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Estimation and Scaling of the Near-Saturated Hydraulic Conductivity

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The hydraulic conductivity in structured soils is known to increase drastically when approaching saturation. Tension infiltration allows *in situ* infiltration of water at predetermined matric potentials, thus allowing exploration of the hydraulic properties near saturation. In this study, the near saturated ($\psi \geq -0.15$ m) hydraulic conductivity was estimated both in the top- and sub-soil of three Norwegian soils. *A priory* analysis of estimation errors due to measurement uncertainties was conducted. In order to facilitate the comparison between soils and depths, scaling analysis was applied. It was found that the increase in hydraulic conductivity with increasing matric potentials (increasing water content) was steeper in the sub-soil than in the top-soil. The estimated field saturated hydraulic conductivity was compared with laboratory measurements of the saturated hydraulic conductivity. The geometric means of the laboratory measurements was in the same order of magnitude as the field estimates. The variability of the field estimates of the hydraulic conductivity from one of the soils was also assessed. The variability of the field estimates was generally smaller than the laboratory measurements of the saturated hydraulic conductivity.

Introduction

Generally, clay soils are thought to have a lower saturated hydraulic conductivity than soils with a coarser texture. However, soil structure may change this drastically, especially macrostructure induced by biological activity such as earthworms and plant roots, and physically processes such as shrinkage. Jarvis and Messing (1995), for instance, measured higher field saturated hydraulic conductivity in loam soils than in sandy soils. Reliable estimates of hydraulic properties are important for prediction of such processes as infiltration and redistribution together with migration of nutrients, pesticides and contaminants in the unsaturated zone. Still such knowledge is scarce for Norwegian agricultural soil conditions.

Determining the hydraulic conductivity is time consuming and tedious and therefore expensive. A common way is thus to indirectly determine this property by measuring and parameterizing the water retention curve and then to infer a statistical pore space model and calculate a relative hydraulic conductivity (Mualem 1986). To obtain the absolute hydraulic conductivity curve at least one measured hydraulic conductivity value is needed. Lately, tension infiltrometers have become common in field measurements of the hydraulic conductivitiy (Watson and Luxmore 1986; White and Sully 1992; Logdson and Jaynes 1993; Jarvis and Messing 1995; Lin *et al.* 1998; Shouse and Mohanty 1998). Tension infiltrometers allow *in situ* infiltration of water at specified matric potentials. It is an advantage to measure the near-saturated conductivity in the field to avoid any disturbance of the soil structure. An excellent review and explanation of the technique was given by White *et al.* (1992).

Measurements of the hydraulic conductivity is subject to considerable variability (Warrick *et al.* 1977). This variability depends on the inherent spatial and temporal variability, the scale at which the measurements are made, and the precision of the measurement technique. The concept of similar media introduced in soil physics by Miller and Miller (1956) has lead to a simplified description of the variability called scaling (Warrick *et al.* 1977) or functional normalisation if the method is based on regression analysis (Tillotson and Nielsen 1984). Hopman (1987) discussed several methods of scaling of soil hydraulic properties. Scaling analysis on tension infiltration data was used both by Jarvis and Messing (1995) and Shouse and Mohanty (1998). They encouraged to further research on this subject.

The intention with this investigation is to find reliable near-saturated hydraulic conductivities for three Norwegian loam soils and explore variations in the hydraulic properties with depth. In order to facilitate the comparison between different soils and depths, scaling analysis was used on the tension infiltration data.

Material and Methods

Theory

Wooding's (1968) classical solution of the Richards equation under steady state conditions for infiltration from a circular pond can be approximated well by

$$q = K(\psi)\left(1 + \frac{4}{\pi \alpha R}\right) \tag{1}$$

Here q is the steady state infiltration rate (volume/area × time), $K(\psi)$ the hydraulic conductivity at soil water matric head ψ (energy per weight) and R the radius of the circular infiltration area. The equation is derived under the assumption that the hydraulic conductivity is proportional to its rate of change, *i.e.* $K=\alpha^{-1}dK/d\psi$, which has the solution $K(\psi)=K^*\exp(\alpha(\psi-\psi^*))$ with K^* and ψ^* integration constants ($K^*=K(\psi^*)$). This corresponds to Gardner's (1958) model of the unsaturated hydraulic conductivity. Usually the matric head ψ^* is chosen as zero and K^* is then the field saturated hydraulic conductivity. The parameter α is a measure of the influence of the capillary force relative to the gravitational force on water transport in a soil (White and Sully 1992). Large values indicates that gravity dominates over capillary forces.

