EXPLORING GEOLOGIC CONTROLS ON INFILTRATION AND GROUNDWATER RECHARGE ON AN EPHEMERAL RIVER: A COUPLED GEOPHYSICS AND MODELING APPROACH

by

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ABSTRACT

Ephemeral streams are the main surface-water source in arid climates, and streambed leakage in these systems is an important component of groundwater recharge. Subsurface geology influences the extent and location of streambed leakage and therefore changes streamflow, impacts groundwater levels, and has the potential to influence confined aquifer recharge. This study looks at exploring geological controls on recharge from an ephemeral river through scenario evaluation with numerical models, constrained by geophysical observations. Drone magnetometer, electrical resistivity, hammer seismic, and hydrologic data were collected along the Alamosa River in southeastern Colorado, USA to constrain and parameterize a MODFLOW-SFR model based on the field system consisting of three layers: 1) an alluvial aquifer, 2) a confining unit, and 3) a confined aquifer. Four scenarios were explored beyond a base case to quantify controls on subsurface geology: 1) adding a fault; 2) changing the alluvial aquifer’s hydraulic conductivity; 3) changing the thickness of the streambed; and 4) removing the confining unit. A fifth scenario, adding a pumping well, was used to explore the role of this human influence on ephemeral river systems.

The drone magnetometer and resistivity data indicated that the middle portion of the study area was geologically distinct from the surrounding area, and the resistivity and seismic results indicate the presence of heterogeneity in the subsurface. Modeling scenarios that changed hydraulic conductivity values resulted in the most notable changes to the river’s hydrologic system, altering streamflow, leakage, and deep aquifer recharge. Streambed thickness proved to be an unimportant parameter. Results here suggest that the extent to which streambed leakage changes is proportional to the ratio of alluvial aquifer hydraulic conductivity ($K_1$) to streambed conductivity ($K_s$), and that this in turn controls the impacts on streamflow. This research suggests that subsurface heterogeneities are a fundamental control on ephemeral rivers’ hydrogeologic system and are key to their resilience under climate change.
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CHAPTER 1
INTRODUCTION

Between 1962 and 1990 the percentage of regions categorized as arid and semi-arid has increased by 7% and this trend is expected to continue (Huang et al., 2016). It has been well documented that climate change and increasing global mean temperature are known to impact the duration, frequency, intensity, and the form of precipitation (e.g., Barnett et al., 2008; Cayan et al., 2007; Stewart et al., 2005; Trenberth, 2011). Warmer winter temperatures result in decreased snow accumulation (e.g., Brown & Robinson, 2011) and increased precipitation as rain (e.g., Mote et al., 2005), which causes earlier snowmelt (e.g., Barnett et al., 2008) and a decrease in overall runoff (e.g., Kundzewicz & Döll, 2009). For montane environments in arid and semi-arid regions of the world, the melting of winter snowpack provides the primary water source (e.g., Serreze et al., 1999) and is vital for sustaining summer baseflow in rivers (e.g. Markovich et al., 2016; Tague et al., 2008). A decrease in snowpack is predicted to result in declining streamflow (e.g., Cayan et al., 2007) and the transition of rivers from perennial to intermittent or ephemeral (Gutierrez-Jurado et al., 2019).

The USEPA (2015) defines the following stream or reach types: 1) ephemeral, which flow for a short time—hours or days—following a precipitation event; 2) intermittent, which flow continuously at certain times of year, such as during spring snowmelt runoff; and 3) perennial, which flow year-round. In this thesis, all streams defined by the USEPA as either intermittent or ephemeral will be referred to as ephemeral streams. Ephemeral streams currently account for over 50% of rivers in the world (Gutierrez-Jurado et al., 2019) and for a much higher percentage in arid and semi-arid regions. In the American southwest, the percentage of ephemeral streams is much higher (i.e. Arizona (94%), Utah (79%), and New Mexico (88%)) (USGS, 2008). Although ephemeral streams are the dominant stream type in arid and semi-arid environments, they are understudied relative to perennial streams (Stromberg et al., 2017).

Ephemeral streams are important for a variety of reasons. In arid climates, ephemeral streams are typically the locations of the greatest biodiversity and density of vegetation (e.g. Kassas & Imam, 1954; Scott et al., 2014). Stromberg et al. (2017) found the volume of vegetation in the riparian zone of ephemeral streams to be significantly higher than in
surrounding area, describing the riparian zone as a hot spot for productivity. The riparian zone is home to most plants in an arid climate as it is the only location that provides a wet enough environment for plants to grow (Stromberg et al., 2017). The riparian zone and floodplain input organic matter and nutrients into the stream (Steward et al., 2012), while providing terrestrial organisms food, cover, nesting and breeding grounds, as well as a movement corridor (Goodrich et al., 2018; Steward et al., 2012). These flow events also benefit aquatic biota and aid in sustaining baseflows along the main stem. Ephemeral tributaries also mitigate the success of invasive species, as they have variable flows that invasive organisms are less equipped to manage (Goodrich et al., 2018).

Ephemeral streams are also an important location for recharge to aquifers (Coes & Pool, 2005; Goodrich et al., 2004; Shanafeld & Cook, 2014). Ephemeral streams are typically underlain by alluvial aquifers that are recharged as water seeps through the streambed and down through the aquifer material to the water table. If there is an intact confining unit below the alluvial aquifer, a deeper, confined aquifer will not be able to be recharged by the river. However, if there is a fracture or fault in the confining unit that provides a conduit for flow, a deep confined aquifer can also be recharged by a river (e.g. Bense & Van Balen, 2004). This thesis looks at how subsurface geology influences surface water, groundwater and the interaction between them. Streamflow measurements and geophysical data were collected along the Alamosa River in south-central Colorado to parameterize and inform the construction of a simplified ephemeral river model. The model was built using MODFLOW, the U.S. Geological Survey’s modular finite-difference flow model, with the streamflow routing package (SFR) and was used to test a variety of subsurface geologic scenarios and to explore the following research question: how does the geology beneath an ephemeral river system control streamflow, streambed leakage, groundwater levels, and recharge to a confined aquifer?
CHAPTER 2
EXPLORING GEOLOGIC CONTROLS ON INFILTRATION AND GROUNDWATER RECHARGE ON AN EPHEMERAL RIVER: A COUPLED GEOPHYSICS AND MODELING APPROACH

2.1 Abstract

Ephemeral streams are the main surface-water source in arid climates, and streambed leakage in these systems is an important component of groundwater recharge. Subsurface geology influences the extent and location of streambed leakage and therefore changes streamflow, impacts groundwater levels, and has the potential to influence confined aquifer recharge. This study looks at exploring geological controls on recharge from an ephemeral river through scenario evaluation with numerical models, constrained by geophysical observations. Drone magnetometer, electrical resistivity, hammer seismic, and hydrologic data were collected along the Alamosa River in southeastern Colorado, USA to constrain and parameterize a MODFLOW-SFR model based on the field system consisting of three layers: 1) an alluvial aquifer, 2) a confining unit, and 3) a confined aquifer. Four scenarios were explored beyond a base case to quantify controls on subsurface geology: 1) adding a fault; 2) changing the alluvial aquifer’s hydraulic conductivity; 3) changing the thickness of the streambed; and 4) removing the confining unit. A fifth scenario, adding a pumping well, was used to explore the role of this human influence on ephemeral river systems.

The drone magnetometer and resistivity data indicated that the middle portion of the study area was geologically distinct from the surrounding area, and the resistivity and seismic results indicate the presence of heterogeneity in the subsurface. Modeling scenarios that changed hydraulic conductivity values resulted in the most notable changes to the river’s hydrologic system, altering streamflow, leakage, and deep aquifer recharge. Streambed thickness proved to be an unimportant parameter. Results here suggest that the extent to which streambed leakage changes is proportional to the ratio of alluvial aquifer hydraulic conductivity ($K_1$) to streambed conductivity ($K_s$), and that this in turn controls the impacts on streamflow. This research suggests that subsurface heterogeneities are a fundamental control on ephemeral streams’ hydrogeologic system and are key to their resilience under climate change.
2.2 Introduction

Ephemeral streams are the main surface-water source in arid and semi-arid climates (Stromberg et al., 2017). Consequently, humans living in these climates tend to rely heavily on groundwater to supplement the limited surface water supply, which has resulted in groundwater overexploitation and aquifer depletion (Bajjali & Al-Hadidi, 2006; Kundzewicz & Döll, 2009; Llamas & Martínez-Santos, 2005). Simultaneously, water use in arid climates is increasing due to population growth and migration (e.g. Vörösmarty et al., 2000) and water use is shifting from agriculture to urbanization, which decreases irrigation return flows (e.g., Dinatale et al., 2008) and leads to a reduction in aquifer recharge (e.g. De Graaf et al., 2014). Water supplies are further impacted by climate change, which is increasing the number of in-stream sections that are dry for long periods in the year (Downing et al., 2012; Luce & Holden, 2009; Meehl et al., 2007; Steward et al., 2012). A large component of groundwater recharge in arid regions is streambed seepage (Coes & Pool, 2005; Goodrich et al., 2004; Shanafield & Cook, 2014). As more river reaches remain dry longer, alluvial aquifer recharge will be reduced (Meixner et al., 2015; Taylor et al., 2013; Kuszewicz & Döll, 2009).

Quantifying ephemeral river seepage is key to exploring groundwater recharge in arid regions – as groundwater recharge is difficult to measure – and to assessing the resilience of these streams under a changing climate and increased usage, where resilience is defined as the degree to which these river systems can withstand naturally occurring and human induced perturbations (e.g. Folke et al., 2005). Recharge tends to be highly variable in time (e.g. Pool, 2005) and space (e.g. Goodrich et al., 2018), as evapotranspiration and wetted channel evaporation contribute strongly to river water loss (e.g. Goodrich et al., 2004). Since it is difficult to quantify recharge directly, infiltration or transmission losses are often measured instead to approximate recharge. Infiltration is typically measured at point(s) at or just below the river surface; transmission losses are the quantification of the reduction in streamflow between two points and include streambed seepage, evaporation, and river bank losses (e.g. Shanafield & Cook, 2014). Field-based techniques that allow for the quantification of infiltration and transmission losses include isotopic, chemical, (Gooseff & McGlynn, 2005; Huang et al., 2016; Subyani, 2004) and heat tracers (e.g. Constantz et al., 2002; Shanafield & Cook, 2014), the chloride mass-balance technique (e.g. Scanlon et al., 2006; Taylor et al., 2013), piezometric data
(e.g. Besbes et al., 1978), controlled infiltration experiments, monitoring changes in water content (e.g. Stewart-deaker et al., 2000), a water balance approach (e.g. Sorman & Abdulrazzak, 1993), which often takes the form of estimating seepage loss or gains by taking the difference of flows at several locations along a river (e.g. Riggs, 1972), and flood wave tracking (e.g. Noorduijn et al., 2014; Shanafield et al., 2012). It is also possible to make direct estimates of recharge using groundwater mounding as measured in monitoring wells (e.g. Goodrich et al., 2004) and by groundwater dating (e.g. Shanafield & Cook, 2014). These techniques are costly, often necessitating the installation of several monitoring wells. Furthermore, these measurements often do not allow for an exploration of small-scale heterogeneity controls. Variations in subsurface geology surrounding surface water influences the extent and location of recharge (e.g. Goodrich et al., 2018). A river flowing over highly porous media, karst, or fractured bedrock can virtually disappear over a short distance (e.g. Dvory et al., 2018; Goodrich et al., 2018).

