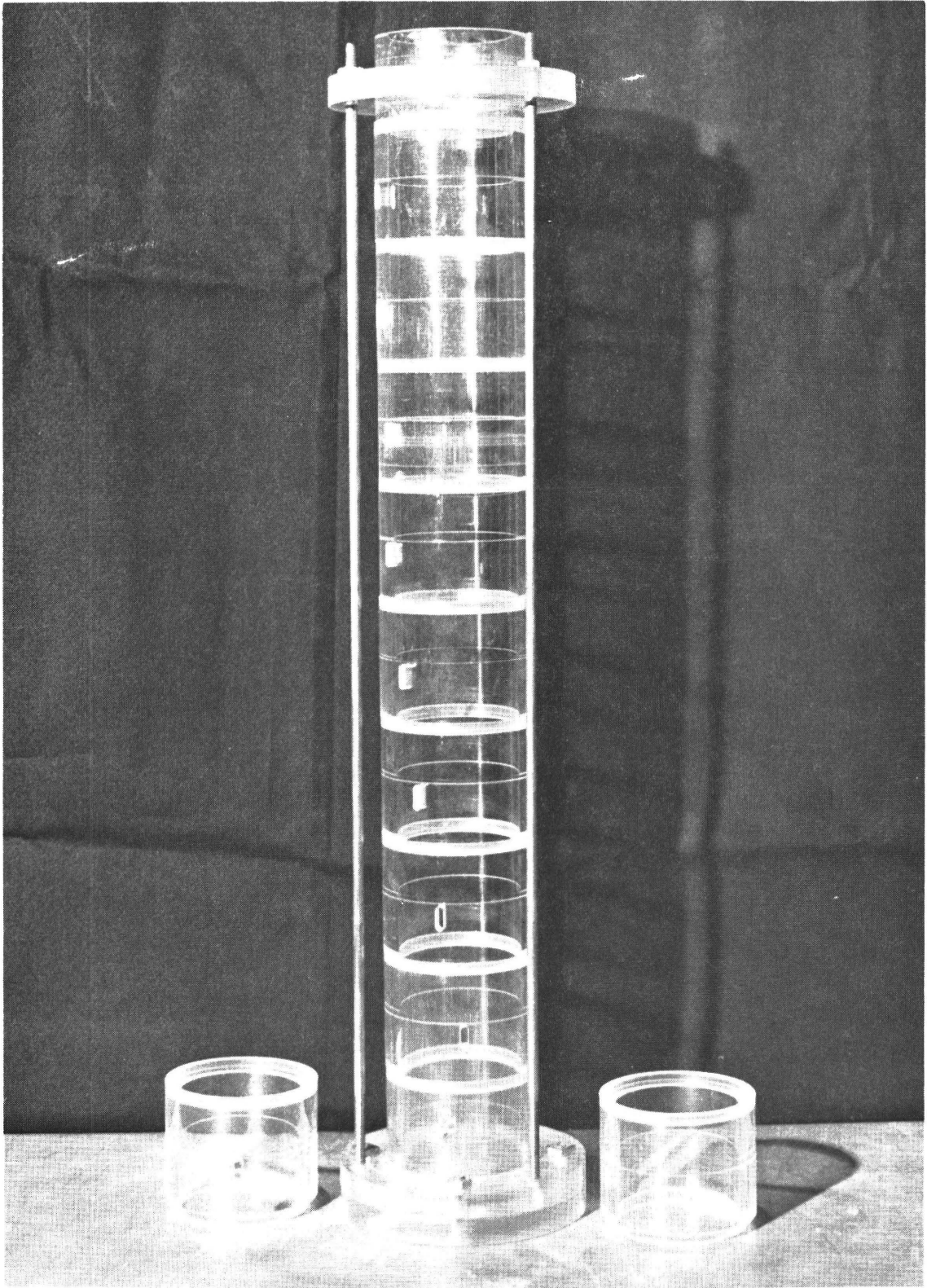


Figure 7. Soil container with porous plate attached to its base.

5 centimeters in height. By superimposing varying numbers of rings (sections), columns of different depths could be made. A maximum of 18 rings were used in the investigation thus making a soil column of 90 centimeters.

A slot, elongated hole, 3 mm wide and 9 mm high, was drilled at the center down the side of each ring. The purpose of these holes was to provide openings for moisture tensiometers and to function as air vents to reduce the entrapped air inside the soil, thus to satisfy assumption 4. Figure 7 shows an empty lucite cylinder with the holes described above. A stainless steel porous plate, which had bubbling pressure in excess of 500 centimeters of water, was attached to the base of the bottom ring to provide a means for drainage. The remainder of the drainage apparatus, to be used if needed, consisted of a plastic base which was fastened to the porous plate. A circular ridge had been cut from the plastic base, leaving a small chamber between the plate and its base. This chamber was vented by means of a rigid plastic tube, and a constant vacuum could be applied to this outlet.

Water Applicator Apparatus (Rain Simulator)

The water applicator used in the experiment was designed to uniformly sprinkle the soil surface at any constant application rate between 0.20 and 10.0 ± 0.05 centimeters per hour. Essentially it consisted of a water supply reservoir, rotating lids, and pumps.

The water supply reservoir had a one gallon capacity and was provided with a constant head syphon which connected, through a laboratory pump, to a rotating lid by means of flexible tubing and a swivel connection.

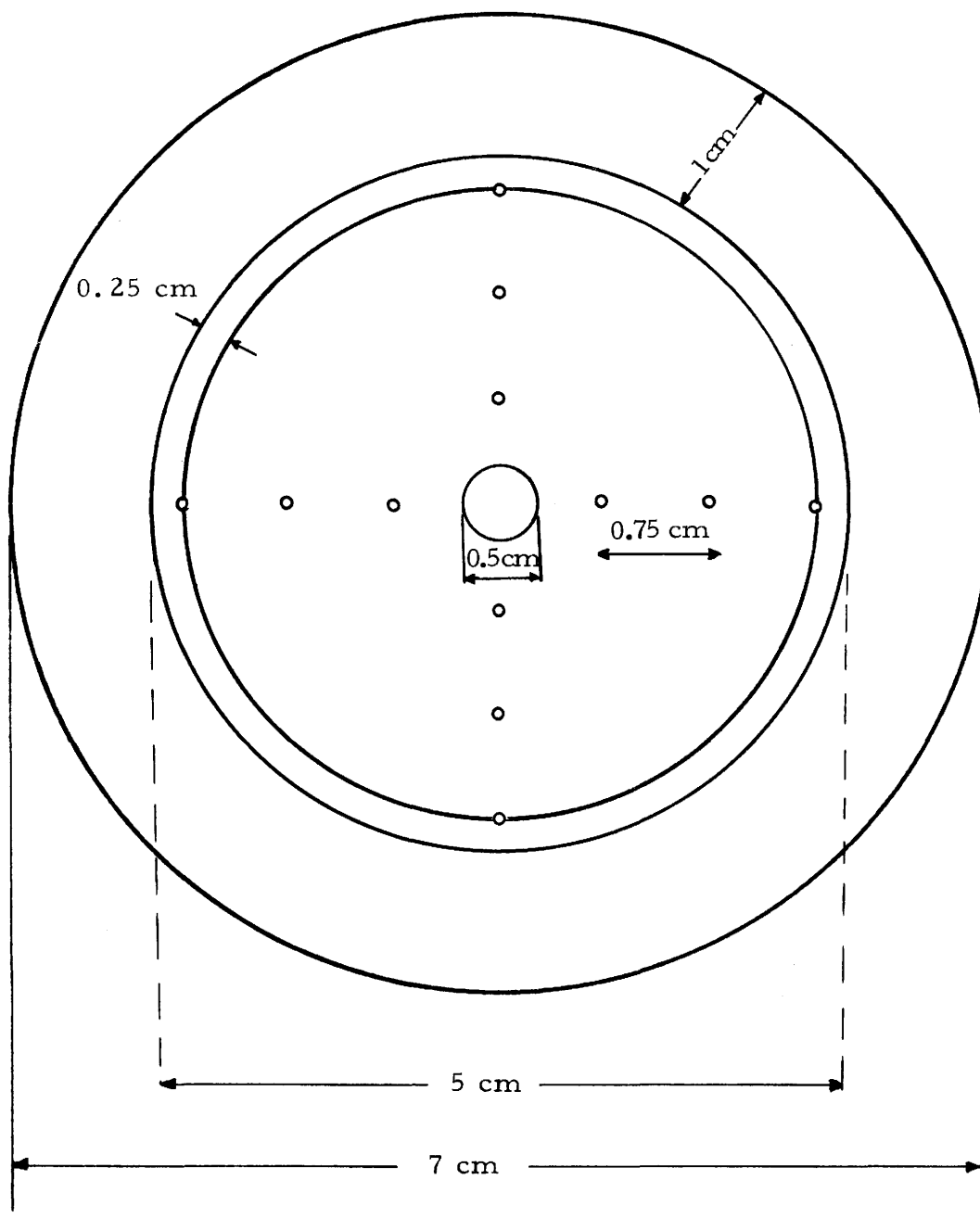
The laboratory pump was a standardized version of the (rotating and reciprocating) piston pump concept, designed to obtain flow at various constant rates, ± 0.05 cm/hr, by moving the flow rate indicator along the flow rate scale, thus assumption number 5 was satisfied. The clear plastic rotating lid, shown in Figure 8, was provided with 12 polyethylene capillary tubes of 0.279 millimeters inside diameter, 0.610 millimeters outside diameter, and 30 centimeters long. These capillary tubes were held one centimeter above the soil surface producing a minimum impact to the soil surface. In order to assure uniform water distribution, the lid was rotated at one rpm during sprinkling by a means of a rotating pump.

Figure 9 shows a soil-filled lucite cylinder in position under the rotating water applicator lid.

Attention is directed to the tiny inside diameter (0.279 mm.) of the polyethylene capillary tubes and the little height (1 cm) at which these capillary tubes were held above the soil surface have reasonably satisfied theory assumption number 6.

Flooding Apparatus

The flooding apparatus was designed to make the collection of accurate infiltration data as a function of time possible throughout the experiment. A burette was used for irrigating the soil column from above. The burette was closed at the top, and air was admitted below the surface of the water by the bubbler (Mariotte) tube. This arrangement made it possible to maintain the surface of the irrigation water at a constant level in the cylinder.



○ → hole has diameter 0.610 mm

Figure 8. Top view of the lucite lid for the water applicator.

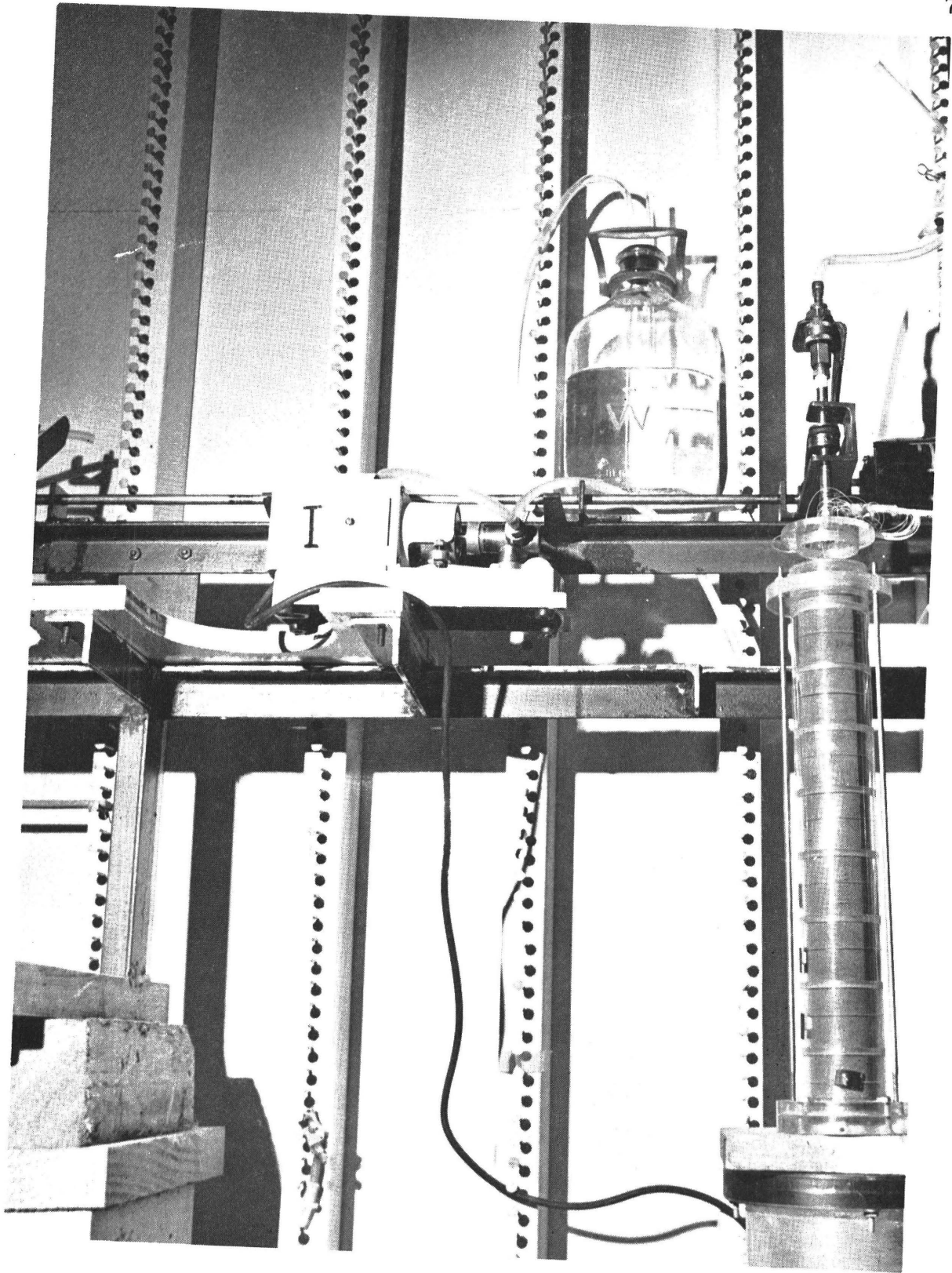


Figure 9. A complete soil filed cylinder in position under a rotating water application lid. I is the LAB pump (for controlling water application rate).

The flooding apparatus was provided by a valve which controlled the depth of water over the soil surface from the first moment of beginning the experiment and kept it constant until the end. The valve, shown in Figure 10, mainly consists of two lucite cylinders, one fits into the other. The holes in the inner cylinder are concomitant with the holes in the outer cylinder. By moving the inner cylinder around, one can match both the holes letting the water run through and can dismatch the holes holding the water above. The depth of water over the soil surface was held at approximately one centimeter. A complete soil-filled cylinder in position under a constant head apparatus is shown in Figure 11.

The Soil Moisturizing Technique

Many determinations in infiltration, chemistry, and microbiology are subject to unknown factors resulting from variation in moisture distribution. Adding water to the soil in alternate increments is unsatisfactory because of the moisture gradient which cannot be avoided unless considerable time is allowed for the water to become uniformly distributed.

The difficulty of making up a soil to a desired initial moisture content has been recognized by many investigators. The method presented here, which was introduced by Shaw (1948), can produce a uniform moisture distribution. It involves the addition of water as ice scorings or snow to the dry soil at -10°C . The desired amount of soil of known initial moisture content and uniformly mixed is passed through a 2-mm screen. The soil is then taken to a deep-freeze room and spread out on a paper so that it will cool rapidly. While the soil is cooling, an amount of

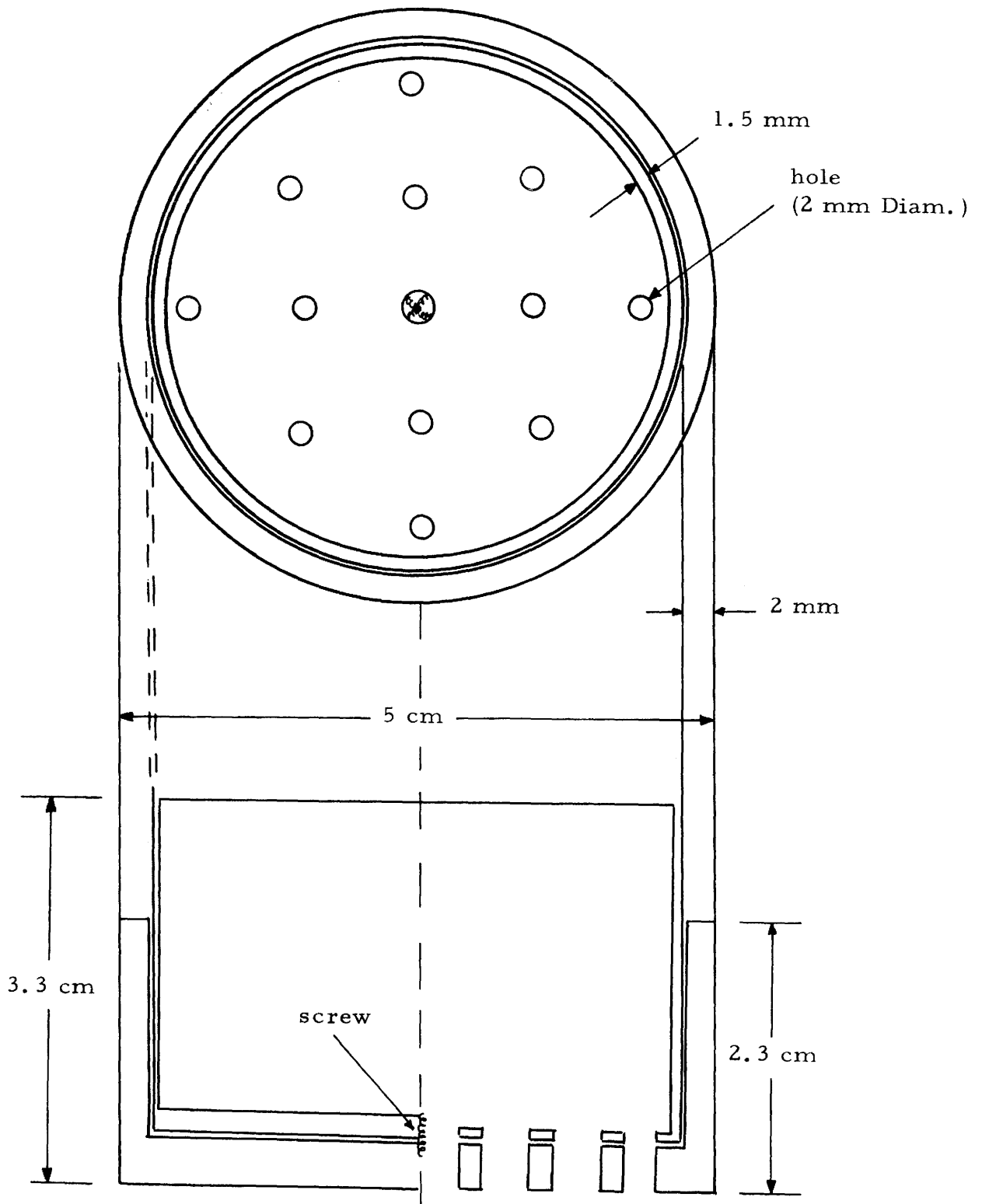


Figure 10. A diagram showing a valve for controlling water supply and how its holes could be matched.

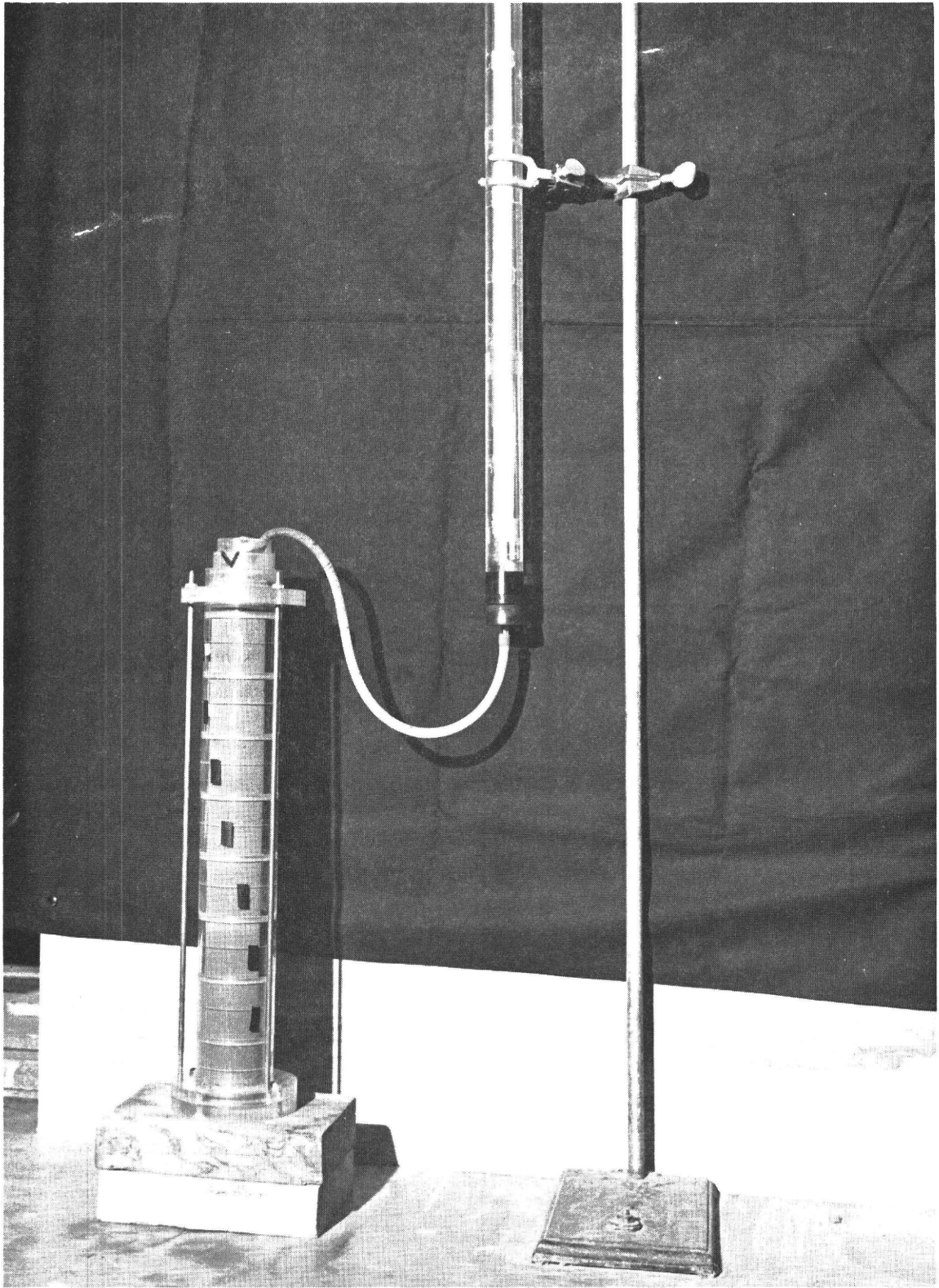


Figure 11. A complete soil filled cylinder in position under a constant head apparatus. V is the flooding valve.

snow or ice from ice-scoring operations equivalent to the desired moisture content is passed through a 2-mm screen. The snow-soil mixture is then carefully rolled until it is as uniform as possible, then left a few hours in the deep-freeze room. The sample is transferred to a tight container and returned to the laboratory and kept overnight to warm to room temperature. The sample is mixed thoroughly again and the moisture content is determined to check the desired one.

The soil has been successfully prepared in a stable, unpuddled condition from air-dry to field capacity moisture levels. The following table shows the moisture content of ten 15-25 gm. samples taken at random from a 5-10 Kgm. lot in which the weight of snow added was chosen to bring the samples to moisture contents of 6.5 and 18 percent on a dry weight basis.

Table 2 shows that the maximum spread is equal to 0.20 percent. These results indicate an unusual degree of uniformity of moisture distribution in view of the evident differences in particle sizes that

Table 2. Moisture content distribution by using snow moisturizing technique.

Sample No.	% Moisture	% Moisture
1	6.40	18.02
2	6.38	18.00
3	6.57	17.97
4	6.57	17.99
5	6.44	17.88
6	6.43	17.82
7	6.36	17.92
8	6.45	17.98
9	6.58	18.01
10	6.38	17.85
σ_x	\pm	\pm

exist in a soil sample screened through a 2-mm sieve. Thus theory assumption 3 is very satisfied.

Experimental Procedure

No soil is homogeneous in the sense that all pores are the same size. The narrower the distribution of pore sizes, the more nearly homogeneous in the microscopic sense is the soil. In a macroscopic sense, however, a soil may be homogeneous, that is, each given element of volume of soil, if chosen large enough, will be identical with every other element of volume in that pore size distributions of these elements will be the same. The homogeneity, in this treatment, was assumed in this macroscopic sense.

The soil was passed through a 2-mm screen, brought up to the desired initial moisture content by the snow moisturizing technique and then packed into the lucite cylinders to a bulk density 1.25 gm/cm^3 .

The packing of the columns proved to be a critical procedure, for when columns were not homogeneous, reproducible results could not be achieved. The method adopted was to pack each column in 5-cm. increments, weighing out enough soil to give a bulk density of 1.25 gm/cm^3 ; thus the homogeneity assumption was reasonably satisfied.

De-aerated Logan City tap water was applied to the soil surface in all the experimental runs. The electrical conductivity of the water, which was always at room temperature when applied, was 0.35 millimhos per centimeter and was nearly identical with the irrigation waters that had been used on this soil in the field. The experiment was conducted in a laboratory where the temperature was $23.0 \pm 1.0^\circ\text{C}$. The temperature

variation during any one experimental run was less than $\pm 0.5^{\circ}\text{C}$. Four levels of initial moisture content, drier than field capacity, of 0.065, 0.10, 0.15, and 0.18 on weight basis were used. Five levels of sprinkler intensities of 1.75, 2.30, 3.15, 4.15, and 5.20 each ± 0.05 centimeters per hour were applied for the four levels of initial moisture contents. All the twenty possible combinations of the five sprinkler intensities and the four initial moisture contents were conducted. The advance of wetting front was recorded with time until the occurrence of surface ponding (appearance of small water pits on the glistening soil surface), then the experiment was cut off. The moisture profile was determined, by sampling at each 2.5 cm increment of depth, and oven-drying at a temperature of 105°C .

The four levels of initial moisture content were also used for flooded infiltration (Figure 11). The cumulative infiltration was recorded as a function of time until the wetting front, its advance also recorded, reached 40 cm depth for all the flooded experimental runs. Then the experiment was terminated and the moisture profile was determined by sampling at each 2.5 cm increment of depth.

At the time when the experiment was cut off in any experimental run, there was at least 10 cm soil-depth above the bottom of the column at its initial moisture content, thus the assumption of a semi-infinite profile was satisfied. Each experiment was repeated four times under flooding, and at least twice under sprinkling.

