VARIATION OF SATURATED HYDRAULIC CONDUCTIVITY
WITH DEPTH FOR SELECTED PROFILES OF
TILLMAN-HOLLISTER SOIL

by

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Niamey, Niger

1974

Submitted to the Faculty of the Graduate College
of the Oklahoma State University
in partial fulfillment of the requirements
for the Degree of
MASTER OF SCIENCE
May, 1977

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ACKNOWLEDGEMENTS

The author wishes to express his sincere thanks to Dr. David L. Nofziger, his Major Advisor, for his guidance and advice throughout the research and academic studies. He also wishes to thank Dr. L. I. Croy and Dr. J. E. Garton for serving on his advisory committee.

Appreciation and thanks are also due to Mr. Harold Gray and Paul D. Kruska for their kind assistance during the study.

Thanks are also expressed to Dr. J. F. Stone and Dr. E. W. ChinChoy for their suggestions in the course of the research.

The author wishes to express appreciation to his parents Mr. and Mrs. I. M. Ouattara and to his wife Sahadatou for their encouragement and patience during his stay in the United States.

Acknowledgement is due to the Agency for International Development for financial support.

Special thanks are extended to Ms. Holesko for her help in the preparation of this manuscript.

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CHAPTER I

INTRODUCTION

Increasing areas of soil in arid and semi-arid regions are being irrigated. To prevent soil salinity and alkalinity problems some water applied to the surface must move through the soil and carry the excess salts out of the root zone. Also high water tables must be prevented. Near Altus, Oklahoma, large areas of Tillman-Hollister clay loam soils have been put under irrigation. Irrigation water of rather low quality is provided by the Lugert Lake. Portions of this irrigated land have gone out of production due to high salt concentrations. Studies are being conducted to determine the origin and magnitude of the salt problem. Ground-water movement is being studied as a first step in developing methods to prevent the development of salinity problems and to reclaim unproductive areas.

The hydraulic properties of the soil and the nature and existence of stratification within the soil are very important in understanding the movement of water in the soil. Ground water under artesian pressure implies soil stratification exists. This study was an attempt to investigate this stratification. The specific objectives of this study were:

1. to assess the validity of Darcy's equation for describing water movement in this clay loam soil.

2. to determine the saturated hydraulic conductivity of this soil at different depths in the soil profile so as to ascertain the existence and location of any stratification.

This information may provide insight into the processes of formation of saline areas in the region and will help in the design of irrigation and drainage systems.

CHAPTER II

LITERATURE REVIEW

Darcy's Law

Water movement in soil takes place under the influences of diverse forces caused by differences in pressure, osmotic, gravitational, thermal, and electrical potentials within the soil. In most soil-water systems, differences in pressure and gravitational potentials are much greater than differences in osmotic, thermal and electrical potentials. Therefore, the influence of osmotic, thermal, and electrical potentials on water flow is commonly assumed to be negligible. The gravitational and pressure potentials frequently are combined and called the hydraulic head.

An early study by Darcy published in 1856 constitutes the basis for most mathematical models of water flow in soil. Darcy studied sand columns in which water was flowing downward at different rates (Hubert, 1956). For each rate the difference in hydraulic head between the inlet and outlet boundaries was measured. Darcy found the volume flow of water per unit time was proportional to the cross-sectional area of the sand column and to the difference in hydraulic head. He also found the flow rate was inversely proportional to the length of the column. Mathematically the findings can be expressed as:

$$Q = -KA\frac{h_2 - h_1}{L} \qquad (1)$$

where Q is the flux of water or the volume of water flowing through the column per unit time,

K is a constant of proportionality and is called the saturated hydraulic conductivity,

A is the cross-sectional area of the column,

h_1 is the hydraulic head at the inlet boundary,

h_2 is the hydraulic head at the outlet boundary, and

L is the length of the sand column.

Equation (1) is frequently written in the form:

$$v = Ki \qquad (2)$$

where $v = Q/A$ is the flux density or the volume of water flowing through the column per unit time per unit area, and

$i = -(h_2-h_1)/L$ is the gradient in hydraulic head.

The hydraulic conductivity K in Darcy's equation received much attention because of its importance for characterizing the behavior of water moving into a saturated porous medium. The saturated hydraulic conductivity has been described by the International Society of Soil Science (1976) as "the constant of proportionality between the flux density and the total driving force in Darcy's law". It may also be defined as the flux density caused by a unit driving force. For a given porous medium, the hydraulic conductivity will depend on texture, structure, and porosity of the porous medium. Cracks, worm holes, and decayed root channels will also affect the saturated hydraulic conductivity of a soil. The hydraulic conductivity was found to be proportional to the density of the fluid flowing through the porous medium and inversely proportional to the viscosity of the fluid. If one assumes that there is no interaction between the fluid and the porous medium the saturated hydraulic conductivity can be expressed as:

$$K = \frac{\rho g}{\mu} k \tag{3}$$

where g is the acceleration of gravity,

ρ is the density of the fluid,

μ is the viscosity of the fluid, and

k is the intrinsic permeability of the porous medium.

The intrinsic permeability reflects the contribution of the porous medium to the saturated hydraulic conductivity.

Darcy's equation has been supported by a large body of experiments. The law has been extended and made applicable to more complex problems. Expressed in a differential form for flow in one direction, Darcy's equation can be written as:

$$v = -K\frac{dh}{ds} \tag{4}$$

where v is the flux density,

K is the saturated hydraulic conductivity,

h is the hydraulic head, and

s is the position coordinate in the direction of flow.

Slichter in 1899 (Hillel, 1971) proposed an extension of equation (4) for three dimensional flow. In this case the equation becomes:

$$v = -K\nabla h \tag{5}$$

where ∇ is the del operator of vector notation and

v, K and h have been defined above.

According to Hubert (1956) equation (5) means that

> at each point into a space where flow is taking place there must be a particular value of a scalar quantity h defined as hydraulic head. The ensemble of such values give rise to a scalar field in the quantity h. In such a field water will flow in direction perpendicular to the surface of equal potential h.

Although it is not of primary interest for this study, Darcy's equation was extended to describe water movement in unsaturated soils at the beginning of this century by Buckingham followed by Gardner and Widtsoe in 1921 and Richards in 1931 (Gardner, 1972; Hillel, 1971).

There are some limitations to the application of Darcy's equation. Inertial forces must be negligible compared to viscous forces before Darcy's equation is valid (Hubert, 1956). At high flow velocities inertial forces become predominant. This causes turbulence and distortion in the flow lines and the flux increases less than proportionally to the gradient in hydraulic head. Klute (1965) reported that most of the gradients generally encountered in nature fall in the range where Darcy's equation is valid for silts and finer textured soils. However, in sands and gravels it may be necessary to restrict hydraulic gradients to values equal to or less than one for Darcy's equation to apply. Deviations from Darcy's equation at low gradients for some porous media have been reported (Swartzendruber, 1962). The causes of such deviations are not yet understood.

Methods of Measuring the Hydraulic Conductivity

Several methods have been used for measuring the hydraulic conductivity of a soil. These methods can be classified as laboratory methods or field (in situ) methods.

