Effect of Rivers on Groundwater Temperature in Heterogeneous Buried-Valley Aquifers: Extent, Attenuation, and Phase Lag of Seasonal Variation

Nathan Lee Young
Wright State University

Follow this and additional works at: http://corescholar.libraries.wright.edu/etd_all
Part of the Earth Sciences Commons, and the Environmental Sciences Commons

This Thesis is brought to you for free and open access by the Theses and Dissertations at CORE Scholar. It has been accepted for inclusion in Browse all Theses and Dissertations by an authorized administrator of CORE Scholar. For more information, please contact corescholar@libraries.wright.edu.
Effect of Rivers on Groundwater Temperature in Heterogeneous Buried-Valley Aquifers: Extent, Attenuation, and Phase Lag of Seasonal Variation

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science

By
NATHAN LEE YOUNG
B.A., Earlham College, 2012

2014
Wright State University
I HEREBY RECOMMEND THAT THE THESIS PREPARED UNDER MY SUPERVISION BY Nathan Lee Young ENTITLED Effect of Rivers on Groundwater Temperature in Heterogeneous Buried-Valley Aquifers: Extent, Attenuation, and Phase Lag of Seasonal Variation BE ACCEPTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF Master of Science.

Robert W. Ritzi Jr., Ph.D.
Thesis Director

David F. Dominic, Ph.D.,
Chair, Department of Earth & Environmental Sciences

Committee on Final Examination

__________________________
Robert W. Ritzi Jr., Ph.D.

__________________________
David F. Dominic, Ph.D.

__________________________
Chris Barton, Ph.D.

__________________________
Robert E. W. Fyffe, Ph.D.
Vice President for Research and Dean of the Graduate School
ABSTRACT

Young, Nathan L, MS Department of Earth and Environmental Science, Wright State University. 2014. Effect of Rivers on Groundwater Temperature in Heterogeneous Buried-Valley Aquifers: Extent, Attenuation, and Phase Lag of Seasonal Variation.

The temperature of groundwater in aquifers is relatively stable when compared to the water temperature in surface-water bodies. However, in aquifers that are hydraulically connected to rivers that have water flux into the aquifer, the local aquifer temperature can show seasonal variation. This project focused on the thermally-altered, near-river zone of such an aquifer, and used numerical methods to examine the extent of seasonal variation in temperature into the aquifer, and the attenuation and phase shift of the signal with distance from the river. The results show that the extent of alteration by diffusive heat flow is negligible compared to the advective component of heat flow. Therefore, because heat transport is driven primarily by advection, the extent of seasonal variation in temperature into the aquifer, as well as the attenuation and phase lag of the signal are significantly dependent on the hydraulic gradient between the river and aquifer. Furthermore, the extent, attenuation, and phase lag of seasonal variation in temperature within the aquifer was found to be strongly dependent on heterogeneity. Considerable differences in the expression of the seasonally varying temperature signal were found to occur as a result of the local presence of high and/or low hydraulic conductivity material. Finally, for the Miami Valley aquifer (which the models used in this study were based upon), seasonal variation
in groundwater temperature is expected only within a lateral distance of about 135 meters from the river and there only within a depth of about 25 meters.
# Table of Contents

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chapter 1</td>
<td>Introduction</td>
<td>1</td>
</tr>
<tr>
<td>Chapter 2</td>
<td>Methods</td>
<td>11</td>
</tr>
<tr>
<td>Chapter 3</td>
<td>Results</td>
<td>19</td>
</tr>
<tr>
<td></td>
<td>3.1 Results of the 1-D diffusion experiment</td>
<td>19</td>
</tr>
<tr>
<td></td>
<td>3.2 Experiments with 3-D flow, both advection and diffusion,</td>
<td>21</td>
</tr>
<tr>
<td></td>
<td>a homogenous aquifer, and no pumping</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3.3 Experiments with 3-D flow, both advection and diffusion,</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>a heterogeneous aquifer, and no pumping</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3.4 Experiments with 3-D flow, both advection and diffusion,</td>
<td>32</td>
</tr>
<tr>
<td></td>
<td>a heterogeneous aquifer, and pumping</td>
<td></td>
</tr>
<tr>
<td>Chapter 4</td>
<td>Discussion</td>
<td>34</td>
</tr>
<tr>
<td>Chapter 5</td>
<td>Conclusions</td>
<td>38</td>
</tr>
<tr>
<td>References</td>
<td></td>
<td>40</td>
</tr>
<tr>
<td>Appendix A</td>
<td></td>
<td>42</td>
</tr>
</tbody>
</table>
List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Schematic diagram of the impact of geothermal pumping, the phase lag and attenuation of the seasonal temperature signal</td>
<td>2</td>
</tr>
<tr>
<td>1.2</td>
<td>Coefficient of performance</td>
<td>3</td>
</tr>
<tr>
<td>1.3.a</td>
<td>Schematic diagram of river thermal influence - August</td>
<td>4</td>
</tr>
<tr>
<td>1.3.b</td>
<td>Schematic diagram of river thermal influence – February</td>
<td>5</td>
</tr>
<tr>
<td>1.4</td>
<td>Temperature of groundwater with lateral distance from the river, Homogenous, heterogeneous, pumped and non-pumped scenarios</td>
<td>9</td>
</tr>
<tr>
<td>1.5.a</td>
<td>Contour map of aquifer temperatures, heterogeneous pumped scenario from Grigsby (2012)</td>
<td>10</td>
</tr>
<tr>
<td>1.5.b</td>
<td>Contour map of aquifer temperatures, homogenous pumped scenario from Grigsby (2012)</td>
<td>10</td>
</tr>
<tr>
<td>2.6</td>
<td>Temperature signal of the river</td>
<td>12</td>
</tr>
<tr>
<td>2.7</td>
<td>Histogram of hydraulic conductivities</td>
<td>17</td>
</tr>
<tr>
<td>3.1.8</td>
<td>Analytical and numerical solutions</td>
<td>20</td>
</tr>
<tr>
<td>3.1.9</td>
<td>Temperature distribution of the seasons</td>
<td>20</td>
</tr>
<tr>
<td>3.2.10</td>
<td>Flow net, relative gradient 2</td>
<td>21</td>
</tr>
<tr>
<td>3.2.11</td>
<td>Time series, relative gradient 2</td>
<td>22</td>
</tr>
<tr>
<td>3.2.12a</td>
<td>Flow net, relative gradient 1</td>
<td>23</td>
</tr>
<tr>
<td>3.2.12b</td>
<td>Flow net, relative gradient 0.5</td>
<td>24</td>
</tr>
<tr>
<td>3.2.13a</td>
<td>Time series, relative gradient 1</td>
<td>24</td>
</tr>
<tr>
<td>3.2.13b</td>
<td>Time series, relative gradient 0.5</td>
<td>25</td>
</tr>
<tr>
<td>3.2.14a</td>
<td>Signal attenuation, depth of $5 , L_D$</td>
<td>25</td>
</tr>
<tr>
<td>3.2.14b</td>
<td>Signal attenuation, depth of $25 , L_D$</td>
<td>26</td>
</tr>
<tr>
<td>3.3.15a</td>
<td>Realization of heterogeneity and permeability</td>
<td>27</td>
</tr>
<tr>
<td>3.3.15b</td>
<td>Heterogeneous flow net, relative gradient 2</td>
<td>28</td>
</tr>
<tr>
<td>3.3.16a</td>
<td>Time series, sample location A</td>
<td>28</td>
</tr>
<tr>
<td>3.3.16b</td>
<td>Time series sample location B</td>
<td>29</td>
</tr>
<tr>
<td>3.3.16c</td>
<td>Time series, sample location C</td>
<td>29</td>
</tr>
<tr>
<td>3.3.17a</td>
<td>Signal attenuation, relative gradient 2</td>
<td>30</td>
</tr>
<tr>
<td>3.3.17b</td>
<td>Signal attenuation, relative gradient 0.5</td>
<td>31</td>
</tr>
<tr>
<td>3.3.18a</td>
<td>Heterogeneous flow net, depth of $25 , L_D$, relative gradient 2, and pumping</td>
<td>33</td>
</tr>
<tr>
<td>3.3.18a</td>
<td>Time series comparison, with and without pumping</td>
<td>33</td>
</tr>
</tbody>
</table>

