Estimating Hydraulic Conductivity for Models of Soils with Macropores

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Abstract: A simple and efficient method was developed to determine soil macroporosity and hydraulic conductivity for dual-porosity models from measurements of unconfined infiltration rates. The utility of this method was demonstrated by analyzing unconfined infiltration tests conducted with a tension infiltrometer at ponded conditions and at negative 3, 6, and 15 cm of water-supply pressure. The conductivity of soil macropores (pores > 1 mm in diameter) was 3.6 times the conductivity of the soil matrix. This contrast in the magnitudes of the hydraulic conductivity may justify the use of dual-porosity models of water flow and solute transport. Positive but small correlation between soil macroporosity and hydraulic conductivity of the soil matrix was identified. Soil macroporosity remained constant near the surface but it decreased with soil depth. The narrow range of 0 to −15 cm of water pressure may govern water flow and contaminant transport under field conditions because of the rapid decrease of conductivity with water pressure, which was reflected by a short macroscopic-length scale. The method described in this paper relied on measurements of unconfined infiltration rates to minimize soil disturbance. This method provided several advantages over measuring the macroporosity with double-ring infiltrometer, which requires driving a ring into the soil to establish one-dimensional flow.

INTRODUCTION

Measurements of the soil hydraulic conductivity are needed for modeling the transport of solutes and contaminants in particulate form in the unsaturated zone. The transport of these contaminants is enhanced particularly by the presence of soil macropores, e.g., root channels, worm holes, and fractures, that form a secondary porosity besides the primary matrix porosity. Because water infiltrating through a macropore channel encounters little resistance as compared to water infiltrating through the soil matrix, elevated water velocities as reported by Mosley (1982) and Pilgrim et al. (1978) or even turbulent flow (Peterson and Dixon 1971) might be initiated in macropores channels. Hence, even though macropores usually comprise a small fraction of the total porosity of the soil, they accelerate the leaching of contaminants in the unsaturated zone as suggested by Thomas and Phillips (1979).

Ahuja and Hebson (1992), Jarvis (1991), and Beven and Clarke (1986), among others, developed dual-porosity models to account for the influence of soil macropores on the transport of contaminants. In a dual-porosity model, the flow of water is subdivided into two domains: (1) The macropore flow domain, which represents the structural porosity embedded in the soil; and (2) the micropore flow domain that constitutes the bulk of the soil matrix. Techniques developed to measure soil hydraulic conductivity should allow proper characterization of the hydraulic conductivity K_{mac} (LT^{-1}) and the macroporosity θ_{mac} (L^3/L^3) of the macropore domain as well as the hydraulic conductivity of the micropore or soil-matrix domain. The hydraulic conductivity of the soil matrix should be characterized, at least, by its saturated conductivity K_{sm} (LT^{-1}) and its macroscopic length scale λ (L). The macroscopic length scale is an integrated measure of the relationship between the unsaturated hydraulic conductivity of the soil matrix and soil-water pressure (White and Sully 1987), and can be thought of as the representative capillary fringe height above a water table. It is also equivalent to the wetting front suction in the Green and Ampt infiltration model (Morel-Seytoux et al. 1994; Ahuja and Hebson 1992). The importance of λ in soil physics and its connection to several unsaturated hydraulic conductivity–soil-water pressure relationships were elucidated by White and Sully (1987), and Warrick and Broadbridge (1992). For example, in Gardner's exponential relationship (Gardner 1958), in which the hydraulic conductivity $K(h)$ is given as follows:

$$K(h) = K_{sm} \exp(\alpha h) \quad (1)$$

where h (L) = soil-water pressure, the parameter α (L^{-1}) = inverse of the macroscopic length scale, i.e., $\alpha^{-1} = \lambda$.

