Characterization of the Stream-Aquifer Hydrologic Connection in the Elkhorn River Basin

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CHARACTERIZATION OF THE STREAM-AQUIFER HYDROLOGIC
CONNECTION IN THE ELKHORN RIVER BASIN

by

Zhaowei Wang

A THESIS
Presented to the Faculty of
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In this study, the hydrologic relationship between the Elkhorn River and the surrounding aquifer at eight study sites were analyzed. Firstly, Geoprobe log data and in-situ permeameter test data were combined to calculate the equivalent hydraulic conductivity \( K \) and unit-length streambed conductance \( C \) at eight sites, where the greatest values of \( K \) and \( C \) were found at Neligh and the lowest values were found at Hadar. Secondly, the calculated hydraulic gradient showed that the adjacent aquifer provided baseflow to the Elkhorn River all year. Reversed gradient was only found at Atkinson site during pumping seasons from 2008 to 2010. Thirdly, the result of cross correlation analysis on the lag effect between stream stage and groundwater level was consistent with the indication of \( K \) and \( C \).

In addition, a geological model of multimillion grids was built based on well log data of test holes and registered wells using IDW interpolation method. The grids were grouped into eight hydrofacies and their corresponding hydraulic conductivity values were assigned based on empirical values from books and former studies in this area. Then the geological model was upscaled to form a hydrostratigraphic model of three aquifer units by coarser grid using an averaging technique (Li et al., 1999) and bound method (Cardwell, 1945). The hydrostratigraphic model showed the low \( K \) surface at northeast
bound and southwest bound of model area. The aquifer is thicker in west and becomes thinner while it extends to the east region.

Finally, a groundwater flow model (ULEN) was built using MODFLOW to calculate the stream depletion ratio at the Neligh and the Hadar site based on the hydrostratigraphic model. A hypothetical well was created and pumped water at a rate of 1000 GPM in June, July and August under three scenarios: the well is located from the river at 1000 ft (304.8 m), 1 mile (1.61 km) and 3 miles (4.83 km), respectively. The modeling results were consistent with the data analysis which showed that the stream depletion ratio was more pronounced at the Neligh site than the Hadar site.
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Chapter 1 Introduction

1.1 Objective and Scope

Irrigated agriculture is vital to the livelihood of communities in the Elkhorn River Basin of Nebraska, and groundwater is used to irrigate most of the cropland (Peterson, 2008). A large number of irrigation wells have been constructed in the High Plains Aquifer and the glacial deposits in this area. Potential reductions of groundwater storage and streamflow depletion are important issues for water resources management in the basin. For better management and usage of the precious surface water and groundwater resources, the hydrologic connection between the Elkhorn River (including its tributaries) and adjacent aquifer must be studied and determined.

The objective of this research is to quantify the levels of the hydrologic connectedness of the river, streambed, and aquifer which will be accomplished by finishing the following three sub-objectives:

A. To determine the hydrologic connectedness by analyzing the groundwater-level data and stream stage data under different scenarios. The research activities include the following aspects.

1) Determination of the streambed conductance,
2) Calculation of the hydraulic gradient between the river and the aquifer,
3) Evaluation of the effects of groundwater irrigation on stream stages,
4) Evaluation of the response of groundwater levels to flooding events,
5) Determination of the hydraulic connectedness between shallow alluvial aquifer and the deep Ogallala Aquifer.
B. To construct a hydrostratigraphic model and analyze the spatial variation of the High Plains Aquifer in selected areas within the basin.

C. To develop a regional groundwater flow model (ULEN) for evaluation of the potential impact of groundwater pumping on streamflow depletion

1.2 Background

1.2.1 Previous studies

U. S. Geological Survey (Newport, 1957) generated a report to summarize existing data that pertains to the ground-water resources in the Elkhorn River Basin. The report includes a brief description of geology and topography, a summary of the chemical analyses of 29 samples of groundwater and 14 samples of surface water, and a discussion of the occurrence and quality of the water. Hereinafter, since the mid 1990’s, the Upper and Lower Elkhorn Natural Resources Districts (NRDs), University of Nebraska Conservation and Survey Division (CSD) and Nebraska Department of Natural Resources (NDNR) have conducted a number of research projects in the Basin (Shafer, 2006; Chen et al., 2009; Chen and Lackey, 2010; Lackey and Chen, 2010); a large quantity of data have been collected. These data sets include streambed electrical conductivity logs and hydraulic conductivity measurements; groundwater level and water quality measurements from a multiple-level groundwater monitoring well network constructed near the bank of the Elkhorn River and its tributaries; stream discharge and river stage data collected from USGS and DNR gage stations. Their studies provided valuable information of the streambed hydraulic conductivity which are important in analyzing water quantity exchange between a stream and its sediments and are also
valuable in determining the streambed conductance. Data from many of these were used in this project.

Other studies have been conducted by the USGS Stanton et al. (2010) who recently developed a regional groundwater model to simulate the effect of groundwater pumping on the stream base flow depletion which covered the Loup River Basin and part of the Elkhorn River Basin. This report suggested that the simulated pumpage within the Lower Elkhorn NRD should be reduced by about 49% and pumpage within the Upper Elkhorn NRD should be reduced by about 17% in order to meet an in-stream flow criterion for the Platte River (A–17331 criterion) for 70% of the time. This model provided a good insight for the stream-aquifer interaction in the Elkhorn River Basin in an overall prospective. However, due to the data limitation and the size of the study domain, this model is limited in its ability to provide more precise information on the connectedness between the Elkhorn River and adjacent aquifers at eight study sites that were interested in this project.

Besides the physically based research approach, GIS has also been incorporated to assess the groundwater pollution risk in the Elkhorn River Basin (Li, 2012). He modeled the groundwater vulnerability based on the geology, climate and soil data. This information will be referenced in this study for the evaluation of the groundwater discharge component in the ULEN model.

1.2.2 Current interest in the Elkhorn River Basin

According to Nebraska’s law, a river basin, sub basin, or reach will be deemed fully appropriated if the Department of Natural Resources determines that then-current uses of hydrologically connected surface water and groundwater in the river basin cause
or will cause the surface water supply to be insufficient to meet the agricultural use (Department of Natural Resources, 2005). For purposes of determining whether or not a river basin is fully appropriated, the Nebraska Department of Natural Resources considers thin wells within the 10% / 50-year boundary are hydrologically connected to the river. The hydrologically connected area is defined as a geographic area within which pumping a well for 50 years will deplete the river by at least 10% of the amount pumped in that time which is also called 10% /50-year area. In the Elkhorn River Basin, there is no sufficient numeric groundwater model available to determine the 10/50 area. The 10/50 area was determined using the Jenkins methodology (Shafer, 2006) based on transmissivity maps, specific yield maps, stream reaches and GIS raster points data.

In this study, a hydrostratigraphic model will be built and will be presented in Chapter 3 as skeleton for the construction of a regional ground model that covers part of Upper and Lower Elkhorn Natural Resource District (ULEN) presented in Chapter 4. A hypothetical well will be placed near the representative study sites at Neligh and Hadar and the amount of stream depletion will be calculated under three scenarios when the well keeps pumping over the next 30 years.

1.3 Site Description

1.3.1 Location and surface hydrology

The study area is within the Elkhorn River drainage basin, located in northeastern Nebraska and includes Cuming, Stanton, and Wayne counties and portions of Antelope, Boone, Brown, Burt, Cedar, Colfax, Dakota, Dixon, Dodge, Douglas, Garfield, Holt, Knox, Madison, Pierce, Platte, Rock, Sarpy, Thurston, Washington, and Wheeler counties (Figure 1-1).
The area of the Elkhorn River Basin is approximately 7,000 square miles (18,000 square kilometers). The major stream in the Basin is the Elkhorn River. This river, originating in the northeastern Sand Hills of Nebraska, flowing 466.71 km, is located in northeast and north-central Nebraska and flows in a southeasterly direction until confluencing with the Platte River near Gretna, Nebraska. Major tributaries to the Elkhorn River include South Fork Elkhorn River, North Fork Elkhorn River, Logan Creek and Maple Creek.

![Figure 1-1 The Elkhorn River Basin and its location in Nebraska.](image)

### 1.3.1 Climate

The climate of the basin is continental with a rather wide range in temperature between winter and summer; generally, it is well suited to raising livestock and growing of feed and grain crops. The spring months are cool and have considerable rain; the
summer months are warm and have moderate precipitation; the autumn months are pleasant with only occasional rains; and the winter months are characterized by frequent low temperature that are usually accompanied by snow. The range in topographic relief is insufficient to cause appreciable climatic differences from place to place (Newport, 1957). The average annual precipitation from 1949 to 2003 ranges from 23.4 inches (594.36 mm) at O’Neill in the northwestern portion of the Basin to 29.9 inches (759.46 mm) at Fremont in the southeastern corner of the Basin. The average growing season ranges from 11.9 (302.26 mm) inches at O’Neill to 15.2 inches (386.08 mm) at Fremont (Shafer, 2006).

1.3.2 Hydrogeology

The surficial Quaternary and underlain Tertiary deposits are the focus of this research. The topography in the western part of the basin is in the Sand Hills region of Nebraska, whereas the eastern part of the basin is within the loess plains region of the Great Plains province. The hydrogeology of the basin is complex due to the wide range of depositional environments, from eolian sands in the west to glacial sediments in the east (Diffendal and Voorhies, 1994). Nearly 40% of the basin has glacial deposits. The surface geology is unconsolidated sedimentary deposits underlain by Pleistocene fluvial silt, sand, gravel and clay deposits. The sediments of Tertiary age consist of thin, interfingerling lenses of gravel, sand, silt, and clay, moderately cemented in some places, and the Cretaceous formations consist of alternating layers of sandstone, limestone, and shale.

The principal aquifer unit in the basin is well known as the High Plains Regional Aquifer which mainly consists of Quaternary alluvial along with shallow aquifers and the deep aquifers of the Tertiary Ogallala Group. The bedrock aquifers range in age from
Tertiary to Cretaceous and supply a small amount of water compared to the other aquifers. The saturated thickness of the principal aquifer unit ranges from 0 to approximately 770 ft (234.70 m). Groundwater tends to move from the upland to the river, and streamflow in the basin is driven by groundwater discharge as baseflow and by precipitation runoff (Chen, 2010).
Chapter 2 Time Series Data Analysis

2.1 Objective

There are eight study sites in this project. Six sites along the Elkhorn River are named: Atkinson, Ewing, Neligh, Meadow Grove, Norfolk and Winslow (Figure 2-1). The remaining two are Pierce along Willow Creek and Hadar near the North Fork Elkhorn River. The research objectives are described as following:

1. Determination of the streambed conductance: Streambed conductance is a key parameter in simulating streamflow depletion. In the Elkhorn River Basin, the EC logs and sediment cores from the Elkhorn River and tributaries was collected to calculate the streambed conductance. Previous data from experiment and calculation do not uniformly show a low permeability at the channel surface.

2. Calculation of the hydraulic gradient between the river and the aquifer: Streamflow depletion is calculated based on Darcy’s law such that $Q = I \times Ce$ where $Q$ is the water infiltrated from the river to the aquifer if the river stage is higher than the elevation of hydraulic head in the aquifer, $I$ is the hydraulic gradient between the river and the aquifer, and $Ce$ is the equivalent streambed conductance.

3. Evaluation of the impact of groundwater irrigation on stream gauges: During each irrigation season, irrigation wells pump a large quantity of groundwater, resulting in groundwater level declines in the aquifer along both sides of the river. The relationship between groundwater level and the stream stage will be determined to see whether there is a pattern in stream stage during each irrigation season.

4. Analysis of the response of groundwater level to flooding events: During each flooding event, the stream stage rises and it may be higher than the groundwater levels. During
these flooding events, the river recharges the aquifer system. The magnitude and the time lag of the response of groundwater level in the aquifer will reflect the level of hydrologic connectedness between the river and the aquifer.

5. Characterization of the hydrologic relations between the shallow aquifer and the deep Ogallala Aquifer: In the top of the Ogallala Aquifer, there commonly exists a silt and clay layer in this study area. This layer can reduce the hydrologic connectedness between the deep Ogallala Aquifer and the shallow alluvial aquifer. Analysis of the relationship between the two aquifers will be done and it is expected that the hydrologic relations between the shallow aquifer and the deep Ogallala Aquifer will vary from one monitoring site to another.

2.2 Available Datasets

2.2.1 Streambed tests

Streambed tests have been conducted at 14 study sites along the Elkhorn River and its tributaries (Chen and Lackey, 2010; Lackey and Chen, 2010). At each testing site, in-situ permeameter tests were conducted to determine the vertical hydraulic conductivity of the upper part of the streambed. The Geoprobe direct-push techniques were used to log electrical conductivity of channel sediments. The EC logs at Atkinson site (Figure 2-1) clearly show the contrast values of the electrical conductivity for silt/clay and sand/gravel to the depth of 50 ft (15.24 m) (15.24 m). The low value of Electrical Conductivity in the figure indicates sand/gravel deposits and large value of Electrical Conductivity in the figure represents silt/clay deposits (Schulmeister et al., 2003). Additionally, the cores of channel sediments were collected by the Geoprobe and tests were made on these cores to provide the vertical hydraulic conductivity of channel sediments at depth. The electrical
conductivity values of three graphs in Figure 2-1 all show that there is a low $K$ layer displayed at 20 ft (m) (6.01 m) under the stream channel surface which is consistent with the $K_v$ results from falling head method (Chen, 2000; Butler et al., 2007; Lu et al., 2011). Chen (2008) contrasted the ranges of $K_v$ determined using permeameter tests (60 $K_v$ values) of sediment cores from the Chapman, Wood River, MSEA, and Killgore Island sites and using pumping tests (65 $K_v$ values). The $K_v$ results that generated from both tests were statistically distributed in the same range. This suggests that the $K_v$ values of the sediments cores collected from Geoprobe is reliable to represent the vertical hydraulic conductivity of streambed in this study.

![Figure 2-1 Comparison of EC logs and corresponding hydraulic conductivity values at Atkinson site (Data source: (Chen and Lackey, 2010)).](image)

**2.2.2 Groundwater level data**

In 1997 the Conservation and Survey Division began the design of monitoring well networks for the Upper Elkhorn and the Lower Elkhorn NRDs as part of cooperative agreements with each district. Well sites were selected near stream gages so that a future assessment of the interaction between the surface and groundwater resources could be
made. The pressure transducers were set to record groundwater levels every eight hours. Sites that include at least one well in the Quaternary aquifer and at least one well in the Ogallala Aquifer include Stuart, Atkinson, Neligh, Meadow Grove, Willow Creek, and Hadar. The water level data were provided by Sue Lackey, the research hydrogeologist of UNL.

2.2.3 Stream stage data

Among the eight study sites, six stream gages are available and are operated by the NDNR and the other two are operated by USGS at Hadar and Norfolk. The streamflow discharge rates and stream stages can be obtained for the analysis. At each groundwater monitoring network site, there is a nearby stream gage that records streamflow discharge and stream stage. Figure 2-2 shows the locations of the groundwater monitoring sites and stream stage sites.
2.3 Methods

2.3.1 Determination of the streambed conductance

Streambed conductance is a lumped mathematical term that describes the transmittal capability of stream water to the channel sediments when the level of stream stage is higher than the elevation of groundwater under the river (Equation 2-1).

\[
C_e = \frac{K_{equ} \cdot L \cdot W}{M} \quad \text{(Eq. 2-1)}
\]

where \( C_e \) is the streambed conductance \((L^2/T)\), \( M \) \((L)\) is the thickness of the low-permeability streambed, plus the thickness of the sediment above this low permeability streambed, \( L \) and \( W \) are the length and width of submerged channel, \( K_{equ} \) \((L/T)\) is the...
equivalent vertical hydraulic conductivity of the streambed which can be calculated by:

\[ K_{equ} = \left( \frac{L_1 + L_2 + L_3 + \cdots + L_N}{K_{v1} + L_2 K_{v2} + \cdots + L_N K_{vN}} \right) \]  
(Eq. 2-2)

where \( L_1 \) is the sediment length in the sample collected from the ground surface to 5 ft (1.52 m) deep in the streambed, \( K_{v1} \) is the vertical hydraulic conductivity of \( L_1 \) sediment. The rest can be done in the same manner until to \( L_N \) of the sediment length in the sample taken out from the \( Nth \) layer 5 ft (1.52 m) deep in the ground.

2.3.2 Calculation of the hydraulic gradient between the river and the aquifer

Groundwater levels from the monitoring network and the stream stage records from the gage stations will be used to determine the daily hydraulic gradient between the river and the aquifer. The direction of hydraulic gradient (upward or downward) between the river and the aquifer will indicate whether the river gains water from or loses water to the aquifer. The hydraulic gradient patterns at each monitoring site are expected to differ and to vary between irrigation and non-irrigation seasons. The gradient was calculated by the equation (Equation 2-3)

\[ I = \frac{(H_{stream} - H_{gw})}{L} \]  
(Eq. 2-3)

where \( I \) \((L/L)\) represents the hydraulic gradient, \( H_{stream} \) \((L)\) is the stream stage elevation and \( H_{gw} \) \((L)\) the elevation of the groundwater, \( L \) stands for the perpendicular distance from groundwater monitoring point to the stream. \( I < 0 \) indicates a gaining river that loses water to the aquifer, while \( I > 0 \) indicates a losing river that gains water from aquifer.

Since the stream stage does not always locate near the groundwater monitoring site, for the purpose of accuracy, the stream stage level will be linearly interpolated to the point which has the shortest distance from the river to the location of the groundwater.
monitoring well. It is because the stage level of the five gages along the Elkhorn River has an approximately linear relationship (Figure 4-9).

2.3.3 Stream-aquifer connection analysis

In the evaluation of stream-aquifer connectivity there is a need to develop simple tools using readily available datasets (Brodie et al., 2008) and several available tools have been discussed by Brodie et al. (2007b). Lag effect in a time series is an important parameter to evaluate the level of connectedness between surface water and groundwater. This phenomenon was observed and discussed by Hvorslev (1951). Kelly (2001) used a regression analysis to discuss the lag effect during the stream-aquifer interactions. The cross-correlation was introduced into the time series data analysis by Bloomfield (1976). After that, this method has been widely applied to various subjects such as climate, hydrology, geology and others. Changnon (1988) studied the relations between groundwater level and precipitation using cross correlation analysis. Kuo et al. (1990) explored the coherence between atmospheric carbon dioxide and global temperature. Brodie et al. (2007a) developed Q-lag method to understanding stream-aquifer connectivity. Ghanbari and Bravo (2011) applied the spectral analysis that was introduced from electrical engineering to evaluate the correlations between groundwater level fluctuations and lake level fluctuations. In this study, the cross-correlation analysis was chosen to evaluate the connectedness between surface water and groundwater.

