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A method to estimate hydraulic conductivity while drilling

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Abstract

A field test and analysis method has been developed to estimate the vertical distribution of hydraulic conductivity in shallow unconsolidated aquifers. The field method uses fluid injection ports and pressure transducers in a hollow auger that measure the hydraulic head outside the auger at several distances from the injection point. A constant injection rate is maintained for a duration time sufficient for the system to become steady state. Exploiting the analogy between electrical resistivity in geophysics and hydraulic flow two methods are used to estimate conductivity with depth: a half-space model based on spherical flow from a point injection at each measurement site, and a one-dimensional inversion of an entire dataset.

The injection methodology, conducted in three separate drilling operations, was investigated for repeatability, reproducibility, linearity, and for different injection sources. Repeatability tests, conducted at 10 levels, demonstrated standard deviations of generally less than 10%. Reproducibility tests conducted in three, closely spaced drilling operations generally showed a standard deviation of less than 20%, which is probably due to lateral variations in hydraulic conductivity. Linearity tests, made to determine dependency on flow rates, showed no indication of a flow rate bias. In order to obtain estimates of the hydraulic conductivity by an independent means, a series of measurements were made by injecting water through screens installed at two separate depths in a monitoring pipe near the measurement site. These estimates differed from the corresponding estimates obtained by injection in the hollow auger by a factor of less than 3.5, which can be attributed to variations in geology and the inaccurate estimates of the distance between the measurement and the injection sites at depth.

Keywords: Hydraulic conductivity; Injection test; Slug test; Pumping test; Instruments; Geophysical logging

1. Introduction

Traditionally the hydraulic conductivity of the subsurface, K, is determined using pumping, injection or slug tests, depending on the purpose. Other components of aquifer characterization include grain size analysis and permeameter tests of disturbed samples. New developments of dipole-flow tests (Zlotnik and Zurbuchen, 1998) are promising. The relative merits and failings of the various techniques can be debated (Hess et al., 1992), but there seems to be no decisive conclusion as to the superiority of one method over another. The choice of a method is usually determined by the hydrological problem at hand and the predisposition of the investigator. However, these methods require the emplacement of a drill hole, which is expensive and can act as a conduit for contaminant transfer at a polluted site, hence alternative techniques

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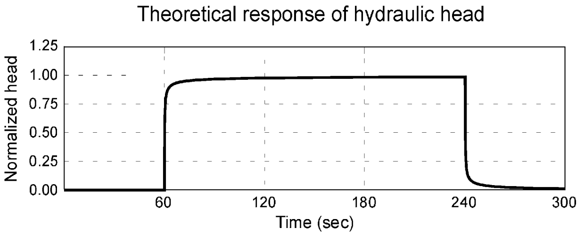

Fig. 1. Theoretical response for normalized hydraulic head with $K = 10^{-3}$ m s^{-1}, $S_s = 10^{-4}$ m^{-1}, $r = 1.0$ m and $Q = 3 \times 10^{-3}$ m^3 s^{-1} for $t = 60$–240 s.

are attractive. The method introduced here does not require a permanently installed drill hole. Measurements are made in a hollow auger, and when the auger is pulled, the hole is backfilled with bentonite through the drill stem. Hence the procedure is rapid, efficient, cost-effective, and safer than traditional techniques.

Pumping tests, performed by extracting substantial volumes of water from the aquifer over a long period of time, provide conductivity estimates averaged over a large aquifer volume. Based on the same concept, water is introduced into an aquifer in an injection test. Therefore pumping or injection tests are inadequate for detailed vertical mapping of the hydraulic conductivity. These tests are expensive because the system needs to be in a steady state over a large area that requires multiple wells, areas of low permeability can be problematic (Lebbe and Van Meir, 2000), and the data are arduous to interpret as can be seen by the extensive literature on the subject (de Marsily, 1986; Fetter, 1994). Other problems such as surging or backwash induced in the gravel pack around the intake port prior to testing may be reflected as increased conductivity; and measurements can be inaccurate if a screen is corroded or clogged as in an older monitoring well (Freeze and Cherry, 1979).

Slug tests provide in situ K estimates representative of a small volume of porous media in the immediate vicinity of the tip of a piezometer, or pressure transducer. A known quantity of water, or a displacement mass, is introduced or withdrawn into a borehole and the transient, or time for the hydraulic head to reach equilibrium, is measured usually at one level in the same hole. Multi-level techniques have been developed (Zlotnik and McGuire, 1998), and the single well test has been extended to multiple wells (Mas-Pla et al., 1997; Belitz and Dripps, 1999; Butler et al., 1999). Slug tests are relative rapid, but estimates of the hydraulic conductivity can be significantly influenced by many factors including: the packers, damaged zones around the well (Brown and Narasimhan, 1995), the shape of the cavity where the test is made (de Marsily, 1986), and the geometry of the well and screens with respect to the aquifer boundaries (Fetter, 1994; Yang and Gates, 1997; Binkhorst and Robbins, 1998; Butler et al., 1999). Thereby suggesting that unless great care is taken estimates of K obtained from slug tests can be highly inaccurate. Hinsby et al. (1992) and Butler et al. (2000) have developed methods employing direct-push technology for detailed vertical profiling, but the direct push technology has a limited depth of penetration of roughly 30 m.

We present a method for detailed in situ measurement of K in unconsolidated soil to a depth of 80 m using a hollow stem auger. Once the measurement site is reached, drilling is halted for approximately 15 min for the injection and hydraulic head measurements. The novelty of this method lies in the fact that K is determined while the drill string is in the ground and the change in hydraulic head is monitored in the same drill stem. A constant flow of water is injected into the formation through ports next to the cutting head. Four pressure transducers, with different vertical offsets to the injection ports, monitor the variation in the hydraulic head in the formation. A stationary level of the hydraulic head is reached after a few minutes of maintained injection. Assuming a Darcian flow, the hydraulic conductivity is determined from the net increase in the hydraulic head. Tests are performed at vertical intervals of approximately 1 m. This relatively dense vertical sampling and the application of four transducer offsets provide a detailed resolution of the vertical variability in hydraulic conductivity.

