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Identifying A Mechanism For An Infiltration Threshold From The Sunflower River, Ms To The Underlying Alluvial Aquifer

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IDENTIFYING A MECHANISM FOR AN INFILTRATION THRESHOLD FROM THE SUNFLOWER RIVER, MS TO THE UNDERLYING ALLUVIAL AQUIFER

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ABSTRACT

Long-term groundwater level and stream stage measurements at a USGS coupled groundwater stream-gaging station located on the Sunflower River in Sunflower, MS show an apparent stage-threshold for infiltration to the underlying alluvial aquifer. This site is located near the center of a large regional groundwater cone of depression in the Mississippi River Valley alluvial aquifer. The USGS well (termed well 3 in this study) was thought to be completed in the regional shallow aquifer, though often recording anomalously high water levels relative to other wells in the region. The purpose of this research was to identify the responsible mechanism for the apparent stage-threshold for surface-groundwater communication. Two possible mechanisms were considered: (1) scour of infiltration-limiting fine-grained bottom sediments during high flow-rate events corresponding to higher stage, (2) and lateral infiltration at high stage into more permeable coarse grained sedimentary layers intersecting the stream channel at higher elevation.

A channel bed sediment survey was conducted over 100 km of the river at high stream stage. The stream bottom was composed of cohesive, fine-grained sediments, eliminating the first hypothesis as a viable mechanism. Entrainment of bottom sediments at the higher velocities did not expose coarser-grained, higher-conductivity bottom sediments. Cores were taken throughout the west bank of the river near the USGS well (well 3) along vertical and horizontal transects to measure variations in grain size. A more coarse grained layer (higher sand content) was identified at an elevation of 29.2-31.6 m (msl), consistent with the second hypothesis.

Two additional monitoring wells (wells 1 and 2) were installed 1 km upstream, 12 m deeper than well 3. Water levels from these wells were generally 7 m lower than in well 3, were more consistent with regional groundwater levels, and showed no response to short-term changes in stream stage. Well 3 appears to be screened within a perched aquifer which is in connection with the Sunflower River at high stream stage through coarse-grained layers intersecting the stream channel at the higher elevations. The two deeper monitoring wells are screened within the regional aquifer, with river-recharge limited to gradual drainage from the perched aquifer.

The results have important implications for groundwater assessment and management for the Delta region of Mississippi, especially concerning the role that streams play as potential sources or sinks for the Mississippi Valley Alluvial Aquifer.

LIST OF ABBREVIATIONS AND SYMBOLS

mm	millimeter
cm	centimeter
m	meter
km	kilometer
m^3	cubic meters
km^3	cubic kilometer
m/d	meter per day
s	second
m/s	meter per second
bls	below land surface
msl	mean sea level
K	Hydraulic Conductivity
MRVA	Mississippi River Valley alluvial aquifer
MERAS	Mississippi Embayment Regional Aquifer System
ADCP	Acoustic Doppler Current Profiler
USGS	United States Geological Survey
YMD	Yazoo Mississippi Delta Joint Water Management District

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER I

INTRODUCTION

A threshold is defined as a point that must be exceeded to begin producing a given effect or to elicit a particular response. In hydrologic systems, various threshold responses have been observed, such as minimum precipitation rates to initialize overland flow (Froehlich and Leszek, 1994), minimum soil saturation for surface water infiltration (Vogelmann et al., 2012), and minimum precipitation that must be exceeded to commence recharge to an aquifer (Jones and Banner, 2003).

In a contiguous or connected stream-groundwater system where a stream discharges to an underlying aquifer, the magnitude of changes in water level within the groundwater should be proportional to changes in the hydraulic gradient created by changes in the stream stage. Initiation of groundwater recharge from the stream should not be dependent on exceeding a threshold stream-stage unless a physical barrier exists that inhibits flow at the lower stage. An apparent stage-threshold, however, has been observed in a multi-year U.S. Geological Survey (USGS) record of groundwater levels and stream stage along a stretch of the Sunflower River, MS.

The Sunflower River is located on the Mississippi River floodplain, in northwestern MS (locally known as the “Delta”; Figure 1) and overlies the Mississippi River Valley Alluvial aquifer (MRVA). Groundwater and surface water hydrographs from a USGS stream gaging station and monitoring well (USGS NWIS) in the town of Sunflower, MS, show an apparent stage-dependent threshold before a response is observed in the groundwater (Figure 2). Water levels in a monitoring well located 40 m from the river appear to be independent of stream flow until the stage exceeds approximately 30 m mean sea level (msl). For flows above 30 m, groundwater levels visibly rise and fall with rising and falling stage. At the same time, high water levels in the well relative to other wells in the region suggested that anomalously high recharge was occurring from a localized stretch of the Sunflower River. The goal of the current study was to investigate the possible mechanisms controlling the observed behaviors.

The study site contained an existing USGS monitoring well and stage recorder on the bank of the river near the Sunflower Bridge on the west side of Sunflower, MS. Historical data from a five month interval in 2012 is shown in Figure 2 to illustrate the observed phenomena. Increases in water levels in the monitoring well followed closely in time after increases in river flow that exceeded a stage of 30 m, with no apparent groundwater response for changes in river flows that fell below the 30 m threshold. Two possible mechanisms that could produce a threshold effect were considered. The first proposed mechanism is scour of the riverbed during high discharge events. This process would remove the fines/silts settled on the riverbed during periods of low flow, entraining them in the suspended load, and potentially exposing coarser grained, more permeable sediments below. Only during these high stream stage events will the velocity be high enough to flush these fine-grained materials from the channel bottom (Bryant, 1992). After returning to low flow conditions, redeposition of these fines will settle back creating the ‘plug’ once again.

The second proposed recharge mechanism calls for lateral flow of surface water during high stream-stage events into more permeable, coarser grained sediments deposited higher on the banks of the channel. This process is primarily dependent on the presence of heterogeneity in strata upon the banks of the Sunflower River. If zones of higher permeability are encountered, surface water from the river will likely infiltrate laterally, then vertically to the underlying alluvial aquifer. Examining the near-surface geology will assess the cogency of this recharge mechanism.

Additional factors considered included possible mechanical issues such as biofouling of the monitoring well that could have biased the groundwater head data. Bio-films clogging a well screen can affect well yields, efficiency, and produce inaccurate data. Biofouling as a result of biofilm growth is a problem in many different fields, causing damage of product or interfering with production processes of industrial, scientific, and electronic equipment (Flemming et al., 1998).

CHAPTER II

GEOLOGIC SETTING

Sand, gravel, and silt with minor clay deposits of Quaternary age make up most of the Mississippi River Valley alluvial aquifer (MRVA). The aquifer ranges from 8 to 45 meters in thickness, with an average thickness of approximately 40 meters. The aquifer consists of braided channel deposits of gravel and coarse sand that is overlain by a finer sequence (confining unit) of sand, silt, and clay that was mostly deposited by a meandering river system (Renken, 1998). The fine-grained surface layers vary from 6 to 9 meters thick. These layers are often referred to as Sharkey clays and impede downward movement of water to the coarser sand and gravel beneath. These fine grained sediments "essentially occlude the porosity and form a 'seal' which hinders flow from the river to the aquifer" (Bryant, 1992). These clays were formed from fine-textured flood deposits from the Mississippi River, and are locally termed buckshot or gumbo clay" (Powell and Keenan, 1959). In the study area near the town of Sunflower, the bottom of the stream channel is contained within the surficial confining layer.

The MRVA, as well as the middle Claiborne aquifer, is a part of the larger Mississippi Embayment Regional Aquifer System (MERAS). The shallow MRVA is the primary source of groundwater for irrigation, while the deeper middle Claiborne is the primary source of drinking water. The wells that penetrate the MRVA are pumped by volume more than nine times the amount of the drinking water quantity via the Claiborne aquifer (Mason, 2010). Pumping for irrigation and aquaculture (catfish farming) has impacted the groundwater level in the alluvial aquifer significantly (Renken, 1998). A cone of depression has formed in the central portions of the Mississippi Delta around Sunflower County from groundwater withdrawal within the MRVA (Figure 1). The groundwater level in the Mississippi Delta from the MRVA has declined significantly, with an average depletion of approximately 0.45 m per year. The Yazoo Mississippi Delta Joint Water Management District (YMD) estimates the aquifer in the cone of depression has experienced a net loss of over 4 km^3 of water from 1987 to 2009 (Barlow and Clark, 2011). On average, withdrawals from the MRVA exceed 2.4 million m^3 during the 5-month growing season (Mason, 2010).

