Spatial variability of specific yield and vertical hydraulic conductivity in a highly permeable alluvial aquifer

Xunhong Chen\textsuperscript{a,∗}, Jinxing Song\textsuperscript{a,b}, Wenke Wang\textsuperscript{c}

\textsuperscript{a}School of Natural Resources, University of Nebraska–Lincoln, Lincoln, NE 68506-0996, United States
\textsuperscript{b}College of Urban and Environmental Sciences, Northwest University, Xi’an 710069, China
\textsuperscript{c}College of Environmental Science and Engineering, Chang’an University, Xi’an 710054, China

1. Introduction

Specific yield ($S_y$) is an important parameter for evaluation of water resource storage in a given unconfined aquifer. The main source of this water is drainage from above the declining water table (Bevan et al., 2005). The magnitude of specific yield affects well yields, solute transport, and rates of water level decline (Said et al., 2005). It is also a critical parameter in the estimation of groundwater evapotranspiration from diurnal water-level fluctuation readings (White, 1932; Lott and Hunt, 2001). The sparseness and low reliability of available $S_y$ values often cause problems in quantifying three-dimensional regional aquifer storage changes (Gehman et al., 2009). Streamflow depletion that is potentially caused by groundwater pumping is a major water resource management issue in the High Plains Region of the United States. Specific yield can be a sensitive parameter in the calculation of streamflow depletion induced by a pumping well which is constructed in a thin aquifer.

The most commonly used methods for estimation of specific yield include drainage measurements of aquifer materials and pumping tests. Johnson (1967) summarized a number of laboratory methods for the determination of specific yield. One of the earliest publications on determination of specific yield of alluvial aquifers from pumping tests was Wenzel (1942). This pumping test was conducted at a site located in the Platte River valley, Nebraska and had 83 observation wells. The publication of Nwankwor et al. (1984) renewed more interest in the groundwater hydrology community in the estimation of specific yield from pumping test data because the specific yield value determined from pumping test data differed much from that determined from other methods.

During pumping tests, the decline of the water table in permeable alluvial aquifers is often very small. According to a pumping test in the Platte River valley of Nebraska (Wenzel, 1942), the drawdown at observation wells from 8 to 20 m from the pumping well ranged from 1.22 to 0.8 m even if a large pumping rate (about 2 m$^3$ per minute) continued for more than 48 h. During the pumping, drainage occurs only for the aquifer materials above the declining water table (Bevan et al., 2005). Thus, pumping tests of a limited duration (for example, several days) often provide the

\begin{itemize}
  \item Specific yield ($S_y$)
  \item Alluvial aquifer
  \item Permeameter test
  \item Drainage experiment
  \item Direct-push equipment
\end{itemize}

\textsuperscript{∗} Corresponding author. Tel.: +1 402 472 0772.
E-mail address: xchen2@unl.edu (X. Chen).

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estimation of the specific yield for the limited thickness of the aquifer materials only near the water table. In practice, pumping can last months. Thus, estimation of the specific yield for the aquifer materials in a larger thickness is very valuable for water resources evaluation.

Although Kollet and Zlotnik (2005) noted that specific yield has highly variable values for a given test site, precise documentation of such variation is scarce. This is mainly due to the fact that construction of observation wells is costly and the small number of observation wells for a pumping test prevents spatial characterization of specific yield. Therefore, a less-expensive approach for determination of the spatial variation of specific yield in unconfined aquifers is truly needed. In fact, Nwankwor et al. (1992) concluded that short-term pumping tests are not reliable methods for determination of specific yield and small-scale laboratory tests may give reasonable values of specific yield. The specific yield determined from laboratory drainage by Nwankwor et al. (1984) has been used as a bench mark value for the Borden aquifer. This value, however, was determined from repacked aquifer materials. During sampling and re-packaging, the original sedimentary structures and particle framework were destroyed. It is ideal to determine specific yield of aquifer materials which have very little disturbance during coring, sampling, or laboratory testing. Furthermore, because of the large spatial variation in \( S_y \) values, a larger number of cores should be collected from the aquifer at a given site for drainage experiments. Some sediment cores must be collected from points closely co-located with observation wells used in a pumping test if \( S_y \) values determined from the two methods (pumping test and drainage experiments using cores) are to be compared.

