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UNSATURATED HYDRAULIC CONDUCTIVITY OF SANDY SOIL COLUMNS PACKED TO DIFFERENT BULK DENSITIES AND WATER UPTAKE BY PLANTROOTS

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C O N T E N T S

1. INTRODUCTION

Knowledge of hydraulic conductivity $\underline{K}$ at different suction values $\underline{\psi}$ is fundamental whenever a water balance has to be done in a soil to predict the amount of water available for the crop.

On the one hand the need of knowing accurately $\underline{K}$ as a function of $\underline{\psi}$ on the other the great difficulties for determining it. Many attempts to determine $\underline{K}(\underline{\psi})$have been carried out by several authors using both mathematical approaches and experimental measurements.

The first approach comprehends various empirical formules (WIND, 1955; WESSELING, 1957; GARDNER, 1958; RIJTEMA, 1965) relating hydraulic conductivity and suction, and also some computational methods mainly based on the pore interaction model (CHILDS & COLLINS-GEORGE, 1950; MARSHALL, 1958; KUNZE et al. 1968; GREEN & COREY, 1970).

Direct measurements can be performed in s i t u or in laboratory (on disturbed or undisturbed soil samples) under steady-state or transient flow conditions. The available data and comparative results from the literature are summarized in WESSELING & WIT, 1969; GREEN & COREY, 1970; WESSELING, 1974; BOELS et al. 1978.

This paper describes a laboratory method (BOELS & VAN HEMMEN, 1975) used to determine both the soil moisture retention curve and the unsaturated hydraulic conductivity in soil columns under transient flow conditions during evaporation.

The soil properties thus determined have been used in calculating through a numerical approach the capillary rise for different flow velocities. Afterwards the water uptake by a crop has been measured in both falling and steady water table conditions in 120-cm-long artificially packed soil columns.

2. EXPERIMENTAL PROCEDURE

2.1. Filling up of the soil samples

The soil used in this experiment is a fine sandy soil of eolic origin (particle size between 200 µ and 300 µ) of an experimental field in the 'Vredepeel'. Values of K, ψ, θ were measured directly in the field during the crop growing season as well as during the winter. In order to realize conditions likely to be found in the field the soil was packed up with different bulk densities ranging from 1.6 to 1.8 g cm^{-3}, values that can represent the one-meter-deep conditions measured in the field (VERHAEGH, personal communication).

Twenty-cm-long rigid PVC cylinders with a diameter ϕ10 cm, with one-cm holes for the tensiometers at 5 cm intervals were used. During the filling-up the lower end of the cores was sealed to a metallic plate and the holes were covered.

As is well known, the soil bulk density depends on the moisture content, provided the pressure on it is constant (fig. 1).

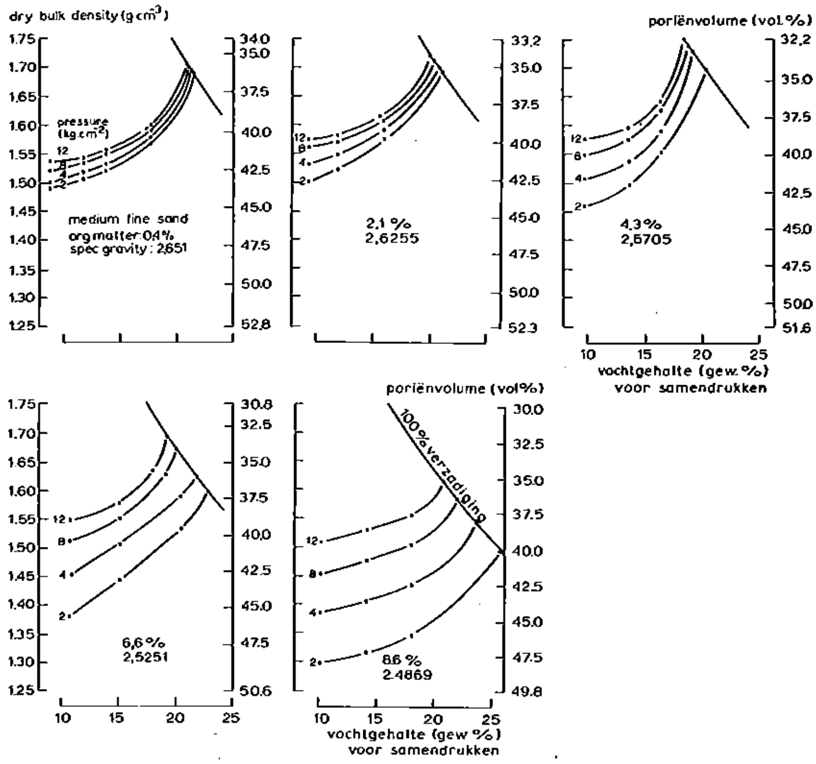


Fig. 1. Bulk density *versus* moisture content at different pressure of sandy soil

Working with different pressures at different soil moisture contents produces various bulk density values. In the sandy soil, these figures have come out:

Moisture content (% d.w.)	Pressure (kg cm^{-2})	B.D. (g cm^{-3})
14	1	1.60
19	3.5	1.65

The pressure was applied with a hydraulic press on the cores whilst taking care to prevent layered profiles. - The soil thereby was made to enter into the cylinders in such an amount that the thickness of the layer - after each one-minute-long pressure application - was increased by 2 cms; the pressed surface was scraped to prevent crust formation.

Due to the narrow range of moisture content in sandy soils, irrespective of the applied pressure B.D. values greater than 1.65 could not be reached.

For the two densiest samples vibrational force was used. In one case with saturated soil (20% d.w. water content) the B.D. was 1.7; in the other case the soil was oven-dried and the B.D. was 1.8. The packed columns were allowed to saturate in tapwater for a few days. - The satuaration conditions were controlled through tensiometers inserted in the top of the samples.

2.2. Set-up of measurement

The here applied evaporating method consists in the simultaneous continuous measurements of both suction at different positions along the core profile and weight of the whole sample in successive times starting from the saturation, under transient flow conditions.

The saturated columns, sealed of the bottom to a plate with flexible aquarium plastic to allow only top evaporation, were placed on a strain-gauge load cell weighing with an accuracy of 1 g.

Ceramic cup tensiometers (BAKKER, 1975) were inserted in the holes at the depths of 2.5 cm; 7.5 cm; 12.5 cm; 17.5 cm connected through a nylon pipe to a scanivalve regulated by a solenoid controller which turns, after pre-selected time, each tensiometer to a pressure transducer. Heigths and tensiometers outputs were recorded on a paper tape.

The tensiometer reading time was chosen on the basis of moisture content i.e. stage of drying off, kind of soil and cup resistance (VEERMAN, 1976). In the suction range between saturation and $\psi = 100$ cm of water column, the reading time was 10'. Above 100 cm the time was turned off to 15'.

The experiment was performed in a dark climatized room at a constant air temperature of $25^{o}C \pm 1^{o}C$ and relative humidity of 30% ± 2%. A fan (wind speed 16.10^{3} mh^{-1}) one meter distant from the samples allowed a faster evaporation.

The measurements were stopped when the change in weight was less than 1 g per day.

At this time the samples were cut in layer, each of them weighed, dried in oven at $105^{o}C$ to check both final soil moisture and bulk density.

3. THEORY

Hydraulic conductivity is the proportionality factor between flow and hydraulic gradient in the Darcy's law extended also to the unsaturated flow conditions:

$$\vec{v} = - K(\psi) \quad \vec{\nabla} H \qquad (1)$$

where:

$\vec{v}$ = flow

H = hydraulic head

ψ = matric suction head

$K(\psi)$ can be derived from (1) knowing the flow and the gradient.

In one-dimensional case it is

$$v = - K(\psi) \frac{dH}{dz} = - K(\psi)\left(\frac{d\psi}{dz} + 1\right) \qquad (2)$$

$H = \psi + z$

z = component of the gravitational head in the direction z v can be calculated integrating the continuity equation

$$\frac{dv}{dz} = - \frac{d\theta}{dt} \tag{3}$$

t = time

θ = volumetric moisture content

provided known the relationship between θ and t.

In the evaporating columns the suction $\underline{\psi}$ was measured in each layer continuously, so ψ(z,t) was known. To obtain the correspondent θ(z,t) it is necessary to have a very accurate relationship between θ and ψ. If θ(ψ(z,t)) is known it is possible then to integrate (3)

$$v = - \int_{z=-D}^{z=0} \frac{d\theta(z)}{dt}\, dz \tag{4}$$

The relationship between θ and ψ was determined in this experiment following two different ways.

