Kerstin Berglund

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LIC. AVHANDLING

SATURATED HYDRAULIC CONDUCTIVITY AS RELATED TO MACROPOROSITY IN CLAY SOILS

Ingmar Messing

Institutionen för markvetenskap
Avdelningen för lantbrukets hydroteknik

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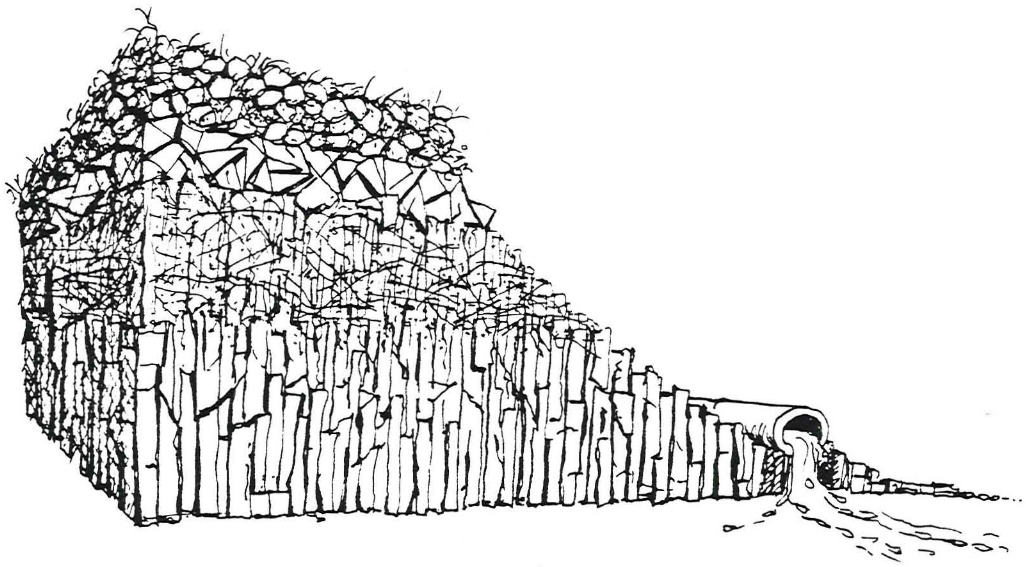

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PREFACE

This report is a thesis for the degree of "Agronomie licentiat". This degree is an autonomous part of a doctoral programme, consisting of 80 credit points, which is equivalent to two years of full time study beyond the M.Sc. in Agriculture. It is awarded on the basis of both course work and a dissertaion. Two papers are presented:

I. *Messing, I. 1989. Estimation of the saturated hydraulic conductivity in clay soils from soil moisture retention data* (in press in *Soil Science Society of America Journal*).

II. *Messing, I. 1989. Field-saturated hydraulic conductivity as a measure of structural variations in heavy clay soils* (submitted to *Journal of Soil Science*).

A general introduction to the subject and summaries of the papers with concluding remarks are given in the first section of the report.

Financial support by the Royal Swedish Academy of Agriculture and Forestry, is thankfully acknowledged. I want to thank Prof. Waldemar Johansson and Prof. Janne Eriksson for their support. Special thanks to Dr. Nicholas Jarvis for valuable advice and suggestions and for correcting my English.

The front page illustration was drawn by Sanna Krevers, Stockholm, and the figures in the text by Hans Johansson.

Uppsala, April 1989

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SATURATED HYDRAULIC CONDUCTIVITY AS RELATED TO MACROPOROSITY IN CLAY SOILS

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INTRODUCTION

The great impact of large pores (macropores), in terms of their size and configuration, on the magnitude of the saturated flow of water in soil is well recognized. Many theoretical and experimental studies have been carried out in an attempt to establish useful relationships between some measurable pore parameter on the one hand, and some parameter quantifying water flow in soil on the other. This task is particularly difficult in swelling and shrinking soils since the pore system changes continuously with time.

In this report two papers are presented with the aim of determining empirical relationships between macroporosity and the saturated hydraulic conductivity (K_s). The first paper relates an effective porosity (ϕ_e), which accounts for the pore volume fraction which contributes most to K_s, with laboratory K_s-values on completely swollen soil core samples. ϕ_e is defined as the difference between the porosity (ϕ) and a fixed point on the soil moisture characteristic curve, corresponding to a given arbitrary lower size limit for macropores. The second paper relates an estimated value for the total macroporosity (e_t), defined as the difference between ϕ and the field soil moisture content, with the measured field-saturated hydraulic conductivity (K_{fs}).

These simple relationships require very few soil parameters to predict K_s or K_{fs}, and therefore may be very useful in the development of two-domain drainage models for clay soils. The theoretical backgound to this subject is given below, followed by short summaries of the two papers.

Theoretical background

The saturated flow of water in soil is determined by the hydraulic gradient, which is an expression of the driving force, as well as by the shape and continuity of the pore system, which is the medium through which the water will pass. A general way to approximate the pore system is to regard the soil as consisting of a bundle of capillary tubes of equal or differing radii. The discharge in each capillary tube (**Q**) can be calculated from Hagen-Poiseuilles relation:

$$Q = \rho g/\eta \quad \pi r^4/8 \quad h/l \qquad (1)$$

where ρ= the density of the water, η= the viscosity of the water, **r**= the radius of capillary tube, and **h/l**= the hydraulic gradient. By knowing the distribution of capillary tube radii, the total flow in a capillary bundle can be calculated by summation.

In a real soil, however, the total flow of water is influenced by a complex geometry including phenomena such as "pore necks", "dead-end pores", and "pore tortuosity". A French engineer, Henri Darcy, by investigating the flow of water through sand filters (Darcy, 1856), derived a macroscopic flow vector (**Q'**) which represents the sum of all microscopic flow vectors in soil:

$$Q' = K_s \, A \, h/l \tag{2}$$

where K_s is the proportionality factor, named saturated hydraulic conductivity, and **A** is the cross-sectional area for flow. Eqs. (1) and (2) form today the general conceptual background for a quantitative treatment of water flow in soil. K_s is the general parameter characterizing the mean velocity of flow under a unit hydraulic gradient. For the interested reader, a general discussion concerning this subject is given in Hillel (1980, p. 166-172).

There have been many attempts to relate K_s to some easily measured soil properties. One of the most widely accepted theories to analytically relate pore parameters to K_s was developed by Kozeny (1927). He derived K_s for a capillary tube model from Hagen-Poiseuilles relation (Eq. (1)) and gave it a more general form, thereby obscuring the fact that it was, by derivation, applicable only to the capillary tube model. He introduced the concept of the specific area of the conducting medium (**S**), namely the total surface area of the solid fraction divided by the volume of the solid fraction. K_s may then be interpreted as (Ahuja et al., 1984):

$$K_s = C\phi^3 / S^2 \tag{3}$$

where **C**= a constant, and ϕ= the porosity. Carman (1937,1956) refined this equation by replacing **C** with C'/T^2 where **T** is the tortuosity, i.e. the actual microscopic flow path length of the fluid divided by the apparent macroscopic path length so that:

$$K_s = C'\phi^3 / (TS)^2 \tag{4}$$

This equation with T constant has been found to work well for uniform-grain sands but fails for materials with a wide range of pore sizes (Corey, 1977, p. 93). The most striking failure of formulas of the Kozeny-Carman type (Eq. (4)) occurs in soils with well developed structure. Large pores (i.e. macropores such as fissures, earthworm channels and root channels) developed in an otherwise dense matrix may increase the actual K_s considerably, whereas the increase in ϕ is small and the increase in S negligible. Eq. (4) would thus predict a much lower K_s than the actual. A more thorough discussion of the derivation and applicability of relations of the Kozeny-Carman type is given in Childs (1969, p. 179-201).

It may be shown from Eqs. (1) and (2) that for a pore system approximated by a bundle of capillary tubes of equal size, K_s is proportional to the square of the tube radius (e.g. Childs, 1969; p. 180). In a capillary bundle with a range of tube sizes it is thus the wider tubes that will contribute most to K_s. Likewise, in a real soil, it is the larger pores, the macropores, which determine the order of magnitude of K_s (e.g. Bouma & Wösten, 1979; Germann and Beven, 1981). To account for this, Ahuja et al. (1984) generalized the Kozeny-Carman equation (Eq. (4)), introducing the concept of an effective porosity (ϕ_e) which mainly contributes to the flow of water when the soil is saturated (Brooks and Corey, 1964). It was assumed that T and S in Eq. (4) decrease with increases in ϕ_e in proportion to some power of ϕ_e so that:

$$K_s = B\, \phi_e^n \qquad (5)$$

where B and n are empirical constants.

Brutsaert (1967) and Corey (1977, p. 42) have suggested that ϕ_e is equal to the porosity (ϕ) minus field capacity of a soil. The value of field capacity is not a precisely defined parameter. Germann and Beven (1981) gave an interpretation in which Ø of the soil is viewed as a two domain system, with a macropore domain superimposed on the micropore domain (Fig. 1). The water in macropores (mode 1) flows under gravity in the absence of capillary forces. In the micropores (mode 2), the water flow is not independent of capillary forces. The threshold between micro- and macropores, however, is a matter of convenience rather than of physical principle.

