EFFECTS OF ANTHROPOGENIC RIVER STAGE FLUCTUATIONS ON SURFACE WATER/GROUND WATER INTERACTIONS ALONG THE DEERFIELD RIVER, MASSACHUSETTS

A Thesis Presented

by

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To Catharine Elizabeth Fleming, my wife.
ACKNOWLEDGMENTS

I want to express my gratitude to my advisor Dave Boutt, only with his guidance and encouragement is this thesis possible. I thank Steve Mabee for sparking my interest in ground-water science. I thank Dave Ahlfeld for broadening my understanding of hydrology. I am thankful to Kathleen Plourde, Patrick Diggins, and Alex Manda for assisting in field work and thoughtful discussions. I would also like to thank John Sweeney for his technical expertise in my field work design and Byron Stone for his thoughtful discussions on surficial geology.
ABSTRACT

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Understanding the connection of surface waters to ground-water systems is important when evaluating potential water resources. In the past surface waters and ground-water have been viewed as two different sources of water but more commonly now they are viewed as one connected resource (Winter et al, 1998). The nature of connection between surface and ground-waters varies depending on climatic and geologic settings, as well as anthropogenic influences such as ground-water pumping and manipulation of river flows by dams. This thesis takes advantage of daily stage changes in the Deerfield River to investigate surface water interactions with ground-water in Charlemont, MA. Two dimensional transient numerical models are constructed to simulate ground-water response to river stage changes. These models are coupled to hypothetical mass transport models to investigate mixing mechanisms of conservative solutes under varying hydraulic scenarios. These simulations support the hypothesis that daily stage fluctuations cause a pumping mechanism which drives solutes into ground-
water systems adjacent to a river at rates higher than normal flow conditions, or even under certain flood conditions.

Riverbed pore-water temperature responses to diurnal temperature fluctuations are measured at two sites along the Deerfield River exposed to the same daily stage changes caused by dams. Temperature and stage data are collected at two sites with differing geologic settings. These data are used to calibrate simple two dimensional models of ground-water flow and heat transport to site specific riverbed hydraulic conductivities. It is suggested that due to the differing depositional environments of the two field sites, hydraulic conductivity of riverbed materials differ, which affects the exchange flux between surface water and ground-water. Understanding the exchange between surface and ground waters under varying hydraulic and geologic conditions is vital to characterizing local water resources and determining ecosystems health.
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CHAPTER 1

IMPLICATIONS OF ANTHROPOGENIC RIVER STAGE FLUCTUATIONS ON
MASS TRANSPORT IN A VALLEY FILL AQUIFER

1.1 Abstract

In humid regions, a strong coupling between surface water bodies and ground-
water systems may exist. In these environments, the exchange of water and solute
depends primarily on the hydraulic gradient between the reservoirs. It is hypothesized
that daily changes in river stage associated with anthropogenic water releases (such as
those from a hydroelectric dam) cause anomalous mixing in the near stream environment
by creating large hydraulic head gradients between the stream and adjacent aquifer. Field
observations of hydraulic gradient reversals in a shallow aquifer are presented. Important
physical processes observed in the field are explicitly reproduced in a physically-based 2-
dimensional numerical model of groundwater flow coupled to a surface water boundary
condition. Under oscillating stage conditions mass transport simulations of a
conservative solute introduced into the surface water are performed and examined
relative to a stream condition without stage fluctuations. Simulations of 20 days for both
fluctuating river stage and fixed high river stage show that more mass is introduced into
the aquifer from the stream in the oscillating case even though the net water flux is zero.
Enhanced transport by mechanical dispersion leads to mass being driven away from the
hydraulic zone of influence of the river. The modification of local hydraulic gradients is
likely to be important for understanding dissolved mass transport in near-stream aquifer
environments and can influence exchange zone processes under conditions of high-
frequency stream stage changes.

1.2. Introduction

Time-dependent variations in surface water stage in both fresh and salt water
environments are acknowledged to strongly influence local groundwater flow [Cooper, 1959]. Variations in surface water stage can arise from many natural and anthropogenic sources including: precipitation and flood events, tidal oscillation, wave induced
displacement, dam releases, and associated reservoir drawdown. These stream stage
fluctuations are known to influence hydraulic gradients in the region surrounding the
stream. In fact, time varying surface water stage is frequently used to estimate aquifer
hydraulic diffusivity [Ferris, 1952; Rowe, 1960; Pinder et al., 1969; Reynolds, 1987;
Swamee and Singh, 2003]. More specifically, shallow ground water and surface water in
glaciated regions are often strongly linked due to the relative position of the groundwater
table to the land surface [Winter et al., 1998]. In these settings, changes in ground water
head or surface water head are transmitted to the adjacent reservoir. For example, river
stage increases during spring runoff events have been observed to cause a corresponding
increase in groundwater elevations and frequently results in reversal of the stream from
gaining to losing conditions [Squillace, 1996]. This process, resulting in shallow
localized water flow into the stream banks, is often termed “bank storage” [Todd, 1955]
and has been acknowledged to play a role in reducing flood peaks in addition to serving
as a storage reservoir for contaminants. The maximum distance that bank-storage water
migrates is a function of river bed and river bank hydraulic conductivity as well as the
hydraulic gradient between the surface water body and aquifer.
Short-term fluctuations of surface water bodies are an important control on groundwater flow in coastal environments [Glover, 1959; Reilly and Goodman, 1985] and possibly in riverine environments too. Tidal fluctuations of ocean and estuarine waters have been observed to modify coastal discharge of ground-water, especially in environments where the aquifer material is thin compared to the magnitude of tidal fluctuation [Ataie-Ashtiani et al., 1999]. Short term fluctuations such as wave-induced displacement, hyporheic flux, and tidal oscillation have been observed to induce nutrient release from sediment pore fluids during low flow summer conditions in estuarine environments [Linderfelt and Turner, 2001]. In a study investigating the interaction between shallow groundwater, saline surface water, and contaminant discharge at an estuarine boundary, Westbrook et al. [2005] studied the influence of tidal oscillation on mixing at the interface. They observed ground-water flow reversals induced by a tidally influenced river which led to groundwater seepage into river sediments and freshening of interface water on outgoing tidal cycles. Yim and Mohsen [1992] developed a 1-D numerical model to simulate mass transport from a contaminant plume to a tidal estuary. They compare models with and without tidal fluctuation and find that although concentrations near the interface are less than in the simulation with tidal fluctuations, the velocity gradients are much greater, and allow more mass exchange from the aquifer to the estuary. In managed surface water settings (e.g. dammed rivers typical of the Northeast US) frequent fluctuations of surface water bodies have been observed to modify adjacent aquifer hydraulic heads on a time scale proportional to the fluctuations in head of the surface water body. The periodicity of the head fluctuations in managed surface water bodies is dependent on water use, but is of a much shorter time scale (and
often more regular) than natural fluctuations in stream stage. In some cases, reversals of a stream from gaining to losing have been observed on a daily basis [Friesz, 1996] due to regular releases of water from upstream reservoirs. It is unknown what effects these reversals have on important ground water surface water processes, such as mixing in the hyporheic zone [Ryan et al., 2004; Arntzen et al., 2006] and nutrient cycling [Kim et al., 1992]. Similarly, changes to the geochemistry of either reservoir can also be transmitted across the boundary at rates proportional to the water flux. In a detailed study of bank storage in Iowa, Squillace [1996] investigated the geochemical response of bank-storage water to a spring-melt runoff event. Gradient reversals between the Cedar River and the adjacent alluvial aquifer were observed, creating losing river conditions. After the flood subsided, gaining river conditions returned. A modeling study showed a 5-times increase of ground-water flux to the river during the first three weeks after the runoff event. Elevated levels of a groundwater contaminant, atrazine, were observed in the bank storage water.

In this paper, we focus on small-amplitude daily oscillation of stream stage through explicit numerical modeling of field conditions. Chemical flux into and out of the groundwater system is assumed to be a function of the magnitude of stream stage fluctuation and may actually be magnified due to dispersive-driven transport processes. A large body of literature exists concerning the impacts of enhanced dispersive mixing in transient flow fields [Dentz and Carrera, 2005; Cirpka and Attinger, 2003; Dentz and Carrera, 2003; Naff, 1998; Bellin et al., 1996; Dagan et al., 1996; Goode and Konikow, 1990]. Transient flow fields are acknowledged to locally increase dispersive transport by creating velocity variations that drive dispersion. Kim et al. [2000] studied an oxygen
isotope plume in an aquifer, where significant fluctuations in the velocity field are caused by temporal changes in recharge rate and lake levels. In this system heterogeneity, fluctuations in recharge rate, and distance from the transient boundary stresses had a significant influence on the vertical transverse dispersion of the plume, while dispersion caused by fluctuations in lake levels alone were found to have a relatively small effect on enhanced transport. In this work we begin with reviewing the impact of daily (i.e. short term) stream stage fluctuations on groundwater flow patterns and investigate the magnitude of short-term fluctuations on mass transport in a hypothetical contaminated surface water to clean ground water scenario. We find that under conditions of oscillating river stage, significant mass may be introduced to the aquifer by dispersive processes, even though no net water flux to the aquifer occurs over time.

