Comparison of alternative methods for deriving hydraulic properties and scaling factors from single-disc tension infiltrometer measurements

Diederik Jacques[1]

Institute for Land and Water Management, Katholieke Universiteit Leuven, Leuven, Belgium

Binayak P. Mohanty[2]

U.S. Salinity Laboratory, USDA-ARS, Riverside, California, USA

J. Feyen

Institute for Land and Water Management, Katholieke Universiteit Leuven, Leuven, Belgium

Received 9 April 2001; revised 28 February 2002; accepted 7 March 2002; published 25 July 2002.

[1] Analysis of single-disc tension infiltrometer data is commonly based on the interpretation of the steady state infiltration rate derived from large time measurements. In this study, three analytical models describing the cumulative infiltration curve were used to estimate the sorptivity and the steady state infiltration rates. Hydraulic conductivity was estimated from (1) two steady state infiltration rates, (2) a sorptivity value and a steady state infiltration rate, and (3) two sorptivity estimates. In total, seven different methods were used to calculate the hydraulic conductivity. We compared the spatial variability of sorptivity, steady state infiltration rates, unsaturated hydraulic conductivity, and scaling factors of these seven methods. Cumulative infiltration curves were measured at 50 locations and three pressure heads along a 40 m long transect on a silty loam Eutric Regosol. In a first step the amount of water and the time needed to wet the contact sand under the disc was successfully filtered from the raw data using nonlinear regression techniques. The analytical models described the infiltration curves very well but gave different estimates of the sorptivities and steady state infiltration rates. Method 2 using both sorptivity and steady state infiltration rate resulted in the largest estimates of the mean unsaturated hydraulic conductivity, whereas the other two methods (1 and 3) resulted in similar mean values. Linear scaling analysis described well the observed variability in the hydraulic conductivity. Overall, different methods produced quite different unsaturated hydraulic conductivity estimates at specific locations, while their field mean or scaled properties were better comparable. Results of this study may prove to be important for deciding appropriate disc infiltrometer data analysis procedure while addressing water flow and chemical transport behavior at different spatial scales. *INDEX TERMS:* 1875 Hydrology: Unsaturated zone; 1894 Hydrology: Instruments and techniques; 1842 Hydrology: Irrigation; *KEYWORDS:* tension infiltrometers, scaling, hydraulic conductivity, field methods, near-saturated

1. Introduction

[2] Tension infiltrometers are frequently used to determine spatial [e.g., *Mohanty et al.*, 1994; *Vandervaere*, 1995; *Jarvis and Messing*, 1995; *Shouse and Mohanty*, 1998; *Zavattaro et al.*, 1999] and temporal [e.g., *Murphy et al.*, 1993; *Logsdon and Jaynes*, 1996; *Angulo-Jaramillo et al.*, 1997] variability of unsaturated hydraulic conductivity. More specifically, they are used to assess the impact of macropores on infiltration characteristics [e.g., *Thony et al.*, 1991; *Buttle and McDonald*, 2000], to analyze the effect of tillage or other management factors on surface conductivity [e.g., *Casanova et al.*, 2000], and to estimate solute transport parameters [e.g., *Clothier et al.*, 1992, 1995; *Jaynes et al.*, 1995; *Angulo-Jaramillo et al.*, 1996; *Casey et al.*, 1997, 1998]. An overview of the theory behind single-disc infiltrometers and their applications is given by *Angulo-Jaramillo et al.* [2000]. In general, to derive the unsaturated hydraulic conductivity, sorptivity and/or steady state rates are estimated from the measured cumulative infiltration curves [e.g., *White and Perroux*, 1989; *Ankeny et al.*, 1991; *Reynolds and Elrick*, 1991; *White et al.*, 1992].

[3] Steady state rates are derived from the linear regression of the measured cumulative infiltration rates at medium or large times. *Zhang* [1998] suggested that reaching steady state infiltration rates may require several hours to days resulting in an overestimation of the soil hydraulic conductivity for methods based on Wooding's

[1]Now at SCK-CEN, Mol, Belgium.

[2]Now at Biological and Agricultural Engineering Department, Texas A&M University, College Station, Texas, USA.

0043-1397/02/2001WR000595$09.00

Table 1. Overview of the Different Methods

Case	Data	Derivation S	Derivation Q_s	Conductivity[a]	Code
1	large time	–	linear regression	RE	Q1
2	large time	–	buried source model	RE	Q2
3	early and large time	Vandervaere	linear regression	W	SQ1
4	early and large time	Vandervaere	buried source model	W	SQ2
5	early and large time	Horton	Horton	W	SQ3
6	early time	Vandervaere	–	WP	S1
7	early time	Lin. dif. model	–	WP	S2

[a] RE, *Reynolds and Elrick* [1991]; W, *White et al.* [1992]; WP, *White and Perroux* [1989].

equation [*Wooding*, 1968] such as *Ankeny et al.* [1991] and *Reynolds and Elrick* [1991]. However, infiltration rates may fluctuate during long-lasting infiltration experiments since the expanding wetting front may reach different soil layers or structural elements in heterogeneous soils [*Logsdon*, 1997]. On the other hand, based on the early time infiltration equation of *Philip* [1969], the sorptivity is derived from the slope of the cumulative infiltration per unit area versus the square root of time [*White et al.*, 1992]. Due to the contact layer between the soil surface and the disc infiltrometer, this method may overestimate the sorptivity [*Minasny and McBratney*, 2000] resulting in incorrect hydraulic conductivity values. Models describing the complete infiltration curve offer an alternative way to obtain estimates of the sorptivity and/or the steady state infiltration rate. These models may be empirical [e.g., *Warrick et al.*, 1992; *Zhang*, 1997], simplified (e.g., neglecting gravity [*Warrick*, 1992]) or physically-based [*Haverkamp et al.*, 1994] in nature. *Hussen and Warrick* [1993] identified some of the most appropriate models for determining sorptivity and steady state infiltration rates.

[4] In this study, tension infiltrometers were used to characterize the spatial variability of soil hydraulic properties. The spatial variability is described by scaling factors of the pressure head and the hydraulic conductivity relating hydraulic properties at a specific location to a reference curve using simple linear relations [*Vogel et al.*, 1991]. Since different infiltration models are based on different assumptions, it is reasonable to expect differences in estimated conductivity values and, consequently, in the statistical and geostatistical properties of the scaling factors.

[5] Traditional interpretations of tension infiltrometer data relied mostly on steady state infiltration rates and Wooding's equation [*Wooding*, 1968], whereas more information present in the cumulative infiltration curve can be used. Recently, *Vandervaere et al.* [2000b] defined four new methods to analyze transient flow data from disk infiltrometers based on the infiltration equation of *Haverkamp et al.* [1994]. Their methods based on measurements with one disk radius and one pressure head (single test method) will be tested in this paper by using alternative infiltration equations. Their conclusions were based on repacked materials and laboratory measured hydraulic properties. Complementary to their methods, an additional method is proposed here in which transient infiltration data is measured with one disc radius and several pressure heads at the same location. The objective of this study is to compare the statistics of sorptivities, steady state infiltration rates, unsaturated hydraulic conductivities and scaling factors derived from the cumulative infiltration curve measured with a single-disc tension infiltrometer using different analytical infiltration models. In this paper, three information levels are used: (1) medium or large time measurements resulting in the steady state infiltration rate, (2) both early and medium or large time measurements resulting in sorptivity and steady state infiltration rate, and (3) early time measurement resulting in sorptivity.

2. Theory

[6] The determination of the near-saturated hydraulic conductivity is based on information in the cumulative infiltration curves from single-disc tension infiltrometers. Three methods are used in this study to estimate the hydraulic conductivity. The first method is based on the steady state infiltration rate obtained from medium or large time. The approximate solution of *Wooding* [1968] is used to relate the steady state infiltration rates to the hydraulic conductivity. Hydraulic conductivities are calculated from the steady state infiltration rate at two pressure heads using the approach of *Reynolds and Elrick* [1991]. In the second method, the sorptivity and the steady state infiltration rate, estimated from a single cumulative infiltration curve, are combined to determine the conductivity with the method of *White et al.* [1992]. Thus both early time and medium/large time data are used. The third method estimates the conductivity from two measured sorptivities for early time infiltration [*White and Perroux*, 1989]. As per the recommendation of *Hussen and Warrick* [1993] we used the linear diffusion model and the modified Horton model for estimating the sorptivity. *Hussen and Warrick* [1993] also stated that the buried source model and the modified Horton model give reliable estimates of the steady state infiltration rates. In addition, the approach of *Vandervaere et al.* [1997] to derive the sorptivity and the linear regression technique for deriving the steady state infiltration rate are used in this study. An overview of the different approaches used in this study is given in Table 1.

2.1. Sorptivity Determination

2.1.1. Standard method

[7] Early time infiltration from a tension infiltrometer can be described by [*Philip*, 1957]:

$$i = St^{0.5} \tag{1}$$

where i [L] is the cumulative infiltration depth of water into the soil under a constant pressure head at the soil surface, t [T] is time, and S [L T$^{-0.5}$] is the sorptivity at the infiltration surface. The sorptivity is determined from the slope of linear regression of i as a function of $t^{0.5}$ for small t. Some problems are associated with this standard method: (1) equation (1) is only valid when the infiltration process is one-dimensional and flow is capillarity-driven, i.e., when t is smaller than the geometry time. The geometry time is defined as the time when the system overrides the initial one-dimensional character [*Philip*, 1969; *White and Sully*, 1987]. However, in reality, equation (1) may be applicable to times much smaller than the geometry time (e.g., less than 100 s) resulting in only a few data points to estimate S [*Philip*, 1969; *Warrick*, 1992; *Cook and Broeren*, 1994]. (2) The contact layer of sand between the tension disc and the soil surface has an effect on the infiltrated amount of water and time [*Quadri*, 1993; *Vandervaere*, 1995] and thus on the estimation of S. (3) The choice of the measurement points used for sorptivity estimation is subjective [*Vandervaere et al.*, 2000b]. In this paper, we will estimate the amount of water and the time needed to wet the contact layer using the approach of *Vandervaere et al.* [1997, 2000a] to correct the measured infiltration curves for errors due to the contact layer. In this way, analytical models can be used to estimate S using all available measurements in an objective manner.