2. Theory for measurements

Transient flow in a saturated porous media is governed by the diffusion equation (Freeze and Cherry, 1979). For the case of a homogeneous, isotropic, infinite space the solution for the hydraulic head,

Table 1
Analogy between hydraulic and electrical parameters

Parameter	Hydraulic	Electrical
Potential	Hydraulic head, h	Electrical potential, Φ
Distance	Between the injection and observation points, r	Between the current and potential electrodes, r
Source	Injected water, Q	Injected current, I
Formation property	Hydraulic conductivity, K	Electrical conductivity, σ

h, due to a point injection of uniform flow, Q, is given by:

$$h(r,t) = \frac{Q}{4\pi rK}\,\mathrm{erfc}\left(\sqrt{\frac{S_s r^2}{4Kt}}\right) + h_0, \tag{1}$$

where S_s is the specific storage, r is the distance between the injection point and the point of observation, h_0, the initial head, t, the time, and erfc, the complementary error function. Eq. (1) can be used to determine the specific storage and hydraulic conductivity of the media for transient flow.

Fig. 1 shows the theoretical response of the hydraulic head for an infinite space with parameters: $K = 10^{-3}$ m s^{-1}, $S_s = 10^{-4}$ m^{-1}, $r = 1.0$ m and $Q = 3 \times 10^{-3}$ m^3 s^{-1}. The injection starts at $t = 60$ s and stops at $t = 240$ s. At $t = 240$ s, the argument of the complementary error function is 0.01 and the function equals 0.99 (Abramowitz and Stegan, 1972), so the hydraulic head is essentially in a steady state, and Eq. (1) becomes

$$h(r,\infty) = \frac{Q}{4\pi rK} + h_0. \tag{2}$$

Eq. (2) is the solution to Poisson's equation for potential fields, where the potential is the hydraulic head, $h(r)$, which can also be developed from Darcy's law subject to appropriate boundary conditions. This is analogous to the electrical resistivity (inverse of conductivity) case developed from Ohm's law with parameters assigned as in Table 1. The analogy of Darcy's law and Ohm's law are the basis for many electrical analog methods for groundwater systems (Todd, 1980; DeWiest, 1965; Viessman et al., 1989). Because the solution to Poisson's equation is the same for both cases, given appropriate assignment of parameters, concepts and tools developed in electrical methods geophysics are directly applicable to solve the hydraulic problem. Specifically, we use the method of images (Telford et al., 1995) to account for the presence of the water table, and a forward and inverse numerical modeling program to estimate the one-dimensional conductivity structure (Christensen and Auken, 1992). The modeling code can be used for interpretation and to investigate the behavior of the system in a stratified earth.

To estimate the variations of K with depth we begin by rewriting Eq. (2) so that

$$K(r) = \frac{Q}{4\pi r[h(r) - h_0]}. \tag{3}$$

Eq. (2) is valid for a homogeneous, infinite space; therefore K is the infinite-space estimate. Because of the assumption of an infinite space, the accuracy of the estimates of K is dependent not only on the value of K at the transducer location, but on the average value within the entire regime influenced by the injection.

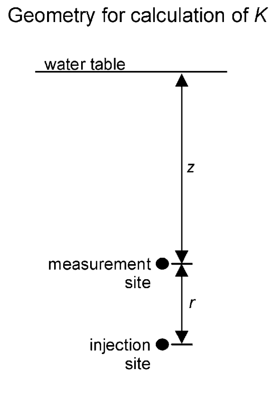

Fig. 2. Geometry used to calculate K as a function of depth. The distance z is the depth to the measurement point, and r is the distance between the injection and measurement sites.

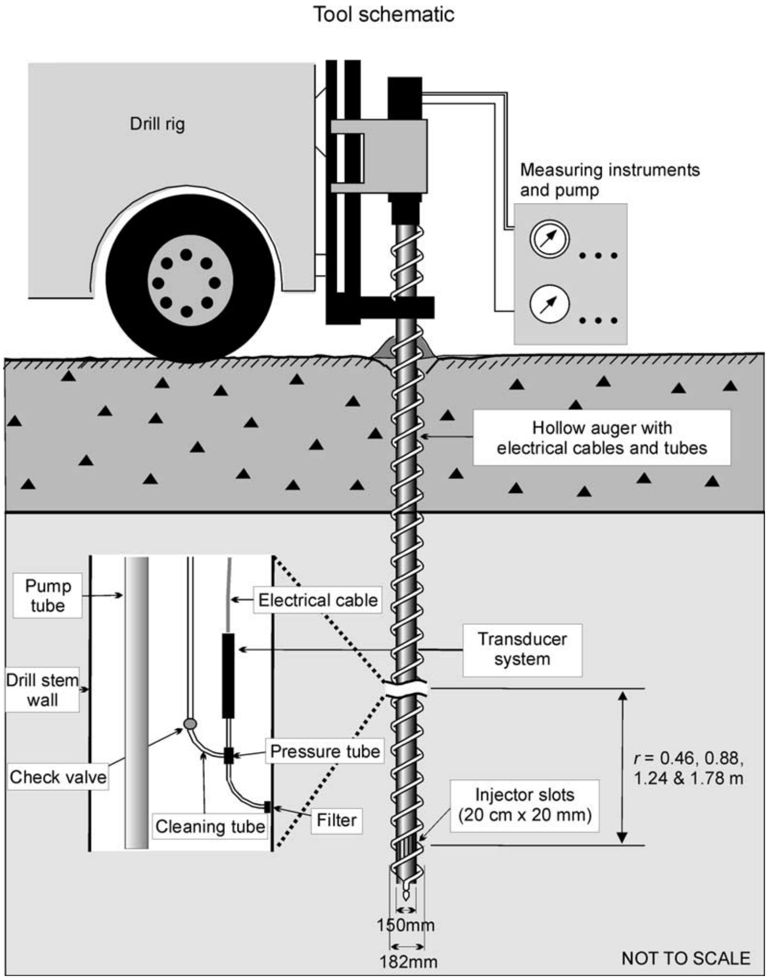

Fig. 3. Schematic of the tool used to measure the hydraulic head while drilling in a hollow auger.

