RIVER LOSSES AT A KARST ESCARPMENT DURING NORMAL FLOW AND FLOOD CONDITIONS AND IMPLICATIONS FOR CARBONATE WEATHERING

By

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A THESIS PRESENTED TO THE GRADUATE SCHOOL
OF THE UNIVERSITY OF FLORIDA IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS FOR THE DEGREE OF
MASTER OF SCIENCE

UNIVERSITY OF FLORIDA

2012

To Stanley – The best cat that ever lived

ACKNOWLEDGMENTS

I would like to thank everyone in the world I have ever met for allowing me to conclude this glorious manuscript of awesome and amazing. I would also like to thank the people who came up with the idea for groundwater flow modeling, for without them I would not be borderline alcoholic. I would also like to thank my adviser, Elizabeth Screaton, who had the patience and understanding to not kick me out of the program after countless follies; and who also appreciated my impeccable timing to disturb her during her lunch hour. I also appreciate the help of Jonathan Martin who took pity on a girl who had grown up in South Florida and not even set foot in the Bahamas. He brought me along for a hell of a ride and I would like to think my future partly lies in those islands. I also thank the lab group for putting up with me and not complaining….too much. I would also like to thank the guy I'm dating for showing me some amazing things along the way - I can't wait to see more. I also thank my mom, ya know, for giving birth to me and stuff. And for being more than supportive when she had no idea what I was ranting about. Praise Jesus.

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May 2012

Chair: Elizabeth Screaton
Major: Geology

Surface water intrusion into karst aquifers can play a major role in carbonate weathering processes. At the Cody Scarp, a karst escarpment which marks the transition from the confined to unconfined Floridan aquifer in northern Florida, these surface and groundwater interactions include both temporary losses of river water to the aquifer during floods and losses of river water to the aquifer under normal (recession) conditions. The losses during floods include reversals of springs, in which river water flows into springs that normally discharge water. This research investigates the losses of water during floods and under normal flow conditions within the Madison Blue springshed near the Cody Scarp. Losses during normal flow are obtained through discharge data available through the United States Geological Survey (USGS) and flowpaths are investigated with numerical modeling. During floods, spring discharge gages become unreliable, and numerical modeling is employed to constrain volumes of water that intruded into Madison Blue spring during 2009 and 2010. These volumes were used in conjunction with existing chemistry data to determine the relative contributions of river loss and diffuse recharge to subsurface carbonate weathering.

Under non-flood conditions, river loss at the scarp contributed the largest part to the carbonate weathering budget. Dissolution during floods contributed a significant amount of subsurface carbonate weathering, however the short timescales of the reversals limit the amount of dissolution that occurs. Simulated residence times of intruded water are shorter than expected equilibration times, consistent with observations that water is undersaturated with respect to calcite once it begins discharging from the spring. As a result, calculations that assume equilibrium result in overestimation of dissolution. In contrast, water lost from the river under normal conditions will likely reach equilibrium in the subsurface. Thus, although reversals aid in conduit dissolution, contributions of water lost under normal conditions should not be ignored. With time, losses of water at the scarp could develop a sink rise system.

CHAPTER 1
INTRODUCTION

At the Cody Scarp, which is the transition between confined and unconfined conditions for the upper Floridan aquifer in Florida (USA), intensive development of karst geomorphic features and complex flow dynamics exist (Smith et al. 2002; Upchurch 2002). The extensive karst development at this transition zone is partially the result of the chemistry of surface water coming off the confining unit. Since river water upstream of the scarp is hydraulically decoupled from the upper Floridan aquifer and thus doesn't come into contact with much limestone, it is generally undersaturated with respect to calcite and therefore thermodynamically capable of dissolving limestone. The flow paths of this water therefore control where dissolution will take place. Under normal conditions, all rivers lose water after they cross the Cody Scarp; several rivers have their entire flow captured and some only have a fraction of river flow diverted underground by siphons while the remainder of losses may occur through diffuse flow through the river banks. Furthermore, when the river discharge increases, temporarily increased losses occur to the banks and springs which normally discharge into the river (Ford and Williams 2007; Gulley et al. 2011). Bank storage and spring reversals occur after a prolonged rain event when allogenic runoff from upstream of the escarpment causes water levels in the rivers to rise faster than hydraulic heads in the aquifer. These reversals are short-lived, and losses that occur due to normal conditions at the scarp prevail for a longer period of time.

In this investigation, we compare the estimated dissolution due to losses during normal flow conditions and due to bank storage and spring reversals during floods. This work will test the hypothesis that the amount of dissolution as a result of the more

prolonged losses from the river under normal conditions will contribute more to subsurface carbonate weathering than floods, which occur infrequently and are short lived.

The magnitude and extent of dissolution is related to the flux of undersaturated water that enters the subsurface as well as the limestone area exposed to the undersaturated water. Reversals as a result of floods that occur in deeply buried, compacted karst regions such as near Mammoth Cave Kentucky can introduce thousands of cubic meters of water into the aquifer (White and White 1989). However these systems, termed telogenetic karst, have low matrix hydraulic conductivity (10^{-6} m/day). Furthermore, the conduits may not be full during normal conditions and provide storage space for the incoming river water. As a result, the majority of the river water remains within the conduit (Vacher and Mylroie 2002). In contrast, the Floridan aquifer system is an eogenetic karst system (Vacher and Mylroie 2002) that has not undergone deep burial and uplift. It comprises a triple porosity system which includes flow through the interstitial pore spaces, small fractures, and conduits. Since it has not undergone deep burial or uplift, the eogenetic limestone has a matrix hydraulic conductivity that is several orders of magnitude higher than more deeply buried telogenetic counterparts (White 2002). In Florida, estimated matrix hydraulic conductivity is on the order of 10^{-1}-10^{1} m/day, and therefore a greater contribution of spring flow results from groundwater in the limestone matrix than in telogenetic karst (Florea et al. 2006; Moore et al. 2009; Bailly-Comte et al. 2010). This higher hydraulic conductivity will allow higher intrusion rates of river water into the matrix, and therefore a higher amount of dissolution. In addition, many conduits in north Florida are below the water table. As a result, river

intrusion into the conduits displaces water into the surrounding matrix and raises the water table.

The volume of water that intrudes into the aquifer during floods can be difficult to quantify. The loss of water during normal flow (conditions where river stage is below flood stage and/or there is no reversal of water into the spring) can be quantified using available discharge data from streams. In contrast, spring-run gages can become unreliable due to water levels exceeding maximum stage for accurate readings at the gage or become inoperable due to high flow that can dislodge gaging stations or break the equipment. In addition, velocity meters can be affected by turbulent eddies during reversals. Gulley et al. (2011) used monitoring of specific conductivity to document intrusion of Withlacoochee River water into the Madison Blue conduit system in north Florida during a 2009 flood. Unfortunately, lack of reliable flow data required volumes to be estimated using assumptions of flow velocities based on the timing of a drop in specific conductivity in the conduits. Numerical modeling provides a potentially valuable tool to assess the variations in velocity during an event and thus the volume of intruded water. In addition, numerical modeling can help assess flow paths for river water that entered the aquifer.

In order to accurately address exchange from conduit to matrix and the fate of water once it intrudes into the formation, conduits in the subsurface need to be accurately simulated. Previous karst aquifer simulations have simulated conduits using high hydraulic conductivity zones (Scanlon et al. 2003), however this approach makes it difficult to track the exchange between conduit and matrix. Using high hydraulic conductivity zones to simulate conduits also neglects head loss due to turbulent flow. As

a result, exchange between the conduit and formation might be overestimated during reversals because conduit hydraulic heads will be overestimated. Conduit Flow Processes (CFP) is a new MODFLOW module that simulates turbulent flow in the conduits and tracks exchange between conduits and porous media (Shoemaker et al. 2008).

This research examines surface and groundwater exchanges during flood and normal conditions and their role on subsurface dissolution. As part of the study, numerical modeling is used to constrain the amount of river water intrusion into a spring during floods, and to understand the fate of that water after it intrudes into the aquifer. The numerical modeling uses the MODFLOW code to simulate Darcian flow in the aquifer matrix and the CFP module to simulate turbulent flow through the conduit network. The simulated fluxes are integrated with previously reported chemistry data to estimate rates of dissolution in the aquifer during 2009 and 2010. Dissolution due to losses of water under normal conditions and flood conditions are compared, as well as how these estimates compare to the overall dissolution from diffuse recharge.

CHAPTER 2
BACKGROUND

Study Site

Madison Blue is a first magnitude spring that discharges from the upper Floridan aquifer (UFA) in northern Florida (Figure 2-1). The UFA in this region comprises the Ocala and Suwannee Limestone (Figure 2-2 and 2-3). The upper Ocala limestone is primarily muddy and granular limestone that is generally white and porous, friable and has a micritic matrix (Randazzo and Jones 1997; Budd and Vacher 2002). Borehole logs obtained from the Florida Department of Environmental Protection (FDEP) indicate that the thickness varies from 20-60 meters in within the springshed (FDEP Website http://www.fdep.org, borehole locations shown on Figure 2-1). The lower unit is partially or completely dolomitized and is not influential in groundwater flow in this area (Randazzo and Jones 1997). The Ocala limestone crops out near the Withlacoochee and Suwannee Rivers' confluence, and is present in the shallow subsurface elsewhere in the springshed. Estimates of hydraulic conductivity from permeameter tests for Ocala Limestone from West-central Florida range from 0.73-5.4 m/day (Budd and Vacher 2002). Estimated Ocala Limestone hydraulic conductivity from slug tests in O'Leno State Park ranged from 1.5-19 m/day (Langston et al. in press). Also at O'leno State Park, Martin et al. (2006) estimated hydraulic conductivities for the Ocala Limestone in Florida of about 10-2000 m/day; however the high end estimate of 2000 m/day likely incorporates conduit flow.

The Suwannee limestone is more prominent at the surface in the springshed, frequently outcropping on the banks of the Withlacoochee River (Scott 1990; Randazzo and Jones 1997). This Oligocene aged limestone is variably vuggy and muddy and is

fossil rich and white to cream colored (Budd and Vacher 2002). Borehole logs indicate Suwannee Limestone thickness ranges between 10- 40 meters throughout the springshed (FDEP http://www.fdep.org) (Figure 2-3) Overlying these strata in the northern portion of the springshed is the Miocene age Hawthorn Group. This formation is made up of siliciclastic sediments, primarily clay and phosphatic sand, and serves as the confining unit for the underlying UFA. In the Madison Blue springshed, the Hawthorn Group is estimated to be less than 5 meters in thickness based on borehole log data and does not form a continuous layer.

Madison Blue spring is the discharge point of a cave system that has been mapped by cave divers to a distance of 1869 meters from the entrance (Figure 2-5). The average diameter of the conduit system has been estimated to be 3 meters based on cave diver surveys (Gulley et al. 2011). Madison Blue spring is an estavelle and frequently reverses (Farrell and Upchurch 2004). Reversals are primarily indicated by tannic water observed in the springs and closure of the swimming and diving area (Park Ranger at Madison Blue Springs State Park Personal Communication 2010). Based on long-term data, Madison Blue Spring contributes about 30% of the discharge to the Withlacoochee River (Farrell and Upchurch 2004). The headwaters of the Withlacoochee River are in southeastern Georgia and the river flows for 17 kilometers before reaching the Cody Scarp north of Pinetta (Figure 2-1). The Withlacoochee River then flows for another 10 kilometers before entering the Suwannee River (Figure 2-1). Several springs discharge into the Withlacoochee River downstream of Madison Blue Spring, however only two springs, a second and third magnitude spring, lie within the boundaries of the springshed (Figure 2-1) and both are unmonitored. Results of dye

tracing conducted by cave divers in 2001 suggest that these springs are connected to the Madison Blue cave system (Farrell and Upchurch 2004) (Figure 2-1).

Where the Withlacoochee River transitions from confined to unconfined conditions, Fennels Funnel, a siphon large enough for cave divers to enter but not yet explored due to the flow (Bob Schulte Personal Communication 2011) diverts a portion of river water underground. Fennels Funnel is the only known siphon on the Withlacoochee River between Pinetta to Madison, and is known to be connected to another cave system in the area known as M2 Blue based on reports of cave divers connecting a conduit to the river (Robert Schulte, Personal Communication, 2011). Though the siphon diverts river water, additional losses may also occur through the stream bed and banks. Stage at the Withlacoochee River near the Cody Scarp varies from ~16.5 meters above sea level (masl) at low flow to the highest recorded value, 26 masl, in 2009. Flood stage at Pinetta is 24 masl and action stage is 18 masl (Figure 2-7).

Following Grubbs (1998), the UFA in this region is primarily under unconfined and poorly confined conditions, consistent with the thin and discontinuous Hawthorn Group recorded in boring logs. Land use is primarily agricultural. Two cities in the springshed, Madison and Lee, use the UFA for public water supply, and about 400 private wells exist in the area with pumping estimated at less than 1.7×10^5 m^3/day (Farrell and Upchurch 2004). In addition to these withdrawals, the Nestle Corporation also pumps from the UFA at a site around 1 kilometer from Madison Blue Spring for bottled water with a maximum allowable pumping rate of 7,570 m^3/day for both wells (Trento 2009).

Previous Work

Spring Reversals

Intrusion of surface water into springs (or reversals) during floods been previously recognized; however most work concerned contaminant transport and groundwater quality issues and the consequences of carbonate erosion from these processes were not investigated. Quinlan (1973) investigated reversing springs and addressed the impacts of pollutant and sewage transport as a result of reversals along the Green River in Kentucky. Alberic et al. (2003) recognized that locally heavy rainfall would cause frequent (10 times in 4 years) backflooding into a spring in France on the order of tens of cubic meters per event, and examined how the influx of organic matter into these systems might affect groundwater quality in the aquifer. Studies in Florida on the effects of river intrusion into the aquifer documented chemistry changes at wells in the aquifer that occurred due to river reversals from floods. Crandall and Katz (1999) found that wells up to 1.2 kilometers away from the Santa Fe River were undersaturated with respect to calcite and rich in organic matter. This research exemplified the importance of these surface and groundwater exchanges on karst aquifer chemistry and hydrology.

Madison Blue Spring

Gulley et al.(2011) estimated rates of carbonate weathering due to spring reversals during floods along the Withlacoochee and Suwannee Rivers in Florida. That research focused on the role of reversals in conduit enlargement. Based on specific conductivity monitoring within the cave system, Gulley et al. (2011) estimated the amount of river water intrusion into Madison Blue Spring to be about 2.5×10^5 m^3 over 7.5 days of reversal during the 2009 flood event. The river and spring gaging stations

were inoperable for the event, and the volume of water intrusion was estimated by using a conduit diameter of 3 m and calculating velocities using the observed specific conductivity pulse moving from the entrance to monitoring points deeper in the cave system. Flow velocity was assumed to linearly decrease velocity after the observed specific conductivity pulse occurred. The linear decrease in velocity resulted in a linear decrease in flow rate into the aquifer through time. A few water chemistry samples were obtained to estimate carbonate dissolution during this time, and potential dissolution was estimated assuming the intruded water reacted to reach equilibrium. Based on these results and observations of cave wall scallops, Gulley et al. (2011) argued that backflooding was a significant contributor to cave enlargement. Further work on the Madison Blue Spring by Brown et al. (2011) studied the effects of river water intrusion on redox chemistry changes in the spring during a small event in spring 2011. Brown et al. (2011) found that the redox changes due to the intruding river water brought trace metals into the aquifer leading to precipitates, and mobilized trace metals already within the aquifer.

Low Flow River Losses

Surface water intrusions also occur during low flow conditions. As rivers flow above the confining unit in Florida, interaction with the UFA is inhibited by the siliciclastic layer. When the rivers flow across the Cody Scarp, they can lose some or all of their water to the aquifer (Smith et al. 2002; Upchurch 2002). Upchurch 2002 recognized the possibility of intense karstic development at this transition, and sinking streams occur along the Cody Scarp. Moore et al. (2009) investigated conduit enlargement and the hydrochemical consequences imposed by these scarp losses at the Santa Fe River sink-rise system in central Florida’s O’Leno State Park. Water

sampled at the Santa Fe River sink and rise was undersaturated with respect to calcite during all 15 sampling periods, and it was determined that dissolution of calcite in the system could be observed along the conduit for 5 out of 15 times of water sampling. This dissolution was observed to occur during high flow events that pushed water into the matrix by increasing the hydraulic head in the conduit. Following the event, water was late released from storage and calcite dissolution was observed in the water chemistry at the rise. The lack of dissolution seen in 2/3 of the sampling was suggested to result during low flow conditions when the inflow of equilibrated water from the high permeability matrix inhibits contact of undersaturated river water with the conduit walls (Martin and Dean 2001, Moore et al. 2009).

Previous Numerical Modeling

Numerical modeling of North Florida has been focused on regional groundwater flow. Bush and Johnston (1988) modeled the Upper Floridan Aquifer in order to understand groundwater flow and the impact of development on the aquifer. The steady-state model simulated confining units as well as surface water bodies, and incorporated the interaction between all distinct sources of water, but the scope of the model was too broad to look at surface and groundwater interactions on a local scale. Planert (2007) simulated groundwater flow of the Upper Floridan Aquifer in the Suwannee River Basin. This one layer model used MODFLOW to look at regional, steady state groundwater flow and to understand the groundwater system and surface and groundwater interactions. Schneider et al. (2005) constructed a three dimensional model that examined at the interactions between the Surficial, Intermediate and Floridan aquifer systems. This was a steady state simulation to understand the hydraulic properties and dynamics of flow between these aquifers, as well as be the basis for

evaluating permits for consumptive use with respect to imposed minimum flows and levels. The extent and regional nature of all three models required combining fracture, diffuse and conduit flow into representative hydraulic conductivities. The cell sizes ranged from 1524 x 1524 meters to 13,000 x 13,000 meters, and the hydraulic conductivity zones specified in the models combined the unconfined portion of the springshed with adjacent confined areas. Values of hydraulic conductivity for the Madison Blue spring area ranged from 1000-3000 m/day (Bush and Johnston 1988, Schneider et al. 2005).

Grubbs and Crandall (2007) constructed a transient flow model to investigate the exchange of water between the Santa Fe and Suwannee Rivers and the Floridan Aquifer for a period of two years. Due to the scope of the model, localized conduit flow was not included. The investigation focused primarily on how groundwater affects river flow based on climatic conditions, and therefore investigations of the implications of river reversals into springs during floods were not detailed.

Dissolution

In order to determine whether a solution is in chemical equilibrium with a mineral phase such as calcite, the saturation index of that mineral phase is estimated using water chemistry data and thermodynamic constants. The resulting saturation index is calculated by the following equation:

$$SI_{cal} = \log(\frac{IAP}{K_{sp}}) \tag{2-1}$$

Where

SI_{cal}= Saturation index of calcite

IAP = Ion activity product (moles)

K_{sp} = Solubility product (moles)

Where IAP is the product of the activities of calcium and carbonate measured in the system in moles (M), and K_{sp} is the solubility product based on the Gibbs Free Energy of the system in question, K_{sp} is representative of the amount of calcium and carbonate that would be in the system if it were in equilibrium with calcite at a given temperature. If the value of SI_{cal} is less than 0, then the solution is thermodynamically capable of dissolving calcite in order to reach a steady state. If it is greater than 0, the solution is supersaturated with respect to calcite and may precipitate calcite in order to reach the equilibrium (SI = 0).

Though the saturation index will establish whether or not a solution is in equilibrium, it gives no information about the rate of reaction. Carbonate dissolution rates are affected by the saturation state of the water as well as the transport of solute away from the reaction site. After about 80-90% saturation, the reaction orders shift from first to fourth order (Plummer et al. 1979; Eisenloher et al. 1998; Dreybrodt et al. 2005), and the rate of calcite dissolution decreases. To understand the kinetics of calcite dissolution in natural settings, it is important to know the hydrodynamic mass transport properties as well as the heterogeneous surface of the mineral reaction site (Plummer et al. 1979).