Whenever long soil columns (65 cm) with initial moisture contents of 0.10, 0.15, and 0.18 on weight basis were used, different application rates for obtaining the capillary conductivity-moisture relationship for each initial moisture content were also used. The application rates used were

less than the value of the saturated hydraulic conductivity, 2.0 cm/hr, to avoid the occurrence of surface ponding. The experiment was allowed to continue for long times, especially with extremely low application rates, until the wetting front reached 50 cm depth; thus approximately a constant-shaped moisture profile was developed which was determined by sampling at each 2.5 cm increment depth. The method used for calculating the capillary conductivity as a function of moisture content was suggested by Youngs (1964) and Childs (1969) as follows. A unit gradient was assumed to exist in the approximately constant shaped moisture profile, and the value of the constant moisture was taken as the average of moisture contents at 10, 15, 20, 25, 30, 35, 40 centimeters depth. This average moisture content had a capillary conductivity equal to the water application rate applied on the surface.

Short soil columns (25 cm) with the four levels of initial moisture content were used also under different application rates for determining the moisture characteristic curve of the soil which represents the tension-moisture relationship. To achieve this purpose, the experiment was cut off when the wetting front was in the vicinity of the bottom of the column, then the column was covered with plastic sheets to prevent evaporation, and the tensiometers were inserted in each section for measuring the tension of the soil. The soil column was left until the equilibrium obtained indicated by no change in tensiometer readings. Usually it took from 12-24 hours to reach equilibrium condition, then the moisture was determined by sampling.

The mercury manometers used with the tensiometers were extremely narrow (1.0 mm inside diameter) to give quick response and accurate

reading. Figure 12 shows the manometers connected to the tensiometers which were inserted in the holes along the column.

Although the moisture characteristic curve obtained above was the absorption curve needed in calculating the infiltration rates, an experiment was conducted to obtain the desorption curve by using a pressure-cooker and a pressure plate apparatus (used by Richards, 1947) to get an idea about the hysteresis in the soil.

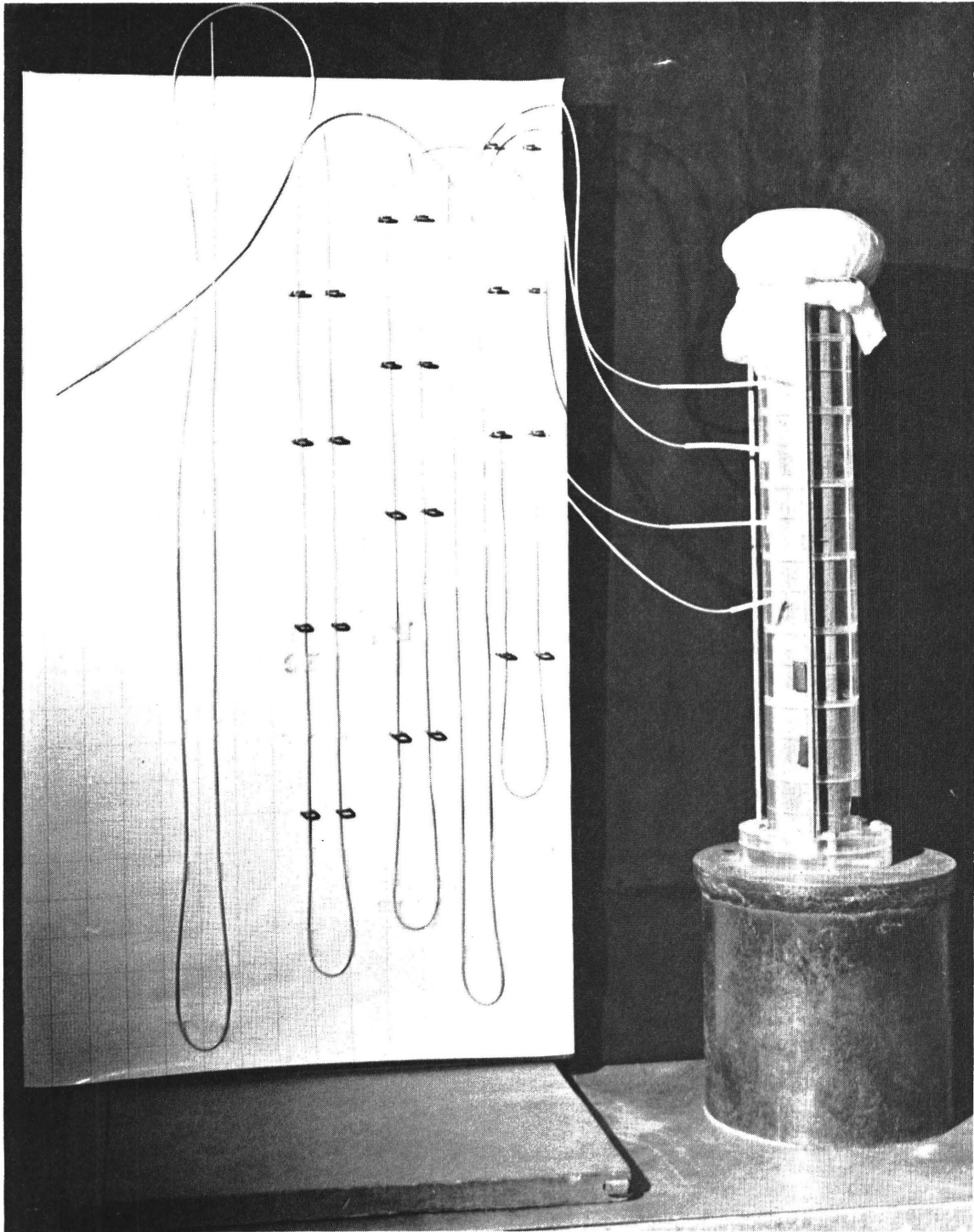


Figure 12. The mercury manometers are shown connected to the tensiometers which were inserted in the holes along the column.

CHAPTER 5
RESULTS AND DISCUSSION

Flooded (Ponded) Infiltration

Measurements of infiltration were made using homogeneous soil columns with initial moisture contents of 6.5, 10, 15, and 18 percent on a weight basis and a bulk density of 1.25 gm/cm^3 . The measurements were continued until the same depth of wetting front (40 cm) was achieved in each case under the same constant head on the soil surface, of approximately one centimeter.

Figure 13 is a log-log plot of the cumulative infiltration, Y , as a function of time, t , at the four initial moisture levels. Each cumulative infiltration curve represents a linear regression line for four replicates. (See Appendix C.) Figure 13 also shows that all the cumulative infiltration curves have almost the same slope, a , approximately equal to 0.60.

By differentiating the cumulative infiltration, Y , with respect to time, t , the intake rate (instantaneous infiltration rate), I , can be obtained which is presented on a log-log plot by Figure 14 which in turn shows that all the intake rate curves have the same slope, n , approximately equal to -0.40.

Figure 15 presents the infiltration rate as a function of time on a rectangular plot. Figure 15 also shows that all the intake rate curves are approaching, after a long time, the same constant infiltration rate (base intake rate), and the narrower the difference in initial moisture content between any two curves, the sooner the approachment of the base infiltration.

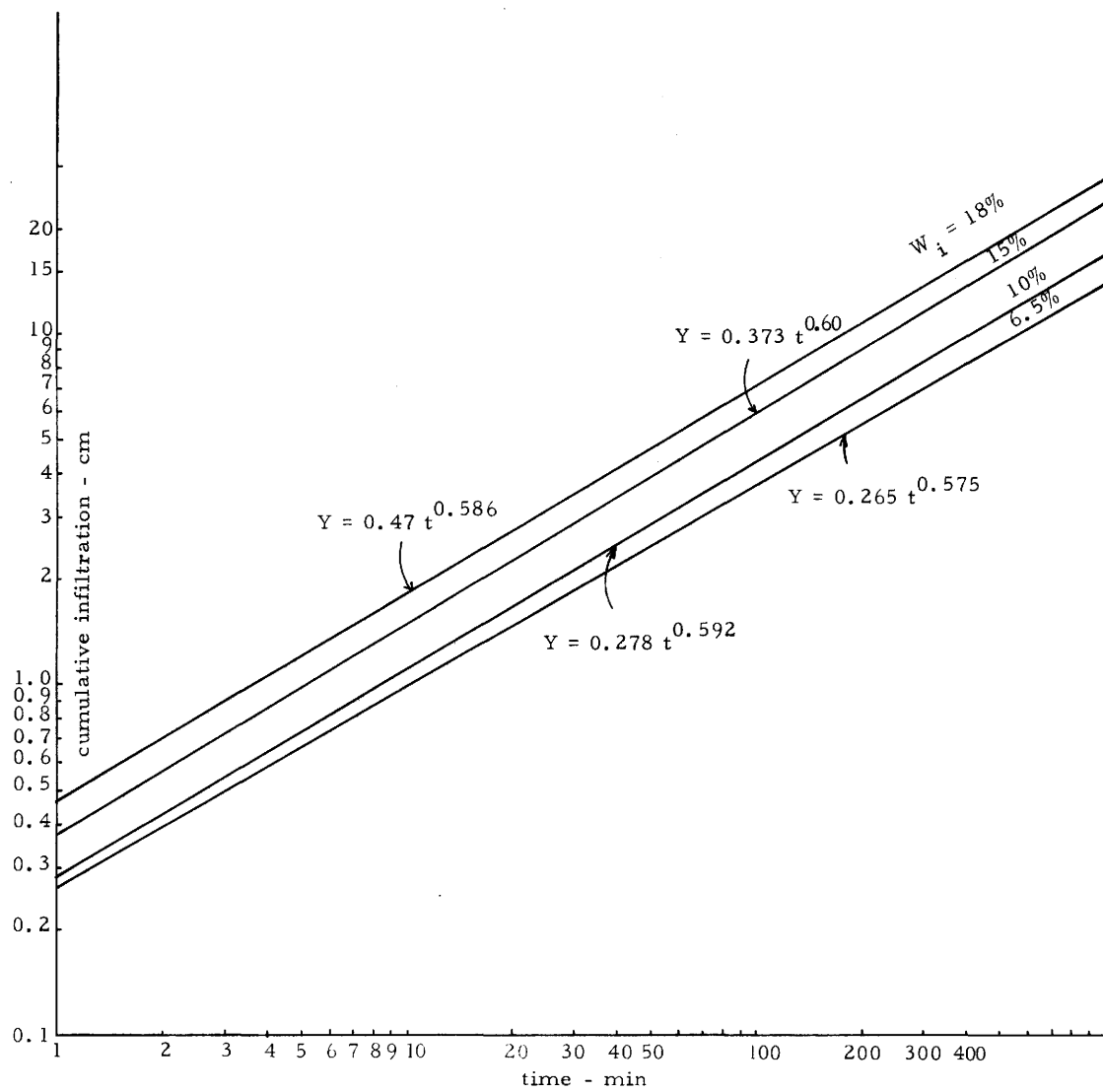


Figure 13. Measured cumulative infiltration at different initial moisture levels. Each curve represents a linear regression line for four replicates for Nibley silty clay loam soil.

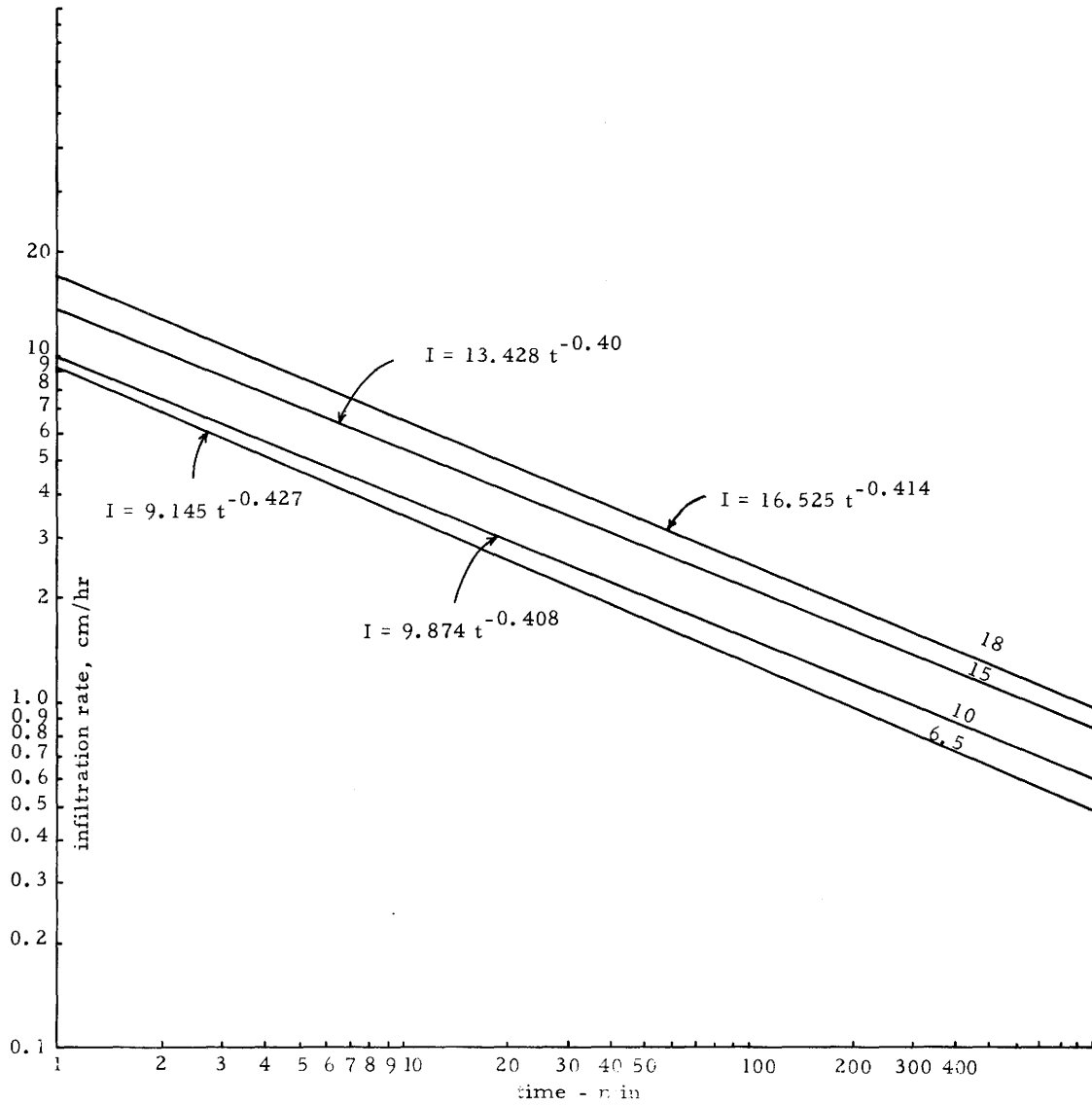


Figure 14. Instantaneous infiltration rate at different initial moisture levels. The numbers labeling the curves indicate the initial moisture content, W_i , as percent for Nibley silty clay loam soil.

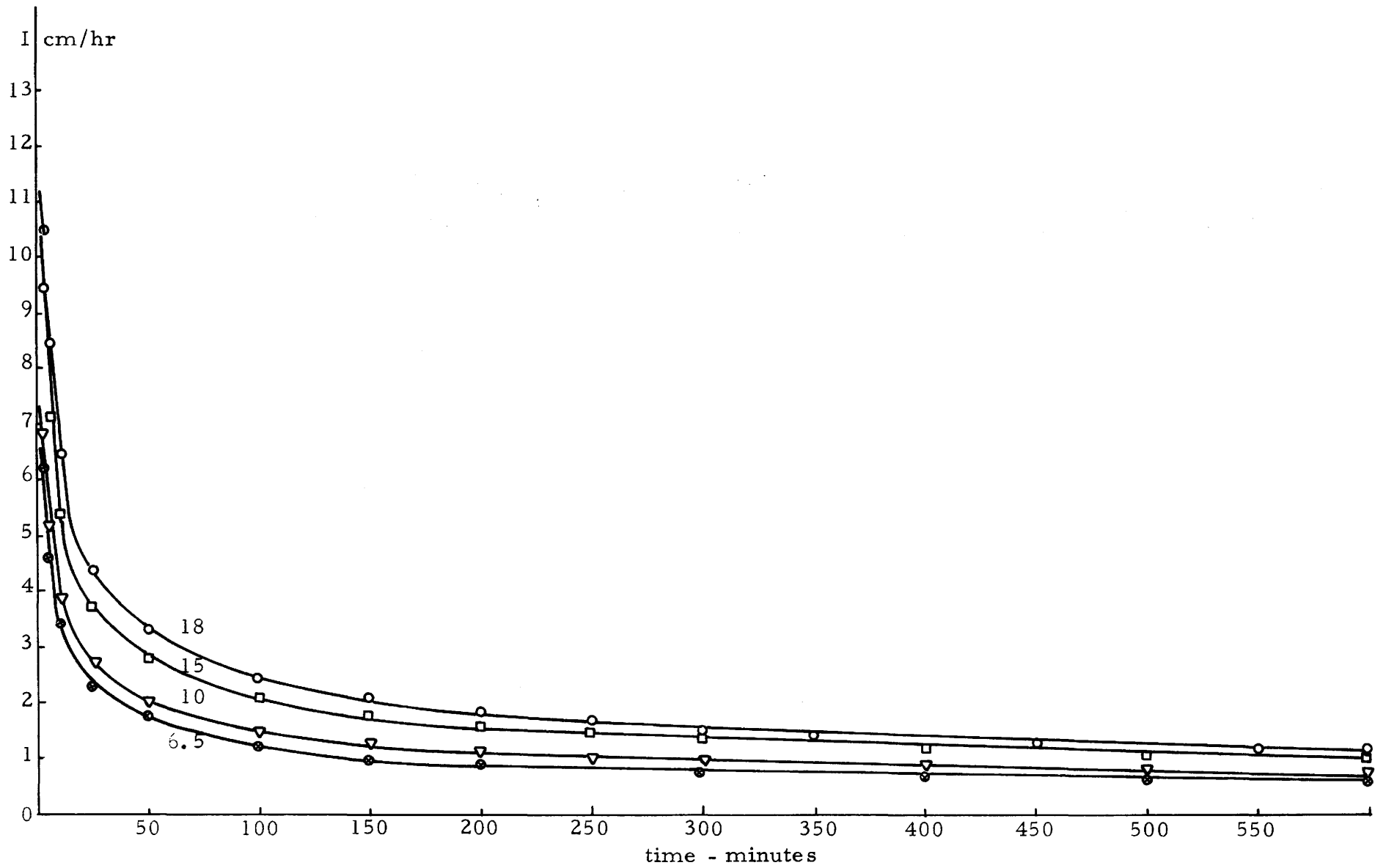


Figure 15. Influence of initial moisture content, W_1 , on infiltration rate, I . The numbers labeling the curves indicate the percent of the initial moisture content, W_1 , for Nibley silty clay loam soil.

It can be noted from Figures 14 and 15 that the infiltration rate increases with increasing the initial moisture content in this dry range. This result, of course, is opposite to the general statement (rate of entry in moist soils is less than in drier soils) introduced by many authors. But careful examination of the previous work on the effect of initial moisture content on infiltration rate by many authors (Van Duin, 1955; Philip, 1957e; Hanks, 1965; Skaggs et al., 1969, and others), one can see that their comparison based on a wide range of initial moisture such as dry, moist, and wet or dry and wet which are two extreme conditions.

However, Fletcher (1960) presented a figure to show the indirect effect of soil moisture on infiltration in two granitic upland soils in southern Arizona. Fletcher's figure indicated higher mass infiltration for the moist soil (initially received several rains) than the dry soil (initially received no moisture for several months). Fletcher stated that the moist soil gave the microflora a chance to grow giving better aggregation.

In this investigation only dry range of initial moisture is dealt with, and this point should be emphasized. The effect of four initial moisture contents are being studied, all of which are in the dry range (initially drier than field capacity of the soil). One can observe the remarkable resistance against wetting applied by the dry soil particles, which reduce the entry of water, since the angle of contact of the menisci at the wetting front approaches 90° (observed by the author and Van Duin, 1955). However, careful examination of Figure 14 indicates that the intake rate curve at initial moisture content, W_i , 6.5 percent will intersect the intake rate curve at W_i equal to 10 percent, at the early

few seconds of the beginning of infiltration. Accordingly, it may be expected that the intake rate at zero percent of initial moisture will intersect the 6.5 percent initial moisture curve and probably the 10 or 15 percent initial moisture curve, implying that the intake rate will be higher at the early time of infiltration (due to the existence of high hydraulic gradient caused by this very dry initial moisture) then decline at a latter time.

Another point to be emphasized here, that the bulk density, 1.25 gm/cm^3 , was the same in each initial moisture run and remained almost constant during the experiment since the soil (Nibley silty clay loam) was a well aggregated stable soil which did not show any settlement during watering.

A numerical solution for solving the non-linear partial differential equation (Equation 7) was used to obtain the theoretical cumulative infiltration from the physical properties of the soil. The Crank-Nicholson implicit technique, adapted to include gravity, was used in the numerical solution.

Figure 16 shows the capillary potential (suction)-moisture content relationship during the adsorption cycle since the infiltration is an adsorption process. Figure 17 shows the capillary conductivity-moisture content relationship for each initial moisture content run. Figure 17 indicates that the soil does not possess a unique relation between capillary conductivity, K , and moisture content, W , and the clay minerals in the soil reacting differently for each initial moisture content. Of course, one smooth curve may be drawn through the scattered points which has been done by many investigators (Bruce and Klute, 1956; Flocker et al., 1968; and others), but the theoretical calculation will not be able to distinguish

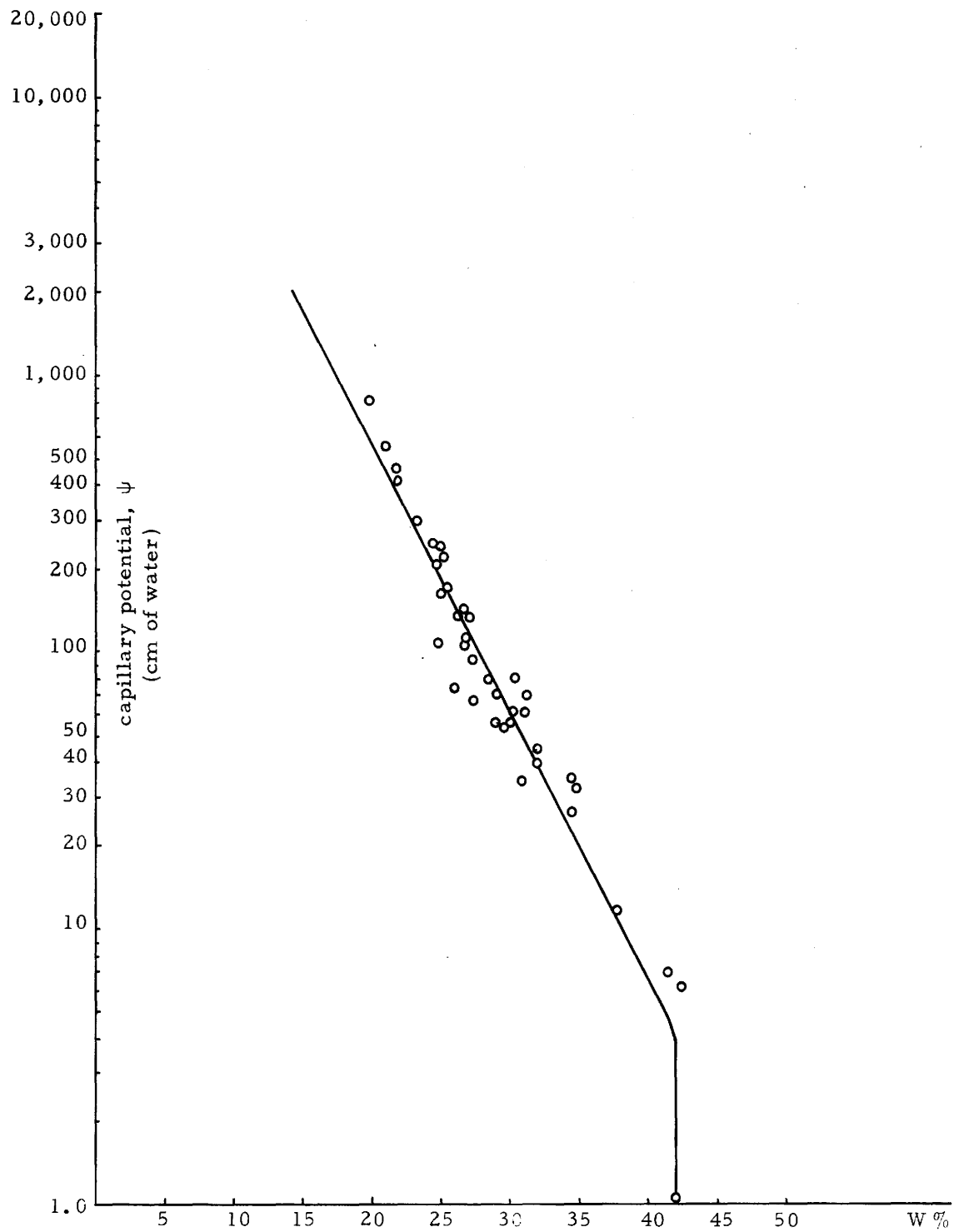


Figure 16. Capillary potential (suction) versus moisture content for Nibley silty clay loam soil.

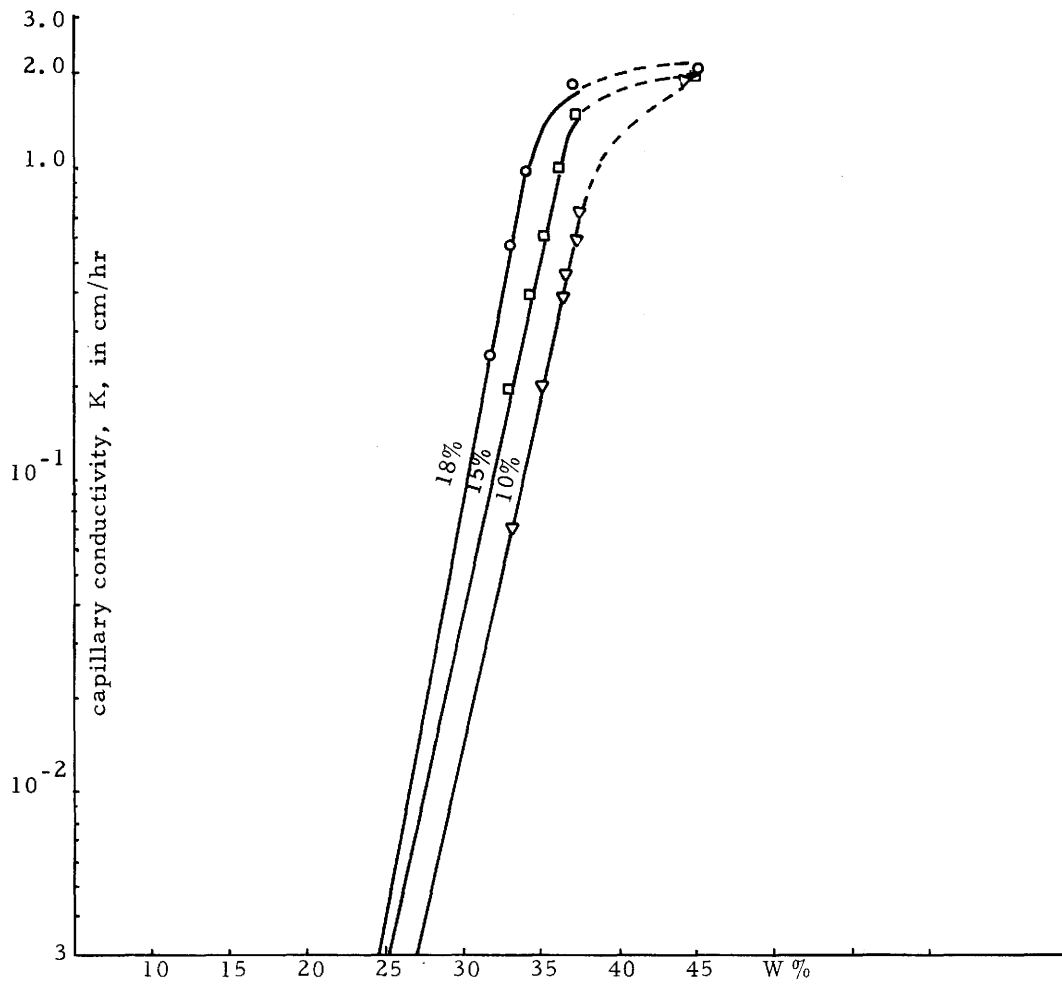


Figure 17. Capillary conductivity versus moisture content. The numbers labeling the curves indicate the initial moisture content, W_i , used in the run for Nibley silty clay loam soil.

between the effect of initial moisture content on infiltration rate within a narrow range, especially the dry range. Hanks (1965) applied the mentioned numerical technique using one smooth curve for capillary conductivity versus moisture content and computed the cumulative infiltration in Sorpy loam soil; his computation distinguished the effect of dry and wet soil but failed to distinguish the effect of moist from wet soil on the cumulative infiltration.