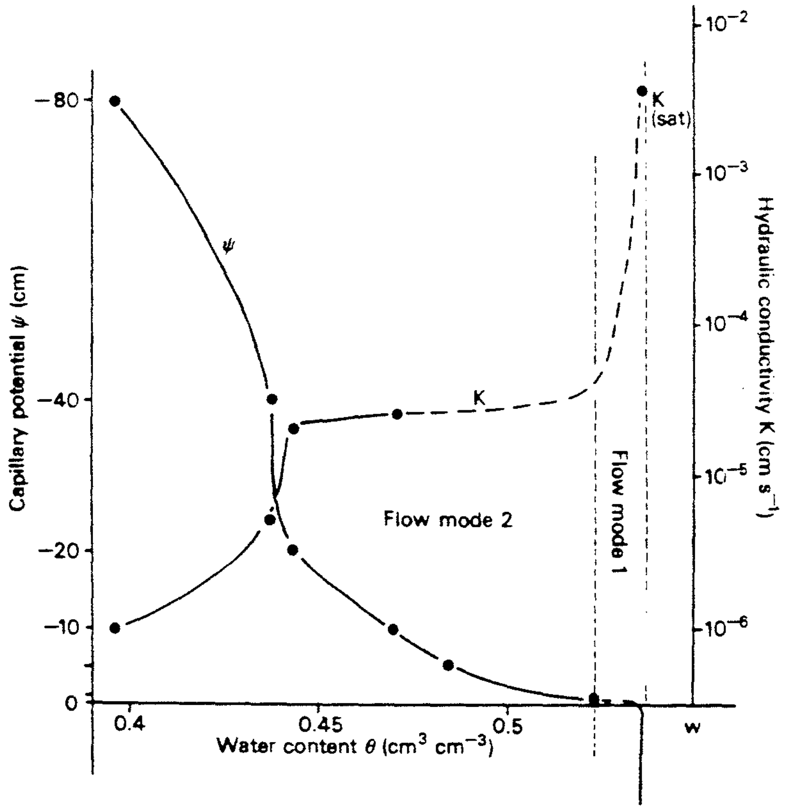

Fig. 1. The two proposed modes of water (Source: Germann and Beven, 1981).

Summary of paper I

Paper I investigates whether appropriate values can be determined for the exponent **n** in Eq. (5). Samples from around 500 horizons from 60 clay soil profiles within Sweden were investigated. Two statistical methods were utilized; i) Linear regression analysis to see whether **individual values** of K_s can be estimated with any degree of confidence from ϕ_e, and ii) comparison of cumulative frequency distributions of scaling factors for K_s with scaling factors for ϕ_e to see whether the **spatial distribution** of K_s can be estimated from ϕ_e. Results are presented for two choices of the definition of ϕ_e (ϕ minus soil water content at -30 kPa and -10 kPa respectively), and for the main clay soil groups as defined by Ekström (1927) (i.e. 15-25 %, 25-40 %, 40-60 %, and >60 % clay content).

In general, **n** seemed to vary from 2 to 3 when -30 kPa was chosen as the lower limit for ϕ_e. With -10 kPa as the lower boundary, **n** varied between 1.5 to 2.5. In the study of Ahuja et al. (1984) on silt loam, silty clay loam and silty clay soils, **n**-values varied between 3 and 4 with a similar, although not identical, experimental set-up. The variation in **n**-values may be partly due to the various pore size distributions and pore geometries in the different clay soil groups as affected by the soil structure, but may also be a result of the experimental techniques themselves. The relatively small size of the soil core samples, which may not account for the continuity of the macropore system **in situ** and which is not always representative of the size of the structural features, may limit somewhat the application of the results. It is thus believed that before Eq. (5) may be applied in practice on structured clay soils, the empirical parameters in the relationship must be derived on larger soil samples.

The results from the cumulative frequency distribution analysis agreed remarkably well with Ahuja et al. (1984) with **n** lying around 4 to 5. Where data on the spatial distribution of ϕ_e exists, a few measurements of K_s at sites having scaling factors near 1.0 would therefore yield an average scaled mean hydraulic conductivity (K_{sm}) and the standard deviation associated with this value (Ahuja et al., 1984). To compare the effects of two different land uses or tillage systems, say on fields I and II of a soil series, a ratio of scaled mean K_s, K_{sm} in terms of ϕ_e for site i (ϕ_{ei}) may be used:

$$\left[\left(\sum_1^N \phi_{ei}^{n/2}\right)^2\right]_I \Big/ \left[\left(\sum_1^N \phi_{ei}^{n/2}\right)^2\right]_{II} \qquad (7)$$

However, the entire distribution of scaling factors as shown in Fig. 2 in paper I may be just as important for comparative purposes (Ahuja et al., 1984).

Summary of paper II

In paper II, the dynamic properties of the macroporosity were recorded in two clay soils over one season by means of measurements of infiltration into auger holes. The term field-saturated hydraulic conductivity (K_{fs}) was used instead of K_s to characterize the steady infiltration rates, since field methods do not generally measure the hydraulic conductivity in completely saturated soil. Knowledge of K_{fs}-values is often preferred, since few natural or man-made infiltration processes in the field produce complete saturation (Reynolds and Elrick, 1985).

The dynamic nature of the macropore system in clay soils is characterized by shrinkage and swelling in response to climatic conditions. The above discussion in the sections **Background** and **Summary of paper I** mainly concerns soils in an extreme condition, that is saturated and fully swollen, so that the hydraulic conductivity may be regarded as a true saturated value (K_s). In this state, most planar cracks are closed and the remaining conducting pores (fissures and biotic macropores) represent the stable macroporosity (e_s). If drying (e.g. evaporation, root water uptake, drainage) influences the soil, then planar cracks will form, opening up the system, connecting otherwise separate (i.e. dead-end) biotic macropores, and hence increasing the effective macroporosity. According to a simple model of Jarvis and Leeds-Harrison (1987), the total macroporosity (e_t) depends on the moisture content (θ) of the bulk soil:

$$e_t = e_s + (\theta_f - \theta) \tag{8}$$

where θ_f is the field capacity. In this model normal shrinkage is assumed for the aggregates and vertical shrinkage is neglected. In paper II, e_t (derived using Eq. (8)) is related to the measured field-saturated hydraulic conductivity (K_{fs}) using an expression very similar to Eq. (5):

$$K_{fs} = B' e_t^{n'} \tag{9}$$

where B' and n' are empirical constants.

It is noteworthy that the value of the exponent n' in Eq. (9) obtained in the experiments described in paper II was 2.8, a value which lies within the range for n-values in Eq. (5) in paper I (i.e. 2 to 3). In addition, the intercepts B and B' were also within a similar range.

General conclusions and recommendations for future research

It seems fruitful to continue to relate macroporosity to K_s and K_{fs} as in Eqs. (5) and (9). The determination coefficients for the relationships, though, are fairly low. In paper I, between 25 and 50 % of the variance in K_s was accounted for by ϕ_e alone. The relatively low correlation is partly explained by the small size of the soil core samples, which therefore are not representative of the size of the structural features. It is also partly explained by the geometric complexity of the pore system which affects K_s and ϕ_e respectively in different ways. For this reason it is desirable in the future to perform measurements on larger soil core samples which give values appropriate to the representative elementary volume of soil. This would decrease the variance and possibly justify the relationships. Methods for quantifying the geometric complexity would also be desirable, although very difficult.

In paper II, 53 % of the variance of K_{fs} was accounted for by e_t. Here, a more appropriate method to measure infiltration rate from an auger hole into the surrounding soil would be desirable. The so called Guelph permeameter (Reynolds and Elrick, 1985) would better correspond to the requirements for determining K_{fs} than the inversed auger hole method used here. Recording of the actual soil moisture tension by means of tensiometers around the auger hole at the time of measurements would possibly be a more useful measure of soil moisture status than the initial soil moisture content as used in paper II.

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ESTIMATION OF THE SATURATED HYDRAULIC CONDUCTIVITY IN CLAY SOILS FROM SOIL MOISTURE RETENTION DATA

INGMAR MESSING

In press in *Soil Science Society of America Journal*

ABSTRACT

A generalized Kozeny-Carman equation was employed to relate saturated hydraulic conductivity (K_s) to effective porosity (ϕ_e) in clay soils. The equation is of the form $K_s = B\ \phi_e^n$, where **B** and **n** are empirical constants. ϕ_e was defined as porosity (ϕ) minus soil water content at -30 kPa. Around 500 horizons from 60 soil profiles in Sweden with clay content >15 % were analysed. The value of **n** was tested with two statistical methods. According to the first method, a regression test on the log-log relationship between K_s and ϕ_e with least-squares lines fitted through the data, the exponent **n** was found to lie around 2 for soils with high clay content (>40 %), and 3 for soils with lower clay content (15-40 %). The correlation coefficients were moderate, ranging from 0.51 to 0.71 for different clay soil groups. In the second method, the cumulative frequency distributions of scaling factors for K_s determined from measured values of K_s were compared with scaling factors for K_s determined from ϕ_e. Reasonable agreements were obtained with **n**=4 to 5 for soils with clay content >40 %, and with **n**=5 to 6 for soils with clay content 15-40 %. The second method gave **n**-values which correspond well with other recent published work. The results indicate that the spatial distribution of K_s is estimated reasonably well from ϕ_e-values, whereas the individual values of K_s are not estimated as well.

