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MEASUREMENT OF UNSATURATED SOIL HYDRAULIC CONDUCTIVITIES USING A CERAMIC CUP TENSIOMETER

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The objective of this study was to develop and evaluate a simple field method to determine unsaturated hydraulic conductivities using measurements of water flux into a tensiometer. The tensiometer consists of a ceramic cup glued to one end of a piece of plastic tubing. A suction is first applied to the inside of the tensiometer, which is closed to the atmosphere. The reduced pressure in the tensiometer causes water to flow into the tensiometer from the soil. As the water flows into the tensiometer, the volume of air in the tensiometer decreases, and pressure increases. The rate of water flow into the tensiometer, water flux, is calculated from the measured pressures using a form of the ideal gas equation, PV = Constant, and its full differential, $PdV/dt+VdP/dt = 0$, where P is the measured pressure, and V is volume. The water flux is obtained from the change in volume with time, $-dV/dt$. The parameters for the unsaturated conductivity equation are determined by using a two-dimensional finite element soil model (2DSOIL) coupled with a Marquardt-Levenberg algorithm to fit calculated fluxes to measured ones. For comparison purposes, unsaturated hydraulic conductivities were also determined for the same soil within 25-cm-diameter rings from measured water contents and matric potentials during drainage for two locations. Fitted and measured fluxes agreed well. Unsaturated hydraulic conductivities obtained from the tensiometer inflow data, however, were much less than unsaturated hydraulic conductivities measured during drainage. We attributed the differences to anisotropy and scale effects although clogging of the tensiometer pores by fine soil material could also be a contributing factor. The method is relatively quick, uses inexpensive materials, provides consistent results and is not limited greatly by the conductivity of the cup. (Soil Science 1998;163:625–635)

Key words: One-step method, 2DSOIL, unsaturated flow.

TENSIOMETERS are used commonly to measure soil matric potential. The device is robust and precise. Soil physicists, hydrologists, and many other specialists in research and industry have a great deal of experience working with tensiometers. In this paper we show that a slight modification in tensiometer design and data processing may expand the utility of tensiometers in determining soil hydraulic properties.

Tensiometers are usually operated under the assumption that water potentials inside and outside the ceramic cup are near equilibrium and water flux through the ceramic cup is negligible during measurements. Suppose, however, that the pressure is perturbed from equilibrium such that the pressures inside and outside the tensiometer are no longer equal. Then water will flow through the ceramic cup. The pressure inside the tensiometer can be made less than atmospheric pressure by applying a vacuum. The pressure difference will then cause water to flow from the soil into the tensiometer. The capillary barrier in the porous ceramic cup will seal the interior of the tensiometer from the atmosphere provided the magnitude of the suction is not greater than the air entry pressure of the ceramic cup. Because the

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tensiometer remains sealed during flow, the air within the tensiometer, under isothermal conditions, will obey the ideal gas law in the form:

$$P(0)V_a(0) = P(t)V_a(t) \tag{1}$$

Here $P(0)$ and $V_a(0)$ are pressure and volume of air, respectively, in the tensiometer at the initial time when the vacuum has been increased, and $P(t)$ and $V_a(t)$ are pressure and volume of air in the tensiometer at some time, t, after the vacuum has been increased. Since the volume of the tensiometer tube is constant (and equal to $V_a(0)$, tensiometer initially empty of water), the volume of air in the tensiometer $V_a(t)$ will be decreasing by the volume of the total amount of water $V_w(t)$ that has flowed into the tensiometer:

$$V_a(t) = V_a(0) - V_w(t) \tag{2}$$

Using this scheme, a series of measurements of pressure within the tensiometer over time can also provide values of the cumulative water accumulation ($V_w(t)$) in the tensiometer by monitoring the pressure and calculating the air-filled volume:

$$V_w(t) = V_a(0) - \frac{P(0)V_a(0)}{P(t)} \tag{3}$$

The flux at any point in time, t, can be calculated as

$$Q(t) = \frac{dV_w(t)}{dt} \tag{4}$$

The initial volume ($V_a(0)$) was the air filled volume of the empty tensiometer, the initial pressure ($P(0)$) was the pressure recorded inside the tensiometer immediately after introducing a vacuum.