vi
List of Tables

<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1 River Gradients</td>
<td>11</td>
</tr>
<tr>
<td>2.2 Heat transport parameters for MT3DMS</td>
<td>13</td>
</tr>
<tr>
<td>2.3 Proportions and hydraulic conductivities of sedimentary units</td>
<td>17</td>
</tr>
</tbody>
</table>
1. Introduction

Groundwater temperature has been shown to remain fairly constant at depths below 10 meters from ground surface, despite changes in surface temperature (Anderson 2005). In many areas of the world, this ambient temperature is colder than surface temperature in the warmer months and warmer than surface temperature in colder months, making groundwater a desirable medium for heat exchange in geothermal heat-pump systems.

In aquifer systems that are hydraulically connected to river systems, as in Figure 1, groundwater temperature can be affected by hyporheic exchange (Anderson 2005; Molina-Giraldo 2011). For cases where flux is from the river into the surrounding aquifer, the groundwater temperature can express a seasonally varying temperature signal as found in the river. Field research conducted in the American Southwest has demonstrated that if the overlying stream has a high degree of temperature variation between the warmer and colder months, the extent of thermal alteration resulting from hyporheic exchange can be as great as 24 degrees Celsius at depths shallower than 5 meters, and as great as 8 degrees Celsius at depths 10 meters or deeper (Bartolino and Niswonger 1999). Seasonal temperature variation has been shown to be attenuated with depth as well as distance from the river (Shin et al. 2010; Molina-Giraldo 2011). Phase lag between the seasonally varying temperature signal within the aquifer and in the river has not been examined in these studies. Additional field work has shown that the process
of extracting water for geothermal use can result in an increased amount of hyporheic exchange, and the formation of “plumes” of warmer stream water emanating from the river near pumping zones, as shown in Figure 1 (Anderson 2005; Shin et al. 2010).

Figure 1: Schematic diagram of a plume of thermally altered water emanating from a river, enhanced by geothermal pumping. Graph illustrates the seasonal temperature change in the river, and the attenuation and phase lag of the temperature signal in the aquifer.

Understanding the propagation of the seasonally varying temperature signal may have practical value in a number of areas. One area, as mentioned above, is geothermal heat exchange. As shown in Figure 2, the efficiency of open-loop geothermal systems depends on groundwater temperature.

The City of Dayton, Ohio, lies above an expansive gravel aquifer, and the opportunity for businesses to save on heating and cooling costs by implementing geothermal heating/cooling systems that withdraw water from the aquifer is seen as an economic development issue for the city (Heapy Engineering 2011). Huntsman (2008) examined the thermal impact of the seasonal temperature change in the Great Miami
River on the shallow subsurface near downtown Dayton, Ohio. He showed a seasonal
temperature change in the aquifer at depths to 5 meters below the river channel (Figure
3). Changes below this depth or further away from the river have not been studied.

Figure 2: Coefficient of performance (COP) for geothermal heating/cooling systems as a
function of intake water temperature. A higher coefficient of performance indicates a
more efficient system (Grigsby 2012).

The extent, attenuation, and phase lag of the river temperature signal found within
the aquifer need to be better understood in order to construct geothermal systems that
preform optimally. Certain locations near the river may have temperature signals within
the aquifer that are 180 degrees out of phase with the river temperature, allowing for
increased system efficiency. However if the temperature signal is in phase with the river,
the system will be extracting water that is warmer than desired for cooling, and cooler
than desired for heating, resulting in a loss in system efficiency.

