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Abstract

Five methods (tidal damping, tracer, slug, permeameter, and grain-size analysis) for the determination of hydraulic conductivity (K) were compared in both upland and shoreline regions of a sandy outwash aquifer adjacent to a shallow coastal embayment. The 7 m tracer test and tidal damping methods appeared to determine in situ K over sufficient horizontal scales to integrate the spatial variability of K within the aquifer, with the tracer test giving the most reproducible K estimates for the upland, 109 m/d. The tidal K measurements were similar to the tracer results, 106 m/d, integrating K over greater distances (30 to 85 m), but were highly variable due to the large aquifer storativity, small tidal amplitude, and irregular tides.

The smaller scale methods yielded lower upland K estimates with the lowest value from permeameters, 29 m/d, only about one-third of the other four methods. Estimates of upland K by both grain-size and slug tests were comparable, 86 m/d and 71 m/d, respectively. However, the K derived from grain-size methods was dependent not only upon the sediment but on the method selected, with consistently higher values of K generated by the Hazen equation than the Krumbein and Monk equation. Verification by in situ tests may be useful in removing this bias.

Significant differences in K were found between shoreline and upland regions, though the relative magnitude of the difference was not constant: among the four methods where comparative measurements were possible, shoreline K ranged from 42% to 75% of upland values. Slug tests yielded the largest differences, 30 m/d and 71 m/d, with permeameter, 18 m/d and 24 m/d, the lowest. The lower K of shoreline sediments appears to be the result of processes associated with coastal embayments and should be addressed in the measurement and modeling of coastal ground-water discharge.

Introduction

A major focus of current coastal research is the role of ground water in transporting nutrients to coastal embayments and harbors. In areas with highly permeable soils, ground water is a major pathway for the transport of terrestrially derived nutrients (Johannes, 1980; Johannes and Hearn, 1985; Weiskel and Howes, 1992). The interest in nutrient transport stems from the degrading impacts of progressive eutrophication of nearshore ecosystems (Johannes, 1980; Nixon, 1986; Kemp et al., 1990; Nixon, 1992), and the resultant losses of economic resources. Acquiring the quantitative data to develop an understanding of this problem has led to an emphasis upon multidisciplinary research involving hydrologists, biogeochemists, ecologists, and environmental managers (Weiskel and Howes, 1991).

One of the more rigorous methods for quantifying ground-water flow compatible with the requirements of ecosystem-scale studies is based upon Darcy's law. The advantages of this approach are that both the rates and spatial distribution of discharge are described. Ground-water discharge is calculated using Darcy's equation (Fetter, 1988):

$$Q = -KA\ dh/dl \qquad (1)$$

where Q = volume discharged to the coast (m^3); K = average aquifer hydraulic conductivity (m/d); A = cross-sectional area through which ground water passes (m^2); dh/dl = hydraulic gradient (dimensionless). While the hydraulic gradient and the aquifer cross section of discharge can be determined through relatively direct, simple field measurements, the determination of the aquifer hydraulic conductivity (K) is more difficult, and the quality of the available techniques less certain. Difficulties in the measurement of K stem from spatial heterogeneity, errors introduced by measurement techniques, and most importantly, scaling problems resulting from the need to extrapolate from measurements made at centimeter and meter scales, to watershed scales (Sudicky, 1986; Melville et al., 1991; Bjerg et al., 1992). While a variety of methods can be used to estimate watershed K, in the present study we evaluated five methods appropriate for use in a Darcian estimation of ground-water discharge to a shallow coastal embayment. We emphasize "coastal aquifer" because of the additional approaches for measuring K that are available where a tidal source is present, the added problems of using other techniques under these conditions, and the possibility of consistent spatial variations in K between inland (upland) and nearshore (shoreline) aquifer zones. The five methods sampled K over a range of spatial scales, and were used in parallel under two sets of aquifer conditions, shoreline and upland.

Methods

Study Site: The study was conducted in the unconfined aquifer within the watershed of Little Pond, a shallow coastal embayment in Falmouth, Massachusetts (41°32.33

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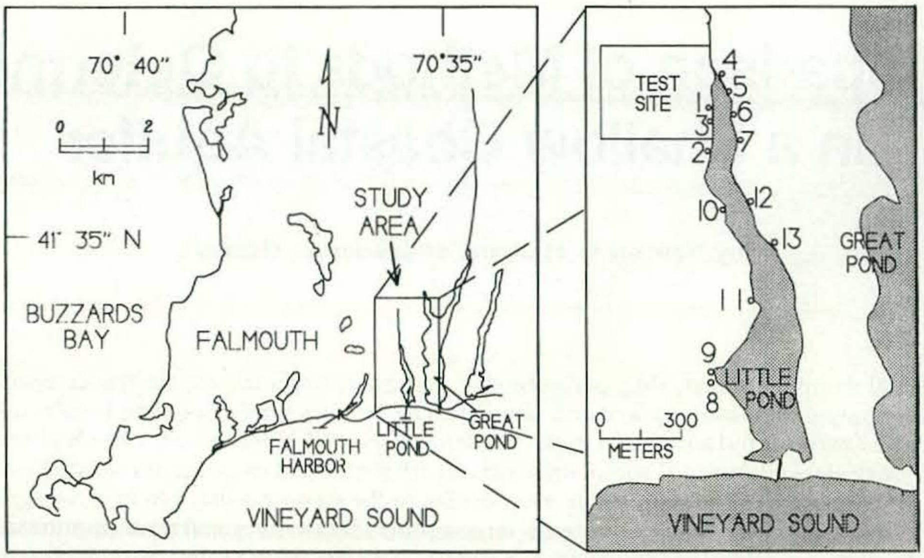

Fig. 1. Location of the study area, Little Pond, Falmouth, Cape Cod, Massachusetts. Inset shows test site location. Numbers denote locations of shoreline vibracores. Sites 10-13 also had slug tests.

