

HYDRAULIC CONDUCTIVITY, POROSITY, AND PARTICLE SIZE DISTRIBUTION OF CORE SAMPLES OF THE UPPER GLACIAL AQUIFER: LABORATORY OBSERVATIONS

Johnson N. Olanrewaju and Teng-fong Wong
 Department of Earth and Space Sciences
 State University of New York at Stony Brook
 Stony Brook, NY 11794-2100

Abstract: Core samples at depth interval of 60' to 152' were obtained from Mill Lane, north central Suffolk County of Long Island. Measurements of the porosity, grain size distribution and hydraulic conductivity were conducted in the laboratory under room conditions. While the porosities of these core samples (34% to 41%) do not change significantly, the grain size and sorting show appreciable variation with depth. The effective grain size D_{10} ranges from 0.04 to 0.15 mm, and the uniformity coefficient C_u ranges from 2 to 10. There is an overall trend for both parameters to decrease with depth. The permeameter measurements range from 0.01 to 0.2 cm/s, which correspond to the 1- and 80-percentile values of 1,357 pumping test measurements compiled by McClymonds and Franke (1972). This implies that our core samples from a single well spanning a depth interval of 92' has spatial heterogeneity comparable to field scale measurements throughout the upper glacial aquifer. To the extent that the maximum and minimum laboratory measurements represent conductivities of the sandy matrix and interbedded fine-grained material, we can predict theoretically the large-scale anisotropy induced by layered heterogeneity in the upper glacial aquifer. The theoretical predictions are in good agreement with horizontal and vertical conductivities inferred from calibration of Buxton and Modica's (1992) groundwater flow models recently developed for Long Island. Among the microstructural variables, the grain size exerts the most significant control over the magnitude of conductivity. Correlation of the permeameter measurements and effective grain size follows an approximately quadratic trend.

INTRODUCTION

Groundwater flow has important control over many geological and environmental problems (Bredehoeft and Norton, 1990). Contaminant transport and its fate within a formation depend on the hydrogeological characteristics. Since we depend on the Long Island aquifers as the sole source of water supply for agricultural, industrial, and domestic uses, the management and conservation of this valuable resource require detailed knowledge of the key hydrologic parameters. In an effort to measure these parameters under controlled conditions, we obtained core samples from the upper glacial aquifer and determined their porosities, grain size distributions, and hydraulic conductivity in the laboratory. In this paper, our laboratory data are summarized and the influence of grain size and its distribution on the hydraulic conductivity is highlighted. The permeameter measurements are compared with published field results, and implications on the anisotropy of groundwater flow in the upper glacial aquifer are discussed using a simple conceptual model.

STUDY AREA AND METHODS OF INVESTIGATION

The core samples were obtained by the Suffolk County Water Authority from a well located at Mill Lane, north central of Suffolk county, Long Island, New York (Figure 1). Specifically the well is located at the intersection of Mill Rd. and Middle Rd. in Southold, at a topographical high of about 30'. This part of Long Island is underlain by upper Pleistocene glacial deposits of the Wisconsinan glacial stage, which form the upper glacial aquifer. Samples were obtained by the split spoon method at depth of between 60' and 150' at intervals of 2'-5'. We do not have detailed information on the stratigraphy and hydrogeology of this exact location, but they are expected to be qualitatively similar to a cross-section prepared by Soren and Stelz (1984) for the Jamesport area, which is to the west of our study area. As indicated by the schematic cross-section reproduced in Figure 2, previous studies have suggested that interstadial deposits of mainly clay and silt lie just below the depth interval from which we obtained our samples.