Suppose, we simulate the flow of soil water to the tensiometer using a numerical solution of the Richards equation for the soil surrounding the tensiometer cup. The dependence of the pressure within the tensiometer, $P(t)$, with a correction for the pressure decrease through the wall of the ceramic cup, serves as a boundary condition on the cup-soil interface. The calculated cumulative flux of water to the tensiometer $\hat{Q}(t)$ can be found from these simulations. This cumulative flux will depend on the soil hydraulic conductivity used in the simulations. If the soil hydraulic conductivity used in simulations differs significantly from the actual one, the simulated flux, $\hat{Q}(t)$, will differ from the actual flux $Q(t)$ calculated from the pressure according Eqs. (3) and (4). On the other hand, if the soil hydraulic conductivity used in simulations is close to actual soil hydraulic conductivity, we should expect a reasonable closeness between '$Q(t)$' and '$\hat{Q}(t)$' dependencies. It follows from this reasoning that measurements of the pressure inside the tensiometer after the pressure drop, along with values of measured water flux into the tensiometer, can provide data sufficient to find the soil hydraulic conductivity. Indeed, we can assume an equation relating the soil hydraulic conductivity to the soil matric potential and set up an automated computer search for the parameter values in this equation that will provide the '$\hat{Q}(t)$' dependence closest to the '$Q(t)$' dependence.

The use of a ceramic cup tensiometer to measure soil hydraulic conductivity was proposed as long ago as 1937 by L.A. Richards (Richards 1937). Previous techniques used either a positive pressure to move water out of the tensiometer into the soil (Upchurch 1980) or both positive or negative pressures (Hayashi et al. 1997) to measure the rate at which water is lost from (or gained by) the tensiometer. Both analysis methods treated the resistance of the ceramic cup as a resistance in series with the soil resistance. As a result, the conductivity of the ceramic cup had to be less than the conductivity of the soil.

The objective of this paper was to develop and to test (i) the above outlined technique of using the change in pressure inside a tensiometer, after introducing a vacuum, to determine the flux of water into a tensiometer and (ii) the feasibility of using these fluxes to determine the unsaturated soil hydraulic conductivity vs. matric potential dependency of a soil.

MATERIALS AND METHODS

Soil Properties and Field Water Retention

Two 25-cm-diameter rings were installed in a field, the soil in which was a Rumford loamy sand (Coarse-loamy siliceous thermic Typic Normudult). Two water-filled tensiometers were fitted with pressure transducers and installed at 7.5- and 15-cm depths in each ring. Two 30-cm-long time domain reflectometry (TDR) probes were installed horizontally from a trench at depths of 7.5 and 15 cm. The TDR and pressure transducers were monitored using a CR10 data logger (Campbell Scientific, Logan, UT) and a Tektronix 1502B cable tester (Tektronix Corp., Beaverton, OR). A calibration for the soil dielectric constant vs. soil water content relationship from a previous study at this site was used (Timlin and Pachepsky 1996). The soil was ponded until steady state was reached and then covered with plastic. Water content and matric potential were recorded for 5 days. The moisture retention curve was obtained from paired water content-matric potential measurements during redistribution. Unsaturated hydraulic conductivities (K(h)) were calculated from drainage data using the redistribution method of Green et al. (1986) and were represented by the Campbell equation (Campbell 1974):

$$K(h) = \begin{cases} K_s \left(\frac{h_b}{h}\right)^n, & h<h_b \\ K_s & h\geq h_b \end{cases} \qquad (5)$$

K_s is the saturated hydraulic conductivity (cm $\min^{-1}$), h is the absolute value of matric potential (cm), h_b is the air entry potential (cm), and n is an exponent.

Measurements of the Pressure in and the Flux into Tensiometers

A tensiometer having a high-flow ceramic cup (Soil Moisture Corp) with a 100-kPa air entry value was fitted with a rubber septum and a pressure transducer (Omega Series PX136, Omega Engineering, Stamford, CT) (Fig. 1). Pressures were recorded using a CR10 data logger (Campbell Scientific, Logan, UT) wired to the pressure transducer. The diameter of the ceramic cup was 22 mm. The tensiometer was installed into a hole in the soil within the ring made with a coring device with an inside diameter similar to the tensiometer outer diameter, and a slurry made from soil removed from the hole was used to insure a tight fit. The depth at the center of the cup was about 10 cm. A water-filled tensiometer was installed 10 cm from the flux measurement tensiometer to monitor soil matric potential. This tensiometer was placed in the same hole as the 7.5-cm-deep tensiometer used during drainage measurements but installed at a 10-cm depth. The 15-cm-deep tensiometer used during redistribution measurements was placed far enough away (>15 cm) to have negligible effect on the tensiometer inflow measurements.

The tensiometer used for measurement of flux was initially empty of all water. Air was partially removed from inside the tensiometer by using a syringe fitted with a hypodermic needle. After removing the air, initial pressures (relative to atmospheric pressures) were in the range of −27 to −22 kPa. Pressure measurements were complete when the pressure differential neared equilibrium. Accumulated water inside the tensiometer was recovered using a tube attached to a syringe after pressure measurements were complete. A single measurement required less than 25 minutes to complete. During this time, the tensiometer remained shaded to avoid temperature changes. Because the pressure transducer gave pressure relative to atmospheric pressure, the absolute pressure, P(t), was calculated as $P(t) = P(m) + P_{atm}$, where P(m) is measured pressure and P_{atm} is atmospheric pressure (assumed to be equal to approx. 100 kPa).