Geophysical methods have been used to differentiate subsurface lithologies both laterally and at depth, which has the potential to identify zones of differential infiltration (e.g. Shanafield et al., 2020) and soil moisture (e.g. Robinson et al., 2012). Aeromagnetic surveys have been used to identify kilometer scale lithologic changes and the presence of faults and other structural features over entire river basins (Drenth et al., 2011; Fitterman & Grauch, 2010; Grauch et al., 2013). At the smaller scale, electromagnetic induction (Burrell et al., 2008; Shanafield et al., 2020) and electrical resistivity tomography (Clifford & Binley, 2010; Martinez-Segura et al., 2019; Stonestrom et al., 2003) have been used in dry streambeds to differentiate meter-scale subsurface lithology laterally; ground penetrating radar has been used to estimate streambed thickness (e.g. Burrell et al., 2008) and to differentiate lithologies at depth (e.g. Gu et al., 2019). Callegary et al. (2007) used electromagnetic induction measurements along with measurements of channel and vegetation characteristics to estimate the recharge potential of ephemeral stream channels. Time-lapse electrical resistivity tomography has been used to image preferential flow paths beneath ephemeral streams and changes in subsurface resistivity over time (e.g. Koehn et al., 2018). Historically, geophysics has been underutilized in parameterizing zones of differential seepage within these systems.

Numerical modeling approaches have also been used to estimate streambed seepage (Dillon & Liggett, 1983; Niswonger et al., 2008; Niswonger et al., 2005; Noorduijn et al., 2014;
Shanafield et al., 2012). To date, most ephemeral river modeling has focused on canals, where the channel geometry is well known and systems can be replicated by a simple model (e.g. Ghobadian & Fathi-Moghadam, 2014), or along rivers where extensive data collection has been possible (e.g. Niswonger et al., 2008; Wang et al., 2017). Extensive measurements allow for the hydraulic conductivity of the streambed along each river reach to be determined by model calibration (e.g. Noorduijn et al., 2014; Shanafield et al., 2012). Hypothetical ephemeral stream hydrologic models have been used to run scenarios in which one or multiple model input parameters are changed to assess the extent to which those changes impact the model results. For example, Reid & Dreiss (1990) modeled a series of fluvial depositional scenarios—including a homogeneous subsurface, a system with low permeability streambed sediments, and both a narrow and extensive low permeability layer within the alluvial aquifer—to assess the impact to recharge timing and amount. Most of the studies done at the surface assume either that the system is homogeneous or that the volume of flow is the main control on recharge. Some studies have explored shallow heterogeneities using geophysics to define zones of differential streambed seepage (e.g. Shanafield et al., 2020), others look at near surface (< 5 m below ground surface) fluvial heterogeneities (Reid & Dreiss, 1990), or use numerical modeling approaches to estimate streambed seepage (Noorduijn et al., 2014). These approaches are often done in isolation rather than taking a coupled approach and combining methodologies.

Here, we develop a simplified, integrated surface water-groundwater model to explore how geology is a driver of streambed recharge, constrained by field hydrological and geophysical measurements. We used the Alamosa River in the San Luis Valley of south-central Colorado, which is representative of managed ephemeral streams in agricultural regions within semi-arid environments, as a test case (Figure 2.1A on page 8). Data collection included the installation of stilling wells and the collection of streamflow measurements and geophysical data, including a drone magnetometer survey, surface electrical resistivity, and seismic refraction. This field data was then used to parameterize a generalized ephemeral river model using the U.S. Geological Survey’s modular finite-difference flow model along with the streamflow routing package (MODFLOW-SFR) (Harbaugh, 2005; Niswonger & Prudic, 2010) as a base case. We then conducted a scenario evaluation of a series of subsurface geological scenarios to examine the extent to which geologic heterogeneity impacts 1) streamflow, 2) streambed leakage, 3)
groundwater levels, and 4) recharge to the confined aquifer. We specifically explored 1) a fault, as low permeability faults form barriers to groundwater flow and high permeability faults may connect otherwise disconnected aquifers and enhance groundwater flow (e.g. Bense & Van Balen, 2004); 2) unconfined aquifer properties, which impact groundwater flow and river discharge (e.g. Hinton et al., 1993); 3) streambed properties, which influence the rate of streambed leakage (e.g. Niswonger & Prudic, 2010); and 4) the role of confining units, which limit vertical flow (e.g. Mulligan et al., 2007). We also assessed the impact of a human-induced perturbation through the installation of a pumping well within our model, which was expected to increase instream depletions (e.g. Butler et al., 2001).

2.3 Site Description

The San Luis Valley is a basin covering an area of 5,030 km$^2$ with an average elevation of 2,350 m (Emery et al., 1969). The valley is bounded to the west by the San Juan Mountains and to the east by the Sangre de Cristo Mountains (Figure 2.1A). The San Juan Mountains are composed of volcanic flows, tuffs, and breccias, whereas the Sangre de Cristo Mountains are made up of igneous, metamorphic, and sedimentary rocks. The valley floor is bordered by alluvial fans. The valley is underlain by up to 9,000 m of alluvium, clay, volcanic debris, and interbedded volcanic flows and tuffs that were deposited between the Oligocene and the Holocene (Emery et al., 1969).

The San Luis Valley drains a snowmelt-fed watershed area of 7,560 km$^2$ (Figure 2.1A). The northern part of the valley is a closed basin that is drained internally by evapotranspiration; the southern portion is drained by the Rio Grande River and its tributaries, which are the main water source for the residents of the San Luis Valley (Emery et al., 1969). Well logs obtained from the Colorado Division of Water Resources (CDWR) indicate the presence of the unconfined aquifer, the confining unit, and the upper confined aquifer in the study area; none of the wells are drilled deep enough to reach the lower confined aquifer. This suggests that the study area would best be modeled as a three layer hydrologic system. Recharge to the confined aquifers predominantly occurs along the edge of the basin, especially in the San Juan Foothills, which drives water from the basin edge down into the center of the basin (Figure 2.2; Harmon, 1987). Shallow aquifer recharge occurs in the center of the basin due to mountain stream seepage into the alluvial fans along the valley floor during periods of snowmelt (Emery et al., 1969).
Deeper aquifer recharge occurs in the center of the basin due to fractures, faults, and jointing that bring all aquifers into hydraulic connection with one another (Hanna & Harmon, 1980).

Figure 2.1 A) Site map locating the study area in the San Luis Valley, Colorado between the San Juan Mountains to the west and the Sangre de Cristo Mountains to the east; B) study area just northwest of the town of Capulin shaded in yellow, showing the locations of hydrological and geophysical data collection: stilling wells as dark blue dots, head gates as grey dots, drone aeromagnetic survey lines in white, seismic lines in red, and the resistivity lines in purple.
Figure 2.2 Simplified San Luis Valley cross sections detailing subsurface geology and generalized groundwater flow paths (modified from Hawley & Kernodle, 1999 and Wasiolek, 1995). Study area is designated by the star. Not to scale.
Of the valley’s three principal aquifers—the unconfined unit, the upper confined aquifer and the lower confined aquifer—the first two units are the most productive. The unconfined unit is composed of the upper sands of the Alamosa Formation and typically ranges between 12 and 30 m thick. The hydraulic conductivity of the unit ranges from 10 to 70 m/day, increasing towards the west. Both the blue-grey lacustrine clays at the base of the Alamosa Formation and the volcanic deposits of the Hinsdale Formation form the upper confining unit (Figure 2.2), which reduces the vertical hydraulic conductivity, leading to the stratification of the unconfined and upper confined aquifer (Harmon, 1987b). The upper confined aquifer is made up of the Lower Alamosa, Santa Fe, and Los Pinos Formations and varies in thickness from 150 to 300 m (Figure 2.2). The hydraulic conductivity ranges from 0.5 to 60 m/day. The lower confined unit ranges from 750 to 4500 m deep, with hydraulic conductivity and water quality decreasing with depth (Hanna & Harmon, 1980). Kilometer scale aeromagnetic data (Bankey et al., 2005) indicates the presence of a north-south trending fault to the north of the study area (personal communication, V.J.S Grauch & B. Drenth, April 8, 2020), which Harmon & Seitz (in press) hypothesized may extend into study area and bring the aquifers into hydraulic connection.

The Alamosa River, one of many ephemeral tributaries to the Rio Grande River, originates in the San Juan Mountains as snowmelt runoff and runs through the Terrace Reservoir, which is located approximately 40 kilometers west-southwest of Alamosa in south-central Colorado (Figure 2.1A). Inflows flow naturally into the reservoir year-round and outflows are managed by the Colorado Division of Water Resources (CDWR). From April to November, outflows are set to mimic inflows by releasing water at the rate of the previous day’s inflows along with any requested storage releases. The City of Capulin, individuals, and other water entities own shares in the Terrace Reservoir and can request to have those shares released for the purposes of irrigation at any time. From December to March, no water is released from the reservoir and the Alamosa River downstream of the reservoir is fed by precipitation and is typically dry for a significant portion of this time period. When the river is flowing, most of the water is diverted by an extensive network of irrigation ditches. Thus, during the late summer/fall the river does not reach Capulin, which is 24 kilometers downstream of the reservoir (Figure 2.1A).
Just upstream of Capulin is our field site/study area (Figure 2.1B), a 4.5 km stretch of the Alamosa River, which runs from 0.44 kilometers west of the Gunbarrel Road river crossing to the County Road 8 river crossing; the model domain runs from 0.1 km east of the Gunbarrel Road river crossing to the County Road 8 river crossing. This reach is heavily managed and contains seven head gates used for measuring river diversions. Farmers have described this reach as containing an area where the river goes dry over 1 km during the end of the irrigation season and Ford & Skidmore (1993) noted 13-15% seepage losses in this area. To quantify the amount of water available for use in the region, the state developed the Rio Grande Decision Support System (RGDSS), a collection of numerical models including consumptive use (StateCU), groundwater (MODFLOW), surface water (StateMOD) and water budget (StateWB) models. The RGDSS has struggled to predict accurate groundwater levels for this reach of the Alamosa River for the past decade (Harmon & Seitz, in press), which is problematic because Colorado relies on these models to calculate available groundwater and to predict groundwater movement in the area. The RGDSS also aids in the fulfillment of the 1939 Rio Grande Compact, which promises substantial water deliveries to New Mexico and Texas.