If the specific yield can vary largely in its magnitude for an unconfined aquifer, the hydraulic conductivity will be heterogeneous as well. Documentation of spatial variation of horizontal hydraulic conductivity \( (K_h) \) commonly exists in the literature (Butler et al., 2002); but the analysis of the spatial variation of the vertical hydraulic conductivity is scarce. In the past several decades, the pumping tests at the Borden site (Nwankwor et al., 1984, 1992; Akindunni and Gillham, 1992; Bevan et al., 2005; Endres et al., 2007) and the Cape Cod site in the USA (Moench, 1995) often have been used to illustrate the procedures for determination of specific yield or for demonstrating the drainage processes during the decline of the water table during pumping. At the Borden site, the unconfined aquifer is composed primarily of medium-grained sand of glacio-deltaic or glacio-fluvial deposits (Bevan et al., 2005, 2007). At the Cape Cod site, the unconfined aquifer consists of highly permeable sand and gravel of glacial outwash (Moench, 1994). From the reported results, the average horizontal and vertical hydraulic conductivity at the Borden site (Endres et al., 2007) is 5.5 and 2.7 m/d, respectively. These values were determined using the Neuman (1974) solution. The ratio of the horizontal to vertical hydraulic conductivity \( \left( \frac{K_h}{K_v} \right) \) is only 2.25. In the simulation of the pumping test at the Borden site, Akindunni and Gillham (1992) used the \( K_h \) (=5.7 m/d) and \( K_v \) (=3.6 m/d) values, which were determined from permeameter tests by Sudicky (1986). However, in the experiments of Sudicky (1986), the sediment samples were re-packed during the permeameter tests. This re-packing process can alter the original sedimentary structure and thus affect the results. Furthermore, these permeameter tests did not produce the values of the vertical hydraulic conductivity of the aquifer materials. Akindunni and Gillham (1992) derived the equivalent vertical hydraulic conductivity from the permeameter test results. The permeameter test results from Sudicky (1986), however, may be smaller than the horizontal hydraulic conductivity but larger than the vertical hydraulic conductivity of the aquifer materials because of the sample remixing. Thus, the \( K_v \) value used in the simulation by Akindunni and Gillham (1992) may be larger than the actual \( K_v \) value of this aquifer. Similarly, the average horizontal and vertical hydraulic conductivity at the Cape Cod site (Moench, 1995) is 103.7 and 56.8 m/d, respectively. The ratio of \( K_h/K_v \) held to be 2 in the analysis of the pumping test data (Moench, 1995).

In contrast to the small ratio of \( K_h/K_v \) at the Borden and Cape Cod sites, pumping test results in unconfined aquifers from other areas often have a large ratio of \( K_h/K_v \). This ratio can be up to 10 and as large as 30 for sediments in Vekol Valley, Arizona (Marie and Hollett, 1996), in the Platte River valley from western to eastern Nebraska (Chen and Ayers, 1997, 1998; McGuire and Kilpatrick, 1998; Chen et al., 2003; Kollet and Zlotnik, 2005; Cheng and Chen, 2007), and at three test sites in the Republican River valley, Nebraska (Chen et al., 1999). During the pumping, the release of water from the zone above the declining water table moves predominantly downward. Thus, the magnitude of \( K_v \) can largely affect the release process of gravity drainage. This can be seen from the mathematical expression used to describe the instantaneous release of water from the pores

\[
K_v \frac{\partial h}{\partial z} = -S_y \frac{\partial h}{\partial t}
\]

where \( K_v \) is the vertical hydraulic conductivity for the aquifer materials near the water table, \( S_y \) the specific yield, \( h \) the hydraulic head, and \( z \) is the vertical axis of the cylindrical coordinate. Eq. (1) assumes that vertical infiltration (recharge) does not exist and the horizontal hydraulic gradient is very small compared to the vertical hydraulic gradient near the water table. Regardless of whether the water release process in the zone of the moving water table is instantaneous or delayed, the association of the vertical hydraulic conductivity and the specific yield can not be neglected. According to Chen et al. (1999), the correlation between \( K_v \) and \( S_y \) can be up to 0.59 for a pumping test conducted in the Republican River valley, Nebraska. Thus, overly simplified treatment of \( K_v \) during the analysis of pumping tests data may bring uncertainties to the estimates of other hydraulic parameters. Therefore, careful examination of \( K_v \) may also provide useful information for the determination of \( S_y \). As has been reported by other researchers, \( K_v \) varies from one to another observation location at a pumping test site. The difference between the minimum and maximum \( K_v \) values within a pumping test site can vary by a factor of 5 (Chen et al., 1999; Kollet and Zlotnik, 2005). In the alluvial sediments below the Platte River, Nebraska, Chen et al. (2008) reported that \( K_v \) values at varied depths can differ by several orders of magnitude.

