Reprinted from the *Soil Science Society of America Journal*
Volume 58, no. 2, March-April 1994
677 South Segoe Rd., Madison, WI 53711 USA

Comparison of Three Field Methods to Characterize Apparent Macropore Conductivity

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ABSTRACT

The objective of this study was to investigate some field-oriented methods designed to characterize the flow rates of both the soil matrix and macropores near saturation. An apparent macropore conductivity was obtained by the difference between saturated and unsaturated hydraulic conductivities at − 0.6 kPa of soil water pressure. Saturated conductivities were measured in double-ring infiltrometers with tensiometers. Soil matrix conductivities were calculated from measurements of soil water pressures and drainage rate during redistribution and were also measured with a thin sand-cement crust. Another set of data for saturated and unsaturated conductivities was obtained from unconfined ponded measurements in 76.2-mm-diam. rings and a tension infiltrometer. Unsaturated conductivities and calculated apparent macropore conductivities from the crust and redistribution methods were comparable. Unsaturated conductivities and apparent macropore conductivities from the tension infiltrometer and unconfined ponded measurements were greater than those from the infiltration-redistribution method. We concluded that apparent macropore conductivity for the profile can be characterized by using ponded infiltration rates and matrix conductivities near saturation determined from redistribution data. This method accounts for the continuity of the macropore flow system in subsurface layers and involves a larger area for macropore infiltration measurement.

There is growing interest in the preferential movement of surface-applied chemicals through the soil to groundwater. One mechanism for this phenomenon is via continuous wormholes, root channels, or interaggregate pores, known as macropores. Watson and Luxmoore (1986) defined macropores as pores having an effective diameter of ≥ 1.0 mm. However, the choice of a lower size limit for macropores is somewhat arbitrary; values in the range of 0.03 and 3.0 mm have been used by other investigators (Beven and Germann, 1982). Flux of water in such macropores has been reported to be one to four orders of magnitude greater than flow in the soil matrix (Beven and Germann, 1981). If these high flux rates are to be realized beyond the Ap horizon, and significant "short circuiting" of surface-applied chemicals is to occur in the field, the macropores must also be continuous (Smettem, 1987).

Characterization of macropore hydraulic conductivity is necessary to determine the potential for movement of surface-applied chemicals through these channels. The maximum macropore flow rate is an important input parameter for some solute and water transport models that simulate preferential flow in macropores, such as the Root Zone Water Quality Model (Ahuja et al., 1991).

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Published in Soil Sci. Soc. Am. J. 58:278-284 (1994).

One means of characterizing hydraulic conductivity of macropores is to measure saturated hydraulic conductivity under ponded conditions and unsaturated conductivity at a specified small negative pore water pressure (Dixon, 1975). The negative pressure corresponds to the lower limit of the effective diameter defined for macropores. For a lower limit of 1.0 mm (Watson and Luxmoore, 1986), this pressure is −0.3 kPa. For a lower limit of 0.5 mm, the pressure is −0.6 kPa. The difference between the two flow rates is caused by the hydraulic conductivity of continuous pores with effective diameters greater than the selected lower limit. Theoretically, such pores would not contribute to water flow at pore water pressures less than the pressure value corresponding to this smaller diameter.

Flow rates at a small negative pore water pressure can be measured in situ using a tension infiltrometer (Watson and Luxmoore, 1986), a disk permeameter (Smettem, 1987; Clothier and Smettem, 1990), or a surface crust to restrict flow rates into the soil (Bouma et al., 1983; Booltink et al., 1991). These methods, however, have limitations. Only surface horizons or the top of an excavated horizon of a layered soil can be characterized using the permeameter or tension infiltrometer. The surface crust method, as described by Booltink et al. (1991), requires a pedestal of soil to be dug out and also generally provides unsaturated conductivity of only the top soil. Further field experimentation after conductivity measurements using these devices is impossible where the soil has been disturbed to obtain soil columns or subsurface measurements. Another important characteristic of macropores—their continuity—cannot be determined when measurements are made on individual horizons. Furthermore, the permeameters and tension infiltrometers generally sample only a small area (7.5-20-cm diam.), which may not meet the criterion of a minimum representative area for a macroporous soil (Bouma, 1983).

