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Oliver A. Opara-Nadi

A comparison of some methods for determining the hydraulic conductivity of unsaturated soils in the low suction range.

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IMPORTANT SYMBOLS

q	=	flux ($cm^3 \cdot cm^{-2} \cdot day^{-1}$)
$k_{(\psi m)}$	=	hydraulic conductivity as a function of matric suction ("unsaturated hydraulic conductivity") (cm/day)
D	=	soil moisture diffusivity as a function of water content (cm^2/min)
H	=	hydraulic head (cm of water column)
ψ	=	tension or suction "negative (or) subatmospheric pressure" (cm of water column) or simply ("cm water")
ψm	=	matric suction (cm water)
ψg	=	gravitational potential (cm water)
θ	=	volumetric water content (cm^3/cm^3)
W	=	volume water content (cm^3)
C	=	specific water capacity ($cm^3 \cdot cm^{-3} \cdot cm^{-1}$)
Q	=	quantity of water flowing per unit time (cm^3/t)
A	=	cross-sectional area of soil sample (cm^2)
L	=	lenght of the soil sample (cm)
V	=	volume of the soil sample (cm^3)
t	=	time (min or day)
μ	=	micron
X	=	distance in the horizontal direction (cm)
Z	=	distance in the vertical direction (cm)

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1. INTRODUCTION

The ability of an unsaturated soil to transmit water is measured by its conductivity. According to the theory of flow of water in unsaturated soil, the hydraulic conductivity as a function of either water content or suction is needed for the quantitative description of soil water behaviour. Soil water behaviour can be of interest under different aspects such as:

1) The determination of the components of the water balance or water conservation in the field which gives the overall mass conservation for rainfall falling in any period. Most of the components in the water balance are dependent upon the soil. These include infiltration, evaporation, transpiration, net surface runoff, subsurface flow, seepage and the increase or decrease in the soil water storage.

2) The processes which involve the soil-water flow in the rooting zone of most plant habitats and therefore the supply of water to plant roots.

In order to understand these flow processes, knowledge of the hydraulic conductivity of the soils is required. This work is mainly intended to deal with a particular problem which is centered on the investigations of the water balance in the field, or in other words, on the determination of the components of the water budget of ecosystems. As a consequence, interest is focussed on flow processes in the soil occuring in the low suction range. This is because as will be shown later and as is well known from past experience and literature, the hydraulic conductivity of most soils drops sharply over suction range of $0 < \psi < 100$ (cm of water) to very low values. Consequently the hydraulic conductivity becomes less significant with regards to the contribution of the corresponding volumetric flux to the water budget. By an annual rain fall of 700 mm (as is often encountered in Central Europe) as an example, an increase in the soil water storage, that is, a water balance of the order of 30 mm may be regarded negligible since this amount is about the magnitude of unavoidable error in the other components. Applying this to the rate of seepage across the bottom of soil profile, it becomes evident that hydraulic conductivity values below about 0.01 cm/day are of little interest within the outlined scope. This allows then the hydraulic conductivity determinations to be limited to the suction range mentioned above.

The principles and methods of measurement of hydraulic conductivity of unsaturated soil have long been known and given in the literature. The techniques, however, are difficult and subject to improvement in many ways (Klute, 1965b). Many of the methods often require a long time before experimental data are obtained or data obtained relatively quickly often involve calculations which are complicated and time consuming (Weeks and Richard, 1967; Renger et al., 1972; Becher, 1975). As a result, none of these methods has brought a break through for practical use in the determination of the hydraulic conductivity of unsaturated soils in the laboratory. Rather a combination of different methods are used especially when the hydraulic conductivities of the soil in the higher suction range are also to be determined.

It is therefore the purpose of this work to study the application of some of these laboratory techniques or methods on some soils and to compare the results obtained using statistical and other methods. Such comparisons allow conclusions to be made on which method or methods is/are best suited for the routine determination of hydraulic conductivity. Since some of the methods found in the literature are subject to assumptions, which may restrict the applicability of these methods, and/or boundary conditions that may have different effects on different soils, it was felt necessary and advisable to test different methods by comparing their results with a reference method. The steady-state method using short samples was chosen as a reference method since it was felt to be the "safest" in that it requires no assupmtions as well as allowing a close control of the flow conditions by the use of tensiometers on both ends of a homogeneous sample. A disadvantage of this method is that it is time consuming and requires complicated equipments. Therefore it is found necessary to look for another method which may be a good substitute for the steady-state method. It can also be seen that another factor which plays a role in defining the objective of this work is related to the equipments available for the determination of the hydraulic conductivity.

Since the problem of the spatial variability of sample results (particularly in field investigations) plays a role and is therefore necessary to deal with a number of replicates large enough to satisfy statistical needs, there is a strong desire to make use of methods which are quick. And as such this comparison is intended especially to help in deciding which method is an alternative to the steady-state methods with small (homogeneous) samples.

A final spect of this work involves the determination of the hydraulic conductivity of large (inhomogeneous) samples as is often encountered when soil samples from two soil horizons and their boundary layer are taken. Secondly, very often and particularly under forest stands, stony soils, which do not allow soil sampling in the usual way, have to be investigated hydrologically. To get undisturbed soil samples in this case, a special technique has been developed (Benecke et al., 1976) using polyester asbestor-fibre to coat the samples in situ. The procedure is described later (see section 4.3.3.). In this case too a steady-state method has been used to determine the hydraulic conductivity of such large soil samples. This is even more time consuming than the small sample versions. Therefore there is a strong desire to look for a more effective, that is, a quicker method.

As was mentioned earlier, several methods and methodical versions have been proposed by many authors. It is therefore necessary to review some of the most important methods found in the literature as to identify the principles, advantages and disadvantages of these methods. Since an understanding of the principles of these methods depends to a large extent upon the knowledge of the theoretical fundamentals, the latter will be discussed first.

2. THEORY

According to Klute (1972) methods for determining the relationship between water condictivity of unsaturated soils and water content (θ) or suction (ψ) may be classified as steady-state, transient or unsteady-state, and computational methods, based on pore-size distribution data. In the steady-state flow system, flux, gradient, and water content remain constant with time, while in the transient flow, they vary.

In steady-state flow methods, the volumetric flux and hydraulic gradient are measured in a system of time-invarient one-dimensional flow, and the Darcy equation is used to calculate the hydraulic conductivity. The value of the conductivity obtained is associated with the suction head and/or water content at the position and time at which the flux and gradient are measured. In the unsteady-state the volumetric flux and the water content change simultaneously and by these methods, the time dependence of some aspect of the behaviour of the flow system is used to obtain the conductivity.

The concept of soil-water diffusivity was originally suggested in a paper by Childs and Collis-George (1950) and like the hydraulic conductivity, it can be obtained from both the steady and transient state flows.

Since it is the aim of this study to test some methods and concepts for determining the hydraulic conductivity of unsaturated soils, further theory is dealt with only as far as it is necessary to understand the theoretical principles of those methods which have been used in this work. In particular it is advisable to examine the assumptions and the initial and boundary conditions associated with each of the methods, and to have idea about the particular physical conditions to which flux is conisdered to be subjected and therefore to see if they can provide a clue explaining deviations of the results.

The hydraulic conductivity is defined by the Darcy equation extended for use in unsaturated soils by assuming that the conductivity is a function of the water content or the soil water suction. In a one-dimensional vertical flow system, the volumetric flux q(z,t) at a position z and time t is given by

$$q_{(z,t)} = -K_{[\psi m(\theta)]} \frac{\delta\ m}{\delta z} + 1 \qquad (1)$$

It is matter of choice to express the hydraulic conductivity k as a function of the soil suction ψm or the water content θ. But since it is often easier to control the water status as a function of location and time by tensiometers rather than measuring the water content, the hydraulic conductivity is subsequently mainly given as a function of soil suction. When the water content is needed, it is inferred from the soil water characteristic $\psi m(\theta)$. In order to avoid hysteresis effects, methods of the desorption type only have been examined. The disadvantage of tensiometers, having a limited suction range (0-800 cm of water), is considered not to be of great significance since mass flow of water, which contributes substantially to the field water turnover, occurs well within this range. In fact, as mentioned in the introduction, it suffices to know in most cases the $k_{(\psi m)}$ curve for $\psi < 100$ cm water column, because seepage rates below the root zone of soils normally drop to negligible values once the suction approaches about 100 cm water column.

The effect of gravity is considered in equation (1) by "+ 1" (or $d\psi g/dz$, where ψg is the gravitational potential and changes at the same numerical rate as z, the distance along the flow path. The + sign means that z is taken positive upwards. Darcy's law as stated in equation (1) can be applied directly to methods based on a steady-state volumetric water flux through the soil sample. As will be explained later, these methods require complicated devices and often a long time is required for steady-state conditions to be reached in a soil sample. Their advantage is that no simplifications or assumptions need to be made and the flow conditions can well be controlled.

Nevertheless, a number of unsteady-state methods have been developed. The advantage of these methods lies predominantly in the possibility to obtain experimental data relatively quickly over a wide range of suction and with relatively simple devices. Furthermore, an unsteady-state concept is needed to carry out measurements in the field, where steady-state flow conditions are rarely, if ever, realized.

The basic equation describing unsteady-state water flow through unsaturated soils is the equation of continuity, often referred to as a form of the law of conservation of matter

$$\frac{\delta\theta}{\delta t} = - \frac{\delta q}{\delta z} \qquad (2)$$

This equation gives the volumetric balance for soil water by relating the change of volumetric flux over a (very short) distance to the change in water content with time within this distance.

Combining equation (1) and (2) the one-dimensional form of the general flow equation results

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z} \; K_{(\psi m)} \; \frac{\delta\psi m}{\delta z} + 1) \qquad (3)$$

In order to change this nonlinear differential equation, for which solutions are hard to find, into a form, that can be solved mathematically, the matric suction gradient can be expanded by the chain rule

$$\frac{\delta\psi m}{\delta z} = \frac{\delta\psi m}{\delta\theta} \; \frac{\delta\theta}{\delta z} \qquad (4)$$

where $\delta\psi m/\delta\theta$ is the slope of the water characteristic curve or the reciprocal of the specific water capacity C

$$C = \frac{\delta\theta}{\delta\psi m} \tag{5}$$

and $\delta\theta/\delta z$ can be called wetness grandient. The diffusivity D is given as

$$D_{(\theta)} = K_{(\theta)} \frac{\delta\psi}{\delta\theta} = \frac{K}{C} \tag{6}$$

Substituting equation (6) in (3) the soil-water diffusivity form of the flow equation is obtained

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z} D \frac{\delta\theta}{\delta z} + \frac{\delta K}{\delta z} \tag{7}$$

The term $\delta K/\delta z$ accounts for the effect of gravity. This can therefore be neglected in a horizontal flow system or if the suction gradient is large compared to the gradient due to gravity. Equation (7) then becomes

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta x} D \frac{\delta\theta}{\delta x} \tag{8}$$

where z is replaced by x, to indicate that gravity is not considered. Assuming that D is constant over a certain range of water content or suction, equation (8) can be written in a form analogous to Fick's second law of gaseous diffusion

$$\frac{\delta\theta}{\delta t} = D \frac{\delta^2\theta}{\delta x^2} \tag{9}$$

Solutions to this now linear differential equation are known and some methods for determining the diffusivity of soil water to be described in the literature are based on this theoretical analysis. Characteristically, they require that the equations to be used be limited to small intervals of water content or suction for which D can be assumed to be approximately constant.

The mathematical derivations of the equations are given in detail by Kirkham and Powers (1972). These authors show that the method of separating variables is used for the analysis.

The difficulty of limiting the D values to small values of water content or suction is eliminated by another analysis also be given by Gardner (1962) and on which the method for determining diffusivity as a function of water content given by Doering (1965) is based. This analysis again starts with equation (8) and also uses the method of separating variables. In order to do this, the following assumption is neccessary

$$D_{(x,t)} = D_x(x)\ D_t(t) \tag{10}$$

that is, it is assumed that the diffusivity can be represented by the product of two components D_x and D_t, where D_x is a function of x (distance) only and D_t is a function of t (time) only.

These assumptions and the following initial and boundary conditions were used by Gardner

$$\begin{aligned} \psi m &= \Delta\psi m && \text{for } z > 0,\ t = 0 \\ \frac{\delta\psi m}{\delta z} &= 0 && \text{for } z = L,\ t > 0 \\ \psi &= 0 && \text{for } z = 0,\ t > 0 \end{aligned} \tag{11}$$

where $\Delta\psi m$ is the applied pressure or suction on the top surface of a wet soil sample and L is the top of the soil sample. In the above conditions, the effect of gravity is neglected and Gardner derives his equation (5) (Gardner, 1962) which reads

$$\frac{1}{D(\theta - \theta f)} \frac{d\theta}{dt} = \frac{-\pi^2}{4L^2} \tag{12}$$

where θf is the final equilibrium water content. Rearranging equation (12) and multiplying θ by the volume of the sample, the equation used by Doering was obtained

$$D = \frac{4L^2}{\pi^2(W-Wf)} \frac{dW}{dt} \tag{13}$$

where L is the length of the sample, W the volume water content when the instantaneous outlow rate dW/dt is determined, and Wf the final equlibrium volume water content.

Another mathematical way to deal with equation (8) is the use of the Boltzmann transformation for a semi-infinite flow system. The Boltzmann transformation is defined as

$$\lambda = xt^{-0.5} \tag{14}$$

where λ is a function of x and t. Since θ is also assumed to be a function of x and t, θ can also be expressed as a function of λ

$$\theta = f[\lambda_{(x,t)}] \tag{15}$$

Gardner (1959) outlines the mathematical solution to the evaporation technique for a semi-infinite soil column which was used by Arya et al. (1975) to propose and test a method for determining soil-water diffusivity. Gardner's development has been for horizontal evaporation flow, but the development can, in many cases, also apply to vertical evaporation flow as Gardner points out. He says, referring to vertical flow, "The effect of gravity is negligible in the early stages of evaporation when the capillary conductivity is uniform throughout the medium and again in the later stages when the gravitational gradient is small compared with the tension or suction gradient." (Kirkham and Powers, 1972). Kirkham and Powers give a somewhat more detailed description of Garnder's theoretical analysis. According to them equation (8) is solved for the following initial and boundary conditions

$$\begin{aligned} \theta &= \theta_i \quad \text{for } x > 0,\ t = 0 \\ \theta &= \theta_o \quad \text{for } x = 0,\ t > 0 \end{aligned} \tag{16}$$

where θ_i is the initial water content, assumed to be constant throughout the semi-infinite soil sample at time $t = 0$, θ_o is the constant water content at the dry surface of the soil and x is the distance.

With the help of the Boltzmann transformation equation (8) is changed to an ordinary differential equation. Accordingly, the initial and boundary conditions change to

$$\begin{aligned} \theta &= \theta_i \quad \text{for } \lambda = \infty \\ \theta &= \theta_o \quad \text{for } \lambda = 0 \end{aligned} \tag{17}$$

After some lengthy mathematical derivation in which it is important to note that Gardner's assumption requires that D can be expressed as

$$D_{(\theta)} = D_o \exp\ \beta \frac{\theta - \theta_o}{\theta_i - \theta_o} \tag{18}$$

where D_o is the diffusivity at the boundary with a moisture content θ_o (at dry soil surface, that is, controlling the flux across the surface) and β is an emperical constant.
As a result of the theoretical analysis, the flux q across the soil surface (evaporation) is given by

$$q = \frac{1}{2} \frac{D_o}{t} (\theta_i - \theta_o) \left.\frac{dc}{d\lambda}\right._{\lambda} = 0 \tag{19}$$

where $c = (\theta - \theta_o)/(\theta_i - \theta_o)$, the relative moisture content.

Equation (18) can be tested by plotting the cumulative evaporation against the square of time. This is also an experimental condition for the application of the evaporation method given by Arya et al. The calculation of the diffusivity as a function of moisture content is based on an analysis of the plot of the water content distribution versus distance. As is expected, the water content at the sealed end or bottom of the sample must be equal to θ_i at the end of the evaporation, that is the sample must behave as if it were semi-infinite. The diffusivity as a function of water content is calculated by using the equation given by Bruce and Klute (1956)

$$D_{(\theta x)} = - 1/2\, t\, \frac{dx}{d\theta} \int_{\theta_i}^{\theta_x} x d\theta \qquad (20)$$

The continuity equation (2) may be applied to one-dimensional flow and integrated to yield:

$$q_{(z_2,t)} = q_{(z_1,t)} - \int_{z_1}^{z_2} \frac{\delta\theta}{\delta t}\, dz \qquad (21)$$

where $q(z_2,t)$ and $q(z_1,t)$ are the fluxes at positions z_2 and z_1 and time t. If the water content distribution $\theta_{(z,t)}$ is known the integral can be evaluated and if either $q(z_2,t)$ or $q(z_1,t)$ is known the other flux can be calculated. If the hydraulic head distribution $H_{(z,t)}$ is also known, the hydraulic gradient at a given position and time can be evaluated. Equation (1) can then be used to calculate the hydraulic conductivity at any given position and time. The method can be applied to soil columns in the laboratory and to soil profiles in the field. This method 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. The known flux that is required at one position may be obtained by 1) closing one end of the flow system (q = 0), 2) finding positions in the flow column at which the hydraulic gradient and hence the flux is zero or 3) measuring the flux crossing a boundary. The second case arises in a situation where evaporation and downward drainage occur simultaneously as in the soil profile. Under this condition and where the position of the "zero flux" or "water divide" plane is known, the hydraulic conductivity can be obtained. As the soil dries, the plane of "zero flux" moves downwards. The hydraulic conductivity at any depth may be calculated as follows: Let z_o, z_a and z_b represent the position of the "zero flux" plane, the soil depth above and the soil depth below the "zero flux" plane, respectively. With depth z taken positive upwards conductivities at depths z_a and z_b are biven by

$$K(_{\psi za}) = \int_{z_o}^{z_a} (\delta\theta/\delta t)\,dz \;\Big/\; \frac{\delta\psi m}{\delta z} + 1 \;{}_{z_a} \qquad (22a)$$

and

$$K(_{\psi zb}) = \int_{z_o}^{z_b} (\delta\theta/\delta d)\,dz \;\Big/\; \frac{\delta\psi m}{\delta z} + 1 \;{}_{z_b} \qquad (22b)$$

respectively.

3. REVIEW OF LITERATURE

3.1. Methods based on steady-state flows

Laboratory techniques for determining hydraulic conductivity based on the steady-state flow systems have been given by Richards and Moore (1952), Nielsen, Kirkham and Perrier (1960) and Nielsen and Bigger (1961). Other methods based on the steady-state were proposed in addition, but most of them make use of disturbed samples (Moore, 1939; Staple and Lehane, 1954; Bruce and Klute, 1956; Childs and Collis-George, 1950; Young, 1960 and 1964). When the conductivity data of disturbed soil samples are used for the calculation of moisture transport in a well-defined soil profile, with its typical sequence of layers, each one of them with a pronounced structure, appreciable difference with what actually happens in the profile will be found. According to Butijn and Wesseling (1959), the only way out of this case is, to determine the hydraulic conductivity of each layer of the profile on an undisturbed soil sample. The general level of the water content and pressure head may be controlled by hanging water columns (Richards, 1931), by use of a bubble tower arrangement (Rose, 1966) or by application of a controlled gas phase pressure greater than atmosphere to the sample (Elrick and Bowman, 1964; Richards and Moore, 1952).

The long column version of the steady-state method has been given by Childs and Collis-George (1950). In this method, the pressure head could be measured with one appropriately placed tensiometer. By starting at saturation and proceeding through a series of progressively decreasing flow rates, one can determine a series of points on the drainage $K_{(\psi)}$ function. As a variation to the above method, Klute (1972) suggested that a series of tensiometers at convenient intervals along the column could be used to determine the head difference across each interval and thus obtain the conductivity function for each section without making the assumption of uniformity of the sample.

3.2. Transient or unsteady-state flow methods

3.2.1. Measurement of the hydraulic conductivity

The direct determination of the hydraulic conductivity using unsteady-state flow system has been given by Gardner (1956). The conductivity values were calculated from the pressure plate outflow data. Alternatively and as is mostly the case, one can determine diffusivity (Childs and Collis-George, 1950) and convert it to conductivity using equation (6).