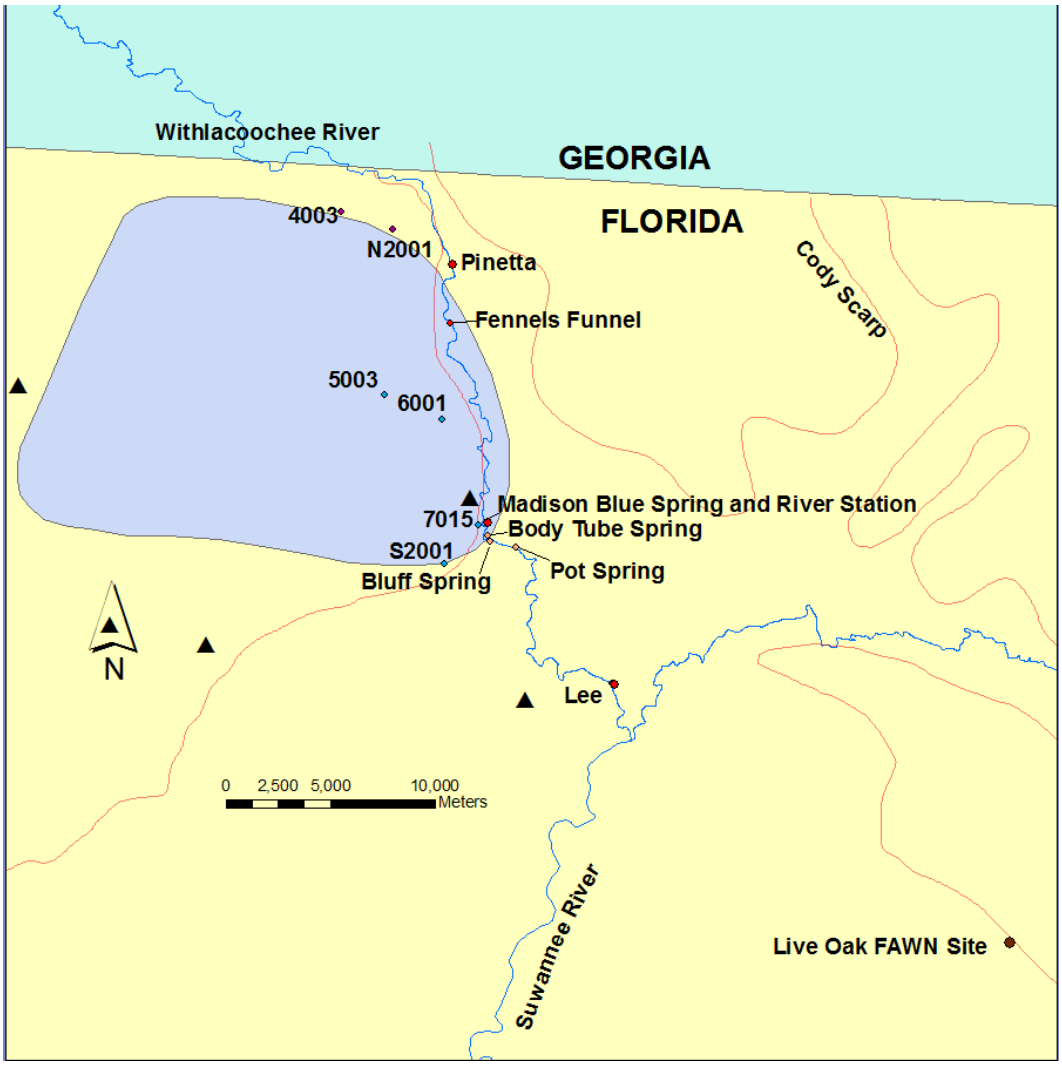

Figure 2-1. Location map of Madison Blue springshed. Madison Blue springshed outlined shaded in blue. Well locations are indicated by diamonds. Red circles show the location of gaging stations on the Withlacoochee River. Triangles indicate borehole log locations used for estimation of formation thickness. Abbreviations for wells are as follows: Well N011022001 (N2001), N011002001(S2001) and N031034003 (4003) are outside the designated springshed boundaries. Wells N011117015 (7015), N021036001 (6001) and N021035003 (5003) within the springshed were used for model calibration. Dye from the dye tracing study was observed at Body Tube Spring, Bluff Spring, and Pot Spring. Dye was added by cave divers inside the Madison Blue cave system (Farrell and Upchurch, 2004).

Series	Hydrostratigraphic Unit	Lithostratigraphic unit
Holocene	Surficial aquifer system	Undifferentiated Pleistocene-Holocene Sediments
Pleistocene		
Pliocene		
Miocene	Confining unit (Lenses of the Intermediate aquifer system)	Miccosukee Formation
		Cypresshead Formation
		Nashua Formation
		Hawthorn Group
Oligocene	Upper Floridan aquifer	Suwannee Limestone
Eocene		Ocala Limestone
		Avon Park Formation
		Oldsmar Formation
		Cedar Keys Formation

Figure 2-2. Hydrostratigraphic framework of the springshed.

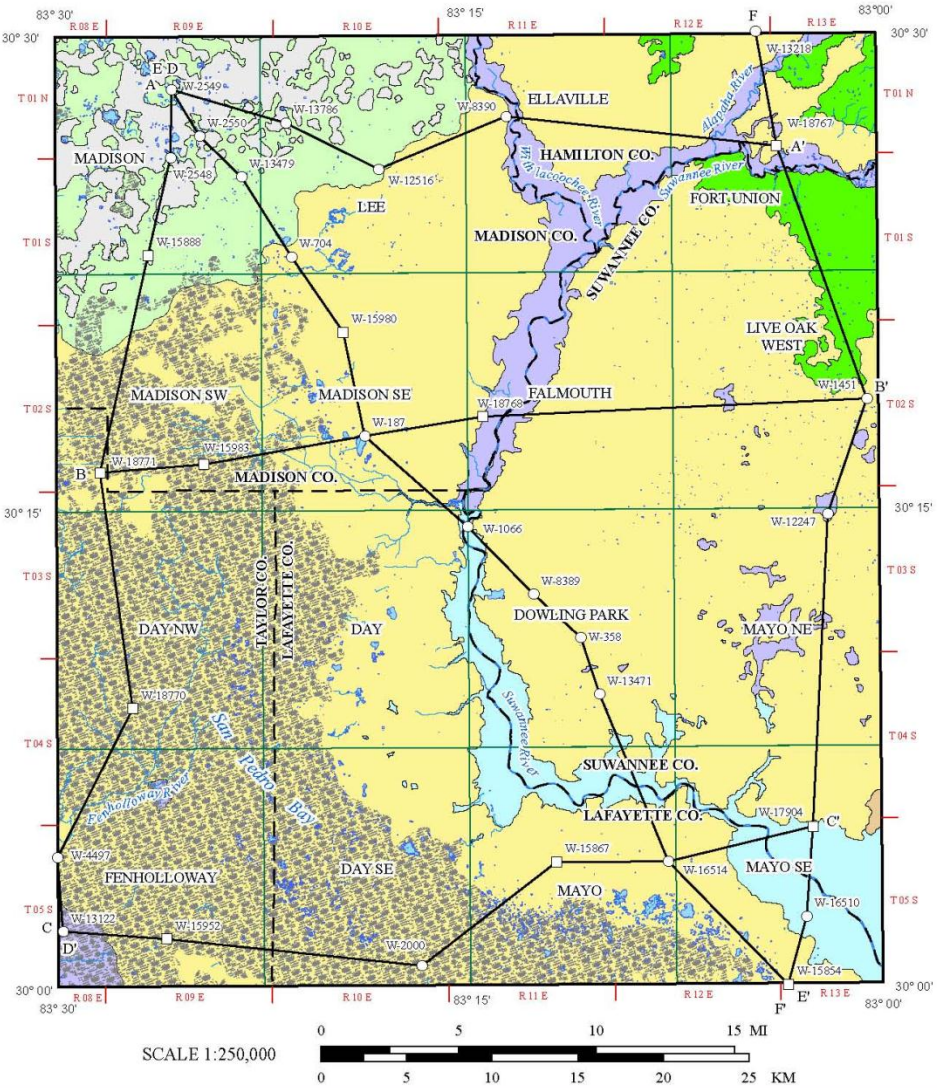

Figure 2-3. Location of cross section. Cross section (A – A') represents the Madison Blue springshed. (Used with permission from Florida Department of Environmental Protection http://www.fdep.fl.us)

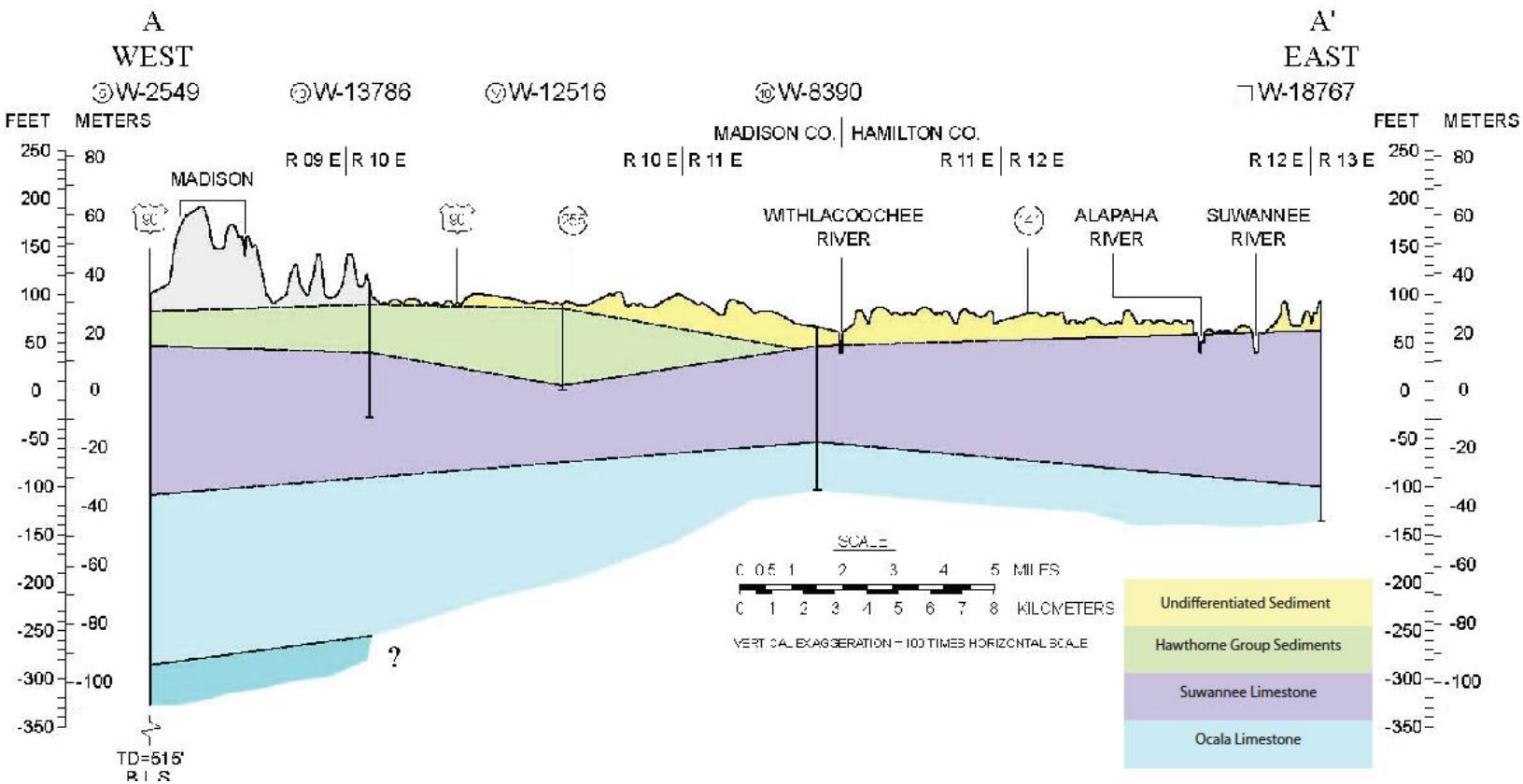

Figure 2-4. Cross section of Madison Blue springshed. The cross section trends west across Madison County in the Madison Blue springshed.(Used with permission from Florida Department of Environmental Protection http://www.fdep.fl.us)

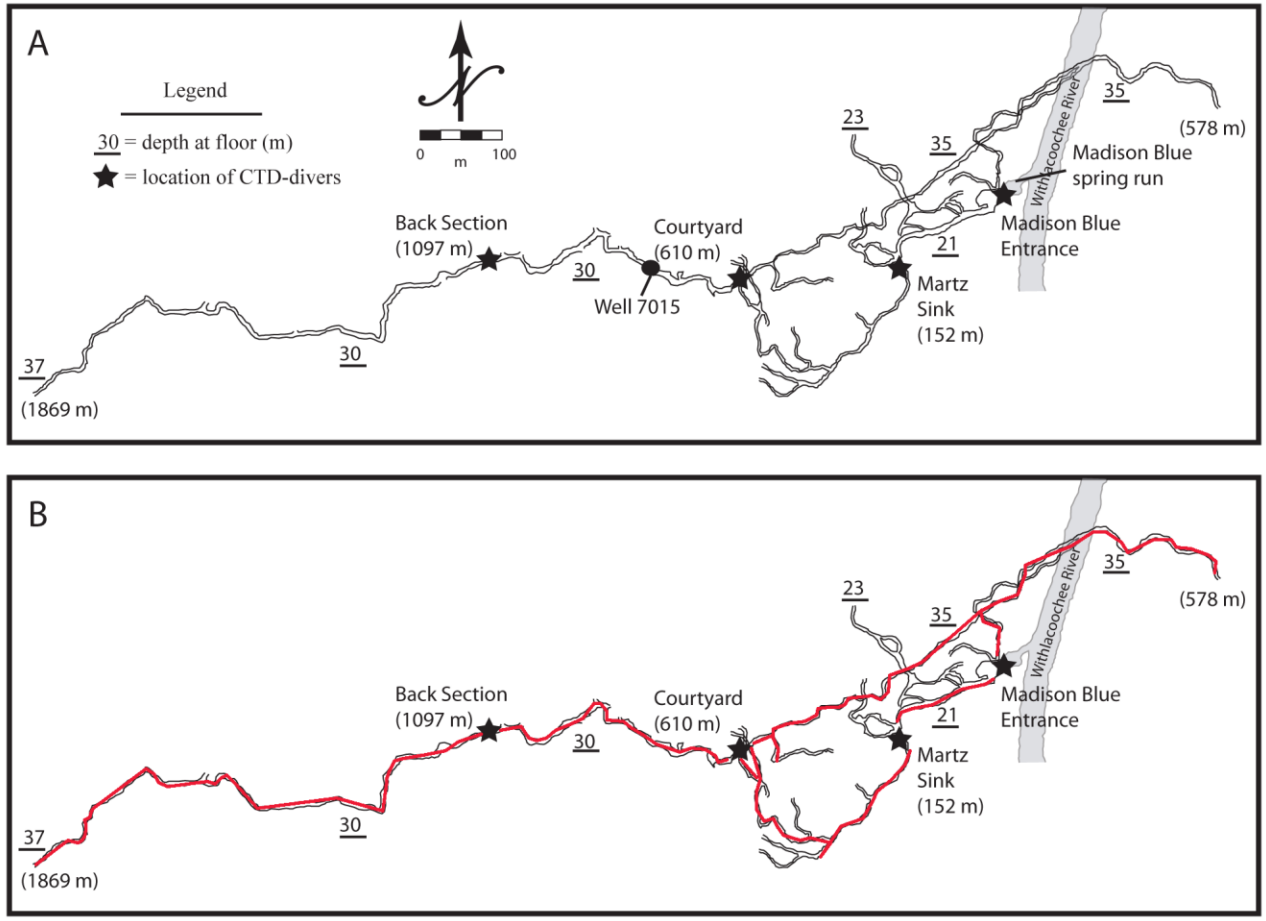

Figure 2-5. Madison Blue Cave map and overlay. (A) Madison Blue cave system current map. Breaks in the system indicate unexplored passageways. (B) Simulated cave system overlaying mapped system. The simulated conduit network was simplified to decrease computational time and maintain stability of CFP.

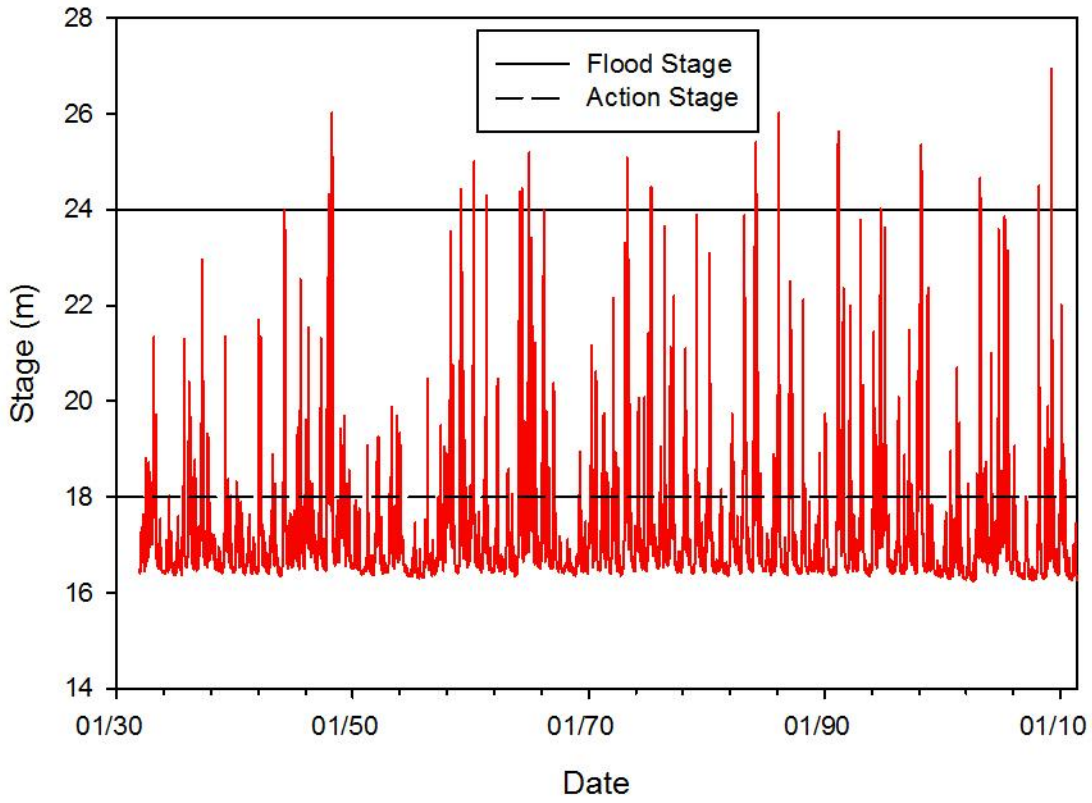

Figure 2-6. Stage at Pinetta station on the Withlacoochee River. The record spans from 1/1/1939- 5/30/2011. Solid line indicates flood stage (24 m) where major floods occurred. Dotted line indicates action stage (18 m) where water is considered to be much greater than average.

Figure 2-7. Map of confined and semiconfined portion of the Floridan aquifer. Reproduced with permission from Grubbs et al. (2003). The location of Madison Blue springshed is shown in red and is under unconfined and semi-confined conditions.

CHAPTER 3
METHODS

Data Collection

Rainfall and Evapotranspiration Data

Rainfall data was obtained by the Suwannee River Water Management District (SRWMD) at the site located in Madison Blue Spring state park (Figure 2-1). Evapotranspiration (ET) data from the Live Oak station (Figure 2-1) was obtained from the Florida Automated Weather Network (FAWN) website (http://fawn.ifas.ufl.edu/). Potential evapotranspiration provided by the FAWN sites was calculated using the Penman-Montieth method:

$$PET_0 = \frac{\Delta(R_n - G) + \frac{86.400\rho_a C_p(c_s^0 - c_a)}{r_a v}}{\Delta + \gamma(1 + \frac{r_s}{r_a v})} \quad (3\text{-}1)$$

Where

PET = Potential Evapotranspiration (m/day)

ρ_a = Density of air (kg/m^3)

Δ = Rate of change of saturation specific humidity with air temperature (Pa K^{-1})

γ = Psychometric constant (66 Pa K^{-1})

C_P = Specific heat of dry air (1.101 x 10^{-3} MJ/kg*C)

e_0= Mean saturated vapor pressure (kPa)

e_a = Mean daily ambient vapor pressure (kPa)

r_{av} = Bulk surface aerodynamic resistance for water vapor (s/m)

r_s = Canopy surface resistance (s/m)

Legacy Discharge and Stage Data

Withlacoochee River stage and discharge data at Pinetta, Madison, and Lee stations and stage and discharge data at Madison Blue spring were collected by the USGS and obtained through their Water Watch Network (USGS http://www.waterwatch.org). Stage and discharge was recorded at 30 minute intervals at the Madison and Lee USGS stations, and at hourly intervals at the Pinetta station (Figure 2-1). The Pinetta station is located 10 kilometers upstream from Madison Blue Spring and has been continually monitored since 1934. The Madison station is located on the Withlacoochee River~ 10 meters upstream from the outlet of the Madison Blue spring run and has been monitored since 1946. The Lee station is located 15 kilometers downstream and has been monitored since 2000. Discharge at all river stations is computed by applying a stream flow rating for the stream to records of water level elevation (Verdi et al. 2006). Madison Blue spring has a gaging station located in its spring run where discharge, velocity, and stage are recorded. Discharge at the spring is measured by using a current meter to measure velocity and relating velocity measurement to discharge. According to the USGS, discharge data from the Madison Blue Spring run are considered unreliable during reversals .

The gage at the Withlacoochee River at the Madison station was not always operational during the 2009 flood. Comparison with the downstream station at Lee indicates that the gaps occurred on relatively linear portions of a rising and falling limb (Figure 3-1). Thus, linear interpolation was used to fill data gaps of the Withlacoochee-Madison river stage. The largest gap in river stage was 16 days (4/17/2009-5/1/2009) on the falling limb of the recession curve.

Stage and discharge data at Madison Blue Spring were not available during 4/5/2009-5/8/2009. An average difference of 0.03 meters between the stage at Madison Blue spring and Madison station from the existing data set was used to approximate the stage at Madison Blue when the stage was inoperable (Figure 3-1).

