A new calculation procedure and simple set-up for the evaporation method to determine soil hydraulic functions

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ABSTRACT

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Keywords: unsaturated hydraulic conductivity, water retention characteristic, evaporation method, calculation procedure, set-up.

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Contents

Preface

Wind's evaporation method has been used over 30 years by the ICW and the SC-DLO. The merit of this method is the simultaneous determination of both the water retention characteristic and the unsaturated hydraulic conductivity on one soil sample. During the years several improvements and changes have been made. Also the set-up has been improved. The latest development is an automated, computer controlled set-up for measurements on 10 samples.
The goal of this report is to document in detail how the evaporation method is used today.

Another goal is to describe a relative cheap and very simple set-up for this method. This type of battery powered set-up can be used in remote sites or sites with a poor infra structure.

Summary

The evaporation method is used to determine the water retention characteristic and the unsaturated hydraulic conductivity of the same soil sample simultaneously. Desorption characteristics are obtained from homogeneous soil samples in the laboratory under relatively natural conditions. The water retention characteristic can be obtained in the whole tensiometer range, i.e. from 0 to -900 cm. The range of the conductivity determination is dependent on the soil type. For sandy soils the range is approximately from -90 to -500 cm and for loamy and clay soils from -5 to -800 cm. Only measurements of time, pressure heads and average water content are needed. A simple set-up has been designed to enable measurements by hand. The set-up consists of a scale and a relatively low-cost pressure transducer with read-out unit, stopcocks and tensiometers. An efficient calculation procedure has been developed to enable automatic calculations. The measurements are processed by a set of computer programs running on a personal computer.

1 Introduction

Studies on environment, agriculture and ecology require ample use of models describing water movement in soils. Knowledge of the hydraulic properties is of vital importance to these kind of quantitative studies. The need for methods to determine the water retention characteristic and the hydraulic conductivity increases with the progress in these numerical simulation models. The still decreasing costs for computation enable more sophisticated and accurate models which need accurate input data.

The determination of the water retention characteristic is generally not seen as a problem. Various standard methods are available, both for laboratory and field use (e.g. Klute, 1986). However, many methods require an equilibrium to be reached, which makes them time-consuming. Determining the hydraulic conductivity is more difficult. The available methods are either time-consuming, or dependent on the hydrological conditions for field methods, or they require simplifications and assumptions.

Laboratory determinations can be divided into steady-state and transient methods. Steady-state determinations of the hydraulic conductivity have been used for a long time (e.g. Richards, 1931; Dirksen, 1991). In general they are accurate, but it may take a long time before steady state is reached, especially when the soil becomes drier, and the hydraulic conductivity becomes very low. Transient methods are faster, but often various assumptions must be made that cannot be met in practice. The so-called "hot air method" (Arya *et al.*, 1975) is one example where simplicity and speed of measurements are advantageous, but serious objections against the assumed isothermal conditions can be raised (Van Grinsven *et al.*, 1985). With some more effort and a more complicated set-up, the flux-controlled sorptivity method (Dirksen, 1979) can be used. However, a tendency of instability and fingering of the wetting front in pure sands may cause problems (Dirksen, pers. commun., 1991). Pressure plate outflow (Gardner, 1956) and one-step outflow (Doering, 1965) methods were widely used, but generally failed to yield satisfactory results (Dirksen, 1991). The outflow methods were revived when the theoretically elegant inverse-modelling approach was introduced (Kool *et al.*, 1985). This method estimates the soil hydraulic properties, but does not always converge or result in unique solutions (Dirksen, 1991; Feddes *et al.*, 1988; Tamari, 1992; van Dam *et al.*, 1992).

Gardner and Miklich (1962) introduced an evaporation method. They used measurements of two tensiometers in a soil sample evaporating at one side. Series of constant fluxes were imposed on the initially closed and equilibrated sample. The flux had to be sufficiently small in order to assume a constant hydraulic conductivity and diffusivity. Becher (1971) simplified this method by letting the soil sample evaporate continuously. He used the measured flux density at the top instead of the required flux density in between the tensiometers. This and other oversimplifications were corrected by a set of soil-dependent factors (Becher, 1975). Recently Wendroth *et al.* (1993) improved this method. It gives good results for loamy and silty clay soils, assuming a linearly decreasing water content in the soil sample. Malicki *et al.* (1992) avoided oversimplifications by measuring the water content, using TDR (time-domain reflectometry), and pressure heads at five depths.

