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Indirect estimation of near-saturated hydraulic conductivity from readily available soil information

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Abstract

Application of process-based water flow and solute transport models is often hampered by insufficient knowledge of soil hydraulic properties. This is certainly true for dual- or multi-porosity models that account for non-equilibrium flow of water in macropores, where the saturated 'matrix' hydraulic conductivity is a particularly critical parameter. Direct measurement is possible, but this is impractical for larger scale studies (i.e. catchment or regional), where estimation methods (pedotransfer functions) are usually required. This paper presents pedotransfer functions for hydraulic conductivity at a pressure head of -10 cm, K_{10}, based on measurements of near-saturated hydraulic conductivity made with tension infiltrometers in 70 soil horizons at 37 different sites in

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13 different countries. Pedotransfer functions were developed using texture classes, the geometric mean particle size, organic carbon content, bulk density and effective porosity as predictor variables. The pedotransfer functions explained no more than 12% to 29% of the variation in K_{10} for the complete dataset. Some important sources of unexplained variation in K_{10} may include errors and uncertainty in the (indirect) method used to measure K_{10}, differences in the way the tension infiltrometer is used, and also temporal changes in hydraulic conductivity due to tillage and/or surface sealing. The importance of tillage was emphasized by the fact that excluding arable topsoils from the analysis gave improved predictions (r^2 values between 26% and 44%) for pedotransfer functions based on texture classes, mean particle size and effective porosity.

Keywords: Pedotransfer function; Near-saturated hydraulic conductivity; Tension infiltrometer; Geometric mean particle size; Effective porosity

1. Introduction

The use of detailed process-based simulation models for management purposes, especially at regional scales, is often hampered by a lack of information concerning model input parameters such as soil hydraulic properties. One solution to this problem is to make use of statistical estimation routines (pedotransfer functions) to derive the required parameter values from more widely available soil survey information (e.g. Rawls et al., 1982; Tietje and Tapkenhinrichs, 1993). Pedotransfer functions can serve as a useful means of parameterizing complex models, providing that the level of accuracy achieved is adequate in a functional sense, and that the range of applicability of the functional relationships are known and respected (Wösten et al., 1990; Mayr and Jarvis, 1999). Pedotransfer functions are particularly useful for catchment and regional scale applications of models since the availability of measured hydraulic properties is inevitably limited across large areas. To date, most attention has focused on developing estimation procedures for soil water retention and saturated hydraulic conductivity, while the unsaturated hydraulic conductivity function has received much less attention (e.g. van Genuchten et al., 1999).

In many field soils, saturated hydraulic conductivity K_{sat} is largely controlled by soil structural features, or macropores. Although the effects of macropores on hydraulic properties are difficult to quantify, some successful attempts have been made to derive pedotransfer functions for K_{sat} based on soil morphological features such as the frequency and size of structural pores or FAO soil structure description (e.g. McKenzie et al., 1991; McKeague et al., 1982). In contrast, the many attempts to relate measured saturated hydraulic conductivity K_{sat} to bulk soil properties such as textural composition, organic carbon content and bulk density have met with limited success (e.g. Tietje and Hennings, 1996; Wösten et al., 1998).

The complete unsaturated hydraulic conductivity function is particularly difficult and time-consuming to measure directly. Therefore, in many model applications, reliance is often placed on predictions of unsaturated conductivity based on measurements of soil water retention and K_{sat} (van Genuchten and Nielsen, 1985). However, the use of measured K_{sat} as a 'matching point' in such estimation procedures may lead to large overestimates in unsaturated conductivity in the dry range due to the dominating effects of soil structure on water flow at and near saturation (Jarvis et al., 1999). Indeed, it is commonly observed that hydraulic conductivity decreases by ca. 1 to 3 orders of magnitude across a small pressure head range near saturation (zero to -10 cm pressure head), due to the effects of structural macropores (e.g. Clothier and Smettem, 1990; Jarvis and Messing, 1995). Therefore, by excluding the effects of macropores, the use of near-saturated hydraulic conductivity as a matching point should give more reliable predictions of unsaturated hydraulic conductivity in the dry range (Luckner et al., 1989).

Tension or disc infiltrometers are perhaps the most widely used technique for measuring near-saturated hydraulic conductivity (White et al., 1992). The method is inexpensive and simple and, most importantly, can be applied to undisturbed soils in the field. Direct measurements made with tension infiltrometers have been widely used to estimate parameters in dual- and multi-porosity models that explicitly account for the effects of macropores on water flow and solute transport (Gerke and van Genuchten, 1993; Jarvis, 1994; Gwo et al., 1995; Mohanty et al., 1997), but reliable estimation methods are lacking for predicting near-saturated hydraulic conductivity in the absence of such measurements.