We carried out 15 trials at each location (without moving the tensiometer) over a range of soil water matric potentials (−1.6 to −4.0 kPa). The data were rejected if the recovered and calculated

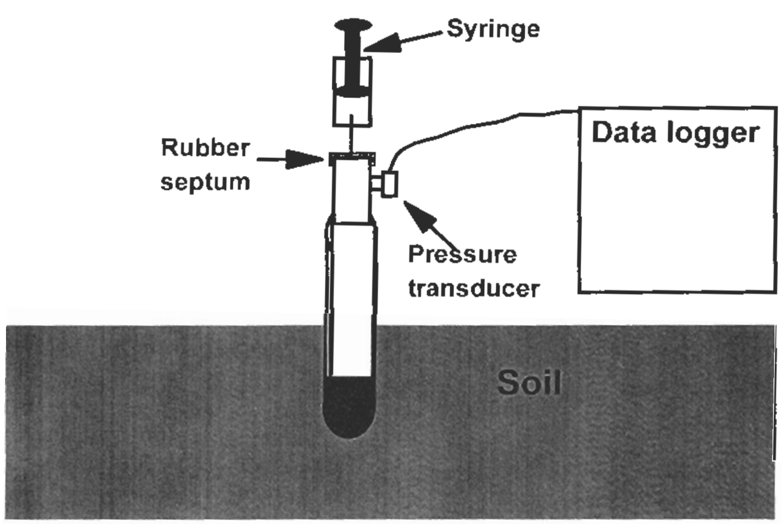

Fig. 1. Schematic of tensiometer setup for measuring water flux into a tensiometer.

volumes of water did not agree within a reasonable tolerance (about 10%). In all cases, the discrepancy was attributed to leakage that was repairable in the tensiometer-transducer connection. One value was discarded from Location 1 and five values were discarded from Location 2.

Calculation of Flux from Pressure Measurements

The pressures (as a function of time, t) were smoothed by fitting an equation of the form:

$$P(t) = \frac{a+bt}{c+dt+et^2} \quad (6)$$

The form of Eq. (6) was chosen because it provided a satisfactory fit to the type of curvature seen in the pressure vs. time relationships we observed. It is important ro reduce the noise in data used for an objective function in an optimization method because noise can lead to convergence difficulries and nonunique parameters (Russo et al. 1991). The smoothed pressures ($P(t)$) were used to calculate $V_w(t)$ using Eq. (3). Time dependent fluxes were calculated using Eq. (4) and these pressures. Inasmuch as $V_w(t)$ is a function of pressure, which is also a function of time, the derivative in Eq. (4) was calculated using the chain rule: $dV/dt = (dV/dP)(dP/dt)$:

$$Q(t) = \frac{P_0 V_0}{P(t)^2}\frac{dP}{dt} \quad (7)$$

dP/dt in Eq. (7) was estimated as a simple difference of smoothed pressures (Eq. (6)) and time for two successive time steps.

Inverse Method to Solve for Unsaturated Hydraulic Conductivities.

The purpose of the data analysis was to obtain values for water flux into the tensiometer and soil matric potential along the outside wall of the tensiometer from the pressures measured inside the tensiomerer. These fluxes and boundary pressures were used as input for an inverse method to optimize the unsaturated hydraulic conductivity vs. matric potential relationship.

We used a two-dimensional finite element representation of the Richards equation, 2DSOIL (Timlin et al. 1996; Šimůnek et al. 1994) coupled with a modified Marquardt-Levenberg algorithm (van Genuchten 1981) to minimize the sum of squared differences between the calculated and measured fluxes of water moving from the soil into the tensiometer. The grid and the flow domain used for the problem serup are shown in Fig. 2. The model was configured to simulate radial flow. The outer boundary of the flow domain was

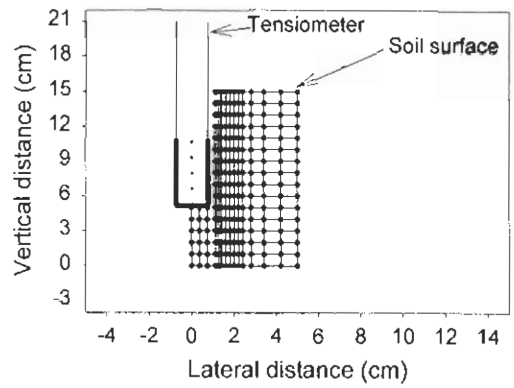

Fig. 2. Grid used in the numerical simulations by 2DSOIL. The ceramic cup is outlined by the bold lines.

set at a radius of 5 cm from the center of the tensiometer. This value was based on preliminary flow simulations that showed little change in matric potentials beyond 5 cm from the center of the tensiometer.