2.1.2. *Vandervaere et al.* [1997, 2000a] (VAN)

[8] The initial infiltration from the tension single disc infiltrometer is not only determined by the soil properties but also by the small contact layer of sand placed between the disc surface and the soil surface. Therefore the physically-based expression of *Haverkamp et al.* [1994] for the three-dimensional cumulative infiltration depth without contact layer, written here in a simplified form:

$$i = S\sqrt{t} + (A + B)t \tag{2}$$

where A [L T^{-1}] and B [L T^{-1}] are constants, was adapted by *Vandervaere et al.* [1997] to describe the three-dimensional infiltration depth versus time in the soil with a contact layer, as:

$$i' = i_0 + S\sqrt{t' - t_0} + (A + B)(t' - t_0) \tag{3}$$

where i_0 and t_0 are the depth of water and the time needed to wet the contact layer, respectively, and i' and t' are the actual measured infiltration (with contact layer) and experimental time, respectively. In the remainder of the paper, i and t denote the infiltration rate versus time for infiltration in the soil corrected for the initial infiltration in the contact layer. The derivative of equation (3) with respect to $t^{0.5}$,

$$p = \frac{\partial i'}{\partial\sqrt{t'}} = S\sqrt{\frac{t'}{t' - t_0}} + 2(A + B)\sqrt{t'} \tag{4}$$

is not influenced by i_0 and the effect of t_0 decreases sharply as time increases. Consequently, when the first (i.e., early time) points of p (where p decreases with $t^{0.5}$) are neglected, the parameters S and $(A + B)$ are estimated by simple linear regression [*Vandervaere*, 1995]. To estimate i_0 and t_0, equation (3) is then fitted to the observed cumulative infiltration curves with fixed values of S and $(A + B)$. Finally, i_0 and t_0 are subtracted from i' and t' to obtain the corrected infiltration curve of i versus t [see also *Jacques et al.*, 1999]. The corrected observations are then used to estimate the parameters of the modified Horton model, the linear diffusion model, and the buried source model. A numerical study in conjuction with a field experiment of *Minasny and McBratney* [2000] indicated that this approach resulted in the most appropriate estimates of the sorptivitiy.

2.1.3. Modified "Horton" Model (MHM)

[9] *Hussen and Warrick* [1993] showed that the empirical relation:

$$i = St^{0.5} + \frac{Q_s}{\pi r^2}\left[t + \frac{a}{c}(1 - \exp(-ct))\right] \tag{5}$$

with r the radius (L), and a and c empirical constants, gives the best simultaneous estimates of S (sorptivity) and Q_s (the steady state infiltration rate, L^3 T^{-1}) compared to other methods discussed in their paper. Due to the similarity of equation (5) and the empirical one-dimensional infiltration equation of *Horton* [1940], this model is referred to as the modified "Horton" model. In this form, four parameters (a, c, S, Q_s) are estimated by nonlinear optimization.

2.1.4. Linear Diffusion Model (LDM)

[10] An approximation for the cumulative infiltration for linear diffusion from a disc source in soils with a constant diffusivity (D [L^2 T^{-1}]) was given by *Warrick* [1992] and *Warrick et al.* [1992] [see also *Hussen and Warrick*, 1993]. If the constant diffusivity is expressed as [*Philip*, 1986]:

$$D = \frac{\pi S^2}{4\left(\theta_{wet} - \theta_{dry}\right)^2} \tag{6}$$

the cumulative infiltration is described as:

$$i = 1.27\left(\theta_{wet} - \theta_{dry}\right)r\left[T + 0.5\pi^{0.5}T^{0.5} - 0.054[1 - \exp(-4.01t)]\right] \tag{7}$$

with θ_{wet} and θ_{dry} the water content corresponding to the applied and initial tension, respectively, and $T = Dt/r^2$ where r [L] is the radius of the disc. In this model, D is the only unknown. The application of this model is limited to small times and/or small disc sizes [*Hussen and Warrick*, 1993]. Numerical simulations of *Warrick* [1992] showed that the linear diffusion model tends to be reasonably good for time as large as three hours.

2.2. Steady State Infiltration Rate Determination

2.2.1. Linear Regression Method (SRM)

[11] The value of Q_s is obtained by linear regression of the measured cumulative infiltration curve at large times. It is assumed that steady state infiltration rates were obtained within the experimental time. Although in many studies, steady state infiltration rates are apparently reached, some studies [e.g., *Zhang*, 1998] have indicated that it may take several hours to days before they are actually reached.

2.2.2. Modified "Horton" Model

[12] As indicated in equation (5), Q_s is also obtained by nonlinear regression.

2.2.3. Buried Source Model (BSM)

[13] *Pullan* [1992] found an approximate solution for the time-dependent infiltration rate with constant diffusivity and an exponential hydraulic conductivity curve from a disc source. The cumulative infiltration for his solution is:

$$i = \frac{Q_s t}{\pi r^2} + Bf(Ct) \tag{8}$$

where

$$B = 2\lambda_c^2 K_0 / D \tag{9}$$

$$C = D / \left(4\lambda_c^2\right) \tag{10}$$

$$f(u) = 0.5 - (u + 0.5)erfc\left(u^{0.5}\right) + (u/\pi)^{0.5}\exp(-u) \tag{11}$$

where *erfc* is a complementary error function, D the constant diffusivity, K_0 the hydraulic conductivity at the applied pressure head [L T^{-1}], and λ_c macroscopic capillary length [L]. In this paper, the parameters B, C and Q_s are estimated by minimizing the sum of squared errors between the measurements and the model predictions. As discussed by *Hussen and Warrick* [1993], this model gives accurate estimates for Q_s.

2.3. Hydraulic Conductivity Determination

2.3.1. Based on two measurements of Q_s

[14] *Reynolds and Elrick* [1991] used the approximation of *Wooding* [1968] for steady state infiltration rate from a disc source of radius r:

$$Q_s = \pi r^2 \left[K + \frac{4}{\pi r}\phi\right] \tag{12}$$

where

$$\phi = \int_{\psi_i}^{\psi_0} K(\psi) d\psi \qquad \psi_0 \le 0 \tag{13}$$

and ψ_i and ψ_0 the initial and applied pressure head [L], for an exponential hydraulic conductivity relationship [*Gardner*, 1958] ($\psi \le 0$):

$$K(\psi) = K_s \exp(\alpha\psi) \tag{14}$$

where K_s is the saturated hydraulic conductivity [L T^{-1}] and α [L^{-1}] the inverse of the macroscopic capillarity length scale λ_c [*Philip*, 1985]. *Reynolds and Elrick* [1991] assumed that the unsaturated hydraulic conductivity between ψ_1 and ψ_2 ($\psi_1 < \psi_2$) is described with an exponential model (equation 14) with two parameters $K_{1,2}^{p}$ (for K_s) and $\alpha_{1,2}^{p}$ (for α) valid between ψ_1 and ψ_2. Assuming that the soil is initially sufficiently dry ($K(\psi_i) \ll K(\psi_1) < K(\psi_2)$), the parameters $\alpha_{1,2}^{p}$ and $K_{1,2}^{p}$ (i.e., the piecewise exponential parameters between ψ_1 and ψ_2) are easily obtained from the measured q_s (= $Q_s/\pi r^2$) measurements at ψ_1 and ψ_2 (see *Reynolds and Elrick* [1991] for details). In this study, the unsaturated hydraulic conductivity at the highest applied pressure head is derived from the steady infiltration rates measured at the highest and middle applied pressure head, whereas the unsaturated hydraulic conductivities at the middle and lowest applied pressure head use the infiltration rates at the middle and lowest applied pressure head. In this study, two variants of this approach to derive the hydraulic conductivity are tested (Table 1). Q1 is the more traditional interpretation of single-disc tension infiltrometer measurements using only steady state flow conditions. Q2 is based on transient flow measurements. Compared with the methods defined by *Vandervaere et al.* [2000b], Q2 forms an additional method to interpret disc infiltrometer data.

2.3.2. Based on one Q_s and one S measurement

[15] *White and Sully* [1987] showed that λ_c [*Philip*, 1985] is related to the sorptivity and the conductivity as:

$$\lambda_c = \frac{bS^2}{(\theta_0 - \theta_i)(K_0 - K_i)} \tag{15}$$

where b is function of the shape of the soil-water diffusivity. The limits of b are between 0.5 and $\pi/4$ with 0.55 as a typical average . If it is assumed that $K_i \ll K_0$, then $\Delta K \approx K_0$ and equation (12) can be written as:

$$\frac{Q_s}{\pi r^2} = K_0 + \frac{4bS^2}{(\theta_0 - \theta_i)\pi r} \tag{16}$$

In contrast to the methods of *Reynolds and Elrick* [1991] and *White and Perroux* [1989], the hydraulic conductivity is obtained from the measurement of the cumulative infiltration at one applied pressure head since it uses information at both small and intermediate/large infiltration times. Comparing the applied methods based on *White et al.* [1992] (Table 1) with the seven methods defined by *Vandervaere et al.* [2000b], SQ1 is similar to their WS (White-Sully method: one disc, one pressure head, steady state flow). The other two variants (SQ2 and SQ3) are alternatives to the single test method (ST method) of *Vandervaere et al.* [2000b].