Numerical Modeling of Groundwater Flow

MODFLOW is a three dimensional finite difference groundwater flow modeling code created by the USGS (Macdonald and Harbaugh 1983) to simulate groundwater flow. The equation for three dimensional groundwater flow in an unconfined aquifer under transient conditions is:

$$\frac{\partial}{\partial x}\left(hK_x\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(hK_y\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(hK_z\frac{\partial h}{\partial z}\right)=S_y\frac{\partial h}{\partial t}-R \qquad (3\text{-}2)$$

Where

$K_{x,y,z}$ = Hydraulic conductivity along the x, y, and z coordinate axes (m/d)

$\partial h/\partial(x,y,z)$ = Hydraulic gradient along the x, y, z coordinates (Dimensionless)

R = Recharge (source or sink of water) (m/d)

h = Piezometric head (m)

S_y = Specific yield(dimensionless)

t = Time (d)

Equation 3-2 assumes fully saturated flow and constant fluid density and viscosity and is valid only in porous medium where flow obeys Darcy's law:

$$Q = KA\frac{\partial h}{\partial l} \qquad (3\text{-}3)$$

Where

Q = Discharge through the porous medium (m^3/d)

K = Hydraulic conductivity (m/d)

$\frac{\partial h}{\partial l}$ = Hydraulic gradient (Dimensionless)

For flow within conduits, CFP checks whether flow is laminar or turbulent by calculating the dimensionless Reynolds number, which is the ratio of inertial to viscous forces (Equation 3-4). A lower Reynolds number of 2000 was used as the boundary for at which turbulent flow becomes laminar flow and an upper Reynolds number of 5000 was used for the transition from laminar to turbulent flow. Due to the conservation of momentum, turbulent flow tends to remain turbulent and laminar flow tends to remain laminar. Therefore, the larger the transition zone between the upper and lower critical Reynolds numbers, the more of a change is required for the flow to change from turbulent to laminar.

$$Re = \frac{\rho v L}{\mu} \tag{3-4}$$

Where

ρ = Density of water (kg/m^3)

L = Hydraulic diameter of pipe (m)

μ = Dynamic viscosity (kg/(m*d))

v = Velocity (m/d)

If flow is laminar, it is simulated using Darcy's law. If flow is turbulent, CFP calculates hydraulic head within conduits using the Darcy Weisbach equation.

$$\Delta h = f \frac{\Delta l v^2}{d 2 g} \tag{3-5}$$

Where:

f = Friction factor (dimensionless)

Δl = Length of pipe segment (m)

v = Velocity (m/d)

d= Hydraulic diameter of the pipe (m)

g = Acceleration due to gravity (m/d^2)

The required parameters for the pipe network consist of pipe diameter, pipe conductance, and wall roughness height. The wall roughness height is used to solve for the friction factor using the Colbrook-White Equation below:

$$\frac{1}{\sqrt{f}} = -2\log\left(\frac{k_c}{3.71d} + \frac{2.51}{Re\sqrt{f}}\right) \tag{3-6}$$

Where:

f = Friction factor (dimensionless)

Re = Reynolds number (dimensionless)

k_c = Roughness height (m)

d= Hydraulic diameter of the pipe (m)

Based on images of the cave system, roughness height was estimated to be about 0.1 meters. Because the conduit walls are permeable, CFP tracks exchange and uses the updated conduit discharge rate to calculate velocity.

MODPATH

MODPATH is a particle tracking program that computes flowpaths using the hydraulic head distribution from the solution of a steady state or transient MODFLOW simulation (Pollock 1994). It determines advective transport, and dispersion is not considered. It uses the effective porosity of the aquifer and the cell by cell flow

calculated by MODFLOW to determine travel velocity and thus distance traveled over a time step:

$$v = \frac{q}{n_e} \tag{3-7}$$

Where

v = Velocity (m/d)

n_e = Effective porosity (dimensionless)

q = Flow between cells (m/d)

Effective porosity, which is the pore space that is interconnected, was assigned as 0.25 based on values from the Ocala limestone (Budd and Vacher 2004).

Grid design and domain

The model domain was based on the demarcation of the Madison Blue Springshed from the Florida Department of Environmental Protection GIS shapefile (Florida Department of Environmental Protection; http://www.fdep.org). The model domain covered an area of 3.2 x 10^8 square meters. The model cell dimensions varied from 10 x 10 meters around the Madison Blue Spring area to 550 x 1000 meters at the farthest boundaries (Figure 3-2). The model was one layer to simulate the interaction between the Floridan Aquifer, the conduit system, and the Withlacoochee River. Based on Grubbs (1998) and FDEP report, the upper Floridan aquifer was assumed to be unconfined. A thickness of 100 meters was used based on examination of boring logs in the springshed that suggested the thickness of the Ocala Limestone to be 20 to 60 meters and Suwannee Limestone to be 10 to 40 meters.

The simulated conduit system (Figure 2-4) was based on the map of the Madison Blue Cave system. An average conduit diameter of 3 meters (Gulley et al. 2010) was used throughout the entire network. However observations from cave divers indicate that the diameter can vary a couple meters over a short distance. Narrow constrictions exist at the edges of the mapped cave system, and offshoot passages are known to exist but have not been mapped due to constrictions. Sensitivity to an extended conduit system was tested by simulating a conduit system with twice the overall length.

Boundary conditions

No flow cells were assigned outside of the springshed (Figure 3-2). The Withlacoochee River was simulated using MODFLOW's River Package. The River Package is used to simulate flow between the aquifer and overlying source reservoir (Anderson and Woessner 1992). Exchange between the river and the aquifer is simulated to be a function of the hydraulic head in the river and aquifer, as well as the conductance of the river bed.

$$Q = C_{riv}\left(h_{riv} - h_{aq}\right) \tag{3-8}$$

Where

Q = Exchange (m^3/day)

h_{riv} = Hydraulic head of river (m)

h_{aq} = Hydraulic head of porous medium (m) in the cell beneath the river

River conductance is computed by the following equation:

$$C_{riv} = \frac{KLW}{M} \tag{3-9}$$

Where

C_{riv} = River conductance (m^2/d)

K = Vertical hydraulic conductivity of streambed sediments (m/d)

W = Width of channel (m)

L = Length of the river within the model cell (m)

M = Thickness of streambed sediment (m)

MODFLOW's RIVER package alters the calculation when the simulated aquifer water level falls below the assigned river bottom. This did not occur at any point in the simulation.

Average stream width was assigned to be 50 meters based on Google Earth distance measurements. To assess sediment thickness, a sediment probe consisting of a metal rod with measured increments was inserted at several-meter intervals across the river channel adjacent to the three river gaging station locations. The sediment was very mobile upon insertion of the sediment probe. The sediment thickness varied between 0 and 1 m with the majority of probed areas having less than 0.1 meters of sediment. As a result, a thickness of 0.1 m was assigned for the modeling. Samples taken of the sediment were obtained and it was visually determined to be mostly silt-sized grains.

River stage data from the Pinetta, Madison and Lee Stations on the Withlacoochee River were used for the model, and river stage was linearly interpolated between each station. Madison Blue Spring was simulated as a specified head node. To allow exchange from river to constant head, and mimic a spring run, the cell between these the specified head node representing the spring and the conduit network was assigned a high hydraulic conductivity of 1.0×10^6 m/day.

Conduit pipe conductance

CFP calculates the exchange between the conduit and matrix:

$$Q_{ex} = C_p(h_n - h_m) \tag{3-10}$$

Where

Q_{ex}= Volume of exchange (m^3/d)

C_p= Pipe conductance (m^2/d)

h_n = Hydraulic head at the conduit node (m)

h_m = Hydraulic head at the node representing the aquifer matrix (m)

The pipe conductance value can be calculated from (Shoemaker et al. 2008):

$$C_p = \alpha K_m A_p \tag{3-11}$$

Where

K_m = Hydraulic conductivity of the porous medium (m/d)

A_p = Surface area of the pipe (m^2)

α = Exchange parameter (1/m)

The alpha exchange parameter is adjusted during calibration. A minimum value of 1 was initially used and the value adjusted during the steady-state calibration. A value of 1 suggests that exchange is simply a function of the hydraulic conductivity of the porous medium and the surface area. A larger exchange value represents a greater surface area for exchange than calculated for a cylindrical pipe. A lower exchange value would represent reduction of exchange due to sediment layers or wall coatings.

Steady State Modeling

A steady state model was run in order to estimate conduit and aquifer parameters and to simulate starting hydraulic heads for a subsequent transient simulation of the 2007-2011 time period. Conditions during September 2007 were chosen for the steady state simulation because it had stable spring and river stage and the best record of well and discharge data.

Recharge and River Mass Balance

Steady state recharge was estimated based on a water budget, assuming that over the long-term, outflow at Madison Blue Springs reflects the sum of diffuse recharge plus/minus net losses/gains of the Withlacoochee River.

$$Q_{sum} = Ax + y \tag{3-12}$$

Where

Q_{sum} = Average daily discharge at Madison Blue and estimated river gains and losses (m^3/d)

A = Area of the springshed (m^2)

x = Recharge rate (m/d)

y = Net river loss and gain (m^3/d)

Q_{sum} is the sum of the average discharge at Madison Blue Spring plus the net river loss from Pinetta to Madison station and proportional river gain from Madison to Lee. Two unmonitored springs discharge into the river from Madison to Lee. Because the springs are unmonitored, there is no way to determine exactly how much water is

contributed to the Withlacoochee from these springs. Therefore, a percentage of the total gain based on the percentage of river modeled in the simulation was used to estimate how much spring discharge would contribute to river gain in the model. The amount estimated to enter into the reach at the end of the model was determined by multiplying the total gain by the percentage of the distance represented in the model.

$$G_{rs} = G_r\left(\frac{L_m}{L_w}\right) \tag{3-13}$$

Where:

G_{rs} = The gain at the lower river reach in the model (m^3/d)

G_r = The total gain from Madison to Lee (m^3/d)

L_m = Length of river reach in model (m)

L_w = Length of the river from Madison to Lee (m)

Recharge was assumed to be uniform. Using this method, the recharge rate for the steady state model was 9.50×10^{-5} m/day. The initial value of hydraulic conductivity was 1400 m/day based on a previous simulation (Planert 2007), and this parameter was adjusted during model calibration. The average stage at Madison Blue Spring was 12.44 m, during September 2007.

Transient Simulation

The hydraulic heads resulting from the September 2007 steady state model were used as initial conditions for transient simulation of groundwater flow in the springshed from October 2007 to May 2011, a time period that includes two events, one in 2009 and one in 2010, that were examined in detail. MODFLOW discretizes time through stress periods. During each stress period, changes to external stressors such as

recharge and hydraulic head at a boundary were specified. Each stress period represented one-day during the March 28th 2009-May 30th 2009 and January 1st 2010-April 16th 2010 events. One month stress periods were used during the remainder of the transient simulation.

Storage and transient recharge

Specific yield, which is defined as the fraction of water that will drain from an unconfined aquifer due to gravity, ranges between 0.05-0.45 in the UFA (Palmer 1991). To determine the amount of recharge that would replenish the aquifer during the simulated event, evapotranspiration and rainfall values obtained from FAWN sites were obtained and recharge was assumed to occur if the soil moisture content was exceeded for that stress period (Grubbs and Crandall 2006). Soil thickness was assumed to be 0.10 meters (Ritorto et al. 2009). The ending moisture content from each day was used as the starting moisture content for the next day. The start of the analysis for soil moisture content was set as January 2007 and initial moisture content was assumed to be 0.05 due to moderately wet conditions prior to this date.

Transient stream conductance

During the transient simulations, stream conductance was adjusted to account for the Withlacoochee River coming into contact with more limestone of the river banks as the water rises. This was done by assigning a multiplier for every meter rise in stage. This multiplier was determined by assuming that the river was only in contact with sediment for the steady-state simulation. For every meter of stage rise, a new conductance was calculated by using a weighted average of the calibrated porous media hydraulic conductivity and the river sediment hydraulic conductivity. This new hydraulic conductivity was used to solve for the new conductance.

Water Chemistry

Surface and ground water chemistry data, which was needed for estimates of saturation indices, were obtained from the Suwannee River Water Management District (Suwannee River Water Management District http://www.srwmd.state.fl.us). Additional surface water samples from the Withlacoochee River near Pinetta were collected for chemical analysis during low flow conditions in October 2010 and January 2011. The samples were analyzed for cations, anions, dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), and alkalinity. To extract the water from the river, a peristaltic pump was used and polyvinyl chloride (PVC) tubing was extended into the river about 1 meter below the surface. A YSI model 566 multi-parameter field meter was used to ensure that specific conductivity, temperature, dissolved oxygen, pH and ORP had stabilized before sampling. Each collection vial was rinsed three times with river water before filling and capping. Samples for DIC were unfiltered and collected in 15 mL square glass bottles and preserved with mercuric chloride to kill any organisms that may be present. The rest of the samples were filtered. Samples for cations were collected in acid washed 20 mL plastic scintillation bottles and preserved with 0.1 M nitric acid to prevent metal precipitation. Water collected for DOC was preserved with 0.1 M hydrochloric acid and stored in 30 mL amber vials to inhibit light to inhibit the breakdown of organic molecules. Alkalinity was collected in 60 mL bottles and put on ice where it was transported back to the lab and titrated within 24 hours of sampling. Alkalinity was titrated using the Gran method. Total dissolved inorganic carbon (DIC) was measured using a UIC (Coulometrics) 5011 CO2 coulometer coupled with an AutoMate automated carbonate preparation device (AutoMateFX.com). Approximately 5 mL of sample is weighed into septum top tubes and placed into the AutoMate carousel. A double needle

assembly was used to purge the sample vial of atmospheric gas using CO_2 free nitrogen carrier gas. Acid was then injected into the sample vial and evolved CO_2 was carried through a silver nitrate scrubber to the coulometer where total C is measured. Major element concentrations (cations and anions) were analyzed on a Dionex model 500DX ion chromatograph at the Department of Geological Sciences, University of Florida.

Geochemical Modeling

The dissolution potential of the surface water was simulated using PHREEQC geochemical modeling software. PHREEQC is a geochemical modeling program that performs low temperature aqueous geochemical calculations. The version of PHREEQC used for this study was PHREEQC 2.17.5.4791. (Parkhurst and Appelo 1999)

Water chemistry data was input into PHREEQC and reacted with 10 moles of calcite to determine the amount of calcite needed to dissolve in order to bring the solution to equilibrium.

To determine the rate at which limestone would dissolve in the aquifer, PHREEQC Kinetic Rate Modeling was used. PHREEQC uses the standard Plummer, Wigley and Parkhurst (1978) (PWP) rate law which is given below:

$$r = k_1[H^+] + k_2[H_2CO_3] + k_3[H_2O] - k_4[Ca^{2+}][HCO_3^-] \quad (3\text{-}14)$$

where $k_{1\text{-}4}$, represent the rate constants for each of the reactions represented. The total rate, r, is multiplied by the area to volume ratio (A/V) to account for how much of the water comes into contact with limestone. The first term is the reaction of the hydrogen proton on calcite dissolution. This reaction is:

$$CaCO_3 + H^+ \rightarrow Ca^{2+} + HCO_3^- \quad (3\text{-}15)$$

This term is negligible for pH > 6, and is the dominating reaction in waters with a pH < 3.5, which is highly acidic. In river waters in Florida, pH is generally above 6.

The second reaction is the function of CO_2, or carbonic acid on carbonate dissolution and is represented by the following equation:

$$CaCO_3 + H_2CO_3^* \rightarrow Ca^{2+} + 2HCO_3^- \quad (3\text{-}16)$$

The third reaction is hydrolysis of calcite and is represented as

$$CaCO_3 + H_2O \rightarrow Ca^{2+} + HCO^{3-} + OH^- \quad (3\text{-}17)$$

The fourth and final reaction is the backwards precipitation reaction which is

$$Ca^{2+} + HCO^{3-} \rightarrow CaCO_3 + H^+ \quad (3\text{-}18)$$

Since water is flowing into a cave system, it is assumed that the system will be closed with respect to carbon dioxide (CO_2), meaning that CO_2 will not be replenished. This is accounted for in the modeling by initially equilibrating the water sample with atmospheric CO_2 in PHREEQC, and then neglecting a source of CO_2 so that it will be used up in carbonate dissolution.

Inputs for the simulation in PHREEQC require the surface area to volume ratio (A/V) and the exponent for surface rate of change for the kinetic rate calculation. The surface area to volume ratio was assumed to be 50.0 m^{-1} for matrix reaction rates and was determined based on an upper representative A/V ratio for small fissures (Eisenlohr et al. 2007) which would be similar to conditions encountered in the matrix. A value of 1.3 m^{-1} was used as the A/V ratio for the conduit and this was determined by assuming a cylindrical conduit with a 1.5 m radius and obtaining the volume of the cylinder input into the model, and the surface area of that same cylinder and taking the ratio of those values.

The exponent for surface rate of change or $(m/m_0)^n$ where n is the exponent of surface rate of change, is a standard value that defines the rates of change of surface area with dissolution. A value of 0.67 value was obtained from Parkhurst and Appelo (2005).

The calcite dissolution rates assume no other chemical inhibition or facilitator of dissolution which can be available in these systems such as trace metals (Lea et al. 2001) microbial activity (Brown et al. 2011) or coatings that could inhibit water to mineral contact. Water in calcite equilibrium was determined by looking at the output data and determined to be the time when the saturation index was at 0. No adjustments for change in temperature during reaction were made.

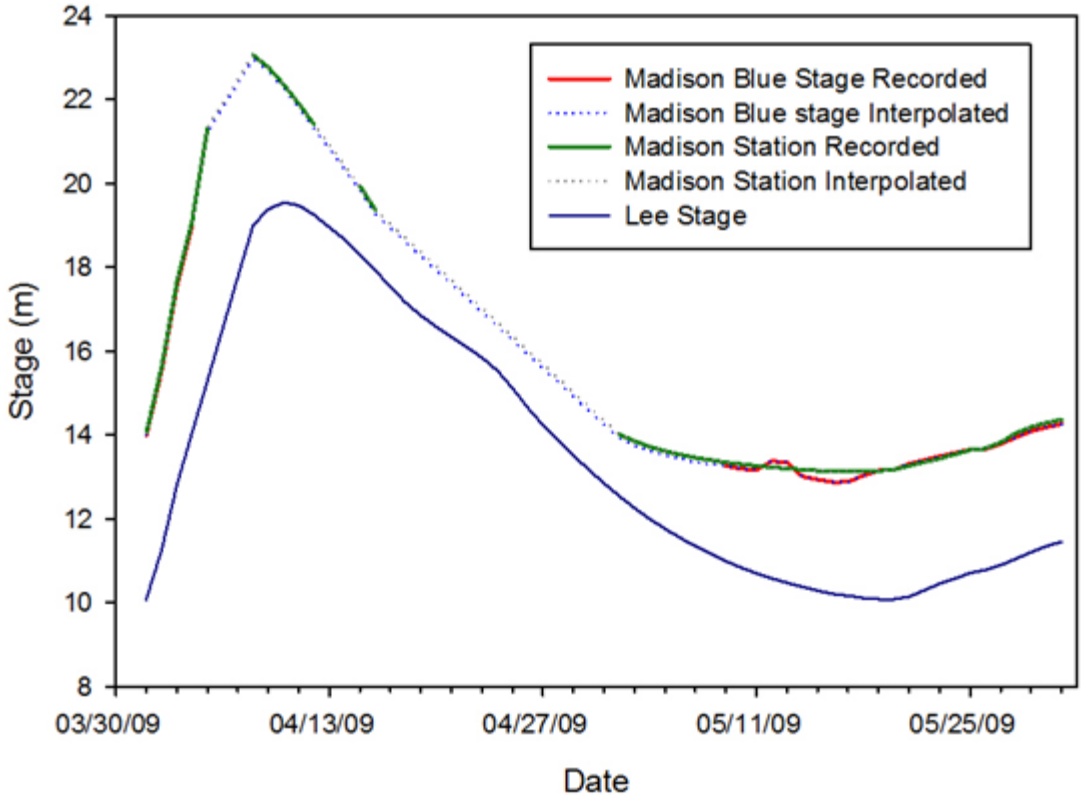

Figure 3-1. Interpolated stage at Madison Blue during 2009 flood.

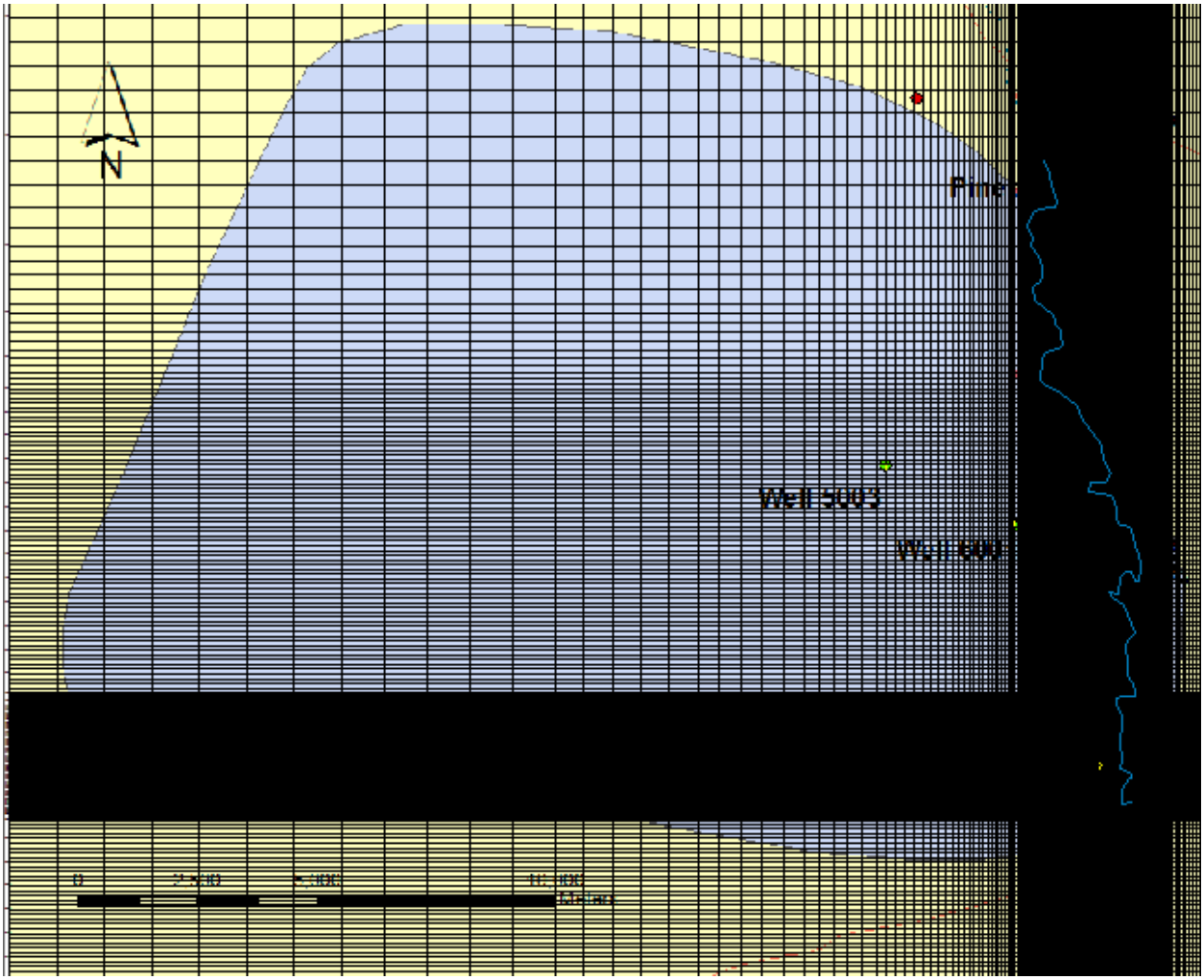

Figure 3-2. Finite-difference grid used for the model. The grid is refined to 10x10 m around the spring, and maximum grid spacing was 550 x 1000 m at the far edges of the model domain.Model cells were inactive outside the springshed in the region shaded in yellow.