INTRODUCTION

The occurrence of a continuous system of medium- to large- size pores, here called macropores, is of special importance for the initial drainage of clay soils. In water-saturated soils the water flow in the macropores can be characterized by the saturated hydraulic conductivity (K_s), which may be highly variable spatially both in vertical and horizontal dimensions. The possibilities of using the soil moisture retention curve for determination of K_s is of general interest since retention data is fundamental in many soil studies. Ahuja et al. (1984) stressed the need for approximate methods for hydrological characterization of spatially variable soils. They related the macroporosity distribution in the field to the distribution of K_s in two different soils with the use of scaling theory. The macroporosity was characterized by the effective porosity (ϕ_e), which can be determined from soil moisture retention data. The objective of the study outlined below is to test this approach by applying it to a large number of clay soil profiles distributed over the whole of Sweden. The study reflects the general applicability of the theory rather than taking into account spatial variability over specific fields. In particular, a number of factors which affect the value of the exponent in the K_s-ϕ_e relationship are investigated. These factors include i) choice of statistical method, ii) definition of ϕ_e, and iii) clay content.

BACKGROUND

Bouma and Wösten (1979) showed with breakthrough-curves and dye-tracing experiments on two soils with different macrostructure, that most of the flow of water in saturated soil occurred in macropores which represented only a small proportion of the total pore volume. Brooks and Corey (1964) introduced the concept of an effective porosity (ϕ_e). ϕ_e accounts for the pore volume fraction which contributes most to the flow of water when the soil is saturated. Ahuja et al. (1984) used this concept and generalized the Kozeny-Carman equation (Carman, 1937 and 1956) to

$$K_s = B\, \phi_e^n \tag{1}$$

where **B** and **n** are empirical constants. A simple capillary bundle model in which ϕ_e consists of pores of equal size, would predict that K_s increases with the square of ϕ_e (Germann and Beven, 1981). For the more general case, when a range of pore sizes and a complex geometry of the pore system occurs, the value of **n** will be more difficult to determine. Brutsaert (1967) stated from a theoretical viewpoint, that the value of the exponent corresponding to **n** for unsaturated conductivity as a function of effective saturation increased as the range of pore sizes increased. Corey (1954) showed experimentally that the value of this corresponding exponent varied for different consolidated porous rocks within a narrow range. A typical value was 4. Ahuja et al. (1984) assumed that, as an extension of these findings, the exponent **n** in Eq. (1) also varies within a similar narrow range.

The spatial distribution of K_s can be characterized in terms of scaling factors (Warrick et al., 1977; Simmons et al., 1979), based on an extension of the concept of similar media (Miller and Miller, 1956)

$$K_{si} = K_{sm}\, \alpha_i^2 \tag{2}$$

where K_{si} is K_s at a certain horizon at site **i** within a soil type, K_{sm} is a scaled mean hydraulic conductivity over the soil type, and α_i is a scaling factor for site **i**. Ahuja et al. (1984) derived the expressions for two corresponding scaling factors from Eq. (1), Eq. (2), and expressions for K_{sm} as

$$\alpha_i = N\,(K_{si})^{1/2} \;/\; \sum_{1}^{N}(K_{si})^{1/2} \qquad (3a)$$

and

$$\alpha_i = N\,\phi_{ei}^{n/2} \;/\; \sum_{1}^{N}\phi_{ei}^{n/2} \qquad (3b)$$

where **N**= number of sites, and $\phi_{ei} = \phi_e$ for site **i**. It should be noted that Eq. (3b) contains only the constant **n** of Eq. (1). They compared the frequency distribution of scaling factors for K_s obtained from K_s-measurements (Eq. (3a)) with the distributions of scaling factors for K_s determined from ϕ_e (Eq. (3b)) in two widely different soils with different soil horizons. They concluded that Eq. (1), with the exponent **n** either 4 or 5, was successful in characterizing the spatial distribution of K_s from ϕ_e-measurements. The results originated from measurements on undisturbed soil core samples, 7.5 cm high and 7.5 cm in diameter.

MATERIALS AND METHODS

The soil data used in this study was taken from Andersson and Wiklert (1977) and Wiklert et al. (1983). Sixty sites from glacial clay soils distributed over the whole of Sweden were chosen (Typic, Fluventic, and Aquic Eutrochrepts; Typic, Fluventic, and Aquic Dystrochrepts). Swedish clay soils have been classified into four groups according to Ekström (1927). The heavy clay soils, 40-60% clay, are generally strongly structured and highly permeable in the subsoil due to regular frost and drying actions. However, a plough pan may reduce infiltration rate during rainy periods and the topsoil may be compacted during parts of the season by heavy traffic. The light clay soils and to a certain extent the medium clay soils, 15-25% and 25-40% clay respectively, have a weaker structure due to the lower content of colloidal clay. This will decrease the permeability in the subsoil. In the very heavy clay soils, >60% clay, swelling and shrinking processes govern the permeability during the season. Long dry periods will induce excessive drainage, whereas long rainy periods will give waterlogged conditions. It is thus clear that clay content has a strong influence on soil structure and therefore is also likely to affect the nature of the relationship between ϕ_e and K_s.

Undisturbed soil core samples, 10 cm high and 7.2 cm in diameter, were taken at each site at 10 cm intervals from the surface to a depth of one meter. The arithmetic mean of 2 to 4 core samples at each level was determined. Most of the cores were sampled during the late summer and autumn after a period of wetting by rain. K_s was measured in the laboratory after prolonged wetting with a constant head permeameter under a unit hydraulic gradient (Andersson, 1953). Water retention at various pressure heads including -10 kPa and -30 kPa was determined using tension table apparatus. The porosity (ϕ) was calculated from the dry bulk density and the density of solids.

Macroporosity was in this study characterized by the effective porosity (ϕ_e). ϕ_e was defined as the total porosity (ϕ) minus soil water content at -30 kPa. In order to investigate the effect of changing this definition, ϕ_e was additionaly calculated with -10 kPa as lower limit, the latter being regarded as the field capacity value in Sweden. Denning et al. (1974) criticized the procedure of estimating the pore size distribution from soil moisture retention data and stated that the capillary model was irrelevant for clay soils. Bouma and Wösten (1979) stated that the equivalent pore size distribution was not necessarily the same as the effective pore size distribution. The latter is determined by means of breakthrough curves and reflects the pore sizes which are effective for water flow better than the concept of equivalent pore sizes. Nevertheless, the definition of ϕ_e from the soil moisture retention curve is of great interest due to its widespread applicability and usefulness and was therefore utilized in this study.

Data from 495 soil layers from the 60 soils were processed statistically (light clay, 67 soil layers; medium clay, 148; heavy clay, 182; very heavy clay, 98). A few negative or zero-values of $\emptyset_e$ and K_s were eliminated in the calculations. Two statistical methods were utilized in order to determine the value of **n** in Eqs. (1) and (3b). First, least-squares line regression tests were carried out on the log-log relationship between K_s and $\emptyset_e$. Confidence intervals (P<0.05, Student's t, unequal variances) were calculated for the differences between the means of pairs of clay soil groups for **n** and **log** B respectively (Eq. (1)). Second, the cumulative frequency distribution of scaling factors for K_s obtained from K_s-measurements using Eq. (3a) was compared with the distributions of scaling factors for K_s determined from $\emptyset_e$ (Eq. (3b)). The results are presented in fractile diagrams with the theoretical probability variable $(Y-\bar{Y})/\sigma_y$. The symbol **Y** stands for the transformed values of the scaling factor (α) (e.g., $\log\alpha$ or others) that would have a normal distribution. The $\bar{Y}$ and σ_y stand for mean and standard deviation.

RESULTS AND DISCUSSION

Regression analysis showed that K_s was positively related to $\emptyset_e$ (P<0.001) when the lower limit of $\emptyset_e$ was defined by -30 kPa (Fig. 1). The correlation coefficients ranged from 0.51 to 0.71 for different clay soil groups. The relationship was strongest for light and medium clay soil groups. The slopes of the least-squares line fitted through the data, the **n**-constants, ranged from 1.78 for heavy clay to 3.01 for light clay soil. The differences between clay soil groups in the values of the **B**- and **n**-parameters were tested and the results are presented in Table 1. Concerning the intercept **B**, the mean log-values were not different (P<0.05). The **n**-parameters, however, seemed to be affected by the variation in clay content. For example, values for the light and medium clay soils (mean values close to 3) differed from the heavy clay soil (mean value close to 2) ($P<0.05$).