The objective of this study was to determine spatial variation of \( S_y \) and \( K_v \) for an unconfined alluvial aquifer at a test site in the Platte River valley of Nebraska. The methods include sediment coring and laboratory experiments on sediment cores. Continuous sequences of sediment cores were collected near each of six observation wells of a pumping test site. These sediment cores were collected in a fashion with very little disturbance of sedimentary structures. The drainage method was used to determine the specific yield and permeameter tests were conducted to determine the vertical hydraulic conductivity of these sediment cores. This paper analyzes the variations of \( S_y \) and \( K_v \) for several vertical profiles of up to 21 m in depth. In addition, the relationship between the specific yield and the vertical hydraulic conductivity is analyzed. The effectiveness of the vertical hydraulic conductivity on the drainage process of sediment cores is also discussed.

2. Methods

2.1. Study site

The study site is in the Platte River valley of Nebraska where the alluvial sediments consist mostly of Quaternary unconsolidated...
sand and gravel. This alluvial aquifer extends more than 500 km in this valley from west to east across Nebraska. A large number of irrigation wells have been constructed in the aquifer to pump groundwater for agricultural production. The study site is about 5 km southwest of Chapman (Fig. 1); it is about 6 km northeast from the pumping test site of Wenzel (1942). The saturated thickness of the alluvial aquifer is about 24.0 m. The depth to the water table is about 1.7 m. Thus, the thickness of the unsaturated zone compared to the saturated aquifer is relatively small. At this site, a pumping test was conducted using a pumping well and six observation wells. Sediment cores were collected near these observation wells for this study. Fig. 2 shows the orientation of these observation wells and the distances between these wells.

2.2. Electric conductivity logging and sediment coring

Direct-push equipment was used to generate electrical conductivity logs and to collect sediment cores at locations co-located with the observation wells in the study site. Direct-push equipment employs both static force and percussion to force sampling and logging tools into the subsurface. A Geoprobe® Systems SC400® soil conductivity probe was used in this investigation. During the penetration of the sediments, the probe can measure the electrical conductivity (EC) of the porous matrix surrounding the probe to a radius of about 5–10 cm. In the saturated zone, pore fluid and matrix properties are the major factors affecting electrical conductivity. Sand and gravel with dominant particles such as quartz, mica, and feldspar are generally nonconductive and thus have a lower EC value. In contrast, clay-sized particles, such as phyllosilicates, humic substances, and iron and manganese oxides and oxyhydroxides, tend to be highly conductive and have relatively large EC values (Schulmeister et al., 2003). EC logs are thus useful in identifying hydrostratigraphic units. A total of six EC logs were produced from the direct-push technique. The logging location was co-located with each observation well.

After an EC log was produced, a Geoprobe® Systems MacroCore® soil sampler was used to collect sediment cores of the alluvial aquifer materials about 1 m from the EC spot. This sediment sampler collected the sediment core in a plastic liner, which was about 1.5 m in length and 4.2 cm in inner diameter. First, the Macro-Core sampler was advanced 1.5 m beneath the land surface to collect the first sediment core. The sampler was then retracted to the surface. The core was removed from the sampler, and the two ends of the core were sealed by caps to prevent water from escaping from the core. A new plastic liner was inserted into the sampler, which was then pushed to the next depth interval of 1.5 m to collect a second sediment core. Repeating this sampling procedure produces a continuous sequence of sediment cores. The plastic liner is transparent; the laminations and beds in the sediments can be visually examined and were cross-checked with the EC log in the field. A continuous sequence of sediment cores was collected near each of these observation wells. EC logging and sediment coring reached to the depth of about 21–24 m below the land surface.

2.3. Permeameter test for determination of Kv

The sediment columns collected in transparent plastic tubes were transferred to the laboratory for measuring specific yield and vertical hydraulic conductivity. Permeameter tests were conducted first to determine the $K_v$ values. The testing approach followed the principles of falling head permeameter tests. Prior to the permeameter tests, the sediment cores were resaturated by submerging them in a water tank. During the test, the plastic sediment core liner acted as a container for the permeameter tests. Thus, it avoided the re-packing process that often occurred in permeameter tests in the laboratory. Re-packing of sediments destroys the original orientation of sediment particles and bedding structures. The sediment core tube was held vertically using a tripod. The bottom end of the tube was covered by several layers of fine aluminous screen to prevent sediments from falling through the lower end of the tube. A water bucket fully filled with water was placed under the sediment core. Water was continuously
added to fill the upper end of the tube until it was full. The water inside the tube started to fall as soon as water stopped being added to the tube. A series of head readings were recorded. The test will stop before the water falls below the top of sediments within the tube. The \( K_v \) was calculated based on Darcy’s Law such that

\[
K_v = \frac{L_v}{(t_2 - t_1)} \ln(h_1/h_2).
\]

where \( L_v \) is the length of sediment core in the tube; \( h_1 \) and \( h_2 \) are hydraulic head above the edge of the water bucket recorded at times \( t_1 \) and \( t_2 \), respectively. A permeameter test on cores of sand and gravel often takes about 10–15 min. A permeameter test for low-\( K \) cores can take hours or days. For each test, all the available readings (more than two readings) of \( h \) were used simultaneously for the computation of the \( K_v \) value.