The MRVA is incised by several small rivers within the Mississippi alluvial plain, potentially establishing a hydraulic connection between ground and surface water. If in connection, water can move freely between the river and the alluvial aquifer. The connection between them depends on the permeability of the river bed strata and the extent to which the river penetrates the regional confining unit and shallow aquifer (Renken, 1998). Streams that are in connection either discharge or recharge the aquifer, thus delineating them as gaining or losing, respectively. This is dependent on seasonal variations in stream stage elevations. During periods of high precipitation in winter/spring, stream flow generally increases; surface water seeps into the river banks ultimately recharging the aquifer. In contrast, during summer/fall when stream flow is low, groundwater stored in the aquifer and river banks is discharged to the river. Because of the large amounts of groundwater withdrawals near Sunflower County, the levels in the aquifer have fallen well below the streambed of the Sunflower River. For those reaches where the Sunflower penetrates the regional confining layer, water is assumed to leak downward to the aquifer year round (Renken, 1988).

CHAPTER III

METHODS

The “stream bed scour” hypothesis was tested by conducting channel bed sampling for grain size analyses during a high flow event along a 100 km stretch of the river (50 km upstream and downstream of the primary study site, Figure 1). Sampling was conducted on March 12-16, 2012 at an average river stage of approximately 32 m msl (4.8 m water depth). A Ponar dredge was employed to take samples from the center of the river channel every kilometer.

For river flows above the apparent threshold stage, an initial simplistic approach was used to measure the characteristics of groundwater response observed in the USGS monitoring well (well 3 in this study) in relation to the stream stage of the Sunflower River. A cross correlation analysis was first used to determine an average lag time between peaks in river flow and corresponding peaks in well 3 groundwater levels. River stage at peak levels were then plotted against the change in groundwater level over the subsequent lag period. In principle, this approach could be used to quantify a threshold stage, below which there is no groundwater response, but it fails to capture the influence of smaller peak river-flows that diminish the rate of decline in groundwater levels (without creating an increase or peak in groundwater level).

If recharge to the groundwater occurs primarily via lateral flow along an isolated stretch of the river, rising stream stage will produce a disproportional recharge response in the groundwater level due to a simultaneous increase in the cross-sectional area available for inflow, even without stream bank heterogeneity. To test whether the groundwater response during high flow was greater than expected simply due to increased cross-sectional area (due, for example, to reaching a more permeable zone at higher elevation) a more rigorous flow model was employed (developed by Dr. Robert Holt, see Appendix C).

The inflow / outflow groundwater model was created to further investigate a method of identifying expected vs. observed results via change in the groundwater system and its relation to stream stage. It was developed for conditions of a perched aquifer of limited extent, receiving recharge derived from discharge through the banks of an overlying stream. The perched aquifer was assumed to discharge to the lower alluvial aquifer. The bottom of the stream was assumed to be impermeable, reflecting properties of the fine sediment bed layer. The stream bank stratum was assumed to be permeable and homogenous. Hourly stream-gage data were used from the Sunflower River with missing values linearly interpolated. By assuming the perched aquifer's recharge coincided with stream bank discharge, the model attempted to simulate the observed head fluctuations in groundwater from well 3 using a homogeneous stream bank lithology possessing a uniform hydraulic conductivity (K). If heterogeneity exists with higher permeable sediments at higher elevations, the modeled groundwater levels should underestimate groundwater response only at higher stream-stage.

To test the hypothesis that high river stage brings surface water in contact with more permeable zones, a series of shallow cores were collected for particle size analysis along 5 vertical transects spread horizontally over approximately 100 m. Points of equal elevation were flagged during low and high flow events before sampling. An elevation survey was also conducted using a GPS total station at approximately 100 stations along the west bank (same side as well 3). A total of 15 cores were collected, each 1 m in depth (Figure 3). The Sunflower River bank is terraced, with four benches at the study site at elevations of 35.2, 33.9, 31.6, and

29.2 m msl. Samples were collected between and on top of all four benches. Each core was mixed and sub-sampled (100 gram homogenized samples) to measure an average grain size at each sampled location. Grain size was determined by wet-sieve analysis followed by laser particle analysis of the finest fraction (Micromeretics Sedigraph).

Two monitoring wells (wells 1 & 2) were drilled 1 km north of the existing USGS monitoring well (well 3), on the west side of the river channel, in early 2013 (Figure 4). The purpose for the installation of new monitoring wells was (1) to determine if the results from well 3 were reproducible in additional wells and (2) to determine if the groundwater gradient orthogonal to the stream responded to changes in stream flow. Wells 1 and 2 were drilled to depths 24 meters below land surface (bls). Water levels were recorded every hour with dedicated data-logging pressure transducers. During the drilling, sediment samples were taken from every 1.5-meter interval. Stratigraphic columns were then produced for both boreholes to compare the subsurface lithologies. Elevations of the wellheads were surveyed to normalize all water levels relative to sea level.

CHAPTER IV

RESULTS

Stratigraphy

Subsurface stratigraphy for all three wells based on drill cuttings is shown in Figure 5. Wellheads 1 and 2 are at approximately the same elevation; wellhead 3 sits approximately 2 m lower. Well 1, located adjacent to the river 1 km north of well 3, was drilled to a depth of 24 m bls. The top 14 m is clay, followed by coarser grains with clay clasts from 14 to 18 m, and coarse sands and gravels from 18 to 24 m depth. Well 2, located 330 m away from the river to the west of well 1, was also drilled to a depth of 24 m bls. The top 9 m is clay, followed by sands with minor zones of clay clasts from 9 to 20 m, and coarse sand and gravels from 20 to 24 m. Well 3 was drilled to a depth of 13 m bls. The top 3 m is clay, followed by clay-rich sands from 3 to 6 m, and coarse sand and gravel from 6 to 13 m. The transition from clay to sand and gravel occurs at a much higher elevation than in wells 1 and 2.

Average water levels

A disparity was observed between the average groundwater elevations in the three wells. At the time of study, wells 1 and 2 were completed approximately 6 m below the water table, with water levels averaging near 21 m msl. The bottom of well 3 sits at an elevation of over 25 m msl, 4 m above the average water table measured in wells 1 and 2, yet the well contained groundwater with levels that ranged from 26.7 to 32.8 m msl over the course of the study.

Hydrographs

Figure 6 displays the Sunflower River gage height vs. groundwater elevations of wells 1, 2, and 3 (24 months for the river and well 3, 13 months for wells 1 and 2). Figure 7 shows the expanded dataset of gage height and well 3 for a period of 4 years. Water levels in well 1 and 2 track closely. Steady increases in water levels in both wells are observed when average stream flow is high, and decreases when average stream flow is low. Long term changes in water levels thus correlate positively with stream flow, but groundwater levels in these wells do not appear to respond to short term increases and decreases in stream flow. Water levels in wells 1 and 2 varied by less than 1 m over the period of study; well 3 water levels varied by more than 6 m.

An abrupt decrease in water level variation in wells 1 and 2, starting at 4/6/2013, corresponds with the date of flushing of the well screens (Figure 6). The difference in groundwater elevations between the two dropped from 15 cm to approximately 5 centimeters. The flushing process likely removed residual drilling mud present within the screen openings and rendered a better connection with the aquifer. With less obstructions within the openings, the well required less time to equilibrate with pressures from the aquifer, ultimately reflecting more accurate head levels. During the winter months when groundwater levels were rising, gradients were fairly low (9.0×10^{-5}) and sloped away from the river. During the summer months the groundwater level dropped approximately 1 meter. The gradient reversed, sloping toward the river (4.5×10^{-4}). The numerous irrigation wells on the east side of the river are likely attributed to the groundwater level sloping towards the river.

<u>Groundwater response in well 3</u>

A cross correlation analysis of collected data found the best fit between river stage and water level in well 3 with a lag time of 100 hrs. A plot of river stage at peak flow events vs. the increase in water level in well 3 over a subsequent 100 hr period is shown in Figure 8. Peak river flows only produced peaks in groundwater level when stream stage exceeded 30 m. Above 30 m, a best fit line is non-linear. A separate cross-correlation was run on the paired groundwater and stream stage hydrograph data above and below this apparent 30 m threshold elevation (with recognition that this does not capture additions to groundwater that may have slowed the decline in groundwater levels without creating a groundwater peak). A correlation coefficient of 0.78 was obtained above threshold level, and 0.43 below.