2.3.3. Based on two S measurements

[16] Combining the definition of the macroscopic capillary length [*Philip*, 1985]:

$$\lambda_c = \frac{1}{K(\psi_0) - K(\psi_i)} \int_{\psi_i}^{\psi 0} K(\psi) d\psi \tag{17}$$

with equation (15) gives:

$$\int_{\psi_i}^{\psi_0} K(\psi) d\psi = bS^2 / \Delta\theta \tag{18}$$

White and Perroux [1989] differentiated equation (18) with respect to the applied pressure head. To derive K from S, and measured θ at two pressure heads, *White and Perroux* [1989] derived the following finite difference equation:

$$K(\bar{\psi}) = \frac{\bar{S}\Delta S}{\Delta\bar{\theta}\Delta\psi} - \left(\frac{\bar{S}}{2\Delta\theta}\right)^2 \frac{\Delta\theta}{\Delta\psi} \tag{19}$$

where $\bar{\psi} = (\psi_1 + \psi_2)/2$, $\bar{S}$; $= (S(\psi_1) + S(\psi_2))/2$, $\Delta\theta = (\bar{\theta}(\psi_1) + \theta(\psi_2))/2 - \theta_i$, $\Delta S = S(\psi_1) - S(\psi_2)$, and $\Delta\psi = \psi_1 - \psi_2$. *Vandervaere et al.* [2000b] proposed this approach to derive K from infiltration measurements under multiple pressure heads for either one disc radius at different locations or several disc radii at different locations. In this study, the *White and Perroux* [1989] approach is applied for one disc radius and one location (S1 and S2 in Table 1).

2.4. Accuracy of Hydraulic Conductivity Estimation

[17] In a recent paper, *Vandervaere et al.* [2000b] proposed criteria to indicate conditions to accurately estimate the sorptivity and the hydraulic conductivity using single-disc tension infiltrometers. To obtain accurate values of the sorptivity, the vertical capillary term of equation (2) ($St^{0.5}$) should be large compared to the gravity and the lateral capillary term of equation (2) $(A + B)t$ at the beginning of the infiltration experiment. The ratio of S to $(A + B)$ is maximal when the sorptivity is equal to:

$$S_{opt} = \sqrt{\frac{r\Delta\theta}{\gamma}\frac{2-\beta}{3}K} \tag{20}$$

with β a constant between 0 and 1, and γ a constant equal to 0.75 [*Haverkamp et al.*, 1994]. If the actual sorptivity of the soil, S, equals S_{opt}, both S and K can be accurately estimated. If S is smaller than S_{opt} (i.e., soils in the so-called gravity domain), estimates of K are accurate whereas the estimation of S is less precise. Estimates of S are very accurate if S is larger than S_{opt} (i.e., soils in the so-called lateral capillarity domain), whereas K estimates are then sensitive to measurement errors. Using the hydraulic properties of seven different soils, *Vandervaere et al.* [2000b] showed that the actual sorptivity was always larger than S_{opt} and that methods based on multiple Q_{s}-measurements (e.g., the method of *Elrick and Reynolds* [1991]) or on mulitple S-measurements (e.g., the method of *White and Perroux* [1987]) were preferable. However, their conclusion was based on repacked materials and hydraulic properties measured in the laboratory. Therefore they stated that the values of S and S_{opt} still has to be investigated for field soils.

3. Materials and Methods

3.1. Experimental Methods

[18] In-situ infiltration measurements were made along a transect on an experimental field located at Bekkevoort, Belgium. The soil is defined as a typic Udifluvent [*Soil Survey Staff*, 1994]. The soil profile exhibits three horizons: Ap (0–25 cm) and C1 (25–55 cm) horizons both developed in colluvial material, and a C2 (55–100 cm) which is a textural B layer. More information about the soil profile is given by *Mallants et al.* [1997a] and *Jacques et al.* [2002]. Within two rows of trees, infiltration measurements with 0.08 m diameter single-disc infiltrometers were made on the soil surface at 50 locations along a 40 m long transect. Infiltration measurements were made at 1-m spacings for the first five and the remaining measurements at 0.8-m spacings. At each location the grass was removed and the soil surface was leveled. A thin sand contact layer ensured good contact between the infiltrometer and the soil surface. Infiltration measurements were done at three different pressure heads in a decreasing sequence. The applied negative pressure heads were different for different locations and ranged between 0 and −10 cm. Cumulative infiltration was manually recorded during the 90 min infiltration experiment. A one-day drainage period was allowed between the first and second or second and third applied pressure head. In this way, it was possible to measure three complete infiltration curves at each location.

3.2. Spatial Variability: Scaling Analysis

[19] Spatial variability of soil hydraulic properties are described in various ways. Using scaling analysis the spatial variability of the hydraulic conductivity is described by a set of scaling factors relating the location-specific soil hydraulic properties to a reference set of soil properties as [*Vogel et al.*, 1991]:

$$\begin{aligned} K(\boldsymbol{x},\psi) &= \alpha_K(\boldsymbol{x})K^*(\psi^*) \\ \psi &= \alpha_\psi(\boldsymbol{x})\psi^* \end{aligned} \tag{21}$$

where ψ, and K are the pressure head [L], and the hydraulic conductivity [L T^{-1}], respectively; $\boldsymbol{x}$ is the vector with the spatial coordinates; $K^*(\psi^*)$ is the reference hydraulic conductivity curve; and α_{K}, and α_ψ are the scaling factors for hydraulic conductivity, and the pressure head, respectively. In this approach, no a-priori correlation between the two different scaling factors is assumed. The reference hydraulic conductivity curve is based on the *van Genuchten* [1980] formulation:

$$K^*(\psi^*) = K_S^*\frac{\left[1+|\beta^*\psi^*|^{n^*-1}\left(1+|\beta^*\psi^*|^{n^*}\right)^{-m^*}\right]^2}{\left(1+|\beta^*\psi^*|^{n^*}\right)^{l^*m^*}} \tag{22}$$

where K_{s}^* is the saturated hydraulic conductivity [L T^{-1}], β^* [L^{-1}] and n^* are fitting parameters, $m^* = 1 - 1/n^*$, and l^* is the pore-connectivity and tortuosity factor that is fixed at 0.5 in this study. The scaling factors (α_{K}, and α_ψ for each location) and the parameters of the reference curve (K_{s}^*, β^*, and n^*) are determined by minimizing the sum of squared errors between observed hydraulic conductivity and predicted hydraulic conductivity by equation (21). In our study the derivatives of this sum of squared errors with respect to all scaling factors and the parameters of the reference curve were set equal to zero. As a normalization condition, the average values of the scaling factors were fixed at unity [*Warrick et al.*, 1977; *Clausnitzer et al.*, 1992]. To solve the system of nonlinear equations, the Newton-Raphson iteration algorithm described by *Carnahan et al.* [1969] was used. All calculations were done using *MathSoft* [1995].

3.3. Spatial Variability: Statistical Methods

[20] Box plots provide a simple graphical tool to show the distributional properties of a sample [*Rice*, 1995]. The middle horizontal line in the box plot represents the median (see Figure 5). The bottom and top lines of the box are the 25 and 75 percentiles or the lower and upper quartiles. The narrowing in the box show an approximate 95% confidence interval of the median [*Velleman and Hoaglin*, 1981]:

$$Median \pm 1.58\left(\frac{IQR}{\sqrt{N}}\right) \tag{23}$$

where IQR is the interquartile range or the difference between the 75 and 25 percentile and N is the number of observations. The top and bottom vertical lines define the extreme data points within a distance of 1.5 of the upper and

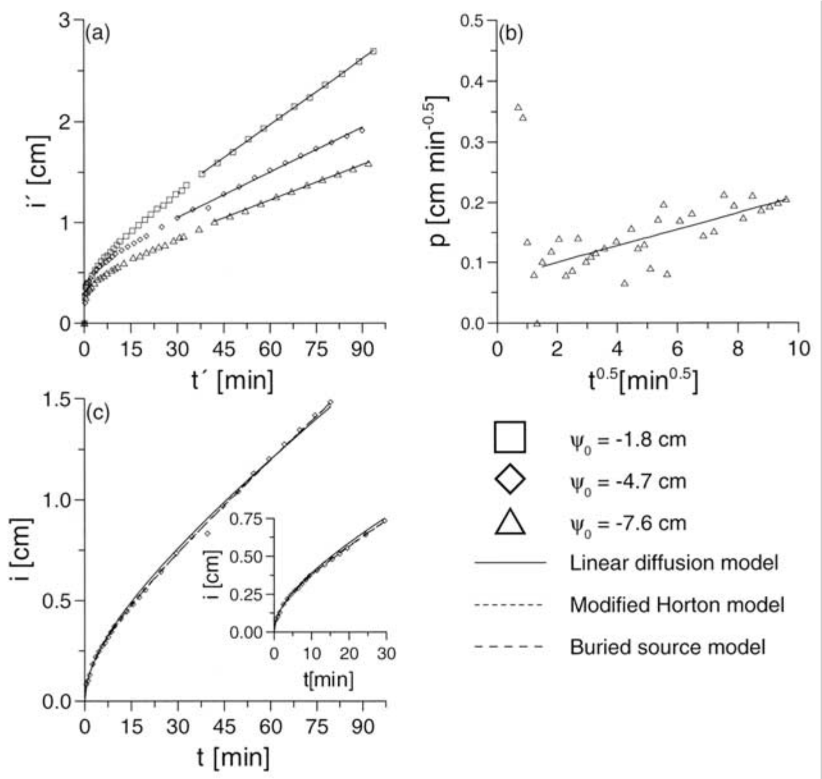


Figure 1. (a) Measured cumulative infiltration versus time at three applied pressure heads at an arbitrarily location along the transect. The solid lines are fitted by linear regression lines using the measurement points at medium or large infiltration time. (b) p versus $t^{0.5}$. The solid line is the linear regression line neglecting the first five data points. (c) Cumulative infiltration versus time corrected for the initial infiltration in the contact layer together with fitted linear diffusion model, modified Horton model, and the buried source model.

lower quartiles, respectively. Data points larger or smaller than the upper or lower adjacent value, respectivley, are indicated with a dot.