2.4 Methods

2.4.1 Hydrological Field Methods

Two stilling wells were installed in the early spring of 2019 to track the river stage in the Alamosa River through snowmelt runoff and the irrigation season at two points above and below the recharge area: SW-1 was placed 0.4 km upstream of the Gunbarrel Road river crossing and SW-2 was placed 0.8 km upstream of the County Road 8 crossing (Figure 2.1B). At each stilling well, velocity profiles were collected on August 6-8, 2019 to estimate river discharge. Given that velocity profiles were only collected at this one time, due to unsafe high flows earlier in the season and no flows shortly thereafter, it was not possible to develop a rating curve. We consequently have a stage record for six months and an estimate of river discharge only from August 6-8, 2019. The discharge data from SW-1 are used to calculate the modeled streamflow rate at the upstream end of the model. The discharge data from SW-2 are used to determine model parameters by matching modeled streamflow to the measured streamflow at SW-2.
2.4.2 Geophysical Field Methods

Geophysical measurements were collected to explore subsurface controls on the geology near the Alamosa River and to parameterize zones within the model. On November 30, 2018, a drone magnetometer survey was conducted by Juniper Unmanned (www.juniperunmanned.com) covering 5.5 km from 1.4 km north of the Gunbarrel Road Alamosa River crossing to 0.2 km East of the County Road 8 crossing (Figure 2.1B). The area was covered by three flight lines. The data were processed by Juniper Unmanned and provided as total magnetic intensity (TMI) for the surveyed area, which is a function of both induced and remnant magnetization. High TMI values indicate geologic features with strong magnetizations, such as volcanic rock (Salem et al., 2007). Contrasts in TMI data have been used previously in the San Luis Valley to locate abrupt changes in subsurface geology (Grauch et al., 2013).

Additionally, a 240 m electrical resistivity profile with forty-eight electrodes spaced 5 m apart was conducted from 1 km upstream of the County Road 8 river crossing trending west (Figure 2.1B). The dipole-dipole survey was collected parallel to the river using an IRIS Syscal Pro Resistivity meter, which transmits a low-frequency alternating current to a pair of electrodes and measures a voltage difference at another pair of electrodes. Each four-electrode measurement is called a quadripole, and our data set consisted of 1159 quadripoles. Three replicate quadripole sets were collected, which took approximately 40 minutes each. Resistance values for the three runs were averaged. The average values were then assigned data weights based on the standard deviations between the three data sets and inverted using R2 (version 3.1, http://www.es.lancs.ac.uk/people/amb/Freeware/R2/R2.htm). A square mesh with 1.25 m spacing in x and z was used. The lower boundary was set at a depth of 210 m and the right and left boundaries were set 1925 m away from the profile so as not to impact the area of interest. The upper boundary was constrained by topography. A resolution matrix was calculated to indicate where the model was constrained by data (Binley & Kemna, 2005).

Hammer seismic refraction surveys were conducted on October 10 and November 21, 2019. The surveys were conducted at six locations spaced out along the area covered by the drone aeromagnetic survey at a range of orientations, where a total of 10 lines were collected using a Geometrics Geode Seismograph (Figure 2.1B). At each location, 25 geophones were set up at 3 m spacing, resulting in a survey length of 75 m. Each line contained 3 shot locations, 1.5
m ahead of the first geophone, in the middle of the line at 37.5 m, and 1.5 m after the last geophone. Each shot was exported and analyzed by manually picking first arrivals of the primary wave (p-wave) present on the time-distance plot output from the seismograph. This approach assumes multilayered system in which the top layer has the slowest velocity and each subsequent layer has a faster velocity. In an alluvial system, these velocity changes can be caused by the water table interface or by the geologic unit contacts, which both result in acoustic impedance changes (e.g. Gálfi & Palos, 1970). When the contact between the first and second layers is flat over the survey distance, the results of the 3 shots should be nearly identical. Under these circumstances, the thickness of the top layer \( h_1 \) for each shot can be calculated using the following equation:

\[
h_1 = \frac{t_i}{2} \frac{V_2 V_1}{(V_2^2 V_1^2)^{1/2}}
\]

where \( V_1 \) = the velocity of the top layer (m/s), \( V_2 \) = the velocity of the second layer (m/s), and \( t_i \) = the intercept of the \( V_2 \) line.

2.4.3 Modeling with MODFLOW

Niswonger & Prudic (2010) developed the streamflow routing package (SFR2) as an add-on to MODFLOW that simulates stream and aquifer interactions by routing flow through a network of surface water channels. SFR2 has been used for sensitivity analyses to determine the relative importance of model input parameters including streambed conductivity \( (K_s) \), stream cross sectional area \( (A) \), channel slope \( (S_0) \), stream width at bankfull, and Manning’s n \( (\text{e.g. Niswonger et al., 2008}) \), as well as applied to field sites \( \text{(e.g. Milzow et al., 2009; Niswonger et al., 2008)} \).

Here, a MODFLOW-SFR model was designed to resemble a highly simplified version of the Alamosa River’s geologic and hydrologic system (Figure 2.3; Table 2.1); that is, a river above unconfined and confined aquifers, separated by a confining unit. The model is 4 km long by 2 km wide, with a cell size of 40 m by 40 m and three layers which represent the Alamosa River Alluvium (Layer 1); the Alamosa Formation Clay or the Hinsdale Basalt (Layer 2), a confining unit; and the Lower Alamosa, Santa Fe, and Los Pinos Formations (Layer 3), which comprise the upper confined aquifer. The components of the upper confined aquifer have
hydraulic conductivity values within a similar range and thus were modeled as a unit. The thickness of Layers 1 and 2 are 18 m and 5 m, respectively. The thickness of Layer 3 ranges from 77 m at the upstream boundary to 55 m at the downstream boundary. The layer thicknesses were determined from well logs drilled along the reach that were obtained from CDWR; the well logs indicate the presence of the Alamosa Formation Clay in most wells in the modeled area and the existence of the Hinsdale Basalt in the eastern portion of the reach. The hydraulic conductivity values for Layers 1 and 3 were derived from Hanna & Harmon (1980) and the nearest order of magnitude value was used. The streambed is filled with gravel and cobbles and streambed hydraulic conductivity ($K_s$) values were taken from the literature (Niswonger et al., 2008).

Figure 2.3 Conceptual model detailing the MODFLOW-SFR setup for the base case (scenario 0) and modeling scenarios including the fault location (scenario 1), variations in properties along the recharge segment (scenarios 2-4), and the pumping well location (scenario 5).

The elevations of the start and end of the modeled reach were taken from LiDAR data (Giffin, 2011) to be approximately 100 and 73 m. The river and associated aquifer were modeled as a sloping plane. The upstream and downstream model boundaries were set as constant-head boundaries with values of 91 and 62 m, estimated from water levels measured in wells near the upstream and downstream ends of the model respectively. The same constant-head boundary values were used for all three layers. The northern and southern model boundaries, parallel to flow, are no-flow boundaries. The river is modeled as a straight reach
3920 m long and 10 m wide through the middle of the model domain. The river and each modeled layer are homogenous and thus one set of aquifer parameter values is used for each (Table 2.1). The river was divided into three segments based on the drone aeromagnetic data signatures, where the upstream and downstream buffer segments represents the high TMI zones, and the central segment represents the low TMI zone thought to be a recharge area (Figure 2.3). The buffer segments separate the recharge segment from the upstream and downstream constant-head boundary conditions and thus data from the buffer segments were not analyzed. The models were run as steady-state flow models.

Table 2.1 Base case river and aquifer properties.

<table>
<thead>
<tr>
<th>Value</th>
<th>Variable and Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 x 10(^{-3})</td>
<td>(K_1) – Layer 1 hydraulic conductivity (m/s)</td>
</tr>
<tr>
<td>1 x 10(^{-6})</td>
<td>(K_2) – Layer 2 hydraulic conductivity (m/s)</td>
</tr>
<tr>
<td>1 x 10(^{-4})</td>
<td>(K_3) – Layer 3 hydraulic conductivity (m/s)</td>
</tr>
<tr>
<td>0.04</td>
<td>(n) – Manning’s roughness coefficient (metric units)</td>
</tr>
<tr>
<td>1 x 10(^{-4})</td>
<td>(K_s) – Hydraulic conductivity of the streambed (m/s)</td>
</tr>
<tr>
<td>3.38</td>
<td>(F_s) – Streamflow at upstream model extent (m(^3)/s)</td>
</tr>
<tr>
<td>0.3</td>
<td>(ST) – Thickness of streambed material (m)</td>
</tr>
<tr>
<td>99.73</td>
<td>(ELEVUP) – Elevation of the top of the streambed (m) at the upstream end of the model</td>
</tr>
<tr>
<td>73.28</td>
<td>(ELEVDN) – Elevation of the top of the streambed at the downstream end of the model</td>
</tr>
<tr>
<td>10</td>
<td>(WIDTH) – Width of the river (m)</td>
</tr>
</tbody>
</table>

A base case model (scenario 0) was simulated using realistic model parameters with the assumption that the flow regime is the dominant factor in the extent and location of recharge; this base case was then compared to five scenarios, outlined below. Alluvial hydraulic conductivity data suggest Layer 1 hydraulic conductivity (\(K_1\)) (Hanna & Harmon, 1980) and streambed hydraulic conductivity (\(K_s\)) (Niswonger et al., 2008) can reasonably vary over three orders of magnitude from \(1 \times 10^{-3}\) – \(1 \times 10^{-5}\) m/s; thus, nine models were tested with a range of values for both parameters from \(1 \times 10^{-3}\) – \(1 \times 10^{-5}\) m/s for the base case to see what best matched measured flow data on August 7, 2019 at the end of the recharge segment.

In addition to the base case model, a scenario evaluation was conducted to look at how subsurface geological heterogeneities as well as pumping along the recharge segment impacted
modeled output metrics, specifically: 1) streamflow, 2) streambed leakage, 3) groundwater levels, and 4) net recharge to Layer 3 (the confined aquifer). These output metrics were chosen because they highlight key processes that were expected to be impacted by subsurface geology. The five scenarios are detailed below and in Table 2.2 and include scenario 1) adding a fault; 2) varying Layer 1 hydraulic conductivity \( K_1 \); 3) changing the thickness of the streambed; 4) removing the confining unit (Layer 2); and 5) adding a pumping well.

Table 2.2 The five modeling scenarios, with subscenario variations.

