# QUANTIFICATION OF THE MATRIX HYDRAULIC CONDUCTIVITY IN THE SANTA FE RIVER SINK/RISE SYSTEM WITH IMPLICATIONS FOR THE EXCHANGE OF WATER BETWEEN THE MATRIX AND CONDUITS

By

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Rapid influx of surface contaminants to the subsurface through dissolution features makes karst aquifers especially vulnerable to contamination. Quantifying mixing rates between conduit water and matrix water will provide valuable insight into methods for protecting karst groundwater resources. Determining matrix hydraulic conductivity is an important factor for determining mixing rates between the matrix and conduits.

The Santa Fe River is a sinking stream in north central Florida. Water flows underground at the River Sink and travels for approximately 8 km through conduits before re-emerging as a first magnitude spring named River Rise. Temperature and water levels were collected from the River Sink, seven intermediate karst windows, the River Rise, and monitoring wells between 2001 and 2003. These data were used to estimate the water velocity through the subsurface between the Sink and the Rise, the volume of water

lost or gained from the conduits, and hydrologic properties of the conduits. Data from monitoring wells and conduits allowed analyses of the matrix groundwater gradient fluctuations in response to recharge pulses, and clarification of the mixing between matrix and conduit water. Analyses of head gradients revealed that the slope of the matrix groundwater gradient correlates with the volume of water lost or gained from the conduit system, indicating that water from the conduit moves between the conduit and matrix. Transmissivity (T) quantified using passive monitoring methods was calculated between 950 and 550,000 $m^2$/d. Hydraulic conductivity (K), calculated using an aquifer thickness of 275 m, was between 4 and 2000 m/d. T and K values for wells within 400 meters of the conduit are likely low due to the partial penetration of the conduit. Scale may also affect values of T and K calculated within small distances of the conduit. With increasing scale, preferential pathways through the matrix tend to dominate a larger percentage of groundwater flow, increasing average transmissivity. A transect or profile of head on specific days during the March 2003 flood was constructed for Wells 1 and 4. Using Darcy's law and an effective porosity estimate, average linear velocities were determined along the calculated transect. A water particle was traced as it left the conduit using the calculated velocities. During particle tracking simulations for Wells 1 and 4, conduit water migrated between 0.45 and 8.5 m into the matrix, and returned to the conduit in approximately 20 days. Simulations suggest that conduit water is temporarily stored in the matrix and does not enter regional groundwater flow. Preferential flow paths within the matrix as well as the effects of diffusion and dispersion could allow conduit water to migrate further into the matrix than particle tracking simulations suggest, and illustrate the need for further investigation.

# CHAPTER 1
# INTRODUCTION

## Background

The importance of understanding hydrologic processes in karst aquifers is apparent if one considers that more than a quarter of the world's population lives on, or obtains its water from, karst aquifers. In the United States, approximately 20 percent of the land surface is karst (Figure 1-1) and 40 percent of the groundwater used for drinking comes from karst aquifers (Quinlan and Ewers, 1989). Karst aquifers can supply large volumes of fresh water, but the water is not uniformly distributed throughout the subsurface. Three types of porosity control flow through karst aquifers: intergranular, fracture, and conduit. Intergranular porosity, also called primary porosity, can be high in clastic carbonate rocks, whereas chemically precipitated rocks often have very low porosity. Portions of the aquifer where intergranular porosity occurs are referred to as matrix. Secondary porosity within a carbonate aquifer forms from the preferential flow of water through fractures and conduits, which in turn are further enlarged by dissolution processes, resulting in higher hydraulic conductivity than the surrounding matrix. The relative proportions of these different types of porosity within an aquifer can cause permeability and flow rates to vary by several orders of magnitude (Martin and Dean, 2001). Fractures have apertures in the range of 50 μm, to 1 cm, while conduits are typically greater than 1 cm wide (White, 2002). Because it is difficult to distinguish

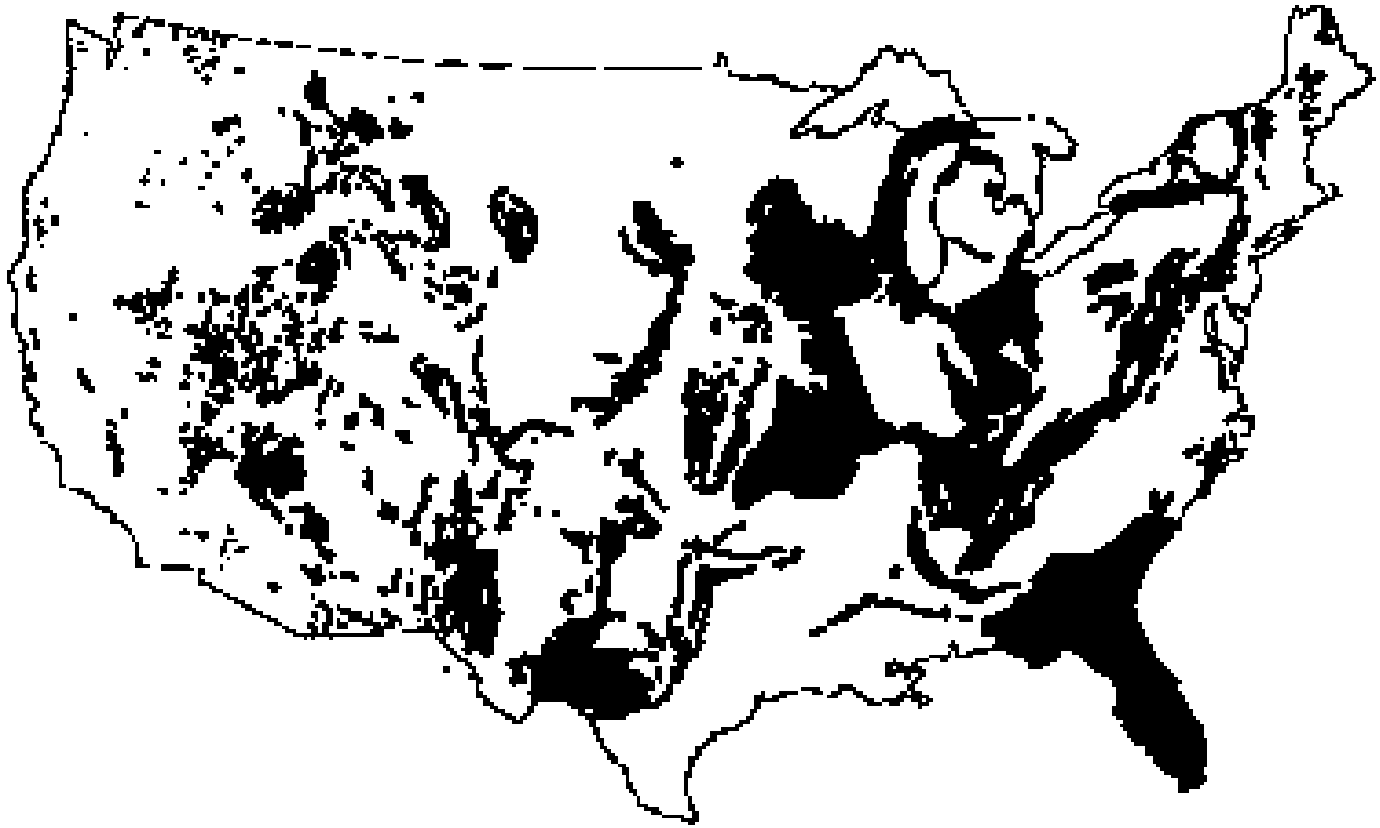

Figure 1-1. Karst regions of the contiguous United States (Davies et al., 1984).

between intergranular and fracture flow, for the purposes of this study, both will be referred to as matrix flow.

Karst hydrology research has undergone significant development in the last forty years. From the initial idea that caves were hydrologically isolated from the flow field, karst hydrology came to signify only the hydrologic properties of conduits during the 1970s and 1980s (White, 2002). In the last decade, researchers have begun to realize that an accurate representation of the hydrologic system must include conduit, fracture and matrix flow components and describe the relationship among them (White, 2002).

Understanding the hydrologic relationship between matrix and conduit systems is crucial for protecting and maintaining groundwater quality in karst regions. The amount of mixing between conduit and matrix water is an important factor for determining the susceptibility of groundwater reservoirs to surface contaminants traveling through conduits. Mixing rates between conduit and matrix water depend on several factors including matrix porosity, transmissivity, and groundwater gradient. Subsurface openings, such as fractures and conduits, allow surface water to travel long distances in a short amount of time with little or no filtration. When surface runoff containing contaminants flows only through conduits, karst springs will have high amplitude, but relatively brief periods of water quality degradation (Ryan and Meiman, 1996). If water is exchanged between conduits and matrix, contaminated surface water may infiltrate groundwater reservoirs. The infiltration of contaminants into the matrix surrounding a conduit can also provide a long-term source of contamination, as contaminants slowly diffuse back out of the matrix and into conduit water.

An important control on the exchange of water between the matrix and conduits is the transmissivity of the matrix. Transmissivity is rate at which water is transmitted through a unit width of the full-saturated thickness of the aquifer for a unit hydraulic gradient. The majority of research on karst has been in regions where extensively recrystallized Paleozoic limestones form the matrix, resulting in little to no movement of water between the matrix and conduit (White, 1999). The relatively young Cenozoic limestones of the Floridan Aquifer have high primary porosity and transmissivity, allowing hydraulic conductivity in the matrix up to four orders of magnitude greater than in Paleozoic limestones (Palmer, 2002). Hydraulic conductivity is transmissivity divided by the full-saturated thickness of the aquifer.

Differences in the hydrologic properties of conduits and matrix rocks make quantifying the transmissivity of karst aquifers difficult. In aquifers where conduit flow dominates or controls a significant portion of groundwater movement, porous media flow theory cannot be applied, at least not on a local scale (Bush and Johnston, 1988). Traditional methods of determining aquifer transmissivity include laboratory tests, which determine transmissivity over distances of centimeters, while single and multiple well slug or pumping tests can provide information over tens of meters (Huntsman and McCready, 1995). In contrast, passive monitoring of water level fluctuations in karst aquifers uses naturally occurring aquifer and conduit fluctuations in response to rain events in combination with analytical methods to determine transmissivity. Passive monitoring, the method used in this study, can provide transmissivity values averaged over a distance of kilometers as well as offer an inexpensive alternative to pumping tests (Huntsman and McCready, 1995).

The Santa Fe River Sink/Rise conduit system, located in the unconfined Floridan Aquifer, provides a relatively controlled study area where water level fluctuations in the conduit and matrix can be monitored for long periods and conduit inflow and outflow can be readily determined. This study examines the relationship between a carbonate aquifer with high hydraulic conductivity and conduits. Physical properties such as water temperature, head gradients, and discharge were used in combination with analytical modeling to determine matrix hydraulic conductivity and to describe the movement of water between the matrix and conduits.

## Study Area

### Location, Physiography, and Climate

The Santa Fe River basin, which is a tributary basin to the Suwannee River, covers an area of approximately 3583 $km^2$ in north-central Florida (Hunn and Slack, 1983). The Santa Fe River originates in the plateau region of North Central Florida from Lake Santa Fe. River discharge is increased by outflow from several lakes, including Lake Altho, Lake Hampton, and Sampson Lake, which have direct surface outlets to the river and by Lake Butler and Swift Creek Pond which drain into tributaries (Skirvin, 1962). The Santa Fe River flows southwest for approximately 50 km until it reaches the Cody Scarp. At the edge of the escarpment, the river sinks and flows underground for approximately 5 km through conduits, reappearing intermittently at several karst windows such as Sweetwater Lake (Figure 1-2), before re-emerging at a first magnitude spring called the River Rise (Martin and Dean, 2001).

O'Leno State Park is located in the Santa Fe River basin, near the border between Alachua and Columbia Counties, Florida. The park consists of approximately 6,000 acres and encompasses nearly all of the Santa Fe River Sink/Rise system.

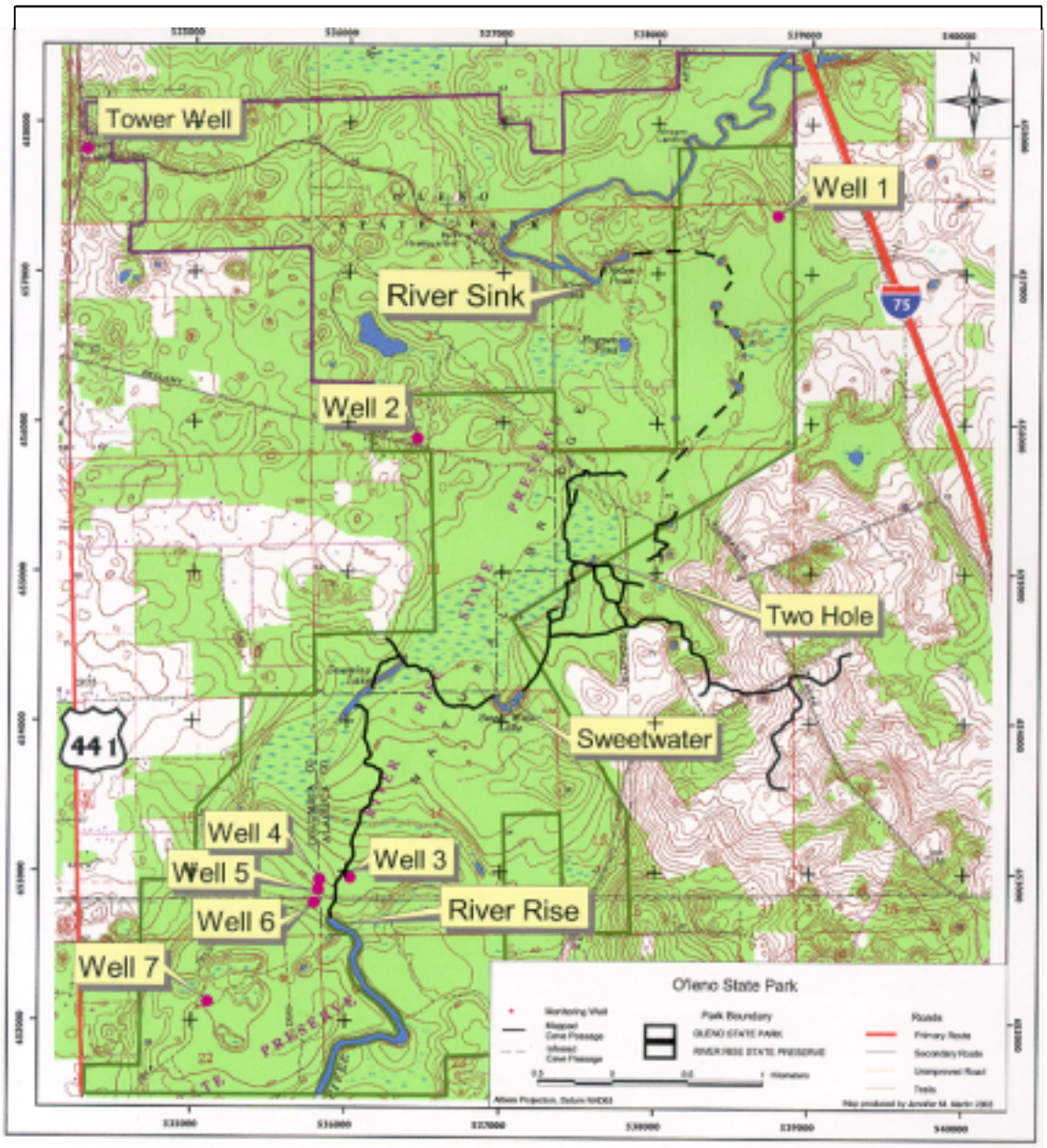


Figure 1-2. Study area including the River Sink, intermediate karst windows, River Rise, and mapped conduits.

O'Leno State Park lies within the Marginal Zone physiographic province. The Marginal Zone (also known as the Cody Scarp) is approximately 2 to 11 km wide and ranges from 15 to 30 meters above mean sea level. The Marginal Zone marks the boundary between the Northern Highlands and the Western Lowlands (Hisert, 1994). The Northern Highlands are plateau-like and are distinguished by elevations in excess of 30 meters and numerous surface streams. The Western Lowlands are typically less than 15 meters in elevation and are characterized as a sinkhole plain with a noticeable lack of surface streams.

**Geologic Background**

The Floridan Aquifer is composed of Oligocene and Eocene carbonate rocks that are between 300 and 800 feet thick in the Santa Fe River Basin (Hunn and Slack, 1983). The Floridan aquifer covers an area of about 100,000 $mi^2$, and underlies all of Florida and parts of Georgia, Alabama, and the southern most part of South Carolina (Bush and Johnston, 1988).

Surficial sediments of Pleistocene and Recent Age, composed of white to gray fine sand approximately 10 feet thick, typically cover the bedrock in the Santa Fe Basin. Where present, the Miocene Hawthorn Group, composed primarily of siliciclastic rocks, acts as a confining unit above the Floridan Aquifer. The erosional edge of the Hawthorn Formation is known as the Cody Scarp and represents the physical division between the confined and unconfined Floridan Aquifer. To the northeast of the scarp, where the Hawthorn Formation is present, the Floridan is confined and surface water is abundant. Southwest of the scarp, where the Hawthorn Formation is eroded away, the Floridan is unconfined or semi-confined and there are few surface streams and numerous karst features such as sinkholes, springs, and disappearing streams. At the edge of the scarp,

streams either flow into sinkholes, as does the Santa Fe River, or become losing streams, discharging a portion of their flow to the ground directly from the streambed.

Except in parts of north Florida and southwest Georgia, the Floridan is divided into Upper and Lower aquifers by a less permeable layer of carbonate rocks belonging to the lower Avon Park Formation (Bush and Johnston, 1988). The Upper Floridan (Table 1-1) is composed of three highly permeable carbonate units: the Suwannee Limestone (Oligocene), Ocala Limestone (upper Eocene), and the upper part of the Avon Park Formation (middle Eocene) (Bush and Johnston, 1988).

The Ocala limestone is the uppermost unit in the unconfined portion of the Santa Fe River Basin, which includes the Santa Fe River Sink/Rise system. The thickness of the Ocala limestone is approximately 275 m near O'Leno State Park (Hisert, 1994). The Ocala is a white to yellow colored bioclastic limestone that is typically soft and friable (Skirvin, 1962). Common fossil fauna found in the Ocala include the orbitoid foraminiferan *Lepidocyclina*, and various echinoids, bryozoans and mollusks (Skirvin, 1962).

The Lower Floridan is composed of the lower part of the Avon Park Limestone (Eocene), the Oldsmar Formation (Eocene), and the Cedar Keys Formation (Paleocene). The Lower Floridan typically contains brackish or saline water, and largely remains undeveloped because the Upper Floridan is so productive.

Table 1-1. Geologic and hydrogeologic units of the Santa Fe River Basin. Hunn and Slack (1983), Scott (1992), and Hisert (1994).

| Series | Stratigraphic Unit | Hydrogeologic Unit | Lithologic Description | Thickness (m) |
|---|---|---|---|---|
| Holocene Pleistocene | Undifferentiated sediments | Surficial Aquifer | Sinkhole fill, fluvial terraces, and thin surficial sand | 0-24 |
| Pleistocene to Miocene | Alachua Formation | Intermediate Aquifer/Upper Confining Unit | Reddish-white sands with clays, sandy clays, and phosphate pebbles | 0-30 |
| Middle to Lower Miocene | Hawthorn Group | | Phosphatic clayey sand-sandy clay with varying amounts of Fullers Earth and carbonate | |
| Oligocene | Suwannee Limestone | Upper Floridan Aquifer | Very pale yellow, moderately indurated, porous, fossil-rich calcarenite | 0-100 |
| Eocene | Ocala Ls. Avon Park Ls. Oldsmar Ls. | | Very permeable white to yellow bioclastic limestone | 250-300 |
| | | Lower Floridan Aquifer | Dolomitic limestone & dolomite | 300-? |
| Late Paleocene | Cedar Keys Formation | | Limestone, some evaporites and clay | |

## Previous Investigations in the Santa Fe River Basin

The first scientific study of the Sink/Rise system was conducted by Skirvin (1962). He noted that there was a change in color between the dark brown tannic water upstream of the Sink and the clearer water discharging from the Rise during low river stage. He measured higher levels of silica, calcium, sulfate, and bicarbonate ($HCO_3$) in water discharging from the Rise indicating that groundwater was entering the conduit (Skirvin, 1962).

Hunn and Slack (1983) described the quantity and quality of surface and groundwater resources of the Santa Fe River Basin. They noted that the potentiometric

map shows groundwater flow toward the river, and therefore assumed that the conduit has a variable subsurface component of discharge between the River Sink and River Rise.

A detailed study by Hisert (1994) used $SF_6$ as a natural tracer to map the groundwater flow of the Santa Fe River through O'Leno State Park. He established that there was a connection between O'Leno Sink and Sweetwater Lake, and between Sweetwater Lake and the River Rise, and found an average flow rate of 1.0 to 3.4 km/day, confirming conduit flow between the River Sink and River Rise. He concluded from tracer studies that between the River Sink and Jim Sink there was one main conduit carrying water flow, and that after Jim Sink the conduit split into two or more main channels.

Kincaid (1998) used the natural tracers Radon-222 ($^{222}Rn$) and $\delta^{18}O$ to quantify the exchange of water in the Devil's Ear cave system located in the western Santa Fe River basin. He demonstrated that the exchange of water between matrix and conduit is not a direct function of river stage, but a result of head differences between the aquifer and the conduit.

Later studies by Dean (1999) and Martin and Dean (1999) demonstrated that temperature could be used as a high-resolution natural tracer. They confirmed flow rates through the conduit of the same magnitude as Hisert's (1994) and found that velocities increased with increasing river stage. Martin and Dean (2001) used changes in discharge between the River Sink and River Rise along with variations in the chemical composition, to quantify the proportions of surface water and groundwater discharging from the Rise. They found that as discharge at the Sink increased the proportion of

surface water in discharge at the Rise increased. Conversely, as discharge dropped, the fraction of groundwater discharging from the Rise increased.

Water levels and temperatures of the Sink, River Rise, and intermediate karst windows were collected during the year prior to this study to estimate water velocity through the subsurface between the Sink and the Rise. The estimated velocity of approximately 3000 m/d during a March 2002 storm event, confirmed conduit flow (Ginn, 2002). By treating the conduit as a closed pipe, Ginn calculated an average velocity of 0.012 m/s and an average conduit area of 375 $m^2$ during the March event. Assuming a circular conduit, the average diameter of the conduit would be 22 m. It should be noted that the previous calculations were for one rain event only, and that velocity changes proportionally with discharge.

Screaton et al. (in press) quantified the volume of water lost to the matrix and conduit in the Santa Fe River Sink/Rise system during the peak of three high flow events between August 2001 and August 2002 by comparing the simultaneous discharge rates at the River Sink and the River Rise. These data were used to calculate conduit area and develop a prediction for the relationship between discharge and velocity. At discharge rates above 14 $m^3/s$, Screaton et al. (in press) observed that calculated conduit water velocities from a previous study (Dean, 1999) are lower than predicted values. This suggests that the closed pipe flow model may not accurately describe flow through the conduit at high discharges (Screaton et al., in press).