4. Results and Discussion

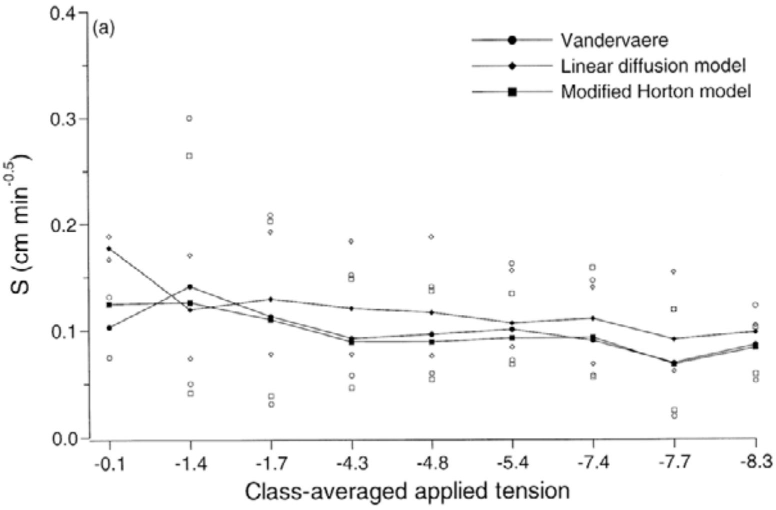
4.1. Sorptivity, Steady State Infiltration, and Hydraulic Conductivity

[21] The estimation of the initial infiltration in the contact layer and the fitting of the linear diffusion model, the modified Horton model and the buried source model are illustrated in Figure 1 for the three applied pressure heads at an arbitrarily chosen location along the transect. Figure 1a shows the original measured cumulative infiltration curves i' and the points used to estimate the steady state infiltration rates with the linear regression technique. Figure 1b illustrates the procedure proposed by *Vandervaere et al.* [1997] to estimate S and (A + B) by simple linear regression. The intercept equals S and the slope is related to (A + B). As mentioned above, p is not influenced by i_0 and the effect of

Table 2. Coefficients of Variation (CV) for S and Q_s for Each Class[a]

				S, cm min$^{-0.5}$			Q_s, cm^3 min^{-1}		
Class	Class Interval, cm	$\langle\psi\rangle$[b], cm	N	VAN	MHM	LDM	LRM	MHM	BSM
1	0 −2	−0.05	2	38.9	144.2	16.2	17.1	37.4	34.3
2	−2 −1.5	−1.35	20	40.8	35.2	27.4	39.9	25.9	27.1
3	−1.5 −2	−1.7	27	43.2	44.8	24.1	24.4	28	26.5
4	−2 −3	−2.6	1	–	–	–	–	–	–
5	−3.9 −4.5	−4.23	12	28.8	36.7	28.7	25.8	25.7	25.9
6	−4.5 −5	−4.79	29	26.7	27.1	24.1	18	20.2	23.1
7	−5 −6.1	−5.38	9	31	25.5	21.1	13.7	27.5	18.4
8	−7 −7.5	−7.42	9	36.2	40.6	20.8	4.8	30.8	25
9	−7.6 −8	−7.72	36	33	30	23.3	19.7	23	22.2
10	−8.1 −9	−8.28	5	35.3	23.7	7.9	12.5	18.1	18.9

[a] VAN, method of Vandervaere; MHM, modified Horton model; LDM, linear diffusion mode; LRM, linear regression method; BSM, buried source model.
[b] Arithmetic average of pressure heads in a given class.



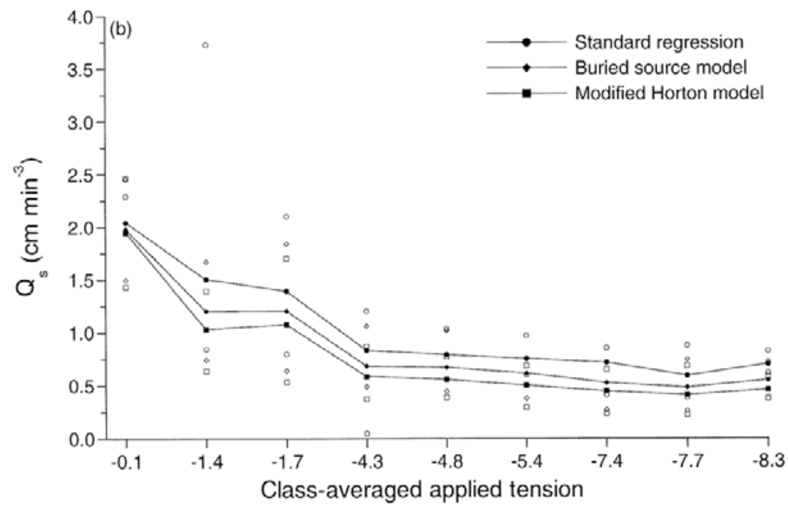


Figure 2. Class-averaged sorptivities (solid symbols) for the method of Vandervaere, the linear diffusion model, and the modified Horton model. (b) Class-averaged steady state infiltration rates for the linear regression, the buried source model, and the modified Horton model. Open symbols are the maximum and minimum value measured in a given class.

t_0 vanished rapidly when t increases. In Figure 1b, a plot of $t^{0.5}$ versus p for a single cumulative infiltration curve is shown. Although there were some scatter, the first two points are obviously larger than the others for $t^{0.5}$ smaller than 2 min$^{0.5}$. For this particular case, the fifth point had a value of zero. Therefore the first five points are excluded from the estimation based on a simple linear regression. Once S and $(A + B)$ are known, i_0 and t_0 are estimated by minimizing the sum of squared errors between the measured i' and the predictions from equation (3). The corrected infiltration curves ($i = i' - i_0$ versus $t = t' - t_0$) are then used to estimate the parameters of the linear diffusion model, the modified Horton model and the buried source model by minimizing the sum of squared errors between the corrected infiltration curve and the model predictions using the SAS NLIN-procedure [*SAS Institute*, 1989]. Results for one particular infiltration curve are shown in Figure 1c. Model fits were very good for the different models. For this particular case, no apparent difference between the modified Horton model and the buried source model was observed. The averages of the sum of squared errors (SSE) for all infiltration curves are 0.0622, 0.00367, and 0.0044 for the linear diffusion, the modified Horton, and the buried source model, respectively. The decrease of average SSE is directly related to the number of unknown parameters in the three different models.

[22] Resulting estimates of the sorptivity, the steady state infiltration rate, and the hydraulic conductivity are summarized in ten pressure head classes (Table 2). The statistics of the first, the fourth and the tenth class are not representative due to the small amount of observations (Table 2) and will be ignored in futher discussion. Class-averaged sorptivities and steady state infiltration rates are plotted in Figure 2. Note that the x-axis is a class-axis. Three estimates of the sorptivity (Figure 2a) and the steady state infiltration rate (Figure 2b) are plotted. The sorptivities estimated with the linear diffusion model were somewhat larger than those estimated with the other two approaches. Furthermore,

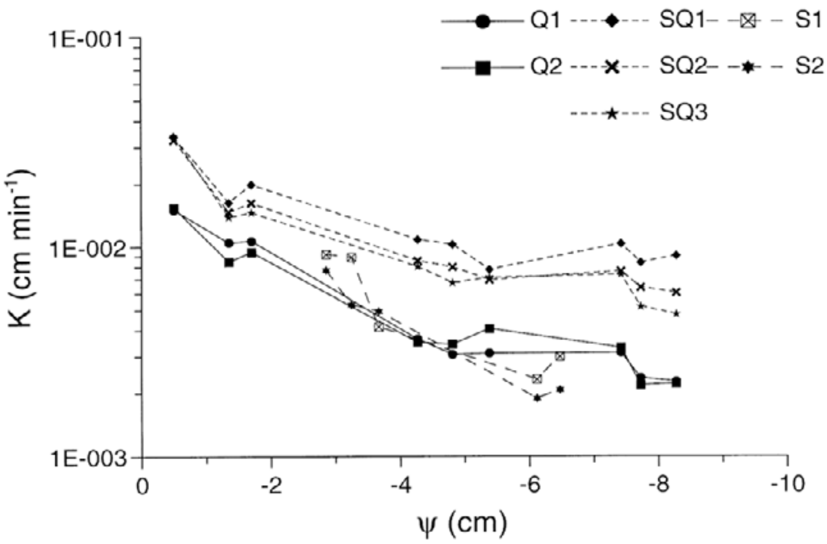


Figure 3. Class-averaged hydraulic conductivities obtained with the seven methods.

almost no decrease of the sorptivities with decreasing pressure head was observed for the linear diffusion model. Within a given class, the S estimates of the linear diffusion model were not correlated with the two other estimates at the 5% significancy level. The steady state infiltration rates estimated with the linear regression method were systematically higher than those of the buried source model. The empirical modified Horton model gives the lowest estimates of Q_s for all pressure heads. At a significancy level of 5%, the mean Q_s of the linear regression approach was significantly higher than the mean of the modified Horton model for each class. Within each class, correlation coefficients between the three estimates of Q_s are significant and close to 1 indicating that the three approaches produced similar trends. Although Figure 1a might indicate that the steady state infiltration rate was reached within the experimental time and, consequently, the linear regression technique may result in good estimates, but in reality, the model-based approaches resulted in lower estimates of Q_s than the linear regression based technique. This corresponds with the experimental and numerical results of *Hussen and Warrick* [1993]. Also, *Zhang* [1998] showed that hydraulic conductivities are significantly overestimated when steady state rates were overestimated and the method of *Ankeny et al.* [1991] was used. Similar overpredicitons are expected when the method of *Reynolds and Elrick* [1991] is used. The combination of sorptivity and overestimated steady state infiltration rates will also result in overprediction of the unsaturated hydraulic conductivity (equation (16)).

