ACCOUNTING FOR STREAM BANK STORAGE FOR A SEASONAL GROUNDWATER MODEL

by

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ABSTRACT

One of the main sources of water in the semi-arid and arid region of the world is flood driven recharge. In recent research on groundwater and surface water interaction, attention has focused on the study of water exchanges between the near-stream aquifer and stream. One of the important near stream processes is bank storage. During flood events, there is a hydraulic gradient from stream to groundwater, which induces a net flux into the aquifer. This water is known as “bank storage”. This water will slowly release back to the stream when the stream water level drops and the gradient is towards the stream.

The aim of this thesis is to document the procedure required to develop a bank storage model that can be linked into a MODFLOW groundwater model. For this purpose a three dimensional, three-season groundwater model was built for the hypothetical Dry Alkaline Basin. A MATLAB code that can simulate bank storage process was developed. These two models were linked through the well package of MODFLOW and water was routed through the SFR package.

Different stage hydrograph scenarios were generated to simulate the effect of bank storage on groundwater. The results of this study indicate that the number of stage rise and shape of stage hydrograph entering to stream system, when they have the same average stream stage, produced similar net flux of water between surface water and groundwater. In addition, the results show that reaches, which were gaining during normal flow of the stream network, can become a losing stream during high flow periods. This flood recharge process can be a key to evaluating the ecological structure of stream systems and for stream-restoration and riparian-management efforts.

1-INTRODUCTION

Riparian ecosystems are valued for their role in sustaining biodiversity, improving water quality, retaining sediment, and providing a bird migration stopover, all of which hinge on sustaining flow regimes (Leenhouts et. al, 2005, Pool and Dickinson, 2006, Stromberg et. al, 2010). The services from riparian ecosystems in arid and semi-arid climates derive from the presence of water and a web of physical, biological and human processes. Riparian ecosystems are influenced by climate change, and variability through non-linear hydrologic interactions, and groundwater pumping. Riparian ecosystems in the southwest United States rely on shallow groundwater for phreatic vegetation and base flow for aquatic plants (Baillie et. al, 2006). The impact of floodwater infiltration on riparian groundwater during floods and flood recession can have a large impact on both the quantity and quality of river base flow and riparian groundwater.

The geochemical signature of floodwater is found both near and distant from the river (Baillie et al, 2007). Hydrologists use the term “bank storage” for the volume of water that is stored during floods and released during stream flow recession. The aquifer near the river is recharged during high summer monsoon flows and these waters are released back to the river during lower flow conditions (Scott Simpson, 2007). Chemical and isotopic composition indicates that riparian groundwater with a distinct component of flood recharge during the summer monsoon can be detected at great distances from the river edge long after flood waters recede (Baillie et al., 2007).

An aim of this study is to link a representation of bank storage to a groundwater model. Groundwater and surface water commonly form a linked system. However, they are often studied in isolation of each other, even computer models such as MODFLOW for groundwater modeling or HEC-RAS for surface water are written to count groundwater system or surface water respectively as the main system of concern, with minimal reference to the other (Carolyn Dragoo, 2004). In recent years there have seen many efforts to incorporate and generate models linking surface water and groundwater (Sophocleous, 2002; Dragoo, 2004; wake, 2008; Valerio, 2008). In groundwater models, surface water is linked to groundwater by using one of the stream packages (e.g., River package and SFR package). In this approach there is an assumption of uniform vertical flux between streambed and aquifer over a given section of the stream (i.e., reach). However, water is exchanged laterally between groundwater and surface water due to the bank storage process. In addition, stream packages have limitations related to the fact that these packages are designed to model long-term changes that occur from months to hundreds of years. In fact, these packages are not designed to simulate exchange of water between stream and shallow groundwater for a short period of time, which a flood by definition is a short duration event. Thus, an approach is needed to simulate the amount of water exchanged between surface water and ground water. Such an approach might involve calculations outside a groundwater model and then link to the groundwater model through use of a MODFLOW package that can simulate a specified flux.

Intense precipitation and the limited infiltration capacity of desert soils results in floods. Improved understanding of groundwater-stream water interaction during these floods and ensuing base flow conditions is necessary in order to understand hydrology, water quantity and quality of arid and semiarid rivers. Streams can be both gaining and losing with respect to their surrounding aquifers (J. Wake, 2008). Water stage rise due to flood events can convert a gaining reach of a stream to a losing one and increase the flux of water recharging into the aquifer, in the losing reaches along the river. The effect of flood driven recharge in different seasons and different locations implies that floodwater infiltrating during a flood event can have significant effects on quantity of stream base flow and riparian groundwater throughout the basin, for all seasons.

The magnitude and route of the flux exchanged between surface water and groundwater is determined by the hydraulic gradient between the river and the underlying aquifer (Rassam, 2011). A key factor that shapes hydraulic gradients between a river and groundwater is the stage of water in the stream. Different stage hydrograph characteristics (i.e., shape, number of peak, same volume of water with different stage hydrograph and average stage rise) effects should be investigated on the volume of water exchanged between surface water and groundwater.

In order to protect, save and maintain current biodiversity of riparian area species, better understanding of the hydrologic processes of the system is needed. To this purpose, a model that can simulate bank storage during the rise of water due to flood and decreeing recession was created. The goal was to link this model with a MODFLOW groundwater model. A bank storage package would introduce a new tool to link surface water and ground water and address questions such as:

1) How to distribute additional water due to flood recharge into a three-season groundwater model?

2) What is the effect of flood driven recharge on groundwater?

3) Is stage hydrograph shape and number of stage rise peaks important in affecting the distribution of water between groundwater and surface water?

4) What is the effect of the same volume of water entering to stream, with a different stage hydrograph, on the flux of water exchanged between stream and groundwater due to bank storage processes?

2-METHODOLOGY APPROACH

A surface water model and a groundwater model are needed in order to create a tool that can simulate the effect of flood driven recharge to the near stream, water storage and then rerelease water during the low stage conditions. A daily, river-aquifer model known as STAQ developed by Simpson (2007) was used as the surface water model in this research. Based on the STAQ surface water model, three stream locations were selected for this study, a losing stream, a gaining stream and a neutral one. Parameters related to surface water (e.g., transmissivity, diffusivity, aquifer storativity and stage-discharge curves) were transferred from STAQ to the bank storage model developed in this study.

A three-season groundwater model for a hypothetical basin known as the Dry Alkaline valley was used to simulate the effect of flood driven recharge on a semi-arid unconfined aquifer. The Dry Alkaline Valley is over 518 km^2, and consists of 12 rows and 20 columns with a uniform cell size of 1610 m on a side. This hypothetical model was previously used in Ajami et al. (2011). The aim of this model was to address questions of linking the bank storage to an aquifer-stream connected system. Although the model is simple, the goal was to represent the important flood driven recharge processes in a semi-arid aquifer. The model has three seasons, which are pre-flood season (dry summer season), flood season (wet summer season) and post-flood season (winter season). Most of the annual discharge of streams in southern Arizona occurs during the summer due to short-duration, high intensity events characteristic of the North American Monsoon. This same situation will be true in the hypothetical Dry Alkaline valley studied here.

STR-5 software (Maddock and Knight, 2011), which is software that assists groundwater modelers in creating the SFR package, was used to create SFR package inputs for Dry Alkaline valley groundwater model. This model contains one losing, one gaining and one neutral stream segment. Along the stream network, the riparian ecosystem was simulated using the EVT package. Winter precipitation, which is less intense with longer storms, contributes the majority of annual mountain-front recharge (Wahi, 2005). Mountain front recharge (MFR) was simulated as a constant flow in the current model. The well package was used to simulate MFR.

Bank storage processes were added to the Dry Alkaline valley groundwater model with use of the well package. The magnitude and direction of the exchange flux between surface and groundwater is mainly determined by the hydraulic gradient between a river and the underlying aquifer (Rassam, 2011). A form of the Darcy equation that can calculate the amount of water exchange laterally between groundwater and surface water due to difference between surface water stage and groundwater stage was used. Finally, the result of the bank storage package was compared to the SFR package result.

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3-FLOOD DRIVEN RECHARGE BACKGROUND

3.1 Background

Recharge in semi-arid basin and aquifer systems (e.g., San Pedro and Tucson basins) does not happen all year long and in all location inside these basins. Recharge is usually viewed as the sum of four distinct processes known as mountain front recharge, mountain block recharge, diffuse recharge and ephemeral channel recharge (Figure 1) (Philips et al., 2004). Because of Arizona's arid and semi-arid climate, on average, recharge to groundwater is estimated to be 2% to 3 % of average annual rainfall (Uhlman, 2005).

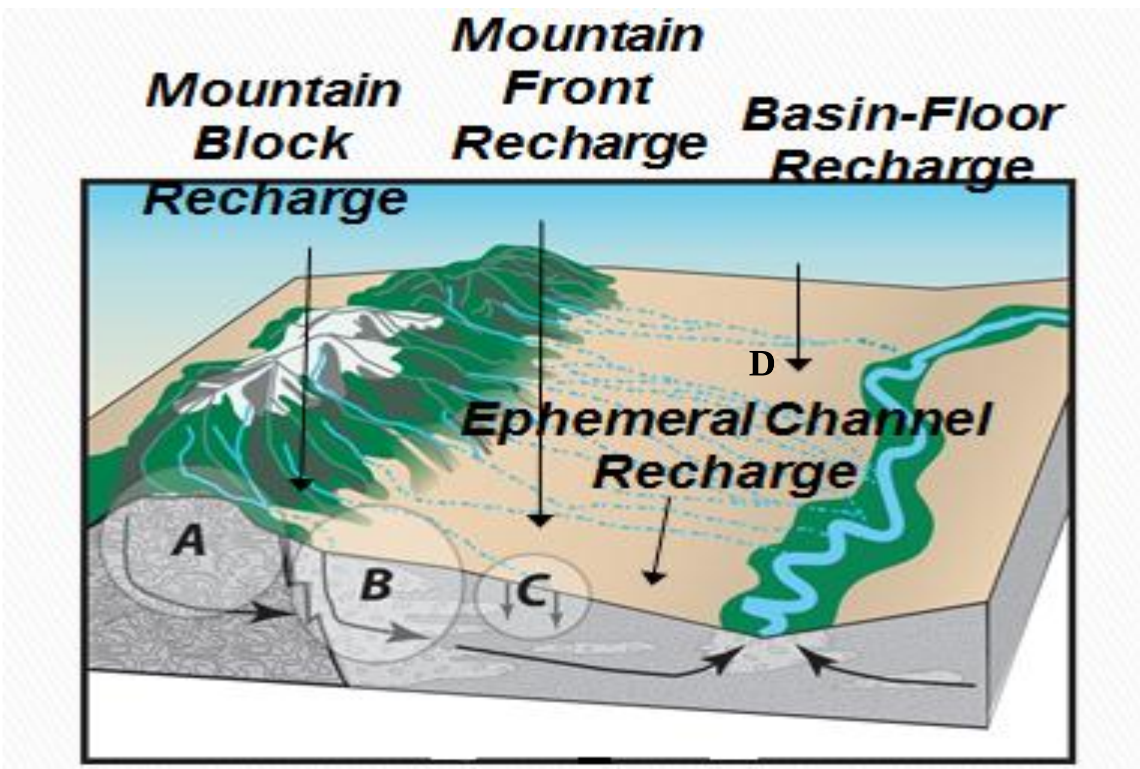

Figure 1: Four distinct recharge process in semiarid regions, A-Mountain block recharge, B- Mountain front recharge, C-Ephemeral channel recharge, D- Basin floor recharge.

Mountain block recharge is viewed as the water that recharged through the bedrock of mountains into the aquifer (Figure 2). Mountain front recharge is the contribution of mountain regions to the recharge of aquifers in adjacent basins (Figure 2). Water can be recharged to the watershed groundwater aquifer from the saturated zone under the mountains and through the unsaturated zone at the mountain front. MFR is an important source of recharge to basins in arid and semiarid regions (Wilson and Guan, 2004; Wahi, 2005). This importance is because more rainfall, due to the orographic effect, falls in the mountains than on the basin floor. In addition, the fractured rock of the mountain and the coarser grained materials along the margins of the alluvial basin allow water to infiltrate rapidly. Combined with lower temperature, lower evaporation, and thinner soils, that reduce loss of water, so, more water is available to recharge to basin groundwater. MFR is directly related to precipitation (Dragoo, 2004). A portion of water from precipitation that falls over mountain finds its way through alluvium, to recharge basin groundwater.

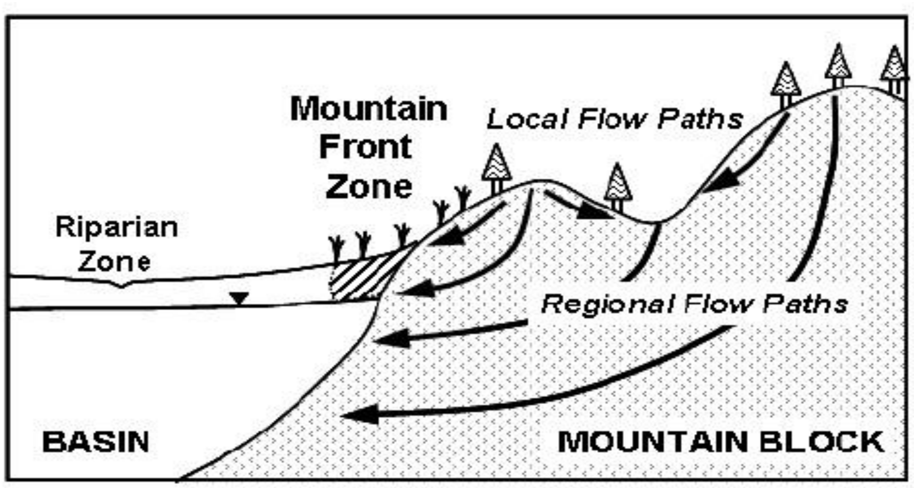

Figure 2: Cross section diagram show flow path of MFR and MBR (Wilson and Guan, 2004).

Intense storms and the limited infiltration capacity of desert soils results in overland flow, near river flood recharge, and ephemeral channel recharge. During flood events, the stage in the river increases which creates a local gradient away from the river. These gradient changes, cause water to infiltrate into the aquifer (M. Baillie, 2005). Previous studies in semiarid regions indicate that 10% of stream flow becomes recharge in an ephemeral stream (Izbicki, 2002, Phillips et al., 2004). Near river flood recharge, provides an important water source for riparian vegetation and wildlife along the river (Morin et al., 2009).

Diffuse recharge is defined as water added to the groundwater reservoir in excess of soil-moisture deficits and evapotranspiration on the basin floor, by direct vertical percolation of precipitation through the unsaturated zone (M. Sophocleous, 2000). Diffuse recharge is spatially distributed and results from widespread percolation through the entire vadose zone. Since, in the semi-arid regions of southern Arizona, mean annual precipitation is less than mean annual potential evaporation, the amount of diffuse groundwater recharge is considered a small component of total recharge to the basin (Scott et al., 2000), thus, in this study we assume that there is no diffuse groundwater recharge.

3.2 Flood driven recharge

One of the main sources of water in semi-arid and arid regions is flood driven recharge (E. Morin et al., 2009). Floods originate as snowmelt or intense precipitation events lasting for several days. Flood events can have damaging effect such as destroying infrastructure near streams, causing erosion, and changes in river morphology, but floodwater can be a valuable water source. Flood driven water has an important role in hydrological processes in the riparian area of rivers (Simpson, 2011). The hydrograph characteristics of stream flows show streambed infiltration losses that result in groundwater recharge (Parissopoulos and Wheater, 1991). Floods have a significant role in long-term geomorphology, hydrology and ecology of streams (Cooper and Rorabaugh, 1963).

Although flow in semi-arid streams is rare, the floodwater infiltrated contributes some of the water necessary for maintaining human settlements, riparian vegetation and wildlife along rivers (E. Morin et al., 2008). Flood recharge happens during monsoon season in the southwestern United States and is important for base flow and riparian groundwater (Baillie et al., 2007; Simpson, 2011). The geochemical signature of flood driven water is found in both near and distant from the river. Riparian areas are impacted hydrologically and geochemically by floods with different sizes and duration (Simpson, 2011). Simpson (2011) in his study indicated that the summer monsoon with its intense convective thunderstorms that create flash floods, accounted for 70% of recharge to the riparian aquifer of the San Pedro River.

Flood hydrologic and geochemical composition can influence riparian areas for a long time after floods recede. Simpson (2011) indicated that larger floods result in much more flood recharge and longer residence time in the groundwater system. Chemical and isotopic composition indicates that riparian groundwater with a distinct component of flood recharge during the summer monsoon can be detected at great distances from the river’s edge long after flood waters recede, providing support for the importance of floodwater (Baillie et al., 2007). Baillie (2005) in his study stated that maintaining flow of monsoon floodwater to riparian areas is important in order to maintain health of the riparian area because 45% to 100% of base flow in the river is from monsoon flood driven recharge.

An increase of high intensity precipitation could be one of the results of climate change (Allan and Soden, 2008; Meehl et al., 2007). More intense precipitation should result in more flash floods. With climate projections indicating decreases in annual precipitation in the southwestern United States (Seager et al., 2007), and persistent droughts (Karl et al., 2009), flood driven recharge will be increasingly important for water resources management in the southwestern United States.

Streams and groundwater interact in distinctly different ways during flood versus base flow periods (Simpson, 2007). The rise of floodwater not only maintains losing segments of a river, but also can make gaining sections become losing ones, resulting in changes of flow to be from the river into the riverbed and resulting in groundwater recharge. Water stored in the stream bank during rise in stream level is known as bank storage (H. Li et al., 2008; Cooper and Rorabaugh, 1963; Pinder and Sauer, 1971). For the San Pedro, riparian groundwater in gaining reaches is almost entirely based on basin groundwater, whereas losing reaches are dominated by prior stream flow. This condition indicates the important role of floodwater infiltration during high flow for riparian areas. Indeed, data collected along the San Pedro River suggests the role of flood recharge as a post-flood nutrient source for riparian areas (Meixner et al., 2007).

3.3 Interaction between groundwater and stream water

Groundwater and surface water are not isolated, they are hydraulically connected. The water flux between groundwater and stream water impacts the water balance of the stream and near stream associated groundwater and has an effect on water quality and the ecosystem health of near stream vegetation (Rassam, 2011). Streams that are connected hydraulically to an underlying aquifer act as a drain when groundwater level is high and act as a source when groundwater level is low. These processes control the elevation of the aquifer's water table (K. Blasch and T. Ferre, 2004). Groundwater extraction, aquifer recharge, bank storage and evaporation are processes that control the flux between groundwater and surface water.

If the surface level of water in a stream is lower than the water table of the aquifer, then there is a situation where water moves from aquifer into the stream, or a so-called "gaining stream"(Figure 3-A). If the groundwater table drops below the level of the river the flow of water is from the river to the aquifer; this situation know as a "losing stream" (Figure 3-B).

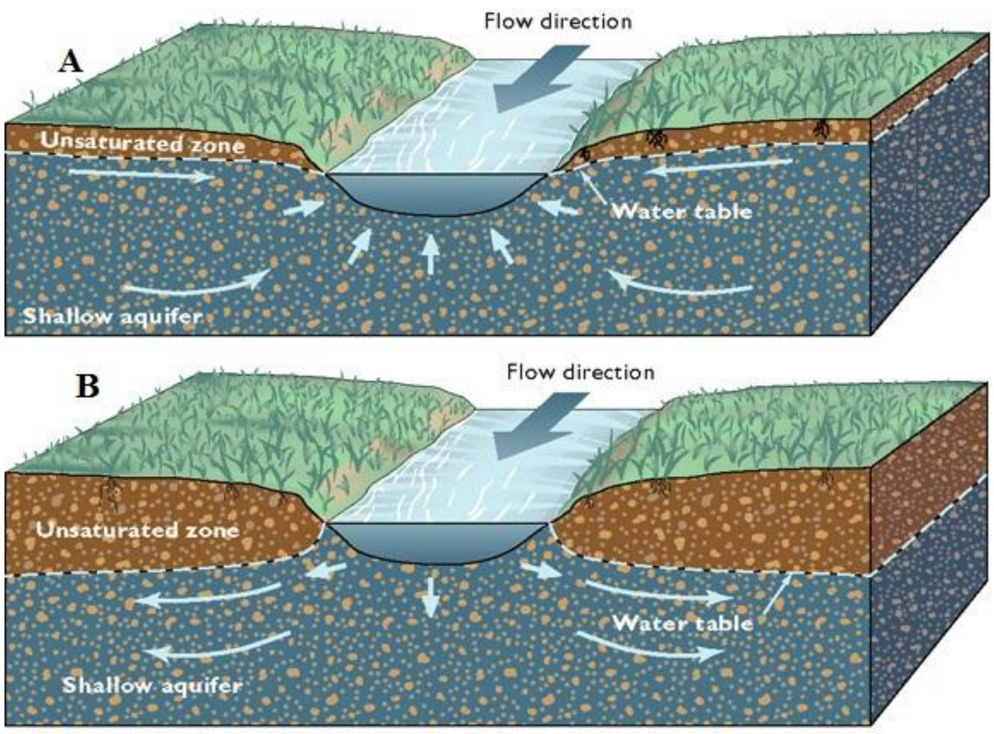

Figure 3: A- A gaining portion of stream. B- A losing portion of stream (Winter et. al, 1998).

