2 HYDROGEOLOGICAL METHODS FOR ESTIMATION OF SPATIAL VARIATIONS IN HYDRAULIC CONDUCTIVITY

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2.1 Introduction

2.1.1 CHAPTER OVERVIEW

Virtually every hydrogeologic investigation requires an estimate of hydraulic conductivity (K), the parameter used to characterize the ease with which water flows in the subsurface. For water-supply investigations, a single estimate of K averaged over a relatively large volume of an aquifer will usually suffice. However, for water-quality investigations, such an estimate is often of limited value. A large body of work has demonstrated that spatial variations in K play an important role in controlling solute movement in saturated flow systems (e.g., Sudicky and Huyakorn, 1991; Zheng and Gorelick, 2003). Numerous studies have shown that information about such variations is required to obtain reliable predictions of contaminant transport and to design effective remediation systems. Varieties of methods have been used in efforts to acquire this information. The primary purpose of this chapter is to describe these methods and assess the quality of the information that they can provide. Later chapters will discuss how geophysics can augment the information obtained with these approaches.

In this chapter, three classes of methods will be discussed. The first class, designated here as "traditional approaches," consists of methods that have been used for a number of decades to acquire information about hydraulic conductivity for water-supply investigations. The second class, designated as "current generation," consists of approaches that have been developed in the last decade or two for the specific purpose of acquiring information about spatial variations in K. The third class, designated as "next generation," consists of methods that are currently in various stages of development. For each method, the underlying principles of the approach will be described, followed by a brief discussion of its major advantages and limitations. Each method will be illustrated with data collected at an extensively studied field site. The use of data from the same site facilitates the discussion of the relative advantages of the various approaches, as well as the type of information each can provide. Although the focus of this chapter will be on methods for investigations in shallow unconsolidated aquifers, many of these methods are also of value for investigations in consolidated materials and in units that would not be classified as aquifers. Techniques primarily used in other hydrogeologic settings are briefly discussed at the end of this chapter.

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2.1.2 FIELD SITE

The methods described in this chapter were evaluated at a research site of the Kansas Geological Survey. This site, the Geohydrologic Experimental and Monitoring Site (GEMS), is located in the floodplain of the Kansas River just north of Lawrence, Kansas, in the central portion of the United States (Figure 2.1). GEMS has been the site of extensive research on flow and transport in heterogeneous formations (e.g., McElwee et al., 1991; Butler et al., 1998a,b, 1999a,b; 2002; Bohling, 1999; Schulmeister et al., 2003).

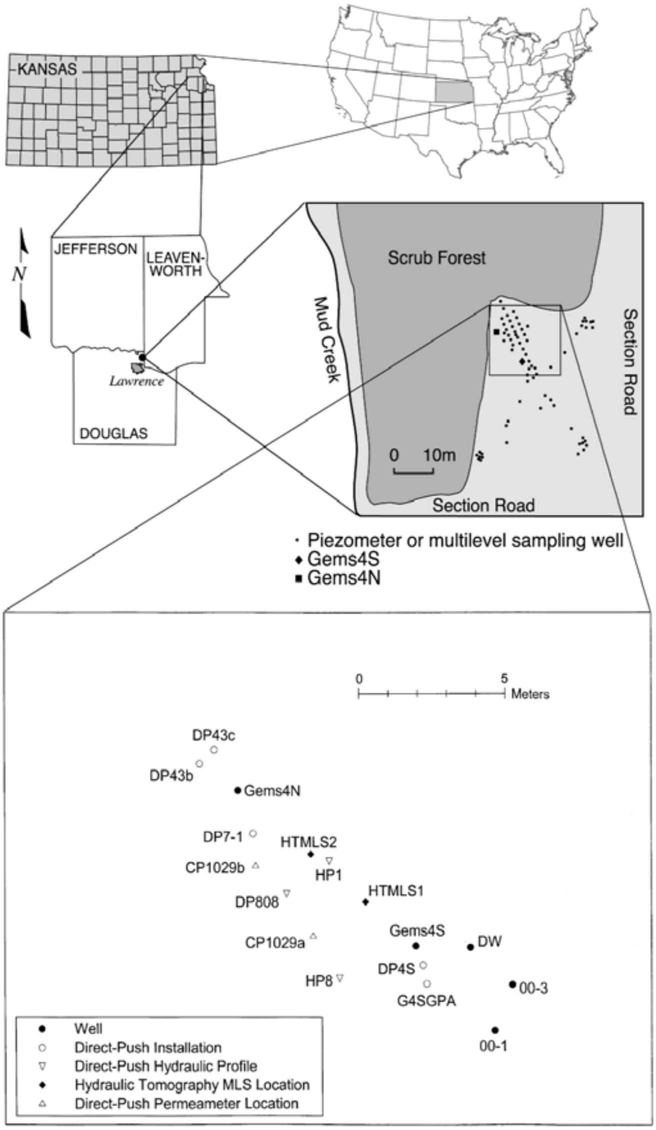


Figure 2.1. Site location map for GEMS(only locations referred to in text are shown in inset)

This work enables the techniques discussed here to be evaluated in a relatively controlled field setting. The shallow subsurface at GEMS consists of 22 meters of unconsolidated Holocene sediments of the Kansas River alluvium that overlie and are adjacent to materials of Pennsylvanian and late Pleistocene age, respectively. Figure 2.2 displays a cross-sectional view of the shallow subsurface with electrical conductivity logging data obtained using a direct-push unit (Butler et al., 1999b), and a geologic interpretation from core and logging data. As shown in that figure, the heterogeneous alluvial facies deposits at GEMS consists of 11.5 m of primarily clay and silt overlying 10.7 m of sand and gravel. The methods described in this chapter were applied in the sand and gravel interval, which is hydraulically confined by the overlying clay and silt.

The subarea of GEMS used in this work is depicted in the inset of Figure 2.1. This inset displays the locations of conventional wells (Gems4N, Gems4S, DW, 00-1, and 00-3), multilevel sampling wells (HTMLS1 and HTMLS2), and various direct-push activities.

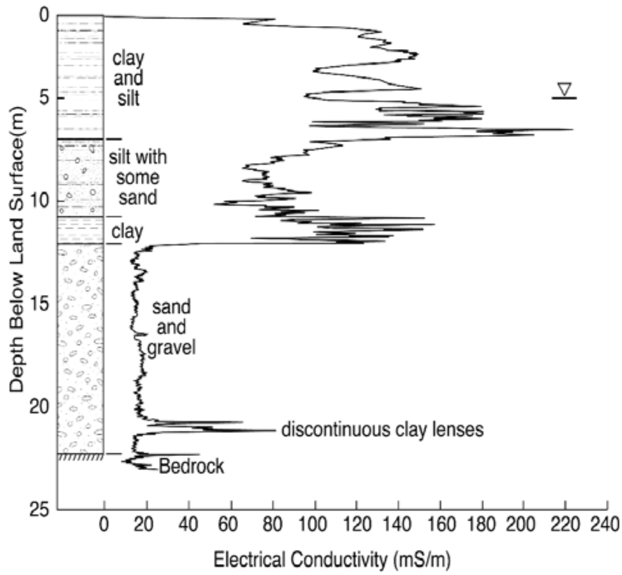


Figure 2.2. Generalized GEMS stratigraphy with electrical conductivity log from G4SGPA (after Butler et al., 1999b)

The conventional wells were all constructed out of PVC and installed with hollow-stem augers, with 0.168 m outer diameter (OD) flights for Wells 00-1 and 00-3 (both wells—0.051 m inner diameter [ID]) and 0.286 m OD flights for other wells (Gems4N and Gems4S–0.102 m ID, DW–0.127 m ID). A natural filter pack was used in all cases, and all wells were grouted across the overlying silt and clay interval. The multilevel sampling wells (MLS) were constructed out of extruded PVC (seven-channel) pipe and installed with a direct-push unit (OD of direct-push pipe–0.083 m; OD of MLS–0.041 m). Various configurations were used for the direct-push activities, as will be discussed in Sections 2.3 and 2.4.

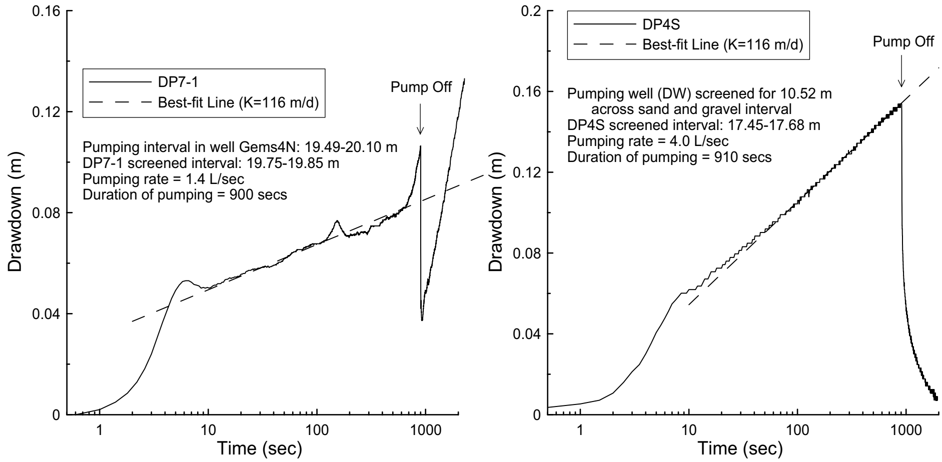
2.2 Traditional Approaches

A variety of methods have been used to obtain information about hydraulic conductivity for water-supply investigations. Many of these methods have also been utilized to assess spatial variations in K. Those methods are described in this section.

2.2.1 PUMPING TESTS

The constant-rate pumping test is the most commonly used method to obtain information about the transmissive nature of an aquifer for water-supply investigations. In this approach, a central well is pumped at a near-constant rate, while changes in water level are measured at that and nearby wells. The changes in water level, termed drawdown, are analyzed using various models of the well-formation configuration (Streltsova, 1988; Kruseman and de Ridder, 1990; Batu, 1998). Several investigations (Butler and Liu, 1993; Meier et al., 1998; Sánchez-Vila et al., 1999) have shown that a pumping test will yield a hydraulic conductivity estimate that represents an average of K over a relatively large volume of an aquifer. Thus, little information can be gleaned about variations in K on the scales of interest for solute-transport investigations.

A series of short-term pumping tests performed at GEMS can be used to illustrate the limitations of this approach for assessment of spatial variations in K. Figure 2.3a depicts the drawdown at direct-push installation DP4S produced by a constant rate of pumping at Well DW. The Cooper-Jacob method (Cooper and Jacob, 1946) is used to fit a straight line to the latter portions of the drawdown record. The resulting hydraulic conductivity estimate of 116 m/d is a reasonable value for the average K of a sand and gravel sequence (Freeze and Cherry, 1979). Figure 2.3b depicts the drawdown at direct-push installation DP7-1 produced by a constant rate of pumping at Well Gems4N. Although a backpressure adjustment between 100 and 200 seconds and the commencement of pumping at a nearby well at approximately 600 seconds complicate the analysis, the Cooper-Jacob method can still be used to fit a straight line to an extensive portion of the drawdown record. The resulting K estimate is again 116 m/d. The similarity in the K estimates from the two tests, which is in keeping with the findings of the previously cited studies, demonstrates that little information about variations in hydraulic conductivity can be obtained using K estimates from analyses of conventional pumping tests performed in nearby wells. Although K estimates will vary little, estimates of the specific storage parameter (S_s) can vary considerably. For example, the S_s estimates, obtained from analyses of drawdown at DP4S and DP7-1, differ by a factor of 36 (3.81×10^{-5} m^{-1} and 1.06×10^{-6} m^{-1}, respectively). This variation in S_s is produced by a number of factors, including variations in the K of the material between the pumping well and the observation point (Butler, 1990; Schad and Teutsch, 1994; Sánchez-Vila et al., 1999). However, given the uncertainty about the factors contributing to the variation in S_s, it is extremely difficult to extract information about variations in K from it.



