UNSATURATED HYDRAULIC CONDUCTIVITY

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I. INTRODUCTION

The unsaturated zone plays an important role in the hydrological cycle. It forms the link between surface water and ground water and has a dominant influence on the partition of water between them. The hydraulic properties of the unsaturated zone determine how much of the water that arrives at the soil surface will infiltrate into the soil, how much will flow off overland causing floods, erosion, etc. In many areas of the world, most of the water that infiltrates into the ground is transpired by plants or evaporated directly into the atmosphere, leaving only little water to percolate deeper and join the ground water. Surface runoff and deep percolation may carry pollutants with them. Then, it is important to know how long it will take for this water to reach surface or ground water resources.

Besides providing water for plants to transpire, the unsaturated zone also provides oxygen and nutrients to plant roots, thus having a dominant influence on the production of food, fiber, etc. Water content also determines soil strength, with many implications for anchoring of plants, root penetration, compaction by cattle and machinery, tillage operations, etc. To mention just one other role of the unsaturated zone, its water content has a great influence on the heat balance at the soil surface. This is well illustrated by the large diurnal temperature variations in deserts.

To understand and describe these and other processes, the hydraulic properties which govern water transport in the soil must be quantified. Of these, the unsaturated hydraulic conductivity is, if not the most important, certainly the most difficult to measure accurately. It varies over many orders of magnitude not only between different soils, but also for the same soil as a function of water content. Much has been published on the determination and/or measurement of the unsaturated hydraulic conductivity, including good reviews [1 - 7] There is no single method that is suitable for all soils and circumstances. Methods which require taking "undisturbed" samples are not well suited for soils with many stones or with a highly developed, loose structure. It is better to select an in situ method for such soils. Hydraulic conductivity for relatively dry conditions cannot be measured in situ when the soil in its natural situation is always wet. It is then necessary to take samples and dry them first. The latter process presents problems if the soil shrinks excessively on drying. These and other factors which influence the choice between laboratory and field methods are discussed separately in section IV.

Selection of the most suitable method for a given set of conditions is a major task. The literature is so exhaustive that it is neither necessary nor possible to give a complete review and evaluation of all available methods. Instead, I have focused on what I think should be the selection criteria (section III) and described the most familiar types of methods (in sections VI to IX) with these criteria in mind. This includes some very recent work. The need for and selection of a standard method is discussed separately in section V.

There are two soil water transport functions which, under restricting conditions, can be used instead of hydraulic conductivity, namely hydraulic diffusivity and matric flux potential. Diffusivity can be measured directly in a number of ways which are easier and faster than the methods available for hydraulic conductivity. Moreover, the latter can also be derived from the former. The same is true for yet another transport function, the sorptivity, which can also be measured more easily than the hydraulic conductivity. At the outset I have summarised the theory and transport coefficients used to describe water transport in the unsaturated zone (section II). Theoretical concepts and equations associated with specific methods are given with the discussion of the individual methods. Readers who have little knowledge of the physical principles involved in unsaturated flow and its measurement can find these discussed at a more detailed and elementary level in soil physics textbooks [8 - 10] and would be advised to consult one of these before attempting this chapter.

Apparatus for determining unsaturated hydraulic conductivity is not usually commercially available as such. However, many of the methods involve the measurement of water content, hydraulic head and/or the soil water characteristic, and methods and commercial supplies of equipment to determine these properties are given in chapters 1, 2 and 3, respectively. Where specialised or specially constructed equipment is required, this is indicated with the discussion of individual methods.

In general, it is difficult if not impossible to measure the soil hydraulic transport functions quickly and/or accurately. Therefore, it is not surprising that attempts have been made to derive them indirectly. The derivation of the hydraulic transport properties from other, more easily measured soil properties is discussed in section X and the inverse approach of parameter optimization in section XI.

II. TRANSPORT COEFFICIENTS

A. Hydraulic Conductivity

In general, water transport in soil occurs as a result of gradients in the hydraulic potential [10]

$$H = h + z \tag{1}$$

where H is hydraulic head, h is pressure head, and z is gravitational head or height above a reference level. These symbols are generally reserved for potentials on weight basis, having the dimension J/N = m. Although h is called a pressure head, in unsaturated flow it will have a negative value with respect to atmospheric pressure and can be referred to as a suction or tension. In rigid soils there exists a relationship between water content (usually expressed as volume fraction, θ (m^3/m^3)) and pressure head, called the soil water retention characteristic, θ[h] (see Chapter. 3). Here, as well as throughout this chapter, square brackets are used to indicate that a variable is a function of the quantity within the brackets. The function θ[h] often depends on the history of wetting and drying; this phenomenon is called hysteresis.
Water transport in soils obeys Darcy's law, which for one-dimensional, vertical flow in the z-direction, positive upward, can be written as

$$q = - k[\theta]\ \delta H/\delta z = - k[\theta]\ \delta h/\delta z - k[\theta] \tag{2}$$

where q is water flux density (m^3/m^2 s = m/s) and k[θ] is the hydraulic conductivity function (m/s). k is in the first place a function of θ, k[θ], since water content determines the fraction of the sample cross-sectional areas available for water transport. Indirectly, k is also a function of pressure head. k[h] is hysteretic to the extent that θ[h] is hysteretic. Hysteresis in k[θ] is of second order and is generally negligible. Determinations of k usually consist of measuring corresponding values of flux density and hydraulic potential gradient, and calculating k with Eq. (2). This is straightforward and can be considered as a standard for other, indirect measurements.

B. Hydraulic Diffusivity

For homogeneous soils in which hysteresis can be neglected or in which only monotonically wetting or drying flow processes are considered, $h[\theta]$ is a single-valued function. Then, for horizontal flow in the x-direction, or when gravity can be neglected, Eq. (2) yields

$$q = - D[\theta]\ \delta\theta/\delta x\ , \qquad D[\theta] = k[\theta]\ (dh/d\theta)[\theta] \tag{3}$$

where $D[\theta]$ is the hydraulic diffusivity function (m/s^2). Thus, under the above stated conditions the water content gradient can be thought of as the driving force for water transport, analogous to a diffusion process. Of course, the real driving force remains the pressure head gradient. Therefore, $D[\theta]$ is different for wetting and drying. There are many methods to determine $D[\theta]$, some of which will be described later. They usually require a special theoretical framework with simplifying assumptions. Once $D[\theta]$ and $h[\theta]$ are known, the hydraulic conductivity function can be calculated according to

$$k[\theta] = D[\theta]\ (\delta\theta/\delta h)[\theta] \tag{4}$$

Because of hysteresis, one should only combine diffusivities and derivatives of the soil water retention characteristic which both are obtained either by wetting or by drying. Since $k[\theta]$ is basically non-hysteretic, the $k[\theta]$ functions obtained along the two ways should agree closely.

C. Matric Flux Potential

Water transport in soils in response to pressure (matric) potential gradients can also be described in terms of the matric flux potential [11, 12]:

$$\Phi = \int_{-\infty}^{h} k[h]\ dh = \int_{0}^{\theta} D[\theta]\ d\theta \tag{5}$$

Equation (3) then becomes

$$q = - \delta\Phi\ /\ \delta z \tag{6}$$

The matric flux potential integrates the transport coefficient and the driving force; it has the dimension m^2/s. In homogeneous soil without hysteresis, the horizontal water flux density is simply equal to the gradient of Φ. This formulation of the water transport process offers distinct advantages in certain situations, especially in the simulation of water transport under steep potential gradients [12 - 14]. It also allows obtaining analytical solutions for steady state, multi-dimensional flow problems, including gravity, when the hydraulic conductivity is expressed as an exponential function of pressure head [15, 16]. Like k and D, Φ also is a soil property which characterises unsaturated water transport and is a direct function of θ and only indirectly of h. A method for measuring Φ directly [13] is described in section VI.D.

D. Sorptivity

Sorptivity is an integral soil water property that contains information on the soil hydraulic properties $k[\theta]$ and $D[\theta]$, which can be derived from it mathematically. Generally, sorptivities can be measured more accurately and/or more easily than $k[\theta]$ and $D[\theta]$, so it is worth considering to determine the latter in this indirect way [17, 18].
One-dimensional absorption (gravity negligible), initiated at time $t = 0$ by a step-function increase of water content from θ_0 to θ_1 at the soil surface, $x = 0$, is described [17, 19] by

$$i = S[\theta_1,\theta_0]\ t^{½} \tag{7}$$

where i is cumulative absorbed volume (m) at any given time t, and sorptivity S ($m/s^{1/2}$) is a soil property which depends on the initial and final water content, usually saturation. Saturated sorptivity characterises ponding infiltration at small times, as it is the first term in the infiltration equation of Philip [19] and equal to the amount of water absorbed during the first time unit. With the flux-controlled sorptivity method [17] the dependence of S on θ_1 at constant θ_0 is determined experimentally. From this $D[\theta]$ can be derived algebraically (subsection VIII.F, Eq. (27)). The $t^{½}$-relationship of Eq. (7) has also been used for scaling soils and estimating hydraulic conductivity [20] and diffusivity [21] of similar soils (section X.B).

III. SELECTION FRAMEWORK

A. Types of Methods

Many methods have been reported in the literature to determine soil water transport properties. There is no single method best suited for all circumstances. Therefore, it is necessary to select the method most suited to any given situation and time spent on this selection is well used. Table 1 lists various types of methods which have been proposed and presents an evaluation of these methods according to the 5 gradations of the selection criteria listed in Table 2. These tables form the nucleus of this chapter. In subsequent sections the various methods are reviewed in varying detail. In general, the theoretical framework and/or main working equations are described and other pertinent information is added to help substantiate the scores given for the various criteria in Table 1. Of the more familiar methods mostly only evaluating remarks are made; some experimental details are given also for the less familiar and newest methods. The scores are a reflection of my own insight and experience and are not (and cannot be) based solely on the information provided. For lacking information the reader is advised to consult the listed references.

A major division is made between steady state and transient measurements. In the first category, all parameters are constant in time. For this reason, steady state measurements are almost always more accurate than transient measurements, usually even with less sophisticated equipment. Their main disadvantage is that they take much more time, often prohibitively so. Therefore, the choice between these two categories usually involves balancing needed costs, available time, and required accuracy. The methods are divided further into field and laboratory methods, the choice of which is discussed in section IV. Methods for measuring soil water transport coefficients can also be divided in those that measure hydraulic conductivity directly and all other methods (column A). From what follows it should become clear that one should measure hydraulic conductivity whenever possible. The distinction made between wetting and drying flow regimes (B) is important because the hysteretic character of soil water retention may affect any application where hydraulic diffusivity or hydraulic conductivity are required as a function of pressure head.

B. Selection Criteria

The criteria on which the methods listed in Table 1 are evaluated are (see Table 2): the degree of exactness of the theoretical basis (C), the experimental control of the required initial and boundary conditions (D), the inherent accuracy of the measurements (E), the propagation of errors in the experimental data during the calculation of the final results (F), the range of pressure heads over which the method can be used (G), the time (duration) required to obtain the particular transport coefficient function over the indicated pressure head range (H), the necessary investment in workshop time and/or money (I), the skill required by the operator (J), the operator time required while the measurements are in progress (K), the potential for measurements to be made simultaneously on many soil samples (L), and the possibility for checking during and/or after the measurements (M). Depending on the particular situation, only a few or all of these criteria must be taken into account to make a proper choice. For example, accuracy will be a prime consideration for detailed studies of water transport processes at a particular site, whereas for a study of spatial variability the ability to make, in a reasonably short time, a large number of measurements is mandatory. These often do not have to be very accurate. If the absolute accuracy of a newly developed method must be established, the most accurate method already available should be selected, since there is no "standard" material with known properties available with which the method can be tested. The need for the selection of a "standard method", as alternative, is discussed in a separate section. When facilities for routine measurements must be set up, the last four criteria are particularly pertinent. Finally, there may be particular (difficult) conditions under which one method is more suitable than others, and these conditions may dominate the choice of method. Such criteria are not covered by Table 1, but are mentioned with the description of individual methods when appropiate.

The 5 gradations used with the selection criteria (Table 2) are mostly self-explanatory and will become clearer with the discussion of the individual methods. At this stage only a few general remarks are made about accuracy (relating to criteria C - F) and the range of application (G) which, out of practical considerations, is associated with pressure heads. For examples, reference is made to methods which are described later in more detail.

C. Accuracy

Direct measurements of weight, volume of water and time, made in connection with the determination of soil hydraulic properties, are simple and very accurate (maximum score 5). An exception is measuring very small volumes of water while maintaining a particular experimental set-up, for example a small hydraulic head gradient. Although the mass and water content of a soil sample can usually be accurately measured, the water content may not conform to what it should be according to the theoretically assumed flow system. For example, for Boltzmann transform methods a water content profile must be determined after an exact time period of wetting or drying. It is not possible to do this instantaneously and during sampling for gravimetric determinations, water contents will change due to redistribution and evaporation of water and due to manipulation of the soil. Indirect water content measurements can be made non-destructively and thus repeatedly during a flow process, but the accuracy of these measurements is normally not very good. Extensive calibration under identical conditions can improve the accuracy, but usually this is not possible or takes too much time.

Derivation of hydraulic properties from other measured parameters introduces two kinds of errors. Firstly, the theoretical basis of the method may not be exact, either because it involves simplifying assumptions or because the theoretical analysis of the water flow process yields only an approximation of the transport property. Secondly, errors in the primary experimental data are propagated in the calculations required to obtain the final results. Mathematical manipulations have each their own inherent inaccuracies, a good example being differentiation. Another common source of error is that the theoretically required initial and/or boundary conditions can not be attained experimentally. For example, it is impossible to impose the step-function decrease of the hydraulic potential at the soil surface under isothermal conditions, as is assumed with the hot air method.

Hydraulic potential measurements are relatively difficult and can be very inaccurate. Water pressures inside tensiometers in equilibrium with the soil water around the porous cup can in principle be measured to any desired accuracy with pressure transducers, but such measurements can become very inaccurate due to temperature variations. Mercury manometers are probably least sensitive to large errors, but their accuracy is limited to about ± 2.5 cm (see Ch. 2). In steady state measurements near saturation, water manometers appear to be most accurate. Beyond the tensiometer range, soil water potentials are mostly determined indirectly from soil water characteristics or by measuring the electrical conductivity, heat diffusivity, etc. of probes in equilibrium with soil water, with all the inaccuracies associated with indirect measurements. Direct measurements can be made with psychrometers (which also measure the osmotic component of the soil water potential) but these can only be used by experienced workers with sophisticated equipment and are at best accurate to about ± 500 cm. However, for many studies, such as that of the soil-water-plant-atmosphere continuum, such accuracies are acceptable, because hydraulic conductivities in this dry range are so low that hydraulic head gradients must be very large to obtain significant flux densities.

