THESIS

EFFECTS OF LONG-TERM PUMPING ON RECHARGE PROCESSES IN AN ALLUVIAL-BEDROCK AQUIFER SYSTEM

Submitted by

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ABSTRACT

EFFECTS OF LONG-TERM PUMPING ON RECHARGE PROCESSES IN AN ALLUVIAL-BEDROCK AQUIFER SYSTEM

The response to pumping in multi-aquifer systems involves complex processes which can significantly affect regional water budgets. Particularly where long-term pumping has occurred, drawdown might take decades to propagate regionally. Failure to incorporate changes caused by long-term pumping into regional hydrogeologic conceptual models can lead to mischaracterization of critical water budget components like recharge, inter-aquifer fluxes, and groundwater-surface water exchange. Accurate description of these budget components is necessary for managing water resources and making predictions about future water supplies.

This study analyzes long-term changes in an area of the Denver Basin aquifer system with high historical groundwater withdrawals to characterize the effects of long term pumping on recharge, inter-aquifer fluxes, and groundwater-surface water exchange. An evaluation of historical water level data (1960s to 2010s) documents large hydraulic head declines (>50m in some areas) and a deepening bedrock water table relative to the stream and alluvial aquifer. Results indicate a muti-decade transition from upward to downward hydraulic gradients in the vicinity of major streams, a change that affects the water budget of bedrock aquifers. Implications for regional water budgets are evaluated using a 2D variably saturated finite-difference model which quantifies fluxes across stream, alluvium, and bedrock interfaces in a vertical sequence. Modeling results demonstrate that long-term head decline can produce complex saturation conditions beneath the alluvial aquifer including a transition period of partial desaturation and ultimately a perched saturated zone in the alluvium underlain by an unsaturated region in the bedrock aquifer. The results illustrate how inter-aquifer fluxes eventually stabilize, with no further changes caused by additional lowering of the bedrock water table. Saturation levels and fluxes across interfaces are strongly dependent on geologic heterogeneity, particularly with respect to hydraulic conductivity contrasts between and within aquifers and the location and connectivity of channelized sandstones. Modeling results demonstrate the importance of considering heterogeneity and saturation when managing aquifers that have undergone long term pumping. The results of this study provide insight into the mechanics of long-term water budget change, including controls on the transition to induced recharge and recharge rates. This has important implications for assessing the aquifer response to ongoing and future stresses.

ACKNOWLEDGEMENTS

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I thank the Town of Castle Rock, Colorado for extending their mission to make Colorado groundwater use more sustainable by sponsoring this research. This experience has provided a unique opportunity to consider real challenges faced by groundwater managers in the State of Colorado.

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CHAPTER 1 – INTRODUCTION

Groundwater is a critical source of water for agricultural, industrial and drinking purposes around the world. With groundwater increasingly utilized to meet rising global water demands, it is important to consider the consequences of its development. Many aquifers are already facing the effects of over-pumping and groundwater depletion including decreased well yields, rising pumping costs, deteriorating water quality and subsidence of the overlying land (e.g., Konikow and Kendy, 2005; Wada et al., 2010). The interaction between groundwater and surface water resources poses additional challenges as diminished surface flows caused by groundwater pumping can negatively impact biological habitats, surface water availability, recreation, and water rights. Combatting mounting challenges to meet rising water demands requires management practices which seek to comprehensively understand the aquifer systems at hand. Optimal use of groundwater requires evaluating the response of water budgets and aquifer system processes to development.

The response of an aquifer system to development is commonly evaluated through the use of mathematical models. To be useful, mathematical models require a solid foundation in the form of a conceptual hydrogeologic model. For aquifers subject to development, the conceptual model may include processes that are only relevant after development is initiated or continued for some time. Failure to identify these processes reflects an inaccurate or incomplete conceptual model and can lead to substantial error in model predictions. Recently, authors have found that conceptual model uncertainty can contribute up to 30% of total uncertainty in models, making it one of the largest sources of error in hydrogeological modeling (Rojas et al., 2008). However, there are several reasons why modelers may disregard alternative conceptualizations of a problem. Beven (1993, 2005) discusses the concept of equifinality whereby many combinations of model structures and parameter sets can provide equally good reproductions of observed system responses. If a model produces reasonable results, why consider alternatives? Bredehoeft (2003, 2005) identified four additional reasons for the lack of conceptual alternatives, including the propensity for modelers to regard conceptual models as immutable, the added effort and time, the tendency for good calibration to mask weaknesses in conceptual models, and the singular focus on parameter uncertainty. As hydrogeologists around the world continue to document the increasing effects of groundwater development, the need for new conceptual models will grow with our refined understanding of aquifer processes. Whatever the reasons, the need to consider alternative conceptual models has been urged by many (e.g., Troldborg et al., 2007; Rojas et al., 2008; Bredehoeft, 2003, 2005; Beven, 1993, 2005).

The construction of alternative conceptual models is a separate challenge for which limited guidelines exist. In particular, regional scale models may take years to construct and calibrate with one conceptual model. Adding alternative conceptualizations without substantial evidence to support alternatives might be unwarranted given time and budget constraints. Further, the development of alternative conceptual models might conflict with parsimony and add computational burdens. For these reasons, it may be necessary to justify the evaluation of alternative conceptualizations and processes at smaller or localized scales before applying them to comprehensive, regional groundwater modeling efforts.

This thesis investigates the hydrogeologic framework and flow processes within to the Denver Basin aquifer system in central Colorado which has been significantly impacted by development over the past century and for which an immutable conceptual model may be hindering sustainable management practices. The complex, multi-aquifer system is comprised of a series confined and unconfined bedrock aquifers overlain by an alluvial aquifer system that drains the adjacent Rocky Mountain Front Range (Robson, 1987). The bedrock aquifers have exhibited large hydraulic head declines of up to 50 meters in heavily pumped regions, and water managers are imminently challenged with supplying water to a growing population.

The current conceptual model for the system assumes a small amount of recharge to the bedrock aquifers through processes which include precipitation and agricultural and urban return flows. However, drastic changes to the aquifer system from long term pumping may challenge this conceptual model assuming limited recharge. Historically, bedrock aquifer hydraulic heads were commonly well above the alluvial aquifer and streams. Bedrock aquifers often exhibited artesian pressures and made significant contributions to alluvial groundwater and surface water flows (Robson, 1984; Paschke et al., 2011). However, rising groundwater extraction rates since the 1950's have substantially altered the relationship of bedrock hydraulic heads to the overlying alluvial aquifer and surface streams. This changing relationship could fundamentally alter processes occurring in the aquifer system like groundwater-surface water exchange and bedrock aquifer recharge.

While several modeling efforts have evaluated the aquifer system's response to development (e.g., Robson, 1987; Banta, 1989; Watts, 1995; Paschke et al., 2011), few have considered changes in recharge rates and exchanges between the alluvial and bedrock aquifers. Previous models have assumed that diffuse precipitation and irrigation return flows are the only recharge sources to the bedrock aquifer system. Additionally, the highly heterogeneous aquifers have been represented by individual, anisotropic layers and conceptual models have assumed a consistent hydraulic connection between the alluvial and bedrock aquifers despite declining bedrock water levels in relation to the alluvium. While this conceptual model may have once been accurate for representing the system, it fails to account for processes which are induced from development like increased rates of stream seepage, induced groundwater recharge, and decreased bedrock discharge. These components of the water budget are critically important for making predictions about future supplies. For example, additional recharge to bedrock aquifers may increase the supply of groundwater resources in a region challenged by rising water demands and complex water legislation.

The most recent and comprehensive model of the Denver Basin aquifer system created by the U.S. Geological Survey (USGS) evaluated basin wide changes over the past century in response to development (Paschke et al., 2011). Extensive data assimilation and calibration efforts went into the production of the model and, overall, results were considered to provide a reasonable representation of the system (Paschke et al., 2011). However, in some areas, including portions of Douglas County, CO, unrealistically low hydraulic conductivities were required to calibrate the model, and large head residuals (>50m) were still observed. This suggests that the model is not capturing some aspect of the hydrologic system. Douglas County has been identified as a particularly heavily pumped region of the Denver Basin aquifer system characterized by large declines in bedrock aquifer hydraulic heads (Barkmann et al., 2015). The response of the aquifer system to development in this area has not been studied in detail, despite the urgent need to quantify capture mechanisms and evolving water budgets. Thus the Denver Basin aquifer system offers a unique opportunity to explore whether an alternative conceptual model might better characterize surface water, alluvial and bedrock groundwater interactions so that future regional scale models may constrain uncertainty and accurately represent system processes.

This thesis investigates the hydrogeologic system of the Denver Basin aquifer system in Douglas County, Colorado, with specific emphasis on the stream, alluvial and bedrock aquifer interfaces to evaluate whether alternative conceptual models and processes may exist for surface water, alluvial and bedrock groundwater exchanges. Specific objectives include:

- Resolve the heterogeneity and hydrogeologic architecture of alluvial and bedrock aquifer interfaces to determine if this affects rates of groundwater exchange
- Characterize surface water, alluvial and bedrock groundwater interactions in a representative study reach.
- Evaluate changes in hydrologic interactions between the alluvial and bedrock aquifer through time.
- Develop a 2-dimensional groundwater model to evaluate water budgets and saturation states for a data-conditioned hydrogeologic setting.
- Formulate a conceptual hydrogeologic model that describes the relationship between alluvial and bedrock aquifers and considers the potential for pumping-induced hydraulic disconnection of the aquifers.

CHAPTER 2 – HYDROGEOLOGIC SETTING

2.1 General Hydrogeology

The Denver Basin aquifer system is an asymmetric, bowl shaped structure comprised of Late Cretaceous through Paleogene-age sandstones, mudstones and shales which together form a series of confined and unconfined bedrock aquifers known as the Denver Basin aquifer system (Figure 2-1) (Robson, 1987). For administrative purposes, the aquifer system is divided into four distinct aquifers, termed from oldest to youngest, the Laramie-Fox Hills, Arapahoe, Denver and the Lower and Upper Dawson aquifers. While administrative designations do not necessarily represent the complex geologic architecture, the terminology is still widely used by hydrogeologists and water administrators (Raynolds, 2002). The aquifer system is bounded on the west by the central and southern Front Range uplift, and on the east by a gentle onlapping relationship onto the mid-continent platform. Stratigraphic units outcrop at aquifer margins and dip towards the south central downwarped axis (Raynolds, 2002).

Older strata reflect a nearshore environment of the retreating Cretaceous Western Interior Seaway and are composed of aggradational sandstone bodies, marine shale and coal which grade into overbank mudstone beds and fluvial channel sandstones. Younger strata reflect two distinct pulses of uplift, erosion and sedimentation associated with the Laramide orogeny (Raynolds, 2002). The two synorogenic pulses are identified as the older "D1" and younger "D2" sequences, which are characterized by deposits from fluvial systems draining the Front Range. Sediments from the D1 and D2 include coarse, clastic alluvial deposits, fluvial channel sandstones and overbank mudstones which rapidly grade into finer grained, paludal and lignitic mudstone and coal deposits. The sequences are bounded by unconformities and separated by a mature paleosol series characterized by red and purple iron oxide, intense mottling and gravelly channel fills (Raynolds, 2002). The D1 sequence contains the Arapahoe, Denver, and Lower Dawson aquifers and the D2 sequence contains the Upper Dawson aquifer. Surface drainage networks flowing across and among the basin since the Late Tertiary epeirogenic uplift include modern tributaries of the South Platte and Arkansas Rivers (Raynolds, 2002).

2.2 Site Hydrogeology

The hydrogeologic setting for this study includes the upper portions of the D1 and D2 sequences, administratively termed the Lower and Upper Dawson Aquifer. These units crop out at the surface and are incised by the South Platte alluvial aquifer in Douglas County (Figure 2-1). The D1 alluvial, fluvial and paludal sediments include fine-grained mudstone beds and coarse sandstone beds. At the base of the D2 sequence is a mature paleosol series overlain by coarse, multistoried channel sandstone beds separated by overbank mudstones and occasional paleosols (Raynolds, 2002). The hydraulically conductive aquifer material primarily includes the coarse grained, poorly to moderately consolidated sandstone beds. Individual sandstone beds are commonly lens shaped and range in thickness from a few centimeters to as much as 60 meters (Robson and Romero, 1981).

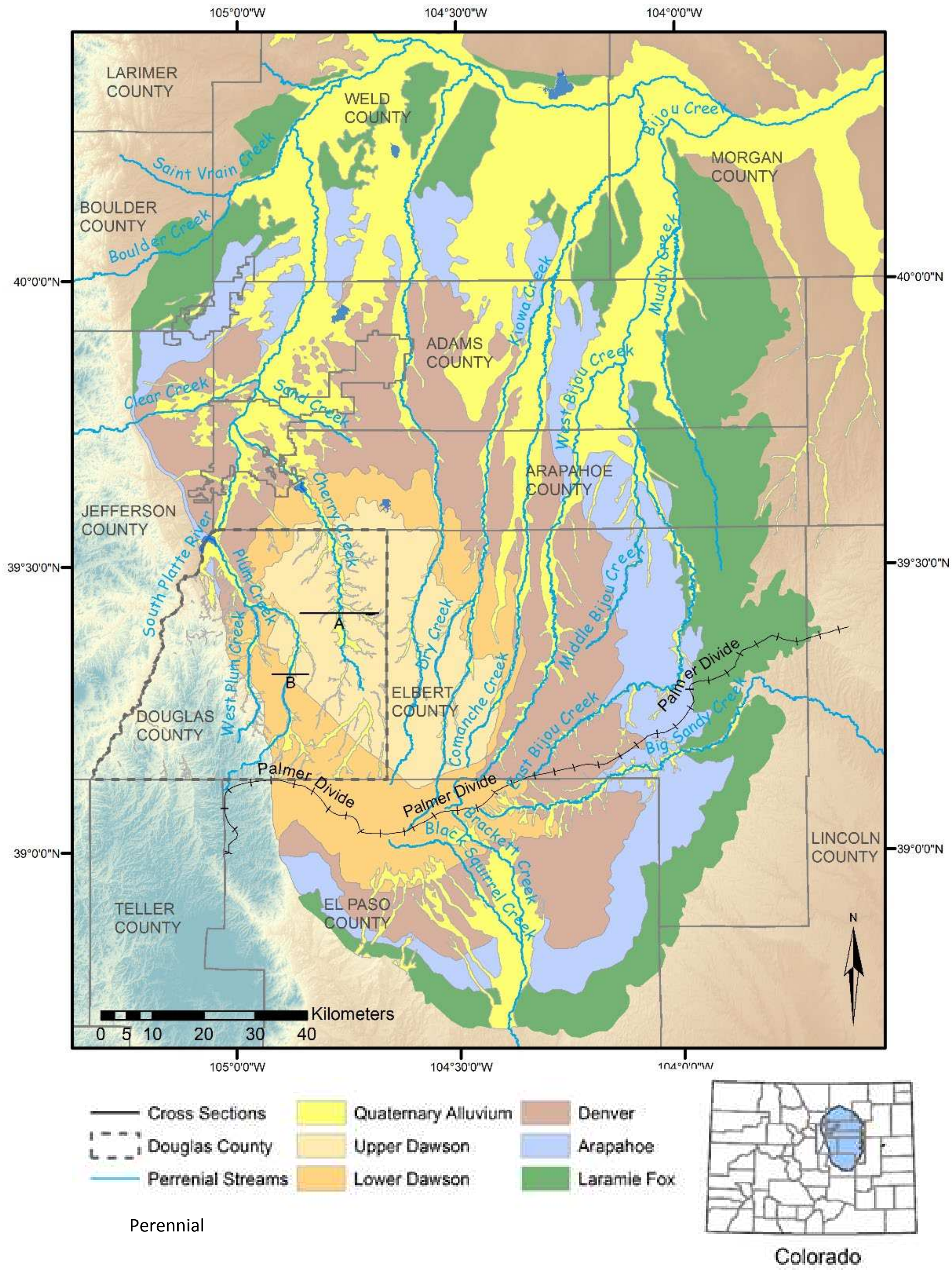

Figure 2-1 Map of the Denver Basin Bedrock and Alluvial Aquifer Systems. Bedrock aquifer delineation by Everett (2014), alluvial aquifers by SPDSS (2004) and Barkmann, et al. (2015). Universal Transverse Mercator, Zone 13, NAD 1983.

