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Determining Soil Hydraulic Properties using Tension Infiltrometers, Time Domain Reflectometry, and Tensiometers

Dong Wang,* S. R. Yates, and F. F. Ernst

ABSTRACT

Tension infiltrometers have become a popular instrument for field determination of soil hydraulic properties. To develop and test different models for parameter estimation based on tension infiltrometer measurement, we obtained simultaneous measurements of transient tension infiltration rate, soil water content, and tension using small time domain reflectometry (TDR) probes and tensiometers installed at fixed locations relative to the infiltrometer disk. Infiltration was made with 10- and 20-cm-diam. disks under 1 and 5 cm of water supply tensions. The soil is an Arlington fine sandy loam (coarse-loamy, mixed, thermic Haplic Durixeralf). Wooding's steady-state approximate solution for water flow from a surface circular pond was used to estimate the saturated hydraulic conductivity (K_s) and an empirical parameter (α_G) used in Gardner's exponential hydraulic conductivity function. These two parameters (i.e., K_s and α_G) were then independently estimated using an integral form of the steady-state Darcy–Buckingham flux law. A sorptivity method was also proposed as an alternative to Wooding's steady-state approach. Calculated K_s and α_G with the Darcy–Buckingham flux law method was in good agreement with estimates using Wooding's steady-state approximation. The sorptivity method produced K_s estimates that were statistically similar to those obtained with Wooding's method. The $K(h)$ inferred from measured $\theta(h)$ underestimated the conductivity close to saturation compared with estimates obtained from the infiltrometer measurements.

OBTAINING SOIL HYDRAULIC PROPERTIES representative of field soil conditions is an important step in understanding the dynamic processes of water and solute movement in the soil. Variables that are commonly used to describe water flow in the soil include infiltration rate, soil water content, and tension. Characteristics of infiltration and the soil water content–tension relationship are controlled by soil hydraulic properties such as saturated hydraulic conductivity, sorptivity, and some empirical parameters for describing these characteristics.

Methods available for determining these hydraulic parameters are often difficult to use and time consuming. Tension infiltrometers (Ankeny et al., 1988; Perroux and White, 1988) are useful instruments that offer a simple and fast means of estimating soil hydraulic properties and structural characteristics based on infiltration measurement at the soil surface, when combined with appropriate theoretical principles or procedures. The most widely used method for parameter estimation based on tension infiltrometer measurement is to use the approximate steady-state solution of water flow from a surface circular source by Wooding (1968). Other methods include the determination of sorptivity and a macroscopic capillary length (White et al., 1992) and numerical inversion (Vogeler et al., 1996; Simunek and van Genuchten, 1996). While Wooding's method requires the tension infiltration to reach the steady-state rate, other methods need accurate measurements of transient infiltration rate for a preselected tension.

Experimentally, the recognition of steady-state flow may be prone to subjective decisions and sometimes it is limited by the amount of water available in the water supply tube of a tension infiltrometer, such as in soils with large water intake rates. For soils with fine textures and small infiltration rates, infiltrometers with automated recording mechanisms, such as the one described by Ankeny et al. (1988), may be required because of the extended time needed to reach steady-state flow. Besides the drastic differences in infiltration rate for different types of soil, the size of the infiltrometer disk and supply tension also affect the time needed to approach steady-state conditions. The use of automated recording can also provide more detailed and accurate measurement of the transient infiltration process, which would enable the use of non-steady-state methods of parameter estimation.

Attempts have been made to assess the accuracy of soil hydraulic parameters estimated with tension infiltrometer measurements using Wooding's approximate solution. Estimated values of hydraulic conductivity for saturated soils were found to be within 5 to ≈300% of those found using numerical simulations or other laboratory measurements (Reynolds and Elrick, 1991; White and Perroux, 1989; Ankeny et al., 1991). Besides the large variation in estimated conductivity using different methods, comparison of the steady-state infiltrometer method (or Wooding's approach) with direct field measurements is not available in the literature and merits investigation to further validate the technique for estimating soil hydraulic properties.

Recent development and application of TDR has provided an accurate, rapid, in situ method of measuring the volumetric soil water content. Theoretical principles of measuring soil water content with TDR have been thoroughly studied and discussed by Topp et al. (1980) and Dalton (1992). Baker and Allmaras (1990) described an automated multiplexing system that can provide continuous measurement of soil water content at many locations in the soil. Field determination of water flow or the change in soil water content under two- or three-dimensional flow regimes has been difficult because of its highly dynamic nature. Using TDR, Kachanoski et al. (1990) measured water redistribution from a surface circular source and found no significant difference by using either straight or curved wave guides. More recently, Vogeler et al. (1996) obtained simultaneous measurements of infiltration rate with a tension infiltrometer, and water and solute content with small TDR probes placed underneath the infiltrometer

U.S. Salinity Laboratory, Soil Physics and Pesticides Research Unit, 450 West Big Springs Road, Riverside, CA 92507-4617 *Corresponding author (dwang@ussl.ars.usda.gov).

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disk. The use of small TDR probes offers a means of nondestructively measuring soil water content in a vicinity that approximates point measurements. This is very useful in characterizing water flow in a two- or three-dimensional flow problem. Detailed discussions on the sensitivity of small TDR probes to water content change can be found in Peterson et al. (1995).