D. Range of Application

The range of application of a particular method depends to a large extent on whether and, if so, how soil water potentials are to be measured. Out of convenience and based on practical experience, therefore, the range of application is described with somewhat vague terms, which are identified further by approximate ranges of pressure head, even for methods in which only water contents or flux densities are measured. Tensiometers can theoretically be used down to pressure heads of about -8.5 m, but in practice air intrusion usually causes problems at much higher values. Fortunately, hydraulic transport properties need not be known in the drier range, except where water transport over small distances is concerned (e.g. evaporation at the soil surface, and water transport to individual plant roots). Water transport over large distances occurs mostly in the saturated zone (or as surface water), for which the saturated hydraulic conductivity must be known. However, there are some exceptions, such as saline seeps which are caused by unsaturated water transport over large distances during many years. Although unsaturated water transport normally occurs over short distances, it plays a key role in hydrology as mentioned in the introduction. The unsteady, mostly vertical water transport in soil profiles is only significant when the hydraulic conductivity is in the range from the maximum value at saturation to values down to about 0.1 mm/day, since precipitation, transpiration and evaporation can generally not be measured to that accuracy. This corresponds with a range in pressure head between 0 and -1.0 to -3.0 m, depending on the soil type.

The pressure head range over which hydraulic transport properties must be known should be carefully considered and be a major consideration in the selection process. It makes no sense, for instance, to determine hydraulic conductivities with the hot air method (which yields very inaccurate results over the entire pressure head range) when the results are only required for use in the hydrological range, for which much better methods are available. Conversely, it is dangerous to select an attractive method suitable only in the wetter range and to extrapolate the results to a dryer range. In practice, the range of application of a particular method depends also on the time required to attain appropiate measurement conditions. Criterion G and H are dependent: the time needed to measure the soil water property function often increases exponentially with increases in the pressure head range towards drier conditions.

E. Alternative Approaches

Because measurements of the soil water transport properties leave much to be desired in terms of their accuracy, cost, applicability, and time, it is not surprising that other ways to obtain these soil properties have been investigated. The most extreme of these approaches is not to make any water transport measurements, but to derive the water transport functions from other, more easily measured soil properties (e.g. particle size distribution or the soil water characteristic). These procedures are usually based on a theoretical model of the relationship [5, 6],but they can also be of a purely statistical nature [22, 23], in which case their application is limited to the range of soils used to derive the relationship. An intermediate approach is the so-called inverse approach, which has recently received renewed attention as the "parameter optimization technique" [7, 24, 25]. To be able to decide how the hydraulic transport functions can best be determined in a given situation, the possibilities and limitations of these alternative approaches should also be considered (section X and XI).

IV. LABORATORY VERSUS FIELD METHODS

A. Working Conditions

A major division between available methods is that of laboratory versus field methods. Laboratory measurements have many advantages over field measurements. In the laboratory all the usual facilities (e.g. electricity, gas, water, and vacuum) are available and temperature variations are usually modest and can be controlled, if necessary. Standard equipment (e.g. balances and ovens) is also more readily available than in the field. Expensive and delicate equipment can often not be used in the field because of weather conditions, theft, vandalism, etc. One can usually save much time by working in the laboratory. Samples from many different locations can then first be collected and measurements carried out consecutively or in series. Considering all these advantages, it would seem good practice to carry out measurements in the laboratory, unless there are overriding reasons to perform them in situ. For hydraulic conductivity measurements, this will normally only be the case if one needs the hydraulic properties of a strongly layered soil profile as a whole or if, due to heterogeneity and instability of soil structure, it is very difficult if not impossible to obtain large enough, undisturbed soil samples and transport them to the laboratory.

B. Sampling Techniques

Because the hydraulic conductivy of soil is very sensitive to changes in soil structure due to sampling and/or preparation procedures, these operations should be carried out with utmost care. Fractures formed during sampling which are oriented in the direction of flow are disastrous for saturated hydraulic conductivity determinations, but have very little influence on unsaturated hydraulic conductivities. Fractures perpendicular to the direction of flow have the very opposite effect on both types of measurements. Soil columns consisting of entire soil profiles can be obtained by driving a cylinder supplied with a sharp, hardened steel cutting edge into the soil with a hydraulic press. If the stroke of this press is smaller than the height of the sample, care should be taken that with each stroke the press is lined up exactly the same. We have been able to accomplish this easily and satisfactorily by pushing a sample holder hydraulically against a horizontal cross-bar anchored firmly by four widely spaced tie lines (Fig. 1). To reduce compaction of the soil inside the cylinder due to the friction between the cylinder wall and the soil, the diameter of the cylinder should be kept large and/or a sampling tool with a moving sleeve should be used [26]. Driving cylinders into the ground by repeated striking with a hammer should not be tolerated for quantitative work, not even for short samples, because of the lateral forces which are likely to be applied. A compromise between a hammer and a hydraulic press is a heavy metal cylinder that is dropped repeatedly onto a sampleholder while being constrained by a steady vertical rod attached to the sampleholder. For measurements of hydraulic conductivity of packed soil columns, it is essential that the packing is done systematically to attain the best possible reproducibility and uniformity. At the moment this appears to be more an art than a science.

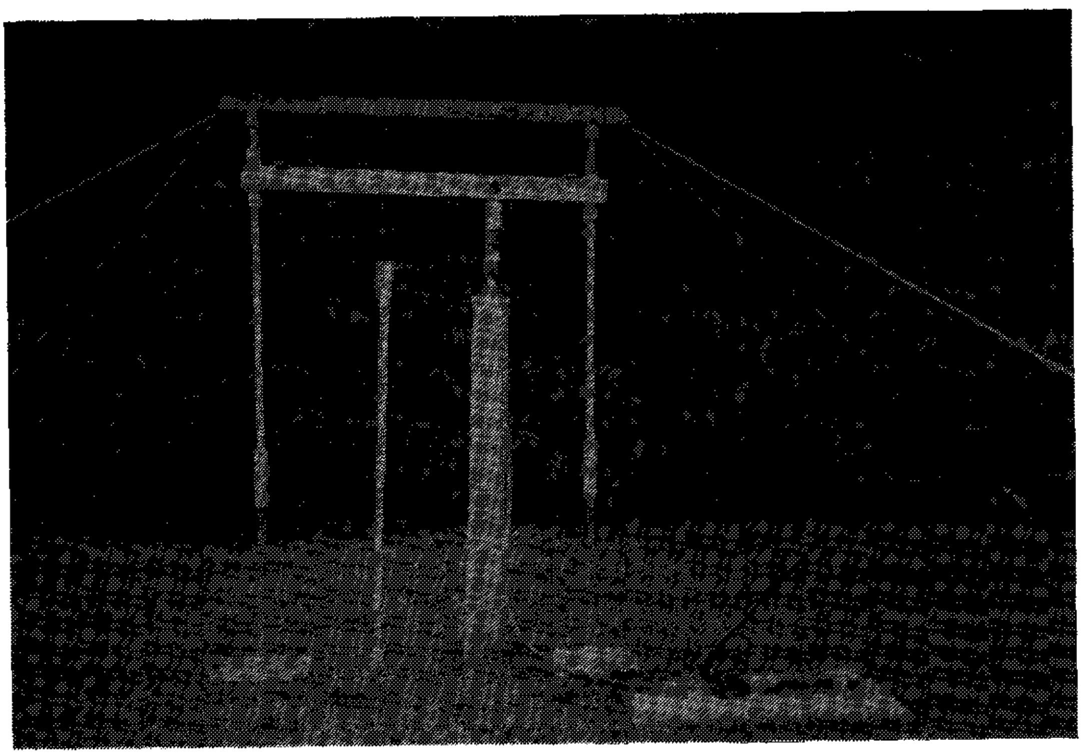

Fig. 1. Hydraulic apparatus for obtaining short (left) and long (right) "undisturbed" soil columns. The apparatus is stabilised by a cross-bar and four widely anchored tie lines.

C. Sample Representativeness

Other important aspects of soil sampling are the size and number of samples required to be representative in view of soil heterogeneity and spatial variability. The development and size of the natural structural units (peds) dictate the size of the sample needed for a particular measurement. If a soil property were measured repeatedly on soil samples of increasing size, the variance of the results normally would decrease until it reached a constant value, the variance of the method alone. The smallest sample for which a constant variance of a specific soil property is obtained is called the Representative Elementary Volume (REV) for that property [27]. Assuming that a soil sample should contain at least 20 peds to be representative, Verlinden and Bouma [28] estimated REV's for various combinations of texture and structure. These varied from the commonly used 50-mm-diameter (100 cm^3) samples to characterize the hydraulic properties of field soils with little structure, to 10^5 cm^3 soil samples for heavy clays with very large peds or soils with strongly developed layering. The desirable length of (homogeneous) soil samples depends on the particular measurement method that is used.

Considering the number of soil samples needed, Warrick and Nielsen [29] list the unsaturated hydraulic conductivity under the category of soil properties with the highest coefficient of variation. They reported that about 1300 independent samples from a normally distributed population (field) were needed to estimate mean hydraulic conductivity values with less than a 10% error at 0.05 significance level. The recently developed theory of regionalised variables or geostatistics [30] provides insight into the minimum number and spatial distribution of soil samples required to obtain results with a certain accuracy and probability. Of course, the same applies to the required number and locations of sites for in situ measurements.

V. STANDARD METHOD

A major problem associated with the determination of soil hydraulic transport properties is that there are no unchanging, uniform soils or other porous materials with constant, known transport properties which can serve as standard reference materials with which to establish the absolute accuracy of any method. It is impossible to pack granular material absolutely reproducibly and consolidated porous materials (e.g. sandstone) are not suitable for most of the methods used on soil materials. Also, repeated wetting or drying of a soil sample to the same overall water content does not lead to the same water content distribution and hydraulic conductivity. Lacking these possibilities, hydraulic transport properties are almost always presented without any indication of their accuracy. Only the method used to determine them is described and sometimes, for good measure, a comparison between the results of two methods is given. Agreement between two methods is still not a guarantee that both are correct. Often the results of two methods are said to correspond well, when in fact they differ by as much as an order of magnitude over part of the range. There is no way to decide which is the most accurate. The only recourse left is to evaluate the available methods on their potential accuracy based on: theoretical exactness, inherent accuracy of the required measurements, possibility of experimentally attaining the theoretically required initial and boundary conditions, error propagation in the required calculations, etc. In this way, instead of a standard material with accurately known properties, a "standard reference method" would be chosen.

In searching for such a standard method, it should be realised that hydraulic conductivity is theoretically the most correct parameter for characterizing water transport in soils, since it is directly associated with the driving force for the movement of water, the hydraulic potential gradient. Moreover, it can be measured more directly and probably more accurately than any of the other parameters characterising water transport, especially when measured during steady state conditions. From this it follows that steady state measurements of hydraulic conductivity in vertical soil columns between two porous plates, in which purely gravitational flow (no pressure head gradient) is established, approach most closely to the requirements for a "standard method" (Fig. 2). Since the pressure head is everywhere the same, the water content and thus the hydraulic conductivity are uniform throughout the column. Therefore, there is no question (error) as to which water content and/or pressure head the obtained hydraulic conductivity should be associated with. Because the contact resistances between the soil column and the porous plates are often too large and unpredictable to rely on measurement of the externally applied hydraulic gradient, the hydraulic head gradient should be measured within the soil column with accurate tensiometer equipment. To assign the status "standard" to this method, the influx and outflux should both be measured until they have become equal. These fluxes can be measured accurately down to very low values by observing the movement of air bubbles in thin glass capillaries.

Once this experimental set-up is assembled, it can be used at various pressure heads. The range of pressure heads is theoretically limited to that of tensiometers, approximately 0 to -8.5 m water. Another limitation of the two-plate method is the time needed to reach a steady

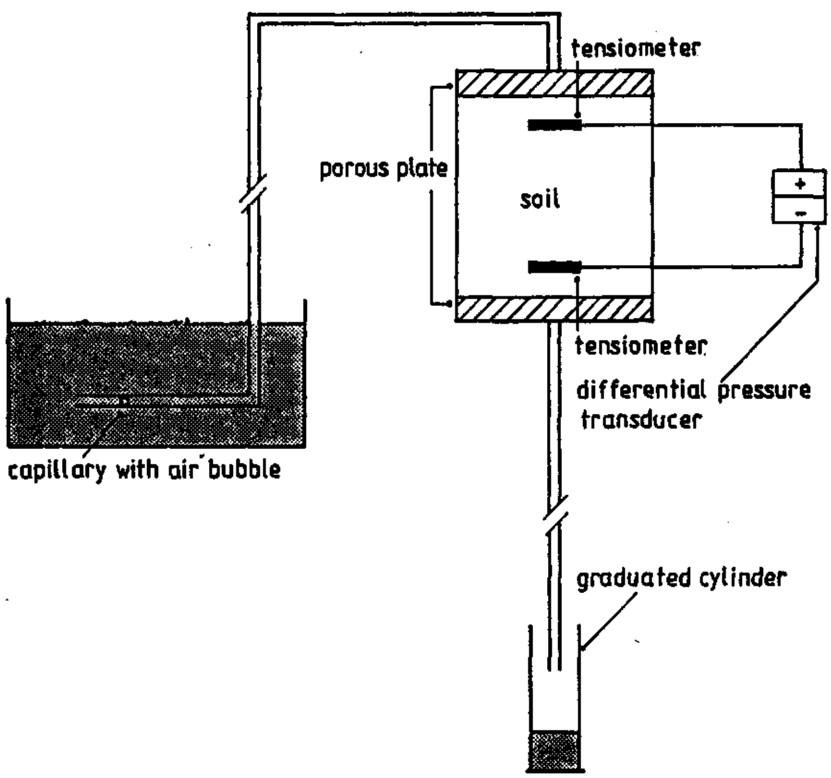

Fig. 2. Diagram of "standard reference method" (head - head).

state. This can become prohibitively large, either due to practical considerations or because long term effects (e.g. microbial activity and loss of water through tubing walls) reduce the overall accuracy to an unacceptable level. Therefore, the practical range is probably to not much below a pressure head of -3.0 m. This is sufficient for characterisation of water transport over relatively large distances. However, for analyses of water transport to plant roots, and of evaporation near the soil surface, etc., hydraulic conductivities for much lower pressure heads and water contents are needed. These can be determined only with other, usually indirect methods. Selection of a standard method for this higher tension range seems as yet not possible. For field measurements, steady infiltration over a large surface area (with tensiometer measurements in the center) with a sprinkling infiltrometer approaches most closely to the requirements for a "standard method". Further comments about these methods follow in the next section.