Farnham and Kraus (2002) and Raynolds (2002) have done considerable work to characterize the D1 and D2 sequences. The schematic diagram shown in Figure 2-2 is extrapolated from their work and illustrates the vertical and lateral relations of channel sandstones, mudstones, lignite beds and paleosols. The interpretations by Raynolds (2002) and Farnham and Kraus (2002) and boring log data obtained from the Colorado Division of Water Resources (CDWR) were used to create site specific cross sections which are part of the conceptual hydrogeological model (Figure 2-3 and 2-4)

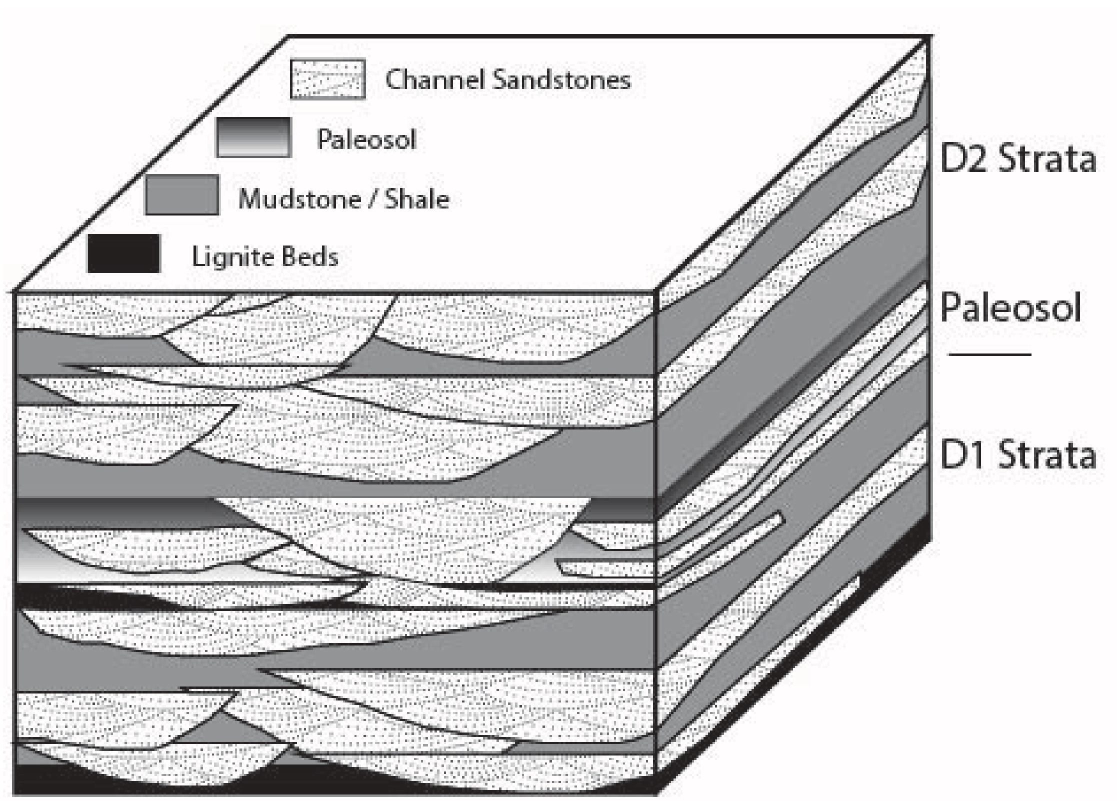

Figure 2-2 Schematic diagram of D1 and D2 fluvial architecture. Paleosol interpretation from Farnham and Kraus (2002) and D1 and D2 interpretation from Raynolds (2002).

In Douglas County, Quaternary aged alluvium follows modern drainage patterns of South Platte River and its tributaries including major traces along Cherry and Plum Creeks. To a lesser extent, alluvium exists as gravel deposited by paleo stream systems and sand and loess deposited by wind (Barkmann et al., 2015). Alluvium ranges from 1 to 90 meters in thickness and from a few hundred to over 2,000 meters in width. Alluvial traces generally widen towards the main channel of the South Platte River which lies to the north and west of Douglas County (Barkmann et al., 2015). Stream-deposited alluvium consists of poorly sorted mixtures of unconsolidated sand, gravel and clay lenses. Bedrock underlying the alluvium is generally less permeable, creating a boundary with contrasting hydraulic properties at the base of the alluvium. Groundwater in the alluvium is hydraulically connected to the surface water flows of the South Platte River and its tributaries. The alluvial aquifer hydraulic conductivity varies between 30 and 600 m/day (Alzraiee et al., 2017).

Two cross sections were constructed along transects perpendicular to Cherry Creek and East Plum Creek (Figure 2-1) to resolve the hydrogeology of the upper alluvial and bedrock aquifers. Geologic data was obtained from boring logs available through the Colorado Division of Water Resources (CDWR) well permit database, and the interpretation of sedimentary architecture was based on Raynolds (2002) and Farnham and Kraus (2002) as described above. Appendix A includes an example boring log and a table of boring logs used in the creation of the cross sections (Table A-1). National Map elevation data with 10-meter resolution (USGS, 2018) were used to generate the land-surface elevation profile and to verify boring log ground surface elevations. Cross section "A" shown in Figure 2-3 spans 15,000 meters west to east across Cherry Creek and the surrounding alluvial aquifer and Upper Dawson bedrock aquifer and covers 375 meters in elevation from 1,625 to 2,000 meters above sea level (m amsl). The orientation of the cross section is perpendicular to modern channel flows, similar to Figure 2-2. Cross section "B" spans 7,000 meters across East Plum Creek and the surrounding alluvial aquifer and Upper and Lower Dawson bedrock aquifers and covers roughly 300 meters in elevation from 1700 to 2,000 m amsl (Figure 2-4)

The cross sections reveal several important features that influence the site hydrology and groundwater flow. Both cross sections portray an unconsolidated alluvial aquifer incised into a highly heterogeneous, sandstone and mudstone bedrock aquifer. The majority of each alluvial aquifer base is bounded by low permeability mudstone rather than sandstone. This suggests a large contrast in hydraulic conductivity at the alluvial and bedrock aquifer interface that could influence alluvial-bedrock groundwater exchange. Channelized sandstone units appear vertically and laterally discontinuous, although greater connectivity of the sandstone is likely to occur in three dimensions. The complex arrangement of channels, packaged within less permeable mudstone, indicates that both horizontal and vertical groundwater flow are influenced by lithologic heterogeneity.

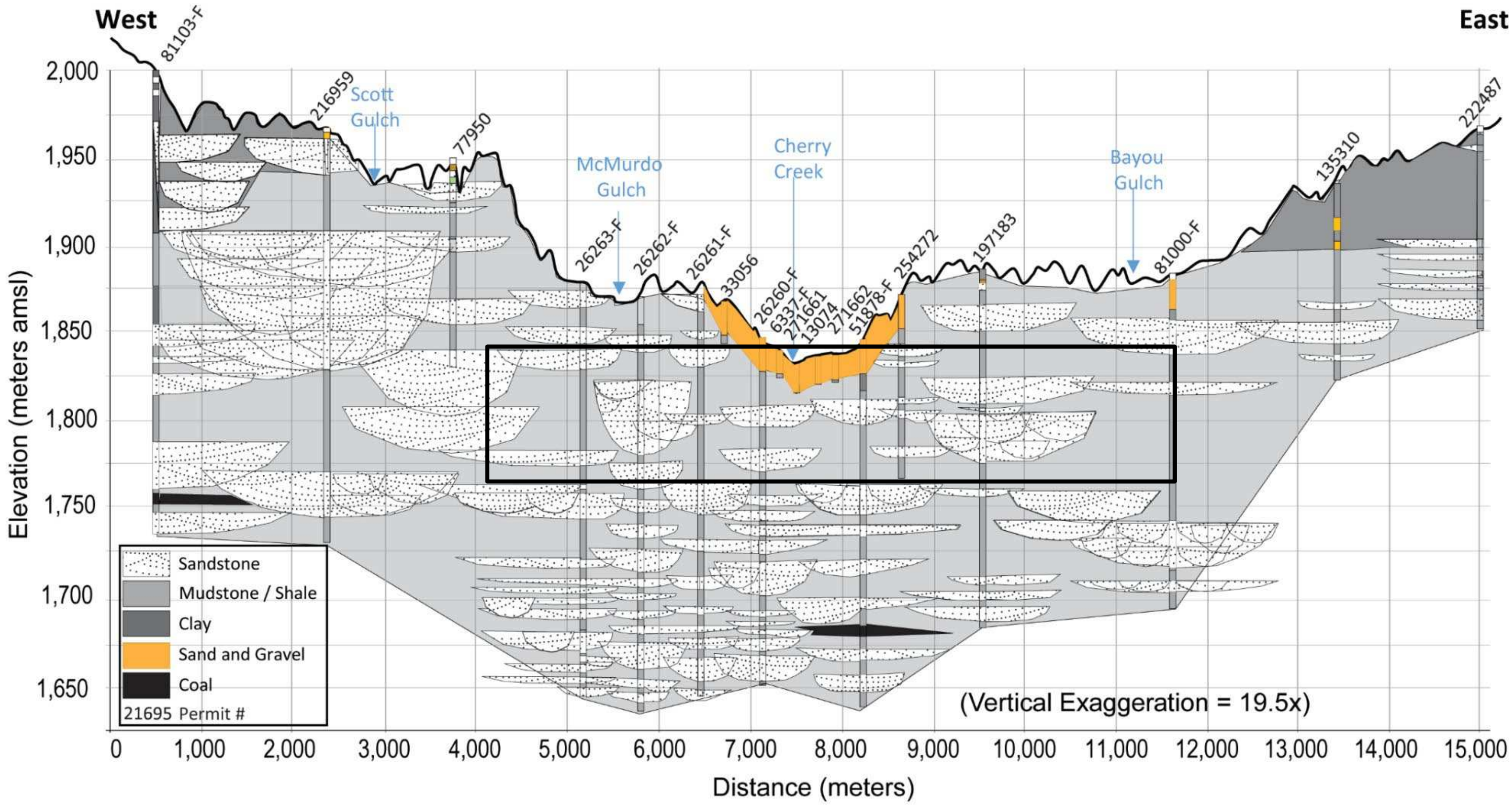

Figure 2-3 Geologic cross section for transect A-A' across Cherry Creek Alluvium and Upper and Lower Dawson Aquifers. The black box depicts the modeled area described later in this study.

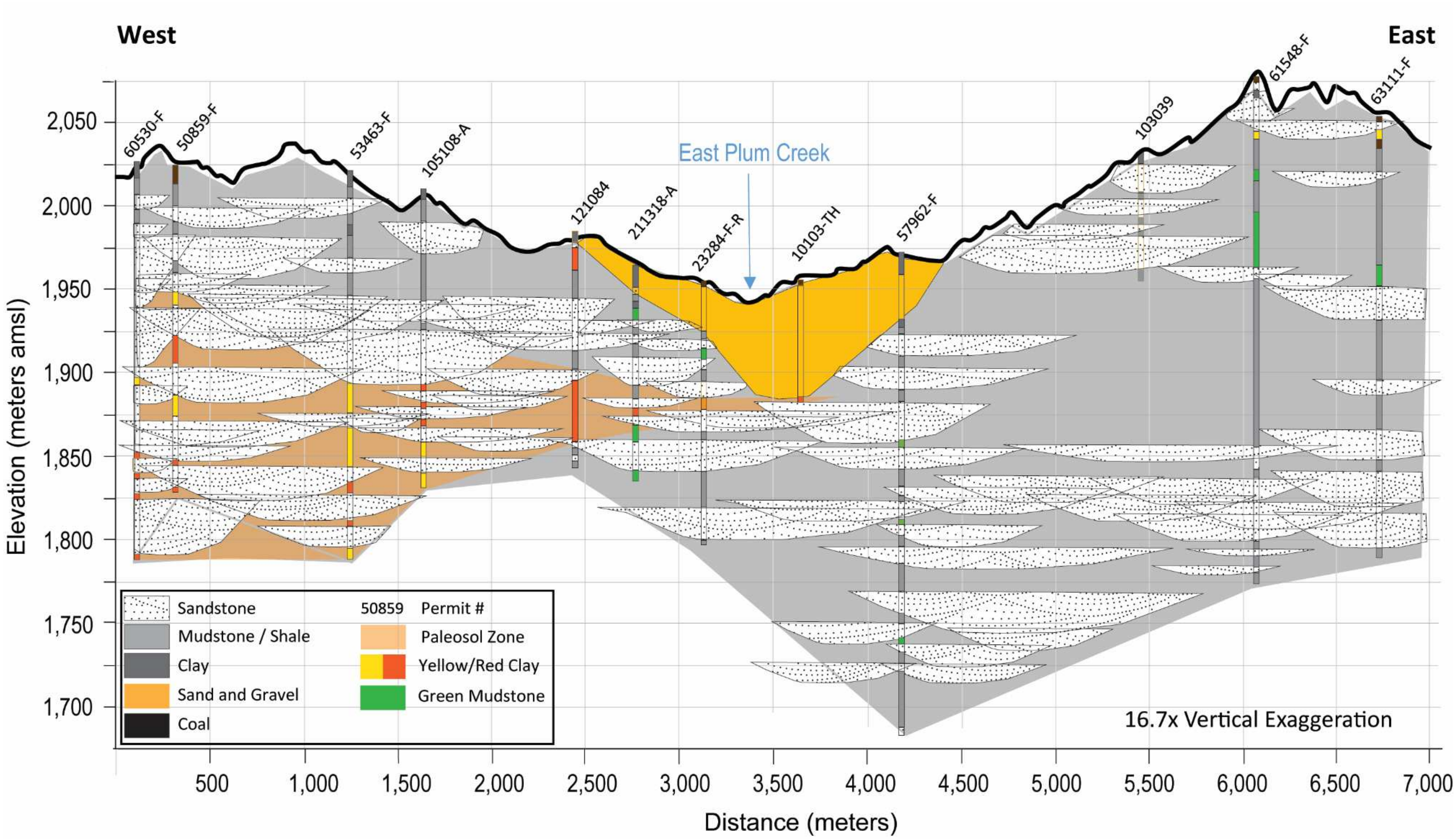

Figure 2-4 Geologic cross section for transect B-B' across East Plum Creek Alluvium and Upper and Lower Dawson Aquifers

2.3 Sources of Aquifer Recharge and Discharge

The hydrologic water budget for the regional groundwater system is influenced by the semi-arid climate, mountain front setting, and present and historic groundwater pumping within the basin. Inputs, or recharge, to the groundwater system include precipitation, seepage from streams, irrigation return flows, and mountain block recharge from fractured crystalline aquifers (Paschke et al., 2011; Barkmann et al., 2015). Outflows, or discharge, from the system includes flow to surface streams and springs, evapotranspiration, and for the last 130 years, pumping from wells (Paschke et al., 2011). Additionally, each aquifer has an individual water budget that accounts for inter-aquifer flows. The uppermost bedrock aquifer discharges to the alluvial aquifer, typically in topographic lows, where bedrock hydraulic heads are greater than alluvial hydraulic heads. Likewise, in regions where bedrock hydraulic heads are lower than alluvial aquifer heads, the alluvium recharges bedrock aquifers. As pumping has lowered regional hydraulic heads in the Denver Basin aquifer system, net recharge from the alluvial to bedrock aquifers has increased while discharge from the bedrock to alluvial aquifer has decreased. This effect was previously identified in the basin-wide numerical modeling performed by Paschke et al. (2011) (Figure 2-5). A conceptual illustration of water budget components for the Denver Basin bedrock aquifers is presented in Figure 2-6.

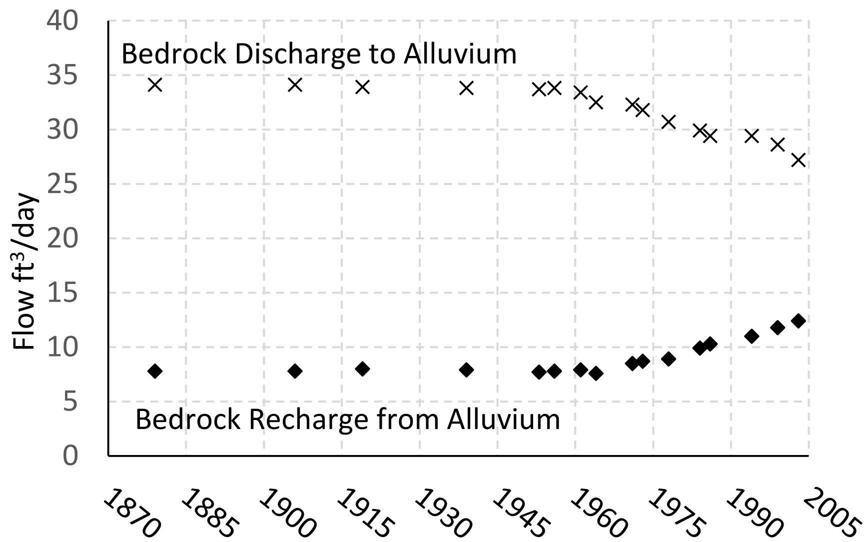

Figure 2-5 Simulated alluvial-bedrock inter-aquifer flow (Paschke et al., 2011).