Let us also assume that K is a function of depth, z_i, so that

$$K(z_i, r) = \frac{Q}{4\pi r[h(r) - h_0]} \tag{4}$$

is determined at various levels, $i = 1, 2, \ldots, N$ that approximate the vertical variation in K (Dam et al., 1996).

Measurements are not made in an infinite space, but can be influenced by the location of the water table. A distortion of K_∞ is introduced if the injection point is close to the water table. Exploiting the electrical analogy the distortion can be estimated by using the geometric factor, G, for the pole–pole configuration in a homogeneous, conducting half-space (Telford et al., 1995). The transducer and the point of injection are located at depths below the water table of z and $z + r$, respectively, as shown in Fig. 2. The factor G is determined with an image pump source located at $z + r$ above the water table. Let h_∞ be the infinite-space hydraulic head, and h_{wt} the hydraulic head observed in the presence of the water table so that

$$h_{\mathrm{wt}} = Gh_\infty = \left(1 - \frac{r}{2z + r}\right)h_\infty. \tag{5}$$

In Eq. (4) $h(r) - h_0 = h_\infty$. Replacing h_∞ with h_{wt} in Eq. (4) gives the corresponding water table hydraulic

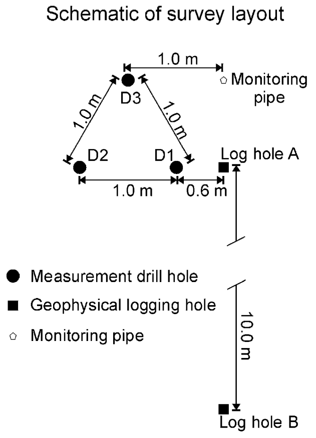

Fig. 4. Survey layout showing the location of the measurements sites D1, D2 and D3 logged drill holes, A and B, and the monitoring pipe used to injected water as a test of an independent source.

conductivity,

$$K_{wt}(z,r) = \frac{Q}{4\pi r[h(r) - h_0]}\left(1 - \frac{r}{2z + r}\right). \quad (6)$$

It is assumed that the water table is stationary during the injection. Eq. (5) shows that the hydraulic head goes to zero as the measurement point approaches the water table. Without the use of G, K_∞ will overestimate K_{wt}.

3. Methodology

The measurement tools are integrated in a hollow auger as depicted in Fig. 3. Through ports next to the cutting head, a constant flow of water of 0.20–0.40 $\mathrm{l\,s^{-1}}$ is injected into the formation. The ports are slots with a height of 0.20 m that are equally spaced around the drill stem at intervals of 0.02 m. The drill stem is 150 mm in diameter. An array of transducers, with vertical offsets of 0.46, 0.88, 1.24, and 1.78 m to the center of the injection slots, is used to measure the rise in the hydraulic head while water is injected into the formation. The transducer system is mounted inside the drill stem and cables run through the auger to a collector with slip rings at the top of the drill head. The hydraulic head in the formation is conveyed to the transducer system through a pressure tube entering a small opening in the drill stem. A filter is placed in front of the opening to prevent obstruction of the tube.

Before measurements are made, the filter is cleaned by pumping water through a cleaning tube connected to the pressure tube. A check valve is mounted to prevent pressure variations in the cleaning tube from distorting the measurements. After cleaning, the cleaning tube is emptied, and the valve is closed. Inside the drill stem, a pump tube runs from ground level to a chamber located just above the injector ports allowing a controlled flow to be injected into the formation at a constant rate. An inflatable packer, acting as a valve, is located in the chamber and controlled from the surface by means of an air tube.

The disturbance of the soil is, minimized by the small flight of the auger (32 mm). When the drill head is within approximately 1 m of the measurement level, the drilling speed is adjusted so that no material is removed with the auger and the injection ports are well seated in the soil. Drilling is halted for approximately 15 min for the injection and hydraulic head measurements. In order to achieve a constant flow rate throughout the injection, the pump tube is filled with water before the injection is started. The inflatable packer restrains the water pressure from the pump, which is started prior to injection start. The injection is initiated by releasing an electrical valve placed next to the inflatable packer, after which a constant flow rate is obtained immediately. The measurement begins 1 min before the injection starts and is maintained for 2 min after the injection is completed. A stationary level of the hydraulic head is reached at the transducer locations within a few minutes after the injection begins.

Measurements of the hydraulic head and the flow rate are acquired at a time interval of 0.063 s with a PC, which is connected to the transducer systems through electrical cables confined in the drill stem. In order to obtain a satisfactory signal-to-noise ratio, the hydraulic head data are digitized at the transducer and pre-processed before being sent to the PC. The drilling is halted for about 15 min while measurements are acquired. The methodology allows determination of K at any level within the saturated zone. Measurements can be made at vertical intervals of 1 m, allowing for a relatively detailed profile of the hydraulic head.