1.3. Ground water response to daily stream stage fluctuations in Charlemont, MA

Friesz [1996] reported on the hydrology of stratified drift and stream flow in the Deerfield River Basin of Western Massachusetts (Figure 1.1), and made measurements of groundwater head during 0.5 m fluctuations in river stage resulting from water releases from an upstream dam. Aquifer materials in the West Charlemont Basin consisted of medium-grained sands to boulders in the alluvium and fine silt to medium sand of glacial origin in the deeper, portion of the aquifer. The Deerfield River here is approximately 40 m wide. On the north side, normal to the river there are two terraces, the lower one being the floodplain which is a meter above the riverbed, and the second 50 meters to the northeast, which is 5 meters above the floodplain. Continuing on the same perpendicular transect toward the northeast, the unconsolidated sediments extend another 500 m to the
surrounding bedrock hills. The wells from the Friesz (1996) report are located on the 5m terrace. The aquifer is geometrically constrained by the surrounding bedrock valley. The depth to bedrock at the study site is 33 meters, determined by geophysical surveys conducted by the Fleming and others in 2006 (Figure 1.2). Piezometers were installed in two clusters on a site adjacent to the Deerfield River 5.5 km downstream from the Fife Brook hydroelectric dam. The ‘riverbank’ cluster was located 3 m from the bank, with 3 piezometers 2.4, 8.5 and 17.1 meters deep. The ‘far’ well cluster was located 38.4 meters from the Riverbank cluster (41.4 m from the river), with 3 wells 3.4, 10.3, and 22.3 meters deep. Water levels were recorded in the piezometers every 5 min for a 72-hour period during which the released flow from Fife Brook Dam fluctuated between 3.5 m3/d and 27.5 m3 /d. Large releases from the dam lasted 10 and 12 hours while smaller releases lasted 15 and 13.5 hours. The river stage fluctuation is a maximum of 0.45 meters during the larger of the releases. Unfortunately, the river stage at the location of the piezometers was not measured, but flow was recorded at USGS gaging station 01168500 located 4.6 km downstream in Charlemont, MA. The response of the aquifer system to stream stage fluctuations measured in three riverbank piezometers is presented in (Figure 1.3). Also presented on this figure is the flow in the stream as measured at the gauging station downstream from the piezometers.

The time series depicted begins with a recession in stream flow and then an increase at 186.5 Julian days. Since the stream gauge is located 4.6 km downstream, approximately a 1 hr lag exists between the responses of the aquifer hydraulic head relative to the downstream measured stream flow. The heads in the riverbank cluster piezometers all rise in response to an increase in stream flow (stage). The head in the
shallow piezometer increases gradually while heads in the medium and deep piezometers respond by first rising sharply then dropping and then rising gradually again. The initial response of the medium and deep piezometers is interpreted to record poroelastic loading of the aquifer by an increase in stress on the aquifer (due to the mass of increased stream flow). The deep piezometers follow a similar pattern as the shallow piezometer by rising until about 186.9 Julian days when a recession in stream flow occurs. The change in hydraulic head is much slower during recession of stream flow, a characteristic previously shown to be indicative of bank storage. The cycle of hydraulic head rise and fall in relation to stream flow repeats itself again showing a similar behavior. Hydraulic heads in the far cluster (not shown) have similar trends compared with the riverbank cluster, but importantly they do not indicate gradient reversals as in the river-bank suggesting a fairly narrow zone of hydraulic influence of the river. The change in head level corresponding with the river stage attenuates with depth for both the river-bank cluster and the far cluster. The pulse dissipates horizontally where changes in head (amplitude attenuation) in the far cluster wells are less than in the riverbank piezometers. Additionally, the response of the water table wells is quite different from that in the deeper wells because the water table responds to a free surface (atmospheric pressure) and the deeper wells are interpreted to be weakly confined to confined by finer grained silts in the alluvium. The data presented in Figure 1.3 demonstrate three important aspects controlling processes in the aquifer-river system as a result of changes in stream stage: (1) a strong connection between the river and shallow flow system in the aquifer exists, (2) vertical gradients in the river bank piezometer change direction from upward at low river stage to downward at higher river stage, switching from a gaining stream to a
losing stream as the water is released from the dam, and (3) these processes are repeated on a short-term (daily) basis. These three components of the system are explicitly represented in a physically-based numerical model of groundwater flow to delineate the impacts of stream stage changes on mass-transport processes in a shallow aquifer.

1.4. Modeling approach

A hydraulic model is built to simulate the oscillatory nature of hydraulic head in an aquifer caused by daily river stage fluctuation. The diurnal stage changes caused by upstream dam releases raise the stage of the stream by 0.5 m. Groundwater flow models are developed to simulate hydraulic head, while reproducing the observed vertical gradient reversals (Figure 1.3). Using the modeled transient hydraulic head distribution and corresponding velocity field, conservative advective-dispersive transport simulations are performed using MT3D/MS [Zheng and Wang, 1998]. A comparison of modeled fixed high stage and oscillating stages is performed and analyzed to investigate the role of time-dependent river stage changes on dilution/concentration of solutes in near-stream environments of an aquifer as characterized by mass exchange between the two reservoirs.

1.4.1. Hydraulic modeling of Stream Stage Fluctuations

1.4.1.1. Base Model geometry and properties

A transient, 2-dimensional groundwater flow model is built using the finite difference code MODFLOW [McDonald and Harbaugh, 1988]. The model domain, 500 m wide, 33 m deep consists of 9 layers, representing glacial and alluvial deposits in a
symmetric bedrock river valley. The bottom and right side of the model are no flow boundaries, the left side, a symmetry boundary, is also a no flow boundary, while the top is modeled as a free (water table) surface. Columns are spaced one m apart on the left side of the model and coarsen to 50 m on the right. Grid spacing is refined near the river at the top left corner of the model (Figure 1.4). A river boundary condition is modeled using the MODFLOW river package (RIV5) and is assigned to the 20 cells adjacent to the symmetry boundary in layer 1, for a total width of 20 m. The river bottom elevation is fixed at 30 m with a 0.2 m thick river bed. There are no other hydraulic sources or sinks present in the base model.

Three hydrostratigraphic units are present in this model. These units explicitly reproduce observed hydrostratigraphy at the Charlemont field site. Unit 1, consists of model layers 1-5 and is designated as unconfined units. The unconfined unit represents the upper 5m of stratigraphy, a mix of cobbles and sands, at the Charlemont site. Unit 2 corresponds to layers 6 and 8, and unit 3 consist of layers 7 and 9. Both Units 2 and 3 are modeled in MODFLOW as confined units because the are below a silt layer and represent the lower 28 m of stratigraphy. The hydraulic properties of these layers are calculated from the hydraulic model and listed in Table 1. These values represent model calibration for the two large stage releases presented in Figure 1.3 and are referred to as our base case models. Hydraulic observation points (Figure 1.4) are positioned at elevations of 31 m, 26 m, and 12 m above the model datum. These observation points are located 3m from the edge of the river and are chosen to match the positions of the screened intervals in the field data discussed previously. A value of 31.25 m is used for the initial hydraulic heads in the transient model to allow for gradient reversals near the river, as observed in the
field data. In the base modeling case, river stages are varied instantaneously at half day increments from 31.5 to 31.0 m with half day stress periods divided into 10 time steps of 1.2 hours each repeated for a 20 day period. All of the oscillating simulations begin with a high head (i.e. 31.5 m).

1.4.1.2. Base model results

Model results of hydraulic head at 3 observation locations for the simulation parameters discussed above are presented in Figure 1.5A. As in the field data (Figure 1.5B) the low hydraulic head in the shallow observation location corresponds to low stream stages and high hydraulic head corresponds to high stream stages. The shallow aquifer heads fluctuate above and below the medium and deep observation locations heads, inducing the stream to switch from a gaining stream at low stage to a losing stream at high stage on a daily basis. As anticipated, the hydraulic head fluctuations in the model mirror that of the stream stage with an amplitude reduction being a strong function of distance from the stream. The head amplitude (difference between high and low head during a stream stage reversal) in the deep well is the smallest of the three observation points due to its relative distance (depth) from the fluctuating stream. Theory predicts that a phase shift should also be apparent, but for the properties of the aquifer being simulated here and the time step chosen a shift is not observed. Simulations with refined time steps (10 min) showed small phase shifts (20 min) between the shallow and deep piezometers with the amplitude difference between the two simulations negligible. The general trends in the observed data (Figure 1.5B) are captured in the modeled heads, namely the head gradient reversals. The field data show more complex temporal patterns, such as poroelastic loading of the aquifer, that are not captured with this first-order model of
hydraulic head fluctuations. Poroelastic loading effects are observed only in deeper parts of aquifer (the confined units) as the fluid is compressed by the increase in total stress on the aquifer skeleton. In general the head change induced by loading is on the order of 10’s of cm and lasts for a short time before being masked by the diffusive pressure wave associated with the stream stage change. The fact that this transient is short and that it is swamped by the diffusive wave leads us to believe that its effect on the overall hydraulics of the system is negligible for the problem explored in this paper. The response of porous media to modifications in hydraulic boundary conditions is known to cause modifications in fluid pressure that can persist for long spatial and temporal scales [Neuzil, 2003]. During fluctuations in stream stage the response of the aquifer is observed to lag behind the changing stream stage conditions (Figure 1.6). In the base case model, contours of hydraulic head indicate that the hydraulic head field is vertically and horizontally heterogeneous, with zones (dashed lines) of localized hydraulic head anomalies. These anomalies buffer the stream-aquifer interface and represent the outer reaches of the aquifer affected by the stream stage head. It is assumed that these zones will grow in size and duration proportional to change in stream stage head and aquifer and streambed parameters.