Cave divers have explored and mapped the conduits between the Sink and Rise since 1995. They were able to identify connections between several of the downstream sinks and the Rise, and between several upstream sinks and the River Sink, but have not

yet found the direct physical connection between Sink and Rise. The Old Bellamy Cave Exploration Team has mapped more than 12.4 kilometers of submerged passageways including a large feeder conduit system entering the Santa Fe system from the east (Old Bellamy Cave Exploration Team, unpublished report, 2001). The team reported passage diameters as large as 45 meters that agree with Ginn's calculations, which predicted large conduits (Old Bellamy Cave Exploration Team, unpublished report, 2001).

## Current Investigations

There are currently two additional studies near completion on the Santa Fe River Sink/Rise system. Brooke Sprouse (UF masters student) is studying the exchange of water between matrix and conduit using natural tracers including $Sr^{2+}$, $^{87}Sr/^{86}Sr$, and $\delta^{18}O$. Lauren Smith (UF masters student) is studying the use of radon-222 as a natural tracer in groundwater.

# CHAPTER 2
# METHODS

## Data Collection

The objectives of this project, which were to determine matrix hydraulic conductivity and to describe the movement of water between conduit and matrix, were met by recording physical measurements of the conduit system, monitoring wells, and hydrologic system. Measurements included monthly precipitation, river stage, water temperature, and water levels. The Santa Fe River stage was recorded by the staff of O'Leno State Park and obtained through the Suwannee River Water Management District (SRWMD). Stage measurements were read from a staff gauge located approximately 0.5 km upstream from the River Sink. Monthly precipitation data, obtained from the Southeast Regional Climate Center, were collected from High Springs, Florida, located 10 km south of the River Sink. Precipitation data from nearby High Springs were used because records at the O'Leno State Park station were incomplete.

Seven monitoring wells were installed into the matrix at various distances from mapped conduits (Figure 1-2) during early 2003. Wells were constructed of 2-inch PVC and were approximately 100 ft deep each. Wells were located along the conduit length between the River Sink and River Rise. An existing well, Tower Well, located near the main entrance to O'Leno State Park, was also monitored to help determine how far from the conduit effects of large recharge pulses were occurring. Well data are located in Table 2-1.

Table 2-1. Monitoring well summary.

| | Completed Depth (ft) | Screened interval (ft) | Depth to bedrock (ft) |
|---|---|---|---|
| Well 1 | 75 | 75-55 | 56 |
| Well 2 | 100 | 100-80 | 20 |
| Well 3 | 93 | 93-73 | 10 |
| Well 4 | 97 | 97-77 | 15 |
| Well 5 | 98 | 98-78 | 18 |
| Well 6 | 102 | 102-82 | 16 |
| Well 7 | 98 | 98-78 | 18 |

**Temperature**

Temperature data were collected using Onset© Optic StowAway waterproof digital thermometers with an accuracy of $\pm 0.2^{o}$C, Van Essen - Diver© loggers with an accuracy of $\pm 0.1^{o}$C or Van Essen CTD Divers with an accuracy of $\pm 0.1^{o}$C. Readings were taken every 10 minutes. Data were downloaded from the loggers every four to five weeks. Loggers placed in monitoring wells were positioned in the center of the screened interval to ensure adequate circulation of ground water. Divers© placed at the Sink, Rise, and karst windows were located within 2-inch PVC stilling tubes, while Onset© loggers were lowered directly into the water and tethered by plastic coated steel wire. It was not necessary to calibrate temperature loggers since temperature maxima and minima were used to correlate between locations and not temperature magnitude. Data logger locations are listed in Table 2-2. Periods lacking data represent times when loggers malfunctioned or were not installed.

**Water Levels**

Water levels were taken at the Sink, Rise, karst windows and monitoring wells by one of three automatic water level recorders (Global Water WL14 Water Level Logger with an accuracy of +/- 0.01m, Van Essen – Diver with an accuracy of +/- 0.005m, or Van Essen CTD Divers with an accuracy of +/-0.03m) or measured using an electronic probe. Loggers located at the Sink, Rise, and karst windows were installed within 2-inch PVC pipe stilling wells. Loggers at monitoring wells were attached to the well cap by a plastic-coated stainless steel wire. Water levels were recorded at 10-minute intervals. Data were downloaded from the loggers every four to five weeks. For each recording interval, water pressures from the loggers were corrected using the ambient barometric pressure (if necessary) recorded by a Baro Diver© (±0.0045m) and then referenced to the water elevation surveyed at the time the data were downloaded or measured in the wells. Original elevations at the Sink, Rise, and karst windows were surveyed by Jonathan B. Martin and Lauren Smith in 2001 using a Sokkia Automatic Level Model B21. Original survey points were marked with a nail in a tree and monthly reference water level measurements were measured in reference to the known elevation. Wells 1, 2, 3, 4, 6, and 7 were surveyed by Britt Surveying of Lake City, Florida. Well 5 was surveyed by Elizabeth J. Screaton and Jennifer M. Martin with a Sokkia Automatic Level Model B21 using the elevation at Well 4 as a benchmark. Inaccuracies caused by survey errors, movement of the logger or instrument drift, and logger accuracy were estimated to be less than ± 0.07 m for the River Sink and Sweetwater, ± 0.08 m for the River Rise, and ± 0.05 for other karst window sites. Water level discrepancies at the karst window sites were examined by comparing measurements between recording intervals at each site and were typically less than ±0.03 m. These discrepancies are included in the total estimated

errors. The total estimates are upper bounds because the observed discrepancies are likely to overlap with the instrumental error. Water level errors at the monitoring wells are expected to be lower than at the karst windows because pressure transducer movement, a suspected source of error at the surface water sites, is likely to be minimal within the monitoring wells. Summing of survey error, water level reading error and instrument error suggests total errors at the monitoring wells of ± 0.02 m for manual readings and ± 0.03 m for automated readings.

### Specific Conductivity

Specific conductivity was collected from the Sink, Sweetwater Lake, and the Rise using Van Essen CTD Divers © with an accuracy of 50 μS/cm. Loggers were placed inside PVC stilling tubes at each location.

## Data Analysis

### Sink and Rise Discharge

Discharge rates of the River Sink were calculated by converting water levels to discharge using a rating curve (Figure 2-1) developed by the Suwannee River Water Management District (Rating No. 3 for Station Number 02321898, Santa Fe River at O'Leno State Park). Rise discharge rates were calculated by Screaton et al. (in press) by creating a rating curve based on the relationship between water level elevations and unpublished discharge data from SRWMD (Figure 2-2). The curve was constructed by plotting recorded discharge measurements for a variety of water levels. Using the best-fit curve of the data points, it was possible to infer discharges for all water levels within the range of measured values.

Table 2-2. Locations and dates of data collection. O = Onset logger (temperature), V = Van Essen Diver, B=Van Essen Barometric Diver, C=Van Essen CTD Diver, Periods of no data represent time when loggers were either malfunctioning or not installed.

| | 4/12/02-5/14/02 | 5/14/02-6/18/02 | 6/18/02-7/8/02 | 7/8/02-8/8/02 | 8/8/02-9/12/02 | 9/12/02-10/17/02 | 10/17/02-11/14/02 | 11/14/02-12/13/02 | 12/13/02-1/22/03 | 1/22/03-2/26/03 | 2/26/03-3/27/03 | 3/27/03-5/14/03 | 5/14/03-7/24/03 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Black | O | O | O | O | | O | O | O | O | O | O | O | |
| Rise | C | C | C | C | C | C | C | C | C | C | C | | C |
| Sweetwater | C | C | C | C | C | C | C | C | C | C | C | C | C |
| Two Hole | | | V | V | V | V | V | V | V | V | V | V | V |
| Hawg | V | V | V | V | V | V | V | V | | | | | |
| Paraners | OG | OG | OG | OG | OG | OG | OG | OG | O | O | O | O | O |
| Ogden | V | V | V | V | | | | | | | | | O |
| Jim | | | | | | | | | | | | | O |
| Jug | OG | OG | OG | OG | OG | OG | OG | OG | OG | O | O | | O |
| Big | VB | VB | V | V | | B | B | B | B | B | B | | |
| Sink | C | C | C | C | C | C | C | C | C | C | C | C | C |
| Tower | | V | V | V | V | V | V | V | V | V | V | | |
| Well #1 | | | | | | | | | | G | G | | OG |
| Well #2 | | | | | | | | | | V | V | V | OV |
| Well #3 | | | | | | | | | | | O | O | O |
| Well #4 | | | | | | | | | | V | V | V | OV |
| Well #5 | | | | | | | | | | | | B | OB |
| Well #6 | | | | | | | | | | G | G | OG | O |
| Well #7 | | | | | | | | | | G | | | |

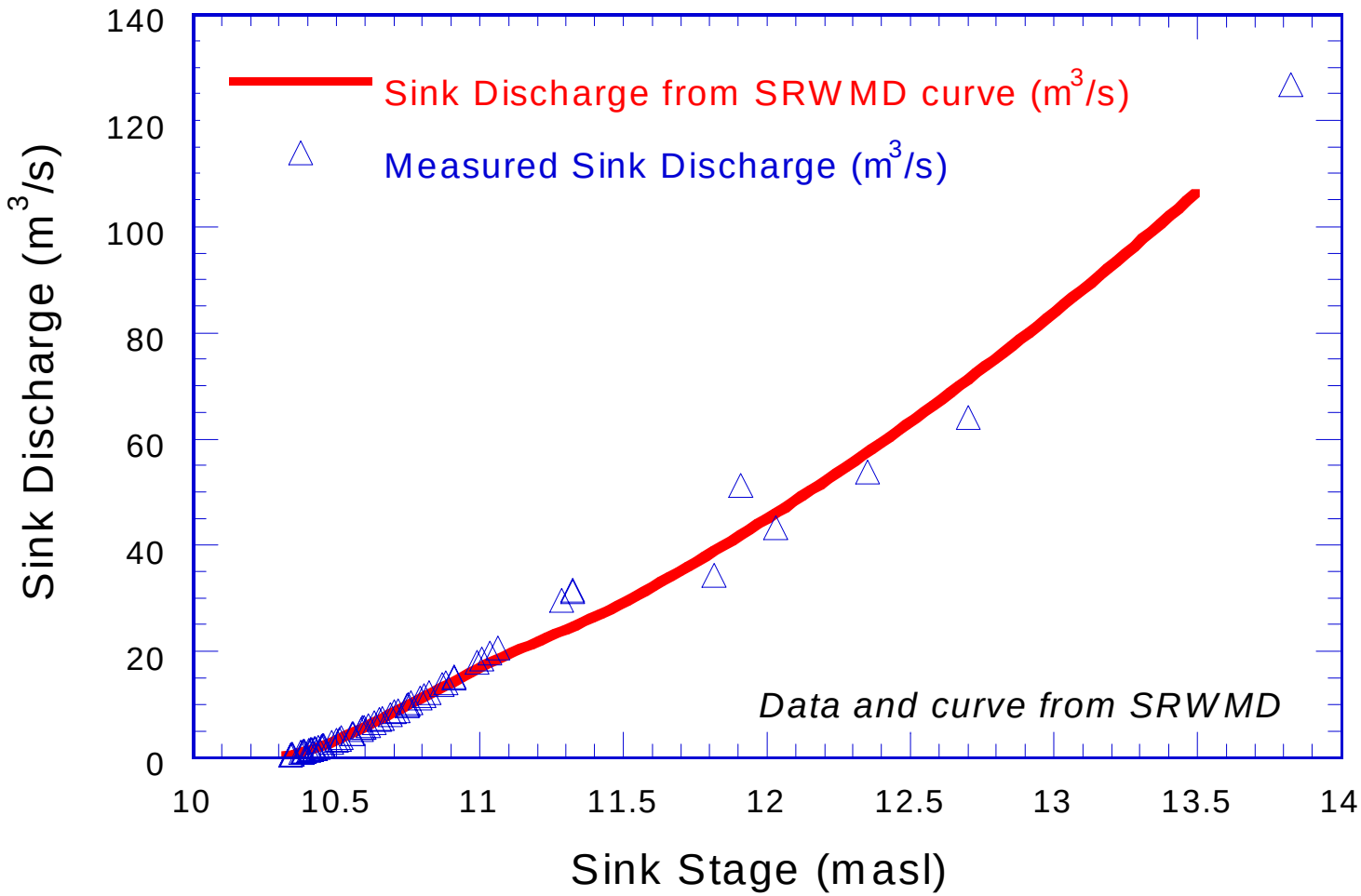


Figure 2-1. River Sink rating curve produced by the Suwannee River Water Management District.

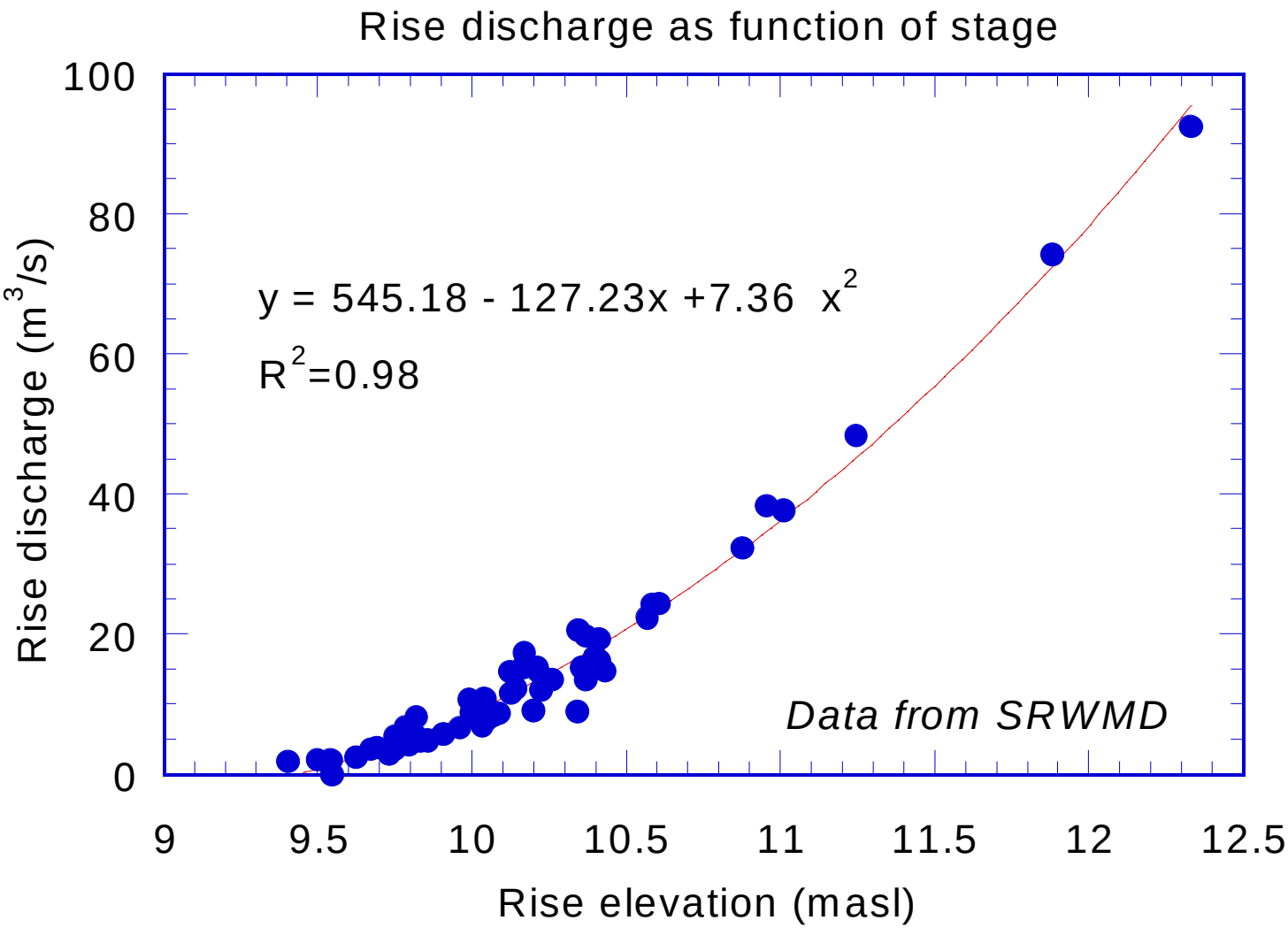


Figure 2-2. River Rise rating curve (Screaton et al., in press).

**Conduit Water Velocity**

Water velocity through the conduit was determined by using temperature as a natural tracer between the Sink and Rise. Temperature data revealed several maxima or minima that could be correlated among the Sink, karst windows, and Rise (Figure 3-8). Temperatures are assumed to remain consistent between the River Sink and River Rise due to sufficiently high flow velocities during correlated events. Benderitter et al. (1993) documented temperature variations in conduits from Guichy, France and determined that the temperature of recharge pulses remain relatively consistent during high velocity flow, but may be slightly delayed during very low velocity flow due to thermal exchange with surrounding rock.

The travel time from the Sink temperature maximum to corresponding Rise maximum divided by the total estimated conduit length (8000 m) equals the average water velocity as it flows through the conduit. Distances between the Sink, karst windows, and Rise were originally estimated by Hisert (1994) based on straight-line distances between locations. Portions of conduit surveyed by cave divers were digitized then overlaid onto digital topographic maps of the area using ArcView GIS v3.2 (Fig. 1-2). By utilizing this new information, more accurate estimations of conduit length and distance between locations were obtained, which allowed a refinement of the velocity calculation.

## Conduit Properties

Conduits located in the Floridan Aquifer can be visualized as leaky pipes transporting water in the subsurface. In order to understand the interaction of the conduit with the hydrologic system, physical properties of the conduit were determined using pipe flow equations. The conduits were assumed to be flowing under “pipe-full”

conditions due to their depth below the water table. Simplifying the system by treating the conduit as a closed circular pipe flowing under pipe-full conditions allowed the application of fluid mechanics equations for pipe flow.

**Reynolds Number**

The Reynolds number relates several factors that determine whether flow will be laminar or turbulent (Fetter, 2001). Velocities calculated from the time lag data were used to calculate the Reynolds number for each of the correlated temperature peaks. Density and viscosity values were calculated for an average groundwater temperature of $20^{o}C$ (Fetter, 2001).

$$Re = \frac{\delta v d}{\mu}$$

Re=Reynolds number, dimensionless
$\delta$=density of water ($kg/m^3$)
v=velocity (m/s)
d=diameter of the conduit (m)
$\mu$=viscosity of water (kg/s-m)

**Darcy-Weisbach Equation**

The Darcy-Weisbach equation is frequently used to determine head loss in pipes, but has been used in the study of karst conduits (Atkinson, 1977; Gale, 1984). Given that the head loss between the Sink and Rise is already known from direct water level measurements, the equation was used to calculate the friction factor (f), also called the resistance coefficient of the conduit. The friction factor value is an indicator of friction losses, most of which occur at a few isolated constrictions or collapses within the conduit system (Wilson, 2001).

$$h_l = f\left(\frac{LV^2}{d2g}\right)$$

$h_l$=head loss (m)
f=friction factor, dimensionless
L=length of conduit (m)
V= average flow velocity (m/s)
d=diameter of the conduit (m)
g=acceleration due to gravity ($m/s^2$)

**Absolute Roughness**

In fluid mechanics, the Colebrook and White (1937) equation is used for determining the necessary pipe size to deliver a specified flow rate under given conditions. By using the friction factor (f) from the Darcy-Weisbach equation, the diameter of the conduit, and the Reynolds number, the absolute roughness of the conduit (e) can be calculated.

$$\frac{1}{\sqrt{f}} = -2\log\left(\frac{e}{3.7d} + \frac{2.51}{Re\sqrt{f}}\right)$$

f=friction factor, dimensionless
e=absolute roughness of the conduit (m)
d=diameter of the conduit (m)
Re=Reynolds number, dimensionless

## Matrix Transmissivity

Accurate estimations of transmissivity are necessary to predict aquifer response to various hydrologic stresses (Pinder et al., 1969). Transmissivity is difficult to determine in karst aquifers due to their heterogeneous nature. Three different methods were utilized to quantify the matrix transmissivity in the Santa Fe River Basin near O'Leno State Park.

**Stage Ratio Method**

The Stage Ratio Method for calculating transmissivity (Ferris, 1963) relates the ground water fluctuation at a well in response to changes in river (conduit) stage. Estimates of porosity for the Floridan Aquifer range from 0.1-0.45 (Palmer, 2002). A value of 0.20 was chosen as a reasonable estimate of storativity for all three methods because storativity in an unconfined aquifer is controlled by specific yield, and specific yield cannot exceed porosity.

$$T = \frac{x^2 \pi S}{-\left[\ln\left(\frac{S_r}{2S_o}\right)\right]^2 t_o}$$

T=transmissivity ($m^2/s$)
x=distance from well to conduit (m)
S=storativity (assumed to be 0.20)
$S_r$=amplitude of the fluctuation at the well
$S_o$=amplitude of the fluctuation at the river
$t_o$=period of the fluctuation

**Time Lag Method**

The Time Lag Method (Ferris, 1963) relates transmissivity to groundwater stage maxima or minima at a well and the timing of corresponding stage maxima or minima in a conduit.

$$T = \frac{x^2 S t_o}{4\pi t_l^2}$$

T=transmissivity ($m^2/s$)
x=distance from the well to the conduit (m)
S=storativity (assumed to be 0.2)
$t_o$=period of the fluctuation
$t_l$=time lag (s)

Both of these methods are typically used for tidal fluxes, but may be used for events with a single maximum or minimum (Ferris, 1963). Six assumptions or simplifications are made when applying the Stage Ratio and Time Lag Methods; (1) the aquifer is homogenous and of uniform thickness, (2) there is an immediate release of water from the aquifer with a drop in pressure, (3) the observation well is located at a sufficient distance from the conduit that the effect of vertical flow can be ignored, (4) the fluctuation at the well is a small percentage of the saturated thickness of the aquifer, (5) the water level fluctuation is sinusoidal, and (6) the conduit fully penetrates the entire thickness of the aquifer (Ferris, 1963).

It should be noted that not all of these assumptions are met in this analysis. There may be affects of vertical flow and partial penetration, especially for wells close to the conduit. However, the methods used in this study provide a first approximation of the hydrologic properties of the system. A much more in depth analysis, most likely using numerical modeling, would be necessary to address these limitations.

**Pinder et al. (1969) Method**

Pinder et al. (1969) proposed a method that does not assume a sinusoidal groundwater fluctuation curve, as do the previous two methods. Except for not assuming a sinusoidal curve, all other simplifications and limitations listed for the Ferris (1963) methods apply. This method allows a flood stage hydrograph of any shape to be used. Because the Floridan Aquifer System in the Santa Fe River Basin is not known to be bounded by impermeable materials, the Pinder et al. (1969) equation for a semi-infinite aquifer was used. The input signal, which is the conduit water level, is broken into increments, and then the incremental change in head is calculated.

$$\Delta h_m = \Delta H_m \operatorname{erfc}\left(\frac{x}{2\sqrt{vt}}\right)$$

$\Delta h_m$=change in head of well per time step (m)
$\Delta H_m$= change in head of conduit per time step (m)
x=distance from well to conduit (m)
v=diffusivity (T/S) ($m^2$/d) (S assumed to be 0.2)
t=time step (days)

The total change in head is calculated by summing the increments. Observed changes in head at a well were compared to theoretical calculated changes in head. Transmissivity was adjusted until the observed curve best matched the calculated curve. It was assumed that the magnitude of the March event was large enough to disregard any differences in antecedent head conditions based on two observations. There was less than 0.01 m/d of head decrease at each location before the event, which is very small when compared to the head changes (up to 0.63 m/day) occurring during the event.

