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Experimental Study on Determining Unsaturated Property of Soil

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Synopsis

This paper deals with the experimental study of hydraulic properties of unsaturated soil. In treating unsaturated zone, a great deal more data are required than are required for the saturated zone, but these properties of soils must be known to apply the finite element approach to actual groundwater flow problems. The purposes of this paper are to propose a rational basis of getting experimental relationships between prossure head(ψ) and hydraulic conductivity(K) and between pressure head(ψ) and volumetric moisture content(θ) with "the instantaneous profile method" in a laboratory. An apparatus was constructed and test procedures were developed to measure pressure head and volumetric moisture content by using pressure transducers and low-energy gamma ray attenuation.

The technique of a low-energy gamma radiation apparatus does provide a means for accurate measurement of water content without disturbing the system into which water is moving. Furthermore rapid measurement of water content becomes possible at any position in a soil so that water content changes with time may easily be followed. The tensiometer-transducer system provides a most valuable means of measuring pressure head with rapid response and with provision of a complete record of the pressure head changes with time.

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1. Introduction

The hydrologic behavior of soils is to a large extent determined by how the hydraulic conductivity of partially saturated soil varies widely as a function of the volumetric moisture content and/or the pressure head which is also a function of the volumetric moisture as shown in Fig.1. These properties of soils must be known if the analysis of finite element method on flow through porous media is to be applied in field situations.

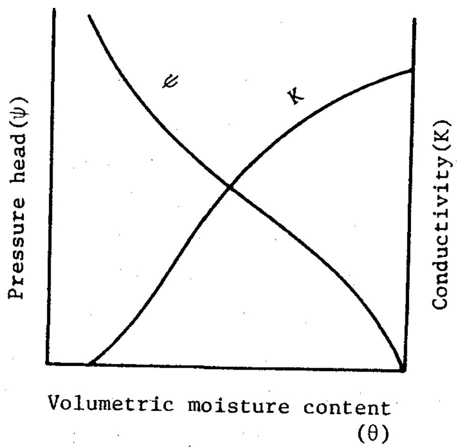


Fig.1 Unsaturated property of soil

Since there is no reliable way to predict these values from more fundamental soil properties, these properties must be measured experimentally. A number of techniques for measuring hydraulic conductivities have been reported in the literature. Most methods are based on the solution of the one-dimensional continuity equation for steady and unsteady flow, with or without the gravity term. In steady flow systems, flux, gradient, and water content are constant with time, while in transient flow systems, they vary. The most commonly used methods are based on steady flow. While many such methods are described in the literature [1-5], they are all based on essentially the same procedure. The tested sample, or core, is mounted either in a plexiglass tube or in a pressurized rubber sleeve, and a steady flow of water is established through it. A steady flow, i.e., inflow equals outflow, is reached within 2-40 hours, depending on the sample's permeability and the method used. At this stage the pressures at either end of the sample, the rates of flow and the saturation are determined. But steady-state methods have the disadvantage of requiring relatively long times to establish steady flow. During this time, changes can occur in the hydraulic properties of the sample and the porous end barriers. The use of mercuric chloride, phenol, or thymol in the water to reduce biological activity is customary. If the water that is used in the measurements differs markedly in ionic strength and composition from the soil solution initially present in the pores of the sample, there may be considerable change in the conductivity of soils containing clay during the flow. Therefore steady-state methods are primarily laboratory methods.

Now a widely used transient-flow methods (unsteady-state methods) for measurement of conductivity and diffusivity in the laboratory may be grouped into : (1) outflow methods and (2) instantaneous profile methods.

The outflow method has been used to obtain estimates of the measurement of the volume of water outflow as a function of time from a sample placed in a pressure cell [6,7]. The slab of soil is placed on a porous plate or membrane and is brought to equilibrium with a certain gas-phase pressure in the cell. By increasing the gas (e.g., air) pressure in the cell by a small increment, water is forced out through the membrane. The volume of water draining out of the sample is measured as a function of time. The procedure is then repeated by further increasing the gas pressure. These data are then introduced into the solution of the unsteady, one-dimensional continuity equation in which the effect of gravity is neglected. By fitting the experimental outflow-time curve to the theoretical curve, one can obtain a value of the soil water diffusivity that can be associated with the mean water content (or pressure head) of the soil sample during the increment of outflow.

But in many cases the theoretical and experimental outflow curves cannot be matched [7], indicating that the assumption of negligible membrane resistance in the analysis is not valid. Jackson has examined the method and concluded that it is not practical to use pressure increments small enough to validate the assumption of constant K and C [8], and that it is extremely difficult to obtain replicated results with the method.

Laboratory measurements of conductivity can also be made on long columns of soil, not only on small samples contained in cells. In such a column, steady-state flow can be induced [9]. If the column is long enough to allow the measurement of pressure gradients and of water-content gradients, the $K(\theta)$ and $K(\psi)$ relationship can be obtained for a considerable range of θ with a single column. If periodic pressure and wetness profiles are measured, the flux values at different time and space intervals can be evaluated by graphic integration between successive moisture profiles. This procedure has been called the "instantaneous profile" method, and it can be applied in the field as well [10]. The theory does not assume uniformity of the hydraulic properties of the flow system, and the boundary conditions do not need to be constant, or known in detail. Methods of this type seem to offer the best possibility for hydraulic characterization of sample or field soils. In the laboratory, pressures are measured by pressure transducer tensiometers [11-14]. Moisture content may be measured by a gamma-ray attenuation system [11-15], by a neutron scattering device [16].

The purposes of this paper are to propose a rational basis of getting experimental relationships between moisture content(θ) and hydraulic conductivity $K(\theta)$ and between pressure head (ψ) and volumetric moisture content(θ) by the instantaneous profile method by using the source of low-energy gamma-ray attenuation and pressure transducer tensiometer.

2. Instantaneous Profile Analysis

In essence, the method of instantaneous profiles consists of determining down the column the profiles of the macroscopic flow velocity, the potential gradient, and the volumetric moisture content at any instant of time after the commencement of drainage or infiltration. Once these are known for a particular time, it is then possible to find the instantaneous hydraulic conductivity for each elevation by dividing the appropriate velocity value by the potential gradient value.

The continuity equation may be applied to one-dimensional flow system as

$$\left(\frac{\partial \theta}{\partial t} \right)_z = -\left(\frac{\partial V}{\partial z} \right)_t \qquad (1)$$

where V is the flow velocity (cm/sec), θ is the volumetric moisture content (cc/cc), and z is the elevation above the datum plane defined as positive in the upward direction as sketched in Fig.2. In the case of drainage, the curves presented in Fig.3. which will be obtained from the experimental information give the variations of water content with time for several column elevations. Using this information, it is a straightforward matter to find the relation between $\partial\theta/\partial t$ and z at several required times as shown in Fig.4.

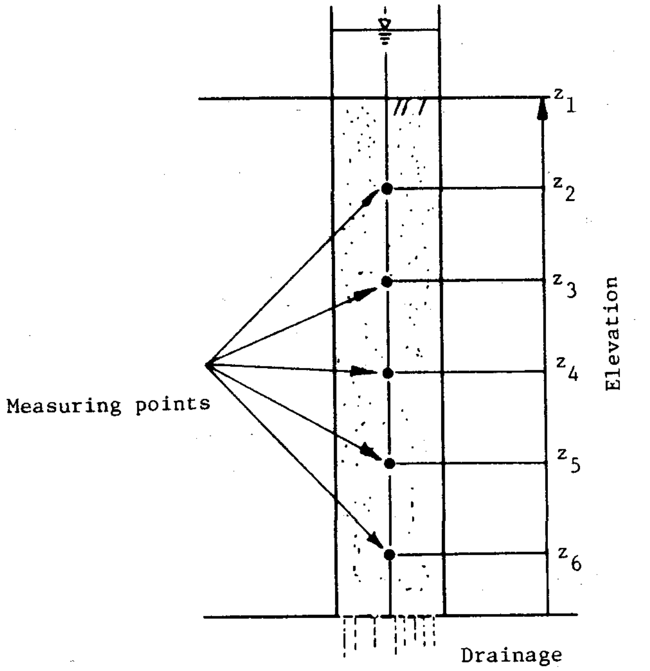


Fig.2 Schematic picture of vertical drainage

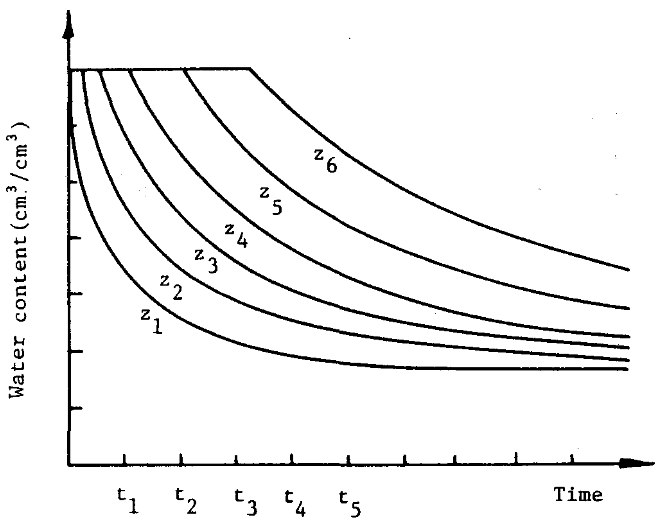


Fig.3 Variation of moisture content with time at several column elevations (after Watson [11])

