HYDRAULIC CONDUCTIVITY IN UPLAND BLANKET PEAT – MEASUREMENT AND VARIABILITY

J. Holden[1]* and T.P. Burt[2]

[1]School of Geography, University of Leeds, Leeds, LS2 9JT, UK.

[2]Department of Geography, University of Durham, Science Laboratories, South Road, Durham, DH1 3LE, UK.

*Corresponding author: Tel: +44 (0) 113 233 3317, Fax : +44 (0) 113 233 3308, Email: j.holden@geog.leeds.ac.uk

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ABSTRACT

A key parameter used in wetland hydrological and landform development models is hydraulic conductivity. Head recovery tests are often used to measure hydraulic conductivity but the calculation techniques are usually confined to rigid soil theory. This is despite reports demonstrating the misapplication of rigid soil theory to non-rigid soils such as peats. While values of hydraulic conductivity calculated using compressible techniques have been presented for fenland peats these data have never, to the authors' knowledge, been compared to such calculations in other peat types. Head recovery tests (slug withdrawal) were performed on piezometers at depths ranging from 10 cm to 80 cm from the surface on north Pennine blanket peats. Results were obtained using both rigid and compressible soil theories allowing comparison of the two techniques. Compressible soil theory gives values for hydraulic conductivity that are typically a factor of five times less than rigid soil calculations. Hydraulic conductivity is often assumed to decrease with depth in upland peats but at the study site in the northern Pennines it was not found to vary significantly with depth within the range of peat depths sampled. The variance within depth categories was not significantly different to the variance between depth categories showing that individual peat layers did not have characteristic hydraulic conductivity values. Thus large lateral and vertical differences in hydraulic conductivity over short distances creates problems for modelling but may help account for the high frequency of preferential flow pathways within what is otherwise a low matrix hydraulic conductivity peat. Hydraulic conductivity was found to vary significantly between sampling sites demonstrating that hillslope or catchment-scale variability may be more important than plot-scale variability. Values for compressibility of the peats are also reported. These generally

decline with depth and also vary significantly between sampling sites. There are implications for the way in which measurements of hydraulic conductivity and other properties of blanket peat are interpreted as the effects of environmental change in one part of a peat catchment may be very different to those in another.

KEYWORDS

Hydraulic conductivity, Runoff, Piezometers, Consolidation, North Pennines, Peat

INTRODUCTION

The movement of water in peats is important for ecology, catchment hydrology and even in determining the shape of raised mires (Ingram, 1982). In particular, hydraulic conductivity is a key parameter used in predictive hillslope or floodplain hydrological models. However, there are few detailed measurements of how hydraulic conductivity in blanket peats varies with depth or between slopes in a catchment. This would clearly be of importance for spatially distributed modelling of catchment hydrology. Hydraulic conductivity measurements are often made using head recovery tests where slugs of water are either added to or removed from piezometers and the recovery to the original water level in the instrument is recorded. In poorly humified peats these tests give results consistent with the behaviour expected from incompressible or rigid soils (Rycroft *et al*., 1975). In humified peat, however, reports have suggested that hydraulic conductivity was dependent on the size of the head difference between the piezometer and the surrounding peat. This has been attributed to non-Darcian flow processes within the peat (Rycroft *et al*., 1975; Waine *et al*., 1985) while others have suggested that the effect can be explained by matrix compression and swelling which causes variable water storage within the peat (Brown and Ingram, 1988; Hemond *et al*., 1984, and

Hemond and Goldman, 1985). The effect of compression and swelling of peat on head recoveries is not well understood (Baird and Gaffney, 1994). Therefore it is important to use both rigid and compressible soil theory on piezometer tests in peatlands. Baird and Gaffney (1994) applied this technique to a fenland peat and found that compression and swelling did affect the course of head recovery in the 16 piezometers they tested. However they found that both rigid and compressible methods gave values of hydraulic conductivity that were too high. Nevertheless, they advocated use of compressible soil theory in peats to allow a standard comparison of hydraulic and storage properties between different peat types.

Since Baird and Gaffney's paper there have, to the authors' knowledge, been very few applications of compressible soil theory in peatlands so that comparison of hydraulic and storage properties between peat types remains difficult. Thus the wider representativeness of results presented becomes difficult to establish. Such information is of particular importance, for example, for development of runoff production models in peatlands, for wetland restoration strategies and for process analysis and prediction of common slope failures in upland peats (Dykes and Kirk, 2001). This paper will compare rigid and compressible techniques applied to humified blanket peat and compare results to those obtained by Baird and Gaffney (1994) for a poorly decomposed fenland peat. Variation in hydraulic conductivity within the blanket peats will also be analysed demonstrating that depth and individual peat layers are not significant controls, but differences between hillslopes may be important.