[23] In Table 2, the CV of S and Q_s are given for each class. No real trend in coefficient of variations, CV, (of S and Q_s) was observed with the pressure head. The CV of S for the linear diffusion model is smaller than the CV of the other two approaches. At the higher pressure heads, the CV of Q_s for the three approaches are quite similar. At the lower pressure heads and near saturation, however, the CV of Q_s for the SRM is somewhat smaller than the other two methods.

[24] As shown in Table 1, seven different methods were used to derive the unsaturated hydraulic conductivity. These cases are grouped into: (1) Q group using only Q_s estimates, (2) SQ group combining estimates of S and Q_s, and (3) S group based on S estimates only. A total of 150, 129, 145, 133, and 129 estimates of K were available for Q1, Q2, SQ1, SQ2, and SQ3, respectively. Missing values were due to a negative K estimate or to nonavailable S or Q_s estimates. For the S group, 67 and 92 K estimates of the 100 possible estimates (two replicates for each location) were obtained for S1 and S2, respectively. Class-averaged conductivities are shown in Figure 3 and statistics are summarized in Tables 3a and 3b. The SQ group gave higher estimates of K compared to the S and Q groups. The trend which was found in the Q_s estimates (based on SRM > BSM > MHM) is preserved in the class-averaged K estimates (SQ1 > SQ2 > SQ3). Mean estimates of K based on S and Q groups are close to each other. Although it is somewhat difficult to observe due to the small number of data points, the slope of the conductivity curve is somewhat steeper near saturation indicating gravity-dominated flow. This was also observed by others [e.g., *Jarvis and Messing*, 1995; *Mohanty et al.*, 1997] at other soil/hydrologic scenarios. The CV of the K estimates for the different methods were comparable. Only for the S group, CVs of S1 were consistently larger than those of S2 except for class 3 (Table 3b). Note also that the minimum K values of SQ2 and SQ3 are substantially lower than those of SQ1 for most pressure heads.

[25] According to the procedure described by *Vandervaere et al.* [2000b], S_{opt} was calculated for each of the 150 cumulative infiltration curves, assuming $\beta = 0.6$ and S and K derived from equation (2), since:

$$A + B = \frac{2-\beta}{3}K + \frac{\gamma S^2}{r\Delta\theta} \tag{24}$$

There were 79 cases that fell in the gravity domains ($S_{opt} > S$) indicating that an accurate estimation of K is possible. For the remaining 71 cases, the estimation of K may be less precise since they were situated in the lateral capillarity domain. The theoretical study of *Vandervaere et al.* [2000b] indicated that soils covering a wide range of textural properties were in the lateral capillarity domain, making

Table 3a. Statistics of the *K* Estimates (cm min^{-1}) for the Q and SQ Groups

	Class									
	1	2	3	4	5	6	7	8	9	10
⟨ψ⟩, cm	−0.05	−1.35	−1.7	−2.6	−4.2	−4.8	−5.4	−7.4	−7.7	−8.3
					Mean					
Q1	15	10.5	10.7	4.8	3.6	3	3.1	3.1	2.4	2.3
Q2	15.4	8.5	9.4	2.4	3.5	3.4	4	3.3	2.2	2.2
SQ1	33.8	16.3	19.8	3.5	10.8	10.2	7.6	10.3	8.4	9
SQ2	32.4	14.7	16.2	–	8.5	8	6.9	7.6	6.4	6
SQ3	34	13.9	14.6	3.9	8	6.7	7.1	7.3	5.2	4.8
					CV					
Q1	33.2	71.3	40.4	–	41.8	30.6	28.7	25.1	29.3	18.7
Q2	50.5	40.8	51.3	–	64.1	58.7	23.4	29.4	36.1	31.7
SQ1	36.4	41.1	41.5	–	29.9	41.6	56.4	41	27.1	44.4
SQ2	57.9	37.1	53.7	–	43	64.6	57.1	55.8	38.7	69.8
SQ3	62.7	41.8	62.4	–	41.6	58	41.6	40.8	40.8	70.8
					Minimum					
Q1	11.5	4.0	3.1	–	0.5	1.6	1.8	1.5	0.5	1.7
Q2	9.9	2.8	0.3	–	1.1	0.7	2.7	2.0	0.6	1.8
SQ1	25.1	2.8	6.4	–	6.8	3.4	1.6	5.0	3.2	3.9
SQ2	19.2	4.4	1.4	–	1.5	0.3	0.9	0.2	0.5	1.6
SQ3	18.9	4.8	0.2	–	4.3	0.1	3.0	2.3	0.1	2.3
					Maximum					
Q1	18.6	40.3	23.1	–	6.5	4.7	4.0	4.1	4.2	2.9
Q2	20.9	13.6	20.5	–	9.9	8.0	5.2	1.5	3.5	3.0
SQ1	42.4	24.8	41.8	–	17.9	24.2	12.5	17.3	15.1	14.2
SQ2	45.8	24.2	36.7	–	15.4	24.0	10.9	12.8	12.5	12.2
SQ3	49.0	22.4	33.1	–	14.9	16.1	10.5	10.8	11.0	9.3

the *K*-determination with tension infiltrometers questionable. However, this study showed that field soils include both the gravity and the lateral capillarity domains within a single field plot.

4.2. Comparison of the References Curves with Other Hydraulic Conductivity Curves Measured at the Experimental Field

[26] For each method, the hydraulic conductivities were scaled to a reference curve. Results of scaling and the parameters of the reference curves are reported in Table 4. The number of locations in the scaling analysis differs between the methods since the negative values of *K* were discarded. Locations where only one *K* estimate was available were omitted from the scaling analysis. In general, scaling was very successful (i.e., SSE reduced by 85–92%) to describe the variability of the hydraulic conductivity.

[27] The seven *K*-ψ reference curves based on the seven methods are plotted in Figure 4. Except for case S1 for which K_s* is approximately three times higher than for the other 6 cases, K_s* values are close to each other. Similar n*-parameters were obtained for the Q and SQ groups whereas larger n* values were estimated for the S group. The β* values were the lowest for the SQ group and the highest for the S group. In the applied pressure head range (between −1 and −10 cm) the hydraulic conductivity of the 7 reference curves differ by almost one order of magnitude. At a pressure head of −20 cm, differences in *K* are almost three orders of magnitude due to the steep decrease in *K* in the S group. Neglecting the S group, *K* differs by one order of magnitude at −20 cm. This difference increases slightly with decreasing pressure head. These observations have a large impact on practical applications and numerical methods, since, starting from the same set of measurements, significantly different hydraulic properties can be derived depending on the interpretations and models used.

[28] As a comparison with other techniques, two hydraulic conductivity curves estimated on the same soil using different techniques are also plotted in Figure 4. The first curve was derived from scaling water retention data measured on

Table 3b. Statistics of the *K* Estimates (cm min^{-1}) for the S Group

	Class				
	1	2	3	4	5
⟨ψ⟩	−2.8515	−3.252	−3.672	−6.1125	−6.516
			Mean		
S1	9.2	8.9	4.1	2.3	3
S2	7.7	5.3	4.9	1.9	2.1
			CV		
S1	68	131.1	15	94	75.2
S2	45.7	48.2	18.9	36.7	35.3
			Minimum		
S1	1.6	0.3	3.7	$7.8\ 10^{-3}$	0.05
S2	2.4	1.5	4.3	0.3	0.8
			Maximum		
S1	22.5	47.6	4.6	8.2	7.5
S2	14.6	11.3	5.6	3.2	3.5

Table 4. Results of Scaling Analysis and Parameters of Reference Curves

	Scaling Results			Reference Curve			
	SSE Unscaled	SSE Scaled	Reduction, %	β^*, cm^{-1}	n^*	K_s^*, cm min^{-1}	l^*
Q1	58.66	4.59	92.2	0.0627	1.168	0.095	0.5
Q2	39.19	5.85	85.1	0.0676	1.202	0.0689	0.5
SQ1	17.38	2.55	85.4	0.0123	1.159	0.0827	0.5
SQ2	45.57	4.56	90	0.0344	1.18	0.0873	0.5
SQ3	33.05	2.67	91.9	0.0451	1.195	0.0926	0.5
S1	114.8	16.97	85.2	0.442	1.498	0.329	0.5
S2	20.66	2.96	85.7	0.167	1.478	0.074	0.5
Vanderborght et al. [1997]				0.019	1.288	0.0264	20
Jacques et al. [2002]				0.0137	1.56	0.0027	0.395

small soil samples (100 cm^3) to obtain β^* and n^* and hydraulic conductivity data as a function of water content derived from an internal drainage experiment on undisturbed soil columns (100 cm long, 30 cm diameter) to obtain l^*. Finally, K_s^* was estimated as the geometric mean of the measured saturated hydraulic conductivities on the small undisturbed soil samples (parameters are listed in Table 4 [*Vanderborght et al.*, 1997]). The second hydraulic conductivity curve was obtained from *Jacques et al.* [2002] who measured water content (using TDR), pressure heads (using tensiometers) and solute concentration (using TDR) at 5 depths and 12 locations along a 6 m long transect under natural boundary conditions during the 384 days. Based on the combination of these field measurements with a numerical water flow and solute transport model (HYDRUS-1D [*Šimůnek et al.*, 1998]), soil hydraulic parameters were obtained by an inverse optimization algorithm. Resulting parameters estimates for the top 0 – 15 cm are given in Table 4 [*Jacques et al.*, 2002]. The saturated hydraulic conductivities of these two curves were smaller than those obtained from scaling the unsaturated hydraulic conductivities from tension infiltrometer measurements (Table 4). Note that K_s^* for the curve of *Vanderborght et al.* [1997] was obtained as the geometric mean of measured K_s values on 180 small undisturbed soil samples collected at 10, 50 and 90 cm depth [*Mallants et al.*, 1996]. The samples collected at 10 cm depth gave a geometric mean K_s of 0.031 cm min^{-1}. At each location, subsequent to applying the tension infiltrometer, a small undisturbed soil sample was taken and the saturated hydraulic conductivity was measured in the laboratory using a constant head method. The geometric mean of the laboratory K_s was 0.059 cm min^{-1} which was remarkably close to the different model estimated K_s^*. As observed at this experimental field, K_s tended to decrease with increasing sampling volume [*Mallants et al.*, 1997b]. This may be one of the factors explaining the lower K_s^* obtained by *Jacques et al.* [2002] where they used water content data from a 6 m long transect. Another factor of lower K_s in *Jacques et al.*