By measuring the steady state infiltration at two heads one can solve Eq. (1) with Gardner's model of the hydraulic conductivity and find an expression for α

$$\alpha = \frac{\ln(q_1) - \ln(q_2)}{\Delta\psi} \tag{2}$$

Here q_i is the steady state infiltration at matric head ψ_i and $\Delta\psi=\psi_1-\psi_2$. K^* can also be found by solving Eq. (1). The variance of the calculated parameters is estimated from ordinary error analysis (Squires 1988). The variance of a function of many variables, $y=f(x_1,x_2,..,x_n)$, is $\sigma_y{}^2=\Sigma_i^n(\partial f/\partial x_i)^2\sigma_x{}^2$ assuming uncorrelated errors, thus in our case

$$\sigma_\alpha^2 \equiv \frac{2}{\Delta\psi^2}\left((\frac{\sigma_q}{q})^2 + (\alpha\sigma_\psi)^2\right) \tag{3}$$

$$\sigma_{K^*}{}^2 \equiv K^{*2}\left((\frac{\sigma_q}{q})^2 + (\alpha\sigma_\psi)^2\right) \tag{4}$$

Here σ_i^2 with subscript $i=\{\alpha,K^*,q,\psi\}$ is the variance of the respective parameters. We see that increasing Δ_ψ reduces the variance of α and that the variance of the conductivity is proportional to the conductivity squared. Both are proportional to the relative error in the infiltration readings (σ_q/q) and in the matric head control ($\alpha\sigma_\psi$).

In order to obtain reference states of spatial varying conductivity scaling analysis was performed. In a Miller-similar media the hydraulic conductivity and the water retention is scaled by a microscopic characteristic length. The ratio of the microscopic length scale to a reference length scale enables us to relate the matric head and the hydraulic conductivity of any site j to a reference site r

$$\psi_j = \frac{\psi^r}{\lambda_j} \tag{5}$$

$$K_j = \lambda_j^2 K^r \tag{6}$$

The reference state and the probability density function of λ will determine (the stochastic property of) the hydraulic conductivity. However, Miller similtude implies constant porosity which is seldom met in real soils. When the scaling coefficients for the soil water potential and the hydraulic conductivity is completely independent and found by empirical methods, as for instance regression, the scaling analysis is called functional normalisation (Warrick *et al.* 1977; Tillotson and Nielsen 1984).

The procedure adopted here is to scale K by K_s and ψ by functional normalisation with the empirical scaling factor λ_j. The analysis is performed on logarithmic transformed data and combine elements from Jarvis and Messing (1995) and Shouse and Mohanty (1998).

The scaling parameter λ_j was found by minimising the total sum of squares from all replicates within one site, $SS = \Sigma_{j=1}^{m} SS_j$, where the sum of square from one replicate (SS_j) is taken over n measurements.

$$SS_j = \sum_{i=1}^{i=n} \left(\ln(K_{ij}) - \ln(K_i^r) - \ln(K_{(s)j}) + \ln(K_s^r)\right)^2 \qquad (7)$$

Here K_i^r is the reference hydraulic conductivity evaluated at the reference matric potential ψ_i^r, $K_{(s)j}$ is the saturated conductivity at site j and K_s^r the reference saturated conductivity. Initial estimates of K_i^r are obtained from linear regression on the original conductivity data. The scaling parameter λ_j is used to find the scaled matric potentials. A new reference curve is then determined using linear regression on the scaled data. This is repeated until further reduction in the total sum of squares is negligible. The scaling factors are normalised to have a mean value of one.

Soil Description and Experimental Procedure

Steady state infiltration rates were measured at three soils and two depths during the summer season in 1996 and 1997. The soils are located at Øsaker near Sarpsborg (soil 1), Bjørnebekk in Ås (soil 2) and at the field station at the Agricultural University of Norway (soil 3). All sites are situated in south-eastern Norway and the texture are clay loam, silt loam and loam soil respectively. Soil 1 is classified as a Typic Endoaqualf, soil 2 as a Typic Glossaqualf according to Soil Survey Staff (1994) and soil 3 as a Typic Haplaquept according to Soil Survey Staff (1975). Soil texture and content of organic carbon are shown in Table 1. Soils 1 and 2 were cropped with ce-

Table 1 – Texture of the three soils.