Some physico-empirical approaches to estimate K_{sat} have been proposed, based on either geometric mean particle diameter (Campbell, 1985), 'effective porosity' (Ahuja et al., 1984), or soil water retention curve parameters (Laliberte et al., 1968). However, these methods may be more suitable for predicting near-saturated hydraulic conductivity, excluding macropore effects (Jarvis et al., 1999). For example, Smettem and Bristow (1999) recently showed that hydraulic conductivity measured at -4 cm in 20 Australian soils was well predicted by a modified form of Campbell's (1985) approach based on clay content rather than the geometric mean particle diameter. Rawls et al. (1993) proposed a method to estimate saturated matrix hydraulic conductivity from effective porosity and air-entry pressure, but predictions were not compared to experimental measurements. It can be noted that although estimation methods based on water retention curve parameters such as 'matrix' air-entry pressure and pore size distribution have the advantage of being physically based, such data may not be generally available for larger areas (Tietje and Hennings, 1996). Also, they may not be so robust, since saturated matrix conductivity should in theory be inversely proportional to the square of the (matrix) air-entry pressure (Laliberte et al., 1968), which itself is not always well defined by measurements, nor easy to predict from other surrogate variables (Mayr and Jarvis, 1999).

In this paper, we investigate the utility of readily available soil property data (particle size distribution, bulk density, organic carbon content, effective porosity) to predict hydraulic conductivity near saturation. Simple pedotransfer functions are derived from a database of tension infiltrometer measurements of hydraulic conductivity made in 70 soil horizons at 37 different sites in 13 different countries. The functions derived in this paper may prove useful in helping parameterize the hydraulic conductivity function of simulation models, including single-, dual- and multi-porosity models, in the absence of any direct measurements.

2. Materials and methods

2.1. The database

Table 1 shows the complete database of 70 soil horizons used in this study to derive pedotransfer functions for near-saturated hydraulic conductivity. Fig. 1 shows that the soils provide a reasonably good coverage of the texture triangle diagram, with 8 of the 11 USDA textural classes represented (sandy clay, sandy clay loam and silt classes are missing). Table 1 shows that bulk densities vary from 0.83 g cm^{-3} in a 38 year continuous blue-grass pasture to 1.68 g cm^{-3} in a silt loam subsoil. Organic carbon contents vary from 0.1% in some sandy subsoils to more than 6% under deciduous woodland. The soil horizons also represent a range of different land uses (Table 1). Arable land dominates ($n=50$), but the database also includes measurements made in forest soils ($n=11$) and grassland (including natural bush and pasture land, $n=9$). The measurements were also made predominantly in the topsoil ($n=58$), most often directly at the soil surface, but some data ($n=12$) are available for subsoil horizons (Table 1).

2.2. Field measurements of hydraulic conductivity and data interpolation

For all soils included in the database, near-saturated hydraulic conductivity was estimated from three-dimensional steady-state infiltration rates measured from tension infiltrometers using the method originally proposed by Ankeny et al. (1991) and Reynolds and Elrick (1991) and further described by Messing and Jarvis (1993). Different infiltrometer designs were used at the different sites, but in all cases, infiltration was measured manually, and the supply tensions determined by calibration in the laboratory against known pressures. The range of supply tensions used also varied from site to site, so that some interpolation of the paired values of hydraulic conductivity and supply tension was required in order to derive a comparable and uniform dataset. Jarvis and Messing (1995) showed, for six soils of contrasting texture, that an exponential relationship between conductivity and pressure head (Gardner, 1958) could be made to fit measured near-saturated conductivity data, but only for a limited range of supply tensions. Therefore, in this study, we derived hydraulic conductivity at the fixed pressure head of -10 cm H_2O (hereafter termed K_{10}) using Gardner's exponential equation to interpolate data points measured at either two or three adjacent pressure heads varying between -3 and -15 cm H_2O. Data measured at pressure heads larger than -3 cm H_2O were excluded from the analysis, since hydraulic conductivity is known to increase rapidly towards saturation in this range of pressure heads (Clothier and Smettem, 1990; Jarvis and Messing, 1995). One average measure of K_{10} was obtained from replicate measurements made at each site in Table 1, giving a total of 70 K_{10} values, varying over three orders of magnitude, from 0.05 to 107.2 mm h^{-1}. That we define saturated 'matrix' hydraulic conductivity at a pressure head of -10 cm is, of course, a subjective choice, but one that is based on pragmatism and experience. For example, Seyfried and Rao (1987) and Jardine et al. (1993) demonstrated that steady-state solute breakthrough experiments carried out at pressure heads larger than -10 cm showed the early breakthrough and long tailing characteristic of preferential flow in macropores, while experiments run at pressure heads less than or equal to -10 cm did

Table 1
Site details and soil characteristics (K_{10} is the hydraulic conductivity at a pressure head of -10 cm)