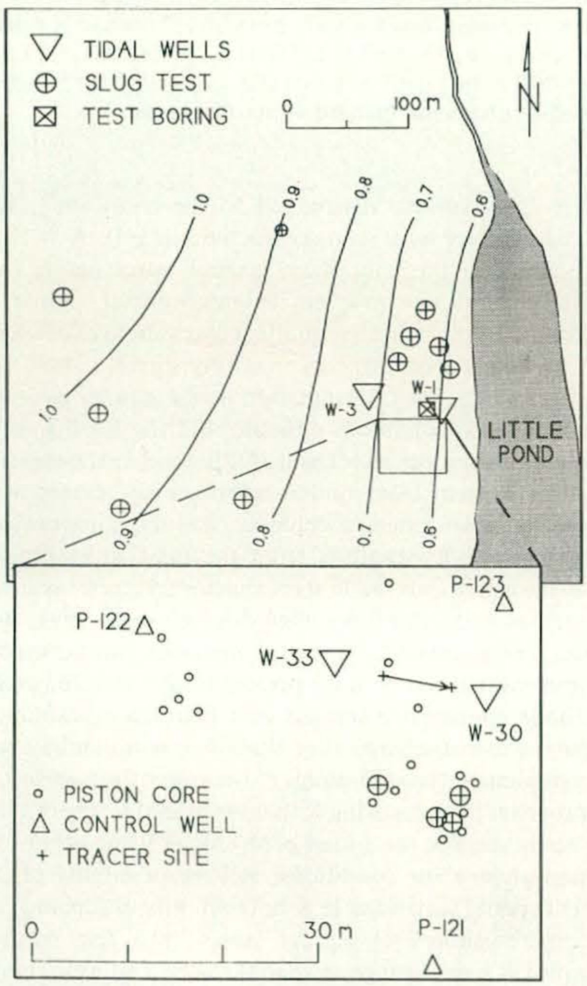

Fig. 2. The primary test site with inset showing the location of the tracer test and additional slug test wells. Inverted triangles are wells where tidal damping measurements were made. Upright triangles are 1.2 cm diameter water-table wells for tracer hydraulic-gradient measurements. The water-table contour interval is 0.1 m. The test boring was located 1 m from tidal well W-1. All wells shown are "upland" as opposed to shoreline sites which are shown in Figure 1.

N, 70°37.34 W; Figure 1). Tidal exchange with Nantucket Sound is through a small inlet at the southern end which attenuates the tidal range from 50 to about 20 cm. Little Pond is brackish, with a salinity gradient of 10 ppt at the upper end to 30 ppt near the inlet. The aquifer consists of deposits of glacial outwash (Oldale and Barlow, 1986). Vertical exposures of the sediment at a gravel pit 2.4 km northeast of the test site, (elevation $\approx$ 8 m NGVD) display abrupt grain-size variations typical of glaciofluvial sediments, and are similar to nearby exposures (9 km north, described by LeBlanc et al., 1991). Significant heterogeneity exists both laterally and parallel to the direction of deposition. Beds of medium sands are typically underlain, overlain, and/or found adjacent to, lenses of coarse sand and gravel. Outwash material in the aquifer around Little Pond is primarily coarse pebbly sand (mean grain-size 0.73 mm, n = 88). At the test site (Figure 1) some fining downwards in the upper 10 m of the aquifer was observed in split-spoon samples (d_{10} = 0.28 to 0.19 mm) taken during a boring which penetrated both the full thickness of the outwash aquifer (to −15.6 m elev. NGVD) and the deltaic glaciolacustrine deposits beneath.

The primary test site for both shoreline and upland sampling was within an abandoned gravel pit (Figure 1), located in the area of highest ground-water discharge to the embayment as determined from monthly local water-table maps and measurements of ground-water seepage at the shoreline (Vanek, 1993; Millham and Howes, in press). Ground water at the test site flows southeasterly towards Little Pond (Figure 2). Additional measurements were made elsewhere along the shoreline of the embayment to increase the size of the nearshore sample set (Figure 1).

K Determinations

Parallel measurements of K in upland and nearshore aquifer sediments were conducted using three in situ field methods, a natural gradient tracer test, tidal damping in wells, and slug tests, and two laboratory methods, permeameter tests, and grain-size analysis. The scale of spatial integration ranged from 0.05 to 80 m. The largest scale integrations of K, 30 to 80 m, were obtained by the tidal damping method. Although natural-gradient tracer tests have been conducted at spatial scales similar to those of tidal determinations (e.g. LeBlanc et al., 1991), tracer tests are also used to integrate K over smaller scales (5-10 m). In contrast, slug test measurements in highly permeable aquifers generally integrate over a scale of less than 1 m, due to the high specific yield and the small volume of water displaced by such tests. Both laboratory procedures, permeameter and grain-size analysis, usually require sample removal by coring, which limits their spatial integration of K to a very small scale (cm).