Soil water tension describes the energy status of soil water and, when combined with water content, provides the characteristic water retention relationship that is a fundamental property for describing dynamic processes of water and solute transport in the soil. The use of tensiometers for measuring soil water tension is the most popular technique in soil physics and irrigation research. A detailed discussion of the theory and application of tensiometers for field determination of soil water tension can be found in Cassel and Klute (1986). To depict soil water status in two- or three-dimensional flow conditions, tensiometers are needed that possess the following properties: (i) they should provide an accurate tension measurement with a sensitivity preferably in the range of ± 1 mm of water tension; (ii) the response time should be small in order to capture the dynamics of water flow; (iii) the size of the buried porous section needs to be small enough not to interfere with water flow; and (iv) the tension range that can be measured without breaking the tension should be reasonably large so that soil water status during a complete infiltration event (i.e., from initial dry soil to steady state near saturation) can be adequately measured. Data collection from tensiometers has been a time-consuming process when they are read manually. Lowery et al. (1986) described an automated electrical readout system that can be used for taking tensiometer readings very efficiently. The use of sensitive pressure transducers and multichannel data logger systems has further enabled the frequent and multilocational application of tensiometers.

Our goal was to provide a simple and accurate determination of soil hydraulic parameters based on tension infiltrometer measurement. The overall objective of this study was to compare soil hydraulic parameters estimated from a different combination of measured variables including infiltration rate under different supply tensions, and soil water contents and tensions during the tension infiltration process. Combinations of information include: (i) tension infiltration alone, the Wooding's method; (ii) tension infiltration and soil water tension measurements, the Darcy–Buckingham flux law method; (iii) tension infiltration and soil water content data, the sorptivity method; and (iv) measured soil water content and tension pairs, fitting the retention curve.

THEORY

Soil Water Retention and Hydraulic Conductivity Functions

The relationship between water content and tension is a fundamental hydraulic characteristic for any soil. Many other hydraulic properties are derived from this basic relationship. To relate the two basic soil hydraulic variables, van Genuchten (1980) applied the following form:

$$\theta(h) = \theta_r + (\theta_s - \theta_r)\left[\frac{1}{1 + (\alpha_{VG}h)^n}\right]^{(1-1/n)} \quad [1]$$

where $\theta(h)$ is the water content ($L^3\ L^{-3}$) at tension h (L), θ_r and θ_s are residual and saturated water content, respectively ($L^3\ L^{-3}$), and α_{VG} (L^{-1}) and n (dimensionless) are fitted parameters that control the shape of the $\theta(h)$ curve. The rate of water and solute transport in an unsaturated soil is greatly affected by the hydraulic conductivity relationship, $K(h)$ ($L\ T^{-1}$) under a given soil moisture regime. Based on Mualem (1976), van Genuchten (1980) used the following form to describe $K(h)$ as a function of water tension:

$$K(h) = K_s\frac{\{1 - (\alpha_{VG}h)^{n-1}\ [1 + (\alpha_{VG}h)^n]^{(1/n)-1}\}^2}{[1 + (\alpha_{VG}h)^n]^{(n-1)/2n}} \quad [2]$$

where K_s is the hydraulic conductivity of a saturated soil ($L\ T^{-1}$). This form is useful because the same sets of parameters, i.e., α_{VG} and n, can be used to describe both the water retention and the hydraulic conductivity functions for a given soil.

An alternative approach for $K(h)$ was used by Gardner (1958), using an exponential expression:

$$K(h) = K_s \exp(-\alpha_G h) \quad [3]$$

where α_G (L^{-1}) is an empirical fitting parameter. This is also a very useful relationship because at steady state Richards' water flow equation can be linearized. Subscripts VG and G for the fitting parameter α are used in Eq. [2] and [3], respectively, to distinguish the two different forms of hydraulic conductivity functions.

Tension Infiltration Models

Wooding's Method

Water flow from a tension infiltrometer disk is a three-dimensional flow system. Temporal changes in soil water content can be described with Richards' equation using initial and boundary conditions defined for geometric and hydraulic parameters specific to the infiltrometer. Because there is no exact analytical solution to such a transient three-dimensional water flow equation subject to the initial and boundary conditions of a tension infiltrometer, numerical inversion has been used to solve for transport parameters based on known flow variables such as transient water content or infiltration rate. Under steady state, Wooding (1968) solved for the infiltration rate from a shallow circular pond of radius r_o (L) based on Gardner's exponential hydraulic conductivity function, i.e. Eq. [3], and found the following approximate solution:

$$q(h_o) = K_s\left(1 + \frac{4}{\pi r_o \alpha_G}\right) \exp(-\alpha_G\ h_o) \quad [4]$$

where $q(h_o)$ is the steady-state water flux density ($L\ T^{-1}$) under a given supply tension h_o (L). Because the only unknowns in this equation are K_s and α_G, they can be solved by making measurements at a fixed disk radius with multiple supply tensions or at a fixed tension with disks having variable radii. Detailed procedures of solving the equation for K_s and α_G can be found in Hussen and Warrick (1993).

Darcy–Buckingham Flux Law Method

Because the supply tension is fixed and uniform across the infiltrometer disk during an infiltration event, water flow from a tension infiltrometer can be approximated as a one-dimensional vertical flow system in regions next to the soil surface and directly under the infiltrometer disk. If we know the water tension at a given depth (not too far from the surface), such as z_1 (L) shown in Fig. 1, we can write the steady-state Darcy–Buckingham flux law as

$$q(h_o) = K(h)\,\frac{(h_1 - h_o) + (z_1 - 0)}{(z_1 - 0)} \rightarrow K(h)\left(\frac{dh}{dz} + 1\right) \quad [5]$$

where h_1 (L) is the water tension at depth z_1. Rearranging Eq. [5] and integrating with respect to h (from h_o to h_1) and z (from 0 to z_1), we have

$$z_1 = \int_{h_o}^{h_1} \left[\frac{1}{\dfrac{q(h_o)}{K(h)} - 1}\right] dh \quad [6]$$

Substituting in Eq. [3] for $K(h)$ and solving for $q(h_o)$, we obtained the following form:

$$q(h_o) = K_s\left\{\frac{\exp[\alpha_G\,(h_1 - h_o + z_1)] - 1}{\exp(\alpha_G\, z_1) - 1}\right\}\exp(-\alpha_G\, h_1) \quad [7]$$

Equation [7] has a similar form to the Wooding's approximate solution, i.e. Eq. [4], where the unknowns K_s and α_G can be solved for by making measurements at a fixed disk radius with multiple supply tensions.

