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Understanding of the Hydrologic Connections Between Wide-channel and Adjacent Aquifers Using Numerical and Field Techniques

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Understanding of the Hydrologic Connections Between Wide-channel and Adjacent Aquifers Using Numerical and Field Techniques

by

Cheng Cheng

A DISSERTATION

Presented to the Faculty of

The Graduate College at the University of Nebraska

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Streambed sedimentary structure plays a vital role in controlling the interactions between surface water and groundwater. In the study, three rivers—the Big Blue River, the Little Blue River, and the lower reach of the Platte River in Nebraska were chosen to characterize the shallow streambed for the two types of rivers (braided and meandering rivers) and investigated the variations of the streambed electrical and hydraulic conductivities with depth. In-situ and laboratory permeameter tests were conducted to determine streambed hydraulic conductivity up to 20 m below the channel surface in the three rivers. Additionally, the electrical conductivity logs were obtained using Geoprobe direct-push technique to characterize the hydrostratigraphy of streambed sediments.

Although the tributaries of the Big Blue River have low-permeability sediments lining beneath the stream bottom which generate smaller \( K_v \) values, the \( K_v \) values in the top 1-m of the streambed sediments are usually greater than 5 m/d in the three rivers, indicating very permeable streambeds. Therefore, shallow streambeds are permeable over the gaining reaches of braided and meandering rivers despite their differences on the watershed size, channel width, topographic reliefs, etc. In addition, the Big and Little
Blue Rivers have more fine-grained sediments deposited in the deep streambed than the Platte River. Furthermore, streambed $K_v$ values in the three rivers exhibit a tendency to decrease with depth in the depth of 0 to 6 m below the channel surface.

The constant head boundary is proposed to be an alternative solution in the simulation of stream-aquifer interactions. This approach is applied in a regional groundwater flow model to evaluate the impact of groundwater irrigation on the streamflow in the lower reach of the Platte River. Additionally, the model provides an accurate estimation of the streambed leakage of the Platte River using numerical and field techniques. Furthermore, the statistical distribution of $K_v$ values of shallow streambed sediments along a 300-km segment of the Platte River is also examined. It was found that they are normally distributed; this finding differs from the widely accepted concept that hydraulic conductivity in aquifers is log-normal distribution.
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Preface

The dissertation discusses the variations of streambed electrical and hydraulic conductivities in three rivers (the Platte River, the Big Blue River, and the Little Blue River) of Nebraska. Also, the statistical distribution of streambed hydraulic conductivity of the shallow streambed in the Platte River of Nebraska is investigated. Finally, a three-dimensional groundwater flow model is developed to evaluate the effects of groundwater pumping on the streamflow in the lower reach of the Platte River. The dissertation is organized as follows:

(1) The first chapter summarizes the previous research on two different channel patterns, braided and meandering rivers, and their differences in shape and sediment load and typical streambed hydraulic conductivities \( K \). Then, the importance of streambed \( K \) in the study of the interactions between the surface water and groundwater is noted, and the possible effects of hyporheic processes on the shallow streambed \( K \) are demonstrated. Finally, this chapter proposes five different hypotheses for the dissertation, and the methods to testify these hypotheses are introduced.

(2) The second chapter characterizes the variations of streambed electrical and hydraulic conductivities with depth in three rivers (the Big Blue River, the Little Blue River, and the lower reach of the Platte River) in Nebraska. The Big and Little Blue Rivers are examples of the meandering rivers, whereas the Platte River is an example of the braided river. The electrical conductivity logs were generated by Geoprobe direct-push technique, and streambed hydraulic conductivities were obtained from in-situ and laboratory permeameter tests. The characterization helps identify whether a low-permeability clogging layer exists at the channel surface for both the meandering and
braided rivers, existence of which was assumed in many modeling studies. Also, it helps understand the hydrostratigraphy of streambed sediments and streambed hydraulic conductivities in the deep streambed of the three rivers. This chapter testifies the first, second, and fourth hypothesis in the dissertation.

(3) The third chapter discusses the statistical distribution and spatial variation of vertical hydraulic conductivity ($K_v$) of the shallow streambed at 18 different sites along a 300-km segment of the Platte River. At each site, 20 to 200 permeameter tests were conducted to determine streambed $K_v$ of the shallow channel sediments, and then the normality tests were performed on these streambed $K_v$ values to determine if they are normally distributed at each site. The approach testifies the fifth hypothesis in the dissertation. Furthermore, the effects of the tributaries of the Platte River on the streambed $K_v$ variability are addressed.

(4) The fourth and final chapter presents a regional three-dimensional groundwater flow model which is developed to simulate the interactions between the lower reach of the Platte River and its adjacent aquifers in light of intensive groundwater irrigation. The field measurements of streambed hydraulic conductivities were incorporated in the model. The regional model provides insights in the evaluation of streamflow depletion of the Platte River due to groundwater irrigation; it helps estimate the value of streambed leakance, which is useful in assessing the stream-aquifer interactions when an analytical solution is employed.
Chapter 1  Introduction

1.1 Stream Channel Patterns and Their Streambed Hydraulic Conductivities

Braided and meandering rivers are two typical types of channel patterns, and meandering rivers are more common (Ikeda and Parker, 1989). The braided rivers usually have a high stream gradient and an abundant supply of bed load sediment (Leopold and Wolman, 1957). This type of stream channel tends to be wide and shallow, and the stream banks are easily to be eroded. In contrast, the meandering rivers usually have deep and narrow channels, and the stream banks are resistant to erosion (Leopold and Wolman, 1957). Also, floodplain (overbank) deposits were better deposited in the meandering rivers (Schumm and Kahn, 1972). The sediment load is primarily carried in bed load for braided rivers whereas the sediment load is primarily carried in suspended load for meandering rivers, and an increase in suspended sediment can allow for the deposition of fine-grained material (Schumm and Kahn, 1972). Consequently, the streambed sediments are associated with coarser materials in the braided rivers but with finer materials in the meandering rivers, thereby leading to a higher streambed hydraulic conductivity for the braided river than that for the meandering river (Kondolf et al., 1987). Brunke and Gonser (1997) reviewed the connectivity between river and groundwater ecosystems, and they noted that even the intense hydrological interactions may take place in the meandering river segments, the fine particulate load can cause clogging of the sediments. In contrast, a braided river system has a high transport capacity for bed load, and the rapid lateral channel migration can induce high permeability of the sediments.

Schubert (2006) noted the clogging process caused by the operation of riverbank filtration wells is present in the Rhine River of Germany, which is a meandering river.
Thus, the riverbed sediments are considered impermeable. Furthermore, the Mississippi River is a classic meandering river, which is an example of a suspended load fluvial system (Larkin and Sharp, 1992), thereby having a low riverbed hydraulic conductivity. Ruhl and Cowdery (2004) calibrated the streambed vertical hydraulic conductivity \( (K_v) \) of the Mississippi River as 0.02 m/d between Brainerd and Little Falls in central Minnesota.

The Gash River in eastern Sudan is considered as a braided river (Alredaisy, 2011). The fluvial deposited sediments in the Gash River consist of coarser materials (sand and gravel) on the upstream and finer materials (clay) on the downstream (Alredaisy, 2011). Alredaisy (2011) also noted that the hydraulic conductivity of the bed load sediments is about 36 to 105 m/day. Additionally, the lower reach of the Yellow River in China is also a braided river (Xu, 1996). Shu et al. (2005) reported that the streambed horizontal hydraulic conductivity \( (K_h) \) is about 1.81 m/day and streambed \( K_v \) is from 0.19 to 0.71 m/day at different test sites in the Yellow River using in-situ standpipe tests.

In the contiguous United States, the Platte River is a notable example of braided river (Huntzinger and Ellis, 1993). Over the past 10 years, numerous permeameter tests have been conducted in determining streambed \( K_v \) in the Platte River of Nebraska (Landon et al., 2001; Chen, 2004; Chen et al., 2008; Cheng et al., 2011), which revealed that the shallow streambed of the Platte River is usually permeable. Hence, the braided rivers are generally characterized by coarse bed load sediments, and finer materials may be deposited locally onto coarser sediments downstream or during a flood event; whereas the meandering rivers are more characterized by fine-grained sediments.
1.2 Importance of Streambed $K_v$ on Groundwater-Surface Water Interactions

A stream may either contribute water to the corresponding aquifer, or act as a groundwater discharge zone when a hydraulic gradient occurs between the stream and the groundwater system depending upon the direction of that gradient. Streambed conductance has been shown to be the most important hydraulic parameter in modeling stream-aquifer interactions (Sophocleous et al., 1995). Streambed conductance is defined as follows (McDonald and Harbaugh, 1988):

$$C_{riv} = \frac{K_v L_{riv} W}{M}$$

where $K_v$ is the vertical hydraulic conductivity of the streambed sediments, $L_{riv}$ is the length of the stream channel, $W$ is the width of the stream channel, and $M$ is the thickness of the streambed sediments.

The flow between the stream and the corresponding aquifer at a certain location is equal to:

$$Q = C_{riv} (h_s - h_a)$$

where $Q$ is the flow between the stream and the aquifer, $h_s$ is the stream level, and $h_a$ is the hydraulic head in the aquifer. A positive value of $Q$ means that the stream contributes water to the aquifer, whereas the stream gains water from the aquifer when the value of $Q$ is negative.

Streambed $K_v$ is a key parameter to know or determine when quantifying stream-aquifer interactions. In general, a higher rate of streamflow depletion is a result of a higher streambed conductance, a thinner streambed, and more permeable stream sediments, or the combination of these three components (Chen and Shu, 2002). Sun and Zhan (2007) noted that one of the most important factors controlling the interaction of an
aquifer with two parallel streams is the hydraulic conductivity ratio of the two streambeds, especially when a low-permeability streambed exists. Moreover, streambed $K_v$ heterogeneity could affect hyporheic zone fluxes and groundwater discharge (Salehin et al., 2004; Kalbus et al., 2009; Kennedy et al., 2009). Therefore, knowledge of streambed $K_v$ is essential to characterize hydrologic connections between a stream and its adjacent aquifers, and better understanding of the hydrostratigraphy of streambed sediments is beneficial to the integrated water resources management.

1.3 Streambed Hydraulic Conductivity at Channel Surface

1.3.1 Assumption of Clogging at Channel Surface in Modeling Groundwater-Surface Water Interactions

Streambed conductance is used widely to integrate the interactions between surface water body and groundwater system in modeling studies, which represents the resistance to flow between the surface water body and the groundwater caused by streambed (McDonald and Harbaugh, 1988). The concept is usually based on the assumption that there is a low-permeability clogging layer (or semi-permeable layer), the hydraulic conductivity of which is smaller than that of the underlying aquifer at the channel surface of the streambed sediments (Figure 1; Sophocleous et al., 1995; Hunt, 1999; Osman and Bruen, 2002; Akylas and Koussis, 2007; Rushton, 2007; Hu et al., 2007; Sun and Zhan, 2007; Intaraprasong and Zhan, 2009).
Figure 1.1 Diagrams from previous modeling studies of stream-aquifer interaction showing the occurrence of a clogging layer (low-permeability layer) at the near-channel and in-channel interface presented by (a) Hunt (1999); (b) Osman and Bruen (2002); (c) Hu et al. (2007); and (d) Rushton (2007).

The clogging process is considered ubiquitous in the streambed and reduces the water exchange at the sediment-water interface during low-flow conditions (Brunke and Gonser, 1997; Battin and Sengscmmitt, 1999; Blaschke et al., 2003). Bouwer (2002) noted that several different processes may be attributed to the occurrence of clogging as water moves through the surface and subsurface soil layers, which include physical (particle settling), chemical (precipitating or gas entrapment), and biological (algae or a biofilm formation) processes. If clogging is present, it may perch the stream and shallow streambed and induce desaturation in the deep streambed (Treese et al., 2009). Furthermore, the presence of clogging can reduce the streambed hydraulic conductivity.
significantly, and thus restrain the interactions between the stream and its adjacent aquifers (Rehg et al., 2005; Treese et al., 2009). Younger et al. (1993) noted that a clogged streambed may act as an intrusion barrier to prevent the polluted surface water to enter into the groundwater system. Kalbus et al. (2009) suggested that implement a heterogeneous distribution of streambed hydraulic conductivity can help avoid the underestimation of peak flows when streambed clogging is present. On the basis of stream ecology, clogging can affect benthic stream communities (Bo et al., 2007), and it may have impacts on the renewal of groundwater through streambed infiltration and the development and colonization of epigean as well as the hypogean invertebrates and fish (Brunke and Gonser, 1997).

In addition, laboratory experiments have been conducted to investigate the effects of clogging. Rehg et al. (2005) conducted experiments in a laboratory flume to identify the effects of fine sediment deposition on hyporheic exchange. They found that the kaolinite clay deposition forms a highly clogged near-surface layer, thereby decreasing the effective permeability and porosity of the streambed which reduces hyporheic exchange. Packman and Mackay (2003) conducted laboratory flume experiments to observe the deposition of kaolinite clay in a sand bed. Their results showed that the clogging of inflow regions produces heterogeneous subsurface clay deposits even when the bed is initially homogeneous, and the clogging of the streambed surface can isolate deeper regions of the bed from the streamflow. Ren and Packman (2007) also noted that larger particles are removed from mixtures and there is a fining of the mixed suspensions over time based on the laboratory flume observations of suspended sediment deposition.

Sophocleous (2002) and Brunner et al. (2010) reviewed the latest research on
groundwater and surface water interactions, and they both noted that the clogging layer is generally present on the channel bottom. Herein, most modelers take this concept for granted of the presence of clogging layer in simulating surface water-groundwater interactions. However, streambed conductance was either arbitrarily chosen or calibrated using numerical simulations in the above studies, thus the uncertainties of the existence and magnitude of the low-permeability clogging layer at the channel surface of streambed can lead to erroneous estimations of stream-aquifer interactions.

1.3.2 Effects of Hyporheic Exchange on Shallow Streambed \( K \)

The exchange of water at the near-channel and in-channel interface usually involves the studies of streambed sediments and hyporheic zone. The hyporheic zone refers to an area beneath a stream channel where the stream water infiltrates and flows through the streambed sediments and returns to the stream after relatively short pathways (Cardenas et al., 2004). Valett et al. (1994) noted that the hydrologic exchange between the hyporheic and surface system can affect the hyporheic zone’s physical (substrate composition), chemical (nutrient environment), and biological (population density and distribution) conditions. The hyporheic zone is of importance for ecosystems, providing nutrients and dissolved gas for microbial communities and organisms.

Cardenas and Zlotnik (2003) suggested that the sediments near the streambed surface have different deposition mechanisms with the deep streambed sediments. Also, the upward and downward water flux or seepage is believed to exist in the hyporheic zone (Packman et al., 2004; Song et al., 2007; Leek et al., 2009; Chen et al., 2009; Rosenberry and Pitlick, 2009), which can induce a higher streambed \( K \) in the hyporheic
zone compared to the deep streambed. Rosenberry and Pitlick (2009) suggested that $K$ values of shallow streambed sediments can increase with upward seepage and decrease with downward seepage, and $K$ values may increase for both upward and downward seepage with the increasing surface water velocity when the bed is fully mobile.

Song et al. (2007) investigated streambed $K_v$ with two connected depths in three major rivers in Nebraska. They found that streambed $K_v$ in the upper sediment layer (50-60 cm below the channel surface) is higher than that in the lower sediment layer (60-90 cm below the channel surface), and they reasoned that this is due to three factors. First, the bigger pore spaces and a more unconsolidated structure of sediments occur in the upper layer caused by the water exchange through upwelling and downwelling zones can cause $K_v$ to increase. Then, redox processes can result in gas production and diffusion in the sediments and the gas moves upward to loosen the upper sediment layer. And finally, invertebrate activities can induce larger pore spaces and increase the permeability of the upper sediments (Song et al., 2007).

Ryan and Boufadel (2006) used a tracer experiment to study the solute exchange within the hyporheic zone in the Indian Creek in Philadelphia, Pennsylvania. They found that higher hydraulic conductivity values were present for the shallow depth of a streambed (7.5-10 cm below channel surface) as compared to those for the lower bed sediments (10–12.5 cm below channel surface). Leek et al. (2009) conducted in-stream slug tests to determine the hydraulic conductivities of a streambed at two test sites (the upper and lower sites) in the Touchet River in Washington. They also found that the mean and median of $K$ for the 0.3-0.45 m depth interval are significantly greater than those values for the 0.6-0.75, 0.9-1.05, and 1.2-1.35 m depth intervals at the lower site.
Besides the hyporheic process, Nogaro et al. (2006) noted that the invertebrates such as tubificid worms can reduce sediment clogging and result in a higher streambed hydraulic conductivity, thereby increasing water-sediment exchanges.

### 1.4 Hypotheses of the Dissertation

#### 1.4.1 Shallow Streambed of the Braided and Meandering Rivers

Although the braided and meandering rivers may have different types of streambed sediments, streambed sediments have their own characteristics due to post-environmental activities near the water-sediment interfaces other than the depositional process for the aquifer sediments, especially when the stream is gaining. At the gaining reaches, baseflow from groundwater can form an uplift force at the channel surface of streambed which may hamper the deposition and settling of fine particles or sediments, resulting in a relatively permeable layer at the streambed surface, even for the meandering rivers which are generally considered to have fine materials at the stream bottom. Consequently, we propose the following hypothesis:

**Hypothesis 1:** Shallow streambeds are permeable over the gaining reaches of the braided and meandering rivers despite their differences on the watershed size, channel width, topographic reliefs, etc.

Furthermore, the assumption of the presence of a clogging layer at the channel surface is adopted widely in numerous analytical and numerical analyses of stream-aquifer interactions (Sophocleous et al., 1995; Hunt, 1999; Osman and Bruen, 2002; Sophocleous, 2002; Akylas and Koussis, 2007; Rushton, 2007; Hu et al., 2007; Sun and Zhan, 2007; Intaraprasong and Zhan, 2009; Brunner et al., 2010). This research suggests
and will demonstrate that this assumption is not always correct for the gaining reaches of the braided and even meandering rivers in simulating the interactions between groundwater and surface water. Thus, we have the second and third hypotheses:

**Hypothesis 2:** If #1 is true, the common assumption of a clogging layer at the channel surface in modeling groundwater-surface water interactions is invalid particularly, for gaining reaches.

**Hypothesis 3:** Thus, in cases of #1 where #2 has been inappropriately applied, a constant-head boundary approach in numerical simulations of streams lacking a clogging-layer is applicable to evaluate the stream-aquifer interactions and has more flexibility for dealing with unclogged streambeds.

In this study, three different rivers in Nebraska were investigated. The Platte River is a typical type of braided channel pattern (Huntzinger and Ellis, 1993); whereas the Big and Little Blue Rivers are typical types of meandering channel pattern (Johnson and Keech, 1959; Mundroff and Waddell, 1966). The Platte River is across the state of Nebraska, but the Big and Little Blue Rivers are more contained locally. Both the Blue Rivers have relatively narrower stream channels and high depths of incision, whereas the Platte River is flatter and more well-defined. Also, the Platte River has higher streamflow discharge and stream velocity than the Blue Rivers. All these differences can lead to an impression that the Platte River has different shallow streambed sediments than the Big and Little Blue rivers.

In-situ permeamter tests were performed to determine the $K_v$ of the shallow streambed sediments for the three rivers, and those $K_v$ values can testify that the shallow streambed is permeable for all the three rivers. In addition, if the shallow streambed is
permeable and thus there is no clogging layer at the channel surface, constant-head boundary approach is hypothesized to be used to simulate stream-aquifer interactions. A hypothetical example of a stream-aquifer system with clogging streambed is developed. A constant-head boundary is used to represent the stream. Meanwhile, the analytical solution provided by Hunt (1999) and the numerical simulation using the River package in MODFLOW (McDonald and Harbaugh, 1988) are also conducted. The results from the three approaches are compared, which can verify the applicability of constant-head boundary approach in modeling the interactions between groundwater and surface water.

1.4.2 Deep Streambed of the Braided and Meandering Rivers

The meandering river usually has more suspended load sediments and a deep and narrow channel allowing for the deposition of fine-grained material at streambed than the braided river. Following the first hypothesis, groundwater flow dynamics may have more effects on the shallow streambed, leading to a mainly permeable streambed surface in both the braided and meandering rivers. However, Cardenas and Zlotnik (2003) suggested that the sediments near the streambed surface exhibiting a bend-flow pattern which may be due to deposition under modern flow regime; while the deeper streambed sediments were considered to have deposited under different flow conditions. Therefore, we propose the following hypothesis:

**Hypothesis 4:** Although the shallow streambed of the braided and meandering rivers is mainly permeable at gaining reaches, the meandering river has a higher content of fine-grained sediments at deep streambed than the braided river.

Numerous studies have presented the measurements of hydraulic conductivity of
the shallow streambed sediments using permeameter testing (Chen, 2000; Landon et al. 2001; Chen, 2004; Genereux et al., 2008; Cheng et al., 2011), slug/bail tests (Cardenas and Zlotnik, 2003; Leek et al., 2009), and pumping tests (Kelly and Murdoch, 2003). Most of these studies focused on the streambed sediments at the depth of 0 to 1.0 m below the channel surface, which may be due to the inherent difficulties in field measurement of streambed hydraulic conductivity (Cardenas and Zlotnik, 2003). The stratification patterns of deep streambed sediments are seldom reported, despite the importance of including such knowledge about the vertical profile of streambed sediments in analyzing the interactions between the streambed and adjacent aquifers.