The objective function was the χ^2 statistic (Press et al. 1994):

$$\chi^2 = \sum_{i=1}^{N} \frac{[Q(t_i) - \hat{Q}(t_i)]^2}{s(t_i)^2} \quad (8)$$

where $Q(t_i)$ and $\hat{Q}(t)$ are measured and calculared fluxes at time t_i, respectively, and s_i^2 are the estimated variances of the data points. We did not have independent estimates of variances s_i^2, and estimated them from the dependence of the differences between actual and smoothed flux values on time. This dependence derived from data of all measurements was:

$$s_i^2 = \frac{0.56}{0.056 + \sqrt{\frac{t_i}{t_f}}} \quad (9)$$

where t_f is the time of the last measurement. When s_i^2 from Eq. (9) was used in Eq. (8), the squared residuals of the first and the last measurements had weights of approximately 0.1 and 1.9, respectively.

The model used in the inverse method (2DSOIL) required the time-dependent values of soil matric potential at the outside wall of the tensiometer as boundary values. The matric potential outside the tensiometer wall was calculated from the pressure inside the tensiometer and the conductivity of the ceramic cup. The Darcy flow equation was rearranged to solve for soil matric potential:

$$P(t)_{soil} = P(t) + \frac{Q(t)\, l}{AK_{cm}} \quad (10)$$

Here $P(t)_{soil}$ is the soil matric potential, $P(t)$ is the pressure inside the tensiometer, $Q(t)$ is flux, A and K_{cm} are the area and conductivity of the tensiometer cup, respectively, and l is the thickness of the cup wall. The values for A, K_{cm}, and l were obtained from specification tables for the ceramic cups given by Soil Moisture Corp (Santa Barbara, CA) and were assumed to be constant. It is assumed here that the pressure potential on the outside of the cup is the same as at the soil-cup interface and that the pressure drop through the cup wall is linear. The measured soil matric potential measured by the water filled tensiometer placed 10 cm from the flux measurement tensiometer was used as the initial soil matric potential and the matric potential at the outer boundary.

The soil moisture retention curve measured during drainage from θ(h) pairs was used to represent the dependency of water content on soil matric potential. The following equation (van Genuchten 1980) was used:

$$\theta = \theta_r + \frac{\theta_s - \theta_r}{[1 + (\alpha h)^n]^m} \quad (11)$$

Here θ_r is the residual water content, θ_s is the saturated water content, h (cm) the absolute value of matric potential, and α (cm^{-1}), n, and m are parameters.

RESULTS

The measured soil water retention curves, the parameters for van Genuchten's equation (Eq. (11)), and the predicted water retention curves are shown in Fig. 3. Generally, the moisture re-

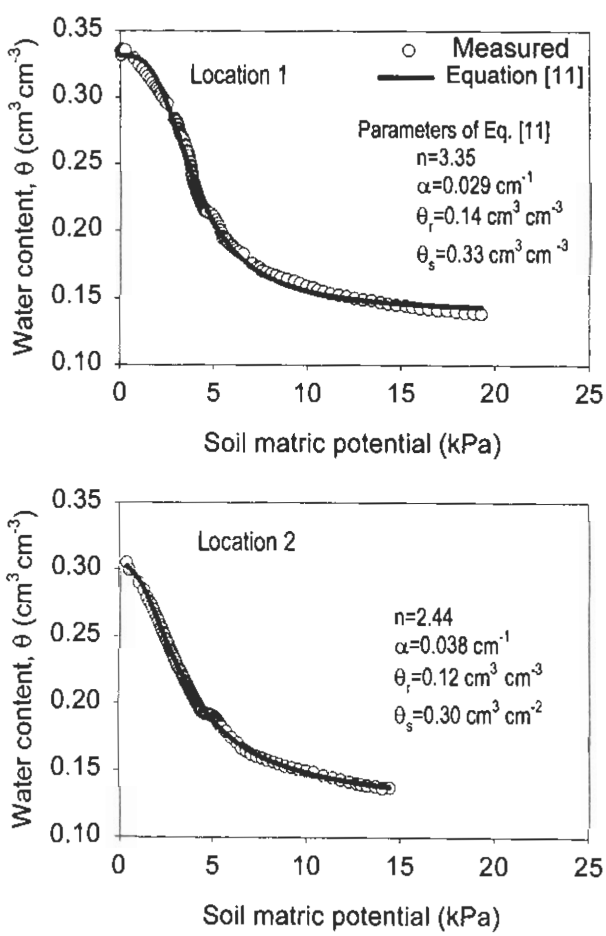

Fig. 3. The measured soil moisture retention curves for the two locations and the curves fit to Eq. (7).

lease curves show a sharp decrease in water content, with matric potential between saturation and a matric potential of 5 kPa (50 cm tension) (Fig. 3). The soil in Location 2 shows a larger decrease in water content between saturation and 5 kPa matric potential than the soil in Location 1. The value of the residual water content (θ_r) was high (0.14 and 0.12 cm^3 cm^{-3}) because we measured θ-h pairs for only a limited range of matric potentials in the wet range during drainage.