3.2.2. Measurement of the diffusivity

The diffusivity can be measured using pressure-plate or pressure membrane outflow data. Gardner (1956) was first to publish a method for calculating D from pressure-plate outflow data. He assumed membrane and plate impedance to be negligible. This method has a drawback in that the resistance to water flow of the flow membrane of the apparatus affects the determination of D for the soil. Miller and Elrick (1958) extended Garnder's technique to include the case of nonnegligible membrane impedance. The techniques of Gardner and of Miller and Elrick were further extended by Rijtema (1959) and by Kunze and Kirkham (1962) to include contact as well as membrane impedance and to eliminate a separate experimental estimation of the impedance. As was mentioned in the theory, in Gardner's technique as well as in these other modifications of this technique, it is assumed that for a small enough suction change and moisture content change, D can be considered constant. Jackson et al. (1963) and Davidson et al. (1966) have examined the Gardner method and concluded that it is not practicable to use pressure increments small enough to validate the assumption of constant diffusivity.

Methods for determining D which do not make use of small increments of pressure, that is, no constant diffusivity is assumed, have also been given. Gardner (1962) and Doering (1965) devised a way to determine D using only one pressure increment. This technique requires the estimation of only the instantaneous rate of outflow, the moisture content and the sample geometry to determine D, but negligible membrane impedance is assumed.

Bruce and Klute (1956) used the Boltzmann transformation and calculated D from a moisture distribution curve plotted from data obtained from the addition of water to horizontal sand columns. Using the same basic approach and a slightly different analysis, Gardner (1959) showed that a diffusivity-water content relationship could be obtained by evaporating water from one end of artificially packed soil columns initially at uniform water content throughout their length. Diffusivities were also calculated from a water content distribution.

Arya et al. (1975) developed and tested this evaporating technique on undisturbed soil cores. These authors used equation (20) with the initial and boundary conditions, equation (16). Recently Ehlers (1976) also tested the method by Arya using undisturbed soil samples taken from tilled and untilled plots.

3.3. Instantaneous profile methods for hydraulic conductivity determination

Methods for this type [equation (21)] for determining the hydraulic conductivity in the field have been given by Richards et al. (1956), Ogata and Richards (1957), Rose et al. (1965), van Baval et al. (1968), Nielsen et al. (1962) and Renger et al. (1970). By these methods the known flux which was required at one position, was obtained from either covered soil surface (Ogata and Richards, Nielsen et al.) or from depth at which hydraulic gradient was zero (Richards et al., Renger et al.) or from estimated evaporation rate (Rose et al.).

4. MATERIALS AND METHODS

4.1. Research Methodic

Generally, the determination of the hydraulic conductivity of unsaturated soils in the laboratory can be divided into three main steps:

- soil sampling
- measurement
- calculation.

Three different soil sampling techniques were used and these will be described in detail later in this chapter.

The different experimental procedure as well as the calculation involved for the different methods will be given later for each method. In order to compare the different laboratory methods used in this work, a grey brown podzolic soil derived from loess was used. The methods used were based on steady-state and transient-flow systems (as was given also in the review of literature). The steady-state methods included:

Short column - small increment version
Long column - small increment version.

The transient-state methods used were outflow types. These included:

Short column - small increment version
Long column - small increment version
Short column - large increment version (one-step method).

Another method used involved the evaporation of water from vertical soil cores and will be referred to as the evaporation method.
The terminology used for the methods and methodical versions are given in the literature by Klute (1972).
Using the steady-state method (short column version) as a reference method, the applicability of the evaporation method

was tested using three soil types (described in the next section) containing the three predominant classes of soil textures (sand, silt and clay). Furthermore, the variability in results due to the methods themselves were tested. For this purpose, a homogeneous material (in this case, quartz powder) was used.

For the determination of the hydraulic conductivity under field conditions, the instantaneous profile method was used. The technique was the "zero flux" or "water divide" boundary technique and the measurements were done on grey brown podzolic soil on a plot adjacent to where the samples for the laboratory determinations were taken.

4.2. Profile description

The description follows the notation as used in the German "Kartieranleitung" (Arbeitsgemeinschaft Bodenkunde, 1971).

"Parabraunerde"

Grey brown podzolic soil (derived from loess)

Depth (cm)	Horizon	Description
0-28	Ap	tilled horizon, dark greyish brown, loamy silt consisting of a mixture of weak clods and granular, friable aggregates, clear boundary.
28-52	A_{11}	pale brown loamy silt, subangular blocky, very friable, many earthworm channels, gradual boundary.

52-90	A_{12}	light yellowish brown loamy silt, subangular blocky, slightly increased but still low compaction, many earthworm channels, gradual boundary.
90-110	A_1B_t	transitory boundary from A_{12} to B_t-horizon.
110-150	B_{t1}	yellowish brown silty loam, slightly mottled, strong blocky and prismlike stable structure, increased compaction, very many earthworm channels, gradual boundary.
150-175	B_{t2}	yellowish brown silty loam, coarse mostly prismlike structure, many earthworm channels.

"Pelosol"

"Pelosol", according to the German soil classification, is a clayey soil derived from more or less metamorphic plate-like clay as parent material. The dominating feature is the high (> 40 %) clay content, normally showing a well-developed polyhedral structure that changes from fine to coarse, more prismlike aggregates with depth. Only one horizon was chosen from a "Pelosol" derived from the geological formation Röt (belonging to the Trias) in 40-80 cm depth that exhibits average physical properties. The colour is reddish grey, the structure medium to coarse polyhedral, plastic, well developed and uncompacted.

Gley podzol

Podzols are characterized by the presence, just below the surface, of an ashy-coloured horizon and a B-horizon in which illuvial humus and iron (and aluminium) have accumulated. In this way the sandy, original single grain structure is changed into a more or less coherent but otherwise structureless material. Samples were taken from this horizon (between 50-70 cm depth), consisting of reddish-brown to brownish black medium coarse and practically free of silt and clay (< 3 %).

4.3. Soil sampling techniques

Benecke (1963 and 1966) reported that since the hydraulic conductivity of a soil depends to a large extent on the structure and compaction of the soil, much emphasis should be placed on the soil sampling technique. The attempt is at obtaining or using soil sampling techniques which would avoid to alter the structure of the soil as far as possible. To this end, a soil pit was dug which served also for the profile description. Samples for the laboratory determinations were obtained from the different horizons or representative depths using three main sampling techniques.

4.3.1. Short metal cylinder samples

The cylinders which were of two sizes (5 cm long, 8 cm I.D. and 10 cm long, 8 cm I.D.) were pushed into the soil pit at representative depths using a core sampler. A weight was used on the core sampler for pushing the cylinder into the soil with a possible impact (due to vibration) on the soil structure. This effect may show up in the result. The samples were carefully removed and the adhering soil scraped off with a knife. Because this sampling technique was not difficult, more replicate samples (7 to 10) per horizon were taken.

4.3.2. Samples in 30 cm long acrylic glass cylinders

To obtain larger soil samples, acrylic glass cylinders 30 cm long, 19 cm I.D. and 3 mm thick were used. Such large cylinders are used in order to determine the conductivity of two horizons and the boundary region between the two horions (Benne, 1977). A soil column, having a diameter a little wider than that of the cylinder and with the base still attached to the underlying soil was prepared. The cylinder was placed on the soil column. A weight was used on the cylinder to make it slide gently through the soil column until the latter protruded at the top of the cylinder (Fig. 1).

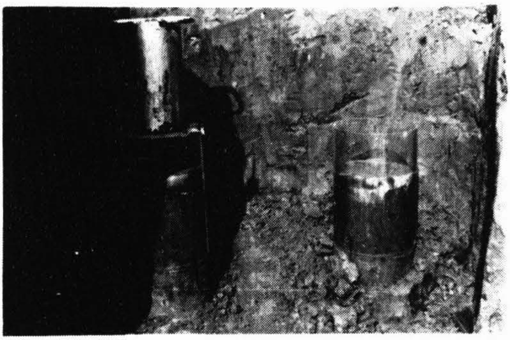


Fig. 1: Soil sampling using acrylic glass cylinder

The sample was then carefully cut off from the underlying soil and trimmed at both ends to the same level as the cylinder. Two such samples were collected from each of the depths 10-40 cm, 55-85 cm, 90-120 cm, 125-155 cm, and 160-190 cm. Due to the difficulty encountered with this sampling technique, many replicate samples could not be taken from each depth. An advantage of the acrylic glass is that it is transparent and tensiometers can be installed at appropriate depths to correpond to soil horizons and transition layers.

4.3.3. Synthetic coated soil monoliths

This soil sampling technique has been proposed by Benecke et al. (1976) and has found practical use especially in soils with stony materials. A small cubical soil monolith about 30 x 30 x 30 cm with its base still in contact with the undisturbed soil, was isolated from the surrounding soil at the required depth. The sample was covered with a thin PVC cellophane held into place with rubber bands and was then finally coated on all sides and top with polyester-asbestos fibre paste (Homeyer et al., 1974). There were no spaces between the soil sample and the polyester-asbestos fibre coating, In soil having stony material, such spaces can occur and should first be filled with the polyester-asbestos fibre paste before a general coating is done (Benecke et al., 1976). The PVC cellophane cover prevented the paste from penetrating into the soil sample. After hardening, the monolith was carefully removed from its underlying soil. The lower surface was then similarly coated with the polyester-asbestos fibre paste. Figure 2 shows the main stages involved in obtaining such monoliths. Benecke et al. found out that even without a PVC cellophane cover, the polyester-asbestos fibre paste could only penetrate about 1 mm into the sample. Since the polyester contracts slightly upon harding, a close contact is established between the sample and the polyester coat. The outer 1 mm thick wall was subtracted from the weight and volume of the sample. The components used to make the polyester-asbestos fibre paste are: Alpolite 303 as polyester resin, 1.5 - 3.5 % MEKP-solution (methylethylketon-peroxide) as solidifier, 1 % cobalt solution to accelerate hardening and asbestos fibre. The mixing proportion recommended by the manufacturer to give a hardening time of 15 minutes at room temperature or above is 17.5 g MEKP, 5.3 g cobalt solution and 1 kg Alpolite 303. Under lower temperature less than 15°C or when the samples are very wet, it is advisable to use a heater to quicken the hardening of the paste. The weight of the samples was between 50 and 70 kg.

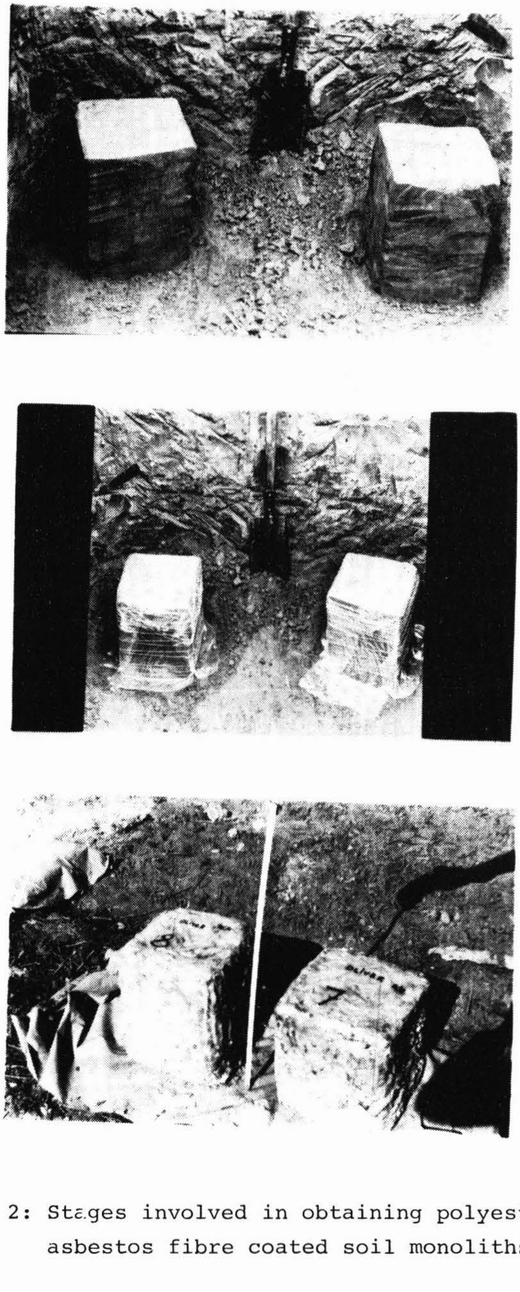


Fig. 2: Stages involved in obtaining polyester-
asbestos fibre coated soil monoliths

4.4. Soil-moisture characteristic

The soil moisture characteristic or soil-moisture-retention curve, also known as the pF-curve, is a graphic representation of the amount of water remaining in the soil at equilibrium as a function of the matric suction. The pF defines the logarithm of the negative suction (pressure or tension) in centimeters of water column.

The soil-moisture characteristic curve was determined in the laboratory by a method based on the method by Richards (1941). The instrument used by Richards is referred to by Rose (1966) as the pressure membrane apparatus. An experimental variation to the method by Richard introduced here was the use of both suction plates for suction range 0 to 330 cm water column and pressure plate extractors for suction of 1000, 3000 and 15000 cm water column. A further variation was the use of the whole undisturbed samples at the lower suction range and disturbed samples at the higher suction range. The 250 ml soil samples, collected by the soil sampling technique described (section 4.3.1.) and from representative soil layers were used for measurement. The samples were first evacuated and then saturated by capillary action, after which they were left to drain freely for 15 to 30 minutes. The samples were weighed and then placed on a ceramic suction plate under a suction of 50 cm water (using a bubble tower arrangement; Rose, 1966). After one week, the samples were weighed and placed on another suction plate subjected to a suction of 100 cm water column and at the third week, the samples were subjected to 330 cm suction. Water loss at each suction level was determined by weighing the samples. After the 330 cm suction level, further desorption was carried out in pressure plate extractors using air-pressures of 1, 3 and 15 atmospheres. Eight small samples were taken from each of the original cylinder samples; two of which were used for water content determination. The remaining six samples were pressed into

PVC-rings (5 mm thick and 20 mm inner diameter). These samples were saturated again and two samples were put in each of three pressure plate extractors and desorption was allowed to occur again for one week. After this period, the water content of each sample was determined gravimetrically (oven dry at 105°C for 24 hours). From this determination, paired values of volumetric water content and suction were obtained and from these, the soil-moisture characteristic for the different soil layers were obtained.

The use of undisturbed and disturbed samples was based on the theoretical fundamentals of the soil-moisture suction relationship. The amount of water retained in the soil at the lower suction levels depends primarily upon the capillary effect and the pore-size distribution, and hence is strongly affected by the structure of the soil. Hence the use of undisturbed soil samples is practically essential. At the higher suctions, water retention is due increasingly to adsorption and is thus influenced less by the structure and more by the texture and specific surface area of the soil material. Thus disturbed soil samples placed in rings were used. Fig. 3a, 3b and 3c show the soil-moisture characteristic of a) artificial quartz powder, b) six depths of the grey podzolic soil and c) one depth of each of the gley podzol and the "Pelosol". In Fig. 3a, it could be seen that the air-entry suction, that is, the critical suction at which the largest pores begin to empty is 100 cm water (pF 2). From pF 2, the shape of the curve is very flat and about 24 % by volume of the pores are drained between pF 2 and 3. The total porosity is 37 % by volume. Fig. 3b shows that these curves are generally steeper (between pF 2 and 3) than the curve shown in Fig. 3a. The total porosity was generally between 41 % and 44 % by volume for all the depths except the 95-105 cm depth, which had a total porosity of 46 % by volume. The shape of the curve from the 10-20 cm depth was generally flatter than all the other curves in the pF range 0.7 to 2.52. In the pF range 0.7 to 1.7 (5 to

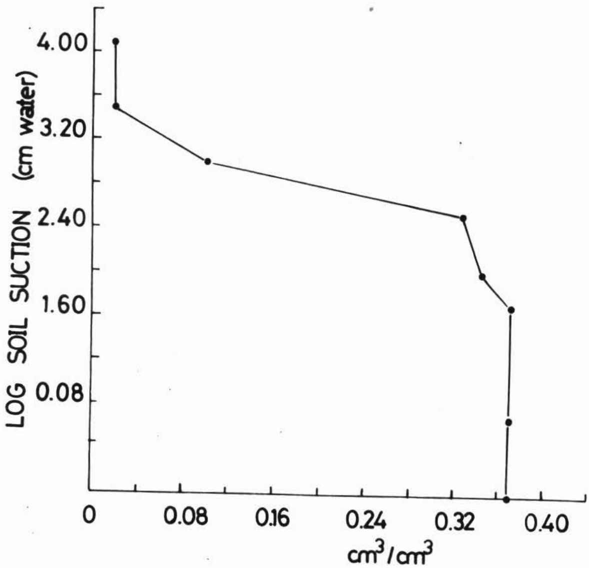


Fig. 3a: Soil-moisture characteristic of artificial quartz powder

50 cm water suction), the curves for the 30-40 cm, 60-70 cm, 125-135 cm and 165-175 cm depths are very steep. This means that a small change in water content corresponds to a very large change in suction. In the 95-105 cm, the curve is flatter than in the four depths mentioned above. Fig. 3c shows again that the shape of the curve is very steep for the clayey soil. On the other hand the curve for the gley podzol is very flat. When the shape of these two curves are compared with those obtained in the grey podzolic soil, the three soil types could be placed in an order according to the steepness of the curves: "Pelosol" > grey podzolic soil > gley podzol.

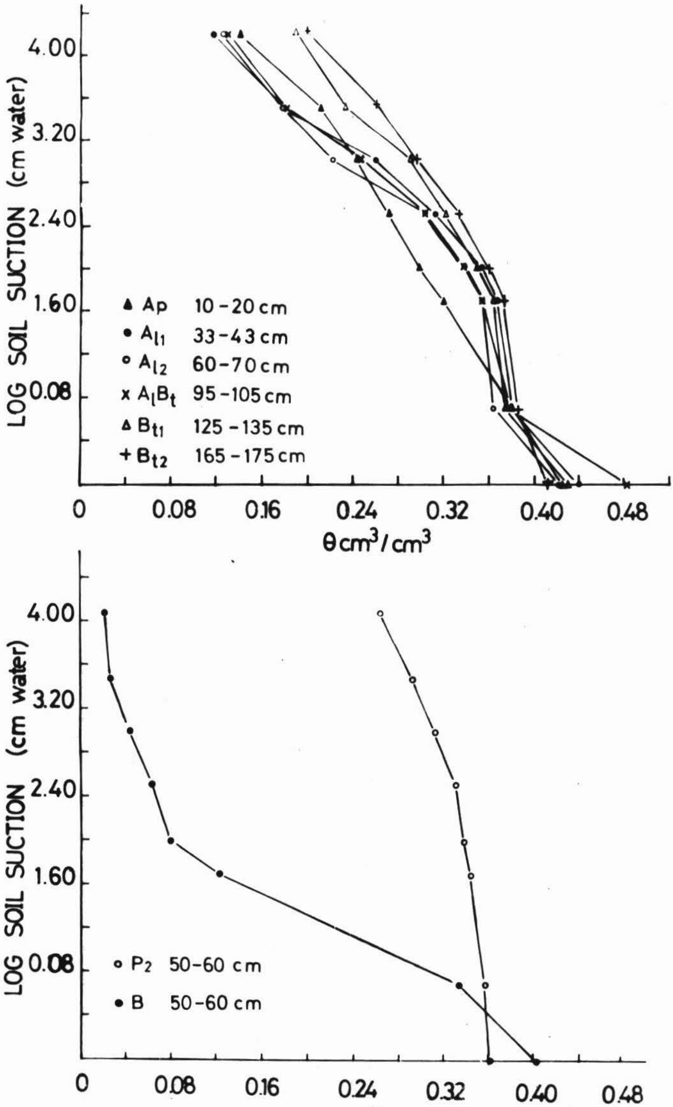


Fig. 3b and 3c: Soil-moisture characteristic of b (above) six depths of the grey brown podzolic soil and c (below) one depth of the gley podzol and one depth of the "Pelosol"

4.5. Use of artificial quartz powder as a contact material

The use of porous plates as end barrier in the determination of the hydraulic conductivity was not without problems. One of the problems which was reported by Klute (1972) is the separation of the soil from the end barrier. This is due to shrinkage and settling and possibly holes as the soil is drained during the course of measurement. In order to avoid this problem, the end barriers could be spring loaded (Elrick and Bowman, 1964) or loaded with a counterweight (Richards and Wilson, 1936). In this work, and as was reported earlier by Benecke et al. (1976), the use of artificial quartz powder as a contact material between the sample and the end plates and/or between the tensiometer and sample (in the case of sandy soils) has found practical acceptance. The moisture characteristic curve and the particle size distribution of the quartz powder was given in Fig. 3a and Tabel 1, respectively, its hydraulic conductivity will be given later in the results. It was found to be high enough as not to hinder the free movement of water during the flow process.

clay	Fine silt	Medium silt	Coarse silt	Fine sand	Medium sand	Coarse sand
< 2	2-6	6-20	20-60	60-200	200-600	600-2000
0.4	4.4	8.0	45.4	38.2	0.2	0.1

Table 1: Particle size distribution (μ) of artificial quartz powder expressed in percentage

4.6. Laboratory methods

4.6.1. Evaporation method

The evaporation method for determining the diffusivity and hydraulic conductivity of unsaturated soils developed by Arya et al. (1975) was used. The theory of this method was given in section 2. The equation given by Bruce and Klute (20) was used with the initial and boundary conditions (16).

