Hydraulic Conductivity as Influenced by Stoniness in Degraded Drylands of Chile

Infiltration measurements in arid, stony soils are notoriously variable in visually homogeneous areas, and have been reported to be influenced by embedded stone fragments. This study aimed to identify the influence of rock fragment contents, orientation, and position within a small arid watershed on hydraulic conductivity in northern Chile. Two different measurement techniques were used, a single-ring infiltrometer with constant ponding head and a tension infiltrometer, which were applied at both an undisturbed field site (44 locations along three transects) and on the disturbed <2-mm soil fraction from the same locations. Variations in saturated and unsaturated hydraulic conductivities were observed when using different calculation methods, adding to the observed variability. For saturated conditions, only small differences in conductivities were observed between two calculation methods, whereas unsaturated hydraulic conductivities calculated by five different methods showed more important variations. Stone fragment content correlated significantly with both saturated and unsaturated conductivities, probably due to a positive correlation between stone content and coarse lacunar pore space. Slope orientations with higher amounts of stone fragments gave higher infiltration rates, as well as transects with steeper slopes and more, but smaller, rock fragments. When evaluating differences in infiltration rates observed along three transects in the watershed, variability could be mainly attributed to stone fragment content influences.

Soil degradation is a common feature in the arid and semiarid regions of Chile, where limited water resources in combination with a high population pressure and overgrazing lead to severe deforestation and increased erosion hazards (Soto, 1999). Due to slow soil-forming processes and high erosion rates, soils in those regions are often shallow and characterized by the occurrence of stone fragments near the soil surface (Alcayaga and Narbona, 1977). These stone or rock fragments are soil particles with a diameter >2 mm and up to 2 m (Poesen and Lavee, 1994). The spatial variability of stone fragments in soils can be very large (Childs and Flint, 1990) and has been observed to vary depending on the direction concerned (Webster, 1985).

Embedded stone fragments affect soil hydraulic properties. According to Mehues et al. (1975), they reduce the soil’s porosity and increase the tortuosity of water flow paths, restricting overall water movement. On the other hand, Sauer and Logsdon (2002) observed slightly positive relations between stone fragment content and saturated hydraulic conductivity, in agreement with the observations of Ravina and Magier (1984), who reported that rock fragment content increases tortuosity, but also created new voids, thereby positively influencing infiltration rates. Fiès et al. (2002) confirmed that stones do create extra porosity as a result of incomplete filling by fine earth of the spaces between rock fragments or because the larger particles prevent smaller ones from compacting (Poesen and Lavee, 1994; Stewart et al., 1970). Additionally, in stony soils with a clay content >30%, shrinkage of the fine phase leads to the formation of coarse lacunar pores due to cracking (Fiès et al., 2002; Towner, 1988).

An additional source of variability when considering infiltration measurements is the measurement method and the subsequent data analysis. Reynolds et al. (2000) compared...
three methods (pressure infiltrometer, tension infiltrometer, and the constant-head soil core method; note that we use the same terminology as in Dane and Topp, 2002, p. 797–1047) to determine the saturated hydraulic conductivity and found very little correlation among the methods used. Mohanty et al. (1994) observed similar differences when comparing the constant-head well (Guelph) permeameter, falling-head well permeameter, tension infiltrometer, concentric ring infiltrometer, and constant-head soil core methods. Gómez et al. (2001) found better correlations among four different measurement methods (falling-head well permeameter, pressure infiltrometer, tension infiltrometer, and rainfall simulator) and were able to detect significant differences in infiltration rates between and under olive (Olea europaea L. subsp. europaea) trees in southern Spain with each of the methods used.

Apart from different measurement instruments, different calculation techniques have become available in recent years to obtain the hydraulic conductivity from cumulative infiltration curves, dividing those that use steady-state infiltration rates from those using the transient approach. The former use a derivation of Wooding’s analysis (Ankeny et al., 1991; Reynolds and Elrick, 1991; Wooding, 1968), whereas the latter models are based on the whole infiltration curve and can be empirical (e.g., Horton, 1939), be physically based (e.g., Haverkamp et al., 1994), or use inverse modeling techniques (Simunek and van Genuchten, 1996). Each of these calculation techniques results in a different estimation of the hydraulic conductivity, adding to the uncertainty of the real value.

The objective of this study was to evaluate infiltration in the heterogeneous stony soils of northern Chile and to answer the following questions:

1. What is the influence of different infiltration measurement techniques and calculation methods on the hydraulic conductivity?
2. Is it possible to clearly identify the effect of embedded stone fragments on infiltration?
3. What variability in infiltration rates can be expected in a small watershed that is characterized by heterogeneous, stony soils, but with only small textural changes between locations?

MATERIALS AND METHODS

Field Site

The field site considered in this study is located in the north of Chile (30°17′31.1″ S, 71°16′65.9″ W) and consists of a small watershed (65 ha) that forms part of the larger Las Cardas Experimental Watershed. The study area has an arid Mediterranean climate, characterized by: (i) an average annual precipitation of 138 mm, of which >70% is produced in the southern winter season; (ii) moderate temperatures, with an absolute minimum of 2°C (June) and an absolute maximum of 30°C (March); (iii) a high relative humidity (80%) with frequent cloudiness; (iv) an average annual solar radiation of 170.7 MJ m−2 d−1, resulting in a water deficit of 800 to 1000 mm yr−1 (Miller, 1976). The climate is categorized as arid, using the aridity index proposed by Middleton and Thomas (1997), and scarce precipitation is often concentrated in short bursts of high-intensity rainfall, leading to a high Modified Fournier Index (Fournier, 1960, also called climatic aggressivity) in 10% of the years (1970–2000). Due to these climatic conditions, vegetation cover in the area is limited and mainly composed of shrubs, herbs, and cacti (Miller, 1976; Olivares and Squeo, 1999), leading to a high exposure to runoff and erosion risks. In the watershed, alcaparra shrubs [Senna cumingii (Hook. & Arn.) H. S. Irwin & Barneby] and Coquimbo shrubs [Atriplex virgata (Vogel)] H. S. Irwin & Barneby] dominate the terrain, which is a strong indicator for overgrazing practices by goat (Capra hircus) flocks (Azócar and Lailhacar, 1990).