Transmissivity: Tidal Damping Method: Transmissivity was determined by two approaches: the damping of the tidal signal with increasing distance from the shoreline; and the time delay in the arrival of tidal maxima or minima at the inland wells. The relationship between the amplitude of water level fluctuation in wells and the tidal amplitude is, after Ferris (1963):

$$s_r = 2s_0 \exp[-x\sqrt{(\pi S/t_0 T)}] \qquad (2)$$

where s_r = range of fluctuation of ground-water level (m); s_0 = rise or fall of the tidal source, e.g., Little Pond (m); S = specific yield of the aquifer (dimensionless); t_0 = period of the tidal signal (d); T = aquifer transmissivity (m^2/d); and x = distance from tidal source (m). Specific yield (0.23) was based upon the reported results of a pump test conducted in outwash sands approximately 9 km north of the test site (Moench et al., 1993). Water level fluctuations at paired monitoring wells W-1 and W-3 (Figure 2) were measured with water level floats and chart recorders. Wells W-30 and W-33, instrumented with vented pressure transducers and a data logger, were located to bracket the path of the tracer experiment (Figure 2). Tidal levels were obtained from Little Pond by a float and chart recorder at a gaging station. The second tidal approach calculates transmissivity from the time delay of the arrival of tidal maxima or minima at inland wells expressed as (Ferris, 1963):

$$T = (x^2 S t_0)/(4\pi t_1^2) \qquad (3)$$

where T, x, S, t_0 are as in equation (2); and t_1 = time delay to successive maxima or minima, between embayment and water-table wells (d). Both tidal methods calculate K from its relationship to transmissivity (Fetter, 1988):

$$T = Kb \qquad (4)$$

where K = hydraulic conductivity (m/d), and b = aquifer thickness (m). In the primary test site, b was determined by a test boring to be about 15.6 m. Differences in saturated thickness b, due to the hydraulic gradient between the well pairs, was about 0.6% of total aquifer thickness and was ignored in the calculations.

Equations (2) and (3) were developed for confined conditions but can be used in unconfined conditions if s_r is small compared to b (in this case $s_r < 0.10$ m while b is 15.6 m), and if the observation well is far enough from the submarine outcrop of the aquifer to avoid vertical flow (in this case the closest well was > 30 m from the shore).

Results: An analysis of 58 flood and ebb tides by the tidal damping method [equation (2)] yielded erratic results. It appears that the signal amplitude at the water-table surface was highly damped due to the large specific yield of the unconfined aquifer (Erskine, 1991), and that the applicability of this method in unconfined sand-and-gravel aquifers may be limited. Due to the inconsistent results, we discontinued this method.

Thirty flood tides were analyzed by the time delay method [equation (3)]. The mean K (± s.d.) calculated for the well (W-1) closest to the shore, 41 (12) m/d, was lower than the more distal well, 86 (22) m/d (W-3; Table 1). While the same spatial trend was seen in the second well pair 98 (23) m/d and 126 (26) m/d (W-30 and W-33; 16 flood tides), the values of K were higher than for the first pair. In both cases more inland wells (W-3 and W-33) were significantly higher (independent t-test, $p < 0.05$); however, the differences between W-3 and W-30 were not significant ($p < 0.05$). Some of the variation between the well pairs may be ascribed to heterogeneity in the aquifer; however, the higher absolute transmissivities of the W-30/W-33 pair may be due to seasonal differences in the soil moisture conditions at or above the capillary fringe: data from well pair W-1/W-3 were collected at a time of low recharge, in the summer and early fall, while pair W-30/W-33, were collected at a time of high recharge in the late winter and early spring. Variability was fairly high for all four wells with coefficients of variation of 21 to 29%. This was partly due to low signal-to-noise ratios (small tidal range) and partly to the variable nature of the tidal regime (primarily in ebb durations) in Little Pond.

Tracer Tests: A natural-gradient tracer test was conducted at the test site using sodium chloride as the conservative tracer. Three 1.27 cm I.D. water-table wells provided local triangulation for determination of the hydraulic gradient (P-121-123, Figure 2). The injection well consisted of a length of 0.32 cm I.D. polyethylene tubing fitted to an aluminum sampler tip (slotted length 3.1 cm; Kerfoot and Soderberg, 1989), positioned 50 cm below the water table, upgradient of a fence of six multilevel samplers (MLS; Smith et al., 1991). Each MLS (2.5 cm O.D.) had five sampling ports spaced vertically at 40 cm intervals, consisting of 0.32 cm I.D. polyethylene tubes with nylon screens.

The injectate was made by dissolving 40 g of Cl^- as NaCl first in one liter of distilled water, which was then mixed in the field in a glass carboy with 18 liters of ground water drawn from the injection well. The injectate chloride concentration was 2105 mg/l.

The initial fence of MLS was placed 3 m downgradient of the injection well. As the injectate front advanced, additional fences of multilevel samplers were installed at distances of 4.5 m and 7.05 m. Disturbance of the aquifer by MLS installation was minimized by the small MLS diameter (2.5 cm) and hand installation. Ground water was collected daily for chloride analysis with 60 ml syringes. The riser tubes were flushed by withdrawing and discarding 25 ml of water (2-3 tube volumes), before withdrawing each 50 ml sample. Samples were stored in 60 ml polyethylene bottles. Chloride concentration was determined with an ion-specific electrode (Orion #9617B). Electrode sensitivity was approximately 1 mg/l Cl^-; standard curves were run for each analytical series; breakthrough was taken as the time of arrival of maximum solute concentration. Hydraulic conductivity was calculated from a form of Darcy's law (Fetter, 1988):

$$v = \frac{-K\ dh/dl}{n_e} \tag{5}$$

where: v = average liner velocity (m/d); K = hydraulic conductivity (m/d); n_e = effective porosity (dimensionless); dh/dl = hydraulic gradient (dimensionless). Effective porosity was estimated from the porosity of sections of piston cores (n = 15) collected from the test site. The mean (± s.d.) porosity was 0.339 (0.0054), which compares well with effective porosity of 0.33 to 0.41, estimated for outwash sands 9 km north of the test site (Garabedian et al., 1991).