(a) (b)

Figure 2.3. Drawdown versus logarithm of time plots for two short-term pumping tests (after Butler et al., 2002): (a) 3/16/99 test at Well DW; (b) 8/13/99 test at Well Gems4N (all depths in this and following figures are with respect to the top of casing at Well Gems4S)

2.2.2 SLUG TESTS

The slug test is a commonly used method for acquiring information about K for both water-supply and water-quality investigations. This approach, which is quite simple in practice, consists of measuring the recovery of head in a well after a near-instantaneous change in head at that well. Head data are analyzed using various models of the well-formation configuration (see Butler [1998] for a detailed description of field and analysis procedures). The slug test provides a K estimate that is heavily weighted towards the properties of the material in the immediate vicinity of the screened interval (Beckie and Harvey, 2002). Thus, at a site with an extensive network of wells, the slug test can be a valuable tool for describing spatial variations in K (Yeh et al., 1995). However, considerable care must be taken in all stages of the test process to obtain reasonable K estimates (Butler et al., 1996; Butler, 1998). In particular, the quality of the K estimates is heavily dependent on the effectiveness of well-development activities (Butler and Healey, 1998; Butler et al., 2002). Failure to give appropriate attention to well-development procedures will often result in K estimates that bear little resemblance to reality.

The potential and pitfalls of this approach can be illustrated using a slug test performed at GEMS. Figure 2.4 is a normalized head (deviation from static head normalized by the initial head change) versus time plot from a test performed in Well 00-1. The data were analyzed using a high-K extension of the Hvorslev (1951) model that incorporates the inertial mechanisms that give rise to the oscillatory head data (Butler, 1998; Butler et al., 2003). The resulting K estimate of 224 m/d is close to twice the large volumetric average obtained from the pumping tests, demonstrating that slug tests can be used to assess the K of discrete zones that may act as preferential pathways for or barriers to solute movement. However, as stated above, appropriate attention must be given to all phases of the test procedure to obtain reliable K estimates. In highly permeable intervals, such as the test interval at Well 00-1, a number of issues must be considered.

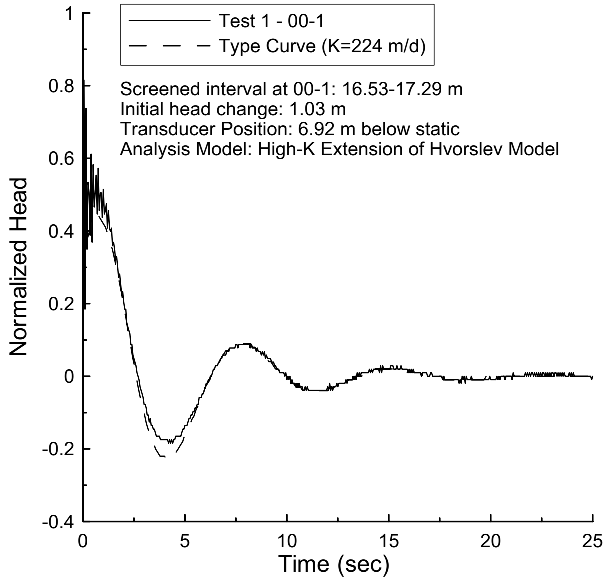


Figure 2.4. Normalized head versus time plot for 3/18/94 slug test #1 at Well 00-1 (data from Butler, 1998)

For example, the pressure transducer at Well 00-1 was located 6.92 m below static, so, as shown by McElwee (2001) and Zurbuchen et al. (2002), water-column acceleration effects on transducer readings must be considered. Furthermore, as shown by Butler et al. (2003), analysis using a model that neglects inertial mechanisms, such as the conventional forms of the Hvorslev (1951) and Bouwer and Rice (1976) models, can result in a significant overestimation of K. Butler et al. (2003) discuss these and related issues, and provide a series of guidelines for the performance and analysis of slug tests in highly permeable intervals. More general guidelines are provided in Butler et al. (1996) and Butler (1998). Adherence to these guidelines is necessary to obtain reliable K estimates. Note that the deviation between the test data and the best-fit type curve in the vicinity of the trough at four seconds on Figure 2.4 is likely a product of the nonlinear responses discussed by Butler (1998) and McElwee and Zenner (1998).

2.2.3 LABORATORY ANALYSES OF CORE SAMPLES

Laboratory analyses of samples collected during drilling is a common method for acquiring information about the properties of a formation. Estimates of the saturated hydraulic conductivity are obtained using various permeameter methods or relationships based on particle-size analyses. Permeameter methods involve running water through a core under either a constant or variable hydraulic gradient. The constant-gradient (constant-head) permeameter is primarily used for materials of moderate or high K, while the variable-gradient (falling-head) approach is primarily used for materials of low K. In either case, considerable care must be taken in all phases of the experimental procedures. Entrapped air, mobilization and redeposition of fine fractions, non-Darcian flow, and use of a permeant fluid at different temperature, pressure, and biochemical conditions than the natural *in situ* fluid will often lead to artificially low estimates of K (Klute and Dirksen, 1986). In most cases, water is run parallel to the vertical axis of the core, so the K estimate is for the vertical component of hydraulic conductivity. Permeameter experiments can be performed on either the original sample, if relatively undisturbed, or on a repacked sample. Hydraulic conductivity estimates obtained from the original samples tend to be lower than those obtained from the repacked cores, a difference that is commonly attributed to the greater structural control in the original samples.

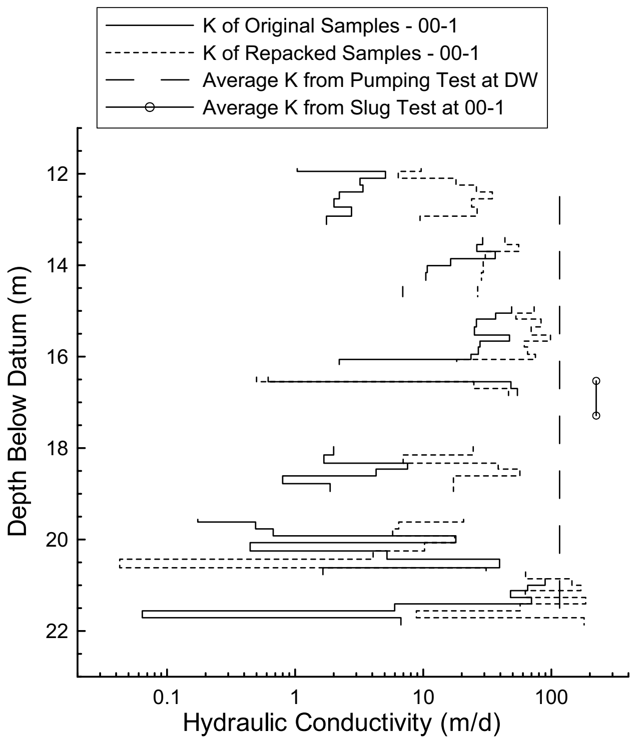


Figure 2.5. Hydraulic conductivity versus depth plots at Well 00-1 (core data from Butler and McElwee, 1996)

Results of analyses of core samples from Well 00-1 can be used to illustrate the limitations of permeameter-based approaches. Figure 2.5 is a plot of K estimates obtained from permeameter analyses of the original and repacked core samples. In addition, the average K obtained from the pumping test at Well DW and the K from the slug test at Well 00-1 are shown for reference. The averages for the permeameter analyses of the original and repacked cores are 16% and 39%, respectively, of the K determined from the pumping test. Every K estimate obtained from the original cores is below the average value from the pumping test, a situation that has often been reported by others (e.g., Table 1 of Rovey, 1998). The increase in K by a factor of 2.4 between the original and repacked cores is undoubtedly caused by the repacking process removing thin layers of low-K material that exert a strong influence on the original core estimates. The underprediction of K in the repacked samples with respect to the pumping-test estimate is likely caused by a combination of imperfect laboratory procedures, heterogeneity, and incomplete sample recovery in zones of high K. This latter possibility is reflected in the incomplete recovery in the zone opposite the screened interval of the well, which the slug test results indicate is of a relatively high K. Recovery of relatively intact core samples is a difficult task in highly permeable intervals. Although specialized devices have been developed for this purpose (e.g., Zapico et al., 1987; McElwee et al., 1991; Stienstra and van Deen, 1994), complete recovery is rarely possible. For example, the recovery at Well 00-1 was 72%.

The second approach commonly used for estimating K from core samples is based on relationships between hydraulic conductivity and various physical properties of the samples. A large number of empirical and theoretical relationships have been developed for estimation of K from particle-size statistics, which are assumed to be a reflection of the pore-size distribution. The most common empirical relationships are based on some measure of effective grain size. For example, the relationship developed by Hazen (Freeze and Cherry, 1979) uses d_{10}, the particle diameter at which 10% of the grains by weight are less than this diameter and 90% are greater ($K = Cd_{10}^2$ where C is a coefficient depending on grain size and sorting). Theoretical relationships have been developed from the Navier-Stokes equations and models of the porous medium. For example, the relationship developed by Kozeny-Carman (Bear, 1972) uses d_m, the geometric mean grain size calculated from $\sqrt{d_{84}d_{16}}$, and porosity (n) to estimate hydraulic conductivity ($K = [n^3/(1-n)^2][d_m^2/180]$ for K in m/d and d_m in mm). Although procedures for calculating particle-size distributions and porosity are well developed (Gee and Bauder, 1986; Danielson and Sutherland, 1986), determining the appropriateness of a given relationship for a particular site can be difficult.

Results of the laboratory analyses of the core samples from Well 00-1 help illustrate the uncertainty inherent in these empirical and theoretical relationships. Figure 2.6a is a plot of the permeameter results for the repacked samples, the K estimates determined from the Hazen equation (C = 864 for a well-sorted sand with K in m/d and d in mm), and the K estimates from the pumping and slug tests. The average K from the Hazen analysis is 147 m/d, 27% larger than that from the pumping test, a difference that could be attributed to heterogeneity, the vastly different scales of the two approaches, and uncertainty regarding the appropriate value for the coefficient C. The Hazen K estimates approach the slug-test K for the screened interval at Well 00-1, but incomplete recovery prevents a full comparison. Although the Hazen K values appear reasonable, use of additional relationships casts doubt on these values. Figure 2.6b compares results obtained using the Hazen and Kozeny-Carman equations. The average K value from the Kozeny-Carman equation is 110 m/d, which is within 6% of that obtained from the pumping test. Use of other relationships (not shown here) reveals a continuing lack of consistency between methods. For example, the empirical relationship of Bedinger (1961) produces an average K of 111 m/d, but in certain intervals, values differ significantly from those obtained using the Kozeny-Carman equation.