STUDY SITE

The experiments were performed at the Moor House National Nature Reserve (NNR), North Pennines, UK (54^o 65' N, 2^o 45' W). Moor House NNR is an area of moorland which straddles the summit ridge of the northern Pennines, a chain of hills running north-south in central northern England. A series of alternating, almost horizontal, beds of limestone, sandstone and shale of Carboniferous age provide a base for a boulder clay on top of which lies approximately 2 m of blanket peat. Peat formation was initiated at the Boreal-Atlantic transition about 7500 years ago when rainfall increased markedly and the presence of glacial boulder clay caused impeded drainage and waterlogging. Mean annual rainfall is 1982 mm with an average of 244 precipitation days per year (Holden and Adamson, 2001). Climate on the reserve can be classified as sub-arctic oceanic (Heal and Smith, 1978). The dominant vegetation type on the blanket peat is a *Calluna-Eriophorum-Sphagnum* association. The upper 5 cm of the intact vegetated soil consists of poorly humified (H2-H3 on the Von Post (1922) scale) black brown coloured peat with living roots and a crumb structure. Below this to 10 cm the peat tends to be brown and slightly humified (H3-H4) overlying a darker brown *Eriophorum-Calluna-Sphagnum* peat (H4). The soil then very gradually becomes more humified with depth. By 1.5 m into the profile the peat is highly humified with decomposition almost complete (H9). Occasionally the peat deposits contain distinctive yellow/orange layers dominated by *Sphagnum* remains. While the structure of these layers depends on the dominance of particular vegetation species, generally no significant differences in dry bulk density (DBD) or throughflow runoff production can be discerned from these layers when compared to surrounding layers. Dry bulk densities range from 0.15 g cm^{-3} at the surface to 0.18 g cm^{-3} at 20 cm depth. The DBD gradually increases to 0.27 g cm^{-3} by 50 cm into the peat mass. Further details on the lower layers of the blanket peat at the study site, including pollen analysis can be found in Johnson and Dunham (1963) and Heal and Perkins (1978).

METHODOLOGY

Calculations

Full treatment of the calculations and derivations used are provided by Brand and Premchitt (1982) and Baird and Gaffney (1994) and so only a summary is provided below. Hvorslev's (1951) used the basic differential equation that describes saturated flow through a falling head permeameter to produce a solution to the pore pressure equalisation process between a piezometer system and an incompressible soil. The solution is:

$$\frac{u-u_0}{u_\infty-u_0}=1-\exp\left(\frac{-Fkt}{V\gamma_w}\right) \qquad [1]$$

in which u_0 is the initial pressure head; u_∞ is the equalisation pressure head; u is the pore pressure in a piezometer at time t after equalisation begins pressure head in a soil of hydraulic conductivity k; γ_w is the unit weight of water and V is the volume of water required to flow into or out of the piezometer system to equalise a unit pressure difference between the piezometer and the surrounding soil. In a standpipe piezometer V is numerically equal to the cross-sectional area of the piezometer (Baird, 1995). F is the shape factor (dimensions of length) which describes the flow field geometry around the piezometer (Kirkham, 1945; Hvorslev, 1951; Youngs, 1968; Brand and Premchitt, 1980). For the present study, the shape factor (units of length) has been determined from the equation of Brand and Premchitt (1980):

$$F = 7\,d + 1.65\,l \qquad [2]$$

where d is the diameter of the tip (25 mm) and l the tip length (50 mm). Compression and swelling of soil around a piezometer may play a major part in piezometer response if the soils are compressible and equation 1 may not adequately describe the equalisation process (Baird and Gaffney, 1994). To analyse compression and swelling on head recovery in a piezometer the effective stress equation can be used:

$$\sigma' = \sigma_T - u \qquad [3]$$

where σ' is the effective stress and σ_T is the total stress. This accounts for the change in volume of the voids of a soil that will occur under if the soil undergoes a change in state of stress. Equation 3 describes the state of stress in a soil. Immediately after slug withdrawal there will be an increase in effective stress around the piezometer tip as pore water pressure decreases while the total vertical stress remains the same. As the water level recovers, effective stress will decline causing more water to enter storage and increase the rate of head recovery (Baird, 1995). For cylindrical piezometers in compressible soil the rate of pressure head recovery is given by the Laplace consolidation equation in axisymmetrical cylindrical coordinates r and z (Al-Dhahir and Morgenstern, 1969):

