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Spatial variability of hydraulic conductivity of an unconfined sandy aquifer determined by a mini slug test

Poul L. Bjerg[a], Klaus Hinsby[b], Thomas H. Christensen[a] and Peter Gravesen[b]

[a] *Department of Environmental Engineering/Groundwater Research Centre, Bldg. 115, Technical University of Denmark, DK-2800 Lyngby, Denmark*

[b] *Geological Survey of Denmark, Thoravej 8, DK-2400 Copenhagen NV, Denmark*

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ABSTRACT

Bjerg, P.L., Hinsby, K., Christensen, T.H. and Gravesen, P., 1992. Spatial variability of hydraulic conductivity of an unconfined sandy aquifer determined by a mini slug test. J. Hydrol., 136: 107–122.

The spatial variability of the hydraulic conductivity in a sandy aquifer has been determined by a mini slug test method. The hydraulic conductivity (K) of the aquifer has a geometric mean of 5.05×10^{-4} m s^{-1}, and an overall variance of ln K equal to 0.37 which corresponds quite well to the results obtained by two large scale tracer experiments performed in the aquifer. A geological model of the aquifer based on 31 sediment cores, proposed three hydrogeological layers in the aquifer concurrent with the vertical variations observed with respect to hydraulic conductivity. The horizontal correlation length of the hydraulic conductivity has been determined for each of the three hydrogeological layers and is found to be small (1–2.5 m). The asymptotic longitudinal dispersivity of the aquifer has been estimated from the variance in hydraulic conductivity and the horizontal correlation length, to be in the range of 0.3–0.5 m compared with a value of 0.42 m obtained in one of the tracer tests performed.

INTRODUCTION

The basis for performing reliable groundwater modelling of pollutant spreading and effects of remedial actions is a sound physical representation of the aquifer in question. The physical parameter of most concern is usually the hydraulic conductivity of the aquifer, partly because it may vary several orders of magnitude according to the geology of the area, and partly because the methods available for determination of hydraulic conductivity are costly or questionable with respect to accuracy. This data acquisition problem is in

Correspondence to: Poul L. Bjerg, Department of Environmental Engineering/Groundwater Research Centre, Bldg. 115, Technical University of Denmark, DK-2800 Lyngby, Denmark.

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particular severe in contaminant modelling, where recent investigations (e.g. Sudicky, 1986; Freyberg et al., 1986) have shown that the spatial variability of the local hydraulic conductivity controls the advection and macrodispersion in solute transport.

In the literature very few investigations are reported on the spatial variability of hydraulic conductivity, where tracer experiments have also been performed allowing for evaluation of the accuracy of the obtained data. At the Borden Air Force site, Canada, Sudicky (1986) found for the logtransformed hydraulic conductivity data a variance ($\sigma^2_{\ln K}$) of 0.38, a horizontal correlation length (λ_h) of 2.8 m and a vertical correlation length (λ_v) of 0.12 m. Hess et al. (1991) found results in the same range at Cape Cod, USA. Substantially larger values were reported for the Columbus Air Force site, USA: $\sigma^2_{\ln K} = 2.8$, $\lambda_h = 5.3$ m, $\lambda_v = 0.7$ m (Boggs and Rehfeldt, 1991).

In the laboratory, the hydraulic conductivity can be determined on sediment samples, e.g. by permeameter tests (e.g. Sudicky, 1986) and by determining the sediment grain size curve and employing empirical equations relating grain size to hydraulic conductivity (e.g. Freeze and Cherry, 1979; Masch and Denny, 1966; Taylor et al., 1987).

In the field, the hydraulic conductivity can be determined, e.g. by flow meter methods, pumping tests and slug tests. These methods are usually very demanding if detailed information on the spatial variability is needed and, in actual cases with contaminant plumes, performing these tests may disturb the contaminant spreading and the interpretation of the contaminant plume. Recently, a mini slug test method has been developed (Hinsby et al., 1991) allowing for a detailed mapping of hydraulic conductivity with a minimum of disturbance on the local groundwater quality. Furthermore, the method allows for groundwater sampling at the same depth making it attractive in groundwater pollution investigations.

The purpose of this study is to present and discuss the results of a spatial mapping of hydraulic conductivity in an unconfined sandy (glaciofluvial) aquifer by help of a newly developed mini slug test method. The results are compared with two large scale natural gradient tracer experiments and the geology of the research site.