These conditions can both occur in different sections of a stream. Streams and groundwater interact in distinctly different ways during flood versus base flow periods (Simpson, 2007). A given portion or stream reach can become a losing one during periods of high flow and flood, but become a gaining reach during low flow. In arid and semi-arid area streams, alternate gaining and losing condition are common (Rushton, 2007). These conditions can influence near stream riparian vegetation and wildlife. The direction of exchange of waters between the river/near-stream zone and the riparian aquifer is determined by river surface elevations and water table. The amount of water that contributes to the riparian system is related to the local occurrence of gaining versus losing river segment conditions. Losing segments along a river are more dependent on flood water than gaining segments because losing segments are the place along the river that result in groundwater recharge.

3.4 Bank storage

In recent research on groundwater and surface water interaction, much attention has been paid to water exchanges between near-channel and in-channel water. These waters are key to evaluating the ecological structure of stream systems and are important to stream-restoration and riparian-management efforts (Valerio, 2008). Near stream, aquifer systems are complex due to the difficulties associated with calculation of the volume of flows into and out of the aquifer, the complicated nature of groundwater and surface water interaction and the uncertainties associated with stream and aquifer properties (Rassam, 2011; Sophocleous,2010).

One of the most important processes in near stream hydrology is bank storage (Rassam, 2011). During flood periods, there is a hydraulic gradient in gaining streams from stream to groundwater, which induces a net flux into the aquifer (

Figure 4). This water is stored for a short period in the floodplain and is released back to the stream when the stream water level drops and the gradient is towards the stream (Rassam and Werner, 2008). Thus, bank storage is the water stored in the bank of the stream, due to stream level rise during a flood. This water can move laterally back to the stream during low flow and can move horizontally into the groundwater. Thus, bank storage is a dynamic process, which can recharge the groundwater aquifer during stream stage rises, discharge back to the river, and contribute base flow during low flow. The net flux that results from this phenomenon can cause an overall losing or gaining river condition (Rassam, 2011).

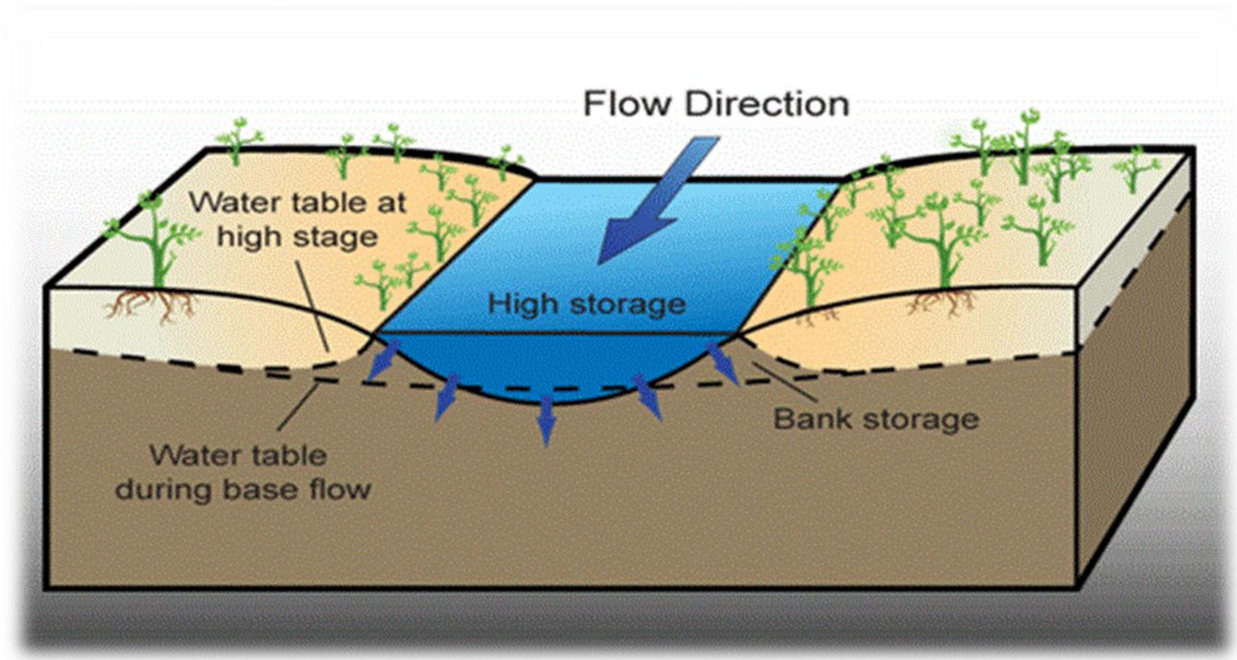

Figure 4: Rise of water due to flood event that make stream a losing one (*Commonwealth of Australia 2006*).

A quick stream rise could cause water to flow into an aquifer (Brater, 1940) and a considerable time may be needed to drain bank storage (Cooper and Rorabaugh, 1963). Cooper and Rorabaugh (1963) indicated that during a storm event, bank storage diminishes and delays flood peaks. In some studies (Chen et al., 2006; squillace, 1996) the bank storage zone includes areas adjacent and beneath the streambed (Figure 5). The amount of water that can be stored as bank storage depends on the stage of groundwater, stage of stream reach, hydraulic conductivity of stream bank materials and sufficient volumes of permeable bank material (Rassam and Werner, 2008). Riverbed conductance has a is a key parameter in reducing the propagation of flood waves entering into an aquifer (Rassam and Werner, 2008). Pinder and Sauer (1971) indicated that bank storage water volume depends mostly on the hydraulic conductivity of bank material and much less on alluvial aquifer width.

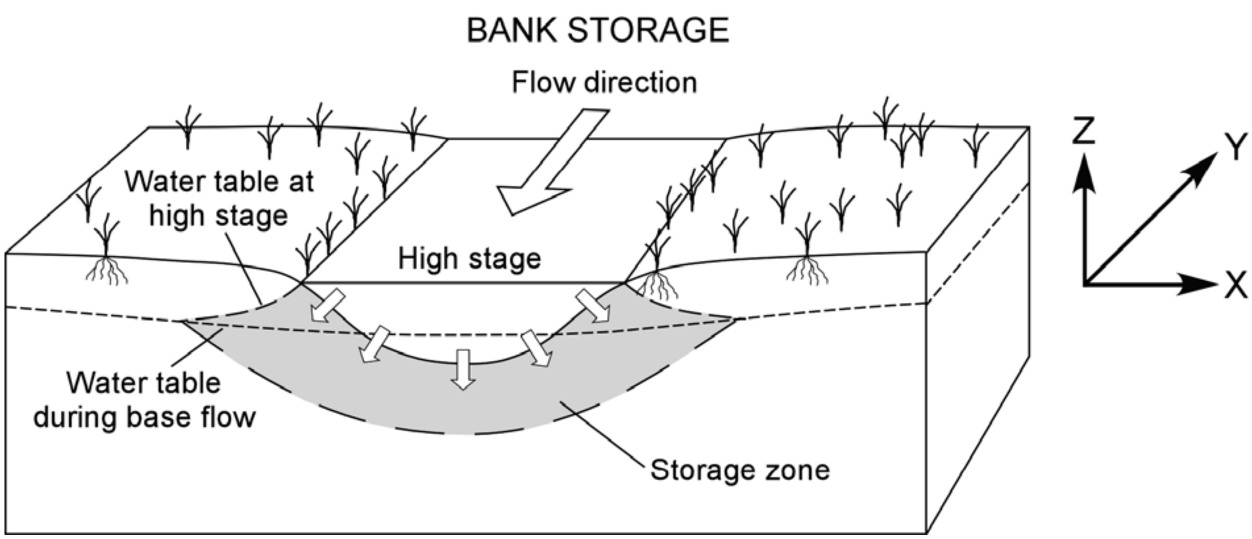

Figure 5: Bank storage zone (Chen, 2008)

Bank storage processes have been described using analytical solutions (e.g., Cooper and Rorabaugh, 1963; Rorabaugh, 1963; Moench et al., 1974; Morel-Seytoux, 1975; Dever and Cleary, 1979; Hunt, 1990; Hung and Zlotnik, 1999, Hantush et al., 2002)) or with numerical models (e.g. Chen and Chen, 2003; Lautz and Siegal, 2006; Li et al., 2008). By the early 1990's considerable literature had been developed on analytical solutions to bank storage interactions (Jolly and Rassam, 2009). In one of these studies, Hunt (2005) modeled a 2D analytical solution. The evaluation of his study indicated that the Dupuit solution approximation is more accurate for two-dimensional solutions for long domains with small anisotropy ratios. Knight and Rassam (2007) developed an analytical solution that the stream head fluctuations can be a random time series instead of having to be represented in a functional form, which must be evaluated numerically. In many analytical solutions, assumptions like the Dupuit-Forcheimer condition is made that can cause erroneous results (Sharp, 1997; Chen et al., 2006).

Numerical flow models were also used to simulate movement of bank storage water in the alluvial aquifer (Chen et al., 2008). Pinder and Sauer (1971) used a 2D numerical groundwater model coupled to a 1D surface water model to simulate how bank storage modifies the flood wave in a basin groundwater system. Lautz and Siegel (2006) used a 3D groundwater model with MT3D to simulate bank storage in a semiarid river. Li et al. (2008) simulated numerically a variably saturated, homogenous and anisotropic aquifer. They found out that surface water enters the stream bank more easily when the capillary effect is weak. Chen and Chen (2003) built a 3D groundwater model to illustrate the effects of the water exchange between a stream and aquifer, the volume of bank storage, and the storage zone. Their study revealed the effects of stream-stage fluctuation, aquifer properties, and hydraulic conductivity of streambed sediments, hydraulic gradients, and recharge rates on the bank storage zone. Chen and Chen (2003) also indicated in their study that water exchanged between the stream and the groundwater could cause intercontamination between stream water and groundwater. Therefore, any study on bank storage processes should be concerned with not only the rate and volume of stream infiltration into groundwater, but also on the volume of aquifer that has become dominated by stream water.

4-GROUNDWATER MODEL

4.1 Groundwater model approach

A simplified groundwater model for a hypothetical Dry Alkaline Basin was used in this study. This model was modified from a previous version used by Ajami et al. (2011). The aquifer is assumed homogeneous and anisotropic. The aim of this model was to create a hypothetical semiarid basin, three-season model that contains mountain front recharge processes, riparian area and a stream network. This groundwater model was built in two steps. First, a conceptual model of the groundwater system was developed containing the location and value of boundaries, stream, evapotranspiration (ET), wells and one layer and their parameters (e.g., hydraulic conductivity and strotivity). The units used in this study are, meter (m) for length and second (s) for time.

Second, a numerical approximation to the conceptual model was constructed. By converting the conceptual model, with all its conceptual coverage, onto a three-dimensional finite difference grid. This model has built using MODFLOW 2005. MODFLOW is a computer program developed by the U.S. Geological Survey (USGS) and numerically solves the three-dimensional ground-water flow equation for a porous medium by using a finite-difference method (McDonald & Harbaugh, 1988). MODFLOW computes the hydraulic head for each cell within a finite difference grid.

4.2 Discretization

4.2.1 Spatial grid

For the Dry Alkaline basin groundwater model, horizontal grid discretization spacing is 1610 m × 1610 m. This resulted in 12 nodes in rows and 20 nodes in columns for this model grid (Figure 6). The model has one layer, with top elevation at 1164 meter and bottom of this layer at 1005 meter, which resulted in a layer thickness of 159 meter. The basin is underline by a single unconfined aquifer with uniform hydraulic conductivity distribution equal to 0.0003 m/s. Specific yield of this model is 10^{-2}. The surface elevation of the model was used for starting head.

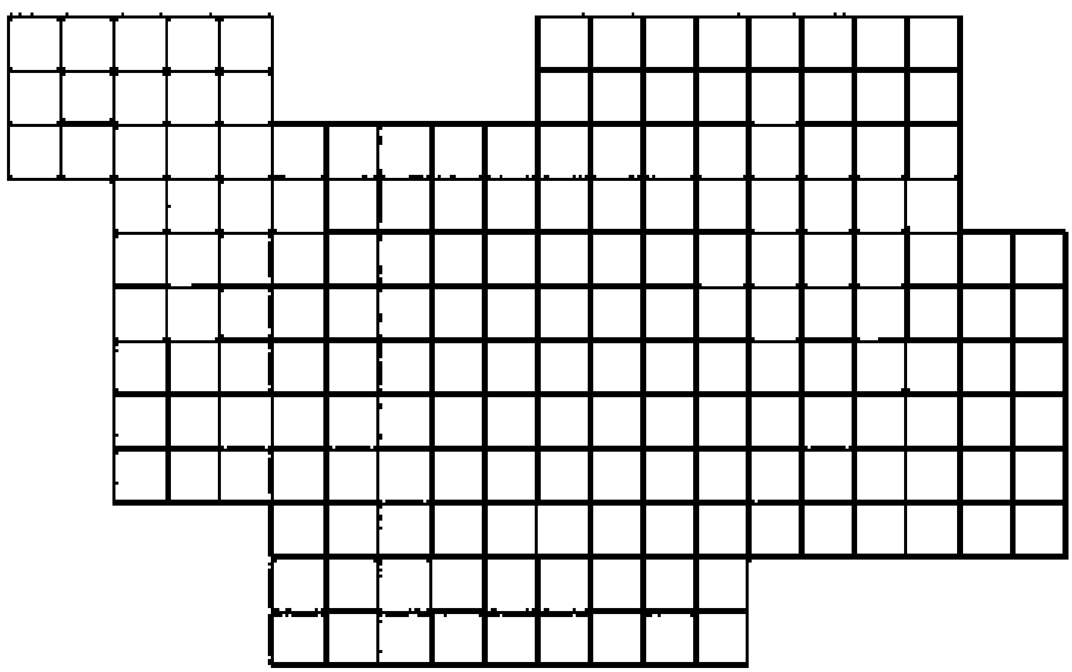

Figure 6: Dry Alkaline groundwater grid.

4.2.2 Temporal

One of the most important components of any hydrological study is the meteorology of the study zone. The Dry Alkaline Valley was assumed to have a climate similar to the San Pedro basin climate condition. Based on climate of semiarid location similar to the San Pedro, a three-season model was created. The rainfall patterns within this semiarid basin are bimodal, having summer and winter precipitation interrupted by spring and summer dry seasons. The monsoon precipitation is known for intense convective thunderstorms, with short duration rainfall (Goode and Maddock, 2000; Pool and Dickinson, 2007). The monsoon season is from July through September (Garfin et al., 2007). The North American Monsoon can bring torrential rain during these months (Baillie et al., 2007; Garfin et al., 2007). A second precipitation season from November through February (Garfin et al., 2007) has infrequent, low intensity multiple-day storms.

Based on this climate, a three-season groundwater model was constructed for the Dry Alkaline Basin for a period of 30 years. These three seasons are a wet summer season containing June, July and August, a winter season from September through February, and a dry summer season that contains March, April and May. Most of the annual discharge of the basin occurs during the summer due to short-duration, high intensity events characteristic of the North American Monsoon (Baillie et al., 2007; Garfin et al., 2007; Simpson et al., 2011). However, winter precipitation, with its less intense and longer storms, contributes the majority of annual mountain-front recharge (Wahi et al., 2008, Simpson et al., 2011). A three-season model allows for a better analysis of the various seasons' contributions to annual recharge.

4.3 Model Boundary

In general, a boundary value problem is designed to determine how the conditions that are imposed on the boundary of a domain affect the state of the system. To be more precise, a boundary condition is a fixed condition at a point for a fixed period.

The north and south boundaries of the model are bound by mountains and are considered no flow boundaries (Figure 7). A zero value in the IBOUND of the BCF package indicates that particular cells are assigned as no flow boundaries. The east and west boundary are treated as constant head boundaries.

One of the main recharge types in a semiarid area is MFR. Usually modelers treat MFR as a specific flux boundary condition. With a specific flux boundary condition, the internal discharge rate Q is specified, as opposed to the default Qs=zero. When representing recharge at a rate of N (L/T), a discharge equal to $N \times \Delta x \Delta y$ must be added to Qs in the uppermost saturated block of the model (Fitts, 2002). In this study, as the model domain includes only the basin fill and not the mountaintops of the basin, a constant flux boundary condition of $Qs > 0$ representing mountain-front and mountain block recharge was applied along the model boundary cells. Inside MODFLOW, the well package was used to simulate MFR as a specific flow condition. Six cells (Figure 7) on the northwest side of the domain contain wells. Based on Goode and Maddock, (2002) model result, recharge rate equal to 0.0082 m/day was assigned as MFR rate. MFR value was added to groundwater model by using the well package.

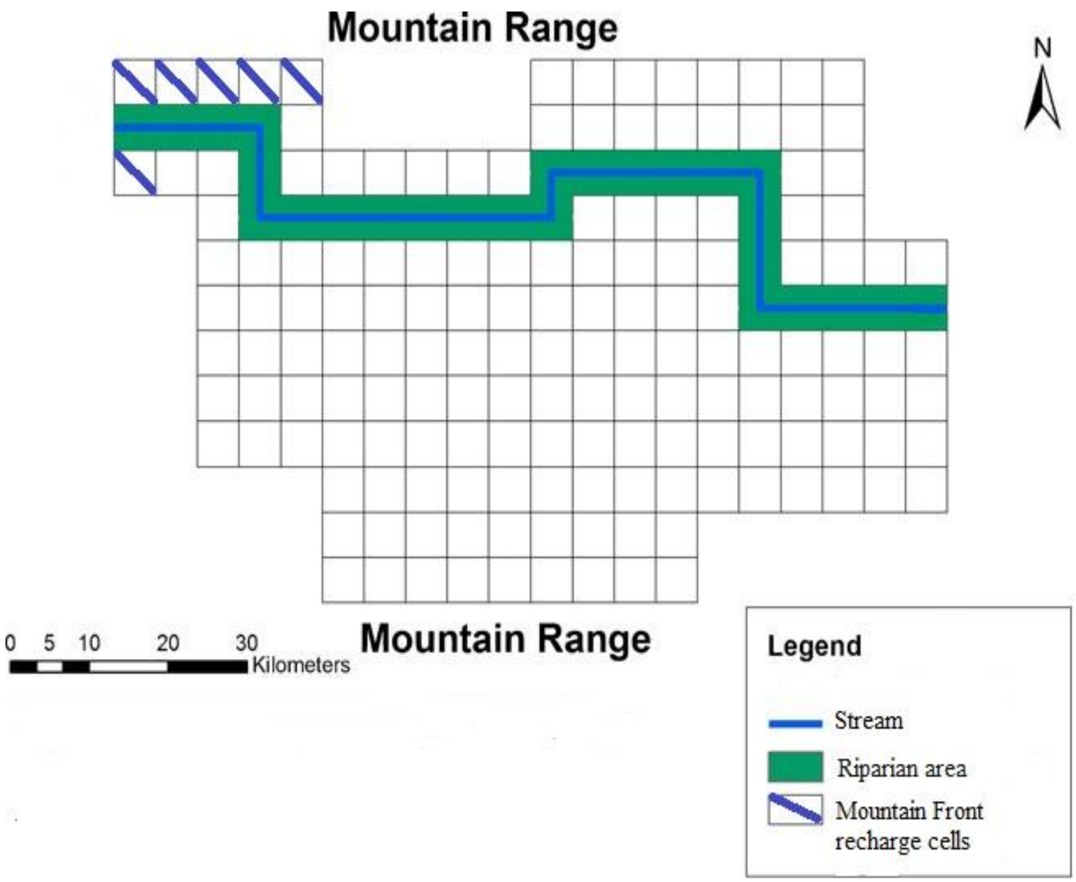

Figure 7: Location of mountain range and mountain front boundary (modified from Ajami et al., 2011)

4.4 Steady Oscillatory

When seasonality is a modeling issue, the use of a single initial hydraulic head and boundary conditions leads to water budget miscalculation. In this case, a type of groundwater solution known as steady oscillatory is used. Steady oscillatory is an intermediate solution between a steady state solution and a transient solution. In this type of groundwater solution, some or the entire variables change through cyclical stress prides, but repeat from year to year (Maddock and Vionnet, 1998). For determining the solution, the model was treated as a transient groundwater model, with imposed periodic boundary condition for a large number of stress periods (Maddock and Vionnet, 1998).

In order to build a steady oscillatory solution, in the current study evaporation and stream flow are the periodic values needed. Seasonality values for evaporation and stream flow were assumed by using factors of the annual average values of evaporation and stream flow into seasonal values. The product of a seasonal scale factor times the annual average rate provides the seasonal rate.