$$c\left(\frac{\partial^2 u}{\partial r^2} + \frac{1}{r}\frac{\partial u}{\partial r} + \frac{\partial^2 u}{\partial z^2}\right) = \frac{\partial u}{\partial t} \qquad [4]$$

where r is the radial distance from piezometer tip mid-point, z is the vertical distance from piezometer mid-length, and c is the coefficient of 'consolidation' that accounts for both compression and swelling. Brand and Premchitt (1982) used a numerical solution to equation 4 to show that the soil - piezometer system was well represented by a control parameter:

$$\lambda = \frac{4\pi a^2 bm}{V} \qquad [5]$$

where a is the outside radius, b the half length of the piezometer tip, and m the coefficient of volume compressibility of the soil. The shape of the head recovery is characterised by λ for which there is a unique ratio between t_{90} (time taken for the head to recover to 90 % of initial head difference between piezometer and soil) and t_{50} (Premchitt and Brand, 1981; Brand and Premchitt, 1982). Using λ as a control parameter, Brand and Premchitt (1982) derived equalisation monographs based on t_{50} and t_{90} that can be used to calculate the hydraulic conductivity and the coefficient of consolidation. These have been used in the following analysis.

Field instrumentation

Networks of thin PVC piezometers with inside diameter of 14 mm and porous plastic tips of 25 mm outside diameter and tip lengths 50 mm, were installed at three sites (S1, S2 and S3) on the Moor House reserve. S1 was an area of intact blanket peat with an *Eriophorum, Calluna* and *Sphagnum* vegetation cover. The peat was around 80 cm deep with a mean slope of 0.09 m m^{-1}. At a distance of approximately 200 m from S1, the second hillslope (S2) had similar surface cover characteristics, with a mean slope of 0.07 m m^{-1} and a mean peat depth of 1.2 m. S3 was located approximately 400 m from S1 near the summit of Burnt Hill which is an area of eroded peat. The hydrology of Burnt Hill was examined in a water balance approach by Conway and Millar (1960). S3 was located in intact peat close to the head of a gully network. The peat was approximately 230 cm deep with a mean slope of 0.03 m m^{-1}. The slope had been subject to severe burning in 1950 but is now fully revegetated with an *Eriophorum and Calluna* cover. *Sphagnum* was present in bog pools on the slope. Ten piezometer nests were installed at each of the three sites. Each nest (coded A-J) consisted of piezometers with tip mid-points at 10, 20, 35, 60 and 80 cm depth. A thin borehole was created with a screw auger and the tubes slotted into position. The piezometers were in position for at least six months before the tests were performed in order to ensure stress-adjustment lags caused by the installation were minimal (Baird and Gaffney, 1994).

RESULTS

Comparison of rigid and compressible techniques

Figure 1 shows head recoveries from two of the piezometers. As found by Baird and Gaffney (1994) all of the recoveries deviated to a greater or lesser extent from rigid soil theory. For example, the results from piezometer S2 C60 correspond quite closely to the response described by Hvorslev (1951); most of the data points (closed circles) are close to the curve. However, S2 C20 shows pronounced deviation from the curve with the data points (open triangles) forming a more gently sloping curve than that described by Hvorslev (1951). Table 1 presents mean values from the experiments for each site and depth using both equation 1 and values calculated from the nomograph of Brand and Premchitt (1982). Values of hydraulic conductivity tend to be higher at S1 than the other two sites. Typically the compressibility of the peats at S1 are also greater. Mean values of hydraulic conductivity using t_{50} were greater than the mean values calculated using $t_{90.}$ In fact, this was the case in all but two of the piezometers (Figure 2) such that Student's t-test (on logarithmically transformed data) indicates that log k_{50} is significantly greater than log k_{90} at $p < 0.0001$ ($T = 5.19$). This is because equation 1 fails to account for variable storage and release of water giving an apparent increase in hydraulic conductivity early in the head recovery (Baird, 1995). Hvorslev's (1951) theory appears to be invalid for all the piezometers.