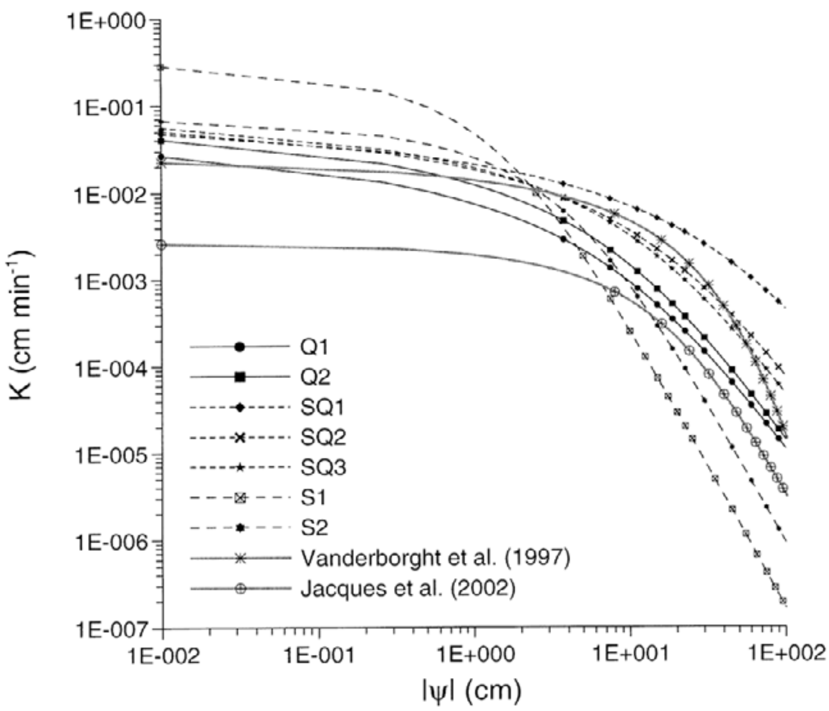


Figure 4. References curves obtained after scaling the estimated hydraulic conductivities from the seven methods and two hydraulic conductivity curves measured on the same soil with parameters from *Vanderborght et al.* [1997] and *Jacques et al.* [2002].

Table 5. Summary Statistics of Scaling Factors

Case	N	Mean[a]	CV[a]	Minimum	Maximum	Normal		Lognormal	
						W[b]	P	W	P
				α_ψ					
Q1	50	2.23 (2.12)	122.5 (112.9)	0.217	11.628	0.623	0	0.953	0.083
Q2	42	3.18 (3.02)	140.6 (155.9)	0.358	21.74	0.65	0	0.923	0.01
SQ1	46	1.91 (1.84)	104.1 (97.7)	0.344	8.85	0.691	0	0.944	0.042
SQ2	42	3.21 (3.59)	97.6 (156.4)	0.103	13.7	0.835	0	0.975	0.604
SQ3	40	3.78 (3.90)	135.7 (177.1)	0.113	25.64	0.674	0	0.983	0.859
S1	2	2.19 (1.73)	212.4 (97.2)	0.425	23.8	0.33	0	0.82	0
S2	445	1.88 (1.80)	103.4 (97.5)	0.489	7.5	0.686	0	0.845	0
				α_K					
Q1	50	1.00 (1.12)	115.9 (144.6)	0.024	7.56	0.626	0	0.933	0.01
Q2	42	0.99 (1.05)	79.1 (109.1)	0.113	4.15	0.862	0	0.95	0.096
SQ1	46	1.00 (1.00)	50.9 (50.3)	0.246	2.73	0.867	0	0.984	0.877
SQ2	42	1.00 (1.04)	72.1 (89.7)	0.065	3.37	0.878	0	0.967	0.358
SQ3	40	1.00 (1.05)	72.7 (100.7)	0.115	2.82	0.896	0	0.954	0.153
S1	24	1.00 (1.16)	55.8 (113.0)	0.044	2.32	0.957	0.3672	0.81	0
S2	45	1.00 (1.24)	79.4 (176.8)	0.048	3.29	0.916	0	0.886	0

[a] Values inside the parentheses are for a lognormal distribution (mean = $\exp(\mu_{\ln} + 0.5\,\sigma^2_{\ln})$, CV = $(\exp(\sigma^2_{\ln})-1)^{0.5}$) with $\mu_{\ln}$ and $\sigma^2_{\ln}$ the mean and variance of the $\log_e$-transformed scaling factors.
[b] W, Shapiro-Wilk W statistics at probability level P.

[2002] is that unsaturated water flow conditions prevailed most of the time during the field experiment.

[29] In the applied pressure head range (between −1 and −12 cm), the reference curve of *Vanderborght et al.* [1997] is close to the reference curves obtained in this study (neglecting method S1). At smaller pressure heads, however, the decrease of K is steeper for *Vanderborght et al.* [1997] compared to the other reference curves. This may be due to the significant difference in measurement techniques: a drainage experiment measuring K as a function of θ versus infiltration measurements determining K versus ψ. The optimized conductivity curve of *Jacques et al.* [2002] had

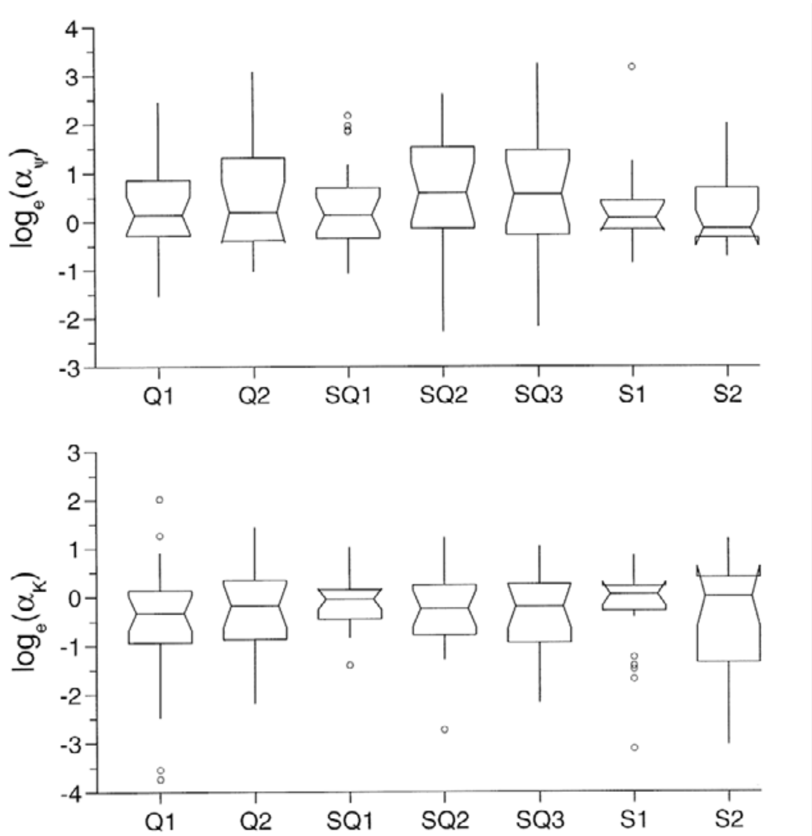


Figure 5. Box plots of $\log_e$-transformed scaling factors of the pressure head and the hydraulic conductivity for the seven methods.

a similar decrease in the drier pressure head range, but significantly smaller conductivities near saturation. Compared to the curves obtained in the other studies, the curves obtained here seem to stress more the near-saturated flow conditions. This is reflected in somewhat higher K_s^* and β^* estimates. From this comparison, the S method is not appropriate to estimate the hydraulic conductivity in the drier pressure head range, and except for Q1, the other methods are quite consistent.

4.3. Statistical Analysis of the Scaling Factors

[30] Summary statistics of the seven sets of scaling factors are given in Table 5. Figure 5 shows the box plots of scaling factors for each method. The Shapiro-Wilk W statistic [*Shapiro and Willk*, 1965] was used to test normality of the original and the $\log_e$-transformed scaling factors (Table 5). For the Q and SQ groups, both the $\log_e$-transformed α_ψ and α_K were better described by the normal distribution than the untransformed values. Few exceptions include $\log_e(\alpha_\psi)$ of Q2 and SQ1 and $\log_e(\alpha_K)$ of Q1 which were not normally distributed at the 5% significancy level. Also α_ψ and $\log_e(\alpha_\psi)$ were not normally distributed at the 5% significancy level for cases S1 and S2, although the W-statistics were somewhat larger for $\log_e(\alpha_\psi)$. In contrast, α_K of cases S1 and S2 were better approximated by the normal distribution than the $\log_e(\alpha_K)$. These observations for the S group are not in correspondence with the results from the Q and SQ group. Furthermore, statistics of the scaling factors of the S group are significantly different from those reported in other studies [e.g., *Warrick et al.*, 1977; *Jarvis and Messing*, 1995; *Jacques et al.*, 1997; *Kasteel*, 1997; *Mallants et al.*, 1997a; *Shouse and Mohanty*, 1998; *Zavattaro et al.*, 1999].