Maddock and Vionnent (1998) indicated that a numerical solution is achieved, when model processes through the cycles and the non-periodic portion are small. Finally, the periodic solution becomes more dominant. In this stage, the seasonal heads and flux are identical thru annual cycle. In the case of current study, steady oscillatory solution was achieved in the 25th year.

4.5 Evaporation Package

Simulating connections and effects between vegetation and groundwater are commonly made through use of the evaporation package for groundwater models. In this package, water is removed from each model cell that is assigned an ET value. In this study, the EVT package was used. It requires elevation, ET extinction depth and maximum ET rate values. When water table rises above the ET elevation, the evapotranspiration occurs at the maximum ET rate and when water table drops below the ET extinction depth, evapotranspiration stops. Between these two points, evapotranspiration from the water table varies linearly with water table elevation.

For this study, riparian vegetation along the stream was assigned as the cells where ET occurred (Figure 8). There are 26 riparian cells in the model domain. ET extinction depth values were used ET values similar to the Pool and Dickinson (2006) model for the Upper San Pedro River. The maximum ET rate (LT^{-1}) was the value that changed for each season. The lowest value was assigned for winter season (0.1E-30 (m/s)), and the highest value was assigned for summer season (1.6087e-9 (m/s)). The maximum ET rate for dry summer season was 8.8430e-10 m/s. These three maximum ET rates were constant for the entire 30 years of this model simulation.

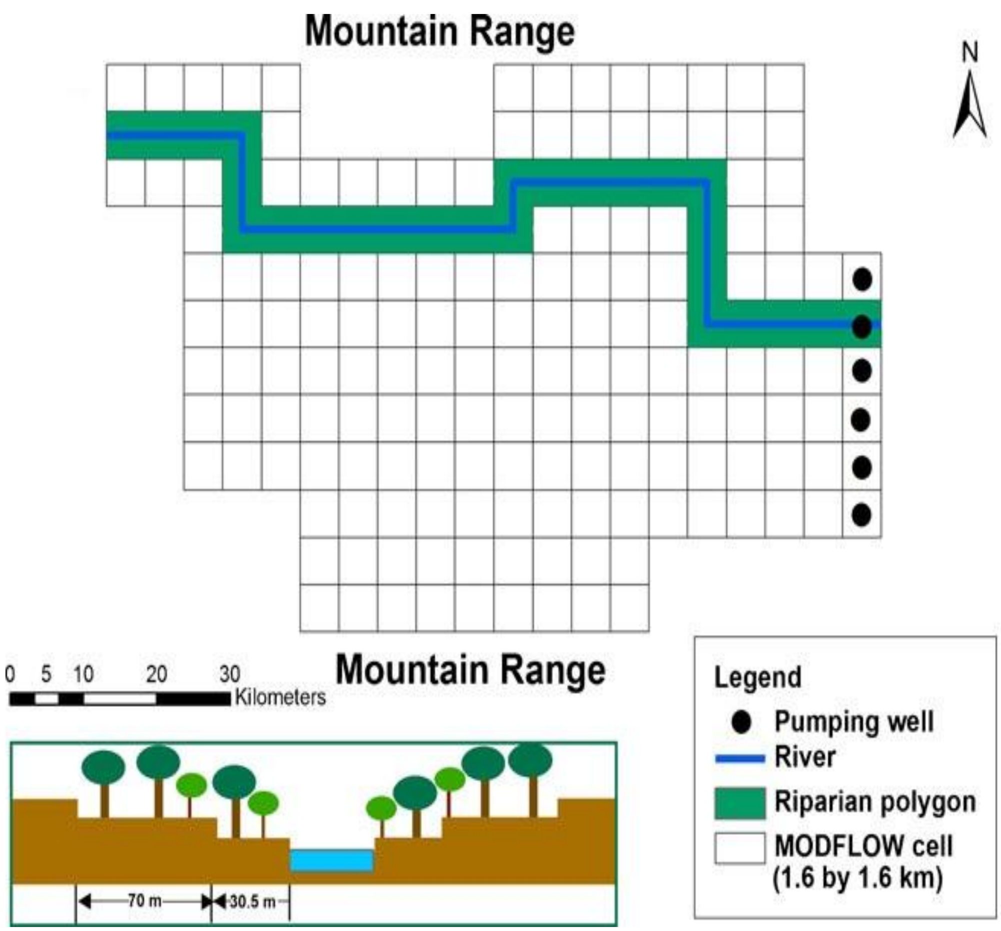

Figure 8: Aerial view of the hypothetical Dry Alkaline Valley and cross-section of the riparian area (From Ajami et al., 2011)

4.6 Well package

The well package was used to add bank storage water into the groundwater model. Each flux calculated by the bank storage model, was multiplied by the length for each reach, which is 1610 m for entire model reaches. This approach resulted in a specified flux for each individual reach in units of m^3/day. The sign of each Q must be designated. A positive value, indicated recharge. A negative value indicated discharge (pumping). Thus, wherever Q was negative water was moving from stream water to groundwater, and this value is positive for the groundwater.

Additional water flux caused by floods during the monsoon season was added to the next season, which is the winter season. Flood season duration for the basin was 90 days. Duration of winter season was 180 days. Thus, the flux of water needed to be divided by two, so that the same amount of water was stored in the system. As described a steady oscillatory solution is used to deal with the groundwater model. For this type of groundwater solution, the fluxes of water in the well package were set to be changed through each season, but repeated from year to year for all 30 years of the simulation.

In this study, the well package was used to simulate mountain front recharge (MFR) and as the outflow point of water from the basin. These outflow wells allow water to move out of the basin. These wells were located at the eastern boundary (Figure 8) and were assumed the same for all the seasons. As mentioned in “Model Boundary” section six cells (Figure 7) in the northwest side of the domain contain wells that simulate MFR.

4.8 Other packages

The Strongly Implicit Procedure (SIP) package was used as the solver package. SIP solves the finite difference equations in each step of a MODFLOW stress period (Harbaugh et al., 2000). SIP provides the solution algorithm for the steady oscillatory solutions. The maximum number of times through the iteration loop in one time step in order to attempt to solve the system of finite-difference equations is 500 for this study. Five iteration variables were used in this study. The SIP convergence factor was 0.001.

5 SURFACE WATER

5.1 STAQ surface water model

The surface water model used for one of the stage hydrograph scenarios and for the verification of bank storage, model code was the coupled hydrological-solute surface water model that Scott Simpson (2007) developed for the Upper San Pedro River known as STAQ. The main goal of this model was to define the amount of water exchanged between the river and riparian groundwater, and investigating the importance of monsoon floods for the riparian system in Upper San Pedro River. Simpson wrote the code in MATLAB software[1], based on water balance and solute balance of the region. In this model, time step is per day and for a period of 12 years from 1995-2007. In addition, parameters such as diffusivity and storativity were taken from the STAQ model in the current study. In one of the stage hydrograph scenarios of the current study, water stage data from the STAQ model was scaled to the Dry Alkaline basin. After running the code, and generating the output, MODFLOW stream package was used to link Simpson's surface water model to the groundwater model.

5.1.1 Model structure[2]

Stromberg et al. (2005) developed a model that classified hydrologic conditions along the San Pedro River. This classification was confirmed by M. Baillie's (2005) geochemical analysis. The vegetation map along the river was used by Scott Simpson to divide the river into nine segments. Each model segment represented a gaining or losing reach in each time step.

[1] Scott Simson,2007, appendix H: model code, p 127-148
[2] This section is based on Scott Simpson (2007) thesis.

Conceptually, this model can be considered a bucket model that contains two buckets, River channel and Riparian groundwater (Figure 9). Each model segment consists of these two reservoirs and water is exchanged between them.

The state of the riparian aquifer system controls the hydrologic response of the system to the change of the river discharge. Quantity and route of basin groundwater exchange to the riparian aquifer, stream to aquifer and groundwater losses to phreatophyte transpiration are the processes that affect riparian groundwater condition. This model permits gradient-based exchange of water between the stream and near stream aquifer. For exchange of water between the riparian aquifer and the groundwater basin, there is an assumption of a constant rate depending on direction of water movement in each segment.

In the model code after exchange of water between basin and groundwater, the water table in the riparian groundwater reservoir is recalculated for each segment. The main processes controlling the water balance of the stream and riparian groundwater are evaporation of stream flow, transpiration by phreatophytes and shallow water evaporation, basin groundwater exchange with the riparian aquifer and river-groundwater exchange.

The flux of water between the riparian aquifer and the river is driven by the elevation of the water table in RGW reservoir and the river water surface. River water surface is calculated based on the river bottom elevation and discharge curve specific for each river segment. The amount of water moved between the river and RGW reservoir is determined by an equation that is a form of Darcy's laws:

$$\boldsymbol{q} = \boldsymbol{T} \times \frac{\boldsymbol{dh}}{\boldsymbol{dl}} \text{(Eq. 1)}$$

Where:

q   is the volume of water gained/lost per meter of river length per day ($\frac{m^2}{day}$)

dh   is (groundwater elevation-river surface elevation) (m)

dl   is half the width of RGW reservoir (m)

When q is positive that section of river gains water from the near stream zone, and when q is negative that segment is losing flow to the near stream zone. In both cases the same amount of flow that is lost or gained from the stream zone, is lost or gained by the riparian aquifer.

Finally, for calculating the amount of water gained or lost along each segment, per-meter flux of water is multiplied by segment length. This calculation would give flow volume in m^3/day. The change in flow volume is then calculated and counted for flow entering the next segment.

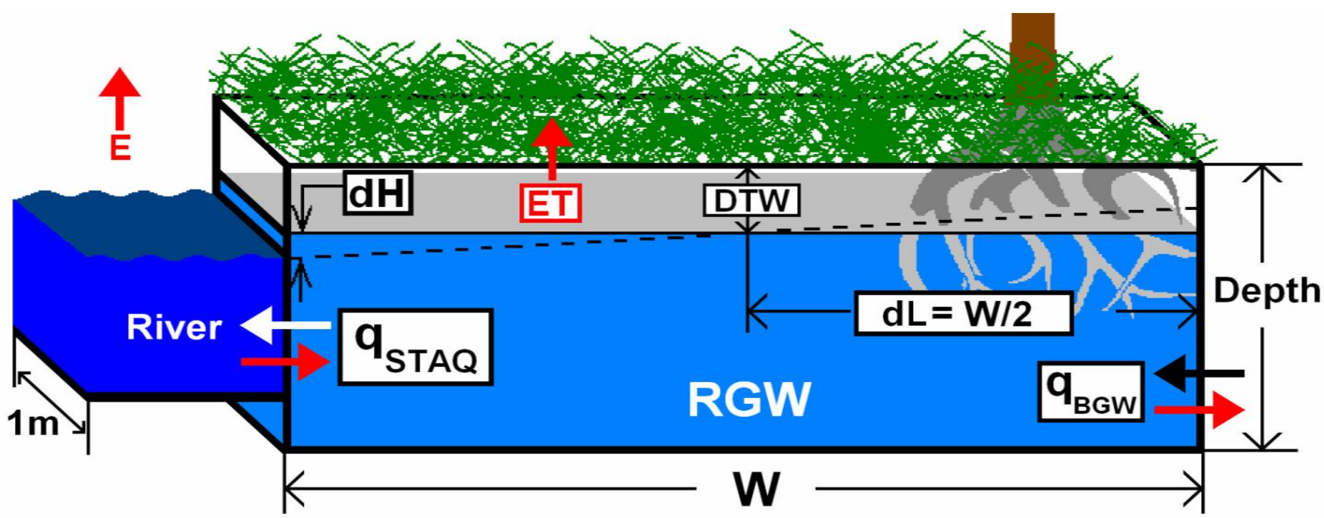

Figure 9: Conceptual cross-section of STAQ model (Scott Simpson, 2011)

5.2 Surface water routing

MODFLOW has three packages that can simulate river processes and link it to groundwater. These three packages are River package, STR package and SFR package. The purpose of these packages is to simulate the effects of flow between surface-water features and groundwater systems. In the current study, the SFR package was used to route water through the stream network of the Dry Alkaline basin. The SFR package was derived from the STR package. The SFR package is able to rout flow of water through the stream network (such as stream, tributaries, diversion and lake). The program limits the amount of groundwater recharge to available stream flow, allows the streams to go dry, and allows the streams to rewet (Prudic et al., 2004).

5.2.1 Creating SFR package

A network of channels along the Dry Alkaline River was simulated with the SFR package. The river is the primary drainage for the Dry Alkaline Basin (Figure 10). Dry Alkaline River is a west-to-east-running River. The basin aquifer and the river both discharge to the east side of the basin. Stream inflow is assumed periodic and changed for each season. These seasonal inflows are determined by factoring the annual average volumes of a steady state model for the Upper San Pedro (Pool and Dickinson, 2006).

Based on the STAQ model (Simpson, 2007) three stream locations were picked, a losing stream, a gaining stream and a neutral one. The aim was to use a simulated stream network similar to situation of a river in a semiarid area. Through this package stream, flow was introduced into the groundwater model.

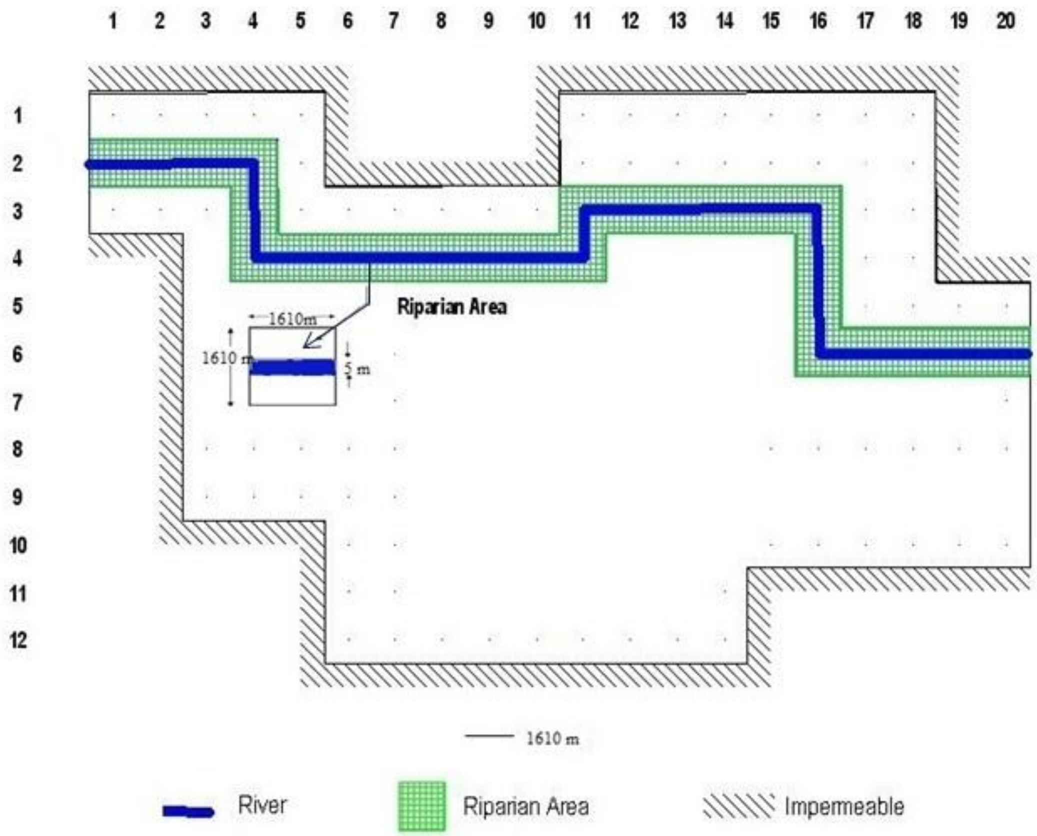

Figure 10: Stream location on Dry Alkaline Basin.

The SFR package requires a network of reaches and segments. The stream reach is a section of the stream network contained inside a grid cell of a MODFLOW model. A single cell may contain multiple reaches. A segment is made of groups of reaches that have the following four characters: 1- uniform rates of overland flow and precipitation. 2- Uniform rate of ET, 3- Linear hydrologic properties and 4-Tributary flow and diversions assigned to the first reach (Prudic et al., 2004). The simulated stream network was divided into 26 reaches depending on the length of the stream associated with a particular cell. These reaches were grouped into three segments. Segments were numbered upstream to downstream. The reaches are numbered from upstream to downstream for each particular segment.

After creating all the necessary input files, through using STR-5 Software (Maddock and Knight, 2011), SFR package for MODFLOW 2005 was created. STR-5 reads spreadsheet data for various inputs in order to create the SFR package. This spreadsheet data include Structure Names, River Cell Information, and Channel Geometry with parameters, ICALC Channel Geometry, Unique Stream flows, Specified Depth and Width, Main Inflows, Treated Effluent Outfalls, Diversions to Canals, Drains, Farm Delivery Requirements, Farm Module Requirements, and Stress Period Print Flags (Maddock and Knight, 2011).

5.2.1.1 River Cell Information

"River Cell Information" started with three columns containing information on, layer number, row number and column number. This value shows the location of each cell that contained a stream. The next column belonged to the number of stream segments (ISEC) that each of the reaches is located in. For the next column, a sequential integer value showing the reach number (IREACH) cell in the segment was assigned. This value starts from number one, in the upstream. Starting from upstream, numbering of segment and reaches inside each segment was observed by eye and added to the excel spreadsheet. Lengths of stream assume to be same as the length of each cell and equal to 1610 meter (Figure 10). By having layer, row, column, segment, reach and structure number for each of the river cells and the length of the river within the model, this data set was completed.

5.2.1.2 Channel Geometry

The Channel Geometry data set was used to input information about the geometry and hydraulic properties of stream segment into the SFR package. First column on this data set is segment number. Segment number is ordered from upstream to downstream.

Next value belonged to an indicator for stage calculation known as ICALC. For keeping this model simple, ICALC equal to one was used. Thus, Manning's equation was used to calculate water flow along the stream network, by choosing ICALC equal to one. In order to have ICALC equal one, a table containing information on roughness coefficient of streambed and stream width for first and last reach of each segment is needed (Maddock and Knight, 2011). For this study, the roughness coefficient was taken from Goode and Maddock (2000). All three segments of Dry Alkaline River were assumed to have a Manning roughness coefficient of 0.022. Stream width was assumed five meter for entire stream network.

Next column contains OUTSEG. This parameter is an integer value of downstream segment that receive inflow from the last downstream reach of the segment. OUTSEG value was assigned to each segment from upstream to downstream. OUTSEG value of the last segment of model assigns zero that means there is no downstream segment receiving inflow from this segment. IUPSEG column is the next column, which is an integer value showing the upstream segment from where water is diverted. The value of this column is assigned a zero because there is no diversion from the stream network for current study (Maddock and Knight, 2011).

Finally, step channel geometry data was calculated the streambed elevation and hydraulic conductivity for the upstream and downstream of each segment. In this study, streambed elevations were extracted for each segment upstream and downstream based on the elevation of the ground surface. The hydraulic conductivity is assumed to be 0.625m/day and was taken from Thomas Goode (2000).

5.2.1.3 Main Inflows

Flows into a segment from external sources such as dams or rivers crossing into the model area are known as the main inflow (Maddock and Knight, 2011). A main inflow chart is created for model segments that have non-zero flow values. Main inflow input, contains value of the structure name, which flow is released into it, the segment number for the release point and the unique stream flow values. Stream inflow in this study changes for each season. Inflow for winter season is 3.9 m^3/s, for dry summer is 1.89 m^3/s and for wet summer is 2 m^3/s.

5.2.2 Building SFR package with help of STR-5

After building input data set spreadsheet, STR-5 was used to convert the data set into the SFR package. STR-5 reads and creates spreadsheet data for various inputs required by SFR package. The CSV files were constructed in an analogous way. STR-5 allows constructing the CSV files externally to the program or within the program. STR-5 constructed the SFR MODFLOW-2005 data set once all the appropriate input data was entered into the software.

6 BANK STORAGE MODEL DEVELOPMENTS

The magnitude and route of the flux exchanged between surface water and groundwater is mainly determined by the hydraulic gradient between the river and the underlying aquifer (Rassam, 2011). The hydraulic properties and capacity of near stream soil is another factor that governs near stream water/aquifer flux exchange processes.

A form of Darcy's equation previously used by Simpson, 2007, was used in this study. The bank storage model allows gradient-based transfer of water between stream and groundwater. The difference between groundwater stage and stream stage produced the gradient that causes water flux exchange between groundwater and surface water. Transmissivity (T) is a parameter counting for near stream soil hydraulic properties.

$$\boldsymbol{q = T \times \frac{dh}{dl}} \quad \textbf{(Eq. 2)}$$

Where:

q is the volume of water gained/lost per meter of river length per day ($\frac{m^2}{day}$)

T is transmissivity (L^2T^{-1})

dh is difference between groundwater elevation and river surface elevation (m)

dl is half the width of riparian area that water can exchange between groundwater and surface water (m)

The stream is assumed to penetrate the full thickness of the aquifer and has a rectangular shape. Thus, the stream bank was perpendicular to the streambed. In addition, the stream is assumed to be connected to the groundwater system in order to making the calculation simpler. In addition, overbank and vertical infiltration that may occur during floods was neglected because overbank events are relatively rare (Simpson, 2007).