<u>Laboratory methods</u>. Laboratory methods involve collecting soil samples, taking them to the laboratory and measuring their hydraulic conductivity. Samples used may be disturbed or undisturbed. However, because disturbance may change the hydraulic characteristics of a soil, undisturbed samples are usually preferred. Constant-head and falling-head methods have been used for measuring the hydraulic conductivity of samples brought to the laboratory. In the constant-head method, hydraulic heads of water are maintained at the inlet and at the outlet boundaries. The volume of water flowing through the soil per unit time is determined either by measuring the inflow at the inlet or the outflow at the outlet as a function of time. In the falling-head method the difference in head between the inlet and outlet is allowed to change with time. Laboratory methods are well suited for detecting stratifications in soil and for measuring the hydraulic conductivity for truly one-dimensional flow.

Different types of apparatus may be used for measuring the hydraulic conductivity of undisturbed soils in the laboratory. Wit (1967) proposed a permeameter complying with both constant- and falling-head methods and which can be used to measure either vertical or horizontal hydraulic conductivity for a great number of samples at one time. Childs and Poulovassilis (1960) developed a method based on the principle of oscillating hydraulic head which causes a limited volume of flow to oscillate to and fro through the soil. The advantage of their method is that the same quantity of water is flowing all the time through the soil. McNeal and Reeve (1964) proposed an apparatus for reducing the effect of boundary flow between the permeameter wall and the soil core. Strain gages and other pressure transducers have been used for measuring very small hydraulic conductivities (Overman, et al., 1968; Nightingale and Bianchi, 1970).

Field methods. Field methods are in situ methods which use the rate of flow of water in some special conditions to estimate the hydraulic conductivity of the soil. Field methods for measuring hydraulic conductivity may be divided into methods requiring a water table and methods applicable in abscence of a water table.

For measuring hydraulic conductivity below a water table, the auger method and piezometer method are often used. The auger method involves drilling a hole below the water table with the minimum of disturbance. The rate of rise of water level in the cylindrical hole is measured. Several equations, empirical and theorotical, have been suggested for calculating the saturated hydraulic conductivity from the basic measurements (Luthin, 1957; Boersma, 1965a). The piezometer method proposed by Kirkham in 1946 (Luthin, 1957) has been designed for stratified soils. It consists of a metal pipe or tube installed vertically into the soil. The rate of flow of water into the lower end of the tube is measured. A formula taking into account the radius of the tube, the level of water in the tube at two different times, and a parameter for the shape of the cavity is used for calculating the hydraulic conductivity. Because it is difficult to determine the area and geometric shape of flow into the auger or piezometer hole, the accuracy of these methods is limited.

Where a water table is permanently or temporarily absent, other in situ methods have been proposed for measuring hydraulic conductivity. The double-tube method proposed by Bouwer (1961) consists of saturating a limited soil region below an auger hole in which two concentric tubes are placed. Hydraulic conductivity is calculated from measurements of the rate of change of the water level in the inner tube. Dimensionless parameters that occur in the equations are determined from a resistance network analog. Other methods used for measuring hydraulic conductivity above a water table include the shallow-well pump-in method and the permeameter method. The shallow-well method consists of measuring the rate of water intake from a lined or unlined auger hole while a constant head of water is maintained in the hole (Boersma, 1965b). The permeameter method is based on the rate of outflow of water from a cylinder placed into a hole. It measures vertical hydraulic conductivity (Boersma, 1965b). An air entry permeameter has also been developed by Bouwer (1966). It enables relatively rapid field measurements of saturated hydraulic conductivity in initially saturated soils.

Factors Influencing the Laboratory Measured Hydraulic Conductivity

Numerous authors have reported that the saturated hydraulic conductivity of an active soil as measured in the laboratory may change continuously with time and nature of the solution used in the percolating process. Christiansen (1947), for example, describes the general characteristics of laboratory determined hydraulic conductivity as follows:

> when long time permeability tests are made on agricultural soils, characteristic permeability curves are obtained. During the first phase of the test, there is usually a decrease in permeability to a minimum somewhat below the initial rate... During the second phase permeability increases... During the third phase, the permeability again decreases, somewhat to very low rate...

The mechanisms causing this variation in hydraulic conductivity are not completely understood. Three mechanisms which have been suggested are the progressive deterioration of soil aggregation due to interaction between the soil and the flowing solution, the action of soil microorganisms, and the effect of entrapped air.

Deterioration of soil aggregation. Although some points still need clarification, this aspect of the problem has been studied in greatest detail. The problem of soil aggregate deterioration due to cations present in flowing solution is often associated with saline and alkali soils. Christiansen (1947) defined saline soils as soils containing soluble salts in concentrations as to affect plant growth (more than 0.2% approximately) and he defined alkali soils as sodic soils having a pH in excess of 8.5 and generally dispersed with very low permeabilities. Numerous studies on hydraulic conductivity of saline and alkali soils showed that the conductivity cannot be dissociated from the salt content and the cation exchange complex of the soil and the electrolyte content of the percolating water. Some saline soils tend to swell and disperse to some extent as salts are removed and this results in a reduction in the effective size of the pores. Reeve and Tamaddoni (1965) stated, for example, that "where low-salt content waters are used for reclaiming sodic soils, even with the application of ample calcium amendments, reclamation often proceeds so slowly as to be impractical, once the soil has developed low permeability as a result of the dispersion effect of low electrolyte water". However, even at low salt concentrations, it is generally accepted that the presence of divalent ions such as Ca and Mg on the ion-complex of the soil frequently stabilizes or increases soil hydraulic conductivity.

There is some controversy on the limit above which the adverse effects of sodium become evident. McNeal, et al. (1966), reported that for most agricultural soils, exchangeable sodium percentage (ESP) values of 15 or greater can generally be tolerated without serious reductions in hydraulic conductivity, provided the percolating solution concentration exceeds 3 mg/liter. The nature of the soil, however, has an important effect on the interaction between exchangeable sodium, electrolyte, and hydraulic conductivity.

Usually high interaction is reported with clays. Rhoades and Ingvalson (1969) stated that vermiculites appear to be capable of withstanding higher ESP values and lower electrolyte concentrations before their hydraulic conductivity becomes appreciably reduced than montmorillonitic soils. The interaction between electrolyte concentrations, exchangeable sodium content, and hydraulic conductivity is also considerably dependent on the structural condition of the soil and the organic matter content.

In order to minimize the decrease in hydraulic conductivity due to the interaction between the percolating water and exchangeable sodium content of the soil, several authors have proposed the use of a solution with a high salt concentration. This suggestion is based on a large body of experimental results. Fireman (1944) reported that a high and constant hydraulic conductivity was obtained when using a 800 ppm $CaCl_2$ solution while the hydraulic conductivity decreased sharply when distilled water was used. Mojallali and Dregne (1968) found that a calcareous Bruno sandy loam subsoil saturated with exchangeable sodium, potassium or ammonium became completely impermeable when leached with distilled water. Zawadzki and Olszta (1971) reported differences in saturated hydraulic conductivity when using ground water, tap water, and distilled water.

The use of too high a concentration of electrolyte generally creates problems in the measurement of hydraulic conductivity. In presence of electrolyte solutions, the soil mass sometimes contracts and gives way to leakage between the permeameter wall and the soil. This problem has been emphasized by Reeve and Tamaddoni (1965) and McNeal and Reeve (1964).

Effect of soil microorganisms. The development of microorganisms in soils has been suggested as one reason for the variation of hydraulic conductivity with time. Allison (1947) reported that hydraulic conductivity changes may be due to biological clogging of soil pores with microbial cells and their synthetized products, slimes or polysaccharides. The more recent work of Gupta and Swartzendruber (1962) also strongly suggested the development of soil microorganisms may be an important factor in hydraulic conductivity changes with time, particularly in sandy soils where electrolyte leaching of soil cations does not have a great importance. Gupta and Swartzendruber (1962) found that the microbial effect on hydraulic conductivity seemed to be correlated to the bacteria number in the soil. However, the manner in which microorganisms may decrease hydraulic conductivity is not understood. The small volume they occupy (even when they are in great numbers) is extremely small compared to the volume of soil pores.