1.4.1.3. Hydraulic model sensitivity analysis

The efficiency with which stream stage changes influence hydraulic gradients (and hence groundwater velocity distribution) in an aquifer depends on (1) degree of connection of the stream to the aquifer, (2) magnitude and frequency of the stream stage fluctuation, and (3) hydraulic properties of the aquifer itself [Hsieh et al., 1987; Wang
and Davis, 1996]. We now examine the sensitivity of modeled heads to items 1) and 3) by varying stream bed conductance and aquifer diffusivity, respectively. Stream bed conductance is a lumped model parameter in the MODFLOW river package that induces head loss between the stream and the aquifer proportional to stream bed width and bed thickness and controls the degree of connection of stream to aquifer. The river bed conductance values are modified by 5 orders of magnitude and the measured hydraulic response of the aquifer system (diurnal head amplitudes) is plotted against river bed conductance for the shallow and deep observation points (Figure 1.7A). Diurnal head amplitude refers to the difference between maximum and minimum head values for each screened interval within the observation well in one day. Larger magnitudes of river bed conductance increase the diurnal head amplitude for both observation points. The shallow aquifer head amplitudes are more responsive than deep head amplitudes, and with increased river bed conductance the difference between shallow and deep amplitudes increases. The base model cases are the middle set of points in Figure 1.7A. At low values of conductance, the aquifer is essentially disconnected from the stream (i.e. no aquifer response). As the riverbed conductance approaches large magnitudes, a change in river stage is transferred almost instantaneously to the aquifer (mimicking a constant head boundary). At these large values, the response of the aquifer is solely controlled by aquifer diffusivity. Aquifer diffusivity, the ratio of saturated hydraulic conductivity to the specific storage (or specific yield for unconfined aquifers), controls the transient hydraulic response of the aquifer. Knowledge of aquifer diffusivity and the length-scale over which hydraulic head changes allows for the calculation of the time it takes for a head change at an observation point. Hydraulic diffusivity for hydrostratigraphic unit 1 is
varied by five orders of magnitude to investigate model sensitivity to this parameter.

Diurnal head amplitudes for shallow and deep observation points are plotted as a function of unit 1 diffusivity. Increases in diffusivity correspond to bounded increases in diurnal head amplitude in both shallow and deep observation points. The rate of increase in the shallow observation point is greater than that of the deep observation points. As unit 1 diffusivity is increased the head amplitude reaches a maximum, which is controlled by the base case river bed conductance parameter. Like river bed conductance, low aquifer diffusivities will hydraulically disconnect the aquifer from the stream thus reducing the influence of these high frequency stream stage changes on hydraulic head. We now investigate the influence of these hydraulic conditions on mass-transport mechanisms in the above described aquifer.

1.4.2. Mass transport modeling of stream stage fluctuations

Using the mass transport code MT3D-MS and TVD solver coupled with MODFLOW, a conservative solute is introduced to the aquifer from the stream using a point source and its fate is explored. The point source is assigned to the river cells to simulate effects of surface water contamination on the underlying groundwater system. The point source is not a boundary condition in the formal sense and uses the stream-aquifer flux term, \( q_s \) (m/day), to calculate the mass flux entering or leaving the system, \( q_s/n \) Cs, where \( n \) is the aquifer porosity and Cs (mg/L) is a fixed concentration of the solute in the stream. This is different than a specified head/ specified concentration (SHSC) boundary condition which is discussed later in the paper. The fixed solute concentration (Cs) in the stream is set to 100 mg/L. The initial concentration in the
aquifer is 0 mg/L and all boundaries in the model are assigned to have no mass flux. In each set of transport simulations, we combine the transport model with either the oscillating stage models (discussed above) or a hydraulic model with a fixed high stage to test mass transport behavior under different stream management conditions. No molecular diffusion is simulated in any of these models due to the short time scale of the simulations and presence of high groundwater velocities (and hence high mechanical dispersion) near the stream. Preliminary simulations with non-zero molecular diffusion coefficients showed no significant difference in simulations. Thus, the two modeled mass transport mechanisms are advection and mechanical dispersion. Simulations of mass transport from stream to aquifer are performed at longitudinal dispersivity values ranging from 0 m (advection only) to 10 m and run for a period of 20 days. Dispersivity, a term whose magnitude controls the degree of hydrodynamic dilution as a fluid flows through a porous media, is well known to be a strong function of spatial scale [Zheng and Bennett, 1995] and these values bound the range likely to be representative of most field conditions at this scale. A plot of the mass introduced into the aquifer as a function of time for different values of dispersivity and stage conditions is presented in Figure 1.8. Results from only one simulation of fixed high stage conditions is presented as the total mass introduced into the aquifer under fixed stage conditions is not sensitive to changes in dispersivity. Despite this, the solute distribution within the aquifer is sensitive to dispersivity even though the total mass input into the system is identical. This phenomena is further observed by comparing the mass introduced into the aquifer during the first half day of simulation time (when both conditions have a fixed stage) for oscillating and high stage cases (Figure 1.8A). All lines are coincident until the first stage change (at 0.5 days)
when oscillating and high stage simulations diverge from one another. Under this condition, flux into the aquifer through the point source is only dependent on $q_s$ and thus dispersive flux is zero. As the simulations progress further in time (Figure 1.8B) the discrepancy between the mass introduced into the aquifer becomes greater. The 10 m dispersivity oscillating case introduces the most mass into the aquifer after 20 days or 20 stream stage oscillation events. The oscillating stage condition at 1 m and 10 m dispersivity actually introduce more mass into the aquifer than fixed high stage case. This is non-intuitive as the 20 day total water flux into the aquifer from oscillating cases is zero while fixed high stage simulations introduce significant (non-zero) water to the aquifer. The fixed high stage case has a high head for 20 days of simulation time and oscillating cases have a high stage only half of the time (10 days). Instead of plotting mass introduced into the aquifer as function of time we plot the mass versus cumulative amount of water moving through the stream during the 20 day simulation period by calculating the volume of water in the stream for both the oscillating and high fixed stage. This indicates (Figure 1.8C) that for an equivalent amount of water in the stream, the oscillating case at 10 m dispersivity introduces almost twice the amount of mass in the aquifer than the fixed high stage case. Additionally, the 1 m dispersivity case introduces more mass into aquifer than a high fixed stage. We now examine the mechanisms driving excess mass transport in the oscillating cases.

1.4.2.1. Analysis of Mass Transport Modeling Results

When the stream stage in the oscillating case is lowered, water in the near-stream environment begins to flow back into the stream, carrying with it a portion of the mass introduced during the previous high stage. Not all mass introduced during the high stage
(which is identical for all three dispersivity cases) flows back into the stream during the low stage due to dispersive transport. Over the time period of simulations presented here the advective case (0 m dispersivity) is most efficient in removing the previously introduced solute mass (Figure 1.9). Simulations with increasingly larger dispersivities (1 m and 10 m) have more mass retained in the aquifer system. As the number of oscillations of the stream stage increase the simulations show more total mass stored in the aquifer as the storage effects are additive. The largest discrepancy between mass input into the aquifer and mass removed is at early times when (1) the hydraulic gradients and (2) the concentration gradients between the stream and aquifer are largest. The difference in mass-in versus mass-out decreases over time and results in late-time flattening of the curves as presented in Figure 1.8B. During the transition from high stage to low stage, mass transported into the aquifer is driven back into the stream at a rate proportional to the water flux. During early times, all oscillating simulations input more mass than is removed. This is due to the hydraulic head gradients reaching equilibrium with the oscillating stream stage condition. The late-time flattening of the curve in Figure 1.8B of the advection only case is due to the reduced hydraulic head gradients in the hydraulic zone of influence. As time progresses the advection only case (as evidenced by the steep slope in Figure 1.9) becomes more efficient in removing mass until it approaches a steady condition where mass introduced during a high stage is removed during a low stage. The 1 and 10 m dispersivity simulations show the same overall trend but their trajectories are quite different, and thus, the time it takes for these simulations to reach a steady state is longer for higher dispersivity. This difference is attributable to mechanical dispersion within the aquifer itself.
1.5. Discussion of Mass Transport Results