Understanding how aquifer temperature variation occurs temporally and spatially
may also have practical value for aquiculture operations conducted in repurposed quarry
Aquaculture is the fastest growing segment within the agricultural sector in the United States, and there has been significant investor interest in increasing the 0.15 percent national market share that Ohio aquaculture currently occupies (OSU-SC 2010). Many varieties of fish raised in aquaculture projects, such as the Yellow Perch, are very sensitive to changes in water temperature (Tidwell et al. 2010). Because of this, old aggregate quarries mined below the water table, which now occur as lakes and have water temperatures that are buffered by groundwater, are particularly desirable for use in aquaculture projects. With interest rising with regards to repurposing these old quarry lakes for aquaculture, understanding how the water temperature within a given lake will vary, including the understanding of heat transport between hydraulically linked aquifers and rivers, could be important for future development of the industry.

Figure 3a: Schematic diagram of the influence of river-water temperature on the subsurface temperature regime associated with the August river temperature maximum. Blue hemisphere represents the approximate location of the river (adapted from Huntsman 2008).
This project will use numerical methods to study heat transport from a river into a surrounding aquifer, and the resulting effect on groundwater temperature variation with time. It will focus on heterogeneous sedimentary aquifers typical of the North American midcontinent. These aquifers are large preglacial valleys that have been filled with a mixture of highly permeable sand and gravel outwash, and generally have dimensions that range between 2 and 3 kilometers in width, and 50 and 100 meters in depth (Ritzi et al. 2000; Kontis et al. 2001; Sheets 2007). The role of diffusive heat flow relative to advective heat flow, as well as the extent, attenuation, and phase lag of the seasonally varying temperature signal as a function of gradient between the river and aquifer will be examined. Additional consideration will be given to the impact of a geothermal extraction well, and whether the presence of pumping near a seasonally-varying river will
result in changes to the expression of the seasonally varying temperature signal relative to areas without pumping.

The following equations describe the flow of groundwater and heat. Groundwater motion is given by Darcy’s law:

$$\mathbf{v} = \frac{k \rho_w g}{n \mu} (-\nabla h)$$  \hspace{1cm} (1)

where \(\mathbf{v}\) is the seepage velocity vector \([L/t]\), \(k\) is the permeability tensor \([L^2]\), \(\rho_w\) is the density of water \([M/L^3]\), \(g\) is gravitational acceleration \([L/t^2]\), \(n\) is the effective porosity of the porous medium \([D]\), \(\mu\) is viscosity \([M/t*L]\], and \(\nabla h\) is the hydraulic gradient vector \([D]\). When combined with an equation of mass balance, the result is a partial differential equation for head in an aquifer (Freeze and Cherry 1979):

$$\nabla \cdot \left( \frac{k \rho_w g}{\mu} \nabla h \right) = S, \frac{\partial h}{\partial t} \pm Q \pm \omega (h_R - h)$$ \hspace{1cm} (2)

where \(S\), is specific storage \([1/L]\), \(t\) is time, \(Q\) is a pumping rate \([L^3/t/L^3]\] and \(\omega (h_R - h)\) is a river source/sink term \([L^3/t/L^{-3}]\] with a prescribed stage value \((h_R)\) and a leakance term \((\omega)\).

Assuming the temperature of water and aquifer sediment are the same, and that there is no net transfer of heat between them, the heat transport equation can be written as (Hecht-Méndez et al. 2010):

$$\nabla \cdot (D_h + \alpha \mathbf{v}) \nabla T - \nabla (\mathbf{v} T) = R \frac{\partial T}{\partial t} + \frac{q_k}{n \rho_w c_w}$$ \hspace{1cm} (3)
where $R$ is the specific heat storage retardation factor of the saturated media $[\ell]$, $T$ is temperature in units/dimensions of kelvin $[K]$, $D_h$ is the thermal diffusivity $[L^2/t]$, $\alpha$ is aquifer dispersivity $[L]$, $q_h$ is heat injection or extraction $[W/L^3]$, and $c_w$ is the specific heat capacity of water $[J/M/K]$.

The thermal diffusivity, relating conduction to the temperature gradient, is computed as the ratio of the effective thermal conductivity of the medium ($\lambda_e$) and the volumetric heat capacity of the fluid:

$$D_h = \frac{\lambda_e}{n\rho_w c_w}$$ (4)

Mechanical dispersion, $\alpha$, is a fitting parameter sometimes used to account for heat spreading and consequent mixing caused by the otherwise unrepresented tortuosity of groundwater flow paths within a porous matrix. The heterogeneous models used in this study represent this mixing directly through variation in hydraulic conductivity, and thus the $\alpha$ fitting parameter is not used.

The retardation factor, $R$, represents the ratio between the effective volumetric heat capacity of the saturated porous medium ($\rho_e c_e$) and the volumetric heat capacity of the fluid ($n\rho_w c_w$):

$$R = \frac{\rho_e c_e}{n\rho_w c_w}$$ (5)

The effective heat capacity of the saturated medium, $\rho_e c_e$, is the weighted arithmetic mean of the heat capacities of the solid ($c_s$) and fluid ($c_w$), as expressed in equation 5 (Anderson 2005; Hecht-Méndez et al. 2010):
\[
\rho_e c_e = n \rho_w c_w + (1 - n) \rho_s c_s
\]  

Equations 2 and 3 can be solved analytically, but only for very simple models with simple boundary conditions. Numerical models are required for the simulation of complex boundary conditions, aquifer heterogeneity, and more extensive series of sources and sinks, all of which need to be accounted for in models meant to approximate realistic scenarios. Anderson (2005) and Hecht-Méndez et al. (2010) have provided a framework for accurately simulating the flow of heat with public-domain groundwater transport codes.

