NUMERICAL MODELING OF WATER FLUX INTERACTION BETWEEN BRAZOS RIVER ALLUVIUM AQUIFER AND BRAZOS RIVER: TESTING OF ALTERNATIVE CONCEPTUAL MODELS

A Thesis

by

TAYYAB MEHMOOD

Submitted to the Office of Graduate and Professional Studies of
Texas A&M University
in partial fulfillment of the requirements for the degree of

MASTER OF SCIENCE

Chair of Committee,	Gretchen R. Miller
Committee Members,	Peter Knappett
	Huilin Gao
Head of Department,	Robin Autenrieth

August 2018

Major Subject: Civil Engineering

ABSTRACT

Quantitative characterization of the dynamics of water exchange fluxes between rivers and aquifers is necessary for water resources management, water quality, environment and ecology of the river-aquifer systems. The main uncertain factors for predicting river–aquifer exchange fluxes are aquifer and riverbed properties. In this study, we characterize the flux exchange dynamics between Brazos River Alluvium Aquifer and Brazos River, TX, USA, using alternative conceptual models. Six alternative conceptual models for the connection between the river and the aquifer, having varying aquifer lithology and river incision levels and incorporating processes such as river bed clogging and seepage face flow, are numerically modeled in HYDRUS 2D using small-scale, high-resolution transects across the river. Modeled results are tested against observed heads in three wells and finally a best-fit conceptual model is used to quantify river-aquifer flux exchange dynamics. Additionally we focused on how factors such as aquifer lithology, river channel incision, water table conditions, seepage face boundaries, and low-conductivity river-bed effect hydraulic head distribution and the corresponding flux exchange dynamics. Our results demonstrate that only a small portion of the aquifer close to the river channel is well-connected with the river and a major portion of the aquifer is disconnected. The proposed conceptual model predicts a) much frequent flux reversals (changes between gaining and losing conditions) and b) much smaller amount of recharge and discharges compared to that of the conceptual model which has been assumed by earlier studies; a reduction of 151% in recharge and 116% in discharges. These results suggest that the magnitude and dynamics of water flux exchange between the river and the aquifer are independent of the hydraulic gradients in the wider disconnected aquifer and are determined by the hydraulic gradients in the connected aquifer close to the river. The results

also demonstrate that river-aquifer flux exchange is sensitive to aquifer lithology, river incision depth, and river-bed clogging. While different settings of aquifer lithology and river incision can produce very similar heads in the wider aquifer, the hydraulic head distribution close to the river and hence the river-aquifer flux exchange varies quite drastically from model to model. River-bed clogging decreases the magnitude of fluxes and effects hydraulic head in the aquifer, especially in the vicinity of the river channel, depending upon the gaining and losing river conditions. Furthermore, seepage face flow could be of the same order as that of flows through river-bed depending upon aquifer lithology and corresponding river incision depth.

ACKNOWLEDGEMENTS

I would like to thank my committee chair, Dr. Gretchen Miller, and my committee members Dr. Peter Knappett and Dr. Huilin Gao, for their guidance, support, and motivation throughout the course of this research.

I also wish to extend thanks to my friends and peers in the Miller Geoecohydrology Research Group for their encouragement throughout my research work.

To my parents, Tahir Mehmood and Abida Shamim, thank you for your special late-night prayers and guidance throughout.

CONTRIBUTORS AND FUNDING SOURCES

This work was supervised by a thesis committee consisting of Dr. Gretchen Miller (advisor) and Dr. Huilin Gao of the Department of Civil Engineering and Dr. Peter Knappett of the Department of Geology and Geophysics.

Portions of this research were conducted with the computing resources provided by Dr. Gretchen Miller's research group Geoecohydrology at Texas A&M University. Most of the river stage and hydraulic head data used in the study was provided by Dr. Peter Knappett of the Department of Geology and Geophysics at Texas A & M University. Stream discharge measurements were conducted by Kimberly Rhodes of the Water Management and Hydrological Science program at Texas A & M University. Saturated and Unsaturated soil hydraulic parameters used in the study were estimated by Jingqiu Zhang of the Department of Civil Engineering.

All other work for the thesis was completed by the student, under the advisement of Dr. Gretchen Miller of the Department of Civil Engineering.

This work was funded in part by the National Science Foundation – CAREER Grant (CBET 1351558). The contents of this thesis are solely the responsibility of the author and do not necessarily represent the official views of Texas A&M University, the Texas A&M College of Engineering, nor the National Science Foundation.

NOMENCLATURE

K	Saturated Hydraulic Conductivity
θ_r	Residual Water Content
θ_s	Saturated Water Content
α	Inverse of the air-entry value (or bubbling pressure)
n	Pore-size distribution index
l	Pore-connectivity parameter
K_h	Horizontal Component of Saturated Hydraulic Conductivity
K_v	Horizontal Component of Saturated Hydraulic Conductivity
K_{cb}	Hydraulic Conductivity of River Channel Bed
h	Hydraulic Head
q	Specific Flux
Q	Volumetric Flux
RC	River Co-efficient
r^2	Linear Regression Coefficient
RMSE	Root Mean Square Error

NSE	Nash Sutcliffe Efficiency

TABLE OF CONTENTS

Page

LIST OF FIGURES

Page

LIST OF TABLES

CHAPTER I

INTRODUCTION AND LITERATURE REVIEW

1.1 Background

Efficient development and management of groundwater and surface-water resources as single unit has grown important in the face of diminishing freshwater supplies versus increasing population demands - a problem facing governments and water managers around the globe (Wrachien & Fasso 2002; Winter et al., 1998). Consequently, it has become more important to accurately characterize the dynamics of water exchange between rivers and aquifers (Osman & Bruen, 2002). This flux not only impacts water management in a basin; it also has impacts biogeochemical and ecological processes that occur at the interface (Bartsch, 2014). A broad range of methods exists to quantify aquifer–river exchange flux (Kalbus et al., 2006); however, the use of physically based numerical models has been one of the most effective strategies, as an increasing number of studies have shown (Fleckenstein, 2006; Bartsch et al., 2014; Eddy-Miller et al., 2009).

Brazos River Alluvium Aquifer (BRAA) is an alluvial aquifer located in Southeastern Texas, USA. It is an important source of freshwater and is predominantly pumped for localized agricultural uses. The aquifer is hydraulically connected to Brazos River along 365 river-km length (Cronin & Wilson, 1969). It is underlain by many major aquifers throughout its length (Figure 1.1). Alluvial aquifers commonly have a high degree of heterogeneity, with hydraulic conductivity values ranging several orders of magnitude (Miall, 1996). The interaction between an alluvial aquifer system and river is influenced by the spatial arrangement of hydrofacies at the interface between river and the underlying aquifer (Woessner, 2000). This in turn also impacts the degree of connectivity within the system i.e. between river and the aquifer. Different flow regimes (floods, droughts etc.) also influences short- and long-term exchange fluxes through different phenomenon, such as river-bed clogging, hydraulic gradient reversals etc. These and many other challenges associated with alluvial aquifers becomes more complex when large systems must be modeled under stressed conditions with different hydrological processes occurring simultaneously.

Being prone to droughts (TWDB, 2017), the state of Texas has developed a robust system to develop, manage and protect its groundwater resources that can be used during low-flow times (O'Rourke, 2000). Texas Water Development Board (TWDB) has thus developed a physically based numerical model for BRAA to facilitate its state and regional water planning and decision making processes (Ewing, 2016). The model covers entire extent of the aquifer and takes into account some hydrological processes occurring within aquifer and between aquifer and its surroundings (see section 1.2.2 for details).

As noted by many in the literature (Rushton, 2006; Fleckenstein, 2006; Sophocleous, 2002, while developing large-scale river-aquifer models, the geometries and properties of rivers and aquifers cannot be represented in detail because of large mesh sizes and computational limitations. Consequently, many small scale processes and river and aquifer characteristics (that will be discussed later) are either oversimplified or neglected in simulations, that otherwise are expected to have significant effects on flux exchange dynamics under certain conditions (see section 1.2.1). Simplifications of actual systems are inherent in numerical models that focus on questions of regional-scale or state-scale water management with complex aquifer and river features (Fleckenstein, 2006). Several studies on small-scale interactions between BRAA and Brazos River were performed at the TAMU Hydrogeological Farm Site (Alden & Munster, 1998; Chakka &

Munster 1997; Munster et al., 1996). These studies were designed to serve specific purposes, such as, measuring transport of pesticides from agricultural activity or seasonal water table fluctuations and have adopted analytical or observation-based approaches rather than numerical models which make their applicability limited to understanding the timing and magnitude of volumetric fluxes between the Brazos River and the BRAA.

It is therefore important to quantify the influence of many small-scale processes that have not been taken into consideration in previous models and to model representative processes that significantly impact the exchange under different modes of aquifer and river conditions; such that these can be effectively incorporated in large-scale regional-level models. Such an assessment will not only help improve our understanding about the BRAA system but will also complement existing large scale models.

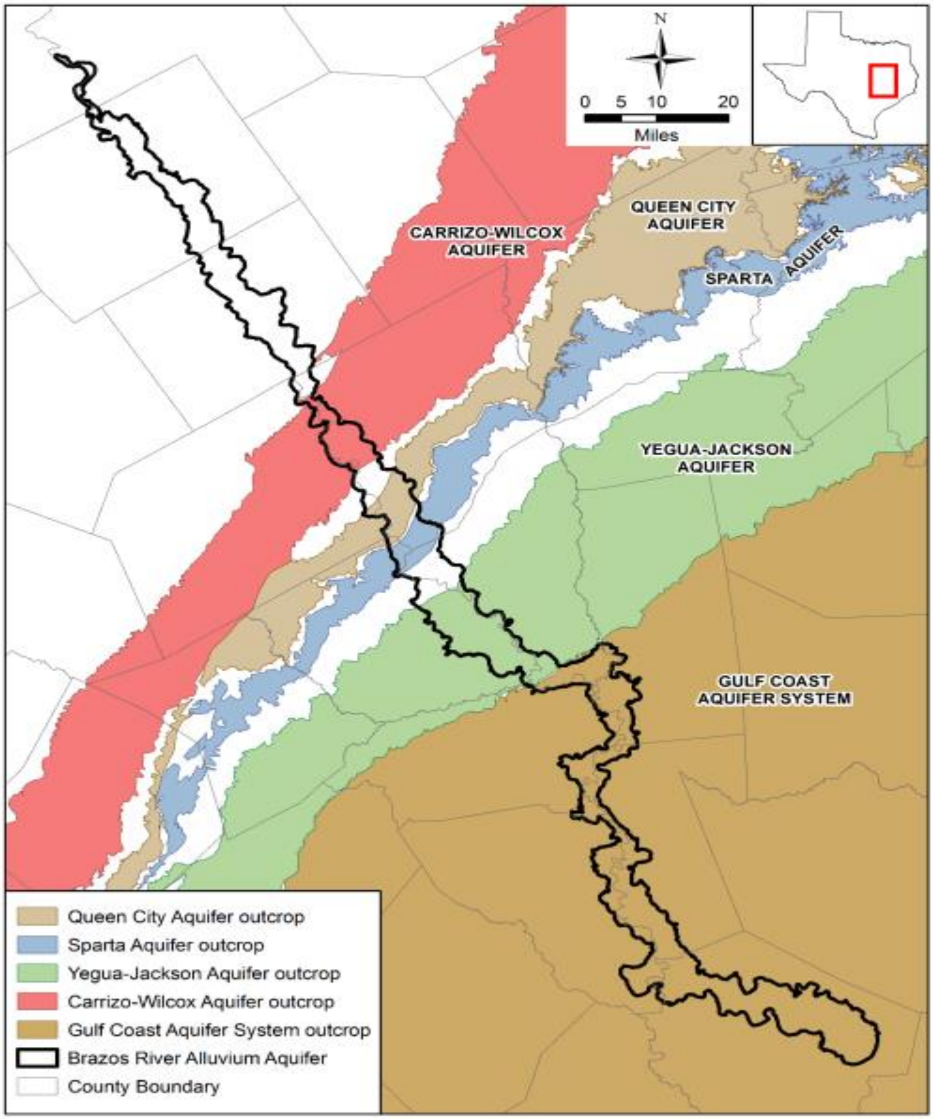


Figure 1.1: BRAA map, along with outcrops of major and minor aquifers (after Ewing, 2016)

1.2 Literature Survey

1.2.1 River-Aquifer Interaction

Water exchange between river and aquifer depends upon a myriad of factors that affect flux exchange differently under different conditions and thus have to be incorporated effectively into numerical models to accurately quantify the amount (Woessner, 2000). Moreover, the processes is spatially and temporally dynamic. A large body of literature exists that highlights the importance and need for accurate characterization and quantification of GW–SW exchange fluxes at increased spatial and temporal resolutions (Bartsch, 2014; Eddy-Miller et al., 2009; Schornberg, 2010; Fleckenstein et al., 2010; Fleckenstein, 2006; Sophocleous, 2006). These factors have been covered in the following sub-sections.

1.2.1.1 Aquifer Geology and Hydrogeology

Alluvium aquifers have high degree of geological heterogeneity (Miall, 1996). Many studies suggests that the distribution and magnitude of hydraulic conductivity (K) within alluvium aquifer is one of the most important factors controlling flux exchange (Fleckenstein et al., 2010; Sophocleous, 2006; Fleckenstein et al., 2006; Rushton, 2006). On the other hand, in regional scale models, aquifers are often represented as comprised of extensive layers with uniform hydrogeological properties (Fleckenstein et al., 2010). To incorporate spatial variability of K values, especially associated with alluvium aquifers, Fleckenstein et al., (2010) suggested the use of Transitional Probability Geo-Statistical Simulation (TPROGS) in a MODFLOW based numerical models. This approach considers modeled aquifer to be comprised of cells with random hydrogeological properties rather than homogenous layers of uniform properties. Fleckenstein et al. (2010) simulated 6 different regimes of K-distribution within the aquifer to assess their impacts on river seepage. River seepage and groundwater heads were found to be highly sensitive to the distribution of K values (Figure 1.2).

Modeling studies have shown that the hydraulic properties of the aquifer adjacent to the river will greatly impact the volumetric flux of water exchanged. Rushton (2006) conducted a model-based experimental study to assess the impact of different hydraulic conductivity zones in relation to water table and river channel incision. He increased K value of the first layer in a fine-grid model---the layer extends from top of aquifer to the bottom of river channel. When K was increased by a factor of 10 this resulted in a 154% increase in magnitude of flux. He further examined the influence of degree of anisotropy in the aquifer. By changing magnitudes of horizontal (K_h) and vertical aquifer (K_v) conductivities, change in magnitude of flux was estimated. Results showed that when $K_v = 0.5K_h$, $K_v = 0.25K_h$ and $K_v = 1.0K_h$; flux (in terms of $K_{aquifer}$ and ($h_{aquifer}$-h_{river})) was 1.06 m^2/d/m, 0.95 m^2/d/m and 1.42 m^2/d/m respectively. The models with vertical anisotropy predicted much less flux exchanged with the river. Thus, Rushton (2006) found that more isotropic, higher K aquifers should have greater flux exchange with rivers.

1.2.1.2 River Channel Geology and Geometry

Hydrogeology, geometry and depth of river-channel incision in the aquifer are among some of the important factors controlling flux exchange. Moreover, in many cases, river channel (both bed and side walls included) hydraulic conductivities (K_{cb}) are not same as that of aquifer materials. Furthermore, K_{cb} changes with flow regime in the river: low-flow conditions cause riverbed clogging resulting in a decrease in K_{cb} values, followed by an increase during high-flows due to scouring (Sophocleous, 2006). In many large-scale models, river channel hydraulic values are assumed to be same as that for the aquifer (K), which is far from reality and could seriously effect simulation results (Rushton, 2006; Osman & Bruen, 2002). Presence or absence of a clogging layer in the river channel causes major differences on the overall flow system in the alluvium (Spalding & Khaleel, 1991).

In large-scale groundwater models, such as MODLFOW, river-aquifer interaction is modeled by using a river-coefficient (RC) which is assumed to be dependent upon (1) hydraulic conductivity (2) geometry of the river channel (McDonald & Harbaugh, 1988).

$$\mathrm{RC} = K_{cb}\frac{LW}{M} \tag{1}$$

Where L, W and M are length, width and thickness of the channel bed/clogging layer. RC as calculated above is then input into equation (2) to estimate flux exchange.

$$q = \mathrm{RC}\,(h_{aquifer}-h_{river}) \tag{2}$$

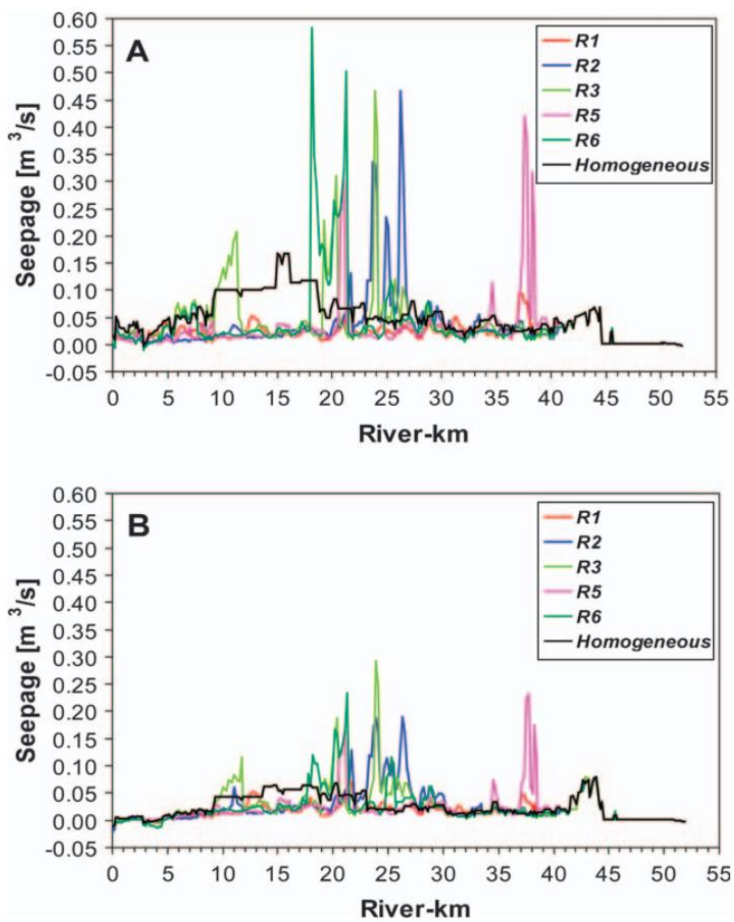


Figure 1.2: Simulated Seepage rates along river channel for six different realizations of the geostatistical models for hydraulic conductivity distribution in the aquifer (after Fleckenstein et al., 2006)

Where $h_{aquifer}$ is the head in the cell representing aquifer below the river channel and h_{river} represents the head in the river cell.

In this approach, flux exchange is assumed to be controlled entirely by the river bed properties. While, it may potentially provide appropriate flux estimates when the direction is from river to aquifer, however, in case of flow direction from aquifer to river, it does not take into account the properties of aquifer and horizontal flows. Furthermore, MODFLOW stream package assumes the river to be comprised of sections each linked to a cell in aquifer model lying below. The river stage is assumed to be same throughout the section and constant during a stress-period - a period of time during which all model stresses remain constant e.g. recharge, groundwater abstraction or discharge to rivers. This leads to the assumption that the river has enough water to maintain flows to aquifer without decrease in river stage.

Rushton (2006) suggested another approach to estimate RC. He calculated exchange flux using a fine-grid small-scale 2D model. He then plotted the relationship between exchange flux and head for a variety of recharge values, under prescribed boundary conditions, Figure 1.3. In the figure 1.3, Q_r and K_h represents flow to the river from aquifer and horizontal hydraulic conductivity, whereas, q represents specific flux (Q_r per unit length of the river).

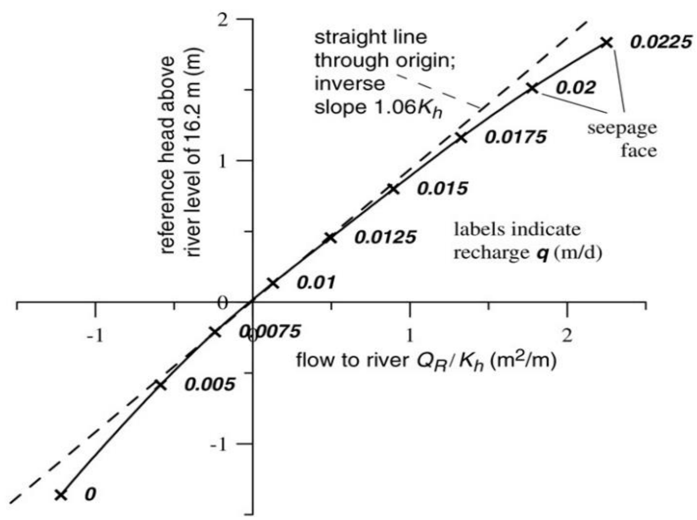


Figure 1.3: Rushton's approach to calculate river coefficients for case (a) K_{cb} = K (after Rushton, 2006)

He then calculated RC as the inverse slope of flux versus head curve at the origin. Figure 1.3 shows his approach. Results for three simulations with K_{cb} being equal to (a) K (b) reduced to 0.15 and (c) reduced to 0.05 m/d reports, although a small increment in flux amount but a significant increment in head distributions in the aquifer which consequently leads to differences in RC's. RC's for three cases were as follows 1.06, 0.9 and 0.77 m^2/d/m respectively. He concluded that RC does not only depends upon river geometry and K_{cb} values, as assumed by MODFLOW, rather it is also dependent upon aquifer K distribution, recharge amounts, head and river stage, and thus cannot be assumed constant.

He further examined the impact of channel geometry by changing channel width and river stage in model. RC decreased from 1.06 to 1.04 m^2/d/m when channel width was decreased from 10 m to 6 m. When river stage was dropped from 1 m to 0.4 m, RC decreased from 1.06 to 0.97 m^2/d/m. He concluded that as long as the river is not substantially wider than the thickness of the aquifer---in which case there is a significant effect on RC---RC is insensitive to channel width and river stage. River bed hydraulic conductivity, on the other hand, has a significant impact on flux.

1.2.1.3 River-Incision, Saturated and Unsaturated Flow Conditions

The degree to which a river channel incised or penetrates through an alluvium has impacts on flux exchange. When a channel fully penetrates, in either losing or gaining river conditions, the flux exchange occurs predominantly through the saturated zone. However, when a river partially penetrates an aquifer, an unsaturated medium could be formed between the channel bottom and deep water table in the aquifer (Winter at al., 1998). Significant flow occurs through the unsaturated zone in such conditions and hence assumptions relying on saturated conditions are not met (Osman & Bruen, 2002). In such cases the use of models designed for variably saturated flow are recommended by many authors.

Osman & Bruen (2002) examined the case of a partially incised river channel with bed clogging using SWMS-2D, a variably saturated code, and compared it with MODFLOW. As explained earlier, in MODLFOW, once the water table falls below river bed, seepage is assumed to be dependent entirely upon river stage and no unsaturated conditions are taken into account. Variably saturated code on the other hand, takes into account unsaturated flows below the river bed. The adequacy of the results from variably saturated code was also tested by a third model (named MOBFLOW) develop on the basis of theory of river-aquifer interaction. In this model, flow through a clogging bed is represented by an equation that assumes fully-saturated, one-dimensional steady-state conditions. This model assumes that a suction head (negative pressure head) is formed outside the clogging layer and controls the exchange, and is calculated using Bouwer's (1969) approach with some improvements (Osman & Bruen, 2002). The results for three models are reproduced in Figure 1.4.

In the graph below, seepage flow from MOBFLOW and SWMS-2D match with each other very well, whereas, MODFLOW seepage rates becomes constant once water table falls below the river bed channel and does not takes into account unsaturated conditions. The maximum seepage flow computed by MOBFLOW, SWMS-2D and MODFLOW are 22050, 20801 and 6750 cm^3/d/cm with serious underestimations by MODFLOW.

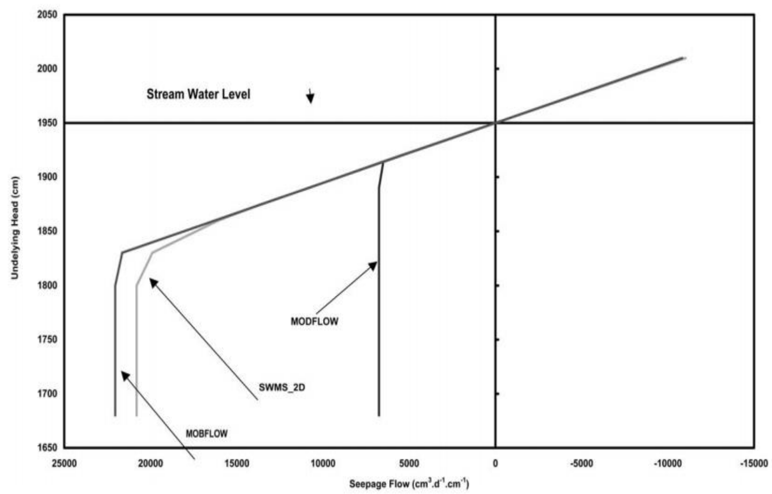


Figure 1.4: Seepage flows from MODFLOW, MOBFLOW and SWMS-2D (after Osman and Bruen, 2002)

1.2.1.4 Seepage Face

Sophocleous (2002) and Franke et al. (2000) have emphasized the importance of seepage face considerations in flux amount estimations. Most large-scale models usually cannot simulate seepage face boundary conditions. Rushton (2006) examined the effect of seepage face by comparing flux amount from his small-scale fine grid model with that of MODFLOW, which does not takes into account seepage face. The results are shown below in Figure 1.4. The differences in head and flux between the two models occur for higher recharge values; when water table is higher than river stage and additional flow occurs through seepage face.