Effect of entrapped air. Van Schaik and Laliberté (1969) compared different techniques of saturating soils and found the saturated hydraulic conductivity was notably less for samples saturated at atmospheric pressure than for samples saturated in a vacuum. They attributed this effect to entrapped air in the soil and explained that "when a porous medium imbibes a wetting fluid, air bubbles become entrapped in isolated regions within the pore spaces of the sample, so that part of the pore volume is unattainable by the wetting fluid". Gupta and Swartzendruber (1964) reported that Wyckoff and Botset suggested that flowing water containing dissolved gas can release some gas which accumulates in the porous medium and reduces hydraulic conductivity. However, Gupta and Swartzendruber (1964) did not find any effect due to air entrapment in their study on sand. Peck (1969) explained the entrapped air can produce significant changes in hydraulic conductivity by its change in volume. The change in volume of the entrapped air can be produced by fluctuating temperatures, external atmospheric pressure changes, liquid pressure variations and diffusion of gas through the liquid. The effect of entrapped air on laboratory determined hydraulic conductivities may be of little practical importance if Van Schaik and Laliberté (1969) are correct in suggesting that field drainage problems often involve partially saturated soils with hydraulic conductivities close to that of undisturbed soils saturated at atmospheric pressure rather than soils saturated under vacuum.

CHAPTER III

METHODS AND MATERIALS

Samples for this study were taken at the Oklahoma Agricultural Experiment Station Irrigation Research Station at Altus on one field west of the highway. The field was about 500 meters long and 180 meters wide. The soil was the Tillman-Hollister clay loam (fine, mixed, thermic-pachic Paleustoll). The profile description is given in the Appendix. The field was plowed to approximately 20 cm after the samples on Site I were taken. Therefore, samples on Site II and Site III were taken on a plowed field.

Sampling

The field was divided in three parts, an area or site was chosen from each part. The sites were about 180 meters apart. Two samples at each depth were taken at each site. The following depths were sampled: 20 cm to 30 cm, 40 cm to 50 cm, 60 cm to 70 cm, 80 cm to 90 cm, 100 cm to 110 cm, and 120 cm to 130 cm.

Undisturbed soil samples were taken using a Gidding company soil sampling machine mounted on a truck. The same stainless steel soil tube (7.62 cm x 121.92 cm) was used during the whole sampling process. Therefore, all the samples were considered to have the same diameter of 7.6 cm; the vertical length of the samples varied from 8 cm to 11.5 cm.

Handling and Casing of Samples

At the time of sampling, all the samples at one site were trimmed and immediately encased in polyolefin heat-shrinkable insulation tubing as described by Bondurant, et al. (1969). Plexiglas end-caps coated with silicone rubber seal were used to prevent breakage of the core end and also to provide connection for the water inlet and outlet tubes. After encasing the samples, ring clamps were firmly tied around the Plexiglas end-caps to prevent leakage. The samples were taken to the laboratory and the wetting process was started immediately.

Handling of undisturbed soil samples after they had been collected was very important because samples had to be brought to the laboratory without breaking and without disturbance of soil aggregation. Good encasing of samples was crucial because improper encasing permitted water to flow between the soil and the permeameter wall.

Wetting Process

A 1500 ppm calcium chloride ($CaCl_2$) solution was prepared for percolating the samples. Samples were wetted from the bottom, at atmospheric pressure and room temperature. Increasing heads of water were applied to the samples. Heads of 2 cm, 6 cm, and 10 cm of water were each applied for about 12 hours. Finally a 14 cm head was applied until water covered the top of the sample. All heads were measured from the bottom of the soil samples.

Measurement and Calculation of Hydraulic Conductivity

After the wetting period, samples were installed on a permeameter as shown in Figure 1. The inlet of the permeameter was connected to a constant head of water maintained by the use of a Mariotte bottle and the outlet was connected to an horizontal calibrated capillary tube. The position of the meniscus was recorded at different times. In this way the volume of water flowing out of the soil was precisely measured. The hydraulic conductivity value was calculated from equation (1) rewritten as:

$$K = -\frac{1}{A}\frac{QL}{h_2-h_1} \tag{6}$$

where K is the saturated hydraulic conductivity (cm/min),

Q is the volume of water flowing through the sample per unit time (cm^3/min),

L is the length of the soil column (cm),

A is the cross sectional area of the sample (cm^3),

h_2 is the hydraulic head at the outlet boundary (cm), and

h_1 is the hydraulic head at the inlet boundary (cm).

The magnitude of h_2-h_1 is represented by Δh on Figure 1.

Six permeameters were used for measuring hydraulic conductivities of samples from the six wells. Therefore, a 6 x 6 Latin square design was used. Rows represented the permeameters, columns represented the sampling wells, the depths of sampling were considered as treatments.

A special study was made to assess the validity of Darcy's equation for this soil. One soil sample was subjected to different gradients.