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The development of different types of infiltrometer provided means to estimate the hydraulic properties of macropores and soil-matrix domains (Watson and Luxmoore 1986; Hussen and Warrick 1993). However, field measurements with infiltrometers can be difficult and time consuming. In particular, soil macroporosity, which reflects the developed soil pore structure at the measuring scale, is affected by soil disturbance resulting from the installation of the infiltrometer device. Watson and Luxmoore (1986) and Dunn and Phillips (1991) used double-ring infiltrometers to measure the ponded infiltration rate and to estimate the macroporosity. However, measurements with double-ring infiltrometers require driving a ring into the soil to establish one-dimensional flow. This procedure is time consuming and may generate cracks inside the soil or may result in the collapse of the naturally occurring large pores, thus affecting the estimate of macroporosity. On the other hand, the Guelph infiltrometer, adopted by Elrick et al. (1988), compares measurements of infiltration rates for different radii surface discs and does not require driving a ring into the soil. However, this method inherently assumes soil homogeneity at two measuring scales, an assumption that is unlikely to prevail in the presence of soil macropores.

In contrast to the field methods just discussed, this paper introduces a simple and efficient method to estimate soil macroporosity and to study the soil-hydraulic characteristics of macropores and soil-matrix domains without driving a ring into the soil or varying the surface area of the tested soil. This method was demonstrated by analyzing 32 unconfined tension infiltrometer tests. The unconfined infiltration rates were measured at ponded conditions and at 3, 6, and 15 cm of negative water-supply pressure. The technique suggested by Ankeny et al. (1991) to estimate hydraulic conductivity was modified to account for the possible presence of macropores in soils. The soil macroporosity and the hydraulic conductivity of macropores were determined.

The primary objective of this study is to develop a simple technique to determine soil macroporosity and hydraulic conductivity parameters for dual-porosity models based on measurements of unconfined infiltration rates. Measuring the unconfined infiltration rate is less disruptive to the soil structure than methods requiring the establishment of one-dimensional flow. Another motivation for this study is that we know little about the continuity and connectivity of macropores with depth and the relationship between the soil-water characteristics of micropore and macropore domains (Beven and Germann 1982). Hence, a second objective for this study is to compare the hydraulic properties of the two flow domains and to investigate the variability of macroporosity with depth.

MATERIALS AND METHODS

Instrument Installation and Infiltration Tests

As shown in Fig. 1, the major components of the tension infiltrometer are a bubbling tube that controls the water supply pressure at the soil surface, a water reservoir that empties as water flows into the soil, and a disc that contains a porous membrane to establish hydraulic continuity with the soil. The polycarbonate frame strengthens the infiltrometer and is used in carrying the device. After preparing the soil surface with a pointing trowel, three layers of cheesecloth were placed on the surface to avoid slaking of soil into the macropores. Slightly moistened silica sand was placed as contact material with the surface. The infiltrometer was then placed onto the center of the sand and pressed gently onto the sand. The device was anchored by securing four sharpened threaded rods at the corner of the base.

Negative water-supply pressure was controlled at the soil surface by the air-entry ports in the bubbling tube. Each of the air-entry tubes can be calibrated to set a different negative water-supply pressure. In this study, in addition to the ponded condition at zero pressure, supply pressures of -3, -6, and -15 cm for each plot were used. Measurements of infiltration rates proceeded from ponded condition to lower water-supply pressures to minimize the time to reach steady state. Tension in the air pocket at the top of the reservoir is proportional to the height of column in the reservoir. Thus infiltration rates were monitored by recording tension changes with a tensicorder. When steady state was reached, the unconfined volumetric infiltration rate, Q ($L^3\,T^{-1}$), was determined from the readings of the falling water level in the filling tube.

The unconfined infiltration tests described were performed at the four supply pressures on 32 plots. Fourteen plots were on the soil surface and the rest were distributed at the depths shown in Table 1. The measured Q will be used to determine the unsaturated hydraulic conductivity.

Determining the Unsaturated Hydraulic Conductivity

Depending on whether or not the soil contains macropores, two approaches are suggested to determine hydraulic conductivity from the measured unconfined infiltration rates.