Calculated cumulative flux of water into the tensiometer and volume of water collected from the tensiometer agreed well (Table 1). This suggests that the pressure-volume relationship (Eqs. (2) and (3)) estimates water influx into the tensiometer correctly. Values of mean flux as a function of the matric potential of the soil surrounding the tensiometer at the starting time of the measurement are shown in Fig. 4 for both locations. Each data point represents a separate measurement. The data in this figure suggest that the rate of movement of water into a tensiometer was affected by the matric potential of the soil surrounding the tensiometer. The cumulative fluxes were generally higher in Location 2 than in Location 1 for the same soil matric potentials.

The measured pressures within the tensiometers as a function of time for two representative trials are shown in Fig. 5. These pressures ranged from −26 kPa when suction was applied to −11 kPa (relative to atmospheric) when the pressure and fluxes were nearing equilibrium. Although the pressures appeared smooth, there was enough noise to result in appreciable variation in the fluxes calculated using Eq. (4) and raw pressures (Fig. 3). The use of smoothed pressures in Eq. (7) resulted in a pressure-flux relationship with little noise. The differences in the flux vs. time dependencies for two trials shown in Fig. 5 are caused by differences in initial conditions and the soil properties at the two sites. The water flux into the tensiometer is seen to decrease more rapidly in Fig. 5b. Figure 4 shows that at Location 2, the mean fluxes were higher than at Location 1 for the same matric potential.

We did not encounter any problems with convergence during the optimizations using the smoothed data. Overall, the parameters for Eq. (5) for the data from the different trials were not very different from each other (Table 2). The residual root mean square errors for the differences between measured and calculated fluxes were low, indicating a good fit to the flux data (Table 2). We calculated unsaturated conductivity values for each set of tensiometer inflow measurements using the parameters given in Table 2 for the range of matric potentials that were measured during drainage. This gave a separate K(h) curve for each trial. In Fig. 6, we show the average of the predicted conductivities as well as the error bars that represent one standard deviation. The uniformity of the optimized parameters at Location 1 was less than at Location 2.

The magnitudes of the optimized conductivities were up to two orders of magnitude lower than the conductivities measured during drainage

TABLE 1

Measured and calculated final volume of water in the tensiometers, soil matric potential, and initial pressure in the tensiometer for the field trials

Location 1				Location 2			
Cumulative volume		Soil matric potential	Initial pressure in tensiometer	Cumulative volume		Soil matric potential	Initial pressure in tensiometer
Measured	Calculated			Measured	Calculated		
mL	mL	kPa	kPa	mL	mL	kPa	kPa
6.0	6.2	−3.5	−29.4	6.5	6.3	−2.9	−29.3
4.0	3.9	−1.4	−22.6	7.6	8.5	−3.4	−29.9
5.0	5.3	−2.3	−26.1	6.9	6.2	−4.0	−30.2
4.5	4.4	−2.9	−22.4	5.5	5.4	−2.1	−26.3
6.8	6.9	−4.0	−30.0	6.8	7.1	−2.9	−30.1
3.5	3.9	−3.7	−24.2	6.7	6.8	−1.6	−30.6
5.0	5.0	−3.5	−25.5	6.5	6.6	−1.9	−29.2
7.3	6.7	−3.6	−24.1	5.6	5.9	−2.4	−25.9
2.2	2.1	−3.7	−21.5	6.1	5.9	−2.5	−30.5
5.6	5.6	−3.5	−29.2	5.8	5.9	−2.4	−29.2
7.0	6.9	−2.1	−30.3				
4.0	3.8	−2.9	−22.7				
4.9	5.0	−2.6	−24.9				
7.2	7.2	−2.3	−26.4				

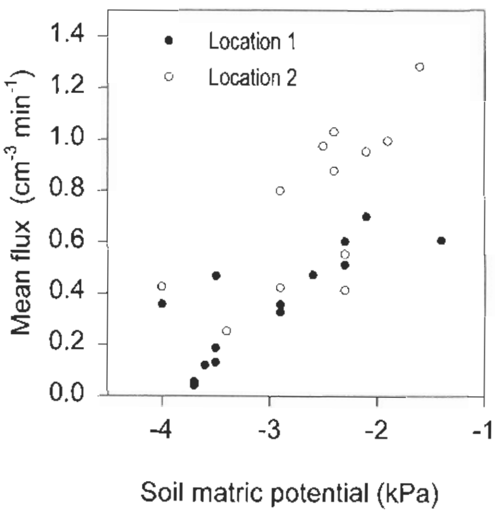

Fig. 4. Soil matric potential and mean flux of water into the tensiometer for individual trials and the two locations.