<table>
<thead>
<tr>
<th>Scenario</th>
<th>Parameter(s) Changed</th>
<th>Parameter Value(s)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>—</td>
<td>—</td>
<td>Base case model</td>
</tr>
<tr>
<td>1</td>
<td>( K_f )</td>
<td>1 m/s</td>
<td>Fault introduced at 1980 m</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1 x 10^{-7} m/s</td>
<td>High permeability fault</td>
</tr>
<tr>
<td>2</td>
<td>( K_1 )</td>
<td>5 x 10^{-3} m/s</td>
<td>Increasing ( K_1 ) from base case</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5 x 10^{-4} m/s</td>
<td>Decreasing ( K_1 ) from base case</td>
</tr>
<tr>
<td>3</td>
<td>( ST )</td>
<td>1.0 m</td>
<td>Increasing ( ST ) from base case</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.5 m</td>
<td>Increasing ( ST ) from base case</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.1 m</td>
<td>Decreasing ( ST ) from base case</td>
</tr>
<tr>
<td>4</td>
<td>( K_2 )</td>
<td>( K_1=1 \times 10^{-3} ) m/s</td>
<td>No confining unit</td>
</tr>
<tr>
<td></td>
<td>( K_3=1 \times 10^{-4} ) m/s</td>
<td>Layer 1 thickened to include Layer 2</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>( PR )</td>
<td>1 x 10^{-1} m(^3)/s</td>
<td>Pumping well introduced at 1980 m</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1 x 10^{-2} m(^3)/s</td>
<td>High volume production well</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1 x 10^{-3} m(^3)/s</td>
<td>Low volume production well</td>
</tr>
</tbody>
</table>

For scenario 1, a fault was introduced at the midpoint, equal to the width of one model cell, crossing the model domain in a vertical north-south orientation and penetrating all three model layers (Figure 2.3), approximating the Alamosa River fault hypothesized by Harmon (in press). To simulate a high permeability fault, a fault conductivity \( K_f \) of 1 m/s was tested (scenario 1.1); to explore a low permeability, mineralized fault, a low \( K_f \) value of 1 x 10^{-7} m/s was tested (scenario 1.2). In scenario 2, \( K_1 \) was increased and decreased along the recharge segment to test the influence that it had on the ephemeral river system over an order of
magnitude. A $K_1$ of $1 \times 10^{-3}$ m/s was used in the base case, which is at the upper range of known values for alluvium in the San Luis valley; here, $K_1$ was increased to $5 \times 10^{-3}$ m/s (scenario 2.1) and decreased to $5 \times 10^{-4}$ m/s (scenario 2.2). For scenario 3, the streambed thickness (ST) was increased and decreased. The ST parameter value used in the base model was 0.3 m and values of 1.0 m (scenario 3.1) 0.5 m (scenario 3.2) and 0.1 m (scenario 3.3) were analyzed. Scenario 4 explored the role of a confining unit in an ephemeral hydrologic system by removing it. The confining unit, Layer 2, was first replaced by thickening Layer 1, $K_2=1 \times 10^{-3}$ m/s (scenario 4.1) and then the confining unit was replaced by thickening Layer 3, $K_2=1 \times 10^{-4}$ m/s (scenario 4.2). Scenario 5 looked at the impact of introducing a single pumping well, screened in the unconfined aquifer, at the midway point of the model at 1980 m. Three pumping rates we were tested: $1 \times 10^{-1}$ m$^3$/s (scenario 5.1), $1 \times 10^{-2}$ m$^3$/s (scenario 5.2) and $1 \times 10^{-3}$ m$^3$/s (scenario 5.3). These rates mimicked the rates of a large-scale production well, a small-scale production well, and a domestic well; CDWR well records indicate the presence of domestic wells in the study area and municipal wells just downstream in Capulin; the presence of sprinkler pivot suggests the presence of irrigation wells (Figure 1B).

2.5 Results

2.5.1 Hydrological Field Results

Velocity profiles were collected at SW-1 and SW-2 on August 6-8, 2019. On both August 6th and August 7th, streamflow at SW-1 was 5.15 m$^3$/s; on August 7th and August 8th, streamflow at SW-2 was 2.38 m$^3$/s and 2.40 m$^3$/s respectively, and 2.77 m$^3$/s and 2.79 m$^3$/s when accounting for diversions. From the SW-1 data, the predicted inflow to the model domain was calculated, taking into account known diversions that day to irrigation ditches as well as 13-15% seepage loss, the typical amount along this reach of the Alamosa River (Ford & Skidmore, 1993). On both August 6th and August 7th inflow to the model was calculated to be 3.88 m$^3$/s. These data were then used to update the model by matching modeled streamflow to the measured streamflow at SW-2, ensuring that the model accurately represented streamflow.

2.5.2 Geophysical Field Results

The drone magnetometer survey measured variability in total magnetic intensity (TMI) in nanoteslas along the surveyed reach (Figure 2.4); these data constrain subsurface geology. Well construction reports indicate the presence of the Hinsdale Basalt in the high TMI zone to the
east, which suggests that the high TMI zone could be related to the presence of volcanics. There are little constraining geologic data in the western high TMI zone. The area of greatest interest is the zone of low TMI situated between two zones of high TMI. Similar to the results of Grauch et al. (2013), our data suggest that the TMI contrast present between the low TMI and high TMI zones may represent a fault crosscutting the Alamosa River.

Figure 2.4 Drone magnetometer survey results and the location of the resistivity line. The colormap indicates total magnetic intensity (TMI) values measured in nanoteslas; warm colors indicate high TMI and cool colors show low TMI.

The surface geophysical data support the airborne measurements. The resistivity survey, conducted on the riverbank, shows a resistive layer at a depth of about 3 m that remains consistent over the profile, but varies slightly with depth (Figure 2.5A), and is well constrained by data (Figure 2.5B). This layer may be the cobbly river bottom. The western portion of the
resistivity profile, which overlaps with the eastern portion of the zone of lowest TMI in the drone aeromagnetic data, is more resistive than the eastern portion of the profile; volcanic rocks are typically associated with high TMI (e.g. Drenth et al., 2011) and low resistivity. Both the drone aeromagnetic data and the resistivity data suggest that the low TMI zone is geologically different from the surrounding area.

![Resistivity profile and log10 values of the diagonal of the resolution matrix](image)

Figure 2.5 A) Resistivity profile and B) log10 values of the diagonal of the resolution matrix from west (0 m) to east (250 m). Higher resistivity values shown in yellow indicate more resistive geology. Values near 1 in (B) indicate where the model resolves the subsurface based on data.

The six hammer seismic locations were spread along the reach to cover the three TMI zones. The hammer seismic results indicate an acoustic impedance change at approximately 3 m depth regardless of location. This matches the depth of the change in the resistivity survey, and also likely represents the cobbly river bottom. Water level measurements from CDWR, in close proximity to the seismic lines, indicate that the alluvial water table ranges in depth along the reach of interest from 9 to 15 m below ground level, suggesting that the p-wave first arrivals do not represent the unsaturated/saturated zone interface.
### 2.5.3 Modeling Results

#### 2.5.3.1 Scenario 0: Base Case Model

A $K_1$ value of $1 \times 10^{-3}$ m/s and a $K_s$ value of $1 \times 10^{-4}$ m/s resulted in a streamflow of 2.70 m$^3$/s, which best matched the measured streamflow of 2.77 m$^3$/s at SW-2 on August 7, 2019, resulting in a percent difference of only -2% (Table 2.3; Figure 2.6). Layer 1 groundwater levels ranged from 0.28 m above ground level to 0.26 m above ground level; these were notably more shallow than measured water level measurements of 9-15 m below land surface. A $K_1$ value of $8 \times 10^{-3}$ m/s and a $K_s$ value of $1 \times 10^{-5}$ m/s were able to achieve the observed maximum groundwater level of 9.6 m below ground level (Table 2.4; Figure 2.7); increasing $K_1$ further did not result in a deeper Level 1 groundwater table. However, a $K_1$ value of $8 \times 10^{-3}$ m/s is not within the measured range of $K_1$ values for the Alamosa River Alluvium in the San Luis Valley (Harmon, 1987a). Given that we could not match measured groundwater data using realistic input parameters, we chose to match streamflow with a $K_1$ value of $1 \times 10^{-3}$ m/s and a $K_s$ value of $1 \times 10^{-4}$ m/s, which were used for the base case model going forward.

Table 2.3 Modeled streamflow at SW-2 for the base case, scenario 0.

<table>
<thead>
<tr>
<th>$K_1$ (m/s)</th>
<th>$K_s$ (m/s)</th>
<th>Streamflow at SW-2 (m$^3$/s)</th>
<th>Percent Difference from Measured Streamflow (%)$^\text{§}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>$1 \times 10^{-3}$</td>
<td>$1 \times 10^{-3}$</td>
<td>2.97</td>
<td>7%</td>
</tr>
<tr>
<td><strong>$1 \times 10^{-3}$</strong></td>
<td><strong>$1 \times 10^{-4}$</strong></td>
<td><strong>2.70</strong></td>
<td><strong>-2%</strong></td>
</tr>
<tr>
<td>$1 \times 10^{-3}$</td>
<td>$1 \times 10^{-5}$</td>
<td>2.62</td>
<td>-6%</td>
</tr>
<tr>
<td>$1 \times 10^{-4}$</td>
<td>$1 \times 10^{-3}$</td>
<td>3.22</td>
<td>16%</td>
</tr>
<tr>
<td>$1 \times 10^{-4}$</td>
<td>$1 \times 10^{-4}$</td>
<td>3.22</td>
<td>16%</td>
</tr>
<tr>
<td>$1 \times 10^{-5}$</td>
<td>$1 \times 10^{-3}$</td>
<td>3.33</td>
<td>20%</td>
</tr>
<tr>
<td>$1 \times 10^{-5}$</td>
<td>$1 \times 10^{-4}$</td>
<td>3.33</td>
<td>20%</td>
</tr>
</tbody>
</table>

$^\text{§}$ a positive value means modeled streamflow $>$ measured streamflow.

* measured streamflow on August 7, 2019

**Bold** = pair that best matched measured streamflow
Figure 2.6 Sensitivity of streamflow to parameters $K_1$ and $K_s$, which were each varied between $1 \times 10^{-3}$ and $1 \times 10^{-5}$ m/s, for scenario 0, the base case. We varied $K_s$ assuming A) $K_1=1 \times 10^{-3}$ m/s; B) $K_1=1 \times 10^{-4}$ m/s; and C) $K_1=1 \times 10^{-5}$ m/s. We then varied $K_1$ assuming D) $K_s=1 \times 10^{-3}$ m/s; E) $K_s=1 \times 10^{-4}$ m/s; and F) $K_s=1 \times 10^{-5}$ m/s. Lower $K_1$ and $K_s$ values resulted in higher streamflow, with $K_1$ exerting a greater influence on streamflow than $K_s$. 

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For scenario 0, boundary conditions show little impact on groundwater heads along the recharge segment, where the potentiometric surface mirrors the elevation of the river base. The Layer 1 head is above the bottom of the stream, but below the river stage; consequently, the river is a net losing river throughout the model domain (Figure 2.7A). Groundwater head in Layers 2 and 3 remains below the base of the stream. Leakage is the highest at the upstream end of the reach and generally declines along the reach as the flow rate in the river decreases; the bowing pattern in scenario 0 leakage (Figure 2.7B) is the result of the river gaining minimally from groundwater, which reduces the net leakage in the middle of the recharge area. Total streambed leakage along the recharge segment for the base case is 0.11 m$^3$/s. We note that in these models, a positive leakage means more water leaks from the river and into groundwater (i.e. a losing river segment) and a negative leakage means more groundwater is discharging into the river than is leaking through the streambed (i.e. a gaining river segment). Layer 3 is recharged at a rate of 0.035 m$^3$/s, which is driven by downward flow from Layer 2 (Table 2.4). There is no upward flow from Layer 3 into Layer 2.