MATERIALS AND METHODS

Research site

The research site is located in the western part of Denmark, Fig. 1. The aquifer is unconfined and the water table is approximately 4–5 m below land surface. At the site, two large scale natural gradient tests were carried out in

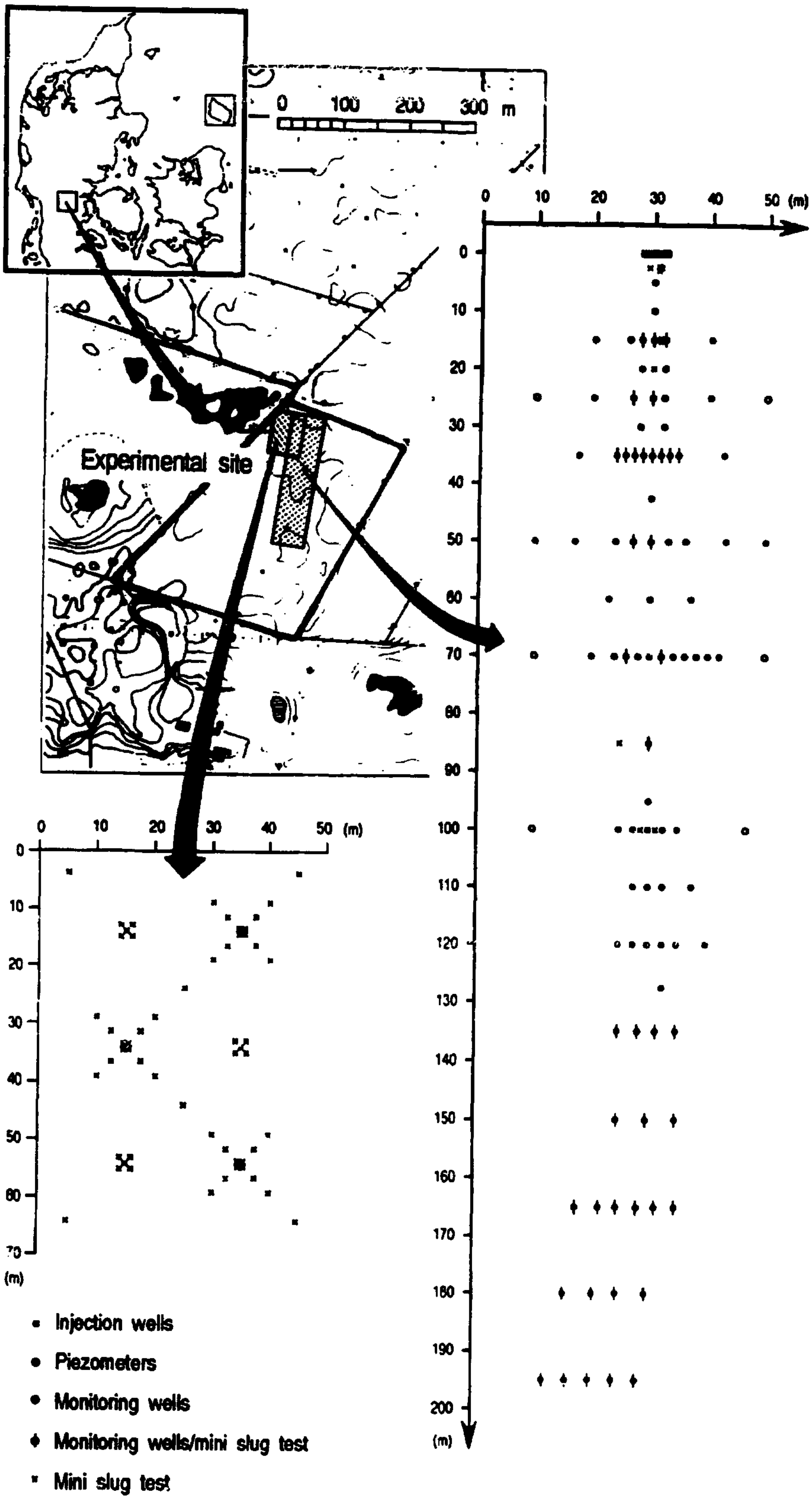

Fig. 1. Map showing the location of the research site and the areas where the slug tests were performed.

1989 and 1990. The first experiment used tritium as a tracer and the purpose was to investigate transport and dispersion processes (Jensen et al., 1992). The second experiment focused on ion exchange processes and used chloride as a tracer (Bjerg and Christensen, 1992).

The mini slug test

The slug tests were performed in 1 inch iron tubes supplied with a 0.25 m screen at the tip. The iron tube was driven into the ground by a motor driven hammer. After the well had been installed the slug test was performed. The water level was raised in the well by a vacuum pump. The vacuum is controlled by a vacuum-meter and the water level in the well by a small pressure transducer. After equilibration, the vacuum was released and the falling head is recorded as a function of time on a computer or a chart-recorder; (see Hinsby et al. (1992) for more details about the method). The locations of the mini slug tests are shown in Fig. 1. A total of 334 slug tests were performed at 110 locations in 1–10 depths.

Sediment sampling

Thirty-one sediment cores were taken out in aluminium tubes and divided into 25 cm or 50 cm segments in the field. The principle of the core sampler is described by Starr (1988). The sediment samples were subject to geological description.