Wind (1968) improved and extended the evaporation method with simple equipment. He used Bouyoucos' electrical nylon units (Bouyoucos & Mick, 1948) to measure the pressure heads and a scale for the total weight of the sample. With a graphical iterative procedure, he calculated, by hand, the water retention characteristic from measurements of only pressure heads and the average water content of a homogeneous soil sample. The hydraulic conductivity was obtained from graphical differentiation of the measured pressure head profile and changes in the water content profile. The latter was calculated from the measured pressure heads, using the calculated water retention characteristic. The merit of Wind's method, in contrast to many other methods, is the simultaneous determination of both soil hydraulic functions on one sample.

Boels *et al.* (1978) avoided Wind's iterative calculation of the water retention characteristic. In order to use this direct method they had to describe the retention characteristic with a polygon. The flux densities were calculated from the change of the water content profile, described by an analytical function. The parameters of this function were iteratively calculated.

Using a polygon to fit the water retention characteristic does not always yield a physically realistic description. Therefore we automated Wind's iterative procedure. This includes the development of algorithms to substitute the soil physical intuition needed by Wind's manual method. The automation avoids also the subjective interpretation of the experimenter. We did not use one or another smooth function to describe the water content profile as both Wind and Boels *et al.* did. Simulations showed that this is not necessary in the pressure head range used (Tamari *et al.*, 1993).

This paper describes the new calculation procedure and a simple set-up to determine simultaneously the water retention characteristic and unsaturated hydraulic conductivity using the evaporation method in a relatively simple and cost-effective manner. Only measurements of time, hydraulic heads and average water content are needed. Computer programs have been developed to calculate the water retention characteristic and hydraulic conductivity from these measurements. These developments enable determinations in less well-equipped laboratories and the use for e.g. land evaluation in remote sites, using simple, portable, battery-operated equipment. The validation and accuracy of the method has been described by Tamari *et al.* (1993). The method and set-up were illustrated by determining soil hydraulic characteristics of two different soils.

2 Methods and materials

2.1 Soil samples

A loamy fine sand and a loam to silty loam were used in the experiments described here. The loamy fine sand was taken from the B horizon at a depth interval of 25 to 35 cm under grass. The soil was a Gleyic Podzol (FAO-UNESCO, 1974). The loam to silty loam was taken from the A horizon at a depth of 15 to 25 cm in an arable (marine) Calcaric Fluvisol.

2.2 Experimental procedures

Undisturbed soil samples were taken in PVC sleeves (inner diameter 110 mm, height 80 mm). Each sleeve was equipped with an iron cutting head. The sleeves were gently pushed into the soil using a hydraulic jack. The sleeves were dug out and the cutting rings removed. The projecting soil was removed with a straight knife.

The samples were saturated by standing them in a few centimetres of water for at least two weeks. Four tensiometers (outer diameter 6 mm, length 55 mm) were horizontally placed into pre-bored holes at heights of 10, 30, 50 and 70 mm from the bottom of the soil sample. The bottom was closed with a lid to prevent evaporation. The tensiometers were connected to stopcocks and a pressure transducer (see below, *Tensiometer box*). The soil sample was placed on an electronic scale (Mettler PM 4800; resolution 0.1 g; accuracy ± 0.1 g). Figure 1 shows the set-up (cf. Krahmer, 1987). The top of the sample was covered and the saturated sample and tensiometers equilibrated for at least 12 h. The top lid was removed at the beginning of the experiment. The soil sample could evaporate only from the top in a controlled environment ($T = 20 \pm 2$ °C, relative humidity = 50 ± 5 %). Under these conditions the evaporation was approximately 2.5 mm d^{-1}. At regular intervals the hydraulic heads h_h and the total weight of the soil sample were measured. Averaging h_m over too long time intervals (e.g. more than 8 h) should be avoided, since this could lead to bias in $\Delta h_m/\Delta z$ and thus in the calculation of K.