Results: The tracer cloud was captured by sampler fences at distances of 3 m, 4.5 m, and 7.05 m, with peak $C/C_o = 0.18$ (Figure 3). K was determined from equation (5), using the measured ground-water velocity (from the breakthrough of chloride; Figure 3), effective porosity, and hydraulic gradient (0.0014). Mean K (± s.d.) for the three fences (Table 1) was 109 m/d (10), and replication of calculated K was good at each distance (multiple ports were hit within a fence). Maximum concentrations exhibited at the deeper sampling ports at the 4.5 and 7 m fences indicate slight sinking of the tracer cloud during transport due to density contrasts and recharge to the water table, while lower concentrations at the two closest fences indicate non-interception of the cloud core. Background concentrations of Cl^- in the aquifer were low, 2.5 to 6 ppm, only a few percent of the breakthrough concentrations.

Slug Tests: Slug tests were conducted in 5.1 cm I.D. PVC wells with 0.025 cm slotted screens, 0.48 m and 1.25 m long, with at least two replicate tests per well. Slug tests were conducted in 10 preexisting test-site monitoring wells, in 9 newly installed wells at the test site, and in 4 wells along the shore. All 23 wells were installed without sand packs

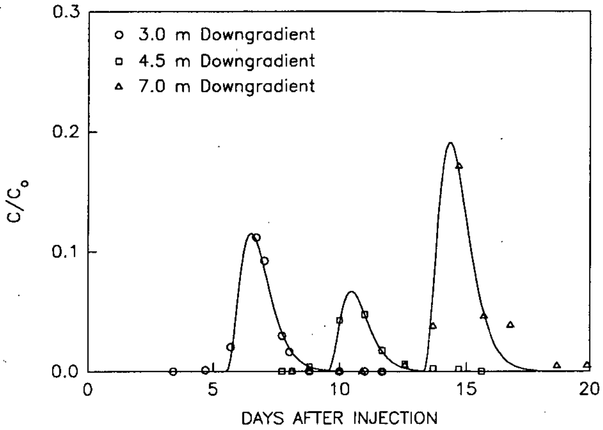

Fig. 3. Arrival of the tracer cloud at sampler fences 3.0 m, 4.5 m, and 7.0 m downgradient of injection point. Maximum concentrations at each fence were found at 1.6, 1.6, and 2.0 m depths below the water table, respectively. The solid curves fitted to the actual data were used to refine the time of arrival of maximum solute concentration for calculating ground-water velocity. Lower concentrations at the 3.0 and 4.5 m fences were due to the plume core missing the samplers.

(Figures 1, 2). Shoreline slug test locations were selected to have minimal vertical ground-water flow. The new wells were installed in hand-augered or vibracore boreholes in order to minimize disturbance to the aquifer matrix. The recovery rate of water in the well when a slug (usually 1.94 liters) was withdrawn was recorded by a pressure transducer (Thor #DXPE 01B) and a digital data logger (Unidata Model 7000). K was calculated using the method of Bouwer and Rice (1976) which was designed for unconfined aquifers and accounts for aquifer thickness and partial penetration of the aquifer by the well. The data were also analyzed by the Hvorslev method (1951) which, although derived for confined aquifers, can be used in unconfined aquifers where the saturated thickness is much greater than the drawdown in the well. The Hvorslev method, however, does not account

Table 1. Mean Hydraulic Conductivities in Upland and Shoreline Regimes

Method	*n*	*Upland mean (sd) (m/d)*	*n*	*Shoreline mean (sd) (m/d)*	*Upland/ shoreline Ratio*	*References*
Tidal						
W3-W1	28	86 (22)	30	41 (12)*	2.1	Ferris, 1963
W33-W30	16	126 (26)	16	98 (23)*	1.3	
Tracer						
	9	109 (10)				Smith et al., 1991
Slug						
	19	84 (21)	4	37 (12)	2.3	Hvorslev, 1951
	19	71 (18)	4	30 (10)	2.4	Bouwer and Rice, 1976
Permeameter						
All	25	29 (14)	34	18 (10)	1.6	Heath and Trainer, 1981
Piston up	14	33 (14)		**		
Vibe-up	11	24 (12)	34	18 (10)	1.3	
Grain size						
	37	104 (64)	40	68 (31)	1.5	Hazen, 1893
	37	86 (73)	39	55 (28)	1.6	Krumbein and Monk, 1942

*These wells do not represent true shoreline conditions.
**There were no piston cores taken in the shoreline.

for aquifer thickness or partial penetration by the test well. Well and aquifer geometries for the tests were appropriate for both the Bouwer and Hvorslev analyses.

Results: Slug tests by the Bouwer and Rice method yielded a mean K of 71 (18) m/d in the upland sites (Table 1), and 30 (10) m/d at the shoreline sites. In both regimes, when the same recovery data were analyzed using the Hvorslev method, mean K was significantly higher (18-23%), 84 (21) m/d in the upland and 37 (12) m/d in the shoreline (paired t-test $p < 0.05$). The higher values are probably due to the lack of a correction for partial penetration of the aquifer by the well in the Hvorslev method. Contrasts between the two regimes of 2.4 and 2.3 for the Bouwer and Hvorslev methods respectively (Table 1), were significant (independent t-test, $p < 0.05$).