2.4. Draining method for estimation of \( S_y \)

After the permeameter tests, the sediment cores were resaturated again by placing them in a water tank. After being saturated, the top end of the tube was covered by a rubber cap to disconnect it from the atmosphere. Then, the tube with sediments was moved from the tank and held vertically in a bracket. A dry transparent plastic bucket was placed under the sediment core. The rubber cap that covered the top of the tube was removed and the top opening was sealed with several layers of plastic membrane punctured with small holes to prevent water loss by evaporation from the top end of the sediment core but to maintain the atmospheric pressure. In addition, the top of the water bucket was also sealed with several layers of plastic membrane to prevent water loss by evaporation. In order to make the water fully drain from the sediments under the force of gravity, the drainage period for each sediment column continued 120 h or longer even after the water depth in the bucket became constant. However, the sediment columns with a part or full tube of clay or of clay and silt mixtures were hardly suitable for specific yield tests because the sediments had a very low hydraulic conductivity and the volume of drained water was too small to be determined accurately. During the drainage experiment, the water depth in the bucket was recorded. If the water level in the bucket kept constant for several days, the drainable water from the sediment cores was considered fully released.

In order to prevent the build-up of a capillary fringe at the lower end of the sediment column during the draining process, the tube was held vertically and its lower end was several centimeters above the bottom of the water bucket. Thus, there was no contact between the drained water and the lower end of the sediment core.

For most cores, the sediments did not fully fill the entire length of the tube during the coring process. During the re-saturation process, some water entered above the top of the sediment core. This amount of water was drained to the water bucket during the draining experiment and it slowed the draining process only slightly because that amount of water was usually small and most of the sediment cores were relatively permeable. Fig. 3 shows the simple devices used for the draining experiments. After the gravity drainable water was fully released from the sediment core, the water in the bucket was poured into a measuring cup to determine the water volume drained. The specific yield was calculated from

\[
S_y = \frac{V_{sw} - V_{sr}}{V_s},
\]

\[
V_{sw} = \pi r^2 L_w
\]

\[
V_s = \pi r^2 L_v
\]

where \( V_{sw} \) is the water volume in the bucket; \( V_{sr} \) the water volume above the sediments in the tube; \( V_s \) the sample volume; \( r \) the radius of the tube; \( L_w \) the depth of water above the sediments in the tube; and \( L_v \) the length of sediments in the tube (see Fig. 3).

3. Results and discussion

3.1. \( S_y \) values determined from drainage experiments

\( S_y \) values were determined from six sets of sediment cores using the drainage method. There were two core sets collected near OW-1. The first set has 14 cores; the second set has six cores. The purpose of using two core sets near OW-1 was to see spatial variability of \( S_y \) within a relatively small distance. Fifteen cores were collected near OW-5. However, only one core from this core set was used for determination of \( S_y \). After the permeameter tests, most cores near this well and some cores near OW-4 were cut into short segments to re-run permeameter tests for examination of vertical variation of \( K_v \) in short intervals (the results for the shortened cores are not reported in this paper). Table 1 summarizes the number of sediment cores for each core sequence, the range of the thickness of measured sediments formed in the tube, and the specific yield values. The average values of \( S_y \) for the sediment cores vary from 0.06 to 0.12 (Table 1).

Fig. 4 shows the \( S_y \) profile for these six sets of cores. As shown in Fig. 4, the value of specific yield varies among each sediment core. \( S_y \) values for sediment cores collected near OW-3 and OW-6 with the depth >3.0 m are all greater than 0.05 (Fig. 4d and f). In contrast, about half of the sediment cores collected near OW-1 and OW-2 have \( S_y \) values smaller than 0.05 (Fig. 4a–c). As shown in Fig. 2, the horizontal distance between OW-2 and OW-3 is only 12.2 m, yet, the vertical profiles of \( S_y \) from the two locations have quite different distribution patterns.