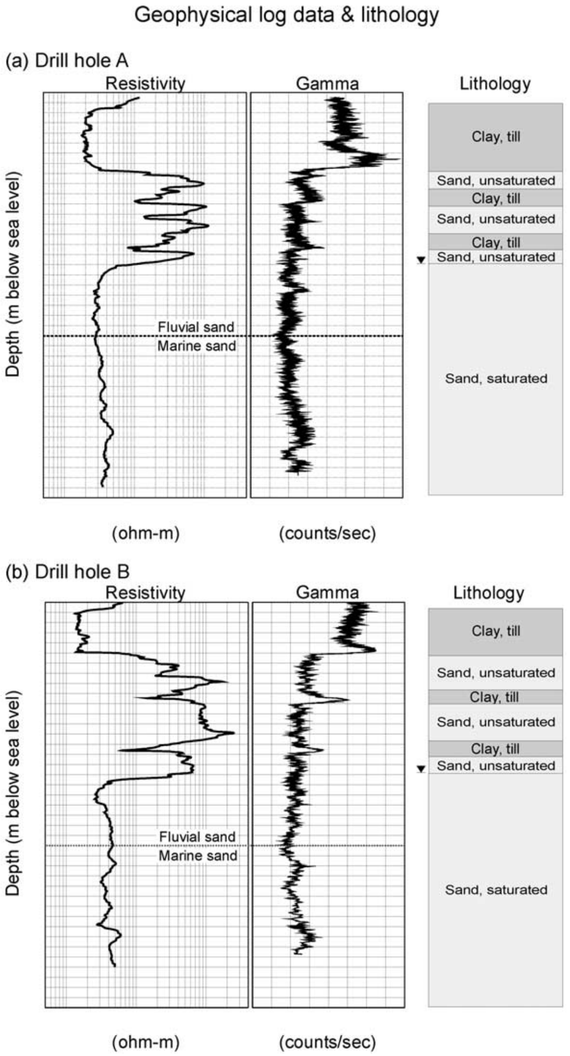

Fig. 5. Resistivity and gamma logs acquired with the Ellog method in (a) borehole A located about 1 m from sites D1, D2, and D3 and (b) borehole B located about 10 m away from A, and corresponding lithology.

4. Field tests

Field tests were performed in three closely spaced drill sites in the Beder aquifer south of the City of Århus, Denmark. The field layout is shown schematically in Fig. 4. Measurements while drilling were carried out at sites D1, D2 and D3, denoted by circles, at 1 m intervals in a depth range of 10 m. Measurements were repeated three times at each level. Independent tests were made in D3 by injecting water through screens in the monitor pipe at a distance of 1 m. Geophysical logs were made in the drill holes A and B, marked as squares.

4.1. Geology

The geology of the field site was determined from an extensive number of drilling reports and surface geophysical surveys. In addition, two resistivity logs and gamma logs, shown in Fig. 5, were acquired at the field site using the Ellog method (Sørensen, 1989). With the Ellog method resistivity and gamma measurements are made in an auger drilling system similar to the one described in this paper. The lithology was interpreted from the geophysical logs based on the knowledge of the regional geology.

The Beder aquifer consists of a sandy Quaternary deposit filling a valley eroded into Tertiary marine clays. The erosion probably occurred by advancing ice during the Saalian and Weichselian glaciations (Harder, 1908; Houmark-Nielsen, 1987) that were subsequently filled with Quaternary deposits. The Quaternary deposits are grouped into three major sedimentary units: (1) the Marine Unit, (2) the Peri-Glacial Unit, and (3) the Glacial Unit.

The Marine Unit, which overlies Pre-Quaternary deposits, was deposited in a shallow marine environment, such as a fjord or bay. The sediments are dominated by medium to coarse grain sand, with gravel as a minor fraction. The formation is very homogeneous, and no clay layers have been identified in the formation. The roughly 5–10 m thick Peri-Glacial Unit overlies the Marine Unit. The sediments were deposited in fluvial, eolian, and lacustrine sub-environments. Fluvial sand is the dominant lithology at the field site. The sand is medium to coarse grain, and well sorted, however, there are several zones with layers of silt and clay. These fine-grained layers are 2–5 cm thick and have a lateral extent of generally less than a few metres. On a local scale, the vertical flow may be controlled by the presence of these layers.

The uppermost Glacial Unit overlies the Peri-Glacial Unit. It is about 15 m thick in the study area, and consists of several layers of clay till 2–10 m thick, alternating with layers of melt-water sand 2–3 m thick. The hydraulic conductivity of the

Table 2
Repeatability of K estimates obtained from three measurements in D2 for equivalent depths from transducer offsets of 0.455, 0.88, 1.235, and 1.78 m

Transducer 1 $r = 0.455$ m			Transducer 2 $r = 0.880$ m			Transducer 3 $r = 1.235$ m			Transducer 4 $r = 1.780$ m		
Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)
21.20	2.5606	1.1	20.78	2.0514	0.2	20.38	2.1723	0.9	20.97	1.3006	3.1
22.22	2.5948	2.5	21.80	2.0772	0.4	21.39	1.7326	0.5	21.97	3.1723	5.6
23.22	6.0497	3.3	22.80	4.4235	2.5	22.39	4.6243	45.5	22.97	4.2924	1.0
24.22	6.6357	1.1	23.80	5.0691	1.0	23.39	5.2939	0.5	23.97	3.4173	20.7
25.22	10.730	0.8	24.80	12.136	1.2	24.39	4.8453	1.6	24.97	7.1481	2.3
26.22	9.8897	2.0	25.80	9.1742	3.8	25.39	11.140	3.0	25.97	5.0402	2.3
27.22	5.7614	0.1	26.80	9.7921	1.4	26.39	7.1055	1.3	26.97	9.8548	8.9
28.22	10.258	2.3	27.80	8.7085	2.4	27.39	10.886	3.5	27.97	11.217	7.0
29.22	12.500	9.7	28.80	11.939	10.3	28.39	12.179	11.6	28.97	7.2254	2.9
30.22	11.977	1.3	29.80	9.8424	3.5	29.39	7.9596	0.9	29.96	6.4318	23.0