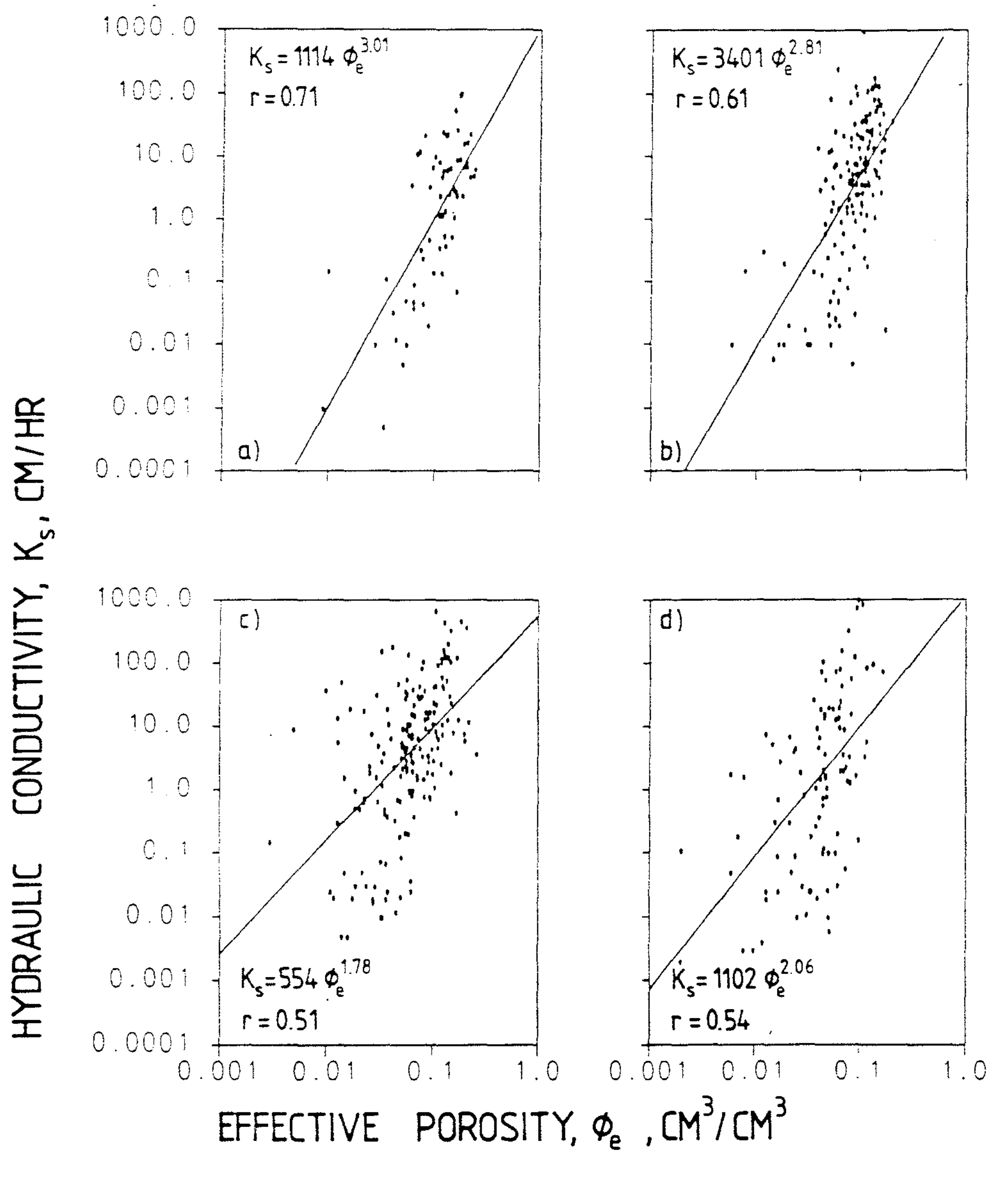

Fig. 1. Saturated hydraulic conductivity as a function of effective porosity for soil cores of Swedish clay soils. a) 15-25% clay, b) 25-40% clay, c) 40-60% clay, d) >60% clay.

Table 1. Differences between clay soil groups in the parameters **n** and **log B** in Fig. 1

	Confidence intervals (P<0.05)	
	n	**log B**
Light clay versus medium clay	0.20 ± 0.94	0.48 ± 1.00
Light clay versus heavy clay	1.23 ± 0.85	0.31 ± 0.93
Light clay versus very heavy clay	0.95 ± 0.96	0.01 ± 1.19
Medium clay versus heavy clay	1.03 ± 0.73	0.79 ± 0.87
Medium clay versus very heavy clay	0.75 ± 0.86	0.49 ± 1.14
Heavy clay versus very heavy clay	0.29 ± 0.77	0.30 ± 1.08

The reasons for the variable **n**-values may be partly explained by the various pore size distributions and pore geometries in the different clay soil groups as affected by the structure. **n**-values obtained by regression analysis often lie between 2 and 3 for many field soils, which is close to the theoretical value of 2 expected for discrete macropores of equal size (e.g. Germann and Beven, 1981). However, the values of the derived parameters may also be affected by the experimental techniques themselves. For example, in the laboratory the soil core samples swell when subjected to initial water saturation and then shrink during the subsequent suction steps and the ultimate drying at 105 ^{o}C. The errors introduced by ignoring such volume changes when determining ϕ_e will be relatively larger at low

$\emptyset_e$-values, and also at higher clay contents since the swelling and shrinking processes become more pronounced as clay content increases.

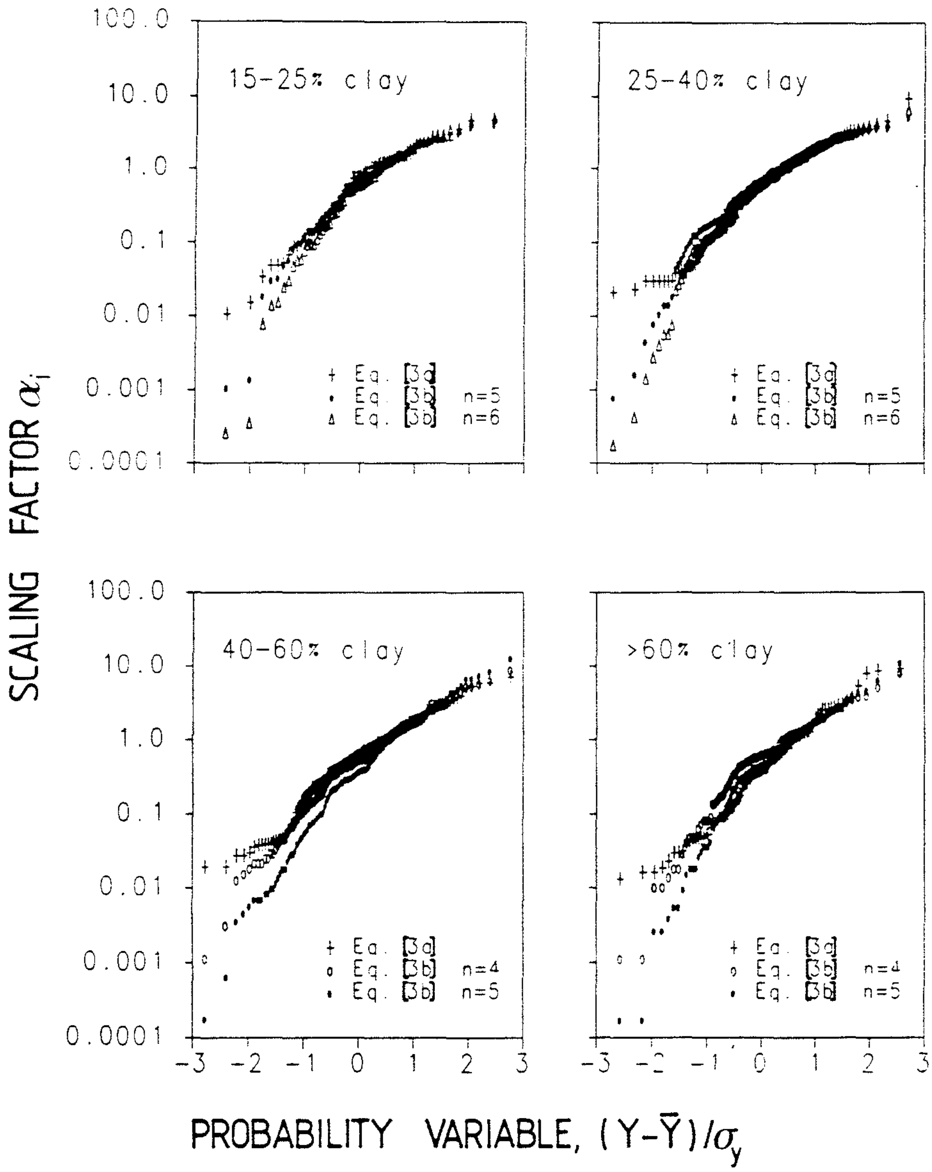

Fig. 2. Fractile diagrams of the logarithms of K_s scaling factors compared with that of the $\emptyset_e$-derived scaling factors for K_s of Swedish light, medium, heavy and very heavy clay soils. See text for the choice of the value of **n** and for the definition of the probability variable.

In Fig. 2, cumulative frequency distributions of the logarithm of scaling factors (α_i) obtained from ϕ_e by using Eq. (3b) are compared with the distribution of the logarithm of α_i obtained from K_s-measurements (Eq. (3a)). The best obtained fits are shown here. It is clear that the heavy and very heavy clay soil groups give good results for **n**=4 to 5, whereas **n**=5 to 6 is more appropriate for light and medium clay soil groups. The divergences in the left-hand part of Fig. 2 between the distributions, can to some extent be explained as errors in the calculation of fractiles, which can be especially large at either end of the distribution (Hald, 1952, p. 138-139). Additionally, errors in ϕ_e will be relatively larger at low ϕ_e-values than at high values due to the swelling and shrinking processes in the laboratory (see above discussion). Apparently, the cumulative frequency distributions yield higher **n**-values than the regression equations. Ahuja et al. (1984) indicated that the slope of a log-log fit is very sensitive to scatter in the data and to outlying data points, as well as to the range and number of the data points. The regression approach is consequently less reliable, whereas the cumulative frequency distributions yield stable **n**-values around 4, 5, and 6 when ϕ_e is defined by -30 kPa.