Figure 2.7 Scenario 0, the base case, and scenario 0 – perturbed, where $K_1$ and $K_s$ values were altered to match groundwater levels – $K_1 = 8 \times 10^{-3}$ m/s and $K_s = 1 \times 10^{-5}$ m/s; A) river stage, river base elevation, and groundwater head in Layer 1; B) streambed leakage. For scenario 0 – perturbed, leakage linearly decreases in response to decreasing streamflow; scenario 0, exhibits a bowing leakage pattern as a result of the river gaining minimally from groundwater, which reduces the net leakage in the middle of the recharge area.
<table>
<thead>
<tr>
<th>Scenario</th>
<th>Parameter Changed</th>
<th>Parameter Value</th>
<th>Streamflow at SW-2 (m³/s)</th>
<th>Max Layer 1 SWL* (m)</th>
<th>Total Streambed Leakage (m³/s)</th>
<th>Net Recharge to Layer 3 (m³/s)</th>
<th>Flow from Layer 3 (m³/s)</th>
<th>Flow to Layer 3 (m³/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>—</td>
<td>—</td>
<td>2.70</td>
<td>—</td>
<td>0.10</td>
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<tr>
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<td>1 m/s</td>
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<tr>
<td>3.2</td>
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<td>2.70</td>
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<td>0.035</td>
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</tr>
<tr>
<td>3.3</td>
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<td>2.70</td>
<td>—</td>
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<tr>
<td>4.1</td>
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<tr>
<td>5.1</td>
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<td>0.035</td>
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<tr>
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<td>2.70</td>
<td>—</td>
<td>0.11</td>
<td>0.035</td>
<td>0.000</td>
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</table>

*SWL = static water level measured below ground level (BGL)
— indicates that the water table in Layer 1 never dropped below ground level
$^4$upstream constant head boundary = 87 ft; downstream constant head boundary = 58 ft
The modeling outputs for scenario 0 provide a basis from which to explore how geologic changes and groundwater pumping affect the modeled hydrologic system. Percent difference values were used to compare each scenario to the base case. For streamflow, changes of $\geq 3\%$ were considered notable; for streambed leakage and net recharge to Layer 3 along the recharge segment, changes of $\geq 10\%$ were considered important. Groundwater level changes of $\geq 0.3$ m at any point along the recharge segment are noted. While these values may ideally be chosen by uncertainties in data, because of limited subsurface measurements, these thresholds were determined by exploring variability within the outputs of the models themselves. Smaller changes were often seen with parameters changes in our models, but did not notably change processes.

2.5.3.2 Scenario 1: Fault Introduced at 1980 m along the River Reach

When a high permeability fault is introduced in scenario 1.1, streamflow decreases sharply from 60 m upstream of the fault through 60 m downstream of the fault, resulting in the streamflow dropping below the base case at 1960 m and reaching a minimum of 0.07 m below the base case at 2060 m; farther downstream, the streamflow flattens out, decreasing at a much slower rate, slower than the base case (Figure 2.8A). This flowrate pattern is caused by changes in streambed leakage along the reach. From ~160 m upstream of the fault to the fault, streambed leakage increases an order of magnitude, reaching peak leakage at the fault location; between the fault and downstream of it by ~80 m, leakage returns to approximately base case levels (Figure 2.8B). However, total streambed leakage from scenario 1.1 was 50% greater than scenario 0, suggesting that the presence of a high permeability fault does not just alter the leakage pattern, but also enhances streambed leakage. Scenario 1.1 also resulted in a decrease in Layer 1 groundwater head below the base of the river in the immediate vicinity of the fault ± 100 m (Table 2.4; Figure 2.8C), resulting in all three model layers coming into hydraulic connection with one another. This is evidenced by the presence of both upward and downward flow at the Layer boundaries, with a net recharge to Layer 3 that is 60% greater than scenario 0. Scenario 1.1 resulted in notable changes to streambed leakage and net recharge to Layer 3. Layer 1 groundwater head changed laterally ± 100 m of the fault. Although, scenario 1.1 results in a streamflow pattern over the recharge segment that is distinctly different from the base case, by SW-2 the streamflow rate is only -1.5% less than the base case.
Figure 2.8 Scenario 1 sensitivity of A) streamflow and B) streambed leakage to fault hydraulic conductivity $K_f$ along the recharge segment. Groundwater head in model layers 1, 2, and 3 along the reach for C) high permeability fault (scenario 1.1) and D) low permeability fault (scenario 1.2). Scenario 1.1 results in increased leakage near the high permeability fault, which leads to a reduction in streamflow; the high permeability fault brings the model layers into hydraulic connection allowing the Layer 1 groundwater level to decrease. Scenario 1.2 results in the formation of a groundwater mound upstream of the fault, which leads to negative leakage and increased streamflow; downstream of the fault there is an abrupt drop in groundwater levels, leakage becomes positive, and streamflow decreases.

The addition of a low permeability fault in scenario 1.2 results in a much larger disruption to the streamflow from the base case. The streamflow increases until the fault is reached and then decreases downstream of the fault, related to the fact that fault is blocking groundwater flow downgradient resulting in streambed leakage that is both negative and decreasing until 40 m upstream of the fault, indicating that the river is gaining water from
groundwater rather than losing water to groundwater. Immediately downstream of the fault, groundwater head drops, streamflow decreases, and leakage increases. Although the leakage pattern for scenario 1.2 is distinctly different from scenario 0, it does not result in notably less total leakage or more total recharge to Layer 3 (Table 2.4). Scenario 1.2 resulted in notable changes to groundwater levels with a distinct groundwater mound upstream of the fault. The streamflow pattern over the recharge segment was visibly different from the base case but resulted in only a 4% increase in streamflow from scenario 0 by SW-2. The changes to streambed leakage and net recharge to Layer 3 are small at approximately -8% and 5%, respectively.

2.5.3.3 Scenario 2: Increasing and Decreasing Layer 1 Hydraulic Conductivity ($K_1$) along the Recharge Segment

When $K_1$ is increased to $5 \times 10^{-3}$ m/s in scenario 2.1 from the base case value of $1 \times 10^{-3}$ m/s, streamflow remains less than scenario 0 and decreases until reaching a minimum of 2.3 m$^3$/s at 2140 m, which is 17% less than scenario 0 (Figure 2.9A). After the minimum streamflow is reached, the streamflow increases through the end of the recharge segment due to the transition of the river from a losing river to a gaining river at that point. Increasing $K_1$ leads to leakage increasing until 1460 m, then decreasing until 2500 m and increasing again until 2520 m (Figure 2.9B). Upstream of 2100 m, leakage is positive, and the river is losing; downstream, leakage is negative, and the river is gaining. The Layer 1 groundwater level reaches a minimum of 0.15 m below the ground surface at 1460 m (Figure 2.9C). After that point, groundwater levels steadily increase due to the $K_1$ decrease at 2520 m (Figure 2.9C). Thus, the impact of the higher $K_1$ zone in the recharge segment is limited by the lower $K_1$ values in the buffer segments. Despite this, total leakage is noticeably higher in scenario 2.1 (0.17 m$^3$/s) than scenario 0 (0.10 m$^3$/s) as leakage is enhanced by the higher $K_1$ value along the recharge segment. Net recharge to Layer 3 is similar to scenario 0, with an increase of ~ 4%. Increasing $K_1$ increased streambed leakage by ~67% when compared with scenario 0, and groundwater head visibly changed, dropping 0.15 m below the streambed. Streamflow was noticeably different than the base case along the recharge segment and the change near SW-2 was notable at ~7%. Net recharge to Layer 3 only increased ~ 4%.
Figure 2.9 Scenario 2 sensitivity of A) streamflow and B) streambed leakage to $K_1$ along the recharge segment. Groundwater head in model layers 1 and 2 for C) increasing $K_1$ (scenario 2.1). Scenario 2.1 resulted in reduced streamflow, where the bowing pattern is caused by the transition of the river from losing to gaining; leakage increases abruptly at near the boundaries of the buffer segments because $K_1$ is increased only along the recharge segment. The opposite effect is noticeable for scenario 2.2, where $K_1$ is decreased.

When $K_1$ is decreased to $5 \times 10^{-4}$ m/s in scenario 2.2, the streamflow is consistently higher than scenario 0 (Figure 2.9A). This is because leakage for scenario 2.2 is lower than scenario 0 upstream of 2300 m. There was no notable change in Layer 1 groundwater levels from the base case (Figure 2.9C), which never dropped below the base of the stream. Total leakage for scenario 2.2 was 27% less than scenario 0 (Figure 2.9B; Table 2.4) and net recharge to Layer 3 was not notably different (Table 2.4). Streambed leakage changed notably from 0.10
m$^3$/s in the base case to 0.08 m$^3$/s in scenario 2.2. Streamflow and net recharge to layer 3 did not visibly change from the base case, with percent changes of <5%.

### 2.5.3.4 Scenario 3: Increasing and Decreasing Streambed Thickness (ST) along the Recharge Segment

When streambed thickness is increased to 1.0 m in scenario 3.1 and to 0.5 m in scenario 3.2 from 0.3 m in the base case, the streamflow remained similar, with values slightly above those in the base case (Figure 2.10A). This is because $K_s$<$K_1$, so when $K_s$ is applied to a thicker streambed, it results in a slight decrease in leakage (Figure 2.10B). Layer 1 groundwater levels and net recharge to Layer 3 were not notably impacted by the increasing ST and total streambed leakage only decreased by ~4% for scenario 3.1 and 2% for scenario 3.2 (Table 2.4).

When ST was decreased to 0.1 m in scenario 3.3, the streamflow rate remained just below the base case (Figure 2.10A) because the thinner streambed reduced the influence of $K_s$, allowing for a slight increase in leakage (Figure 2.10B). Similar to scenarios 3.1 and 3.2, Layer 1 groundwater levels and net recharge to Layer 3 were not visibly impacted by the decrease in ST (Table 2.4).

![Figure 2.10 Scenario 3 sensitivity of A) streamflow and B) streambed leakage to streambed thickness along the recharge segment.](image)

Scenarios 3.1 and 3.2, which increase ST, produce similar streamflow and recharge results and are thus indistinguishable on the plots. The abrupt decrease in leakage for scenarios 3.1 and 3.2 at the boundaries with the buffer segments is caused by the fact that streambed thickness is increased only along the recharge segment, resulting in this abrupt change; the opposite effect is noticeable for scenario 3.3, where ST is decreased.
2.5.3.5 Scenario 4: Removal of the Confining Unit along the Recharge Segment

Scenario 4.1 involves replacing the confining unit with $K_1$, $1 \times 10^{-3}$ m/s, which represents a two layer system, with a thickening of the alluvium (Layer 1) underlain by the confined aquifer material (Layer 3 in the base case) and scenario 4.2 involves replacing the confining unit with $K_3$, $1 \times 10^{-4}$ m/s, which represents a two layer system with the alluvium (Layer 1) overlying a thickened confined aquifer (Layer 3 in the base case). Both scenarios 4.1 and 4.2 resulted in a decrease in streamflow from scenario 0, by 0.04 m$^3$/s and by 0.01 m$^3$/s respectively at SW-2, as more water leaks through the streambed without the presence of the confining unit (Figure 2.11A). Scenario 4.1 resulted in a greater reduction in streamflow and more total streambed leakage than scenario 4.2, because scenario 4.1 assigned a higher K value to the area where the confined unit had been. Streambed leakage for scenario 4.2 was always above scenario 0 along the entire river length (Figure 2.11B), whereas scenario 4.1 had leakage larger than the scenario 0 until 2260 m, where it drops (Figure 2.11B). This decrease in leakage is likely caused by the drop in streamflow rate, the increase in groundwater levels (Figure 2.11C&D), and approaching the boundary of the buffer segment where the confining unit returns. Scenarios 4.1 and 4.2 resulted in notable changes to streambed leakage with values of 0.13 m$^3$/s and 0.12 m$^3$/s, respectively, relative to the base case value of 0.10 m$^3$/s. Net recharge to Layer 3 also notably increased relative to the base case for both scenarios with volumes of 0.066 m$^3$/s and 0.064 m$^3$/s relative to the base case value of 0.035 m$^3$/s. Groundwater levels increased in all three layers, but there was no substantial change in streamflow.