The experiment ended when air entered the top tensiometer. The measurements took 8.5 d for the loamy sand. The average water content of the sample decreased from 0.34 to 0.09 $m^3\,m^{-3}$ during that period. For the loam, the experiment took 5.6 d and the average water content decreased from 0.41 to 0.24 $m^3\,m^{-3}$. The lowest recorded pressure or tensiometer heads for the loamy sand and the loam were -886 and -909 cm respectively. The soil was removed from the sleeve, weighed and dried at 105 °C to determine the average water content at the end of the experiment. From these data and the measured weights the average water contents $\bar{\theta}_t$ (t = 1, 2,..n; measurement scan index) during the experiment were calculated. The pressure (or

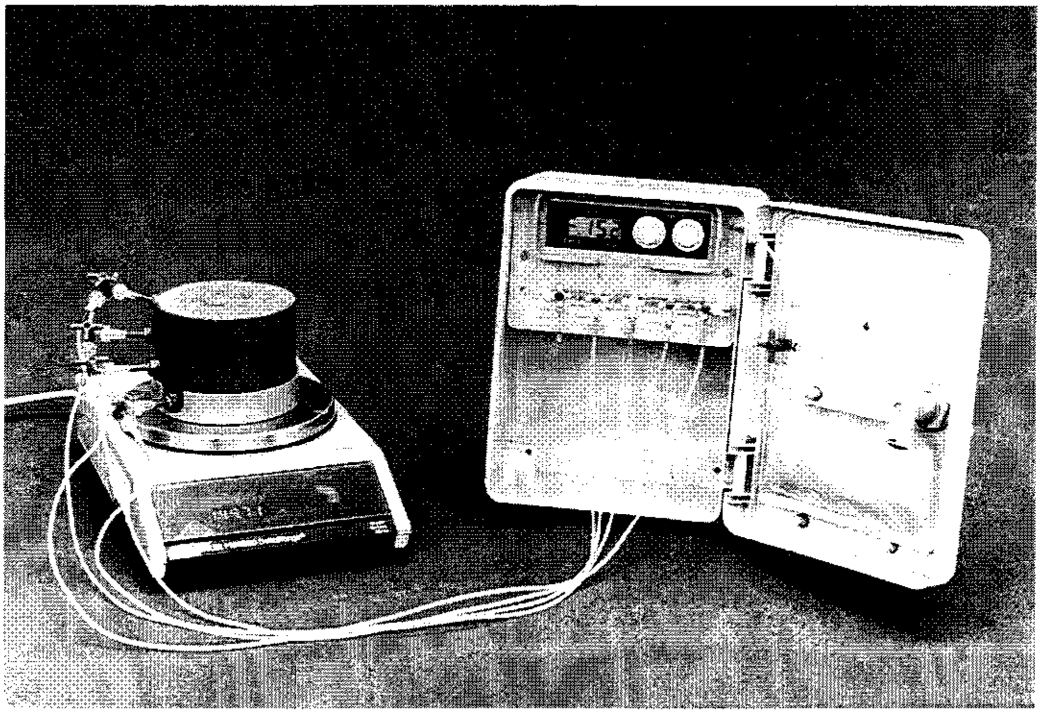

Fig. 1 Set-up of the evaporation method. The unsaturated hydraulic conductivity and the water retention characteristic are calculated from pressure heads and average water contents of a soil sample evaporating in the laboratory. The average water content is derived from the weight loss of the sample, measured with a balance (left). The pressure heads are derived from tensiometer measurements, using a 'tensiometer box' (right). The 'tensiometer box' consists of a pressure transducer with read-out unit, valves, tubing and a water supply bag.

tensiometer) heads, h_p, were calculated from the measured hydraulic heads, h_h, and the gravitational heads, h_g. The latter were given by the position of the tensiometer and the pressure transducer above the reference level. Other heads, such as osmotic and pneumatic heads were assumed to be negligible in the soil samples used. Therefore the matric head h_m equals h_p:

$$h_m = h_p = h_h - h_g \tag{1}$$

The time of the measurement, the average water content and the four pressure heads were entered into a sequential ASCII file of a personal computer. These data were checked by plotting the measurements against time. Outliers and non-reliable readings (e.g. due to too slow reaction time of tensiometers at the end of the experiment caused by air bubbles) were removed interactively. From these screened data the K-h_m-θ relationships were calculated (see below, *Computer programs*).