Data Processing

The wells were constructed at eight sites and the groundwater level was monitored at different depths in these wells. Normally the shallow well was screened in the Quaternary aquifer and the middle and deep wells monitored water levels in the
Ogallala Aquifer. Pressure transducers were installed to measure groundwater levels every 8 hours. The three records in each day were then averaged to one daily value for time series analysis. Groundwater level elevation was calculated using surface elevation minus the depth to groundwater table. The streamflow data were acquired from DNR and USGS gage stations and was also averaged to one value per day.

Cross-correlation analysis

Taken the description from Nagpaul (2005), cross-correlation is a measure of similarity between two time series. Mathematically speaking, it is the linear correlation coefficient between two time series as a function of time lag between the two series. Consider $N$ pairs of observations on two time series $x_t$ and $y_t$. Cross-covariance function is given by Equation 2-4:

$$
\Gamma_{xt}(k) = \frac{1}{N} \sum_{t=1}^{N-k} (x_t - x_{mean})(y_{t+k} - y_{mean}) \text{ for } k = 1, 2, 3, \ldots, N-1
$$

and

$$
\Gamma_{yt}(k) = \frac{1}{N} \sum_{t=1}^{N} (x_t - x_{mean})(y_{t+k} - y_{mean}) \text{ for } k = -1, -2, -3, \ldots, -(N-1)
$$

(Eq. 2-4)

where

$r_{xt}(k)$ is the cross-covariance when $y_t$ lags $x_t$

$r_{yt}(k)$ is the cross-covariance when $x_t$ lags $y_t$

$N$ is the series length

$x$ and $y$ are the sample means

$k$ is the lag.
Cross-correlation is the cross-covariance scaled by the variances of the two series (Equation 2-5)

\[ \Gamma_{xy}(k) = \frac{\Gamma_{xy}(k)}{\sqrt{\Gamma_{xx}(0)\Gamma_{yy}(0)}} \]  

(Eq. 2-5)

where \( \Gamma_{xx}(0) \) and \( \Gamma_{yy}(0) \) are the sample variances of series \( x \) and \( y \). Thus, this cross correlation analysis would calculate the largest correlation coefficient between groundwater level and stream stage at lag \( k \) days.

2.4 Results

2.4.1 Determination of the streambed conductance

The values of \( K_v \), \( M \), and \( W \) can be found from the two reports (Chen and Lackey, 2010; Lackey and Chen, 2010) that were submitted to the Upper and Lower Elkhorn Natural Resources districts. The two reports summarized the streambed test results. The streambed conductance at each site was listed in Table 2-1

### Table 2-1 Calculated streambed conductance at seven study sites.

<table>
<thead>
<tr>
<th>Site Name</th>
<th>Streambed Length (ft)</th>
<th>Stream Channel Width (ft)</th>
<th>Streambed Thickness (ft)</th>
<th>Equivalent ( K_v ) (ft/d)</th>
<th>Streambed Conductance (ft²/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atkinson</td>
<td>1</td>
<td>26.73</td>
<td>5</td>
<td>0.1120</td>
<td>0.599</td>
</tr>
<tr>
<td>Ewing</td>
<td>1</td>
<td>40.50</td>
<td>20</td>
<td>0.0117</td>
<td>0.024</td>
</tr>
<tr>
<td>Neligh</td>
<td>1</td>
<td>116.35</td>
<td>35</td>
<td>0.2441</td>
<td>0.811</td>
</tr>
<tr>
<td>Meadow Grove</td>
<td>1</td>
<td>178.00</td>
<td>10</td>
<td>0.0128</td>
<td>0.228</td>
</tr>
<tr>
<td>Norfolk</td>
<td>1</td>
<td>105.00</td>
<td>45</td>
<td>0.0203</td>
<td>0.047</td>
</tr>
<tr>
<td>Hadar</td>
<td>1</td>
<td>41.31</td>
<td>25</td>
<td>0.0052</td>
<td>0.009</td>
</tr>
<tr>
<td>Pierce</td>
<td>1</td>
<td>15.80</td>
<td>20</td>
<td>0.1143</td>
<td>0.090</td>
</tr>
</tbody>
</table>

2.4.2 Calculation of the hydraulic gradient between the river and the aquifer.

**SITE Atkinson**

This site is located in the upstream of the Elkhorn River in Holt County. Figure 2-3 shows that the Elkhorn River flowing by this site is a gaining river. The red line
represented the gradient which was below the zero gradient line (in green). The values ranged from -0.02 to 0.02. However, the gradient values were smaller than 0 for 98% of the time. It means the groundwater level was higher than the stream stage and the river was gaining water from the aquifer. There were several spikes where $I > 0$ and the most pronounced one happened in June 2010. The reason was that an obvious rain event (the blue line indicates precipitation) occurred just before this spike, which reversed the gradient between the stream and aquifer by raising the stream stage elevation. A few days after, without external hydrologic stress, the stream level recessed back to the normal level and the groundwater continued discharging to the stream as baseflow.

It is also observed that the hydraulic gradient was in response to the groundwater hydrograph. The green line represented the groundwater level in well 3-S. During the pumping season, the groundwater level dropped and then gradually recovered. The hydraulic gradient also became closer to zero and then increased back to the normal value. As we can see from the graph, the gradient became closer to 0 when the groundwater level dropped during the irrigation season.
SITE Ewing

The Ewing site is located about 35 miles (56.33 km) southeast of the Atkinson site. The stream stage is located downstream on the South Fork of the Elkhorn River, close to the confluence point with the Elkhorn River. The South Fork of the Elkhorn River at the Ewing site is a gaining river. From Figure 2-4 we can clearly see that the gradient in blue line was smaller than 0 from 2008 to 2010 every day and the gradient values ranged from -0.0016 to 0.0004. It was suggested that the river was gaining water from the adjacent aquifer all year.

The hydraulic gradient was in response to the groundwater hydrograph. The black line represented the groundwater level in well 16-S. During the pumping season, the
groundwater level dropped and then gradually recovered. The hydraulic gradient also decreased in absolute value to respond to the irrigation pumping period.

![Time series data of groundwater level in well 16S, hydraulic gradient and stream stage at Ewing.](image)

**SITE Neligh**

This site is located in Antelope County in the Upper Elkhorn NRD, about 18.5 miles (29.77 km) southeast of the Ewing site along the Elkhorn River. Figure 2-5 indicates that the Elkhorn River at this site is a gaining river. The blue line in the graph represented the gradient. The values ranged from -0.005 to 0 which were all smaller than 0 from 2007 to 2010, meaning that the groundwater level was always higher than the stream stage and the river was gaining water from the aquifer.

The hydraulic gradient was in response to the groundwater hydrograph. The black line represented the groundwater level in well 17-M. During the pumping season, the groundwater level dropped and the absolute value of hydraulic gradient then dropped
quickly since the groundwater level became lower. There were also several spikes of gradient during the non-pumping season. These were due to precipitation events that resulted in the lifting of stream stage.

Figure 2-5 Time series data of groundwater level in well 17M, hydraulic gradient and stream stage at Neligh.

SITE Meadow Grove

This site is located at Madison County in the Lower Elkhorn NRD, about 16 miles (25.75 km) downstream of the Neligh site along the Elkhorn River. Figure 2-6 indicates that the Elkhorn River at this site is a gaining river. The gradient (in blue line) ranged from -0.001 to -0.004 which were all smaller than 0 from 2007 to 2010. This means that the groundwater level was always higher than the stream stage and the river was gaining water from the aquifer. The stream stage values used in the gradient calculation were linearly interpolated between gage station 06798500 and gage station 06799000.
The hydraulic gradient was in response to the groundwater hydrograph. The black line represented the groundwater level in well 12-S. During the pumping season, the groundwater level dropped and the absolute value of hydraulic gradient then dropped quickly since head difference between stream stage and groundwater level became smaller. There were also several spikes of gradient during the non-pumping season. These were due to the precipitation event that resulted in the rise of the stream stage.

![Figure 2-6 Time series data of groundwater level in well 12S, hydraulic gradient and stream stage at Meadow Grove.](image)

**SITE Norfolk**

This site is located in the middle portion of the Elkhorn River in Madison County. Figure 2-7 shows that the Elkhorn River at this site is a gaining river. The blue line in the graph represented the gradient, which was almost below the zero gradient line (the green line). The values ranged from -0.004 to 0.0003. However, in 98% of time, the gradient
values were smaller than 0. It means that the groundwater level was higher than the stream stage and the river was gaining water from the aquifer. There were several spikes where \( I > 0 \). The first reason was attributed to rain events and the second reason was the heavy pumping activity from June to September, both factors leading to the reversion of gradient between the stream and aquifer and turned the stream to a losing river. Once the pumping stopped, the groundwater level returned to the normal level and kept providing baseflow to the river.

The hydraulic gradient was in response to the groundwater hydrograph. The red line represented the groundwater level in well 16-M. During the pumping season, the groundwater level dropped and then gradually recovered. The hydraulic gradient also got close to zero and then increased back to the normal value. As we can see from this figure, the gradient became closer to 0 when the groundwater level dropped during the irrigation season.
FIGURE 2-7 Time series data of groundwater level in well 16M, hydraulic gradient and stream stage at Norfolk.

SITE Winslow

This site is located at Dodge County in the Lower Elkhorn NRD, about 35 miles (56.33 km) upstream to the outlet point of the Elkhorn River Basin. Data at this site are only available from July 2010 to November 2011. Figure 2-8 indicates that the Elkhorn River at this site is a gaining river. The blue line in the graph represented the gradient and the value ranged from -0.001 to -0.005 which were all smaller than 0 during these 4 months in 2010. It showed that the hydraulic head in the aquifer was always higher than the stream stage and the river was gaining water from the aquifer.
SITE Hadar

This site is located in the middle portion of the North Fork Elkhorn River which is a tributary of the Elkhorn River in Pierce County. Figure 2-9 shows that the North Fork Elkhorn River flow at this site is a gaining river. The blue line in the graph represented the gradient. The value ranged from -0.06 to 0.04. However, 91% of the measured period, the gradient values were smaller than 0. This means that the groundwater level was higher than the stream stage and the river was gaining water from the aquifer. There were several spikes where $I > 0$. The rain event and the heavy pumping from June to September were two main reasons that contributed to the reversion of gradient between

Figure 2-8 Time series data of groundwater level in well 36S, hydraulic gradient and stream stage at Winslow.
the stream and aquifer and turned the stream to a losing river. Once the pumping stopped, the groundwater level recovered back to the normal level and provided baseflow to the river.

The hydraulic gradient was in response to the groundwater hydrograph. The red line represented the groundwater level in well 04-S. During the pumping season, the groundwater level dropped and then gradually recovered. The hydraulic gradient also got close to zero and then increased back to the normal value. As we can see from the graph, the gradient became closer to 0 when the groundwater level dropped during the irrigation season.

Figure 2-9 Time series data of groundwater level in well 04S, hydraulic gradient and stream stage at Hadar.
SITE Pierce

This site is located in Pierce County. The river is called Willow Creek which is a tributary to the North Fork Elkhorn River. Figure 2-10 indicates that the Willow Creek flow by this site is a gaining river. The stream stage used in the gradient calculation was interpolated from the stream level at gage 06799080 near Foster and gage 06799100 near Hadar. The blue line in the graph represented the gradient and the value ranged from -0.001 to -0.004, all smaller than 0 from 2004 to 2010. This showed that the groundwater level was always higher than the stream stage and the river was gaining water from aquifer.

The hydraulic gradient was in response to the groundwater hydrograph. The black line represented the groundwater level in well 06-S. During the pumping season, the groundwater level dropped and the absolute value of hydraulic gradient then dropped quickly after the head difference between stream stage and groundwater got smaller. There were also several spikes of gradient during the non-pumping season. These were due to the precipitation events that caused a rise of stream stage.
2.4.3 Stream-aquifer interactions

SITE Atkinson

The Atkinson streambed test site was about 4.07 miles (6.55 km) southeast of the DNR stream gage station (Figure 2-11). The test hole, 1-UE-99 (Figure 2-12), was drilled to 60 ft (18.29 m) below the land surface. The upper 20 ft (6.10 m) of sediments consisted of unconsolidated Quaternary sand. Below the Quaternary sediments was the Tertiary Ogallala Group from the depth of 40 to 60 ft (12.19 ~ 18.29 m). Another test hole, 2-UE-99 (Figure 2-13) was drilled to a depth of 400 ft (121.92 m). According to the log of the test hole, the upper 30 ft (9.14 m) consisted of the undifferentiated Quaternary sand with some gravel at the base. The Tertiary Ogallala Group was found at depths of 30 ~ 400 ft (9.14 ~ 121.92 m). A groundwater monitoring well 2S was constructed about 468 ft (142.65 m) north of the Elkhorn River and another two groundwater monitoring
wells were located about 300 ft (91.44 m) south of the Elkhorn River includes which were well 3S (shallow well) and 3D (deep wells) that were screened at different depths of 26 ft (7.92 m) and 221 ft (67.36 m), respectively.

Figure 2-11 Plan view of monitoring wells, gage station, streambed test and test holes location at Atkinson site.
Figure 2-12 Lithology and stratigraphy of test hole 1-UE-99.

Figure 2-13 Lithology and stratigraphy of test hole 2-UE-99.
The Elkhorn River was gaining water from the aquifer at this site. The graph (Figure 2-14) showed that the stream level was in response to the groundwater level variation at both shallow wells. The screened depths of both wells were at the Quaternary aquifer which consisted mostly of sand-sized sediments. It was likely that there was a relatively strong degree of connection between the aquifer and the river.

![Figure 2-14 Comparison of stream stage and groundwater level screened at shallow and deep depth.](image)

The data also (Figure 2-15) indicated that the groundwater levels in the shallow well 3S and deep well 3D had a similar response pattern during both irrigation and non-irrigation seasons. During the non-irrigation season, the water level was higher in the deep well than in the shallow well. The water level difference between the two wells was often greater than 2 ~ 3 ft (0.61 ~ 0.91 m). This hydraulic head difference suggested that groundwater from the Ogallala Aquifer was moving upward and into the lower shallow
Quaternary aquifer during the non-irrigation season. In contrary, during the pumping season, the water was reversely moving downward.

The groundwater hydrograph contains five irrigation seasons from 2005 to 2010. The majority of the irrigation wells in this area are screened in the Ogallala Aquifer. Only a few withdraw water from both the shallow and deep aquifers. During the pumping season, the hydraulic head in the deep well dropped sharply with the average about 10 ft (3.05 m) in the Ogallala Aquifer which resulted in the decline of the water level about 3 ~ 4 ft (0.91 ~ 1.23 m) in the Quaternary aquifer by reducing the upward flow.

The cross-correlation analysis showed that the largest correlation coefficient was 0.605 between groundwater level in well 3S and stream stage when the lag time was 1 day; the largest correlation coefficient was 0.832 between groundwater level in well 2S and stream stage when lag time was 16 days (Figure 2-16 and 2-17). The blue dashed line represents the 95% confidence interval (in all cross-correlation analysis graphs). Considering that the surface water levels are also susceptible to climate, wind speed and the spatial variability of geologic condition between monitoring well and stream gage, such correlation coefficient is remarkable when the time series data were used to carry out the cross-correlation analysis. This result further proves the strong connection between the stream and the Quaternary aquifer at the Atkinson site.
The hydraulic effect of the water level drop between the two aquifers was examined by doing cross-correlation analysis (Figure 2-14). The largest correlation coefficient was 0.613 when lag time was 15 days. It means there were 15 days needed for the water level in the shallow well responding to the water level variation in the deep well.
Groundwater temperature data (Figure 2-18) were also collected at this site. For shallow groundwater well, the temperature ranged from 48 F to 57 F in well 3S and from 47 F to 57 F in well 2S. In Nebraska, the pumping usually starts in June when the groundwater level begins to decrease from the highest state. The declining trend lasts 3 months until the end of August. The graph showed that the groundwater temperature was lowest in the early of June and reaches the peak value at the beginning of November. For the deep well 3D, the annual temperature variation ranged from 51 F to 53 F. Although the variation was small, the yearly regularity was still obvious. For example, from 2006 to 2007, the smallest groundwater temperature at the deep well occurred at July 23 and then the temperature began to increase until March 19, 2007.

Compared to that the temperature difference in well 3D (screened at 210 ft (64.00 m)) was smaller than 2 F, the groundwater temperature difference in 3S was about 10 F with a screen depth at 20 ft (6.10 m). The different extent of the temperature variation is
due to the well-known facts that only very shallow groundwater exhibits appreciable temperature changes in response to the seasonal variations in the amount of solar energy reaching the Earth’s surface (Lovering and Goode, 1963). It is showed that the effective perturbation depth of temperature fluctuations at the Earth’s surface is on the order of 10 m. Thus, one explanation for the larger variation in the shallow well could be attributed to that the shallow groundwater is more susceptible to the air temperature than the deep well.

It is also observed in the shallow well that the groundwater temperature reached the peak value in November while in the deep well the highest temperature occurred in the next March. This lag effect of temperature can be explained by the discipline of heat transfer. Heat can be transported from point to point in a porous media by three processes: conduction, convection and radiation (Domenico and Schwartz, 1990). The effect of radiation and convection is negligible when they are compared to conduction process.

The conductive transport can be described in part by Fourier’s Law (Equation 2-6) where the temperature moves from high region to low region. In porous media, both the fluids and the solids were taken as conductors.

\[
H_e = -K_e \cdot \text{grad} T
\]

Eq. 2 – 6

where \(H_e\) is an effective heat flux, \(K_e\) is an effective thermal conductivity (cal/m s °C), and \(\text{grad} T\) symbolizes the average temperature of the solid and fluid mass. Domenico and Schwartz (1990) reported the thermal conductivities of rocks in which quartz (sand) is 2 (cal/m s °C), clay is 0.2~0.3 (cal/m s °C) and water is 0.11 (cal/m s °C). Apparently, with the same temperature gradient, the higher the thermal conductivity, the bigger the heat flux. Thus, it is expected that the difference of lag time from well to well and site to
site is due to the heterogeneity of sediments at different locations. Besides the
temperature difference between the ends of porous media and the heat conductivity, it is
obvious that the distance between the ends of the porous media would also affect the
velocity of heat transfer.