6.1 Transmissivity

Transmissivity in the Darcy equation accounts for the impact of near stream soil properties on the water flux exchanged between an aquifer and a stream. The transmissivity is a parameter that measures how much water can be transmitted between streams and aquifers. Transmissivity is helpful in predicting the response of an aquifer system to stream water fluctuations (Pinder et al., 1969). Cooper and Rorabaugh (1963) used transmissivity to solve their analytical solution for groundwater movement and bank storage due to a flood. Transmissivity in the current study is based on the aquifer diffusivity calculated by using the iterative curve matching method developed by Pinder et al. (1969) $\boldsymbol{T}=\boldsymbol{D}*\boldsymbol{S}$ (Eq. 3)

$$\boldsymbol{T} = \boldsymbol{D} * \boldsymbol{S} \text{ (Eq. 3)}$$

Where:

T is transmissivity [L^2T^{-1}].

D is Diffusivity, which is the ratio of transmissivity to storage coefficient [L^2T^{-1}].

S is the dimensionless aquifer storativity that for unconfined aquifer, specific yield is used.

Specific yield values can be determined from an aquifer test used in conjunction with physical boundaries of the aquifer in the groundwater model (Pinder et al., 1969). In this study, based on Roeske and Werrells (1973) estimation of specific yield 0.05 to 0.25 for the Upper San Pedro basin, specific yield equal to 0.1 and .16 was used.

Diffusivity (D) is the ratio of transmissivity to storage coefficient. Diffusivity value, from the STAQ model (Simpson, 2007) was used in this study. Simpson used Pinder et al. (1969) method to calculate Diffusivity. Pinder et al. (1969) proposed their method to determine diffusivity based on the comparison between observed changes of groundwater head in response of the aquifer to fluctuation in river stage, with theoretical head value of stage fluctuation in river.

6.2 Water gradient term

The term $\frac{dh}{dl}$ in equation 7, represent the head change in the direction of l (half the width of riparian area). Head h and l coordination both have a length unit, so $\frac{dh}{dl}$ is dimensionless. This gradient consists of three terms groundwater elevation, stream water elevation, and the width of the riparian area.

The Dry Alkaline groundwater model calculated groundwater elevation was used for this study. Target season of the model is wet summer. Groundwater head elevation of previous season, the dry summer season, was used as groundwater stage elevation. Therefore, there is an assumption of constant groundwater head elevation throughout the target season. This assumption is because groundwater travel time is slow and can range from days to hundreds or thousands of years (Fitts, 2002).

The stage of water was needed for the bank storage model. In the case the surface water model gives flow discharge, curves that generate river stage from daily discharge can be used (Figure 11). The product of adding stage (m) value, to the riverbed elevation was the surface water stage elevation, which was used for the bank storage equation.

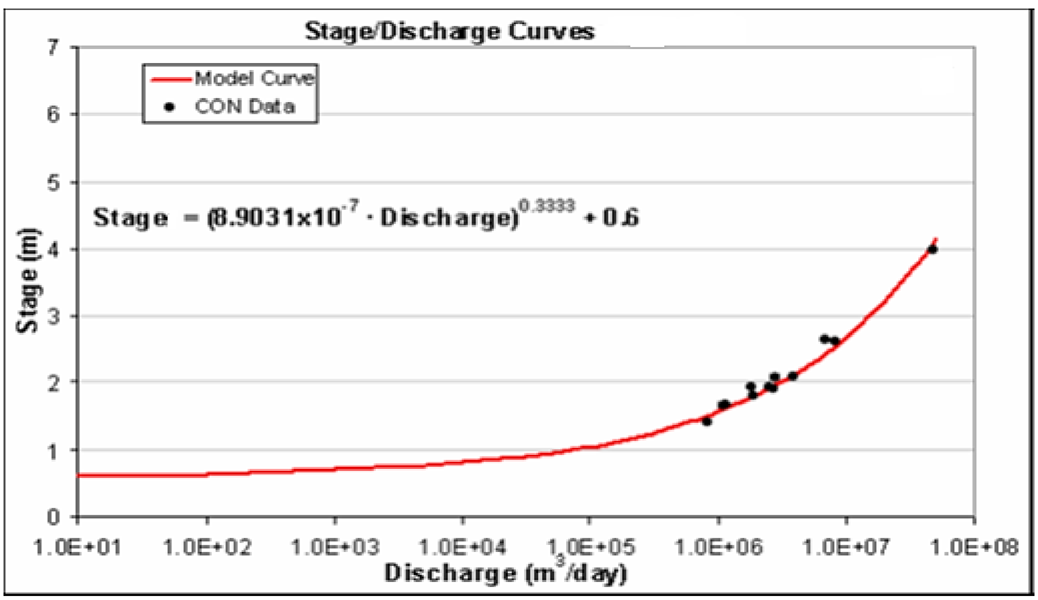

Figure 11: Model curves for generating river stage from daily discharge (Simpson, 2011)

The current study was for a hypothetical basin. Surface water stage was simulated using hypothetical surface water stages. A symmetric stage hydrograph for a maximum flood rise of 2.5 meter for a period of 90 days is shown (Figure 12).

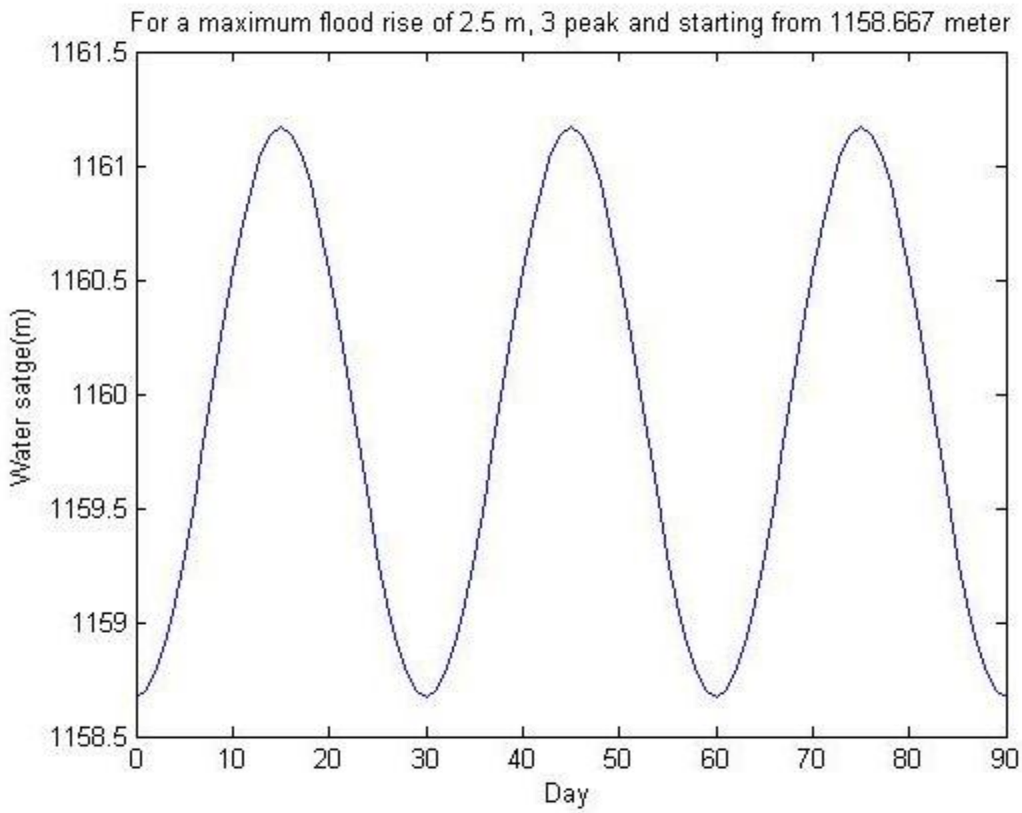

Figure 12: Stage hydrograph generated to simulated flood wave.

Based on the riparian definition which refers to transitional areas between terrestrial and aquatic ecosystems that depend on the existence of free flowing stream and/or shallow groundwater table (Figure 13) (Leenhouts et. al, 2005), an assumption made that dl is equal to the half of width of riparian area. The reason of assumption is to simplify the system. Simpson, 2007 indicated that this simplification can describe the overall process of water exchanged with near stream groundwater. In the STAQ model, width was allowed to change from 60-300 meters. This value is based on Scott et al., 2005 study on range of floodplain width along the San Pedro River.

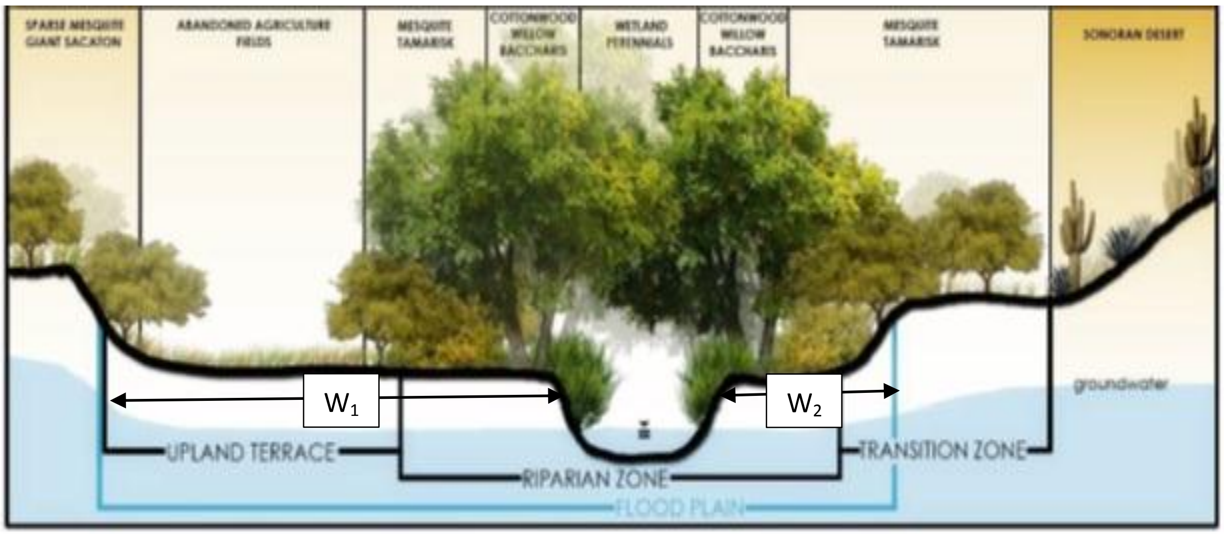

Figure 13: Cross section view of riparian area along stream and the width of flood plan where water can exchange between groundwater and surface water. W_1 is width of a cottonwood and mesquite dominant riparian region and W_2 is Sonora desert dominant riparian area (modified from: Shannon Hatch, 2011)

6.3 Output of model

Output of Darcy equation (Eq. 1) is the volume of water gained or lost per meter of river (q). When groundwater level was lower than surface water elevation, q was negative and water was lost to near stream groundwater. Vice versa, when groundwater level was higher compared to surface water level that section of stream gained water from groundwater and q was positive.

The per-meter flux of water into or out of the river (q) is multiplied by the reach length (l=1600 (m)) to give the flow volume (m3/day) gained or lost along the entire reach (Eq. 10). In a given system, a reach may be losing during high stream flow events and became gaining as flow decreased. This additional water flux caused by floods was added to groundwater for the next season with the use of the Well package.

$$Q = L_s \, . \, q \text{ (Eq. 4)}$$

Where:

L_s Length of stream segment (m)

q is the volume of water gained/lost per meter of river length per day ($\frac{m^2}{day}$)

6.4 Stream packages vs. Bank storage package

Stream packages (SFR, STR) and river package use the same assumption that measurable head losses between the stream and the aquifer are limited to those across the streambed. For the purpose of flux exchange between surface water body and aquifer, these packages use a form of Darcy low equation (Prudic et al., 2004):

$$q = \frac{K * w}{m}(h_s - h_a) \text{ (Eq. 5)}$$

Where

q is the volume of water gained/lost per meter of river ($\frac{L^2}{T}$)

K is the hydraulic conductivity of streambed sediments, in unit of length per time

W is width of stream (L)

M is thickness of the streambed (L)

h_s is the head in stream determined by adding stream depth to the elevation of streambed (L)

h_a is the head in the aquifer beneath the streambed (L)

In this equation, leakage across the streambed changes, with the change of stream head and the aquifer head calculated during each time step. There is an assumption of uniform vertical flux exchange between streambed and aquifer over a given section of the stream with the use of hydraulic conductivity term in this equation (Figure 17) (Prudic et al., 2004). Thus, these packages just deal with vertical flow exchange from/to the streambed. In addition, surface water stage is computed inside these stream packages. For example, in the SFR package the rise of water stage in the stream network was due to inflow of water and routing of water through reaches from upstream to downstream.

In the bank storage model, the water exchange process, between groundwater and surface water improved, by using a type of Darcy Equation that accounts for lateral flow movement from streambed to or from groundwater (Figure 15). In addition, stream stage for the bank storage model can be calculated by using real data with use of a surface water model such as STAQ.

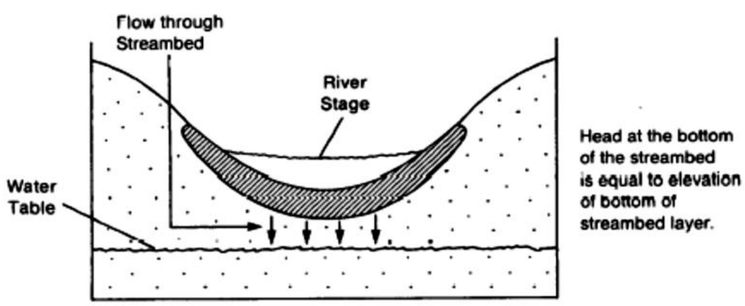

Figure 14: Water exchange with stream vertically.

Other stream package limitations are related to the fact that these packages are designed to model long-term changes that occur from months to hundreds of years in using averaged stream flows (Prudic et al., 2004). Extreme events such as floods are rapid hydrologic process that may happen within duration of hours to days. Therefore, stream packages are not capable to model flux of water exchange between stream water and groundwater aquifer due to rapid change of surface flow. The bank storage model is capable of accounting flux exchange due to flood events in a daily time step. Therefore, the bank storage model is filling in for limitation of stream packages, but still these packages are needed to route water through the stream network inside a groundwater model.

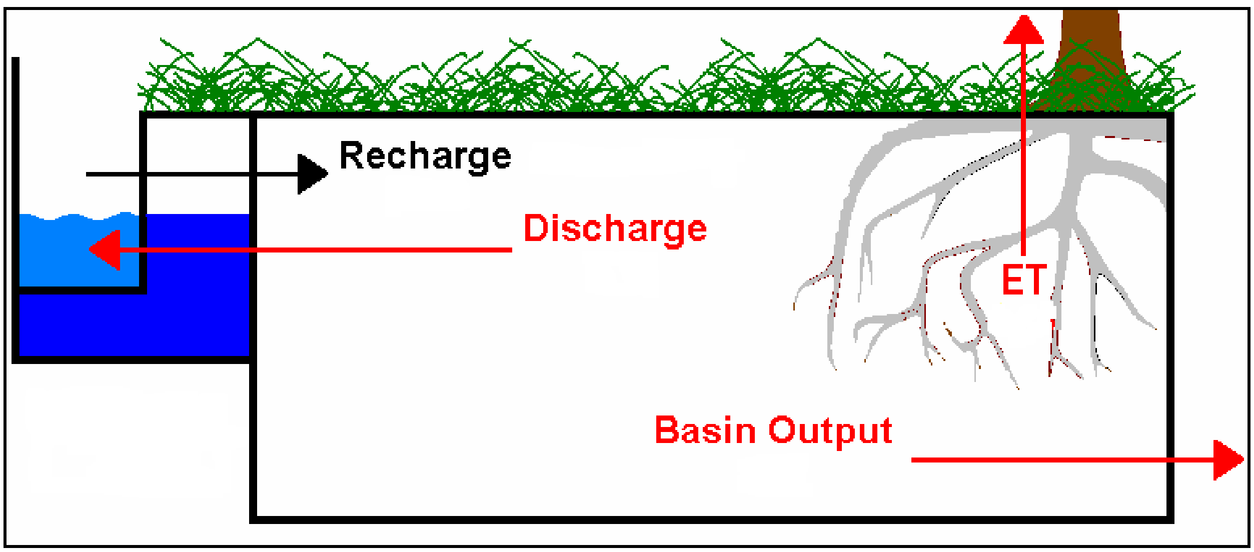

Figure 15: Lateral water exchange between near stream aquifer and stream (Simpson, 2007).

7- RESULTS AND DISCUSSION

7.1 Effect of bank storage for a stream reach

7.1.1 A gaining reach that remains gaining

Reach number two was chosen from the gaining segment of the Dry Alkaline hypothetical river, in order to investigate the effect of flood driven recharge on this gaining reach. Groundwater head of pre flood season for this reach was 1160 (m). This head value was used as groundwater stage for entire duration of model simulation for the wet summer season. Streambed elevation in this reach was 1158.5 (m). A stage hydrograph with three stage rise peaks was created (Figure 16). Maximum stream stage rise was 2.5 (m).

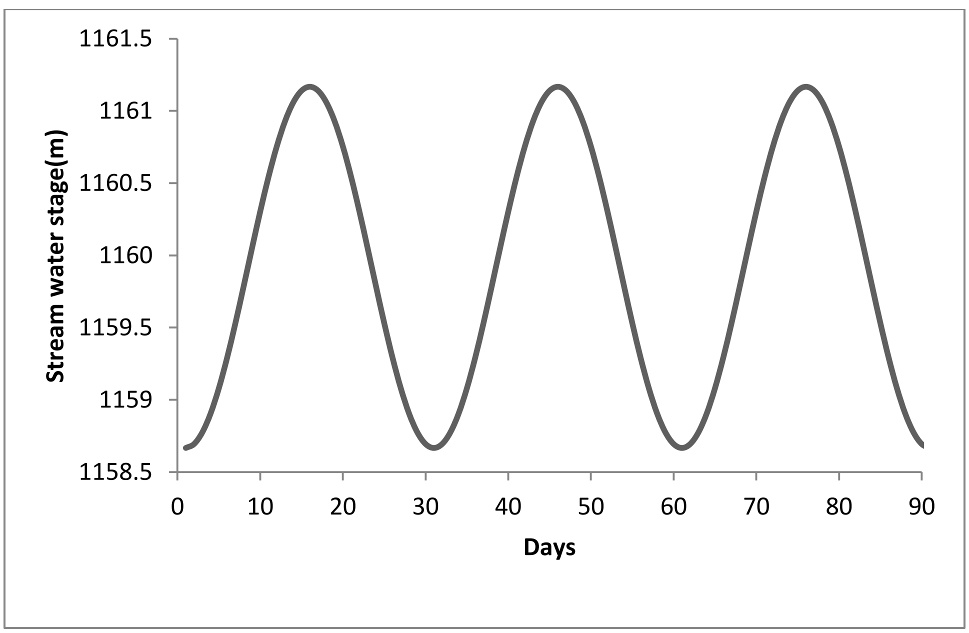

Figure 16: Stream stage hydrograph used for reach two.

In this example the diffusivity value was 1760 (m^2/day) and specific yield was 0.16. These values resulted in a transmissivity equal to 281.6 (m^2/day). The width of the riparian area was chosen to be 50 (m) for the length of distance that water can exchange between groundwater and surface water (dl).

Flux exchanged between groundwater and surface water for each day was calculated with the use of the Darcy equation (Figure 17). Positive q indicated that the river gained water and negative indicated that the stream was losing water. Net volume of water, which is the sum of all volume of water gained or lost for the flood season equaled 51.8 m^2/day. Since each reach of river had length equal to 1610 m, Q total for this section of model was equal to 83336 m^3/day or 0.96 m^3/sec. Thus, this segment remained a gaining reach during flood season.

