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EXPERIMENTAL AND THEORETICAL ANALYSES OF
CAPILLARY RISE THROUGH POROUS MEDIA

presented by

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EXPERIMENTAL AND THEORETICAL ANALYSES OF CAPILLARY RISE THROUGH POROUS MEDIA

By

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ABSTRACT

EXPERIMENTAL AND THEORETICAL ANALYSIS OF CAPILLARY RISE THROUGH POROUS MEDIA

By

George C. Zalidis

The objectives of this study were 1) to examine the process of capillary rise through soils, 2) to test numerical solutions of Richards' equation in describing upward water flow in soils and 3) to use the experimental and theoretical analyses in a case study. For this purpose capillary rise experiments were performed under near constant temperature conditions using Metea sandy loam soil (Arenic Hapludalts) packed in columns of different bulk density.

Adsorption soil moisture characteristic curves were obtained from these experiments. The data were analysed and information for understanding the physics of the phenomenon was gained. All curves exhibited a theta-straight region, followed by a parabolic curve flattened at higher suctions. The parabolic curve could be separated into three different zones. The different zones suggested transitions in the water content-potential relationships. These transitional zones can be associated with relative percentages of filled pores and thickness of water films. It was found necessary to use a power function in each of the three zones to successfully fit the adsorption data. Conductivity and diffusivity functions were determined from adsorption data.

FINDIT was tested against a finite-difference scheme and experimental data. When moisture characteristics, cumulative infiltrations and fluxes were used as criteria, the agreement between the two models and experimental data was very good. FINDIT performed excellently using internally generated derivatives and predicted extremely accurate experimental profiles. It agreed well with a simple algebraic equation for calculating cumulative infiltration.

To test the influence of a wetting agent in capillary rise, both treated and untreated soils were used. When experimental and simulated wetting curves of these soils were analysed, the treated soil exhibited higher cumulative infiltration and rise of wetting front but lower water capacity. A change in the kinetics of capillary rise was observed when wetting films approach a critical thickness which resulted in a significant slow down of the wetting process. At equilibrium, water content gradients serve as the upward driving force but are countered by gravity. The wetting curves obtained from capillary rise can be used to determine capacity and intensity factors in addition to proportionality coefficients and flux factors of soil water.

To my parents

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I owe special thanks to my parents and my brother for their support and encouragement which began from my youth, to all my friends and their wives and especially to Alex Kopatsis and his wife whose support, encouragement and inspiration went beyond the limits of our friendship. I also owe many thanks to my fiancee Connie for her unlimited love and understanding.

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LIST OF SYMBOLS

Symbols	Meaning
a, b	arbitrary constant
D	soil-moisture diffusivity (cm^2/min)
$D(\theta)$	D as function of θ
g	acceleration due to gravity (cm/sec^2)
i	cumulative infiltration (cm)
i^m	matric component of cumulative infiltration
i^g	gravitational component of cumulative infiltration
$i+1/2, j+1/2$	indices referring to average quantities
i, j	summation indices
K	unsaturated hydraulic conductivity (cm/min)
$K(\theta_s)$	saturated hydraulic conductivity (cm/min)
$K(\theta_i)$	calculated conductivity for a specified moisture content (cm/min)
K_s	measured saturated conductivity (cm/min)
K_o	calculated saturated conductivity (cm/min)
K_i	hydraulic conductivity at moisture content θ_i (cm/min)
$K(\theta)$	K as a function of θ
K_n	hydraulic conductivity at initial moisture content
q	flux density (m^3/m^2 min)
S	sorptivity ($cm/min^{1/2}$)

t	infiltration time (min)
t_o	total infiltration time (min)
$t^{1/2}$	square root of time
V_o, $\partial i/\partial t$	infiltration rate (cm/min)
V_o^m	matric component of infiltration rate
V_o^g	gravitational component of infiltration rate
z	wetting distance in vertical direction (cm)
z'	wetting distance in horizontal direction (cm)
y	vertical component of flow attributed to gravity
θ_o, θ_s	volumetric moisture content at saturation (m^3/m^3)
θ, θ_v	volumetric moisture content (m^3/m^3)
θ_m, θ_g	gravimetric moisture content (cm^3/g)
θ_n, θ_i	initial moisture content (m^3/m^3)
$\theta(x, t)$	θ as a function of x and t
C	specific water capacity
dh/dz	hydraulic gradient in one-dimensional space
$\partial h/\partial x$, $\partial h/\partial y$, $\partial h/\partial z$	hydraulic gradients in x-, y- and z-directions
$\partial \theta/\partial z$	volumetric moisture content gradient
$\partial \theta/\partial t$	volumetric moisture content rate of change with time
$\partial K/\partial z$	hydraulic conductivity gradient
h_e	air entry potential
h	total potential (cm or kpa)
h_m	matric potential

h_g	gravitational potential
∇h	matric potential gradient in three-dimensional space
h_o	pneumatic pressure head
∇	vector differential operator
Δ	small change in flux distance or time
λ	Boltzmann transformation ($cm/min^{1/2}$)
$\lambda(\theta)$	λ as a function of θ
χ, Ψ, ω, ---, fm	parameters which are single-valued functions of θ
ρ_b	density of water (g/cc)
γ	surface tension ($dynes/cm = g/sec^2$)
$\int$	integral sign (operator)
sfl	specific flux
v_{av}	average flux
$\sum$	summation sign

INTRODUCTION

The upward movement of water into and through soil is of great importance and concern to the agricultural and petroleum industries. Knowledge of the rise of water above a water table and the changes of soil water content in sub-irrigation of plants, evaporation, and rise of the water table are necessary for good water and land management.

Mathematical equations are helpful in the prediction of water content changes and rise of the wetting front in the soil profile. The complex nature of the soil-water system makes it difficult to specify directly the forces acting on soil water. The description of soil water movement depends not only on the forces residing in the soil but also on the amount of water present. Both diffusion and convection may occur in liquid and gaseous phases while the solid matrix determines the path length of diffusing molecules and the cross-sectional area available for these processes. Understanding the mechanism by which water moves upward through soil is extremely important for promoting good soil-plant relationships. Capillary rise of water in soil has been observed for a long period of time; however, only few experimental and theoretical investigations have been published.

Like infiltration and drainage, capillary rise of water in soil is a process which is driven towards equilibrium. Equilibrium conditions are ideal for measuring physical parameters of the soil like proportionality coefficients, fluxes, water capacity and soil water potential. Under such conditions these parameters are more representative in characterizing the hydraulic properties of a given soil than measurements taken prior to equilibrium conditions.

The mathematical equation for describing liquid water movement combines the continuity and Darcy equations and was derived by Richards (1931). Because of the strong non-linearity of this partial differential equation, a general analytical solution has not been found. Numerical solutions, however, may be used to obtain solutions of the partial differential equation. An iterative numerical method based on flux terms (FINDIT) and a finite-difference implicit scheme were used to obtain the solutions in this study.

The objectives of this study were:

1). To examine the mechanism of capillary rise, an adsorption process, and compare it with the desorption process of water retention in soils

2). To determine the conductivity and diffusivity functions which are necessary inputs to both FINDIT and finite-difference equations to model capillary rise.

3). To examine how well the two numerical methods based on Richards' equation describe upward water movement.

4). To use capillary rise as an experimental method to test the efficacy of wetting agent treated soils to improve water retention and water movement through the soil matrix.

CHAPTER I

CAPILLARY RISE IN METEA SOIL

1. INTRODUCTION

Capillary rise is defined as the wetting of a soil from a lower elevation; hence wetting of soil is opposed by the force of gravity. Wollny (1885) was the first to publish his experiments on capillary rise. The well known equation for the height of rise, h, in capillary tubes is:

$$h - 2\gamma \cos(\varphi)/ g\rho r \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [1]$$

where γ is the surface tension, ρ the density of the liquid, g the acceleration due to gravity, r the radius of the tube and φ the angle of wetting. This formula explains the rise of water in soils, the smaller the pore diameter the higher the rise of water. Pores in coarse soils are larger than in fine ones, thus water rise will be lower than in fine textured soils with smaller pores. The wetting of soils through capillary rise is frequently characterized by the pF-curve, i.e, the relation between the negative log of tension in water films and the water content. Peerlkamp and Boekel (1960) gave some examples of soils where the moisture content of the pF-curve at saturation remained the same with increasing negative pressure (tension). McLaughlin (1924) observed that the zone of maximum moisture content was appreciably above the water table, frequently referred to as theta-straight zone. Thereafter, the moisture content decreased gradually with increasing tension.

Although the capillary tube theory is applicable to most soil systems, the energetic theory of soil moisture as introduced by Buckingham (1907) and improved by Gardner (1922) is more appropriate. Buckingham (1907) attributed the rise of water in soil to differences in the surface curvature of soil water masses lying between adjacent soil grains. The resulting increasing tension in the water masses is supposed to lift water from the reservoir at the bottom of the soil column, move it through the connecting films, and use part of it at least, in increasing the radius of curvature of the water masses immediately above the last wetted grain. Water ceases to rise when the upward and downward forces in the wetted region are in equilibrium.

According to Keen (1924) the following three rather distinct moisture horizons exist in a capillary system: "the first of these should be marked by complete saturation, the next by a gradual decrease of water content and the last by a region of incomplete and decreasing moisture content approaching the initial condition of the soil."

Wadsworth (1931), plotted log-rise of the wetting front against log-time and suggested that capillary rise is a compound process dominated in the first phase by an action which may result in complete filling of pore spaces, and in the later phase by an action which results in incomplete filling of pore spaces and a continuous, uniform decrease in moisture content with height. When coarse materials are used, the transition between these two processes can be noted by inspection of the log-log plot of rise versus time curve. With fine materials, inspection of the moisture distribution is required. Subsequent further movement of water into the dry zone is at a very slow rate some of which may be attributed to vapor movement.

Swartzendruber (1955) used Poiseuille's analysis, assuming an idealized soil model consisting of parallel, non-interconnected capillary tubes. He found that it is possible within limits to predict from the intersection of behavior of capillary rise curves the behavior of the capillary adsorption coefficient (sorptivity) as it is affected by changes in pore radii, surface tension and wetting angle. Because of favorable qualitative comparisons between the capillary tube theory and experimental data, he suggested that the capillary tube theory can be refined to determine the wetting angle which is now a difficult but determinable soil physical property.

Temperature plays an important role in all water flow phenomena in soils. Surface tension decreases with increasing temperature. Therefore, an unsaturated soil at a given potential holds less water at higher than at a lower temperature. This results in an increased rate of water uptake and lower pressure potentials throughout the column when temperature decreases. A rise in temperature causes the water potential to increase whereas a drop in temperature has the reverse effect. Nielsen, et. al. (1962) found deviations from a linear square root of time versus distance relationship when wetting soil in a horizontal mode. Contrarily, the results of Stockinger, et. al. (1965) taken in a constant temperature room showed linearity and they concluded that fluctuations in temperature are an additional driving force for water movement,causing the deviations from linearity in Nielsen's work.

The macroscopic definition of hydraulic conductivity stems from Darcy's law:

$$q = -K\nabla h \qquad [2]$$

where q is the specific discharge vector and ∇h is the gradient of the hydraulic head. For unsaturated soils ($\theta<\theta_s$), Equation [1] is of a limited applicability in the field (Braester et. al., 1971). The dependence of K upon the fluid and matrix properties and the sequence of wetting and drying processes is more complex than in the saturated state. Darcy's law, however, is still generally used as a working tool and applied to a wide range of unsaturated flow problems. Because water potential is difficult to measure, measurements of water potential distributions are not always made, and instead, water content measurements are made and are frequently recorded in the literature. However, to use Darcy's law with such data the moisture characteristic must be generated from models or other characteristics of the soil.

The quality of water used when measuring water content profiles in soil columns is a major concern. Childs and Collis-George (1950) reported that it was impossible to maintain the initial rate of flow in saturated soils when tap water was used because the dissolved air coming out of solution in the soil pores reduced the moisture content and permeability of the soil. The trouble did not arise with unsaturated soils because the continuous air filled pore space provided a leak path for the escape of the momentarily trapped air soon after its emergence from solution. In unsaturated soils the use of boiled water is not necessary. On the other hand, for unsaturated experiments which last for long periods of time, like capillary rise experiments, the use of deaired water can create differences in fluxes with time as the deaired water will change to become aerated with time. Thus the use of regular tap, not deaired water, is recommended. Precautions should be taken to prevent water loss or evaporation through acrylic plastic and other impermeable construction materials. This is particularly important for small values of soil-water flux (Willis, et. al. 1965, Nielsen, et. al. 1972).

Soil water content distributions may be measured nondestructively using gamma-attenuation techniques (Ferguson and Gardner, 1963; Rawlins and Gardner, 1963; Davidson, et. al., 1963) or neutron attenuation (Gardner and Calissendoff, 1967). Gravimetric sampling has been successfully used but many samples are required. To obtain a soil column with sufficient uniformity offers the greatest difficulty (Cassel, et. al. 1968). The soil packer developed by Jackson, et. al. (1962) allows several columns to be packed to the same average bulk density and to nearly the same bulk density distribution.

It is very useful to describe water flow using water content as a driving force. Combining Darcy's law with the continuity equation and applying the chain rule for the one dimension horizontal flow yields:

$$\partial\theta/\partial t = \partial/\partial z\ [K(\theta)\ \partial h/\partial\theta\ \partial\theta/\partial z] \ldots\ldots [3]$$

Childs and Collis-George (1950) defined diffusivity, *D*, as:

$$D(\theta) = K(\theta)\ dh/d\theta \ldots\ldots [4]$$

Using [4], [3] takes the form:

$$\partial\theta/\partial t = \partial/\partial z\ [D(\theta)\ \partial\theta/\partial z] \ldots\ldots [5]$$

and [2]:

$$q = -D(\theta)\nabla\theta \ldots\ldots [6]$$

Both unsaturated conductivity and diffusivity functions decrease as water content decreases with the former showing a sharper reduction. This results from the fact that as water content is reduced, the larger pores empty first. According to the Hagen-Poiseuille law, small pores conduct water much less readily than large pores. In addition, the path of the flow becomes more tortuous as the soil desaturates. These factors have been successfully incorporated in models which give very good descriptions of unsaturated hydraulic conductivity and diffusivity as functions of water content (Brooks and Corey, 1966; Campbell, 1974) or as a function of soil water potential (Gardner and Mayhugh, 1958).

The relationship between matric potential and water content is described by the retention curve after equilibrium is established and the wetting front ceases its upward movement. For a matric potential h, where $h < h_e$, the air entry potential, the retention curve is described by the function (Campbell, 1975):

$$h = h_e(\theta/\theta_s)^{-b} \quad \text{[7]}$$

Here θ is the water content, θ_s is the saturated water content and b is the slope of the log(h) vs. $\log(\theta/\theta_s)$ plot. Hence, the height above the water table versus the corresponding moisture content data are used to find h_e and b values. By plotting moisture retention data on a log-log scale and fitting a straight line to the data, the values of h_e and b are found.

Most of the models characterizing the unsaturated hydraulic conductivity function use semi-empirical formulas. Campbell (1974) used the formula:

$$K = K_s(\theta/\theta_s)^{m} \quad \text{[8]}$$

and he suggested that $m = 2b+3$. This is identical to that of Brooks and Corey (1966). Equation [8] combined with [7] gives:

$$K = K_s(h_e/h)^{n} \quad \text{[9]}$$

where $n = 2+3/b$.

Values of diffusivity can be obtained through water characteristic data and known conductivity values:

$$D_{i+1/2} = K_{i+1/2} (h_{i-1} - h_i)/(\theta_{i-1} - \theta_i) \ldots\ldots\ldots\ldots [10]$$

Equation [10] is the difference form of equation [4]. Simple forms have been derived to express diffusivity as a function of water content. Gardner and Mayhugh (1958) used an expression for the diffusivity D of the form:

$$D = D_n \exp[b\ (\theta-\theta_n)] \ldots\ldots\ldots\ldots [11]$$

where D_n is the diffusivity at the lower boundary with a moisture content θ_n and b is an empirical constant.

Haverkamp, et. al. (1977) used analytical forms for unsaturated conductivity as a function of matric potential h given as:

$$K = A\ K_s/(A + h^B) \ldots\ldots\ldots\ldots [12]$$

and water content:

$$\theta = a(\theta_s - \theta_r)/(a + h^b) + \theta_r \ldots\ldots\ldots\ldots [13]$$

where the parameters A, B, a, and b are obtained through a non-linear parameter estimation. Experimental methods for measuring the functions are also available. The disadvantages of the experimental methods are: a) The functions are soil specific and their measurements are difficult and time consuming and b) different methods yield different results. Thus it is unclear which method best represents the hydraulic characteristics of the field (Ragab, et. al., 1981).

Bruce and Klute (1956) first solved for D with Equation [5], introduced by Childs and Collis-George (1948). Using Boltzmann's transformation (Boltzmann, 1894), they let θ be given by:

$$\theta = f(\lambda) \qquad [14]$$

where λ is a function of distance x and time t defined by:

$$\lambda = xt^{-1/2} \qquad [15]$$

with initial and boundary conditions for the moisture flow in the horizontal tube:

$$(1)\quad \theta = \theta_i \quad \text{for } x \rightarrow \infty \qquad [15a]$$

$$(2)\quad \theta = \theta_s \quad \text{for } x = 0 \qquad [15b]$$

the solution of [5] with respect to D is given by:

$$D_{(\theta)} = 1/(d\theta/d\lambda)_\theta \; (-1/2) \int_{\theta_i}^{\theta x} \lambda d\theta \qquad [16]$$

Methods for determining D(θ) based on the nonsteady-state and the Boltzmann transform are described by Bruce and Klute (1956). The water content gradient, determined by gravimetric sampling at a fixed time in a horizontal infiltration flow system, was used to calculate the diffusivity function. Whisler, et. al. (1968) described a method in which the water content was measured as a function of time at a fixed position. This procedure required a nondestructive method of determining the water content in the soil column, such as gamma attenuation. The sorptivity method (Dirksen, 1975) entails the determination of the soil water diffusivity function D(θ) by means of a series of one-dimensional adsorption experiments. All the above described methods determine diffusivity for wetting soils and are applicable to capillary rise and other adsorption experiments. Gardner (1956) devised a method for computing the diffusivity D(θ) and conductivity K(θ) functions using water outflow from soil placed in a pressure plate apparatus applicable only to desorption experiments.

2. MATERIALS AND METHODS

2.1 MATERIALS

The Metea sandy loam soil (Arenic Hapludalts; sandy over loamy, mixed, mesic[1]) was used in this study. The A-horizon of the Metea soil is a dark, sandy loam, approximately 10 cm thick. It has a very rapid permeability in the upper profile and moderate in the lower profile. The water holding capacity is described as moderate.

Disturbed samples of soils were taken between 0 and 10 cm depth from Michigan State University Soils Research Farm in East Lansing, located in the north central portion of Ingham County between 42 and 43 degree latitude and 84 and 85 degree longitude. The particle density of the soil is 2.624 Mg m^{-3}. To minimize the problems during the packing procedure, a soil was selected with low clay content. The cumulative particle size distribution is shown in Figure 1.

2.2 PREPARATION OF THE FLOW SYSTEM

2.2.1 SAMPLE PREPARATION

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[1] Soil Survey of Ingham County, Michigan, United States Department of Agriculture and Soil Conservation Service in cooperation with Michigan Agricultural Experimental Station. 1977.

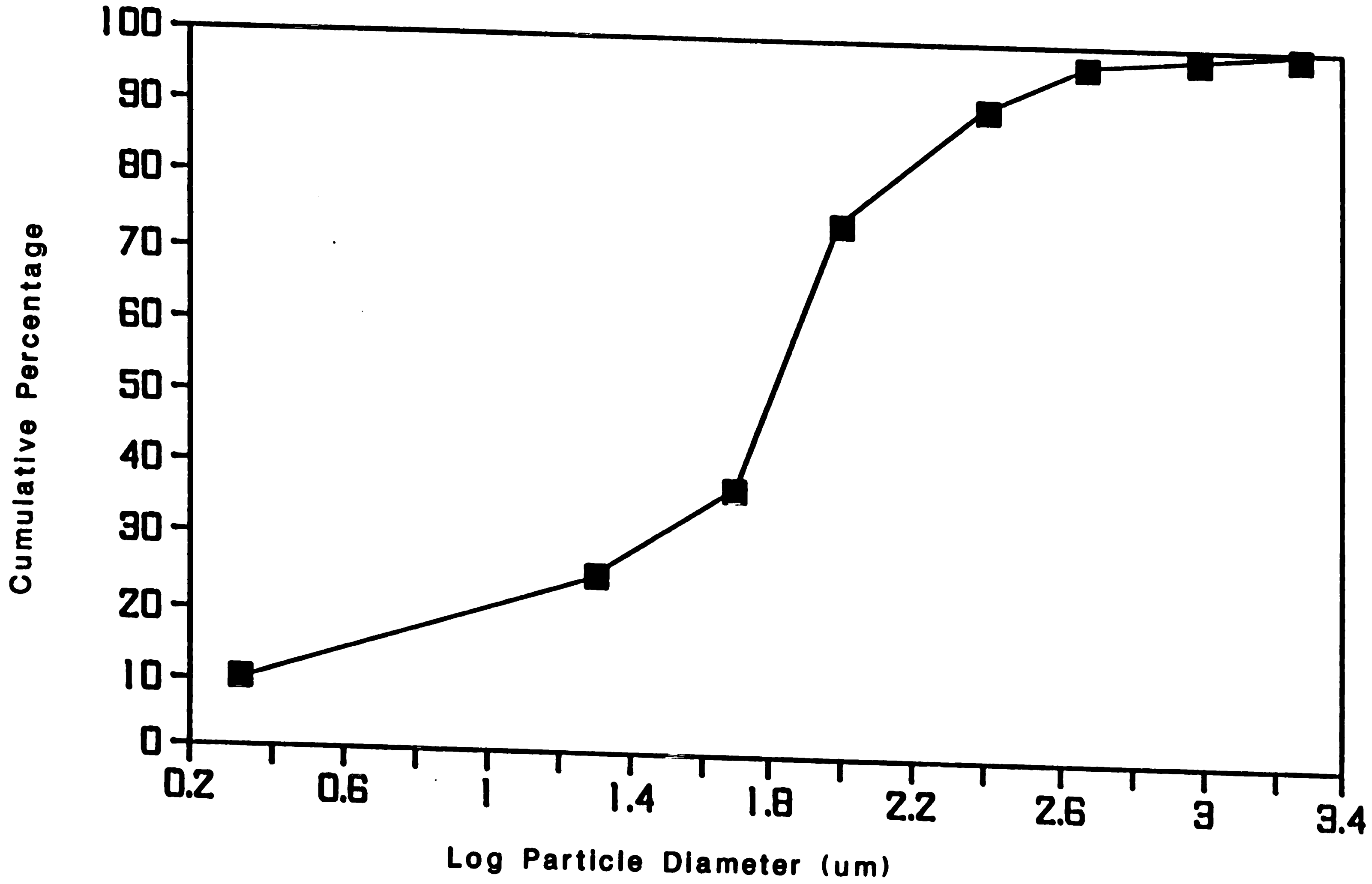

Fig. 1. Particle size distribution curve for Metea sandy loam soil

Soil samples were evenly spread over laboratory benches to obtain air-dryness and later screened through a 2-mm sieve followed by a screening through a 1-mm sieve. The screening was done to help obtain the same average bulk density between soil columns and a uniform bulk density distribution within each column.

2.2.2 PACKING OF THE SOIL COLUMNS

Fifty or more two-cm cylindrical glass sections were fastened to each other by tape to form a long cylinder with an additional 10 1-cm cylindrical sections fastened to one end. The sections were numbered and arranged in sequential order forming a cylinder one meter or more in length. The column was packed in a vertical position and closed from the bottom with a rubber stopper. To overcome the difficulty of uniform packing, several techniques were tested, all relating to the addition of soil to the column.

A special packing device, designed by Dr. A. T. Corey of Colorado State University and modified by H. M. Kar-Kuri (1983), was first tested without satisfactory results. A major problem was the continuous clogging of the screen and the unsteady flux of soil into the column. This prompted a modification of the device. A sketch of the device is shown in Figure 2. The upper part of the receptacle consisted of a storage vessel with a moveable door to control the flux of the soil. The 45 degree angle helped to keep the flux constant and to prevent clogging of the screen. The 2-mm diameter screen in the lower part of the device helped in the random distribution of soil particles by scattering particles after striking the screen.