VI. STEADY STATE LABORATORY METHODS

A. Head-controlled (Head - Head)

This method, featured in most soil physics textbooks, involves steady state measurements on a soil column in which the pressure head is controlled at both ends (usually by two porous plates) such that it is uniform over the entire length (Fig. 2). Principles, apparatus, procedures, required calculations and general comments are given in great detail by Klute and Dirksen [3]. In the previous section the method has been identified as most suitable for use as a "standard method". This is reflected in the maximum scores in Table 1 for theoretical basis (C), control of initial and boundary conditions (D), and error propagation in data analysis (F). Tensiometric measurements generally are tedious and error-prone, but can be very accurate when done carefully with good equipment (this is indicated by the additional score within parentheses in column E). Also, the ease with which fluxes can be measured accurately decreases with their magnitude. The installation of the tensiometers and the porous plates in good contact with the soil column may take considerable time. The time required to reach steady state at unit hydraulic gradient (i.e. gravitational flow) increases rapidly with decreasing hydraulic conductivity. Therefore, while theoretically the entire tensiometer range can be covered, this method will in practice probably not be used at pressure heads below-2.0 to -3.0 m. If the hydraulic conductivity is to be measured over an extensive range of water contents (warranted when the method is used as a standard to establish the accuracy of another method) the measurements will take much longer than 1 month (parentheses for criteria G and H).

Near saturation, one such measurement takes little time for all but the least permeable soils. For this reason, and the inherent accuracy of the measurements I use this method to obtain the one hydraulic conductivity value (at about $h = -0.1$ m) normally used to correct hydraulic conductivities derived theoretically from other data, e.g. the soil water characteristic (see section X.A). Most often, the saturated hydraulic conductivity is used as such a correction (matching) factor. This is often the worst possible choice. Saturated hydraulic conductivities of different samples of the same soil can vary tremendously due to imperfections in the sampling procedure, worm and root channels, structural cracks and fissures, etc. If present, these large pores are at saturation filled with water and completely dominate water transport through the soil sample, yet they have little if any relation with the properties of the soil matrix from which the hydraulic conductivity function is derived. However, even at small suctions, all these large spaces are empty and the then prevailing hydraulic conductivity is a truer reflection of the soil matrix.

B. Flux-controlled (Flux - Head, Head - Flux, Regulated Evaporation)

Hydraulic conductivities can also be measured at steady state by controlling the flux density rather than the hydraulic head at one end of a vertical soil column [3]. If the water flows towards a water table at the bottom ("flux - head"), the range of pressure heads that can be covered is limited to the height above that water table. The range can be extended by maintaining a controlled suction at the bottom of the soil column, either with a porous plate or another soil column with a water table at some depth. Steady state can also be attained when the water flows upward from a water table or a water supply at constant negative pressure head and is evaporated at the soil surface at a constant rate ("head - flux"). In this latter case, it is no longer possible to have a measuring zone with uniform pressure head and water content. As the soil becomes drier, the hydraulic gradient will become larger and more difficult to measure accurately. The derived hydraulic conductivity then will be for some kind of average of a range of water contents and the correct water content to which it should be assigned will be uncertain.

A slightly different experimental arrangement was used by Gardner and Miklich [31]. Their soil column was closed at one end, which makes it theoretically impossible ever to reach a steady state. Nevertheless, they claimed that various constant fluxes could be attained by regulating evaporation from the other end of the column according to the size and number of perforations in a cover plate ("regulated evaporation"). This would seem to require a lot of manipulation. The rates of water loss were determined by weighing the entire column. The hydraulic gradient was measured with two tensiometers and for each evaporation rate, k and θ were assumed constant between the tensiometers. The hydraulic conductivity is then approximated by

$$k = (x_1^2 - x_2^2) q / 2 L (h_1 - h_2) \tag{8}$$

where x_1, x_2 are the positions of the tensiometers and L is the length of the soil column. These rather severe assumptions limit the applicability of the method and the method has not been frequently used.

C. Long Column Infiltration

When a constant water flux density of water is applied to a long dry vertical soil column, the flow system can reach a "quasi" steady state [32, 33]. True steady state, of course, will never be attained because, although the potentials on both ends of the flow system are constant, the distance between these ends keeps increasing with time. As a result, the pressure head gradient keeps diminishing with time. Eventually, it may become small enough to be negligible with respect to the constant, unit gravitational potential gradient. Then, a "quasi" steady state is attained. If the soil column is sufficiently long for a zone to develop at the top of the column in which the hydraulic gradient can be assumed unity, the hydraulic conductivity there is then equal to the externally imposed known flux density. Thus, tensiometers are not needed and if the hydraulic conductivities are assigned to measured water contents, the pressure head range of the method can theoretically extend beyond the tensiometer range. Whilst this method does not present problems with contact resistances between soil and porous plates, it does require a device to deliver small fluxes uniformly over the soil surface [see e.g. 34, 35]).

D. Matric Flux Potential

The configuration of a controlled evaporative flux from a short soil column in which the pressure head at the other end is controlled (section VI.B) was used by Ten Berge et al. [13] in a steady state method for measuring the matric flux potential as function of water content. They assumed that the matric flux potential function has the form

$$\Phi[\theta] = - A / (x + B), \qquad x = 1 - (\theta / \theta_0) \tag{9}$$

where A is a scale factor (m^2/s) and B is a dimensionless shape factor, both typical for a given soil, and θ_0 is a reference water content, experimentally controlled at the bottom of the soil column. Whereas Ten Berge etal. use the earlier [36] proposed diffusivity function

$$D[\theta] = a (b - \theta)^{-2} \tag{10}$$

where a and b are constants, the method can be used with any set of two-parameter functions of $\Phi[\theta]$ and $D[\theta]$.
After a small soil column is brought to a uniform water content (pressure head) and weighed, it is exposed to artificially enhanced evaporation at the top, while the bottom is kept at the original condition with a Mariotte-type water supply. When the flow process has reached steady state, the flux density is measured, as well as the wet and oven dry weight of the soil column. From these simple, accurate experimental data the parameters A and B, and thus $\Phi[\theta]$ and $D[\theta]$, can be evaluated by assuming that gravity can be neglected. In this case the matric flux potential at steady state decreases linearly with height so that this method does not suffer from any ambiguity (generally associated with upward flow) in the assignment of appropiate values of water content and pressure head to the calculated values of the water transport parameter.

It is better not to start from saturation, but at a small negative pressure head to reduce the influence of gravity and be able to meet the theoretically required upper boundary condition (θ=0). The method is rather slow and covers a limited range of θ and h, but the measurements require little attention while in progress. The major source of errors appears to be that the theoretically prescribed initial and boundary conditions are hard to obtain experimentally. Furthermore, the theoretical basis involves a number of assumptions. However, direct measurement of $\Phi[\theta]$ is likely to be more accurate than methods involving separate measurements of $D[\theta]$ and $h[\theta]$ for flow processes involving steep gradients, thin, brittle soil layers, etc. For an analysis of the propagation of errors, see Ten Berge, et al. [13].

VII. STEADY STATE FIELD METHODS

A. Sprinkling Infiltrometer

Analogous to the long column measurements in the laboratory (section VI.C), hydraulic conductivies can be measured directly in the field under quasi steady state conditions with a sprinkling infiltrometer [4, 37]. It is the closest counterpart to the two-plate laboratory method as a "standard reference method" for the field. In that application it is warranted to use the very elaborate sprinkling equipment, which normally must be attended whenever it is in operation. This may extend over days or even weeks, depending on the range of water contents to be covered. This range is technically limited by the ability to reduce the sprinkling rate while retaining uniformity. This can best be done by intercepting an increasing proportion of the artificial rain, rather than reducing the discharge from a nozzle [35, 38, 39]. Green etal. [4] give as a practical lower limit for the flux density 1 mm/h. To prevent hysteresis, the flux density of the applied water should be increased monotonically with time. Because soil profiles are frequently inhomogeneous and because the possibility of lateral flow, the hydraulic gradient cannot be assumed to be unity and it should be measured when a high accuracy is required. Sprinkling infiltrometers are used frequently for soil erodability studies. Then, the impact energy of the water drops emitted by the sprinkling infiltrometer should be as equal to that of natural rain drops as possible [40], since changes of the physical soil properties due to structural breakdown of the soil (e.g. crust formation) have a great effect on the erosion process [41, 42]. In contrast, for hydraulic conductivity measurements the soil surface generally should be protected against crust formation as much as possible, e.g. by covering the soil surface with straw.

Field measurements of hydraulic conductivity with a sprinkling infiltrometer may take a long time, during which large temperature variations may occur. Temperature changes and gradients may have a significant influence on the water transport process, especially for small water flux densities and/or hydraulic head gradients near the soil surface. Therefore, it is good practice for all field measurements to minimize temperature changes as much as possible, for example by shielding the soil surface from direct sunlight.

B. Isolated Soil Column

Analoguous to the long column method, a soil column can be isolated in situ by carefully excavating the surrounding soil. Although not strictly necessary for unsaturated conditions, usually a plaster of Paris jacket is cast around the soil column and cylinder assembly for protection, transportation and/or subsequent saturated conductivity measurements. Use of such a truly undisturbed soil column is especially suitable for soils with a well developed structure, since large scale "undisturbed" samples which are easily damaged during transport would otherwise be required.

Usually, the pressure head, rather than the flux, has been controlled, for example with a crust [43, 44]. After smoothing the soil column surface at the desired depth, a close fitting cylinder is pushed into the top of the column. A crust of uniform thickness and composition (usually a mixture of hydraulic cement and sand) is applied inside the cylinder. After the crust is cured, normally 24 hours, the cylinder is sealed off and water is applied to the soil column via the crust at constant head with a Mariotte device. Supposedly, the crust soon causes the flow density to attain steady state at unit hydraulic gradient, after which time the hydraulic conductivity is equal to the prevailing flux density. Measurement of the pressure head in the soil just below the crust with a single tensiometer provides the pressure head corresponding to this value of hydraulic conductivity. However, because the assumption of unit hydraulic gradient is often invalid, the hydraulic gradient should be measured with at least two tensiometers. By using different values of the controlled pressure head and/or crust resistance, a number of points on the hydraulic conductivity function can thus be obtained. In doing this, one should proceed from dry to progressively wetter conditions (by replacing higher resistent crusts with progressively less resistent ones) since the wetter wetting fronts will quickly overtake each other. Letting the soil first dry before applying a smaller flux density takes much time and introduces hysteresis into the measurements. The minimum pressure head that can be attained with crusts appears to be, practically, not much lower than -50 cm.

In comparison with ponding infiltration, the claim that crusts enhance the attainment of a steady state is correct. The hydraulic head loss across the relatively less permeable crust decreases the pressure head difference between either end of the extending zone of wetted soil. Thus the pressure head gradient will become negligible with respect to the constant, unit gravitational potential gradient more quickly with a crust. I suspect, however, that often the final measurements with the crust method are made before a "quasi" steady state is reached. The crust does not add to the speed of attaining steady state in comparison to the application of a non-saturating, constant water flux to a soil column (the previous method). On the contrary, it may well be slower and it also introduces other experimental problems. Crust resistances have proved to be quite unpredictable, often non-uniform and unstable in time. Making and replacing good crusts is tedious work and curing of the crusts takes time. It may also add chemicals to the soil solution which alter the hydraulic conductivity. I would advocate, therefore, that the crust method in its present form no longer be used.

The isolation of a soil column is an attractive feature that can be retained, but the water should be applied uniformly over the soil surface at easily changed, constant rates which can be verified. We have been exploring application of water from a reservoir with hypodermic needles suspended just above the isolated soil column (Fig. 3). When the water is applied with a pulsating pump, each needle can be made to release just one water drop per pulse down to fairly low average flux densities of about 2 mm/day. The uniformity of water supply can be determined easily by placing a rack of reaction tubes in the same pattern under the needles. Additional study is needed to see whether flux density can be reduced further by decreasing the pulse frequency and/or the needle density without unduly effecting the flow process by the inhomogeneous water application. When electricity is not available, a constant head (Mariotte) water supply can be used, but the water application becomes non-uniform at flux densities less than about 10 cm/day. This variant of the isolated soil column method appears to be a very attractive, much simplified version of the sprinkling infiltrometer.

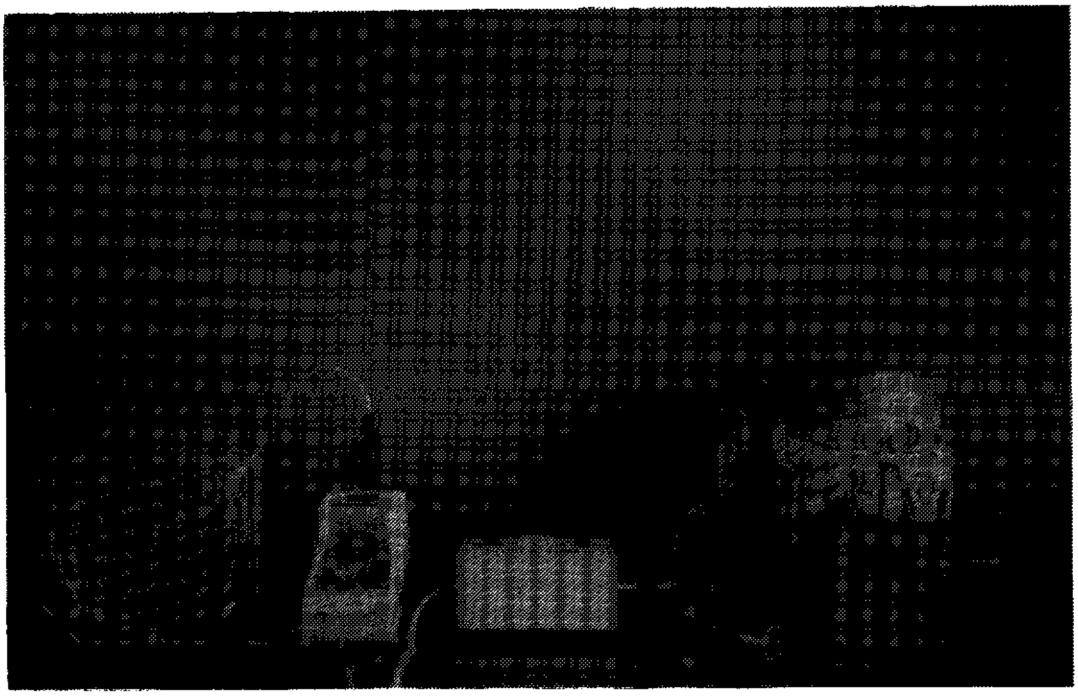

Fig. 3. Isolated soil column method. Water supply via hypodermic needles is regulated by stroke and frequency of pulsating pump. Tensiometers are hydraulically switched to pressure transducer with digital volt meter.