In view of the large possible errors in the calculation of diffusivity $\left(K \frac{\partial \Psi}{\partial \theta}\right)$, Bruce and Klute (1956) concluded that it is not possible to say, for example, whether or not the three curves calculated for a soil at different initial moisture content are different.

Although Figure 16 is the adsorption curve needed in the infiltration calculation, the writer presents Figure 18 to show how much hysteresis this soil has.

Figures 16 and 17, with the initial conditions of the soil were introduced as input data in the computer program (Appendix C), to calculate the cumulative infiltration. Figure 19 presents a comparison of calculated and experiment-measured cumulative infiltration at three different initial moisture levels (drier than field capacity). Examination of Figure 19 reveals the excellent agreement between the calculated and measured values, due to the satisfaction of the theory's assumptions.

The computational method used permits derivation of other information in addition to cumulative infiltration. Since the method requires that Ψ and W as a function of depth, Z , and time, t , be calculated in order to compute cumulative infiltration, the program's output can supply the cumulative advance of wetting front and the moisture content as a function of depth. Figure 20 shows the excellent agreement between the calculated

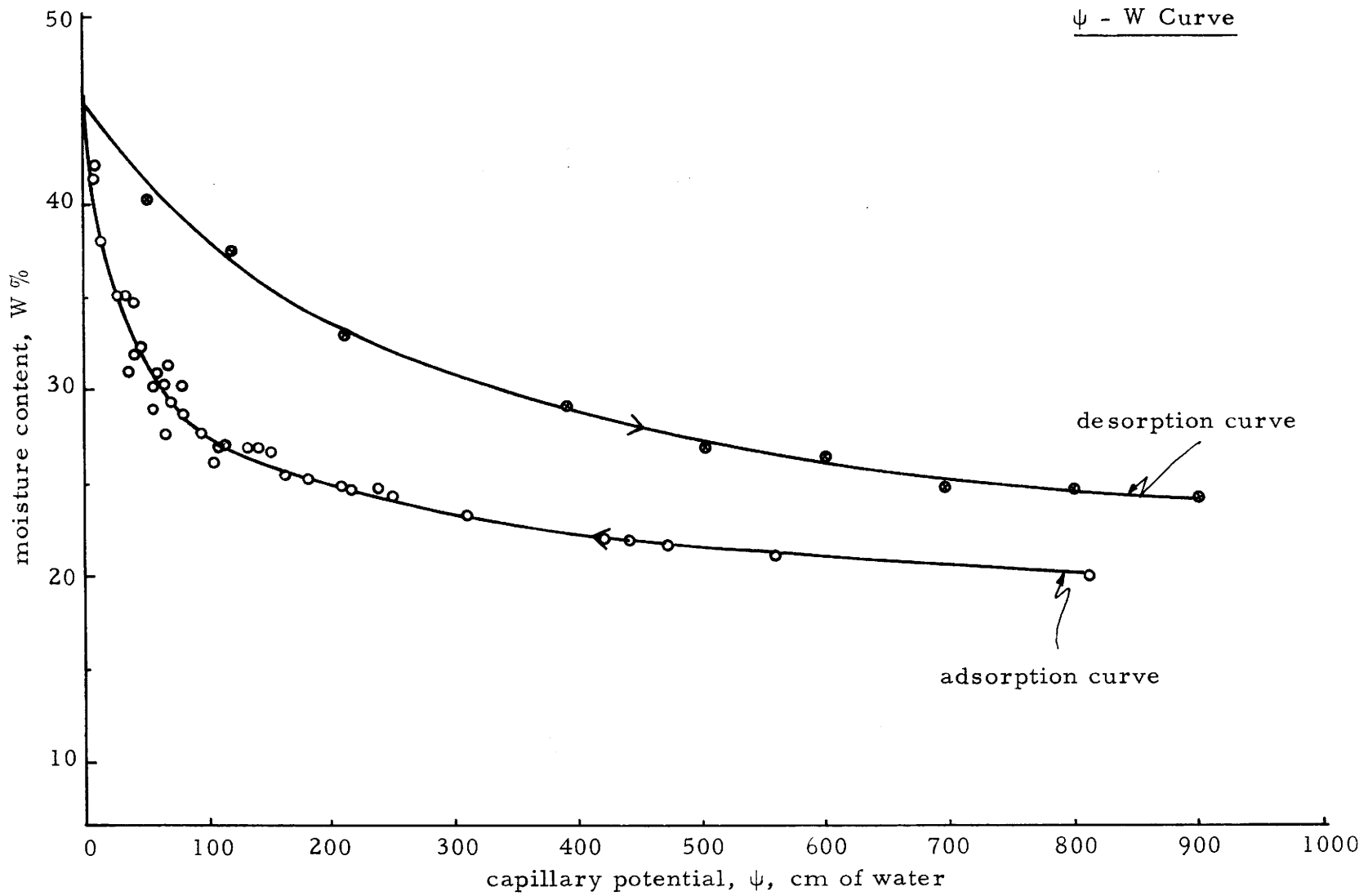


Figure 18. Soil moisture characteristics showing hysteresis. The arrows indicate direction of water content for Nibley silty clay loam soil.

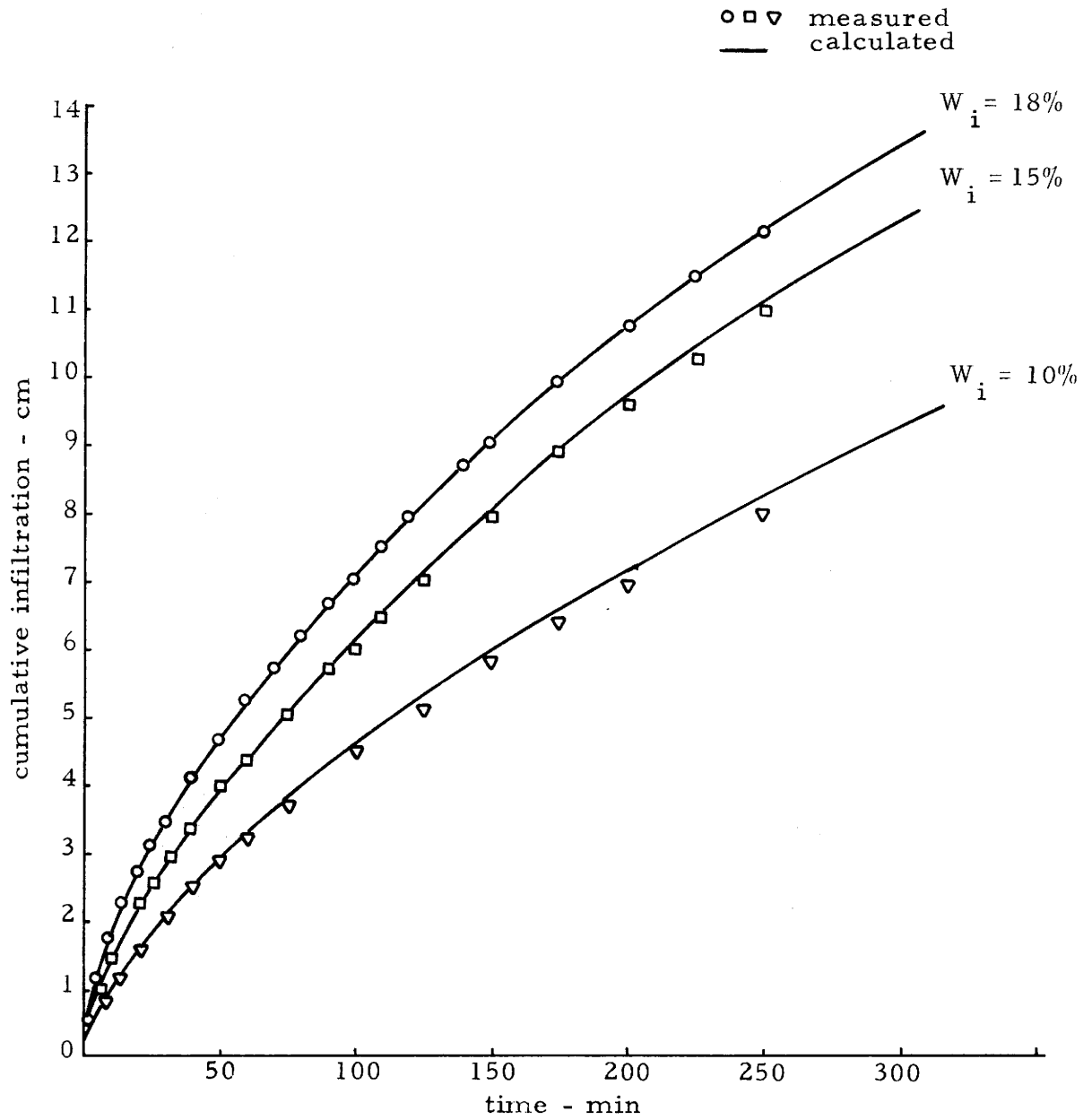


Figure 19. Influence of initial moisture content, W_i , on cumulative infiltration on soil. Comparison of measured and theoretical values.

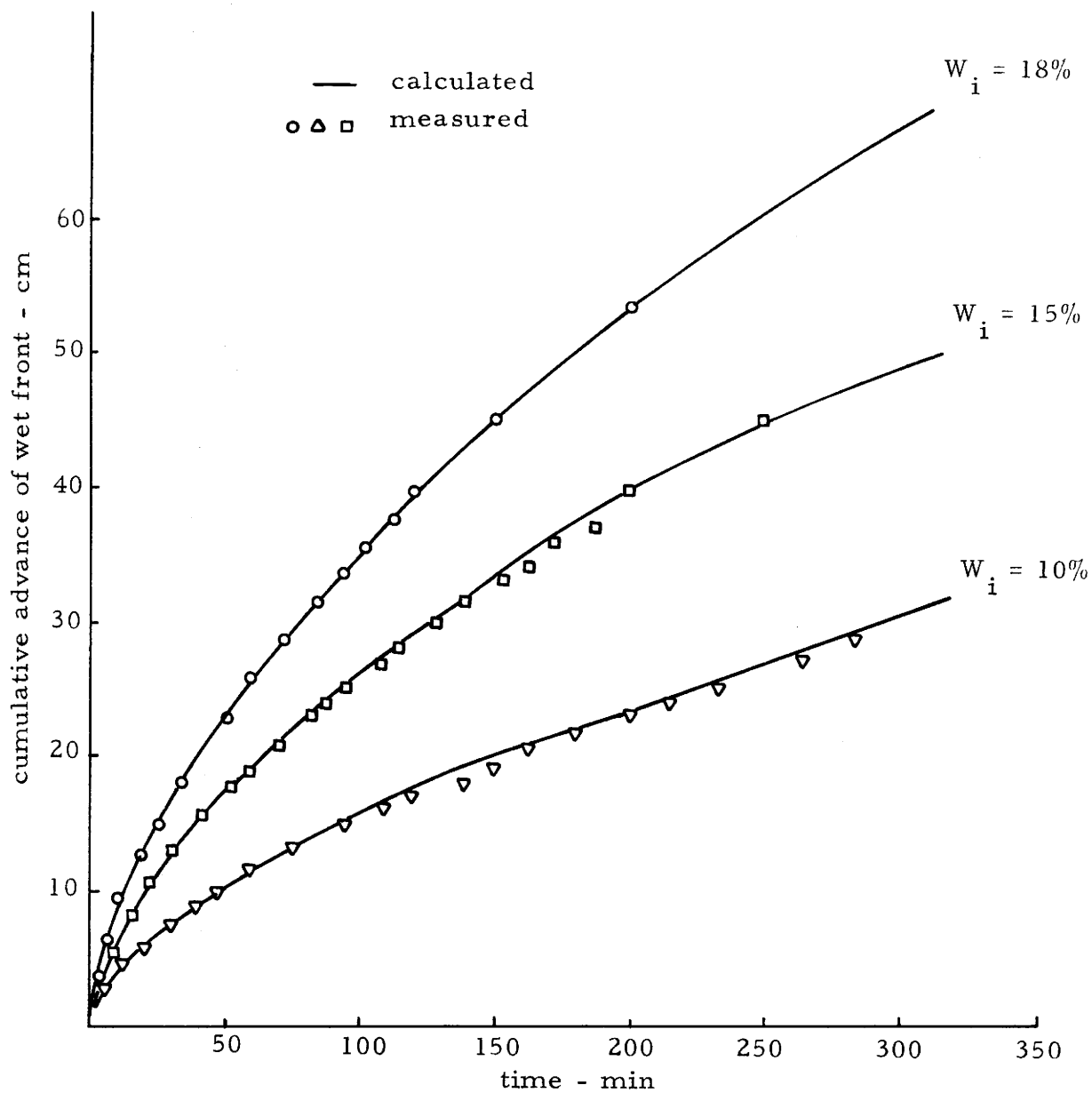


Figure 20. Influence of initial moisture content, W_i , on wetting front advance during flooded infiltration on Nibley silty clay loam soil. Comparison of measured and theoretical values.

and measured cumulative advance of wetting front at three different initial moisture levels (10, 15, and 18 percent). Figure 20 shows that the rate of penetration of the wetting front increases with increasing moisture content in the dry range.

Figures 21, 22, 23, and 24 present the moisture profiles at the same depth of penetration (40 cm) for the different initial moisture levels. One also can notice the good agreement between the calculated and measured moisture profiles from Figures 22, 23, and 24.

Figure 25 combines the moisture profiles for the different initial moisture contents. One can conclude from Figure 25 that under flooded (ponded) infiltration, no matter what the initial moisture content is, and the same depth of penetration is achieved (after enough time to approach a steady state condition), one obtains the same transmission zone for each initial moisture content drier than the field capacity.

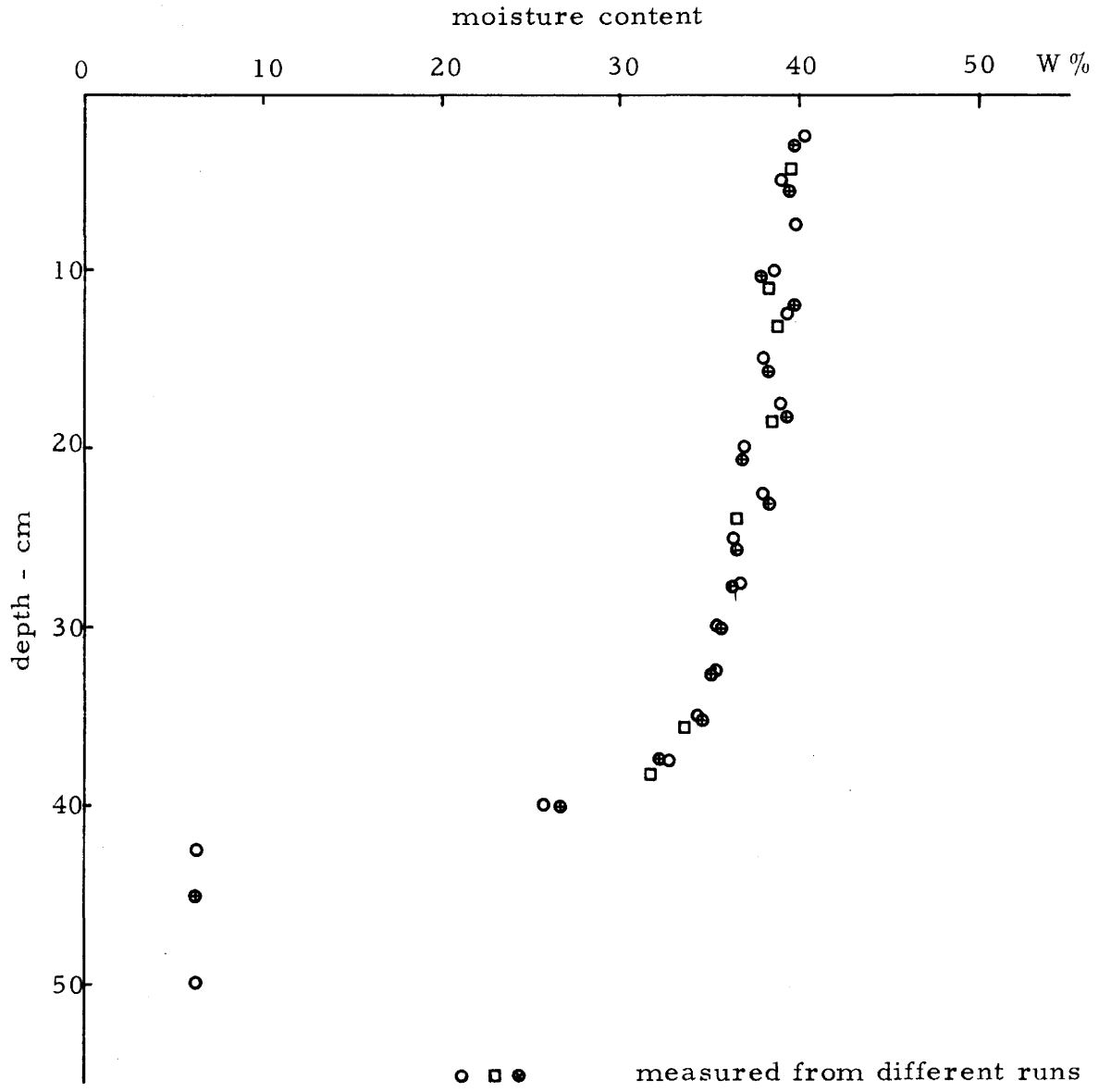
Figure 26 demonstrates the reduction of water entry as the depth of penetration increases.

Rain (Sprinkler) Infiltration

Effect of initial moisture content and sprinkler intensity on the rate of wetting front advance

Sprinkled infiltration experiments were conducted on all the possible combinations of the five constant water application rates of 1.75, 2.30, 3.15, 4.15, and 5.20 centimeters per hour and the four levels of initial moisture contents of 0.065, 0.10, 0.15, and 0.18 on weight basis.

Figures 27, 28, 29, and 30 present the wetting front advance, Z , as a function of time, t , prior to the occurrence of surface ponding



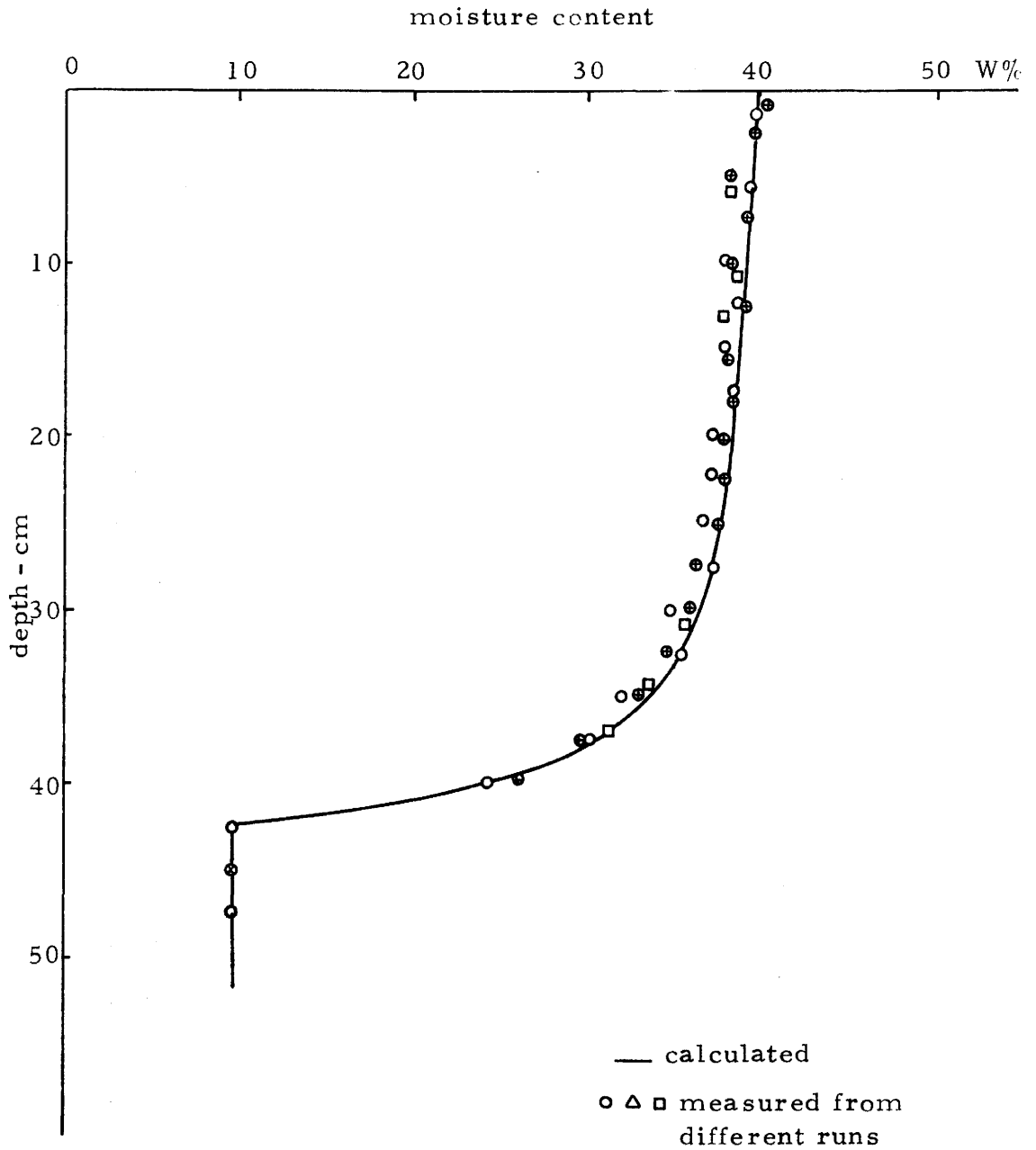
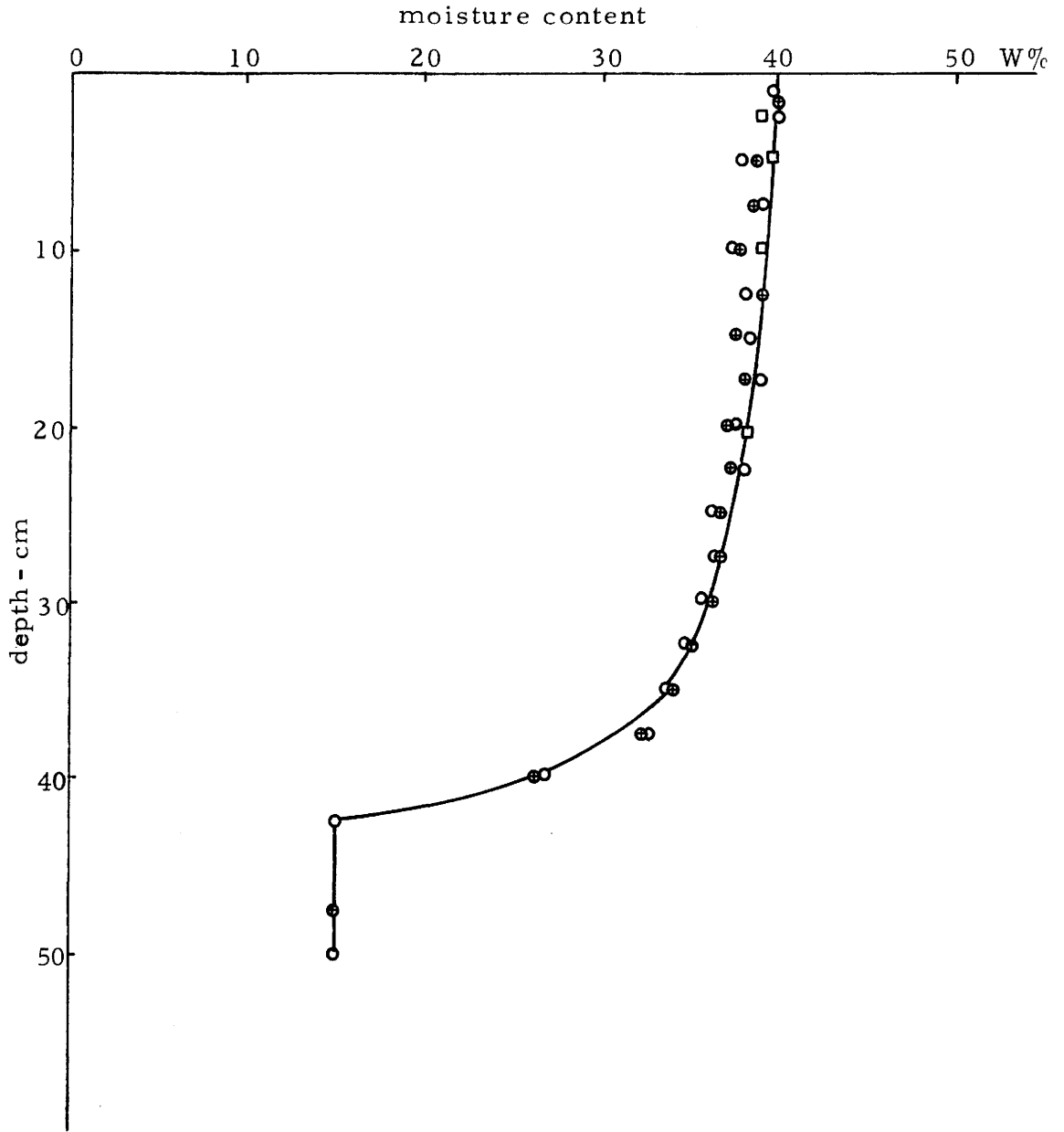
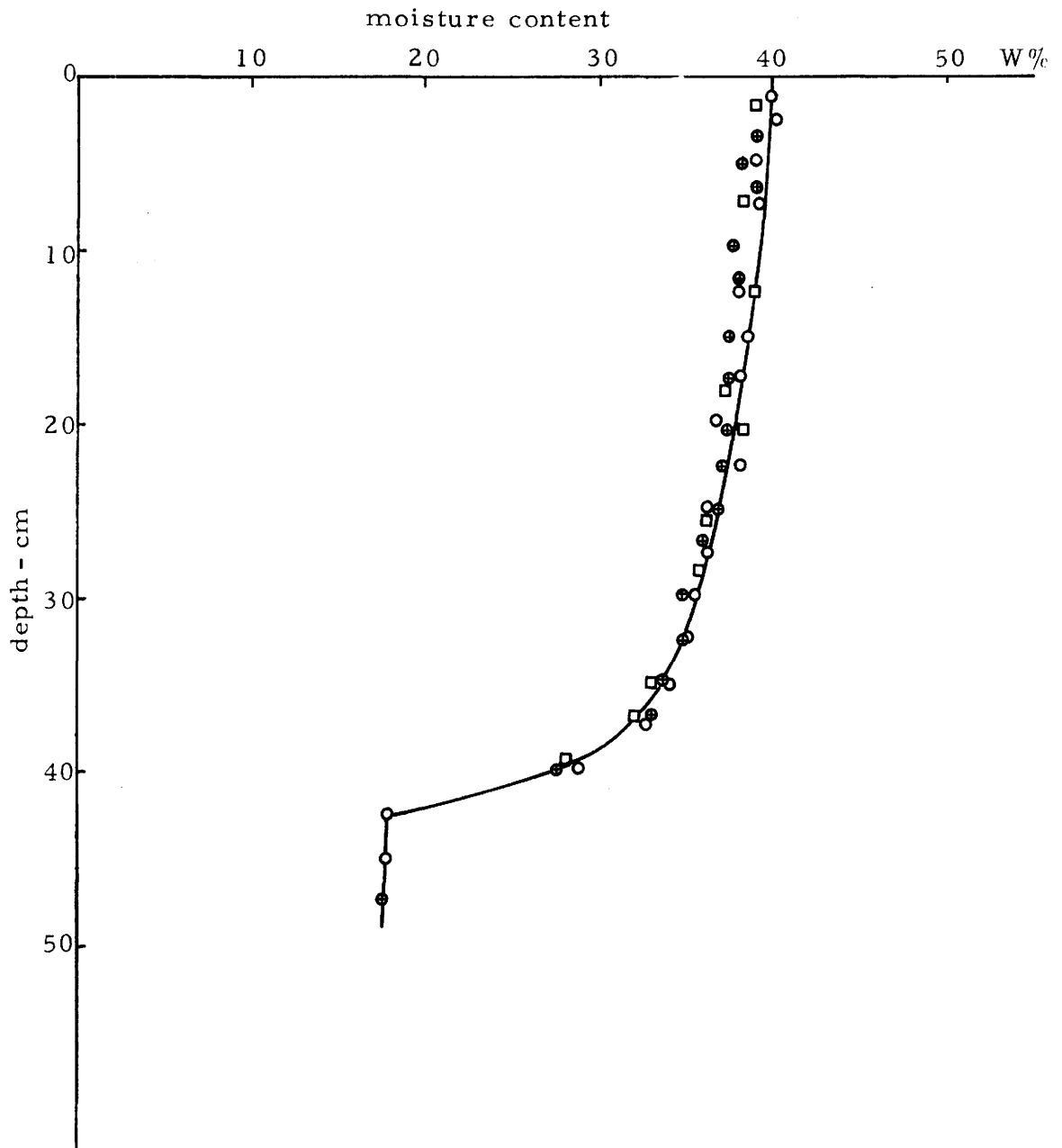


Figure 22. Moisture profile under flooded infiltration for $W_i = 10\%$ for Nibley silty clay loam soil. Comparing measured and calculated values.