Ponar dredge samples

At high stage (32 m msl at the Sunflower gaging station), the Sunflower River bottom consisted of fine-grained sediments. Cohesive clay was encountered at all sampling points along the 100 km stretch. The dredge had difficulty penetrating the compacted bottom sediments, typically yielding less than 20 g of sediment. At many locations, particularly between 149-170 km (downstream of Sunflower), the dredge came up empty with minor amounts of clay sticking to the points of impact.

River-bank core samples

The shallow cores collected from the river bank near well 1(SW1-SC1 to -SC3) contained a high percentage of fines and a low percentage of sands (Table 1). The sand percentages ranged from 5-9%, with corresponding fines up to 91%. From the laser fine-particle analysis on these samples, up to 65% were finer than 1.7 microns. These were taken along a vertical transect, pulling a core from each exposed bench. Only three samples were collected and analyzed from the well 1 location. Because the groundwater data were not reflecting the localized threshold phenomenon, the remaining shallow core samples were focused near well 3 (USGS-C1 to -C13).

At the top of the river channel near well 3, there were relatively low sand and high fines percentages. Core #7 was taken between bench 3 and 4 and contained a fines percentage upwards of 98%. At low stream stage, near bench 2, the sand percentage increased up to 40% in two samples. The samples with the highest percentage of sands were located on or adjacent to bench 2. Core #9 contained the highest sand percentage at 48%. The majority of the sand grains (27%) were approximately 0.125 millimeters; termed a very-fine to fine sand. Of the smaller grain sizes, 18% were silt sized, and 32% were finer than 1.7 microns. Cores taken below bench 2 (USGS-C6, -C10) were again dominated by fines in excess of 70%.

Figure 3 shows an aerial and cross sectional view of the coring locations from the west bank of the Sunflower River, near well 3. Clay-rich facies bounded the sandy lens above and below. The grain size analyses relative to their spatial location show that a sandy lithology was present proximal to bench 2. The elevation of this sand-rich zone was 29.2 – 31.6 meters (msl).

Dynamic water-balance model

The dynamic water-balance model, developed by R.M. Holt from the University of Mississippi, was simulated to determine if the observed data from well 3 could be modeled for lateral infiltration through bank sediments with a constant hydraulic conductivity (no stage-dependent threshold). The model assumed an impermeable bottom, and accounted for increasing cross-sectional inflow area through the banks as the stream stage increased. The dimensions and hydraulic properties of the perched aquifer were optimized to find the best possible fit with the observed well 3 data (see Appendix C). Figure 9 shows a relatively good agreement between observed and modeled results at low stream stage, but significantly underestimated groundwater elevation response at a higher stream stage (above approximately 30 m).

CHAPTER V

DISCUSSION

The Yazoo Management Joint Water District (YMD) conducts an annual groundwater elevation survey within the Mississippi Delta. Figure 10 shows the potentiometric surface of wells 1, 2, and 3 in relation to the regional water table along a north-south transect. Wells 1 and 2 lie on the YMD's regional groundwater elevation profile, whereas well 3's elevation was much higher by an average of approximately 7.5 meters.

The anomalously high water levels in well 3 indicate that this well was drilled into a perched aquifer that comes increasingly into communication with the river during high flow events. Wells 1 and 2 appear to be completed in the regional aquifer, with water levels consistent with surrounding monitoring wells. Increases in water level in wells 1 and 2 corresponding to periods of higher stream flow may be caused by direct recharge from the river via minor river-bed leakage, indirect recharge from the river via drainage from the perched aquifer (steadily declining water levels in well 3 during periods of low stream flow indicates that drainage from the perched aquifer occurs readily when not actively resupplied), or by regional infiltration following periods of precipitation events that also produce higher river flow. Declining groundwater levels in wells 1 and 2 are observed in the summer months when irrigation withdrawals are high and precipitation levels declined. Figure 4 includes the location of irrigation wells in the study area. The reversal in the groundwater gradient in the summer months suggest that pumping for irrigation was heavier on the East side of the river during the period of study.

The remainder of the discussion focuses on the apparent stage-dependent threshold for communication between the river and water level in well 3. The thick surficial clay sequences encountered in wells 1 and 2 that extend to or below the river bottom, coupled with the presence of cohesive clay in the river bottom during high flow, is evidence that the bottom-scour hypothesis is not valid. The strength of bonds on clay particles are strong and require greater force than available in order to break them (Vanoni, 2006). Vertical infiltration into the MRVA will be severely limited without exposing permeable channel bottom sediments.

The non-linear relationship between river stage and groundwater level shown in Figure 8 is consistent with flux through an increasing cross-sectional area on the river bank as the river rises. In principle, a similar non-linear relationship would be generated for vertical flow with a stream bottom that widens as the river rises. However, the presence of clay on the river bottom effectively limits the possible explanation to infiltration through the river banks.

The absence of groundwater peaks for stream-flow peaks below 30 m (Figure 8) does not represent a genuine stage threshold for infiltration because it ignores groundwater responses to increased stream flow that fail to produce *increases* in groundwater level, but that may slow the rate of groundwater decline. If a threshold does exist, the results of Figure 8 serve only to set a maximum possible value (~30 m msl).

A more rigorous analysis is made possible by the dynamic water balance model. The simulation results provide a reasonable match to the observed well head levels in well 3 at lower river stage (Figure 9). It mimics the perched aquifer's response to small, short duration stream stage events but under-predicts them at large and longer duration events. These results lend support to the proposed mechanism 2, with (1) increasing hydraulic conductivity higher on the bank, or (2) rising stream stage encountering an increasing number of higher permeability zones resulting in a greater than expected groundwater response. The agreement between the model and observed water levels at lower river stage indicates that a hard threshold does not exist, where recharge is entirely prevented until the threshold is passed. Rather, a "soft" threshold is indicated where lateral recharge occurs at all stages, but above a threshold of approximately 29 m, zones of higher permeability are encountered that enhance recharge beyond what would otherwise be expected.

This result is consistent with the grain size distributions found in bank sediment cores. Figure 11 shows the inferred subsurface lithology of the west bank of the Sunflower River near well 3. The core results show that there were zones of high sand percentages between bench 1 and bench 2, correlating to river stages of 29.2 – 31.6 m. The perched aquifer is variably connected to the riverbank, possibly pinching out laterally, with aquifer overflow likely discharging to the regional alluvial aquifer and contributing to the seasonally rising groundwater levels in wells 1 and 2.

CHAPTER VI

CONCLUSIONS

The water bearing formation penetrated by well 3 is a perched aquifer that sits approximately 7 meters above the regional alluvial aquifer (MRVA). In contrast, wells 1 and 2 penetrate an aquifer reflecting average groundwater levels of the MRVA. The presence of thick cohesive clays on the river bottom at high stream stage argue against the bottom-scour hypothesis for exposing permeable channel bottom sediments. The presence of a 'hard' threshold where recharges to the perched aquifer occurs only at a specific bank elevation is not supported by the data. The dynamic water-balance model results match the observed well 3 head levels for continuous recharge at lower stream stage through a stream bank with a uniform hydraulic conductivity (K), but significantly underestimate well head levels at stages exceeding 29 m msl. Rather, a 'soft' threshold seems to exist; low rates of recharge occur below a stage of 29 m, and higher rates of recharge occur above, due to surface water encountering zones of higher permeability on the channel banks above 29 m. Sediment analysis from the west bank shallow cores are consistent with this finding, with sand-rich, higher permeability zones present from elevations 29.2 – 31.6 m msl.

The results from the research highlight the importance of groundwater assessment and management within the 'Delta' region of Mississippi and the potential significance of surface-groundwater interaction in this region.

