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SPRINKLING OF GRASSLAND

II. Fundamentals of soil water flow at the experimental field

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C O N T E N T S

1. INTRODUCTION

In 1981 a study was started focussed on water use and production of grassland with different levels of water and nitrogen supply. In the framework of that study field experiments were carried out at the experimental farm Heino during the period 1981-1984. The organisation and set-up of these experiments have been described previously (VAN BOHEEMEN and HUMBERT, 1983).

In this note attention is paid to the fundamentals of soil water flow at the experimental field in Heino. Knowledge of these fundamentals is required for analyzing the soil water measurements done in the field as well as for simulation and prediction purposes.

Chapter 2 gives an equation describing soil water movement in general terms. It contains coefficients defined for characterizing basic hydraulic properties of the soil, viz. water retentivity and hydraulic conductivity.

Chapter 3 shows how the soil profile at the experimental field has been schematized to a system of homogeneous and isotropic layers. The water retentivity of the various layers has been expressed in functions. These functions are given in Chapter 4. The functions expressing the corresponding hydraulic conductivity follow in Chapter 5.

Chapter 6 gives a summary of the note.

2. BASIC FLOW EQUATION

The kinitic energy being of no importance in soil water movement, the discussion can be restricted to potential energy. According to the concepts in the thermodynamics, potential energy is seen here in terms of differences in partial free energy between soil water and water in a reference state.

Force fields generally affecting soil water originate from the attraction of water by the soil matrix and from the presence of hydrostatic pressure, gravitation, external gas pressure, and solutes solved in the soil water. The resulting total potential which characterizes the relative energy status of soil water, can be seen as the sum of the potentials caused by each of the separate force fields. In formula:

$$\psi_t = \psi_p + \psi_g + \psi_a + \psi_o \tag{1}$$

where ψ_t = total potential ($J \cdot kg^{-1}$)

ψ_p = potential resulting from matric forces and hydrostatic pressure ($J \cdot kg^{-1}$)

ψ_g = gravitational potential ($J \cdot kg^{-1}$)

ψ_a = potential resulting from external air pressure ($J \cdot kg^{-1}$)

ψ_o = osmotic potential ($J \cdot kg^{-1}$)

The sum of the pressure and gravitational potentials is called hydraulic potential ψ_h, hence

$$\psi_h = \psi_p + \psi_g \tag{2}$$

The potential is considered here traditionally, namely in terms of energy per unit mass ($J \cdot kg^{-1}$), having the dimension of $L^2 \cdot t^{-2}$.

Starting point in the present study is that no water transport occurs as a consequence of differences in osmotic potential, hence the introduction of the statement:

$$\psi_o = 0 \tag{3}$$

Furthermore, it is assumed that the atmospheric pressure forms the only external pressure and variations in place and time of that pressure are negligible. Accordingly we have for the external air pressure:

$$\psi_a = 0 \tag{4}$$

Here no distinction is made between the potential in the unsaturated zone due to the physical affinity of water to soil-particle surfaces and capillary pores and the potential of water in the saturated zone because of the presence of hydrostatic pressure. This means, that there is a single continuous potential, the so-called pressure potential, prevailing in both the unsaturated and saturated part of the soil profile.

When defining the pressure potential it is usual to take the atmospheric pressure as reference. Because a pressure below the atmospheric is needed to withdraw water from the unsaturated zone, the pressure potential of water in that zone is considered to be negative. Hydrostatic pressure causes that water in the saturated zone has a pressure higher than atmospheric and therefore the corresponding potential is considered to be positive.

The gravitational potential depends on the position of the soil water in the gravitational force field. Here the soil surface is taken as a reference and the height above soil surface as positive. Hence, the formula

$$\psi_g = gz \tag{5}$$

where g = gravitational acceleration ($m \cdot s^{-2}$)

z = depth below soil surface (m)

Adding the expressions for the different partial potentials gives:

$$\psi_t = \psi_h = \psi_p + gz \tag{6}$$

expressing that the hydraulic potential can be seen as the representative for the total potential. Multiplying the different terms by density ρ of water and introducing of new symbols leads to:

$$H = h + \rho g z \tag{7}$$

where H = pressure equivalent of hydraulic potential (Pa)

h = soil water pressure (Pa)

ρ = density of soil water ($kg \cdot m^{-3}$)

Eq. (7) gives the energy status of soil water in terms of energy per unit volume ($J \cdot m^{-3}$ = Pa), having the dimension of $M \cdot L^{-1} \cdot t^{-2}$. The energy status can also be seen in terms of energy per unit weight ($J \cdot N^{-1}$), having the dimension L. The weight equivalent for the hydraulic potential is called the hydraulic head, because it is expressed usually in terms of a head of water. The so-called pF, introduced by SCHOFIELD (1935) to avoid large numbers for the 'head', is the logarithm of the head equivalent for the pressure potential of water in the unsaturated zone, expressed in centimeters water column.

For analyzing flows where only differences in pressure and gravitational potentials play a part and the movement is rather slow and laminar, the following differential equation can be used:

$$q = -\frac{k}{\rho g} \nabla H \qquad (8)$$

where q = flux density (flux) c.q. volume of water passing a unit cross-sectional area per unit time ($m \cdot s^{-1}$)

k = hydraulic conductivity ($m \cdot s^{-1}$)

∇H = gradient of hydraulic potential in the three-dimensional space ($Pa \cdot m^{-1}$)

This equation has the form of the equation derived by SLICHTER (1899) from Darcy's law (DARCY, 1856) for a three-dimensional flow in a saturated porous medium. RICHARDS (1931) proved that Darcy's law can also be used for describing flows in an unsaturated porous medium, if the hydraulic conductivity is treated as a function of soil water pressure.

The occurrence of the product ρg at the right side of the equation relates to the preference to give hydraulic conductivity the same dimension as flux, thus $L \cdot t^{-1}$.

Besides Darcy's law the law of conservation of matter should hold. When expressed in the equation of continuity this gives:

$$\frac{\delta \theta}{\delta t} = -\nabla \cdot q - S \qquad (9)$$

where θ = volume of soil water per unit of soil volume (-)

t = time (s)

S = volume of soil water extracted per unit of time from a unit of soil volume by vegetation, pumping, etc. ($m^3 \cdot s^{-1} \cdot m^{-3} = s^{-1}$)

The expression $\nabla \cdot q$ indicates the so-called divergence q.

Combining (8) and (9) yields:

$$\frac{\delta \theta}{\delta t} = \nabla \cdot \{\frac{k}{\rho g} \nabla H\} - S \qquad (10)$$

Substituting of (7) leads to:

$$\frac{\delta\theta}{\delta t} = \nabla\cdot\{\frac{k}{\rho g} \quad \nabla(h + \rho gz)\} - S \tag{11}$$

Because ∇z is equal to 0 in the horizontal plane and to 1 in the vertical, the last equation can be written as:

$$\frac{\delta\theta}{\delta t} = \nabla\cdot\{\frac{k}{\rho g} \nabla h\} + \frac{\delta k}{\delta z} - S \tag{12}$$

This equation has two dependent variables, namely θ and h, which complicates mathematical treatment. It is more comfortable to start with an equation having only one dependent variable. Hence, the introduction of the term:

$$C = \frac{\delta\theta}{\delta h} \tag{13}$$

where C = differential water capacity (Pa^{-1})

The relationship between soil water pressure h and soil water content θ cannot always be described by a single-valued function due to hysteresis, hence the expression $\frac{\delta\theta}{\delta h}$ instead of $\frac{d\theta}{dh}$.

Substituting eq. (13) into (12) yields an equation with only one dependent variable:

$$C\,\frac{\delta h}{\delta t} = \nabla\cdot\{\frac{k}{\rho g} \nabla h\} + \frac{\delta k}{\delta z} - S \tag{14}$$

With aid of the term hydraulic diffusivity D (CHILDS and COLLIS GEORGE, 1950) being

$$D = \frac{\frac{k}{\rho g}}{C} \tag{15}$$

whereby D is in $m^2 \cdot s^{-1}$, there can also be derived an equation with θ as the only dependent variable;

$$\frac{\delta\theta}{\delta t} = \nabla\cdot\{D\nabla\theta\} + \frac{\delta k}{\delta z} - S \tag{16}$$

Use of the equation with the soil water pressure h as dependent variable has, however, advantages if both saturated and unsaturated situations in the soil are considered (PHILIP, 1958).

The grass at the experimental field in Heino has a root system whereby the distribution pattern at different depths is homogeneous. This means that horizontal soil water movement in the unsaturated zone hardly takes place there and practically only flow in vertical direction is present. Limiting eq. (14) to this specific case and assuming isotropy of the soil lead to:

$$C \frac{\delta h}{\delta t} = \frac{\delta}{\delta z} \left\{\frac{k}{\rho g} \frac{\delta h}{\delta z}\right\} + \frac{\delta k}{\delta z} - S \qquad (17)$$

a second order, non-linear partial differential equation. Such an equation is normally treated with numerical techniques as the finite difference and the finite element techniques. A finite difference technique is easier to program, but not suitable for flow problems with complex geometries.

For studying situations with a rather simple geometry as the case at the experimental field in Heino a finite difference technique is normally preferred. A solution of the flow equation for a specific problem can, however, only be found if the valid boundary conditions, and in cases of unsteady flow the initial conditions too, are known. Another requirement is that for the different layers in the soil profile the interrelationship between soil water pressure, soil water content and hydraulic conductivity has to be known, for example in sets of $h(\theta)$- and $k(\theta)$-functions.

3. PROFILE SCHEMATISATION

The soil profile of the experimental field in Heino has been divided into six homogeneous and isotropic layers (Table 1) on the base of data about texture and dry bulk density (VAN BOHEEMEN and HUMBERT, 1983).

Table 1. Textural characteristics and density of the different layers in the soil profile of the experimental field

Number of soil layer	Depth (cm below surface)	Dry bulk density ($t \cdot m^{-3}$)	Organic matter (weight percents)	Loam (weight percents of total mineral fraction)	Median of sand fraction (μm)
1	0 - 12.5	1.49	6	10	160
2	12.5- 82.5	1.33	6	10	160
3	82.5- 97.5	1.46	2	5	160
4	97.5-160	1.58	0.2	3	175
5	160 -230	1.65	0.2	6	145
6	230 -320	-	0.2	20	150

4. WATER RETENTIVITY OF DIFFERENT SOIL LAYERS

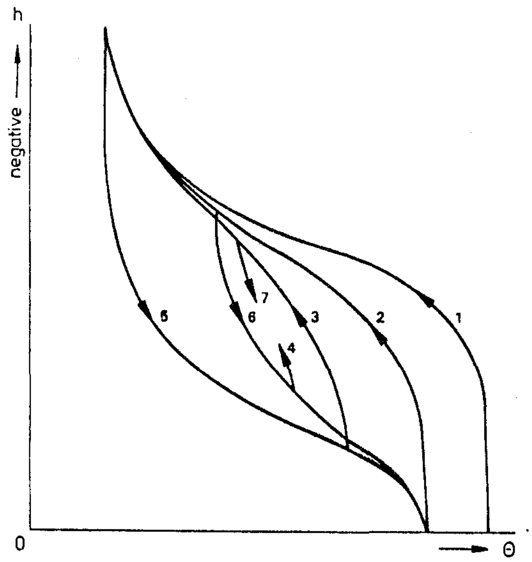
For determining the function $h(\theta)$, normally called soil water characteristic or soil water retention curve, several techniques have been developed. In this chapter a review of these techniques is given, followed by a presentation of the results obtained with them. Before that the phenomenon hysteresis is discussed as far as it affects the $h(\theta)$-relation.

4.1. Hysteresis

4.1.1. General

The soil water content corresponding with a certain water pressure depends on the previous history of the soil. In case the equilibrium is reached after wetting higher values are found than in the situation preceeded by a drying process. This phenomenon, already studied by HAINES (1930), is called (capillary) hysteresis.