Sorptivity Method

Because the determination of steady-state flow rate can be subjective and limited by experimental conditions, methods using early-time infiltration data to estimate soil hydraulic properties become attractive. The following procedures are based on early-time infiltration data and water content increase during an infiltration event.

The first step is to solve for the soil sorptivity S_o (L T$^{-1/2}$) under supply tension h_o from the transient tension infiltration data using an approximate infiltration equation by Warrick (1992):

$$q(h_o,t) \approx S_o/2\sqrt{t} + D_e S_o/r_o \quad \text{for small times} \quad [8]$$

where $q(h_o,t)$ is the transient infiltration rate (L T^{-1}) under supply tension h_o, t is time (T), and D_e is a fitting parameter (L T$^{-1/2}$) that is a constant for a given set of h_o and r_o. Both S_o and D_e are fitted with a nonlinear best-fit program.

The second step is to solve for K_s from a relationship developed by Youngs (1987) using S_o and the measured water content increase for small times

$$K_s = 342.25\,\frac{\eta\rho g}{\sigma^2(\theta_o - \theta_i)^2}\,S_o^4 \quad [9]$$

where η is water viscosity (M L^{-1} T^{-1}, assuming isothermal), ρ is the density of water (M L^{-3}), g is the gravitational acceleration (L T^{-2}), σ is the surface tension of water (M T^{-2}), and θ_o and θ_i are the final and initial water contents, respectively (L^3 L^{-3}), at the supply surface.

The final step is to solve for α_G from White and Sully (1987) using the estimated S_o and K_s values:

$$\alpha_G = \frac{(\theta_o - \theta_i)K_s}{bS_o^2} \qquad b \approx 0.55 \quad [10]$$

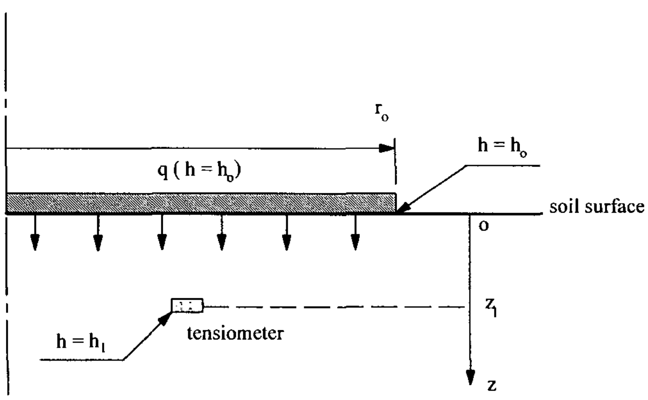

Fig. 1. Schematic of water flow under a tension infiltrometer disk next to the soil surface. $q(h = h_o)$ is the steady-state water flux density when the supply tension (h) equals h_o; r_o is the radius of the infiltrometer disk. The tensiometer measures a steady-state water tension of h_1 at depth z_1 in the soil.

MATERIALS AND METHODS

The Soil

A field experiment was conducted to obtain simultaneous measurements of transient tension infiltration rate, soil water content, and tension. The soil is an Arlington fine sandy loam (coarse-loamy, mixed, thermic Haplic Durixeralf) with an Ap horizon for the surface 10 cm. Within this depth, the particle-size distribution consists of 63.5% sand, 29.7% silt, and 6.8% clay. From soil core measurements, the residual and saturated soil water contents were 0.077 ± 0.006 and 0.371 ± 0.014 cm^3 cm^{-3}, respectively. Soil bulk density within this depth was 1.53 ± 0.03 g cm^{-3}. Since no definable structure can be observed, the soil is considered massive (R.C. Graham, 1997, personal communication).

Tension Infiltrometer Measurements

Water infiltration was measured with a tension infiltrometer (Soil Measurement Systems, Tucson, AZ) that had the infiltration disk attached to the water supply reservoir and tension control tubes via a flexible tubing. This design was considered advantageous since the disk alone is much lighter than the conventional design of attaching a disk to the bottom of reservoir tubes; this design would therefore reduce the chance of compacting the soil under the infiltration surface. A level was used to assure that the disk and infiltrometer base were at the same level or with a zero relative distance, so that the head between the bubbling outlet at the bottom of the water supply tube and the disk membrane was constant. We used a 21X datalogger (Campbell Scientific, Logan, UT) and two pressure transducers (MICRO SWITCH, Honeywell, Fort Washington, PA) to measure and record transient infiltration rate in a setup similar to that described by Ankeny et al. (1988). A layer of about 1 mm of no. 60 silica sand (diameter ≈ 250 μm; $K_s \approx$ 283 m d^{-1}; water entry ≈ 22 cm) was used between the disk membrane and the smoothed soil surface to improve hydraulic contact. The measurements were made with combinations of two disk diameters (10 and 20 cm) and two supply tensions (1 and 5 cm). Sufficient time (≥48 h) was given for the soil to hydraulically equilibrate between measurements, so that a similar initial condition was obtained for each measurement.