Ten replicate samples of the 10 cm long cylinders were taken from each of the following depths: 10-20 cm, 33-43 cm, 60-70 cm, 95-105 cm, 125-135 cm and 165-175 cm of a grey brown podzolic soil by the sampling method described (section 4.3.1.). Seven replicate samples were also taken from the 50-60 cm depth of a gley podzol and similarly in the 50-60 cm depth of the "Pelosol". The samples were placed on a porous plate in an exicator and air was evacuated from them for 24 hours. They were saturated with water by capillary rise. Samples were protected against evaporation at both ends and allowed to equilibrate in a horizontal position for a few days. After this, the samples were opened at one end, put on a balance and hot air (> 100°C) was directed on the wet surface at a constant rate. The sample was weighed at intervals to determine the water loss and the cumulative evaporation was plotted against the square root of time. To satisfy the theory, the cumulative evaporation must be equal to the square root of time. The requirement cannot normally be met for the first few minutes. The time required to achieve the desired proportionality was strongly influenced by the initial wetness of the soil and by the potential evaporation rate. On the other hand, the actual evaporation rate depends on the conductivity of the soil. At the end of evaporation, which was usually in 16-25 minutes, the bottom seal of the sample was removed and the sample was pushed out and cut into 12-15 segments as quickly as possible. The water content distribution and bulk density were determined. Since the water content gradient was greatest towards the dry end of the soil core, thinner segments were cut from this end. The initial water content was calculated from the difference between the initial weight of the sample and weight of the dry sample. The second theoretical and test condition was considered satisfactory if the water content at the sealed end was equal to the initial water content of the sample. This means the original water content of the soil sample at the closed end remained unchanged during the evaporation procedure. The distance of the hot air outlet and the air temperature had to be varied until the time law (cumulative evaporation α $t^{1/2}$) was attained. The temperatures of the warm air and the distances from the air outlet to the evaporating surface selected for each depth of the different soil types are shown in Table 2.

After weighing and oven drying, the volume of each soil segment was calculated by dividing its dry weight by the average bulk density of the soil sample. Segment length was obtained by dividing the segment volume by the cross sectional area of the cylinder. The moisture content distribution of each sample was plotted. From this plot, the derivative and integral expressions in equation (20) were graphically evaluated, and the diffusivity as a function of water content was calculated for different levels of water content.

These diffusivities were used to calculate hydraulic conductivities using equation (6). Table 3 shows how the initial water content in the samples was calculated.

Soil type Horizon	Depth at which sample was taken (cm)	Distance from evaporating surface to air outlet (cm)	Temperature of warm air (°C)
Grey brown podzolic soil (Loess) (Parabraunerde)*			
Ap*	10-20	12	216
Al_1	33-43	—**	—**
Al_2	60-70	9	230
A_1B_t	95-105	—**	—**
Bt_1	125-135	16	185
Bt_2	165-175	16	185
Gley podzol (Gleypodzol)*			
B_h	50-60	6	241
Clayey soil (Pelosol)*			
P_2	50-60	28	106

* Soil type and classification according to the German system

** Experimental conditions could not be accomplished

Table 2: Some selected temperatures of warm air and distances from the evaporating surface to air outlet for three soil types using core samples of 500 ml volume and 8 cm I.D.

Replicate	Total wet soil g	Total dry soil g	Total water g	Volume of soil core cm^3	Initial volume of water cm^3/cm^3
1	949.20	754.03	195.17	501.72	0.389
2	900.80	694.41	206.39	498.14	0.414
3	962.70	771.40	191.30	494.32	0.387
4	942.30	752.85	189.45	499.65	0.379
5	970.00	767.36	202.64	500.35	0.405
6	939.00	732.81	206.19	497.64	0.414
7	950.20	746.29	206.91	496.19	0.417
8	961.30	753.30	208.00	501.20	0.415
9	987.50	799.79	187.71	500.15	0.375
10	978.70	789.56	189.14	498.39	0.380

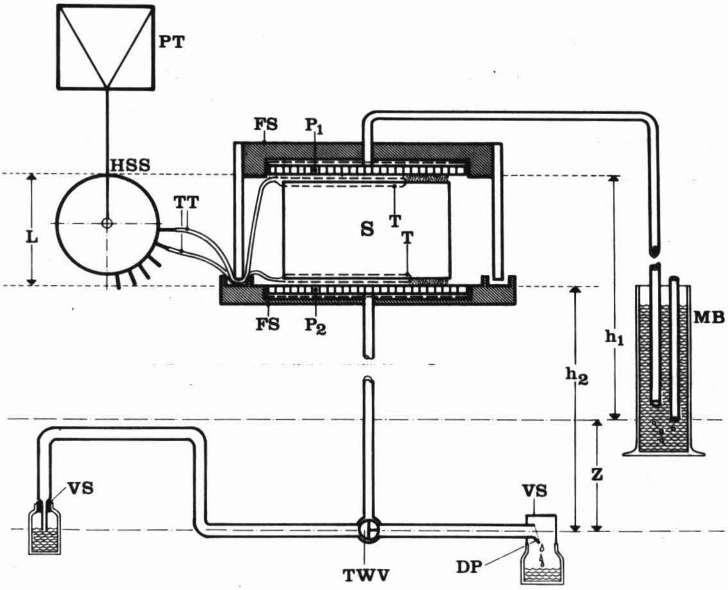
Table 3: Example of calculations for determining initial water content of the soil (Ap 10-20 cm depth) of the grey podzolic soil.

4.6.2. Outflow methods

4.6.2.1. Short column - small increment version

The method used in this section is based on a graphical analysis of the outflow data obtained by stepwise small increment of suction applied to the suction plate at the bottom of the soil sample. The graphical procedure is explained later in this section.

The experimental setup shown in Fig. 4 was designed by Benecke (1977) and was used to measure the diffusivity and hydraulic conductivity of short soil cores during both the steady-state and transient-state flow processes. Values of hydraulic conductivity could be measured over the matric potential range of 0 to 150 cm water column. This experimental setup resembles the double-plate or double-membrane apparatus given in the literature by Vetterlein (1964) and Henseler et al. (1969). The upper end cap (FS) with the attached Mariotte bottle (MB) as shown in Fig. 4 was used in the steady-state flow measurements only. Undisturbed soil cores 5 cm long and about 8 cm I.D. to give a volume of 250 ml taken from representative depths of a grey brown podzolic soil were evacuated for 24 hours and then saturated by capillary rise using tap evacuated water. The samples were placed with the lower ends on ceramic plates cemented to transparent acrylic glass end caps. The end caps had radial-running water channels (not shown) bored into them. Water could move freely from the sample through these channels between the end caps and the porous plate and eventually out through the drip point. Good contact between the samples and the porous plates was achieved using a thin paste of artificial quartz powder which was placed on the porous plate before the samples were placed in position. This quartz powder helped also to prevent separation of the soil samples from the plates in the case of shrinking soils. Two tensiometers 8 cm long and 0.7 cm in diameter were in-



PT = pressure transducer

HSS = hydraulic selector switch (Scanivalve)

TT = tensiometer tabing

L = sample length

FS = end cap with cover against evaporation

P_1, P_2 = pressure plate

S = soil

T = tensiometer

M = Mariotte bottle

VS = measuring cylinder

DP = drip point

TWV = three-way valve

Fig. 4: Schematic illustration of the apparatus used in determining the diffusivity and hydraulic conductivity of small soil core samples

stalled horizontally above and below the sample. This was to measure directly hydraulic head difference in the sample. Thus the uncertainties introduced by the variable head loss across the end plates (Kramer and Meyer, 1968) and the contact resistance between the plate, quartz powder and soil (Renger et al., 1972) could be avoided. The air entry value (bubbling pressure) of the tensiometer cups was given as 4000-5000 cm water and the conductivity was between 0.1 and 0.3 cm/day. The response time of the tensiometer (given as the time interval between the switching on of a particular tensiometer and the time the reading was taken) was about 1/2 to 1 minute. The two tensiometers (T) were connected to two different pressure transducers (PT) (sensitivity to the order of $\pm$ < 0.001 cm water) by means of a hydraulic selector switch (HSS) (Scanivalve). The output of the pressure transducers was registered with digital voltmeters (not shown) and the conversion factor was given as (0.1665 mV = 1 cm water). The volumetric outflow at different time intervals was collected from a drip point in two ways: 1) in a glass and then weighed, 2) when the outflow volume was very small, accurate measurement was done through a narrow glass pipe with the tip about 2 mm under water in a small glass container. The container with the water was weighed before and after each measurement. Suction levels of 10, 20, 40 and 70 cm of water starting at saturation were applied to the water below the porous plate by lowering the hanging water column attached to the lower plate. Drainage of the sample occurred. For each suction level, measurements of volumetric outflow and tensiometer readings were made at various time intervals. Hydraulic equilibrium was established for each suction level in order to obtain the total volume of outflow and to have defined conditions at the start of the next suction level. Short time intervals were chosen at the start of drainage and the interval was increased accordingly as drainage progressed.

The cumulative outflow versus time and the tensiometer readings versus time were plotted graphically. These curves were smoothed by eye-fitting. Darcy's equation (1) was used to calculate the unsaturated hydraulic conductivity. In the Darcy equation, the flux is given by

$$q = \frac{d(W/V)}{dt} \qquad (23)$$

where W is the volume water content and V is the volume of the soil sample. The flux was derived as the tangent of the cumulative outflow versus time curve for any particular time for which the hydraulic gradient was calculated, whereby the hydraulic gradient is given by

$$\frac{d\psi m}{dz} + 1 \;\hat{=}\; \overline{\Delta H/\Delta z} = \frac{\psi^{a}m(t) - \psi^{b}m(t)}{z^{a} - z^{b}} + 1 \qquad (24)$$

where $\overline{\Delta H/\Delta z}$ = the mean hydraulic gradient

ψm^{a} = matric potential above (tensiometer reading)

ψm^{b} = matric potential below (tensiometer reading) (cm water column)

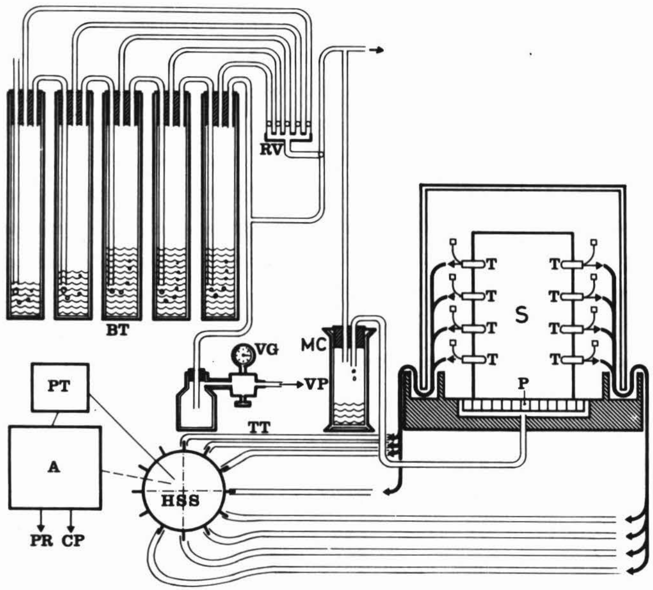
t = time (minute)

z^{a} and z^{b} = two positions where ψm^{a} and ψm^{b} were taken (cm).

The calculated conductivity is associated with the mean suction $\psi m^{a} + \psi m^{b} / 2$ at any particular time for which the flux was measured. The method of data analysis is explained further with aid of the results (section 5.2.2.).

4.6.2.2. Long column - small increment version

The hydraulic conductivity of larger soil columns (section 4.3.2.) and synthetic coated cubical soil monoliths (section 4.3.3.) were determined by the outflow method using the apparatus shown in Fig. 5. This apparatus is similar to that described for the short column version and therefore only variations would be mentioned here. Before the determinations were carried out using the cubical monoliths, the upper and lower surfaces were sawed off to expose the soil samples. These surfaces were cleaned and smoothened off. Holes for tensiometers were drilled in all samples on both sides at distances of 2, 10, 18, and 26 cm from the bottom corresponding to soil parameters such as horizons, transition boundaries etc. Similar tensiometer cups described for the short column version were installed horizontally in the holes. The series of tensiometers at intervals in the samples were to determine the head difference across each sample interval and thus obtain the hydraulic gradient for each section without making the assumption of a straight gradient in the samples. The porous plate was sealed to a water chamber on its underside. Adequate contact between the sample and the porous plate was maintained as in the short column version. The four pairs of tensiometers were connected to four pressure transducers by means of automatic selector switches. The output of the transducers, recorded directly in mm of water column, was read by a system Hartmann and Braun Digitron automatic read-out instrument and then recorded by both a paper tape printer and a data puncher instrument. The samples, while on the plates, were saturated for days by capillary action; tensiometer readings were used to control when the samples were saturated. Equilibrium was achieved when each tensiometer reading corresponded to the height of the tensiometer above the porous plate. Suctions of 20, 50, and 100 cm water were increasingly applied to the porous plate using a bubble tower (Rose, 1966, p. 196). Drainage of the samples was allowed to occur for each suction



S	= soil sample	MC	= measuring cylinder
T	= tensiometer	PT	= pressure transducer
P	= pressure plate	A	= digitron automatic read-out instrument
TT	= tensiometer tubing		
HSS	= hydraulic selector switch	PR	= paper tape printer
VG	= vacuum gauge	CP	= data puncher
VP	= vacuum pump	BT	= bubble tower
		RV	= regulating valve

Fig. 5: Schematic illustration of apparatus used in determining the hydraulic conductivity of large soil samples

level until equilibrium was attained. The volumetric outflow was measured in a graduated cylinder. All readings were taken at one minute intervals at first and then progressively increased to a maximum of 8 hourly intervals as drainage progressed. A total of 5 x 24 tensiometers could be read in one operation. Evaporation from the upper surfaces of the samples was prevented by enclosing each sample in a casing. Hydraulic conductivity was calculated for each sample interval using the Darcy equation. The method of calculation is given below. Each K value was associated with a mean suction obtained from the upper and lower tensiometers and at any particular time interval.

Curves of the tensiometer reading (at the four depths) versus time were plotted. The cumulative outflow was also plotted as a function of time. The moisture content-suction relationship was obtained from data using other samples. From the matric suction-time curves, six time intervals (Δt) were selected and the suction as a function of depth was obtained for the different times. The suction profiles as a function of time for suction levels of 20, 50 and 100 cm water are shown in Fig. 6 for the 120-150 cm depth of the grey brown podzolic soil. The curves were prolonged at both ends since there were no tensiometers intalled directly above and below the samples. The mean hydraulic gradient $\overline{H/\ z}$ for any time interval and for any sample interval was given by

$$\overline{H/\ z} = {}^{a}m(t_1) + {}^{a}m(t_2) \Big/ 2 + \ z$$

$$- \ {}^{b}m(t_1) + {}^{b}m(t_2) \Big/ 2 \qquad (24)$$

where t_1 and t_2 = the consecutive times for which outflow volume and tensiometer reading were taken.

In the Darcy equation, the flux q has been replaced by the expression $Q/A\Delta t$ (Richards and Weeks, 1953) where Q is the volume of water (cm^3) flowing through a unit cross-

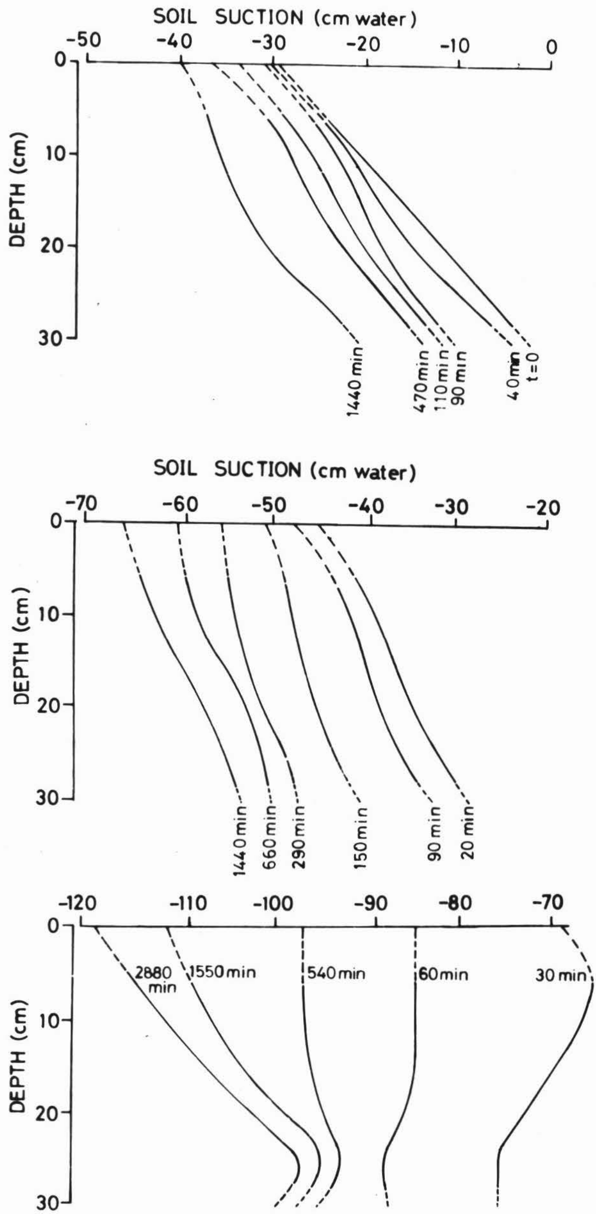


Fig. 6: Soil moisture profiles in the synthetic coated soil monolith (10-40 cm depth) during a desorption process under suctions of 20 cm (above), 50 cm (middle) and 100 cm (bottom) of water column starting from saturation.

sectional area A (cm^2) in the time interval Δt (minute). To obtain the quantity of water Q flowing through each sample interval use was made of the tensiometer readings and the average pF-curve for each sample interval. Using these tensiometer readings (suctions), the corresponding water content as a function of time and depth was obtained from the pF-curve. The quantity of water flowing through each sample interval was calculated by multiplying the change in water content $\Delta\theta$ in this interval by the equivalent volume of the cylinder interval. The sum of the Q^*s (ΣQ^*) for all the different intervals for any time interval Δt should be equal to the experimentally measured outflow volume (Q) for that particular time interval. For most cases, the experimentally measured outflow volume Q was less than the calculated volume (ΣQ^*). The reason for this could be that water loss by evaporation from the samples, the measuring cylinder as well as the outlet tubings could not be completely avoided. A correction factor given as $\Sigma Q/\Sigma Q^*$ was used to multiply all the Q^* values for the different intervals to obtain the actual Q values which were then used for the calculation of the $K_{(\psi)}$ values. The volume of the sample was determined at the end of the experiment using homogeneous quartz sand of known bulk density. The weight of sand used to fill the acrylic glass container or the polyester fibre coat was divided by its bulk density to give the sample volume. The volume of each compartiment of the sample was then obtained from the total volume of the sample. The cross sectional area of the sample was calculated by dividing the volume by the length of the sample.

4.6.2.3. Short column - large increment method (one-step method)

In this method, one large suction increment (by use of hanging water column) was applied to a soil core on a porous plate and the rate of outflow was measured. Thus this method differs in principle from the pressure outflow method described by Doering (1965) and based on analysis by Gardner. The theoretical fundamentals are however also based on the method by Doering of which the theory of this method was given (section 2). Equation (13) was used to calculated diffusivities and the conductivities were calculated using equation (6). Soil samples used for both the steady-state and outflow methods (short column) were saturated while still on the experimental setup (Fig. 4). One suction increment (in this case 150 cm water column) was applied to the soil sample by lowering the hanging water column 150 cm below the sample. The volumetric outflow was measured. Equilibrium water content was supposed to be attained when the outflow volume could no longer be measured. The final water content was determined by oven-drying of subsamples of the soil.

Before drainage started, the suction in the sample was measured with two tensiometers. This variation from the original method by Doering was introduced in order to be able to determine the initial water content in the sample from the water content-suction relationship.

