DATA REDUCTION METHODS FOR FIELD ESTIMATED HYDRAULIC PROPERTIES

by

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This dissertation is dedicated to the spirit of my father, who guides me in every step of my life, and to the enduring love of my mother and my family for their patient support and encouragement.

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A B S T R A C T

Three $K(\Theta)$ functions were used to describe hydraulic conductivity data from a layered field soil. The functions were $K_0(\Theta/\Theta_m)^\beta$, $K_0((\Theta-\Theta_c)/(\Theta_m-\Theta_c))^n$ and $K_0\exp\{\alpha(\Theta-\Theta_m)\}$ where $K_0 = K_0(z) = g(z)K_m$, $g(z)$ is a scaling factor that varies with depth z, K_m a constant and Θ_c, Θ_m, β, n, and α are parameters. For each function three cases were considered: Case 1 fit discrete values by depth to $g(z)$ and α, β and n: The second case treated the scaling and exponential parameters as continous functions of depth: And the third case fit scaling factor as a discrete function of depth but held β, n and α constant. More variation in r^2 and MSE was found between cases than between functions.

INTRODUCTION

Most soil physics field studies to date have been directed at obtaining detailed descriptions of the various soil horizons or alternatively the course of action has been to ignore layering and treat the soil as being a uniform body. Describing a soil horizon by horizon is a tedious task with no clearly defined stopping point. Warrick et al. (1977) used scaling to reduce the amount of effort required to describe the hydraulic properties of small laboratory cores. The objective of this paper is to evaluate scaling as a technique for describing hydraulic properties of a layered field soil.

THEORY

Spatially-varying hydraulic conductivities can be approximated as the product of a function of depth and a function of volumetric water content (Warrick et al., 1977), i.e.:

$$K(\Theta, z) = g\tilde{K}(\Theta) \qquad [1]$$

where $K(\Theta,z)$ is the spatially-varying hydraulic conductivity, $g=g(z)$ is the scaling factor expressed as a function of depth (z), and $\tilde{K}(\Theta)$ is the nominal hydraulic conductivity and depends only on water content. Taking the log of both sides of Eq. [1] results in

$$\log\{K(\Theta, z)\} = \log\{g(z)\} + \log\{\tilde{K}(\Theta)\} \qquad [2]$$

Choose

$$\tilde{K}(\Theta) = K_m(\frac{\Theta}{\Theta_m})^{\beta} \quad [3]$$

$$\tilde{K}(\Theta) = K_m(\frac{\Theta - \Theta_c}{\Theta_m - \Theta_c})^{n} \quad [4]$$

$$\tilde{K}(\Theta) = K_m \exp\{\alpha(\Theta - \Theta_m)\} \quad [5]$$

where K_m is the hydraulic conductivity at the reference water content Θ_m. Θ_c, β, n, and α are parameters. Eq. [3], [4] and [5] were taken from Watson (1967), Brooks and Corey (1964) and Davidson et al. (1963), respectively. Substitution of one of Eq. [3], [4], or [5] into Eq. [2] results in

$$\log\{K(z,\Theta)\} = \log\{K_0(z)\} + \beta \cdot \log(\frac{\Theta}{\Theta_m}) \quad [6]$$

$$\log\{K(z,\Theta)\} = \log\{K_0(z)\} + n \cdot \log(\frac{\Theta - \Theta_c}{\Theta_m - \Theta_c}) \quad [7]$$

$$\ln\{K(z,\Theta)\} = \ln\{K_0(z)\} + \alpha(\Theta - \Theta_m) \quad [8]$$

where $K_0(z) = g(z)K_m$.

While no formal statistical test is known for choosing the best g(z) or functional form of $\tilde{K}$, a qualitative appraisal can be made on the basis of r^2 and mean square errors. Three cases for each of Eq. [6], [7], and [8] will be considered.

CASE 1:

$$\log\{K_0(z)\} = \log\{K_0(z_i)\}$$
$$\beta_i = \beta(z_i);\ n_i = n(z_i);\ \text{and}\ \alpha_i = \alpha(z_i)$$

where z_i denotes a depth where measurements were made and g_i, β_i, α_i and n_i are discrete estimates. This case required fitting functions at each individual depths.

CASE 2:

$$\log\{K_0(z)\} = \log\{K_0(z)\}$$
$$\beta = \beta(z);\ n = n(z);\ \text{and}\ \alpha = \alpha(z)$$

where $\log\{K_0(z)\} = a_0 + a_1 z + a_2 z^2 + a_3 z^3 + a_4 z^4 + a_5 z^5$

$$\alpha(z), \beta(z) \text{ or } n(z) = b_0 + b_1 z + b_2 z^2 + b_3 z^3 + b_4 z^4 + b_5 z^5$$

and $a_0, a_1, \ldots, a_5$ and $b_0, b_1, \ldots, b_5$ are regression coefficients.

CASE 3:

$$\log\{K_0(z)\} = \log\{K_0(z_i)\}$$

β, n and α are constant over all depths.

Case 3 results in a single relation between K and Θ, $\tilde{K}(\Theta)$, which will be hereafter referred to as the nominal hydraulic conductivity function.

Materials and Methods

The field study was carried out on Muir silt loam (fine- silty, mixed, mesic Pachic Haplustoll) located on Kansas State University Ashland Research Farm about 10 km south of Manhattan, Kansas. The Muir series consists of deep nearly level soils on river and creek terraces formed in deep alluvium. The 3 by 4 m test plot was bermed with soil and sand bags. An aluminum access tube, 4.13 cm outside diameter and 183 cm long, was installed to a depth of 168 cm in the center of the plot 4 months before starting infiltration on 14 July 1983. Mercury-manometer tensiometers were arranged in a circular pattern, 80 cm in diameter around the access tube at the 0, 20, 40, 60 , 80, 100, 120, 140, and 160 cm depths. Water was maintained at a depth of 5 to 6 cm on the plot until soil water pressure and Θ appeared constant with time. The plot was then covered with plastic and 2 to 3 cm of soil to prevent evaporation.

Neutron moisture observations were made at 12 depths 10, 20, 30, 40, 50, 60, 70, 80, 100, 120, 140, and 160 cm as the mean of two 64s counts. During the early drainage phase, readings were taken every 2 to 4h for the first 12h then less frequently at larger times. Soil water content measurements were made immediately after the tensiometer readings. Hydraulic conductivities were estimated using a modified instantaneous profile method described by Rose et al. (1965).

Percentages of sand, silt, and clay were determined using the hydrometer method (Day, 1965). Except for the upper 15 cm and the lower 30 cm (135 -165 cm) where the soil texture is silt loam the soil profile is predominantly a silty clay loam. Average particle-size distribution for each depth, given in Table 1, shows that the mean fraction of clay, silt, and sand were 28, 54, and 18%, respectively. Table 1 also shows that the average bulk density value for the top 100 cm is 1.4 $Mg \cdot m^{-3}$; this value decreased to a minimum at 140 cm with the lower 60 cm of the profile having an average density of 1.32 $Mg \cdot m^{-3}$.

The neutron probe was calibrated by regressing volumetric water content vs. neutron probe count using data from borings obtained from the plot area following the infiltration-drainage experiment. To extend the range of the field calibration to the wet end the area was reflooded and soil samples taken. Bulk densities and volumetric water content were estimated from 7.6 by 7.6 cm undisturbed cores taken with a thin-walled hydraulic probe. There were a few exceptions where the average bulk density was used to convert gravimetric water to a volume basis. The narrow range of the clay content and bulk density (Table 1) suggested using one neutron probe calibration curve for all depths (Fig. 1). Observations at 10 cm depth, where the calibration was affected by the restricted soil volume near the surface, were deleted from analysis.