— calculated
 ○ ○ □ measured from different runs

Figure 23. Moisture profile under flooded infiltration for $W_i = 15\%$ for Nibley silty clay loam soil. Comparing measured and calculated values



— calculated

○ ● □ measured from different runs

Figure 24. Moisture profile under flooded infiltration for $W_i = 18\%$ for Nibley silty clay loam soil. Comparing measured and calculated values.

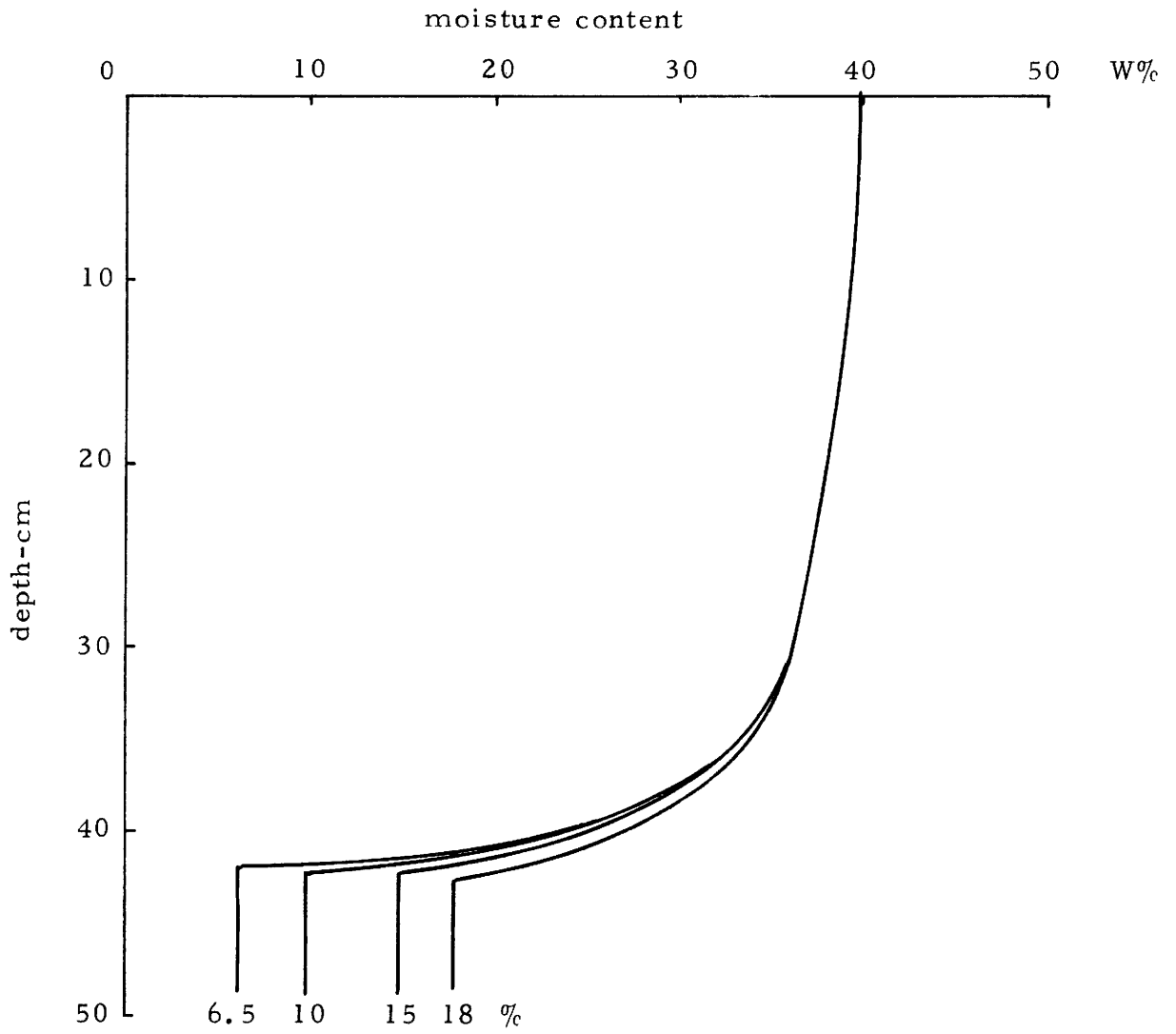


Figure 25. The moisture content profiles at the end of flooded infiltration at the same depth for different initial water contents as labeled.

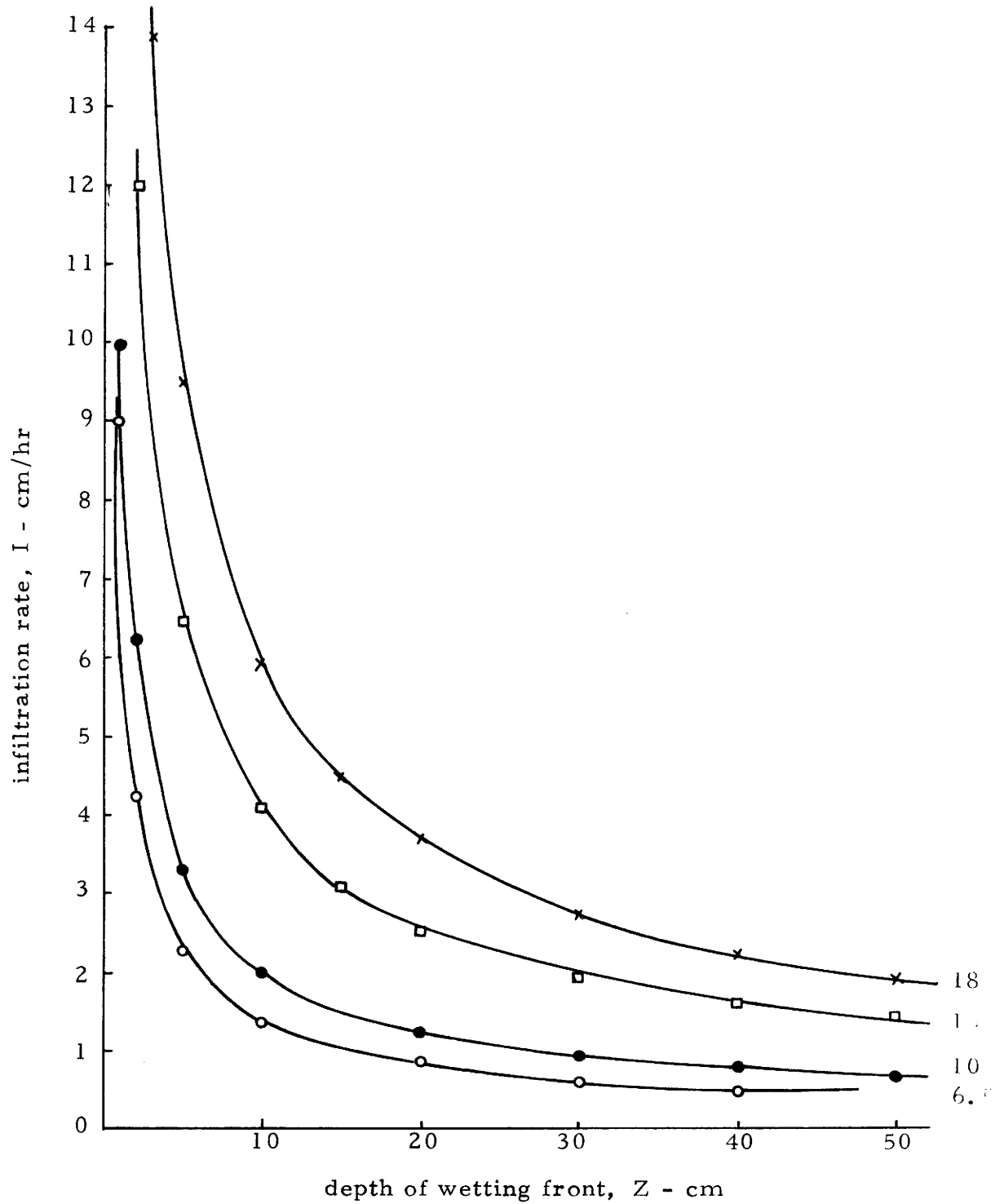


Figure 26. Infiltration rate vs. depth of wetting front. The numbers labeling the curves indicate the percent of initial moisture content, W_i .

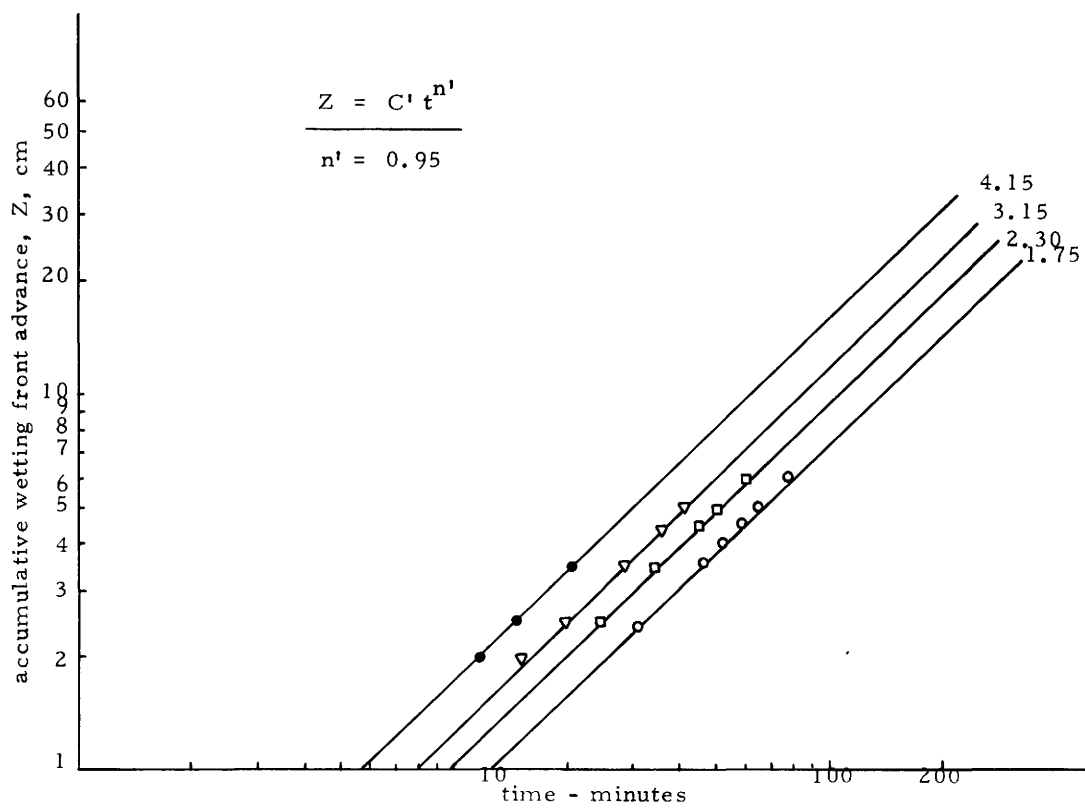


Figure 27. Wetting front advance for different water application rates, $W_i = 6.5\%$. The numbers labeling the curves indicate the rate of application in cm/hr. for Nibley silty clay loam soil. Comparing measured values to theoretically computed values.

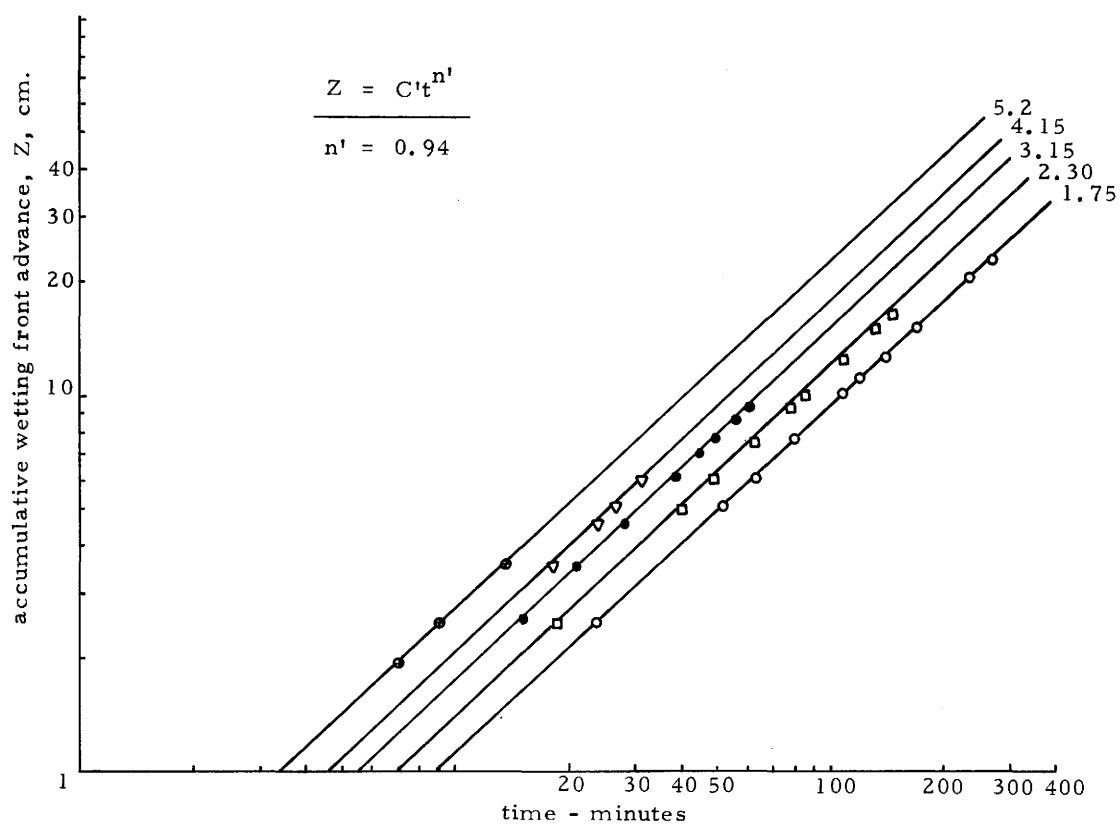


Figure 28. The wetting front advance for different water application rates, $W_i = 10\%$. The numbers labeling the curves indicate the rate of application in cm/hr. for Nibley silty clay loam soil comparing measured and theoretical values.

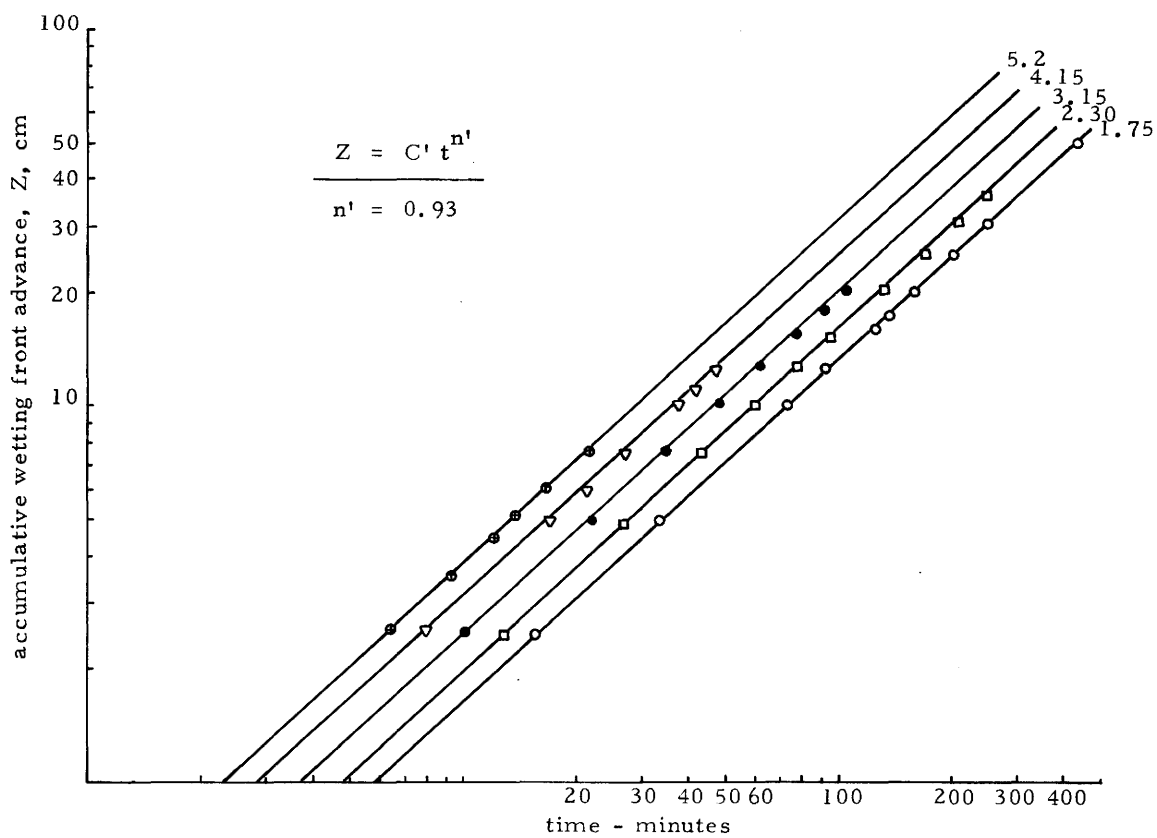


Figure 29. The wetting front advance for different water application rates, $W_i = 15\%$. The numbers labeling the curves indicate the rate of application in cm/hr. for Nibley silty clay loam comparing measured and theoretical values.

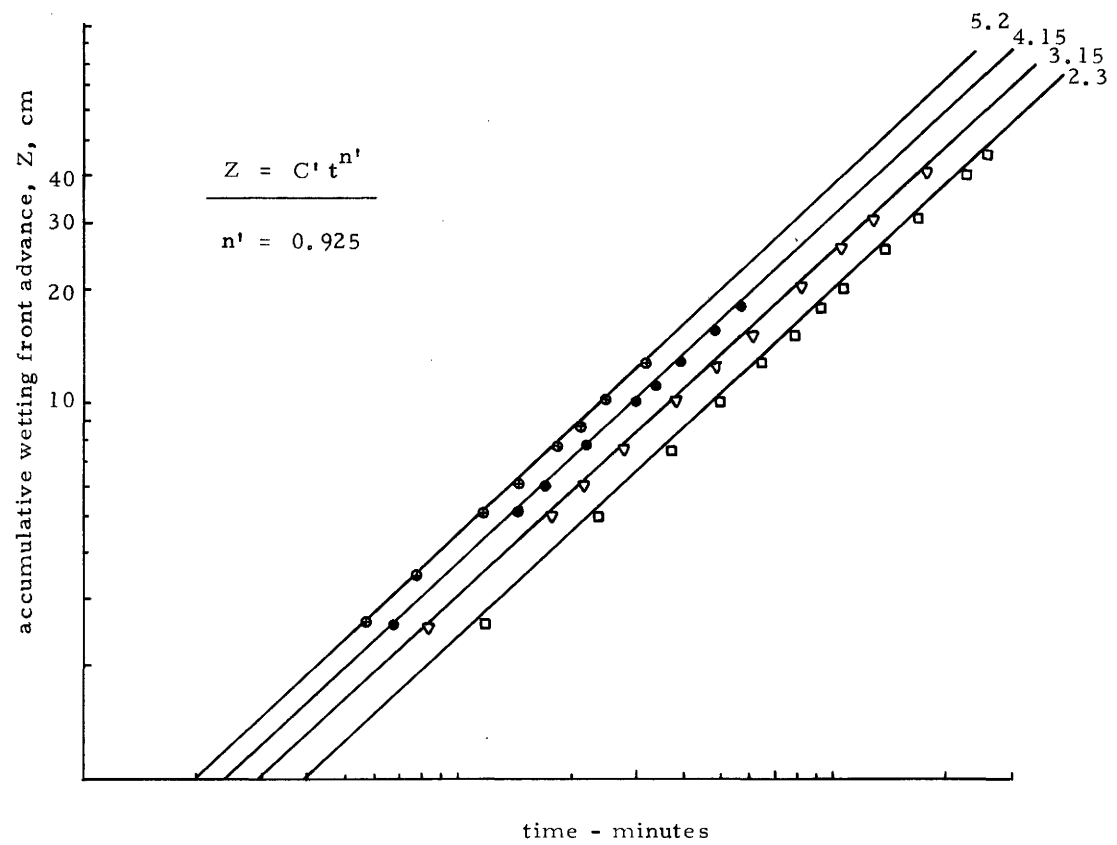


Figure 30. The wetting front advance for different water application rates, $W_i = 18\%$. The numbers labeling the curves indicate the rate of application in cm/hr. for Nibley silty clay loam soil comparing measured and theoretical values.

for different water application rates at different initial moisture content, W_i . One can conclude from these curves that for a constant initial moisture content, the advance of the wetting front increases as the rate of application, R , increases, and for a constant application rate, the advance of wetting front, Z , increases as the initial moisture content, W_i , increases.

Figures 27, 28, 29, and 30 also show how well the data fit the derived exponential Equation (83).

$$Z = C' t^{n'} \dots \dots \dots (83)$$

on log-log plot giving values for the slope, n' , varying between 0.93 to 0.95.

Table 3 gives values of the constant C' , which are dependent on the rate of application, R , and the initial moisture content, W_i , of the soil.

Figure 31 presents the constant C' as a function of R at different initial water contents, W_i . The constant C' in Equation (83), which describes the cumulative advance of wetting front, Z , can be obtained easily under any condition of water application rate, R , and initial moisture content, W_i , by using Equation (84).

$$C' = A'(\theta_i) R^m \dots \dots \dots (84)$$

Table 3. The values of constant, C' , in Equation (83).