C. Spherical Cavity

The previous discussions make it clear that in one-dimensional flow, steady state can only be achieved when there are two controlled, steady boundaries, either potentials or flux densities. Both features are inconvenient under field conditions, particularly when measurements must be repeated many times. It is not too difficult to force the flow to be one-dimensional by isolating a soil column, either as practiced with the crust method or by making vertical trenches, covering the vertical walls with plastic sheet and refilling the trenches with soil. However, it requires a major experimental effort to impose a steady boundary condition at the bottom of a flow system in the field. The practical solution is usually to perform measurements in a deep uniform soil profile in the center of a larger area wetted by a sprinkling infiltrometer, allowing the "quasi" steady state of a constant-shape wetting front moving downward at constant velocity. This is then due to the action of gravity. Without gravity (i.e. in a horizontal direction or when the pressure head gradient is sufficiently large for the effect of gravity to be neglected), the wetting front advances according to $t^{1/2}$, as long as water is applied at the soil surface. This process is often referred to as adsorption.

In contrast, three-dimensional infiltration from a point source reaches a "large-time steady state" with and without the influence of gravity [19]. The influence of gravity is much smaller in three-dimensional than in one- or two-dimensional flow. Without gravity, three-dimensional infiltration from a point source is spherically symmetric. Raats and Gardner [11] showed that the hydraulic conductivity can be derived from a series of such steady flows. This presents a very attractive set of conditions for measuring hydraulic conductivity, especially in situ because: 1) only one controlled boundary is required, 2) the influence of gravity, which must be neglected, is especially small, 3) steady state measurements are inherently accurate. For these reasons, I have explored the possibilities of this "spherical cavity" method and analysed the influence of gravity [45]. Water is supplied to the soil (which needs to be initially at uniform pressure head) through the porous walls of a spherical cavity maintained at a constant pressure head until both the flux, F, and the pressure head, h_a, at any radial distance $r = a$ from the center of the spherical cavity, have become constant. This is repeated for progressively larger (less negative) controlled pressure heads in the cavity. Hydraulic conductivity can then be calculated according to

$$k[h_a] = (dF / dh_a) / a \tag{11}$$

which is simply the slope of the graphs in Fig. 4 at any desired pressure head, divided by the radial distance of the particular measuring point. In this way hydraulic conductivities down to h = - 700 cm were obtained in about two weeks, with each tensiometer and the cavity yielding its own result. This overlap provides an internal check. Note that the pressure head range can be expanded downward easily by increasing the radial distance of the measuring point. Of course, the time required to reach steady state increases then also. It is possible to use the regulated pressure head in the cavity as the only "tensiometer" data. This reduces the experimental operations to a minimum. The resistance between the water supply and the soil (porous walls and soil

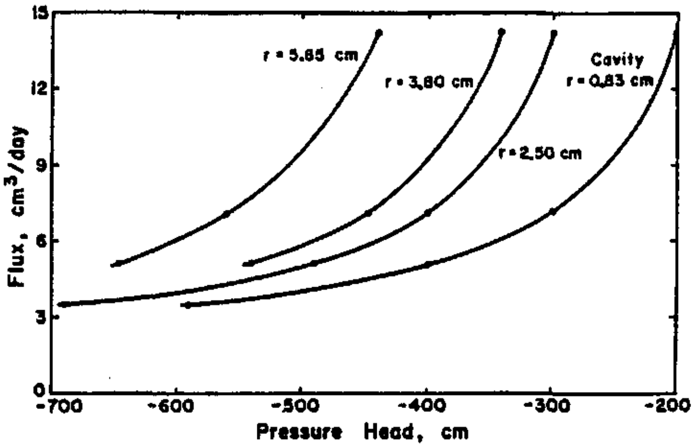

Fig. 4. Steady fluxes from a spherical cavity versus steady pressure heads in cavity and in three tensiometers at indicated radial distances. (From Ref. 45)

ceramic interface) must then be negligible. The effect of gravity is minimized when tensiometers, if used, are placed directly below the cavity. The method has only been demonstrated in the laboratory, with some exploratory measurements in the field. Because of its very attractive features, especially as an in situ method, the method is worth of further investigation. If tensiometer measurements can be omitted, placement of the spherical cavity without undue contact resistance with and disturbance of the soil presents the only great experimental challenge.

D. Ponded Disk

After a complicated mathematical analysis, wherein he assumed

$$k = k_s \exp \alpha h \tag{12}$$

where k_s is saturated hydraulic conductivity and α is a constant characterising different soils, Wooding [46] obtained a simple, linear equation for the steady infiltration of water from a shallow circular pond

$$q = \alpha\Phi_s + 4\Phi_s / \pi r \tag{13}$$

or

$$q = k_s + 4k_s / \pi\alpha r \tag{14}$$

where Φ_s is the matric flux potential. The first term is the contribution of gravity, the second that of the matric potential gradient. Scotter et al. [47] used this result to determine k_s, Φ_s, and S, the latter by assuming that soils have a delta-function diffusivity. When (average) steady infiltration flux densities, q, are measured with shallow rings of two different radii, r, then

$$k_s = (q_1 r_1 - q_2 r_2) / (r_1 - r_2) \tag{15}$$

$$\Phi_s = (\pi/4) (q_1 - q_2) (1/r_1 - 1/r_2) \tag{16}$$

$$S = [2 \Phi_s (\theta_s - \theta_n)]^{1/2} \tag{17}$$

From the same results the parameter α in the exponential hydraulic conductivity function can also be derived

$$\alpha = [4 (q_1 r_1 - q_2 r_2)] / [\pi (r_1 r_2) (q_1 - q_2)] \tag{18}$$

Strictly speaking, these are saturated measurements and belong in the previous chapter. However, because of the pre-assumed functional relationships, they yield hydraulic properties of unsaturated soil. It seems appropiate, therefore, to review a few details of the experimental aspects. The measurements are clearly simple enough to be carried out in great number. Apart from the flux measurements, only volumetric water contents before and immediately after each infiltration run must be determined.

Scotter et al. presented equations for the standard deviations of k_s and S, whether they are normally or log-normally distributed. They performed sufficient measurements (from 4 to 25 per ring) to investigate the spatial variability of k_s and S. The rings, with radii ranging from 25 to 204 mm ($r_1 > 2r_2$), were gently pushed into the soil only about 10 mm, keeping disturbance to a minimum and making the method suitable for a wide range of soils. The ponding depth, also about 10 mm, was maintained with a Mariotte device or by hand. Measurements were continued for an hour after steady state appeared to have been reached, which occurred after elapsed time periods ranging from 5 to 100 minutes (in soils ranging from sandy loam to silt loam). However, Scotter et al. warned that this time may be much longer and cited an example where it took 14 hours. They also suggest plotting q versus (log t) rather than t to judge whether steady state has been reached.

E. Dripper

Shani et al. [48] used the same theoretical basis as the previous method for estimating the hydraulic conductivity function. Instead of confining the saturated zone at the soil surface with rings and waiting until the flux has become steady, they used commercially available drippers, used for drip irrigation, to apply water at different steady discharge rates and waited until the diameter of the ponded area at the soil surface had become steady. They stated that this usually occurred within 15 minutes. They dubbed this the "dripper" method. Also, rather than substituting average values of q in Eq. (14), they estimated first k_s from the intercept of a linear regression of q versus 1/r and then determined α from the slope of the linear regression equation, b, according to

$$\alpha = 4k_s \;/\; b\pi \tag{19}$$

These saturated measurements yield unsaturated results only due to the pre-assumed functional relationships. Therefore, the results can not be better than the degree to which these relationships hold. It should also be realised that these functions are based on measurements in the wet range. They can easily be extrapolated to lower pressure heads, but there is no guarantee that this is valid.

Shani et al. [48] used the same data also to determine the parameters of the Brooks and Corey [49] relationship for hydraulic conductivity

$$k = k_s \;(\; h_w \;/\; h \;)^{\mu} \tag{20}$$

Because of the inter-relationship between the Brooks and Corey equations, this also yields the soil water characteristic. Equation (20) contains two soil parameters: μ, which is related to a pore size distribution index, and the air-entry or bubbling head, h_w. Both can be determined from the dripper measurements if, again, the sorptivity is also measured. Shani et al. did this by measuring the horizontal wetting front advance from the steady ponded zone perimeter at the soil surface as a function of time. They checked their results by, among others, measuring the air-entry head directly [50] but this is not unambiguous, especially in structured soils. The determinations of the pore size distribution index and residual saturation, required for the Brooks and Corey equations, are also not always straightforward. Brooks, Corey, and their co-workers invariably tested these equations with the hydrocarbon fluid "Soltrol", which has altogether different soil wetting properties than water. There is, therefore, some doubt whether these equations are valid for soil - water systems. Van Schaik [51] found large internal discrepancies, even for studies which have been claimed to yield the best results for the Brooks and Corey equations. For these reasons, I would caution against the use of these equations.

VIII. TRANSIENT LABORATORY METHODS

A. Instantaneous Profile

In contrast to the steady state methods, most transient laboratory methods yield in the first place hydraulic diffusivities. $k[\theta]$ must then be derived from $D[\theta]$ with the soil water characteristic (see section II.B). The one major exception is the instantaneous profile method. In its many variants it is probably the most used method to determine non-destructively the hydraulic conductivity of laboratory columns in which other water transport processes are studied for which $k[\theta]$ must be known. Often, quite sophisticated equipment, such as automated gamma attenuation scanners and multiple tensiometer apparatus [52], is already available which allow more complete and/or accurate determination of $k[\theta]$ than is normally the case. This is reflected in the scores for the various criteria for this method as a laboratory method, in comparison with the scores as a field method. This method is especially suited to be used in situ; it is discussed in more detail in the next section.

B. Pressure Plate Outflow

Gardner [53] proposed the pressure-plate outflow method. A soil sample at hydraulic equilibrium on a porous plate is subjected to a step decrease in the pressure head in the porous plate (e.g. a hanging water column) or a step increase in the air pressure. The resulting outflow of water is measured with time. The step decrease/increase must be small enough that the hydraulic conductivity can be assumed constant and that the water content is a linear function of pressure head. The experimental water outflow as function of time is matched with a theoretical solution, yielding after many approximations

$$\ln (Q_0 - Q) = \ln (8 Q_0 / \pi^2) - (\pi/2L)^2 Dt \qquad (21)$$

where Q is the cumulative outflow at time t, Q_0 is the total outflow, and L is the length of the soil sample. The diffusivity for the mean pressure head can be derived from the slope of a plot of $\ln (Q_0 - Q)$ versus t. This is repeated for other step increases in pressure, which must only be initiated after a new state of hydraulic equilibrium has first been reached. The pressure increments must be small enough for the assumptions to be valid, but large enough to allow accurate measurement of water outflow, while the more steps, the more time it takes to cover the desired range of water content. This method was initially widely used, but generally failed to yield satisfactory results. Much effort was spent to improve it, especially with respect to the correction for the resistance of the porous plate or membrane, without much success.

C. One-step Outflow

Doering [54] proposed the one-step variant of the previous method, which is much faster and not very sensitive to the resistance of the plate or membrane. If uniform water content in the soil column is assumed at every instant, diffusivities can be calculated from instantaneous rates of outflow and average water content

$$D[\theta] = - 4 L^2 / [\pi^2 (\theta - \theta_f)] . d\theta / dt \qquad (22)$$

where L is the length of the soil sample, θ is the average water content when the outlow rate is $d\theta/dt$, and θ_f is the final water content. These can be determined by measuring the cumulative outflow and the final weight. Doering found the results as reliable as those obtained with the original version (VIII.B) and there were large time savings. Gupta et al. [55] showed that the analysis of one-step outflow data according to Gardner [56] and used by Doering can be in error by a factor 3. They improved the analysis by first estimating a weighted mean diffusivity. This does not require the assumption of a constant diffusivity over the pressure increment, nor over the length of the soil sample and also reduces the effect of membrane impedance. Passioura [57] obtained about the same improvement in accuracy with a much less complicated calculation procedure (with detailed stepwise instructions) by assuming that the rate of change of water content at any time is uniform throughout the entire soil sample. He also estimated that a 60-mm long soil sample will take about 5 weeks to run and a 30-mm sample about 1 week. Measurements have been automated recently for up to 16 samples [58]. The one-step outflow method is attractive for its experimental simplicity; the theoretical analysis of the data remains its weakest point. Since this limitation does not apply to the simulation of the flow process, it is not surprising that recently the same measurements were selected as basis for the parameter optimization approach [section XI].

D. Boltzmann Transform

There are 3 variants of the transient, so-called Boltzmann transform methods. The theoretical framework on which these methods are based is well known and can be found in soil physics textbooks [10, 59]. By neglecting gravity (e.g. using horizontal columns), the flow equation can be expressed in the diffusivity form of Eq.(3). For a step-function increase/decrease of the water content at the adsorption/desorption interface of an effectively semi-infinite uniform soil column, this partial differential equation can be transformed into an ordinary differential equation using the Boltzmann variable $\tau = x/t^{1/2}$, where x is the distance from the sample surface and t is time. Integration of this equation for the also transformed initial and boundary conditions yields the diffusivity as

$$D[\theta'] = 1 / 2 \; . \; (d\tau / d\theta)_{\theta'} \int_{\theta'}^{\theta_1} \tau[\theta] \, d\theta \tag{23}$$

where θ_1 is the final water content at the adsorption/desorption interface, θ' is the water content at which D is evaluated, and θ is the water content as function of x and t. Thus the diffusivity at any water content is equal to half the product of the slope and area indicated in Fig. 5. The function $\tau\,[\theta]$ can be determined experimentally in two ways: by measuring either the water content distribution in a soil column at a fixed time [60] or the change of water content with time at a fixed position [61]. The first is often done gravimetrically, the latter needs to be done non-destructively with specialised equipment, e.g. gamma attenuation, capacitance sensors. Gravimetric measurements must be done very quickly to minimize redistribution and evaporation of water during sampling. The main drawback of the fixed-time method is the sensitivity of the calculated diffusivities to irregularities in the bulk density and water content in the soil column and the consequent propagation of errors from errors in the water contents. At first thought, the fixed-position method would seem to eliminate most of these problems. However, indirect, non-destructive water content measurements are inherently less accurate and the propagation of errors is therefore similar in both cases. A comparative study of the two variants [62] yielded similar errors.