These limitations can be overcome by the use of an infiltration-redistribution method. The combined saturated hydraulic conductivity of the continuous macropores and the soil matrix is obtained from measurements during ponded infiltration in a large ring or plot. The hydraulic conductivity of the soil matrix near saturation (at a small known negative pressure) can be obtained from redistribution data using the unsteady drainage-flux (instantaneous profile) method (Green et al., 1986). The difference between the two values, the apparent macropore conductivity, reflects the contribution of continuous macropores. This method, of course, provides characterization of the unsaturated hydraulic properties of a layered soil matrix as well.

The objective of this study was to compare this method with two other suggested methods of measuring macropore conductivity: (i) ponded infiltration and infiltration through a thin sand-cement crust and (ii) unconfined ponded infiltration and unsaturated infiltration using a tension infiltrometer. For the infiltration–redistribution method, we used the double-ring infiltrometer–multiple-depth tensiometer setup (Ahuja et al., 1976; Green et al., 1986).

MATERIALS AND METHODS

The study site is located at the USDA-ARS National Agricultural Water Quality Management Laboratory (Durant, OK). The soil at this site, Bosville fine sandy loam (fine, mixed, thermic Albaquic Paleudalf), has been in grass cover for >15 yr. Textural layering and bulk density with depth for the eight sites are shown in Fig. 1.

The sequence of the experimental work was to characterize the soil for saturated hydraulic conductivity by ponding water and then allowing the soil to drain to obtain unsaturated conductivities. These drainage measurements were followed by infiltration through a crust and, finally, measurements were collected using unconfined ponded and tension infiltrometers in the same rings. A dye was then applied in a pulse of water, and the soil was sampled to obtain cores for the moisture characteristic. These experiments are described below separately.

Double-Ring Infiltrometer Measurements

During the summer of 1989, eight double-ring infiltrometers, separated by a distance that varied from 3 to 5 m, were randomly located in the study site. The inner ring of the infiltrometer was 0.5 m in diameter and the outer ring 0.9 m in diameter. Each ring was uniformly driven into the soil to a depth of at least 0.10 m. One multiple-depth tensiometer was installed in the center of each ring by boring a hole with a soil auger slightly larger in diameter than the tensiometer. Prior to installation of the tensiometer, the hole was filled with a clay–silt slurry. Each tensiometer contained five ceramic cups located at 0.1-m intervals to a depth of 0.5 m. As part of a related experiment, the herbicide glyphosate [isopropylamine salt of *N*-(phosphono-methyl) glycine] was applied within six of the rings to kill the grass cover. Herbicide was not applied to the soil around and within Rings 1 and 5.

Ponded infiltration was begun by adding water to a known ponding depth to both rings, typically 40 to 50 mm. The soil surface inside the rings was protected with blocks of wood as the water was poured into the rings to minimize soil disturbance. Water was maintained at a constant level in each ring with a float and valve connected to a water reservoir. The water height inside the reservoir was measured using a sight glass tube. When the tensiometers no longer registered measurable changes for at least 1 h, steady state was assumed to have been reached. The time to reach this stage varied from 1 to 5 h depending on initial water content. Saturated hydraulic conductivities in different layers were calculated from the measured infiltration rates and the hydraulic gradients. For the 0- to 0.2-m layer, the infiltration rate used in the calculation was the one measured at the time when this layer first became saturated (as indicated by tensiometers) and the layer below was beginning to wet. For the second layer (0.2–0.4 m), the infiltration rate measured at the time when this layer was completely saturated was assumed to be the harmonic mean of the saturated conductivities of the first and second layers. Thus, the ponded K_s of the second layer could be estimated from the above measurements as

$$K_S^{(2)} \approx \frac{L^{(2)}}{(L^{(1)} + L^{(2)})/i - (L^{(1)}/K_S^{(1)})} \qquad [1]$$