Prior to development, the Denver Basin aquifer hydrologic system existed in a state of relative equilibrium (Paschke et al., 2011). The majority of precipitation either entered surface water and the alluvial aquifer system or was lost to evapotranspiration. A small amount recharged bedrock aquifers in upland outcrop areas through direct recharge or seepage from alluvial streams and aquifers. Bedrock groundwater moved laterally towards lower discharge areas, such as stream valleys and alluvial aquifers, or flowed vertically downward through leaky confining layers into deeper bedrock aquifers. Groundwater that entered the alluvial system eventually returned to surface streams where it was transported out of the Denver Basin aquifer system or lost to evapotranspiration (Robson and Romero, 1981; Paschke et al., 2011). Many of the aquifer hydraulic heads were artesian at this time, forming springs that discharged water to shallow aquifer units along natural joints and faults. Alluvial aquifers and stream valleys were generally situated below bedrock water tables and received groundwater discharge.

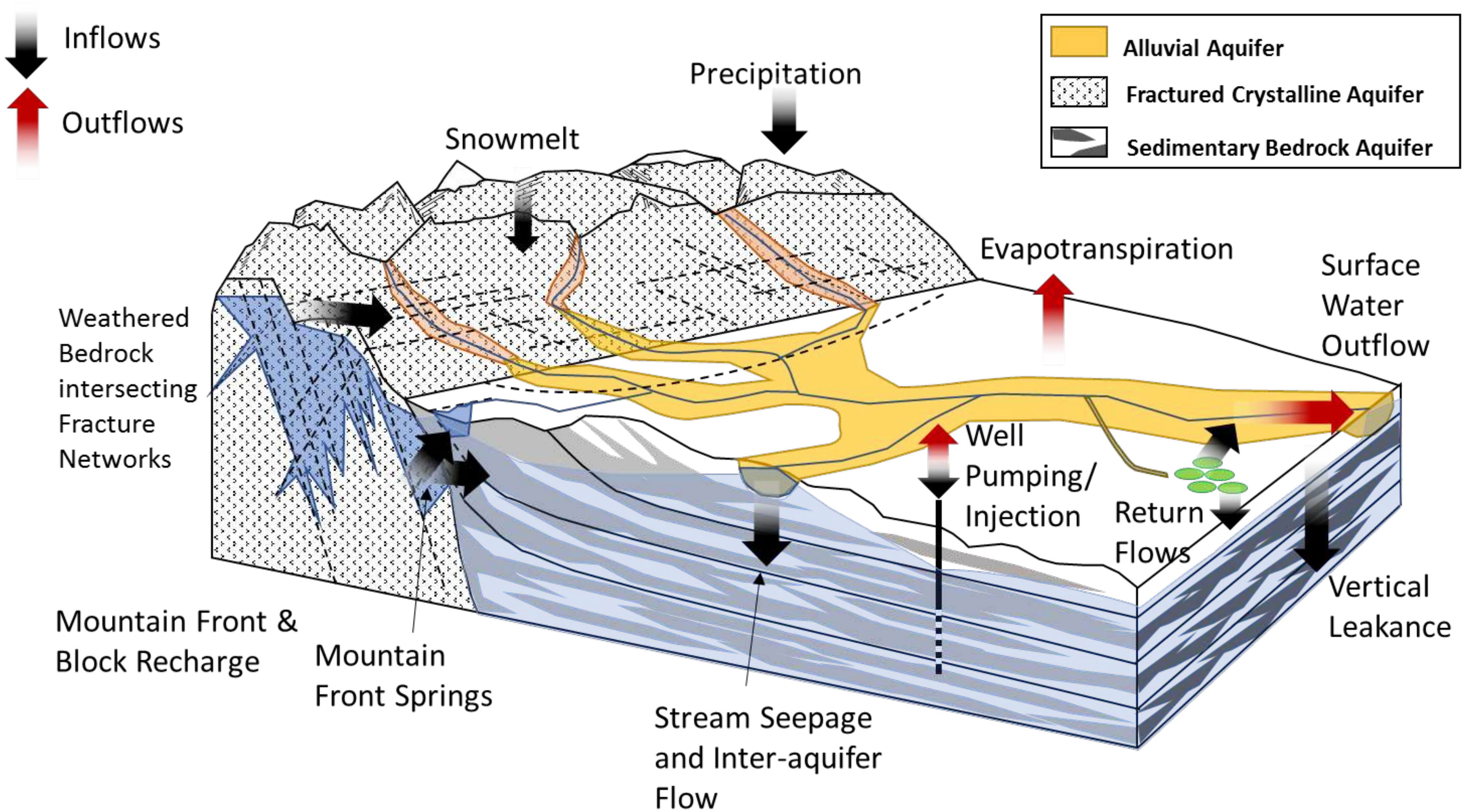

Figure 2-6 Conceptual illustration of water budget components for the Denver Basin bedrock aquifers.

By the late 1800's, just a decade after the first Denver Basin aquifer system water wells were drilled, artesian pressures were mostly eliminated due to pumping (Emmons et al., 1896). The following decades were marked by exponential population growth, the sweeping availability of electricity, and advancements in drilling and pumping technologies which accelerated groundwater utilization. By 1981, increased groundwater pumping had caused bedrock hydraulic heads to decline as much as 50 meters in heavily pumped regions (Robson and Romero, 1981; Paschke et al., 2011).

The commencement of development marked a significant transition away from equilibrium conditions in the Denver Basin aquifer system to a new system influenced by the transient stresses of rising groundwater pumping rates. Bedrock groundwater still flows vertically and laterally as before, however, it also flows toward pumping wells which act as localized discharge areas (Paschke et al., 2011). Reduced hydraulic head gradients have diminished flows from bedrock to alluvial aquifers, and substantially lowered hydraulic heads have induced recharge from the alluvial to bedrock aquifer system. As bedrock aquifer drawdowns persist, the hydraulic head difference between the alluvial aquifer and bedrock aquifers is increasing. The mechanisms, changes in recharge, and state of hydraulic connection between the two aquifers remains poorly understood.

CHAPTER 3: METHODS

3.1 Groundwater-Surface Water Interaction

Groundwater-surface water interactions were investigated through pressure data collected from a series of nested piezometers along a stream reach in Douglas County, CO to characterize surface water-alluvial groundwater interactions in a representative study reach.

3.1.1 Piezometer Nests and Vertical Hydraulic Head Gradients

Four pairs of nested shallow and deep piezometers were installed along a 700 meter stretch of East Plum Creek near Cross Section B (Figure 3-1) to evaluate vertical hydraulic gradients and seepage fluxes along the streambed. Piezometer nests were constructed of 3-cm diameter steel pipes with 0.2-cm horizontal slots spanning 5- to 10-cm from the pinched bottom of the pipe. Each pipe was connected by two steel cross bars and a locking bar to prevent tampering and enhance stability in the streambed as shown in Figure 3-2. The shallow piezometer screen (slotted interval) was manually set to a depth just below the streambed, and the deep piezometer slotted interval was set roughly 1 m into the streambed such that the midpoints of these intervals (the measurement depths) were separated by a known distance and the tops of the pipes were level as shown in Figure 3-3

Unvented pressure transducers (Solinst Model 3001 Levelogger) were installed in each piezometer and set to record an absolute pressure reading every 10 minutes. Raw absolute pressures were used to calculate vertical gradients between the shallow and deep piezometer using the equation:

$$\frac{\Delta h}{\Delta z} = \frac{h_{shallow} - h_{deep}}{\Delta z} = \frac{(\psi_{shallow} + z_{shallow} + p_{atm}) - (\psi_{deep} + z_{deep} + p_{atm})}{\Delta z} \quad \text{(Eq. 3-1)}$$

The elevation head (z), fluid pressure head (ψ), atmospheric pressure head (p_{atm}), and total hydraulic head (h) are in length units relative to an arbitrary datum (Figure 3-3). Atmospheric pressure is assumed equal for each piezometer and cancels out.

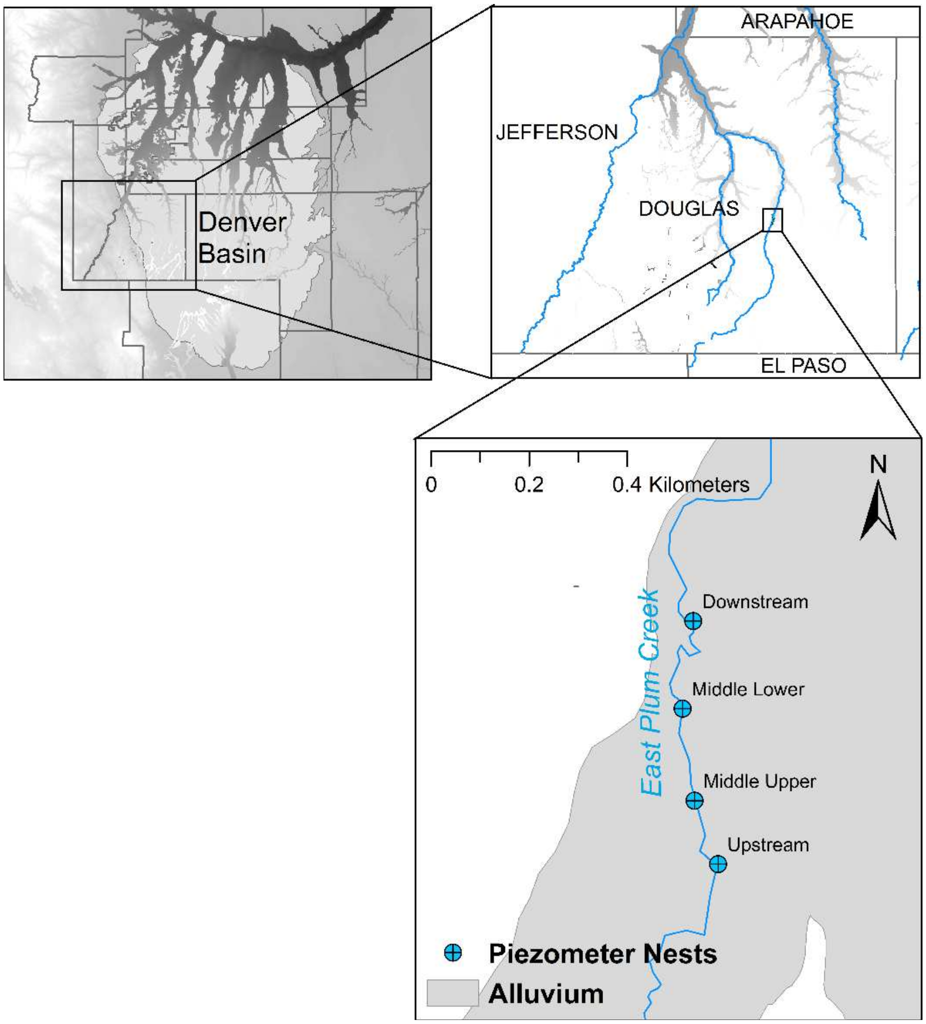

Figure 3-1 East Plum Creek Study Area showing piezometer nest locations and surficial extent of the alluvium.

Figure 3-2 Photograph of piezometer nest deployed in stream.

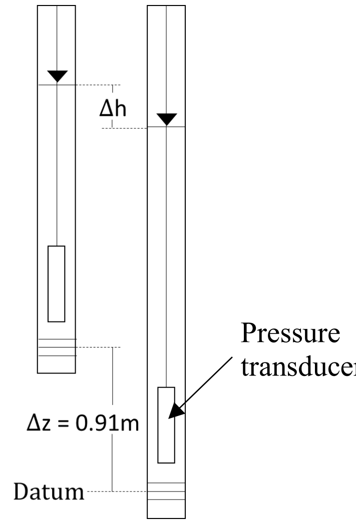

Figure 3-3 Piezometer nest configuration with pressure transducers.

3.1.2 Darcian Flux Calculations

Water exchange between the stream and alluvium was evaluated using field measurements and a 1-D Darcian flux calculation. The Darcian flux (L/T) provides an estimate of specific discharge, or volumetric discharge per unit area, at the streambed. Specific discharge values were subsequently used to evaluate groundwater-surface water exchange and to constrain the groundwater model. Specific discharge (q) was calculated using the equation:

$$q = K_{sb} \frac{dh}{dz} \qquad \text{(Eq. 3-2)}$$

where K_{sb} is the streambed vertical hydraulic conductivity, and dh/dz is the vertical gradient as described in Section 3.1.1. Streambed hydraulic conductivity measurements were taken from Ronayne et al. (2017) who employed falling-head tests to estimate K_{sb} at each of the piezometer nest locations. The average K_{sb} value along the East Plum Creek Study Reach was estimated as 25 m/d.

3.2 Spatiotemporal Analysis of Bedrock Aquifer Water Levels

To evaluate changes in hydrologic interactions between the alluvial and bedrock aquifer through time, an analysis of bedrock aquifer water levels through time was conducted for Douglas County, CO. Well water levels in the unconfined regions of the Upper and Lower Dawson aquifers in Douglas County were compared to mapped base elevations of the alluvial aquifer at decadal intervals from 1970 through 2010. Only unconfined regions of the bedrock aquifers were evaluated because confined regions are inherently separated from the alluvium by a confining unit. Unconfined regions were considered to be areas where the aquifers outcropped at the surface or were not overlain by a confining unit. Although the Denver aquifer crops out in the northwest part of Douglas County (Figure 2-1), it was not considered in this analysis due to insufficient water level data from wells. Water levels measured during spring months (February, March and April) were used to best characterize the annual period when hydraulic heads are recovered from the effects of late spring through fall pumping. Figure 3-4 shows a typical hydrograph for a Denver Basin aquifer system observation well with annual trends in hydraulic heads.

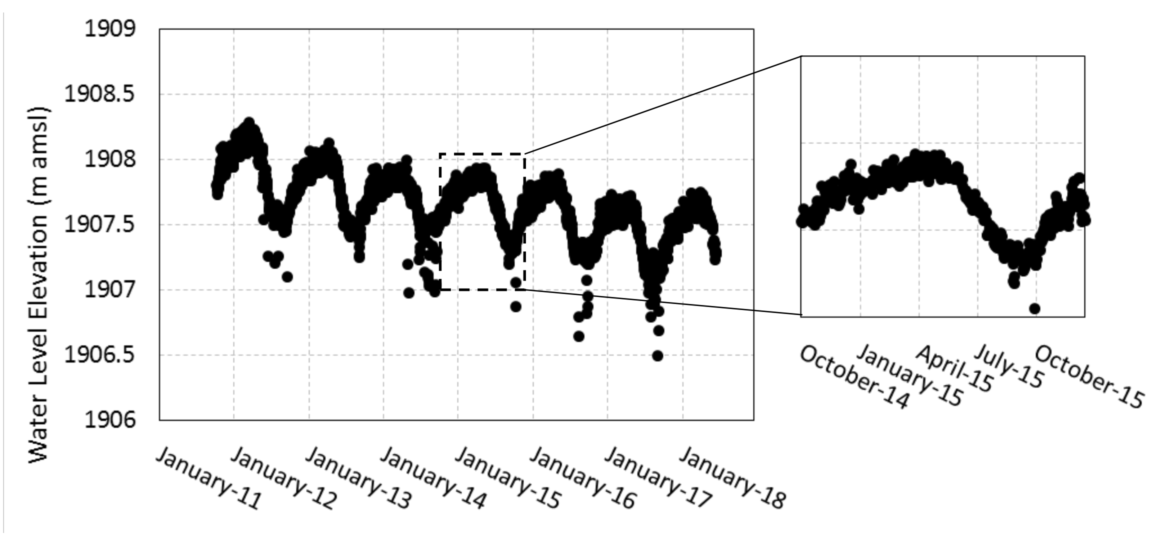

Figure 3-4 Typical water level hydrograph for a Denver Basin aquifer system observation well (USGS Site ID: 392412104434201) with annual peak water level occurring in the spring.