1.2.1.5 Impacts of Underlying Aquifer

Alluvial aquifers are usually shallow and in most cases they are connected to other aquifers to some degree. Along with many properties of the alluvium itself (some of which were discussed in preceding sections), hydrogeology and water table in bottom formations are the two most important factors that influence inter-aquifer exchange. Woessner (2000) have described that usually two aquifers are separated by a low-permeability material layer and water table in the underlying formation remains in confined conditions and hence has impacts on upward or downward leaking. Rushton (2000) in his numerical model tested the influence of hydraulic conductivity (K_b) and head (h_b) of bottom aquifer on river-aquifer exchange in top alluvium. He found out that flux amount increases with increase in K_b. Changing h_b impacted the exchange in a rather different manner. With increasing h_b, aquifer to river flows were increased but river to aquifer flow were decreased: vice versa for decreasing h_b.

1.2.1.6 Effect of River Flow Regimes

The dynamics of river-aquifer exchange is also effected by rainfall; through subsequent runoff flows to rivers and recharges to aquifers. Precipitation strongly effects the temporal exchange pattern. Bartsch et al., (2014) has examined the effect of short-duration and high-intensity extreme rainfall events on the dynamics of exchange fluxes using small-scale fine-grid numerical model and observed data. The river channel was surveyed before and after the event and resulted river channel profile indicates a maximum increment of 30 cm in depth of channel because of erosion caused by rushing water. Two models reflecting the channel conditions before and after the event were used to analyze the effect of river-bed clogging on fluxes.

The results show that flux exchange reversed rapidly, within hours after the rainfall event from losing river conditions to gaining conditions. Under pre-event conditions, the downward fluxes were evident nearly to depth of 0.7 m below the bed, below which the flow field continued to flow in the direction of outlet point, lying below the river to the south-east end of the modeled domain, undisturbed by the exchange. A small rainfall event before the extreme event, temporarily increased the river stage further above the water table and vertical flux exchange deepened (> 0.7 m). A sudden reversal in flux direction occurred upon the arrival of heavy rainfall that followed the small rainfall event. Around this time, vertical exchange flux to river was confined between the depths of 0 and 0.3 m. As small events continued to follow, the recharge to groundwater increased and heads in the aquifer steadily rose. However, the river stage quickly reverted to low levels soon after the events, resulting in maximum flux from aquifer to river during this period. The flows from aquifer to river continued to dwindle until further small rainfall event caused short-term increase in river stage resulting in short-term sporadic flow reversals. The results are reproduced in Figure 1.6 below. The results demonstrate that highly variable hydrologic conditions (intense monsoonal precipitation events) within the monsoon season result in high temporal and spatial variability in river-aquifer exchange fluxes. The results were different than found by an earlier study (Barlow & Coupe, 2009). Barlow and Coupe (2009) investigated the flux reversal in Bogue Phalia River in Florida caused by high-flow events i.e. high river stages. They demonstrated that exchange through the riverbed and a critical stage for flow reversals can be determined by using heat as a tracer even with little available hydraulic head data. They determined one critical river stage after which flux reversals occur in the system. Bartsch et al., (2013) however found out that the reversal of fluxes is rather dynamic and can occur at different river stages depending upon the groundwater flow field. They have argued that heat tracer alone is insufficient and hydraulic head data at sufficient temporal resolution is very important in flux exchange assessments.

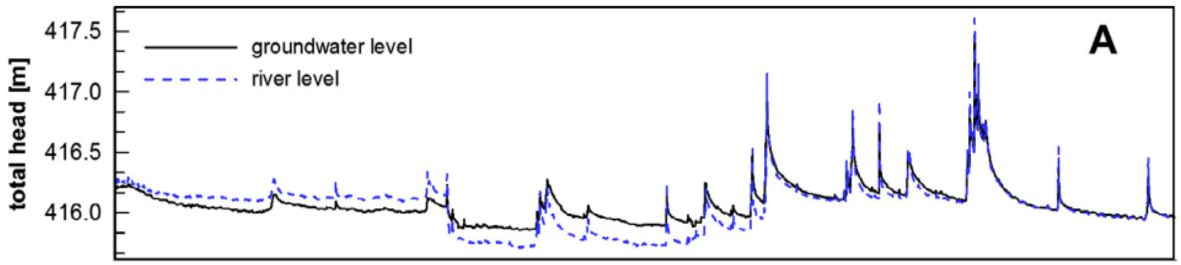


Figure 1.5: Temporal variation in river-aquifer exchange in response to monsoon rainfall (after Barstch et al., 2014)

1.2.2 Previous Numerical Models for BRAA

Texas Water Development Board (TWDB) has developed a 3D numerical model, Groundwater Availability Model (GAM), for the entire alluvium using MODFLOW-USG (Ewing, 2016). The purpose of the model is to facilitate state groundwater development and management. The model is capable of simulating saturated flows and solves the general groundwater equation in 3D by finite difference approximation scheme (Harbaugh, 2005). The model consists of three layers having uniform hydrogeological properties, and the model grid is composed of square grid cells ranging from 1/8-mile to 1-mile in size. A conceptual model adopted in the GAM has been reproduced in Figure 1.7. It shows the processes taken into consideration during simulations.

A groundwater model for a central portion of the BRAA in Milam, Robertson, Burleson, and Brazos counties was prepared by O'Rourke (2006) for TWDB. The purpose of this model was to evaluate a proposed project where the BRAAwould be used to store excess water during high flows of the Brazos and Little Brazos rivers for later use during relatively low-flow periods. The model used a single layer which had uniform hydraulic properties and a 500-foot grid size. Upward vertical leakage from underlying aquifers was acknowledged as a possibility by the author, but the base of the aquifer was modeled as a no-flow boundary due to the lack of reliable data to account for this process.

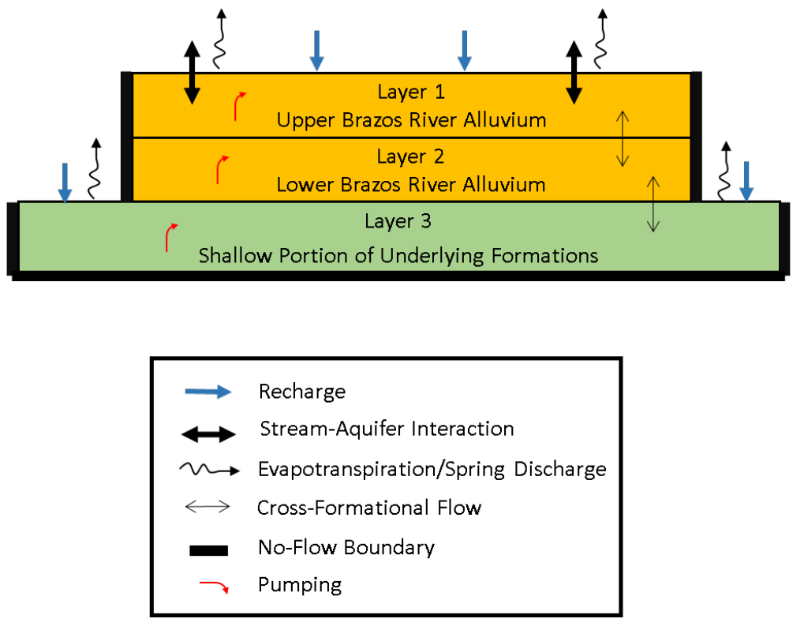


Figure 1.6: Conceptual model for BRAA as adopted in GAM (after Ewing, 2016)

1.2.3 Geological and Hydrogeological Studies for BRAA

The earliest detailed geological study was conducted by Cronin and Wilson (1969) covering geology and hydrogeological aspects of the aquifer. Using wells and bore-logs record, the study provides a detailed description on: lithology and thickness of the aquifer; relationship between the aquifer and underlying formations; water table conditions and movement of water in the aquifer; recharge conditions; hydraulic properties of the aquifer and water quality conditions. The study reports great heterogeneity in aquifer geology, and in general defines a dominating upward fining pattern. Later a study by USGS and TWDB (Shah et al., 2002) employed electromagnetic and electric resistivity methods to assess subsurface geology at a specific location. Along with upward fining of material, they also found out lateral distribution of low and high conductivity materials around the river channel, with most conductive portion of the aquifer separated from the river channel by lower conductivity fine material. The general lithology of the aquifer can be described as follows: gravel at the bottom, to coarse sand & sand in the middle, to clay at the top. Laboratory and field tests were conducted by Cronin & Wilson (1969) from 70 wells to determine aquifer properties over its entire length. The terrace alluvium bordering the floodplain alluvium, where connected hydraulically, yields water to the floodplain alluvium. But hydraulic properties, such as permeability, of the terrace formation is rather low, indicating no significant exchange between them. Therefore, the terrace alluvium is not included in lateral extent of the aquifer. A representative simplified conceptual model for the aquifer is shown in figure 1.8.

Hydrogeological properties of the aquifer sediments are available from earlier works, such as Cronin and Wilson (1969). The values for coefficient of permeability range from 0.001 gpd/ft^2 for samples composed mainly of clay, to as high as 3400 gpd/ft^2 for samples composed of gravels. Also the values change with the depth of sample and proportion of each material in the mix. Based on the information from paper, Table 1.1 summarizes porosity and coefficient of permeability values for different depths and material mix. Studies conducted by Alden and Munster (1997a), Wrobleski (1996), Hvorslev (1951), Bouwer (1989) and Alden & Munster (1997b) at TAMU farm site have used slug tests, long and short duration pump tests and flow sensors to estimate saturated hydraulic conductivities (K). A summary of these values at two spatial locations within TAMU farm site is shown in Table 1.

Table 1.1: Summary of Hydrogeological properties for the entire BRAA (after Cronin & Wilson 1969)

Sampling Site	Dominant Material	Depth Sampled	Porosity (%)	Specific Yield (%)	Coefficient of Permeability (gpd/ft^2)
1	Clay	11 to 12	---	---	0.006
2	Sand and Clay mix	38 to 39	27.3	23.1	2.000
3	Clay	19 to 20	38.5		0.002
4	Medium to Coarse Sand	31 to 32	36.5	35.4	800.000
5	Silt	22 to 23	43.5	27.9	15.000
6	Sand and Clay mix	36 to 37	34.8	31.3	150.000
7	Sand and Clay mix	20.5 to 21	24.7	16.9	290.000
8	Silt and Clay	21 to 22	48.5	4.4	0.800
9	Silt	42 to 43	35.4	31	29.000
10	Gravel	32.5 to 32.75	40.1	26.6	3400.000
11	Silt	27 to 28	35.8	28.3	9.000
12	Clay	46	59.5	18.7	0.001
13	Gravel	River Bank	28.6	24.8	2600.000
14	Gravel	River Bank	32.3	21	14000.000
15	Gravel	River Bank	34.7	22.3	18000.000

Table 1.2: Summary of K values as found in literature

TAMU Farm Location	Study	Method Used	Average Depth (m)	Average K (m/day)
1	Alden and Munster (1997b)	Flow Sensor	13.7	28.9
	Wroblesi (1996)	Pump Test	14.9	60.6
	Bouwer (1989)/Alden and Munster (1997a)	Slug Test	14.9	19
	Hvorslev (1951)/Alden and Munster (1997a)	Slug Test	14.9	32.3
2	Alden and Munster (1997b)	Flow Sensor	18.3	16.5
	Wroblesi (1996)	Pump Test	18.8	58.2
	Bouwer (1989)/Alden and Munster (1997a)	Slug Test	18.8	3.2
	Hvorslev (1951)/Alden and Munster (1997a)	Slug Test	18.8	3.6

Texas A&M University (TAMU) under its Texas Water Observatory (TWO) program has developed a hydrogeological field site covering 21 acres of BRAA to study various aspects of hydrologic cycle, Figure 1.9 (for details about the site see Wrobleski, 1996).

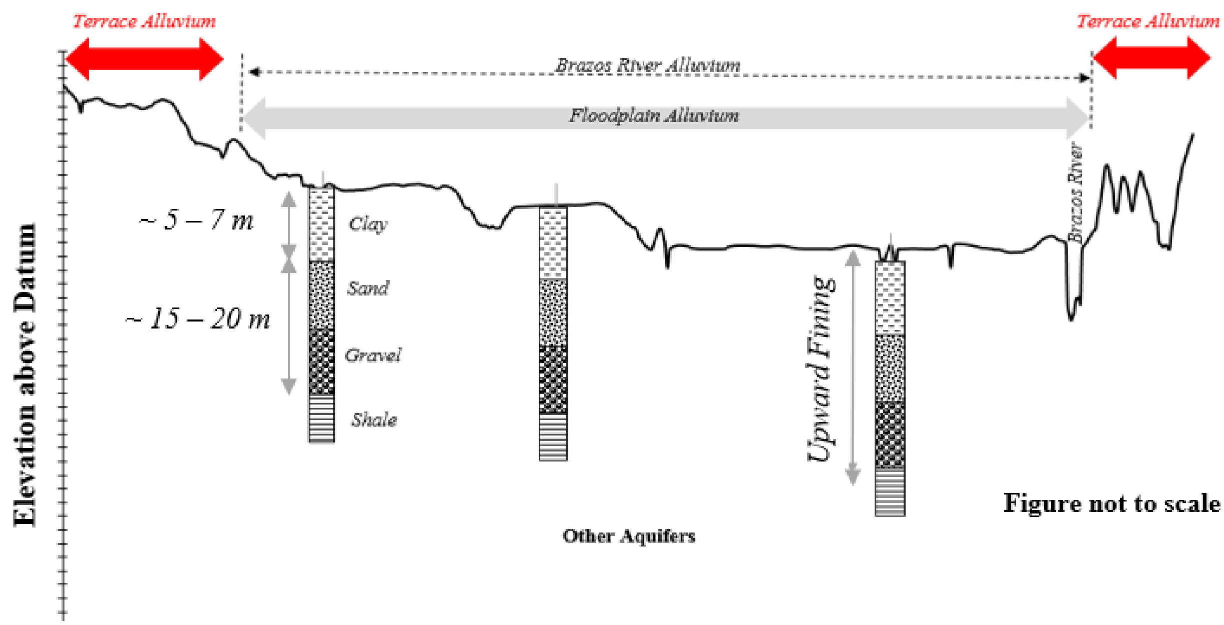


Figure 1.7: Idealized conceptual model for general BRAA geology at the study site

Notable work has been done by researchers at TAMU in order to better understand the river-aquifer relation at different scales (Alden & Munster, 1998; Chakka & Munster, 1997; Munster et al., 1996;; Proffitt, 2015; Rhodes et al., 2017). Some of these studies verified and improved the earlier estimates of hydrogeological properties as proposed by Cronin & Wilson (1969).

Cronin & Wilson, (1969) suggests that principal recharge to the aquifer occurs through rainfall on the aquifer top which infiltrates to the water table sometimes alongside runoff from upstream catchments. On the other hand, Alden & Munster (1997) suggests that the principal source of recharge within the Texas A&M farm site is streamflow from Brazos river, that discharges into the aquifer during high river stage. Water table is mostly in unconfined conditions (∽ 2.5 m to 3 m below the bottom of top clay layer); however, locally confined conditions are also possible as the clay layer varies in thickness and also because of the presence of clay lenses.

1.3 Study Objectives

The high-degree of geological heterogeneity and hydraulic-connection with the Brazos River associated with BRAA has led to varied hypotheses about its groundwater flow system, specifically about the connection between the river and the aquifer. The investigation by Cronin & Wilson (1969) suggested a very well-connected system, with the river incising in the middle of a conductive, sand and gravel portion of the aquifer. Wrobleski (1996), Alden & Munster (1998), Chakka & Munster (1997) and Munster et al., (1996) based their assessment of river-aquifer interaction and contaminant transport using the same conceptual model. Later a study by USGS and TWDB (Shah et al., 2002) employed electromagnetic and electric resistivity methods to assess subsurface geology at specific locations. Recently, Rhodes et al., (2017) carried out an extensive field-based investigation to understand interactions at high-frequency over an 8-month period along a 24 Km river reach. By measuring subsurface fluxes and identifying their sources, they conclude that only a small portion of the aquifer in the vicinity of the river is well-connected with the river, whereas the broader alluvium is disconnected.

The main objective of this study is to test alternative conceptual models, as informed by above cited studies and subjective judgement of the site, to narrow down to a representative conceptual model that will quantify the flux exchange between the river and the aquifer. To this end, six alternative conceptual models with varying settings of aquifer lithology and river incision, incorporating processes such as river-bed clogging and seepage face flow were numerically modeled using small-scale fine grid 2D-transects to the river. Modeled results were tested against observed water heads in three wells and finally a best-fit conceptual model was used to quantify river-aquifer flux and off-site flux into the aquifer. This approach provides a systematic way to narrow down important factors that controls river-aquifer interaction at the site and should result in a parsimonious model. This study will not only deepen our understanding about flux exchange dynamics of the natural system but will also result in a range of guidelines for large-scale modeling and management endeavors.

1.4 Study Area and Site

The Brazos River begins in Stonewall County, Texas, at the confluence of the Salt and Double Mountain Forks. It meanders southeast through Texas for 1,352 river-km (TPWD, 1974) before it discharges into the Gulf of Mexico. The BRAA consists of floodplain alluvium spread around the Brazos River for 563 Km (George, 2011), starting from Bosque County in the northwest to Fort Bend County in the southeast Texas. Studies report great heterogeneity throughout the aquifer, and in general is comprised of an upward fining pattern from gravel, to coarse sand & sand, to clay (TCEQ, 2015; Cronin & Wilson, 1969). Thickness of the aquifer ranges from negligible to 51 m, with an average of 15-24 m (Shah et al., 2007). The water table is mostly under unconfined conditions, however, due to overlying clay layer, local confined conditions are also possible. Throughout its length, the aquifer is underlain by many major aquifers which dip in the same direction as the river, but strike laterally (Ewing, 2016). Low permeability terrace alluvium bordering the floodplain alluvium, marks the lateral extent of the aquifer (Cronin & Wilson, 1969). The aquifer primarily discharges to the Brazos River and is also pumped for irrigation uses (Ewing, 2016; O'Rourke, 2006). The potential sources of recharge includes rainfall, the Brazos River, and fluxes from underlying major aquifers (Chakka & Munster, 1997; O'Rourke, 2006; Turco et al., 2007; Wong, 2012)

The study area (Figure 1.9) is a small portion of the aquifer located 0.5 Km downstream of the intersection of the river with State Highway-60 (SH60) and lies in the Texas A&M University Hydrogeological Farm Site (TAMU Site) located in the Burleson County (longitude -96° 25' 24.98", latitude 30° 32' 11.39"). Within the extent of TAMU site, upward fining is evident with the thickness of top clay layer averaging around ∽ 6 m above the ∽ 15 m thick sand and gravel underlaid by the Yegua-Jackson aquifer. The shallow portion of Yegua-Jackson is a shale layer that separates it from the bottom of BRAA and acts as an impermeable bed-rock. The water table is mostly in unconfined conditions lying ∽ 10 meter below natural ground (Munster et al, 1996). Alden & Munster (1997) suggests that the principal source of recharge to the aquifer at the site is river flow that discharges into the aquifer during high river stage.

CHAPTER II

METHODOLOGY AND MODELING

2.1 Introduction

Sustainable development and management of groundwater and surface-water resources as single unit has grown important in the face of diminishing freshwater supplies versus increasing population demands (Wrachien & Fasso, 2002; Winter et al., 1998). Information on water exchange between rivers and aquifers is needed for fair allocation of water rights (Morel-Seytoux et al., 2015) and for protection from environmental and ecological threats to water resources (Smith et al 2009; Brunke & Gonser, 1997). To address these concerns, we must first quantitatively characterize the dynamics of water exchange between rivers and aquifers at increased spatial and temporal resolutions (Bartsch et al., 2014; Osman & Bruen, 2002; Eddy-Miller et al., 2009; Schornberg et al., 2010). A broad range of methods exist to assess river-aquifer flux exchange (Kalbus et al., 2006); however, the use of physically based numerical models has been one of the most effective strategies, as an increasing number of studies have shown (Fleckenstein, 2006; Bartsch et al., 2014; Eddy-Miller et al., 2009). Accurate quantification and characterization of river-aquifer flux exchange require that conceptual and numerical models of river-aquifer systems take into account important factors that can have wide-ranging effects under varied settings (Woessner, 2000).

Alluvial aquifers have high degree of geological heterogeneity (Tang et al., 2015; Miall, 1996). Many studies suggest that the lithology, distribution and magnitude of hydraulic conductivity (K) within the alluvial aquifer and in the vicinity of the river channel is among the most important factors controlling the flux (Fleckenstein et al., 2010; Tang et al., 2015; Sophocleous, 2002; Rhodes at al., 2017). Rushton (2006) conducted a modeling study to assess the impact of aquifer lithology on the magnitude of fluxes. Multiple layers with different hydraulic conductivities were simulated under confined and unconfined water table conditions. The study demonstrated that the flux between river and aquifer is sensitive to K values of the individual layers. For example, increasing the hydraulic conductivity of the top layer by 10 times, resulted in 154% increase in the cumulative flux amount.

Conventionally, alluvial aquifers are considered connected to their river systems, with the connection assessed in terms of the water table and river stage only; a water table higher than the river bed implies a well-connected system (Brunner et al., 2009 and 2011; Larkin & Sharp, 1992). However, deposition of fine materials by meandering rivers over time may lead to the depositional patterns surrounding river channel causing disconnection with broader alluvial aquifer (Rhodes et al., 2017; Jung et al., 2015). In such instances, the groundwater flow field as assessed by the piezometeric surface in the aquifer far from the river channel may not represent flow field in the vicinity of the river channel and hence may mislead flux exchange dynamics, and especially flux reversal patterns. Furthermore, thicknesses of layers vary across the aquifer which in turn changes the water table conditions, for example from confined to unconfined or partially confined conditions, which may in turn effect the groundwater flow field across the aquifer.

River channel morphology and feedbacks between open-channel and porous media hydraulic properties, are also important factors that determine the connectivity between rivers and aquifers. In many cases, river bed hydraulic conductivities (K_{cb}) are not the same as that of aquifer materials and can change based on the river's current flow regime: low-flow condition causes riverbed clogging resulting in decreased K_{cb} values, followed by an increase in K_{cb} values following high-flow events causing scouring (Sophocleous, 2006; Pavelic et al., 2011). The presence or absence of a clogging layer in the river channel causes major differences on the overall flow system in the alluvium (Spalding & Khaleel, 1991). Pholkern et al., (2015) showed that low energy flows, especially in rivers with high turbidity, can decrease river bed conductivities by a factor of 100. Increases in channel flow velocity only caused partial removal of the clogging and initial K_{cb} values were not recovered. The depth to which a river channel is incised into the underlying alluvium also impacts river-aquifer flux exchange through the formation of saturated and unsaturated zones. When a channel fully penetrates, in either case of losing or gaining river conditions, the flux exchange occurs predominantly through the saturated zone. When the river bed is deeply incised relative to the water table, additional flow occurs through seepage face formed along the river banks (Sophocleous, 2002; Franke et al., 2000). In contrast, when a river partially penetrates an aquifer, an unsaturated medium could be present between the channel bottom and deep water table in the aquifer (Winter at al., 1998). Significant flow then occurs through the vadose zone, hence saturated assumptions fall apart (Osman & Bruen, 2002). An example of this, is that unlike in a fully connected river-aquifer system, in this “disconnected” system hydraulic gradients within the aquifer will not predict the direction of flow between the river and the aquifer.

Owing to the interplay of above factors, there may be a range of impacts of each factor across various settings, making the system complex. It is therefore imperative that these factors and the processes which they impact are investigated through the formulation of alternative conceptual models of the system under a range of realistic settings. These models must then be tested against field observations before one final paradigm is assumed appropriate to explain the given data (Bredehoeft, 2005). The methodology of constructing alternative conceptual-numerical models and then selecting the best among them, or alternatively rejecting those that are deemed to provide an inadequate fit to the observations has been endorsed and adopted by several authors for different groundwater objectives (Bredehoeft, 2005; Carrera & Neuman, 1986; Sun et al., 1995; Neuman, 2003; Tsai et al., 2003a, b; Knappett et al., 2014). This study adopts a similar modeling strategy to assess the dynamics of water exchange between Brazos River Alluvium Aquifer (BRAA) and Brazos River (BR), Texas USA (Figure 2.1).