Permeameter Tests: Sediment cores were collected in the field by both piston and vibracorers. Piston cores were collected from the water table downward, using either thin-wall aluminum (5 cm I.D.) or polycarbonate (6.5 cm I.D.) tubes (Munch and Killey, 1985). Vibracores were collected using aluminum tubes (7.62 cm I.D.) with stainless steel core catchers to improve core retention. Vibration was conducted at minimum levels and for the shortest time necessary to drive the tubing to refusal, usually 1 to 2.5 m. Shoreline cores were obtained from sand and gravel shoreface areas where ground-water discharge had been observed (Figure 1). Areas of marsh and peat, which generally impede the vertical movement of ground water (Redfield, 1959) were not sampled.

Upon return to the laboratory, cores were placed in a constant-head permeameter apparatus (Heath and Trainer, 1981; Klute and Dirksen, 1986). Piston core tubes were run on permeameters using their full length, usually 50 cm, while vibracore tubes were cut to lengths of 25 cm. To prevent the formation of gas bubbles which can reduce flow in the core tube, inflow water was heated (49° C), equilibrated with the atmosphere, collected in a closed tubing coil and cooled in a constant-temperature bath to $\approx 20°$C, before entering the permeameter. Hydraulic head was measured to 1 mm with manometers. Discharge volume was measured by collecting the outflow in a graduated cylinder. Cores were run at several hydraulic heads (0.016-0.136) which were comparable to field conditions of vertical flow at some of the shoreline sites and the results averaged. Hydraulic conductivity was calculated as follows (Fetter, 1988):

$$K = \frac{CVL}{At\Delta h} \tag{6}$$

where C = correction factor for temperature effects upon density and viscosity of the water (dimensionless); V = volume of discharge (cm^3); L = length of core sediment (cm); A = cross-sectional area of the core (cm^2); t = time interval of discharge (s); Δh = change in hyduaulic head (cm). For results from laboratory methods to be comparable to field measurements we corrected both permeameter and GSA data to reflect the approximate local ground-water temperature of 10° C.

Results: Permeameters gave the lowest estimates of both upland and shoreline K of any of the methods, although the range of ratios of upland to shoreline K, 1.3 to 1.6 were similar to other methods (Table 1). Upland sites had significantly higher K than shoreline sites, 29 (14) m/d and 18 (10) m/d, respectively ($p < 0.05$, independent t-test; Table 1). The high standard deviation of the shoreline cores reflects the occurrence of organic matter in some of the samples which lowered flow rates; however, cores with peat layers were excluded. We had concerns that use of vibracores would bias the permeameter results. K measurements from upland piston and upland vibracores were not significantly different, 33 and 24 m/d, respectively. Permeameter results from both vibracores and piston cores were more variable than results from the in situ methods (Table 1), and as a result upland and shoreline permeameter K from vibracores were not significantly different. It appears that K as determined by permeameter may be lowered by core collection by vibracoring.

Grain-Size Analysis: Beginning with Darcy, investigators have examined unconsolidated sands to identify a physical characteristic which accurately predicts in situ hydraulic conductivity. Empirical expressions relating statistical parameters of grain-size distribution in sands to permeameter experiments using presieved, uniform, unstratified sediment columns have been developed by Hazen (1893), Krumbein and Monk (1942), and Masch and Denny (1966). We compared K calculated using the expression of Hazen with that of Krumbein and Monk. According to Hazen (1893),

$$K = a(D_{10})^2 \tag{7}$$

where K is expressed in m/d; a = 1000 at 10°C, or the approximate ground-water temperature, when D_{10} is expressed in mm; and D_{10} = the 10th percentile of grain size (10% of the sample is finer by weight). The equation of Krumbein and Monk (1942):

$$k = 760(GM_d)^2 e^{-1.31\sigma} \tag{8}$$

where k = intrinsic permeability (mm^2); GM_d = the geometric mean grain diameter (mm); and σ = the standard deviation in phi units ($-\log_2$ [dia. in mm]). To convert intrinsic permeability, k, to hydraulic conductivity K, k was multiplied by the density of the water at 10°C, by the gravitational constant, and the result divided by the dynamic viscosity at 10°C.

Grain-size distributions were determined on subsamples of the cores following permeameter tests. For vibracores, the complete 25 cm length was homogenized and then split several times to obtain a workable sample size ($\approx$ 200 g). Piston cores were subsampled vertically during extrusion from their core tubes. Grain-size distributions were determined either by mechanical sieving or by settling tube analysis (Schlee, 1966). Results were expressed in terms of cumulative weight percent (Folk, 1974), and used to estimate K from equations (7) and (8).