TDR and Tensiometer Measurements

To measure soil water content and tension, TDR probes and tensiometers were installed prior to the infiltrometer measurements. As shown in Fig. 2, two 10-cm three-rod probes (TDR #1 and #2, rod diameter = 1.6 mm, spacing between rods = 10 mm) were installed horizontally at 2.5-cm depth and at opposite directions for the radial distance of 0 to 10

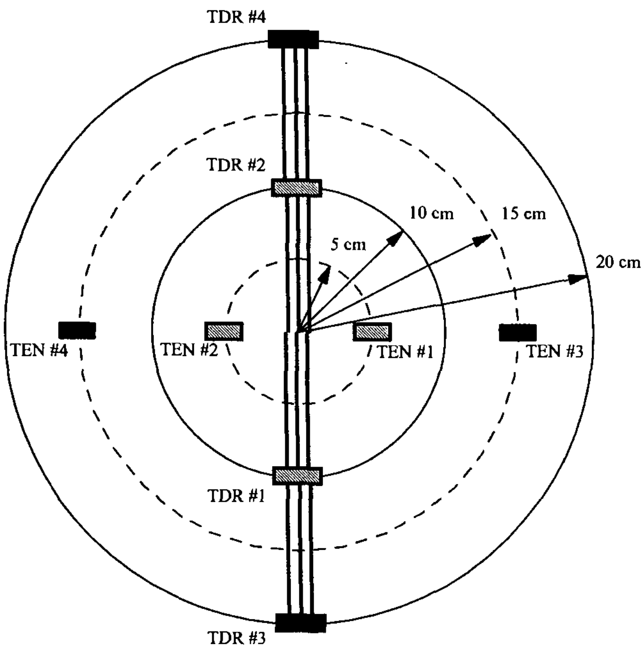

Fig. 2. Location of time domain reflectometry (TDR) probes and tensiometers (TEN) relative to the location of tension infiltrometer measurements; z = depth from the soil surface.

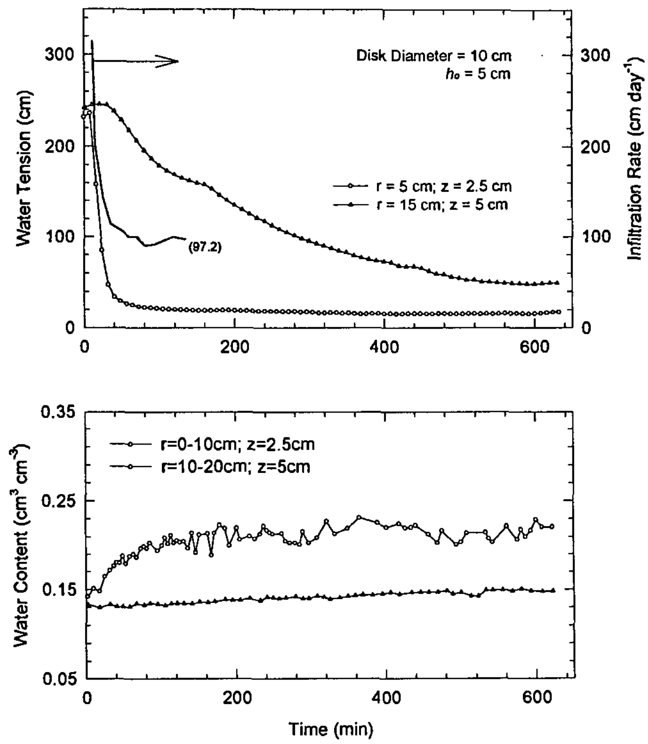

Fig. 3. Infiltration flux from a 10-cm-diam. infiltrometer disk under 5-cm supply tension (h_0), and measured soil water potential and water content at fixed locations relative to the center of the disk; r = pond radius and z = depth from the soil surface.

cm from the center of the infiltrometer disk. A pair of tensiometers (TEN #1 and #2, 2 cm long and 1-cm o.d., pore diameter = 2.2 μm, Coors Porcelain Co., Golden, CO) were also installed horizontally at 2.5-cm depth at an average radial distance of 5 cm from the center of the infiltrometer disk. Because of radial symmetry and to avoid interference with the TDR readings, the tensiometers were placed in line with the center of the infiltrometer disk and in a direction perpendicular to the TDR probes. To observe water content and tension change at a larger distance away from the source, we installed another set of probes (TDR #3 and #4) and tensiometers (TEN #3 and #4) at 5-cm depth and at an average radial distance of 15 cm from the center of the infiltrometer disk. The TDR probes were installed by excavating 3-cm-wide trenches to 2.5- and 5-cm depths and inserting the probes into the undisturbed soil from the end of the little trenches. The tensiometers were installed in a similar fashion except that we used a drill (diameter ≈ 9.5 mm) to create a cavity at the end of the little trenches where the tensiometers were inserted. The trenches were backfilled to conditions similar to the original soil and left for 24 h to attain hydraulic equilibrium before initiating the first infiltration experiment.

We used a 1502 B Tektronix cable tester (Tektronix, Beaverton, OR) with a Campbell SDM1502 communication interface and a SDMX50 multiplexer to read the four TDR probes with a CR10 datalogger (Campbell Scientific). The TDR setup was calibrated in the laboratory against gravimetric water content measurements using field soil packed to an average bulk density of 1.53 g cm^{-3} and under five volumetric moisture contents. Because the salinity of the soil was very low (<0.5 dS m^{-1}), we assumed that laboratory calibration with deionized water can be used to represent field conditions.