Near OW-1, two sequences of sediment cores were collected. The first sequence reached to a depth of 21 m and the second to a depth of 9 m. The separation between two coring locations was only about 2 m. The specific yield varied from 0.01 to 0.15. In the longer core sequence, the values of \( S_y \) for sediments from the depth of 7.5–12 m below the land surface were smaller than \( S_y \) values of cores from other depths (see Fig. 4a). The core from the depth of 6–7.5 m had the largest \( S_y \) value (\( S_y = 0.15 \)). The second sediment core sequence has only six cores. The vertical profile of \( S_y \) in this core has a similar distribution pattern to that of the first core sequence for the same depth. In the second core sequence, the core
from the depth of 6–7.5 m also had the largest $S_y$ value ($S_y = 0.18$). Although the two cores are only about 2 m apart, the $S_y$ values of sediment from the same depth can still differ slightly. The average $S_y$ value for the top six cores in the first sequence was 0.071, and the average $S_y$ for the second core was 0.071 as well.

For sediment cores near OW-2 (Fig. 4c), the $S_y$ has a wide range from 0.01 to 0.13. For the sediment cores from the depth of 3–21 m, five of the 11 cores have the $S_y$ values smaller than 0.05. The two cores from the depth of 3–4.5 m and 6–7.5 m are greater than 0.1. The average $S_y$ value is 0.06 and standard deviation is 0.043.

For sediment cores collected near OW-3 (Fig. 4d), the $S_y$ for sediment cores from depths greater than 3 m were greater than 0.05. Compared to the $S_y$ values for cores near OW-2, the $S_y$ values from the cores near OW-3 varied less with depth. The average $S_y$ near OW-3 was 0.087 and standard deviation was 0.028. Standard deviation reflects the dispersion of data with respect to the mean value. The standard deviation for these $S_y$ values was smaller than the standard deviation of $S_y$ from the cores near OW-2.

A total of 15 sediment cores were collected near OW-4, but draining experiments were conducted only on five cores. Their $S_y$ values are shown in Fig. 4e. The largest $S_y$ value is for the core from

<table>
<thead>
<tr>
<th>Coring sites</th>
<th>Thickness of sediment cores (m)</th>
<th>Specific yield</th>
<th>Number of tested sediment cores</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Range</td>
<td>Average</td>
<td>Range</td>
</tr>
<tr>
<td>OW-1</td>
<td>0.26–1.01</td>
<td>0.59</td>
<td>0.01–0.15</td>
</tr>
<tr>
<td>OW-1</td>
<td>0.33–1.07</td>
<td>0.7</td>
<td>0.02–0.18</td>
</tr>
<tr>
<td>OW-2</td>
<td>0.42–0.86</td>
<td>0.65</td>
<td>0.01–0.13</td>
</tr>
<tr>
<td>OW-3</td>
<td>0.69–1.36</td>
<td>1.05</td>
<td>0.02–0.12</td>
</tr>
<tr>
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<td>0.02–0.09</td>
</tr>
<tr>
<td>OW-6</td>
<td>0.49–0.98</td>
<td>0.72</td>
<td>0.06–0.13</td>
</tr>
</tbody>
</table>

Table 1
Thickness of sediment cores and ranges of specific yield values for each location.
the depth of 3–4.5 m and it is 0.09. A total of 15 cores were collected near OW-5, but a drainage experiment was conducted for one core for those collected near OW-5. The $S_y$ value for the core at the depth of 3–4.5 m was 0.12. All of these cores collected near OW-4 and OW-5 were tested for $K_v$ values, which will be presented in a later section.

For sediment cores near OW-6, the $S_y$ varies from 0.06 to 0.13. In the saturated zone, $S_y$ varies from 0.07 to 0.13, with the average of 0.12 and standard deviation of 0.02. The largest $S_y$ value was for the sediments at the depth of 3–4.5 m and was 0.13. Among the $S_y$ values for the cores below the depth of 3 m, eight out of the 11 $S_y$ values were greater than 0.1.

Among the 62 $S_y$ values determined from the draining experiments, two of them were smaller than 0.01, 17 were between 0.01 and 0.05, 23 were between 0.05 and 0.1, 19 were between 0.1 and 0.15, and only one was greater than 0.15 (Fig. 5). Performance of Jarque–Bera test of normality (Jarque and Bera, 1987) using MATLAB indicated that these $S_y$ values are normally distributed at the 95% confidence level. $S_y$ values for the cores 0–1.5 m below the land surface were generally small because of a high content of clayey soil.

### 3.2. $K_v$ values of sediment cores

$K_v$-depth profiles for six sequences of sediment cores collected from spots collocated with observation wells of a pumping test. The values of $K_v$ vary between cores. The variation can reach several orders of magnitude.