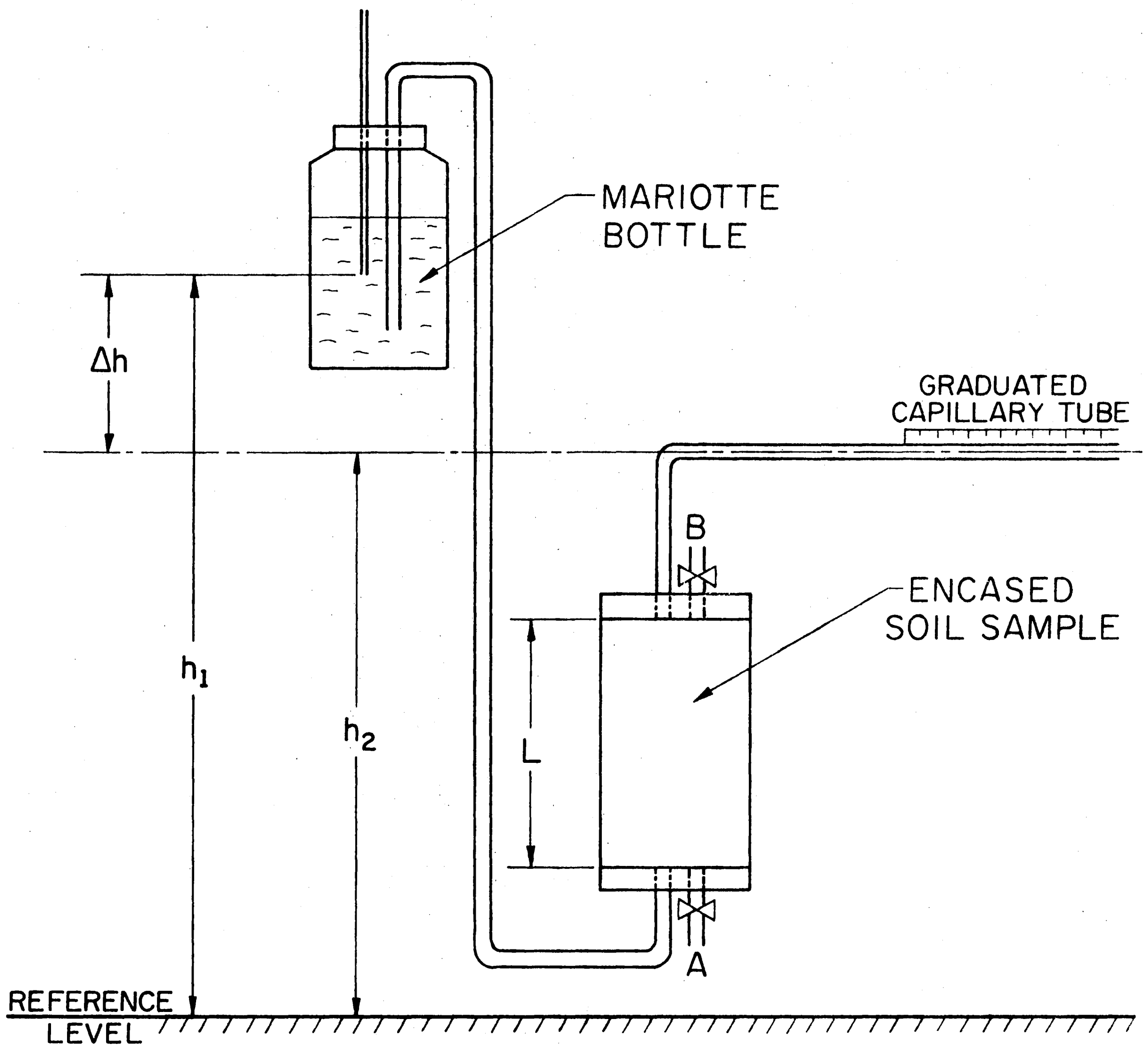

Figure 1. Permeameter Used for Measuring the Saturated Hydraulic Conductivity

The differences Δh between outlet and inlet hydraulic heads were successively given values of 8 cm, 24 cm, 40 cm, and 58.5 cm. For the sample used, these values of Δh corresponded, respectively, to gradients of 1, 3, 5, and 7.3 centimeters of water per centimeter. At each gradient the flow was measured. The validity of Darcy's equation was then determined.

CHAPTER IV

RESULTS AND DISCUSSION

Darcy's Equation Test

The validity of Darcy's equation was investigated using one soil sample from a depth of 20 cm to 30 cm. Water flux was measured and flux density was calculated for different hydraulic gradients. Gradients of 1, 3, 5, 7.3, 5, 3, 1, 3, 5, and 7.3 cm of water/cm were successively applied to the soil sample. The results shown in Figure 2 indicate the data points corresponding to a particular gradient fall very close together, except for one point corresponding to a decreasing gradient of 3. The close agreement of data for both increasing and decreasing gradients implies that time and volume of flow had little, if any, effect on the hydraulic properties of the soil during the time of measurement. The measurements were made over a period of four hours which correspond to a volume flow of 5.25 cm^3. Water had been flowing through the sample for 12 days before these measurements were begun.

Results shown in Figure 2 indicate the flux density increased more than proportionally with the gradients. This implies that Darcy's equation is not strictly valid for the soil studied. No attempt was made in this study to determine the cause of these deviations from proportionality. Similar deviations have been reported in the literature. Russel and Swartzendruber (1971) reported non-proportional

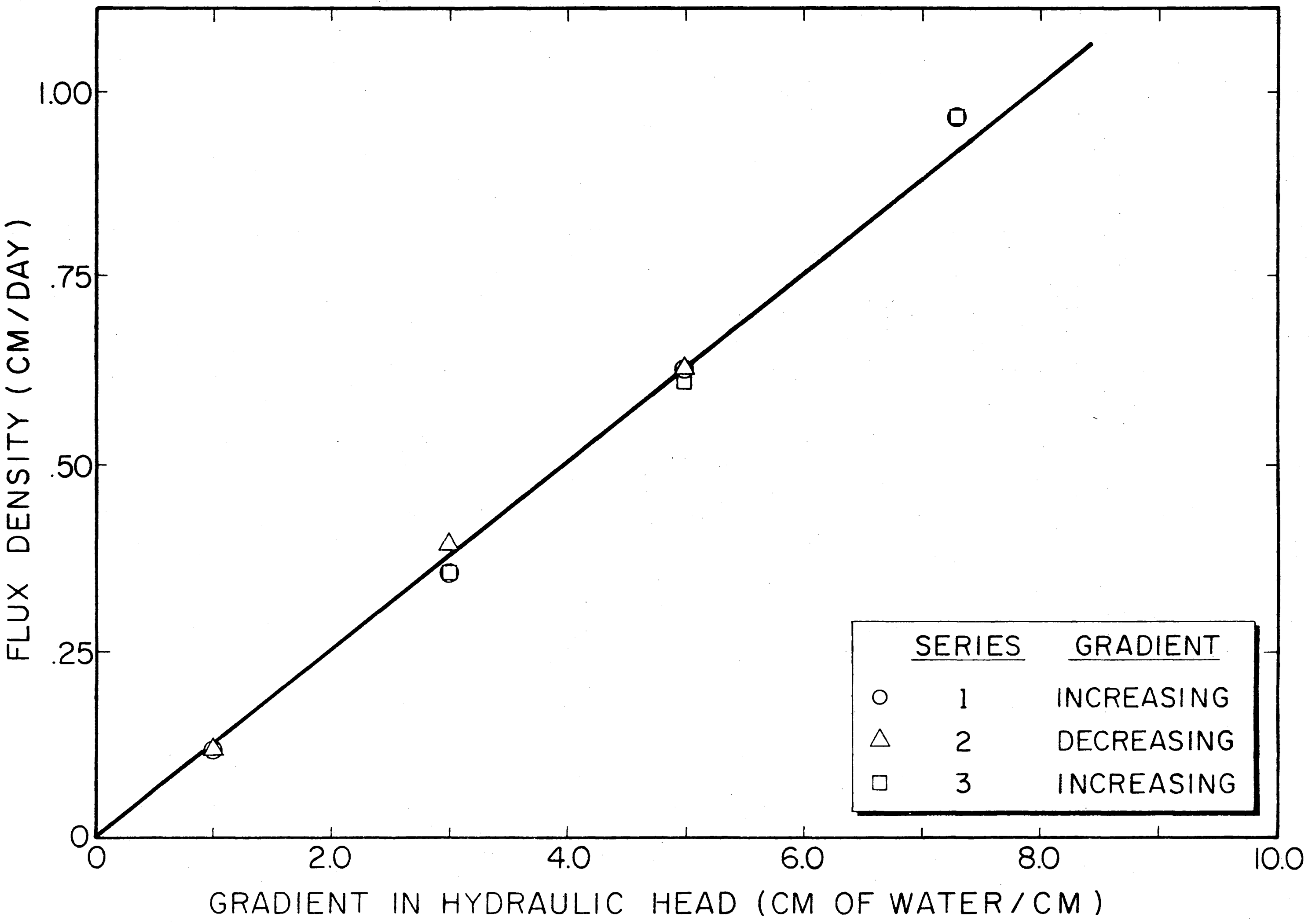

Figure 2. Flux Density Versus Gradient in Hydraulic Head

flux-gradient curves of a general sigmoidal shape when using a porous medium containing swelling clay. Miller and Low (1963) observed greater than proportional flow with a lithium-clay. Von Engelhardt and Tunn, Lutz and Kemper, Kutilek, and Paez, have all been credited with reporting greater than proportional flows by Russel and Swartzendruber (1971). Reversible reorientation of particles along the stream lines and existence of a range of pore sizes in which some threshold gradients are reached as gradients are increased have been proposed as two possible explanations of this behavior.