Altogether 11 samples were analyzed. Table 1 lists the depths at which our samples were obtained. The porosities were determined using the *volumetric method*. A known volume of water (V_{water}) was added to a dry sample until it was completely saturated. If the bulk volume of the saturated sample is denoted by V_{sat} , then the *porosity* is given by $\eta = V_{water}/V_{sat}$. We also used the gravimetric method to determine the porosity, and data from the two techniques are comparable. Since the gravimetric method depends on an assumed value for the solid grain density, we prefer to use data from the volumetric method. As shown in the compilation in Table 1, the porosities

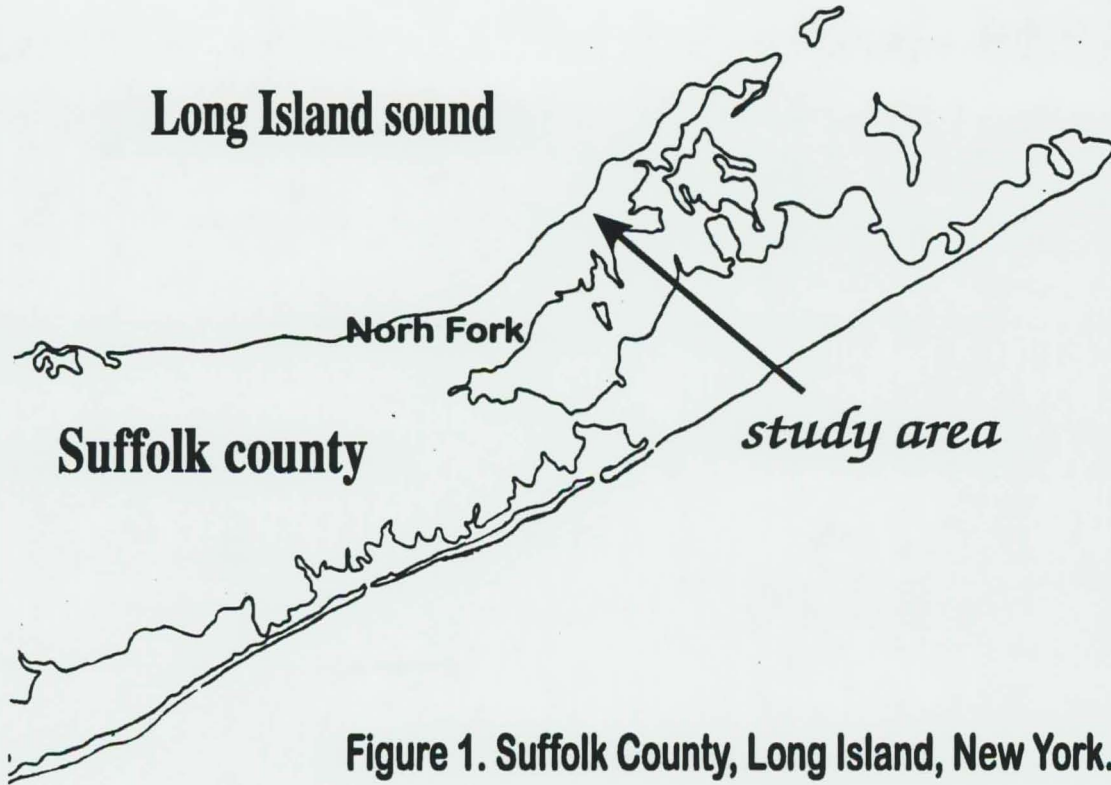


Figure 1. Suffolk County, Long Island, New York.