This project was originally motivated by considerations of developing open-loop geothermal systems, as promoted by the city of Dayton, Ohio. Hence the aquifers represented are similar to the Miami Valley aquifer underlying the Great Miami River and downtown Dayton. Results of a previous numerical study of this area were presented in Grigsby (2012). This work focused on the impact of a network of open-loop geothermal pumping wells located near a stream of constant temperature that is losing water to the surrounding heterogeneous aquifer. Grigsby (2012) showed that river influence in the absence of pumping extends between 100-650 meters laterally into the surrounding aquifer, depending on heterogeneity (Figure 4). When a single pumping well was introduced, this area of river-heat influence increased to 200-700 meters, and groundwater temperatures were warmer due to the increase in river-water infiltration. Grigsby (2012) also found that an extensive network of geothermal pumping wells will increase the amount of exchange between the river and the surrounding aquifer. This subsequently results in a warming of the hyporheic zone due to the increase in warmer river water entering the aquifer (Figure 5). Note that the river temperature in Grigsby
was the summer maximum temperature of 30 degrees Celsius, and only one river-aquifer gradient was examined. Therefore, this work did not take into account seasonal variability of river temperature or the influence of river-aquifer gradients.

This project will expand on previous research by examining the impact of river-aquifer gradients and aquifer heterogeneity on the travel distance of the seasonally varying river temperature signal. Models created for this study will be constructed within the zone of river-heat influence presented in Grigsby (2012). Results of these smaller-scale models will then be normalized so that they can be presented dimensionlessly, with the hope that they can then be scaled to represent any aquifer system. Emphasis will be placed on investigating the extent, attenuation, and phase lag of the seasonally varying river temperature signal within the aquifer.

![Temperature of groundwater with lateral distance from the river in both homogenous and heterogeneous aquifers, with and without pumping as presented in Grigsby (2012).](image)

**Figure 4:** Temperature of groundwater with lateral distance from the river in both homogenous and heterogeneous aquifers, with and without pumping as presented in Grigsby (2012).
Figure 5a: Map view of steady-state temperature values for 30 feet of depth with 25 pumping wells for a heterogenous aquifer. Blue line represents column containing river and crosses and squares represent well locations (Grigsby 2012).

Figure 5b: Map view of steady-state temperature values for 30 meters depth with 25 pumping wells for a homogenous aquifer (Grigsby 2012).
2. Methods

Groundwater flow was modeled using the numerical method of finite differences to solve equation 2 (MODFLOW code, McDonald and Harbaugh 1988; 1996). The model grid contained 10 layers, 50 rows and 150 columns with a constant grid spacing of 10 meters by 30 meters by 30 meters respectively. Aquifer thickness, hydraulic conductivity, and hydraulic head values were taken from Grigsby (2012). Rows 1 and 50 of each layer are prescribed as having constant head in order to create a constant down-valley gradient of 0.0001, an appropriate value for the region being examined (Sheets 2007).

Column 75 contains the river. Head in the river was prescribed as higher than the surrounding aquifer, in order to create a losing stream. Three different river gradients were examined, with magnitudes based on the range of river gradients observed by Sheets (2007). This difference in hydraulic head per unit thickness of the stream bed is called the river gradient hereafter. Table 1 lists the three different river gradients.

<table>
<thead>
<tr>
<th>Table 1: River gradients used in this study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low</td>
</tr>
<tr>
<td>Medium</td>
</tr>
<tr>
<td>High</td>
</tr>
</tbody>
</table>
The models created for this study are intended to represent one reach of a longer river-aquifer system, and the river is assumed to be locally straight within this reach. Hydraulic head is assumed to be at steady state. Finally, the density of water and viscosity of water were treated as constant, as the amount of model error associated with small changes in these parameters has been demonstrated to be relatively trivial (Hecht-Méndez et al. 2010).

The flow of heat was simulated by solving equation 3 with a third-order finite-volume method (MT3DMS code, Zheng and Wang 1999; Hecht-Méndez et al. 2010). The model grid was the same as used for the groundwater flow model. Constant concentration boundaries were prescribed along rows 1 and 50 in each layer. Temperature in the river was sinusoidally varied over 12 discrete stress periods through the sink/source mixing package. The mean, amplitude and phase of the river signal were taken from Huntsman (2008). The temperature signal is illustrated in Figure 6.

![Figure 6: Temperature signal of the river.](image)
Parameters used for the modeling the flow of heat are the same as used by Grigsby (2012) and are listed in Table 2, below. These values were originally given by Hecht-Mendez (2010) as appropriate for sedimentary aquifers.

Table 2. Parameters used for heat transport simulation with MT3DMS code

<table>
<thead>
<tr>
<th>Reaction Package</th>
<th>Volumetric heat capacity of water ((\rho_w c_w))</th>
<th>(4.18 \times 10^6 \frac{J}{m^3 K})</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Specific heat capacity of solids ((c_s))</td>
<td>(880 \frac{J}{kg K})</td>
</tr>
<tr>
<td></td>
<td>Heat retardation factor ((R))</td>
<td>(2.59)</td>
</tr>
<tr>
<td></td>
<td>Bulk density ((\rho_b))</td>
<td>(1961 \frac{kg}{m^3})</td>
</tr>
<tr>
<td>Dispersion Package</td>
<td>Effective thermal conductivity ((\lambda_e))</td>
<td>(2.33 \frac{W}{mk})</td>
</tr>
<tr>
<td></td>
<td>Thermal diffusivity ((D_h))</td>
<td>(1.86 \times 10^{-6} \frac{m^2}{sec})</td>
</tr>
</tbody>
</table>

Heat advection within the mass transport model was simulated through the use of the Total Variation Diminishing (TVD) package in the MT3DMS code, per Hecht-Mendez (2010). This gave an acceptably small numerical error in the heat budget (less than one hundredth of one percent of heat flux). However, this still gave results with a small amount of spurious numerical error in mean temperature in the river cells (i.e. slightly higher than mean input temperature). This small numerical error (2.6 percent error in temperature degrees Kelvin) was subtracted from the raw model temperature.
values, and the results are then plotted as degrees Celsius above or below the corrected mean.