(Fig. 6). However, slopes of the measured and optimized log(K) vs. Log(h) curves were similar only at Location 2 (Table 2). The optimized unsaturated hydraulic conductivities from the tensiometer inflow measurements for the two locations were similar, whereas the field-measured conductivities showed larger differences when compared with each other (Table 2).

DISCUSSION

The tensiometer inflow method we have proposed here gave consistent results over a range of initial conditions and trials. The optimized parameters were fairly uniform over the trials within a location. The relative 'flux vs. matric potential' relationships for the two locations were also similar to the differences measured by drainage. As a result we have confidence that the dara we obtained reflected the hydraulic properties of the locations where the measurements were taken.

It is difficult to explain why the magnitude of the optimized conducrivities should be so much less than that we measured during drainage. We can only conjecture reasons for the differences in magnitude of conductivity values. We tried to account for poor contact between the tensiometer wall and the soil by decreasing the pressure drop between the inner wall of the tensiometer and the soil. The results, however, were not very differenr from the results obtained without considering poor contact. We believe that it is unlikely that poor contact with the tensiometer wall is a major contributor to the differences seen here. We used a slurry of the resident soil to ensure a tighr fit around the tensiometer, and the thickness of this slurry was much less than 1 mm. Furthermore, we conjecture that small soil particles will be drawn to the tensiometer walls along with rhe water and fill in small gaps during influx of water. When we re-

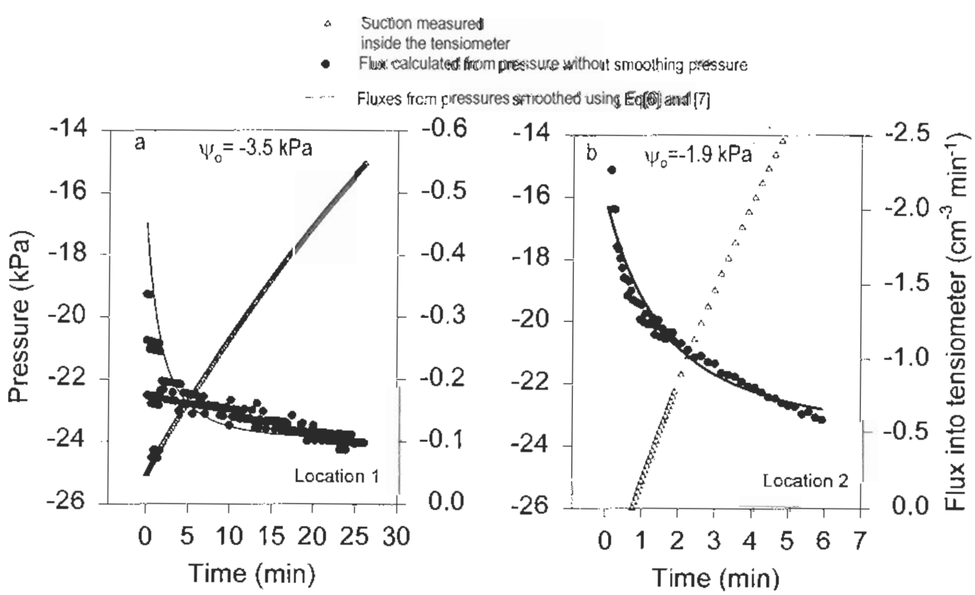

Fig. 5. Pressures measured inside the tensiometer, and fluxes calculated from raw and smoothed pressures. The data are from two of the trials with different initial soil matric potentials, $\psi = -3.5$ kPa (a) and $\psi = -1.9$ kPa (b).