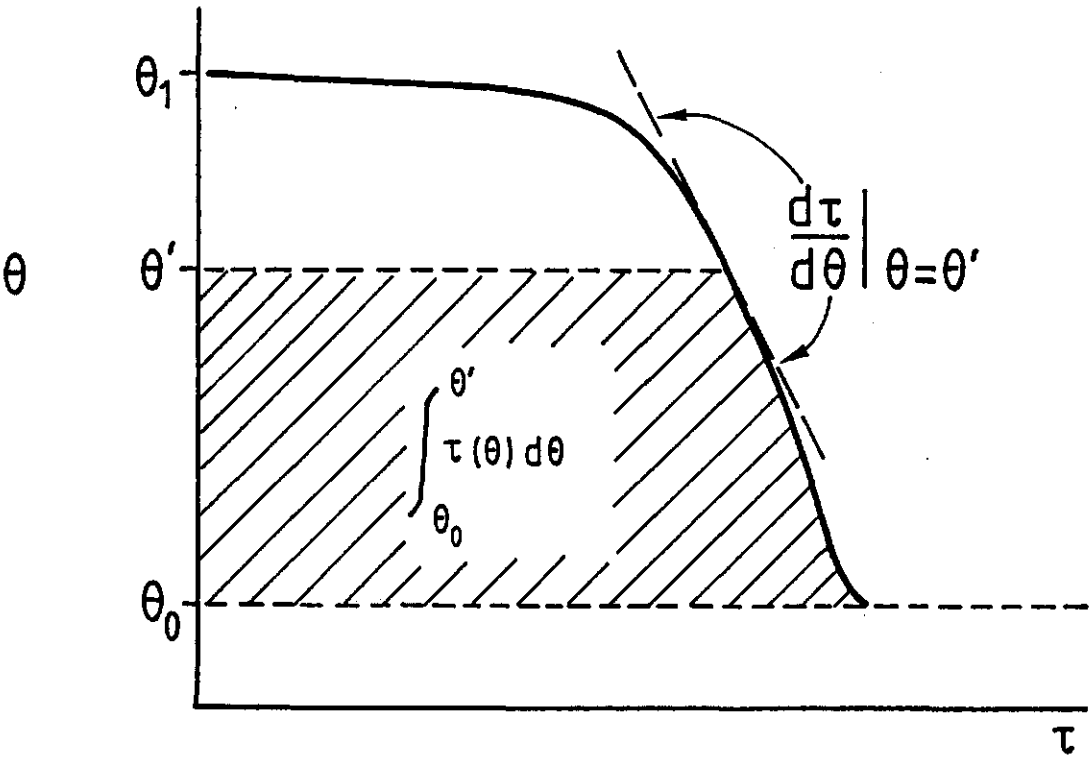

Fig. 5. Graphical solution of Boltzmann transform Equation (23).

Derivation of a $D[\theta]$ function from experimental $\tau[\theta]$ data according to Eq. (23) involves differentiating experimental data with scatter, which is inherently inaccurate and yields poor results, especially near saturation where the water content profile is quite flat [63, 64]. Clothier et al. [64] showed that it is much better to find a value for a parameter p by fitting the experimental $\tau[\theta]$ data to the function

$$\tau[\theta] = \epsilon\,(1 - \Theta)^p\,, \qquad p > 0 \tag{24}$$

where ϵ is a parameter which can be derived from p and the sorptivity. Θ is the dimensionless soil water content

$$\Theta = (\theta - \theta_0) / (\theta_1 - \theta_0) \tag{25}$$

where θ_1 is the final water content at the adsorption/desorption interface and θ_0 is the initial water content. The corresponding equation for the diffusivity is then

$$D[\theta] = p(p+1)\ S^2\ [(1-\theta)^{p-1} - (1-\theta)^{2p}]/[2(\theta_1-\theta_0)^2] \tag{26}$$

This analysis of the experimental data ensures correct integral properties of the obtained $D[\theta]$ function, because it is fitted to the primary data set $\tau[\theta]$ and the measured value of the sorptivity. Moreover, it never leads to physically nonsensical $D[\theta]$ functions which decrease with increasing θ, as least squares fitting of $\tau[\theta]$ can do. Instead, it yields S-shaped diffusivity curves with infinite diffusivity at saturation (Fig. 6), as observed for many soils [65]. More details on this recently proposed improved data analysis can be found in the original publication [64].

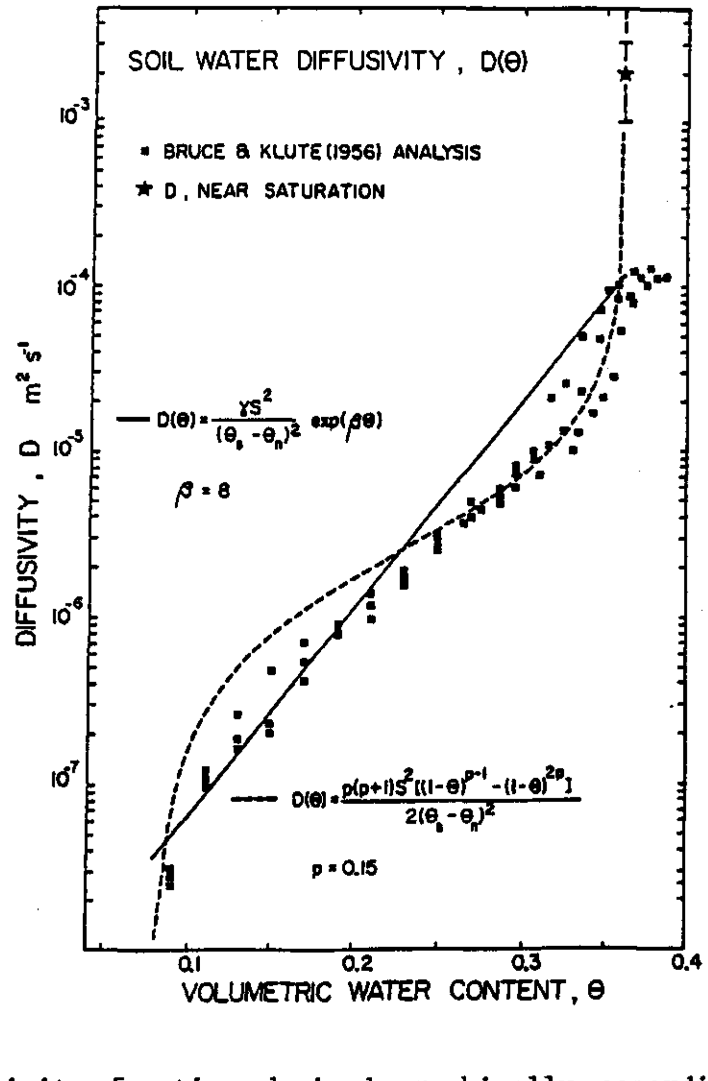

Fig. 6. Diffusivity function derived graphically according to Fig. 5 and derived from fit to Eq. (26), for p = 0.15, and diffusivity measured near saturation. (From Ref. 64)

E. Hot Air

A third variant of the Boltzmann transform method was reported by Arya et al. [66]. As the "hot air" method, this variant has become quite popular in some areas, undoubtedly due to the simplicity and speed of the required measurements, and the large range of θ over which $D[\theta]$ values are obtained. It is the drying counterpart of the Bruce and Klute variant. It not only has all the disadvantages of this variant, but also many others. Whereas the required boundary condition of a step-function change in potential (water content) can be attained easily in the case of wetting, a drying step-function is experimentally nearly impossible. It is imposed by a stream of hot air directed at the soil surface, while the rest of the soil column (usually 10 cm long and 5 cm diameter) is shielded from it as much as possible. Air temperatures of up to 240 °C have been required for sandy soils. Even then it takes normally a few minutes to dry the soil surface, while the total evaporation period normally lasts from 10 to 15 minutes. Whereas temperatures in excess of 90 °C have been measured in the soil [67] the data can be analysed only by assuming isothermal conditions. The effects of temperature on viscosity, surface tension, etc. and of any water transport due to the thermal gradient are significant, but must be ignored. Because the soil is hot, there is significant loss of water during sampling due to evaporation. Finally, the measurements are usually performed on initially saturated, vertically oriented soil columns. This introduces errors due to gravity during a run and loss of water at the wet end due to compaction during sampling. This can be reduced by equilibrating the soil column at a moderate negative pressure head (around -50 cm).

Without arbitrary manipulation of the water content profile of the sample, the data often yield diffusivities decreasing with water content. This is physical nonsense. To prevent this, computer programs have been devised [68] which keep the analysis within the theoretically acceptable framework, but the results are still based on very dubious experimental measurements. When the method appears to yield useful results, this may be accidental; several sources of errors appear to cancel each other [67]. I feel, therefore, that the hot air method should be abandoned. It may be possible to find a way to impose the boundary condition by using hygroscopic agents, eliminating the temperature effects, but in view of all the other objections this does not seem worth the effort. In this connection, it should be pointed out that it is not necessary to dry the soil instantaneously at the surface; only a constant water content or pressure head must be imposed. This does not need to go beyond the range over which the diffusivity or conductivity function is required.

F. Flux-controlled Sorptivity

The sorptivity method is related to the Boltzmann transform methods in that the same transformation is used in the derivation of the working equation [17, 69]

$$D[\theta_1] = \frac{\pi S^2}{4(\theta_1-\theta_0)^2} \left[\frac{(\theta_1-\theta_0)}{(1+\gamma)\log e} \frac{d}{d\theta_1} (\log S^2[\theta_1,\theta_0=\text{const}]) - \frac{1-\gamma}{1+\gamma} \right] \quad (27)$$

where γ is a constant which can be varied between 0.50 and 0.67 without significant effect [70]. Detailed information on required experimental apparatus and a step by step description of the experimental procedure of the sorptivity method can be found in Klute and Dirksen [3]. Experimentally, the method entails the determination of $S[\theta_1, \theta_0 = \text{constant}]$, the sorptivity as function of the water content at the adsorption interface, θ_1, for constant initial water content, θ_0 (see Eq. (7)). This can be accomplished by means of a series of one-dimensional absorption runs, each yielding one set of (S,θ_1) values. Rather than regulating θ_1 via h_1, each sorptivity is controlled by mechanically controlling the supply of water to the adsorption interface according to the $t^{½}$-relationship of Eq. (7). Then, after each run a single soil sample is required for gravimetric determination of θ_1. This takes only about 10 seconds which virtually eliminates errors due to evaporation and redistribution during sampling. Moreover, near the soil surface θ changes neither with time ("pseudo" steady state) nor with position. With proper functioning of a somewhat complex apparatus, experimental errors are thus limited to a minimum, and thus any propagation of errors in the calculation of $D[\theta]$ according to Eq. (27) is also minimised. The required differentiation is performed algebraically on a polynomial regression of $\log S^2$ in terms of θ_1. Depending on the desired accuracy, a diffusivity function can be obtained from 1 to 3 soil samples of 10 cm length. By first drying these samples the required uniform initial water content is easily guaranteed and a large water content (pressure head) range can be covered. For each run a new dry soil surface must be carefully prepared. The effect of non-uniformity of soil samples on the final results still requires further investigation. The theoretical basis of Eq.(27), although not rigorously exact, appears to be accurate [18, 69, 71]. Although water is applied through porous plates, diffusivities well beyond the "tensiometer" range have been obtained. This is possible, because the individual runs need to be continued for only a few minutes near saturation to a maximum of 1 hour when the final water content is very low. A complete diffusivity function can be determined in 1 day.

During sorptivity measurements in the wetter range, h_1 could be measured by an isolated small tensiometer, slightly protruding in the center of the porous plate, and a pressure transducer which needed virtually no water displacement for a full scale measurement (zero-balance principle). Later tests yielded the best pressure transducer response with tensiometers of only 1.5 mm diameter. Such simultaneous pressure head measurements allow immediate determination of $k[\theta]$ which is convenient because wetting $h[\theta]$ functions are not normally available. The line in Fig. 7 indicated by 'sorptivity method' was in the wetter region obtained with such simultaneous measurements. Only 7 sorptivity runs each lasting from 6 to 12 minutes yielded $k[\theta]$ values for water contents less than $\theta = 0.10$. The results with the instantaneous profile method, obtained on the same packed soil before the samples for the sorptivity measurements were taken, required several weeks and still yielded only $k[\theta]$ values for water contents larger than 0.20. The experimental results presented here and in Dirksen [17] were all obtained with apparatus fabricated in our own machine shop. More versatile apparatus is commercially available [3] as indicated in Table 1 between parentheses.

G. Other Methods

Several other methods have been proposed in the literature, which fall in the category of transient, laboratory methods. Without being exhaustive, and without evaluating them in Table 1, a few of these will be mentioned.