Statistical methods

Basic statistics and statistical testing were performed with the help of the STATGRAPHICS software package (Statgraphics, 1989). The GEO-EAS software package was used in the geostatistical analysis (Englund, 1988). The principles of the geostatistical analysis were based on Journel and Huijbregts (1978). The variogram function $2\gamma(h)$ was calculated by an estimator $2\gamma^*(h)$ being the arithmetic mean of the squared differences between two experimental measurements $[z(x_i), z(x_i+h)]$ at any two points separated by the vector $\boldsymbol{h}$:

$$2\gamma^* = \frac{1}{N(h)} \sum_{i=1}^{N(h)} [z(x_i) - z(x_i+h)]^2$$

where $N(h)$ is the number of experimental pairs $[z(x_i), z(x_i+h)]$

Some important features of a variogram are briefly described below. Often a variogram reaches a constant value called a sill value. The range is the minimum distance between points where no correlation exists. Different theoretical models may be applied to an experimental variogram, e.g.

spherical, gaussian or exponential models (see Journel and Huijbregts, 1979). The exponential variogram function is defined as:

$$\gamma(h) = C[1 - \exp(-h/a)]$$

where C is a constant corresponding to the sill value, h is the distance between measurement points, and a equals the integral scale or correlation length.

In case of a discontinuity at the (0, 0) in the calculated variogram a nugget effect is recognized. This may be the result of measurement errors or of a variability on a scale smaller than the smallest spacing of the measurement points.

GEOLOGY OF THE RESEARCH SITE

The geology at the site has been described based on information from thirty-one sampled sediment cores. Fourteen geological sections were constructed using textural, structural and genetic information to correlate the beds between sediment cores (Fig. 2). From these sections a geological model has been proposed as described in the following paragraphs. The geology at the site is typical for the glacial outwash plains in the western part of Denmark.

The bottom of the aquifer consists of silt and clay deposited in a meltwater lake in the Lower Weichselian of the Quaternary Period. The upper surface of the bottom silt–clay unit is undulating between 30 m.a.s.l. and 31 m.a.s.l. This is partly the result of erosion of the surface by the flow of meltwater rivers (except in an area 140 m downstream of the injection wells, where the clay layer is situated deeper).

The aquifer is a sandy aquifer with a thickness of approximately 9–10 m. The groundwater table is situated at 35 m.a.s.l. (5 m below land surface). The glaciofluvial sediments are composed of medium- and coarse-grained sand with gravel, which is deposited in a braided river on a sandur or in small lakes in the Late Weichselian of the Quaternary Period. The aquifer can be subdivided into three different structural and textural types.

(1) Between 30 m.a.s.l. and 31 m.a.s.l. the medium- and coarse-grained sand is deposited in the eroded channels in the silt–clay surface. The sand contains probably clay and silt clasts. In this horizon small lake basins are filled with silt in the centre of the basins and with fine- to medium-grained delta sand along the margins.

(2) From 31 m.a.s.l. to 33 m.a.s.l. the beds are dominated by medium-grained sand intercalated by medium- and coarse-grained sand. The sand is planar cross-bedded and is deposited on banks and sand waves. The stream directions of the river have been towards the west. Silty inhomogeneities can