To capture the range of variable soil conditions within the watershed, three transects were laid out perpendicular to the (dry) riverbed from hilltop to hilltop in a northwest direction (Fig. 1), with a mean slope of 0.18, 0.18, and 0.29 m m−1 for the first, second, and third transect, respectively. On each of them, 15 locations were selected to perform the infiltration measurements in such a manner that each point would be representative of the surrounding area, forcing the measurement sites to be unevenly spaced. Preference was given to measurement locations with at least some slope microrelief as possible, while compacted goat paths were avoided. One measurement location on the third transect was discarded due to soil perturbation by intense animal activity (the Chilean rose tarantula, Grammostola rosea Walckenaer), reducing the total number of data points to 44.

Field and Laboratory Measurements

At each measurement site, superficial litter and loose stones were removed without altering the soil surface to preserve topsoil characteristics. Consequently, a metal ring with an inner diameter of 0.28 m and a height of 0.25 m was driven in the soil for at least 0.03 m, and a contact material consisting of 99% sand (>50 μm) with a saturated hydraulic conductivity of 0.277 mm s−1 and an air-entry value of −0.3 kPa was used to level the area within the metal ring. Since the saturated hydraulic conductivity of the contact material exceeded that of the measured hydraulic conductivities, the effects of the sand layer can be neglected (e.g., Bagarello et al., 2000; Vandervaere et al., 2000; Bodhinayake et al., 2004).

First, a measurement with a single-ring infiltrometer with constant ponding head was performed by maintaining a constant head of 0.03 m, using the Mariotte system of the Model 2800 Guelph permeameter (Soilmoisture Equipment, Santa Barbara, CA). Readings of the cumulative infiltration were done manually every minute for at least 30 min or until the infiltration rates of three successive time intervals remained constant. During sunny days, the reservoir was covered with a white cloth to avoid temperature influences on the vacuum chamber.

After reaching steady-state ponded infiltration, a Model 2825 tension infiltrometer adaptor module (Soilmoisture Equipment, Santa Barbara, CA) with a diameter of 0.20 m was attached to the Mariotte system of the Guelph permeameter and pressed firmly on top of the contact material inside the ring. Three successive negative pressure heads were applied, −0.29, −0.59, and −1.18 kPa, for at least 15 min or until the infiltration rate of three consecutive time intervals was constant. Although no correction was made for the variable thickness of the contact sand layer between different measurement sites, calculations using the approach formulated by Reynolds (2006) indicated that contact sand layer effects on hydraulic conductivities were limited, with mean errors of 1.0, 0.69, and 0.3 μm s−1, respectively, for the successive pressure heads applied. The cumulative infiltration data of both saturated and unsaturated measurements were used to determine the saturated (KSat) and unsaturated (Kvp) hydraulic conductivity of the chosen field sites. All field measurements were conducted between...
14 Aug. and 8 Oct. 2006, during which no rainfall events were observed.

At each measurement location, an undisturbed sample of the topsoil was taken using standard sharpened steel 100-cm$^3$ Kopecky rings with a ring depth of 0.05 m and a radius of 0.025 m. On these samples, the soil water retention curve was determined using a sand box apparatus (Eijkelkamp Agrisearch Equipment, Giesbeek, the Netherlands) for matric potentials between −0.98 and −9.81 kPa, with pressure chambers (Soilmoisture Equipment, Santa Barbara CA) for matric potentials between −19.61 and −1471.01 kPa. Bulk density was measured using the excavation method (Blake and Hartge, 1986) on a representative elementary soil volume of 3.38 × 10$^{-3}$ m$^3$ (a cube with sides of 0.15 m) and is represented in Fig. 2 for the 44 measurement locations in this study. The soil was oven dried at 105°C for 24 h, after which the dry soil mass ($M_t$) was weighed. For each of the samples, the rock fragment content was evaluated using three different characteristics. The fine earth fraction was separated from the rock fragments by wet sieving through a 2-mm mesh. After drying the remaining fragments, the rock fragment mass was weighed, from which the gravimetric stone fragment content (or gravel content) ($R_m$) was determined as the rock fragment mass per total mass of the soil sample. The stone fragment volume was obtained by water displacement using the law of Archimedes, and was used to calculate the volumetric stone fragment content ($V'_p$), expressed as the rock fragment volume per total volume of the soil sample. The volume ($V'_f$) and mass ($M'_f$) of the <2-mm fraction could then easily be obtained as the difference between the values determined before wet sieving ($V'_t$ and $M'_t$) and those measured for the retained rock fragments ($V'_{rf}$ and $M'_{rf}$). The rock fragment distribution was measured using nine different sieve meshes: 2.36, 4.75, 9.5, 19, 25, 37.5, 50, 63, and 75 mm, allowing the determination of the mean weight diameter (MWD) of the rock fragments present at each measurement location and for each sample size. Using the calculated stone fragment content characteristics, the coarse lacunar pore space ($V_p$) was determined using the formula proposed by Fiès et al. (2002):

$$V_p = \frac{1}{\rho_{b,t}} - \frac{1 - R_m}{\rho_{b,f}} - \frac{R_m}{\rho_{b,rf}}$$

where $1/\rho_{b,t}$ is the volume of a unit mass of soil with bulk density $\rho_{b,t}$, $(1 - R_m)/\rho_{b,f}$ is the volume of a mass $(1 - R_m)$ of the <2-mm fine soil fraction with bulk density $\rho_{b,f}$ and $R_m/\rho_{b,rf}$ is the volume occupied by a mass $R_m$ of rock fragments of particle density $\rho_{b,rf}$. Thus $V'_p$ represents only the coarse lacunar pores that are left due to the presence of the rock fragments (Fiès et al., 2002). The value of $\rho_{b,f}$ was estimated by filling a Kopecky ring with the fine <2-mm earth fraction from each sample location, whereas the $\rho_{b,t}$ was determined as $M'_t/V'_t$ and $\rho_{b,rf}$ as $M'_{rf}/V'_{rf}$.