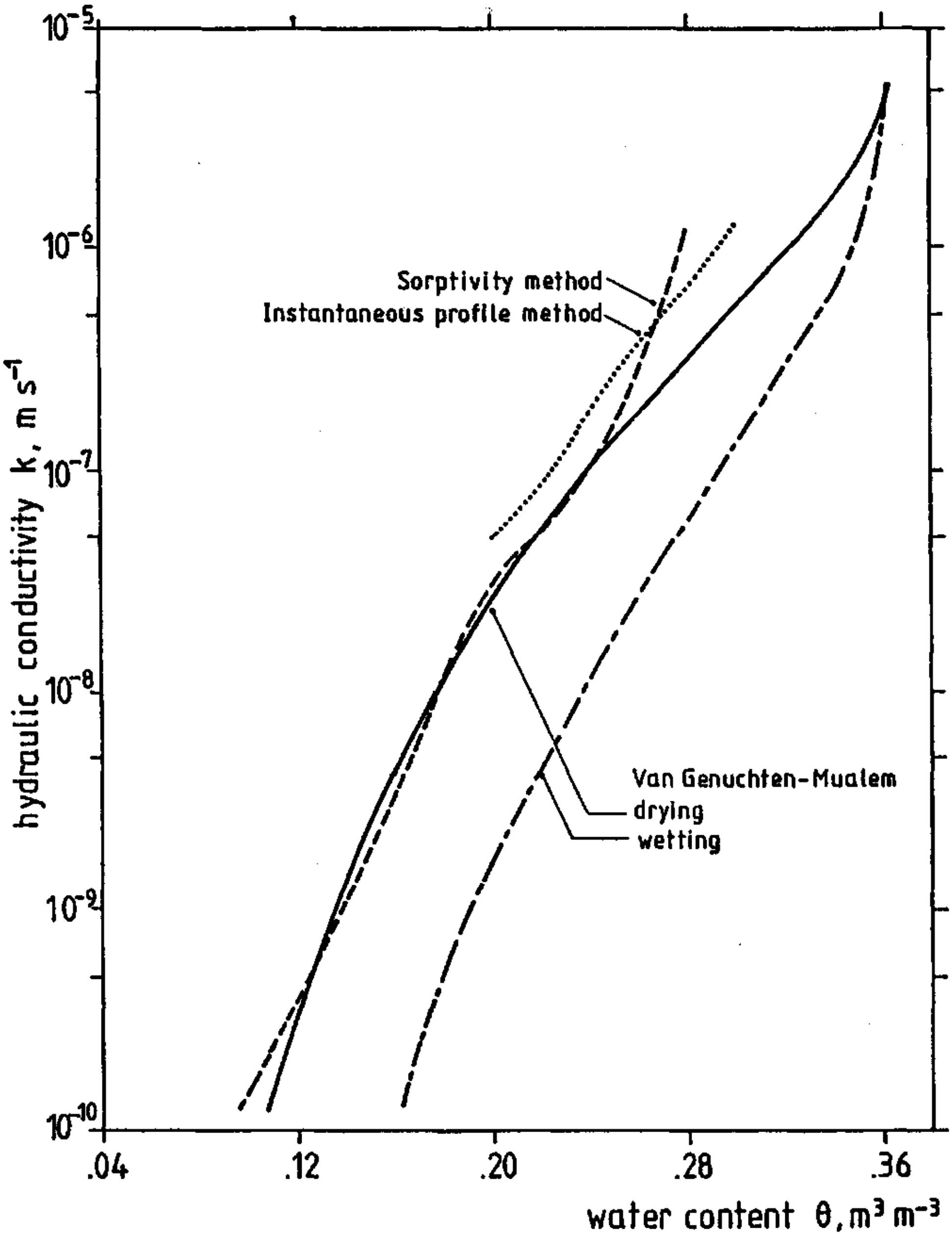

Fig. 7. Hydraulic conductivity functions of Pachappa sandy loam measured with flux-controlled sorptivity method (and simultaneously measured pressure heads) and with instantaneous profile method. The Van Genuchten - Mualem functions (Eq. (39)) are based on the fitted soil water retention characteristics in Fig. 10.

Wind [72] proposed a modified instantaneous profile method to measure simultaneously the water retention characteristic and the hydraulic conductivity of the same soil sample. An initially saturated and homogeneous sample is allowed to evaporate at the top. The total weight and the pressure heads at at least two depths are recorded. From these data he calculated the water retention characteristic with an iterative method. Knowing this, he could determine the flux densities at the bottom (zero), at the top (measured evaporation rate) and in between the depths where the pressure heads were recorded. The calculation is then further the same as for the instantaneous profile method (next section). Boels et al. [73] designed an automatic recording system for these measurements on many soil samples. They also proposed a direct calculation method by approximating the soil water retention characteristic by a polygon. The data of these experiments can also be used for the inverse parameter optimization approach (section XI), allowing a comparison between the two approaches [74]. All this has been improved and automated further to the point that it is now the major method at their institute (Halbertsma, ICW, Wageningen, pers. comm.).

Ahuja and El-Swaify [75] determined the soil hydraulic properties by measuring one-step cumulative inflow or outflow from short soil cores through high-resistant plates at one end and measuring the pressure head at the other end. They obtained good results for pressure heads down to -150 cm. Scotter and Clothier [75] claimed, without referring to the previous authors, that it is better to analyse the results of a series of small pressure head changes than of one large change, because it obviates the difficult task of measuring small flow rates. The accuracy relies mainly on the time delay of the outflow and not on the shape of the outflow curve.

IX. TRANSIENT FIELD METHODS

A. Instantaneous Profile

The relative merits of laboratory and field measurements were discussed in section IV. Especially for layered soils or soils with a well developed structure, the unsaturated hydraulic conductivity function can best be determined in situ. For drying conditions, this is done most frequently with the instantaneous profile method, also called the unsteady drainage flux method [1, 4, 52, 77, 78]. Water contents and hydraulic potentials are measured as function of time and depth during drainage of an initially saturated, bare soil profile. When the water flux density, q, is known for all time at one depth, z_0, the flux density can be calculated for any depth and time from the water contents

$$q[z,t] = q[z_0,t] - \int_{z_0}^{z} (\delta\theta / \delta t) [z,t] \, dz \qquad (28)$$

This equation assumes only vertical transport, without root uptake. The boundary condition $q[z_0,t]$ is usually set as a zero flux at the soil surface, obtained by covering the surface to prevent evaporation. Hydraulic conductivities can then be calculated from calculated flux densities and measured hydraulic potentials obtained for a set of times and depths (if needed after smoothing and interpolation) from

$$k[\theta,z] = q[z,t] / (\delta H / \delta z) [z,t] \qquad (29)$$

Unless the draining surface area is very small, water contents can be determined gravimetrically by taking soil samples with an auger. This is accurate and does not take all that much time. Often, however, water contents are measured indirectly and non-destructively e.g. by neutron scattering, gamma attenuation, or capacitance sensors. Hydraulic potentials should be measured directly with tensiometers, using mercury manometers or pressure transducers. Hydraulic conductivities can thus be obtained for any layer between two tensiometers. Within the range of the experimental data a soil water characteristic can also be constructed for each distinct soil layer from the values of θ and h already measured.

The range of water contents that can be covered is limited on the wet end by the degree of saturation that can be attained by ponding the water on the soil surface. This is often no more than 90 % of the available pore volume because air tends to be entrapped by the wetting front. At the drier end, the water content range is limited by the drainage characteristics of the particular soil in its hydrological setting. At first, near saturation, θ and H should be measured as frequently as possible, because they vary so quickly that it is hard to obtain accurate results without automated data collection. After the first few days, further accurately measurable differences in water contents will take days and even weeks and even then yield only k values for pressure heads that usually do not go below - 200 cm. This is the main disadvantage of the method, namely the rather limited range of θ and h over which $k[\theta]$ can be determined. This is reflected in the concept of field capacity which still appears to be useful in practice in spite of theoretical misgivings.

An analysis of the error propagation of this method [79] is not very encouraging; especially towards the dry end, errors can be very large (Fig. 8). At small times tensiometer errors predominate, while later water content measurements introduce the largest errors.

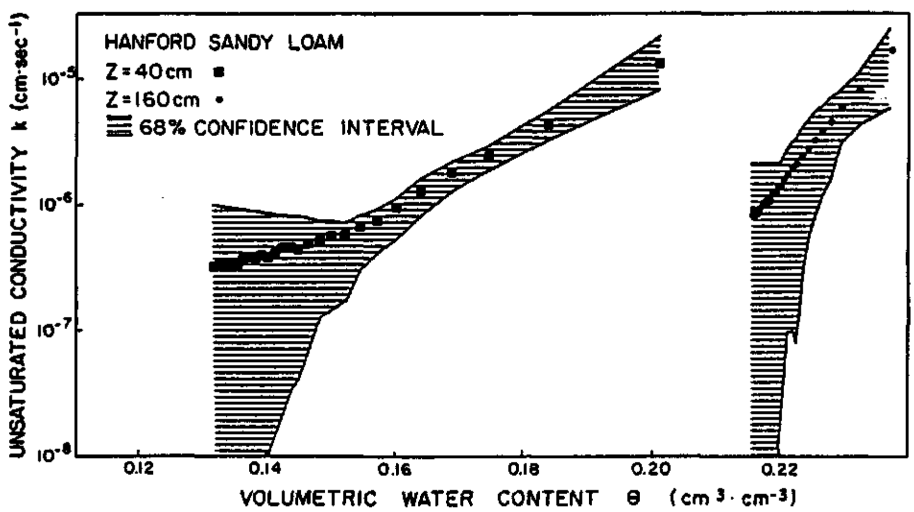

Fig. 8. Confidence intervals (68%) for hydraulic conductivity at two depths due to error propagation in instant profile calculations. (From Ref. 79)

To reduce errors in fine textured soils, water content measurements should be intensified; in coarse-textured soils it is better to increase the number and/or frequency of tensiometer measurements. Contrary to usual laboratory conditions which allow only non-destructive indirect soil water content measurements in soil columns, it is often quite possible to make repeated direct (gravimetric) soil water content determinations in instantaneous profile measurements in the field. Since this will improve the accuracy of the final results, if enough soil samples are taken, this is indicated between parentheses in Table 1. The h-range can be expanded by allowing evaporation from the soil surface and determining the zero-flux plane from the tensiometer data [80]. However, the overall results will be even less accurate. The same is true, if only, either water contents or hydraulic potentials, are measured and the others are derived from an independently determined soil water characteristic.

B. Unit Gradient With Prescribed k-Function

With the present emphasis on studying the spatial variability of soil hydraulic properties, there is a need for simple in situ measurements. Tensiometric measurements are much less convenient for this purpose than water content measurements, especially when the latter are performed with neutron probes. A simplified version of the instantaneous profile method involving only water content measurements was recently used by Jones and Wagenet [81]. They installed 100 neutron access tubes in a 50 x 100 m fallow field and wetted the soil around them by ponding water in 37-cm-diameter rings inserted 15 cm into the soil. When water contents were steady down to 120 cm, the access tube sites were covered and redistribution was followed for 10 days. At the end gravimetric samples were taken to back up the neutron measurements. The results were analysed in five somewhat different ways, all assuming the hydraulic gradient unity at all times and exponential hydraulic conductivity functions

$$k[\theta] = k_0 \exp [\beta (\theta - \theta_0)] \qquad (30)$$

where k_0 and θ_0 are values measured during steady ponded infiltration, sometimes called 'satiation'. All five analyses yielded values of the constants k_0 and β, with their mean and variance, for selected depths.

The difference between the analyses mostly concerned further assumptions on the water content distributions. For instance, in one analysis, already proposed by Libardi et al. [82], the average water content θ^* to depth z is assumed to be a linear function of the water content θ at depth z

$$\theta^* = a\ \theta + b \tag{31}$$

This leads, for larger times, to

$$\theta - \theta_0 = 1/\beta \ln t + 1/\beta \ln (\beta k_0 / z a) \tag{32}$$

Thus, for each depth a plot of $(\theta - \theta_0)$ versus ln t yields β as the reciprocal slope and the intercept, given a, yields k_0.

Jones and Wagenet concluded that the five approximate analyses will be most useful in developing relatively rapid, preliminary estimates of soil water properties over large areas , but are not as useful when k_0 and β at a particular location need to be known precisely.

C. Simple Unit Gradient

In an even more simplified version, uniform water content and pressure head (and thus unit hydraulic gradient) are assumed throughout the draining profile [4]. This implies that the increase of k with depth, which is needed to accomodate the increasing flux density with depth, is assumed to occur with a negligible increase of θ. The hydraulic conductivity is then

$$k[\bar{\theta}] = L (d\bar{\theta} / dt) \tag{33}$$

where $\bar{\theta}$ is the average water content of the profile above depth L. With a single tensiometer at depth L and making the same assumptions, the diffusivity can be determined analogously [83] as

$$D[h] = L (dh / dt) \tag{34}$$

Unless the soil profile is highly uniform, it is doubful that these versions can yield results better than an educated guess.

D. Sprinkling Infiltrometer

If hydraulic properties must be known for wetting conditions, the instantaneous profile analysis could be used on transient data obtained with a sprinkling infiltrometer. However, this equipment is much more elaborate (see section VII.A) than that needed simply to saturate a soil profile and, it normally must be attended whenever it is in operation.

E. Sorptivity Measurements

Sorptivity is the first term in the Philip infiltration equation [19] and is a function of θ_1 and θ_0 (see section II.D). This function contains composite information on other soil hydraulic transport properties [18, 71], which can be obtained mathematically. Saturated sorptivity is measured easily in the field [84]. To prevent macropores from dominating saturated sorptivity measurements, Clothier and White [85] measured "saturated" sorptivity under very small negative pressure heads. Dirksen [69] proposed the apparatus in Fig. 9 to measure sorptivities in situ over a large range of pressure heads. This was used by Russo and Bresler [50], with other measured parameters (saturated hydraulic conductivity, air-entry value and residual water content) to determine the probability density functions of k[θ] and h[θ] for statistical analysis.

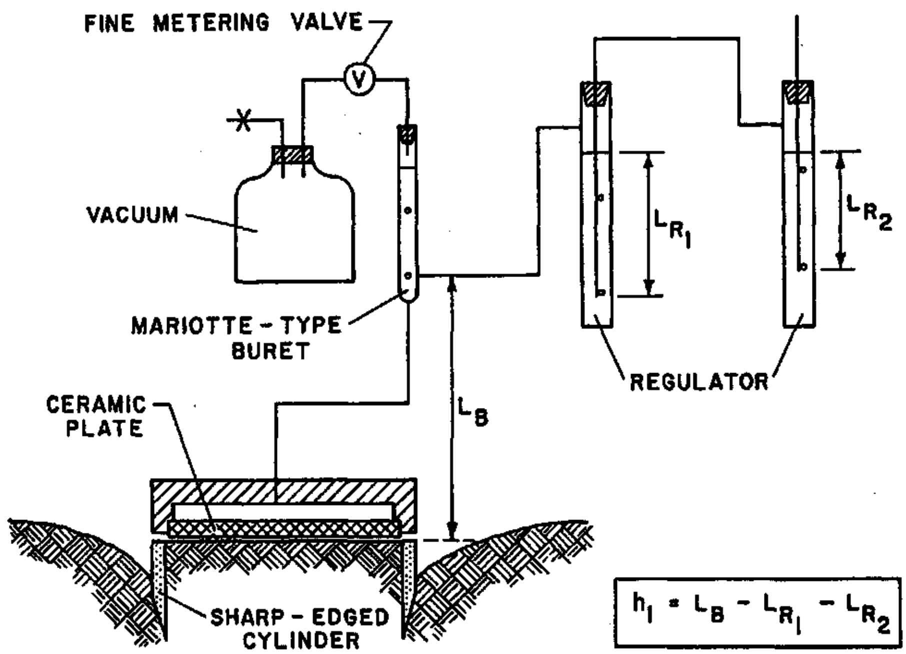

Fig. 9. Diagram of apparatus for in situ unsaturated sorptivity measurements. (From Ref. 69)

X. DERIVATION FROM OTHER SOIL PROPERTIES

A. Soil Water Retention Characteristic

Physical measurements of soil hydraulic conductivities and other transport parameters are time-consuming and tedious, and therefore expensive. Moreover, despite considerable effort, the accuracy most often is very poor. With the tremendous variability of these soil properties, both in space and in time, the practical value of such measurements is difficult to estimate. It is worthwhile, therefore, to consider the possibility of deriving these properties from more easily measured soil properties. The soil water retention characteristic is most often used for this purpose because, at least in the range of water contents where the capillary binding of water is predominant, it reflects the geometry of the pores and this geometry, in turn, determines to a large extent the hydraulic transport properties. The pressure head difference across an air - water interface is [10]

$$h = 2\,\sigma \;/\; \rho\, g\, R \tag{35}$$

where σ is the surface tension of the air - water interface (N/m), ρ is the density of water (kg/m^3), g is the gravitational constant (N/kg), and R is the equivalent radius of the interface (m). If the soil material is perfectly hydrophylic (i.e. zero angle of contact), then R is equal to the (equivalent) radius of the pore at the interface and the soil water retention characteristic can be converted into an equivalent pore size distribution: since the water content at any given pressure head is equal to the porosity contributed by the pores that are smaller than the equivalent diameter corresponding to that pressure head (measured with respect to atmospheric pressure) as given by Eq. (35).