Soil	Soil 1		Soil 2		Soil 3	
Depth (cm)	0-20	20-40	0-20	20-40	0-20	20-40
Sand (2-0.06 mm)	24	15	11	4	28	8
Silt (0.06-0.002 mm)	40	36	62	60	45	71
Clay (<0.002 mm)	36	49	27	36	27	21
Organic C (%C)	2.2	1.9	1.4	0.4	4.3	1.6

Table 2 – Number of tension steps and replicates.

Soil	Soil 1		Soil 2		Soil 3	
Depth (cm)	0	30	0	30	0	30
Tension (n)	3	3	3	3	3	3
Replicates (m)	3	2	3	2	7	4

reals and soil 3 was fallow. Normal soil management is autumn ploughing and harrowing and sowing in the spring.

The infiltration was measured with tension infiltrometers from Soil Measurement Systems (AZ, USA) at predefined matric heads. The measurements were selected at ψ_1=-0.15 m, ψ_2=-0.06 m and ψ_3=-0.03 m, which were calibrated in the laboratory before and regularly during the measurement period. To ensure good contact between the soil and the infiltrometer disk a thin layer of sand on the surface was used with area equal to the infiltration disk (R = 10 cm). When cracks or large pores were open to the surface a permeable piece of clothing (Eijkelkamp geomembrane) was put on the surface in order to prevent filling the pores with sand. Table 2 shows number of tension infiltration measurements and number of sites for each soil and depth. At soil 3 the investigations were extended with measurements at 10 cm intervals down to 60 cm depth.

The hydrological parameters K^* and α are determined according to Eqs. (1)-(2). At each site α is calculated between ψ = -0.15 m and -0.06 m and between -0.06 m and -0.03 m and K^* is thereafter calculated at ψ^* = -0.15 m, -0.06 m and -0.03 m. The parameter uncertainty is estimated for each parameter according to Eqs. (3)-(4).

Undisturbed soil samples (100 cm^3) was taken from 2-7 cm depth and from 25-30 cm depth from all sites and the saturated hydraulic conductivity and the water retention curve were measured in the laboratory. Five replicates from soil 3 was used and four from the other soils. Saturated hydraulic conductivity was determined by the method of constant head (Klute and Dirksen 1986) with a pressure head of 0.025 m. The water retention curve was measured using a sandbox with matric heads of 0 m, -0.050 m, -0.20 m and -0.50 m and in a pressure cell with pressure heads of 1.0 m, 5.0 m, 10 m and 30 m. From the water retention measurements the parameters in the van Genuchten parameterization (Eq. (8)) of the water retention curve was determined by non-linear parameter estimation using the RETC program (van Genuchten *et al.* 1991)

$$\theta(h)-\theta_r = (\theta_s-\theta_r)(1+(\alpha_{vg}h)^n)^{-m} \tag{8}$$

Here $\theta(h)$ is the water content at suction $h(=-\psi)$ and θ_s and θ_r are saturated and residual water content respectively. The van Genuchten parameter α_{vg} is distinguished from the previous α by a subscript and m and n are other fitting parameters. In the fitting procedure, m ■ 1-1/n and all parameters are treated as fitting parameters.

The variability of the hydraulic conductivity is reported as the coefficient of variation (*CV*) of the scaled and the measured saturated hydraulic conductivities. We assume that both the hydraulic conductivity and the scaling factors are log-normal distributed. Parkin *et al.* (1988) presented a comparison of three methods to calculate mean, standard deviation and *CV* of log-normal distributed data. The method best suited to calculate *CV* is chosen here (the uniformly minimum variance unbiased estimators method). Only *CV*'s from soil 3 are calculated.

Results and Discussion

An attempt to measure the infiltration of free water was not very successful. The sand plate between the infiltration disk and the soil had to be absolutely horizontal. A small slope caused the water to funnel out on the sides. Further, when this condition actually was met, it was difficult to define a steady state infiltration, maybe because of small fluctuations in the atmospheric pressure. Hence the survey on the other soils were conducted only at sub atmospheric pressures (negative heads). Nevertheless, one interesting feature was discovered. Infiltration on a site with ants and ant burrows open to the surface was compared to a site close by without ant burrows. A major difference was observed only for free infiltration (atmospheric pressure). This difference between infiltration with and without biological macropores is one reason for the large variability of the saturated conductivity. Lodgson and Jaynes (1993) considered large macropores as not being predictive of hydraulic properties of the bulk soil matrix.

