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QUANTIFYING AND MODELING STREAM-AQUIFER INTERACTIONS IN THE ELKHORN RIVER BASIN, NEBRASKA

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QUANTIFYING AND MODELING STREAM-AQUIFER INTERACTIONS IN
THE ELKHORN RIVER BASIN, NEBRASKA

by

Can Liu

A THESIS
Presented to the Faculty of
The Graduate College at the University of Nebraska
In Partial Fulfillment of Requirements
For the Degree of Master of Science

Major: Natural Resources Sciences

Under the Supervision of Professor Xun-Hong Chen

Lincoln, Nebraska
March, 2014
This study combined statistical analyses, field investigations and numerical groundwater flow modeling to quantify the connectivity between the Elkhorn River and its adjacent aquifers in Nebraska. The Mann-Kendall trend tests were conducted to detect increasing or decreasing tendencies on the time series data of streamflow, which were collected from eighteen gauging stations in the Elkhorn River and its tributaries. Decreasing trends were not found in the annual streamflow data.

Field investigation of streambed hydraulic properties was performed in the Elkhorn River near Winslow and Norfolk. Vertical hydraulic conductivities ($K_v$) of the shallow streambed sediments were obtained by in-situ permeameter tests. Geoprobe® techniques were employed for electrical conductivity (EC) logging and sediments coring. The $K_v$ profiles of the sediment cores were determined using in-lab permeameter tests. The $K_v$ values of streambed sediments were generally higher at the Norfolk site than at the Winslow site.

Two numerical groundwater flow models were developed for the Upper Elkhorn River Basin near Atkinson and the Lower Elkhorn River Basin near Winslow using
MODFLOW-2005. The aquifer system was divided into five hydrostratigraphic units and the external hydrologic processes were incorporated in the models. The two models were calibrated under transient conditions based on the observation records from the U.S. Geological Survey groundwater level database. Hypothetical wells having different locations, pumping schedules and screen depths were placed in the model for streamflow depletion analyses. The results suggested that a well which was located closer to the river with shallow screen was more susceptible in causing stream depletion within the Atkinson model. The thick layer of silt/clay/till deposits in the Winslow area may reduce the hydrological connection between the river and aquifers and result in low and steady stream depletion rate. The stream depletion caused by the spatially allocated wells with the same distance to the river varied, which could result from the anisotropy and the heterogeneity of the aquifer system. This study would aid the administrative authorities to make informed decisions on integrated water resources management.
ACKNOWLEDGEMENT

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Finally, thanks to my mom, Yingmian Li. Thank you for giving birth to me, raising me and supporting me whatever I do. Thank my fiancé, Yao Li. Your endless love has been accompanying me every moment and everywhere.
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CHAPTER 1 INTRODUCTION

1.1 Background

Integrated water resources management is an important issue in the Elkhorn River Basin. The streamflow and the sub-surface aquifers are vital for the aquatic life conservation, hydropower production, agricultural irrigation, and municipal water use. Since 1930’s, the groundwater withdrawal for cropland irrigation became greater due to the expansion of irrigation acreages. The rapid growth of groundwater irrigation wells has altered the original water flow by intercepting groundwater flow which originally discharged into surface water bodies, or by increasing the rate of water movement from the surface water body into an aquifer. The significant drawdown of groundwater table in some areas during dry years has caused the local authorities concern about the sustainability of in-stream flow in the Elkhorn River. The recent drought in 2012 amplified this concern and the possible needs to integrate surface and sub-surface water resources management. To reach a balance of high agricultural productivity and long-term water use, comprehensive understanding of the hydrological connectivity between the streams and the aquifer system is needed.

1.2 Objectives

The objective of this study was to quantify the hydrologic connectivity among the stream, the shallow alluvial aquifer and the deep aquifer, and to simulate the potential streamflow depletion caused by agricultural pumping. Non-parametric tests were used to
examine whether there was a decreasing trend of the streamflow data over decades. Field study of streambed hydraulic properties provided a basic understanding of the dynamic stream-aquifer system at the Norfolk and Winslow sites. Two numerical groundwater flow models were developed and calibrated to simulate the local groundwater system in the Atkinson and Winslow areas. The Atkinson area was located in the upper Elkhorn River Basin and the Winslow area was located in the lower Elkhorn River Basin; the principal aquifer units differed in the two areas. The model outputs revealed the changes of groundwater storage before and after the widespread irrigation. Model simulations were used to evaluate the effect of hypothetical pumping wells on the streamflow depletion at different locations, with varied pumping schedules and well screen depths in the Atkinson and Winslow areas.

1.3 Previous work

A previous study by Newport (1957) provided a broad understanding of geology, hydrology, and chemistry in the Elkhorn River Basin above Pilger. He asserted that the groundwater reservoir was recharged by local precipitation, and perennial streamflow was formed by the subsurface movement of groundwater, eastwards and southeastwards. In 1997, the Conservation and Survey Division (CSD) began the design of monitoring well networks for the Upper and Lower Elkhorn Natural Resources Districts (NRDs). Groundwater levels were recorded by transducers for every eight hours at these wells. U.S. Geological Survey (Peterson, 2008) developed a numerical model using only one hydrostratigraphic unit to simulate the cumulative effects of groundwater irrigation on base flow in the Elkhorn and Loup River Basin. Simulations were constructed at the pre-
development (before 1940), historical development (1940-2005) and future hypothetical development (2006-2045/2055). Results revealed that the groundwater pumping-caused stream depletion rate in the future 50 years was greater than 10% when a hypothetical well was located within 7 or 8 mi from the stream (Peterson, 2008). In 2010, a phase-two study was extended to the model (Stanton et al., 2010) for new data updates, calibration improvement, and additional simulation analyses. Chen et al. (2009) proposed a new method for quantifying the vertical seepage flux across the stream-sediment interface by measuring the hydraulic gradient and hydraulic conductivity in the Elkhorn River. The magnitude and direction of the flux in the hyporheic zone varied greatly over a short distance (Chen et al., 2009). Chen and Lackey (2010) and Lackey and Chen (2010) conducted the streambed tests in the Upper and Lower Elkhorn River to determine the extent of streambed and aquifer connectivity. Wang (2012) characterized the hydrologic connectedness between the Elkhorn River and its surrounding aquifers at eight study sites through cross-correlation analyses, hydraulic gradient determination and streambed studies. He also developed a numerical groundwater flow model to calculate the stream depletion ratio at the Neligh and Hadar sites by inserting one hypothetical pumping well.

However, most of the previous groundwater flow models covered a spacious district area where some specific, detailed and spatially varied hydrogeology settings were not taken into account. A localized stream-aquifer study within a small district unit is rare but would be helpful for managing water resources and making decisive water administration policies.
CHAPTER 2 NONPARAMETRIC TREND TESTS ON STREAMFLOW DATA

2.1 Previous studies

Mann-Kendall tests (Mann, 1945; Kendall, 1975) have been widely used in detecting monotonic trend in random hydrologic and climatologic data (Hamed and Rao, 1998; Yue and Wang, 2004; Hirsch and Slack, 2010). It is a non-parametric approach which doesn’t require the assumption of normality and is resistant to the influence of extremes (Lins and Slack, 1999). With the help of the Mann-Kendall test, Lettenmaier et al. (1994) examined four monthly hydro-climatological variables for the continental United States for the period 1948-1988: averaged temperature, precipitation, streamflow and averaged range of daily temperature. Strong increasing trends in streamflow in the period of November-April was found. Lins and Slack (1999) evaluated the secular trend in streamflow over the conterminous United States with selected quantiles of discharge. Increasing trends were detected for a broad section of the United States and decreasing trends were only found in parts of the Pacific Northwest and the Southeast. Previous studies focused on detecting hydrological trends in the broad area and localized study was rare.

Wen and Chen (2006) found a temporal decreasing trend using Mann-Kendall tests in the flow rate of the Republican River in Nebraska. No significant trend was found in the Elkhorn River. According to Wen and Chen (2006), the intensive groundwater withdrawal might be a primary factor causing the reducing trend given the fact that no
temporal trend was found in precipitation data. Seven stream gauging stations in the main Elkhorn River were tested in Wen and Chen’s study. The latest data used in their analyses was from the year of 2003. Since the study of Wen and Chen (2006), more streamflow data has become available and streamflow trends should be updated. To examine the streamflow trends in the entire Elkhorn River Basin, eleven more gauging stations in the tributaries of the Elkhorn River were added in our analysis, including North Fork, South Fork, Holt Creek, Clear Water, Willow Creek, Union Creek, Pebble Creek, Maple Creek, and Logan Creek. The streamflow data for all the gauging stations were collected either from the historical surface water database of U.S. Geological Survey (USGS) or from the surface water database of the Department of Natural Resources in Nebraska (NDNR). Meanwhile, the dataset of annual streamflow rate was updated to 2012, except for Station 06796978 and Station 06798300 which no data was collected after the year of 1989 and 1991. Compared to the study of Wen and Chen (2006), our analysis has advantages in judging the trend of streamflow specifically in the Elkhorn River and its tributaries. The result would provide an initial understanding of the potential stream depletion.