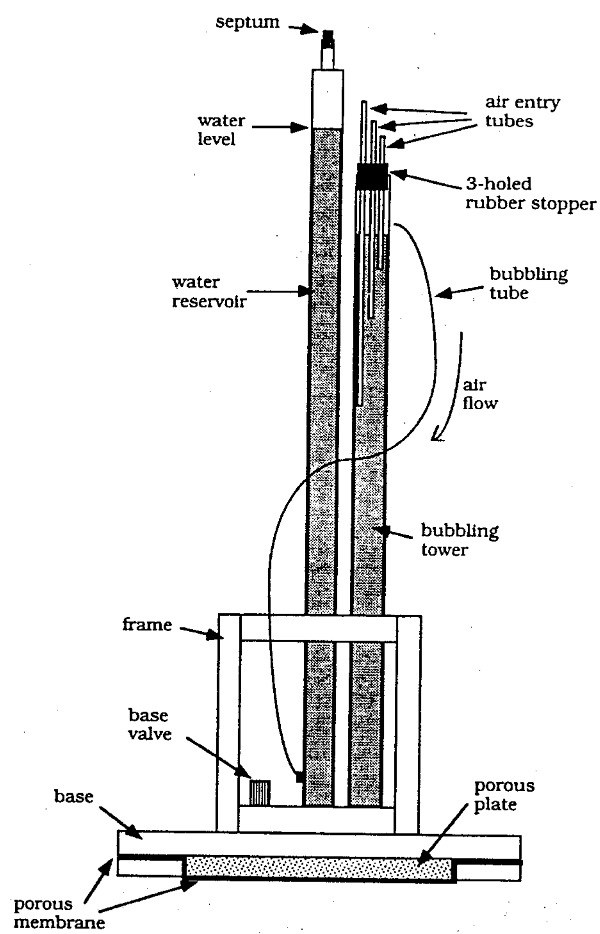

FIG. 1. Schematic Diagram of Tension Infiltrometer

One-Domain Interpretation (Micropores Only)

For steady infiltration from a circular disc, Wooding (1968) provided an algebraic solution based on an assumed exponential relationship between hydraulic conductivity and water-supply pressure, described by the following:

$$\frac{Q(h_0)}{\pi r^2} = K(h_0) + K(h_0)\frac{4}{\pi r \alpha} \tag{2}$$

where $Q(h_0)$ $(L^3 \cdot T^{-1})$ = measured volumetric infiltration rate at the applied water pressure h_0 (L); r (L) = radius of the infiltrometer; and α $[L^{-1}]$ = parameter of the Gardner's exponential model given in (1). Philip (1969) recognized α as a measure of the relative influence of capillarity and gravity in soils. Coarse soils, where gravity tends to dominate, have large values for α; fine-textured soils, where capillarity is important, have small values of α. The second term on the right-hand side of (2) reflects the influence of flow geometry and capillarity on the dynamics of flow. When the radius of the infiltrometer is large, or when the impact of capillarity is negligible, i.e., α is large, this term tends to zero and the flow becomes one-dimensional, driven by gravity alone.

If the exponential hydraulic conductivity–water-pressure relationship is assumed adequate (Philip 1969, 1985), then the substitution of (1) into (2), after applying the logarithmic transform, yields the following:

$$\ln(q) = \alpha h_0 + \ln\left[K_{sm}\left(1 + \frac{4}{\pi r \alpha}\right)\right] \tag{3}$$

where $q(h_0) = Q/(\pi r^2)$. The parameter α = slope of the linear regression of $\ln(q)$ on h_0 as illustrated in Fig. 2. The infiltration rate at zero pressure is included in the one-domain model and excluded in the two-domains model; the foregoing equation refers to the one-domain interpretation. The saturated conductivity K_{sm} is determined from the intercept, I, of this line, with the vertical axis using the following relationship:

TABLE 1. Summary of Data Analysis

Test number (1)	Depth from surface (cm) (2)	K_{sm}[a] ($cm \cdot hr^{-1}$) (3)	α[b] (cm^{-1}) (4)	K_{mac}[c] ($cm \cdot hr^{-1}$) (5)	$\theta_{mac} \times 10^2$[d] (cm^3/cm^3) (6)
1	0	5.48	0.27	21.44	0.175
2	0	4.79	0.17	11.09	0.090
3	0	0.85	0.13	20.63	0.168
4	0	8.16	0.32	16.08	0.131
5	0	2.93	0.15	8.83	0.072
6	0	3.92	0.17	24.64	0.201
7	0	6.03	0.26	22.80	0.186
8	0	8.08	0.26	20.17	0.164
9	0	4.81	0.18	20.12	0.164
10	0	3.97	0.17	12.67	0.103
11	0	7.21	0.20	11.96	0.098
12	0	9.49	0.21	15.20	0.124
13	0	—[e]	—[e]	5.61	0.046
14	0	6.84	0.22	—[e]	—[e]
15	10	3.09	0.16	15.40	0.126
16	15	3.30	0.22	19.04	0.155
17	15	4.54	0.22	16.72	0.136
18	15	9.16	0.25	21.16	0.173
19	15	3.10	0.20	22.52	0.184
20	15	3.64	0.20	17.56	0.143
21	20	0.81	0.13	7.22	0.059
22	25	1.53	0.22	15.40	0.126
23	30	1.46	0.32	15.25	0.124
24	30	5.20	0.31	16.67	0.136
25	30	4.41	0.38	12.86	0.105
26	30	0.25	0.14	17.49	0.143
27	35	1.27	0.12	1.59	0.013
28	35	1.35	0.35	18.28	0.149
29	37	6.84	0.29	8.22	0.067
30	39	2.65	0.25	3.22	0.026
31	45	2.70	0.31	4.93	0.040
32	46	0.53	0.11	13.28	0.108