In this study, the streambed electrical conductivity (EC) logs and sediment cores were obtained at the Platte River, and the Big and Little Blue Rivers using Geoprobe to a depth up to 20 m below the channel surface. EC logs are valuable for assessing the stratification of streambed sediments, as sand and gravel have significantly different EC values as compared to silt and clay. Sand and gravel have a lower EC value compared to silt and clay, since sand and gravel have a larger value of resistivity than silt and clay, and an electrical log is the inverse of a resistivity log (Schulmeister et al., 2003; Sellwood et al., 2005; Chen et al., 2008). Schulmeister et al. (2003) reported that the EC values are about 27 mS/m for sand and gravel and greater than 130 mS/m for silt and clay based on an EC log of the Kansas River floodplain. Chen et al. (2008) reported that the EC values of the streambed sediments in the Platte River were about 20 to 30 mS/m for sand and gravel, 40 to 60 mS/m for fine sand, and greater than 80 mS/m for silt and clay. Also, the sediment cores were collected in transparent tubes, so the laminations and other components of the sediment cores could be visually identified for comparison with the
EC logs. Permeameter tests were performed on the collected sediment cores to determine their $K_v$ values. Consequently, the variations of streambed $K_v$ and EC values with depth for the three rivers are characterized, which can provide insights for the hydrostratigraphy of streambed sediments to testify the hypotheses whether the Big and Little Blue Rivers have more distributions of fine-grained sediments at deep streambed than the Platte River.

### 1.4.3 Statistical Distribution of Shallow Streambed of Braided Rivers

The hydraulic conductivity of aquifer materials is typically assumed to be log-normally distributed in stochastic groundwater analysis (Freeze, 1975). Field investigations generally support the concept of log-normal hydraulic conductivity in aquifers. Bjerg et al. (1992) used a mini slug test to determine the $K$ values of an unconfined sandy aquifer in the western part of Denmark, and they found that these $K$ values can be characterized by a log-normal distribution on the 90% confidence level based on 334 measurements. Hess et al. (1992) conducted flowmeter and permeameter tests to obtain nearly 1500 measurements of $K$ values in the sand and gravel aquifer at Cape Cod, Massachusetts. Their results indicated that the 668 $K$ values obtained from flowmeter tests are log-normally distributed as well as the 825 $K$ values calculated by permeameter tests. Log-normal distribution for $K$ values of aquifer sediments were also reported by other researchers (Sudicky, 1986; Woodbury and Sudicky, 1991; Rehfeldt et al., 1992). The log-normal concept is very commonly used in generation of hydraulic conductivity realizations in stochastic simulations of groundwater flow and solute transport in porous media. For the study of stream-aquifer interactions, the log-normal concept is inherited for characterization of streambed hydraulic conductivity. For
example, Irvine et al. (2012) assumed that streambed sediments are in log-normal distribution in their simulations of surface water–groundwater infiltration flux with a heterogeneous streambed for losing connected, losing transitional, and losing disconnected streams.

Aquifer systems were deposited by a more diverse depositional environment. For example, an alluvial aquifer may consist of the interbedded layers of channel, point bar, levee, and floodplain deposits. Combination of these depositional processes results in very heterogeneous aquifer sediments that have a wide range of hydraulic conductivity. Thus, log-normal distribution seems to be a good representation for aquifers. The flow dynamics within the river channels are less diverse. More importantly, streambeds near water-sediment interface undergo unique hyporheic processes that winnow away fine particle. Therefore, the hydraulic conductivity in shallow streambeds can have a narrow range, and its statistical distribution is unlikely to be log-normal. Instead, normal distribution can represent the statistical characteristics of the less heterogeneous hydraulic conductivity of shallow streambeds. Here, we propose the fifth hypothesis of the dissertation:

**Hypothesis 5:** Shallow streambed sediments of braided rivers are less heterogeneous as compared to the underlying aquifer materials, thus their hydraulic conductivities are not log-normally distributed as typical aquifer hydraulic conductivities.

Several previous studies have investigated the statistical distribution of the horizontal or vertical hydraulic conductivity of the shallow sandy streambed sediments at one or several adjacent sites in a river. Springer et al. (1999) suggested a bimodal distribution for the $K_h$ of the sediments within several reattachments in the Colorado
River of Grand Canyon. Cardenas and Zlotnik (2003) used multilevel constant-head injection tests to collect streambed $K_h$ values at one test site in the Prairie Creek of Nebraska, and their results indicated that streambed $K_h$ is normally distributed based on 456 measurements. Ryan and Boufadel (2007) conducted slug tests using a portable falling-head permeameter to estimate the streambed $K_h$ in two different depths of the Indian Creek in Philadelphia, PA. They noted that $K_h$ is log-normally distributed within each sediment layer but not for the combined dataset of two sediment layers. Genereux et al. (2008) carried out in-situ permeameter tests to obtain 487 measurements of streambed $K_v$ over a year in the West Bear Creek in North Carolina. They found that streambed $K_v$ values are neither normally nor log-normally distributed but show a little bimodal. Above all, these researchers did not develop a statistical distribution analysis of streambed $K_h$ or $K_v$ at distant sites along a large braided river.

The Platte River is a typical braided river and is across the state of Nebraska, and lots of attention has been paid to the study of its interactions with the adjacent aquifers, and thus the statistical distribution of streambed hydraulic conductivity in the Platte River is of importance to understand the interactions for water quantity and quality issues. In-situ permeameter tests were performed within the top 1 m of the shallow streambed at 18 sites along a 300-km (180-mile) segment of the Platte River to determine the streambed $K_v$ values. At each site, 8 to 200 different permeameter tests were conducted. Four different normality tests, Jarque-Bera (J-B), Kolmogorov-Smirnov (K-S), Lilliefors, and Shapiro-Wilk (S-W) tests (Sprent, 2001), were applied at the 0.05 significance level to testify whether the original or log-transformed datasets of streambed hydraulic conductivities are in normal distribution.
Chapter 2  Variations of Streambed Electrical and Hydraulic Conductivity with Depth in Three Rivers of Nebraska

2.1 Introduction

Streambed sediments and their hydraulic conductivities play an important role in controlling stream-aquifer interactions, especially at the near-channel and in-channel interface. The braided and meandering rivers may have different types of streambed sediments due to their differences in stream gradient, source of sediment load, width, etc (Leopold and Wolman, 1957; Schumm and Kahn, 1972). However, streambed sediments have their own characteristics due to post-environmental activities near the water-sediment interfaces other than the depositional process for the aquifer sediments, especially when the stream is gaining. At gaining reaches, baseflow from groundwater can form an uplift force at the channel surface of streambed which may hamper the deposition and settling of fine particles or sediments, resulting in a relatively permeable layer at the streambed surface, even for the meandering rivers which are generally considered to have fine materials at the stream bottom. In this study, three different rivers in Nebraska are investigated, including the lower reach of the Platte River as an example of the braided river (Huntzinger and Ellis, 1993), and the Big and Little Blue Rivers as examples of the meandering river (Johnson and Keech, 1959; Mundroff and Waddell, 1966). All these rivers or reach are mainly gaining, and an analysis of the hydraulic conductivity values of the shallow streambed sediments can help understand whether they are permeable in order to testify the first hypothesis. Moreover, the electrical and hydraulic conductivities in deep streambed are also investigated, which can help
understand the hydrostratigraphy of streambed sediments of whether the meandering rivers have a greater distribution of fine-grained sediments at deep streambed than the braided river, which is beneficial to the integrated water resources management.

In addition, the lower reach of the Platte River in Nebraska is chosen because in addition to that it is a braided river; this area has challenging water resources management issues. Besides groundwater irrigation, the water supply wells of Lincoln and Omaha are located nearby the Platte River and they pump a large amount of groundwater. A superfund contaminant site is located within the study area as well, for which a pump-and-treat system was constructed to assist remediation system. Therefore, knowledge of the hydrostratigraphy of streambed sediments is helpful in regional analysis of the interactions between the reach of the Platte River and its adjacent aquifers.

The Big and Little Blue River Basins are located in southeastern Nebraska, and the hydrologic connectivity of stream-aquifer systems in the two basins has been studied for decades. Emery (1966) first constructed an electric analog model to determine the effects of groundwater withdrawals on the streamflow of the Big and Little Blue Rivers. His model predicted that the base flow depletion induced by the maximum groundwater withdrawals between 1962 and 2002 is very low when compared to the total streamflow. Since the study of Emery (1966), the interest in stream-aquifer interactions in the Blue River Basin has continued. The post-audit analysis of Emery’s (1966) analog model by Alley and Emery (1986) concluded that the decline of water-levels was overestimated but the streamflow depletion was underestimated. These authors used nonparametric statistical tests to analyze the trends in streamflow based on streamflow records from five stations in the Big and Little Blue river Basins. Their results indicated that the average
streamflow depletion rates are an order of magnitude greater than those estimated by Emery (1966). Furthermore, the aquifer storage coefficient was underestimated in the analog model, and a number of factors were omitted from the model such as groundwater recharge from surface water irrigation and groundwater evapotranspiration (Alley and Emery, 1986). Also, Bredehoeft (2005) listed the conflict of the analog’s predictions and the post-audit of actual development in his review of the conceptual model problem. Tabidian and Pederson (1995) found a slightly increasing base flow in the Big Blue River as a result of shutting off irrigation wells, and they also noted that streambed conductance is an important factor in controlling model calibration for the Big Blue river Basin. Furthermore, trend analysis of streamflow by Wen and Chen (2006) indicated that streamflow is stable for the Little Blue River, and that the streamflow decline trend is insignificant based on six streamflow stations with discharge records from 1950 to 2003. The inconsistent results from these studies completed during different time periods suggested that streambed information is urgently needed to be characterized to investigate the stream-aquifer relationships in the Big and Little River Basins more accurately.

2.2 Study Area

The Platte River has its headwaters in the Rocky Mountains and flows through Wyoming and Colorado before entering Nebraska. The North Platte River from Wyoming and the South Platte River from Colorado merges into one river (the Platte River) in North Platte of Nebraska. It is usually a braided, sand-bottom stream with many islands (Huntzinger and Ellis, 1993). A total of ten sites (Figure 2.1) were selected
Figure 2.1 Map showing the locations of the test sites of streambed EC logging and collection of sediment cores for permeameter tests. P1 is designated as the first site in the Platte River; B1 is designated as the first site in the Big Blue River; and L1 is designated as the first site in the Little Blue River.
in the reach of the Platte River. The Platte River is usually wider than 200 m, and becomes even wider toward downstream and can be as wide as 400 m near the city of Ashland. The average streamflow discharge at a United States Geological Survey (USGS) gauge station near North Bend was about 128 m³/s from 1949 to 2008 (http://water.usgs.gov) (Table 2.1). Additionally, the Platte River has low riverbank and low water depths as well. The principal aquifer consists of saturated unconsolidated sediments and alluvium of Quaternary age, and the Tertiary Ogallala Group (http://www.dnr.ne.gov/LB962/AnnualReport_2006/LowerPlatteReport.pdf). However, the alluvial deposits were deposited mainly during the Quaternary time so that they are unconsolidated.

<table>
<thead>
<tr>
<th>Nearby Stream Gauge Station</th>
<th>Platte River 06976000</th>
<th>Big Blue River 06881000</th>
<th>Little Blue River 06883000</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean Stream Level (m, amsl)</td>
<td>386.0</td>
<td>402.4</td>
<td>498.5</td>
</tr>
<tr>
<td>Stream Level Date Range</td>
<td>1989 to 2008</td>
<td>1954 to 2008</td>
<td>1954 to 2008</td>
</tr>
<tr>
<td>Mean Stream Discharge (m³/s)</td>
<td>128</td>
<td>11.1</td>
<td>4.0</td>
</tr>
<tr>
<td>Discharge Date Range</td>
<td>1949 to 2008</td>
<td>1954 to 2008</td>
<td>1954 to 2008</td>
</tr>
<tr>
<td>Average Channel Width (m)</td>
<td>200 to 400</td>
<td>3 to 27</td>
<td>1.5 to 19</td>
</tr>
<tr>
<td>Test Date</td>
<td>June and July 2008</td>
<td>November 2006</td>
<td>November 2006</td>
</tr>
</tbody>
</table>

The Big and Little Blue River Basins are not far away from the Lower Platte River valley. However, the Big and Little Blue River Basins are in a separate watershed, and neither river is as wide as the Platte River as they are rarely wider than 50 m. The
landscape in the Blue River Basin has more rolling hills and narrower valleys compared to the wider and flatter Platter Valley in eastern Nebraska. A total of twelve sites (Figure 2.1) were chosen in the two rivers for investigation of streambeds. At the test sites, the channel widths vary from 3 to 27 m for the Big Blue River, and from 1.5 to 19 m for the Little Blue River. Accordingly, the streamflow discharge in the two rivers is much smaller. The average streamflow discharge was about 11.1 m$^3$/s within the Big Blue River from 1954 to 2008, and about 4.0 m$^3$/s in the Little Blue River from 1954 to 2008 (http://water.usgs.gov) (Table 2.1). The study area consists mostly of gently rolling loess (wind-deposited silt) upland of low relief dissected by small meandering rivers occupying wide shallow valleys (Verstraeten et al., 1998). In the Big and Little Blue River Basins, the principal aquifer consists of Pleistocene alluvial aquifers filling paleovalleys combined in places with small areas of Tertiary Ogallala Group bedrock.  

2.2 Methods

2.2.1 Electrical Conductivity Logging and Coring of Streambeds

Electrical conductivity logs are valuable for assessing the stratification of streambed sediments, as sand and gravel have significantly different EC values as compared to silt and clay. Sand and gravel have a lower EC value compared to silt and clay, since sand and gravel have a larger value of resistivity than silt and clay, and an electrical log is the inverse of a resistivity log (Schulmeister et al., 2003; Sellwood et al., 2005; Chen et al., 2008). Schulmeister et al. (2003) reported that the EC values are about 27 mS/m for sand and gravel and greater than 130 mS/m for silt and clay based on an EC
log of the Kansas River floodplain. Chen et al. (2008) reported that the EC values of the streambed sediments in the Platte River were about 20 to 30 mS/m for coarse sand and gravel, 40 to 60 mS/m for fine sand, and greater than 80 mS/m for silt and clay.

In this study, the EC logs generated by direct-push techniques using a Geoprobe were collected at ten test sites within the Platte River during the summer of 2008 and 2010, at six test sites within the Big Blue River and its two tributaries, Turkey Creek and Swan Creek in November 2006, and at six test sites in the Little Blue River and its two tributaries, Spring Creek and Big Sandy Creek in November 2006 (Figure 2.1). Figure 2.2 shows an example of the field measurements of the EC logs generated by the direct-push technique performed by Geoprobe.

Figure 2.2 Field measurements of electrical conductivity logs and collection of sediment cores using Geoprobe in the Platte River. The channel is wide and some streambed was exposed due to low water level in the river.
A Geoprobe® Systems SC400® soil conductivity probe consisting of a four-electrode Wenner array with an inner-electrode spacing of 2 cm was used. When the probe is pushed through the streambed sediments, an imposed current also passes through the sediments, which can calculate the electrical conductivity (Schulmeister et al., 2003). At each test site, EC logs were recorded every 1.5 cm as the probe was being pushed through the channel sediments. This low EC measurement spacing provided a high-resolution stratigraphic profile of sediments.

In addition to the EC logs, sediment cores were also collected using the Geoprobe® Systems Macro-Core® soil sampler. Cores were collected in polycarbonate tubes every 1.5 m in length that were placed inside the metal core barrel. The polycarbonate tubes were about 4.2 cm in diameter and transparent, so the laminations of the sediment cores could be visually identified for comparison with the EC logs. After the sediment cores were removed from the soil sampler, the two ends of the cores were covered with plastic caps, to minimize impact the impacts of transport on sediment structure and to prevent possible dewatering of the sediments from leakage or evaporation. EC logging and coring of streambed were often to the depth of about 20 m.

2.2.2 In-situ and Laboratory Falling-head Permeameter Testing

Both the in-situ and laboratory falling-head permeameter tests were used to determine the vertical hydraulic conductivity ($K_v$) of the streambed sediments in the tube. In-situ permeameter tests usually only measure the $K_v$ of the shallow streambed sediments ($<1$ m) because of the length of the tube and the difficulty of pushing the tube to a deeper depth. However, the sediment cores collected using Geoprobe® Systems
Macro-Core® soil sampler can reach 20 m below the stream bottom for the test sites in the study, and they were performed laboratory permeameter tests which can provide the vertical profile of streambed $K_v$ to a deeper depth.

Figure 2.3 shows an example of in-situ permeameter test in the river. An in-situ permeameter test using the falling head method usually involves inserting a standpipe into channel sediments (Figure 2.4).

![Figure 2.3 Example of in-situ permeameter test in the Platte River.](image)

In this case, transparent polycarbonate tubes were used for all the tests, and the heads inside the tubes can be easily observed. The tube is 1.5 m in length and 5.0 cm in diameter, and is pressed vertically into the channel sediments. The wall of the tube is about 1 mm thick, thus its effects of disturbance on streambed sediments would be expected to be minimal. After the tube was pressed into a desired depth, the tube
remained in the channel for an appropriate length of time to allow the hydraulic head inside the tube to reach equilibrium due to the slight compaction of the streambed sediments inside the tube. After the head inside the tube equilibrated, the surface water-level at the streambed surface was considered as the initial hydraulic head at the measurement point. However, side-wall leakage, preferential flow within the core, changes in the natural sediment pore-structure and sediment fabric, and simulation at pressures that are unrealistic for the in-situ conditions may be a problem with these tests.

Water was then added from the top of the tube. The hydraulic head in the tube began to fall and the head was recorded in different time steps. In the study, water levels were recorded more than 10 times for each permeameter test. Any pair of measurements from the in-situ permeameter tests can be used to calculate the $K_v$ value using the
equation of Hvorslev (1951):

\[
K_v = \frac{\pi D + L_v}{11m(t_2 - t_1)} \ln\left(\frac{h_1}{h_2}\right)
\] (1)

where \(L_v\) is the length of sediment in the tube; \(h_1\) and \(h_2\) are hydraulic head inside the tube measured at times \(t_1\) and \(t_2\), respectively, \(D\) is the interior diameter of the tube, and \(m = \sqrt{K_h/K_v}\). \(K_h\) is the horizontal hydraulic conductivity of the channel sediment around the base of the sediment core.

\(K_h\) in the equation (1) indicates the possible existence of the horizontal bypass flow along the sides of the tube. Due to the unknown \(K_h\) in the equation (1), a nonlinear regression method was used to determine the streambed \(K_v\). In the computation of \(K_v\), \(m\) must be arbitrarily chosen. Chen (2004) noted that the estimation errors of \(K_v\) based on different arbitrary \(K_h\) values (thereby arbitrary values of \(m\)) is less than 5% when the ratio of \(L_v\) to \(D\) is greater than 5, which indicates that the horizontal bypass flow is insignificant as long as the length of sediment in the tube is large enough compared to the diameter of the tube. All tubes were 5.0 cm in diameter in this study. Also, for the in-situ permeameter tests, only sites P4 to P10 (Figure 2.1) were investigated. The \(L_v\) for each of the in-situ permeameter tests at the test sites ranged from 42 to 50.8 cm. Therefore, the ratios of \(L_v\) to \(D\) are all greater than 5 for the in-situ permeameter tests at all sites.

Figure 2.5 shows the setup of the laboratory permeameter test and a schematic diagram of how to record the measurements.
The lower cap was removed from the polycarbonate tube filled with the sediment and the tube was placed vertically in a tank full of water. The bottom end of the tube was covered by several layers of fine screen to prevent sediment from falling out and to allow water to pass freely through the tube. Water was added to the top of the tube. The hydraulic head inside the core tube begins to fall, and the rate of falling depends on the hydraulic properties of the sediment in the tube and the hydraulic head differences between the tube and the tank. Unlike the in-situ permeameter testing, the horizontal bypass flow along the side of the tube is minimal because the sediments only exist in the tube and the horizontal bypass flow originating from nearby streambed sediments for in-situ permeameter testing is hardly present. According to Darcy’s equation, the vertical hydraulic conductivity is calculated by
\[ K_v = \frac{L_v}{(t_2 - t_1)} \ln(h_1 / h_2) \]  

where \( K_v \) is the vertical hydraulic conductivity of the sediment (L/T), \( L_v \) is the length of the sediment column in the tube (L), and \( h_1 \) and \( h_2 \) (L) are the hydraulic head differences between the tube and the tank at time \( t_1 \) and \( t_2 \) (T) since the permeameter test begins.