The effect of an alternate wetting and drying process on the soil water characteristic can be described schematically as shown in Fig. 1. Curve 1, here called the 'first drying curve', represents the relation in case of drying of the soil after a complete saturation. The lower part of the first drying curve has a vertical direction till the point



1 first drying curve
2 main drying curve
3 primary drying scanning curve
4 secundary drying scanning curve
5 main wetting curve
6 primary wetting scanning curve
7 secundary wetting scanning curve

Fig. 1. Hysteretic relationships between soil water pressure h and soil water content θ

where the soil water pressure becomes equal to the so-called air entry value.

If wetting occurs at the moment the water content has a very low value, a relation like curve 5 is found which coincides in the beginning the first drying curve, but later diverges. This relation is called the main wetting curve. The water content of the soil found at a soil water pressure equal to zero can be lower than that at complete saturation, because, as a consequence of entrapped air, it takes some time before all pores have been filled with water.

A drying process following this wetting will give a relation like curve 2, called the main drying curve. This curve together with the main wetting curve form the main hysteretic loop.

When the main drying process changes into wetting, one of the primary wetting scanning curves is obtained (curve 6). From a primary wetting scanning curve a secundary drying scanning curve can depart (curve 4).

The processes underlying hysteresis are not yet understood completely. Here the most important aspects are mentioned. A water molecule has no net charge, but, as a consequence of the specific arrangement of the electrons belonging to the hydrogen and oxygen atoms, it acts like an electrical dipole. A hydrogen atom of a water molecule can be bond strongly to an oxygen atom of an adjacent water molecule (resulting in

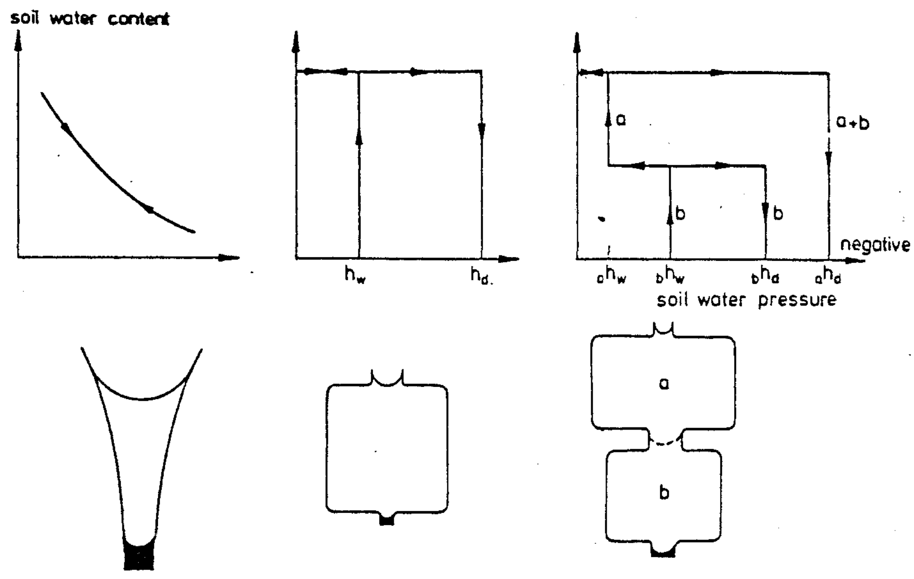


Fig. 2. Hysteretic effect of some pore configurations on water retention curve (after POULOVASSILIS, 1962)

cohesion) but it can also be attracted by an oxygen atom at the surface of a soil particle (resulting in adhesion). Through the combined effect of adhesion and cohesion forces, water enters or leaves a soil pore till the radius of the curvature at the air-water interface corresponds with the prevailing physical conditions. The interrelationship between the factors determining the equilibrium can be described as:

$$h = \frac{2\sigma \cos \alpha}{r} \tag{18}$$

where h = soil water pressure (Pa)

σ = surface tension of the soil water ($Pa \cdot m^{-1}$)

α = contact angle, angle between the air-water interface and the solid-water interface (-)

r = radius of the curvature at the air-water interface (m)

The phenomenon mentioned above causes a part of the hysteresis in the $h(\theta)$-relation. Fig. 2 shows schematically its effect for a few situations occurring in a soil.

The contact angle α, the angle between the air-water and the solid-water interfaces, also shows hysteresis. For pure water on clean and smooth, inorganic soil surfaces, the angle α is generally zero, but where the surface is rough or coated with organic substances having a low content of oxygen atoms, the contact angle can be considerably great and further time-dependent (DE BANO, 1983). A relative large value is found at sites where a dry soil surface with a water repellent character is wetted. When the same soil water pressure is found there after a drying process, the contact angle may be lower, because in a wet situation the difficult soluable substances may have been solved. The higher soil water contents corresponding with the latter case can also been caused by the occurrence of bridges between easily wettable sites whereby water transport in a vapor phase has played a part (DE BANO, 1983).

Other factors causing hysteresis in the h(θ)-relation are re-arrangement of soil particles during wetting and drying, and entrapping of air occurring if entrances of a pore have been closed by the wetting fluid. Entrapped air can only disappear by diffusion.

4.1.2. Magnitude

A lot of experiments for showing hysteresis in the soil water characteristic have been done with cores consisting of

- glass beads and uniform sands (POULOVASSILIS, 1962; TOPP and MILLER, 1966; LEES and WATSON, 1975);
- repacked natural soil. Repacking often took place after drying, crushing and screening (JACKSON et al., 1965; STAPLE, 1966; CARY, 1967; TOPP, 1969; STAPLE, 1969; POULOVASSILIS, 1970; TALSMA, 1970; POULOVASSILIS and CHILDS, 1971; TOPP, 1971a; VACHAUD and THONY, 1971; CARY, 1975; GILLHAM et al., 1976; CLAUSNITZER, 1978).

Fig. 3 gives results obtained by STAPLE (1966) and CARY (1975) with repacked natural soils. Staple found that about 7 percent of the soil volume was not saturated during rewetting because of entrapped air.

With regard to hysteresis in undisturbed natural soils minor information is, however, available. ILNICKI (1982a,b) did laboratory experiments with undisturbed samples of humous sand and peat. He saturated the samples, lowered the soil water pressure until -10 kPa and then increased the soil water pressure again until zero. For each sample the maximum difference in soil water content during drying and wetting was determined, further the corresponding soil water pressure (Table 2).

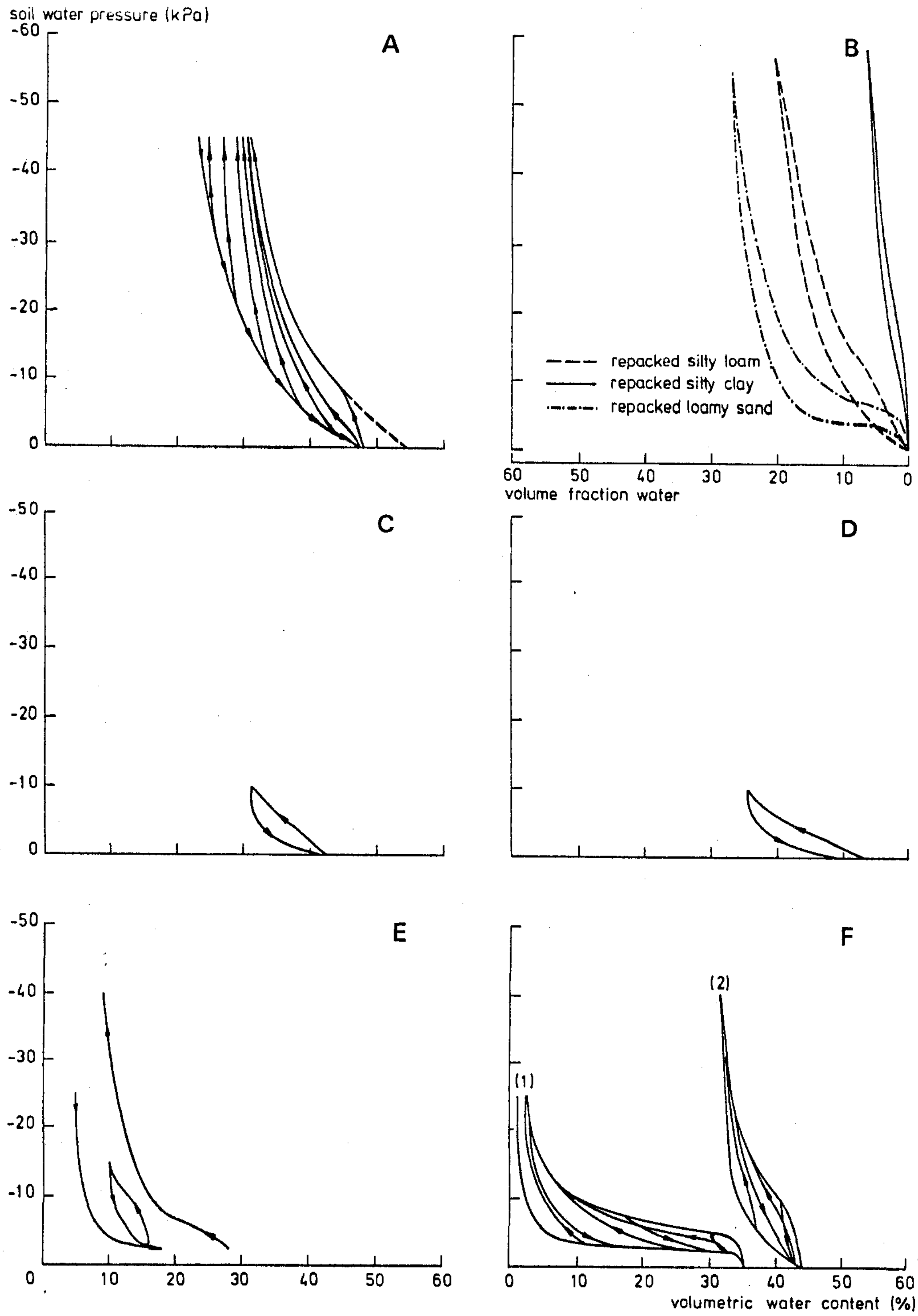


Fig. 3. Hysteretic water retention curves found in literature. A. repacked silt loam, after STAPLE (1966); B. after CARY (1975); C. undisturbed sand with 5 percent organic matter and density of 1.47 $t \cdot m^{-3}$, after ILNICKI (1982); D. ibid with 9 percent organic matter and density of 1.14 $t \cdot m^{-3}$, after ILNICKI (1982); E. in situ fine sand, after ROYER and VACHAUD (1975); F. in situ loess (1) and gravelly sand (2), after RENGER et al. (1974)

Table 2. Maximum difference in soil water content at the same soil water pressure, due to hysteresis (after ILNICKI, 1982a)

Soil	Organic matter content (weight %)	Number of cores	Dry bulk density ($t \cdot m^{-3}$)	Maximum difference in soil water content (vol. %)	Corresponding soil water pressure (kPa)
Sand	7	2	1.31	6.2	-2 to -5
Sandy peat	31	2	0.71	3.5	-3 to -6
Peat	43	3	0.54	2.4	-2 to -5
Peat	86	7	0.33	5.2	-2 to -5
Peat	93	6	1.29	7.6	-1 to -3

Fig. 3 shows the results obtained for two kinds of humous sand. Ilnicki noted that a further lowering of the soil water pressure below -10 kPa lead to a greater maximum difference in soil water content.

Ilnicki has also summarized the results of experiments occurring in 15 publications (Table 3). It concerns mainly experiments with uniform sand and repacked natural soils, whereby the soil water pressure lowered until -25 to -100 kPa. The maximum differences in soil water content found for sand varied strongly (5 to 26 volume percents), the values of the corresponding soil water pressures had only a small variation (-3 to -5 kPa).