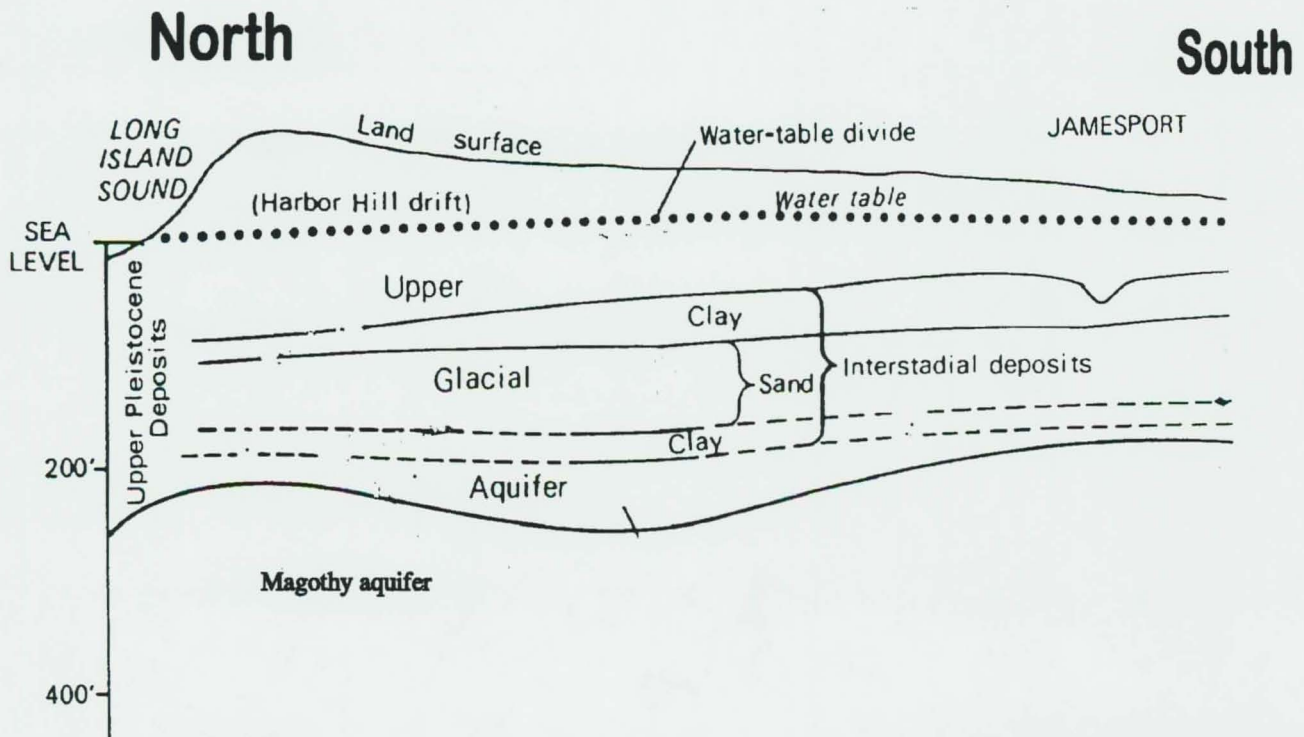


Figure 2. Cross section showing stratigraphy and hydrologic units in the Jamesport area. This cross section is to the west of our study area in Mill Lane.

span a relative narrow range from 34% to 41%. There seems to be a slight trend for porosity to increase with depth.

The *sieve technique* was used to determine the grain size distributions. The *effective grain size* was characterized by the coefficient D_{10} , the value corresponding to 10% finer by weight. We characterized the sorting by the *uniformity coefficient* defined to be $C_u = D_{60}/D_{10}$ where D_{60} is the grain size corresponding to 60% finer by weight (Table 1). If $C_u > 4$, the sample is commonly referred to as “poorly sorted”, and if $C_u < 4$ the sample is “well sorted” (Fetter, 1994). Figure 3 shows the two end-members with the maximum and minimum uniformity coefficients. The sorting of the shallowest sample (from the depth interval 60’-62’) was extremely poor, with a relatively high value of $C_u = 10$. In contrast, the sample from depth interval 135’-137’ was very well sorted with $C_u = 2$.