# CHAPTER 3
# RESULTS

## Precipitation, Potential Evapotranspiration, and Recharge

Average precipitation in the Santa Fe River Basin is 137 cm/year with most precipitation occurring June through September (Hunn and Slack, 1983). Although not used for estimation of recharge due to incompleteness, daily precipitation records from O'Leno State Park for the hydrologic year June 2002 through May 2003 are shown in Figure 3-1. Due to a moderate El Niño year, there was higher than normal precipitation during the winter/spring of 2003. Increased precipitation led to flooding in the region after a series of storms produced >10 cm of rain between March 1 and 9, 2003. Annual precipitation (June 2002 through May 2003) at High Springs, FL, located ten kilometers south of the River Sink, was 172.0 cm (Southeast Regional Climate Center), 35 cm above the annual average of 137 cm.

Potential evapotranspiration (PET) was calculated using the Thornthwaite method (Thornthwaite and Mather, 1957) to estimate the annual amount of water that could be lost to the atmosphere.

$$PE_m = 16\ N_m\ [10\ T_m / I]^a \text{ mm}$$

$PE_m$ = monthly potential evapotranspiration
$N_m$ = monthly adjustment factor related to hours of sunlight
$T_m$ = the mean monthly temperature in degrees C
I = heat index for the year given by:
$I = S\ i_m = S\ [T_m/5]^{1.5}$ for each month (m = 1, 2, 3, … 12)
$a = (6.7e\text{-}7)\ (I^3) - (7.7e\text{-}5)\ (I^2) + (1.8e\text{-}2)\ (I) + 0.49$

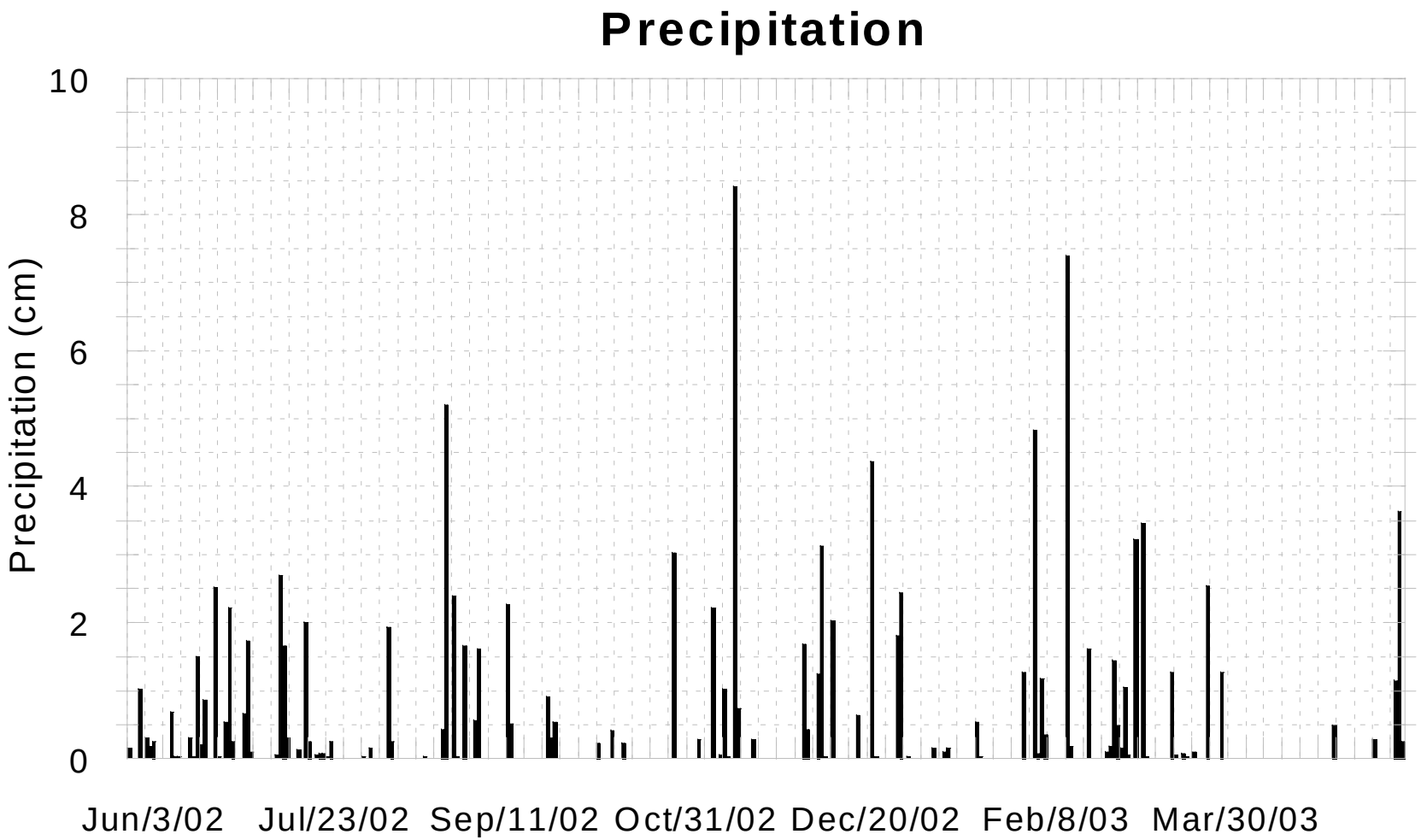


Figure 3-1. Precipitation records from O'Leno State Park.

This method uses the mean monthly air temperature, latitude, and mean daily duration of sunshine hours to calculate PET. Mean monthly air temperature values were recorded at the High Springs, FL station and obtained from the Southeast Regional Weather Center web page. The monthly adjustment factor related to hours of sunlight is from a USDA chart in Watson and Burnett (1995). This method assumes that the only effects on evapotranspiration are meteorological conditions and ignores the density of vegetation. Despite simplifications, this method gives a reasonable approximation of PET, and is especially suited for humid regions such as Florida (Watson and Burnett, 1995). Calculated average PET near O'Leno State Park is 105 cm/yr. This value agrees with Thornthwaite's (1948) average annual estimate of 105-115 cm for this region. Other calculations of PET in north-central Florida include Gordon (1998) who calculated PET of 107 cm/yr for June 1996 through May 1997 in the Ichetucknee River basin, and Jacobs (2001) who reported a PET value of 111 cm/yr for Gainesville, FL, located approximately 40 km south of O'Leno State Park. Figure 3-2 shows the relationship between precipitation and potential evapotranspiration in the study area.

Diffuse recharge was estimated based on the difference between total annual precipitation and calculated annual potential evapotranspiration. The dry season, which typically spans from November to May, was unusually wet in 2002 and 2003 due to El Niño. Monthly estimates for recharge are shown in Figure 3-3. In the unconfined portions of the basin where water can percolate directly into the Floridan Aquifer, average recharge is estimated at approximately 46 cm/year (Hunn and Slack, 1984). Diffuse recharge in the Santa Fe River Basin from May 2002 through June 2003 was estimated at 67 cm/yr.

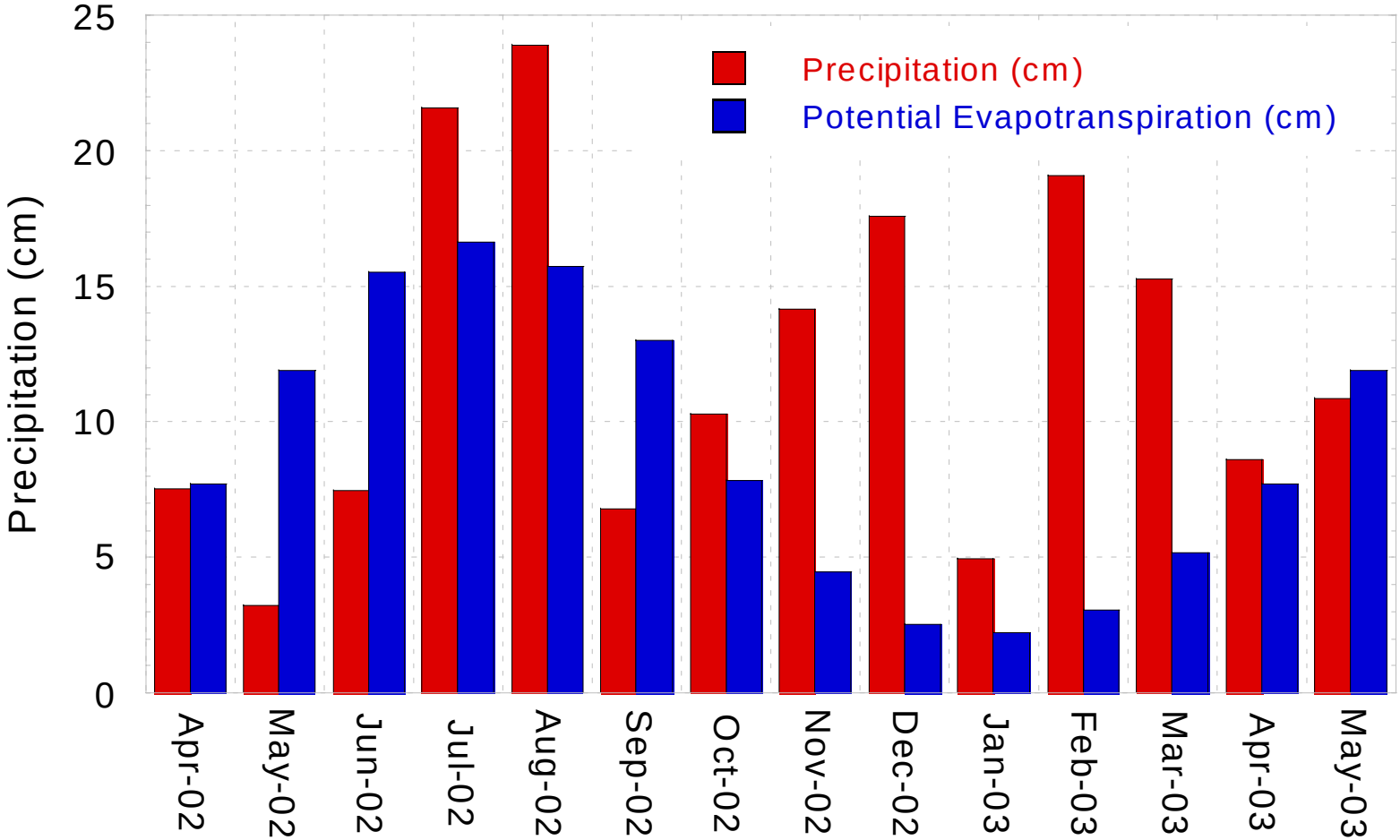


Figure 3-2. Precipitation compared to potential evapotranspiration at O'Leno State Park.

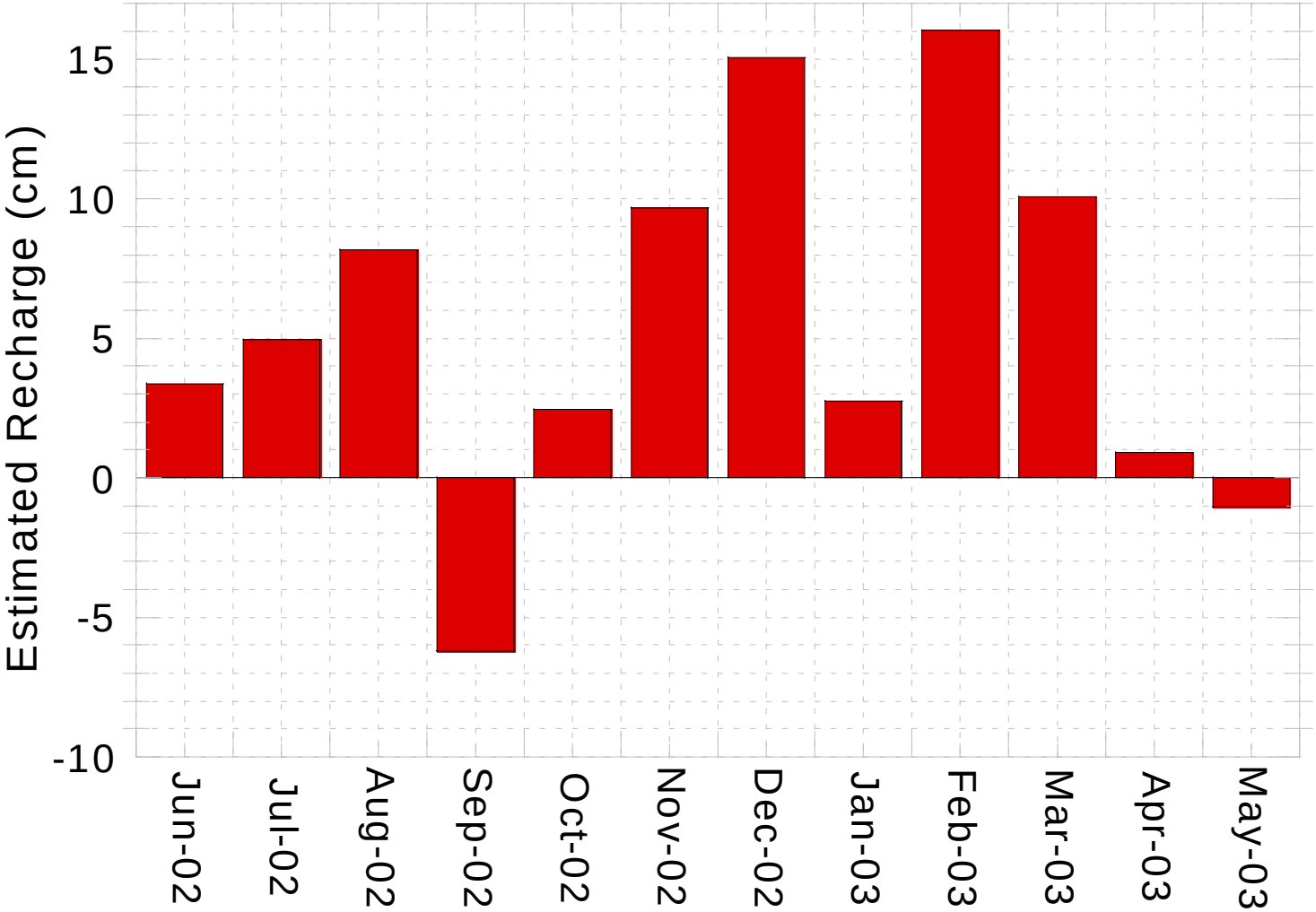


Figure 3-3. Estimated recharge in O'Leno State Park.

## Water Levels

### Conduit

Conduit water levels were measured at the Sink, intermediate karst windows, and the Rise (Figure 3-4). Differences between the Sink water level and the corresponding Rise water level signify head change along the length of the conduit. As discharge increases the gradient between the Sink and Rise increases.

### Monitoring Wells

Water levels in monitoring wells were measured between January and March 2003 (Figure 3-5), with the exception of Tower Well, which was measured between May 2002 and March 2003. During the monitoring period, there was a large discharge/flood event in the conduits with which to compare matrix water levels. Water level fluctuations in the matrix are too large to be explained by diffuse recharge alone. For example, the water level at Well 4 fluctuated approximately 2 meters between 2/8/03 and 3/2/03. However, during the same period, total rainfall was only 19 cm. Additional water from the conduits must be contributing to the matrix to account for the rise in head at the wells. For comparison, the River Sink (conduit) water level is shown in Figure 3-5. The conduit water level fluctuates much more rapidly than the matrix water levels except at Well 1. There is a noticeable time lag between water level maxima in the conduit and water level maxima in the matrix. The time lag is a reflection of the matrix transmissivity and storage between the conduit and each well.

## Sink Stage

The Santa Fe River stage was recorded at O'Leno State Park approximately 0.5 km upstream from the Sink from June 2002 to May 2003 (Figure 3-6). During this period, the stage ranged from 9.48 masl on 31 July 2002 to a maximum of 14.3 masl on 13

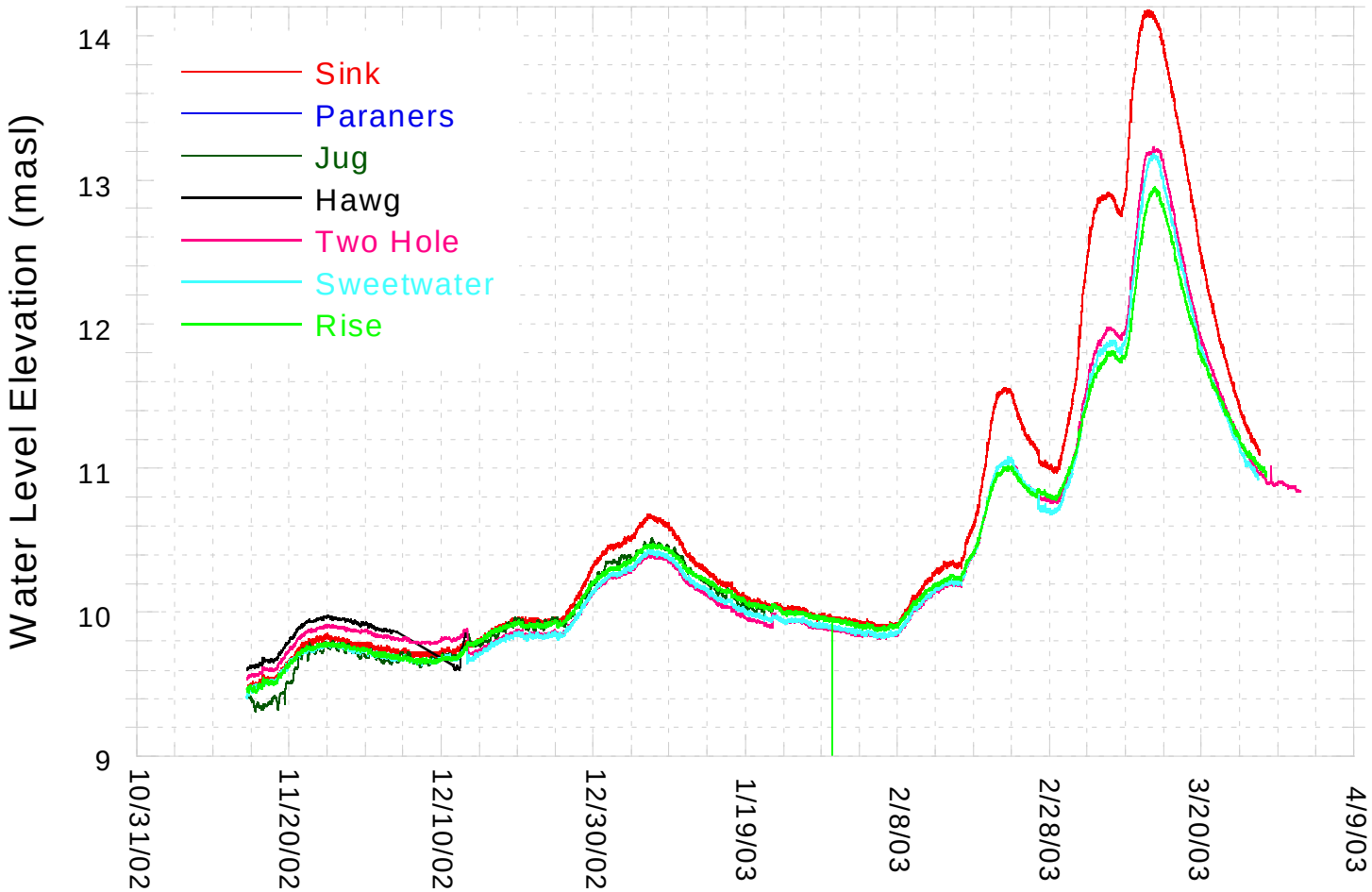


Figure 3-4. Conduit water level elevations.

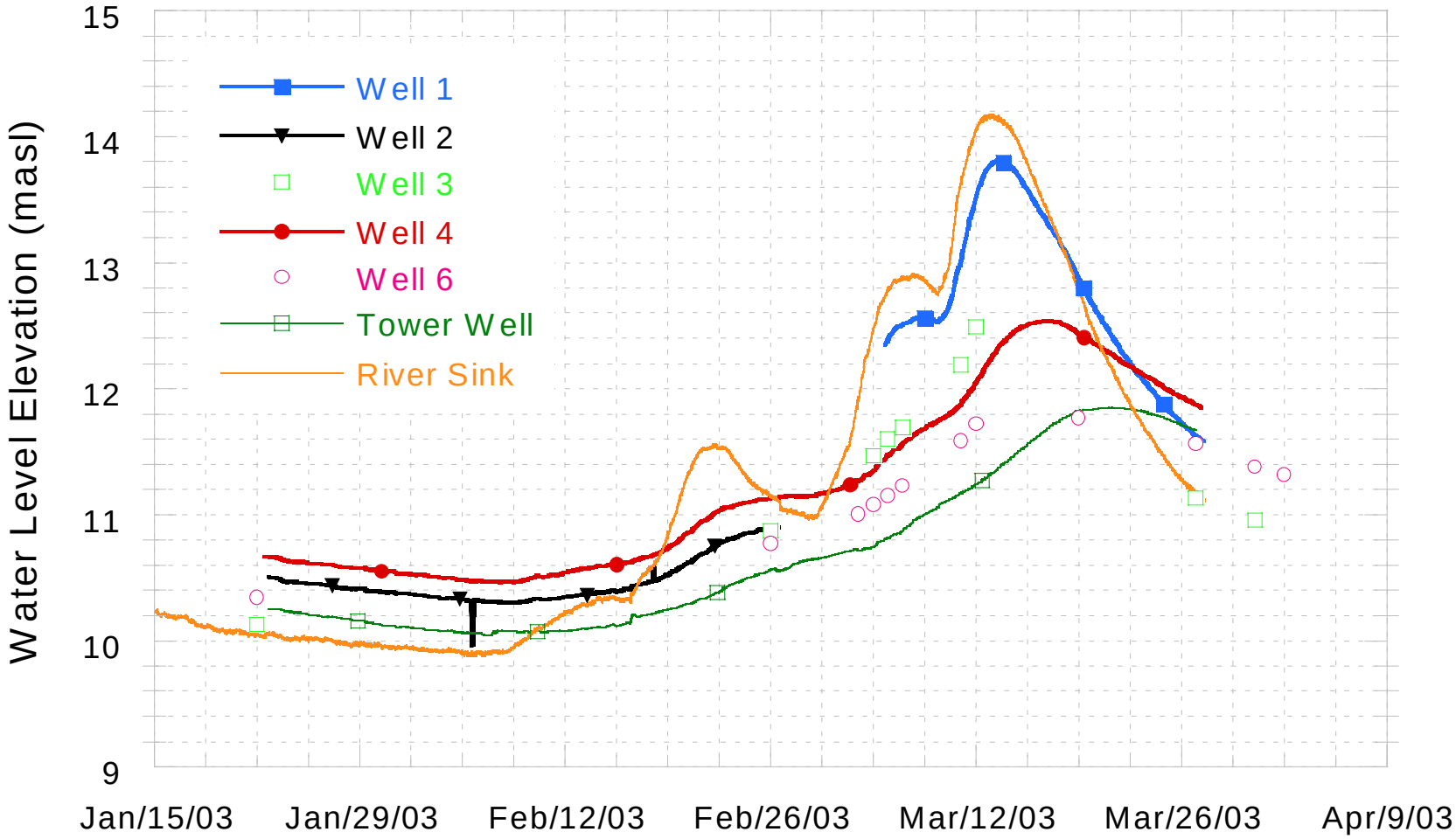


Figure 3-5. Monitoring well water levels with River Sink water level included for comparison.