The mass retention mechanism described for this setting involves dissolved mass being transported into the aquifer away from the river boundary, thereby reducing the available mass adjacent to the river to be removed from the system (Figure 1.10A and B). Simulations with high dispersivity values have concentration profiles in Unit 1 that show the longest transport distances after one stage reversal (Figure 1.11A) and 10 stage reversals (Figure 1.11B). Simulations with no dispersive transport mechanisms (dispersivity of 0 m) show the most compact and localized concentration profiles have more mass located near the river boundary. However, over time this mass is transported further from the river boundary and is essentially disconnected from the short term influence of the river (Figure 1.10A and B). The end result is that more mass is input to the aquifer as a function of transport parameters, namely dispersion, and diffusion (although not explicitly simulated here). In oscillating cases, (Figure 1.10A and 1.10B) the river imparts a zone of influence in the aquifer, whose size is dependent on the head gradient and the aquifer diffusivity. When mass is introduced into the river, it is transported by advection into this zone, as the stage drops back down, mass is transported back into the river from the aquifer, however some mass remains. This is due to mechanical dispersion driving mass outside the river's zone of influence. As dispersion is decreased, the oscillating heads are more efficient in removing mass from the aquifer, which means less mass leaves the zone of influence and therefore less total mass is introduced into the aquifer. Dispersion is a factor on transport in the oscillating case because flow velocities remain high compared to a fixed high stage (Figure 1.10C),
where solute transported by dispersion is controlled by the decreasing head gradients. While the above mechanisms explain the dependence of mass input into the aquifer on dispersivity, it does not specifically address the differences in the oscillation versus fixed stage river conditions. As shown in Figure 1.6, gradients in hydraulic head during stream stage changes in the region surrounding the stream are dynamic and therefore must influence mass transport processes during these events. High stage oscillation events deliver a relatively constant amount of mass of over time (for the simulation parameters in this study 0.7 kg) compared to the fixed high stage that inputs a maximum of 0.4 kg and decreases following an exponential trend to 0.05 kg over 20 days (Figure 1.12). We attribute these differences to temporal evolution of hydraulic head gradients in the oscillating versus fixed stage cases. Figure 1.13 illustrates the nature of horizontal hydraulic head gradients between the stream and aquifer for both an oscillating and fixed stage conditions. The local hydraulic gradients near the stream are enhanced following a stream stage reversal resulting in a large mass flux term (qsCs) and the addition of more mass to the aquifer. During the first 0.5 days of simulation time both cases have the same hydraulic head and the gradients are identical. As the stage reverses the horizontal gradients oscillate with the changing stage and the fixed high stage raises the overall water table, reducing hydraulic gradients in the near-stream environment. No net change in water table position is observed for the oscillating cases, consistent with the fact that no net water flux enters the aquifer under the oscillating case compared to significant water flux in the fixed cases. This mechanism of hydraulic gradient modification and the influence of transport properties (dispersion) on mass retention in the aquifer, explains the pumping mechanism responsible for enhanced transport. This brings us to the non-
intuitive conclusion that more net-mass is input into the aquifer under a condition of zero net-water flux compared to conditions of significant water flux in the fixed high stage case. We now briefly examine the dependence of mass input to the aquifer to the magnitude of stream stage oscillation.

1.5.1. Influence of magnitude of stream stage oscillation on mass flux

To evaluate the influence of large stream stage oscillations on mass flux into the aquifer we performed a number of simulations holding longitudinal dispersion constant and increasing river stage levels (Figure 1.14). In all simulations, mass flux increases into the aquifer with increased stage heads. As discussed above, the total mass flux for fixed stage simulations are not sensitive to changes in dispersivity, therefore, fixed stage results are plotted against the oscillating stage results for 10m, 1m, and 0m longitudinal dispersion (Figure 1.14). The oscillating case delivers more mass into the aquifer than the fixed case at lower stage amplitudes for all values of dispersion. However, with increasing stage heads the fixed case will eventually transport more mass into the aquifer than the oscillating case. This is because the water table in the aquifer takes longer to equilibrate to the high stage, allowing hydraulic gradients to stay large over a greater period of time and inducing more mass flux. The difference between the maximum head and initial head (termed here max-initial head) where this occurs is controlled by the dispersivity of the oscillating stage cases (Figure 1.14).
1.5.2. Sensitivity of model results to numerical boundary conditions

To evaluate the influence of different numerical representations of the surface water/ground water interface on mass transport, we created parallel models. One model uses the river package and point source (RPPS) and the other uses specified head and specified concentration boundaries (SHSC). The boundary values of both models were the same, heads of 31.5 m, and specified concentrations of 100 mg/L for the high stage comparisons. These cases represent two end members; one with zero dispersive flux into the aquifer from the surface water (RPPS) and a case with maximum dispersive flux (SHSC). During high stage base RPPS simulations, the amount of mass transmitted into the aquifer after 20 days is unchanged for dispersion values of 10 m, 1 m, and with dispersion shut off (0 m). However, with SHSC, the amount of mass in the aquifer after 20 days increases with greater dispersion values. The total mass into the aquifer after 20 days for zero dispersion is identical for both boundary conditions, indicating that dispersive flux from stream to aquifer is significant for the SHSC case. The reason for the difference between these two simulations is the specified concentration boundary condition. The RPPS river bed conductance term regulates how much water will pass from the river into the aquifer; acting as a water flux control. With RPPS, solute flux from the river to aquifer is solely water flux dependent whereas, the SHSC is dependent on both the hydraulic gradient and concentration gradient. Therefore, with greater dispersion, the spatial distribution of solute plume is larger because of the specified boundary concentration of 100 mg/L and the initial condition of 0 mg/L creating larger concentration gradients and allowing more mass to enter the aquifer. Oscillating simulations (results not shown) with the SHSC show similar trends to those presented
with RPPS but the SHSC simulations all show higher mass input into the aquifer, due to the larger dispersive flux from surface water feature to aquifer. To this end, a case with zero dispersive flux into the aquifer from the surface water (RPPS) and a case with maximum of dispersive flux (SHSC) both show the overall trends of more net mass input into the aquifer despite having significantly less net water flux for simulations lasting 20 stream stage cycles.

1.6. Conclusion

Mass transport from a surface water body to groundwater under conditions of fluctuating river stage shows enhanced mixing and significant mass transport under conditions of zero-net water flux. The mass flux into the aquifer increases for a number of values of aquifer dispersivity as well as surface water model representations. In this work we present field data documenting anthropogenic impacts of surface water fluctuations on groundwater heads that lead to significant velocity variations within the subsurface aquifer materials. Modeled impacts of this surface water-induced velocity variations significantly impact the sub-surface transport of dissolved mass compared to control simulations. These modeling results suggest a pumping mechanism induces net mass flux of solute into the aquifer under conditions of zero net water flux. Results suggest that dispersive transport is an important factor in driving mass into and out of surface water bodies. This study is the first to document these enhanced transport mechanisms as applied to high frequency surface water fluctuations caused by anthropogenic water releases.
Groundwater velocities in the near-stream environment appear to be the most important factor controlling mass flux. Dispersive properties and concentration gradients of the aquifer act to transport this mass away from the stream. Ground water velocities increase with higher head in the river in both the oscillating and fixed cases. Results demonstrated here are consistent with the enhanced transport conditions observed in transient flow fields [Dentz and Carrera, 2005; Cirpka and Attinger, 2003; Dentz and Carrera, 2003; Naff, 1998; Bellin et al., 1996; Dagan et al., 1996; Goode and Konikow, 1990]. Although the work by Kim et al. [2000] showed in a lake-aquifer system that mixing of conservative tracer was influenced more by recharge transients compared to surface water fluctuations, in a river dominated setting we show that stage fluctuations can be quite significant. This is especially enhanced when the source of the contaminant is located in the fluctuating surface water body. We show here that strongly connected surface and ground water systems can induce significant hydraulic transients that influence mass transport through enhanced dispersion. These anthropogenic high frequency stage changes have the potential for disrupting nutrient and geochemical processes in the near-stream environment.