[a]Determined from (4) using the intercept of the regression line in (3); infiltration rates at negative supply pressure of 3, 6, and 15 cm were used to fit the regression line.
[b]Slope of the linear regression in (3).
[c]Eq. (5) is source.
[d]Eq. (7) is source.
[e]Infiltration rates were not measured at four supply pressures.

$$K_{sm}\left(1 + \frac{4}{\pi r \alpha}\right) = \exp(I) \quad (4)$$

In the one-domain interpretation, the measured $q(h_0)$ at four supply pressures are used to fit the regression line in (3).

Two-Domains Interpretation (Macroporous Soils)

There is growing evidence that the fitting of Gardner's exponential model provides poor results, especially close to saturation. Ankeny et al. (1991), who suggested the use of a different α for different ranges of water-supply pressure, reported unusually large values of α in the range 0 to -3 cm of water pressure (Table 1). Increasing values of α in this range of water pressure are consistent with gravity-driven macropore flow. Theoretically, α is infinitely large for macropores. The Darcian principle, used in developing Wooding's (1968) solution, fails to describe the rapid infiltration through macropore channels (Beven and Germann 1982). This infiltration is assumed to occur at zero pressure, because the macropores will desaturate and may not conduct water at negative supply pressures (Watson and Luxmoore 1986; Dunn and Phillips 1991). Hence, to account for macropore flow, the measured unconfined infiltration rate at zero supply pressure, $q(h_0 = 0)$, is decomposed into a flow rate through macropores, q_{mac}, and a flow rate through the soil matrix, q_{sm}. Symbolically

$$q(h_0 = 0) = q_{mac} + q_{sm} \quad (5)$$

Beven and Germann (1982), who listed several pore classification schemes to distinguish between macropores and micropores, suggested that any cutoff size between the two domains is arbitrary. Measurements with infiltrometers have assumed that any pore size larger than 1 mm in diameter can be considered a macropore (Watson and Luxmoore 1986; Luxmoore 1981). Pores of this