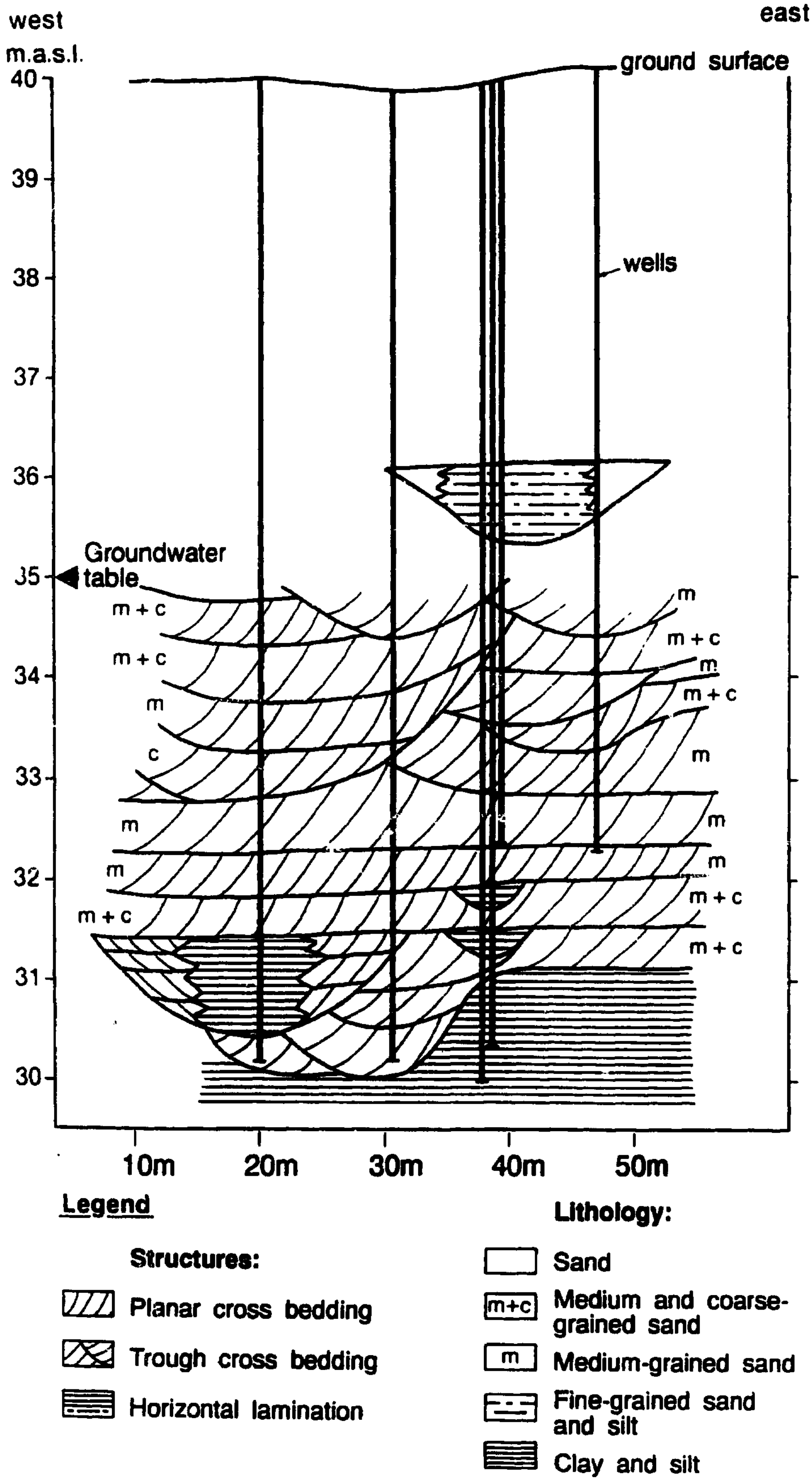

Fig. 2. Geological cross section at the research site.

be found along the bedding planes and between 32.3 m.a.s.l. and 32.8 m.a.s.l. many clay–silt clasts and silty sand clasts are deposited in the sand.

(3) Between 33 m.a.s.l. and 35 m.a.s.l. the structural framework is changed as the beds of medium- and coarse-grained sand are trough formed. Trough cross-bedding and channel fill structures show that the sand is deposited as dunes and in small channels in the river with a stream direction towards the west. Silt inhomogeneities occur in the bedding planes and as slough plug deposits. Between 33.8 m.a.s.l. and 34.3 m.a.s.l only few silty sand clasts are found in the sand, while above 34.3 m.a.s.l. many clay–silt clasts and silty sand clasts are normal sediments.

RESULTS AND DISCUSSION

A total of 334 mini slug tests have been performed at the research site. Figure 3 shows an example of hydraulic conductivity profiles determined in three adjacent locations with a horizontal distance between the locations of 0.7 m. The profiles show some similarity, but also some differences are

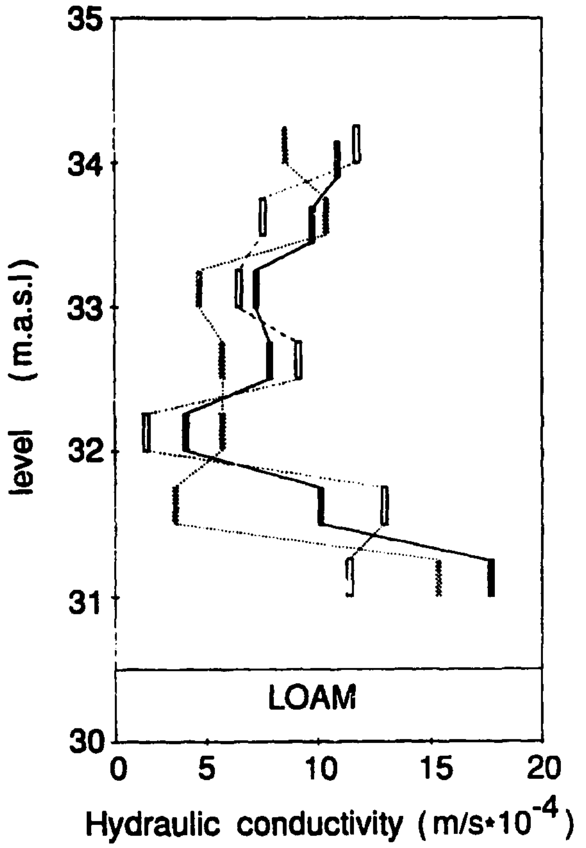

Fig. 3. Vertical profiles showing the hydraulic conductivity at three adjacent locations with a horizontal distance of 0.7 m.