Soil/horizon	Location	Land use/treatment	Texture	K_{10} (mm h^{-1})	Organic carbon content (%)	Bulk density (g cm^{-3})	Particle size distribution (%)		
							clay (< 2 μm)	silt (2–50 μm)	Sand/gravel (> 50 μm)
Mellby A_p	Sweden, 56°29′ N, 13°00′ E	Arable (spring barley at time of measurements)	Loamy sand	1.9	3.4	1.23	10.4	8.4	81.2
B_g			Sand	11.8	0.2	1.55	2.9	1.9	95.2
Kjettslinge A_p	Sweden, 60°10′ N, 17°38′ E	Arable (spring barley at time of measurements)	Loam	0.1	2.0	1.38	19.0	29.0	52.0
Lanna A_p	Sweden, 58°21′ N, 13°08′ E	Arable (oats at time of measurements)	Clay	0.2	2.6	1.16	46.5	42.2	11.3
Ultuna A_p	Sweden, 59°49′ N, 17°39′ E	Arable (oats), June	Clay	0.18	1.7	1.17	44.0	38.0	18.0
		August	Clay	0.05	1.7	1.28	44.0	38.0	18.0
		Arable, non-trafficked and sewage sludge amended (40 years)	Clay	0.09	2.4	0.91	44.0	38.0	18.0
Säby A_p	Sweden, 59°49′ N, 17°39′ E	Arable (oats at time of measurements)	Loam	0.1	3.2	1.13	17.2	44.8	38.0
Nåntuna A_p	Sweden, 59°49′ N, 17°39′ E	Arable (spring barley at time of measurements)	Loamy sand	1.1	0.7	1.40	7.7	3.7	88.6
Silsoe Farm A_p	England (UK), 52°00′ N, 00°25′ W	Fallow	Sandy loam	0.4	1.9	1.44	13.0	10.8	76.2
Silsoe College A_p	England (UK), 52°00′ N, 00°25′ W	Recently rotovated bare soil	Clay	0.2	2.0	1.12	48.0	18.0	34.0
Needham A_p	England (UK), 52°37′ N, 00°10′ E	Fallow, previously under strawberries	Loam	3.1	5.8	1.42	9.0	39.0	52.0
Hyytiälä, Site 1	Finland, 61°48′ N, 24°19′ E	Pine forest, surface	Sand	3.0	0.7	1.19	2.0	8.7	89.3
		7 cm depth	Sand	3.0	1.2	1.25	1.3	7.5	91.2
		20 cm depth	Sand	47.4	0.4	1.41	0.8	3.2	96.0
		50 cm depth	Sand	107.2	0.1	1.48	0.8	1.1	98.1

(*continued on next page*)

Table 1 (*continued*)

Soil/horizon	Location	Land use/treatment	Texture	K_{10} (mm h^{-1})	Organic carbon content (%)	Bulk density (g cm^{-3})	Particle size distribution (%)		
							clay (< 2 µm)	silt (2–50 µm)	Sand/gravel (> 50 µm)
Hyytiälä, Site 2	Finland, 61°48′ N, 24°19 ′ E	Pine forest, surface	Loamy sand	3.9	0.8	1.22	2.0	13.9	84.1
		7 cm depth	Sand	3.7	1.5	1.10	1.4	10.8	87.8
		20 cm depth	Sand	33.2	0.5	1.37	0.7	9.6	89.7
		50 cm depth	Sand	75.1	0.1	1.48	0.8	7.2	92.0
Vilan 2	Sweden, 59°49′ N, 17°39′ E	Arable	Sandy loam	1.04	1.4	1.48	12.0	31.0	57.0
Tetto Frati	Italy	Maize	Loam	1.9	1.2	1.43	8.5	43.1	48.4
		Permanent meadow	Silt loam	0.6	1.2	1.18	7.5	66.3	26.2
Beltempo	Italy	Maize	Sandy loam	1.2	2.7	1.32	4.0	42.8	53.2
		Permanent meadow	Silt loam	0.2	2.1	1.44	6.8	56.5	36.7
Isolabella	Italy	Maize	Loam	0.2	1.0	1.43	11.6	43.5	44.9
		Permanent meadow	Silt loam	1.0	1.9	1.44	3.0	51.4	45.6
Colombero	Italy	Maize	Sandy loam	1.2	1.4	1.37	7.0	40.3	52.7
		Maize after barley	Silt loam	4.2	1.6	1.53	7.9	37.8	54.3
Acquasana	Italy	Maize	Loam	1.6	1.1	1.50	12.0	48.0	40.0
Øsaker	Norway, 59°19′ N, 11°01′ E	Spring cereals, surface	Clay loam	2.3	2.2	1.15	36.0	33.0	31.0
		30 cm depth	Clay	1.2	1.9	1.45	49.0	30.0	21.0
Bjørnebekk	Norway, 59°39′ N, 10°50′ E	Spring cereals, surface	Silt loam	3.6	1.4	1.35	27.0	52.0	21.0
		30 cm depth	Silt loam	1.7	0.4	1.54	36.0	51.0	13.0
Ås, AUN	Norway, 59°39′ N, 10°46′ E	Fallow, surface	Clay loam	4.7	4.3	1.33	27.0	37.0	36.0
		30 cm depth	Silt loam	1.7	1.6	1.68	21.0	59.0	20.0
Vilan	Sweden, 59°49′ N, 17°39′ E	Vegetables, 25 cm depth	Clay	0.42	0.7	1.46	42.0	30.0	28.0