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LIST OF APPENDICES

APPENDIX A: FIGURES

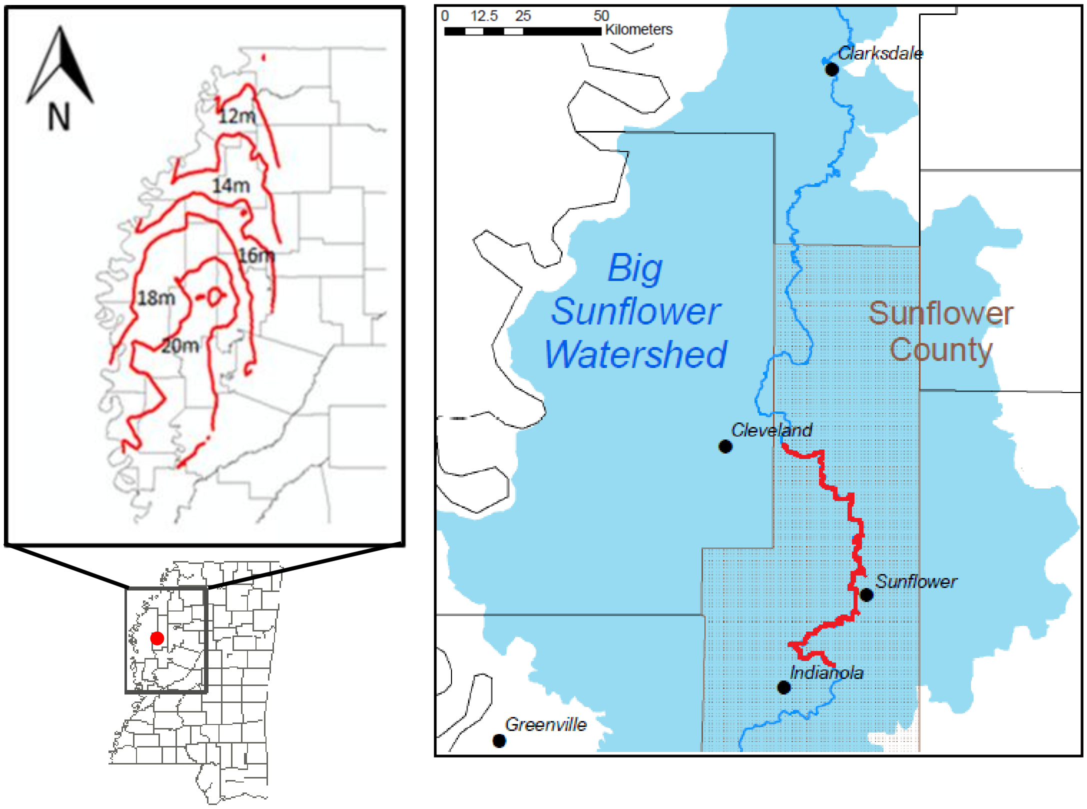


Figure 1. Study area showing the extent of the groundwater cone of depression (values are meters below surface), the Big Sunflower watershed, and the 100 km channel bed survey (red) of the Sunflower River (YMD regional groundwater survey, 2012).

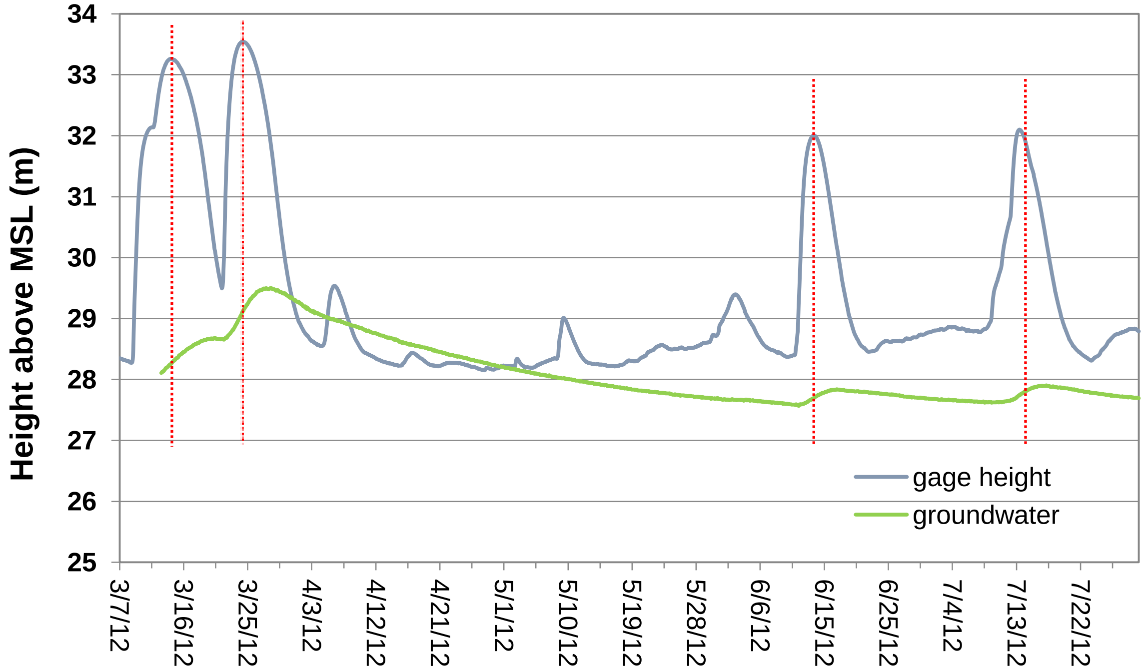


Figure 2. Sunflower River gage height and groundwater elevation illustrating the apparent threshold for infiltration. Obvious groundwater responses to stream flow in this data set occur only when stage exceeds 30 m (USGS NWIS).

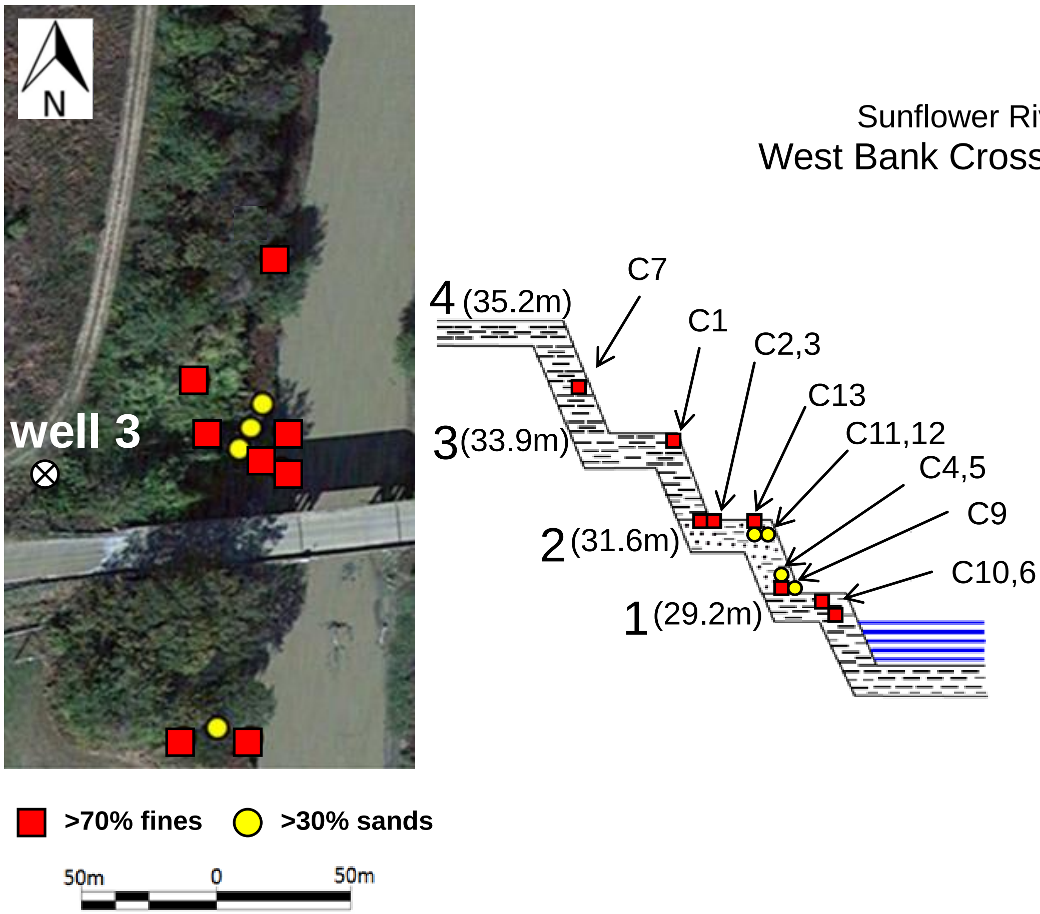


Figure 3. Shallow coring locations on the west bank of the Sunflower River, west of the town of Sunflower, MS. Numbers 1, 2, and 3 on the cross section refer to the flat benches present along the bank. Number 4 is the land surface.