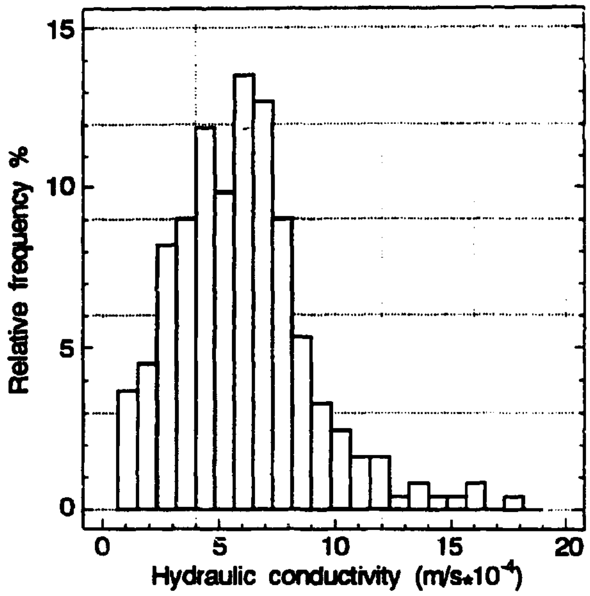

Fig. 4. Frequency plot of the hydraulic conductivity measurements by the mini slug test method.

revealed. The average vertical distance between tests is 0.5 m. In the horizontal direction the distance varies from 0.45 m to 200 m.

Frequency distribution on hydraulic conductivity

A frequency plot of the hydraulic conductivities obtained by the mini slug test is shown in Fig. 4. A χ^2 test, performed on a subset of the log normal transformed data, showed that a log normal distribution can be assumed on a 90% significance level. This result is in accordance with Lachassagne et al. (1989) stating that hydraulic conductivity data in most cases can be described by a log normal distribution. The hydraulic conductivity has a geometric mean of 5.05×10^{-4} m s^{-1} and the overall variance of $\ln K$ is 0.37.

Vertical variation in hydraulic conductivity

In order to evaluate the variations in hydraulic conductivity over depth in view of the geology of the area, as previously described, the data set was divided into subgroups for each 0.5 m of depth increment. The size of the wells screens (0.25 m) and the size of the data set do not allow for a more detailed vertical discretization. The statistics of these (7) subgroups are shown in Table 1. Two important facts are revealed. (1) The variance of the hydraulic conductivity is larger for levels 31–33 m than for levels 33–34 m. (2) Some indications on differences of the average value for the hydraulic conductivity exist. These differences are in good accordance with the geological model for the area

TABLE 1

Hydraulic conductivities determined by mini slug tests according to 0.5 m depth intervals

Level (m.a.s.l.)	N^a	$K_g{}^b$ (10^{-4} m s^{-1})	Range (10^{-4} m s^{-1})	$\sigma^2_{\ln K}{}^c$
31.0–31.5	26	6.64	1.7–22.5	0.29
31.5–32.0	30	5.50	1.6–12.1	0.28
32.0–32.5	42	3.96	0.8–10.0	0.36
32.5–33.0	42	4.92	0.8–12.3	0.46
33.0–33.5	30	4.80	2.4–7.5	0.12
33.5–34.0	37	5.36	1.3–10.4	0.23
34.0–34.5	37	6.78	1.9–15.7	0.18

[a] N is the number of mini slug tests performed at this depth.
[b] K_g is the geometric mean of the hydraulic conductivity.
[c] $\sigma_{\ln K}$ is the variance of $\ln K$.

which identified 2 layers with different bedding structure: a layer from 31 to 33 m.a.s.l. showing cross bedding and a layer from 33 to 35 m.a.s.l. showing through bedding.

The layer from 31 to 33 m.a.s.l. contains a larger number of small silt or clay clasts (between level 32.3–32.8 m.a.s.l.) which locally may yield lower hydraulic conductivities. This may be reflected in the slightly lower mean value for the layer from 32–33 m.a.s.l. A good accordance exists between the geological model and the values for the hydraulic conductivity. Based on these observations a hydrogeological model consisting of three layers is proposed (Table 2). This model will be used as follows.

TABLE 2

Hydraulic conductivities in 3 layers of the aquifer determined by the mini slug tests

Level (m.a.s.l.)	N^a	$K_g{}^b$ (10^{-4} m s^{-1})	Range (10^{-4} m s^{-1})	$\sigma^2_{\ln K}{}^c$
31.0–32.0	56	6.05	1.6–22.5	0.29
32.0–33.0	84	4.41	0.8–12.3	0.41
33.0–34.5	104	5.64	1.3–15.7	0.20

[a] N is the number of mini slug tests.
[b] K_g is the geometric mean of the hydraulic conductivity.
[c] $\sigma^2_{\ln K}$ is the variance of $\ln K$.

Comparison with tracer tests

The average value for the hydraulic conductivity can be compared with results obtained by tracer tests performed with chloride and tritium at the site. A tritium tracer slug was injected at a depth of 33.40–33.90 m.a.s.l. and the plume travelled primarily in a level from 33 to 34.5 m.a.s.l. (Jensen et al., 1992). Chloride was injected 0.5 m lower than tritium and, owing to an initial sinking of the plume, the plume travelled primarily just above the bottom of the aquifer corresponding to level 30.5–33 m.a.s.l. (Bjerg and Christensen, 1992). The average tracer velocity for tritium was 0.77 m day^{-1}, and was fairly constant between sampling points. The velocity of chloride showed an average value of 0.70 m day^{-1}.