BARANOWSKI and PABIN (1975) presented hysteretic effects determined on undisturbed cores of medium sand and loamy black earth, TOPP and ZEBCHUK (1979) gave results for clay and sandy loam cores.

ROYER and VACHAUD (1975) reported on in situ measurements in fine sand and chalked clay. Fig. 3 contains the results for the fine sand. In situ measurements were also done in clay loam by WATSON et al. (1975) and in loess (silty loam) and underlying gravelly sand by RENGER et al. (1974). The results of the latter work are given in Fig. 3 too.

Laboratory measurements on an undisturbed soil monolith have been described by BEESE and VAN DER PLOEG (1976) and TZIMAS (1979). Beese and Van der Ploeg took a monolith from a grey brown podzolic soil, Tzimas from so-called lower greensand occurring near Cambridge (England).

In the summary made by Ilnicki the results mentioned in the last six publications except those of Royer and Vachaud have not been incorporated.

Table 3. Maximum difference in soil water content at the same soil water pressure, due to hysteresis. Data derived from literature by ILNICKI (1982a)

Soil	Number of cores	Average dry bulk density ($t \cdot m^{-3}$)	Maximum difference in soil water content (vol. %) average	variation	Corresponding soil water pressure (kPa)
Sand	13	1.69	13.4	4.8 to 26.5	- 3 to - 5
Silty loam	4	1.19	5.6	4.0 to 7.2	-13.5 to -17.5
Loam	2	-	2.2	2.0 to 2.5	-23 to -35
Clayey loam, Clay	3	-	4.0	3.0 to 6.0	- 5.5 to -14
Loess	3	1.42	5.7	1.5 to 12.0	- 5 to -10
Humous sand	3	1.35	6.0	2.6 to 7.8	- 2.5 to - 7.5
Peat	1	-	10.0	-	- 1 to -10

4.1.3. Models

In the past twenty years several models have been developed to calculate hysteretic effects. An important step was the application of the independent domain theory to soil physics by POULOVASSILIS (1962). About a decade earlier, using Néel's diagram (NEEL, 1942) the independent domain theory had been developed for describing generally hysteretic processes (EVERETT and WHITTON, 1952; EVERETT and SMITH, 1954; EVERETT, 1954, 1955; ENDERBY, 1955).

Poulovassilis proposed to divide the total amount of water draining out and re-entering the soil when the main hysteresis loop is followed, into small elements of which each is completely specified by a pair of small water pressure ranges, e.g.

- $\delta h_d(i)$ with $h_d(i)$ as average, representing the water pressure range over which the element drains out of the soil during a drying process;
- $\delta h_w(j)$ with $h_w(j)$ as average, representing the water pressure range over which the element re-enters the soil during a wetting process.

The existence of the elements was expressed in a domain diagram as occurring in the horizontal h_d,h_w-plane of Fig. 4. All elements can

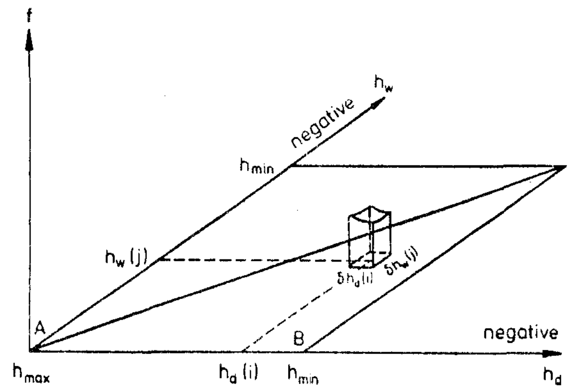


Fig. 4. Graphical presentation of volume f corresponding to the soil water element draining out of the soil at water pressure change $\delta h_d(i)$ and re-entering at water pressure change $\delta h_w(j)$ (after POULOVASSILIS, 1962)

be placed in the triangle ABC because for no element $h_d(i)$ exceeds $h_w(j)$. The terms h_{max} and h_{min} noted along the h_d- and h_w-axes are the water pressures at the two ends of the main hysteresis loop. The elements indicated on the diagonal AC are those which drain out and re-enter at one and the same water pressure range.

The vertical coordinate f has been introduced for indicating the volumes of water corresponding with the elements in the domain diagram. This means that

$$\delta V = f(h_d,h_w)\delta h_d \delta h_w \tag{19}$$

where δV = volume of the element

f = distribution function

h_w = water pressure when the element drains out

h_d = water pressure when the element re-enters

Dependent on the hysteresis in the soil the function $f(h_d,h_w)$ varies from point to point of the triangle ABC, while it is zero outside it.

Fig. 5 shows schematically the processes occurring when the main wetting and main drying curves are followed. According to Fig. 5C the wetting process can be seen as a re-entering of the elements with h_d,h_w-values passed by the line moved in the h_d,h_w-plane parallel and towards to the h_d-axis. The amount of water re-entering when the water pressure is increased from h_{min} to $h_w(k)$ equals to:

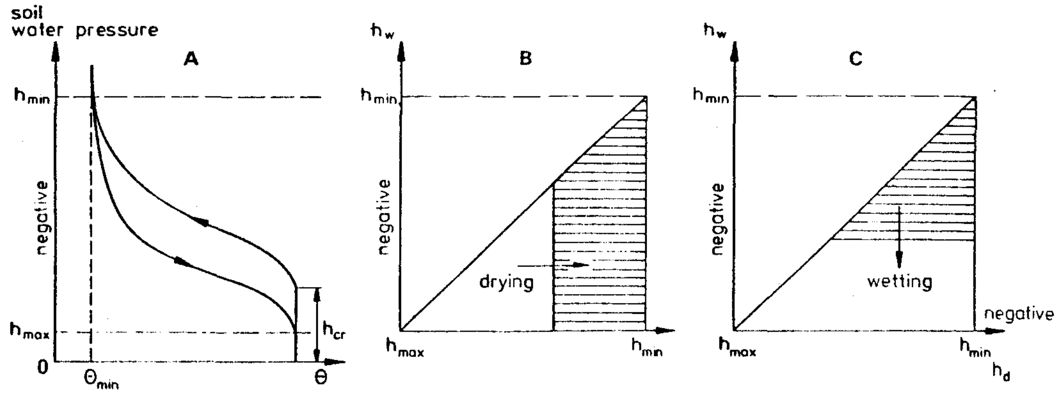


Fig. 5. A. a main drying and main wetting curve, B. main drying process; the elements with h_d,h_w-values in the arced region will be drained out, C. main wetting process; the elements with h_d,h_w-values in the arced region have been re-entered

$$\int_{h_{min}}^{h_w(k)} \int_{h_{min}}^{h_{max}} f(h_d,h_w)\,\delta h_d \delta h_w \tag{20}$$

The processes occurring when primary scanning curves are followed, have been described schematically in Fig. 6.

The domain diagram presented in Fig. 7 has been derived by analysis of a main drying curve and a set of wetting scanning curves determined

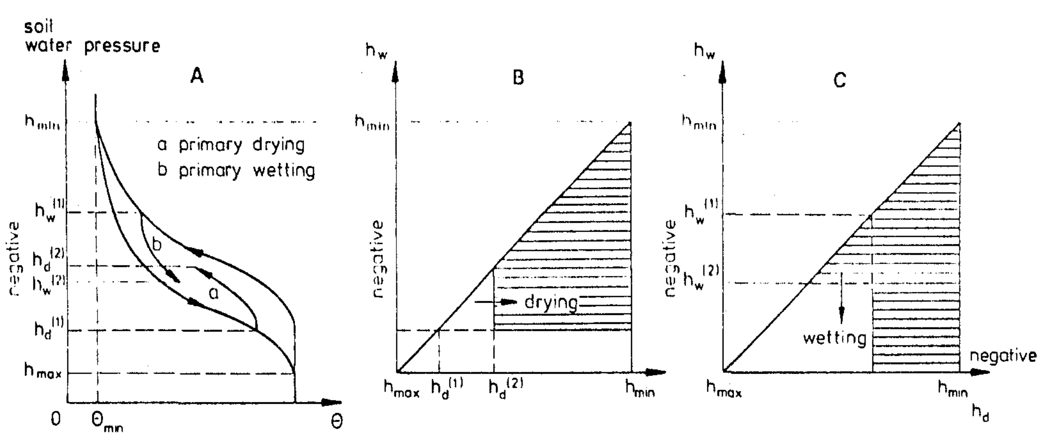


Fig. 6. A. a primary drying and primary wetting scanning curve, B. primary drying process; the elements with h_d,h_w-values in the arced region will be drained out, C. primary wetting process; the elements with h_d,h_w-values in the arced region have been re-entered

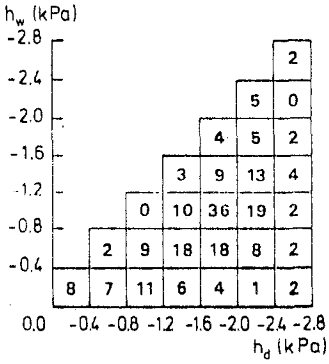


Fig. 7. Domain diagram determined for a glass-bead structure, showing the volumes of the elements (cm^3) draining out and re-entering the medium of different water pressure ranges within the main hysteresis loop (after POULOVASSILIS, 1962)

for a glass-bead structure. The numbers in the diagram indicate the volumes of the different elements. With such a diagram it is possible to predict the state of the concerning medium after any series of changes in water pressure. The diagram shows, for example, that the water content of the medium increases with 0 + 10 + 36 + 19 + 2 = 67 cm^3 when during the main wetting process the water pressure increases from -1.2 to -0.8 kPa.

The results obtained with the above theory varied. In some cases a good agreement was found between calculated and measured curves (POULOVASSILIS, 1970; TALSMA, 1970). However, several applications delivered a poor agreement (TOPP and MILLER, 1966; TOPP, 1969, 1971a; VACHAUD and THONY, 1971). The failures of the independent domain theory are attributed to the fact that it does not account for the fact that drainage of pores can be dependent on the state of neighbouring pores, in the sense that they are not always accessible to air during drainage at their characteristic values. This is severe if soils are considered whereby a major portion of the hysteretic loop is in the range of soil water pressures higher than air entry value.

By introducing the similarity hypothesis the applicability of the Néel diagram could be simplified and the agreement between measured and calculated curves even improved (PHILIP, 1964; MUALEM, 1973). In the approach of Mualem the distribution function $f(h_d,h_w)$ is assumed to equal the product of two independent functions:

$$f(h_d,h_w) = p(h_d)\ q(h_w) \qquad \text{if } h_w > h_d \tag{21}$$

According to this hypothesis the course of the function f along each line parallel to the h_d-axis is identical apart from a constant factor $q(h_w(j))$. The same is the case along each line parallel to the h_w-axis.

The similarity hypothesis permits to derive in a relative simple way any hysteretic path inside the main loop for soils of which only the two main curves are available. Later (MUALEM, 1977) the similarity hypothesis was extended to equality between $p(h_d)$ and $q(h_w)$, leading to:

$$f(h_d,h_w) = p(h_d)\ p(h_w) \qquad \text{if } h_w > h_d \tag{22}$$

This means that, if only one branche of the main hysteresis loop has been measured, predicting of the scanning curves can be performed.

A further improvement of the independent domain theory could be reached by using of a variant of the Néel diagram, further called the Mualem diagram (MUALEM, 1974, 1977). However, the improvements in modeling did not lead to satisfactoring results for soils with an important part of the hysteretic loop in the range of soil water pressures higher than air entry value.