[31] The box plots in Figure 5 and the CVs in Table 5 show how much the variability of the scaling factors differs among the seven methods. For the Q and SQ groups, CVs of $\log_e(\alpha_\psi)$ are larger than those of $\log_e(\alpha_K)$. In most other scaling studies, the reverse was observed. However, those studies derived α_K from the saturated hydraulic conductivity and then subsequently scaled water retention data [e.g., *Jacques et al.*, 1997] or (near-saturated) hydraulic conductivities [e.g., *Shouse and Mohanty*, 1998]. Also when water retention and hydraulic conductivity data were simultaneously scaled using the relations of *Vogel et al.* [1991], the reverse was observed [e.g., *Kasteel*, 1997].

[32] The less variable sets of scaling factors were obtained for case SQ1, although the interquartile range IQ (i.e., the 75 percentile minus the 25 percentile) of S1 is smaller. However, the distributions for S1 showed some outliers. Furthermore, the interpretation of the distributional characteristic of the S1 group is somewhat suspicious due to the small sample size. For $\log_e(\alpha_\psi)$, IQ of Q1 and S2 are similar but smaller than the IQ of Q2, SQ2 and SQ3. For $\log_e(\alpha_K)$, IQ of S2 is larger than the relatively similar IQs of Q1, Q2, SQ2 and SQ3. And SQ1 and S1 have the lowest IQ for $\log_e(\alpha_\psi)$ and $\log_e(\alpha_K)$.

[33] A significant correlation between α_ψ and α_K at the 5% significancy level was found for the seven methods, except for S1. Correlation coefficients varied between -0.400 (for Q1) and -0.672 (for S2). Only five (out of 21) significant correlation coefficients for α_ψ between the different cases were found at the 5% significancy level (between Q1-Q2, Q2-SQ2, Q2-S1, SQ1-SQ2, SQ2-SQ3) varying between 0.345 and 0.760. For α_K, eight significant correlation coefficients were found at the 5% significancy level between Q1-Q2, Q1-S2, Q2-SQ2, Q2-SQ3, Q2-S2, SQ1-SQ2, SQ1-SQ3 and SQ2-SQ3 with correlation coefficients ranging between 0.336 and 0.768. Experimental variograms of the scaling factors show pure nugget (results not shown) for each method.

5. Conclusions

[34] In this study, a number of methods based on different infiltration models were used to derive the unsaturated hydraulic conductivity using cumulative infiltration curves measured with single-disc tension infiltrometers across a 40-m long transect on a silty loam Eutric Regosol. Based on our results, following conclusions were drawn:

1. Using the approach of *Vandervaere* [1995], it is possible to estimate the amount of water and the time needed to wet the contact layer. These two parameters are needed to correct the measured cumulative infiltration curve for the initial wetting of the contact layer. After subtracting i_0 and t_0 from the measured infiltration and time, the parameters of the analytical models can be estimated.

2. To derive the unsaturated hydraulic conductivity based on the measurement of the steady state infiltration rate only, measurements of the cumulative infiltration curve at two different pressure heads are required. This leads to difficulties for the precise determination of the 'hydraulic conductivity-pressure head' couple. A similar remark is valid for the methods based on two sorptivity measurements. Therefore methods based on the measurement of a single infiltration curve are preferable, i.e. the SQ methods. According to *Hussen and Warrick* [1993], the modified Horton model gives good estimates of both S and Q_s, and therefore SQ3 might be an appropriate method to derive the unsaturated hydraulic conductivity from a single-location, single-pressure head infiltration experiment. The results of all (statistical) analyses in this study for the SQ2 method are consistent with those for the SQ3 method, and thus SQ2 forms an alternative for SQ3. From a practical point of view, the experimental burden is the same for both methods.

3. The methods based on the determination of two sorptivity values failed to give acceptable results for this soil. Various indications for this are the relatively small number of positive estimates of K for S1, the significant differences between the reference curves for S1 and S2 compared with the other reference curves and the other hydraulic conductivity curves measured at the experimental field, the statistics of their scaling factors, and insignificant correlation between the scaling factors of the S group with the other methods. One possible reason, as suggested by *Vandervaere et al.* [2000b] the sorptivity is only accurately estimated when the soil is in the lateral capillarity domain. In that case, however, the determination of K is less accurate. On the other hand, when the soil is in the gravity domain, the determination of S is subject to large uncertainties, resulting in large uncertainties on K estimation (based on S only).

4. A remarkable observation was that the geometric mean of the saturated hydraulic conductivity measured on small undisturbed soil samples (100 cm^3) taken along the same transect and same locations (underneath the disc) was close to the different estimated K_s^* values. This observation

was rather unique given the large variability and uncertainty of these measurements. However, the applicability of this geometric mean as an appropriate value to define K_s^* of the reference near-saturated hydraulic conductivity curve should be tested for other soils before application. Overall, results of this study indicate that collocated sorptivity and steady state infiltration rates should be jointly used for the estimation of soil hydraulic conductivities near saturation and their spatial variability.

[35] **Acknowledgments.** The first author wants to acknowledge the financial support of the Flemish Institute for the Encouragement of Scientific-Technical Research in the Industry (IWT). We also thank Dave Elrick for his helpful suggestions. Thanks are due to the Department of Land Resource Science, University of Guelph, Canada, and LETHE, Université J. Fourier, France, for lending their infiltrometers. We acknowledge the Deckers family, owners of the experimental field, for the permission to perform these experiments on their property. The assistance of T. Deleus, L. Feyen, A. Timmerman, M. Vanhoorick, and Z. Wang in the field is highly appreciated. We also acknowledge the constructive comments of the anonymous reviewers.

References

Angulo-Jaramillo, R., J.-P. Gaudet, J.-L. Thony, and M. Vauclin, Measurement of hydraulic properties and mobile water content of a field soil, *Soil Sci. Soc. Am. J.*, *60*, 710–715, 1996.

Angulo-Jaramillo, R., F. Moreno, B. E. Clothier, J.-L. Thony, G. Vachaud, E. Fernandez-Boy, and J. A. Cayuela, Seasonal variation of hydraulic properties of soils measured using a tension disc infiltrometer, *Soil Sci. Soc. Am. J.*, *61*, 27–32, 1997.

Angulo-Jaramillo, R., J.-P. Vandervaere, S. Roulier, J.-L. Thony, J.-P. Gaudet, and M. Vauclin, Field measurements of soil surface hydraulic properties by disc and ring infiltrometers: A review and recent developments, *Soil Tillage Res.*, *55*, 1–29, 2000.

Ankeny, M. D., M. Ahmed, T. C. Kasper, and R. Horton, Simple method for determining unsaturated hydraulic conductivity, *Soil Sci. Soc. Am. J.*, *55*, 467–470, 1991.

Buttle, J. M., and D. J. McDonald, Soil macroporosity and infiltration characteristics of a forest podzol, *Hydrol. Processes*, *14*, 831–848, 2000.

Carnahan, B., H. A. Luther, and J. O. Wilkes, *Applied Numerical Methods*, 604 pp., John Wiley, New York, 1969.

Casanova, M., I. Messing, and A. Joel, Influence of aspect and slope gradient on hydraulic conductivity measured by tension infiltrometer, *Hydrol. Processes*, *14*, 155–164, 2000.

Casey, F. X. M., S. D. Logsdon, R. Horton, and D. B. Jaynes, Immobile water content and mass exchange coefficient of a field soil, *Soil Sci. Soc. Am. J.*, *61*, 1030–1036, 1997.

Casey, F. X. M., S. D. Logsdon, R. Horton, and D. B. Jaynes, Measurement of field soil hydraulic and solute transport parameters, *Soil Sci. Soc. Am. J.*, *62*, 1172–1178, 1998.

Clausnitzer, V., J. W. Hopmans, and D. R. Nielsen, Simultaneous scaling of soil water retention and hydraulic conductivity curves, *Water Resour. Res.*, *28*, 19–31, 1992.

Clothier, B. E., M. B. Kirkham, and J. E. McLean, In situ measurement of the effective transport volume for solute moving through soil, *Soil Sci. Soc. Am. J.*, *56*, 733–736, 1992.

Clothier, B. E., L. Heng, G. N. Magesan, and I. Vogeler, The measured mobile water content of an unsaturated soil as a function of hydraulic regime, *Aust. J. Soil Res.*, *33*, 397–414, 1995.

Cook, F. J., and A. Broeren, Six methods for determining sorptivity and hydraulic conductivity with disc permeameters, *Soil Sci.*, *157*, 2–11, 1994.

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

Haverkamp, R., P. J. Ross, P. J. Smettem, and J. Y. Parlange, Three-dimensional analysis of infiltration from the disc infiltrometer, 2, Physically based infiltration equation, *Water Resour. Res.*, *30*, 2931–2935, 1994.

Horton, R. E., An approach toward a physical interpretation of infiltration capacity, *Soil Sci. Soc. Am. Proc.*, *5*, 399–417, 1940.

Hussen, A. A., and W. A. Warrick, Algebraic models for disc tension permeameters, *Water Resour. Res.*, *29*, 2779–2786, 1993.

Jacques, D., J. Vanderborght, D. Mallants, B. P. Mohanty, and J. Feyen, Analysis of solute redistribution in heterogeneous soil, I, Geostatistical approach to describe the spatial scaling factors, in *geoENV I—Geostatistics for Environmental Applications*, edited by A. Soarez, J. Goméz-Hernandez, and R. Froideveaux, pp. 271–282, Kluwer Acad., Norwell, Mass., 1997.