1. Polynomial method

This method (BOELS, 1973) consists in subdividing the continuous function ψ(θ) in an $\underline{m}$ number of intervals in which it is sensible to assume a linear relationship between the change in moisture content Δθ and the change in suction Δψ:

$$\Delta\theta = \beta\Delta\psi \tag{5}$$

The angular coefficient β can be determined from

$$\Delta w = F\int_{t_1}^{t_2} \int_{z=-D}^{z=0} \beta\psi(z,t)dz\, dt \tag{6}$$

being:

Δw = the measured change in weight (water lost during evaporation) in the t_2-t_1 time interval

F = area of the crossection

D = the height of the sample

z = the vertical coordinate

$\psi(z,t)$ = the measured suction

The integral (6) is the sum of the contributions of each $\psi(z)$ between t_2 and t_1 multiplied by the β_s. It can be written as

$$\Delta w_j = \sum_{i=1}^{m} \beta_i \, A_{ji} . F \tag{7}$$

where A_{ji} is (in cm) the area comprehended between two successive section profiles (ψ_j and ψ_{j-1}) into the chosen interval i. Each Δw_j is measured, each A_{ji} can be determined on a graph basis. From fig. 2 an example for the sample with B.D. 1.6 g cm^{-3}. Taken the w_5, i.e. the measured change in weight between the suction profiles at times t_5 = 9957 and t_4 = 9367, we have:

$$\Delta w_5 = \sum_{i=2}^{5} \beta_i \, A_{5,i}$$

$$0.31 = \beta_2 \, A_{5,2} + \beta_3 \, A_{5,3} + \beta_4 \, A_{5,4} + \beta_5 \, A_{5,5}$$

similar polynomials can be written for each w_j; that allows to have a number n of equations with m (<n) unknowns i.e. the β_is. These values introduced in (5) yield the knowledge of $\theta(\psi)$.

2. Reiteration method

WIND (1969) used a simple method for determining the soil moisture retention curve based on the correction of the moisture content in relation to the suction reading, assuming valid a first moisture retention curve.

The moisture values read on that curve are multiplied by the quotient between the mean moisture content of the whole column and the calculated moisture content. These new moisture contents plotted versus suction give a second moisture retention curve. From this curve for each value of the suction along the profile and during

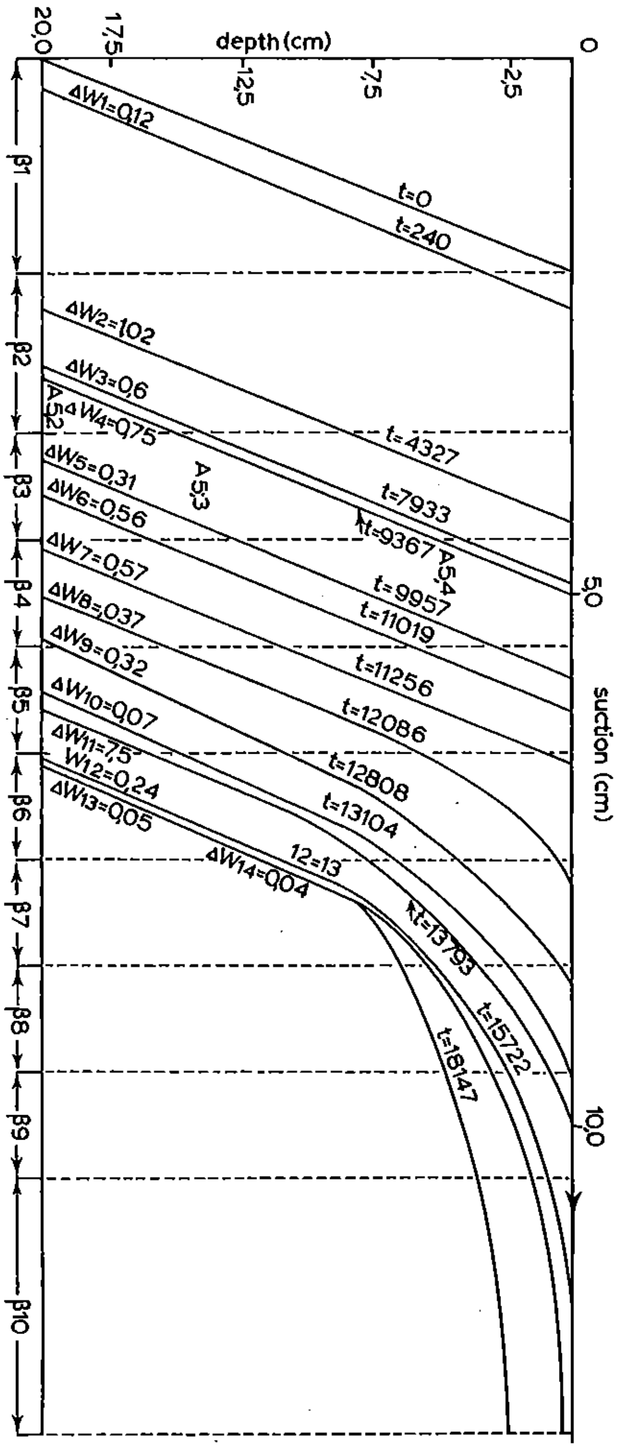


Fig. 2. Suction profiles of a drying soil column. This sample has B.D. 1.6 g cm^{-3}

the time are derived new values of moisture content which are added for every time and the sum is divided by the mean moisture content of the whole column. Each moisture content value is multiplied by the second quotient and θ again plotted versus suction. Through these points the second moisture retention curve of approximation is drawn.

Fig. 3 shows the curves obtained with this method for 1 column.

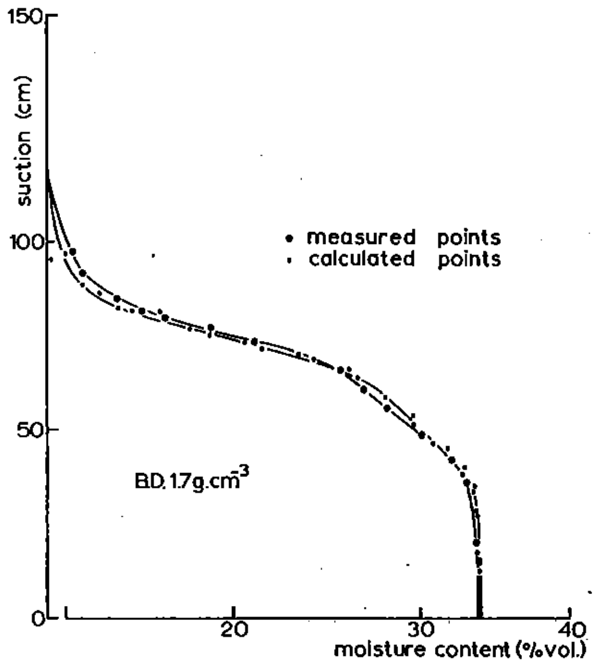


Fig. 3. Soil moisture retention curve determined in a 20-cm-high soil column. The continuous line (——) is obtained directly from laboratory measurements; the dotted line (---) is the second correction obtained with the reiteration method

The reiteration has stopped when the deviation of the points from the smooth line through them are not greater than 1% of moisture content.

In this case due to the accuracy of the measurements the reiteration was stopped already at the second step.

In fig. 4 the soil moisture retention curves for the sample at B.D. 1.65 g cm^{-3} calculated with the two different methods, polynomial and reiteration.