There are two approaches to calculating soil hydraulic conductivities from soil water retention characteristics. One was originated by Childs and Collis-George [32] and later modified [86, 87]. The other, based on the generalized Kozeny equation, had its origin in the oil industry and was introduced into the soil literature by Brooks and Corey [49]. This approach will not be discussed here further; for a good summary of the theory and the final working equations, see Laliberte et al. [88].

Childs and Collis-George assumed that the soil consists of randomly distributed pores of various sizes, which can be divided into a number of size classes. If two imaginary cross-sections of a soil were to be brought into contact with each other, the hydraulic conductivity of the assembly would depend on the number and sizes of pores on each side that connect up with each other. The chance of pores of two sizes connecting is proportional to the product of the relative contributions of their respective pore size classes to the total cross-sectional area. Childs and Collis-George assumed further that, since according to the law of Poiseuille the flow of water through a pore is proportional to the square of its diameter, the flow through two matching pores is determined by the smallest of the two. By dividing the soil water retention characteristic into a number of pore size classes, based on Eq. (35), they finally obtained

$$k = F \frac{\rho g}{\eta} \sum_{\Gamma=0}^{\Gamma=R} \sum_{\delta=0}^{\delta=R} \delta^2 f(\Gamma)\, dr \quad f(\delta) \quad dr \tag{36}$$

where F is a correction (matching) factor to match the calculated hydraulic conductivity at a single water content to a measured value at the same water content, η is the viscosity of water (Pa s), and $f(\Gamma)dr$ and $f(\delta)dr$ are the partial areas occupied by pores of radii Γ to $\Gamma+dr$ and δ to $\delta+dr$, respectively.

With this equation the hydraulic conductivity for a selected water content can be obtained by carrying out the calculations up to the value of r for which the pores are still just water-filled. Jackson [89] reviewed and summarized the various versions of this equation and, since the calculations were quite cumbersome, proposed a simpler procedure without making basic changes. For a complete example of the required calculations according to Jackson, see Hillel [8, p. 223]. Many experimental verifications of this approach have been reported, [e.g. 89 - 92]. In all these the matching factor F (based on measured saturated hydraulic conductivities) was unpredictable and varied between 2.0 and 0.004. Often, the shape of the theoretical and experimentally determined curves for $k[\theta]$ also differed substantially.

Mualem [93] introduced a few basic changes to the theory of Childs and Collis-George [32]. For instance, he calculated the contribution to the hydraulic conductivity of a larger pore (radius r_1) following a smaller one (radius r_2). Assuming that the length of a pore is equal to its diameter, allowed him to define an equivalent radius of the two pores as $(r_1 r_2)^{½}$. Combining his theory with elements of the model of Brooks and Corey [49] for the soil water characteristic and of Burdine [94] for the relative hydraulic conductivity he found that, based on a comparison with experimental results of 45 soils, the relative hydraulic conductivity was described best by

$$k_r[\Theta] = \Theta^{1/2} \left[\int_0^{\Theta} (1/h)\, d\Theta \Big/ \int_0^{1} (1/h)\, d\Theta \right]^2 \tag{37}$$

where $k_r = k/k_s$ is the relative hydraulic conductivity, Θ is a dimensionless water content (see Eq. (25)), and θ_r is the residual water content, which is that water content at which the hydraulic conductivity becomes negligibly small.

Van Genuchten [95] proposed as approximation for the soil water retention characteristic

$$\Theta = [\ 1 + (\ -\alpha\ h\ )^n\ ]^{-m} \tag{38}$$

where α, n, and m are fitting constants. He then combined Eq. (38) with the model of Mualem (Eq. 37)

$$k_r[\Theta] = \Theta^{1/2}\ [1 - (1 - \Theta^{1/m})^m\ ]^2, \qquad m = 1 - (1/n) \tag{39}$$

By substituting into Eq. (39) the parameter values obtained in fitting Eq. (38) to a soil water retention characteristic, a relative hydraulic conductivity function is obtained without additional measurements. For absolute hydraulic conductivities, the hydraulic conductivity must be determined for one water content. Figure 10 shows the fits of Eq. (38) to experimental wetting and drying soil water retention characteristics of Pachappa fine sandy loam. The corresponding absolute hydraulic conductivity functions according to Eq. (39) are given in Fig. 7. The absolute values were obtained with an independently determined hydraulic conductivity at 'satiation', $\theta = 0.36$. The comparison with the experimental hydraulic conductivity data is very good for drying, especially in the drier range, but very poor for wetting. The reason for this is

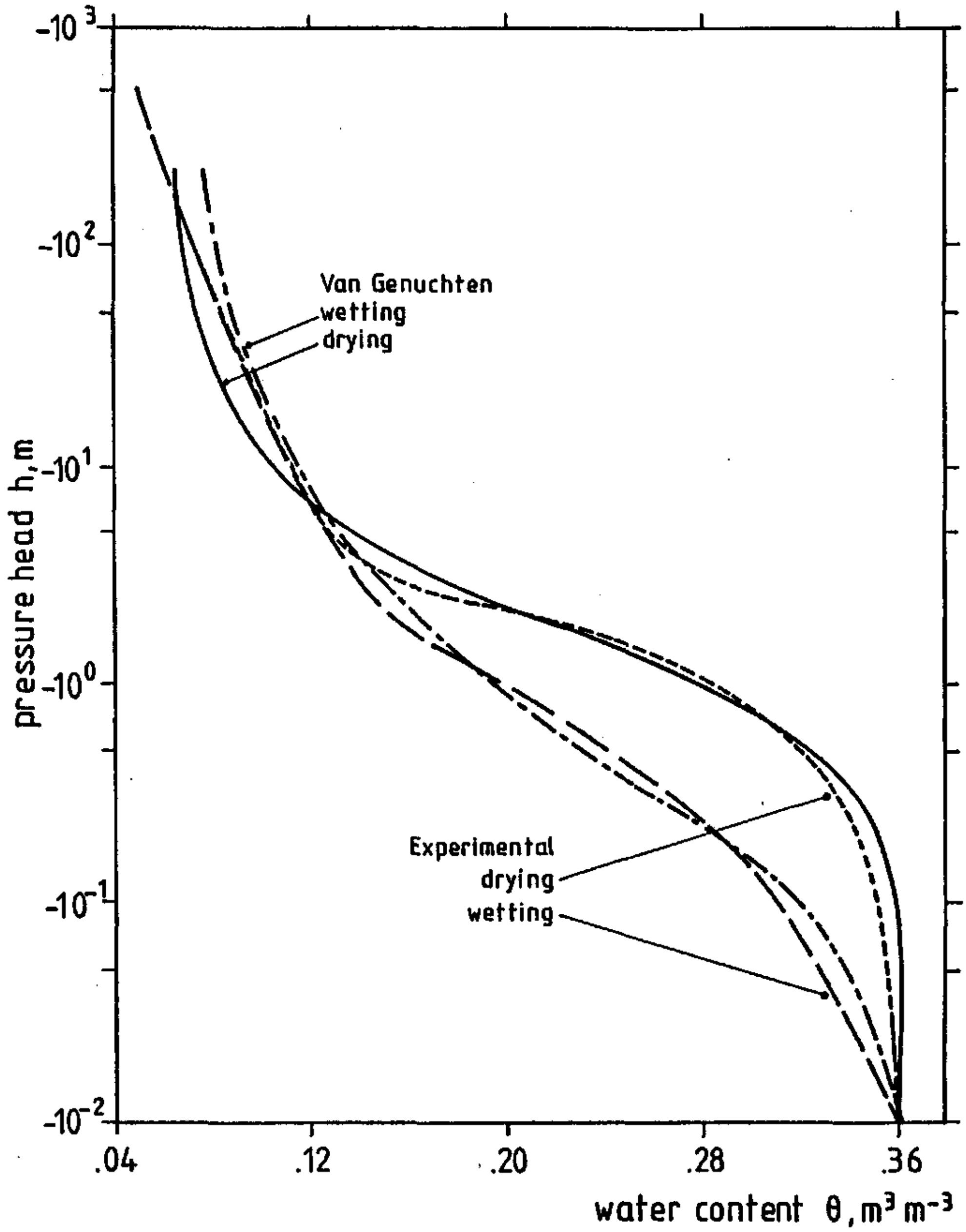

Fig. 10. Soil water retention characteristics of Pachappa sandy loam composed of various experimental data, and the fits of these to Eq. (38). The corresponding hydraulic conductivity functions according to Van Genuchten - Mualem are shown in Fig. 7.

not clear, nor whether this result can be expected generally. For a more extensive review of this and other models to calculate hydraulic conductivities, see Van Genuchten and Nielsen [6].

It is common practice to use measured saturated (or 'satiated') hydraulic conductivities to match calculated and measured values. In general, this is about the worst choice one can make. The standard deviation of such measurements is normally very large since they can be totally dominated by wormholes, old root channels, fractures resulting from poor sampling procedures, etc. More importantly, such features have no relation with the pore size distribution of the soil matrix. At small negative pressure heads, all large spaces not associated with the soil matrix are empty and do not conduct water. Therefore, I recommend that hydraulic conductivities measured at small negative pressure heads be used in the calculation procedure outlined above. These can be measured accurately and fast with the "head-head" technique (section VI.A).

The determination of θ_r, especially, is problematic. Van Genuchten developed a procedure to determine the parameters θ_r, α, n, and m simultaneously with a least squares curve-fitting algorithm of the soil water characteristic. This is used by many investigators and already has earned a certain reputation. More recently, Van Genuchten has developed a program with which up to 7 parameters (the 4 mentioned above, plus θ_s, k_s, and the exponent of Θ which in eq. (39) has the value ½) can be optimised based on differently weighted experimental data of $h[\theta]$ as well as $k[\theta]$. If desired, even the relationship between n and m, given with eq. (39), can be left out.

B. Scaling

If scaling relationships of Miller and Miller [96, 97] are assumed, soil hydraulic properties can often be determined with much less work than otherwise required. For example, Reichardt et al. [98] measured hydraulic diffusivities of 12 different soils with the fixed-time Boltzmann method [60] and converted these to hydraulic conductivities according to Eq. (4). When these hydraulic conductivities were scaled according to the square of a characteristic microscopic length, λ, the data coalesced nicely into one relationship (Fig. 11). The solid line in Fig. 11 can, for k in cm/s, be described by [20]

$$k[\Theta] = 1.942 \times 10^{-12}\, m^4 \exp\left(-12.235\, \Theta^2 + 28.061\, \Theta\right) \tag{40}$$

λ was assumed proportional to the square of the slope, m, of the linear relationship between advance of wetting front and square root of time during horizontal infiltration (see Eq. (7)) and is listed for each soil

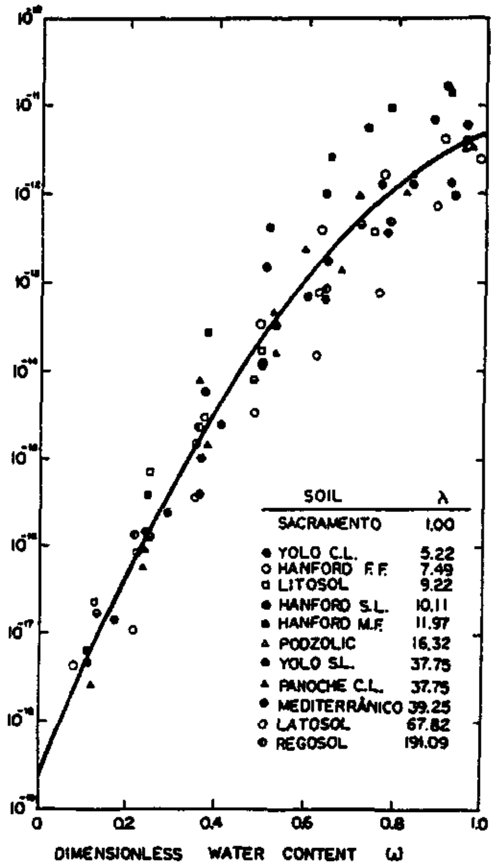

Fig. 11. Hydraulic conductivities of 12 soils scaled according to λ^2 (or m^4) versus dimensionless water content. (From Ref. 20)

in Fig. 11 as a ratio with the standard soil. If a soil belongs to the group for which this assumed scaling relationship is valid (which normally will not be known beforehand and needs to be verified) the hydraulic conductivity function can be obtained with Eq. (40) and just one simple, short infiltration run to measure m, θ_1, and θ_0.

Miller and Bresler [21] showed that the experimental data of Reichardt et al. [98] on which Eq. (40) is based, can be transformed to what they suggest to be a "universal" equation for the diffusivity

$$D[\theta] = \alpha\, m^2 \exp\, [\beta\, \Theta], \quad \text{with} \quad \alpha = 10^{-3} \text{ and } \beta = 8. \qquad (41)$$

Bresler et al. [99] derived a relationship for the hydraulic conductivity from the same experimental data

$$k[\theta] = 0.27\, m^4\, \Theta^{7.2} \qquad (42)$$

C. Texture

Hydraulic conductivities have also been correlated with soil textural data. These are more abundantly available than soil water retention characteristics and, therefore, attractive. However, the results do not have a physical basis and the observed relationships can only be of a statistical nature. They must be verified by measurements on a large number of soils, while it remains uncertain whether they can be extrapolated to soils outside the group used to obtain this relationship. If such correlations are shown to be reliable predictors, a lot of work could be saved.

Bloemen [22] defined a particle size distribution index

$$f = \sum_{i=1}^{n} [(p_{i+1} - p_i) \log(p_{i+1}/p_i) / \log(S_{i+1}/S_i)] / \sum_{i=1}^{n} (p_{i+1} - p_i) \quad (43)$$

where p_i is the cumulative weight percentage and S_i is the corresponding particle size class boundary.