Jacques, D., J. Feyen, and D. Mallants, Determination of hydraulic properties using tension infiltrometer data and inverse optimization, in *Characterization and Measurement of the Hydraulic Properties of Unsaturated Porous Media*, edited by M. T. van Genuchten, F. J. Leij, and L. Wu, pp. 749–760, Univ. of Calif., Riverside, 1999.

Jacques, D., J. Šimůnek, A. Timmerman, and J. Feyen, Calibration of Richards' and convection-dispersion equations to field-scale water flow and solute transport under rainfall conditions, *J. Hydrol.*, *259*, 15–31, 2002.

Jarvis, N. J., and I. Messing, Near-saturated hydraulic conductivity in soils of contrasting texture measured by tension infiltrometers, *Soil Sci. Soc. Am. J.*, *59*, 27–34, 1995.

Jaynes, D. B., S. D. Logsdon, and R. Horton, Field method for measuring mobile/immobile water content and solute transfer rate coefficient, *Soil Sci. Soc. Am. J.*, *59*, 352–356, 1995.

Kasteel, R., Solute transport in an unsaturated field soil: Describing heterogeneous flow fields using spatial distribution of hydraulic properties, *Diss. ETH 12477*, 108 pp., Eidg. Tech. Hochsch., Zurich, Switzerland, 1997.

Logsdon, S. D., Transient variation in the infiltration rate during measurements with tension infiltrometers, *Soil Sci.*, *162*, 233–241, 1997.

Logsdon, S. D., and D. B. Jaynes, Spatial variability of hydraulic conductivity in a cultivated field at different times, *Soil Sci. Soc. Am. J.*, *60*, 703–710, 1996.

Mallants, D., D. Jacques, M. Vanclooster, J. Diels, and J. Feyen, A stochastic approach to simulate water flow in a macroporous soil, *Geoderma*, *70*, 299–324, 1996.

Mallants, D., D. Jacques, P.-H. Tseng, M. T. van Genuchten, and J. Feyen, Comparison of hydraulic measurement methods using three sizes of soil cores, *J. Hydrol.*, *199*, 295–318, 1997a.

Mallants, D., B. P. Mohanty, A. Vervoort, and J. Feyen, Spatial analysis of saturated hydraulic conductivity in a soil with macropores, *Soil Technol.*, *10*, 115–131, 1997b.

MathSoft, Mathcad user's guide, Mathcad 6.0, Mathcad PLUS 6.0, Cambridge, Mass., 1995.

Minasny, B., and A. B. McBratney, Estimation of sorptivity from disc-permeameter measurements, *Geoderma*, *95*, 305–324, 2000.

Mohanty, B. P., M. D. Ankeny, R. Horton, and R. S. Kanwar, Spatial analysis of hydraulic conductivity measured using disc infiltrometers, *Water Resour. Res.*, *30*, 2489–2498, 1994.

Mohanty, B. P., R. S. Bowman, J. M. H. Hendrickx, and M. T. van Genuchten, New piecewise-continuous hydraulic functions for modeling preferential flow in an intermittent-flood-irrigated field, *Water Resour. Res.*, *33*, 2049–2063, 1997.

Murphy, B. W., T. B. Koen, B. A. Jones, and L. M. Huxedurp, Temporal variation of hydraulic properties of some soils with fragile structure, *Aust. J. Soil Res.*, *31*, 179–197, 1993.

Philip, J. R., The theory of infiltration, 4, Sorptivity and algebraic infiltration equations, *Soil Sci.*, *84*, 257–264, 1957.

Philip, J. R., Theory of infiltration, *Adv. Hydrosci.*, *5*, 215–296, 1969.

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

Philip, J. R., Linearized unsteady multidimensional infiltration, *Water Resour. Res.*, *22*, 1717–1727, 1986.

Pullan, A. J., Linearized time-dependent infiltration from a shallow pond, *Water Resour. Res.*, *28*, 1041–1046, 1992.

Quadri, M. B., L'infiltrométrie multi-disques pour la caractérisation hydrodispersive des sols non saturés: Modélisation des transferts et applications, thèse de doctorat. 226 pp., l'Univ. Joseph Fourier, Grenoble, France, 1993.

Reynolds, W. D., and D. E. Elrick, Determination of hydraulic conductivity using a tension infiltrometer, *Soil Sci. Soc. Am. J.*, *55*, 633–639, 1991.

Rice, J. A., *Mathematical Statistics and Data Analysis*, 602 pp., Duxburry, Belmont, Calif., 1995.

SAS Institute, SAS/STAT user's guide, version 6, 4th ed., vol. 2, 864 pp., Cary, N. C., 1989.

Shapiro, S. S., and M. B. Willk, An analysis of variance test for normality (complete samples), *Biometrica*, *52*, 591–611, 1965.

Shouse, P. J., and B. P. Mohanty, Scaling of near-saturated hydraulic conductivity measured using disc infiltrometers, *Water Resour. Res.*, *34*, 1195–1205, 1998.

Šimůnek, J., M. Šejna, and M. T. van Genuchten, The HYDRUS-1D software package for simulating the one-dimensional movement of water, heat, and multiple solutes in variably-saturated media, version 2.0, *IGWMC-TPS-80*, 202 pp., Int. Ground Water Model. Cent., Colo. Sch. of Mines, Golden, 1998.

Soil Survey Staff, *Keys to Soil Taxonomy*, 6th ed., U.S. Gov. Print. Off., Washington, D. C., 1994.

Thony, J.-L., G. Vachaud, B. E. Clothier, and R. Angulo-Jaramillo, Field measurements of the hydraulic properties of the soil, *Soil Technol.*, *4*, 111–123, 1991.

Vanderborght, J., D. Jacques, D. Mallants, P.-H. Tseng, and J. Feyen, Comparison between field measurements and numerical simulation of steady-state solute transport in a heterogeneous soil profile, *Hydrol. Earth Syst. Sci.*, *4*, 853–871, 1997.

Vandervaere, J. P., Caractérisation hydrodynamique du sol in situ par infiltromètrie à disques: Analyse critique de régimes pseudo-permanentes, méthodes transitoires et cas de sols éncroutès, thèse de doctorat, 329 pp., Univ. Joseph Fourier, Grenoble, France, 1995.

Vandervaere, J. P., C. Peugot, M. Vauclin, R. Angulo Jaramillo, and T. Lebel, Estimating hydraulic conductivity of crusted soils using disc infiltrometers and minitensionmeters, *J. Hydrol.*, 188–189, 203–223, 1997.

Vandervaere, J. P., M. Vauclin, and D. E. Elrick, Transient flow from tension infiltrometers, I, The two-parameter equation, *Soil Sci. Soc. Am. J.*, *64*, 1263–1272, 2000a.

Vandervaere, J. P., M. Vauclin, and D. E. Elrick, Transient flow from tension infiltrometers, II, Four methods to determine sorptivity and conductivity, *Soil Sci. Soc. Am. J.*, *64*, 1272–1284, 2000b.

van Genuchten, M. T., A closed-form equation for predicting the hydraulic conductivity of unsaturated soils, *Soil Sci. Soc. Am. J.*, *44*, 892–898, 1980.

Velleman, P. F., and D. C. Hoaglin, *Applications, Basics and Computing of Exploratory Data Analysis*, Duxburry, Belmont, Calif., 1981.

Vogel, T., M. Cislerova, and J. W. Hopmans, Porous media with linearly variable hydraulic properties, *Water Resour. Res.*, *27*, 2735–2741, 1991.

Warrick, A. W., Models for disc infiltrometers, *Water Resour. Res.*, *28*, 1319–1327, 1992.

Warrick, A. W., G. J. Mullen, and D. R. Nielsen, Scaling field-measured soil hydraulic properties using a similar media concept, *Water Resour. Res.*, *13*, 355–362, 1977.

Warrick, A. W., P. Broadbridge, and D. O. Lomen, Approximations for diffusion from a disc source, *Appl. Math. Modelling.*, *16*, 155–161, 1992.

White, I., and K. M. Perroux, Estimation of unsaturated hydraulic conductivity from field sorptivity measurements, *Soil Sci. Soc. Am. J.*, *53*, 324–329, 1989.

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

White, I., M. J. Sully, and K. M. Perroux, Measurement of surface-soil hydraulic properties: Disc permeameter, tension infiltrometers and other techniques, in *Advances in Measurement of Soil Properties: Bringing Theory in Practice*, *Spec. Publ. Soil Sci. Soc. Am. 30*, pp. 69–103, Soil Sci. Soc. of Am., Madison, Wiss., 1992.

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

Zavattaro, L., N. Jarvis, and L. Persson, Use of similar media scaling to characterize spatial dependence of near-saturated hydraulic conductivity, *Soil Sci. Soc. Am. J.*, *63*, 486–492, 1999.

Zhang, R., Determination of soil sorptivity and hydraulic conductivity from the disk infiltrometer, *Soil Sci. Soc. Am. J.*, *61*, 1024–1030, 1997.

Zhang, R., Estimating soil hydraulic conductivity and macroscopic capillary length from the disk infiltrometer, *Soil Sci. Soc. Am. J.*, *62*, 1513–1521, 1998.

J. Feyen, Institute for Land and Water Management, Katholieke Universiteit Leuven, Vital Decosterstraat 102, B-3000 Leuven, Belgium.

D. Jacques, SCK-CEN, Boeretang 2000, B-2400 Mol, Belgium. (djacques@sckcen.be)

B. P. Mohanty, Biological and Agricultural Engineering Department, Texas A&M University, College Station, TX 77843-2117, USA.