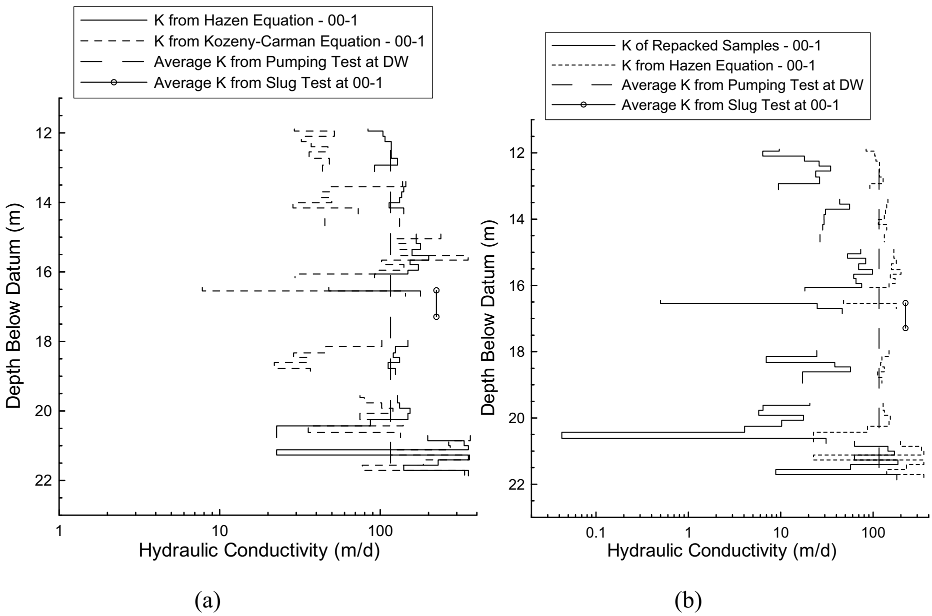


Figure 2.6. Plots of hydraulic conductivity versus depth (Butler and McElwee, 1996; previously unpublished KGS data): (a) Estimates from permeameter and particle-size analyses; (b) Estimates from particle-size analyses

2.2.4 GEOPHYSICAL LOGGING

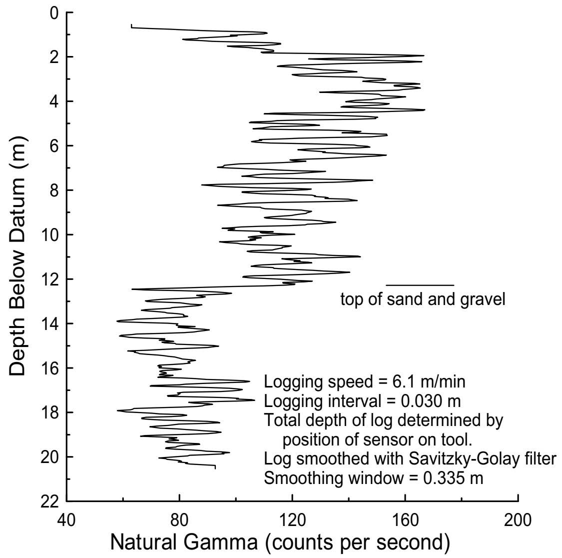
Geophysical logging is a common means of acquiring information about relative variations in hydraulic conductivity (Keys, 1990; also, see Chapter 10 of this volume). In unconsolidated formations, variations in hydraulic conductivity are often produced by variations in clay content, which can be detected using a variety of logging tools. However, the relatively large averaging volume of wellbore logging tools used in unconsolidated formations limits their utility. Thus, although large variations in clay content between a clay and sand unit are readily detectable, characterization of smaller variations within a single unit is rarely possible. Recently, geophysical logging tools have been incorporated into direct-push equipment (Christy et al., 1994). The averaging volume of these direct-push logging tools can be significantly smaller than conventional wellbore logging tools, enabling valuable information to be obtained about small-scale variations in clay content.

The two most common types of geophysical logs used to characterize variations in clay content in unconsolidated formations are natural gamma and electrical conductivity. Natural-gamma logs provide a record of natural-gamma radiation versus depth. This radiation is quantified by counting the gamma particles passing through a scintillation crystal in a certain time interval. A high natural-gamma reading is generally associated with clay-rich intervals, while a low reading is generally associated with sands and gravels (Keys, 1990, 1997). The stochastic nature of the radiation process, coupled with the speed at which logs are normally run, introduces a great deal of noise into the natural-gamma data. Although various filtering approaches (e.g., Savitzky-Golay algorithm (Press et al., 1992)) can reduce this noise, they often do so at the cost of a loss in vertical resolution. The spatial averaging produced by filtering can largely be avoided by running the logs at very low speeds (<1 m/min). That approach, however, is rarely implemented in practice.

Electrical-conductivity or resistivity logs are commonly used to detect variations in clay content, fluid-filled porosity, and water chemistry. When variations in groundwater chemistry and porosity are small, changes in electrical conductivity are primarily a function of changes in the clay-sized fraction. Although conventional logging tools have relatively large averaging volumes, electrical-conductivity sensors incorporated in direct-push equipment can detect layers as small as 0.025 m in thickness (Schulmeister et al., 2003). Unlike natural-gamma logs, electrical-conductivity logs are not affected by logging speed.

A series of wellbore and direct-push geophysical logs have been performed at GEMS to assess variations in clay content in the unconsolidated sequence. Figure 2.7 (top) is a plot of natural-gamma radiation versus depth at Well 00-3. The high-frequency fluctuations observed on Figure 2.7 (top) are most likely noise, as discussed above, and not related to variations in clay content. Although the log in Figure 2.7 (top) was run at a standard logging speed, natural-gamma logs at GEMS have been run at speeds as low as 0.15 m/min. The high-frequency fluctuations are dramatically reduced at the very low speeds, indicating that the fluctuations are primarily artifacts of the radiation process and logging speed. Figure 2.7 (bottom) supports this interpretation, because the natural-gamma fluctuations at Well 00-3 are not strongly correlated with variations in the fine fraction at Well 00-1 (2.3 m from Well 00-3). In addition, the high-resolution electrical-conductivity log at G4SGPA (2.9 m from 00-3) shows little variation except near the top and bottom of the sand and gravel interval. Comparison with Figure 2.6 indicates that K variations in the sand and gravel are, for the most part, not a product of variations in the fine fraction. In such cases, natural-gamma and electrical-conductivity logs may be of little use for characterizing relative variations in K.

Hydraulic conductivity variations can also be the product of variations in porosity, which may be assessed using a variety of conventional geophysical logs. However, the interpretation of those logs in terms of K is difficult when the particle-size distribution and clay fraction vary as well (see Chapter 10 of this volume).



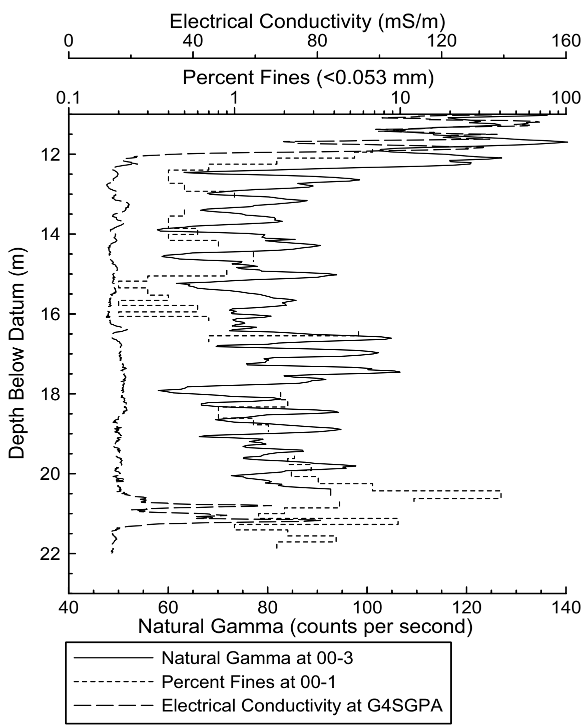


Figure 2.7. (top) Plot of natural gamma vs depth, Well 00-3; (bottom) Expanded view of the sand and gravel interval.

2.2.5 SUMMARY OF TRADITIONAL APPROACHES

The traditional approaches appear to have difficulty providing K estimates at the level of reliability and detail needed to characterize spatial variations in K for water-quality investigations. Pumping tests do produce reliable K estimates, but the estimates are large volumetric averages. Laboratory analyses can provide information at a very fine scale, but there are many questions about the reliability of the K estimates obtained with those analyses. Geophysical logs can provide valuable information about relative K variations produced by changes in clay content and, to a lesser extent, porosity, but are of limited value otherwise. Although the slug test has the most potential of the traditional approaches for detailed characterization of K variations, most sites do not have the extensive well network required for effective application of this approach. The use of slug tests for characterizing K variations is more fully explored in the following sections where modifications of the traditional approach are described.

2.3 The Current Generation

Casual scrutiny of sedimentary strata in outcrop reveals a pronounced anisotropy in the degree of continuity of most observable media characteristics. In general, strata tend to have significant continuity in the lateral direction (parallel to bedding), but very little in the vertical (perpendicular to bedding). Experimental studies in unconsolidated sedimentary sequences have found that hydraulic conductivity varies in a similar manner (e.g., Smith, 1981; Sudicky, 1986; Hess et al., 1992). Given this apparent large anisotropy in the continuity or correlation structure of hydraulic conductivity, much valuable information about K variations at a site can be gained through a limited number of profiles of hydraulic conductivity versus depth. In an effort to obtain such profiles, a variety of techniques have been developed to characterize the vertical variations in K along the screened interval of a well. These techniques, which have been increasingly applied in recent years, are reviewed in this section.

2.3.1 DIPOLE-FLOW TEST

The dipole-flow test (DFT), first described by Kabala (1993), is a single-well hydraulic test involving use of the three-packer tool shown in Figure 2.8 (left). A pump in the central pipe of the middle packer transfers water from the upper chamber to the lower, setting up a recirculatory pattern in the formation. Pressure transducers placed in the upper and lower chambers (transducers labeled UT and LT in Figure 2.8 (left), respectively) measure the head change in each chamber. A control transducer (CT in Figure 2.8 [left]) above the tool is used to detect short-circuiting through fittings in the upper packer or along a near-well disturbed zone. Zlotnik and Ledder (1996) and Zlotnik and Zurbuchen (1998) developed the theory, equipment, and field methodology for the steady-state form of the DFT. They found that the radial component of hydraulic conductivity (Kr) can be estimated from the total head change in the two chambers at steady state (Δh; Figure 2.8 [right]) using an approximate equation:

$$K_r = \frac{Q}{2\pi(\Delta h)\Delta} \ln\left[\frac{4a\,\Phi(\lambda)\Delta}{e\,r_w}\right] \quad (2.1)$$

where a = anisotropy ratio, $(K_r/K_z)^{0.5}$; K_z = vertical component of hydraulic conductivity; and $\Phi(\lambda)$ = dipole shape function, ranging from 0.5 to 1.0 depending on λ (L/Δ ratio). Zlotnik and Ledder (1996) show that Equation (2.1) is a reasonable approximation under conditions met in most field applications. Using this formula, the vertical variation in K_r can be estimated through a series of DFTs between which the tool is moved a short distance in the well. An estimate of the anisotropy ratio, which is rarely known, is required to calculate K_r. In most cases, however, an anisotropy ratio of one is assumed for lack of better information. Butler et al. (1998a) discuss a number of additional practical issues that must be considered for successful application of the DFT. Transient forms of the DFT have been proposed (Kabala, 1993), but have not proven successful in field applications (Hvilshøj et al., 2000). Such approaches have the most potential in units of moderate to low K, where the time to steady state is greater than the few to tens of seconds found in sand and gravel intervals.