In humid regions, where the water table is near the ground surface, surface waters and ground waters often interact with each other. There is a range of conditions and environments (coastal, estuarine, and riverine) where oscillating stream stage changes can create a potential for enhanced transport when compared to steady stream stages of the same magnitude. Using ground water flow and mass transport modeling techniques, the dominating parameter for mass transport in this setting is groundwater velocity (qs).
which remains relatively high (near the stream) under oscillating conditions compared to fixed high stage conditions. The larger velocities in oscillating stream stage models allow for greater hydrodynamic dispersion compared to a high stage model creating excess mixing and hence more transport even under conditions of zero net flux of water (i.e. no long term storage of water) into the aquifer. Additional factors, such as recharge, evapotranspiration, and regional groundwater gradients will control the spatial extent of the velocity reversal zone and the magnitude of groundwater velocities in the vicinity of the stream-aquifer interface. As water table gradients are increased toward the surface water body (strongly gaining stream), the efficiency of this “pumping” mechanism is likely to be decreased. In the reverse case (a losing stream), this mechanism is likely to be more efficient. The controlling factors for simple mass transport in hydro-dynamic stream water/ground water settings are oscillating stage frequency, the magnitude of the stream stage change, and the shallow subsurface material properties. These should be the primary parameters to define when investigating solute movement through ground water and surface water interfaces in relatively short temporal and spatial scales. Stream width, although not explicitly considered, is another parameter that merits additional consideration in this process. This pumping mechanism is likely to be important to consider when investigating exchange zone processes under conditions of transient stream stage changes.
Figure 1.1 Massachusetts portion of the Deerfield River Watershed with the Charlemont study area boxed out in red.
Figure 1.2 The West Charlemont Basin, Charlemont, MA showing well locations, geophysical surveys, and cross sections.
Figure 1.3 Measured response to aquifer system to fluctuations in stream stage. Data from piezometers are relative to masl and recorded at 15 minute intervals. Symbols are used to distinguish between different piezometers and stream flow data. The timing of stage changes and stream flow data lag each other because gage is 4 km downstream from Piezometers. Modified from Friesz [1996]
Figure 1.4 Grid geometry and hydrostratigraphic units for hydraulic flow model of stream-stage reversals. Location of river boundary cells is indicated with solid black dots. Layer shading (from white to dark gray) indicates specific hydrostratigraphic units.
Figure 1.5 Simulated (A) and observed (B) head responses to stage fluctuations in shallow, medium, and deep observation wells. Important hydraulic phenomena captured in the simulations are the reversals in the vertical gradients of hydraulic heads.
Figure 1.6 Contours of hydraulic head at two times following a switch in stream stage from a low to high stream stage. Grey shading indicates un-confined portion of aquifer. Time lag in fluid flow creates vertically (A) and horizontally (B) distinct zones of (dashed regions) hydraulic head adjacent to the stream. These zones border the stream and represent transient features associated with stream stage reversals. Contour intervals are 0.01m.
Figure 1.7 Model produced sensitivities of head fluctuations in shallow (triangles) and deep (squares) observation locations to changes in river bed conductance (A) and changes in unit 1 diffusivity (B). Base case simulations are the middle data point in each figure.
Figure 1.8  Mass introduced into the aquifer through a fixed high stream stage (solid line) and oscillating stream stage (various dashed lines) as a function of aquifer dispersivity (value indicated in parentheses). Data is plotted versus time (A,B) and the sum of the flow in stream per unit length of stream (C). All cases depict a rapid increase in mass transfer and a gradual reduction as concentration gradients become smaller. When plotted against the volume of water in the stream (C), 1 and 10 m dispersivity oscillating simulations show more mass input into aquifer during the 20 day simulation period. High stream stage simulations are not sensitive to changes in aquifer dispersivity.
Figure 1.9 The amount of mass removed during stage reversals as a function of dispersivity values (indicated in parentheses). In early times, mass removal is low during the half day low stages but quickly increases as hydraulic gradients stabilize within the aquifer. The advection only case (0 m dispersivity) is the most efficient at removing mass from the aquifer.
Figure 1.10  Schematic showing solute front at high and low stage of oscillating case, (A and B), and constant high stage of fixed stage case (C).  (A) River stage is high, the vertical line represents the hydraulic zone of influence, and the 2 dashed curves represent the contributions of advection and the combined effects of advection and dispersion on solute transport into the aquifer.  (B) Low stage where groundwater flow directions are reversed. Inside the hydraulic zone of influence, solute is transported back to the stream due to advection, but beyond the zone of influence, velocity gradients force solutes deeper into the aquifer.  (C) Fixed high stage with the plume of solute transport as a function of time represented by dashed lines. Notice lines get closer together as time increases.
Figure 1.11  Horizontal concentration profiles in unconfined portion of aquifer (unit 1) for first high and low stages (1st reversal) (A) and concentrations after 10 reversals (B). Dispersivity values are indicated in legend. During the first stage reversal (A) hydraulic and concentration gradients are high and cause significant amount of mass to enter the aquifer. Simulations with high dispersivity values are efficient in transporting this mass beyond the hydraulic influence of the stream. At later times (B) solute mass that has moved beyond the hydraulic influence of the stream is not influenced by the stream stage.
Figure 1.12 A plot comparing the mass input into the aquifer for a high fixed stage and oscillating stage. Mass introduced into the aquifer during high stages (for all time) of the oscillating case is relatively constant whereas the mass input over time for a fixed high stage decreases significantly. This decrease is directly related to the fact that net water flux is directed into the aquifer.
**Figure 1.13** Horizontal head profiles for the first high/low (1st reversal) and fixed stage (A) and stage reversal/fixed stage at 9.5 and 10 days (B). During the first stage reversal (A) hydraulic gradients in the high fixed and oscillating stage simulations are similar. At later times (B) hydraulic head profiles in the oscillating case are similar to early time, whereas the fixed high stage are less steep, broader and explain the differing mass flux into the aquifer presented in Figure 1.8
Figure 1.14 Mass introduced into aquifer from stream as a function of magnitude of high stage, and magnitude of oscillation from initial head values (maximum stage height - initial aquifer head). Simulations with highest dispersivity values introduce most mass into aquifer until hydraulic gradients become large enough such that fixed high stage introduces more mass. This transition is a function of aquifer dispersivity.
Table 1: Properties of hydrostratigraphic units in the base hydraulic model

<table>
<thead>
<tr>
<th>Unit</th>
<th>Layers</th>
<th>K (m/s)</th>
<th>S (1/m)</th>
<th>D (m^2/s)</th>
<th>Anisotropy</th>
<th>Type</th>
<th>Riverbed Conductance (m^3/d)</th>
</tr>
</thead>
<tbody>
<tr>
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<td>1-5</td>
<td>1.4E-03</td>
<td>2.1E-01</td>
<td>6.6E-03</td>
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<td>6.5E+00</td>
</tr>
<tr>
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<td>6,8</td>
<td>8.8E-05</td>
<td>6.0E-05</td>
<td>1.5E+00</td>
<td>1\10</td>
<td>Confined</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>7,9</td>
<td>3.5E-04</td>
<td>6.0E-05</td>
<td>5.8E+00</td>
<td>1\10</td>
<td>Confined</td>
<td></td>
</tr>
</tbody>
</table>
CHAPTER 2

ESTIMATING HYDRAULIC CONDUCTIVITY OF RIVERBED MATERIALS USING TEMPERATURE AS A TRACER AND COUPLED HEAT TRANSPORT AND GROUND-WATER NUMERICAL MODELS

2.1. Abstract

The magnitude of surface water and ground water interaction is known to be highly dependent on the character of riverbed sediment and aquifer substrate. Therefore evaluation of the geologic substrate beneath river reaches is important when determining the magnitude of interaction. Two study sites along the Deerfield River, which lie above different sediment packages only a few kilometers (km) apart, are hypothesized to have different subsurface responses to similar river stage changes caused by upstream dam releases. Taking advantage of these upstream dam releases, which cause the Deerfield River stage to change by a half meter daily, vertical temperature profiles are measured in the riverbed at two field sites within the Charlemont basin. To estimate surface water and ground water flux through riverbed materials these temperature observations are used to calibrate two-dimensional coupled models of ground-water flow and heat transport. Model results calibrated to field data demonstrate higher hydraulic conductivities of riverbed materials at the northern study site and lower hydraulic conductivities at the southern study site which differ in geologic substrate. This implies that there is greater potential for surface water interaction with ground-water at the northern site than at the southern site. These initial results support the hypothesis that differences in geologic substrate can influence the flux of water through the surface water / ground-water interface, even over small spatial scales within the same geologic basin.
2.2. Introduction

In river environments, surface water interactions with ground-water control chemical, physical and biological gradients. These interactions can be important to estimates of recharge in arid environments (Stonestrom et al, 2007), solute transport (Squillace, 1993) and overall ecosystem health (Stanford and Ward, 1988). Geomorphologic factors controlling chemical and physical gradients have been investigated in fluvial systems and are found to be highly variable over small spatial scales (Bencala and Walters 1983, Cardenas and Zlotnik, 2003 Cardenas and Wilson, 2007). In controlled laboratory investigations, flumes have been used to better understand sediment structure controls on surface water interactions with ground-water (Packman and Salehin, 2003).

In many settings with significant ground-water velocities, temperature can be used as a conservative tracer for ground-water flow (Anderson, 2005). Shallow ground-water temperature responses to diurnal air temperature fluctuations have been investigated to estimate water flux using coupled models of ground-water flow and heat transport (Hatch et al, 2006, Constantz and Stonestrom, 2003). However, little work has been done using basin scale depositional architecture to predict surface water / ground-water exchange.

In a study on the Santa Clara River in California, a comparison of bromide, (a conservative solute) and heat as ground-water tracers was undertaken (Constantz et al 2003). The authors combined field measurements of temperature and bromide concentrations with numerical models to estimate stream channel properties and water
flux. They concluded that results from both bromide and temperature were comparable to one another.