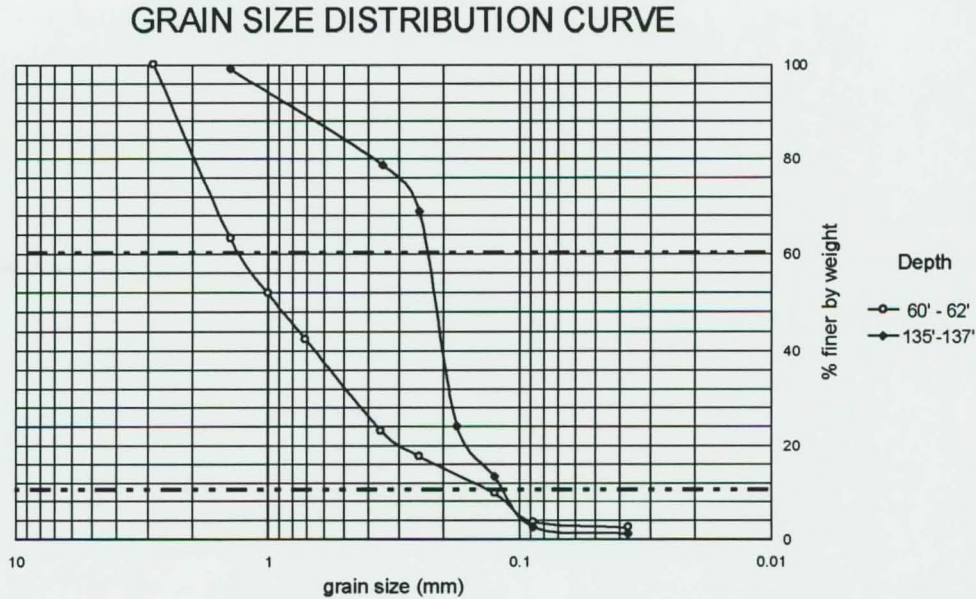


Figure 3. Particle size distribution of two samples (from depths of 60’-62’ and 135’-137’) determined by sieve analysis.

Figure 4 summarizes the variation of sorting characteristics with depth. There is an overall trend for the uniformity coefficient to decrease with increasing depth, implying that the deeper samples are better sorted. As is indicated by the size distribution shown in Figure 3, the shallowest sample (from a depth of 60’-62’) shows the poorest sorting.

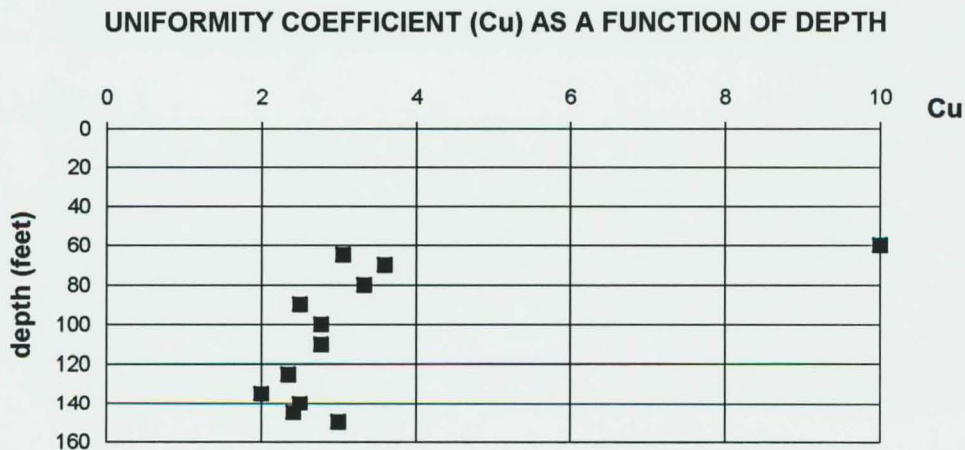


Figure 4. Sorting characteristics of samples as a function of depth. There is an overall trend for the deeper samples to be well sorted.

Figure 5 shows the effective grain size as a function of depth. There is an overall trend for grain size to decrease with depth. There is a dramatic decrease in grain size at a depth of 150'-152', in the close vicinity of the interstadial bed of clay and silt in the North Fork (Figure 2).