2.3 Tensiometer box

The hydraulic heads were derived from tensiometer readings obtained with a 'tensiometer box'. This equipment is an improved version of Bakker's system (Bakker, 1978). It can be used both in the field and in the laboratory. It consists of (i) a battery-operated pressure transducer with read-out unit (Druck DPI 700, range 0 to -0.1 MPa, resolution 1 hPa, accuracy better than ±1 hPa); (ii) a block of five cheap stopcocks (Mallinckrodt 91084); (iii) watertight and airtight flexible tubing; and (iv) a water supply bag. These parts were assembled in a standard plastic box (see Fig. 1). The tubing consists of a polyamide (nylon; type 6) inner tube (inner and outer diameters 1.0 and 1.8 mm respectively). This tube has a small compliance, no creep and is airtight, but rather permeable to water vapour. Therefore the polyamide tube was placed in a polyethylene outer tube (inner and outer diameters 2.0 and 3.0 mm respectively). This tube has not the favourable properties of polyamide, but it is not permeable to water vapour. The inner and outer tube were melted together. In this way airtight and watertight flexible tubing with high mechanical compliance was obtained. Switching from one tensiometer to another resulted in a fast response. This response time depends also on soil type and soil water content and was always smaller than 30 s. The water supply bag was used to flush the system. It was made of an aluminium laminate with outer layers of polyethylene and polypropylene. The whole system remained air-free during several months. The total cost for the parts used was approximately 1250 Dutch guilders.

2.4 Calculation procedure

The method can be used for soil samples that are homogeneous in their hydraulic properties. It is assumed that (i) the osmotic and pneumatic heads are negligible and (ii) the water flow can be described with a generalization of Darcy's equation:

$$v = -K(h_m)\frac{\partial h_h}{\partial z} = -K(h_m)(\frac{\partial h_p}{\partial z}+1) = -K(h_m)(\frac{\partial h_m}{\partial z}+1) \tag{2}$$

where v = flux density in cm d^{-1}; $K(h_m)$ = unsaturated hydraulic conductivity in cm d^{-1} as a function of the matric head h_m in cm; h_h = hydraulic head in cm; h_p = pressure (or tensiometer) head in cm; and z = the vertical coordinate in cm (positive upwards). The hydraulic reality is simplified as follows: We assume that the sample can be divided into compartments around the tensiometers. The matric heads and water contents are both assumed to be the same in the whole compartment. The calculation procedure needs at least two compartments. In this paper, we have used four compartments.

The calculation procedure can be divided into two parts: the iterative procedure to calculate $h_m(\theta)$, and the calculation of $K(h_m)$ and $K(\theta)$.

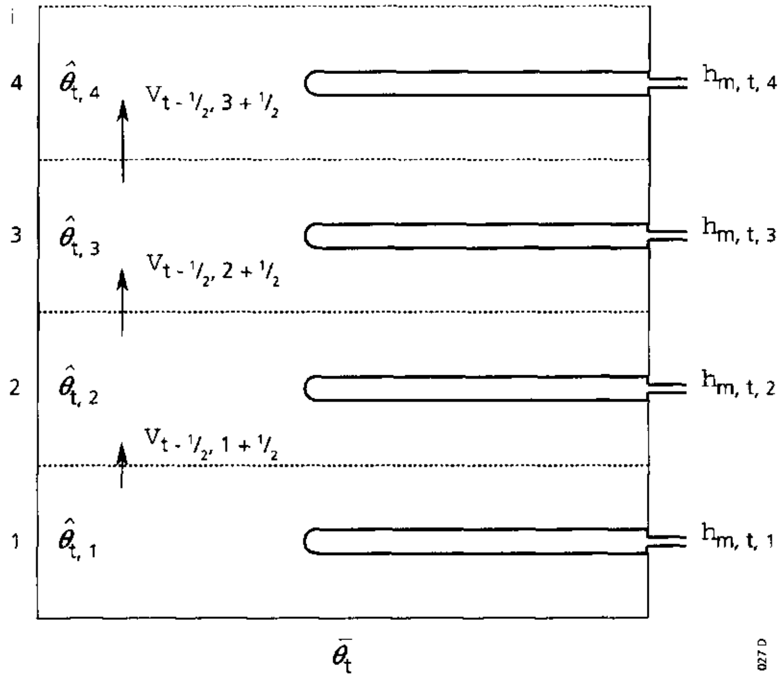

Fig. 2 Diagram of the soil sample. The sample is divided into four compartments. A tensiometer is placed in every compartment. The sample evaporates from the top. The four matric heads, $h_{m,t,i}$, and average water content, $\bar{\theta}_t$, of scan t are measured. The water contents, $\hat{\theta}_{t,i}$, of the compartments are estimated. The flux densities, $v_{t-1/2,i+1/2}$, in between scans t - 1 and t and in between compartments i and i+1 are calculated from the estimated water contents.