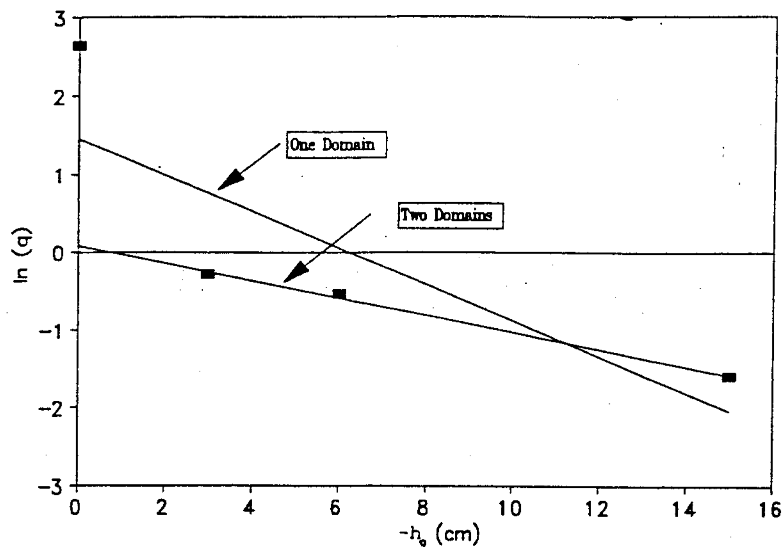

FIG. 2. Regression Lines Producing Saturated Matrix Conductivity and α

size or larger are excluded from the infiltration at supply pressure smaller than −3 cm. Hence the procedures to determine the parameters of the two domains are as follows. The hydraulic conductivity of the soil matrix, K_{sm}, and α are estimated first from the nonzero pressure infiltration rates using (3) and (4) (see Fig. 2). The macropore infiltration rate q_{mac} is the difference between $q(h_0)$ and q_{sm}, approximated by (4). When the macropores are flowing full under the sole influence of gravity, the hydraulic gradient at steady state is unity. Hence, the hydraulic conductivity, k_{mac}, of the macropores is the velocity of flow, q_{mac}. Also at steady state, Poiseuille law for laminar flow is adopted to determine the number of macropores per unit area, M (Watson and Luxmoore 1986). Mathematically

$$M = 8\mu \frac{K_{mac}}{\pi d g r_p^4} \tag{6}$$

where μ = viscosity of water ($M\ L^{-1}\ T^{-1}$); d = density of water ($M\ L^{-3}$); g = acceleration due to gravity ($L\ T^{-2}$); and r_p (L) = 0.5-mm minimum pore radius (Watson and Luxmoore 1986). Knowing the number of macropores per unit area, the macroporosity is determined using the following expression:

$$\theta_{mac} = M\pi r_p^2 \tag{7}$$

Statistical Distributions of Soil Parameters

To compare the hydraulic properties of the two domains, the parameters K_{mac}, θ_{mac}, K_{sm}, and α were first estimated using (3)–(5) and (7), and the procedure described previously on the 32 plots. It is essential to determine the statistical distributions of these parameters to decide whether arithmetic or geometric means should be used in the comparison. The mean, standard deviations, skewness, and kurtosis coefficients of the parameters K_{mac}, θ_{mac}, K_{sm}, and α, and of their logarithmic transforms were evaluated. The skewness and kurtosis were used to verify the statistical distribution of the data. The skewness coefficient is closer to zero for a normally distributed variable. The kurtosis coefficient, which is a dimensionless measure of flatness of a distribution, is 3 for a normally distributed random variable. Hence, the closer the calculated kurtosis is to 3, the greater is the tendency toward normality. The Kolmogorov-Smirnov test is another measure of the goodness of fit of a statistical distribution. The basic procedure for this statistical test is to compare the experimental cumulative frequency distribution and the assumed theoretical cumulative frequency distribution. If the discrepancy is large with respect to what is expected, the theoretical model of frequency distribution is rejected. Application of the Kolmogorov-Smirnov test to the analysis of soil properties is well documented by Rao et al. (1979) who recommended the use of the alternate hypothesis, i.e., to test the probability of wrongly claiming "the distribution of measurements does not have a specified functional form" as true, at 0.15 probability level. They concluded that when the coefficient of variation of a random variable is less than or equal to 0.4, the normal and lognormal distribution may be equally adequate. In such cases, the arithmetic and geometric means of the random variable are close, and accepting normal distribution as the appropriate distribution when the true distribution is lognormal will result in an error between 6 and 20%.

RESULTS AND COMMENTS

The installation of the infiltrometer used in this study was simple, and the infiltration tests at any particular location were completed in approximately two hours. Infiltration tests at supply pressure of -15 cm in the B horizon took longer time to reach steady state. This was due to the increase in clay content of the soil with depth. The downloading of the tensicorder data to a graphing software was helpful in determining visually the time to steady state to terminate the tests.

Table 1 shows K_{mac}, θ_{mac}, K_{sm}, and α, parameters of the two domains estimated from the 32 plots. As illustrated in Fig. 2, α, the slope of the regression line in absolute value, decreased in a two-domains interpretation. This decrease in α, which was observed at 27 plots, was anticipated because large α is associated with gravitational flow in large pores. Hence the exclusion of macropores should yield a smaller value for α. Regression analysis using (3) yielded r ranging from 0.87–1 with a mean of 0.96.