Water level data were obtained from the Colorado Department of Natural Resources (CDNR, 2018) and the USGS National Water Information System (USGS, 2018). For each well location, available spring water levels within each decade were averaged to produce a set of spring values for that particular decade. For the 1970's decade, 1960's water level records were also utilized due to sparse data coverage. While most well locations had data coverage for each decade, several locations had only one or two decades of data coverage. A visual sensitivity analysis confirmed that removing wells with less than 20 years of coverage did not significantly affect results, but substantially reduced the spatial extent of the data coverage. For this reason, even locations with a single decade of data were used in generating figures and performing subsequent analyses. Final water levels, well locations, and well identification numbers used in the spatiotemporal analysis are reported in Appendix B. Average decadal spring water levels were interpolated using the inverse distance weighting (IDW) method to produce a raster surface of the average bedrock aquifer water table position for each decade (1970-2010). The interpolation utilized a grid with 50-meter by 50-meter cell sizes.

The water table elevation for each decade was then subtracted from the base of the alluvial aquifer across the extent Douglas County. Elevation contours for the base of alluvium were obtained from the Douglas County Groundwater Study georeferenced contour datasets (Barkmann et al., 2015) and converted to a raster surface with 100-meter by 100-meter cell sizes. A raster subtraction function was used to generate a raster surface representing the distance between the bedrock water table and the base of alluvium for each decade. This process is illustrated in Figure 3-5.

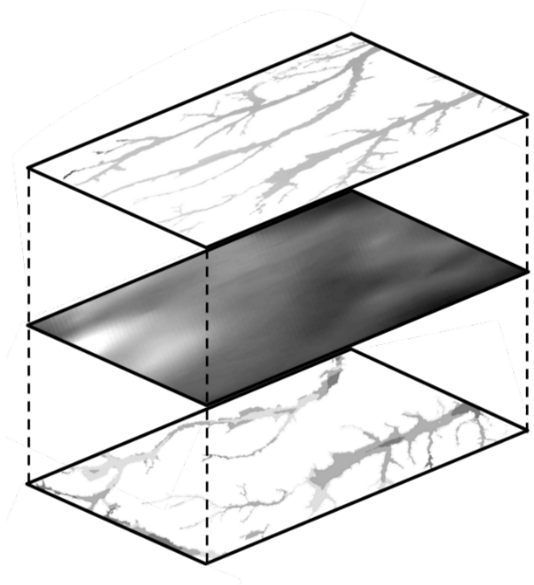

Figure 3-5 Schematic illustrating raster subtraction method.

3.3 Groundwater Model

Variably-saturated flow modeling was performed to further investigate groundwater flow within the stream-alluvium-bedrock aquifer system. The two-dimensional (2D) numerical model was oriented along a cross section perpendicular to the streamflow direction and was designed to (i) explore the influence of subsurface heterogeneity on saturation states and groundwater flow directions; and (ii) quantify how regional head declines (over decades) impact alluvial-bedrock aquifer interaction and bedrock aquifer recharge.

3.3.1 Model Construction

The groundwater flow model was developed using the finite-difference, variably saturated groundwater flow modeling package MODFLOW-SURFACT (HydroGeoLogic, Inc., 2002) with the commercially available graphical user interface Groundwater Vistas version 6 (Rumbaugh and Rumbaugh, 2011) used as a pre- and post-processor. Finite difference models use iterative procedures to solve for hydraulic heads throughout a computational domain by discretizing a governing mathematical equation over a spatial grid. To evaluate how the degree of saturation beneath the alluvium would affect recharge rates, variably saturated flow equations were required. Variably saturated flow models require more advanced numerical computations than fully saturated flow models because hydraulic conductivity and aquifer storage properties become functions rather than constants. MODFLOW-SURFACT formulates relationships between pressure head, relative permeability and saturation to solve the governing variably saturated flow equation expressed as (Huyakorn et al., 1986):

$$\frac{\partial}{\partial x}\left(K_{xx}k_{rw}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}k_{rw}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}k_{rw}\frac{\partial h}{\partial z}\right)-W=\Phi\frac{\partial S_w}{\partial t}+S_wS_s\frac{\partial h}{\partial t} \qquad \text{(Eq. 3-3)}$$

where K_{xx}, K_{yy}, and K_{zz} are principal components of saturated hydraulic conductivity (L/T) along the x, y, and z axes (L); k_{rw} is the relative permeability, a dimensionless value that ranges from 0 to 1 depending on the water content; h is the hydraulic head (L); W is a volumetric flux per unit volume representing sources and sinks (T^{-1}); Φ is the porosity; S_w is the degree of saturation of water, S_s is the specific storage of the porous material (L^{-1}); and t is time (T).

It is also necessary to specify the relationship between relative permeability, the water phase saturation, and the pressure head in order to solve for saturation and relative permeability in the governing equation (Eq. 3-3). MODFLOW-SURFACT simplifies these relationships using the following functions (van Genuchten, 1977, 1980):

$$k_{rw}=S_e^{1/2}\left[1-\left(1-S_e^{1/\gamma}\right)^{\gamma}\right]^2 \qquad \text{(Eq. 3-4)}$$

And

$$S_e=\frac{S_w-S_{wr}}{1-S_{wr}}=\begin{cases}\frac{1}{\left[1+(a\psi)^{\beta}\right]^{\gamma}} & for\ \psi<0\\ 1 & for\ \psi\geq 0\end{cases} \qquad \text{(Eq. 3-5)}$$

where S_e is the effective water saturation, α (L^{-1}), γ (-) and β (-) are empirical parameters, ψ is the pressure head (L), and S_{wr} is the residual water phase saturation. The parameters β and γ are related by $\gamma=1-1/\beta$. Together, the above equations allow for hydraulic head to be solved at each node in the model domain.

3.3.2 Model Domain and Grid

The modeled area is centered across the stream, alluvial, and bedrock aquifer sequence represented in the Cherry Creek cross section presented in Chapter 2 (Figure 2-3). The cross-sectional model domain spans 8,000m (west to east), and 200m vertically below the streambed. The model grid is partitioned into 20 layers, 85 columns and 1 row for a total of 1,700 active model cells. Grid cell spacing along the x-axis ranges from 25 to 120m with the closest spacing centered on the stream. All model layers are evenly spaced at 5m (Figure 3-6).

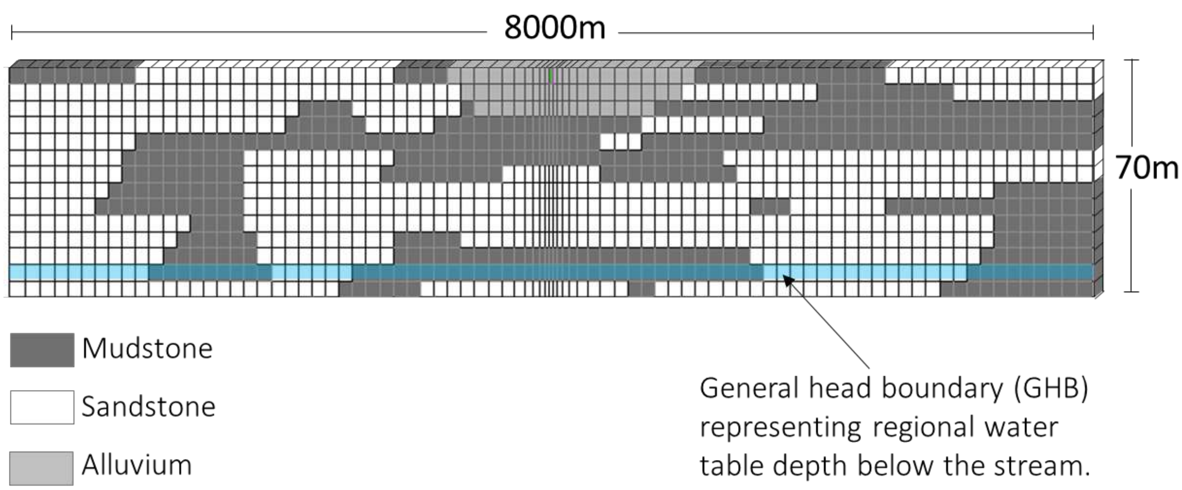

Figure 3-6 Model domain and boundary conditions showing lithologic heterogeneity assigned from binary image and manually adjusted connectivity.

3.3.3 Boundary Conditions

The lateral edges of the model (x=0 and x=8000m) were treated as no-flow boundaries given the expectation of regional flow normal to the modeled section, without significant cross gradients that would produce horizontal inflow or outflow in those distal areas. The model was constructed to analyze flow directions in the channel and stream alluvium; the lateral boundary conditions were placed at a sufficient distance such that they did not influence the simulated gradients near or underneath the alluvium.

The stream was modeled as a perennial stream using a head-dependent flux boundary condition as implemented in MODFLOW's River Package (Harbaugh et al., 2000). The simulated flux into or out of the "river" cell is proportional to a stream width, assigned stream stage and streambed hydraulic conductivity, and the head difference between the stream and adjacent alluvial aquifer. The streambed hydraulic conductivity and width were assigned from field measurements as described in Section 3.1. Total fluxes across the stream-alluvium interface were checked against field calculations.

The lower boundary condition followed the bedrock aquifer regional water table, and was modeled using MODFLOW's General Head Boundary (GHB) Package (Harbaugh et al., 2000). The GHB package allows model cells to be assigned a head-dependent boundary condition, causing the flux entering the model to vary proportional to a head difference between an external source and the model boundary cell. In this case, the external-source head was the elevation of the regional water table. To evaluate the cumulative effects of many pumping wells beyond the extent of the model domain, regional hydraulic head declines were simulated by lowering the water table (GHB external heads) for successive simulations representing distinct time periods.

3.3.4 Model Parameters

Hydrogeologic input parameters for each model cell were assigned based on lithology type. Bedrock lithology was assigned directly from the cross section (Figure 2-3) by converting the cross section image to an integer text file and performing a nearest neighbor interpolation function on the lithology and grid matrices in Matlab (R2017b). Lithology was assigned as either Sandstone or Mudstone, and the alluvium was distinguished manually. The alluvial aquifer thickness was manually altered to account for the valley topography which initially rendered the alluvial thickness in excess. Model parameter inputs and sources are summarized in Table 3-1.

The water retention parameters α, γ, β, and residual water content (θ_r), were assigned from published values. Note that θ_r values are input to the model for the calculation of effective saturation, S_e, using the following equation (HydroGeologic Inc., 2002):

$$S_e = \frac{\theta - \theta_r}{\theta_s - \theta_r} \qquad \text{(Eq. 3-6)}$$

where θ is the volumetric water content, θ_r is the residual water content, θ_s is the saturated water content assumed equal to porosity (Φ).

Hygiene sandstone parameters estimated by van Genuchten (1980) were assigned to modeled sandstone units. Sand and clay values estimated by Carsel and Parrish (1988) were assigned to the alluvium and mudstone, respectively. Saturated water content, θ_s was assigned from Paschke et al. (2011). Considerable effort was devoted to estimating the Dawson Aquifer sandstone parameters using water retention data from Dawson core samples reported by Lapey (2001). However, the water retention data coverage was deemed insufficient to fully characterize the water retention curve, and resulting parameters were not comparable to published values. Attempts and results are summarized in Appendix C.

3.3.5 Steady-State Modeling Scenarios

Groundwater flow was simulated for different regional water-table positions using steady-state solutions. Without a time component, the right side of Equation 3-3 cancels out and the variably saturated flow equation becomes:

$$\frac{\partial}{\partial x}\left(K_{xx}k_{rw}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}k_{rw}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}k_{rw}\frac{\partial h}{\partial z}\right)-W=0 \qquad \text{(Eq. 3-7)}$$

Steady-state simulations were run to produce a matrix of saturation states and hydraulic heads for 13 scenarios that imposed different regional water-table positions. Additionally, mass balance results were analyzed to determine the simulated flow rates between the alluvium and bedrock. Given that the modeling domain is a 2D vertical slice, these inter-aquifer flows are reported as volumetric rates per unit length ($L^3T^{-1}L^{-1}$).

Table 3-1 Numerical model input parameters and sources.

Parameter (unit)	Unit	Alluvium	Dawson Aquifer	
			Sandstone	Mudstone
Saturated Hydraulic Conductivity				
K_x	m/d	130 [1]	0.381 [2]	0.00381 [2]
K_y	m/d	130 [1]	0.381 [2]	0.00381 [2]
K_z	m/d	13 [1]	0.381 [3]	0.00381 [3]
van Genuchten Parameters				
θ_s	-	0.4 [2]	0.4 [2]	0.4 [2]
θ_r	-	0.045 [4]	0.15 [5]	0.068 [4]
α	m^{-1}	14.5 [4]	0.79 [5]	0.8 [4]
γ	-	2.68 [4]	10.4 [5]	1.2 [4]
β	-	0.627 [4]	0.903 [5]	0.083 [4]
Stream Parameters				
Streambed Conductivity	m/d	25.2 [6]		
Channel Width	m	3.5 [7]		

1. SPDSS Task 43 Phase 3
2. Paschke et al., 2011
3. Barkmann, 2004
4. Carsel and Parrish, 1988 (Alluvium as Sand & Dawson Mudstone as Clay)
5. vanGenuchten, 1981 (Dawson Sandstone as Hygene Sandstone)
6. Ronayne et al., 2017
7. Google Earth satellite imagery

CHAPTER 4 – RESULTS

4.1 Groundwater-Surface Water Interaction

Pressure data analyzed from piezometer nests revealed downward vertical gradients between 0.1 and 0.4 (m/m) at the Upstream, Middle Upper and Middle Lower piezometer nests throughout the measurement period of November, 2016 to January, 2018. Substantially lower gradients, between 0.1 and -0.1 (m/m) were observed at the Downstream station. The Middle Lower Piezometer nest (Figure 4-1) was flooded during June of 2017 by a beaver dam or wood jam roughly 20m downstream from the station location. Hydraulic gradient data from both the Middle Lower and Downstream stations indicate a potential response to the ponding of water. Data collection was suspended at the Middle Lower station following submergence.

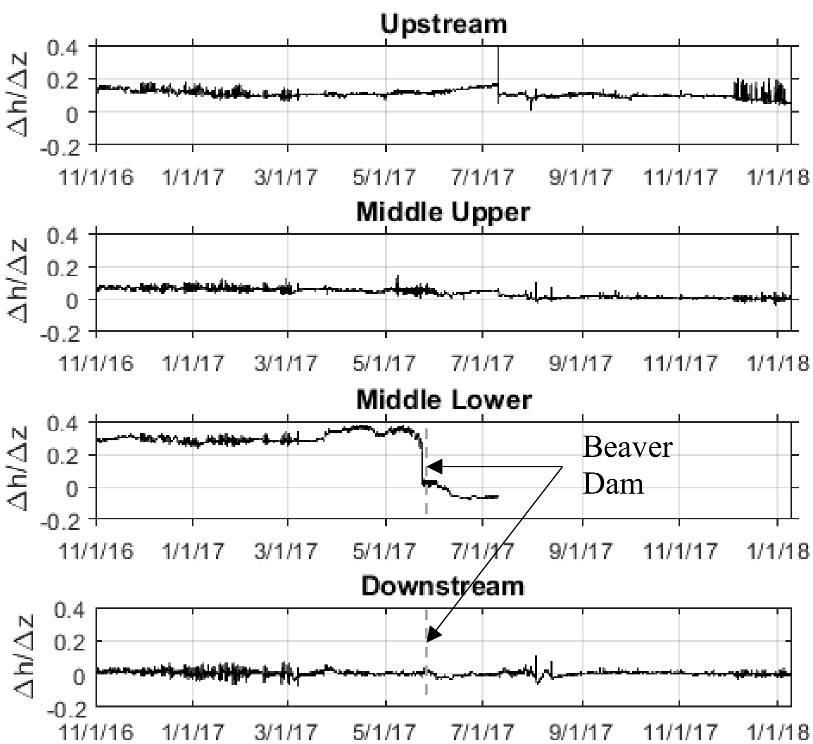

Figure 4-1 Vertical Hydraulic Gradients along East Plum Creek. Positive values indicate downward gradients.

Previously measured vertical streambed hydraulic conductivity (K_{sb}) values ranged from 11 to 43 m/d along the East Plum Creek Study Reach (Ronayne et al., 2017; Table 4-1). The average K_{sb} of 25 m/d was used as the streambed hydraulic conductivity input for the 2D groundwater model. Calculated specific discharge (q) values ranged from 0.07 to 3 m/d at point locations during the study period. The average specific discharge during the study period for all locations was 1.9 m/d. A seepage rate was also calculated by multiplying the specific discharge by the stream width to estimate the volumetric rate of seepage per length of stream (Table 4-1).