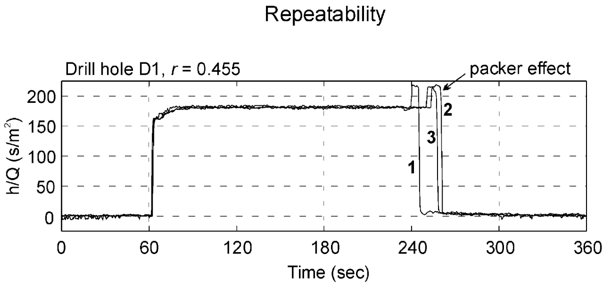

Fig. 6. Three repeated measurements (1, 2, and 3) for transducer offset $r = 0.455$ m made at the same level of the formation at site D1. The hydraulic head is normalized to the applied flow rate. Note that the lengths of the time series are not identical. The spikes seen at the end of the injection are caused by the inertia of the water, which increases the flow rate and hence the hydraulic head when the packer is inflated to stop the flow.

clays is not established, but is assumed to be very low. The uppermost layer, pertaining to the Glacial Unit, consists of 5–8 m of clay till. In the depth range of 16 m below sea level (mbs) to 39 mbs, the end of the log, the sediments consist of sand. The water table, delineated by a drop in resistivity, is situated at 17 mbs. The limits of the Peri-Glacial Unit cannot be recognized from the logs.

The boundary between fluvial sand (Peri-Glacial Unit) and marine sand (Marine Unit) at 24 ± 2 mbs was established by a mineralogical analyses, including grain size distribution by Larsen (1996). As we shall see, inversion of the data estimates the boundary at 26 mbs. The grain size distribution shows that the fluvial sand are fine to medium grain, whereas the marine sand are medium to coarse grain with a minor fraction of fine gravel. These findings suggest that the hydraulic conductivity of the marine sand may be significantly higher than that of the fluvial sand.

Drill holes A and B are spaced 10 m apart and differences are observed in the character of geophysical logs of Fig. 5, particularly in the unsaturated zone. It is not unusual to see variations in layer thickness in fluvial and eolian deposits over this scale, as can be seen in the alternating sand and clay layers. The presence of thin layers of 2–5 cm of clay and silt with small lateral extent in the Peri-Glacial unit can cause variations between the boreholes.

4.2. Repeatability

Repeatability tests conducted at 10 vertical levels of the formation demonstrated a standard deviation in general of less than 10%. Table 2 shows the mean value of estimates of K, and the corresponding standard deviation, based on three repeated measurements conducted at each level in drill site D2. Tests were made in such a way as to minimize the disturbance to the soil. After a measurement was made, the auger was rotated in both direction 2 or 3 times to reseat the drill head. Of course a full test would entail re-drilling, but that would change either the conditions at the site or the testing location, and hence the K estimates.

A typical example of three repeated measurements carried out at the same level in drill site D1 is shown in Fig. 6. The measurements were made at a constant flow rate of 0.3 l s^{-1}. Comparison of Fig. 5 with Fig. 1 shows the field response is similar to the theoretical response of Fig. 1. The stationary level is well-defined and reached within less than 240 s for which the injection is maintained. The spikes seen at the end of the injection are caused by the inertia of the water, which increases the flow rate and hence the hydraulic head when the packer is inflated to stop the flow.

4.3. Reproducibility

Reproducibility tests were conducted in the three closely spaced drilling sites—D1, D2 and D3. The mean K and standard deviation of the estimates determined at the four transducer offsets at the three sites are presented in Table 3. The standard deviation is computed from the mean values of the repeated measurements at each site. It is seen that for any of the transducer offsets the standard deviation is less than 38%, and for most depths less than 10%. The

Table 3
Reproducibility of the mean value of K obtained in the three test sites—D1, D2 and D3, for equivalent depths from transducer offsets of 0.455, 0.880, 1.235, and 1.780 m

Transducer 1 $r = 0.455$ m			Transducer 2 $r = 0.880$ m			Transducer 3 $r = 1.235$ m			Transducer 4 $r = 1.780$ m		
Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)	Depth (m)	K (m s^{-1}) $\times 10^{-4}$	SD (%)
21.22	2.2788	18.6	20.80	1.9765	1.6	20.39	2.0367	8.1	20.97	2.1757	24.6
22.22	2.4905	9.9	21.80	3.4110	21.6	21.39	2.8603	21.3	21.97	6.1204	20.0
23.22	2.6250	35.6	22.80	2.6389	37.7	22.39	6.0565	22.4	22.97	3.4324	10.3
24.22	5.7330	2.5	23.80	4.2494	3.2	23.39	4.5074	7.8	23.97	3.6872	3.0
25.22	6.3660	2.4	24.80	5.5370	5.7	24.39	4.7485	7.0	24.97	6.9798	1.0
26.22	10.495	1.3	25.80	10.095	9.7	25.39	10.564	2.4	25.97	5.0667	3.0
27.22	9.5355	1.6	26.80	9.0203	6.1	26.39	7.4020	2.0	26.97	8.7321	9.5
28.22	5.5676	4.1	27.80	9.3005	12.4	27.39	10.569	1.6	27.97	10.902	1.4
29.22	10.340	5.9	28.80	8.4374	4.6	28.39	11.395	3.2	28.97	6.8966	8.1
30.22	12.184	2.3	29.80	12.039	1.4	29.39	8.8572	4.2			

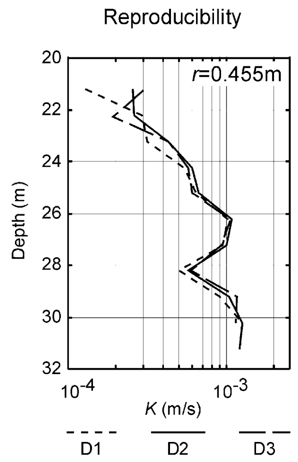

Fig. 7. Measured values of K for transducer $r = 0.455$ m obtained at sites D3 (short dash) D2 (solid), and D3 (long dash) demonstrating reproducibility in closely spaced sites.