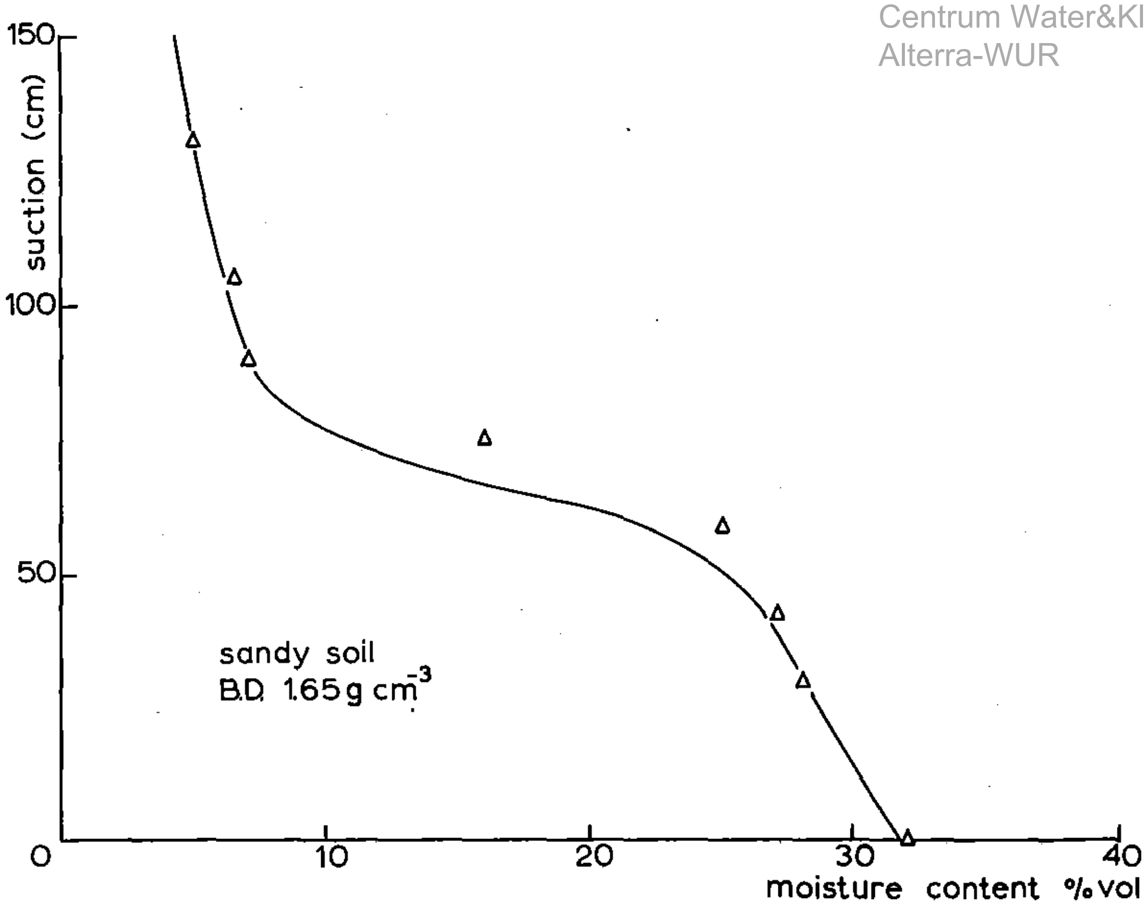


Fig. 4. Soil moisture retention curve (——) obtained with reiteration method; the points (Δ) are calculated with the polynomial method.

In fig. 5 the soil moisture retention curves of the sandy soil at the different bulk densities, used for the K(ψ) calculation.

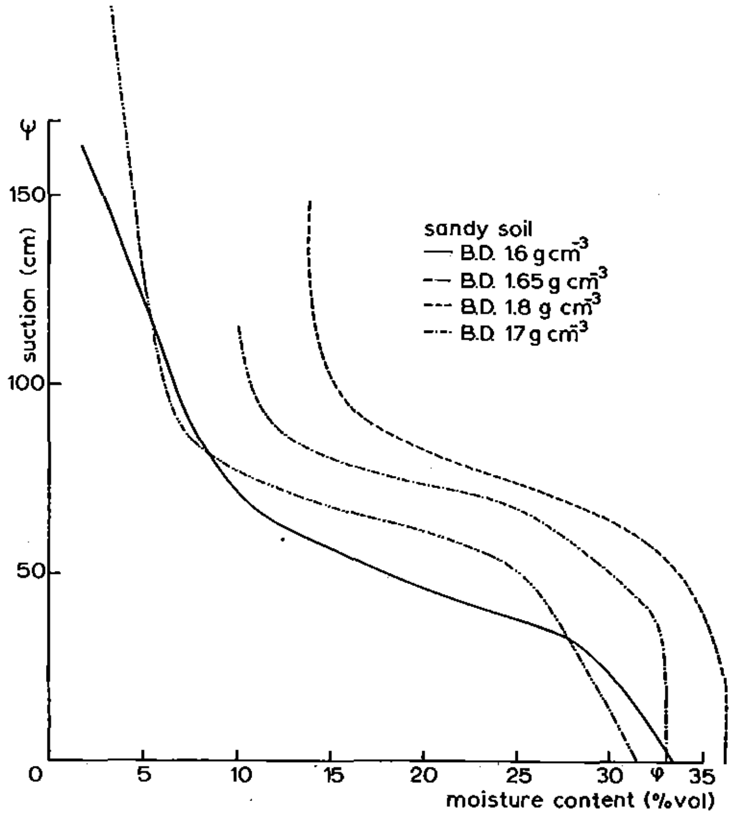


Fig. 5. Soil moisture retention curves for the sandy soil at different B.D.

3. Determination of the unsaturated hydraulic conductivity

Calculation of K(ψ)

Knowledge of θ(ψ) allowed the calculation of the flow velocity v from (4).

In the semplification used here, the (4) has been written as

$$v = \frac{\sum_{i=1}^{3} \Delta\theta_i}{\Delta t} \times d \qquad (8)$$

being $\Delta\theta_i$ the change in moisture content ($cm^3\ cm^{-3}$) in the i th layer, during Δt; the summatory is extended from the bottom layer to the surface s_3 in between the two uppest tensiometers; Δt is the time interval between 2 successive measured suction profiles; d (cm) is the thickness of the layer.

$K(\psi)$ can be derived now from (2), where the suction gradient $d\psi/dz$ has been substituted with the finite ratio $\Delta\psi/\Delta z$

$$K(\psi) = -v \left(\frac{\overline{\Delta\psi}}{\Delta z} + 1\right)^{-1} \tag{9}$$

where $\overline{\Delta\psi} = \dfrac{\Delta\psi_{t+1} + \Delta\psi_t}{2}$

and

$\overline{\Delta\psi}$ is the suction difference (cm) between the two uppest tensiometers. The flow was also calculated in the following way:

$$\sigma = \frac{\Delta w - \Delta\theta_4 .d}{\Delta t} \tag{10}$$

where Δw is (in cm) the change in weight during Δt

$\Delta\theta_4$ is the change of the soil moisture in the uppest layer during Δt.

In table 1 are represented the figures of the change in moisture content (mm) calculated with the two different methods.

The great differences may be attributed to the fact that the computation with the (10) equation overestimates the flow, being the term $\Delta\theta_4$ not representative for the whole layer. In the uppest layer, in fact, the suction gradient is abrupt and particularly high in the surface between the mulch layer and the part below it. Suction in the top can reach values ranging between 15-16 bars while the 1 cm deep readings given values of 0.8 bar and the 2.5 cm deep values < 0.10 bar. In these mulch conditions where the water is lost only by diffusion limited by the relative humidity in the air, the $\Delta\theta_4$ is too small to represent the change in water content in the 5 upper cms.

Table 1. Measured change in weight of the samples during the time (evaporation) and computed ones

B.D. 1.6													
Δt(min)		3915	4024	1665	1415	1067	722	1866					
$\Delta\theta_4$	(mm)	3.3	2.5	2.5	1.5	0.95	0.6	0.5					
$\sum_1^3 \Delta\theta i$	"	9.0	10.0	17	12	8.0	5	4.0					
evaporation	"	6.6	10.8	8.4	7.8	5.1	3.2	4.0	1.9	0.1	0.1	0.5	
B.D. 1.65													
Δt(min)		3915	4024	1665	1415	1067	2238	1046	1237	1407			
$\Delta\theta_4$	(mm)	2.4	3.0	2.8	1.8	0.7	0.4	0.3	0.5	1.3			
$\sum_1^3 \Delta\theta$	"	10.5	8.1	12.0	12.5	7.0	7.8	2.1	2.4				
evaporation	"	7.1	6.1	10.6	7.0	3.6	2.9	7.0	0.3	0.8			
B.D. 1.7													
Δt(min)		425	935	1810	2105	875	720	1870	1040	1460	1265	2360	3230
$\Delta\theta_4$	(mm)	0.2	0.5	0.6	0.9	0.8	0.8	2.4	1.9	0.8	0.7	0.5	0.7
$\sum_1^3 \Delta\theta_i$	"	0.9	2.0	3.2	5.2	4.0	5.2	5.5	8.7	5.4	4.1	5.5	
evaporation	"	1.1	1.6	3.2	3.7	2.3	2.8	9.3	4.6	5.0	2.5	2.6	3.7
B.D. 1.8													
Δt(min)		425	935	1810	2105	875	720	1870	1040	1460	1265	2360	3230
$\Delta\theta_4$	(mm)	0.6	0.6	0.7	1.0	1.1	0.8	2.2	0.9	1.2	0.6	0.3	0.5
$\sum_1^3 \Delta\theta$	"	2.4		2.8		4.1	2.9	11.7	8.7		5.7	5.6	5.4
evaporation	"	2.2	1.1	2.2	2.6	2.5	2.6	7.8	3.9	7.2	2.7	3.4	3.8

In table 2 the measured K(ψ) values calculated with the two different flow, for each sample with different B.D.