2.3 Results

2.3.1 Streambed \( K_v \) from In-situ Permeameter Tests

The measured streambed \( K_v \) values for the shallow streambed sediments in the three rivers are summarized in Table 2.2. More permeameter tests were performed in the Platte River than both the Blue Rivers. In the Platte River, the tested depth of shallow streambed ranged from 42 to 50.8 cm, and the average streambed \( K_v \) values ranged from 23.4 to 45.2 m/d, which suggested that the shallow streambed is highly permeable at all test sites in the Platte River. In the Little Blue River, the average streambed \( K_v \) values ranged from 17.9 to 82.3 m/d in the depth of 54.8 to 61.2 cm below the channel surface. These large \( K_v \) values also indicated that the Little Blue River has a permeable shallow streambed. However, in the Big Blue River, the \( K_v \) values of the shallow streambed can be smaller. At test sites B5 and B6, the \( K_v \) values of the shallow streambed in the depth of 57.8 to 67.2 cm below the channel surface were less than 0.2 m/d. Fine-grained sediments (silt and clay) were found in the shallow streambed at the two sites. Note that the two sites were located in the two tributaries (Turkey Creek and Swan Creek) of the meandering Big Blue River, and they were narrower than the Big Blue River. The suspended load sediments may deposit in the two creeks under low-flow conditions and the uplift force from the baseflow at the channel surface of streambed is not intensive as
other test sites. Overall, the shallow streambed in the three rivers is mainly permeable, which supported the first hypothesis that the gaining reaches of the braided and meandering rivers have a permeable shallow streambed despite their differences on the watershed size, channel width, topographic reliefs, etc.

Table 2.2 Average $K_v$ values of the shallow streambed sediments at the test sites in the three rivers (the Platte River, Big and Little Blue Rivers).

<table>
<thead>
<tr>
<th>Test Site</th>
<th>Test Date</th>
<th>River</th>
<th>Number of Tests</th>
<th>Average $L_v$ (cm)</th>
<th>Average $K_v$ (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P4</td>
<td>June and July 2008</td>
<td>Platte River</td>
<td>64</td>
<td>42.0</td>
<td>31.4</td>
</tr>
<tr>
<td>P5</td>
<td></td>
<td></td>
<td>48</td>
<td>50.8</td>
<td>33.3</td>
</tr>
<tr>
<td>P6</td>
<td></td>
<td></td>
<td>20</td>
<td>51.9</td>
<td>32.4</td>
</tr>
<tr>
<td>P7</td>
<td></td>
<td></td>
<td>49</td>
<td>50.8</td>
<td>29.8</td>
</tr>
<tr>
<td>P8</td>
<td></td>
<td></td>
<td>49</td>
<td>50.8</td>
<td>37.8</td>
</tr>
<tr>
<td>P9</td>
<td></td>
<td></td>
<td>64</td>
<td>50.8</td>
<td>45.2</td>
</tr>
<tr>
<td>P10</td>
<td></td>
<td></td>
<td>40</td>
<td>50.8</td>
<td>23.4</td>
</tr>
<tr>
<td>B1</td>
<td>November 2006</td>
<td>Big Blue River</td>
<td>4</td>
<td>84.9</td>
<td>7.5</td>
</tr>
<tr>
<td>B2</td>
<td></td>
<td></td>
<td>4</td>
<td>87.7</td>
<td>70.7</td>
</tr>
<tr>
<td>B3</td>
<td>November 2006</td>
<td>Big Blue River</td>
<td>5</td>
<td>79.2</td>
<td>26.2</td>
</tr>
<tr>
<td>B4</td>
<td></td>
<td></td>
<td>4</td>
<td>72.7</td>
<td>1.3</td>
</tr>
<tr>
<td>B5</td>
<td></td>
<td></td>
<td>2</td>
<td>57.8</td>
<td>0.05</td>
</tr>
<tr>
<td>B6</td>
<td></td>
<td></td>
<td>4</td>
<td>67.2</td>
<td>0.16</td>
</tr>
<tr>
<td>L1</td>
<td>November 2006</td>
<td>Little Blue River</td>
<td>8</td>
<td>56.0</td>
<td>17.9</td>
</tr>
<tr>
<td>L2</td>
<td></td>
<td></td>
<td>8</td>
<td>61.2</td>
<td>32.5</td>
</tr>
<tr>
<td>L3</td>
<td>November 2006</td>
<td>Little Blue River</td>
<td>1</td>
<td>54.8</td>
<td>41.2</td>
</tr>
<tr>
<td>L4</td>
<td></td>
<td></td>
<td>8</td>
<td>59.2</td>
<td>33.0</td>
</tr>
<tr>
<td>L5</td>
<td></td>
<td></td>
<td>4</td>
<td>57.5</td>
<td>22.0</td>
</tr>
<tr>
<td>L6</td>
<td></td>
<td></td>
<td>8</td>
<td>56.7</td>
<td>82.3</td>
</tr>
</tbody>
</table>

2.3.2 Electrical Conductivity Logs with Depth in the Three Rivers

The relationship between the grain size and hydraulic conductivity for different rock types is shown in Table 2.3 (Freeze and Cherry, 1979). Grain-size analysis was not performed in this study to classify the different sediment types in the tubes collected in
the streambed. However, the sediment cores were stored in transparent tubes, so the sediments inside the tubes were visualized and cross-checked with the EC values since sand and gravel have significantly lower EC values compared to silt and clay (Schulmeister et al., 2003; Sellwood et al., 2005; Chen et al., 2008).

Table 2.3 The relationship between grain size and hydraulic conductivity for different rock types (Freeze and Cherry, 1979).

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Grain size (mm)</th>
<th>Hydraulic Conductivity $K$ (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay</td>
<td>0.0005-0.002</td>
<td>$10^{-8}-10^{-2}$</td>
</tr>
<tr>
<td>Silt</td>
<td>0.002-0.06</td>
<td>$10^{-8}-1$</td>
</tr>
<tr>
<td>Fine Sand</td>
<td>0.06-0.25</td>
<td>1-5</td>
</tr>
<tr>
<td>Medium Sand</td>
<td>0.25-0.50</td>
<td>5-20</td>
</tr>
<tr>
<td>Coarse Sand</td>
<td>0.50-2</td>
<td>20-100</td>
</tr>
<tr>
<td>Gravel</td>
<td>2-64</td>
<td>100-1000</td>
</tr>
</tbody>
</table>

2.3.2.1 The Platte River

The EC logs for the ten sites in the Platte River are shown in Figure 2.6. Meanwhile, the sediments collected in the polycarbonate tubes are visually examined to cross-check the relationship between the EC values and the sediment types. Site P1 (Figure 2.6a) and site P4 (Figure 2.6d) have the same pattern of EC values with depth. The EC values in the top 4 m of channel sediments are about 10 to 30 mS/m, and the sediment cores show that this part of channel sediments consists mainly of coarse sand. Below the coarse sand, the channel sediments at the three sites consist mainly of silt and clay, with an EC value of more than 100 mS/m and even more than 200 mS/m at site P4.
(b)
Figure 2.6 Streambed EC logs produced by Geroprobe and $K_v$ values determined by laboratory permeameter tests of streambed cores at the ten sites in the Platte River (a) site P1; (b) site P2; (c) site P3; (d) site P4; (e) site P5; (f) site P6; (g) site P7; (h) site P8; (i) site P9; and (j) site P10.
Similarly, site P2 (Figure 2.6b), site P7 (Figure 2.6g), and site P8 (Figure 2.6h) have similar pattern of EC values with depth. The EC values in the top 4 m of channel sediments at sites P2 and 7 and in the top 8 m of the channel sediments at site P8 are about 40 to 60 mS/m with a few high values of more than 100 mS/m at sites P2 and P8, and the sediment cores show that this part of channel sediments consists mainly of fine sand with some thin layers of clay and silt. Below the fine sand sediments, the EC values increase to about 80 to 200 mS/m, indicating the existence of silt and clay in deep streambed based on the visualization of the sediments in the tubes.

The EC patterns at sites P5, P6, P9, and P10 (Figure 2.6e, 2.6f, 2.6i, and 2.6j) are very similar. The streambed sediments consist mainly of coarse- to fine-grained sand to more than 10 m below the channel surface, with some very thin layers of silt and clay. The average EC value for the sand is about 20 mS/m, whereas the silt and clay sediments have the EC value of 60 to 120 mS/m. Consequently, at the three sites, sand is the main sediment type in the streambed. Moreover, site P3 (Figure 2.6c) has a similar EC pattern to sites P2, P7, and P8. The difference is that sand sediments occur again below the silt and clay at deep streambed.

In summary, coarse- to fine-grained sand sediments are the main component of the streambed sediments in the Platte River. At several sites, thin layers of silt and clay may appear within the sand sediments in the deep streambed.

2.3.2.2 The Big Blue River

The EC logs for the six sites in the Big Blue River are shown in Figure 2.7. Site B2 (Figure 2.7b), site B3 (Figure 2.7c), site B4 (Figure 2.7d), and site B5 (Figure 2.7e)
have similar EC patterns with depth: fine or coarse sand – silt and clay – fine and coarse sand from top to bottom. The sand sediments in the shallow streambed have an average EC of 20 mS/m at a depth between 0 and 2 m for sites B2, B3, and B4 and at a depth between 0 and 12 m for site B5. Then, a layer of silt and clay sediments appears beneath the sand with an EC value of 80 to 120 mS/m. Fine sand sediments occur again below the silt and clay sediments, which is interbedded with thin layers of silt and clay.

Site B1 (Figure 2.7a) and site B6 (Figure 2.7f) have similar EC patterns. Silt and clay (or with interbedded sand) appear at a depth between 0 and 2 m of the shallow streambed, with an EC value of 60 to 80 mS/m. Fine sand sediments with interbedded silt occur below the silt and clay sediments.

In summary, sand sediments are found in the deep streambed at all the sites and they also appear in the shallow streambed at several sites. EC values of the sand sediments are between 10 and 20 mS/m. The shallow streambed in the Big Blue River also consists mainly of sand sediments, although site B1 and B6 have thin layers of silt and clay at the channel surface. Silt and clay sediments with higher EC values (60 to 120 mS/m) occurred at depths below 2 m at several sites. Compared to the Platte River, the Big Blue River has apparently more silt and clay sediments in the top 5 m of streambed.
(a) Silt and clay
Fine and coarse sand with interbedded silt

(b) Coarse sand
Silt and clay
Fine sand with interbedded silt
(c) Fine sand with interbedded silt

(d) Coarse sand

Fine sand with interbedded silt
Figure 2.7 Streambed EC logs produced by Geroprobe and $K_v$ values determined by laboratory permeamter tests of streambed cores at the six sites in the Big Blue River: (a) site B1; (b) site B2; (c) site B3; (d) site B4; (e) site B5; and (f) site B6.
2.3.2.3 The Little Blue River

The EC logs for the six sites in the Little Blue River are shown in Figure 2.8. Site L2 (Figure 2.8b) and site L3 (Figure 2.8c) have similar EC patterns. At the two sites, sand sediments appear at the channel surface and silt and clay sediments are the main component in the deep streambed. The average EC value for the shallow streambed is about 15 mS/m. The depth of the silt and clay sediments can reach up to 12 m at site L2 and 18.0 m at site L3. The average EC values of the silt and clay sediments are 95 mS/m at site L2 and 65 mS/m at site L3.

At sites L4, L5, and L6 (Figures 2.8d, 2.8e, and 2.8f), the streambed is composed mainly of fine sand sediments with several very thin layers of interbedded silt and clay sediments. The average EC value for the fine sand sediments at the three sites is about 20 mS/m, whereas the largest EC value for the silt and clay sediments at the three sites is 160 mS/m.

At site L1 (Figure 2.8a), sand sediments occur in the shallow streambed with an average EC of 15 mS/m at depths between 0 and 3 m. A thick sediment layer composed of silt, clay, and fine sand appears at a depth between 3 and 7 m, and the maximum EC value for this layer was 180 mS/m. Fine sand sediments occur again below the silt and clay sediments.

In summary, sand sediments are the primary component at the channel surface in the streambed at the six sites in the Little Blue River, although silt and clay sediments may appear in the deep streambed at sites L2 and L3. The EC value of the sand sediments is between 10 and 20 mS/m, and the EC value of the silt and clay sediments is between 80 and 180 mS/m. Compared to the Big Blue River, the Little Blue River
(a) Fine and coarse sand
Silt and clay with sand
Fine sand with interbedded silt

(b) Coarse sand
Silt and clay
Figure 2.8 Streambed EC logs produced by Geroprobe and $K_v$ values determined by laboratory permeameter tests of streambed cores at the six sites in the Little Blue River (a) site L1; (b) site L2; (c) site L3; (d) site L4; (e) site L5; and (f) site L6.
usually has the sand sediments at the shallow streambed. Because grain-size was not performed, the estimated hydraulic conductivities of the sediment cores can provide the comparison for the sediments in the deep streambed of the three rivers.

2.3.3 Streambed Vertical Hydraulic Conductivity with Depth in the Three Rivers

2.3.3.1 The Platte River

The streambed in the Platte River is composed mainly of coarse- to fine-grained sands, thus the streambed’s \( K_v \) values are usually greater than 1 m/d (Figure 2.6). There are a total of 60 streambed \( K_v \) measurements from the collected sediment cores. The average streambed \( K_v \) value of all the sand sediments is 11.1 m/d, whereas the average streambed \( K_v \) value of all the silt and clay sediments is 0.3 m/d.

For the shallow streambed at the depth of 0 to 3 m below the channel surface for all of the collected sediment cores, the streambed \( K_v \) values are larger than 10 m/d at sites P1, P4, P7, and P9, and larger than 1 m/d at sites P2, P3, P6, and P10. At sites P5 and P8, very thin layers of silt and clay sediments appear to occur in the shallow streambed at depths between 0 and 1 m, thereby yielding relatively lower \( K_v \) values, which are less than 1 m/d, but are still larger than 0.1 m/d.

At the depth of 3 to 6 m below the channel surface among all the sediment cores, the streambed \( K_v \) values exhibit a wider range of variation up to six orders of magnitude. Sites P1, P2, P3, and P4 all show the same \( K_v \) pattern with depth that the occurrence of silt and clay sediments decreases the streambed \( K_v \) values by several orders of magnitude. The average streambed \( K_v \) value at depths of 4.5 to 6 m is about 0.0005, 0.0003, 0.0008, and 0.0006 m/d at sites P1, P2, P3, and P4, respectively. From sites P5 to P10, the
streambed $K_v$ values at the depth of 3 to 6 m below the channel surface are all larger than 1 m/d, except for that the streambed $K_v$ values are less than 0.01 m/d at site P8. In addition, at sites P6, P7, and P9, the streambed $K_v$ values at depths of 3 to 6 m are less than the corresponding values at depths of 0 to 3 m. Therefore, the streambed $K_v$ values decrease with depth from the channel surface down to 6 m in the Platte River.

From a depth of 6 m below the channel surface among all the sediment cores, there is only one streambed $K_v$ measurement at sites P1, P2, P3, P7, P8, and P9. At sites P1 and P7, the streambed $K_v$ values are less than 1 m/d; whereas streambed $K_v$ values are all larger than 1 m/d at sites P2, P3, P8, and P9. There are no $K_v$ measurements at site P4 below a depth of 6 m in the streambed because Geoprobe encountered silt and clay sediments and then stopped. Moreover, the streambed $K_v$ values range from 1.8 to 36.0 m/d with an average of 16.0 m/d among these measurements at sites P5, P6, and P10 from the depth of 6 m below the channel surface.

Figure 2.9 (a) shows the varied streambed $K_v$ values with depth at all the ten sites in the Platte River. It is apparent that the streambed $K_v$ values at depths from 0 to 3 m below the channel surface are higher than those at depths from 3 to 6 m. The average streambed $K_v$ value is about 9.2 m/d at depths of 0 to 3 m compared to 7.8 m/d at depths of 3 to 6 m (Table 2.4). In addition, because the range of variation of streambed $K_v$ values at different sites is usually more than one order of magnitude, the geometric mean of the streambed $K_v$ values is also calculated. The geometric mean streambed $K_v$ value is about 4.9 m/d at depths from 0 to 3 m compared to 0.34 m/d at depths from 3 to 6 m (Table 2.4). From a depth of 6 m below the channel surface, the streambed $K_v$ values also exhibit a tendency to decrease with depth (Figure 2.9a; Table 2.4).
Table 2.4 The streambed $K_v$ values relative to different depths in the three rivers (the Platte River, the Big and Little Blue Rivers).

<table>
<thead>
<tr>
<th>River</th>
<th>Depth (m)</th>
<th>Number of Measurements</th>
<th>Average Streambed $K_v$ (m/d)</th>
<th>Geometric Mean Streambed $K_v$ (m/d)</th>
<th>Median Streambed $K_v$ (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>The Platte River</td>
<td>0-3.0</td>
<td>19</td>
<td>9.2</td>
<td>4.9</td>
<td>7.4</td>
</tr>
<tr>
<td></td>
<td>3.0-6.0</td>
<td>19</td>
<td>7.8</td>
<td>0.34</td>
<td>5.6</td>
</tr>
<tr>
<td></td>
<td>6.0-9.0</td>
<td>12</td>
<td>7.8</td>
<td>2.3</td>
<td>5.8</td>
</tr>
<tr>
<td></td>
<td>9.0-12.0</td>
<td>5</td>
<td>8.3</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>12.0-15.0</td>
<td>5</td>
<td>2.4</td>
<td>0.4</td>
<td>1</td>
</tr>
<tr>
<td>The Big Blue River</td>
<td>0-3.0</td>
<td>8</td>
<td>7.0</td>
<td>0.46</td>
<td>2.0</td>
</tr>
<tr>
<td></td>
<td>3.0-6.0</td>
<td>12</td>
<td>0.95</td>
<td>0.04</td>
<td>0.08</td>
</tr>
<tr>
<td></td>
<td>6.0-9.0</td>
<td>12</td>
<td>2</td>
<td>0.19</td>
<td>0.27</td>
</tr>
<tr>
<td></td>
<td>9.0-12.0</td>
<td>11</td>
<td>2</td>
<td>0.75</td>
<td>0.99</td>
</tr>
<tr>
<td></td>
<td>12.0-15.0</td>
<td>10</td>
<td>1.9</td>
<td>0.67</td>
<td>0.7</td>
</tr>
<tr>
<td>The Little Blue River</td>
<td>0-3.0</td>
<td>12</td>
<td>28.7</td>
<td>2.2</td>
<td>23.9</td>
</tr>
<tr>
<td></td>
<td>3.0-6.0</td>
<td>12</td>
<td>0.27</td>
<td>0.04</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td>6.0-9.0</td>
<td>11</td>
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<td>0.21</td>
<td>0.36</td>
</tr>
<tr>
<td></td>
<td>9.0-12.0</td>
<td>10</td>
<td>3.4</td>
<td>0.15</td>
<td>1.8</td>
</tr>
<tr>
<td></td>
<td>12.0-15.0</td>
<td>7</td>
<td>9.3</td>
<td>0.39</td>
<td>3.41</td>
</tr>
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<td>15.0-18.0</td>
<td>5</td>
<td>7.9</td>
<td>0.4</td>
<td>2.93</td>
</tr>
</tbody>
</table>

![Histogram of Streambed $K_v$ values](a)
Figure 2.9 Distributions of streambed $K_v$ values with depth from the ten sites in the Platte River (a), from the six sites in the Big Blue River (b), and from the six sites in the Little Blue River (c).
2.3.3.2 The Big Blue River

Compared to the Platte River, fine-grained sand sediments in the Big Blue River are more interbedded with silt and clay sediments, which is not surprising given that the Big Blue River is a meandering river while the Platte River is a braided river, and thus both rivers have different sediment sources and loads and channel patterns. The streambed $K_v$ values at all the depths below the channel surface in the Big Blue River are less than those in the Platte River (Table 2.4). There are a total of 53 streambed $K_v$ measurements from the collected sediment cores. The average streambed $K_v$ value of all the sand sediments is 4.4 m/d, whereas the average streambed $K_v$ value of all the silt and clay sediments is 0.08 m/d.

For the shallow streambed from the depth of 0 to 3 m below the channel surface among all the sediment cores, the streambed $K_v$ values exhibit a wider range of variation with four orders of magnitude. They may be permeable at depths of 0 to 1.5 m, but they are not as permeable as the Platte River.

At the depth of 3 to 6 m below the channel surface among all the sediment cores, the streambed $K_v$ values are also small and are between 0.005 and 6.4 m/d. Note that the maximum streambed $K_v$ value at a depth of 3 to 6 m is smaller than that at a depth of 0 to 3 m. Moreover, the average streambed $K_v$ value is about 4.7 m/d at a depth of 0 to 3 m compared to 0.95 m/d at a depth of 3 to 6 m.

From the depth of 6 m below the channel surface among all the sediment cores, the streambed $K_v$ values become larger than those at the depth of 3 to 6 m. At sites B1, B4, B5, and B6, the streambed $K_v$ values are usually between 0.1 and 10 m/d. At site B3, silt and clay sediments occur at a depth of 6 to 10 m below the channel surface, which yields
a lower \( K_v \) value (<0.1 m/d) at this depth. Fine-grained sand sediments appear below this depth, and thereby result in a relatively larger \( K_v \) value (between 0.1 and 10 m/d). At site B2, there are three streambed \( K_v \) measurements below a depth of 6 m. Although the EC values are about 20 mS/m, the corresponding streambed \( K_v \) values are lower with an average value of 0.008 m/d at a depth of 6 to 9 m and 1.1 m/d at a depth of 9 to 10.5 m. EC logs can only provide a general idea of how permeable the sediments are. If there is a very thin layer of low-permeability sediments inside the dominantly permeable sediments, it can decrease the \( K_v \) values significantly because the equivalent \( K_v \) value is mainly affected by the lowest \( K_v \) value for different layers of sediments; however, the EC log may not observe this thin layer of low-permeability sediments.