R cm/hr	$W_i = 6.5\%$	$W_i = 10\%$	$W_i = 15\%$	$W_i = 18\%$
1.75	0.081	0.134	0.206	-
2.30	0.103	0.165	0.250	0.295
3.15	0.130	0.210	0.316	0.378
4.15	0.170	0.260	0.390	0.445
5.20	-	0.32	0.475	0.524

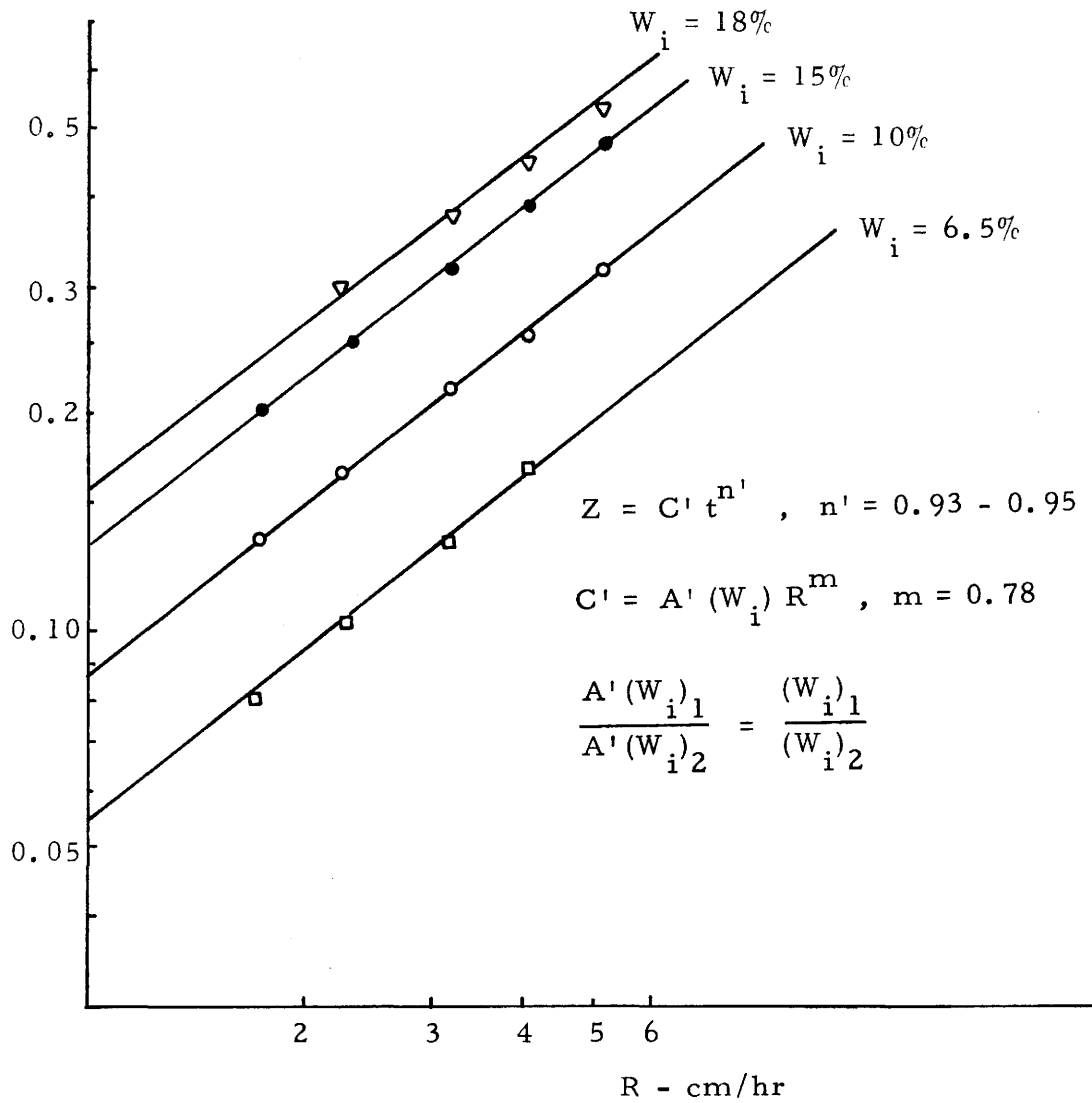


Figure 31. Evaluation of constant, C' , in Equation (83).

or

$$C' = A'(W_i) R^m \dots \dots \dots (84a)$$

and Equation (85)

$$\frac{A'(\theta_i)_1}{A'(\theta_i)_2} = \frac{(\theta_i)_1}{(\theta_i)_2} \dots \dots \dots (85)$$

or

$$\frac{A'(W_i)_1}{A'(W_i)_2} = \frac{(W_i)_1}{(W_i)_2} \dots \dots \dots (85a)$$

The value of the constant $A'(W_i)$ in Equation (84a) can also be obtained from Figure 31 by taking the value of C' at a unit application rate, R .

The following example illustrates how the general derived equation

$$Z = A'(\theta_i) R^m t^{n'} \dots \dots \dots (86)$$

can predict the advance of wetting front as affected by the initial moisture content, W_i , and sprinkler intensity, R .

Example. Suppose one had conducted a quick experiment on a homogeneous soil which has an initial moisture content, W_i , equal to 15 percent, to determine the advance of the wetting front, Z , under three or four different application rates. By the end of his experiment, he would be able to figure out the values of slope n' and m in addition to $A'(15)$. From Figures 29 and 31, the values are 0.93, 0.78, and 0.13 respectively. Now he would like to predict theoretically the corresponding equations for the same soil with 10 percent initial moisture, W_i , under application rates of 2.30 and 4.15 cm/hr.

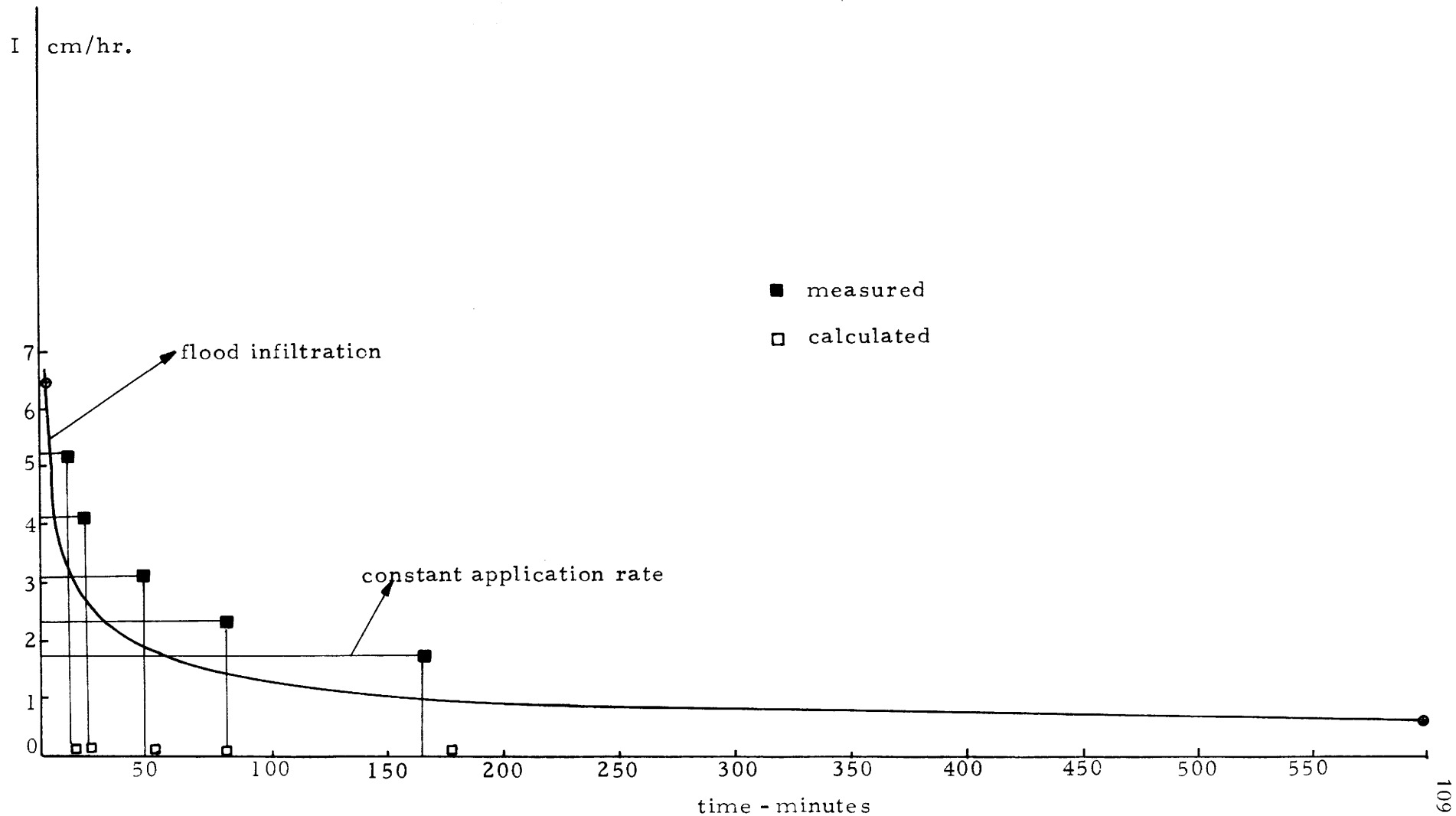


Figure 32. Effect of sprinkler application rate on surface ponding, t_s , for $W_1 = 6.5\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

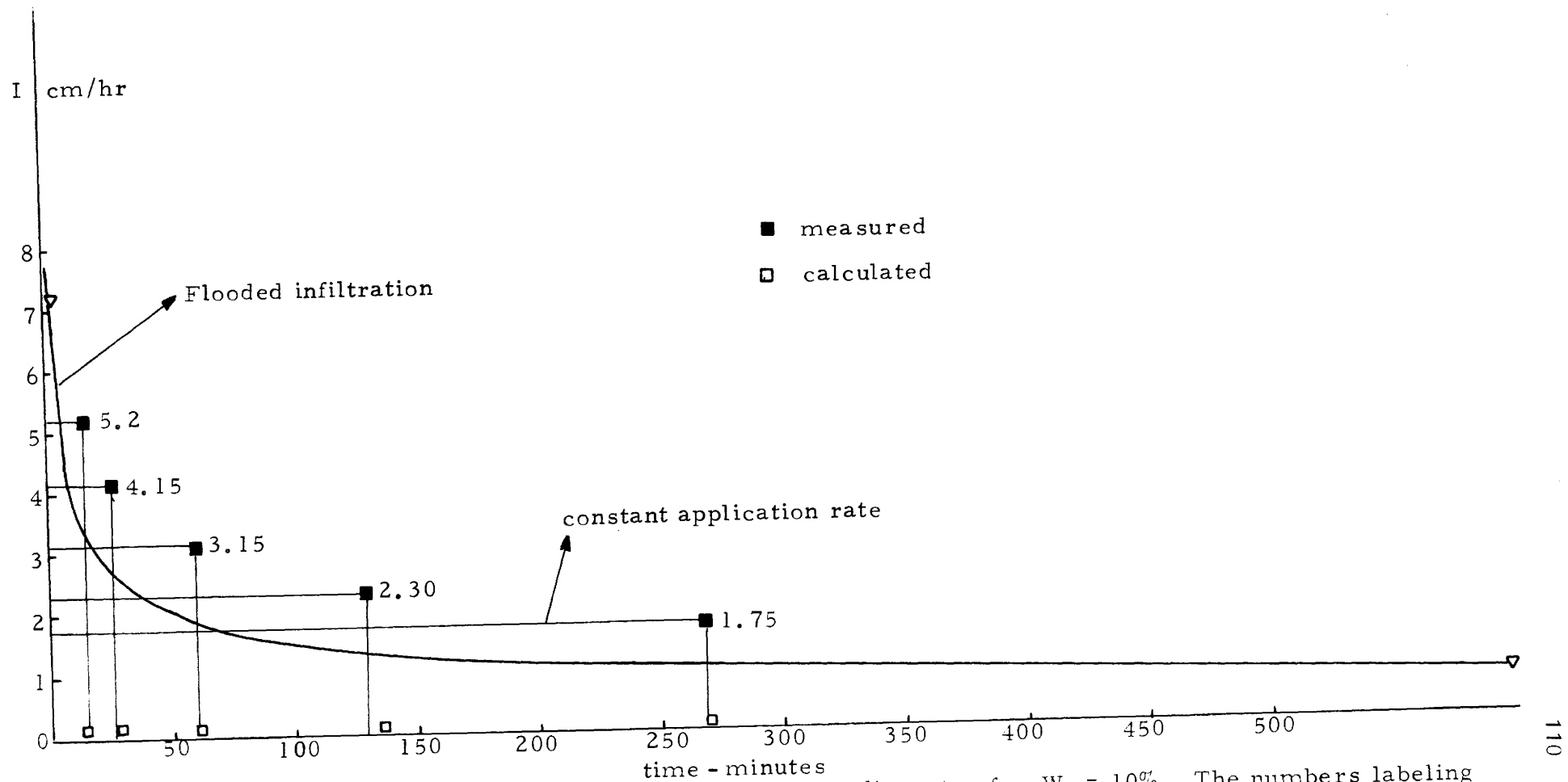


Figure 33. Effect of sprinkler application rate on surface ponding, t_s , for $W_i = 10\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

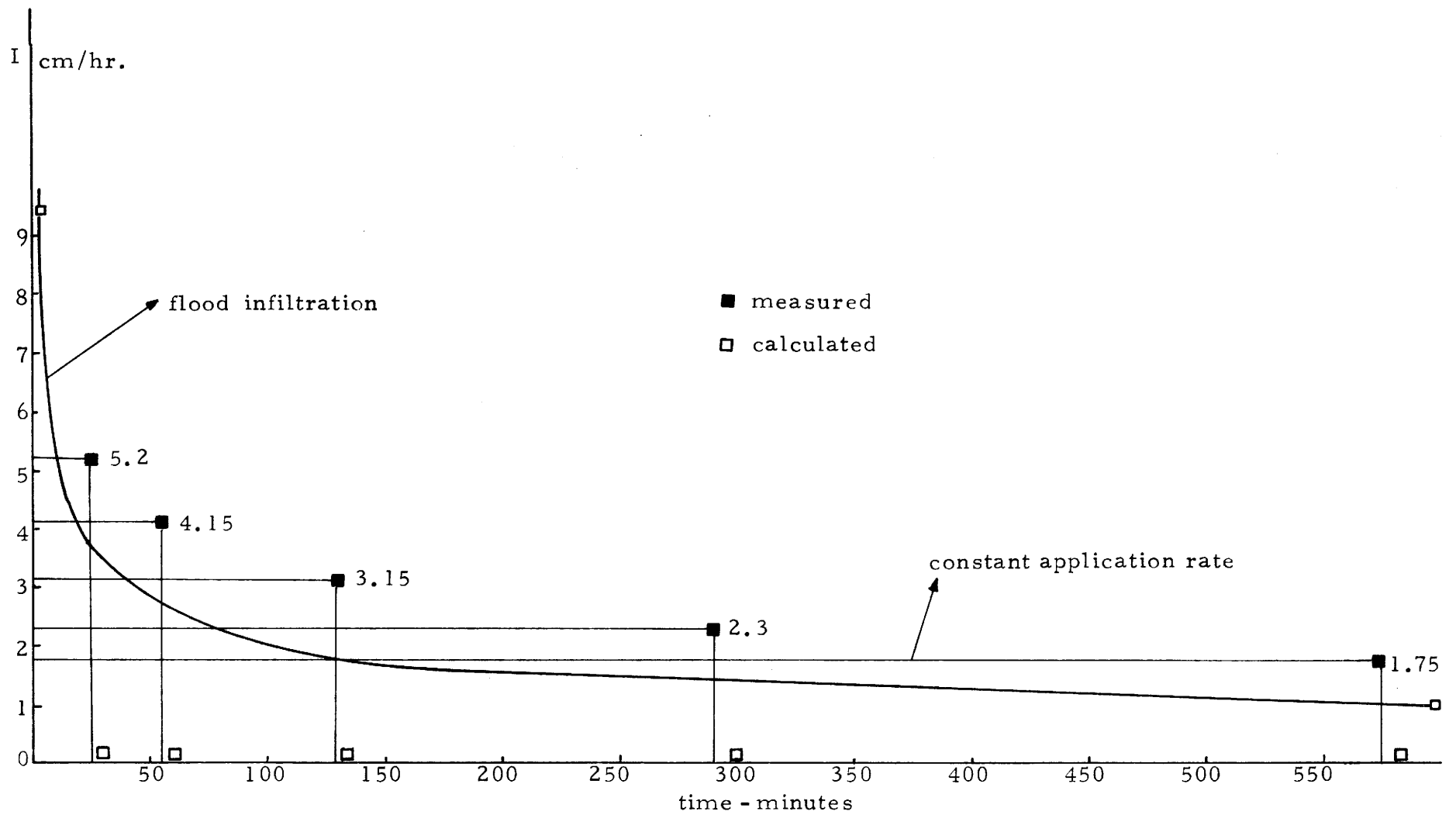


Figure 34. Effect of sprinkler application rate on surface ponding, t_s , for $W_i = 15\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

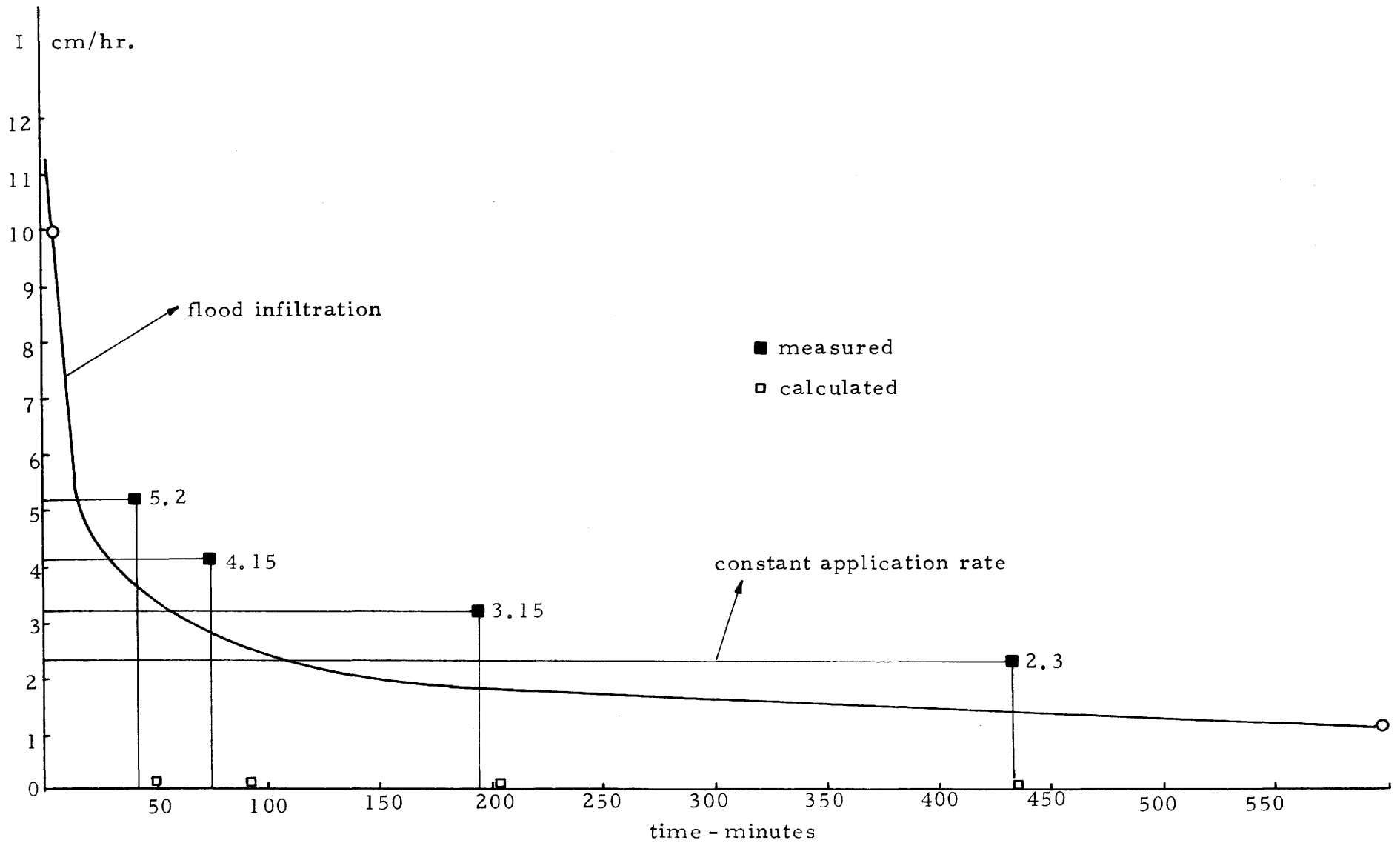


Figure 35. Effect of sprinkler application rate on surface ponding, t_s , for $W_i = 18\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

initial moisture contents (6.5, 10, 15, and 18 percent) are in good agreement. The curves also show that the theory can predict the time of surface ponding, t_s , within an error not more than 10 percent. In general, the error percent increases as the sprinkler application rate increases at wetter initial moisture content. A study of the curves (Figures 32, 33, 34, and 35) is of great interest.

The curves to imply that the water application rate and the initial moisture content of the soil (drier than its field capacity) have a pronounced effect on the time, t_s , at which surface saturation will occur.

The curves show that at the same initial moisture content, that t_s will take place sooner by increasing the rate of application, R . On the other hand, the drier the soil (initially drier than its field capacity) the sooner t_s will take place under the same rate of application. The reason for these results can be explained by two resistance forces (among other forces such as weight or gravity force, water pressure forces, and others) acting together on the soil particles. The first resistance force, F_1 , which is important as the soil becomes initially drier, is the resistance applied upward by the dry soil particles against wetting, since the angle of contact of the menisci at the wetting front approaches 90° (Van Duin, 1955).

The second resistance force, F_2 , which plays a big role as the rate of application, R , increases, can be called seepage force as introduced by Polubarinova-Kochina (1962). Considering a unit mass of water,

$$F_2 = - \frac{g}{K} R.$$

From the above discussion, one expects that the intermediate water application rate and initial moisture content can result in smaller runoff losses and the best uniformity of application. Figures 32, 33, 34, and 35 can predict the best condition of the initial moisture content (in the dry range) of the soil at which to be sprinkled by a specific rate of application to produce the best irrigation level.

The following example shows how simply one can obtain t_s from the intake rate curve of the soil.

Example. Suppose one likes to determine t_s under sprinkler intensity, R , equal to 3.15 cm/hr and initial moisture content, W_i , equal to 15 percent.

Using the intake rate Equation (54)

$$I = N t^n \quad (54)$$

at $W_i = 15\%$. Hence,

$$I = 13.428 t^{-0.40}, n = -0.40$$

and by integrating Equation (54) one obtains Equation (29)

$$Y = C t^a \quad (29)$$

and

$$Y = 0.373 t^{0.60}, C = 0.373$$

By applying Equation (91), t_s can be determined

$$t_s = (R/C)^{1/n} \quad (91)$$

$$t_s = (3.15/0.373)^{1/-0.40} = 135 \text{ minutes}$$

which is in good agreement with the measured one, 130 minutes (Figure 33).

And by applying Equation (89), rain infiltration (Y_R) can be determined

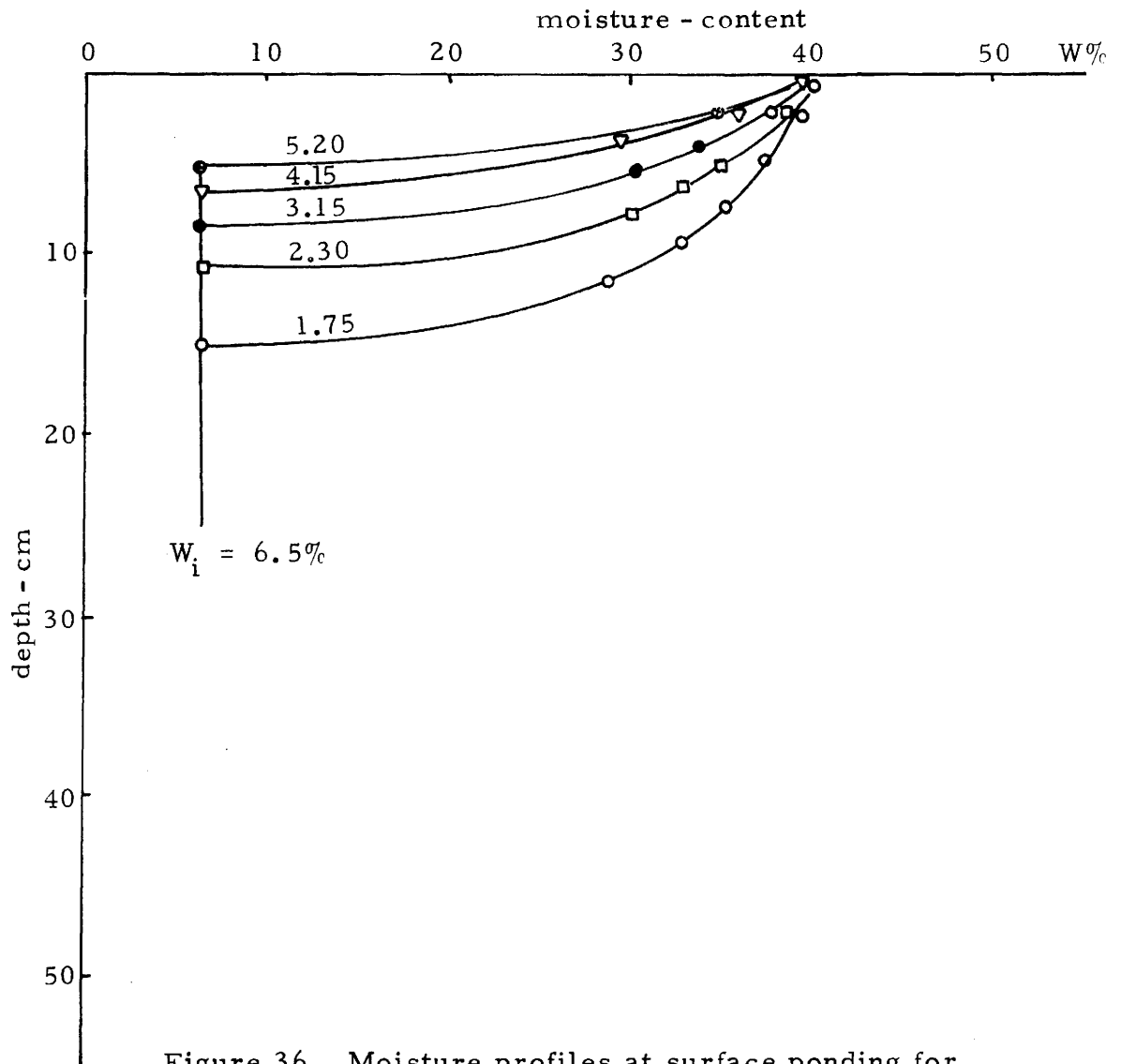


Figure 36. Moisture profiles at surface ponding for different application rates and $W_i = 6.5\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