The instantaneous outflow rates were determined graphically from a plot of the accumulated outflow versus time. This curve was first smoothened off by eye-fitting. The values of dW/dt (where dW/dt = outflow rate / sample volume) in equation (13) are obtained from the smooth curves whereby the outflow rate is given as the tangent of the curve at any particular time for which dW/dt was measure. From the accumulated outflow versus time curve both the values of W and W_f (equation 13) can be obtained. The W value is given by the value of the corresponding value on the ordinate axis. Accordingly the W_f can also be given as the total value on this axis. To obtain $(W - W_f)$, the cumulative outflow at that time and for which dW/dt is measured is substracted from the final or total outflow. The calculated initial water content was compared with that obtained using the pF-curve. Values of $(W - W_f)$ were obtained at the different times for which dW/dt was determined and equation (13) was then used to calculate the diffusivities. The water content-suction relationship was obtained from different samples since this methods involves the destruction of the samples and secondly it was necessary to use the slopes of the same curves for this method and the evaporation method to obtain the conductivity values.

4.6.3. Steady-state methods

4.6.3.1. Short column version

The experimental setup (Fig. 4) described and used for the outflow short column version was used for this method. The saturated soil sample was held between the two porous plates P_1 and P_2. A Mariotte bottle attachted to P_1 was used as a constant head water supply system. The hanging water column which also served as the drip point was connected to the porous plate P_2 (as in the outflow method). The flow of water was through all parts of the plates, that is, water moved freely through the plates. The Mariotte bottle was not calibrated and so the volumetric flux was controlled and measured at the outflow end only. One Mariotte bottle was used to supply five samples through a five-way glass system (not shown). The experimental principle could be summarized as follows: Water at a hydraulic head ($h_1 = \psi m^a$) was supplied from the Mariotte bottle to the top of the porous plate P_1. Flow occurred through this plate, the soil sample and the porous plate P_2. Water was maintained at a constant hydraulic head ($h_2 = \psi m^b$) in a space below plate P_2 by the allocation of the drip points. The difference between the hydraulic heads h_1 and h_2 was equal to the length (L) of the soil.

Undisturbed soil cores 5 cm long and 8 cm I.D. were taken from the grey podzolic soil, from the gley podzol and from "Pe-osol". Short sample lengths were generally preferred since the sample length influenced the time required to proceed from one steady state to another (Klute, 1965b). Suction levels of 10, 20, 40 and 70 cm of water were applied one after the other. In order to obtain the required matric suction in the sample and for flow to occur in the downward direction the inflow point (Mariotte bottle) and the outflow point (drip point) were lowered together and maintained below the sample; the sample position did not change. In this way the hydraulic heads h_1 and h_2 were increased to correspond to the actual matric potentials ψm^{a*} and ψm^{b*} in the sample below and above the ceramic plates. When the height difference between the Mariotte bottle and the drip point was equal to the length of the sample (z = L) as in Fig. 4, then the matric potentials in the sample were equal ($\psi m^{a*} = \psi m^{b*}$). But since it is the hydraulic potential and not the matric potential that is responsible for the flow and of course the direction of flow, water movement occurred downwards, that is, towards a region of lower gravitational potential. The water movement was in two stages: 1) drainage which arose from a desorption process because water originally under equilibrium at a lower suction level was again subjected to a higher suction level and 2) the steady-state flow which was then later attained. When the steady-state flow was attained, that is, the flux and the matric suction remained constant with time, measurements of Q (quantity of water measured in a time interval Δt) and tensiometer readings were taken. Darcy equation was then used to calculate the hydraulic conductivity.

The calculated hydraulic conductivity was associated with the mean matric potentials in the sample. In equation (1) and as has already been shown for the outflow methods (small increment versions), $d\psi m/dz$ was replaced by $(\psi m^a - \psi m^b) / \Delta z$, ψm by $(\psi m^a + \psi m^b)/2$ and now the suction heads ψm^a and ψm^b by ψm^{a*} and ψm^{b*}, respectively.

These experimental conditions were only achieved when the porous plates offered negligible resistance to the flow of water, that is, there was no contact resistance. By negligible plate resistance, the tensiometer readings where then equal to ψm^{a*} and ψm^{b*}, respectively, or deviate very narrowly from these. If the flow of water was hindered by other resistances outside the sample resistance, then the tensiometer readings were appreciably smaller than ψm^{a*} and ψm^{b*}. The distance z was increased by moving the Mariotte bottle and the drip point away from each other until the tensiometer readings corresponded again to ψm^{a*} and ψm^{b*}. Evaporation of water from the sample was further checked by enclosing the sample in a casing and by having a water pond in the inner side of the casing. Very minute volumes of outflow water were measured using an analytic balance with five decimal places in terms of grams.

4.6.3.2. Long column version

The apparatus shown in Fig. 5 and described in section 4.6.2.2. was used for the steady-state measurements on the same samples which were used in the outflow long column. The experimental principles were the same as those of the steady-state short column version described in section 4.6.2.1.

The Mariotte bottle was graduated and each sample had a separate Mariotte bottle. Thus at steady-state, the fluxes at the inflow and outflow ends were controlled and these were equal. The outflow and the steady-state methods were carried out as a continuous measurement at each suction level. This was to minimize the time required for the experiment. Any soil column at saturation or equilibrium water content would drain to a condition of steady-state downward flow. Thus the desorption process was allowed to occur first for each suction level with the Mariotte bottle closed until the equilibrium water content was attained. Mariotte bottle was then opened, the existing pressure head was first introduced into the Mariotte bottle before the supply to the soil sample was opened. Water at a hydraulic head h_1 was supplied to the top of the ceramic plate P_1 and outflow occurred under a constant hydraulic head h_2 in a space below plate P_2 by the allocation of the drip point. The hydraulic heads h_1 and h_2 were introduced from the bubble tower and so it was not necessary to have the Mariotte bottle and the drip point below the sample. In order that the water movement should occur in the region of lower gravitational potential, the Mariotte bottle was maintained at a height ($z = L$) above the drip point. The distance z was increased in both directions when the tensiometer readings deviated from the expected matric potential in the sample. When steady-state flow conditions were attained, and for any suction level, consecutive measurements of Q were made (usually at one hourly interval). Since the flux was equal at all points in the sample, the volume of water flowing through each sample compartment during each time interval was equal to the total outflow volume from the sample for the same time.

4.7. Field method

The experiment was conducted from April to October 1977 on the grey brown podzolic soil on an untilled fallow plot. The plot was about 5 meter away from the profile in which samples for the laboratory measurements were taken. The method used here was based on the method by Richards et al. (1956). The general principle of this method is summarized and given by Klute (1972). Field measurements consisted of soil water content determinations and tension readings.

4.7.1. Measurement of soil water content

Soil water content measurements were obtained by two methods namely the neutron method and gravimetric moisture determinations.

4.7.1.1. Neutron method

The procedures of this method have been described in detail by van Bavel and Stirk (1967), Long and French (1967), Richter (1974), Babalola (1978) and many other authors, and therefore would be only briefly outlined here.

Three thin-walled steel pipes 3.6 mm I.D. and 2.4. meter long used as access tubes were installed vertically at about 1 meter distance from the tensiometer installations. Measurements of soil water content were made from 10 to 200 cm in 10 cm increments using 30 seconds counts at each depth. Measurements were made twice weekly and at least one neutron measurement was made on the same day when the gravimetric moisture measurements were done.

4.7.1.2. Gravimetric moisture determinations

Gravimetric soil moisture samples were taken from three locations at distances that would not interfere with the sphere of influence of the neutron probe or the tensiometers, usually within a radius of 30-100 cm. Soil samples were collected once a week, at 10 cm segments (0-180 cm depth) using an auger. Three replicates were collected (corresponding to the three neutron probe/tensiometer sites) and these were weighed, and oven-dried at 105°C to constant weight. Soil moisture content for the different depths was calculated on a volumetric basis using the soil bulk density for these depths.

4.7.2. Measurement of hydraulic head

Three tensiometers were installed at each of the following depths: 10, 20, 30, 40, 50, 60, 80, 100, 120, 140 and 160 cm. The tensiometers were read daily and were purged with deaerated water air bubbles were noticed.

4.7.3. Method of data analysis

The hydraulic conductivity of the different soil layers were calculated using the "water divide" technique of which the theory is given in section 2. A sample calculation as shown by Ehlers and van der Ploeg (1976) is presented in Table 4, in which the hydraulic conductivity data are calculated at July 13, using suction and water content measurements of 13th and 15th July 1977. The hydraulic conductivity for the depths above and below the "water divide" were calculated using equations (22a) and 22b), respectively.

Table 4: Calculation of unsaturated hydraulic conductivity at July 13, 1977

	July 13 (t_1)			July 15 (t_2)			Calculations for July 13		
Z (cm)	ψ (cm)	θ (cm^3/cm^3)	$\Delta H/\Delta Z$ (cm/cm)	ψ (cm)	θ (cm^3/cm^3)	$\Delta H/\Delta Z$ (cm/cm)	$\frac{\frac{\Delta H}{\Delta z}_{t1} + \frac{\Delta H}{\Delta Z}_{t2}}{2}$ (cm/cm)	$\Delta\theta/\Delta t$ ($cm^3/cm^3/day$)	K (cm/day)
0									
10	- 504			- 543					
		0.232	+ 15.60		0.220	+ 20.30	+ 17.95	- 0.006	6.69×10^{-4}
20	- 338			- 330					
		0.243	+ 3.50		0.239	+ 1.80	+ 2.65	- 0.002	2.26×10^{-3}
30	- 293			- 302					
		0.255	+ 1.40		0.254	+ 1.20	+ 1.30	- 0.0005	3.08×10^{-3}
40	- 269			- 280					
		0.254	+ 1.00		0.252	+ 0.90	+ 0.95	- 0.001	3.68×10^{-3}
50	- 249			- 261					
		0.259	+ 0.70		0.257	+ 0.60	+ 0.65	- 0.001	3.85×10^{-3}
60	- 232			- 245					
		0.266			0.265			- 0.0005	
70			+ 0.30			+ 0.35	+ 0.33		6.06×10^{-3}
		0.269			0.269			0.000	
80	- 205			- 218					
		0.267			0.266			- 0.0005	
90			+ 0.10			+ 0.20	+ 0.15		6.67×10^{-3}
		0.276			0.275			- 0.0005	
100	- 183	xxxxxxxxxx	xxxxxxxxxx	- 194			xxxxxxxxxxxxxxxxx		
		0.282			0.278			- 0.0020	
110			- 3.25			- 2.55	- 2.90		1.55×10^{-3}
		0.293			0.288			- 0.0025	
120	- 228			- 225					
		0.307			0.302			- 0.0025	
130			- 7.80			- 8.40	- 8.10		9.88×10^{-4}
		0.303			0.302			- 0.0005	
140	- 364			- 373					
		0.282			0.284			+ 0.0005	1.59×10^{-3}
150			- 4.70			- 5.35	- 5.03		
		0.272			0.272			0.000	
160	- 438			- 460					

xxxxx position of "water-divide"

5. RESULTS AND DISCUSSION

5.1. General concept

Though it is the main objective of this study to compare different methods and approaches for determining the conductivity functions of unsaturated soil layers, it seems advisable firstly to discuss in some detail the results of each of the methods. This is to allow for a comprehensive comparison of the methods later and in particular to examine which one would fit specified circumstances and tests. In discussing the results, the conductivity and diffusivity functions have been given using different scales for the different soils as well as the different soil layers of the same soil. Results were not obtained from all the methods and in all depths. This was due to two main reasons: 1) The evaporation method (as will be seen later) did not yield results in all the depths, and 2) the 250 ml samples were used for three different determinations (namely steady-state, one-step and outflow methods) and perhaps some of these samples were damaged and consequently did not yield results in the later determinations.

5.2. Results of the different laboratory methods

5.2.1. Evaporation method

Fig. 7 shows that required linearity between cumulative evaporation and the square root of time was reached between three and six minutes. Arya et al. (1975) and Ehlers (1976) reached linearity after 2 minutes and between 2.5 to 4 minutes respectively. According to Ehlers, the time differences might be explained partially by the geometry of the core. Arya used 80 ml cores with an inner diameter of 3.65 cm and an air temperature of about 90^{o}-$100^{o}C$. On the other hand, Ehlers used 200 ml

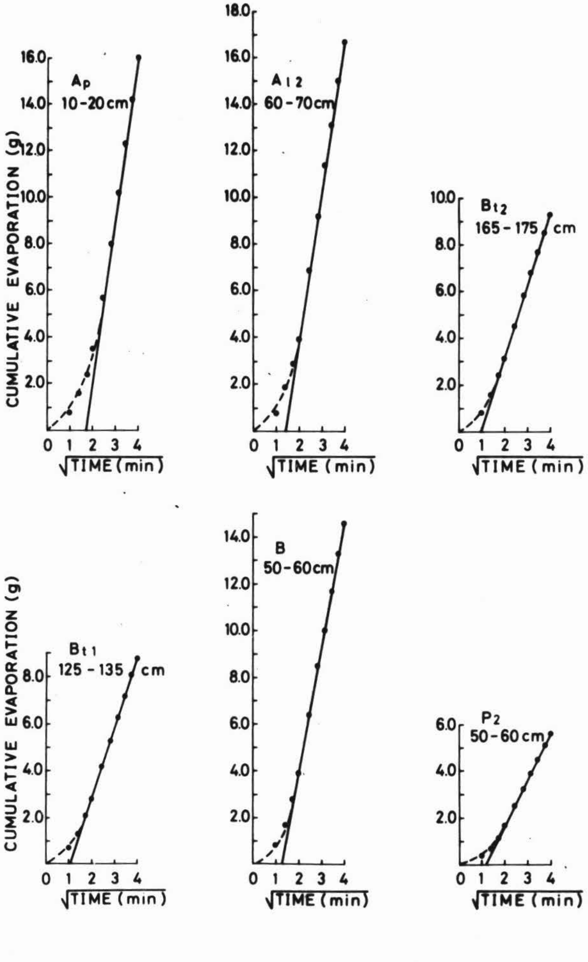


Fig. 7: Linearity of evaporation to square root of time for the different soil depths

cores with an inner diameter of 5.4 cm and the air temperature was 130^{o}C. In this work, 500 ml cores with an inner diameter of 8 cm were used. The much larger surface area of the cores used here will require an increased temperature necessary to overcome the initial phase of low evaporation rates. Table 2 shown earlier (Materials and Methods) shows that higher air temperatures were used as compared with the temperatures used by Arya and Ehlers. In this table, it can also be seen that the air temperatures differed from soil to soil, as well as from horizon to horizon. Another factor which may influence the length of time required to reach linearity may be the hydraulic properties of the soils. Generally, the time required to achieve the required proportionality may be influenced by one or a combination of some factors such as the size of the core, the initial wetness of the soil, the energy available, the hydraulic properties of the soil and sometimes, when not controlled, the external evaporative conditions.

After 16-25 minutes of evaporation, the cores no longer behaved as if they were semi-infinite. For the cores to behave this way, the water content at the sealed end of the cores must remain the same. Fig. 8 shows the water content distributions in the samples from the different layers of the soils with evaporation proportional to the square root of time. It could be seen from these curves that the water content of the samples at the sealed end did not deviate significantly from the initial water content though the small deviations as will be explained later, lead to important consequences. The steep gradients of the water content with distance near the soil surface are the result of a very rapid drying of the soil and suggest that the soil rather than the external conditions controlled the flow of water. The water loss was mostly from the first few centimeters of the soil. In the Ap 10-20 cm of the grey brown podzolic soil and the B 50-60 cm of the gley podzol, the water content distribution curves are not very steep. In these two horizons, a larger samples length of 4-5 cm contributed to the water loss.

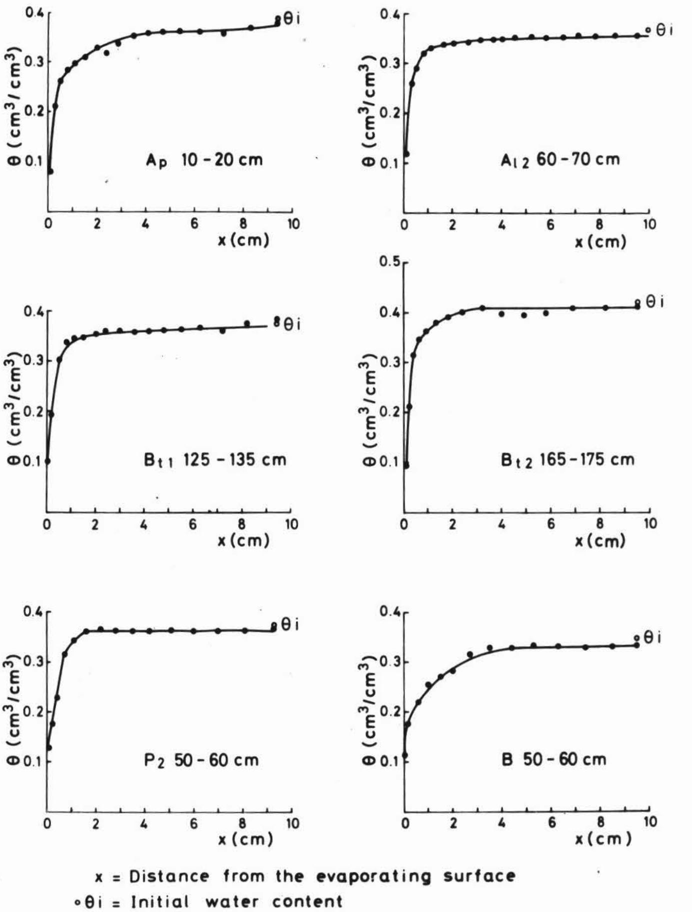


Fig. 8: Water content distribution in the soil cores with evaporation proportional to the square root of time.

In the Al_2 60-70 cm, Bt_1 125-135 cm, Bt_2 165-175 cm (grey brown podzolic soil) and the P_2 50-60 cm of the "Pelosol", water loss is entirely restricted to first two centimeters of the soil. This may perhaps be attributed to the conductivity of the soils.

The diffusivity as a function of moisture content of the different soil layers is shown in Fig. 9. The arithmetic means and the confidence intervals (95 % confidence level) of the means are also given. The narrow confidence intervals (except in the lower moisture content of the Bt_2 165-175 cm depth) show that the diffusivities for the replicate samples do not deviate so much from each other. In most cases the diffusivity curves have a gentle slope. This shows perhaps that the form and slope of the moisture characteristic curve of the soils determine to a larger extent the form and steepness of the conductivity function (see Fig. 10).

Diffusivity functions were not obtained for the following depths: Al_1 33-43 cm and A_1B_t 95-105 cm of the grey brown podzolic soil. The two conditions (linearity and constant θ at lower end) could not be met for these depths. Secondly, the water content versus distance curve did not give smooth curves. These observations have been reported earlier by Ehlers for the 30-40 cm depth of an untilled plot of the same soil. He thought that the high hydraulic conductivity of the soil might be responsible for this. The applicability of the method on quartz powder was not possible. In this case also, the water content distribution did not yield smoot curves. This may b- related to the pore-size distribution (Fig. 3a) of this sample.

The hydraulic conductivity functions of the soil layers are shown in Fig. 10. The conductivity function obtained for the 10-20 cm depth of the grey podzolic soil agree with results published by Ehlers (1976) for the same soil and depth. As can be seen from Fig. 10, the hydraulic conductivity data at the lower suction range (< 70 cm water) were obtained only

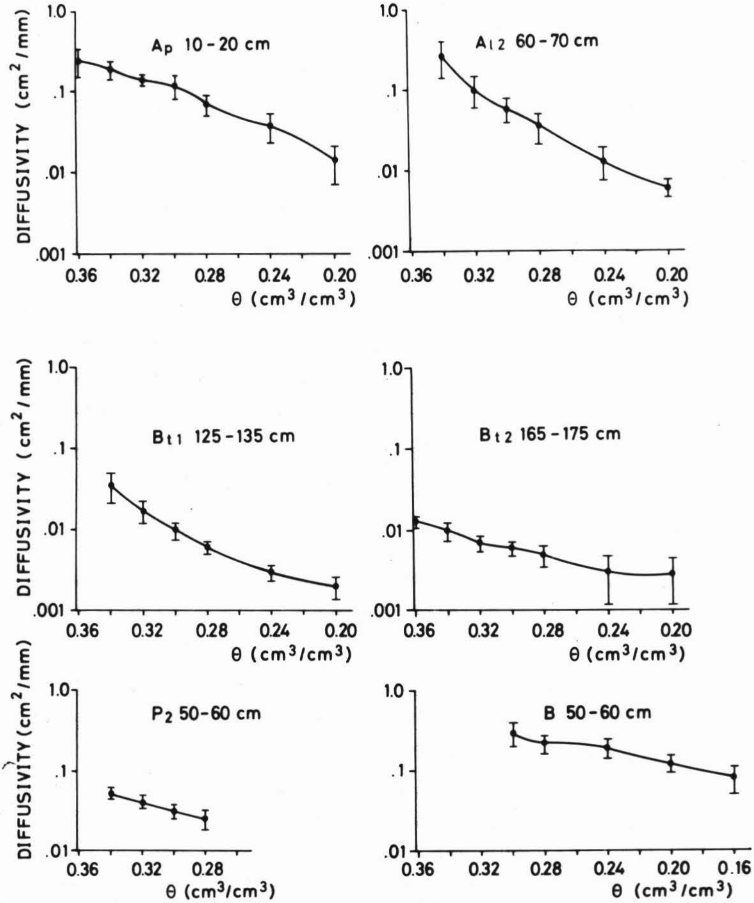


Fig. 9: Diffusivities as a function of the volumetric water content in four layers of the grey brown podzolic soil, one layer of the gley podzol and one layer of the "Pelosol".