Figure 2-18 Time series of groundwater temperature data and groundwater level data in well 3D, well 3S and well 2S.

SITE Ewing

The Ewing streambed test site was about 3 miles (4.82 km) southeast of the DNR
stream gage station (Figure 2-19). The test hole, 27-A-59 (Figure 2-20), was drilled up to
350 ft (106.68 m) below the land surface and was 2 miles (3.22 km) away from the well
in southeastern direction and used as an indicator of stratigraphy of the middle and deep
well. The upper 49 ft (14.94 m) of sediments consisted of Quaternary unconsolidated sand. Below the Quaternary sediments was the Ogallala Group up to 304 ft (92.66 m) which was the Tertiary series. A middle well 19M and a deep well 19D were constructed about 7600 ft (2316.48 m) south of the South Fork Elkhorn River. The well 19M and 19D at this site were screened at the depth of 82 ft (24.99 m) and 220 ft (67.06 m), respectively.

Figure 2-19 Plan view of monitoring wells, gage station, streambed test and test holes location at Ewing site.
During the non-irrigation season, the water level was higher in the deep well than in the shallow well. The water level difference between the two wells was greater than 1 ft (0.3 m). This hydraulic head difference suggested that this was a groundwater discharge area where the groundwater from the lower Ogallala Aquifer was capable of moving upward into the shallow Ogallala Aquifer. During the pumping season, the hydraulic head in the deep well dropped about 4 ft (1.22 m) and the water level drawdown in the middle well dropped only 2 ft (0.61 m). These data suggested that the
groundwater pumping reversed the gradient and turned the upward flow into downward flow.

The stream stage data from 2008 to 2010 (Figure 2-21) are available. The water level rose in both the shallow and the deep wells in response to those spikes in the streamflow hydrograph. Therefore, it was likely that there was some degree of hydrologic connection between the river and the upper Ogallala Aquifer.

![Figure 2-21 Times series data of groundwater level in well 19M, well 19D and stream stage.](image)

The cross-correlation analysis (Figure 2-22) showed that the largest correlation coefficient was 0.495 between groundwater level in well 19M and stream stage when the lag time was 11 days; Considering that the surface water levels were also susceptible to
climate, wind speed and other factors, and the well was screened in the upper Ogallala Aquifer, such correlation coefficient was remarkable when using long time series data to carry out the cross-correlation analysis. This result indicates a possible connection between the river and the Ogallala Aquifer.

![Cross-correlation of 19M GW Level and Stream](image)

**Figure 2-22 Cross correlation of groundwater level in well 19M and stream stage.**

The hydraulic effect of the water level drop at the two wells in the Ogallala Aquifer was examined by doing cross-correlation analysis (Figure 2-23). The largest correlation coefficient was 0.946 when lag time was 1 day. It was reasonable that the correlation was high because both wells were screened in the same hydrogeologic system and thus a quicker response can be expected when one of them changes.
Groundwater temperature data (Figure 2-24) were also collected at this site. The temperature ranged from 52.5 F to 54.5 F in well 19M and from 54 F to 55.2 F in well 19D. Although the variation was small, the yearly regularity was still obvious. Accordingly, for well 19M, the graph suggested that the groundwater temperature was the lowest in the middle of June and reached the peak value at the end of December.
Figure 2-24 Time series of groundwater temperature data and groundwater level data in well 19M and well 19D.

SITE Neligh

The Ewing streambed test site was about 2 miles (3.22 km) southeast of the DNR stream gage station (Figure 2-25). The test hole, 15-A-57 (Figure 2-26), was drilled to 350 ft (106.68 m) below the land surface which was 2 miles (3.22 km) away from the well in northwestern direction (used as an indicator of stratigraphy of the middle and deep well). The upper 91 ft (27.74 m) of sediments consisted of the Quaternary unconsolidated sand. Below the Quaternary sediments was the Ogallala Group up to 320
ft (97.54 m). A middle well 17M and a deep well 17D were constructed about 1550 ft (472.44 m) south of the Elkhorn River. The well 17M and 17D at this site were screened at the depth of 74.5 ~ 84.5 ft (22.71 ~ 25.76 m) and 227.5 ~ 237.5 ft (69.34 ~ 72.39 m).

On the north bank of the Elkhorn River, two groundwater monitoring wells were drilled 5 miles (8.05 km) away from stream gage. The well 18S and 18M at this site were screened at the depth of 159.5 ~ 169.5ft (48.62 ~ 51.66 m) and 259 ~ 269 ft (78.94 ~ 81.99 m), respectively. The test hole, 16-A-57 (Figure 2-27), was drilled to 430 ft (131.06 m) below the land surface which was 3 miles (4.83 km) east of well 18S and 18M. The upper 186.5 ft (56.85 m) of sediments consisted of Quaternary unconsolidated sand. Below the Quaternary sediments was the Ogallala Group up to 421 ft (128.32 m) which was the Tertiary series.
Figure 2-25  Plan view of monitoring wells, gage station, streambed test and test holes location at Neligh site.
Figure 2-26 Lithology and stratigraphy of test hole 15-A-57.

Figure 2-27 Lithology and stratigraphy of test hole 16-A-57.
The hydrograph showed that a number of downward spikes (Figure 2-28) were in response to the pumping season from June to August when the groundwater levels in the middle well and deep well decreased. Since the shallow aquifer was unconfined and consisted of mostly sand-sized sediments, it was likely that there was a relatively strong degree of connection between it and river. The cross-correlation analysis showed that the largest correlation coefficient was 0.762 between groundwater level in well 17M and stream stage when the lag time was 1 day (Figure 2-29).

Figure 2-28 Stream flow hydrograph and groundwater hydrograph in well 17M and well 17D.
The data also indicated that the groundwater levels in the middle well 17M and deep well 17D had a similar response pattern during both irrigation and non-irrigation seasons. The water level was higher in well 17D than in well 17M. The water level difference between the two wells was greater than 6 ~ 7 ft (1.83 ~ 2.13 m). This hydraulic head difference suggests that groundwater from the Ogallala Aquifer was moving upward into the lower shallow Quaternary aquifer during either non-irrigation season or irrigation season.

The hydrograph contains three irrigation seasons from 2007 to 2010. During the pumping season, the hydraulic head in the deep well dropped sharply with the average about 6 ft (1.83 m) in the Ogallala Aquifer which resulted in the decline of the water level about 5-6 ft (1.52 ~ 1.83 m) in the Quaternary aquifer by reducing the upward flow. The hysteresis effect of the water level drop between the two aquifers was examined by doing cross-correlation analysis (Figure 2-30). The largest correlation coefficient was 0.762 when the lag time was 0 day.
The connection between the Quaternary and Tertiary aquifer was different at the north bank of the Elkhorn River at Neligh site. The data also (Figure 2-31) indicated that the groundwater levels in the middle well 18S and deep well 18M had a similar response pattern during both irrigation and non-irrigation seasons. The water level was higher in well 18S than in well 18M. The water level difference between the two wells was greater than 1 ~ 10 ft (0.30 ~ 3.05 m). This hydraulic head difference suggested that groundwater from the lower shallow Quaternary aquifer was moving downward into the Ogallala Aquifer during either non-irrigation season or irrigation season.
The hydrograph contains three irrigation seasons from 2006 to 2010. During the pumping season, the hydraulic head in the deep well dropped sharply with the average about 15 ft (4.57 m) in the Ogallala Aquifer which resulted in the decline of the water level about 3-4 ft (0.91 ~ 1.22 m) in the Quaternary aquifer by inducing the water migration downward. The hysteresis (Figure 2-32) effect of the water level drop between the two aquifers was examined by cross-correlation analysis. Although the largest coefficient was at 300 days, the spike at 22 days (when the coefficient is 0.192) was more reasonable by visual judgment. Although the variation pattern of water levels in the shallow aquifer matched the regularity in deep aquifer, this correlation coefficient
suggested that in this area groundwater irrigation had more impact on the groundwater levels in the Ogallala Aquifer than on the groundwater levels in the Quaternary aquifer. Therefore, the aquifers were stratigraphically separate systems and the water levels within each system reacted differently to groundwater pumping stresses.

![Cross-correlation of GW Level 18M and GW Level 18S](image)

*Figure 2-32 Cross correlation between groundwater level in well 18M and groundwater level in well 18S.*

Groundwater temperature (Figure 2-33) showed that in well 19M, the temperature ranged from 52.5 F to 54.5 F and in well 19D the temperature ranges from 53 F to 54.5 F. Although the variation was small, the yearly regularity was still obvious. For well 19M, it was observed from the graph that the groundwater temperature was the lowest at the end of August and reaches the peak value on April 18th in next year. For well 19D, the groundwater level started the lowest value in August and reached the highest value in the middle of May. It was because that it took longer time for the heat energy arriving at the deep aquifer unit.
SITE Meadow Grove

The Meadow Grove streambed test site was about 5 miles (8.05 km) east of the DNR Tilden stream gage station (Figure 2-34). The test hole, 10-LE-99 (Figure 2-35), was drilled to 239 ft (72.85 m) below the land surface. The upper 44.5 ft (13.56 m) of sediments consisted of the Quaternary unconsolidated sand with silt/clay at top 5 ft (1.52 m). The Tertiary Ogallala Group was beneath the Quaternary sediments and penetrated the depth interval from 44.5 to 516 ft (13.56 ~ 157.27 m). Below the Ogallala Group was the Cretaceous Colorado Group at this site. Three wells (the shallow, middle and deep...
wells) were constructed in May 1999 (Groundwater level data in deep well are not available which were located about 4200 ft (1280.16 m) north of the Elkhorn River. The shallow well was designed to monitor the water levels in the Quaternary aquifer at a depth of 31.5 ~ 41.5 ft (9.60 ~ 12.65 m). The middle and deep wells were screened at the Ogallala Aquifer from 127 to 132 ft (38.71 ~ 40.23 m) and from 174 to 179 ft (53.04 ~ 54.56 m), respectively. A second groundwater monitoring site was located about 2300 ft (701.04 m) south of the Elkhorn River. A well (13S) was constructed on June 1999 and was screened at the Quaternary aquifer at a depth of 20 ~ 30 ft (6.10 ~ 9.14 m). The middle well at this site was screened at the depth of 32 ~ 42 ft (9.75 ~ 12.80 m). (Data are not available for this well). A test hole, 11-LE-99, was drilled to a depth of 160 ft (48.77m). According to the log of the test hole, the upper 69.5 ft (21.18 m) consisted of the undifferentiated Quaternary sand with some gravel at the base. The Tertiary Ogallala Group was beneath the Quaternary sediments and penetrated the depth interval from 69.5 to 135 ft (21.18 ~ 41.15 m) deep. The shale was beneath the Ogallala Group sediments up to 160 ft (48.77 m).
Figure 2-34 Plan view of monitoring wells, gage station, streambed test and test holes location at Meadow Grove site.

Figure 2-35 Lithology and stratigraphy of test hole 10-LE-99.
Figure 2-36 Lithology and stratigraphy of test hole 11-LE-99.

Full analysis was difficult to be carried out at this site due to the missing data of water levels in shallow and middle wells and stream stage. But available data still showed general patterns from 2002 to 2010 (Figure 2-37). Before May 2002, the water level was higher in the deep well than in the shallow well during the non-irrigation season. The water level difference between the two wells was greater than 1 ft (0.30 m). This hydraulic head difference suggested that this was a groundwater discharge area where groundwater from the lower Ogallala Aquifer was capable of moving upward into the shallow Quaternary aquifer. During the pumping season, the hydraulic head in the deep well sharply dropped about 10 ft (3.05 m) and the water level drawdown in the shallow well was within 4 ft (1.22 m). These data suggested that groundwater pumping reversed the hydraulic gradient and turn the upward flow into downward flow. It also indicated
that in this area groundwater irrigation had more impact on the groundwater levels in the Ogallala Aquifer than it on the groundwater levels in the Quaternary aquifer. Therefore, the aquifers were stratigraphically separate systems and that the water levels within each system reacted differently to groundwater pumping stresses.

The stream stage data are available from 2007 to 2010. During this period, the water level rose in both the shallow and deep wells in response to those spikes in the stream hydrograph. Therefore, it was likely that there was some degree of hydrologic connection between the river and the Quaternary aquifer and the Ogallala Aquifer. Additional data were needed for the cross-correlation analysis.

Groundwater temperature data (Figure 2-38) were collected at this site. In shallow groundwater well 12S, the temperature ranged from 51.5°F to 55°F. In 2008, the
groundwater temperature was the lowest at the end of June and reached the peak value at the beginning of December. This variation was also believed due to the change of the air temperature through the year and the lag effect was attributed to the extra time that the aquifer took to transfer the heat to the groundwater.

Figure 2-38 Time series of groundwater temperature data and groundwater level data in well 12S.

SITE Norfolk

The Norfolk streambed test site was about 4.07 miles (6.55 km) southeast of the USGS stream gage station (Figure 2-39). The test hole, 15-LE-99 (Figure 2-40), was drilled to 70 ft (21.34 m) below the land surface. The upper 52 ft (15.85 m) of sediments consisted of the Quaternary unconsolidated sand. Below the Quaternary sediments was the Cretaceous Colorado Group up to 70 ft (21.34 m). A middle well was constructed
about 600 ft (182.88 m) north of the Elkhorn River. The middle well was constructed on June 23, 1999 and was designed to monitor water levels in the Quaternary aquifer at a depth of 35.5 ~ 45.5 ft (10.82 ~ 13.87 m). A second groundwater monitoring site was located about 3100 ft (944.88 m) south of the Elkhorn River. The middle well at this site was screened at the depth of 32 ~ 42 ft (9.75 ~ 12.80 m). A test hole, 14-LE-99 (Figure 2-41) was drilled to a depth of 80 ft (24.38 m). According to the log of the test hole, the upper 47 ft (14.33 m) consisted of the undifferentiated Quaternary sand with some gravel at the base. The Colorado Group was beneath the Quaternary sediments up to 80 ft (24.38 m).

Figure 2-39 Plan view of monitoring wells, gage station, streambed test and test holes location at Norfolk site.
Figure 2-40 Lithology and stratigraphy of test hole 15-LE-99.

Figure 2-41 Lithology and stratigraphy of test hole 14-LE-99.
The streamflow hydrograph (Figure 2-42) showed a number of spikes during the high flow periods. The water levels rose in the two middle wells in response to these events. Since the screened depths of both wells were at the Quaternary aquifer which consisted mostly of sand-sized sediments, it was likely that there was relatively strong degree of connection between it and the river.

![Groundwater elevation from well 16M and well 17M and stream stage at Norfolk stream gage.](image)

The cross-correlation analysis showed that the largest correlation coefficient was 0.639 between groundwater level in well 16M and stream stage when the lag time was 1 day; the largest correlation coefficient was 0.683 between groundwater level in well 17M and stream stage when lag time was 8 days (Figure 2-43 and 2-44).
Figure 2-43 Cross correlation of groundwater level in well 16M and stream stage.

Figure 2-44 Cross correlation of groundwater level in well 17M and stream stage at Norfolk stream gage station.
Groundwater temperature (Figure 2-45) showed that in well 16M the temperature ranged from 52 F to 54 F and in well 17M the temperature also ranged from 52 F to 54 F. Although the variation was small, the yearly regularity was still obvious. For well 16M, the groundwater temperature was the lowest at the end of August and reached the peak value at beginning of March in next year. Similar pattern was also found between groundwater level in 17M and relative groundwater temperature.

![Figure 2-45 Time series of groundwater temperature data and groundwater level data in well 16M and well 17M.](image)

**SITE Winslow**

The groundwater monitoring well was about 10.1 miles (16.11 km) southwest of the DNR stream gage station (Figure 2-46). The test hole, 08-LE-08 (Figure 2-47, stratigraphy data are not available at this test hole), was drilled to 293 ft (89.30 m). The test-hole geophysical log at this location indicated that there were 120 ft (36.58 m) of
Quaternary sand and gravel with interbedded layers of silt/clay. The Tertiary Ogallala Group was beneath the Quaternary sediments and penetrated the depth interval of 120 ~ 260 ft (36.58 ~ 48.77 m). This geologic unit consisted of interbedded sand, silt, and clay with varying degrees of cementation. The Cretaceous Pierre shale underlies the Ogallala Group. Two wells (the shallow and deep wells) were constructed in June 2008 and they were located about 1300 ft (396.24 m) east of the DNR Winslow stream gage. The shallow well was designed to monitor water levels in the Quaternary aquifer at a depth of 23 ~ 43 ft (7.01 ~ 13.11 m). The deep well was screened at the bottom of the Ogallala Aquifer from 205 to 215 ft (62.48 ~ 65.53 m) deep.

Figure 2-46 Plan view of monitoring wells, gage station, streambed test and test holes location at Winslow site.
Figure 18 showed the hydrographs of the groundwater levels recorded in the shallow and deep wells. The groundwater level in the shallow well was about 7 ft (2.13 m) higher than the water level in the deep well during the non-irrigation season. During the irrigation season in 2010, the maximum water level decline was 2 ft (0.61 m) and 20 ft (6.1 m) within the shallow and deep wells, respectively. This suggested that in this area the groundwater irrigation had more impact on the groundwater levels in the Ogallala
Aquifer than it on the groundwater levels in the Quaternary aquifer. Therefore, the aquifers were stratigraphically separate systems and the water levels within each system reacted differently to groundwater pumping stresses. The test-hole geophysical log at this groundwater monitoring site indicated the silt/clay deposits between 60 ft (18.29 m) and 200 ft (60.96 m) which may be acting as confining layers that reduce the degree of hydrologic connection between the two aquifers (Figure 2-47).