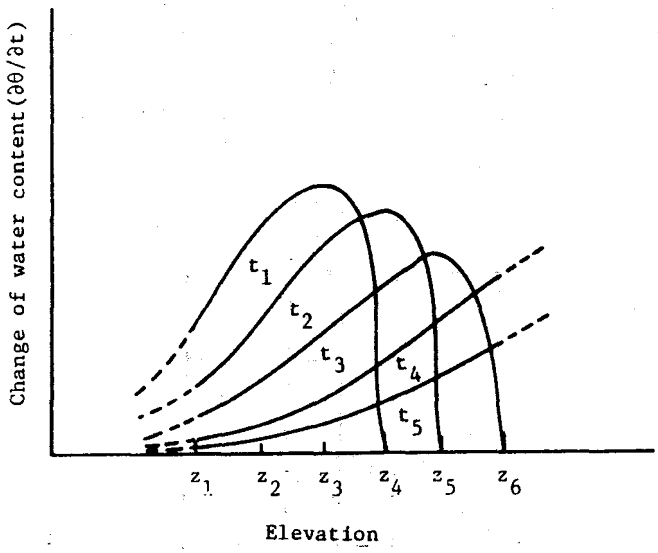


Fig.4 Changes of water content at each elevations with time

Integration of Eq.(1) with respect to z yields:

$$\int \frac{\partial \theta}{\partial t} dz = -V + C(t) \qquad (2)$$

It should be noted the velocity is zero at the surface, then Eq.(2) becomes

$$-V(z,t) = \int_{z_s}^{z} \frac{\partial \theta}{\partial t} dz \qquad (3)$$

where V(z,t) is the velocity at position z and z_s is the elevation of surface. The velocity profiles are obtained by integrating graphically with respect to z-∂θ/∂t profile curves, and the resulting velocity profiles are given in Fig.5. These profiles represent the instantaneous velocities down the column at time stated.

The first part of the method is the determination of the pressure head(ψ) profile with time at several elevations in the column as shown in Fig.6. Since the total potential h is equal to the negative pressure head(ψ) (in the case of drainage) and the gravitational component z, the total potential profiles at each time may be readily plotted and are presented in Fig.7. These curve may then be differentiated graphically to give the positive potential gradient (∂h/∂z) profiles as shown in Fig.8.

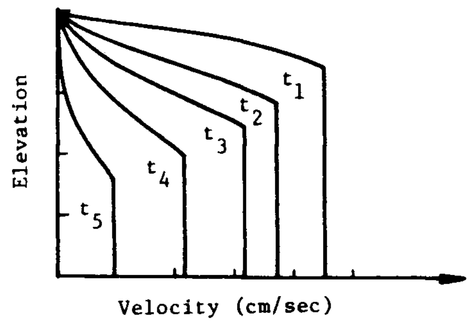


Fig.5 Instantaneous velocity profiles

From Fig.5 and Fig.8, it is a relatively simple matter to determine the instantaneous hydraulic conductivity for any elevation and time from Darcy's law

$$K = \frac{(V)_{z,t}}{(\frac{\partial h}{\partial z})_{z,t}} \qquad (4)$$

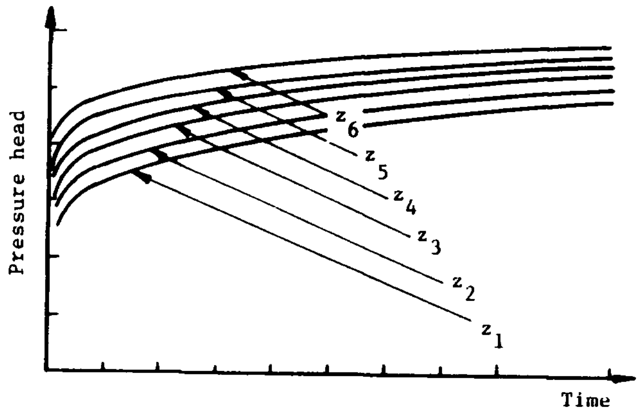


Fig.6 Variation of soil water suction with time at several column elevations

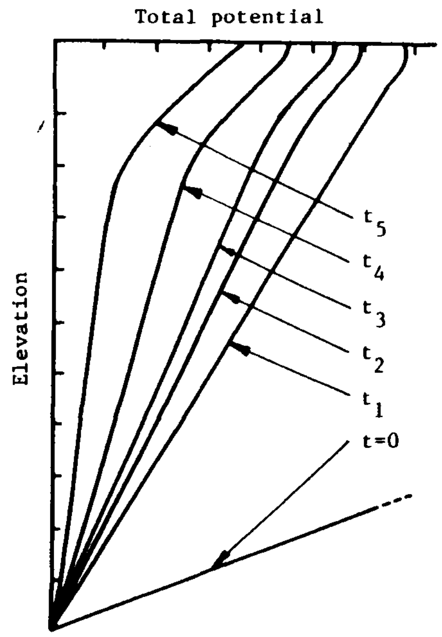


Fig.7 Instantaneous total potential profiles

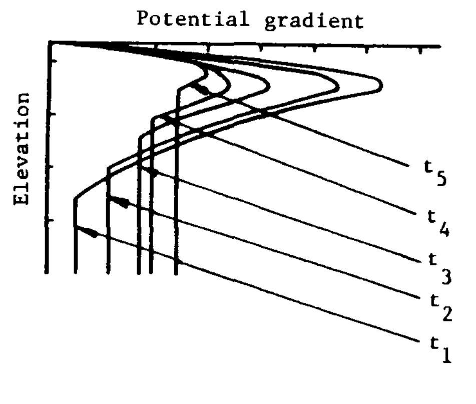


Fig.8 Instantaneous potential profiles

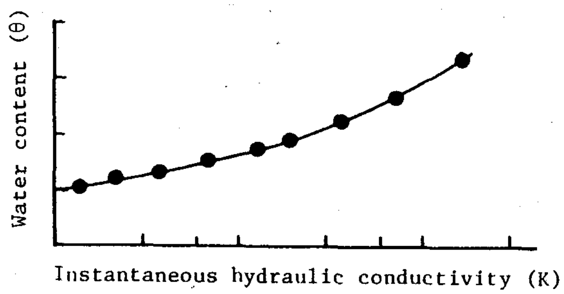


Fig.9 The water content-instantaneous hydraulic conductivity relation showing the computed values

The last step of the method is the plotting of the curve relating the values of the instantaneous hydraulic conductivity and the water content. This relationship is given in Fig.9, where the conductivity scale is logarithmic to enable the lower conductivities to be presented accurately.

3. Experimental Apparatus and Procedure

3.1 Measurement of Moisture Content by Gamma Ray Attenuation

In the determination of unsaturated permeability of columns of soil, some experimental methods of measuring the distribution of soil water content along the length of the column have been reported. Gravimetric sampling is the common method used [17,18]. This disturbs the system so that information about water distribution as a function of time must be put together from measurements made on replicated systems. Because of undetected density gradients, cracks and other flaws which may occur, true replicated systems are difficult to produce; thus, serious errors may result in water distribution-time studies.

Then, some methods of nondestructive measurement are desirable. Methods involving measuring electrical [19] or thermal properties of porous blocks are secondarily applied, however, these methods have the following limitations:

(1) lag in following water content changes

(2) insufficient range

To conquest these limitations, two radioactive techniques have been used for measuring soil water content, the neutron-scattering method and the gamma ray absorption method. In the neutron-scattering method, the region of neutron scattering is very wide, and so it is difficult to get an accurate coordinate of measuring point. While, the gamma ray method enable accurate measurement of soil density [20] and water content at given positions in a column of soil, based on the fact that scattering and absorption of gamma rays are related to the density of matter in their path.