The remaining <2-mm fine earth fraction from each measurement location was used to fill a circular metal column with a diameter of 0.22 m using the previously determined bulk density of the corresponding field site, ranging from 1.40 to 2.20 g cm$^{-3}$. This was done by filling the cylinder in layers of 0.05-m height and tamping down the soil until the appropriate density was reached. In agreement with the field experiments, both the single-ring infiltrometer with constant ponding head and the tension infiltrometer were applied to determine the saturated ($K_s$) and unsaturated hydraulic conductivity ($K_{upp}$). Disturbed soil material from each measurement point was used to determine the soil texture (Fig. 3) using the pipette method (Gee and Or, 2002) and for organic matter content determination (Fig. 2).
Calculation Methods

To determine the saturated and unsaturated hydraulic conductivity, different calculation methods were applied and compared. With regard to the saturated hydraulic conductivity, the method of Reynolds and Elrick (1990) and the method formulated by Wu and Pan (1997) were considered. The former method, referred to here as the RE1 method, consists in calculating the saturated hydraulic conductivity \( K_s \) from data obtained with a single-ring infiltrometer at steady state. According to this method, the saturated hydraulic conductivity \( K_s \) follows from

\[
K_s = \frac{q_s}{\pi R_s^2} \left( \frac{H}{C_1d + C_2 R_s^2} + \lambda_c \right) + 1
\]

where \( q_s \) [L^3 T^{-1}] is the steady-state flow rate, \( R_s \) [L] is the ring radius, \( H \) [L] is the steady-state depth of ponded water in the ring, \( \lambda_c \) [L] is the macroscopic capillary length, and \( d \) [L] is the depth of ring insertion into the soil. The coefficients \( C_1 = 0.316 \pi \) and \( C_2 = 0.184 \pi \) are dimensionless quasi-empirical constants that apply for \( d \geq 0.03 \) m and \( H \geq 0.05 \) m. Nevertheless, only very small changes are observed in the form factors for \( 0.02 \) m < \( H < 0.05 \) m (Reynolds and Elrick, 1990).

A more generalized solution to describe infiltration from single-ring infiltrometers was developed by Wu and Pan (1997) using scaling techniques (the WP method). Their infiltration equation can be written as

\[
i = \frac{i_c}{t_c} = a + b \left( \frac{t}{T_c} \right)^{-0.5}
\]

where \( i \) [L T^{-1}] is the infiltration rate, \( i_c \) [L T^{-1}] is the characteristic infiltration rate, \( t \) [T] is time, \( T_c \) [T] is the characteristic time scale, and \( a \) and \( b \) are dimensionless constants (\( a = 0.9084, b = 0.1682 \)). The value of \( K_s \) can be calculated by fitting the simplified integrated form of Eq. [3]:

\[
I = At + B\sqrt{t}
\]

against the obtained infiltration data, yielding the coefficients \( A \) and \( B \), from which \( K_s \) can be obtained using

\[
K_s = \frac{\Delta \theta \lambda_c}{T_c}
\]

where \( \Delta \theta = \theta_s - \theta_i \), \( \theta_s \) is the saturated water content, and \( \theta_i \) is the initial water content. The parameters \( \lambda_c \) and \( T_c \) are a function of the ponded depth in the ring, the ring insertion depth, the radius of the ring infiltrometer, form factors \( a \) and \( b \) (Eq. [3]), and regression coefficients \( A \) and \( B \) from Eq. [4] (Wu et al., 1999).

Wu and Pan (1997) found the \( K_s \) values obtained with their method to be comparable with those calculated using the single-head method of Reynolds and Elrick (1990) and close to the true values determined for three different types of soils using numerical tests.

The unsaturated hydraulic conductivity and its relation to pressure potential were obtained from tension infiltrometer measurements using three types of calculation methods. The first type of method is based on the solution of Wooding’s equation for unconfined steady-state infiltration from a circular pond (Wooding, 1968):

\[
\frac{q_s}{\pi R^2} = K(\psi) \left( 1 + \frac{4}{\pi R K_s} \right)
\]

where \( q_s \) is the steady-state flow rate [L^3 T^{-1}], \( R \) is the radius of the disk [L], \( K(\psi) \) is the hydraulic conductivity [L T^{-1}] and \( \kappa \) [L^{-2}] is a fitting parameter. The two unknowns \( K(\psi) \) and \( \kappa \) can be found from tension infiltrometer measurements. This steady-state approach was used by Ankeny et al. (1991), Reynolds and Elrick (1991), and Logsdon and Jaynes (1993), using different pressure potentials.

The simultaneous approach of Ankeny et al. (1991), further denoted as the AN method, consists in solving a set of three equations simultaneously, for two different pressure heads, \( \psi_1 \) and \( \psi_2 \):

\[
K_{\psi_1} = \frac{q_s}{\pi R^2 + 2(\psi_1 - \psi_2) R \left( \frac{q_s}{q_{\psi_1}} \left( 1 + \frac{q_{\psi_2}}{q_{\psi_1}} \right) / \left( 1 - \frac{q_{\psi_2}}{q_{\psi_1}} \right) \right)}
\]

\[
K_{\psi_2} = \frac{q_s K_{\psi_1}}{q_{\psi_1}}
\]

\[
\kappa = \frac{2(K_{\psi_1} - K_{\psi_2})}{(\psi_1 - \psi_2)(K_{\psi_1} + K_{\psi_2})}
\]
The conductivity corresponding to these suctions can be found by the IV method allowing indirect determination of the hydraulic parameters from transient tension infiltrometer data. In this approach, the unknown hydraulic parameters are estimated from observed cumulative infiltrometer data by numerical inversion of the Richards (1931) equation, which can be written for radial symmetric Darcian flow in terms of pressure potential \( \psi \) as

\[
\frac{\partial \theta}{\partial t} = \frac{1}{r} \frac{\partial}{\partial r} \left[ r K(\psi) \frac{\partial \psi}{\partial r} \right] + \frac{\partial}{\partial \psi} \left[ K(\psi) \frac{\partial \psi}{\partial \psi} \right] + \frac{\partial K(\psi)}{\partial \psi} \tag{17}
\]

This equation is subject to the following initial and boundary conditions (Warrick, 1992):

\[
\theta(r, \psi, t) = \theta_0 \text{ or } \psi(r, \psi, t) = \psi_0 \text{ at } t = 0
\]

\[
\psi(r, \psi, t) = \psi_0 \text{ with } 0 < r < r_0 \text{ at } \psi = 0
\]

\[
\frac{\partial \psi(r, \psi, t)}{\partial \psi} = -1 \text{ with } r > r_0 \text{ at } \psi = 0
\]

\[
\psi(r, \psi, t) = \psi_1 \text{ at } r^2 + \psi^2 = \infty
\]

where \( r \) [L] and \( \psi \) [L] are the radial and vertical distances, respectively, \( \psi_0 \) [L] is the initial pressure potential, \( \theta_0 \) [L^3 L^{-3}] is the initial water content, and \( \psi_0 \) [L] is the imposed supply pressure potential. Solving these equations is possible using the finite element code HYDRUS-2D developed by Šimunek et al. (1999a), which uses the parametric models of van Genuchten (1980) for the \( \theta(\psi) \) relation and of Mualem (1976) for the \( K(\psi) \) relation:

\[
K(\psi) = K_\infty \left( 1 - \left( \frac{\psi}{\psi_0} \right)^{n} \right)^{\lambda} \tag{19}
\]

\[
\theta(\psi) = \begin{cases} 
\theta_0 + \left( \frac{\theta_s - \theta_0}{1 + \alpha \psi} \right) & b < 0 \\
\theta_s & b \geq 0
\end{cases} \tag{20}
\]

\[
S_e = \frac{\theta - \theta_0}{\theta_s - \theta_0} \tag{21}
\]

where \( \theta_0 \) and \( \theta_s \) are the residual and saturated water content, respectively [L^3 L^{-3}], \( \alpha \) is the inverse of the air-entry value [L^{-1}], \( n \) is a pore size distribution index > 1, \( \lambda \) is a pore-connectivity parameter (dimensionless), \( S_e \) is the effective water content (dimensionless), and \( m = 1 - 1/n \).