Bush Estate	Scotland (UK)	Arable, surface	Sandy loam	15.86	2.5	1.21	16.0	16.0	68.0
(Penicuik)	55°51′ N, 3°13′ W	lower topsoil	Sandy loam	11.04	2.5	1.33	16.0	16.0	68.0
		subsoil, 50 cm depth	Sandy loam	4.99	1.2	1.32	8.0	14.0	78.0
Glencorse Mains Farm	Scotland (UK)	Arable, surface	Loam	15.87	3.1	1.25	24.0	29.0	47.0
(Penucuik)	55°51′ N, 3°13′ W	lower topsoil	Loam	4.15	3.1	1.34	24.0	29.0	47.0
		subsoil, 50 cm depth	Loam	1.68	0.7	1.43	23.0	32.0	45.0
Crichton Royal Farm,	Scotland (UK)	Grassland, surface	Silt loam	10.53	2.2	1.33	26.0	53.0	21.0
Dumfries	55°4′ N, 3°37′ W	lower topsoil	Silt loam	3.03	2.2	1.39	26.0	53.0	21.0
		subsoil, 50 cm depth	Silty clay	0.88	1.0	1.39	44.0	47.0	9.0
German Greve Silva, Santiago	Chile, 33°28′ S, 70°50′ W	Natural pastoral land	Loam	0.55	1.0	1.52	20.4	33.0	46.6
Kabete, Nairobi	Kenya, 1°15′ S, 36°44′ E	Cleared natural bush, first year Maize	Clay	1.5	3.0	1.04	52.0	28.0	20.0
Boone, IA	USA	Maize rows	Loam	6.3	2.5	1.35	22.8	35.2	42.0
Näsbygård, Skåne	Sweden	Spring wheat, hilltops	Loam	0.4	1.2	1.42	25.1	34.5	40.4
		Midslopes	Loam	1.1	1.4	1.38	18.1	25.5	56.4
		Hollows	Loam	0.8	5.0	1.13	22.2	25.7	52.1
Bekkevoort	Belgium	Bare soil	Silt loam	1.0	2.2	1.43	15.4	50.6	34.0
Brookston A_p	Canada, 42°13′ N, 82°44′ W	14-year maize–soybean, mouldboard plough	Clay loam	0.36	2.2	1.37	37.0	35.0	28.0
Brookston A_p	Canada, 42°13′ N, 82°44′ W	14-year maize–soybean, no tillage	Clay loam	0.33	2.2	1.33	37.0	35.0	28.0
Brookston A_h	Canada, 42°13′ N, 82°44′ W	Virgin woodlot, deciduous trees and native grasses	Clay loam	0.50	6.8	0.88	37.0	35.0	28.0
Brookston A_h	Canada, 42°13′ N, 82°44′ W	38-year continuous blue-grass sod	Clay loam	0.21	3.9	0.83	37.0	35.0	28.0
Guelph A_p	Canada, 43°38′ N, 80°11′ W	17-year maize–soybean–winter wheat, ploughed	Loam	1.18	2.3	1.35	16.0	48.0	36.0
Guelph A_p	Canada, 43°38′ N, 80°11′ W	9-year maize–soybean–winter wheat, ploughed	Loam	0.76	2.7	1.20	16.0	48.0	36.0

(*continued on next page*)

Table 1 (*continued*)

Soil/horizon	Location	Land use/treatment	Texture	K_{10} (mm h^{-1})	Organic carbon content (%)	Bulk density (g cm^{-3})	Particle size distribution (%)		
							clay (< 2 μm)	silt (2–50 μm)	Sand/gravel (> 50 μm)
Guelph A_h	Canada, 43°38′ N, 80°11′ W	Virgin woodlot (>50 years), deciduous trees and native grasses	Loam	1.59	5.1	0.89	16.0	48.0	36.0
Fox A_p	Canada, 42°52′ N, 80°31′ W	15-year maize–soybean–winter wheat, ploughed	Sand	8.03	0.7	1.52	5.0	5.0	90.0
Fox A_p	Canada, 42°52′ N, 80°31′ W	6-year maize–soybean–winter wheat, no-till	Sand	2.79	0.8	1.54	5.0	5.0	90.0
Fox, A_p	Canada, 42°52′ N, 80°31′ W	Virgin woodlot (>50 years), deciduous trees and native grasses	Sand	1.17	2.3	1.10	5.0	5.0	90.0
Herceghalom 1 A_p	Hungary, 47°30′ N, 18°45′ E	Maize	Silt loam	4.4	1.5	1.38	24.9	65.3	9.8
Herceghalom 2 A_p	Hungary, 47°30′ N, 18°45′ E	Alfalfa	Silt loam	1.6	1.6	1.43	24.5	59.6	15.9
Gödöllõ 1 A_p	Hungary, 47°30′ N, 19°22′ E	Maize, loosened and ploughed (March)	Sandy loam	8.6	0.9	1.37	17.1	20.6	62.3
Gödöllõ 2 A_p	Hungary, 47°30′ N, 19°22′ E	Maize, direct drilled (August)	Loamy sand	3.56	0.8	1.50	17.7	4.5	87.8
Kaposvár A_p	Hungary, 46°22′ N, 17°50′ E	Maize	Silt loam	0.57	0.9	1.47	20.6	68.7	18.1
Székelyszabar A_p	Hungary, 46°24′ N, 18°20′ E	Maize	Silt loam	2.23	0.8	1.52	28.4	65.7	5.9
Görcsöny A_p	Hungary, 45°58, 18°9′ E	Maize	Silt loam	3.42	0.8	1.63	26.0	62.6	11.4