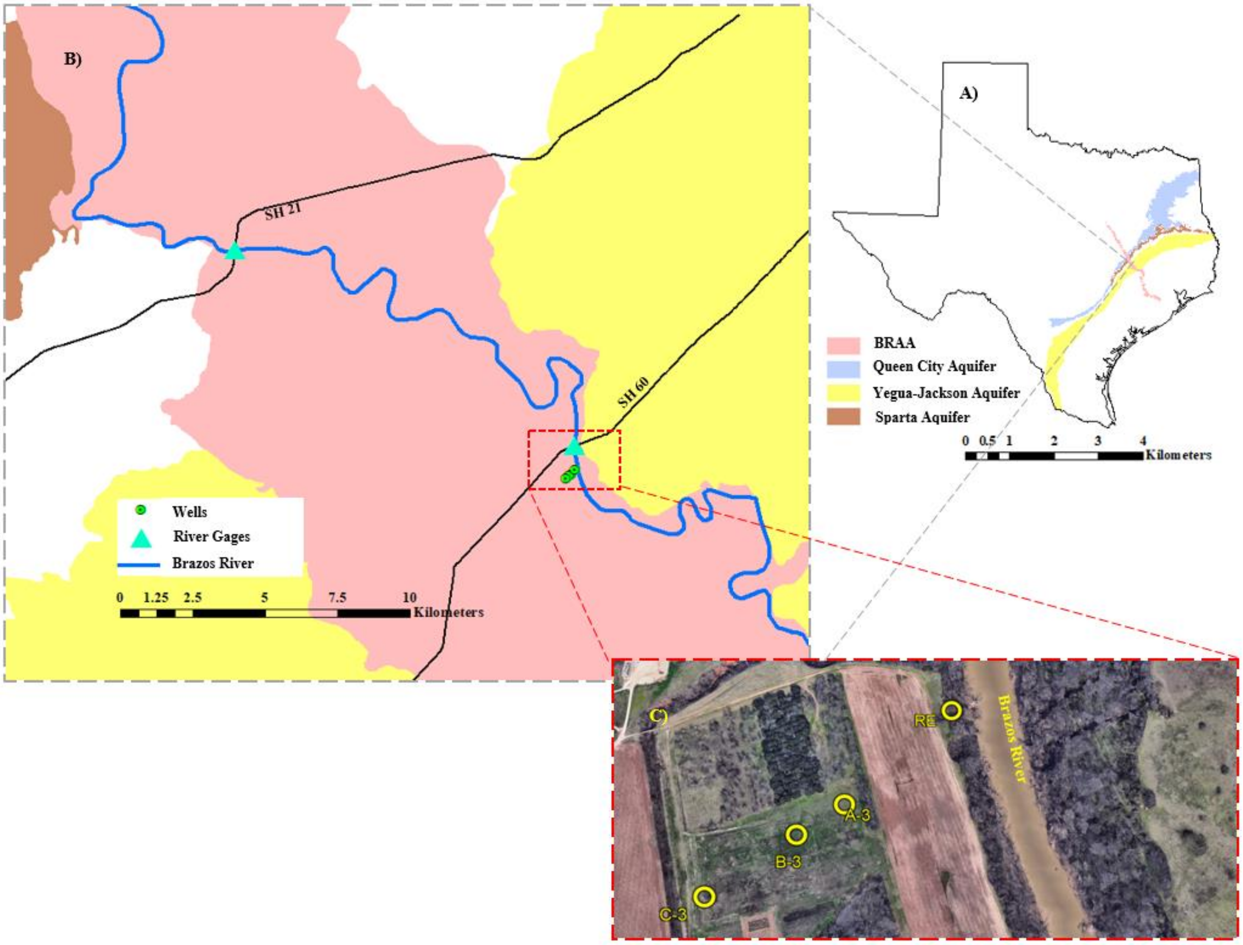


Figure 2.1: Study area map. A) Location of BRAA in Texas, USA, along with underlying major aquifers. B) Location of TAMU Hydrogeological Farm Site, monitoring wells and river gaging station (SH60) within BRAA. C) Transect to the river at TAMU Farm Site used in the study and locations of Well C-3, B-3, A-3 and RE.

The high degree of geological heterogeneity associated with the BRAA and its hydraulic-connection with the Brazos River has led to several hypotheses about its groundwater flow system. Field-based investigative studies conducted at the site suggest different conceptual models for the connection between the river and the aquifer. Cronin & Wilson (1969) suggest a very well-connected system, with the river incising in the middle, most conductive, sand and gravel portion of the aquifer. Wrobleski (1996), Alden & Munster (1998), Chakka & Munster (1997) & Munster et al., (1996) based their assessment of river-aquifer interactions and contaminant transport on the same conceptual model. Later, researchers with the United States Geological Survey (USGS) and Texas Water Development Board (TWDB) (Shat et al., 2007) employed electromagnetic and electric resistivity methods to assess subsurface geology near the study site. Along with upward fining of materials, they also found that the most conductive potion of the aquifer was separated from the river channel by low conductivity fine material.

Recently, Rhodes et al., (2017) carried out an extensive field-based investigation to understand the interaction along a 24-km- river reach. By measuring subsurface fluxes and identifying their sources, they concluded that only a small portion of the aquifer, bank storage zone, is well-connected to the river, whereas the broader alluvium is disconnected. Prior to measuring water fluxes, Rhodes et al. (2017) performed a preliminary assessment of the connectivity of the aquifer-river system by constructing geologic cross-sections using publically available geologic borehole data from the Texas Water Development Board (TWDB). This analysis suggested a regional, continuous, basal gravel layer overlying the bedrock, incised by the River. The water fluxes, however, defied this conceptual model. One possibility is that the public database is biased towards high K alluvium creating an impression of higher K sediments than are really typical (Rhodes et al., 2017). If this is true, then models based on public geologic borehole databases, may generally overestimate connectivity of rivers and alluvial aquifers.

The main objectives of this study are: a) to test a range of alternative conceptual models, as informed by above cited studies and subjective judgement of the site, b) determine the most representative model based on fit to observed, high-frequency water levels over multiple seasons, and c) use that model to quantify the flux exchange between the river and the aquifer during normal, drought, and flood conditions. To this end, six alternative conceptual models are numerically modeled in HYDRUS 2D using small-scale, high-resolution transects across the river. The models have varying aquifer lithology and river incision depths and incorporated processes such as river bed clogging and flows through seepage face boundaries. Modeled results are tested against observed water heads in three wells and finally a best-fit conceptual model is used to quantify river-aquifer flux exchange dynamics. Following three types of river-aquifer fluxes were simulated: a) portion of baseflow (from aquifer to river) that occurs across the seepage face boundary, hereto referred as seepage, b) portion of baseflow that occurs across the river-bed, hereto referred as discharge and c) flow from river to aquifer, i.e. losing conditions, hereto referred as recharge. This approach provides a systematic way of determining which factors exert the most control on river-aquifer interactions in this system and which parameters are truly essential for capturing system behavior. This study will help us better understand flux exchange dynamics between the aquifer and the river at the study site and identify important controlling factors and their impacts on the system.

2.2 Materials and Methods

2.2.1 Study Area and Site

The Brazos River begins in Stonewall County, Texas, at the confluence of the Salt and Double Mountain Forks. It meanders southeast through Texas for 1,352 river-km (TPWD, 1974) before it discharges into the Gulf of Mexico (Figure 2a). The BRAA consists of floodplain alluvium distributed around the Brazos River for 563 km (George, 2011), starting from Bosque County in the northwest to Fort Bend County in the southeast Texas. Studies report great heterogeneity throughout the aquifer, and in general defines a dominating upward fining pattern from gravel, to coarse sand & sand, to clay (TCEQ, 2015; Cronin and Wilson, 1969). Thickness of the aquifer ranges from negligible to 51 m, with most of it lying around 15-24 m (Shah et al., 2007). The water table is mostly in unconfined conditions, however, due to overlying clay layer, local confined conditions are also possible. Throughout its length, the aquifer is underlain by the outcrops of five significant aquifers which strike orthogonal to the river and dip towards the Gulf Coast (Ewing, 2016); these are, in order from oldest to youngest, the Queen City, the Yegua-Jackson, and the Carrizo-Wilcox. Low permeability terrace alluvium, bordering the floodplain alluvium, marks lateral extent of the aquifer (Cronin & Wilson, 1969). The aquifer primarily discharges to the Brazos River and is also pumped for irrigation uses (Ewing, 2016; O’Rourke, 2006). The potential sources of recharge include rainfall, the Brazos River, and fluxes from underlying major aquifers (Chakka & Munster, 1997; O'Rourke, 2006; Turco et al., 2007; Wong, 2012)

The study area is a small portion of the aquifer located 0.5 Km downstream of the intersection of the river with State Highway 60 (SH60) and lies in the Texas A&M University Hydrogeological Farm Site (TAMU Farm Site) located in the Burleson County (longitude -96° 25' 24.98", latitude 30° 32' 11.39"). Within the vicinity of the TAMU site, upward fining was recorded by Wrobleski (1996), with a top clay layer averaging around 6 m (a vertisol), followed by a 15-m thick sand and gravel layer, underlain by the Yegua-Jackson aquifer. The top portion of the Yegua-Jackson formation is a shale layer that separates it from the bottom of the BRAA and creates a nearly impermeable barrier. The water table is mostly in unconfined conditions lying below the top clay layer (Wrobleski, 1996), at a depth ∽ 8.5 to 9 meters below the natural ground, however, this can change during prolonged dry or wet periods. Alden and Munster (1997) suggest that the principal source of recharge within the site is river flow discharging into the aquifer during high river stages, but given that the clay is prone to cracking under dry conditions, significant infiltration may be possible.

2.2.2 Data Collection

The piezometric surface in the broader alluvium was measured at 20-minute intervals in four wells (RE, A-3, B-3 and C-3) along a transect across the river. The wells are 30 m, 80 m, 200 m, and 420 m away from the western edge of the river respectively (Figure 2.1 c). These wells, which are 18.3 m deep with 15-cm long screens, were installed in the late 90's, during the initial construction of the field site (Chakka & Munster, 1997; Wrobleski, 1996). Water levels were converted to elevations based on a Real-Time Kinematic GPS survey completed by Rhodes et al (2017). Water level readings were recorded by pressure transducers (Levelogger, Solinst Ltd., Canada) and were corrected for atmospheric pressure variations by using a barometric pressure sensor (Barologger, Solinst Ltd., Canada) located < 10 km from the wells and river gages. Net aquifer discharges to the river between SH21 and SH60 were measured in a separate study (Rhodes et al., 2017) and were used here to help choose the correct model, assuming the site is representative of other parts of the river between SH21 and SH60. To compare the simulations with the field data, measured volumetric discharges (in m^3/s) were converted to specific fluxes ($m^3/hr/m$) by dividing by 24500 m, which is the length of the river reach between SH21 and SH60. This assumes the average specific flux is something that should occur at a given field site. Measured river stages at SH60 are available for first 247 days of the record from Rhodes et al., (2017). The stages were re-constructed for the rest of the record (from April 01, 2016 to February 07, 2017) using hydraulic routing in HEC-RAS 1D (see appendix for details) along with the record from USGS gage number USGS 08108700 located 24.5 km upstream (USGS, 2018).

2.2.3 Modeling Approach

A range of probable conceptual models revolving around different settings of aquifer lithology and the connection between the river and the aquifer were formulated and numerically modeled using small-scale, high-resolution 2D-transects in variably saturated code HYDRUS. In all models, the domain is 26 meter thick and spans 420 m on each side of the 180 m wide river, which has a simplified elliptical channel shape (Figure 2.2) with seepage face boundaries. Well C-3 marks the lateral boundary of the modeled domain. A 50 cm thick layer around the river channel is used to simulate low-conductivity river-bed conditions in applicable models. Since estimates of K_{cb} are not available for the site, based on Pholkern et al., (2015), K values of aquifer materials reduced 100 times are used as K_{cb}. Due to unavailability of the water table data on the east side of the river, aquifer symmetry is assumed around the river channel. Numerically simulated well heads based on each conceptual model are tested against the observed well heads in three wells B-3, A-3 and RE. Visual inspection, linear regression coefficients (r^2), and RMSE (root mean square error) and NSE (Nash Sutcliffe efficiency) statistics (Nash and Sutcliffe, 1970) are used to assess the fit between the modeled and the observed data. Instead of model calibration, a range of alternative conceptual models are tested against observed data in all three wells. The best-fit model is then used to estimate flux exchange between the river and the aquifer. This approach, where a set of small-scale, simple models are tested has several advantages: 1) it involves less parameterization and allows for use of representative K values, 2) the effects of confounding factors, such as recharge to the aquifer from top and inter-aquifer flux exchange from bottom formations, can be excluded; and 3) it provides a higher-level view of the overall behavior of flux exchange between the river and the aquifer.

2.2.3.1 Basis for Alternative Conceptual Models

Six conceptual models (CM) were developed (Figure 2.2), each based on previous studies, pre-existing borehole logs, and/or knowledge of fluvial processes typical in low-energy rivers. The models have different lithological configurations and depths of river incision, but maintained the same material properties (e.g., if a clay layer was present, it was assumed to have the same hydraulic conductivity as the clay layers in the remaining models). In all cases, an impermeable shale layer marks the bottom of the aquifer. When describing these models, the term "well-connected river-aquifer system" implies that the fluctuations in river stage are propagated proportionately in the entire aquifer modeled domain, whereas, a "poorly-connected system" implies that river stage fluctuations are significantly to completely attenuated as they travel into the aquifer. Conceptual models are explained below:

- *CM 01:* Hypothetical homogenous aquifer comprised of high-conductivity sand and gravel material above the bottom shale. Suggestive of a very well-connected river-aquifer system.
- *CM 02:* Hypothetical three-layered aquifer comprised of uniform clay layer on top of sand and gravel. River is assumed to incise within the top clay layer. Suggestive of a poorly connected system.
- *CM 03:* As suggested by Wrobleski (1996) and predominantly assumed. Three-layered aquifer comprised of clay layer on the top of sand and gravel. River incises through the top two layers. Presence of thin, discreet sloping clay lenses close to the river.
- CM 04: Hypothetical case of three layered aquifer comprised of clay top, sand and gravel middle and shale bottom. River incises through to the bottom shale layer.

- *CM 05:* As suggested by Shah et al., (2007). This model assumed a poorly-connected system with homogeneous, low-conductivity material surrounding the river channel.
- *CM 06:* As suggested by Rhodes et al (2017). Only a small portion of the aquifer close to the river is well-connected with the river, whereas the majority of the aquifer is disconnected from the river because of the presence of an abandoned channel separating the two.

Table 2.1: Van Genuchten parameters used in the study

Material	**θ_r (-)**	**θ_s (-)**	**α (1/m)**	**n (-)**	**K_s (m/d)**	**l (-)**
Top Clay	0.108	0.450	0.018	1.250	0.046	0.500
Sand & Gravel	0.053	0.300	0.035	3.177	80.00	0.500
Bottom Shale	0.068	0.380	0.800	1.090	0.0002	0.500

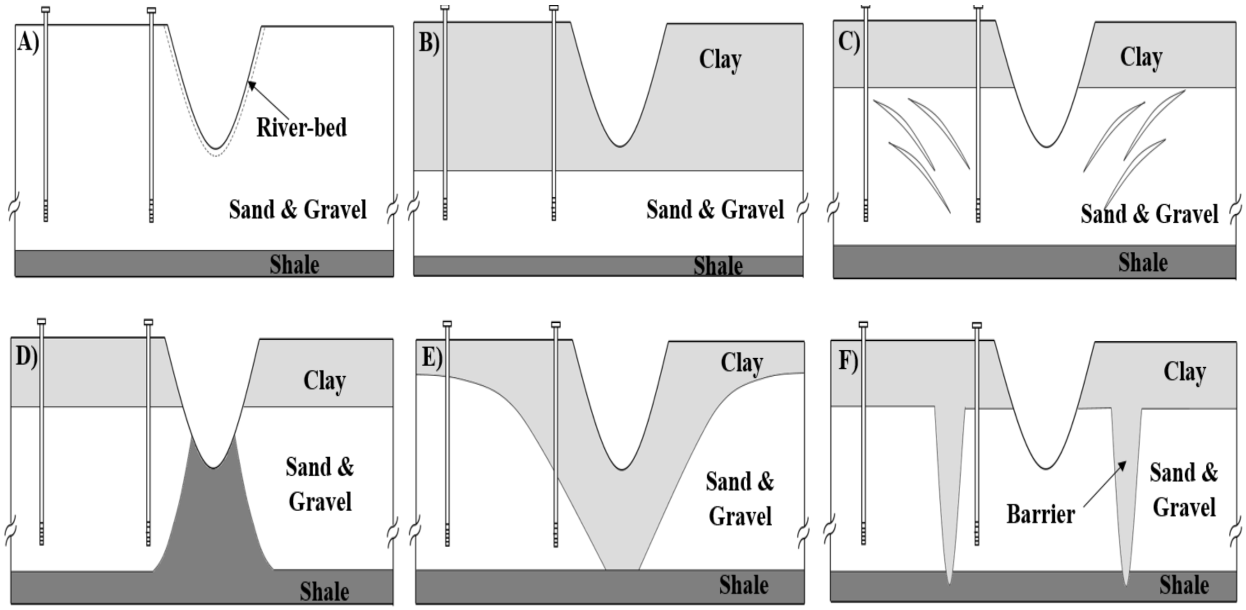


Figure 2.2: Six alternative conceptual models not drawn to scale. All the conceptual models have same dimensions and material properties. A) Conceptual model 1. B) Conceptual model 2. C) Conceptual model 3. D) Conceptual model 4. E) Conceptual model 5. F) Conceptual model 6. A proximal and a distal monitoring well are shown, where the proximal well represents the approximate location of the river edge well.

2.2.3.2 Numerical Models

HYDRUS 2D/3D (version 2.103 PC-Progress, Prague, Czech Republic) was used for numerical modeling. It solves the partial differential form of Richards' equation using a finite element approach to describe transient subsurface flow in variably-saturated porous media (Šimůnek et al., 2011). Well C3 was used to proscribe a variable head (first-type) boundary condition (BC) on either side of the domain. The land surface adjacent to the channel and the bottom of the domain was described as no-flow BC. The continuously measured river stage served as the variable head BC (first-type boundary) for the defined river channel nodes which switches to a seepage face BC when the specified nodal head is negative (Šimůnek et al., 2011). All time-variable input data were implemented as hourly values. To obtain reasonable initial conditions for the transient simulations, first the models were run for a 21 day spin-up period with forcing from the first 21 days of the data record. Hydraulic properties of the materials are adopted from earlier work (Zhang, 2014) and are shown in Table 2.1. The Van Genuchten-Mualem (1980) formulation was used for the unsaturated flow simulations, which includes saturated hydraulic conductivity (K), Residual Water Content (θ_r), Saturated Water Content (θ_s), α is the inverse of the air-entry value (or bubbling pressure), n is a pore-size distribution index, and l is a pore-connectivity parameter.

2.3 Results

2.3.1 Measurements - Groundwater Table, River Stage, and Discharges

Figure 3 shows hydraulic head, river stage and discharges for two separate periods used in the study. From April 28, 2015 to January 29, 2016 (period 1), continuous hourly records of river stage at SH60 and heads for Well C-3 and B-3 are available for 247 days while for well A-3 it lasts for 199 days (Figure 2.3 a). Measured discharges were available for selected eight periods, a total of 102 days (Figure 2.3 b). From April 01, 2016 to February 07, 2017 (period 2), re-constructed river stages at SH 60 and measured hourly records for the newly added well R-E including C-3, B-3 and A-3 are available for 347 days (Figure 2.3 a).

During period 1, the river stage ranged from 67.8 masl (meters above sea level) during historic flood in late-April, to 54.8 masl at the end of drought in late-September. In contrast, the water table in the wider aquifer, at well C-3, only varied slightly from 60.4 masl, in mid-January, to 58.1 masl in late-April. This pattern was also evident during period 2, when the river stage dropped by 11.4 m in contrast to a drop of 3.78 m in well C-3. Well B-3 and A-3 behaved in a similar fashion.

Throughout the record, the piezometric surface in the wider aquifer (C-3, B-3, A-3) sloped towards the river (gaining river conditions), except during high river stages (grey patches in Figure 2.3a) when the hydraulic gradient slightly shifted away from the river (losing river conditions) for a short time period and began to reverse back as the river stage recedes. This shows that the effects of large river changes are attenuated considerably in the wider aquifer. The average hydraulic gradient during gaining conditions is approximately 0.002. In contrast to wider aquifer, piezometric surface in RE well is very responsive to the river stages, as can be seen during period 2, with sudden slope reversals happening over short time periods. This shows that the effects of even small river changes propagates rapidly in the closer aquifer. Discharges varies from a high of 197.0 m^3/day, during the beginning of recession period in mid-June, to almost 0 m^3/day, towards the end of drought in late-September. This pattern was also evident elsewhere when discharges dwindled as the river stage continued to drop (Figure 2.3b).

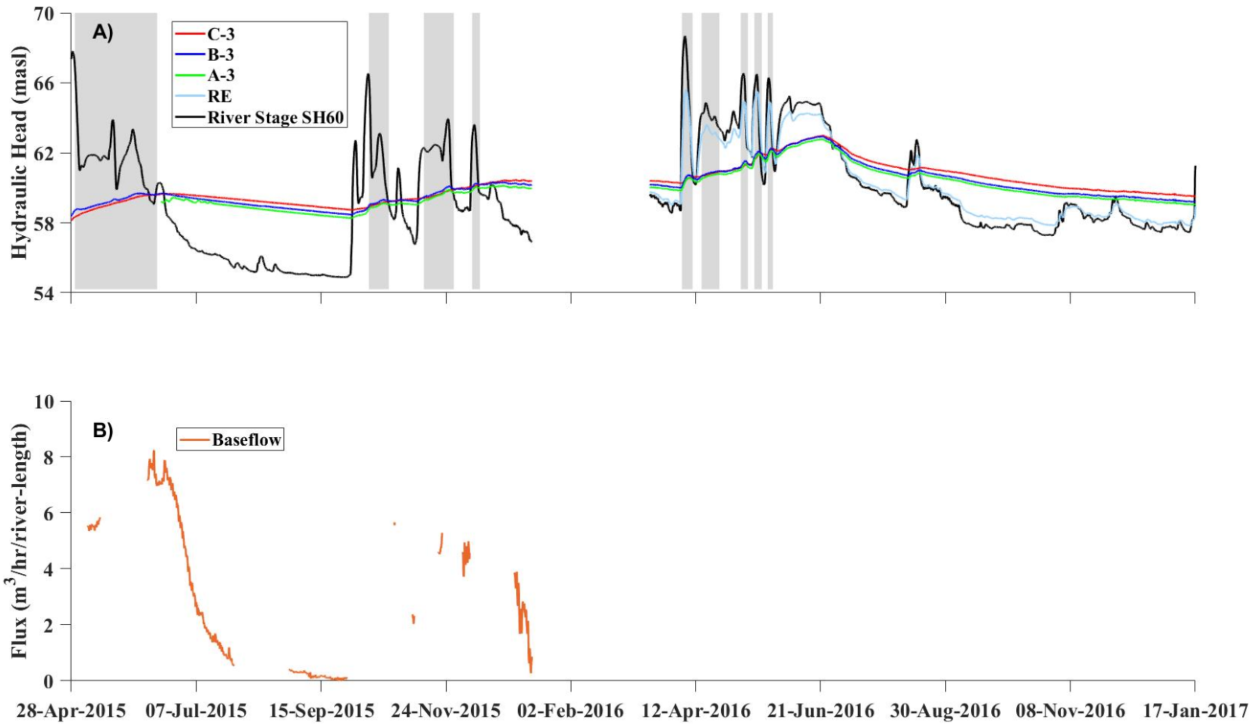


Figure 2.3: Measured data. A) River stage at SH60 and hydraulic heads in wells, C-3, B-3, A-3 and RE, corrected to the same datum. Data is shown in two continuous periods, period 1 from April 28, 2015 to January 29, 2016 and period 2 from April 01, 2016 to February 07, 2017, separated by gap where no data was available. Gray patches show periods when water table in the farther aquifer fall below the river stage B) Net discharges measured in the reach between SH21 and SH60 were available only during period 1. The net discharges values are shown per unit length of the reach between SH21 and SH60 for comparison with 2D modeled fluxes.

2.3.2 Model Results

In the discussion below the results from the numerical models that are conceptually similar have been grouped together for clarity.

2.3.2.1 CM 1, CM 3 and CM 4

The models CM 1, CM 3 and CM 4 all represent a well-connected river-aquifer system but vary by degree. All the three models can explain the piezometric heads near the river (Well RE - Figure 2.4a, c, d), with r^2 = 0.979, 0.975 and 0.981, NSE = 0.970, 0.971 and 0.976 and RMSE =

0.373, 0.365 and 0.335 (Table 2.2) respectively. For Well RE (green line in Figure 2.4), simulated piezometric levels were in close agreement with the observed piezometric levels throughout the simulation period in all the three models. Models CM 3 and CM 4 explain Well RE slightly better than CM 1, as evident by the improved statistics. However, the models fail to explain the water table in the well in the center of the profile (Well B-3), with r^2 = 0.866, 0.887 and 0.884 and NSE = 0.719, 0.776 and 0.791 and RMSE = 0.561, 0.500 and 0.483 respectively. For Well B-3 (blue line in Figure 2.4), the models over-predict the magnitude of fluctuations in the piezometric surface. Although during period 2, the difference between simulated and observed head is smaller as compared to period 1, overall, the models fail to explain the rather attenuated response of aquifer at Well B-3. These simulations suggest that as the river stage fluctuates, proportionate fluctuations propagate in the aquifer throughout the modeled domain, a well-connected system. While this assumption holds true for Well RE, it does not explain Well B-3. Overall, CM 1, CM 3 and CM 4 explains the piezometric surface in the aquifer close to the river but fails as we move further into the aquifer.