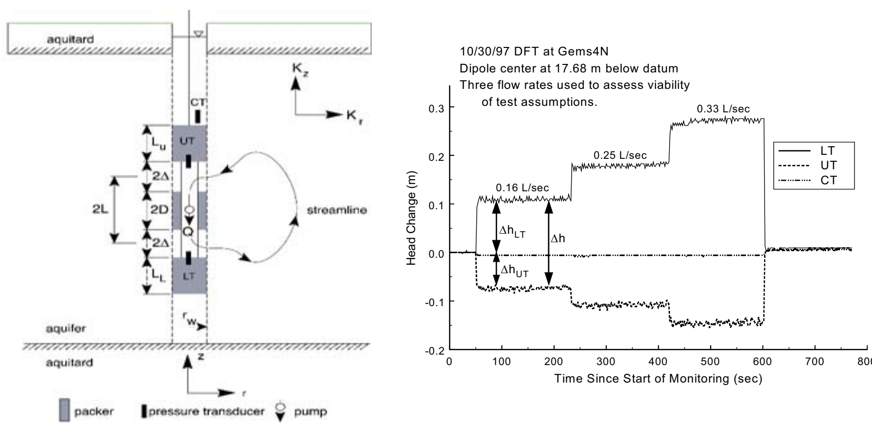


Figure 2.8. (left) Schematic of dipole-flow test; (right) Plot of head change versus time for a dipole-flow test at Gems4N (after Butler et al., 1998a)

A series of DFTs performed at Wells Gems4N and Gems4S can be used to demonstrate the advantages and limitations of this approach. Figure 2.9a is a plot of the hydraulic conductivity profiles obtained at these two wells using the DP_2 configuration of Zlotnik and Zurbuchen (1998). The significant difference between the Gems4N and Gems4S profiles at the top of the aquifer is consistent with differences in natural-gamma logs from the two wells (Butler et al., 1998a). The average K obtained from the two DFT profiles is considerably less than that obtained from the pumping tests (dashed vertical line). For example, the average K from the DFT at Gems4S is 73 m/d, which is 63% of that from the pumping tests. This difference could be a result of a number of factors, including insufficient testing near the bottom of the interval where the K appears highest (Figures 2.5 and 2.6), heterogeneity, and incomplete well development. In this case, the first two factors appear to be the most probable contributors to the difference. Insufficient testing, which is undoubtedly the most important contributor, occurs because of the length of the DFT tool (3.19 m) and the termination of the screened interval above the bottom of the sand and gravel. As will be shown later (Section 2.3.2), complete testing across this interval could have produced an average K_r estimate much closer to that obtained from the pumping test.

The steady-state DFT analysis is based on the assumption of a homogeneous formation. However, the profiles displayed in Figure 2.9a indicate that this assumption may not be appropriate at Gems4N and Gems4S. Thus, an analysis approach more appropriate for heterogeneous formations is needed. Zlotnik and Ledder (1996) have demonstrated that head gradients are largest near the two chambers, as would be expected in a convergent flow system. The hydraulic conductivity of the portions of the formation in the vicinity of a chamber should therefore have the greatest influence on the head changes in that chamber. Thus, a more appropriate analysis method for the DFT in heterogeneous formations would be to estimate K_r using the head changes in a single chamber. Zlotnik and Ledder (1996) have derived a single-chamber form of the DFT formula:

$$K_r = \frac{Q}{4\pi(\Delta h_{sc})\Delta} \ln\left[\frac{4\,a\,\Phi(\lambda)\Delta}{e r_w}\right] \tag{2.2}$$

where Δh_{sc} is the head change in a single chamber (Δh_{LT} or Δh_{UT}, Figure 2.8b). Zlotnik et al. (2001) and Zurbuchen et al. (2002) have demonstrated the increased level of detail that can be obtained with the single-chamber formula. This increased level of detail was also observed at Gems4S (Figure 2.9a). Note that the upper chamber should be used for this analysis to avoid an underestimation in K_r resulting from the inadvertent injection of fine materials into the lower chamber (i.e., formation of a low-K skin during the DFT).

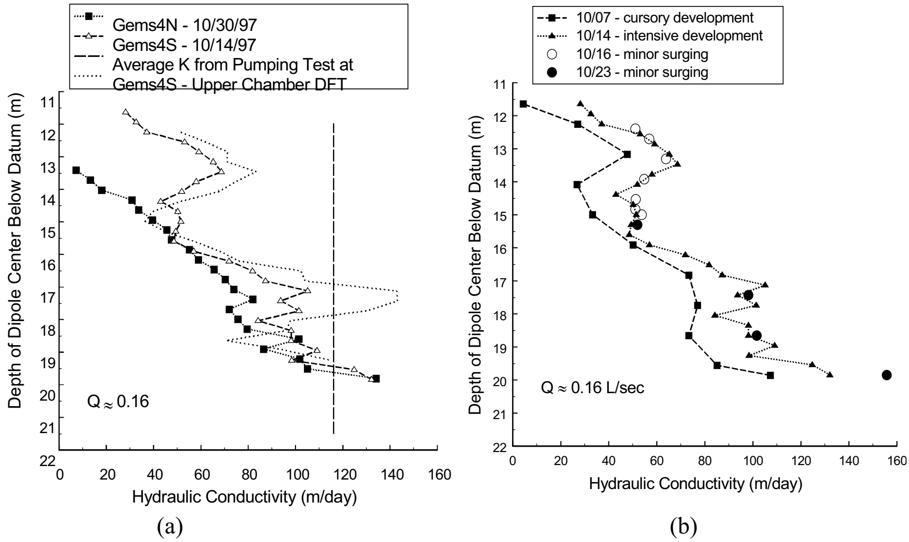


Figure 2.9. (a) Hydraulic conductivity profiles from DFTs at Gems4N and Gems4S; (b) Impact of well-development activities on DFT profiles at Gems4S (after Butler et al., 1998a)

Zlotnik and Zurbuchen (1998) emphasize several important advantages of the DFT for field applications. These include (1) no water is added or removed from the well during a test program, (2) the scale of the region influenced by the test can be readily defined and controlled, (3) K_r estimates can be obtained using a simple steady-state formula, and (4) relatively low flow rates can be used in high K media, so that the well losses associated with other methods are avoided. In addition, the dipole tool can be configured so that the influence of a high-conductivity zone created by the filter pack is minimized (Peursem et al., 1999). However, as with any single-well hydraulic test, the results are highly dependent on the effectiveness of well-development activities. During the drilling process, a considerable amount of fine debris will be concentrated in the near-well portions of the formation. One of the primary goals of well-development activities is to remove this drilling-generated material, so that K_r values representative of the formation can be obtained. Figure 2.9b displays the results of DFT surveys performed at Gems4S after varying degrees of well development. The cursory development consisted of pumping at a constant rate (1.3 L/s) until an approximately clear stream of water was obtained (20 min). The intensive development consisted of stressing discrete intervals via pumping and surging. The final two phases of development consisted of a small amount of surging followed by pumping to remove the debris brought into the well. At Gems4S, the intensive development produced an increase in the magnitude of the K_r estimates (profile average increased by 33%) but minimal change in the profile shape. The minor surging produced little change except at the bottom of the well, where fine material that had accumulated during previous development activity was removed. Note that the change in the average K_r was much greater at Gems4N (increase by a factor of 6.7) as a result of clay being smeared across the sand and gravel interval during drilling. Clearly, proper attention must be given to the development of wells at which the DFT, or any other single-well hydraulic test, is to be applied. The stability of K_r profiles before and after a period of development is the most convincing demonstration of the sufficiency of the development activities.

2.3.2 MULTILEVEL SLUG TEST

The multilevel slug test (MLST) is an extension of the traditional slug test specifically developed for characterizing vertical variations in hydraulic conductivity along the screened interval of a well (Melville et al., 1991; Butler et al., 1994). This approach involves the performance of slug tests in a portion of the screen isolated with a two-packer tool (Figure 2.10 [left]). A number of tests are performed in each isolated interval (Figure 2.10 [right]) to assess the viability of test assumptions (Butler et al., 1996; Butler, 1998). Test data are analyzed as discussed in Section 2.2.2. A K profile is obtained by repeating this process as the tool is moved in steps through the screened interval. The tool depicted in Figure 2.10 (left) was specifically developed for slug tests in highly permeable aquifers, so the pneumatic method is used for test initiation to minimize the noise introduced by noninstantaneous initiation. Other initiation methods (solid slug, etc.) can be used in less permeable intervals. Note that packer circumvention can affect test results in certain conditions. Butler et al. (1994) demonstrate those conditions and recommend measures for diminishing the potential for circumvention (use of additional or longer packers).

A series of MLSTs performed at Gems4N and Gems4S can be used to demonstrate the potential of this approach. Figure 2.11 is a plot of the K profiles obtained at the two wells using the tool depicted in Figure 2.10. The average K from the multilevel slug tests at Gems4N and Gems4S is 77% and 81%, respectively, of the average K from the pumping tests. This difference from the pumping-test K is likely a result of the incomplete testing of the bottom portion of the aquifer. As shown on Figure 2.11, the bottom 1.5 m of the aquifer is not tested at either well because both wells terminate above the bottom of the aquifer. If the K value of the deepest interval tested at each well is assumed to represent the untested interval for that well, the average K value from the MLST is 90% (Gems4N) and 93% (Gems4S) of the pumping-test K. Heterogeneity and anisotropy could easily account for the remaining difference. As discussed by Butler (1998), the anisotropy ratio cannot be estimated from a slug test in the absence of observation wells, so a value for it must be assumed in the analysis. Isotropy was assumed here for lack of better information.