While specific units were used for grid dimensions and parameters in the model, the results were normalized in order to make them more general. The river gradient was normalized by the down-valley gradient:

\[ \text{Relative gradient} = \frac{\nabla h_r}{\nabla h_{dv}} \]  \[7\]

where \( \nabla h_r \) is the gradient from the river into the surrounding aquifer, and \( \nabla h_{dv} \) is the down-valley gradient. The three relative gradients used in these experiments were 0.5, 1, and 2. Distance was also normalized to give results in dimensionless length units, \( L_D \).

\[ L_D = \frac{2000L}{fK_g} \]  \[8\]

where \( L \) is distance \([L]\), \( f \) is the once-per-year frequency of temperature variation \([t^{-1}]\), and \( K_g \) is the geometric mean hydraulic conductivity \([L/t]\). Note \( L_D \) is then scaled the factor of 2000 to allow for easier presentation. Well pumping rate was normalized, \( Q_D \), as:

\[ Q_D = \frac{Q}{4\pi K_g \Delta z \Delta r} \]  \[9\]

where \( Q \) is pumping rate \([L^3/t]\), \( \Delta z \) is grid cell thickness \([L]\), and \( \Delta r \) is grid cell row width \([L]\).
To build confidence in the results of the numerical heat flow model, it was first compared to an analytical solution. This was possible for simple 1-D heat diffusion with a periodic boundary condition. The analytical solution is given by:

\[ T(x) = \bar{T}_r + A e^{-x \sqrt{\frac{\pi c_m \rho_m}{t_0 \lambda_m}}} \cos(\omega t - x \sqrt{\frac{\pi c_m \rho_m}{t_0 \lambda_m}}) \]  

[10]

Where \( T(x) \) is aquifer temperature at location \( x \) [K], \( \bar{T}_r \) is the average river temperature [K], \( x \) is the distance from the river [L], \( c_m \) is the specific heat capacity of the porous media [M·L²/t²/L/T], \( \rho_m \) is the density of the porous media [M/L³], \( \lambda_m \) is the effective thermal conductivity of the porous media [M·L²/t³/L/T], \( t_0 \) is the period of oscillation (365.25), and \( \omega \) is angular frequency \( \frac{2\pi}{t_0} \).

For the test of the numerical approach, a single-layer, single-row model was created with 1001 columns and constant grid spacing of 1 by 1 by 1 meters. One boundary cell was specified to vary in temperature through the time-variant specified concentration option of the sink/source mixing package. Thermal parameters used were identical to those used in the models with advection, with the exception of the value used for thermal diffusivity (\( D_h \)). For the diffusion-only models, the effective thermal conductivity of the porous media, \( \lambda_m \), was used in place of thermal diffusivity in order to simulate heat flow as a result of diffusion alone. The results are presented in section 3.1.

When three-dimensional heat flow with advection was numerically simulated, models of homogenous aquifer composition were first constructed to provide a basis of comparison when evaluating the heterogeneous model. The homogenous models have a uniform hydraulic conductivity of 0.0008 meters per second, the geometric mean
hydraulic conductivity for the Miami Valley Aquifer (Conrad et al. 2008). No geothermal pumping was simulated in these models. When results were analyzed, only the region of the model domain from column 75 to column 150 and row 30 to row 45 was used in order to remove boundary effects. The results are presented in section 3.2.

In order to better assess how the seasonally varying temperature signal will propagate through an aquifer under realistic conditions, a final set of models was created to represent the heterogeneity of a buried valley aquifer system. Sedimentary architecture was simulated in the model through the same methodology as Grigsby (2012). This method uses a Markov-chain approach to simulate aquifer stratigraphy, and then maps permeability into grid cells from statistical distributions defined for each strata type (Carle 1999). Two different lithofacies were represented, a high-conductivity sandy gravel, and a low-conductivity mud and diamicton unit. Proportions of high and low conductivity units were taken from Ritzi et al. (2000), while their respective geometric mean hydraulic conductivities were taken from Conrad et al. (2008). The proportions used and their respective hydraulic conductivities are summarized in Table 4. Figure 7 shows a histogram of the distribution of hydraulic conductivity within each lithofacies. The results from these experiments are presented in section 3.3.

Gradients and thermal parameters used in the heterogeneous models were identical to those in the homogenous models. For heterogeneous models that included a geothermal extraction well, a single pumping well was included in row 35, layer 3, column 80, withdrawing at a rate of $0.6 Q_D$ (equivalent to $500 \frac{m^3}{\text{day}}$, 91.7 gallons/minute). This is the estimated requirement for cooling a 10-story office building utilizing an open-
loop geothermal system (Grigsby 2012; Heapy Engineering 2011). The results are presented in section 3.4.

Table 4: Proportions of sedimentary units, and their respective hydraulic conductivities

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Number of Cells</th>
<th>Percentage</th>
<th>Geometric Mean Hydraulic Conductivity (in ln(cm/s))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand and Gravel</td>
<td>63642</td>
<td>0.85</td>
<td>-2.30</td>
</tr>
<tr>
<td>Mud and Diamicton</td>
<td>11358</td>
<td>0.15</td>
<td>-11.71</td>
</tr>
</tbody>
</table>

Figure 7: Histogram of the hydraulic conductivities of the two lithofacies used in the homogenous models.