The flux between the river and the aquifer (Figure 2.5) alternates from being towards the aquifer when the river stage is higher than the piezometric surface (recharge), to towards the river when the piezometric surface is higher than the river stage (discharge and seepage). During the long recession periods (Figure 2.3 a), from 17th Jun., 2015 to 25th Sep. 2015 and 18th Aug., 2016 to 15th Jan., 2017, the aquifer continued to discharge into the river which reverses abruptly to recharge as the river stage rises above the piezometric surface. Additionally, seepage also occurs whenever the piezometric surface is well above the river stage. Cumulative discharge and recharge are highest in CM 1 (6540 and 2958 m^2 respectively), followed by CM 3 (5069 and 2363 m^2

respectively) and CM 4 (1929 and 1900 m^2 respectively). In contrast, seepage is highest in CM 4 (1936 m^2), followed by CM 3 (288.0 m^2) and CM 1 (171.0 m^2). In general, these models suggests that entire modeled domain exchanges water with the river.

2.3.2.2 CM 2 and CM 5

CM 2 and CM 5 essentially represent a poorly-connected system, but again vary by degree. Contrary to the well-connected conceptual models, both CM 2 and CM 5 can explain the piezometric heads in the in the wider aquifer (Well B-3 - Figure 2.4 b, e), with r^2 = 0.979 and 0.984, NSE = 0.964 and 0.958 and RMSE = 0.198 and 0.216 respectively (Table 2.2). For Well-B-3, the simulated piezometric heads were in close agreement with the observed piezometric heads throughout the simulation period. Model CM 2 better captured the magnitude of fluctuations in Well B-3, hence high NSE and low RMS values, whereas CM 5 slightly under-predicted the fluctuations. However, the two models failed to explain the water table in Well RE, with r^2 = 0.953 and 0.723, NSE = 0.826 and 0.507 and RMSE = 0.899 and 1.415 respectively for CM 2 and CM 5 (Table 2.2). These models suggest that the river stage fluctuations are considerably attenuated in the aquifer throughout the modeled domain, a poorly connected system. While this holds true for Well B-3, it fails to explain the higher magnitude of fluctuations observed at Well RE. For Well RE, the models under-predict the water table fluctuations. Overall, models CM 2 and CM 5 explain the piezometric surface in the wider aquifer but falls apart as we move closer to the river.

The flux exchange between the river and aquifer is very small for CM 2 and negligible for CM 5 (Figure 2.5). The river channel incises completely through the top clay layer in CM 2 and is completely surrounded by low conductivity material through to the bottom of the aquifer in CM 5 producing even decreased flux exchanges. As a result, CM 2 produces higher cumulative discharge, seepage and recharge (1372, 119.0 and 706.0 m^2 respectively) than CM 5 (94.00, 42.00 and 126.0 m^2 respectively).

2.3.2.3 CM 6

Unlike above conceptual models, CM 6 well explains the hydraulic heads in both Well RE and Well B-3 (Figure 2.4 f). The value of r^2, NSE and RMSE for Well RE are 0.984, 0.967 and 0.396, and for Well B-3 are 0.987, 0.972 and 0.177 respectively (Table 2.2). The statistics for individual wells are better than any other conceptual model. This model suggests that fluctuations in the river stage propagate only up to a certain small distance into the aquifer (<80 m). Before they reach Well A-3, located ∽ 200 m from the western river edge, they are attenuated by the low-conductivity barrier. Overall, CM 6 explained the response of both the wider and the closer aquifer and therefore is the only conceptual model that explains the response of entire aquifer. Consequently, the cumulative discharge and recharge in CM 6 (2608 and 1369 m^2 respectively) are greater than that of a poorly-connected system, say CM 2, and are smaller than that of a well-connected system, say CM 1. However, the seepage in CM 6 (32.25 m^2) is smaller than any of the other conceptual models.

Table 2.2: Model evaluation: Results of the statistical measures for each of the six conceptual models for Well B-3 and RE. The slope and intercept p-values are associated with statistical significance of r^2 values. Green and blue highlighted rows represent conceptual models that explain Well RE and Well B-3 respectively.

Conceptual Model	Well	r^2	p-value (Slope & Intercept)	Statistically Significant	NSE	RMSE
CM1	Well B	0.866	< 0.00100	Yes	0.719	0.561
	Well RE	0.979	< 0.00100	Yes	0.970	0.373
CM2	Well B	0.979	< 0.00100	Yes	0.964	0.198
	Well RE	0.953	< 0.00100	Yes	0.826	0.899
CM3	Well B	0.887	< 0.00100	Yes	0.776	0.5
	Well RE	0.975	< 0.00100	Yes	0.971	0.365
CM4	Well B	0.884	< 0.00100	Yes	0.791	0.483
	Well RE	0.981	< 0.00100	Yes	0.976	0.335
CM5	Well B	0.984	< 0.00100	Yes	0.958	0.216
	Well RE	0.723	< 0.00100	No	0.507	1.415
CM6	Well B	0.987	< 0.00100	Yes	0.972	0.177
	Well RE	0.984	<0.00100	Yes	0.967	0.396

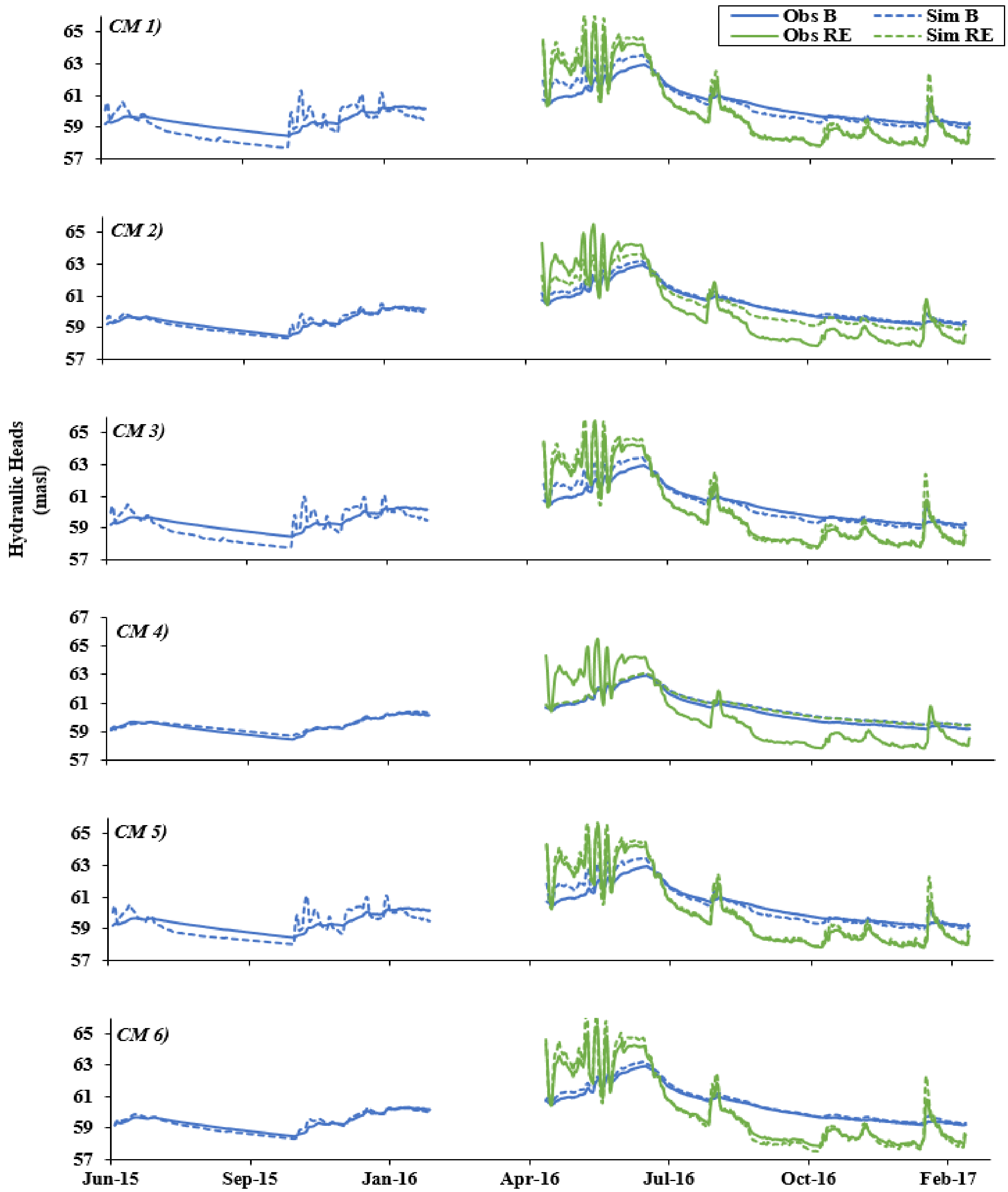


Figure 2.4: Simulated heads for six conceptual models for the entire period. Solid green and blue lines represent observed hydraulic heads in Well RE and Well B-3 respectively, whereas, dashed lines represent simulated hydraulic heads for the respective wells.

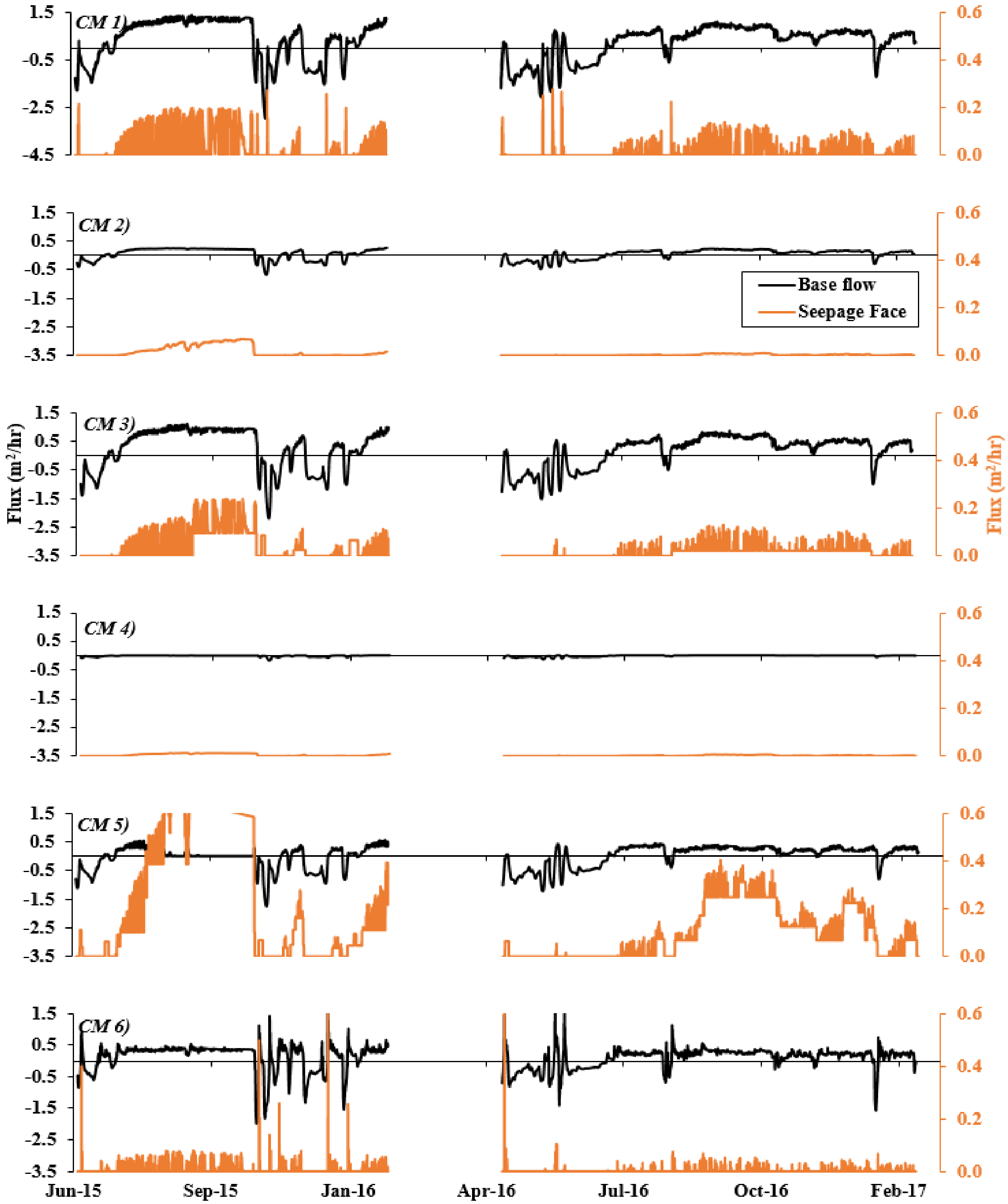


Figure 2.5: Simulated fluxes for six conceptual models for the entire period. On left side axis, plotted are the recharge and discharge fluxes. On right side axis plotted are the seepage fluxes. Positive fluxes represent either seepage or discharge from aquifer to river, whereas, negative fluxes represent recharge from river to aquifer.

2.3.2.4 *Simulated vs Measured Net Discharge*

To compare the measured discharges with simulated discharges from the conceptual models, measured volumetric discharge rates (in m^3/s) were converted to specific fluxes ($m^3/hr/m$) by dividing by 24500 m, length of the river reach between SH21 and SH60. It is important to note that an inherent assumption underlies in computing the specific fluxes per unit length of the reach that the width of well-connected portion of the aquifer is the same throughout the 24500 m reach of the river between SH21 and SH60. Even the most well-connected case, CM 1 that produces highest fluxes, hugely underestimates the amount as well as slope of the discharge curve (Figure 2.6). This difference shows that the underlying assumption is not true. While the width of the well-connected zone may be small at the study site (close to SH60), it widens out further into the aquifer as we move along the river and hence the net discharge measured over the reach is higher than predicted by the simulations at SH60. From this we hypothesize that the width of the well-connected zone varies along the river length; it is small at places where the river is at or close to the edges of the aquifer, for example SH60 and SH21 locations in Figure 2.1 b), and is considerably widened at places where the river is close to the center of the aquifer, for example midway between SH60 and SH21 in Figure 2.1 b). The difference in slopes between the simulated and measured discharge curves may be attributed to the combined effect of the size of well-connected zone as well as head difference between the river stage and the piezometric surface in the aquifer that may exist at other sites along this reach.

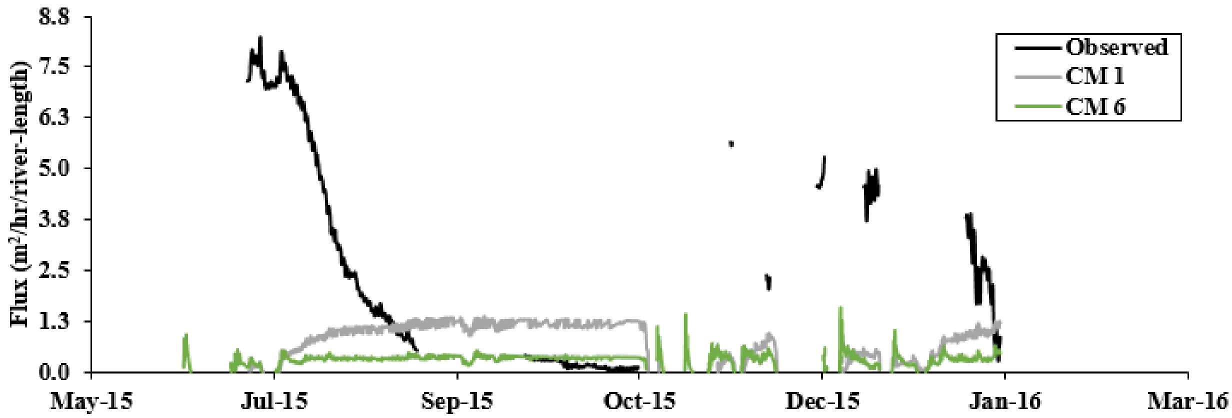


Figure 2.6: Observed vs. simulated discharges for select periods. Simulated discharges are shown for conceptual model with the highest fluxes i.e. CM 1, and the conceptual model proposed in the study i.e. CM 6 only.

2.3.3 Sensitivity Analysis

To take into account the effect of potential uncertainties in estimated parameters and input data on the overall selection criteria for best-fit conceptual model, a sensitivity analysis was performed. Saturated hydraulic conductivity and river stage values were deemed as the potential sources of uncertainty in this study. Based on judgement and experimentation, we determine that the outputs, simulated head, discharge and recharge, are sensitive to K values and the impact of van-Genuchten parameters is negligible. This is due to the fact that only the top clay layer and small portions of upper sand & gravel layer seldom undergo unsaturated during the simulation period, and hence for the most part, the flux exchange occurs through the saturated zone. Another potential source of uncertainty is the river stage, here used as variable-head BC. This uncertainty is caused by the potential datum conflict between wells in the aquifer and gage in the river channel. The objective here is to simply determine if uncertainties in K and river stage change the overall best-fit conceptual model selection, based on the criteria established earlier. For this purpose, we increase and decrease both K and river stage values by ±20 % of their respective base values (values used earlier to find the best-fit conceptual model) and the effect on simulated head and statistical measures were recorded. No perceptible change in the hydraulic head was observed when K values were increased or decreased. However, fluxes through the river-bed, discharge and recharge, are relatively sensitive to the K values (Figure 2.7 a, b) and the sensitivity of fluxes to K varies with the conceptual model. River stage effects hydraulic heads directly. Whereas, in case of fluxes, river stage also effects flux reversal patterns along with the flux magnitude (Figure 2.7 c, d). For example, during the period from August 11, 2015 to August 15, 2015, increasing the river stage changes the direction of flux from recharge to discharge as compared to the base model (Figure S1 through S4 Appendix).

Nevertheless, small changes in magnitudes of hydraulic head, the overall statistical measures between measured and observed hydraulic head repeat the same pattern as that shown by the conceptual models with base values (Table 2) and still only CM 6 explains the response of entire aquifer. Hence, it can be established that the model selection criteria is neither sensitive to K nor to the river stage values.

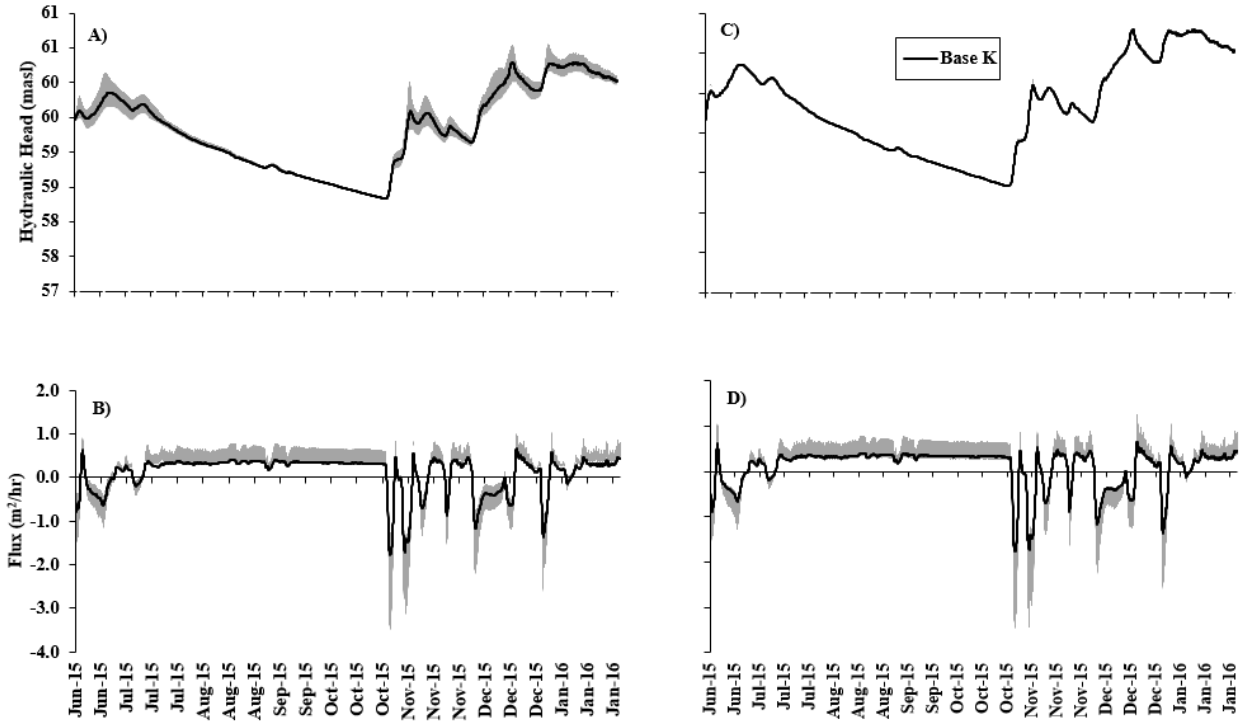


Figure 2.7: Sensitivity Analysis. A) Effect of River stage uncertainty on hydraulic heads in Well B-3. B) Effect of river stage uncertainty on flux exchange. C) Effect of K uncertainty on hydraulic heads and D) Effect of K on flux exchange. Results are shown for the proposed conceptual model, CM 6, only.

2.4 Discussion

2.4.1 Effect of Aquifer Lithology and River Incision on Head and Flux

All the six conceptual models have different configurations of aquifer lithology and river incision, which in turn results in different predicted water table conditions. In CM 1, the water table always remains unconfined. In CM 3 and CM 4, the water table at the furthest well, Well B-3, remains unconfined during most of the simulation period (~2.5 m below the clay bottom) except during the high flow seasons, for example from 10th May, 2016 to 8th Aug, 2016 (~1.4 m above the clay bottom). The water table at closer well, Well RE, remains confined during the period from 20 Apr, 2016 to 14th Jul, 2016 (~ 0.9 m above the clay bottom) and is unconfined from 18th August, 2016 to onwards (∽2.6 m below the clay bottom). Although CM 3 and CM 4 explains Well RE slightly better than CM 1, as evident by the improved statistics, the overall results show that different settings of aquifer lithology and river incision does not have a pronounced effect on the water table heads. However, these conditions have affected the fluxes between the river and the aquifer quite evidently. Together, the top clay layer and clay lenses in CM 3 have reduced total volume of discharge by 29 %, recharge by 25 % and has increased seepage by 40% compared to CM 1. The decrease in discharge and recharge in CM 3 is majorly attributed to the decrease in perimeter of the river channel connected with the conductive sand and gravel layer. It is important to note here that based on the earlier work (Wrobleski, 1996; Alden & Munster, 1998; Chakka & Munster 1997; Munster et al., 1996) information about the location of clay lenses was not available. Hence, in CM 3, discreet clay lenses, in shapes and sizes as suggested by the above cited studies, were arranged hypothetically between the river and the wells to create maximum possible disconnection between the river and the aquifer in order to test their potential as a barrier. The result show no significant decrease in the fluxes, which rules out the possibility of the potential of thin discreet clay lenses to cause a disconnection of the sort found in observed data.

In CM 4, the discharge volume has reduced by 239 %, recharge by 55.0 % and seepage has increased by 1032 % compared to CM 1 as a result of top clay layer and river incising in bottom shale. The decrease in discharge and large increase in seepage in CM 4 is attributed to the fact that only a small perimeter of river channel lying in the middle was in connection with the sand and gravel portion. As a result, during low river stages, for example from 17th Jun, 2015 to 25th Sep 2015, water level in the river was within the lower portion of the river channel surrounded by shale and essentially zero flux exchange occurred through the river bed during the time period and most of the aquifer flows into the river came through the seepage face boundary. In CM 4, seepage amount is of the same order as that of discharge and recharge. In going from CM 1 to CM 3 and to CM 4, seepage increases. This is attributed to the presence of low-conductivity high-water-holding-capacity clay around the river bed which provides more sustained supplies to fluxes compared to high-conductivity low-water-holding-capacity sand-gravel. From this follows the importance of seepage consideration in river-aquifer flux exchange assessments.

In CM 2, due to the presence of thick clay layer on top of the aquifer, water table always remain in confined conditions throughout the modeled domain (∽ 3 m above the bottom of top clay), whereas in CM 5, it is always in confined conditions close to the river channel and unconfined conditions as moves away from the river. Although CM 2 better explains water table in Well B-3, evident by improved statistics, the overall impact of aquifer lithology is not pronounced in heads in the aquifer. However, there is a large decrease in fluxes in CM 5 compared to CM 2, which is attributed to the fact that in CM 2, below the river bottom lies conductive sand and gravel layer, which allowed flux exchange with the river, whereas in CM 5, there is low conductivity material below the river through to the bottom of the aquifer, hence allowing almost no exchange between the aquifer and the river.

From above discussion, it follows that while the flux exchange between river and aquifer is very sensitive to aquifer lithology and river incision depth, the effect of these factors on well heads is not pronounced. Different settings of aquifer lithology and river incision depth can produce very similar heads in the wells, but the hydraulic head distribution in the aquifer close to the river and hence the river-aquifer flux exchange varies quite drastically between them. Careful representation of aquifer lithology especially surrounding river channel is very important in assessment of river-aquifer flux exchange. Models lacking thereof while constrained only by the hydraulic gradients in the wider aquifer may not represent the real flux exchange.

2.4.2 Effect of River Stage on Flux Reversals

For conceptual model 1 through 5 (Figure 2.5 a to f), the results show that as the river stage gets higher than a critical value above the water table, flux reversal occurs simultaneously in all of the conceptual models. For example, at the start of period 1 on Jun 17, 2015 0:00, the river stage was 3.238 m higher than the water table at the Well C-3 and the flux direction was into the aquifer which suddenly reversed to discharges into the river on Jul 8, 2015 15:00 as the river stage gets 0.048m below the water table. The largest continuous supply of discharges came during the two prolong droughts in period 1 and period 2 (Figure 2.3a). However, in conceptual model 6, the flux reversal is more frequent than the rest as it is independent of the water table in the farther aquifer. It rather depends upon the relative height of piezometric surface close to the river. For example, in period 2 from Nov 9, 2016 5:00 to Nov 16, 2016 11:00, although the water table at Well C-3 was ∽1.969 m higher than river stage, the river still continued to discharge fluxes to the aquifer. This is because the water table close to the river, at Well RE, was ∽ 0.2 m below the river stage (Figure 2.8).