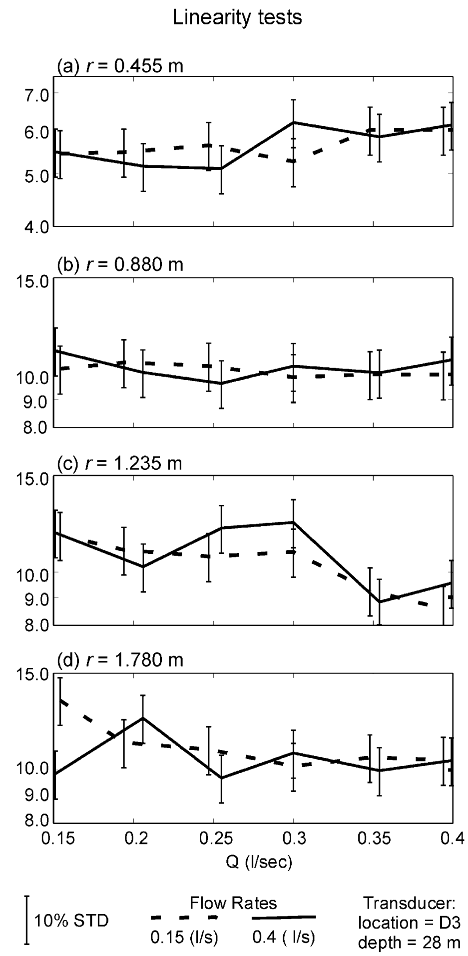

Fig. 8. Measured values of K for two flow rates for transducer offsets (a) $r = 0.455$ m, (b) $r = 0.880$ m, (c) $r = 1.235$ m and (d) $r = 1.780$ m at D3 showing linearity of flow within the measured errors. The injection site was located at 28 mbs.

highest standard deviation are found at the top of the profile in the fluvial where lateral variations in hydraulic conductivity are more likely, as discussed earlier.

Fig. 7 shows the mean value of K obtained at transducer offset $r = 0.455$ m, based on the three repeated measurements conducted at each level. The estimates exhibit a high reproducibility, particularly at depths below 24 mbs. The K increases by nearly an order of magnitude from 2.0×10^{-4} to 1.0×10^{-3} m s^{-1} from the top to the bottom of the profile, with a thin layer of less than 2.0×10^{-4} m s^{-1} at 28 mbs. The increase in K is consistent with the change in grain size identified as the boundary between the fluvial and marine sand.

4.4. Linearity

Linearity tests were made to examine the dependence of the method on injection flow rates. Fig. 8 show estimates of the hydraulic conductivity obtained from two data series measured at different flow rates in successive measurements. Measurements were made at two flow rates, 0.15 and 0.40 l s^{-1}, to examine flow rate bias. The measurements were performed at one specific level corresponding to transducer location 28.0 mbs in site D3. On the basis of the repeatability tests, a 10% standard deviation of the estimates is assumed, as shown by the error bars. Data are plotted on a linear-log scale so that the error bars are the same size in all plots. Note that the estimates for K in Fig. 8(a) range from 4 to 7×10^{-4} m s^{-1}, while those for Fig. 8(b)–(d) are from 8 to 15×10^{-4} m s^{-1}. Taking the 10% error into account estimates of K are essentially independent of the applied flow rate in the range from 0.15 to 0.40 l s^{-1}.

4.5. Independent estimates of K

Injections through screens in a nearby monitoring pipe were made to obtain K estimates by an independent means. By injecting outside the auger site it is possible to isolate circumstances related solely to

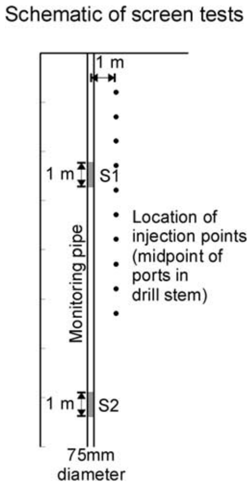

Fig. 9. Schematic of the layout with depth for tests that uses an independent source for water injection. The horizontal separation, x, between the measurement sites and the monitoring pipe is not accurately known at depth.

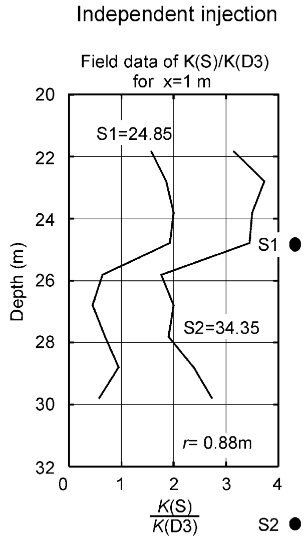

Fig. 10. Hydraulic conductivity for injection through screens, K(S), in the monitoring pipe normalized by values obtained via injection at the site D3, K(D3), for field data computed with $x = 1$ m.

injection through the drill stem, especially the possibility of water flowing along the drill stem when injecting through the drill stem. Estimates of K were obtained from measurements made by the transducer arrays mounted on the drill stem at each level in site D3 for water injected through three different sources: the drill stem, screen 1 (S1), and screen 2 (S2), as shown schematically in Fig. 9. The slotted screens in the monitoring pipe, about 1 m from D3 and 1 m in depth extent, are centered at $z = 24.85$ m and $z = 34.35$ m. The pipe is about 75 mm in diameter.

Estimates of K from the screen injections are normalized by the corresponding estimates from drill stem injection data to emphasize departures from the drill stem injection results. There are three reasons that can account for differences between the normalized values for S1 and S2, and their departure from the value of one. First, the aquifer material is not homogeneous. Second, the horizontal distance, x, between D3 and the monitoring pipe is not accurately known at depths of over 20 m. The distance is necessary to compute the hydraulic conductivity geometric factor. And lastly, the water may be flowing up the drill stem from the injection site in D3. Analysis of synthetic response and the field data are used to estimate the relative contribution.

Fig. 10 shows the normalized values of hydraulic conductivity for screen 1 and screen 2 at transducer offset of $r = 0.88$ m. The departures between the two curves are as great as a factor of 3.5 near the water table and 0.5–2.5 deeper in the section. To account for this separation in the curves, several numerical experiments were conducted.