The inverse modeling technique proposed by Šimunek and van Genuchten (1996) is based on the minimization of an objective function \( \Phi \), describing the discrepancies between the simulated and observed values, which is performed in HYDRUS-2D using the Levenberg–Marquardt algorithm (Levenberg, 1944; Marquardt, 1963). To obtain a unique solution for the unknown parameters, it is advised to combine multiple-tension cumulative infiltration data with measured values of the initial and final water contents (Šimunek and van Genuchten, 1997). The potential of using inverse modeling techniques was supported by Ramos et al. (2006), who found this method to be a reliable alternative method for determining \( K(\psi) \) and \( \theta(\psi) \) curves on the basis of tension infiltrometer data.

For the IV method, two different parameter estimation sets were selected: a first one including all parameters \( \theta_0, \theta_s, \alpha, n, \lambda \), and \( K_\infty \) (the IV1 method) and a second one where \( K_\infty \) was not optimized and was taken equal to the value calculated by the WP method from the single-ring infiltrometer measurements (the IV2 method).

As an alternative for the equation of Wooding (1968), Šimunek and van Genuchten (1996) proposed an inverse modeling approach (the IV method) allowing indirect determination of the hydraulic parameters from transient tension infiltrometer data. In this approach, the unknown hydraulic parameters are estimated from observed cumulative infiltration data by numerical inversion of the Richards (1931) equation, which can be written for radial symmetric Darcian flow in terms of pressure potential \( \psi \) as

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This equation is subject to the following initial and boundary conditions (Warrick, 1992):

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\theta_0 + \left( \frac{\theta_s - \theta_0}{1 + \alpha \psi} \right) & b < 0 \\
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where \( \theta_0 \) and \( \theta_s \) are the residual and saturated water content, respectively [L^3 L^{-3}], \( \alpha \) is the inverse of the air-entry value [L^{-1}], \( n \) is a pore size distribution index > 1, \( \lambda \) is a pore-connectivity parameter (dimensionless), \( S_e \) is the effective water content (dimensionless), and \( m = 1 - 1/n \).

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For the IV method, two different parameter estimation sets were selected: a first one including all parameters \( \theta_0, \theta_s, \alpha, n, \lambda \), and \( K_\infty \) (the IV1 method) and a second one where \( K_\infty \) was not optimized and was taken equal to the value calculated by the WP method from the single-ring infiltrometer measurements (the IV2 method).
1999b) to reduce parameter autocorrelation. Since measurements with the tension infiltrometer were performed from wet to dry, estimates of the initial water content were obtained as 95% of the total porosity, whereas the water content at −1.18 kPa obtained from the measured soil water retention data was used as an initial estimate of the final water content in the inverse analysis. Initial model parameters for the van Genuchten–Mualem model were taken from Carsel and Parish (1988) based on the textural characteristics of each measurement location.

To evaluate the differences between the calculation methods and fitting procedures, and to allow comparison of the different approaches, three evaluation indicators were used, the mean error (ME), the RMSE, and the correlation coefficient ($\rho$):

\[
\begin{align*}
ME &= \frac{1}{u} \sum_{z=1}^{u} (y_{z1} - y_{z2}) \\
RMSE &= \sqrt{\frac{1}{u} \sum_{z=1}^{u} (y_{z1} - y_{z2})^2} \\
\rho &= \frac{\text{cov}(y_{z1}, y_{z2})}{\text{var}(y_{z1}) \text{var}(y_{z2})}
\end{align*}
\]

where $u$ (dimensionless) is the number of observations, $y_{z1}$ corresponds to the $z$th data pair of Data Set 1, and $y_{z2}$ corresponds to the $z$th data pair of Data Set 2.

### RESULTS AND DISCUSSION

#### Infiltration Measurements

**Ponded Head Infiltration**

Table 1 shows the mean values of the indicators used to evaluate the fitting procedure of the WP method for the undisturbed field tests and the <2-mm fraction measurements. It can be seen that the correlation coefficients for the WP model are very high for both data sets, with a mean overestimation <0.01 mm, indicating very good fits between modeled and observed data and a maximum error of 0.3%. Both the ME and RMSE were, on average, smaller for infiltration measurements conducted on the disturbed <2-mm fraction, which can be attributed to the homogeneous structure of the soil column and the absence of rock fragments.

Figure 4 compares the saturated hydraulic conductivity determined by the WP and RE1 methods. The $K_s$ values calculated with both methods agreed well, especially when considering $K_{s,t}$. The $K_{s,t}$ and $K_{s,f}$ values determined using the RE1 method differed, on average, by 6 and 36%, respectively, from the values obtained using the WP method, but no significant differences between the two calculation methods were found. When looking at absolute values, however, differences between the RE1 and WP methods were significantly smaller at the 95% confidence interval for the measurements on the fine fraction, with a maximum of 5 μm s$^{-1}$, compared with the field data set, where maximal deviations of up to 26 μm s$^{-1}$ were observed. This leads to the observation that, on average, only small differences between the RE1 and WP methods were observed, although the boundary conditions of the RE1 method were not met. Wu et al. (1999) also reported close agreement between the RE1 and WP methods for a series of simulated infiltration curves, as well as for a set of measured infiltration curves using a single-ring infiltrometer on a sandy loam soil.
**Tension Infiltrometry**

Table 2 shows the quality of fit between the observed and simulated cumulative infiltration data obtained with the tension infiltrometer for the V, IV1, and IV2 methods. The IV methods yielded better fits, with less error, for both data sets. The V method often yielded negative sorptivity values, causing measurements (39% for the field data set and 57% of the disturbed <2-mm fraction) to be discarded and thereby reducing its applicability. Differences between the two IV methods were small, but lower ME and RMSE values were observed for the IV1 method, which could be attributed to the increased flexibility in parameter optimization, with a better fit as a result. Nevertheless, the small differences in indices between the IV methods suggests that the saturated hydraulic conductivity calculated by means of the WP method were similar to the values estimated with the IV1 method using tension infiltrometer data. This is confirmed by Fig. 5, which depicts scatterplots of $K_s$ values obtained by the WP and IV1 methods. The mean absolute difference in $K_s$ of both methods was 5.97 and 2.29 μm s$^{-1}$ for measurements in the field and on the soil column, respectively. Although this difference was significant at the 95% confidence interval, differences were found acceptable given the fact that they were determined by using different measuring and calculation methods. Satisfactory agreement among $K_s$ values estimated by the IV1 method and the constant-head permeameter method was also reported by Ramos et al. (2006).