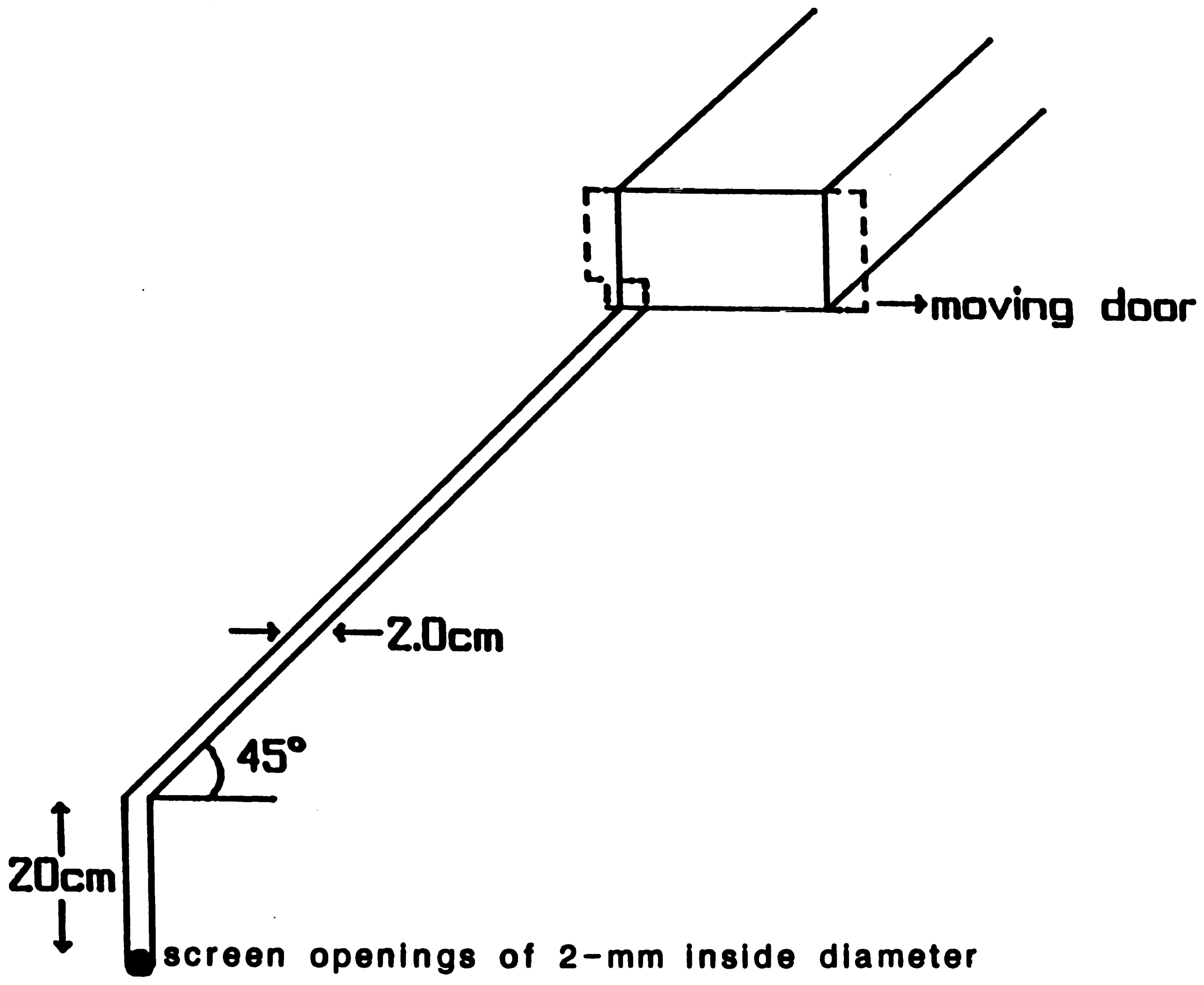

Fig. 2. Schematic diagram of the soil column packer.

Before packing began, the screen was positioned approximately 20 cm above the rubber stopper at the end of the empty cylinder. This position was maintained constant by moving the cylinder with an elevating jack throughout the filling process.

The purpose of this extensive effort in filling the tube was to attain uniform bulk density within and between different columns. Through the height of fall and the striking of the screen by soil particles, the packing device maintained a continuous and uniform flow of soil from the source into the column and thereby produced a homogeneous distribution of particles across the entire soil column surface area.

Once the column was filled, the desired bulk density was obtained by dropping the column on the stoppered end from a height of 10 cm, 50 to 150 times, depending on the desired bulk density. The greater number of drops produced a higher bulk density. As the soil consolidated, additional soil was added. Because the variability of bulk density was larger at the ends of the column, a 20-cm section was removed from each end after the consolidation procedure was complete.

The observation of identical wetting curves is the strongest evidence of similar bulk density distibutions and bulk density averages - similar soils generate similar wetting curves. Experiments based on similar wetting curves were conducted to test the hypothesis that soil columns packed in the same manner will exhibit similar bulk density distributions and bulk density averages. If the hypothesis is accepted, the two columns exhibiting the same water characteristic curves are considered a pair of replicates. In these experiments the desired average bulk density was achieved by controlling the energy applied to the system through the number of drops. In a prelimenary experiment when using 48 rings and striving for an average bulk density of 1.36 Mg m^{-3}, the standard deviation was ±0.02 within the rings of column. An F-test at the 0.01 level showed no significant statistical difference. Out of fifteen capillary rise experiments which were run in the laboratory the following three pair of replicates shown in Table 1 below were arbitrarily selected for this study and are identified as experiments 4, 5, 6, 7, 12 and 13 in the Appendix.

TABLE 1. Characterization of soil columns used for capillary rise experiments.

Soil	Replicate	No. of drops	No. of rings	Average Bulk density Mg m^{-3}	Standard deviation
Soil A*	1	100	48	1.331	±0.039
	2	100	48	1.332	±0.045
Soil B*	1	150	48	1.432	±0.020
	2	150	48	1.406	±0.024
Soil C**	1	50	48	1.354***	±0.055
	2	50	48	1.370***	±0.057

* Soils in the following text are going to be referred to by their names as defined above

** Soil which had gone through one cycle of wetting and drying in the cylinder

*** The bulk density was calculated after the wetting and drying cycle

Soil columns labeled C had gone through one previous cycle of wetting and drying in cylinders before the capillary rise experiment to test the hypothesis that reorientention of soil particles may change the bulk density and the flow dynamics. Discontinuities in the soil matrix can be minimized following the first wetting and drying cycle because particles are more firmly attached to neighboring particles.

2.2.3 THE FLOW SYSTEM

The apparatus used for the capillary rise experiments is shown in Figure 3. The water supply system contained a constant-head burette of 250-ml capacity and divisions of 1 ml facilitating measurement of water uptake by the soil. The soil column consisted of 2-cm and 1-cm sections of clear glass tubing 3.60 cm in diameter taped together to form a 1-meter long cylinder. Alignment of the column was accomplished with a trapezoidal-shaped notch cut into a block of wood. A meter stick was taped to the column to measure wetting front positions. A ceramic plate with relatively large pores (air entry value less than 5 cm of suction) and low resistance to the water flux was attached to the bottom part of the column and positioned to contact a free water surface exactly at the interface between the plate and the soil. Hence, the ceramic plate was aligned at $x=0$ and zero pressure was applied. The upper part of the packed column was plugged with a small amount of glass wool and a rubber stopper with a hole in it, allowing air to escape as the water front advanced through the column. When the water front stopped moving, equilibrium was assumed. Then, the negative matric potential at any point in the wetted column was assumed to be equal to the distance of that water above the water table.

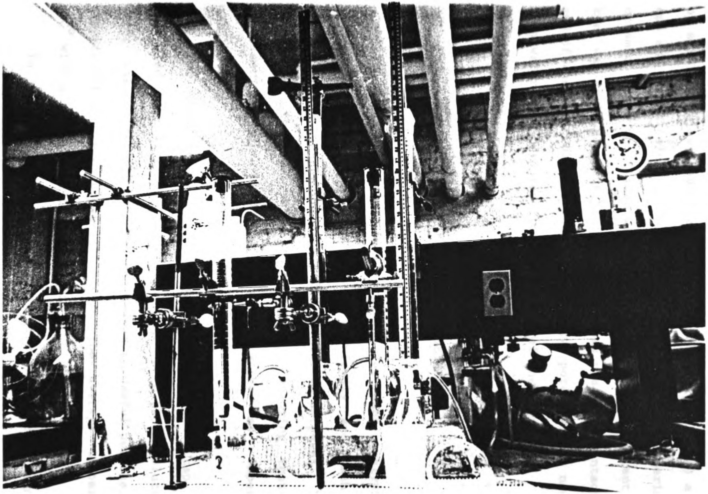

Fig. 3. Experimental apparatus used for capillary rise experiments.

Horizontal infiltration experiments were also run to determine the diffusivity function, applicable to capillary rise. During these experiments the potential was maintained at -0.14 kpa by a fritted glass head plate (Nielsen and Philips, 1958) attached to one end of the column and placed at x=0. To start the experiment, the ceramic plate was filled with water and the desired potential (-0.14 kpa) was applied to the plate before it was placed in contact with the end of the soil column. The small negative potential was necessary to keep water from dripping from the bottom of the plate.

2.3 EXPERIMENTAL METHODS

Two soil columns were prepared for each analysis. Fifteen columns were prepared as outlined previously and each pair of replicates was run simultaneously for comparison purposes. Thus, both were under the same temperature regime, ensuring that the two experimentally generated profiles were as alike as experimentally possible.

For this study the wetting agents used were tap water and distilled water with $CaSO_4$ added to make a 0.01 N solution. The water was not deaired because this served no useful purpose. Water temperature was recorded at all times. The average temperature was $22C^o$ with a standard deviation of $\pm 2C^o$.

In both horizontal and capillary rise experiments, water entering the columns was measured volumetrically in the constant head burette for a given time and distance to the wetted front was measured at the same time with the meter stick. When flow had proceeded for the desired time or distance (in capillary rise until equilibrium was established) the water supply was cut off and the column quickly segmented into its 2- and 1-cm sections. The water content of each section was determined gravimetrically (θ_g) and converted to volumetric values (θ_v) by using the bulk density of the corresponding section.

Measurement of wetting front distances versus the square root of time and soil moisture distribution profiles were made for both experiments. Soil water diffusivity values (defined in equation [16]) versus water content were obtained by the method of Bruce and Klute (1956). Soil water potential values needed for generating capillary conductivity functions were obtained from the capillary rise experiments. Experimental desorption characteristic curves were obtained using pressure plate and pressure cooker equipment following the standard procedure.

An ATT 6300 personal computer with an 8087 math co-processor and a 20-megabyte hard disk was used for data analysis. All the programming was done using Microsoft FORTRAN 77, version 3.2, 1984, and the FORTRAN IMSL math library.

3. RESULTS AND DISCUSSION

Six of the fifteen experiments, as indicated in Table 1, were selected for discussion. Equation [7] was used to fit the adsorption characteristic data for each experiment. Figures 4 and 5 show the experimental wetting curves for two replications of two soils with different bulk densities. One can see that: 1) the upper part is parabolic and flattens out at the end; 2) the wetting front stopped rising at a height of about 60 cm even though experiments continued for more than 7000 minutes; and 3) the bulk densities of the two replications are in good agreement as could be inferred from the similarity of the two curves and agrees well with data shown in Table 1. This supports the hypothesis that columns packed in the same manner will exhibit similar wetting curves. By fitting the parabolic part of the curves or segments of it on a log-log scale to a straight line, the values for the air entry pressure h_e and the slope b were found. Using these values, smoothed analytical wetting curves were generated and compared with experimental data. Figure 6 shows this fitting procedure for soil B.

As the literature suggests (Campbell, 1985), if Figure 6 had been generated from desorption data, it would have been one slope with the matric potential decreasing monotonically by the same factor (slope) as the water content decreased. In capillary rise three different slopes could be extracted from each curve, which means matric potential decreased monotonically but by different

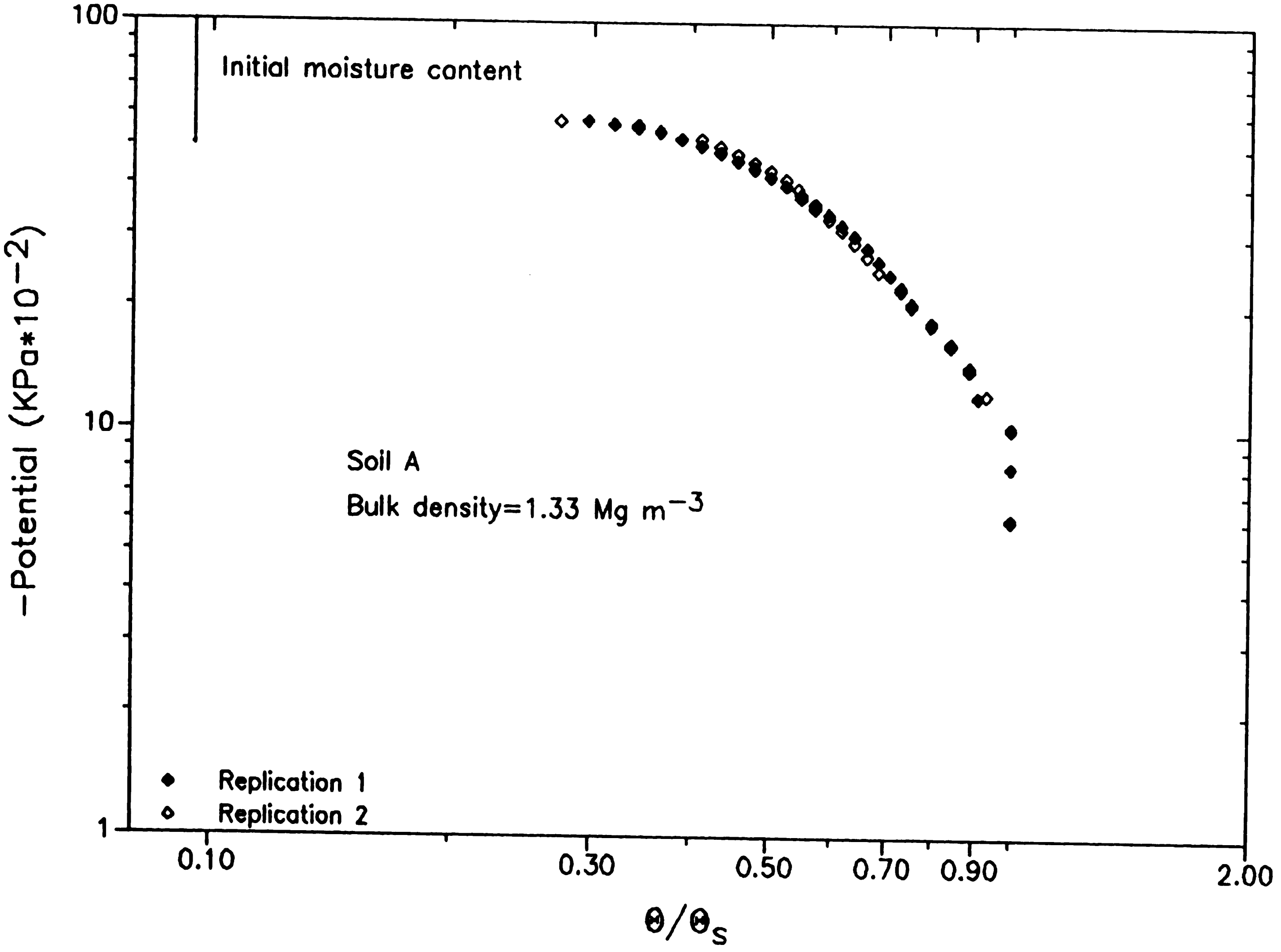

Fig. 4. Experimental water adsorption characteristic curve of soil A.

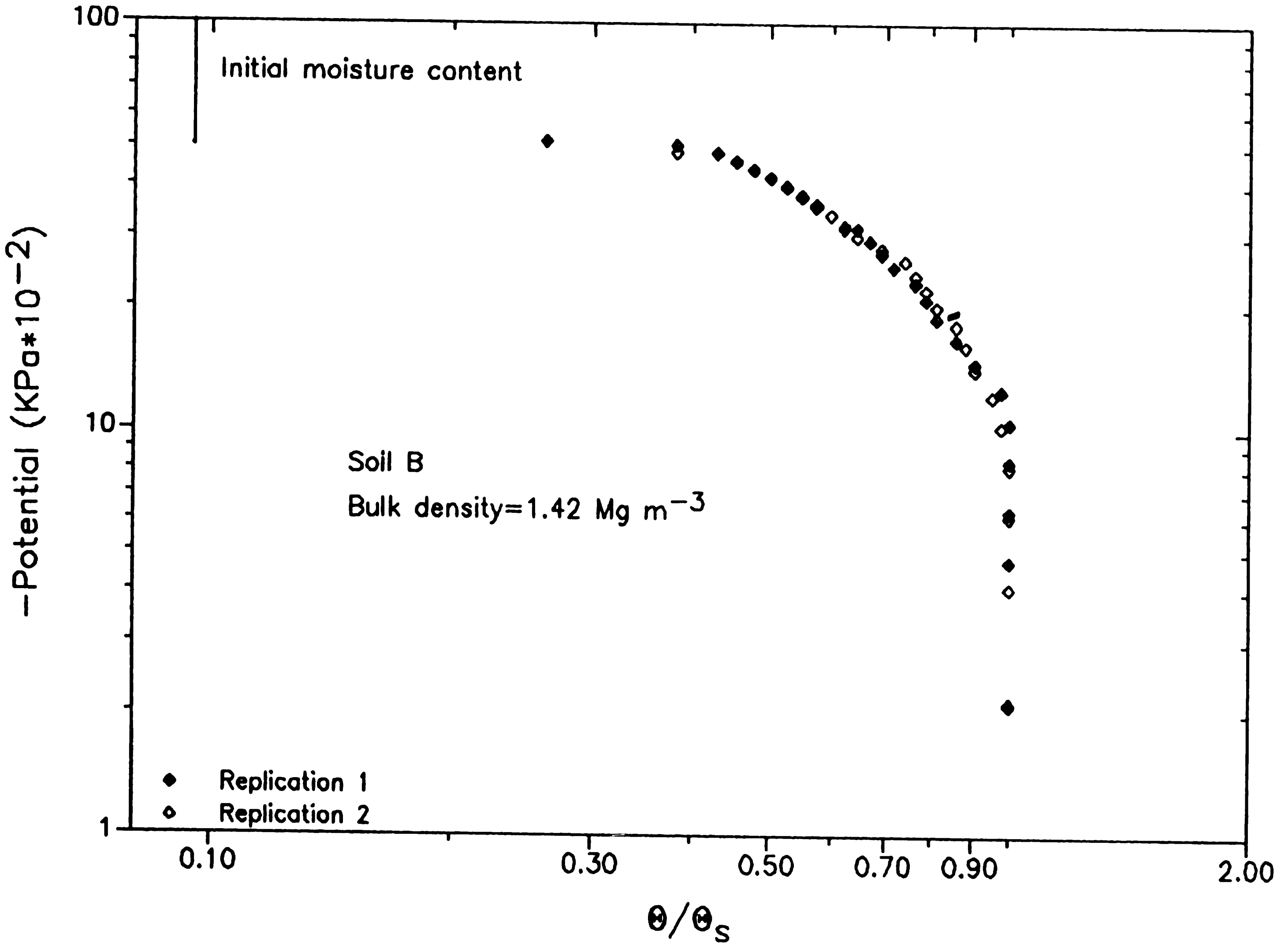

Fig. 5. Experimental water adsorption characteristic curve of soil B.

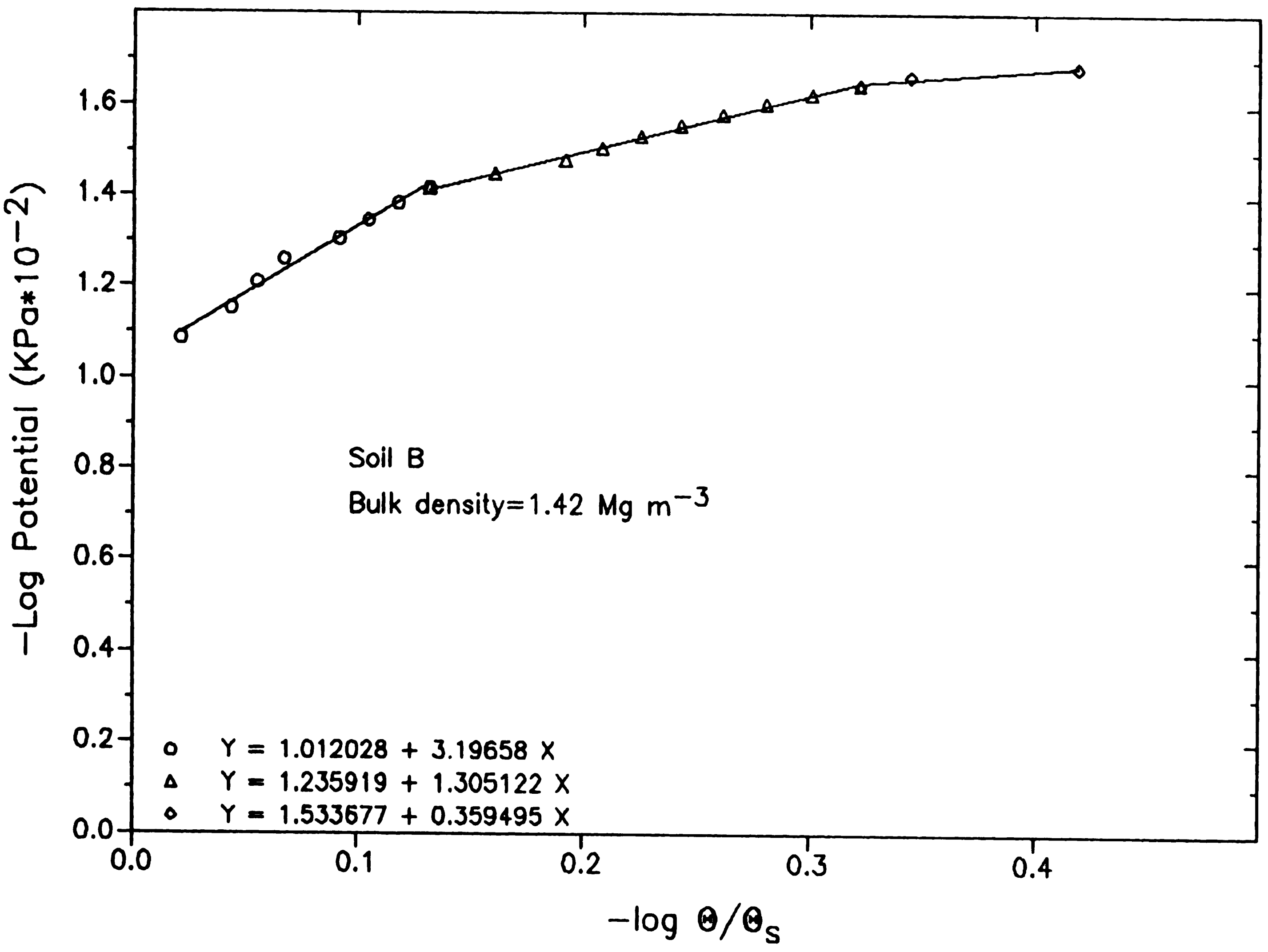

Fig. 6. Fitting capillary rise data on a log−log scale.

scales (slopes) as the water content decreased. From looking at Figures 4 and 5 one can say that during capillary rise two obvious changes in the slope of pF curves are taking place: 1) in the theta-straight portion there is a change in potential but no change in water content; 2) in the parabolic portion there is both a change in potential and water content. By plotting results from Figure 4 on a larger scale in Figure 6, three transitional stages of wetting were detected in the parabolic portion and Equation [7] was applied to each of the three different regions. These transitional stages are associated with the number and size of soil pores filled and with the thickness of water films on soil particles. When the films become critically small, the wetting process stops because contact is not made with adjacent soil particles.

Table 2 shows the values for air entry pressure, h_e, and slopes b_i, obtained using Equation [7] in three different zones. The minimum, maximum, mean and standard deviation values of h_e and b_i for the first nine experiments were calculated to delineate bounds due to sample variation and experimental error. The other six experiments were omitted because additional treatments were applied to the soil columns. If two or more profiles were to be compared, the agreement of the parameters for a given moisture range was considered acceptable when the values were within two standard deviations. When the parameters were outside of this bounding range, a significant difference in profile shape was detected. Because soil column densities were varied purposely, the standard deviations were larger than they would be if densities were uniform.

Figures 7 and 8 show experimental data when compared with 1)

TABLE 2. Values for air entry pressure h_e and slopes b_i calculated with Equation [7] for three different zones.

Experiment #	Average Bulk density $Mg\ m^{-3}$	Log_{10} h_e cm	b_1	b_2	b_3
1	1.3860	1.2669	1.7147	0.6380	0.4205
2	1.3400	1.0446	2.0293	0.8521	0.4677
3	1.3450	1.0638	1.8875	0.9501	0.2035
4*	1.4320	1.0120	3.1965	1.3051	0.3595
5	1.4060	1.0842	2.1471	1.0905	0.1101
6	1.3320	1.0444	2.0874	0.4520	0.1669
7	1.3310	1.0293	2.2571	0.9496	0.3459
8**	1.1600	0.9725	1.4760	1.1361	0.3569
9	1.1500	1.0101	1.6967	1.2438	0.2437
10	1.2000	0.4855	2.4446	1.2446	0.6352
11	1.2300	0.4422	2.4836	1.3244	0.5892
12***	1.3700	0.7828	1.8802	0.6395	0.3021
13	1.3540	0.7738	1.8888	0.5132	0.4034
14	1.3300	0.6276	2.2208	1.1260	0.5916
15	1.3400	0.7441	2.1128	1.5021	0.4008
Minimum****	1.1500	0.9725	1.4760	0.4520	0.1101
Maximum	1.4300	1.2669	3.1965	1.3051	0.4677
Mean	1.3166	1.0586	2.0547	0.9574	0.2971
Std. Dev.	±0.0849	±0.0797	±0.4668	±0.2627	±0.1143

* Experiment numbers 4 and 5 are labeled soil B, 6 and 7 are labeled soil A and 12 and 13 are labeled soil C

** Experiments 8 and 9 were soil columns described in Chapter 3

*** Experiments 12, 13, 14 and 15 experienced one cycle of wetting and drying in the cylinder

**** The following data refer to experiments 1-9 only fitted data using Equation [7] for the entire curve and 2) fitted data using Equation [7] by segmenting the curve into three different slopes as suggested in Figure 6. The curves with the square symbols in Figures 7 and 8 are typical for fitting desorption data; hence Equation [7] when used only as a single slope curve is unable to fit experimental capillary rise data. These figures illustrate that with three slopes the agreement between fitted and experimental data is excellent.