All the measurements with the tension infiltrometer were conducted during the summer season and the estimated hydraulic parameters only represent this period. No significant trends were found during the summer. However extreme conditions as draught or soil frost will surely affect the structure and thus the hydraulic conductivity. In addition neither of these models take hysteresis into account.

Fig. 1 shows steady state infiltration rates at different matric heads at depth intervals of 10 cm at one site on soil 3. We see that the infiltration generally decreases with depth. The measurements indicate a difference between top- and sub-soil. This was also reported by Lin *et al.* (1998) who explained it with decreasing biological activity, cracking and pedal development with increasing depth. Long-term cultivation may also induce differences between top- and sub-soils.

The estimated parameters in the van Genuchten model of the water retention curve are shown in Table 3. This model fitted well to all measurements (r^2 ranging from 0.93 to 0.98). The average saturated hydraulic conductivity measured on 100 cm^3 soil cores in the laboratory are shown in the same table. Both the arithmetic and geometric mean are given together with the range of measured values. Some samples from the subsoil hardly conducted any measurable quantity of water at all which results in low geometric means. The low values could be a result of weaker aggregate

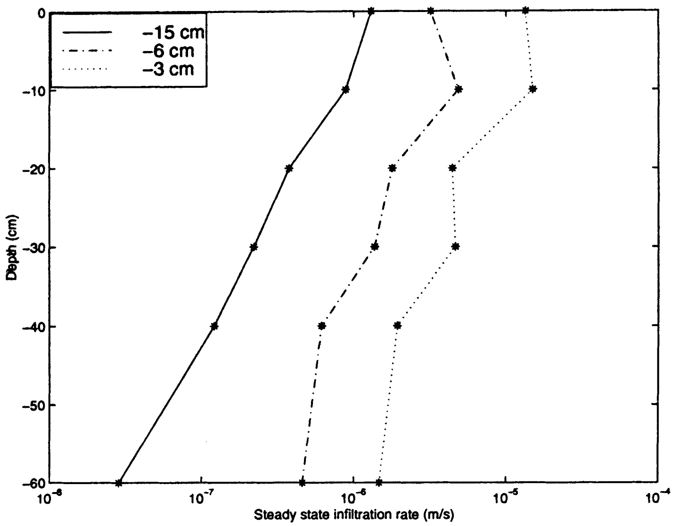

Fig. 1. Steady state infiltration rates at different matric potentials (m) as a function of depth in soil 3.

stability in the subsoil and perhaps breakdown of the aggregates. In addition we used a rather small gradient (2.5 cm) which perhaps not exceeds the threshold gradient required in some clay soils. At soil 2 and soil 3 the saturated hydraulic conductivity in the top-soil (2-7cm) are much higher than the conductivity measured in the subsoil (25-30 cm). The measurements varied a lot indicating the existence of macropores, but these pores were not necessarily open to the surface.

The large variation in the saturated hydraulic conductivity, Table 3, will clearly influence estimated hydraulic conductivities based on parameterizations as for instance the Mualem-van Genuchten model (Mualem 1986). Bouma (1985) estimates a representative elementary volume in clay loam to 10^3-10^4 cm^3 depending on soil

Table 3 - Measured saturated hydraulic conductivity (AM = arithmetic mean and GM = geometric mean) and the van Genuchten parameters for the water retention curve.

Soil	Soil 1		Soil 2		Soil 3	
Depth (cm)	2-7	25-30	2-7	25-30	2-7	25-30
K_s (μm/s) AM	61	77	69	37	8.0	5.7
K_s (μm/s) GM	6.7	14	8.0	0.71	4.2	0.64
Range K_s (μm/s)	0.041-291	0.26-157	0.29-531	0.028-139	3.60-16	0.028-25
θ_r	0.34	0.33	0.28	0.23	0.26	0.16
θ_s	0.58	0.51	0.48	0.43	0.46	0.37
α_{vg} (m^{-1})	13	8.3	11	3.9	5.8	9.6
n	1.56	1.20	1.34	1.13	1.55	1.23

structure. The area of infiltration with our tension infiltrometer is πR^2=314 cm^2 and the depth of influence is found by TDR measurements to be from about 10 to 20 cm during the measurement period. The volume measured by the tension infiltration is thus in the same range as the estimated representative elementary volume. This should reduce the spatial variability of the tension infiltration data compared to the laboratory data measured on small samples.