The slope of a curve through the data points in Figure 2 is equal to the hydraulic conductivity of the soil. If the deviations from proportional flow are neglected and hydraulic conductivity K of this soil is evaluated from the slope of the straight line shown in Figure 2, one finds that K = 0.126 cm/day. Using a curve drawn through actual data points yields hydraulic conductivity values of .119 and .132 cm/day at gradients of 1 and 7.3, respectively. This difference in hydraulic conductivity of approximately 10% is relatively small as will be seen from additional results presented in this study. From this one can conclude that Darcy's equation is a good first approximation for water movement in this soil.

Variation of Hydraulic Conductivity With Time

Variation of hydraulic conductivity with time was observed for almost all the samples studied. Table I shows some hydraulic conductivity values calculated after different periods of flow. The variation is also indicated by the change in the slope of the curve obtained when

TABLE I

SATURATED HYDRAULIC CONDUCTIVITY VALUES (CM/DAY) AFTER DIFFERENT DURATIONS OF FLOW (SITE III, SET 1)

Depth (cm)	Cumulative Time of Flow (Min) 24	512	1625
20–30	.16	.15	.09
60–70	.11	.04	.02
80–90	.20	.09	.07

the position of the meniscus in the capillary tube is plotted as function of time. Typical curves for the study are shown in Figures 3 and 4. The change in slope is directly proportional to the change in hydraulic conductivity. The curves indicate that the most rapid change in hydraulic conductivity occurs at the beginning of the measurements. This can also be observed on Figure 5 which shows the calculated hydraulic conductivity versus time. Although the decrease in hydraulic conductivity was continuous with time, no evidence of complete sealing of the sample was found for periods of flow as long as 12 days. Some samples reached a relatively constant hydraulic conductivity value after long periods of time.

The change of hydraulic conductivity with time has been reported by several authors (Christiansen, 1947; Poulovassilis, 1972; Gupta and Swartzendruber, 1962). The change has been attributed to swelling or dispersion of soil due to the leaching of cations during the flow, to microbial development in the soil, or to air entrapment. If leaching of cations is the predominant mechanism of hydraulic conductivity decrease, the volume of water flowing through the soil should be more important than the duration of the flow. Figure 6 shows the hydraulic conductivity plotted versus volume outflow (represented by the position of the meniscus in the capillary tube) for some samples of Site III. The curves in Figure 5 and 6 are very similar. One can not determine from this information whether the change in hydraulic conductivity was primarily due to time or due to volume of flow.