The method is not specific for water, as in the case of the neutron-scattering method. However, it has the advantage that gamma rays may be collimated by suitable geometry and shielding to a narrow beam which gives resolution in position at which readings are taken. As neutrons are not readily collimated, and as the counting technique commonly used counts the scattered neutrons rather than the unscattered, the neutron method is more suited to water-content measurements of a large bulk of soil.

To use the gamma ray method to determine water content, the soil density must be known to the same degree of accuracy. This may readily be determined at a fixed value of water content and used for subsequent determinations of water content provided the soil density remains unchanged thereafter. Then, the method has one major limitation which restricts its use in measuring soil water content. A change in density which occurs in the soil because of shrinking and swelling will affect gamma absorption. Hence, use of the method is restricted to conditions where bulk density changes are negligible compared with changes in moisture density to be measured, or where independent measurements permit correction for changes in bulk density. Then, in this study, the water content was measured by using gamma ray attenuation, following the procedure of Davidson et al. [21], Watson [11]. However, in contrast to previous works as indicated in Table 1, low-energy gamma radiation (100μ curies of cobalt 60) rather than cesium 137 or americium 241 was used as the source of gamma photons.

The arrangement of the source and detector is shown schematically in Fig.10. The C_o^{60} source holder was made of lead bricks arranged in a cube about 6.0cm on a side. The source was placed in the center of one of the block faces in such a way that it was about 2cm from the face.

Table 1 Summary of gamma attenuation method

Reference	Gamma ray source	
Gurr.C.G. [22]	25 millicuries of C_s^{137}	1962
H.Ferguson et al. [23]	20 millicuries of Cesium^{137}	1962
Davidson et al. [21]	200 millicuries of C_s^{137}	1963
Topp G.C et al. [24]	200 millicuries of Cesium^{137}	1966
G.C.Topp et al. [25]	100 millicuries of Cesium^{137}	1967
Yen et al. [26]	200 millicuries of C_s^{137}	1968
Topp [27]	200 millicuries of C_s^{137}	1969
G.Vachaud et al. [13]	100 millicuries of americium^{241}	1971
R.S.Saksena [28]	100 millicuries of C_s^{137}	1974

The total thichness of the lead collimator was 5.0cm, and the collimator slit was 7.0cm high and 0.6cm wide. The detector was inserted in a lead annular ring with the front of the probe flush against the rear of the collimator so that the collimation slit was centered on the face of the probe. Thus, with minor exception, the probe detects only those gamma rays which pass directly from the source through the sample and collimation hole to the probe.

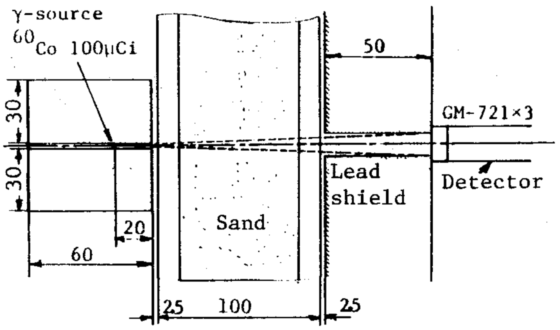


Fig.10 Schematic position of the gamma beam

(Schematic representation of the gamma source holder, collimation and probe)

A small conveyor was placed on a vertical guide track between the source holder and the collimator in such a way that the conveyor could be moved vertically through the stationary gamma beam. The guide track maintained the conveyor at a constant position with regard to the collimator face. During a run, soil through which water was moving was placed on the conveyor in such a way that the gamma beam passed through the soil in a direction perpendicular to the direction of water movement. By moving the conveyor it was possible to pass the beam through any vertical section of soil.

It was necessary to calibrate counts against moisture content. The various content soils were packed with same dry density (γ_d = 1.5g/cm^3) in acrylic rectangular boxes 7.0cm high, 10.0cm wide, and 20.0cm long. The curve as shown in Fig.11 was obtained by plotting the record of the counts per 10seconds obtained at a particular moisture content. This graph of moisture content against counts was used to transform the counts obtained during a run to moisture content.

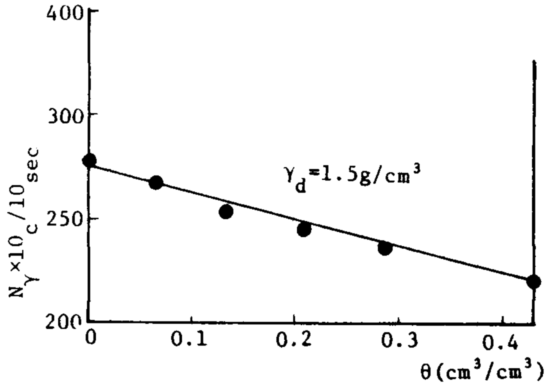


Fig.11 Calibration curve for sand

3.2 Measurement of Pore Water Pressure

To know the relation of moisture content and pore water pressure, it is necessary to measure pore water pressure at the same time. The satisfactory use of a pressure transducer for pressure head measurements requiring rapid response has been previously reported by Klute [29,30], and Watson [31].

Then, in this experiment, pressure head was measured at the same time for 5 different depths in the column by using tensiometers and transducers (the range of measuring from $-100g/cm^2$ to $+100g/cm^2$), each tensiometer was connected to its own transducer. Five tensiometers were arranged vertically along the column at 10cm intervals as shown in Fig.12. The active area of each tensiometer was a ceramic plate which has a bubbling pressure of approximately 2000cm of water. The tensiometer units, illustrated in Fig.13, are screwed directly into the acrylic box. Ecah transducer was connected by a multichannel data acquisition system.

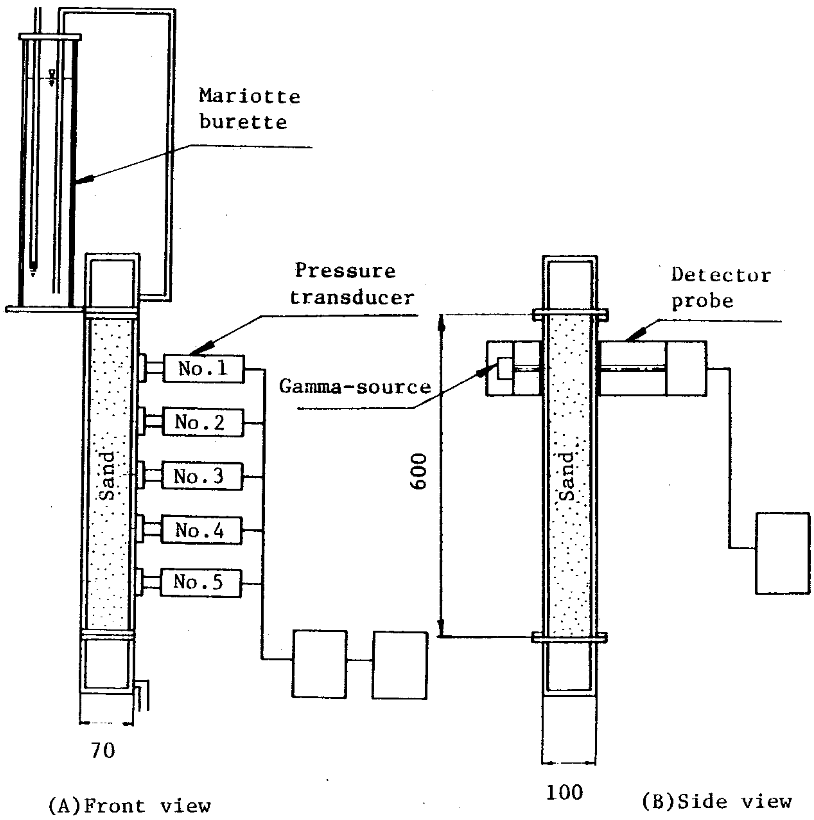


Fig.12 A schematic representation of the experimental apparatus: (A) front view showing relative position of the gamma system, (B)side view showing relative positions of tensiometers

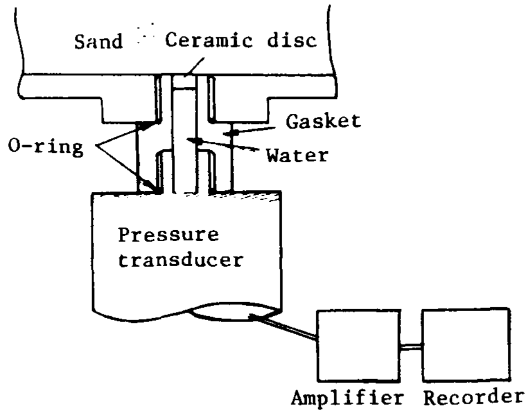


Fig.13 Pressure transducer assembly details

3.3 Experimental Procedure

The soil sample used in the experiment was Toyoura standard sand with a specific gravity of 2.65. The material was carefully packed into the acrylic rectangular box 7.0cm heigh, 10.0cm wide, and 60.0cm long with a particular dry density (γ_d = 1.50g/cm^3) using a tremie to prevent segregation and tamped frequently to produce a tight packing. The mode of packing was identical in all experiments. Both ends of column were made of fine screen through which water moved into soil.