Chemical analysis of the well water used in this study was as follows: 0.6 mmol L^{-1} Na^{+}, 0.1 mmol L^{-1} K^{+}, 15.2 mmol L^{-1} Ca^{2+}, 3.6 mmol L^{-1} Mg^{2+}, 0.21 mmol L^{-1} Cl^{-}, 1.4 mmol L^{-1} SO_4^{2-}, 0.2 mmol L^{-1} NO_3^{-}, 10 mmol L^{-1} HCO_3^{-}, pH=7.08, and EC = 0.85 $dS \cdot m^{-1}$.

Results and Discussion

Water content profiles at 0.0, 0.2, 0.5, 1.4, 3.4, 7, 12, 18.3, and 28.4d are presented in Fig. 2. Measured water contents are indicated by the solid circles. Average water content change above 70 cm was 0.03 $m^3 \cdot m^{-3}$ during drainage. A larger change, 0.06 $m^3 \cdot m^{-3}$, occured below 70 cm. The rate of change of water content, $\partial\theta/\partial t$, during early drainage phase was about 80 times greater than the rate of change of water during late drainage phase.

Figure 3 shows total hydraulic head plotted against depth. During the 28-day drainage period, the pressure head at deeper depths, 100, 120, 140, and 160 cm, changed as much as twice that of the upper depths. Hydraulic head gradients approached a magnitude of 3 near the end of the drainage phase.

Three $K(\Theta)$ were fitted at each depth for Case 1. Θ_c in the Brooks and Corey Model was estimated from curve fitting total water above z (W) data,

$$W = W(z,t) = \int_0^z \Theta dz \qquad [9]$$

where t is time in days. For the Brooks and Corey Model and assuming a unit gradient Sisson et al. (1980) expressed W as,

$$W = W(z,t) = \Theta_c z + (1-1/n) z (\Theta_m - \Theta_c) \left(\frac{z}{At}\right)^{1/(n-1)} \qquad [10]$$

here Θ_m was estimated as the average of $\Theta(z,t)$ after 16 days of ponding. Note that the n here is 1/n in Sisson et al., (1980). A is given by $A = K_m/[(\Theta_m-\Theta_c)/n]$, K_m was estimated from the final infiltration rate at the soil surface after 16 days of ponding, and Θ_c and n were estimated by non-linear least squares fitting to Eq. [10] (PROC NLIN procedure provided by Helwig and Council, 1979).

The three $K(\Theta)$ functions (Eq. [6], [7] and [8]) were fitted for Case 1. The results for the 140 cm depth are shown in Fig. 4 and were considered typical of most depths. Estimates of the parameters are given in Table 2. The three functions performed equally well in that similar r^2 values were obtained at each depth. The results from other depths tended to follow the sigmoidal shape shown in Fig. 4. None of the $K(\Theta)$ functions cosidered here could mimic such a shape. The eleven regression lines resulting from fitting Eq. [8] for Case 1 are shown in Fig. 5.

The slope of each regression curve obtained from fitting the three different $K(\Theta)$ functions for individual depths, was tested against all other slopes and the results of this test are presented in Table 3. This test was a t-test dependent on the difference between two slopes as well as individual standard errors. This test gave the same ranking of exponential parameters (slopes) regardless of the $K(\Theta)$ function. Slopes from the 120 and 60 cm depths had the largest differences (Table 2). The ranking of exponents remain the same regardless of the $K(\Theta)$ function. It may be concluded from Table 3 that significant differences existed among exponential parameters at the one percent probability level.

Case 2 fit the slopes of Eq. [6], [7] and [8] as polynomials of z. The maximum r^2 improvement technique was used to fit polynomial models to the field data (Helwig and Council, 1979). Estimates of the polynomial coeffients are given in Table 4. The continuous polynomial functions produced higher MSE and lower r^2 values than Case 1 (Table 5). The continuous polynomials offer the advantage of estimating hydraulic conductivity at any depth over the 0 - 160 cm range.

Case 3 requires fitting curves of the same slope to Eq. [6], [7] and [8]. The fitting was done using the method of dummy variables (Draper and Smith, 1981). The results of this fitting are shown in Fig. 6 for the Davidson model. A numerical comparison of cases is presented in Table 5. MSE values of the single slope model were about twice that of the variable slope models. Cases 1 and 2 also produced r^2s that were 9 to 15% higher than the single slope model. The parallel curve models explained more than 74% of the variation of log{K} or ln{K} around the means (Table 5). The basic advantage of the parallel model is that only one function needs to be estimated to describe a soil.

Case 3 allows a single $\tilde{K}(\Theta)$ for the whole profile, once the scaling factor g(z) has been estimated. Assuming the Davidson Model K_m for the nominal hydraulic conductivity was estimated as the log mean of $K_0(z)$ over all depths and is shown in Fig. 7 as the solid curve. Hydraulic conductivity data from all depths were adjusted by the difference between $\ln K_0(z)$ and $\ln K_m$ and also plotted on Fig. 7. A regression of the data in Fig. 7 indicated an r^2 of 0.91+ considereably higher than most of the r^2s given in Table 2 and any of the r^2s in Table 5. If the r^2s in Table 5 had not been adjusted for the mean of the dependent variable they would also exceed 0.90.

Further field work would be required to establish how reliably the single curve predicts hydraulic properties of the Muir silt loam other locations.

<u>Summary and Conclusions</u>

1. Three K(Θ) were tested their ability to describe field measured K(Θ) data. Exponents in the three K(Θ) functions were allowed to vary with depth in two tests and were fixed to a single value in a third test. All three K(Θ) functions performed equally well as determined by r^2, although the r^2 for the variable and fixed models were different.

2. When the exponents in the three K(Θ) functions were fitted as discrete and polynomial functions of depth, the discrete functions yielded higher r^2 (0.89). When one exponent was used for all depths, the r^2 ranged from 0.74 - 0.78. The advantage of the polynomial procedure was at points between the depths studied could be interpolated. The advantage of fixed slope model is that only one exponent needed to be determined for all eleven depths. The discrete model had the highest precision.

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Fig. 1. Neutron probe calibration for the Muir silt loam.

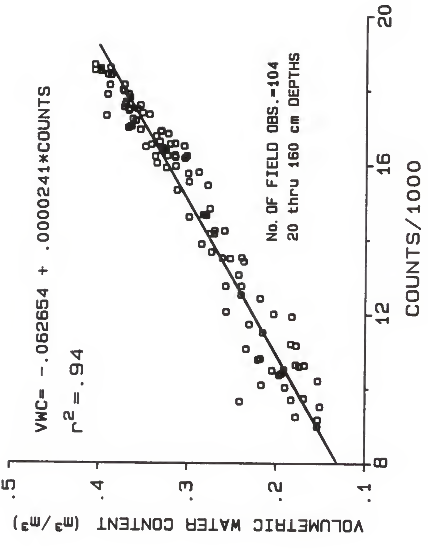

VWC= -.062654 + .0000241*COUNTS
r^2 = .94
No. OF FIELD OBS.=104
20 thru 160 cm DEPTHS
COUNTS/1000
8
12
16
20
VOLUMETRIC WATER CONTENT (m^3/m^3)
.1
.2
.3
.4
.5

Fig. 2. Soil water content profiles of Muir silt loam during drainage cycle.

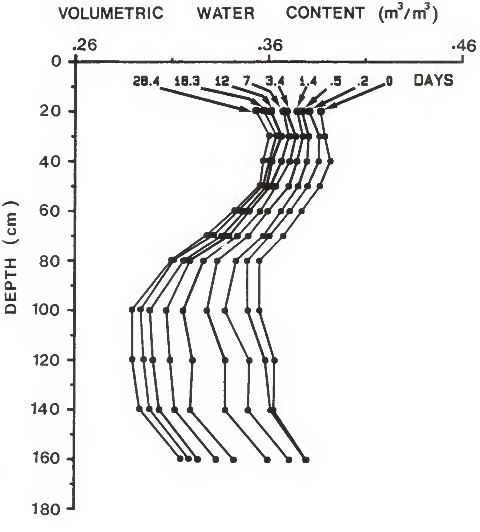

VOLUMETRIC WATER CONTENT (m³/m³)
.26
.36
.46
28.4
18.3
12
7
3.4
1.4
.5
.2
0
DAYS
DEPTH (cm)
0
20
40
60
80
100
120
140
160
180

Fig. 3. Hydraulic head profiles of Muir silt loam during drainage cycle.