Hydraulic Conductivity Values

Gradients from 2.7 to 3.7 cm of water/cm were used for measuring

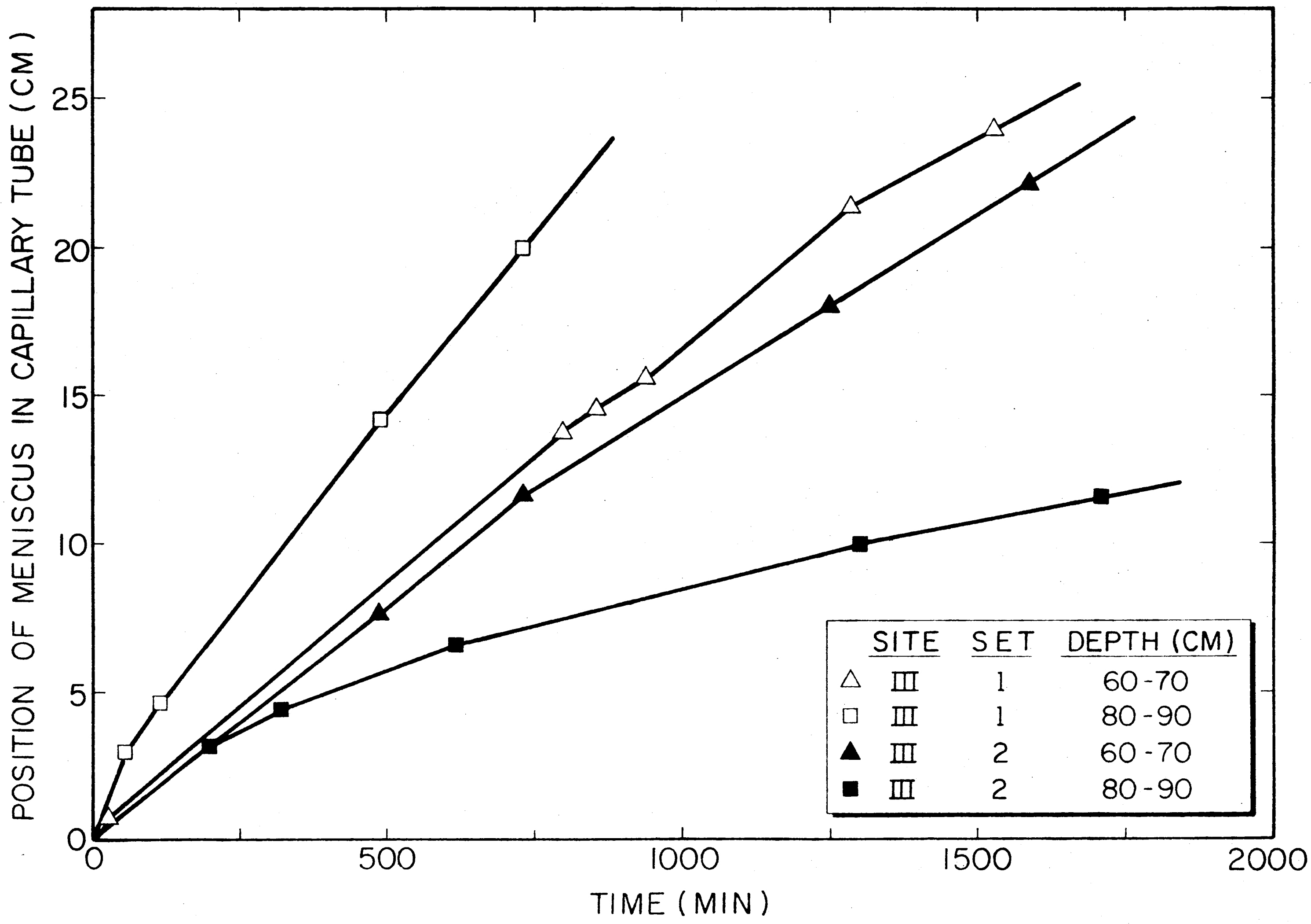

Figure 3. Variation of Flow with Time

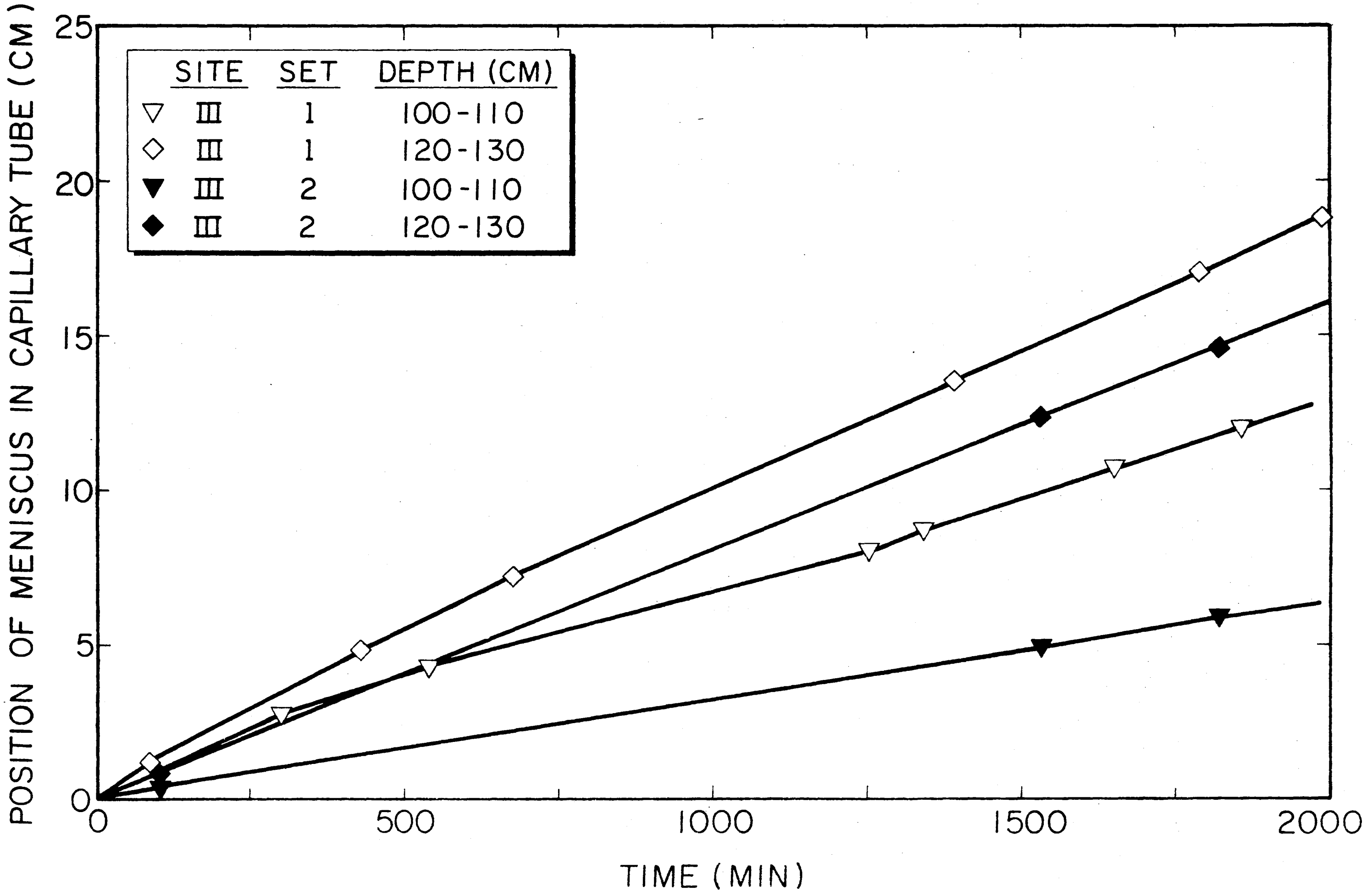

Figure 4. Variation of Flow with Time

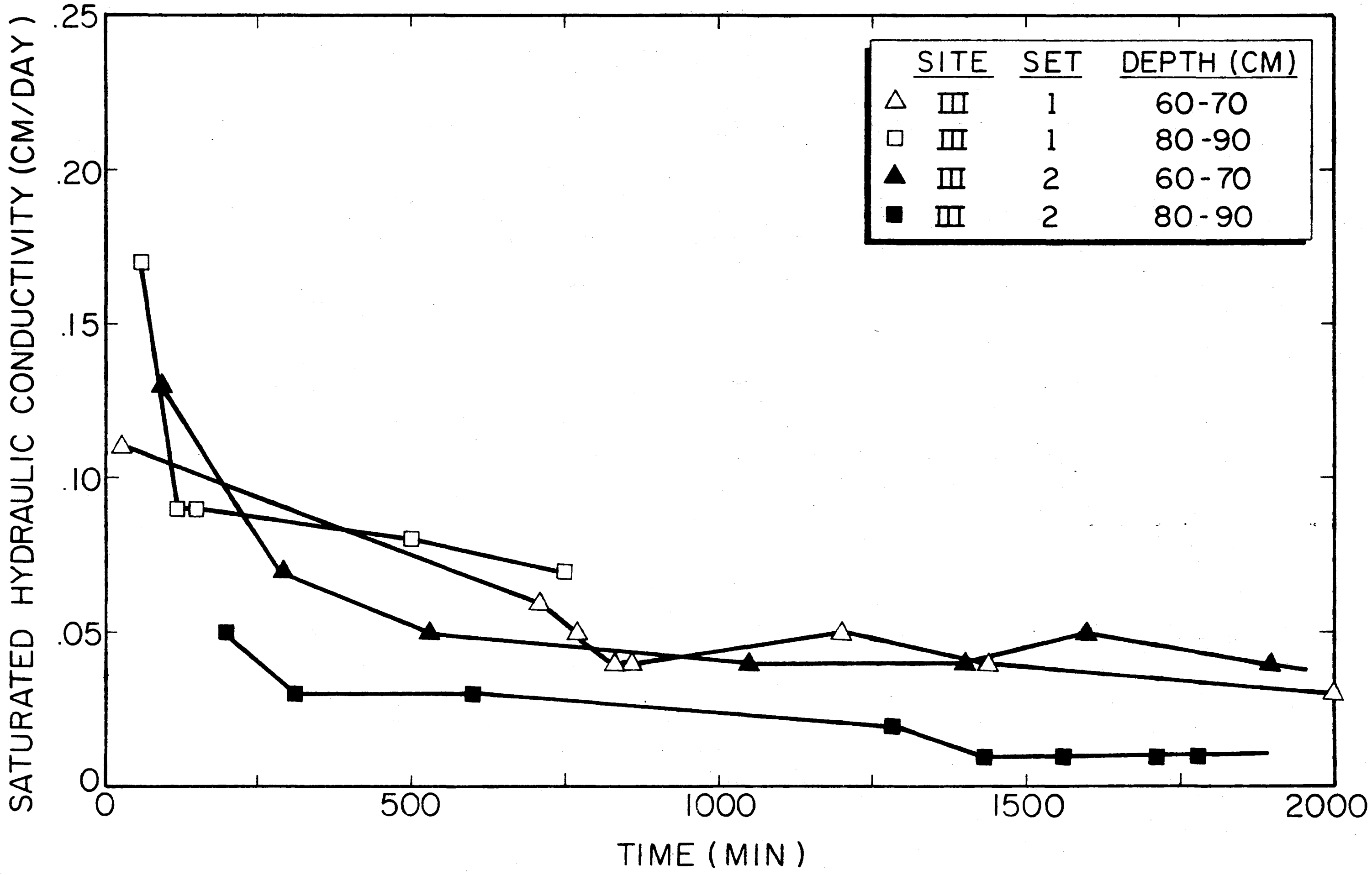

Figure 5. Change in Saturated Hydraulic Conductivity with Time

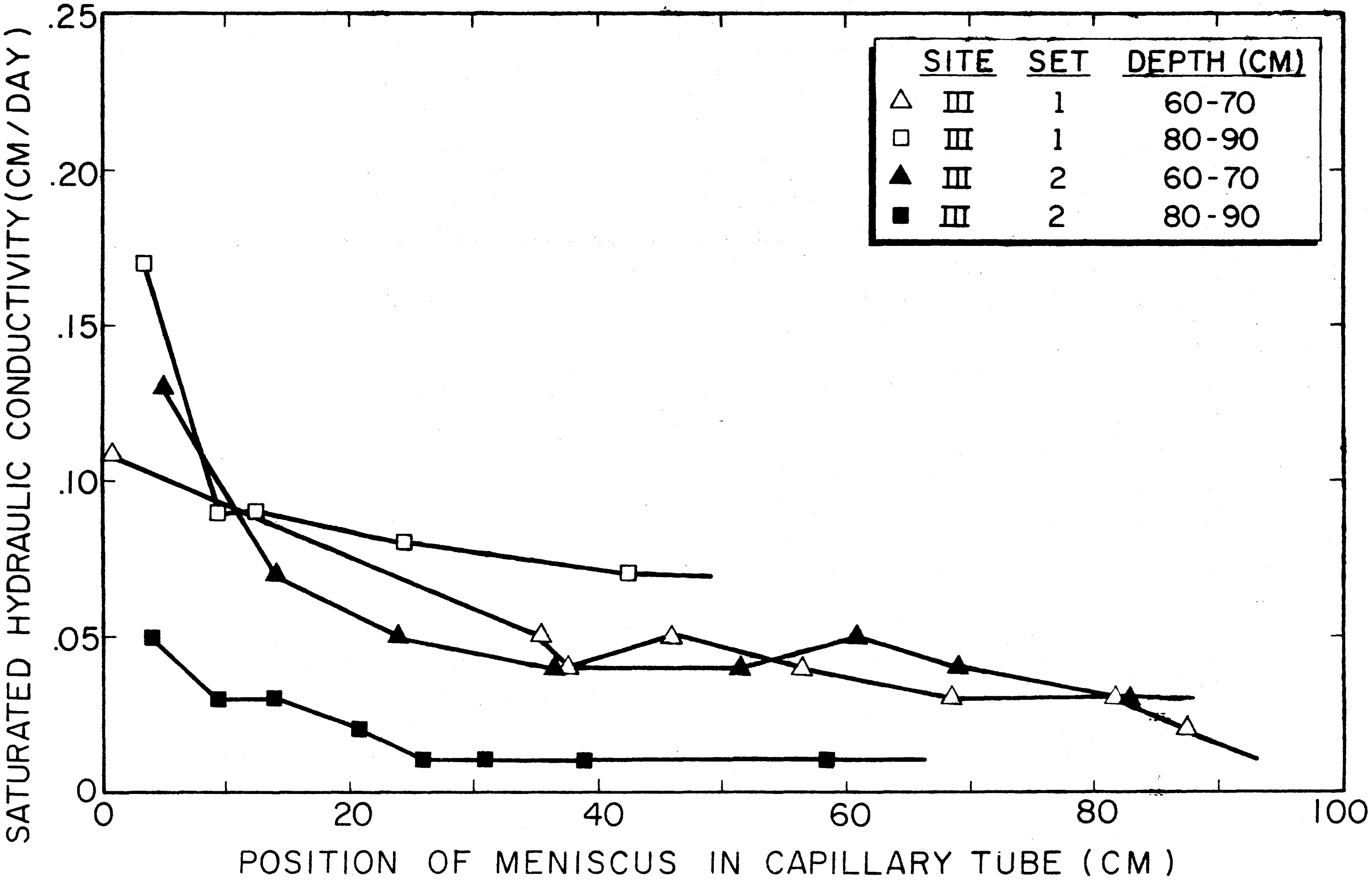

Figure 6. Change in Saturated Hydraulic Conductivity with Volume of Outflow

all the hydraulic conductivities reported below. Table II shows hydraulic conductivity values for each site, set and depth sampled. These hydraulic conductivity values were calculated for the first sixty minutes of flow. This arbitrary period of time was chosen to obtain hydraulic conductivity values representative of the field, while recognizing the dependence of the conductivity upon time. Figure 7 shows hydraulic conductivity values averaged over the sets for each site and each depth.

Analysis of variance of the values did not show any significant difference among sites or among depths. Therefore, no statistically important stratification was found for the profiles investigated in this study. However the depth of profile investigated was relatively shallow because of difficulties encountered in sampling undisturbed samples below 130 cm with the