Table 2. Hydraulic conductivity values in relation to suction (cm), calculated using the flow through the 5-cm-deep surface (K_1) and the flow through the top-surface (K_2)

B.D. 1.6			B.D. 1.65			B.D. 1.7			B.D. 1.8		
ψ	K_1	K_2	ψ	K_1	K_2	ψ	K_1	K_2	ψ	K_1	K_2
105	0.001	-	146	0.001	-	102	0.05	0.02	105	0.06	0.10
87	0.08	0.05	106	0.02	-	95	0.13	0.12	97	0.06	0.28
79	0.12	0.19	96	0.03	0.22	90	0.30	0.25	92	1.01	0.54
71	1.43	0.76	89	0.10	0.07	87	0.80	0.66	83	2.66	2.95
64	1.54	1.60	81	0.31	0.27	84	0.80	1.19	75	8.5	8.65
57	4.01	4.27	74	0.67	0.56	81	1.70	1.24	70	9.0	7.36
49	4.90	3.40	66	1.30	1.70	75	0.90	2.13	66	7.4	4.64
39	18.0	29.70	56	0.58	0.44	70	3.47	2.67	62	1.3	2.10
25	17.30	12.10	36	1.93	1.75	64	6.60	5.04	39	5.9	5.60
						57	3.60	3.82			
						51	2.60	3.52			
						44	3.10	2.28			
						33	3.20	6.10			

Determination of the unsaturated hydraulic conductivity from the soil moisture retention curve

Unsaturated hydraulic conductivity can be predicted from the pore-size distribution. Several equations have been developed and many experiments have been carried out to test the equations against the experimental data. Many tests suggest the superiority of those methods which use a matching factor.

The method here used is mainly based on the GREEN & COREY 1971 method, modified by VERHAEGH & WIEBING, 1977.

It is assumed that is valid the Poiseulle's law, that the water flow in the soil depends on pore size, continuity of pores, total volume of the pores, which are assumed to be cylindrical.

The pores of different diameters may constitute continuous tubes (of varying diameters) with a certain probability limited by the smallest diameter. According to Poiseulle's law $K \simeq c.r^2$, when r is the radious of a pore - the relation between r and the maximum possible suction $\underline{\psi}$ in a pore is:

$$r = \frac{\rho g}{\sigma} . \psi^{-1} \qquad (11)$$

where:

ρ = water density

g = gravity acceleration

σ = cynematic viscosity of water

That yields:

$$K \simeq \frac{c\rho g}{\sigma} \psi^{-2} = c^1 \psi^{-2} \qquad (12)$$

Introducing now the probability that a pore with a certain radius r_j constitutes a continuous system with the pores of different radius in the a d j e c e n t surface, the conductivity is:

$$K_i = p\, c^1 \sum_{j=i}^{m} (2j + 1 - 2i)\, \psi_j^{-2} = K_o \sum_{j=i}^{m} (2j + 1 - 2i) \psi_j^{-2} \qquad (13)$$

$i = 1 \; .... \; n$

where p is the probability of pores continuity
n is the number of interval of equal $\Delta\theta$in which the coordinate θ has been subdivided and m is the number of classes of pores.

In the (13) it has been assumed that the conductivity of the (n + 1)th interval is 0.

The relationship between K and ψ is in this case determined by measuring a value (for instance at the saturation) of K.

If it is not sensible to assume zero the K of the (n+1)th interval, it is then necessary to know the K value of this interval and solve the equation in respect to the constant pc, being known a $K(\psi_i)$.

In the present experiment n was chosen equal 10 (KUNZE et al. 1968) the conductivity of the (n+1)th interval equal to 0; the constant K_o was found using one of the $K(\psi)$ measured as matching factor.
In table 3, the figures of the $K(\psi)$ and of the K/Ko for the samples.

Table 3. Hydraulic conductivity values (K) in relation to suction (ψ) and soil moisture content (θ) calculated with a pore size distribution method
M.F. = matching factor   K/Ko = proportionality factor

Soil sample with B.D. 1.6 ($g\ cm^{-3}$)					Soil sample with B.D. 1.65 ($g\ cm^{-3}$)				
θ(%vol)	ψ(cm)	K/Ko	K($cm\ d^{-1}$)		θ(%vol)	ψ(cm)	K/Ko	K($cm\ d^{-1}$)	
2.0	160	0.0009	0.06		2.0	240	0.0005	0.007	
5.4	118	0.0045	0.31		5.2	125	0.0036	0.05	
8.8	80	0.0136	0.93		8.4	81	0.0130	0.18	
12.2	64	0.0323	2.20		11.5	74	0.0331	0.46	
15.6	56	0.0644	4.40	M.F.	14.7	69	0.0652	0.9	M.F.
19.0	48	0.1464	7.83		17.9	64	0.1121	1.53	
22.4	42	0.1888	12.89		21.1	59	0.1740	2.39	
25.8	37	0.2942	20.09		24.2	53	0.2564	3.52	
29.2	29	0.4455	30.43		27.4	35	0.3749	5.15	
32.6	8	1	68.30		30.6	8	1	13.75	

Soil sample with B.D. 1.7($g\ cm^{-3}$)					Soil sample with B.D. 1.8($g\ cm^{-3}$)				
θ	ψ	K/Ko	K		θ	ψ	K/Ko	K	
13.5	83	0.0064	0.08		15.0	108	0.0035	0.23	
15.8	79	0.0263	0.34		18.2	87	0.0160	1.07	
18.3	75	0.0612	0.80		20.2	83	0.0399	2.66	M.F.
20.5	73	0.1121	1.46		22.2	78	0.0771	5.10	
22.8	71	0.1802	2.34		25.4	73	0.1281	8.50	
25.0	67	0.2668	3.47	M.F.	28.8	67	0.1963	13.03	
27.3	61	0.3751	4.88		30.5	63	0.2832	18.09	
29.5	53	0.5111	6.65		33.6	50	0.3970	26.5	
31.8	42	0.6878	8.95		35.8	24	0.5987	39.9	
33.0	20	1	13.01		36.0	18	1	66.6	

Fig. 6 shows the relationship between K and respectively θ(fig. 6a) and ψ(fig. 6b).

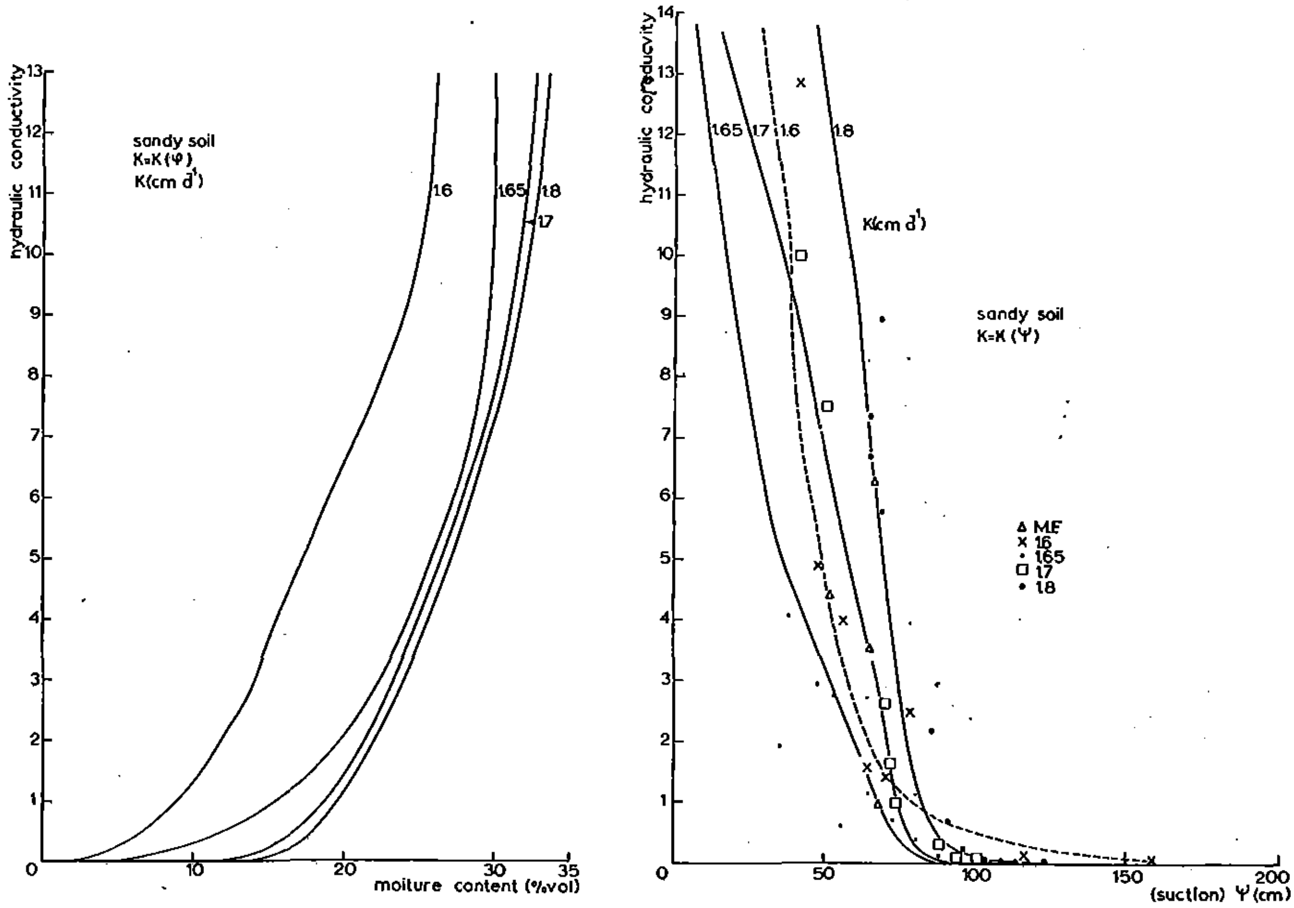


Fig. 6. Unsaturated hydraulic conductivity for the soil at different bulk density values, calculated with the pore size distribution method
a: $K = K(\theta)$; b: $K = K(\psi)$