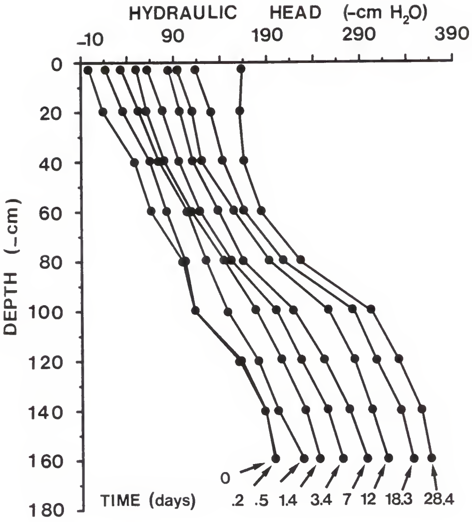
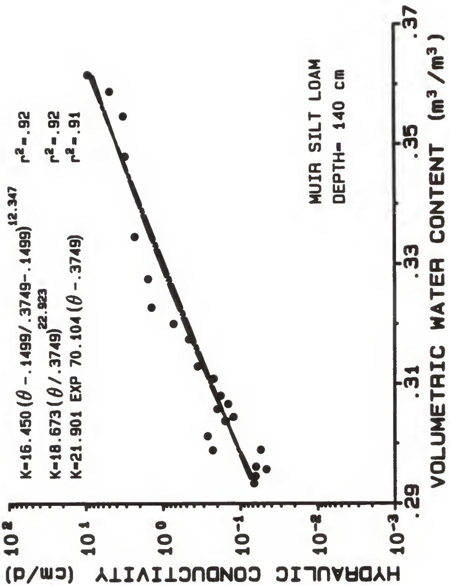

HYDRAULIC HEAD (-cm H_2O)
-10
90
190
290
390
DEPTH (-cm)
0
20
40
60
80
100
120
140
160
180
TIME (days)
0
.2
.5
1.4
3.4
7
12
18.3
28.4

Fig. 4. Hydraulic conductivity volumetric water content relations of the 140 cm depth. Solid lines were obtained from fitting the discrete slope and intercept functions to field measured K-θ data and the circles refer to instantaneous profile method.



$K=16.450(\theta-.1499/.3749-.1499)^{12.347}$ $r^2=.92$
$K=18.673(\theta/.3749)^{22.923}$ $r^2=.92$
$K=21.901$ EXP $70.104(\theta-.3749)$ $r^2=.91$
MUIR SILT LOAM
DEPTH= 140 cm
10^2
10^1
10^0
10^{-1}
10^{-2}
10^{-3}
HYDRAULIC CONDUCTIVITY (cm/d)
.29
.31
.33
.35
.37
VOLUMETRIC WATER CONTENT (m^3/m^3)

Fig. 5. Hydraulic conductivity volumetric water content relations for all depths. Solid lines were obtainad from fitting Eq. [8] as a discrete model to the field measured K-θ data.

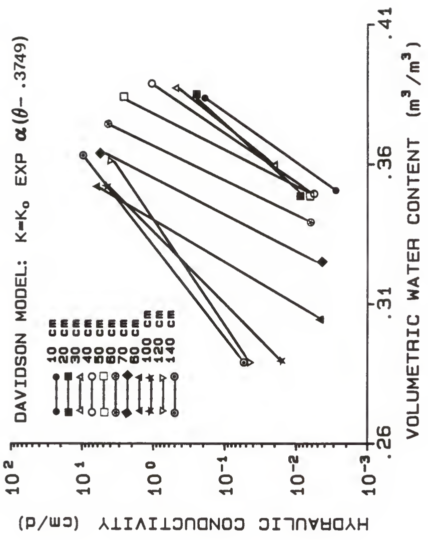
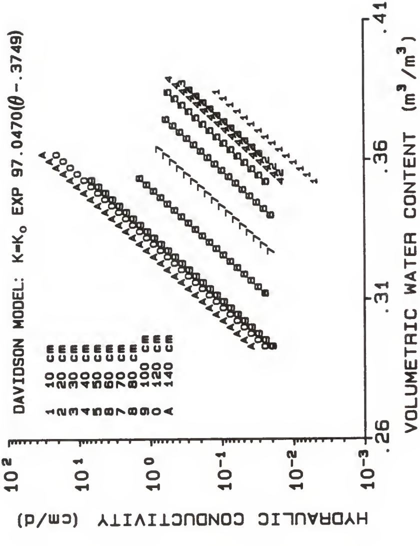
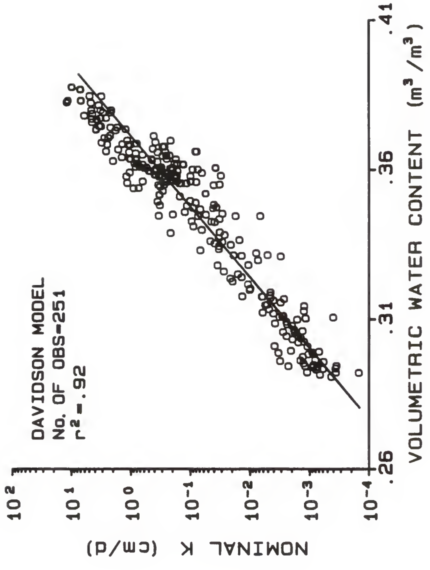

DAVIDSON MODEL: K=K$_o$ EXP α(θ− .3749)
10 cm
20 cm
30 cm
40 cm
50 cm
60 cm
70 cm
80 cm
100 cm
120 cm
140 cm
HYDRAULIC CONDUCTIVITY (cm/d)
10^2
10^1
10^0
10^{-1}
10^{-2}
10^{-3}
VOLUMETRIC WATER CONTENT (m^3/m^3)
.26
.31
.36
.41

Fig. 6. Hydraulic conductivity volumetric water content relations for all depths. Solid lines were obtained from fitting Eq. [8] as a fixed exponent model to field measured K-θ data.



DAVIDSON MODEL: K=K₀ EXP 97.0470(θ-.3749)
1 10 cm
2 20 cm
3 30 cm
4 40 cm
5 50 cm
6 60 cm
7 70 cm
8 80 cm
9 100 cm
0 120 cm
A 140 cm
HYDRAULIC CONDUCTIVITY (cm/d)
10²
10¹
10⁰
10⁻¹
10⁻²
10⁻³
.26
.31
.36
.41
VOLUMETRIC WATER CONTENT (m³/m³)

Fig. 7. Nominal hydraulic conductivity based on Eq. [8].


DAVIDSON MODEL
No. OF OBS=251
r^2=.92
NOMINAL K (cm/d)
10^2
10^1
10^0
10^{-1}
10^{-2}
10^{-3}
10^{-4}
.26
.31
.36
.41
VOLUMETRIC WATER CONTENT (m^3/m^3)

LIST OF TABLES

Table 1. Average particle-size distribution and bulk density of Muir silt loam.