A large body of literature on river-aquifer interactions (Fleckenstein et al., 2010) has demonstrated that exchange fluxes tend to be highly variable in time (e.g. Anibas et al., 2009; Schmidt et al., 2012). And an improved quantification of this variability is an important prerequisite for a better understanding of the biogeochemical processing of nutrients (e.g. Zarnetske et al., 2012; Krause et al., 2013) and pollutants (e.g. Greenberg et al., 2002; Schmidt et al., 2011) in the transition zone between ground- and surface water. Barlow and Coupe (2009) investigated the flux reversal in Bogue Phalia River in Florida caused by high-flow events i.e. high river stages. They demonstrated that exchange through the riverbed and a critical stage for flow reversals can be determined by using heat as a tracer even with little available hydraulic head data. They determined one critical river stage after which flux reversals occur in the system. Bartsch et al., (2013) found out that the reversal of fluxes is rather dynamic and can occur at different river stages depending upon the groundwater flow field. They have argued that heat tracer alone is insufficient and hydraulic head data at sufficient temporal resolution is very important in flux exchange assessments. Our results demonstrate that flux reversals are sensitive to the spatial head distribution in the aquifer in addition to temporal changes. Because of the hydrogeological heterogeneity, the piezometric surface as measured in the wider aquifer may not represent the groundwater flow field close to the river. This underlines the importance of adequately capturing the groundwater flow field in the entire aquifer at sufficient spatial resolution, in addition to temporal resolution, for adequate characterization of flux exchange between aquifer and river.

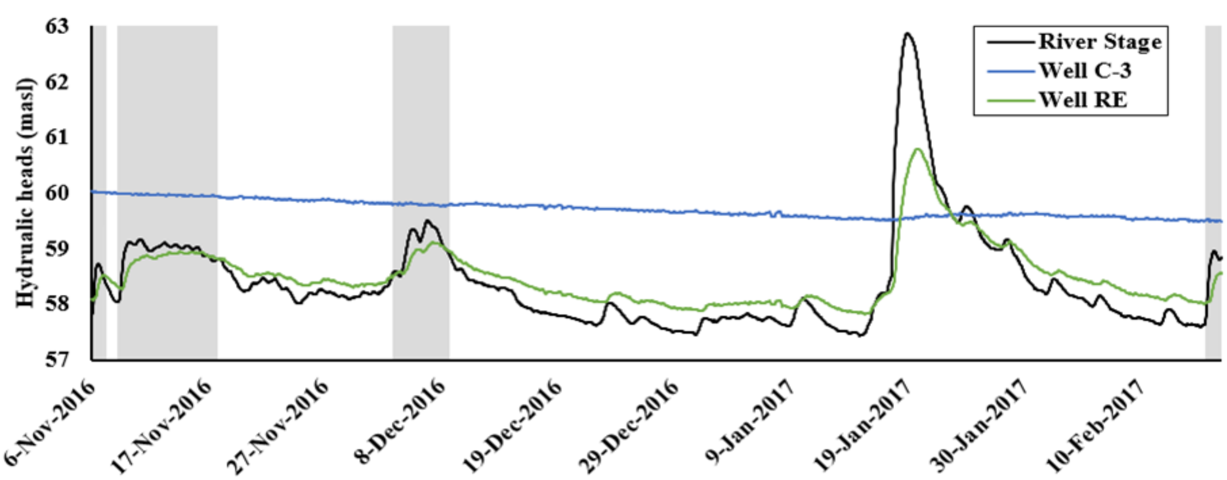


Figure 2.8: Relationship between river stage and water table at Well C-3 and Well RE. Grey highlight show periods when river stage was lower than water table at Well C-3 but higher than Well RE and the flux direction was towards the aquifer.

2.4.3 Effect of River-bed Clogging on Flux Exchange

River-bed clogging decreases the magnitude of flux exchange and effects hydraulic heads depending upon the gaining and losing river conditions (Figure 2.9). During high river flows i.e. loosing conditions, for example from Apr 27, 2016 to Jun 5, 2016, heads in the aquifer are attenuated (by ∽ 0.173 m in this case) and recharge is reduced in the post river-bed clogging simulation. This is because the presence of low-conductivity layer surrounding river channel reduces fluxes from river towards aquifer and in turn decreases heads in the aquifer. During low flows i.e. gaining conditions, for example from Sep 7, 2016 to Nov 6, 2016, heads in the aquifer are increased (by ∽ 0.2 m in this case) while the discharges from the aquifer is reduced in the post river-bed clogging simulations. This time the presence of low conductivity layer reduces the discharges from the aquifer to the river and hence allows for higher heads in the aquifer. The seepage, however, increases contrary to the recharge and the discharges. The low-conductivity layer surrounding the river channel allows for more sustained supply of fluxes through the seepage face. The cumulative recharge and discharge have decreased by 11% and 4% respectively, while seepage has increased by 34%. The inclusion of a low conductivity river bed has slightly improved the overall quality of simulation. The value of r^2 increases from 0.979 to 0.982, RMSE decreases from 0.373 to 0.340 and NSE increases from 0.970 to 0.974, as shown in Figure 2.8. From the discussion above, it also follows that although low river-bed conductivities can cause attenuation of heads in the aquifer during losing conditions, their effect is not significant to cause a disconnection between the river and the farther aquifer of the sort found in observed data.

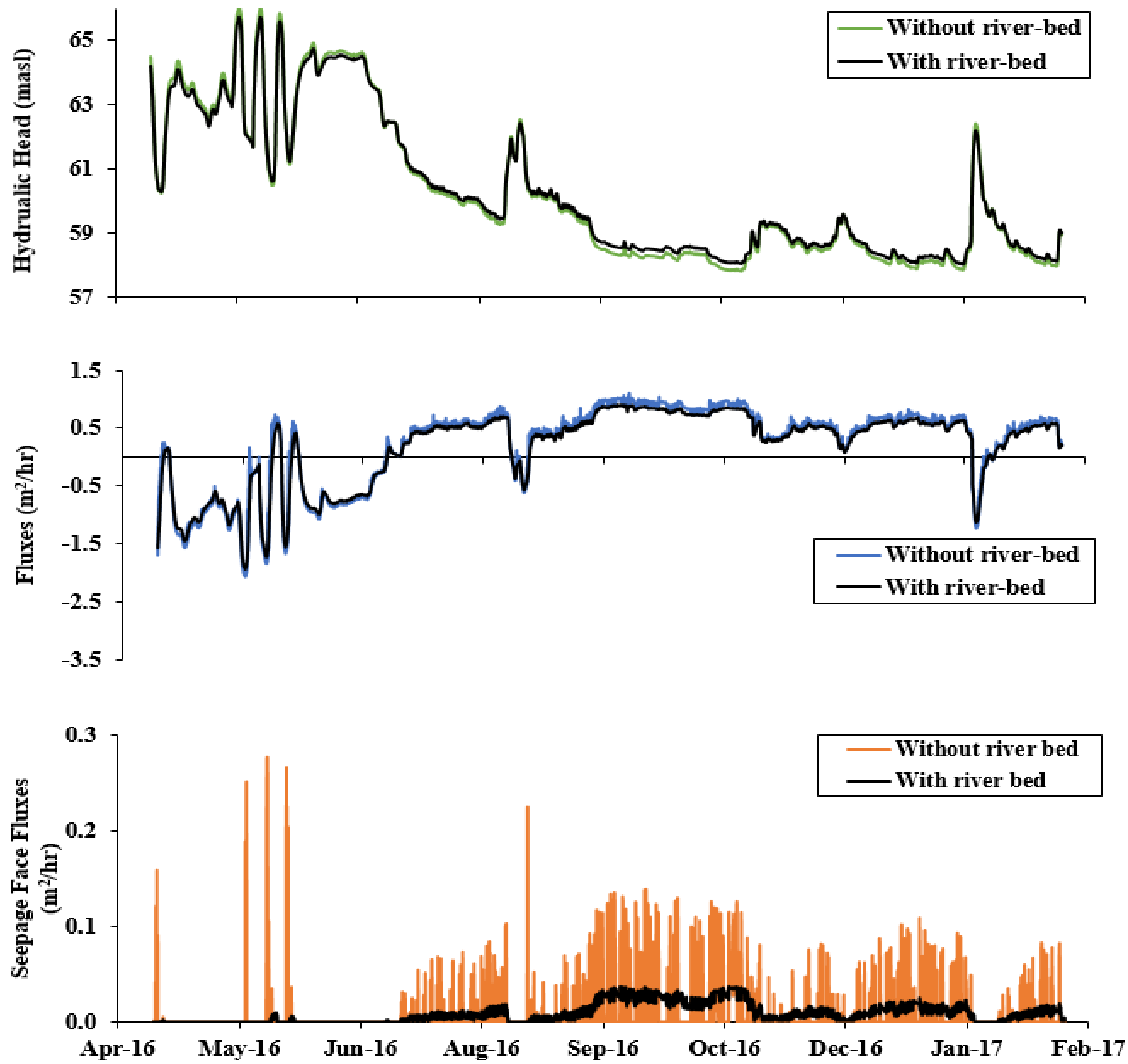


Figure 2.9: Effect of low conductivity river-bed for period 2. Each plot shows simulated output with and without low-conductivity river-bed for: A) Hydraulic heads, Well RE. B) Discharges (positive flux) and recharge (negative flux). C) Seepage.

2.4.4 Implications of Proposed Conceptual Model for Water Management

The results of the study demonstrate that only a small portion of BRAA, lying close to the Brazos River, is well-connected with the river, whereas, the greater aquifer lies disconnected by some sort of geological barrier. The flux exchange between the river and the aquifer is restricted to this relatively small zone. When compared to the generally accepted conceptual model CM 3---which assumes that entire aquifer is well-connected---the cumulative recharge and discharge in CM 6 are smaller by 151 % and 116 % respectively. This finding has important implications for water management in the area especially from the point of view of conjunctive use of surface- and groundwater. On one hand, CM 3 over-estimates the amount of water available to the river from the aquifer, especially in prolonged dry periods which are not uncommon in Texas. On the other hand, it also over-estimates recharge to the aquifer which has consequences to downstream surface water right users. Furthermore, the study finding also has important consequences for the regional BRAA Groundwater Availability Model (GAM) (Ewing, 2016). In the GAM, the discharges amounts are estimated based on hydrograph separation analysis (Ewing, 2016) which does not take into account the suggested compartmentalization of the aquifer. Also, losing and gaining reaches of the Brazos River were identified based on hydraulic heads in the aquifer within 2 miles of river gaging stations. These assumptions by GAM model imply that the entire aquifer is connected with the river and hence over-estimate the amount and time period of available discharge supplies. Finally, the proposed conceptual model also suggests that flux exchange within the two compartments of the aquifer is insignificantly small to be taken into account in the river-aquifer flux exchange process. This has important consequences for the disconnected portion of the aquifer about potential sources of recharge and discharge. For example, rainfall recharge from top and inter-aquifer flux exchange with the bottom aquifers may be more important processes to consider in large-scale regional models covering entire aquifer (O’Rourke 2006).

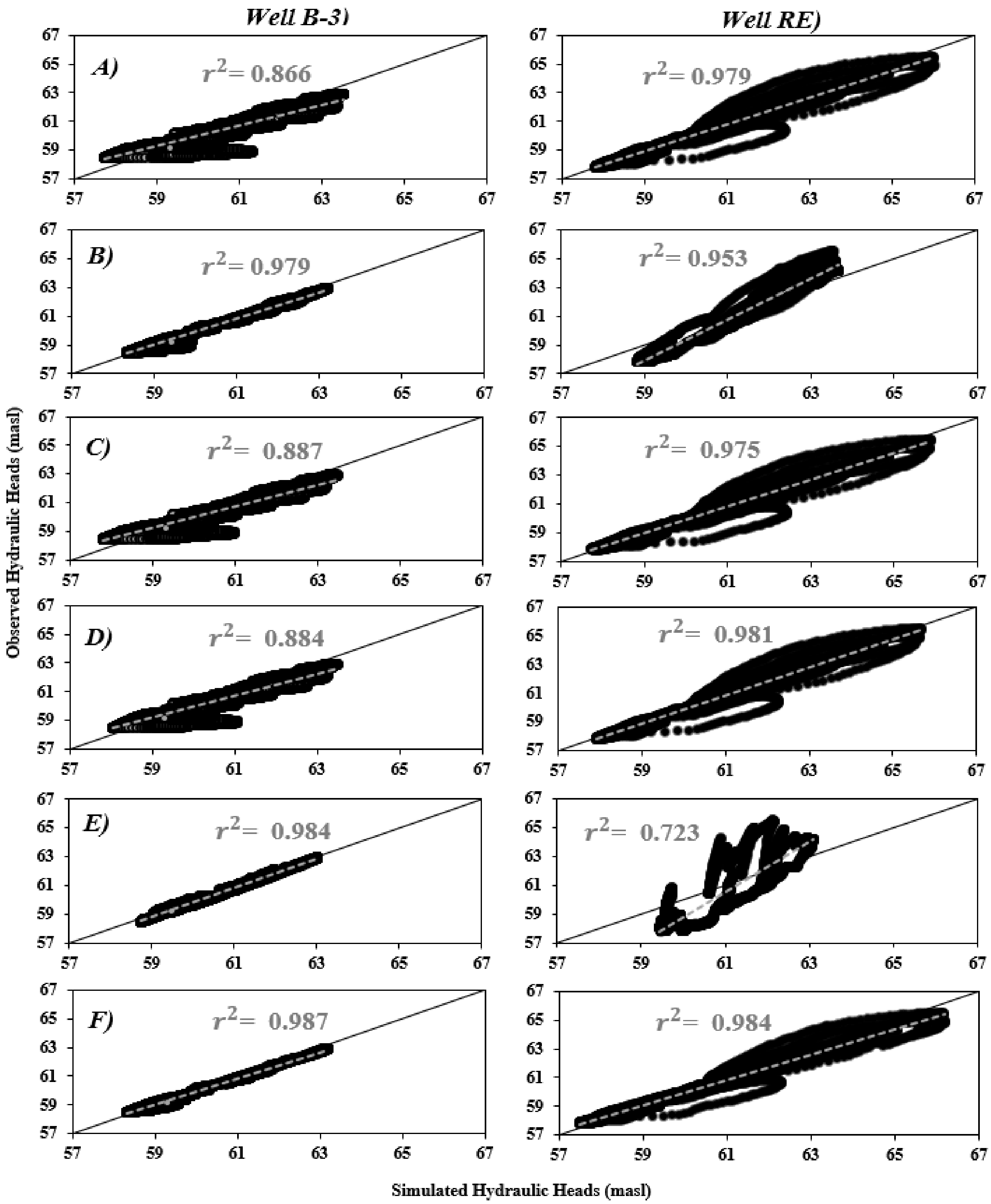


Figure 2.10: Simulated vs. Observed hydraulic heads. Row A) through F) represents conceptual model 1 through 6 respectively. Column B-3 and RE represents Well B-3 and Well RE respectively. Dotted grey line represents linear regression fit to the data, whereas, black solid line through the center of the plot represents 45° line.

2.5 Conclusions

The main focus of this study was to investigate several alternative conceptual models about the degree of connection between BRAA and Brazos River to determine the one that best explains the river-aquifer interaction and then to quantify and characterize corresponding flux exchange dynamics. Six alternative conceptual models were tested against continuously monitored hydraulic head in the aquifer at two wells, located proximal and distal from the river channel, to accurately capture the effect of geological heterogeneity on the groundwater flow field. We additionally focused on how different factors, such as, aquifer lithology, river channel incision, water table conditions, seepage face boundaries and low-conductivity river-bed effect hydraulic head distribution and the corresponding flux exchange dynamics.

With respect to the conceptual model for BRAA and Brazos River, our results demonstrate that only a small portion of the aquifer is well-connected with the river, whereas, the greater aquifer is disconnected due to geological heterogeneity in between (Figure 2f). The flux exchange is restricted to this small zone lying close to the river channel and hence the water table slopes in the farther aquifer does not accurately define the flux exchange magnitudes and flux reversals patterns. While the width of connected zone may be small at our study site SH60 (< 80 m from the river banks), large yet finite volumes of observed discharges show that the zone further spreads out into the aquifer as it moves alongside the river reach (Figure 6).

With respect to the identification of important factors and processes that control the dynamics of river-aquifer flux exchange, our results demonstrate that:

- Flux exchange magnitudes and flux reversal patterns are very sensitive to a) aquifer lithology close to the river channel and corresponding river incision and b) groundwater flow field close to the river channel.
- Due to geological heterogeneity, spatial distribution of hydraulic head may not be the same in the entire aquifer. Hence, piezometric head data measured at adequate spatial resolution should be used in river-aquifer interaction studies. Incorporation of measured discharges data in model development is also highly recommended.
- Seepages, which are often neglected in model development, can be of the same order as that of flow through the river bed depending upon aquifer lithology and corresponding river incision.
- Low conductivity river bed attenuates water table during high-river stages, increases water table during low-river stages, decreases flux magnitudes through the river bed and increases. This may have important consequences on the exchange of nutrients and biogeochemical processes in the transition zone and should be incorporated in such studies.

CHAPTER III

CONCLUSION

3.1 Summary of the Findings

This study determines the degree of connection between Brazos River Alluvium Aquifer and Brazos River and quantifies and characterizes the flux exchange dynamics at TAMU Farm Site. Alternative conceptual models, each based on previous studies, preexisting bore logs, and/or knowledge of fluvial processes typical in low-energy rivers, were tested against observed water levels in the aquifer. Instead of model calibration, a range of alternative conceptual models were tested against observed data at two locations in the aquifer. The best-fitting model was then used to estimate flux exchange between the river and the aquifer. Additionally, the study focused on how different factors, such as, aquifer lithology, river channel incision, water table conditions, seepage face boundaries and low-conductivity river-bed effect hydraulic head distribution in the aquifer and the corresponding river-aquifer flux exchange dynamics.

With respect to the conceptual model for BRAA and Brazos River, our results demonstrated that only a small portion of the aquifer was well-connected with the river, whereas, the greater aquifer was disconnected due to geological heterogeneity in between (Figure 2f). The flux exchange was restricted to this small zone close to the river channel and hence the hydraulic gradient in the wider aquifer did not define the river-aquifer flux exchange magnitudes and flux reversals patterns. The river-aquifer flux exchange was rather sensitive to the hydraulic gradient in the connected zone. While the width of connected zone may be small at our study site, SH60 (<

80 m from the river banks), large yet finite volumes of observed discharges show that the zone further spreads out into the aquifer as it moves alongside the river reach (Figure 6).

With respect to the identification of important factors and processes that control the dynamics of river-aquifer flux exchange, our results demonstrated that:

- Flux exchange magnitudes and flux reversal patterns were very sensitive to a) aquifer lithology close to the river channel and corresponding river incision and b) groundwater flow field close to the river channel. However, hydraulic head were shown to be much less sensitive to the above factors.
- Due to geological heterogeneity, spatial distribution of hydraulic head may not be the same in the entire aquifer. Hence, piezometric surface measured at adequate spatial resolution should be used in river-aquifer interaction studies. Incorporation of measured discharges data in model development is also highly recommended.
- Seepage, which are often neglected in model development, can be of the same order as that of flow through the river bed depending upon aquifer lithology, piezometric surface and river incision depth.
- Low conductivity river bed attenuated piezometric head during high-river stages, increased water table during low-river stages, decreased flux magnitudes through the river bed and increased seepage. This may have important consequences on the exchange of nutrients and biogeochemical processes in the transition zone and should be incorporated in such studies.

3.2 Directions for Future Research

In this study, the interaction between Brazos River Alluvium Aquifer and Brazos River is investigated at only one site, SH 60. Future research could use similar alternative conceptual model strategy to determine the best-fitting model at other places in the aquifer along the reach between SH 21 and SH 60. Two ideal locations would be 1) in the middle of the reach and 2) at the SH 21 site. This might explain the rather large volume of discharges measured in the field which cannot be explained by the modeled results in this study. It might also be helpful in identifying a general pattern on how the width of connected zone varies along the river throughout the entire aquifer length. Together, it could provide an effective way to incorporate geomorphological aspects of alluvium aquifers in regional-scale numerical models. Almost all the groundwater models available commercially lack the ability to explicitly incorporate geomorphology of rivers and their alluviums.

Furthermore, this study covers a small portion of the aquifer (< 420 m on each side of the river). Future studies could include wells located in the wider aquifer, covering the full lateral extent. Consequently, more processes, such as, recharge from top, evapotranspiration and inter-aquifer flux exchange from bottom aquifers could be added on top of the factors included in this study. From this, potential sources of aquifer recharge and discharge could be determine for the entire aquifer and the water budget models for the BRAA system could be improved. This will be particularly important from the point of view of the portion of BRAA disconnected with the river, where inter-aquifer leakage could be an important player in the recharge and discharge dynamics of not only BRAA but other underlying major aquifers. Finally, based upon the outcomes of above studies, a 3D regional-scale level model could be constrained between SH 21 and SH 60 and compared with the BRAA GAM outputs. Future research could also be devoted to determine effective ways to incorporate important factors and processes into regional-scale groundwater models.

The outcomes of this study suggest that seepage could be a significant portion of the river-aquifer water exchange and hence should be incorporated in the flux exchange assessment studies. While many studies recognize the need and importance of seepage face boundaries, there is no information available on field based methods to measured flux volumes entering the river through the seepage face boundary. Future research could be devoted to field measure seepage.

REFERENCES

1. Alden, A. S. and C. L. Munster. (1997). Assessment of river-floodplain aquifer interactions. Environmental & Engineering Geoscience, 3(4), 537-548.
2. Anibas, C., Fleckenstein, J. H., Volze, N., Buis, K., Verhoeven, R., Meire, P., & Batelaan, O. (2009). Transient or steady-state? Using vertical temperature profiles to quantify groundwater-surface water exchange. Hydrological Processes, 23(15), 2165-2177. doi:10.1002/hyp.7289
3. Barlow, J. R. B., & Coupe, R. H. (2009). Use of heat to estimate streambed fluxes during extreme hydrologic events. Water Resources Research, 45(1). doi:10.1029/2007wr006121
4. Bartsch, S., Frei, S., Ruidisch, M., Shope, C. L., Peiffer, S., Kim, B., & Fleckenstein, J. H. (2014). River-aquifer exchange fluxes under monsoonal climate conditions. Journal of Hydrology, 509, 601-614. doi:10.1016/j.jhydrol.2013.12.005
5. Bredehoeft, J. (2005). The conceptualization model problem---surprise. Hydrogeology Journal, 13(1), 37-46. doi:10.1007/s10040-004-0430-5
6. Brunke, M., & Gonser, T. O. M. (1997). The ecological significance of exchange processes between rivers and groundwater. Freshwater Biology, 37,1–33.
7. Brunner, P., Cook, P. G., & Simmons, C. T. (2009). Hydrogeologic controls on disconnection between surface water and groundwater. Water Resources Research, 45(1). doi:10.1029/2008wr006953
8. Carrera, J., & Neuman, S. P. (1984). Adjoint state finite element estimation of aquifer parameters under steady-state and transient conditions. In Finite Elements in Water Resources (pp. 199-211): Springer.

9. Chakka, K. B., and Munster, C. L. (1997). Atrazine and nitrate transport to the Brazos River floodplain aquifer. Transactions of ASAE, 40(3), 615–621

10. Cronin, J. G. and C. A. Wilson. (1967). Ground water in the flood-plain alluvium of the Brazos River, Whitney Dam to vicinity of Richmond, Texas. Austin, TX: Texas Water Development Board.

11. De Wrachien, D., & Fasso, C. A. (2002). Conjunctive use of surface and groundwater: overview and perspective. Irrigation and Drainage, 51(1), 1-15

12. Eddy-Miller, C.A., Wheeler, J.D., & Essaid, H.I. (2009). Characterization of interactions between surface water and near-stream groundwater along Fish Creek, Teton County, Wyoming, by using heat as a tracer. (Sci. Invest. Rep. 2009–5160, 53 pp.). Reston, VA: U.S. Geological Survey.

13. Ewing, J. E., Harding, J. J., & Jones, T. L. (2016). Final conceptual model report for the Brazos River alluvium aquifer groundwater availability model (514 pp.). Austin, TX: Texas Water Development Board

14. Fleckenstein, J. H., Krause, S., Hannah, D. M., & Boano, F. (2010). Groundwater-surface water interactions: New methods and models to improve understanding of processes and dynamics. Advances in Water Resources, 33(11), 1291-1295. doi:10.1016/j.advwatres.2010.09.011

15. Fleckenstein, J. H., Niswonger, R. G., & Fogg, G. E. (2006). River-aquifer interactions, geologic heterogeneity, and low-flow management. Ground Water, 44(6), 837-852. doi:10.1111/j.1745-6584.2006.00190.x

16. Franke, O.L., Reilly, T.E., & Bennett, G.D. (1987). Definition of boundary and initial conditions in the analysis of saturated ground-water flow systems-an introduction.

Techniques of Water-Resources Investigations of the United States Geological Survey Chapter B5. Washington, USA: U. S. Geologic Survey.