2.2 Mann-Kendall test

A prerequisite for performing the Mann-Kendall test is that the serial data are independent from each other. According to Hamed and Rao (1998), potential correlations existing in the data increased the possibility of detecting trends when actually none existed, and vice versa. The streamflow data may have serial correlations which would impact the accuracy of identifying trends. To reduce the risk of biased assessment, a pre-whitening method was utilized in this study, as first proposed by von Storch and Navarra
(1995) to eliminate or mitigate the effect of serial correlation. The pre-whitened series $x^*$ was generated by the following equation:

$$x^* = x_t - r^2 x_{t-1}$$

(1)

where $x_t$ and $x_{t-1}$ were the data values at time $t$ and $t-1$; $r^2$ (Table 1) was the lag-1 serial correlation coefficient, which was calculated using the auto-correlation function in statistical package “R”. The Mann-Kendall tests were performed after the pre-whitening process.

The research hypothesis was that there was a trend in the long series of stream gauging data and the null hypothesis was that the trend didn’t exist. Firstly, all the observation values were ranked by time series. Then, the difference of each successive data was calculated and the Mann-Kendall statistic ($MK$) was determined by the sum of the signs.

$$MK = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} sgn(X_j - X_i)$$

(2)

where $n$ was the number of records; $X_i$ and $X_j$ were the successive observation values.

$$sgn(X_j - X_i) = \begin{cases} 
1, & \text{if } X_j - X_i > 0 \\
0, & \text{if } X_j - X_i = 0 \\
-1, & \text{if } X_j - X_i < 0 
\end{cases}$$

(3)

$Z$ values were utilized for trend detection. Mann (1945) stated that the $MK$ values were symmetrical and normally distributed when the null hypothesis stood. Given the possibility of ties in $X$ values, the variance of $MK$ was calculated using the Equation 4 (Kendall, 1975). Subsequently, $Z$ values were determined by Equation 5 based on the
variance of $MK$ values. The positive value of $Z$ indicated an upward trend in the data series and the negative value represented a downward trend.

\[
\text{var}(MK) = \frac{1}{18} \left[ n(n - 1)(2n + 5) - \sum_{p=1}^{q} t_p(t_p - 1)(2t_p + 5) \right]
\]  

(4)

where $n$ was the number of records, $q$ was the number of tied groups and $t_p$ was the number of data in the $p^{th}$ group (Gilbert and Richard, 1987).

\[
Z = \begin{cases} 
\frac{MK - 1}{\sqrt{\text{var}(MK)}}, & \text{if } MK > 0 \\
0, & \text{if } MK = 0 \\
\frac{MK + 1}{\sqrt{\text{var}(MK)}}, & \text{if } MK < 0 
\end{cases}
\]  

(5)

2.3 Streamflow trend tests in the Elkhorn River

Fig. 1 Locations of the eighteen stream gauging stations in the Elkhorn River Basin
Before the test, stations with less than ten years of records were eliminated from the tests, to guarantee the temporal sample size was big enough. The streamflow data were gathered from eighteen river gauging stations which were spread out evenly in the Elkhorn River Basin (Fig. 1). The annual streamflow rate in each water year was selected to conduct the trend test, which was the period from October 1st of the current year to September 31st of the next year. The hydrographs of the eighteen stations are shown in Figure 2. The locations and time periods of the eighteen gauging stations are summarized in Table 1.
Fig. 2 Hydrographs of annual streamflow rate of the 18 gauging station in the Elkorn River

The confidence level for the trend test was 0.05. When the $p$-value was less than the confidence level, the null hypothesis was rejected and a trend was proved to exist judged by the sign of $Z$ values. Otherwise, there wasn’t enough evidence to reject the null
hypothesis and no trend was detected. The statistical outputs of the $r^2$, $Z$ and $p$-values are listed in Table 1.

As the statistic results indicated, there were no negative $Z$ values with $p$-values that were less than 5%. In other words, significant decreasing trends were not detected in the eighteen sets of the annual flow rate data. For fifteen gauging stations, no trends were found in the annual streamflow data. This was consistent with the Wen and Chen’s conclusion on the streamflow trend of the five gauging station in the Elkhorn River. Nonetheless, streamflow rates of the other three gauging stations, of which the site numbers were Station 06800500, Station 06799450 and Station 06799500, presented upward tendencies. The stations concentrated on the Lower Elkhorn Basin, especially in Logan Creek area. This differed with Wen and Chen’s findings with Station 06799500 and Station 06800500 that the trends were insignificant. It was reasonable for the diverse conclusions because the time series in this study was longer.

To explore the connection between the streamflow trend and precipitation trend, the Mann-Kendall test was also performed on the annual precipitation series in the Lower Elkhorn area. The data were gathered from the Fremont climate station which was in the Lower Elkhorn Basin and the time series ranged from 1929 to 2012. The $Z$ value was 0.07 and $p$-value was 0.35, which indicated no trend was found in the precipitation data. However, the stream depletion caused by agricultural pumping may still exist even though no decreasing streamflow trends were found from the Mann-Kendall tests. The annual streamflow rate is affected by several hydrologic mechanisms, which could result in the increasing annual streamflow trend, such as precipitation, vegetation cover,
evapotranspiration, etc. Also, the annual flow rate trend test is not able to reveal the seasonal variation of streamflow. The intensive pumping in summer season may cause significant streamflow reduction.
Table 1 Statistic outputs of the Mann-Kendall tests for the time series of streamflow data

<table>
<thead>
<tr>
<th>Site number</th>
<th>Location</th>
<th>Period</th>
<th>$r^2$</th>
<th>Sample size</th>
<th>$Z$</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>06796973</td>
<td>Elkhorn River near Atkinson</td>
<td>1983-2012</td>
<td>0.27</td>
<td>30</td>
<td>-0.015</td>
<td>0.93</td>
</tr>
<tr>
<td>06796978</td>
<td>Holt Creek near Emmet</td>
<td>1979-1989</td>
<td>0.28</td>
<td>11</td>
<td>0.022</td>
<td>1.00</td>
</tr>
<tr>
<td>06797500</td>
<td>Elkhorn River at Ewing</td>
<td>1948-2012</td>
<td>0.25</td>
<td>65</td>
<td>0.070</td>
<td>0.41</td>
</tr>
<tr>
<td>06798000</td>
<td>South Fork at Ewing</td>
<td>1948-2012</td>
<td>0.26</td>
<td>65</td>
<td>0.026</td>
<td>0.79</td>
</tr>
<tr>
<td>06798300</td>
<td>Clear Water at Clearwater</td>
<td>1962-1991</td>
<td>0.50</td>
<td>17</td>
<td>0.050</td>
<td>0.82</td>
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<tr>
<td>06798500</td>
<td>Elkhorn River at Neligh</td>
<td>1931-2012</td>
<td>0.46</td>
<td>82</td>
<td>0.148</td>
<td>0.05</td>
</tr>
<tr>
<td>06799000</td>
<td>Elkhorn River at Norfolk</td>
<td>1946-2012</td>
<td>0.30</td>
<td>67</td>
<td>0.065</td>
<td>0.45</td>
</tr>
<tr>
<td>06799080</td>
<td>Willow Creek near Foster</td>
<td>1976-2012</td>
<td>0.54</td>
<td>37</td>
<td>0.133</td>
<td>0.26</td>
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<tr>
<td>06799100</td>
<td>North Fork near Pierce</td>
<td>1961-2012</td>
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<td>52</td>
<td>0.148</td>
<td>0.13</td>
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<tr>
<td>06799230</td>
<td>Union Creek at Madison</td>
<td>1979-2012</td>
<td>0.06</td>
<td>34</td>
<td>0.095</td>
<td>0.45</td>
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<tr>
<td>06799315</td>
<td>Elkhorn River at Pilger</td>
<td>2002-2012</td>
<td>0.42</td>
<td>11</td>
<td>0.200</td>
<td>0.47</td>
</tr>
<tr>
<td>06799350</td>
<td>Elkhorn River at West Point</td>
<td>1973-2012</td>
<td>0.46</td>
<td>40</td>
<td>0.144</td>
<td>0.20</td>
</tr>
<tr>
<td>06799385</td>
<td>Pebble Creek at Scribner</td>
<td>1979-2012</td>
<td>0.18</td>
<td>34</td>
<td>-0.03</td>
<td>0.82</td>
</tr>
<tr>
<td>06800000</td>
<td>Maple Creek near Nickerson</td>
<td>1952-2012</td>
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<td>61</td>
<td>0.132</td>
<td>0.14</td>
</tr>
<tr>
<td>06800500</td>
<td>Elkhorn River at Waterloo</td>
<td>1929-2012</td>
<td>0.49</td>
<td>83</td>
<td>0.183</td>
<td>0.01</td>
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<tr>
<td>06799445</td>
<td>Logan Creek at Wakefield</td>
<td>2003-2012</td>
<td>0.34</td>
<td>10</td>
<td>0.167</td>
<td>0.60</td>
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<tr>
<td>06799450</td>
<td>Logan Creek at Pender</td>
<td>1966-2012</td>
<td>0.46</td>
<td>47</td>
<td>0.231</td>
<td>0.02</td>
</tr>
<tr>
<td>06799500</td>
<td>Logan Creek near Uehling</td>
<td>1942-2012</td>
<td>0.49</td>
<td>71</td>
<td>0.175</td>
<td>0.04</td>
</tr>
</tbody>
</table>
CHAPTER 3 FIELD STUDY OF STREAMBED IN THE ELKHORN RIVER BASIN NEAR WINSLOW AND NORFOLK