Figure 4. Monitoring well and irrigation well locations, north of the town of Sunflower, MS; well 3 represents the USGS monitoring well.

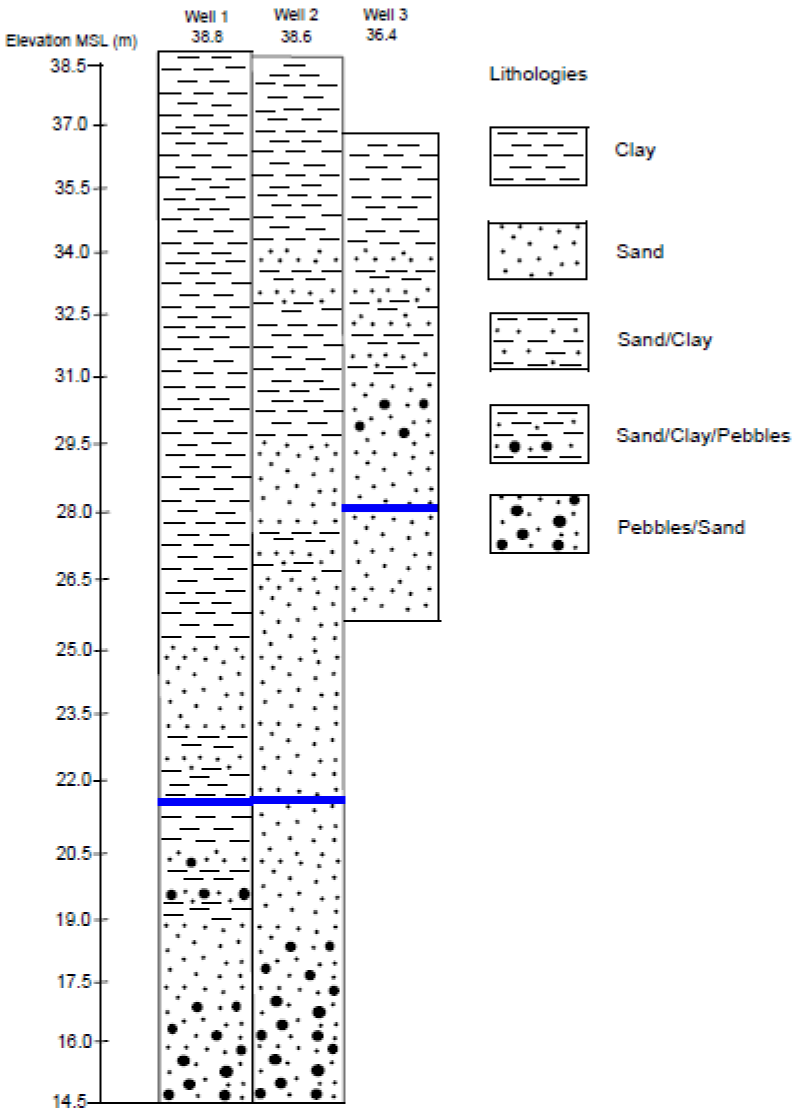


Figure 5. Wells 1, 2, & 3 subsurface stratigraphy and average water table elevations. The bottom of the river channel is approximately 27 m msl (well 3 data from Jeannie Barlow, USGS).

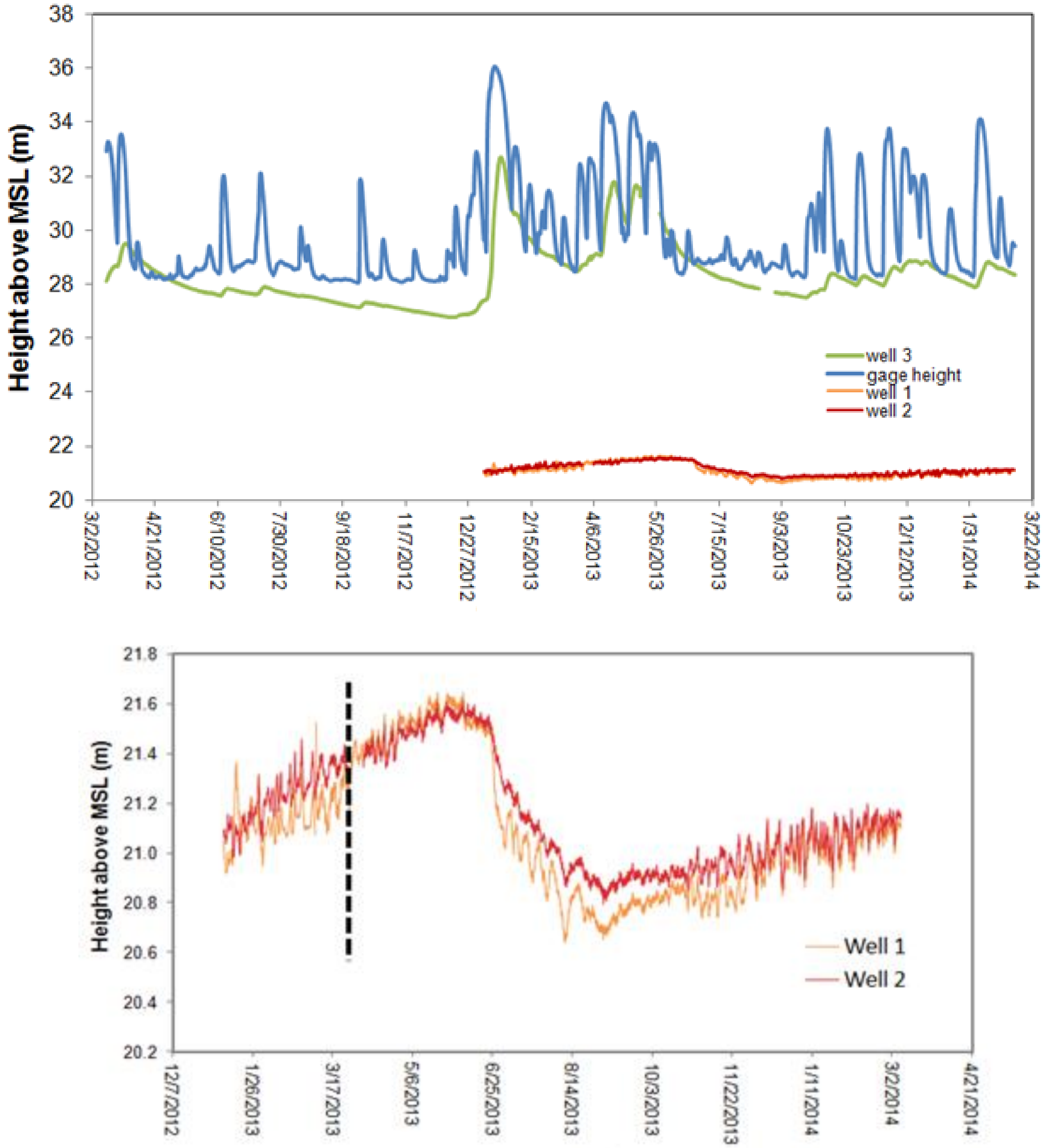


Figure 6. Sunflower River gage height and water level in wells 1, 2 and, 3 from 3/2/2012 – 3/22/2014 with expanded view of wells 1 and 2 to show changes in gradient direction. Vertical dashed line marks the date of flushing to clean screen of fines.