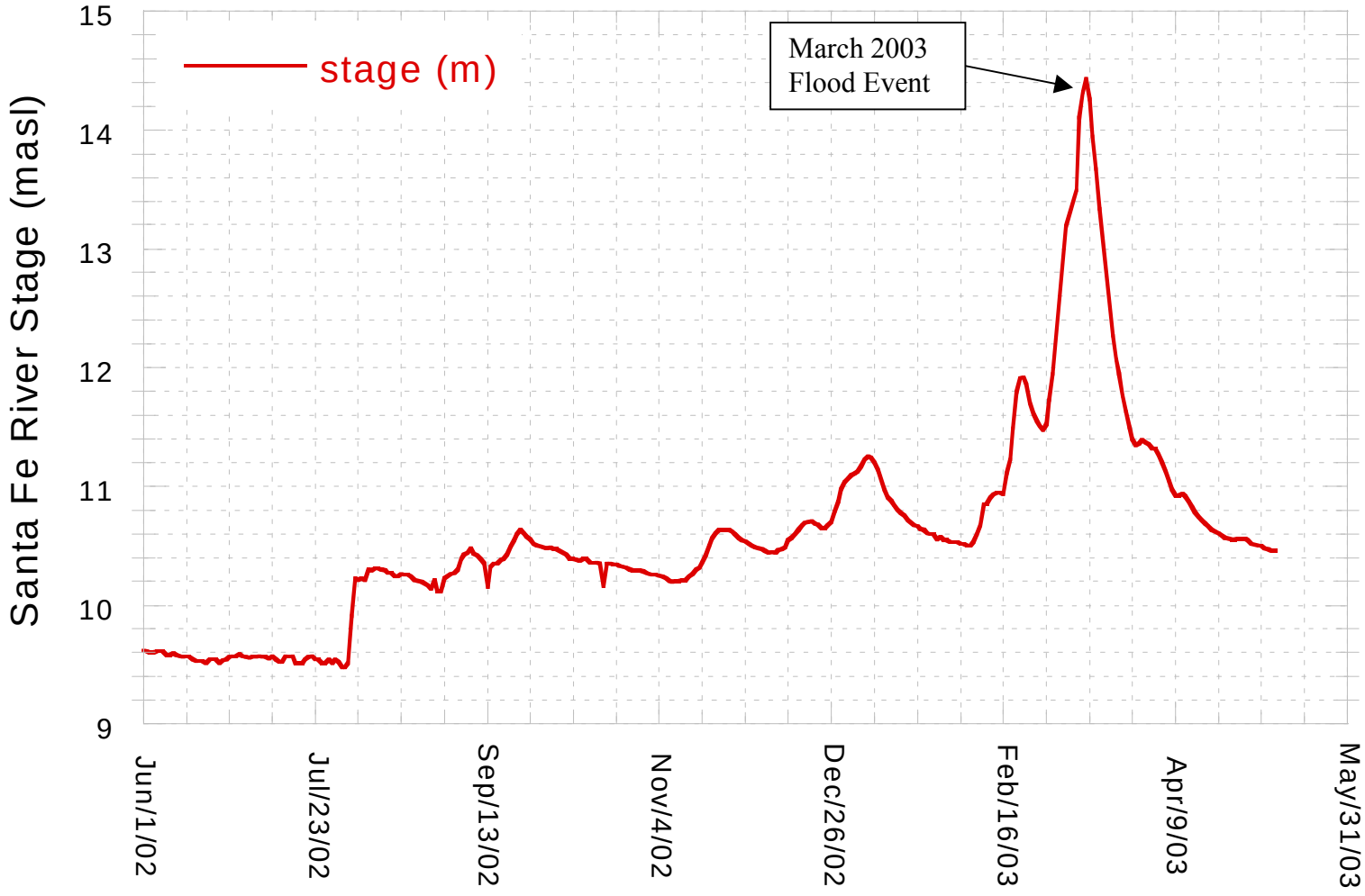


Figure 3-6. Two year stage records for the Santa Fe River at O'Leno State Park.

March 2003. Dean (1999) noted during a January 1999 flood event that when the Santa Fe River stage reaches approximately 14.3 masl it overflows its banks and the Sink, karst windows, and Rise become connected by overland flow. The March 2003 flood event reached Dean's (1999) minimum estimated threshold for overland flow. It caused partial overland flow in limited areas and significant flooding in large portions of the park.

## Sink and Rise Discharge

Plotting the River Sink discharge versus the River Rise discharge (Figure 3-7) illustrates that during high river stage the discharge into the Sink frequently exceeds the discharge out of the Rise. If more water is entering at the Sink than is exiting at the Rise, there is a quantifiable volume of water lost from the conduit system. Conversely, during times of low flow, the Rise discharge often exceeds the Sink discharge. At these times, the conduit is gaining water along its underground flow path.

## Water Temperature

### Conduit

Conduit water temperature was recorded between November 2002 and March 2003 (Figure 3-8). Nine temperature maxima or minima were correlatable between the Sink, five karst windows (sinkholes), and the Rise between 12/30/02 and 3/27/03. Temperature records at three sinkholes, Hawg, Two Hole, and Jug (Figure 1-2), between 11/20/02 and 1/22/03 are inconsistent with water temperature at other locations during the same period. This is most likely the result of insufficient logger depth during periods of low water, resulting in temperature records from non-circulating water. In Black Lake (Figure 1-2), temperature maxima and minima occur at the same time as maxima and minima from the Sink, indicating that changes in water temperature are caused by precipitation and changes in ambient air temperature, not by a connection with the conduit system.

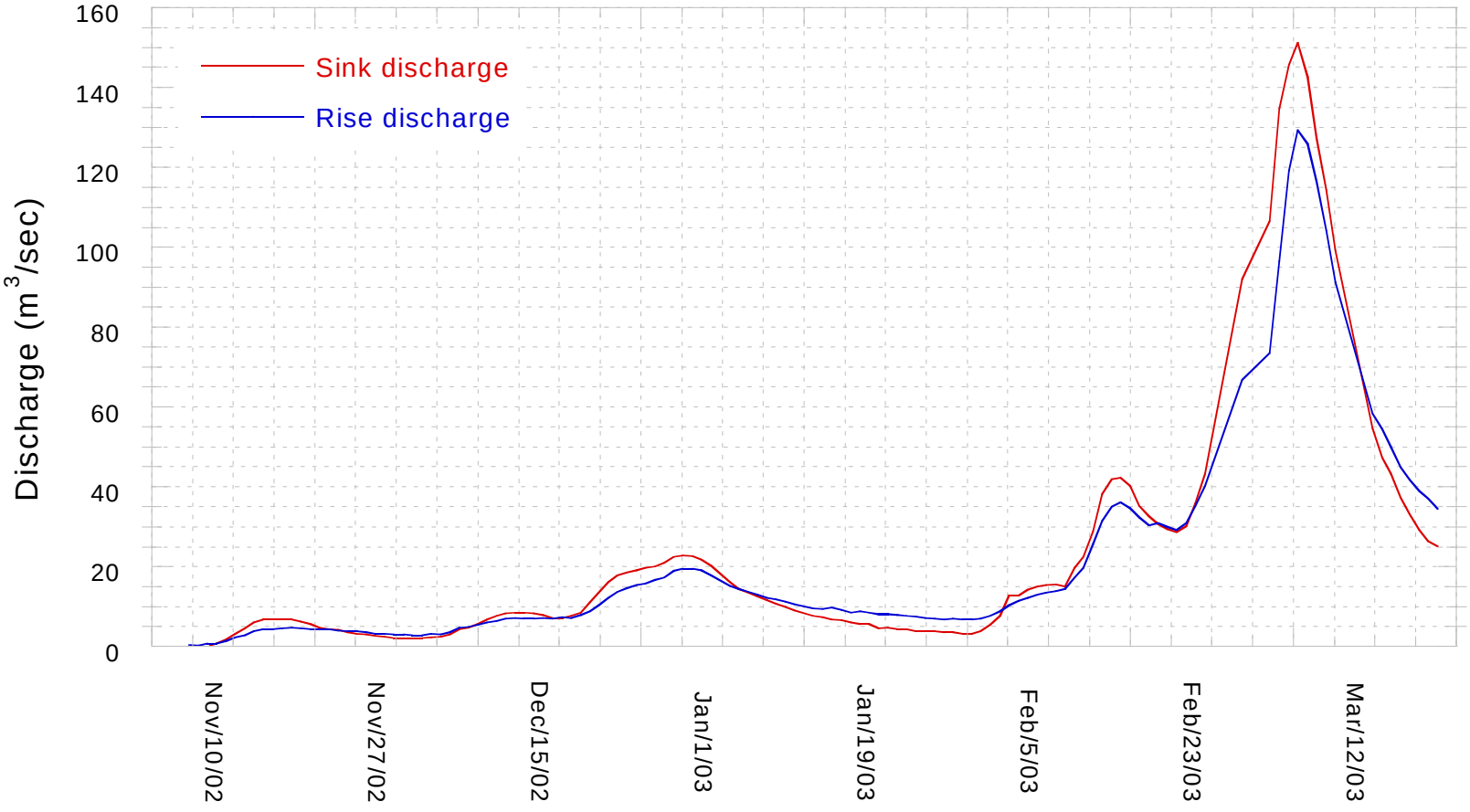


Figure 3-7. River Sink and River Rise discharge comparison.

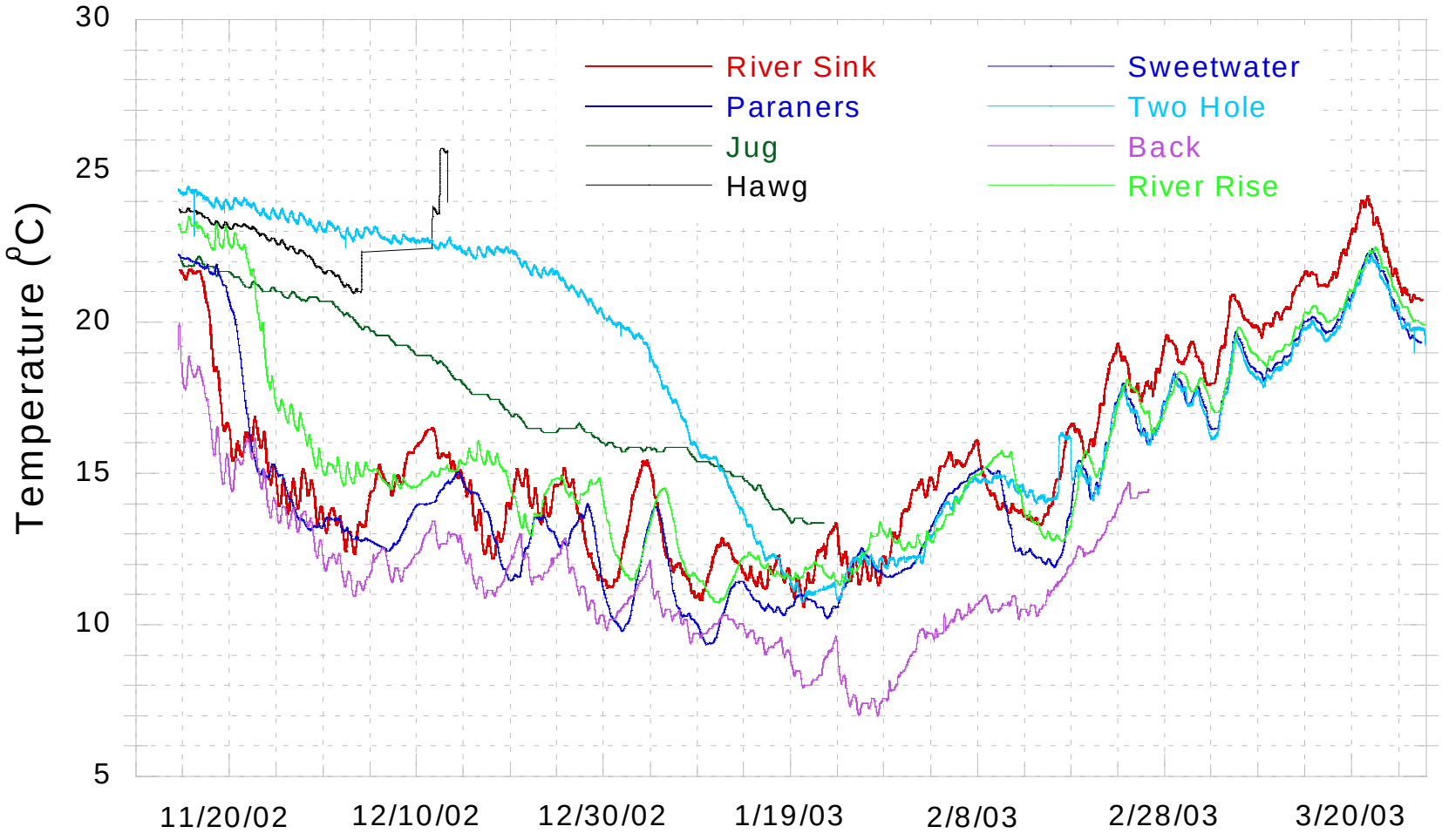


Figure 3-8. Water temperature records from the River Sink, intermediate karst windows, and River Rise.

Chemical analyses of water from Black Lake indicate it is not connected to the Sink/Rise system, but appears to be a perched lake (B. Sprouse, personal communication).

### Monitoring Wells

Temperature records from monitoring wells remained relatively constant (Figure 3-9) during the sampling period. Rapidly fluctuating temperatures would be an indication that surface water was reaching the well by preferential flow paths, fractures, or conduits. Temperature stability is an indication that the monitoring wells are receiving groundwater from diffuse flow, and are therefore representative of the matrix.

## Specific Conductivity

Specific conductivity was measured in the Sink, Sweetwater, and the Rise (Figure 3-10). Conductivity records were erratic at the Sink and Rise. Groundwater in the area typically has a maximum specific conductivity of approximately 0.5 mS/cm, whereas surface water tends to be lower. Several readings from the Sink, Sweetwater, and the Rise were consistently higher than 0.5 mS/cm suggesting measurement problems. The loggers were cleaned and tested with standard solutions and did not need recalibration. Stagnant water or accumulating sediment near the logger sensor could account for the high readings. Due to difficulties with measurement of specific conductivity, correlations between the River Sink, Sweetwater, and River Rise could not be made for this study.

## Conduit Properties

### Conduit Water Velocity

Conduit water velocity increases with river stage (Fig 3-11 and Table 3-1). Velocities were calculated between each location along the conduit length and then plotted. Velocity remains relatively linear with increasing distance from the River Sink. Distances between karst windows located closer to the River Sink are not well

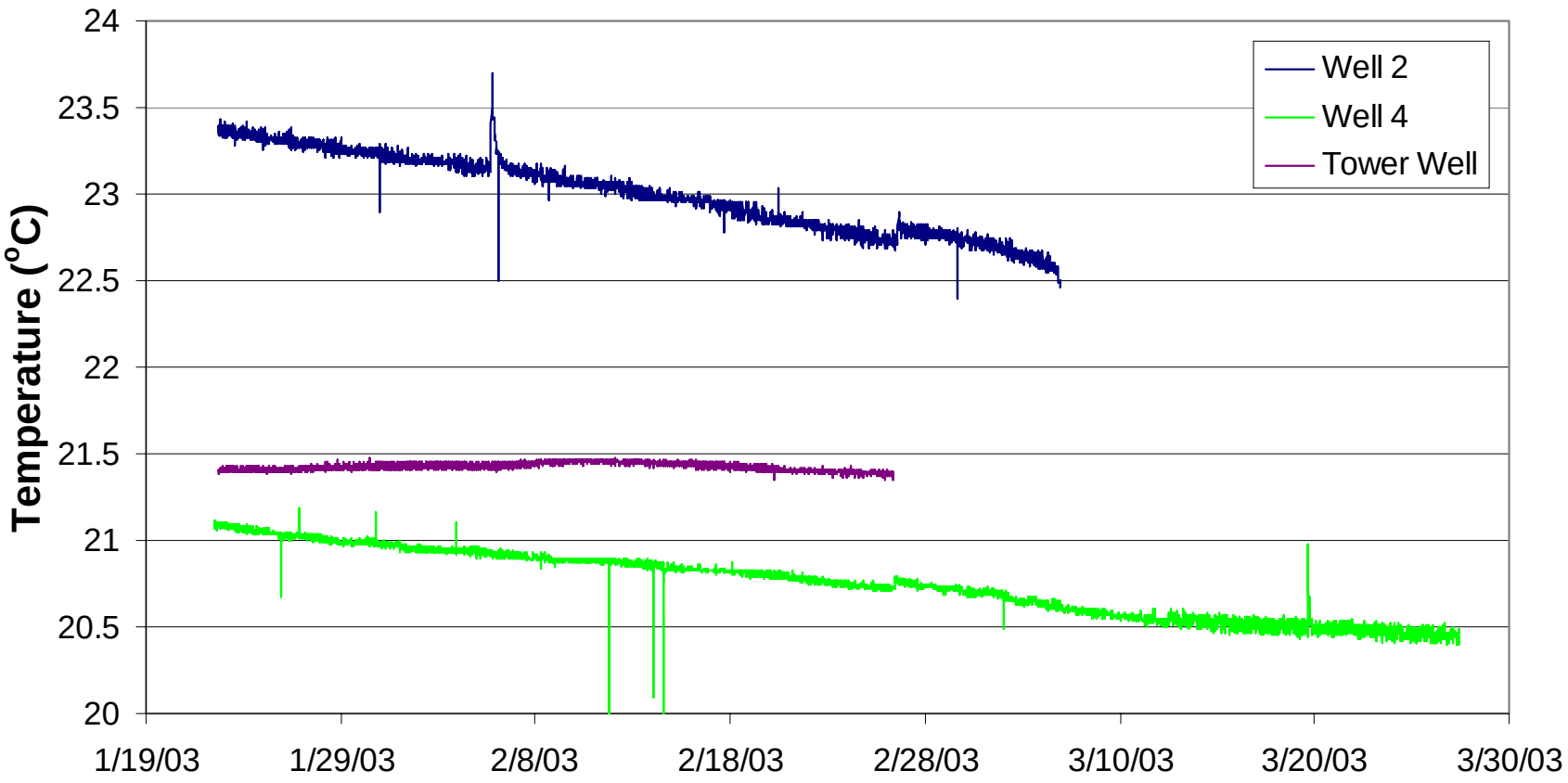


Figure 3-9. Monitoring well water temperatures.

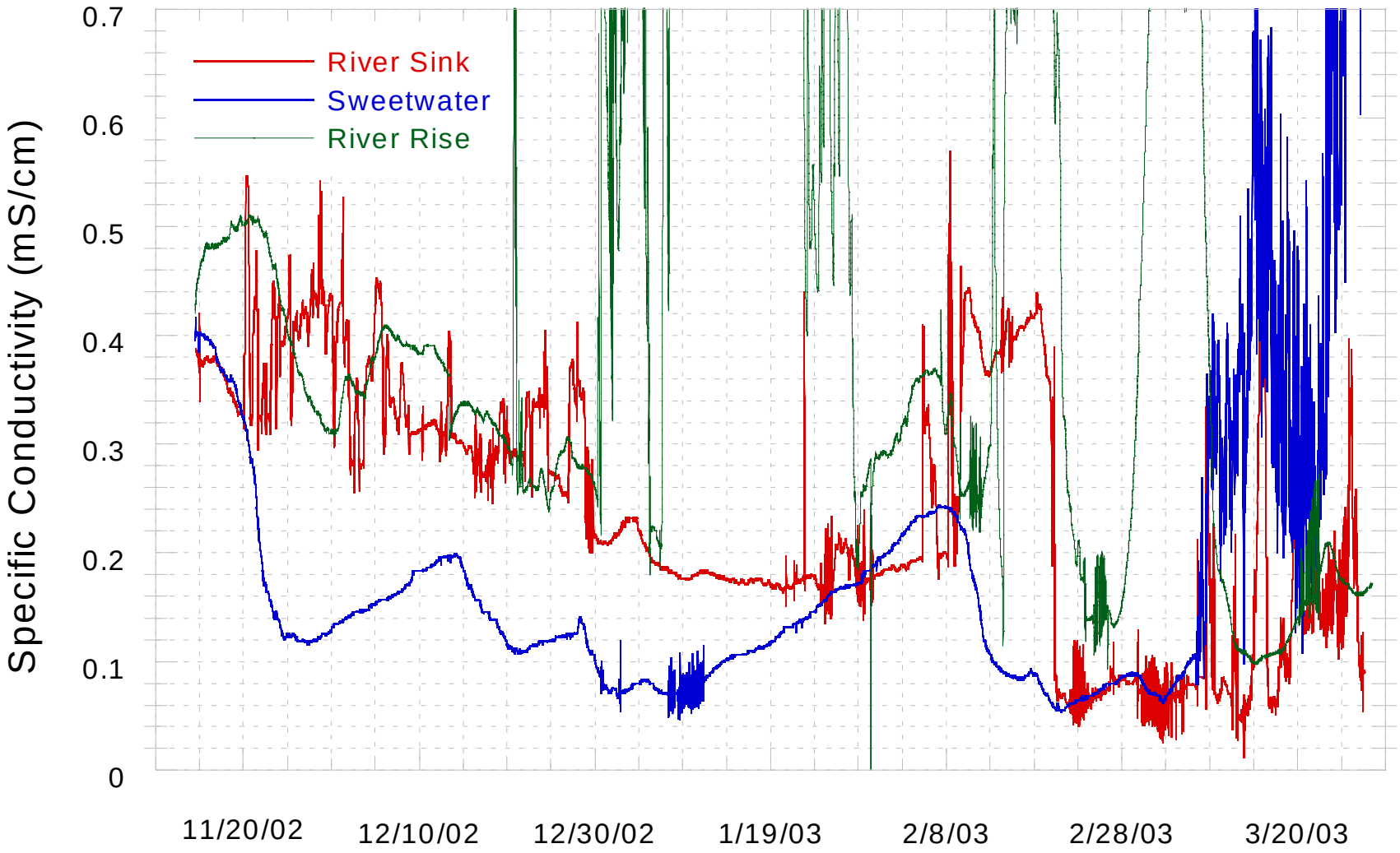


Figure 3-10. Specific conductivity measurements from the conduit system.