The stream stage hydrograph (Figure 2-48) showed several spike from 2010 to 2011, but the groundwater levels were stable through the same period. Since both the streamflow data and groundwater level data were monitored less than a year at this site, additional data were required for the cross-correlation analysis of the response of groundwater level to flooding events and the groundwater temperature analysis.
SITE Hadar

The Hadar groundwater monitoring well was about 700 ft (213.36 m) northwest of the USGS Hadar stream gage station (Figure 2-49). The test hole, 1-LE-99 (Figure 2-50), was drilled up to 220 ft (67.06 m) below the land surface. The upper 105 ft (32.00 m) sediments consisted of the Quaternary unconsolidated sand with interbedded silt/clay. The Tertiary Ogallala Group was beneath the Quaternary sediments and penetrated the depth interval from 105 to 192 ft (32.00 ~ 58.52 m). Below the Ogallala Group was the Colorado Group at this site. The groundwater monitoring site were located about 500 ft (152.4 m) west of the North Elkhorn River. Three wells (the shallow, middle and deep wells) were constructed in May 1999 (Groundwater level data in the deep and middle...
The shallow well was designed to monitor water levels in the Quaternary aquifer at a depth of 20 ~ 30 ft (6.09 ~ 9.14 m). The middle and deep wells were screened at the Ogallala Aquifer from 66 ~ 76 ft (20.12 ~ 23.16 m) and 151 ~ 156 ft (46.02 ~ 47.55 m) deep, respectively.

Figure 2-49 Plan view of monitoring wells, gage station, streambed test and test holes location at Hadar site.
In Figure 2-51, the groundwater level in 04S dropped regularly from May to September during the pumping season. The maximum water level decline was 2 ft (0.61 m), in response to the stream stage decline of 2 ft (0.61 m). The cross-correlation analysis (Figure 2-39) shows that the largest correlation coefficient between groundwater level and stream stage was 0.457 when a lag was 100 days.
Groundwater temperature (Graph 2-52) showed that in well 04S the temperature ranged from 47 F to 54 F. Compared to the former sites, the variation of groundwater level in well 04S was pronounced. In the graph, the groundwater temperature was the lowest at the beginning of June and reached the peak value at the end of December.
The Pierce streambed test site was about 3.65 miles (5.87 km) northwest of the DNR stream gage station (Figure 2-53). The test hole, 5-LE-99 (Figure 2-54), was drilled up to 319 ft (97.23 m). The test-hole geophysical log at this location indicated that there were 165 ft (50.29 m) of Quaternary sand and gravel with interbedded layers of silt/cay. The Tertiary Ogallala Group was beneath the Quaternary sediments and penetrated the depth interval of 165 ~ 319 ft (50.29 ~ 97.23m) deep. This geologic unit consisted of interbedded sand, silt, and clay with Niobrara fragments at the bottom. The groundwater monitoring well were located about 4100 ft (1249.68 m) south of the Willow Creek. Three wells (shallow, middle and deep wells) were constructed in May 1999 (Groundwater level data in deep well are currently not available).
the middle well were designed to monitor water levels in the Quaternary aquifer at a depth of 56 ~ 66 ft (17.07 ~ 20.12 m) and depth of 149 ~ 159 ft (45.42 ~ 48.46 m), respectively. The deep well was screened at the bottom of the Ogallala Aquifer from 275 ~ 280 ft (83.82 ~ 85.34 m) deep. Water levels obtained from the shallow well were above the top of the aquifer, indicating that unconfined conditions existed within the Ogallala Aquifer at this site.

Figure 2-53 Plan view of monitoring wells, gage station, streambed test and test holes location at Pierce site.
The data indicated that the groundwater levels in the shallow and middle well had a similar response pattern during both irrigation and non-irrigation seasons. Figure 2-55 showed the hydrograph of the groundwater levels in the shallow and the middle wells at the south of the river. During the non-irrigation season the water level was higher in the shallow well than in the middle well. The water level difference between the two wells was greater than 5 ft (1.52 m). This hydraulic head difference suggested that this was a groundwater discharge area where groundwater from shallow Quaternary aquifer was moving downward and into the lower Ogallala Aquifer.
The hydrograph contains six irrigation seasons from 2004 to 2009. The majority of the irrigation wells in this area are screened in the Ogallala Aquifer. Only a few withdraw water from both the shallow and deep aquifers. During the pumping season, the hydraulic head in the deep well dropped sharply with the average about 10 ft (3.05 m) which resulted in the decline of the water level about 2 ~ 3 ft (0.61~ 0.91 m) in the Quaternary aquifer by inducing the water migrating downward into the Ogallala Aquifer. The hysteresis effect of the water level drop between the two aquifers was examined by
the cross-correlation analysis (Figure 2-56). The largest correlation coefficient was 0.739 when lag time was 17 days.

Figure 2-56 Cross correlation between the groundwater levels in well 06S and well 06M.

The stream stage hydrograph (Figure 2-57) showed a number of spikes during the six irrigation seasons. The water levels in the shallow and deep wells were higher than the stream stage. The stream gage was 1660 ft (505.97 m) above sea level which was located downstream about 3.74 miles (6.02 km) away from the monitoring well where the land surface elevation was 1711 ft (521.51 m). Taking account of the factor that the stream gage and groundwater monitoring well were not located at the same position, the water level difference between the wells and stream stage was about 29 ft (8.84 m) and 19 ft (5.79 m), respectively. The number of spikes in the stream stage was a reflection of natural discharge, irrigation pumping, natural stresses and variations in climatic conditions such as temperature and precipitation. Thus it is foreseen that the stream stage and groundwater level would not match perfectly due to the more variation characteristics.
of stream stage data. The cross-correlation analysis (Figure 2-58 and 2-59) showed the
largest correlation coefficient was 0.424 and 0.292 between groundwater level in middle
well and shallow well and stream stage when the stream stage lagged for 19 and 2 days.

Figure 2-57 Groundwater elevation from the shallow and middle well and stream stage recorded at
the Foster gage station.
Figure 2-58 Cross correlation between the groundwater levels in well 06M and stream stage.

Figure 2-59 Cross correlation between groundwater levels in well 06S and stream stage.
The groundwater temperature (Figure 2-60) variation reflected unique characteristics at this site. From the graph, the groundwater temperature was relatively stable which ranged from 53.4 F to 53.8 F from 1998 to 2003 and ranged only from 53.8 F to 54.0 F after 2003. The yearly pattern of the groundwater temperature at this site was flat and differed from the variation pattern observed at other sites.

Figure 2-60 Comparison of groundwater temperature and groundwater level in well 06M.
Chapter 3 Hydrostratigraphic model

3.1 Objective

Traditionally, the study of river aquifer interactions is from regional scale where the representation of the hydrogeological condition is crude and hydraulic parameters are often assumed to be homogeneous (Heinz et al., 2003). However, the models used to support local water resource management or environmental safety would need to account for small scale patterns and dynamics of river-aquifer exchange (Dahl et al., 2007). Fleckenstein and Fogg (2008) proposed the TPROGS algorithm to create geostatistical heterogeneity model and upscaled the hydraulic parameters. Another difficulty of characterizing subsurface heterogeneity is normally the datasets of wire line logs, driller's lithological logs may provide excellent information on the vertical variability of the sediments but only limited information about the lateral distribution and variability of the deposits (Weissmann et al., 1999).

The objective of this chapter is to build a reasonable three dimensional geologic model that will consider the spatial variation of the aquifer deposits in terms of hydrofacies and then upscale the geological model into hydro stratigraphic model which could be imported into a groundwater flow model.

3.2 Model Area

The DEM area in Figure (3-1) is the groundwater model area that covers part of Upper Elkhorn NRD and Lower Elkhorn NRD which includes five study sites: Neligh, Tilden, Norfolk, Hadar and Pierce. However, the hydrostratigraphic model area is within the red rectangle which is designed to be slightly larger than the groundwater model since
it will provide better estimation of the hydrogeology condition at the boundary of groundwater model area during interpolation.

![Figure 3-1 Hydrostratigraphic model area.](image)

### 3.3 Geology Data

#### 3.3.1 Test hole data

The Nebraska statewide test-hole database contains information for about 5,500 test holes that were drilled since 1930 by the Conservation and Survey Division (CSD), School of Natural Resources (SNR), University of Nebraska, and cooperating agencies. The database includes: test-hole location, lithological descriptions, stratigraphic interpretations and geophysical log records and usually a geologist was present who
logged the lithology during drilling. A total of 348 test holes (Figure 3-1 blue triangle) were used in this study.

3.3.2 Registered well logs

Besides test hole data, another geological data source were derived from the registered wells which was created and was maintained by Department of Natural Resources. Only the registered wells which are within the groundwater model area are used for future interpolation. These registered wells are categorized in: irrigation, municipal use, drinking water source. During well construction, boring data were interpreted by well drillers. The quality of the database varies from place to place depending on well drillers. Thus, a small portion of the registered well was excluded due to the concern of quality. A total of 7885 registered wells were used and were showed as red dots in Figure 3-2.

Figure 3-2 Registered wells used in the construction of hydrostratigraphic model.
3.4 Interpreting Well Log Description

The quality of the test hole data are higher than registered wells. Since the test hole data were interpreted and recorded by professional geologists from Conservation and Survey Division at UNL, and those registered well reports were submitted by private well drillers who had insufficient professional training, and some logs were compiled after well completion. Another problem is that the descriptions were based on predefined intervals rather than based on the subtle differences in sediment. Consequently, the quality of the data varies considerably. The third concern is the using of professional language, a term such as “hard layer” refer to limestone, however it is also possible referring clayey silt or sandstone. A lot of effort has been made to be consistent and accurate when reading and translating the test hole log and well log into geologic categories.

3.4.1 Eight hydrofacies

Mickelson et al. (2008) proposed a hydrostratigraphic model which has four hydrofacies. In this hydrostratigraphic model, sediments were categorized into eight hydrofacies: clay, silt/till, sand, gravel, limestone, siltstone, and sandstone. The log description and its corresponding hydrofacies with hydrofacies number in the model were listed in Table 3-1. The gravel, sand, sandstone and limestone facies formed the aquifer, while the clay, silt/till and siltstone formed aquitard layers. The gravel facies mainly consisted of cobbles and pebbles; coarse sand was also included into gravel facies due to its similar $K$ value as gravel. Except for the unconsolidated sedimentary materials, limestone, siltstone and sandstone facies were used to represent the sedimentary rocks that formed part of the Tertiary Ogallala Aquifer or the upper part of the Cretaceous
group. Table 3-2 gives an example of translation from the well log description into the grouped hydrofacies log.

3.4.2 Hydraulic conductivity

Literature values of hydraulic conductivity were assigned to each of the eight hydrofacies. There are several available sources of representative values of hydraulic properties. The most popular one refers to the textbook of Domenico and Schwartz (1990). Field derived values of hydraulic conductivity from the literature were used rather than laboratory derived values because they better account for larger scale features such as weathering horizons and fractures (Mickelson et al., 2008). Dugan (1983) did some research in the hydraulic conductivity values at south of LPNRD. Since the LPNRD is geographically adjacent to Elkhorn River Basin, these values were used to assign the hydraulic conductivity value of unconsolidated deposits in the model.
Table 3-1 Categorization of the log description into eight hydrofacies and its corresponding hydraulic conductivity value.

<table>
<thead>
<tr>
<th>Hydrofacies</th>
<th>Test hole and Well log description</th>
<th>Shale</th>
<th>Clay</th>
<th>Silt/Till</th>
<th>Sand</th>
<th>Gravel</th>
<th>Sandstone to sandstone, sand and sandstone to sandy sandstone</th>
<th>Siltstone to sandstone, sandstone to sandy sandstone</th>
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<td>4</td>
<td>5</td>
<td>6</td>
<td>7</td>
<td>8</td>
<td></td>
</tr>
<tr>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>Silt/Till</td>
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<td>Sand</td>
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<table>
<thead>
<tr>
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<th>100</th>
<th>400</th>
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<td>4</td>
<td>100</td>
<td>400</td>
<td>0.2</td>
<td>0.02</td>
</tr>
</tbody>
</table>

| Test hole and Well log       | 1 | 0.001 | 4 | 100 | 400 | 0.2 | 0.02 |

| Test hole and Well log       | 1 | 0.001 | 4 | 100 | 400 | 0.2 | 0.02 |

| Test hole and Well log       | 1 | 0.001 | 4 | 100 | 400 | 0.2 | 0.02 |
Table 3.2 Well log of registered well and its corresponding hydrofacies in the hydrostratigraphic model; hydraulic conductivity values of each facies was referenced from Domenico and Schwartz (1990) and Digan (1983).

<table>
<thead>
<tr>
<th>ID</th>
<th>WellID</th>
<th>FromDepth</th>
<th>ToDepth</th>
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<tr>
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<td>119</td>
<td>130</td>
<td>SAND</td>
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</table>
The hydraulic conductivity value $K$ of hydrofacies 1 was determined mainly from Domenico and Schwartz (1990) who gave a range of $10^{-11}$ to $10^{-9}$ m/s, which is from $10^{-6}$ to $10^{-3}$ ft/day. However, most cited data were around or smaller than $10^{-3}$ ft/day. Thus, the value of $10^{-3}$ ft/day was chosen for these deposits. Silt/till deposits have values ranging over several magnitude from $10^{-12}$ to $10^{-5}$ m/s, Dugan (1983) measured the $K$ values from 1 to 15 ft/d (0.35 ~ 4.57 m/d) at Lower Platte NRD. Because most silt is very sandy at this study site, 4 ft/d (1.22 m/d) was chosen for hydrofacies 2 which is recommended by Dugan (1983). Sand and gravel $K$ values were also determined based on Dugan’s study, in which the $K$ for sand is 100 ft/d (30.48 m/d) and for gravel is 300 ft/d (91.44m/d). Hydrofacies 6, 7 and 8 were mainly determined by Domenico and Schwartz (1990), middle value was picked as 0.2 ft/d (0.06 m/d), 0.02 ft/d (0.006 m/d) and 1 ft/d (0.35 m/d), respectively.

3.5 Geological Modeling Software.

After comparing several commercial softwares packages, Rockworks 15 (Rockware, 2012) was chosen to construct the hydrostratigraphic model. Rockworks was developed by Rockware Company which was aimed to serve the environmental and mining industry. The reason that Rockworks was selected was because of its ability to import and visualize the borehole data, to draw cross sections and to create 3D solid model which is important for the model upscaling. Rockworks is also powerful in visualizing the well raw data (Figure 3-3) which is very useful for the modeler to determine how many model layers should be built for the groundwater flow simulation.
3.6 Geologic Model Design

A grid spacing of 2640 ft (804.67 m) by 2640 ft (804.67 m) was used, creating 118 nodes in the X direction and 162 nodes in the Y direction. The vertical grid was spaced in 2 ft (0.61 m) to create 476 layers in the Z direction. The final hydrostratigraphic model had a total of 9,099,216 nodes.

3.6.1 Raw data revision and cross section

All the well data and test hole data were grouped into hydrofacies and were assigned number as showed in Table 3. This number was associated with the hydraulic
conductivity value when Rockworks created the solid model. All the hydraulic conductivity values were transformed into natural logs ($lnK$) because the $K$ usually ranges over several orders of magnitude. The transformation would help to prevent the creation of anomalous regions by using just the normal values of $K$.

After the revision on the raw data, the lithology data with hydrofacies number and associated $lnK$ value were imported into rockworks. Spatial locations of the test holes and registered wells referred to the coordinate information from DNR web site. Then the lithology of those wells was displayed as cylinders in 3D space. Visualizing the raw data in 3D was a helpful way to develop an intuitional and general sense of the geology in the region, before allowing Rockworks to create a solid model (Mickelson et al., 2008).

First, we can observe that the top sediments along northeastern bound and along southwestern bound of the model region consist mostly of clay and silt by looking from the top point (Figure 3-4) of view. In cross section map (Figure 3-5 through Figure 3-12), the noticeable feature is that generally the sand and gravel unit is thicker in the west region and becomes gradually thinner from west to east by looking at cross section A-A’, B-B’ and C-C’. The graph also indicates the missing clay deposit between the Quaternary aquifer and the Ogallala Aquifer at some region. This feature suggests that the low $K$ aquitard is not continuous in the model region (Figure 3-5 through Figure 3-7).
Figure 3-4 Distribution of deposit at the surface from top point of view.
Figure 3-5 Cross section location map.
Figure 3-6 Cross section A-A'.
Figure 3-7 Cross section B-B'.
Figure 3-8 Cross section C-C'.
Figure 3-9 Cross section D-D'.
Figure 3-10 Cross section E-E'.
Figure 3-11 Cross section F-F'.
Figure 3-12 Cross section G-G'.

Cross-Section G-G'
3.6.2 Interpolation algorithm

Since the huge amount of nodes will be used for computation, the Inverse-Distance Weighting method was selected as the interpolation method in the solid modeling for the consideration of simplicity and time saving. Taken the description from Rockworks help manual (Rockware, 2012) for inverse distance method: The Inverse-Distance Weighting modeling method is one of the "flavors" of the Inverse-Distance algorithm. In general, with this method, the value assigned to a grid node is a weighted average of either all of the data points or a number of directionally distributed neighbors. The value of each of the data points is weighted according to the inverse of its distance from the grid node where an interpolated value is generated, taken to a user-selected power (Equation 3-1). (An exponent of "2" = Inverse-Distance squared, "3" = Inverse-Distance cubed, etc).

\[ Z_{node} = \frac{\sum (Z_{point})}{\sum (1/d^n)} \]  \hspace{1cm} \text{Eq. 3-1}

where \( Z \) is the attribute value, \( d \) is the distance from the node that is being solved for, and \( n \) is an exponent. The greater the value of the exponent, the less influence distant control points will have on the assignment of the node value. The greater the value of the exponent you specify, the more localized the gridding since distant points will have less influence on the value assigned to each grid node. The advantage of IDW is that this method produces a smooth and continuous grid and will not exaggerate its extrapolations beyond the given data points. The range of grid values will be smaller than the data point range. The highest grid value will be less than the maximum data point, and the lowest grid value will be greater than the minimum data point.
Rockworks also lets the user define the maximum number of data points that are to be used when computing the grid node value. The default number in Rockworks is 8; increasing the number of points will decrease bulls-eyes (concentric closed contours around control points) but slow down the gridding process.

### 3.6.3 Model parameter selection

The Inverse-Distance Weighting method can use either all of the available data points when computing a node’s value or it can search for specific points. And, instead of automatically using a weighting exponent of "2", the program allows the user to assign different weighting exponents to control points oriented vertically versus horizontally from the node. The greater the exponent you enter, the less influence those data points will have.

The default values at horizontal and vertical direction are both 2. The biggest value that is allowed in Rockworks is 12 and the smallest value is 0. In natural condition, lateral continuity of geologic facies is usually significantly stronger than the vertical continuity which has been proved by stratigraphy. This standard will be used as a criterion in selecting the weighting component that will result in a more flat formation and reduce “bull-eye” effect.