In this study we investigate the relationship of the geologic setting to potential surface water interaction with ground-water in a glacially deposited valley fill environment. A conceptual depositional model of the Charlemont basin is constructed (Figures A4 and A5), supported by well logs, seismic refraction surveys, and field mapping. Pore water temperature is measured at two field sites, located in different depositional environments and models are used to estimate the hydraulic conductivity of riverbed materials at each site. Model results support the hypothesis that surface water exchange with ground-water will vary significantly at two different geologic settings within the Charlemont basin under similar hydraulic conditions.

2.2.1 Geologic Setting

In the glaciated region of the northeast United States, many modern river channels were carved by melt waters from the retreating ice sheet. Many of these drainage networks incised valleys into the underlying bedrock which then filled with sediments carried by ice and melt water. In the Deerfield River watershed (Figure 2.1), ice retreated in a north-northwest direction as interpreted by the orientation of drumlins in the area (Mabee et al 2007). This northwest retreat of the ice in the Charlemont basin left behind a temporary glacial lake in front of the ice. Well logs (Figure A1) from the basin show a fining upward sequence of sediments in the lower portions of the basin, from coarse sand and gravel to lacustrine clays up to 15 m thick. Logs to the northwest also show the same pattern but include a thinner clay unit. Finally, in the northern most logs, the fining upward sequence is replaced with coarser, poorly sorted sediments. The deepest sections
of the sediment package are 45 meters (m) thick as estimated by seismic refraction surveys in the northern basin. Using well and borehole logs, existing surficial geologic maps and supplemental surficial mapping as well as the principals of morphosequences (Randall, 2001) a conceptual sediment architecture was created (Figures A3-A7 in Appendix).

The Deerfield River and adjacent tributaries eroded through some of the glacial deposits in the basin. The path of the river has meandered across much of the valley creating several terraces that exist today. The terrace closest to the present day river is the floodplain terrace, and is only a few meters above the river. A second terrace varies in height and distance throughout the valley, but is 3 to 5 m higher than the first. It is observed that the riverbed materials change from cobbles and coarse sand in the north to cobbles with fine silts and clays mixed with some sands to the south. Given the variation in geologic and geomorphological conditions throughout the basin, two sites were selected to investigate whether the depositional model can be used to estimate relative differences in magnitude of surface water interaction with ground-water.

2.3. Field Methods

The Deerfield River in Charlemont, MA is subjected to regular stage changes by upstream dams. These stage changes reverse ground-water flow gradients near the river, changing the reach from gaining to losing on a daily basis (Friesz 1996). To better understand the magnitude of water flux through the surface water / ground-water interface, pore-water temperature measurements, along with surface water temperature and pressure measurements were taken at two sites on the Deerfield River. The purpose
of the data are to calibrate numerical models of ground-water flow coupled with heat transport in order to estimate the hydraulic conductivity of the riverbed materials, and flux through these materials. Previous work using temperature as a ground-water tracer (Constantz 1998, Constantz et al 2003, Johnson et al 2005) has shown heat to behave similarly to conservative chemical tracers under high velocity ground-water flow conditions.

2.3.1 Field Overview

The design of the temperature collection is modified from both (Johnson et al 2005 and Constantz 2003). Steel pipes were constructed into drive point piezometers in order to penetrate the coarse gravels and cobbles of the riverbed at the study sites (Figure 2.3). The piezometers were made out of 1.67 m galvanized steel pipe sections with an inside diameter of 3.2 cm. To each 1.67 m section, a series of 4, .64 cm diameter holes were drilled every 5 cm for 1.2 meters up the pipe. These holes were drilled 90 degrees from each other. The holes were tapered open and de-burred as was the inside of the pipe. The perforated pipe was attached to steel drive points and driven into the riverbed with a 9 kg slide hammer. A cast iron drive cap was screwed on top of the piezometer during installation to prevent the threaded end of the pipe from becoming deformed.

Inside the perforated piezometers a baffle system was installed to avoid preferential vertical flow through the pipe. Between the baffles, IBUTTONS (model DS1921Z), small temperature data loggers, were placed with a resolution of 0.125 degrees Celsius programmed to store temperature measurements every 15 minutes. The piezometer and baffle system was used to measure pore-water temperatures in the riverbed, Solinst level loggers (3001 Gold F5) with a resolution of 0.3 cm were used to
measure pressure and temperature in the River.

At the Mohawk site (Figure 2.2), the multiport monitoring well was instrumented with a temperature probe and several druck mini pressure transducers. These were all attached to a Campbell Scientific CR1000 data logger and programmed to collect readings every 15 minutes. The data logger was powered by 8 D-cell alkaline batteries, and in the summer, the batteries needed replacing every few months. Data quality appeared good during the warm summer months, but as the temperatures became colder, the equipment fell out of calibration. The battery life decreased to only several weeks and readings from the druck pressure transducers became erratic. The equipment was taken back to the lab for recalibration, but was not deployed again. Pressures and temperatures from the well equipment through the fall are considered reliable, but from December on, data is not used.

The USGS real-time gage (01168500) on the Deerfield River is several miles downstream of the Mohawk site, and about 1 mile from the Route 2 site (Figure 2.2). This gage keeps stage height and discharge data every 15 minutes. Depending on the flow of the river, the lag time for responding to stage changes on the Deerfield River from the Mohawk site is about 4 hours, and about 1 hour from the Route 2 site. Under most conditions it is observed that the magnitude of the stage change at each site corresponds with stage observed at each station. However, during several high stage events over the winter, the USGS gage was much more responsive than the pressure transducer deployed at the Mohawk site. It is possible that because the unregulated Cold and Chickley Rivers meet the Deerfield River downstream of the Mohawk site and
upstream of the USGS gage these two sites undergo different hydraulic conditions under high flow conditions.

### 2.3.2 Field Results

Data collection efforts including riverbed pore-water heads and pore-water temperatures are represented in Figure 2.4 for the northern Mohawk Trail State Forest site and Figure 2.5 for the Route 2 rest stop site in Charlemont. Each figure shows both riverbed pore-water and stream water temperatures in degrees Kelvin on the left Y axis. The relative stage in the stream is shown in cm on the right Y axis. Time, in units of days is shown on the X axis. In Figure 2.4, at the Mohawk site, the dashed line is the river temperature, the blue (x) line is the pore-water temperature 30 cm below the riverbed surface and the solid black line is the stream stage change in meters. The river temperature fluctuates by as much as 8 degrees (K) over the course of a day. The cause of the daily changes in temperature is solar heating during the day and cooling in the evenings. The erratic nature of the temperature fluctuations is believed to be due to the river characteristics at the northern site: large exposed cobbles and a shallow, fast moving river.

The riverbed pore-water temperature response at the northern site is clearly correlated to the stream stage change where warmer surface water is being flushed into the riverbed. In contrast, the decrease in surface water temperature with the stage change is not as easily understood. Surface water temperatures decrease at the same time the pore-water temperature spikes, yet at the upstream site it is unclear whether the decrease is a result of the diurnal cooling or instead a result of cooler surface waters being flushed down stream with each dam release. However, there is a small stage increase near Day 5
that does not correspond with a increase in surface water temperature. This supports the idea that a major cause of the surface water temperature variation is the influx of dam released water.

In Figure 2.5, observed data is plotted for the downstream Route 2 site. The dashed line is surface water temperature, the blue (x) line is pore-water temperature at 80 cm depth in the riverbed and the solid black line is the relative stage change of the river in meters. Unlike Figure 2.4, the peak in surface water temperature is matched with the moment just before the stream level change. At this site the pore-water temperature at 80 cm below the riverbed is slightly later in time, corresponding to the peak in stage. Over the course of 5 stage changes, the same sequence repeats itself.

Several site characteristics must be considered to explain the significant difference in the temperature responses between the northern and southern sites. The datasets for the two sites are not concurrent, though both were observed in the summer months. The depths of the pore-water temperature measurements below the riverbed are different: 30 cm at the northern site and 80 cm at the southern site. Measurement depths for each site were determined by the nature of the riverbed and the longevity of the pore-water temperature sensors. The primary difference between the datasets is the nature of the surface water temperatures. Part of this difference is attributable to the position of the pipes in a spot with more shade upstream and less shade downstream. Also significant, is the northern site’s proximity to the upstream dam and the resulting potential exposure to colder water after release. The Route 2 site is several kilometers downstream, plus is downstream of the unregulated Cold River the combination of which could cause a more even temperature distribution in the surface water.
2.4. Modeling Methods

To estimate the hydraulic conductivity of the riverbed sediments, a simple two-dimensional transient model of ground-water flow coupled with heat transport was designed using the commercial code Comsol Multiphysics. A cross-section measuring 1 m wide by 2 m deep was constructed with a finite element mesh (Figure 2.6). Models are run at steady state, and then the measured river temperature and head are input into the top of the model as boundary conditions. Thermal properties of the riverbed material are taken from literature, as are thermal fluid properties. The bottom of the model is a specified head and temperature boundary. The sides are both no flow and no heat flux boundaries. The only manipulated variable is the hydraulic conductivity of the sediments. Observation points were inserted at the same locations as the temperature sensors. The hydraulic conductivity values were adjusted until an approximate match of temperature responses was made. Table 2 lists the thermal properties for each model, while table 3 lists the hydraulic properties of each model.