Therefore, several efforts have been made to eliminate the assumption of independence, resulting in the development of the dependent domain theory (EVERETT, 1967; TOPP, 1971b; POULOVASSILIS and CHILDS, 1971; POULOVASSILIS and EL GHAMRY, 1978; MUALEM and DAGAN, 1975; MUALEM, 1976b, 1977; MUALEM and MILLER, 1979). Everett introduced a weighting function in order to account for blockage of access to air, representing the ratio between the volume of pores actually emptied and the volume which could have been emptied if all the pores were independent, i.e. guaranteed access to air from the neighbouring pores. Topp extended the theory of Everett by defining a second weighting function for the wetting process as well, thus for blockage against water entry. Mualem simplified these models by using the similarity hypothesis. In this way he developed models allowing prediction of all scanning curves for soils from which, besides the main curves, one drying and one wetting scanning curve (required for characterizing of blockage to air and water entry respectively) are available.

Application of the models discussed above requires availability of one or more experimental curves. NAKANO (1976a,b, 1980) went a quite different way. He defined a theoretical pore volume distribution function according to probability theory, proposed a new soil-pore model and presented a method for predicting theoretically the first drying curve and the two main curves. As far as known the work has not (yet) been extended for predicting scanning curves.

Some of the factors underlying hysteresis are time-dependent. This is often not considered in measuring and modeling of hysteresis effects and tacitly ranked as a second order problem. The same happens with the indications that the h(θ)-relationship depends on the state of flow, thus the h(θ)-relationship for static, steady-state and unsteady-state situations are different (TOPP, KLUTE and PETERS, 1967; VACHAUD, VAUCLIN and WAKIL, 1972; ESFALTAWAY and MANSELL, 1975).

4.2. Determination

4.2.1. Available techniques

Usually the water retention curve of a soil is determined on laboratory samples by controlling the pressure deficit across air-water interfaces and allowing the water content to adjust until an equilibrium is obtained. Thereby, the water content is measured for each pressure difference created, either by measuring the water outflow from the samples or by weighing of the sample. Finally the sample is dried at 105°C and weighed for determining the absolute values of the water content in the different measuring steps.

Two types of devices can be distinguished, i.e. pressure cells and tension plates. The pressure cell method, first described by GARDNER et al. (1922), involves placing a soil sample in contact with an artificial porous medium that has pore sizes small enough to remain completely filled with water when a substantial pressure difference is imposed across the air-water interfaces of the medium. That pressure difference is usually controlled by increasing or decreasing the air pressure in a sample chamber on one side of the barrier and allowing the water in the barrier to remain at or near atmospheric pressure.

The tension plate method, popularized by HAINES (1930), forms a variation on the pressure cell method. Here, the air in the sample chamber remains at atmospheric pressure, while the water pressure in the barrier is reduced below atmospheric pressure by means of an outflow siphon (maximum reduction 100 kPa).

For determining soil water contents at soil water pressures below -3 MPa, whereby water is mainly transported in vapour phase, samples can be brought into chambers with a body of a low water potential (vapor pressure technique). The water potential occurring after reaching an equilibrium is then computed from the relative humidity of the air in the chamber, as proposed by SCHOFIELD (1935) using the theory of thermo-dynamics. The corresponding water content of the soil is obtained gravimetrically.

A quite other approach, developed primarily for field application, consists of simultaneously measuring soil water pressures by tensiometers or other devices like thermocouple-psychrometers, installed at different depths in the soil, and water contents, determined either gravimetrically or by neutron scattering c.q. gamma ray absorption techniques. A similar procedure has been developed for application in the laboratory on undisturbed soil columns. In the latter situation changes of the water content in the soil columns can also be measured by weighing of the total samples. Different levels of soil water content are created then by a drying process in the form of evaporation or a draining process of the soil columns.

Especially for studies spanning large areas, prediction techniques of the soil water characteristic have been developed, mainly by relating data on textural and structural properties of the soil to data on water retentivity obtained with the techniques mentioned above. The simplest one consists of equations giving water contents at specified soil water pressures as a function of soil texture, organic matter content and dry bulk density (JAMAISON and KROTH, 1958; SALTER et al., 1966; HUSZ, 1967; HALL et al., 1977; GUPTA and LARSON, 1979; GOSH, 1980; DE JONG et al., 1983). For soils occurring in the Netherlands similar equations have been derived; for clay soils by POELMAN and VAN EGMOND (1979) and for sandy soils by KRABBENBORG et al. (1983).

Also an approach exists based on the use of a power curve with parameters for which values have been found empirically (RUBIN et al., 1964; BROOKS and COREY, 1964; KING, 1965; VISSER, 1966; ROGOWSKI, 1971; 1972; FARELL and LARSON, 1972; FINCK and JACKSON, 1973; CLAPP and HORNBERGER, 1978; VAN GENUCHTEN, 1980).

More advanced techniques are developed by NAKANO (1976a,b, 1980) as discussed in Par. 4.1.3 as well as by ARYA and PARIS (1981) and d'HOLLANDER (1979). These authors based their concepts on the thought that the soil water characteristic is essentially a pore size distribution function. Therefore efforts have been made in finding a pore volume and a representative pore radius corresponding to each pore size c.q. particle size fraction. This means a more physical approach in order to account for the effects of texture and packing characteristic of the soil. Here the work done by STUYT (1982) can also be mentioned. He proposed to fix the soil water characteristic by numerical simulation of the desorption of a porous medium derived from data about the particle size distribution of the concerning soil with the aid of the probability theory.

4.2.2. Results

All measuring techniques mentioned for determination of $h(\theta)$-relations have been applied, except the vapour pressure technique because the corresponding measuring range is limited to soil water pressures below -3 MPa. Furthermore, the prediction method developed by KRABBENBORG et al. (1983) for Dutch sandy soils has been used.

Pressure cell technique

The pressure cell technique has been used for determination of water contents at soil water pressures of -250 and -1600 kPa. The soil samples were saturated before placing in the cells. The measuring procedure followed here was described by STAKMAN et al. (1969a). Table 4 gives the final results being averages of values obtained for the different samples from the same soil layer.

KRABBENBORG et al. (1983) related a large number of pressure cell measurements for sandy soils to textural characteristics by applying multiple regression. They used the following model:

$$\theta(h) = b_0 + \sum_{i=1}^{4} b_i x_i + e \qquad (23)$$

where θ = volumetric soil water content

h = soil water pressure

x_1 = organic matter content of the soil

x_2 = fraction particles <50 μ in the soil

x_3 = median of the fraction particles >50 μ

x_4 = reciproke of the dry bulk density

$b_0 \ldots b_4$ = regression coefficients

e = stochastic variable with a normal distribution

Table 4. Water content (volume percents) of different soil layers at the experimental field at water pressures of -250 and -1600 kPa, measured in pressure cells and predicted according to KRABBENBORG et al. (1983)

Soil layer	1	2	3	4
Depth (cm -surface)	0-12.5	12.5-85.5	82.5-97.5	97.5-160
Number of soil samples	3	6	3	5
Water content at -250 kPa				
measured	11.9	11.4	5.4	1.9
predicted	12.0	10.9	6.6	1.2
Water content at -1600 kPa				
measured	9.7	8.6	4.2	1.3
predicted	7.6	7.1	3.8	0.9

Table 4 also contains soil water contents obtained with equations given by Krabbenborg et al. for soil layers distinguished at the experimental field. Nearly all differences between the predicted and measured values are smaller than the standard deviations found in deriving the regression equations. This means that the measured values correspond with those gathered for soils of the same type elsewhere in the Netherlands.

Tension plate technique

The tension plate technique was applied on soil samples in Kopecky-rings (100 cm^3). These samples were taken in vertical direction; till 70 cm below surface two per 5 cm depth and below that level two per 10 cm depth. In the laboratory the samples were saturated from the lower side and then placed on tension plates. Water contents were determined at soil water pressures equal to -0.3, -1.0, -3.2, -6.3, -10, -20 and -50 kPa. The procedure followed during these experiments is according to STAKMAN et al. (1969b).

Fig. 8 illustrates the variation in the results obtained for the layers 12.5-82.5 and 97.5-160 cm below surface. The broken lines indicate the porosity of the samples, the plotted points the water contents at -0.3, -6.3 and -50 kPa respectively. The water contents at water

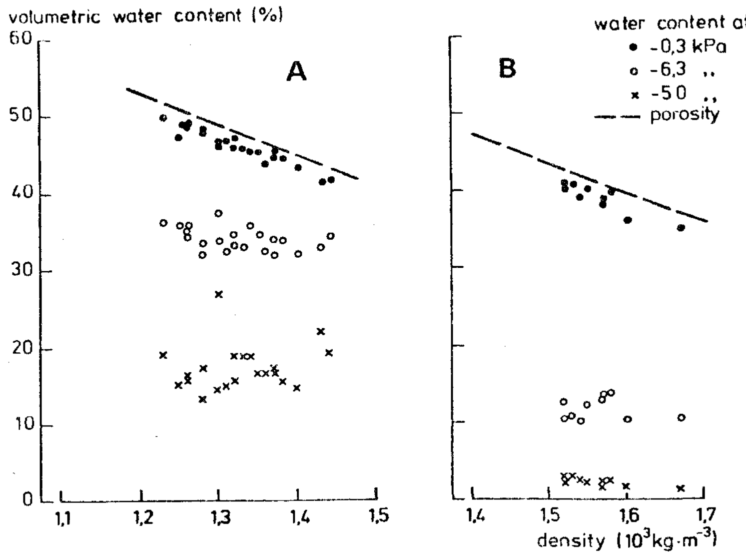


Fig. 8. Results of tension plate measurements on initially saturated samples from the soil layers 12.5-82.5 (A) and 97.5-160 (B) cm below surface, related to the dry bulk density of the samples

pressures higher than -6.3 kPa (pF = 1.8) proved to be correlated negatively with the dry bulk density.

The water retention curves derived from the tension plate and pressure cell measurements are given in Fig. 9. The points corresponding

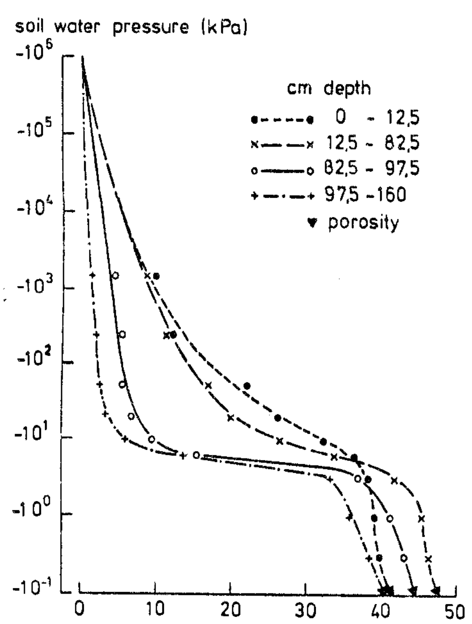


Fig. 9. Water retention curves derived from pressure cell and tension plate measurements on initially saturated soil samples from the experimental field