Cumulative frequency distributions were also determined with the lower limit for ϕ_e defined by soil water content at -10 kPa pressure head instead of -30 kPa. In this case the distributions showed agreement with lower **n**-values (light and medium clay soil groups **n**=4, heavy clay soil group **n**=3, very heavy clay soil group **n**=5). Regression tests yielded **n**-values ranging from 1.44 to 2.24 for the different soil types, and correlation coefficients between 0.53 and 0.66 (Table 2). This may be compared with Fig. 1 where the values of **n** were higher and the correlation coefficients within the same range. Regression analyses and cumulative frequency distributions thus gave similar results in that a broader definition of ϕ_e yielded a higher value of **n**. For clay soils of low K_s, the pores between -10 and -30 kPa may indeed be contributing significantly more to K_s (in relative terms) than for soils of high K_s, where both the relative and even the absolute contribution of these pores is probably much less. This may explain why the slope **n** had lower values when -10 kPa was used as the lower limit instead of -30 kPa.

Table 2. Parameters in Eq. (1) when $\emptyset_e$ is defined as porosity ($\emptyset$) minus soil water content at -10 kPa pressure head

Clay content %	Number of soil layers	B	n	r
15-25	69	468	2.24	0.66^{***}
25-40	155	490	1.81	0.60^{***}
40-60	185	372	1.44	0.56^{***}
>60	92	1950	2.05	0.53^{***}

*** Correlation coefficient significant at P=0.001.

CONCLUSIONS

The two statistical procedures which were utilized in order to define an appropriate value of **n** in Eqs. (1) and (3b) gave different results. The first method, where least-squares lines were fitted to plots of log K_s versus log $\emptyset_e$ showed **n**-values lying around 3 for soils with 15-40 % clay content, and around 2 for soils with >40 % clay content, when the lower limit for $\emptyset_e$ was defined by -30 kPa (Fig. 1). These values may be underestimated due to the difficulties in determining low $\emptyset_e$-values on swelling and shrinking soils. The **n**-values obtained may also be strongly affected by outliers in the data. With the second method, in which cumulative frequency distributions of scaling factors from Eqs. (3a) and (3b) were compared (Fig. 2), **n** varied between 4 and 6 for different clay soil groups. This corresponds well to the values proposed by Ahuja et al. (1984) where **n** ranged between 4 and 5. However, when the lower limit for ϕ_e was defined by -10 kPa, **n**-values were generally lower (between 3 and 5 when determined from the cumulative frequency distributions, and between 1.5 and 2.5 when determined from the regression analyses).

An evident disadvantage for the application of the results in this paper lies in the relatively small size of the core samples, which may not account for the continuity of the macropore system **in situ** and which is not always representative of the scale of the structural features. Hence, Eq. (1) could not be directly applied on any soil core sample individually. If, however, an appropriate number of samples are taken over a specific area or areas, an average K_{sm}-value and its standard deviation value can be calculated from frequency distributions of scaling factors of ϕ_e (Eq. (3b)), providing a few field measurements of K_s are carried out, preferably at sites with α_i-values near 1. If direct relationships between K_s and ϕ_e are to be calculated for clay soils similar to those reported here, it is believed that the size of each soil core sample should be larger in order to account for the totality, including the continuity, of the macropore system.

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FIELD-SATURATED HYDRAULIC CONDUCTIVITY AS A MEASURE OF STRUCTURAL VARIATIONS IN HEAVY CLAY SOIL

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SUMMARY

A simple method (the inversed auger hole method) for measuring changes in field-saturated hydraulic conductivity (K_{fs}) and also indirectly the structural fluctuations in clay soils over the season was investigated. Measurements were carried out at three depths in two Swedish alluvial heavy clay soils (45-60 and 65-80 % clay content respectively) in the spring, summer and autumn. It was inferred that the structural fluctuations were greater both in time and with depth for the heavier clay soil. It was also shown that the logarithms of the K_{fs} values were positively correlated (r=0.73) with estimates of the total macroporosity, the latter being determined from a simple model of soil swelling/shrinkage. The inversed auger hole method was concluded as being a useful tool in extensive studies of seasonal changes in K_{fs} and soil structure.

Additional index words: Inversed auger hole method, structure, macropores.

INTRODUCTION

The soils in Sweden can undergo several drying and wetting cycles during the growing season. In the spring after thawing of the soil profile, a drainage equilibrium with field capacity in the topsoil may be obtained after some time. During the summer a vegetated clay soil undergoes swelling and shrinkage which is governed by the balance between precipitation and evapotranspiration. These processes induce changes in soil structure. The higher the clay content, the more extensive are these structural variations. During the autumn, which is normally high in precipitation and low in potential evapotranspiration, the swelling processes dominate and normally proceed until soil freezing.

The drainage efficiency of the soils is strongly influenced by the structural fluctuations. A very heavy clay soil is almost impermeable after a long period of rain, whereas it becomes excessively permeable during dry periods due to the creation and enlargement of cracks. It is this macropore system which is the primary source for rapid transport of excess water within the soil profile (Beven & Germann, 1982). Thus, the volume of macropores is often closely related to the saturated hydraulic conductivity (K_s) (Bouma & Wösten, 1979; Ahuja et al., 1984). In the field, prewetting of initially unsaturated clay soil may cause swelling processes to proceed for months (Bouma et al., 1976) which results in constantly decreasing values of K_s. It may also be assumed that due to air entrapment during the infiltration process, the field methods do not measure the hydraulic conductivity in a completely saturated soil. Therefore, the term field-saturated hydraulic conductivity (K_{fs}) is preferred instead of K_s. It can be argued that K_{fs} is actually the preferred value since few natural or man-made infiltration processes in the field produce complete saturation (Reynolds & Elrick, 1985).

The objective of the study outlined in this paper was to evaluate a simple method for measurement of K_{fs}, and to investigate whether the values obtained could be related to seasonal variations in soil structure. In this study the inversed auger hole method was used to estimate K_{fs} due to its procedural simplicity (Kessler & Oosterbaan, 1974). The variation in K_{fs} as a function of depth and time at two sites with differing clay contents was measured and then interpreted with the aid of a simple model of soil shrinkage.

MATERIALS AND METHODS

Theory

K_{fs} can be determined from the flow out of an auger hole into surrounding soil either for constant head (e.g. constant head well permeameter method, Amoozegar & Warrick, 1986) or falling head conditions (e.g. inversed auger hole method, Kessler & Oosterbaan, 1974). In the past, these methods have been avoided mainly due to drawbacks in their theoretical development. Talsma (1960) noted that approximations in the well permeameter theory (Zangar, 1953) may cause misleading values. Other problems such as silting up of the well wall during the outflow process were also mentioned. Colombani et al. (1973) criticized the constant head well permeameter method on the grounds that the water volume necessary for the measurements is large as is the time needed for establishing K_{fs}. This practical limitation was eliminated by the development of an "in-hole" Mariotte-type permeameter (Talsma & Hallam, 1980) which was further developed by Reynolds et al. (1983,1984).

Flow out of a shallow well into unsaturated soil is properly regarded as a three-dimensional infiltration process and should, as predicted by Philip

(1968,1969), achieve steady-state rapidly and within a finite wetting region. Several solutions exist for estimating K_{fs} values in this situation. The most simple, the Porchet solution (Porchet & Laferre, 1935; Kessler & Oosterbaan, 1974), neglects the functional effect of the pressure head due to the water column in the auger hole. A number of other solutions which are derived from the Glover solution (Zangar, 1953), however, include the effect of the positive pressure potential in the water (Reynolds et al., 1983). Reynolds et al. (1985) and Reynolds & Elrick (1985) extended this theory further by also accounting for the effects of the initial pressure head of the unsaturated soil. They proposed in accordance with Stephens & Neuman (1982) that the wetted region around the well consisted of a small, field-saturated zone adjacent to the hole encased within a much larger unsaturated envelope. The steady-state flow from the well is influenced by the unsaturated envelope so that the capillary properties of this region must affect the calculated value of K_{fs}. In this context, the original theory of Reynolds et al. (1983) was thought to be only slightly in error for soils exhibiting weak capillarity, but could be substantially in error for soils exhibiting strong capillarity. Also Stephens et al. (1987) tried solutions which accounted for capillarity and which were derived from a regression analysis of numerical simulations. These provided good agreement with other solutions following measurements in a uniform sand profile.

The simple Porchet solution (as well as other solutions) assumes a homogeneous, isotropic and rigid porous soil. If the real soil is not uniform the calculated K_{fs} reflects the conductivity of the most permeable layer. When the Porchet solution is used, measurements should thus be made with a small head of water column above the hole bottom so that a homogeneous and isotropic soil in the flow area can be assumed. Colombani et al. (1973) showed how K_{fs} could be determined for each layer in a multi-layered soil with the inversed auger hole method for a falling water head. In this way a more realistic mean K_{fs} for a layered soil can be calculated.