sampling machine available. Therefore, stratification may still exist in the profile. Statistical regression analysis showed a cubic function variation of hydraulic conductivity with depth, with a peak at depth 40-50 cm. Table II shows a relatively high hydraulic conductivity for depth 40-50 cm on Site I , Set 1. This value is suspected to be apparent. It is believed that in some cases the soil sample shrank, creating a flow path between the soil and the tubing used for encasing the sample. Such shrinkage may be due to a possible decrease in water content of the sample before the saturation process or to the effect of the solution flowing through the sample. Reeve and Tamaddoni (1965) and McNeal and Reeve (1964) discussed such shrinkage and attributed it to contraction of soil clay packets in presence of high salt waters.

TABLE II

SATURATED HYDRAULIC CONDUCTIVITY
VALUES (CM/DAY)

Depth (cm)	Site I		Site II		Site III	
	Set 1	Set 2	Set 1	Set 2	Set 1	Set 2
20-30	.20(6)*	.25(1)	0 (2)	0 (3)	.15(4)	.13(5)
40-50	1.14(4)	--(2)	.22(6)	.37(1)	--(5)	.84(3)
60-70	.11(1)	.04(5)	.35(3)	.94(4)	.08(2)	.11(6)
80-90	.11(5)	--(6)	.11(1)	.29(2)	.23(3)	.05(4)
100-110	.13(2)	--(3)	.09(4)	.01(5)	.01(6)	.01(1)
120-130	.10(3)	--(5)	.04(5)	.09(6)	.02(1)	.01(2)

*Numbers in parenthesis indicate the apparatus used for measuring the hydraulic conductivity.