A. Series 1 (Water applied on top of the column)

Test 1-1. The air-dried column of sand at an initial uniform water content (0.01cm^3/cm^3) had water applied on top of the surface by keeping a head of water 8.5cm in height as shown in Fig.12. The quantity of water entering the column was measured by using a Mariotte burette as shown in Fig.12.

Pore water pressure was measured by pressure transducers. The use of this equipment permitted accurate (±0.5mm of water) pressure measurements to be made on a dynamic system.

To measure moisture content 10-second counts was made at each

10cm interval along the column. The conveyor was then positioned to the place where the first water measurement was desired water flow into the soil was started, and a 10-second count was made.

Both the count and the time at which it was started were recorded. Because of the rapid entry of water into the air-dry soil, counts were made as frequently as possible during the early part of the run. As water moved into the soil, the conveyor was moved to whatever position a water determination was desired. In this way a record of counts per 10seconds as a function of time was obtained at various positions along the soil column.

Infiltration occurred for 16minutes after which the wetting front reached the bottom (z=0). Infiltration examined for a period of 25minutes.

Test 1-2. At the end of Test 1-1, to obtain the hydraulic conductivity K_s ($S_r \doteqdot 100\%$) in saturated state constant-head permeability test was performed. Fig.14 shows the setup. The lowest elevation in the sand column was thus a fixed piezometric surface; this surface was taken as datum in gradient measurements. Under these conditions the hydraulic gradient was known and, from measuring the rate of volume outflow from the bottom of the column, the saturated hydraulic conductivity (K_s) could be determined.

Test 1-3. In the third test of experiments, the columns were first allowed to become saturated by infiltrating water, keeping the surfaces saturated, until water drained from the bottom. Since the flow condition was that of drainage of the saturated column to atmosphere at its base, the base of the column was constructed so that air at atmospheric pressure was always maintained during an experiment on the under side of the screen supporting the sand. The drainage process commenced the moment the ponded water disappeared through the upper soil surface. Drainage examined for a period of 35 hours measuring the variations of pressure head and volumetric moisture content to obtain the relationship during the drying process.

Test 1-4. A fourth test of experiments was initiated at the end of Test 1-3. The soil column was submitted to another cycle of infiltration and redistribution to obtain a scanning curve. On rewetting, water was ponded constantly on the soil surface to a depth of 8.5cm, but the initial water content distribution in the soil column was that obtained at the end of the Test 1-3 redistribution and was very close to a static equilibrium water profile. Infiltration occurred for 22 minutes.

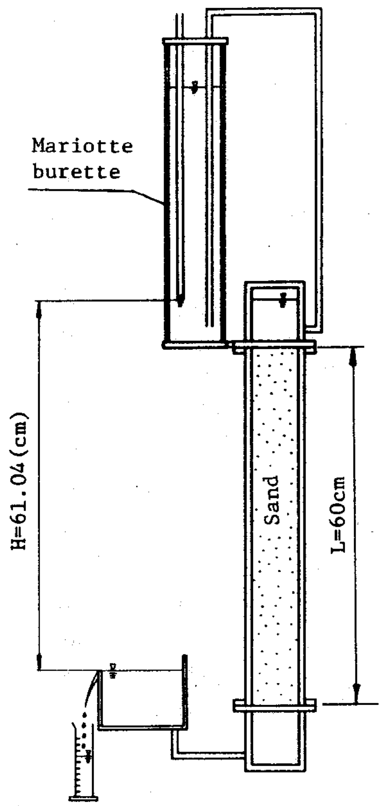


Fig.14 Setup for constant-head permeability test

B. Series 2 (Water applied from the base of the column)

Test 2-1. The air-dried column of sand at an initial uniform water content ($0.01cm^3/cm^3$) had water applied from the base of the column by a constant head of water 31.5cm. in height as shown in Fig.15. So the lower end of the column was then immersed under a free water surface. The rate of infiltration was reduced to slower constant rates by means of capillaries fed from a constant head, and the moisture profiles were allowed to attain their equilibrium conditions. In this test, the variations of pressure head and volumetric moisture content were measured.

Test 2-2. At the end of Test 2-1, the initial water content distribution in the soil column was obtained. The potential head was suddenly lowered to a head of water 15.2cm in height at the bottom. The pressure head and volumetric moisture content was changed by this process. These variations were measured for a long period of time.

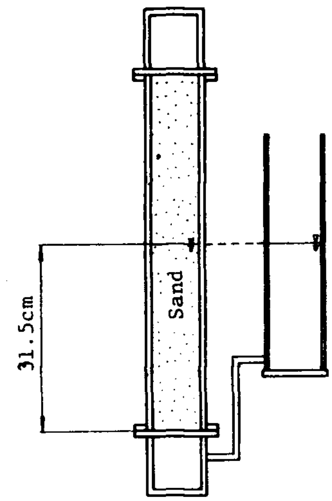


Fig.15 Setup for water applied from the base

4. Experimental Results and Discussions

4.1 Relationships between hydraulic conductivity (K), volumetric moisture content (θ) and pressure head (ψ)

The results relative to Test 1-1 are reported in Fig.16. Fig.16 indicates the variation of the volumetric moisture content (θ) with time at five elevations in the column for the 25 minutes of infiltration. These values of the moisture content (θ) were obtained by using the calibration curve in Fig.11.

As it is clear from Fig.16, the moisture content at the wetting front changed so rapidly that it is difficult to obtain hydraulic conductivity at low moisture contents. Therefore the results relative to drainage process (Test 1-3) was mainly used to determine the relationships between hydraulic conductivity (K) and volumetric moisture content (θ).

The experimental information obtained from Test 1-3 has been summarized in Figs.17, 18, and 19. Fig.17 indicates the volumetric moisture content variation with time for each depth. If these lines are replotted with time as the parameter, as shown in Fig.18, useful soil-water relations may readily by calculated. It is possible to calculate moisture flux through each depth increment by integrating moisture-time curve with respect to depth. The slope ($d\theta/dt$) is measured at particular points in time.

In a similar manner, Fig.19 gives pressure head changes with time at the same column elevations. From this results hydraulic head profiles can be calculated by adding pressure head to depth for each tensiometer. By using Eq.(4), the hydraulic conductivity can be calculated by dividing fluxes by the corresponding hydraulic gradient values. The gradients are obtained by measuring the slopes of hydraulic head versus elevation. Correlating each time the value of K with the mean water content $\overline{\theta}$ obtained in z_j during the mean time $(t_{K+1}+t_K)/2$ permitted us to obtain the $K(\theta)$. The results are given in Fig.4.20. The hydraulic conductivity which was obtained from the results of constant-head permeability test of Test 1-2 was $K_s=2.084\times10^{-2}$ cm/sec.

In Fig.20, the relation between K-θ is shown. On the other hand, the relations between K and ψ was sometimes used heretofore. Various empirical equations for that relation of conductivity to pressure head (ψ) or moisture content (θ) have been proposed by many researchers as indicated in Table.2. In Table 2, the empirical parameters are depend upon the liquid, the soil, and the capillary pressure history of the system and the values of the parameters must be determined experimentally. And then no fundamentally based equation of general validity is available for the relation, and existing knowledge does not allow the reliable prediction of unsaturated conductivity from basic soil properties.

There is general agreement that the relation of conductivity to pressure head depends upon hysteresis, and is thus different in a wetting than in a drying soil [15,24]. Namely, if Fig.21 is compared with Fig.22, of which results are obtained in the same condition, it is clear that the relation of conductivity to moisture content is affected by hysteresis to a much lesser degree. One may thus neglect any hysteresis in $K(\theta)$ and use a unique relationship.