3.1 Study location

The Elkhorn River is one of the largest tributaries of the Platte River in Nebraska. The river flows generally from the west-northwest to the east-southeast and the flow rate in the channel varies with the changes in the precipitation, snow melt and irrigation load. Its major tributaries include South and North Forks of the Elkhorn River, Logan Creek and Maple Creek. The river originates from the northeastern Sand Hills of Nebraska and ends with the confluence with the Platte River near Gretna, Nebraska. The mean annual precipitation within the basin changes from around 22 in in the west to about 30 in in the east (Huntzinger and Ellis, 1993). The principal aquifer system in this basin is composed of a paleovalley alluvial aquifer, the alluvial shallow aquifer, and the extensive Tertiary Ogallala aquifer (Diffendal and Voorhies, 1994). Groundwater tends to move from upland to the river, and the river is recharged by both base flow and precipitation. However, 66 percentage of the annual flow in the Elkhorn River is derived by the groundwater discharge (Peterson, 2008). The close hydrologic connection between the Elkhorn River and its adjacent aquifers should not be neglected.
In 2010, Chen and Lackey (2010) performed streambed hydrology tests in the Upper Elkhorn River at five sites near Stuart, Atkinson, Ewing and Neligh. Following that study, Lackey and Chen (2010) continued the tests in the Lower Elkhorn River and its tributaries at ten sites by Meadow Grove, Norfolk, Pilger, West Point, Willow Creek, North Fork, Taylor Creek, Maple Creek and Logan Creek. Information about the channel width, the thickness and layering of streambed sediments, and the vertical hydraulic conductivity of channel sediments were collected. In order to complete the streambed investigation in the Elkhorn River, two study sites were chosen and they are pointed out by red triangles in Figure 3.
The field investigation was accomplished in September, 2012. The Norfolk site was located at the border of the Upper and Lower Elkhorn Natural Resources Districts, about 7 mi southeast of Norfolk. From our measurements, the width of the river was 216 ft and the water depth varied from 2 to 9 in. The site lay in a loess-mantled plain (Newport, 1957) and the dominant land cover was lowland tallgrass prairie. The Ogallala group occupies most part of the deep aquifer system. Poorly consolidated and sorted sand, gravel and conglomerate are the major components.

The Winslow site was near the confluence of the Elkhorn River with the Platte River. It was on the east direction of Winslow within 2 mi. The width of the river was 140 ft and the depth to water varied from 4 to 20 in. Located at the farther east of the Elkhorn River Basin, the Winslow site lay in the glaciated-drift region (Newport, 1957) and upland tallgrass prairie. The Dakota group is the predominant geological unit in the deep aquifer, which consists mostly of white to red fine grained sandstones.

3.2 In-situ permeameter tests

In-situ permeameter test was proposed and conducted by Chen (2000) to determine the hydraulic conductivity and anisotropy of the streambed. It had unique advantages in measuring streambed hydraulic conductivity directly in river channels. This experiment was flexible for measuring in-situ sediment anisotropy without destroying the original channel texture.
Figure 4 is a schematic diagram showing the in-situ permeameter test. It began with a transparent tube inserted into the submerged sediment in the streambed. The hydraulic head and the sediment length were measured both inside and outside of the tube. Then, water was filled to the top of the tube and it would begin to drop owing to the hydraulic head difference. The rate of the declining water in the tube was mainly controlled by the $K_v$ value of the inner sediments and the hydraulic gradient. During each test, we recorded 6-8 hydraulic head readings and the elapsed time, which gave us information about how fast the water level within the tube declined. The hydraulic gradient was inferred from the hydraulic head measurements. Thus, the $K_v$ values of the unconsolidated sediments inside the tube were calculated using Hvorslev’s equation (1951):

$$K_v = \frac{\pi D + L_v}{11m} \ln \left( \frac{h_1}{h_2} \right)$$ (6)
where \( L_v \) was the length of the sediment column in the pipe, \( D \) was the inner diameter of the pipe, \( h_1 \) and \( h_2 \) were the hydraulic heads measured inside the pipe at elapsed times \( t_1 \) and \( t_2 \), and \( m = \sqrt{K_h / K_v} \), which was often unknown at the time of computation. If it was assumed that the thin layer of sediment at the surface of a streambed was isotropic (i.e., \( K_h = K_v \)), then Equation (6) could be simplified by setting \( m = 1 \). This simplification led to an overestimation of the \( K_v \) value for anisotropic stream sediments where \( K_h > K_v \). Chen (2000) noted that as long as the value of \( L_v \) was four times or more greater than the value of \( D \), use of Equation (6) will result in a very small error. For the tests in this study, the diameter of the tube \( (D) \) was 2 in. The sediment length in the tube \( (L_v) \) ranged from 18.8 to 23.5 in, which was nine times greater than \( D \) values. Thus, it was safe to use Equation 6 without overestimating \( K \) values.

In-situ permeameter tests were conducted mostly on the right bank of the river where the water was shallow. It was impossible to continue the tests across the channel to the left bank where the stream water was too deep to set up the equipment. Ten permeameter tests were conducted by the right bank and four more tests were carried out along a transect across the river (Fig. 5) at the Norfolk site. The tests for two neighboring points were 10 ft apart and the results are listed in Table 2. In addition to these in-situ permeameter tests, sediments cores and three EC logs were collected from the sandbar which was near the right bank of the Elkhorn River, by means of Geoprobe® direct push techniques.
Fig. 5 In-situ permeameter tests in the Elkhorn River streambed at the Norfolk site

Table 2 $K_v$ values of the top streambed sediments at the Norfolk site

<table>
<thead>
<tr>
<th>Point</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_v$ (in)</td>
<td>23.5</td>
<td>22.1</td>
<td>22.8</td>
<td>22.1</td>
<td>22.8</td>
<td>22.5</td>
<td>22.9</td>
<td>22.5</td>
<td>21.5</td>
<td>22.5</td>
<td>20.3</td>
<td>21.4</td>
<td>22.5</td>
<td>22.5</td>
</tr>
<tr>
<td>$K_v$ (ft/d)</td>
<td>77.5</td>
<td>15.6</td>
<td>40.5</td>
<td>28.0</td>
<td>18.5</td>
<td>29.4</td>
<td>67.6</td>
<td>16.0</td>
<td>10.5</td>
<td>10.3</td>
<td>10.1</td>
<td>49.6</td>
<td>101</td>
<td>72.3</td>
</tr>
</tbody>
</table>

Similar to the Norfolk site, seven in-situ permeameter tests were performed along the river and another three permeameter tests were conducted on a transect at the Winslow site (Fig. 6). The distance between every two adjacent test point was also 10 ft and Table 3 lists the $K_v$ values of the top streambed sediments. Five EC logs and one set of sediment cores were also gathered using Geoprobe®.
Fig. 6 In-situ permeameter tests in the Elkhorn River streambed at the Winslow site

Table 3 $K_v$ values of the top streambed sediments at the Winslow site

<table>
<thead>
<tr>
<th>Point</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_v$ (in)</td>
<td>19.5</td>
<td>18.5</td>
<td>20.25</td>
<td>19.5</td>
<td>18.5</td>
<td>18.8</td>
<td>19</td>
<td>20.5</td>
<td>21.1</td>
<td>21.25</td>
</tr>
<tr>
<td>$K_v$ (ft/d)</td>
<td>7.9</td>
<td>12.9</td>
<td>12.7</td>
<td>21.5</td>
<td>9.6</td>
<td>9.4</td>
<td>3.6</td>
<td>2.4</td>
<td>2.4</td>
<td>0.70</td>
</tr>
</tbody>
</table>

3.3 In-lab permeameter test and EC logs

The sediment cores collected by Geoprobe® techniques with plastic liners were brought back to the laboratory for further examination. The Geoprobe® equipment produced the subsurface sediment cores down to 40 ft for the Norfolk site and 25 ft for the Winslow site. The subsurface sediments in every 5 ft were contained in one core, so the different laminations, beddings and orientation could be visualized through the transparent pipes. Prior to the test, one end of the tube was covered by layers of fine
plastic screen with a clamp where water could get through without sediments falling out and it was submerged into a large water tank for sediment re-saturation. After that, permeameter tests were conducted on the sediment core. A schematic diagram of this test (Fig. 7) was referred from Cheng and Chen (2007) for illustration purpose. After re-saturation, the pipe was held vertically by a tripod with one end inundated in a water bucket. Water was continuously added to the upper end until the tube overflowed to ensure fully saturated sediment during the test. The water level within the tube fell as the test began and the drawdown was recorded for calculating $K_v$ based on Equation 7 which was derived from Darcy’s Law:

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

(7)

where $L_v$ was the length of sediment core in the tube; $h_1$ and $h_2$ were hydraulic heads at elapsed times $t_1$ and $t_2$, respectively. This test maintained the undisturbed sediment structure under the streambed and provided a $K_v$ profile in 5 ft increments. With the help of the EC logs, the layered sediment texture was characterized and the streambed conductance was calculated.
For the Norfolk site, three sets of EC logs (Fig. 8) were collected for the sediments below the land surface. The EC values of the sediments were impacted by water content, solution salinity, formation factors, soil mineralogy, etc. Generally, a low-permeable layer of silt/clay would give higher EC values than a layer of fine sand and gravel, which was shown as a spike curve in the EC log. It provided a remarkably different curve pattern in the EC log, so the layering of the subsurface sediments was easily shown the by EC log. Meanwhile, the in-lab permeameter tests produced $K_v$ values of the continual sequence of sediments cores. The $K_v$ values of sediment cores varied significantly, even with several orders of magnitude (Fig. 8). According to our observation, the streambed sediments were not homogeneous and the EC and $K_v$ values varied significantly with depth. From the EC logs and the $K_v$ profile, we could infer that the subsurface sediments in the up 38-40 ft of the Norfolk site were mostly permeable materials, mostly sand and gravel. At the depth of 40 ft, the bedrock showed up and sediment cores were not collected due to the high resistance of the sediments.
Fig. 8 Three EC logs and the $K_v$ profile of the subsurface sediments at the Norfolk site

Five EC logs were collected at the Winslow site, which were shown together with the $K_v$ profile generated by in-lab permeameter tests (Fig. 9). As indicated in this figure, the upper 10-15 ft of the subsurface sediments had high EC values and $K_v$ values, which resulted in high permeability. The low-permeability layer occurred at the depth of about 15 ft and had a thickness of about 10 ft as indicated by the EC logs. They were mixture of...
silt, clay, and till with traces of fine sands. The $K_v$ values for the last two sediment cores were extremely small (less than 0.01 ft/d), which would reduce stream-aquifer interaction at this site.
Fig. 9 Five EC logs and the $K_v$ profile of the subsurface sediments at the Winslow site.
4.1 Overview

4.1.1 Purpose and scope

The groundwater flow model development aimed at studying and quantifying the hydrologic connection between the Elkhorn River and its adjacent aquifer system at two areas, one area in the upper Elkhorn River near Atkinson, the other area in lower Elkhorn River near Winslow. The construction of the groundwater flow models incorporated not only the internal structure of the aquifer system but also the external hydrological processes including streams, recharge, evapotranspiration and groundwater pumping. The two models were calibrated under transient conditions. After the calibration, the two models were utilized for simulating various hypothetical irrigation scenarios to understand the effect of groundwater withdrawal on streamflow.

4.1.2 Model domain

The land surface of the modeling area in the upstream Elkhorn River was dominated by the Sand Hills features; the downstream model area was at the confluence of the Elkhorn River and the Platte River where irrigated crops were the major land covers. The model domains were 20 mi in length and 15 mi in width for each modeling area (Fig. 10).
Fig. 10 Location map showing the Atkinson and Winslow model areas in the Elkhorn River Basin

4.1.3 Climate, land use and hydrogeology

The Elkhorn River Basin lies in a typical continental climate zone with seasonal variation in temperature and precipitation. Most of the rainfall concentrates in the summer season (May – September) with relatively high temperature and little precipitation occurs in the winter season (October – April) with relatively low temperature (Fig. 11 and Fig. 12). The mean annual precipitation from 1952 to 2012 in the O’Neil weather station, which was located in the Atkinson model area, was 23.8 in and the mean annual precipitation from 1960 to 2010 in the Fremont weather station was 30.1 in, which was located in the Winslow model area. The distribution of monthly
precipitation for the two weather stations is shown in Figure 12. As shown in Figure 11, the temperature differences between the two study sites are not noticeable, whereas the precipitation in the Winslow area is slightly greater than that in the Atkinson area.

![Fig. 11 Monthly temperature distribution graph of the O'Neil and Fremont weather station](image1)

![Fig. 12 Monthly precipitation distribution graph of the O’Neil and Fremont weather station](image2)

The elevation for the Elkhorn River Basin was generally higher in the west (the Sand Hills) and lower eastward. The main features around the Atkinson area were the
scattered sand dunes of various sizes with dried hay meadow and other grasses, where livestock raising and dryland farming were commonplace (Fig. 13). In contrast, the Winslow area lay in a tall-grass prairie (Fig. 14). Low-permeability glacial-till deposits occurred at the Lower Elkhorn River Basin where the wet meadows, river valleys and marsh plains were dominant (Huntzinger and ellis, 1993). Soils in the upstream area of the Elkhorn River are more coarse-grained as compared to the east Elkhorn Basin, resulting in little runoff and high recharge rate to the groundwater (Keech and Bentall, 1971).

Fig. 13 Landscape map of the Atkinson model area
Ground water in the Elkhorn River Basin generally flowed from west to east with an average water-table slope of about 10 ft/mi, with a larger hydraulic gradient in the upstream area and a lower value for the eastern area (Peterson, 2008). The saturated thickness of the principal aquifers in the Elkhorn River Basin varied from 0 to 800 ft (Li, 2012). According to observation records of USGS monitoring wells, the depth to the water table ranged from 0 to more than 44 ft in the Atkinson area, and 0 to more than 73 ft in the Winslow area.

4.2 The groundwater and surface hydrology

4.2.1 Hydrostratigraphic units in the Atkinson model

A Nebraska geologic map was obtained through USGS online spatial dataset (U.S. Geological Survey, 2013). The description of the major lithological units in the Atkinson area is summarized in Table 4. For the Atkinson site, the alluvial sediments from the
Quaternary period were widespread in the upper layer of the land. Most of the sediments were very coarse to fine-grained unconsolidated sand, gravel and wind-deposited silt with top soil. The thickness of this layer varied from 20 to 150 ft. Sediments from the Ogallala Group of the Tertiary period existed below the alluvial deposits, which were mostly light olive very fine to fine sand, sandstone, siltstone, interfingered calcareous silt, and locally poor consolidated conglomerate. The sediments in the Ogallala Group tended to be more fine-grained than the alluvial deposits, but had larger saturated thickness (Peterson, 2008). The thickness of the Ogallala aquifer in the Atkinson area ranged from 250 ft to more than 440 ft. Thus, a large number of groundwater irrigation wells were screened in the Ogallala aquifer and pumped intensively in the growing season. The Montana Group from the Upper Cretaceous acted as the bedrock of the local aquifer system. It consisted of gray, yellowish or black, clayey and slightly calcareous shale. Based on the numerous test holes and registered well log data from the Conservation and Survey Division (CSD), the local groundwater system in the Atkinson area was divided into five hydrostratigraphic units. Figure 15-a and Figure 15-b were examples showing how the hydrostratigraphic units were interpreted. The red multiply signs in Figure 15-a showed the spatial locations of the registered wells and test holes. Figure 15-b was a sketch depicting the five layers of the aquifer system in the Atkinson model area. Layer 1, the upmost layer, represented the Quaternary alluvial deposits with high permeability. Layer 2 and 4 stood for the low-permeability sediments that commonly existed among the alluvial and Ogallala aquifers. Layer 3 and Layer 5 were used to simulate the main Ogallala aquifer components of fine sand, silty sand, sandstone, etc. The shale from the Montana Group was used as the baseline of Layer 5.
Table 4 Summary of the lithological units in the Atkinson area

<table>
<thead>
<tr>
<th>Location</th>
<th>System</th>
<th>Series</th>
<th>Group</th>
<th>Lithological Units</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atkinson</td>
<td>Quaternary</td>
<td>-</td>
<td>-</td>
<td>Sand, gravel with topsoil</td>
<td>Fine to coarse alluvial sand, fine to coarse gravel with top soil; Depth varies from 20 to 125 ft.</td>
</tr>
<tr>
<td>Tertiary</td>
<td>Miocene</td>
<td>Ogallala</td>
<td></td>
<td>Sand, gravel sandstone, conglomerate interbedded with silt/clay</td>
<td>Mostly light olive very fine to fine sand, sandstone, interfingered with calcareous silt, and locally poor consolidated conglomerate, siltstone; Depth varies from 250 to 440 ft.</td>
</tr>
<tr>
<td>Cretaceous</td>
<td>Upper C</td>
<td>Montana</td>
<td>Shale</td>
<td></td>
<td>Gray, yellowish or black, clayey, slightly calcareous shale</td>
</tr>
</tbody>
</table>


Fig. 15-a Map showing the registered wells and test holes in the Atkinson model area.