The term σ_q/q in Eqs. (3)-(4) is the relative error in the readings of the steady state infiltration rate. This is assumed to be small compared to the error in the pressure control $\alpha\sigma_\psi$. The infiltrometer was calibrated regularly in the laboratory, but the precision of pressure control in the field is limited by pressure fluctuations of about ± 1 cm (Jarvis and Messing 1995) which is taken as an estimate of σ_ψ. The relative errors in the hydraulic conductivity ($\sigma_{K*}/K*$) was thereby typically in the range of 0.05 to 0.5. In order to reduce these errors due to pressure fluctuations automatic logging of the applied pressure with pressure transducers could be advantageous.

Several authors have suggested that α is a function of ψ (Parlange and Hogarth 1985; Reynold and Elrick 1991; Jarvis and Messing 1995). In our investigations the calculated α from the top-soil tended to increase close to saturation (*i.e.* the interval between ψ=-0.06 m and ψ=-0.03 m). However the error in these estimates also increased (see Eq. 3). Most of the α from one interval was within one standard deviation from the other interval and all were within two standard deviations. The increasing α is also noticed in soil 3 (Fig. 4), but because of the increasing error and few data points a model with more parameters, as for instance a two-exponential model proposed by Jarvis and Messing (1995) or the piecewise continuous function in Mohanty *et al.* (1997), is not introduced.

The estimated parameters of the scaled reference curve, the logarithm of the field saturated conductivity (K^r_{fs}) and α^r together with its estimated uncertainty from the linear regression, are shown in Table 4. Further the goodness of fit of the reference curve, measured as the fraction of the total variance accounted for (r^2) is shown. Except for one case the reduction in sum of squares is considerable (from 86.4% to 99.7%). The soil with a low reduction in sum of squares (13.8%) has in fact little variations without scaling analysis. The results of the scaling analysis is shown graphical in Figs. 2 to 4.

Table 4 – Parameters of the reference hydraulic conductivity curve (ln(K^r_{fs}) and α^r with uncertainty estimates together with r^2 and the reduction in the total sum of squares (SS).

Soil	Soil 1		Soil 2		Soil 3	
Depth (cm)	0	30	0	30	0	30
ln(K^r_{fs}) (μm/s)	-12.9±0.1	-12.2±0.4	-12.8±0.1	-12.8±0.1	-11.8±0.1	-12.4±0.1
α (m^{-1})	15.3±0.9	26.8±4.1	11.2±1.1	18.8±0.3	17.9±0.7	21.5±0.8
r^2	0.975	0.913	0.952	0.999	0.996	0.988
Reduction in SS (%)	95.1	13.8	93.2	99.7	86.4	93.7

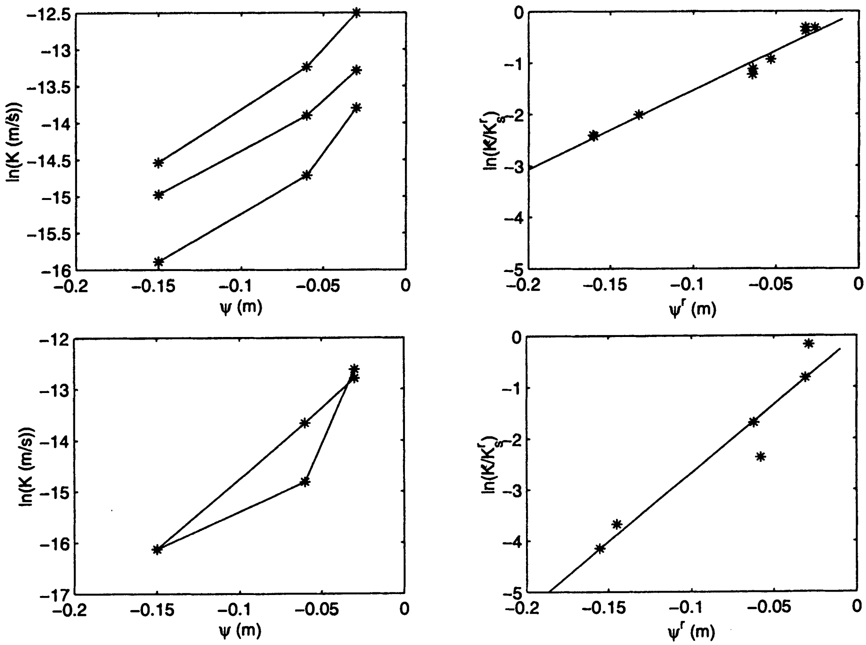

Fig. 2. Unscaled and scaled conductivity from soil 1.