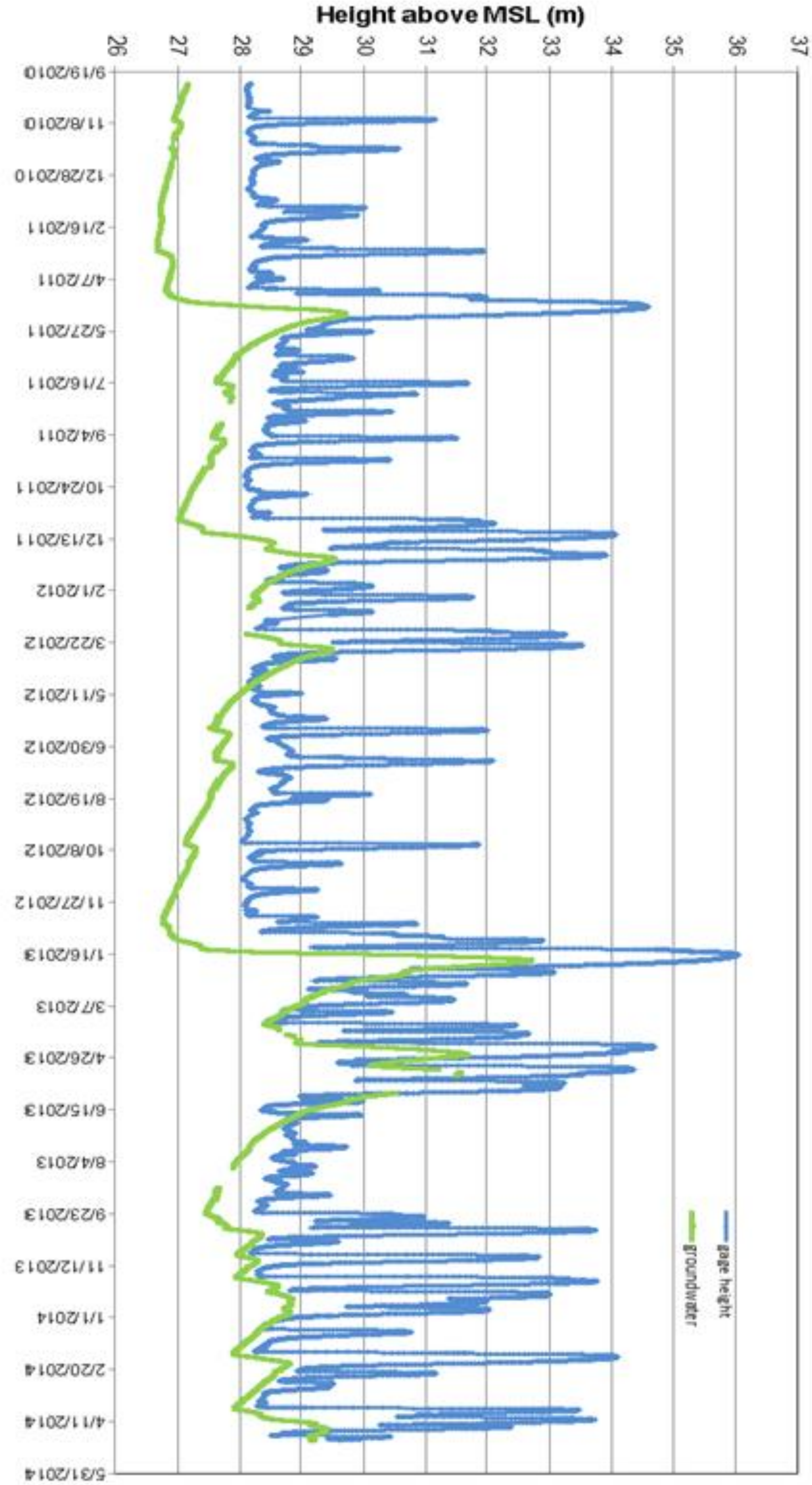
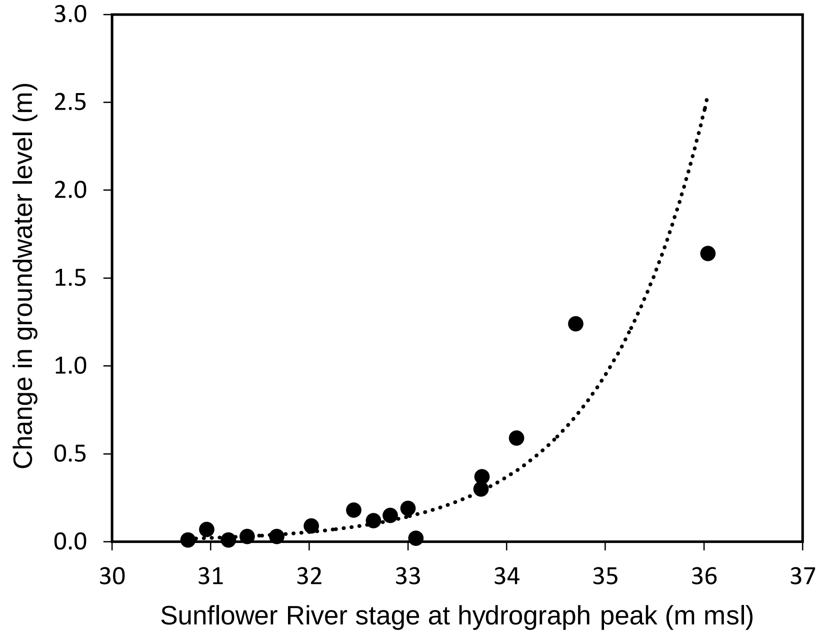


Figure 7. Sunflower River stage and well 3 groundwater levels from 9/1/2010 - 4/12/2014.



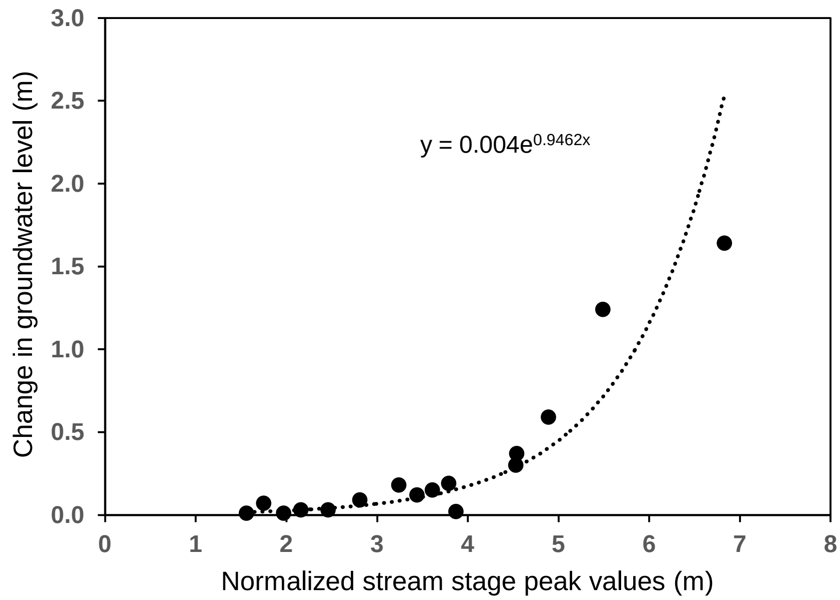


Figure 8. Plotted Sunflower River stream stage peaks vs. change in groundwater level for well 3 over the subsequent 100 hour lag time. Secondary graph depicts the stream stage peak values normalized to represent change above the apparent threshold stage of 30 m with an exponential best fit line.