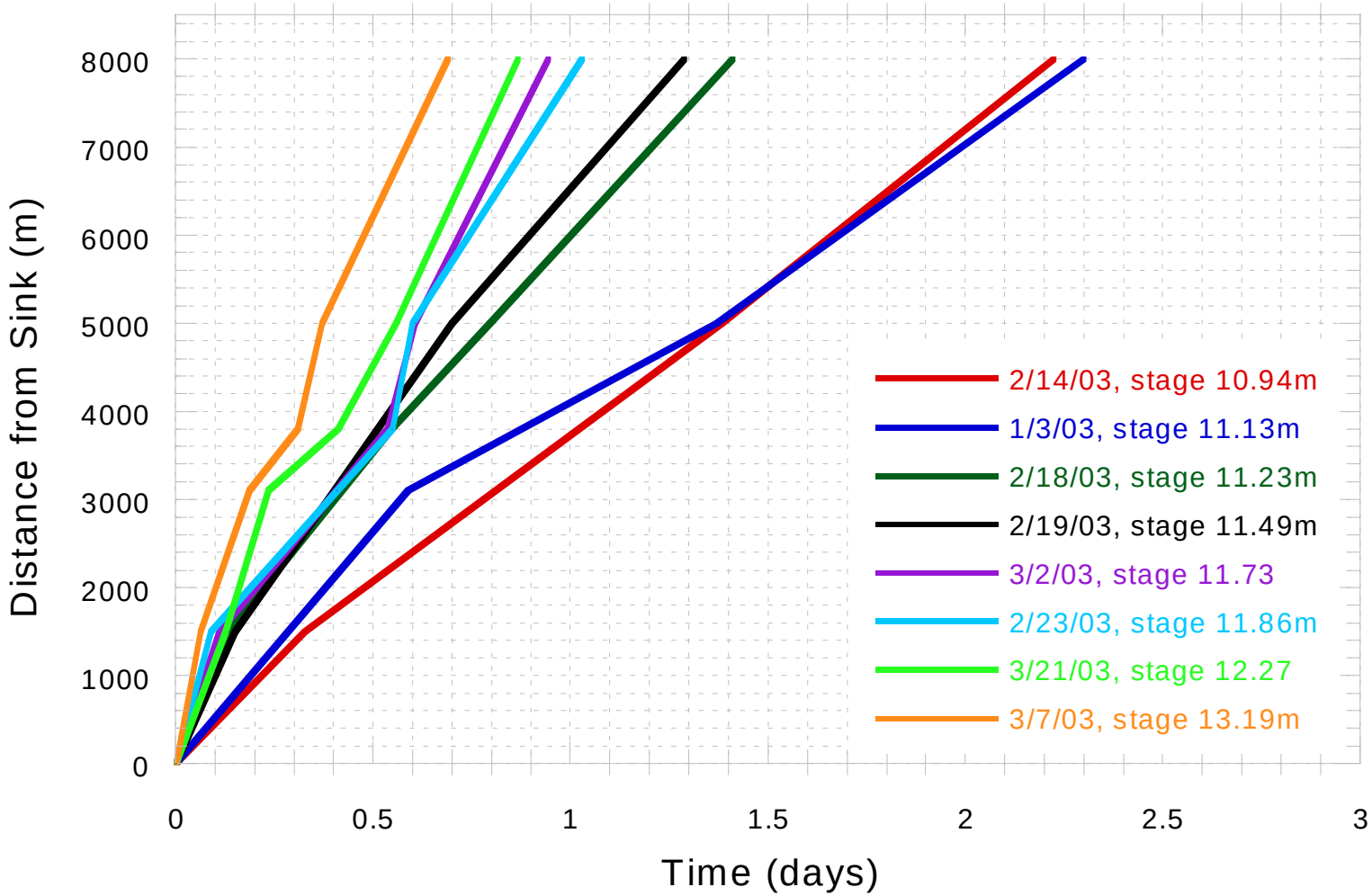


Figure 3-11. Distance from the River Sink versus temperature signal lag time relationship. The slope of the line is equal to conduit water velocity. Conduit water velocity increases with increasing river stage.

constrained due to lack of physical mapping of the conduit and were estimated using straight-line distances. This may result in small velocity calculation errors between these locations.

**Reynolds Number, Colebrook and White, and Darcy-Weisbach Equation**

The Reynolds number determines whether flow will be laminar or turbulent (Fetter, 2001). Reynolds numbers calculated for the Santa Fe River Sink/Rise system (Table 3-1) are >4000, confirming turbulent flow through the conduits during each period. For diffuse groundwater flow, the Reynolds number is usually less than one, but Darcy's Law remains valid up to a Reynolds number of 10. In pipes, flow will be laminar up to a Re of 2000. From 2000-4000, flow is in transition from laminar to turbulent, and above 4000, flow is completely turbulent. The friction factor (f) of the conduit that was calculated using the Darcy-Weisbach equation yielded values ranging from 6.2-18.5 (Table 3-1). Typical values for the absolute roughness of the conduit (e), are less than 0.2 for man made pipes. Calculated values of e (Table 3-1) (Colebrook and White, 1937) ranged from 49 m to 61m, with an average of 55 m.

Table 3-1. Average conduit water velocity, Reynolds number (Re), friction factor (f), and absolute roughness (e) results for each event.

| Date | River Stage | Velocity (m/s) | Re | f | e (m) |
|---|---|---|---|---|---|
| 12/30/02 | 11.03 | 0.034 | 750,580 | 8.55 | 54.4 |
| 1/3/03 | 11.13 | 0.03 | 662,276 | 12.55 | 58.2 |
| 1/9/03 | 11.14 | 0.038 | 838,833 | 10.7 | 55.9 |
| 2/14/03 | 10.94 | 0.031 | 662,276 | 10.7 | 56.7 |
| 2/18/03 | 11.23 | 0.05 | 1,103,793 | 6.212 | 49.0 |
| 2/19/03 | 11.49 | 0.054 | 1,192,097 | 7.55 | 53.0 |
| 2/23/03 | 11.86 | 0.068 | 1,479,083 | 8.03 | 53.7 |
| 3/7/03 | 13.19 | 0.1 | 2,207,587 | 7.98 | 53.6 |
| 3/21/03 | 12.27 | 0.08 | 1,766,069 | 18.48 | 61.6 |

## Matrix Properties and Exchange between Conduits/Matrix

### Transmissivity Results

Transmissivities calculated using the methods of Ferris (1963) and Pinder et al. (1969) ranged from 140 to 550,000 $m^2/d$ (Table 3-3). Transmissivity results for Wells 3 and 6 using the Stage Ratio method and the Time Lag method could not be calculated due to lack of head data during the water level peak. Curve matching using the Pinder et al. (1969) method required two different transmissivities to match amplitude and time lag for Tower Well (12/13/02 to 2/6/03) (Figures 3-16 and 3-17). Due to lack of peak water level data for wells 3 and 6, amplitude could not be matched. Curve matches between actual head change and head change calculated using the Pinder et al. (1969) method are shown for Well 1 (Figure 3-12), Well 3 (Figure 3-13), Well 4 (Figure 3-14), Well 6 (Figure 3-15), Tower Well (1/6/03) (Figures 3-16 and 3-17), and Tower Well (3/15/03) (Figure 3-18).

Since storativity was estimated at 0.2, sensitivity tests were conducted to determine the effects of variations in storativity on transmissivity using the minimum and maximum ranges of effective porosity (Palmer, 2002) as upper and lower limits for storativity since storativity cannot exceed porosity. Lowering storativity to 0.1 reduced transmissivity by approximately half, while raising storativity to 0.45 approximately doubled transmissivity. Distances used for calculating transmissivity can be considered a maximum value since they were measured from the well to the nearest known conduit location. If the distance to the nearest conduit is less than the value used to calculate transmissivity than transmissivity will decrease. For example, if the distance to the nearest conduit is reduced by half, transmissivity will be four times smaller than estimated.

Table 3-2. Transmissivity results.

| | | Transmissivity ($m^2$/day) | | |
|---|---|---|---|---|
| Location and Dates of fluctuation. | Distance from Conduit (m) | Stage Ratio Method | Time Lag Method | Pinder, et al. (1969) Method (both time lag and amplitude match unless otherwise noted) |
| Sink to Tower Well 12/13/02-2/6/03 (55 day period) | 3750 | 109000 | 250000 | Time lag match 120000 Amplitude match 550000 |
| Sink to Tower Well 2/8/03-3/27/03 (47 day period) | 3750 | 78000 | 153000 | 160000 |
| Sink to Well 1 2/5/03-3/27/03 (50 day period) | 475 | 4200 | 396000 | 97000 |
| Rise to Well 4 2/5/03-3/27/03 (50 day period) | 115 | 140 | 960 | 950 |
| Rise to Well 6 2/8/03-3/27/03 (47 day period) | 85 | NA | NA | Time Lag match 900 Amplitude match NA |
| Rise to Well 3 2/8/03-3/27/03 (47 day period) | 30 | NA | NA | Time Lag match 5000 Amplitude match NA |

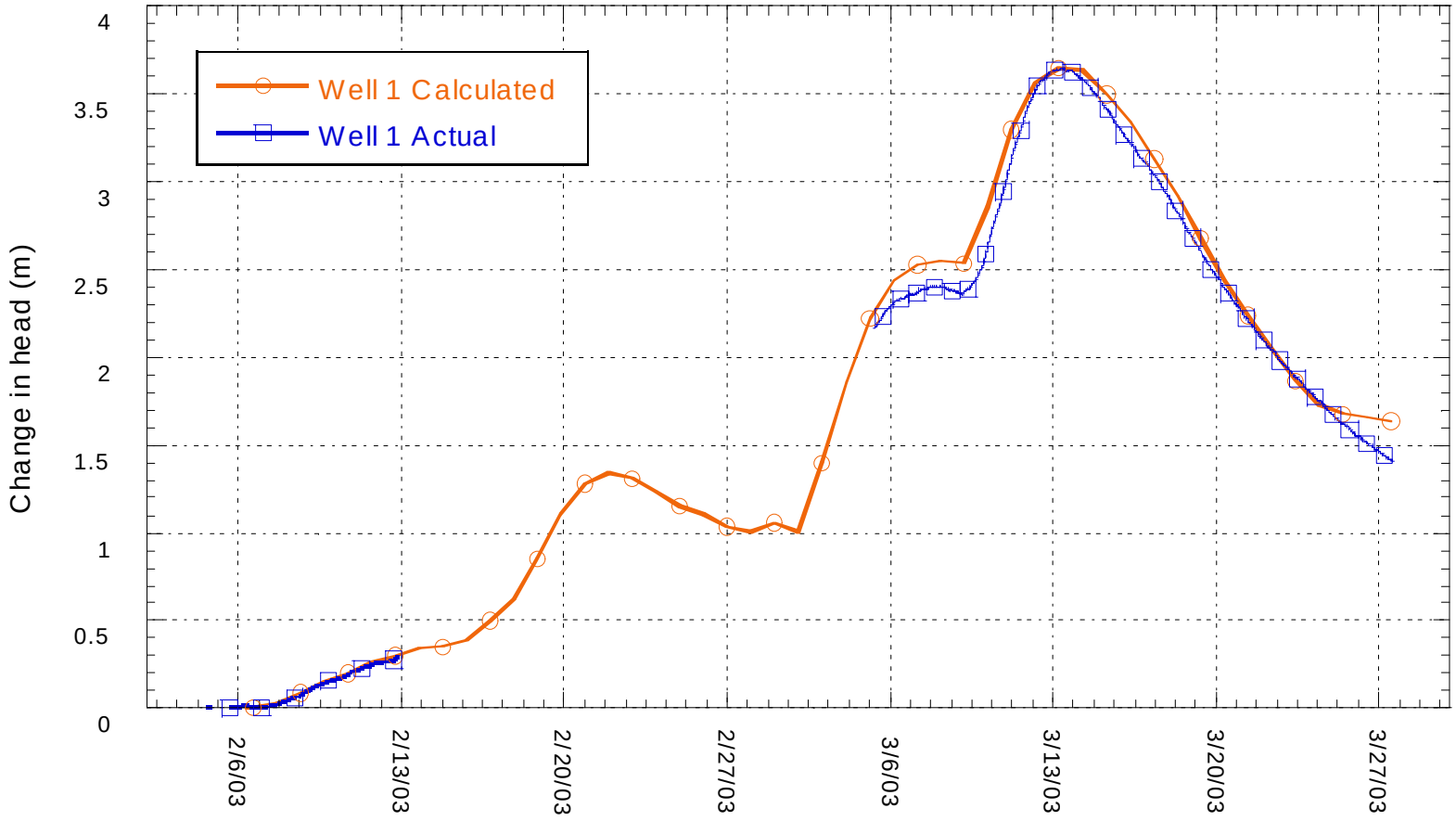


Figure 3-12. Curve matching for Well 1 using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

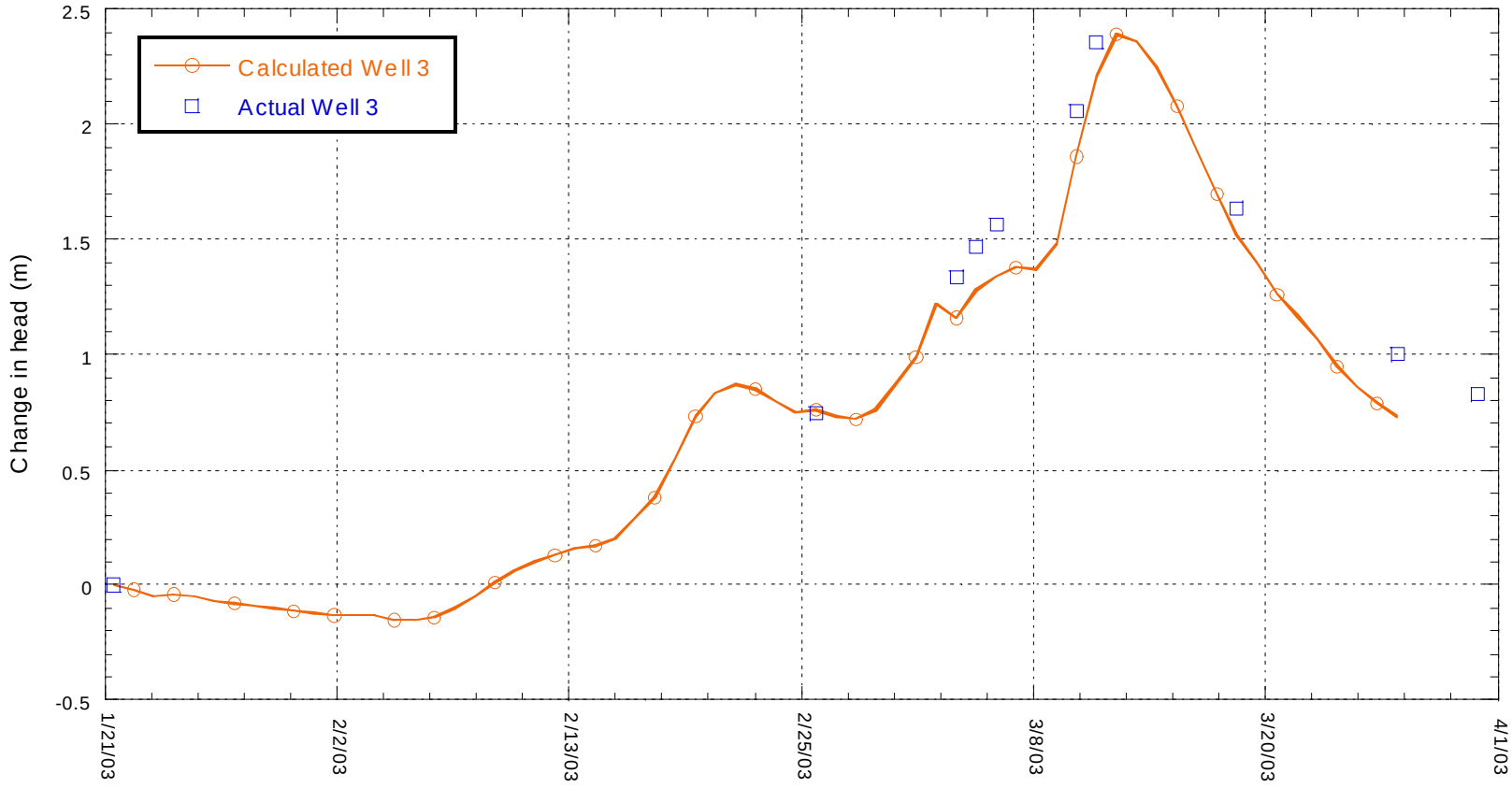


Figure 3-13. Curve matching results for Well 3 using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

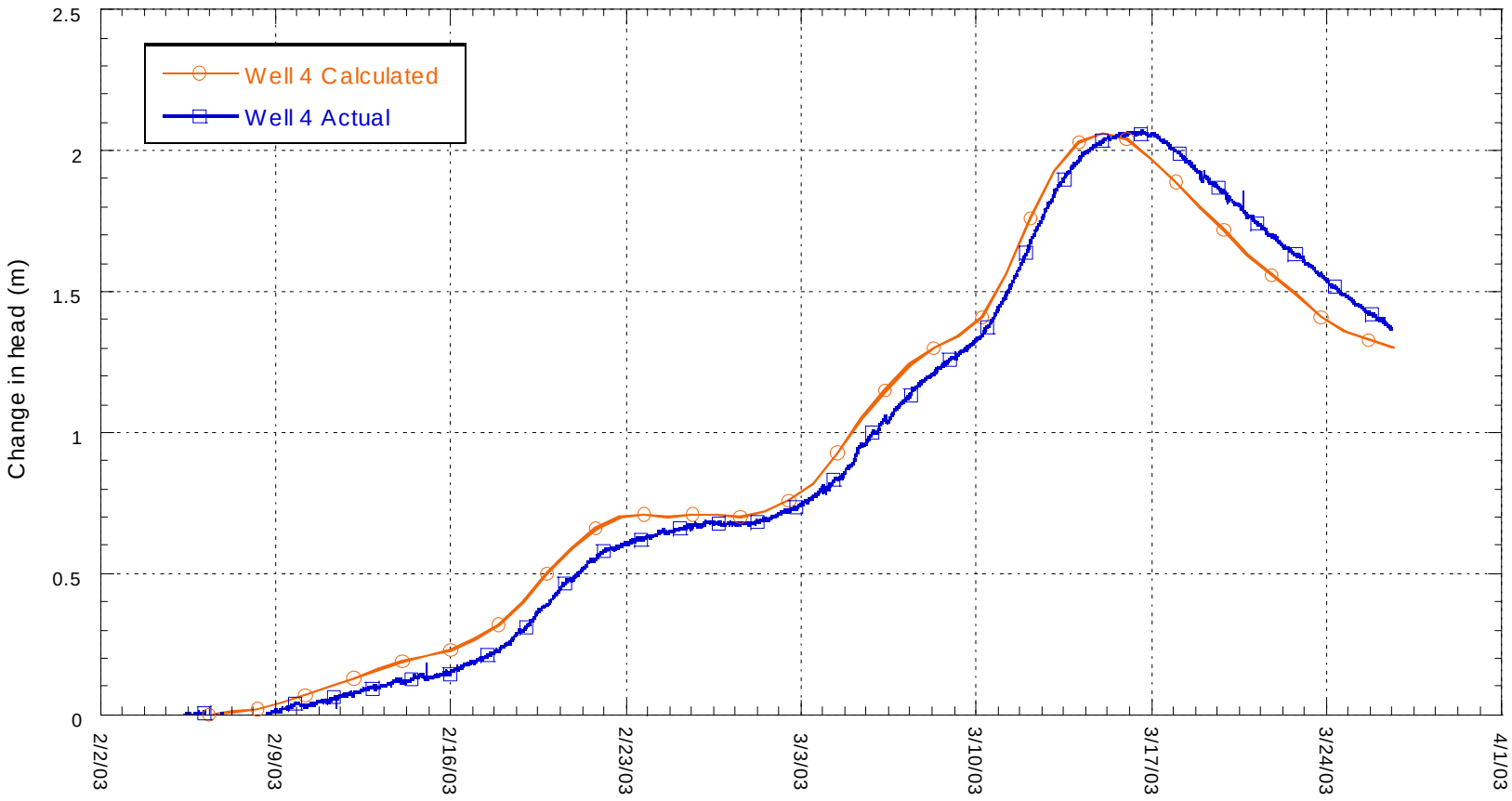


Figure 3-14. Curve matching results for Well 4 using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

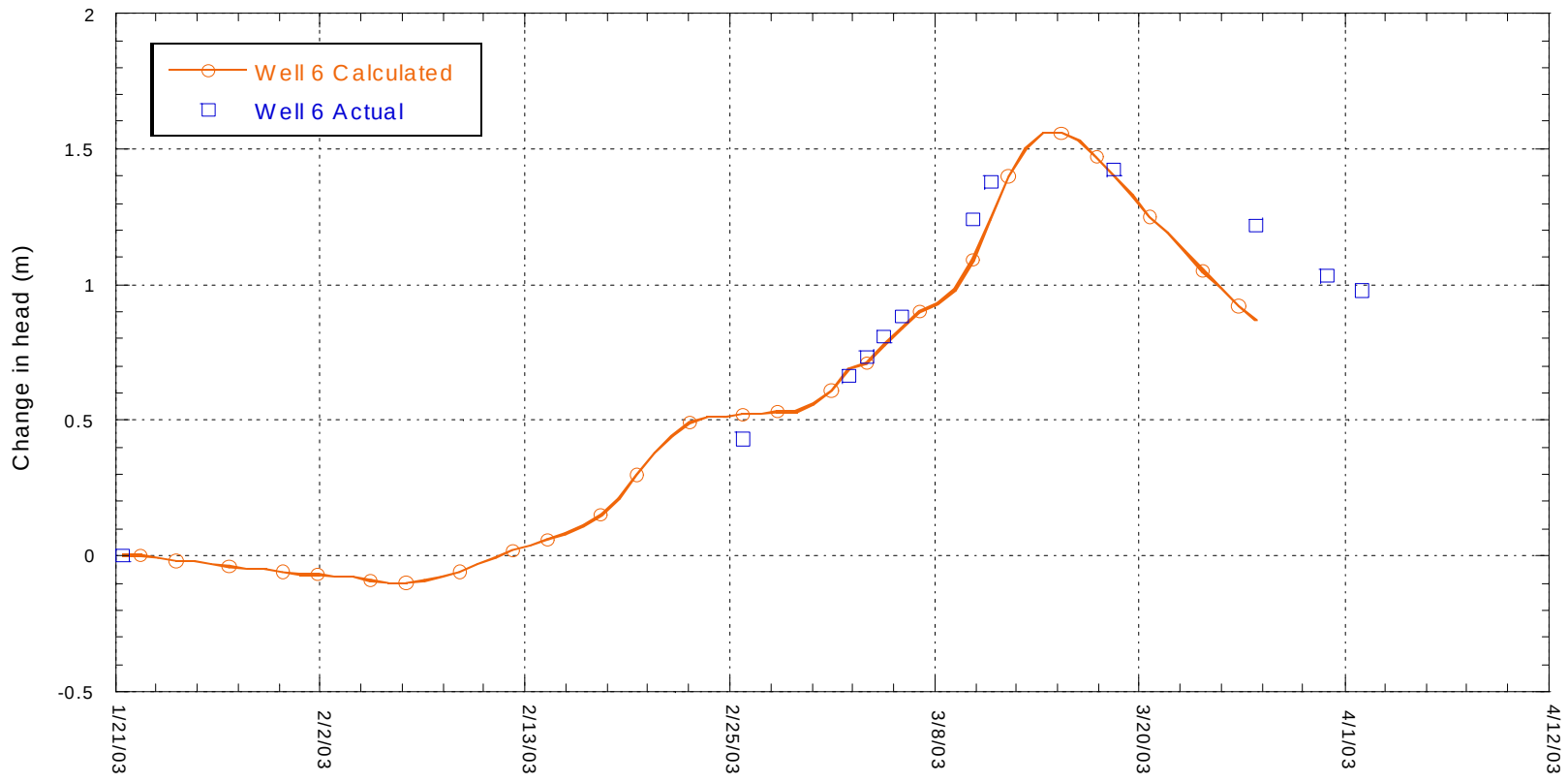


Figure 3-15. Curve matching results for Well 6 using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