![Fig. 6. $K_v$-depth profiles for six sequences of sediment cores collected from spots collocated with observation wells of a pumping test.](image-url)
Among the 73 $K_v$ values determined from the sediment cores near each of the observation wells (cores from the depth of 0–3 m are excluded for this analysis), the average value is 4.7 m/d, and standard deviation is 4.5 m/d. 45 out of 74 $K_v$ values fall in the range of 1–10 m/d. Fig. 7 shows the number of cores in different $K_v$ ranges. Performance of Jarque–Bera test of normality in MATLAB indicated that the 73 $K_v$ values showed neither a normal distribution nor a log-normal distribution. Sudicky (1986) showed that the hydraulic conductivity of sediment samples for the Borden aquifer was a log-normal distribution. Note that the values of hydraulic conductivity reported in Sudicky (1986) do not necessarily represent the vertical hydraulic conductivity because of sample remixing during the permeameter tests. In contrast, the hydraulic conductivity in this study is the vertical hydraulic conductivity. The statistical distribution of horizontal or vertical hydraulic conductivity for an aquifer may be site specific and it deserves further study.

Correlation coefficients were calculated for the cores from which both $S_y$ and $K_v$ were determined. The correlation between $S_y$ and $K_v$ was 0.495. After the natural logarithmic transform, the correlation coefficient between $\ln(S_y)$ and $\ln(K_v)$ increased to 0.595. Fig. 8 is the plot of $\ln(K_v)$ versus $\ln(S_y)$. Here, the value of $S_y$ is expressed as a percentage (%). The correlation between the two variables was higher for some locations. For example, the correlation coefficient for sediments near OW-6 was 0.62. After logarithmic transform of $S_y$ and $K_v$, the correlation coefficient reached 0.8 at this location.

The coring techniques have some limitations in producing the $S_y$ and $K_v$ profiles. The lengths of the sediment cores for sand and gravel are often smaller than the length of the sampling liner. In other words, part of the sediments was not recovered during the coring process. For fine sand, silt, and clay, the sediment recovery in the tube was nearly 100%. Sand and gravel seem to have higher tendency to be lost during coring. As a result, the test results could possibly bias to the low side for some cores. Nevertheless, the $S_y$ and $K_v$ profiles generated in this study show a strong spatial
variation in this alluvial aquifer, and the values of $K_v$ range in several orders of magnitude.

3.3. Water release process of sediment cores

In the laboratory drainage experiments, the drainage times for individual cores were recorded. Fig. 9a shows the drainage process for the core at the depth of 3–4.5 m near OW-2. The sediment core was 86.6 cm long. The determined $S_p$ value for this core was 0.13 and the $K_v$ value was 3.2 m/d. After 2.2 h of drainage in laboratory, about 77% of the water within the sediments drained. The water level in the bucket was not recorded between 2.2 and 24 h after the test began. But after 24 h of drainage, 100% of the drainable water by gravity from the core was released. Thus, the time of duration for the full drainage can be less than 24 h. Before the test, the water that remained above the top of the sediment core was 0.6 cm in thickness. This small amount of water can prolong the drainage process slightly.

The length of sediment core may affect the drainage process. For the sediment core at the depth of 6–7.5 m near OW-1, its length was 100 cm, and the $S_p$ was 0.15. Fig. 9b shows the drainage process. After 2 h, about 76% of gravity drainable water was released to the water bucket and after 17.3 h, 94% of gravity drainable water was released to the bucket. However, the last 6% of water took a much longer time to drain. After 41.5 h since the test began, 100% of the water was released from the core (Fig. 9b). The $S_p$ was 0.15 and $K_v$ was 4.3 m/d for this core. Prior to the draining test, the water column that remained above the top of the sediment core was 1.7 cm.

For the sediment core from the depth of 4.5–6.0 m near OW-6, the length of the core was 74.6 cm, and the $S_p$ value was 0.13. After 1.2 h of drainage, 76.5% of gravity drainable water was released. After 17.5 h, 100% of the water was released. Readings of the water level were not recorded between 1.2 and 17.5 h. Before the test, the water column that remained above the top of the sediment core was 0.6 cm.

Slow drainage occurred in some sediment cores. The core from the depth of 9–10.5 m near OW-6 was 84.5 cm long. After the first 26.3 h of drainage, only was 37.5% of drainable water was released (Fig. 9c). After 71.5 h, the percentage of drained water was 75%, and after 113.7 h, 100% of drainable water was released. The $S_p$ was 0.07 and $K_v$ was 0.38 m/d. This slow drainage process is apparently due to a small vertical hydraulic conductivity value. Before the test, the water column that remained above the top of the sediment core was 0.2 cm.