CHAPTER 4
RESULTS

The study focuses on a recession period in 2007, a major flood in 2009, and two smaller events in 2010. Total rainfall during 2007 was 1.0 meter which was less than the average of 1.14 meters for the data available at the station (2003-2011) Rainfall was 1.29, 1.30, and 1.30 for 2008, 2009 and 2010 respectively (Figure 4-1). The Withlacoochee River stage at Pinetta reached historic lows during 2007 (Figure 4-2). During 2007, the average loss of the Withlachoochee River between Pinetta and Madison was 2.67 x 10^5 m^3/day (Figure 4-3) and the reach on the Withlacoochee River from Madison to Lee gained an average of 9.42 x 10^5 m^3/day (Figure 4-3).

High discharge events occurred in 2008, 2009 and 2010 (Figure 4-2) associated with significant rainfall (Figure 4-1). In 2008, Tropical Storm Fay brought 0.05 meters of rain to the Madison Blue station and a stage increase of 2.7 meters at Withlacoochee River at Pinetta. However no chemical or monitoring data were available for this time period, so this event was not studied in detail. Rainfall in the 9 days (3/26/2009-4/3/2009) prior to the 2009 flood totaled 0.3 meters, and total estimated recharge was 0.18 meters. The 2009 flood had the highest stage (26 m) on record for the Withlacoochee River at the Pinetta station. During this flood, specific conductivity meters within the conduit system documented intrusion of dilute river water (Gulley et al. 2011).

During the 2010 event, there was no specific conductivity or temperature monitoring within the cave system. Total rainfall during the January 2010 event was 0.10 meters, and estimated recharge was 0.08 meters. Two pulses of increased discharge were observed on the Withlacoochee River from the beginning of January

2010 to the end of February 2010 (Figure 4-4) with the first being larger at 5 meters. Although discharge data from Madison Blue spring run indicate outflow throughout 2010, visual observations by park rangers indicated flow of tannic river water into the spring (Personal Communication, Madison Blue Spring Park Ranger, 2010*).* In addition, samples collected from the spring during February 2010 indicate the presence of undersaturated water, supporting the reversal observations (Figure 4-5).

Comparison of water levels with the Madison Blue spring run, the Withlacoochee River at Madison Blue, and a monitoring well (7015) located near the conduit system provides additional information about hydraulic head gradient and thus about reversals during 2010. Figure 4-6 shows the hydraulic head difference between Madison Blue spring run and Well 7015. Between events, the hydraulic head differences are positive, consistent with water flowing from the aquifer to Madison Blue spring. However during events, the hydraulic head differences are negative indicating that river water will flow into the aquifer.

Intrusions during floods do not just occur at the Madison Blue spring, but also elsewhere along the river. During events more water is lost to the subsurface between Pinetta to Madison stations than during recessions (Figure 4-3). Losses increase as the flood pulse migrates downstream, and are followed by gains along the reach.

While events like the 2009 flood are rare, events similar in magnitude to the 2010 event are more common. Data from head differences between Madison Blue spring and well 7015 show that during the period when head data was available at the well, gradient reversals occurred every year for the last 8 years (Farrell and Upchurch 2004). For the full record at Pinetta from 1939, Pinetta was at or beyond Action Stage (the

stage defined as when the national weather service needs to take action to prepare for a flood) 48% of the time. Stage reached or exceeded the levels seen during the 2010 event 6% of the entire period of record.

Due to the high flood waters, some of the differences between discharge at the Pinetta and Madison stations during events may be due to increase or decrease of water stored in the channel or as water overflowing the banks. An estimate of the daily change in volume of surface storage was done by taking the difference in stage height multiplied by the average width (50 meters) and the distance from Pinetta to Madison (10 km). This change in surface storage was used to estimate gain from or loss to the aquifer from the measured discharge differences. An unknown additional amount of surface storage occurs during the 2009 event as the river overflows its banks. As a result, the estimated losses during the 2009 event are likely overestimates.

Numerical Simulation

Calibration of the steady-state simulation of September 2007 conditions was based on head observations at 3 monitoring wells, measured discharge at Madison Blue spring, and calculated losses of the Withlacoochee River between Pinetta and Madison stations. The calibration yielded a hydraulic conductivity of 2000 m/day, a river conductance of 2000 m^2/day, and a pipe conductance of 550,000 m^2/day. Hydraulic head contours and well residuals of the calibrated steady state model can be seen in Figure 4-7. Table 4-1 shows the target values for wells and simulated results as well as mass balance from the river losses.

Hydraulic conductivity of 2000 m/day provided the best match for Madison Blue spring discharge and hydraulic heads in the springshed for the steady-state calibration (Table 4-1). Increasing the hydraulic conductivity to 2500 m/day increased discharge at

the spring by 7.1, x 10^4 m^3 and decreasing it to 1500 m/day decreased discharge by 5.0 x 10^4 m^3.

The calculated losses of water between Pinetta and Madison during the calibration period required a river conductance of 2000 m^2/day. This river conductance implies a vertical hydraulic conductivity of river sediments of 0.4 m/day (Equation 3-9), which is within range for the hydraulic conductivity of clayey/silty sediment (3.0 x 10^{-2} m/day–3 m/day; Bear, 1973) such as found in the channel. Increasing the river conductance also increased discharge at Madison Blue spring, as water lost from the river contributes to simulated Madison Blue discharge; however this value did not impact the discharge as much as changes in the porous medium hydraulic conductivity. Discharge at the spring with a river conductance of 2500 m^2/day was 2.3 x 10^5 m^3/day (an increase of 1,867 m^3/day), whereas a decrease to 1500 m^2/day resulted in a decrease in discharge to 1.8 x 10^5 m^3/day. Changes in river conductance also affected hydraulic heads of the target wells, with Well 6001, the closest to the river, being the most sensitive to changes in river conductance. When conductance was changed to 2500 m^2/day, head residual at Well 6001 was 0.15 m, and at 1500 m^2/day, the residual was 0.19 m.

The steady-state simulation provided reasonable values for hydraulic heads at most monitoring wells, with a root-mean-square error (RMSE) of 0.02 meters. Two additional wells, N2001 and N4003 outside the springshed were compared to the simulation results. N4003 matched results reasonably, but at N2001, the simulated head was 2 meters less than estimated by extrapolating contours out beyond the boundary. Matching extrapolated head at this well required reducing hydraulic

conductivity to 200 m/day, which significantly increased the error in heads elsewhere in the springshed and decreased the amount of water lost to the aquifer from Pinetta to Madison by 2 orders of magnitude. It is possible that hydraulic conductivity differs in this region or that the UFA near Well N2001 may be hydraulically connected to the Withlacoochee River outside of the simulated region. Because the Withlacoochee River is not simulated north of Pinetta, contributions from the river to the aquifer are neglected and aquifer heads may be underestimated in this region.

Calibrated pipe conductance was 550,000 m^2/day, which implies an alpha parameter of 2.32 (Equation 3-11). This allows 2.32 times the exchange than would occur in a perfectly smooth cylindrical pipe. Reducing pipe conductance to 1.4×10^5 m^2/day (an alpha value of 1), increased head residuals near the spring by 0.22 meters and discharge was at the spring was ~1.0×10^5 m^3/day lower than the target value. Increasing the pipe conductance did not greatly affect discharge. An alpha value of 5.3 increased discharge at the spring by 5.0×10^3 m^3/day to 2.0×10^5 m^3/day.

Calibrated transient model

Specific yield, which is not used in the steady-state calibration, was calibrated during the transient simulation. Water levels at the wells were more sensitive to changes in the specific yield than was discharge at the spring (Figure 4-8, 4-9 and 4-10). Through calibration, a final value of 0.10 for specific yield was determined to be the best fit for the springshed (Figure 4-8 and 4-9, Table 4-1).

The simulated well heads from the calibrated transient model matched reasonably well to observed head over the specified time interval of simulation. Spring discharge also matched relatively well during low flow (Figure 4-10). However, during

events, the reported and simulated discharge at Madison Blue spring diverged considerably, as is expected given the known issues with the discharge measurements.

Simulated river losses

Simulated river losses on the Withlacoochee River between Pinetta to Madison stations were compared with observed values to check if the model was accurately simulating river losses. Simulated losses matched well during the 2009 event, and also toward the end of the first reversal in 2010 (Figure 4-11). There were large differences between simulated and observed values during the peak of both events, however these differences are likely the result of surface water storage in the channel. A comparison of volumes between estimated surface water storage and the difference in observed and simulated losses are on the same order of magnitude for both years (10^7 m^3) suggesting that this is the likely cause of this discrepancy.

Simulated River Water Intrusion

The volume of simulated river water intrusion was obtained by summing the volume of inflow at the constant head boundary representing Madison Blue spring. For the 2009 event, the calculated total of river inflow into the spring was 9.25 x 10^6 m^3.This result is significantly greater than previously calculated (2.8 x 10^5 m^3; Gulley et al. 2011). The simulated reversal began on 4/1/2009, and the reversal time in the simulation was 7-8 days. Both results are consistent with specific conductivity recorded in the spring. Modeled flow velocity within the conduit on April 1st 2009 was 0.03 m/s (Figure 4-12), in close agreement with values calculated from tracking the low specific conductivity signal from the spring entrance to Back Section (Gulley et al. 2011) which was also 0.03 m/s. However, instead of decreasing after April 1st, as was assumed by Gulley et al. (2011), simulated velocity increased by an order of magnitude until April 5th

(Figure 4-13). This difference between simulated velocities and the decrease assumed by Gulley et al. (2011) explains the significantly higher intrusion volume calculated here.

Though there were 2 reversals in 2010 with one having a longer duration of simulated intrusion than in the 2009 event, the total volume of water that intruded into the aquifer in 2010 was smaller (2.5×10^6 m^3) than estimated for 2009. The first simulated reversal occurred on January 18th (Figure 4-6) and the simulated reversal time was 13 days. Total river water intrusion over this time was 2.4×10^6 m^3. The pulse that generated this reversal was about 5 meters lower than the peak of the 2009 flood. The second reversal occurred on February 7th and had a simulated intrusion volume of 1.2×10^5 m^3 over 3 days. The timing of both simulated reversals is consistent with the head differences between the Madison Blue spring gage and Well 7015 (Figure 4-6).

Sensitivity to conduit extent

To test the impact of uncertainties in conduit extent, additional simulations were conducted with a conduit system twice as long as that mapped (Figure 4-15). The hydraulic conductivity was then re-calibrated with this conduit system geometry, and the best fit for the 2007 steady-state simulation was achieved with a hydraulic conductivity of 1500 m/day. The resulting Madison Blue spring discharge, losses between Withlacoochee River between Pinetta and Madison stations, and match to observed heads were similar to the base simulation (Table 4-1). To test the impacts on the estimates of volumes of inflow, the transient simulation was then run with the extended conduit and lowered hydraulic conductivity. The simulated exchange of water from surface to aquifer during the 2009 flood was 12% less with the extended conduit than the simulated exchange for the base model. In contrast, simulated exchange for the

extended conduit model was 30% greater for the 2010 event than that from the base model.

Sensitivity to stream conductance

In the base simulation, river conductance was changed with stage to reflect the increased contact with upper Floridan aquifer limestone as the river rises. Because the calibrated hydraulic conductivity of the aquifer was 2000 m/day, significantly larger than the estimated value of the stream bed sediments (0.4 m/day) this results in a stream conductance that increases with stage. To test the effect of this increase in stream conductance, the transient model was re-run with constant stream conductance. This resulted in longer reversal times in the 2009 event by 4.5 days, and river losses during the event were significantly lower than estimated losses when the gages were still working on the river stations between Pinetta and Madison on the Withlacoochee River. Simulated river water intrusion into Madison Blue spring during the 2009 event increased by 14% from 9.25×10^6 m^3 to 1.07×10^7 m^3.

In 2010, the constant stream conductance resulted in only 1 reversal instead of 2 during the simulation, with a decrease in the duration of the first event by 3 days, contrary to evidence from head differences between Madison Blue spring and Well 7015. The simulated volumes of river water entering in the aquifer at Madison Blue spring decreased from 2.52×10^6 m^3 to 2.31×10^6 m^3.

Simulated Exchange Between Conduits and Matrix

The volumes of inflow and outflow to the conduit system simulated by the base model provide a rough approximation of residence times. For the 2009 event, the inflow velocity was approximately twice the outflow velocity (Figure 4-13), and the time it took for an equivalent volume of river water that intruded during the 8 day reversal period to

discharge was 22 days (Figure 4-16). In 2010 the inflow and outflow velocities were similar (Figure 4-14) first reversal had a rate of inflow of 13 days, and outflow for the same amount of water was 6 days (Figure 4-17). The second reversal was much shorter with infiltration time of 3 days and outflow of the same amount required 2 days (Figure 4-18).

Relative to the volume estimates, particle tracing with MODPATH provides a more complex picture of inflow and outflow to the surrounding formation during the 2009 flood and 2010 events (Figure 4-19 and 4-20). During the 2009 event (Figure 4-19) particle tracking indicates outflow along the entire simulated conduit system, with penetration into the surrounding aquifer averaging 92 meters. The average depth of particles decreased along the length of the conduit except at the termination, where increases in formation penetration were seen. Particles placed at the termination of the conduit system continue in the general trajectory of the conduit and migrated up to 110 meters into the aquifer by the end of the simulation. Exchange between conduit and formation during the 2009 event was variable, with some nodes showing water entering the conduit network from the formation while the spring was still reversing (Figure 4-21).

Despite the much lower volume of simulated inflow, the observed travel distance seen in the first 610 meters of the cave system in 2010 is similar to the magnitudes seen in 2009, with an average penetration depth of 71 meters. However, during the 2010 event, there was almost no infiltration into the formation beyond the location of Courtyard. The overall exchange over different sections of the cave normalized to exchange per unit area can be seen in Figure 4-23. As also shown by the particle tracking, inflow to the aquifer matrix in 2009 was roughly uniform throughout the entire

conduit network, with a slight decrease toward the back of the conduit. In 2010, the average inflow to the matrix was greatest near the entrance, and much less toward the back of the mapped cave system. The exchange estimates also show lesser exchange between the entrance and Courtyard than between the entrance and Martz Sink.

For the 2009 event, of the particles that returned, the longest residence time was 119 days to return to the conduit and the shortest time was 3 days. Particles that never returned were near the front of the cave system and migrated toward the river, which affected the local hydraulic head distribution. In 2010, all particles that left the conduit were near the front of the cave and none returned to the conduit during the simulation. Many that left the conduit eventually discharged to the spring or river.

Carbonate Equilibrium

Calculations in PHREEQC provided a rate for dissolution due to river water intrusion in the conduit and during matrix flow. For 2009, a representative water sample from Pinetta taken on 4/15/2009 was used to represent the composition of intruding river water. Reaction rates obtained for matrix flow where the A/V was assumed to be 50 m^{-1} were 4.6 days to reach equilibrium. For reactions within the conduits, the assumed A/V was assumed to be 1.3 m^{-1} and the resulting reaction rate is slower. Calculations indicated that 90% saturation would occur in 24 days, and equilibrium would be reached in 98 days. For 2010, a water sample from 2/10/2010 was used to represent water chemistry that could have intruded into the aquifer for the second reversal on 2/7/2010. It took 5 days for water to reach saturation in the matrix. For conduit flow, it took 30 days to reach 90% saturation and 120 days for equilibrium.

Dissolution

To calculate calcium removed from the aquifer due to diffuse recharge, the calculated diffuse recharge volumes were multiplied by the difference in moles of calcium between average rainwater values obtained by the National Atmospheric Deposition Program (http://www.nadp.org) and average normal flow water collected by the University of Florida at Madison Blue Spring (1.042 mM). Since the dissociation of calcite results in 1 mole of calcium and 1 mole of carbonate, calcium can be used as a proxy for 1 mole of calcite. The difference in moles represents how many moles would have to dissolve to reach the value observed in the discharge from Madison Blue spring. The calculations were carried out assuming a molar volume of calcite of 3.694 x 10^{-5} m^3/mole, a density of calcite of 2710 kg/m^3 and a porosity of 0.25. Estimated dissolution from diffuse recharge totaled 9.5 x 10^3 m^3 during 2009 and 1.1 x 10^3 m^3 during 2010.

Dissolution due to Withlacoochee River losses between Pinetta and Madison

Calcium concentrations of the Withlacoochee River water were estimated based on a relationship between calcium and discharge on the Withlacoochee River at Pinetta Station developed from 103 USGS sample results obtained between 1992-2011 (Figure 4-24). The Pinetta station was used because it had the most extensive data set and changes in chemistry are generally negligible between Pinetta to Madison stations. Using this relationship, the amount of calcium in the river water was calculated during 2009 and 2010 for each day. This value was subtracted by the average concentration of calcium that was obtained at Madison Blue during normal flow periods (1.042 mMol) and the difference, which was assumed to represent calcium added by dissolution of

limestone, was multiplied by the volume of river water lost between Pinetta and Madison.

The dissolution due to losses between Pinetta and Madison stations was calculated separately for recession periods and for the large 2009 and 2010 events. For the recession periods of 2009 and 2010, the molar volume of calcite that would be dissolved was calculated to be 4.4 x 10^3 m^3 in 2009 and 4.0 x 10^2 m^3 in 2010. For the events, the amount of calcite that could have potentially been dissolved from river losses from Pinetta to Madison in 2009 were 4.5 x 10^3 m^3 and over the course of the total 2010 event, total potential dissolution was 1.26 x 10^3 m^3. These volumes were corrected for surface storage.

Dissolution due to river intrusion into Madison Blue spring

An upper bound on dissolution due to river intrusion into Madison Blue Spring was estimated by assuming that the intruding water (with estimated calcium values shown on 4-24), reacted to the normal flow Madison Blue Spring calcium concentration. This is an oversimplification because not all the water that enters the conduit system will react to reach normal flow values, as indicated both by the evidence of outflow of undersaturated water and by comparing the duration of simulated reversals and the estimated time to equilibrium. For the 2009 and 2010 events, the calcium concentration of river water flowing into the spring was estimated from the discharge (Figure 4-24) and was multiplied by the amount of water that the simulations calculated intruded into the aquifer. The total amount of calcite that was estimated to potentially be dissolved from river water intrusion into Madison Blue Spring from the flood in 2009 was 7.11 x 10^2 m^3 of calcite (Figure 4-25). For 2010, the amount of potential dissolution from the calcium mass balance was estimated for each of the two recorded intrusions. For the first

intrusion, total potential calcite dissolved was 3.55 x 10^2 m^3. For the second, smallest event, it was 5.28 x 10^1 m^1. The total calcite that could have been dissolved for each reversal in 2010 was 4.10 x 10^2 m^3.

CFP Limitations

Though CFP was valuable in interpreting conduit flow patterns, the simulations took significantly longer than with a porous media model in which the conduits were represented by high hydraulic conductivity (14 hours versus 1 hour for the porous media model). In addition, the exchange and flow direction was highly variable in the conduits, particularly during events. This could not be resolved by adjusting time step length and could reflect numerical issues with the large pipe conductance value used.

Table 4-1. Table with calibrated values for Steady State Model

Value	Target Value	Calibrated Value	Extended Conduit Value	Transient RMSE Sy = 0.05	Transient RMSE Sy = 0.1	Transient RMSE Sy = 0.4
Estimated Loss – Pinetta to Madison	199,022 m^3/day	221,523 m^3/day	220,012 m^3/day			
Constant Head Boundary (Madison Blue)	155,864 m^3/day	172,260 m^3/day	177,550 m^3/day			
Well 7015	12.62	12.62	12.66 m	0.15 m	0.14 m	0.18 m
Well 6001	13.92	14.04	14.08 m			
Well 5003	14.33	14.26	14.28 m	1.11 m	1.0 m	1.48 m
Well 2001N	17.52	14.69	14.67 m			

Table 4-2. Chemistry data from Pinetta to Madison during normal flow conditions

Date and Location	T (ºC)	SpC (µS/cm)	DO (mg/L)	pH	Na mM	K mM	Mg mM	Ca mM	F mM	Cl mM	SO4 mM
11/20/2010											
Pinetta	19.09	441	7.28	6.7	2.3412	0.0793	0.2750	1.0585	0.0045	0.2574	0.9749
Madison	18.27	423	8.34	6.68	2.1607	0.0732	0.2760	1.0929	0.0037	0.2467	0.9116
1/28/2010											
Pinetta	12.19	323	8.54	7.93	1.6194	0.0833	0.1543	0.4344	0.0049	0.2795	0.6731
Madison	13.58	362	9.93	7.48	1.6809	0.0764	0.1952	0.6323	0.0024	0.3092	0.7351

Table 4-3. Legacy Chemistry data from 2010 at Madison Blue spring.