Soil depth cm	%Sand	%Silt	%Clay	Bulk density $Mg.m^{-3}$
5 - 15	20.0	55.0	25.0	1.40
15 - 25	16.0	54.5	29.5	1.43
25 - 35	15.0	54.0	31.0	1.38
35 - 45	14.0	54.0	32.0	1.34
45 - 55	15.5	54.5	30.0	1.35
55 - 65	17.0	55.5	27.5	1.40
65 - 75	16.0	56.0	28.0	1.42
75 - 85	16.0	55.5	28.5	1.37
95 - 105	14.0	57.0	29.0	1.33
115 - 125	18.0	54.0	28.0	1.32
135 - 145	22.0	51.0	27.0	1.29
Means	16.7	54.6	28.7	1.37

Table 2. Values of r-square, intercept (K_o), and slope for all depths obtained from fitting the three $K(\theta)$ functions as discrete models to in situ measured $K-\theta$ data.

Depth cm	θ_o $m^3.m^{-3}$	Discrete function: Watson r^2	Watson K_o $cm.d^{-1}$	Watson β	Brooks & Corey r^2	Brooks & Corey K_o $cm.d^{-1}$	Brooks & Corey n	Davidson r^2	Davidson K_o $cm.d^{-1}$	Davidson α
10	0.387	0.72	0.061	46.60	0.72	0.061	27.57	0.72	0.061	126.9
20	0.387	0.59	0.104	36.12	0.58	0.104	21.34	0.59	0.104	98.56
30	0.389	0.63	0.110	41.74	0.63	0.111	24.95	0.63	0.109	111.9
40	0.391	0.83	0.166	48.84	0.83	0.168	29.13	0.83	0.164	131.3
50	0.386	0.89	0.557	65.01	0.89	0.561	38.55	0.89	0.550	176.4
60	0.377	0.89	4.35	66.30	0.89	4.20	38.87	0.88	4.58	185.5
70	0.367	0.92	25.7	60.25	0.92	22.6	34.04	0.91	30.9	174.6
80	0.355	0.94	132	50.07	0.94	99.2	27.25	0.94	193	151.9
100	0.355	0.91	23.4	28.47	0.92	18.3	15.05	0.90	32.0	89.05
120	0.363	0.86	7.16	19.84	0.87	6.40	10.64	0.85	8.27	60.90
140	0.363	0.92	18.7	22.92	0.92	16.5	12.35	0.91	21.9	70.10

Table 3. All possible significant differences between slopes of the fitted regression curves of the discrete functions.

Depth (cm)	120	140	100	20	30	10	40	80	70	50	60
120			**		**	**	**	**	**	**	**
140						**	**	**	**	**	**
100	**						**	**	**	**	**
20								**	**	**	**
30	**								**	**	**
10	**	**									**
40	**	**	**								**
80	**	**	**	**							
70	**	**	**	**	**						
50	**	**	**	**	**						
60	**	**	**	**	**	**	**				

** Depths are significantly different at the 0.01 level.

Table 4. Values of the final fitted parameters of the slope and intercept as functions of depth obtained from fitting the three $K(\theta)$ functions as continuous polynomials of depth to the field measured $K-\theta$ data.

Regression coefficient	Estimated value based on fitting as a polynomial: Watson model	Brooks & Corey model	Davidson model
a_o	$-2.7011 * 10^{0}$	$-2.6901 * 10^{0}$	$-6.2376 * 10^{0}$
a_1	$2.3789 * 10^{-1}$	$2.3854 * 10^{-1}$	$5.4337 * 10^{-1}$
a_2	$-1.1495 * 10^{-2}$	$-1.1505 * 10^{-2}$	$-2.6306 * 10^{-2}$
a_3	$2.4081 * 10^{-4}$	$2.3993 * 10^{-4}$	$5.5425 * 10^{-4}$
a_4	$-2.0490 * 10^{-6}$	$-2.0370 * 10^{-6}$	$-4.7294 * 10^{-6}$
a_5	$5.9942 * 10^{-9}$	$5.9535 * 10^{-9}$	$1.3850 * 10^{-8}$
b_o	$4.1250 * 10$	$2.5424 * 10$	$1.0632 * 10$
b_1	00000000000	00000000000	00000000000
b_2	$7.8687 * 10^{-3}$	$3.6032 * 10^{-3}$	$2.7696 * 10^{-2}$
b_3	00000000000	00000000000	00000000000
b_4	$-1.8934 * 10^{-6}$	$-9.7421 * 10^{-7}$	$-6.0079 * 10^{-6}$
b_5	$1.0320 * 10^{-8}$	$5.4055 * 10^{-9}$	$3.2146 * 10^{-8}$

Table 5. Values of r-square, sum square error (SSE), and mean square error (MSE) obtained from fitting the three $K(\theta)$ functions as discrete, continuous polynomial and and fixed exponent models to field measured $K-\theta$ data.

$K(\theta)$ function	Model								
	Discrete			Polynomial			Fixed exponent		
	r^2	SSE	MSE	r^2	SSE	MSE	r^2	SSE	MSE
Watson	0.89	15.02	0.0656	0.86	18.69	0.0836	0.77	31.02	0.1298
Brooks & Corey	0.89	14.75	0.0644	0.86	18.68	0.0775	0.74	32.80	0.1373
Davidson	0.88	15.40	0.0672	0.86	18.78	0.0779	0.78	29.12	0.1218

Appedices

Appendix A

Literature Review

Understanding of the water movement into and through soil profile is of significant importance because it is the basic input for irrigation and drainage systems design, soil management practices, crop selection and land use, and environmental aspects. These practices are all, to a certain extent, dependent upon the hydraulic properties of soil which include the water and conductivity chracteristic that relates soil water content with pressure potential and hydraulic conductivity, respectively. Water infiltration and redistribution are time dependent and essential for solute transport through soil. Gardner and Widtsoe (1921) developed an equation to relate the mean values of water content following irrigation with time, T, in days

$$W = 14.6 + 7.6\ e^{-0.02T} + 7.6\ e^{-0.04T}$$

where W represents the mean value of water contents measured to a depth of 183 cm and expressed as a percentage on a dry weight basis. While considering the physical processes involved in loss of water from a field plot of a uniform sandy loam after irrigation, Richards et al. (1956) found that the total water, W, above a given depth could be closely related to time by an equation of the form

$$W = aT^{-b}$$

where a and b are costants, and that the rate of loss of soil water was inversely proportional to time, T, i.e;

$$dW/dT = -abT^{-b-1} \qquad = -bW/T$$

Using the same equation, Wilcox (1959) noted that straight lines were obtained when log W was plotted vs. log T. Also he reported that the fit was quite close except during the early drainage phase and that in any one soil equal water contents are accompanied by equal rates of drainage. Richards and Weeks (1953) suggested a method to calculate the hydraulic conductivity from data obtained during transient changes of water content and tension in soil column. The basis of their calculation was the Darcy equation

$$q=Ki$$

where q=flux, K is the hydraulic conductivity, and i is the hydraulic head gradient. They rewrote the equation in the form

$$1/A * \partial Q/\partial T = Ki$$

where $\partial Q/\partial T$ is the rate of flow of water past a given cross section area, A, of the soil column. The derivative $\partial Q/\partial T$ was evaluated using a method suggested by Richards (1938) to calculate the volume of water, Q, in a soil column as follows:

$$Q = A \int_0^z \rho_b g dz$$

where ρ_b is the soil bulk density, A and z are the cross sectional area and length of the soil column, respectively, and g is the gravimetric water content. This integral was evaluated at different times during desorption along the soil column to calculate the flux ($\partial Q/\partial T$ = q), then the flux was divided by the hydraulic gradient, obtained from tensiometers, to calculate the hydraulic conductivity. The method of Richards and Weeks was modified by Richards et al. (1956) and Ogata and Richards (1957) for the analysis of field observations to determine the relation of water content to suction and hydraulic conductivity of a fine sandy loam of uniform profile. However, they expressed reservations on the possibilty of in situ measurement of the hydraulic conductivity of unsaturated soil profile that was not uniform.