2.4.1 Iterative calculation of $h_m(\theta)$

As a *first step* in the iteration procedure, $h_m(\theta)$ is approximated from the measured average water content $\bar{\theta}_t$ (t = 1, 2, ...n; measurement scan index) and the mean, $\bar{h}_{m,t}$, of the measured matric heads, $h_{m,t,i}$ (i = 1, 2,..4; compartment index, bottom compartment = 1) of each scan t. A curve is fitted through this set of n data points. This can be any appropriate curve. We use a fifth-order polynomial to start the iterations. The polynomials are calculated with multiple linear regression, where the errors in θ are assumed to be much smaller than those in h_m.

In the *second step* of the iteration procedure, the four values of the measured $h_{m,t,i}$ of each scan t are inserted into the polynomial. This results in four estimated water contents, $\hat{\theta}_{t,i}$, for the four compartments for each scan. From these four $\hat{\theta}_{t,i}$, the mean estimated water content $\hat{\bar{\theta}}_t$ of the sample is calculated. This $\hat{\bar{\theta}}_t$ is compared with the measured average water content $\bar{\theta}_t$. All the four estimated water contents $\hat{\theta}_{t,i}$ are multiplied by $\bar{\theta}_t/\hat{\bar{\theta}}_t$. This is done for all n scans. This results in a new set of $h_{m,t,i}$ and updated $\hat{\theta}_{t,i}$. A new curve is fitted through this data set. Then we use a sixth-order polynomial that describes the $h_m(\theta)$ accurately for almost all natural soils. When the slope is not physically realistic, i.e. increasing θ with decreasing h_m, the order of the polynomial is decreased.

In the *next step* of the iteration procedure the calculations of the second step are repeated. For each scan t the four values of the measured $h_{m,t,i}$ are inserted into the updated polynomial. From the estimated water contents $\hat{\theta}_{t,i}$ of the compartments the mean estimated water content $\hat{\theta}_t$ is calculated again. The $\hat{\theta}_{t,i}$ are updated and a new polynomial is fitted to the data set. These iterations continue until there is statistically no significant improvement (F-test on the variances of the residues of the old and updated regression curves). The procedure is efficient, usually it leads to a solution in three to five iterations.

2.4.2 Calculation of K

For all the scans t - 1 and t (t = 2, 3,..n) the decrease in water content in compartment i (i = 1, 2, 3) is calculated from $\hat{\theta}_{t-1,i}$ - $\hat{\theta}_{t,i}$. From these decreases, the compartment height and the time in between the scans the three flux densities $v_{t-½,i+½}$ in between the compartments are calculated. We used a zero flux at the bottom of the soil sample as the lower boundary condition:

$$v_{t-1/2,i+1/2} = \frac{1}{\tau_t - \tau_{t-1}} \sum_{j=1}^{i} a_j(\hat{\theta}_{t-1,j} - \hat{\theta}_{t,j}) \tag{3}$$

where $v_{t-1/2,i+1/2}$ = the flux density in between compartment i and i + 1 and in between scan t - 1 and t; τ_t = the moment of time of scan t and a_j = the height of compartment j.

The gradients are calculated from the geometric mean matric head of scans t - 1 and t between two adjacent compartments i and i + 1:

$$\frac{\partial h_m}{\partial z} \approx \left(\frac{\Delta \bar{h}_{m,t-½,i+½}}{\Delta z}\right) = \left(\frac{-\sqrt{h_{m,t,i+1}\, h_{m,t-1,i+1}} + \sqrt{h_{m,t,i}\, h_{m,t-1,i}}}{z_{i+1} - z_i}\right) \tag{4}$$

The geometric mean is used to average h_m, because the strong non-linear distribution of h_m within the sample. Geometric means give better results than arithmetic means in those cases. This is analogous with the calculation of mean values of K (Schnabel & Richie, 1984).