Date	T	SpC	DO	pH	Na	K	Mg	Ca	F	Cl	SO4	Alkalinity
	ºC	µS/cm	mg/L		mM	mM	mM	mM	mM	mM	mM	mM
2/3/2010	11.98	82	7.56	6.57	0.206	0.070	0.051	0.176	0.0024	0.230	0.062	0.236
2/10/2010	11.12	68	9.01	6.83	0.192	0.063	0.044	0.114	0.0023	0.218	0.052	0.142
2/18/2010	11.2	123	5.4	6.76	0.196	0.050	0.080	0.353	0.0028	0.225	0.072	0.631
2/25/2010	12.63	142	-	6.58	0.234	0.052	0.058	0.281	0.0026	0.241	0.077	0.591
3/5/2010	13.13	120	-	7.55	0.271	0.051	0.065	0.309	0.0026	0.253	0.087	0.671

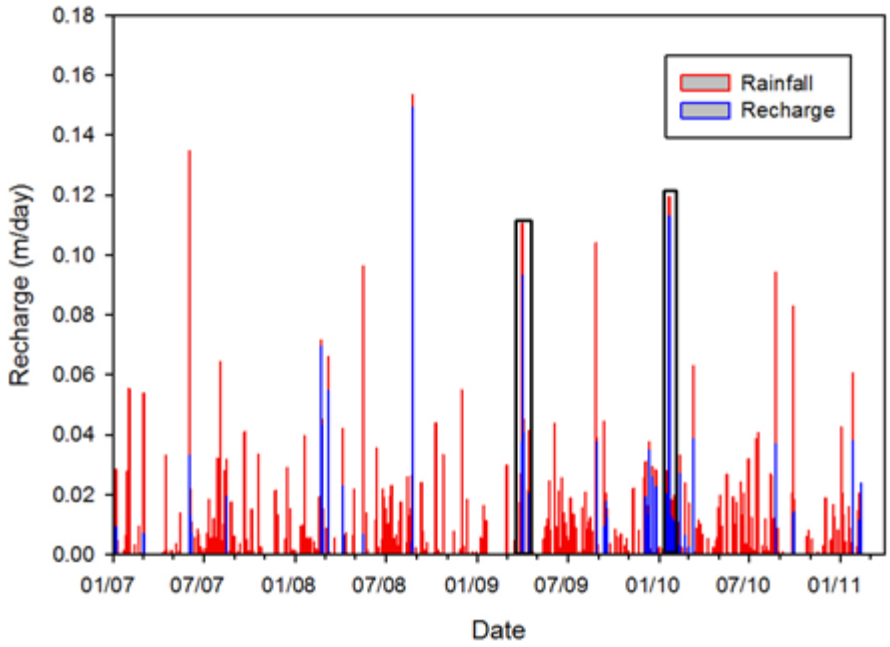

Figure 4-1. Recharge and rainfall in the Madison Blue springshed. Rainfall is in red and estimated recharge is indicated in blue. Black boxes indicate the 2009 flood and 2010 event.

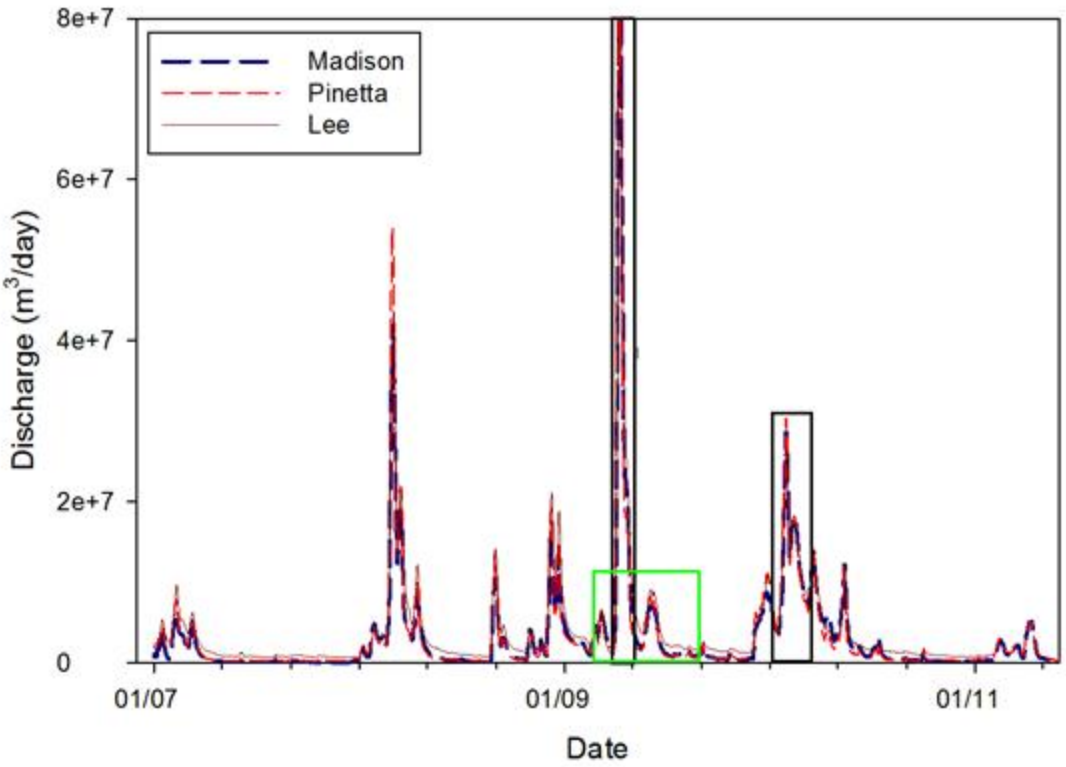

Figure 4-2. Discharge of the Withlacoochee River at the Pinetta, Madison, and Lee stations. The period of record spans from January 2007-May 2011. Black boxes indicate the 2009 flood and 2010 event. Green box indicates time period specific conductivity loggers were in the Madison Blue Cave system.

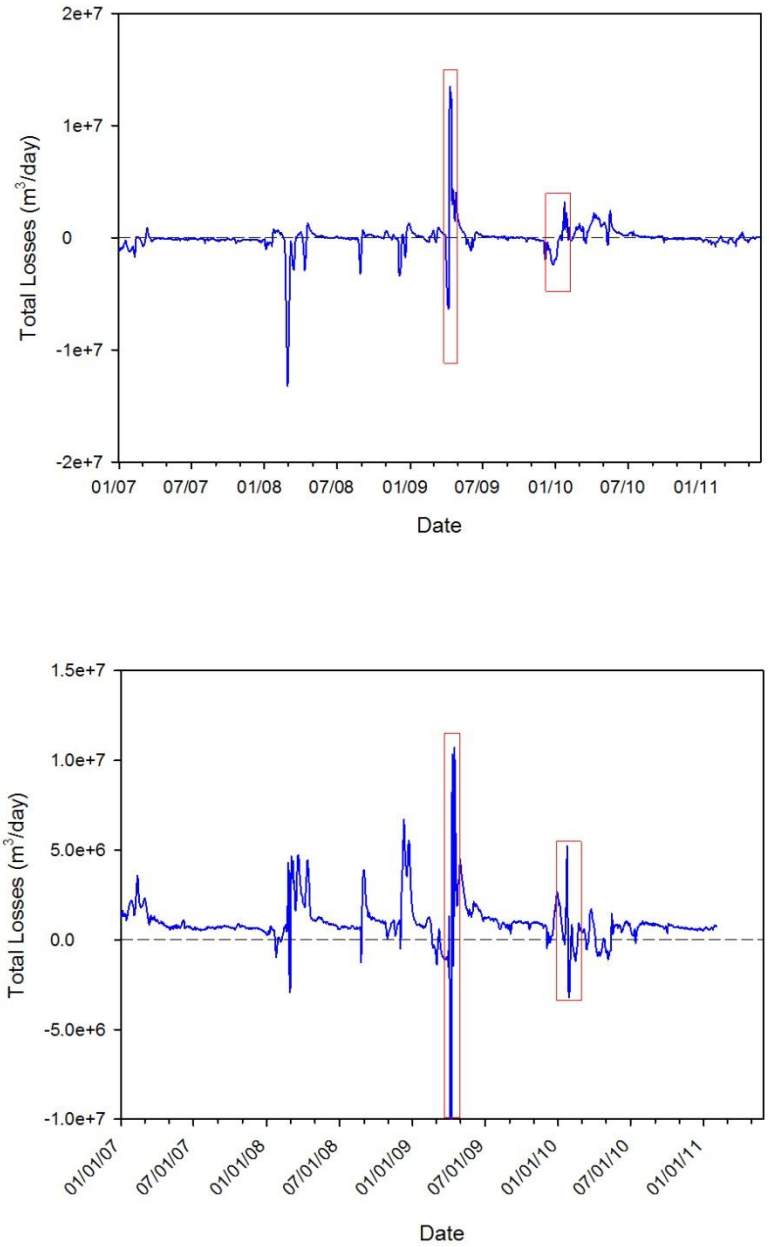

Figure 4-3. River losses and gains on the Withlacoochee River between the USGS gaging stations. (TOP) Withlacoochee River losses and gains between Pinetta and Madison stations. (Bottom) Withlacoochee River losses between Madison and Lee stations. Positive values indicate the river is gaining water and negative values indicate that the river is losing water. Red boxes indicate the 2009 flood and 2010 event.

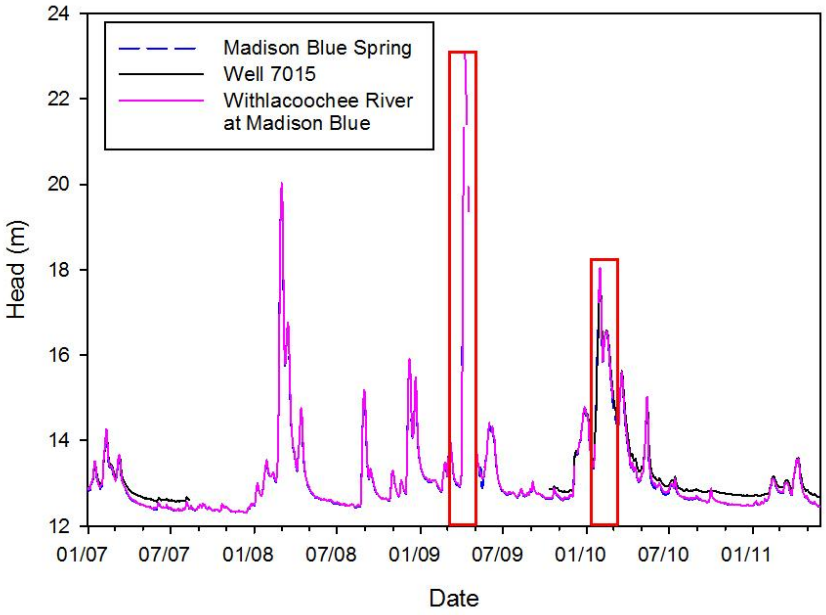

Figure 4-4. Stage at Madison Blue spring run, Withlacoochee River at Madison Blue, and Well 7015. Period represented is from January 2007-May 2011. Red boxes indicate the 2009 flood and 2010 event.

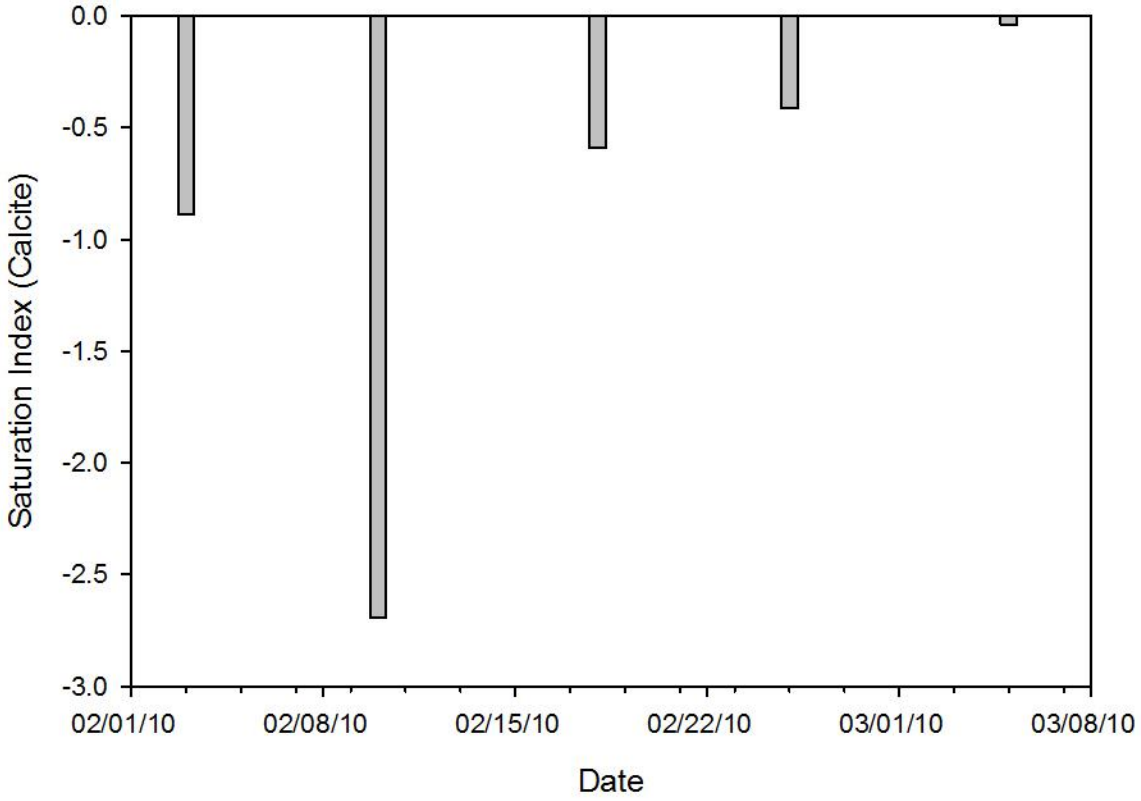

Figure 4-5. Calculated calcite saturation indices for samples collected from Madison Blue Spring. Values are calculated from water chemistry samples collected from 2/3/2010-3/05/2010.

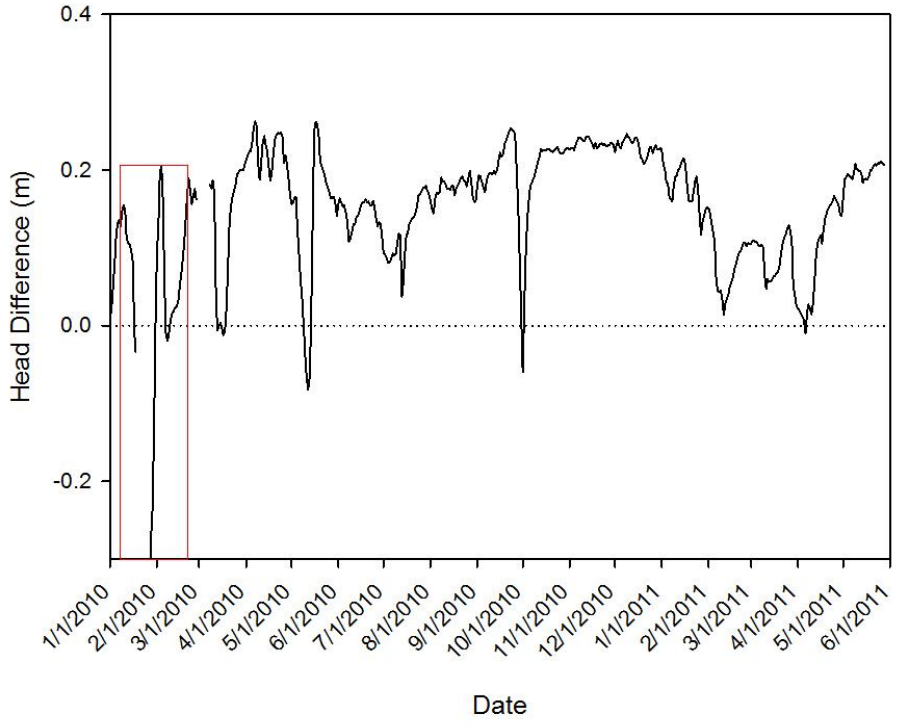

Figure 4-6. Head difference between Madison Blue spring run and Well 7015 in 2010 and 2011. Reversals discussed in text are marked by red boxes.

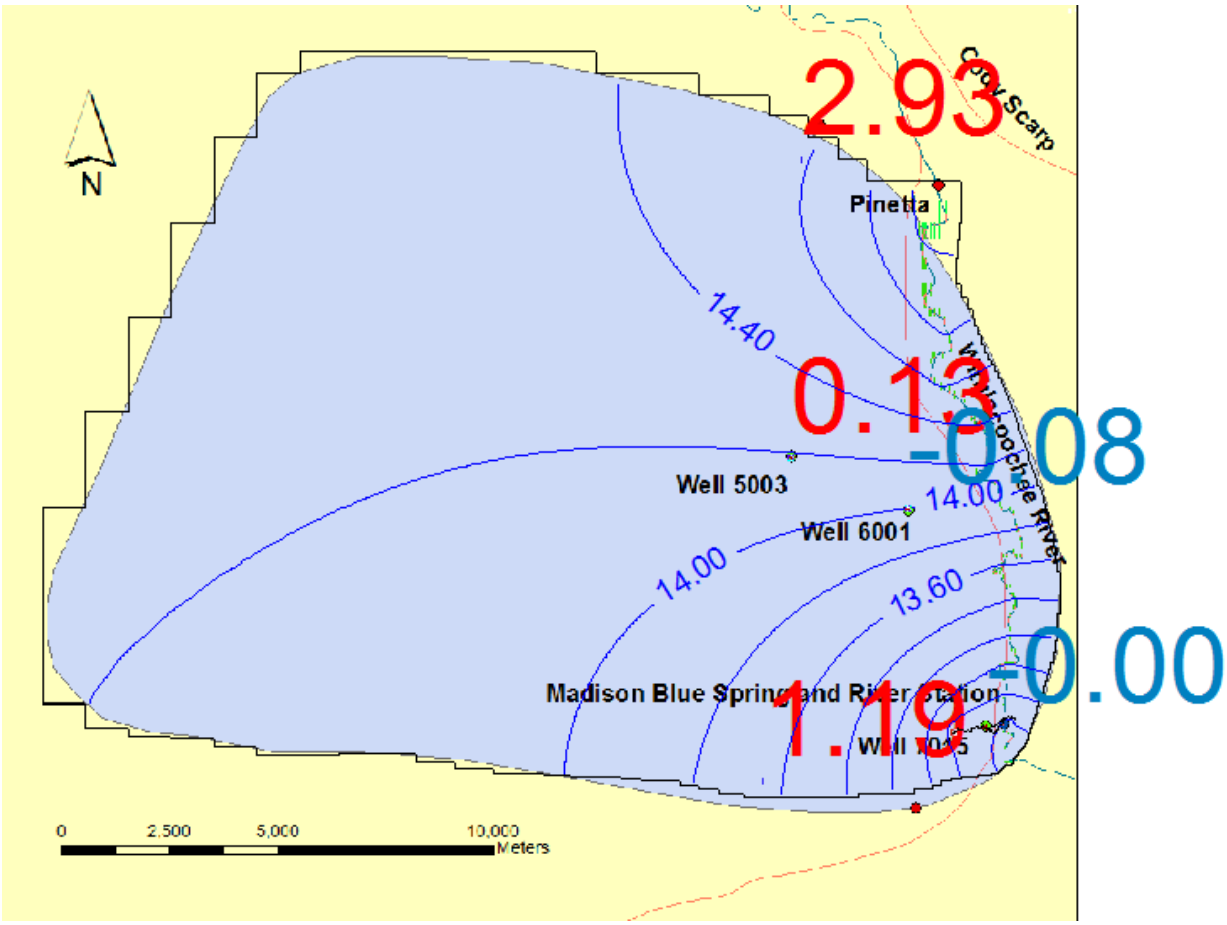

Figure 4-7. Steady State simulation head contours and residuals. Residuals are reported in meters. The 2.81 m residual reflects Well N0001 which is not part of the springshed and not quantitatively used for the calibration.

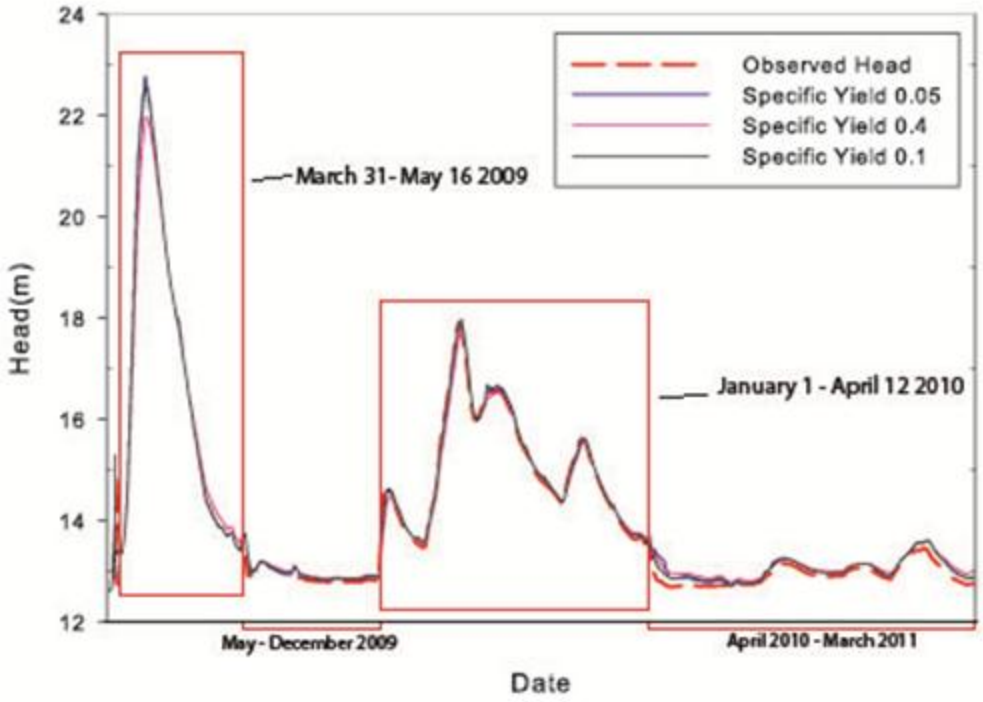

Figure 4-8. Specific yield sensitivity at Well 7015. Red boxes indicate time frame in which flood occurred and daily stress periods were used.