During the late 50's the most reliable work, concerning the rate of drainage following irrigation, was that in which plots of bare soils were covered to prevent evaporation and to prevent rainfall from falling, due to the unavailability of an independent measurement of evaporation. Wilcox (1959) studied this phenomena on bare soils which were irrigated with sufficient water to wet the soils to field capacity and then covered to prevent evaporation. The results indicated that the finer textured the soil the greater were the rates of moisture loss, however, slope of the line which relates the rate of moisture loss to time was much less with a clayey than with a sandy soil. He also noted that the moisture content following desorption increased progressively with depth because with increasing depth each successive layer of soil received more and more water from above.

While tensiometers are the most common and satisfactory devices to measure soil water tension and hence the hydraulic head gradient, the neutron method of measuring soil water has the advantages of precision and cost compared to gravimetric sampling because it satisfies the most common requirement, the nondestructive attribute, by allowing repeated sampling in the same access hole when measuring changes in stored water. This technique was used by Burrows and Kirkham (1958) and Nielsen et al. (1959) to measure changes in stored water of a soil profile.

Rose et al. (1965) represented a theory based on the water conservation equation for a given volume of vegetation-free soil to determine the hydraulic conductivity in the field as a function of depth over the entire range of water contents on a soil of non-uniform profile. This method is known in the literature as the instantaneous profile method. On using this method, there are three options with respect to the measurement of soil water content and soil water pressure head profiles: i) in situ measurement of water content and pressure distributions, ii) in situ measurement of water content distribution and inferred pressure head from water retention data , and, iii) in situ measurement of pressure head and inferred water content from retention data. The first option is inherently the best choice Klute (1972), because the in situ determined Θ(h) curves often disagree with Θ(h) curve determined on undisturbed core samples collected from the same site. This disagreement has been more experienced with fine-textured soils (Luxmoore et al. 1981), but it also occurs for the coarse-textured soils (Dane, 1980). At one location, reasonably accurate and precise values can be obtained after infiltration during the redistribution and drainage of water within and from soil profile where measurements were made for both soil water contents and soil water pressure (Nielsen et al. 1964 ; van Bavel et al. 1968a).

Wilcox (1960) investigated the effect of the rate of drainage following irrigation on the determination of consumptive use. He assumed that the drainage rate from a given depth of a soil is a function only of water content and independent of the rate of extraction of water by roots. He underestimated the consumptive use when the drainage rates from a covered plot was deducted from the total loss of a nearby cropped plot, and over estimated the consumptive use when the total moisture lost from a cropped plot was used as estimate of the consumptive use. The theory of the instantaneous profile method, referred to hereafter as the IPM, was modified by Rose et al., (1967) to permit the calculation of water withdrawal by plant roots as a function of depth and time with out neglecting water movement in the soil. Using this method to separate between water redistribution and uptake by plant roots it requires:

1. Knowledge of change in water storage which is given by

$$\int_{0}^{z}\int_{T_1}^{T_2} (\partial\theta/\partial T)\, dz\, dT$$

2. An estimate of the vertical flux q which is given by the Darcy equation

3. An estimate of evaporation which is distinct from transpiration

$$\int_{T_1}^{T_2} E\, dT = - \int_{T_1}^{T_2} q_o\, dT$$

where q_0 is the upward flow at the soil surface which occured in a liquid phase and that the vapor flow is small in comparison.

4. Amount of water which is withdrawal by roots from a soil with an upper boundary at the soil surface, a lower boundary at depth z is given by

$$\int_0^z r(z)\,dz = \int_{T_1}^{T_2} (I-q-E)\ dT - \int_0^z \int_{T_1}^{T_2} (\partial\theta/\partial T)\,dz\,dT$$

where I=rate of irrigation including precipitation, and r(z) = time-averaged rate of water withdrawal by roots over the period T1 to T2.

They concluded that the method can be used succesfully to determine in situ the pattern of water withdrawal from the soil by a growing crop.

The instantaneous profile technique may also be applied to laboratory flow columns and has the advantage of comparing the flux in the still saturated zone at selected profile times with the outflow rate per unit area of the same times as measured from the volume outflow at the base of the column. Also the arrangement of measuring the soil water content and soil water suction with the gamma rays absorption and tensiometer pressure tranceducer respectively, permits the simultaneous nondestructive and rapid response measurement of soil water content and soil water suction and provide a record of these changes against time. Watson (1966) proposed this technique and applied it to an initially saturated sand column where measurements of soil water content and pressure head as a function of depth and time were made. In his investigation, Watson (1966) used the one dimensional continuity equation for unsaturated materials to calculate the flux at given time intervals by integrating graphically $\partial\Theta/\partial T$ with respect to depth, then the gradient was determined by differentiating graphically the total potential with respect to depth at the same time intervals. The instantaneous hydraulic conductivity was then calculated by dividing the flux by the gradient.

Before 1967, a problem associated with the studies of separating the drainage and consumptive use (root absorption plus evaporation) for an actual field situation was that in none of these studies was there an independent measure of evaporation rate. Van Bavel et al. (1968a) determined the evaporation losses from three precision weighing lysimeters in which the surface and environment were closely identical to those of the test plots. They used the IPM (Watson 1966) to calculate the flux and the hydraulic conductivity from in situ measured water content and pressure potential by determining $\partial\Theta/\partial T$ and $\partial H/\partial z$ at a selected group of time intervals using a graphical technique and integrating $\partial\Theta/\partial T$ for depth increments under consideration to get the flux, which was divided by the hydraulic gradient to get the hydraulic conductivity at a given depth. The rate of root extraction, r(z), was then calculated using

$$r(z) = \partial\Theta/\partial T - \partial q/\partial z$$

Using this method, van Bavel et al. found that the calculated root extraction rates agreed reasonably with the independent lysimetric measurements of the water loss from the surface to the atmosphere.

In bare soils, infiltration ,evaporation, and deep percolation depend, in some measure, upon the water conent of soil profile. Black et al. (1969) showed that the cumulative evaporation for a bare Plainfield sand at any stages was proportional to the square root of time following each heavy rainfall and can be calculated from the diffusivity measurements, and that the drainage rate was an exponential function of total water stored above a given depth of soil profile. Miller et al. (1971) studied the effect of evaporation rate on drainage losses at different depth. They found that drainage losses increased and extended over longer times as the evaporation decreased, and that the drainage losses at 120 cm were greater and extended over a longer period of time than that at 70 cm depth .

Arya et al. (1975) described another field method to determine $K(\Theta)$ and $\Theta(h)$ which requires, as well as the IPM, either direct measurements of both soil water content and soil water pressure head profiles or the direct measurements of either one of these variables and the indirect determination of the other variable through the separately determined water retention curve. In this method the boundary between the upward and downward movement of water, the plane of zero flux, was positioned as it moved down the soil profile with evaporation and drainage occurring simultaneously. Arya et al. graphically evaluted the hydraulic head gradient from total potential, measured in the field using tensiometers, plotted as a function of depth at various times. Then the graphically determined values of the hydraulic head gradient were plotted against time to determine the position of zero flux plane. This method has the same basis of calculating the flux as that suggested by Richards (1956) and the same basis of inferring the hydraulic head gradient as that suggested by Watson (1966).

Olsson and Rose (1978) used the IPM to determine the hydraulic conductivity characteristics of a soil profile that is subjected to volume changes with changes in water content, swelling soils, from in situ measurements of water content and suction during the redistribution of water through the profile of a red-brown earth which exhibits swelling properties. They noted that at a given water suction, the hydraulic conductivity was generally lower in subsoil where micropores dominate these layers imposing a high resistance on water flow and thus reducing the bulk velocity of water for a given potential gradient.