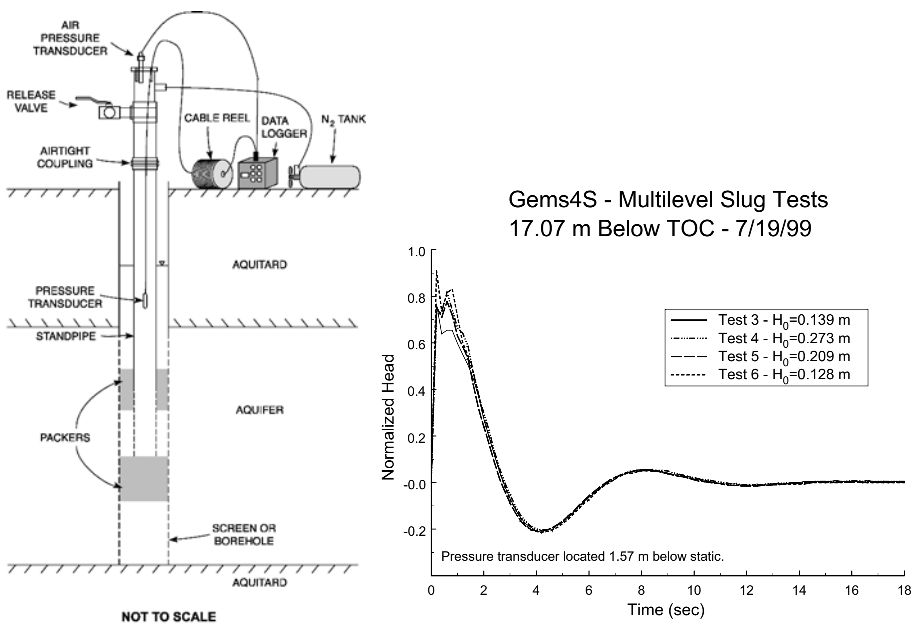


Figure 2.10. (left) Schematic of multilevel slug test configuration (after Butler, 1998); (right) normalized head versus time plot from one test interval

The upper-chamber DFT profile from Gems4S is also plotted on Figure 2.11 to demonstrate the similarity between profiles obtained with two different techniques. Note that the upper two DFT K values plotted on Figure 2.11 are questionable because of concerns about leakage through the top DFT packer during those tests. The excellent agreement between the MLST and upper-chamber DFT profiles is strong evidence of the viability of these two approaches (Zlotnik and Zurbuchen, 2003). In addition, the similarity of the average K from each profile to the average K from the pumping test, when the untested region is considered, is further evidence for the viability of these methods.

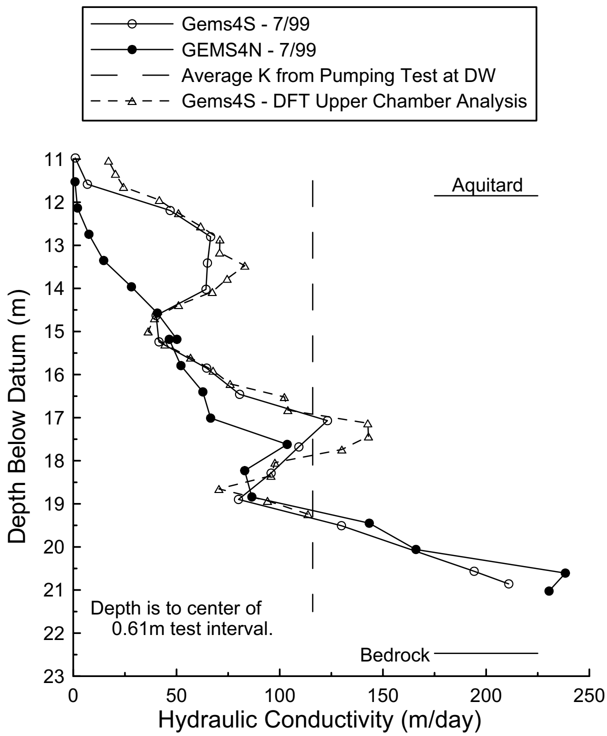


Figure 2.11. Hydraulic conductivity profiles for MLSTs at Gems4S and Gems4N and DFTs at Gems4S (Butler et al., 1998b; previously unpublished KGS data)

2.3.3 BOREHOLE-FLOWMETER TEST

The borehole-flowmeter test (BFT) is the most efficient, in terms of both field and analysis procedures, among the current generation of techniques for estimation of spatial variations in K. This approach involves pumping a well at a constant rate while measuring the vertical flow within the well using a downhole flowmeter (Figure 2.12). The flowmeter is initially positioned at the bottom of the screen and then systematically moved up the well while pumping continues. The flowmeter is kept at each depth until a stable reading is obtained (usually a few minutes). A test is completed when the flowmeter reaches the water table or the top of the screen. The record of cumulative vertical flow versus depth (Figure 2.12 and Figure 2.13a) is then processed to obtain a profile of hydraulic conductivity versus depth (Molz et al., 1989; Molz and Young, 1993). This processing has two primary steps. First, the flow between successive flowmeter positions (ΔQ) is calculated by subtracting the lower flowmeter reading from the upper one and taking any ambient flow into account. Second, hydraulic conductivity is calculated from the ΔQ record using several approaches. The simplest and most defensible of the analysis approaches is based on a study of pumping-induced flow in perfectly layered aquifers by Javandel and Witherspoon (1969). In that study, it was shown that, when the pumping well is fully screened across the aquifer, the ΔQ of an interval (ΔQ_i) will be proportional to the hydraulic conductivity (K_i) and thickness

(Δb_i) of that interval. The ratio of K_i over the average K for the aquifer ($\overline{K}$) can therefore be calculated as:

$$\frac{K_i}{\overline{K}} = \frac{\Delta Q_i / QP}{\Delta b_i / B} \tag{2.3}$$

where QP is the total pumping rate and B is the aquifer thickness. A record of K_i versus depth can then be obtained by multiplying this ratio by the average hydraulic conductivity of the aquifer near the well. Although the average K from a pumping test at the well is often used for $\overline{K}$, this procedure can introduce error into the K_i estimates, because the average K in the immediate vicinity of the well may differ from that determined from a pumping test in a laterally heterogeneous aquifer. Use of the K from a slug test at the same well should be a better approach. Young and Pearson (1995) describe different types of flowmeters and conclude that the electromagnetic flowmeter is the best for use in aquifers. Boman et al. (1997) describe field applications of the electromagnetic borehole flowmeter and discuss important practical issues. Chapter 10 of this volume discusses flowmeter use in fractured or multi-aquifer systems.

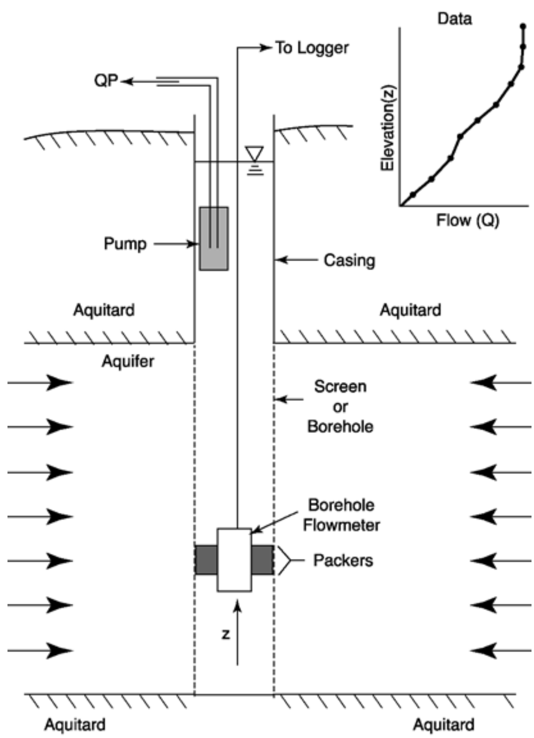
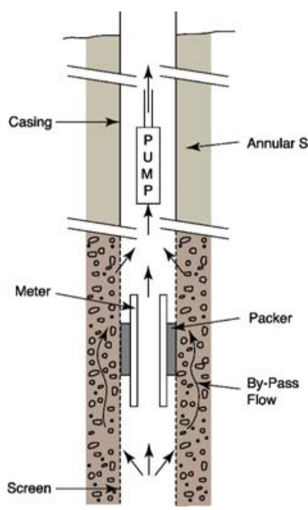
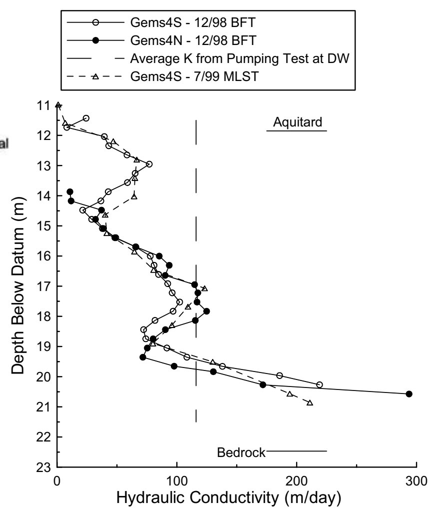


Figure 2.12. Schematic of borehole- flowmeter test (after Molz et al., 1989)

A series of BFTs performed at Wells Gems4N and Gems4S demonstrate both the potential and pitfalls of this approach. Figure 2.13a is a plot of cumulative discharge versus depth obtained at Gems4S using the 0.025 m ID electromagnetic flowmeter described in Young and Pearson (1995). Mechanical packers were attached to the device to direct flow through the meter as shown in Figure 2.12. The flow rate was measured every 0.305 m, and ΔQ was calculated from two successive readings as discussed earlier. Figure 2.13b, a plot of ΔQ versus depth at Gems4S, reveals some of the problems that can arise in highly permeable aquifers. Note the large increase in flow at the top of the screen as well as the intervals of negative ΔQ (those to left of dashed vertical line at zero). Both of these features are a result of flow bypassing the flowmeter (Figure 2.14a), a phenomenon that often occurs in highly permeable formations because the head loss through the flowmeter is comparable or greater than that produced by bypass flow in the formation (Dinwiddie et al., 1999). The increase in flow at the top of the screen is a result of the flowmeter moving into the casing where bypass flow no longer is possible. An attempt was made to minimize head losses through the flowmeter by using a small flow rate as recommended by Arnold and Molz (2000), but bypass flow is difficult to avoid in highly permeable units. An unweighted 1.52 m moving average was employed to remove the zones of negative flow on Figure 2.13b. Processing then proceeded using the smoothed data. The processed data were multiplied by the average value of the multilevel slug tests for the tested interval at Gems4S and Gems4N to produce a record of K versus depth. Figure 2.14b compares the BFT K profiles from Gems4S and Gems4N with the MLST K profile from Gems4S. Given the problems introduced by bypass flow, the agreement between the BFT and MLST profiles is quite good. Although these BFT experiments were performed by pumping water from the well, the test could also be performed in an injection mode. However, considerable care must be used to avoid injecting entrained air and sediments that can artificially lower the K in the immediate vicinity of the screen (Crisman et al., 2001).