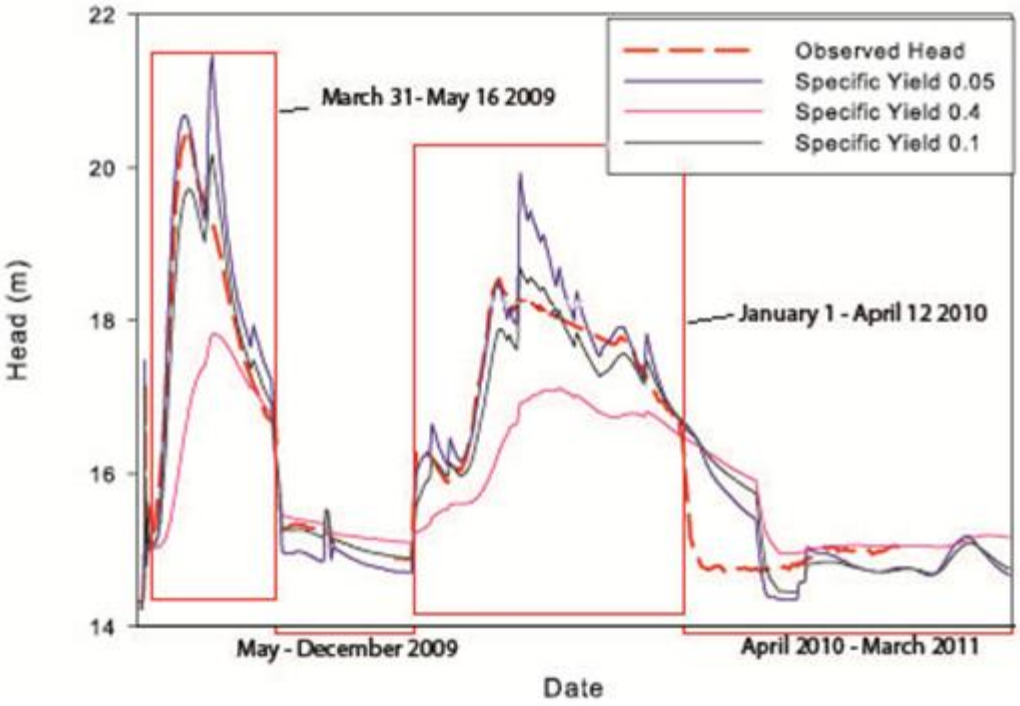

Figure 4-9. Specific yield sensitivity at Well 5003. Red boxes indicate time frame in which flood occurred and daily stress periods were used.

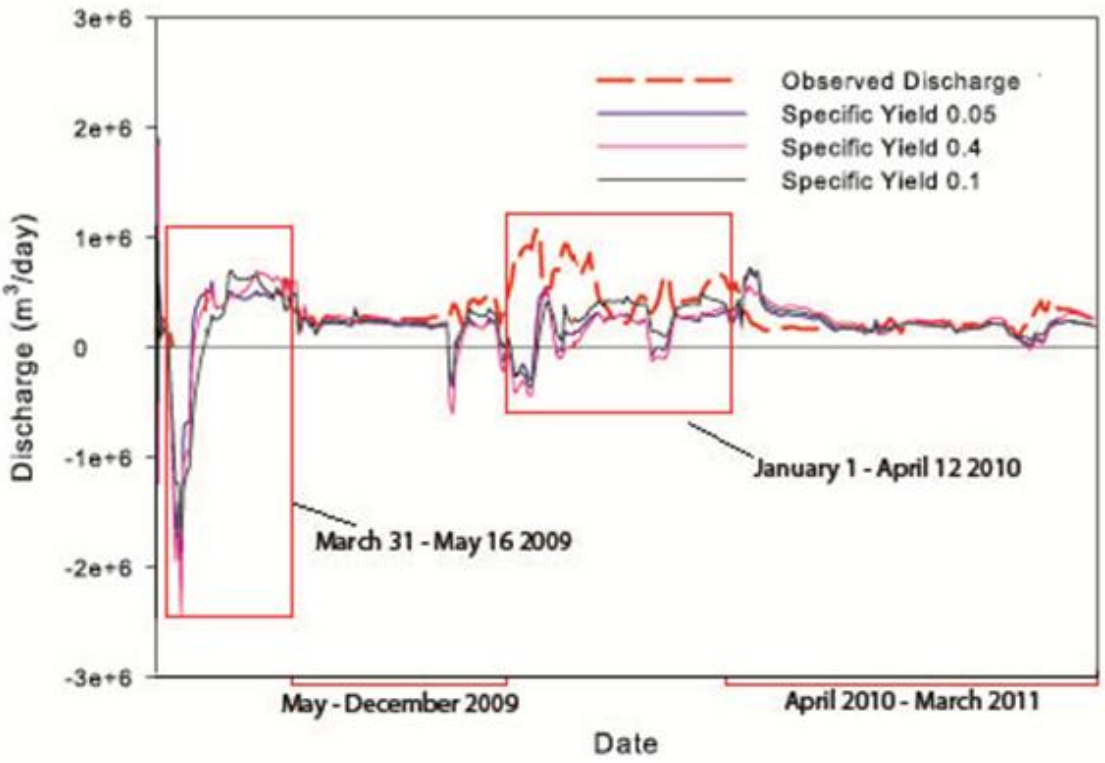

Figure 4-10. Specific yield sensitivity at Madison Blue spring. Red boxes indicate time frame in which flood occurred and daily stress periods were used.

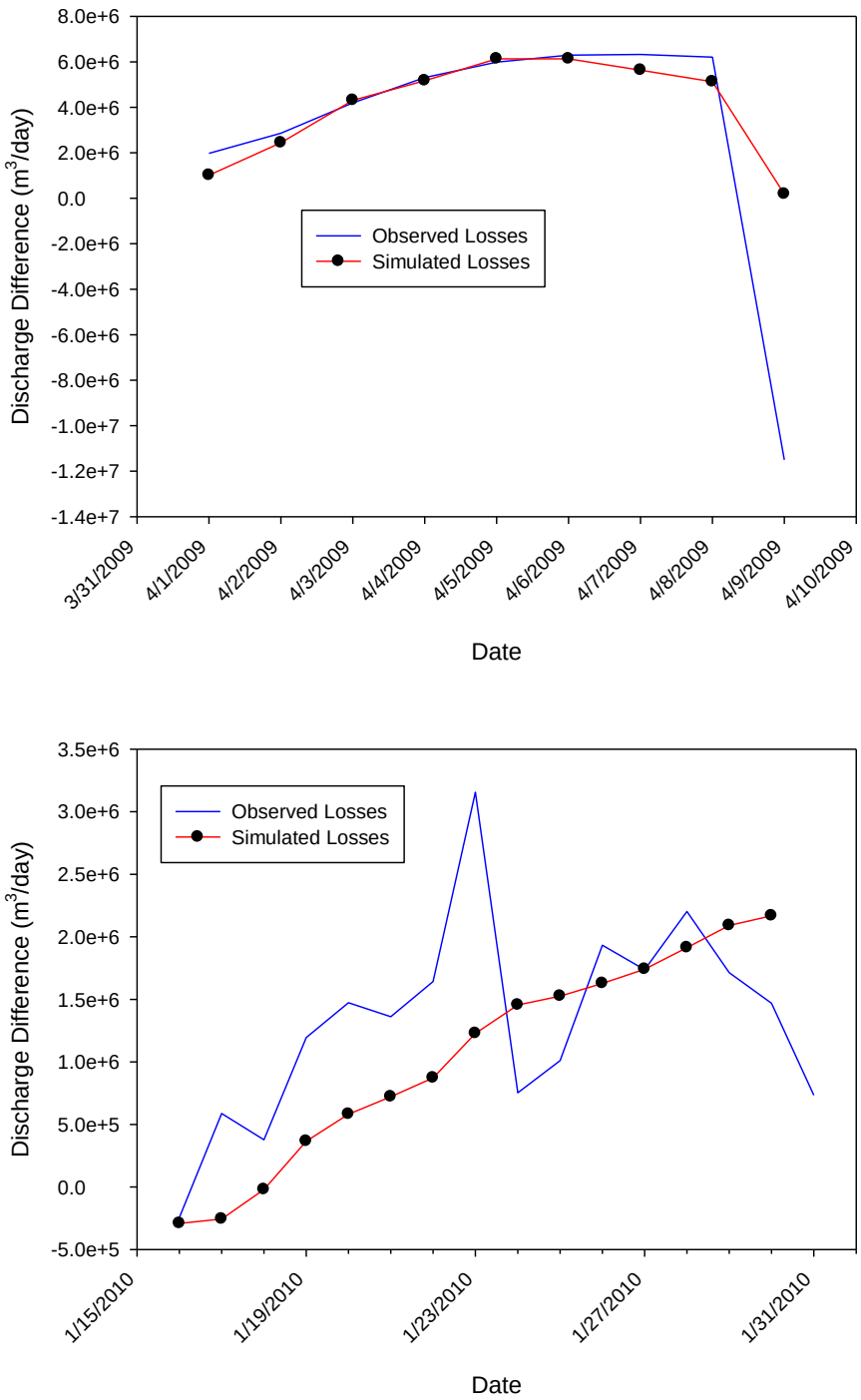

Figure 4-11. Simulated Withlacoochee River losses between Pinetta and Madison compared to observed losses. (TOP) Losses during 2009 event (BOTTOM) Losses from first reversal during the 2010 event. Observed losses are the discharge difference between the stations corrected for estimated changes in surface storage.

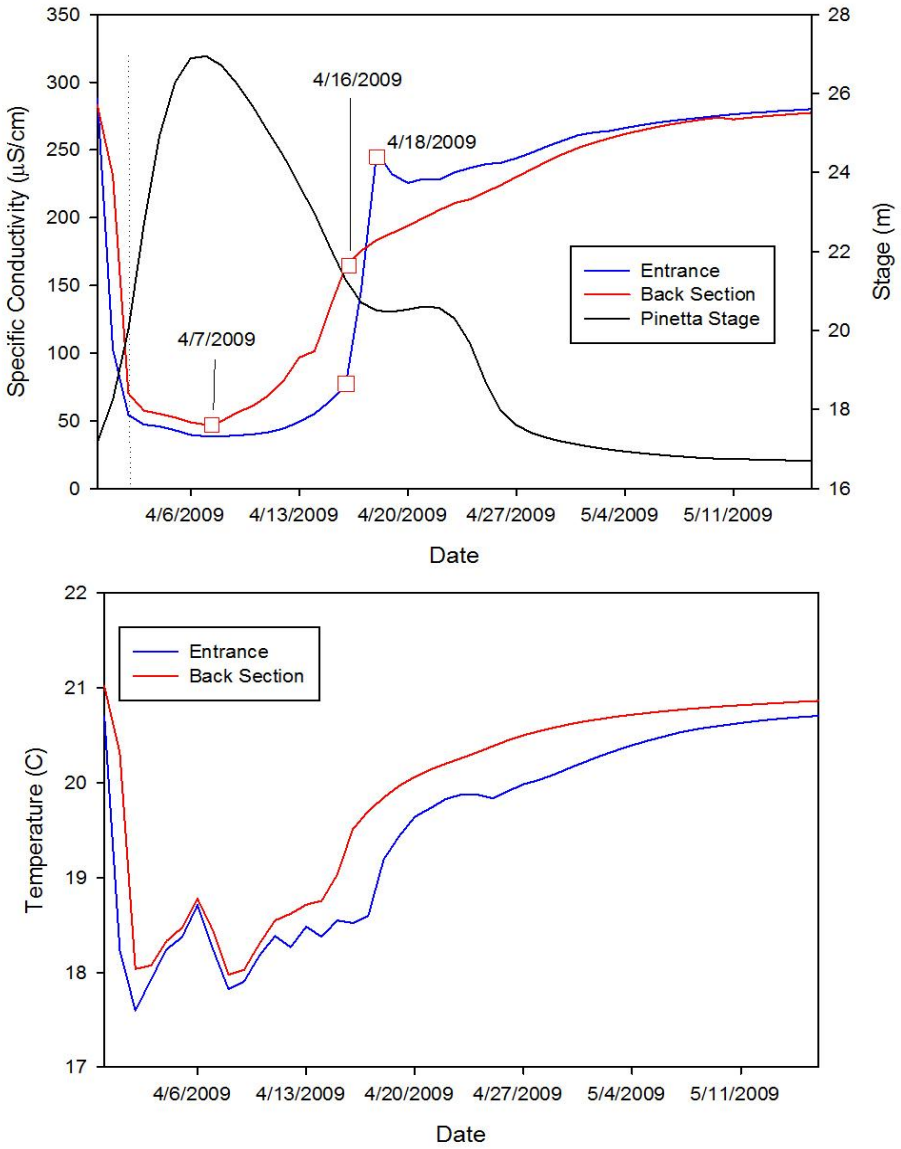

Figure 4-12. Specific conductivity and temperature signal in Madison Blue cave system. Data was obtained from sensors that were placed at Back Section and the entrance to Madison Blue Spring. Period of record is 3/31/2009-5/16/2009. Cited dates noted by red boxes. Solid line indicates start of simulated reversal. Refer to Figure 2-5 for locations.

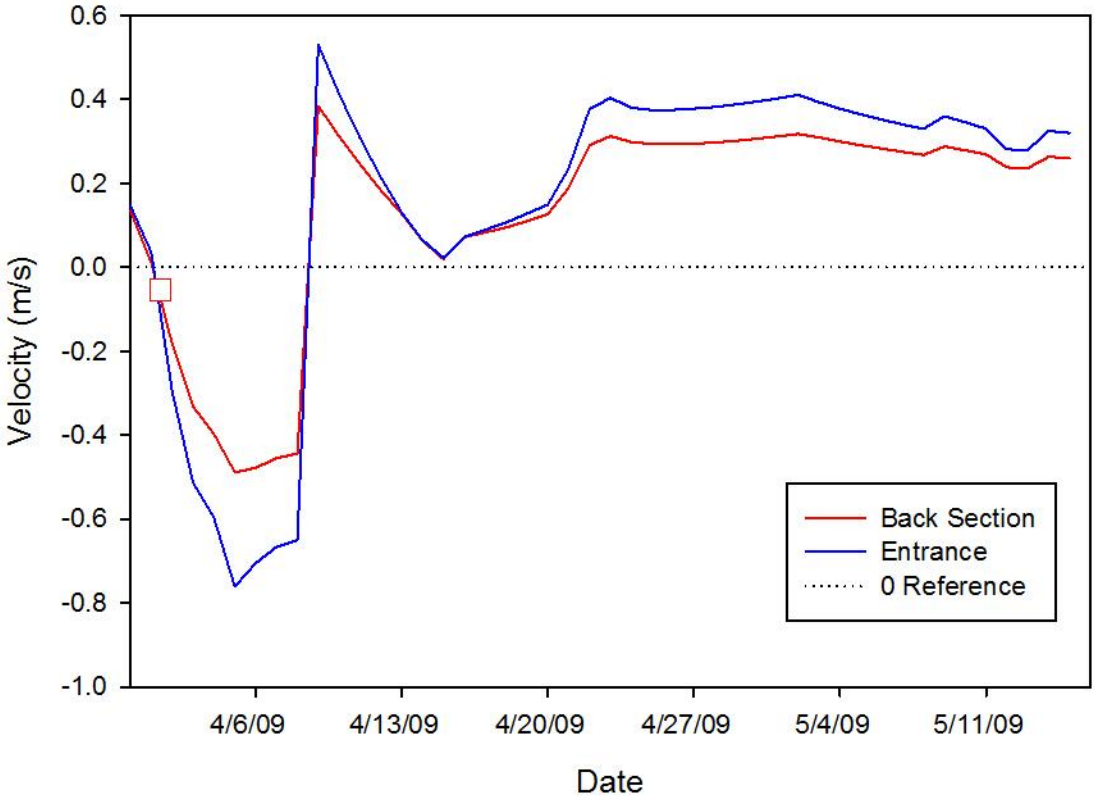

Figure 4-13. Simulated velocity profile in the Madison Blue cave system. Period of record is for 3/31/2009-5/16/2009. The velocity inferred by Gulley et al. (2011) marked by a red box. Positive velocities indicate flow toward Madison Blue spring, and negative velocities indicate flow away from Madison Blue spring (reversal).

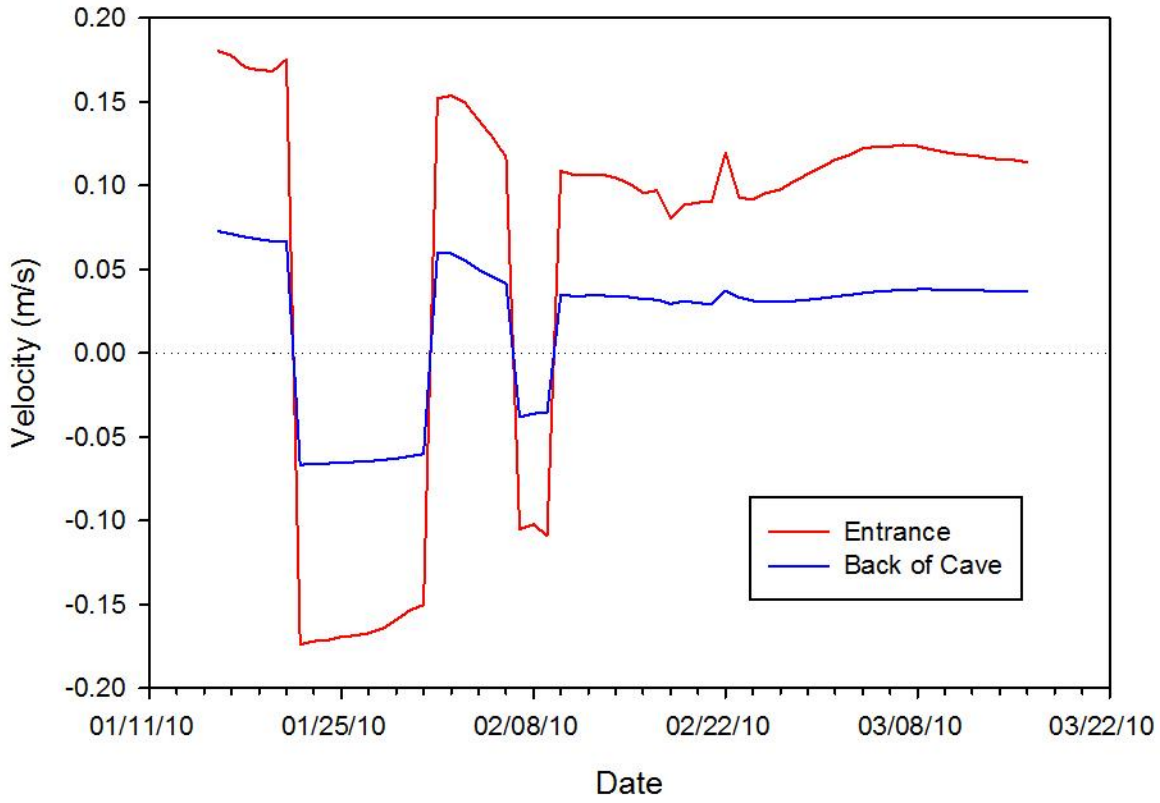

Figure 4-14. Simulated velocities at Madison Blue entrance and back of cave. Period of simulation was from 1/16/2010-3/15/2010

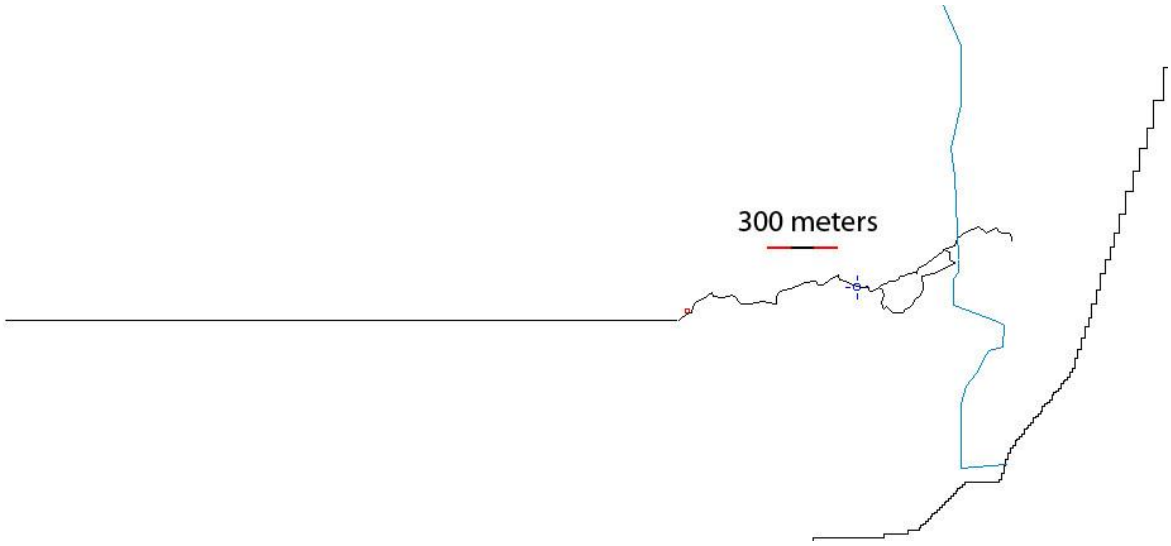

Figure 4-15. Simulated extended conduit system. Red dot indicates the end of the mapped cave system and blue line indicates the simulated Withlacoochee River.