4. Calculation of the capillary rise

From the Darcy's law

$$v = -K\ (\psi)\left(\frac{d\psi}{dz} + 1\right) \tag{2}$$

where v = flow velocity

$K(\psi)$ = unsaturated hydraulic conductivity

ψ = suction

z = vertical coordinate

keeping v constant, we have:

$$dz = -\left(\frac{v}{K(\psi)} + 1\right)^{-1} d\psi \tag{14}$$

this differential equation can be integrated

$$\int_{o}^{z} dz = -\int_{\psi(o)}^{\psi(z)} \left(\frac{v}{K(\psi)} + 1\right)^{-1} d\psi \tag{15}$$

Due to the complexity of the relationship between K and ψ, eq. (14) is integrated numerically

$$z_j = \sum_{i=1}^{j} \Delta z_i = -\Delta\psi \sum_{i=1}^{j} \left(\frac{v}{K(i\Delta\psi)} + 1\right)^{-1} \tag{16}$$

where $\psi j = j\Delta\psi$

The initial conditions are $z_o = 0$ $\psi_o = 0$
The increment $\Delta\psi$ has to be not too large, for realising an acceptable accuracy.

In the hereby calculation it has been taken the following convention: the sign of the flow is positive if the flow is in the direction of the positive z-coordinate; the suction has a negative sign; the increment $\Delta\psi$ is 2 cm of H_2O.

In tables 4 the data relative to the different samples and the same data are represented graphically in fig. 7.

Table 4. Height (z) of the capillary rise in relation to flow velocity and suction of sandy soil at different bulk density values

a. Bulk density 1.60 g cm^{-3}

v(cm d^{-1})	0.5	0.3	0.2	0.1	0.05	0.025	0.010
ψ(cm)	z(cm)						
20	18.85	19.90	19.94	19.97	19.98	19.99	20.00
40	39.47	39.68	39.79	39.90	39.95	39.97	40.00
60	57.94	58.72	59.13	59.56	59.77	59.88	60.00
80	73.08	75.49	76.84	78.34	79.15	79.57	79.98
100	85.04	89.72	92.59	95.96	97.87	98.91	99.96
120	94.14	101.33	106.07	112.04	115.70	117.76	119.90
140	100.46	110.02	116.76	125.96	132.11	135.78	139.82
160	103.89	115.08	123.39	135.60	144.61	150.51	157.66

b. Bulk density 1.65 g cm^{-3}.

v(cm d^{-1})	0.5	0.3	0.2	0.1	0.05	0.025	0.010
ψ(cm)	z(cm)						
20	19.38	19.63	19.75	19.87	19.94	19.97	19.98
40	37.93	38.72	39.14	39.56	39.78	39.89	39.96
60	55.31	57.06	58.00	58.98	59.48	59.74	59.90
80	66.70	70.58	73.01	75.99	77.82	78.85	79.53
100	71.62	77.61	81.97	88.35	93.09	96.17	98.36
120	74.84	82.45	87.52	98.09	106.15	111.95	116.42
140	76.61	85.23	92.33	104.61	115.98	125.13	132.98
160	78.04	87.51	95.57	110.19	124.70	137.27	148.86
180	79.17	89.33	98.17	114.80	132.19	148.16	163.85
200	79.99	90.66	100.10	118.60	138.15	157.33	177.42
220	80.48	91.46	101.28	120.53	142.15	163.97	188.46
240	80.64	91.72	101.66	121.27	143.56	166.55	193.53

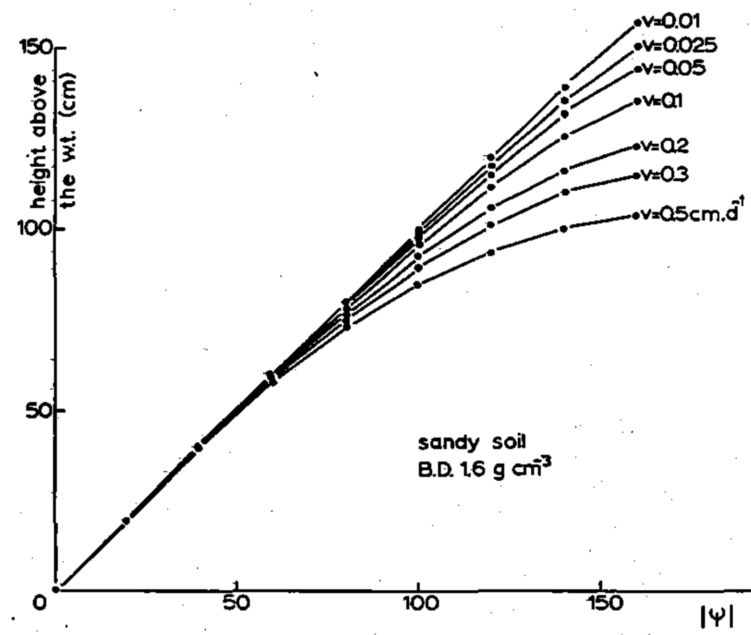
c. Bulk density 1.70 g cm^{-3}

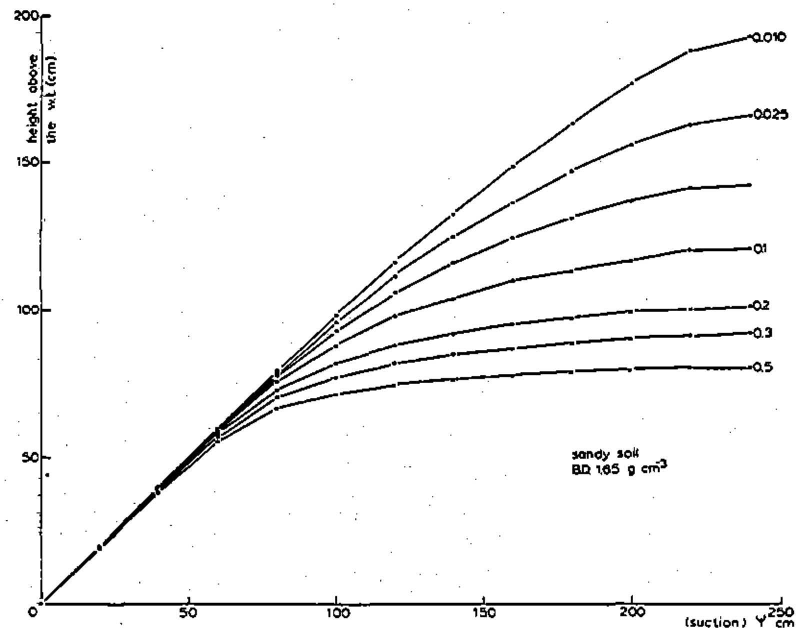
v(cm d^{-1})	0.5	0.3	0.2	0.1	0.05	0.025	0.010	0.001
ψ(cm)	z(cm)							
20	19.78	19.87	19.93	19.96	19.98	19.99	20.0	20.0
40	38.91	39.33	39.55	39.77	39.89	39.94	39.98	40.0
60	57.55	58.50	58.99	59.49	59.74	59.87	59.95	60.0
80	72.31	74.78	76.25	77.94	78.91	79.44	79.77	80
100	74.91	78.74	81.61	86.32	90.65	94.19	97.26	99.70

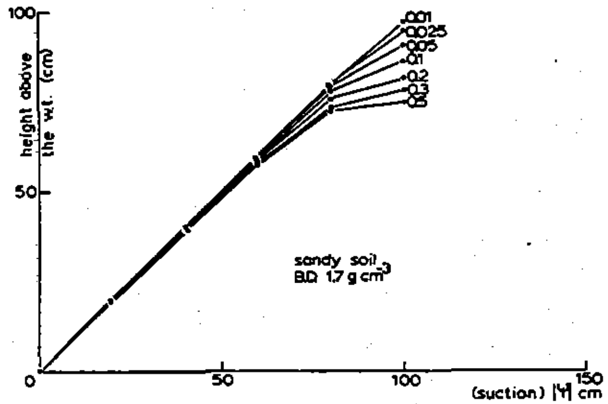
Table 4.