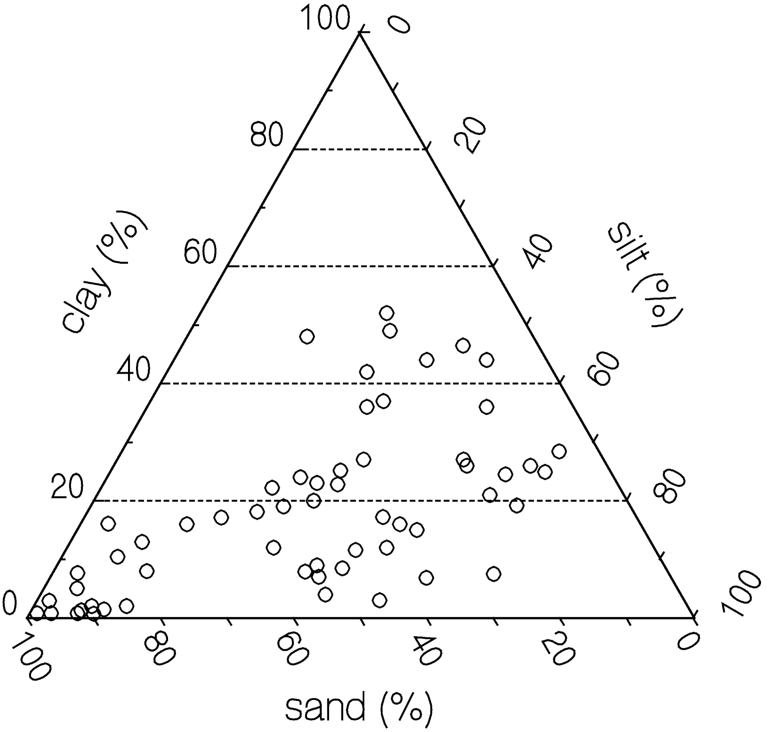

Fig. 1. Particle size distribution of the soils included in the database.

not. In applications of dual-porosity models to field data collected under natural rainfall boundary conditions, we have obtained best results when setting the boundary between pore systems in this range (i.e. −10 cm), rather than closer to saturation, at say −3 or −4 cm (e.g. Larsson and Jarvis, 1999). This implies that the preferential flow region includes not only large macropores (> 1 mm diameter), but also the smaller macropores (> 0.3 mm).

2.3. Statistical analysis

Apart from fundamental soil properties such as the particle size classes clay, silt and sand (< 2, 2 to 50 and > 50 μm), organic carbon content (either measured directly or estimated from organic matter content) and bulk density, two additional calculated or derived variables were also tested for their ability to explain variations in K_{10}, namely the geometric mean particle size (Shirazi and Boersma, 1984) and the effective porosity (Ahuja et al., 1984).

Assuming a log-normal particle size distribution, the geometric mean particle diameter d_g is given by (Shirazi and Boersma, 1984):

$$d_g = \exp(\Sigma m_i \ln d_i) \quad (1)$$

where m_i and d_i are the mass fraction and the arithmetic mean diameter of particle class i, respectively. For the size fractions used in this study, values for d_i are 0.001, 0.026 and

1.025 mm for clay, silt and sand, respectively. Campbell (1985) proposed a physico-empirical model for K_{sat} as a linear function of d_g based on the observations (i) that the air entry-pressure in the Brooks and Corey (1964) equation (with the residual water content assumed zero) appeared to be inversely related to the square root of d_g for two UK datasets (Hall et al., 1977; Bache et al., 1981), and (ii) that from a physical point of view, K_{sat} should be inversely related to the square of the air entry-pressure (Laliberte et al., 1968). Campbell's method is tested in this study for its utility in predicting K_{10}.

A generalized Kozeny–Carman equation is often used to predict saturated hydraulic conductivity from effective porosity, ε (Ahuja et al., 1984; Messing, 1989; Minasny and McBratney, 2000). Adapting the method to predict near-saturated hydraulic conductivity, we have:

$$K_{10} = B\varepsilon^n \tag{2}$$

where B and n are empirical coefficients, and the effective porosity is defined in this study as the water content at a pressure head of -10 cm H_2O minus the water content at -50 cm H_2O. The choice of the lower limit for pores effectively contributing to hydraulic conductivity is, of course, arbitrary.

In a similar way to that described above for K_{10}, water contents at -10 cm and -50 cm were estimated (if not directly measured) by interpolation from the available soil water retention measurements using the van Genuchten (1980) function. Suitable water retention data to determine effective porosity was only available for 47 of the 70 soil horizons. For the remaining 23 horizons in the database, effective porosity was estimated from soil particle size distribution, organic carbon content and bulk density using the pedotransfer functions for the parameters of the van Genuchten (1980) water retention function developed from the HYPRES database (Wösten et al., 1998). The HYPRES database consists of soil physical and hydraulic properties of ca. 5500 soil horizons collected from 18 institutes in 10 European countries.

Pedotransfer functions for K_{10} were obtained by regression analysis with each predictor variable investigated singly and also in combination (using backwards stepwise multiple regression). Only functions with significant and uncorrelated independent variables ($p < 0.05$) were accepted. Both K_{10} and the predictor variables d_g and ε were log-transformed, while the remaining independent variables tested (bulk density, organic carbon, textural fractions) were not. For the linear physico-empirical model proposed by Campbell (1985), the regression line with log–log transformed data passes through the mean values of $\log(K_{10})$ and $\log(d_g)$, with a fixed slope of unity, and the coefficient in Campbell's linear model is given by the intercept (Webster, 1989).