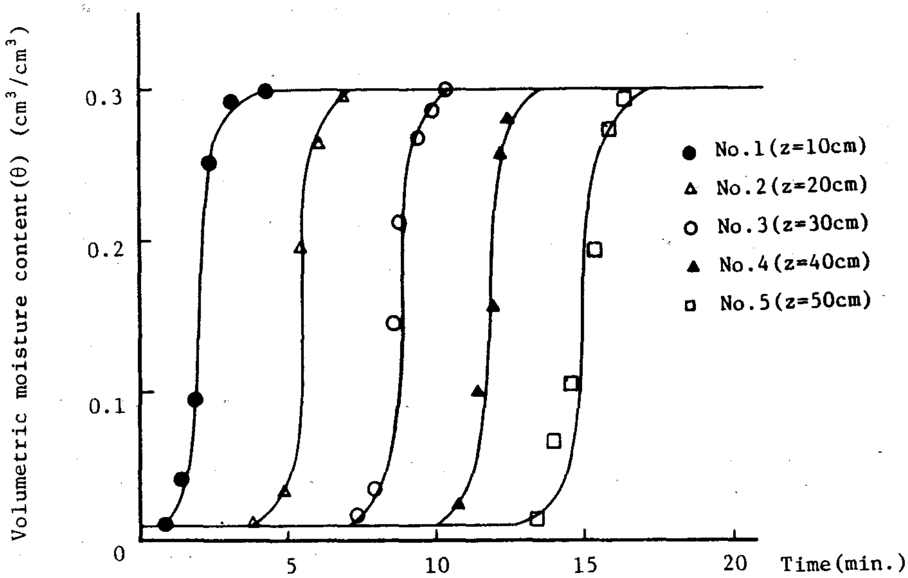


Fig.16 (a) Change of moisture content with time during infiltration (Test 1-1)

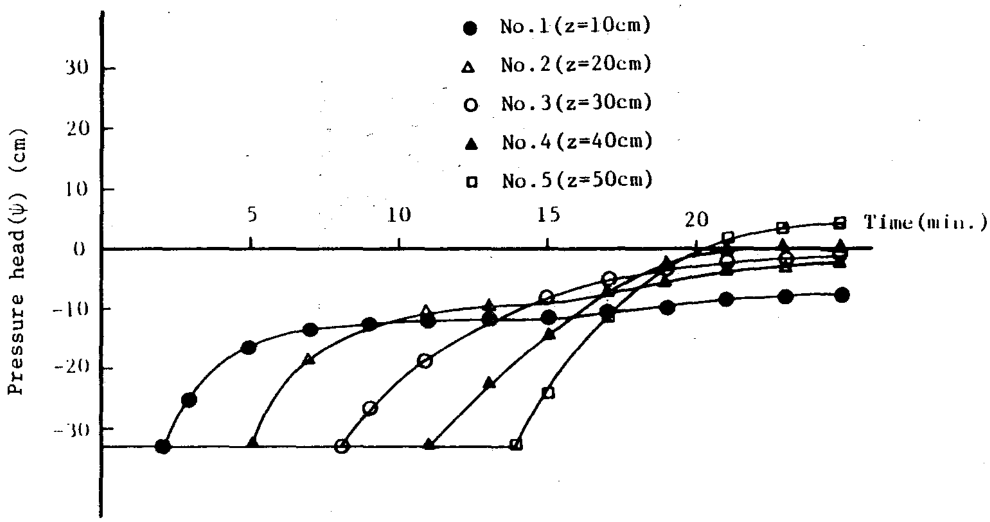


Fig.16 (b) Change of pressure head with time at five column elevations during infiltration (Test 1-1)

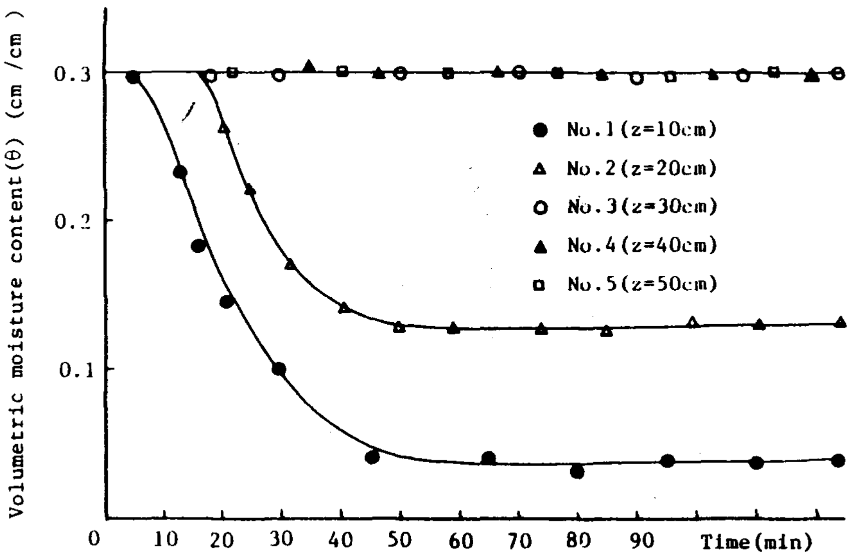


Fig.17 Change of moisture content with time during drainage (Test 1-3)

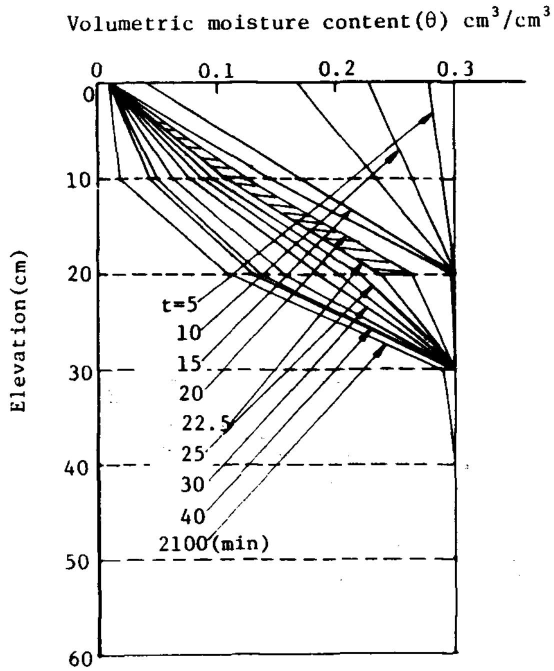


Fig.18 Distribution of water content for sand during drainage (Test 1-3)

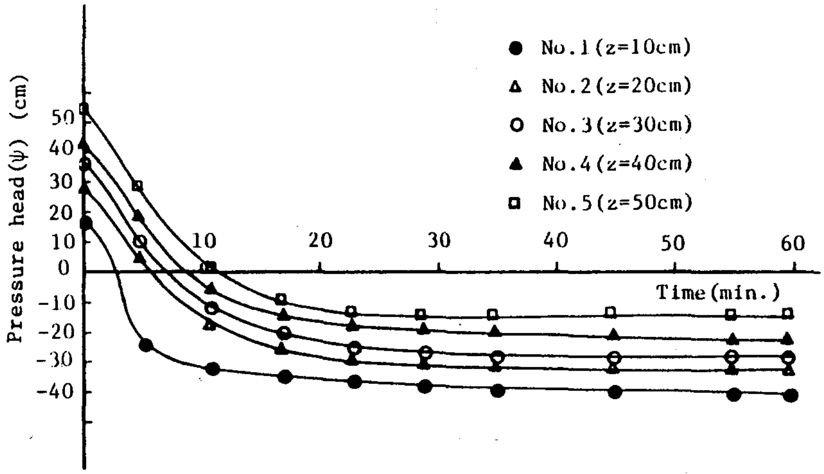


Fig.19 Change of pressure head with time at five column elevations during drainage (Test 1-3)

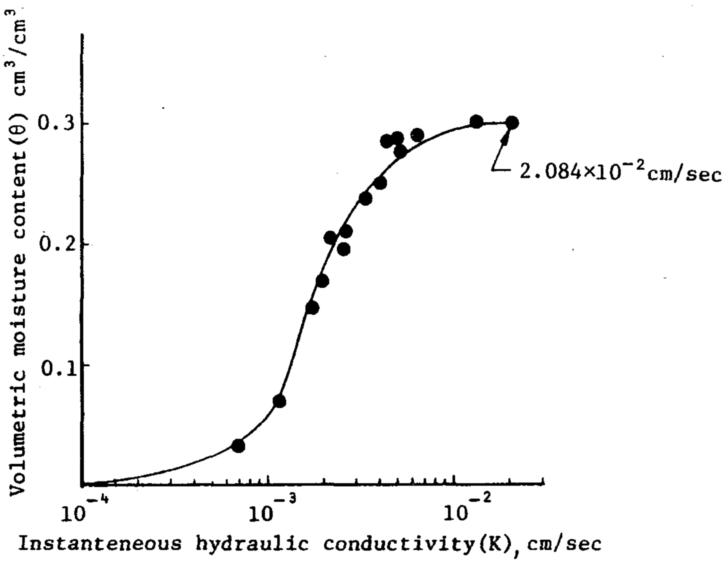


Fig.20 The moisture content - instantaneous hydraulic conductivity relation for sand (Test 1-3)