In calculating the geometric factor for K the midpoint of the screens and the horizontal separation, estimated at $x = 1$ m, were used. K(S1)/K(D3) for field data plotted for $x = 0.5$, 1.0 and 2.0 m in Fig. 11(a) show variations as great as a factor of 4 can be present. To further investigate the effect of horizontal separation, the sensitivity of the geometric factor to various values of x in an infinite-space is shown in Fig. 11(b). For overestimates of the true value of x the distortion is seen at distances greater than $+/-4$ m from the injection point. For underestimates of the true value the distortion is felt $+/-1$ m from the injection point. Miscalculation of the geometric factor can cause variations as much as a factor of 4.

To test the effect of layering the one-dimensional electrical modeling program of Christensen and Auken (1992) is used as an analog for the hydrological problem. The program computes the forward or inverse responses for point injection and point measurements anywhere in a layered aquifer.

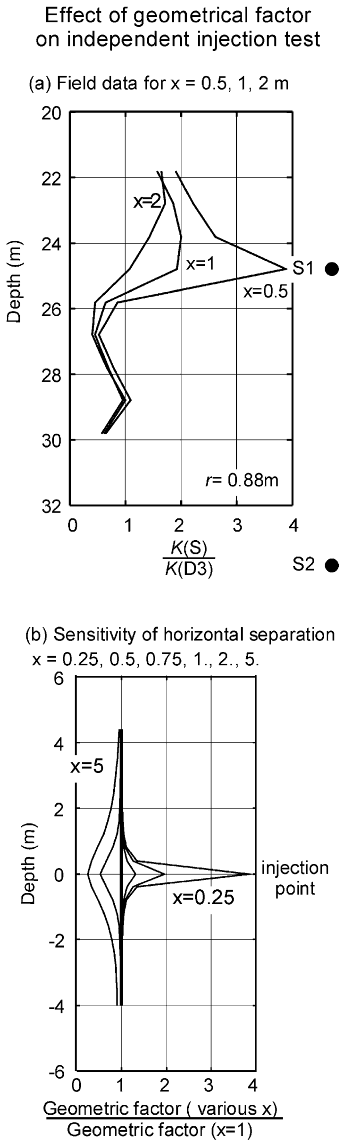

Fig. 11. (a) Hydraulic conductivity for injection through screen 1, *K*(S1), in the monitoring pipe normalized by values obtained via injection at the site D3, *K*(D3), for field data computed with $x = 0.5$, 1, 2 m. (b) Theoretical sensitivity function for the geometric factor for the screen injection geometry for various values of *x*.

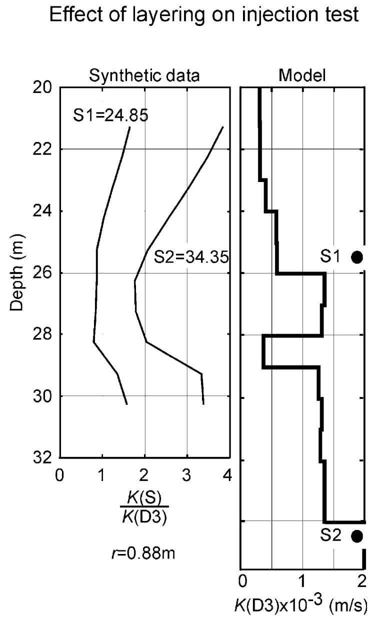

Fig. 12. Theoretical values of *K*(S1)/*K*(D3) and *K*(S2)/*K*(D3) for a layered earth model shown to the right of the response curves.

Responses, shown in Fig. 12, were simulated for injection and transducers at a separation of $r = 0.88$ with a 1 m spacing downhole for D3, and injections at 24.85 m and at 34.35 m with transducers every 1 m in D3 for S1 and S2. The model, shown to the right of the responses is based on an inversion discussed in Section 5. Layering can lead to a separation between the two curves and a departure from the value of one. Fig. 12, showing the *K*(S1)/*K*(D3) and *K*(S2)/*K*(D3) synthetic response for multiple layer model, demonstrates that layering splits the curves by a factor similar to that seen in the field data. Layering outside of the measured section can have a significant influence on the response if a layer is in close proximity to an injection site.

The analysis shows that the departure from one of the normalized screen curves in Fig. 10(a) can be accounted for by the inability to accurately estimate the horizontal separation between D3 and the monitoring pipe, which is used in the computation of the geometric factor, and layering in the subsurface. Hence, we can conclude that the presence of the drill stem does not seem to influence the *K* estimates when the injection is performed through the drill stem.

5. Interpretation

Detailed information can be gleamed by inspection of the *K* curves. However, sensitivity to the aquifer material varies for different source–transducer

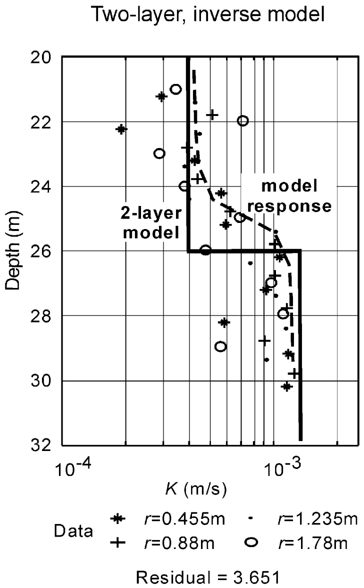

Fig. 13. Two-layer inverse model, response, and data for $r = 0.455$ in D3. Data were inverted to estimate the marine-fluvial boundary.

separations; processing the data with an inversion algorithm removes the system response. Inversion of data to recover earth properties is a well-known method in geophysical data processing. Menke (1989) describes a damped, least-square, non-linear, iterative approach that is implemented in the one-dimensional electrical resistivity algorithm of Christensen and Auken (1992). The downhole data, transformed to K, are used as the input. A 10% standard deviation is assumed. Various parameters, such as the damping term, are set to stabilize the inversion process.