In Fig. 6, $K(\psi)$ values are compared for the different calculation methods and for both the field data and the soil column data sets. It should be noted that the V method was excluded from these graphics due to its much larger range, which was 9 to 11,517 μm s$^{-1}$ at −0.29 kPa, 94 to 2693 μm s$^{-1}$ at −0.59 kPa, and 26 to 994 μm s$^{-1}$ at −1.18 kPa. For the field data set, the A, LJ, and IV methods produced similar results for all pressure potentials, whereas the RE2 method resulted in a larger range for pressure potentials of −0.29 and −0.59 kPa. The results for the disturbed <2-mm soil fraction data set showed much less variation between calculation methods, and lower $K(\psi)$ values were obtained at each location after extracting the rock fragments. A paired-sample t-test ($\alpha = 0.05$) between the field data set and the disturbed data set revealed significant differences in all pressure potentials and all calculation methods, with the exception of the $K_s$ value calculated with the V method at the −0.29-kPa pressure potential due to its much larger range.

Since the real unsaturated hydraulic conductivity at the applied pressure potentials was not known, values calculated by the different methods were compared relative to the IV1 method and are presented in Table 3. From this table, it is clear that the largest ME and RMSE and lowest correlation coefficients were observed between the V and IV1 methods, resulting in $K(\psi)$ values obtained with the V method that differed fairly strongly with values obtained by the other methods. This was

<table>
<thead>
<tr>
<th>Method</th>
<th>Mean error $\times 10^{-5}$ m$^3$</th>
<th>RMSE $\times 10^{-5}$ m$^3$</th>
<th>$\rho$</th>
<th>n/rt</th>
</tr>
</thead>
<tbody>
<tr>
<td>V</td>
<td>2.628</td>
<td>3.020</td>
<td>0.99</td>
<td>27–35</td>
</tr>
<tr>
<td>IV1</td>
<td>0.266</td>
<td>1.309</td>
<td>0.99</td>
<td>41</td>
</tr>
<tr>
<td>IV2</td>
<td>0.627</td>
<td>1.839</td>
<td>0.99</td>
<td>41</td>
</tr>
<tr>
<td>Disturbed &lt;2-mm fraction soil column</td>
<td>0.814</td>
<td>0.866</td>
<td>0.99</td>
<td>19–32</td>
</tr>
<tr>
<td>IV1</td>
<td>0.230</td>
<td>0.828</td>
<td>0.99</td>
<td>43</td>
</tr>
<tr>
<td>IV2</td>
<td>0.257</td>
<td>0.968</td>
<td>0.99</td>
<td>43</td>
</tr>
</tbody>
</table>

*For V, minimum and maximum number of data points is given.*

![Fig. 5](image-url)
probably due to the occurrence of the negative sorptivity values obtained for many of the measured data sets (between 9 and 17 data sets for the field measurements, and 12 to 25 negative values for measurements on the fine fraction), resulting in a lower number of pairs on which the indices for the V method were calculated. Furthermore, if the sorptivity was positive, it often was more than a factor of 10 larger than the corresponding saturated hydraulic conductivity calculated with the WP method.

Among the other methods, the IV2 method produced the smallest differences and highest correlation coefficients, as could be expected due to its close relation to the IV1 method. The LJ method gave $K(\psi)$ values similar to the ones produced by the IV2 method, especially for the field data set. Evaluation indices for the $K(\psi)$ values calculated using steady-state methods (AN, RE2, and LJ) for measurements conducted on the <2-mm fraction soil column were similar, indicating that these methods performed comparably. Only for the field data set, the RE2 method seemed to produce $K(\psi)$ values that deviated rather strongly from those found with the IV1 method. Since differences between methods proved to be small, except for the V method, in the remainder of the study the IV method was preferred, since it determines the hydraulic conductivities on the whole transient infiltration data set instead of using only the steady-state infiltration rates (as in the case of the AN, RE, and LJ methods). The IV1 method, in turn, was chosen over the IV2 method because only small differences were observed between the two methods and because the IV2 method needs additional independent measurements to estimate the $K_s,t$ value.

Good agreement between the IV1 and the A method was also observed by Ramos et al. (2006), who determined $K(\psi)$ values on plots with sandy loam, loamy sand, sandy clay loam, and loam soils. Šimunek et al. (1998) compared the IV methods with a Wooding-based analysis, such as the A, RE2, and LJ methods, and found good agreement, indicating that the IV method results in calculated hydraulic conductivities in correspondence with the more traditional calculation methods.
was also confirmed by significantly higher infiltration rates at


<table>
<thead>
<tr>
<th>Method</th>
<th>Mean error</th>
<th>RMSE</th>
<th>$\rho$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$-0.29$ kPa</td>
<td>$-0.59$ kPa</td>
<td>$-1.18$ kPa</td>
</tr>
<tr>
<td>A</td>
<td>1.64</td>
<td>1.84</td>
<td>1.79</td>
</tr>
<tr>
<td>RE2</td>
<td>$-14.34$</td>
<td>$-4.53$</td>
<td>1.20</td>
</tr>
<tr>
<td>LJ</td>
<td>0.12</td>
<td>1.30</td>
<td>1.13</td>
</tr>
<tr>
<td>V</td>
<td>$-261.12$</td>
<td>$-52.35$</td>
<td>$-15.24$</td>
</tr>
<tr>
<td>IV2</td>
<td>$-0.41$</td>
<td>0.53</td>
<td>0.23</td>
</tr>
</tbody>
</table>

**Influence of Stoniness**

### Saturated Hydraulic Conductivity

The influence of stoniness on the infiltration process was investigated by means of three different identifiers for stoniness. Figure 7 shows that the best correlation was found for gravimetric rock fragment content, $R_{mt}$, when plotted against both $K_{s,t}$ (using the WP method) ($\rho = 0.57$) and the steady-state infiltration data ($q_{s,t}$) ($\rho = 0.63$), indicating a strong positive influence of embedded stone fragments on the infiltration process. A similar response was observed for the volumetric rock fragment content, resulting in slightly lower correlation coefficients ($\rho = 0.51$ and $\rho = 0.58$, respectively). These lower coefficients were attributed to small under- or overestimations in the total volume of the excavated soil due to diffuse boundaries when dealing with loose soil under sloped conditions. Correlation with the MWD of the stone fragments proved to be negative but not significant ($\rho = -0.14$ and $\rho = -0.19$, respectively), suggesting that the saturated hydraulic conductivity was lower at measurement sites containing larger rock fragments. Sauer and Logsdon (2002) found similar trends for the relation between infiltration and stoniness, but their correlation coefficients were less pronounced.