Figure 2.9 (b) shows the variation of streambed \( K_v \) values with depth at the six sites in the Big Blue River. The streambed \( K_v \) values at the depth of 0 to 3 m below the channel surface are apparently higher than those at the depth of 3 to 6 m. However, from the depth of 6 m below the channel surface, statistical analysis showed there is no apparent trend in the streambed \( K_v \) values relative to depth (Figure 2.9b; Table 2.4). Streambed \( K_v \) values are lower at all depths in the Big Blue River than those in the Platte River, which is due to that the two rivers belong to different channel patterns.

### 2.3.3.3 The Little Blue River

The streambed in the Little Blue River is composed mainly of fine-grained sediments interbedded with silt and clay. Compared to the Big Blue River, the silt and clay sediments occur primarily in the deep streambed. There are a total of 57 streambed \( K_v \) measurements made on segments of the collected sediment cores. The average
streambed $K_v$ value of all the sand sediments is 18.3 m/d, whereas the average streambed $K_v$ value of all the silt and clay sediments is 0.005 m/d.

For the shallow streambed from the depth of 0 to 3 m below the channel surface for all of the sediment cores collected, the streambed $K_v$ values are generally very large near the channel surface. The streambed $K_v$ values are all larger than 10 m/d at depths of 0 to 1.5 m at the six sites. However, at sites L1, L2, and L4, the occurrence of silt and clay sediments at the depth of 1.5 to 3 m yields smaller streambed $K_v$ values of less than 0.01 m/d. Streambed $K_v$ values are still larger than 10 m/d at the depth of 1.5 to 3 m at sites L3 and L6, while they are about 1 m/d at site L5. Overall, the shallow streambed below the channel surface in the Little Blue River is similar to that in the Platte River and is more permeable than that in the Big Blue River.

At the depth of 3 to 6 m below the channel surface among all the sediment cores, the streambed $K_v$ values are much smaller than those at depths of 0 to 3 m. At sites L1, L2, and L3, the streambed $K_v$ values are usually less than 0.01 m/d; whereas the streambed $K_v$ values are usually between 0.1 m/d and 1 m/d at sites L4, L5, and L6. The average streambed $K_v$ value is about 28.7 m/d at a depth of 0 to 3 m compared to 0.27 m/d at a depth of 3 to 6 m, and the geometric mean value of streambed $K_v$ is about 2.2 m/d at a depth of 0 to 3 m compared to 0.04 m/d at a depth of 3 to 6 m (Table 2.4).

From the depth of 6 m below the channel surface observed in all the sediment cores, the streambed $K_v$ values are also very small at sites L2 and L3 and are usually less than 0.01 m/d. At sites L1, L4, and L6, streambed $K_v$ values are usually larger than 1 m/d, and can be as large as 42 m/d. At site L5, streambed $K_v$ values are all larger than 0.1 m/d (Figure 2.9c).
Figure 2.9 (c) shows the variation of streambed $K_v$ values with depth at the six sites in the Little Blue River. The average and geometric mean streambed $K_v$ values at depths of 0 to 3 m below the channel surface are higher than those at depths of 3 to 6 m. From a depth of 6 m below the channel surface statistical analysis showed there is no apparent trend in the streambed $K_v$ values relative to depth (Figure 2.9c; Table 2.4). Compared to the Platte River, the average streambed $K_v$ values in the Little Blue River can be higher at some depths. For instance, streambed $K_v$ values are higher at the depth of 0 to 3 m, 6 to 9 m, and 12 to 15 m below the channel surface. However, the geometric means of streambed $K_v$ values are lower at all depths in the Little Blue River than those in the Platte River, which is also due to that the two rivers belong to a different channel pattern. Compared to the Big Blue River, the Little Blue River has higher streambed $K_v$ values, which implies that the Little Blue River has a more permeable streambed.

2.4. Discussion

2.4.1 $K_v$ Values of Shallow Streambed Sediments

From the in-situ permeameter tests, the average $K_v$ value is about 33.3 m/d for the top 50-cm streambed in the Platte River, 17.7 m/d for the top 75-cm streambed in the Big Blue River, and 38.2 m/d for the top 60-cm streambed in the Little Blue River (Table 2.2). All indicate very permeable streambeds. However, at the two tributaries (Turkey Creek and Swan Creek) of the Big Blue River, there are low-permeability sediments lining beneath the stream bottom which generates a smaller $K_v$ value (Table 2.2). From the laboratory permeameter tests at the depth of 0 to 1.5 m below the channel surface, the average $K_v$ value is about 6.7 m/d in the Platte River, 17.3 m/d in the Big Blue River, and
38.7 m/d in the Little Blue River. Therefore, the shallow streambed at the three rivers is mainly permeable based on the streambed $K_v$ values at the depth of 0 to 1.5 m below the channel surface from the in-situ and laboratory permeameter tests, which testifies the first hypothesis in the dissertation.

The braided river usually has a sandy stream bottom because the sediment load is primarily carried in bed load, thus it is not surprising that the shallow streambed in the Platte River is highly permeable. The Big and Little Blue Rivers are meandering rivers therefore they usually have a less permeable streambed (compared to the Platte River. Note based on the $K_v$ values, the top layer of the streambeds in the three rivers are all very permeable) because the sediment load is primarily carried in suspended load and low-flow conditions can create an environment for fine-grained sediments to be deposited. For example, Hatch et al. (2010) found the streambed $K_v$ values are lower in the dry season for the reach of the Pajaro River in central coastal California as the fine-grained sediments were deposited on the streambed during low flow conditions. The measurements in the Big and Little Blue Rivers were conducted in November 2006 under low-flow conditions; however, although the tributaries of the Big Blue River have low-permeability sediments beneath the stream bottom, the shallow streambed is mainly permeable for the two rivers. Hyporheic process is believed to exist at the channel surface. The upward and downward water flux or seepage exists in the hyporheic zone (Packman et al., 2004; Song et al., 2007; Leek et al., 2009; Chen et al., 2009; Rosenberry and Pitlick, 2009), which can loosen the streambed sediments in the hyporheic zone and result in a high streambed hydraulic conductivity. Moreover, at the gaining reaches, baseflow from groundwater can form an uplift force at the channel surface of streambed
which may hamper the deposition and settling of fine particles or sediments, resulting in a relatively permeable layer at the streambed surface, even for the meandering rivers generally characterized with fine materials at the stream bottom.

Furthermore, an assumption of the occurrence of a clogging layer at the channel surface is adopted widely in numerous analytical and numerical analyses of stream-aquifer interactions (Sophocleous et al., 1995; Hunt, 1999; Osman and Bruen, 2002; Sophocleous, 2002; Akylas and Koussis, 2007; Rushton, 2007; Hu et al., 2007; Sun and Zhan, 2007; Intaraprasong and Zhan, 2009; Brunner et al., 2010). As discussed above, the shallow streambed is usually permeable at the gaining reaches of the braided and meandering rivers; therefore this assumption is not always correct, which supports the second hypothesis in the dissertation. Consequently, it is important to perform field measurements of shallow streambed hydraulic conductivity before adopting such assumptions in numerical simulations.

### 2.4.2 Variation of Streambed $K_v$ Values with Depth

In the three rivers, the streambed $K_v$ values near the channel surface are generally larger than 5 m/d, which indicates that the shallow streambed is highly permeable in the three rivers. However, at some sites in the Big and Little Blue Rivers, the streambed $K_v$ values are between 0.001 and 0.1 m/d at the depth of 1.5 to 3 m below the channel surface, and the corresponding sediment cores show the presence of thin layers of silt and clay sediments. In general, the Platte River is wider and has higher stream flow discharge and flow velocity compared to the Big and Little Blue Rivers (Table 3.1). The low flow in the Big and Little Blue Rivers results in the fine-grained sediments (e.g. silt and clay)
more easily deposited within the pores of sand sediments and not washed away with the streamflow. Moreover, considering the lower streambed $K_v$ values (<0.1 m/d) in the deep streambed (in the depth of 1.5 m below the channel surface) in the three rivers, there are 18 measurements in the Big Blue River, 25 measurements in the Little Blue River, but only 10 measurements in the Platte River. Also, the total measurements of streambed $K_v$ values in the Platte River are more than those in the two Blue Rivers. Therefore, the apparent difference between the Platte River and both the Blue Rivers shows the characteristics of a meandering river for both the Blue Rivers: even though the very shallow streambeds of the two rivers are mostly permeable, they have more fine-grained sediments deposited in the deep streambed than the braided river. It confirms the fourth hypothesis in the dissertation.

Cardenas and Zlotnik (2003) suggested that the sediments near the streambed surface exhibiting a bend-flow pattern which may be due to deposition under modern flow regime; while the deeper streambed sediments were considered to have deposited under different flow conditions. In this study, streambed $K_v$ values in the three rivers exhibit a tendency to decrease with depth in the depth of 0 to 6 m below the channel surface (Figure 2.9). Hyporheic process is believed to be a main reason for this phenomenon (Ryan and Boufadel, 2006; Song et al., 2007; Leek et al., 2009). The upwelling and downwelling water exchanges occurring in the hyporheic zone could result in unconsolidated sediment structure. Additionally, redox process and bioturbation activities are believed to affect the streambed hydraulic conductivity as well (Nogaro et al., 2006; Song et al., 2007), and the mechanism of these processes decreases with depth.
2.4.3 Variation of Streambed EC Values with Depth

Direct-push EC logging is used widely by researchers to characterize the site hydrostratigraphy (Schulmeister et al., 2003; Wilson et al., 2005; Sellwood et al., 2005; Zlotnik et al., 2007). Schulmeister et al. (2003) noted that there is an agreement between peaks in the EC profiles and increases in the clay content of the sampled layers, and they also pointed out that higher EC value generally reflect fine-grained material. In this study, for the sediments at depths between 0 and 3 m below the channel surface, the EC values are usually 10 to 40 mS/m with a few outliers of more than 60 mS/m in the three rivers. Chen et al. (2008) reported that the EC values in the top 2 m streambed are about 20-40 mS/m between Kearney and Columbus in the Platte River. Since the streambed sediments in the near-channel surface in the three rivers are mainly composed of coarse-to fine-grained sand, the EC values are accordingly small. For the sediments from a depth of 4 m below the channel surface, the EC values increase to more than 100 mS/m at several sites in the three rivers, which is likely due to the presence of fine-grained material, e.g., silt and clay sediments.

Schulmeister et al. (2003) found that the transition from unsaturated to saturated conditions could be abrupt for coarse-grained material and become more gradual for fine-grained material. Note that the streambed sediments in the three rivers are all saturated, thus the changes of EC values truly reflect the variations of sediment content in the streambed. Also, variations in water chemistry and porosity can also have a major impact on EC (Schulmeister et al., 2003), whereas the factors affecting streambed $K_v$ could be more complex, including the sediment texture and sorting, and especially the grain-size diameter. Although grain-size analysis can provide additional information on the
sedimentation of the streambed, the variations of EC values with depth characterize the hydrostratigraphy of streambed sediments, which is useful in numerical simulations of stream-aquifer interactions. For instance, if there is a high EC value (>100 mS/m) at the channel surface, which usually indicates the existence of a low-permeability layer (or clogging layer) and thus those analytical solutions (Hunt, 1999; Zlotnik and Huang, 1999; Butler et al., 2001) with an assumption of a clogging layer are applicable to represent the stream-aquifer interactions. The possibility of clogging is much higher in the meandering rivers than the braided rivers. However, if there is a low EC value (<40 mS/m) at the channel surface, the shallow streambed is permeable, which is very common in the braided rivers. In this case, there may be a low-permeability layer in the deep streambed, or this low-permeability layer may not be present at all. Here, the depth of well and the length of well screen should be taken into account for pumping-induced interactions. For example, if the well depth is below the low-permeability layer, pumping-induced stream depletion might be still low because the low-permeability layer still acts as a barrier to prevent streamflow infiltration to the aquifer system. Also, if the low-permeability layer does not occur at all in the streambed, the stream-aquifer interaction can be much higher, thus the assumption of a clogging layer underestimates the stream-aquifer interactions. Overall, the Geoprobe-generated EC logs with depth depict the hydrostratigraphic sediment layers in the streambed, which can characterize streambed sedimentary structure and provide references for simulating stream-aquifer interactions.

2.4.4 Relationship between Streambed $K_v$ and EC

The EC values are averaged every 1.5 m in correspondence to the streambed $K_v$. 
values, which are calculated every 1.5 m in the streambed. Both the streambed $K_v$ and EC values are transformed based on log$_{10}$. The relationship between the streambed $K_v$ and EC in the three rivers is shown in Figure 2.10.

![Figure 2.10 The relationship between the EC values and streambed $K_v$ values for all the sediment cores in the three rivers (the Platte River, the Big and Little Blue Rivers).](image)

Although the estimated R-square is only 0.39, there is a trend that the streambed $K_v$ values are somewhat inversely correlated to the EC values in the three rivers. If there is a very thin layer of low-permeability (low-$K$) sediments inside the dominantly permeable sediments in the 1.5-m tube, the low-$K$ sediments can decrease the $K_v$ value significantly because the equivalent $K_v$ value for the 1.5-m sediments is mainly affected by the lowest $K_v$ value for different layers of sediments; however, the arithmetic mean EC values are calculated for every 1.5-m sediments, so they cannot reflect this thin layer of low-permeability sediments. Hence, grain-size analysis is suggested for future investigations.
of streambed hydrostratigraphy to provide additional information.

2.5 Summary and Conclusions

The vertical profile of streambed electrical conductivity (EC) and vertical hydraulic conductivity ($K_v$) values are presented in three rivers of Nebraska: the Platte River, the Big Blue River, and the Little Blue River. The Platte River is a braided river, whereas the Big and Little Blue Rivers are meandering rivers. The EC logs of streambed sediments were obtained using Geoprobe® up to a depth of about 20 m below the channel surface in the three rivers, and then the sediments cores were collected into polycarbonate tubes every 1.5 m in length using the Geoprobe® Systems Macro-Core® soil sampler. Laboratory permeameter tests were performed on these sediment cores to determine the $K_v$ values of the shallow and deep streambed sediments, and in-situ permeameter tests were performed as well to determine the $K_v$ values of the shallow streambed (<1 m).

Streambed $K_v$ values near the channel surface are generally larger than 5 m/d in the three rivers according to the in-situ permeameter tests and the laboratory permeameter tests on the sediment cores at the depth of 0 to 1 m below the channel surface, which indicates that the shallow streambed is highly permeable. However, at the tributaries of the Big Blue River, there are low-permeability sediments lining beneath the stream bottom which generates a smaller $K_v$ value. The Big and Little Blue Rivers are meandering rivers therefore they usually have a less permeable streambed because the sediment load is primarily carried in suspended load and low-flow conditions can be easy for fine-grained sediments to be deposited. The upward and downward water flux or seepage exists in the hyporheic zone can loosen the streambed sediments in the hyporheic zone and result in a
high streambed hydraulic conductivity. Moreover, at the gaining reaches, baseflow from groundwater can form an uplift force at the channel surface of streambed which may hamper the deposition and settling of fine particles or sediments, resulting in a relatively permeable layer at the streambed surface, even for the meandering rivers generally characterized with fine materials at the stream bottom. Consequently, the assumption of the presence of a clogging layer at the channel surface adopted widely in numerous analytical and numerical analyses of stream-aquifer interactions is not always correct, especially at the gaining reaches.

The two Blue Rivers have more lower $K_v$ values in the deep streambed than the Platte River, which indicates that both the Blue Rivers show the characteristics of a meandering river for both the Blue Rivers: even though the very shallow streambeds of the Big and Little Blue Rivers are mostly permeable, they have more fine-grained sediments deposited in the deep streambed than the Platte River. Furthermore, streambed $K_v$ values in the three rivers exhibit a tendency to decrease with depth in the depth of 0 to 6 m below the channel surface. Previous studies suggested that the sediments near the streambed surface exhibiting a bend-flow pattern which may be due to deposition under modern flow regime; while the deeper streambed sediments were considered to have deposited under different flow conditions. Hyporheic process is believed to be a main reason for this phenomenon. Additionally, redox process and bioturbation activities are believed to affect the streambed hydraulic conductivity as well, and the mechanism of these processes decreases with depth.

The variations of EC values with depth characterize the hydrostratigraphy of streambed sediments, although grain-size analysis can provide additional information on
the sedimentation of the streambed. Sand and gravel have a lower EC value compared to silt and clay, since sand and gravel have a larger value of resistivity than silt and clay, and an electrical log is the inverse of a resistivity log. In this study, for the sediments at depths between 0 and 3 m below the channel surface, the EC values are usually 10 to 40 mS/m with a few outliers of more than 60 mS/m in the three rivers, which can also suggest that the shallow streambed is mainly permeable. For the sediments from a depth of 4 m below the channel surface, the EC values can increase to more than 100 mS/m at several sites in the three rivers, which is likely due to the presence of fine-grained material, e.g., silt and clay sediments. Overall, streambed $K_v$ values are inversely correlated to the EC values in the three rivers.
Chapter 3  Statistical Distribution and Spatial Variation of Streambed Vertical Hydraulic Conductivity in the Platte River of Nebraska

3.1 Introduction

Streambed vertical hydraulic conductivity ($K_v$) is a key parameter to know or determine when quantifying stream-aquifer interactions. Heterogeneity of streambed $K$ could affect hyporheic zone fluxes and groundwater discharge (Salehin et al., 2004; Kalbus et al., 2009; Kennedy et al., 2009). A number of researchers discussed the methods for the determination of streambed $K_v$, which include the permeameter test (Hvorslev, 1951; Chen, 2000; Landon et al., 2001; Chen, 2004; Chen, 2005; Genereux et al., 2008; Kennedy et al., 2009), slug/bail tests (Springer et al., 1999; Landon et al., 2001; Ryan and Boufadel, 2007; Leek et al., 2009), grain-size analysis (Chen, 2000; Landon et al., 2001), and pumping test (Kelly and Murdoch, 2003). Generally, slug and bail tests can only provide streambed horizontal hydraulic conductivity ($K_h$) values. Grain-size analysis cannot evaluate the anisotropy of $K$ values because the sediment structure is destroyed during sampling (Chen, 2000; Kalbus et al., 2006; Cheng and Chen, 2007). In contrast, permeameter tests can provide streambed $K_v$ values which are more accurate than grain-size analysis and less expensive than those determined using pumping tests. However, side-wall leakage, preferential flow within the core, changes in the natural sediment pore-structure and sediment fabric, and simulation at pressures that are unrealistic for the in-situ conditions may be a problem with these tests. Above all, the $K_v$ value of shallow streambed sediments is a crucial factor in controlling the interactions between surface water and groundwater, and it is beneficial to better understand its statistical distribution and spatial variability at different sites in a large braided river.
The spatial and temporal variations of streambed $K_h$ and $K_v$ have been analyzed and discussed by many researchers (Springer et al., 1999; Cardenas and Zlotnik, 2003; Chen, 2005; Ryan and Boufadel, 2007; Genereux et al., 2008). They found that the shallow streambed sediments may have a wide range of variations for a given site, e.g., they are in bimodal, normal, or log-normal distributions. In addition, the hydraulic conductivity of aquifer materials is typically assumed to be log-normally distributed in stochastic groundwater analysis (Freeze, 1975; Bjerg et al., 1992; Hess et al., 1992; Sudicky, 1986; Woodbury and Sudicky, 1991; Rehfeldt et al., 1992). For the study of stream–aquifer interactions, the log-normal concept is inherited for characterization of streambed hydraulic conductivity. For example, Irvine et al. (2012) assumed that streambed sediments are in log-normal distribution in their simulations of surface water–groundwater infiltration flux with a heterogeneous streambed for losing connected, losing transitional, and losing disconnected streams. However, the flow dynamics within the river channels are less diverse compared to aquifer aquifers, which may consist of the interbedded layers of channel, point bar, levee, and floodplain deposits. More importantly, streambeds near water-sediment interface undergo unique hyporheic processes. Therefore, the hydraulic conductivity in shallow streambeds can have a narrow range, and its statistical distribution is unlikely to be log-normal. Instead, normal distribution can represent the statistical characteristics of the less heterogeneous hydraulic conductivity of shallow streambeds.

Over the past 10 years, numerous in-situ and laboratory permeameter tests have been conducted in determining streambed $K_v$ in the Platte River of Nebraska (Landon et al., 2001; Chen, 2004; Chen, 2005; Song et al., 2007; Chen et al., 2008). Landon et al.
(2001) performed in-situ permeameter tests to investigate the streambed $K_v$ in the Platte River near Brady. They concluded that in the top 25-cm of the streambed, $K_v$ is usually greater than 50 m/d. Chen (2004) reported streambed $K_v$ values in three rivers (the Platte, Republican, and Little Blue Rivers) in south-central Nebraska. The average $K_v$ ranges from 15 to 47 m/d with an $L_v$ (length of sediments in the tube) of 40 cm for sandy streambed. The average streambed $K_v$ is 40.2 m/d at seven test sites between Kearney and Central City in the Platte River (Chen, 2005). Song et al. (2007) reported that the average streambed $K_v$ is about 34.4 and 48.2 m/d for two sites between Grand Island and Central City in the Platte River.