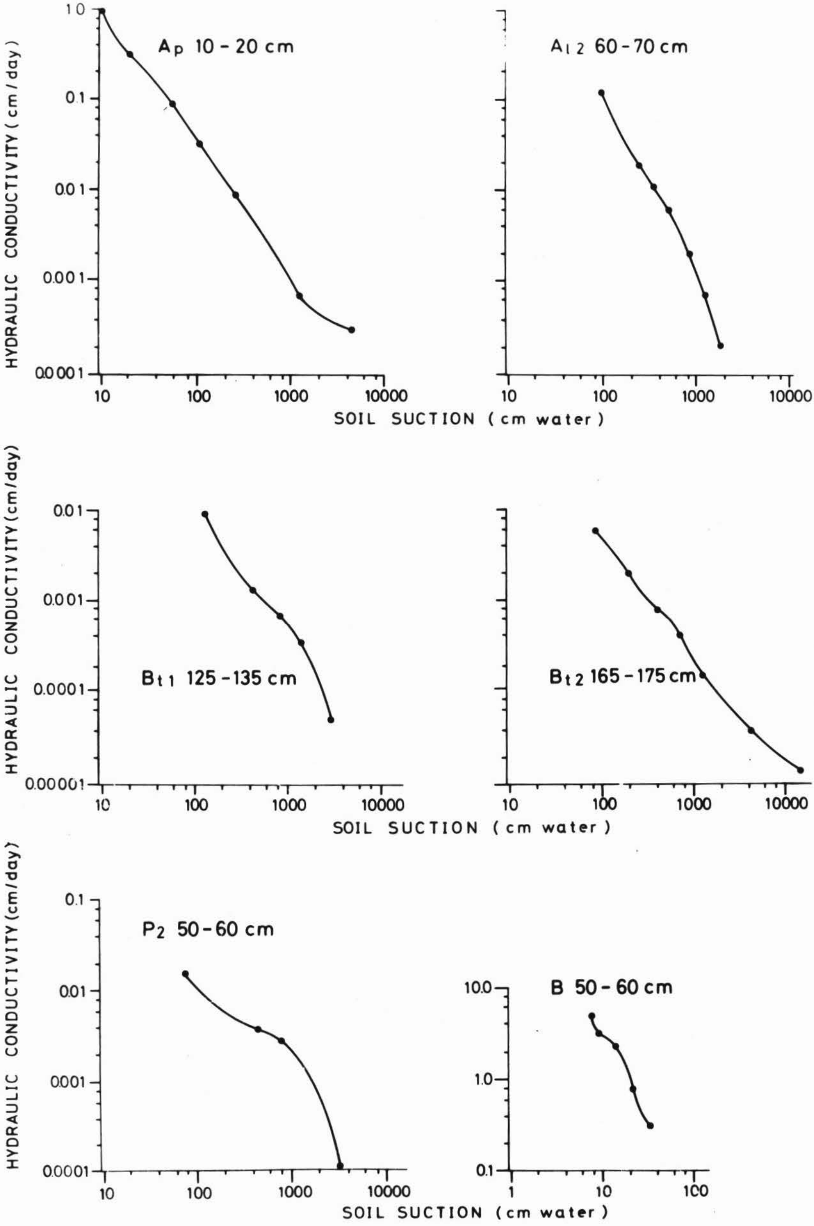


Fig. 10: Hydraulic conductivity as a function of soil suction obtained by the evaporation method (same soils as in Fig. 9).

in the Ap 10-20 cm and by the B 50-60 cm depth of the gley podzol. The inability to obtain values at lower suction for the other depths may be explained as follows: The initial water content θ (measured) is generally about 1 % by vōlume lower than the calculated θ_i. The suction equivalent to θ was obtained form the moisture characteristic curve in the suction range between pF 0.7 and 1.7 (Fig. 3b and 3c). In this suction range, the shape of the curve is so steep that a small difference in water content gives a very large difference in suction. Therefore, the difference in water content of the order of 1 % by volume means a big difference in suction. Thus in view of the small error of the data, the suction values for which $K_{(\psi)}$ values were obtained may be much lower than the suction at the saturation water content. This seems to be the case in the Al_2 60-80 cm, Bt_1 120-140 cm and Bt_2 160-180 cm depths of the grey brown podzolic soils as well as the P_2 50-60 cm depth of the "Pelosol". In the Ap 10-20 cm and the B 50-60 cm depths, the shape of the pF-curves are flat and a small decrease in water content means a smaller increase in suction. In these curves, a far better pairing of water content and suction values could be made. It could be concluded that the evaporation method yielded more reliable results for soils whose moisture characteristics are not too steep.

The attempt to calculate the variance of the $K_{(\psi)}$ values by the evaporating method was not practicable within the scope of this work. As shown in equation (6), which was used to calculate the $K_{(\theta)}$ values by the evaporating method, and the outflow method (short column - large increment version), the variance of the hydraulic conductivity $K_{(\theta)}$ is a product of the variance of the diffusivity $D_{(\theta)}$ and that of the slope of the water content-suction relationship, given as $(d\theta/d\psi)_\theta$. The error in $K_{(\theta)}$ is described statistically as an error propagation from $D_{(\theta)}$ and $(d\theta/d\psi)_\theta$. The mathematical solution of the equation used to calculate $K_{(\theta)}$ requires the knowledge of the variance in $D_{(\theta)}$ and $(d\theta/d\psi)_\theta$ as well as the covariance $(D, d\theta/d\psi)$. The covariance $(D, d\theta/d\psi)_\theta$ shows the dependence between $D_{(\theta)}$ and $(d\theta/d\psi)_\theta$. Since the diffusivity and the soil moisture characteristics were determined from two separate sample populations, this dependence between D and $d\theta/d\psi$ can not be evaluated. Samples were taken randomly in the field and two different sample populations were used for the two determinations, precluding the possibility to put the samples into pairs in a meaningful way.

5.2.2. Outflow method: Short column - small increment version

Fig. 11a shows the suction readings versus time curves for 40 and 70 cm suction levels applied to the sample. By the 40 cm suction level, the lower tensiometer reached the end reading in about 30 minutes after the suction was applied. The time taken by the lower tensiometer to reach the end reading was shorter (about 20 minutes) by the 70 cm suction level. The upper tensiometer at the 70 cm suction level on the other hand, took longer time to approach the end readings than at the 40 cm suction level. The point of interception of the upper and lower tensiometers indicates that the matric suction at these two points were equal ($\psi m^a = \psi m^b$). This again indicates that at this point the flow of water was due only to the difference in the gravitational potentials between the two levels of the soil sample where the matric suctions (tensiometer readings) were measured. Consequently, the hydraulic conductivity is equal to the magnitude of the flux at this point in the flow system.

Fig. 11b shows the hydraulic gradient as a function of time. The gradient decreased with increased time and approached zero as the drainage reached an equilibrium. Higher gradients were measured by the 70 cm suction level. The decrease of

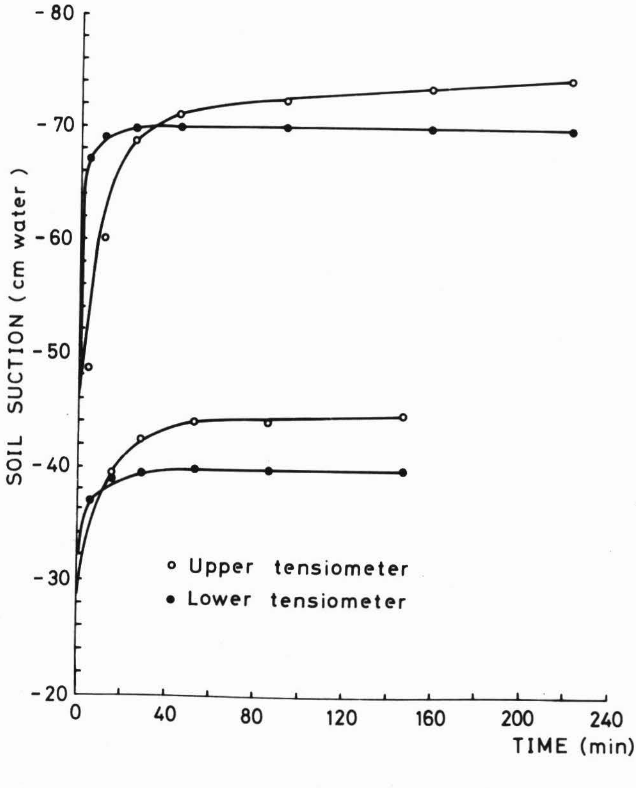


Fig. 11a: Suction reading as a function of time for the grey brown podzolic soil at 125-135 cm depth under suctions of 40 and 70 cm water column applied to the bottom of the soil sample.

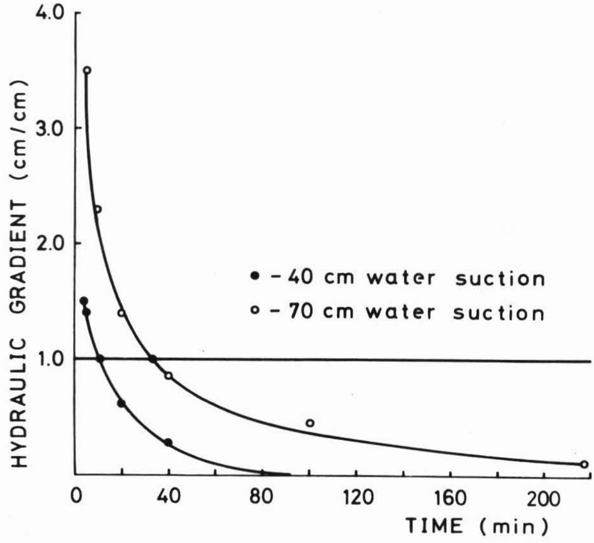


Fig. 11b: Hydraulic gradient as a function of time (see Fig. 11a).

hydraulic gradient with time was faster by the 40 cm water suction. For the calculation of conductivity values, hydraulic gradients equal to or near the value one were used. Fig. 11b illustrates this and as shown in this figure, the gradient values at both ends of the suction-time curves were respectively high and low and so conductivity values thus calculated from these gradient values were less reliable. This could be supported by the fact that the measured flux was very small. For example, at the 70 cm suction level only 2 cm^3 on the average came out from the sample in the entire time. Thus the rate of flow was very small even at the first stages of drainage. Since the hydraulic conductivity is a quotient of the flux and the hydraulic gradient, extreme values of the latter may give unreliable conductivity values. The flux was given by the tangent of the accumulated outflow versus time curve and for any particular time for which the hydraulic gradient was calculated.

Calculated hydraulic conductivity as a function of matric suction is shown in Fig. 12 for four depths of the grey podzolic soil. The points shown in the curves are the conductivity values from 5 to 10 replicate samples for each depth. In most cases, conductivities were calculated only at one point (usually by hydraulic gradient equal to one). In most cases, this point corresponded to the same suction for replicate samples. From the curves, it can be seen that for each suction level, the suction was the same and only the conductivity values varied.

From the correlation coefficients it can be seen that the scattering of data was very small.

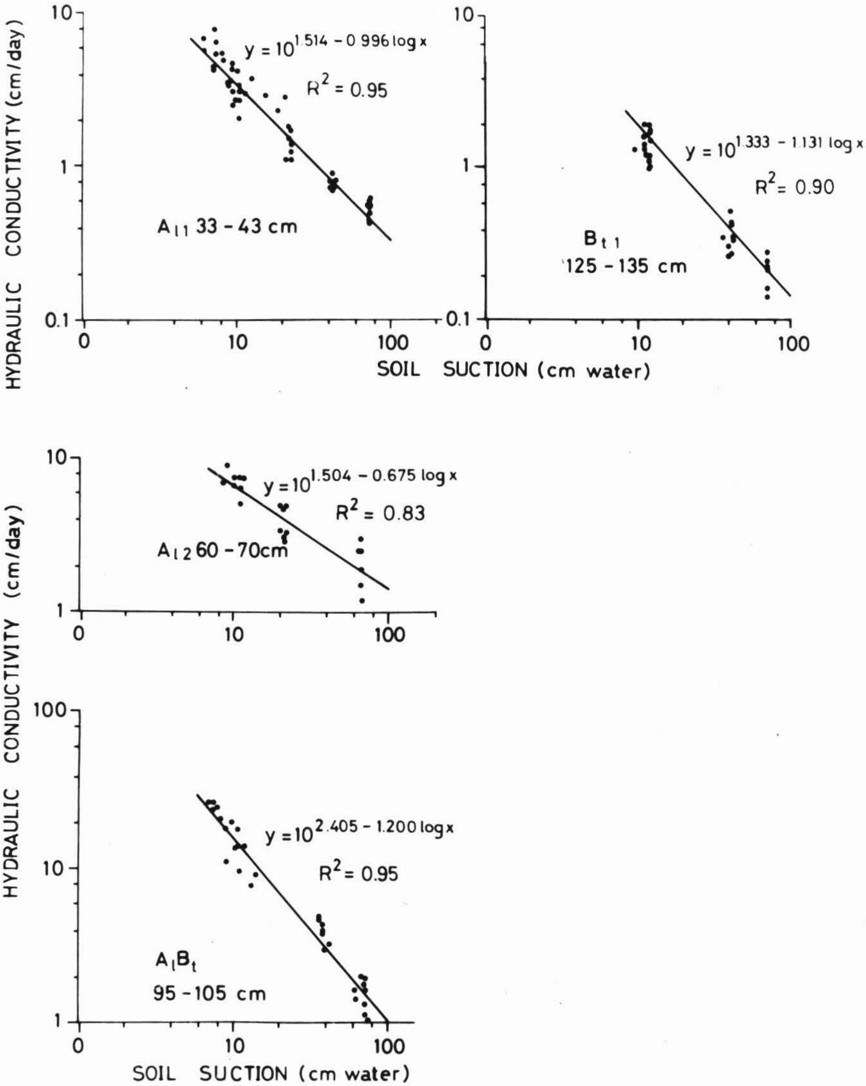


Fig. 12: Hydraulic conductivity-soil suction relationship obtained by the outflow method (short column version).

5.2.3. Outflow method: Long column - small increment version

In this section, the results obtained from both the samples in acrylic glass and the synthetic coated samples will be compared. Since both samples were handled experimentally the same way, it is the main objective to find out whether or not one can utilize one or the other of these sampling techniques. This question is of considerable practical interest because it requires more labour and time to prepare the polyester coated samples. This sampling technique could then be restricted to stony soils if results obtained using both sampling techniques do not show significant differences. Fig. 13a shows the hydraulic conductivity data of two replicate acrylic glass samples from each of the depths 60-80 cm. 95-115 cm, 130-150 cm and 160-180 cm. Fig. 13b shows the comparison for the synthetic coated samples for the depths 95-115 cm, 130-150 cm and 160-180 cm. To test if there was any significant difference between the results from replicate samples, the homogeneity of regression was used. This test says, if there is any difference between two population regression coefficients, that is, if the conductivity data from both sampling techniques could be represented by a single regression line. At 5 % confidence level, there was no significant difference between replicate samples taken with acrylic glass. For the synthetic coated samples, significant difference was found at one depth, 130-150 cm only. From this test, it could be concluded that the long column - small increment version could give reproducible results using either acrylic glass sample or synthetic coated soil samples. Results obtained using acrylic glass samples were further compared with those from synthetic coated samples. Fig. 14 shows the hydraulic conductivity functions obtained from both synthetic coated and acrylic glass samples for 95-115 cm, 130-150 cm and 160-180 cm depths. Significant difference was found between the results of both techniques at depths of 130-150 cm and 160-180 cm but there was none at the 95-115 cm depth. From Fig. 14, it can be seen that the

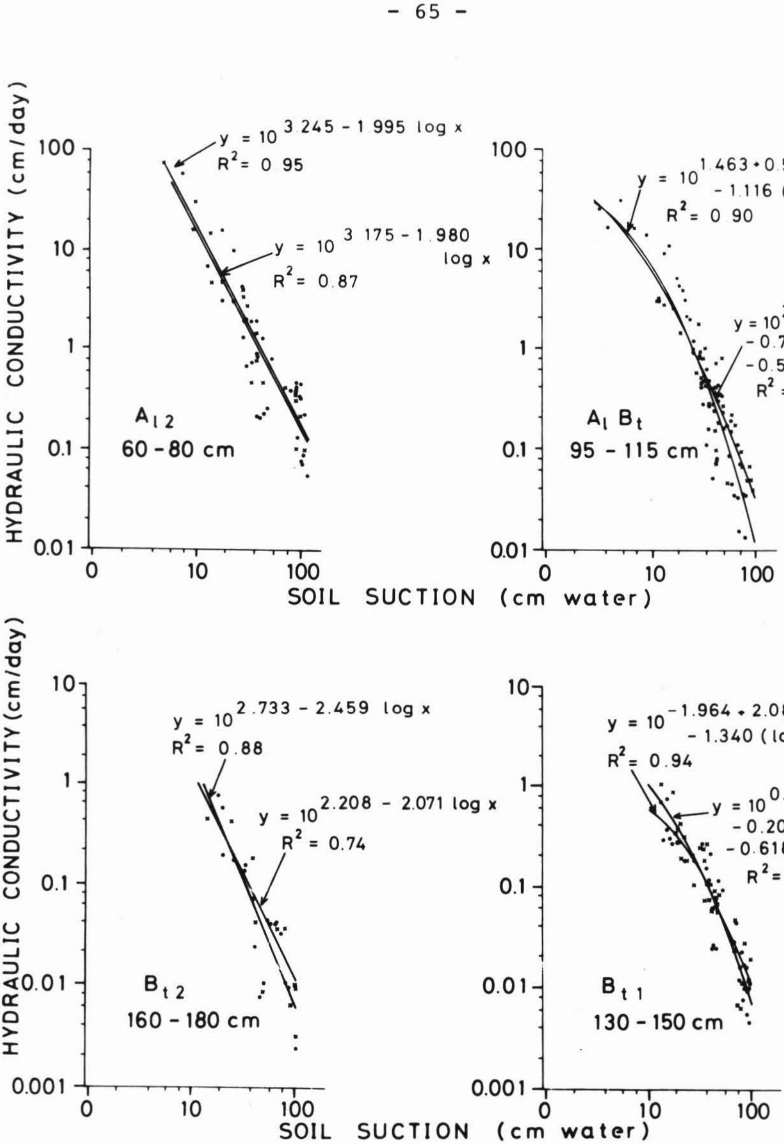


Fig. 13a: Hydraulic conductivity-soil suction relationship of two replicate samples in acrylic glass for the 95-115 cm, 130-150 cm and 160-180 cm depths.