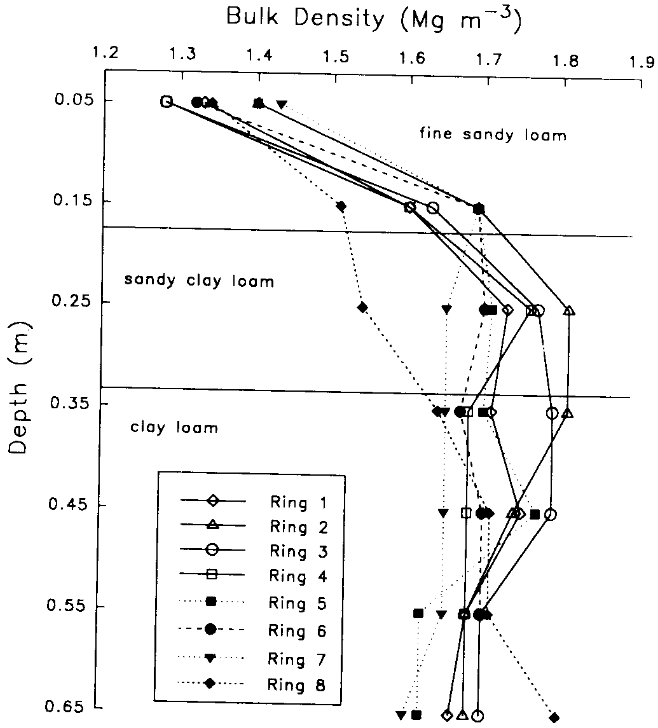

Fig. 1. Soil bulk density and soil texture with depth at eight locations where measurements were taken.

where $K_S^{(1)}$ and $K_S^{(2)}$ are the saturated conductivities of the first and second layers, $L^{(1)}$ and $L^{(2)}$ are the thicknesses of the first and second layers, and i is the infiltration rate. This approximation assumes that the gradient in the wetted profile is unity, which it approximately was at that time.

Redistribution Measurements

After the ponded infiltration experiments were completed, the soil surface within both the inner and outer rings was covered with plastic sheeting and hay to prevent evaporation and reduce temperature fluctuations. The rings were then covered with slotted plywood sheets to allow installation around the tensiometers. Tensiometer readings were recorded at 1- to 2-min intervals initially. The measurement frequency decreased to twice daily after the sixth day as the drainage rate decreased. Readings continued until the drainage rate slowed to near zero, typically after 20 to 30 d.

Measurements with Crust Method

The infiltration measurements with the crust were carried out in the same rings after the completion of the infiltration-redistribution measurements. A crust for the unsaturated conductivity measurements was made from a mixture of quick-setting cement and fine sand mixtures (Bouma et al., 1983; Bouma and Denning, 1971). The ratio of sand to cement (v/v) varied from 30:70 to 35:65 depending on the flux rate required to maintain unsaturated conditions. The soil surface was protected by placing two layers of cheesecloth on the soil surface and adding a 20- to 40-mm layer of fine sand. The cheesecloth facilitated removal of the sand (by vacuuming) after the measurements were completed. The fine sand provided a level surface for the crust, because the soil surface was uneven, and allowed placement of a uniform crust thickness. The crust was applied only in the inner ring because prior tests had