2.5.3.6 Scenario 5: Pumping Well Introduced at 1980 m along the River Reach

Introduction of a pumping well, screened in the alluvial aquifer, in scenarios 5.1, 5.2, and 5.3 unsurprisingly resulted in less streamflow than scenario 0; streamflow decreased more rapidly than the base case until approximately 100 m downstream of the pumping well and then continued decreasing at a rate similar to the base case, resulting in a small change by SW-2 (Figure 2.12A). The decrease in streamflow was proportional to the pumping rate. There was also an increase in leakage in the vicinity of the pumping well, proportional to the pumping rate and resulting in greater leakage than scenario 0 (Figure 2.12B). For scenarios 5.1, 5.2, and 5.3 leakage remained positive along the reach and the river remained a losing stream. Total leakage
for scenario 5.1 was ~90% greater than total leakage of scenario 0; scenario 5.2 and 5.3 resulted in a 9% and 1% increase in leakage relative to scenario 0 respectively. Net recharge to layer 3 was not notably changed by the presence of a pumping well.

![Graphs showing sensitivity of streamflow and streambed leakage to K2 along the recharge segment.](image)

Figure 2.11 Scenario 4 sensitivity of A) streamflow and B) streambed leakage to K2 along the recharge segment. Groundwater head in model layers 1, 2, and 3 for C) replacing the confining unit with K1 (scenario 4.1) and D) replacing the confining unit with K3 (scenario 4.2). Both scenarios 4.1 and 4.2, where K2 is increased, resulted in decreased streamflow and increased groundwater levels. For the beginning of the recharge segment, both scenarios resulted in increased leakage; scenario 4.1 decreased below the base case as the downstream buffer segment boundary was approached due to the fact the K2 was only changed along the recharge segment; scenario 4.2 leakage never decreased below the base case, but similar boundary condition effects were observed at a smaller scale.

Despite adding a pumping well, Layer 1 groundwater head never dropped below the base of the stream. Groundwater head for all three layers for scenario 5.2 and 5.3 closely matched
scenario 0 (Figure 2.12C). For scenario 5.1, Layer 1 groundwater head decreases, approaching the river base, from 50 m upstream of the pumping well to 50 m downstream of the pumping well (Figure 2.12D). Head in Layer 2 also decreases slightly in response to well pumping. Leakage for scenario 5.1 results in a notable increase with a value of 0.20 m$^3$/s compared to the base case value of 0.10 m$^3$/s; there was a visible increase in leakage for scenario 5.2, but not for scenario 5.3. Groundwater head, streamflow, and net recharge to Layer 3 did not notably change as a result of scenarios 5.1, 5.2, and 5.3.

Figure 2.12: Scenario 5 sensitivity of A) streamflow and B) streambed leakage to the presence of pumping wells (PW) at different pumping rates. Groundwater head in model layers 1 and 2 for C) adding a high-volume production well (scenario 5.1) and D) adding a low-volume production well (scenario 5.2). Groundwater levels for adding a domestic well (scenario 3.3) are indistinguishable from scenario 5.2 and are not shown separately. All three scenarios produced proportional decreases in streamflow and groundwater levels as well as proportional increases in leakage.
2.6 Discussion and Conclusions

The drone magnetometer data were used to divide the study area into the recharge segment—based on the low TMI zone—and the buffer segments. These data, along with the resistivity data, suggest that the recharge segment is geologically distinct from the buffer zones. The resistivity and seismic data indicate the presence of the cobbly riverbed and suggest the presence of subsurface heterogeneities at depth, leaving open questions about the nature of the subsurface geology. To explore the geologic controls on the streamflow, streambed leakage, groundwater levels, and net recharge to layer 3, multiple scenarios were explored, specifically looking at five scenarios: 1) introduction of a fault; 2) a higher or lower $K_1$; 3) a thicker or thinner streambed; 4) the absence of the confining unit (Layer 2); and 5) a pumping well.

The addition of a high permeability fault (scenario 1.1) created both upward and downward vertical flow within the fault and allowed all three model layers to come into hydrologic connection with one another, resulting in enhanced recharge into the confined aquifer (Figure 2.13B), as has been seen in fractured aquifer scenarios, for example, by Bense & Van Balen (2004) in the Rhine River. The introduction of a low permeability fault (scenario 1.2) inhibited flow and resulted in the formation of groundwater mound upstream of the fault (Figure 2.8D), transitioning the system from a losing river to a gaining river upstream of the fault. Low permeability faults have led to gaining streams in the field, for example in the work Folch & Mas-Pla (2008).

Changing the hydrogeologic characteristics of the recharge segment also impacted the hydrologic results. Increasing $K_1$ (scenario 2.1) resulted in the Layer 1 groundwater level reaching a minimum of 0.15 m below the ground surface at 1460 m (Figure 2.9C), which is 20 m downstream of the start of the higher $K_1$ zone. This result suggests that increasing $K_1$ along the recharge segment allows for the groundwater to move downgradient faster than upgradient groundwater can replace it, thus resulting in a drop in groundwater level. Increasing $K_1$ also resulted in increased leakage (Figure 2.13B) and decreased streamflow (Figure 2.13A), whereas decreasing $K_1$ (scenario 2.2) led to the opposite response, as would be expected. The effect of changing $K_1$ was more important than that of streambed thickness (scenario 3). As streambed thickness increased (scenarios 3.1 & 3.2), streambed leakage decreased slightly (<4%) due to the fact that the $K_s < K_1$. Streamflow, net recharge to Layer 3, and groundwater levels did not notably
change when ST was varied, suggesting that ST does not have much influence on the river system. We also replaced the confining unit in the recharge zone (scenario 4) by a thickened Layer 1 (scenario 4.1) and by a thickened Layer 3 (scenario 4.2), mimicking a spatially discontinuous confining unit, as the confining unit remained in place in the buffer segments. Both scenarios resulted in increased vertical flow and net recharge to Layer 3 (Figure 2.13C), as well as enhanced leakage (Figure 2.13B) and increased groundwater levels. Interestingly, upward flow was minimal in scenarios 4.1 and 4.2, but was a much greater factor in the fault scenarios (scenarios 1.1 and 1.2). These results would be expected from removing the low permeability confined unit, as were seen in Mulligan et al. (2007), who found that vertical flow was increased along paleochannels crosscutting confining units offshore Wrightsville Beach, North Carolina.

Lastly, a well was introduced to test the impact of pumping at different rates, including that of a large-scale production well, a small-scale production well, and a domestic well, all of which are commonly located near ephemeral streams in arid climates. The introduction of pumping increased streambed leakage and decreased groundwater levels in Layer 1 and Layer 2, proportional to the pumping rate.

In both the base case (scenario 0) and in the modeling scenarios, the river was modeled as a straight reach, where the length of the modeled river along recharge segment equaled the length along the recharge segment in the field as evidenced by the drone magnetometer data. In reality, the Alamosa River contains meanders. Previous research suggests that differential seepage occurs in meandering streams with the highest seepage along the outside of a meander (e.g. Balbarini et al., 2017; Eekhout et al., 2013); this is partially due to the distribution of streamflow (Bennett et al., 2002) as well as differential streambed conductivity (Sebok et al., 2015). Thus, the presence of meanders could alter the location of streambed seepage on the order of meters (Balbarini et al., 2017; Stonedahl et al., 2010). This suggests that modeling the Alamosa River as a straight river likely alters the precise recharge location; however, more research needs to be done to determine to what extent net recharge is altered over a kilometer scale.

Overall, the output metrics were the most influenced by scenarios that involved changing K values, including scenario 1, scenario 2, and scenario 4. These scenarios resulted in notable changes to streambed leakage and groundwater levels. Net recharge to layer 3 was notably
enhanced by scenarios that provided a conduit through the confining unit, including a high permeability fault (scenario 1.1) and removing the confining unit along the recharge segment (scenario 4). Streamflow was notably impacted by scenarios where $K_1$ was altered along the recharge segment (scenario 2). This suggests, unsurprisingly, that the $K$ values are the most important model parameters tested, as they control the movement of groundwater (e.g. Bense & Van Balen, 2004; Mulligan et al., 2007; Niswonger & Prudic, 2010).

Figure 2.13 Percent difference between scenario 0 and the various scenarios for A) flowrate at SW-2, B) total streambed leakage along the recharge segment, and C) net recharge to Layer 3 along the recharge segment. Scenarios are 1.1: adding a high permeability fault; 1.2: adding a low permeability fault, 2.1: increasing Layer 1 $K$ ($K_1$); 2.2: decreasing $K_1$; 3.1: increasing streambed thickness (ST) to 1 m; 3.2 increasing ST to 0.5 m; 3.3: decreasing ST to 0.1 m; 4.1: removing the confining unit ($K_2$) and replacing it with $K_1$; 4.2: removing the confining unit ($K_2$) and replacing it with $K_3$; 5.1: adding a high volume production well; 5.2: adding a low volume production well; 5.3: adding a domestic well. Pumping wells is designated by PW.

The output metrics were least influenced by changing the thickness of the streambed (scenario 3) and by introducing a pumping well (scenario 5). Scenario 3 and the introduction of either a low-volume production well (scenario 5.2) or the introduction of a domestic well (scenario 5.3) resulted in no notable changes to any of the output parameters; the introduction of a high-volume pumping well (scenario 5.1) resulted in a notable increase in leakage, but no notable change to streamflow or net recharge to Layer 3. Scenario 5.1 resulted in a notable decrease in water levels near the pumping well, but scenario 5.2 resulted in no notable change in groundwater levels. This suggests that changing the thickness of the streambed and introducing domestic well have more minimal impacts on a river’s hydrologic system. Niswonger & Prudic (2010) found the streambed thickness to result in greater changes than we found to leakage and
Layer 1 groundwater levels; however, increasing leakage and groundwater levels are also correlated to an increase in streamflow, so it is possible that the influence of streambed thickness increases as streamflow increases.

As noted above, changing $K$ had the greatest impact on the output metrics; however, increasing and decreasing $K$ did not result in equally proportional changes. For scenario 2 where $K$ was both increased and decreased, increasing $K$ led to a greater increase in leakage than a decrease of the same amount did. This may be due to the relationship between the $K_1$ value and the $K_s$ value. For all scenarios $K_s$ remained $1 \times 10^{-4}$ m/s. In scenario 2.1, a $K_1$ value of $5 \times 10^{-3}$ m/s was used, and in scenario 2.1, a $K_1$ value of $5 \times 10^{-4}$ m/s was used. Thus, the ratio of $K_1$:$K_s$ for scenario 1.1 is 50:1 and the ratio for scenario 1.2 is 5:1. The larger $K_1$:K_s ratio for scenario 1.1 may explain why the increase in leakage from scenario 1.1 was greater than the decrease from scenario 1.2. We note that the streamflow output metric is the best constrained by our field system, as the base case $K_1$ and $K_s$ parameters were chosen to match streamflow at SW-2. Streambed leakage is inversely related to streamflow; consequently, streambed leakage values are also thought more meaningful. Impacts on groundwater levels and net recharge to Layer 3 are only estimates of what could be occurring; we do not have constraints on these estimates from the field.