With these data the value of the unsaturated conductivity K is calculated with Equation (2). The quotient of the flux density and the hydraulic head gradient gives the conductivity. We observed slightly decreasing flux densities during the experiments. However, the absolute value of the hydraulic head gradient is small ($\approx$ 0) in the beginning of the experiment and increases gradually. When this gradient is small, its measuring noise has a large influence on K (Tamari *et al.*, 1993). Therefore only gradients are used that have an absolute value above the measuring noise. A threshold of two times the standard deviation of the gradient reduces the noise in K considerably (Tamari, 1992). The resolution of the 'tensiometer box' (see above) used is 1 cm, which acts as a sufficient threshold of 0.5 for the measuring noise in the hydraulic head gradient (Δz = 2 cm for the experiments described).

The matching water content of the K value is calculated from the arithmetic mean of the water contents of the adjacent compartments i and $i+1$ of scans $t-1$ and t. Therefore:

$$\bar{\theta}_{t-½,i+½} = \frac{1}{4}(\hat{\theta}_{t,i} + \hat{\theta}_{t,i+1} + \hat{\theta}_{t-1,i} + \hat{\theta}_{t-1,i+1}) \tag{5}$$

where t = 2, 3..n and i = 1, 2, 3.

The matching matric head is calculated from the geometric mean of the matric heads of the adjacent compartments i and $i+1$ of scans $t-1$ and t. Therefore:

$$\bar{h}_{m,t-½,i+½} = -\sqrt[4]{h_{m,t,i}\ h_{m,t,i+1}\ h_{m,t-1,i}\ h_{m,t-1,i+1}} \tag{6}$$

where t = 2, 3..n and i = 1, 2, 3.

This gives maximally $(n-1)(i-1)$ data points for the $K(\theta)$ and $K(h_m)$ relationships. These data sets can be described with a suitable curve.

2.5 Computer programs

A set of computer programs has been developed to calculate the $h_m(\theta)$, $K(h_m)$ and $K(\theta)$ relationships on an IBM (compatible) personal computer.
The measurements are stored, together with experimental data (e.g. dimensions of the sample, number and placing of the tensiometers, comments), in a formatted sequential access ASCII file (= DOS or ASCII file). The data input program reads this file and checks the data for realistic values and gradients. The measurements are retrieved from this file and stored in an unformatted direct access file for further calculations. The plot program is used to inspect and, if necessary, correct the data interactively. The data are displayed on a graphic screen. Outliers can be deleted or replaced by an interpolated value. The water retention program calculates $h_m(\theta)$ following the iterative procedure described. Graphical and numerical results are given. A hard copy can be obtained on an EPSON (compatible) printer. The numerical data are stored in an ASCII file. They can be used, together with additional data, to fit Van Genuchten parameters (Van Genuchten, 1980) on this data set. The conductivity program calculates the $K(h_m)$ and $K(\theta)$ relationships. Graphical output is displayed and hard copies can be obtained. Numerical output of the K-h_m-θ data triplets is stored for further processing, such as the output of the water retention program.
The programs work under MS-DOS, version 3.3 or higher, and are written in Microsoft FORTRAN, version 5.0, and Microsoft C, version 5.1. An EGA or VGA graphic adapter is needed for graphical output. The programs can be obtained from the authors.

3 Results

The water retention characteristic, $h_m(\theta)$, and the unsaturated hydraulic conductivity, $K(h_m)$ and $K(\theta)$, were determined for two different undisturbed soil materials: a loamy sand and a loam. The evaporation method was used with a simple set-up (Fig. 1). The data of the measurements were entered into a personal computer and the K-h_m-θ relationships were calculated, using a new calculation procedure. The results are presented in Figure 3.

The water retention characteristic was calculated for the whole range of measurements, i.e. the tensiometer range. The $h_m(\theta)$ curve of the loamy sand was described with two polynomials (Fig. 3a). A fit on all measurements resulted in a third-order curve that is unable to describe the $h_m(\theta)$ curve adequately. Higher-order curves could not be used since they result in a physically unrealistic shape. A description with Van Genuchten parameters (Van Genuchten, 1980) does not yield satisfactory results for this soil because four parameters give too less degrees of freedom (Durner, 1992). This problem was solved by using a third-order curve for the measurements between -5 and -25 cm and a sixth-order curve for those between -25 and -886 cm. The $h_m(\theta)$ curve for the loam was described with a fifth-order polynomial, since its steep shape resulted in oscillations for sixth- and higher-order polynomials (Fig. 3d).