Comparing Hydraulic Properties of Macropores and Soil Matrix

The difference between K_{mac} and K_{sm} varied from 20.72 $cm \cdot hr^{-1}$ (at test 6 on the soil surface), where the macropores comprised 0.2% of the soil volume, to 0.32 $cm \cdot hr^{-1}$ (at test number 27), where the macropores were only 0.013%. Statistical moments of the conductivity of macropores K_{mac} and soil matrix, K_{sm}, and their logarithmic transforms, ln K_{mac} and ln K_{sm}, are shown in Table 2 for the data from the 32 plots. The skewness and kurtosis of K_{sm} and K_{mac} were closer to those of a normal distribution than the skewness and kurtosis of their logarithmic transforms. This result concurred with the findings of the Kolmogorov-Smirnov tests summarized in Table 3. Hence, it was concluded that K_{mac} and K_{sm} were more likely to be normally distributed than lognormally distributed for these data. The arithmetic mean of K_{mac} (14.8 $cm \cdot hr^{1}$) was 3.6 times the arithmetic mean of K_{sm} (4.1 $cm \cdot hr^{-1}$). This result was consistent with the findings of Watson and Luxmoore (1986) and Dunn and Phillips (1991) who suggested that between 73–83% of the flow is conducted through macropores under ponded conditions. This result substantiated the claimed difference between the hydraulic properties of the two flow domains and may justify the need for dual-porosity models.

To investigate the variation of soil macroporosity with depth, the values of macroporosity estimated in Table 1 were averaged in 15-cm intervals with depth. As shown in Table 4, the averaged macroporosity decreased by a factor of 2 in the depth range of 30–46 cm. This decrease reflected possible discontinuity of macropores at larger depths probably due to short grass roots at this site. Hence, dual-porosity models that assume a uniform soil macroporosity in a soil profile may overestimate the role of macropores in channeling water flow.

TABLE 2. Summary of Statistics of Conductivities of Soil Matrix and the Macropores

Variable (1)	Mean (2)	Standard deviation (3)	Skewness (4)	Kurtosis (5)
K_{mac} ($cm \cdot hr^{-1}$)	4.14	2.62	0.38	2.08
ln K_{mac}	2.56	0.62	−1.65	5.44
K_{sm} ($cm \cdot hr^{-1}$)	14.77	6.08	−0.49	2.25
ln K_{sm}	1.14	0.89	−0.99	3.35

TABLE 3. Summary Results of Kolmogorov-Smirnov Statistical Test

Variable (1)	Maximum difference between observed and theoretical distribution[a] (2)	Probability level for wrongly claiming alternate hypothesis[b] (3)
K_{sm}	0.55	0.2
K_{mac}	0.80	0.1
ln K_{sm}	0.94	0.025
ln K_{mac}	1.24	—

[a]Determined using (9) of Rao et al. (1979).
[b]The alternate hypothesis is "the distribution of measurements does not have the specified normal form" (Rao et al. 1979).

TABLE 4. Averaged Macroporosity in 15 cm Intervals

Depth interval (cm) (1)	Averaged macroporosity × 10^2 (cm^3/cm^3) (2)
0–15	0.137
16–30	0.115
31–46	0.067

Variability in macropores properties are attributed to heterogeneities in pores formed by plant roots, soil fauna, or cracks and fissures; however, Rawls and Brakensiek (1988) associated the variability in the saturated matrix conductivity with heterogeneities in soil texture and structure. The correlation coefficient, ρ, between K_{mac} on K_{sm} was evaluated to investigate the relationship between these two types of natural variabilities in soils. The correlation coefficient was 0.26. The null hypothesis, $\rho = 0$, was rejected at 0.2 probability level using the t-statistic test of Snedecor and Cochran (1980). The identified correlation coefficient is small and might be of little use in estimating parameters for dual-porosity models.

The parameter α of the exponential model given in (1) varied between 0.11 and 0.38 cm^{-1} with a mean of 0.22 cm^{-1} and a coefficient of variation of 0.32. This range of variation was consistent with the range defined by Elrick et al. (1989), who suggested that α lies between 0.12, for unstructured sandy loams, and 0.36, for coarse and gravelly sands. The parameter α had a skewness of 0.35 and a kurtosis of 2.20. It also passed the Kolmogorov-Smirnov normality test at 0.2 probability level. However, because of its small coefficient of variation, this test was inconclusive about the probability-density function of α (Rao et al. 1979). Because the parameter α in (1) measures the decrease of the hydraulic conductivity with negative water pressure, the range 0 to -15 cm water pressure, in which the infiltration tests were conducted, may govern water flow and solute transport. Indeed, if the mean of α is used in (1), the hydraulic conductivity of the soil matrix, and hence the potential for the transport of contaminant by advection, will decrease twenty-sevenfold for a change in water pressure between 0 and -15 cm.