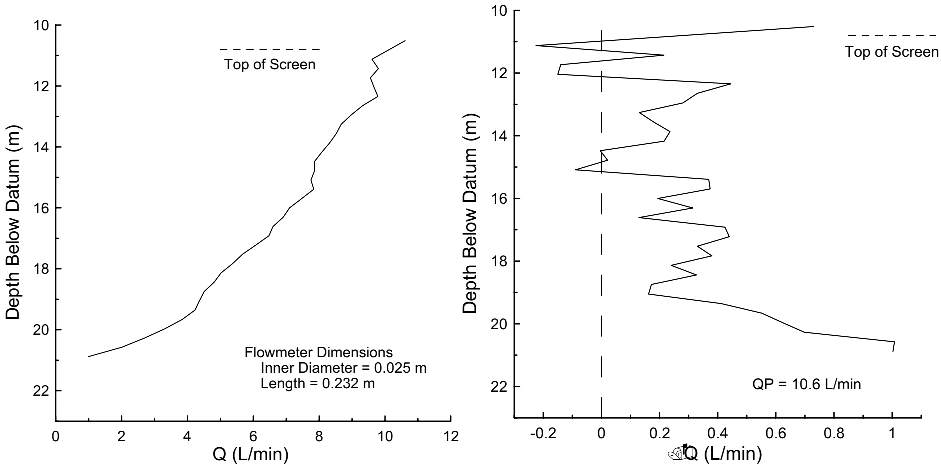


(a)



(b)

Figure 2.13. (a) Plot of cumulative flow versus depth at Well Gems4S; (b) Plot of ΔQ versus depth at Gems4S (after Butler et al., 1998b)



(a) (b)

Figure 2.14. (a) Schematic of bypass flow (after Boman et al., 1997); (b) K profiles from borehole-flowmeter tests at Gems4S and Gems4N and multilevel slug tests at Gems4S (Butler et al., 1998b; previously unpublished KGS data)

2.3.4 SUMMARY OF THE CURRENT GENERATION

When appropriate attention is given to the details of the field and analysis procedures, the current generation of methods for estimation of vertical variations in hydraulic conductivity are capable of providing reliable estimates of K variations along the screened interval of a well. Given the differences in the theoretical bases of the three techniques that were the primary focus of this section, the agreement between the K profiles obtained with the various methods (Figure 2.15) is very strong evidence of the viability of these approaches (Zlotnik and Zurbuchen, 2003). The similarity between the profile averages and the average K from the pumping tests further demonstrates the quality of the information that can be obtained with these approaches. The decision regarding which approach to use in a particular investigation may not be straightforward. The borehole-flowmeter test is the fastest in terms of both field and analysis procedures, but in-well hydraulics (bypass flow) can introduce considerable uncertainty in highly permeable aquifers, and the removal (pumping mode) or addition (injection mode) of water may conflict with regulatory restrictions at some sites. An important limitation of this approach is the requirement that the well be screened across the entire aquifer. The dipole-flow test is well suited for use in highly permeable aquifers, has a well-defined scale, and does not require the addition or removal of water. However, the complexity and size of the multi-packer tool can limit its utility. In addition, a relatively long-screened well is required to obtain information about K variations. The multilevel slug test is the most flexible of the three techniques in terms of the K range over which it can be applied, the size of the test interval, and the length of the well screen—but the analysis procedure is considerably more involved.

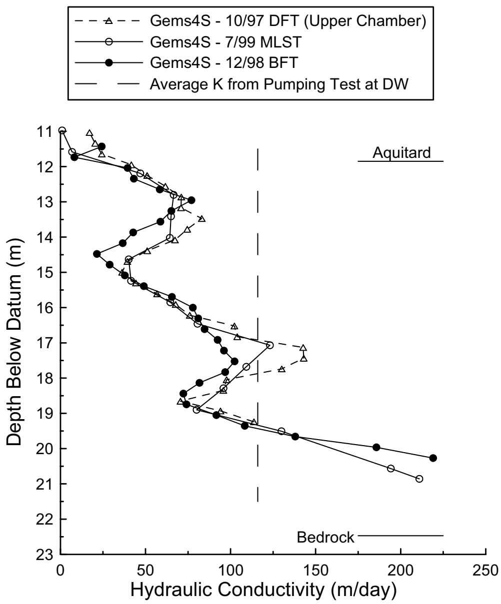


Figure 2.15. Hydraulic conductivity profiles obtained at Gems4S using the dipole-flow test, the multilevel slug test, and the borehole-flowmeter test.

2.4 The Next Generation

There are two major limitations to the current generation of methods for estimation of spatial variations in hydraulic conductivity. First, these techniques can only be used in wells, which often must be screened across a relatively large portion of the aquifer. Second, these techniques only provide information about conditions in the immediate vicinity of the well in which they are used. New techniques are being developed that provide information about K variations outside of the immediate vicinity of existing wells. These methods are described in this section.

2.4.1 DIRECT-PUSH METHODS

Most of the methods discussed so far in this chapter have involved procedures performed in wells. The restriction of these techniques to existing wells limits their utility for characterizing spatial variations in K. Over the last two decades, direct-push technology has become a widely used alternative to conventional well-based methods for site investigations in unconsolidated formations. A variety of methods have been developed that exploit the unprecedented access to unconsolidated formations provided by this technology, to obtain information about spatial variations in K without the need for permanent wells. For example, empirical relationships have been developed for estimating K from sediment classification information produced by cone penetrometer (CPT) surveys (Farrar, 1996). The resulting K values, however, are, at best, order-of-magnitude estimates for the hydraulic conductivity of the formation. Pore-pressure dissipation tests accompanying CPT surveys have been used to estimate K from a relationship between hydraulic conductivity and the consolidation properties of the formation (Baligh and Levadoux, 1980). Lunne et al. (1997), however, caution that this relationship results in relatively poor approximations of K. Pitkin and Rossi (2000) describe a method for estimation of relative variations in K by monitoring the rate and pressure of water injected during the advancement of an unshielded well point. Dietrich et al. (2003) have demonstrated that this approach can yield semi-quantitative K estimates by utilizing regressions with existing K data. A more promising direction for quantifying actual K variations is the performance of various types of hydraulic tests in direct-push equipment. Two classes of these methods will be discussed here.

2.4.1.1 *Direct-Push Slug Tests*

Slug tests can be performed in direct-push pipe using small-diameter adaptations of conventional methods (Butler et al., 2002). Systems have been developed that allow tests to be performed at one or multiple levels in a single probehole. The earliest approach was that of Hinsby et al. (1992), in which slug tests are performed in small-diameter pipe attached to an unshielded well point that is driven from the surface. This approach may be reasonable in sand and gravel sequences with little clay, but it is of questionable effectiveness in more heterogeneous settings where the buildup of fine material on the well screen can affect test responses. Butler and coworkers (Butler et al., 2002; Butler, 2002) describe a method for performing slug tests in direct-push equipment in which a screen is driven within protective steel casing to the target depth, and then exposed to the formation for the tests. If the screen is exposed in a series of steps, the vertical variations in K over the screened interval can be assessed (Figure 2.16a). The screen cannot be reshielded downhole, so the equipment has to be brought to the surface before another level can be tested. McCall et al. (2002) have developed a shielded-screen approach that enables multiple levels to be tested in a single probehole. Sellwood et al. (in review) increased the efficiency of that approach and added an electrical-conductivity probe for lithologic determination. In intervals of high hydraulic conductivity, frictional losses in the small-diameter direct-push pipe can introduce error into K estimates. Butler (2002), however, presents a procedure for accounting for those losses. Direct-push extensions of conventional pumping tests can also be utilized for estimation of vertical variations in K. Cho et al. (2000), for example, describe a constant-head pumping test method for direct-push equipment. Injection-based variants of their approach are the most promising, because there is no suction-depth restriction, and higher flows can be obtained. However, the time to steady state and other logistical issues may limit the utility of these methods in many systems. Regardless of which approach is used, the quality of K estimates is critically dependent on the efficacy of development procedures. Henebry and Robbins (2000) and Butler et al. (2002) describe development procedures designed for use in small-diameter direct-push installations.

A series of direct-push slug tests (DPSTs) performed at GEMS can be used to demonstrate the potential of this approach. Figure 16a presents a comparison between direct-push slug tests performed at DP43b and DP43c (0.84 m apart), and multilevel slug tests performed at Gems4N (DP43b and DP43c are 1.85 m and 2.16 m, respectively, from Gems4N). At both direct-push installations, three sets of slug tests were performed as the screen length was progressively increased from 0.15 m to 0.61

m. As explained by Butler (2002), the test data can be analyzed to obtain a detailed view of the K variations over the 0.61 m interval tested at the two locations. The agreement between the DPST and the MLST estimates over the same interval is quite good (within 6% at the bottom of the interval and within 12% at the top), demonstrating that direct-push slug tests can provide reliable K estimates.

Figure 2.16a shows that the DPST can be used to acquire detailed information about K variations over a single interval. However, as discussed above, the approach can also be implemented through a profiling procedure in which multiple levels are tested in a single probehole (McCall et al., 2002; Sellwood et al., in review). Figure 2.16b presents three series of direct-push slug tests performed in profiling mode between Gems4S and Gems4S (DPST test interval of 0.31 m in all cases). The MLST profile from Gems4S is also presented for comparison purposes. As shown in the figure, the DPST and MLST profiles are not in good agreement in the upper portion of the aquifer. The trough and peak in the MLST record is shallower than that in the DPST records at points A and B. This difference cannot be explained by the dissimilar length of the test intervals (0.31 m DPST versus 0.61 m MLST). Heterogeneity is undoubtedly the cause of this difference, as it is for the difference between the MLST K profiles at Gems4N and Gems4S (Figure 2.11). The DPST and MLST profiles are in much better agreement in the lower portion of the aquifer, as the trough and peak in the vicinity of C and D, respectively, are at similar depths. In this case, the difference between profiles may primarily be a function of the length of the test interval. Note that the average K from each of the DPST profiles is within 4% of that obtained from the pumping test at DW.

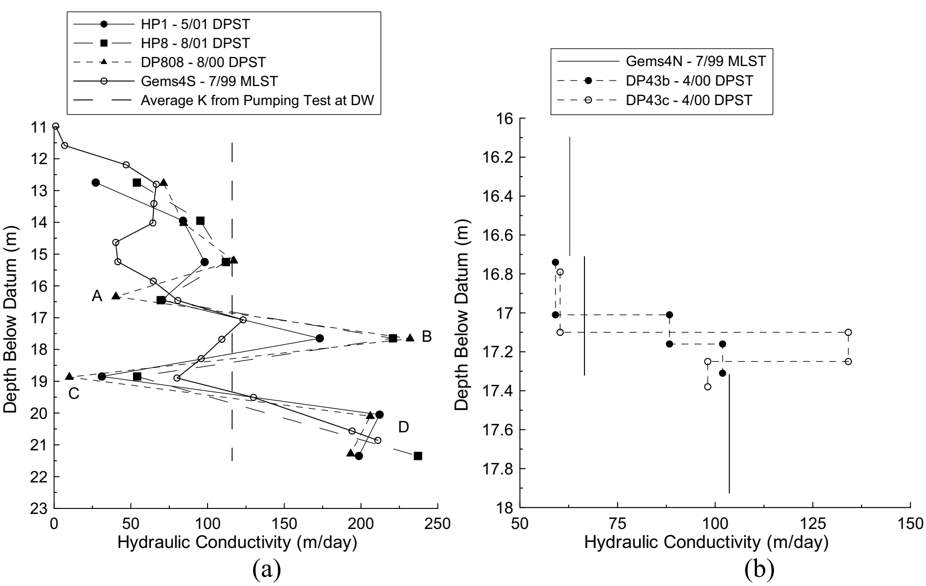


Figure 2.16. (left) Comparison of MLST K estimates from Gems4N with DPST K estimates from nearby direct-push installations (after Butler, 2002); (right) K profiles from DPST at three locations and from MLST at Gems4S (McCall et al., 2002; Sellwood et al., in review)