The overall positive relationship between $R_{mt}$ on the one hand and $q_{s,t}$ and $K_{s,t}$ on the other hand was attributed to an increase in porosity due to stone fragment content. Although not significant, a positive correlation ($\rho = 0.18$) between the coarse lacunar pore volume and $R_{mt}$ was observed (Fig. 8) suggesting that rock fragments create additional drainage paths that positively influence infiltration rates, as seen in Fig. 7. This was also confirmed by significantly higher infiltration rates at the undisturbed field sites in comparison with the infiltration rates measured on the <2-mm soil fraction. Other researchers, such as Ravina and Magier (1984) also found that stone fragment content could contribute significantly to the infiltration process, which was confirmed by Fié et al. (2002), who found a clear increase in lacunar pore volume with increasing stone content.

Figure 7 further shows higher scatter for higher stone fragment contents, indicating that rock fragments could also limit infiltration in some cases. This could be explained by a reduction in total soil volume through which water transport can take place and an increase in tortuosity. It is suggested that from a certain rock fragment content on, this decrease becomes more important than the extra coarse pore volume related to the rock fragments, and the observed saturated hydraulic conductivity will drop beyond this threshold. This could explain the fairly low $q_{s,t}$ and $K_{s,t}$ values sometimes observed at sites with a high (>0.30 kg kg$^{-1}$) rock fragment content (Fig. 7).

### Unsaturated Hydraulic Conductivity

Correlation coefficients for $K(\psi)$ (using the IV1 method) vs. rock fragment content and soil characteristics are presented in Table 4 for both the undisturbed measurements and those conducted on the disturbed <2-mm fine fraction for each of the applied pressure potentials. The patterns observed agreed very well with the pattern depicted in Fig. 7 for saturated water flow, although correlation coefficients between $R_{mt}$ and $K_{\psi,f}$ weakened as the pressure potential decreased. This indicates that the influence of rock fragments in the creation of additional drainage paths and increased (coarse lacunar) pore spaces was of greatest importance under saturated conditions and their role in conducting water diminished rapidly as the pressure potential dropped. Following the capillarity equation, it can be assumed that these pore spaces rapidly became too large for water transport as the matric potential became more negative. Sauer and Logsdon (2002) reported similar trends, but observed even negative correlations between rock fragment content and infiltration rates at the most negative pressure potential (~1.18 kPa), which contrasts with our observations.

Correlation coefficients between $K_{\psi,f}$ at all pressure potentials and $R_{mt}$ were significantly positive and higher than with $K_s$ (Table 4). The size of the rock fragments in terms of MWD correlated negatively to $K_{\psi,f}$, although not significantly, which is in agreement with the poor negative relationship found between the MWD and $K_s$. When comparing the relation between textural classes and both $K_{\psi,f}$ and $K_{\psi,\theta}$ contrasting patterns were observed. The value of $K_{\psi,\theta}$ had a
significantly positive correlation with the silt content at all applied suctions and a negative correlation with the sand content, whereas the $K_{\psi_{p,f}}$ had a mostly negative significant correlation with silt content and a positive correlation with sand content. The correlations between both $R_m$ and silt (negative) and silt and $K_{\psi_{p,t}}$ (positive) seem to suggest that the silt content plays an important role in stony soils at unsaturated pressure potentials. A common feature in these soils is coarse pores, created by stone fragments, contributing to conductivity under saturated conditions. At more negative pressure potentials, however, the smaller pores created by the fine earth fraction, especially silt, prove to be important in providing continuity.

Fig. 7. Field-saturated hydraulic conductivity and steady-state infiltration rate in relation to gravimetric rock fragment content, volumetric rock fragment content, and mean weight diameter. *Correlation coefficients were significant at the 95% level.
and conductivity. This would explain the increasing correlation between the silt fraction and the conductivity as the pressure potential becomes more negative, in clear contrast with the stone content, which exhibits a decreasing correlation with decreasing pressure potentials. The importance of the silt fraction on the unsaturated conductivities in stony soils was already mentioned by Sauer and Logsdon (2002), a conclusion strengthened by these results.

The relationship observed between \( R_m \) and \( K(\psi) \) seems to be in contrast with the findings of Mehuys et al. (1975), who reported only little differences between stony and non-stony samples when hydraulic conductivity was expressed as a function of \( \psi_p \). Their conclusion was drawn, however, considering \( \psi_p \) values between −5 and −5000 kPa, which is outside the range we measured. Furthermore, their hydraulic conductivity curves seemed to diverge as \( \psi_p \) approached −5 kPa, indicating differences in conductivity between stony and non-stony samples at the least negative pressure heads.

**Infiltration Variability in a Small Watershed**

**Saturated Hydraulic Conductivity**

Above, variability in infiltration rates was related to differences in the measurement technique as well as the calculation method used. Nevertheless, infiltration variability is also due to the sample location and its position in the landscape, and to the soil physical characteristics, such as textural influences and stoniness, as indicated above.

Saturated hydraulic conductivities (using the WP method), rock fragment content, and MWD are shown in Table 5 grouped by transect, textural class, and gravimetric rock fragment class. The \( K_{s,t} \) was higher at locations with a loam texture, but not significantly different from those with a sandy loam texture. This could only be explained by the slightly higher porosity in these soils (total porosity of loam soils was higher porosity in these soils) and, more importantly, the smaller dimension of the rock fragments and higher stone fragment content (Table 5). Mean observed \( K_{s,t} \) values were found to be significantly higher in the third transect than the first two, although textural differences were not different, and differences were therefore attributed to the significantly higher stone fragment content (both \( R_m \) and \( R_s \)) in the third transect, but with smaller stone fragment dimensions (smaller MWD).