d. Bulk density 1.80 (g cm^{-3})

v(cm d^{-1})	0.5	0.3	0.2	0.1	0.05	0.025	0.001
ψ(cm)	z	z	z	z	z	z	z
20	19.89	19.93	19.96	19.99	19.99	19.99	20.00
40	39.62	39.77	39.85	39.92	39.96	39.98	40.00
60	59.2	59.53	59.69	59.84	59.92	59.96	60.00
80	78.20	79.90	79.26	79.63	79.81	79.91	80.00
100	90.17	92.93	94.71	96.96	98.34	99.13	99.96
116	91.79	95.39	98.05	102.21	105.83	108.78	114.00







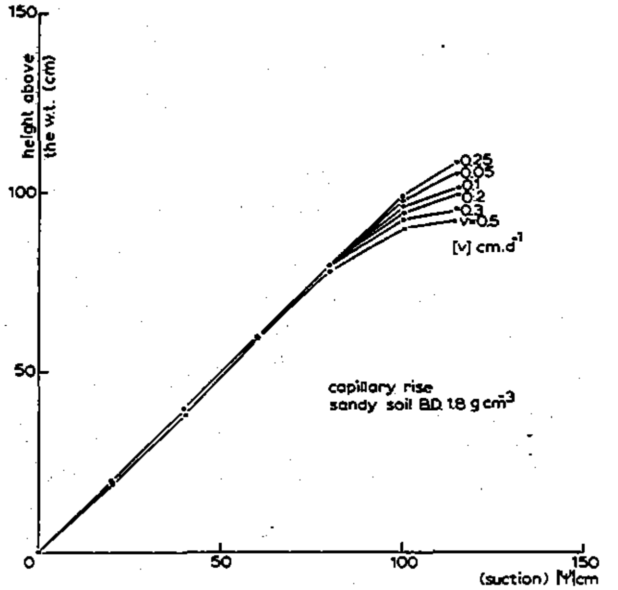


Fig. 7. Capillary rise computed for different flow velocities

4. WATER UPTAKE BY THE ROOTS

Two PVC pipes (120 cm long with diameter of 10.5 cm) were filled up from the bottom: 1.8 cm of gravel; 2 mm-thick layer of glass fiber; 108 cm of sandy soil at B.D. 1.8 g cm^{-3}; 10 cm of sandy soil (from the experimental field Sinderhoeve) at B.D. 1.0.

Holes for tensiometers (ϕ 1 cm) were drilled at the depths of: 5; 12; 17; 22; 32; 52; 82 cm along the wall of the pipes. At the bottom a hole for the water supply/drainage-Grass (lollium perenne) was sowed in the first cm of the columns soil at the moisture content of 18% of volume, suction 200 cm (STAKMAN et al. 1963) -

Fertilizers and water have been added. The columns were settled in a fully climatized room with constant air temperature (25^{o}C) relative humidity (30%) light intensity ($\pm$ 2%). When the grass covered completely the surface and reached the height of 5 cm, the measurements started. Tensiometers and pressure measurements devices were the same used in the determination of K(ψ) in the 20-cm-long columns.

4.1. Falling water table

The column was allowed to saturate connecting the bottom with top water; when the sandy soil at B.D. 1.8 was completely saturated (to 10 cm from the top of the column) the water supply was stopped and the falling water table has become into being. As the suction in the soil profile was increased the tensiometers were inserted in each hole and the measurements started.

In fig. 8 are given the suction profiles at different time intervals.

The moisture content, pertaining to the measured suction at a certain depth is read from the retention curve, thus converting the suction profile into a moisture content-profile. The total moisture content at each moment in the column is calculated. In table 5 the value of the moisture content during the evaporation.

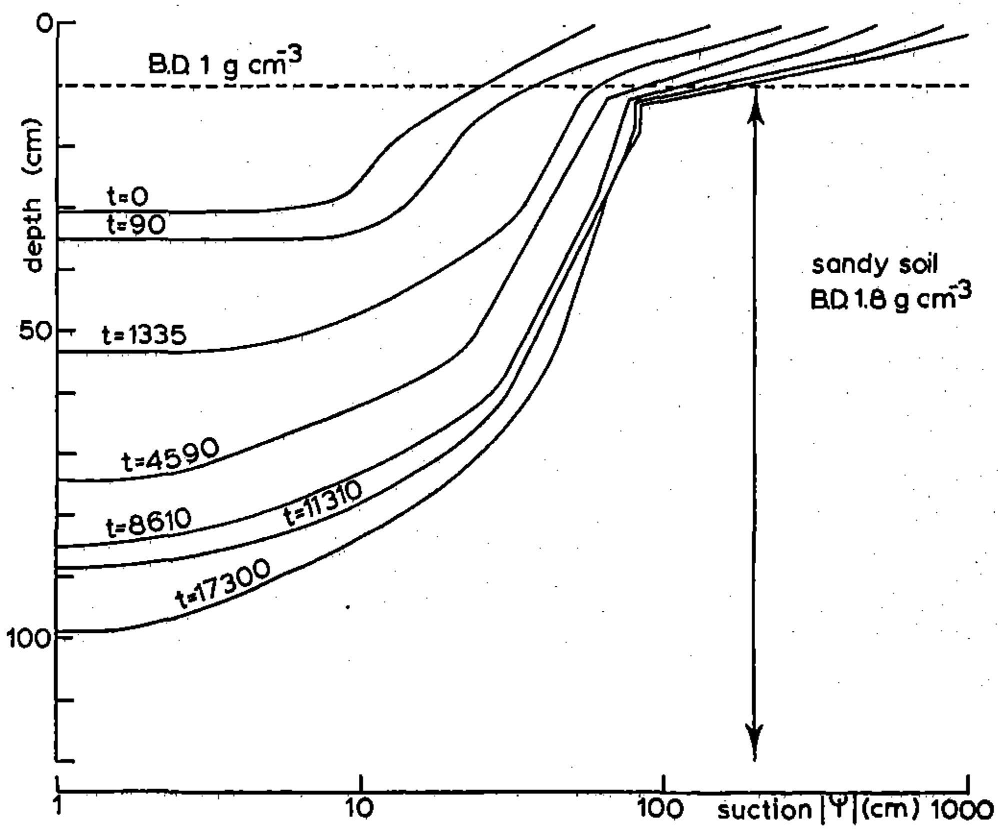


Fig. 8. Suction profiles in the 120-cm-long column during the falling down of water table

Table 5. Moisture content in the subsoil (θ_s) and in the root zone (θ_r)

t(min)	θ_s (cm)	θ_r (cm)
0	34.38	2.75
90	34.18	2.15
1 335	33.50	2,00
4 590	32.03	1.85
8 610	31.29	1.75
11 310	29.43	1.55
17 310	29.28	1.50

4.2. Steady state water table

The column has been connected to a Mariotti's bottle, the depth of the water table has been fixed at 70 cm below the soil surface. When the suction readings along the soil profile were constant the water table level has been lowered 10 cm. The suction profile for the two equilibrium situations are given in fig. 9.

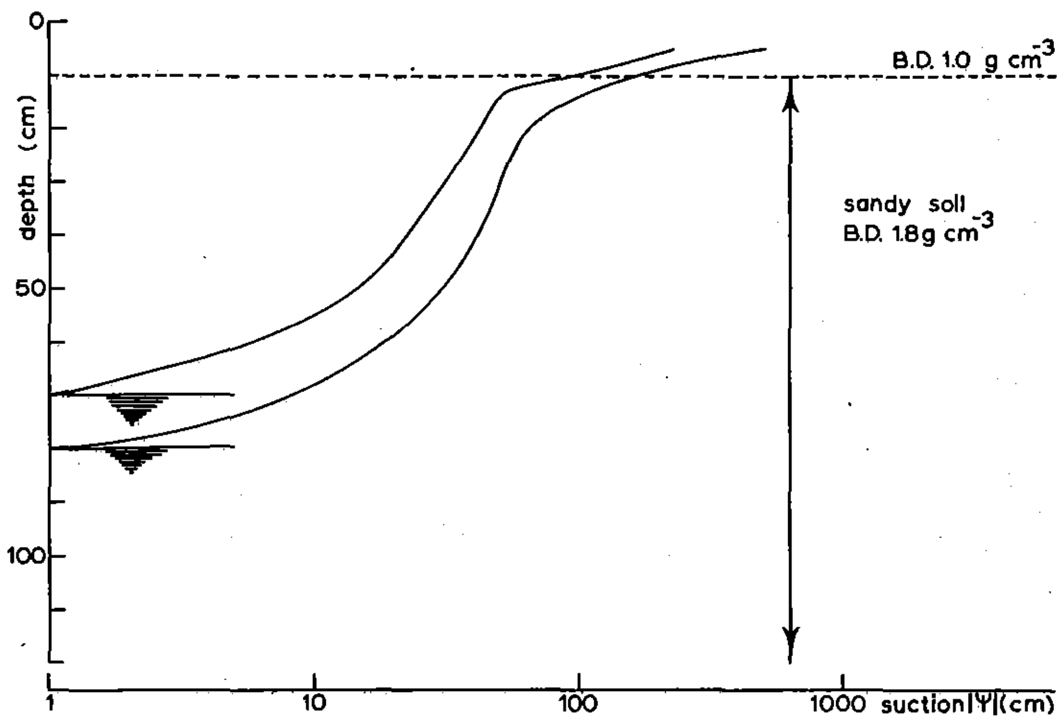


Fig. 9. Suction profiles in the 120-cm-long column equilibrium situations at two constant water table levels

The water supplied by the bottle has been measured too, the maximum evaporation rate has been determined - during the time between the two profiles -. The amount of water lost from the soil below the rooting zone was 8.1 mm d^{-1}, from the rooting zone 0.8 mm d^{-1}; the water supplied by the bottle 2.3 mm d^{-1}.