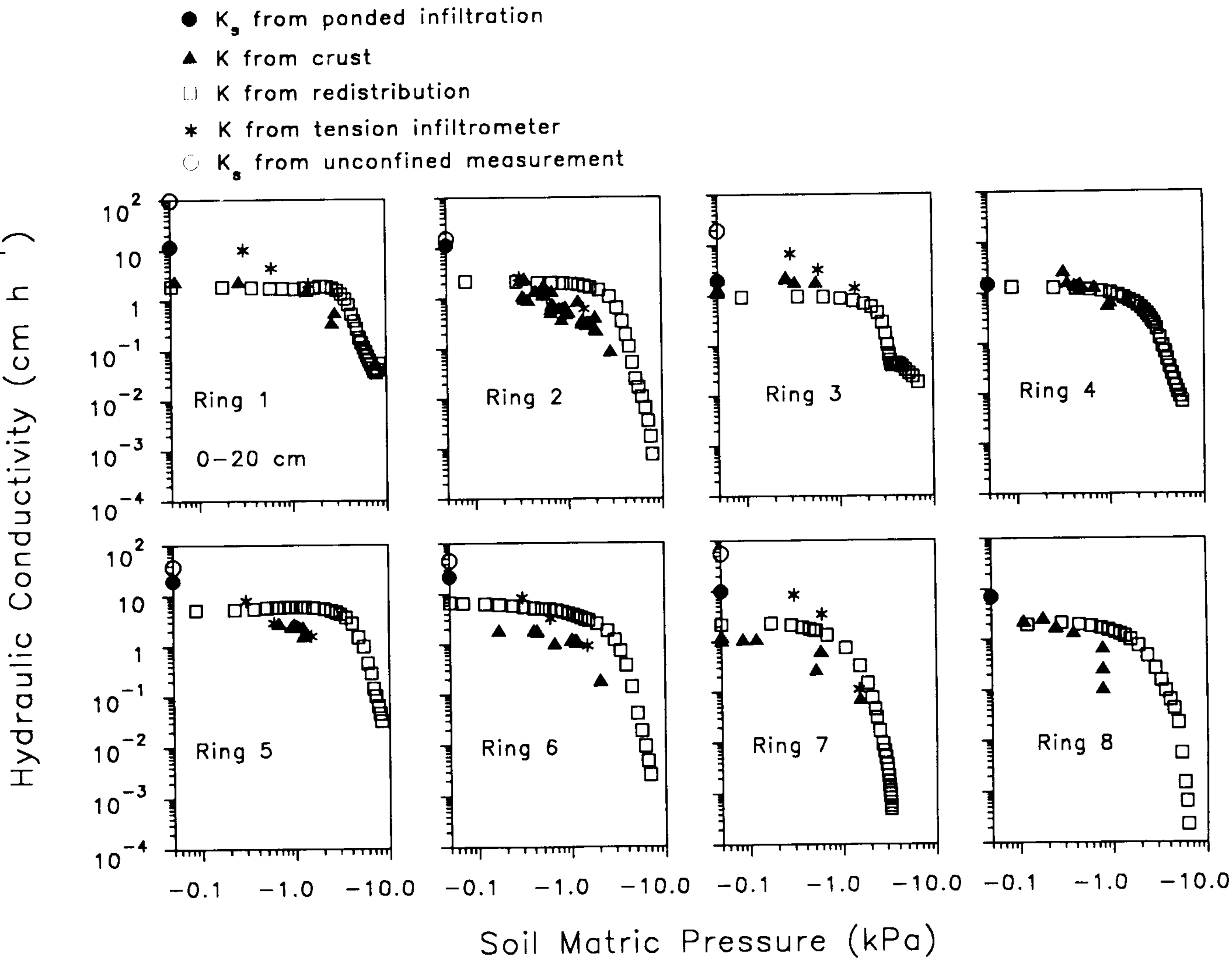

Fig. 2. Hydraulic conductivity vs. matric pressure determined from crust data, redistribution data, and the tension infiltrometer for soil within eight rings at 0- to 0.2-m depth.

indicated that lateral flow was minimal under unsaturated conditions when the outer ring was kept moist by a light spraying of water. A water-filled tensiometer was also placed near the multiple-depth tensiometer at a depth of 50 mm. Flow rates through the crust were controlled by varying ponding depth over the crust, typically from 5 to 30 mm. These different flow rates changed the soil water suction in the profile below the crust and, thus, allowed measurement of hydraulic conductivity at several different suctions close to saturation. Flux rates through the soil profile to a 0.4-m depth were assumed to be steady when the tensiometers to that depth no longer registered any changes for a period of 30 min. The time required to reach this stage varied from 1 to 2 h. The gradients for the 0- to 0.2-m depth were calculated from the slope of the hydraulic head vs. depth data calculated from the tensiometer measurements at the 0.05-, 0.1-, and 0.2-m depths. Hydraulic heads from the 0.20-, 0.3-, and 0.4-m tensiometer measurements were used for the 0.2- to 0.4-m depth gradients. Unsaturated hydraulic conductivities were calculated from the ratios of the steady flux rates through the crust and the gradients.

Unconfined Infiltrometer Measurements

After the crust infiltration experiments, unconfined ponded infiltration and tension infiltrometer experiments were conducted on the same locations using smaller (76.2-mm-diam.) rings. Four replicate measurements were taken at the soil surface in both the inner and outer ring areas at six of the eight locations. Steady ponded infiltration rates were first measured within 76-mm-diam. rings that were inserted ≈10 mm into the soil. After ponding measurements, tension infiltrometers (Ankeny et al., 1991) were used to obtain steady flow rates, in sequence, at −0.30, −0.60, and −0.15 kPa of suction (30, 60, and 150 mm negative pressure head). It required about 2 h to complete all measurements at each location (ring). At Rings 4 and 8, only measurements in the outer ring were taken. Hydraulic conductivities were calculated by the method of Ankeny et al. (1991).