Results: At the upland sites, mean K calculated by Hazen [equation (7)] and Krumbein [equation (8)], were 104 (64) and 86 (73) m/d, respectively, and at the shoreline sites were 68 (31) and 55 (28) m/d, respectively (Table 1). K determined by both methods was significantly higher in the upland versus the shoreline sites (independent t-test, p <

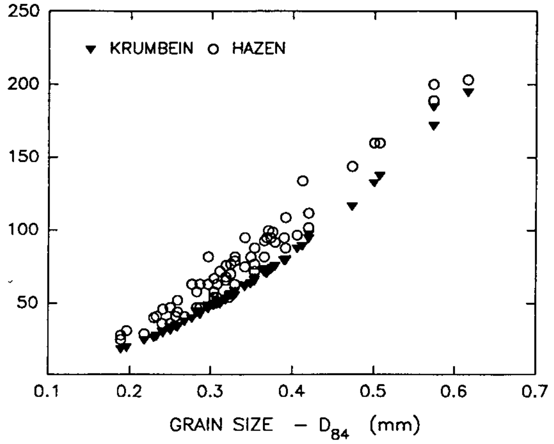

Fig. 4. Hydraulic conductivities of upland piston and vibracore sediment samples (n = 67), calculated using the Krumbein and Monk and Hazen methods, arranged by effective grain size, D_{84}. Krumbein and Monk values are solid inverted triangles, and Hazen values are open circles. K calculated for the same sample is dependent upon the method chosen and the method biases the results in a systematic way. The smooth curve of the Krumbein data as opposed to the scatter of the Hazen data is due to the use of D_{84} in the Krumbein graphic standard deviation δ, $(D_{84} - D_{50})$ for the computation of K.

0.05), but Krumbein yielded consistently lower values in both regimes (paired t-test; $p < 0.05$). The different K values from the same sample is likely the result of the use of different parameters, GM_d and D_{10}, as well as the use of standard deviation in equation (8) (Figure 4).

The significant differences in shoreline versus upland K from both GSA methods supports the suggestion that vibracoring affects K as measured by permeameter. The GSA analyses were from sediments from the permeameter vibracores which showed no significant difference between shoreline and upland K. Therefore, some of the reduction in permeameter K in vibracores is most likely due to the effect of vibrational disturbance on natural sediment packing, rather than grain-size differences.

Discussion

Upland-Shoreline Contrasts: There was consistently higher K measured by all methods in the upland as compared with shoreline settings (Table 1; Figure 5). The highest ratio of upland to shoreline K was for the Bouwer and Rice slug tests, 2.4, and the lowest, 1.3, was for the tidal well pair W-30/33 and vibracore permeameters (Table 1). Several conditions found at the ground-water-marine interface can contribute to lower shoreline K. Organic material occurs as dispersed detritus within the aquifer matrix and as layers of peat. In three sample shoreline cores taken from high ground-water discharge areas near the test site, dispersed organic material measured as loss on ignition ranged from 2% of dry weight near the surface to 0.5% at 0.4 m depth (Figure 6). For comparison, values for loss on ignition for samples of aquifer material taken from the upland aquifer (n = 10) were less than 0.25%. While these amounts are relatively small, they are several times less dense than the inorganic sediment fraction and occur in small particle sizes which occupy pore spaces in the aquifer matrix. Where peat occurred as buried discontinuous layers, observed ground-water discharge was greatly reduced, because peat typically has very low permeability (Redfield, 1959). The effects of buried peat on shoreline ground-water seepage has been noted by Bokuniewizc (1980). Around the shore of Little Pond, we obtained many vertical field profiles of interstitial water taken with a small diameter well-point sampler. In some profiles buried peat was manifested as a zone from which samples could not be drawn. Buried peat layers were also found in six shoreline vibracores, and we interpret them to be evidence of submerged shore environments buried as the result of transgressive Holocene sea level rise.

Shoreline sediments are affected by postdepositional processes which affect K. Ferrous iron precipitates where anoxic ground water discharges to oxygenated shore environments (Chambers and Odum, 1990) and thus may also act to decrease effective porosity. Ground water discharging along the shoreline near the test site was anoxic (< 2 mg/l vs 10 mg/l O_2 in overlying pond waters) and FeO(OH)

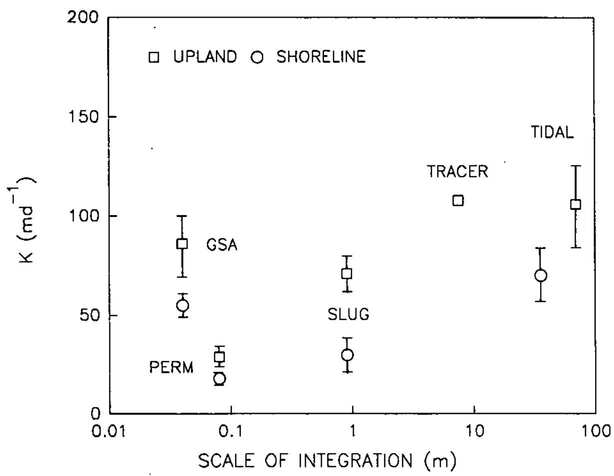

Fig. 5. Comparison of K determined by five methods in upland (squares) and shoreline (circles) aquifer settings versus the approximate distance integrated by each method (GSA data is from the Krumbein method and tracer results are only for the upland). GSA, permeameter, and slug tests all yielded significantly higher hydraulic conductivity in the upland. For purposes of graphical clarity values are shown as mean ± standard error [s.e. = s.d./($\sqrt{n-1}$)].