In the end of the experiment the columns were cut in slices and the soil moisture of each layer was measured (oven-drying at 105°C) and related to the suction measured at each depth with the tensiometers. The soil moisture retention curve is given in fig. 10.

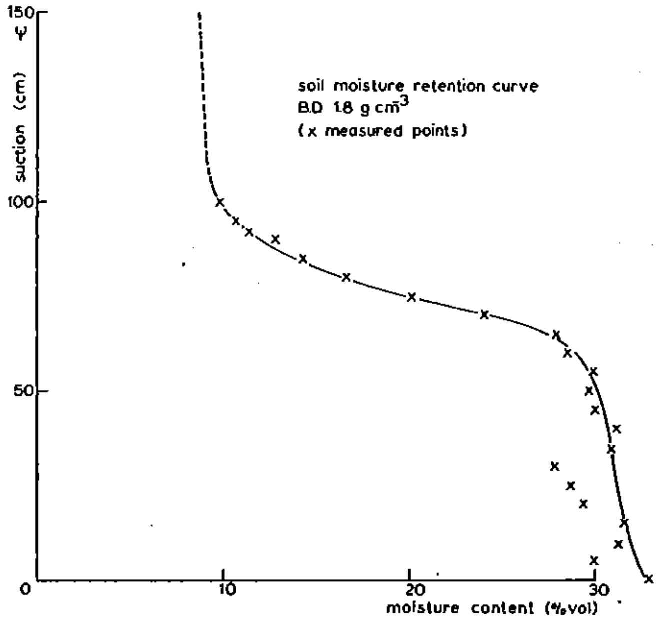


Fig. 10. Soil moisture retention curve for the sandy soil determined by soil samples taken out from the 120-cm-long column

The total amount of water, extracted from below the root zone in the experiments with a falling water table was 66 mm.

4.3. Water uptake by plants simulated with a simple method

Introduction

Relationship between water table level and water available for the crop in the soil, is not easily predictable because of the complexity of the solution of the flow equation. Very complete models can be used, which take in account the greatest possible number of environmental conditions (FEDDES en ZARADNY, 1977) in order to have a detailed quantitive description of the process of water uptake by the root zone.

If the requested knowledge of water withdrawel is relative only to a high permeable soil with high water table level, can be used a simple simulation method (DE LAAT, 1976) which describes the flow by a sequence of steady-state solutions. Here this simulation technique has been compared with experimental data obtained in artificially packed of column covered with grass.

Method

The soil column is subdivided in 2 homogeneous layers: the effective root zone (in the sense of RIJTEMA, 1969) and the subsoil i.e. the layer between root zone and saturated zone. The method used is based on the following assumptions (DE LAAT, 1976)

1. no vertical-flow in the root zone
2. the water uptake rate is constant with the depth in the root zone
3. water uptake from below the root zone occurs at the interface rootzone-subsoil
4. crop evapotranspiration equals the potential evaporation untill the wilting point in the root zone is reacted
5. reduction in evapotranspiration occurs if the wilting point is reacted, while from this moment the evaporation is determined by the capillary flow to the root zone
6. at each moment exists a constant flow-velocity between water table and root zone (the so called pseudo steady-state conditions).

For each day (or different time period) the water uptake rate balance can be written as:

$$E_o = E_r + V_w + \left(\frac{\Delta W_s + P}{\Delta t}\right) \qquad (1)$$

where

E_o cm d^{-1} = potential evapotranspiration

E_r cm d^{-1} = water uptake rate from the root zone

V_w cm d^{-1} = infiltration (+) or seapage (-) at the level of the water table

ΔW_s cm = change in available water content between root zone and water table

P cm = water supplied from above (rainfall or irrigation)

Δt d = time step

Under field conditions the potential evaporation, E_o, can be calculated according to RIJTEMA, 1965.

The initial conditions are that at a given water table a situation of equilibrium exists. Nor seapage neither infiltration are taken in account.

Data needed for the application of this method are:

1. initial water table depth and suction- or moisture content profile
2. capillary rise in relation to flow velocity and suction
3. soil moisture retention curve for both rootzone and subsoil
4. Maximum evapotranspiration rate during each period.

Iteration procedure (scheme of DE LAAT, 1976). First period -

i - assume an initial water uptake per day from the rootzone ($E_{i,r}$)

ii - calculate the available water in the root zone W_r

$$W_{i,r} = W_{i-1,r} + (E_o - E_{i,r})\ \Delta t$$

iii - From the fig. 11 read the value of $\psi_{i,r}$ corresponding to the value W_z:

$$W_r = \tfrac{1}{2}\ (W_{i,r} + W_{i+1,r})$$

iiii - Find (fig. 12) the height of the capillary rise $z_{i,r}$ for those given $\psi_{i,r}$ and $E_{i,r}$

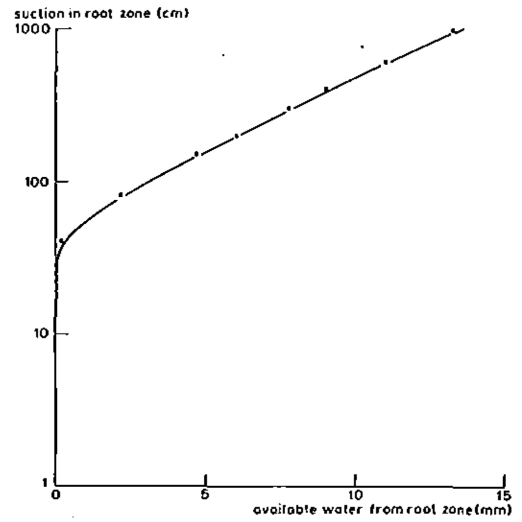


Fig. 11. Available water in the root zone from the initial suction 'up to the wilting point'

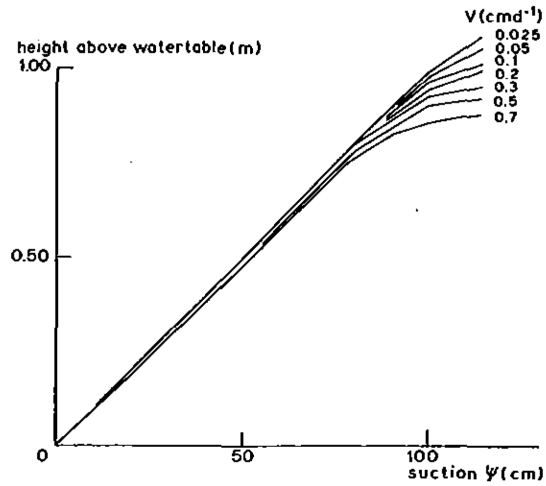


Fig. 12. Capillary rise in the subsoil. BD = 1800 kg m^{-3}

v - Read from fig. 13 for $z_{i,r}$ and $E_{i,r}$ the amount of water available in the subsoil $W_{i,s}$

vi - check if $W_{i,s}$ satisfies the water balance equation

$$W_{i,s} = W_{i-1,s} + E_{i,r}\,\Delta t$$

if it does not, start once again the iteration with a new $E_{i,r}$, chosen in such a way that:

if $W_{i,s} < W_{i-1,s} + E_{i,r}\,\Delta t$, start with $E^{II}_{i,r} < E^{I}_{i,r}$

or:

if $W_{i,s} > W_{i-1,s} + E_{i,r}\,\Delta t$, start with $E^{II}_{i,r} > E^{I}_{i,r}$

The iteration for the first period is stopped when the water balance equation (v_i) is satisfied. From the values of $W_{i,r}$ and $W_{i,s}$ calculated for the first period go on for the successive periods.

Experimental procedure

Column (length 120 cm, diameter 10.5 cm) has been filled with sandy soil at the bulk density of 1800 kg m^{-3} up to 0.10 m below the top. This high value of bulk density avoided the penetration of the roots in it satisfying the assumption of two separated layers namely root-zone and subsoil. The upper 0.10 m was filled up with sandy soil (humus content 4%) at the bulk density of 1000 kg m^{-3}, in which grass (Lollium perenne) was sowed at 0.5 cm of depth. Tensiometers were inserted in the column (from the top) : 0.05 m (in the rooting zone) 0.12, 0.17, 0.22, 0.32, 0.52, 0.82 m (in the subsoil). Water was supplied from the bottom till the subsoil was saturated; at this moment the water supply was interrupted and the successive levels of the falling-down water table were measured.