TABLE 2

Parameters for Eq. (7) fit by the optimization method for each trial

Location 1				Location 2			
K_s	n	h_b	χ^2 (Eq. (8))	K_s	n	h_b	χ^2 (Eq. (8))
cm min^{-1}		cm	cm^3 min^{-1}	cm min^{-1}		cm	cm^3 min^{-1}
5.47e-02	3.68	5.5	0.326	4.65e-02	3.05	7.5	0.090
2.20e-01	3.49	2.4	0.088	4.43e-02	3.55	7.7	0.192
3.76e-02	3.22	7.2	0.235	1.63e-01	3.63	5.0	0.254
4.59e-02	3.89	4.0	0.039	8.51e-02	2.96	4.3	0.042
1.22e-02	2.94	10.2	0.177	3.35e-03	2.96	12.2	0.220
6.39e-02	3.93	4.5	0.253	1.40e-01	3.60	5.2	0.136
7.12e-02	3.96	6.1	0.148	1.64e-01	3.11	4.6	0.085
1.99e-01	2.66	2.4	0.104	5.31e-02	3.43	7.2	0.037
1.70e-01	3.55	4.4	0.151	9.83e-03	3.07	10.8	0.107
4.31e-01	3.61	3.5	0.092	8.30e-02	3.27	6.3	0.042
8.58e-02	3.67	5.4	0.059	6.40e-03	3.01	13.8	0.056
4.58e-02	3.61	6.5	0.067				
5.94e-02	3.24	6.8	0.057				
8.75e-02	3.57	5.4	0.031				
Overall†							
1.36e-02	3.35	8.6		1.72e-02	3.23	9.4	
Parameter for drainage data							
5.91e-02	2.40	10.6		9.10e-02	3.10	13.8	

†Parameters for Eq. (7) fit to K (h) pairs predicted using parameters from individual trials.

moved the tensiometer we noticed strong cohesion between the soil and the tensiometer cup. Of course this layer may also have reduced the conductivity at the cup-soil interface.

The conductivity of the ceramic cup may also change over time. This problem is probably common to all one-step methods that use a porous ceramic plate. The conductivity of the ceramic material should probably be measured before and after experiments with soil. We used a pressure boundary condition between the tensiometer wall and the soil, and, thus, the conductivity of the ce-

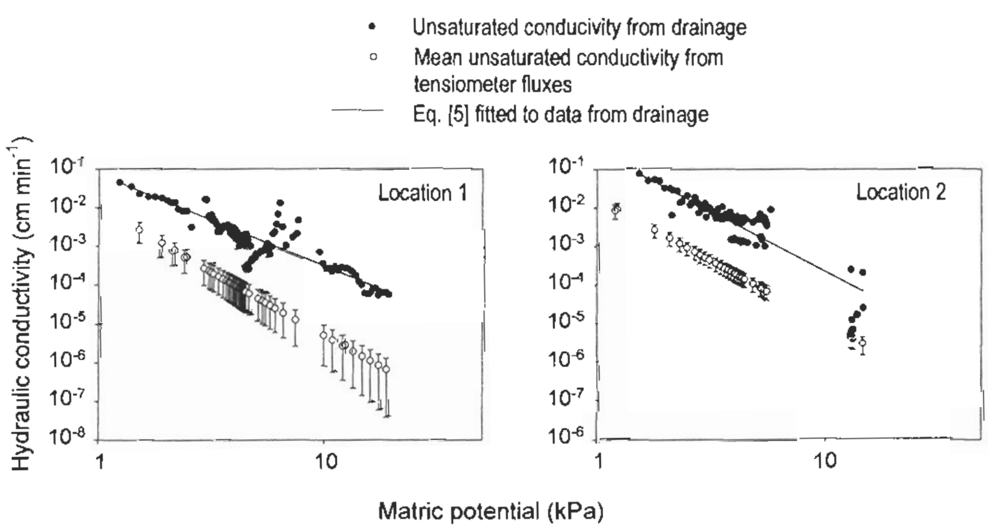

Fig. 6. Unsaturated hydraulic conductivities calculated from drainage data and from the tensiometer inflow data.

ramic cup was not a strongly limiting factor in the simulations. Differences in cup conductivity would be manifested as differences in boundary pressures. If the actual cup conductivity is lower than the specified value, the boundary pressure used in the simulations will be less. Numerical simulations showed that changes in cup conductivity of as much as one order of magnitude did not result in differences in optimized conductivities as high as two orders of magnitude.

We considered errors caused by the presence of vapor pressure in the tensiometer. There will be some water vapor in the tensiometer because the ceramic cup is saturated with water. The Gibbs equation shows that the vapor pressure increases as the total pressure on the liquid water increases (Castellan 1983, p. 270). The increase is proportional to the ratio of the molar volume of the liquid to the molar volume of the vapor. Since the molar volume of the liquid water is very much less than the molar volume of the vapor water, the rate of increase of vapor pressure with total pressure will be small. The differences in water vapor pressure between any two successive pressure measurements will be also small, and there will be negligible condensation under isothermal conditions. The magnitude of the vapor pressure is approximately 1.4 kPa at 20°C compared with the total air pressure in the tensiometer of about 60 to 100 kPa. As a result, the effect of vapor pressure on the measurements and volume calculations will not be significant.