One of the largest challenges faced in this project was the very long model run times required to “spin up” the models from the arbitrary initial conditions to a state of dynamic equilibrium with the imposed river forcing. These long run times resulted from
the high specific heat of water, which caused the models to equilibrate very slowly. For example, the model runs with a homogenous aquifer presented in section 3.2 took a week to reach dynamic equilibrium. Furthermore, introducing additional complexity into the models, such as heterogeneity, pumping, or greater river forcing, resulted in significantly longer run times. The model runs presented in sections 3.3 and 3.4 are an example of this, as these runs took between two and three weeks to reach dynamic equilibrium.
3. Results

3.1 Results of the 1-D Diffusion Experiment

The results of the 1-D diffusion experiment are shown in Figures 8 and 9. No appreciable increase in the travel distance of the temperature signal was seen whether the temperature of the model was uniformly prescribed to be the river average temperature of 289.93 degrees Kelvin, or if the temperature in cell 1001 was set to the far field average temperature of 286 degrees Kelvin. Figure 8 provides a comparison between the numerical model and the exact analytical solution at distances of 5 and 20 $L_D$ (5 m and 20 m in specific model units) from the river. The close comparison of the numerical model with the analytical solution builds confidence in the ability of the numerical model to properly compute heat diffusion into the aquifer. Figure 9 shows the numerical result at four times: summer, when river temperature is at its maximum; winter, when river temperature is at its minimum; and spring and fall, when river temperature is closest to the annual average temperature. Results indicate that the diffusion of heat bearing the seasonally fluctuating temperature signal of the river does not extend more than 10 $L_D$ laterally into the aquifer. Additionally, Figure 9 shows that there is “residual” heat storage in the aquifer from summer in the fall, and a heat “deficit” in the spring from the winter minimum.
Figure 8: Analytical and numerical solutions at distances of 5 and 20 $L_D$ from the river.

Figure 9: The temperature distribution for each of the four seasons in the numerical simulation of 1-D diffusive heat flow.
3.2 Experiments with 3-D flow, both advection and diffusion, a homogenous aquifer, and no pumping.

When both advective and diffusive flow are analyzed in a homogenous 3-D model, relationships between the expression of temperature signal in the aquifer and the magnitude of the gradient from the river emerge. Figure 10a, the flow net with a relative gradient of 2, shows how water is infiltrating into the aquifer from the river under the highest relative gradient used. The stream tubes from the river are broader than under smaller gradients (Figure 12). Note that with increased distance from the river, the water in the aquifer is flowing from stream tubes that originate progressively farther upstream.

![Flow net for a homogenous aquifer at depth of 5 \( L_D \) and the highest relative river gradient, 2.](image)

Results were collected at sample location B, 15 \( L_D \) from the river (15 meters in specific model units) as shown in Figure 10. Figure 11 compares the results at 5 and 25
$L_D$, under the highest relative gradient, 2. At a depth of 25 $L_D$, the amplitude of thermal alteration is smaller and the phase lag is larger when compared to results gathered at a depth of 5 $L_D$.

Figure 11: Time series plot of mean removed temperature vs time at depths of 5 and 25 $L_D$ at sample location B, in a homogenous aquifer with a river gradient of 2. Figures 12a and 12b show the flow nets for relative gradients of 1 and 0.5. Note that with a smaller gradient from the river into the surrounding aquifer, the stream tubes are not as broad due to lower advective flow from the river.

Figures 13a and 13b show that lower advective flow from the river into the surrounding aquifer results in a weaker seasonally varying temperature signal at location B due to a reduction in advective heat transport. These figures also show that the lower the gradient, the greater the attenuation and phase lag in the signal at point B.
The relationships between attenuation, lateral distance, and depth are conveyed in Figure 14. Figure 14a shows that the amplitude of the seasonally varying temperature signal decays rapidly with distance from the river, and is filtered out beyond 135 $L_D$ from the river. Figure 14b shows that attenuation is greater with depth. In a homogenous system, the strongest expressions of the seasonally varying temperature signal will be seen at shallow depths, close to the river.

Figure 12a: flow net for a homogenous aquifer. at a depth of 5 $L_D$ and a medium relative river gradient of 1.
Figure 12b: flow net for a homogenous aquifer at a depth of 5 $L_D$ and the lowest relative river gradient of 0.5.

Figure 13a– Time series plot of mean removed temperature vs time at depths of 5 and 25 $L_D$ for sample location B, with a river gradient of 1.
Figure 13b– Time series plot of mean removed temperature vs time at depths of 5 and 25 $L_D$ for sample location B, with a river gradient of 0.5.

Figure 14a: Homogenous aquifer signal attenuation with dimensionless lateral distance from the river at a depth of 5 $L_D$ at location B.
Figure 14b: Homogenous aquifer signal attenuation with dimensionless lateral distance from the river at a depth of 25 $L_D$ at location B.
3.3 Experiments with 3-D flow, both advection and diffusion, a heterogeneous aquifer, and no pumping.

The distribution of facies and corresponding variation in permeability in the heterogeneous model are show in Figure 15a. The effect of heterogeneity on the groundwater flow field is shown in Figure 15b. Results from the heterogeneous simulation were collected from 3 different points along the river in order to observe differences in the impact of local heterogeneity. Results were gathered 15 $L_D$ from the river at depths of 5 and 25 $L_D$ at three different locations labeled A, B, and C in Figure 15a.

Figures 16a-16c show time-series plots for sample locations A, B and C, for the
Figure 15b: flow net at a depth of 5 $L_D$ and the highest relative river gradient, 2, showing the effect of heterogeneity on the groundwater flow field.

Figure 16a: Time series plot of mean removed temperature vs time for depths of 5 and 25 $L_D$, at location A for a gradient of 2.
Figure 16b: Time series plot of mean removed temperature vs time for depths of 5 and 25 $L_D$, at location B for a gradient of 2.