Reliable estimates of unsaturated hydraulic conductivity are difficult to obtain due to extensive variability in the field and cost. For this reason equations have been developed by Childs and Collis-George (1950), Marshal (1958), and Millington and Quirk (1959, 1960, 1961) to calculate the unsaturated conductivity from pore size distribution which can be characterized easily by the standard measurements of water content vs. pressure. These methods have been tested with some success by Jackson et al. (1965), Kunze et al. (1968), Green and Corey (1971), and Campbell (1974). In developing these equation it is assumed that the soil is isotropic with respect to hydraulic conductivity, an assumption of uncertain validity in field soils. van Genuchten (1978, 1980) derived a closed form analytical solution based on both Burdine theory (1953) and Mualem theory (1976a) for predicting the hydraulic conductivity from the soil moisture curve. This model contains two or three independent parameters which may be obtained from soil moisture retention data by a non-linear least- squares curve-fitting method with the aid of digital computers. van Genuchten showed that the soil water content, θ, as a function of pressure head, h, is given by:

$$\theta = \theta_r + \frac{\theta_s - \theta_r}{[\;1+(\alpha h)^n]^m}$$

where α, n, and m are unknown parameters that may be determined from $\theta(h)$ curve, and θ_r and θ_s are the residual and saturated value of θ which can be determined experimentally. θ_r may be considered as the volumetric water content at a large value of pressure head (assuming h is inversely proportional to θ) where $dH/d\theta \approx 0$ which on this basis assigns θ_r a positive value and a physical basis when either Burdine or Mualem three parameters model is considered. However, Ward et al. (1983) indicated that θ_r should be considered as a fitting parameter rather than a soil property, which allows θ_r to be less than zero. Also Stephens et al. (1985) found that the lower limit of the 95% confidenc intervals of θ_r was less than zero, which is physically impossible. Van Genuchten model was tested by van Genuchten (1980), Dane (1980), Ward et al. (1983), and Stephens et al. (1985) and found to be convenient and sufficiently accurate for field application, but the reliability of the model may depend upon a reasonable estimate of θ_r and θ_s.

The IPM is perhaps the most reliable method for determining the saturated as well as the unsaturated hydraulic conductivities for field conditions. Basis of this method is the Darcian-based flow theory. However, Application of this theory to field is time consuming, expensive to characterize the water storage and conductivity functions, and the uncertainity of representation of measurements to large areas owing to the inherent spatial variability of a field soil (Rogowski, 1972). Technical problems and soil profile charactristics may limit the application of the theory of the IPM (Baker et al. 1974).

Field hydraulic properties have been measured, by Black et al. (1969); Davidson et al. (1969); Nielsen et al. (1973); Luxmoore et al. (1981); Libardi et al. (1980); Chong et al. (1981); Sisson et al. (1980); and Jones et al. (1984) assuming that the moisture content and suction over a substantial length are uniform which implies that the potential gradient is only gravitational. In these methods the soil is wetted deeply and allowed to drain while evaporation is prevented. Under this assumption, the method is known in the literature as the the unit gradient method. Using this method for covered and non-vegetated uniform or weakly layered soil proflie with a shallow water table, where the hydraulic gradient is nearly unity and water content is a function of time and independent of depth, it can be shown that the average hydraulic conductivity, K, for a soil layer between the soil surface (z_0 =0) and depth of interest (D=-z_1) is given by

$$K = D * (\partial\Theta/\partial T)$$

where Θ is the average volumetric water content above the depth D and K is the hydraulic conductivity evaluated at the water content Θ and depth D. One of the advantages of this method is that the soil water content profiles, measured as a function of time following steady state infiltration conditions in a field soil, provide the only data necessary to estimate K(Θ).

Libardi et al. (1980) developed a method to obtain K(Θ) only assuming a unit hydraulic gradient, the hydraulic conductivity is an exponential function of soil water content, and that the average soil water content between soil surface and depth L is a function of soil water content. This method was developed by integrating Richards equation similar to that integration done by Nielsen et al. (1973). They concluded that the calculated K(Θ) may differ from reality at each site especially with soils having distinct layers of horizons of greatly differing hydraulic conductivity. The limitations of this method are only K(Θ) information is obtained and the necessity to assume unit gradient (Dane et al., 1983). The calculated K(Θ) values using Libardi method seemed to correspond fairly well with the corresponding K(Θ) from IPM at intermediate values of Θ, but it did not correspond at all at higher values of Θ (Dane, 1980). This behaviour may be explained by the fact that during the early drainage period (which corresponds to low absolute value of pressure potential, h, and higher Θ values), $\partial H/\partial z$ <0, but after several days of drainage (which corresponds to higher absolute value of h and lower Θ values), $\partial H/\partial z$ >0 (Chong et al. 1981). However, when K(Θ) evaluated using Libardi method was compared with K(Θ) detrmined using the IPM , Schuh et al. (1984) noted that Libardi method worked well on coarse and fine textured, homogeneous materials underlying stratified soil materials, and adequate slopes of K(Θ) curves were obtained within layered soil materials but the calculated K(Θ) as a matching values were often inadequate.s suggesting that the fit between calculated and measured values of hydraulic conductivity could be improved by using field-saturated K(Θ) values.

Jones et al. (1984) compared five different methods, including Libardi method, to estimate K(Θ) from in situ measurements assuming a unit hydraulic gradient. They concluded from their study that the five approximate methods were useful in developing a rapid and rough estimate of soil water properties over large areas, but the methods were not as useful for a particular location where soil water properties need to be precisely known.

Ahuja et al. (1980) proposed a method to determine the hydraulic conductivity as well as the soil water characteristic by analyzing of the drainage phase tensiometric data, combined with the field measured value of near saturated hydraulic conductivity and one soil moisture sampling during drainage. This method, when applied for data obtained at five sites chosen for maximum variation of materials between and within sites, resulted in a good estimate of K(h) as determined with the IPM (Schuh et al. 1984). Unless an accurate determination of saturated hydraulic conductivity, from steady state infiltration prior to drainage, is used as a matching value as suggested by Ahuja et al. (1980) the matching value should be representative of the actual log K vs. log h (where h is the pressure potential) to avoid inaccurate results (Schuh et al. 1984).

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Appendix B

Table 1. Volumetric water content vs. time during flooding cycle.