The objective of this chapter is to determine the statistical distribution and spatial variation of streambed $K_v$ values at 18 test sites between Kearney and Ashland, about 300 km apart, in the Platte River from south-central to eastern Nebraska (Figure 3.1). This study can provide a detailed picture of site-by-site statistical distribution of streambed $K_v$ along a 300-km segment of the Platte River, and present the possible influences of tributary in controlling streambed permeability at a large scale.

### 3.2 Study Area and Test Sites

The study sites are located along the Platte River from south-central to eastern Nebraska (Figure 3.1). The Platte River has its headwaters in the Rocky Mountains and flows through Nebraska from west to east, and the Loup River and the Elkhorn River merge with the Platte River in eastern Nebraska. The Platte River is usually a braided, sand-bottom stream with many islands (Huntzinger and Ellis, 1993). The primary land uses in the basin consist of dry cropland, irrigated cropland, and pastureland. Dense
vegetation including trees, shrubs and grasses occur in the riparian zone, and cottonwood is the dominant tree (http://www.dnr.ne.gov/LB962/ AnnualReport_2006/LowerPlatteReport.pdf). The Platte River is usually wider than 200 m, becomes wider downstream and can be as wide as 400 m at Ashland. However, the Platte River is generally shallow and the water depth is less than 1 m. The Platte River is an important habitat for a number of endangered river species. In recent years, stream depletion in the Platte River attributed to the extensive use of groundwater for irrigation has become an important issue because it may threaten river habitats.

Figure 3.1 Map showing the study sites. In-situ permeameter tests were performed at 18 test sites (from sites A to R) between Kearney and Ashland, square dots indicating the nearby city or town names.

Five USGS (U.S. Geological Survey) gauge stations record stream stage and streamflow rate in the Platte River within the study area. The stations are USGS 06770200 near Kearney, USGS 06770500 near Grand Island, USGS 06774000 near Duncan, USGS 06796000 near North Bend, and USGS 06801000 near Ashland,
respectively (Figure 3.1). The average stream level and stream discharge for the five stations are shown in Table 3.1. Furthermore, the higher streamflow discharge rate at the North Bend and Ashland stations is a result of the contribution of streamflow from the Loup and Elkhorn Rivers.

Table 3.1 Average stream levels and stream discharge of the Platte River at five USGS gauge stations.

<table>
<thead>
<tr>
<th>Station Location</th>
<th>USGS Code</th>
<th>Mean Stream Level (m)</th>
<th>Stream Level Date Range</th>
<th>Mean Stream Discharge (m$^3$/s)</th>
<th>Discharge Date Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kearney</td>
<td>06770200</td>
<td>651</td>
<td>1987 to 2008</td>
<td>38</td>
<td>1985 to 2008</td>
</tr>
<tr>
<td>Grand Island</td>
<td>06770500</td>
<td>559</td>
<td>1986 to 2008</td>
<td>44</td>
<td>1942 to 2008</td>
</tr>
<tr>
<td>Duncan</td>
<td>06774000</td>
<td>451</td>
<td>1997 to 2008</td>
<td>50</td>
<td>1941 to 2008</td>
</tr>
<tr>
<td>North Bend</td>
<td>06796000</td>
<td>386</td>
<td>1989 to 2008</td>
<td>128</td>
<td>1949 to 2008</td>
</tr>
<tr>
<td>Ashland</td>
<td>06801000</td>
<td>322</td>
<td>1992 to 2008</td>
<td>185</td>
<td>1988 to 2008</td>
</tr>
</tbody>
</table>

Streambed $K_v$ values at 10 of the 18 sites between Kearney and Ashland in the Platte River were presented by Chen (2004), Chen (2005), and Song et al. (2007). Eight new test sites between Schuyler and Ashland (Figure 3.1), about 100 km apart along the Platte River in eastern Nebraska, were selected to perform in-situ permeameter tests in June and July 2008. The eighteen sites were designated as sites A to R between Kearney and Ashland in the Platte River (Figure 3.1). Note that sites J, K, L, N, O, P, and Q correspond to sites P4, P5, P6, P7, P8, P9, and P10 in the chapter 2, respectively. Here, site names J to Q are used in order to make the reading more easily. At each site, 20 to 200 measurements of in-situ permeameter tests were conducted to characterize the streambed variability. Near the City of Fremont, two test sites were selected. One was site M where the permeameter tests were conducted close to the north bank of the Platte River, and the other one was site N where the permeameter tests were conducted near the south bank of the Platte River. At site M, the nearest measurements were only 3.0 m
from the river bank due to the deep water depth. Hence, the measured streambed $K_v$ values at this site may be affected by lower flow conditions compared to site N. Similarly, near the City of Ashland, the tests conducted at site Q in this study were located in the eastern half of the Platte River, while the tests at site R conducted by Chen (2005) were located in the western half of the river. Overall, most of the permeameter tests were conducted in sandy streambed sediments and were over 50 m away from the river bank, thus they could represent the general streambed characteristics in the Platte River.

3.3 Methods

3.3.1 In-situ Permeameter Test

The method and limitations of in-situ falling-head permeameter test are introduced in section 3 of chapter 2 (see pages 29-32). In this study, the number of permeameter tests, the grid spacing between test points, the average $L_v$, and the average water depth are summarized for each of the eight new test sites (sites J to Q) between Schuyler and Ashland in the Platte River, which are shown in Table 3.2. All tubes were 5.0 cm in diameter in this study. The $L_v$ for each measurement of in-situ permeameter tests ranged from 42 to 50.8 cm. Therefore, the ratios of $L_v$ to $D$ are all greater than 5 for the in-situ permeameter tests at all sites, and thus the estimation errors of $K_v$ based on equation 1 in chapter 2 is insignificant.
Table 3.2 Average streambed $K_v$ values, average $L_v$, average water depth, and grid spacing at the eight test sites (sites J to Q) from Schuyler to Ashland in the Platte River in eastern Nebraska.

<table>
<thead>
<tr>
<th>Test Site</th>
<th>Test Date</th>
<th>Number of Tests</th>
<th># of rows, distance between each test point</th>
<th># of columns, distance between each test point</th>
<th>Average $L_v$ (cm)</th>
<th>Average $K_v$ (m/d)</th>
<th>Standard Deviation of $K_v$</th>
<th>Average Water Depth (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>J</td>
<td>June 19, 2008</td>
<td>64</td>
<td>8, 3.0 m</td>
<td>8, 3.0 m</td>
<td>42.0</td>
<td>31.4</td>
<td>5.0</td>
<td>46.8</td>
</tr>
<tr>
<td>K</td>
<td>June 26, 2008</td>
<td>20</td>
<td>4, 30 m</td>
<td>5, 10 m</td>
<td>51.9</td>
<td>32.4</td>
<td>4.8</td>
<td>35.8</td>
</tr>
<tr>
<td>L</td>
<td>July 3, 2008</td>
<td>48</td>
<td>8, 1.5 m</td>
<td>6, 1.5 m</td>
<td>50.8</td>
<td>33.3</td>
<td>7.1</td>
<td>34.4</td>
</tr>
<tr>
<td>M</td>
<td>July 11, 2008</td>
<td>200</td>
<td>4, 1.5 m</td>
<td>50, 1.5 m</td>
<td>50.8</td>
<td>17.7</td>
<td>4.7</td>
<td>44.8</td>
</tr>
<tr>
<td>N</td>
<td>July 10, 2008</td>
<td>49</td>
<td>7, 1.5 m</td>
<td>7, 1.5 m</td>
<td>50.8</td>
<td>29.8</td>
<td>6.2</td>
<td>21.5</td>
</tr>
<tr>
<td>O</td>
<td>July 21, 2008</td>
<td>49</td>
<td>7, 1.5 m</td>
<td>7, 1.5 m</td>
<td>50.8</td>
<td>37.8</td>
<td>4.7</td>
<td>24.4</td>
</tr>
<tr>
<td>P</td>
<td>July 9, 2008</td>
<td>64</td>
<td>8, 1.5 m</td>
<td>8, 1.5 m</td>
<td>50.8</td>
<td>45.2</td>
<td>7.7</td>
<td>34.3</td>
</tr>
<tr>
<td>Q</td>
<td>July 17, 2008</td>
<td>40</td>
<td>4, 1.5 m</td>
<td>10, 1.5 m</td>
<td>50.8</td>
<td>23.4</td>
<td>9.7</td>
<td>16.0</td>
</tr>
</tbody>
</table>
3.3.2 Normality Test and $t$-Test

Normality tests are used to determine whether a set of measurements comes from a normal distribution population. In this study, Jarque-Bera (J-B), Kolmogorov-Smirnov (K-S), Lilliefors, and Shapiro-Wilk (S-W) tests (Sprent, 2001) were applied at the 0.05 significance level. The K-S and S-W tests are commonly used, and the Lilliefors test is an adaptation of the K-S test. The S-W test has requirements for the sample size $N$ (7≤$N$≤2000), while the K-S and Lilliefors tests are preferable to apply for a large sample size $N$ (N≥2000). The J-B test is not good for distributions with short tails, and the K-S and Lilliefors tests are also less powerful than the S-W test. These tests were used to determine whether streambed $K_v$ at each test site is normally distributed. Furthermore, a $t$-test with unequal variance was used to compare the $K_v$ values at two test sites, and this test can determine whether streambed $K_v$ differ significantly between different test sites.

3.3.3 Determination of Independent Samples using an Exponential Model

Streambed $K_v$ values at one site may be dependent on each other since streambed sediments move along the flow direction. Therefore, in addition to the normality test on all the streambed $K_v$ values at one site, the independent samples were determined and the four normality tests were also used to testify the independent sub-datasets of streambed $K_v$ values. Rehfeldt et al. (1992) noted that a population of samples can be reduced to independent samples by taking out the spatially correlated samples, and they introduced a reduction factor to identify both the horizontal and vertical correlation. In this study, an exponential model was used to fit the experimental semi-variogram along the flow direction at each test site. The fitted model provided the correlation scale (Hess et al.,
1992; Genereux et al., 2008; Zhao et al., 2010). If the correlation scale is smaller than the sampling spacing, then the measurements of streambed $K_v$ were regarded as independent; otherwise, the measurements within the correlation scale were eliminated from the sample and thus the remaining streambed $K_v$ values were considered to be independent. Furthermore, because streambed sediments move along the flow direction and then the streambed $K_v$ values across the flow direction were regarded as independent in this study.

The exponential model used in this study did not include a nugget effect, and is written as

$$\gamma = C \left( 1 - \exp \left( \frac{-h}{\lambda} \right) \right)$$

where $\gamma$ is the semi-variogram statistic, $C$ is the variogram sill value, $h$ is the lag distance, and $\lambda$ is the correlation scale.

3.4. Results

3.4.1 Streambed $K_v$ Values between Schuyler and Ashland in the Platte River

Previous studies showed that the vertical hydraulic gradient (VHG) may vary spatially horizontally at the same depth across the streambed at nearby locations (Chen et al., 2009; Leek et al., 2009). Chen et al. (2009) noted that the positive and negative VHG values occur between two locations only several meters apart for the streambed sediments in the Elkhorn River of Nebraska, which indicates the significant presence of downward and upward flux at water-streambed interface. In this study, the VHG values at the test sites in the Platte River are very small (all less than 0.02), thus using the surface water-level at the streambed surface as the initial hydraulic head at the measurement point does
not affect the accuracy of the estimation of streambed $K_v$ values greatly.

The average streambed $K_v$ values and standard deviation of $K_v$ at each of the eight new test sites are shown in Table 3.2. The average $L_v$ from sites J to Q in this study is about 49.8 cm, which is slightly larger than that from sites A to I and site R (40.7 cm) from previous studies (Chen, 2004; Chen, 2005; Song et al., 2007). At sites J, K, and L, the average streambed $K_v$ values were similar. The streambed $K_v$ values at site M are lower than those at site N, which may be due to that site M is closer to the river bank and the presence of the riparian trees and vegetation can induce nutrients at streambed surface and thereby reducing the streambed $K_v$. Also, site N was closer to the center of the Platte River than site M, thus the flow velocity was higher at site N, and thereby affecting the streambed $K_v$ values at two the sites. Furthermore, the average streambed $K_v$ values at the two sites were lower than those values at sites J, K, and L; whereas the average streambed $K_v$ values at sites O and P were higher than those values at other test sites.

The test site near Ashland (site Q) in this study is different from the Ashland site (site R) of Chen (2005). He performed in-situ permeameter tests along four transects on the west half of the Platte River, while the permeameter tests in this study were conducted on the east half of the Platte River, about 200 m apart from the test locations of Chen (2005). The streambed $K_v$ values range from 2.9 to 41.9 m/d with an average $K_v$ of 23.4 m/d, while Chen (2005) reported that the average streambed $K_v$ is 16.8 m/d at 40 test points. The average $K_v$ value at site Q is lower than those values at other new test sites in this study except for site M.

Out of the eight new test sites, the $K_v$ values at site Q have the largest standard deviation, while the $K_v$ values at site M have the smallest standard deviation (Table 3.2).
Large standard deviation of $K_v$ indicates that the streambed $K_v$ values can vary significantly within the same site, especially at sites J, P, and Q (Table 3.2). On the whole, the standard deviations of $K_v$ values at the eight sites are slightly different, and they are smaller than those from sites A to I between Kearney and Central City in the Platte River (Chen 2004; Chen, 2005; Song et al., 2007; Table 4.3). This difference is probably because (1) a larger number of permeameter tests are conducted at the eight new sites in this study and (2) the tests for these eight new sites are conducted in regularly spaced but closely located points compared to the tests along the across-channel transect for the sites A to I between Kearney and Central City (Chen 2004; Chen, 2005; Song et al., 2007). Streambed sediments mainly move along the flow direction and thus it can be anticipated that larger heterogeneity in streambeds exists along a transect across the channel and less heterogeneity in streambeds of smaller scale plots.

### 3.4.2 Statistical Distribution of All Streambed $K_v$ Values at Each Test Site along the Platte River

The histograms of the streambed $K_v$ values and the cumulative distributions on normal probability plots at the eight new sites (sites J to Q) from Schuyler to Ashland in the Platte River are shown in Figure 3.2 (a-h). The J-B and S-W tests indicated that the streambed $K_v$ values are in normal distribution at the eight test sites at the 0.05 significance level except for site N (Table 3.3), while the Lilliefors and K-S tests implied non-normal distribution of streambed $K_v$ at sites J and Q as well as site N. At site N, all four tests suggested that the streambed $K_v$ values are not in normal distribution at the 0.05
Table 3.3 Average streambed \(K_v\) values and length of tested streambed sediment at the eighteen test sites. Normality tests by the Jarque-Bera (J-B), Kolmogorov-Smirnov (K-S), Lilliefors, and Shapiro-Wilk (S-W) tests indicate whether streambed \(K_v\) values at these sites are in normal distribution (‘Yes’ means streambed \(K_v\) is normally distributed while ‘No’ implies not).

<table>
<thead>
<tr>
<th>Test Site</th>
<th>Number of Tests</th>
<th>Average (K_v) (m/d)</th>
<th>Average (L_v) (cm)</th>
<th>Standard Deviation (K_v)</th>
<th>Normal Distribution by J-B Test</th>
<th>Normal Distribution by K-S Test</th>
<th>Normal Distribution by Lilliefors Test</th>
<th>Normal Distribution by S-W Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>8</td>
<td>32.5</td>
<td>40.0</td>
<td>23.1</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>B</td>
<td>10</td>
<td>32.7</td>
<td>38.1</td>
<td>9.6</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>C</td>
<td>9</td>
<td>38.1</td>
<td>38.2</td>
<td>10.7</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>D</td>
<td>16</td>
<td>45.9</td>
<td>42.4</td>
<td>14.9</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>E</td>
<td>21</td>
<td>40.7</td>
<td>40.0</td>
<td>14.9</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>F</td>
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<td>34.4</td>
<td>48.0</td>
<td>16.7</td>
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<td>Yes</td>
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<td>Yes</td>
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<tr>
<td>G</td>
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<td>48.2</td>
<td>50.3</td>
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<td>Yes</td>
</tr>
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<td>8</td>
<td>46.7</td>
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<td>Yes</td>
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<tr>
<td>A to I</td>
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<td>42.7</td>
<td>16.6</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>J</td>
<td>64</td>
<td>31.4</td>
<td>42.0</td>
<td>5.0</td>
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<td>No</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>K</td>
<td>20</td>
<td>32.4</td>
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<td>4.8</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>L</td>
<td>48</td>
<td>33.3</td>
<td>50.8</td>
<td>7.1</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>M</td>
<td>200</td>
<td>17.7</td>
<td>50.8</td>
<td>4.7</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>N</td>
<td>49</td>
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<td>50.8</td>
<td>6.2</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>N excluding two outliers</td>
<td>47</td>
<td>29.0</td>
<td>50.8</td>
<td>4.7</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>O</td>
<td>49</td>
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<td>50.8</td>
<td>4.7</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>P</td>
<td>64</td>
<td>45.2</td>
<td>50.8</td>
<td>7.7</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>J to P</td>
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<td>28.3</td>
<td>49.7</td>
<td>11.3</td>
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<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>Q</td>
<td>40</td>
<td>23.4</td>
<td>50.8</td>
<td>9.7</td>
<td>Yes</td>
<td>No</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>R</td>
<td>48</td>
<td>16.8</td>
<td>40.0</td>
<td>8.7</td>
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<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Q and R</td>
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<td>19.8</td>
<td>44.9</td>
<td>9.7</td>
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<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
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<tr>
<td>All Test Sites</td>
<td>689</td>
<td>29.2</td>
<td>48.0</td>
<td>13.4</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
</tbody>
</table>
Site J

Site K

Site L

Site M

Normal Probability Plot (Site J)

Normal Probability Plot (Site K)

Normal Probability Plot (Site L)

Normal Probability Plot (Site M)
Figure 3.2 Histograms and normal probability plots of streambed $K_v$ from sites J to Q between Schuyler and Ashland in the Platte River (a) site J; (b) site K; (c) site L; (d) site M; (e) site N; (f) site O; (g) site P; and (h) site Q.
significance level, since the $p$-values are all smaller than 0.05. There are two outliers of larger $K_v$ values at site N, probably due to the presence of coarser sand sediments locally. If the two largest $K_v$ values were considered as outliers and eliminated from the sample, the remaining 47 $K_v$ values at site N are normally distributed from all four normality tests, which indicates that the streambed sediments are mainly homogeneous at this site. Furthermore, the 48 streambed $K_v$ values at site R determined by Chen (2005) were normally distributed according to the Lilliefors and K-S tests. When the streambed $K_v$ values at both sites Q and R near Ashland are combined as a single dataset, the 88 $K_v$ values were still normally distributed according to all four normality tests (Table 3.3).

Chen (2005) noted that streambed $K_v$ values are normally distributed for the combined dataset at sites A, B, C, D, E, H, and I between Kearney and Central City in the Platte River (Figure 3.1). In this study, all four normality tests are performed for streambed $K_v$ values at each of the eight test sites (sites A to I) except for site H, which had only 4 measurements (Chen 2004; Chen, 2005; Song et al., 2007). The results indicate that streambed $K_v$ is normally distributed at these individual sites (Table 3.3). Song et al. (2007) reported the streambed $K_v$ values at sites F and G, but they did not perform a statistical distribution analysis of $K_v$ values. All four normality tests illustrate that streambed $K_v$ values are in normal distribution at sites F and G as well (Table 3.3). When the $K_v$ values obtained from sites F and G were combined with those values reported by Chen (2005), the new dataset of streambed $K_v$ values from sites A to I (Chen 2004; Chen, 2005; Song et al., 2007; Figure 3.1) was in normal distribution, which is attributed to the fact that the Platte River has no tributaries between Kearney and Central City (Figure 3.1) and thus the streambed sediments within this river reach are well
distributed and belonged to a single population of hydraulic conductivity values.

When all the \( K_v \) values from sites J to P are combined, the normality tests indicated that the data are not in normal distribution (Table 3.3), which may be a result of different hydrogeological processes, including geological conditions, geomorphic history, and physical transport processes (Hoey and Bluck, 1999; Rice and Church, 1998), controlling the structure of channel sediments at individual sites. Streambed \( K_v \) is also not normally distributed for the combined data of all the 689 measurements in the Platte River from sites A to R between Kearney and Ashland which is concluded by the four normality tests (Table 3.3).