For a comparison, the core from the depth of 1.5–3 m near OW-1 was 86.4 cm in length and had $S_p$ of 0.07. After 2 h of drainage, 76.9% of drainable water was released. The next measurement was 17.3 h after the test began. The released water accounted for 84.6% of the drainable water (Fig. 9d). At the time of 41.3 h after the test, 100% of water had been released. The $K_v$ of this core was 1.44 m/d. Before the test, the water column that remained above the top of the sediment core was 4 cm.

Generally, after several hours of drainage, more than 50% of drainable water was released from the sediment cores. For the sediment cores near OW-6, the released water accounted for 38–100% for nine of 11 cores during the early drainage time of 0.4–2.2 h (Fig. 10). The average length of sediment cores from the location near OW-6 was 72 cm. The above test results suggest that the appreciable effect of delayed yield in this permeable alluvial aquifer is unlikely to last longer than 24 h. According to the sensitivity analysis of the specific yield by Chen et al. (2003), a 24 h pumping test using a high capacity pumping well seemed to provide good data for the determination of reliable specific yield in a permeable alluvial aquifer.

3.4. Sedimentary properties of the alluvial aquifer

Seven sediment samples from the saturated zone, which were collected during the test-hole drilling at OW-1, were sent to Geo-technical Services Inc. in Grand Island, Nebraska for grain size analysis. Table 2 lists the percent of gravel, sand and silt–clay by weight, as well as the porosity and sorting. In the sieving, 0.075 mm was used as a particle grade. Here for a convenience, particles <0.075 mm were placed into the silt–clay group. In sediment text books, for example in Davis (1983), particles <0.0625 are classified into silt and clay. The grain-size distribution data indicate that the alluvial aquifer at this site consist largely of sand and gravel with a small amount of silt–clay (0.6–13.2% with an average of 6.6%). A sample from the depth of 0–1.5 m (in the unsaturated zone) consists of 57.1% of sand and 42.9% of silt and clay (<0.075 mm). This sample does not contain gravels.

The porosities of the sediment samples were derived based on the grain-size distribution data (Vukovic and Soro, 1992). The average porosity of the seven sediment samples is 0.31 (Table 2). The pumping test site of Wenzel (1942) is about 6 km from our study site, and the average porosity determined from 19 sediment samples was reported to be 0.29 at the site.

The sediment porosity reflects the storage capacity of water in the aquifer. However, the magnitude of specific yield is dependent on both porosity and field capacity (or specific retention) (Nachabe, 2002). Field capacity of sediments indicates the amount of water which is not able to drain from the sediment under the action of gravity. According to Ratliff et al. (1983) and Meyer and Gee (1999), the value of field capacity is about 0.12 for sand and 0.19 for loamy sand. If we use the value of 0.15 as field capacity, the specific yield for these sediment samples will be 0.16 (porosity minus field capacity). Additionally, the in situ sediments in the

| Table 2 |
| Porosity, sorting, and percent of sediments in three grain-size groups by weight. |
| --- | --- | --- | --- |
| Depth (m) | >2.0 mm | 2–0.075 mm | <0.075 mm | Porosity | Sorting |
| 10.7–12.2 | 18.3 | 75.1 | 6.6 | 0.33 | MS |
| 12.2–13.7 | 30.9 | 68.5 | 6.0 | 0.32 | PS |
| 13.7–15.2 | 23.4 | 71.4 | 5.2 | 0.33 | PS |
| 15.2–16.8 | 19.6 | 67.2 | 13.2 | 0.29 | MS |
| 16.8–18.3 | 20.2 | 70.1 | 9.7 | 0.3 | MWS |
| 18.3–19.8 | 22.4 | 69.2 | 8.4 | 0.3 | MS |
| 24.4–25.6 | 31.2 | 66.2 | 2.6 | 0.32 | PS |
| Average | 23.7 | 65.7 | 6.6 | 0.31 | |

* MS = moderately sorted.  
* PS = poorly sorted.  
* MWS = moderately well sorted.
aquifer are compacted, but drilling destroyed the original lamination and beddings in the sediment and made the sediment samples loose. There is a possibility that the porosity derived from sieving analysis may be slightly larger than the real value of the in situ sediments. The level of sediment sorting can affect the magnitude of pore size in sediments, which affects hydraulic conductivity and specific yield. Davis (1983) proposed seven sorting classes of sediments based on standard deviation of grain-size distribution. We used a graphic standard deviation (Davis, 1983, p. 10) for the determination of sorting classes. Among the seven sediment samples, one sample is moderately well sorted, three samples moderately sorted, and three poorly sorted (Table 2).