We note that many models could be produced to explore the system, and we only consider a few of the infinitely many possibilities that could have been modeled. We chose the scenarios we did based on our field system. The fault scenario (scenario 1) is supported by the drone aeromagnetic data, and was hypothesized by Harmon (in press) based on previous work. The existence of pumping wells in the study area is documented by CDWR, specifically domestic wells (i.e., scenario 5.3) and aerial imagery suggests that wells are also being used for irrigation (Figure 1B), which would pump at a higher rate resembling a low volume production well (scenario 5.2). For scenario 2, a higher or lower $K_1$; scenario 3, a thicker or thinner streambed; and scenario 4, the absence of the confining unit; there are no data present that suggest these exact conditions along the Alamosa River; however, it is expected that heterogeneities of these sorts occur in most river systems.

Our results can be used to help update the RGDSS model in this complicating part of the system. Based on our results, mapping hydraulic conductivity is important; however, because
real streams are spatially heterogeneous, capturing these data in sufficient resolution may be difficult. Geophysics has the potential to map heterogeneities relatively quickly and inexpensively, and can constrain the location of facies such as confining units that control flow and transport in some systems (e.g., Fitterman & Grauch, 2010; Gu et al., 2019; Malenda et al., 2019; Martinez-Segura et al., 2019). For the system here, shallow constraints on facies may be best found using electromagnetic induction (e.g. Shanafield et al., 2020), whereas resistivity (e.g. Clifford & Binley, 2010; Rodriguez et al., 2007) and seismic profiles (e.g. Mulligan et al., 2007) or downhole logs (e.g. Jimoh et al., 2018) may be more useful in constraining deeper heterogeneities. However, geophysical data are most accurate when they can be ground truthed with borings or well logs. Our detailed investigation of the drillers log data in the vicinity of Capulin indicates that these data could be used to update layer thicknesses in the RGDSS model to more accurately reflect the geology, which currently is a 5 layer model constructed using generalized basin-scale cross sections (Harmon & Seitz, in press).
CHAPTER 3
FUTURE WORK

Major conclusions from this thesis were that the drone magnetometer indicate that the recharge area is geologically distinct from the surrounding area and the resistivity and seismic data suggest the presence of heterogeneities at depth. The modeling results indicate the importance of hydraulic conductivity values in influencing the river’s hydrologic system by altering streamflow, leakage, and deep aquifer recharge as well as the unimportance of the streambed thickness parameter. The model suggests that the extent to which streambed leakage changes is proportional to the ratio of alluvial aquifer hydraulic conductivity ($K_1$) to streambed conductivity ($K_s$).

If this project faced neither time nor budgetary constraints, I would suggest collecting more field data iteratively over time. First, I recommend collecting an airborne time-domain electromagnetic (TDEM) survey over an area of 10 km$^2$ including the study area. This would allow me to better explore the extent of features imaged in the December 2018 drone magnetometer survey, while achieving greater depth resolution. Based on those data, I recommend drilling six monitoring wells along the Alamosa River: two in the recharge zone and two in each buffer segment; during drilling, split spoon samples should be collected every meter to detail changes in the lithology. In each segment one well would be screened in the unconfined alluvial aquifer and the other well would be screened in the confined aquifer. I would conduct pumping tests in each aquifer to determine hydraulic conductivity of Layers 1 and 3 along each segment. In addition to using the monitoring wells to learn about the subsurface heterogeneities, the wells could also be used to track groundwater levels in both the unconfined alluvial aquifer and the confined aquifer. Pressure transducers would be installed in the wells so that groundwater levels can be tracked continuously over time.

At the stilling wells, additional streamflow measurements should also be collected, and a rating curve developed to track the streamflow over the course of the irrigation season. During the winter when the riverbed is dry, conducting constant-head permeameter tests along each segment would help to constrain the hydraulic conductivity of the streambed.
To correlate geologic units from the lithologic logs with their geophysical signatures, wireline logs would be collected at each monitoring well including resistivity, gamma, acoustic televIEWer, heat-pulse flow meter, and fluid conductivity logs. If most heterogeneities noted in the wells were < 10 m below land surface, electromagnetic induction data would be collected over the full study area using the DualEM, a 4-m long instrument that is carried at hip height. If heterogeneities occurred at depths > 10 m, a series of resistivity profiles would be collected in the recharge zone and along the boundaries of the recharge zone and the buffer segments, as it has the potential to see down 10s of meters, depending on conditions. The geophysical data would be correlated with the associated signature from the well geophysical logs and thus lithologic unit. Then using the spatial distribution of the geophysical signatures, the lithologies could be mapped and the model layering could be updated accordingly, which would allow for a more realistic base case model to be created. This would likely allow for the development of a base model that is better able to match both streamflow and groundwater levels, while using realistic parameter values.

Given a more realistic base case model, the scenarios considered in this thesis should be rerun by amending the new base model with more accurate layer thicknesses and hydraulic conductivity values. Furthermore, more complicated model scenarios could be evaluated; two previously tested variables could be changed at once and the combined effect of those changes could be analyzed. This would allow for the exploration of the relationships between model parameters, such as the K1:Ks ratio, as well as the impact of a high permeability fault next to a pumping well. These complex scenarios could allow for a further assessment of the extent to which subsurface geology constrains ephemeral river recharge.
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APPENDIX A
HOW TO BUILD A MODFLOW-SFR FILE

The Streamflow-Routing (SFR2) Package is an add-on to MODFLOW and simulates stream-aquifer interaction with the option of including unsaturated flow beneath streams. This document details the process to set up a MODFLOW SFR file with the following assumptions: 1) the model will be run at steady state, 2) the stream channel is assumed to be rectangular, 3) stream reaches are defined and stream properties are assigned to the upstream and downstream ends of the reach (a linear interpolation is done between these points), and 4) unsaturated flow is ignored. I do briefly discuss how the SFR file needs to be modified to deviate from these assumptions, but further explanation can be found in both the SFR1 (Prudic et al., 2004) and SFR2 (Niswonger & Prudic, 2010) documentation.

Building the SFR File

The SFR Package requires a single input file, a text file with the .sfr file extension as shown in Figure A.1, where each line is numbered to assist with explaining the purpose of each line as well as each of the input values. For ease of explanation, each input value will be assigned a letter in alphabetical order from left to right. For example, from line 1, 98, is assigned the value 1a and 1 is assigned the value 1e. Table A.1 correlates the line numbers to the corresponding sections in the SFR2 documentation.

Table A.1 Relation between Figure A.1 section number and the SFR2 documentation in Appendix 1 Section Number

<table>
<thead>
<tr>
<th>Figure A.1 Section Number</th>
<th>SFR2 Documentation – Appendix 1 Section Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>3</td>
<td>5</td>
</tr>
<tr>
<td>4</td>
<td>6a</td>
</tr>
<tr>
<td>5</td>
<td>6b</td>
</tr>
<tr>
<td>6</td>
<td>6b</td>
</tr>
</tbody>
</table>
Figure A.1 SFR example file delineated with numbers to show the file input format. Dotted lines indicate where entries have been removed to simplify the figure.
Section 0 is an optional line for comments, the # symbol must be placed in column 1 and the variable description is placed in column 2. The character variable can be up to 199 characters. Line 0 can be repeated multiple times.

Section 1 includes the basic stream network description, where the value in parentheses was the value used in the simulations in this thesis:

- 1a. NSTRM (98) – the number of stream reaches (finite difference cells) in the model. This value also specifies the number of entries in line 2 on Figure A.1. NSTRM is positive when unsaturated flow is ignored and negative when unsaturated flow is simulated.
- 1b. NSS (3) – the number of stream segments, which each consist of one or more reaches that share stream properties (i.e. have the same bed slope gradient). If any stream property changes a new segment must be defined.
- 1c. NSFRPAR (0) – number of stream parameters that must be defined. This parameter must be equal to zero when unsaturated flow is simulated.
- 1d. NPARSEG (0) – an integer value that is greater than or equal to the NSFRPAR.
- 1e. CONST (1) – a conversion factor. If ICALC ≠ 1 or 2 (defined later). Set equal to 1 if the units are m³/s; set to 1.486 for ft³/s. If days are the time unit instead of seconds, the factor must also be multiplied by 86,400.
- 1f. DLEAK \(10^{-4}\) – the tolerance level of the stream depth used in computing leakage; \(10^{-4}\) is recommended when units of ft or m are used.
- 1g. ISTCB1 (-1) – a flag indicating where stream-aquifer leakage values are printed; a negative value indicates that these values are printed in the .lst file.
- 1h. ISTCB2 (0) – a flag for writing all information on inflows and outflows to a separate formatted file. 0 indicates that this information is written to the .lst file and not to a separate file

NOTE: additional parameters not used in this example file are needed when unsaturated low is simulated; see Niswonger & Prudic (2010) for instructions on simulating unsaturated flow. Unsaturated flow cannot be simulated with a steady state model.
Section 2 includes stream reach descriptions. Each stream reach (model cell that includes a portion of the stream) needs to be listed here* and match the NSTRM value:

- 2a. KRCH (1) – Layer number of the cell containing the stream reach; streams should always be modeled in Layer 1 (the top model layer)
- 2b. IRCH (26) – Row number of the cell containing the stream reach. For this thesis, the stream was modeled as a straight channel along row 26 of the model domain.
- 2c. JRCH (2-99) – Column number of the cell containing the stream reach. As the stream was modeled as a straight channel along row 26, reaches range in column value from 2-99 where each reach represents a different column.
- 2d. ISEG (1-3) – the number of the stream segment in which the reach is located; this model contains 3 segments, so a given reach can be located in segments 1, 2, or 3. A reach cannot be located in more than one segment for a given model simulation.
- 2e. IREACH (1-36) – sequential number in a stream segment, where the farthest upstream reach is given a number of 1, with values increasing downstream.
- 2f. RCHLEN (40) – the physical length of the stream channel within the stream reach. Since this stream is being modeled as a straight channel along row 26, the physical length of the stream channel is equal to the width of the model cell, which is 40 m. For a meandering stream the channel length within each reach would vary.

*NOTE: not all Line 2 items are listed in Figure A.1; dotted lines indicate where items have been removed to simplify the figure.

Section 3 is a read and print flag for each stress period. Since this model is steady state this line appears only once:

- 3a. ITMP (3) – if ITMP is > 0, ITMP is the number of stream segments defined below using the format of Item 6, which assumes that stream reaches are defined and stream properties are assigned to the upstream and downstream ends of the reach (a linear interpolation is done between these points). In this case 3 stream segments are defined using item 6.
- 3b. IRDFLG (0) – indicates that input data for this stress period will be printed
• 3c. IPTFLG (0) – indicates that streamflow routing results for this stress period will be printed

• 3d. NP (0) – not read when NSFRPAR = 0; 0 used as placeholder

Section 4 includes general stream input data, which assumes that stream properties are defined by stream segment and not by stream reach. The descriptions below pertain to the input data for Segment 1:

• 4a. NSEG (1) – identifies the stream segment for which information is given to calculate inflow, outflow and stream depth. A separate section is provided for each stream segment.