Table 4-1 Summary of streambed properties, piezometer data, and estimated seepage rates for measurement stations on East Plum Creek.

Stream Segment	Average Streambed Hydraulic Conductivity (m/d)	Average Vertical Gradient dh/dz	Specific Discharge (m/d)	Average Stream Width (m)	Seepage rate per length of stream ($m^3d^{-1}m^{-1}$)
Upstream	26	0.12	3.0	2.7	8.3
Middle Upper	22	0.04	0.86	3.0	2.6
Middle Lower	11	0.36	3.7	2.1	7.7
Downstream	43	0.002	0.068	2.5	0.17

4.2 Spatiotemporal Analysis of Bedrock Aquifer Water Levels

Significant changes occurred from 1970 to 2010 in the relationship between the alluvial and bedrock aquifer (Figure 4-2). During the 1970's, the water table in the unconfined bedrock aquifer was well above the base of the alluvial aquifer (0-50m) in the vicinities of East Plum and Cherry Creeks. This implies the potential for flow from the bedrock to the alluvial aquifer and surface streams. As early as the 1980's, those same alluvial traces exhibited bedrock water levels at or well below (100m) the base of the alluvial aquifer. This trend continued and increased through the 1990's, 2000's, and 2010's as the bedrock aquifer water levels declined further in relation to the alluvial aquifer. During the 2010's, the bedrock aquifer water levels were 100-160m below the base of the alluvial aquifer in places. Trends were documented both vertically and laterally as depicted in Figure 4-2. Areas that started off with bedrock aquifer water tables above the alluvial aquifer transitioned to having bedrock water tables increasingly below the alluvial aquifer. This transition describes vertical gradients changing from upward to increasingly downward gradients and indicates a vertical trend. Areas that transitioned to downward gradients also expanded through time to cover greater areas of the alluvium. This indicates a lateral trend in the bedrock and alluvial aquifer relationship.

4.3 Groundwater Model

The groundwater model was developed to evaluate the effects of declining bedrock water levels on hydraulic heads, fluxes, and saturation states in a stream-alluvial-bedrock aquifer sequence. Results from the groundwater model are in the form of pressure heads, effective saturation values, and simulated mass balance for 13 steady state model runs.

4.3.1 Pressure Heads

Model simulations illustrate changes in hydraulic heads and flow directions that occur in response to declining water levels within the bedrock aquifer. As described in Section 3.3.3, each successive model simulation incorporates a lower bedrock aquifer water table through the external boundary head specification in the MODFLOW GHB Package. In this chapter, the "regional water table position" refers to the modeled elevation of the bedrock aquifer water table at relatively large lateral distances (> 1 km) away from the alluvial aquifer.

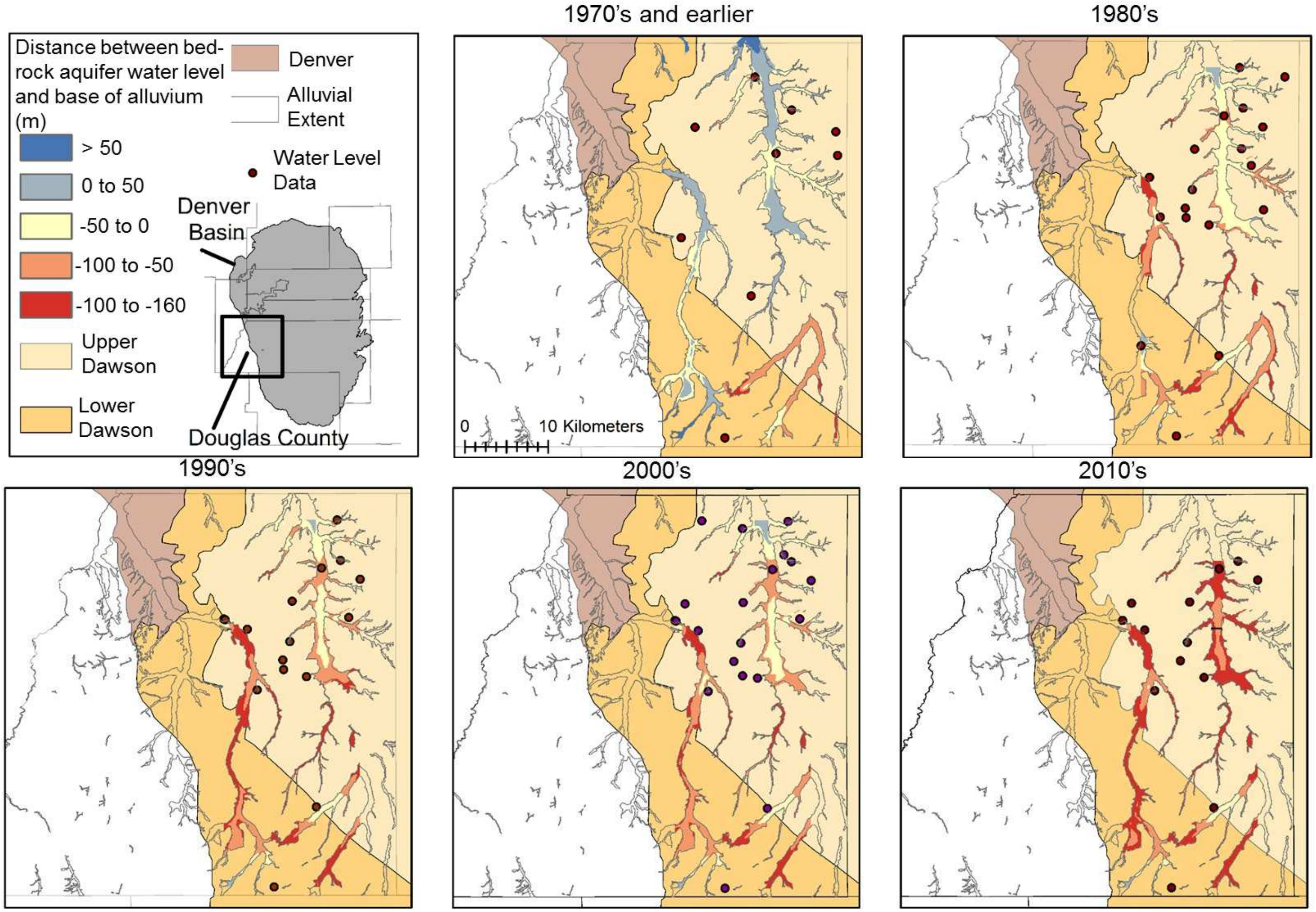

Figure 4-2 Spatiotemporal analysis of bedrock aquifer water levels in relation to base of stream alluvium. Positive values (blue colors) indicate that the water table elevation within the bedrock aquifer is above the base of alluvium. Negative values (yellow/orange/red) indicate that the bedrock aquifer water table is below the base of alluvium, highlighting the potential for unsaturated conditions within the bedrock.

Pressure head results from the steady state groundwater model show the development of elevated pressure heads and mounding below the recharge sources (stream and alluvial aquifer) when the water table position lowered from the stream elevation to 5m below the alluvial aquifer base (Figure 4-3). When the water table position reached 20m below the alluvial aquifer base, negative pressure heads developed between the alluvial and bedrock aquifers. When the regional water table position reached 35m below the alluvial aquifer, a consistent zone of negative pressures was present between the alluvial and bedrock aquifers. These results indicate increasing downward vertical gradients below the alluvial aquifer until the point where negative pressure heads develop between the two saturated zones. They also demonstrate that perched aquifer conditions may be induced as bedrock water levels drop below the alluvial aquifer. Pressure head results are plotted in Figure 4-3.

4.3.2 Saturation States

Saturation results from the steady state groundwater model (Figure 4-4) closely mimic the pressure head results. However, these results also demonstrate how lithologic heterogeneity and pressure head control the degree of water saturation and, in turn, fluxes through the system. When the bedrock water table position is lowered relative to the alluvium, mudstones tend to retain their water while sandstones desaturate. This is largely due to a difference in water retention properties between mudstone and sandstone as depicted in Figure 4-5. Sandstones have relatively “steep” water retention curves which leads to greater loss in saturation over a smaller range of changes in hydraulic heads. Conversely, mudstones have “gradual” water retention curves and tend to hold water across large changes in pressure head.

Effective saturation results also show that the alluvium remains mostly saturated despite its steep water retention curve predicting desaturation. This is due to the relatively low hydraulic conductivity of the bedrock aquifer at the base of the higher hydraulic conductivity alluvial aquifer. In this way, the hydraulic conductivity contrast between the alluvial and bedrock aquifers acts as a rate limiting factor for bedrock recharge.

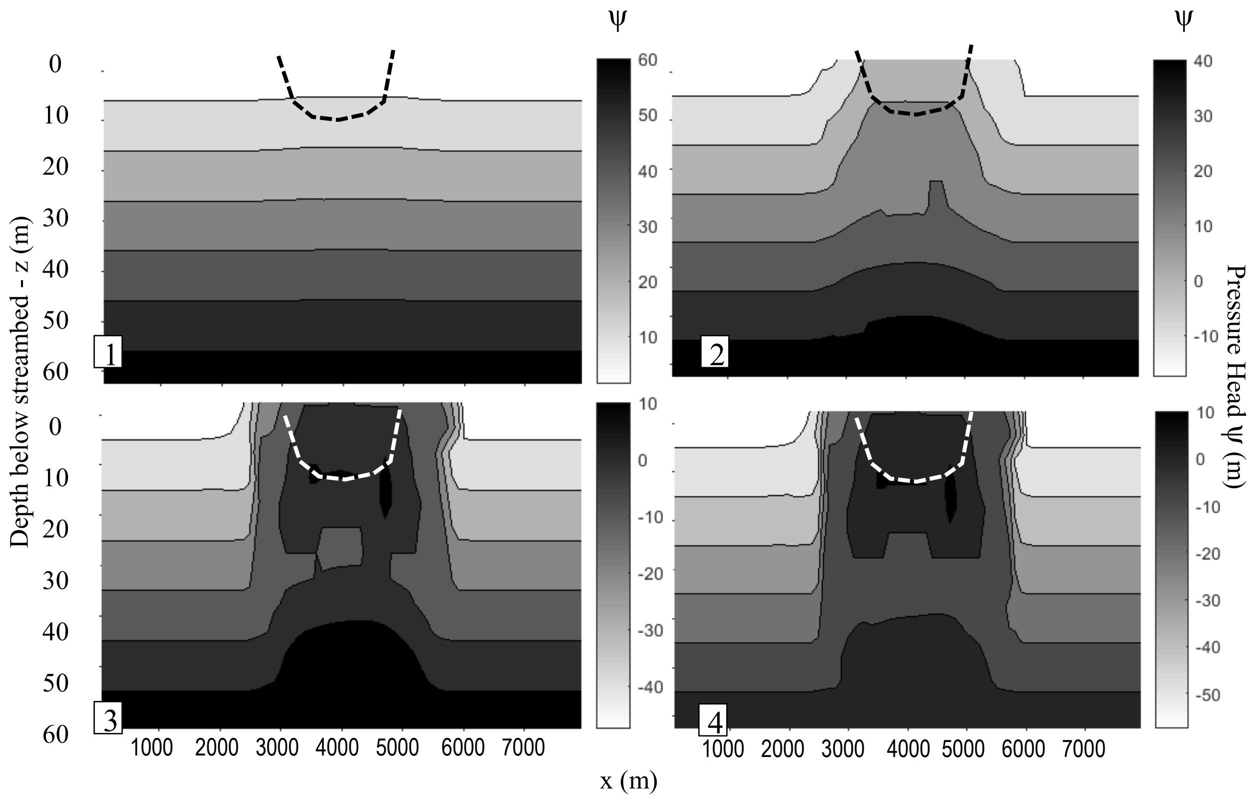

Figure 4-3 Pressure head results from the steady state model plotted for 4 regional water table positions. Alluvium-bedrock interface depicted as dashed line. The stream is located near center of alluvium at surface (z=0). (1) Regional water table at stream elevation (0m). (2) Regional water table position 5m below base of alluvial aquifer (20m). (3) Regional water table position 35m below base of alluvial aquifer (50m). (4) Regional water table position 45m below base of alluvial aquifer (60m).

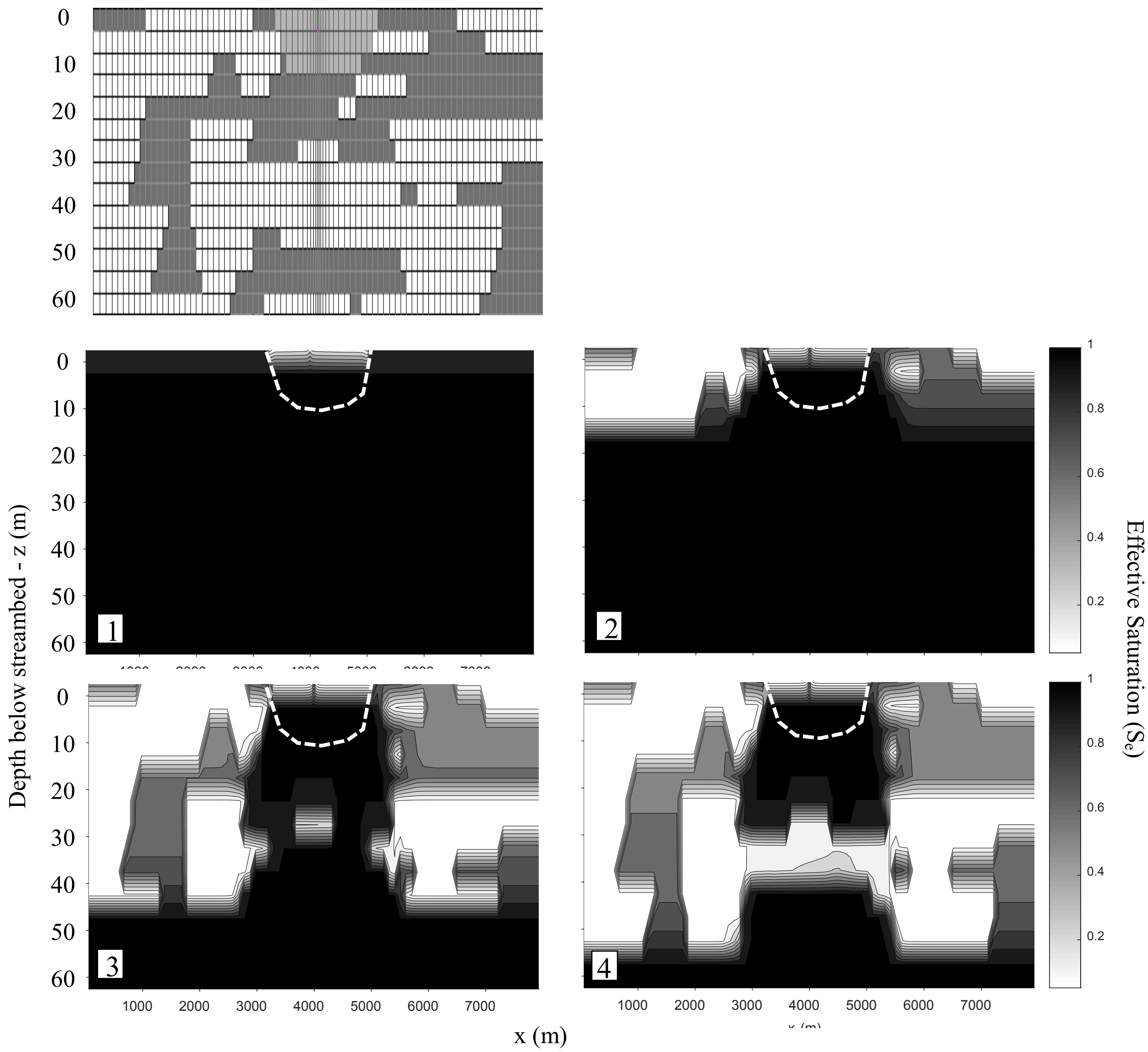

Figure 4-4 Model grid for reference and effective saturation results from steady state model plotted for 4 regional water table positions. Alluvium-bedrock aquifer interface is depicted as a dashed line. The stream is centered in the alluvium at the surface (z=0). (1) Regional water table at stream elevation (0m). (2) Regional water table position 5m below base of alluvial aquifer (20m). (3) Regional water table position 35m below base of alluvial aquifer (50m) with development of complex saturation patterns. (4) Regional water table position 45m below base of alluvial aquifer (60m).