### 3.4.3 Statistical Distribution of Independent Streambed \( K_v \) Values at Each Test Site along the Platte River

The experimental semi-variograms of \( K_v \) and fitting exponential models along the flow direction at the eight new test sites (sites J to Q) in the Platte River are shown in Figure 3.3. At site L, the measured semi-variograms of \( K_v \) increase with the lag distance gradually and the fitted exponential model showed that the values of \( \lambda \) (correlation scale) and \( C \) (variogram sill value) cannot be determined uniquely when all the semi-variogram values were used. Thus, we chose the first three measured semi-variograms of \( K_v \) to fit an exponential model, and the fitted values of \( C \) and \( \lambda \) were about 45.2 and 2.1 m, respectively (Figure 3.3c). At sites K, O, P, and Q, the correlation scale was less than 1.5 m which corresponds to the distance between the test points. However, the correlation scale is larger than the sampling spacing at each of sites J, L, M, and N. The \( K_v \) values within the correlation scales were taken out and normality tests were performed on the
remaining independent sets of streambed $K_v$ values. At sites J, L, and M, the correlation scales are less than twice the sample spacing at each site, two independent sets of $K_v$ values can be generated after the correlated samples were removed; while at site N, three independent sets of $K_v$ values are generated.

All the datasets of independent streambed $K_v$ values are in normal distribution except for one dataset at site J which was identified by the K-S and Lilliefors tests and one dataset at site N which was identified by the J-B and S-W tests (Table 3.4). However, when the largest $K_v$ value in this dataset at site N is regarded as an outlier and taken out, the remaining 13 values are normally distributed determined by all four normality tests. Therefore, the streambed $K_v$ values at each of the eight test sites (sites J to Q) between Schuyler and Ashland in the Platte River can be regarded as normally distributed, which indicated that the streambed $K_v$ values cluster around the average $K_v$ value and using the average $K_v$ value obtained from a large number of measurements to represent the streambed $K_v$ characteristics was appropriate at these sites in the Platte River.
Figure 3.3 Semi-variogram of $K_v$ along the flow direction from sites J to Q between Schuyler and Ashland in the Platte River (a) site J; (b) site K; (c) site L; (d) site M; (e) site N; (f) site O; (g) site P; and (h) site Q.
Table 3.4 Average streambed $K_v$ values of the independent datasets at sites J, L, M, and N. Normality tests by the Jarque-Bera (J-B), Kolmogorov-Smirnov (K-S), Lilliefors, and Shapiro-Wilk (S-W) tests indicate whether streambed $K_v$ values at these sites are in normal distribution (‘Yes’ means streambed $K_v$ is normally distributed while ‘No’ implies not).

<table>
<thead>
<tr>
<th>Test Site and Set of Independent Samples</th>
<th>Number of Samples</th>
<th>Average $K_v$ (m/d)</th>
<th>Normal Distribution by J-B Test</th>
<th>Normal Distribution by K-S Test</th>
<th>Normal Distribution by Lilliefors Test</th>
<th>Normal Distribution by S-W Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sub-dataset 1 at site J</td>
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<td>32.3</td>
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<td>No</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 2 at site J</td>
<td>32</td>
<td>30.4</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 1 at site L</td>
<td>24</td>
<td>32.3</td>
<td>Yes</td>
<td>Yes</td>
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<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 2 at site L</td>
<td>24</td>
<td>34.3</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 1 at site M</td>
<td>100</td>
<td>17.8</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 2 at site M</td>
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<td>17.6</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 1 at site N</td>
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<td>28.3</td>
<td>Yes</td>
<td>Yes</td>
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<td>Yes</td>
</tr>
<tr>
<td>Sub-dataset 2 at site N</td>
<td>14</td>
<td>30.9</td>
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<td>Yes</td>
<td>Yes</td>
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<tr>
<td>Sub-dataset 2 at site N (excluding the largest value)</td>
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<td>29.4</td>
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<tr>
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<td>14</td>
<td>31.0</td>
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</tr>
</tbody>
</table>
3.5 Discussion

3.5.1 Spatial Variation of Streambed $K_v$ Values along the Platte River

At the eight new test sites in this study, the $t$-test suggest that the $K_v$ values are similar between sites J, K, and L with that all the $p$-values were larger than 0.05. Site N also had similar $K_v$ values with sites J and K but not site L. The $K_v$ values at sites M and Q were significantly lower than those at all other sites ($p<0.0001$). Site M has the lowest average streambed $K_v$ value, while the average streambed $K_v$ value at site P is the highest among the eight test sites.

Comparing the average streambed $K_v$ values from sites A to I between Kearney and Central City (Figure 3.1) in the Platte River (Chen, 2004; Chen, 2005; Song et al., 2007), larger average streambed $K_v$ values occurred at sites D, E, G, H, I, and P, which were all greater than 40 m/d (Figure 3.4). The average streambed $K_v$ values at the eight new test sites (sites J to Q) between Schuyler and Ashland in the Platte River are all lower than those between Kearney and Central City, except for sites O and P, which may be attributed to localized coarse streambed sediments where the permeameter tests were conducted. Nevertheless, the average streambed $K_v$ value from sites J to Q in the Platte River was 27.1 m/d, which is lower than that from sites A to I in the Platte River (41.0 m/d; Chen, 2004; Chen, 2005; Song et al., 2007). The increased presence of agricultural crop close to the stream and the presence of loess and till in eastern Nebraska are an important factor for the decreasing tendency of streambed $K_v$ along the Platte River.

In addition, the characteristic of normal distribution of streambed $K_v$ in the Platte River is different from the distribution of $K_v$ reported in the West Bear Creek in North Carolina (neither normal nor log-normal) (Genereux et al., 2008), from the log-normal
distribution of streambed \( K_v \) in the Indian Creek in Philadelphia, Pennsylvania reported by Ryan and Boufadel (2007), from the bimodal distribution of streambed \( K_h \) in the Colorado River (Springer et al., 1999), and from the log-normal distribution for aquifer materials (Freeze, 1975). Consequently, streambed hydraulic conductivities have different ranges of values at different rivers, and their statistical distribution is different than the aquifer materials (log-normal). For a large braided river, normal distribution can represent the statistical characteristics of the less heterogeneous hydraulic conductivity of shallow streambeds.

Figure 3.4 Boxplot of streambed \( K_v \) values at the 18 test sites (from sites A to R) between Kearney and Ashland in the Platte River in Nebraska (Chen 2004, 2005; Song et al. 2007). Box indicates the upper and lower quartile, the dash horizontal line indicates the median value, and the solid horizontal line indicates the mean value.
3.5.2 Effects of Tributaries on Streambed $K_v$ Variability

Usually the overall grain size of bulk streambed sediments declines with the distance downstream due to abrasion and sorting, and selective transport (Surian, 2002; Frings, 2008) and a downstream gravel-sand transition occurs (Singer, 2008). Additionally, sediment sources of the tributaries play a significant role in controlling the grain-size pattern change for river bed sediments (Rice, 1998). In this study, two other major rivers merge with the Platte River, which can induce additional effects on the distribution of streambed $K_v$ values because they drain in the significantly different regions in terms of sediment type, distribution, and erodability (Huntzinger and Ellis, 1993; Peterson et al., 2008). The Loup River merges with the Platte River near Columbus, which is upstream of site J, while the Elkhorn River merges with the Platte River at where it is only 10 km upstream of sites Q and R near Ashland. The Loup River originates from the Nebraska Sand Hills, which has about 49695 km$^2$ in drainage. Previous studies suggested that the Nebraska Sand Hills is an important factor for the distribution of loess deposits downwind of the Sand Hills. The Sand Hills either serves as a sediment transport pathway and allows loess deposits to be carried away (Mason, 2001), or generates silt sized sediments by abrasion and ballistic impacts under strong winds (Muhs, 2004). The Loup River thus carries these fine-grained sediments, which can mix with the sediments moving downstream in the Platte River and result in lower streambed $K_v$ values at the eight test sites.

At site Q near Ashland, the streambed $K_v$ values were lower than those values at other sites in this study and similar results were reported by Chen (2005). The Elkhorn River is a meandering river (http://www.agiweb.org/environment/publications/mapping/
graphics/Nebraska1.pdf), so it might have lower streambed $K_v$ values than the Platte River. First, Huntzinger and Ellis (1993) noted that low-permeability glacial-till deposits occur in the subsurface of the Elkhorn River; and second, Song et al. (2009) found that streambed $K_v$ values at the West Point site in the Elkhorn River (about 67 km upstream of the confluence with the Platte River) were on average 20.7 m/d. These relatively fine sediments may be carried from the Elkhorn River downstream, and deposited in the Platte River thus resulting in additional damped effects on streambed $K_v$ values as well as the Loup River at sites Q and R near Ashland.

### 3.5.3 Streambed $K_v$ Variability across the Channel

Streambed $K_v$ values can vary laterally across the river channel. At site R, Chen (2005) conducted permeameter tests along four transects on the west half of the Platte River, and he found that the $K_v$ values tend to increase towards locations that are further from the river bank. Also, Genereux et al. (2008) reported that the $K_v$ value at the center of the channel is usually higher than the values close to the river bank. In this study, streambed $K_v$ values at site M were much lower than those at site N, given that both sites were located at different sides of the Platte River. However, site M has the nearest measurements only 3.0 m to the river bank where flow velocity was small, while site N was over 50 m away from the river bank. The center of the river usually has higher flow velocity than the sides of the channel. A larger $K_v$ value occurs in the channel sediments where the flow velocity is generally higher, since fine-grained sediments can be washed away by higher flows and they may deposit again in the area with lower flow velocity.
3.6 Summary and Conclusions

In-situ permeameter tests were conducted at eighteen test sites between Kearney and Ashland in the Platte River from south-central to eastern Nebraska. The streambed $K_v$ values at the eighteen sites (sites A to R) may be categorized into three groups in terms of downstream fining and tributary inputs. The first group is the streambed $K_v$ values from site A to I between Kearney and Central City in the Platte River. Within this area, the Platte River has no tributaries and the average streambed $K_v$ value was 41.0 m/d. The second group is the streambed $K_v$ values from site J to P in the Platte River, and the average streambed $K_v$ value was 28.3 m/d within this area. The Loup River merges with the Platte River near Columbus, which is upstream of site J. The Loup River originates from the Nebraska Sand Hills and can carry the fine-grained sediments generated by the Sand Hills to the Platte River. The mixing of the fine-grained sediments from the Loup River and sediments from the downstream Platte River could explain the lower-$K$ streambed sediments that occur downstream of the confluence. Also, the increased presence of agricultural crop close to the stream and the presence of loess and till in eastern Nebraska are another important factor for the decreasing tendency of streambed $K_v$ along the Platte River. The third and final group is the streambed $K_v$ at sites Q and R near Ashland in the Platte River. Another river, the Elkhorn River, as a meandering river, merges with the Platte River at where it is only 10 km upstream of both the Ashland test sites. Low-permeability glacial-till deposits occur in the subsurface of the Elkhorn River (Huntzinger and Ellis, 1993), and permeameter tests suggest a lower $K_v$ value for streambed sediments in the Elkhorn River (Song et al., 2009), which may contribute additional fine-grained (lower-$K$) sediments deposited downstream of the Platte River.
The average streambed $K_v$ value was 19.8 m/d for sites Q and R, which was much lower than those values at other sites, except for site M.

Streambed $K_v$ values were normally distributed at nearly each test site in the Platte River from south-central to eastern Nebraska, except for site N. However, when the two largest $K_v$ values were regarded as outliers of locally coarser sand sediments and eliminated from the sample, the remaining streambed $K_v$ was in normal distribution at site N. Additionally, when the correlated $K_v$ values were removed from the datasets collected from gridded sampling plots, the remaining independent sub-datasets of streambed $K_v$ values were still in normal distribution at each of the eight new test sites. Moreover, the combined dataset of streambed $K_v$ values from site A to I between Kearney and Central City, about 200 km apart along the Platte River, were normally distributed, which is due to the fact that the Platte River has no tributaries in-between and thus the streambed sediments were well distributed in the Platte River in this reach and belonged to a single population of hydraulic conductivity values. On the other hand, the combined dataset of streambed $K_v$ values from site J to R between Schuyler and Ashland, about 100 km apart along the Platte River, were not in normal distribution. Within this lower reach, the mixture of three sediment sources from the upstream Platte River, the Loup River, and the Elkhorn River leads to a wide range of variations in streambed vertical hydraulic conductivity.

Overall, this chapter testified the fifth hypothesis: shallow streambed sediments of braided rivers are more uniform as compared to the underlying aquifer materials, thus their hydraulic conductivities are not log-normally distributed as are typical aquifer hydraulic conductivities.
Chapter 4  Field and Numerical Study of Stream-Aquifer Interactions Across an Unclogged Streambed in the Lower Platte River Basin, Nebraska

4.1 Introduction

Groundwater withdrawal near a stream can lower the stream stage and affect the streamflow discharge when the stream and aquifer are hydrologically connected. Extensive research has been conducted to investigate the effects of groundwater abstraction, especially in semi-arid or dry environmental regions (Rood et al., 2003; Stromberg et al., 2005; Zekster et al., 2005; Rodriguez et al., 2006; Wen and Chen, 2006; Peterson et al., 2008; Zume and Tarhule, 2008). Streamflow depletion has become an important issue because extensive use of groundwater may threaten river habitats, and thus it is necessary to evaluate the potential effects of groundwater extraction on streamflow for water resources management and prediction.

In the regional High Plains aquifer, groundwater pumping wells are densely located and a large amount of water has been withdrawn for irrigation use. As a result, streamflow depletion may induce conflicts over water use demands between surface water and groundwater user groups among different states. Wen and Chen (2006) applied nonparametric techniques to analyze the impacts of groundwater irrigation on the streamflow of major rivers in Nebraska. They found that the decreasing trends of streamflow in these rivers have a simple inverse relationship with the increasing trend of irrigation wells and the decreasing trend of groundwater levels. Furthermore, they noted that streamflow depletion in the Republican River Basin is attributed to the fact that a large amount of groundwater was withdrawn for irrigation uses in the neighboring states.
of Nebraska such as Kansas and Colorado. Peckenpaugh et al. (1995) developed a three-dimensional groundwater flow model to simulate the effects of groundwater withdrawals on water levels in the High Plains aquifer in the Upper Republican Natural Resources District of Nebraska. The model was calibrated under both the steady-state and transient conditions, and then they used two different pumping scenarios to predict the water-level declines for the period of 1989-2030. The first scenario was that pumpage was constant at a rate necessary to supply a crop’s consumptive irrigation requirement; whereas the second scenario was that pumpage was constant at the rate necessary to apply 13 inches of water on irrigated crops during the irrigation season. They found that the simulated water-level declines were larger for the second scenario and it can be as much as 90 ft by 2030 in northwestern Chase County of Nebraska. Peckenpaugh and Dugan (1983) developed a two-dimensional groundwater flow model to simulate the impacts of groundwater withdrawals for irrigation in the Central Platte and Lower Loup natural resources districts of Nebraska. They examined three management alternatives and found that water-levels would decline even without additional groundwater development. However, their model did not incorporate the streambed characteristics, and instead assumed that the entire groundwater system was an isotropic and unconfined aquifer. Peterson et al. (2008) developed a regional groundwater flow model to simulate the effects of irrigation pumping on baseflow in the Elkhorn and Loup River Basins in Nebraska. They noted that these effects were minimal before 1970 but increased steadily after 1970. This may be a result of the fact that irrigation using groundwater became more common throughout the entire study area after 1970 as compared to the limited areas near the Platte River before 1970. Consequently, a more accurate streamflow
analysis in the Lower Platte River basin is beneficial to the understanding and quantifying stream-aquifer interactions with the help of the new findings added to a groundwater flow model.

A variety of studies have reported on streamflow depletion analysis, which includes analytical solutions by Theis (1941), Glover and Balmer (1954), Hantush (1965), Hunt (1999), Butler et al. (2001), Lough and Hunt (2006), and Sun and Zhan (2007). Also, analog models were used to simulate the effects of groundwater withdrawals on the streamflow (Emery, 1966). In more recent research, numerical models have been used to quantify the streamflow depletion instead of using analytical solutions (Sophocleous et al., 1995; Chen and Shu, 2002; Chen and Chen, 2003; Hunt and Scott, 2005; Rodriguez et al., 2006; Zume and Tarhule, 2008). Numerical models can provide more flexibility in dealing with real-world hydrologic conditions considering groundwater evapotranspiration (ET), aquifer heterogeneity and anisotropy, partial penetrating streams, heterogeneous hydraulic properties of streambed sediments, etc. Finite element and finite difference methods are two mostly used numerical approaches to simulate stream-aquifer interactions. MODFLOW, a finite-difference groundwater flow model approach, is often employed to evaluate the impacts of groundwater exploitation on baseflow variations. The river package and the stream package embedded in MODFLOW are two alternative packages to quantify streamflow depletion (Sophocleous et al., 1995; Rodriguez et al., 2006; Zume and Tarhule, 2008).

Generally, the concept of streambed conductance (Sophocleous et al., 1995) or river coefficient (Rushton, 2007) is used to characterize stream-aquifer interactions. They are both calculated as (McDonald and Harbaugh, 1988):
$C_{riv} = \frac{K_v L_{riv} W}{M}$

where $K_v$ is the vertical hydraulic conductivity of the streambed sediments, $L_{riv}$ is the length of the stream channel, $W$ is the width of the stream channel, and $M$ is the thickness of the streambed sediments. The streambed leakance is a vital parameter within the construct of streambed conductance, which is equal to $K_v/M$. The streambed leakance is the most uncertain parameter in streambed conductance because the width and length of the stream channel can be measured directly.

Streambed conductance is used widely to integrate the interactions between surface water body and groundwater system in modeling studies, which represents the resistance to flow between the surface water body and the groundwater caused by streambed (McDonald and Harbaugh, 1988). The concept of streambed conductance is usually based on an assumption that there is a low-permeability streambed clogging layer whose hydraulic conductivity is smaller than that of the underlying aquifer. However, as discussed in chapter 2, the shallow streambed is mainly permeable over the gaining reaches of the braided and meandering rivers. Consequently, the assumption of a clogging layer at the channel surface in the numerical modeling approaches can yield inaccurate estimations of streambed leakance and then induce errors in evaluating stream-aquifer interactions.

Use of analytical solutions such as Hunt (1999) solution is still a popular approach for calculating stream depletion caused by groundwater abstraction, thus it is important to be able to estimate a reasonable streambed leakance value to describe the stream-aquifer interaction. In most previous studies, streambed leakance is estimated based on an aquifer test (Hunt et al., 2001) or calibrated using stream-aquifer modeling (Chen and
Chen, 2003; Zume and Tarhule, 2008; Rodriguez et al., 2006). However, large over or under estimation of the hydrologic connectivity is still possible due to the insensitivity of streambed leakance to hydraulic head in a regional model and the lack of characterization of channel sediment stratification.

This study aims to evaluate the impacts of groundwater irrigation on streamflow in the Platte River, and also to estimate the streambed leakance of the Platte River using numerical and field techniques. The streambed leakance value is useful to assess the interactions between the Platte River and its adjacent aquifers when analytical solutions are employed. In this study, the stratification of channel sediments in the lower reach of the Platte River was determined based on the obtained electrical conductivity logs using Geoprobe direct-push technique and streambed hydraulic conductivities from in-situ permeameter tests, which provided new information in determining the stream-aquifer interactions in this area more accurately. Subsequently, a three-dimensional groundwater flow model was developed using Visual MODFLOW to simulate the interactions between the Platte River and its adjacent aquifers. A constant head boundary was used to represent the Platte River in the model. The model was calibrated using a trial-and-error method according to the changes of groundwater levels in 40 observation wells for the period from January 1994 to December 2004. Then, the calibrated hydraulic parameters were applied to calculate the exchange between the Platte River and its adjacent aquifers and further to determine the stream depletion rate. Finally, the River package with different streambed leakance values was simulated instead of the constant head boundary to represent the Platte River in the model, using the cumulative baseflow as the calibrated target to determine the appropriate streambed leakance value.
4.2 Study Area

4.2.1 General Information

The study area (Figure 4.1) is part of the Lower Platte River Basin, which is located in eastern Nebraska. The Platte River is the major river flowing through the study area (Figure 4.1). The study area covers about 2,090 km$^2$, within which numerous groundwater wells have been developed for irrigation use. The primary land uses in the study area consists of dry cropland, irrigated cropland, and pastureland (Figure 4.2). The physiographic of the study area consists of the valleys, plains, rolling hills, bluffs and escarpments (Figure 4.3). A topographic elevation map (DEM) of the study area is shown in Figure 4.4, which shows the pronounced difference in elevation between the Platte River valley and upland areas. In addition, the elevation of ground surface decreases from the west to the east in the study area. The poorly drained soil distributes in the Platte River valley whereas the well drained soil in the upland area (Figure 4.5). Figure 4.6 shows the depth to water table in the study area, which is apparent that the Platte River valley has the shallowest depth to water table while the upland area has the deepest depth to water table.