Comparing the hydraulic conductivity values calculated using both theories shows that the mean value of k^* (the hydraulic conductivity calculated using the response time charts derived numerically by Brand and Premchitt (1982)) was much lower than k_{90}. Baird and Gaffney (1994) reported that both rigid and compressible soil theories gave values of hydraulic conductivity for each piezometer installation at their fenland site within a factor or two of each other. This is not the case for the Moor House blanket peats where generally the difference is a factor of around five (in 28 cases, N = 82) but can be as high as a factor of ten (in two cases). A t-test on logarithmically transformed data suggests that log k_{90} is significantly greater than log k^* at $p < 0.0001$ ($T = 27.57$). Hvorslev (1951) suggests that reliable estimates of hydraulic conductivity in compressible soils can only be calculated using equation 1 when exchanges to and from storage are nearly complete at the end of the head recovery process (e.g. $t_{99.9}$). Baird and Gaffney (1994), however, in their fenland peat study found that values of k^* were often closer to k_{50} than k_{90} and concluded that both Hvorslev's (1951) and Brand and Premchitt's (1982) theories give values of hydraulic conductivity that are too high. Results from blanket peat at Moor House do not indicate that Brand and Premchitt's method gives hydraulic conductivity values that are too high because k^* values are much closer to k_{90} than k_{50}. The fact that k^* is much lower than k_{90} at Moor House is one of the major differences between the results from the north Pennine blanket peat and those from the fenland peat reported by Baird and Gaffney (1994).

Comparison of hydraulic conductivity values by depth and site

The hydraulic conductivity values from 10 cm to 80 cm depth in blanket peat are generally an order of magnitude lower than those measured in the Somerset Levels (Baird and Gaffney, 1994). Analysis of variance (ANOVA) suggests that, for depths equal to or greater than 10 cm, depth is not a significant control on k^* (Table 2). Even at 10 and 20 cm depth hydraulic conductivity can be as low as 3.43 x 10^{-7} cm s^{-1} and 1.78 x 10^{-7} cm s^{-1} respectively. There is often an assumption that k decreases gradually with depth (or decomposition) in peat (e.g. Ingram, 1983). Rycroft *et al.* (1975) extensively reviewed reported hydraulic conductivity values from peats (using rigid soil theory and hence comparison must be treated with caution) and their table of values suggests k for blanket peats ranging from 1.1 x 10^{-5} cm s^{-1} at 30 cm depth (Galvin and Hanrahan, 1967) to 6 x 10^{-8} cm s^{-1} at 1 m (Ingram, 1967). Values in other peats tend to be slightly higher (e.g. Dai and Sparling, 1973; Neuman and Dasberg, 1977). In poorly decomposed fenland peats values as high as 5 x 10^{-3} cm s^{-1} have been reported at 1 m depth (Rycroft *et al.*, 1975). In the Moor House blanket peats mean k^* between 10 and 80 cm depth is 2.9 x 10^{-6} cm s^{-1} (although this would have been calculated at 1.28 x 10^{-5} cm s^{-1} if rigid soil theory was used at t_{90}). There is no evidence therefore to suggest that k^* decreases significantly with depth (or decomposition) at Moor House between 10 and 80 cm. In fact k^* at 80 cm depth can sometimes be greater than k^* at 10 depth within the same piezometer nest. Figure 3 shows that mean k^* is slightly greater at 10 and 20 cm depth than at other depths but there is a significant amount of overlap such that there is no significant decrease with depth. Hydraulic conductivity at 80 cm depth at S2 ranged over almost two orders of magnitude from 9.70 x 10^{-8} to 6.32 x 10^{-6} cm s^{-1}. Hence single peat layers cannot be characterised by typical hydraulic conductivity values. Lateral variation in hydraulic conductivity can often be just as high as vertical variation. Applying hydraulic conductivity measurement to hydrological modelling therefore becomes problematic.

Holden and Burt (2000) showed that the blanket peats in the north Pennines were dominated by flow within the top 10 cm of the peat mass. Thus, the low hydraulic conductivity of the peat below 10 cm depth is responsible for rapid development of saturation within the near-surface peat during a rainfall event resulting in the production of saturation-excess overland and subsurface (throughflow) stormflow. It is only in the upper few centimetres of the peat mass that hydraulic conductivities are sufficiently high to allow rapid throughflow generation. Holden *et al.* (2001) showed that infiltration-excess overland flow was a rare occurrence in these blanket peats given high hydraulic conductivities at the peat surface. This near-surface peat layer, the acrotelm (Ingram, 1978), is much thinner than found in many other peatlands where an active acrotelm may be as deep as 80 cm (see Ingram, 1983). This may be a result of the much wetter conditions (2000 mm precipitation per annum) in the north Pennines such that high water tables are more readily maintained than at sites with lower rainfall totals (Evans *et al*., 1999).