Table 2. Empirical equations for the relation of hydraulic conductivity of unsaturated soil to pressure head or moisture content

Empirical equation	Reference
$K = \dfrac{a}{(-\psi)^m + b}$	Gardner,W.R [32]
$K = K_0 (a/\psi)^m$	Schleusener,R.A. and A.T.Corey [33] Scott,V.H. and A.T.Corey [34]
$K = a \left[\dfrac{\cosh[(\psi/b)^m] - 1}{\cosh[(\psi/b)^m] + 1} \right]$	King,L.G. [35]
$K = K_0 \exp(m\psi)$	Gardner,W.R. [36] Philip,J.R. [37]
$K = a\theta^m$	Ahuja,L.R. [38]
$K = K_0 \left(\dfrac{\theta - \theta_r}{\theta_s - \theta_r} \right)^3$	Kroszynski,V. [39]
$K = K_0 + a\theta + b\theta^2$	Bruch,J.C. and G.Zyvoloski [40]

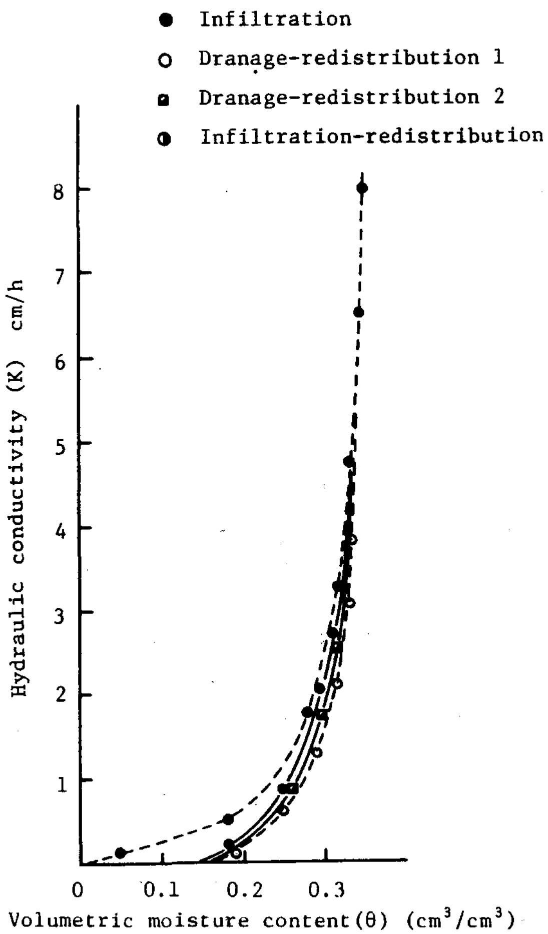
where

a,b and m : Empirical parameters depending upon the liquid, the soil, and the capillary pressure history of the system.

K_0 : The hydraulic conductivity in saturated state.

θ_s : Volumetric moisture content in saturated state

θ_r : Residual volumetric moisture content.



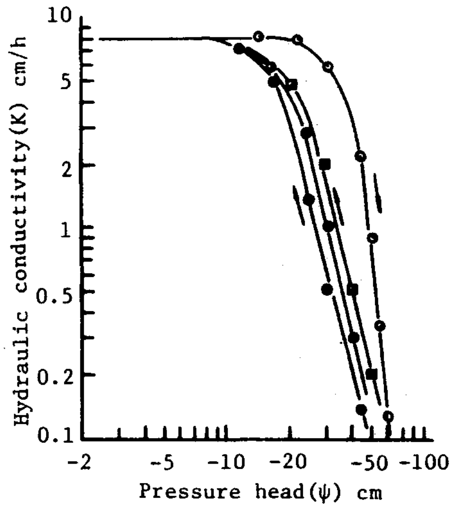


Fig.22 K(ψ) relationship for different flow conditions (after Varchard) [15]

Fig.21 K(θ) relationship for different flow conditions (after Varchard) [15]

4.2 Relationships between Pressure Head (ψ) and Volumetric moisture Content (θ)

To get the relationships between pressure head (ψ) and volumetric moisture content (θ), the variation of the volumetric moisture content and that of the pressure head with time are used.

The results from Test 1-1 are reported in Fig.16 (a) and Fig.16 (b) and those from Test 1-3 are shown in Fig.17 and Fig.19. Using the these figures, it should be noted that a relationship (θ-ψ) is obtainable by plotting, for any elevation, the pressure head at successive times. When this process was carried-out for different elevations, the relationship (θ-ψ) on the sand was defined as shown in Fig.23.

In Fig.23, data points (closed squares) for the scanning curves which is obtained from selections of data from Test 1-4 for rewetting, the tensiometer position at 20cm from the top of the column, are added and line was drawn by eye as a best fit.

In soil science, the curves in Fig.23 are called "retention curves", as they show how water is retained in the soil by capillary forces against gravity. Some authors refer to the drying retention curve as a "desorption curve" and to the wetting curve as a "sorption curve".

It is evident from Fig.23 that hysteresis in the relationship between pressure head and volumetric moisture content exises in soils. The reasons why the hysteresis loop exists in the retention curve may be as follows [41].

(1) The geometry of the void space of soil with many bottlenecks, i.e., the ink bottle effect.

(2) The angle of contact α_c is a function of the direction of the displacement, α_c may have different values if equilibrium is approached by advancing or receding over a face. Fig.24 shows this phenomenon for an air-water interface.

(3) The air in the void space may be trapped in the process of water advancing.

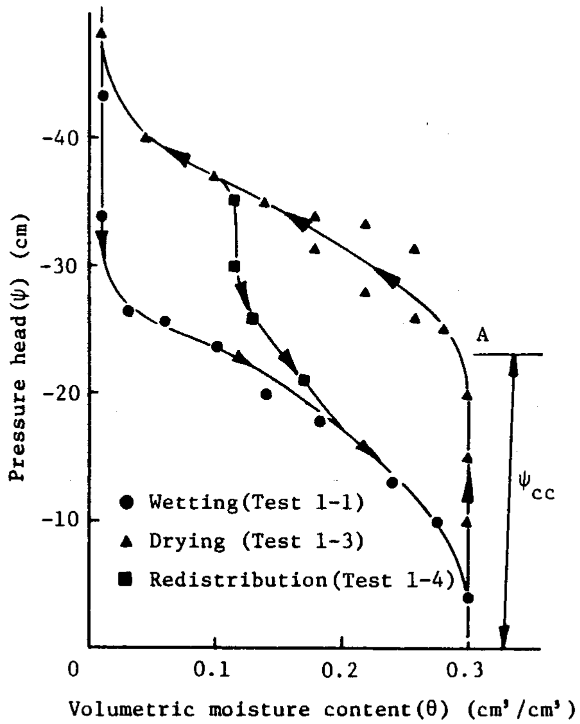
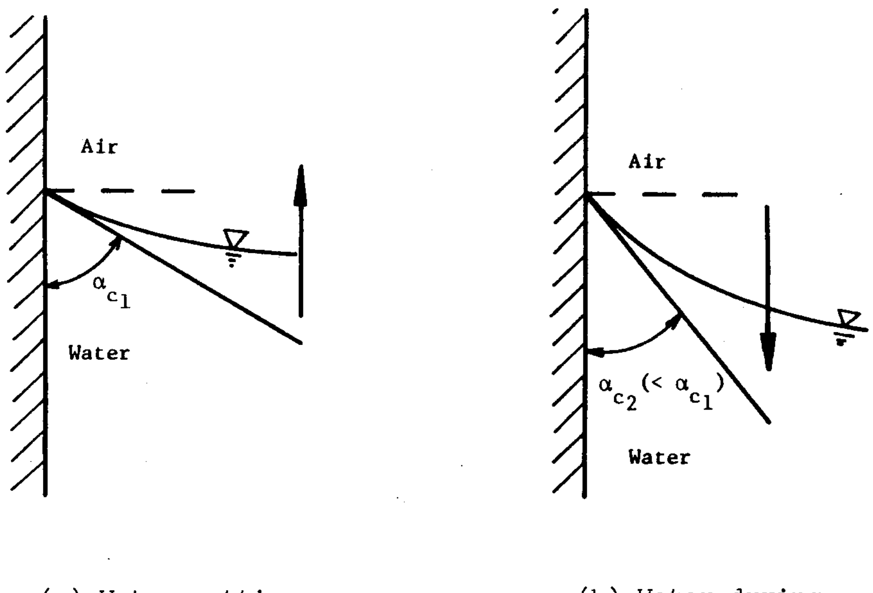


Fig.23 Relationships between volumetric moisture content and pressure head



(a) Water wetting (b) Water drying

Fig.24 Contact angle between a water-air interface and a solid

Point A in Fig.23 is the critical capillary head ψ_{cc}. If the drainage process is start from a saturated sample, no water will leave the sample until the critical capillary head is reached. As the value of ψ_{cc} is increased, the initial small reduction in θ is associated with the retreat of the air-water meniscii into the pores at the external surface of the sample. Then, at the critical value ψ_{cc} the larger pores begin to drain.