The conductivity data were plotted as individual points. These clouds of points were approximated with second-order polynomials. The $K(\theta)$ and $K(h_m)$ curves for loam (Figs 3e and 3f respectively) covered almost the whole range of measurements. This range was slightly reduced due to averaging h_m and θ between compartments and scans (see Materials and Methods). The situation for the loamy sand was different. A clear gradient in h_m started to develop below -90 cm. Above this value no gradients could be measured with the simple equipment used. Therefore no K could be calculated (Figs 3b and 3c). In another experiment the gradients were enhanced by increasing the evaporation rate with a fan above the soil sample. An increase in the evaporation rate from 2.5 to 5 mm d^{-1} resulted in the development of a gradient below -80 cm. Disadvantages of an increased evaporation rate are the quick development (-90 cm h^{-1}) of h_m and the large gradients (-300 cm cm^{-1}) at the end of the experiment. Averaging h_m and gradients for calculating $K(h_m)$ resulted in a reduced accuracy and a smaller range. Also deviations from the static water retention characteristic can be expected (Vachaud *et al.*, 1972).

The decrease in the values of K for loamy sand (Fig. 3c) diminishes with lower h_m values. This phenomenon has also been observed in other experiments (Feddes *et al.*, 1988; Wösten & Van Genuchten, 1988). This effect is too large to be caused by measuring errors or averaging effects. This can be explained by the steep curved behaviour of the relationship between $\log(|h_m|)$ and θ (Fig. 3a) and the more or less linear behaviour of the relationship between $\log(K)$ and θ (Fig. 3b) at small θ values.