Holden and Burt (2000) and Holden *et al.* (2001) showed that runoff could be produced through preferential routes in the deeper peat layers. Water moving along lines of weakness in the peat may eventually result in the development of soil pipes which are common upland humid soils (Bryan and Jones, 1997). Jones (1981) noted that piping is often found where there is a sudden change in soil properties, hydraulic conductivity in particular. In the peats at Moor House measurements suggest that sudden changes of hydraulic conductivity of up to two orders of magnitude within a few centimetres both laterally and vertically are common. Thus preferential flow and pipe development are likely to be dependent upon local hydraulic gradients and the connectivity of the pathways. Indeed piping is common in the peats of the North Pennines, often producing more than 10 % of catchment runoff (Holden and Burt, in press). Thus large variations in hydraulic conductivity at such small-scales may be extremely important in determining pathways for a large proportion of runoff. This has crucial implications for runoff and water quality modelling and for our understanding of slope stability in these environments.

Hydraulic conductivity does vary significantly with site (Table 2). Figure 3 shows that S1 tends to have greater $k*$ values than S2 which also tends to have greater $k*$ than S3. With S1 and S2 located only 200 m apart and with very similar vegetation, slope and peat depth characteristics, this indicates the difficulties of generalising catchment-scale hydraulic conductivity based on only a few measurements. Hillslope-scale and catchment-scale variability may be more important than plot-scale variability. The lower $k*$ values on the gullied slope (S3) may be because dense peat dissection lead to a lowering of the water table and enhanced decomposition of the surficial peat. It may also be that the peat itself naturally has different properties to those a few hundred metres away as seen in the case of S1 and S2. It may be for these reasons that water balance approaches to blanket peat hydrology have provided conflicting evidence for the effects of land use change in wetland areas (e.g. moorland drainage – c.f. Conway and Millar, 1960 with Burke, 1975). Simply, the local properties of the peat itself may result in different responses to the same land use change in different locations. Thus process-based approaches are required to help us predict the effects of future climate and land use change and associated wetland remediation techniques.

Peat consolidation

With water contents ranging from 75% to 98% by volume peat is an extremely compressible material (Hobbs, 1986). In a compressible soil the ratio t_{50}/t_{90} will always be greater than 3.322 (rigid soil t_{50}/t_{90} = 3.322) and will increase with the volume of compressibility of soil (Premchitt and Brand, 1981). All of the t_{50}/t_{90} ratios (essentially a measure of the effect of compressibility on the head recovery) were above 3.322 for the blanket peat piezometers (e.g. see Table 1). Values of the coefficient of consolidation, *c*, could be important for wetland management. Price and Schlotzhauer (1999), for example, concluded that most peatland water balances should take account of storage changes associated with peat volume changes, and that peat volume changes may increase water limitations to plants. This may be more important on damaged peatlands than on intact sites. Values of *c* could also be important in modelling water flow in peats which are subject to rapid changes in pore water pressures. This may occur when heavy rainfall follows a prolonged dry period resulting in changes to effective stress and soil water storage. Most slope failures of peats in the north Pennines are associated with this sort of hydrometeorological condition. The nine values of *c* determined by Baird and Gaffney (1994) ranged from 0.56 at 2 m to 13.23 at 1.2 m depth for a poorly humified fenland peat. Their values fall within the three orders of magnitude variation found at Moor House where *c* varied from 0.03 at S3 C80 to 90.25 at S1 C10. ANOVA demonstrates that both depth and site are genuine controls on *c* (Table 3). Figure 4 shows that c generally declines with depth (although values for c tend to be greater at 35 cm depth than at 20 cm). These values are skewed by the high *c* values found in the 35 cm layer at S1. It seems that occasionally some layers can have particular characteristics demonstrating that layering of the peat may be more important than depth for some parameters. The coefficient of consolidation is significantly greater at S1 than S2 and *c* at both S1 and S2 is significantly greater than S3. As with hydraulic conductivity, this suggests that hillslope-scale variability may be more important than plot-scale variability and this may have implications for wetland management schemes. A restoration strategy on one hillslope may not necessarily work on the next hillslope because the peat properties (e.g. hydraulic conductivity, compressibility, storativity) may be very different.