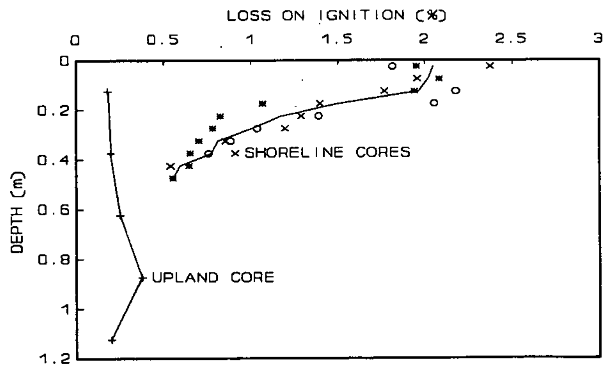

Fig. 6. The relative amount of dispersed organic material as percent sediment dry weight in three shoreline piston cores and one upland piston core. Solid line is the mean of shoreline measurements. The shoreline cores were collected from areas of active ground-water discharge (i.e., there were no buried peat layers). Loss on ignition in shallow and deep upland cores was typically between 0.15 and 0.20% by weight.

precipitates were observed in this and other shoreline areas of high ground-water discharge. The concentration of FeO(OH) in three cores taken from these areas was less than 0.2% by dry weight of sediment and averaged only 0.06% by dry weight of sediment ($n = 27$). Though the concentrations of FeO(OH) are relatively small, similar to organic matter under in situ flow conditions, their role may be significant because of their relatively large surface areas (Crosby et al., 1983). Given the measured dry weights it appears that the potential role of organic material in lowering K is likely much greater than that of ferric oxyhydroxide deposition.

Systematically lower permeameter measurements of K in the upland may be due to the fact that permeameter cores are usually obtained in a vertical orientation. The fluvial genesis of outwash sediment fabric typically results in anisotropies (Freeze and Cherry, 1979). Ratios horizontal to vertical K in outwash sands from a tracer test at a site 9 km north of Little Pond were estimated to range from 10:1 to 2:1 (LeBlanc et al., 1991), and have been supported by a comparison of vertically averaged permeameter results and borehole flow (horizontal K) measurements (Wolf et al., 1991). The latter measures yielded geometric mean K values of 105 m/d and 32 m/d (ratio 3.3) for borehole flow-meter and permeameters, respectively (Wolf et al., 1991). We found the same ratio for the mean of tidal, tracer, and Bouwer slug tests (95 m/d) to mean permeameter (29 m/d) measurements found at the upland sites around Little Pond. The contrast between permeameter and in situ flow measurements suggests that permeameter data should only be relied upon as a relative measurement of in situ K. On the other hand, ground-water flow often has vertical components at the shoreline which the vertically oriented permeameters may better approximate.

The two tidal wells, W-1 and W-30, located in the upland 31 and 42 m from the shore respectively, had lower K than the two more inland wells. The degree to which lower K along the adjacent shoreline influenced these two wells is difficult to assess since the resultant K from these wells includes both upland and shoreline components.

The slug and GSA methods had similar values for mean upland K, 71 and 86 m/d, respectively, while either of the two GSA shoreline results, 55 and 68 m/d, were twice the shoreline slug tests, 30 m/d. The similarity of the slug and GSA methods in the upland and the contrast in the shoreline GSA-slug results may be due in part to the insensitivity of the GSA method to factors (dispersed organic particles, peat, and FeO(OH) precipitates) which can lower in situ effective porosity (and thus lower slug results) without significantly altering the size distribution. For instance, organic particles expand when hydrated, lowering K, but given their low density, their existence may have a minor effect on the dry weight grain-size distributions used by the GSA methods.

Methods Evaluation: The mean hydraulic conductivity determined from all the methods spanned a range of about one-half an order of magnitude within each environment (Figure 5), and for some purposes any one of the five methods might give satisfactory results. However, for quantitative estimation of inputs of fresh-water or ground-water transported solutes using Darcian calculations, the range in K translates to a proportionately large range in estimated discharge. Clearly, where quantitative discharge estimates are sought, it is necessary to refine the approaches for determining K. However, it appears that the appropriate method is related not only to the scale of integration but also the setting (upland vs shoreline).

Of the five methods tested in the upland aquifer only the permeameter gave results which were widely different. Excluding the permeameter results the range of upland K is only twofold (71 to 126 m/d; Table 1). Of the remaining four methods, the larger scale of measurement recommends the tidal and tracer approaches. Although tidal data may be confounded where tidal signals arrive from multiple directions (e.g., peninsulas) or where the stratigraphy is complex, the economy and simplicity of tidal methods and their assessment of overall aquifer transmissivity (Ferris, 1963) support their general application. In Little Pond, the small and irregular tides required that records be collected over relatively long periods (lunar cycles) in order to acquire sufficient data for calculation of K. It is likely that use of tidal methods in coastal watersheds adjacent to embayments with larger tidal ranges (> 0.5 m) and use of deeper piezometers will improve results by decreasing the variability of measured delay times (Pandit et al., 1991).

The tracer test had a major advantage because ground-water flow was measured in situ using the natural hydraulic gradient. Results from the tracer test were the least variable of all methods, with good replication from separate sampling ports at the same multilevel sampler. K calculated from the tracer test is an average of the portion of aquifer through which the tracer cloud travels. Data from the most distant sampler therefore gives the most inclusive estimate of aquifer K. The mean K of the three horizontal segments (0-7 m) of 109 m/d was very close to the mean K for the last segment (4.5-7 m; 3 ports) of 106 m/d. This relatively small variation in measured K, when contrasted with the substantial heterogeneity over short distances in the aquifer, suggests that the scale of the tracer test (7 m) may have been sufficient to integrate the range of horizontal heterogeneity.