Based on data for a large number of Dutch soils Bloemen found

$$k_s = 0.02\ M_d^{1.93}\ f^{-0.74} \quad \text{(cm/day)} \quad (44)$$

$$h_a = 2914\ M_d^{-0.96}\ f^{0.79} \quad \text{(cm)} \quad (45)$$

$$n = 1.4 + 4.536\ (e^{0.3f} - 1) - 0.75\ f^{1.6} \log OM \quad (46)$$

$$k\ [h] = k_s\ (h_a/h)^n \quad (47)$$

where M_d is the median particle size, h_a is the pressure head at air entry (cm), n is an empirical coefficient, and OM is the organic matter weight percentage.

It is doubtful that these results can be extrapolated to soils in other parts of the world. Schuh and Bauder [23] did a similar study on a number of soils in the USA. They found particularly good correlations between n and the sand to silt ratio.

XI. PARAMETER OPTIMIZATION

Recently, the so-called inverse approach has received renewed attention in the form of a parameter optimization technique. First proposed around 1970 [100, 101] the inverse approach requires a relatively simple experiment with inherently accurate measurements to be performed. Subsequently, assuming algebraic forms of the hydraulic property functions, the water transport process is simulated on a computer, starting with guessed values of the parameters in the transport functions and then repeated with the newly estimated values until the simulated results agree with the experimental results to within the desired degree of accuracy. Thus the problem is reduced to optimising the parameters in the transport functions. Optimization is a specialised mathematical process for which computer programs are available [102]. Mathematical details will not be discussed at this point. The technique appears to have been improved recently such that it has become attractive for solving soil water flow problems. To be able to decide how the hydraulic transport functions can best be determined in a given situation, the merits of this inverse approach should be appreciated. Only a few aspects of it will be discussed here. Further details can be found in the references. An up-to-date review is given in Kool etal. [7].

Whereas in principle many flow systems with different initial and/or boundary conditions can be used for the parameter optimization, the one-step outflow method is especially suitable [25, 103]. It only requires inherently accurate measurements of cumulative (external) outflow as function of time from an initially saturated short soil column as a result of a step-increase of the air pressure in a pressure plate apparatus. It allows a large water content range to be covered in a reasonably short time. The influence of the resistance of the porous plate on the outflow, which complicates the traditional analysis of the experimental results, is easily accounted for in the simulation. A draining soil column in which water content profiles must be measured at different times [24, 104, 105] is less attractive experimentally and can cover a much smaller water content range. Sir etal. [106] used one-dimensional infiltration as the flow process for optimization. The remarks in the following paragraphs specifically apply to optimization of the parameters in the Van Genuchten - Mualem functions (Eqs. 38 and 39), based on the experimental one-step outflow data of Parker etal. [103]. The same authors were also able to evaluate hysteresis in the hydraulic functions by solving the inverse problem consecutively for outflow and inflow on the same soil column [107].

A major aspect of the inverse approach is **convergence**. The first guess of the parameter values may be so far off from the actual values, that the optimization procedure can not yield the correct values or do this only after prohibitively long computing time. As first guess for medium textured soils the "average" values $\alpha = 2.50\ m^{-1}$, $n = 1.75$ and $\theta_r = 0.150$ may be taken, with suitable adjustments for differently textured soils. Convergence also may be a problem when the information contained in the input data is too scanty. Therefore, the input data should cover as large a range of water contents, time, etc. as practical. To prevent undue use of computer time a maximum number of function evaluations may be set. If the solution fails to converge within this number, a new solution can be started with different initial parameter values.
Another aspect of the inverse approach is **uniqueness**: there may be more solutions to the problem as stated and the solution obtained may not be the correct one. This is not expected to be a serious problem with the one-step outflow measurements, if the pressure step and the time period are kept relatively large. However, the obtained solutions should be verified and again,in case of doubt, the optimization process should be repeated with different initial estimates of the parameters.
The **accuracy** of the optimised parameters is dependent on the accuracy of the experimental data used as input in the optimization procedure. The sensitivity for this source of errors is different for each combination of flow process and parametric function and deserves further study. Of course, if the pre-selected algebraic functions are incapable of describing the actual soil hydraulic properties accurately, even a perfect optimization process will not yield an accurate result.

XII. SUMMARY AND CONCLUSIONS

Water transport in soils which are not fully saturated with water plays an important role in hydrology, water uptake by plant roots, irrigation management, transport of pollutants through the environment, etc. This transport is to a large extent characterized by the dependence on volume fraction of water, θ, of hydraulic conductivity, k, diffusivity, D, matric flux potential, Φ, and sorptivity, S. For a given soil, these soil water transport functions, $k[\theta]$, $D[\theta]$, etc. vary over several orders of magnitude and can differ by orders of magnitude between soils. Measuring these functions is a difficult task, on which much time and effort continues to be spent. Many methods have been proposed, but no single method is suitable for all conditions and/or purposes. Most methods lack accuracy, take a prohibitively long time and/or are costly. In general, steady state methods are more accurate than transient methods, but they take a lot more time and are therefore more expensive. There is also the choice to be made between laboratory and field measurements. The former have many advantages, which are spelled out in a special section, but they require the acquisition of undisturbed soil samples and the transport of these to the laboratory.

The absolute accuracy of any given method cannot be established by using it on a "standard" porous medium with very accurately known hydraulic properties. As a result, it is standard practice to compare between the results obtained by two (or more) different methods, without knowing the accuracy of either of them separately. It is necessary, therefore, to evaluate the available methods on their inherent features and potential accuracy. Various types of methods are described and evaluated in Table 1 with respect to a number of criteria and gradations, given in Table 2. Where the highest accuracy is required, methods should be selected according to: soundness of theoretical basis (criterion C), control of initial and boundary conditions (D), inherent accuracy of the required measurements (E), and error propagation (F). On these criteria, "head-head" measurements on undisturbed soil cores between two porous plates score the highest. It is proposed, therefore, in view of the lack of a "standard" material, to elevate this method to the status of "standard method", against which other available methods could/should be evaluated. A disadvantage of this method is that it can be used conveniently only over a pressure head range from saturation down to about -2.5 m (G). This is normally more than sufficient for hydrological studies. With special effort (parentheses in Table 1) a larger pressure head range can be covered at the expense of more time (H) and better equipment (I). This is justified when a "standard" measurement is needed. Of the other laboratory methods, the "flux - head" variant, long column infiltration, and flux-controlled sorptivity methods score the highest for criteria C to G.

As for field methods, the instantaneous profile method might seem to have only one big disadvantage, namely the very limited pressure head range over which it can yield results even after rather long time periods. Unfortunately, the error analysis of Fluhler et al. [79] shows that, even with directly measured pressure heads and using only Darcy's law, the accuracy of the final results can be very poor. Use of the sprinkling infiltrometer under steady state conditions at least eliminates large errors introduced when fluxes are calculated from indirectly measured water contents. Therefore, the sprinkling infiltrometer appears to be the strongest candidate for "standard field method". Operation of this equipment is very cumbersome and time-consuming. However, if accuracy is of overriding importance, criteria of required time (H), investments (I), skill (J), and operator time (K) should play a secondary role.

When accuracy is not as important as speed and minimizing cost, criteria H to K, as well as the potential for simultaneous measurements (L), become dominant. When many simultaneous measurements are made, it is also important (especially when these are carried out by unskilled workers) that some check on the quality of the work is possible (M). The recently proposed matric flux potential and ponded disk/dripper methods score quite high on these criteria. Also the hot air method is very attractive with respect to these criteria. However, the theoretical basis, control of boundary conditions, error propagation and limitations on measurement accuracy are in my opinion so totally unacceptable that the hot air method should no longer be used. The other Boltzmann-type methods do not have the disadvantage of poor boundary control and non-isothermal conditions, but the inaccuracy of the measurements and the analysis thereof are serious disadvantages. The spherical cavity method has a number of attractive features which appear to deserve further investigation. The pressure plate outflow method in its one-step variant is not good as a direct method due to the approximate nature of the analysis of the experimental data. As a basis for the inverse approach of parameter optimization, however, it is very attractive owing to the simple, accurate measurements involved.

The unpredictability and non-uniformity of the conductivity of the crusts, as they are presently being made for the crust method, makes its use questionable as to its potential accuracy, while the pressure head range is very small. The crust method is too cumbersome and too time-consuming to be suitable for routine measurements at many sites. The use of hypodermic needles with a pulsating pump as a substitute for the crust promises to eliminate or improve most of these limiting factors. This makes it a small, much simplified version of the sprinkling infiltrometer which may well prove to be very useful.

Derivation of the water transport functions from other soil properties may be a good alternative to direct measurements, particularly when the absolute accuracy is not of primary importance but many results are required (e.g. studies of spatial or temporal variability as such). Often, the required input data are already available. The Van Genuchten -Mualem model appears to have an edge on other alternatives. It has an adequate theoretical basis, is generally available in user-friendly PC programs and is, therefore, widely used, and has given good results for many studies. The same model is also used for the parameter optimization technique. In this "inverse" approach, those values of the parameters of the model are sought which give the best agreement between measured and numerically simulated quantities. It would seem that, as the mathematical procedure is further improved in terms of convergence, uniqueness and accuracy, this approach should be used more and more. This will be true particularly, if the selected experimental flow system can be tailored to the actual situation and conditions in which the results will be used.

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TABLE 1. EVALUATION OF METHODS TO MEASURE SOIL WATER TRANSPORT PROPERTIES ACCORDING TO CRITERIA AND GRADATIONS IN TABLE 2.

	CRITERIA												
STEADY STATE METHODS	A	B	C	D	E	F	G	H	I	J	K	L	M
Laboratory methods													
Head - head	k	w/d	5	5	3(5)	5	3(4)	2(1)	3(2)	3(2)	4	4	4
Flux - head (infiltration)	k	w	5	5	3	5	3	2	3	3	4	4	4
Head - flux (evaporation)	k	d	3	3	3	3	3	2	3	3	5	5	4
Regulated evaporation	k	d	2	2	3	3	2	2	3	3	4	2	4
Long column infiltration	k	w	4	4	4	5	4	1	3	3	5	4	4
Matric flux potential	Φ	d	3	3	5	3	3	3	3	4	5	5	4
Field methods													
Sprinkling infiltrometer	k	w	5	4	2	5	3(4)	2(1)	1	2	1	1	3
Isolated column	k	w	4	3	3	2	2	3	3	3	2	2	3
Spherical cavity	k	w	4	3	3	3	4	2	4	2	3	4	3
Ponded disk / dripper	k	w	2	3	4	3	3	4	4	4	5	4	4
TRANSIENT METHODS	A	B	C	D	E	F	G	H	I	J	K	L	M
Laboratory methods													
Instantaneous profile	k	d	5	5	2	2	3	2	2	2	2	2	2
Pressure plate outflow	D	d	2	4	5	3	2	2	3	4	3	4	3
One-step outflow	D	d	2	4	5	3	2	3	3	4	3	4	3
Boltzmann, fixed time	D	w	4	5	2	1	5	4	3	4	5	3	3
Boltzmann, fixed pos.	D	w	4	5	2	1	5	5	1	2	4	2	2
Hot air	D	d	4	1	2	1	5	4	4	4	4	3	2
Flux-controlled sorptivity	D	w	4	4	4	3	5	4	3(1)	4	3	2	4
Field methods													
Instantaneous profile	k	d	5	4	2(4)	2	3	2	2	2	2	2	2
Unit gradient, prescribed	D	d	2	3	2	2	3	3	4	2	2	4	2
Unit gradient, simple	k/D	d	1	1	4	2	3	2	4	3	3	4	2
Sprinkling infiltrometer	k	w	4	3	2	2	3	2	1	1	1	1	2

TABLE 2. SELECTION CRITERIA AND GRADATIONS FOR METHODS TO MEASURE SOIL WATER TRANSPORT PROPERTIES.

A. PARAMETER MEASURED
   k. Hydraulic conductivity
   D. Hydraulic diffusivity
   Φ. Matric flux potential

B. FLOW REGIME
   w. Wetting
   d. Drying

C. THEORETICAL BASIS
   5. Simple Darcy law or rigorously exact
   4. Exact, with minor simplifying assumptions
   3. Quasi exact, with simplifying assumptions
   2. Major simplifying assumptions
   1. Minimal theoretical basis

D. CONTROL OF INITIAL / BOUNDARY CONDITIONS
   5. Exact - no requirements
   4. Indirect and accurate
   3. Approximate
   2. Approximate part of the time
   1. Little control, if any

E. ACCURACY OF MEASUREMENTS
   5. Weight, (external) volume of water, and time
   4. Water content measurements, direct
   3. Pressure head measurements
   2. Indirect measurements, and/or other sources of error
   1. Approximate measurements without calibration

F. ERROR PROPAGATION IN DATA ANALYSIS
   5. Simple quotient (Darcy law)
   4. Accurate algebraic operations with accurate data
   3. Inaccurate operations with accurate data
   2. Accurate algebraic operations with inaccurate data
   1. Inaccurate operations with inaccurate data

G. RANGE OF APPLICATION (PRESSURE HEADS)
   5. Saturation to wilting point (0 to -160 m)
   4. Tensiometer range (0 to -8.5 m)
   3. Hydrological range (0 to -2.5 m)
   2. Dry range (-2.5 to -150 m)
   1. Wet range (0 to -0.5 m)

H. DURATION OF METHOD
5. 1 hour
4. 1 day
3. 1 week
2. 1 month
1. More than 1 month

I. EQUIPMENT
5. Standard for soil laboratory
4. General purpose, off the shelf
3. Easily made in average machine shop
2. Special purpose, off the shelf
1. Special purpose, custom-made

J. OPERATOR SKILL
5. No special skill
4. Some practice
3. General measuring experience
2. Special training of good experimentalist
1. Highest degree of specialisation

K. OPERATOR TIME
5. Simple and fast manipulations only at beginning and end
4. Elaborate manipulations at beginning and/or end
3. Simple and fast operations at regular time intervals
2. Elaborate operations at regular time intervals
1. Operator required during entire measuring period

L. SIMULTANEOUS MEASUREMENTS
5. No limit
4. Large number, at significant costs
3. Small number, at little costs
2. Small number, at substantial costs
1. No potential

M. CHECK ON MEASUREMENTS IN PROGRESS OR AFTERWARDS
5. Continuous monitoring of all parameters possible
4. Verification easy at any time
3. Each verification requires considerable effort
2. Single check is major effort
1. Check not possible