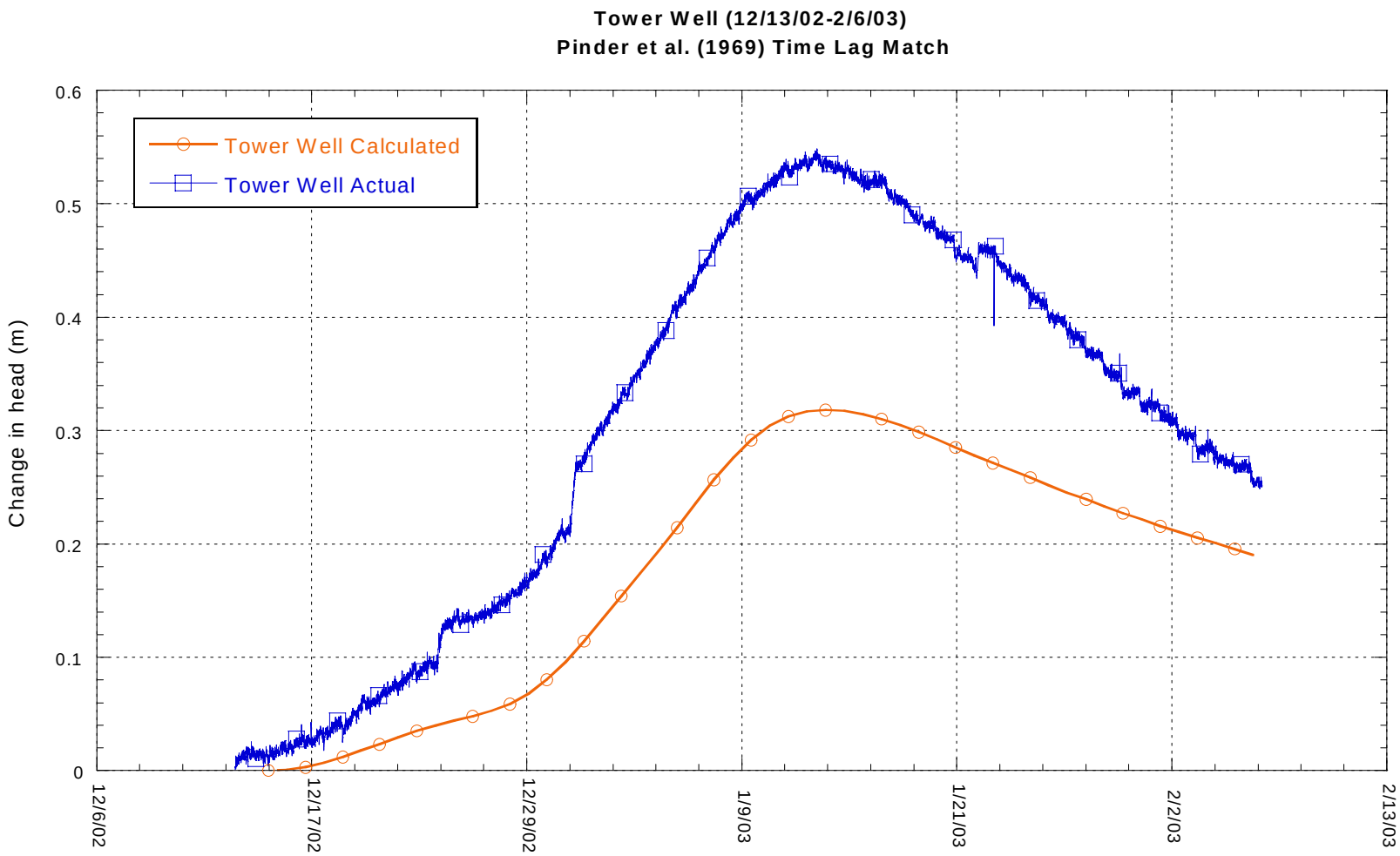


Figure 3-16. Curve matching results for Tower Well (12/13/02-2/6/03) time lag using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

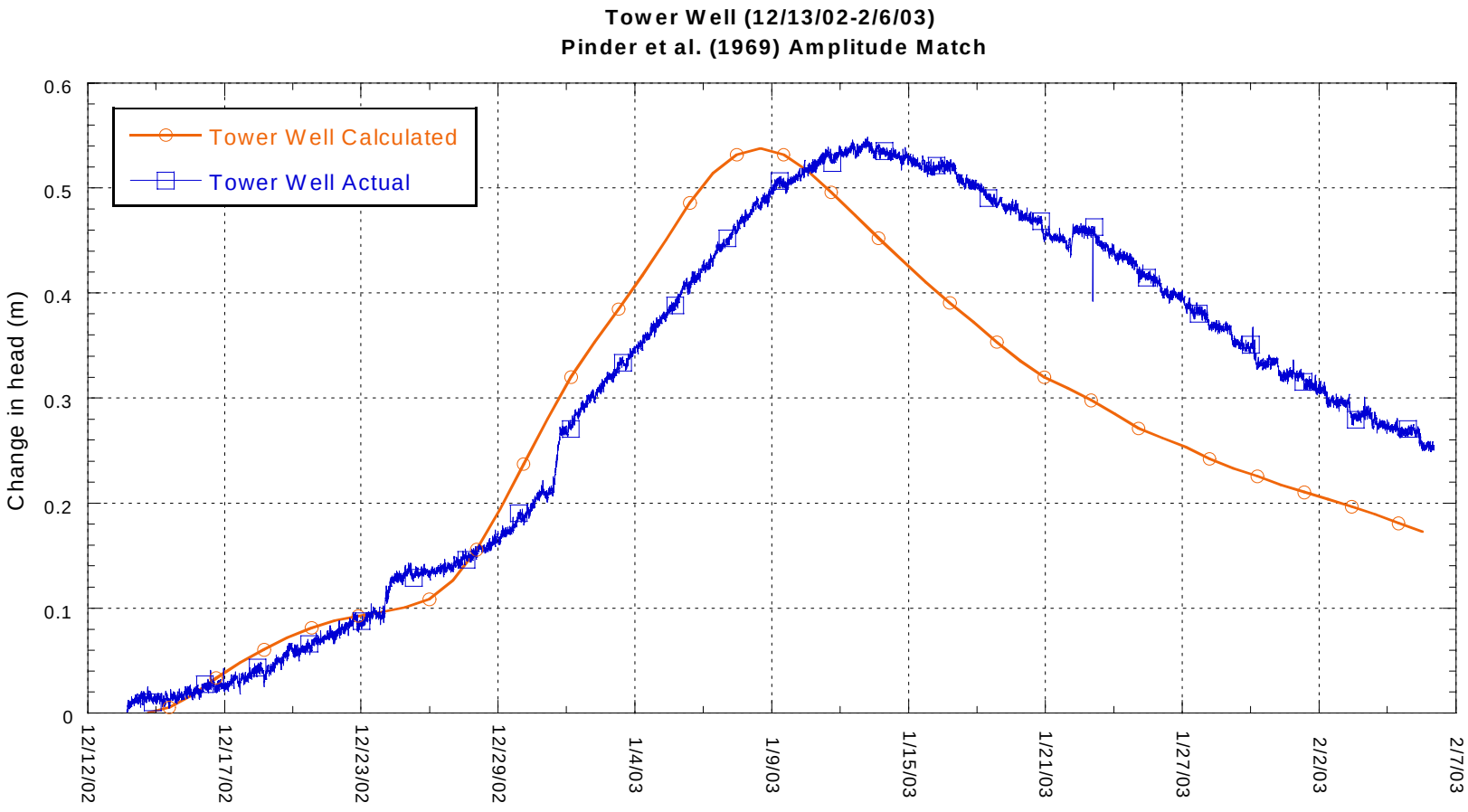


Figure 3-17. Curve matching results for Tower Well (12/13/02-2/6/03) amplitude using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

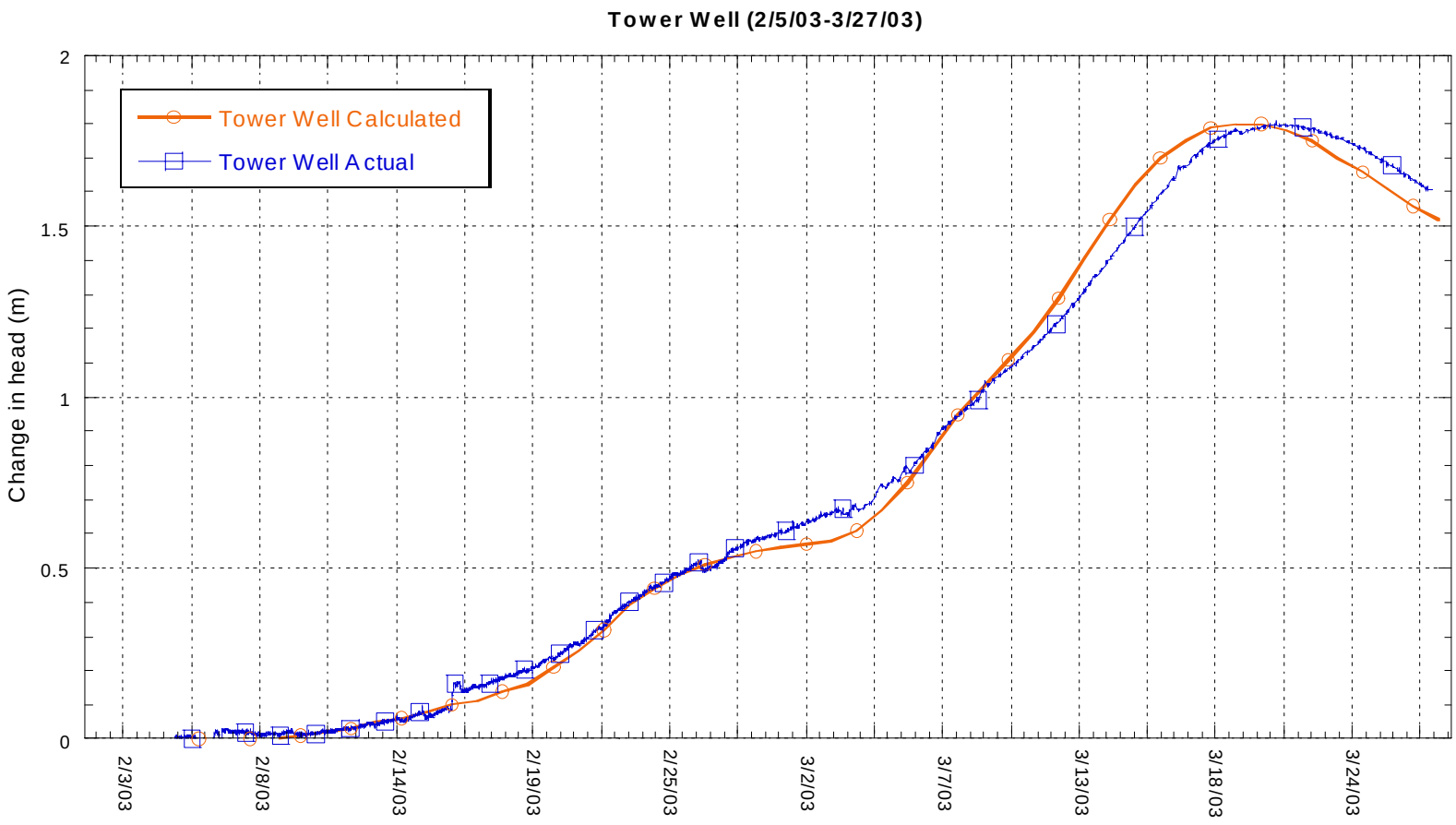


Figure 3-18. Curve matching results for Tower Well (2/5/03-3/27/03) using the Pinder et al. (1969) method. The change in head refers to the difference between the water level on a given day and the water level on the first day.

# CHAPTER 4
# DISCUSSION

## Conduit Properties

Characterizing the hydrologic properties of submerged conduits is a difficult task. Direct observation and mapping of caves such as the Santa Fe River Sink/Rise system can be performed only by divers. Understanding the hydrologic properties of the conduit system as well as the hydrologic properties of the matrix is crucial for determining the interrelationship between them. Characterizing conduit properties such as the Reynolds number, friction factor, roughness factor, and conduit water velocity, helps to clarify the interaction between conduits and the surrounding matrix.

Friction factor (f) results (Table 3-1) are similar to values in three conduit systems in the Mendip Hills, Somerset, U.K. that ranged from 24 to 340 (Atkinson, 1977). Bloomburg and Curl (1974), used artificial laboratory flumes to calculate f, and Gale (1984) studied a segment of Fissure Cave in northwest England. The latter two studies both reported f values less than one, far lower than the results reported by Atkinson 1977. There are two possible explanations for the discrepancy between the results. Atkinson's results and those from this study were obtained using the entire length of the conduit, which may vary significantly in diameter over its length. In contrast, Gale (1984) used short sections of conduit to calculate f values. Similarly, the experimentally produced flutes and scallops of Bloomburg and Curl (1974) involved short artificial conduits.

Atkinson (1976) determined that the absolute roughness was about three times the diameter of the conduits he studied in the Mendip Hills, Somerset, UK. Likewise, the

average calculated absolute roughness of the O'Leno State Park Sink/Rise system is 55m, nearly three times the estimated conduit diameter of 20 m (Ginn, 2002).

**Matrix Properties**

Differences in calculated transmissivity among the Stage Ratio, Time Lag, and Pinder et al. (1969) methods varied by up to three orders of magnitude (Table 3-2). The Stage Ratio and Time Lag methods may be in error because calculations requiring sinusoidal groundwater fluctuation curves to were applied to non-sinusoidal events. The Pinder et al. (1969) method does not assume a sinusoidal curve, and because it matches actual water level fluctuations to calculated water level fluctuations, it almost certainly yields the best calculation of transmissivity for the study area of the three methods. It is unclear precisely why the calculated and actual head change curves between the River Sink and Tower Well from 12/13/02 to 2/6/03 event did not match with the transmissivity chosen for the best time lag match (Figures 3-16 and 3-17). Because the total head change during this event was much smaller than during the February-March event, the water levels are likely to be more affected by measurement errors, instrument disturbances, or diffuse recharge. For example, a comparison of the actual head change curve with the calculated change in head (Figure 3-16) shows sudden rises in the measured head on 12/24/02 and 12/31/02, which could possibly be due to recharge. Thus, the transmissivity determined from this event is expected to be less reliable than from the February-March event.

Bush and Johnston (1988) estimated the transmissivity of the Upper Floridan Aquifer in the region. They estimated a value of 93,000 $m^2$/day based on calibration of a quasi-three-dimensional finite difference model with a cell size of 165 $km^2$. This is

consistent with transmissivities calculated with the Pinder et al. (1969) method for Wells 1 and Tower Well, which are more than 475 m from the conduit.

Slug tests performed on wells 3, 4, 5, 6, and 7 near the River Rise yielded transmissivity values ranging from 270 $m^2/d$ - 550 $m^2/d$ (Hamilton, 2003). Slug test results are one-fifth the value of transmissivities calculated for wells near the River Rise using the Pinder et al. (1969) method. When comparing the Pinder et al. (1969) results to very small-scale slug tests, lower values of transmissivity are expected for slug tests. Due to their limited effective radius, slug test results typically underestimate transmissivity by 30% to greater than 100% (Weight and Sonderegger, 2001).

**Matrix Hydraulic Conductivity**

Transmissivities calculated from the Pinder et al. (1969) method were converted to hydraulic conductivity (K) using an aquifer thickness of 275m (Table 4-1) (Hisert, 1994). Hydraulic conductivity is the rate water is transmitted through a cross sectional area of the aquifer. Because the conduit partially penetrates the aquifer, it cannot be assumed that the full-saturated thickness of the aquifer is participating in flow. A simple calculation was used to determine the radius beyond which the effects of partial penetration can be ignored. Effects are limited to a radius equal to 1.5 (horizontal hydraulic conductivity ($K_h$) / vertical hydraulic conductivity ($K_v$))$^{1/2}$ times the saturated thickness of the aquifer (Anderson and Woessner, 1992). The minimum radius can be calculated by ignoring the anisotropy ($K_h/K_v$), which is likely to be greater than 1, and the effective radius becomes 413 meters. This means that the effects of conduit partial penetration cannot be ignored for wells at a distance less than approximately 400 meters from the conduit. Therefore, using an aquifer thickness of 275 meters to calculate

hydraulic conductivity for wells less than approximately 400 meters from the conduit may result in artificially low values.

Table 4-1. Matrix Hydraulic Conductivity based on an aquifer thickness of 275 m.

| | | Hydraulic Conductivity (m/day) |
|---|---|---|
| Location and dates of fluctuation. | Distance from Conduit (m) | Pinder et al. (1969) Method (both time lag and amplitude match unless noted) |
| Sink to Tower Well 12/13/02-2/6/03 | 3750 | Time lag match 440<br>Amplitude match 2000 |
| Sink to Tower Well 2/8/03-3/27/03 | 3750 | 580 |
| Sink to Well 1 2/5/03-3/27/03 | 475 | 350 |
| Rise to Well 4 2/8/03-3/27/03 | 115 | 4 |
| Rise to Well 6 2/8/03-3/27/03 | 85 | Time lag match 3<br>Amplitude match NA |
| Rise to Well 3 2/8/03-3/27/03 | 30 | Time lag match 18<br>Amplitude match NA |

## Scale Effects

Averaging numerous small-scale tests of hydraulic conductivity in a karst aquifer will result in lower results than the average of a large-scale test in the same area (Bradbury and Muldoon, 1990; Rovey and Cherkauer, 1994c). Calculating hydraulic conductivity over larger distances increases the likelihood that water is finding preferential paths through the matrix. With increasing scale the preferential pathways tend to dominate a larger percentage of groundwater flow, thus increasing average hydraulic conductivity (Rovey, 1994). In karstic carbonates such as the Upper Floridan Aquifer, hydraulic conductivity increases proportionally with the amount of dissolution within the aquifer (Rovey, 1994). In addition to the possible effects of the partially

penetrating conduit, scale effects are another likely reason that hydraulic conductivity values calculated for distances between a well and the conduit of greater than 475 m are two orders of magnitude greater values calculated for wells proximal to the conduit. Increasing values of hydraulic conductivity with distance from the conduit may reflect the highly heterogeneous nature of the Floridan Aquifer.

## Mixing of Conduit and Matrix Water

### Comparison of Discharge between Sink and Rise

Discharge differences between the Sink and Rise show whether water is either entering or leaving the conduit system (Fig. 3-7). When Sink discharge exceeds Rise discharge, the conduit is losing water to the matrix, and when Rise discharge exceeds Sink discharge, the conduit is gaining water from the matrix. One of the largest contributors to groundwater entering the conduit during low river stages is a feeder conduit entering the main conduit from the East (Fig. 1-2) (Old Bellamy Cave Exploration Team, unpublished report, 2001). There are no obvious surface water sources supplying water to the eastern system, suggesting that it is recharged by groundwater from the Floridan Aquifer (Screaton et al., in press).

There is not a linear relationship between river stage and the change in discharge between the Sink and Rise (Fig. 4-1), but there are linear trends depending on whether the river stage is rising or falling. As river stage rises, the conduit begins to lose water to the matrix. As more water moves out of the conduit, matrix heads begin to rise. When river stage begins to fall, the change in discharge becomes increasingly more positive. Because matrix heads have increased with increasing river stage due to water outflow from the conduit, when the river stage begins to fall, they do not follow the same path as

when stage was rising. This illustrates the complexity of the hydrologic system and demonstrates the importance of matrix head on the mixing of conduit and matrix water.

**Gradients between Conduits/Matrix**

To demonstrate the movement of water between the matrix and conduit system, the water table gradient between the wells and conduit was estimated at varying river stages (Table 4-2). Because the head in the conduit closest to the wells was unknown, an assumption was made that the head within the segment of the conduit was the same as at the Sink or Rise, whichever was closer to the well. Results indicate that when the water table gradient measured in the matrix is higher than the gradient in the conduit water is flowing from the matrix into the conduit. Conversely, when the gradient in the conduit is higher than the gradient in the matrix, water is leaving the conduit and entering the matrix. When compared to discharge data, the observed gradients agreed with times when the conduit was gaining or losing water. This shows that water is in fact moving into the matrix when the conduit is losing water and leaving the matrix when the conduit is gaining water.

Table 4-2. Water level elevation comparison between monitoring wells and the Sink and Rise. Missing data indicate unavailable water level data. Head differences of less than 0.11 m may not be significant due to potential water level errors.

| Date | Sink wl (m) | Tower Well (m) | Well 1 (m) | Well 2 (m) | Well 3 (m) | Well 4 (m) | Well 5 (m) | Well 6 (m) | Rise (m) |
|---|---|---|---|---|---|---|---|---|---|
| 1/22/03 | 10.03 | 10.25 | 10.33 | 10.31 | 10.13 | 10.64 | | 10.34 | 10.04 |
| 2/26/03 | 11.11 | 10.55 | 11.26 | 10.9 | 10.88 | 11.13 | | 10.77 | 10.84 |
| 3/5/03 | 12.67 | 10.79 | 12.36 | 11.17 | | | | 11.08 | 11.57 |
| 3/6/03 | 12.86 | 10.84 | 12.46 | 11.24 | 11.60 | 11.53 | | 11.15 | 11.69 |
| 3/7/03 | 12.89 | 10.92 | 12.54 | | 11.70 | 11.63 | | 11.23 | 11.76 |
| 3/11/03 | 13.86 | 11.20 | 13.28 | | 12.19 | 11.97 | | 11.59 | 12.35 |
| 3/12/03 | 14.14 | 11.27 | 13.67 | | 12.49 | 12.13 | 12.14 | 11.72 | 12.74 |
| 3/27/03 | 11.12 | 11.66 | 11.62 | | 11.13 | 11.84 | 11.79 | 11.56 | 11.02 |
| 5/14/03 | | 10.12 | 10.18 | 10.71 | 9.99 | 10.61 | 10.54 | 10.33 | 9.94 |

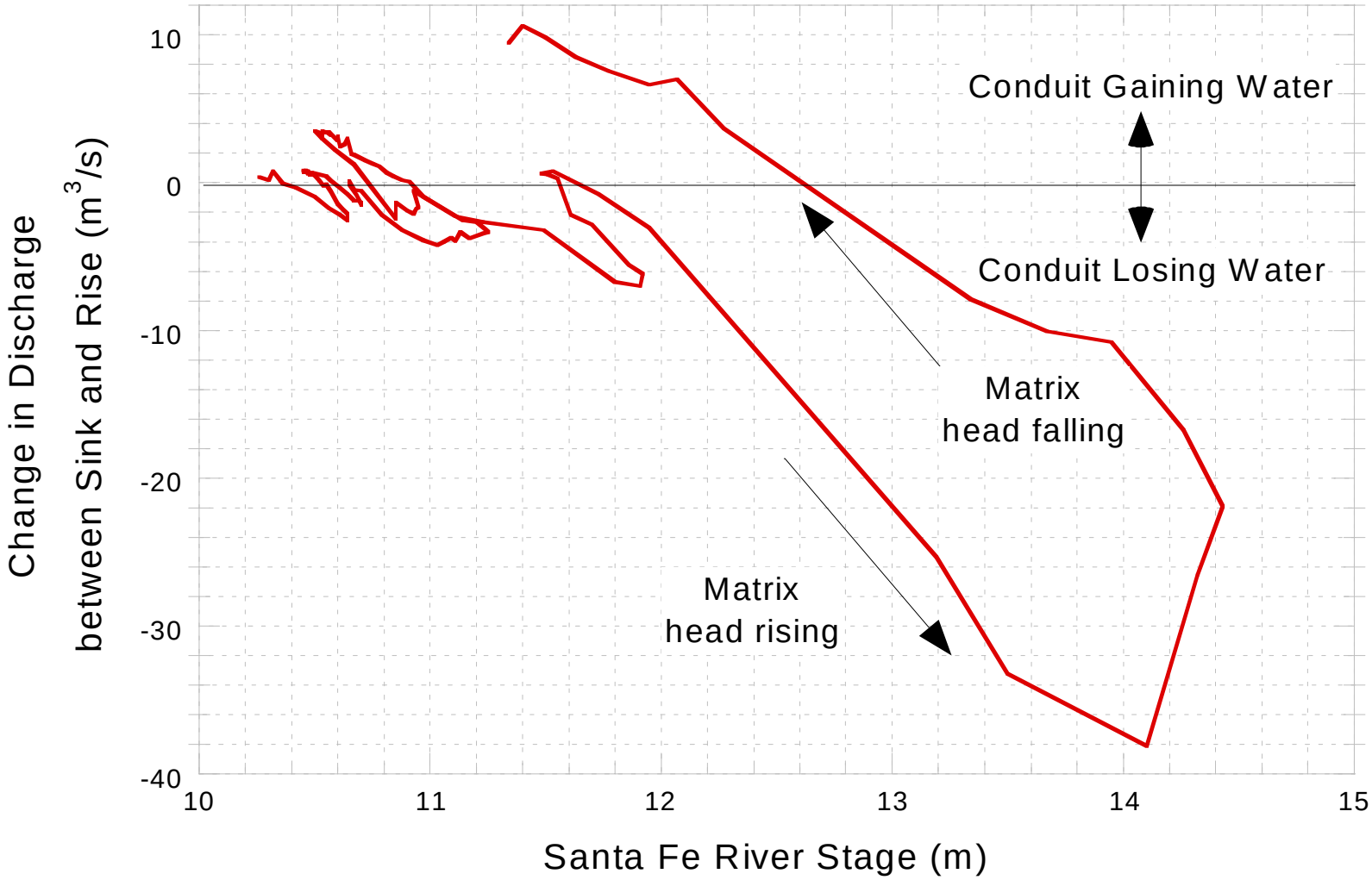


Figure 4-1. Relationship between changes in discharge and river stage.