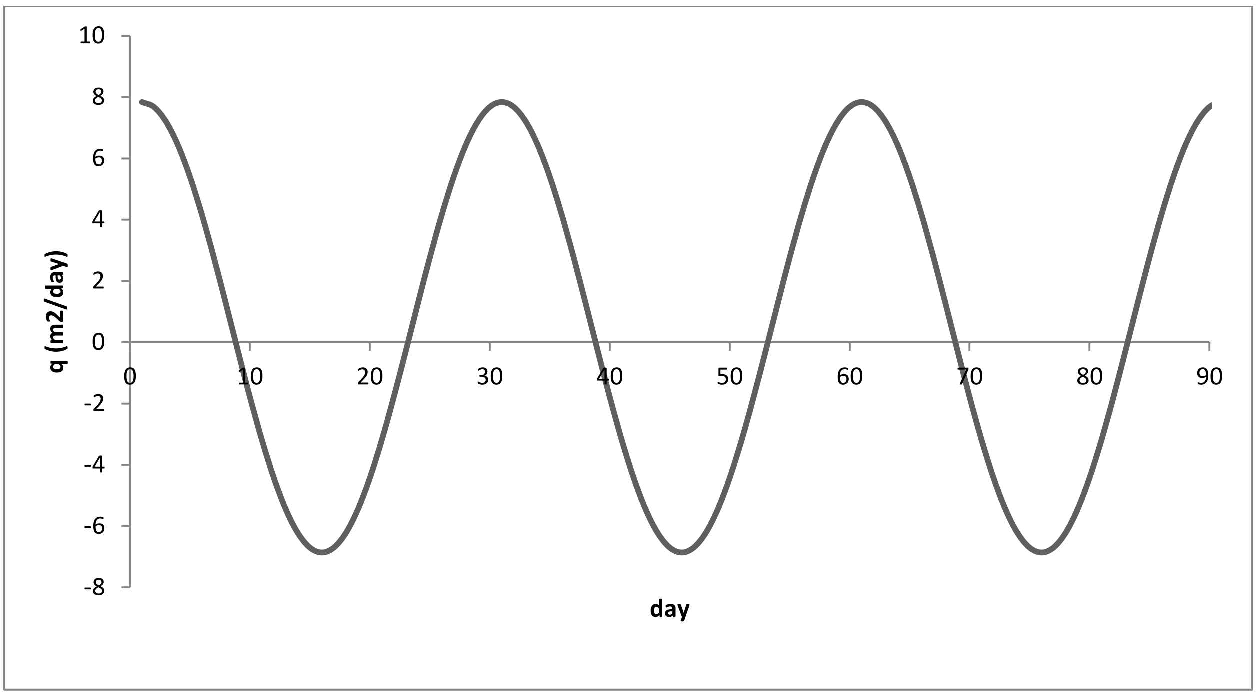

Figure 17: Flux resulted from Darcy equation for reach number 2, during flood season, for 90-day period.

7.1.2 A gaining reach that becomes losing

In order to illustrate that a reach of stream network can be gaining during a low flow period, but becomes losing during the high flow stage, reach number three was chosen. Based on SFR package calculations, this reach gained 0.98 (m^3/day) before incorporating flood recharge. Groundwater head for pre flood season was 1159 (m) in this reach. Same diffusivity, specific yield and width of riparian area as the above simulation were used. Streambed elevation for this reach was 1158.3 (m). A stage hydrograph with three stage rise peaks were created for this example (Figure 18). Maximum stage rise of water during peak flow was 2.5 (m).

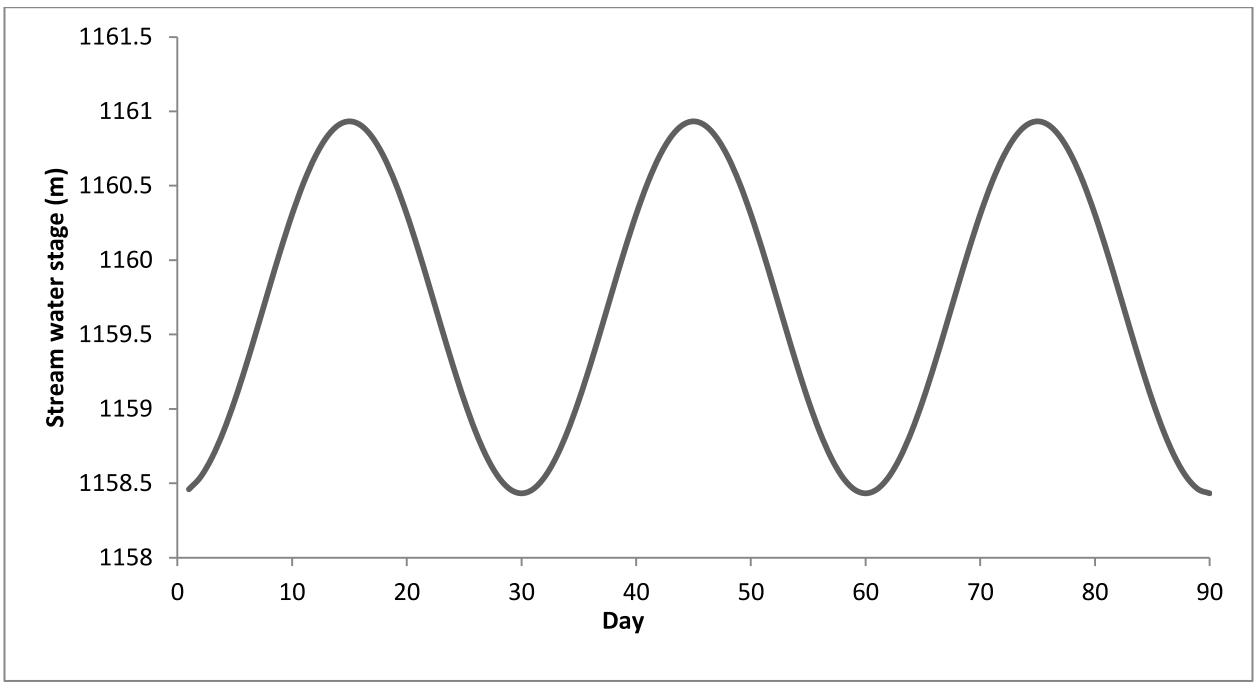

Figure 18: Stream stage hydrograph for reach number three with 2.5 (m) maximum stage rise.

Flux of water exchanged between groundwater and surface water for each day was calculated by using the Darcy equation (Figure 19). Positive volume of water (q) indicated that rivers gain water and negative values indicated that the river was losing water. Net volume of water, which is the sum of all water gained or lost for the flood season was equal to -359.4 (m^2/day). Since each reach of river has a length equal to 1610 (m), Q total for this model reach was equal to -577990 (m^3/day) or -6.7 (m^3/sec).

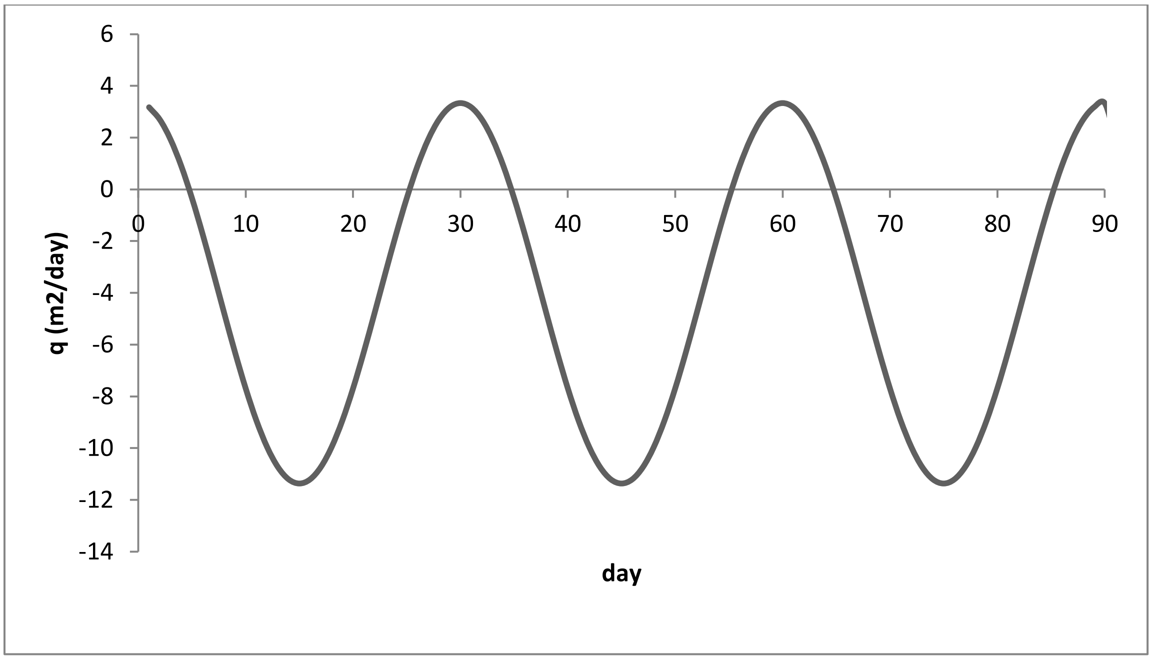

Figure 19: Flux resulted from Darcy equation for reach number 3, during flood season, for 90-day period.

7.2 STAQ model Results

Diffusivity and specific yield are parameters that incorporate soil properties to the bank storage model. For the hypothetical model of this study, the same diffusivity values as Simpson (2007) were used. Simpson used a series of nine flood pulses observed at the Lewis Springs transect between Aug. 4, 2006 and Sept. 8, 2006 to calculate diffusivity. Based on his study the value of diffusivity ranged from 1109 to 4475 (m^2d^{-1}). Simpson assumed constant diffusivity values applied to all model segments and for the entire model domain

7.2.1 Model code verification

STAQ model was used for verification of the bank storage program code. STAQ model was run with diffusivity of 4483.8 (m^2/day) and specific yield of 0.101. These parameters resulted in transmissivity equal to 452.9 (m^2/day). These values were assumed constant for entire simulation. Segment number two from STAQ model was used for code verification. The length of this segment was 5900 (m). Width of riparian area for this segment was equal to 157 (m). Groundwater elevation was constant and equal to 1259.6 (m). Streambed elevation in this point of model was 1258 (m). Stream water data for first 120-day simulation of model was used to verify bank storage code (

Figure 20). With the above input data STAQ was used to simulated flux of water (q) exchanged between groundwater and surface water (Figure 21).

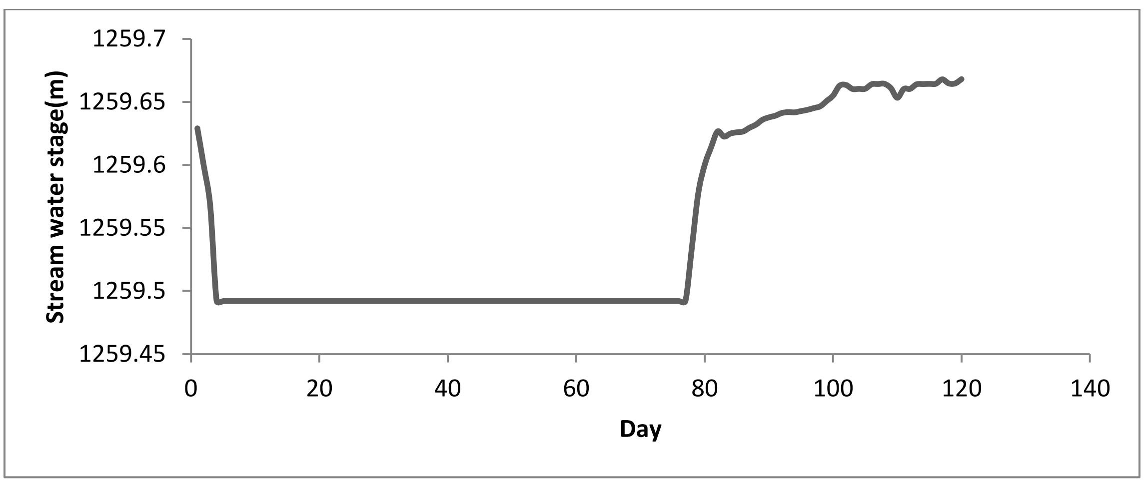

Figure 20: Stream water stage hydrograph for first 120 days of segment 2.

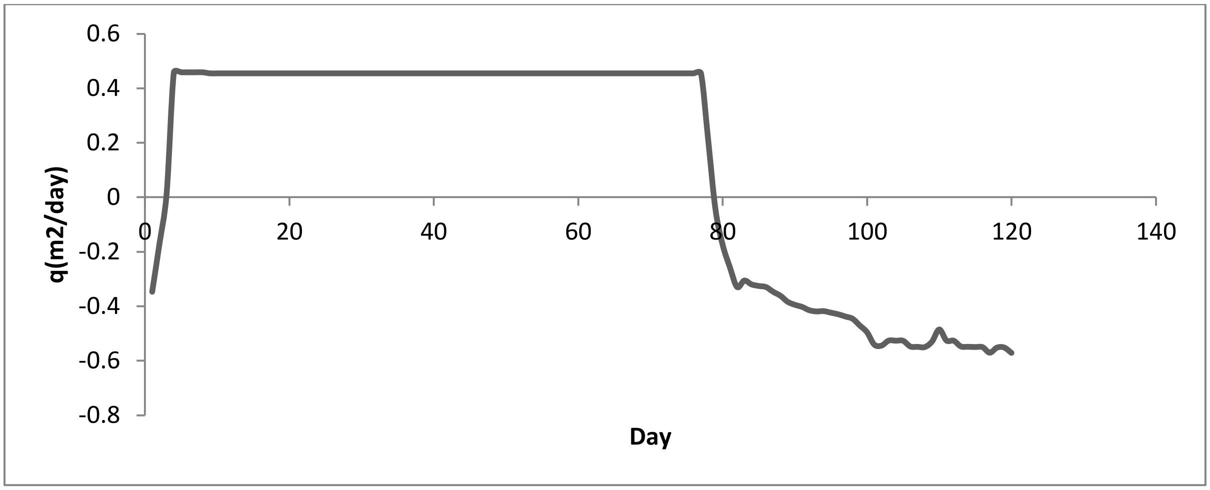

Figure 21: Water flux exchanged between groundwater and surface water from STAQ for first 120 days period of simulation for segment number two.

The calculation was repeated for the bank storage model. The same input data was used for stream elevation, diffusivity, specific yield, Width, streambed, groundwater elevation and stream water elevation.

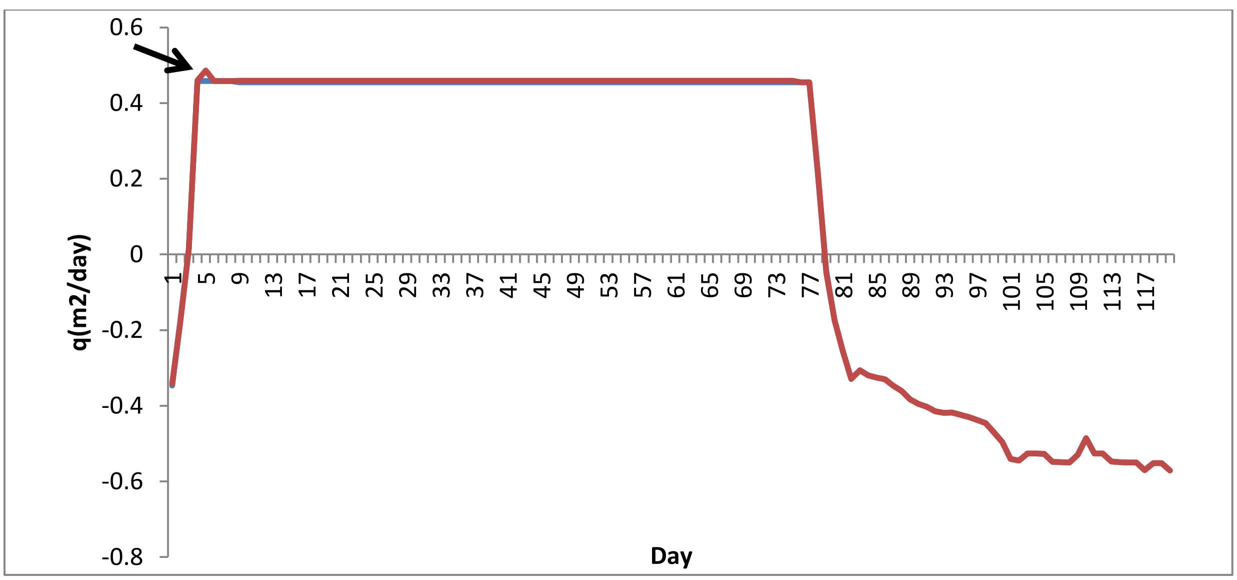

Figure 22: Result of bank storage model (red curve) almost fit on the result of STAQ model (blue curve)

The resulting water flux between groundwater and surface water from both models for same duration and location was plotted in same figure (Figure 22).There is only one location (an arrow in the figure 24) that result was not the exact same. Therefore, based on the results a conclusion can be made that STAQ validated the performance of bank storage model code .

7.3 Effect of different stage hydrograph characteristics

7.3.1 Effect of stage hydrograph shape

In this investigation, all other factors of each reach other than shape of stage hydrograph was the same. Normal stage hydrograph was created with 2.5 m maximum water stage rise, three-stage rise, and for a period of 90 days (Figure 23). Based on the area under normal stage hydrograph, a damping stage hydrograph that had 66% damping at the end of simulation period was created (Figure 24).

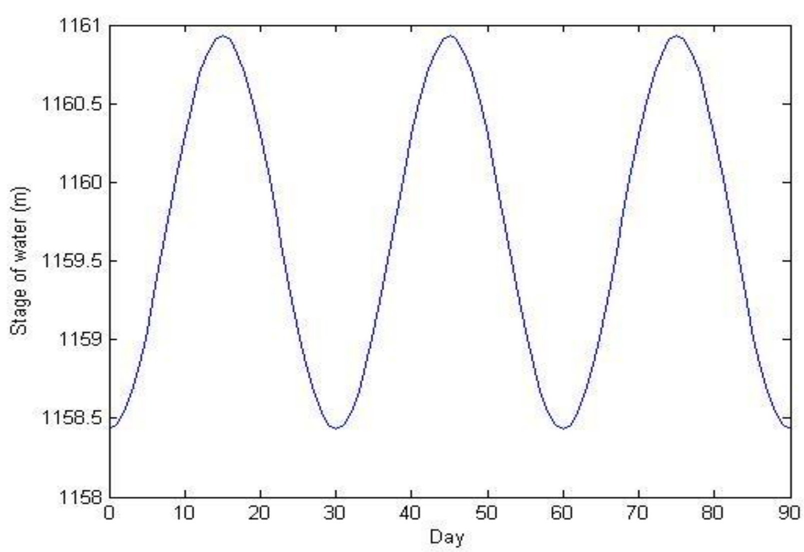

Figure 23: A normal stage hydrograph with base flow of 1158.4 m, peak rise of 2.5 m for a duration of 90 days.

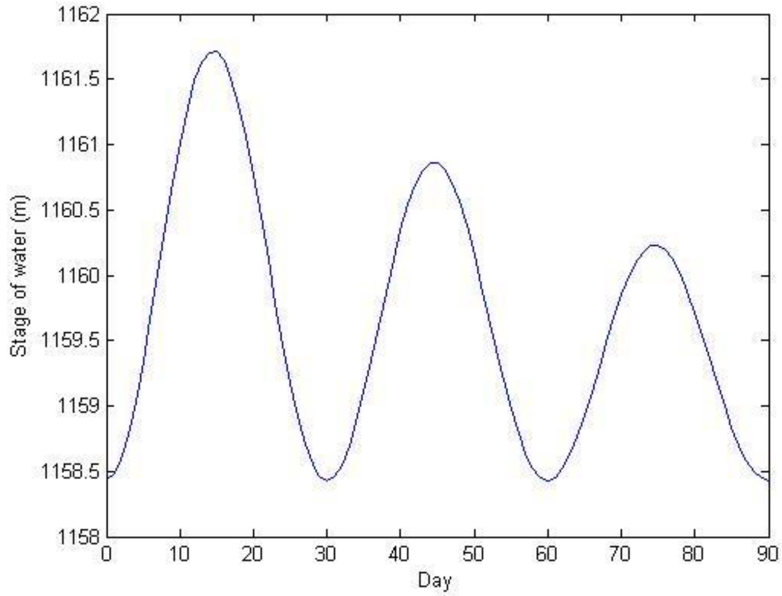

Figure 24: A stage hydrograph with 66% damping. Base flow of 1158.4 m for duration of 90 days.

With a diffusivity value equal to 1760 m^2/day and specific yield of 0.16, resulting transmissivity value was 281 m^2/day. The width of the riparian area was chosen to be 100 (m) for the length of distance that water can exchange between groundwater and surface water (dl). Groundwater level of each reach, came from the MODFLOW simulation for dry summer season. The model was run two times for each reach, one time in order to get flux of water for normal hydrograph and one time for the hydrograph with damping (Table 1).