CONCLUSIONS

In line with the findings of Baird and Gaffney (1994) who examined a fenland peat, head recoveries in all the piezometers deviated from rigid soil theory. Rigid soil theory gives values of hydraulic conductivity that are too high in blanket peats. Baird and Gaffney (1994) suggested that the technique of Brand and Premchitt (1982) also gave values of k^* that were too high because values of k^* were often closer to k_{50} than k_{90}. There is no evidence to suggest that this is the case in blanket peats since k^* values were much closer to k_{90} than k_{50}. Nevertheless, the magnitude of difference between rigid and compressible calculations was far greater in the blanket peat of the northern Pennines than in the fenland peat of the Somerset Levels. Generally, values of k^* in blanket peat were a factor of five to ten times less than values of k_{90} compared to a factor of two for poorly decomposed fenland peat. This suggests that it is even more important to use compressible soil theory when calculating hydraulic conductivity and other soil properties in these upland peats. Within the blanket peat mass, hydraulic conductivity values do not differ significantly with depth (at least between 10 and 80 cm depth), but they do vary significantly between sampling sites. The coefficient of consolidation decreased significantly with depth and also varied significantly between sampling sites. The results for k^* and c suggest that peat properties can be significantly different on intact peat slopes with similar slope angles and surface cover within a short distance of each other. The peat properties at the head of a gully network on an eroded slope were also significantly different with lower k^* and c values than the other sites. It is therefore not only important to apply compressible soil theory in peatlands (including wetland riparian zones) but also important to establish the spatial controls on peat properties so that process-based assessment of the effects of land use and climate change can be made. This information will also allow hydrological and slope stability modelling to be properly informed.

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Table 1. Mean values of hydraulic conductivity calculated using rigid and compressible soil theories, standard deviations given in brackets.

		Hvorslev (1951)		Brand and Premchitt (1982)	
Site/depth, cm	t_{90}/t_{50}	k_{50} x 10^{-6} cm s^{-1}	k_{90} x 10^{-6} cm s^{-1}	k^* x 10^{-6} cm s^{-1}	c x 10^{-3} cm^2 s^{-1}
S1					
10	3.9 (0.5)	73.5 (25.4)	57.5 (40.9)	9.8 (4.4)	46.4 (38.1)
20	12.3 (2.1)	58.6 (47.9)	25.9 (32.7)	4.3 (3.62)	5.9 (8.6)
35	3.8 (0.4)	14.3 (14.5)	12.0 (13.8)	2.4 (2.7)	42.6 (37.6)
60	3.6 (0.2)	16.3 (13.0)	18.6 (10.7)	2.7 (1.9)	10.8 (11.4)
80	4.7 (1.7)	1.9 (1.8)	1.2 (0.8)	0.2 (0.1)	1.2 (1.4)
Mean	6.0 (4.5)	33.1 (35.9)	23.9 (28.8)	4.0 (4.1)	23.0 (29.8)
S2					
10	3.4 (0.1)	0.9 (0.1)	0.8 (0.1)	0.2 (0.0)	2.9 (0.5)
20	4.0 (0.2)	1.8 (0.2)	2.5 (0.2)	0.3 (0.1)	3.2 (1.9)
35	3.9 (0.0)	16.3 (21.9)	2.3 (1.8)	0.5 (0.3)	4.7 (0.7)
60	13.6 (3.2)	24.0 (21.6)	6.0 (1.4)	0.9 (0.4)	2.3 (3.3)
80	3.9 (2.2)	0.6 (1.2)	0.5 (0.1)	0.1 (0.1)	0.5 (2.8)
Mean	6.1 (6.9)	10.4 (18.8)	2.5 (2.3)	0.5 (0.4)	3.1 (4.7)
S3					
10	4.0 (0.6)	39.4 (50.9)	32.6 (41.8)	10.4 (12.9)	39.5 (52.3)
20	11.0 (1.5)	16.5 (25.2)	6.9 (6.5)	2.2 (1.7)	0.8 (0.5)
35	3.6 (1.1)	0.6 (0.1)	0.6 (0.6)	0.1 (0.1)	0.8 (0.5)
60	4.1 (1.1)	0.5 (0.2)	0.4 (0.1)	0.1 (0.0)	0.7 (0.7)
80	7.6 (1.6)	7.6 (10.4)	3.2 (4.2)	0.4 (0.4)	0.5 (0.5)
Mean	6.6 (4.5)	11.5 (22.3)	8.5 (18.2)	2.5 (5.8)	7.7 (22.8)
Mean	6.2 (5.2)	19.6 (29.2)	12.7 (22.3)	2.4 (4.2)	12.3 (23.6)

t_{90}/t_{50} = 90% equalisation time divided by 50 % equalisation time, k_{50} = hydraulic conductivity calculated using 50 % equalisation time and equation 1, k_{90} = hydraulic conductivity calculated using 90 % equalisation time and equation 1, k^* = hydraulic conductivity calculated using the response time chart of Brand and Premchitt (1982), c = coefficient of consolidation