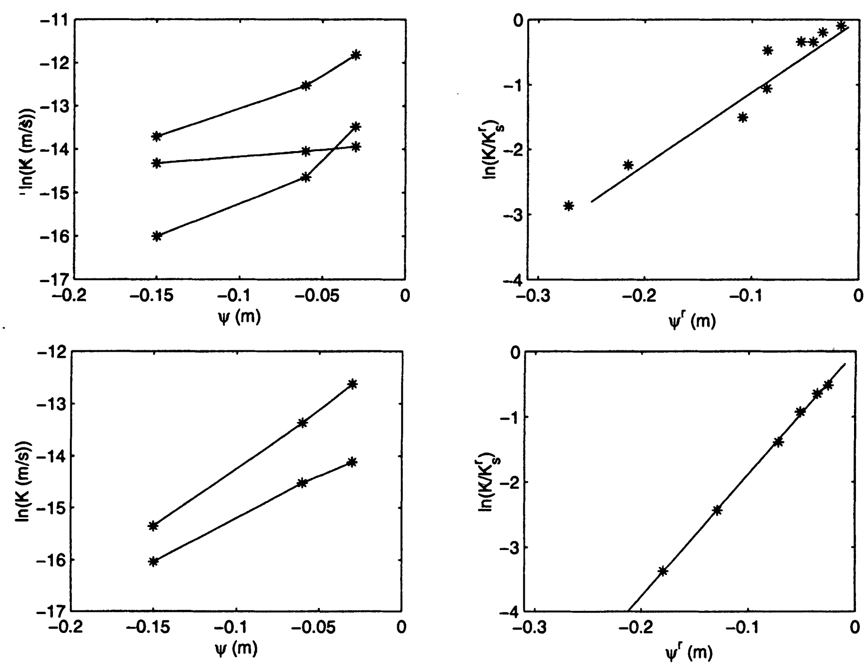

Fig. 3. Unscaled and scaled conductivity from soil 2.

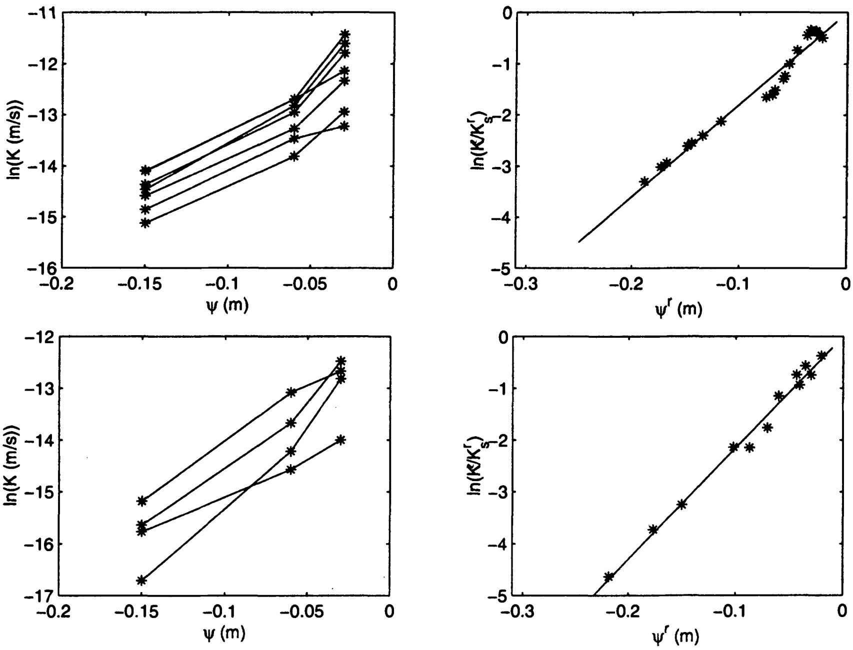

Fig. 4. Unscaled and scaled conductivity from soil 3.