In the shoreline environment the choice of suitable methods is more limited. Permeameters and slug tests appeared to be the best methods available for shoreline K measurement. The good replication of upland slug test K with upland tracer and tidal K along with relatively strong contrasts of slug K between the upland and shoreline regimes, suggests that slug test K is the best measure of K in the shoreline. In contrast to the upland, the vertical ground-water flow conditions in the shoreline zones of discharge made the vertical flow conditions of the permeameters more realistic and is likely the reason for the better results. Permeameters and slug tests have opposite conditions for best application since the slug tests can only be used in shoreline areas with little or no upward flow. GSA methods yielded shoreline K's about two times that of the other two small scale methods, most likely due to their insensitivity to additions of low density fine-grained materials, like organic matter.

The tidal damping and tracer test had disadvantages which limited their use in a shoreline setting. The tidal damping method cannot be used in an area of vertical flow, a condition typical in shoreline ground-water discharge zones, and in areas of large tidal range the range of water-table fluctuation near the shore may be too large relative to the aquifer thickness. Likewise, measurements of ground-water velocities and hydraulic gradients required for tracer tests are complicated by tidally induced fluctuations of the water table in the nearshore, and alternates to Cl^- as a tracer are necessary due to the high ambient Cl^- conditions of the marine environment.

Our results do not show that either Hazen or Krumbein GSA equations are better predictors of aquifer K, as the difference between them when compared to the range of K occurring in the upland aquifer is relatively small. However, the disparate results of the GSA equations (Figure 4) are significant in that they represent a technical "bias" and thus suggest that K obtained by GSA be verified by in situ techniques. GSA methods continue to find widespread application because disaggregated sediment samples obtained by hand augers or split-spoon samplers are often the only locally available source of data from an aquifer. For example, in a sandy Nigerian aquifer of large extent, the Hazen and Masch and Denny methods have been re-calibrated using pump test data and then applied to other areas with limited in situ data (Uma et al., 1989). Unfortunately, in outwash grain size changes can occur over short vertical or horizontal distances, and samples obtained for GSA may be a composite of two distinct layers. We found that small scale vertical changes in grain size were common, as in one piston core where GM_d and the resultant calculated Krumbein K varied from 0.212 mm (35 m/d) to 1.32 mm (133 m/d), over 23 cm of core length (Figure 7). Identifying and sampling many such layers greatly improve K determinations using GSA (Wolf et al., 1991).

The method which gave the best results in *both* upland and shoreline regimes was the slug test. Although K determined by a slug test is limited to a small radius (< 1 m) around the well and vertically to the length of the screen, these distances appeared to be sufficient to integrate some of the smaller scale heterogeneities in the aquifer. Evidence for this may be seen in the fact that Bouwer slug test K and

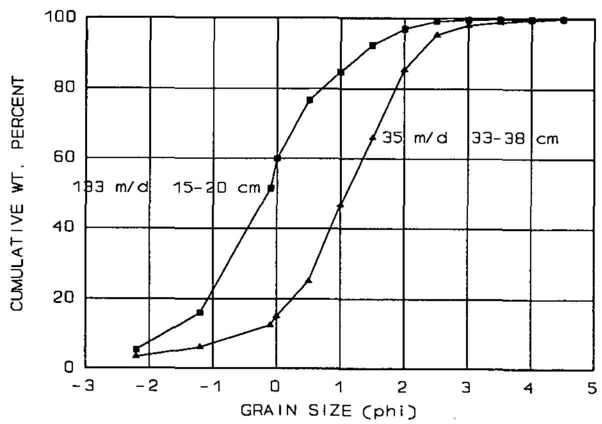

Fig. 7. Changes in grain-size distribution and thus hydraulic conductivity in a piston core (cumulative frequency; grain size decreases to the right). Hydraulic conductivity is shown by the number adjacent to each curve, determined by the GSA-Krumbein method. Estimated hydraulic conductivity changes nearly four-fold over 23 cm of core.

Krumbein GSA K in the upland are similar, 71 and 86 m/d, yet the standard deviation of the slug test is relatively low (18 m/d) compared to that of Krumbein, 73 m/d. Because of the rapid recovery in highly permeable sediments, slug tests offer the advantage that they can be repeated quickly in the same well and a large number of wells can be tested in a short time.

Conclusions

Our investigation of hydraulic conductivity in a coastal aquifer indicates that for estimation of discharge, tidal transmissivity and tracer test methods are preferable to point-specific field (slug tests) and laboratory methods (permeameter and grain-size analysis) for two reasons; they yield integrated K values for larger volumes of the aquifer, and they measure horizontal K. In order to characterize K over the spatial scale required for a Darcian calculation of ground-water discharge to coastal waters, the two larger scale in situ methods also appear to be more efficient approaches than the requisite large number of small scale measurements. In the upland the similarity of slug test and GSA results to those of the larger scale methods suggests that they are acceptable alternatives for K measurement, while the slug test appeared to be reliable in the shoreline as well. In the upland environment, permeameter results were substantially lower, apparently reflecting a threefold horizontal to vertical K differential.

The aquifer setting significantly affected measured K. The slug, permeameter, and GSA methods indicated lower K in the shoreline setting, with the largest differences in K determined by slug test. Significant differences in the character of the aquifer matrix exist between the upland aquifer and the shoreline discharge zone which may be caused by increased dispersed organic matter and peat layers. While some of these matrix alterations have been previously described, their hydrologic significance is often overlooked. Permeameter results were the lowest of all methods in both environments in part due to handling artifacts, but both permeameters and slug tests appeared to reflect the in situ conditions which lowered shoreline K.

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