Date and time	Depth (cm) 20	30	40	50	60	70	80	100	120	140	160
	Volumetric water content $m^3 \cdot m^{-3}$										
07 14 1983 1405	0.256	0.281	0.281	0.285	0.292	0.286	0.267	0.250	0.235	0.254	0.267
07 14 1983 1413	0.336	0.307	0.307	0.302	0.306	0.300	0.293	0.322	0.250	0.251	0.267
07 14 1983 1430	0.335		0.308		0.312		0.296	0.326	0.257	0.252	0.274
07 14 1983 1454	0.340	0.313	0.310	0.310	0.311	0.306	0.295	0.326	0.263	0.252	0.271
07 14 1983 1500	0.342	0.317	0.313	0.305	0.312	0.310	0.296	0.331	0.267	0.252	0.264
07 14 1983 1515	0.349	0.314	0.311	0.303	0.312	0.310	0.296	0.328	0.260	0.258	0.269
07 14 1983 1530	0.354	0.326	0.309	0.305	0.310	0.310	0.299	0.327	0.279	0.254	0.271
07 14 1983 1630	0.357	0.332	0.312	0.307	0.310	0.305	0.297	0.329	0.282	0.252	0.271
07 14 1983 1830	0.364	0.340	0.315	0.310	0.312	0.310	0.295	0.320	0.308	0.253	0.270
07 14 1983 2130	0.372	0.351	0.325	0.312	0.310	0.305	0.295	0.307	0.284	0.254	0.270
07 15 1983 830	0.376	0.376	0.364	0.341	0.314	0.306	0.299	0.306	0.271	0.257	0.270
07 15 1983 2030	0.376	0.380	0.376	0.370	0.347	0.328	0.297	0.299	0.270	0.259	0.269
07 16 1983 830	0.380	0.380	0.379	0.377	0.361	0.350	0.323	0.300	0.272	0.259	0.266
07 16 1983 1630	0.400	0.402	0.401	0.394	0.361	0.353	0.336	0.309	0.268	0.256	0.268
07 17 1983 0000	0.378	0.379	0.380	0.377	0.366	0.353	0.340	0.326	0.277	0.260	0.268
07 17 1983 830	0.378	0.381	0.382	0.375	0.363	0.353	0.342	0.328	0.297	0.261	0.268
07 17 1983 1630	0.398	0.403	0.401	0.395	0.385	0.375	0.360	0.347	0.329	0.269	0.272
07 18 1983 830	0.383	0.384	0.382	0.377	0.367	0.357	0.341	0.332	0.328	0.300	0.274
07 18 1983 1630	0.377	0.379	0.379	0.376	0.365	0.353	0.343	0.336	0.333	0.316	0.284
07 19 1983 830	0.379	0.380	0.381	0.376	0.367	0.357	0.343	0.335	0.340	0.328	0.319
07 19 1983 1630	0.379	0.381	0.383	0.376	0.365	0.357	0.343	0.336	0.342	0.333	0.335
07 20 1983 830	0.379	0.382	0.380	0.380	0.369	0.355	0.344	0.337	0.342	0.336	0.348
07 20 1983 1630	0.379	0.383	0.381	0.378	0.366	0.355	0.343	0.337	0.343	0.338	0.352
07 21 1983 830	0.385	0.382	0.383	0.382	0.372	0.357	0.348	0.338	0.349	0.343	0.363
07 22 1983 830	0.382	0.388	0.379	0.378	0.365	0.358	0.347	0.339	0.345	0.344	0.361
07 23 1983 830	0.384	0.384	0.385	0.380	0.369	0.357	0.344	0.345	0.349	0.347	0.364
07 24 1983 1100	0.385	0.386	0.388	0.377	0.369	0.358	0.347	0.347	0.346	0.348	0.366
07 25 1983 1115	0.388	0.388	0.389	0.385	0.368	0.358	0.348	0.343	0.350	0.350	0.368
07 26 1983 1130	0.387	0.389	0.388	0.385	0.370	0.363	0.351	0.346	0.353	0.350	0.371
07 28 1983 1400	0.386	0.389	0.390	0.384	0.372	0.364	0.351	0.349	0.357	0.351	0.371
07 29 1983 1130	0.382	0.389	0.389	0.385	0.371	0.364	0.352	0.350	0.358	0.357	0.372
07 30 1983 1130	0.385	0.391	0.391	0.388	0.374	0.366	0.357	0.352	0.364	0.359	0.381
07 31 1983 850	0.387	0.388	0.391	0.386	0.377	0.367	0.355	0.355	0.363	0.363	0.380

Table 2. Tensiometric pressure head vs. time during flooding cycle.

Date and time	Depth (cm) 0	20	40	60	80	100	120	140	160
	Pressure head (−cm H_2O)								
07 14 1983 1400	832.8	300.2	157.9	203.4	473.2	443.1	372.7	386.7	370.5
07 14 1983 1830	−1.2	1.1	13.5	24.7	179.6	206.2	206.4	272.1	363.0
07 15 1983 800	−1.2	6.1	3.4	19.6	303.1	399.0	250.5	383.0	360.4
07 15 1983 2030	−1.2	3.6	6.7	15.8	53.6	435.6	270.7	385.5	360.4
07 16 1983 830	−1.2	6.1	11.7	15.8	27.2	399.0	278.2	380.4	357.9
07 16 1983 1630	−1.2	2.3	18.0	13.3	27.2	419.2	287.0	328.8	350.4
07 17 1983 0000	−1.2	2.3	18.0	13.3	24.6	367.5	279.5	375.4	355.4
07 17 1983 830	−1.2	4.8	15.5	12.1	27.2	125.6	244.2	372.9	350.4
07 17 1983 1630	−1.2	2.3	20.5	9.5	28.4	82.8	173.6	331.3	344.1
07 18 1983 830	−1.2	3.6	20.5	9.5	28.4	81.7	86.7	320.0	346.6
07 18 1983 1630	−1.2	1.0	20.5	8.3	30.9	81.7	77.4	211.6	330.2
07 19 1983 830	−1.2	1.0	20.5	7.0	30.9	74.6	45.1	112.1	150.0
07 19 1983 1630	−1.2	2.7	21.8	13.1	32.2	65.4	67.8	94.4	99.6
07 20 1983 830	2.5	8.6	21.8	11.6	29.7	53.4	57.7	80.6	73.2
07 20 1983 1630	2.5	7.3	23.1	14.4	29.7	57.2	57.7	74.3	66.9
07 21 1983 830	1.2	2.7	21.8	18.1	24.6	50.4	52.7	71.7	58.0
07 22 1983 830	−1.2	1.0	23.1	13.1	23.4	40.9	53.9	65.4	58.0
07 23 1983 830	2.5	7.3	21.8	10.6	20.9	40.9	51.4	62.9	51.7
07 24 1983 1100	−1.2	8.6	21.8	13.1	29.7	40.9	48.9	60.4	51.7
07 25 1983 1115	−1.2	7.3	21.8	13.1	32.2	23.0	22.4	57.9	48.0
07 26 1983 1130	−1.2	8.6	21.8	6.8	29.7	18.0	45.1	50.3	44.2
07 27 1983 1200	−1.2	2.3	20.5	8.1	30.9	10.5	45.1	51.6	32.9
07 28 1983 1400	−1.2	−1.0	15.5	3.1	17.1	13.0	42.6	51.6	19.1
07 29 1983 1200	−0.1	−1.5	14.2	1.8	22.1	14.3	42.6	51.6	25.3
07 30 1983 1730	−1.2	−2.3	14.2	6.8	12.0	14.3	38.8	49.1	26.5
07 31 1983 945	−1.2	−6.1	13.0	5.0	19.6	12.8	40.1	47.8	19.0

Table 3. Volumetric water content vs. time during drainage cycle.