4.2.2 Geology

In the study area, the principal aquifer consists of saturated unconsolidated sediments and alluvium of Quaternary age, and the Tertiary Ogallala Group (http://www.dnr.ne.gov/LB962/AnnualReport_2006/LowerPlatteReport.pdf).
Figure 4.1 Map showing the study area, and the locations of the USGS gauge stations, the EC log sites, and the permeameter test sites.
Figure 4.2 Land cover and land use for the study area (land cover shapefile is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).
Figure 4.3 The physiographic of the study area (physiographic shapefile is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).
Figure 4.4 Digital Elevation Model (DEM) surface elevation map of the study area (DEM shapefile is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).
Figure 4.5 The soil drainage of the study area (soil drainage shapefile is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).
Figure 4.6 The depth to water of the study area (depth to water shapefile is obtained from [http://snr.unl.edu/data/geologysoils/index-geologysoils.asp](http://snr.unl.edu/data/geologysoils/index-geologysoils.asp)).
4.2.2.1 Quaternary Age Deposits

Quaternary Age deposits cover almost the entire study area and consist of sand, gravel, silt, and clay. These deposits overlie the Ogallala Group and the Cretaceous shale. The Quaternary Age deposits are unconsolidated due to the occurrence of erosion and soil formation and lack of cementation. The thickness of the deposits varies from a few meters in the Platte River valley area to more than 60 meters in the northern and southern upland areas.

4.2.2.2 Ogallala Group

Test-hole and irrigation well logs show that the Ogallala Group underlies the study area. Most of the groundwater irrigation wells are screened in this group and large amounts of water are pumped in the irrigation season. The Ogallala Group consists mostly of sand and gravel with occasional silt and clay, as well as interbedded limestone and sandstone at some locations. The thickness of the Ogallala Group is greater beneath the upland area, reaching up to as much as 60 meters in the Saunders County. The thickness becomes less in the Platte River valley areas. Moreover, in some areas the Ogallala Group is only a few meters thick due to the higher elevation of the Cretaceous bedrock at these locations. The Ogallala Group lies below the Quaternary Age deposits, and the sediments are partially consolidated due to compaction and cementation. Thus, the hydraulic conductivity of the Ogallala Group is usually lower than that of the alluvial deposits. The base elevation of the Ogallala Group is higher in the west of the study area and lower in the east, which is related to the topography of the underlying Cretaceous bedrock and the Rocky Mountain uplift. The Cretaceous shale underlaying the Ogallala
Group and the Quaternary Age deposits is considered as the confining unit, which contains thicker shale and thinner shaley chalk.

4.2.2.3 Alluvial Deposits

Alluvial deposits occur in the Platte River valley and some other river valleys in the study area, and its thickness of the alluvial deposits varies from a few meters to 20 meters. The alluvial deposits lie above the Cretaceous shale, and they contact laterally with the Ogallala Group and the Quaternary Age deposits. Similar to the Ogallala Group, the alluvial deposits consist mostly of sand and gravel with small percentage of silt and clay sediments. However, the alluvial deposits are mostly unconsolidated because they were not cemented. Therefore, in contrast to the Ogallala Group, the hydraulic conductivity of the alluvial deposits is generally higher due to less compaction and the coarser grain size, and the absence of cementation and iron encrustation as well.

4.2.3 Hydrology

4.2.3.1 Precipitation

Average monthly precipitation measurements at twelve weather stations were obtained from the High Plains Regional Climate Center (http://www.hprcc.unl.edu). The average annual precipitation was about 737 mm/yr (29 in/yr) from 1950 to 2004. The 55-year average monthly precipitation chart for the precipitation stations at David City, Fremont, Schuyler, and Wahoo is shown in Figure 4.7. Monthly precipitation data from the four sites shows that most of the precipitation occurs from April to September, which
occupies about 75% of the total annual amount.

Figure 4.7 Average monthly precipitation at four stations in the study area (precipitation data are obtained from http://www.hprcc.unl.edu).

4.2.3.2 Stream Level and Streamflow Discharge in the Platte River

The Platte River flows through the study area. Streamflow in the Platte River is supplied primarily by precipitation, snowmelt, and groundwater. The Platte River and regional groundwater flow from westerly to southeasterly in the study area. Three USGS (US Geological Survey) gauge stations (Figure 4.8) along the Platte River were selected to obtain the historical stream level and streamflow discharge records. The three stations are USGS 06796000 at North Bend, USGS 06796500 at Leshara, and USGS 06796550 at Venice. The station at Leshara is approximately 39 km downstream of the station at North Bend. The Platte River has no large tributaries between the three gauge stations. The mean stream level was about 386, 350, and 336 m at the stations of North Bend,
Leshara, and Venice from January 1950 to December 2004, respectively. The station at Venice does not have enough streamflow discharge data, whereas the mean annual streamflow discharge rate was about 128 m$^3$/s at the North Bend station between 1949 and 2008 and 202 m$^3$/s at the Leshara station between 1995 and 2008. Figure 4.9 shows the daily streamflow discharge over different time periods at the two USGS gauge stations at North Bend and Leshara.

Figure 4.8 The three USGS stream gauge stations in the study area (shapefile of USGS gauge stations is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).
4.2.3.3 Groundwater Pumping

There are more than 1,500 registered pumping wells in the study area (Figure 4.10), which are active and have a pumping rate of greater than 50 gpm (936 m³/day) (http://www.dnr.ne.gov). More than 90% of these pumping wells are used to irrigation. Usually the irrigation period is seasonal from June to August, and it is also cyclic. Moreover, 40 observation wells (Figure 4.11) located in the study area have historical groundwater level records which were used to calibrate the groundwater flow model. Figure 4.12 shows the water-level elevation in the study area at year 1979 and 1995.
Figure 4.10 The groundwater irrigation wells in the study area (irrigation wells shapefile is obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp).

Figure 4.11 The observation wells in the study area (irrigation wells shapefile is obtained from http://water.usgs.gov and provided by Lower Platte River North Natural Resources District of Nebraska).
Figure 4.12 Map of Groundwater-levels across the study area in (a) 1979; and (b) 1995. (groundwater level shapefiles are obtained from http://snr.unl.edu/data/geologysoils/index-geologysoils.asp)
4.2.4 Previous Studies on Hydraulic Properties in the Study Area

Peckenpaugh and Dugan (1983) developed a two-dimensional groundwater flow model to simulate the impacts of groundwater irrigation in the Central Platte and Lower Loup natural resources districts of Nebraska. They assigned a hydraulic conductivity value to each lithologic unit in the test-hole logs according to the grain size of the material and its degree or sorting and/or silt content. In their model, they divided the aquifer into two different zones: Quaternary Age materials and the Ogallala Group. The hydraulic conductivity ranged from 0.3 to 91 m/d (1 to 300 ft/day) for the Quaternary Age materials and 0.3 to 24 m/d (1 to 79 ft/day) for the Ogallala Formation. Moreover, they suggested that the specific yield was from 0.16 to 0.26 for the Quaternary Age materials and from 0.12 to 0.21 for the Ogallala Formation.

Layne-Western (1981) performed a pumping test in the Platte River valley near Fremont, and used the Jacob Non-Equilibrium method to calculate the aquifer hydraulic properties. The average hydraulic conductivity was about 215 m/d (707 ft/day) and the specific storage was about 0.0064 for the alluvial aquifer (Layne-Western, 1981). Layne Geosciences (2002) constructed a groundwater flow model to determine the impacts of additional 3000 gpm withdrawal on existing wells in the alluvial aquifer near Fremont. In their model they used a saturated hydraulic conductivity of 107 m/d (350 ft/day) to represent the alluvial aquifer $K$.

Chatman & Associates Inc. (2004) performed well field vulnerability analysis in the Fremont Well Field of Nebraska. They developed a numerical groundwater flow model to evaluate the potential contaminant threats to the Well Field. The calibrated hydraulic conductivity value was between 110 to 118 m/d (361 to 388 ft/day) for Wells
55, 56, and 58 and between 224 to 232 m/d (735 to 762 ft/day) for Wells 51, 52, and 54. All these hydraulic conductivity values were for the Platte River valley aquifer sediments.

Woodward-Clyde Consultants (1996) performed pumping tests to estimate the hydraulic properties of the Todd Valley aquifer and the Platte River Valley aquifer at the Operable Unit No. 2, former Nebraska Ordnance Plant in Mead, Nebraska. The Todd Valley aquifer consists of the fine sand and sand/gravel units. The Platte River Valley aquifer includes sandy alluvium, while the Omadi Sandstone aquifer is comprised of the Omadi Sandstone of the Omadi Formation which belongs to the Cretaceous system. They found that the hydraulic conductivity was about 30 m/d (99 ft/day) and the specific yield is about 0.1 for the Todd Valley. For the Platte River Valley, they divided it into two zones: the Platte River alluvium and the underlying Omadi Sandstone. Their results showed that the hydraulic conductivity was about 86 m/d (282 ft/day) and the specific storage was about 0.0003 for the Platte River Valley alluvium, and the hydraulic conductivity was about 15 m/d (50 ft/day) and the specific storage was about 0.00006 for the Omadi Sandstone aquifer.

URS Group Inc. (2007) refined the conceptual model to estimate the horizontal and vertical extent of capture zones in the alluvial aquifer for individual extraction wells at the Operable Unit No. 2, former Nebraska Ordnance Plant in Mead, Nebraska. Their calibrated $K$ values varied from 6 to 152 m/d (20 to 500 ft/day) for the entire model domain. In the Platte River valley, $K$ ranged from 73 to 152 m/d (240 to 500 ft/day); in the Todd valley, $K$ ranges from 24 to 91 m/d (80 to 300 ft/day); in the Wahoo valley, $K$ ranged from 37 to 46 m/d (120 to 150 ft/day); and in the upland aquifers, $K$ was about 21 m/d (70 ft/day). In addition, the specific yield of the Platte River, Wahoo, and Todd
Valleys ranged from 0.2 to 0.25.

Chen and Chen (2003) summarized $K$ values of the alluvial aquifer in the Platte River valley which was located westerly of the study area. The horizontal $K$ was generally about 100 m/day, but the vertical $K$ was not as accurate as horizontal $K$. Cheng and Chen (2007) used a groundwater flow model to analyze a pumping-recovery test in Chapman near the Platte River, and their results suggested that horizontal $K$ of the alluvial aquifer is about 110 m/day and the anisotropic ratio is about 16. They also noted that the horizontal $K$ in the aquifer is the most certain and reliable parameter.

The estimated hydraulic conductivities for different geologic units from the above literature are summarized in Table 4.1, and these values could be helpful in the groundwater flow model calibration.

<table>
<thead>
<tr>
<th>Hydraulic Conductivity (m/day)</th>
<th>Quaternary Age Deposits</th>
<th>Platt River Valley</th>
<th>Todd Valley</th>
<th>Ogallala Group</th>
<th>Omadi Sandstone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pecknaphaugh and Dugan (1983)</td>
<td>0.3 to 91</td>
<td>0.3 to 24</td>
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<tr>
<td>Layne-Western (1981)</td>
<td></td>
<td></td>
<td></td>
<td>215</td>
<td></td>
</tr>
<tr>
<td>Layne Geosciences (2002)</td>
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<td></td>
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<tr>
<td>Woodward-Clyde Consultants (1996)</td>
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<td></td>
<td></td>
<td>86</td>
<td>30</td>
</tr>
<tr>
<td>URS Group Inc. (2007)</td>
<td>21</td>
<td>73 to 152</td>
<td>24 to 91</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cheng and Chen (2007)</td>
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<td></td>
<td></td>
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</table>
4.3. Methods

4.3.1 Using Constant-Head Boundary to Simulate Stream-Aquifer Interactions

As discussed in chapter 2, the shallow streambed in the gaining reaches of the braided and meandering rivers is mainly permeable, thus the general river or stream package in MODFLOW has difficulty to incorporate the absence of the clogging layer, especially to define the streambed hydraulic conductivity and streambed thickness. However, constant-head boundary approach only needs the input for stream level as described by the constant-head. The hydrostratigraphy of the streambed sediments below the constant-head boundary can be obtained from the well logs or electrical conductivity logs, and these streambed sediments can be simulated as aquifer materials. This approach can avoid the arbitrary selection of streambed hydraulic conductivity and streambed thickness. In this section, a hypothetical example of a stream-aquifer system was used to verify the applicability of using a constant-head boundary to simulate the stream-aquifer interactions instead of using the river or stream package of MODFLOW. In the hypothetical stream-aquifer system, a 30-m thick unconfined aquifer was assumed with an isotropic hydraulic conductivity of 100 m/d and a specific yield of 0.2. A 50-m wide stream partially penetrated the aquifer with a semi-previous layer on the channel surface similar to the hydrologic conditions discussed by Hunt (1999), which is shown in Figure 4.13. A pumping well was located 500 m from the stream with a pumping rate of 5000 m$^3$/d. A pumping period of 45-day was considered in this study. The water in the stream was one meter deep and the low-permeability layer was 0.5 m in thickness with a vertical hydraulic conductivity of 0.1 m/d. The stream stage was at the same elevation as the initial water table in the aquifer.
A three-dimensional groundwater flow model using Visual MODFLOW was constructed to calculate streamflow depletion using the Zone Budget package. The dimensions of the model were 10,000 m in length, 10,000 m in width, and 30 m in depth. The model was discretized into 205 columns and 206 rows in the x, y dimensions, and into 4 layers in the z dimension (Figure 4.14).
The stream traverses from west to east across the model. In the first layer, the stream was represented by a constant-head boundary package, which shows that the river stage is 30 m in elevation and 1 m in depth. In the second layer, the grids below the streams were designated by a low-conductivity value of 0.1 m/d, which represented the semi-previous layer in the streambed. The exchange of groundwater flow and streamflow beneath the stream was mostly vertical and the horizontal flow is insignificant. Therefore, in the first layer around the constant-head boundary, a Wall package available in Visual MODFLOW was used to prevent the lateral flow to and from the constant-head boundary grids. A very small hydraulic conductivity value \(10^{-9}\) m/d was assumed in the Wall package. Additionally, another similar groundwater flow model was developed to simulate the hypothetical stream-aquifer system, and the river package was used in this model to represent the stream. The streambed hydraulic conductivity was 0.1 m/d and streambed thickness was 1 m. The Hunt (1999) analytical solution was used as the reference to evaluate the accuracy of the two numerical modeling solutions.

4.3.2 EC logs and Streambed \(K_v\) measurements

Electrical conductivity (EC) logs are valuable in assessing the stratification of channel sediments since sand and gravel have significantly different EC values compared to silt and clay. Schulmeister et al. (2003) noted that there is an agreement between peaks in the EC profiles and increases in the clay content of the sampled layers, and they also pointed out that higher EC value generally reflect fine-grained material. Consequently, sand and gravel have a lower EC value compared to silt and clay, because sand and gravel have a larger value of resistivity than silt and clay, and an electrical log is the
inverse of a resistivity log (Schulmeister et al., 2003; Chen et al., 2008).

In the study area, the EC logs generated by direct-push techniques using Geoprobe® Systems SC400® soil conductivity probe were collected at 4 test sites from North Bend to Woods Landing in the Platte River in the summer of 2008 (Figure 4.1). The four sites were sites P5, P6, P7, and P8, which were consistent with those sites in chapter 2. At each test site, EC logs were recorded every 1.5 cm as the probe was pushed through channel sediments. In addition to the EC logs, sediments cores were collected using the Geoprobe® Systems Macro-Core® soil sampler into polycarbonate tubes every 1.5 m in length and 4.2 cm in diameter. The polycarbonate tube was transparent; therefore the laminations and components of the sediment cores can be visually identified.

In-situ permeameter tests were performed at five sites between North Bend and Woods Landing in the Platte River (Figure 4.1) to determine the vertical hydraulic conductivities of channel sediments. The five sites were sites L, M, N, O, and P, which were consistent with those sites in chapter 3. At each site, 48 to 200 permeameter tests were conducted. The schematic diagram of in-situ permeameter test is shown in Figure 2.4 of chapter 2. The number of permeameter tests, the grid spacing between test points, and the average $L_v$ are summarized for each site in the Platte River (Table 4.2).
Table 4.2 Average streambed $K_v$ values, average $L_v$, average water depth, and grid spacing at the five test sites (sites L to P) in the Platte River of Nebraska.

<table>
<thead>
<tr>
<th>Test Site</th>
<th>Test Date</th>
<th>Number of Tests</th>
<th>Grid Spacing</th>
<th>Average $L_v$ (cm)</th>
<th>Average $K_v$ (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L</td>
<td>July 3, 2008</td>
<td>48</td>
<td>8, 1.5 m</td>
<td>50.8</td>
<td>33.3</td>
</tr>
<tr>
<td>M</td>
<td>July 11, 2008</td>
<td>200</td>
<td>4, 1.5 m</td>
<td>50.8</td>
<td>17.7</td>
</tr>
<tr>
<td>N</td>
<td>July 10, 2008</td>
<td>49</td>
<td>7, 1.5 m</td>
<td>50.8</td>
<td>29.8</td>
</tr>
<tr>
<td>O</td>
<td>July 21, 2008</td>
<td>49</td>
<td>7, 1.5 m</td>
<td>50.8</td>
<td>37.8</td>
</tr>
<tr>
<td>P</td>
<td>July 9, 2008</td>
<td>64</td>
<td>8, 1.5 m</td>
<td>50.8</td>
<td>45.2</td>
</tr>
</tbody>
</table>

4.3.3 Modeling of the Interactions between Groundwater and the Platte River

Visual MODFLOW was used to simulate the stream-aquifer interactions in the study area. The groundwater flow model was divided into 5 layers according to different geologic and hydrogeologic conditions on the basis of the test-hole log reports (http://snr.unl.edu/data/geologysoils/index-geologysoils.asp) and irrigation well log reports (http://dnrdata.dnr.ne.gov/wellssql). The spatial variations in layer thickness were determined using the kriging method based on the elevations at the test-holes and irrigation wells. The first layer contained Quaternary Age silt and clay sediments, which varied in thickness from 1 m in the Platte River valley to more than 35 m in the upland area. The second layer was composed of alluvial sand and gravel deposits in the Platte River valley and Quaternary Age sand and gravel sediments in the upland area, with a few till sediments embedded. The third layer consisted mainly of silt and clay sediments, which was regarded as a confining layer, while the fourth layer was mainly composed of Ogallala sand and gravel sediments. The fifth layer contained mostly silt and clay with a few limestone and sandstone sediments. The sediments below the fifth layer were
considered to be hydrologically disconnected from the Platte River; therefore, they were regarded as a confining unit. Within each layer, the model was subdivided into a horizontal grid of 178 columns and 118 rows. The horizontal grid size varied from 200 by 200 m near the Platte River valley area to 400 by 400 m in the upland area. Figure 4.15 shows the model grids for the three-dimensional groundwater flow model.

![Figure 4.15 Model grids for the three-dimensional groundwater flow model in the Lower Platte River Basin of Nebraska.](image)

The Platte River in this model was simulated using a constant head boundary instead of using the River package in Visual MODFLOW. Use of a constant head boundary still allows the user to prescribe varied river stages at different times. The constant head grids were imposed to the first layer where the Platte River covers. The thickness of the constant head grids was set to be 1.0 m, which can reflect the depth of
water in the Platte River during the irrigation season (Chen et al., 2008). Linear interpolation was used to determine the stream levels in-between and out of the two USGS gauge stations at the constant-head grids. In this study, the exchange of groundwater and streamflow was assumed to occur through streambed so the exchange is mainly vertical. Therefore, in the first layer around the constant-head boundary, a Wall package available in Visual MODFLOW was adopted with a very small hydraulic conductivity value ($10^{-9}$ m/d) to prevent the lateral flow into and out of the constant-head boundary grids.

Eight different hydraulic conductivity ($K$) zones were applied in the model. The sand and gravel sediments in the upland of the Quaternary Age, the alluvial deposits in the Platte River valley of the Quaternary Age, and Ogallala Group had different $K$ values. A $K$ zone was designated to the channel sediments below the Platte River. Moreover, silt and clay, till, limestone and sandstone were also assigned with different $K$ zones. The specific storage and specific yield were grouped into 2 zones. During model calibration, the $K$ values, specific storage, and specific yield values were refined. The sediments below the fifth layer were treated as no-flow boundary. For the west and east sides of the model domain, general head dependent boundaries were applied, which simulated the subsurface flow into and out of the model domain. A general head dependent boundary was also applied to part of the south side of the model domain since the Platte River flows southeasterly in the study area. The model was simulated in transient conditions, thus the initial hydraulic head for each grid was needed. The water levels from the observation wells on January 1994 were obtained, and the kriging interpolation method was used to achieve the initial hydraulic head distribution for the entire study area.
The MODFLOW groundwater recharge package was applied to the active cells in the top layer. Usually one part of groundwater recharge comes from precipitation. Irrigation return flow makes up the remaining part of groundwater recharge. In the irrigated agricultural lands, the practice of irrigation results in higher soil moisture than that in the non-irrigated lands and therefore groundwater return flow rate is higher in the irrigated agricultural lands (Luckey and Cannia, 2006). As a result, the groundwater recharge rate in the irrigated lands was usually greater than that in the non-irrigated lands. Furthermore, groundwater recharge rates depends on the land topography (Luckey and Cannia, 2006), such that the rates are higher over level areas and less over more steep slopes. In the study area, the initial groundwater recharge rates were applied in both the Platte River riparian valley and upland area, and they were refined after model calibration. In terms of groundwater evapotranspiration (ET), it is often specified to occur in the riparian vegetation areas. Chen and Shu (2006) estimated that the groundwater ET rate ranges from 499 to 640 mm/yr for the riparian vegetation and that the cut-off depth for ET is 4.6 m in the central Platte River Valley, Nebraska. In our study, the groundwater ET losses exist mostly in the Platte River riparian valley area. Here, the groundwater ET rate was assumed to be 700 mm/yr, and no groundwater was lost through ET when the extinct depth is greater than 4.5 m.