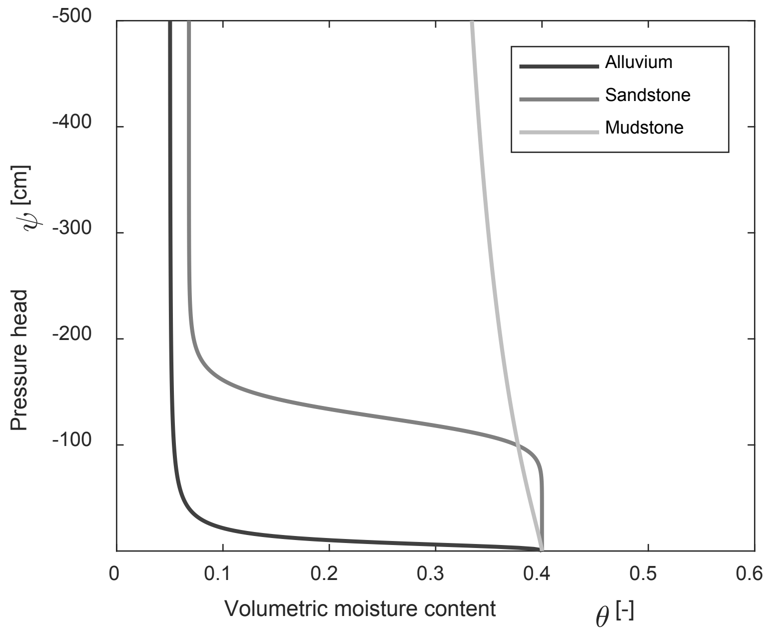

Figure 4-5 Water retention curves for modeled lithologies.

4.3.3 Simulated mass balance and estimation of recharge

The simulated mass balance results were analyzed to estimate recharge to the bedrock aquifer. Recharge rates were obtained for each steady-state model simulation by calculating the flow across the stream-alluvial and alluvial-bedrock interfaces. This was performed using a built-in Groundwater Vistas utility that calculates mass balance components (inflows and outflows) between user specified zones. In this case, two zones were assigned to represent the alluvial and bedrock aquifers. Recharge rates are reported as $m^3d^{-1}m^{-1}$ to represent a volumetric recharge flux per unit length in the regional flow direction. Since the model utilized steady state solutions, stream seepage was equal to flow across the alluvial-bedrock aquifer interface in all cases except the scenario where the regional water table position was above the alluvial aquifer. This exception is a result of the GHB model cells supplying water to the stream.

Simulated recharge rates ranged from 0 to 1.4 $m^3d^{-1}m^{-1}$ and increased linearly as the bedrock aquifer heads were increasingly lowered relative to the streambed and alluvial aquifer. The model-simulated recharge rates are within the range of values calculated from field data collected on East Plum Creek (Table 4-1). When the bedrock aquifer water table reached roughly 40-50m below the streambed, recharge rates started to level off, and the relationship between recharge and water table position started to flatten (Figure 4-6). As seen in the pressure head and saturation results, this depth corresponds to the initiation of desaturation of sandstone units in the bedrock aquifer. When bedrock water levels reached roughly 55m to 60m beneath the streambed ($\geq$ 40m below the base of alluvium), the recharge rate stabilized, or no longer increased, at 1.4 $m^3d^{-1}m^{-1}$. This corresponds to pressure head and saturation results when a zone of negative pressures and desaturation fully separated the bedrock and alluvial saturated zones.

The deviation from the otherwise smooth trajectory when bedrock water levels were roughly 40m below the stream (Figure 4-6) is likely due to effects of heterogeneity on desaturation though it cannot be confirmed from this analysis. However, this highlights the need for more research on the effects of heterogeneity on recharge rates and saturation.

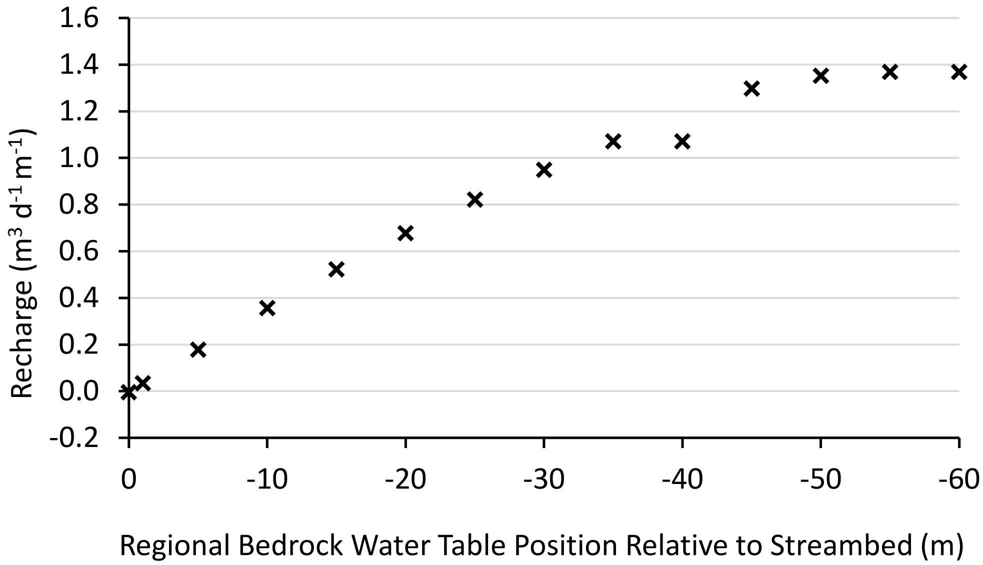

Figure 4-6 Modeled steady state flow rate across the alluvial-bedrock interface for various regional bedrock water table positions. Positive values indicate flow from the alluvium into the bedrock aquifer (i.e., Recharge).

CHAPTER 5 – DISCUSSION

This study characterized the effects of long term pumping on inter-aquifer flow directions and recharge processes in an alluvial-bedrock aquifer system. It identified stream seepage as an active source of aquifer recharge and described how seepage rates, recharge, and inter-aquifer fluxes change through time as a bedrock aquifer water table is lowered from long term pumping. Results lend significant insight into water budgets and flow dynamics in multi-aquifer systems including mechanisms for induced recharge between aquifers, conditions for which seepage and inter-aquifer fluxes may stabilize, and new concepts for saturation affected fluxes with long term pumping. Results from streambed nested piezometers, spatiotemporal analysis of bedrock aquifer water levels, and numerical modeling are now reviewed and synthesized into a conceptual model that characterizes the response to long term pumping in a stream-alluvial-bedrock aquifer sequence.

The finding of persistent downward vertical gradients beneath East Plum Creek streambed indicate year-round seepage to the underlying alluvial aquifer. This identifies a potentially significant source of recharge to alluvial and bedrock aquifers. Since vertical hydraulic gradients are the driving force for recharge to deeper aquifers, the study also evaluated how gradients are changing in response to regional, long term pumping stresses from deeper within the aquifer system. The spatiotemporal analysis of regional water levels revealed a deepening bedrock water table relative to the alluvial aquifer. This suggests the potential for increasing downward vertical hydraulic gradients and increasing seepage and recharge. Evolving alluvial-bedrock relationships also demonstrate a transition from upward to downward hydraulic gradients that occurred over decades. This highlights the delayed response of the aquifer system to regional, long term pumping and suggests that water budget components have changed and continue to change through time. The raster-based GIS method used in this study presents a new method to quantify the effects of long term pumping through time in other aquifer systems. By identifying the magnitude and extent of increased vertical gradients both laterally and vertically, the method captures regional pumping propagation and increased flux potential through time.

The 2D numerical modeling results show how a deepening regional bedrock water table can heavily influence stream seepage, fluxes, saturation, and hydraulic connection states in a multi-aquifer system. As the modeled water table position was lowered relative to the stream and alluvium, stream seepage and inter-aquifer fluxes increased linearly until a distinct point when desaturation was initiated. Initial desaturation occurred as individual sandstone bodies lost water following decreased fluxes through overlying mudstones. Following initial desaturation, a transition period was observed during which seepage fluxes and recharge rates were less affected by further lowering of the water table than if saturation had been maintained. This is indicated by the flattened response of recharge to further lowering of the water table position as depicted in Figure 5-1. As the water table position was increasingly lowered, the zone of desaturation increased until the alluvial and bedrock aquifers were fully separated by an unsaturated zone. At this point, inter-aquifer fluxes stabilized, with no further changes caused by additional lowering of the bedrock water table (Figure 5-1).

Modeling results also demonstrate that vertical hydraulic conductivity contrasts between the alluvial and bedrock aquifer, and between sandstone and mudstone units within the bedrock aquifer, act as rate limiting factors for aquifer recharge fluxes.

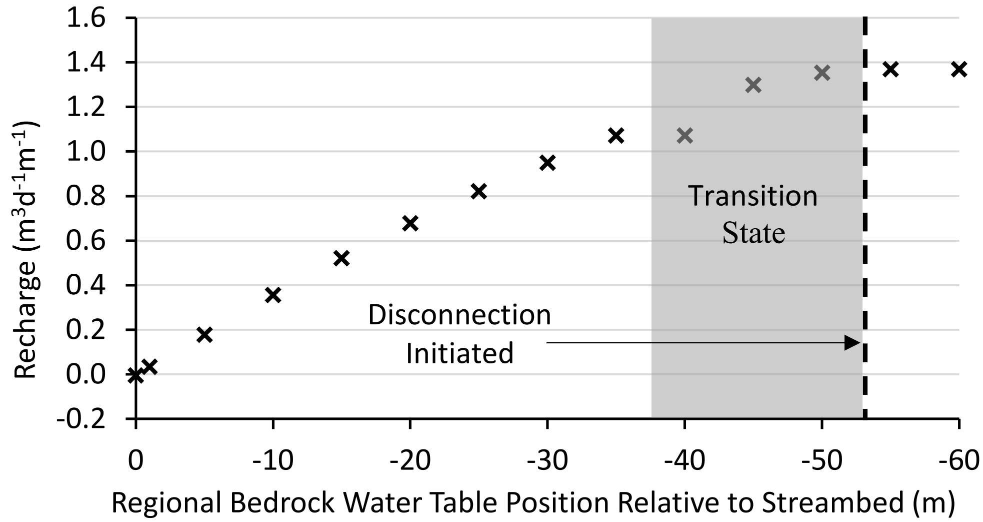

Figure 5-1 Model results delineating transition state where recharge fluxes are less affected by further lowering of the water table, and the initiation of disconnection where recharge fluxes are no longer affected by further lowering of the water table.

The independence of recharge fluxes to additional lowering of the regional water table position is analogous to the disconnected or perched regime that has been identified for stream-aquifer systems (Fox and Durnford, 2003; Schilling et al., 2017). Hydraulic disconnection is defined as the scenario in which the water table below a stream is sufficiently deep, whether naturally or pumping induced, such that further lowering of hydraulic heads does not alter the stream seepage rate. Disconnected regimes have significant implications for water budgets, groundwater-surface water interactions, and recharge (Brunner et al., 2009). As demonstrated in this study, hydraulic disconnection can also result between aquifers in a similar manner. Groundwater model results suggest reduced stream seepage and reduced inter aquifer recharge fluxes compared to a scenario where hydraulic connection is maintained. Note that hydraulic disconnection does not denote physical separation of water, but rather a change in the hydraulic relationship between the two water reservoirs. Recharge fluxes can still be increased through increased gradients elsewhere in the system but not by further lowering of the bedrock aquifer water table. Failure to identify disconnection could lead to inaccurate estimates of water budget components like pumping capture, groundwater recharge, and groundwater-surface water exchange.

The above observations may be synthesized into a new conceptualization for stream, alluvial, and bedrock aquifer systems affected by long term pumping. The conceptual model as shown in Figure 5-2 depicts four successive scenarios with regional bedrock water table positions at increasingly greater depths relative to the stream and alluvium. As the bedrock water table declines relative to the stream and alluvium, a reversal in flow direction followed by increasing downward fluxes across the alluvial-bedrock aquifer interface occurs. With increased lowering, a transition period initiates in which partial desaturation starts to limit fluxes across the alluvial bedrock aquifer interface. Finally, desaturation occurs enough that further lowering of the bedrock aquifer water table no longer increases fluxes across the alluvial-bedrock interface.

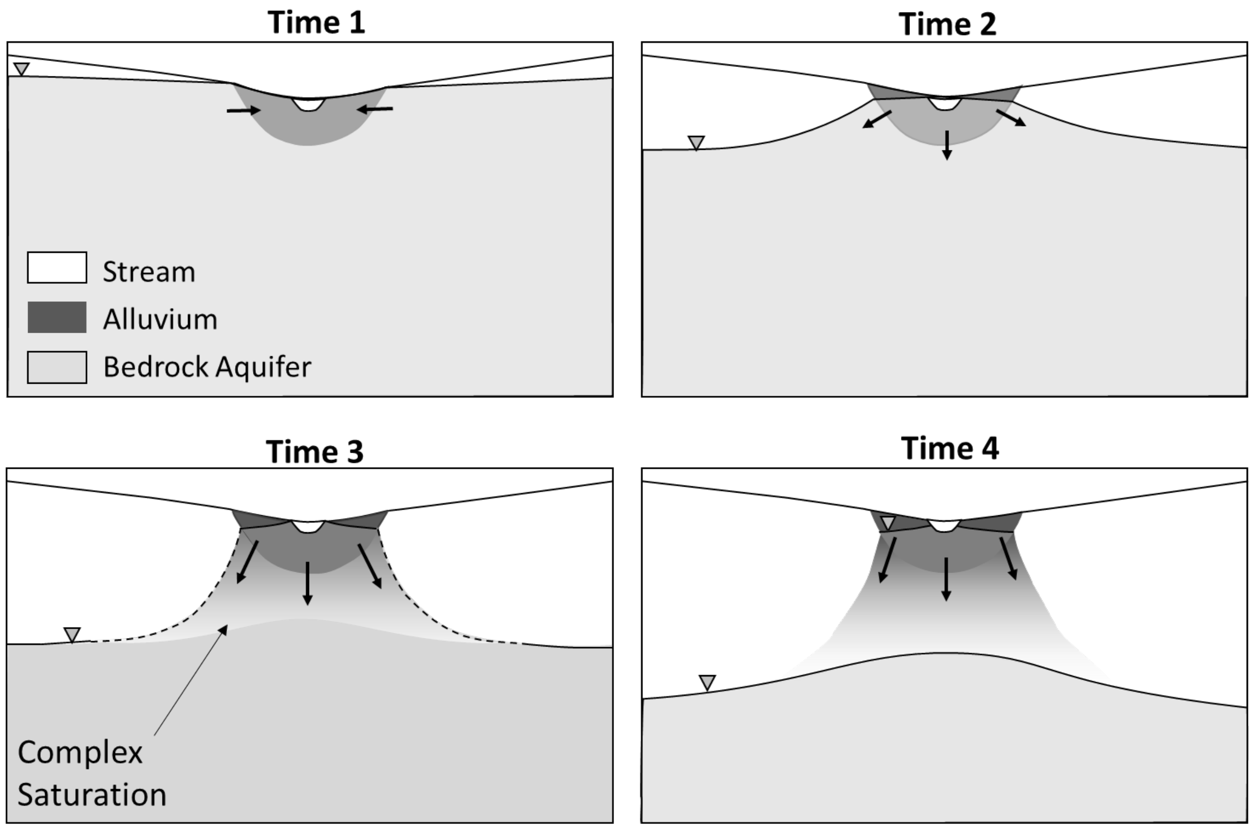

Figure 5-2 Conceptual model for a deepening bedrock water table relative to a stream and alluvium. Time 1 represents an early-time scenario prior to significant pumping; groundwater in bedrock aquifer discharges into alluvium and contributes to streamflow. During Time 2 (with pumping effects accumulated), flow direction has reversed and the alluvial aquifer discharges to the bedrock aquifer. Time 3 depicts the transition state in which partial desaturation of the bedrock aquifer reduces the effects of the lowered water table on alluvial-bedrock fluxes. Time 4 represents a disconnected regime in which further lowering of the bedrock water table position no longer increases flow across the alluvial-bedrock interface.

CHAPTER 6 – CONCLUSIONS AND FUTURE WORK

6.1 Conclusions

This thesis analyzed long-term water level changes in an area of the Denver Basin aquifer system with high historical groundwater withdrawals. It focused on specific processes, e.g. fluxes across stream-alluvial and alluvial-bedrock interfaces, to characterize the effects of long term pumping on important water budget components like recharge and groundwater-surface water exchange; components which are necessary for managing water resources and making predictions about future water supplies.