Relevancy of K_{mac}, θ_{mac}, K_{sm}, and α to Existing Dual-Porosity Models

For soils containing macropores, adopting a two-domain approach provides the necessary information to model the influence of pore size on soil hydraulic properties near saturation. This information is required to account for macropore flow in existing dual-porosity models. For example, the hydraulic conductivity for soil macropores, K_{mac}, can serve as parameter input for the water flow and solute transport model developed by Jarvis (1993). Similarly, these data provided estimates of the number of macropores per unit area, which is an input parameter to the model suggested by Ahuja and Hebson (1992). Estimates of the hydraulic conductivity for the soil-matrix conductivity were also provided by the developed technique. The parameter α in (1) is more than a fitting parameter and can be used in dual-porosity models. For example, Beven and Clarke (1986) and Ahuja and Hebson (1992) adopted the Green and Ampt infiltration model to account for water flow through the soil matrix. In this model, estimates of saturated conductivity of the soil matrix and of the wetting front suction are required. The macroscopic length scale, which is the inverse of α in Table 1, is indeed equivalent to the wetting front suction.

Jarvis (1991) employed Richards' equation and adopted the Brooks and Corey $K(h)$ relationship to model infiltration through the soil matrix. The parameters of the Brooks and Corey $K(h)$ relationship can be determined from the macroscopic length scale using algebraic equations introduced by White and Sully (1987) and Morel-Seytoux et al. (1994).

SUMMARY

A simple field method was suggested to estimate the parameters of dual-porosity models. This method relied on measurements of unconfined infiltration rates with a tension infiltrometer. The field method described here provided an advantage over double-ring infiltrometers, which require driving a ring into the soil to establish one-dimensional flow. Forcing one-dimensional flow is not needed to estimate soil macroporosity, because macropore flow is driven by gravity. Also, driving a ring into the soil is time consuming and may change the soil structure.

The conductivity of the macropores as well as the parameters of the soil-matrix conductivity were estimated by the developed technique. The relevancy of these parameters to existing dual-porosity models was illustrated. The hydraulic conductivity of the macropores was 3.6 times the conductivity of the soil matrix. This result can serve as initial guide to estimate hydraulic conductivities of dual-porosity models. Application of dual-porosity models should consider the variability of macroporosity with depth, which can be controlled by pedogenic factors. For field soils, hydraulic conductivity may decrease rapidly with soil-water pressure, and the narrow range of 0 to -15 cm of water pressure may govern the rate of transport of contaminants in the vadose zone. A positive but small correlation coefficient existed between the conductivities of the soil matrix and the macropores.

ACKNOWLEDGMENTS

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APPENDIX I. REFERENCES

Ahuja, L. R., and Hebson, C. (1992). "Root zone water quality model." *GPSR Tech. Rep. No. 2*, United States Dept. of Agric., Agric. Res. Services, Fort Collins, Colo.

Ankeny, M. A., Ahmed, M., Kaspar, T. C., and Horton, R. (1991). "Simple field methods for determining unsaturated hydraulic conductivity." *Soil Sci. Soc. Am. J.*, 55(2), 467–470.

Beven, K., and Clarke, R. T. (1986). "On the variation of infiltration into a homogeneous soil matrix containing a population of macropores." *Water Resour. Res.*, 22(3), 383–389.

Beven, K., and Germann, P. (1982). "Macropores and water flow in soils." *Water Resour. Res.*, 18(5), 1311–1325.

Dunn, G. H., and Phillips, R. E. (1991). "Macroporosity of a well-drained soil under no-till and conventional tillage." *Soil Sci. Soc. Am. J.*, 55(3), 817–823.