It is obvious in the initial part of the adsorption wetting curves that all soil particles are saturated in the theta-straight region. Experimental data of Figures 4 and 5 show clearly the existence of this region. In the parabolic part where the curves flattened the wetting front stopped rising because of two possible circumstances 1) the net driving force (the difference between the matric and gravitational gradients) approached zero and/or 2) a discontinuity occurred in the soil matrix. Such discontinuities may be formed continuously during the wetting process just ahead of the wetting front. When films are thick, the wetting process may not be restricted. However, as films become thinner, contact cannot be made with adjacent dry particles and the wetting process may be hindered or stopped completely.

Philip (1957a), using Moore's (1939) data, chose a parabolic shaped water profile without any theta-straight part. Green and Ampt (1911) working with downward adsorption data suggested an orthogonal water profile. An orthogonal profile would be one where the theta-straight line is extended for some distance and then intercepted by a perpendicular line to form a rectangle. Figures 4 and 5 suggest that a

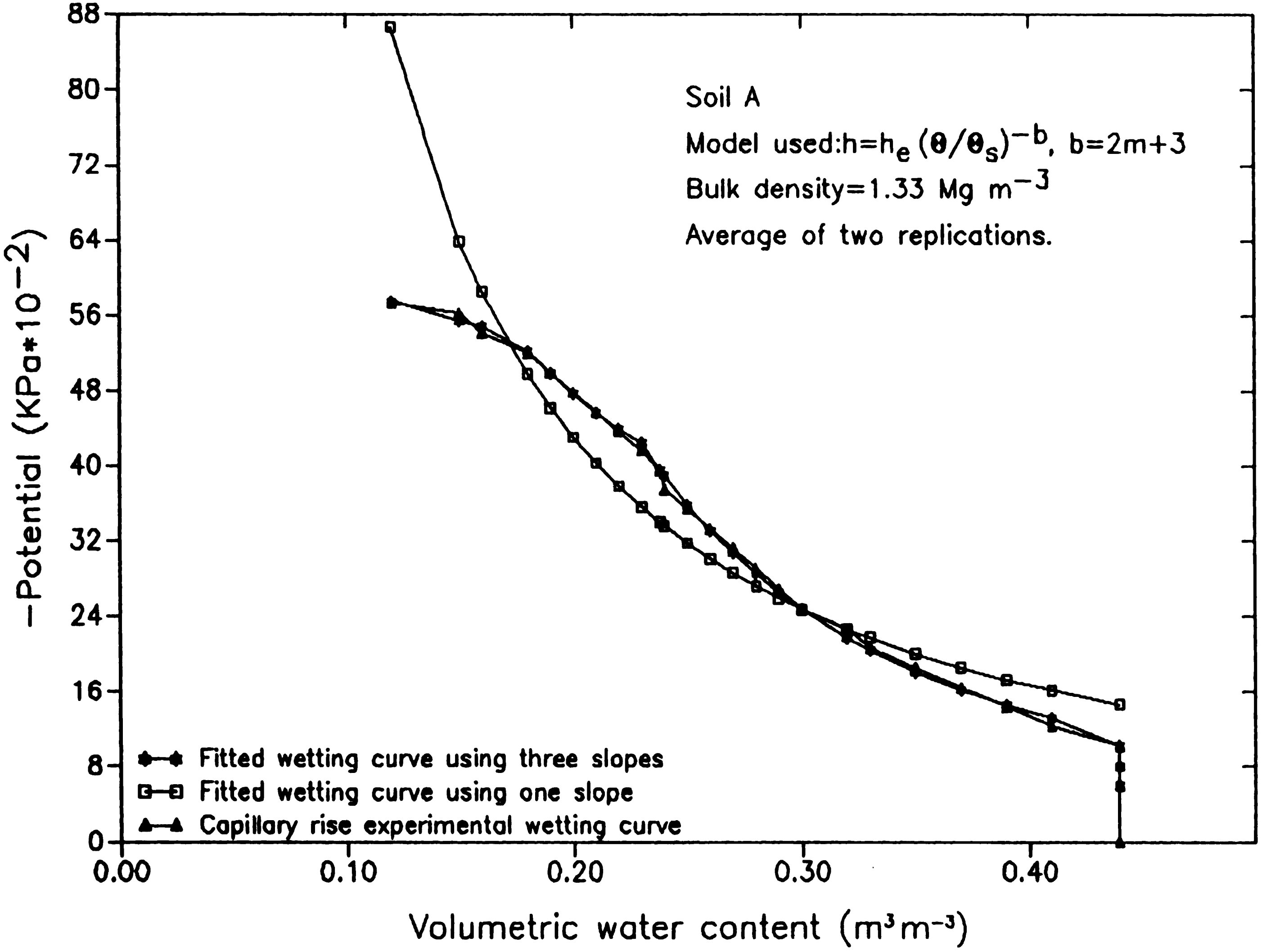

Fig. 7. Comparison between experimental and fitted wetting curves for soil A.

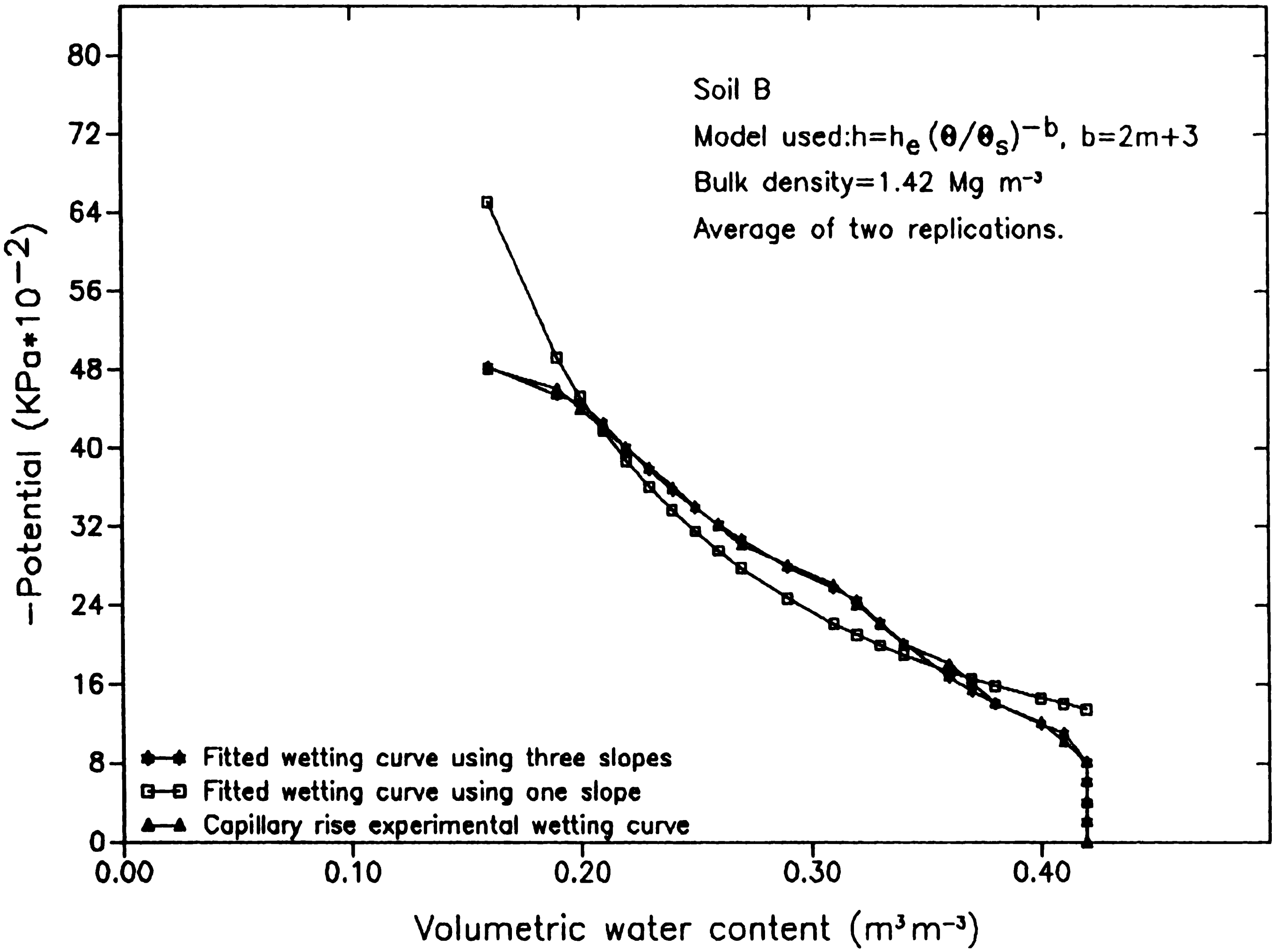

Fig. 8. Comparison between experimental and fitted wetting curves for soil B.

theta-straight region, followed by a parabolic curve flattened at the end, is the most realistic approach to fit capillary rise data.

Both standard wetting and drying procedures for calculating moisture characteristic curves may be done using pressure plate equipment. The adsorption wetting curves from capillary rise experiments do not appear to be compatible with the standard procedure of constructing such curves. Figure 9 shows the incompatibility between retention curves for Metea soil when obtained through desorption experiments and through capillary rise experiments. This incompatibility may be explained by a discontinuity taking place in the soil column during capillary rise.

Conductivity and diffusivity functions may be generated from the experimental wetting curves. For the conductivity function in this study Equation [8] was used where the parameters b_i were taken from Table 2. Campbell (1974) suggested first that a pore interaction term can be included to make $m = 2b+3$ instead of $2b+2$ as used by Brooks and Corey (1966). Also in this study $m = 2b+4$ was tested to account for more tortuosity. Equation [8] worked best for fitting capillary rise data with $m = 2b+3$. The diffusivity function was obtained using Equation [10] where the slope $dh/d\theta$ was obtained from the fitted wetting curves using three different slopes. Figures 10 and 11 show the diffusivity and conductivity functions for two replications of two different bulk density soils. Diffusivity functions, although not similar and smooth, are still sufficiently accurate to reproduce the experimental profiles in Figures 7 and 8. When smooth diffusivity functions with the one slope fitting procedure were used as an input to the model, there was a significant discrepancy

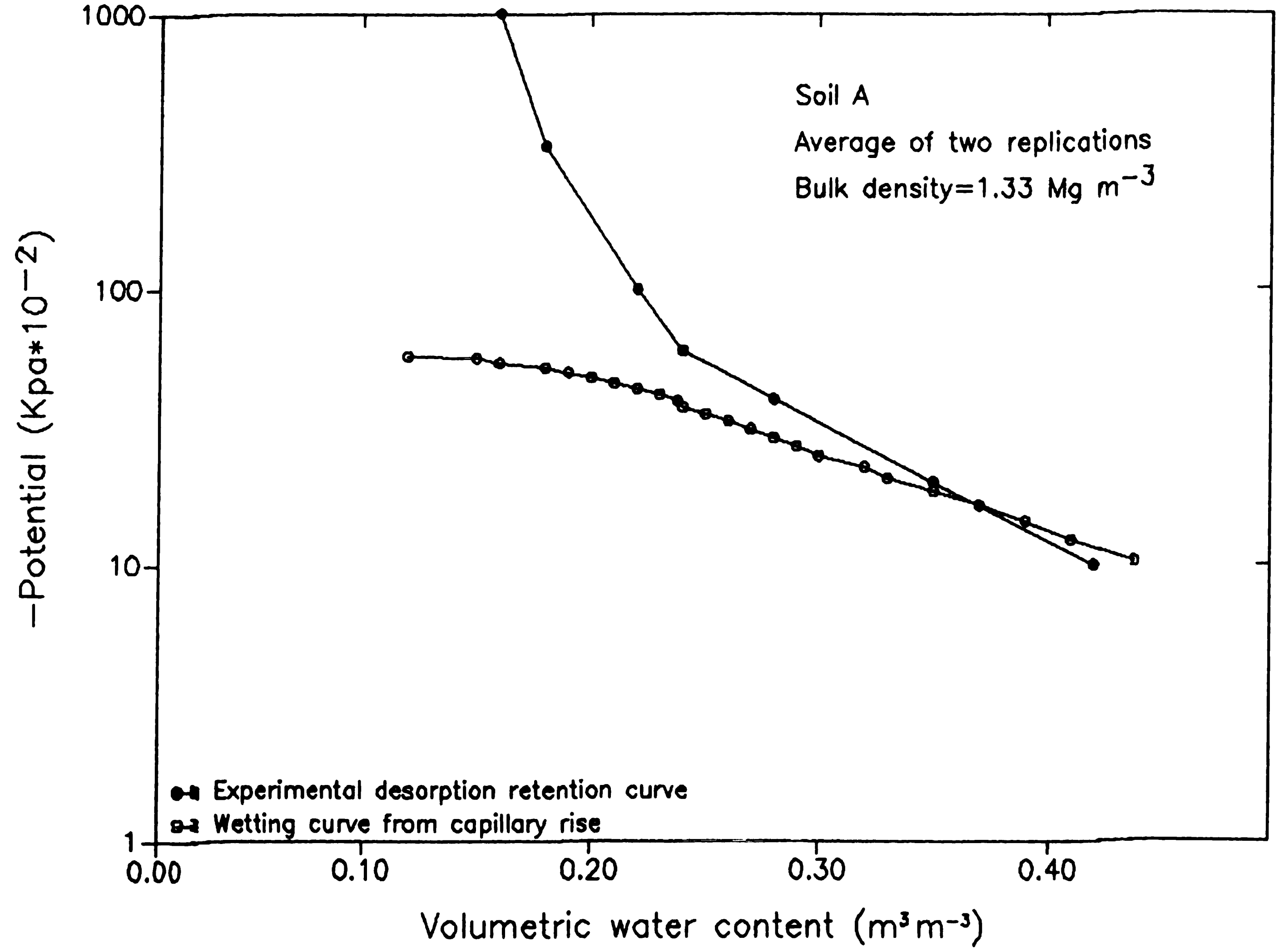

Fig. 9. Experimental retention curves obtained through capillary rise and desorption experiments for Metea soil.

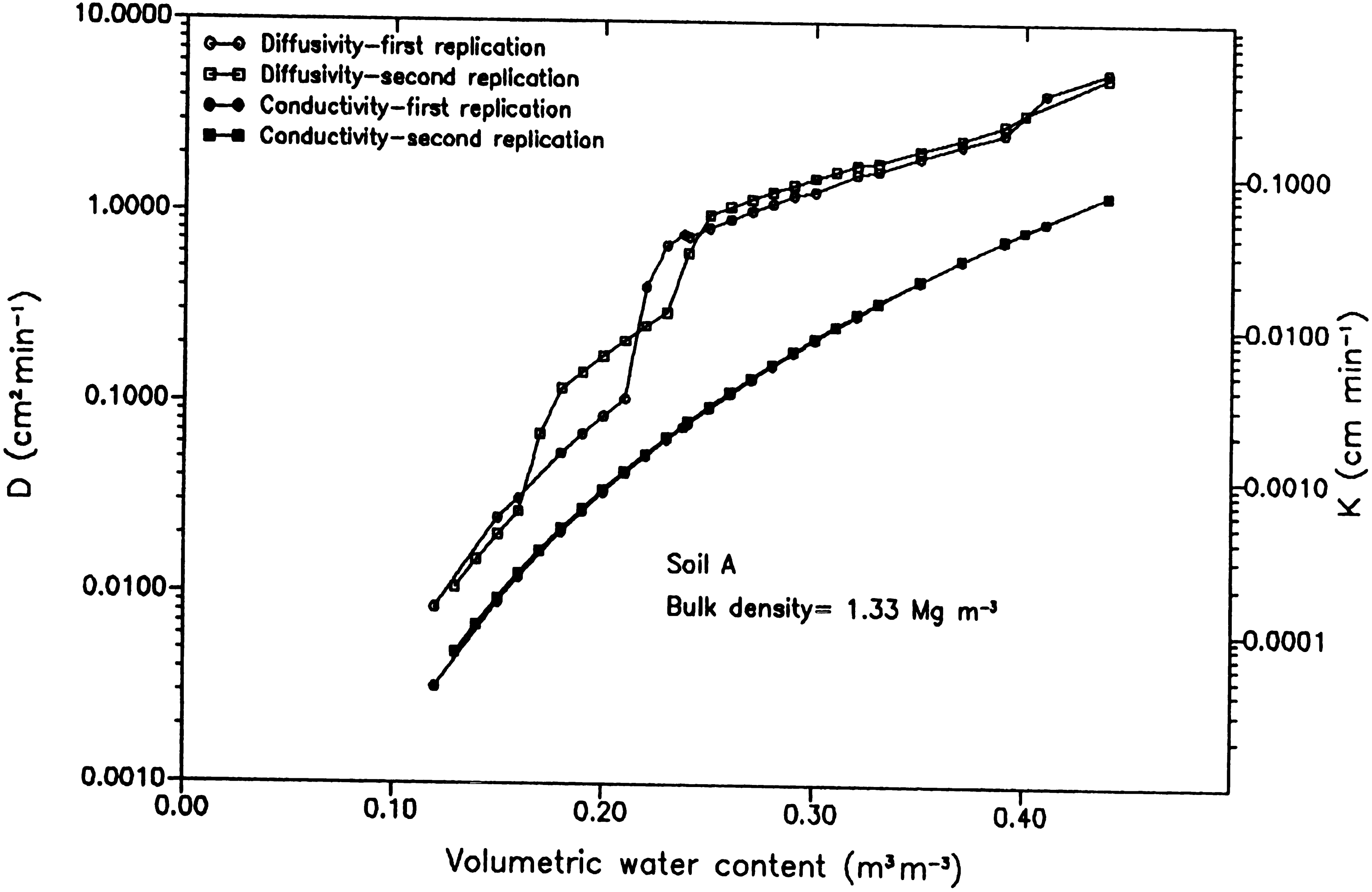

Fig. 10. Diffusivity and conductivity functions for two replications of Soil A.

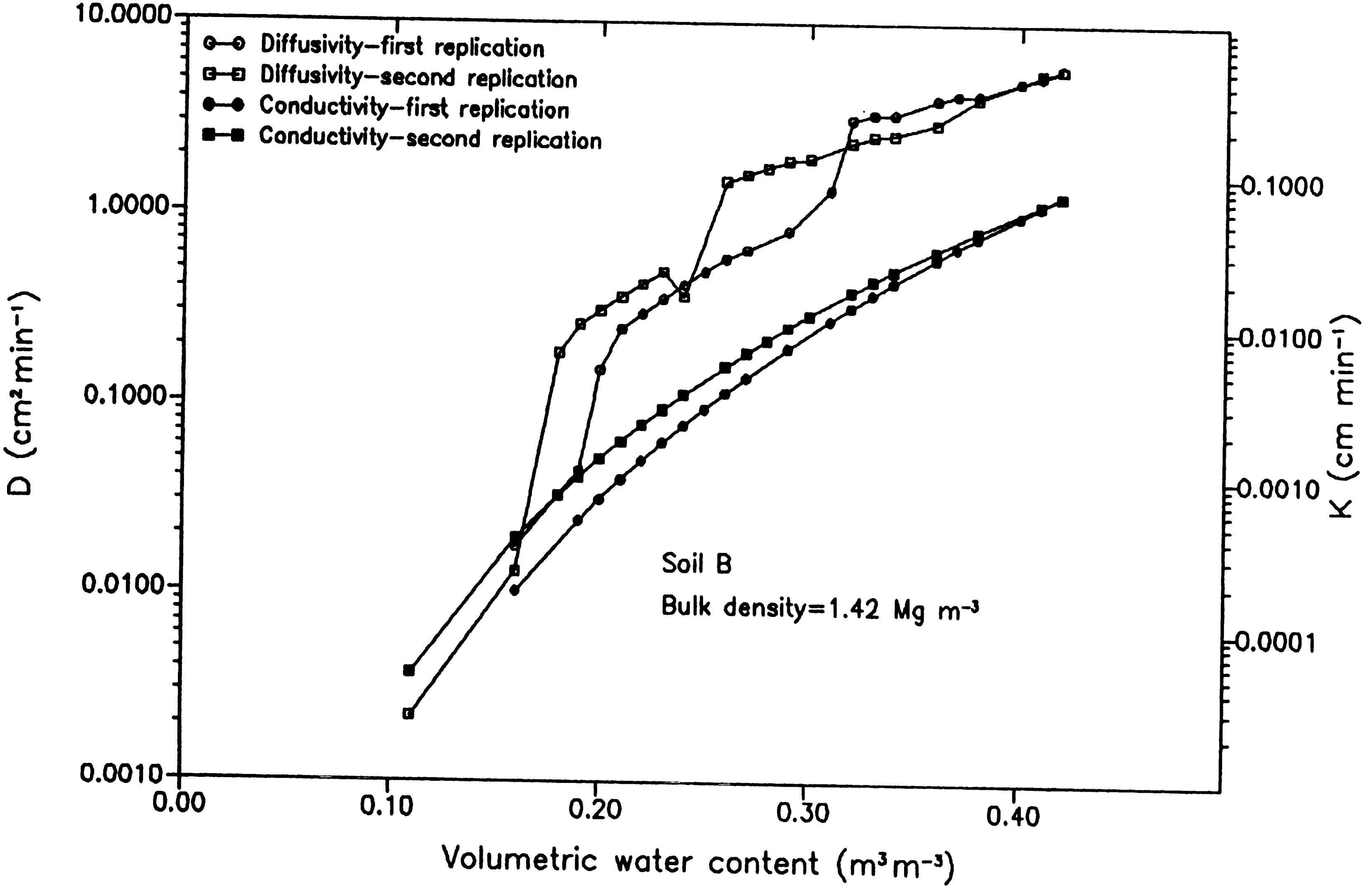

Fig. 11. Diffusivity and conductivity functions for two replications of Soil B.

between experimental and simulated profiles. Diffusivity data from capillary rise experiments did not exhibit an exponential increase with water content. Thus, Gardner and Mayhugh's (1958) semi-empirical model, Equation [11], could not be used successfully to fit capillary rise data, and it is not recommended for use in the capillary rise mode.

Haverkamp et. al. (1977) introduced analytical equations for conductivity and water content as described in Equations [12] and [13]. Through a non-linear parameter determination using Marquardt's least square technique, the parameters a and b for Equations [12] and [13] were determined and used as input in the finite-difference model of one-dimensional water movement in soils developed by D. L. Nofziger (1985). Typical results of this fitting procedure are shown in Figure 12. The goodness of fit differed from experiment to experiment, depending on the spread of the data. Because different equations (8 and 12) were used for calculating K in the FINDIT and finite-difference procedures, it was difficult to get the K functions to agree. This discrepancy between the input functions used by FINDIT and the finite difference-scheme is particularly noticeable in the low moisture range.

Figures 13 and 14 illustrate the experimental and simulated (FINDIT) wetting curves for two replications of two different bulk density soils. Experimental and simulated wetting characteristic curves compared very well because both curves are located within the bounding range. Hence, FINDIT appears to be very capable in modeling capillary rise in soils, but not without difficulty.

Figure 15 shows the experimental profiles at equilibrium for

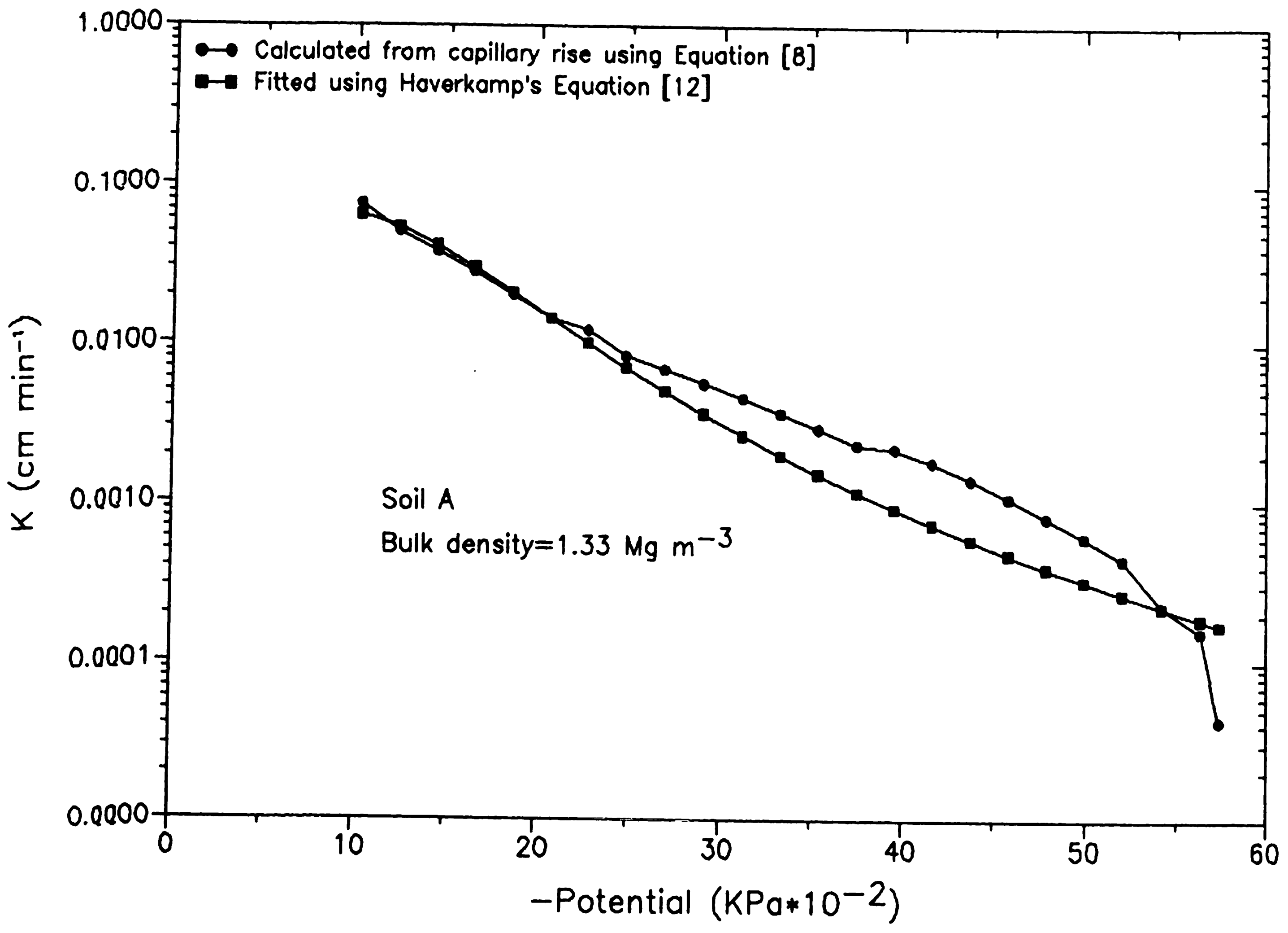

Fig. 12. Comparison of the conductivity function calculated from capillary rise using Equation [8] and Haverkamp's Equation [12].