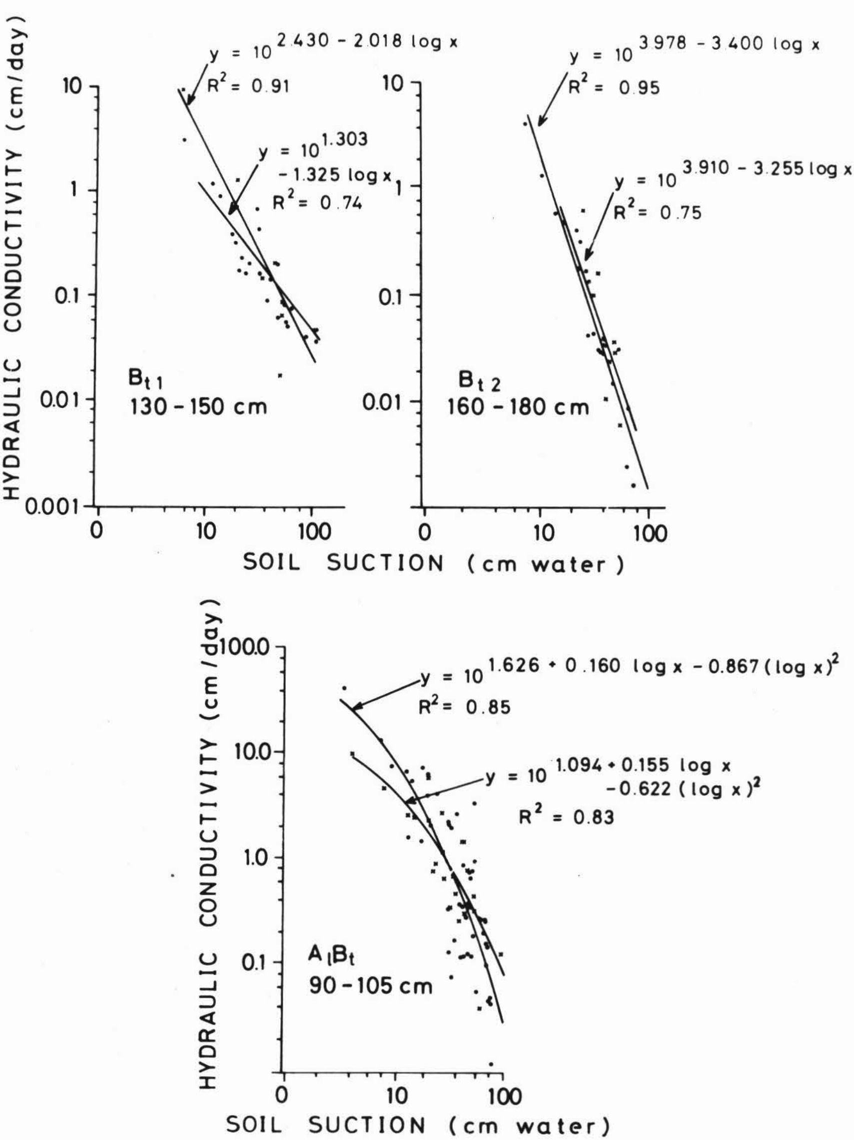


Fig. 13b: Hydraulic conductivity data obtained from two replicates using the synthetic coated samples for the 95-115 cm, 130-150 cm and 160-180 cm depths.

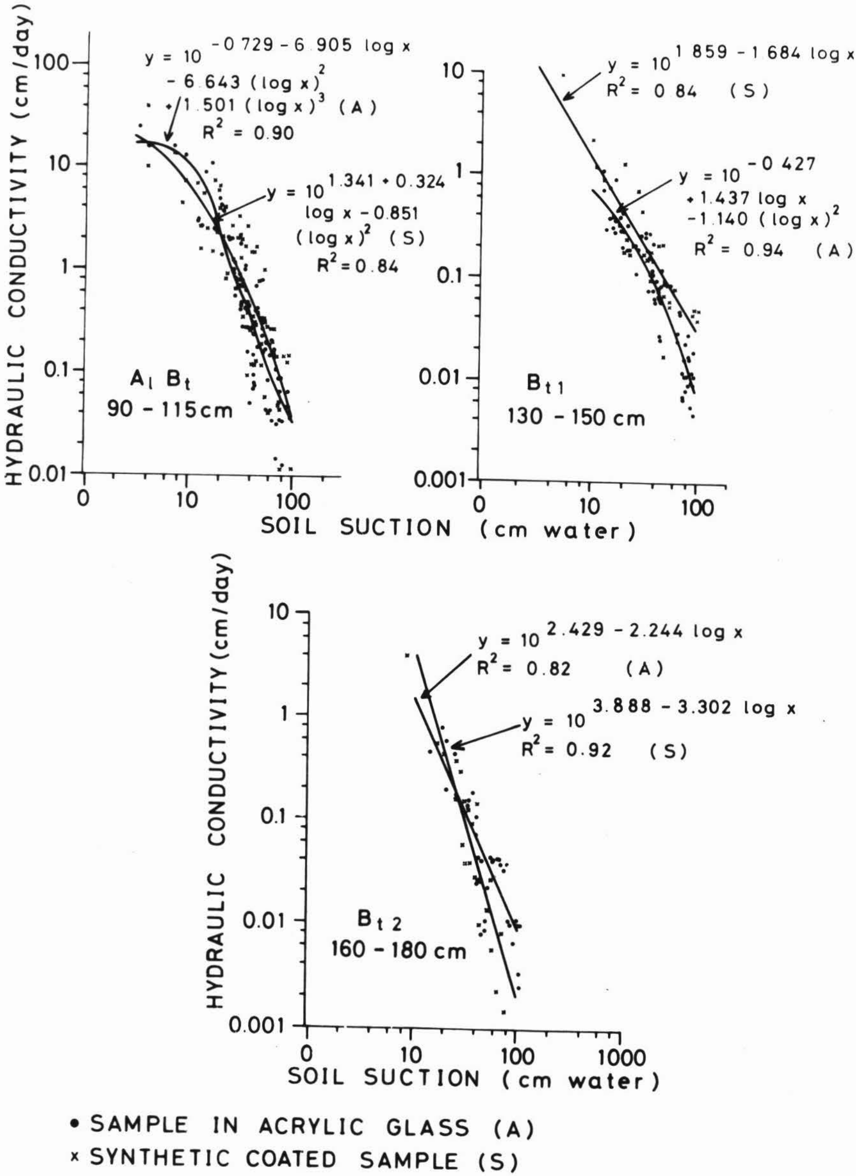


Fig. 14: Hydraulic conductivity data obtained from samples in acrylic glass and synthetic coated samples in similar depths.

synthetic coated samples yield higher K values in the lower suction range in the 95-115 cm and 130-150 cm depths. This result excludes the fact that an imperfect contact between cylinder wall and soil existed by the samples in acrylic glass, since this situation might have resulted in higher conductivity values by the acrylic glass samples at the lower suction range. It seems however that by the sliding action of the cylinder along soil column, a good contact wall might have resulted. The vibrations in the sampling technique using the acrylic glass columns might probably accounted for the higher conductivity values with increased suction by these samples in the 160-180 cm depth. On the whole, it can be seen that the results obtained using both sampling techniques are close.

5.2.4. Outflow method: Short column - large increment version (one-step method)

The diffusivity as a function of water content and the hydraulic conductivity as a function of soil suction are shown for four depths in Figs. 15 and 16, respectively. The arithmetic means and the confidence intervals of the means are also given for the diffusivity-water content relationship. Results were not obtained for the 10-20 cm and 165-175 cm depths because only two samples from each of these depths gave measurable outflow volumes. The other samples might have been damaged during previous experiments in which they were used. The confidence intervals are comparatively wider for the 33-43 cm depth. Although one suction increment (150 cm water) was applied to all the samples to cause drainage, the ranges of suction for which the samples yielded results varied from horizon to horizon. Fig. 16 shows that conductivity values for suction ranges up to and above 100 cm water were calculated for the 60-70 cm depth only. This could be related to the steepness of the moisture characteristic curves in this suction range. So it could be

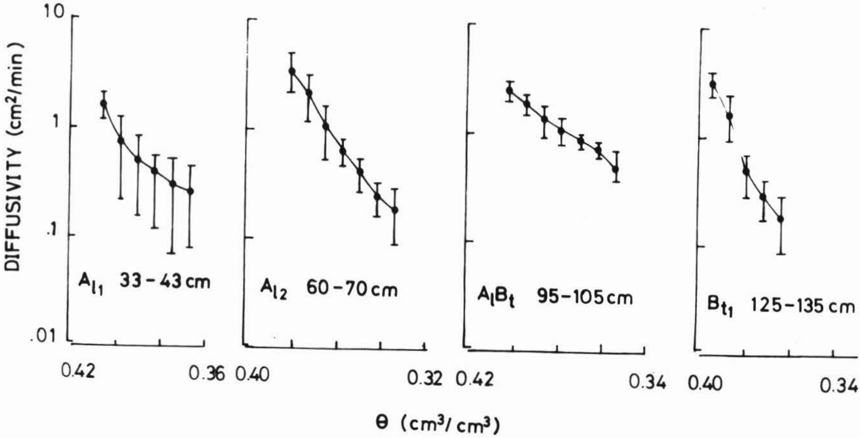


Fig. 15: Diffusivities as a function of water content obtained by the one-step method.

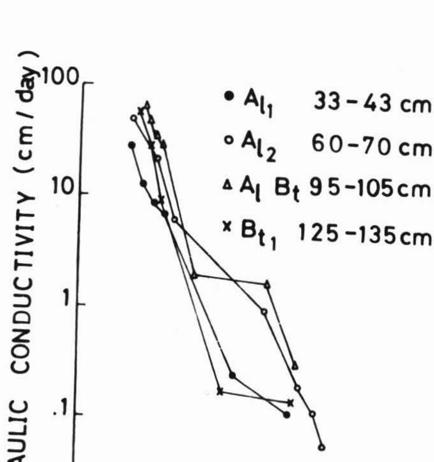


Fig. 16:
Hydraulic conductivity-soil suction relationship obtained by the one-step method.

suggested that a much larger suction increment (200 to 500 cm water) should be applied in order to obtain conductivity values for those suction ranges which are important in the soil water-plant relation. In Fig. 16, it can also be seen that the hydraulic conductivity values measured at the different depths did not deviate appreciably from one another between suctions of 2 cm to 10 cm of water column. Since the diffusivity-water content relations for the different depths yielded smooth curves (Fig. 15), it can also be seen that the form of the hydraulic conductivity function (Fig. 16) obtained are related to the steepness of the moisture characteristic curves.

As is shown by the diffusivity and conductivity function, a small decrease in water content corresponded to a larger increase in suction between 10 cm and 100 cm water suction and for the 33-43 cm, 95-105 cm and 125-135 cm depths.

5.2.5. Steady-state method: Short column version

The variability in results which might be due to this method was tested by determining first the hydraulic conductivity of the quartz powder. Table 5 shows the hydraulic conductivities for the three quartz powder samples. Shown also in Table 5 are arithmetic mean values and the standard deviations at each suction level. The standard deviations are very small, so the hydraulic conductivity values from the three samples are consistent at all suction levels. It can therefore be concluded that the variability in result which may be due to this method itself is very small. One might expect that hydraulic conductivity values from the same method on replicate soil samples from one soil horizon (with little spatial variability) might not deviate appreciably from one another. Fig. 17 shows the graphic representation of the mean conductivity of the three samples of the artificial quartz powder. The conductivity function as is given

Samples	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150
	Hydraulic conductivity (cm/day)														
	Soil suction (cm water)														
1	6.42	6.55	6.90	6.30	6.20	5.89	7.02	5.85	6.17	6.21	5.82	6.42	6.83	6.66	5.95
2	6.45	6.40	6.88	6.32	6.21	6.15	7.00	5.95	6.03	6.13	5.81	6.46	6.79	6.62	5.83
3	6.50	6.46	6.81	6.13	6.30	5.98	6.80	6.03	6.11	6.30	5.90	6.32	6.81	6.65	6.02
Arithematic mean															
$\overline{x}$	6.46	6.47	6.86	6.25	6.23	6.01	6.94	5.94	6.10	6.21	5.84	6.40	6.81	6.64	5.94
Standard deviation															
S_D	0.033	0.062	0.039	0.082	0.057	0.110	0.099	0.074	0.057	0.070	0.040	0.059	0.016	0.017	0.090

Table 5: Hydraulic conductivity data of three replicate samples of industrial quartz powder

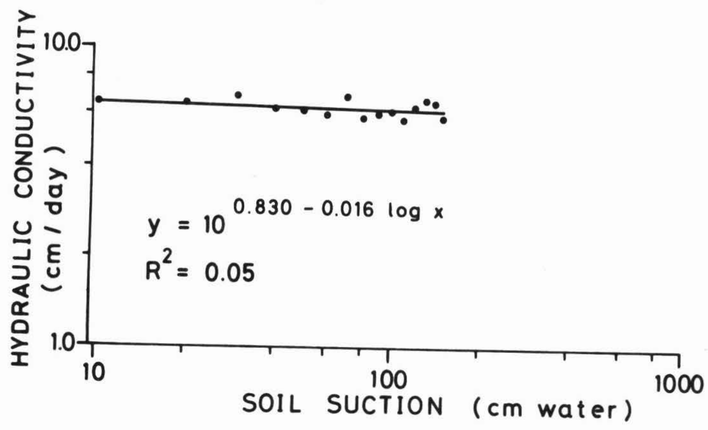


Fig: 17: Hydraulic conductivity data for the artificial quartz powder

by the regression line remained more or less constant over the entire suction range (10 to 150 cm water). This may be said to be consistent with the shape of the soil moisture characteristic shown already. The hydraulic conductivity functions of the three different soils are given in Figs. 18a and b. As a test for the variability in results, the confidence intervals are shown. In the grey brown podzolic soil (Fig. 18a) and gley podzol (Fig. 18b), the upper and the lower values of the confidence intervals differed on the average by a factor of 2-3. The narrowest confidence interval was obtained for the 10-20 cm depth of the grey brown podzolic soil. For the "Pelosol" (Fig. 18b), the upper and the lower values of the confidence interval lie by a factor of 6 apart on the average. The contributing factor could be the polyhedral form of the soil aggregates found in this soil type. The sizes of these aggregates differ appreciably over short distances within a soil horizon. Therefore the somewhat discrete nature of

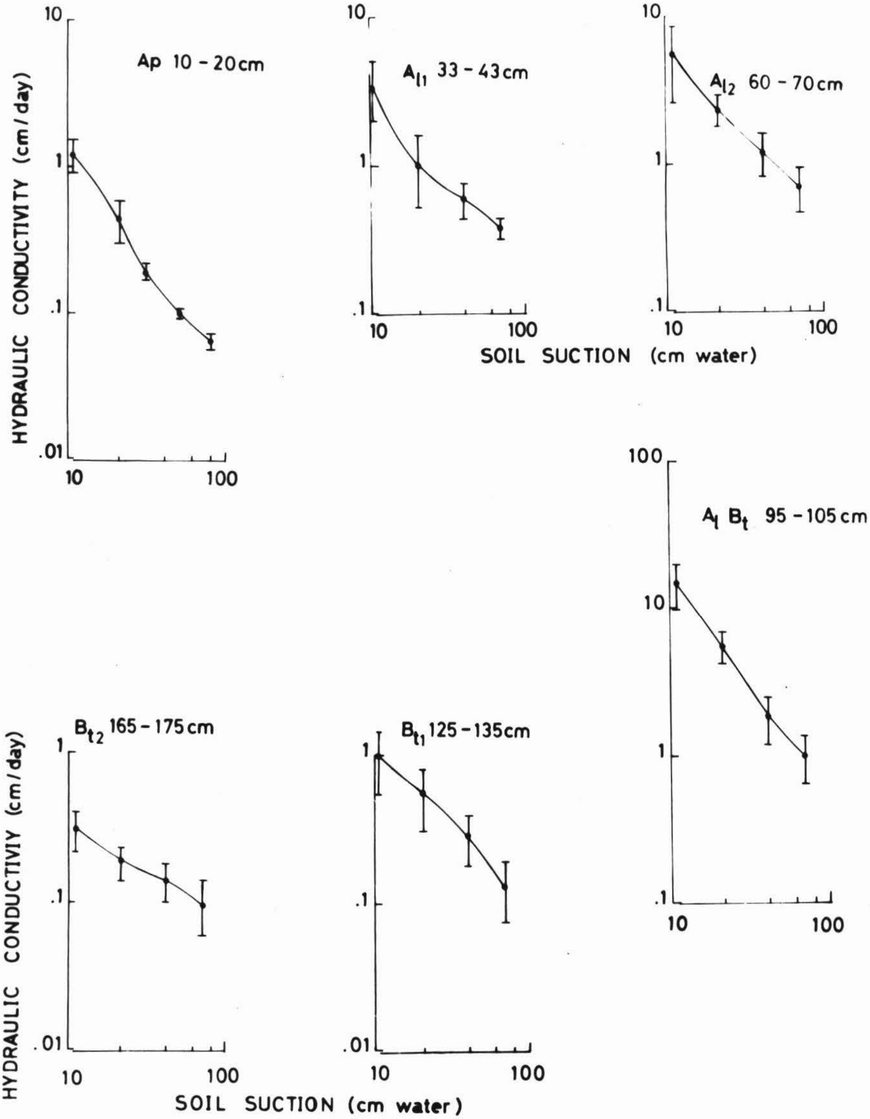


Fig. 18a: Hydraulic conductivity data obtained by the steady-state method (short column version) at six depths of the grey brown podzolic soil.

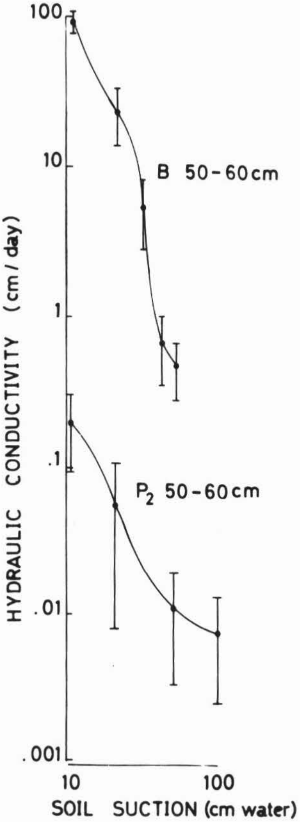


Fig. 18b: Hydraulic conductivity data obtained by the steady-state method (short column version) for one depth of the gley podzol and one depth of the "Pelosol".

soil pores might lead to a variation in the hydraulic conductivity values of replicate samples. The steep decrease of conductivity function over a small increase in suction (50 cm water) by the gley podzol is consistent with the soil-moisture retention.

5.2.6. Steady-state method: Long column version

Fig. 19 shows the hydraulic conductivity-soil suction relationship measured at different depths of the grey brown podzolic soil using both the samples in acrylic glass and

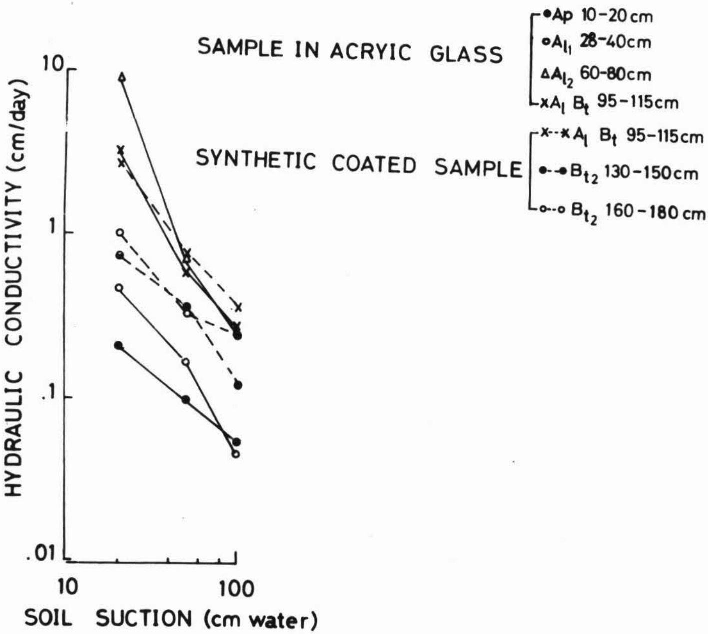


Fig. 19: Hydraulic conductivity data by the steady-state method (long column version) using both samples in acrylic glass and synthetic coated samples.

the synthetic coated samples. The conductivity values given here are results from one sample for each of the depths or sampling technique. As had been mentioned earlier, the difficulty involved in these sampling technique as well as the long time required to reach steady-state by these samples, did not allow for several replicates and measurements to be made. As is shown in the figure, conductivity values for the 95-115 cm depth using both sample types did not deviate much from each other.

Conductivities were higher deeper in the soil profile than in the upper depths. However, the highest conductivity values were measured for the 60-80 cm depth. The highest and lowest conductivity values for the different depths differed by a factor of 45 at 20 cm water suction. This difference was only a factor of 6 at 100 cm water suction. This could be attributed to the sizes and number of water conductivity pores in the different depths and at different suction levels. Because of the structural difference found in the different horizons, the sizes and number of the larger pores vary considerable. The smaller pores, whose sizes are dependent on the texture rather than the structure of the soil do not vary much from horizon to horizon. These smaller pores are important for the conductivity of water at higher suctions.