Another possible reason for the differences could be the effect of the different scales of measurements. The redistribution data was on the scale of the 25-cm-dia. infiltration ring. The scale of the conductivity measurements by the single tensiometer was within a radius of about 5 cm. Rovey (1994) has shown that the saturated hydraulic conductivity increases exponentially with the radius of the measurement area. Ehlers (1976) compared unsaturated conductivities from a drainage-based field method and 5.4-cm-diameter cores. The conductivities from the field method were greater than from the core method near saturation. The differences, however, were not as large in our case. The differences in scale, however, were greater for our study.

Anisotropy may also be a contributing factor as the drainage flux was in a vertical plane while the water flux into the tensiometer was in a horizontal plane. Paige and Hillel (1993) compared conductivities obtained from a drainage method, soil cores, and a Guelph Permeameter. The unsaturated conductivities from the Guelph Permeameter were two to three orders of magnitude less than the values obtained using the other methods. The differences were attributed to anisotropy inasmuch as the horizontal fluxes are dominant in the Guelph Permeameter.

Errors in unsaturated hydraulic conductivities from the drainage method may be as high as one order of magnitude (Flühler et al. 1976). Kim et al. (1992) reported conductivities obtained from a one-step method to be lower than conductivities obtained using drainage and evaporation. The one-step method used suction applied at the bottom of the plate.

There may be other reasons for the differences—for example, flow characteristics under high gradients—that wait for further research. Little is known about water flow between the porous ceramic tensiometer cup and the soil.

The tensiometer inflow method is similar to one-step methods (Parker et al. 1985; van Dam et al. 1994). The difference is that one-step methods require measurement of two variables, flux and pressure. In the method proposed here, only pressure is measured. The use of a ceramic cup tensiometer is an extension that has been investigated by Upchurch (1980) and Hayashi et al. (1997). As with our method, the pressure drop and both pressure and flux dynamics are measured. Both Upchurch (1980) and Hayashi et al. (1997) reported that their tensiometer methods were sensitive to the conductivity of the cup and that the cup conductivity limited the higher range of conductivities that could be measured. These limitations were caused, in part, by the assumptions of their numerical analysis, although the conductivity of the cup can place an upper limit on the resultant flux. In the methods reported, the resistance of the soil and the cup were placed in series and, therefore, it was necessary to require that the conductivity of the cup be higher than the conductivity of the soil.

The method described in this paper does not assume that the cup conductivity is greater than the soil hydraulic conductivity. We separated the cup conductivity from the soil conductivity by calculating a pressure drop through the cup wall and using the resultant pressure on the outside wall of the tensiometer as a boundary condition for the two-dimensional model. This involved assumptions that have their own limitations, and further work is needed to evaluate them. A method that incorporates some of the analytical approaches of Upchurch (1980) and Hayashi et al. (1997) with the approach outlined here may be promising.

The proposed method has several advantages because it can be used *in-situ* and it does not require soil sampling or direct measurements of the fluxes. Rapid flux measurements can be made at different water contents. We used the measured 'matric potential – water content' relationship (θ(h)) to estimate k(h). It may be useful to try the method and optimize for the θ(h) as well as K(h). We anticipate that the method will not be very successful and that the chances of obtaining nonunique parameters are greater.

Our objective in testing this method was to see if such measurements with a tensiometer were feasible. We conclude that this method is feasible, but we cannot say with certainty that we are measuring the same hydraulic conductivities measured by other techniques currently used. If there is a strong scale effect because unsaturated conductivities on the scale of the tensiometer are much smaller than those from other instruments, then this instrument and technique would be very valuable as a model to study the hydraulic characteristics of soil and water flow near plant roots.

SUMMARY

We have proposed a method that utilizes a commonly used tensiometer to determine the unsaturated hydraulic conductivity-matric potential dependency of soil. We measured flux of water from soil to a ceramic cup tensiometer by monitoring pressures inside the tensiometer after applying a vacuum. We coupled a two-dimensional radial flow model and a Marquardt-Levenberg optimization routine to obtain the parameters of the K(h) relationship by minimizing the differences between optimized and measured fluxes. The cumulative flux calculated from pressure measurements was similar to the volume of water recovered from inside the tensiometer after a trial, suggesting the method used to obtain flux from pressure measurements was sound. The optimized parameters for 24 trials (flux measurements) in two locations were uniform within a location. The K(h) relationship was determined by applying the fitted parameters from each trial to a range of matric potentials and calculating mean values of conductivity over all of the trials for each matric potential. We compared K(h) from tensiometer inflow data to K(h) obtained from drainage data. The slopes of the log transformed, optimized K(h) were similar to the ones obtained from drainage data but the optimized values of conductivities were one to two orders of magnitude less. The differences may be attributable to scale effects or anisotropy, and further research is required.

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