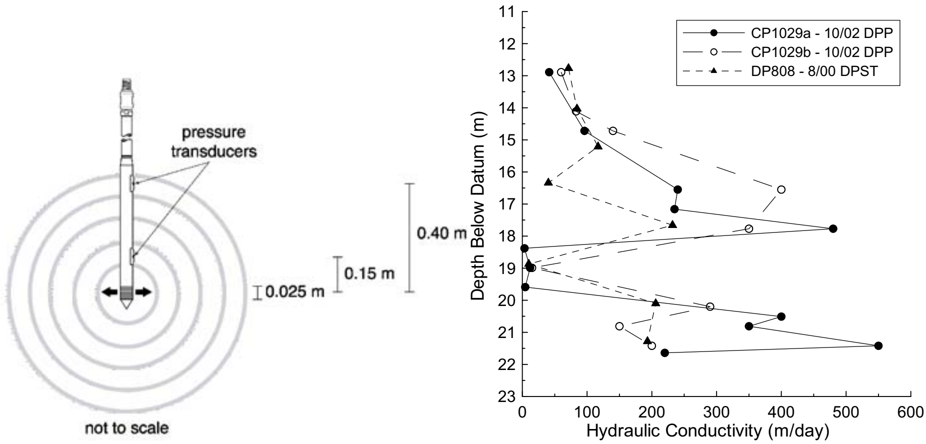
2.4.1.2 *Direct-Push Permeameter*

A major limitation of the direct-push slug test, as with the methods discussed in Section 2.3, is that the quality of the K estimate is highly dependent on the effectiveness of well-development activities. Unfortunately, the uncertainty regarding the effectiveness of development activities can never be completely eliminated. To address this limitation, a method has been developed that is relatively insensitive to the zone of disturbance/compaction created by the advancement of a direct-push tool (Stienstra and van Deen, 1994; Lowry et al., 1999). This method, which is referred to here as the direct-push permeameter (DPP), involves an unshielded-screen tool with pressure transducers inset into the pipe above or below the screen (Figure 2.17a). The tool is pushed into the ground while water is injected at a low rate to keep the screen clear. Upon reaching a depth at which a K estimate is desired, pushing ceases and a pumping test is performed by injecting water through the screen while monitoring pressure changes at the transducer locations. K estimates are obtained using the spherical form of Darcy's law at steady-state (constant hydraulic gradient) conditions:

$$\mathrm{K} = \left(\mathrm{Q}\left(\frac{1}{\mathrm{r}_2} - \frac{1}{r_1}\right) \right) / \left(4\pi (p_2 - p_1) \right) \tag{2.4}$$

where r_i and p_i are the distance to and pressure at transducer i. Although Equation (2.4) assumes an isotropic aquifer, an anisotropic form of this equation can readily be developed for use when the anisotropy ratio is known from other information. Lowry et al. (1999) have shown that a compacted zone of lower permeability along the direct-push pipe will not influence steady-state pressure gradients and thus the K estimate. An important advantage of this approach is that the scale of the test is readily defined and controlled, because the K estimate represents the average over the interval between the transducers.

A field assessment of the direct-push permeameter at GEMS demonstrates the potential of this approach, as well as the need for further refinement (Butler et al., 2004). Figure 2.17b presents DPP K profiles obtained at CP1029a and CP1029b, and a DPST K profile obtained at DP808 (DP808 located 2.24 m and 1.72 m from CP1029a and CP1029b, respectively). As shown in the figure, the profiles are in good agreement in the lower permeability zones at the top of the aquifer and at about 19 m below datum. The agreement is quite poor in the higher K intervals. The poor agreement in the higher-K intervals is most likely a result of the lower signal-to-noise ratio of the pressure measurements in those intervals and of preferential flow along the direct-push pipe. A recent follow-up field assessment in Germany found that the DPP can provide reasonable estimates for K between 0.01 and 100 m/day (Dietrich et al., 2003), in keeping with the range reported by Stienstra and van Deen (1994). Modification of the equipment to allow use at higher K ranges is proceeding. These initial field assessments indicate that this approach should soon be capable of producing reliable estimates over the range of K values expected in most aquifers. The efficiency of the approach is particularly noteworthy. A K profile can be obtained with the direct-push permeameter much faster than with slug-test profiling methods (the DPP profiles of Figure 2.17b were each obtained in two hours, while the DPST profile was obtained in two days). The direct-push permeameter appears to be as efficient as a flowmeter survey, but without many of its limitations. Clearly, this approach has potential for characterizing spatial variations in hydraulic conductivity at a level of detail and efficiency that has not previously been possible. Note that Sørensen et al. (2002) have recently proposed a related approach using hollow-stem augers that may allow testing at greater depths than possible with direct-push technology.



(a) (b)

Figure 2.17. (a) Schematic of direct-push permeameter; (b) K profiles from DPP at CP1029a and CP1029b and DPST at DP808 (McCall et al., 2002; Butler et al., 2004)

2.4.2 HYDRAULIC TOMOGRAPHY

Except for the pumping tests described in Section 2.2.1, all of the methods that have been discussed here provide information about conditions in the immediate vicinity of a well or probehole. Solute transport, however, depends critically on the connectivity of regions of low or high hydraulic conductivity, which may be difficult to determine from a suite of essentially point values. Although multiwell tracer tests can provide information about hydraulic conductivity variations between wells (Freyberg, 1986; Hess et al., 1992), such tests are too expensive in terms of time, money, and effort to be used on a routine basis. Thus, new, more efficient approaches are needed to provide information about spatial variations in K between wells.

Over the last decade and a half, several research groups have begun work on a new approach, hydraulic tomography, that has the potential to yield information on the spatial variations in K between wells at a level of detail that has not previously been possible (Neuman, 1987; Tosaka et al., 1993; Gottlieb and Dietrich, 1995; Butler et al., 1999a; Yeh and Liu, 2000; Liu et al., 2002; Bohling et al., 2002). This method consists of performing a series of short-term pumping tests in which the position of the stressed interval in the pumping well is varied between tests (Figure 2.18a). The sequence of tests produces a pattern of crossing streamlines in the region between the pumping and observation wells, similar to the pattern of crossed ray paths used in seismic or radar tomography (see Chapters 7 and 9 of this volume). Although a large number of drawdown measurements is required to delineate the numerous streamlines produced by the test sequence, new methods for drawdown measurement have been developed that can provide the requisite density of data in a practically feasible manner (Butler et al., 1999a; Davis et al., 2002). Bohling et al. (2002) describe a variety of methods for the analysis of the drawdown data collected during a test sequence. All of these methods involve the numerical inversion of test data, assuming a model of the aquifer structure based on existing site information. Hydraulic tomography extracts information about variations in K from vertical differences in drawdown. These differences, however, diminish as the distance between the pumping and observation wells increases. At distances greater than $2B(K_r/K_z)^{1/2}$, where B is aquifer thickness and K_r and K_z are the average radial and vertical component of K, respectively (Kruseman and de Ridder, 1990), there will rarely be a measurable difference in drawdown in the vertical direction (Bohling et al., 2002).

A series of hydraulic tomography experiments performed at GEMS illustrate the type of information that can be obtained with this approach. In these experiments, Gems4S served as the pumping well, while drawdown measurements were made at multilevel sampling wells HTMLS1 and HTMLS2, which are 2.74 m and 5.74 m, respectively, from Gems4S (Davis et al., 2002). Pumping tests (Q=1.3 L/s) were performed at 15 different intervals approximately equally spaced across the screened interval at Gems4S, each test lasting 900 seconds. Drawdown was measured at three depths in each multilevel well. The depths at which drawdown was measured were offset between the wells to allow more complete coverage across the aquifer. Drawdown records were similar in form to those depicted in Figure 2.3. The data from HTMLS1 and HTMLS2 were numerically inverted to estimate conditions between those two wells, using a seven-layer model based on a crosshole radar survey between wells Gems4N and Gems4S. As shown in Figure 2.18b, the comparison between the numerical inversion of the tomography data and the DPST profile at HP1 is quite reasonable. Differences between the HP1 profile and the inversion results are most likely caused by differences between the actual aquifer structure and the model based on the radar data, and the termination of the screened interval in well Gems4S above the bottom of the aquifer. Note that the average K from the inversion of the hydraulic tomography data is 78% of that obtained from the pumping test at DW, most likely again a result of the undersampling of the permeable lower portion of the aquifer.

Variations of the hydraulic tomography procedure described above are currently under development. For example, ongoing work is exploring the use of additional geophysical methods for better definition of the model of the aquifer structure used for inversion of test data. In addition, direct-push methods are being explored as a flexible means of installing temporary observation wells. (Figure 2.3b is an example of an initial test of that concept.) The combination of geophysical methods for structure definition and direct-push methods for data acquisition has great potential.

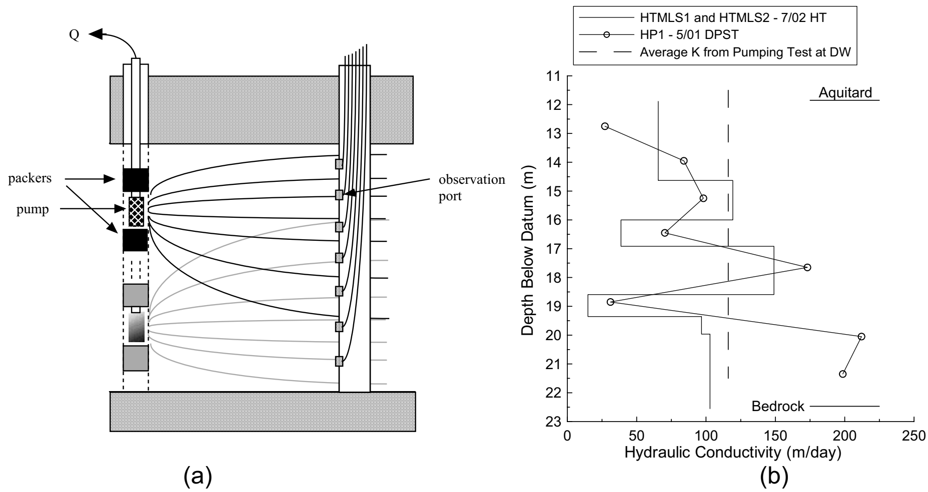


Figure 2.18. (a) Schematic of hydraulic tomography procedure (after Butler et al., 1999a); (b) K profile from hydraulic tomography experiments using HTMLS1 and HTMLS2 and K profile from DPST at HP1 (after Bohling et al., 2003; Sellwood et al., in review)

2.5 Additional Methods

The major methods for obtaining information about spatial variations in hydraulic conductivity in shallow unconsolidated aquifers have been described in this chapter. These, however, are not the only methods that can be used to estimate spatial variations in K. Several additional methods have been employed to estimate spatial variations in K in shallow unconsolidated aquifers. Moreover, a number of techniques have been specifically developed for use in other hydrogeologic settings. In this section, these additional methods are briefly described.