The water table has been extensively monitored during the tracer experiments. The water table showed a clear seasonal fluctuation, but the gradient and flow direction were almost constant over the year and over the field. The approximate gradient was 4.5 ‰.

The effective porosity is estimated from Cone-Penetration tests and packing tests to approximate 0.33. This value is comparable with results found at other tracer experimental sites with similar geological history, e.g. 0.33 at the Borden Site (MacKay et al., 1986) and 0.35 at the Cape Cod site (Garabedian et al., 1988).

Assuming two-dimensional flow parallel to the bedding and the same bedding structure in all horizontal directions for each hydrogeological layer (isotropic case), the effective hydraulic conductivity (K_e) equals the geometric mean of the hydraulic conductivity (Gelhar and Axness, 1983). This allows for estimation of the linear transport velocity (v) for the three hydrogeological layers by $v = K_e \times I/\theta$, where I and θ are the hydraulic gradient and the effective porosity as discussed above, and K_e corresponds to the geometric mean of the slug test results for each layer. Employing the data from Table 2, the linear velocities are estimated to 0.71 m day^{-1}(31.0–32.0 m.a.s.l.), 0.52 m day^{-1}(32.0–33.0 m.a.s.l) and 0.66 m day^{-1}(33.0–34.5 m.a.s.l.). These estimates compare well with the tracer test determined velocities referred to above. For the bottom layer the estimates value (0.71 m day^{-1}) is identical to the observed velocity (0.70 m day^{-1}), while the estimate for the upper layer (0.66 m day^{-1}) is slightly lower (15%) than the observed velocity (0.77 m day^{-1}). The difference between estimated and observed velocity in the upper layer is supposed to be related primarily to the estimate of θ and the determination of K_e, since the gradient (I) and the observed velocity are considered to be relatively less uncertain. Bringing the estimated value closer to the observed value demands a lower value of θ or a higher value of K_e. A lower value of θ (θ should be 0.30 or slightly less) seems unrealistic unless the fraction of coarse gravel and stones is very high in this layer. It seems more likely that the K_e is underestimated in this coarse layer. Very high values of K_e may approach the value of the screen resistance of the filtertip used in this experiment (see Hinsby et al., 1992). The screens have been improved since this experiment and now show a negligible screen resistance. Missing the very high hydraulic conductivity in the slug test yields a low estimate of K_e and hence a low estimate of the linear velocity.

In conclusion, the hydraulic conductivities determined by the mini slug test are in good accordance with the linear flow velocities determined in large scale natural gradient tracer tests.

Horizontal variation

The horizontal variation in hydraulic conductivity is described by the variograms shown in Fig. 5 for the three previously identified hydrogeological layers. The number of distances within each of the lag classes vary between 12 and 294 for the variograms showed, but several other combinations of lag intervals were tested and similar results obtained. The variograms, albeit scattered, show some correlation structure on a short scale. An exponential variogram model was fitted to the variograms, see Fig. 5. The integral scale was found to be approximately 1 m for levels 31–32 m.a.s.l. and 32–33 m.a.s.l. and 2.5 m for levels 33–34.5 m.a.s.l. This is in accordance with the different layering structure of the layers. The variograms are quite uncertain owing to relatively few pairs in each lag class. This is a result of the strategy for establishing the network, which primarily was designed for water sampling purposes. From a geostatistical point of view another strategy would have been desirable.

Another network has been established at the field area at a plot adjacent to the tracer test site in order to confirm the results in Fig. 5. The geology is assumably similar at the two sites. The supplementary network consist of 60 wells with distances of 0.7 m and 60 m between wells. All iron tubes were placed at the same depth corresponding to 33.15 m.a.s.l. Figure 6 shows the distribution of the hydraulic conductivity at this plot.

The hydraulic conductivity at the supplementary site has a geometric mean of 7.6×10^{-4} m s^{-1} and the variance of ln K is 0.17. The average value is higher than at the tracer area (Table 2), but the variance corresponds quite well to the result obtained at depth 33–34.5 m.a.s.l. at the tracer area. A semivariogram is shown for the hydraulic conductivity in Fig. 7. This variogram shows the same structure as the previous variograms in Fig. 5, but owing to a larger number of especially short distances (less than 5 m) the variogram is more