**Discharge/Gradient Relationship**

To further explore how the exchange of water varies with changes in discharge and water table gradient, the change in discharge between the Sink and Rise versus the corresponding head gradient between a well and the conduit was plotted. Tower Well and Well 4 were chosen because of the high number of data points collected from each location. The plots reveal a linear relationship between matrix head gradient and the change in discharge between the River Sink and the River Rise (Figures 4-2 and 4-3). Gradient magnitudes seem to be proportional to the magnitude of the discharge. Ideally the best-fit line should cross the origin, the point where the conduit is neither gaining nor losing water and the head gradient is zero. Tower Well (Fig. 4-2) comes closest to the ideal but is still off by +0.38 meters. A 0.38 meter decrease in the head difference between Tower Well and the River Sink would be required for the best-fit line to cross the origin. The error required for the best-fit line of Well 4 to cross the origin is a 0.35 meter increase in the head difference between Well 4 and the River Rise. Summing of errors at the Sink or Rise and the wells suggest a total error of ± 0.09 to 0.11 m. One possibility for the discrepancies is that the conduit may not gaining/losing water uniformly along its length. For flow that follows Darcy's Law, the slope of the best-fit line derived from the plot of the change in discharge between the Sink and Rise versus matrix head gradient (Figures 4-2 and 4-3) is equal to hydraulic conductivity (K) multiplied by area (A). Since KA = transmissivity (T) multiplied by width (w), dividing the slope by transmissivity values calculated using the Pinder et al. (1969) equation, should equal the width of the

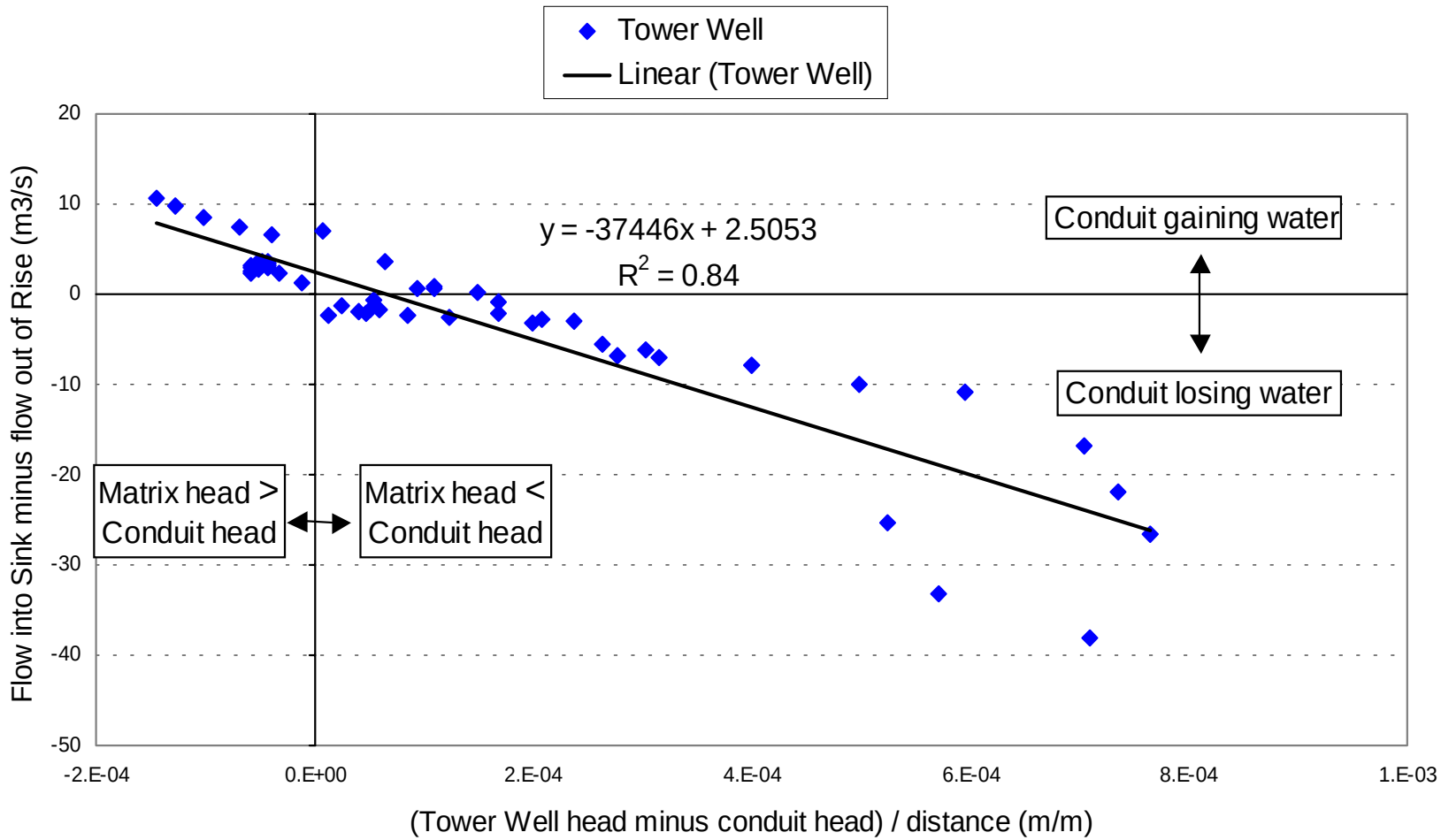


Figure 4-2. Relationship between change in discharge and gradient for Tower Well.

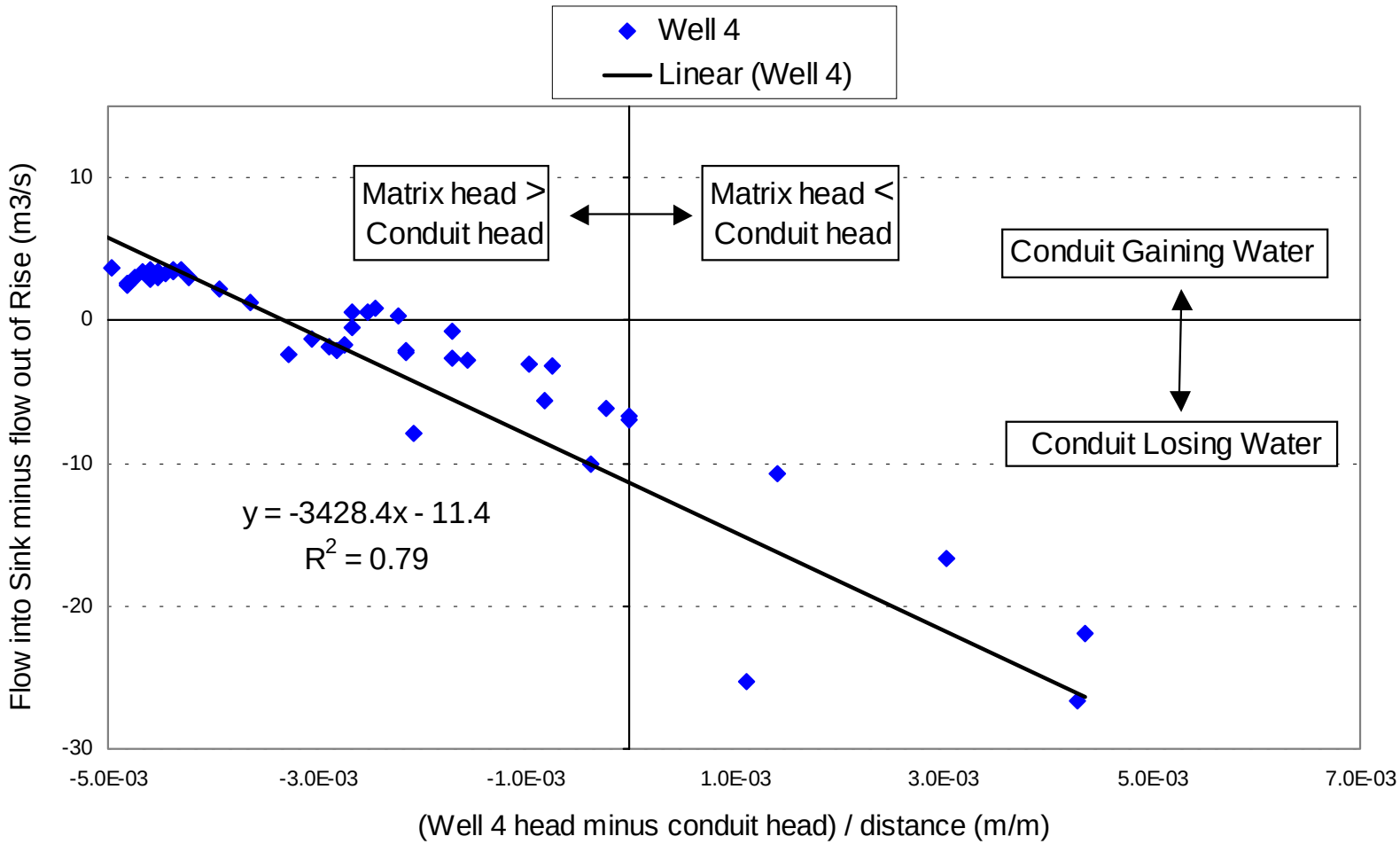


Figure 4-3. Relationship between change in discharge and gradient for Well 4.

conduit interface. The width of the conduit interface divided by 2 (to account for two sides of the fully-penetrating conduit) is estimated to be approximately 8000 meters, the extent of the known conduit.

The area calculations between the conduit and Tower Well using transmissivity values calculated from Pinder et al. (1969) resulted in a conduit length of 3,000 – 13,000 m (Table 4-3), close to the approximately 8,000 meters of known conduit. In contrast, calculations between Well 4 and the conduit using a transmissivity value calculated from Pinder et al. (1969) showed that a conduit interface length of 135,000 meters would be required for a matrix transmissivity of 1100 $m^2/d$. Either the conduit is ten times larger than observed, which seems unlikely, or the transmissivity calculated between Well 4 and the conduit is insufficient to account for the volume of water known to be lost from the conduit. The low value of transmissivity calculated between Well 4 and the conduit is not necessarily wrong, it may be the correct transmissivity of the aquifer when looked at on a very small scale. This demonstrates how scale can affect calculations of transmissivity over small distances such as between Well 4 and the conduit (115 m) and Tower Well and the conduit (3750 m).

Table 4-3. Calculated values for the area of the conduit interface using transmissivities calculated from the Pinder et al. (1969) method.

| Location | Distance from the Conduit (m) | Calculated Transmissivity ($m^2/d$) (Pinder et al. (1969) Method) | (KA) slope of the plot of change in discharge vs gradient (Figures 4-2, 4-3) | Calculated Width of the Conduit Interface (m) |
|---|---|---|---|---|
| Well 4 | 115 | 950 | 3428 | 270,000 |
| Tower Well | 3750 | 120,000 | 37446 | 13,500 |
| Tower Well | 3750 | 160,000 | 37446 | 10,000 |
| Tower Well | 3750 | 550,000 | 37446 | 3,000 |

## Particle Tracking

Particle tracking simulations were conducted in order to better understand and characterize the movement of water between the conduit and matrix. The Pinder et al. (1969) method was previously used to calculate heads as a function of time based on transmissivity and storativity at specified distances from the conduit. The total distance from the conduit was broken into several intervals (Table 4-4). Using this method, multiple spreadsheets were constructed and used to calculate head at varying distances from the conduit.

Table 4-4. Intervals used in the particle tracking simulation.

| Interval | Well 1<br>Interval distance<br>(meters from the conduit) | Well 4<br>Interval distance<br>(meters from the conduit) |
|---|---|---|
| 1 | 475-115 | 115-95 |
| 2 | 115-95 | 95-75 |
| 3 | 95-75 | 75-55 |
| 4 | 75-55 | 55-35 |
| 5 | 55-35 | 35-15 |
| 6 | 35-15 | 15-5 |
| 7 | 15-5 | 5-0 |
| 8 | 5-0 | NA |

A transect or profile of head on specific days during the March 2003 flood was constructed between Well 1 and the conduit and between Well 4 and the conduit. Wells 1 and 4 were chosen for particle tracking because of their proximity to the conduit and their complete water level record. The background gradient of the matrix at the beginning of these simulations is assumed to be toward the conduit based on two facts. First, at the beginning of the simulation, the conduit was gaining water from the matrix based on discharge measurements. Second, the water level in the matrix (measured at wells) was higher than the water level in the conduit. Therefore, assuming a linear gradient between the wells and the conduit, calculated changes in head were

superimposed on the initial heads interpolated between the well and the conduit. Using Darcy's law and an effective porosity estimate, average linear velocities were determined along the calculated transect. Laboratory tests on limestone cores from the Floridan Aquifer have yielded effective porosity values of about 0.17 (Wilson, 2002) and porosity estimates of between 0.10 and 0.45 (Palmer, 2002). Since effective porosity cannot exceed porosity, an effective porosity value of 0.2 was used for average linear velocity calculations.

$$V_x = -\frac{Kdh}{n_e dl}$$

$V_x$ = average linear velocity (m/s)
K = hydraulic conductivity (m/s)
dh/dl = water table gradient (m/m)
$n_e$ = effective porosity (dimensionless)

Using velocities calculated along the transect, a water packet or particle was traced as it left the conduit. Distance calculations for the particle were broken into several shorter time intervals in order not to miss gradient reversals. The residence time in the matrix for water packets was determined by totaling the days the water packet traveled between leaving and returning to the conduit.

The potential distance for a packet of water to travel through the matrix depends on the hydraulic conductivity as well as the head gradient of the matrix. Using the residence time and the distance the water packet traveled it can be determined whether the packets return to the conduit or escape into regional groundwater flow. Advection was assumed the only process affecting particle movement during the simulation. Effects of dispersion and diffusion were ignored.

When the head in the conduit became larger than the head in Well 4 on 3/5/03, the first water particle leaves the conduit (Figure 4-4). The head gradient reversed on 3/16/03 and the water particle began moving back toward the conduit (Figure 4-5). By 3/23/03 the water particle was back in the conduit. The water particle reached its maximum distance of 0.45 meters from the conduit on 3/15/03 (Figure 4-6). The total residence time of the water particle in the matrix was 19 days.

The head in the conduit became larger than the head in Well 1 on 3/3/03 (Figure 4-7). The head gradient reversed after 3/18/03 and the water particle began moving back toward the conduit (Figure 4-8). By 3/25/03 the water particle was back in the conduit. The water particle reached its maximum distance of 8.5 meters from the conduit on 3/18/03 (Figure 4-9). The total residence time of the water particle in the matrix was 21 days.

Although the simulation demonstrated that the water particle did not reach either Well 4 or Well 1 and returned to the conduit in approximately 20 days, this does not mean that it is impossible for conduit water to reach the wells. For the water particle simulation, it is assumed that the particle is traveling through a homogeneous matrix between the conduit and well. It is possible that solutionally enlarged fractures or preferential flow paths exist between the conduit and wells. Hydraulic conductivity could potentially be much higher. For example, if only 1 meter of the estimated 275 m thickness of the Upper Floridan Aquifer is conducting all of the flow, then flow velocity would be 275 times greater than calculated during the particle tracking simulation. The pre-existing groundwater gradient is also likely to be more complex than assumed for the

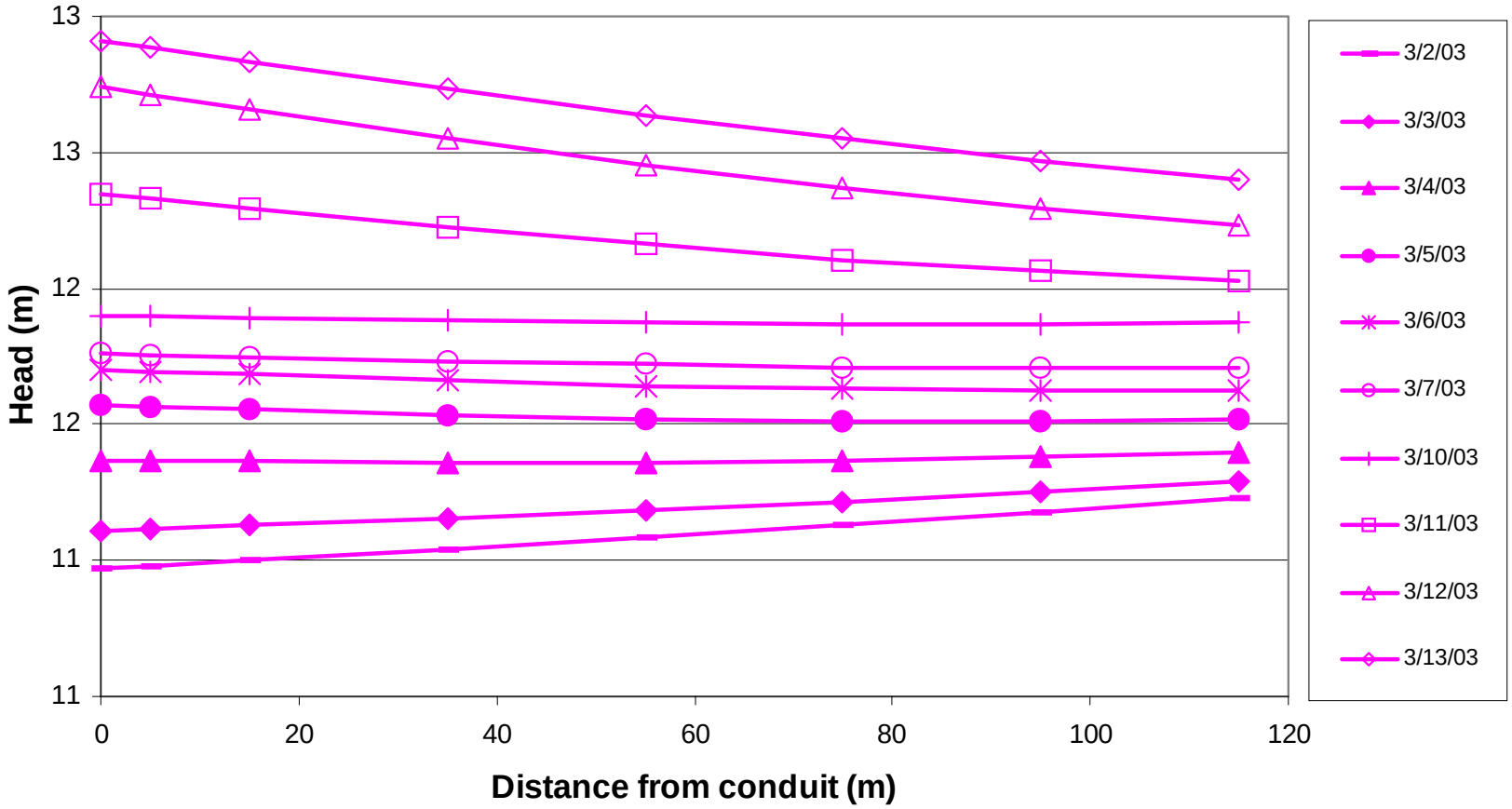


Figure 4-4. Calculated water levels between Well 4 and the conduit system from 3/2/03 to 3/13/03.