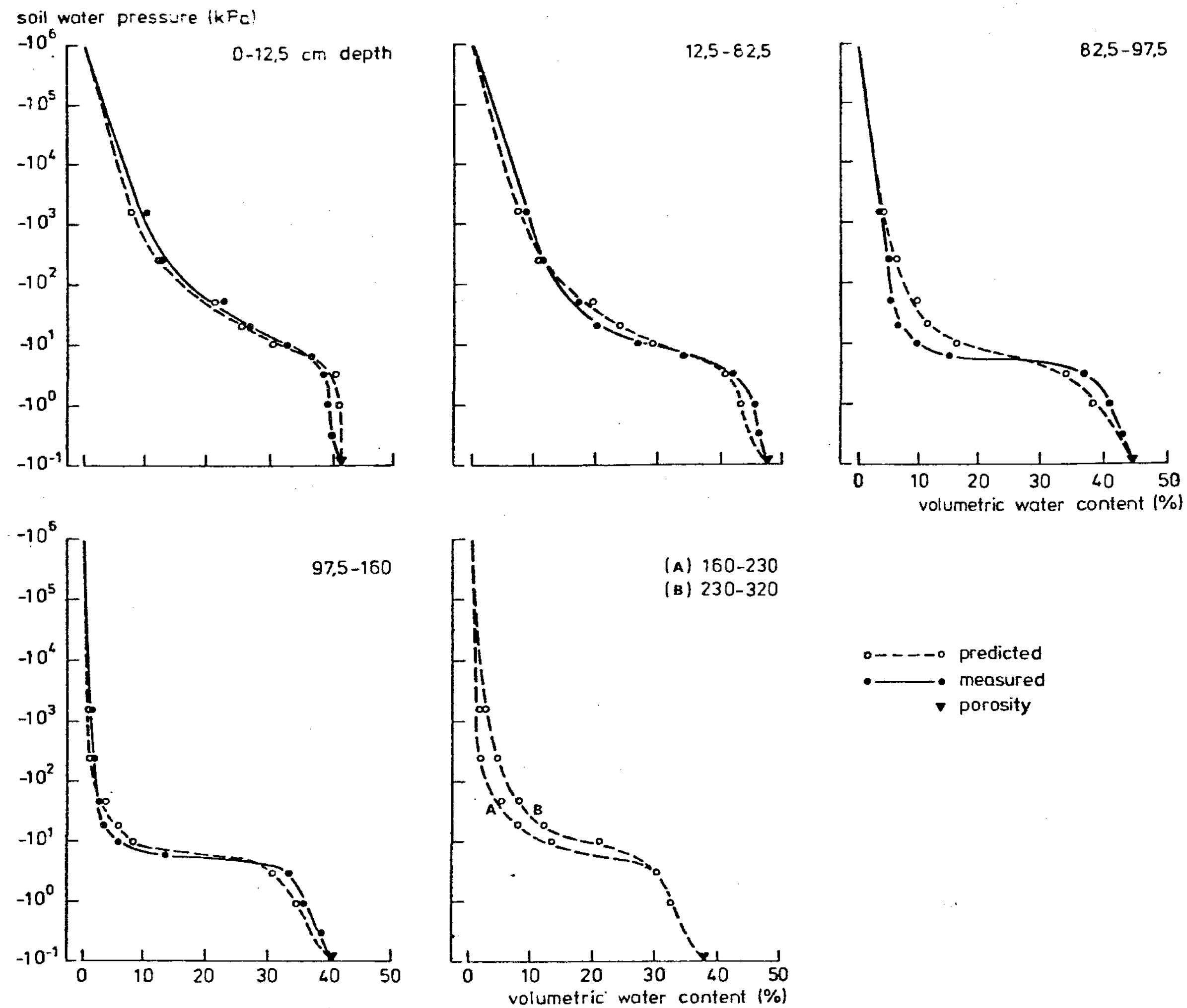


Fig. 10. Predicted and measured results of pressure cell and tension plate methods on initially saturated soil samples from the experimental field

with soil water presures higher than -100 kPa represent values measured on soil samples with the same dry bulk density as the average dry bulk density of the concerning layers. In constructing the curves, the assumption was made that after the saturation process the water content of the samples was equal to porosity of the soil. This is reasonable beaause the water content measured at water pressure -0.3 kPa was about 1.6 volumetric percent lower than porosity. The lines presented in Fig. 9 have been classified as first drying curves.

Besides results of pressure cell measurements, KRABBENBORG et al. (1983) also related results of tension plate measurements to textural characteristics. The water retention curves obtained with the prediction method of Krabbenborg et al. are given in Fig. 10, together with the

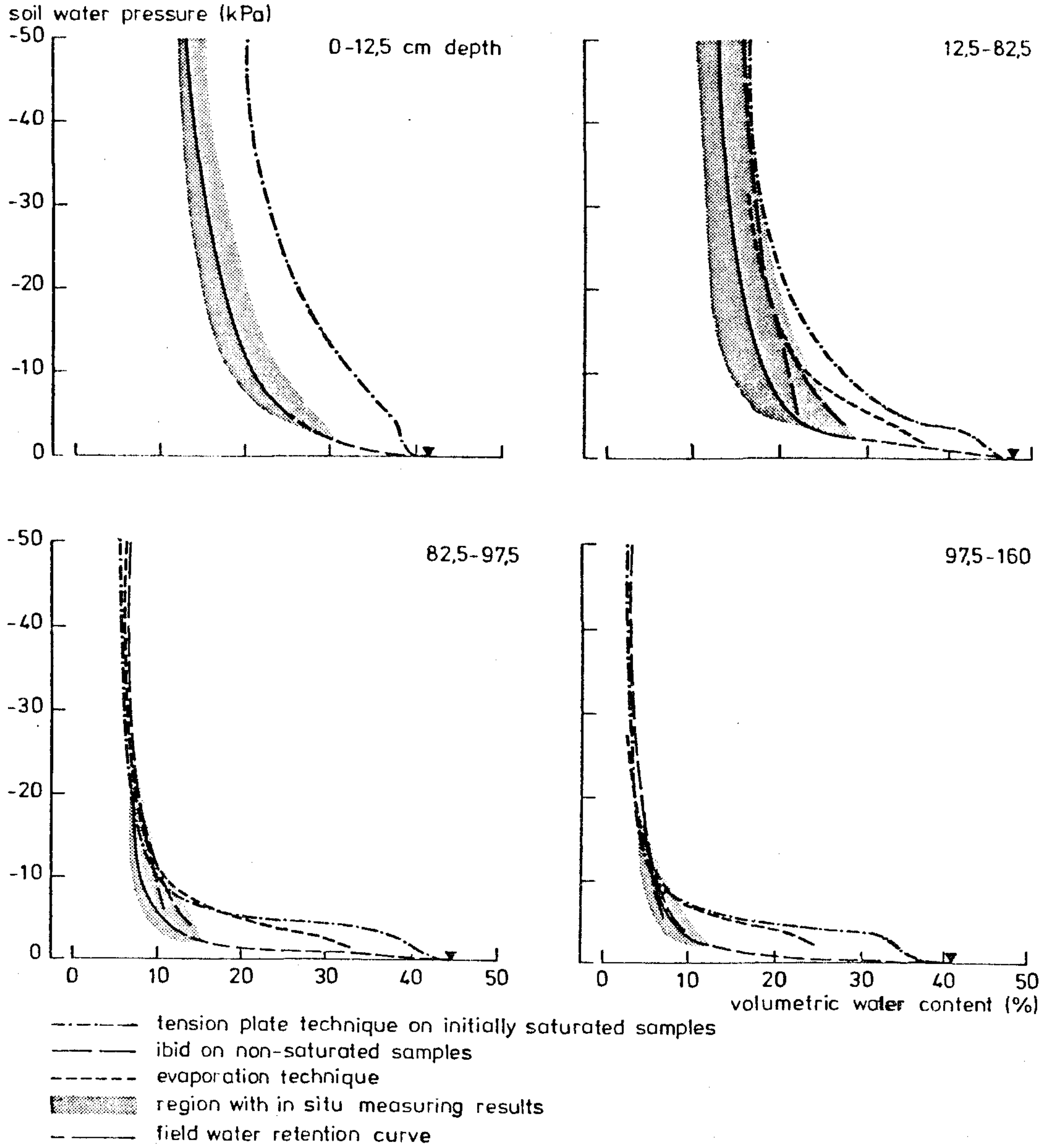


Fig. 11. Water retentivity of different soil layers at the experimental field, measured in situ and in the laboratory

curves presented earlier in Fig. 9. In general, the agreement is well. This indicates that the water retentivity of the soil layers at the experimental field, if measured on initially saturated samples, show no special features. Differences between the calculated and measured values which are significantly larger than the standard deviations found in deriving the regression equations, were only obtained for the layer 82.5-97.5 cm below surface at soil water pressures -10, -20 and -50 kPa. This is assumed to originate from the relative large spatial variability in the layer, the B2-horizon, that affected the accuracy of both the measurements and the predictions.

In order to obtain an insight in the possible presence of hysteresis, tension plate measurements were performed on soil samples which were not pre-wetted. For this purpose two times soil samples were taken in the field when relative high water pressures prevailed. The tensions installed at the start of the measurements corresponded with the soil water pressures measured in the field during sampling. Fig. 11 gives water retention curves derived from the latter measurements. They represent the results obtained on soil samples with a dry bulk density equal to the average dry bulk density of the concerning layers. The lower parts of the curves lie below the first drying curves taken from Fig. 9, the upper parts approach them asymptotically. The obtained curves are therefore considered as drying scanning curves. Those obtained for samples from the layers 82.5-97.5 and 97.5-160 cm below surface approach the corresponding first drying curves at soil water pressures of about -10 kPa. For samples from 12.5-82.5 cm depth this is the case at lower values (about -50 kPa). This is attributed to a stronger drying of the layer 12.5-82.5 cm below surface during dry periods.

After the last drying step on the tension plate was completed, the samples were saturated and then the determinations were repeated. Then $h(\theta)$-relations were obtained which proved to be equal to those found for soil samples saturated immediately after arrival in the laboratory.

So, the water content of the samples at the start of the measurements affected largely the measuring results. This indicates, that the phenomenon hysteresis is of importance.

Evaporation technique

Fig. 11 also shows the $h(\theta)$-relations derived from measurements on undisturbed soil samples according to the evaporation technique described by BOELS et al. (1978) and BEUVING (1982). The soil of the experimental field was sampled in duplo at 30-38, 55-63, 85-93, 115-123 and 140-148 cm depth. Thereby cylinders with a height of 8 cm and a diameter of 10 cm were used.

After wetting in a water bath the lower sides of the cylinders were closed and evaporation from the upper sides was admitted. Periodically the cylinders were weighted and soil water pressures were measured at different depths in the cylinders with tensiometers. Processing of the measuring data gave a $h(\theta)$-relation for each sample. The relations obtained for the different samples from the layer 12.5-82.5 cm below surface hardly differed. The same was the case for the samples from 82.5-97.5 and 97.5-160 cm depth.

At high water pressures the obtained relations have a different course than the corresponding first drying curves (Fig. 11). This is related to the fact that the samples were not saturated in the water bath. At the start of the measurements the water content of the samples from 12.5-82.5 cm depth was about 7 volume percent lower than the calculated porosity. For the samples from 82.5-97.5 and 97.5-160 cm depth the differences were 10 and 15 volume percents, respectively. The curves obtained with the evaporation technique are therefore considered to be drying scanning curves.

The curves for the layers 82.5-97.5 and 97.5-160 cm below surface approach the first drying curves at soil water pressures of about -10 kPa, those corresponding with 12.5-82.5 cm depth at -50 kPa. In discussing the scanning curves derived from the tension plate measurements a comparable phenomenon was mentioned.