Table 1: Result of water exchanged with normal stage hydrograph and damping stage hydrograph

Reach #	Base flow stage(m)	Groundwater stage(m)	Net flux of water for normal hydrograph	Net flux of water for hydrograph with damping	% error
1	1159.10	1161	170.08	169.85	0.14
3	1158.40	1159	-171.50	-170.98	0.30
8	1156.60	1157	-214.29	-214.57	0.11
26	1150.10	1151	-86.16	-86.41	0.28

7.3.2 Effect of number of peak rise of stage hydrograph

Three different stage hydrographs with three, six and nine stage peaks for reach number one, three, eight and twenty-six were generated (Figure 25). These three stage hydrographs had 2.5 maximum raise of stage during peak flow for a period of 90 days. Average stage rise of model was similar for all three of these stage hydrographs.

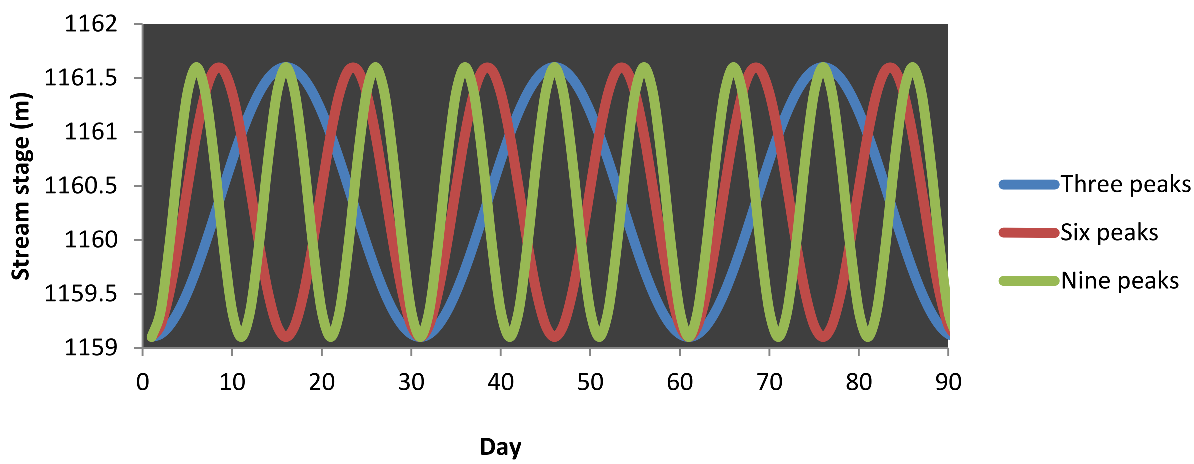

Figure 25: Three stage hydrograph with different number of stage peak

For this investigation, reach parameters such as diffusivity, specific yield and width of riparian area were the same as simulation on different hydrograph shape. The bank storage model was run for each stage hydrograph and for each of the reaches. The resulting net flux of water exchanged between groundwater and surface water (Table 2) was the same for all three hydrographs

.

Table 2: Net value of water exchanged between groundwater and surface water

Reach #	Base flow stage(m)	Groundwater stage(m)	q_{net} for 3 Peaks	q_{net} for 6 peaks	q_{net} for 9 peaks
1	1159.10	1161.00	170.09	170.08	170.09
3	1158.40	1159.00	-171.50	-171.51	-171.51
8	1156.60	1157.00	-214.30	-214.30	-214.29
26	1150.10	1151.00	-86.17	-86.17	-86.17

7.4 Incorporating bank storage result into groundwater model

7.4.1 Generating stage hydrographs

Four different types of stage hydrographs (Figure 26) were created for each of the twenty-six reaches of Dry Alkaline Basin model stream network.

Three types of these four hydrographs were hypothetical normal symmetric hydrograph. The input values in order to generate each of these three types of hydrographs were numbers of peak rise and stage of base flow of pre flooding season for each reaches of stream. All three types of hydrographs that were used in this study had three-peak rise. Stage of base flow was minimum flow from SFR package of groundwater model, during pre-flood season. Maximum stage rise is respectively 1.5 (m), 2.5 (m) and 3.5 (m). Duration of these hydrographs was for 90 days.

Furth type of hydrograph was generated based on the real data, from the STAQ model. This data was extracted from STAQ simulation for reach number two, during monsoon season (June, July and August) of year 1999. Then, the elevation was scaled into the elevation value of

Dry Alkaline Basin. Twenty-six of this type of hydrograph for duration of 90 days also was generated for the entire Dry Alkaline River.

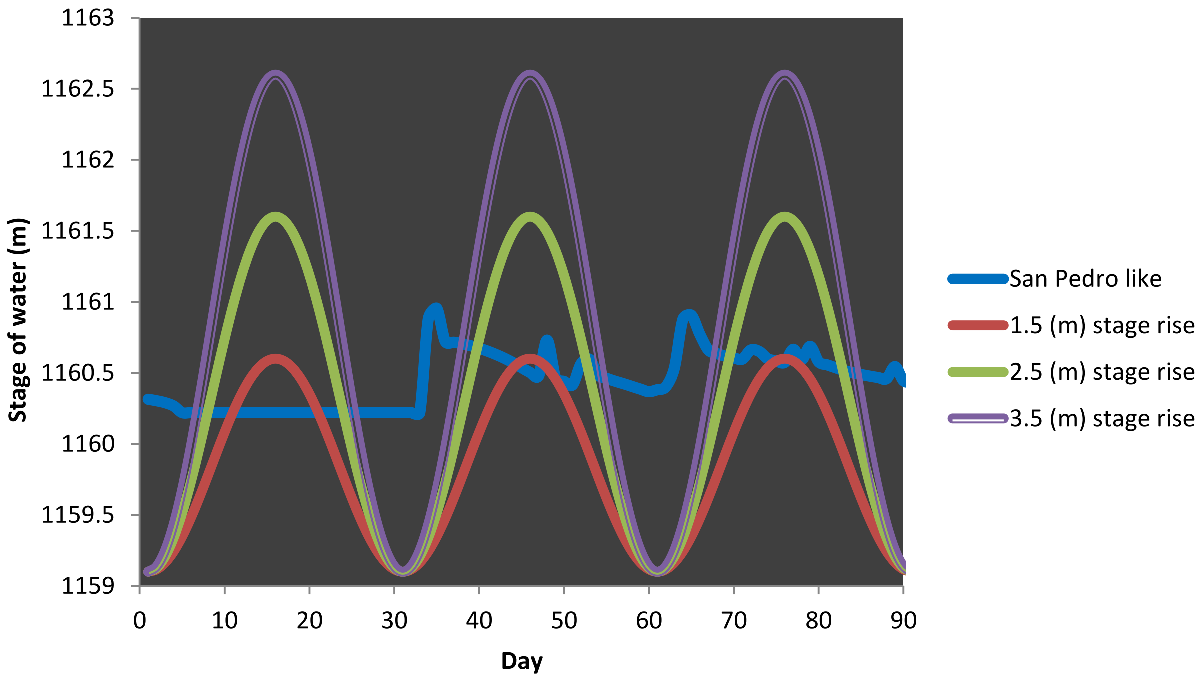

Figure 26: Four different types of stage hydrograph scenarios used in this study

7.4.2 Simulating bank storage effect

Flux of water exchanged between groundwater and surface water was calculated with used of the bank storage model. For each reach of the stream network, parameter such as diffusivity, specific yield, width of riparian area, stage elevation of groundwater and surface water is needed. All Darcy equation parameter remained same as the ones used in section 7.3. Groundwater level of each reach was taken from MODFLOW simulation for the pre flood season. Stream stage data was taken from the stage hydrographs mentioned in the previous section.

With the above parameter, the bank storage model was run for each hydrograph and for each of the reaches of the model in order to simulate flux of water exchanged between groundwater and surface water. Positive flux of water was indicating that stream gained water and negative flux of water was indicated that river was losing water. Finally, Net volume of water, which is the sum of all volume of water gained or lost during the flood season were calculated, and one value of q_{net} was reported (Figure 27).

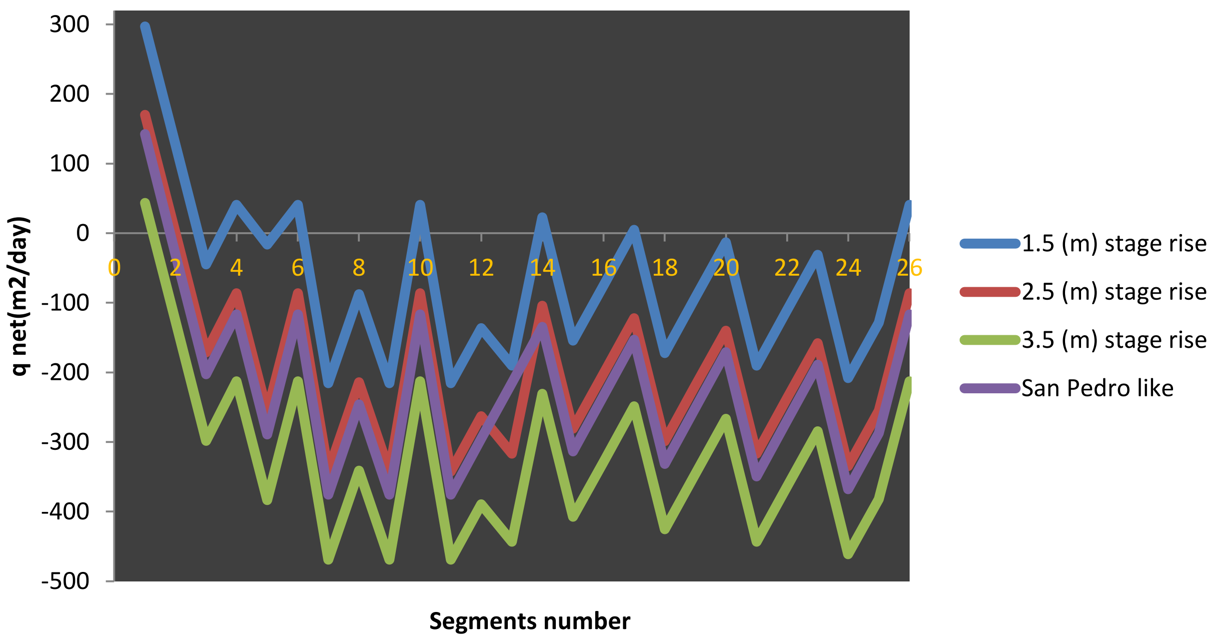

Figure 27: Net value of flux exchanged between groundwater and surface water for each type of hydrograph

7.4.3 Result of linking bank storage into the groundwater model

7.4.3.1 SFR Package Result

CAPT_CALC software was used to extract and plotted flux of water exchange between groundwater and surface water from SFR package. This software was created by Maddock III et al, (2010) and is able to read binary output file of MODFLOW and calculate value of capture from MODFLOW packages such as SFR, STR, and EVT etc.

The result from the SFR package for run of base case with MODFLOW model in the winter season shows how this stream was divided into three segments: gaining, neutral and losing (Figure 28). SFR output results were plotted for winter season after adding bank storage water resulted from different stage hydrographs, by using CAPT_CALC software (Figure 29). Each curve shows the stream condition in term of adding water to the aquifer, or removing water from the aquifer.

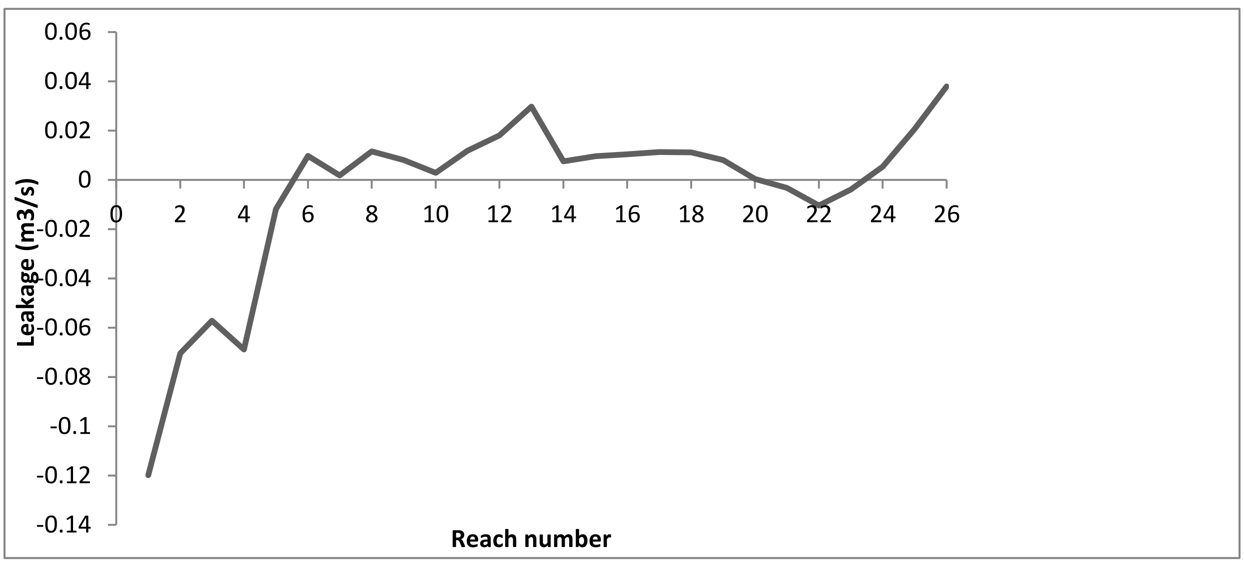

Figure 28: SFR output result for the base case using CAPT_CALC software

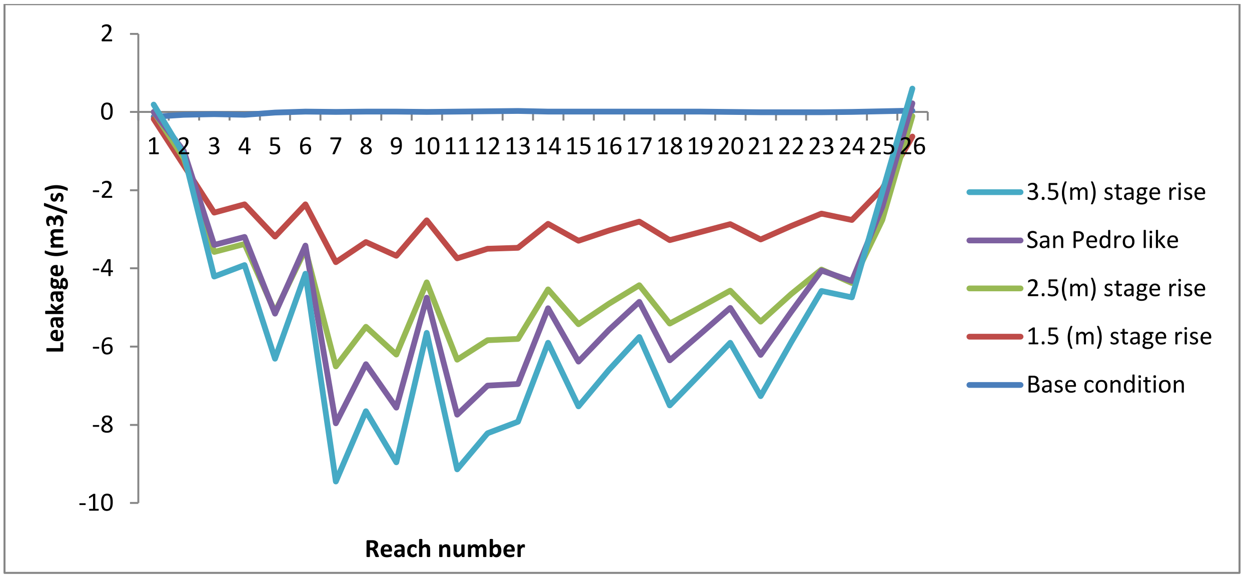

Figure 29: Recharge or discharge result from SFR package for winter season with different stage hydrograph using CAPT_CALC software

7.4.3.2 MODFLOW Result

The MODFLOW groundwater model of Dry Alkaline Basin was run five times for the base case and for each of the four different stage hydrograph scenarios. In order to find the water volume that is directly related to flood recharge for each case, water flux added to the system by Mountain Front Recharge needed to be subtracted from total volume of water that was added to the groundwater model by all wells. MFR was calculated in the base case model, where no flood recharge was added to the groundwater. MFR resulted was equal to 0.26 (M^3/s). By subtracting MFR quantity from total volume of water added to the groundwater model by wells, flood recharge for each stage hydrograph scenarios was calculated (Table 3).

Table 3: Volume of water that was added to the groundwater by wells because of bank storage processes in unit of (m^3/s)

	1.5 (m) stage rise	2.5(m) stage rise	3.5(m) stage rise	San Pedro like
Volume of water add to the system (m^3/s)	21.38	50.06	78.62	27.65

Groundwater head distribution over the basin for each season was next output result generated from MODFLOW. Five different hydrograph scenarios, and three seasons for each of these scenarios, resulted in fifteen different head distributions. Every single head value belonged to the end of each time step and for the middle of each grid cell. In order to show the groundwater head distribution over a basin, raster maps were used. The pixels of the squares were 1610 (m) ×1610 (m) as the cell of groundwater model. Head difference between each of four hydrograph scenarios in compared with base case was generated to illustrate result of adding bank storage processes. Raster maps are shown head difference between 3.5 (m) stage rise hydrograph and base case (Figure 30) and also head difference between San Pedro like hydrograph and base case (Figure 31) for winter season. In addition, to indicate that influence of bank storage can be observed in the dry season, raster map shows head difference between 3.5 (m) stage rise hydrograph (Figure 32) and San Pedro like hydrograph (Figure 33) in compared with base case were produced for dry summer season.

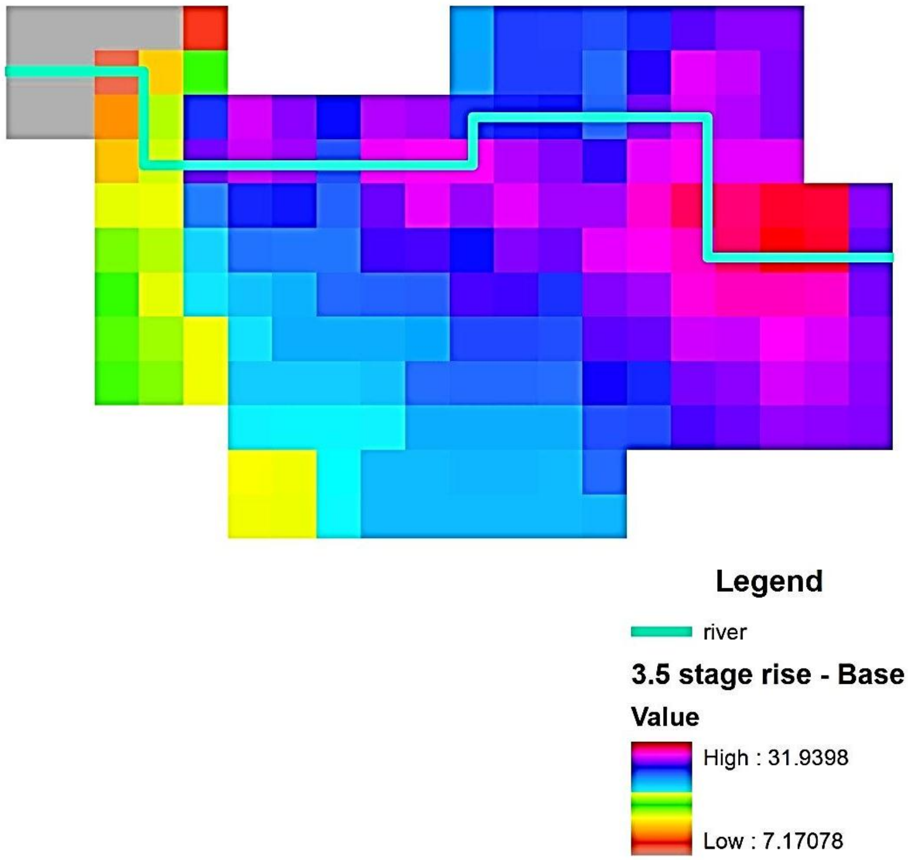

Figure 30: groundwater head difference between 3.5 (m) stage rise hydrograph and base case for winter season

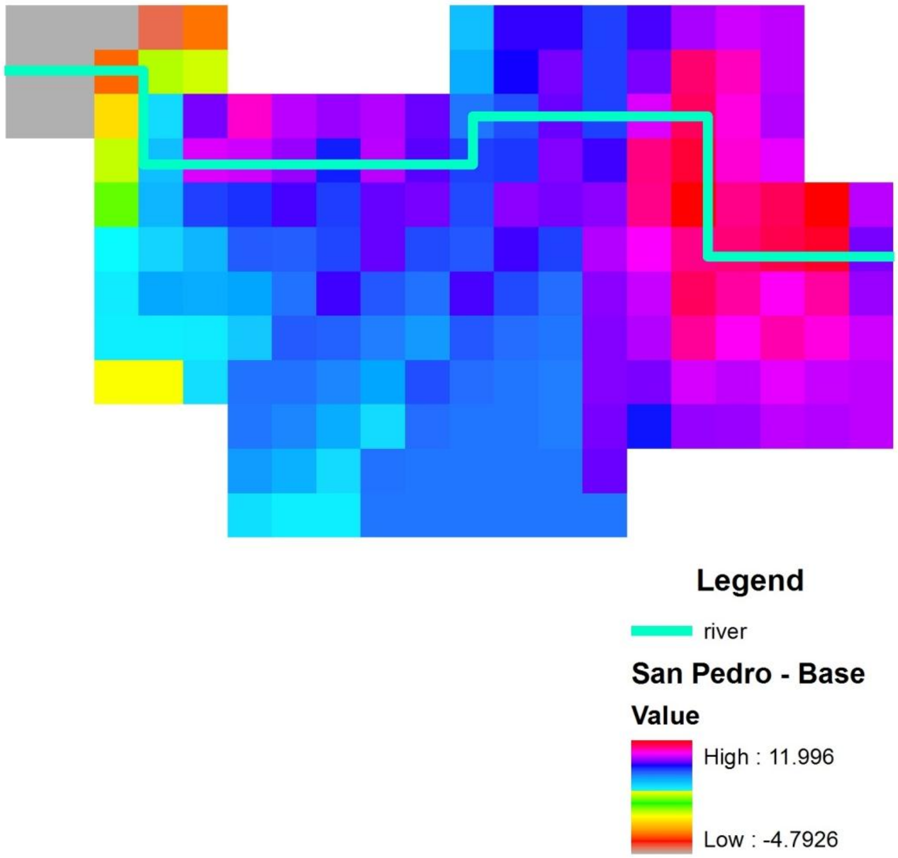

Figure 31: groundwater head difference between San Pedro like hydrograph and base case for winter season.