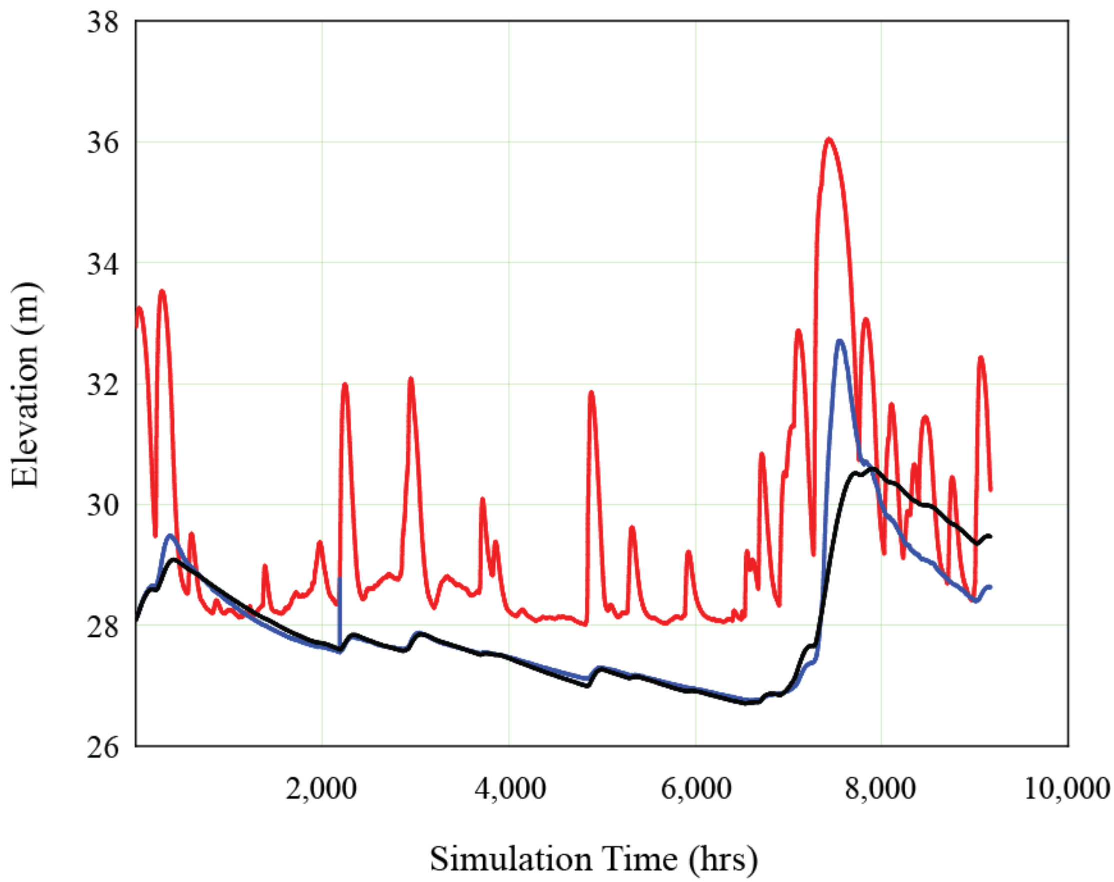


Figure 9. Modeled well 3 results (black line) plotted with observed well data (blue line) and stream stage (red line) (Holt, 2014).

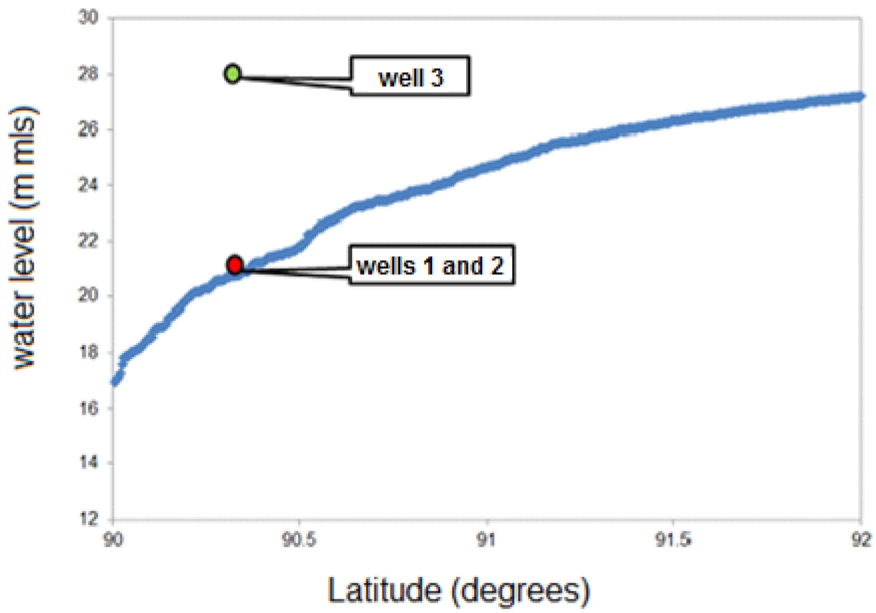


Figure 10. YMD regional groundwater survey showing regional water levels along a north-south transect (between 32.3° -34.9° longitude) (YMD regional survey, 2012), and average water level for wells 1, 2, and 3.

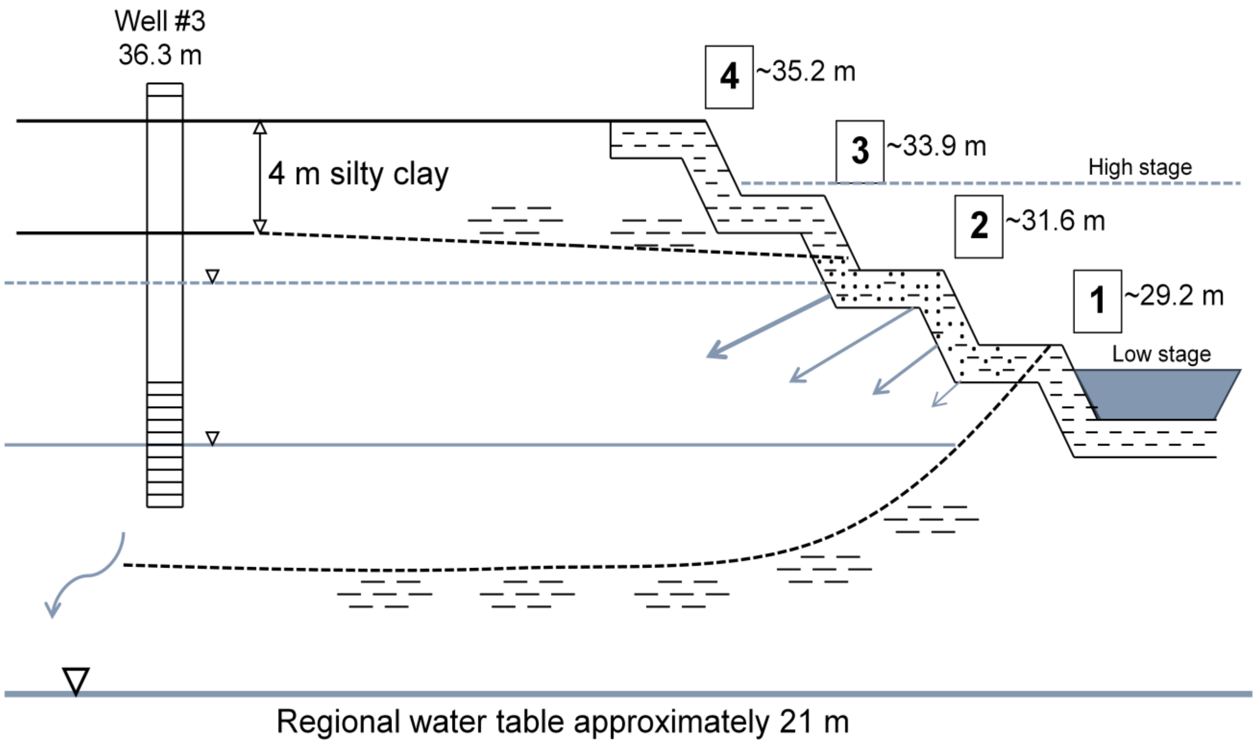


Figure 11. Sunflower River channel west bank: inferred subsurface lithology at well 3 (elevations relative to msl). Thicker recharge arrows to perched aquifer depict enhanced recharge at higher stream bank elevations.

Core	Sample weight (g)	Sands (g)	**Sand (%)**	Fines (g)	**Fines (%)**
SW1-SC1	100.32	8.85	**8.82**	88.24	**87.96**
SW1-SC2	100.02	7.73	**7.73**	91.20	**91.18**
SW1-SC3	99.98	4.61	**4.61**	91.20	**91.22**
USGS-C1	100.34	17.23	**17.17**	83.57	**83.29**
USGS-C2	101.62	27.32	**26.88**	73.47	**72.30**
USGS-C3	100.38	18.61	**18.54**	82.30	**81.99**
USGS-C4	100.64	24.13	**23.98**	77.23	**76.74**
USGS-C5	100.04	36.72	**36.71**	63.31	**63.28**
USGS-C6	100.76	25.88	**25.68**	75.15	**74.58**
USGS-C7	100.55	1.48	**1.47**	98.62	**98.08**
USGS-C9	100.27	48.19	**48.06**	50.66	**50.52**
USGS-C10	100.46	28.35	**28.22**	73.16	**72.83**
USGS-C11	100.82	31.62	**31.36**	67.09	**66.54**
USGS-C12	100.42	33.02	**32.88**	68.84	**68.55**
USGS-C13	100.70	28.10	**27.80**	71.50	**71.00**

Table 1. Shallow core grain size results.

APPENDIX B: STREAM VELOCITY MEASUREMENT AND WELL SLUG TESTS

Additional field work was completed during the course of the study that was not deemed useful in addressing the specific hypotheses for the thesis project, but may be pertinent to other studies and is presented here.