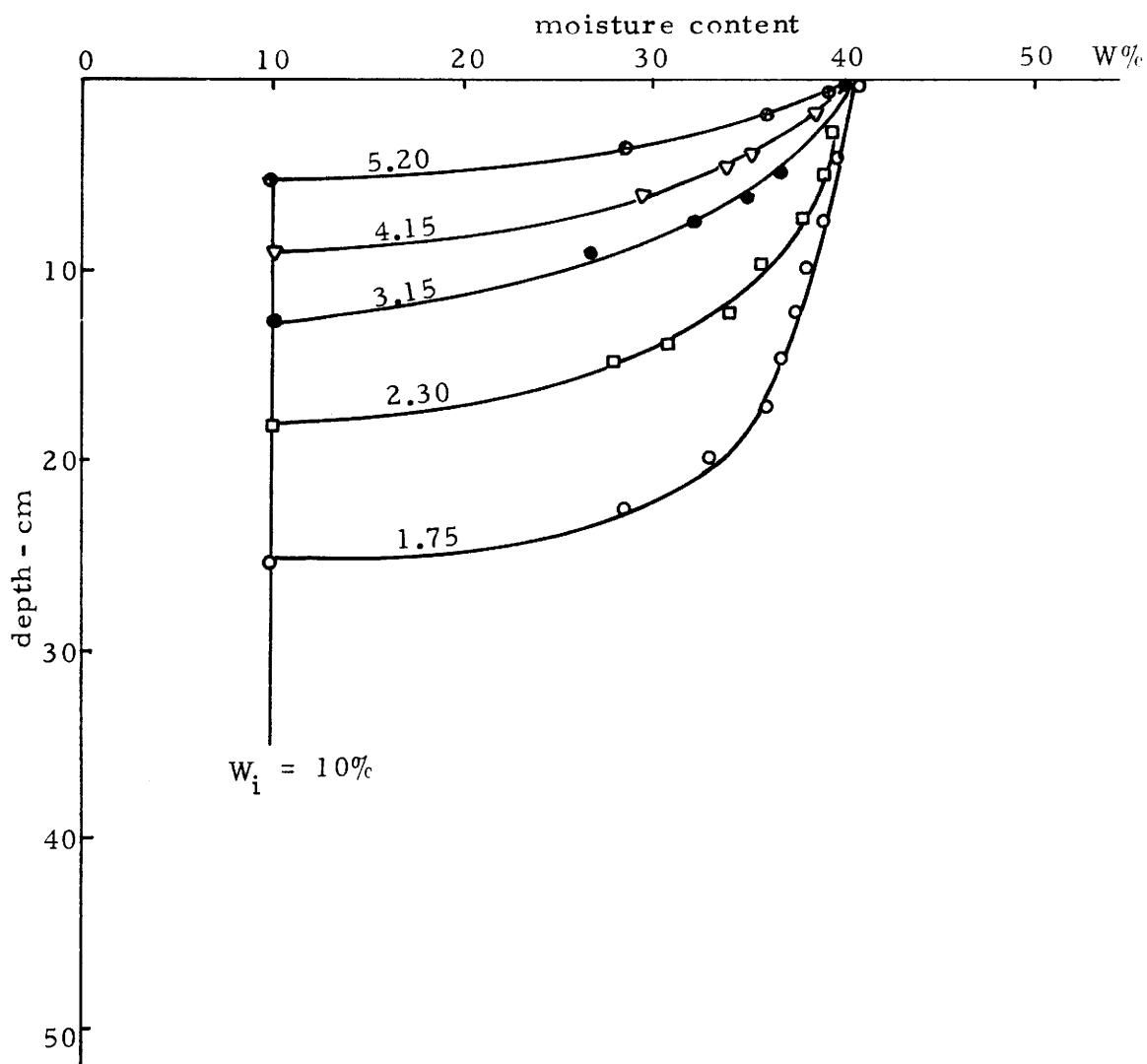


Figure 37. Moisture profiles at surface ponding for different application rates and $W_i = 10\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

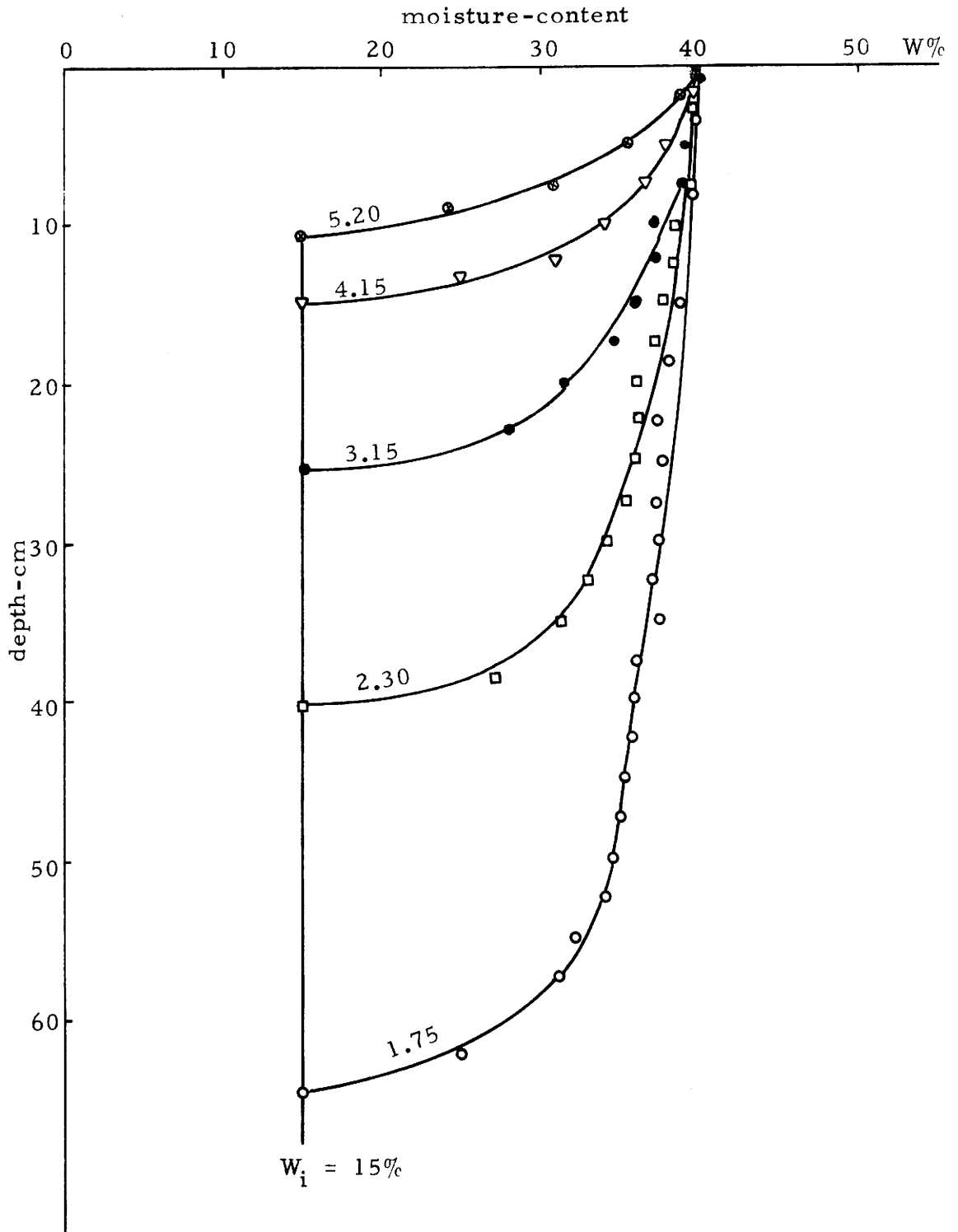


Figure 38. Moisture profiles at surface ponding for different application rates and $W_i = 15\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

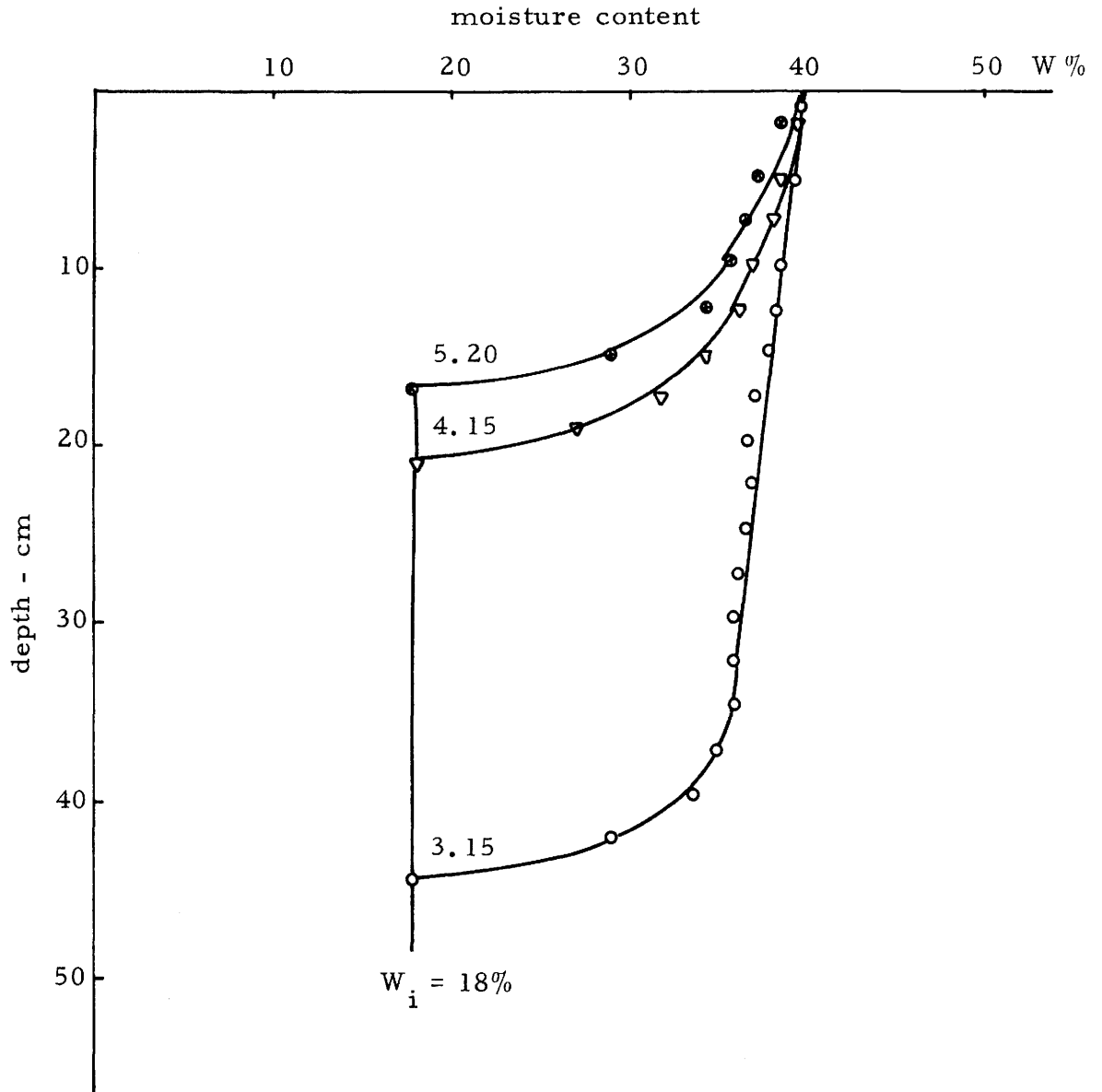


Figure 39. Moisture profiles at surface ponding for different application rates and $W_i = 18\%$. The numbers labeling the curves indicate the rate of application in cm/hr.

CHAPTER 6

SUMMARY AND CONCLUSIONS

Many situations arise in irrigation and hydrology where it is desirable to evaluate infiltration of water into the soil under either ponded or rainfall (sprinkler) conditions. An example is the uniformity of applied moisture in the soil profile can be considerably reduced by runoff on soils due to high application rates or low intake rates of the soils. Since a large percentage of the rainfall usually infiltrates into the soil, the accuracy with which the surface runoff can be predicted is heavily dependent on an accurate evaluation of infiltration.

The purpose of this study was to predict theoretically the time, t_s , at which rainfall (sprinkling) induces surface ponding by using the intake rate (instantaneous flood infiltration rate) of the soil which was obtained under the same initial conditions of the soil involved under rain (sprinkler) infiltration. Although one can achieve the same condition of initial moisture content and bulk density in both cases (flooding and sprinkling), it is difficult to achieve an infinitesimal depth of water on soil surface under flooding infiltration (since this depth is considered to be zero under sprinkler infiltration). However, a specially designed valve with Mariotte tube was used to obtain approximately 1.0 cm depth of water which was considered small enough for this study.

Since most of the irrigation practice takes place when the soil initially has moisture content between air dry and field capacity, the study was restricted to soils initially drier than its field capacity

to fit a straight line through a log-log plot of Y versus t. The constants C, a, N, and n were determined to give the intake rate in cm/hr.

Long soil columns (65 cm) with initial moisture contents of 0.10, 0.15, and 0.18 on weight basis were used under different low application rates (0.07 - 1.50 cm/hr, to avoid ponding) for obtaining the capillary conductivity-moisture relationship. The method used for calculating the capillary conductivity was suggested by Youngs (1964) and Childs (1969). The moisture characteristic curve of the soil (tension-moisture relationship) was also determined under the adsorption cycle. Both $K-\theta$ relationship and $\Psi-\theta$ relationship, in addition to the initial and boundary condition of the soil, were needed for solving Equation (7).

From the analysis and discussion of the flooded (ponded) infiltration, the following conclusions can be drawn:

1. The phenomenon of infiltration depends on complex interactions between initial moisture content, diffusivity, hydraulic conductivity, hydraulic gradient, and many other factors depending on soil type.
2. The infiltration phenomenon in the dry range (especially in soils possessing appreciable amount of clay) is completely different than in the wet range, and the rate of infiltration is increased by increasing the initial moisture content (initially drier than field capacity) in the Nibley silty clay loam soil.
3. The infiltration data fit the empirical equations (Equations 29 and 54) very well at least for the first ten hours.

$$Y = C t^a \quad (29)$$

$$I = N t^n \quad (54)$$

4. Under flooded infiltration, no matter what the initial moisture content (drier than field capacity), and if the same depth of penetration is achieved (after enough time to approach almost a steady state condition), the same transmission zone is obtained for the different initial moisture contents.
5. The moisture flow equation (non-linear partial differential equation) is capable of predicting infiltration rate, advance of wetting front, and moisture profile from the physical properties of the soil if the assumptions made in deriving the equation are satisfied.
6. Since the numerical solution of Equation (7) is greatly complicated by experimental difficulties attendant on measurement of the diffusivity or capillary conductivity of the soil, it is concluded that it will, generally, be more efficient to make experiment infiltration measurements than to determine soil moisture characteristics and then compute infiltration.

Rain (Sprinkler) Infiltration

Wetting front advance

Assuming steady state flow in the vertical direction in a homogeneous soil, the following Equation (86) which predicts the advance of the wetting front, Z , at various constant water application rates, R , and initial moisture contents, W_i (prior to surface ponding) is developed from Darcy's law

$$t_s = (R/C)^{1/n} \dots \dots \dots (91)$$

where C and n are constants to be determined from the intake rate equation.

The foregoing laboratory experiment carried out under sprinkler infiltration provided the data for t_s and moisture profiles at surface ponding.

The experimental results strongly supported the proposed theory and t_s can be predicted within an error of 10 percent. It is concluded that the drier the soil (initially drier than its field capacity) the sooner t_s will occur under the same rate of application, and at the same initial moisture content, the sooner t_s will take place by increasing the rate of application. It is also concluded that t_s decreases as Z decreases under any conditions of water application rate and initial moisture content. It is suggested from the above study that intermediate water application rates and initial moisture contents result in smaller runoff losses and the best uniformity of application.

Finally, it should be emphasized that the prediction equations developed here are valid for estimating cumulative infiltration, Y, wetting front advance, Z, moisture profiles for ponded infiltration, and Z, t_s , Y_R , etc., for various constant sprinkler rates and initial moisture contents on Nibley silty clay loam soil under the laboratory conditions imposed. In other words, it is not known whether or not the prediction equations will apply to all soils under any degree of uniformity regarding density and moisture content. Additional laboratory analyses will be necessary and are, therefore, suggested in order to test the limits and universal nature of the theory proposed.

CHAPTER 7

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APPENDIXES

APPENDIX A

Tables

Table 4. Cumulative flooded infiltration, Y, in centimeters, as a function of time, t, in minutes for three replicates at initial moisture content, W_1 , equal to 6.5 percent for Nibley silty clay loam soil.^{1a}

Elapsed time t	Replic 1		Replic 2		Replic 3	
	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y
0	3.7	0	2.5	0	18.0	0
1	-	-	4.0	0.32	20.3	0.49
2	4.8	0.23	4.7	0.46	21.0	0.63
3	5.2	0.31	5.3	0.59	21.4	0.72
4	5.6	0.40	-	-	-	-
5	6.0	0.48	6.0	0.74	22.10	0.87
6	-	-	-	-	-	-
7	-	-	-	-	-	-
8	7.2	0.74	6.7	0.89	23.0	1.10
9	-	-	-	-	-	-
10	7.8	0.86	7.2	0.99	23.4	1.14
12	8.2	0.95	-	-	-	-
15	8.9	1.09	8.3	1.22	-	-
17	-	-	-	-	24.8	1.43
20	9.8	1.28	9.2	1.41	25.3	1.54
25	10.7	1.47	-	-	-	-
30	-	-	11.1	1.81	26.9	1.88
40	13.0	1.96	12.8	2.17	28.4	2.19
50	14.4	2.25	14.2	2.47	29.7	2.47
60	15.7	2.54	15.3	2.70	-	-
70	-	-	-	-	32.0	2.95
75	17.5	2.91	-	-	-	-
85	-	-	18.0	3.27	-	-
90	18.6	3.14	-	-	-	-
100	19.7	3.37	19.5	3.59	35.2	3.63
125	22.2	3.90	-	-	-	-
130	-	-	22.2	4.16	38.1	4.24
150	24.7	4.43	23.9	4.52	39.9	4.62
180	-	-	26.3	5.02	42.5	5.17
200	29.0	5.34	27.9	5.36	44.2	5.53
230	-	-	30.1	5.82	-	-
250	32.9	6.16	-	-	-	-
290	-	-	-	-	51.0	6.96
300	36.6	6.94	35	6.86	-	-
350	40.2	7.70	-	-	-	-
400	43.6	8.41	-	-	58.4	8.52
430	45.6	8.84	-	-	-	-
450	-	-	-	-	61.3	9.14
460	-	-	44.9	8.95	-	-
500	-	-	47.1	9.41	64.3	9.77

Table 4. Continued.

Elapsed time t	Replic 1		Replic 2		Replic 3	
	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y
510	50.4	9.85	-	-	-	-
550	-	-	-	-	67.5	10.44
580	54.3	10.67	-	-	-	-
600	-	-	52.2	10.49	70.5	11.07
640	57.6	11.37	-	-	-	-
660	-	-	55.2	11.12	-	-
700	60.7	12.02	-	-	76.0	12.24
735	62.4	12.38	-	-	-	-
750	-	-	59.5	12.03	-	-
820	-	-	-	-	83.0	13.72
850	67.8	13.52	-	-	-	-
880	-	-	65.5	13.29	-	-
900	70.0	13.98	-	-	86.5	14.40
970	-	-	69.8	14.20	-	-

$$^a Y = (\text{cylind. reading} - \text{cylind. reading at } t = 0) \times 0.211.$$

Table 5. Cumulative flooded infiltration, Y, in centimeters, as a function of time, t, in minutes for four replicates at initial moisture content, W_i , equal to 10 percent for Nibley silty clay loam soil.^a

Elapsed time t	Replic 1		Replic 2		Replic 3		Replic 4	
	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y
	0	4.0	0.0	4.3	0	6.3	0	5.0
1	4.6	0.13	5.1	0.17	7.6	0.27	-	-
2	5.6	0.34	6.0	0.36	8.2	0.40	7.2	0.46
3	6.4	0.51	-	-	8.7	0.51	-	-
4	-	-	7.3	0.63	-	-	8.4	0.72
5	7.4	0.72	7.8	0.74	9.6	0.70	-	-
6	-	-	-	-	-	-	9.1	0.87
7	-	-	-	-	-	-	-	-
8	8.7	0.99	9.0	0.99	10.7	0.93	9.7	1.00
9	-	-	-	-	-	-	-	-
10	9.4	1.14	9.7	1.14	11.3	1.06	10.2	1.10
15	10.9	1.46	11.4	1.50	12.8	1.37	11.4	1.35
20	12.1	1.71	12.8	1.79	14.0	1.62	-	-
25	-	-	-	-	-	-	-	-
30	14.0	2.11	15.2	2.15	16.2	2.09	14.2	1.94
35	-	-	-	-	-	-	-	-
40	15.1	2.34	17.1	2.55	18.0	2.47	-	-
50	17.1	2.76	18.8	2.91	19.5	2.79	17.2	2.60
60	-	-	-	-	21.0	3.10	-	-
70	19.7	3.31	21.8	3.54	-	-	-	-
80	-	-	-	-	23.7	3.67	20.9	3.4
90	-	-	-	-	-	-	-	-
100	23.5	4.11	26.2	4.47	-	-	-	-
110	-	-	-	-	-	-	-	-
120	25.9	4.62	-	-	-	-	-	-
130	26.8	4.81	29.6	5.19	29.5	4.90	26.0	4.43
140	-	-	-	-	-	-	-	-
150	-	-	32.8	5.87	31.7	5.36	-	-
160	30.1	5.51	-	-	-	-	28.7	5.0
170	-	-	-	-	-	-	-	-
180	32.0	5.91	36.2	6.58	34.8	6.01	-	-
200	-	-	38.5	7.07	-	-	31.7	5.63
230	-	-	41.9	7.79	-	-	-	-
240	-	-	-	-	-	-	34.6	6.25
250	38.3	7.24	44.0	8.23	41.7	7.47	-	-

Table 5. Continued

Elapsed time t	Replic 1		Replic 2		Replic 3		Replic 4	
	Cylind. reading	Accum. intake	Cylind. reading	Accum. intake	Cylind. reading	Accum. intake	Cylind. reading	Accum. intake
		Y		Y		Y		Y
270	-	-	-	-	-	-	36.7	6.69
300	-	-	49.1	9.31	46.2	8.42	-	-
330	44.9	8.63	-	-	-	-	-	-
350	-	-	-	-	-	-	41.7	7.74
370	47.8	9.24	55.5	10.66	-	-	-	-
400	50.0	9.71	58.1	11.35	54.4	10.12	44.6	8.36
430	52.0	10.13	-	-	-	-	-	-
450	-	-	-	-	58.2	10.95	47.3	8.93
470	-	-	63.7	12.53	-	-	-	-
480	55.5	10.87	-	-	-	-	-	-
500	-	-	-	-	-	-	-	-
530	-	-	-	-	63.8	12.13	51.4	9.8
550	60.0	11.39	-	-	-	-	-	-
600	-	-	-	-	68.5	13.12	-	-
700	-	-	-	-	74.9	14.47	-	-
720	-	-	-	-	-	-	60.5	11.71
730	-	-	-	-	80.2	15.59	-	-
750	-	-	-	-	-	-	61.9	12.0

$$^a Y = (\text{cylind. reading} - \text{cylind. reading at } t = 0) \times 0.211.$$

Table 6. Cumulative flooded infiltration, Y, in centimeters, as a function of time, t, in minutes for four replicates at initial moisture content, W_i , equal to 15 percent for Nibley silty clay loam.^a

Elapsed time t	Replic 1		Replic 2		Replic 3		Replic 4	
	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y	Cylind. reading	Accum. intake Y
0	8.2	0	6.8	0	6.0	0	8.5	0
1	-	-	8.5	0.36	-	-	10.0	0.32
2	11.1	0.61	9.6	0.59	8.0	0.42	11.3	0.59
3	11.8	0.76	-	-	8.7	0.57	12.3	0.82
4	-	-	-	-	-	-	13.1	0.97
5	12.9	0.99	11.8	1.05	10.1	0.87	13.8	1.12
6	-	-	-	-	-	-	-	-
7	-	-	-	-	11.4	1.14	15.0	1.37
8	14.2	1.27	13.3	1.37	-	-	15.6	1.50
9	-	-	-	-	-	-	-	-
10	15	1.44	14.2	1.56	13.0	1.48	16.4	1.67
12	15.6	1.56	-	-	-	-	-	-
15	16.6	1.77	16.0	1.94	15.2	1.94	18.3	2.07
19	17.7	2.01	-	-	-	-	-	-
20	-	-	17.6	2.28	16.8	2.28	20.1	2.45
30	-	-	20.3	2.85	19.5	2.85	23.1	3.08
39	22.4	3.0	-	-	-	-	-	-
40	-	-	22.7	3.35	22.1	3.40	-	-
45	-	-	-	-	-	-	27.0	3.9
49	24.4	3.42	-	-	-	-	-	-
50	-	-	24.9	3.82	24.5	3.90	28.5	4.22
60	-	-	-	-	26.6	4.35	-	-
70	-	-	28.7	4.62	-	-	33.3	5.23
79	29.7	4.54	-	-	-	-	-	-
80	-	-	-	-	30.6	5.19	-	-
90	-	-	-	-	-	-	37.9	6.2
100	-	-	34.0	5.74	34.4	5.99	39.8	6.6
109	34.4	5.53	-	-	-	-	-	-
110	-	-	-	-	36.2	6.37	-	-
120	-	-	-	-	-	-	43.8	7.45
125	-	-	-	-	38.7	6.90	-	-
130	-	-	39.0	7.00	-	-	-	-
139	38.7	6.44	-	-	-	-	-	-
140	-	-	-	-	-	-	47.9	8.31
150	-	-	42.1	7.45	42.9	7.79	-	-
160	-	-	-	-	-	-	51.5	9.07
170	42.7	7.28	-	-	46.2	8.48	-	-
180	-	-	-	-	-	-	55.0	9.81
200	-	-	49.7	9.05	51.0	9.50	-	-
229	50.2	8.87	-	-	-	-	-	-
250	-	-	57.1	10.61	-	-	-	-
280	-	-	61.5	11.54	-	-	-	-
300	-	-	64.4	12.15	-	-	-	-

^aY = (cylind. reading - cylind. reading at t = 0) x 0.211.

Table 7. Cumulative flooded infiltration, Y, in centimeters, as a function of time, t, in minutes for four replicates at initial moisture content, W_i , equal to 18 percent.

Elapsed time t	Replic 1		Replic 2		Replic 3		Replic 4	
	Cylind. reading	Accum. intake Y^a	Cylind. reading	Accum. intake Y^a	Cylind. reading	Accum. intake Y^a	Cylind. reading	Accum. intake Y^a
0	17.0	0	20.5	0	5.4	0	8.1	0
1	19.3	0.49	22.7	0.46	7.7	0.49	10.5	0.51
2	20.0	0.63	24.0	0.74	9.0	0.76	12.0	0.82
3	20.5	0.74	25.0	0.95	9.9	0.95	13.10	1.06
4	-	-	-	-	-	-	-	-
5	21.5	0.95	26.6	1.29	11.3	1.24	14.7	1.39
6	-	-	-	-	-	-	-	-
7	-	-	-	-	-	-	-	-
8	22.7	1.20	28.4	1.67	12.8	1.56	16.7	1.81
9	-	-	-	-	-	-	-	-
10	23.6	1.39	29.5	1.90	13.7	1.75	17.7	2.03
15	25.3	1.75	32.0	2.43	15.9	2.22	20.1	2.53
20	27.0	2.11	34.1	2.87	17.7	2.60	22.1	2.95
30	30.1	2.76	37.8	3.65	21.0	3.29	25.5	3.67
40	33.0	3.38	41.1	4.35	23.9	3.90	28.7	4.35
50	35.7	3.95	44.2	5.00	26.6	4.47	31.7	4.98
70	40.9	5.04	49.8	6.18	31.8	5.57	37.1	6.12
80	43.3	5.55	52.4	6.73	34.3	6.10	39.7	6.67
90	-	-	55.1	7.30	-	-	-	-
100	48.0	6.54	57.7	7.85	39.1	7.11	44.6	7.70
110	-	-	60.1	8.36	41.4	7.60	47.1	8.23
120	52.3	7.45	-	-	43.8	8.10	-	-
130	54.3	7.87	-	-	-	-	-	-

$$^a Y = (\text{cylind. reading} - \text{cylind. reading at } t = 0) \times 0.211.$$

Table 8. Wetting front advance, Z , in centimeters, as a function of time, t , in minutes, under flooding infiltration, at initial moisture content, W_i equal to 6.5 percent for Nibley silty clay loam.