The tensiometers were measured with pressure transducers and the 21X datalogger that was also used for the infiltrometer recording. Deaired water was used to fill the tensiometers and thick-wall Tyflon tubing (diam. ≈ 1 mm) that connected the tensiometer porous cups to the pressure transducers. The transducers were calibrated in the laboratory with water manometers prior to field use.

RESULTS AND DISCUSSION

Measured Infiltration, Water Content, and Tension

With the 10-cm-diam. disk under 5-cm supply tension, the measured infiltration rate decreased drastically from an initial 314.3 cm d^{-1} to a quasi-steady-state rate of 97.2 cm d^{-1} in <2 h (Fig. 3). Initial soil water tension was ≈250 cm of water at either 2.5- or 5-cm depth. Corresponding initial soil water content was about 0.14 cm^3 cm^{-3}. Tensiometers and TDR probes at 2.5-cm depth responded promptly to the infiltration event, with measured tension and water content approaching 15 cm and 0.22 cm^3 cm^{-3}, respectively. Tensiometers and TDR probes at 5 cm exhibited a considerable delay in responding to infiltration because of increased radial distances from the water source or the infiltrometer disk, compared with sensors at the 2.5-cm depth.

After water redistribution and drainage (≥48 h), we started another infiltration event with the same disk (10 cm in diameter) but reduced supply tension to 1 cm (Fig. 4). Initial soil water tension reached 500 cm at the 2.5-cm depth with an initial water content of about 0.105 cm^3 cm^{-3}. Measured infiltration rate decreased from an initial maximum of 428.6 cm d^{-1} to a final 128.6 cm d^{-1}. Both the initial and final infiltration rates are larger than the previous measurements, i.e., with 5-cm supply tension. This was caused by the reduced supply tension and higher initial soil water tension, hence larger tension gradient or driving force. Water flow reached a quasi-steady state with the tension approaching 17.3 and 51.7 cm at 2.5- and 5-cm depths. Water content at 2.5 cm reached about 0.23 cm^3 cm^{-3} before starting to decrease due to drainage.

The TDR and tensiometers responded to surface infiltration more rapidly when the size of the infiltrometer disk was increased from a diameter of 10 to 20 cm. With supply tension set at 5 cm, the infiltration rate decreased from an initial 469.1 to 57.1 cm d^{-1} in 50 min (Fig. 5). Measured initial soil water tensions were 350 and 275 cm at 2.5- and 5-cm depths, respectively. Corresponding water contents were 0.085 and 0.110 cm^3 cm^{-3}. The final soil water tension reached 56.1 and 80.9 cm at 2.5- and 5-cm depths, respectively. Final water content averaged about 0.25 and 0.16 cm^3 cm^{-3}.

Reducing supply tension to 1 cm, the infiltration rate started with an initial maximum of 329.1 cm d^{-1} and reached a quasi-steady-state rate of 80.4 cm d^{-1} in about 45 min (Fig. 6). Final soil water tension reached about 10 and 36 cm at 2.5- and 5-cm depths where the water content was about 0.29 and 0.18 cm^3 cm^{-3}, respectively.

Estimated Parameters

Using Wooding's method, estimated K_s ranged from 22.3 to 35.2 cm d^{-1} and α_G from 0.0540 to 0.0856 cm^{-1}, and both parameters had small variations between measurements (Table 1). With the same sets of steady-state infiltration rate or $q(h_o)$ as used in the Wooding's solution and direct tensiometer measurements at 2.5-cm depth for the 20-cm disk, we found K_s = 34.2 cm d^{-1} and α_G = 0.1407 cm^{-1} using the Darcy–Buckingham flux law method. Both of these estimates are in reasonable agreement with estimates using Wooding's method. While Wooding's approximate solution can be used to estimate soil K_s and α_G with the measurement of steady-state infiltration rate, the Darcy–Buckingham flux law method appears to be useful if tension measurements at two fixed locations are available. Use of the Darcy–Buckingham flux law method requires the soil water tension gradient measurement made close to soil surface and near the center of the infiltrometer disk, approximating one-dimensional water flow. Because the total steady-state infiltration flux from a tension infiltrometer disk (three-dimensional) is greater than one-dimensional flow (Smettem et al., 1994), a tension gradient at large distances from the disk may not be used in applying the Darcy–Buckingham flux law method for parameter estimation.

Using early-time infiltration data and soil water content increase, the sorptivity method provided K_s estimates ranging from 16.5 to 76.2 cm d^{-1} and α_G from 0.0200 to 0.0430 cm^{-1} (Table 1). The two parameters were found to be quite variable between different disk radii or tensions used for the infiltrometer measurements. The variation may be caused by the assumption used in the development of the sorptivity method, i.e., a single-term infiltration equation. A statistical mean