Fig. 15-b Stratigraphic columns for the selected wells and test holes (see Fig. 15-a) in the Atkinson model area. The aquifer system was divided into five model layers.
4.2.2 Hydrostratigraphic units in the Winslow model area

The lithological units and their description of the aquifer system in the Winslow area are summarized in Table 5. Similar to the Atkinson model, the upper layer of the Winslow model was from the Quaternary sediments but the principal mantle deposits was capped by loess (Newport, 1957). There was a thick layer of medium brown to black soil/clay/silt mixture commonly occurring in the upper part of the Winslow groundwater system. The Winslow area lay in the glacial-drift region and the Ogallala Group had been eroded away. The Dakota Group from early the Cretaceous period was the dominant lithological unit in this area. Very fine to coarse grained, crossbedded and lenticular sandstone was widely spread in the Dakota Group, with layers of light gray, sandy carbonaceous shale. The maximum thickness of the Dakota Group was around 148 ft. The limestone from the Pennsylvanian-Missourian-Kansas City Group acted as the shale of this groundwater system. It was light brownish gray limestone with fine crystallines, ostracods and cherts with dark gray shale with scattered limestones. Figure 16-a and Figure 16-b were used to reveal how the hydrostratigraphic units were interpreted. The red multiply signs in Figure 16-a showed the spatial locations of the registered wells and test holes. Five hydrostratigraphic units were interpreted and incorporated in the Winslow model area (Fig. 16-b). The first layer (Layer 1) stood for the alluvial sediments from the Quaternary System and Layer 2 and 4 was low-permeable silt/clay/till deposits. The Dakota Group represented by Layer 3 and Layer 5 was permeable sediments.
<table>
<thead>
<tr>
<th>Location</th>
<th>System</th>
<th>Series</th>
<th>Group</th>
<th>Lithological Units</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winslow</td>
<td>Quaternary</td>
<td>-</td>
<td>-</td>
<td>Sand, gravel, clay, silt and till</td>
<td>Alluvial sand and gravel with fine texture, interbedded with thin layer of clay/silt/till. Depth varies from 60 to 209 ft.</td>
</tr>
<tr>
<td>Cretaceous</td>
<td>Lower C</td>
<td>Dakota</td>
<td>Sandstone, shale, sand and siltstone</td>
<td>Very fine to coarse grained, crossbedded, and lenticular sandstone with layers of light gray, sandy carbonaceous shale. Maximum thickness is around 148 ft.</td>
<td></td>
</tr>
<tr>
<td>Pennsylvanian</td>
<td>Missourian</td>
<td>Kansas city</td>
<td>Limestone; shale</td>
<td>Light brownish gray limestone with fine crystalline, ostracods and chert. Dark gray shale with scattered limy zones.</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 16-a Map showing the registered wells and test holes in the Winslow model area.

Fig. 16-b Stratigraphic columns for the selected wells and test holes (see Fig. 16-a) in the Winslow model area. The aquifer system was divided into five model layers.
4.2.3 Recharge

Before the precipitation reached the water table, there were several mechanisms intercepting the water from recharge. It could be captured by the leaf canopy, evaporation, transpiration, surface runoff, plant root uptake, and lateral flow in unsaturated zone. It was complex and difficult to quantify the impact for each parameter on the process of groundwater recharge. Here, a simplified method was utilized to estimate groundwater recharge in this study:

\[ R = \alpha P + \alpha_i Q \]  

(8)

where \( R \) was the groundwater recharge; \( \alpha \) was the ratio of the water reaching the aquifer to the areal precipitation; \( P \) was the areal precipitation; \( Q \) was the groundwater pumpage; \( \alpha_i \) was the percentage of the groundwater pumpage returned to the aquifer.

Monthly precipitation data were employed in the above equation to estimate the groundwater recharge. The recharge ratio \( \alpha \) was a parameter changing with the different land use types, vegetation covers, soil texture, aridity index, etc. Generally, it was highest on the irrigation cropland and lowest on the urban area where low recharge occurred on concrete or asphalt surface. Because the irrigation dimension has been expanded since 1930’s, the returned flow from the groundwater pumping added additional recharge to the aquifer system. An increasing \( \alpha_i \) value was utilized to simulate the enlarged amount of the anthropic recharge over years as pumping continued. Based on the time-dependent number of registered wells within the model simulation period, we manually designated six irrigation periods for the Atkinson area, and five irrigation periods for the Winslow
area, as shown in Fig. 17. Each irrigation period had a unique $\alpha_i$ value representing the return flow towards the aquifer.

![Graph showing the number of registered wells over time]

Fig. 17 Historical increase in the number of wells in the Winslow area with periodical nodes

4.2.4 Evapotranspiration (ET)

Groundwater ET didn’t occur in the system as much as surface water network did. The groundwater ET amount was associated with water table depth, soil types and vegetation covers. Normally, groundwater ET occurred more intensive near the river valleys where the depth to water was small. In the Atkinson area, sand dunes with scattered hay meadows were the typical views where the groundwater table was shallow. Four ET zones were incorporated in the model depending on the land use types and the water table depth. The ET zone for riparian area was determined by buffering the Elkhorn River polyline file for a distance of 3000 ft. However, for the Winslow area, only one ET
zone was developed for the riparian area because few ET occurred in the rest of the Winslow area where the minimum depth to water was 20 ft. In this study, we assumed ET occurred mostly in summer season (May to September) and no groundwater was lost through ET when the depth to water exceeded 15 ft. A maximum ET rate was given for each ET zone (Table 6).

Table 6 Maximum groundwater ET rates for ET zones in the Atkinson and Winslow areas

<table>
<thead>
<tr>
<th>ET zones</th>
<th>ET in Atkinson (in/yr)</th>
<th>ET in Winslow (in/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Irrigated land</td>
<td>2.6</td>
<td>-</td>
</tr>
<tr>
<td>Pasture</td>
<td>1.2</td>
<td>-</td>
</tr>
<tr>
<td>Urban area</td>
<td>0.16</td>
<td>-</td>
</tr>
<tr>
<td>Riparian area</td>
<td>5.4</td>
<td>2.6</td>
</tr>
</tbody>
</table>

4.2.5 Groundwater pumping

The registered wells of high capacity in the model area were grouped based on water use categories, including the commercial/industrial consumption, agricultural irrigation and livestock uses. Crop irrigation wells accounted for more than half of the total registered wells in the Atkinson and Winslow areas. The actual pumping rate of each well cannot be directly determined because the irrigated inches depended on the climate conditions and farmers’ decisions. Thus, the pumping rate of the irrigation wells needed to be estimated based on the irrigation requirement of the crops and the precipitation in the summer season.
The pumping schedule was set to be June to August for each year. Sharma (2011) quantified the net irrigation requirement of corn and soybean for each county in Nebraska. For Dodge County (the Winslow area), corn needed 6.94 in for June to August and soybean needed 4.81 in. For Holt County (the Atkinson area), it was 9.56 in for corn and 7.40 in for soybean. In the Atkinson area, the area percentage of corn cultivation was 61.38% and soybean percentage was 38.2%, whilst the corn percentage in the Winslow area was 53.5% and 46.5% for soybean. According to the area ratio of the cultured crops, the multi-year averaged net irrigation requirement ($\mu_{net}$) was 8.74 in for the Atkinson area and 5.95 in for the Winslow area. Thus, the net irrigation requirement for each year was calculated as follows:

$$Q_{net} = \mu_{net} + \mu_{summer} - P_{summer}$$  \hspace{1cm} (9)

where $Q_{net}$ was the net irrigation requirement of each year for the Winslow and Atkinson area; $\mu_{net}$ was the multi-year averaged net irrigation requirement; $\mu_{summer}$ was the multi-year averaged precipitation in the summer from June to August. $P_{summer}$ was the summer precipitation (June to August) in every year. The pumping rates for the irrigation wells in every year could be estimated by the following equation:

$$Q = \frac{Q_{net}A}{\epsilon T}$$  \hspace{1cm} (10)

where $A$ was the irrigated area which was delineated from the land use map. $\epsilon$ was the application efficiency for the irrigation system. In this case, sprinkler irrigation with center pivot was widely utilized in the area and $\epsilon$ was set to be 0.7. $T$ was the total irrigation time from June 1st to August 31st. The pumping rates for non-agricultural
irrigation wells, such as commercial/industrial use, livestock use and domestic use, were estimated by 1/10 of their pumping capacity.

4.2.6 Stream network

In the Atkinson model area, the Elkhorn River was included in the model and monthly river stage was gathered and interpreted to simulate the water table variation. The Atkinson station was installed in October, 1982, so the stage records before 1982 had to be interpreted from the nearby stream stations. We built up a polynomial relationship between the stream stages from 1982 to 2012 in the Atkinson station with the same time period data in the Ewing station. By the help of the connection of the two stations, the monthly stages from 1952 to 1982 in the Atkinson were inferred from the early water table records in the Ewing station. The monthly river stages from 1952-2012 are shown in Figure 18.

![Fig. 18 Monthly river stages from 1952 to 2012 at the Atkinson site](image)
For the Winslow model area, not only was the Elkhorn River incorporated in the Winslow model but also the two tributaries, Logan Creek and Maple Creek, were included for the analysis of stream-aquifer interactions. The Winslow station was established by Department of Natural Resources on July 11, 2007. Same strategy was used in this case and the monthly river stages were interpreted from the Waterloo station. Due to the relatively steady water table in Maple Creek and Logan Creek, the annual stream stage was used for these two rivers.