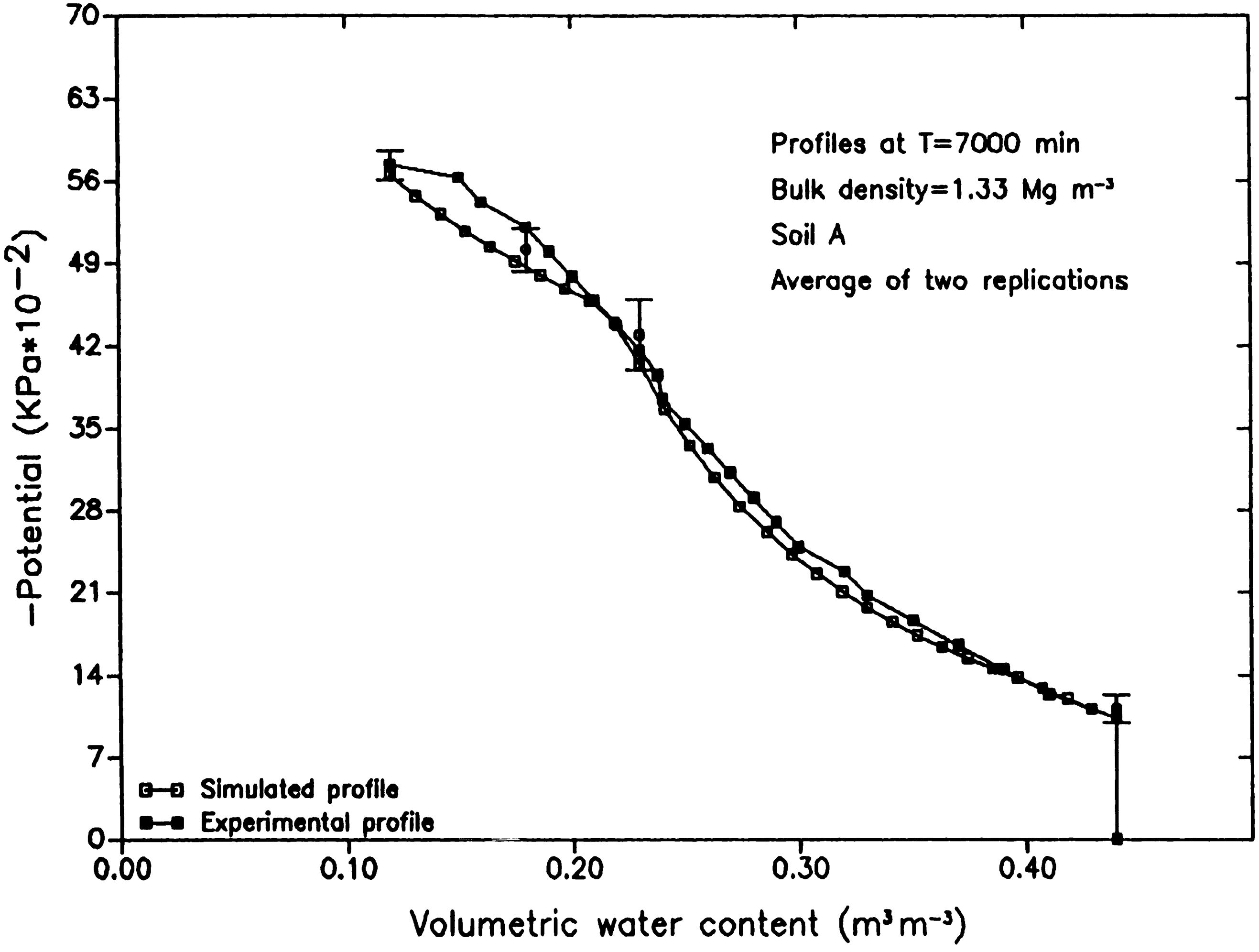

Fig. 13 . Simulated curves generated with FINDIT and experimental wetting curves for two replications of soil A.

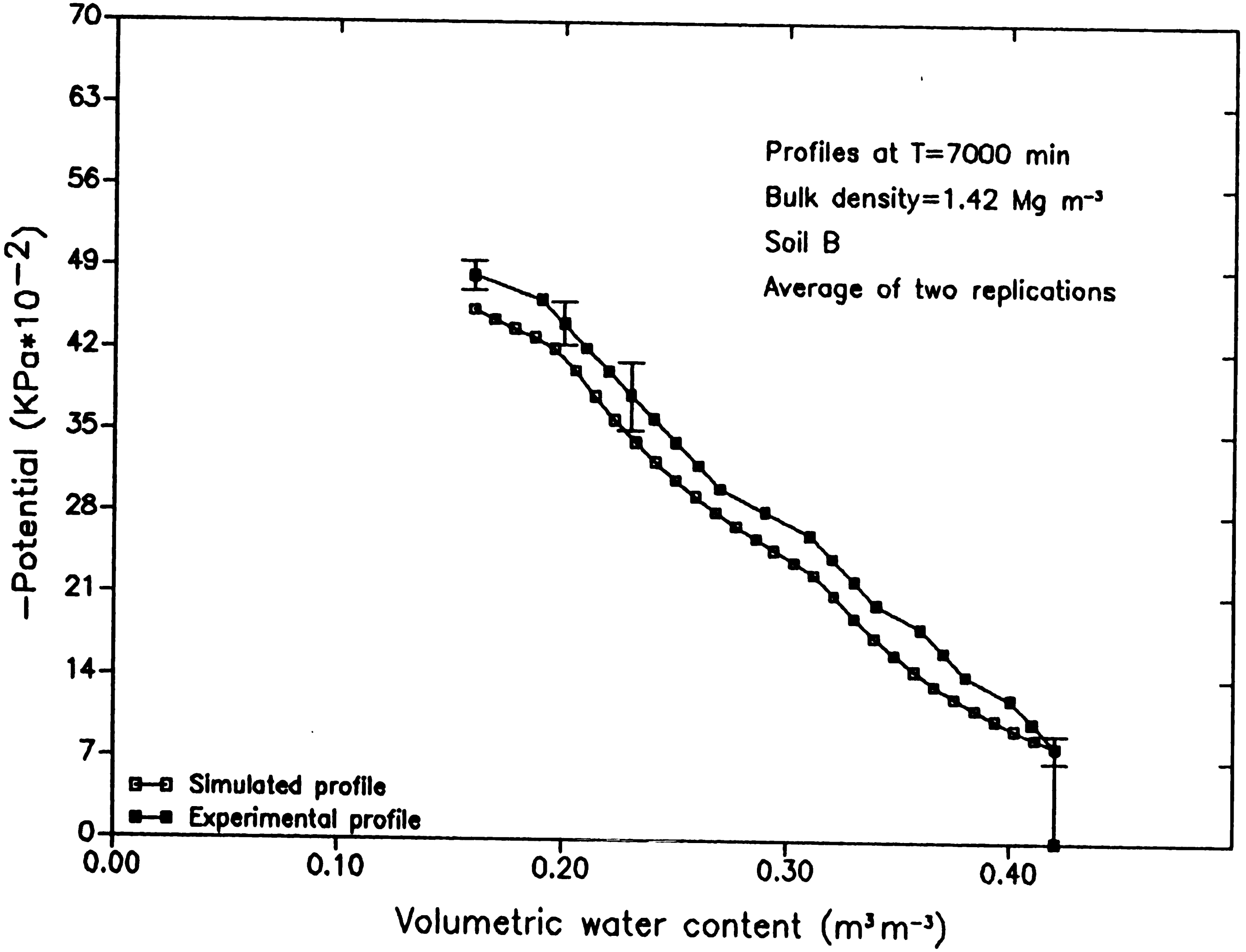

Fig. 14 . Simulated curves generated with FINDIT and experimental wetting curves for two replications of soil B.

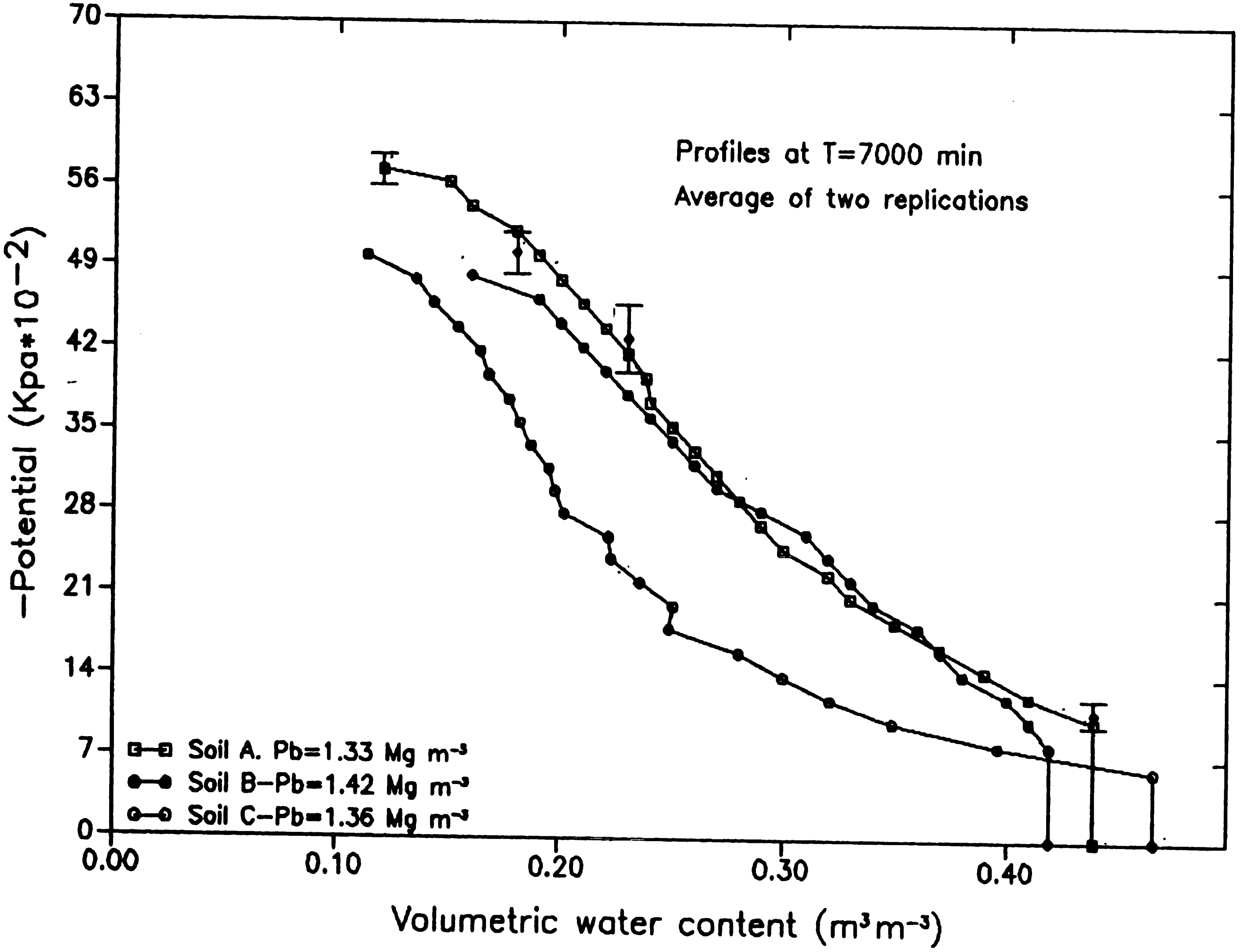

Fig. 15 . Experimental wetting curves for Metea soil for different bulk densities.

different bulk density soils. Soils A and B, although having similar wetting curves, were found to be different when bound analyses was used as criteria. Soil B was always outside the bounds at the four points of interest. Table 2 supports this hypothesis by indicating that soil B had the highest bulk density. Soil C which had gone through one cycle of wetting and drying held more water in the theta-straight region but held significant less water than soils A or B at all other potential values. When compared with the bound analyses,it was found to be a soil of a total different character. Assuming that the available energy (potential) is associated with the number and size of soil pores filled with water, reorientation of soil particles during the first wetting and drying cycle apparently increased the effective number of large pores and decreased the effective number of small pores.

4. SUMMARY

Capillary rise experiments were performed under nearly constant temperature in columns packed at different bulk densities with Metea sandy loam soil. Through these relatively simple experiments the wetting moisture characteristic curves were obtained. All curves exhibited a theta-straight region, followed by a parabolic curve flattened at higher suctions. The parabolic curve when plotted on log-log scale could be separated into three different zones in all 15 samples tested. The different zones suggest transitions in the water content potential relationships. These transition stages can be associated with relative percentages of filled pores and thickness of water films. At equilibrium, the matric and gravitational forces in the wetted soils were equal but of opposite sign. Contrary to desorption data, a power function had to be used in each of the three zones described above to successfully fit the experimental adsorption data. Conductivity and diffusivity functions were determined from experimental adsorption data using Campbell's, Haverkamp's and Gardner's formulas. Simulated (FINDIT) and experimental capillary profiles were compared and showed good agreement.

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CHAPTER II

SIMULATING UPWARD WATER MOVEMENT INTO METEA SOIL

1. INTRODUCTION

Buckingham's (1907) "capillary potential and capillary conductivity" represented two basic steps in the development of modern ideas of soil water movement. The concept of soil water movement as a diffusion phenomenon was explicitly proposed by Childs (1936a) who studied the hypothesis of constant diffusivity. Since the diffusivity varies greatly with moisture content, it is not surprising that results do not accord well with observation.

In 1950, Childs and Collis-George developed concentration-dependent diffusivity. This work implied the existence of a general partial differential equation describing liquid phase water movement in a porous medium which would prove to be the parabolic second order diffusion type. In this case the pressure potential is a single-valued function of the moisture content.

Klute (1952), following Richards' (1931) approach, applied the continuity requirement to the vector form of Darcy's law for a homogeneous medium and derived the equation which is the basis for the mathematical study of moisture flow in unsaturated porous media. This equation, as derived by Klute (1952), expressed in volumetric terms with the liquid density assumed constant can be written as:

$$\partial\theta/\partial t = \nabla(K\nabla h) \quad [17]$$

where ∇ is the vector differential operator. When the total potential is separated into gravitational and matric components, as is assumed throughout the following analysis, Equation [17] may be written:

$$\partial\theta/\partial t = \nabla(K\nabla h) + \partial K/\partial z \quad [18]$$

When diffusivity is introduced into Equation [18], the following equation is derived:

$$\partial\theta/\partial t = \nabla(D\nabla\theta) + \partial K/\partial z \quad [19]$$

Kirkham and Feng (1949) found that the diffusion equation is not an acceptable mathematical model for the movement of water in unsaturated soil when diffusivity is constant.

Gardner and Mayhugh (1958) reported good agreement between experimental data obtained from solutions of Equation [19] when diffusivity was considered a function of moisture content. The authors' success depended largely upon the Boltzmann transformation which converts Equation [18] into an ordinary differential equation and gives transformed boundary conditions which are as simple as the non-transformed boundary conditions. However, the Boltzmann transformation is limited to problems of horizontal flow with a uniform initial moisture content and a semi-infinite uniform medium. Problems still remain with finite media and arbitrary initial moisture concentrations that cannot be handled using the Boltzmann transformation. These problems have not been treated very extensively in the literature mainly because of the difficulties encountered in obtaining analytical solutions for non-linear differential equations. Numerical approaches seem to provide the best method of attack although some analytical and analogue techniques have been used.

As Philip (1967) stated one can use three techniques for solving non-linear equations like Equation [19]: 1) A computer solution using numerical methods, 2) Quasi-analytical and analytical solutions using methods of mathematical analysis to establish the basic form of the solution, and 3) Analytically based estimations of integral properties of the solution. Quasi-analytical solutions of the non-linear Fokker-Plank equation in one dimensional form, subject to restrictive and simple initial and boundary conditions, were developed by Philip (1957a) and more recently by Parlange (1971). The predictions appear generally to be in agreement with short time laboratory observations for situations where the air phase is free to escape in advance of the wetting front. Long time predictions are not yet possible without changing equations and making modifications.

Analytical solutions for non-linear equations have some assumptions which do not always hold under real conditions. Also field description of transient isothermal flow of water into nonswelling unsaturated soil is highly complicated since the initial and boundary conditions are usually not constant while the soil characteristics may vary with time and space. In view of this, most efforts have in recent years been concentrated on seeking numerical solutions. Freeze (1969) presented a list of papers with finite difference solutions for one-dimensional infiltration equations with their initial and boundary conditions. Each of these solutions employs different forms of the non-linear Fokker-Plank equation and different ways of discretization.

Using the specific water capacity C(h), Equation [18] can be transformed into:

$$C(h)\ \partial h/\partial t = \partial/\partial z[K(h)(\partial h/\partial z - \cos(a))] \ldots\ldots\ldots\ldots\ldots [20]$$

where a is the angle between the direction of flow and the vertical downward direction. In the saturated zone Equation [20] becomes Laplace's equation, provided the soil is isotropic and homogeneous. The value of h varies from positive values in the saturated zone to negative values in the unsaturated zone. Equation [20] can be solved using explicit or implicit methods (Haverkamp, et. al., 1977). In the explicit method a series of linearized independent equations is solved directly. In the implicit method, on the other hand, a system of linearized equations has to be solved. For a given grid point at a given time, the values of the coefficients C(h) and K(h) may be expressed either

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from their values at the preceding time step (explicit linearization) or from a prediction at time $(t + 1/2\Delta t)$ using an implicit linearization method described by Douglas and Jones (1963).

Equations [17] to [20] are valid only if one assumes the $h(\theta)$ relationship to be unique. Due to the strong non-linearity, there exists no general analytical solution. A specific solution for Equation [19] was first obtained by Philip (1957b). The solution is valid only for the case of infiltration in a homogeneous semi-infinite column satisfying the boundary conditions:

$$\begin{array}{lll} t \leq 0 & z \geq 0 & \theta = \theta_n \\ t \geq 0 & z = 0 & \theta = \theta_o \end{array} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [20a]$$

In a later paper (Philip, 1958) Equation [19] was solved for the conditions:

$$\begin{array}{lll} t < 0 & z \geq 0 & h = h_n \\ t \geq 0 & z = 0 & h = h_o \end{array} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [20b]$$

where h_o could take positive values corresponding to an infiltration experiment with submersion. Philip's method led to a solution of a power series in $t^{1/2}$, (Philip, 1955, 1957a, 1957b), having the form:

$$z = \lambda t^{1/2} + \chi t + \psi t^{3/2} + \omega t^{4/2} + \cdots + fm(\theta) t^{m/2} \ldots\ldots\ldots [21]$$

where λ, χ, ψ, ω,...,fm are single-valued functions of θ, subject to conditions [19a]. Equation [21] provides a theoretical formula for obtaining values of the vertical distance z versus θ, useful for predicting wetting distances in vertical columns.

The solution of Equation [21] is accurate for short time intervals; however, for long time intervals it fails to converge and loses accuracy. To avoid this, Philip (1957c) used a matching procedure to empirically link the short time solution with that of the long time. The t-range of convergence is dependent upon the soil characteristic and the initial and boundary conditions. Philip (1957d) compared theoretical moisture distribution curves with the experimental curves of Bodman and Coleman (1944) who divided the soil moisture profile into four zones: saturated, transition, transmission and a wetting zone. Philip's analysis predicted all these zones with the exception of the transition zone. Apparently in this zone diffusivity is not a unique function of moisture content due to air entrapment.

By introducing the relation:

$$(\partial z/\partial t)_\theta = -(\partial\theta/\partial t)_z \ (\partial z/\partial\theta)_t \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [22]$$

Equation [19] may be transformed into:

$$\partial z/\partial t + \partial/\partial\theta \ [D\partial\theta/\partial z] = \partial K/\partial\theta \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [23]$$

which is valid only if the K(θ) relationship is unique. Parlange

(1971) has proposed a quasi-analytical solution for Equation [23] with the following initial and boundary conditions:

$$t<0 \quad z\geq 0 \quad \theta=\theta_n$$
$$t\geq 0 \quad z=0 \quad q=k-D\partial\theta/\partial z=q_o \ldots\ldots [24]$$

One of Parlange's (1985) recent solutions for capillary rise has the form:

$$2K_1{}^2t/S^2 = -(1-a)^{-1}[2K_1i/S^2 + \ln(1-a^{-1} + a^{-1}\exp(-2aK_1i/S^2))] \ldots\ldots [25]$$

where K_1 is K_o-K_n, the difference between the conductivities at $\theta=\theta_o$ and $\theta=\theta_n$, respectively, and sorptivity S, which can be estimated when soil properties are known as a function of θ as given by Parlange (1975) is:

$$S^2 = \int_{\theta_n}^{\theta o} (\theta_o+\theta_{j+1/2}-2\theta_n)\ D_{j+1/2}d\theta \ldots\ldots [26]$$

Philip's (1957a) earlier definition of sorptivity was given by:

$$S = \int_{\theta_n}^{\theta o} \lambda d\theta \ldots\ldots [26a]$$

Both Equations [26] and [26a] give very nearly the same results for S.

Cumulative infiltration i is the volume of water that moves into the surface of the soil profile over a specified time and can be expressed as:

$$i = \int_{\theta_n}^{\theta o} z d\theta + K_n t \qquad [27]$$

By reducing [21] to only two terms, Philip attempted to describe an all-encompassing, simple infiltration equation well suited to the needs of applied hydrology. The two term algebraic equation is given by Philip (1957e):

$$i = St^{1/2} + At \qquad [28]$$

where

$$A = A' + K_n$$

$$A' = \int_{\theta_n}^{\theta o} z d\theta \qquad [28a]$$

The parameter A is not easily calculated. Models as presented by Philip (1969) and Kunze (1982) show that A' ranges from 0.378 to 0.67 K_o.

The infiltration rate can be obtained by differentiating Equation [28] with respect to time and setting $V_o = \partial i/\partial t$ to give:

$$V_o = 1/2St^{-1/2} + A \qquad [29]$$

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In both Equations [28] and [29] S is treated as constant while A is treated as variable corresponding to the gravitational component of flux. Sorptivity, is constant in horizontal flow but should not be expected to hold in vertical flow. In the vertical flow one should expect it to be the other way around because the gravitational part of the flux is the result of the constant gravitational field which does not change in our frame of reference.

Philip (1957f, 1958) studied the influence of the initial moisture content and the water depth, h, on the infiltration rate, cumulative infiltration and the shape of the moisture profile. The infiltration rate decreased while the advance of the wetting front increased at higher initial moisture content. He also found an increase in the depth of the saturated zone with larger h. In one of his papers (1958) he introduced the aspect of a tension saturated zone. He defined the zone as one where the volumetric moisture content is equal to that at $h=h_o$, but in which h assumes a non-zero negative value. In this study this zone is going to be called the theta-straight region, but it has also been referred to as "capillary fringe" in research articles and textbooks.

1.1 THEORETICAL DEVELOPMENT - FINITE-DIFFERENCE MODEL

The finite-difference model was developed by D. L. Nofziger (1985) for IBM compatible personal computers. The model is based on Equation [20] and can be used for finite or semi-infinite soil systems. In the finite case, the initial conditions are:

$$h(z,t) = h_{initial} \quad \text{for } t = 0 \text{ and } 0 < z < L \text{.........} [30]$$

where L is the length of the soil system. In the semi-infinite case, the initial conditions are:

$$h(z,t) = h_{initial} \quad \text{for } t = 0 \text{ and } 0 < z \text{............} [31]$$

At time $t=0$ and any later time, one of the following boundary conditions can be imposed at the upper boundary ($z=0$):

BC#1. Constant Potential h_o: $h(0,t) = h_o$

BC#2. Constant Flux Density q_o: $-[K(h)(\partial h/\partial z - \cos(a))] = q_o$ for $z=0$

BC#3. Mixed Type:

$$\begin{aligned} -[K(h)(\partial h/\partial z - \cos(a))] &= q_o \quad \text{for } z = 0;\ t \leq t_o \\ h(0,t) &= h_o \quad \text{for } z = 0;\ t > t_o \text{......} \ [32] \end{aligned}$$

where h_o and q_o represent the specified potential and flux at the soil surface, respectively, and t_o is the time at which the soil at $z=0$ reaches a potential h_o. The rainfall boundary condition #3 has q_o equal to the rainfall rate and h_o equal to zero. The flux is positive for water entering the soil system at $z=0$ and negative for water leaving the soil at $z=0$. For the finite soil system at time

$t=0$, one of the following boundary conditions can be imposed at the lower boundary ($z=L$):

BC#4. Constant Potential h_L: $h(L,t) = h_L$

BC#5. Constant Flux Density q_L: $-[K(h)(\partial h/\partial z - \cos(a))] = q_L$ for $z=L$

BC#6. Mixed Type:

$$\begin{aligned} &-[K(h)(\partial h/\partial z - \cos(a))] = q_L && \text{for } z = L;\ t \leq t_L \\ &h(0,t) = h_L && \text{for } z = L;\ t > t_L \ldots\ldots \quad [33] \end{aligned}$$

where h_L and q_L represent the specified potential and flux at $z=L$, respectively, and t_L is the time at which the soil at $z=L$ reaches a potential h_L.