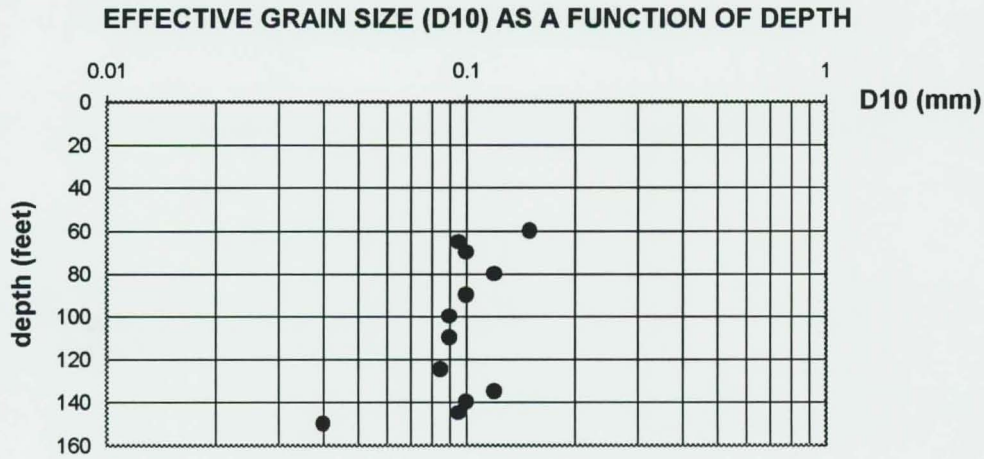


Figure 5. Effective grain size as a function of depth. There is an overall trend for grain size to decrease with depth.

In addition to porosity and grain size distribution, we also use a permeameter to determine the *hydraulic conductivity* under room conditions of temperature and pressure. Fluid flow in porous media is governed by Darcy's law, which states that the specific discharge q is proportional to the hydraulic gradient i with the proportionality constant given by the hydraulic conductivity $K = k(\rho g) / \mu$. The fluid is characterized by the density ρ and viscosity μ , whereas the rock pore space is characterized by the permeability k . In this study, conductivity and permeability measurements are reported in units of cm/s and darcy respectively. (For water under nominal conditions, $k = 1$ darcy $\Rightarrow K \approx 0.001$ cm/s.)

Our permeameter can be operated in two different modes. The *falling head* technique can be implemented for relatively impermeable samples (with $K < 10^{-4}$ cm/s, Freeze and Cherry, 1979). However, since all our samples were relatively permeable, it was more straight forward to use the *constant head* technique. The cumulative discharge V was measured as a function of time t keeping the head difference (dh) across the sample (of length $d\ell = 6$ cm) constant. By linear regression of the discharge-time data, the volumetric discharge Q was determined. Since the hydraulic gradient was kept constant at $i = \frac{dh}{d\ell}$, the conductivity can be evaluated using the relation

$$K = -\left(\frac{Q}{A}\right) \left/ \left(\frac{dh}{d\ell}\right)\right. \quad \text{where } A = 8.74 \text{ cm}^2 \text{ is the cross-sectional area of the sample.}$$

Under room pressure condition, the hydraulic gradient should be less than 0.5 or so to avoid the development of quick sand condition within the sample assembly (Fetter, 1994). In this study, we varied the head difference from 0.7 cm to 3 cm, corresponding to $i = 0.12 - 0.5$. In accordance with Darcy's law, the conductivity was observed to be independent of the hydraulic gradient. However, we did observe a trend for conductivity to decrease somewhat with time if the sample assembly was allowed to be saturated with water for several days. The effect seems to be more appreciable in the fine-grained samples. The phenomenon is probably related to time-dependent physical and chemical processes, such as clogging actions induced by oxidation and compaction creep processes. For each sample, we conducted at least 3 permeameter measurements under different hydraulic gradients. Table 1 lists the range of the experimental data, as well as the arithmetic mean for the conductivity of each sample.