17. George, P. G., Mace, R. E., & Petrossian, R. (2011). Aquifers of Texas. Austin, TX: Texas Water Development Board.
18. Greenberg, M., Burton, G. A., & Rowland, C. (2002). Optimizing interpretation of in situ effects of riverine pollutants: Impact of upwelling and downwelling (Vol. 21).
19. Jung, H. B., Zheng, Y., Rahman, M. W., Rahman, M. M., & Ahmed, K. M. (2015). Redox Zonation and Oscillation in the Hyporheic Zone of the Ganges-Brahmaputra-Meghna Delta: Implications for the Fate of Groundwater Arsenic during Discharge. Appl Geochem, 63, 647-660. doi:10.1016/j.apgeochem.2015.09.001
20. Kalbus, E., Reinstorf, F., and Schirmer, M. (2006). Measuring methods for groundwater – surface water interactions: a review. Hydrol. Earth Syst. Sci., 10, 873-887. doi:10.5194/hess-10-873-2006
21. Krause, S., Tecklenburg, C., Munz, M., & Naden, E. (2013). Streambed nitrogen cycling beyond the hyporheic zone: Flow controls on horizontal patterns and depth distribution of nitrate and dissolved oxygen in the upwelling groundwater of a lowland river. Journal of Geophysical Research: Biogeosciences, 118(1), 54-67. doi:10.1029/2012jg002122
22. Larkin, R. G., & Sharp, J. M. (1992). On the relationship between river-basin geomorphology, aquifer hydraulics, and groundwater-flow direction in alluvial aquifers. Geological Society of America Bulletin, 104(12), 1608–1620
23. Miall, A. D. (1996). The geology of fluvial deposits: sedimentary facies, basin analysis, and petroleum geology. Berlin, Germany: Springer

24. Morel-Seytoux, H. J., Miracapillo, C., & Mehl, S. (2016). Impact of aquifer desaturation on steady-state river seepage. Advances in Water Resources, 88, 21-31. doi:10.1016/j.advwatres.2015.09.012

25. Munster, C. L., C. C. Mathewson, and C. L. Wrobleski. (1996). The Texas A&M University Brazos River hydrogeologic field site. Environmental and Engineering Geoscience, 2(4), 517-530.

26. Nash, J. E., & Sutcliffe, J. V. (1970). River flow forecasting through conceptual models part I—A discussion of principles. Journal of hydrology, 10(3), 282-290.

27. Neuman, S. P. (2003). Maximum likelihood Bayesian averaging of uncertain model predictions. Stochastic Environmental Research and Risk Assessment (SERRA), 17(5), 291-305. doi:10.1007/s00477-003-0151-7

28. O'Rourke, D. (2006). Conjunctive use of the Brazos River Alluvium Aquifer. Report 365, Texas Water Development Board.

29. Osman, Y.Z., & Bruen, M. P. (2002). Modelling stream–aquifer seepage in an alluvial aquifer: An improved loosing-stream package for MODFLOW. Journal of Hydrology, 264, 69-89

30. Pavelic, P., Dillon, P. J., Mucha, M., Nakai, T., Barry, K. E., & Bestland, E. (2011). Laboratory assessment of factors affecting soil clogging of soil aquifer treatment systems. Water Res, 45(10), 3153-3163. doi:10.1016/j.watres.2011.03.027

31. Pholkern, K., Srisuk, K., Grischek, T., Soares, M., Schäfer, S., Archwichai, L., & Wirojanagud, W. (2015). Riverbed clogging experiments at potential river bank filtration sites along the Ping River, Chiang Mai, Thailand. Environmental Earth Sciences, 73(12), 7699-7709. doi:10.1007/s12665-015-4160-x

32. Rhodes, K. A., Proffitt, T., Rowley, T., Knappett, P. S. K., Montiel, D., Dimova, N., & Miller, G. R. (2017). The Importance of Bank Storage in Supplying Discharges to Rivers Flowing Through Compartmentalized, Alluvial Aquifers. Water Resources Research, 53(12), 10539-10557. doi:10.1002/2017wr021619

33. Rushton, K. (2007). Representation in regional models of saturated river–aquifer interaction for gaining/losing rivers. Journal of Hydrology, 334(1-2), 262-281. doi:10.1016/j.jhydrol.2006.10.008

34. Schmidt, C., Martienssen, M., & Kalbus, E. (2011). Influence of water flux and redox conditions on chlorobenzene concentrations in a contaminated streambed. Hydrological Processes, 25(2), 234-245. doi:10.1002/hyp.7839

35. Schmidt, C., Musolff, A., Trauth, N., Vieweg, M., & Fleckenstein, J. H. (2012). Transient analysis of fluctuations of electrical conductivity as tracer in the stream bed. Hydrology and Earth System Sciences, 16(10), 3689-3697. doi:10.5194/hess-16-3689-2012

36. Schornberg, C., Schmidt, C., Kalbus, E., & Fleckenstein, J. H. (2010). Simulating the effects of geologic heterogeneity and transient boundary conditions on streambed temperatures — Implications for temperature-based water flux calculations. Advances in Water Resources, 33(11), 1309-1319. doi:10.1016/j.advwatres.2010.04.007

37. Shah, D. S., Kress, H. W., and Legchenko, A. (2007). Application of surface geophysical methods, with emphasis on Magnetic Resonance Soundings, to characterize the hydrostratigraphy of the Brazos River Alluvium Aquifer, College Station, Texas, July 2006—a pilot study. (Sci. Invest. Rep. 2007-5203, pp.). Austin, TX: U.S. Geological Survey.

38. Šimůnek, J., M. Th. van Genuchten, and M. Šejna (2011), The HYDRUS software package for simulating two- and three-dimensional movement of Water, Heat, and Multiple Solutes in Variably-Saturated Media, Technical Manual, Version 2.0. (258 pp.) Prague, Czech Republic: PC Progress

39. Sophocleous, M. (2002). Interactions between groundwater and surface water: The state of the science (Vol. 10, pg 52, 2002). Hydrogeology Journal, 10(2), 348–348

40. Spalding, C.P., & Khaleel, R. (1991). An evaluation of analytical solutions to estimate drawdowns and stream depletions by wells. Water Resources Research, 27(4), 597–609

41. Sun, N. Z., Jeng, M. C., & Yeh, W. W. G. (1995). A Proposed Geological Parameterization Method for Parameter Identification in Three-Dimensional Groundwater Modeling. Water Resources Research, 31(1), 89-102.

42. Tang, Q., Kurtz, W., Brunner, P., Vereecken, H., & Hendricks Franssen, H. J. (2015). Characterisation of river–aquifer exchange fluxes: The role of spatial patterns of riverbed hydraulic conductivities. Journal of Hydrology, 531, 111-123. doi:10.1016/j.jhydrol.2015.08.019

43. TPWD. (1974). A report on the physical characteristics of rivers, streams and bayous in Texas. Texas parks and Wildlife Department, Texas Agricultural Extension Service, Texas A&M University. Austin, TX: Author.

44. Tsai, F. T. C., Sun, N. Z., & Yen, W. W. G. (2003). A combinatorial optimization scheme for parameter structure identification in ground water modeling. Groundwater, 41(2), 156-169.

45. Tsai, F. T. C., Sun, N.-Z., & Yeh, W. W. G. (2003). Global-local optimization for parameter structure identification in three-dimensional groundwater modeling. Water Resources Research, 39(2). doi:10.1029/2001wr001135

46. Turco, M. J., East, J. W., Milburn, M. S., & Board, T. W. D. (2007). Base flow (1966–2005) and streamflow gain and loss (2006) of the Brazos River, McLennan County to Fort Bend County, Texas. (Sci. Invest. Rep. 2007–5286, 27 pp.). Reston, VA: U.S. Geological Survey.

47. Winter, T. C., Harvey, J. W., Franke, O. L., & Alley, W. M. (1998). Groundwater and Surface Water: A single resource. (U.S. Geological Survey Circular 1998-1139). Denver, CO: U. S. Geologic Survey

48. Woessner, W. W. (2000). Stream and fluvial plain ground water interactions: Rescaling hydrogeologic thought. Ground Water, 38(3), 423–429.

49. Wong, S. (2012). Developing a geospatial model for analysis of a dynamic, heterogeneous aquifer: The Brazos River Alluvium Aquifer, central Texas (88 pp.). Waco, TX: Baylor University.

50. Wrobleski, C. L. (1996). An aquifer characterization at the Texas A&M University Brazos River Hydrologic Field Site, Burleson Co., Texas (Vol. xii, 127 pp.). College Station, TX: Texas A&M University.

51. Zarnetske, J. P., Haggerty, R., Wondzell, S. M., & Baker, M. A. (2011). Labile dissolved organic carbon supply limits hyporheic denitrification. Journal of Geophysical Research, 116(G4). doi:10.1029/2011jg001730

APPENDIX A

<u>Steps to Run Simulation in HYDRUS GUI</u>

Following steps were used in all the conceptual models unless specified.

1) Define Geometry:

The first step is to define geometry of the domain. This includes defining the shape of aquifer and river channel. I used following three mesh sizes or spaces between the nodes to appropriately capture the geometry: a) 0.3 m for majority of river channel nodes; however, for few of the top ones, increased nodal spacing, 0.5 to 0.8 m, was used depending upon the requirement to create finite grid for the entire modeled domain.

2) Define Material Properties:

The three types of materials used in the study are defined in section 2.2.3.1 and their hydraulic properties are shown in Table 2.1.

3) Boundary Conditions:

Different kinds of boundary conditions used have been defined in section 2.2.3.3.

4) Define Initial Conditions:

In all case, following steps were followed before final transient simulations were run:

a) After everything has been defined until step 3), first simulations were run to get realistic initial conditions that would be used later in the transient simulations. For this simulation, pressure head distribution, being equal to the first value in the data record, varying linear from top to bottom of the domain were used as initial conditions. The simulations were then run for the next 21 days of the record until reasonable steady state is achieved. Note that in HYDRUS there is no option to directly calculate steady state unlike MODFLOW. So the simulations were run until fluxes became constant.

b) The final pressure distribution from step a) were then imported as initial conditions in a new set of model keeping everything else the same. The transient simulation were then run for the entire period.

5) Output Processing:

Although HYDRUS allows to get outputs at specified time intervals, the output text files contain random values at time steps in between. The difference in time steps shifts the whole series relative to the observed data which gives a false sense of time lag between simulated and observed values. So especial care should be taken in removing those time steps in outputs before they are compared with observed data.

<u>Data Files</u>

The excel files containing input and output data used in the study are described below. All the files are available on Synology in the folder "***Tayyab_MSThesis_Folder"*** located at **DriveName**:\Student Folders\Tayyab Mehmood\DATA.

1) Input Files:

Excel File Name: **Period_1.xlsx**

The file contains input data for the period 1, April 28, 2015 to January 29, 2016, including discharges, hydraulic heads in all three wells and river stage at SH 60. The original dataset at 20 minutes interval and hourly averaged values---averaged around the center value---both are shown. The sheets are named accordingly.

Excel File Name: **Period_2.xlsx**

The file contains input data for the period 2, April 01, 2016 to February 07, 2017, including hydraulic heads in all four wells (RE was added during this period) and river stage at SH 60. The original dataset at 20 minutes interval and hourly averaged values---averaged around the center value---both are shown. The sheets are named accordingly.

There is a sheet "Both Periods @ Hourly" which contains both the periods combined date-wise.

Excel File Name: **Initial_and_Final_Cond..xlsx**

The files contains information on how to convert the hydraulic heads to pressure heads and all the relative information. Read the comments in the excel file for details.

Excel File Name: **Lag_Between_SH60&SH21.xlsx**

This files contains river stages at SH 21 and SH 60 for the period 1. These are the measured river stages as measured in Rhodes et al., 2017.

2) Output Files

Excel File Name: **Models_results.xlsx**

This file contains 9 sheets and the names are self-explanatory. In sheets CM 1 through CM 6, simulated hydraulic heads and fluxes are there along with observed data. The data is graphed as well. It also contains the values of statistical parameters.

Excel File Name: **Sensitivity.xlsx**

This file contains simulated hydraulic heads and fluxes for sensitivity analysis for all the conceptual models. There are 12 sheets. Each Conceptual model has 2 sheets. For example, for conceptual model 1, the sheets are named as “Case1_K” and “Case1_RS”. Where, ‘Case1’ refers to Conceptual model number 1 and ‘K’ & ‘RS’ refer to sensitivity to changes in K values and River Stage values respectively.

<u>HYDRUS Files</u>

The folder “HYDRUS_Files” contains the HYDRUS simulations for all the conceptual models. There are 12 files altogether in the folder. Each conceptual model has two files. For example for conceptual model 1, files are named as Case1a.h3d2 and Case1b.h3d2. Where ‘Case1’ represents conceptual model number ‘1’ and ‘a’ represents spin-up simulations and ‘b’ represents transient simulations.

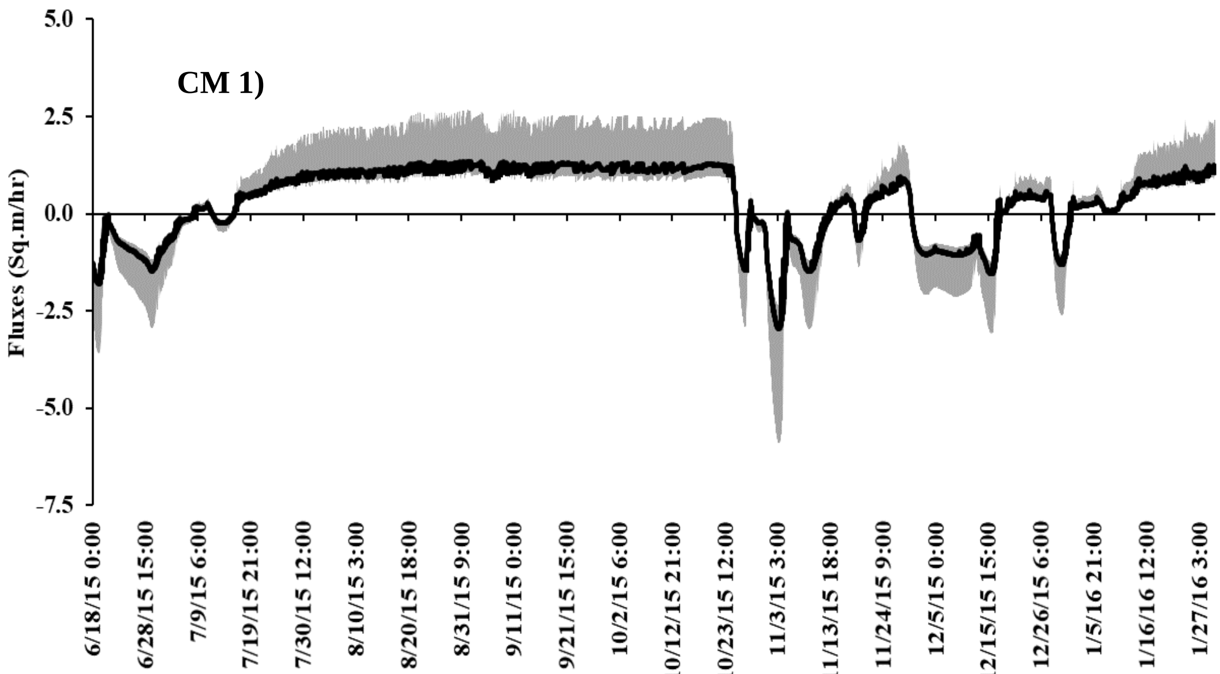
APPENDIX B

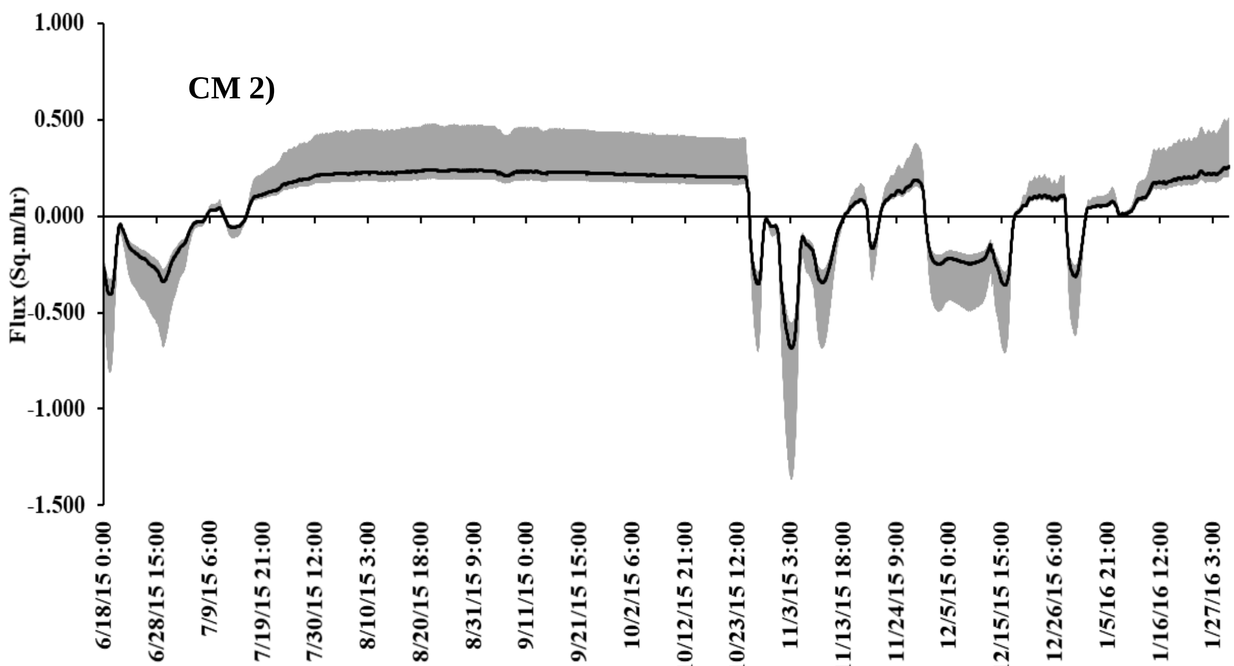
B.1 Results of the Sensitivity Analysis

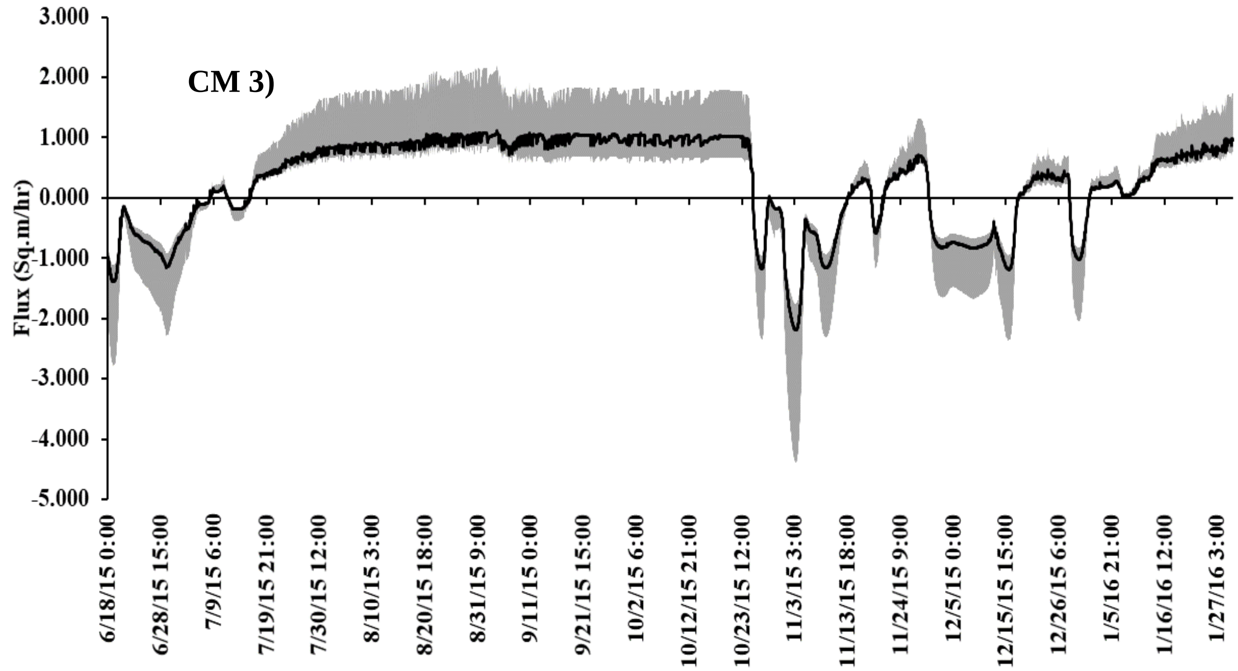
Results are shown only for period 1, from April 28, 2015 to January 29, 2016.

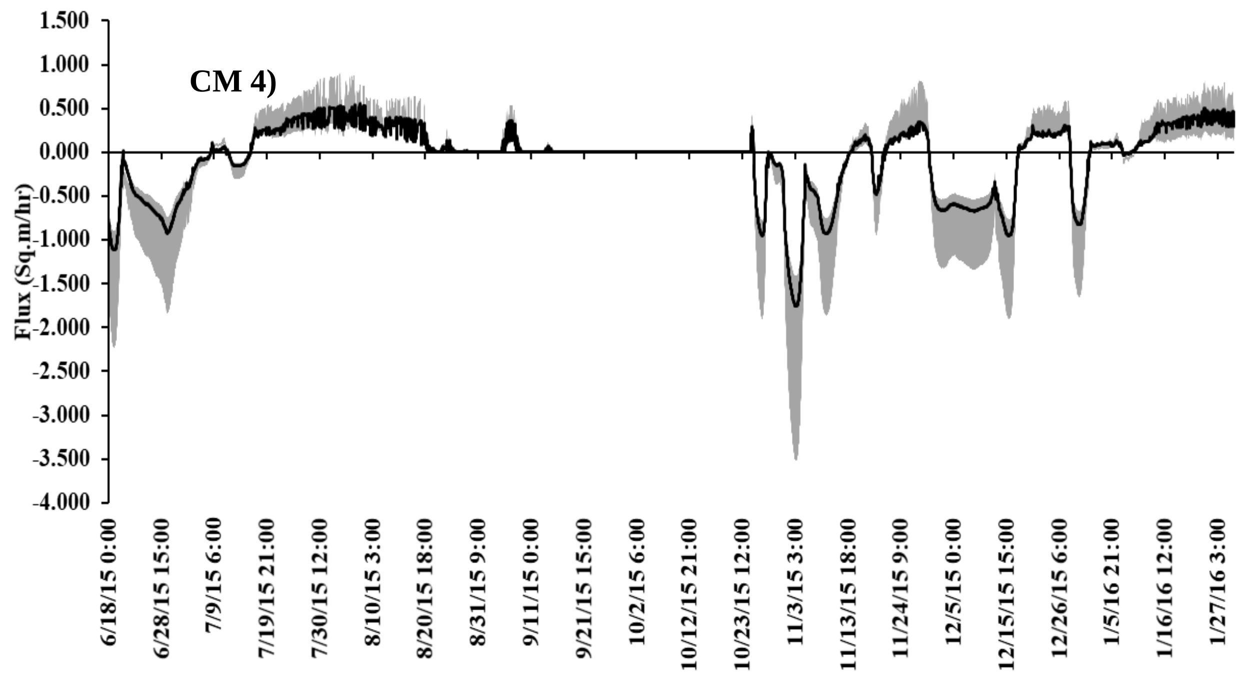
B.1.1 Sensitivity to Saturated Hydraulic Conductivity K

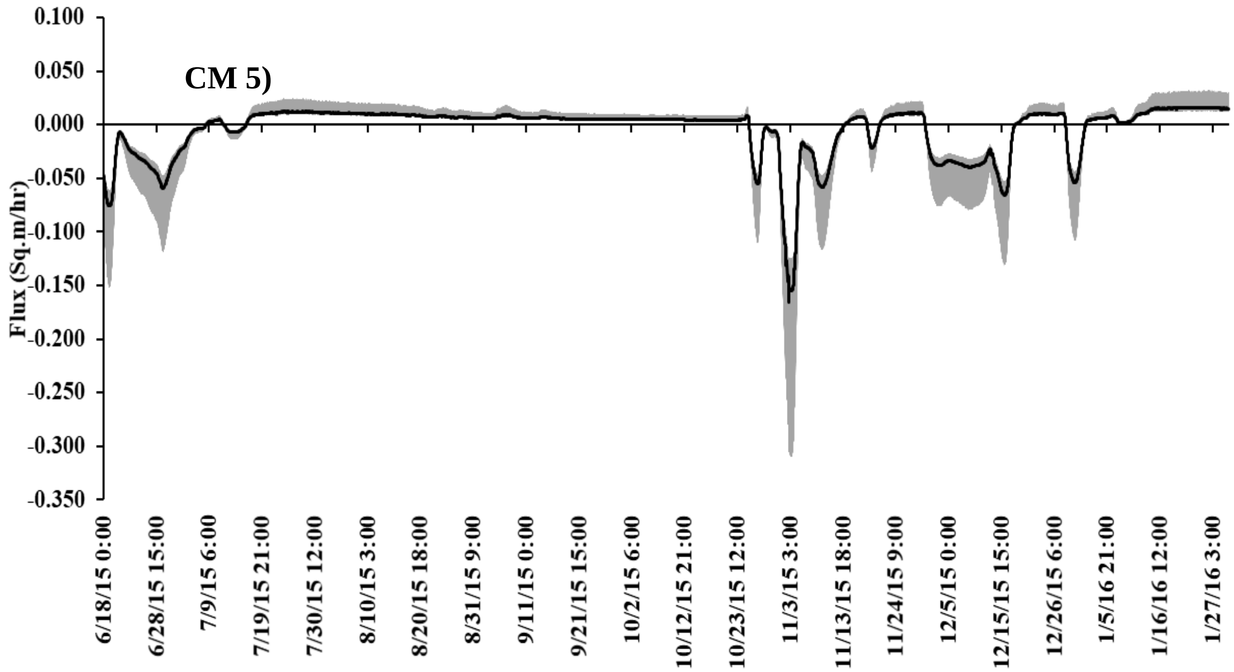
B.1.1.1 Fluxes






CM 3)
Flux (Sq.m/hr)
3.000
2.000
1.000
0.000
-1.000
-2.000
-3.000
-4.000
-5.000
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00