The reason of this phenomenon may be the time lag in air entry into soil. The soil pores are probably smaller at the soil surface than they are in the interior of soil mass. There is a "skin effect" due to the smaller surface soil pores. When the surface soil pores are emptied, the water declines in the capillaries for a considerable distance downward, since the soil pores are larger beneath the soil surface. The soil water tension at the soil surface would very quickly increase. The entry of air into the soil would occur shortly after the water table reached the bottom of the soil column; assuming that the length of the column is such that the air-entry value of the soil is exceeded.

In same manner of the relations (K-θ), various empirical equations for the relations of pressure head to moisture content have been proposed, as indicated in Table 3.

Table 3. Empirical equations for the relation of pressure head to moisture content

Empirical equation	Reference
$\frac{\theta - \theta_r}{\theta_s - \theta_r} = \exp[\,a(\psi - b)]$	Kroszynski,U. [39]
$\theta = a\left[\frac{\cosh[(\psi/b)^m + c\,] - d}{\cosh[(\psi/b)^m + c\,] + d}\right]$	King,L.G. [35]
$\theta = a + b\log_e(\psi - c)$	Rogowski,A.S. [42]

where

a,b,c,d, and m : empirical parameters

The main loops in Fig.23, drying curve and wetting curve which are obtained by making the air-dried soil saturated and the saturated soil dried respectively, are unique for a given soil and can be determined experimentally. However, if, at any point on the drying (or wetting) curve the process is reversed, hysteresis, as shown by the scanning curve in Fig.23, is exhibited. In the wetting process the θ versus ψ relation moves along the scanning curve until it meets the wetting curve and then moves up the wetting curve. The drying and the wetting curves form the boundary of the "hysteresis loop", within which the soil can assume any value of saturation and ψ, depending on the past history of the procass. Although a number of analyses which took into account the phenomenon of this hysteresis have been done for the unsaturated soil [43-46], all of them have used experimental hysteresis data directly and their procedures are very complex.

On the other hand, several attempts to estimate any hysteresis loop from the main drying and wetting curves have been proposed. The most of them are same methods in applying "independent domain theory" which was formulated by Everett [52]. Among them, Mualem's method [51] seems to be the simplest and the most efficient.

4.3 Evaluation and Discussion on Green and Ampt Model and Pressure Distribution at Equilibrium Condition

4.3.1 Green and Ampt model

The Green and Ampt model [53]of infiltration has been the subject of considerable attention in recent literature with the need to model simply the infiltration component in hydrological studies [54-57].

The model is based on the following form of Darcy-type equation [58] for a uniform profile in vertical infiltration as depicted in Fig.25,

$$q = K_s \frac{H_0 - H_c + z_f}{z_f} \tag{5}$$

where q is the rate of the infiltration into the soil; K_s is the hydraulic conductivity (which is a uniform soil under ponding approximates the saturated conductivity); H_o is constant depth of ponded water at the soil surface; z_f is the distance from the surface to the wetting front (i.e., the length of the wetted zone); H_c is negative pressure head at the wetting front.

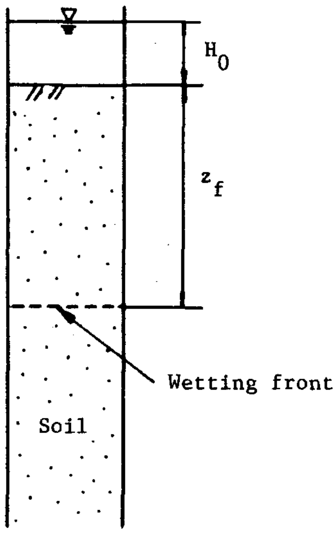


Fig.25 Vertical infiltration

This model assumes that the flow of water occurs in a uniformly saturated region on account of a hydraulic head gradient caused by a constant soil water pressure head at the wetting front and gravity. The advancing wetting front is thus a precisely defined surface boundary above which the soil is saturated and below which the soil is unsaturated. These assumptions simplify and linearize the flow equation, making it amenable to analytical solution.

In order to find the model to be satisfactory for infiltration through an unsaturated soil, the result of the advancing wetting front with respect to Test 1-1 is shown in Fig.26. By using Figs.16(b) and 26, the distributions of total head ($h=\psi+z_f-z$) are shown in Fig.27. It is obvious in Fig.27 the graduate of total head is constant in the wetting zone for each time, and H_c is equal to the initial pressure head of air-dry soil. Fig.28 shows the infiltration (Q) plotted as a function of time. According to Fig.16, it is evident that behind the wetting front, the soil is uniformly wet. Since an uniformly wetted zone can be assumed to extend all the way to the cumulative infiltration Q should be equal to the product of the wetting front depth z_f and the wettness increment $\beta=\theta_t-\theta_i$(where θ_i is the transmission-zone wetness during infiltration and θ_i is the initial profile wetness which prevails beyond the wetting front):

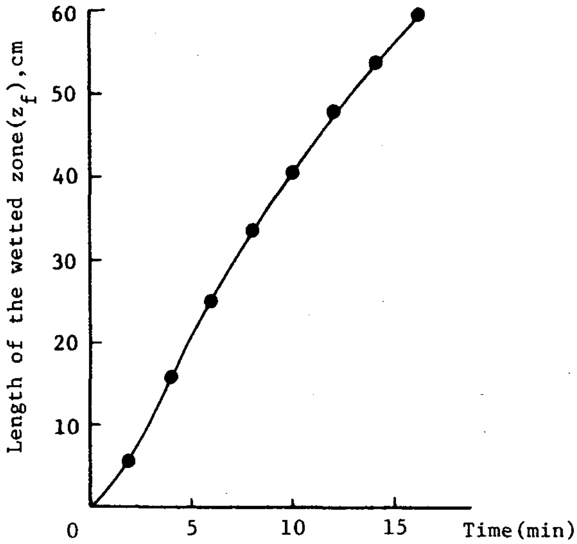
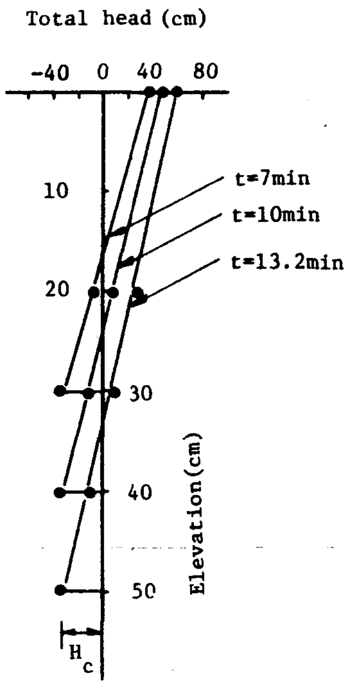


Fig.26 Advancing of wetting front with time



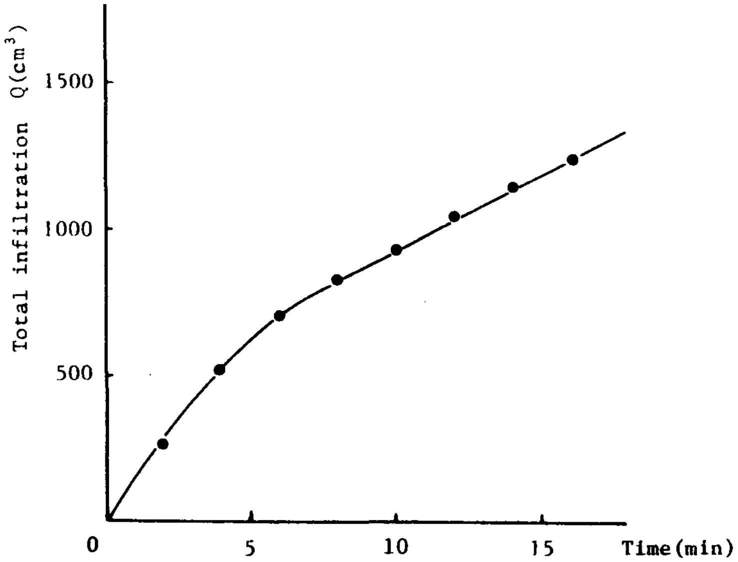


Fig.28 Infiltration Q plotted as a function of time

$$Q = A\beta z_f \tag{6}$$

where A is the area in cross section of soil column. Therefore,

$$\beta = \frac{Q}{Az_f} \tag{7}$$

By using Fig.26 and Fig.28, the average value of β was obtained as 0.31, and this value is nearly equal to the value getted by gamma ray attenuation.