<u>Stream velocity measurement</u>

On 6/07/2013, an Acoustic Doppler Current Profiler (ADCP) was deployed for water velocity measurements. While in motion, the Doppler profiler took real-time water velocity measurements at all depths of the water profile. Velocities were obtained from the bottom of the water column. The ADCP transected the river a total of 10 times, collecting 10 transects of velocity and depth data. For testing the ability of high flows to scour bottom clays, this field work would have been better at high river stages. At the time of measurement, water depth was only 1.1 m. From the ADCP data, maximum velocity measurements near the channel bottom were recorded at approximately 0.8 m/s. Clays and clay properties are too varied to determine a single critical fluid shear stress or water velocity required to erode the cohesive clay-rich bottom sediments (Smerdon and Beasley, 1961; Vanoni, 2006), such as found along the bottom of the Sunflower River. Based on the Ponar dredge samples at higher river stage, the measured velocities were insufficient to erode down to the underlying sands.

<u>Well slug tests</u>

Slug tests were performed on wells 1 and 2 using the Hvorslev method (Butler, 1998). The hydraulic conductivity derived for well 1 (near the river) was 6.8×10^{-3} m/d. The hydraulic conductivity for well 2 (300 meters west of well 1) was 6.4×10^{-2} m/d. These values are lower than expected, given that Bear (1972) characterizes these as representing semi-pervious to impervious systems. They are also inconsistent with a well screen completed in sands and gravel.

APPENDIX C: DYNAMIC WATER-BALANCE MODEL DESCRIPTION

A DYNAMIC WATER-BALANCE MODEL FOR A PERCHED AQUIFER RECHARGED FROM THE BANK OF A STREAM

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The following documents a simple dynamic water-balance model for a perched aquifer that receives recharge from the bank of an overlying stream. A conceptual model for this scenario is shown in Figure B1. The aquifer is conceptualized as a linear reservoir with recharge derived from the discharge of a stream through its bank. The base of the stream is assumed to be impermeable, while the stream bank is assumed to have a layer of material different from that of the underlying aquifer with a thickness B_1 and a hydraulic conductivity K_1. The slope of the stream bank is given by θ. The stream stage is the elevation h_I, and the elevation of the base of the stream is H_B. The volumetric discharge from the stream bank is considered to be the only source of recharge to the aquifer and is given by

$$Q_{in} = wp(h_I)W\frac{K_1}{B_1}\left(h_I - h\right), \tag{1}$$

where W is the unit width of the aquifer/stream system, h is the average hydraulic head in the aquifer, and $wp(h_I)$ is the wetted perimeter of the stream bank given by

$$wp(h_I) = \frac{h_I - H_B}{\sin(\theta)}. \tag{2}$$

The perched aquifer (Figure B1) is considered to be a linear reservoir, where all spatial variations in hydraulic properties and hydraulic head are ignored. The average hydraulic head in the aquifer is h, the perched aquifer discharges to a lower aquifer at some characteristic distance L_a from the stream, and the elevation of the perched aquifer outflow is h_o. The volumetric discharge of the aquifer is given by

$$Q_{in} = A\alpha\left(h - h_o\right), \tag{3}$$

where A is the area of the aquifer and α is an outflow coefficient. Gelhar and Wilson (1974) found that α can be given by

$$\alpha = \frac{\beta T}{L^2}, \tag{4}$$

where T is the mean aquifer transmissivity and L is the distance from the discharge point and a groundwater divide, here assumed to underlie the stream. The coefficient β is a geometry term that typically ranges between 2.5 and 3.5. Gelhar and Wilson (1974) use a value of 3.0, which we also use here.

The change in aquifer storage is due to fluctuations in the recharge from the stream bank can be expressed as

$$AS_y \frac{\partial h}{\partial t} = Q_{in} - Q_{out}, \tag{5}$$

where S_y is the specific yield of the aquifer, here assumed to be equivalent to the average porosity of the aquifer. If we assume that the area of the aquifer is given by

$$A = L_a W, \tag{6}$$

use equations (1), (2), and (3) in equation (5), and divide the result by the area of the aquifer, equation (5) can be expressed as a non-linear, first-order, ordinary differential equation

$$S_y \frac{\partial h}{\partial t} + \left[C\left(h_I - H_B\right) + \alpha \right] h = C\left(h_I - H_B\right) h_I + \alpha h_o, \tag{7}$$

where C is a constant given by

$$C = \frac{K_1}{B_1 L_a \sin(\theta)}. \tag{8}$$

Equation (7) can be expressed as a backwards-in-time finite-difference equation

$$S_y \frac{h^{n+1} - h^n}{\Delta t} + \left[C\left(h_I^{n+1} - H_B\right) + \alpha \right] h^{n+1} = C\left(h_I^{n+1} - H_B\right) h_I^{n+1} + \alpha h_o, \tag{9}$$

where the superscript n refers to the *nth* time step. Equation (9) can be solved for the average aquifer head at the n + 1 time step using

$$h^{n+1} = \left[\frac{\Delta t}{S_y + C\left(h_I^{n+1} - H_B\right) + \alpha} \right] \cdot \left[S_y \frac{h^n}{\Delta t} + C\left(h_I^{n+1} - H_B\right) h_I^{n+1} + \alpha h_o \right]. \tag{9}$$

Equation (9) was applied to part of a data set provided in this thesis by Austin Patton of the University of Mississippi for the Sunflower River near Sunflower, MS. These data consist of hourly stream-gauge measurements and corresponding measurements in a nearby well. The object of the simulation was to determine if the model described above could reasonably mimic the data observed in the well.

The following parameter values were assumed for this simulation:

T = 12.6 m^2/hr,
K1 = 0.018 m/hr,
La = 800 m,
Δt = 1 hr,
Sy = 0.35,
ho = 22.772 m,
HB = 26.762 m,
B1 = 0.8 m, and
θ = 30 degrees.

Hourly stream-gauge data from the period between 3/13/2012 and 3/29/2013 were used to provide the forcing function h_I. Missing data were linearly interpolated to insure that the stream-gauge record for that period was complete. The initial head in the aquifer was assumed to be equal to that measured in the well at midnight on 3/13/2012 (28.107 m).

Simulation results are shown in Figure 9. The simplified dynamic water-balance model provides a reasonable match to the observed head values in the well. The model is particularly adept and mimicking the aquifer response to short duration and small to moderate river-stage events. For longer duration and larger increases in the river stage, the model under-predicts the aquifer response. It is likely that these discrepancies between the model and the observed well data could be resolved by the hydraulic conductivity of the stream bank to increase with elevation.

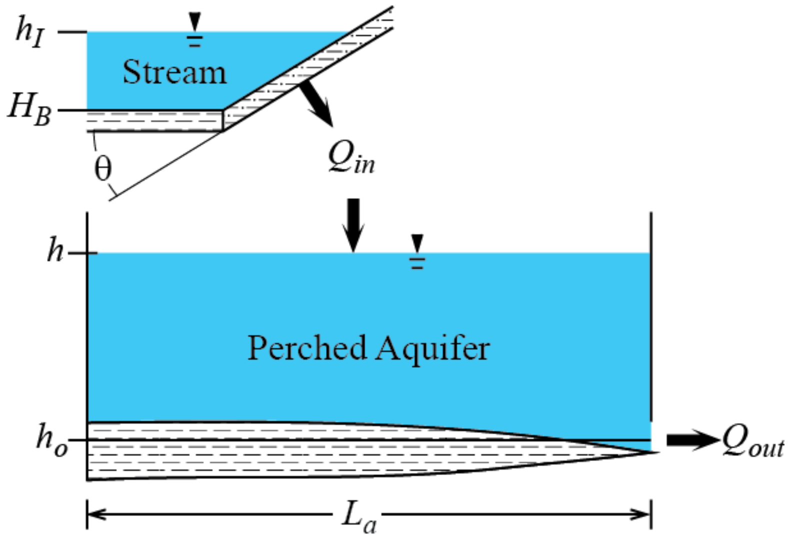

Figure B1. Conceptual depiction of a simple dynamic water-balance model for a perched aquifer that receives recharge from the bank of an overlying stream. The base of the stream is assumed to be impermeable, while the stream bank is assumed to have a layer of material different from that of the underlying aquifer. The slope of the stream bank is given by θ. The stream elevation is h_I, and the elevation of the base of the stream is H_B. The volumetric discharge of the stream into the aquifer is Q_{in}. The average hydraulic head in the aquifer is h, the perched aquifer discharges to a lower aquifer at some characteristic distance L_a from the stream, and the elevation of the perched aquifer outflow is h_o. The volumetric discharge of the aquifer at the outflow point is Q_{out}.

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Dean’s List-Fall 2011
Chancellor’s List-Fall 2010, 2012