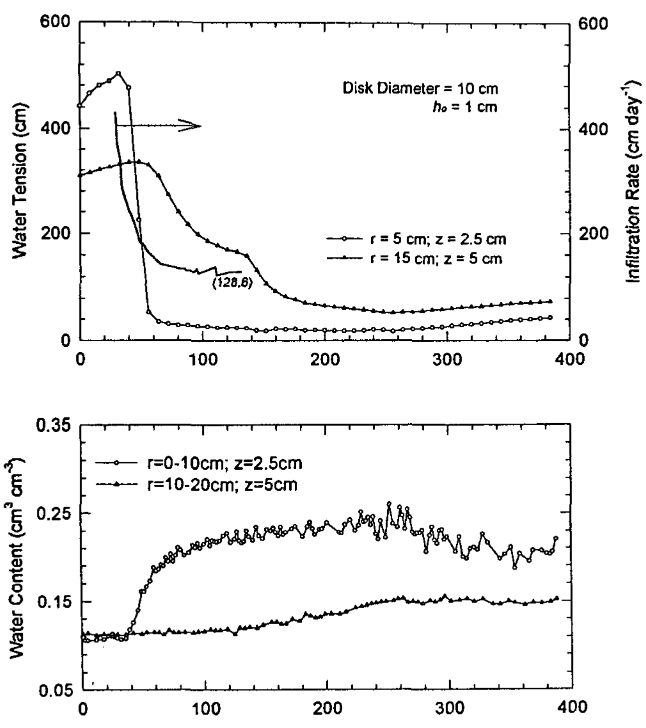

Fig. 4. Measured filtration flux with a 10-cm-diam. disk under 1-cm supply tension (h_o), soil water tension, and water content at fixed locations relative to the center of the infiltrometer disk; r = pond radius and z = depth from the soil surface.

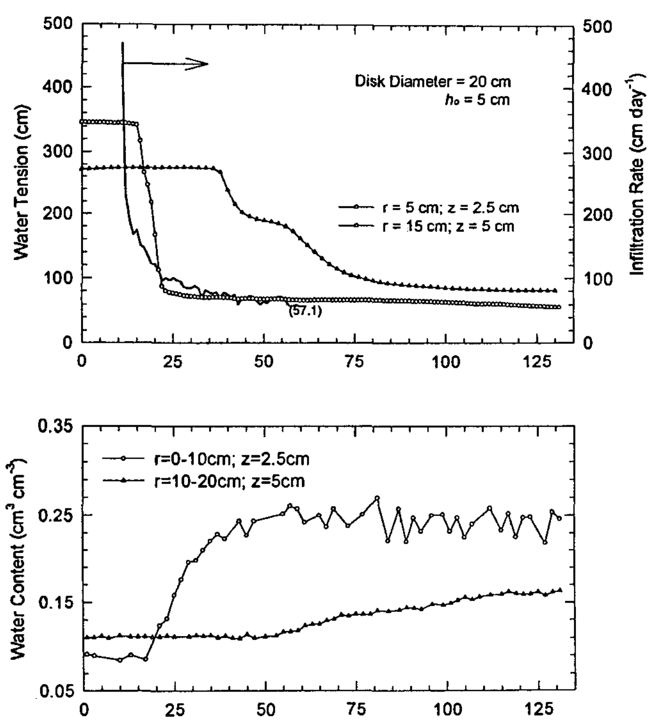

Fig. 5. Infiltration flux from a 20-cm-diam. disk under 5-cm supply tension (h_o), and measured soil water tension and water content at fixed locations relative to the center of the infiltrometer disk; r = pond radius and z = depth from the soil surface.

Table 1. Estimated soil hydraulic parameters with Wooding's steady-state approximate solution, an integral form of steady-state Darcy–Buckingham flux law, and a transient-state sorptivity method†.

Method	Supply tension	Disk diam.	K_s	α_G	S_o	D_e	R^2	$\theta_o - \theta_i$
	cm	cm	cm d^{-1}	cm^{-1}	cm d$^{-1/2}$	cm d$^{-1/2}$		cm^3 cm^{-3}
Wooding	1, 5	10	29.7	0.0700	–	–	–	–
	1, 5	20	35.2	0.0856	–	–	–	–
	1	10, 20	35.1	0.0851	–	–	–	–
	5	10, 20	22.3	0.0540	–	–	–	–
Mean ± SE			30.6 ± 3.04	0.0737 ± 0.0075				
Darcy-Buckingham	1, 5	20	34.2	0.1407	–	–	–	–
Sorptivity	1	10	67.8	0.0405	19.10	30.00	0.827	0.12
	1	20	16.5	0.0200	15.49	19.16	0.984	0.16
	5	10	52.2	0.0356	16.33	11.12	0.914	0.10
	5	20	76.2	0.0430	20.47	4.26×10^{-8}	0.943	0.13
Mean ± SE			53.2 ± 13.2	0.0348 ± 0.0052				

† K_s = hydraulic conductivity of saturated soil; α_G is a fitting parameter used in Gardner's hydraulic conductivity function; S_o = soil sorptivity under supply tension h_o; D_e = a fitting parameter; R^2 = coefficient of determination for S_o; θ_o and θ_i are the final and initial water contents, respectively, under an infiltrometer disk.

comparison (with a t-statistic for two population means with small sample sizes), however, indicated that the estimated K_s values were not significantly different (at the P = 0.05 level) from estimates using Wooding's method. Estimated α_G values were smaller than that from Wooding's method. The sorptivity method is generically different from Wooding's approach, using different parts of the experimental information, i.e., steady-state flow rate for Wooding's method vs. early-time transient infiltration rate and water content increase for the sorptivity approach. The required water content increase in the sorptivity method can be easily determined by taking soil samples prior to and right after the tension infiltration measurements rather than