The streambed conductance was a critical parameter in quantifying groundwater and surface water interaction (Sophocleous et al., 1995). It was determined by the thickness of low-permeability streambed \( M \), length of the stream channel \( L \), the vertical hydraulic conductivity of the streambed \( K_v \) and the width of the submerged channel \( W \) (McDonald and Harbaugh, 1988):

\[
C = \frac{K_v L W}{M}
\]  

(11)

The distance from the channel surface to the detected low-permeability layer, which was determined from the EC logs, was used as \( M \) value. An equivalent \( K_v \) value was needed for quantifying the heterogeneity of each study site. The following equation for calculating equivalent \( K_{equ} \) was adopted from Freeze and Cherry (1979):

\[
K_{equ} = \left( \frac{b_1 + b_2 + b_3 + \cdots + b_N}{\frac{b_1}{K_{v1}} + \frac{b_2}{K_{v2}} + \cdots + \frac{b_N}{K_{vN}}} \right)
\]  

(12)

where \( b_1 \) was the sediment length in the sample collected from the ground surface to 5 ft deep in the streambed, \( K_{v1} \) was the vertical hydraulic conductivity of \( b_1 \) sediment. The
rest could be done in the same manner until to \( b_N \) of the sediment length in the sample taken out from the \( N_{th} \) 5 ft core.

The information in the five EC logs and the \( K_v \) profile (Fig. 9) for the Winslow site was used to estimate the channel sediment thickness \( (M) \) which was 25 ft. The equivalent \( K_v \) value was 0.0027 ft/d calculated by the Equation 12. The streambed conductance \( (C) \) for the Elkhorn River in the Winslow area was 0.015 ft\(^2\)/d when the length of stream channel was set to be 1 ft. The channel width \( (W) \), \( K_v \) values and the streambed thickness \( (M) \) for Maple Creek and Logan Creek were referenced from Lackey and Chen (2010). When the length of the channel was 1 ft, the calculated streambed conductance \( (C) \) for Maple Creek and Logan Creek were 0.086 ft\(^2\)/d and 0.09 ft\(^2\)/d, respectively. According to Wang (2012), the equivalent \( K_v \) values for the streambed in the Atkinson area was 0.112 ft/d and the streambed conductance \( (C) \) was 0.6 ft\(^2\)/d.

4.3 Transient numerical models using MODFLOW

4.3.1 Space and time discretization

The dimension of the Atkinson model was 20 mi \( \times \) 15 mi. It was divided into 120 rows and 160 columns and the size of each grid was 660 ft \( \times \) 660 ft. The surface elevation in the model varied from 1580 to 2343 ft. The model simulation period was from 09/01/1952 to 08/31/2012. Instead of using real time, there were 720 stress periods divided in the whole simulation period; each stress period represented one month. The size of the Winslow model was also 20 mi in length and 15 mi in width and there were 132 rows and 176 columns. The grid size was 600 ft \( \times \) 600 ft and the elevation range was 864 ft to 1423 ft. The modeling period was 50 years from 09/01/1960 to 08/31/2010,
including 600 stress periods. In both models, each stress period was divided into ten time steps with a geometric ratio of 1.2 for smoother simulation curve.

4.3.2 Initial condition

The initial condition was essential for the transient model simulation. The observation wells in both the Atkinson and Winslow models were sparse and unevenly located. For each well, the averaged water table record before the 1970’s were gathered and utilized for initial water table estimation because the groundwater table before 1970’s in this region did not show much decline by pumping. The initial head of groundwater was interpreted using Kriging algorithm, a widely used geostatistical method in hydrogeology (Desbarats et al. 2002; Theodossiou and Latinopoulos, 2006; Gundogdu and Guney, 2007). A linear variogram was incorporated in the Kriging method and the contour maps of the initial water head are shown in Figure 19 and Figure 20, respectively, for the two modeling areas.
Fig. 19 Contour map of the initial hydraulic head in the Atkinson model area

Fig. 20 Contour map of the initial hydraulic head in the Winslow model area
4.3.3 Hydraulic property

In this study, the hydraulic properties including the vertical and horizontal hydraulic conductivity \((K_v\) and \(K_h\)), storage coefficient \((S_s)\) for the confined aquifer, specific yield \((S_y)\) for the unconfined aquifer and porosity \((\phi)\), were estimated based on empirical values and previous research. Domenico and Schwartz (1990) gave a table of representative values of \(K\) values and the \(S_s\) and \(S_y\) value were referenced from Batu (1998). Chen and Ou (2012) developed a groundwater flow model for the Lower Platte North Natural Resources District area and the calibrated hydraulic properties used in their study were considered as well. In reality, hydraulic properties could change with locations in a small scale and it was almost impossible to reflect small-scale variations of the aquifer hydraulic properties in the two models because the grid spacing was greater than 600 ft. Several hydraulic property zones were identified based on the test holes and registered well log data. Same hydraulic property values were given for a group of similar sediments.

4.3.4 Boundary conditions

Because the model area was only part of a hydrogeological system, appropriate boundary conditions must be given to the two models. The boundary conditions described the water entering and exiting the model domain. According to the characteristics of the water flux, there are three types of the boundary conditions: lateral flow, areal flow and internal source and sink.

Lateral flow boundary represented the water fluxes along the physical boundaries of the model, including constant head boundary (CHD) and general head boundary
(GHB). CHD fixes the hydraulic head value on the boundary, ignoring the head changes within the model and acts as an infinite source or sink for the grids near the boundary. It is often used when a river or water reservoir with constant head was located near the boundary. GHB generates the water flow in or out of the boundary by specifying an external source, the distance to the cell and the assigned head value of the source. This type of boundary is generally used in the numerical modeling and avoids unnecessarily extending the model domain to unconcerned area. In both the Atkinson and Winslow models, GHB was employed to simulate all the lateral boundary conditions. The conductance of the boundary was estimated based on the description of the registered well logs. The external boundary head was determined from the map of the water table in 1995 that was produced by the Conservation and Survey Division (Fig. 21).

Fig. 21 Boundary determination by the water table map in 1995 from CSD
The areal boundaries were the distributed flux that entered or exited the surficial model area, which included the recharge and evapotranspiration in the Atkinson and Winslow models. Recharge was applied to layer 1, which is the upper-most layer in the model because the natural recharge occurred near the ground surface. The horizontal plane of the model was composed of three recharge zones, based on the predominant land use types and vegetation covers. They were irrigation crop zone, dryland crop/pasture zone and urban zone. The ratio of the recharge to the water that entered into the ground surface was estimated and calibration was discussion in the next section. The evapotranspiration zones simulated the effect of direct evaporation, crop transpiration and seepage at the ground surface by removing water from the saturated groundwater regime (Schlumberger, 2011). ET occurs when the water table is shallow and near the ground surface. In these two models, the ET zones were set to be the same with the recharge zones. In MODFLOW (McDonald and Harbaugh 1988), the ET rates were assumed to vary linearly with depth and were estimated by the following equations:

\[
Q_e = \begin{cases} 
\text{PET} & h > \text{SURF} \\
\frac{\text{PET}}{\text{EXPD}} \left( \text{SURF} - \text{EXDP} \right) & (\text{SURF} - \text{EXDP}) \leq h \leq \text{SURF} \\
0 & h < \text{SURF} - \text{EXDP}
\end{cases}
\]  

(13)

where \(Q_e\) was the estimated ET rate; \(PET\) was the potential groundwater ET; \(SURF\) was the ET surface elevation which coincided the land surface in the model; \(EXDP\) was the extinction depth below which the groundwater ET ceased; and \(h\) was the groundwater level.
In this study, internal boundaries were associated with streambed seepage and well pumping. The water interchange on the streambed between the river and the aquifers was simulated using the RIV package of the MODFLOW. Given the streambed elevation of the gauging station, the streambed elevation of the simulated river segments was estimated by the topographic gradient. Thereafter, the monthly stage data was collected for the starting and ending point of the river and the stage data for all the river grid was linearly interpolated in Visual MODFLOW. The hydraulic gradient between the Elkhorn River and its adjacent aquifers determined the direction in which the water moved. The groundwater recharges to the river as base flow when the groundwater table was higher, or the river discharges to the aquifer causing stream depletion which could threaten the riparian environment. The pumping wells were incorporated in the models using WEL Package of MODFLOW. Registered well information within the model area was collected from the Nebraska Department of Natural Resources and was organized for model input. The pumping rates were estimated by crop irrigation requirement, summer precipitation and pumping capacity. The pumping schedule was set from June 1st to August 31st. For some wells which had screen depth on record, they were directly used in the model. For the other wells, the screen depth was artificially determined by means of total well depth. In total, there were 711 registered wells in the Atkinson model and 844 ones in the Winslow model.

4.3.4 Model run settings

Both the Atkinson and Winslow models were developed with the Visual MODFLOW Classic software. MODFLOW-2005 was chosen as the numerical engine for
the two groundwater flow models. It is a finite-difference three-dimensional groundwater flow model with independent modules that was developed by the U.S. Geological Survey. Geometric Multigrid Solver (GMG) was utilized because it has been demonstrated to reduce model run time as opposed to other solvers (Schlumberger, 2011). Recharge and evapotranspiration were both applied to the uppermost layer. Layer 1 in the Atkinson and Winslow models was considered as unconfined layer and the other layers were confined. The Zone Budge module was also used in the two models. It calculated the regional and sub-regional water budgets and would be utilized for analyzing the stream depletion.