In situ technique

In the experimental field a large number of gravimetric determinations of soil water content have been done, accompanied by measurements of soil water pressure with tensiometers. The results of both types of determinations have been plotted against each other in Fig. 12. The open points pertain to water content determinations in soil sampled in Kopecky rings so that the volumetric water content could be determined exactly. The full points pertain to determinations for which soil samples with unknown volumes were taken and whereby the volumetric water contents were found by multiplying the water contents in weight

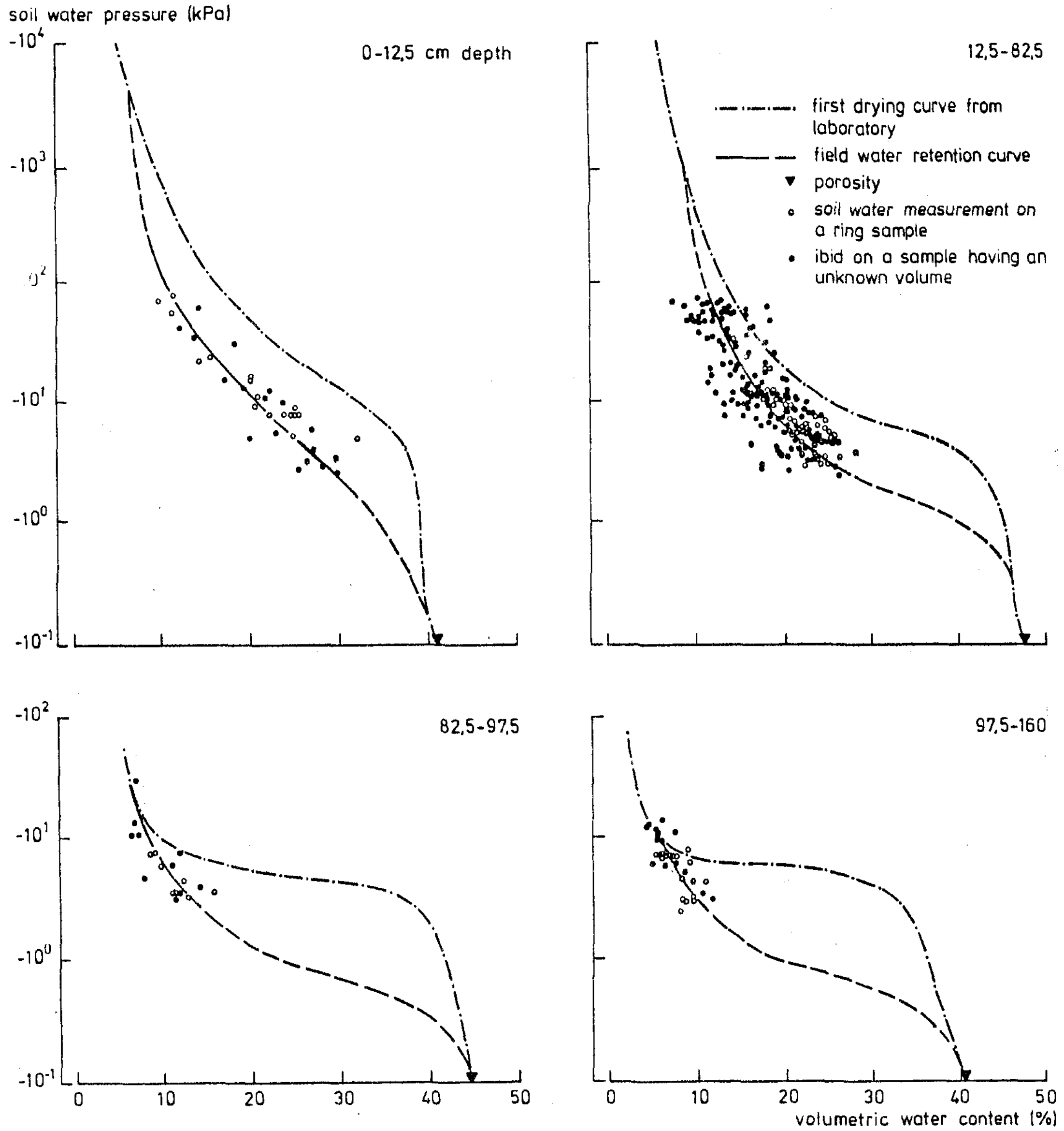


Fig. 12. Water retentivity of different soil layers at the experimental field, according to in situ measuring results

percents by the average values found for the dry bulk density of the soil at the sampling depths (VAN BOHEEMEN and HUMBERT, 1983).

As shown in the figure, soil water pressures higher than -2.5 kPa (pF = 1.4) were not measured. This is due to the good internal drainage of the soil and the rather deep groundwater level. Soil water pressures lower than -85 kPa (pF = 2.9) could not be measured, because tensiometers were applied.

The in situ results proved not to be grouped around the first drying curves based on tension plate measurements. This phenomenon, also noted by RICHTER (1974), FLÜHLER et al. (1976), SILVA (1977) and VAN DER SCHANS

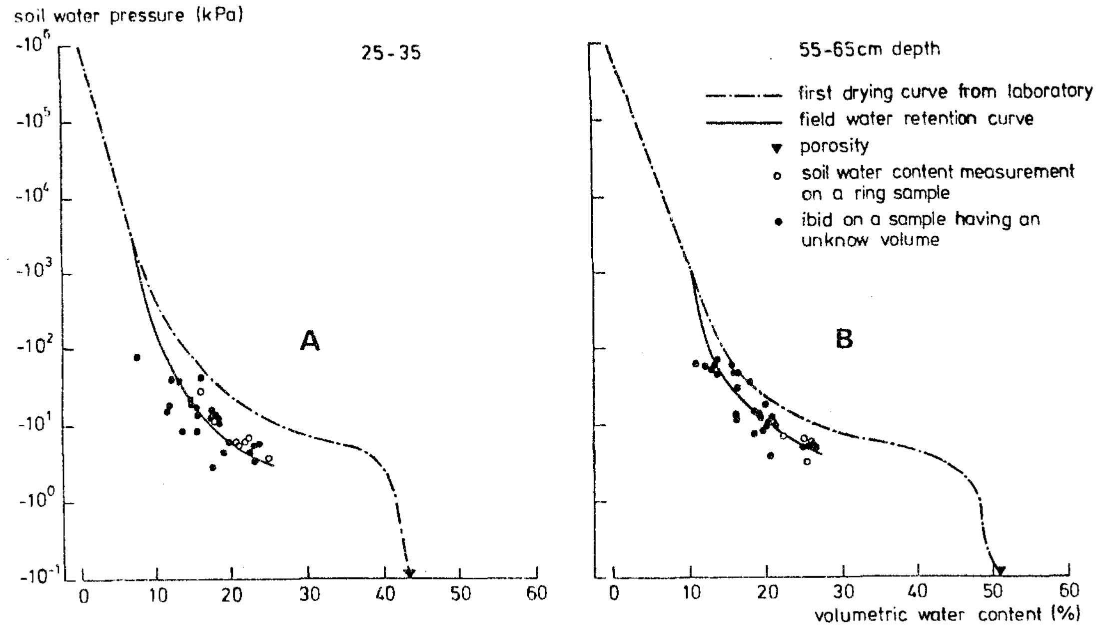


Fig. 13. Water retentivity measured in situ at two levels in the soil at the experimental field

and HELLINGS (1984) is attributed to hysteresis. The in situ results are considered to represent points on the scanning curves followed in the field.

The highest points correspond with situations reached after dry periods and indicate the place of the upper ends of the field drying scanning curves. The lowest points correspond with situations occurring shortly after a strong wetting of the layers and indicate the place of the lower ends of the field wetting scanning curves.

The soil water pressure at which the laboratory and field data approach each other, is relatively low for the shallow layers and relatively high for the deeper ones. This is related to the fact that the deeper horizons do not dry out so strongly. In discussing the tension plate measurements on non-wetted samples and the application of the evaporation technique comparable phenomena were mentioned.

In Fig. 13 the field data of the horizons 25-35 and 55-65 cm below surface are given separately. These two horizons represent the extremes in dry bulk density within the layer 12.5-82.5 cm below surface (ρ_{25-35} = 1.41 and ρ_{55-65} = 1.25 $t \cdot m^{-3}$; VAN BOHEEMEN and HUMBERT, 1983). Furthermore, they are depleted differently during dry periods.

The solid lines drawn through the clouds of points have been extrapolated upwards in a way that they join the first drying curves at soil water contents corresponding with values found in these layers after a long dry period. The broken lines indicating the downwards extrapolations are assumed to be the lowest parts of main wetting curves. They are according to the experimental data on hysteresis given by ILNICKY (1982a) and VACHAUD and THONY (1975) and presented in Fig. 3.

Air entrapment has not been accounted for in the last procedure, because soil samples from the experimental field proved to be saturated practically completely if they stayed only two days in a water bath.

4.3. Evaluation

In literature a lot of data about causes and magnitude of capillary hysteresis have been found. These data show that, especially in sand, the variation in soil water content at the same soil water pressure can be large.

At the experimental field the soil consists of rather uniform sand with an organic matter content of about 6 percents in the layer 0-82.5 cm below surface and 0.2 percents below 97.5 cm depth (VAN BOHEEMEN and HUMBERT, 1983).

The results of the in situ measurements and the laboratory measurements on non-wetted field samples demonstrate that at the experimental field the occurrence of hysteresis is of importance and that the first drying curves determined by desorption of initially saturated soil samples on tension plates and in pressure cells cannot be used as $h(\theta)$-relations for field situations.

The field measuring program was however not so intensive that the results admit determining field main (drying and wetting) curves as well as field scanning curves.

Prediction of the hysteretic effects by modeling is considered to be feasible, if for that purpose the two main curves would be available and some wetting and drying curves for verification. The first drying curves obtained in the laboratory might be seen as main drying curves. Foundation of the prediction on only such curves, which is possible when using one of Mualem's models, has been regarded too risky. Another difficulty, being of second order, is that the variation in soil water content measured at different water pressures in the field is not only caused by hysteresis, but also by experimental errors and spatial variability in the concerning soil layers.

In further analyses the lines drawn in Fig. 12 through the clouds of points, together with the parts of the first drying curves above the junction of both types of curves, will be tried out as representative $h(\theta)$-relations. In this way a good approximation for very dry and wet horizons seems possible. The approximation will be less for less extreme situations like those prevailing during infiltration at a rather low rate and internal drainage of a wetted zone. The inaccuracy introduced thereby in modeling soil water flow macroscopically is not so severe as suggested by the figures 11, 12 and 13, because the differences in these figures between 'measured' and 'schematized' are not only due to hysteresis, but also to experimental errors and spatial variability.

For some horizons of the soil profile at the experimental field water retention curves are given by WÖSTEN (1983). These curves have been based on one or two in situ measurements of water content and water pressure, tension plate measurements on initially saturated soil samples (pressure range 0 to -10 kPa) as well as simultaneous measurements of water content and water pressure in soil columns in the laboratory (-10 to -80 kPa). Latter columns were used firstly for determination of the hydraulic conductivity at rather high soil water pressures and

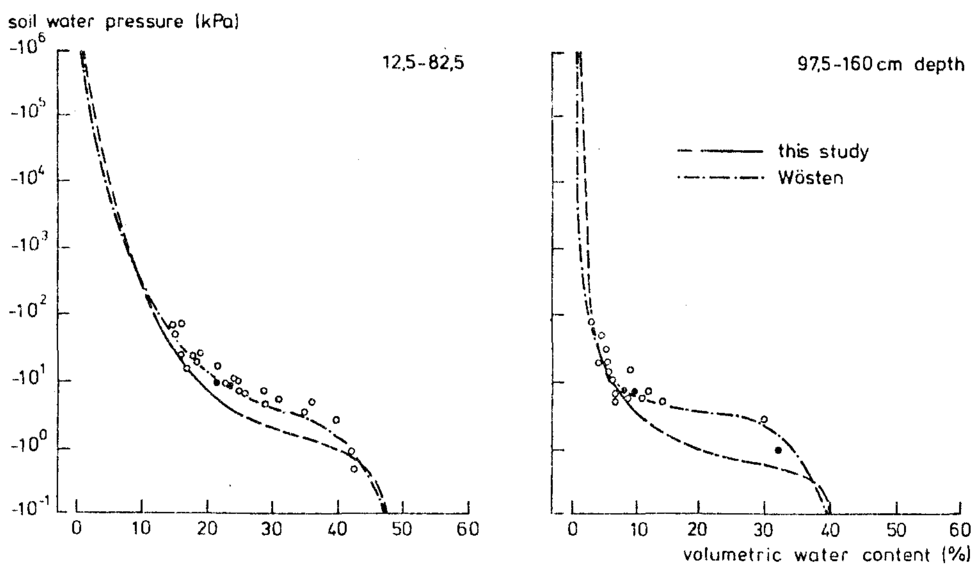


Fig. 14. Water retention curves derived by WÖSTEN (1983) from laboratory (o) and in situ (●) measurements on samples from the layers 12.5-82.5 and 97.5-160 cm below surface, as well as the corresponding curves derived in this study from in situ measurements

during the water retentivity measurements they dried by evaporation from the upper side. For applying the tension plate technique other instruments were used than in this study.