The values of riverbed hydraulic conductivity for each site are estimates based on that specific local area. River environments are extremely heterogeneous, with sediments varying only over a few meters or less. Yet, given the differences in geologic setting between the upstream and downstream sites, significant interpretations can still be made. First, the Mohawk site is located in a steeper part of the bedrock valley, with the river abutting the outcropped schist on the opposite shore from the site. As a result, the water flows much faster in this reach of the river, across riverbed sediments consisting of boulders, gravel, and coarse sand. At the downstream Route 2 site, the river valley is broader, the river wider, and the river flows at a slower rate. The riverbed here still has
cobbles and gravel, but these are in addition to finer sediments mixed in from the down cut riverbank.

These simple two dimensional models begin to confirm the gradual change in depositional environments moving down valley. The resulting estimates of hydraulic conductivity are a useful tool to evaluate relative flux potential in riverbed sediments, but to more accurately estimate the flux of water through the riverbed, a more intensive data collection and modeling effort taking into account the multi-dimensionality of the flow paths is needed.

2.4.1 Modeling Results

Two sets of models were constructed to estimate the hydraulic conductivity of the riverbed at the two study sites. For each model, the model geometry is the same, but boundary conditions of river temperature and river stage are unique to the two sites. Simulated and observed temperature 30 cm below the riverbed are compared in (Figure 2.7). The solid black line is the river stage changing over several days the dashed red line is the change in river temperature over time. The (x) blue line is the observed temperature change 30 cm below the riverbed, and the green crossed line is the simulated temperature response. During the first two river stage pulses, simulated and observed temperatures match well, both do not respond to the third small pulse, but on the fourth pulse, the modeled and observed responses deviate. The model does not respond to the fourth stage pulse, while the field temperature probe does. One possibility is that the simple 2-dimensional models fail to capture possible 3 dimensional flow paths. In the models, there is a fixed lower head boundary through the duration of the simulation, it is
possible that the actual field bank storage could be more transient than the constant head boundary assumed in the 2-dimensional models.

Figure 2.8 compares modeled and observed temperature responses at the southern study site. The observed measurement point is 80 cm below the riverbed as is the modeled temperature. At this site, the boundary conditions are more regular. Solid black and dashed red lines are the river stage and river temperature respectively. The (x) blue line is the observed field temperature response and the crossed green line is the simulated. The magnitude of temperature response is a good fit for the first pulse however; there is a time lag in the simulated response. The time lag occurs at each pulse and the magnitude of the temperature response is within 1.5 degrees K. Over the period of data collection, there was a warming trend in the riverbed which was not captured in the simulated results. Changing hydraulic conductivity values can slightly affect the lag seen in the models, but by doing that the magnitude of the temperature change is overestimated. The models fail to account for potential 3 dimensional flow paths, particularly bank storage which could affect the timing of the temperature response.

2.4.1.1 Sensitivity Analysis

Both observed and best fit modeled temperature responses are shown in Figures 2.9 through 2.14. Figures 2.9-2.10 are sensitivity plots for the downstream Route 2 site while Figures 2.11-2.12 are sensitivity plots for the upstream Mohawk site. Figure 2.13 is a one day plot of model results with different hydraulic conductivities. Although the timing and magnitude of the temperature responses are very close, there are notable differences. In some of the modeled results a time lag appears between observed and
modeled temperature response. Also of note, for some stage pulses, the model is not very responsive, when compared to field data. To investigate these differences, sensitivity analyses were conducted on each of the above mentioned parameters.

As mentioned, Figures 2.11 and 2.12 show plots of observed temperature, best-fit modeled temperature, and modeled temperatures with different hydraulic conductivities at the Mohawk field site. The value of hydraulic conductivity was changed by 1 and 2 orders of magnitude higher and lower. Figure 2.13 is a close up of one stage change and the resulting temperature response observed both in the field and simulated in the model. Increasing the value of hydraulic conductivity causes two things to occur: First, the temperature responds faster to the stage change, lessening the time lag, but increasing the magnitude of the temperature rise. Expectedly, lowering the hydraulic conductivity value decreases the temperature magnitude and delays the temperature response to the stage change. The hydraulic conductivity values one order of magnitude in either direction are still close to observed temperatures, while hydraulic conductivity values two orders of magnitude higher and lower than the base case values result in unrealistic temperature responses.

Variations in the lower head boundary values were changed by a factor of two lower and higher. Base conditions reflect a gaining reach, with a lower head value of 1.3 m where the transient stage varies from .4 to 1.7m. Lowering the head boundary creates temperature variations which mimic the river temperature changes. Increasing the head boundary creates a non-responsive temperature value. It is important to note that the first two stage changes are modeled accurately, but the small third and moderate fourth stage pulses are not simulated at all. This is suspected to be due to a transient lower head
boundary, though field data was not taken to account for this. Lastly, changing values of thermal conductivity were found not to cause a significant change in the model temperature responses.

2.5. Discussion

Each site model has the same general response to the sensitivity analysis, but there are notable outlier temperatures in the analysis for the Mohawk models. One possibility for these results is the river temperature values collected in the field and used as boundary conditions in the model. These data do not coincide as closely as the Route 2 rest stop site data.

There are several differences geologically between the northern and southern sites, and those geologic differences create distinct ground-water responses to river stage changes. For example, the modeled best-fit hydraulic conductivity for the upstream Mohawk site is 1.5e-3 m/s, while at the downstream Route 2 site the best-fit approximation is 4.9e-4 m/s. The sediments upstream at the Mohawk site consist of cobbles, gravels, and coarse sands, while downstream cobbles and gravel are mixed with silts and clays, giving the riverbed sediments an overall lower hydraulic conductivity. Another key difference between the sites is the nature of the valley: upstream at the Mohawk site, the valley is much steeper which was the rationale in using larger lower head conditions in the Mohawk model. The Route 2 site is located in a broader part of the valley, where the Deerfield River flows more gently.

Although important insight can be gleaned from these simple two dimensional transient models and vertical temperature field data sets, there are many aspects of surface water / ground-water interactions in rivers that need to be addressed in three
dimensions. Field data and modeled results are point data, and do not capture the heterogeneity of riverbed systems. Through trail and error methods were found that worked best under our specific field conditions. With advancing technology, new techniques are being developed that may be better suited for larger scale investigations of ground-water / surface water interactions, such as fiber optic distributed temperature sensors which can be used to enhance spatial coverage of temperature response.

2.6. Conclusions

Understanding the sediment architecture of river valleys can be an important tool in predicting the spatial variability of surface water interactions with ground-water along the length of river. Geophysical surveys, field mapping and analysis of well and bore hole logs within the West Charlemont Basin in Summer and Fall of 2006 were used in conjunction with the morphosequence concept to predict sediment architecture and in extension, relative areas of surface water / ground-water flux. Surface water / ground-water exchange was estimated by observing changes in temperature of riverbed pore-waters at two study sites in the Charlemont basin. These observed data were used to calibrate simple two dimensional coupled models of ground-water flow and heat transport. Models of both sites were most sensitive to changes in the lower head boundary and riverbed hydraulic conductivity. Better constraint on the models can be achieved by measuring head at the bottom to the riverbed, as well as by a better distributed temperature logging network. Through these models, hydraulic conductivity of the riverbed materials was estimated at each study site and found to be higher at the upstream Mohawk Trail site than at the downstream Route 2 site. Differences in the
geologic setting within the same basin are the determining factor in the differences in surface water exchange with ground-water for these two sites.
Figure 2.1 The Massachusetts portion of the Deerfield River Watershed with the Charlemont study area boxed out in red.
Figure 2.2 The Deerfield River in the West Charlemont valley, Charlemont, MA.
Figure 2.3 Schematic of temperature sensor field deployment near riverbank. Steel pipe is driven into riverbed and pore-water temperature is measured at discrete depths below the riverbed.
Figure 2.4 Observed river temperature and pore-water temperature 30 cm below the riverbed at the Mohawk Trail State Forest. Black line is the relative change in river stage in meters.
Figure 2.5 Observed river temperature and pore-water temperature 80 cm below the riverbed at the downstream field site.
Figure 2.6 Schematic of the simple 2-dimentional ground-water flow and heat transport model with boundary conditions.
**Figure 2.7** Modeled and observed temperature responses to stream stage fluctuations from the upstream field site.
Figure 2.8 Modeled and observed temperature responses to stream stage fluctuations from the downstream field site.
Figure 2.9  Time series plot of the downstream models sensitivity to hydraulic conductivity.
Figure 2.10 A time series plot of the downstream models sensitivity to changes in the lower head boundary.
Figure 2.11 A time series plot of the upstream models sensitivity to changes in the lower head boundary.
Figure 2.12 Time series plots of the upstream models sensitivity to hydraulic conductivity.
Figure 2.13 A time series plot of the upstream site models sensitivity to hydraulic conductivity during 1 stage fluctuation.
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Table 2  Model properties related to heat transport.
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*Table 3* Parameters investigated in the sensitivity analysis of upstream and downstream models.
APPENDICES

A. GEOLOGIC SETTING

The geology of the northeastern U.S. is characterized by complex metamorphic bedrock packages overlain by glacial sediment, which can be a significant water resource for local communities especially in incised river valleys. Previous work by Gay et al. (1974) mapped the deposits as a medium yield aquifer with potential yields of 51-200 gpm (Figure A1). To assess the water bearing properties of the deposits, a hydrostratigraphic conceptual model needed to be created, and to accomplish this, an investigation including field mapping, seismic refraction surveys, borehole sediment samples and sediment analysis was conducted.