CM 4)
Flux (Sq.m/hr)
1.500
1.000
0.500
0.000
-0.500
-1.000
-1.500
-2.000
-2.500
-3.000
-3.500
-4.000
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00



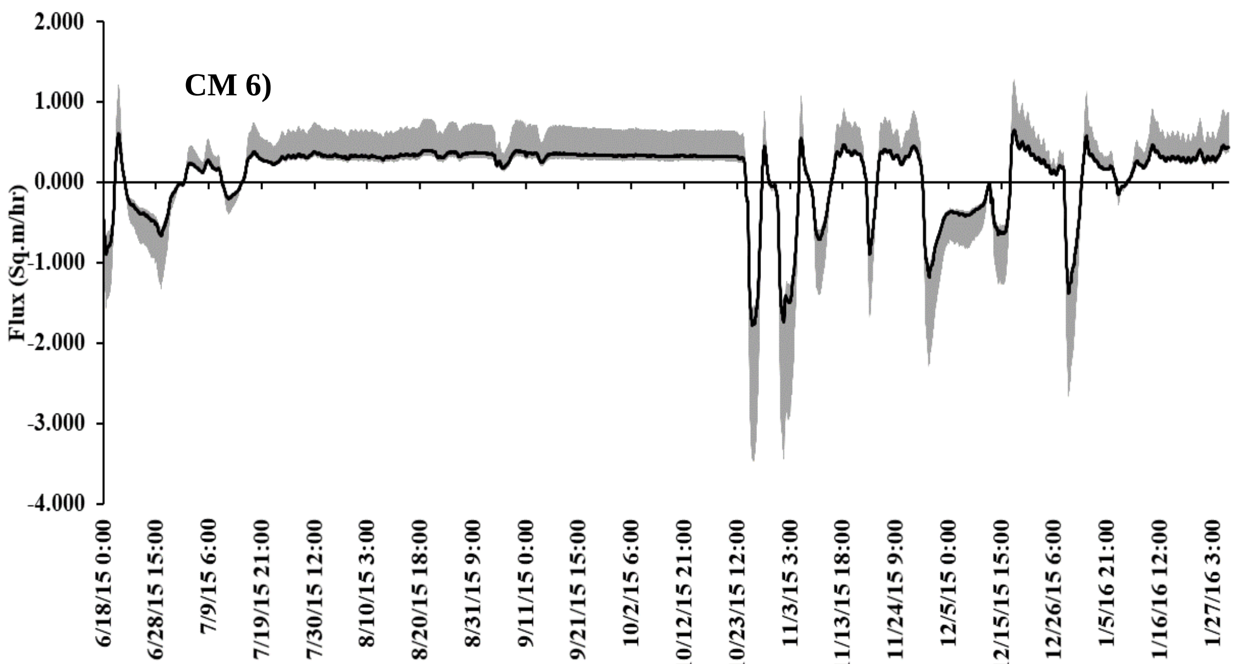
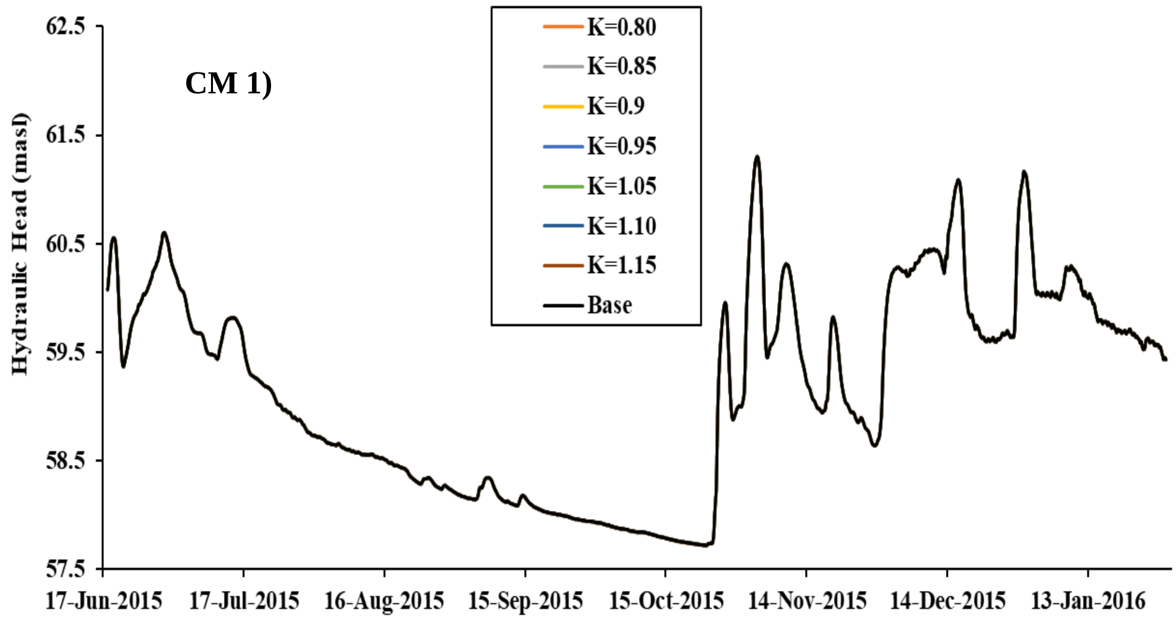
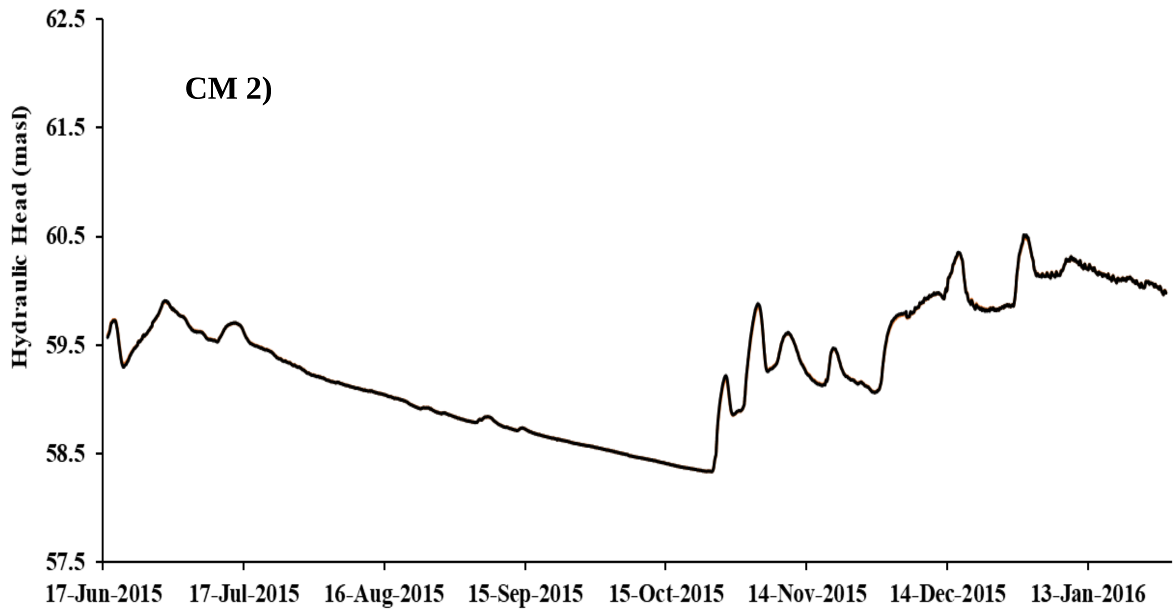
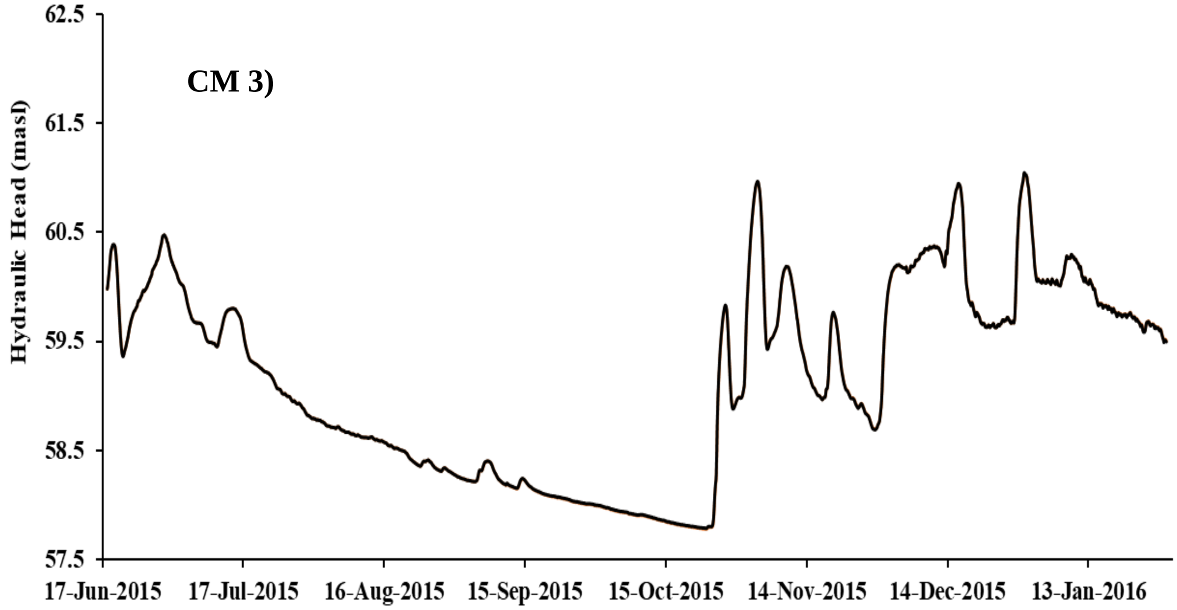


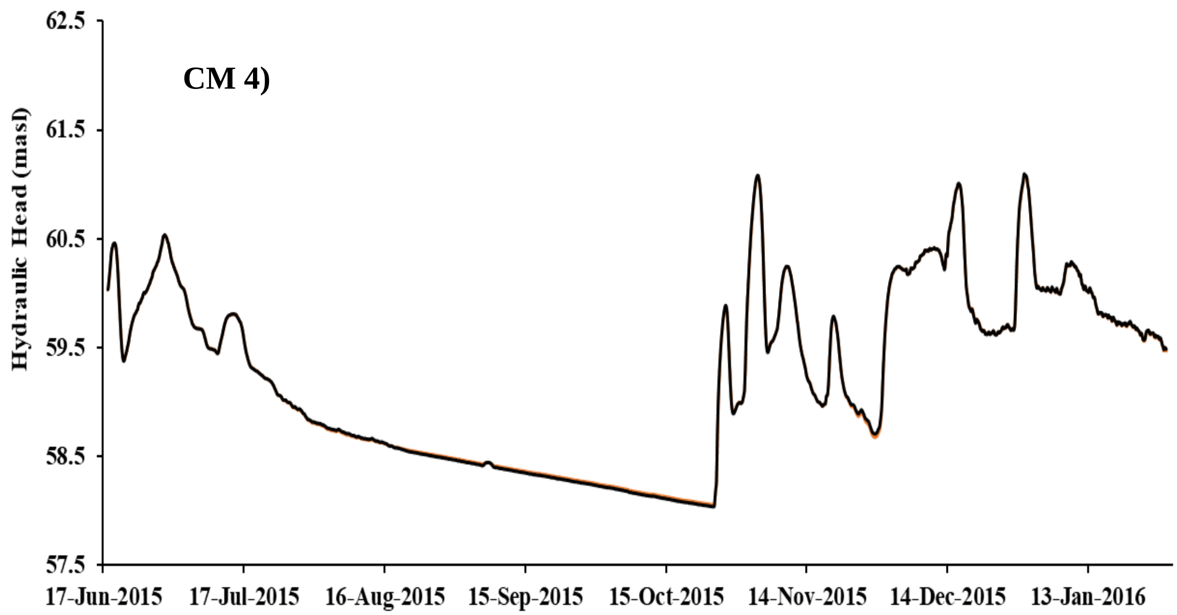
Figure S 1: Sensitivity of discharge and recharge fluxes to the saturated hydraulic conductivity.

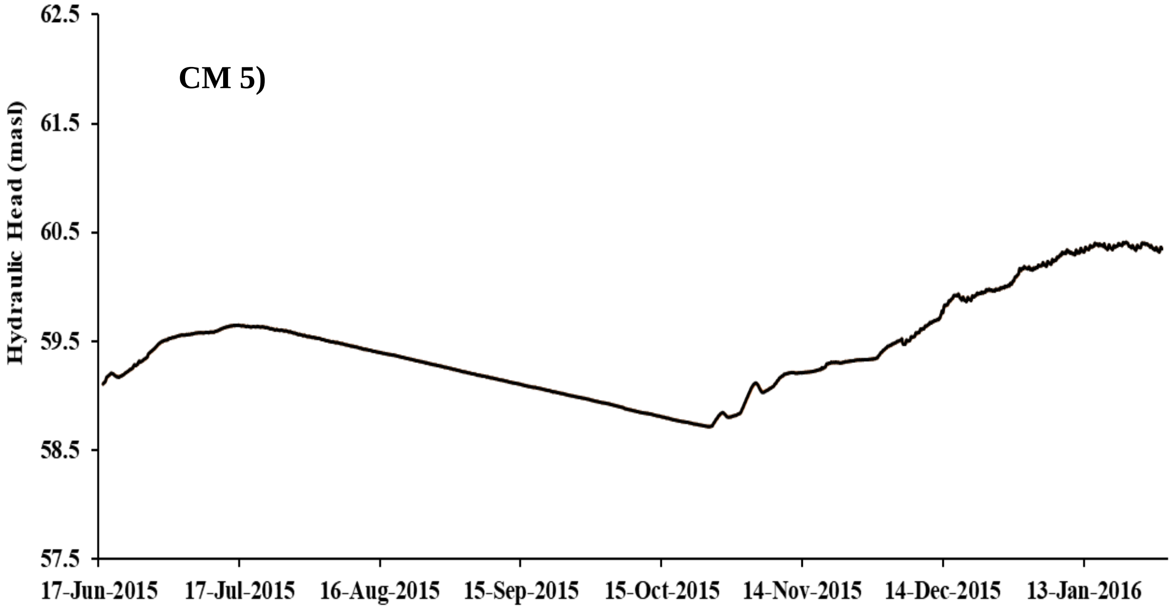
B.1.1.2 Hydraulic Head






CM 3)
Hydraulic Head (masl)
62.5
61.5
60.5
59.5
58.5
57.5
17-Jun-2015
17-Jul-2015
16-Aug-2015
15-Sep-2015
15-Oct-2015
14-Nov-2015
14-Dec-2015
13-Jan-2016


CM 4)
Hydraulic Head (masl)
62.5
61.5
60.5
59.5
58.5
57.5
17-Jun-2015
17-Jul-2015
16-Aug-2015
15-Sep-2015
15-Oct-2015
14-Nov-2015
14-Dec-2015
13-Jan-2016



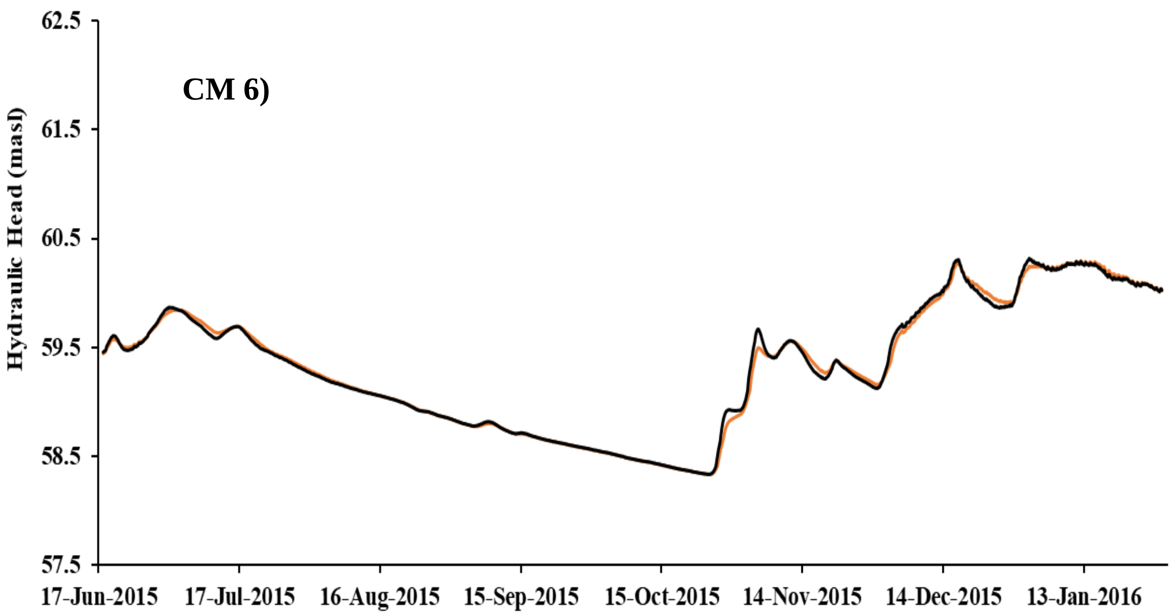
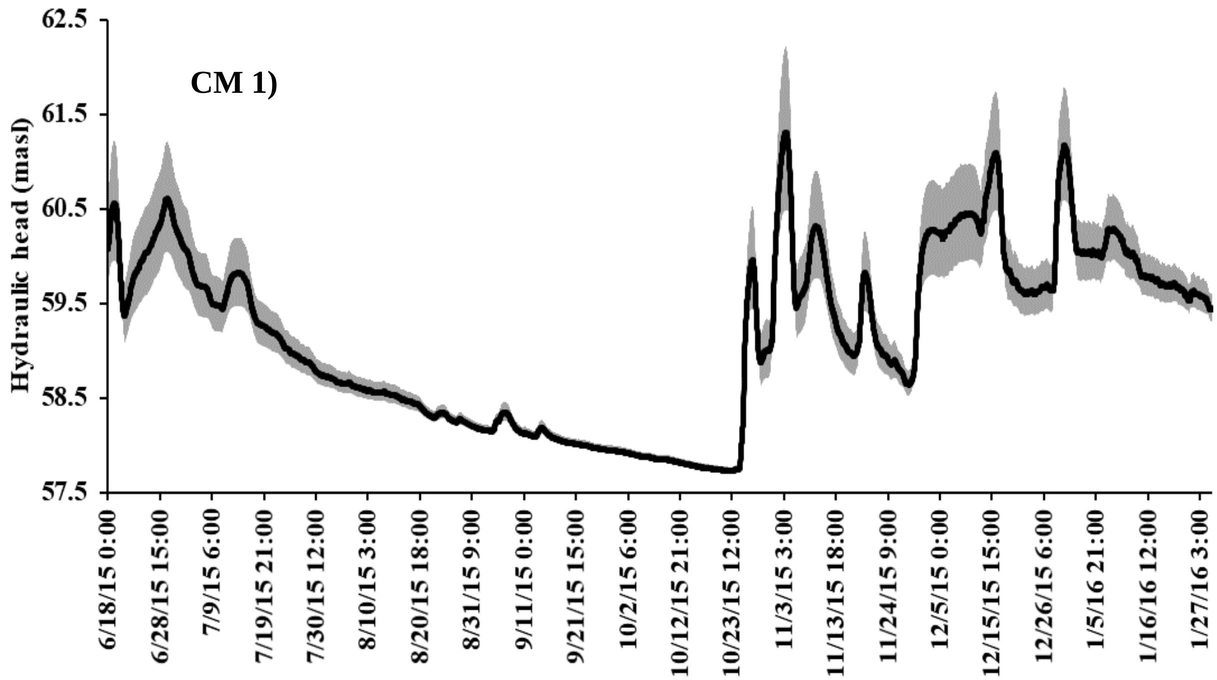
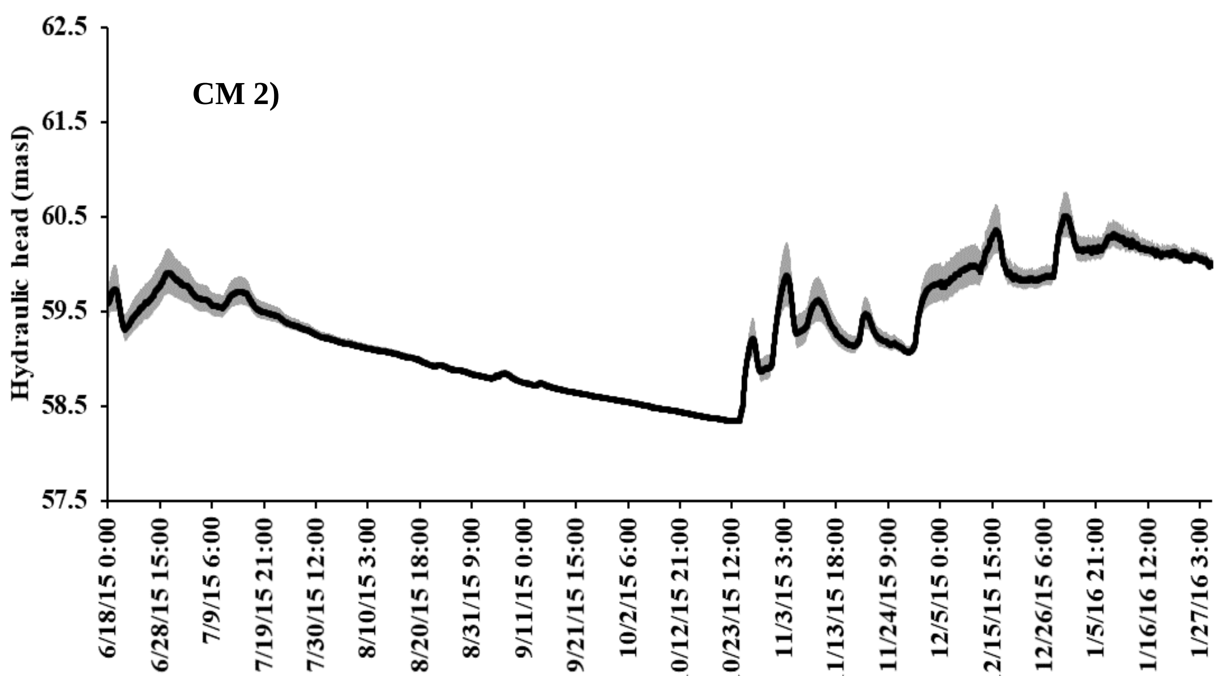


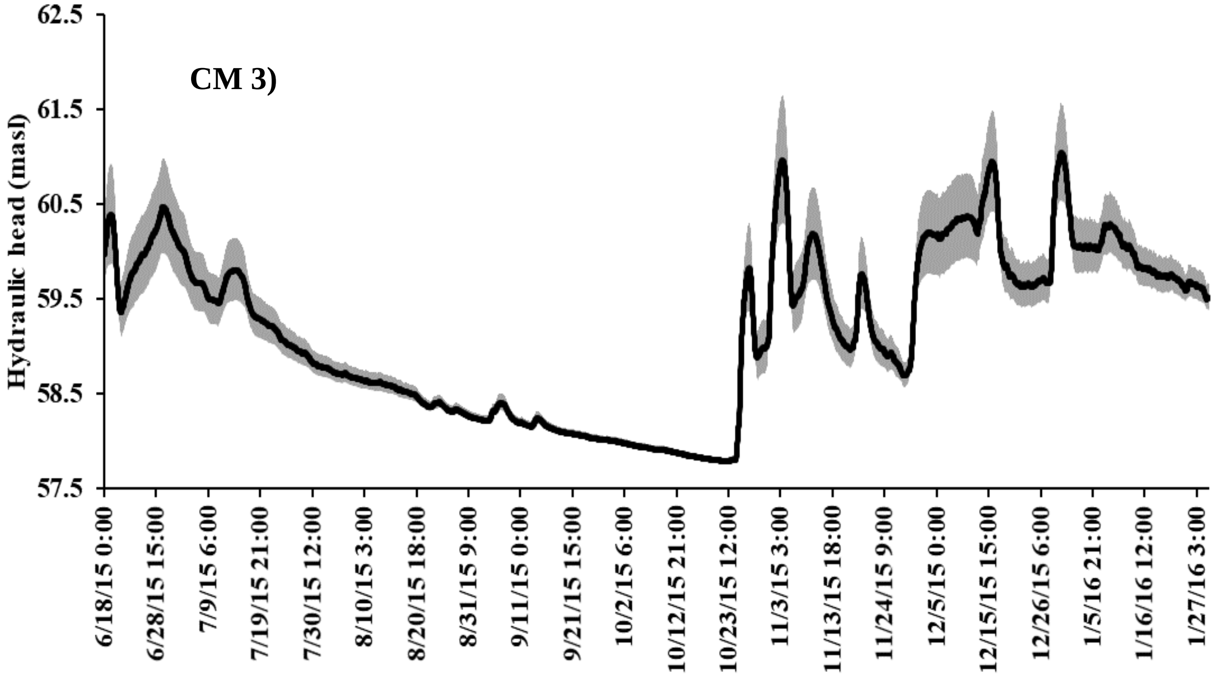
Figure S 2: Sensitivity of hydraulic heads to the saturated hydraulic conductivity.