Z	Replic 1 t	Replic 2 t	Replic 3 t
0	0	0	0
2.5	8	6	6
5	30	27	25
7.5	60	54	52
10	96	91	90
12.5	140	135	133
15	185	188	185
17.5	242	245	-
20	302	308	292
22.5	365	-	-
25	430	-	425
27.5	-	525	499
30	572	607	573
32.5	652	691	650
35	734	785	736
37.5	-	887	820
40	920	985	910

Table 9. Wetting front advance, Z , in centimeters, as a function of time, t , in minutes, under flooding infiltration, at initial moisture content, W_i , equal to 10 percent for Nibley silty clay loam.

Z	Replic 1 t	Replic 2 t	Replic 3 t	Replic 4 t
0	0	0	0	0
2.5	4	4	4	4
5	14	14	16	16
7.5	30	29	30	35
10	54	47	52	58
2.5	82	69	79	89
15	112	95	-	124
17.5	146	123	146	160
20	183	156	185	210
22.5	-	190	221	261
25.0	270	225	264	-
27.5	-	265	305	383
30	372	-	-	455
32.5	423	-	395	526
35	483	389	443	609
37.5	545	435	-	681
40	607	485	549	768
42.5	-	-	602	-
45.0	-	-	670	-
47.5	-	-	732	-
50	-	-	795	-

Table 10. Wetting front advance, Z , in centimeters, as a function of time, t , in minutes, under flooding infiltration, at initial moisture content, W_i , equal to 15 percent for Nibley silty clay loam.

Z	Replic 1	Replic 2	Replic 3	Replic 4
	t	t	t	t
0	0	0	0	0
2.5	2	3	3	2
5	6	8	8	5
7.5	12	15	15	11
10	20	25	23	17
12.5	31	37	34	27
15	43	40	45	36
17.5	56	54	58	48
20	71	70	73	60
22.5	87	86	88	-
25.0	105	103	104	86
27.5	123	121	122	101
30	142	140	139	115
32.5	162	159	159	131
35	182	179	177	148
37.5	202	199	198	-
40	-	-	218	181
42.5	244	239	-	-
45.0	265	260	-	-
47.5	286	283	-	-
50	307	305	-	-

Table 11. Wetting front advance, Z , in centimeters, as a function of time, t , in minutes, under flooding infiltration, at initial moisture content, W_i , equal to 18 percent for Nibley silty clay loam.

Z	Replic 1 t	Replic 2 t	Replic 3 t	Replic 4 t
0	0	0	0	0
2.5	-	1	1	1
5	5	4	4	3
7.5	11	7	7	6
10	17	11	13	11
12.5	25	17	19	16
15	33	23	26	23
17.5	43	30	34	30
20	52	37	43	38
22.5	62	46	52	47
25	72	54	62	55
27.5	84	-	72	65
30	95	73	82	75
32.5	107	82	93	-
35	119	92	104	95
37.5	132	102	116	106
40	145	113	127	116

Table 12. Moisture profile, under flooding infiltration, at different initial moisture content, W_i , for Nibley silty clay loam.

depth cm	Moisture content as percent on weight basis, W							
	$W_i = 6.5\%$		$W_i = 10\%$		$W_i = 15\%$		$W_i = 18\%$	
	Replic 1	Replic 2	Replic 1	Replic 2	Replic 1	Replic 2	Replic 1	Replic 2
2	40.76	40.10	40.02	38.54	39.80	40.0	40.07	39.49
5	39.41	38.89	37.70	37.58	37.36	37.80	36.97	37.05
7.5	40.20	40.46	39.25	39.70	38.84	39.37	38.88	39.27
10	37.96	38.43	37.93	37.58	37.53	37.40	37.25	37.12
12.5	40.03	39.76	39.16	38.52	39.29	38.27	37.94	37.73
15	38.34	38.13	37.96	37.65	37.46	38.22	36.71	36.34
17.5	39.46	39.11	38.57	38.24	38.22	38.88	37.54	37.78
20	37.21	36.80	37.71	36.94	36.68	37.40	36.52	36.36
22.5	37.94	38.11	37.77	36.87	37.41	37.57	36.84	37.85
25.0	36.63	36.31	37.55	36.43	36.51	36.26	36.39	36.10
27.5	37.02	36.74	36.33	34.53	36.50	36.36	35.98	36.18
30	35.64	35.52	35.86	34.81	36.34	35.46	35.06	35.53
32.5	35.49	35.45	34.65	33.08	34.70	34.51	34.87	34.89
35	34.60	34.54	33.05	31.91	33.80	33.64	34.10	33.61
37.5	32.71	33.07	30.02	30.17	32.29	32.50	33.13	32.44
40	27.26	25.99	26.19	24.51	26.39	26.44	27.73	27.85

Table 13. Capillary potential (suction, Ψ , versus moisture content, W, of the soil during adsorption cycle (wetting curve) for Nibley silty clay loam.

W %	Ψ cm of water	W %	Ψ cm of water
34.66	26.30	34.84	34.00
24.96	108.45	30.56	80.00
27.56	93.87	34.60	35.10
28.46	80.62	38.13	12.00
27.43	66.34	24.60	250.0
23.32	309.10	24.78	210.0
26.67	139.18	24.87	215.0
26.48	138.01	21.11	560.0
31.18	34.34	42.50	8.0
26.29	74.01	31.26	70.0
29.36	71.00	20.02	810.0
32.25	44.70	22.08	415.0
31.92	41.30	21.78	470.0
29.85	54.54	24.88	240.0
29.78	54.40	26.73	135.0
26.88	113.54	25.28	160.0
26.79	108.10	25.32	165.0
30.89	60.70	41.47	7.0
30.59	60.00	42.15	6.5
29.01	55.12		

Table 14. Capillary conductivity, K, versus moisture content, W, using soil samples at three different initial moisture contents, W_i , for Nibley silty clay loam.

$W_i = 10\%$		$W_i = 15\%$		$W_i = 18\%$	
W %	K cm/hr	W %	K cm/hr	W %	K cm/hr
33.20	0.070	32.9	0.20	31.60	0.25
35.10	0.20	34.3	0.39	33.0	0.56
36.3	0.38	35.0	0.60	33.80	0.96
36.7	0.46	36.0	1.0	37.0	1.85
37.3	0.59	36.80	1.50	-	-
37.5	0.74	-	-	-	-

Table 15. Wetting front advance, Z , as a function of time, t , for different water application rates, R , in centimeters per hour, at initial moisture content, $W_i = 6.5\%$ for Nibley silty clay loam.

Z cm	t in minutes				
	R 1.75	R 2.30	R 3.15	R 4.15	R 5.20
0	0	0	0	0	0
0.5	-	-	5	3	2.5
1.0	-	-	-	-	5
1.5	-	-	-	-	-
2.0	-	-	15	11	9
2.5	31	25	19	14	12
3.0	-	-	-	-	-
3.5	47	-	28	19	-
4.0	-	-	-	-	-
4.5	-	45	35	-	-
5.0	64	51	41	-	-
6.0	77	60	-	-	-
7.5	99	80	-	-	-
8.5	110	-	-	-	-
9.0	120	-	-	-	-
10.0	135	-	-	-	-

Table 16. Wetting front advance, Z, as a function of time, t, for different water application rates, R, in cm/hr, at initial moisture content, $W_i = 10\%$ for Nibley silty clay loam.

Z cm	t in minutes				
	R 1.75	R 2.30	R 3.15	R 4.15	R 5.20
0	0	0	0	0	0
0.5	-	-	-	3	2
2.0	-	-	-	9.5	7
2.5	24	19	16	12	9
3.5	-	-	21	17	14
4.5	-	-	29	22	-
5.0	52	41	32	25	-
6.0	63	49	39	-	-
7.5	80	64	50	-	-
8.5	-	-	57	-	-
9.5	-	81	-	-	-
10.0	109	87	-	-	-
12.5	141	110	-	-	-
15.0	170	133	-	-	-
17.5	203	-	-	-	-
20.0	236	-	-	-	-
22.5	270	-	-	-	-

Table 17. Wetting front advance, Z , as a function of time, t , for different water application rates, R , in cm/hr, at initial moisture content, $W_i = 15\%$ for Nibley silty clay loam.

Z cm	t in minutes				
	R 1.75	R 2.30	R 3.15	R 4.15	R 5.20
0	0	0	0	0	0
2.5	16	14	10	8	7
3.5	-	-	-	-	9.5
4.5	-	-	-	-	13
5.0	34	29	22	17	14
6.0	-	-	28	22	17
7.5	54	45	35	27	22
10.0	74	63	48	37	-
12.5	95	81	62	48	-
15.0	-	100	76	-	-
17.5	139	117	91	-	-
20.0	161	139	105	-	-
22.5	182	158	120	-	-
25.0	206	178	-	-	-
27.5	-	199	-	-	-
30	253	214	-	-	-
32.5	277	239	-	-	-
35	-	254	-	-	-
40	349	-	-	-	-
50	444	-	-	-	-
55	491	-	-	-	-
60	539	-	-	-	-

Table 18. Wetting front advance, Z , as a function of time, t , for different water application rates, R , in cm/hr, at initial moisture content, $W_i = 18\%$ for Nibley silty clay loam.

Z cm	t in minutes				
	R 1.75	R 2.30	R 3.15	R 4.15	R 5.20
0	0	0	0	0	0
2.5	-	12	8	7	6
3.5	-	-	-	-	-
5.0	-	24	18	15	12
7.5	-	38	28	23	19
10.0	-	51	39	32	26
12.5	-	66	50	41	33
15.0	-	81	62	50	42
17.5	-	96	74	60	-
20.0	-	-	85	-	-
22.5	-	126	97	-	-
25.0	-	141	109	-	-
27.5	-	156	122	-	-
30	-	172	134	-	-
32.5	-	187	146	-	-
35	-	-	158	-	-
37.5	-	218	170	-	-
40	-	232	183	-	-
45	-	263	-	-	-
50	-	295	-	-	-
55	-	326	-	-	-
60	-	373	-	-	-

Table 19. Time of surface ponding, t_s , and depth of penetration, Z , for various sprinkler intensities, R , at different initial moisture contents for Nibley silty clay loam.

R cm/hr	$W_i = 6.5\%$		$W_i = 10\%$		$W_i = 15\%$		$W_i = 18\%$	
	t_s	Z	t_s	Z	t_s	Z	t_s	Z
	min	cm	min	cm	min	cm	min	cm
1.75	165	11.5	270	22.5	575	63.5	-	-
2.30	80	7.5	130	15.0	290	39.5	435	72.0
3.15	44	5.3	58	8.5	127	23.0	195	42.5
4.15	19	3.5	26	5.0	56	13.0	75	18.2
5.20	12	2.6	14	3.5	25	8.0	42	15

APPENDIX B

Figures

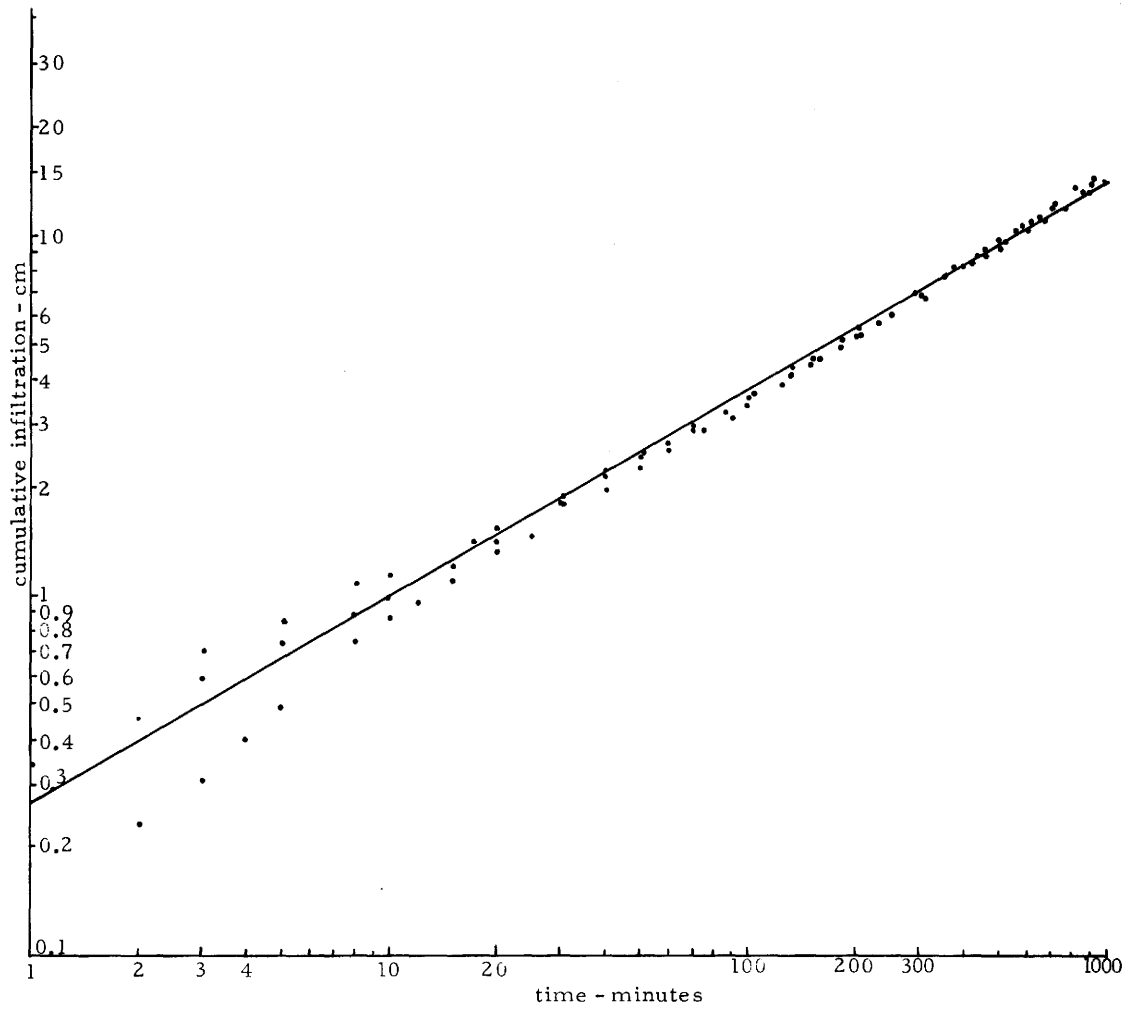


Figure 40. A linear regression line through flooded infiltration data, for three replicates, at $W_1 = 6.5\%$. For Nibley silty clay loam soil.

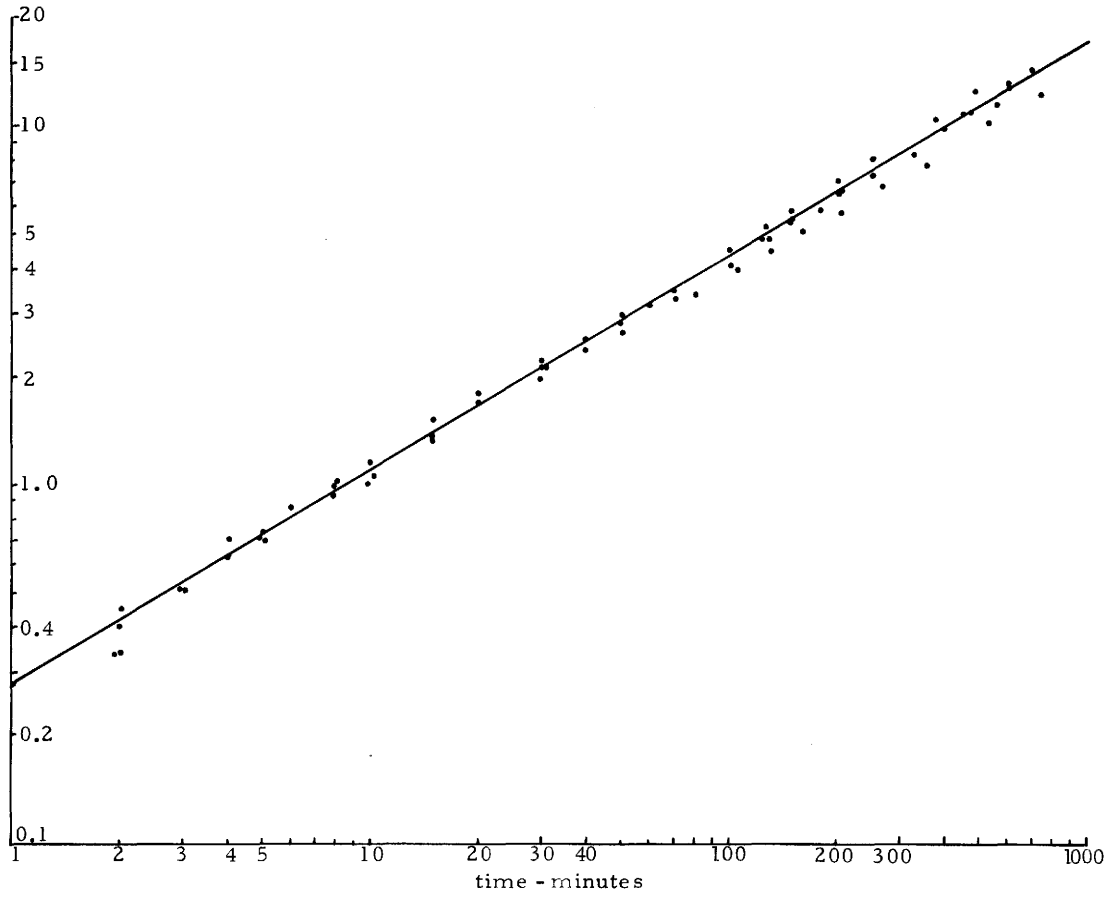


Figure 41. A linear regression line through flooded infiltration data, for four replicates, at $W_i = 10\%$. For Nibley silty clay loam soil.

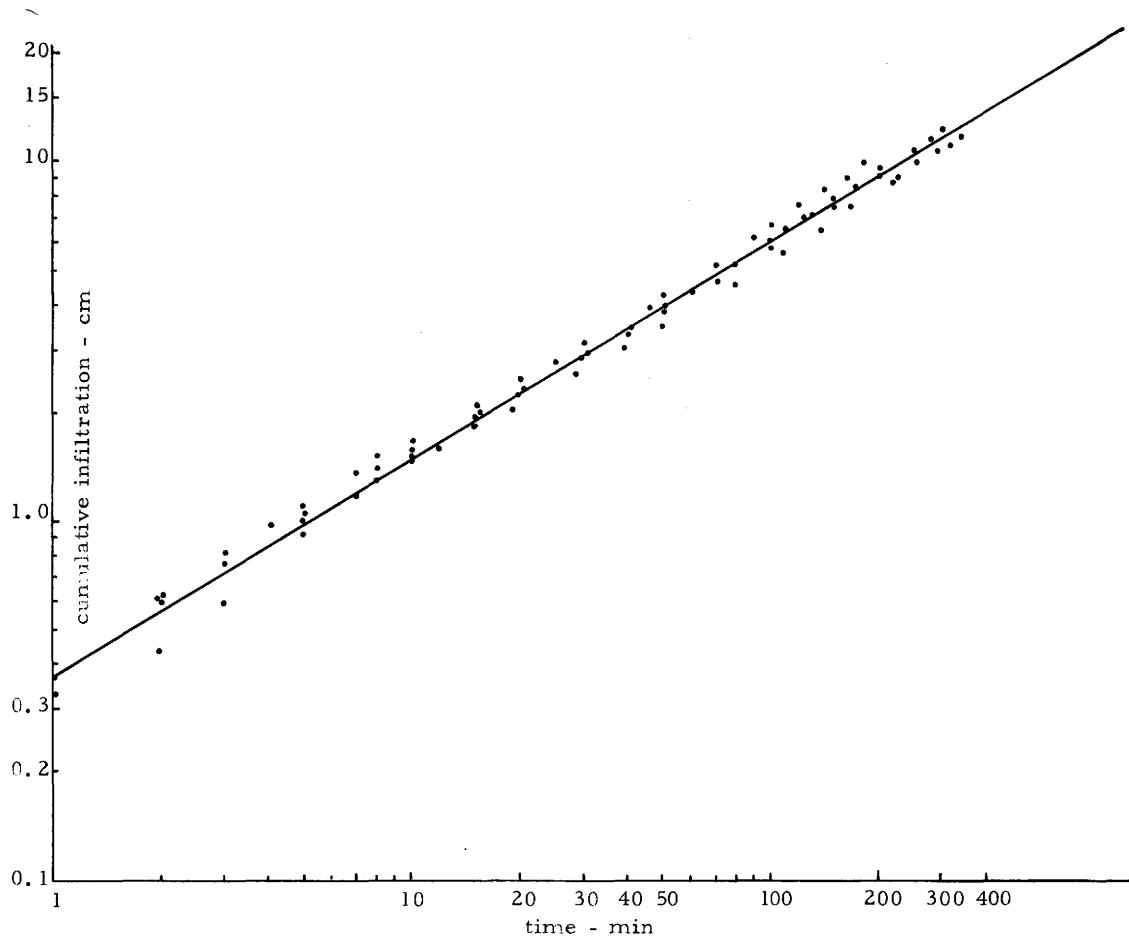


Figure 42. A linear regression line through flooded infiltration data, for four replicates, at $W_i = 15\%$ for Nibley silty clay loam soil.

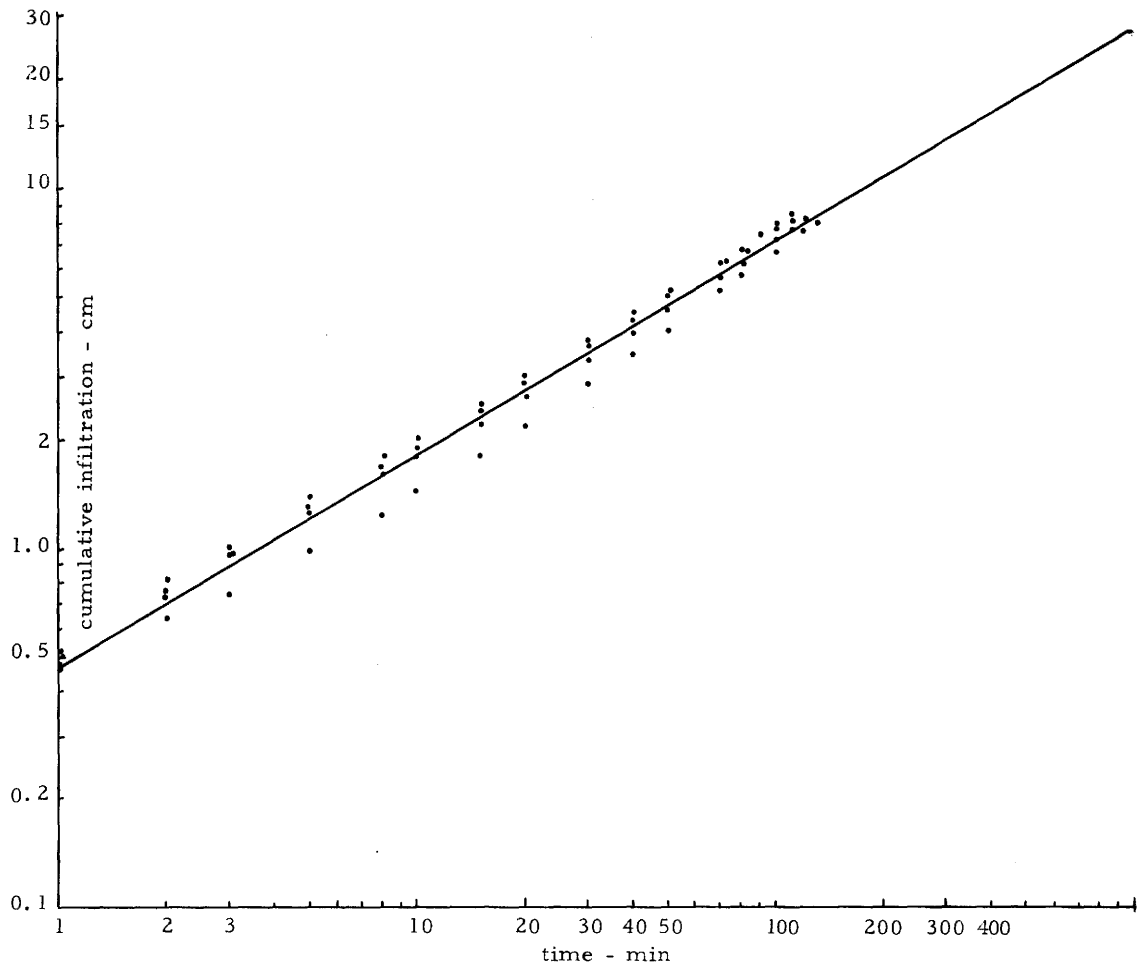


Figure 43. A linear regression line through flooded infiltration data, for four replicates, at $W_1 = 18\%$ for Nibley silty clay loam soil.

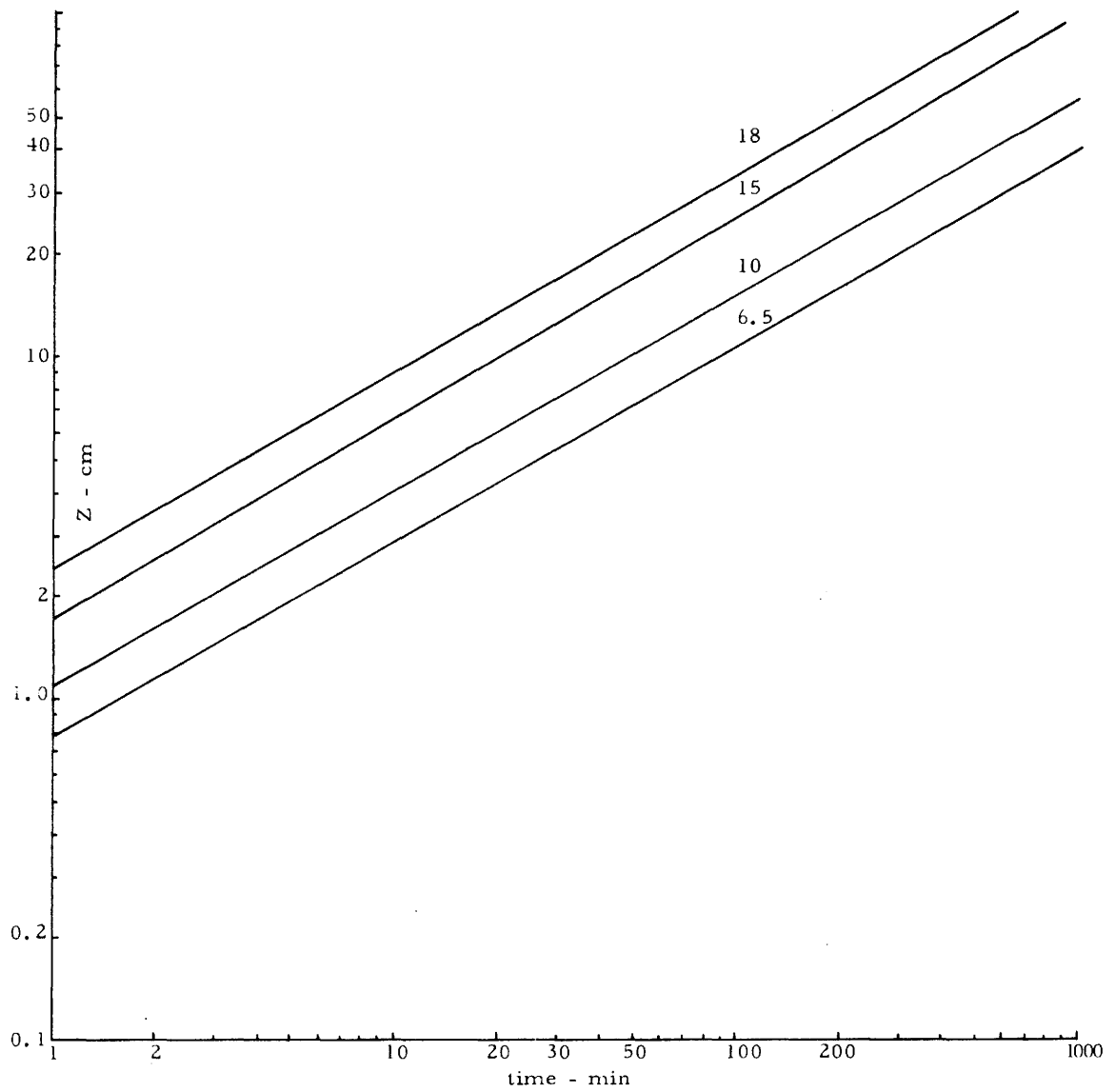


Figure 44. Wetting front advance under flooded infiltration. The numbers labeling the curves indicate the initial moisture content, W_i , as percent. Each curve represents four replicates.