Dye

The soil was dyed to reveal preferential flow paths and their continuity. Methylene blue dye ($C_{16}H_{18}ClN_3S \cdot H_2O$) was applied in a pulse of water at the rate of 0.2 g dye L^{-1} solution after all measurements were complete.

Core Samples for Moisture Release Data

At the end of all experiments, four replicate soil cores, 54 mm in diameter and 60 mm long, were taken in each ring

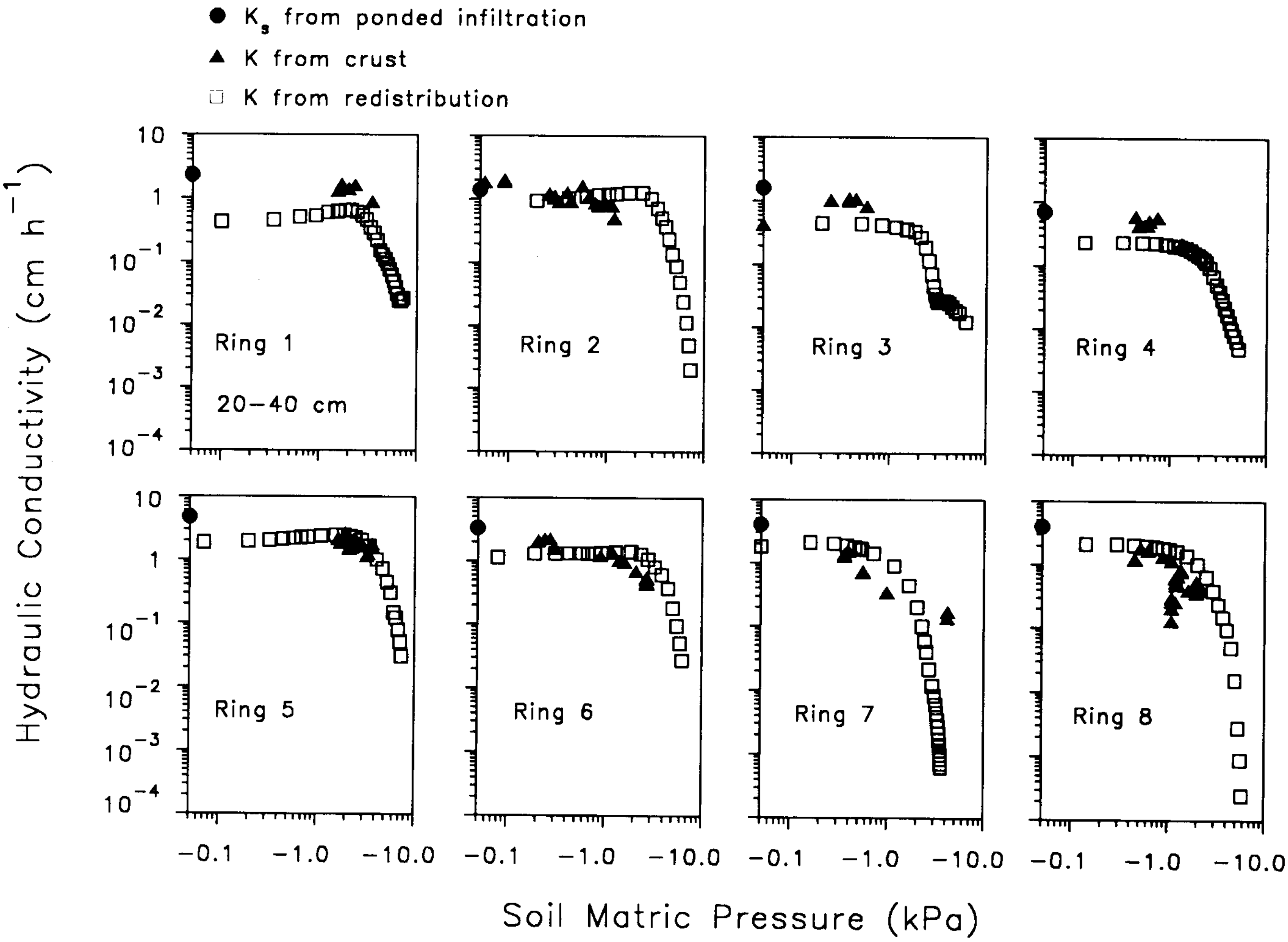

Fig. 3. Hydraulic conductivity vs. matric pressure determined from crust data and redistribution data for soil within eight rings at 0.2- to 0.4-m depth.

from four depths corresponding to observed soil horizonation. The sampling depths varied from ring to ring but were generally from 0 to 0.06, 0.1 to 0.16, 0.24 to 0.30, and 0.44 to 0.5 m. At this time, the soil was removed in 20-mm increments to 0.25-m depth to reveal any macropores stained by the dye.