Table 2. Analysis of variance of hydraulic conductivity, log data.

Source	Degrees of freedom	F ratio	Probability > F
Depth	4	1.29	0.283
Site	2	6.90	0.002

Table 3. Analysis of variance of the coefficient of consolidation, log data.

Source	Degrees of freedom	F ratio	Probability > F
Depth	4	3.71	0.035
Site	2	4.82	0.004

Figure captions

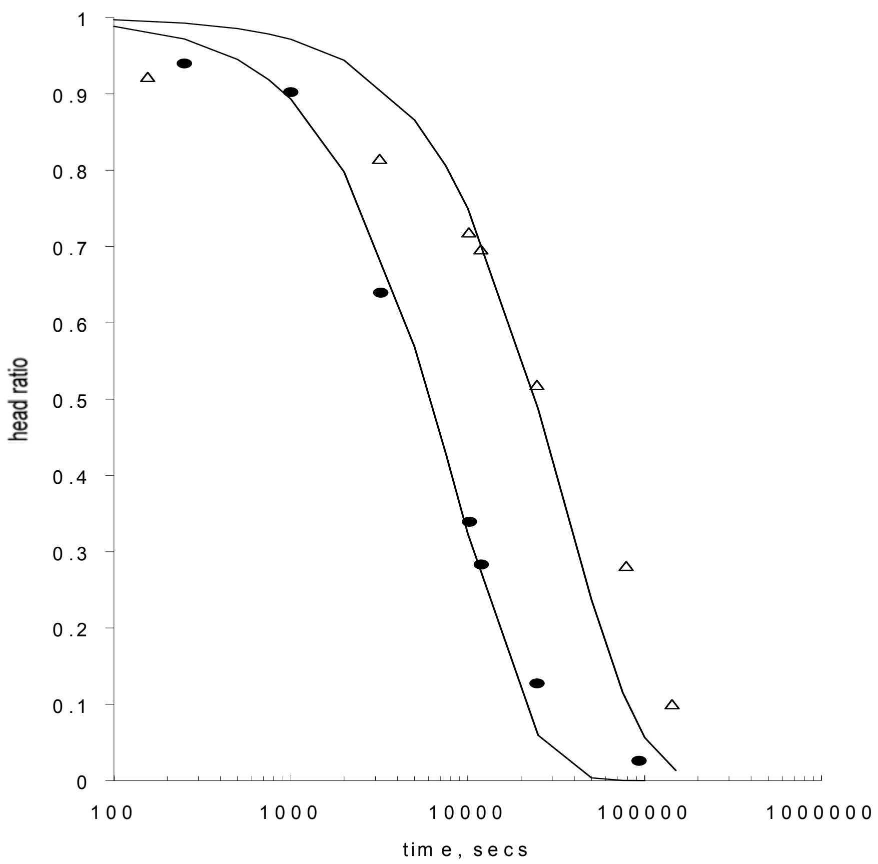
Figure 1. Example piezometer head recoveries. Closed circles for S2 C60, open triangles for S2 C20. The solid lines are fitted responses (least differences) according to equation 1.

Figure 2. Scatterplot of k_{90} against k_{50} showing the consistently lower values of K calculated by equation 1 in early head recovery time for each piezometer.

Figure 3. Geometric mean and 95 % confidence interval of k^* for each category of a) depth and b) site.

Figure 4. Geometric mean and 95 % confidence interval of the coefficient of consolidation for each category of a) depth and b) site.