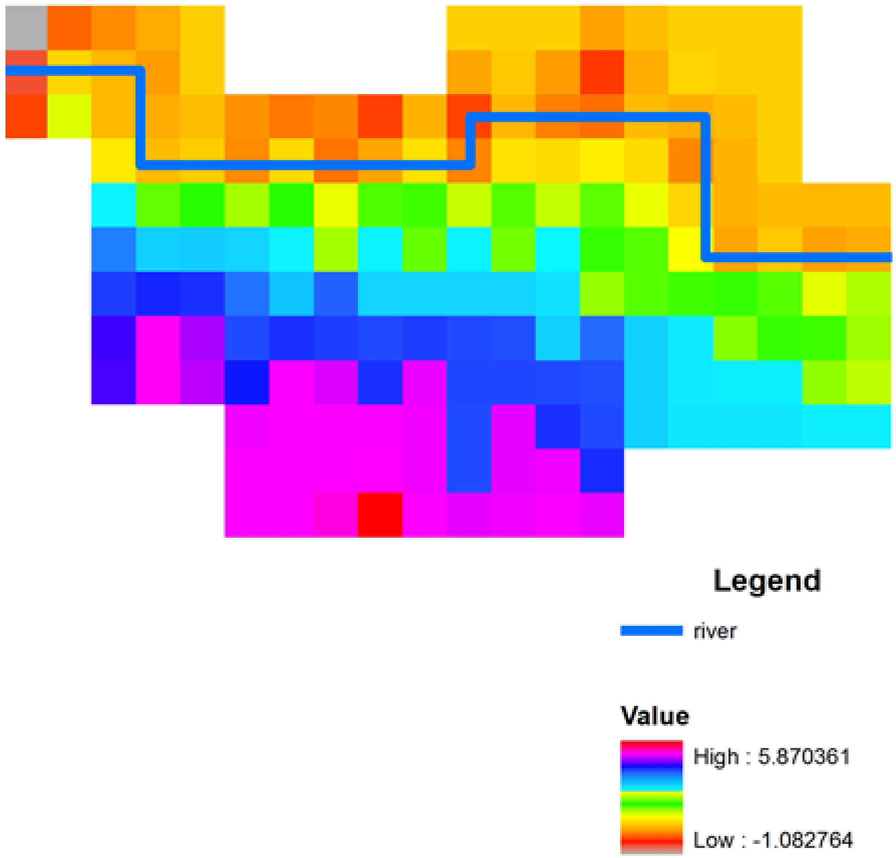

Figure 32: groundwater head difference between 3.5 (m) stage rise hydrograph and base case for dry summer season.

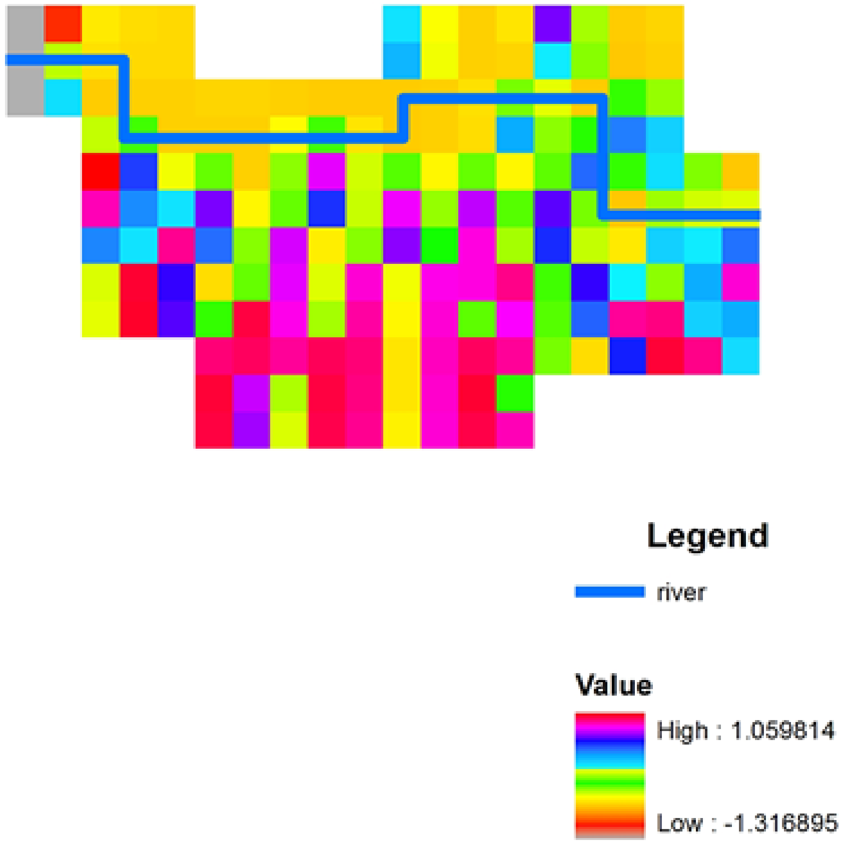

Figure 33: groundwater head difference between San Pedro like hydrograph and base case for dry summer season.

Finally, head differences between the base case and four hydrograph scenarios were compared along the model columns, for reach number seven and fifteen in order to observe the impact of flood driven recharge in north and south side of stream network (Figure 34, Figure 35). Zero value in the x-axis shows the location of stream reach, and positive value indicates north side of the stream and positive value shows the south side of the stream.

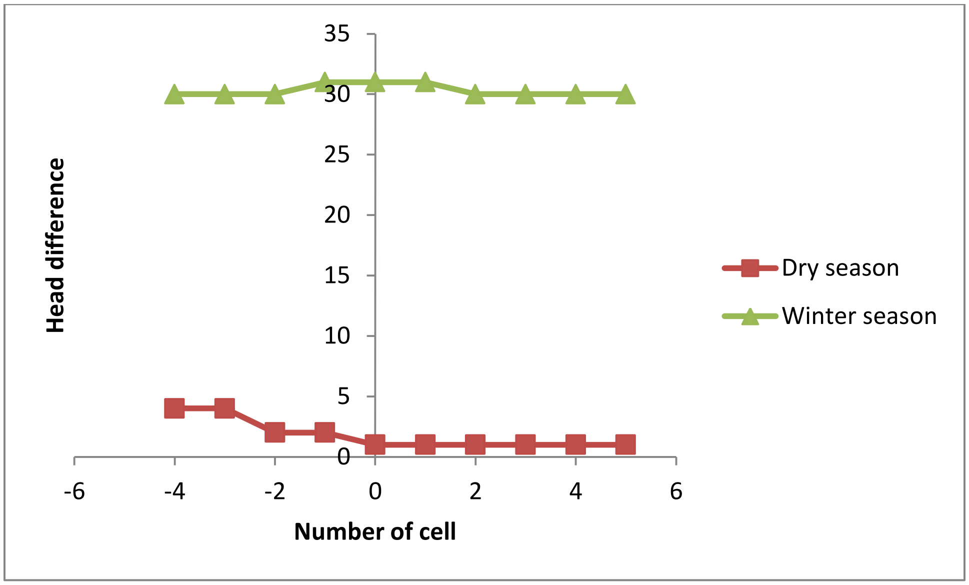

Figure 34: Head difference of 3.5(m) stage rise scenario with base case in segment number 7 for winter season (green line) and dry season (red line). Zero is location of the stream, positive values are cells located in north side of stream and negative values are cells located in southern part of stream

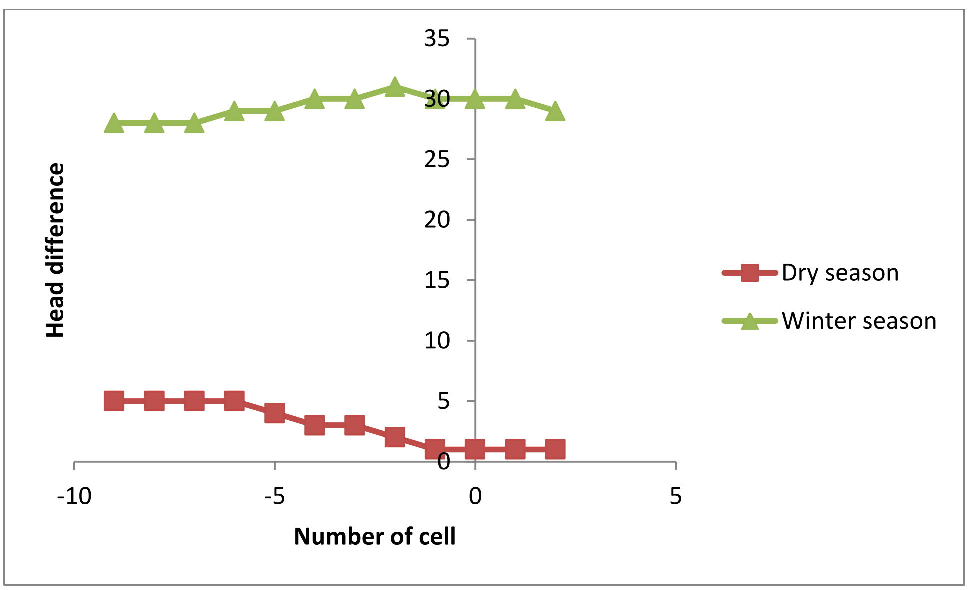

Figure 35: Head difference of 3.5(m) stage rise scenario with base case in segment number 15 for winter season (green line) and dry season (red line). Zero is location of the stream, positive values are cells located in north side of stream and negative values are cells located in southern part of stream

7.5 Effect of same volume of water with different stage hydrograph

In this section, the aim is to investigate effects of same discharge value entering into the stream with different stage hydrograph. For this investigation 2.5-stage rise, hydrograph was chosen for a period of 90 days (Figure 23). An arbitrary hydrograph (Figure 37) was created with use of a stage-Discharge curve (Figure 36) equation:

$$\mathbf{S= (8.9031*10^{-7}\text{- } D)^{0.33}+S_b \ (Eq.\ 6)}$$

Where:

S is stage of water in stream (m)

D is Discharge volume (m^3/day)

S_b is stage of water in base case (m)

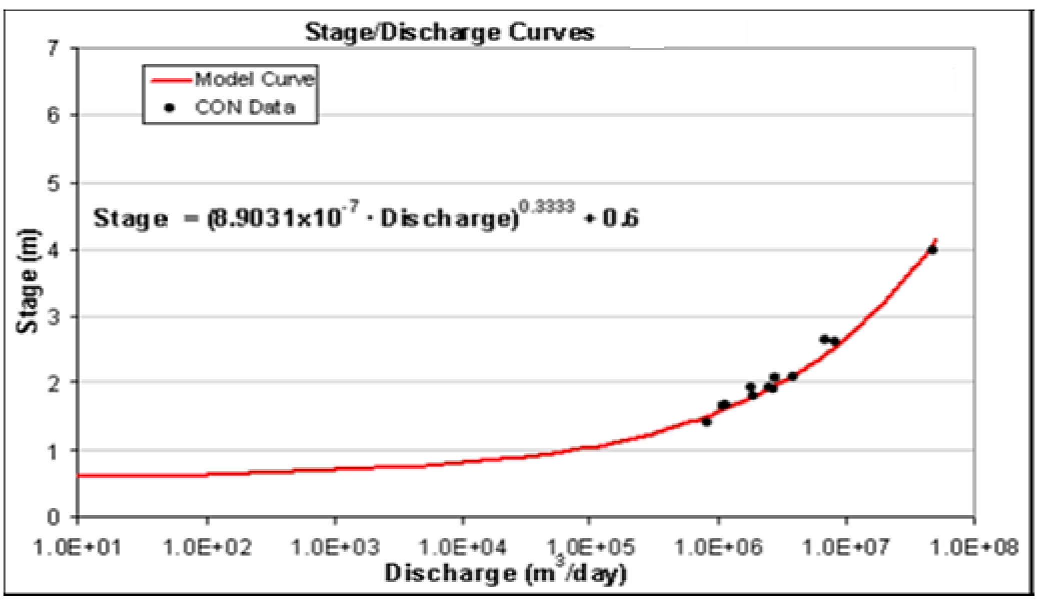

Figure 36: Stage-discharge curves for generating river stage from daily discharge (Simpson, 2011)

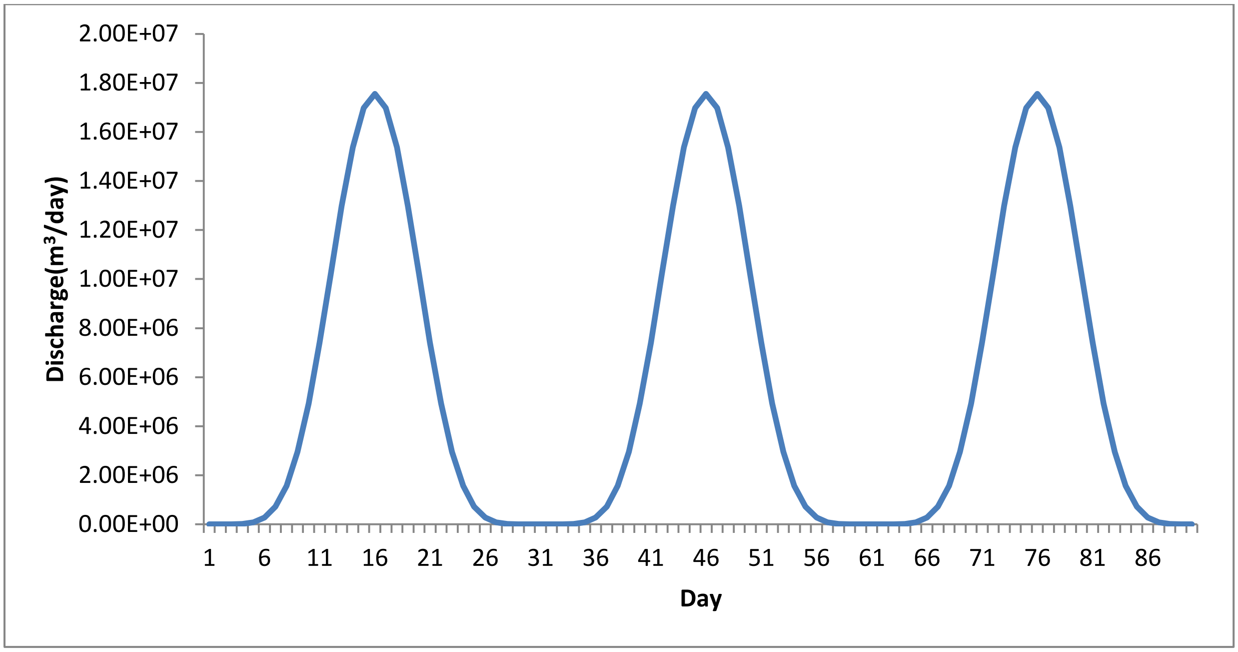

Figure 37: Hydrograph generated by using stage-discharge relation, with base flow of 1159.1 (m)

Volume for this hydrograph (Figure 37) was equal to 4.94E+08 (m^3/day). Based on this volume of water and with use of the same stage-discharge relationship a different adjusted stage hydrograph was created (Figure 38). The bank storage model was run for each stage hydrographs. The resulting net water volume (m^3/s) exchanged between groundwater and surface water (Table 4) was different.

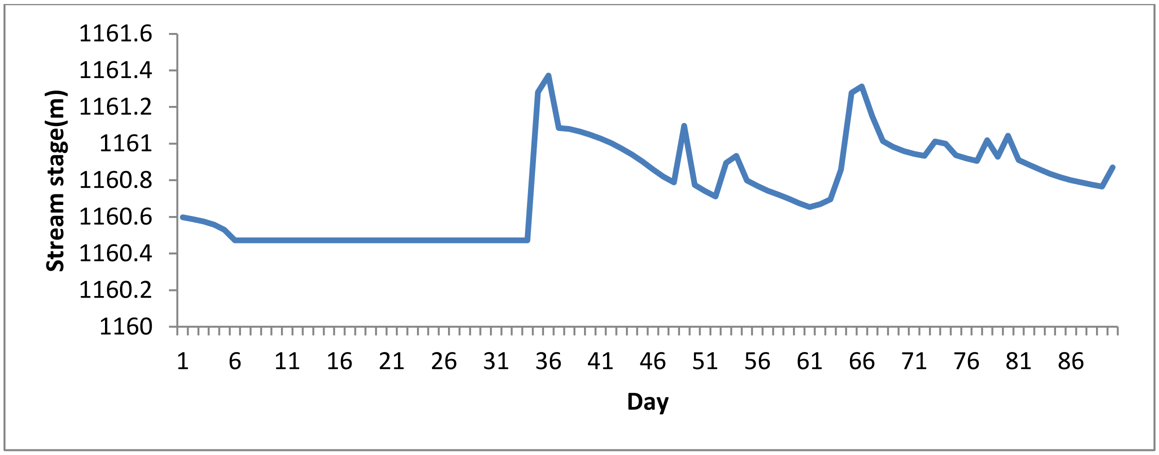

Figure 38: Stage hydrograph created with used of stage-discharge relationship and same volume of water (4.94E+08(m^3/day)) as 2.5 stage rise.

Table 4: Net volume of water exchanged (m^3/s) between groundwater and surface water for each of stage hydrographs for reach number one, three, eight and twenty-six

Reach #	Base flow stage(m)	Groundwater stage(m)	Q_{net} for 2.5 m stage rise	Q_{net} for adjusted stage
1	1159.1	1161	3.16	1.16
3	1158.4	1159	-3.19	-4.27
8	1156.6	1157	-3.99	-4.91
26	1150.1	1151	-1.60	-2.76

MODFLOW groundwater model of Dry Alkaline Basin was run for the 2.5 m stage rise and for adjusted stage hydrograph scenarios. Groundwater head distribution over the basin for each season was calculated with MODFLOW. Head difference between these hydrograph scenarios in compared with each other was calculated to illustrate result of same volume of water entering into the stream with different stage hydrograph on the groundwater. Raster maps shown the head difference between 2.5 (m) stage rise hydrograph and adjusted stage one were created for winter season (Figure 39) and for dry season (Figure 40).