Slope aspect was also found to be an important influence on infiltration rates, resulting in significant differences for measurements located on the western and eastern parts of the watershed and those located in the (dry) riverbed. Apparently, textural differences between the three slope locations were significant for both the sand and silt fractions (data not shown), partially explaining the infiltration differences. Additionally, as was observed with the different transects, stone fragment content (and volume) were also significantly different among the three slope aspects (Table 5). Since stone fragment content

<table>
<thead>
<tr>
<th>Data group</th>
<th>( K_{s,t} )</th>
<th>( R_m )</th>
<th>MWD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transect</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>21.1 ± 15.5 at</td>
<td>209 ± 112 a</td>
<td>26.9 ± 16.9</td>
</tr>
<tr>
<td>2</td>
<td>25.6 ± 23.0 a</td>
<td>262 ± 109 a</td>
<td>24.2 ± 13.7</td>
</tr>
<tr>
<td>3</td>
<td>51.6 ± 37.7 b</td>
<td>312 ± 172 b</td>
<td>20.5 ± 10.5</td>
</tr>
<tr>
<td>Texture</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Loam</td>
<td>38.7 ± 33.5</td>
<td>268 ± 115</td>
<td>20.8 ± 10.5</td>
</tr>
<tr>
<td>Sandy loam</td>
<td>23.3 ± 17.1</td>
<td>232 ± 116</td>
<td>26.8 ± 16.3</td>
</tr>
<tr>
<td>Slope aspect</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>West</td>
<td>16.8 ± 9.8 a</td>
<td>202 ± 111 a</td>
<td>25.0 ± 15.7</td>
</tr>
<tr>
<td>East</td>
<td>39.0 ± 29.6 b</td>
<td>277 ± 98 b</td>
<td>23.0 ± 13.2</td>
</tr>
<tr>
<td>Riverbed</td>
<td>89.8 ± 31.8 c</td>
<td>528 ± 201 c</td>
<td>25.5 ± 9.4</td>
</tr>
</tbody>
</table>

\( \dagger \) Data are mean ± standard deviation. Different letters indicate significant differences between data classes.

\[
28.4 ± 7.2 × 10^{-2} \text{ and } 27.6 ± 8.4 × 10^{-2} \text{ m}^3 \text{ m}^{-2} \text{ for sandy loam soils}
\]

and, more importantly, the smaller dimension of the rock fragments and higher stone fragment content (Table 5). Mean observed \( K_{s,t} \) values were found to be significantly higher in the third transect than the first two, although textural differences were not different, and differences were therefore attributed to the significantly higher stone fragment content (both \( R_m \) and \( R_s \)) in the third transect, but with smaller stone fragment dimensions (smaller MWD).

Slope aspect was also found to be an important influence on infiltration rates, resulting in significant differences for measurements located on the western and eastern parts of the watershed and those located in the (dry) riverbed. Apparently, textural differences between the three slope locations were significant for both the sand and silt fractions (data not shown), partially explaining the infiltration differences. Additionally, as was observed with the different transects, stone fragment content (and volume) were also significantly different among the three slope aspects (Table 5). Since stone fragment content

### Table 4. Correlation coefficients of calculated hydraulic conductivities using the method of Simunek and van Genuchten (1996) with saturated hydraulic conductivity \( (K_s) \) as a free parameter (IV1 method) for both the field and disturbed <2-mm fraction soil column data for pressure potentials of −0.29, −0.59, and −1.18 kPa vs. gravimetric rock fragment content \( (R_m) \), volumetric rock fragment content \( (R_s) \), mean weight diameter (MWD), and different <2-mm texture fractions.

<table>
<thead>
<tr>
<th>Property</th>
<th>Field data set</th>
<th>&lt;2-mm fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(-0.29 \text{ kPa})</td>
<td>(-0.59 \text{ kPa})</td>
</tr>
<tr>
<td>( R_m )</td>
<td>0.50*</td>
<td>0.43*</td>
</tr>
<tr>
<td>( R_s )</td>
<td>0.43*</td>
<td>0.36*</td>
</tr>
<tr>
<td>MWD</td>
<td>−0.15</td>
<td>−0.18</td>
</tr>
<tr>
<td>Sand content, 50–2000 μm</td>
<td>−0.21</td>
<td>−0.24</td>
</tr>
<tr>
<td>Silt content, 2–50 μm</td>
<td>0.33*</td>
<td>0.35*</td>
</tr>
<tr>
<td>Clay content, 0–2 μm</td>
<td>−0.12</td>
<td>−0.09</td>
</tr>
</tbody>
</table>

\* Correlation is significant at the 95% confidence interval.
Table 6. Mean unsaturated hydraulic conductivities \([K(\psi)]\) at different potentials, as calculated by the inverse modeling method of Šimunek and van Genuchten (1996) with saturated hydraulic conductivity as a free parameter (IV1 method), grouped by transect, textural class, slope aspect, and gravimetric stone fragment content \((R_m)\).

<table>
<thead>
<tr>
<th>Data group</th>
<th>(K(−0.29 \text{ kPa}))</th>
<th>(K(−0.59 \text{ kPa}))</th>
<th>(K(−1.18 \text{ kPa}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transect</td>
<td>(\mu\text{m s}^{-1})</td>
<td>(\mu\text{m s}^{-1})</td>
<td>(\mu\text{m s}^{-1})</td>
</tr>
<tr>
<td>1</td>
<td>11.5 ± 6.3</td>
<td>8.4 ± 4.0</td>
<td>4.5 ± 2.2</td>
</tr>
<tr>
<td>2</td>
<td>14.4 ± 7.7</td>
<td>9.9 ± 4.8</td>
<td>5.0 ± 2.3</td>
</tr>
<tr>
<td>3</td>
<td>22.8 ± 14.6</td>
<td>14.9 ± 9.0</td>
<td>7.4 ± 4.1</td>
</tr>
<tr>
<td>Texture</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Loam</td>
<td>19.4 ± 13.4</td>
<td>13.0 ± 8.3</td>
<td>6.5 ± 3.8</td>
</tr>
<tr>
<td>Sandy loam</td>
<td>12.8 ± 6.6</td>
<td>9.0 ± 4.4</td>
<td>4.8 ± 2.4</td>
</tr>
<tr>
<td>Slope aspect</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>West</td>
<td>9.7 ± 4.5 a†</td>
<td>7.0 ± 3.2 a</td>
<td>3.8 ± 1.9 a</td>
</tr>
<tr>
<td>East</td>
<td>20.6 ± 12.2 b</td>
<td>14.1 ± 7.6 b</td>
<td>7.1 ± 3.3 b</td>
</tr>
<tr>
<td>Riverbed</td>
<td>28.8 ± 3.9 b</td>
<td>16.0 ± 2.0 b</td>
<td>6.8 ± 3.0 ab</td>
</tr>
<tr>
<td>(R_m)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>&lt;0.15 kg kg(^{-1})</td>
<td>8.2 ± 4.6 a</td>
<td>5.9 ± 2.9 a</td>
<td>3.2 ± 1.4 a</td>
</tr>
<tr>
<td>0.15–0.30 kg kg(^{-1})</td>
<td>14.7 ± 7.7 b</td>
<td>10.2 ± 5.0 b</td>
<td>5.4 ± 2.5 ab</td>
</tr>
<tr>
<td>&gt;0.30 kg kg(^{-1})</td>
<td>23.6 ± 13.5 c</td>
<td>15.3 ± 8.3 c</td>
<td>7.4 ± 3.8 b</td>
</tr>
</tbody>
</table>