3. Results and discussion

Significant inter-correlations limited the number of useful functional relationships that could be derived from the fundamental soil properties (particle size fractions, bulk density, organic carbon content). Only simple functions based on the clay and silt content gave significant relationships without inter-correlation. However, for the complete dataset, only 29% of the variability was explained by soil textural fractions (Eq. (1) in Table 2). The use

Table 2

Derived pedotransfer functions (K_{10} = hydraulic conductivity at a pressure head of -10 cm in mm h^{-1}, si = % silt, > 2 and < 50 μm, c = % clay, < 2 μm, d_g is the geometric mean particle diameter in mm, ε is the effective porosity in m^3 m^{-3})

Data	Equation	Significance
(1) All soil horizons ($n = 70$)	$\text{Log } K_{10} = 0.8874 - 0.0193\ \text{c} - 0.0089\ \text{si}$	$r^2 = 0.29$, all terms $p < 0.05$
(2) Excluding arable topsoils ($n = 27$)	$\text{Log } K_{10} = 1.111 - 0.0127\ \text{c} - 0.0163\ \text{si}$	$r^2 = 0.44$, all terms $p < 0.05$
(3) All soil horizons ($n = 70$)	$\text{Log } K_{10} = 0.9 + 0.616 \log d_g$	$r^2 = 0.29$, $p < 0.0001$
(4) Excluding arable topsoils ($n = 27$)	$\text{Log } K_{10} = 1.042 + 0.637 \log d_g$	$r^2 = 0.40$, $p < 0.001$
(5) All soil horizons ($n = 70$)	$K_{10} = 21.3\ d_g$	$r^2 = 0.29$, $p < 0.0001$
(6) Excluding arable topsoils ($n = 27$)	$K_{10} = 25.2\ d_g$	$r^2 = 0.40$, $p < 0.001$
(7) All soil horizons ($n = 70$, ε predicted in 23 cases)	$\text{Log } K_{10} = 1.078 + 0.661 \text{ Log } \varepsilon$	$r^2 = 0.12$, $p < 0.01$
(8) Excluding arable topsoils ($n = 27$, ε predicted in 14 cases)	$\text{Log } K_{10} = 1.485 + 0.853 \text{ Log } \varepsilon$	$r^2 =0 .26$, $p < 0.01$

of the geometric mean particle diameter as an independent predictor variable is potentially advantageous since it integrates into one parameter the effects of texture on the water retention and hydraulic conductivity functions, thereby avoiding problems of inter-correlation (Campbell, 1985). However, least squares regression of log(K_{10}) against log(d_g) gave no improvement compared to the use of texture classes (Fig. 2, Eq. (3) in Table 2). This was also true for the method proposed by Campbell (1985), which forces a linear function between K_{10} and d_g through the origin (Eq. (5) in Table 2). The coefficient of the linear relationship for predicting K_{10} from d_g (=21.3 h^{-1}) is roughly seven times smaller than the value (=144 h^{-1}) proposed by Campbell (1985) for predicting K_{sat}.

Fig. 3 shows K_{10} as a function of effective porosity, ε, for the complete dataset. Only 12% of the variability in K_{10} was explained by ε (Eq. (7) in Table 2). Some likely reasons for this relatively poor performance can be identified, and these are discussed below, but errors in estimating effective porosity for those cases ($n = 23$) where it was not measured do not seem to be important. No significant relationships between K_{10} and ε were found for a limited dataset that included only those horizons where ε was directly measured ($n = 47$), which suggests that estimation of ε may be at least as reliable as direct measurement. Thus, ε appears to be no better than d_g as a predictor of K_{10} even though the physical basis of such a relationship ought to be stronger. In contrast to d_g, which is an easy to measure, static and relatively uniform soil property, ε is both highly dynamic and spatially variable. Thus, the estimates of ε may be subject to more error and uncertainty, especially since d_g and ε were not necessarily measured on the same sample as K_{10}, (i.e. directly under the tension infiltrometer), nor necessarily at the same time.

Thus, no more than 29% of the variation in K_{10} for the full dataset of 70 soil horizons can be explained by the predictor variables tested, and the 95% confidence intervals for predictions span over two orders of magnitude (Figs. 2–4). This is little better than