Figure 1.



head ratio
1
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
100
1000
10000
100000
1000000
time, secs

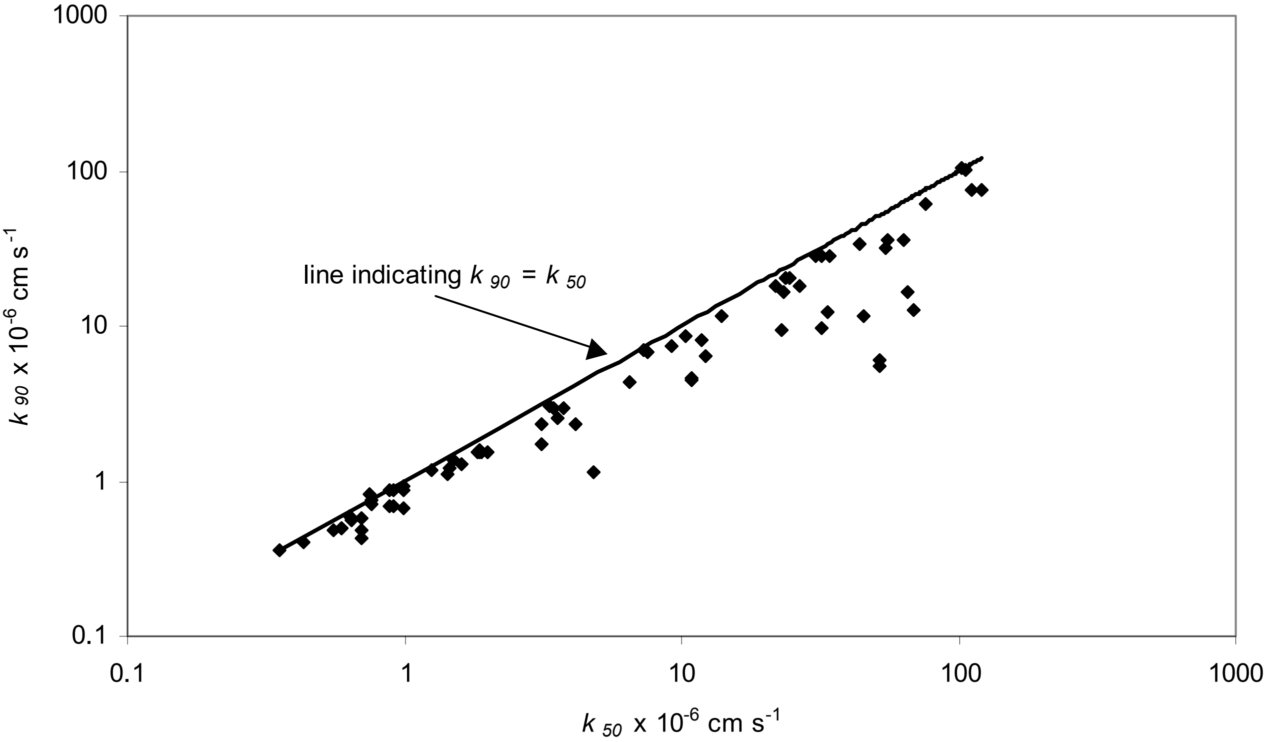

1000
100
10
1
0.1
0.1
1
10
100
1000
line indicating k_{90} = k_{50}
k_{90} x 10^{-6} cm s^{-1}
k_{50} x 10^{-6} cm s^{-1}

Figure 2

Figure 3.

a)

```
-+-------------+---------+-----------+--
              (-----------*-----------) 10 cm
           (----------*---------)       20 cm
    (---------*--------)                35 cm
     (---------*---------)              60 cm
    (--------*---------)                80 cm
-+-------------+---------+-----------+--
 0.1          0.4       1.0         3.0
```

k, cm s^{-1} x 10^{-6} (log scale)

b)

```
-+-------------+---------+-----------+--
                       (------*-------) S1
            (------*-------)            S2
    (------*-------)                    S3
-+-------------+---------+-----------+--
0.1           0.4       1.0         3.0
```

k, cm s^{-1} x 10^{-6} (log scale) a)

```
---+----------+--------+-----------+-
                   (------*------)     10 cm
          (------*-----)               20 cm
                 (------*-----)        35 cm
         (------*------)               60 cm
 (------*------)                       80 cm
---+----------+--------+-----------+-
  0.1        1.0      5.0        50.0
```

coefficient of consolidation(x 10^{-3} cm^2 s^{-1}), log scale

b)

```
---+---------+---------+---------+---
                   (--------*--------) S1
         (---------*--------)          S2
(---------*----------)                 S3
---+---------+---------+---------+---
  0.3       1.0       3.0       10.0
```

coefficient of consolidation(x 10^{-3} cm^2 s^{-1}), log scale