4.4 Results

4.4.1 Applicability of Constant-Head Boundary to Simulate Stream-Aquifer Interactions

For the hypothetical example, the resulting streamflow depletion due to
groundwater pumping obtained from the three methods (the constant-head boundary package, the River package, and the Hunt analytical solution) show good agreement (Figure 4.16), which indicates that using the constant-head boundary in simulating stream-aquifer interactions is reliable for the modeled conditions with a low-permeability layer of sediments at the channel surface. In this hypothetical example, the stream stage is assumed to be constant during the simulation period; however, use of a constant head boundary still allows the user to prescribe varied river stages at different times.

Figure 4.16 Streamflow depletion rates obtained from the three methods (the constant-head boundary and River package in MODFLOW, and the Hunt [1999] solution) for the hypothetical stream-aquifer system with a K of 100 m/d for the aquifer with a low-permeability layer of sediments (K of 0.1 m/d) at the channel surface, a distance of 500 m between the stream and pumping well, and a stream width of 50 m.

Using the river or stream package needs the input of streambed conductance, which is a lumped parameter of stream channel width and length, streambed hydraulic conductivity and streambed thickness. A lot of previous modeling studies assumed that
there is a clogging layer at the channel surface (Sophocleous et al., 1995; Hunt, 1999; Osman and Bruen, 2002; Akylas and Koussis, 2007; Rushton, 2007; Hu et al., 2007), and they used the hydraulic conductivity and thickness of the clogging layer to calculate streambed conductance. However, as discussed before, the assumption of a clogging layer at the channel surface is not always correct, especially for a braided river or the gaining reaches for a meandering river. If there is no clogging layer at the channel surface, those modeling studies usually selected the streambed hydraulic conductivity and streambed thickness arbitrarily when they used the River or Stream package.

The constant-head boundary approach only needs the input for stream level as described by the constant-head, and the hydrostratigraphy of the streambed sediments can be obtained from the well logs or electrical conductivity logs, and these streambed sediments can be simulated as aquifer materials. Hence, this approach can avoid the arbitrary selection of streambed hydraulic conductivity and streambed thickness, and it provides more flexibility to represent the real-world streambed sedimentary structure in the numerical simulations.

4.4.2 Electrical Conductivity logs

The EC logs at the four test sites (P5, P6, P7, and P8) were consistent with those sites in chapter 2. The figures of EC logs for these sites are shown in Figure 2.6 of chapter 2. As discussed in chapter 2, the coarse- to fine-grained sand sediments are the main component of the streambed sediments at these sites in the Platte River.
4.4.3 \( K_v \) values of Channel Sediments

The average streambed \( K_v \), \( L_v \), and water depth values at each of the five test sites are shown in Table 4.2. At each site, the permeameter tests were regularly spaced with the grid spacing of 1.5 m along and cross the river flow direction. In this study, the average \( L_v \) was the same for all the test sites, which was 50.8 cm. The range of average \( K_v \) at these sites was from 17.7 to 45.2 m/d, and the average \( K_v \) value of the five sites was 32.4 m/d, which was used in the groundwater model to simulate the sediments below the constant-head boundary representing the Platte River.

4.4.4 Model Calibration

The model was calibrated using the groundwater levels in a total of 40 observation wells for the period between January 1994 and December 2004. Most of the observation wells had two records of water levels in each year. Of the 40 observation wells, 21 were located in the Platte River valley, 12 in the Todd valley, and 7 in the upland area. In terms of the screens of the 40 observation wells, 26 were screened in the Ogallala Group sand and gravel sediments, 11 were screened in the alluvial and quaternary sand and gravel sediments, and another 3 were screened in the silt and clay sediments. Groundwater levels in the observation wells responded to seasonal irrigation withdrawals of groundwater; however, they did not have an apparent overall declining trend. During model calibration, 90 continuous pumping days were used to represent the pumping period of the groundwater irrigation wells. The time interval of the calibration was divided into 132 stress periods, and each stress period consisted of 10 time steps.

Trial-and-error calibration procedure was used to refine the initially assigned
hydraulic parameters, including the hydraulic conductivity, specific yield, specific storage, groundwater recharge, and groundwater ET values. Four different groundwater recharge zones were assigned, and the monthly recharge rate was calibrated to be from 0.001 to 512 mm/yr in different zones over the calibration periods, which accounted for 0.1% to 22% of the precipitation in the study area. Groundwater recharge values were higher near the Platte River valley, which is in accordance with the assumptions that the land topography was level and more lands were irrigated in the Platte River valley area. Final calibrated $K_h$ (horizontal hydraulic conductivity) values for the Quaternary Age and Quaternary alluvial sand and gravel sediments ranged from 42 to 98 m/d, whereas the $K_h$ value of the Ogallala Group sand and gravel sediments was about 15 m/d. Previous studies have shown that $K$ ranges from 30 to 152 m/d in the alluvial aquifer and from 0.3 to 24 m/d in the Ogallala Group aquifer based on pumping test and grain-size analysis (Peckenpaugh and Dugan, 1983; Chen and Chen, 2003; Cheng and Chen, 2007). The calibrated $K$ values could be smaller than these values due to the representation of equivalent hydraulic conductivity during the development of the conceptual model, but the calibrated $K$ values were in the range of those determined by previous studies. In addition, laboratory permeameter tests in chapter 2 provided the $K_v$ values of streambed sediments up to 20 m below the channel surface, which corresponds to the alluvial sand sediments in the model. The calibrated $K_v$ value of the alluvial sand sediments is 6.7 m/d, which is in the range of the average $K_v$ value for the streambed sediments (2.4 to 9.2 m/d, Table 2.4 in chapter 2). For all aquifers, the specific storage ranges from $1.0 \times 10^{-5}$ to $3.0 \times 10^{-5}$ m$^{-1}$, and the specific yield varies from 0.06 to 0.2. The calibrated hydraulic conductivity values for different zones of sediments are shown in Table 4.3. Of the 40
observation wells in the model, 26 wells (65%) have absolute residuals of the calculated and observed water levels between 0 and 1.0 m, 10 wells (25%) have absolute residuals between 1.0 and 2.0 m, and only 4 wells (10%) have absolute residuals higher than 2.0 m.

The calculated and observed water levels from four representative observation wells (Figure 4.17) are shown in Figure 4.18. The observed hydraulic heads were more damped than the calculated heads for some observation wells (e.g. OB-2 in Figure 4.18), which is due to that the pumping period in the model was assumed to be a continuous 3-month for every year from June to August. However, in reality, groundwater irrigation might be on and off during the summer time depending on the precipitation, which can affect the groundwater level significantly. For the regional mode, it is difficult to define a specific pumping period for each irrigation well because of its uncertainty. Nevertheless, the overall calibration indicates a good trend of variations between the simulated and observed groundwater levels, and most of the observation wells have the absolute residuals of groundwater levels less than 2 m. Consequently, the calibration in the study indicated good agreements between the simulated and observed groundwater levels.

Table 4.3 Calibrated horizontal and vertical hydraulic conductivity values for different sediments in the model area.

<table>
<thead>
<tr>
<th>Sediments</th>
<th>$K_h$ (m/d)</th>
<th>$K_v$ (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Silt/Clay</td>
<td>0.08</td>
<td>0.0035</td>
</tr>
<tr>
<td>Quaternary Sand/Gravel</td>
<td>70</td>
<td>4.5</td>
</tr>
<tr>
<td>Alluvial Sand/Gravel</td>
<td>98</td>
<td>6.7</td>
</tr>
<tr>
<td>Sand/Gravel in Todd Valley area</td>
<td>42</td>
<td>3.5</td>
</tr>
<tr>
<td>Till</td>
<td>3.6</td>
<td>0.25</td>
</tr>
<tr>
<td>Channel Sediments below the Platte River</td>
<td>120</td>
<td>32.4</td>
</tr>
<tr>
<td>Ogallala Group Sand/Gravel</td>
<td>15</td>
<td>1.2</td>
</tr>
<tr>
<td>Limestone/Sandstone</td>
<td>0.7</td>
<td>0.06</td>
</tr>
</tbody>
</table>
Figure 4.17 Location of four representative observation wells.
4.4.5 Analysis of Stream Depletion Due to Groundwater Pumping

We defined the reach between the two USGS stations, North Bend and Leshara, in the Platte River as zone 2 (noted as REACH), the reach upstream of the station North Bend as zone 3, the reach downstream of the station Leshara as zone 4, and the rest of the study area as zone 1. The Zone Budget module in Visual MODFLOW was used to retrieve the flux exchanges between different zones over different time periods. Figure 4.19a shows the stream-aquifer flux across the REACH, where the positive flux indicates the groundwater flow to the REACH (regarded as baseflow) and the negative flux indicates the streamflow leakage from the REACH to the aquifer. Both fluxes fluctuate monthly with different pumping schedules and groundwater recharge rates (Figure 4.19a). With groundwater pumping, baseflow rate ranged from 158 to 72,878 m$^3$/d whereas the streamflow leakage rate ranged from 0.06 to 85,081 m$^3$/d. The cumulative baseflow was $5.9 \times 10^7$ m$^3$, and the cumulative stream leakage from the REACH to the aquifer was 1.2
×10^8 m^3 from January 1994 to December 2004 (Figure 4.19b).

Figure 4.19 The stream-aquifer flux across the REACH with groundwater pumping. The positive flux indicates the groundwater flow to the REACH (baseflow), whereas the negative flux indicates the streamflow leakage from the REACH to the aquifer; and (b) the cumulative baseflow and stream leakage across the REACH from January 1994 to December 2004.
In order to analyze the impacts of groundwater pumping on stream depletion, the groundwater pumping wells are removed from the model, which could characterize the stream-aquifer flux under the baseline condition without any groundwater pumping effects. It is apparent that groundwater contributed more to the aquifer when there were no groundwater withdrawals, and the stream leakage to the aquifer was correspondingly lower (Figure 4.20a). Without any groundwater pumping, the cumulative baseflow was 1.4 ×10^8 m^3, and the cumulative stream leakage from the REACH to the aquifer was 5.4 ×10^7 m^3 from January 1994 to December 2004 (Figure 4.20b).

From January 1994 to December 2004, the difference of cumulative baseflow based on the two conditions (with and without groundwater pumping) was 7.8 ×10^7 m^3, whereas the difference of cumulative stream leakage based on the two conditions was 6.3 ×10^7 m^3. The cumulative stream depletion is the combined results of these two components, which resulted in a total of 1.4 ×10^8 m^3 from January 1994 to December 2004. Meanwhile, the cumulative groundwater pumping at this time period was 8.5 ×10^8 m^3. Accordingly, the stream depletion rate relative to the groundwater pumping for the REACH was about 16%.
Figure 4.20 The stream-aquifer flux across the REACH without groundwater pumping. The positive flux indicates the groundwater flow to the REACH (baseflow), whereas the negative flux indicates the streamflow leakage from the REACH to the aquifer; and (b) the cumulative baseflow and stream leakage across the REACH from January 1994 to December 2004.
4.4.6 Sensitivity Analysis of Streambed $K_v$ in Evaluating Stream Depletion

The range of average $K_v$ at the four sites was from 17.7 to 45.2 m/d, and the average $K_v$ value of the five sites was 32.4 m/d, which was used in the groundwater model. In this section, the minimum and maximum $K_v$ values were assigned to the $K$ zone of the channel sediments and to determine their effects on stream depletion. The effects of different $K_v$ values of the channel sediments on the calibrated hydraulic heads are minimal; however, they can affect the flux exchange between the Platte River and its adjacent aquifers.

When the $K_v$ of the channel sediments was 17.7 m/d, the cumulative stream depletion was the combined results of the two components (stream infiltration and baseflow reduction), which resulted in a total of $1.3 \times 10^8$ m$^3$ for the REACH from January 1994 to December 2004. When the $K_v$ of the channel sediments is 45.2 m/d, the cumulative stream depletion was $1.5 \times 10^8$ m$^3$ for the REACH from January 1994 to December 2004. Accordingly, the stream depletion rate relative to the groundwater pumping was about 15% and 17% for a $K_v$ of 17.7 and 45.2 m/d, respectively. The difference of stream depletion rates from the three $K_v$ values is minimal, which is due to that all three $K_v$ values are still in the same magnitude of order, and the stream depletion rate is more dependent on the low-permeability sediments in deep streambed since most of the groundwater irrigation wells were located in the Ogallala Group.

If a clogging layer was assumed in the model as many modeling studies did, the stream depletion rate can be decreased significantly. For example, if we chose a $K_v$ value of 0.1 m/d and 0.01 m/d for the channel sediments, the corresponding stream depletion rate relative to the groundwater pumping was about 7% and 4%, respectively.
Consequently, careful selection of streambed $K_v$ values is very important in the evaluation of pumping-induced stream depletion, and an arbitrary assumption of a clogging layer at the channel surface underestimates the real-world stream-aquifer interaction if the streambed is mainly permeable.

Because the streambed $K_v$ values are measured by in-situ permeameter tests in point locations, the average values or interpolated values from different locations are often used for a regional groundwater flow model. In this study, the average $K_v$ value was used, which can represent the characteristic of channel sediments in the reach of the Platte River within the model domain.

4.4.7 Estimation of Streambed Leakance using Groundwater Modeling

River package was used instead of the constant head boundary to represent the Platte River, which can be beneficial to determine the streambed leakance values. The river package was imposed to the same grids where the constant head boundary covered. Here, four different streambed leakance values ($\lambda=0.001 \text{ d}^{-1}$, 0.01 $\text{ d}^{-1}$, 0.1 $\text{ d}^{-1}$, and 0.2 $\text{ d}^{-1}$) were testified. The effects of different streambed leakance values on the calibrated hydraulic heads are minimal; however, they can affect the flux exchange between the Platte River and its adjacent aquifers significantly. Figure 4.21 shows the stream-aquifer flux across the REACH for different values of $\lambda$. When $\lambda$ was 0.001 $\text{ d}^{-1}$, the baseflow rate varied from 0.01 to 23552 $\text{ m}^3/\text{d}$, and the cumulative baseflow in the REACH from January 1994 to December 2004 was $1.9 \times 10^7 \text{ m}^3$, which was much smaller and almost one third of that in constant head simulation. When $\lambda$ was 0.01 $\text{ d}^{-1}$, the baseflow rate increases apparently, which was in a range of 142 to 60775 $\text{ m}^3/\text{d}$. The cumulative
baseflow in the REACH from January 1994 to December 2004 was $4.8 \times 10^7$ m$^3$.

Furthermore, the streambed leakage values of 0.1 and 0.2 d$^{-1}$ were used in the simulations, which resulted in a much closer cumulative baseflow in the REACH to that in the constant head simulation. The cumulative baseflow was $5.8 \times 10^7$ m$^3$ and $6.0 \times 10^7$ m$^3$ for the $\lambda$ of 0.1 and 0.2 d$^{-1}$, respectively (Table 4.4). Consequently, the appropriate streambed leakage for the reach between North Bend and Leshara in the Platte River may range from 0.1 to 0.2 d$^{-1}$.

Table 4.4 The cumulative baseflow in the reach between the two USGS stations North Bend and Leshara using different simulation approaches from January 1994 to December 2004 ($\lambda$ is streambed leakance).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Cumulative Baseflow (m$^3$)</th>
<th>Difference with Constant Head Approach</th>
</tr>
</thead>
<tbody>
<tr>
<td>Constant Head Approach</td>
<td>$5.9 \times 10^7$</td>
<td>0</td>
</tr>
<tr>
<td>$\lambda=0.001$ d$^{-1}$</td>
<td>$1.9 \times 10^7$</td>
<td>211%</td>
</tr>
<tr>
<td>$\lambda=0.01$ d$^{-1}$</td>
<td>$4.8 \times 10^7$</td>
<td>23%</td>
</tr>
<tr>
<td>$\lambda=0.1$ d$^{-1}$</td>
<td>$5.8 \times 10^7$</td>
<td>1.7%</td>
</tr>
<tr>
<td>$\lambda=0.2$ d$^{-1}$</td>
<td>$6.0 \times 10^7$</td>
<td>-1.7%</td>
</tr>
</tbody>
</table>

Chen and Chen (2003) estimated the streambed leakage of the Platte River from a stream-aquifer test using the Hunt (1999) solution in Killgore Island, Nebraska. They noted that the streambed leakage is from 0.48 to 2.4 d$^{-1}$ using the observations in individual wells. Peterson et al. (2008) used a streambed conductance of 22.5 to 31.5 ft/day in a unit length for the Loup River of Nebraska. Considering the width of the Loup River is about 200 to 300 ft, thus the streambed leakage value for the Loup River is about 0.1 d$^{-1}$. Although the Platte River has a sandy stream bottom, the streambed leakage value is still low. The low-permeability sediments still occur at deep streambed
based on the four EC logs (Figure 2.6 of chapter 2), e.g., 4 to 6 meters below the channel surface. However, the groundwater irrigation wells were mainly constructed in the Ogallala Group. As discussed in chapter 2, if the well depth is below the low-permeability layer, pumping-induced stream depletion might be still low because the low-permeability layer still acts as a barrier to prevent streamflow infiltration to the aquifer system. Therefore, it is not surprising that the streambed leakance value is not high for the Platte River. Above all, the estimated streambed leakance value based on groundwater flow modeling in this study is compatible to those in the previous studies, which confirms the validity of the streambed leakance of the Platte River, and it can be used in assessing the stream-aquifer interactions when an analytical solution is adopted.

4.5. Summary and Conclusions

Streambed leakance is an important parameter of streambed conductance. However, streambed conductance is usually arbitrarily chosen based on the assumption that a low-permeability clogging layer occurs in the channel surface. This study and previous studies (Chen et al., 2008; Chen, 2010; Cheng et al., 2011) suggested that the Platte River is generally not clogged at the channel surface, thus its streambed leakance value needs to be characterized accurately when an analytical solution is adopted in the evaluation of the interactions between the Platte River and its adjacent aquifer.
Figure 4.21 The stream-aquifer flux across the REACH when (a) \( \lambda = 0.001 \text{ d}^{-1} \); (b) \( \lambda = 0.01 \text{ d}^{-1} \); (c) \( \lambda = 0.01 \text{ d}^{-1} \); and (d) \( \lambda = 0.2 \text{ d}^{-1} \). The positive flux means the groundwater flow to the REACH (baseflow), while the negative flux means the streamflow leakage from the REACH to the aquifer.
In this study, a regional groundwater model was developed to evaluate the impacts of groundwater irrigation on the streamflow in the Lower Platte River Basin, and this model was used to make an accurate estimation of streambed leakance of the Platte River using numerical and field techniques. First, the stratification of channel sediments in the lower reach of the Platte River of Nebraska was studied. The collected EC logs at four sites show that sand sediments occur at the shallow streambed, and silt and clay were found in deep layers of channel sediments. The absence of low-permeability layer at the channel surface in the Platte River may indicate a high hydrologic connectivity between the Platte River and its adjacent aquifers. Second, the \( K_v \) values of channel sediments at five sites in the Platte River were determined by permeameter tests. The average \( K_v \) value of the five sites was 32.4 m/d for the top 50.8-cm sediment, and this value was used in the groundwater flow model.

Following that, a three-dimensional groundwater flow model was developed using Visual MODFLOW to simulate the interactions between the Platte River and its adjacent aquifers. The model was calibrated using trial-and-error method according to the changes of groundwater levels in 40 observation wells for the period from January 1994 to December 2004. About 65% of the observation wells had the absolute residuals of the calculated and observed water levels less than 1 m. Final calibrated \( K_h \) values for the Quaternary Age and alluvial sand and gravel sediments range from 42 to 98 m/d, while the \( K_h \) value of the Ogallala Group sand and gravel sediments is about 15 m/d, and these values were in the range of those determined by previous studies. The calibrated hydraulic parameters were then applied to calculate the exchange between the Platte River and its adjacent aquifers using Zone Budget module in Visual MODFLOW. The
results indicates that the baseflow (groundwater contributes to the Platte River) rate ranges from 158 to 72878 m$^3$/d whereas the streamflow leakage (flow from the Platte River to groundwater) rate ranged from 0.06 to 85081 m$^3$/d, and the cumulative baseflow in the REACH from January 1994 to December 2004 was $5.9 \times 10^7$ m$^3$. The stream depletion rate relative to the groundwater pumping for the REACH was about 16% for the time period from January 1994 to December 2004. Finally, river package was used instead of the constant-head boundary to represent the Platte River in the model. The appropriate streambed leakance for the reach between North Bend and Leshara in the Platte River was between 0.1 and 0.2 d$^{-1}$. Although the Platte River has a sandy stream bottom, the streambed leakance value is still low. It is because that the low-permeability sediments still occur at deep streambed, which still can act as a barrier to prevent streamflow infiltration to the aquifer system. The estimated streambed leakance value in this study was compatible to those in the previous studies, and it can be used in assessing the stream-aquifer interactions when an analytical solution is adopted.
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