Figure 16c: Time series plot of mean removed temperature vs time for depths of 5 and 25 $L_D$, at location C for a gradient of 2.
Collectively, these figures show that significant differences in thermal alteration that can occur as a result of the presence of high or low conductivity material near the river. At location A, the seasonally varying temperature signal is nearly identical at both 5 and $25 L_D$ below the surface. At location B, the attenuation and phase lag are less at depth than near the surface. At location C there is very little attenuation or phase lag in the signal at either depth relative to location A or B. These relationships are the same under relative gradients of 1 and 0.5.

Figures 17a and 17b summarize how attenuation varies between points A, B, and C, and beyond them with further distance perpendicular to the river. The signal is essentially filtered out beyond approximately $135 L_D$ in all cases.

Figure 17a: Decay of the seasonally varying temperature signal with dimensionless lateral distance from the river along lines taken through points A, B and C. Relative gradient of 2.
Figure 17b: Decay of the seasonally varying temperature signal with dimensionless lateral distance from the river along lines taken through points A, B and C. Relative gradient of 0.5
3.4. Experiments with 3-D flow, both advection and diffusion, a heterogeneous aquifer, and pumping.

The effect of a well pumping at $0.6 \, Q_D$, (equivalent to $500 \, \text{m}^3/\text{day}$, or 91.7 gallons/minute) at a lateral distance of $135 \, L_D$ from the river and a depth of $25 \, L_D$ is given in the flow net in Figure 18a.

The effect is small. Accordingly, the temperature signal at any given location is not strongly changed from that seen without the presence of pumping. For example, the time series in Figure 18b taken at sample location B (the sample location closest to the pumping well) does not show significant change from a scenario without pumping, with the exception of a slight decrease in phase lag. The inclusion of pumping does not significantly pull the seasonally varying temperature signal further into the aquifer. The temperature signal is still filtered out by $135 \, L_D$ from the river under extraction of $0.6 \, Q_D$ at $25 \, L_D$ below the surface.
Figure 18a: Flow net of a heterogeneous model at a depth of 25 $L_D$, with a relative gradient of 2 and a pumping well. Red dot indicates location of pumping well. Note that the presence of the pumping well does not have a noticeable impact on the hydraulic head contours.
4. Discussion

In the aquifers represented in this study, the influence of advective heat transport from the river is significantly greater than diffusive heat flux. Because temperature within the zone of river-heat influence is largely controlled by advective heat transport, the extent, attenuation, and phase lag of the seasonally varying temperature signal is strongly dependent on the hydraulic gradient from the river. Figures 11b, 13a, and 13c show that with the largest relative river gradient, 2, the lateral and vertical extent of the seasonally varying temperature signal is largest, and the phase lag and attenuation are smallest. The attenuation and phase lag are also strongly dependent on local heterogeneity, as shown in Figures 16a-16c which show a large degree of variability in signal expression compared to the results for a homogenous aquifer.

Seasonal variation of temperature in the aquifer is a near-river phenomenon that is generally attenuated within a lateral distance of $135 \, L_D$ near the top of the aquifer and less at depth. Because of the short travel distance of the seasonally varying temperature signal, there is little expression of the thermal signal at depths where geothermal extraction wells are typically located. In the Miami Valley aquifer near downtown Dayton, the seasonally varying temperature signal can be expected to be completely attenuated by 135 meters from the river, and at 25 meters depth. The presence of a
geothermal extraction well with a pumping rate of 91.7 gallons/minute (suitable for cooling an average-sized office building in the downtown area), does not appreciably affect these distances. The potential effects from larger numbers of wells and higher extraction rates were not evaluated in this project, but could be studied in future work that builds upon this study.

According to Grigsby (2012), it is preferable to stay at least 750 meters laterally from the river in order to avoid the negative thermal influence of the river on geothermal extraction wells. Results of the experiments presented here show that locating wells closer within this zone of influence and within 135 meters of the river could result in water expressing the seasonal temperature signal from the river. Note that the presence of the seasonally varying temperature could actually be beneficial if the seasonally varying temperature signal is close to 180 degrees out of phase with the river, which may be possible at a specific location near the river (e.g. Figure 16a). However, given the sizable influence that local heterogeneity has on the expression of the seasonally varying temperature signal within 135 meters of the river, it would be nearly impossible to predict a-priori where such locations would occur. When choice exists, it would be better to follow the recommendation presented by Grigsby.

These results compare favorably with the results of Bartolino and Niswonger (1999), Shin et al. (2010), and Molina-Giraldo (2011). Bartolino and Niswonger (1999) demonstrated that the impact of the seasonally varying temperature is greatly attenuated with depth. Their field data showed that the amplitude of the seasonally varying temperature signal was reduced from 11.25 degrees Celsius at ground surface, to 5.5 degrees Celsius at a depth of 10 \( L_D \) in most wells they sampled. Shin et al. (2011)
demonstrated that the gradient from river to aquifer is the primary method of transmitting the seasonally varying temperature into the surrounding aquifer, though in their models this gradient was driven by the creation very large well fields near the river (within 150 $L_D$) and pumping at extreme rates. Their models showed that at these rates of pumping, a seasonal change of 0.25 degrees Celsius could be detected at pumping wells as far as 150 $L_D$ from the river. Molina-Giraldo (2011) showed that the greatest amplitude of the seasonally varying temperature signal will be seen within the first 10 $L_D$ of the river, but that total travel distance could exceed 150 $L_D$ provided that monitoring equipment was sensitive enough to observe very small changes in temperature.

In summary, the results presented here demonstrate that the extent, attenuation and phase lag are significantly dependent on the gradient between the river and aquifer, This point supports the results of Shin et al. (2010) and Molina-Giraldo (2011), which show that larger gradients from the river into the aquifer contribute to the seasonally varying temperature signal being detected at greater distances both vertically and laterally from the river.