The first inversion was computed for a two-layer aquifer to determine the boundary between the fluvial and the marine layers, which is responsible for the primary change of K in the section. Grain size analysis estimated a contact of 24 ± 2 mbs. Data from D3 for all the injection–transducer separations and for only $r = 0.455$ were inverted. The resulting model for all the D3 data is shown in Fig. 13: the thickness of the top layer was found to be 9.04 m below the water table with a total residual of 3.651. The residual is a measure of the goodness of fit between the model response and the field data. When only the $r = 0.455$ m data were used, the thickness was estimated at 9.12 m with a residual of 3.726. Estimates of the hydraulic conductivity for the top layer were the same for both inversions and dropped from 1.2×10^{-3} to 7.5×10^{-4} m s^{-1} for the $r = 0.455$ m data.

Data for all four transducer separations from all the drilling sites, processed with the water table correction term, are shown in Fig. 14(a). It is clear by inspection that there are more than two layers in the section. To show how inversion can be used for interpretation, data from each of the three drilling sites were processed with a multiple layer inversion scheme. In contrast to inverting for conductivity and thickness as in the two-layer model above, the multiple-layer scheme uses layers with a fixed thickness of 1 m, and the inversion is asked to estimate the values of conductivity in each layer. The model results, with residual, for D1, D2 and D3 are shown in Fig. 14(b).

The contact between the fluvial and the marine layer is present at all site at a depth of 26 m. At a depth of 28–29 m there is also a low conductivity layer present at all three sites. This layer is probably somewhat thinner than 1 m, as can be seen in the hydraulic conductivity curves. It is only apparent in the small offset ($r = 0.455$), and therefore must be thin. The consistency of this layer between all three sites is what one would expect in a marine environment. In contrast, the upper part of the section, above the 26 m boundary is quite variable from site to site. In particular there appears to be a low conductivity zone at D2 that is not apparent at D1 and D3. This is consistent with a fluvial environment where clay rich lenses could have a small lateral extent.

6. Conclusions

A single-hole method for in situ estimates of hydraulic conductivity while drilling has been developed and shown to be a rapid, viable alternative to traditional methods. The method was tested in three closely spaced drilling operations at 10 levels in an unconfined aquifer. The tests included repeatability, reproducibility, linearity, and independent injection sources. The repeatability test showed that, in most cases, the standard deviation of three repeated measurements was less than 10%. The reproducibility test showed that the standard deviation of estimates obtained in three drilling sites was less than 20% at most levels. Linearity tests showed that estimates

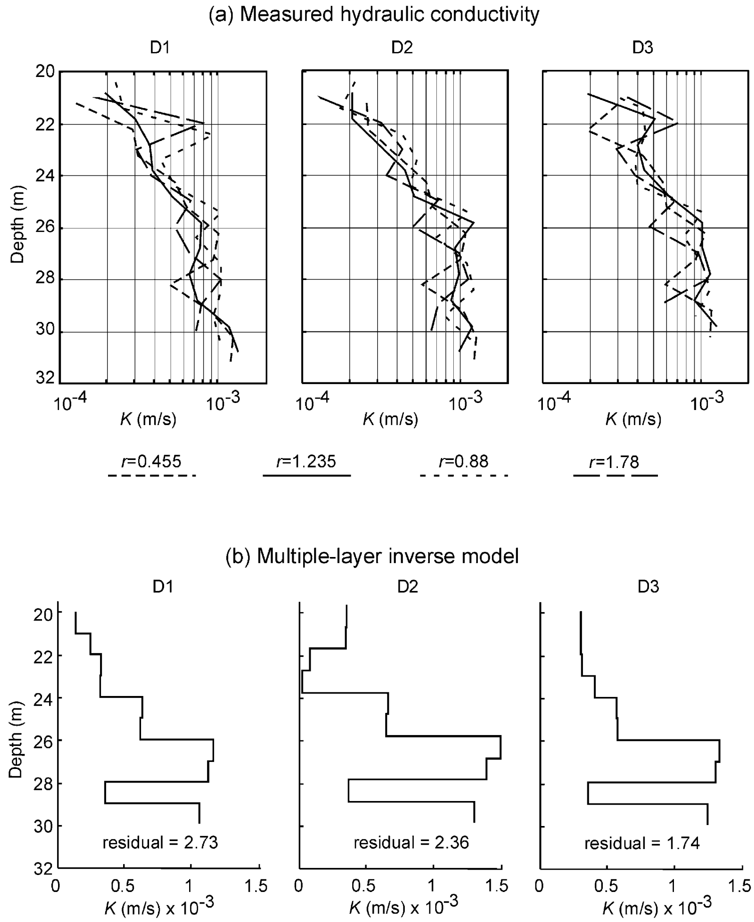

Fig. 14. (a) Hydraulic conductivity at each site and transducer separation computed with the water table correction term, and (b) the corresponding one-dimensional models from a multiple-layer inversion process.

were independent of the flow rate in the range from 0.15 to 0.40 $\mathrm{l\,s^{-1}}$ within the uncertainty determined with repeated measurements.

Variations of estimates of K made with injection from a different source were explained through a numerical modeling experiment that showed these variations can be due to layering and poor estimates of the horizontal distance between the injection and measuring site. Hence water traveling up the drill stem is not of significant concern. Data, converted to hydraulic conductivity and corrected for the water table interface, give detailed information with depth, as opposed to an averaged value obtained in more traditional tests. A layered-aquifer interpretation can be made using inversion software commonly used in geophysical applications. The inversion results are consistent from site to site and with the known geology.

The injection point is always in close proximity to the measurement site, hence a detailed vertical profile of the conductivity can be obtained in contrast to techniques that have a stationary point of injection or average the response over a large area. The measurement while drilling method offers a rapid, robust, inexpensive and relatively non-intrusive method to acquire detailed K estimates of the subsurface.

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