5.3. Results by field method

Count ratios obtained using the neutron moisture meter correlated poorly with the moisture content at the different depths. Thus the calibration curves were found inadequate for use in monitoring the absolute moisture content in the field. Consequently, the soil moisture data from sample (% weight) of 10 cm soil layers were plotted as a function of time and curves during desiccation periods were smoothed out by eye-fitting, so that the water content changed continuously from one day to the next (Ehlers, 1976). From the curves, daily volumetric water content changes above and below the water divide were calculated using the bulk densities of the specific layers. The hydraulic gradients were computed from the tensiometer readings. Fig. 20 shows the calculated hydraulic conductivities for the 10-20 cm, 20-40 cm, 60-80 cm, 80-100 cm, and 120-140 cm soil depths. From the figure, it could be seen that the suction range, at which $K_{(\psi)}$ calculation was possible for all the soil depths was small (usually between 150 and 450 cm of water). This indicates that the moisture content change in the soil profile was small owing to the weather condition of the period (April - September 1977) during which the experiment was carried out, which was very dry. Because the previous year (1976) was also dry, the deficit in the soil moisture in that year could not be made up in the next year. Therefore the soil moisture suction below 60 or 80 cm depth was always more than 100 cm water. The correlation coefficients for the different depths are low when compared for example with the results in section 5.2.3. This indicates that the scattering of data is high. It may be necessary to consider that the data presented here were calculated from measurements, which extended over several months. During this time, hysteresis might have played a significant role, so that the water contents measured for the same suction might have varied considerably during sorption and desorption phases. As was shown by van Bavel et al. (1968), Nielsen et al. (1973) and quoted by Ehlers (1976), the hydraulic properties of the soil under field conditions might hold a considerable spatial variability.

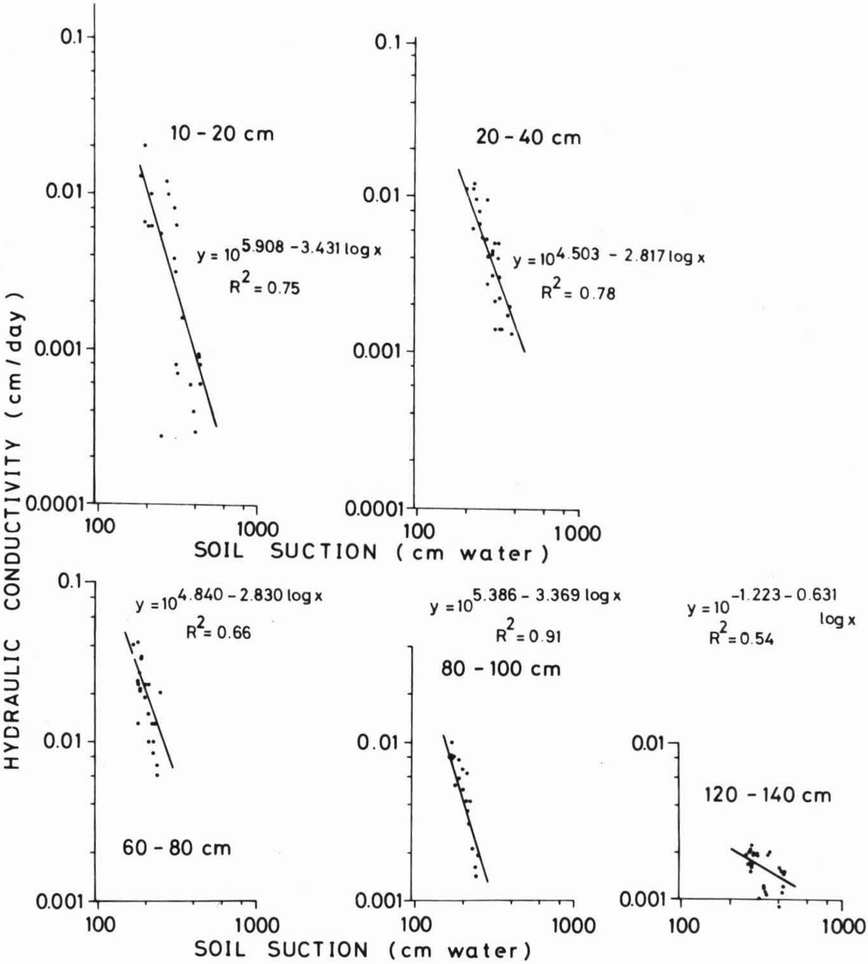


Fig. 20: Field determined hydraulic conductivity data.

5.4. Comparison of methods

5.4.1. Same methods applied to different soils

In this section, the hydraulic conductivity data obtained by the evaporation method and the steady-state method (short column version) will be compared with one another. The evaporation method was chosen because it is a quick method. Secondly, unlike the one-step method, which is also a quick method, and the other laboratory methods, there is no resistance to the flow of water since the method does not involve the use of porous end plates. Figs. 21a and b show the conductivity functions for four depths of the grey brown podzolic soil and the 50-60 cm depth of the gley podzol and the 50-60 cm depth of the "Pelosol". Also shown in these figures are the confidence intervals of the arthmetic means by the steady-state method. Similar calculations of the confidence intervals by the evaporation method was not feasible for reasons given earlier in section 5.2.1. As can be seen from Fig. 21a, the conductivity results obtained by the two methods agree very well in the Ap 10-20 cm depth of the grey brown podzolic soil. Similar agreement can also be said of the "Pelosol" (Fig. 21b) in the suction range where the two methods overlap. On the other hand, the hydraulic conductivity as obtained by the two methods show different results for the B 50-60 cm depth of the gley podzol. As a reason for this deviation, it may be suggested that the conductivity function obtained by the evaporation method is related to the energy supply which might have limited evaporation and especially during the initial stages. Thus the smaller conductivity values by this method were perhaps due to the insufficient energy supply to the wet soil. However, the results from the two methods seem to converge with increased suction. Similar comparison could not be made in the 60-80 cm, 120-140 cm and 160-180 cm of the grey brown podzolic soil. The inability to obtain $K_{(\psi)}$ data in the low suction range (< 70 cm water) has already been explained in section 5.2.1. In these depths,

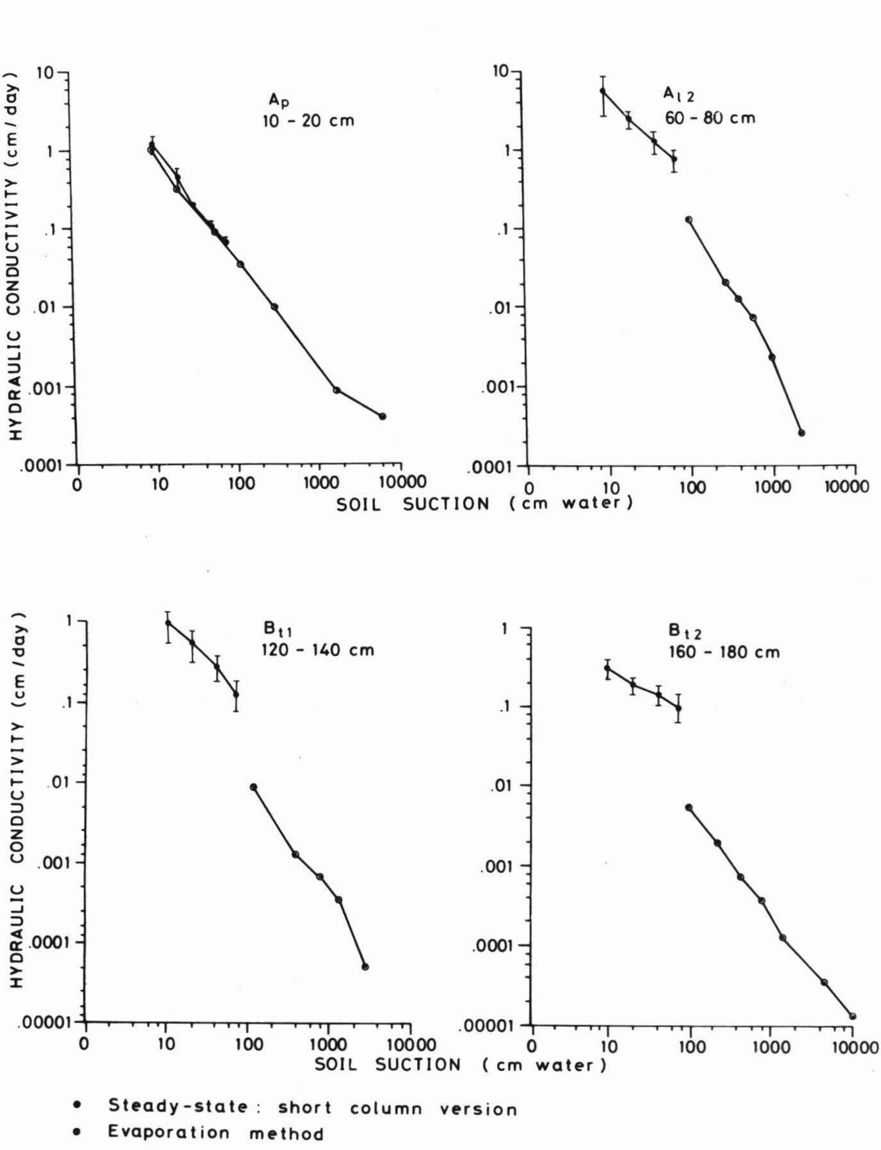


Fig. 21a: Hydraulic conductivity data obtained by the evaporation method and the steady-state method (short column version) at four depths of the grey podzolic soil.

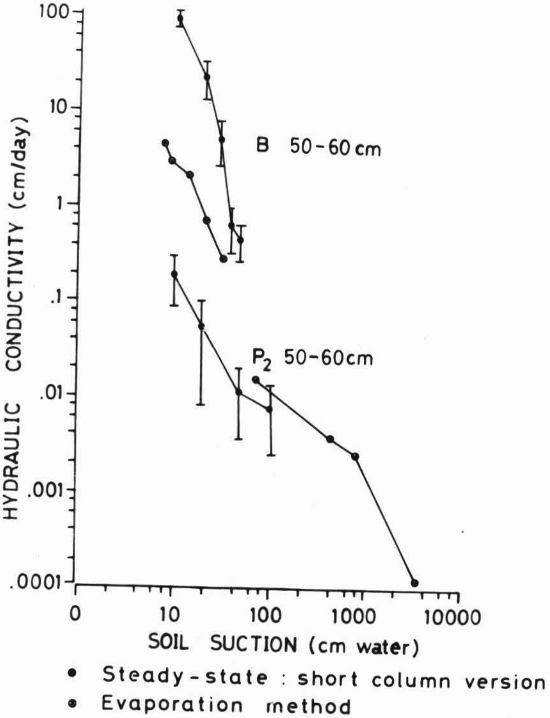


Fig. 21b: Hydraulic conductivity data obtained by the evaporation method and the steady-state method (short column version) for the gley podzol and the "Pelosol".

as well as by the B 50-60 cm depths, there is generally a tendency of lower $K_{(\psi)}$ data by the evaporation method. The opposite result of this tendency is obtained by the P_2 50-60 cm depth of the "Pelosol".

5.4.2. Different methods apllied to the same soil

The conductivity functions obtained by the different laboratory methods on the grey brown podzolic soil will be compared here with the steady-state method (short column version). As was mentioned in the introduction, the short column version of the steady-state method was chosen as a reference method since this method and in general steady-state methods are not subjected to restricting assumptions and/or simplifications. The samples used were considered homogenous since they were much shorter than the thickness of the horizons from which they were taken and the flow conditions could be physically controlled. With the aid of two tensiometers installed on the top and bottom of the sample, the hydraulic gradient was maintained always at 0.95 and 1.05 and therefore there was a linear hydraulic gradient always in the sample.

The hydraulic conductivity-suction relationship as was obtained by the different methods are shown in Figs. 22a and b for six depths of the grey brown podzolic soil.

Conductivity data by the outflow method (short column - small increment version), one-step method and the steady-state method (short column version) were obtained using the same samples (250 ml cores). Similarly, the steady-state method and the outflow method (long column versions) were on the same samples of either in acrylic glass or with synthetic coat.

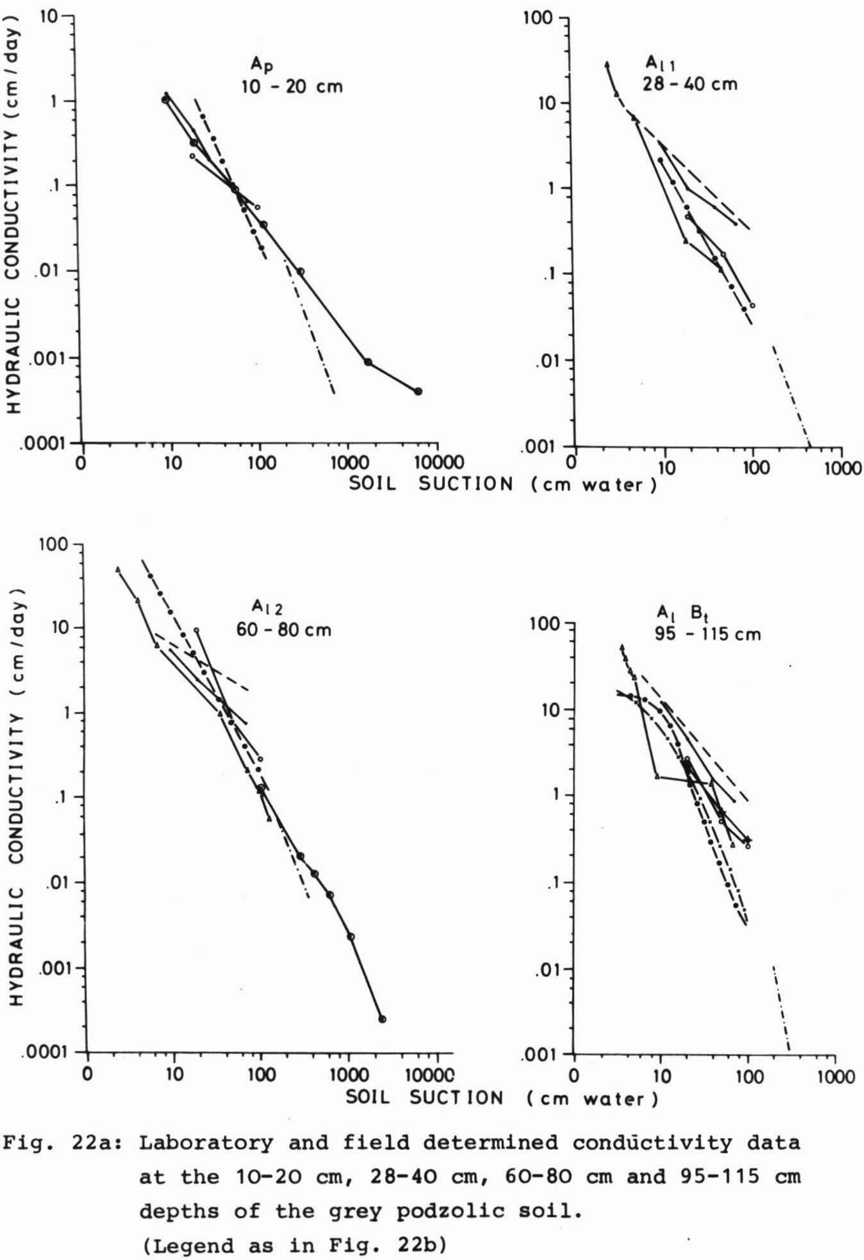


Fig. 22a: Laboratory and field determined conductivity data at the 10-20 cm, 28-40 cm, 60-80 cm and 95-115 cm depths of the grey podzolic soil. (Legend as in Fig. 22b)

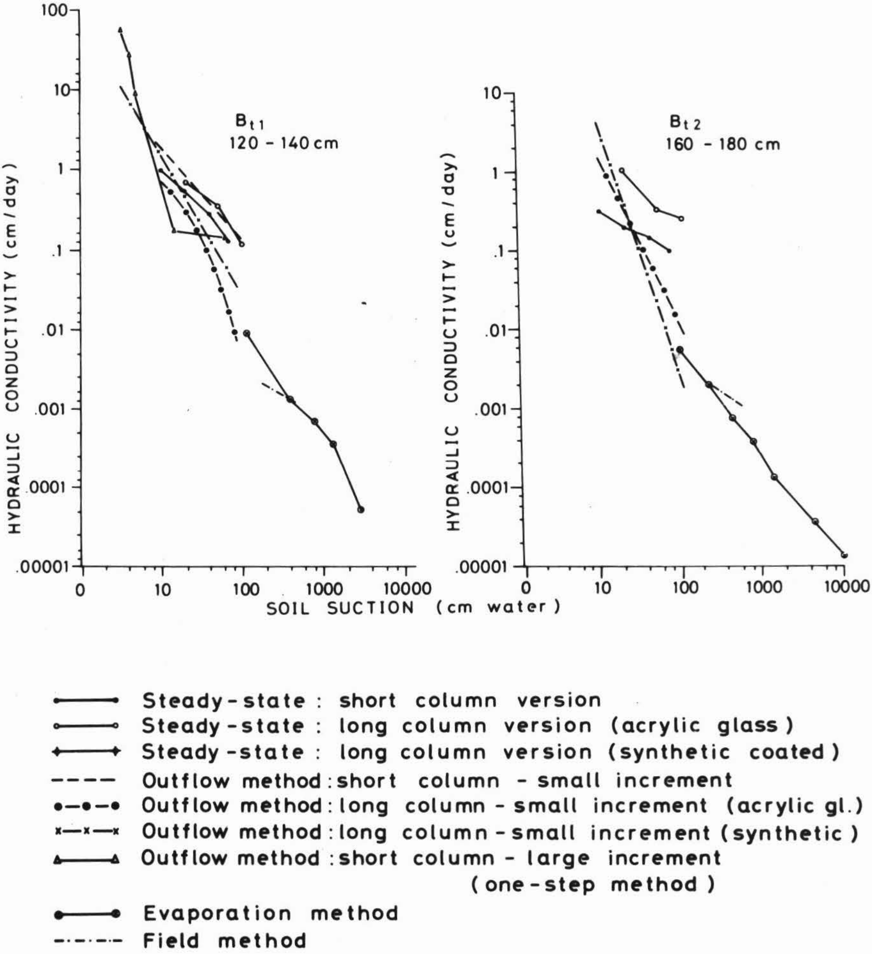


Fig. 22b: Laboratory and field determined conductivity data at the 120-140 cm and 160-180 cm depths of the grey podzolic soil.

In order to have standardized criteria for the comparison of the methods, tables 6a and b were prepared with the same data as in Figs. 22a and b. In the table 6a, the conductivity data by the different methods are compared with the steady-state method (short column version) for each depth. Table 6b shows the data from each method and for all the depths. Comparison is made with respect to the deviation of the methods from the reference method at different suction levels, and their slope characteristic. As a standardized criterion, the difference of logarithm was used because this gives a constant proportionality between the data and the reference k-value. The following classification was used:

Deviation (S)		% Deviation	
0.1	≙	25	very good
0.2	≙	60	good
0.3	≙	100	fair
0.4	≙	150	poor
0.5	≙	320	very poor
> 0.5			not comparable

This proportionality holds always with respect to the smaller value, that is, if the deviation is > 0, the compared conductivity value is as indicated, bigger than the value by the steady-state method (short column version). On the other hand, if the deviation is < 0, then the value by the steady-state method is bigger than the compared value by this number. In the last column of tables 6a and b, an inference based on these criteria has been drawn as to how good or bad the different results agree with the result from the reference method.