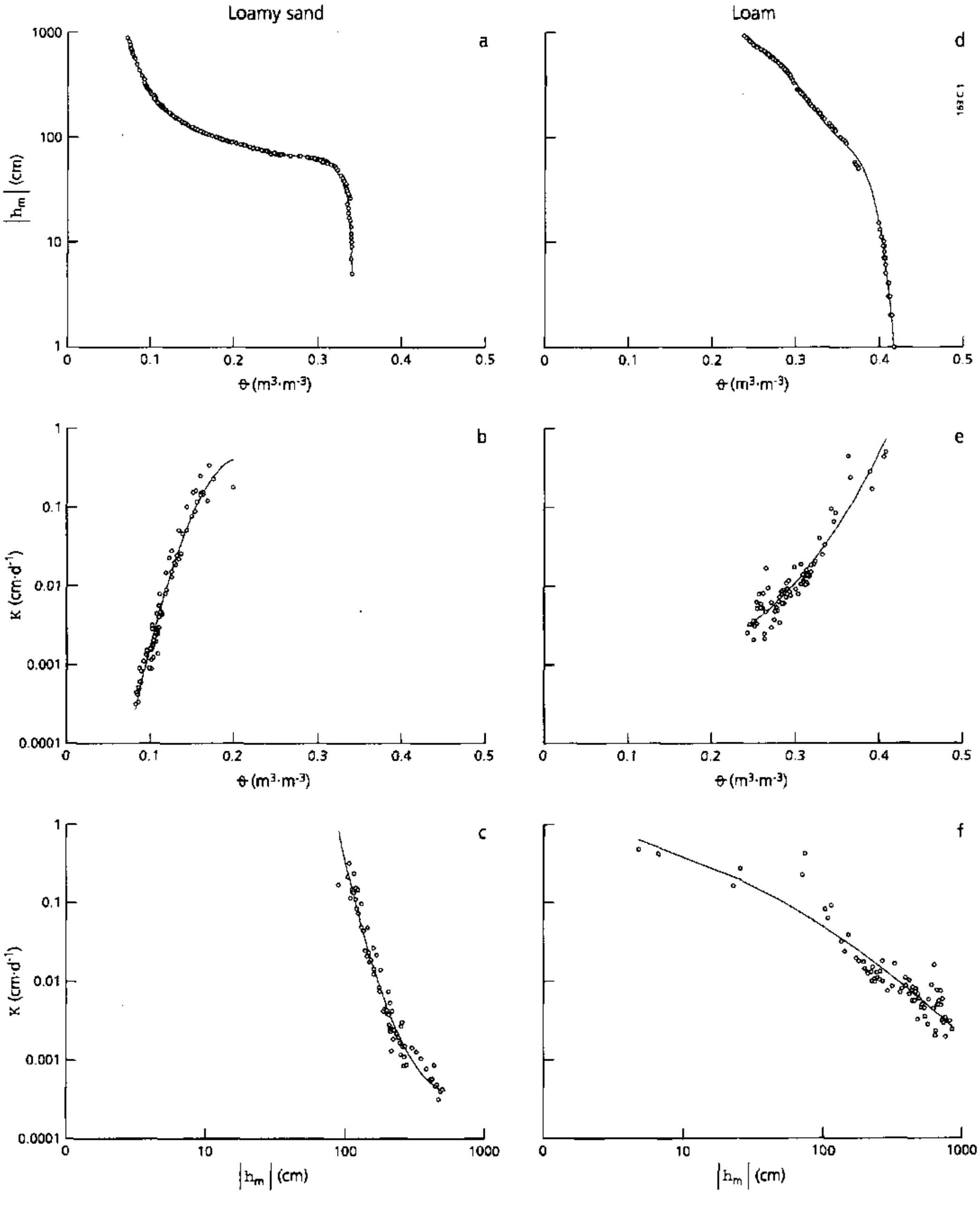

Fig. 3 The water retention characteristics and unsaturated hydraulic conductivities for a loamy sand, (a) - (c), and a loam, (d) - (f), determined with the evaporation method. The water retention characteristics (a and d) were approximated with polynomials. The hydraulic conductivity, K, in cm d^{-1}, has been plotted as a function of the water content, θ, (b and e) and as a function of the absolute value of the matric head, $|h_m|$, (c and f). The clouds of points have been described with second-order polynomials, using multiple linear regression analysis.

Changes in θ, and therefore in $\log(K)$, cause increasingly larger changes in $\log(|h_m|)$ with decreasing θ values.

4 Discussion and conclusions

The evaporation method combined with the new calculation procedure works well, even with rather simple equipment. The water retention characteristic and the unsaturated conductivity can be determined using standard soil-physical equipment: a clock, a scale, tensiometers, a drying oven, and a personal computer for the calculations. The method can be used for all sizes of soil samples. A determination takes between a few days (clay soils) to a few weeks (sandy soils). The assumption must be made that the flow of water can be described with the generalization of Darcy's law. The hydrological reality is simplified by dividing the sample into compartments.

Experiments and simulations showed that the method is relatively accurate. Simulations without 'measurement' noise resulted in differences between 'true' and estimated h_m values of 4% for the water retention characteristic. The differences for the conductivities were 10 to 20%. Adding noise to the simulations did hardly influence the water retention characteristic. The conductivities were clearly influenced, especially at higher conductivities (at low gradients). After proper testing for significance of the gradients, the accuracy of K is bounded by $K/3$ and $3K$ (Tamari, 1992; Tamari *et al.*, 1993). Another advantage of the method is that the experiment is performed under relatively natural conditions, avoiding possible side effects of large temperature or hydraulic gradients.

The method has also disadvantages and limitations. The soil sample must be homogeneous in its hydraulic properties. The method gives only the desorption characteristics, leaving out possible hysteresis effects. The water retention characteristic can be calculated for the whole range of tensiometer measurements. However, the calculation of the hydraulic conductivity is limited to the range with measurable hydraulic gradients. This effect is pronounced for sandy soils, where no conductivities are available for matric heads between 0 and -100 cm. The reason for the absence of measurable gradients is the high hydraulic conductivity of the soil. This is not a problem for loamy and clay soils, since the gradient develops at the start of the experiment. Fortunately stationary methods, such as the crust and sprinkler infiltrometer methods, work well and relatively fast in this region (Dirksen, 1991).

Inverse modelling cannot solve this problem. Any solution with conductivities for the wet range that are higher than those just giving a measurable gradient will work. This may explain why no unique solution could be found for this flow problem (Feddes *et al.*, 1988; Tamari, 1992). Additional data are still needed.

The evaporation method should not be seen as a method for measuring the hydraulic conductivity in the complete tensiometer range. It complements the measurements of the saturated and nearly saturated conductivities.

Determinations close to the investigated areas, even in very remote sites with poor infra-structure, are enabled by the use of battery-operated equipment and relative short measuring times.

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