Rockworks provided an example (Figure 3-13) of the settings for weight exponents. The model on the left within the following diagram was based on horizontal and vertical exponents of 2.0. The model on the right is based on a horizontal exponent of zero and a vertical exponent of 6.

Note the pronounced lenticularity within the model (Figure 3-13). The regularity of the formation is preferred considering the geologic formation in the natural condition. A total of twelve models were tested with different combination of weight component in
horizontal and vertical directions. The default setting 2-2 was first run as a base case. Then the extreme setting was used as contrary examples. From the example above, horizontal weighting component should be set to smaller than 2 in order to achieve continuation.

After the horizontal weighting was fixed, the vertical weighting component was selected through trial. It was found that the “bull eye” dominated when the vertical weighting component went extreme. Mickelson et al.(2008) used a trial method to find parameter combination which generated the reasonable geologic models compared to the raw data. Similar method was used here too. Table 3-3 shows the visualized result of different combinations. Finally horizontal weighting equal to 1 and vertical weighting equal to 6 were chosen, and the maximum data point was selected as the searching method that all nodes would be considered when computing one grid node value. Figure 3-14 shows the solid model that was generated using above the settings.
Table 3-3 Different weighting parameter used for inverse distance modeling.

<table>
<thead>
<tr>
<th>H. VS V. Weight Component</th>
<th>Solid Model Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>2-2</td>
<td>Base case; no weighting; vertical tubes, no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>0-0</td>
<td>Slight horizontal weighting; vertical tubes, no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>12-12</td>
<td>Slight horizontal weighting; vertical tubes, no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>0-6</td>
<td>extreme horizontal weighting; no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>1-6</td>
<td>average horizontal weighting; no extreme problems; no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>1.5-6</td>
<td>Slight horizontal weighting; no extreme problems</td>
</tr>
<tr>
<td>2-6</td>
<td>Slight horizontal weighting; no extreme problems</td>
</tr>
<tr>
<td>1-4</td>
<td>Average horizontal weighting; no extreme problems, no northeastern and southwestern surface clay</td>
</tr>
<tr>
<td>1-8</td>
<td>Average vertical weighting; bull eye effect</td>
</tr>
<tr>
<td>1-10</td>
<td>Extreme vertical weighting; obvious bull eye effect</td>
</tr>
<tr>
<td>1-12</td>
<td>Extreme vertical weighting; no northern and southern surface clay</td>
</tr>
</tbody>
</table>

Figure 3-14 Solid model using horizontal setting equals 1 and vertical setting equals to 6. The scale bar represents the range of lnK value of the model nodes.
3.7 Geologic Model Upscaling to Hydrostratigraphic Model

Modeling groundwater flow requires a realistic description of the spatial distribution of hydraulic conductivity to capture flow paths and to make realistic forecasts of groundwater behavior (Renard et al., 2000). In this chapter, we have used IDW method to describe the subsurface distribution of deposits that produced a very high spatial resolution. It was mentioned in the section 3.5 that there are more than 9 million nodes and 476 layers in the geological model. Due to the lack of meaning in hydrogeology, computational cost and huge difficulties in the model calibration process, such high resolution information cannot be directly used in the groundwater model. Thus it is necessary to upscale the geologic model into hydrostratigraphic model which has coarser grid and fewer layers.

Upscaling is a technique that transforms a detailed geologic model to a coarse-grid simulation model so that the fluid-flow behaviors in the latter one could get as similar as possible to the former one. The upscaling approach consists of two separate steps: gridding and averaging. The former one intends to capture the global geologic features of a geologic model, the latter one focuses on calculating an equivalent property that behaves similarly (Li et al., 1999).

3.7.1 Gridding

A coarse grid 59×81×3 was used to scale up the 118×162×476 grid. Each coarse grid contained 2×2×Zi fine grid cells. The gridding process for the X and Y directions are showing in the graph. The area contained in the red line represents a single coarse grid in a plane view which contains four finer grids (Figure 3-15).
Figure 3-15 Gridding process in the horizontal direction.
Figure 3-16 Gridding process in vertical direction.
Then two layers were used to select the grid in between (Figure 3-16). The layer was generated in grid file that had the same resolution. For each fine grid, a unique number of grids were selected. The average of these four numbers was assigned to Zi in order to assure a cubic shape for each coarse grid. There were a total of four layers used. The first layer represented the land surface, the second layer represented the bottom of the upper aquifer, the third layer represented the bottom of the middle aquifer, and the fourth layer represented the bottom of the lower aquifer. A C++ code was written to carry out this selecting process (Appendix A). At last, three sub-solid models were created through these four layers during the selecting process and each will be processed to calculate the equivalent hydraulic conductivity through averaging process.

3.7.2 Averaging

The averaging process aims to calculate the effective hydraulic conductivity of the coarse simulation grid in a way that preserves fine grid flow characteristics within the coarse grid. A review of past research in the area of upscaling has been conducted (Desbarats, 1992). The actual conductivity of a heterogeneous medium can be obtained by averaging theoretical bounds with an adequate procedure. One possibility is to use some bounds which are closer than the arithmetic mean and the harmonic mean and then to average them. Whatever the averaging procedure, the closer the bounds are, the smaller the expected errors are (Desbarats, 1992; Renard et al., 2000). Theoretically, the arithmetic mean and harmonic mean are the upper and lower bound for up scaled hydraulic conductivity. The approach to calculate the upper bound and lower had been widely discussed in the oil industry. This study adopted an efficient technique proposed by Li et al., (1999) which is designed for scaleup of multimillion-cell geological models.
The first step is to calculate the upper bound and lower bound of $K_X$, $K_Y$, and $K_Z$ using the following (Equations 3-2 through 3-7) as Li et al. showed in their paper:

\[
K_{X+} = \frac{\Delta X}{\Delta Y \Delta Z \sum_{i=1}^{N_x} \frac{\Delta x_i}{\sum_{j=1}^{N_y} \sum_{k=1}^{N_z} \Delta y_j \Delta z_{i,j,k} K_{x,i,j,k}}} \quad \text{Eq. 3 - 2}
\]

\[
K_{X-} = \frac{\Delta X}{\Delta Y \Delta Z \sum_{j=1}^{N_y} \sum_{k=1}^{N_z} \Delta y_i \Delta x_i \Delta z_{i,j,k} K_{x,i,j,k}} \quad \text{Eq. 3 - 3}
\]

\[
K_{Y+} = \frac{\Delta Y}{\Delta X \Delta Z \sum_{i=1}^{N_x} \frac{\Delta y_j}{\sum_{j=1}^{N_y} \sum_{k=1}^{N_z} \Delta z_{i,j,k} K_{x,i,j,k}}} \quad \text{Eq. 3 - 4}
\]

\[
K_{Y-} = \frac{\Delta Y}{\Delta X \Delta Z \sum_{i=1}^{N_x} \sum_{j=1}^{N_y} \Delta x_i \Delta y_j \Delta z_{i,j,k} K_{x,i,j,k}} \quad \text{Eq. 3 - 5}
\]

\[
K_{Z+} = \frac{\Delta Z}{\Delta X \Delta Y \sum_{k=1}^{N_z} \frac{\Delta z_k}{\sum_{i=1}^{N_x} \sum_{j=1}^{N_y} \Delta x_i \Delta y_j \Delta z_{i,j,k} K_{x,i,j,k}}} \quad \text{Eq. 3 - 6}
\]

\[
K_{Z-} = \frac{1}{\Delta Z \Delta Y} \sum_{i=1}^{N_x} \sum_{j=1}^{N_y} \sum_{k=1}^{N_z} \frac{\Delta x_i \Delta y_j \Delta z_{i,j,k}}{K_{x,i,j,k}} \quad \text{Eq. 3 - 7}
\]

Nomenclature

$k_x$, $k_y$, $k_z$ = hydraulic conductivity of fine grid along the $x$, $y$, and $z$ direction, $k_x$ using the value from Table 2 and $k_y=0.8\cdot k_x$, $k_z=0.1\cdot k_x$.

$K_X$, $K_Y$, $K_Z$ = the hydraulic conductivity of coarse-grid along $x$, $y$, $z$ direction

$N$ = number of grid blocks in a specific direction

$N_t$ = number of tubes

$t$ = tube index

$\Delta x_i$ = fine gridblock length in $x$ direction
\( \Delta X \) = coarse gridblock length in \( x \) direction
\( \Delta y_i \) = fine gridblock length in \( y \) direction
\( \Delta Y \) = coarse gridblock length in \( y \) direction
\( \Delta z_{i,j} \) = cumulative thickness of column, \( i, j \).
\( \Delta z_{i,j,k} \) = gridblock thickness
\( \Delta z_k \) = average thickness of layer \( k \)
\( \Delta Z \) = coarse gridblock length in \( z \) direction
\( H \) = in the horizontal direction
\( i \) = index in \( x \) direction
\( j \) = index in \( y \) direction
\( k \) = index in \( z \) direction
\( t \) = tube index
\( x \) = in \( x \) direction
\( y \) = in \( y \) direction
\( z \) = in \( z \) direction
\( + \) = upper bound
\( - \) = lower bound

There are numerous ways to estimate the effective hydraulic conductivity using upper and lower bounds. The approach (Equation 3-8 through 3-11) used here is proposed by Cardwell and Parsons (1945) and improved by Li et al., (1999). An iterative procedure is used to derive the equations to estimate the horizontal directional effective hydraulic conductivity. The geometric average of the initial \( x \)-directional upper and lower bounds gives the first iteration. Because the geometric mean tends to give a result that is
closer to lower bound, then the geometric mean will be used as a new lower bound for the second iteration. The second estimate is considered higher than the effective hydraulic conductivity. Thus the second estimate will be used as upper bound for the third iteration. Thus through several rounds of iteration, the estimate are forced to approach the effective hydraulic conductivity Li et al., (1999).

\[ K_X = K_X + \frac{5}{8} \cdot K_X - \frac{3}{8} \]  
Eq. 3 – 8

\[ K_Y = K_Y + \frac{7}{8} \cdot K_Y - \frac{1}{8} \]  
Eq. 3 – 9

\[ K_Z = (1 - a) \cdot (K_Z^+) + a \cdot (K_Z^-) \]  
Eq. 3 – 10

\[ a = 0.823 + 0.167 \cdot \lambda \]  
Eq. 3 – 11

where \( \lambda = \frac{K_Z^-}{K_H} \), \( K_Z^- \) = the arithmetic mean of the fine-grid z directional hydraulic conductivities within a coarse grid block, and \( K_H \) is the arithmetic mean of the fine-grid horizontal-directional (both \( x \) and \( y \)) hydraulic conductivities within the coarse grid block.

Thus, these three-sub solid models were upscaled and combined to constitute a hydrostratigraphic model. This hydrostratigraphic model had three aquifer units and every single coarse grid in each aquifer was associated with unique effective hydraulic conductivity both in horizontal direction and vertical direction. The whole calculation process was programmed with C++ language (Appendix B).

3.7.3 Categorization in hydraulic property zone

Although the geologic model was scaled up using a coarse grid, the resolution of the hydrostratigraphic model is still too dense for the future calibration of hydraulic conductivity for the groundwater model. So it is necessary to group the coarse grid with adjacent hydraulic conductivity values into the same hydraulic property zone. The lower limit and upper limit was set according to the hydrofacies value which was selected from
literature and local studies. For each zone, the frequency distribution graph was plotted to find out the value having the highest frequency within that zone. This highest frequency value was picked and assigned to every grid whose value fell in that zone. Figure 3-17 shows the number of frequency of the hydraulic conductivity. Eight zones were defined in the upper aquifer unit and their hydraulic conductivity values fell in these ranges: 0.01 to 0.1 ft/d (0.00305 ~ 0.0305 m/d), 0.1 to 1 ft/d (0.0305 ~ 0.305 m/d), 1 to 5 ft/d (0.305 ~ 1.52 m/d), 5 to 10 ft/d (1.52 ~ 3.05 m/d), 10 to 20 ft/d (3.05 ~ 6.10 m/d), 20 to 30 ft/d (6.10 ~ 9.14 m/d), 30 to 50 ft/d (9.14 ~ 15.24 m/d) and 50 to 300 ft/d (15.24 ~ 91.44 m/d). Then the highest frequency value within each zone was picked and assigned as the representative hydraulic conductivity value to that zone. In the end, the three aquifer unit were assigned with KX, KY and KZ value as attributes and converted into polygon shapefile for importing into groundwater flow software for the next step. All these process was done using C++ programming language as listed in the appendix C. Figure 3-18, 3-19, and 3-20 show an example of KX, KY, KY distribution of every grid of layer1 and the K distribution after zonation, where Kx, Ky and Kz is hydraulic conductivity within each K zone after zonation.
Figure 3-17 Number of frequency of hydraulic conductivity for zonation.

Figure 3-18 Calculated KX value with coarser grid and the Kx hydraulic property zone in layer 1.
Figure 3-19 Calculated KY value with coarser grid and the Ky hydraulic property zone in layer 1.

Figure 3-20 Calculated KZ value with coarser grid and the Kz hydraulic property zone in layer 1.
Chapter 4 Upper and Lower Elkhorn Groundwater Model

4.1 Objective

From the evaluation in Chapter 2, the Elkhorn River is a gaining river at all eight sites. It suggests that the baseflow discharge from the adjacent aquifer is an important component of streamflow in the Elkhorn River. The cross-correlation analysis indicates the potential connection between the Elkhorn River and the adjacent aquifer. Thus, it is important to evaluate the effect of groundwater pumping on the baseflow discharge to the Elkhorn River.

Concerns of sustainability using groundwater and surface water urged the Nebraska state legislature to create a law to give DNR an increased responsibility for the administration of water resources. State legislation was enacted in 2004 and claimed that future development of water use will be carefully considered if it is located in the hydrologically connected area. According to the Department of Natural Resources, the hydrologically connected area (10/50 area) is defined as those areas within which pumping of a well for 50 years will deplete base flow by at least 10% of the pumped volume.

4.2 Location, Topography and Climate

The ULEN model area covers part of the Upper Elkhorn NRD and part of the Lower Elkhorn NRD and is approximately 2,000 square miles (5200 km²) including all of Antelope, Pierce, Madison County and portions of Stanton, Wayne and Platte. The biggest city in study area is Norfolk which has a population of 23,272 according to Google Public Data on July, 2009. The boundaries of the model were queried from National Hydrologic Watershed Unit Database and also are topographic divides of
drainage basins. The topography is higher in the west and lower in the east. The highest ground surface elevation is about 2172.7 ft (662.64 m) at southwestern part of Antelope County, and the lower ground surface elevation is about 1443.9 ft (440.10 m) at the downstream end of the Elkhorn River in the Stanton County. The altitude difference is about 728.8 ft (222.14 m) in the study area. The main topographic regions in the area are sand hills, wet meadows and marsh plains, loess hills, river valleys, transitional sandy plains, dissected loess plains, plains, and river breaks.

The area has a humid continental climate with cold but relatively dry winters and hot and occasionally humid summers. Concurrence of heat and rain in the summer provides ideal growing conditions for crops in this area. The average annual precipitation ranges from 27.9 inches (708.66 mm) at Neligh in the western portion of the model area to 28.1 inches (713.74 mm) at Norfolk in the eastern portion of the model area (Figure 4-1). The area receives an average annual precipitation of 28 inches (711.2 mm). The graph shows the monthly average precipitation from 1980 to 2010 during which 63% of annual precipitation occurs between May and September which is also the growing season for crops. There is not big difference between the two stations in monthly average precipitation (Figure 4-2).
Figure 4-1  Yearly average precipitation in 30 years in the study area at Norfolk station (data was downloaded from HPRCC).

Figure 4-2 Comparison of monthly precipitation at Neligh and Norfolk weather stations (data was downloaded from HPRCC).
4.3 Geology and Hydrogeology

Geological strata overlying the upper Cretaceous shale are of importance for this study because they are the principal aquifer systems. These include the Ogallala Group, alluvial deposits in the river valley and the Quaternary deposits in the uplands. The Cretaceous shale is considered as a no-flow boundary in the model. Logs of 348 test holes are available in the model area for characterization of the aquifer systems.

**Quaternary Deposits**

Quaternary-age deposits are composed of wind-deposited silts or fine-grained sands, or alluvial silt, sand, and gravel. Quaternary deposits overly on the Ogallala Group or Cretaceous Shale. Wind-deposited sands of the Nebraska Sand Hills overlie about 6.4% of the study area. Quaternary-age deposits have sufficient saturated thickness to be developed as a source of groundwater in most of the ULEN area, with an average thickness of 237 ft (72.24 m) and can be as much as 430 ft (131.06 m) thick. It is thicker in the northern and southern parts of the ULEN model area and thin or absent in the river or tributary valleys. The Quaternary-age deposits usually are the coarsest deposits found in the study area and can support sustainable pumping rates in excess of 1,000 gallons per minute (gal/min) (0.06 m³/s) (DNR, 2005).

**Alluvial Deposits**

Alluvial deposits occur in the major Elkhorn River Valley, as well as in the North Fork Elkhorn River valley. The river valley area accounts for 8.6% of the study area. Their widths vary from 1 to 2 miles (1.61 ~ 3.22 km). In contrast to the Quaternary deposits in the upland, the alluvial sediments are dominantly sand and gravel and may contain small silt and clay lenses. Beneath the alluvial deposits is the Cretaceous shale. Because of the lack of consolidation and coarse grain sizes, the hydraulic conductivity of
the alluvial sediments is high. The depth to groundwater in river alluvium is relatively shallow.

*The Ogallala Group*

The Ogallala Group deposits are present in most of the study area and are composed of clays, silts, sands, gravels, and poorly consolidated sandstone and siltstone. Test-hole records show that the Ogallala Group is widespread in the study area except at Elkhorn River valley and the southeastern part of the model area where the Ogallala had been eroded away. Generally the saturated thickness of this group is more than 90 ft (27.43 m) at a great distance away from Elkhorn River and become thinner in the vicinity of the River. Maximum Ogallala Group thicknesses described in test holes in the ULEN area were around 422 ft (128.63 m), with an average thickness of about 110 ft (33.53 m). At some points, the saturated thickness of this group is only several feet due to the high elevation of Cretaceous bedrock. The base elevation of the Ogallala Group is higher in the western and lower in the eastern part of the ULEN area with the highest base elevation from 1735.76 ft (529.06 m) to the lowest base elevation 1343.288 ft (409.43 m). Because of compaction and cementation, the Ogallala Group is partially consolidated in the study area which accounts for a lower hydraulic conductivity than the unconsolidated alluvial sediment.