The estimated field saturated conductivities from the tension infiltrometer measurements are almost the same for each soil and depth. However, there is a steeper rate of change with matric heads in the sub-soil than in the top-soil. This can be seen from the estimated αr in the same table.

For soils 1 and 2, the arithmetic mean of the saturated hydraulic conductivity measured in the laboratory (Table 3) are higher than the field estimates of the same parameter from the tension infiltrometer measurements while the geometric means are in the same range as the field estimates. That some laboratory measured conductivities are a lot higher than the field estimates are expected because by extrapolating from ψ = -0.03 m the largest pores are excluded. For soil 3 the estimated field saturated hydraulic conductivities are closer to the arithmetic means. The estimated field saturated hydraulic conductivities are close to the values of saturated hydraulic conductivity of comparable soils reported by Cassel and Parish (1988) and Rawls *et al.* (1982). But they are approximately a factor of 10 lower than the values estimated from tension infiltrometer by Jarvis and Messing (1995) on soils of comparable texture.

The coefficients of variation of the field saturated hydraulic conductivity and of the laboratory measured saturated hydraulic conductivities at soil 3 are shown in Table 5. We see that the coefficient of variation is higher on the measured small soil samples, especially in the subsoil. The variability of the field estimates are in the

Table 5 – Geometric mean (GM) and coefficient of variation (CV) of field saturated reference hydraulic conductivity (Kr_{fs}) and laboratory measured saturated hydraulic conductivity (K_s).

	Field estimates		Laboratory measurements	
Depth (cm)	$GM\ (Kr_{fs})$ (μm/s)	$CV\ (Kr_{fs})$ (%)	$GM\ (K_s)$ (μm/s)	$CV\ (K_s)$ (%)
5	7.5	56	4.2	63
30	4.1	52	0.64	224

same range as reported by Shouse and Mohanty (1998), but higher than in Jarvis and Messing (1995). Lin *et al.* (1998) reported *CV*'s of 25% to 35% on infiltration rates at negative matric potentials on a sandy loam while the *CV* of free infiltration was 57%. The *CV*'s of the the scaling coefficient of the matric potential were 17% and 34% in the sub- and top-soils, respectively. This is smaller than reported by Shouse and Mohanty (1998).

The estimated parameters in the van Genuchten model of the water retention curve are earlier shown in Table 3. With these parameters we can estimate the relative hydraulic conductivity using the Mualem-van Genuchten model (Mualem 1986). However we need to fix the level of this curve by at least one conductivity value. Clothier and Smettem (1990) recommend not to use the saturated hydraulic conductivity as a matching point because of the influence of the macropores and thus the large variability. This was also the conclusion of Jarvis and Messing (1995). Further Mohanty *et al.* (1997) reported several orders of magnitude difference between hydraulic conductivity measured with tension infiltration between 0 and -3 cm head. The reported field hydraulic conductivities are estimated on near saturated measurements and should thus represent the soil excluding the largest pores. This could be used as the matching point in the Mualem-van Genuchten model. The parameterization will then represent the bulk soil without the largest pores.

In areas with rain intensities usually smaller than the field saturated hydraulic conductivity the latter parameter is probably not the most important parameter to know exactly. Jarvis (1991) found that model predictions of leaching was rather insensitive to K_s', but that the hydraulic conductivity at the breakpoint between the macropore- and micropore-domain was a critical parameter.

Conclusion

Tension infiltration is found to be a convenient method to obtain near saturated hydraulic conductivities. The effect of measurement errors on a parametric representation of the unsaturated hydraulic conductivity is investigated. This is used to estimate the near saturated hydraulic conductivities in the top- and sub-soil from three Norwegian soils. Scaling of the measurements at each site and depth decreased the sum of squares of the residual errors significantly. The scaling analysis simplified comparison between soils and depth. It is found that the increase in the near saturated hydraulic conductivity with increasing matric head is steeper in the sub-soil than in the top-soil. The variability in the field estimates of the saturated hydraulic conductivity is smaller than saturated hydraulic conductivities measured on small undisturbed soil cores in the laboratory. The field estimates could also be used as a matching factor in the Mualem-van Genuchten model of the unsaturated hydraulic conductivity.

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