DISCUSSION

Our laboratory measurements of the hydraulic conductivity range from 0.013 cm/s to 0.21 cm/s, (corresponding to values of $k = 13-210$ darcy and $K = 275-4,452$ gpd/ft² or 35.6-575 ft/day). Our data for the Mill Lane core samples are in reasonable agreement with McClymonds and Franke's (1972) compilation of conductivity

inferred from 1,357 wells in the upper glacial aquifer, which range from 100 gpd/ft² to 8,000 gpd/ft² (Table 2) and show a frequency distribution approximately following a log-normal trend. Our laboratory data fall between the 1 and 80 percentile values of the *pumping test* data, implying that our core samples from a single well spanning a depth interval of 92' has spatial heterogeneity comparable to field scale measurements throughout the upper glacial aquifer.

In multi-layered numerical models recently developed for groundwater flow on Long Island, the upper glacial aquifer has been modelled as an *anisotropic* layer. Constraints on the conductivity values and their anisotropy were provided by calibration of the numerical models. In Buxton and Modica's (1992) finite element model, two sets of conductivity values were used in the upper glacial aquifer: $K_{horizontal}=230$ ft/day (.084 cm/s) and $K_{vertical}=23$ ft/day (0.008 cm/s) for the *outwash* sections; and $K_{horizontal}=75$ ft/day (0.027 cm/s) and $K_{vertical}=1$ ft/day (0.0004 cm/s) for the *moraine* sections (Table 2). Similar values had been used previously in the 3-dimensional finite difference and particle tracking study of Buxton et al. (1991). Our laboratory values (Table 1) fall between these ranges.

It is not practical to obtain direct information on conductivity anisotropy in the laboratory. Indeed the anisotropy inferred from numerical model calibration is very difficult to characterize even with extensive pumping tests conducted in the field. The very significant anisotropy is partly attributed to large-scale *layered heterogeneity*. This effect can be analyzed using a simple conceptual model. Consider the upper glacial aquifer to be of uniform thickness L , interbedded by n interstadial layers of relatively impermeable material (clay and silt). If the bed thickness of an individual layer is denoted by ℓ_i , then the total thickness of the interstadial beds is given by $L_b = \sum_{i=1}^n \ell_i$ and the total thickness of the sandy matrix is $L_m = L - L_b$. An estimate of the conductivity of the clay and silty materials is provided by our laboratory measurements for the sample at depth of 150'-152': $K_b \approx 0.01$ cm/s, and similarly we can infer from our laboratory measurements of the shallower samples that the conductivity of the sandy matrix $K_m \approx 0.1$ cm/s. For this type of layered heterogeneity, the effective conductivities in the horizontal and vertical directions are given by (Freeze and Cherry, 1979):

$$K_{horizontal} = \frac{L_m}{L} K_m + \frac{L_b}{L} K_b \quad \frac{1}{K_{vertical}} = \frac{L_m}{L K_m} + \frac{L_b}{L K_b} \quad (1)$$