Electrical conductivity logs of the alluvial aquifer generated near OW-2, OW-3, OW-5 and OW-6 are shown in Fig. 11. The EC values are 20–30 mS/m for sand and gravel and usually >40 mS/m for silt and clay at this site. These EC logs suggest that this alluvial aquifer contains thin layers or lenses of fine-grained sediments (silt and clay). The two EC logs near OW-2 and OW-3 (Fig. 11a and b) exhibit up to five spikes with EC values >40 mS/m in the depth of 2–14 m. The big spike below the depth of 26 m in Fig. 11b is the clay at the base of the aquifer. Although the distance between OW-2 and OW-3 is only 12.2 m (see Fig. 2), the pattern of the two EC logs is not the exactly same, indicating some variation of sedimentary structure between the two locations. Fig. 11c and d show the EC logs of the sediments near OW-5 and OW-6. In contrast to OW-2 and OW-3, the two EC logs near OW-5 and OW-6 display a fewer spikes of silt and clay, indicating less heterogeneity in lithology. These two wells are located along the south–north transect of the observation wells (see Fig. 2). The number, depth and thickness of silt and clay layers as indicted by the EC spikes vary at each location, indicating that an individual silt/clay does not continue widely in the lateral directions. All these EC logs suggested thin silt and clay layers occur within the sand/gravel-dominated alluvial aquifer. These small-scale silt/clay lenses can reduce vertical hydraulic conductivity of the aquifer and slow down the drainage process if the aquifer is to be dewatered.

4. Conclusions

Our study confirms that $S_v$ varies from one to another location in both the horizontal and vertical direction in this alluvial aquifer. The most common $S_v$ values are from 0.05 to 0.15. Thus, it is an overly simplified approach to use a single $S_v$ value to represent the capacity of the extractable water in an unconfined aquifer. In the literature, an imperfect match between calculated and observed data has (for example, drawdowns) been often attributed to flawed models or bad data used for the analysis of pumping test data. However, the effect of a strong variation of $S_v$ and $K_v$ among the observation locations...
may have complicated the analysis of the pumping test data. Therefore, it is common that a curve fitting between observed and calculated drawdowns (using analytical solutions) shows a good agreement at one observation well but becomes poor at another location if single values of $S_h$ and $K_v$ are used for all the observation wells. It is thus recommended that the heterogeneity of $S_h$ and $K_v$ be considered in the analysis of pumping tests.

Because of the high cost for the construction of an observation well, a pumping test often has a limited number of observation wells. Thus, pumping tests are not an economic means for determination of detailed spatial variation of $S_h$. Coring sediments using the direct-push technique and running drainage tests on these cores without re-packing gives a promising approach to characterize the spatial variation of specific yield. Sediment cores can be collected at varied depths and give a more complete picture about the small scale variation of $S_h$ in a three-dimensional field.

$K_v$ also shows a strong variation from one to another observation location among these sediment cores. The $K_v$ values fall in the range of several orders of magnitude. Researchers have reported that hydraulic conductivity (mostly for the horizontal hydraulic conductivity) has a log-normal distribution. However, the $K_v$ values at this test site exhibit neither log-normal distribution nor normal distribution. Statistical distribution of hydraulic conductivity deserves further study.

$K_v$ has some correlation with $S_h$ for individual cores. At one location, the correlation coefficient is 0.62, which is the highest correlation coefficient for the tested cores. The draining experiment suggested that the magnitude of $K_v$ affected the pace of drainage from the sediment cores. Thus, in the analysis of pumping test data, holding the $K_v$ value constant for different locations is an approach that overly simplifies the complex distribution nature of this parameter. It will introduce errors in the estimated $S_h$ values. In addition to examining the delayed release process, careful examination of spatial variation of $S_h$ as well as $K_v$ is needed in the determination of aquifer hydraulic parameters using pumping test data. Because $K_v$ is highly heterogeneous, it can be anticipated that the ratio of horizontal to vertical hydraulic conductivity is heterogeneous as well in an alluvial aquifer. While the alluvial aquifer in the study site is highly permeable, the authors anticipate that $K_v$ and $S_h$ in a less permeable alluvial aquifer are heterogeneous as well.

For permeable sediments, more than 50% gravity drainable water can be released in a couple of hours. The above test results suggest that the appreciable effect of delayed yield in such a permeable alluvial aquifer is unlikely to last longer than 24 h. A pumping test lasting for several days gives a sufficient time for the water above the declining water table to be released by the action of gravity.

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