The water retention curves given by Wösten, which correspond rather well with the curves from the evaporation technique, lie below the first drying curves presented in Fig. 9, but above the curves derived in this study from in situ measurements (Fig. 14). The latter point is attributed to the fact that the curves of Wösten have been based practically only on measurements during drying of saturated samples.

5. HYDRAULIC CONDUCTIVITY OF DIFFERENT SOIL LAYERS

For the determination of the hydraulic conductivity, either as a function of soil water pressure or as a function of soil water content, a number of techniques is available. Some of them have been applied. Before discussing the results, attention is paid firstly to the phenomenon hysteresis.

5.1. Hysteresis

As noted above there are two ways for expressing the hydraulic conductivity, i.e. the functions $k(h)$ and $k(\theta)$. The former one shows significant hysteretic effects if the water retention curve of the concerning soil does.

The discussion about hysteresis of the $k(\theta)$-relation has not yet been closed definitely. YOUNGS (1964) and STAPLE (1965) reported that the conductivity corresponding with a certain soil water content is higher during wetting than during draining. Youngs did measurements on slate dust in the range of conductivity values of 1 to 100 $cm \cdot d^{-1}$ and found a proportionality factor of about 2. Staple studied the conductivity of repacked Grenville silt loam in the range 10^{-4} to 15 $cm \cdot d^{-1}$ and found a proportionality factor of 10. COLLIS-GEORGE and ROSENTHAL (1966) derived a factor 1/10 from experiments on fine Fontainebleau sand, implicating that the conductivity during wetting is smaller than during drying (measuring range 10^{-2} to 1 $cm \cdot d^{-1}$). POULOVASSILIS (1969, 1970) and POULOVASSILIS and TZIMAS (1975) described measurements on repacked, uniform and mixed sand fractions as well as on glass beads. During drying the condctivity was found to be higher than during wetting, the proportionality factor being rather low (<2). The lowest conductivity value measured in these experiments was about 0.5 $cm \cdot d^{-1}$, a value much higher than occurring normally in the field.

A lot of investigators state that the magnitude of hysteresis is zero or so small that it is difficult to say or the differences between the wetting and drying curves results from hysteretic effects or from experimental errors in determining the $k(\theta)$-values (NIELSEN and BIGGAR, 1961; ELRICK and BOWMAN, 1964; GREEN et al., 1964; JACKSON et al., 1965; TOPP and MILLER, 1966; TOPP, 1969; TALSMA, 1970; ROGERS and KLUTE, 1971; VACHAUD and THONY, 1971; TZIMAS, 1979). Two of these papers concern undisturbed natural soils. Green et al. did measurements on undisturbed field cores, Tzimas on an undisturbed soil column.

Thus, only a small part of the literature mentioned above indicates hysteresis of the $k(\theta)$-relation, although the corresponding measurements do not cover a wide range of field situations. On the contrary, many studies showed that hysteresis is absent or negligible. In this study the $k(\theta)$-relation therefore has been regarded to be a single-valued function. Hence, the measuring program was focussed on obtaining $k(\theta)$-relations for the different soil layers at the experimental field and no intensive study has been made of the magnitude of the hysteretic effects on the $k(h)$-relation and the possibilities to simulate and to predict them with aid of models.

5.2. Saturated conductivity

5.2.1. Available techniques

Because of the deep groundwater table at the experimental field techniques suitable for determining the saturated conductivity of soil layers in the unsaturated zone were necessary.

For laboratory use the so-called permeameter has been developed. In the constant-head variant of this method the hydraulic potential gradient in the sample is maintained constant during the experiment, in the constant-flux variant the discharge through the sample. In the falling-head variant both the gradient in hydraulic potential and the discharge vary.

Devices have also been developed for application of the constant-head variant of the permeameter in the field (BAKER and BOUMA, 1976).

Different types of infiltrometer techniques are available for determining the saturated conductivity in the field of soil layers above a water table. The simplest one includes measuring of the infiltration rate in a cylinder placed in the soil at the moment the hydraulic potential gradient in the saturated zone has approached the gradient of the gravitational potential (KESSLER and OOSTERBAAN, 1974). The air-entry method of BOUWER (1966) and the double-tube method, which was also developed by BOUWER (1961, 1962, 1964) require measuring of the infiltration rate and the free water level above the soil surface. In case of applying the double-tube method the infiltration depth must be measured too.

The inverse auger-hole method (KESSLER and OOSTERBAAN, 1974) is a relative rough method involving derivation of the saturated conductivity from the infiltration rate from an auger hole.

For sandy soils as occurring at the experimental field it is also possible to predict the saturated conductivity from data about the pore geometry. A well-known way is the use of the Kozeny - Carman equation (CARMAN, 1939), requiring data about porosity and internal surface of the soil exposed to the water. A comparable equation has been formulated by HOOGHOUDT (1934). BRINKMAN (1949) proposed a different type. Furthermore, several efforts have been made, using Poiseuille's law, to predict the saturated conductivity on the basis of the pore-size distribution function represented by the water retention curve. This technique was proposed first by CHILDS and COLLIS-GEORGE (1950) and modified later by MARSHALL (1958), MILLINGTON and QUIRK (1959), KUNZE et al. (1968), GREEN and COREY (1971), and JACKSON (1972). All the above prediction techniques include use of one or more so-called matching factors.

BLOEMEN (1980) published recently empirical equations for prediction of the saturated conductivity on the basis of data about the particle size distribution and organic matter content.

5.2.2. Results

In this study the permeameter technique has been used because of its high accuracy. Furthermore, the prediction technique of Bloemen was applied because the data required for application were available and the constants in Bloemen's equations are mainly based on data of Dutch soils.

In the experimental field single soil samples were taken at five

Table 5. Saturated conductivity (K_s) of soil layers at the experimental field, predicted according to Bloemen's method and measured with permeameters

Soil layer		Bloemen's method	Constant-flux technique		Constant-head technique	
number	depth	K_s	sampling depth	K_s	sampling depth	K_s
	(cm)	($cm \cdot d^{-1}$)	(cm)	($cm \cdot d^{-1}$)	(cm)	($cm \cdot d^{-1}$)
1	0 - 12.5	252	-	-	-	-
2	12.5- 82.5	283	33- 48 53- 68	218 156	45- 65	180
3	82.5- 97.5	176	81- 96	164	-	-
4	97.5-160	218	118-133 141-152	604 218	125-145	650
5	160 -230	204	-	-	165-185	210
6	230 -320	-	-	-	-	-

depths in cylinders with a height of 15 cm and a diameter of 10 cm. In the laboratory the saturated conductivity of the samples was measured, using the constant-flux variant of the permeameter technique (Table 5).

WÖSTEN (1983) applied the constant-head variant of the permeameter technique on samples with a diameter and a height of 20 cm. His results agree very well with those presented earlier (Table 5). He also found a remarkable high value for the middle of the layer 97.5-160 cm depth.

The conductivity values obtained with Bloemen's prediction method, except the value for the middle of the layer 97.5-160 cm below surface, have practically the same magnitude as the measured ones.

5.3. Unsaturated conductivity

5.3.1. Available techniques

Several techniques involve the analysis of an one-dimensional unsaturated steady state flow. That of CHILDS and COLLIS-GEORGE (1950) and the variation on it proposed by WESSELING and WIT (1969) require the introduction of a constant infiltration rate at the upper side of a soil sample and a free outlet of percolated water at the bottom. Application of the other methods requires the installation of different constant water potentials at the two ends of a sample, for instance in case of the double-membrane technique developed by RICHARDS (1931). The same principle is applied in the technique of YOUNGS (1964) whereby the upper side of a soil column is connected with a hanging water column, and that of HILLEL and GARDNER (1970) whereby infiltration of free water in the soil is controlled by an impeding layer (crust) covering the soil surface. The latter two techniques involve infiltration as a consequence of a constant, relative high potential at the upper side of the soil column and eventually freely draining of percolated water at the bottom. In the evaporation technique proposed by MOORE (1939) water enters at the lower side of the sample and evaporates from the top. Conductivity values are calculated as quotients of the flux and the corresponding gradient in hydraulic potential. They must be referred to the water content of the soil between the points where the hydraulic potential has been measured.

All the above mentioned techniques have been developed for application in the laboratory. The impeding layer (crust) technique has been modified for field application (BOUMA et al., 1971; BOUMA and DENNIG, 1972). Of course, data on steady state flows in the soil profile collected in the field can also be used for determination of the conductivity.

Several ways exist for deriving the unsaturated conductivity by analyzing an one-dimensional transient flow. Some involve measurements on soil samples wherein a drying process has been induced. Among them are the outflow technique (GARDNER, 1956), the one-step technique (GARDNER, 1962; DOERING, 1965) as well as the hot air technique developed by ARYA et al. (1975). The outflow and one-step techniques require the interpretation of data about the falling rate of outflow from a sample in a pressure cell or on a tension plate after the pressure c.q. tension has been increased. The $k(\theta)$-relation can only be determined if the $h(\theta)$-relation prevailing during the measurements is known. If the hot air technique is applied, the soil sample dries due to evaporation from the upper side. By analyzing the water profile in the sample after the evaporation process has been stopped, the $k(\theta)$-relation can be determined. In that case the $h(\theta)$-relation corresponding with the drying process is needed.

DIRKSEN (1979) developed the so-called sorptivity technique leading to wetting instead of drying of the soil. Hereby a set of infiltration processes in downwards direction are created. They are controlled mechanically in a way that the cumulative absorption decreases proportional with the square root of the time elapsed after starting the infiltration. With aid of the $h(\theta)$-relation valid for infiltration processes the $k(\theta)$-relation can be derived from data about the water pressure in the wetted part of the soil. BRUCE and KLUTE (1956) described how, if the water retentivity of the soil is known, the conductivity can be derived from data about the progress of infiltration in a horizontal soil column, induced by maintaining saturation at one end of the column. ROSE (1968) applied the same technique by following the drying process upon evaporation at one side of the column.

All the above mentioned transient flow techniques have been developed for use in the laboratory. The instantaneous profile technique is, however, perfectly suitable for use in the field. It can be applied in different ways (KLUTE, 1972). In principle it comprises of measuring periodically soil water content profiles and soil water pressure profiles in situations where only upwards or downwards flow and no extraction by for example vegetation occur. The fluxes at different time and space intervals can be derived from the successive soil water profiles and related to the corresponding gradients in hydraulic potential. So, simultaneously the $h(\theta)$- and $k(\theta)$-relations of the occurring soil layers can be obtained.

The instantaneous profile technique also forms the basis of different laboratory methods (KLUTE, 1972), The method developed by BOELS et al. (1978) is one of them. Thereby a wetted soil column dries due to evaporation from the upper side. The method provides measuring periodically soil water pressures at different depths and weighing of the column.

Furthermore, there are different models for predicting $k(\theta)$-relations. Two groups can be distinguished, one with an apparent macroscopic and another with an apparent microscopic and statistical approach. The first group consists of power functions based on a generalization of Kozeny's theory for saturated and unsaturated porous media (AVERJENOW, 1950; IRMAY, 1954; MUALEM, 1978). The second group includes more advanced models, based on the thought that the $k(\theta)$-relation can be derived, using Poisseuille's law, from the pore size distribution function represented by the water retention curve. It consists of modifications of the model developed by Childs and Collis-George (par. 5.2.1) as well as the models of BURDINE (1953), WYLLIE and GARDNER (1958) and MUALEM (1976a). A review was given recently by MUALEM and DAGAN (1978). The models of both groups have parameters for which values have to be determined empirically.