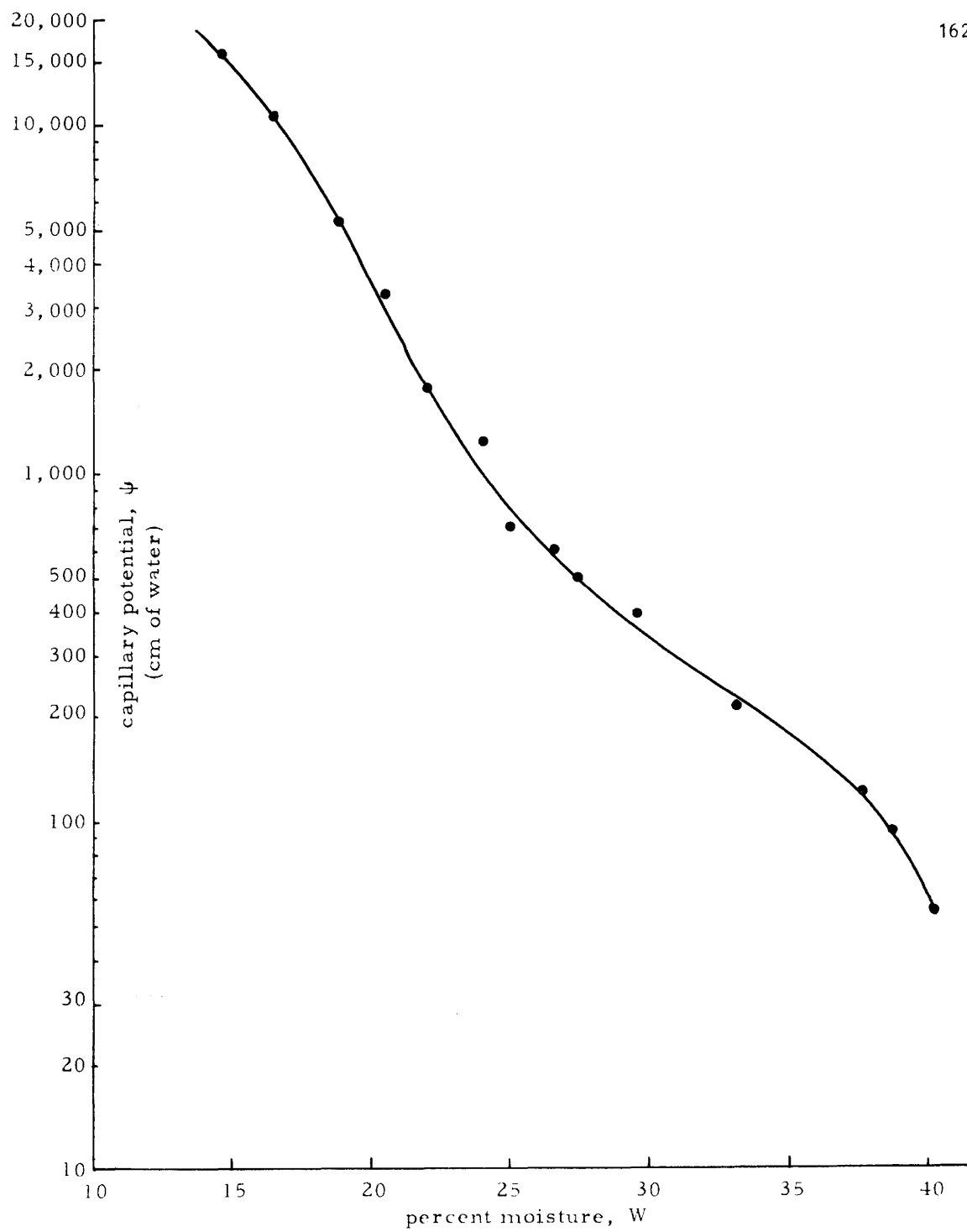


Figure 45. Capillary potential (suction) versus moisture content, W , during desorption cycle for Nibley silty clay loam soil.

APPENDIX C

Fortran Programs

Linear Regression Line

```

1*  C LINEAR REGRESSION LINE
2*      DIMENSION XX(200),YY(200),X(200),Y(200)
3*      N=78
4*      READ(5,10)(XX(I),I=1,N)
5*  10  FORMAT(8F10.5)
6*      WRITE(6,11)(XX(I),I=1,N)
7*  11  FORMAT(10X,F10.5)
8*      READ(5,20)(YY(I),I=1,N)
9*  20  FORMAT(8F10.5)
10*     WRITE(6,21)(YY(I),I=1,N)
11*  21  FORMAT(30X,F10.5)
12*     DO 9 I=1,N
13*     X(I)=ALOG10(XX(I))
14*  9   Y(I)=ALOG10(YY(I))
15*     SUM=0.0
16*     DO 30 I=1,N
17*     SUM=SUM+X(I)
18*  30  CONTINUE

19*     AN=N
20*     XM=SUM/AN
21*     WRITE(6,25)XM
22*  25  FORMAT(1H1,20X,'AVERAGE X=',E16.7)
23*     SN=0.0
24*     DO 40 I=1,N
25*     SN=SN+Y(I)
26*  40  CONTINUE
27*     YM=SN/AN
28*     WRITE(6,26)YM
29*  26  FORMAT(1H1,20X,'AVERAGE Y=',E16.7)
30*     SF=0.0
31*     SG=0.0
32*     DO 50 I=1,N
33*     SF=SF+(X(I)-XM)*(Y(I)-YM)
34*     SG=SG+(X(I)-XM)**2.0
35*  50  CONTINUE
36*     B=SF/SG
37*     WRITE(6,15)B
38*  15  FORMAT(1H1,20X,2HB='F10.5)
39*     A=YM-(B*XM)
40*     A1=2.3026*A
41*     AA=EXP(A1)
42*     WRITE(6,16)A,A1,AA
43*  16  FORMAT(/20X,F10.5)
44*     STOP
45*     END

```

END OF UNIVAC 1108 FORTRAN V COMPILATION.

D *DIAGNOSTIC* MESSAGE(S)

Numerical Solution for the Partial Differential Equation (7)

```

1*  C      COMPUTATION OF INFILTRATION DEVELOPED BY HANKS-BOWERS
2*  C      CWF IS COMULATIVE WATER FLOW(INFILTRATION)
3*  C      K IS NO. OF DELX INCREME T RAIN TO WET,2 IF WET TO DRAIN,3 ETTING
4*  C      IER IS NO. OF V ELEMENTS          S A M ESW A   _KDA ,TING
5*  C      CONDITION CHANGES—ODD          OE      6
6*  C      P IS PRESSURE TABLE ( DR          E      6
7*  C      D IS CONDUCTIVITY TABLE STARTING WITH LOWEST WATER CONTENT VALUE
8*  C      C IS WATER CAPACITY AS A FUNCTION OF DEPTH BEGINNING AT TOP
9*  C      W IS WATER CONTENT AS A FUNCTION OF DEPTH BEGINNING AT TOP
10* C      H IS WATER PRESSURE AS A FUNCTION OF DEPTH BEGINNING AT TOP
11* C      V IS BOUNDARY CONDITIONS AT TOP AND TIMES CONDITIONS APPLY
12* C      DELX IS DEPTH INCREMENT
13* C      DETT IS TIME INCREMENT TO START WITH AND LOWEST TO USE
14* C      GRAVY IS GRAVITY COMPONENT USUALLY THE SAME AS DELX
15* C      CONQ IS SMALLEST WATERCONTENT CHANGE ALLOWED EACH COMPUTATION
16* C      DELW IS WATER CONTENT DIFFERENCE CORRESPONDING TO TABLE INCREMENTS
17* C      T IS WATER CONTENT TABLE HAS EQUAL SPACED INCREMENTS
18* C      TIME IS CUMULATIVE TIME AT START OF COMPUTATION
19* C      TT IS 1.0 FOR LAASONEN AND 0.5 FOR CRANK NICHOLSON
20* C      CUMT IS TIME AT END OF COMPUTATION
21* C — — TAA=1, FOR ZERO FLUX AT BOTTOM, TAA=0 FOR H(K) CONSTANT
22* C      CONSH IS IS VALUE BY WHICH H(I) MUST BE DIFFERENT, OR GREATER
23* C      FROM G(I) OR H(I)=G(I)
24* C      CTM IS LOWEST VALUE OF DELT PERMITTED— IF AS LOW STOPS
25* C      HDRY IS PRESSURE OF LOWEST POSSIBLE WATER CONTENT
26* C      HWET IS PRESSURE OF HIGHEST POSSIBLE WATER CONTENT
27* C      WATL IS LOWEST POSSIBLE WATER CONTENT
28* C      WATH IS HIGHEST POSSIBLE WATER CONTENT
29* C      CB IS A CONSTANT TO MULTIPLY D ARRAY BY—USUALLY 1.0
30* C      WFRU IS WATER FLOW UP
31* C — — K IS NO. OF DELX INCREMENTS,MM NO. OF TIMES H+W PRINTED,KIT NO.OF
32* C — — START HERE FOR A NEW PROGRAM
33* C — — MI IS TO PRINT H,W ARRAYS EACH ITER., IER NO. OF V ELEMENTS
34* C — — MYST IS 0 FOR ALL DRAIANGE,1FOR HYSTERESUS,3 FOR WETTING
35* DIMENSION Y(99), C(99), G(99), B(99), E(99), F(99), V(99), DD(99)
36* DIMENSION H(99),W(99),P(39),D(39),T(39)
37* READ 163, ML
38* LMM=0
39* 1      LMM=LMM+1
40* READ 163, K,MM,IER
41* KK=K+1
42* LL=MM
43* READ 165, P
44* READ 165, D
45* DO 200 I=1,KK
46* W(I)=7.1875
47* 200   CONTINUE
48* READ 105, (V(I),I=1,IER)
49* READ 105, DELX,DETT,GRAVY,CONQ,DELW,TIME
50* READ 105, TT,CUMT,TAA,CONSH,CTM
51* READ 105, HDRY,HWET,WATL,WATH,CB

```

```

52*      WRITE (6,169)
53*      WRITE (6,164) K,MM,IER
54*      WRITE (6,171)
55*      WRITE (6,500)P
56*      WRITE (6,173)
57*      WRITE (6,500)D
58*      T(1)=0.0625
59*      D(1)=(D(1))*(P(2)-P(1))*CP
60*      DO 2 I=2,39
61*      D(I)=D(I)*(P(I)-P(I-1))*CB+D(I-1)
62*      2   T(I)=DELW+T(I-1)
63*      C   D IS NOW DIFFUSIVITY TIMES DELW NOT CONDUCTIVITY
64*      DO 3 I=1,KK
65*      J=(W(I)-T(I))/DELW+1.0
66*      3R=(W(I)-T(J))/DELW
67*      832 H(I)=(P(J+1)-P(J))*BB+P(J)
68*      3   CONTINUE
69*      WRITE (6,166) (H(I),I=1,KK)
70*      WRITE (6,179)
71*      WRITE (6,166) (V(I),I=1,IFR)
72*      WRITE (6,180)
73*      WRITE (6,166) DELX,DETT,GRAVY,CONG,DELW,TIME
74*      WRITE (6,181)
75*      WRITE (6,166) TT,CUMT,TAA,CONSH,CTH
76*      WRITE (6,172)
77*      WRITE (6,166) HDRY,HWET,WATL,WATH,CB
78*      WRITE (6,183)
79*      WRITE (6,166) D
80*      EOR=V(1)
81*      DETT=DETT
82*      TME=1.0-TT
83*      TBB=1.0-TAA
84*      N=KK
85*      YMAX=WATH
86*      DD(1)=0.0
87*      DO 14 I=1,KK
88*      DD(I+1)=DELX+DD(I)
89*      G(I)=H(I)
90*      14  Y(I)=W(I)
91*      NIT=0
92*      PIT=0.0
93*      DO 15 I=2,K
94*      15  PIT=W(I)+PIT
95*      KCC=1
96*      KCK=1
97*      KCH=1
98*      RUNOF=0.0
99*      CUMS=0.0
100*      CUMB=0.0
101*      CUMH=0.0
102*      CALL PLOT (KK,WATH,W,DD)
103*      WRITE (6,166) TIME
104*      C-- --COMPUTATION OF CONDUCTIVITY (R) AND WATER CAPACITY (C)
105*      16  TOP=WATH
106*      BOT=WATL
107*      IF (FOR=0.0) 17,19,18
108*      17  W(1)=WATL
109*      H(1)=HDRY

```

```

110*      GO TO 19
111*  18   W(1)=WATH
112*      H(1)=HWET
113*  19   TWW=(W(1)+Y(1))*0.5
114*      J=(TWW-T(1))/DELW+1.0
115*      BR=(TWW-T(J))/DELW
116*      DIFFA=(D(J+1)-D(J))*BR+D(J)
117*      HI=(P(J+1)-P(J))*RB+P(J)
118*      DO 37 I=1,K
119*      TW=(W(I+1)+Y(I+1))*0.5
120*      J=(TW-T(1))/DELW+1.0
121*      BB=(TW-T(J))/DELW
122*      DIFFB=(D(J+1)-D(J))*BB+D(J)
123*      GI=(P(J+1)-P(J))*BR+P(J)
124*      IF (TW-TWW) 20,32,20
125*  20   B(I)=(DIFFA-DIFFB)/(HI-GI)
126*      IF (I-1) 21,21,33
127*  21   IF (EOR-0.0) 27,31,22
128*  22   ER=(P(1)*((H(1)-H(2)+G(1)-G(2))/2.0+GRAVY))/DELX
129*      IF (ABS(EOR-ER)-APS(0.1*EOR)) 33,33,23
130*  23   IF(KCK-10)305,33,33
131*  305  KCK=KCK+1
132*      IF (EP-EOR) 24,33,26
133*  24   IF (W(1)-WATH) 25,33,33
134*  25   BOT=W(1)
135*      W(1)=(W(1)+TOP)*0.5
136*      GO TO 28
137*  26   IF (W(1)-WATL) 33,33,27
138*  27   TOP=W(1)
139*      W(1)=(W(1)+BOT)*0.5
140*  28   J=(W(1)-T(1))/DELW+1.0
141*      BR=(W(1)-T(J))/DELW
142*      IF (EOR-0.0) 30,31,30
143*  30   H(1)=(P(J+1)-P(J))*RB+P(J)
144*      GO TO 19
145*  31   B(1)=0.0
146*      GO TO 33
147*  32   B(I)=(D(J+1)-D(J))/(P(J+1)-P(J))
148*      IF (I-1) 33,21,33
149*  33   TWW=TW
150*      HI=GI
151*      DIFFA=DIFFB
152*      TW=(W(I+1)+Y(I+1))*0.5
153*      J=(TW-T(1))/DELW+1.0
154*  35   C(I+1)=DELW/(P(J+1)-P(J))
155*  37   CONTINUE
156*      KCC=1
157*      KCK=1
158*  38   POT=DELX*DELX/DELT
159*  C-- --COMPUTATION OF TRIANGULAR MATRIX MAIN BODY
160*      DO 42 I=2,K
161*      BB=C(I)*POT/TT+B(I)+R(I-1)
162*      DA=(C(I)*POT+C(I)+R(I)*(TM*(G(I+1)-G(I))-GRAVY)+R(I-1)*(TM*(G(I-1)
163*      1-G(I))+GRAVY))/TT
164*      IF (I-2) 39,39,40
165*  39   DA=DA+R(I-1)*H(I-1)
166*      F(I)=DA/BB
167*      E(I)=B(I)/BB

```

```

168*      GO TO 42
169*  40  IF (I-K) 41,43,43
170*  41  E(I)=B(I)/(BB-B(I-1)*E(I-1))
171*      F(I)=(DA+B(I-1)*F(I-1))/(BB-B(I-1)*E(I-1))
172*  42  CONTINUE
173*  43  BB=BB-TAA*B(I)
174*      DA=DA+TAA*B(I)*((G(I)-G(I+1))*TM+GRAVY)/TT+TBB*R(I)*H(KK)
175*      H(I)=(DA+B(I)*F(I-1))/(BB-B(I)*E(I-1))
176*  44  I=I-1
177*      H(I)=E(I)*H(I+1)+F(I)
178*      IF (I-2) 45,45,44
179*  45  IF (TAA-1.0) 47,46,46
180*  46  H(KK)=H(KK)+GRAVY
181*  47  IF (EOR-D.0) 49,48,49
182*  48  H(1)=H(2)-GRAVY
183*      GO TO 54
184*  49  IF (W(1)-WATL) 54,54,50
185*  50  IF (W(1)-WATH) 51,54,54
186*  51  WFRD=(B(1)*((H(1)-H(2)+G(1)-G(2))/2.0+GRAVY))/DELX
187*      IF (ABS(WFRD-ER)-ABS(ER*0.1)) 54,54,52
188*  52  IF (DELT-DELT) 54,54,53
189*  53  DELT=DELT*0.5
190*      GO TO 38
191*  54  DO 60 I=2,KK
192*      IF (H(I)-HWET-DD(I)*GRAVY/DELX) 56,56,55
193*  55  H(I)=HWET
194*  56  IF (H(I)-G(I)) 57,59,58
195*  57  IF (0.01*H(I)+(H(I)-G(I))) 60,59,59
196*  58  IF (0.01*H(I)-(H(I)-G(I))) 60,59,59
197*  59  H(I)=G(I)
198*  60  CONTINUE
199*      NN=0
200*  C--- COMPUTATION OF WATER CONTENTS AS A FUNCTION OF PRESSURES JUST COMP
201*      DO 117 I=1,KK
202*      IF (I-1) 61,61,62
203*  61  IF (EOR-D.0) 117,62,117
204*  62  IF (H(I)-G(I)) 65,116,65
205*  65  NHI=39
206*      NLO=1
207*      J=25
208*  66  IF (H(I)-P(J)) 67,72,68
209*  67  NHI=J
210*      GO TO 69
211*  68  NLO=J
212*  69  JT=J
213*      J=(NHI-NLO)/2+NLO
214*      IF (J-JT) 66,70,66
215*  70  IF (H(I)-P(J)) 71,72,72
216*  71  J=J-1
217*  72  WAT=(H(I)-P(J))*DFLW/(P(J+1)-P(J))+T(J)
218*  73  TW=WAT
219*  113 W(I)=TW
220*      GO TO 117
221*  116 W(I)=Y(I)
222*  117 CONTINUE
223*      SUM3=0.0
224*      SUM1=0.0
225*      SUM2=0.0

```

```

226*      SUMA=0.
227*      DO 131 I=2,K
228*      SUM1=W(I)+SUM1
229*      SUM2=Y(I)+SUM2
230*      IF (ABS(SUM1-SUM2)-ABS(SUM3)) 131,131,130
231* 130  SUM3=SUM1-SUM2
232* 131  CONTINUE
233*      IF (ABS(SUM3)-ABS(CON0))134,134,132
234* 132  IF (DELT-DETT) 134,134,133
235* 133  DELT=DELT*0.5
236*      GO TO 38
237* 134  WFRD=(R(1)*((H(1)-H(2)+G(1)-G(2))/2.0+GRAVY))/DELX
238*      CWF=(SUM1-PIT)*DELX
239*      WFRDD=(SUM1-SUM2)*DELX/DELT
240*      WFRU=(R(K)*((H(K)-H(KK)+G(K)-G(KK))/2.0+GRAVY))/DELX
241*      CUMS=WFRD*DELT+CUMS
242*      CUMB=WFRU*DELT+CUMB
243*      CUMM=WFRDD*DELT+CUMM
244*      CWF LX=(SUM1-SUM2)*DELX
245* 700  CONTINUE
246* 706  IF (EOR-0.0) 136,136,135
247* 135  RUNOF=(EOR-WFRD)*DELT+RUNOF
248* 136  TIME=TIME+DELT
249*      IF (LL-MM) 138,137,137
250* 137  CALL PLOT (KK,WATH,W,DD)
251*      WRITE (6,166) (H(I),I=1,KK)
252*      LL=0
253*      WRITE (6,184)
254* 138  WRITE (6,166) TIME,CWF,EOR,DELT,RUNOF,WFRU
255*      IF (SUM3-0.0) 139,141,139
256* 139  TW=ABS(CON0*DELT/SUM3)
257*      IF (TW-CTM) 152,140,140
258* 140  IF (TW-0.1*DETT) 141,142,142
259* 141  TW=DETT*0.1
260*      GO TO 144
261* 142  IF (TW-100.0*DETT) 144,144,143
262* 143  TW=100.0*DETT
263* 144  DELT=TW
264* C---TEST TO SEE IF EVAP OR RAIN INTENSITY (EOR) HAS CHANGED
265*      I=1
266* 145  IF (TIME-V(I+1)) 148,147,146
267* 146  I=I+2
268*      GO TO 145
269* 147  CALL PLOT (KK,WATH,W,DD)
270*      WRITE (6,166) (H(I),I=1,KK)
271*      WRITE (6,166) TIME,CWF,EOR,DELT,RUNOF,WFRU
272*      DELT=DETT
273*      EOR=V(I+2)
274*      W(1)=W(2)
275*      H(1)=H(2)
276*      GO TO 151
277* 148  IF (TIME+DELT-V(I+1)) 150,150,149
278* 149  DELT=V(I+1)-TIME
279*      IF (DELT-CTM) 147,147,150
280* 150  EOR=V(I)
281* 151  LL=LL+1
282*      IF (TIME-CUMT) 153,152,152
283* 152  IF (ML-LMM) 162,162,1

```

```

284* 153 Y (1)=(W(1)+Y(1))*0.5
285* J=(Y(1)-T(1))/DELW*1.0
286* BR=(Y(1)-T(J))/DELW
287* 155 G(1)=(P(J+1)-P(J))*98+P(J)
288* 156 DO 161 I=2,KK
289* TW=(W(I)-Y(I))+W(I)
290* IF (TW-WATH) 157,157,159
291* 157 IF (TW-WATL) 158,160,160
292* 158 TW=WATL
293* GO TO 160
294* 159 TW=WATH
295* 160 Y(I)=W(I)
296* W(I)=TW
297* G(I)=H(I)
298* 161 CONTINUE
299* KCH=1
300* NIT=0
301* GO TO 16
302* 162 STOP
303* C-----
304* 105 FORMAT(7F10.4)
305* 500 FORMAT(E8.2)
306* 163 FORMAT(20I3)
307* 164 FORMAT(20I3)
308* 165 FORMAT(10E8.2)
309* 166 FORMAT(10E12.4)
310* 167 FORMAT(80I1)
311* 168 FORMAT(2X,80I1)
312* 169 FORMAT(11H K MM IER)
313* 171 FORMAT(43H WETTING PRESURE STARTING WITH LOWEST VALUE)
314* 172 FORMAT(53H HDRY HWET WATL WATH CR)
315* 173 FORMAT(41H CONDUCTIVITY STARTING WITH LOWEST VALUE)
316* 179 FORMAT(93H FLUX1 TIME 1 FLUX 2 TIME 2 FLUX
317* 1 3 TIME 3 FLUX 4 TIME 4)
318* 180 FORMAT(66H DELX DETT GRAVY CONG DELW
319* 1 TIME)
320* 181 FORMAT(54H TT CUMT TAA CONSH CTR)
321* 183 FORMAT(45H DIFFUSIVITY DATA--SUMMATION OF D TIMES DELW)
322* 184 FORMAT(68H TIME CWF EOR DELT SUN)
323* 10F WFRU)
324* 185 FORMAT(55H WLO WHI DRUT DTGP SINK)
325* END

```

END OF UNIVAC 1108 FORTRAN V COMPILATION. G *DIAGNOSTIC* MESSAGE(1)

Subroutine Program for Plotting

```

1*      SUBROUTINE PLOT(N,WMAX,WVALUE,XVALUE)
2*      DIMENSION ALINE(101),WVALUE(99),XVALUE(99)
3*      DATA FILL,AXIS,CHAR/1H ,1H ,1HW/
4*      WRITE(6,7) WMAX
5*      DO 1 J=1,101
6* 1     ALINE(J)=AXIS
7*      WRITE(6,8) (ALINE(K),K=1,101)
8*      DO 2 J=1,101
9* 2     ALINE(J)=FILL
10*     ALINE(1)=AXIS
11*     DO 4 L=1,N
12*       J=100.0*(WVALUE(L)/WMAX)+1.5
13*       IF (J.LT.1.OR J.GT.101) GO TO 17
14* 10    ALINE(J)=CHAR
15* 12    WRITE(6,9) XVALUE(L),WVALUE(L),(ALINE(K),K=1,101)
16*     ALINE(J)=FILL
17* 55   ALINE(1)=AXIS
18* 4     CONTINUE
19*     DO 5 J=1,101
20* 5     ALINE(J)=AXIS
21*     WRITE(6,8) (ALINE(K),K=1,101)
22*     RETURN
23* 7     FORMAT (15H XVALUE WVALUE,5X,17H MAX WAT CONT IS,F7.4)
24* 8     FORMAT (31X,101A1)

25* 9     FORMAT (1H ,F6.1,F9.4,7H          ,101A1)
26*     END

```

END OF UNIVAC 1108 FORTRAN V COMPILATION. 0 *DIAGNOSTIC* MESSAGE(S)

VITA

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Ph.D. (Irrigation Science)	Utah State University, 1970

Teaching Experience:

1963-1964	Teaching assistant of soil physics, Soil Science Department, University of Alexandria
1964-1965	Teaching assistant of irrigation and drainage, Institute of Land Reclamation, Alexandria

Research Experience:

1965-1967	University of California, Davis
1968-present	Utah State University

Academic Honors:

Received honor degree with the B.S., University of Alexandria

Honored with an academic achievement medal by the president
of U.A.R. (Egypt) in 1964

Granted a scholarship to complete Ph.D. degree.