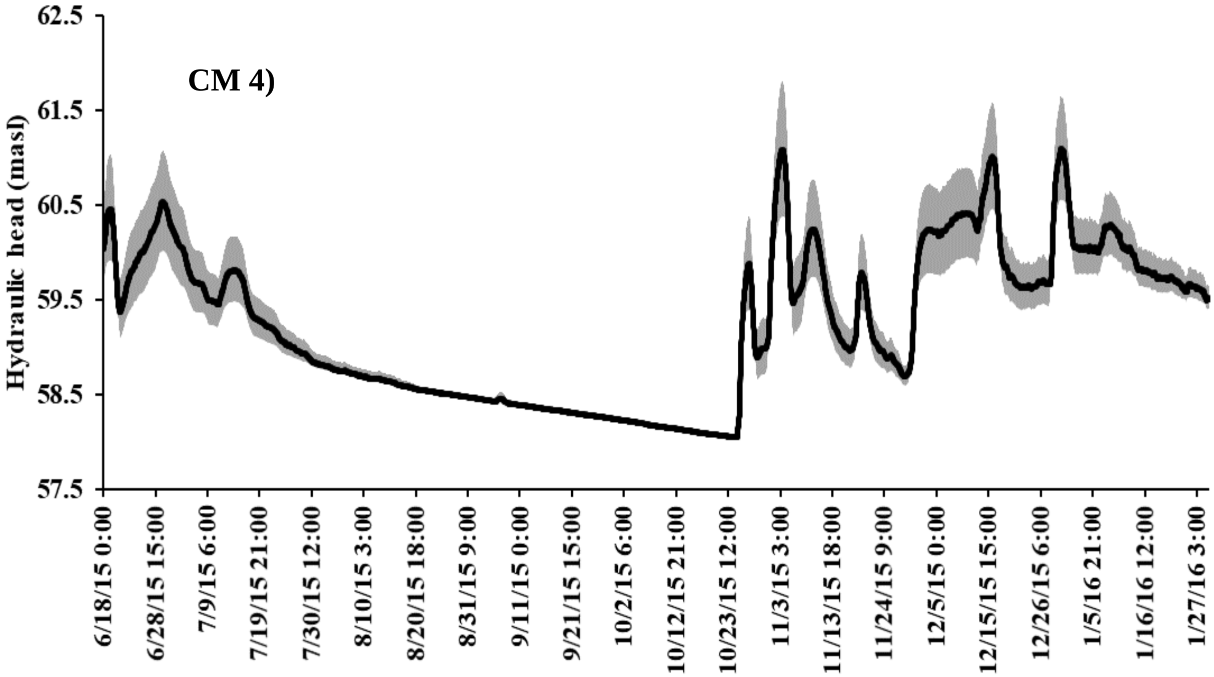
B.1.2 Sensitivity to River Stage

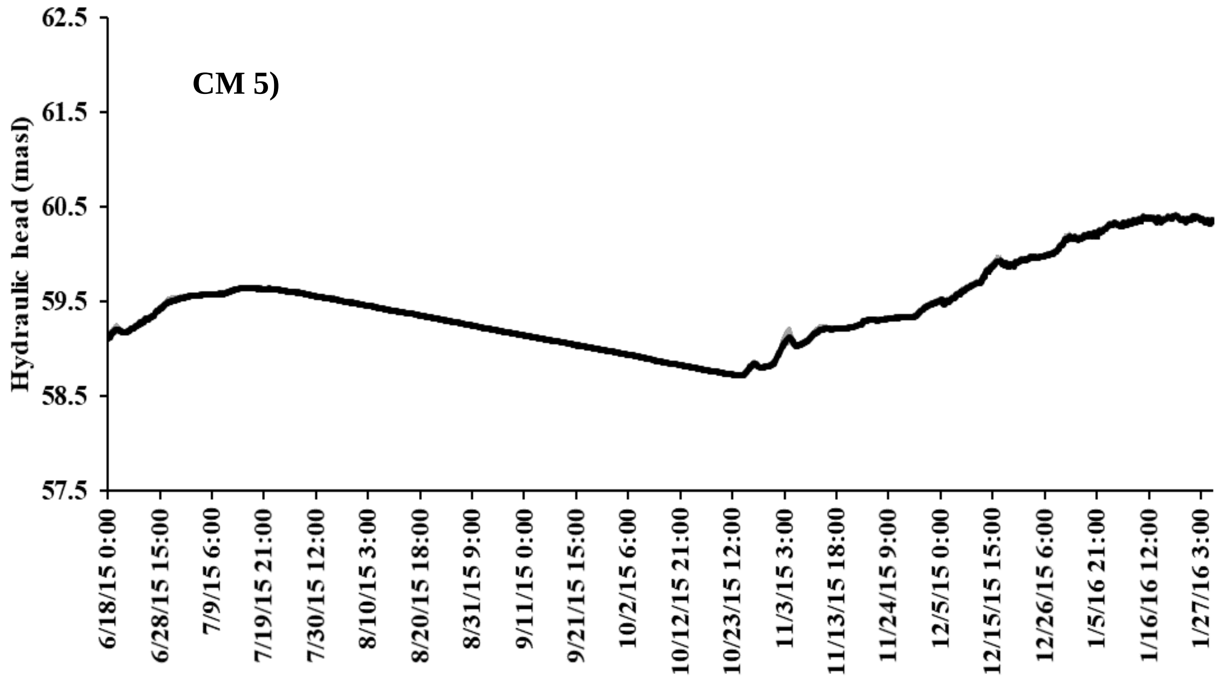
B.1.2.1 Hydraulic Head






CM 3)
Hydraulic head (masl)
62.5
61.5
60.5
59.5
58.5
57.5
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00


CM 4)
Hydraulic head (masl)
62.5
61.5
60.5
59.5
58.5
57.5
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00



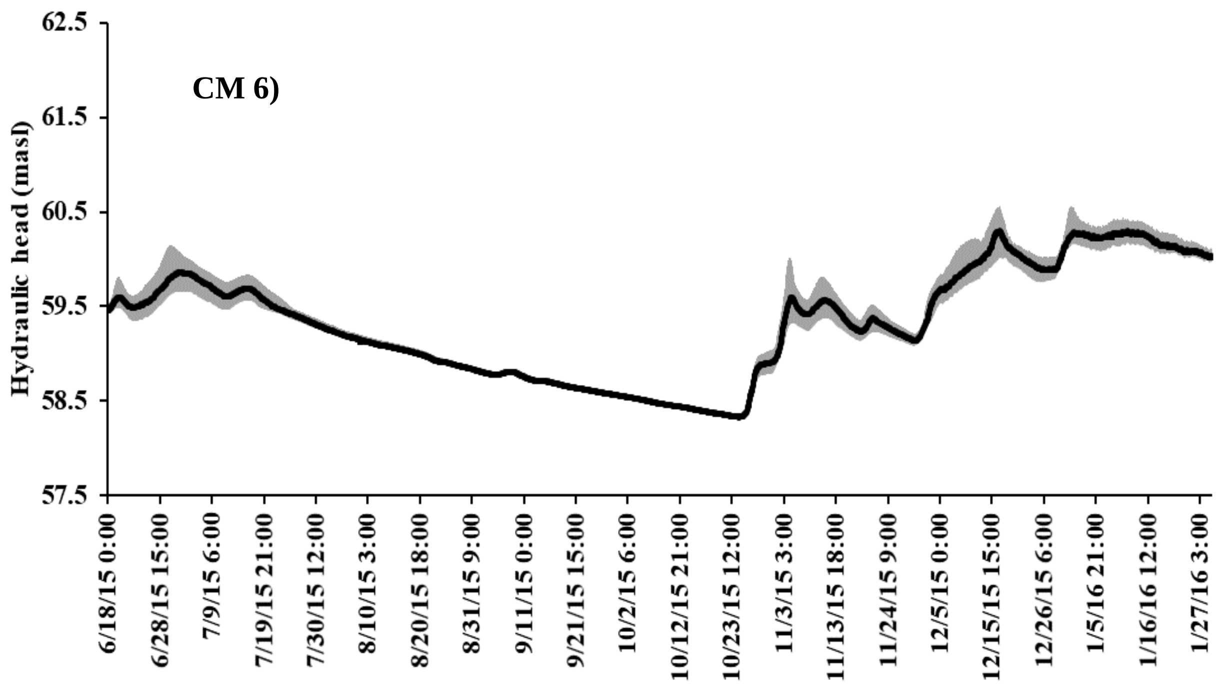
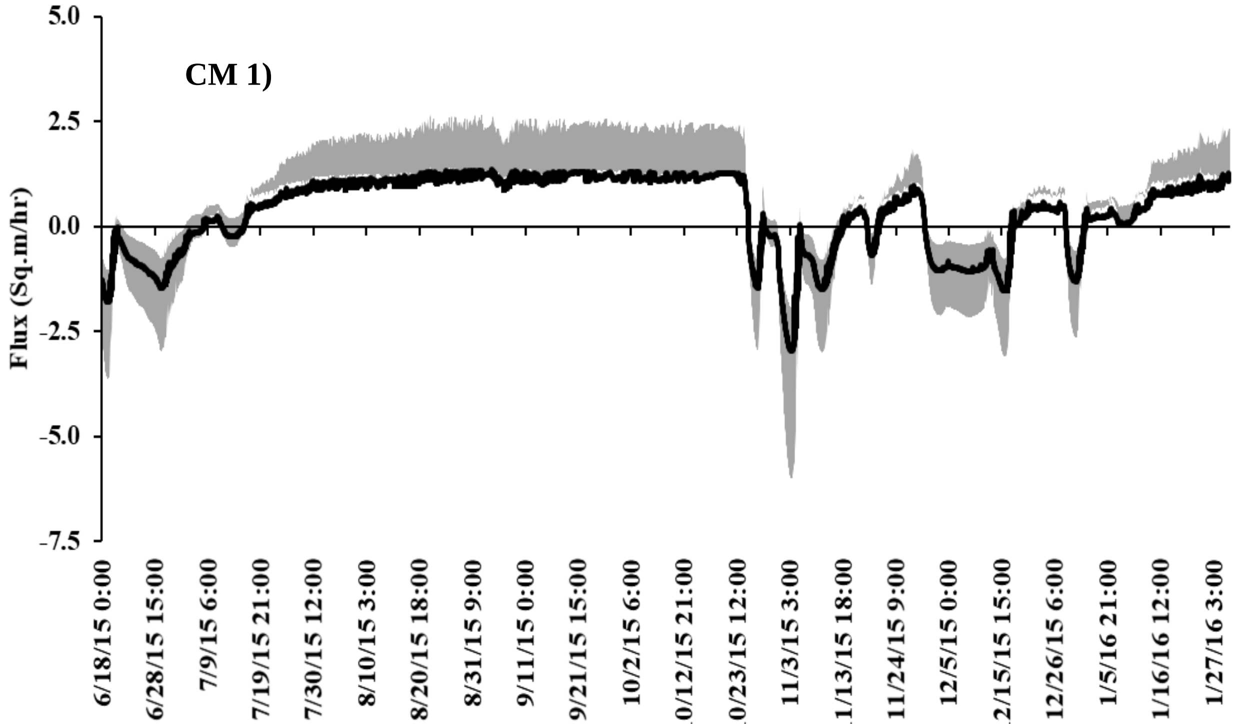
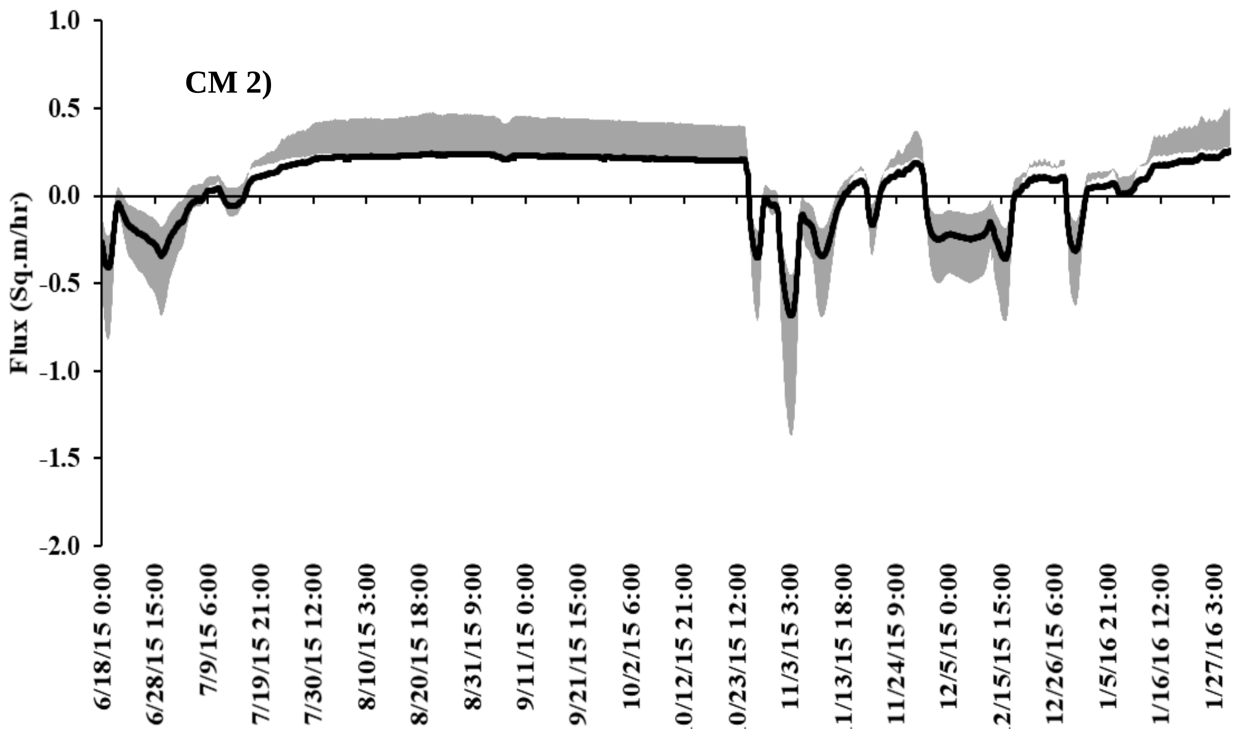
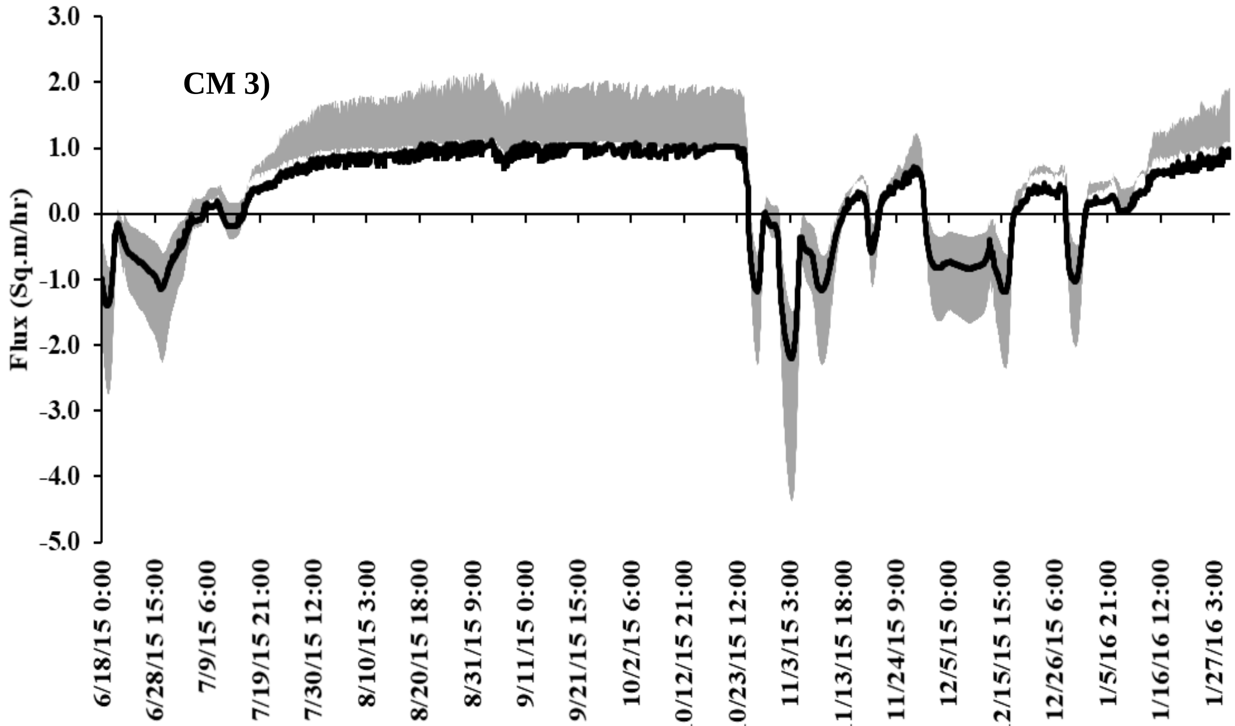


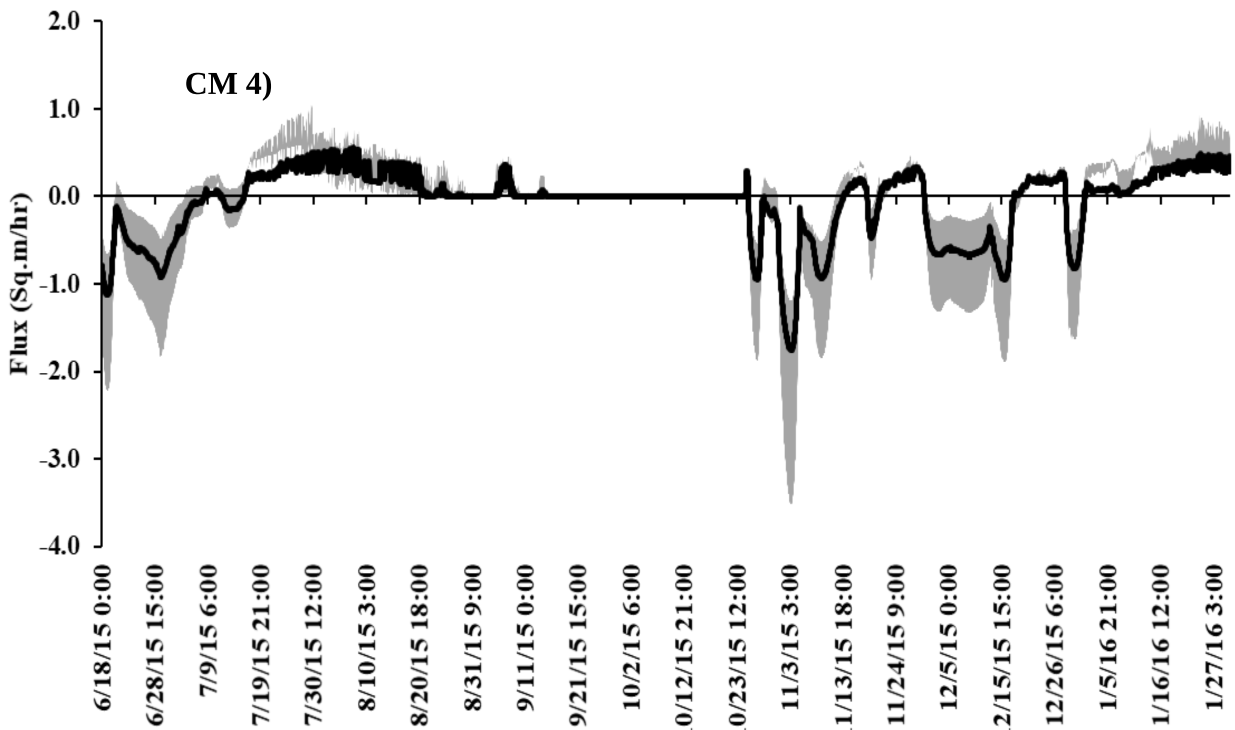
Figure S 3: Sensitivity of hydraulic heads to the river stage.

B.1.2.2 Fluxes


CM 1)
Flux (Sq.m/hr)
5.0
2.5
0.0
-2.5
-5.0
-7.5
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00


CM 2)
Flux (Sq.m/hr)
1.0
0.5
0.0
-0.5
-1.0
-1.5
-2.0
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00


CM 3)
Flux (Sq.m/hr)
3.0
2.0
1.0
0.0
-1.0
-2.0
-3.0
-4.0
-5.0
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00


CM 4)
Flux (Sq.m/hr)
2.0
1.0
0.0
-1.0
-2.0
-3.0
-4.0
6/18/15 0:00
6/28/15 15:00
7/9/15 6:00
7/19/15 21:00
7/30/15 12:00
8/10/15 3:00
8/20/15 18:00
8/31/15 9:00
9/11/15 0:00
9/21/15 15:00
10/2/15 6:00
10/12/15 21:00
10/23/15 12:00
11/3/15 3:00
11/13/15 18:00
11/24/15 9:00
12/5/15 0:00
12/15/15 15:00
12/26/15 6:00
1/5/16 21:00
1/16/16 12:00
1/27/16 3:00

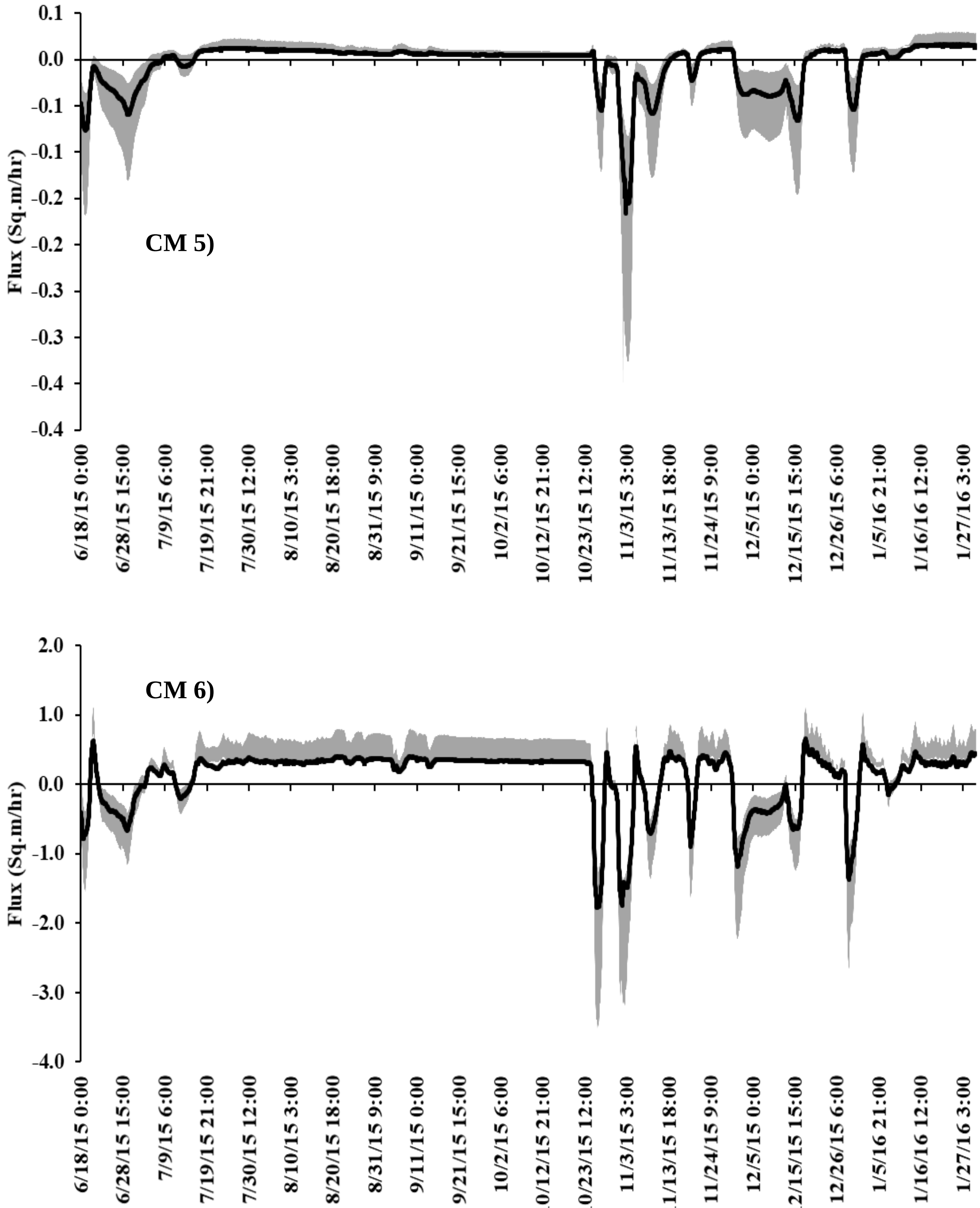


Figure S 4: Sensitivity of discharge and recharge fluxes to the river stages.