Comparison with other methods

Since the inversed auger hole and constant head well permeameter methods have not been commonly used, comparisons with other methods to determine K_{fs} and K_s are sparse. In a study by Sillanpää (1956) a constant head well permeameter method was compared with the auger hole test in which K_{fs} is determined below the ground water table. The ratio of K_{fs} values of the former to the latter ranged from 0.72 to 1.97. Talsma (1960) and Winger (1960) compared the inversed auger hole method with the auger hole test and reported lower values (0.50-0.85) for the former. Loudiere & Fatton (1983) reported 3-fold higher K_{fs} values estimated using a double-ring infiltrometer than for inversed auger hole. Swarzendruber & Olsson (1961) reported similar values for the two methods if a correction factor 0.53 for lateral flow under the double-ring infiltrometer was assumed. In a study by Petersson et al. (1987) in Tunisia, the double-ring infiltrometer yielded 2-3 times higher K_{fs} values than the inversed auger hole method on non-vegetated plots. On vegetated plots no significant differences were found. Stephens et al. (1987) showed K_{fs} values from regression solutions which were consistent with values obtained by an air entry permeameter and ponding tests.

Soil descriptions

Two sites, Ultuna (59°48'N, 17°39'E) and Limsta (59°48'N, 16°34'E), were chosen for measuring the variation of K_{fs} over one season. Both sites consisted of alluvial heavy clay soils situated in Central Sweden about 80 km apart. The crop rotations were dominated by arable crops, although Limsta was cropped with pasture when the study was carried out (1983). Ultuna was sown with spring cereals. The clay content in the Ultuna soil is 45 % in the topsoil increasing to 60 % at 0.4 m and then decreasing steadily to 40

% at a depth of one metre. The Limsta soil is higher in clay content ranging from 65 % clay in the topsoil to 80 % at 0.5 m and 70 % at 0.8 m.

Ultuna has a strong fine and medium granular structure in the subsoil changing with depth into coarser angular and subangular peds, and ending up with coarse angular blocky peds at 1 m depth (FAO classification system; FAO, 1977). Tubular pores formed by earthworms and roots are many, coarse and fine, which implies that the conditions for both root penetration and water movement are good. The topsoil is medium to coarse subangular blocky when moist but becomes massive when dry. A plough pan of 0.03-0.05 m is observed at 0.20-0.30 m depth, although this becomes perforated with cracks during the summer.

Limsta has a less distinct structure in the subsoil below 0.55 m. The peds, 0.03-0.10 m in diameter, are irregular in shape and less clearly defined than at Ultuna. Vertical cracks with 0.03-0.10 m spacing are formed during dry periods. Earthworm channels and other tubular pores are few. The layer at 0.35-0.55 m depth consists of very fine angular peds which can be compared to the granular layer in the Ultuna soil at approximately the same depth. This layer is formed by frost action and, in the case of Ultuna, rich biological activity. In the topsoil and in the plough pan the Limsta soil is massive when dry and cohesive when wet. When the soil is dry, wide cracks at 0.2-0.4 m spacing are created. However, a grass and clover pasture, as during the measurement period, enhances the formation of fine and medium angular peds in these layers.

Experimental procedure and analyses of data

At the two sites K_{fs} and gravimetric soil water content were determined three times during the season; in the spring (20-22nd April), summer (19-

22nd August), and autumn (27th October at Limsta and 24th November at Ultuna). In order to determine $\mathbf{K_{fs}}$, holes of 0.08 m diameter were augered to 0.30, 0.50, and 0.80 m depth at each site, with five replicates at each depth. The distance between auger holes was 1.5 m. New holes were augered on each experimental occasion. Water was then filled up to 0.20 m above the hole bottom at each measurement. In the spring and autumn the rate of fall of the water level from the two first replenishments were averaged for calculating $\mathbf{K_{fs}}$. In the summer the values from the second and third replenishments were averaged. Colombani et al. (1973) noted on heavy clay soil when comparing the first and second replenishments of the auger hole that the flow out of the hole into the surrounding soil was significantly lower the second time. The decrease was less between the second and third replenishments and was supposed to decrease only slowly thereafter. The flow rate is influenced by movement along macropores as well as by water uptake by the peds. Once the peds have been wetted however, "short-circuiting" along soil structural features will dominate the flow (Hoogmoed & Bouma, 1980). The swelling of the soil will thereafter reduce the macropore volume and thus also the flow. This process is very slow and will proceed for a very long time resulting in continuously diminishing flow rate (Bouma et al., 1976). In order to verify these aspects, short time prewetting (2-3 replenishments) was compared with continuous prewetting during three hours at the summer measurements at Ultuna.

The inversed auger hole method with falling water head was utilized to calculate $\mathbf{K_{fs}}$ in accordance with the Porchet solution (Porchet & Laferre, 1935; Kessler and Oosterbaan, 1974). Briefly, the rate of fall of the water level is registered. $\mathbf{h+(r/2)}$, where $\mathbf{h}$=the head of water above the hole bottom at each time-step and $\mathbf{r}$=the radius of the auger hole, is plotted on semi-logarithmic paper as a logarithmic function of linear time. The slope of the line ($\mathbf{tan}\alpha$), is put into the formula

$$K_{fs} = 1.15 \tan\alpha \qquad (1)$$

to find the field-saturated hydraulic conductivity.

Because of the low number of observations extensive statistical treatment of the data has been avoided. The median is thought to represent a better central estimate than the mean since it is less influenced by gross errors (Dixon, 1986). Furthermore the range (**w**) is proposed as a convenient measure of the dispersion. In order to convert the range to a measure of dispersion independent of the number of observations, **w** was multiplied by a factor (K_w) which is tabulated in Dixon (1986). The product $wK_w = s_w$ is thus an estimate of the standard deviation for a small number of observations. The coefficients of variation (CV_w) were estimated from $CV_w = s_w$/median. Using the CV_w value we can compare the variation of each set of observations for different sites, seasons, and depths.

The values of initial gravimetric soil water contents were calculated from three separate augerings as the arithmetic means at each depth. At the autumn experiments at Ultuna only two augerings were carried out due to difficult sampling environment (frozen topsoil). Volumetric soil water contents were calculated from dry bulk densities at the two sites. The dry bulk densities at Limsta ranged between 1.30 g/cm^3 in uncompacted layers and 1.35 in the compacted layer (unpublished data), whereas at Ultuna they ranged between 1.48 and 1.52 g/cm^3 (Wiklert et al., 1983). The total porosity ($\emptyset$) was calculated from the dry bulk density and the density of solids. Water retention at the pressure heads -10 kPa and -1500 kPa was determined using tension table and pressure plate apparatus.

The K_{fs} values determined with the inversed auger hole method at Ultuna are also compared with K_s values determined on small core samples (0.072 m in diameter and 0.10 m in height) in the laboratory. The K_s values were estimated after prolonged wetting, with a constant head permeameter under a unit hydraulic gradient (Andersson, 1953). The K_s values presented are the mean for 11 sites at Ultuna scattered at and nearby the present experimental site but sampled over some years in the past (Wiklert et al., 1983).

Soil structure relationships

In order to interpret variations in K_{fs} in terms of soil structure fluctuations, a way of describing the latter must be defined. A simple model described by Jarvis & Leeds-Harrison (1987) is utilized here. The term field capacity (Θ_f) is taken to imply the water content of a fully swollen clay soil, but with any remaining fissures together with biotic macropores empty of water. These macropores comprise the stable macroporosity (e_s). The total macroporosity (e_t) is dependent on the moisture status of the bulk soil as described by:

$$e_t = e_s + (\Theta_f - \Theta) \tag{2}$$

where Θ is the current soil water content. The term $\Theta_f - \Theta$ may be termed the dynamic macroporosity (e_d). If e_s is defined as soil water content at saturation ($\emptyset$) minus Θ_f, then

$$e_t = \emptyset - \Theta \tag{3}$$

Two important assumptions for the model are: i) normal shrinkage assumed for aggregates (i.e. no air entry into peds), and ii) vertical shrinkage is neglected.

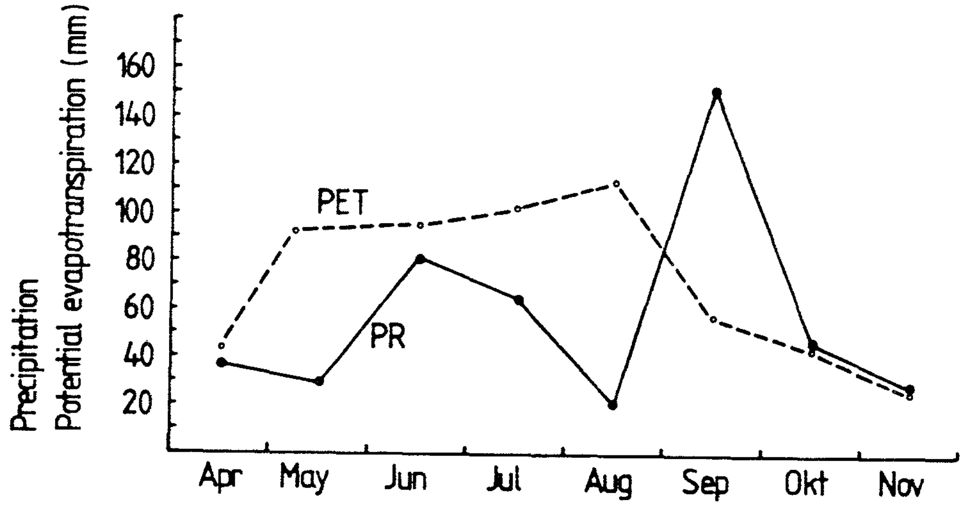

Fig. 1. Monthly values of precipitation (PR) and Penman potential evapotranspiration (PET) at Ultuna 1983.