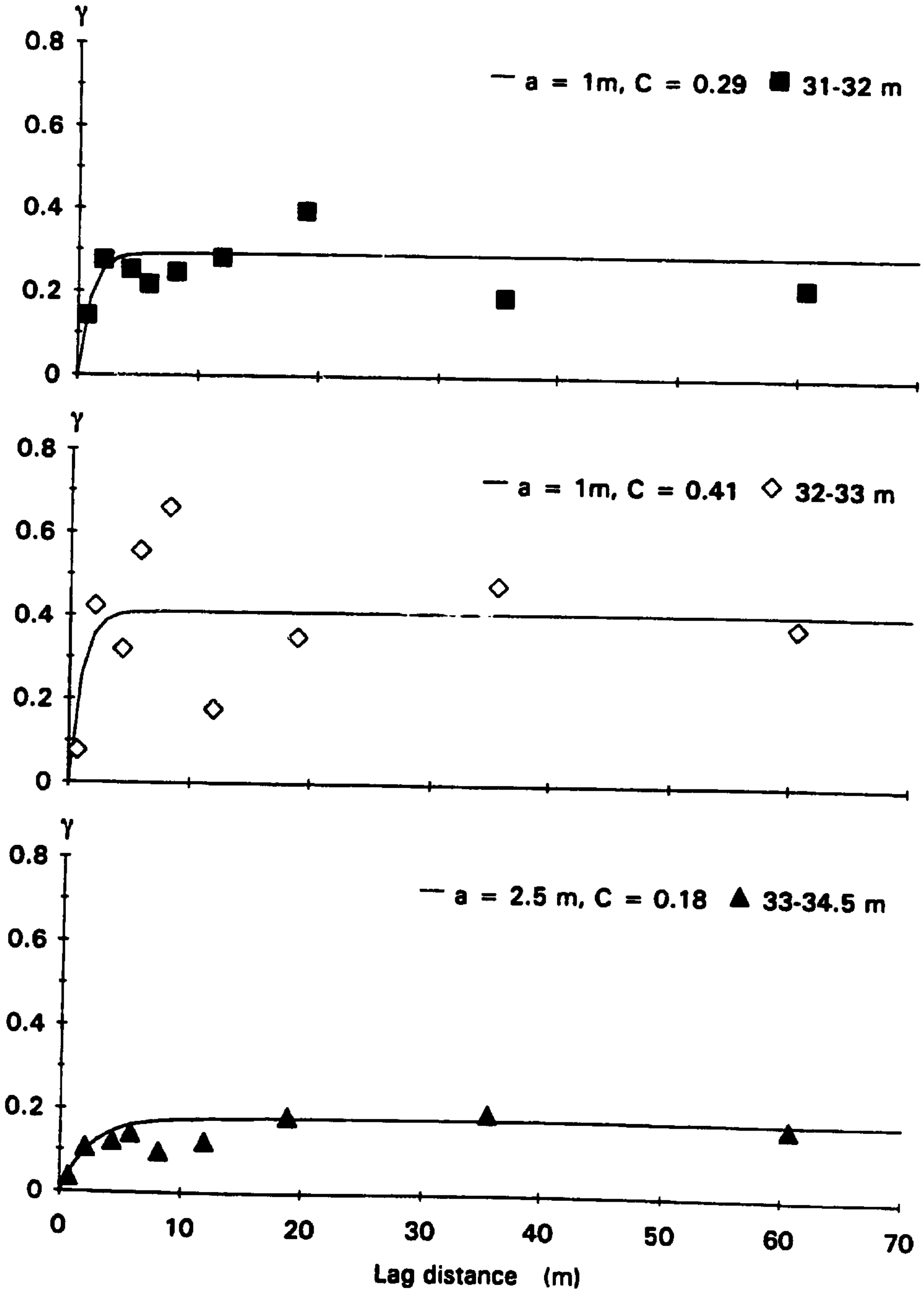

Fig. 5. Variograms and fitted models for the hydraulic conductivity of three layers at the tracer test area.

reliable. An exponential model with an integral scale of 2.5 m is applied (identical to the model used in Fig. 5 for layer 33–34.5 m.a.s.l.).

The Vejen aquifer may be described as having three hydrogeological layers of high hydraulic conductivity, small $\sigma^2_{\ln k}$ and a short horizontal correlation length (1–2.5 m). The results are in good agreement with the Borden site which