The defined soil hydraulic properties equal the $\Theta(h)$ and $K(h)$ functions. The discrete experimental data obtained from capillary rise experiments were fitted using the analytical functions from Haverkamp et. al. (1977) through a non-linear parameter determination as described in the first chapter.

An implicit finite-difference scheme with explicit linearization was used. In this scheme the partial differential equation takes the form:

$$\begin{aligned} &C_{(i,j)}(h_{(i,j+1)}-h_{(i,j)})/\Delta t = \\ &\qquad 1/\Delta z\ [K_{(i+1/2,j)}((h_{(i+1,j+1)}-h_{(i,j+1)}]/\Delta z)-\cos\ (a)) \\ &\qquad -K_{(i-1/2,j)}((h_{(i,j+1)}-h_{(i-1,j+1)}]/\Delta z)-\cos\ (a))] \end{aligned}$$

where

$$h_{(i,j)} = h_{(i\Delta z,j\Delta t)},$$

$$C_{(i,j)} = C(h_{(i,j)}),$$

$$K_{(i+1/2,j)} = [K(h_{(i,j)} + K(h_{(i+1,j)}))]/2,$$

$$K_{(i-1/2,j)} = [K(h_{(i-1,j)} + K(h_{(i,j)}))]/2 \ldots\ldots\ldots\ldots\ldots [34]$$

and Δt and Δz are the mesh sizes in time and depth, respectively.

The finite-difference equation above is used for all interior mesh points of the soil system. Special forms of this equation are used to represent the boundary conditions selected. This results in a system of simultaneous equations which must be solved for each time step Δt. One equation applies to each mesh point in depth. Since each equation involves only three unknowns, the system of equations defines a tridiagonal matrix which is relatively easily solved for up to 200 equations on the personal computer. All floating point calculations are performed in double precision (16 decimal digits). The software contains an algorithm to determine the initial mesh sizes in time and depth. Frequently the mesh size in time and depth can be increased as the time increases. The model contains an algorithm to adjust the mesh size based on the mass balance and the depth of wetting. In the following discussion the term 'implicit scheme' will be used for the finite difference implicit model.

1.2 THEORETICAL DEVELOPMENT - FINDIT

Equation [19] in its one-dimensional form can describe capillary rise with initial and boundary conditions:

$$\begin{aligned} \theta(z,t) &= \theta_n \quad \text{for } z>0,\ t=0 \\ \theta(z,t) &= \theta_o \quad \text{for } z=0,\ t>0 \\ \theta(z,t) &= \theta_n \quad \text{for } z \dashrightarrow \infty,\ t>0 \ldots\ldots\ldots\ldots [35] \end{aligned}$$

At the moment the water surface is brought into contact with the soil, θ_n changes to θ_o at the lower soil boundary, while the rest of the semi-infinite height soil column remains at θ_n.

The calculations of the distance z_j at all θ_j values are given by the following two recursion equations as they were derived by Kunze (1984):

$$z_{j+1} = z_j$$

$$+D_{j+1/2}\ d\theta\ /\ [\partial/\partial t(\int_{\theta_n}^{\theta_{j+1/2}} (z/t)t\ d\theta)+K_{j+1/2}-K_n] \ldots\ldots\ldots\ldots [36]$$

and

$$\int_{\theta_n}^{\theta_{j+1/2}} (z/t)\ d\theta = \int_{\theta_n}^{\theta_{j+1/2}} (z/t)\ d\theta - z_{j+1}/t\ d\theta \ldots\ldots\ldots\ldots [37]$$

For capillary rise the flux at the free water surface V_o can be partitioned into matric and gravitational components (Kunze and Nielsen, 1982, 1983) to give:

$$V_o = V^m{}_o + V^g{}_o \ldots\ldots\ldots\ldots [38]$$

or can be partitioned at any moisture content:

$$V_j = V^m_j + V^g_j \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots [39]$$

where a negative sign proceeding the second term on the right side means that gravitational force opposes the matric force.

Because flux is the time derivative of cumulative infiltration, equation [39] can take the form (see Kunze and Nielsen, 1982, 1983):

$$V_{j+1/2} = \partial/\partial t\ [\int_{\theta_n}^{\theta_{j+1/2}} (z/t)t\ d\theta] \ldots\ldots\ldots\ldots\ldots\ldots [40]$$

where fluxes are now evaluated at the average water content of a θ interval indicated by 1/2 in the subscript and integration limit, consistent with $D_{j+1/2}$ and $K_{j+1/2}$, the average diffusivity and conductivity, respectively. As Kunze and Nielsen (1982) showed, Equation [40], which is part of the recursion formula [39], is broken down and evaluated in terms of two derivative products, namely:

$$\partial/\partial t\ [\int_{\theta_n}^{\theta_{j+1/2}} (z/t)t\ d\theta] = \int_{\theta_n}^{\theta_{j+1/2}} z/t\ d\theta + t\partial/\partial t\ (\int_{\theta_n}^{\theta_{j+1/2}} z/t\ d\theta) \ldots\ldots [41]$$

The actual or specific flux in Equation [41] is calculated from the average flux plus the change of the average flux. This calculation is the central feature of FINDIT procedure. For capillary rise the specific flux is always less than the average flux. Equation [41] must converge for all $\theta_{j+1/2}$ values in the moisture profile before a solution is acceptable. Present criteria require that:

$$\Delta_{j+1/2} < (10^{-4})\ t\partial/\partial t\ (\int_{\theta_n}^{\theta_{j+1/2}} z/t\ d\theta) \ldots\ldots\ldots\ldots\ldots\ldots [42]$$

where $\Delta_{j+1/2}$ is the change in successive iterative calculations of the far right-hand-side term of Equation [41].

1.3 MATERIALS AND METHODS

A series of capillary rise experiments were performed in the laboratory and the experimental data were fitted to obtain the $\theta(h)$, $K(\theta)$ and $D(\theta)$ functions used as input to Equations [34] and [36]. Details of the measuring techniques were reported in the first chapter.

2. RESULTS AND DISCUSSION

Figure 16 shows the moisture profiles for capillary rise calculated with FINDIT over wide range of time values. It is apparent from the curves that : 1) FINDIT can cope with a large range of time values; 2) the wetting process slows down appreciably with time as was shown in the first chapter; and 3) after long periods of time equilibrium was established. FINDIT approximated the experimental moisture characteristic curve very well as was shown in Chapter 1. Kunze et. al. (1985) also showed that FINDIT approximated the wetting characteristic curve calculated from the analytical steady state solution of Richards' equation ie. the solution of Equation [19] without the left hand side term, as was discussed by Philip (1966) and Parlange and Aylor(1972). These two cases are the best evidence that the recursion formulas and the iterative process of FINDIT are not only rigorous, but also highly sensitive to the laws governing upward water flow in soils.

The accuracy in determining the wetting heights is mainly dependent on the accuracy of numerical calculations of the fluxes as given by Equation [41]. This equation essentially breaks down the actual or specific flux (left hand side of the equation) into average flux and the change in average flux or the derivative term. The flux terms have a physical significance. The average flux corresponds to the average velocity of soil water whereas the actual flux corresponds to the velocity at a specific time. The cumulative infiltration at time t may be calculated as a function of average flux, v_{av}, with the

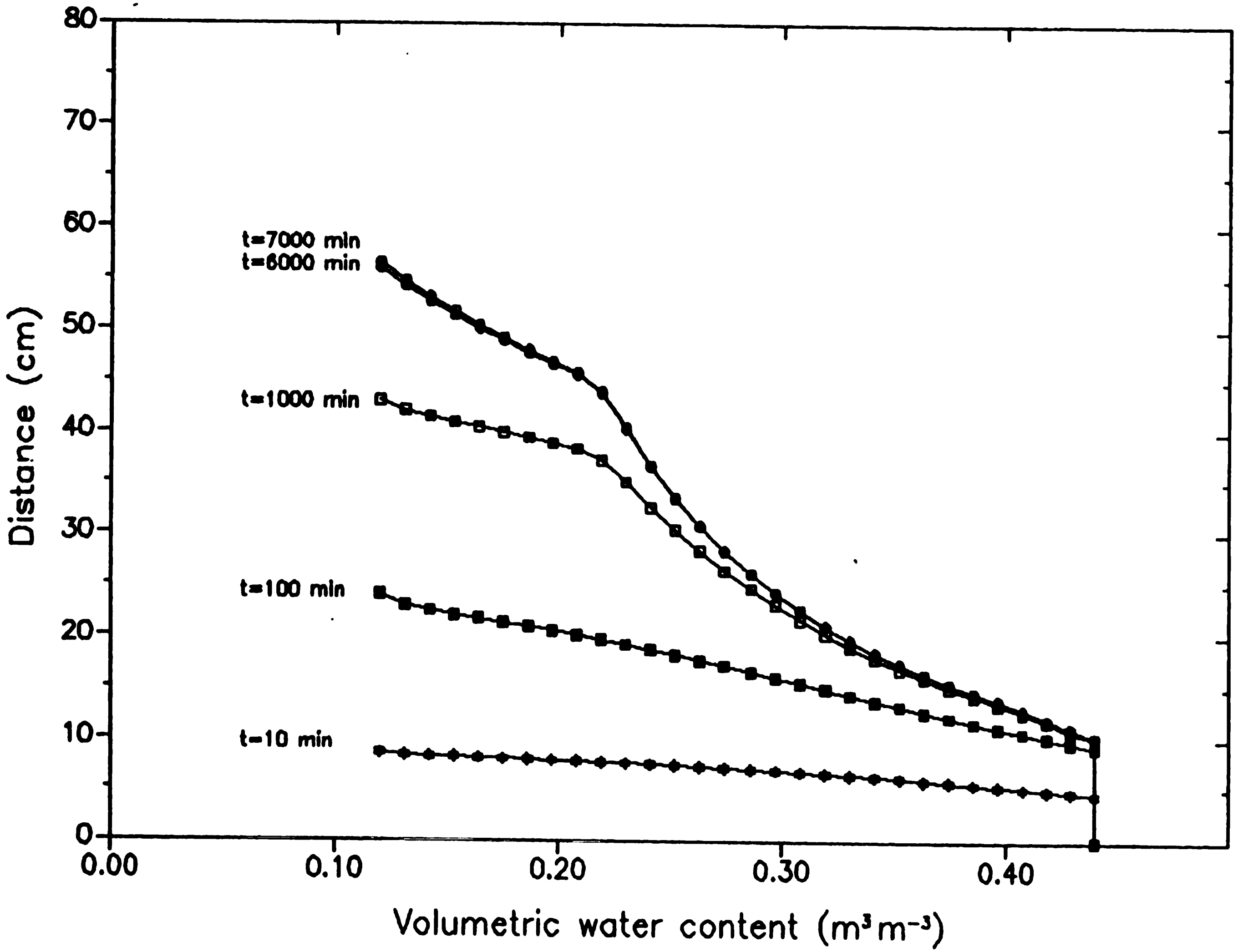

Fig. 16. Moisture profiles calculated with FINDIT for Metea sandy loam soil at indicated time values.

following algebraic equation:

$$i = v_{av} \, t \qquad [43]$$

The derivative term corresponds to change of the matric field since the gravitational field is constant. The second term on the right hand side of Equation [41] was calculated numerically using the three point derivative, ie. at $T+\Delta T$, T and $T-\Delta T$. The derivatives at $T\pm\Delta T$ were determined from the derivative at T multiplied by the respective average flux terms at $T\pm\Delta T$ and divided by the average flux at T. Figure 17 shows the change in the different flux components over time for a given moisture content at the soil surface where the moisture content is saturation. Hence the fluxes for any moisture content could be represented as was done for saturation in Figure 17. The derivative curve is negative, continuous and smooth and has a similar shape as the average flux term which is to be expected based on the method of calculations. As stated earlier, the specific flux is always less than the average flux but both decrease with time. In any case their difference is always a negative quantity represented by the flux derivative.

Figure 18 illustrates the moisture profiles calculated with the implicit finite-difference scheme. The implicit scheme also showed the ability to cope with a wide range of time values. The model was stable for $\Delta T=1x10^{-4}$ min and $\Delta Z=.5$ cm. These time and space mesh sizes were chosen by a special algorithm supplied by the model. Limitations of the finite-difference approach include errors in calculating the mass balance. Here the quantity of water entering or

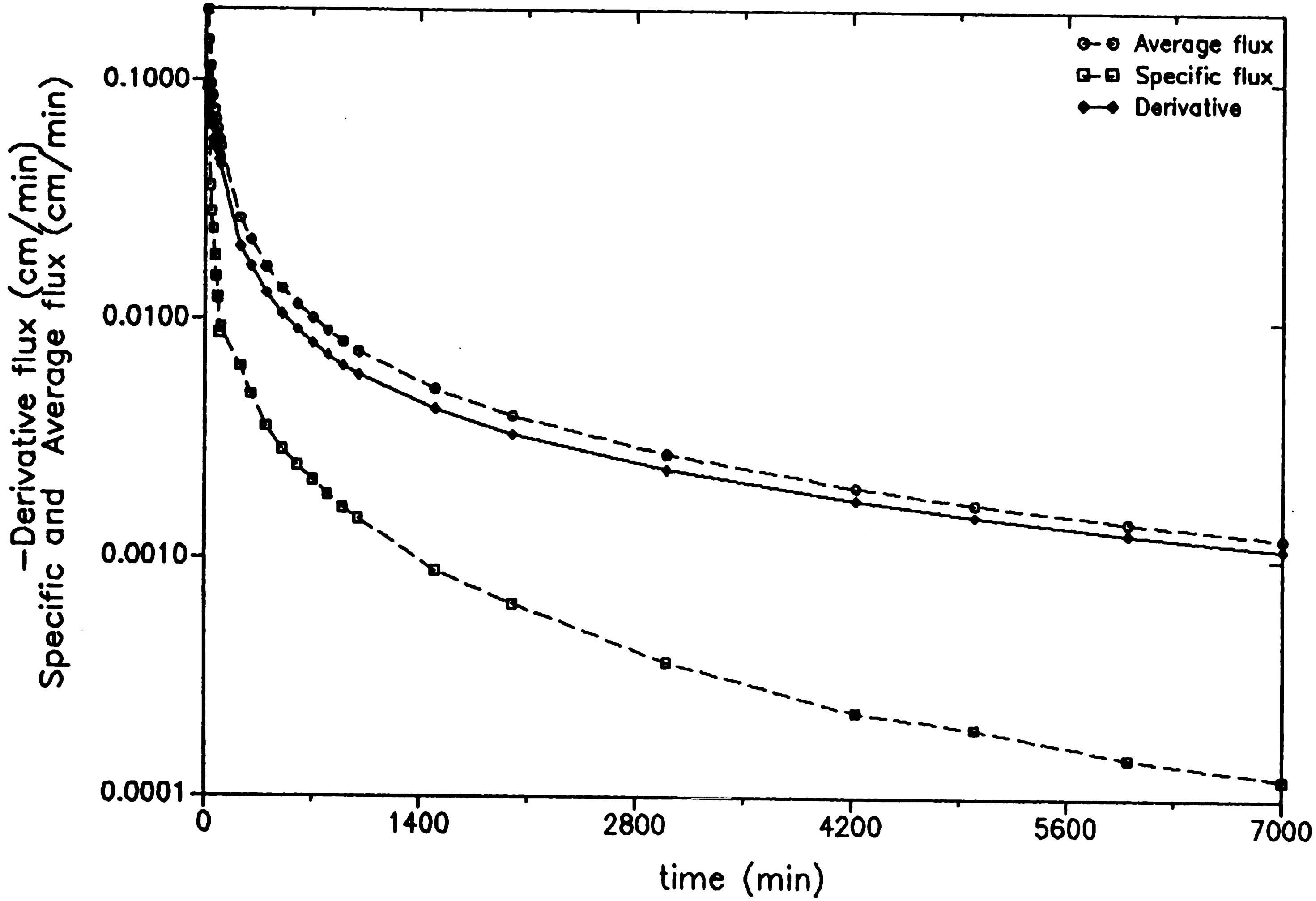

Fig. 17. The specific, average and derivative fluxes calculated at the soil plate interface with FINDIT.

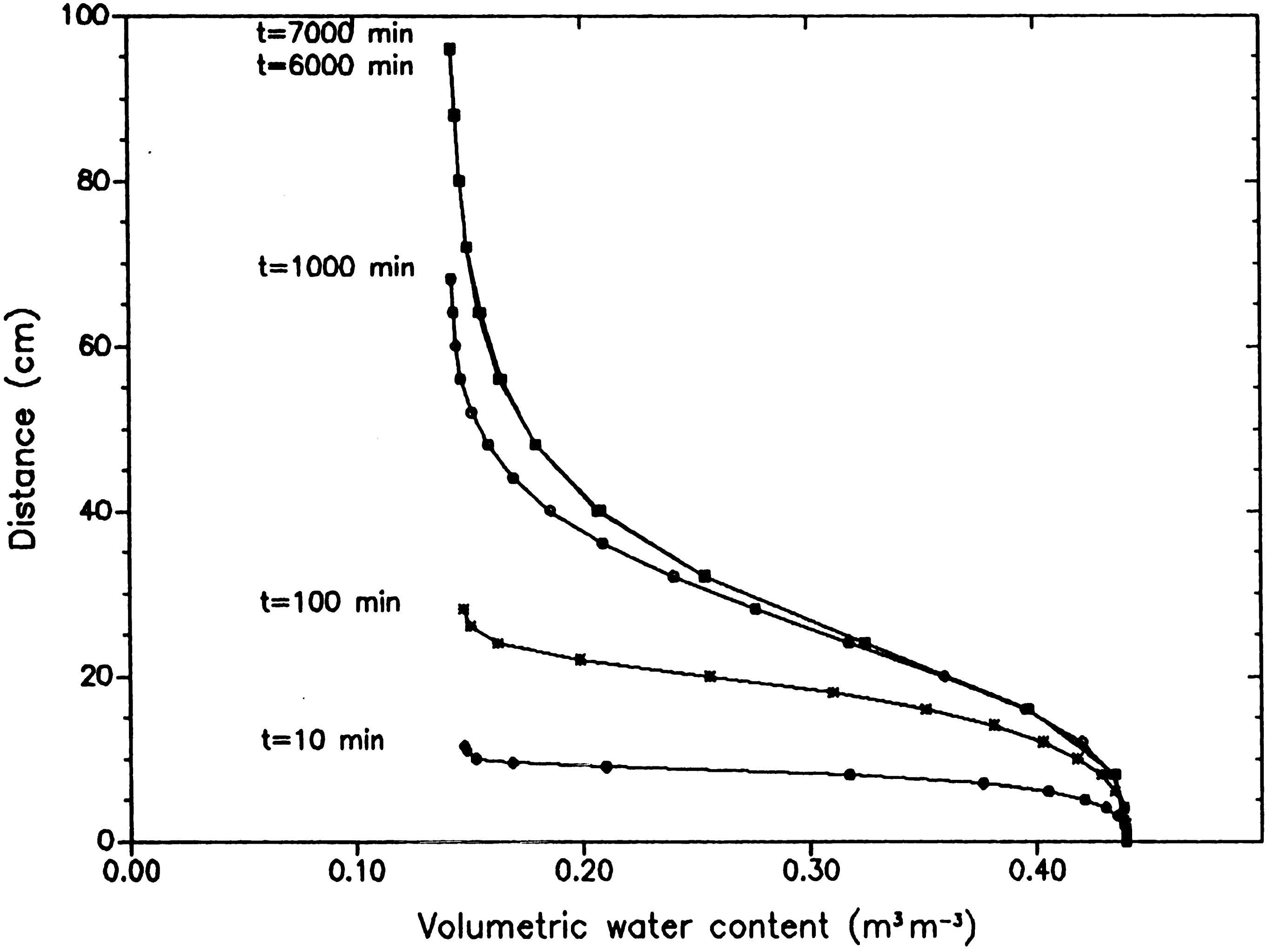

Fig. 18. Moisture profiles calculated with the implicit scheme for Metea sandy loam soil at indicated time values.

leaving the soil surface may not be the same if determined from surface fluxes and from changes in water content especially for short time values. However, no oscillation of the results was observed in the calculations.

The profiles generated by the two models were compared for short and long time values in Figure 19. The agreement was acceptable over most of the moisture range. In the low moisture range the agreement was not as good as expected. Discrete values of the diffusivity and conductivity functions were supplied at mid-moisture intervals for FINDIT. Haverkamp's analytical formulas, Equations [12] and [13], were used to supply the functions for the implicit scheme. The non-linear parameter determination routine used in conjunction with Haverkamp's equation was not able to successfully fit the conductivity function in the low moisture range. Hence, the discrepancy in the conductivity function resulted in the higher wetting front predictions when simulation was done with the implicit scheme. FINDIT approximated the experimental profile very well as was shown in Figures 13 and 14. At equilibrium, the implicit scheme again simulated the experimental profile well in the saturation to 25% moisture range but not very well in the drier region as the bound analysis suggests in Figure 20. The largest discrepancy was in the lowest water content region as expected.

The specific fluxes from FINDIT and the implicit scheme were compared in Figure 21 . The agreement was very good for the whole time range; however, the flux simulated with the implicit scheme was lower than the flux simulated with FINDIT for time values more than 350 min. For shorter times the flux from the implicit scheme was higher than

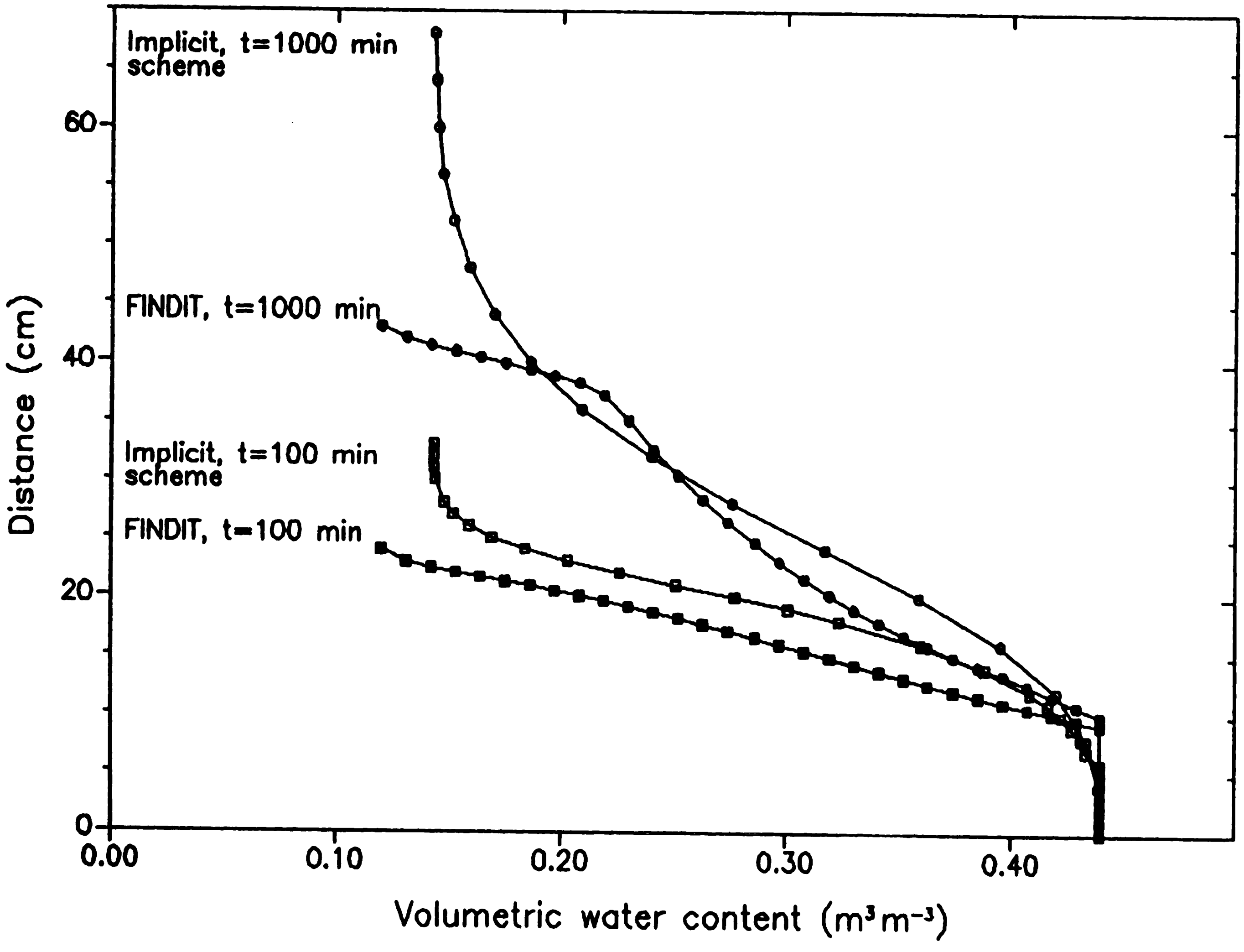

Fig. 19. Comparison of moisture profiles obtained with the implicit scheme and FINDIT at indicated time values.