**4.4 Water Use and Management**

Water resources in the ULEN area are used for a variety of purposes: domestic, irrigation, livestock, mining and industrial. The unit in Table 5 is million gallons per day. Irrigation took the largest portion of the water use in the study area. Groundwater use
took 98% of the total use and surface water took only 1.6% of the total budget (Figure 4-3). It again reflects the importance of groundwater resources in the study area.

Table 4-1 Groundwater use and surface water use in UPLN model (unit for the water use rate is in gallons per day).

<table>
<thead>
<tr>
<th>Percent (%)</th>
<th>Thermo-electric power</th>
<th>Mining</th>
<th>Industrial</th>
<th>Aquaculture</th>
<th>Livestock</th>
<th>Domestic</th>
<th>Irrigation</th>
<th>Public supply</th>
<th>ULEN</th>
<th>Groundwater</th>
<th>Surface water</th>
<th>Total</th>
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<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>Domestic</td>
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<td>0.82</td>
<td>0</td>
<td>0</td>
<td>0.41</td>
<td>0.41</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Irrigation</td>
<td>4.29</td>
<td>0.82</td>
<td>0</td>
<td>0</td>
<td>0.41</td>
<td>0.41</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<tr>
<td>Public supply</td>
<td>4.06</td>
<td>0.82</td>
<td>0</td>
<td>0</td>
<td>0.41</td>
<td>0.41</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<td>0</td>
</tr>
<tr>
<td>ULEN</td>
<td>8.06</td>
<td>0.82</td>
<td>0</td>
<td>0</td>
<td>0.41</td>
<td>0.41</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Groundwater</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>427.5</td>
<td>0.41</td>
<td>434.6</td>
<td>98.4</td>
</tr>
<tr>
<td>Surface water</td>
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<td></td>
<td></td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>434.6</td>
<td>1.62</td>
<td>439.6</td>
<td>100</td>
</tr>
</tbody>
</table>
4.4.1 Surface water use

The surface water mainly consists of river, creeks, and canals in the study area. Three major perennial streams are the Elkhorn River, North Fork Elkhorn River and Willow Creek. There are five stream stations in the study area (Table 4-2):

Table 4-2 List of stream gage stations in ULEN model area.

<table>
<thead>
<tr>
<th>Site</th>
<th>Station Number</th>
<th>River</th>
<th>Agency</th>
</tr>
</thead>
<tbody>
<tr>
<td>Neligh</td>
<td>06798500</td>
<td>Elkhorn</td>
<td>DNR</td>
</tr>
<tr>
<td>Tilden</td>
<td>06798780</td>
<td>Elkhorn</td>
<td>DNR</td>
</tr>
<tr>
<td>Norfolk</td>
<td>06799000</td>
<td>Elkhorn</td>
<td>USGS</td>
</tr>
<tr>
<td>Hadar</td>
<td>06799100</td>
<td>North Folk Elkhorn</td>
<td>USGS</td>
</tr>
<tr>
<td>Pierce</td>
<td>06799080</td>
<td>Willow Creek</td>
<td>DNR</td>
</tr>
</tbody>
</table>

Other creeks are seasonal streams that carrying only intermediate runoff from precipitation to discharge the adjacent aquifer when groundwater level is higher than the bottom of the creeks. As of October 1, 2005, there are approximately 550 surface water appropriations in the basin issued for a variety of uses. The majority of the surface water
appropriations are for irrigation and they tend to be located on the major streams (Bleed, 2005).

4.4.2 Groundwater use and flow

Groundwater in the ULEN (Figure 4-4) area generally flows from west to east with an average water-table slope of about 7 ft /mi (1.26 m/km). The water-table gradient tends to be evenly distributed through the model area. The highest groundwater elevation is around 1900 ft (579.12 m) at southwest corner of the model area and lowest groundwater elevation is about 1450 ft (441.96 m) at the east side where the outlet of the Elkhorn River is. From the water table graph, apparently the Elkhorn River gains water from the aquifer system and the groundwater also discharges to some creeks. Groundwater depth (Figure 4-5) becomes much shallower toward the Elkhorn River and other creeks in the study area. Depth to groundwater affects infiltration and evapotranspiration. Both groundwater level data and depth to water data that were used to generate graph (Figure 4-4 and Figure 4-5) were downloaded from School of Natural Resources web site at University of Nebraska Lincolns:

http://snr.unl.edu/data/geographygis/NebrGISwater.asp#wtable.
Figure 4-4 Contour map of groundwater level in 1995, the number along the contour line has a unit in ft (data source: http://snr.unl.edu/data/geographygis/NebrGISwater.asp#wtable).

Figure 4-5 Configuration of depth to water in 1979 (data source: http://snr.unl.edu/data/geographygis/NebrGISwater.asp#depth).
There are about 7990 (Figure 4-6) registered groundwater wells which are almost evenly spread in the study area. Very few wells are constructed in the north and south bound of the model area, since these areas are the upland area and the principal aquifer thickness is smaller than 100 ft (30.48 m). There is a zone where pumping wells are intensely developed while the aquifer thickness is thin. It is because this area is also located in the river valley area that the pumping wells obtain large amounts of water from the Elkhorn River.

Figure 4-6 Pumping wells used in the well package.
4.5 Conceptual Groundwater Flow Model

A conceptual model of groundwater flow is a qualitative framework upon which data related to subsurface hydrology can be organized and integrated. A conceptual model is indeed a water balance calculation. The basic components of a conceptual model are the physical boundaries and distribution of hydraulic properties in the region and water budget.

Boundary can be divided into lateral boundary which accounts for the water fluxes into or out of the system through the model boundary; areal boundary included in this model are recharge from precipitation and irrigation, and groundwater evapotranspiration from rooted plants located in riparian zones along the Elkhorn River; internal boundary was represented by the river that exchanges water with aquifer depending on the hydraulic gradient and pumping well that extracted groundwater during irrigation season.

4.5.1 Hydrogeologic units

The construction of hydrostratigraphic model has been discussed in detail in Chapter 3. The first unit was defined as unconfined aquifer type which consisted mainly of Quaternary deposits in the study area. The second unit was categorized into an unconfined/confined aquifer type which consisted of low hydraulic conductivity deposits such as loess, till and clay. Low $K$ deposits were not extended through the model domain. The third unit was defined as a confined layer type which was mainly composed of the Tertiary Ogallala Group and upper Cretaceous deposits at several locations. For each unit, the hydraulic conductivity was calculated and grouped into hydraulic property zones.
Here, we will introduce how the elevation of each hydrostratigraphic unit was determined through Kriging interpolation method.

Control points: The observation wells from USGS, registered wells from DNR and test hole from CSD in the study area were selected as control points (Figure 4-7). The information of elevation, coordinates and lithological records to depths were provided from these control points. The elevation of each control points that extracted from DEM were used to interpolate the top elevation of the first unit. CSD test holes also recorded stratigraphic information that described the depth of Quaternary (Q) and Tertiary (T) deposits. The geo logs of registered wells were also grouped to define the separation elevation of Q and T deposits. The separation elevation of these points were used to calculate the separation layer of QT through the model domain. Then elevation of the top layer of second unit was defined by the separation layer of QT plus 10 ft (3.05 m) and the elevation of the bottom layer of second unit minus 10 ft (3.05 m). This was done because the low $K$ unit was not extended through the model area and average thickness of the low $K$ deposits from the geo log was around 20 ft (6.10 m). The low $K$ deposits sometimes are above the QT separation line while sometimes are below the separation line according to the geo log record. The elevation of the bottom layer of the third unit was calculated based on the elevation of the shale deposit in the test hole record.

Variogram: The software Surfer 10 (Golden, 2011) was used to generate variogram and created the surface grid. The spatial variability of the layer elevation was analyzed through fitting the semivariogram of the synthesized elevation results with theoretical variogram models. For a two dimensional dataset, the semivariogram is:

$$\gamma(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} [z(u) - z(u+h)]^2$$  \hspace{1cm} (Eq. 4-1)
where \( \gamma(h) \) is the semivariogram; \( h \) is the distance interval along all or some specified directions; \( N(h) \) is the number of pairs of data points with such distance interval \( h \); and \( z(u) \) and \( z(u+h) \) are a pair of values with the distance interval \( h \).

Figure 4-7 Control points for Kriging interpolation.

Kriging algorithm

The Kriging algorithm was used to interpolate the elevation of each layer based on the values at each control point according to the fitted variogram model of each layer. The fitted variograms (Figure 4-8) for the top elevation of Unit 1, and top and bottom elevation of Unit 2, and bottom of Unit 3 are shown in Figure 8. The black dot line is the
variogram directly calculated with Eq (4-1) based on the elevation at each control points (or experimental variogram). The blue solid line is the fitted theoretical variogram model. The equations of fitted theoretical models are also shown in Figure 5. The models of (a), (b) and (c) in Figure 5 belong to the nonstationary power model, while (d) detrended before the variogram calculation, belong to the stationary exponential model.

![Figure 4-8 Fitted variogram for the four layers.](image)

### 4.5.2 Recharge

Recharge is a hydrologic process where water moves downward from surface water to groundwater until it reaches the regional groundwater table. Recharge in the
conceptual model was considered from two sources: one was from precipitation and the other was from irrigated water on the cropland during pumping activity. Before the water reaches the groundwater table, it undergoes a number of processes in the hydrologic cycle, such as canopy interception, runoff generation, and movement in unsaturated zones, etc. Several software packages were developed to simulate the flow movement in this variably saturated porous media such as Hydrus 1D (Šimůnek 1998). However, for simplicity, this study used an equation to represent the amount of water that reached the groundwater table as a ratio to the total amount of water that presented on the land surface.

\[ R = \alpha \times Q \]  

\text{Eq. 4 - 2}

where \( R \) is the flux of groundwater recharge, \( \alpha \) represents the percentage amount of water that reaches the groundwater table, and \( Q \) is the total amount of water presents at the land surface which is represented by equation \( Q = I + P \); where \( I \) is the irrigation water from pumpage and \( P \) is the precipitation in ft/d. Recharge from precipitation was applied to every cells in the study area, while irrigation water was only applied to the place where cropland presents.

4.5.3 Groundwater evapotranspiration

Groundwater Evapotranspiration (ET) refers to the transpiration of plants and evaporation of groundwater near the land surface. In past studies, groundwater ET was usually neglected due to the minimal effect on the total water budget. However, in ULEN area the groundwater ET can remove large amounts of water where the water table is near to the land surface. Normally the groundwater ET occurred in wetland and riparian areas or along river valleys. It also possibly happens on the other land use type which depends on the water table level. Pederson (2008) assumed a uniform evapotranspiration rate of
Chen and Shu (2006) estimated a groundwater ET rate about 490 ~ 660 mm/year with an extinction depth about 15 ft (4.57 m) in the Republican River valley. In this study, a uniform extinction depth at 15 ft (4.57 m) was used for the whole domain. ET rate at 27 in/yr was set to in the wetland area, an ET rate was referenced from Szilagyi’s research (Szilagyi et al., 2011) who reported an ET rate at 17 in/yr in the Sand Hills regions. This value was assigned to the type of land use of Sand Hills in the study area. ET rate of 14 in/yr from Peterson’s research was set for the other types of land use. Since the groundwater ET rate reached the highest value during summer, each ET rate was accordingly divided into 5 parts from May to September based on the temperature ratio. Assumption of no occurrence of groundwater ET was made in the other seven months.

4.5.4 River

Stream Width and Length

The stream was generated through delineating a 30 meter USGS DEM. The hydrology extension tool in ArcGIS was used to create the stream network and calculate the stream width and stream length. The threshold value was set 20000 acres which meant the river that had a drainage area that were larger than 20000 acres will be delineated.

Stream Stage

In this model, the stream has a drainage area that is larger than 20000 acres are the main stem of the Elkhorn River and North Fork Elkhorn River. Willow Creek was considered since Pierce the study site was located nearby and the stream stage data are also available from DNR. There are five hydrologic gage stations along the main course of the Elkhorn River and one along the North Fork Elkhorn River and the other located
on Willow Creek. In each stress period, the river stage between gage stations is linearly interpolated based on the observed stream stage levels at the adjacent upstream and downstream gage stations. The calculation equation is as follows (Equation 4-3):

\[
H_{RIV} = H_{RIV1} + \frac{L_1(H_{RIV2} - H_{RIV1})}{(L_1 + L_2)}
\]

where \(H_{RIV}\) is the river stage at a grid cell; \(H_{RIV1}\) and \(H_{RIV2}\) are river stages at the adjacent upstream gage and downstream gage, respectively; \(L_1\) and \(L_2\) are the lengths of river sections from the calculated river grid cell to the adjacent upstream gage and downstream gage, respectively. Figure 4-9 shows the river stages along the main course of the Elkhorn River in Dec, 2010. The stages at five gage stations were nearly linearly distributed with the corresponding downstream river lengths from the inlet point of the Elkhorn River at the boundary of the study area. The overall slope of the river stage was 0.992\%.
Streambed hydraulic conductivity

The streambed hydraulic conductivity was assigned to each grid based on the field in-situ permeameter tests and Geoprobe test (Chen and Lackey, 2010; Lackey and Chen, 2010). The equivalent $K$ was calculated using the experimental $K$ value of the core up to 50 ft (15.24 m) deep that was collected by Geoprobe. From Figure 4-10, each color represents a unique hydraulic conductivity value. Thus, the streambed conductance could be calculated as Equation 4-4:

$$C_{RIV} = \frac{KbLW}{M}$$  \hspace{1cm} Eq. 4 – 4

where $C_{RIV}$ is the conductance of streambed [L2/T]; $K$ is the streambed conductivity
$L$ is the stream length [L]; $W$ is the stream channel width [L]; and $M$ is the streambed thickness [L]. The exchange between the stream and aquifer can then be computed on the basis of the conductance and the difference of the stream stage and groundwater level.

Figure 4-10  Lateral boundary condition in the UPLN model, the unit of streambed hydraulic conductivity value is in ft/d.
4.5.5 Pumping well

A total of 7990 pumping wells were developed before 2010 in the study area (Figure 4-6). The well type included the irrigation wells for crops, domestic well for human living, municipal well for city water supply and livestock well for animal life on the farm. Different type of well was set to pump water for different period. For example: I denotes irrigation well which was set to pump water during June, July and August; D denotes domestic well which was set to pump water through whole year. However, logging of the pumping rate start after 2003; it is hard to estimate the real pumpage before this data. Excluding other types of well, more than 97% are irrigation wells in the study area. Thus, the groundwater pumpage will be calculated using crop irrigation requirement and pumping capacity of each well.

The irrigated area and the water pumping discharge capacity of each well can be obtained from DNR’s registered well database. A ratio of the actual pumpage and the capacity could be calculated if the irrigation requirement is known during the crop growing seasons. The amount of groundwater pumped from the aquifer was assumed to be equal to the irrigation requirement, and then we can calculate the pumping ratio as follows (Equation 4-5):

\[ \beta = \frac{\alpha \times (R - P) \times \sum_i^n A_i}{\sum_i^n (C_i T)} \]

where \( \beta \) is the pumping ratio of the actual pumpage to pumping capacity in the study area; \( \alpha \) is a unit conversion factor which is equal to 18.86; \( R \) is the annual irrigation requirement in this area, (inch); \( n \) is the total number of irrigated wells in this area; \( C \) is the average pumping capacity of the \( i \)'th wells in this area (GPM); \( T \) is the time of the irrigation period which we assumed to be 90 days in June, July, August every year; \( A \) is
the irrigated acres of the $i^{th}$ well in this area, (acre). $\beta$ was estimated to be 0.37 using Eq (2) based on the properties of each well in the study area and the annual net irrigation requirement shown in Figure 4-11, which was compiled by DNR (http://dnr.ne.gov/SurfaceWater/CountyMapIrrReq.pdf)

![Figure 4-11 Annual net irrigation requirement of corn in Antelope, Pierce and Madison. The number in the graph has a unit of inch (data source: http://dnr.ne.gov/SurfaceWater/CountyMapIrrReq.pdf).](image)

4.6 Groundwater Flow Model Simulation

4.6.1 Space and time discretization

MODFLOW employs a finite-difference method to solve the partial-differential equation governing the groundwater flow. The method discretizes the space into a grid framework composed of rectangular blocks called cells; and divides simulation time into stress periods, in which the external stresses are constant. The length and time units of the ULEN model are foot and day, respectively.
The ULEN model was horizontally subdivided into 180 rows by 201 columns, and vertically into 3 layers. The first layer represented the Quaternary deposits, the third layer symbolized the Tertiary deposits and the second layer represented a low $K$ layer that was displayed between Quaternary deposits and Tertiary deposits. The cell sizes were 1320 ft (402 m) by 1320 ft (402 m) through all domains. The interpolated elevations of each model layer were assigned to the center of each grid cell. The model was composed of three layers. To account for the change of confined and unconfined conditions at some location where hydraulic head significantly drops, we used Type 3 aquifer layer (McDonald and Harbaugh, 1988) to simulate each model layer. In MODFLOW, for Type 3 aquifer layer, "transmissivity of the layer varies and is calculated from the saturated thickness and hydraulic conductivity. The storage coefficient may alternate between confined and unconfined values. Vertical flow from above is limited if the aquifer desaturates" (Harbaugh, 2005).

The model was developed for transient conditions and the time was divided into discrete time intervals called stress periods. The simulation period extended from January 1, 1980 to January 1, 2010, for a total 30 years and 360 monthly stress periods. Simulated hydrologic stresses, such as recharge, pumpage and river stages, were updated between stress periods. The hydraulic properties, such as conductivities, specific storages and specific yields, were kept unchanging during the whole simulation. Each stress period was subdivided into a number of calculation time steps. In the ULEN model, there were 12 time steps in each stress period. The time steps within each stress period started with a minimum time 0.5 days and increased with a geometric ratio of 1.2 in this model. The
length of the early time step was shorter, because the updated external stress of a new stress period can cause rapid change of the hydraulic heads.