• 4b. ICALC (1) – an integer value indicating the method used to calculate stream depth. A value of 1 indicates that stream depth is calculated using Manning’s equation and assuming a rectangular channel.

• 4c. OUTSEG (2) – indicates the segment immediately downstream of this stream segment that receives tributary inflow from this segment.

• 4d. IUPSEG (0) – set to 0 since flow is not received from a diversion, but simply flows from upstream to downstream. If you are modeling the main stem of a stream or river, make this 0; this value is only needed in a forking system.

• 4e. FLOW (3.88) – streamflow at the upstream end of the segment. For segments 2 and 3 the model will calculate how much water is received from upstream. Only enter a value here for subsequent segments if there is additional inflow (i.e. a tributary).

• 4f. RUNOFF (0) – overland runoff that enters the stream system.

• 4g. ETSW (0) – volumetric rate per unit area of water removed by evapotranspiration from the stream channel (length/time)

• 4h. PPTSW (0) – volumetric rate per unit area of water added by precipitation directly on the stream channel (length/time)

• 4i. ROUGHCH (0.04) – Manning’s roughness coefficient for the channel in this segment; depends on the material of the stream bottom

Section 5 contains stream segment data for the upstream end of each stream segment. The descriptions below pertain to the input data for Segment 1:
• 5a. HCOND1 \((10^{-4})\) – hydraulic conductivity if the streambed (m/s)
• 5b. THICKM1 (0.3) – thickness of the streambed (m)
• 5c. ELEVUP (99.73) – elevation of the streambed (m)
• 5d. WIDTH1 (10) – average width of the stream channel (m)

Section 6 contains stream segment data for the downstream end of each stream segment. The descriptions below pertain to the input data for Segment 1:

• 6a. HCOND2 \((10^{-4})\) – hydraulic conductivity if the streambed (m/s)
• 6b. THICKM2 (0.3) – thickness of the streambed (m)
• 6c. ELEVDN (90.32) – elevation of the streambed (m)
• 6d. WIDTH2 (10) – average width of the stream channel (m)

Note: Sections 5 and 6 must be repeated for each segment

**Updating MODFLOW input files to run the SFR Package**

A generic MODFLOW model without any additional packages requires a series of input files with the same root name (i.e. test): a basic package file (test.bas), a discretization file (test.dis), a layer property flow (LPF) package file (test.lpf), a Name file (test.nam), and a preconditioned conjugate gradient (PCG) package file (test.pcg). The basic package file (.bas) contains the locations of active, inactive, and specified head cells, the head stored in inactive cells, and the initial heads in all cells. The discretization file (.dis) includes the number of rows, columns, and layers, the cell sizes, the presence of quasi-3d confining beds, and the time discretization. The LPF package specifies properties controlling flow between cells. The name file (.nam) contains the names of the input and output files, unit numbers and packages. The PCG package solves the finite difference equations in each step of MODFLOW stress period.

Note: If using Groundwater Vistas to assist with building the model files, you also need an mf2005.nam file, a test.MF2005win32 file, and a mf2005.in file.

To use SFR, the test.nam file needs to be updated by adding a line using the following format:

```
SFR 30 Base_Model_3_1_20.sfr
```

where column 1 = SFR, column 2 = file number of SFR file, column 3 = SFR input file name.
If using Groundwater Vistas, the mf2005.nam file needs to be updated to include the SFR File, by adding a line using the following format:

SFR 30 Base_Model_3_1_20.sfr

where column 1 = SFR, column 2 = file number of SFR file, column 3 = path to SFR input file.
APPENDIX B
GEOLOGICAL HISTORY OF THE SAN LUIS VALLEY

During the Early and Middle Cambrian, the San Luis Valley was located in the Colorado Sag, a lowland basin, which was filled with marine sediments and then was subjected to uplift during the Rocky Mountain Uplift event (Hanna & Harmon, 1980). In the Pennsylvanian and Permian, the eastern portion of the valley was located in the Central Colorado Trough and the western half was part of the Uncompahgre Highlands. Sediment was shed from the highlands into the trough forming fluvial red beds, which make up the present-day deep confined aquifer system. The Laramide Orogeny took place in the Late Cretaceous and formed the Sangre de Cristo Mountains along a series of thrust faults. The Eocene was dominated by erosion and thus uplift due to buoyancy forces (Hanna & Harmon, 1980).

The Rio Grande Rift began in the mid to late Oligocene, forming the beginning of the stratigraphic and structural record visible today. The rift was accompanied by faulting and volcanism, which formed the San Juan Mountains and deposited lava flows, which form the Conejos Formation in the western portion of the valley. The thinning of the valley floor, caused by rifting, led to the formation of three prominent north-south trending features the Monte Vista Graben, the Alamosa Horst, and the Baca Graben (Burroughs, 1981). From the Late Oligocene to the Mid-Miocene, fluvial sediments belonging to the Santa Fe Formation were deposited in the Baca Graben and the Los Pinos formation was deposited in the Monte Vista Graben and the Alamosa Horst. The Sangre de Cristo fault formed along the eastern edge of the valley during the Miocene/Pliocene, causing up to 9,150 m of relief. The Pliocene and Pleistocene were marked by the ending of the Santa Fe Formation deposition and beginning of the Alamosa Formation deposition. The Alamosa formation consists of interbedded fluvial and lacustrine blue/grey clay and silty sand deposits (Hanna & Harmon, 1980). This clay series is typically 3 to 25 m thick and is mostly found in the central and northern part of the valley (Emery et al. 1969).

Due to the presence of both the lacustrine blue-grey clays of the Alamosa Formation and volcanic deposits of the Hinsdale and Conejos Formations the vertical hydraulic conductivity is reduced, limiting deep aquifer recharge (Hanna & Harmon, 1980). However, due to faulting there are three areas in the valley where deep aquifer recharge is known to occur: 1) the headwaters of Saguache and Carnero Creeks due to jointing of the Conejos Formation (Harmon,
1987b), 2) the del Norte High anticline due to heavy fracturing and faulting (Gries & Dyer, 1985), and 3) La Jara Creek near Capulin.
APPENDIX C
RESISTIVITY AND SEISMIC REFRACTION DATA COLLECTION

Resistivity data was collected using an IRIS Syscal Pro Resistivity meter, which transmits a low frequency alternating current (I) to a pair of electrodes (A1 & A2) in the ground “a” distance apart. The voltage (V) is then measured at another pair of electrodes (B1 & B2), also “a” distance apart and “na” distance away from where the current was injected, where n is an integer (Figure C.1). This pattern continues using different pairs of electrodes until all data has been collected. The IRIS was equipped with forty-eight electrodes spaced five meters apart and the data was collected using a dipole-dipole array as shown on Figure C.1.

Hammer seismic refraction data was collected using a Geometrics Geode Seismograph with 25 geophones. Hammer strikes, or shots, were taken 1.5 m before the first geophone, in the middle of the line and 1.5 past the last geophone. After each shot, each geophone produced a seismic trace, where the first arrival or p wave, is marked by a distinctive squiggle. The locations of the first arrivals are traced on the seismograph (Figure C.2A). P waves always take the fastest path, which means that the waves will travel through soil/the unsaturated zone to get to the closest geophones. Eventually, in a two layer system, where the bottom layer has a higher wave velocity, the P waves will refract into and out of the higher velocity zone, allowing for a faster travel time (Figure C.2B; Bullock, 1978).
A two-layer system is created when one layer has a different acoustic impedance value than the other. This could be caused by overburden on top of bedrock or by a water table in alluvial sands and gravels. The acoustic properties of saturated sands and gravels are distinctly different from unsaturated or partially saturated sands and gravels (Gálfi & Palos, 1970).

Figure C.2 Hammer seismic survey setup and data interpretation. Solid purple lines indicate the first arrivals that traveled only through the top layer. The solid yellow lines indicate first arrivals that refracted into and out of the bottom layer. Dashed lines indicate later arrivals.
APPENDIX D
MEASURING VOLUMETRIC FLOW RATE

Two methods were used to measure the volumetric flow rate at along the study area: 1) velocity profiles and 2) dilution gaging. Velocity profiles were taken at SW-1 and SW-2, which involved taking twenty velocity measurements across the river width, generating 19 subsections. Measurements were taken at 2/3 of the river depth. Area was calculated using the midpoint Riemann sum method and discharge (Q) was calculated by multiplying velocity and area (Figure D.1). Pressure transducers were also placed in the stilling wells to track the river stage with the goal of generating a rating curve. However, due to dangerously high flows from spring through mid-summer and the absence of flow midway through August it was not possible to collect enough velocity profiles to develop a rating curve.

To determine changes in volumetric flow rates between diversion points along the Alamosa River, dilution gaging was employed between 0.4 km upstream of Gunbarrel Rd. crossing to the crossing and from 1 km upstream of the Country Road 8 crossing to the crossing. The individual reaches ranged from 37 m to a few hundred m. Dilution gaging involves measuring the dilution of a known conservative tracer, such as NaCl. The method involves injecting a known tracer mass in a nearly instantaneous manner and then measuring concentrations at a fixed location downstream at a short, regular time interval. These

---

Figure D.1 Velocity profile method for calculating river discharge (Q).

\[
Q = \text{Velocity} \times \text{Area}
\]

\[
x = \text{width of subsection (m)}
\]

\[
y = \text{depth of subsection at midpoint (m)}
\]

Area of subsection (AS) = x \times y
Area = \Sigma AS

---
measurements produce a breakthrough curve (Gooseff and McGlynn, 2005; Payne, 2009). Discharge can be calculated from breakthrough-curve data:

\[ Q = \frac{M}{\int_0^T C(t) \, dt} \]

where \( Q \) is discharge (m\(^3\)/s), \( M \) is the mass injected (g), \( C \) is the concentration (g/m\(^3\)), and \( t \) is time (s). The denominator of the right side of the equation is the integration of the area under the breakthrough curve. This is also known as the zeroth moment. The only other variable in the discharge equation is \( M \), which is a known quantity. For dilution gaging to be an effective means of measuring discharge the following assumptions must hold true:

1. All of the injected mass was recovered at the downstream location
2. The tracer is completely mixed when it passes the downstream location

The velocity profiles produced usable data that is detailed in the main section of the report. However, the results of the dilution gaging datasets were inconclusive. As shown in Figure D.2, breakthrough curves were collected, but streamflow calculated from dilution gaging was vastly different from the velocity profiles, with values of 1.71 m\(^3\)/s and 2.38 m\(^3\)/s respectively. This difference in streamflow was most likely caused by the fact that assumption (1) did not hold and all the injected mass was not recovered. The loss in injected mass may have resulted from too long of a reach length, the fact that the volumetric flow rate was too high, or tracer losses through the streambed that did not return to river during our monitoring period.

Figure D.2. Dilution gaging breakthrough curve.