The infiltration rate is thus seen to be inversely related to the cumulative infiltration. Rearranging Eq.(5), next equation is obtained:

$$q = \frac{dQ}{dt} = K_s \frac{H_0 - H_c + z_f}{z_f} A \tag{8}$$

Then the value of K_s for each time can be calculated by using the both results of Q versus t and z_f versus t.

The average value of K_s is obtained as 1.9×10^{-2}cm/sec. This value also well agrees with the value obtained by Test 1-2. Thus it can be understood that attempts to reconcile the Green and Ampt model with the classical flow equation is successful for the infiltration through the air dry sandy soil. However, in actual field conditions, particularly where the initial moisture content is not uniform, H_c may be undefinable. In many real situations, the wetting front is too diffuse to indicate its exact location at any particular time. Then this model remains a quasi-empirical method founded on crude theory.

4.3.2 Pressure distribution at equilibrium condition

In order to perform the nonsteady numerical analysis of flow through porous media, the distribution of initial pressure head ψ_o in the entire flow domain must be used as the initial condition. Test 2-1 was carried out to obtain this distribution and infiltration from the base of the column occurred for 2 days. Fig.29 gives the advances of capillary zone.

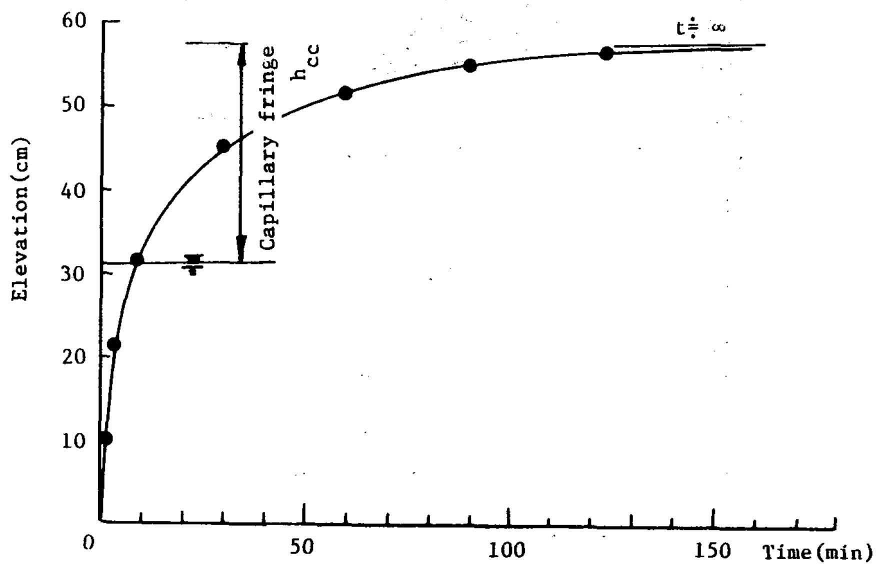


Fig.29 Advances of capillary zone

As this phenomenon is analogous to rise in capillary tube where the water rises to a certain height above the free surface with a full saturated tube below the meniscus, the nearly saturated zone above the free surface is called the capillary fringe. This h_{cc} in Fig.29 is the capillary rise for this soil and its value became about 25cm.

Fig.30 shows the change of moisture distribution above the free surface with respect to time, and after 2 days this change reached the near static equilibrium condition. Fig.31 represents the distribution on pressure head at this condition. In Fig.31, as z increases upward, so decreases ψ at the ultimate straight line was gotton as the relationship (ψ-z) in the capillary fringe. Then, in this zone the total potential (h) may be

$$h = \psi + z = 0 \tag{9}$$

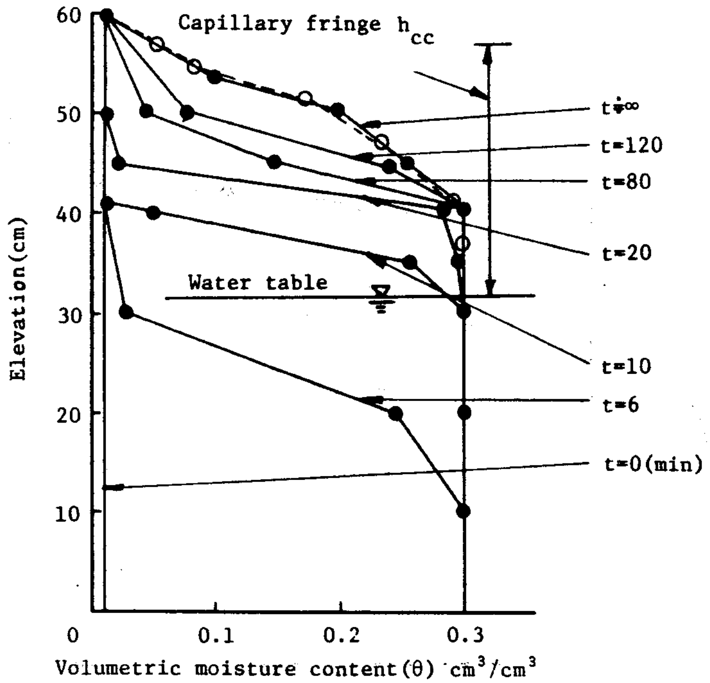


Fig.30 Distribution of water content for standard sand during infiltration from bottom (Test 2-1)

The distribution of pressure head above the capillary fringe must be ψ_0 which is initial pressure head of air dry soil because the volumetric moisture content did not change during this successive time.

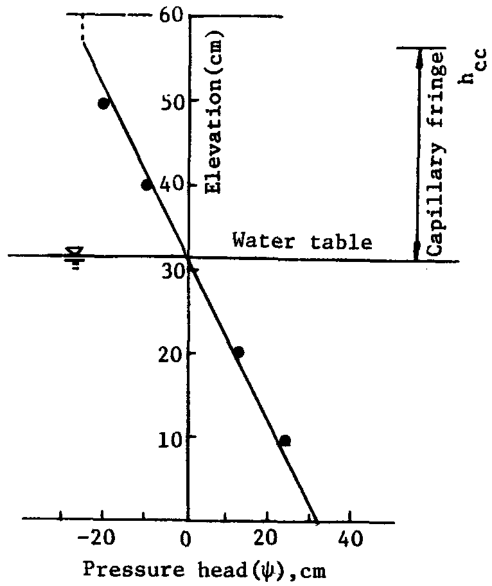


Fig.31 Pressure head distribution in capillary fringe at equilibrium condition

By the way, the open circles points in Fig.30 were plotted of the value of moisture content from the wetting curve of Fig.23 considering the value of pressure head is equivalent to the height above free surface. From this result, the equilibrium moisture profile can therefore be applicable to the retention curve of wetting process.

5. Conclusions

In this paper, the need for determining the hydraulic properties of soil profiles was pointed out and available methods are reviewed. Experimental tests have been performed to determine the hydraulic properties of unsaturated soil during the flow of water in a vertical soil column and a technique for handling the data was systematically described and illustrated. Throughout of this paper, the following conclusions are obtained.

(1) The instantaneous profile method for determining soil hydraulic properties based on simultaneously monitoring the changing moisture content and pressure head profiles during internal drainage is outlined and this method is probably easier to carry out than altanative methods.

(2) A technique for using a low-energy gamma radiation apparatus for measuring water content in the column of soil has been described. Although its use in limited to porous materials which do not shrink or swell upon wetting, it does provide a means for accurate measurement of water content without disturbing the system into which water is moving. Furthermore rapid measurement of water content becomes possible at any position in a soil so that water content changes with time may easily be followed.

(3) The tensiometer-transducer system provides a most valuable means of measuring pressure head with rapid response and with provision of a complete record of the pressure head changes with time.

(4) The hydraulic conductivity in relation to volumetric moisture content of the sand was obtained.

(5) Different flow conditions have been imposed to map out conveniently the hysteresis domain in $\psi(\theta)$.

(6) Applications of the Green and Ampt infiltration equation were discussed and good agreements between the model and experimental data were shown, then it was found that the Green and Ampt approach is satisfactory for infiltration through the air-dry sand column where the initial moisture content is uniform.

(7) The distribution of pressure head and moisture content above the free surface was obtained at the equilibrium condition in order to apply this distribution to the numerical analysis of drainage and infiltration in soil as a initial condition.

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