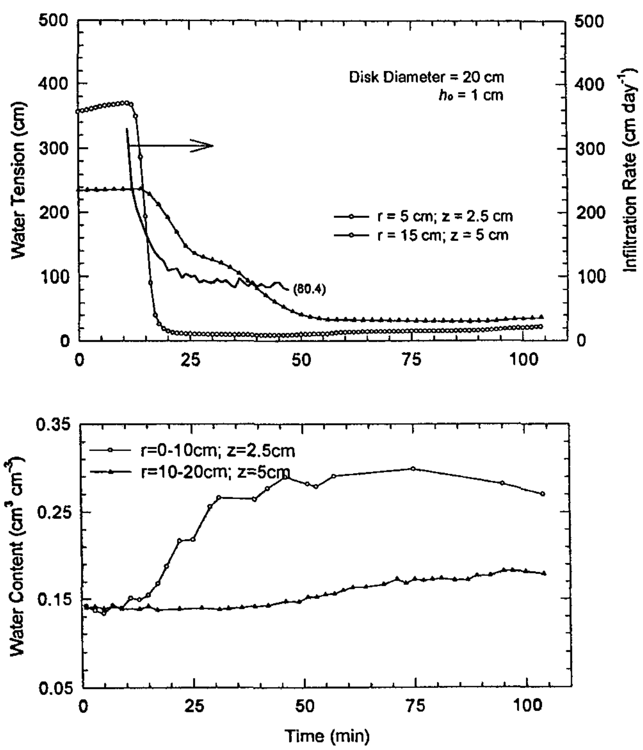

Fig. 6. Measured filtration flux with a 20-cm-diam. disk under 1-cm supply tension (h_o), and soil water tension and water content at fixed locations relative to the center of the infiltrometer disk; r = pond radius and z = depth from the soil surface.

taking continuous measurements with TDR. While Wooding's steady-state method appears to work reasonably well, the Darcy–Buckingham flux law and sorptivity procedures could produce similar results (comparable to their large spatial variability in the field, Wierenga et al., 1991), and may be used as alternatives to Wooding's approach. The sorptivity method may become more useful in fine-textured soils where steady-state infiltration is difficult to reach.

Using measured soil water content and tension pairs, the parameters needed in the water retention function Eq. [1] were determined using a nonlinear regression with the residual and saturated soil water content as fixed values determined from the soil core measurements. As shown in Fig. 7, fitted parameters are n = 1.3731 and α_{VG} = 0.3571 cm^{-1}. These two parameters can be converted to the single parameter α_G because we can equate Eq. [2] and [3] using the relative conductivity

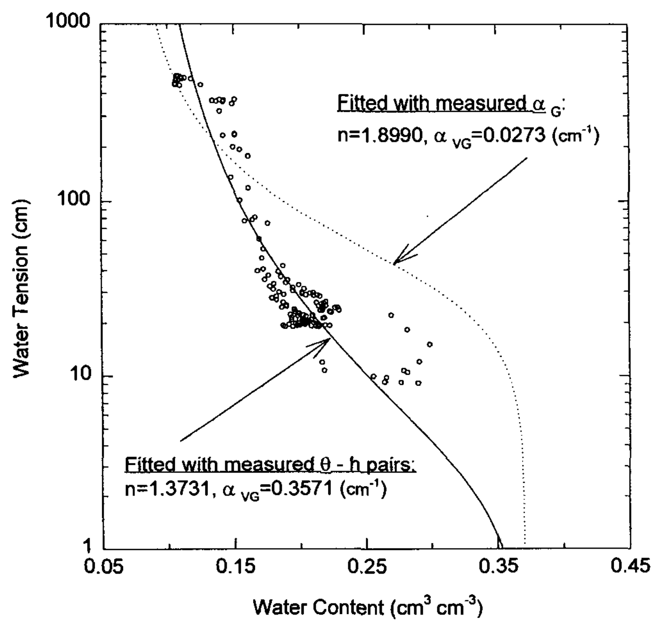

Fig. 7. Fitted water retention functions (α_{VG}) using van Genuchten (1980) and measured water content (θ) and tension (h) data (symbols), where residual soil water content (θ_r) = 0.077 cm^3 cm^{-3} and saturated water content (θ_s) = 0.371 cm^3 cm^{-3}.

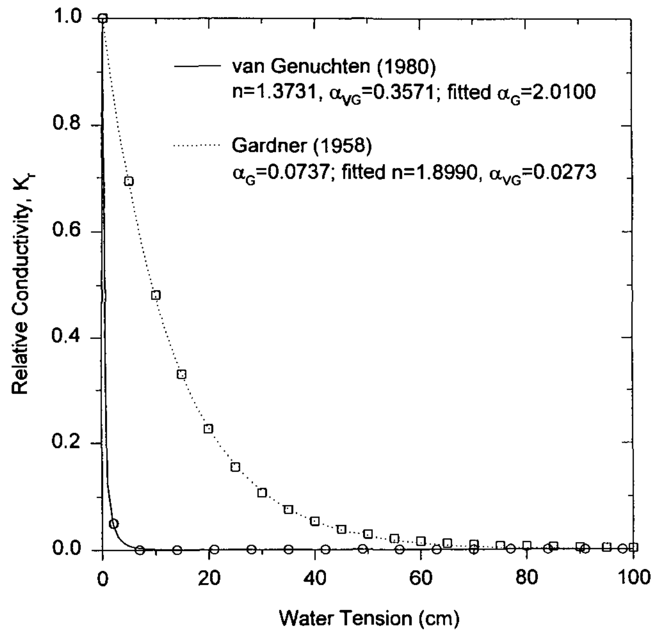

Fig. 8. Comparison of relative hydraulic conductivity functions by Gardner (1958) and van Genuchten (1980) using parameters estimated from infiltrometer measurements, where $K_r = K(h)/K_s$. Symbols are fitted to the predetermined conductivity functions to obtain comparative parameters used in either Gardner (1958) or van Genuchten (1980).