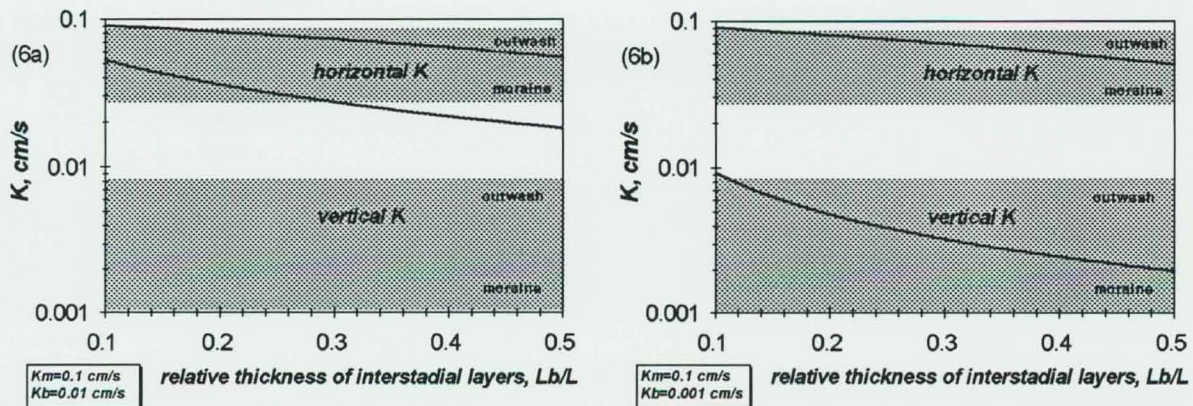


Figure 6. Effective conductivities in the horizontal and vertical directions as a function of interstadial deposit thickness. The upper and lower solid lines represent respectively the horizontal and vertical conductivities calculated by equation (1) assuming isotropic values of K_m and K_b (for the sandy matrix and fine-grained material in the interstadial layers respectively) as indicated in the figures. The grey areas indicate values used by Buxton & Modica (1992) in their numerical model.

The resultant conductivities in the horizontal and vertical directions are shown respectively by the upper and lower solid curves in Figure 6a. For relative thickness L_b/L greater than 0.15, the calculated values for the horizontal K fall between the ranges used in the numerical models, but the calculated values for the vertical K are consistently higher than what were used by Buxton and Modica (1992). One possible explanation for the apparent discrepancy is that since our deepest sample was from a location somewhat higher than the interface with the

interstitial deposits, the laboratory measurement only represents an upper bound on K_b . Therefore the interstitial beds may be more impermeable than what was assumed in Figure 6a. The effect is illustrated in Figure 6b, in which we use a value of $K_b = 0.001$ cm/s, an order of magnitude lower than the value in Figure 6a. Indeed this result is in better agreement with both the vertical and horizontal conductivities assumed in the numerical models.

While the agreements shown in Figures 6a and 6b are encouraging, it is also important to keep in mind the limitations of both the laboratory studies and numerical modelling. Our measurements were made under room conditions, and the *in situ* values should be somewhat lower due to compaction from the overburden pressure. We have only data from a single well, and we need to have a more extensive data base to assess the extent to which our data are representative of different locations on the upper glacial aquifer. On the other hand, it should also be noted that the model calibration of a multi-layered model involves trade-offs among the parameters of the different layers, and therefore the inferred values for the upper glacial aquifer are sensitively dependent on the conductivities assumed for the other aquifers in the multi-layered models.

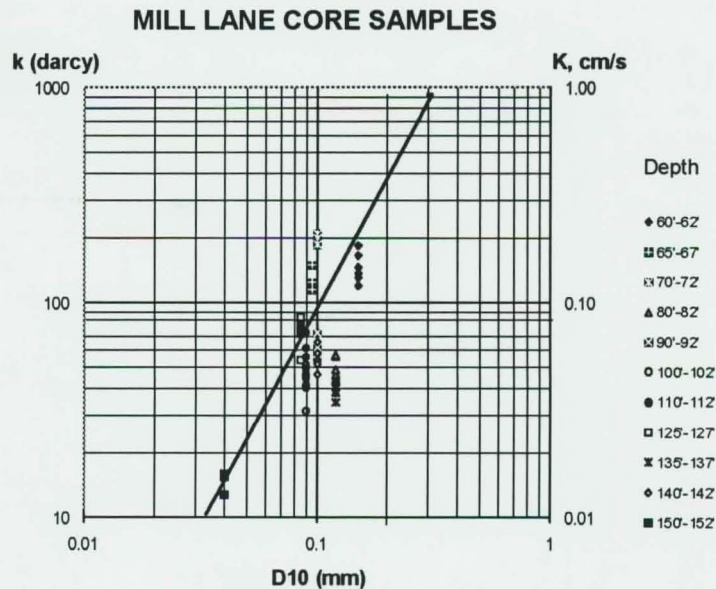


Figure 7. Correlation of conductivity K and permeability k with effective grain size D_{10} . The straight line indicates quadratic dependence