RESULTS AND DISCUSSION

Climate and soil water content

Since Ultuna and Limsta are not situated far apart the climate was similar for the two sites. Data on precipitation and Penman potential evapotranspiration at Ultuna for 1983 are shown in Fig. 1. The spring measurements were carried out about 10 days after thawing. The groundwater level was at that stage at about 1 m depth at both sites, and the soils were thought to be fully swollen and near field capacity. In Fig. 2 the spring values for volumetric soil water content are shown to lie close to the -10 kPa pressure head curves, the latter being regarded as a field capacity value in Swedish soils. Between the spring and summer measurements, 210 mm of rain fell at Ultuna (SMHI, 1983) which was a normal amount, and the cumulative potential evapotranspiration was 360 mm. The potential water deficit was thus 150 mm during that period. The consequent reduction in soil water content was sufficient to create an extensive cracking pattern at both sites. The higher water consumption at Limsta (Fig. 2) was probably due to the pasture which started transpiration and root water uptake earlier in the spring than the spring cereals at Ultuna.

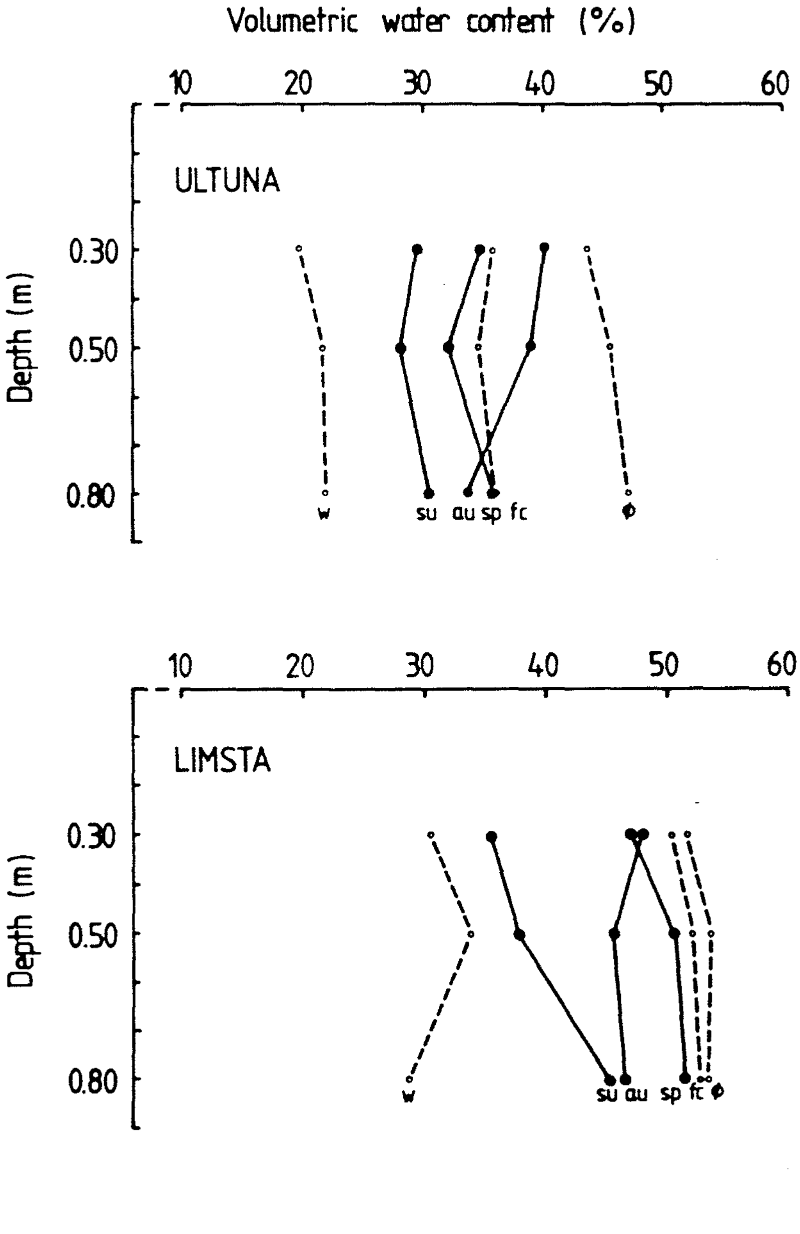

Fig. 2. Initial soil water content for spring (**sp**), summer (**su**), and autumn (**au**) experiments. **w**=wilting point; **fc**=soil water content at -10 kPa pressure head; **Ø**=total porosity.

The period between the summer and autumn measurements was characterized by an excess of rainfall over potential evapotranspiration (Fig. 1). At Ultuna 225 mm of rain fell, which was 95 mm more than the normal (SMHI, 1983). The potential evapotranspiration was 175 mm and hence the potential water surplus was 50 mm. The Ultuna soil in particular was assumed to be fully swollen at the time of the autumn experiments, which were carried out one month later than at Limsta and two months after harvest of the spring cereal crop. The autumn curve for soil water content at Ultuna (Fig. 2) may overestimate the real conditions. Nevertheless, it may be assumed that the curve should (as for the spring curve) lie near the -10 kPa curve due to a fairly constant stable macroporosity (e_s in Eq. (2)).

Variation of K_{fs} and the effect of soil structure

Median K_{fs} values for the two sites are presented in Fig. 3 and Table 1. It is shown that the Limsta soil has lower K_{fs} values than the Ultuna soil with the exception of the summer experiments. It is also shown that for the depths 0.30 and 0.50 m the differences between spring and autumn experiments compared with summer experiments were larger at Limsta than at Ultuna. This implies larger structural fluctuations over the season at Limsta. Perched ground water levels in and above the compacted layer during rainy periods in autumn, winter and spring enhance swelling and reduce K_{fs}. The very low spring values therefore reflect the conditions in wet, fully swollen soil. Some of the remaining macropores may have been smeared by the auger yielding underestimated K_{fs} values. Thus, the zero-values in the spring may reflect infiltration in the clay matrix rather than in the stable larger pores (e_s in Eq.(2)). Extensive cracking during dry periods at Limsta enhanced shrinkage and increased the dynamic macroporosity and hence K_{fs} during the summer in the 0.30 and 0.50 m layers. At 0.80 m the conditions were more constant over the season due to a less variable soil water content (Fig. 2). The high CV_w values in the two upper layers imply the occurrence of wide cracks in a rather dense matrix.

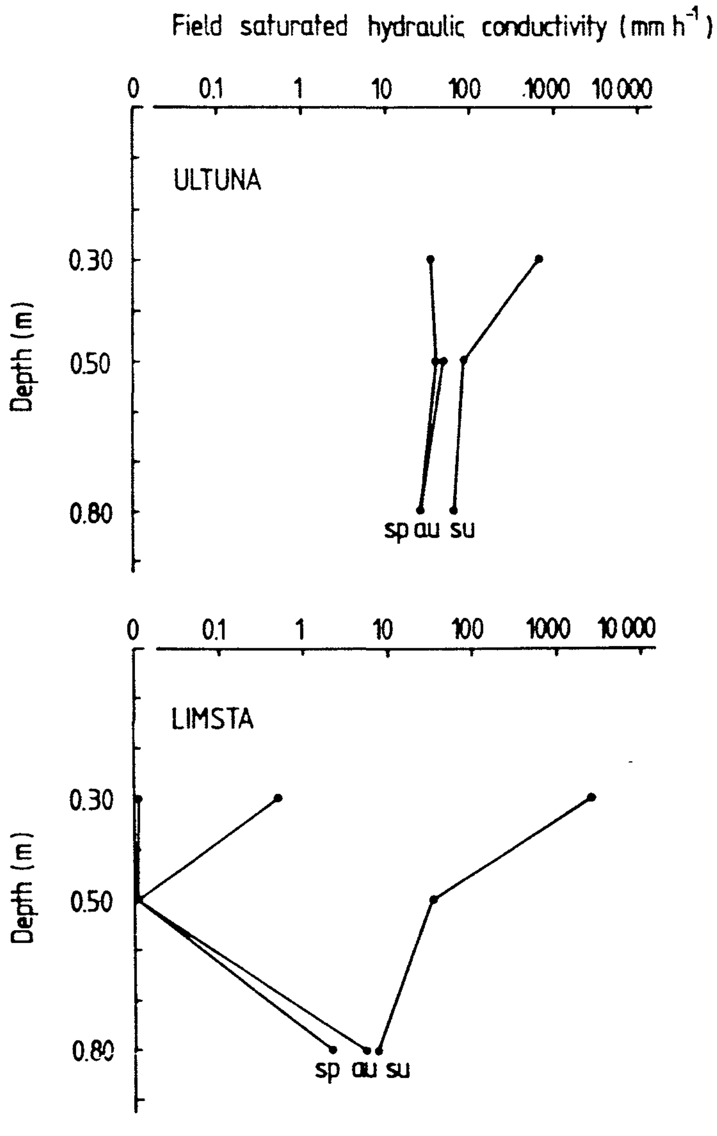

Fig.3. Field-saturated hydraulic conductivity (K_{fs}) for spring (**sp**), summer (**su**), and autumn (**au**) experiments.