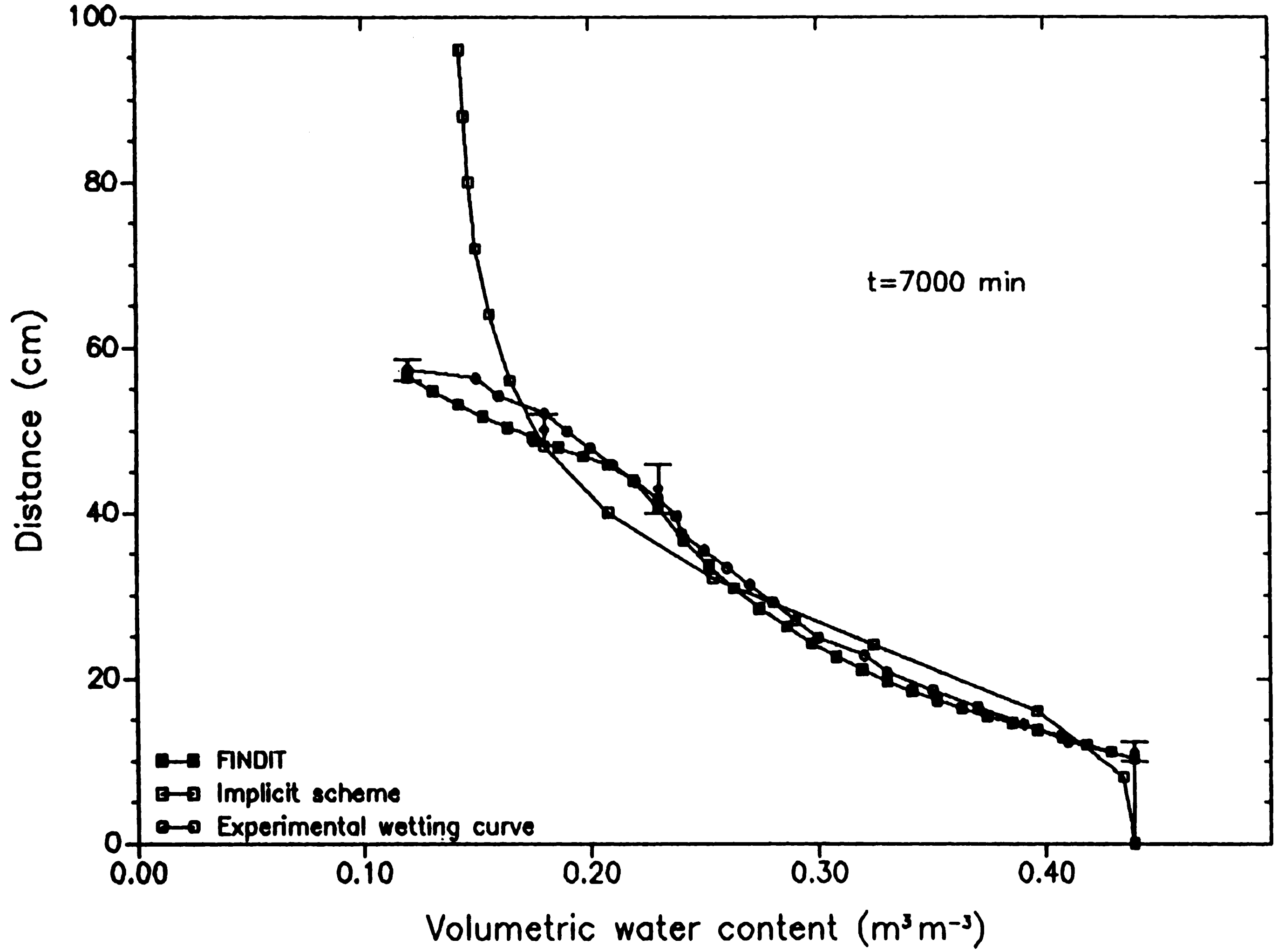

Fig. 20. Comparison of capillary rise experimental data with moisture profiles obtained with the two models for Metea sandy loam soil.

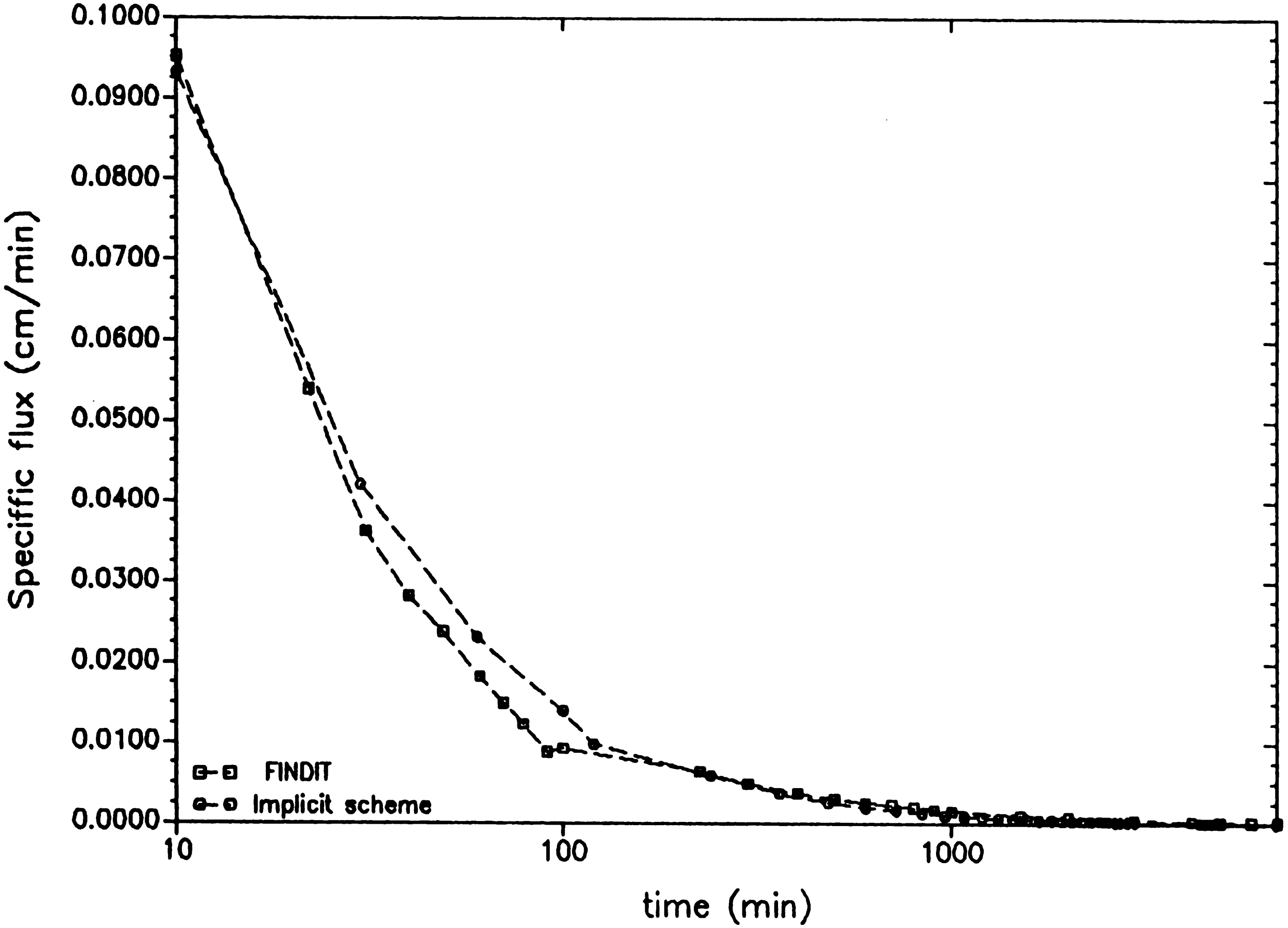

Fig. 21. Ccmparison of specific fluxes at soil plate interface calculated with FINDIT and the implicit scheme.

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from FINDIT. This is in agreement with the cumulative infiltration generated over time for the two models, illustrated in Figure 22. Although cumulative infiltration obtained with FINDIT was always lower, its final value is nearly that of the implicit scheme. Cumulative infiltration could be considered as the area under the wetting profiles as shown in Figure 19. The area under the profile obtained with the implicit scheme was larger than the area obtained with FINDIT resulting in higher cumulative infiltration. The experimental infiltration in Figure 22 was lower than both simulated infiltration profiles; part of this difference may be ascribed to errors in experimental techniques, however FINDIT was able to approximate the experimental infiltration profile better than the implicit scheme. To further explore the agreement between cumulative infiltration simulated with the two models, the cumulative infiltration ratios were plotted as a function of time in Figure 23. Acceptable agreement was observed over both short and long time values but a maximum difference of 12.56% was observed in the mid-time range.

Both models have limitations in describing water movement in soils. The fact that the models are designed for one-dimensional flow in homogeneous soils means that they will not precisely describe flow in heterogeneous soils where flow is likely to be three dimensional. Computational speed may be a limitation mainly with the implicit scheme. This problem was greatly reduced by the use of the 8087 math co-processor. The vapor phase was not incorporated in either models. How much it contributed in the low moisture range has not been assessed.

Piecemeal integration is a routine to check FINDIT for accuracy

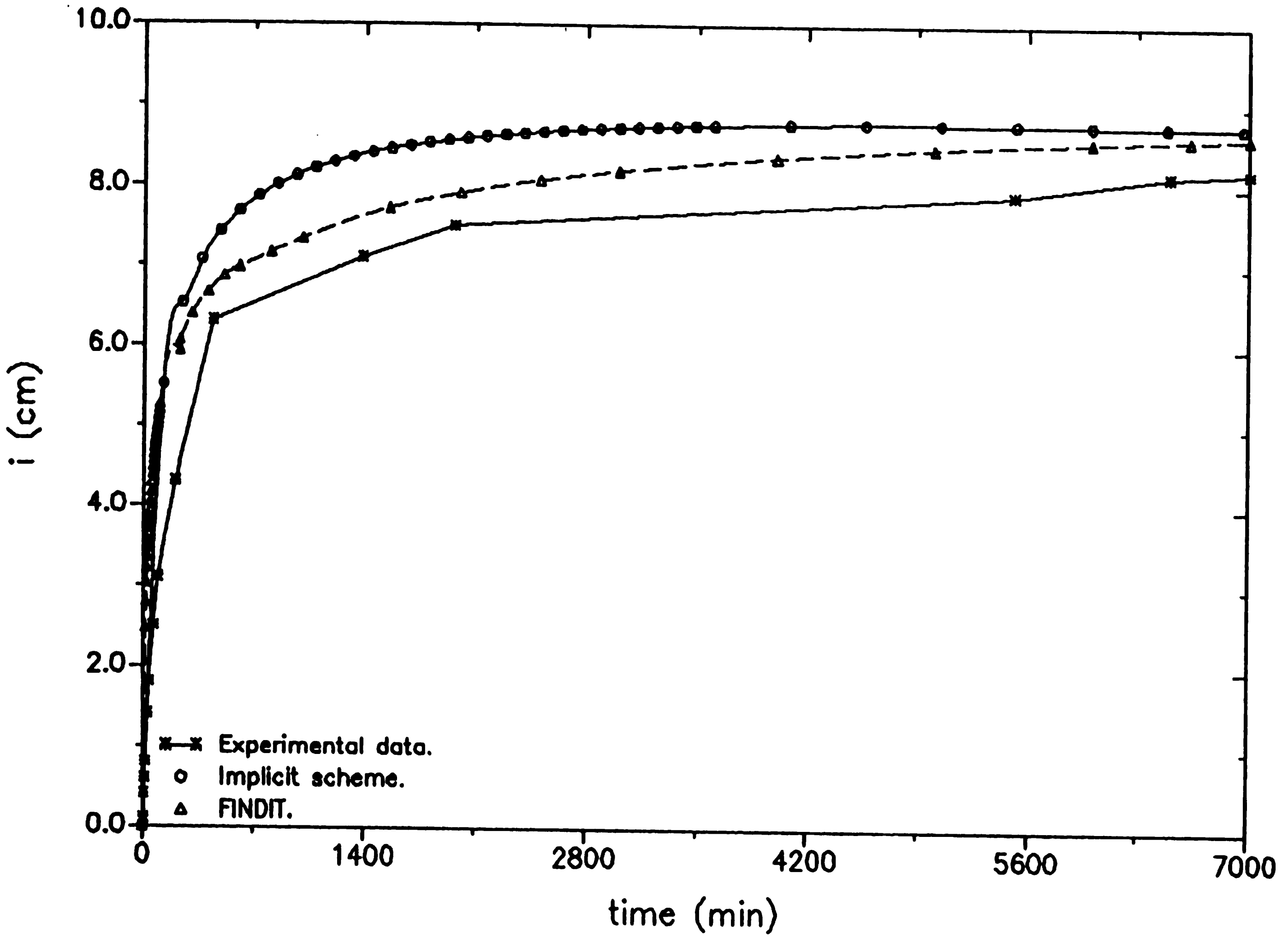

Fig 22. Cumulative infiltration i versus time t obtained with FINDIT, the implicit scheme and experimental capillary rise data.

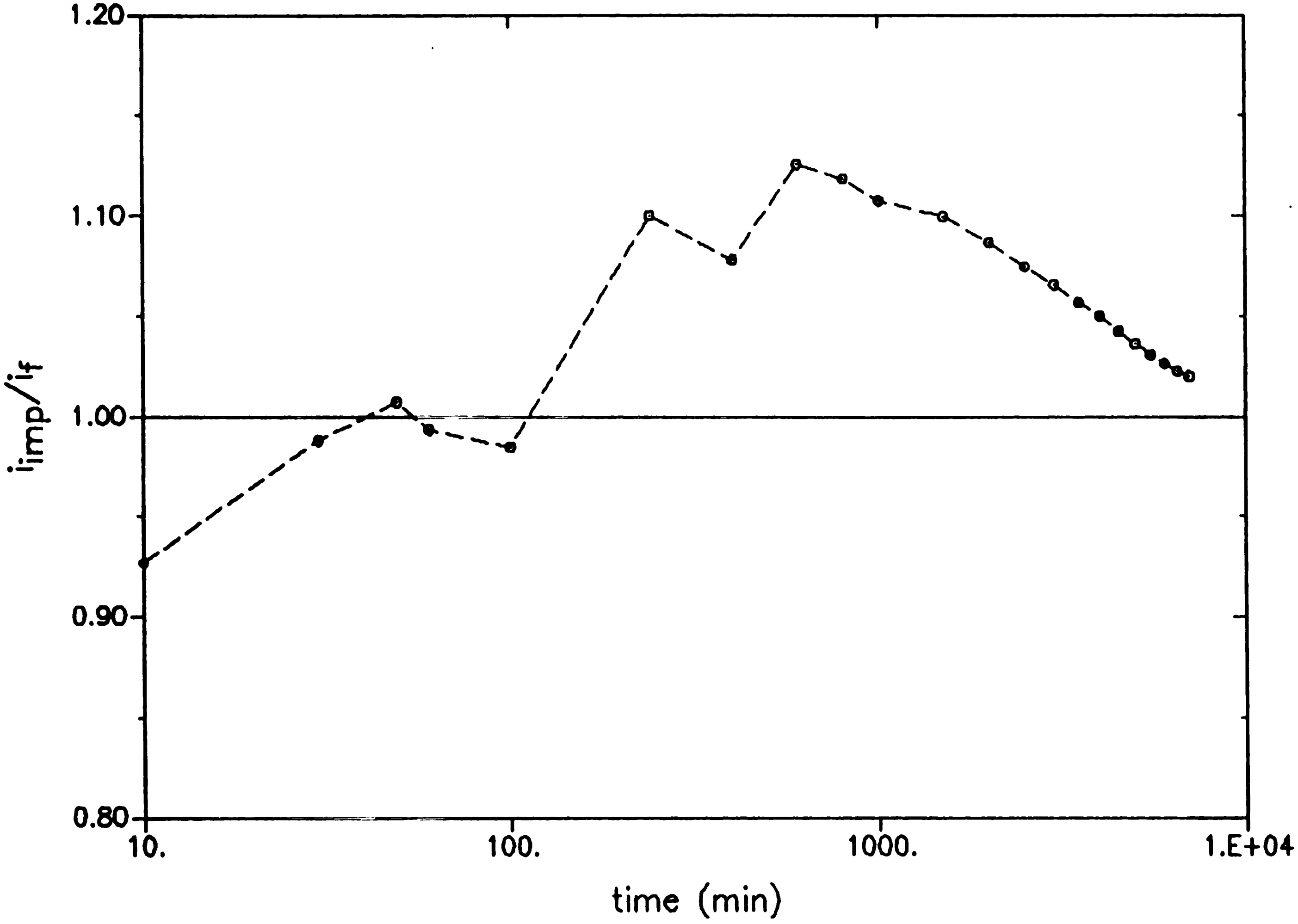

Fig. 23. Cumulative infiltration ratios obtained with FINDIT and the implicit scheme for capillary rise.

TABLE 3. Cumulative infiltration from FINDIT and piecemeal integration, $i_{p,n}$, for Metea sandy loam soil.

Time (min)	FINDIT(cm)	$i_{p,n}$ (cm)	Difference
13	2.4413	2.3811	0.0602
100	5.2632	5.1917	0.0715
240	5.9290	5.8809	0.0481
400	6.5526	6.4871	0.0655
600	6.8721	6.8682	0.0039
800	7.1558	7.1299	0.0259
1000	7.3314	7.3286	0.0028
1500	7.6907	7.6820	0.0087
2000	7.9076	7.9007	0.0069
2500	8.0710	8.0660	0.0050
3000	8.1879	8.1838	0.0041
3500	8.2856	8.2821	0.0035
4000	8.3596	8.3567	0.0029
4500	8.4252	8.4226	0.0026
5000	8.4766	8.4745	0.0021
5500	8.5228	8.5209	0.0019
6000	8.5599	8.5580	0.0019
6500	8.5951	8.5934	0.0017
7000	8.6233	8.6218	0.0015

of the derivative calculations. It presents another way of determining the cumulative infiltration beside the standard cumulative infiltration calculated by FINDIT; however data generated by FINDIT are necessary. If the two do not agree, the derivative calculations are in error. The piecemeal integration routine calculates the cumulative infiltration, $i_{n,p}$, at the present time, T_n, as a function of the specific flux, sfl_p, summed over the interval ΔT and added to the accumulated flux obtained at T_{n-1}:

$$i_{n,p} = v_{av_{n-1}} T_{n-1} + [(1/m) \sum_{p=1}^{m} sfl_p] \, \Delta T \ldots\ldots\ldots\ldots [44]$$

The sfl_p values for the ΔT time interval were calculated using a cubic spline technique. The results obtained are shown in Table 2 for $m=20$. Data illustrate that the agreement between cumulative infiltration obtained with FINDIT and piecemeal integration, $i_{n,p}$, is excellent suggesting that the calculation of the derivative is accurate.

3. SUMMARY

FINDIT supplies the user with two physically based components of the specific flux, the average flux and the flux derivative. It is from the latter two fluxes that the specific flux at any moisture content can be calculated and the wetting profiles generated. Average flux can be used to calculate cumulative infiltration at any time with FINDIT.

In the previous chapter it was shown that profiles obtained with

FINDIT were in excellent agreement with experimental profiles. Here FINDIT is compared with an implicit finite-difference scheme. The profiles from the two models showed fair agreement with FINDIT approximating the experimental profile better at higher suctions. Specific soil water fluxes agreed very well over time. When cumulative infiltration was used as a criterion, the agreement between the two models did not exceed a maximum difference of 12.56%. Hence, the overall agreement of the two models was found to be acceptable for most capillary rise work. Cumulative infiltration from FINDIT and piecemeal integration routines compared excellently. Lack of agreement would have suggested that the derivative function was not calculated accurately.

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CHAPTER III

SOIL WATER CONTENT AND SOIL WATER POTENTIAL AS DRIVING FORCES

1. INTRODUCTION

The potential concept of soil water is of primary importance in determining the state and movement of water in soil. The soil water potential is defined as the amount of work that a unit quantity of water in an equilibrium soil-water system is capable of doing when it moves to a pool of water in the reference state at the same temperature. Buckingham (1907) first introduced the concept of soil water potential without mentioning any relationship between his equation and Darcy's law. W. Gardner and Widtsoe (1922), were the first to include a gravity component in defining the potential for the soil solution. L. A. Richards (1928) defined a total potential, h, in reference to the soil solution as the sum of two components:

$$h=h_m+z \quad \ldots\ldots [45]$$

in which h_m is the capillary potential or matric potential often referred to as tension head and z is the gravitational potential. Richards was the first to recognize that Darcy's equation applies to flow in unsaturated soil and he observed that K depends on moisture content. Edlefsen and Anderson (1943) also noted that the constituent water in a soil solution may be transported in response to concentration gradients as well as in response to gradients of a solution potential.

There are shortcomings in using the total potential concept as Corey and Klute (1985) point out in their review. The most important problem is that the osmotic potential, which refers to the water component only, has been added to pressure and gravitational potentials of the soil solution. This is why the total potential cannot be a valid potential for liquid flow and why the net transport of water may be in any direction with respect to a gradient of the quantity defined. Another problem with the potential concept is that the effects of both gravity and pressure cannot be uniquely defined for a solution of varying density.

Isothermal transport of water in the solution phase of a porous medium can be considered at both the microscopic and macroscopic levels. Convection and diffusion may be formulated in terms of a linear rate law, stating that flux is proportional to the moving force. When convection, also called mass flow, is taking place the entire mass of water is moving from a zone of higher pressure to a zone of lower pressure in the context of total soil solution pressure. In the case of diffusion on the other hand, the moving force is the gradient of concentration or water content, and it causes the molecules of the unevenly distributed water content to migrate from a zone of higher to lower concentration even while the liquid phase as a whole may remain stationary.

According to Corey and Klute (1985) there are three requirements for thermodynamic equilibrium in any system: 1) thermal equilibrium; this requires that the temperature of the system remains uniform, 2) mechanical equilibrium; this requires that there be no net force tending to produce convection acting on any part of the system, and 3) chemical equilibrium; this requires that there be no net diffusion transport or chemical reaction of any components in the system. Thus, the constant soil solution potential is a necessary, but not a sufficient condition for thermodynamic equilibrium of water in a soil-water system.

The relation between soil water content and the soil solution potential is important in characterizing the hydraulic properties of soil as well as solving the flow equation for soil water. This relationship is identified as the water retention function or soil moisture characteristic. The function relates a capacity factor, the water content, θ, to an intensity factor, h, the energy state of the soil water.

The water retention function is primarily dependent upon the texture or particle-size distribution of the soil and the structure or arrangement of these particles. Also, the organic matter content and the composition of the solution phase play an important role in determining the retention function (Salter & Williams, 1965; Richards & Weaver, 1944; Croney & Coleman, 1954).

The traditional method of determining the water retention function involves establishing a series of equilibria between water in the soil sample and a body of water at a known potential. Each data point (θ_v, h_m) constitutes one point on a retention function. In laboratory measurement the soil-water system is in hydraulic contact with the body of water via a water-wetted porous plate or membrane. At each equilibrium, the volumetric water content, θ_v, of the soil is determined and paired with a value of the matric pressure head, h_m, determined from the pressure in the body of water and the gas pressure phase of the soil.

The conventional way to determine the retention function experimentally is by obtaining adsorption or desorption scanning curves. A drainage curve is mapped by establishing a series of drainage equilibria starting from a zero pressure head. A wetting curve is obtained by equilibrating samples, bringing them to some degree of saturation from a lower water content. The retention function is hysteretic, i.e. the water content at any given pressure head for a wetting soil is always less than that for a draining soil (Poulovasilis, 1962; Pavlakis & Barden, 1972).

Soils of relatively rigid structure have retention curves where the water content is nearly constant (theta-straight region) as the matric pressure head is decreased from zero. In this region the slope $d\theta/dh_m$ is zero. In all cases a decrease of water content requires the entry of air (Klute, 1986).

In determining the retention function through a pressure cell apparatus, the total pressure head is given by:

$$h=h_a+h_m \qquad [46]$$

where h_m is the matric pressure head and h_a is the gas phase pressure or the pneumatic pressure head. When the soil sample is at equilibrium, the pneumatic pressure head is positive, and the matric pressure head has an equal negative value. When the gas pressure is released, it is assumed that the matrix pressure is unchanged. Klute (1986) showed that this assumption is not totally valid for two reasons. First, the isolated bodies of entrapped air in the solution phase of the sample will expand when the reduction of the gas phase pressure is transferred to the liquid phase. This expansion of entrapped air will affect the curvature of the interface between the solution phase and the continuous gas phase affecting the matrix pressure head. Second, disturbance of the soil matrix is possible when the sample is removed from the chamber causing the matrix pressure head to increase.

During the upward isothermal infiltration process, commonly called capillary rise, there is a point where the gravitational and matrix components of the potential are equal but of opposite sign. At that time all points in the soil solution are at equilibrium. Here the height above the water table is the matric potential h_m. The matric potential h_m and the corresponding volumetric water content θ_v are the data pair at one point of the retention function. Assuming osmotic potential is negligible, the retention function can be obtained via a simple capillary rise experiment.

Three forces influence the movement of soil solution into a porous medium: a) gravity, which is constant and may be either a positive or negative force relative to a reference elevation, b) adhesion and, c) cohesion. The last two, collectively known as surface tension, are the two constituents of the matrix water potential. Bouyoucos (1925) reported that if columns of dry clay are wetted with either water or kerosene, the organic liquid at the end of three days will have risen four or five times as high as the water in spite of its lower surface tension. He explained this phenomenon as the inability of the colloidal material to absorb the kerosene and consequently reduce the effective pore space. Pekishek, et. al. (1962) and Sunderman and Banbury (1978), have shown that wetting agents have either no effect or an adverse effect on infiltration into wettable, hydrophilic, soils. A wetting agent (surfactant) applied to a hydrophilic soil where the wetting angle is already zero and no further decrease is possible, is going to reduce the effective pore space and thus have an adverse effect. On the other hand, a decrease of the wetting angle and a reduced surface tension causes higher infiltration rates in non-wetting soils. The rules of wetting a soil that apply to infiltration also apply to capillary rise. Hence, when the soil is water repellent (hydrophobic), the addition of a wetting agent thus will increase the amount of water adsorbed and the height that water can rise in the soil matrix (Bond, 1968).