4.4 Model calibration

4.4.1 Parameters adjustment

Because pumping tests were not performed in the study area, the hydraulic properties of the aquifer system was estimated based on empirical values, previous studies and the test hole records. This may result in a significant difference between the real values with the simulated ones. Also, a simplified method was employed to simulate the recharge rate and ET rates and there were no accurate field-measured data to rely on. Thus, the hydraulic properties including hydraulic conductivity, storage coefficient, specific yield and porosity, the recharge rate and ET rates were adjusted using the manual trial-and-error approach. The estimated annual recharge rates for the two models are shown in Figure 22 and the maximum ET rates are identified in Table 6. The calibrated hydraulic properties for the Atkinson and Winslow models are listed in Table 7. Hydraulic properties were assigned to the grids based on the well logs and modeler’s judgment during model calibration. The identification of the recharge rates, ET rates and
hydraulic properties were supposed to provide a reasonable and agreeable fit between the simulated water tables with the observed records.

Fig. 22 Estimated annual recharge rate for the two groundwater flow models
Table 7 Calibrated hydraulic properties in the two groundwater flow models

<table>
<thead>
<tr>
<th>Model</th>
<th>Zone</th>
<th>$K_h$ (ft/d)</th>
<th>$K_v$ (ft/d)</th>
<th>$S_s$ (ft$^{-1}$)</th>
<th>$S_y$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atkinson</td>
<td>Layer 1 (unconfined)</td>
<td>60</td>
<td>10</td>
<td>6E-05</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>Low-permeable sediments in Layer 1</td>
<td>0.5</td>
<td>0.1</td>
<td>5E-06</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td>High-permeable sediments in Layer 1</td>
<td>100</td>
<td>34</td>
<td>9E-05</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>Layer 2 and Layer 4</td>
<td>0.01</td>
<td>0.005</td>
<td>1E-06</td>
<td>0.01</td>
</tr>
<tr>
<td></td>
<td>Layer 3 and Layer 5</td>
<td>10</td>
<td>3</td>
<td>2E-05</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td>Low-permeable sediments in confined</td>
<td>2</td>
<td>0.5</td>
<td>1E-05</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>aquifer (localized)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winslow</td>
<td>Layer 1 (unconfined)</td>
<td>40</td>
<td>8</td>
<td>4E-05</td>
<td>0.06</td>
</tr>
<tr>
<td></td>
<td>Layer 2 and Layer 4</td>
<td>0.5</td>
<td>0.1</td>
<td>1E-06</td>
<td>0.004</td>
</tr>
<tr>
<td></td>
<td>Layer 3 and Layer 5</td>
<td>6</td>
<td>1</td>
<td>1E-05</td>
<td>0.04</td>
</tr>
<tr>
<td></td>
<td>High-permeable sediments (localized)</td>
<td>70</td>
<td>12</td>
<td>5E-05</td>
<td>0.08</td>
</tr>
<tr>
<td></td>
<td>Low-permeable sediments (localized)</td>
<td>0.1</td>
<td>0.05</td>
<td>2E-05</td>
<td>0.01</td>
</tr>
</tbody>
</table>

4.4.2 Simulated water table with observation plot

Twenty-two and thirty-three USGS observation wells in the Winslow and Atkinson models provided historical water level records and were used for the model calibration. The chosen wells had at least ten temporal records of the groundwater table. The calibration process was influenced by the modelers’ expertise and biases. Based on the latest data and our knowledge, this was the best statistical solution that we could get. The quality and accuracy of these groundwater models would be enhanced if certain
aquifer tests were performed in the future or the study area was broaden. Part of the
diagram of the simulated water table with the observation plots in the Atkinson and
Winslow models are shown in the Appendix A and B, respectively.

4.5 Regional water budget

The accumulative volume amount of water entering (in) and exiting (out) the
aquifer in the total simulation period through flow boundary, aquifer storage changes,
hydrologic processes and surface water is summarized in Table 8. The total simulation
period was 60 years for the Atkinson model and 50 years for the Winslow model. Storage
meant the groundwater that reserved in the shallow and deep aquifers. As shown in this
table, there was no significant groundwater storage reduction in the Atkinson and
Winslow model areas. The irrigation well pumpage accounts for 28% of the overall flow
amount. For the two groundwater flow models, the river leakage rate for the flowed out
water was greater than it for the flowed in water, which indicated that the Elkhorn River
and its tributaries received groundwater baseflow from the nearby aquifers. For the
Atkinson and Winslow models, the percent discrepancy between the total inflow and
outflow was less than 0.2%, which meant the numerical solutions was precise.
Table 8 Accumulative water volumes from the Atkinson and Winslow models outputs

<table>
<thead>
<tr>
<th>Items</th>
<th>Atkinson (out)</th>
<th>Atkinson</th>
<th>Winslow (out)</th>
<th>Winslow</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storage (ft³)</td>
<td>2.29E+10</td>
<td>2.45E+10</td>
<td>2.06E+10</td>
<td>1.91E+10</td>
</tr>
<tr>
<td>Wells (ft³)</td>
<td>2.63E+10</td>
<td>0.00E+00</td>
<td>2.81E+10</td>
<td>0.00E+00</td>
</tr>
<tr>
<td>River leakage (ft³)</td>
<td>8.89E+09</td>
<td>3.43E+08</td>
<td>3.16E+09</td>
<td>6.26E+06</td>
</tr>
<tr>
<td>Evapotranspiration (ft³)</td>
<td>1.51E+10</td>
<td>0.00E+00</td>
<td>1.00E+10</td>
<td>0.00E+00</td>
</tr>
<tr>
<td>General head boundary (ft³)</td>
<td>1.97E+10</td>
<td>2.67E+10</td>
<td>2.14E+10</td>
<td>1.69E+10</td>
</tr>
<tr>
<td>Recharge (ft³)</td>
<td>0.00E+00</td>
<td>4.11E+10</td>
<td>0.00E+00</td>
<td>4.74E+10</td>
</tr>
<tr>
<td>Total (ft³)</td>
<td>9.3E+10</td>
<td>9.3E+10</td>
<td>8.3E+10</td>
<td>8.3E+10</td>
</tr>
</tbody>
</table>

4.6 Stream depletion simulation using the calibrated models

4.6.1 Introduction

A large number of studies have modeled stream depletion using analytical solutions where there is groundwater pumping. Theis (1941) was a pioneer in deriving the unsteady state solution for stream depletion under several assumptions, including infinitely long river edge, a fully penetrating stream and homogeneous aquifer sediments. Later, Hantush (1965) solved the problem with a streambed lined with semi-pervious material. Hunt (1999) obtained the solution of leaked streamflow when the stream shallowly penetrated in an infinite aquifer. Zoltnik (2004) and Lough and Hunt (2006) continued to generalize and extended the analytical solutions for calculation of stream depletion. Despite the numerous findings and considerable advances in those studies, the
analytical solutions were not capable to incorporate the realistic features of hydrologic system and external processes flexibly, such as evapotranspiration, recharge, streambed and aquifer hydraulic property heterogeneity, multiple-well pumping schedules, etc. MODFLOW-based numerical modeling approach has been used to construct stream-aquifer systems and to model stream depletion caused by groundwater pumping (Sophocleous et al., 1995; Chen and Shu, 2006; Zume and Tarhule, 2008). The numerical modeling approach can overcome some difficulties in handling aquifer heterogeneity faced by the analytical solutions.

In order to take into account the complex hydrologic situation in the Elkhorn River Basin, the calibrated numerical methods were employed for the streamflow depletion study. Simulation pumping wells were placed in the calibrated Atkinson and Winslow groundwater models, to estimate the stream depletion rate. In the early stage of pumping, the depletion rate was relatively low due to the groundwater storage utilization. As pumping continued, the depletion rate increased because the drawdown of the pumping well intercepted the groundwater inflow to stream, resulting in larger stream depletion rate. Finally, the stream depletion would become stable. The stream depletion rate was calculated by the difference of the river exchange rate over the pumping rate:

\[ D_p = \frac{Q_{riv} - Q_{riv}}{Q_p} \]  

where \( D_p \) was the depletion ratio for the new well; \( Q_{riv} \) and \( Q_{riv} \) were the exchange rates between streams and aquifers after and before adding the new pumping well in the model; \( Q_p \) was the pumping rate of the added hypothetical well.
We examined the stream depletion rate caused by hypothetical pumping wells in different locations. Figure 23 and Figure 24 show the well locations in the Atkinson and Winslow models. The distance between the pumping well and the Elkhorn River was set to be 0.62 mi (1000 m), 1.86 mi (3000 m), and 3.72 mi (6000 m), respectively. The pumping rate was 100 gpm (gallon per minute), which was lower than the average rate of a real well in this area, to avoid the occurrence of dry cells in the top layer of the two models.

Fig. 23 Distribution of the hypothetical wells in the Atkinson model area
4.6.2 Stream depletion curve

The stream depletion rate curve was generated for each pumping well of various pumping scenarios. The depletion rate represented the stream leakage rate for all the simulated rivers in each model. The procedure was described in three phases.