Table 1. Field-saturated hydraulic conductivity (K_{fs}) (mm h^{-1}) for different seasons and depths at Ultuna and Limsta

Site	Season	Depth (m)	n	Median K_{fs} (mm h^{-1})	s_w (mm h^{-1})	CV_w (%)	w (mm h^{-1})
Ultuna	Spring	0.30	5	34.7	33.2	96	7.7-8.5
		0.50	5	41.1	35.3	86	8.7-90.7
		0.80	5	26.1	38.2	146	20.0-108.8
	Summer	0.30	5	672.0	1400.1	208	67.4-3323.4
		0.50	5	84.2	35.2	42	49.8-131.6
		0.80	5	60.4	18.3	30	47.3-89.8
	Autumn	0.30[a]	-	-	-	-	-
		0.50	5	45.7	16.0	35	26.9-64.1
		0.80	5	25.3	10.5	42	19.3-43.8
Limsta	Spring	0.30	5	0.0	0.4	200[b]	0.0-1.0
		0.50	5	0.0	1.8	164[b]	0.0-4.3
		0.80	4	2.2	1.1	50	1.2-3.5
	Summer	0.30	5	2683.1	4283.4	160	38.5-10000[c]
		0.50	5	34.8	61.6	177	24.8-168.0
		0.80	5	7.6	6.4	84	3.4-18.2
	Autumn	0.30	5	0.5	1.1	220	0.0-2.9
		0.50	5	0.0	0.9	225[b]	0.0-2.2
		0.80	5	5.8	3.1	53	2.9-10.2

a) Frozen topsoil.

b) Calculated from antilogged mean of log K_{fs}.

c) Estimated value. Excessive flow rate.

At Ultuna the variation of median K_{fs} values over the season was less pronounced and was evident mainly in the 0.30 m layer (Fig. 3). Deeper in the profile the medians varied within one order of magnitude although the summer values were higher. This difference is clearly reflected in the structural descriptions of the soils, in that Ultuna is better structured than Limsta and has a larger e_s (as derived by Eq. (2)) thus yielding high K_{fs} values even at high soil water contents. The CV_w values were generally moderate at Ultuna, although the high CV_w in the uppermost layer in the summer experiment reflected the more frequent occurrence of wider cracks in a rather dense matrix than for the other depths and seasons. Smearing seemed to be less of a problem at Ultuna since the K_{fs} values were generally high.

The power law relationship between K_{fs} and total macroporosity (e_t as estimated by Eq. (3)) was investigated. In this analysis the zero-values for median K_{fs} in Table 1 were replaced by arithmetic means. The approach is promising (see Fig. 4) since K_{fs} may simply be approximated from only two parameters, porosity ($\emptyset$) and initial soil water content (θ). The exponent 2.80 in Fig. 4 may be compared with Messing (1989), where the relationship between laboratory determined $\emptyset_e$ and saturated hydraulic conductivity was estimated on a large number of cores from Swedish clay soils. The exponent of $\emptyset_e$ varied between 2 and 3.

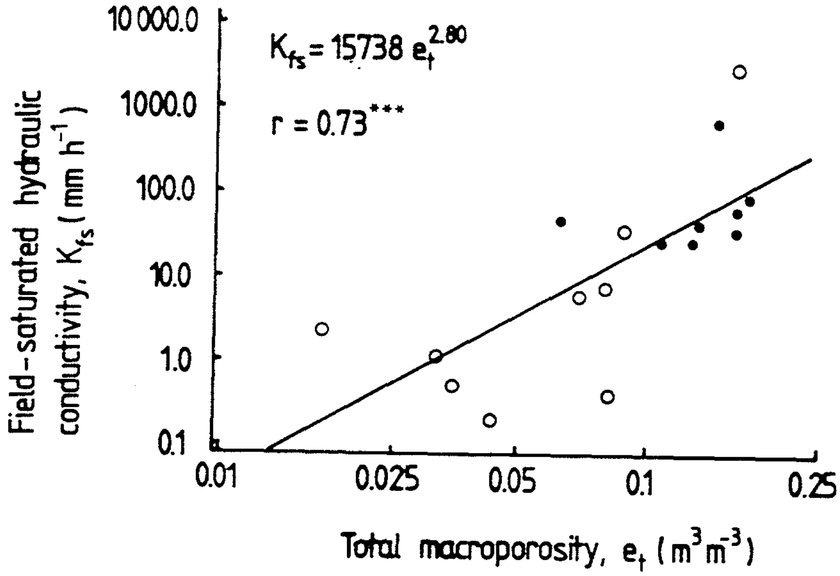

Fig. 4. Field-saturated hydraulic conductivity (K_{fs}) as a function of total macroporosity (e_t). (• = Ultuna; o = Limsta).

Table 2. Field-saturated hydraulic conductivity (K_{fs}) (mm h^{-1}) at summer experiments at Ultuna after three hours prewetting

Depth (m)	n	Median K_{fs} (mm h^{-1})	s_w (mm h^{-1})	CV_w (%)	w (mm h^{-1})
0.30	4	68.8	359.0	521	24.8-763.4
0.50	5	22.0	12.5	57	14.5-43.6
0.80	4	22.6	25.3	112	14.3-66.3

Effect of time of prewetting

In the summer experiments the Ultuna soil was allowed to prewet for three hours by continuous replenishments of the auger holes. The K_{fs} values obtained are presented in Table 2. It was noted that these values were much lower than those calculated after the second and third replenishments (Table 1). They were actually closer to the spring and autumn values. Nevertheless, it was considered that a prewetting of several hours with a continuous water head did not correspond to the purpose outlined in the present study, which was rather to find a way to explain seasonal variations in K_{fs} in terms of changes in soil structure.

Comparison with laboratory K_s values

Mean values for the saturated hydraulic conductivity K_s at Ultuna, determined in the laboratory on undisturbed soil core samples are presented in Table 3. These values may be compared with the field-measured spring and autumn K_{fs} values at Ultuna (Table 1). The K_s values from the soil cores are similar to the K_{fs} values only for the 0.30 m layer. At 0.50 and 0.80 m

Table 3. Saturated hydraulic conductivity (K_s) (mm h^{-1}) determined on soil core samples in the laboratory from nearby sites at Ultuna

Depth (m)	n	Mean[a)] (mm h^{-1})	SD[a)] (mm h^{-1})	CV (%)	**w** (mm h^{-1})
0.30	33	24.6	69.1	281	0.2-1040
0.50	33	110.7	48.7	44	2.0-2530
0.80	32	149.1	104.2	70	1.0-6270

a) Antilogged mean and standard deviation (SD) of log K_s.

depth the K_s values are much higher. Many macropores, particularly the vertical tubular macropores. which are abundant at 0.5 and 0.8 m depth are probably hydraulically passive in the field, especially after prewetting, since they may end in dead-ends. In the laboratory measurements on the other hand, they will contribute much more to the water flow since they are truncated every 10 cm. In the plough pan at 0.30 m, however, the number of tubular macropores is fewer resulting in similar K_s and K_{fs} values.

CONCLUSIONS

Measurements of the field-saturated hydraulic conductivity (K_{fs}) with the inversed auger hole method using the Porchet solution showed that in a structured soil K_{fs} is not a constant, but instead varies with the structural state of the soil at the time of measurement. K_{fs} was shown to be related to the observed structural variations in two clay soils over one season (Fig. 3). For example, for the well structured Ultuna soil with 45-60 % clay content and spring sown cereals, the main structural changes that affected K_{fs} were recorded to a maximum depth of between 0.3-0.5 m. In these upper layers the fluctuations in soil water content were largest (Fig. 2).

Below this depth variations in both K_{fs} and initial soil water content were recorded but were not as obvious. For the heavier soil at Limsta with 65-80 % clay content and under pasture, large changes in K_{fs} were measured deeper into the soil (0.5 to 0.8 m depth). These observations coincided well with visual observations of structure in the field. In the layers where the K_{fs} fluctuations were largest, the occurrence of wider cracks in a rather dense matrix was noted in the summer.

Using the concepts and terminology in Eq. (2), it may also be concluded that the well structured Ultuna soil has a higher stable macroporosity (e_s) than the Limsta soil. K_{fs} was shown to be well related to the total macroporosity (e_t), the latter being defined as in Eq. (3).

Smearing of the hole walls is a problem with the inversed auger hole method which is difficult to detect. The user must be aware that measurements in wet soil may produce underestimated K_{fs} values. However, a more serious drawback concerning this method lies in the theoretical treatment to estimate a real K_{fs} from the flow rate. If the strict limitations in the experimental procedures as presented in **Materials and methods** are observed, conclusions on the relative changes in K_{fs} as well as the structural fluctuations causing the changes can be drawn.

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