Stated objectives were met through multiple methods of investigation. The evolving hydrologic interactions between the alluvial and bedrock aquifer were evaluated through time using water level elevation data and raster subtraction methods. The heterogeneity and hydrogeologic architecture of alluvial and bedrock aquifer interfaces were characterized through the generation of two geologic cross sections representing stream, alluvial, bedrock sequences. This information was synthesized to evaluate surface water, alluvial and bedrock groundwater interactions in a representative study reach. A 2-dimensional groundwater model was developed to evaluate water budgets saturation states for a data-conditioned hydrogeologic setting using the cross sections and regional water level trends. Lastly, a conceptual hydrogeologic model was produced to describe the relationship between alluvial and bedrock aquifers with the potential for pumping-induced hydraulic disconnection of the aquifers.

The study found that pumping induced water level declines can significantly alter water budgets within multi-aquifer systems such as the Denver Basin alluvial-bedrock aquifer system. After identifying stream seepage as a source of alluvial and bedrock aquifer recharge, it was demonstrated how long term pumping stresses from deeper within the aquifer can affect seepage recharge and fluxes between aquifers. A method was developed for evaluating spatiotemporal changes in alluvial-bedrock hydrologic interaction, and potential changes in recharge, using commonly available geologic and water level data; the method is transferrable to other systems which have undergone long term pumping. Finally, numerical modeling performed as part of this study demonstrated the importance of considering the effects of desaturation when calculating water budgets in aquifers that have undergone long term pumping. The modeling demonstrated how water budget components including seepage recharge and inter-aquifer fluxes are sensitive to the bedrock water table position up until the point when desaturation initiates a disconnected regime. Additionally, modeling results demonstrate that hydraulic conductivity contrasts within and between aquifers can be rate limiting factors for recharge in multi-aquifer systems.

6.2 Recommendations for Future Work

While stream seepage and inter-aquifer flow were demonstrated as a source of aquifer recharge to alluvial and bedrock aquifers, surface seepage fluxes were estimated in the field for a limited study reach and depth. Further, the variability in vertical hydraulic gradients along East Plum Creek demonstrate that there are many factors, including stream morphology and surface features, that influence fluxes through the streambed. Characterizing fluxes at greater spatial and vertical extent within the alluvial aquifer and even across the alluvial-bedrock aquifer interface would better constrain recharge fluxes and lend significant insight into recharge processes and the effects of heterogeneity to the bedrock aquifer. Additionally, spatial analysis of diurnal temperature signals could be used to investigate the effects of hyporheic exchange on seepage fluxes at the streambed (e.g., Bhaskar et al., 2012).

While water level trends are consistent with other studies of the Denver Basin aquifers (Robson, 1984; Paschke et al., 2011), the magnitude of declines below the alluvial aquifer identified in this study (>50m) are larger compared to previous efforts. This warrants further investigation. One possible source of error involves the interpolation method used to generate the water table elevation between data points and its ability to capture the effects of surface topography on local water tables. However, the best way to minimize this error would be to have a greater resolution of water level data points. While the CDNR and USGS databases offer exceptional coverage of water level data compared to many other states and countries, the rigorous data sifting methods needed to capture water levels unaffected by pumping at specific depths drastically reduced the available data.

Additional field verification and modeling are needed to characterize the effects of heterogeneity and water retention parameters on desaturation in multi-aquifer systems. This study proposes that desaturation is controlled by heterogeneity of alternating mudstones and sandstones, and that the presence of low-conductivity units reduces vertical fluxes enough to desaturate sandstones. During model construction, there were obvious sensitivities to model inputs including lithologic heterogeneity and hydraulic properties. However, a rigorous evaluation of sensitivities to different lithologic heterogeneities and water retention parameters would better constrain these initial observations. A novel result of this study is the documentation of pumping-induced disconnection between aquifers. However, 3D modeling and collection of field data validating this finding are necessary to gain a better understanding of disconnection dynamics.

REFERENCES

Alzraiee AH, Bailey R, Bau D (2017) Assimilation of Historical Head Data to Estimate Spatial Distributions of Stream Bed and Aquifer Hydraulic Conductivity Fields. Hydrological Processes, 31(7):1527-1538

Banta ER (1989) Hydrologic effects of pumpage from the Denver Basin bedrock aquifers of northern El Paso County, Colorado. U.S. Geological Survey Water-Resources Investigations Report 88–4033

Barkmann PE, Deschesne M, Wickham ME, Carlson J, Formolo S (2011) Cross-sections of the fresh-water bearing strata of the Denver Basin between Greeley and Colorado Springs, Colorado. Colorado Geological Survey Open File Report 11-03

Barkmann PE, Fitzgerald FS, Sebol LA, Curtiss W, Pike J, Moore A, Taylor B (2015) Geology and groundwater resources of Douglas County. Colorado Geological Survey Open File Report 15-10

Beven KJ (1993) Prophecy, reality and uncertainty in distributed hydrological modelling. Adv Water Resourc 16:41–51

Beven KJ (2005) On the concept of model structural error. Water Sci Technol 52(6):167–175

Bhaskar AS, Harvey JW, Henry FJ (2012) Resolving hyporheic and groundwater components of streambed water flux using heat as a tracer. Water Resour Res 48(8)W08524,1-W08524,16

Bredehoeft JD (2003) From models to performance assessment: The conceptualization problem. Groundwater 41:571-577

Bredehoeft JD (2005) The conceptualization model problem-surprise. Hydrogeol J 13(1)37–4

Brunner P, Cook PG, Simmons CT (2009) Hydrogeologic controls on disconnection between surface water and groundwater. Water Resour Res 45:W01422

Carsel RF, Parrish RS (1988) Developing joint probability distributions of soil water retention characteristics. Water Resources Research 24(5):755-69

Colorado Department of Natural Resources (CDNR) (2018) DWR Water level database. https://dnrweb.state.co.us/cdss/GroundWater/WaterLevels/. Accessed 1 Mar 2018

Colorado Water Conservation Board (CWCB) (2006) SPDSS Phase 3 Tasks 44.3 South Platte alluvium region water level technical memorandum. ftp://dwrftp.state.co.us/cdss/-gwm/tm/ SPDSS44-3_20061129.pdf. Accessed 5 Dec 2017

Emmons SF, Cross W, Eldridge GE (1896) Geology of the Denver Basin in Colorado. U.S. Geological Survey Monograph 27

Everett RR (2014) Groundwater levels in the Denver Basin bedrock aquifers of Douglas County, Colorado, 2011–2013. U.S. Geological Survey Scientific Investigations Report 2014–5172

Farnham TM, Kraus MJ (2002) The stratigraphic and climatic significance of Paleogene alluvial paleosols in synorogenic strata of the Denver Basin. Rocky Mountain Geology 37:201-213

Fox GA, Durnford DS (2003) Unsaturated hyporheic zone flow in stream/aquifer conjunctive systems: Advances in Water Resources 26:989-1000

Harbaugh AW, Banta ER, Hill, MC, McDonald, MG (2000) The U.S. Geological Survey's Modular Ground Water Flow Model – User Guide to Modularization Concepts and the Ground Water Flow Process. U.S. Geological Survey Open-File Report 00-92

Huyakorn, PS, Springer EP, Guvanasen V, Wadsorth TD (1986) A three dimensional finite element model for simulating water flow in variably saturated porous media. Water Resour Res 22(12):1790-1808

HydroGeologic, Inc. (2002) MODFLOW-SURFACT: Version 2.1, A comprehensive MODFLOW based flow and transport simulator, User's Manual. Herndon.

Konikow LF, Kendy E (2005) Groundwater depletion: A global problem. Hydrogeol J 13(1):317-320

Lapey LA (2001) Hydrogeologic parameters of the Kiowa research core Kiowa, Colorado. Colorado State University Master's Thesis

Paschke SS (editor) (2011) Groundwater availability of the Denver Basin aquifer system, Colorado. U.S. Geological Survey Professional Paper 1770

Raynolds RG (2002) Upper Cretaceous and Tertiary stratigraphy of the Denver Basin, Colorado. Rocky Mountain Geology 37(2):111–134

Robson SG, Romero JC (1981) Geologic structure, hydrology, and water quality of the Dawson aquifer in the Denver basin, Colorado. U.S. Geological Survey Hydrologic Investigations Atlas HA-643

Robson SG (1984) Bedrock Aquifers in the Denver Basin, Colorado - A Quantitative Water Resources Appraisal. U.S. Geological Survey Open-File Report 84-43 1

Robson SG (1987) Bedrock aquifers in the Denver Basin, Colorado—A quantitative water-resources appraisal. U.S. Geological Survey Professional Paper 1257

Rojas R, Feyen L, Dassargues A (2008) Conceptual model uncertainty in groundwater modeling: Combining generalized likelihood uncertainty estimation and Bayesian model averaging. Water Resour Res 44 W12418

Ronayne M, Sale T, Cognac K, Brown G (2017) 2016-2017 Studies Supporting Sustainable Use of the Denver Basin Aquifers in the Vicinity of Castle Rock. CSU-Castle Rock Annual Report

Rumbaugh J, Rumbaugh D (2011) Groundwater Vistas 7.0, 1996-2011. Environmental Simulation Inc.

Schilling OS, Irvine DJ, Hendricks Franssen HJ, Brunner P (2017) Estimating the spatial extent of unsaturated zones in heterogeneous river-aquifer systems. Water Resour Res 53:10,583-10,602

Troldborg L, Refsgaard JC, Jensen KH, Engesgaard P (2007) The importance of alternative conceptual models for simulation of concentrations in a multiaquifer system. Hydrogeol J 15:843–60

USGS (2015) U.S. Geological Survey (USGS) National Elevation Data Set. http://ned.usgs.gov. Accessed 5 Dec 2017

USGS (2018) National Water Information System (USGS Water Data for the Nation). http://waterdata.usgs.gov/nwis. Accessed 1 Mar 2018

van Genuchten MTh, Pinder GF, Saukin WP (1977) Modeling of leachate and soil interactions in an aquifer, Management of Gas and Leachate in Landfills. U.S. EPA Report EPA-600/9-77-026:95-103.

van Genuchten MTh (1980) A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci Soc Am J 44:892–898

Wada YL, van Beek PH, van Kempen CM, Reckman JWTM, Vasak S, Bierkens MFP (2010) Global depletion of groundwater resources. Geophys Res Lett 37 L20402

Watts KR (1995) Hydrogeology and simulation of flow between the alluvial and bedrock aquifers in the upper Black Squirrel Creek Basin, El Paso County, Colorado. U.S. Geological Survey Water-Resources Report 94–4238

APPENDIX A

BORING LOGS EXAMPLE AND DATA

I. Example boring log
II. Table A-1 Boring logs used in cross sections

I. Example Boring Log

FORM NO. GWS-31 01-93

WELL CONSTRUCTION AND TEST REPORT
STATE OF COLORADO, OFFICE OF THE STATE ENGINEER

For Office Use Only
RECEIVED
MAR 21 2005
GWS 32 94 04

1. WELL PERMIT NUMBER 23284FR

2. OWNER NAME(S) R&M DEVCO, LLC
Mailing Address 200 S WILCOX ST #402
City, St. Zip CASTLE ROCK, CO 80104
Phone 814 8251

3. WELL LOCATION AS DRILLED: SW 1/4 NW 1/4, Sec. 34 Twp. 8S , Range 67W
DISTANCES FROM SECTION LINES:
2400 ft. from NORTH (north or south) Sec. line and 660 ft. from WEST (east or west) Sec. line
SUBDIVISION BELL MTN RANCH LOT ___ BLOCK ___ FILING(UNIT) ___
STREET ADDRESS AT WELL LOCATION:

4. GROUND SURFACE ELEVATION DRILLING METHOD COMBINATION
DATE COMPLETED 2/4/2005 TOTAL DEPTH 500 ft. DEPTH COMPLETED 500

5. GEOLOGIC LOG:

Depth		Description of Material (Type, Size, Color, Water Location)
0	5	brown loam & gravel
5	83	gravel & rocks
83	100	gray shale
100	115	sand
115	144	green shale
144	163	sand
163	187	gray shale
187	221	sand
221	233	brown shale
233	263	sand
263	300	blue shale
300	306	sand
306	431	gray shale
431	495	sand
495	500	gray shale
500		bottom

REMARKS:

6. HOLE DIAM. (in.)

HOLE DIAM. (in.)	From (ft)	To (ft)
8 3/4	0	184
6 1/8	184	500

7. PLAIN CASING

OD (in)	Kind	Wall Size	From(ft)	To(ft)
6 5/8	STEEL	0.190	-1	184
4 1/2	PVC	0.240	20	280

PERF. CASING Screen Slot Size: 0.03

OD (in)	Kind	Wall Size	From(ft)	To(ft)
4 1/2	PVC	0.240	280	500

8. FILTER PACK:
Material SAND
Size 8-12
Interval 280'-500'

9. PACKER PLACEMENT
Type 2 4x6 rubber
Depth 184

10. GROUTING RECORD:

Material	Amount	Density	Interval	Placement
CEMENT	25 SACS	17 GALS	9'-184	POS DIS

11. DISINFECTION: Type HTH Amt. Used 75cups

12. WELL TEST DATA. ☐ Check box if Test Data is submitted on Form No. GWS 39 Supplemental Well Test.
TESTING METHOD BAILER
Static Level 215 ft. Date/Time Measured 2/4/2005 , Production Rate 12 gpm
Pumping Level 250 ft. Date/Time Measured 2/4/2005 , Test Length (hrs.) 4
Remarks AIR TEST 115 gpm @ 500'

13. I have read the statements made herein and know the contents thereof, and that they are true to my knowledge. [Pursuant to Section 24-4-104 (13)(a) C.R.S., the making of false statements herein constitutes perjury in the second degree and is punishable as a class 1 misdemeanor.]

CONTRACTOR Hier Drilling Co. Phone (303) 688-3012 Lic. No. 8
Mailing Address P.O. Box 250 Castle Rock, Colorado 80104

Name/Title (Please type or print)	Signature	Date
George R. Hier, President	George R. Hier	2/22/2005

II. **Table A-1** Boring logs used in cross sections

Cherry Creek (A-A')		Plum Creek (B-B')	
Permit #	**Depth (m bgs)**	**Permit #**	**Depth (m bgs)**
81103-F	263	60530-F	227
216959	238	5089-F	195
77950	124	53463-F	229
26263-F	244	105108-A	171
26262-F	230	106084	151
26261-F	244	211318-A	122
33056	17	23284-F-R	152
26260-F	186	10103-TH	75
6337-F	14	57962-F	332
271661	20	103039	79
13074-F	20	61548-F	307
271662	15	63111-F	255
51878-F	203		
254272	91		
197183	210		
81000-F	191		
135310	110		
222487	146		

APPENDIX B

SPATIOTEMPORAL ANALYSIS WATER LEVEL DATA

I. Table B-1 Data used in spatiotemporal analysis of bedrock aquifer water levels (Section 3.2 / Section 4.2)

Table B-1 Data used in spatiotemporal analysis of bedrock aquifer water levels

Permit # / USGS Site ID *	UTMX	UTMY	Decade	Source	Aquifer	Spring Average Water Level (m)
14435	522404.00	4371557.00	1960	CDNR	UPPER	1780.64
29756	514395.00	4332210.00	1960	CDNR	DAWSON	2205.23
393858104551201	506817.70	4388851.80	1970	NWIS	DAWSON	1783.10
392811104403801	527718.80	4368952.60	1960	NWIS	UPPER	1913.60
393145104472401	518002.40	4375521.20	1970	NWIS	UPPER	1760.54
392647104453901	520533.50	4366340.70	1970	NWIS	UPPER	1804.46
392639104403001	527920.10	4366117.10	1960	NWIS	UPPER	1930.37
392830104522602	510800.80	4369496.30	1960	NWIS	UPPER	1806.29
392120104533601	509143.90	4356238.50	1970	NWIS	UPPER	1895.38
391733104474601	517536.20	4349255.00	1970	NWIS	UPPER	2050.58
14435	522404.00	4371557.00	1980	CDNR	UPPER	1769.89
17931-F	523393.00	4364671.00	1980	CDNR	UPPER	1809.04
17929-F	516574.00	4366642.00	1980	CDNR	UPPER	1800.67
25043-F	515551.10	4358403.10	1980	CDNR	DAWSON	1811.58
26265-F	516301.70	4361768.60	1980	CDNR	UPPER	1812.42
26268-F	515448.70	4359558.60	1980	CDNR	UPPER	1828.03
26271-F	518308.00	4357548.00	1980	CDNR	UPPER	1825.38
29756	514395.00	4332210.00	1980	CDNR	DAWSON	2144.65
30813-F	520107.00	4370630.00	1980	CDNR	UPPER	1688.45
30812-F	520116.00	4370638.00	1980	CDNR	DAWSON	1779.47
31958	524773.00	4369288.00	1980	CDNR	UPPER	1877.23
55018-	519502.00	4341844.00	1980	CDNR	UPPER	2085.69
17752-F	511165.80	4363254.60	1980	CDNR	DAWSON	1584.66
392300104423901	524858.20	4359355.50	1980	NWIS	UPPER	1946.50
391412104525501	510142.20	4343045.90	1980	NWIS	DAWSON	2024.70
392658104442901	522205.60	4366684.50	1980	NWIS	UPPER	1793.44
393137104405001	527409.60	4375302.10	1980	NWIS	UPPER	1868.41
392233104511801	512443.00	4358493.40	1980	NWIS	UPPER	1836.26
393214104443701	521986.90	4376425.50	1980	NWIS	UPPER	1748.93
14435	522404.00	4371557.00	1990	CDNR	UPPER	1766.69
17931-F	523393.00	4364671.00	1990	CDNR	UPPER	1802.99
17929-F	516574.00	4366642.00	1990	CDNR	UPPER	1795.69
25043-F	515551.10	4358403.10	1990	CDNR	DAWSON	1820.73
26265-F	516301.70	4361768.60	1990	CDNR	UPPER	1804.56
26268-F	515448.70	4359558.60	1990	CDNR	UPPER	1817.11
26271-F	518308.00	4357548.00	1990	CDNR	UPPER	1808.93
26505-F	512369.80	4355952.00	1990	CDNR	UPPER	1816.75
29756	514395.00	4332210.00	1990	CDNR	DAWSON	2149.25