The soil water characteristics were measured on these cores. A sand table was used to determine volumetric water content at 0.3, 1.0, and 2.5 kPa soil water suctions relative to the midpoint of the core. A pressure plate apparatus was used for 5.0, 10.0, 20.0, and 50.0 kPa suctions. At the lowest suction (0.3 kPa), the suction in the 6-cm-high core varied from 0.0 to −0.6 kPa. Because the relationship between water content and suction is generally linear in this range, the water content measured at the average value of suction will not be very different from the actual water content. Even though macropores may empty, it will not affect the water content of the matrix. Macropore conductivity properties are obtained from ponded infiltration data and not from the redistribution and core data. Therefore, any underestimation of water content will not affect the estimates of overall macropore conductivity.

Continuous water retention curves were obtained by fitting a two-piece function to the water content vs. suction data averaged from the four replicate cores. The first section, beginning at saturation, was a second-order polynomial. The second section, beginning at an intersection point, was a log–log function. These water retention curves were used in conjunction with the tensiometric data obtained during redistribution to determine the unsaturated hydraulic conductivities, using the procedures described in Green et al. (1986).

Although some soil disturbance is required to obtain cores from the experimental area, the water characteristic can also be obtained using nondestructive methods, such as time domain reflectometry or a neutron probe. Cores can also be obtained from nearby areas or after the experiment is complete.

Calculating Apparent Macropore Conductivities

Unsaturated conductivity at −0.6 kPa of suction was chosen as the representation of matrix saturated conductivity. This value was chosen based on the range of data available from the field experiments. The apparent macropore conductivity was obtained by subtracting this value from the ponded-water hydraulic conductivity.

RESULTS AND DISCUSSION

Saturated and unsaturated conductivities as a function of soil suction for the three methods and two depths are shown in Fig. 2 and 3.

Dye Observations

Staining was mostly limited to the soil surface (0.1-m depth), although there were occasional stained pores from 0.15 to 0.2 m. At the surface and to about 50 to 60 mm, large areas were stained rather than individual pores. There was little staining of root channels below the 50- to 60-mm depth. This indicated that most of the dye moved through small pores with relatively high surface area and was adsorbed by the soil.

Unsaturated Conductivities from Crust and Redistribution Data

The unsaturated hydraulic conductivities near saturation from the crust method were in close agreement with those calculated from redistribution data (Fig. 2 and 3, Table 1). At more negative suction (−1 to −3 kPa), conductivities from the crust method were less than conductivities from the redistribution data. This is most likely caused by hysteresis of hydraulic properties, because redistribution is a draining process whereas infiltration through the crust is primarily a wetting process.

The slopes in some of the conductivity–pressure relationships calculated from redistribution data (Fig. 2 and 3) were negative in the wet range. This may be due to the difficulty in obtaining accurate values of small gradients from Hg tensiometer data during early stages of redistribution. Such errors, or fluctuations, in tensiometric measurements may be minimized by using a more precise tensiometric measuring system such as water-filled tensiometers with a pressure transducer. These fluctuations were, however, not serious given the nature of such relationships.

Overall, the agreement is good between the unsaturated conductivities near saturation determined from the crust and redistribution methods (Fig. 2 and 3, Table 1). Therefore, unsaturated conductivities from the redistribution method can be used in place of conductivities near saturation from the more laborious crust method. A measure of apparent macropore conductivity can, in turn, be easily obtained from the difference between saturated conductivity and conductivity at −0.6 kPa from these redistribution measurements.