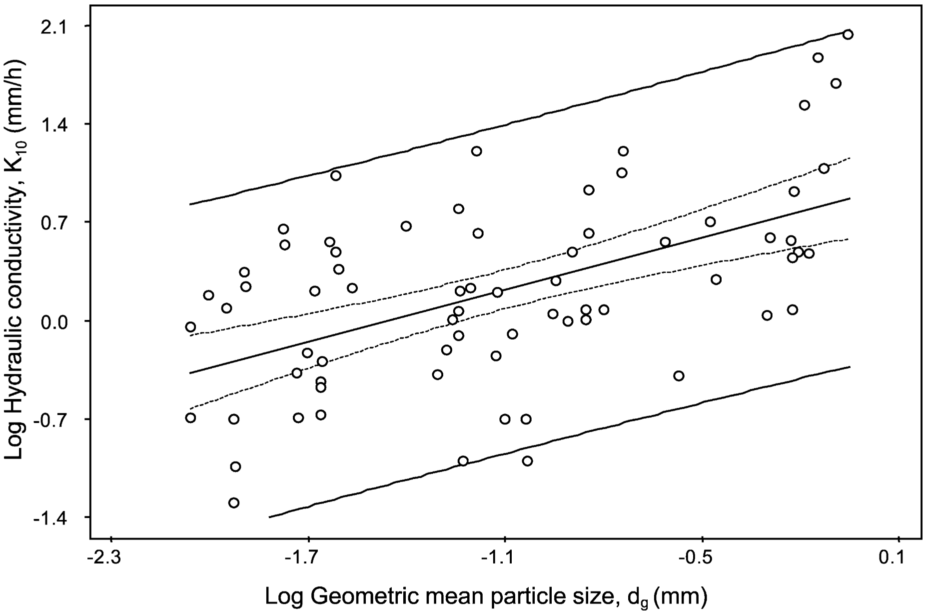

Fig. 2. Hydraulic conductivity at a pressure head of −10 cm (K_{10}) as a function of the geometric mean particle size, d_g ($n = 70$). The straight line in the center represents the fitted line (Eq. (3) in Table 2). The inside curved lines mark the 95% confidence interval for the fitted line, the outside curves mark the 95% predicted interval.

existing pedotransfer functions for K_{sat} (Tietje and Hennings, 1996; Wösten et al., 1998; Minasny and McBratney, 2000), which is somewhat surprising and disappointing. We attempted to stratify the data in order to improve the predictive power of the pedotransfer functions. Some improvements were found when arable topsoils were excluded (Eqs. (2), (4), (6) and (8) in Table 2, Fig. 4). This may be partly attributed to the large temporal changes in hydraulic conductivity due to disturbance by tillage, and in sensitive soils, crust formation due to exposure to rainfall impact when the surface cover is minimal. For example, Messing and Jarvis (1993) showed that hydraulic conductivity at pressure heads between −5 and −10 cm varied over one order of magnitude through one growing season (June to October) on Ultuna silty clay. The value of K_{10} measured directly at the soil surface with tension infiltrometers will be extremely sensitive to the presence of even thin surface crusts of low permeability. It is likely that the different sub-sets included in the study for arable topsoils were measured at different times of the year in relation to tillage, and on soils with different susceptibilities to surface sealing.

The possibility of operator bias was also considered since the complete dataset of 70 horizons consists of a number of sub-sets of data obtained by different individuals. Small, seemingly unimportant, differences in the way the tension infiltrometer is used may give significantly different estimates of K_{10}. The thickness of the contact sand layer may be important, since it influences the actual supply pressure head at the soil surface (Reynolds and Zebchuk, 1996). However, this should be most critical close to saturation, and less important at the supply pressure potential of −10 cm used here. The antecedent moisture

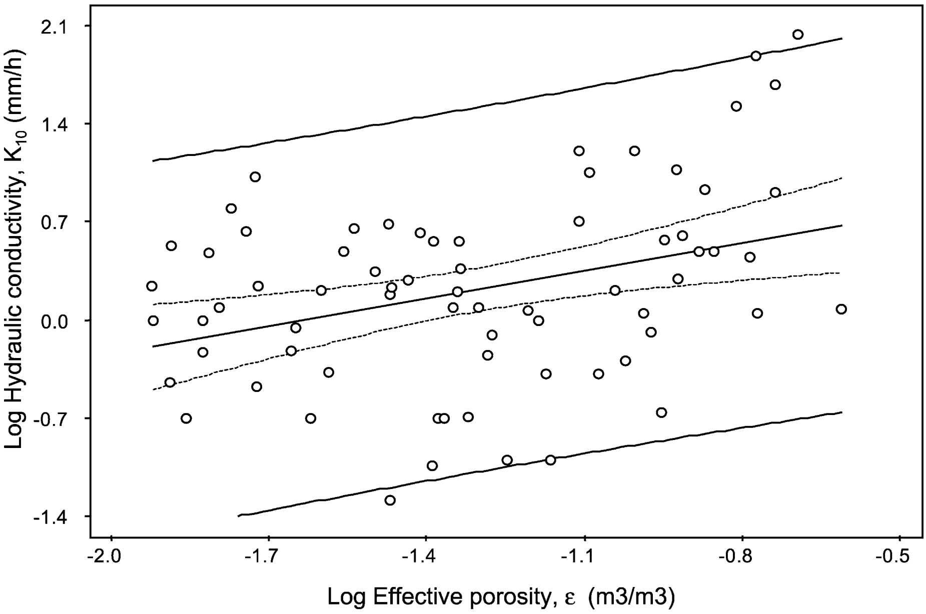

Fig. 3. Hydraulic conductivity at a pressure head of -10 cm (K_{10}) as a function of effective porosity, ε, defined as the water content at a pressure head of -10 cm minus the water content at -50 cm. Complete dataset ($n = 70$). The straight line in the center represents the fitted line (Eq. (7) in Table 2). The inside curved lines mark the 95% confidence interval for the fitted line, the outside curves mark the 95% predicted interval.

content of the contact sand layer is also important. Use of air-dry sand as a contact layer may give larger estimates of hydraulic conductivity compared to pre-wetted moist sand (Smettem and Clothier, 1989), since it may fall into surface-vented macropores, thereby acting as a wick. Poor hydraulic contact between the sand and the infiltrometer, which may occur in windy conditions for example, may also lead to erroneous estimates of hydraulic conductivity that may not be easy to detect (Vandervaere et al., 2000a). The length of time the user is prepared to wait to reach 'steady-state', and the criteria adopted for assessing whether steady-state has been reached, may also vary between users.