Date and time	Depth (cm) 20	30	40	50	60	70	80	100	120	140	160
	Volumetric water content $m^3 . m^{-3}$										
07 31 1983 1100	0.387	0.389	0.391	0.386	0.377	0.367	0.355	0.355	0.363	0.363	0.380
07 31 1983 1130	0.385	0.387	0.390	0.385	0.376	0.365	0.354	0.355	0.362	0.363	0.380
07 31 1983 1530	0.381	0.386	0.386	0.380	0.371	0.360	0.349	0.349	0.358	0.361	0.379
07 31 1983 1730	0.380	0.384	0.385	0.377	0.368	0.356	0.348	0.345	0.358	0.357	0.378
07 31 1983 1930	0.379	0.382	0.383	0.378	0.366	0.358	0.345	0.343	0.357	0.356	0.377
07 31 1983 2130	0.379	0.380	0.381	0.378	0.365	0.356	0.342	0.340	0.353	0.353	0.375
07 31 1983 2330	0.377	0.381	0.380	0.375	0.366	0.357	0.343	0.337	0.350	0.350	0.371
08 01 1983 800	0.376	0.378	0.380	0.372	0.362	0.352	0.336	0.333	0.344	0.344	0.365
08 01 1983 2000	0.376	0.378	0.374	0.370	0.360	0.349	0.333	0.328	0.338	0.338	0.360
08 02 1983 2000	0.373	0.376	0.373	0.365	0.357	0.347	0.330	0.320	0.327	0.326	0.347
08 03 1983 800	0.370	0.374	0.371	0.365	0.355	0.345	0.328	0.318	0.326	0.324	0.342
08 03 1983 2000	0.369	0.375	0.371	0.364	0.356	0.344	0.326	0.316	0.321	0.320	0.342
08 04 1983 800	0.369	0.374	0.369	0.364	0.354	0.342	0.324	0.314	0.320	0.320	0.340
08 04 1983 2000	0.369	0.375	0.369	0.362	0.354	0.342	0.326	0.313	0.317	0.318	0.338
08 05 1983 800	0.369	0.373	0.368	0.361	0.353	0.341	0.324	0.310	0.315	0.315	0.337
08 05 1983 2000	0.369	0.373	0.366	0.361	0.353	0.340	0.323	0.311	0.316	0.315	0.337
08 07 1983 1100	0.367	0.370	0.366	0.361	0.350	0.340	0.319	0.307	0.309	0.312	0.333
08 08 1983 1100	0.365	0.370	0.366	0.360	0.350	0.339	0.319	0.305	0.309	0.310	0.332
08 09 1983 1030	0.365	0.369	0.366	0.360	0.347	0.338	0.317	0.302	0.307	0.307	0.331
08 10 1983 1145	0.363	0.368	0.365	0.361	0.348	0.337	0.317	0.303	0.305	0.306	0.328
08 11 1983 1930	0.362	0.368	0.363	0.360	0.347	0.336	0.317	0.299	0.304	0.305	0.324
08 12 1983 1200	0.361	0.366	0.361	0.358	0.348	0.336	0.316	0.299	0.300	0.304	0.324
08 15 1983 1950	0.359	0.366	0.360	0.358	0.347	0.334	0.315	0.298	0.299	0.303	0.322
08 16 1983 2000	0.358	0.366	0.362	0.358	0.346	0.333	0.315	0.295	0.294	0.300	0.321
08 17 1983 2000	0.358	0.364	0.360	0.358	0.345	0.332	0.315	0.294	0.293	0.299	0.320
08 18 1983 1930	0.357	0.364	0.360	0.357	0.344	0.332	0.311	0.294	0.296	0.299	0.319
08 19 1983 1930	0.356	0.364	0.359	0.357	0.345	0.332	0.314	0.293	0.296	0.297	0.319
08 20 1983 1920	0.356	0.364	0.359	0.357	0.345	0.332	0.311	0.293	0.295	0.296	0.319
08 21 1983 1920	0.356	0.363	0.360	0.356	0.345	0.331	0.311	0.292	0.294	0.296	0.320
08 22 1983 1930	0.355	0.363	0.359	0.356	0.344	0.331	0.311	0.292	0.293	0.295	0.316
08 23 1983 1930	0.355	0.362	0.359	0.356	0.343	0.329	0.310	0.292	0.293	0.294	0.316
08 25 1983 1930	0.355	0.362	0.357	0.356	0.343	0.328	0.310	0.292	0.291	0.294	0.316
08 28 1983 1930	0.353	0.360	0.357	0.355	0.342	0.328	0.310	0.290	0.290	0.294	0.315

Table 4. Tensiometric pressure head vs. time during drainage cycle.

Date and time	Depth (cm) 0	20	40	60	80	100	120	140	160
	Pressure head ($-cm\ H_2O$)								
07 31 1983 1100	−1.2	−6.1	13.0	5.6	19.6	12.2	40.1	47.8	19.0
07 31 1983 1130	0.1	−1.0	13.0	9.4	19.6	13.4	40.1	45.2	15.2
07 31 1983 1530	17.7	16.6	25.6	23.2	23.3	13.4	43.8	49.0	17.7
07 31 1983 1730	20.2	19.1	34.4	29.5	24.6	18.5	50.1	50.3	55.5
07 31 1983 1930	24.0	27.9	34.4	34.6	43.5	24.8	55.2	60.4	61.8
07 31 1983 2130	29.0	31.7	34.4	43.4	44.8	31.1	57.7	60.4	65.6
07 31 1983 2330	34.1	33.0	34.4	44.6	44.8	37.4	60.2	61.6	69.3
08 01 1983 800	37.9	33.0	35.7	44.6	49.8	66.3	70.3	70.4	75.6
08 01 1983 2000	50.5	41.8	40.7	49.7	64.9	78.9	86.7	91.9	87.0
08 02 1983 2000	55.5	48.1	42.0	53.5	64.9	95.3	93.0	100.7	99.6
08 03 1983 800	61.8	53.1	44.5	56.9	58.6	96.6	98.0	98.2	103.4
08 03 1983 2000	63.1	59.4	57.1	59.7	73.7	110.4	108.1	115.8	112.2
08 04 1983 800	68.1	68.3	62.1	66.0	73.7	104.1	110.6	112.0	114.7
08 04 1983 2000	70.6	69.5	65.9	69.8	77.5	115.5	116.9	122.1	121.0
08 05 1983 800	70.6	77.1	70.9	72.3	76.3	114.2	119.4	120.8	122.3
08 05 1983 2000	75.6	73.3	70.9	74.8	85.1	114.2	127.0	136.0	131.1
08 07 1983 1100	84.5	77.1	70.9	78.6	85.1	119.3	132.0	138.5	137.4
08 08 1983 1100	88.3	78.3	74.7	87.4	97.7	138.2	139.6	139.7	143.7
08 09 1983 1030	87.0	80.9	77.2	83.7	95.2	138.2	148.4	148.6	147.5
08 10 1983 1145	93.3	87.2	78.5	90.0	101.5	147.0	153.4	149.8	153.8
08 11 1983 1930	92.0	88.4	73.5	91.2	107.8	157.1	163.5	167.5	161.3
08 12 1983 1200	94.6	90.9	81.0	96.3	114.1	157.1	164.8	157.4	160.1
08 15 1983 1950	100.9	98.5	92.4	101.3	121.6	174.7	178.6	186.4	176.4
08 16 1983 2000	109.7	104.8	101.2	102.6	125.4	178.5	183.7	190.1	181.5
08 17 1983 2000	109.7	108.6	103.7	105.1	127.9	184.8	187.5	193.9	184.0
08 18 1983 1930	113.5	109.8	103.7	106.3	127.9	183.5	187.5	193.9	186.5
08 19 1983 1930	114.7	114.9	106.2	107.6	130.4	183.5	191.2	196.4	187.8
08 20 1983 1920	121.0	118.7	101.2	103.8	126.7	178.5	188.7	190.1	186.5
08 21 1983 1920	126.1	119.9	110.0	110.1	131.7	188.6	195.0	197.7	191.6
08 22 1983 1930	138.7	123.7	111.3	111.4	133.0	188.6	195.0	200.2	191.6
08 23 1983 1930	150.0	130.0	113.8	112.6	135.5	191.1	197.5	201.5	197.9
08 25 1983 1930	160.1	135.0	122.6	122.7	143.0	199.9	205.1	209.0	201.6
08 28 1983 1930	163.8	141.3	126.4	125.2	148.1	203.7	212.7	216.6	206.7

DATA REDUCTION METHODS FOR FIELD ESTIMATED HYDRAULIC PROPERTIES

by

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AN ABSTRACT OF A MASTER'S THESIS

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A B S T R A C T

Three $K(\Theta)$ functions were used to describe hydraulic conductivity data from a layered field soil. The functions were $K_O(\Theta/\Theta_m)^\beta$, $K_O((\Theta-\Theta_c)/(\Theta_m-\Theta_c))^n$ and $K_O\exp\{\alpha(\Theta-\Theta_m)\}$ where $K_O = K_O(z) = g(z)K_m$, $g(z)$ is a scaling factor that varies with depth z, K_m a constant and Θ_c, Θ_m, β, n, and α are parameters. For each function three cases were considered: Case 1 fit discrete values by depth to $g(z)$ and α, β and n: The second case treated the scaling and exponential parameters as continous functions of depth: And the third case fit scaling factor as a discrete function of depth but held β, n and α constant. More variation in r^2 and MSE was found between cases than between functions.