Unconfined and Confined Ponded Measurements

The unconfined measurements of saturated conductivities taken in the 76.2-mm rings were, in some cases, as much as an order of magnitude greater than those from confined infiltration in the larger double-ring infiltrometers (Fig. 2). An explanation for a portion of the difference may be the effect of air entrapment, which acts to reduce initial infiltration rate when water enters a soil containing pockets of air. This is less of a problem with the unconfined measurements because of the radial shape and unconfined nature of the wetting front. Another reason for the differences in our case relates to the strong increase in soil bulk density with depth from 0 to 0.2 m and the corresponding decrease in porosity (Fig. 1). The unconfined ponded measurements were influenced only by soil properties close to the surface (≈0.10–0.12 m), and not by the deeper layers, because of the small diameters of the rings. The confined measurements in the inner ring of the double-ring infiltrometers were affected more by the increase in soil bulk density with depth because

Table 1. Unsaturated conductivities of macropores from redistribution and crust data calculated for a matric pressure of −0.6 kPa.

Ring	Depth interval	Redistribution method	Crust method
	cm	cm h^{-1}	
1	0–20	1.8	2.3
	20–40	0.5	−†
2	0–20	2.1	1.5
	20–40	1.0	1.1
3	0–20	1.0	2.0
	20–40	0.4	1.4
4	0–20	1.2	1.7
	20–40	0.2	1.2
5	0–20	5.7	2.7
	20–40	2.2	−†
6	0–20	5.3	1.7
	20–40	1.3	1.9
7	0–20	1.8	0.5
	20–40	1.8	1.1
8	0–20	2.0	1.6
	20–40	2.0	1.3

† No unsaturated conductivities for pore water pressure greater than −0.8 kPa were available.

of the one-dimensional nature of flow and the larger area of ponding and because measurements were recorded when the front reached the 0.2-m depth. Furthermore, dye stains indicated that most macropores were near the surface, which appears to be reflected in the greater conductivities from the unconfined surface measurements.

Magnitudes of the mean saturated and unsaturated conductivities and within-ring Coefficients of Variation for the unconfined and tension infiltrometer measurements are listed in Table 2. In five of the six measurements taken on the soil in the inner rings, the saturated conductivities show more variability than do unsaturated measurements.

Unsaturated Conductivities from Redistribution Measurements and from Tension Infiltrometer

Unsaturated conductivities calculated from tension infiltrometer data were, generally, greater than those from the redistribution measurements, especially near saturation (Fig. 2). As in the confined and unconfined ponded measurements, the differences are probably related to the strong increase in soil bulk density with depth. The redistribution measurements taken in the larger, double-ring infiltrometers more closely reflect the presence of this layer. The tension infiltrometer data more closely reflect the surface conditions, where there was lower bulk density and more macroporosity. The differences in unsaturated conductivities are less in the drier range (lower suction) presumably because the effects of bulk density and large pores on flow processes decrease at lower potentials.

Apparent Macropore Conductivity

A majority of the rings exhibited evidence of nonzero apparent macropore conductivities using infiltration–redistribution and ponded plus crust measurements (Table 3). The differences between the two methods were small for both depths although apparent macropore con-

Table 2. Mean hydraulic conductivities and CVs of macropores from the unconfined measurements as a function of matric pressure for six of the eight inner rings.

Ring	0 kPa		−0.3 kPa		−0.6 kPa		−1.5 kPa	
	Mean	CV	Mean	CV	Mean	CV	Mean	CV
	cm h^{-1}	%	cm h^{-1}	%	cm h^{-1}	%	cm h^{-1}	%
1	96.7	84.7	10.2	24.9	4.5	21.5	2.1	32.4
2	15.4	74.1	2.7	35.2	1.0	30.6	0.6	36.9
3	19.5	28.4	6.	11.5	3.3	14.2	1.4	12.7
5	37.2	71.2	8.1	46.1	2.9	55.3	1.6	50.3
6	45.4	107.0	8.3	65.9	3.2	62.1	0.9	57.5
7	21.9	15.4	8.5	30.0	3.5	15.9	0.1	82.6

ductivities from the ponded plus crust method tended to be higher for the larger values of apparent macropore conductivity.

The differences in apparent macropore conductivities calculated from unconfined ponded plus tension infiltrometer measurements and calculated from the infiltration–redistribution measurements were especially large in two cases (Rings 1 and 4). These large differences were mainly caused by the very high saturated conductivities measured under ponded, unconfined conditions, which reflected the soil properties of a shallow surface soil layer. This is compared with average soil properties of the 0- to 0.2-m layer that were reflected in the infiltration–redistribution measurements.