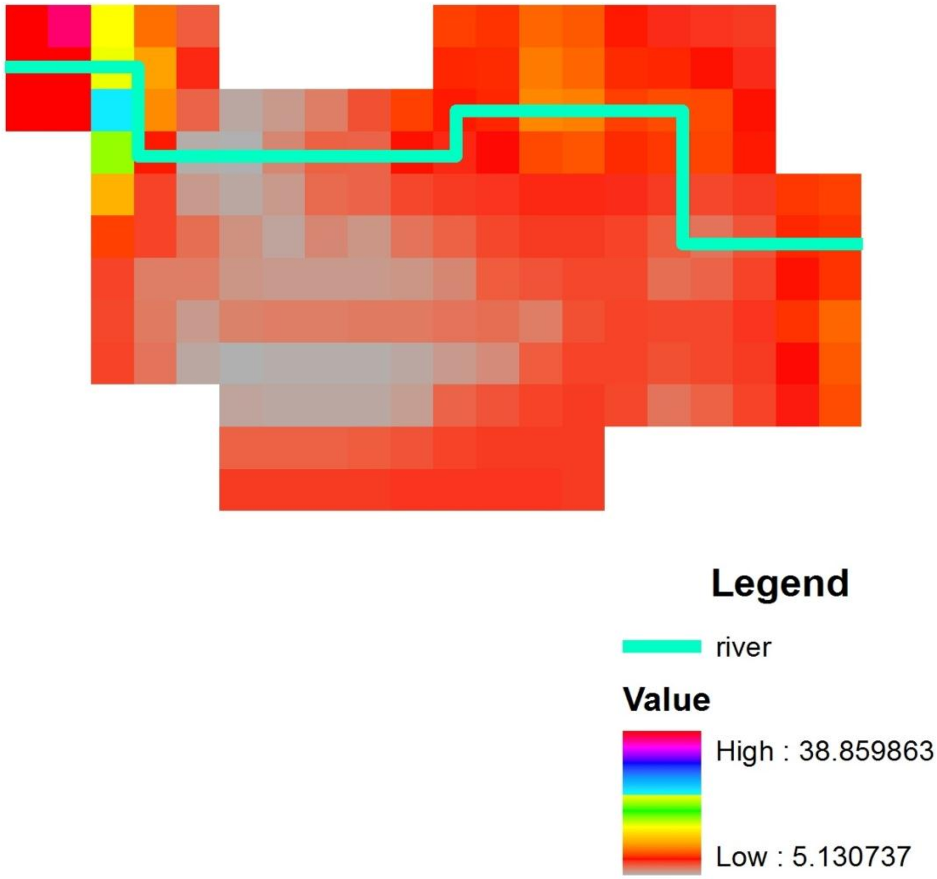

Figure 39: groundwater head difference between 2.5 (m) stage rise hydrograph and adjusted stage hydrograph for winter season.

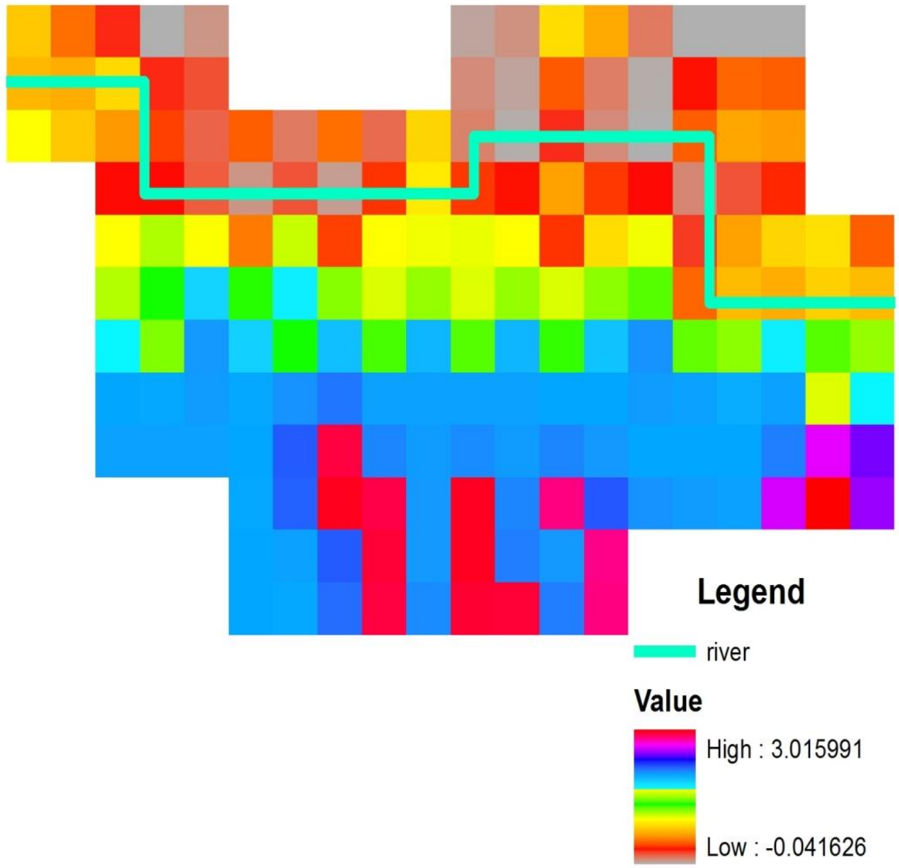

Figure 40: groundwater head difference between 2.5 (m) stage rise hydrograph and adjusted stage hydrograph for dry season.

7.6 Discussion

7.6.1 Adding floodwater to groundwater model

Usually streams linked to groundwater model by using one of the stream packages. There is an assumption of uniform vertical flux exchange between streambed and aquifer over a given section of the stream (Figure 17). In addition, there is a limitation related to the fact that these packages are designed to model long-term changes that occur from months to hundreds of years in using averaged stream flows (Prudic et al., 2004). During the flood event, net volume of water exchanged laterally between ground water and surface water, is stored in the bank area of the stream during the stage rise of stream. In addition, extreme events such as floods are rapid hydrologic processes that may happen within duration of hours to days. Stream packages are not capable to model flux of water exchange between stream water and groundwater due to rapid change of surface flow. The bank storage model produced here offer a way around limitation of stream packages by accounting lateral fluxes due to flood events in a daily time step.

In order to link the amount of flood recharge to a groundwater model, a MODFLOW package that can simulate flux of water as a boundary condition is needed. Two packages for MODFLOW were able to do so, the well package and recharge package.

The Recharge package is usually used to simulate a specified flux distributed over the top of the model and specified in units of L/T. Within MODFLOW, these rates are multiplied by the horizontal area of the cells to which they are applied to calculate the volumetric flux rates. The Well package is used to simulate a specified flux to individual cells and specified in units of L^3/T. Because grid cells size of groundwater was 1610(m) ×1610(m). So if the cell size was refined and contain only the stream and riparian area, the recharge package could be used easily. Well package was the package used to put additional water resulted due to bank storage in to groundwater model.

7.6.2 Effect of flood recharge

Effect of floodwater recharged on each reach of the stream network, depends on the condition of the reach in the base case (Figure 28), but results show that maximum floodwater infiltration occurs, with the maximum rise of flood stage (Figure 17). If in the base case the reach was a losing reach, this reach remained losing, but during floods stream lost more water to the near stream aquifer.

The rise in the stream stage during high flow events (floods) can induce losing stream conditions even along stream reaches that are strongly gaining during low flow conditions. Results from this study indicated that a reach which was gaining during normal flow of the stream network, could became a losing stream during high flow period (Figure 19). This depends on the stage of water entering to the system and condition of the reach in the base case, for example, reach number two (Figure 17) in this study was dominantly gaining and even remains gaining with 2.5 (m) stage hydrograph, but this reach became losing with 3.5 (m) stage hydrograph.

Five different flood hydrograph scenarios were compared to investigate different effects of flood driven recharge (Figure 26). The results of this study indicated that, a stream stage rise in compare with near stream groundwater level during flood events leads to infiltration of stream water into the aquifer (Figure 27). Vice versa, lower stream water stage during low flow period lead to return of the infiltrated water to the stream system.

Head differences between base case in comparison with scenarios that bank storage model were incorporated (Figure 30 and Figure 31) demonstrate that, linking flood recharge in to the groundwater, resulted in an overall groundwater head rise over the entire basin. This head difference depended on the quantity and average flood stage entering to stream network of basin. Influence of flood driven recharge is less as the river flows through gaining streams. Moving to downstream losing reaches, the influence of monsoon floodwater increased. The Results also indicated that the highest impact of floodwater was observed in locations near the stream for this season (Figure 34 and Figure 35). Moving to the left or right hand side of the stream reach, the difference between groundwater head of base case with the flood induced one. That is because of bank storage processes water was exchanged with the near stream aquifer during flood season, and impact can be observed on the head distribution of groundwater in this post flood season.

The effects of flood driven recharge can be observed in dry season as well (Figure 32 and Figure 33). The resulting head difference for this season also indicated head rise over the entire basin like winter season. However, differences between head distribution of winter season and dry summer season are on the location that highest head difference can be observed. This comparison showed that in cells far from the stream, the difference between groundwater heads was higher for the dry season. During dry season, some portion of bank storage had been moved back to the stream. The lowest impact of floodwater during this season was observed in location near the stream, particularly in gaining part of stream. Moving to the north or south side of the stream reaches, the difference between groundwater head of base case with the flood induced one become higher (Figure 34 and Figure 35). Reason for this phenomenon is flood wave effect. Floodwater recharged alters hydraulic head at the river so, that pushes water from near the stream farther from the stream. Therefore, when water enters from stream to the basin groundwater, it created a pressure wave in the aquifer that continued to migrate out throughout the basin after the flood season was completed. Thus, the effect of the floodwater wave caused head differences in the area far from stream to be higher for simulations including flood drive recharge compared with the base case for the dry summer season.

7.6.3 Impact of different stage hydrograph characteristics

In this study, some of the specific stage hydrograph characteristics such as number of peaks and shape of hydrograph were examined.

Based on the area under normal stage hydrograph (Figure 23), a damping stage hydrograph that had 66% damping at the end of simulation period was created (Figure 24). Product of model (Table 1) shows that with very small error (less than 0.3%), these two fluxes of water caused by two different stage hydrographs that had same area under the stage hydrograph, but with different shapes, resulted in the same amount of flux of water exchange between aquifer and stream.

Next hydrograph characteristic that was examined in this research was effect of peak number (Figure 25). The results showed that the number of peak does not have any impact on the net flux of water exchanged between groundwater and surface water due to bank storage process (Table 2). All three-hydrograph scenarios (with three peaks, six peaks and nine peaks) resulted in same net volume of water exchanged between groundwater and surface water. Thus, when the resulting net flux of water exchanged for these three hydrographs linked in to a MODFLOW model, led to same result.

A key aspect in all of the above simulation was, all of the stage hydrographs had same area under the curve, means the same average stream stage. This result illustrated that average stream stage is an important element of the stage hydrograph impact on the volume of water exchange between surface water and groundwater. Higher average stage caused higher amount of water exchange between groundwater and surface water (Figure 27). Reason is the magnitude and route of the flux exchanged between surface water body and groundwater is determined by the hydraulic gradient between the river and the underlying aquifer, and product of higher stream stage resulted a higher hydraulic gradient value. The bank storage model could be simplified by just using the average stage value for the entire duration of the flood season.

7.6.4 Effect of same volume of water with different stage hydrograph

A comparison of same volume of water entering to the stream with different stage hydrograph showed the resulting net water volume (m^3/s) exchanged between groundwater and surface water (Table 4) was different. The water volume exchanged between groundwater and surface water in case of adjusted stage hydrograph was higher. Reason for higher volume of water exchanged between groundwater and surface water was higher average stage. Comparison of Figure 23 with Figure 38 indicates that the adjusted stage hydrograph has higher average stage in comparison to the 2.5 (m) stage rise hydrograph.

Groundwater head distribution over the basin for winter season (Figure 39) indicates that different volumes of water exchanged due to bank storage processes resulted in different head distributions. Since the adjusted stage hydrograph has higher average stage, the overall head distribution was higher. Along the river, higher head differences can be observed. Also for winter season, area on the gaining part of the stream show higher head difference in comparison with the losing section. The reason for this result is, higher stage elevation for adjusted stage hydrograph. So higher difference is observed in cells near the stream.

The effects of flood driven recharge can be observed in the dry season although, the head difference is not as high as winter season result (Figure 40). The outcomes of MODFLOW indicated that groundwater head was higher for the adjusted stage hydrograph in this case as well. This comparison indicated that in cells far from the stream, the difference between groundwater heads was higher. This is again because of flood wave effect. The influence of the floodwater wave caused head differences in the area far from stream to be higher for this simulation. During dry season near stream water return back to the stream, resulted in lower head difference at near stream cells.

The 2.5 (m) stage rise hydrograph consisted of three instantaneous flood events, with stream rise to a maximum value and then decreased to the base flow stage (Figure 23). The maximum stage rise is lower for the adjusted stage hydrograph (Figure 38), in comparison with 2.5 (m) stage rise hydrograph, but the stage was raised gradually and stage of water was higher than base flow during the 90-day simulation. Same volume of water entered the stream with both of these hydrographs, but as indicated, adjusted stage hydrograph resulted in higher volume of water exchanged. So an implication here is, in order to have more water stored in bank storage, higher average stage is needed. Higher average stage happens not with a huge volume of flash flood, but with a consistent flood that last for a longer time. Therefore, if two floods with the same volume of water enter a stream, one with a big flood, over a short time and other one with lower stage rise, but for a longer period, the event that lasts longer may result in more water exchange due to bank storage processes.

8 CONCLUSION

This study investigated flood driven recharge with a bank storage model coupled to a groundwater model. To link the amount of flood driven recharge into a groundwater model, a MODFLOW packages is needed. This package must have the capability of simulating a specified flux as a boundary condition. Two packages for MODFLOW are able to do so, the Recharge package and Well package. In order to make the process of adding recharge driven flux as simple as possible, in this study the Well package was used. Additional water flux caused by floods during the flood season was added to the next season, which in this study is the winter season. In the Well package, flood recharge was treated as an injection well to the middle of cell. Discharge to the stream was simulated by pumping well from the middle of each cell.

The effect of flood driven recharge on each stream reach depends on the condition of the reach in the base case. In this study maximum floodwater infiltration occurred at maximum stream stage rise. If in the base case the reach was a losing reach, this reach remained losing. However, during floods event, stream losses increased. Losing more water leads to higher recharge to the near stream system. This result indicated that there is floodwater available as a recharge source in losing part of stream. The rise in the stream stage during high flow events can also induce losing stream conditions along stream reaches that are strongly gaining during low flow conditions. This phenomenon depends on the quantity of floodwater, the floodwater stage rise, and condition of the reach in the base case. For example, reach number two in this study was dominantly gaining and even remains gaining with a 2.5 (m) stage rise hydrograph, but this reach became losing with a 3.5 (m) stage rise hydrograph. Thus, this study shows that such two-way exchange does occur in a particular stream reach. Results also indicated that higher stage hydrographs means a severe flood can make an entire river system a losing stream.

Groundwater head differences for the entire simulated aquifer system demonstrated that, linking flood recharge into a groundwater model can result in an overall groundwater head rise over the entire basin domain. The influence of flood driven water is minimal for gaining reaches. Losing reaches show much greater recharge.

The effects of flood driven recharge can be observed in the season that water was added to the system and in other seasons of a groundwater model. The results indicated that groundwater head was higher for the 3.5 (m) case, by comparing dry season head difference between the 3.5 (m) stage rise of water and the base case. This comparison showed that in cells far from the stream, the difference between groundwater heads was higher. Floodwater recharge alters hydraulic head at the river. This altered head pushes water from near the stream farther from the stream. Therefore, when water enters from stream to the basin groundwater, it created a pressure wave in the aquifer that continued to migrate out throughout the basin after the flood season was completed. Thus, the effect of the floodwater wave caused head differences in the area far from the stream to be higher for simulations including flood drive recharge compared with the base case for the dry summer season. For this dry season near stream, hydraulic head returns to the stream and resulted in lower head difference at near stream cells.

The highest impact of floodwater during post flood season was observed in locations near the stream. Moving to the north or south side of the stream reaches, the difference between groundwater head of base case with the flood induced one declined. In this hypothetical study, the effect of floodwater is observed in the cell farthest from the stream because the Dry Alkaline basin groundwater model is a high transmissivity case study. In the real situation, the distance of floodwater influence depends on aquifer characteristics, groundwater levels and the size of the flood.

Different shape and number of peak rise of stage hydrograph, when the average stage was the same, resulted in the same net flux exchanged between the stream and the aquifer. This result indicates that the most important element of the stage hydrograph for volume of water exchange between surface water and groundwater is the average stream stage. Higher average stage caused higher amount of water exchange between groundwater and surface water. Thus, bank storage model could be simplified by just using the average stage value for the entire duration of the flood season.

However, simulations of equal volume but different average stage resulted in different recharge fluxes from the surface water. The hydrograph that had higher average stage, resulted in the higher net flux of water exchanged between surface water and groundwater. A big flood that last for a short time cannot make average stage rise for duration of a season. Higher average stage happens with a consistent flood that last for a longer time.

Based on the result of this study it is recommended that a bank storage model needs a surface model, so that real water stage data can be generated, so that these values can be averaged over the flood season. Thus, the amount of water calculated as bank storage would be accurate, given good simulation of the average stage hydrograph. In addition, a recommendation here is to use the result of this study in order to link bank storage effect to the SFR package and create a new package that can simultaneously, simulate the effect of flood recharge and route the water through the stream network.

Finally, in order to protect and maintain riparian systems, water decision makers need to know quantity of water and the source of this water. There are two significant water sources for the riparian zone in a semiarid region, local basin groundwater discharge and local recharge of floodwater during the flood season. The tools developed by this study can be a good means for water managers to account for floodwater effects and the subsequent linking to groundwater models. This bank storage model is applicable in any basin that flood event are important on a seasonal basis. This recommendation is particularly true in rivers with alternating gaining and losing reaches.

APPENDIX A: MODEL CODE

A- BANK STORAGE CODE

```
  % This code calculate bank Storage for each reach.
clc
clear all
close all
status = 0;
while (status==0)
    strD = input('Enter Diffusivity: ', 's'); %to get Diffusivity
    [D, status] = str2num(strD);
end
status = 0;
while (status==0)
    strS = input('Enter specific yield : ', 's');%get sy
    [S, status] = str2num(strS);
end
T = D * S; % calculate transmissivity
status = 0;
while (status==0)
    strW = input('Enter width: ', 's');
    [W, status] = str2num(strW);
end
dl = W/2; %is half the width of area that water can exchange between
groundwater and    surface water (m)
status = 0;
while (status==0)
    strx = input('Enter x(Stream Bed Elev.): ', 's');
    [x, status] = str2num(strx);%stream bed as crtical point
end
status = 0;
while (status==0)
    strH2 = input('Enter groundwater elevation from MODFLOW: ', 's');% this
term is from MODFLOW run
    [H2, status] = str2num(strH2);
end
```

```
status = 0;
while (status==0)
   strL = input('Enter Length of Reach: ', 's');%
   [L, status] = str2num(strL);
end
H1 = dlmread('h.txt');% reading surface elavation from text file
constraint = 1-((H1<x).*(H2<x));%% this term is to make sure, to give q==0
in situation of drought which both groundwater level and surface water level
is below stream bed
q = T * (H2 - H1).*constraint/dl;%Darcy equation calculation
Precision = 7;%0000.000
filename = 'q.txt';
plot(q);
dlmwrite(filename , q, 'delimiter', '\n', 'precision', Precision);
q_Net = sum(q)
Q = L * q_Net
Recharge=Q/(1610*1610)
Qs=Q/(86400)
```

B- HYDROGRAPH WITH DAMPING

```
clc
close all
clear all

% Input
FirstLvl = 1000;
Days = 120;
Increment = 5;
NumberOfPeaks = 3;
Precision = 7;
DampTime = 90; % 66% damp up to this day
filename = 'surface water.txt';
% % % % %

w = 2*pi*NumberOfPeaks/Days;
```

```
t = 0:Days;
x = FirstLvl + Increment * exp(-t/DampTime).*(-1*cos(w*t)+1)/2;
plot(t,x)

dlmwrite(filename, x, 'delimiter', '\n', 'precision', Precision);
```

C- NORMAL HYDROGRAPH

```
clc
close all
clear all

% Input
FirstLvl = 1159.1;
Days = 90;
Increment = 2.5;
NumberOfPeaks = 3;
Precision = 8;
filename = 'h.txt';
% % % % %

w = 2*pi*NumberOfPeaks/Days;
t = 0:Days;
x = FirstLvl + Increment*(-1*cos(w*t)+1)/2;
plot(t,x)

dlmwrite(filename, x, 'delimiter', '\n', 'precision', Precision);
```

D- CREATE A DAMPING SHAPE HYDROGRAPH FROM NORMAL ONE

```
clc
close all
clear all

% Input
FirstLvl =1150.1;
Days = 90;
Increment = 2.5;
NumberOfPeaks = 10;
Precision = 7;
DampTime = 90; % 66% damp up to this day
filename = 'DAMPING.txt';
% % % % %

w = 2*pi*NumberOfPeaks/Days;
t = 0:Days;
x = FirstLvl + Increment * (-1*cos(w*t)+1)/2;
figure(1)
plot(t,x)
Area = sum(x);
x2 = exp(-t/DampTime).* (-1*cos(w*t)+1)/2;
x2 = FirstLvl + Increment * x2/max(x2);
%%%%%%  Increment addjustment
while(sum(x2)<Area)
    Increment = Increment + .01;
    x2 = exp(-t/DampTime).* (-1*cos(w*t)+1)/2;
    x2 = FirstLvl + Increment * x2/max(x2);
end
%%%%%%
figure(2)
plot(t,x2)
AreaUnder = Area - FirstLvl * (Days+1)
dlmwrite(filename, x2, 'delimiter', '\n', 'precision', Precision);

s1=x';
s2=(x2)'
```

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