2. MATERIALS AND METHODS

The soil selected for this study was Metea Ap whose physical properties were described in Chapter 1. The hydrophobic character of the soil was presumed but not assessed. The average bulk densities were 1.15 and 1.16 Mg m^{-3} with standard deviations of ± 0.023 and ± 0.022 for the treated and the untreated columns, respectively, and are listed as columns 8 and 9 in Table 2. The profiles at equilibrium were determined experimentally and simulated using the FINDIT model as described in the first chapter. The diffusivity and conductivity functions were used as input to the model and were obtained as outlined in Chapter 1.

Capillary rise experiments were conducted in the same way as described in the first chapter. In addition, soil which was labeled 'treated' received 2048 ppm of the blended non-ionic soil wetting agent, Aqua-Gro, manufactured by Aquatrols Corporation of America and dried before it was packed in the column. Both columns were wetted only with tap water in the capillary rise experiments.

3. RESULTS AND DISCUSSION

The diffusivity and conductivity functions for the wetting agent treated and untreated soils are shown in Figure 24. The functions are not significantly different for the two soils. Figure 25 shows the simulated wetting curves using the finite-difference scheme and FINDIT at equilibrium. The wetting curves obtained with the two models showed good agreement over the wetter half of the moisture range and they are in agreement with the observations made in Chapter 2, Figure 20. The simulated wetting characteristic curves were higher for the wetting agent treated soil with both models. These results are supported in Figure 26 where simulated water profiles are shown for the treated and untreated soils at 30 and 600 minutes, respectively. The capillary rise was higher for the wetting agent treated soil at all moisture contents except close to saturation. Hence the trend of the wetting characteristics of the two columns observed at equilibrium (7000 minutes) in Figure 25 could already be observed at short times. This suggests that the treated soil had a greater adhesion capability for water during the entire wetting process.

Figure 27 shows treated and untreated experimental capillary rise profiles at 7000 minutes. At this point the soil water system was at equilibrium for all practical purposes and no further movement of the water front would be observed. At each point the gravitational potential was equal but of opposite sign to the matric potential producing the wetting characteristic curve. The

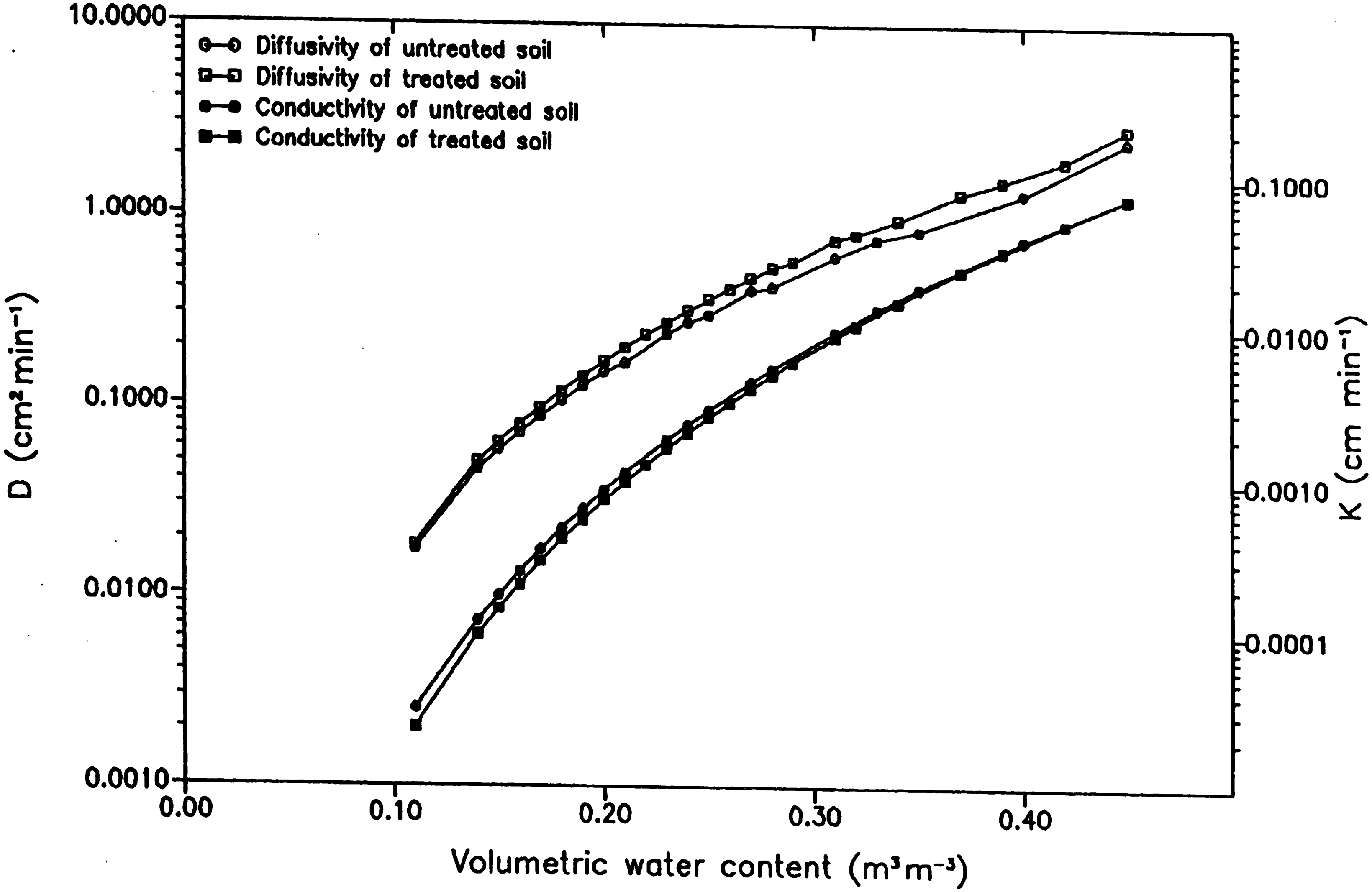

Fig. 24. Diffusivity and conductivity functions for treated and untreated Metea sandy loam soil.

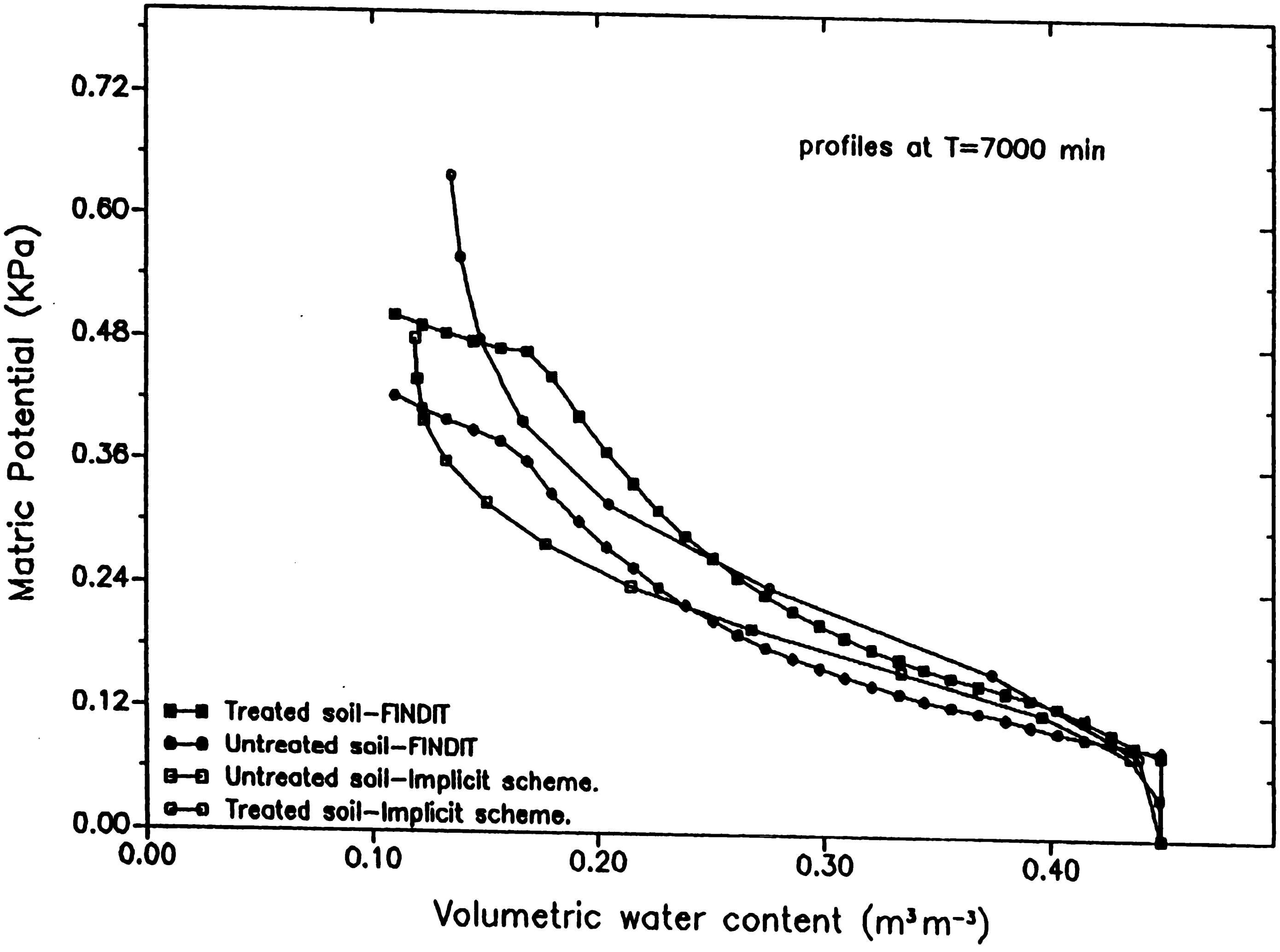

Fig. 25. Simulated wetting curves for treated and untreated soils using the implicit scheme and FINDIT.

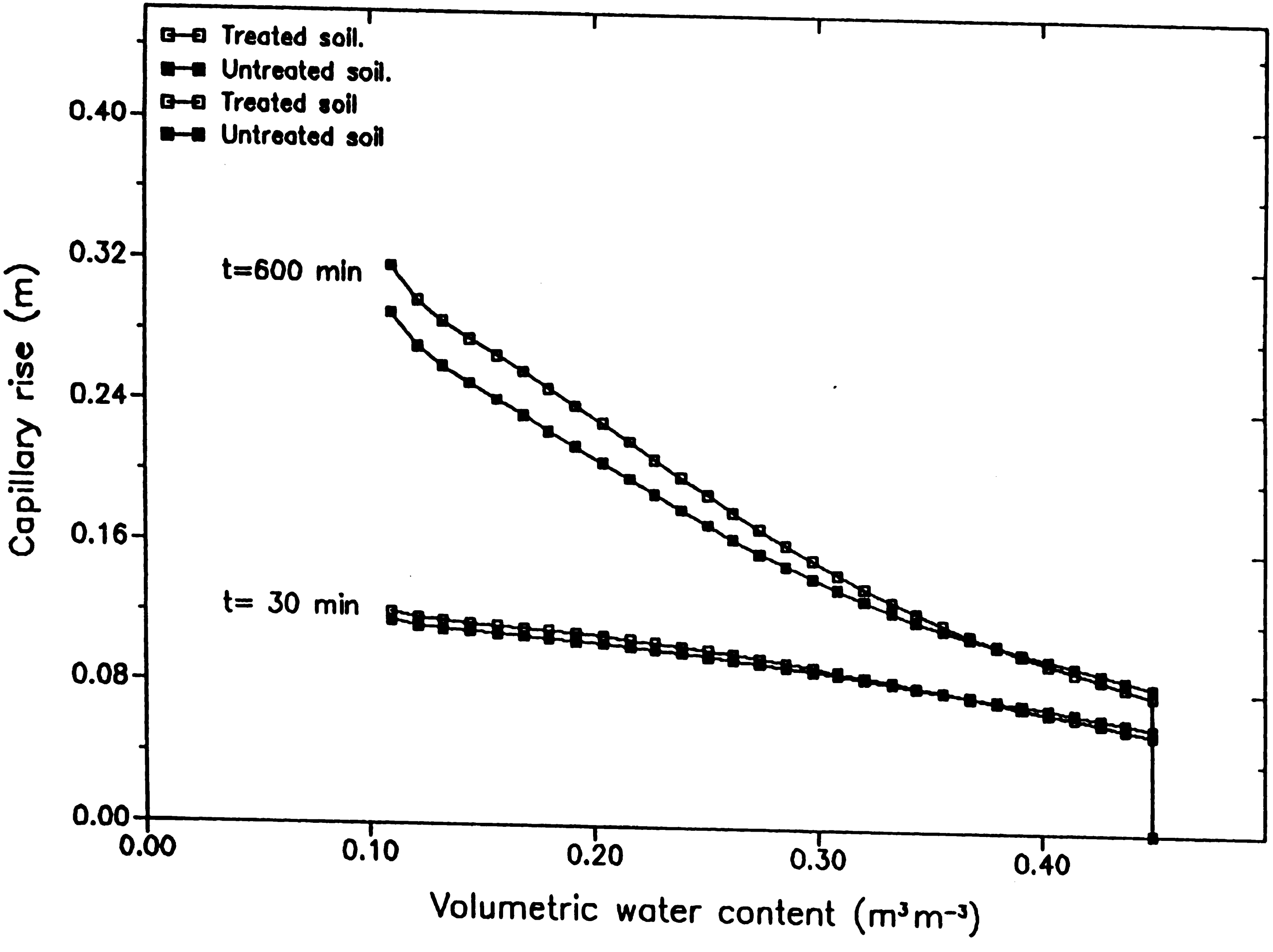

Fig. 26. Simulated water profiles for treated and untreated soils at 30 and 600 min.

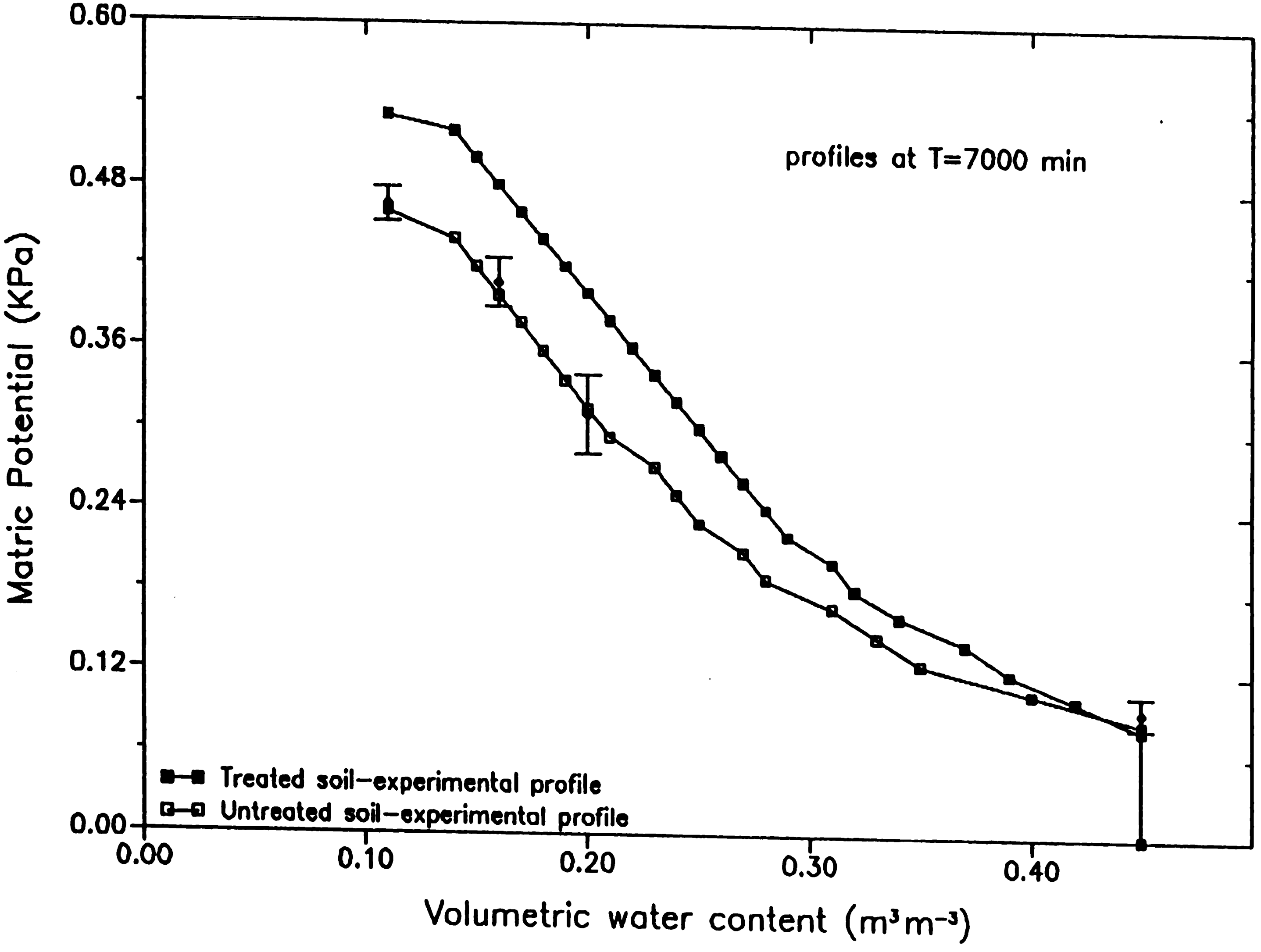

Fig. 27. Experimental wetting curves for treated and untreated soils.

treated soil was wetted to higher elevations than the untreated one over the entire moisture range with the exception of saturation. The difference in the wetting curves may be explained by stating that the treated soil held more water at any given potential. It is known from the literature (Sunderman, undated) that when a wetting agent is used on a hydrophobic soil the soil wets to a higher elevation. The bound analyses shown in Figure 27 suggests that the treated wetting curve is significantly higher than the untreated one although the bound analyses is based on samples of much higher bulk density (1.32 Mg m^{-3}). The bulk densities of the treated and untreated soil in Figure 27 were 1.15 and 1.16 Mg m^{-3}, respectively. The low bulk densities greatly increase the probability of discontinuity formation and reinforces the fact that treatment effect did occur.

The derivatives ($d\theta/dZ$) of the experimental profiles from Figure 27 are calculated and plotted in Figure 28. The resulting gradients are larger for the treated than for the untreated soil indicating that the treated soil had a stronger driving force ($d\theta/dZ$) to move water into soil as described by Equation [6]. This also enabled water to be adsorbed to higher elevations. Figure 28 demonstrates that the driving force for the treated soil was greater over the entire moisture range. The water in the treated soil experienced a stronger driving force ($d\theta/dZ$) throughout the soil profile in spite of the higher moisture content for a given potential. When thermal gradients do not exist, the water content gradient can be interpreted to be the upward driving force although countered by gravity in the case of capillary rise. The contribution of other soil solution constituents is negligible in the total pressure of soil

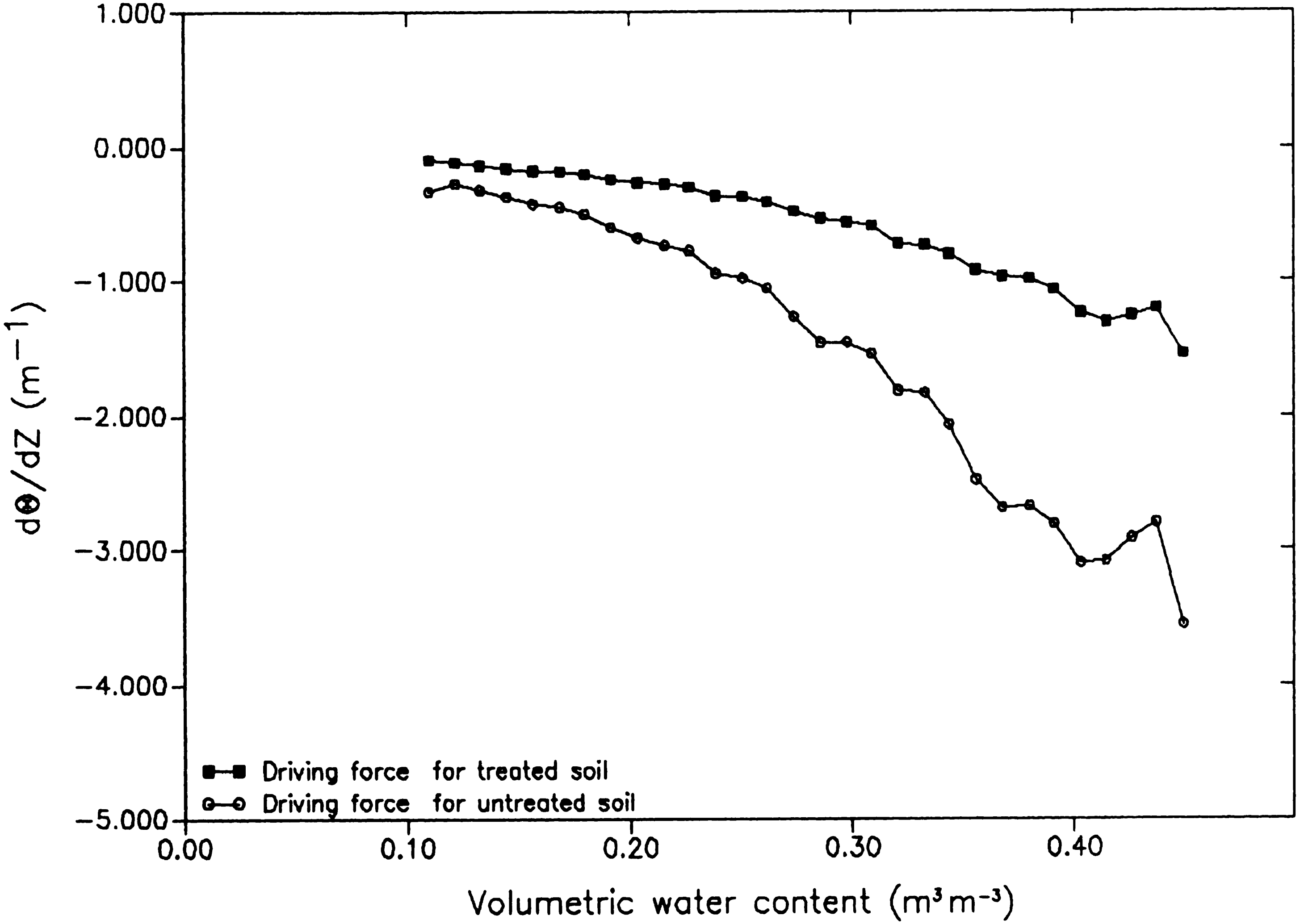

Fig. 28. The water content gradients for treated and untreated Metea sandy loam soil.

water.

Water capacity is defined as the amount of water adsorbed or released by soil per unit change of energy. The lower the water capacity (C) for a given soil the less water is adsorbed by the soil per given amount of energy and the greater the diffusivity (Equation [4]), thus exhibiting a greater flux (Equation [6]). The treated soil exhibited the lower water capacity as it is seen in Figure 29, an indication that it is able to conduct the soil water faster through its profile. The water capacity in the theta-straight region of both profiles is zero which is not shown in Figure 27.

Experimental data in Figure 30 shows the wetting front rises higher and at a faster rate with the wetting agent treated soil. This is in agreement with data from experimental and simulated profiles for all time periods, again an indication that the wetting agent has influenced the contact angle between the water and the soil matrix. Wadsworth (1931) suggested that from the plotting of log capillary rise-log time curves, two processes may be observed. The first, where soil eventually becomes saturated, is suggested by the straight line part on the left hand side of the curve in Figure 30. The straight-line part suggests that the wetting front moved as a function of the square root of time. The second process dominates at large time with a much slower rate of wetting front movement than observed initially. Both the treated and the untreated soils have curves with the same shape. A change in the rate of adsorption took place at the inflection point of the curves in Figure 30 which is close to 15 cm and 80 min resulting in a change of slope at this point. A change in the kinetics of the adsorption reaction can be

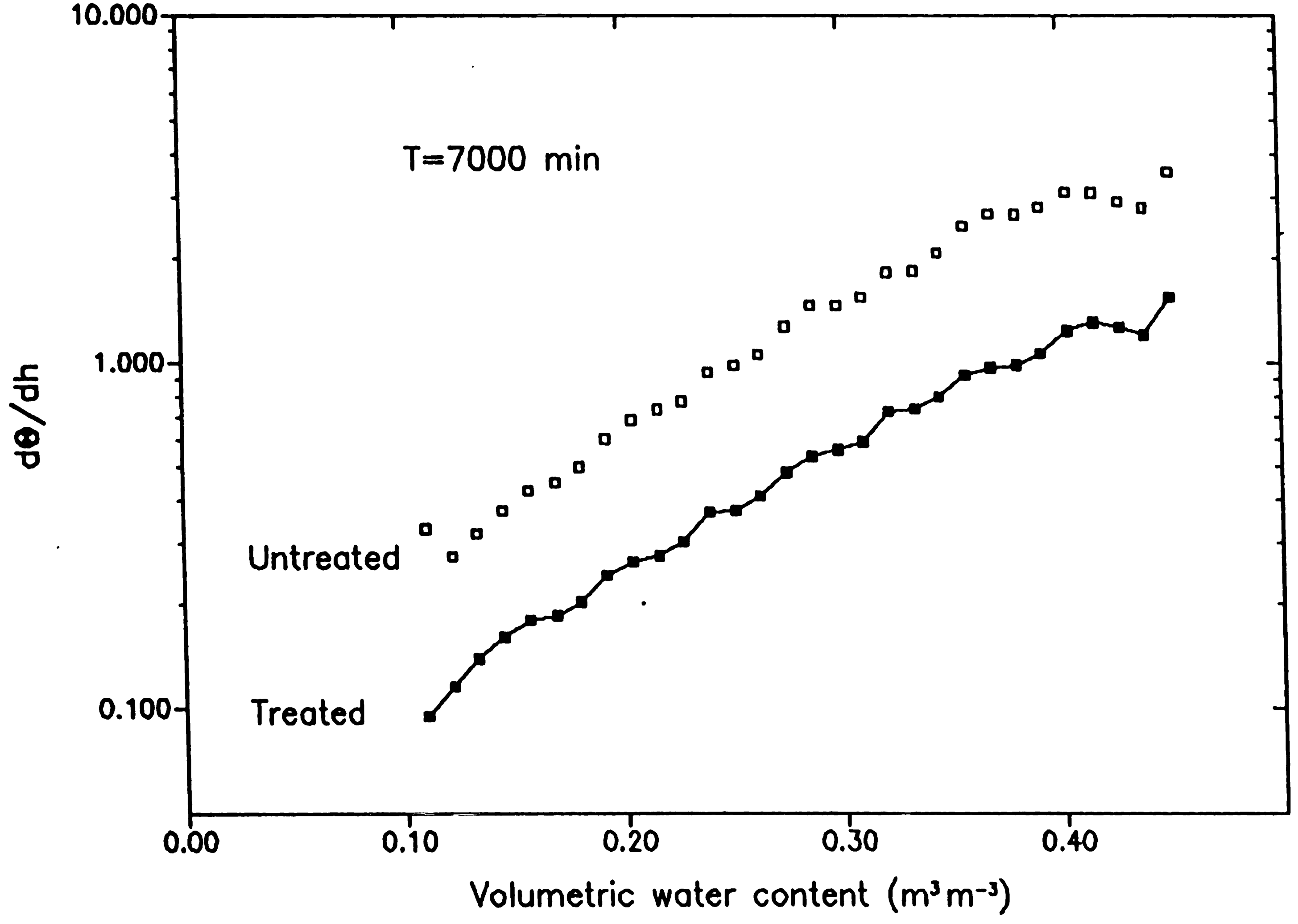

Fig. 29. The change of water capacity with water content for treated and untreated Metea sandy loam soil.