2.5.1 METHODS FOR SHALLOW UNCONSOLIDATED AQUIFERS

The most commonly used method for investigations in shallow unconsolidated aquifers not covered in the preceding sections is the tracer test. Chemical and physical tracers have long been used to obtain information about the permeable nature of subsurface formations. Single-well borehole-dilution and injection-withdrawal tests (Freeze and Cherry, 1979; Leap and Kaplan, 1988; see Chapter 10 of this volume) performed with and without packers have been utilized to determine the relative variation in groundwater velocity along the screened interval of a well. These variations can be related to K variations under certain conditions (e.g., constant hydraulic gradient along the screen). Various other single-well tracer tests have been proposed in the last two decades (Taylor et al., 1990; Sutton et al., 2000), but have seen relatively limited use because of their involved procedures. Multiwell tracer tests performed under natural or induced hydraulic gradients can provide information about K variations between wells (Freyberg, 1986; Hess et al., 1992; Gelhar et al., 1992). Although significant advances in the analysis of multiwell tracer tests have been achieved (e.g., Datta-Gupta et al., 2002), logistical and regulatory constraints still limit their use.

An induced-gradient tracer test, GMSTRAC1, performed at GEMS in the fall of 1994, illustrates the effort required for, the logistical problems associated with, and the potential information that can be obtained from multiwell tracer tests. A bromide tracer was introduced into a steady flow field created by pumping well DW at a constant rate (4.4 L/s). The tracer moved 14.2 m laterally from the point of injection to Well DW through a network of 24 multilevel sampling wells (17 sampling ports/well, ports distributed evenly across the sand and gravel aquifer) located to the immediate northeast of the wells shown in the inset of Figure 2.1. The test lasted 32 days, and over 6,000 samples of bromide were collected and analyzed during that period. An evaluation of the tracer data produced a profile of fluxes that was then transformed into a K profile, using the average hydraulic conductivity from the pumping tests and the assumption of horizontal flow (Bohling, 1999). Figure 2.19 is a comparison of the tracer-test K profile with those obtained from direct-push slug tests performed at locations adjacent to the sampling well network. Although the patterns of the tracer-test and slug-test profiles are in reasonable agreement in the lower half of the aquifer, the K values are not. Bohling (1999) details the problems that likely influenced the K profile obtained from the tracer test. These included the inability to introduce the tracer uniformly over the aquifer (and thus define the tracer distribution immediately after introduction), more rapid-than-expected tracer movement in the lower portion of the aquifer, and the influence of nearby pumping wells. Although the GMSTRAC1 test was impacted by a number of design and logistical problems, the resulting K profile does illustrate the potential of multiwell tracer tests for assessing spatial variations in K. Thus, despite the expense and effort required to perform multiwell tracer tests, they often can provide valuable information about interwell variations in K. However, in many cases, the approaches described in the preceding sections can provide similar information in a significantly more efficient (time, cost, and effort) fashion. To get the most information from tracer tests, attention must be given to all phases of test design and performance, and, whenever possible, tests should be performed in conjunction with geophysical surveys (see Chapter 13 of this volume).

Nuclear magnetic resonance (NMR) methods also have potential for providing information about spatial variations in hydraulic conductivity in shallow unconsolidated settings. NMR logging is commonly used in the petroleum industry to obtain information about vertical variations in K. This approach involves measuring the response of hydrogen protons to a series of imposed magnetic fields using a downhole logging tool (Coates et al., 2001). The response is a function of, among other things, the size and distribution of pores in the immediate vicinity of the borehole. Hydraulic conductivity is estimated from the pore-size distribution information with techniques similar to those described in Section 2.2.3 (White, 2000). Calibration using permeameter analyses of cores from the logged borehole is done to reduce the uncertainty associated with the K estimates. NMR logging could potentially be a useful means for rapid acquisition of information regarding relative variations in K, but logistics (e.g., size of the logging tool) and costs have greatly limited its use for shallow hydrogeologic investigations. In addition, there are concerns regarding the interpretation of NMR data in the near-surface environment owing to the sensitivity of the data to redox conditions (Bryar and Knight, 2002). Yaramanci et al. (2002) have recently demonstrated the potential of NMR surface surveys for obtaining vertical profiles of K. Further discussion of NMR methods is provided in Chapter 16 of this volume.

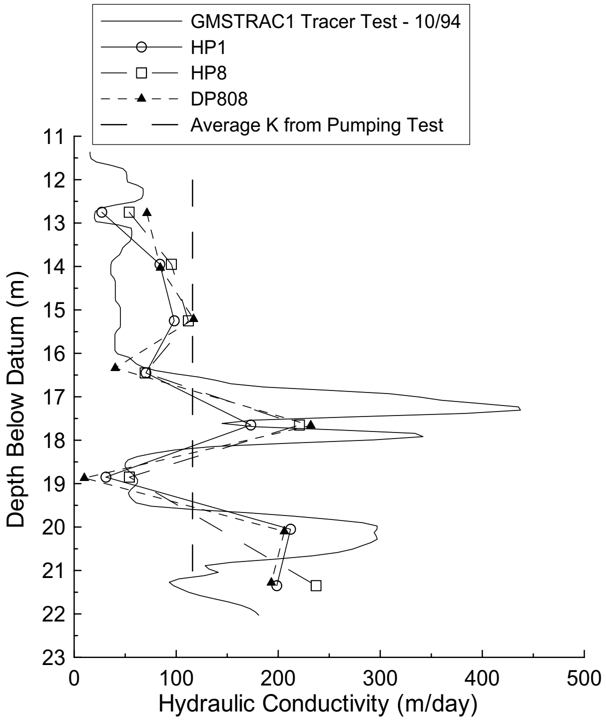


Figure 2.19. Hydraulic conductivity profile for GMSTRAC1 tracer test and DPST K profiles from nearby locations (Bohling, 1999; McCall et al., 2002; Sellwood et al., in review).

2.5.2 METHODS FOR CHARACTERIZING OTHER HYDROGEOLOGIC SETTINGS

Fractures in media of relatively low permeability often play an important role in groundwater flow and solute transport. Thus, efforts to acquire information about spatial variations in K in fractured settings often focus on determining which fractures serve as conduits for flow. Constant-head injection tests (packer tests) performed with a straddle-packer system similar to that depicted in Figure 2.18a are commonly used for this purpose (Doe et al., 1989). Constant-head tests, however, are susceptible to error introduced by low-permeability well skins and the assumption of a steady flow rate (Bliss and Rushton, 1984; Braester and Thunvik, 1984). In addition, such single-well tests cannot provide information about how the identified flow conduits are interconnected in space. Multiwell slug or pulse tests (Novakowski, 1989), and conventional pumping and tracer tests (Karasaki et al., 2000) are commonly used to assess fracture interconnectivity. Traditional interpretations of hydraulic test data, however, may not be viable because of the geometric complexity of the fracture network (Barker, 1988; Doughty and Karasaki, 2002). Multidisciplinary approaches, such as those described by Shapiro et al. (1999) and Karasaki et al. (2000), are often required to improve understanding of K variations in fractured settings. Chapter 10 of this volume describes additional approaches for identifying flow conduits in fractured media.

Many of the methods discussed in this chapter have been developed for use in aquifers. Often, however, there is a need to get information about the transmissive nature of units of relatively low K. Neuzil (1986) describes the many challenges faced by those working to obtain estimates of hydraulic conductivity in low-permeability formations. Van der Kamp (2001) summarizes the major methods for obtaining K estimates in shallow units of relatively low permeability (aquitards) and identifies the slug test as the primary tool for K estimation in that setting. Butler (1998) discusses the variants of the slug test that have been developed for low-K media. Constant-head tests are also commonly used in low-K settings (Tavenas et al., 1990). Van der Kamp (2001), however, questions the utility of constant-head tests, because they are considerably more difficult to perform than slug tests and provide little, if any, additional information.

Estimates of hydraulic conductivity are also needed for the vadose zone, where water is at pressures less than atmospheric and the pore space is usually not completely filled (saturated) with water. Stephens (1996) discusses the additional complexities confronting investigators in unsaturated media and the primary methods used for obtaining information about K in that setting. Reynolds et al. (2002) summarize the major advantages and disadvantages of these methods, while many authors have discussed the factors that can introduce error into the resulting K estimates (Flühler et al., 1976; Campbell and Fritton, 1994). Holt et al. (2002) discuss how such errors in a commonly used technique, the tension infiltrometer, can lead to a bias in the description of the spatial variations of K. The recently proposed tension permeametry approach (Shani and Or, 1995; Or et al., 2000) appears to have much potential for characterizing spatial variations in K. Further work, however, is needed to fully assess that potential.

Given the difficulties associated with efforts to estimate hydraulic conductivity in unsaturated media, many investigators have preferred to use pneumatic-based methods in the vadose zone. In this case, estimates of the air permeability of the media can be obtained using steady-state or transient extensions of conventional hydraulic tests (Baehr and Hult, 1991; Illman and Neuman, 2000, 2001). Although the degree of water saturation is commonly ignored, it should be considered in the calculation of intrinsic permeability from air-permeability estimates when the degree of saturation is above the specific retention (Weeks, 1978). The power of pneumatic approaches has recently been demonstrated by Vesselinov et al. (2001a,b), who used a suite of injection tests to perform pneumatic tomography in an unsaturated fractured tuff. Pneumatic minipermeameters have also been widely used to study permeability variations in outcrops and rock samples (Hurst and Goggin, 1995; Tidwell and Wilson, 1997; Dinwiddie et al., 2003).

2.6 Summary and Conclusions

Varieties of hydrogeologic methods are available for the estimation of spatial variations in hydraulic conductivity in shallow unconsolidated aquifers. A number of those methods were reviewed in this chapter. The potential and limitations of each method were evaluated in a highly controlled field setting. This evaluation demonstrated that the current generation of methods can provide reliable estimates of K variations along the screened interval of a well. The choice of the most appropriate technique for an investigation depends on the particulars (e.g., existing wells, the scale at which estimates are desired, available time) of that investigation. For example, the borehole-flowmeter test is the most efficient of existing techniques, while the dipole-flow test has the best-defined scale of investigation. None of the current generation of methods, however, can provide information away from the vicinity of existing wells, a severe limitation at sites with a sparse well network.

Methods are currently being developed that provide information about K variations away from the vicinity of existing wells. Direct-push methods have the potential to provide information about lateral and vertical variations in K at a level of detail that has not previously been possible. The direct-push permeameter is noteworthy because of its efficiency and its low sensitivity to conditions in the disturbed zone created by the advancement of the direct-push pipe. However, these direct-push methods only provide information about hydraulic conductivity in the immediate vicinity of the probehole. An approach for obtaining information away from existing wells or probeholes is currently under development. This approach, hydraulic tomography, can provide a detailed view of conditions between wells or probeholes in many situations. The integration of direct-push methods, crosshole geophysics, and hydraulic tomography is a particularly promising direction for future research.

Existing or in-development methods can only provide information about K variations in the immediate vicinity of a well/probehole or between relatively closely spaced wells. Additional methods are needed to provide information about the connectivity of regions of high or low hydraulic conductivity under more general conditions. Surface and crosshole geophysical methods have considerable potential in this respect. That potential is explored thoroughly in later chapters of this volume.

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