Permit # / USGS Site ID *	UTMX	UTMY	Decade	Source	Aquifer	Spring Average Water Level (m)
30813-F	520107.00	4370630.00	1990	CDNR	UPPER	1659.27
30812-F	520116.00	4370638.00	1990	CDNR	DAWSON	1774.62
31958	524773.00	4369288.00	1990	CDNR	UPPER	1875.65
38404-F	508355.10	4364480.20	1990	CDNR	DAWSON	1792.20
55018	519502.00	4341844.00	1990	CDNR	UPPER	2086.03
17752-F	511165.80	4363254.60	1990	CDNR	DAWSON	1589.55
393214104443701	521986.90	4376425.50	1990	NWIS	UPPER	1746.82
65988-F	508494.80	4364413.00	1990	NWIS	DAWSON	1789.68
14435	522404.00	4371557.00	2000	CDNR	UPPER	1766.38
17931-F	523393.00	4364671.00	2000	CDNR	UPPER	1808.62
17929-F	516574.00	4366642.00	2000	CDNR	UPPER	1781.65
26265-F	516301.70	4361768.60	2000	CDNR	UPPER	1776.09
26268-F	515448.70	4359558.60	2000	CDNR	UPPER	1791.09
26271-F	518308.00	4357548.00	2000	CDNR	UPPER	1784.48
26505-F	512369.80	4355952.00	2000	CDNR	UPPER	1848.02
29756	514395.00	4332210.00	2000	CDNR	DAWSON	2141.07
30813-F	520107.00	4370630.00	2000	CDNR	UPPER	1593.76
30812-F	520116.00	4370638.00	2000	CDNR	DAWSON	1770.78
31958	524773.00	4369288.00	2000	CDNR	UPPER	1875.22
35171-F	511600.30	4376494.80	2000	CDNR	UPPER	1770.99
38404-F	508355.10	4364480.20	2000	CDNR	DAWSON	1804.57
55018	519502.00	4341844.00	2000	CDNR	UPPER	2085.08
17752-F	511165.80	4363254.60	2000	CDNR	DAWSON	1592.35
393003104450001	521496.90	4372408.40	2000	NWIS	UPPER	1827.99
393214104443701	521986.90	4376425.50	2000	NWIS	UPPER	1746.35
392210104482901	516512.40	4357823.90	2000	NWIS	UPPER	2004.90
65988-F	508494.80	4364413.00	2000	CDNR	DAWSON	1807.28
238516	516496.00	4375561.00	2000	CDNR	DAWSON	1770.91
67803-F	509583.80	4366488.00	2000	CDNR	DAWSON	1793.14
392210104482901	516512.43	4357823.96	2000	NWIS	DAWSON	2003.84
393003104450001	521496.84	4372408.50	2000	NWIS	DAWSON	1827.07
393215104490001	515708.63	4376441.09	2000	NWIS	DAWSON	1761.52
393225104484901	515970.58	4376749.91	2000	NWIS	DAWSON	1760.90
393245104493401	514895.24	4377364.32	2000	NWIS	DAWSON	1762.60
393259104491001	515479.48	4377817.43	2000	NWIS	DAWSON	1756.50
14435	522404.00	4371557.00	2010	CDNR	UPPER	1760.62
17929-F	516574.00	4366642.00	2010	CDNR	UPPER	1764.07
26265-F	516301.70	4361768.60	2010	CDNR	UPPER	1735.57

Permit # / USGS Site ID *	UTMX	UTMY	Decade	Source	Aquifer	Spring Average Water Level (m)
26268-F	515448.70	4359558.60	2010	CDNR	UPPER	1752.75
26271-F	518308.00	4357548.00	2010	CDNR	UPPER	1760.30
26505-F	512369.80	4355952.00	2010	CDNR	UPPER	1847.41
29756	514395.00	4332210.00	2010	CDNR	DAWSON	2136.52
30813-F	520107.00	4370630.00	2010	CDNR	UPPER	1514.77
30812-F	520116.00	4370638.00	2010	CDNR	DAWSON	1711.33
31958	524773.00	4369288.00	2010	CDNR	UPPER	1875.90
55018	519502.00	4341844.00	2010	CDNR	UPPER	2084.90
17752-F	511165.80	4363254.60	2010	CDNR	DAWSON	1581.40
65988-F	508494.80	4364413.00	2010	CDNR	DAWSON	1807.55
238516	516496.00	4375561.00	2010	CDNR	DAWSON	879.80
67803-F	509583.80	4366488.00	2010	CDNR	DAWSON	1790.01
391229104421901	525451.33	4339703.17	2010	NWIS	DAWSON	2086.04
391654104464501	519009.83	4347889.42	2010	NWIS	DAWSON	2008.96
393215104490001	515708.63	4376441.09	2010	NWIS	DAWSON	1758.26
393225104484901	515970.58	4376749.91	2010	NWIS	DAWSON	1760.01
393225104490001	515732.24	4376750.66	2010	NWIS	DAWSON	1756.98
393244104484601	516093.81	4377349.48	2010	NWIS	DAWSON	1757.50
393245104493401	514895.24	4377364.32	2010	NWIS	DAWSON	1762.79
393253104484401	516119.46	4377639.32	2010	NWIS	DAWSON	1762.42
393256104484801	516035.73	4377731.63	2010	NWIS	DAWSON	1759.62
393259104491001	515479.48	4377817.43	2010	NWIS	DAWSON	1758.29
393301104484801	516021.09	4377885.74	2010	NWIS	DAWSON	1757.83
393303104484801	516020.98	4377941.23	2010	NWIS	DAWSON	1754.15

* NWIS sites are identified by USGS Side ID, CDNR sites are identified by permit number.

APPENDIX C

VAN GENUCHTEN PARAMETER ESTIMATION

This appendix documents an attempt to estimate van Genuchten (VG) parameters for Dawson Aquifer sandstones. The attempt was predicated on the availability of water retention data for Dawson Aquifer sandstones and results would have been assigned to sandstone features in the 2D groundwater model.

Following the methodology outlined in McWhorter (1990) and standard ASTM D 2325-68, Lapey (2001) reported volumetric water contents (θ) at various capillary pressures (0 through 13.5 bars) for seventeen Dawson Aquifer sandstone samples from the Kiowa Core. This data included estimates for saturated volumetric water content (θs). Together, these water retention data sets define a range of water characteristic curves for Dawson Aquifer sandstones that can be used to estimate and constrain VG parameters for Dawson Aquifer sandstones.

The empirical van Genuchten equation (1980) which describes the water characteristic curve is commonly expressed as:

$$S_e = \frac{1}{[1+(\alpha h)^n]^m} \qquad \text{Eq. C-1}$$

where α, n, and m are empirical constants (VG parameters) and Se is the effective saturation defined as:

$$S_e = \frac{\theta-\theta_r}{\theta_s-\theta_r} \qquad \text{Eq. C-2}$$

VG parameters were estimated by fitting the van Genuchten equation (1980) to the water retention data using a nonlinear least-squares optimization function (Matlab 2017b, lsqnonlin) which solved the best fit solution that minimized the sum of squares. VG parameter sets, including θ_r, α, n, and m were calculated for 15 of 17 available water retention data sets. The two samples for which VG parameters were not calculated had an insufficient range of reported volumetric water contents and capillary pressures to perform parameter estimation. Results in Table C-1 and generated water retention curves using estimated VG parameters are plotted in Figure C-1.

Resulting parameter values and generated water characteristic curves were compared to published parameters and curves for various lithologies from several sources. Resulting water retention data did not agree well with published data (Table C-1) and water characteristic curves generated using estimated VG parameters did not adhere to expected shape for a sandstone lithology (Figure C-2). Most noticeably, α values were several orders of magnitude greater than typical published values while n, m, and S_r values were within expected ranges. The error in these estimates was not evaluated for significance with statistics because the data were ultimately discarded.

While not confirmed, the error was assumed to result from incomplete characterization of the water retention curve with available data.

Table C-1 van Genuchten parameters; published and estimated values.

Soil Type	θ_s	α (1/m)	n	θ_r	m	Source
Published Values						
Crab Creek Sand	-	0.119	2.451	0.000	0.592	Stephens et al., 1987
Gravelly Sand G.E. 9	-	0.015	2.839	0.079	0.648	
Silt of Nave-Yaar	-	0.072	2.197	0.398	0.545	
Isa Silt Loam (0-15cm)	-	0.090	1.177	0.000	0.150	
Yolo light clay	-	0.027	1.600	0.180	0.375	
Sand	0.43	14.500	2.680	0.045	0.627	Carsel and Parrish, 1988
Silt	0.46	1.600	1.370	0.034	0.270	
Sandy Clay Loam	0.39	5.900	1.480	0.100	0.324	
Clay Loam	0.41	1.900	1.310	0.095	0.237	
Sandy Clay	0.38	2.700	1.230	0.100	0.187	
Clay	0.38	0.800	1.090	0.068	0.083	
Coarse Sandstone 1	0.34	1.003	2.798	0.017	0.083	Parajuli et al, 2017
Coarse Sandstone 2	0.35	0.032	2.115	0.012	0.083	
Fine Sandstone 1	0.04	0.085	1.219	0.000	0.083	
Fine Sandstone 2	0.03	0.012	1.361	0.000	0.083	
Millville Silt Loam	0.44	0.737	1.483	0.023	0.083	
Wedron Sand	0.35	2.262	7.980	0.040	0.083	
Hygiene Sandstone	0.25	0.790	10.400	0.153	0.904	van Genuchten, 1980
Parameter estimation results						
2Aa	0.35	7.84	1.19	0.230	0.163	Results
2Ab	0.37	24.9	1.17	0.230	0.147	
3Aa	0.25	0.255	1.61	0.171	0.377	
3Ab	0.29	251	1.06	0.008	0.056	
4Aa	0.29	985	1.13	0.029	0.111	
4Ab	0.29	173	1.12	0.008	0.106	
5Aa	0.42	10.2	1.48	0.112	0.323	
6Aa	0.31	152	1.09	0.008	0.084	
6Ab	0.31	359	1.09	0.008	0.079	
7Aa	0.6	1.97	1.74	0.252	0.424	
7Ab	0.53	8.16	1.34	0.258	0.252	
8Aa	0.49	8.19	1.35	0.159	0.261	
8Ab	0.26	2.91	1.45	0.171	0.308	
9Aa	0.38	667	1.18	0.176	0.153	
9Ab	0.28	271	1.21	0.156	0.177	

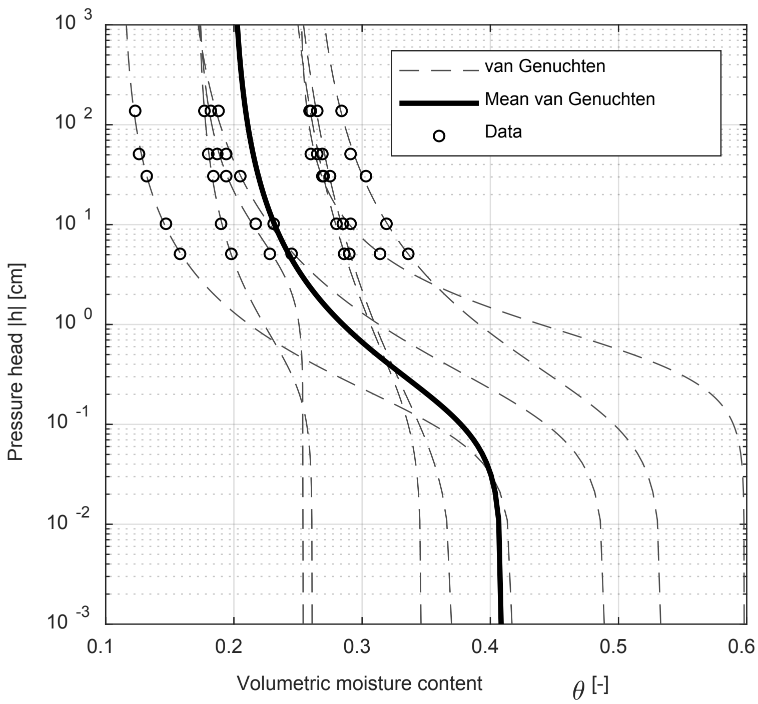

Figure C-1 van Genuchten data and water retention curves fit to VG parameters.

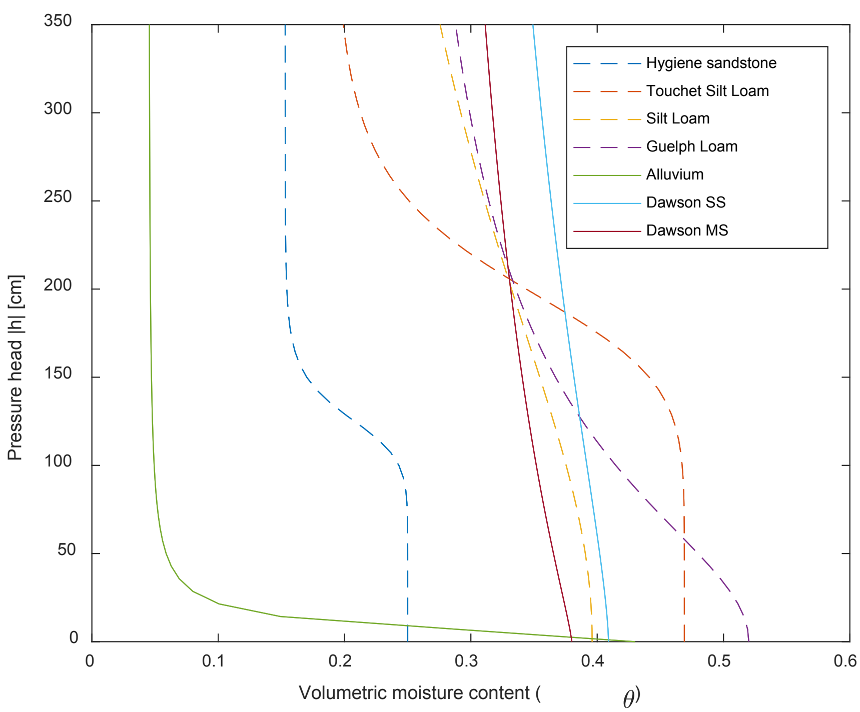

Figure C-2 Water characteristic curves generated using published VG parameters for various lithologies and estimated VG parameters for Dawson Sandstone.

Citations:

ASTM Standard References (1993) Standard Test Method for Capillary-Moisture Relationships for Coarse- and Medium-Textured Soils by Porous-Plate Apparatus. American Society for Testing and Materials D2325-68:184-90

McWhorter D, Garcia A (1990) The concept of specific yield and its evaluation by laboratory measurements. Colorado State University, Fort Collins 80523

van Genuchten MTh (1980) A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci Soc Am J 44:892–898

Lapey LA (2001) Hydrogeologic parameters of the Kiowa research core Kiowa, Colorado. Colorado State University Master's Thesis