Herbicide application to the soil in all rings, except 1 and 5, did not noticeably affect the apparent macropore conductivities by any method. The ranges of values in Rings 1 and 5 are similar to those in the other rings.

Although the infiltration–redistribution method is not designed to detect differences in apparent macropore conductivity as a function of suction, the apparent macropore conductivity at a suction very close to saturation is easily obtained with adequate accuracy. An additional advantage of using infiltration–redistribution data is that apparent macropore conductivity for the deeper layers can also be calculated (Table 3). This could provide a measure of the continuity of macropore flow. In this case, the sizable differences between the two layers reflect the effect of the flow-restricting layer at 0.2 m (Fig. 1) on macropore continuity.

The infiltration–redistribution method does not require restrictive assumptions. One assumption is that averages of gradients and soil moisture spanning a depth interval adequately represent the soil hydraulic state. Some curve fitting and interpolation is required to obtain changes in water content with time from suction data. However, there are no assumptions regarding the shape of the relationships. The time interval chosen for calculations of flux must be small enough that the gradient does not vary too greatly with the time interval. The time interval must also be large enough that errors in calculating differences in water contents are relatively small. The depth intervals must be small enough that soil properties are not averaged from relatively large vertical distances. The infiltration method does assume unit gradient to calculate hydraulic conductivity. With instrumentation, this assumption can be checked.

Table 3. Apparent macropore conductivities for the eight rings and two depths calculated using ponded plus crust, infiltration–redistribution, and unconfined ponded plus tension infiltrometer measurements.

Ring	Depth	Apparent macropore conductivity		
		Redistribution†	Crust†	Tension‡ infiltrometer
	cm	cm hr^{-1}		
1	0–20	9.9	9.4	86.5
	20–40	1.8	–§	
2	0–20	9.3	9.9	12.7
	20–40	0.4	0.3	
3	0–20	1.0	0.0	12.7
	20–40	1.2	0.2	
4	0–20	0.3	0.0	71.1
	20–40	0.5	0.0	
5	0–20	14.2	17.4	29.1
	20–40	2.6	–§	
6	0–20	16.2	19.8	NA¶
	20–40	2.0	1.4	
7	0–20	8.2	9.5	13.4
	20–40	2.2	2.9	
8	0–20	5.0	5.4	NA¶
	20–40	1.9	2.6	

† Apparent macropore conductivity is calculated as $K_{sat} - K_{(\psi > -0.6\ kPa)}$ where K_{sat} is saturated hydraulic conductivity from ponded measurements in the double ring.

‡ Calculated as $K_{sat} - K_{(\psi = -0.6\ kPa)}$ where K_{sat} is from ponded, unconfined infiltration measurement and $K_{(\psi = -0.6\ kPa)}$ is from tension infiltrometer measurements.

§ No unsaturated conductivities for pore water pressure greater than −0.8 kPa were available.

¶ No measurements taken.

In strongly layered soils, as in this experiment, in the initial stage of redistribution, gradients in the surface layers may be quite small and difficult to measure. Although this may result in some uncertainty in calculated unsaturated hydraulic conductivities near saturation, data from the cores are used only to determine properties of the soil matrix. Excluded from the calculations are the unsaturated conductivities of the soil determined from core and redistribution measurements that are in the range of pore sizes where macropores are important. Infiltration measurements under ponded conditions are used to infer macropore flow properties.

CONCLUSIONS

Based on these results, redistribution data along with ponded infiltration rates can be used to determine apparent macropore conductivities to a suction near saturation. When knowledge of conductivities at the soil surface is desired, a tension infiltrometer may be more appropriate. The infiltration–redistribution method, however, has several advantages. The method does not require the soil to be homogeneous with depth. The apparent macropore conductivity can also be determined for subsurface layers without soil disturbance. This gives a measure of macropore continuity. The sampling area can be made as large as required to enclose a representative area of soil containing macropores. Redistribution measurements allow determination of unsaturated conductivities at lower suctions as well. In fact, redistribution is a commonly used and recommended field method for determining the hydraulic conductivity–potential relationship of soils (Green et al., 1986).

ACKNOWLEDGMENTS

The authors thank Mark Prieksat, who helped take the tension infiltrometer measurements, and Dr. Yakov Pachepsky, who offered many helpful comments during the preparation of this manuscript.

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