Another important source of inter-user variability is likely to be the treatment of the soil surface: in nearly all cases the measurements were made directly at the surface, irrespective of any surface seals, but in a few cases, a surface seal was removed before the measurements were made (B. Mohanty, personal communication). Interestingly, the r^2 values are improved slightly if the Swedish sub-set is excluded from the analysis. For given values of d_g or ε, K_{10} values in the Swedish data appear consistently smaller than for the other sub-sets. However, the Italian dataset was obtained using exactly the same methods as those adopted in Sweden (L. Zavattaro, personal communication), which would tend to suggest that user-bias may not have been a significant factor. The most likely explanation is that some of the Swedish soils were sealed at the time of the measurements. Indeed, in several cases, surface seals were noted (e.g. Kjettslinge, hilltops at Näsbygård, Ultuna in August and in sewage-sludge amended soil).

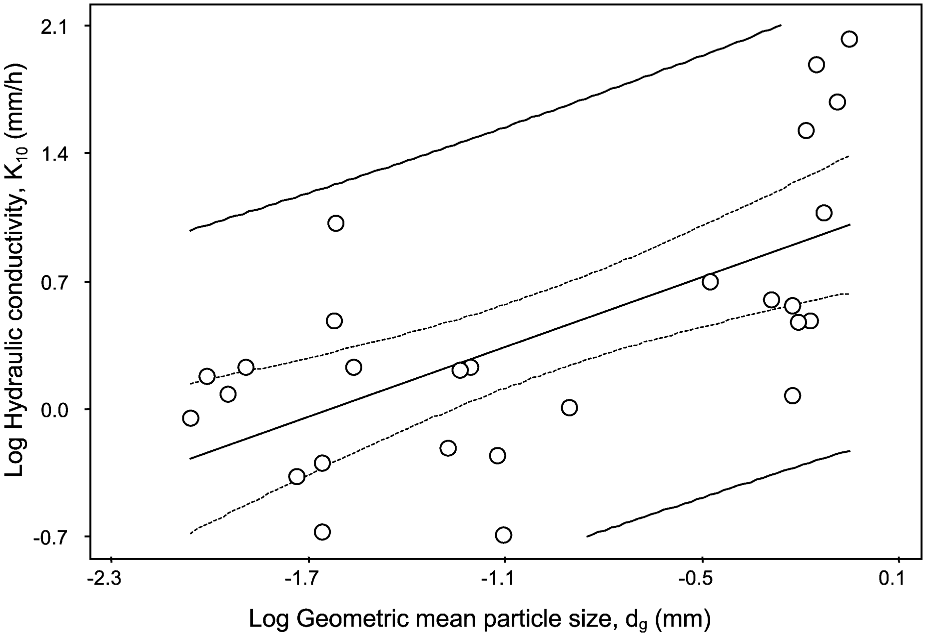

Fig. 4. As Fig. 2, but excluding data for arable topsoils ($n = 27$, the fitted regression line is Eq. (4) in Table 2).

Pedotransfer functions based on more than three measured particle size fractions would most likely explain more of the variability in K_{10}, since the geometric mean particle size would be more reliably estimated (Tietje and Hennings, 1996), and the particle size distribution index or geometric standard deviation of particle size could also be used as a predictor variable (e.g. Bloemen, 1980; Rajkai et al., 1996). However, this kind of function would not be so widely useable, since soil survey data is generally limited to three texture fractions.

Finally, the indirect method used to 'measure' K_{10} may be another important source of error. It relies on the theory outlined by Wooding (1968) to convert three-dimensional steady-state infiltration rate to one-dimensional hydraulic conductivity assuming homogeneous, isotropic, soil at a uniform initial water content. These conditions may not always be fulfilled in the field. In addition, in many soils, lateral capillary flow may dominate the vertical gravity flow component at a supply pressure head of -10 cm, even at steady-state, so estimating K_{10} may be inherently prone to error (Vandervaere et al., 2000b).

4. Conclusions and recommendations

It is clear that predicting near-saturated hydraulic conductivity remains difficult and uncertain. In addition to error in measuring K_{10} itself, we consider that one of the most important sources of unexplained variation in K_{10} is variation in the pore structure of arable soils due to loosening by tillage, subsequent consolidation and the formation of surface seals. For site-specific model applications, direct measurements are therefore both relatively simple to make and more reliable. However, in the absence of any direct measurements, we recommend that Eqs. (4) or (6) in Table 2 should be the preferred methods to predict the hydraulic conductivity at a pressure head of -10 cm. Although there is clearly large uncertainty in predicting K_{10} for any given soil, these functions could nevertheless prove useful in large-scale (regional) modelling applications where direct measurements are not practical, and where it is sufficient to distinguish broad differences in hydraulic behaviour between soil types. These K_{10} values can be used as (i) the 'matching point' hydraulic conductivity for prediction of unsaturated hydraulic conductivity in single domain models, or (ii) the saturated matrix hydraulic conductivity in dual- and multi-porosity models. Although the approach shows promise, it should be tested on a wider data set, and the effects of the prediction uncertainty should also be assessed in a functional sense (Wösten et al., 1990). This may indicate the need for further development and refinement.

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