4.6.2 Initial condition

A total of 187 wells with 29554 observation records were selected for the model calibration and each well had more than 10 observation records. Initial heads were estimated using the same interpolation algorithm as elevation did. The control point for initial head was collected from USGS groundwater monitoring site, DNR groundwater monitoring site and test holes. Only the records from 1980 to 1987 were averaged and used to interpolate the initial groundwater elevation since the number of irrigation wells remained stable during this period. Figure 4-12 shows the contour map of the initial head and the fitted variogram model, which is a nonstationary power model.
4.6.3 Boundary condition

*Latera boundary condition*

Lateral boundary condition (Figure 4-13) includes the constant head boundary (CHD) and no flow boundary. We used CHD to represent the whole model boundary except for the sections where the no-flow boundary was assigned due to the water table contours almost perpendicular to the model boundary. The brown color in the graph represents the CHD boundary while the grey color grid represents the no-flow boundary.
Areal Boundary

Recharge

The study area was divided into a number of recharge zones to estimate the contribution of precipitation plus irrigation to recharge. Each recharge zone was assumed to represent an area with a uniform recharge ratio, that is, $\alpha$ in Eq (4-2). However, considering the land use and irrigation development kept changing during the modeling period, it is not reasonable to use a constant $\alpha$ to represent the infiltration processes throughout the simulation time. Figure 4-13 indicates that the number of the groundwater wells kept growing, implying the increasing development of irrigated agricultural lands in this area. In the study area, the well numbers were stable between 1980~1987 and then kept growing almost at a constant rate. Thus, the modeling time was split into five phases: 1980~1987, 1988~1995, 1996~2000, 2001~2005 and 2006~2010. As a result, each recharge zone had five infiltration ratios corresponding to these phases. These infiltration rates are difficult to predict and thus will be manually calibrated.
Figure 4-13 Accumulative development of pumping wells in the Elkhorn River Basin from 1980 to 2010.

According to the land use and land cover, the recharge zones (RCH) were defined as shown in Figure 4-14. RCH was adjusted subsequently according to the density of groundwater wells. The Precipitation Recharge Zone 1 (RCH-1) represented irrigated cropland such as corn, soybean, potatoes and etc. Through examining the history of groundwater well development on the irrigated farms, there were considerable differences at different phases, thus the infiltration ratio was set to be different through the whole timeline. RCH -2 represented the land type as range, pasture and grass land which has a reported infiltration rate about 2.76 inch/year (70.10 mm/year) (Gurdak et al., 2007). RCH -3 included urban land, road and open water where the infiltration rate is stable and low through the whole simulation time. RCH -4 represented the wetland and riparian forest along the river where the infiltration is considered being the highest. PRZ-
5 represented the non-irrigated cropland such as dry land soybeans and dry land sunflower.

Evapotranspiration

Evapotranspiration was assumed to occur through all domains and the parameter was set according to description in the conceptual model. In the ULEN model, the groundwater evapotranspiration was divided into three zones (Figure 4-15): wetland and riparian (blue color), the Sand Hills (green color) and other types of land use (cyan color) as showed in Figure 4-15. The actual groundwater evapotranspiration was estimated with the evapotranspiration package (EVT) of MODFLOW (McDonald and Harbaugh, 1988).
It assumes that evapotranspiration varies linearly with water-table elevation, and reaches the maximum rate at the land surface (Equation 4-6):

\[ Q_e = \begin{cases} 
PET, & h > \text{SURF} \\
PET \times \frac{h-(\text{SURF}-\text{EXDP})}{\text{EXDP}} & (\text{SURF} - \text{EXDP}) \leq h \leq \text{SURF} \\
0, & h < \text{SURF} 
\end{cases} \quad \text{Eq. 4 - 6}

where \( Q_e \) is the rate of groundwater ET; \( PET \) is the potential groundwater ET; \( SURF \) is the ET surface elevation which coincides the land surface in the model; \( EXDP \) is the extinction depth below which the groundwater ET ceases; and \( h \) is the groundwater table. In this model, an extinction depth of 15 ft (4.57 m) was assigned to all domains.
4.7 Model Results

4.7.1 Water budget

During the model calibration, the calculated head values at those groundwater monitoring wells were manually calibrated to match the observed groundwater levels around study sites: Neligh, Meadow Grove, Norfolk, Hadar and Pierce. In Table 6, the cumulative water budget from 1980 to 2010 was listed in different terms. The “STORAGE” term refer to the water storage change of the aquifer. The percentages are
the ratio of the term and total inflow or outflow. In the inflow terms, recharge is the primary water resource for the groundwater system; while the groundwater pumping and river leakage are most important in the discharge terms. The percent discrepancy between the total inflow and outflow is about 0.03%, indicating the numerical solution preserved good precision.

### Table 4-3 Cumulative water budget of the ULEN model.

<table>
<thead>
<tr>
<th>Term</th>
<th>FLOW IN (FT³)</th>
<th>Percentage</th>
<th>FLOW OUT (FT³)</th>
<th>Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>STORAGE</td>
<td>1.94E+11</td>
<td>0.30</td>
<td>2.45E+11</td>
<td>0.38</td>
</tr>
<tr>
<td>CONSTANT HEAD WELL</td>
<td>2.33E+10</td>
<td>0.04</td>
<td>4.79E+10</td>
<td>0.07</td>
</tr>
<tr>
<td>WELL</td>
<td>0.00E+00</td>
<td>0.00</td>
<td>1.99E+11</td>
<td>0.31</td>
</tr>
<tr>
<td>RIVER LEAKAGE ET RECHARGE</td>
<td>8.25E+09</td>
<td>0.01</td>
<td>1.40E+11</td>
<td>0.22</td>
</tr>
<tr>
<td>ET</td>
<td>0.00E+00</td>
<td>0.00</td>
<td>5.68E+09</td>
<td>0.01</td>
</tr>
<tr>
<td>RECHARGE</td>
<td>4.17E+11</td>
<td>0.65</td>
<td>0.00E+00</td>
<td>0.00</td>
</tr>
<tr>
<td>TOTAL</td>
<td>6.43E+11</td>
<td>1.00</td>
<td>6.43E+11</td>
<td>1.00</td>
</tr>
</tbody>
</table>

Figure 4-16 shows the average monthly water flow rate of each term. The cylinder that has negative values in the graph represents the outflow while positive values are inflow. The seasonal variability of recharge and well pumping is pronounced because of the irrigation season and temporal distribution of rainfall in this area. Inflow and outflow of river leakage are relatively stable due to the small change of river stage and groundwater level.
4.7.2 Groundwater levels

Due to the time limitation, only the wells that near the interested sites was calibrated: Neligh, Norfolk, Meadow Grove, Pierce and Hadar. The mean absolute error between simulated head and observation head are 1.7 ft (0.52 m). Figure (4-17) shows the good fit of these wells.
Figure 4-17 Simulated hydraulic heads vs. observed hydraulic heads.
4.7.3 Streamflow depletion analysis

Another important objective of this study is to analyze the streamflow depletion of the Elkhorn River. The stream depletion is defined as the reduction of streamflow due to induced infiltration of stream water into the aquifer and capture of aquifer discharge to the stream (Bredehoeft, 1997; Sophocleous, 1997). In the early studies, the stream depletion was mainly calculated by analytical solution. Theis (1941) proposed a transient method to evaluate the impact that groundwater pumping on a nearby stream. Studies by Glover and Balmer (1954) and Jenkins (1968) all assumed the streams fully penetrated the aquifers without stream clogging which has many limitations. Because conventional assumption of a fully penetrating stream will lead to a significant overestimation of stream depletion (>100%) in many practical applications (Butler et al., 2001). Glover (1974) proposed a stream-aquifer analytical solutions which are commonly used in water resources management. Recently, Hunt (1999) proposed an analytical depletion method that can be used for streams with a clogging layer that only partially penetrates the aquifer. Numerical modeling techniques were also widely applied to evaluate the stream-aquifer interaction. The code MODFLOWP ((Hill, 1992) was used by Nyholm et al. (2002) to analyze the stream depletion. Chen and Shu (2002) also applied numerical modeling techniques to simulate stream-aquifer interactions from seasonal groundwater pumping.

In this study, the stream depletion was analyzed by MODFLOW based on numerical modeling techniques. A method in Chen and Yin (1999) was used to calculate the stream depletion ratio at Neligh and Hadar sites. At each interested site, a hypothetical well was put in the model at 1000 ft (304.8 m), 1 mile (1.61 km) and 3 miles (4.83 km) away from the river and pumped water at a rate of 1000 GPM for June, July
and August through 30 years. Then the rate of stream leakage, reduction of stream baseflow is extracted from the MODFLOW output file and stream depletion ratio is calculated and compared at each site. The depletion ratio is computed as (Equation 4-7):

\[ D = \frac{Q_{riv_a} - Q_{riv_b}}{Q_p} \quad \text{Eq. 4 – 7} \]

where \( D \) is the depletion ratio of the new well; \( Q_{riv_a} \) is the exchange rate between streams and aquifers before pumping and \( Q_{riv_b} \) is the exchange rate between streams and aquifers after adding the hypothetical well. The exchange rate is indeed the combined effect of stream leakage and reduction of stream baseflow. \( Q_p \) is the total pumpage of depletion well.

In this study, we first run MODFLOW to generate the distribution of groundwater head, and then run the zone budget module to obtain the flow rate and the volume of each component in the groundwater system. Finally we extract these two components from MODFLOW output file to calculate the streamflow depletion ratio.

Figure 4-18 shows that: 1) At the Neligh site, if the pumping well was located about 1,000 ft (304.8 m) away from the river, the maximum stream depletion is less than 15% of the total pumpage; 2) if the distance between pumping well and river increases to 1 mile (1.61 km), the maximum stream depletion is less than 12%; 3) if the distance between wells and the river increases to 3 miles (4.83 km), the stream depletion decreases to about 4% of the pumping rate.

In the contrary, Figure 4-19 shows that: 1) At the Hadar site, if the pumping well was located about 1,000 ft (304.8 m) away from the river, the maximum stream depletion is less than 6% of the total pumpage; 2) if the distance between pumping wells and river increases to 1 mile (1.61 km), the maximum stream depletion is less than 4%; 3) if the
distance between wells and the river increases to 3 miles (4.83 km), the stream depletion decreases to about 2% of the pumping rate.

Figure 4-18 Depletion at the Neligh site under three pumping scenarios.

Figure 4-19 Depletion at the Hadar site under three pumping scenarios.
Chapter 5 Conclusions

The stream-aquifer connection was characterized in the Elkhorn River Basin at eight study sites in this thesis. The hydraulic gradient between stream stage and groundwater level in the adjacent aquifer showed that the Elkhorn River is a gaining river at all study sites. The streambed conductance in upstream of the Elkhorn River is generally higher than the downstream of the Elkhorn River (Figure 5-1). (The separation line of upstream and downstream of the Elkhorn River is same as the boundary between Upper Elkhorn NRD and Lower Elkhorn NRD). This result reflects the depositional environments gradually changing from eolian in the west to the glacial deposits in the east.

![Figure 5-1 Comparison of averaged vertical hydraulic conductivity value between Upper ENRD and Lower ENRD.](image)

Cross correlation analysis between stream stage and groundwater level showed that the highest correlation coefficient is 0.762 when the stream stage lags 1 day at the Neligh site. Two lowest correlation coefficients were found at the two tributaries of the
Elkhorn River at the Pierce and the Hadar site, respectively. This result is consistent with the calculated streambed conductance which is highest at the Neligh site and lowest at the Hadar site. The groundwater temperature variation is in response to the seasonal air temperature change and it is found that the peak value of groundwater temperature is lagged compared to the peak value of air temperature. This phenomenon is mainly attributed to the heat transfer process through the deposits. The average lag time is about 4 months with the longest time of 7 months and shortest time of 3 months.

The geologic model was created by interpolating hydrofacies value from test hole data using Inverse Distance Weighting method. It is found that the aquifer unit is thicker in the west, and gradually becomes thinner toward the east. In some other parts of Nebraska, there is usually a layer of clay and silt deposits displayed between the Quaternary aquifer and the Ogallala Aquifer. However, in the Elkhorn River Basin the geologic model showed that this low $K$ layer does not continuously extend through the model area. The nodes in the geologic model sufficiently represent the heterogeneity of the subsurface condition. During the upscaling zonal process, the equivalent hydraulic conductivity was calculated and categorized into hydrogeological property zone to symbolize the hydrogeological condition. Thus, this geologic model has a good representation of the subsurface deposits in the study area.

The generated hydrogeological property zone was imported into groundwater flow model. The climate data, land use data, stream stage data, groundwater level data and pumping well data were used for the development of the ULEN groundwater flow model. After the completion of the ULEN model calibration, a hypothetical well was added to the model away from the river at 1000 ft (304.8 m), 1 mile (1.61km) and 3 miles
(4.83 km) to analyze the stream depletion at the Neligh site and the Hadar site. It is observed that the stream depletion decreases rapidly as the distance between pumping well and river decreases. The stream depletion can be neglected when the distance between pumping well and river is close to 3 miles (4.83 km) and farther. It is also found that the depletion ratio is larger at the Neligh site than the depletion ratio at the Hadar site when the pumping well is located at the same distance away from the river. The result suggests that the level of connectedness between stream and adjacent aquifer at Neligh site is higher than it at Hadar site which is also consistent with the result of cross-correlation analysis. Considering that the observation well was only calibrated around the study sites and also the associated model uncertainties, the ULEN will need to be further calibrated in the future.
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namespace K
{
    class Program
    {
        static int startIndex = 1, minimumCapacity = 200000;
        const int xMin = 1826920, xMax = 2133160, xInterval = 2640, zInterval = 2;
        const int yMin = 15057920, yMax = 15482960, yInterval = 2640;
        const int gridCntX = 2, gridCntY = 2;

        const string connectionStringK = "Provider=Microsoft
(m).ACE.OLEDB.12.0;Data Source=D:\data\Beta_final8.accdb;Persist Security
Info=False;";
        const string connectionStringLayer = "Provider=Microsoft
(m).ACE.OLEDB.12.0;Data Source=D:\data\layer_grid.accdb;Persist Security
Info=False;";

        static string GetOutputPath()
        {
            string path = "D:\output_" + DateTime.Today.Day + "_" +
                DateTime.Now.Hour + "_" + DateTime.Now.Minute + ".txt";
            return path;
        }

        static DataTable GetNodes(int X, int Y, double minZ, double maxZ)
        {
            //Console.WriteLine("I am getting nodes" + DateTime.Now.ToString());
            OleDbConnection cn = new OleDbConnection(connectionStringK);
            //string command = string.Format("SELECT * FROM Cubic WHERE Field1 =
            //\{0\} and Field2 = \{1\} and Field3 >= \{2\} and Field3 <= \{3\}" , X, Y, minZ, maxZ);
            string command = string.Format("SELECT * FROM Final WHERE Field1 = \{0\}
            and Field2 = \{1\} and Field3 >= \{2\} and Field3 <= \{3\}" , X, Y, minZ, maxZ);
            OleDbDataAdapter da = new OleDbDataAdapter(command, cn);
            DataTable dataTable = new DataTable();
            dataTable.MinimumCapacity = minimumCapacity;
            try
            {
                cn.Open();
                dataTable.BeginLoadData();
                da.Fill(dataTable);
                dataTable.EndLoadData();
            }
            catch (OleDbException e)
{  
    string msg = "";
    for (int i = 0; i < e.Errors.Count; i++)
    {
        msg += "Error #" + i + " Message: " + e.Errors[i].Message + "\n";
    }
    System.Console.WriteLine(msg);
}
finally
{
    if (cn.State != ConnectionState.Closed)
    {
        cn.Close();
    }
    return dataTable;
}
static DataTable GetNodes(List<int> X, List<int> Y, List<double> minZ, List<double> maxZ)
{
    DataTable dataTable = new DataTable();
    dataTable.MinimumCapacity = minimumCapacity;
    string query = "";
    OleDbConnection cn = new OleDbConnection(connectionStringK);
    for (int i = 0; i < X.Count; i++)
    {
        string command = string.Format("SELECT * FROM Final WHERE Field1 = {0} and Field2 = {1} and Field3 >= {2} and Field3 <= {3}", X[i], Y[i], minZ[i], maxZ[i]);
        if (i != 0) query = query + " Union ";
        query = query + command;
    }
    OleDbDataAdapter da = new OleDbDataAdapter(query, cn);
    try
    {
        cn.Open();
        da.Fill(dataTable);
    }
    catch (OleDbException e)
    {
        string msg = "";
        for (int i = 0; i < e.Errors.Count; i++)
        {
            msg += "Error #" + i + " Message: " + e.Errors[i].Message + "\n";
        }
        System.Console.WriteLine(msg);
    }
    finally
    {
        if (cn.State != ConnectionState.Closed)
        {
            cn.Close();
        }
    }
}
static List<double> GetLayers(int x, int y, string top, string bottom) {
    List<double> res = new List<double>();
    OleDbConnection cn = new OleDbConnection(connectionStringLayer);
    string command = string.Format("SELECT * FROM Layer_Updated_4_28 WHERE x={0} and y={1}", x, y);
    OleDbDataAdapter da = new OleDbDataAdapter(command, cn);
    DataTable dataTable = new DataTable();
    try {
        cn.Open();
        da.Fill(dataTable);
    }
    catch (OleDbException e) {
        string msg = "";
        for (int i = 0; i < e.Errors.Count; i++)
            { msg += "Error #" + i + " Message: " + e.Errors[0].Message + "\n"; }
        System.Console.WriteLine(msg);
    }
    finally {
        if (cn.State != ConnectionState.Closed) {
            cn.Close();
        }
    }
    DataRow[] rows = dataTable.Select();
    res.Add((double)rows[0][top]);
    res.Add((double)rows[0][bottom]);
    dataTable.Clear();
    dataTable.Dispose();
    dataTable = null;
    //Console.WriteLine(string.Format("x:{0} y:{1} zmin: {2} zmax: {3}", x,y,res[0],res[1]));
    return res;
}
static List<double> GetAverageLayers(int minX, int maxX, int minY, int maxY, string top, string bottom) {
    List<double> res = new List<double>();
double z = 0, z2 = 0;

for (int i = minX; i <= maxX; i += xInterval)
{
    for (int j = minY; j <= maxY; j += yInterval)
    {
        List<double> tmp = GetLayers(i, j, top, bottom);
        z += tmp[0];
        z2 += tmp[1];
    }
}
res.Add(z / (gridCntX * gridCntY));
res.Add(z2 / (gridCntX * gridCntY));
return res;