† Different letters indicate significant differences between data classes.

 Unsaturated Hydraulic Conductivity

Mean hydraulic conductivities calculated by means of inverse modeling (the IV1 method) at pressure potentials of −0.29, −0.59, and −1.18 kPa are shown in Table 6 for each of the different transects, textural classes, slope aspects, and gravimetric stone fragment content classes. When compared with the results from the saturated conditions, it can be readily observed that the same trends are conserved, although they are not always significant.

The fewest significant differences were found among transects and textural classes. Higher values at all pressure potentials were observed in sites along the third transect, however, which was consistent with the significantly higher saturated hydraulic conductivity observed along this transect (Table 5). With respect to soil texture, the observed \(K(\psi)\) values were, on average, always higher, although not significantly, for sample locations having a loam texture, consistent with saturated conditions. Considering the hillside orientation, values at the two least negative applied pressure potentials differed significantly between the west-oriented slope and the east-oriented slope, which could be attributed to stone fragment content differences. In fact, for these two pressure potentials, the mean \(K(\psi)\) of each of the slope orientations is positively correlated with the stone fragment content (data not shown), whereas for the lowest pressure potential, the differences between hydraulic conductivities become much smaller, reducing the influence of rock fragments. A similar trend can be observed when looking at \(K(\psi)\) influenced by the different rock fragment content classes, indicating significant differences when approaching saturated conditions. This suggests that, as pressure potential decreases, the \(K(\psi)\) of soils with a high \(R_m\) will become similar to those soils with fewer rock fragments.

Variability in calculated hydraulic conductivities under ponded conditions (using the WP method) and at three negative pressure potentials (using the IV1 method) along the three transects is illustrated in Fig. 9. This figure shows that the \(K_{st}\) was, in all but four locations, higher than the \(K(\psi)\) under negative pressure potentials, indicating that measurement results were consistent. Furthermore, patterns in hydraulic conductivities were conserved among different pressure potentials, although they became weaker at lower pressure potentials. This indicates that both the single-ring infiltrometer with constant ponding head and the tension infiltrometer were able to detect the same infiltration-influencing factors, such as the rock fragment content effects. For measurement sites located in the riverbed, however, serious drops in conductivity were observed as soon as water flow became unsaturated, indicating that only under saturated conditions did the largest pores (created by a larger stone fragment content) play an important role in the infiltration process.

CONCLUSIONS

In this study, variability in hydraulic conductivities was investigated, attributing differences to calculation methods, measurement techniques, and soil heterogeneity. Special attention was given to the influence of rock fragments on hydraulic conductivity through the creation of additional infiltration paths and larger pores.

When comparing calculation methods for saturated hydraulic conductivity measurements performed with a single-ring infiltrometer with constant ponding head, it was found that the methods of Reynolds and Elrick (1990) and Wu and Pan (1997) gave comparable results. Six different calculation methods were used to determine the \(K(\psi)\) values from infiltration measurements with a tension infiltrometer. The three steady-state calculation methods, i.e., those by Ankeny et al. (1991), Reynolds and Elrick (1991), and Logsdon and Jaynes (1993), performed similarly, especially for the <2-mm disturbed soil fraction, whereas the RE2 method resulted in slightly higher values for the field measurements. The inverse methods based on the transient infiltration data of Šimunek and van Genuchten (1996), IV1 with \(K_s\) as a free parameter and IV2 with \(K_s\) fixed at the single-ring infiltrometer value, gave similar results, which differed strongly from those obtained from the V method (Vandervaere et al., 2000). The latter proved difficult to apply to our data sets, resulting in
very high differences with other calculation methods and often resulting in negative sorptivity values. The inverse modeling method proposed by Šimunek and van Genuchten (1996) was found to be the best method for analyzing tension infiltrometer data, since it uses the whole measurement data set instead of just the steady-state infiltration rates, and since it also provides reliable estimates of the water retention curve (as discussed in Baetens et al., unpublished data, 2008).

The influence of stoniness under both saturated and unsaturated conditions proved to be positive, increasing hydraulic conductivities under all but the most negative pressure potentials. The same conclusion could be drawn when comparing the field data set to the disturbed data set without stone fragments, resulting in significantly higher saturated hydraulic conductivities for the field measurements due to the presence of stone fragments. This effect was probably due to an increase in coarse lacunar pores and was found to be strongest at sample locations with rock fragments having small MWDs.

The variability of hydraulic conductivities observed along three transects in an arid watershed proved to be significantly related to differences in the stone fragment content. Eastern, drier slopes and the riverbed positions showed higher infiltration rates linked to their higher stone fragment content. Higher stone contents, positively influencing infiltration, were also found on steeper slopes located higher in the watershed, which might be attributed to the more intense soil erosion processes observed at these locations.

This leads to the overall conclusion that stone fragment content is possibly the most important factor influencing infiltration rates in the soils present in our experimental watershed, which show only small textural differences. Since our watershed is representative of a large part of the arid drylands of northern Chile, this conclusion can probably be extended to a much larger region. This implies that stone fragment content should be taken into account when hydrologic processes are evaluated in those regions and when developing pedotransfer functions to predict hydraulic properties for arid soils.

ACKNOWLEDGMENTS

This research was funded by the Flemish Government, Department of Sciences and Innovation/Foreign Policy. We wish to thank Prof. Julio Ponce from the University of La Serena for assistance with the laboratory analysis.

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