5.3.2. Results

Measurements have been done according to the evaporation technique described by BOELS et al. (1978), because thereby conductivities could be determined for an important range of soil water contents occurring in the field and without use of $h(\theta)$-relations based on other measuring procedures. Fig. 15 shows the results obtained from the separate soil samples. The curves derived from these results for different soil layers are given in Fig. 16 (solid lines). In the latter figure also $k(\theta)$-values derived from $k(h)$-values given by WÖSTEN (1983) are presented. Those $k(h)$-values were obtained with the crust technique on soil columns from the concerning layers and have been transformed in $k(\theta)$-values with aid of $h(\theta)$-relations also given by Wösten. This procedure is admitted, because the crust technique was applied involved drying of the soil columns used for measuring the saturated conductivity and the $h(\theta)$-relations used in the transformation were based on measurements during drying of initially saturated soil. The junction of the $k(\theta)$-values from the crust technique to those of the evaporation technique is rather well.

Wösten also presented $k(h)$-values derived from measurements according to the sorptivity technique of DIRKSEN (1979) with aid of the $h(\theta)$-relation mentioned above. Transformation into $k(\theta)$-values lead to large differences with the $k(\theta)$-values obtained with the evaporation technique. This is attributed to the fact that due to hysteresis $h(\theta)$-relations based on absorption measurements would have to be used in analyzing the sorptivity measuring results.

During the spring of the year 1983 a steady state situation was reached at the experimental field as a consequence of a period with a rather constant precipitation rate. For this period $k(\theta)$-values have been calculated, using the percolation intensity in the soil profile, the prevailed soil water pressure profile and the corresponding soil water content profile (Table 6). In this way one point of the $k(\theta)$-relations for the layers 12.5-82.5 and 97.5-160 cm below surface could

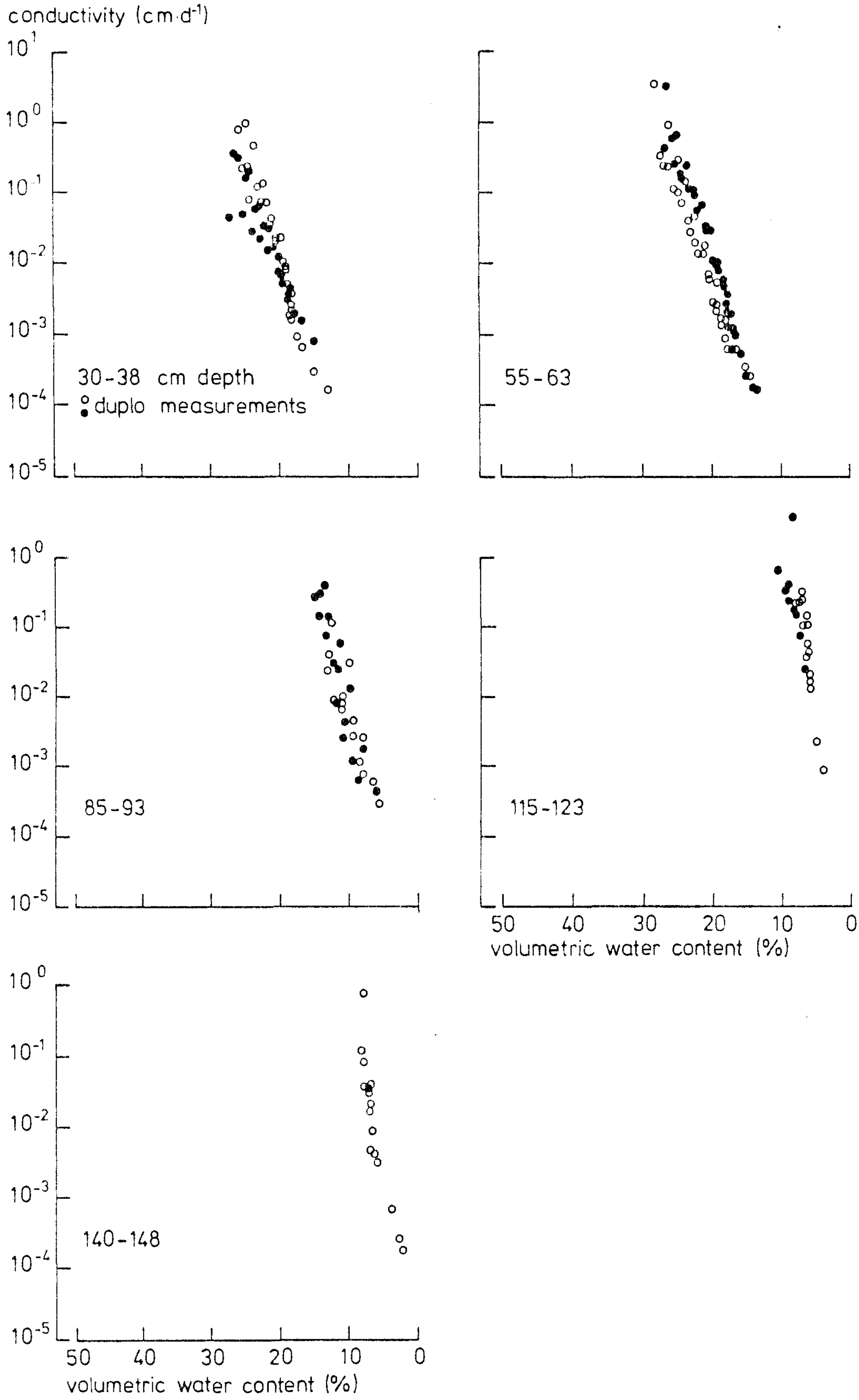


Fig. 15. Hydraulic conductivity of different soil samples from the experimental field measured according to the evaporation technique of BOELS et al. (1978)

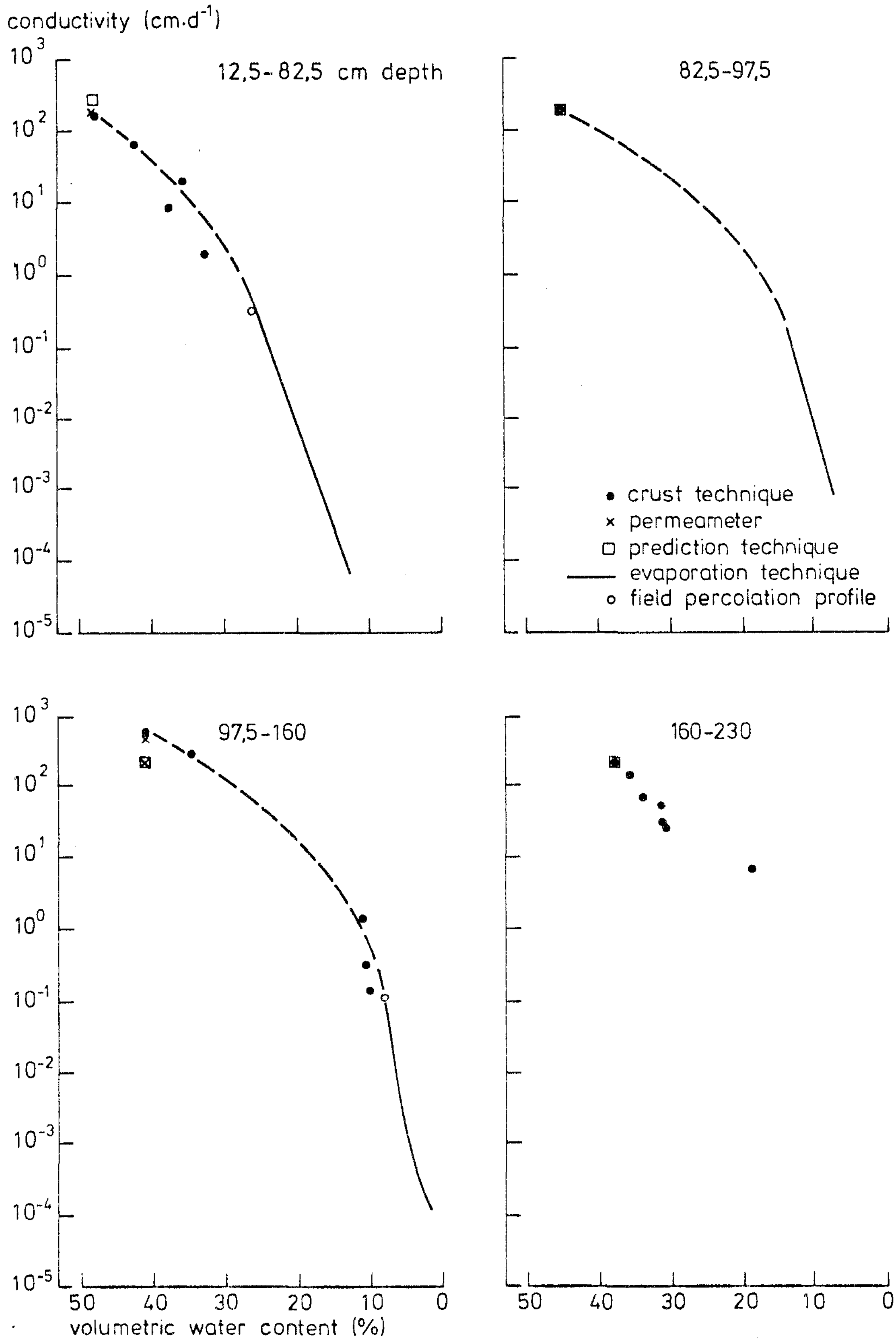


Fig. 16. Hydraulic conductivity of different soil layers at the experimental field

Table 6. Steady state situations found in the field and used for deriving conductivity values

Depth (cm)	Soil water pressure (kPa)	Soil water content ($cm^3 \cdot cm^{-3}$)	Hydraulic potential gradient ($kPa \cdot cm^{-1}$)	Flux ($cm \cdot d^{-1}$)	Conductivity ($cm \cdot d^{-1}$)
60	4.9	26	-0.042	-0.15	0.36
130	3.1	8	-0.125	-0.15	0.12

be obtained. As shown by Fig. 16 the in situ measuring results agree well with the laboratory results.

5.4. Evaluation

K(θ)-relations (Fig. 16) were obtained by applying the evaporation technique described by BOELS et al. (1978) and the crust technique described by BOUMA et al. (1971) and BOUMA and DENNIG (1972). In situ measurements executed in the field gave k(θ)-values which proved to be in close agreement with those k(θ)-relations. Therefore the curves presented in Fig. 16 will be used in further analyses.

6. SUMMARY

A partial differential equation has been formulated for describing soil water flows in the field where sprinkling experiments on grassland were carried out. Thereby parameters have been defined for characterizing water retentivity and hydraulic conductivity of the soil. These properties have been fixed in sets of h(θ)- and k(θ)-functions for the different layers distinguished in the soil profile at the experimental field.

The water retentivity of the soil proved to be affected by capillary hysteresis. The water retention curves determined in the laboratory by desorption of initially saturated soil samples in pressure cells and on tension plates have been classified as first drying curves. Tension plate measurements on non-wetted field samples and in situ measurements demonstrated that the latter curves cannot be used as h(θ)-relations for field situations. Therefore alternative curves have been defined, mainly on basis of the results of the in situ measurements.

The $k(\theta)$-relation of the soil has been regarded to be a single-valued function. For determining it the evaporation technique was preferred, because thereby the $k(\theta)$-relation is found directly for an important range of soil water contents occurring in the field. Results of the impeding layer (crust) technique have been used for defining the $k(\theta)$-relation at high soil water contents. $K(\theta)$-values measured in situ proved to be in agreement with the defined $k(\theta)$-functions.

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