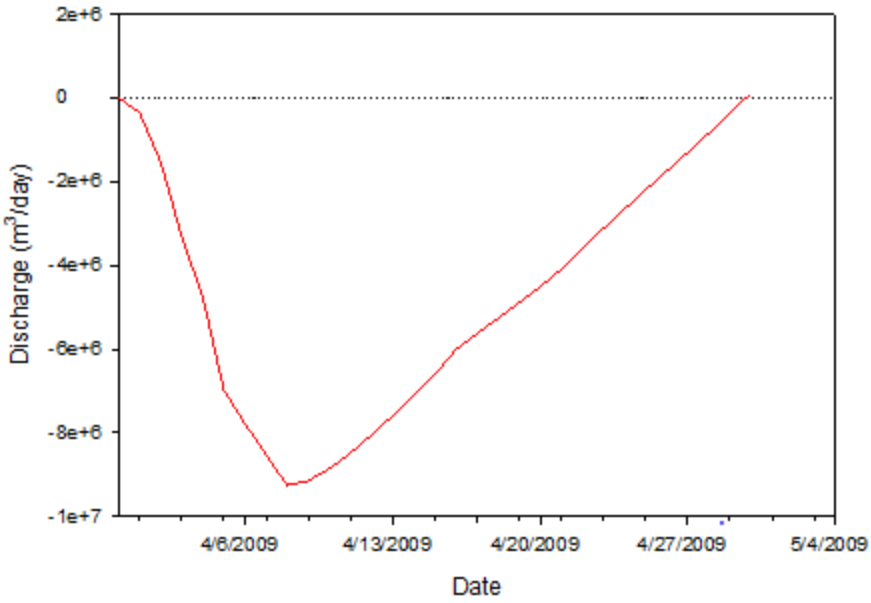

Figure 4-16. Simulated cumulative discharge for 2009. Simulated period is from 3/31/2009-5/16/2009. Negative values indicate flow into Madison Blue conduit system. Positive values indicate outflow.

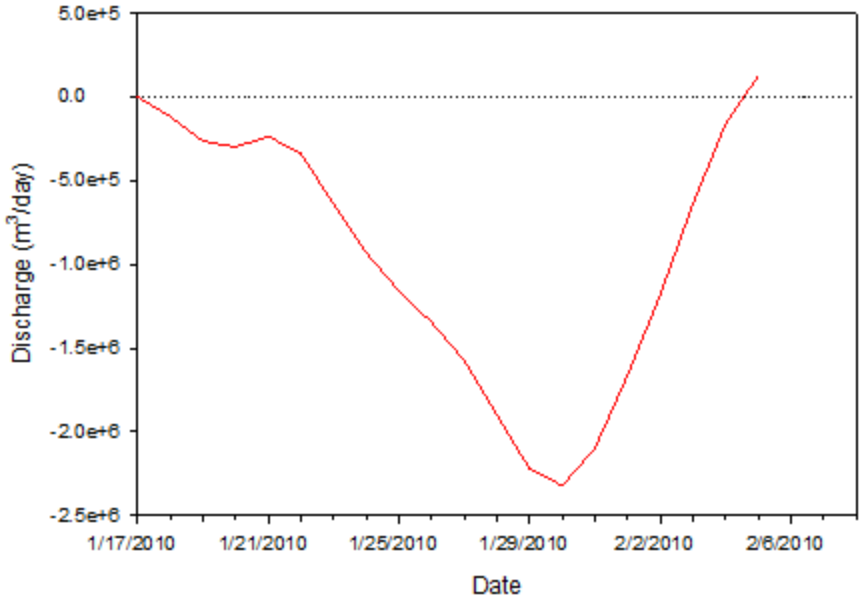

Figure 4-17. Simulated cumulative discharge from 2010. Simulated period is from 1/16/2010-2/5/2010. Negative values indicate flow into Madison Blue conduit system. Positive values indicate outflow

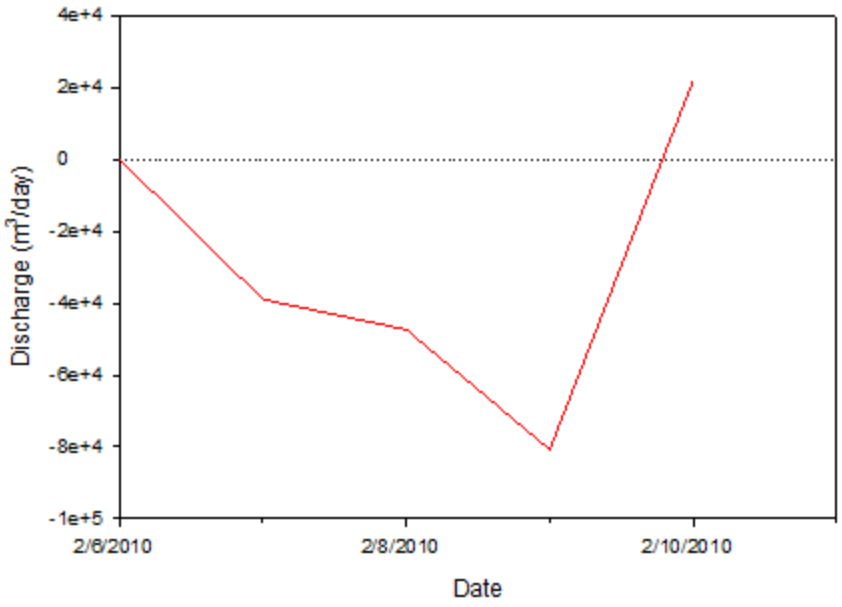

Figure 4-18. Simulated cumulative discharge from second reversal in 2010. Simulated period is from 2/6/2010-2/11/2010. Negative values indicate flow into Madison Blue conduit system. Positive values indicate outflow

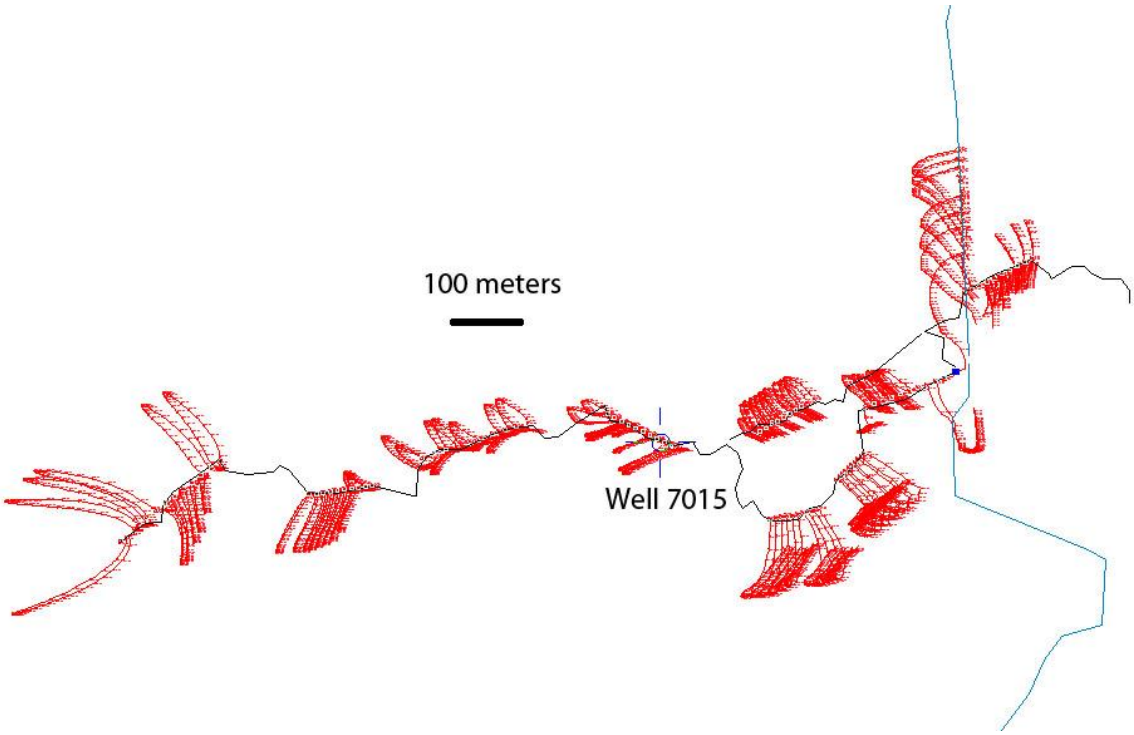

Figure 4-19. Results of particle tracking from 2009 event. A total of 172 days were simulated. Blue circle is Well 7015

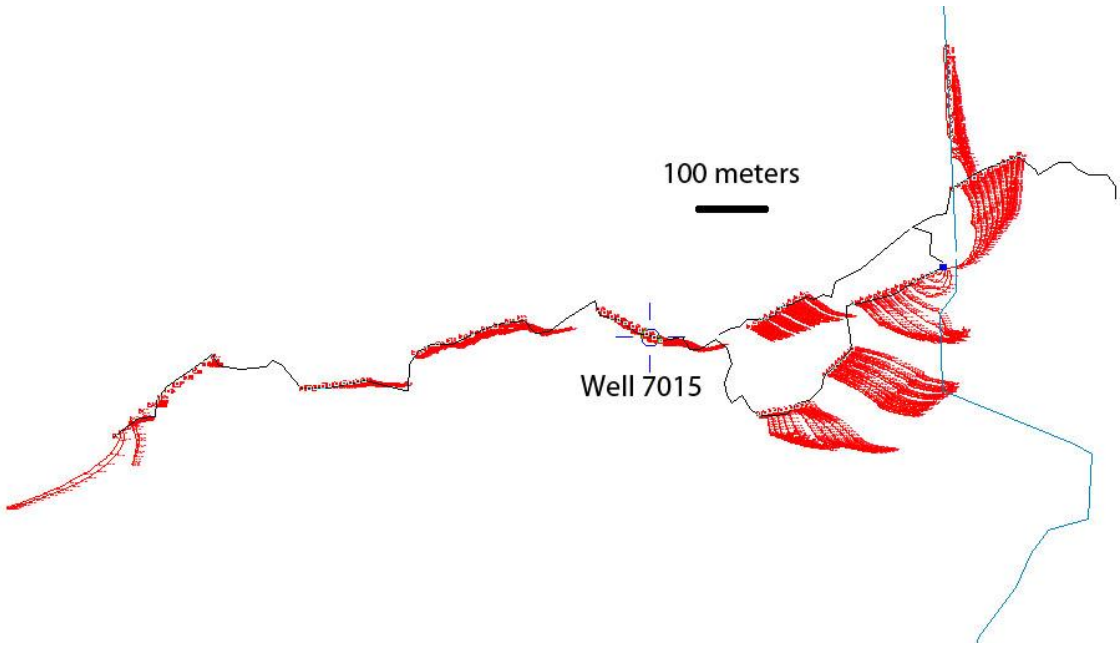

Figure 4-20. Results of particle tracking from 2010 event. A total of 350 days were simulated.

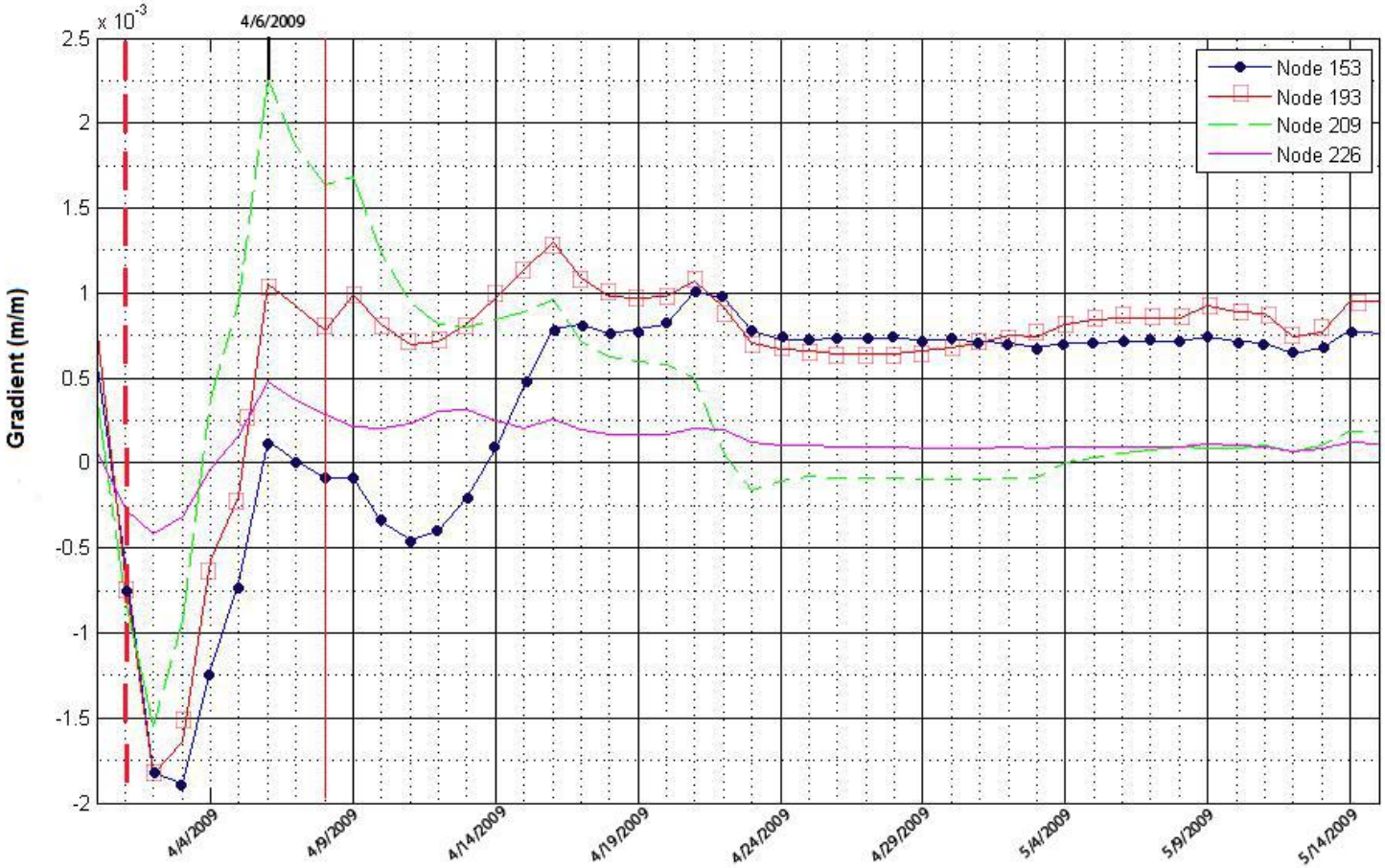

Figure 4-21. Gradient between conduit and formation during 2009 flood. Positive values indicate water is flowing from matrix to conduit, and negative values indicate flow from conduit to matrix. Red dashed line indicates where intrusion begins, solid red line indicates where discharge at the spring begins.

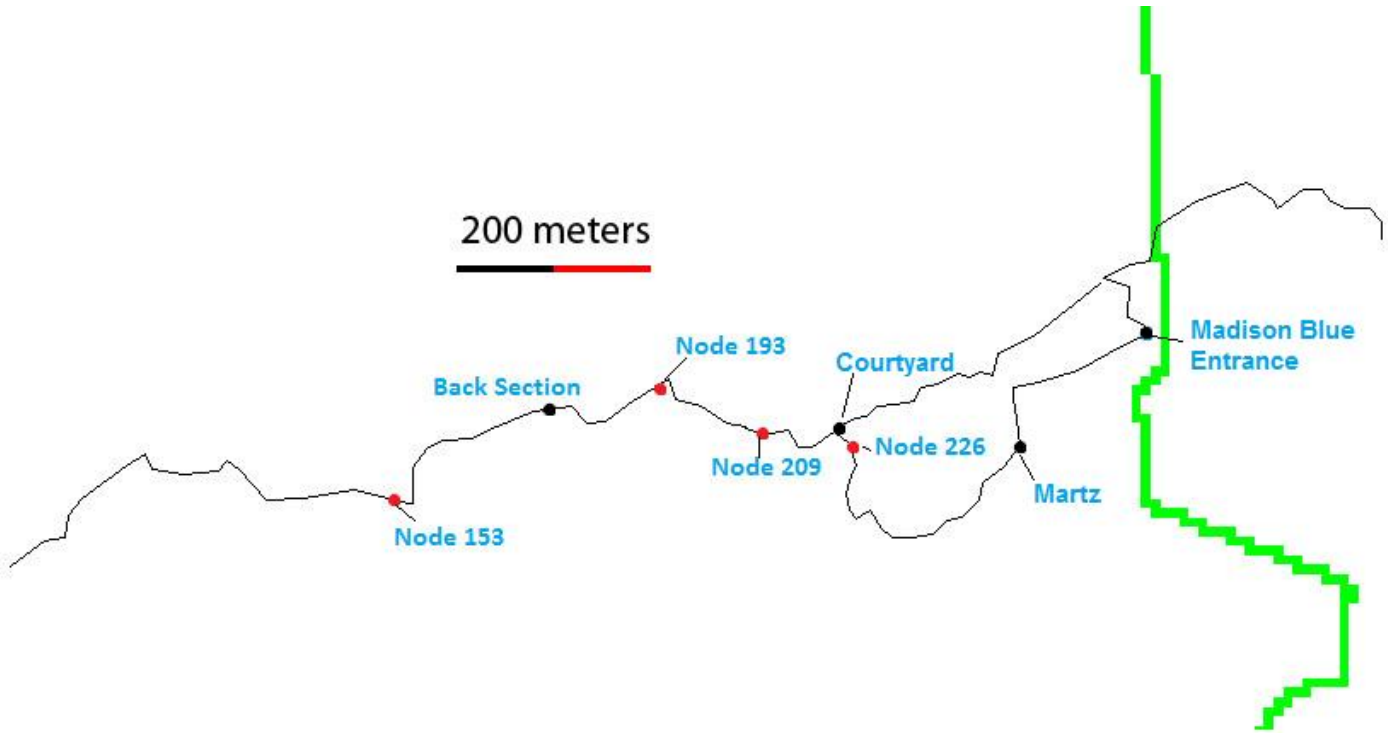

Figure 4-22. Locations for nodes in Figure 4-21.

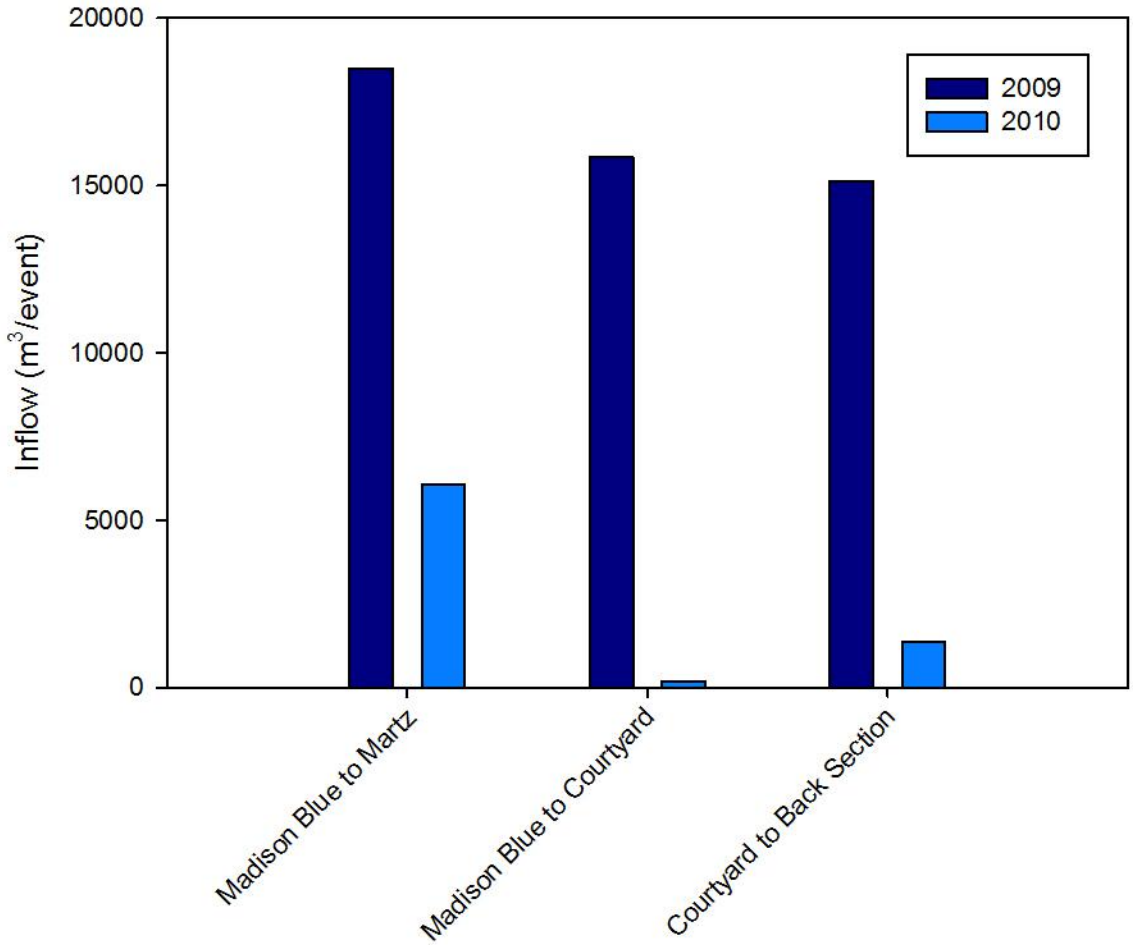

Figure 4-23. Total inflow into the matrix for Madison Blue cave system. Values normalized to segment length. See Figure 4-21 for locations.