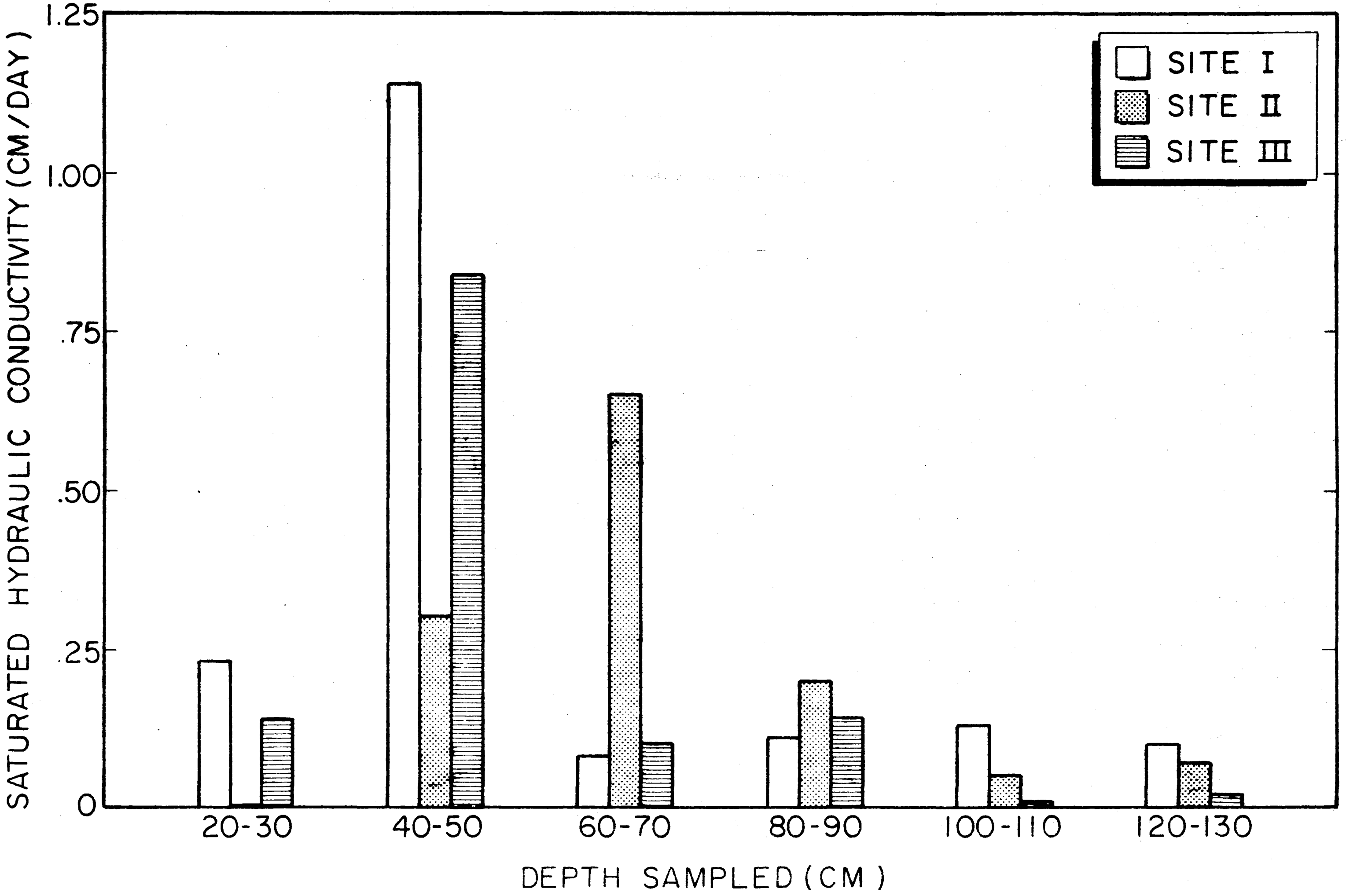

Figure 7. Saturated Hydraulic Conductivity Values at the Different Sites and Depths of Tillman-Hollister Clay Loam

Table III shows the hydraulic conductivity values for each depth averaged over the three sites and adjusted for variations from set to set and from apparatus to apparatus.

The hydraulic conductivity mean was 0.17 cm/day and the standard deviation was 0.211 cm/day. The hydraulic conductivity of the soil is therefore very small and may be classified as very low in the permeability classes of O'Neal (Klute, 1965). This may imply problems for irrigation and drainage of the Tillman-Hollister soil. The saturated hydraulic conductivity measured corresponds to the vertical hydraulic conductivity of the soil because one dimensional flow was studied in the laboratory. The soil may have a greater horizontal hydraulic conductivity

TABLE III

MEAN SATURATED HYDRAULIC CONDUCTIVITY VALUES*

Depth (cm)	Hydraulic Conductivity (cm/day)
20-30	.12
40-50	.48
60-70	.27
80-90	.14
100-110	.07
120-130	.08

*Averaged over sites. Means were adjusted for apparatus and sets.

CHAPTER V

SUMMARY AND CONCLUSIONS

The hydraulic conductivity of a field on the Oklahoma Agricultural Experiment Station Irrigation Research Station at Altus was studied. Depths of 20 to 30 cm, 40 to 50 cm, 60 to 70 cm, 80 to 90 cm, 100 to 110 cm, and 120 to 130 cm were sampled from three sites. A constant-head method was used for measuring hydraulic conductivity in the laboratory. Samples were encased in heat-shrinkable tubing and a 1500 ppm calcium chloride solution was used for flow measurements.

To test the applicability of Darcy's equation to the swelling clay loam soil studied, the flux density was determined for gradients ranging from 1 to 7.3 centimeters of water per centimeter. The flux density was found to increase more than proportionally with the gradient. The deviations from proportionality were small. Darcy's equation was considered to be a good approximation for the range of gradients used in the study.

The saturated hydraulic conductivity was found to decrease with time during the measurement process. The decrease was found to be maximum during the first hours of flow.

Very low saturated hydraulic conductivity values were found for all samples. The hydraulic conductivity mean was .17 cm/day. Statistical analysis of the data did not show any significant difference in hydraulic conductivity values between the sites studied or between the depths. It was therefore concluded that no evident stratification exists for the 130 cm profile studied.

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APPENDIXES

PROFILE DESCRIPTION[1]
TILLMAN CLAY LOAM
(TYPICAL PROFILE)

Horizon	Depth	Description
A_1	0 to 10 inches	reddish-brown (5YR 4/3, dry; 3/3.5, moist) clay loam becoming slightly darker in color below plow depth; slightly crusted surface; weak granular structure; hard when dry, firm when moist; noncalcareous (pH 7.5); clear boundary.
B_2	10 to 28 inches	reddish-brown (5YR 4/3, dry; 3/2,moist) light clay that is slightly lighter in color when crushed; moderate, very fine, blocky structure; very hard when dry, very firm when moist; clay skins apparent, but not pronounced; few small, black concretions; noncalcareous (pH 8.0); gradual boundary.
C_{ca}	28 to 50 inches	reddish-brown (5YR 3/4, dry; 3/6, moist) clay; massive (structureless); very hard when dry, very firm when moist, many soft concretions of calcium carbonate; soil mass calcareous; gradual boundary.
C	50 to 60 inches	yellowish-red (5YR 4/6c dry; 3/6, moist) clay containing less calcium carbonate concretions than above.

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PROFILE DESCRIPTION[1]
HOLLISTER CLAY LOAM
(TYPICAL PROFILE)

Horizon	Depth	Description
A_p	0 to 5 inches	grayish-brown (10YR 5/2, dry; 3/2, moist) clay loam; weak, granular structure; hard when dry, firm when moist; noncalcareous (pH 7.5); abrupt boundary.
A	5 to 9 inches	very dark gray (10YR 3/2, dry; 2/2, moist) clay loam; weak, granular structure; hard when dry, firm when moist; many fine pores; peds have a weak shine; noncalcareous (pH 7.5); gradual boundary.
B	9 to 28 inches	very dark gray (10YR 3/1, dry; 2/2, moist) clay; moderate, medium, subangular blocky structure becoming blocky at 16 inches; very hard when dry, firm to very firm when moist; clay skins apparent; noncalcareous to 20 inches (pH 7.5); gradual boundary.
B	28 to 36 inches	gray (10YR 5/1, dry; 4/1, moist) clay; weak, blocky structure; very hard when dry, very firm when moist; few whitish spots of soft calcium carbonate; calcareous; gradual boundary.
C_{ca}	36 to 44 inches	gray (10YR 5/1, dry; 4/1, moist) clay; weak, blocky structure; very hard when dry, very firm when moist; more compact than layer above; mixture of soft and hard concretions of calcium carbonate strongly calcareous; gradual boundary.
C	44 to 60 inches +	gray (10YR 5/1, dry; 5/2, moist) clay; garding to reddish-brown clay. This is apparently red-bed residuum.

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