static double KXPlus(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KXPlus: ");
    double k = 0;
    for (int x = minX; x <= maxX; x += xInterval)
    {
        string command = string.Format("Field1 = {0}", x);
        DataRow[] rows = dataTable.Select(command);
        //Console.WriteLine("size: {0}", rows.Length);
        double tmpk = 0;
        for (int i = 0; i < rows.Length; i++)
        {
            //Console.WriteLine(string.Format("Field1: {0} Field2: {1} Field3: {2} Field4: {3}", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));

            double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
            tmpk += kx;
        }
        k += 1 / (tmpk * zInterval);
    }
    k = k * deltaZ;
    k = 1 / k;
    return k;
}

static double KXNegative(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KXNegative: ");
    //Console.WriteLine(string.Format("xmin: {0}, y", k));
    double k = 0;
    for (int y = minY; y <= maxY; y += yInterval)
    {
        for (int z = minZ; z <= maxZ; z += zInterval)
        {
            string command = string.Format("Field2 = {0} and Field3 = {1}" , y, z);
            DataRow[] rows = dataTable.Select(command);
            //Console.WriteLine("size: {0}", rows.Length);
            if (rows.Length == 0) continue;

            double kx = Math.Pow(Math.E, (double)rows[0]["Field4"]);
            tmpk += kx;
        }
        k += 1 / (tmpk * zInterval);
    }
    for (int y = minY; y <= maxY; y += yInterval)
    {
        for (int z = minZ; z <= maxZ; z += zInterval)
        {
            string command = string.Format("Field2 = {0} and Field3 = {1}" , y, z);
            DataRow[] rows = dataTable.Select(command);
            //Console.WriteLine("size: {0}", rows.Length);
            if (rows.Length == 0) continue;

            double kx = Math.Pow(Math.E, (double)rows[0]["Field4"]);
            tmpk += kx;
        }
        k += 1 / (tmpk * zInterval);
    }
    k = k * deltaZ;
    k = 1 / k;
    return k;
}
double tmpk = 0;
for (int i = 0; i < rows.Length; i++)
{
    //Console.WriteLine(string.Format("Field1: {0} Field2: {1}
                                    Field3: {2} Field4: {3}\n\", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));
    double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
    tmpk += 1 / (kx * zInterval);
}
//Console.WriteLine(string.Format("tmpk: {0}\n\",tmpk));
k += 1 / tmpk;
}
//Console.WriteLine(string.Format("k: {0}\n\", k));
//Console.WriteLine(string.Format("deltaZ: {0}\n\", deltaZ));
k = k / deltaZ;
return k;
}

static double KYPlus(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KYPlus: ");
    double k = 0;
    for (int y = minY; y <= maxY; y += yInterval)
    {
        string command = string.Format("Field2 = {0}\n\", y);
        DataRow[] rows = dataTable.Select(command);
        //Console.WriteLine("size: {0}\n\", rows.Length);
        double tmpk = 0;
        for (int i = 0; i < rows.Length; i++)
        {
            //Console.WriteLine(string.Format("Field1: {0} Field2: {1}
                                               Field3: {2} Field4: {3}\n\", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));
            double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
            double ky = kx * 0.8;
            tmpk += ky;
        }
        k += 1 / (tmpk * zInterval);
    }
    k = k * deltaZ;
    k = 1 / k;
    return k;
}

static double KYNegative(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KYNegative: ");
    double k = 0;
    for (int x = minX; x <= maxX; x += xInterval)
    {
        for (int z = minZ; z <= maxZ; z += zInterval)
        {
            string command = string.Format("Field1 = {0} and Field3 = {1}\n\", x, z);
            DataRow[] rows = dataTable.Select(command);
            //Console.WriteLine("Field1: {0} Field2: {1}
                                               Field3: {2} Field4: {3}\n\", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));
            double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
            double ky = kx * 0.8;
            tmpk += ky;
        }
        k += 1 / (tmpk * zInterval);
    }
    k = k * deltaZ;
    k = 1 / k;
    return k;
}
//Console.WriteLine("size: {0}", rows.Length);
if (rows.Length == 0) continue;
double tmpk = 0;
for (int i = 0; i < rows.Length; i++)
{
    //Console.WriteLine(string.Format("Field1: {0} Field2: {1} Field3: {2} Field4: {3}", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));
    double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
    double ky = kx * 0.8;
    tmpk += 1 / (ky * zInterval);
}
k += 1 / tmpk;
}
k = k / deltaZ;
return k;

static double KZPlus(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KZPlus:");
    double k = 0;
    for (int z = minZ; z <= maxZ; z += zInterval)
    {
        string command = string.Format("Field3 = {0}", z);
        DataRow[] rows = dataTable.Select(command);
        //Console.WriteLine("size: {0}", rows.Length);
        if (rows.Length == 0) continue;
        double tmpk = 0;
        for (int i = 0; i < rows.Length; i++)
        {
            //Console.WriteLine(string.Format("Field1: {0} Field2: {1} Field3: {2} Field4: {3}", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], rows[i]["Field4"]));
            double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
            double kz = 0.1 * kx;
            tmpk += (kz);
        }
        //Console.WriteLine(string.Format("tmpk: {0}", tmpk));
        //fixme zInterval is inaccurate
        k += zInterval / (tmpk * xInterval * yInterval);
    }
k = k * ((double)xInterval * gridCntX * yInterval * gridCntY);
k = deltaZ / k;
return k;
}

static double KZNegative(DataTable dataTable, int minX, int maxX, int minY, int maxY, int minZ, int maxZ, double deltaZ)
{
    //Console.WriteLine("KZNegative:");
    double k = 0;
    for (int x = minX; x <= maxX; x += xInterval)
    {
        for (int y = minY; y <= maxY; y += yInterval)
        {
            
```
string command = string.Format("Field1 = {0} and Field2 = {1}", 
  x, y);
DataRow[] rows = dataTable.Select(command);
//Console.WriteLine("size: {0}", rows.Length);
double tmpk = 0;
for (int i = 0; i < rows.Length; i++)
{
  //Console.WriteLine(string.Format("Field1: {0} Field2: {1}
  Field3: {2} Field4: {3}", rows[i]["Field1"], rows[i]["Field2"], rows[i]["Field3"], 
  rows[i]["Field4"]));
  double kx = Math.Pow(Math.E, (double)rows[i]["Field4"]);
  double kz = kx * 0.1;
  tmpk += zInterval / kz;
}
k += ((double)xInterval * yInterval * rows.Length * zInterval) / tmpk;
}
k = k / ((double)xInterval * gridCntX * yInterval * gridCntY);
return k;
}
static List<double> GetKs(int minX, int maxX, int minY, int maxY, string top, string bottom)
{
  List<double> Ks = new List<double>();
  DataTable dataTable = new DataTable();
  dataTable.MinimumCapacity = minimumCapacity;
  int minZ = 10000, maxZ = -10000;
  /////////////////////////////////////////////////////////
  int cnt = 0;
  double deltaZ = 0;
  /////////////////////////////////////////////////////////
  List<double> layers = GetAverageLayers(minX, maxX, minY, maxY, top, 
  bottom);
  for (int x = minX; x <= maxX; x += xInterval)
  {
    for (int y = minY; y <= maxY; y += yInterval)
    {
      DataTable tmpTable = GetNodes(x, y, layers[1], layers[0]);
      dataTable.Merge(tmpTable);
      DataRow[] rows = tmpTable.Select();
      /////////////////////////////////////////////////////////
      deltaZ += rows.Length * zInterval;
      cnt++;
      if (layers[1] < minZ)
      {
        minZ = (int)layers[1];
        if (minZ % 2 != 0) minZ = minZ + 1;
      }
      if (layers[0] > maxZ)
      {
        maxZ = (int)layers[0];
        if (maxZ % 2 != 0) maxZ = maxZ - 1;
      }
  
      }
tmpTable.Clear();
tmpTable.Dispose();
tmpTable.Reset();
tmpTable = null;

/////////////////////////////////////////
}

deltaZ = deltaZ / cnt;

//Console.WriteLine("I am calculating KXPlus" +
DateTime.Now.ToString());
Ks.Add(KXPlus(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));
//Console.WriteLine("I am calculating KXNegative" +
DateTime.Now.ToString());
Ks.Add(KXNegative(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));

//Console.WriteLine("I am calculating KYPlus" +
DateTime.Now.ToString());
Ks.Add(KYPlus(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));
//Console.WriteLine("I am calculating KYNegative" +
DateTime.Now.ToString());
Ks.Add(KYNegative(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));

//Console.WriteLine("I am calculating KZPlus" +
DateTime.Now.ToString());
Ks.Add(KZPlus(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));
//Console.WriteLine("I am calculating KZNegative" +
DateTime.Now.ToString());
Ks.Add(KZNegative(dataTable, minX, maxX, minY, maxY, minZ, maxZ, deltaZ));

dataTable.Clear();
dataTable.Dispose();
dataTable.Reset();
dataTable = null;

return Ks;
}
static void Main(string[] args)
{
    FileStream filestream = new FileStream(GetOutputPath(),
FileStream.Create);
StreamWriter streamwriter = new StreamWriter(filestream);
streamwriter.AutoFlush = true;
Console.SetOut(streamwriter);

Console.WriteLine(DateTime.Now.ToString());

if (args.Length > 1)
{
    startIndex = int.Parse(args[1]);
}
int gridIndex = 1;
List<string> layerName = new List<string>();
//layerName.Add("layer1");
//layerName.Add("layer2");
layerName.Add("Surface");
layerName.Add("Lay2");
layerName.Add("Lay3");
layerName.Add("Bottom");
for (int x = xMin; x < xMax; x += gridCntX * xInterval)
{
    for (int y = yMin; y < yMax; y += gridCntY * yInterval)
    {
        if (gridIndex < startIndex)
        {
            gridIndex++;
            continue;
        }
        Console.WriteLine(string.Format("Index {0}", gridIndex++));
        Console.WriteLine(string.Format("{0},{1}", x, y));
        for (int i = 0; i < layerName.Count - 1; i++)
        {
            List<double> res = GetKs(x, x + gridCntX * xInterval - 1,
                                    y, y + gridCntY * yInterval - 1, layerName[i], layerName[i + 1]);
            Console.Write(i + 1);
            foreach (double k in res)
            {
                Console.Write("," + k);
            }
            Console.WriteLine();
        }
    }
    Console.WriteLine(DateTime.Now.ToString());
}
Appendix B

```csharp
using System;
using System.Collections.Generic;
using System.Linq;
using System.Text;
using System.IO;
namespace RawK2K
{
    class Program
    {
        static string input = "D:\Data\Kfinal8.txt";
        const int layerNum = 3;
        static string GetOutputPath(int layer)
        {
            string output = "D:\Data\K3\realK_layer"+layer+".txt";
            return output;
        }
        static void Main(string[] args)
        {
            StreamReader inputReader = new StreamReader(input);
            List<StreamWriter> outputWriter = new List<StreamWriter>();
            for (int i = 0; i < layerNum; i++)
            {
                outputWriter.Add(new StreamWriter(GetOutputPath(i+1)));
            }
            //inputReader.ReadLine();
            string tmp, coordinates, layer;
            while (true)
            {
                if ((tmp = inputReader.ReadLine()) == null) break;
                if ((coordinates = inputReader.ReadLine()) == null) break;
                for (int i = 0; i < layerNum; i++)
                {
                    layer = inputReader.ReadLine();
                    string[] items = layer.Split(',');
                    double kxPlus = double.Parse(items[1]);
                    double kyPlus = double.Parse(items[2]);
                    double kyNegative = double.Parse(items[3]);
                    double kzPlus = double.Parse(items[4]);
                    double kzNegative = double.Parse(items[5]);
                    double kx = Math.Pow(kxPlus, 2.0 / 3.0) * Math.Pow(kxNegative, 1.0 / 3.0);
                    double ky = Math.Pow(kyPlus, 7.0 / 8.0) * Math.Pow(kyNegative, 1.0 / 8.0);
                    double alph = 0.823 + 0.167*(1.0 / 9.0);
                    double kz = (1 - alph) * kzPlus + alph * kzNegative;
                    outputWriter[i].Write(coordinates);
                    outputWriter[i].WriteLine(string.Format(",,{0},{1},{2}", kx, ky, kz));
                }
            }
        }
    }
}
```
{ 
    outputWriter[i].Close();
}
}
using System;  
using System.Collections.Generic;  
using System.Linq;  
using System.Text;  
using System.IO;  

namespace realk_stat  
{
    class Program  
    {
        static string input = "D:\data\K3\realk_layer3.txt";  
        static string output = "D:\data\K3\realk_layer3_stat.csv";
        
        class Node  
        {
            public double interval;  
            public int frequency;
        };
        
        static void identify(double k, List<Node> list)  
        {
            for (int i = 0; i < list.Count - 1; i++)
            {
                if (k > list[i].interval && k < list[i + 1].interval)
                {
                    if ((k - list[i].interval) < (k - list[i + 1].interval))
                    {
                        list[i].frequency++;
                    }
                    else
                    {
                        list[i + 1].frequency++;
                    }
                return;
            }
            }
            Console.WriteLine("Wrooooooooooooooooooooooong!");  
            Console.WriteLine(k);
            Console.WriteLine("Wrooooooooooooooooooooooong!");
        }
        static void Main(string[] args)
        {
            List<Node> intervalListX = new List<Node>();  
            List<Node> intervalListY = new List<Node>();  
            List<Node> intervalListZ = new List<Node>();
            double i = 0;
            int cnt = 0;
            while (i <= 401)
            {
                Node tmp;
                tmp = new Node();  
                tmp.frequency = 0;
                tmp.interval = i;
                intervalListX.Add(tmp);
                tmp = new Node();  
                tmp.frequency = 0;
tmp.interval = i;
intervalListY.Add(tmp);
tmp = new Node();
tmp.frequency = 0;
tmp.interval = i;
intervalListZ.Add(tmp);

if (cnt < 10) i += 0.0001;
else if (cnt < 19) i += 0.001;
else if (cnt < 28) i += 0.01;
else if (cnt < 37) i += 0.1;
else i += 1;
cnt++;

StreamReader reader = new StreamReader(input);
StreamWriter writer = new StreamWriter(output);
while (true)
{
    string line = reader.ReadLine();
    if (line == null) break;
    string[] items = line.Split(' , ');
    identify(double.Parse(items[2]), intervalListX);
    identify(double.Parse(items[3]), intervalListY);
    identify(double.Parse(items[4]), intervalListZ);
}

for (int k = 0; k < intervalListX.Count; k++)
{
    writer.WriteLine(string.Format("{0},{1},{2},{3} ",
intervalListX[k].interval, intervalListX[k].frequency, intervalListY[k].frequency,
intervalListZ[k].frequency));
}
reader.Close();
writer.Close();
}

using System;
using System.Collections.Generic;
using System.Linq;
using System.Text;
using System.IO;

namespace realk2zone
{
    class Program
    {
        static string input = "D:\data\K3\realk_layer3.txt";
        static string output = "D:\data\K3\realk_layer3_zone.txt";

        class Node
        {
            public double left, right;
public double zone;
public double k;
}

static void AddZone(double left, double right, double k, double zone, List<Node> list)
{
    Node node = new Node();
    node.left = left;
    node.right = right;
    node.k = k;
    node.zone = zone;
    list.Add(node);
}

static double IdentifyK(double k, List<Node> list)
{
    for (int i = 0; i < list.Count; i++)
    {
        if (k >= list[i].left && k < list[i].right)
        {
            return list[i].k;
        }
    }
    return -1;
}

static double IdentifyZone(double k, List<Node> list)
{
    for (int i = 0; i < list.Count; i++)
    {
        if (k >= list[i].left && k < list[i].right)
        {
            return list[i].zone;
        }
    }
    return -1;
}

static void Main(string[] args)
{
    List<Node> listX = new List<Node>();
    List<Node> listY = new List<Node>();
    List<Node> listZ = new List<Node>();
    // add zone based on your requirement
    // AddZone(left, right, k, zone, list);
    // left is inclusive, right is exclusive
    AddZone(0.01, 0.1, 0.08, 1, listX);
    AddZone(0.1, 0.5, 0.2, 2, listX);
    AddZone(0.5, 1, 0.8, 3, listX);
    AddZone(1, 5, 4, 4, listX);
    AddZone(5, 10, 6, 5, listX);
    AddZone(10, 15, 12, 6, listX);
    AddZone(15, 20, 18, 7, listX);
    AddZone(20, 40, 30, 8, listX);
    AddZone(40, 300, 50, 9, listX);
    AddZone(0.01, 0.1, 0.08, 1, listY);
    AddZone(0.1, 1, 0.2, 2, listY);
    AddZone(1, 5, 2, 3, listY);
    AddZone(5, 10, 6, 4, listY);
AddZone(10, 20, 11, 5, listY);
AddZone(20, 30, 24, 6, listY);
AddZone(30, 50, 34, 7, listY);
AddZone(50, 400, 50, 8, listY);

AddZone(0.001, 0.01, 0.009, 1, listZ);
AddZone(0.01, 0.05, 0.02, 2, listZ);
AddZone(0.05, 0.1, 0.08, 3, listZ);
AddZone(0.1, 0.5, 0.2, 4, listZ);
AddZone(0.5, 1, 0.6, 5, listZ);
AddZone(1, 300, 2, 6, listZ);

StreamReader reader = new StreamReader(input);
StreamWriter writer = new StreamWriter(output);

string line;
while ((line = reader.ReadLine()) != null)
{
    string[] items = line.Split(',', ' ');
    writer.Write(items[0]);
    writer.Write(' , ');
    writer.Write(items[1]);
    writer.Write(' , ');
    writer.Write(items[2]);
    writer.Write(' , ');
    writer.Write(IndentifyK(double.Parse(items[2]), listX));
    writer.Write(' , ');
    writer.Write(IndentifyZone(double.Parse(items[2]), listX));
    writer.Write(' , ');
    writer.Write(items[3]);
    writer.Write(' , ');
    writer.Write(IndentifyK(double.Parse(items[3]), listY));
    writer.Write(' , ');
    writer.Write(IndentifyZone(double.Parse(items[3]), listY));
    writer.Write(' , ');
    writer.Write(items[4]);
    writer.Write(' , ');
    writer.Write(IndentifyK(double.Parse(items[4]), listZ));
    writer.Write(' , ');
    writer.Write(IndentifyZone(double.Parse(items[4]), listZ));
    writer.WriteLine();
}
reader.Close();
writer.Close();