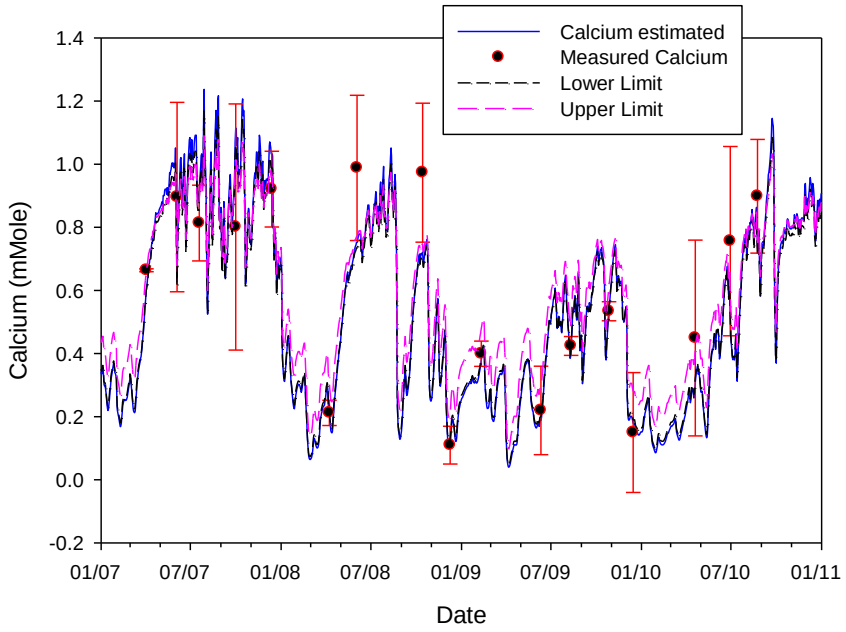

Figure 4-24. Estimated calcium (mMole) through time at the Pinetta station of the Withlacoochee River. Red dots indicate calcium values from USGS sampling results.

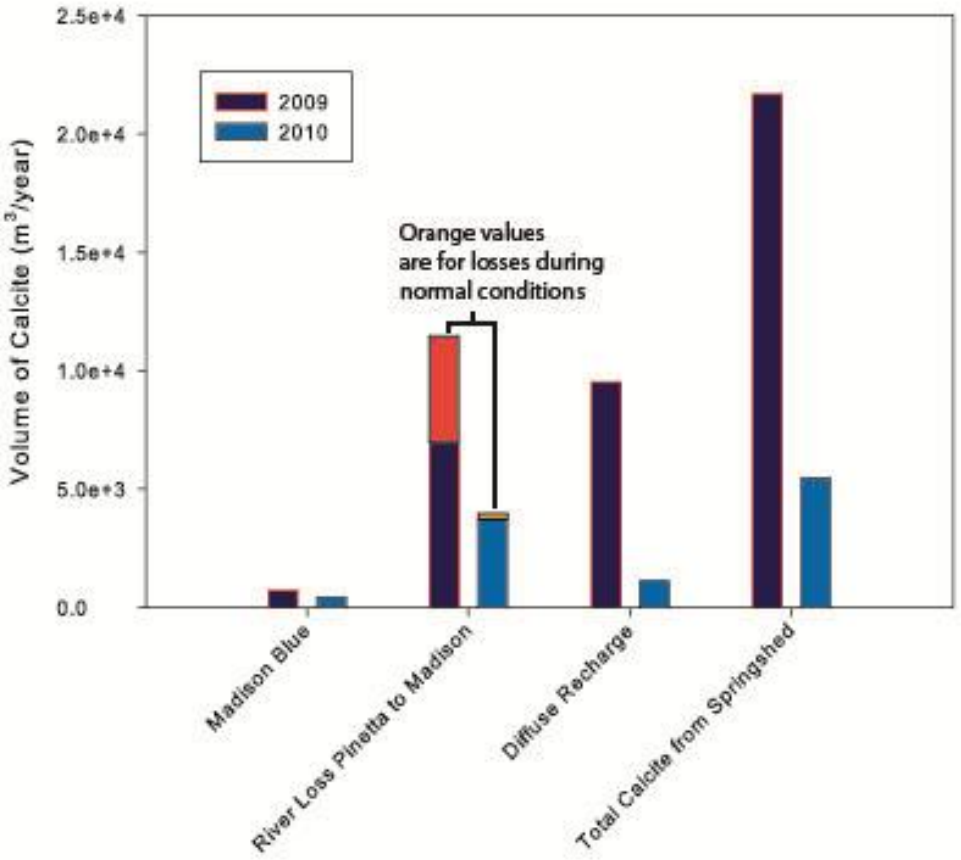

Figure 4-25. Bar chart showing comparative dissolution for each process in 2009 and 2010. Orange values for River Loss Pinetta to Madison are for dissolution under normal conditions, and blue values signify dissolution due to events. Total calcite dissolved is the sum of all the calcite dissolved for each year.

CHAPTER 5
DISCUSSION

Influences on Exchange from the Madison Blue conduit system

The 2009 and 2010 events show very different patterns of exchange, as indicated by particle tracking (Figure 4-19 and 4-20), velocities with the conduits (Figure 4-13 and 4-14), and simulated exchange between the conduit and surrounding aquifer (Figure 4-23). During the 2009 event, water left the conduit network throughout the entire length of simulated conduit, indicating that a large fraction of the intruded water could be in contact with the formation. In contrast, particles in MODPATH show that infiltration into the formation during the 2010 events did not occur past Well 7015. In addition, average penetration depth for 2009 in the front was 115 meters and the average for 2010 was 96 meters.

The magnitude of the pulse most likely played a role in the differences in volumes of estimated infiltration between 2009 and 2010, as the river stage in January 2010 was only 5 meters higher than pre-flood conditions as compared to an increase of 10 meters in April 2009. In addition, the antecedent water table slope toward the river dampens exchange. Thus, if the pulse of water that increases head in the conduit is not increased higher than the background water table, exchange from conduit to matrix will not occur. In December 2009, rainfall was high and the elevated river stage initiated a reversal. This was evident from the Well 7015 to Madison Blue spring head differences. The high water conditions as a result of the December 2009 recharge and reversal may have inhibited the length of the reversal in January 2010.

Recharge during the 2009 event was much greater than in 2010 (0.18 m in 2009 compared to 0.04 m in 2010). Given the estimated specific yield of 0.1, this would yield

a 1.8 m rise in the water table during 2009 and a 0.4 m rise in 2010. It therefore would have reduced the total exchange volume and the reversal times for each event, suggesting that the intrusion during 2009 could have been larger if local recharge had not been so high.

Losses from the Withlacoochee River between Pinetta and Madison stations also impact exchange to the aquifer at the Madison Blue conduits, but that the relationship is complex as shown by the simulations using constant river conductance. In 2009 intrusion volumes increased by 14% when a constant river conductance was applied rather than assuming that the conductance increases with stage. Greater intrusion of water from the river enabled by higher river conductance would increase heads in the aquifer and reduce conduit outflow to the formation. This could have been responsible for a lesser reversal duration when stream conductance was increased during events in 2009. In contrast, intrusion during the 2010 event was 9% lower when stream conductance was held constant. By increasing the stream conductance with stage, total reversal length in 2010 increased by 3 days. More complicated hydrological conditions may have occurred as a result of high water conditions during late 2009 affecting the 2010 event.

Carbonate Equilibrium

The amount of inflow of river water into the conduits and subsequently the matrix will affect the rate and amount of carbonate dissolution in the subsurface. Carbonate dissolution rates are greatly impacted by the amount of contact with limestone (Equation 3-14). Because the ratio of surface area contact to volume of water is higher in the matrix than the conduit, the more water that enters the matrix the faster the rate of

dissolution. This is illustrated by estimated carbonate dissolution rates 90% faster in the matrix than in the conduit.

The complexity of the aquifer makes it difficult to predict the flow paths of the infiltrated river water. Despite the inclusion of ~3 km of mapped conduits in the numerical model, it remained necessary to incorporate a bulk regional hydraulic conductivity that combines conduits and matrix flow. The calibrated porous medium hydraulic conductivity for the model, 2000 m/day, is significantly larger than reported matrix hydraulic conductivities for the Ocala and Suwannee Limestone (0.73 – 6 m/day). This large hydraulic conductivity requires the presence of additional unmapped conduits or fractures. The sensitivity run in which the length of the conduit was doubled allowed the hydraulic conductivity of the porous medium to be dropped by 25% to 1500 m/day. This decrease in hydraulic conductivity is significant, and suggests that if conduit locations are well known, it may be feasible to accurately simulate matrix flow in the porous medium. However, mapping sufficient conduits to reduce the simulated hydraulic conductivity to that of the matrix (<6 m/day) seems unrealistic as it is impossible to know the locations of all the conduits and fractures. Therefore it would be difficult to accurately model these systems. The particle tracking helps to visualize flow directions, but the actual flow field will be more complex. More water will be forced into areas of least resistance (higher hydraulic conductivity) and therefore preferentially exchange with higher hydraulic conductivity zones. As a result, the dissolution will be distributed around an extensive network of conduits rather than only around the mapped conduits (Moore et al. 2010).

Comparison of observations to simulation results provides additional insight to exchange processes. In some portions of the conduit system, aquifer to conduit flow initiates on 4/6/2009 (Figure 4-21) though conduit flow is still reversed. This corresponds to an increase in observed specific conductivity at Back Section, suggesting the potential for some addition of dissolved solids from the matrix. Simulated conduit outflow begins on 4/9/2009 and specific conductivity records indicate very dilute water (specific conductivity < 80 µS/cm at Entrance) expelled from the system until 4/16/2009. This indicates that a significant quantity of intruded water was discharged from the spring prior to equilibration. Calculated rates of dissolution show that water that enters the matrix could reach equilibrium as quickly as 5 days. Therefore, it seems likely that most water expelled prior to 4/16/2009 was stored within conduits, with the increase in specific conductivity due to dissolution of the conduit walls and the rapid increase on 4/16/2009 could be either water coming out of matrix storage that had reacted or pre-flood water.

Chemistry data obtained at Madison Blue spring during the 2010 event also indicate that intruded water did not fully equilibrate with the formation before it discharged from the spring (Figure 4-5) and was still undersaturated 30 days after the onset of reversal. Calculated rates with PHREEQC using a representative water sample from 2010 estimated that water in the conduit would take about 30 days to reach 90% saturation which is consistent with chemistry data obtained from the spring (Figure 4-5). Therefore, it appears as though the majority of water that discharges at the spring likely represents water that only came in contact with the limestone walls of the conduit rather than being within the matrix. The 2010 event did not cause as much intrusion into the

formation as in 2009, and MODPATH estimates indicate that the water that enters the formation could remain there for a significant amount of time, if it comes out of storage to Madison Blue Spring at all.

Calculated Dissolution

We estimated equivalent lowering rates (Figure 5-1) for dissolution from river water intrusion into the aquifer under normal and flood conditions and compared this to total dissolution from recharge for each year. It is hard to know over what area the scarp losses between Pinetta and Madison stations would dissolve under normal conditions, as the losses occur through both the Fennel's Funnel siphon and diffuse river loss. To estimate the area in which the dissolution would occur under normal conditions, particle tracking during normal flow was used to determine how far that particle would travel. Estimated travel distance was ~600 m from the river during the time frame it would take for water to reach equilibrium in the conduit (about 120 days). Using this distance with the estimated conduit diameter of 3 meters, the equivalent lowering rate for water lost at the scarp under normal conditions was 10^{-1} m/year for 2009 and 10^{-2} m/year for 2010.

To determine the equivalent lowering rate around the Madison Blue spring conduit network and at the scarp during the flood events, the area of influence by dissolution was calculated by MODPATH (3.5×10^5 m^2 in 2009 and 2.1×10^5 m^2 in 2010) and carbonate equilibrium for each event was used to constrain rates. The equivalent lowering rate around the Madison Blue cave system during each flood was about 10^{-3} m/year, which is two orders of magnitude higher than regional estimates for Florida (10^{-5} m/year - Opdyke et al. 1984, Denizman et al. 1998, Adams et al. 2010). As discussed above, the outflow of low specific conductivity water in 2009 and

undersaturated water in 2010 indicates that dissolution rates due to river intrusion are overestimated.

For dissolution from river losses between Pinetta and Madison stations, the equivalent lowering rate was 10^{-2} m/year for each event. For the river losses at the scarp during a flood, the same area calculated from MODPATH simulations for the Madison Blue cave system were used to constrain an area of influence. However this area would probably be an underestimate since the larger amount of infiltration from the river above Madison Blue spring could potentially push water further into the matrix.

The dissolution and lowering rate estimates suggest that subsurface weathering around the river at the scarp (between Pinetta and Madison) is happening at a much higher rate than elsewhere in the springshed. The total dissolution from floods (Madison Blue spring losses as well as Withlacoochee River losses between Pinetta to Madison stations during the floods) were significantly less than estimated dissolution from normal losses in 2009. During 2010, flood and normal losses were similar in magnitude. Though the lowering rate is at a maximum at 10^{-1}-10^{-2} m/year for normal river losses, total dissolution estimates suggest this contributes to a significant proportion of subsurface weathering regardless of the uncertainty in areal influence for normal losses. It is possible that these losses could make their way to another discharge point in the river before doing much dissolution in the subsurface; however, this is unlikely since there is a net loss of water from this part of the river throughout the year. Therefore it could remain in the subsurface for a significant amount of time and distance allowing it to reach equilibrium in the springshed.

Although Gulley et al. (2011) focused on upstream dissolution in conduits due to river intrusion during spring reversals the potential magnitude of dissolution from river losses under normal conditions at the Cody Scarp is significantly larger than that contributed by spring reversals. At least one siphon, Fennels Funnel, captures water from the Withlacoochee River. The magnitude of dissolution from Pinetta to Madison during normal and flood conditions suggest the potential for the development of a sink-rise system in this area. If competing surface dissolution in the Withlacoochee River is not as great as that due to dissolution of the funnel, eventually all water from the river could be diverted underground. A larger conduit would increase the volume of water that flows through, but there would be a decrease in the surface area to volume ratio and therefore potentially decrease in dissolution per increase in water volume. This would change during floods, as water would be forced into the matrix. Therefore any development of a sink rise system would alter the amount of dissolution that occurs under varying conditions as well as the overall hydrology in that area.

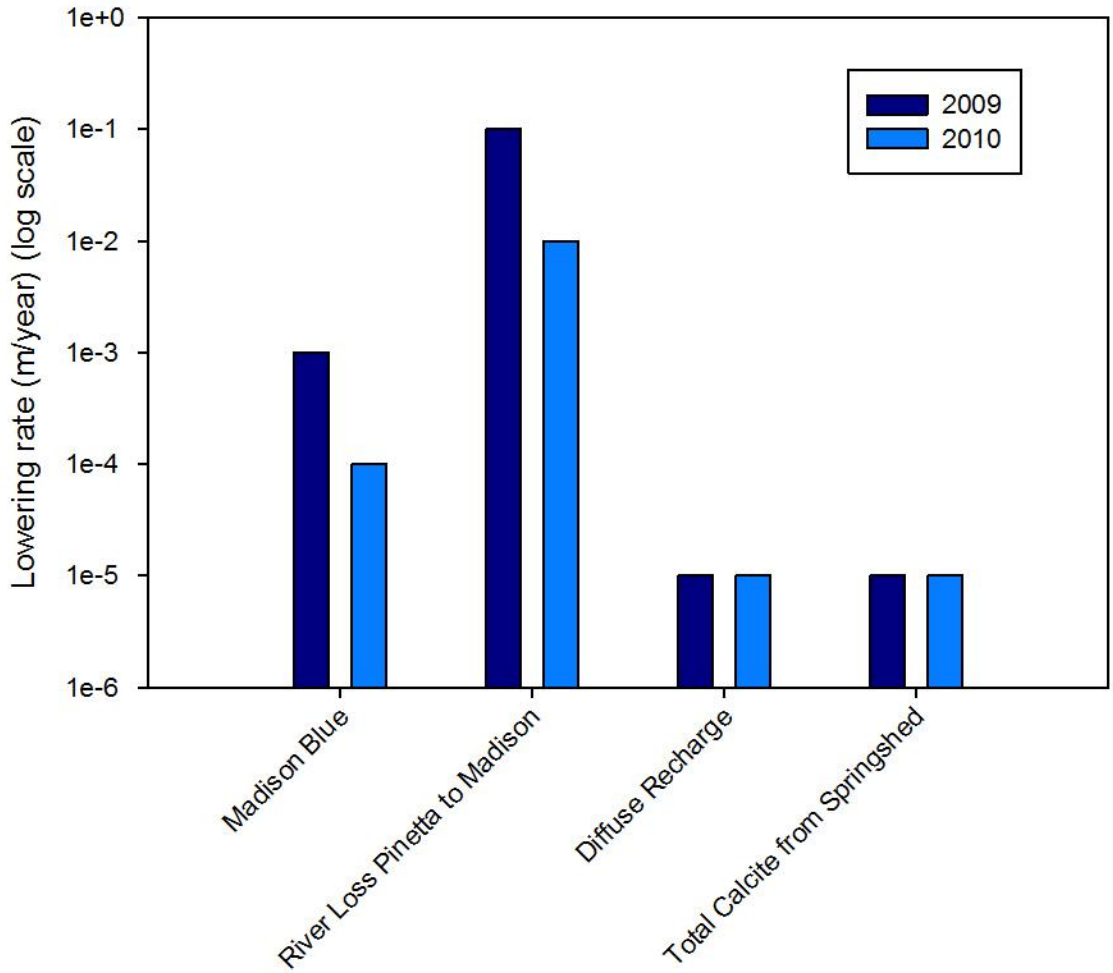

Figure 5-1. Equivalent lowering rate by location for the springshed. Total calcite from springshed is the sum of the all the calcite dissolved (Figure 4-25) and divided by the area of the springshed.

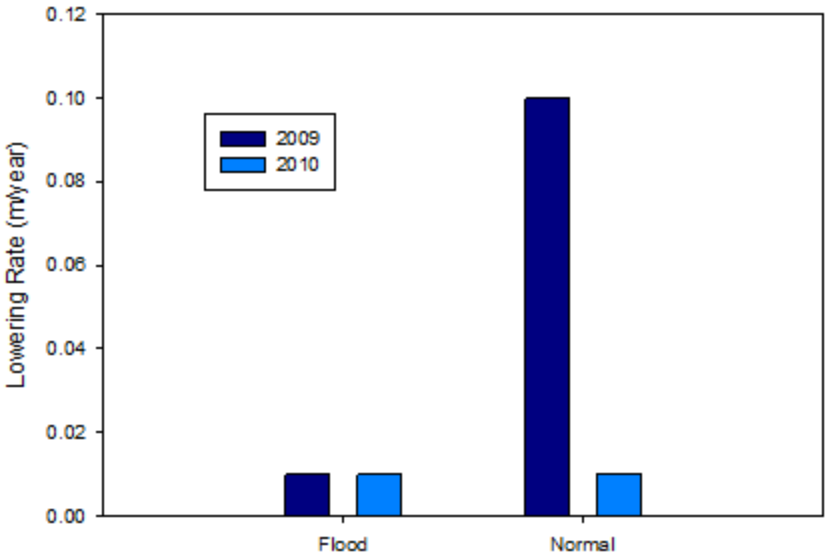

Figure 5-2. Equivalent lowering rate comparing normal to flood conditions.

CHAPTER 6
CONCLUSIONS

Subsurface dissolution around the Cody Scarp occurs due to diffuse recharge, losses during normal (recession) conditions, as well as due to bank storage and river intrusion into Madison Blue spring during high-discharge events on the Withlacoochee River. Numerical modeling simulated inflow and subsequent outflow into the spring during events and infiltration volumes were used to estimate the magnitude of dissolution in the subsurface. Results indicate inflow volumes during reversals larger than previously estimated, however simulated residence times indicate that when water begins to discharge at the spring again it is unlikely to have reached equilibrium. In contrast, dissolution from losses under normal conditions would likely equilibrate in the subsurface within springshed boundaries. The contribution of losses from the river during normal conditions is significant, and contributes to a large part of the overall carbonate weathering budget in the springshed. Carbonate dissolution due to floods plays an important role on impacting the carbonate weathering budget on shorter time scales. An important consequence of river losses between Pinetta and Madison could be the development of a sink rise system for the Withlacoochee River, as has occurred on other Suwannee Basin rivers. The addition of subsurface dissolution due to normal conditions increases the development of this system. If competing dissolution at the surface is not as fast, all river water could eventually be diverted underground, resulting in changes in overall hydrology of the springshed as well as changes in dissolution magnitudes under normal conditions.

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BIOGRAPHICAL SKETCH

Patricia Spellman was born and raised in the small, quiet town of Hollywood, FL. At the tender age of 5, she decided that she wanted to be an astronaut and set out to build the best space ship that would take her and her cats to the moon. However when she turned 12, she had no spaceship and a few less cats, but discovered while hiking in Kings Canyon Australia that maybe a career outdoors was the move for her. So she sailed through school and into a career in the geosciences where she hopes to experience many more adventures…with a little less modeling.