In the Ap 10-20 cm depth, the evaporation method agrees well with the reference method. The results by the steady-state method (long column version) with samples in acrylic glass also agree fairly well with results from the short column version. However, there is tendency of lower k values

Horizon/Depth (cm)	Method	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	**Inference**
Ap 10-20	**Steady-state method: long-column version (samples in acrylic glass)**	-	-0.32	-0.10	-0.05	0	+0.02	+0.02	gentler slope	good
	Evaporation method	-0.11	-0.15	0	-0.01	-0.02	-0.02	-0.02	more or less the same slope	very good
	Outflow method: long column version (sample in acrylic glass)	-	+0.72	+0.67	+0.16	+0.15	-0.14	-0.39	steeper	very poor
	Field method	-	-	-	-	-	-	-	much steeper	
Al_1 28-40	**Steady-state method: long column version (sample in acrylic glass)**	-	-0.26	-0.38	-0.44	-0.47	-0.63	-	steeper	poor
	One-step method	-0.46	-0.64	-0.65	-0.67	-	-	-	steeper	not comparable
	Outflow method: short column - small increment	0	+0.32	+0.17	+0.13	+0.11	+0.08	-	more or less the same	very good
	Outflow method: long column - small increment version (acrylic glass)	-0.20	-0.26	-0.50	-0.60	-0.70	-0.86	-	steeper	not comparable
	Field method	-	-	-	-	-	-	-	very much steeper	

Horizon/Depth (cm)	Method	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	Inference
	Steady-state method: long column version (acrylic glass)	-	+0.59	+0.28	+0.04	+0.12	+0.19	-	very much steeper	fairly good
	One-step method	-0.14	-0.11	-0.16	-0.23	-0.35	-0.56	-	more or less the same	fairly good
	Evaporation method	-	-	-	-	-	-	-	steeper	
Al_2 60-80	Outflow method: short column version	+0.08	+0.26	+0.30	+0.34	+0.37	+0.41	-	gentler slope	fair
	Outflow method: long column - small increment version (acrylic glass)	+0.47	+0.25	+0.06	-0.07	-0.16	-0.31	-	much steeper	good
	Field method	-	-	-	-	-	-	-	very much steeper	
	Steady-state method: long column version (acrylic glass)	-	-0.24	-0.29	-0.32	-o.40	-0.38	-	more or less the same	fairly good
A_1B_t 95-115	Steady-state method: long column (synthetic coated)	-	-0.34	-0.29	-0.26	-0.28	-0.34	-	more or less the same	fairly good
	One-step method	-0.89	-0.50	-0.25	-0.14	-0.31	.	.	more or less the same	poor

Horizon/Depth (cm)	Method	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	**Inference**
	Outflow method: short column - small increment version	+0.03	+0.10	+0.17	+0.22	+0.20	+0.19	-	more or less the same	good
A_1B_t 95-115	Outflow method: long column - small increment (acrylic glass)	+0.14	-0.40	-0.65	-0.78	-0.90	-1.19	-	much steeper	not comparable
	Outflow method: long column - small increment (synthetic coated)	-0.35	-0.40	-0.49	-0.59	-0.68	-0.94	-	much steeper	not comparable
	Field method	-	-	-	-	-	-	-	very much steeper	
	Steady-state method: long column version (synthetic coated)	-	+0.11	+0.14	+0.17	+0.23	+0.20	-	more or less the same	good
	One-step method	-0.19	-0.53	-0.38	-0.28	-0.16	-	-	steeper	fair
Bt_1 120-140	Evaporation method	-	-	-	-	-	-	-	steeper	
	Outflow method: short column - small increment	+0.28	+0.19	-0.15	+0.13	ü0.16	+0.20	-	more or less the same	good
	Outflow method: long column - small increment (acrylic glass)	-0.13	-0.24	-0.36	-0.49	-0.56	-0.76	-	much steeper	poor

Horizon/Depth (cm)	Method	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	Inference
Bt_1 120-140	Outflow method: long column - small increment (synthetic coated)	+0.23	-0.08	-0.20	-0.29	-0.31	-0.36	-	steeper	fairly good
	Field method	-	-	-	-	-	-	-	gentler slope	
Bt_2 160-180	Steady-state method: long column version (synthetic coated)	-	+0.71	+0.57	+0.47	+0.42	+0.47	-	steeper	very poor
	Evaporation method	-	-	-	-	-	-	-	steeper	
	Outflow method: long column - small increment (acrylic glass)	+0.68	+0.22	-0.10	-0.30	-0.47	-0.69	-	much steeper	poor
	Outflow method: long column - small increment (synthetic coated)	+1.10	-0.30	-0.19	-0.56	-0.81	-1.20	-	very much steeper	not comparable
	Field method	-	-	-	-	-	-	-	gentler slope	

log (value considered) $\pm$ S = log (reference value)

Table 6a: Comparison of hydraulic conductivity data by the different methods (laboratory and field) with the data by the steady-state method (short column version) at six depths of the grey brown podzolic soil.

Method	Horizon/Depth (cm)	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	Inference
Steady-state method: long column version (sample in acrylic glass)	Ap 10-20	-	-0.32	-0.10	-0.05	0	+0.02	+0.02	gentler slope	good
	Al_1 28-40	-	-0.26	-0.38	-0.44	-0.47	-0.63	-	steeper	poor
	Al_2 60-80	-	+0.59	+0.28	+0.04	-0.12	-0.19	-	very much steeper	fairly good
	A_1B_t 95-115	-	-0.24	-0.29	-0.32	-0.40	-0.38	-	more or less the same	fairly good
Steady-state method: long column version (synthetic coated sample)	A_1B_t 95-115	-	-0.24	-0.29	-0.26	-0.28	-0.34	-	more or less the same	fairly good
	Bt_1 120-140	-	+0.11	+0.14	+0.17	+0.23	+0.20	-	more or less the same	good
	Bt_2 160-180	-	+0.71	+0.57	+0.47	+0.42	+0.47	-	steeper	very poor
One-step method	Al_1 28-40	-0.46	-0.64	-0.65	-0.67	-	-	-	steeper	not comparable
	Al_2 60-80	-0.14	-0.11	-0.16	-0.23	-0.35	-0.56	-	more or less the same	fairly good
	A_1B_t 95-115	-0.89	-0.50	-0.25	-0.14	-0.31	-	-	more or less the same	poor
	Bt_1 120-140	-0.19	-0.53	-0.38	-0.28	-0.16	-	-	steeper	fair
Outflow method: short column - small increment	Al_1 28-40	0	+0.22	+0.17	+0.13	+0.11	+0.08	-	more or less the same	very good
	Al_2 60-80	+0.08	+0.26	+0.30	+0.34	+0.37	+0.41	-	gentler slope	fair
	A_1B_t 95-115	+0.30	+0.10	+0.17	+0.22	+0.20	+0.19	-	more or less the same	good
	Bt_1 120-140	+0.28	+0.19	+0.15	+0.13	+0.16	+0.20	-	more or less the same	good

Method	Horizon/Depth (cm)	Characteristics of the other methods in comparison with the steady-state method (short column version) Deviation (S) at suction levels (cm water) of								
		10	20	30	40	50	70	80	Slope of curve	**Inference**
Outflow method: long column - small increment (sample in acrylic glass)	Ap 10-20	-	+0.72	+0.67	+0.38	+0.16	+0.14	+0.39	steeper	very poor
	Al$_1$ 28-40	-0.20	-0.26	-0.50	-0.60	-0.70	-0.86	-	steeper	not comparble
	Al$_2$ 60-80	+0.47	+0.25	+0.06	-0.07	-0.16	-0.31	-	much steeper	good
	A$_1$B$_t$ 95-115	-0.14	-0.40	-0.65	-0.78	-0.90	-1.19	-	much steeper	not comparable
	Bt$_1$ 120-140	-0.13	-0.24	-0.36	-0.49	-0.56	-0.76	-	much steeper	poor
	Bt$_2$ 160-180	+0.68	+0.22	-0.10	-0.30	-0.47	-0.69	-	much steeper	poor
Outflow method: long column - samll increment (synthtetic coated sample)	A$_1$B$_t$ 95-115	-0.35	-0.40	-0.49	-0.59	-0.68	-0.94	-	much steeper	not comparable
	Bt$_1$ 120-140	+0.23	-0.08	-0.20	-0.29	-0.31	-0.36	-	steeper	fairly good
	Bt$_2$ 160-180	+1.10	+0.30	-0.19	-0.56	-0.81	-1.20	-	very much steeper	not comparable

Table 6b: Comparison of hydraulic conductivity data (same data as in table 6a) of the different laboratory methods with the data by the steady-state method (short column version) at different depths.

by the long column version in the lower suction range (< 20 cm of water). Deviation in results is obtained by the outflow method (long column version) with the samples in acrylic glass. It can be seen from Fig. 22a that all the methods except the field method give the same k value at about 56 cm of water suction level. In the Al_1 28-40 cm depth, results from the different methods with the exception of the outflow method (short column - small increment version) deviate from the result of the steady-state method (short column version). The deviation is more with increased suction. There is however a tendency for similar results by all the methods in the lower suction range (< 10 cm of water). In the Al_2 60-80 cm depth, the conductivity values as measured by the one-step method agree best with the reference values. From Fig. 22a it can be seen that the k function by one-step method lies below the k function by the reference method. The opposite of this result was obtained by the outflow method (short column version). The conductivity values as were determined by both the steady-state method (long column) and the outflow method (long column) using samples in acrylic glass are higher than the values by the reference method at suction levels < 40 cm and < 30 cm of water column, respectively. Above these suction levels, the two methods give lower results.

In the A_1B_t 95-115 cm depth, the outflow method (short column - small increment version) yields results which agree best with the results by the reference method. As in the 60-80 cm depth, results by the former are higher at all suction levels than the reference method. The other laboratory methods all give k values which are lower than the values by the steady-state method (short column version). However the results by the steady-state method (long column version) with samples in acrylic glass or synthetic coated samples agree better than the results from the same samples but using the outflow method. In the Bt_1 120-140 cm depth, the steady-state method (long column version) with the synthetic coated sample and the outflow method (short column - small increment version) give more or less the similar results. The k values by these two methods are little higher than those of the reference method at the same suction. The k values as were determined by the outflow method (long column version) using the sample in acrylic glass and synthetic coated sample are lower than the k values by the reference method except below the 20 cm water suction where by the synthetic coated sample higher k values are obtained. As in the 95-115 cm depth, results by the outflow method (long column version) deviate more and more from that of reference method with increased suction. Similarly the form of the k function by the one-step method is similar to that of the same method in the 95-115 cm depth.

In the Bt_2 160-180 cm depth, the results by the other methods deviate considerably from the results obtained by the steady-state method (short column version).

As a final observation it can be seen from Figs. 22a and b that the slope of the k function by the evaporation method corresponds to the slopes by outflow method (long column version) in the 60-80 cm, 120-140 cm and 160-180 cm depths.

From Table 6b it can be seen that the hydraulic conductivity data as determined by the steady-state method (long column and using the sample in acrylic glass) are higher than the reference data in the 60-80 cm depth but lower in the 20-40 cm and 95-115 cm depths. In the 10-20 cm depth, lower values are obtained at suction levels $\leq$ 40 cm of water column but higher values at the 70 cm and 80 cm of water column suction levels. On the average, the data by this method deviate by $\pm$ 0.26 from the reference data. The slopes of the curves vary from gentler to very much steeper slope compared with the slope of the reference curve.

Using the same method but the synthetic coated sample, it can be said that similar result as mentioned above are also obtained. In the 95-115 cm depth, the k data are smaller than those of the steady-state method (short column version) whereas in the 120-140 cm and 160-180 cm depths, higher k values are obtained. The average deviation is $\pm$ 0.41 by this sample. The slope of the curves in the 95-115 cm and 120-140 cm depths is almost the same as the slope by the reference curve but a steeper curve is obtained in the 160-180 cm depth.

The one-step method gives data which are lower at all the suction levels and in all the depths. The average deviation is - 0.38. Steeper slopes are obtained in the 28-40 cm and 120-140 cm depths whereas in the 60-80 cm and 95-115 cm depths the slopes of the curves are similar to the reference curve.

By the outflow method (short column - small increment version), the results are opposite to those by the one-step method, that is, the k values are higher at all suction levels and in all the depths. The average deviation is + 0.20. The slopes of the curves are almost the same as those of the steady-state method (short column) except in the 60-80 cm depth where a gentler slope is obtained.

For the samples in acrylic glass column, the conductivity data as was determined by the long column version of the outflow method are higher than those obtained by the reference method in the 10-20 cm depth. The opposite of this result is obtained in the 95-115 cm and 120-140 cm depths. The average deviation is $\pm$ 0.44. The curves are all steeper than the reference curve.

For the synthetic coated sample, the k data are higher in the 95-115 cm depth. The same could be said for the 120-140 cm and 160-180 cm depths except at the 10 cm and 20 cm of water column suction levels, respectively. The deviation is on the average $\pm$ 0.51. The slope of the curves is steeper. From the results shown in table 6b, it can be concluded that the steady-state method (long column version and using samples in acrylic glass) and the outflow method (short column - small increment version) gave results which are in reasonable agreement with the results by the steady-state method (short column version). The best agreement is however obtained by the latter of the two methods.

5.5 CLOSING REMARKS

In judging the usefulness of the different laboratory methods for the routine determination of the hydraulic conductivity of unsaturated soils, it might be concluded that no one method has completely surpassed the others. The different methods have their advantages and disadvantages.

In this study the evaporating method has not been found very useful for the routine determination of hydraulic conductivity since the method seems to be too limited to certain soils, and the suction range at which results may be obtained seems also to depend on the form of the moisture characteristic of the soil. However this method yields results in the higher suction ranges where it is difficult to obtain results from the other methods since the other methods are dependent on the estimation of outflow volumes. For problems dealing with processes which are significant in the higher suction range (which have been excluded in this study) such as water uptake by roots, the usefulness of this method must be evaluated quite differently. The method is rather easy and quick and except the pressure plate apparatus (for determining soil-moisture characteristic), no other special and expensive laboratory equipment is needed. A disadvantage of the method is the calculation involved. Another disadvantage of this method could be the relative inaccuracy of graphical differentiation and integration as discussed by Kirkham and Powers (1972). A further disadvantage is that preliminary tests are always required for each soil and the different depths in order to select the temperature of the warm air and the distance of the air outlet to the soil surface so as to attain the cumulative evaporation-square root of time linearity. As a conclusion, it may be said that this method is best suited to soils with uniform pore-size distribution.

The outflow method generally involves calculations which are complicated and time consuming. Since the flux and the hydraulic gradients are time dependent factors, they need to be measured often at various times. So an experimental setup for an outflow experiment may require expensive laboratory equipments to measure tension or water content changes or both. However, the short column - small increment version as was done in this work may be found very useful in the routine determination of the hydraulic conductivity of unsaturated soils. The method may be found relatively quick since it involves only the measurement of the cumulative outflow and the tensiometer reading. In addition, a simple graphical analysis of data has been employed. The hydraulic gradient and the instantaneous outflow rate are taken only at one point and are obtained directly from the tensiometer reading versus time and the cumulative outflow-time curves, respectively. The results may be as reliable as those produced by the steady-state method (short column version).

The outflow method (short column - large increment version) known as the one-step method has the advantage of greater speed as compared to the small increment version since only one equilibration is needed. The results obtained may not be as reliable as those produced by the small increment version since by the one-step method the slope of the moisture characteristic is additionally needed to calculate the conductivity and thus increase the variability in results. The volumetric measurements are somewhat easier because of the relatively larger volumes to be measured. If the moisture-content-pressure head is known (from separate measurements), the conductivity function may be calculated. Like the evaporation method, the one-step method is only applicable to the determination of the drainage diffusivity function.

Steady-state methods in general have the disadvantage of requiring relatively long times to establish steady flow. The short column version has been used as a reference method since the method can be controlled physically and requires no assumptions and/or simplifications. Because of the relatively easy sampling technique (core sampling) as opposed to the acrylic glass columns or synthetic coated samples, replicates of up to 10 per soil depth may be taken. But when and where laboratory determinations on such large number of samples are quickly required, the steady-state method (short column version) may not be found very useful since it requires a long time to establish steady flow. The method yields directly a conductivity function and the calculations required are quite simple and not time consuming. The inaccuracy in results obtained are generally small since the method does not involve the determination of the soil-moisture characterisitc curves which in themselves have some degree of variability.

The soil sampling technique played an extra role in the laboratory determination. From the results obtained in section 5.2.3. it can be said that the polyester sampling technique could be restricted to stony soils, in which case soil samples can be obtained without distrubing the soil structure. This is because this sampling technique and the acrylic glass soil columns yielded results which are close but the polyester sampling technique requires more labour and time. A further advantage of samples taken in acrylic glass is that the column is transparent and tensiometers can be installed exactly at the boundaries of soil horizons. Nevertheless, both of these sampling techniques are more tedious and require more time than taking samples with short cores. Secondly, longer time is required to determine the moisture transmission of such large columns or blocks. From table 6b it can be concluded that the steady-state method yielded better results than the outflow method using these two sampling techniques.

Among the different laboratory methods tested in the work, the outflow method (short column - small increment version) could be considered as a method which has all the good characteristics of the steady-state method (short column version). In addition the method is quick and the method of graphical analysis of data is simple. As a conclusion, it can be said that the outflow method (short column version) could be a good substitute for the steady-state method (short column version).

6. SUMMARY

Quantitative description of the soil water behaviour in the unsaturated soil zone requires knowledge of the hydraulic conductivity and moisture characteristics of the soils involved. Mathematical models are increasingly used to simulate the behaviour of water in the unsaturated soil zone. This in turn provides a comprehensive means of linking all processes involved in the water turnover of ecosystems and thus determining for instance the components of the water balance equation such as seepage, runoff or evapotranspiration as a function of time. Therefore, the determination of these time dependent components of the water balance equation for ecosystems requires the determination of reliable hydraulic conductivity as a function of water content or suction of the individual horizons of the soil being investigated.

On the other hand, only a limited range on the low suction side of the conductivity function is needed, $0 < \psi < 100$ cm water suction, as long as attention is primarily paid on the calculation of the downward seepage below the root zone. This is because this output variable approaches negligible rates once the suction surpasses values well below 100 cm water column in the region below the root zone. For this reason, the hydraulic conductivity as a function of suction was primarily determined by means of different methods within the mentioned suction interval.

Reviewing the literature, one finds numerous and different concepts and approaches for the determination of the $K_{(\theta)}$ or $K_{(\psi)}$ functions (K = hydraulic conductivity, θ = water content, ψ = suction). But none of these concepts and approaches could be said to be superior to the other for they have their advantages and disadvantages.

Some of the existing methods were tested again and new versions and techniques were also introduced. The criteria for the selection of only some of these numerous methods as well as the introduction of new versions and techniques were based on practical reasons, which were mainly equipment available for the determination, time requirement and sampling techniques. Because of their simple theoretical base and straightness of experimental application, methods based on a steady-state flux have been given preference so far in the already mentioned experimental work. But, nevertheless, the question of verification of the methods remained as well as the desire for more effecitve methods.

Verification was sought by comparing results with corresponding ones of field method. More effective, that is in particular less time consuming methods were expected from non-steady-state methods. Further problems arose with respect to stony and heterogeneous (layered) soils, requiring special sampling techniques and sample volumes. It is essential to answer the question if sampling technique as well as sample size and treatment give similar results as by the steady-state method on "homogeneous" soil samples, that is, small undisturbed samples.

After reviewing the theoretical fundamentals associated with the methods used in this work and some of the numerous publications, the methods were described and the hydraulic conductivity of samples measured. The results obtained from the different methods were discussed under two viewpoints, namely: Firstly, the results obtained from each method were given and discussed; secondly, as the main aspect of this work, the results by the different methods were compared using the steady-state method (short column version) as a reference. In addition to the experimental data, the advantages and disadvantages of the methods based on such criteria as the amount of work and length of time required were also given.

The evaporation method was compared with the reference method using their soil types: a grey brown podzolic soil derived from loess, a gley podzol and a "Pelosol". Conductivity data by the two methods agreed best in the Ap 10-20 cm depth of the grey brown podzolic soil and fairly good in the P_2 50-60 cm of the "Pelosol". Deviation was found by the B 50-60 cm of the gley podzol. Similar comparison could not be made in the other depths of the grey brown podzolic soil because the evaporation method yielded results in the higher suction levels only.

The conductivity functions obtained by the different methods on the grey brown podzolic soil were compared. The deviation (S) of the K data of the method from the reference method at different suction levels was given as: log (value considered) $\pm$ S = log (reference value), whereby if S > 0, the considered value > reference value and if S < 0, then the reference value > than the considered value. The deviation of the results ranged from $\pm$ 0.2 to $\pm$ 0.5 units on the average. The best agreement was obtained from the outflow method (short column - small increment version) in which the average deviation was + 0.2.

Generally it can be said that no method completely surpasses the other since all the methods have their advantages and disadvantages. However, for the routine determination of the hydraulic conductivity in the laboratory, the outflow method (short column - small increment version) could be a good substitute for the steady-state method. As far as results from this work are concerned, it cannot be said with all certainty to what degree the sampling technique, sample size and sample treatment affect the results. The two sampling techniques (which were considered specially in this work - samples in acrylic glass column and the synthetic coated samples) produce results which are quite reproducible and are close with one another. However, the results varied from that of the reference method and depended also on the methods used. It can be said that since the use of polyester coating on soil block is very tedious and time consuming, it may be suggested that this sampling technique be restricted only to stony soils where it may be difficult to obtain undisturbed samples by other sampling techniques.

7. LITERATURE

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