Elrick, D. E., Reynolds, W. D., Baumgartner, N., Tan, K. A., and Bradshaw, K. L. (1988). "In-situ measurements of hydraulic properties of soils using Guelph permeameter and the Guelph infiltrometer." *Proc., Int. Work Land Drain*, 3rd Ed., Ohio State Univ., Columbus, Ohio, 13–23.

Elrick, D. E., Reynolds, W. D., and Tan, K. A. (1989). "Hydraulic conductivity measurements in the unsaturated zone using improved well analysis." *Ground Water Monitoring Rev.*, 9(3), 184–193.

Gardner, W. R. (1958). "Some steady state solutions of unsaturated moisture flow equation with application to evaporation from a water table." *Soil Sci.*, 85(3), 228–232.

Hussen, A. A., and Warrick, A. W. (1993). "Alternative analysis of hydraulic data from disc tension infiltrometers." *Water Resour. Res.*, 29(12), 4103–4108.

Jarvis, N. (1991). *MACRO—A model of water movement and solute transport in macroporous soils*. Swedish Univ. of Agric. Sci., Dept. of Soil Sciences, Uppsala, Sweden.

Luxmoore, R. J. (1981). "*Micro-, meso-, and macroporosity of soil*." *Soil Sci. Soc. Am. J.*, 45(3), 671.

Mosley, M. P. (1982). "Subsurface flow velocities through selected forest soils, South Island, New Zealand." *J. Hydrol.*, 55(1-4), 65–92.

Peterson, A. E., and Dixon, R. M. (1971). "Water movement in large soil pores." *Res. Rep.*, Res. Div., Coll. of Agr. and Life Sci., Univ. of Wisconsin, Madison, Wisc., 1–8.

Philip, J. R. (1969). "Theory of infiltration." *Adv. in Hydrosci.*, Vol. 5, 215–296.

Philip, J. R. (1985). "Reply to comments on 'Steady infiltration from spherical cavities.'" *Soil Sci. Soc. Am. J.*, 49(3), 788–789.

Pilgrim, D. H., Huff, D. D., and Steel, T. D. (1978). "A field evaluation of subsurface and surface runoff, 2: Runoff processes." *J. Hydrol.*, 38(3-4), 319–341.

Rao, P. V., Rao, P. S. C., Davidson, J. M., and Hammond, L. C. (1979). "Use of goodness-of-fit tests for characterizing the spatial variability of soil properties." *Soil Sci. Soc. Am. J.*, 43(2), 274–278.

Rawls, W. J., and Brakensiek, D. L. (1988). "Estimation of soil water retention and hydraulic conductivity." *Unsaturated flow in hydrologic modeling*. H. J. Morel-Seytoux, ed., Kluwer Academic Publishers, Boston, Mass.

Snedecor, G. W., and Cochran, W. G. (1980). *Statistical Methods*. Iowa State University Press, Ames, Iowa.

Thomas, G. W., and Phillips, R. E. (1979). "Consequences of water movement in macropores." *J. Environ. Quality*, 8(2), 149–152.

Warrick, A. W., and Broadbridge, P. (1992). "Sorptivity and macroscopic capillary length relationships." *Water Resour. Res.*, 28(2), 427–431.

Watson, K. W., and Luxmoore, R. J. (1986). "Estimating macroporosity in a forest watershed by use of a tension infiltrometer." *Soil Sci. Soc. Am. J.*, 50(3), 578–582.

White, I., and Sully, M. J. (1987). "Macroscopic and microscopic length scales from field infiltration." *Water Resour. Res.*, 23(8), 1514–1522.

Wooding, R. A. (1968). "Steady infiltration from a shallow circular pond." *Water Resour. Res.*, 4(6), 1259–1273.

APPENDIX II. NOTATION

The following symbols are used in this paper:

d = density of water;
g = acceleration of gravity;
h = soil-water pressure;
K_{mac} = hydraulic conductivity of macropores;
K_{sm} = hydraulic conductivity of soil matrix;
M = number of macropores per unit area;
N = sample size;
Q = volumetric infiltration rate;
q_{mac} = infiltration rate through macropores;
q_{sm} = infiltration rate through soil matrix;
r = infiltrometer radius;
r_p = pore radius;
T_c = t-statistic;
α = parameter in Gardner's model;
θ_{mac} = macroporosity;
μ = viscosity of water; and
ρ = correlation coefficient.