The flow and stage of the Deerfield River is controlled by several dams, used for flood control and hydroelectric power generation. While investigating the site geology, it became apparent that there was a strong relationship between the changing stage of the Deerfield River and the ground water system below. While the geology of the West Charlemont Basin can be separated into bedrock and surficial components; all ground water and surface water interactions investigated in this work take place in the surficial deposits. The bedrock geometry, however, is a controlling factor for the architecture of the surficial deposits, and therefore is an important factor in the architecture of the valley sediments.
B. BEDROCK GEOLOGY

The crystalline bedrock underlying the West Charlemont area consists of two units of Ordovician age: the Moretown formation and the Hawley formation (Chidester et al., 1967; Zen et al., 1983) (Figure A2). Both the Moretown and Hawley formations are strongly foliated, or layered. This foliation is oriented north northeast and dips steeply to the east at about 70° to 80° (Stanley and Hatch, 1988). The western portion of the study area is underlain by the Moretown formation, a light gray-green to buff, fine- to medium-grained schist and granulite (even sized, interlocking granular minerals). The Moretown formation contains a lens of round, reddish garnets, 2 to 5 mm in diameter, near its contact with the Hawley formation. The Hawley Formation, a schist and granulite rock interbedded with darker colored amphibolites, underlies the eastern half. These units have been incised by ancient drainage systems which at present form the Deerfield River Drainage Basin.

The geometry of these valleys is critical to the architecture of the surficial materials deposited within them. The canyons can be as deep as 1000 feet from valley bottom to hilltop. Most of the surficial sediments were derived from glaciers and transported by melt water (Figure A3). Sediments carried by melt water will fall out of suspension as depositional environments decrease in energy. Typically, steep and narrow valleys are high energy environments where coarser materials will be deposited. In contrast, broad, shallow valleys are typically lower energy environments where fine grained materials can settle out of the water column (Randall, 2001). Better constraining the geometry of these bedrock valleys is the first step in developing a conceptual model for the deposition of surficial materials.
The first task in assessing the geology of the study area was to compile a well database. Fifteen well logs and two geophysical profiles from previous work were found in the study area. The next step was incorporating these data into our base maps using a GIS. Areas where data points were especially sparse were identified and additional geophysical profile sites were selected to constrain the geometry of the basin. To better understand the stratigraphy and to gain insight into the depositional patterns within the basin, three borehole locations were chosen. Surficial mapping was conducted to the north and west near Pelham brook to locate potential upstream sediment sources.
C. GLACIAL GEOLOGY

The northeastern U.S. has been covered in ice several times in the Pleistocene. Melt waters from these previous glacial events have incised steep valleys into the crystalline basement rock throughout the northeastern U.S., and this is the case with the proto-Deerfield drainage network in the Deerfield River watershed.

During the last glacial maximum, the terminal moraine of the ice sheet formed at what are now Nantucket, Martha’s Vineyard, Block Island, and Long Island. It is believed that the ice sheet was up to a mile thick over the West Charlemont study area at which time glacial till was deposited over most of the region. These till deposits are generally thinner at the tops of mountains and hills and thicker in valleys.

As the ice retreated northward, the melt water from the glacier traveled through the previously incised drainage networks. Many of these drainage networks became damned with glacial debris, resulting in the formation of glacial lakes. These glacial lake systems are known to have complex sediment packages, with a wide variety of hydraulic properties. To explain the different sediment packages observed in the Charlemont basin, a conceptual depositional model is constructed, and enhanced by well and borehole logs, geophysical surveys and surficial geologic mapping.

Sediments in the study area are primarily of glacial origin and consist of glacial till deposited directly by glacial ice and stratified deposits laid down by melt-water streams or in temporary glacial lakes formed during ice retreat. Glacial till is a poorly-sorted homogeneous mixture of boulders, cobbles, gravel, sand, silt and clay that was deposited beneath glacial ice as the ice advanced over the region. It generally forms a thin veneer over the entire landscape, thinner in the hills and thicker in valley bottoms.
Stratified deposits generally consist of well-sorted, layered sediments deposited by glacial melt-water streams. Post-glacial stream terraces and modern floodplain deposits associated with the Deerfield River cover most of the glacial deposits in the valley. Stratified deposits can range in texture from coarse sand and gravel to clay depending on the depositional environment. Coarse-grained stratified deposits comprised of sands and gravels are most favorable for the development of high-yield water wells. Well yields ranging from 200 gpm to over 1000 gpm are possible. However, in areas containing thick deposits of clay or silty fine sands, the potential yield is greatly reduced and is unsuitable for the development of water wells with yields in excess of a few gallons per minute.

The bulk of the stratified deposits along the Deerfield River are restricted to the low-lying areas adjacent to the river. Aquifers associated with the coarser-grained, highly transmissive sands and gravels are typically located in areas where deeper and wider valleys were cut into the bedrock and are more or less isolated from one other.

In the west Charlemont area, the bedrock valley contains thick deposits of glacial and post-glacial sediments (Figure A3). The distribution and hydraulic characteristics of these sediments are controlled by four factors: the bedrock topography, depositional events, lake levels associated with ice retreat and the post-glacial alluvial processes associated with the Deerfield River. At the end of the last glacial maximum, the region was under several kilometers (km) of ice. During retreat of the ice the higher elevations and hilltops were exposed first followed later by the valleys. As the ice retreated up the Deerfield River valley, melt-water streams carried sediment out from the ice into the bedrock valley depositing the coarsest materials proximal to the ice margin and finer and
finer materials at increasing distances from the ice margin. Several ice contact deposits are found in the study area at Legate Hill and at the confluence of the Deerfield River with Pelham Brook (Figure A3). These deposits are exposed in many of the local gravel pits. As the ice kept retreating into the Vermont portion of the watershed, finer sediments were deposited on top of the coarser sediments creating a vertical sequence of materials that becomes finer in an upward direction. Clays and silts do occur at the site below the surface where the valley is wide and deep. Clays and silts in this abundance usually correspond to a standing body of water. Therefore, it is likely that a temporary glacial lake occupied the area at some time during ice retreat. In the narrower parts of the valley the sediments are coarser and the fine silts and clays are typically absent.

More recently, alluvial processes driven by the Deerfield River reworked the glacial melt-water sediments forming several floodplain terraces. These terraces formed as the Deerfield River down-cut into the melt-water sediments. It is likely that some of the fine-grained silts and clays that may have covered much of the valley floor immediately after deglaciation may have been eroded away during down-cutting. At least three terraces are observed in the study area. These deposits are comprised of coarse sands and gravels with abundant cobbles. In the northern part of the basin, the valley is steep and seismic refraction surveys show sediments to be over 120 ft deep. A well drilled down to 74 ft was sampled with a split spoon every 10 feet. Figure A4 depicts a cross section which incorporates the borehole and the seismic refraction line. Sediments are mostly fine sands and silts with some coarse sands and gravels near the bottom. A combination of boreholes, well logs, and seismic refraction lines were used to create cross sections at mid-basin and in the southern basin, where the valley is widest.
(Figure A5). Progressing southward down the valley a fining upward sequence of sediments starts to become present. At the southern-cross section a varved clay layer is 40 ft thick.

The difference in originally deposited material at the southern and northern cross-sections leads to different floodplain and riverbed materials. The northern site consists of coarse sands, gravels, and cobbles while the southern site, which still has cobbles, is mainly silts and clays. Differences in riverbed materials allow for differences in surface water exchange with ground-water at the two sites. This is the focus of the investigation undertaken in chapter two.

Figure A-1  Site map of the West Charlemont Basin including wells, borings, seismic refraction surveys, and cross sections. The original aquifer extent as mapped by Gay 1974.
Figure A-2 The two major bedrock types underlying the basin. All wells on this map are screened in the surficial deposits.
Figure A-3  Surficial geology of the West Charlemont Basin as mapped by Chidester et al (1967).
Figure A-4 Cross section through the Mohawk Trail State Forest in the northern portion of the West Charlemont Basin. This is the deepest part of the valley, with sediment thickness up to 140 ft. The sediments are mostly poorly sorted silts and sands with very little fines.
Figure A-5  Cross sections B and C in the middle and southern portions of the basin. Notice the thick clay unit in C.
Figure A-6 Rendering of the West Charlemont Basin with till as green, alluvium as red and ice contact deposits in light blue.
Figure A-7  Rendering of West Charlemont Basin with till as green and alluvium removed to show clay unit in the southern portion of the basin in black. Ice contact deposits are in light blue and silty sand is in light green.
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