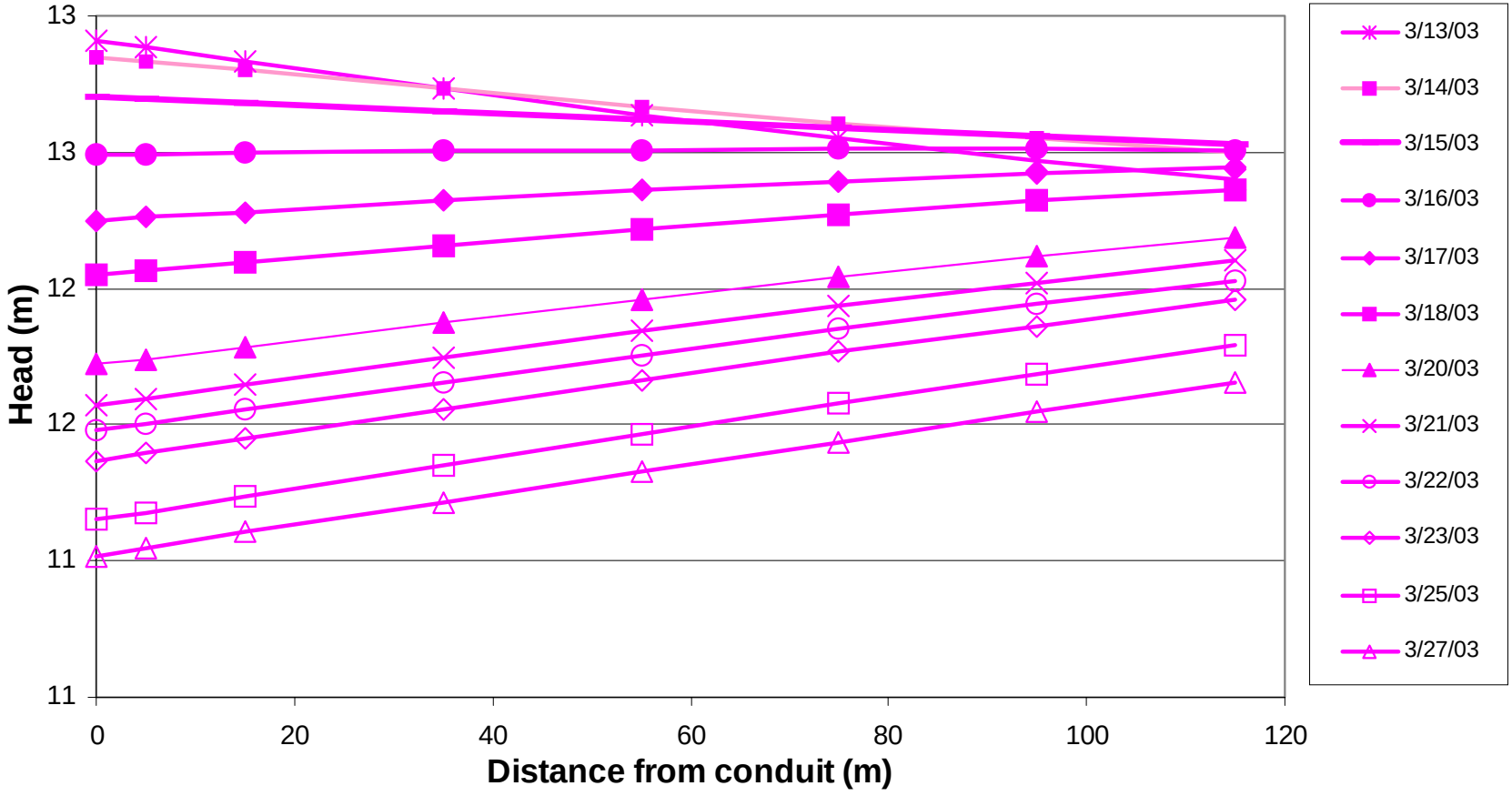


Figure 4-5. Calculated water levels between Well 4 and the conduit system from 3/13/03 to 3/27/03.

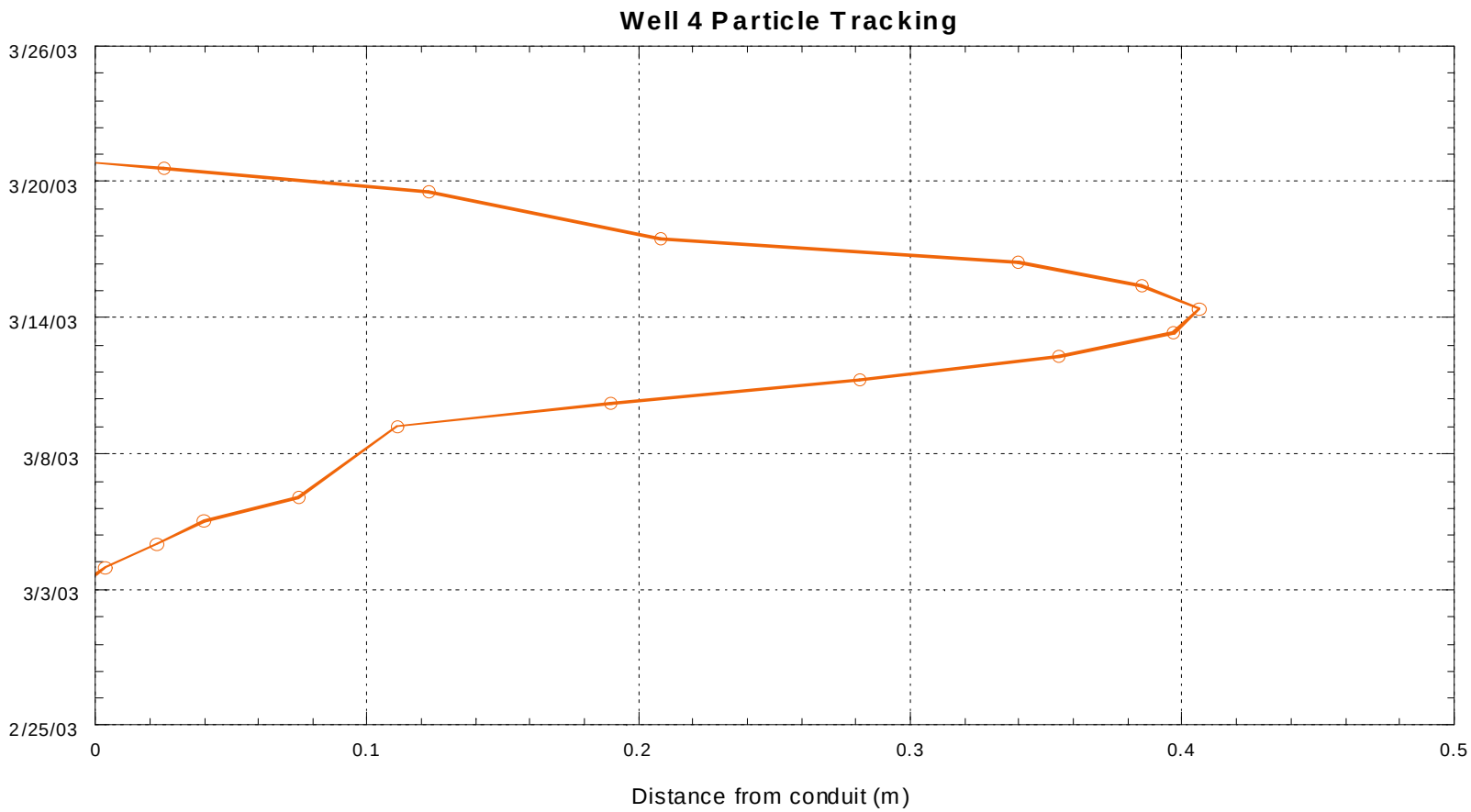


Figure 4-6. Particle tracking between the conduit and Well 4 of a water particle leaving the conduit on 3/4/03.

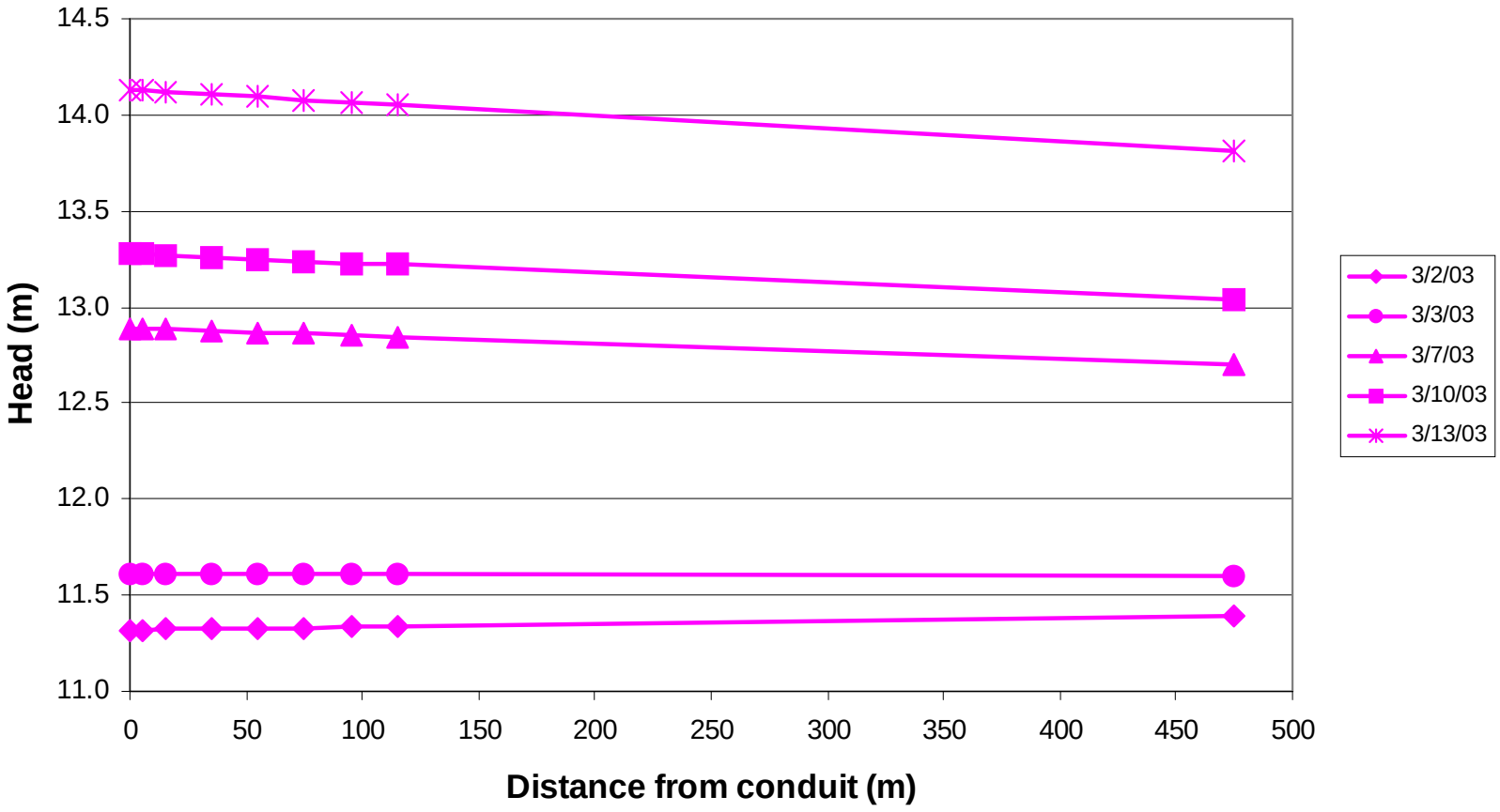


Figure 4-7. Calculated water levels between Well 1 and the conduit system from 3/2/03 to 3/13/03.

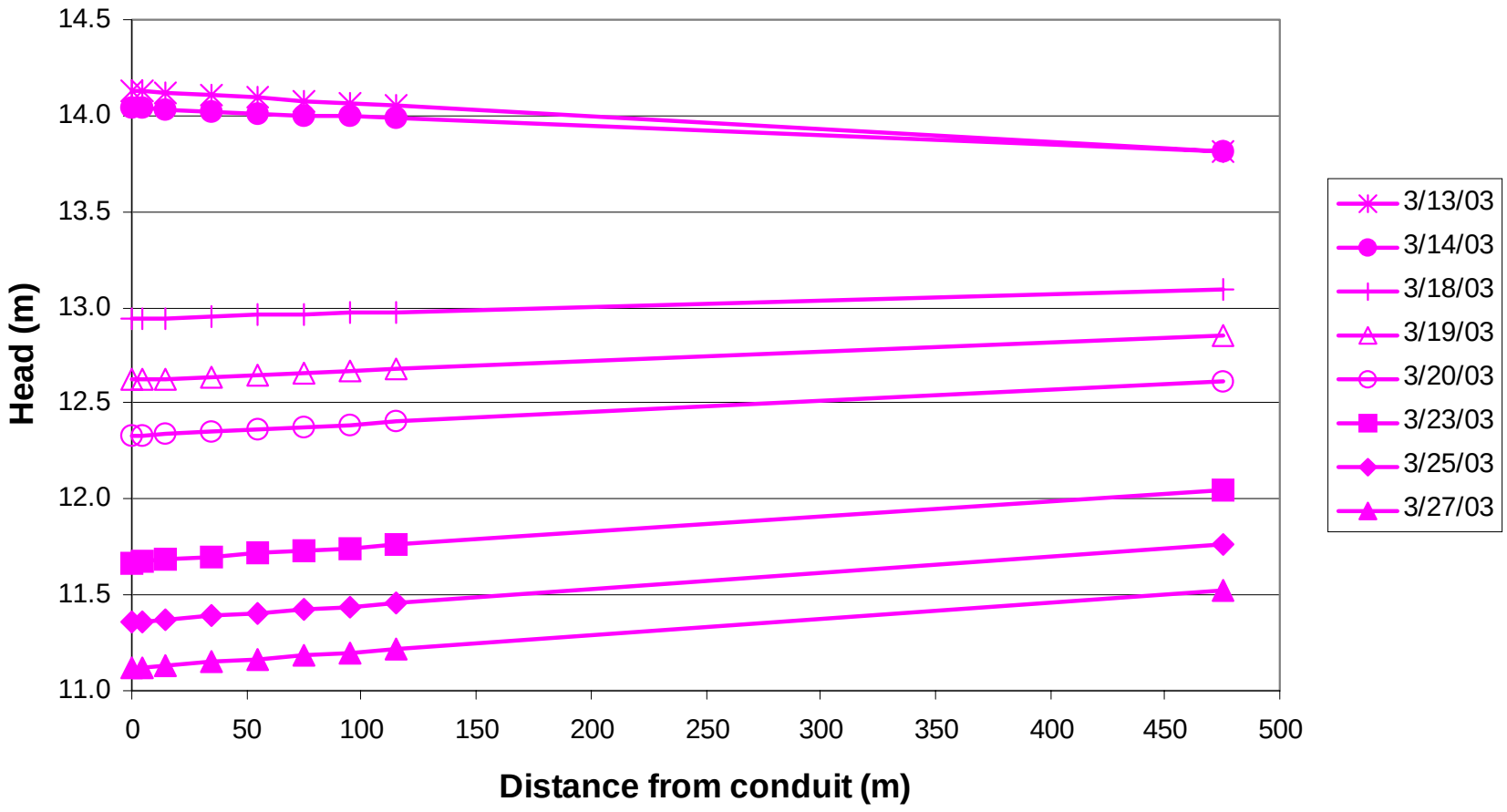


Figure 4-8. Calculated water levels between Well 1 and the conduit system from 3/13/03 to 3/27/03.

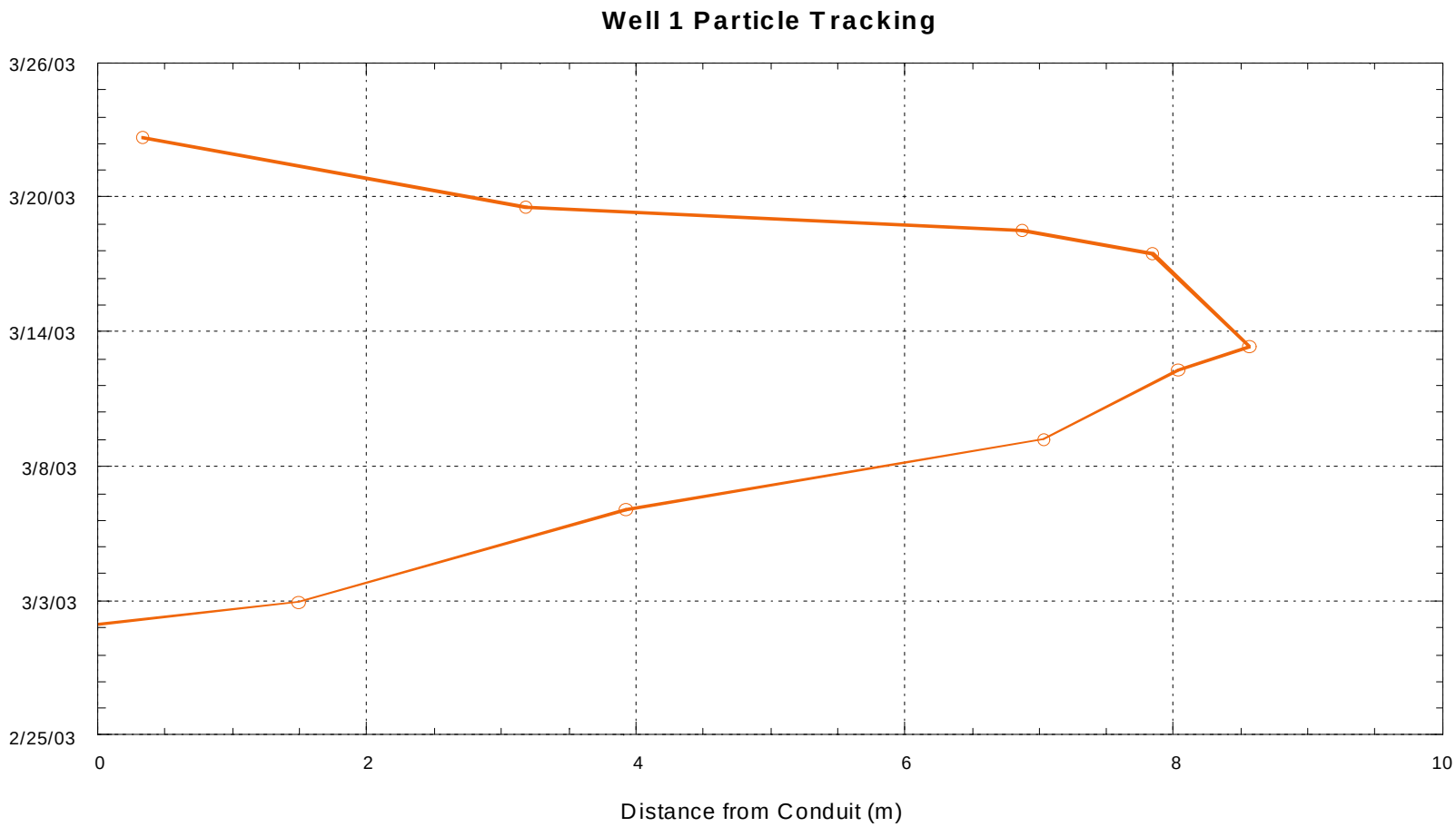


Figure 4-9. Particle tracking between the conduit and Well 4 of a water particle leaving the conduit on 3/2/03.

simulation and could act to increase or retard the movement of water into or out of the conduit.

Variation from the assumed effective porosity value of 0.2 used in the particle tracking calculation will also affect water movement from the conduit to the matrix. Laboratory tests have yielded values of porosity ranging from 0.1 to 0.45 in the Floridan Aquifer (Palmer, 2002). Substituting these values into the calculation for average linear velocity will result in a minimum and maximum distance conduit water could move into the matrix. Sensitivity tests show that if effective porosity is reduced from 0.2 to 0.1 the distance water moves into the matrix will be reduced by half. Likewise, if effective porosity is raised to 0.45, water movement distance is approximately doubled.

Movement of surface water from conduits to wells located in the matrix has been documented in the Upper Floridan Aquifer. Katz et al. (1998) studied the Little River in Suwannee County Florida after its disappearance into a series of sinkholes along the Cody Scarp. Monitoring wells, positioned near karst solution features located using ground-penetrating radar, were used to document the response in the Floridan aquifer after a recharge pulse from the sinking stream. Changes in water chemistry after the recharge pulse were used to determine the fraction of surface water found in wells near the conduit. Katz et al. (1998) determined the proportion of surface water found in the wells after the recharge pulse was between 0.13 and 0.84, using the natural tracers $^{18}O$, deuterium, tannic acid, silica, tritium, $^{222}Rn$, and $^{87}Sr/^{86}Sr$. The close proximity of his wells to conduits or enlarged fractures is the most likely reason he found movement of conduit water into the wells.

During high river stages, Dean (1999) found that chlorine ($Cl^-$), sodium, and sulfur concentrations decreased at the Rise Well, located approximated 1200 m west of the Rise, and closely resembled $Cl^-$ concentrations at the Sink during low river stages. It seems contrary to the particle tracking simulations that water from the main conduit reached the Rise Well. However, new feeder conduits coming into the main conduit from the west are being explored by cave divers (Alan Heck, personal comm.). The possibility exists that the Rise Well could be located closer to a solution feature than wells in this study allowing it to receive surface water rapidly during storm events.

Recent chemical data from this study area suggest that conduit water may reach Wells 2 and 7 (Sprouse, personal communication, 2003). However, hydraulic conductivity could not be estimated for these two wells due to lack of water level data from these wells during the event.

The effects of dispersion and diffusion were ignored during the particle tracking simulations presented in this study. Solutes can move through porous media by diffusion even if there is little or no groundwater gradient (Fetter, 2001). Variations in linear ground water velocity caused by heterogeneities in the aquifer will cause larger effects of hydrodynamic dispersion (Fetter, 2001). Both of these processes can increase the probability that water leaving the conduit migrates further into the matrix than predicted by particle tracking simulations.

# CHAPTER 5
# SUMMARY

Karst aquifers are a significant source of drinking water for millions of people, but are especially vulnerable to contamination by surface water. Understanding the relationship between the exchange of matrix and conduit water will help in determining the best way to protect this valuable resource. Surface water entering the matrix will also have significant effects on the rate of karstification in the Floridan Aquifer. Data from this project describe the relationship between hydraulic conductivity, head gradient, river stage, and the movement of water between matrix and conduits.

Hydrologic properties of the conduit system such as the Reynolds number, the friction factor, and absolute roughness were determined using conduit diameter, head loss, conduit length and average flow velocity. Analytical methods used to describe the physical properties of conduits will help future simulations more accurately represent the conduit system.

Analyses of head gradients between wells and the conduit reveal that the slope of the gradient coincides with the change in discharge between the Sink and Rise, indicating that water from the conduit is moving between the conduit and matrix. Transmissivity quantified using the passive monitoring methods of Ferris (1963) and Pinder et al. (1969), was calculated between 140 and 550,000 $m^2/d$. Because it does not rely on a sinusoidal groundwater fluctuation curve, the higher values calculated using the Pinder et al (1969) method were determined to be the more reliable estimates of transmissivity in the study area.

Hydraulic conductivities calculated using an aquifer thickness of 275 m, were between 0.5 and 2000 m/d. Values of transmissivity (T) and hydraulic conductivity (K)

calculated for portions of the aquifer within 400 meters of the conduit are most likely artificially low due to the partial penetration of the conduit. Calculations of hydraulic conductivity increased with increasing scale, which is indicative of the highly heterogeneous nature of the Upper Floridan Aquifer.

Particle tracking simulations were conducted to determine how far water leaving the conduit could migrate into the matrix and if it returned to the conduit or entered regional groundwater flow. The simulations were conducted for the March 2003 flood event using data from Wells 1 and 4 and hydraulic conductivities calculated using the Pinder et al. (1969) method. Conduit water left and then returned to the conduit in approximately 20 days and migrated between 0.45 and 8.5 meters into the matrix. These simulations suggest that conduit water is temporarily stored in the matrix and does not enter regional groundwater flow. Preferential flow paths within the matrix as well as the effects of diffusion and dispersion could allow conduit water to migrate further into the matrix than particle tracking simulations suggest, and illustrate the need for further investigation.

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## BIOGRAPHICAL SKETCH

Jennifer M. Martin was born in Bowling Green, KY. Family members include parents Dr. J. Glenn and Edith Lohr and sisters Susan and Mary Ellen Lohr. Jennifer received her B.S. degree in geology with a minor in environmental studies from Western Kentucky University in December 1997. She married Craig D. Martin in 2001, and later that year began pursuing her master's degree in hydrogeology at the University of Florida. She currently resides in Queens, NY, with her husband.