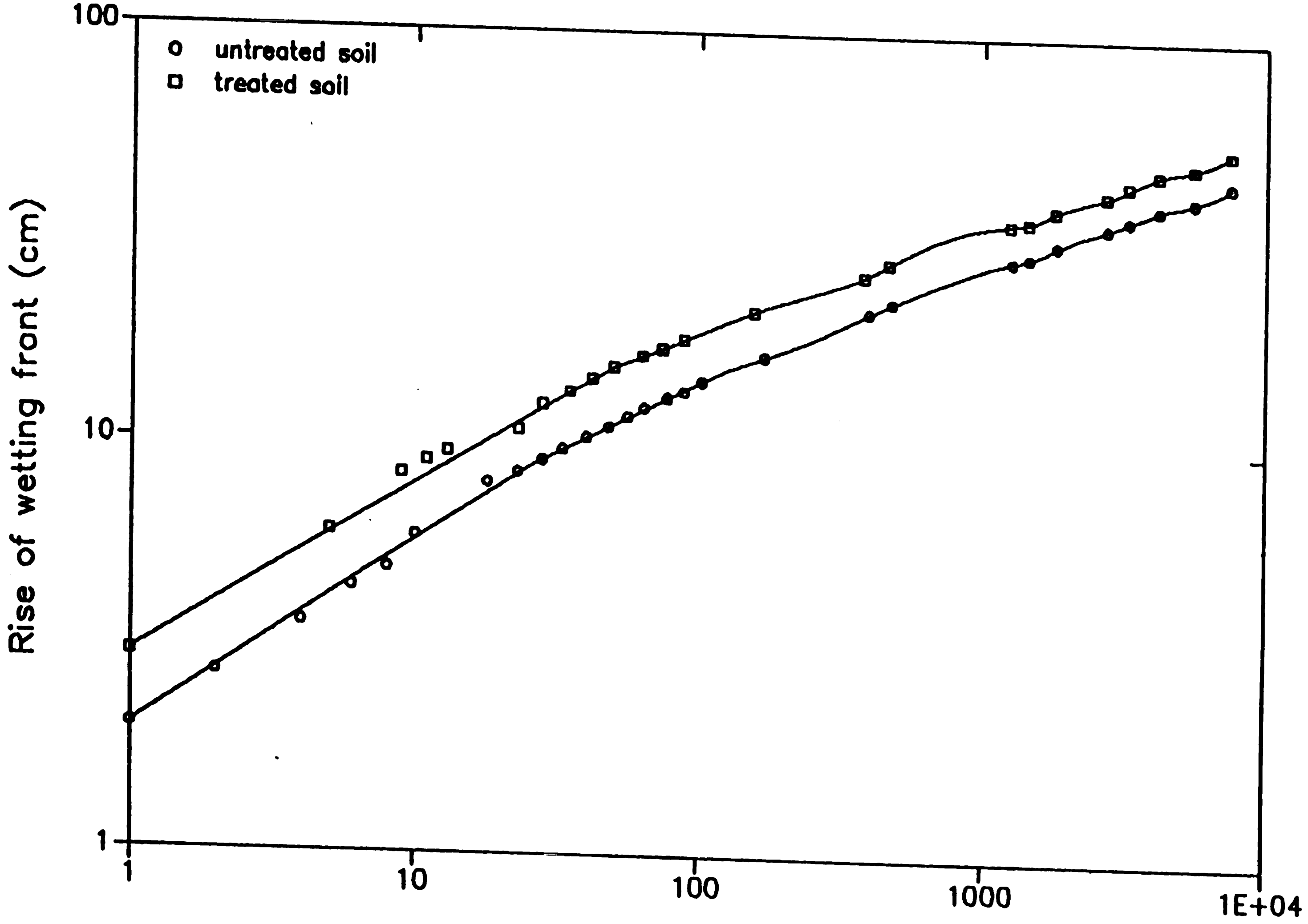

Fig. 30. Experimental wetting front–time curves for treated and untreated soils.

inferred from this change of slope. It is reasonable to assume that there exists a critical energy level-pore size relationship, dependent upon the wetting agent used and the particle size distribution of a given soil. Energy reductions beyond this threshold level caused the capillary rise process to slow down significantly. This agrees well with the changes in the slope of pF curves which were observed in the first chapter where equilibrium data suggested three different zones. Figure 30 supports the existence of different macroscopic flow characteristics in two zones. In the first zone the flow is associated with the linear part of the curve in Figure 30. In the second zone, flow is significantly slowed down as shown by the parabolic curve of the upper arm. In the final stage the wetting process stopped completely after 7000 minutes. From a practical point of view, whenever prediction of upward water flow has to be made, knowledge of the inflection point or the height of the theta-straight region can be very useful.

Figure 31 supports the higher rise of the wetting front for the treated soil illustrated in Figure 30 by showing that water uptake was larger for the wetting agent treated soil. Differences in cumulative infiltration became evident as early as 5 minutes after the capillary rise was initiated. Thereafter the cumulative infiltration differences between the treated and untreated soil continued to increase until equilibrium was achieved.

Diffusivity functions were generated by several different methods as shown in Figure 32 to demonstrate the capability of the different techniques for calculating this parameter. Diffusivity functions generated with Equation [10] compared well with diffusivity functions

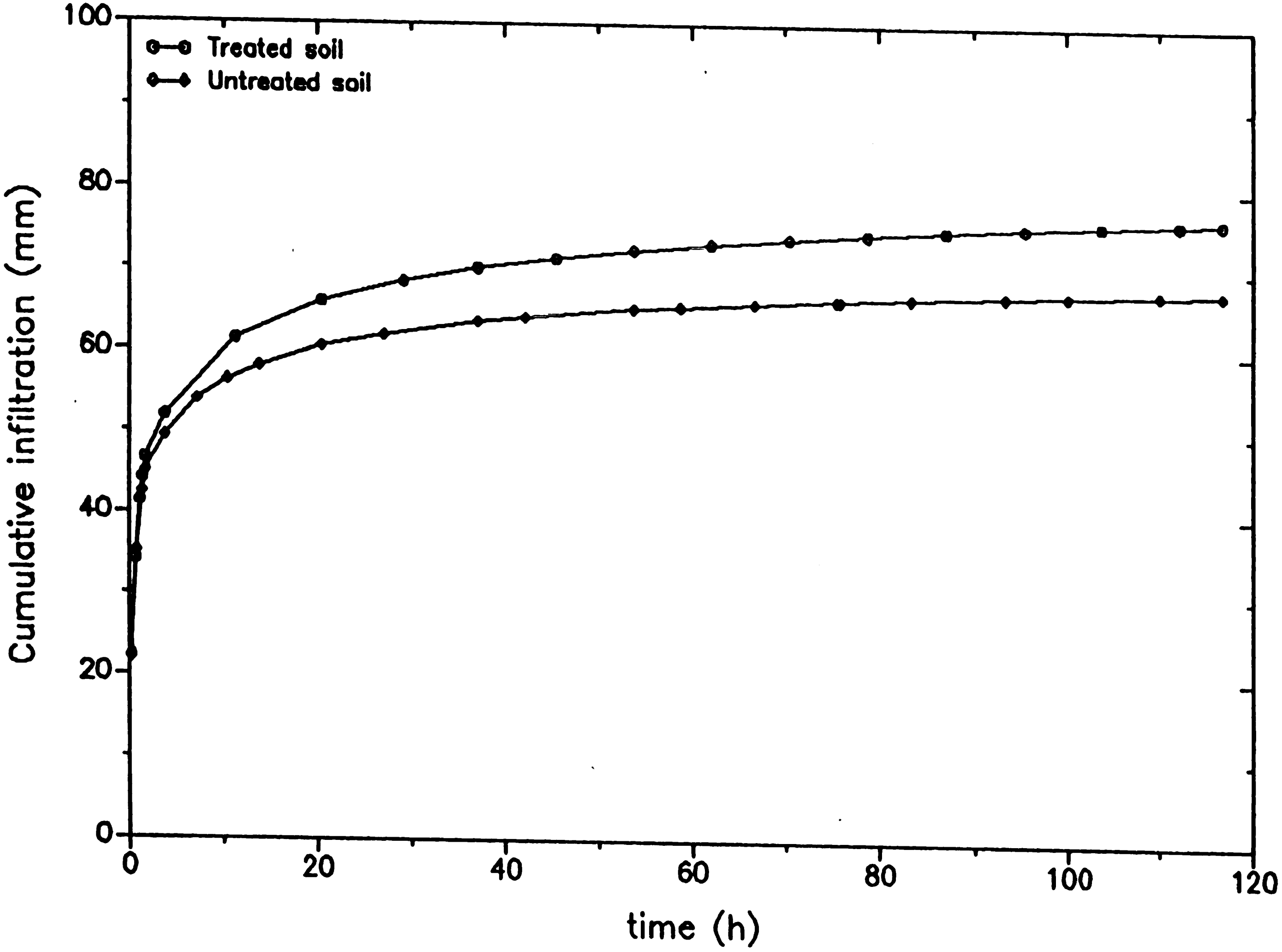

Fig. 31. The time dependence of cumulative infiltration for treated and untreated soils.

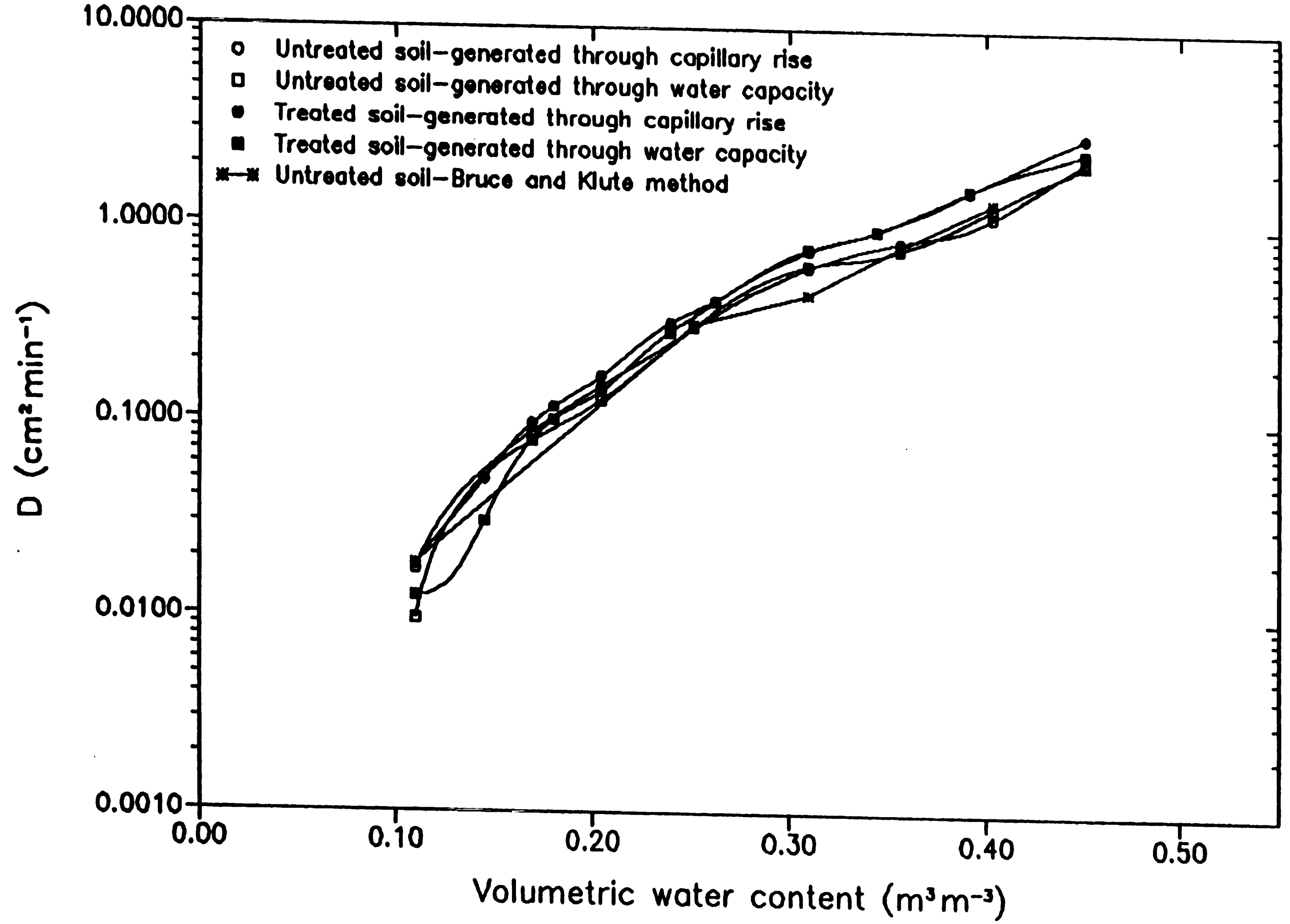

Fig. 32. Comparison of diffusivity functions generated using calculated water capacity and water flow experiments in horizontal and vertical soil columns.

obtained by 1) fitting experimental capillary rise data as described in Chapter 1 and 2) Bruce and Klute's horizontal transient method. Hence, any of these methods appear to be acceptable in generating the $D(\theta)$ function.

It is known from classical physics that physical parameters may be characterized in terms of: 1) capacity, 2) intensity, 3) proportionality coefficient and 4) flux. In heat, these factors are heat capacity, temperature, thermoconductivity and heat flux, respectively. In case of soil water the capacity factor is the water capacity $C(\theta)$, the change in water content per unit change of potential, the intensity factor is the water potential or the energy state of the soil, proportionality constants are the diffusivity and conductivity functions and flux is the water velocity. From the wetting characteristic curves obtained through capillary rise experiments, the specific water capacity, the intensity factor, h, the proportionality constants and the flow velocity were successfully obtained. The importance of these parameters can be recognized in Equations [2] and [6] in addition to characterizing the physical properties of a given soil.

4. SUMMARY

The hypothesis that the use of wetting agent can influence the flow of water into the soil matrix was tested as a case study of capillary rise. The treated soil had a higher cumulative infiltration and held more water at a given potential as shown from the experimental and simulated adsorption characteristic curves. It also conducted water faster because of the presence of stronger adsorption forces resulting from decreased surface tension.

A change in the kinetics of capillary rise was observed which resulted in a significant slow down of the process. Water gradients serve as the upward driving force but are countered by gravity. When equilibrium is attained, these are equal but of opposite sign. The wetting curves obtained from capillary rise can be used to determine capacity and intensity factors in addition to proportionality coefficients and flux factors of soil water.

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APPENDIX

TABLE 4. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 1.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
16.000	0.420
18.000	0.410
20.000	0.392
22.000	0.370
24.000	0.360
26.000	0.350
28.000	0.337
30.000	0.331
32.000	0.292
34.000	0.289
36.000	0.281
38.000	0.279
40.000	0.277
42.000	0.275
44.000	0.255
46.000	0.251
48.000	0.250
50.000	0.240
52.000	0.227
54.000	0.223
56.000	0.220
58.000	0.214
59.000	0.212
60.000	0.210
61.000	0.205
62.000	0.200
63.000	0.195
64.000	0.190
65.000	0.185
66.000	0.180
67.000	0.178
68.000	0.173
69.000	0.165
70.000	0.160
71.000	0.158
72.000	0.150
73.000	0.130
	0.095**

* Average bulk density=1.386 Mg m^{-3}; No of drops=100

** Initial moisture content

TABLE 5. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 2.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
10.000	0.410
12.000	0.370
14.000	0.360
16.000	0.340
18.000	0.330
20.000	0.320
22.000	0.310
24.000	0.280
26.000	0.270
28.000	0.260
30.000	0.250
32.000	0.240
34.000	0.230
36.000	0.210
38.000	0.200
40.000	0.190
42.000	0.170
44.000	0.160
46.000	0.140
47.000	0.120
	0.095**

* Average bulk density=1.340 Mg m^{-3}; No of drops=100

** Initial moisture content content

TABLE 6. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 3.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
8.000	0.420
10.000	0.410
12.000	0.400
14.000	0.380
16.000	0.350
18.000	0.340
20.000	0.330
22.000	0.310
24.000	0.300
26.000	0.290
28.000	0.270
30.000	0.260
32.000	0.250
34.000	0.230
36.000	0.220
38.000	0.210
40.000	0.200
42.000	0.190
44.000	0.180
46.000	0.160
47.000	0.150
48.000	0.120
	0.095**

* Average bulk density=1.345 Mg m^{-3}; No of drops=100

** Initial moisture content

TABLE 7. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 4.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
8.090	0.420
10.210	0.410
12.200	0.400
14.180	0.380
16.190	0.370
18.170	0.360
20.160	0.340
22.160	0.330
24.140	0.320
26.130	0.310
28.110	0.290
30.080	0.270
32.080	0.260
34.090	0.250
36.090	0.240
38.080	0.230
40.100	0.220
42.100	0.210
44.110	0.200
46.130	0.190
48.180	0.110
	0.095**

* Average bulk density=1.432 Mg m^{-3}; No of drops=150

** Initial moisture content

TABLE 8. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 5.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
10.450	0.420
12.560	0.410
14.660	0.380
16.760	0.360
18.870	0.340
20.990	0.330
23.110	0.320
25.260	0.300
27.340	0.290
29.420	0.280
31.550	0.270
31.610	0.260
35.670	0.240
37.800	0.230
39.880	0.220
41.920	0.210
43.990	0.200
46.090	0.190
48.220	0.180
50.360	0.160
51.370	0.110
	0.095**

* Average bulk density=1.406 Mg m^{-3}; No of drops=150

** Initial moisture content

TABLE 9. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 6.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
10.250	0.440
12.320	0.410
14.420	0.390
16.470	0.370
18.570	0.350
20.670	0.330
22.700	0.320
24.800	0.300
26.920	0.290
29.050	0.280
31.190	0.270
33.270	0.260
35.360	0.250
37.480	0.240
39.540	0.238
41.610	0.230
43.730	0.220
45.810	0.210
47.860	0.200
49.920	0.190
52.030	0.180
54.150	0.160
56.290	0.150
57.310	0.120
	0.095**

* Average bulk density=1.332 Mg m^{-3}; No of drops=100
** Initial moisture content

TABLE 10. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 7.*

Elevation (h) (cm)	Water content (θ) ($m^3\ m^{-3}$)
10.150	0.440
12.200	0.400
14.160	0.390
16.290	0.370
18.270	0.350
20.260	0.330
22.260	0.320
24.240	0.310
26.230	0.300
28.230	0.290
30.220	0.280
32.200	0.270
34.190	0.260
36.150	0.250
38.150	0.240
40.160	0.230
42.160	0.220
44.150	0.210
46.170	0.200
48.180	0.190
50.180	0.180
52.200	0.170
54.250	0.160
55.340	0.150
56.470	0.140
57.500	0.130
58.000	0.110
	0.095**

* Average bulk density=1.331 Mg m^{-3}; No of drops=100

** Initial moisture content

TABLE 11*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 10.**

Elevation (h) cm	Water content (θ) $m^3 m^{-3}$
6.010	0.434
8.010	0.292
10.030	0.248
11.990	0.247
13.970	0.246
15.970	0.232
18.000	0.205
19.930	0.202
21.950	0.194
23.960	0.188
25.840	0.172
27.830	0.161
29.790	0.148
31.710	0.126
32.000	0.120
	0.095***

* Average bulk density=1.200 Mg m^{-3}; No of drops=150
** Soil particle diameter greater than 1 mm
*** Initial moisture content

TABLE 12*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 11.**

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
6.070	0.442
8.030	0.279
9.960	0.267
11.950	0.243
13.920	0.244
15.900	0.234
17.880	0.220
19.890	0.209
21.890	0.184
23.900	0.182
25.920	0.183
27.930	0.174
29.970	0.156
31.990	0.129
32.500	0.120
	0.095***

* Average bulk density=1.230 Mg m^{-3}; No of drops=150

** Soil particle diameter greater than 1 mm

*** Initial moisture content

TABLE 13. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 8.*

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
8.490	0.450
10.550	0.400
12.660	0.350
14.700	0.330
16.800	0.310
18.910	0.280
20.930	0.270
23.030	0.250
25.160	0.240
27.280	0.230
29.420	0.210
31.510	0.200
33.590	0.190
35.710	0.180
37.780	0.170
39.840	0.160
41.960	0.150
44.050	0.140
46.090	0.110
	0.095**

* Average bulk density=1.160 Mg m^{-3}; No of drops=50

** Initial moisture content

TABLE 14*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 9.**

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
8.030	0.450
10.080	0.420
12.040	0.390
14.160	0.370
16.150	0.340
18.130	0.320
20.140	0.310
22.120	0.290
24.110	0.280
26.110	0.270
28.090	0.260
30.080	0.250
32.060	0.240
34.030	0.230
36.030	0.220
38.040	0.210
40.040	0.200
42.020	0.190
44.050	0.180
46.050	0.170
48.060	0.160
50.080	0.150
52.120	0.140
53.220	0.110
	0.095***

* Average bulk density=1.150 Mg m^{-3}; No of drops=50

** Soil which was treated with wetting agent

*** Initial moisture content

TABLE 15*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 12.**

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
5.973	0.461
7.938	0.382
9.862	0.360
11.857	0.321
13.821	0.303
15.806	0.269
17.790	0.262
19.794	0.253
21.798	0.238
23.803	0.228
25.827	0.217
27.831	0.207
29.875	0.199
31.899	0.187
33.883	0.184
35.848	0.178
37.832	0.173
39.816	0.165
41.801	0.156
43.825	0.143
45.829	0.133
47.833	0.123
49.857	0.102
	0.095***

* Average bulk density=1.340 Mg m^{-3}; No of drops=50

** Soil which had gone through one cycle of wetting and drying in the cylinder

*** Initial moisture content

TABLE 16*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 13.**

Elevation (h) cm	Water content (θ) m^3 m^{-3}
6.013	0.467
8.017	0.396
10.031	0.349
11.996	0.321
13.980	0.300
15.974	0.280
18.008	0.249
19.933	0.251
21.957	0.236
23.961	0.223
25.846	0.222
27.831	0.202
29.795	0.198
31.710	0.195
33.685	0.187
35.639	0.182
37.624	0.177
39.688	0.168
41.712	0.164
43.736	0.154
45.829	0.143
47.833	0.135
	0.095***

* Average bulk density=1.354 Mg m^{-3}; No of drops=50

** Soil which had gone through one cycle of wetting and drying in the cylinder

*** Initial moisture content

TABLE 17*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 14.**

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
3.949	0.491
6.052	0.399
8.037	0.380
10.081	0.327
12.045	0.317
14.168	0.288
16.153	0.268
18.137	0.254
20.141	0.247
22.126	0.231
24.110	0.226
26.095	0.224
28.079	0.205
30.083	0.201
32.068	0.196
34.032	0.190
36.036	0.180
38.040	0.172
40.045	0.166
42.029	0.148
44.053	0.146
46.057	0.126
47.000	0.120
	0.095***

* Average bulk density=1.370 Mg m^{-3}; No of drops=50

** Soil which had gone through one cycle of wetting and drying in the cylinder

*** Initial moisture content

TABLE 18*. Adsorption soil water content-elevation data for Metea sandy loam soil-Experiment 15.**

Elevation (h) cm	Water content (θ) $m^3\ m^{-3}$
4.247	0.446
6.390	0.403
8.493	0.375
10.557	0.320
12.660	0.314
14.704	0.283
16.808	0.264
18.911	0.246
20.935	0.245
23.039	0.226
25.162	0.217
27.285	0.208
29.428	0.202
31.512	0.187
33.595	0.183
35.719	0.173
37.782	0.171
39.846	0.162
41.969	0.149
43.954	0.121
	0.095***

* Average bulk density-1.354 Mg m^{-3}; No of drops-50

** Soil which had gone through one cycle of wetting and drying in the cylinder

*** Initial moisture content

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