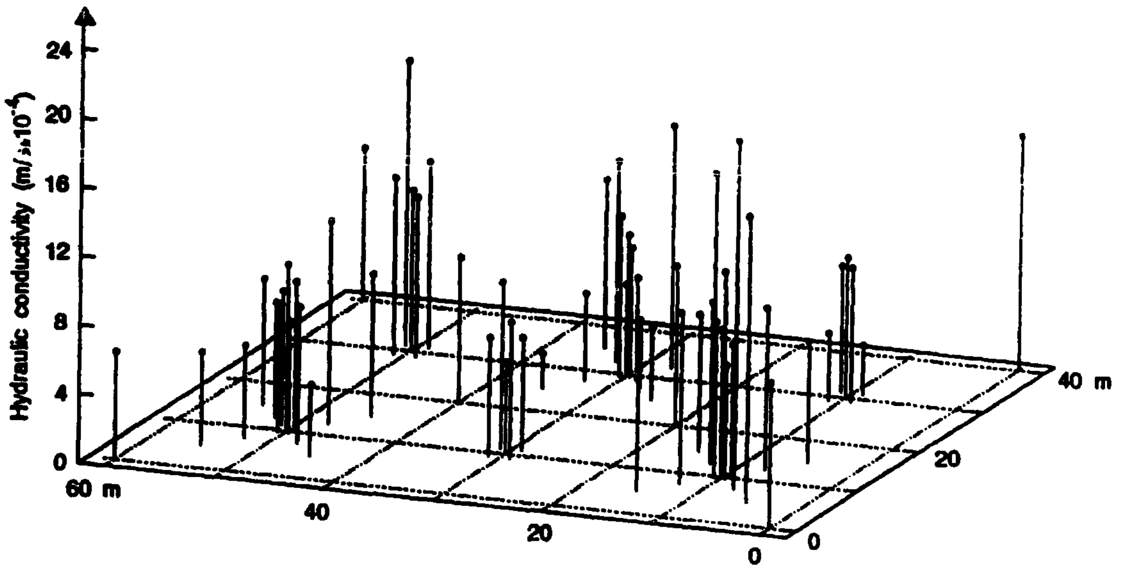

Fig. 6. Areal distribution of the hydraulic conductivity in level 33.15 m.a.s.l. at a plot adjacent to the tracer test area.

has a comparable geological history: $\sigma^2_{\ln k}$ is almost the same and the correlation length is of the same magnitude (Sudicky, 1986).

Dispersion

The correlation length for the hydraulic conductivity is an important parameter in discussing macrodispersion in aquifers. The relation between macrodispersivities and the three-dimensional spatial correlation structure of an aquifer has been developed for several special cases by Gelhar and Axness

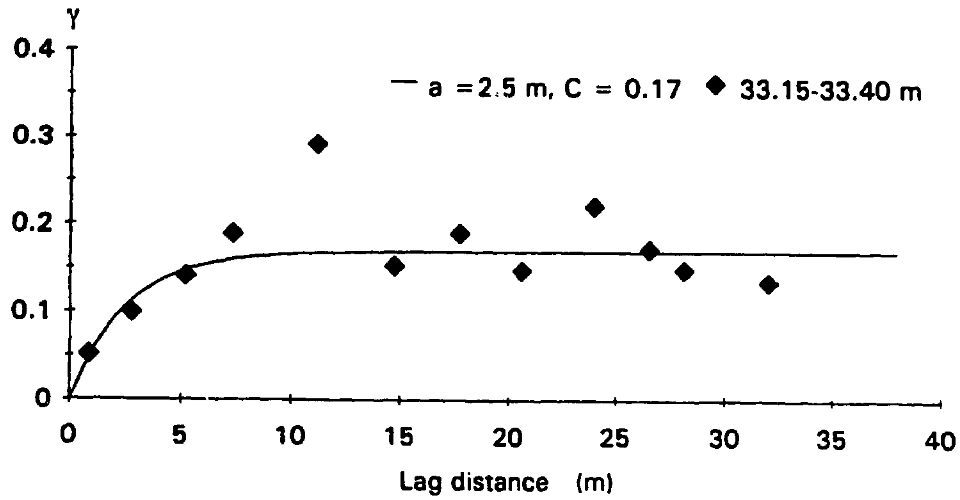

Fig. 7. Variogram and fitted model for the hydraulic conductivity in level 33.15 m.a.s.l. at a plot adjacent to the tracer test area.

(1983). The asymptotic longitudinal dispersivity A_{11} for the two dimensional isotropic case is expressed as:

$$A_{11} = \sigma^2_{\ln k} \times \lambda_h$$

The expected implication for the asymptotic longitudinal dispersivity in the investigated aquifer may be evaluated by help of this formula. Based on the obtained values of λ_h and $\sigma^2_{\ln k}$ the longitudinal dispersivity will be small, within a range of 0.3–0.5 m. Jensen et al. (1992) found in their tracer experiments performed in the same aquifer a longitudinal dispersivity of 0.42 m.

CONCLUSION

A total of 334 determinations of the hydraulic conductivity have been performed in a glacial outwash sand aquifer. The results have been evaluated in view of the geology at the site and analysed by statistical and geostatistical methods in order to describe vertical and horizontal variations. The following conclusions are made.

The hydraulic conductivity has a global geometric mean value of 5.04×10^{-4} m s^{-1} and the variance of ln k is 0.37 based on 274 local hydraulic conductivities at the tracer site. The vertical distribution of the hydraulic conductivity may be ascribed to three hydrogeological layers with high hydraulic conductivity but slightly different mean values. The model for the hydraulic conductivity is in good accordance with the geological description of the site.

The geometric mean values of the hydraulic conductivity for the three hydrogeological layers were compared with two large scale natural gradient tracer tests performed at the site. The tracer velocities showed reasonable accordance with a linear travel velocity based on measurements of the gradient and an estimate of the porosity.

The horizontal correlation of the hydraulic conductivity showed a short correlation length (in a range of 1–2.5 m) at the tracer site. These results were confirmed by mapping the hydraulic conductivity of an adjacent plot with similar geology. Based on these values it is expected that the asymptotic longitudinal dispersivity will be small (approximately in the range of 0.3–0.5 m), which is in good accordance with results obtained at the performed tracer experiments.

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