Soil moisture retention curves of both root-zone soil and subsoil as measured in small samples allowed to convert the suction profiles, successively determined from the tensiometer readings, into

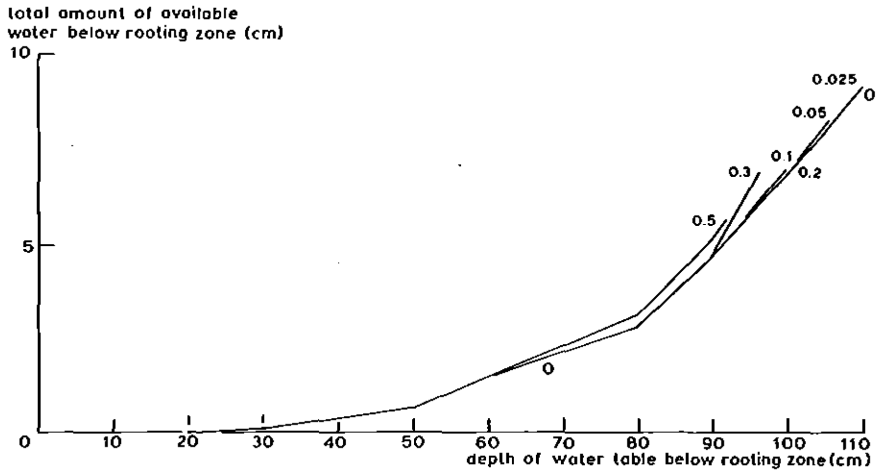


Fig. 13, Total amount of water available in the subsoil, calculated for different flow velocity, at different depths of the water table

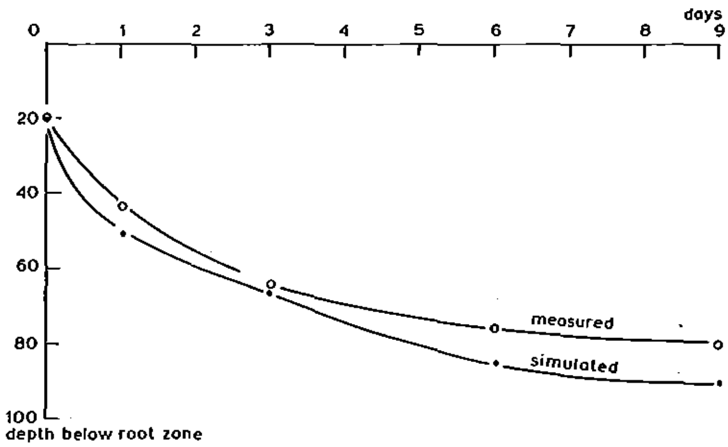


Fig. 14. Trend of the falling water table level

water content. The amount of water available at each time was calculated with a numerical integration. From these values the maximum dayly evaporation rate (water extracted from both subsoil and rootzone) has been found as 0.008 m^{d-1}.

The capillary rise in the subsoil was calculated (fig. 2) at different flow velocities from

$$z = \int_0^{\psi} \frac{k(\psi)}{E_{i,r} + k(\psi)} d\psi$$

with a numerical integration, the relationship between the hydraulic conductivity k and ψ (table 6) for this soil being known (pag. 15 table 3). The amount of water available in the subsoil was calculated in relation to the depth of the water table at different flow velocities (fig. 13).

Table 6.

ψ (cm)	k (cm d^{-1})
0	370
18	67
24	40
50	27
63	19
67	13
73	8
78	5
83	3
87	1
110	0.2

Results

Fig. 14 shows the trends of the measured and simulated water table level during the evaporation of the cropped soil column. It is evident the good fitting of the values.

Table 7.

	Water uptake from the subsoil (mm)	
days	simulated	from tensiometer readings
1	0.007	0.0068
3	0.022	0.0216
6	0.045	0.0289
9	0.064	0.0483

The amount of water withdrawn by the crop during the time is tentatively drown down (fig. 15) as a function of the lowering of water table level.

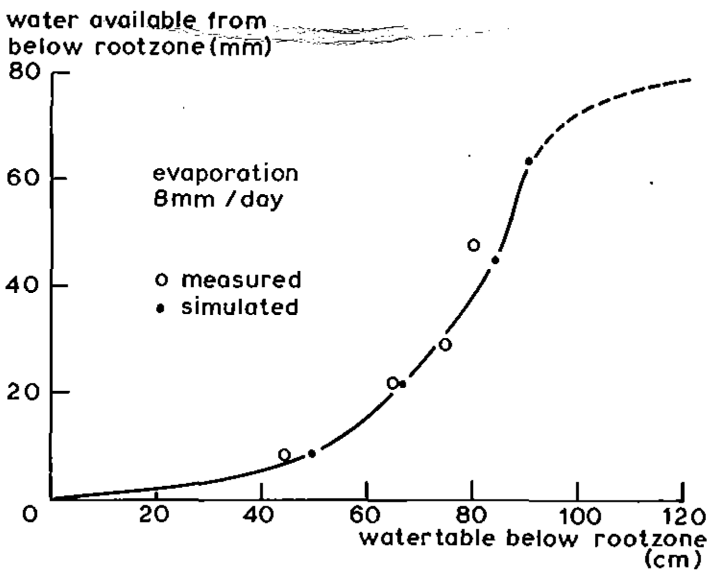


Fig. 15. Water available from the subsoil during the falling down of the water table

From measured water content profiles at the end of the experiment it was found that the total water extraction from below the rootzone was 0.066 m.

LITERATURE

BAKKER, J.W., 1975. Tensiometers, materialen en reactie-snelheden van systemen. Nota 847. Inst. for Land and Water Management Research, Wageningen, the Netherlands

BOELS, D., 1973. Bepaling van het capillair geleidingsvermogen en een deel van de pF-curve in een proefopstelling van het RID. Nota 742 Inst. for Land and Water Management Research, Wageningen, The Netherlands

_____ en H.D.A. VAN HEMMEN, 1975. Het effect van het losmaken van een ploegzool. Nota 858. Inst. for Land and Water Management Research, Wageningen, The Netherlands

_____ ,J.B.H.M. VAN GILS, G.J. VEERMAN, K.E. WIT, 1978. Determining Both Moisture characteristic and unsaturated conductivity of a large number of soil cores with an automatical Recording and controling system - soil science (in print)

CHILDS, E.C. and N. COLLIS-GEORGE, 1950. The permeability of porous materials. Proc. Roy. Soc. London A:201. 392-405

FEDDES, R.A., KOWALIK, P.J., ZARADNY, H., 1978. Simulation of field water use and crop yield. Centre for Agricultural Publishing and Documentation (Pudoc) Wageningen

GARDNER, W.R., 1956. Calculation of capillary conductivity form pressure plate outflow data. Soil Sci.Soc.Am.Proc. 20:317-320

GREEN, R.E. and J.C. COREY, 1971. Calculation of hydraulic conductivity: A further evaluation of some prodicitive methods. Soil Sci. Soc. Am. Proc. 35: 3-8

KUNZE, R.J., G. UEHARA and K. GRAHAM, 1968. Factor important in the calculation of hydraulic conductivity Soil Sci. Soc. Amer. Proc. 32: 760-765

LAAT, P.J.M. DE, 1976. A pseudo steady-state solution of water movement in the unsaturated zone of the soil. Journ. of Hydrology, 30, pp. 20-27, Amsterdam

MARSHALL, T.J., 1958. A relation between permeability and size distribution of pores. J. Soil Sci. 9: 1-8

RIJTEMA, P.E. 1965. An analysis of actual evapotranspiration. Pudoc. Wageningen, 107 pp.

_____ 1969. Soil moisture forecasting. Nota 513. Inst. for Land and Water Management Research, Wageningen. The Netherlands

STAKMAN, W.P., G.A. VALK & G.G. VAN DER HORST, 1969. Determination of soil moisture retention curves I, I.C.W. Wageningen

VEERMAN, G.J., 1976. Onderzoek naar de mogelijkheden van automatisering van de bepaling van capillair geleidingsvermogen en pF-curve. Nota 955. Inst. for Land and Water Management Research, Wageningen, The Netherlands

VERHAEGH, W.B., B. WIEBING, 1978. Improvement of podzolic soils, moisture extraction in soil from 1975 till 1977. Nota 1041 I.C.W., Wageningen

WESSELING, J., 1974. Hydraulic conductivity of natural soil columns. Soil science, 118 no. 1, pp 7-10

_____ and K.E. WIT, 1969. An infiltration method for the determination of the capillary conductivity of undisturbed soil cores. Symposium Water in the unsaturated zone; Vol. I IASH-AISH-Unesco 223-234

WIND, G.P., 1969. Capillary conductivity data estimated by a simple method. Symposium on water in the unsaturated zone, Wageningen, The Netherlands