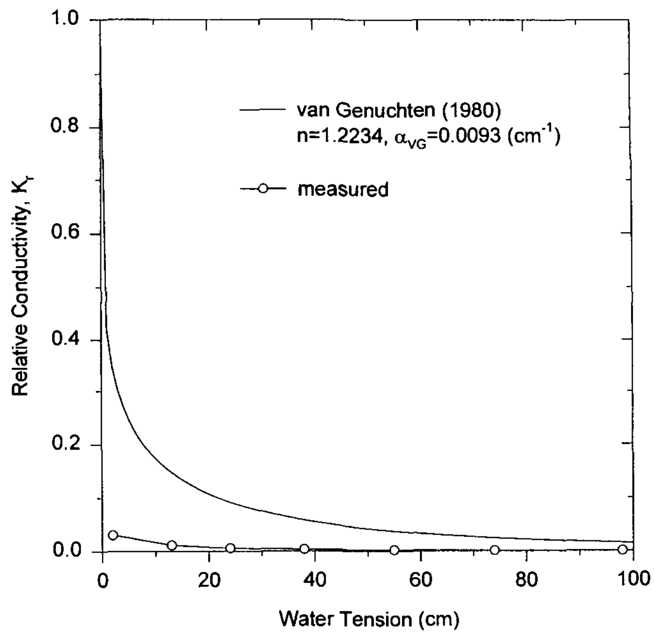

Fig. 9. Comparison of relative hydraulic conductivity between measured and predicted values using parameters fitted with van Genuchten (1980) to water retention data, where $K_r = K(h)/K_s$ and saturated hydraulic conductivity K_s = 69 cm d^{-1}.

$K_r = K(h)/K_s$ as a common factor. The value of α_G (equivalent to $n = 1.3731$ and $\alpha_{VG} = 0.3571$ cm^{-1}) is 2.0100 cm^{-1}. As shown in Fig. 8, it was obtained by fitting the $K_r(h)$ function from Eq. [2] with $n = 1.3731$ and $\alpha_{VG} = 0.3571$ cm^{-1} to the $K_r(h)$ function from Eq. [3] or $\exp(-\alpha_G h)$. Similarly, we fitted the $K_r(h)$ function of Eq. [3] with $\alpha_G = 0.0737$ cm^{-1} to Eq. [2] and found $n = 1.8990$ and $\alpha_{VG} = 0.0273$ cm^{-1}. This provided us a set of parameters (θ_r, θ_s, n, and α_{VG}) for describing the water retention curve based on Wooding's or the Darcy–Buckingham flux law method, independent of the direct soil water content and tension pair measurements (Fig. 7). The difference between the two fitted retention curves is large near saturation. This is caused by the large departure in predicted $K_r(h)$ near saturation (Fig. 8) using the two different methods of parameter estimation or measurements. The Wooding's method of using Gardner's exponential $K(h)$ model (Eq. [3]) estimated larger $K_r(h)$ values, for $h < 100$ cm, than fitting the measured soil water content and tension pairs using van Genuchten's water retention model (Eq. [2]). Using laboratory retention data, we fitted the retention curve with van Genuchten (1980) and obtained a separate estimate of $K(h)$ as a comparison with the field direct measurement (Fig. 9). The fitted $K(h)$ was overestimated compared with the measured values for $h < 100$ cm. The overestimation would be even larger when compared with Wooding's method of using Gardner's exponential $K(h)$ model. In theory, the water retention and hydraulic conductivity functions should be coherent for a given soil so that the parameters fitted to water retention, after conversion, should be able to describe the hydraulic conductivity function as well. However, this study indicates a disagreement or conflict between the two basic hydraulic functions. Since both models have been independently tested and used successfully in many studies, the incoherence may be attributed to the difference in methodology used to obtain parameters for the two functions. It is also possible that preferential flow through macropores might have invalidated the two models to a different degree, as indicated by Clothier and Smettem (1990). The discrepancy from this study may also be attributed to the timing of water content and tension pair measurements, which represented the imbibition part of the water retention curve. Intermittent drying between each infiltration event would create hysteresis that tends to fall within an envelope of drying and wetting cycles, adding to the scattering in the measured water content and tension pairs.

CONCLUSIONS

The use of small TDR probes and tensiometers during field tension infiltrometer experiments provided simultaneous measurements of soil water content, tension, and transient infiltration rate under preselected supply tensions. Based on these measurements, soil saturated hydraulic conductivity (K_s) and parameters (n, α_{VG}, and α_G) used in water retention and hydraulic conductivity functions were estimated using Wooding's approximate solution, an integral form of the steady-state Darcy–Buckingham flux law, a sorptivity method, and by fitting the retention curve with measured water content and tension pairs. The Darcy–Buckingham flux law method provided a K_s and α_G estimate similar to estimates made with Wooding's method. Using early-time transient infiltration rate, the sorptivity method also produced K_s estimates that were statistically (at the $P = 0.05$ level) similar to values obtained with Wooding's method. The estimated hydraulic conductivity function $K(h)$ using the infiltrometer methods (i.e., Wooding's, Darcy–Buckingham flux law, or the sorptivity method) overpredicted the unsaturated conductivity near saturation ($h <$ 100 cm), compared with predictions using parameters derived from fitted water retention function. Discrepancy in $K(h)$ using either the infiltrometer measurements or parameters converted from water retention data was attributed to the difference in the models (i.e., Eq. [2] vs. Eq. [3]), in the methodology used to obtain parameters used in the two models, and possibly to preferential flow through macropores.

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