1. Ran the calibrated model with no hypothetical well with zone budget module; extracted the river leakage rate for each time period (each month).
2. Inserted the hypothetical well; let it pump for 50 year with a constant rate (100 gpm); extracted the river leakage rate for this simulation as well.
3. Combined and compared the river leakage rate for the above two runs; calculated the stream depletion rate for each time period using the above equation; plotted the stream depletion curve for each well.
4.6.3 Various pumping scenarios

For the first phase of simulation, we located the well screen in the deep aquifers which is common for irrigation wells. The well screen of the hypothetical wells was in the Ogallala Group for the Atkinson model and in the Dakota Group for the Winslow model. The hypothetical wells were set to pump continuously for fifty years: from 1952 to 2002 in the Atkinson model and from 1960 to 2010 in the Winslow model.

![Fig. 25 Stream depletion curves for all-year pumping wells located at 0.62 mi to the river in the Atkinson model](image)

Fig. 25 Stream depletion curves for all-year pumping wells located at 0.62 mi to the river in the Atkinson model

To study the spatial variation in stream depletion caused by wells, we put three hypothetical wells (HW1, HW3 and HW6) from the western to the eastern Atkinson model with the same distance to river (0.62 m). Figure 25 shows the stream depletion rate for HW1 was the highest (56%), following by HW6 (53%) and HW3 (49%). Although a uniform streambed conductance was used in the simulated river, the stream depletion rates for the wells in different locations were not same due to the anisotropy of the
aquifer system. In the Winslow model, the stream depletion curves for the wells of HW1, HW4, HW5, HW8, HW9 and HW10 were computed and shown in Figure 26. As the curves implied, the stream depletion rate increased when the wells moved eastwards with the same distance to river (0.62 mi). The stabilized stream depletion rate changed from 5% to 25%. Besides the effect of the hydraulic gradient in stream depletion, the increased depletion rate could also be led by the accumulative effect of Maple Creek and Logan Creek leakage. In the eastern part of the model area, the drawdown of the pumping intercepted the base flow or induced stream infiltration of not only the Elkhorn River but also Maple Creek and Logan Creek.

Fig. 26 Stream depletion curves of all-year pumping wells located at 0.62 mi to the river in the Winslow model
The effect of the distance between the pumping well and river on stream depletion was also observed. In the Atkinson model, as the hypothetical well moved farther from the river, the depletion rate decreased. In contrast, the depletion rate in the Winslow model didn’t decrease dramatically with the increasing distance between the pumping well and river (Fig. 27).

Fig. 27 Stream depletion curves of the hypothetical wells with longer distance to the river (from 0.62 mi to 3.72 mi) in the Atkinson and Winslow models

For the second phase of simulation, the pumping schedule of the hypothetical wells was changed to June to August, which represented the realistic irrigation season. The simulation result for well HW1 in the Atkinson model is shown in Fig. 28 as an example. The stream depletion rate for all-year pumping was greater than seasonal pumping. The calculated stream depletion rate reached the highest during the summer time and then declined to a low depletion rate in non-irrigated season. Although the all-year pumping seemed useless in reality, it was decisive for the local interpreted water
resources management. The extent of an aquifer that is considered to be hydrologically connected is the so-called “10/50” line by the interpreted management process (Department of Natural Resources, 2008). The preliminary consideration was the area within which pumping of a well for 50 years will deplete the river or a base flow tributary thereof by at least 10% of the amount pumped in that time. The all-year pumping simulation with varied distance to the river should be helpful in determining the 10/50 line in the Elkhorn River Basin.

Fig. 28 Comparison of stream depletion rate caused by all-year pumping and season pumping

For the third phase of stream depletion analysis, the well screen was moved to the shallow layer of the groundwater system, which belonged to the alluvial aquifers. Figure 29 and 30 show the simulation curve of the sample wells from the Atkinson and Winslow models, respectively. As the figures indicated that the alluvial aquifer pumping would
cause slightly higher stream depletion rate in the Atkinson model and no observable
difference in the Winslow model simulation. It could be leading by the fact that alluvial
sediments in the Atkinson model consisted of coarse sand and gravel while the Winslow
model upper layer was topped by a thick layer of silt/clay mixture. The coarse-grained
deposits in the Atkinson alluvial aquifer facilitated the water interaction between the
Elkhorn River and the shallow aquifers and resulted in greater stream depletion rate.

Fig. 29 Stream depletion rate differences for wells pumping from the alluvial and
Ogallala aquifers in the Atkinson model
Fig. 30 Stream depletion rate differences for wells pumping from the alluvial and deep aquifers in the Winslow model
CHAPTER 5 DISCUSSION AND CONCLUSIONS

5.1 Streamflow trend tests

No significant decreasing trend of the annual streamflow from 18 gauging stations was found using the Mann-Kendall test. This was consistent with the long-term groundwater level data which was relatively stable. Streamflow time series in three of the eighteen gauging stations showed upward trends. However, the stream depletion caused by agricultural pumping may still exist even though no decreasing trends were found from the Mann-Kendall test results. The annual streamflow rate was affected by several hydro-climatologic factors. Also, the annual flow rate trend test is not able to reveal the seasonal variation of streamflow. The intensive pumping in summer season may induce significant streamflow reduction. The fluctuation of the streamflow within one year was observable, with an increasing tendency from January to June and a decreasing trend from June to October. Figure 31 shows the averaged streamflow rate chart for each month of the seven gauging stations in the Elkhorn River.
5.2 Field study in the Norfolk and Winslow site

The hydraulic conductivity ($K$) values for shallow streambed sediments of the Elkhorn River were generally higher at the Norfolk site than at the Winslow site. $K$ values in vertical direction ranged from 10 to 101 ft/d at the Norfolk site. In contrast, they varied from 0.7 to 22 ft/d at the Winslow site. For the Norfolk site, the averaged $K_v$ value for the ten point measurements from the top layer of the streambed was 39 ft/d and was almost four times larger than the averaged $K_v$ value of the Winslow site (8 ft/d).

The $K_v$ profiles of the subsurface streambed sediments were obtained through the in-lab permeameter tests on the sediment cores. Electrical conductivity (EC) logs were also gained by means of Geoprobe® technique. The EC logs suggested that the bedrock occurred around 40 ft under the channel surface at the Norfolk site and the deposits above the bedrock were composed mostly of sands and gravels. At the Winslow site, a thick
layer of low permeable sediments was observed at the depth of 12 ft to 25 ft, with bedded traces of fine sands.

5.3 Regional groundwater flow model development

Two groundwater flow models were developed near Atkinson and Winslow, one in the upstream area and the other in the downstream area of the Elkhorn River. The hydrogeological structure of the aquifer system was studied using the lithological records of test holes and registered wells. The aquifer system in each model was divided into five hydrostratigraphic units. Time-dependent and spatially varied recharge, stream networks and evapotranspiration were incorporated in the models for monthly simulation. Because of lack of groundwater use records, groundwater pumpage was estimated by crop irrigation requirement and summer precipitation. The numerical models were calibrated to the groundwater level records of 55 observation wells using trial and error approach, 33 wells in the Atkinson model and 22 wells in the Winslow model. The water budget output of the models indicated that the groundwater storage amount remained steady according to our pumping simulation.

5.4 Streamflow depletion analyses using the numerical models

The calibrated numerical models were utilized for calculation of streamflow depletion with changed pumping scenarios. Hypothetical wells were placed in the model at varied locations, with different well screen depths and pumping time. Stream depletion curves were plotted for each well using the results of the water budget exported from model simulations. Although only one streambed conductance value was utilized in each
model, the stream depletion rates caused by spatially allocated wells with the same distance to the river were not same, which could result from the anisotropy and heterogeneity of the aquifer system. In the Winslow model, the stream depletion rate increased when the hypothetical pumping well was moved to the east, which may be associated to the accumulative impact of Maple Creek and Logan Creek leakage as well. Meanwhile, as the distance between the pumping well and river became smaller, the stream depletion rate increased dramatically in the Atkinson model but barely changed in the Winslow model. The thick layer of the low-permeability sediments that existed on the top of the Winslow model area could reduce the connection between the rivers and aquifers, resulting in relatively small stream depletion rate.

Although the agricultural pumping time concentrated in the summer season, two different pumping schedules were used for the simulation. One was all year pumping and the other one was seasonal pumping from June to August. The seasonal pumping output showed that the stream depletion rate within one year reached to approximately 20% in the growing season and reduced to 5% in the winter time. Nevertheless, all year pumping resulted in continuously increased stream depletion rate for a certain period and came to a near stable value at the end of the simulation period. While the all year pumping simulation seemed unrealistic, it could aid the local water management authorities to depict the “10/50” line for individual well regulation.

The impact of well screen depth on the stream depletion was also analyzed by moving the well screen depth from the alluvial layer to the Ogallala or Dakota Group. No difference in stream depletion rate was found for the Winslow model simulation. For the
Atkinson model, when the well screen was located at the alluvial sediments, the stream depletion was slightly higher than that for the deep screen wells. The reason for this result could be that the alluvial sediments were fine-grained sands and gravels compared to the Ogallala Group sediments, which inclined to greater hydrological connection between the river and aquifer.
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APPENDIX A: PLOT OF OBSERVATION POINTS WITH THE MODELED HYDRAULIC HEAD CURVE
APPENDIX B: PLOT OF OBSERVATION POINTS WITH THE MODELED HYDRAULIC HEAD CURVE