Permeability is a manifestation of the size distribution and connectivity of the pore space. Among the various microstructural variables, grain size exerts the strongest control over the hydraulic conductivity and permeability of unconsolidated materials. That our measurements for the conductivity vary by a factor of 20 is primarily due the grain size which varies by a factor of 4. The correlation of our experimental measurements of conductivity K (and permeability k) with the effective grain size D_{10} is illustrated in Figure 7. As expected, conductivity and permeability increase with increasing grain size. Hydrodynamic analysis of laminar flow suggests that the hydraulic conductivity and permeability should have a quadratic dependence on grain size (Hubbert, 1940), and empirical measurements have shown that sediments and soils of comparable sorting and angularity follow Hazen's formula: $K = C D_{10}^2$. On a log-log plot, this formula corresponds to a straight line with a slope of 2, and as shown in Figure 7 our data follow this trend approximately. Alternatively, we can characterize the grain size dependence with the mean grain size D_{50} . Shepherd (1989) recently suggested the power-law $k = c D_{50}^n$, with the exponent n ranging from 1.65 to 1.85. Both n and c are sensitively dependent on sorting characteristics, textural maturity and degree of induration. Although we can attempt to place tighter constraints on the correlation between conductivity and grain size by regression analysis, such a refined approach is probably unwarranted at this stage given the data scatter and the appreciable variation in sorting evident in our samples.

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Table 1. Summary of experimental measurements of effective grain size, sorting porosity and hydraulic conductivity

Depth (ft)	Number of Experiment	Effective Grain Size, D_{10} (mm)	C_u	Porosity η	Range of Permeameter Measurements: k (darcy) K (10^{-3} cm/s)	Arithmetic Mean: k (darcy) K (10^{-3} cm/s)
60' - 62'	7	0.15	10.0	0.34	119.88 - 184.78	147.29
65' - 67'	5	0.095	3.05	0.35	114.61 - 149.56	124.36
70' - 72'	3	0.1	3.60	0.343	187.35 - 210	198.8
80' - 82'	5	0.12	3.33	0.371	38.10 - 56.9	49.14
90' - 92'	5	0.1	2.50	0.357	54.02 - 72.30	63.44
100' - 102'	5	0.09	2.78	0.362	31.23 - 51.44	42.61
110' - 112'	7	0.09	2.78	0.376	40.3 - 72.48	57.10
125' - 127	9	0.085	2.35	0.396	53.95 - 84.90	72.86
135' - 137'	5	0.12	2.00	0.405	34.27 - 42.61	39.22
140' - 142'	5	0.1	2.50	0.403	46.25 - 65.20	55.02
150' - 152'	3	0.04	3.00	0.383	12.75 - 15.97	14.74

Table 2. Magnitude and anisotropy of the hydraulic conductivity of the upper glacial aquifer inferred from pumping tests and numerical model calibrations

Technique	Range of Conductivity	Reference
Pumping tests (total number: 1,357)	100 - 8,000 gpd/ft ² (mean = 2,304 gpd/ft ²) 4.72 - 378 x 10 ⁻³ cm/s (mean = 109 x 10 ⁻³ cm/s)	McClymonds and Franke (1972)
Calibration of 2-dimensional finite-element model of groundwater flow on Long Island	Outwash: $K_{horizontal}$ = 230 ft/day (84 x 10 ⁻³ cm/s) $K_{vertical}$ = 23 ft/day (8.4 x 10 ⁻³ cm/s) Moraine: $K_{horizontal}$ = 75 ft/day (27 x 10 ⁻³ cm/s) $K_{vertical}$ = 1 ft/day (4 x 10 ⁻⁴ cm/s)	Buxton and Modica (1992)