Future work following this project should take two primary directions. First, larger-scale models with multiple pumping wells should be created to investigate the impact of a network of geothermal extraction wells could have on the transmission of the seasonally varying temperature signal from the river into a heterogeneous aquifer. Second, a series of very high resolution models should be created to assess how the phase lag and attenuation increase with distance from the river in detail greater than what was possible with the models developed here. Additionally, an investigation into the variability in river stage could also prove useful. This variability in river stage could
impact the size of the river-aquifer gradients seen within a given season, which in turn could influence extent, attenuation, and phase lag of the seasonally varying temperature signal of the river as seen in the aquifer.
5. Conclusions

The temperature of groundwater in aquifers is relatively stable when compared to water temperature in surface-water bodies. However, in aquifers that are hydraulically connected to rivers that have water flux into the aquifer, the aquifer temperature can show seasonal temperature variation (Bartolino and Niswonger 1999; Anderson 2005; Shin et al. 2010). Grigsby (2012) examined a river-aquifer system in the North American mid-continent region and found that under ambient, non-pumped conditions, the river influence on aquifer temperature may extend as far as 650 $L_D$ laterally from the river (where $L_D$ is distance normalized by hydraulic conductivity times the frequency of the temperature signal). Furthermore, this distance may increase with significant extraction from geothermal source wells. This study examined the region inside this thermally altered zone to see how far seasonal variation in temperature extended into the aquifer, and examined the attenuation and phase shift of the signal with distance from the river.

- The extent of alteration by diffusive heat flow is negligible compared to the advective component of heat flow.
- Because heat transport mostly depends on advection, the extent of seasonal-variation in temperature into the aquifer, as well as the attenuation and phase lag of the signal are significantly dependent on the hydraulic gradient between the river and aquifer.
• The extent, attenuation, and phase lag of seasonal-variation in temperature within the aquifer are also strongly dependent on heterogeneity. Considerable differences in the expression of the seasonally varying temperature signal can occur as a result of the local presence of high and/or low hydraulic conductivity material.

• Aquifer expression of the seasonally varying temperature signal is a near-river phenomenon, and is expected only within about $135 \ L_D$ laterally from the river and a depth of $25 \ L_D$. 
6. References


Carle, S.F., 1999, Transition probability geostatistical software, version 2.1, 76p., University of California, Davis, California


Heapy Engineering, 2011, City of Dayton geothermal feasibility study, prepared by Heapy Engineering. pp 100.


Huntsman, B.E., K.C. Smith, D.J. Wagel, 2008, A thermometric study of the surface water/ground-water interactions along the Great Miami River in Dayton, Ohio, Terran corporation technical report prepared for Michael Ekberg, Miami Conservancy District. pp 57


Ohio State University – South Centers. 2010. Ohio Aquaculture Industry Analysis, Executive Summary 2010. The Ohio Department of Agriculture.


APPENDIX A

Using the SAVELAST accessory program in MT3DMS to create initial conditions for future model runs

Aside from the very long model runtimes detailed at the end of section 2, another challenge faced in this project was managing the extremely large model output files that resulted from these long runtimes. When the sink/source mixing is used, concentrations of the species being modeled are printed into an unformatted text file (called .UCN hereafter) at the end of every stress period. Thus, in the context of the models presented here, the temperatures of every cell within the model grid are recorded at the end of every stress period. Because of the high specific heat of water, the models required an extremely long time in order to reach equilibrium (typically somewhere between 100,000 and 150,000 stress periods were used). Consequently, the unformatted text files generated were typically in excess of 75 gigabytes. These large output files create issues related to both file storage, as one model run will be needed for each of the four seasons, as well as data extraction (done with the PM utility program for MT3DMS).

The problems created by these large file sizes can be avoided with the utilization of the SAVELAST utility program for MT3DMS. This program extracts the final concentration printed in the .UCN and saves it for use as the initial conditions of a new model run. When utilized, this program allows for a single model run to be used to equilibrate the model with the river forcing, while shorter runs can be conducted to examine temperature distributions within the model at different points in time.
Step 1. Saving the last time step

To use the SAVELAST program, copy the SAVELAST executable from the “utilities” file within the MT3DMS directory into the folder containing the original model run. Execute the file within a command window and follow the prompts. Note that the version of MT3DMS used in this project creates two .UCN files: MT3D001.UCN and MT3D001S.UCN. Only MT3D001.UCN should be used. The “S” file includes the sorbed concentration of the species being modeled, and since there is no sorption being examined in these model runs, the file is redundant and can be discarded to save space. Also keep in mind that the larger the input .UCN file being used, the longer SAVELAST will take to run. For .UCN files larger than 75 gigabytes, allow up to 30 minutes for the program to finish running.

Step 2. Restarting

The output of the SAVELAST program is another .UCN file, whose default filename is RESTART.UCN. To use this file as the initial conditions for another model run, the file will need to be called by the BTN package in MT3DMS, and also needs to be included in the .NAM list file. To do this, first make a copy of the BTN package, and within this copy delete the initial concentrations, which can be found after the ICBUND array. Replace the initial concentrations with one copy of the following line for each layer in the model, taking care to maintain proper spacing:

-44 1(20G14.0) -1 A13. Starting concentration in layer 1 for species # 1

Note that everything coming after the A13. is a caption, and is not used by MT3DMS.
Next, the RESTART.UCN file needs to be referenced in the .NAM list file so it will be included in any subsequent model runs. Make a copy of the .NAM file, (for the models used in this project, it is named MT300) and rename it MT300RESTART. After the final line of the list file, add the following line, making sure to maintain proper spacing:

DATA(BINARY)  44  RESTART.UCN

Once these two changes have been made, MT3DMS will use the output of the previous model run as the initial conditions of a new model run. Take care to use the new .NAM file, MT300RESTART, as using the previous .NAM file will result in a re-run of the original model without the use of the .UCN initial concentrations.